THE STRATIGRAPHY AND STRUCTURE OF THE SOUTHERN PART OF THE HARE BAY ALLOCHTHON, N.W. NEWFOUNDLAND

VOLUME 1

W. R. SMYTH

THE STRATIGRAPHY AND STRUCTURE OF THE SOUTHERN PART OF THE HARE BAY ALLOCHTHON, N.W. NEWFOUNDLAND

Volume 1

by

© W. R. Smyth

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

MEMORIAL UNIVERSITY OF NEWFOUNDLAND

1973

TABLE OF CONTENTS

Page

CHAPTER III

Page

Page

TABLES

VOLUME 2

NUMBER

ABSTRACT

The allochthonous sequence south of Hare Bay comprises four distinct tectonic slices which from the structurally lowest to the highest are: the Maiden Point Slice, the Croque Head Slice, the St. Julien Island Slice and the White Hills Slice. The allochthonous sequence was emplaced westwards in Middle Ordovician time over an autochthonous, mainly carbonate, sequence. The top of the autochthon consists of a northeasterly derived flysch that contains detritus derived from the allochthon. This flysch records a decrease in grain size, bed thickness and sand content, from north to south across the area.

The study has shown that the slices of the allochthon display different stratigraphy, different internal deformation and different metamorphic history. The slice contacts are in most places marked by a few tens of meters of black shaley mélange.

The Maiden Point Slice consists mainly of coarse quartzo-feldspathic greywackes and minor mafic volcanic rocks. The Croque Head Slice consists of finer grained greywackes than those in the Maiden Point Slice. The St. Julien Island Slice consists of sandy limestone in fault contact with a polymictic conglomerate. The conglomerate contains volcano-plutonic detritus which was probably derived from an island arc to the east. The White Hills Slice contains a partial ophiolite, structurally underlain by metavolcanic rocks and minor amounts of metasedimentary rocks that record pre-emplacement polyphase deformation and metamorphism. These metamorphic rocks consist of pyroxene-bearing amphibolites, amphibolites, and greenschists, that display decreasing metamorphic grade and intensity of deformation downward and away from the overlying

ultramafic rocks. They are interpreted as a dynamothermal aureole related to obduction and earliest transport of oceanic crust and mantle over previously undeformed and unmetamorphosed supracrustal rocks.

Emplacement of the allochthon in Middle Ordovician time produced a slaty cleavage, rare recumbent folds and low greenschist metamorphism in the lower tectonic slices and in the mélange zones. The autochthonous rocks are in most places unaffected by the emplacement deformation (Taconic orogeny), except at Canada Bay where mélange is absent and an imbricate structure and associated northwest-facing recumbent folds are developed.

After Middle Ordovician emplacement the whole area was affected by the Acadian orogeny. The pre- and syn-emplacement structures are refolded about open, upright to moderately inclined folds and an associated crenulation cleavage is widely developed.

The evolution of the allochthon and autochthon is described in terms of a plate-tectonic model related to the birth and destruction of a continental margin and a marginal ocean basin from late Precambrian to upper Paleozoic times. Comparisons with the other "Taconic" allochthons found along the western margin of the northern Appalachians suggests that similar tectonic processes controlled their evolution.

CHAPTER I

INTRODUCTION

Location, Access and Means of Travel

The Hare Bay Allochthon is located at the tip of the Great Northern Peninsula, the northeastern extremity of the Appalachians in Newfoundland. The Allochthon extends for approximately 120 km along the western shore of White Bay and up to 35 km inland. Hare Bay exposes an excellent cross section across its central part and forms the northern boundary of the map area. Canada Bay affords another well exposed cross section across its southern part. The southern and western extremities of the allochthon south of Hare Bay approximately define the limits of the map area.

Main Brook in Hare Bay and Englee in Canada Bay are served by unpaved roads and a road to Conche was opened in 1970. Access to the small coastal settlements of Canada Harbour (resettled 1971), Croque, Grandois, St. Juliens and Fishot Island is by sea and the area is served by C.N. coastal boat service approximately once a week.

The coastal survey was by means of an 18 ft. open boat, "Moses", the flagship of the Geology Department. A small rubber boat was used on Whites Arm Pond and a 12 ft. aluminum boat with a 9 hp motor on Coles Pond. Base camps were made at Main Brook (1969), Englee and Conche (1970 and 1971) and abandoned houses rented in various settlements along the coast served as temporary camps.

Physical Features

The central core of the Great Northern Peninsula is underlain by gneisses that form the Long Range Mountains (Fig. 1) that rise up to 760 m. Flanking the mountains to the north and west are lowlands (0-120 m) underlain chiefly by carbonate cover rocks. The northeastern part of the peninsula is underlain by allochthonous rocks that form a rolling upland terrane rising up to 260 m in the map area. A prominent north-south trending scarp marks the western extremity of the allochthon and a number of large ponds are located at its western foot. The allochthonous terrane is underlain chiefly by sandstone and consists of rounded, uncovered hills with intervening depressions covered with forest, marsh or ponds. A prominent ridge that runs from Whites Arm to Croque Harbour is the topographic expression of resistant basalt and similar ridges occur north of Coles Pond. The long, linear depression within the uplands in the northern part of the area is bottomed by carbonate and represents a window through the allochthon hereby named Whites Arm Window. The strong northeast grain of the uplands reflects the regional structure and erosion along fault zones.

Drainage in the area is poor and major rivers are only established in the lowlands. Streams flowing off the uplands commonly disappear and flow underground through the carbonates of the lowlands.

Glacial till is patchy and thin. Glacial erratics are common and together with the glacial striations indicate an easterly movement of glacial ice. (Fig. 3).

Exposure along the coast is excellent and Hare Bay, Whites Arm, Croque Harbour and Canada Bay afford valuable cross sections.

The shore line is gentle in Hare Bay but becomes cliffed south of St. Juliens. However, in most places the cliffs are accessible from a small boat except south of Canada Head where landings can rarely be made. Inland, exposure is moderate in the uplands but generally poor in the lowlands.

Geological Setting

The Hare Bay Allochthon lies upon the Western Platform (Kay, 1967), one of three distinct geological provinces of the Newfoundland Appalachians defined by contrasting Precambrian and early Paleozoic depositional and structural history (Williams, 1964).

The Western Platform is bounded to the east by the Central Mobile Belt (Williams, 1964) that consists of two northeast trending marginal orthotectonic belts with a wide paratectonic central portion that consists in part of a typical ophiolite succession (Snooks Arm Group, Lush's Bight Group). The western orthotectonic belt consists of an older Grenvillian (?) gneissic basement (M. de Wit, unpublished), overlain by a lower psammite division, followed by a mixed greywacke and mafic volcanic sequence, in turn overlain by mixed mafic and silic volcanic rocks and cut pre-kinematically by acidic intrusions (Kennedy, 1973a). The entire succession is collectively referred to as the Fleur de Lys Supergroup (Church, 1969).

The geology of the Western Platform is typical of that all along the western margin of the Appalachians and consists of three major tectonic elements outlined in Fig. 1, namely:

(1) a Precambrian crystalline basement deformed during the Grenvillian orogeny;

(2) an autochthonous Cambro-Ordovician carbonate sequence that unconformably overlies the Precambrian rocks and consists of a thin basal clastic-volcanic unit, overlain by a thick dolostone-limestone unit and capped by an easterly-derived flysch unit;

(3) a series of allochthonous thrust slices emplaced over the autochthonous sequence in middle Ordovician times. The transported rocks constitute two disconnected masses, the Humber Arm Allochthon in the south at Bay of Islands and the Hare Bay Allochthon in the north. The lower structural slices of the allochthons consist of Cambrian (?) to lower Ordovician mainly clastic sedimentary rocks, overlain by higher structural slices that consist of igneous and metamorphic rocks. The highest slice contains an ophiolite sequence with an attached aureole of metavolcanic rocks at its base. The slices are generally separated from each other and from the autochthon by black shaley mélange zones.

History of Previous Work and Development of Ideas

Prior to 1937 little was known of the geology of the Great Northern Peninsula. Murray made a cursory coastal examination of White Bay in 1864 and recognized a succession of Precambrian gneisses followed by Cambrian sediments and by limestones and sandstones of presumed Silurian age. Forty years later Howley (1902) spent a week in the Canada

Bay area and generally concurred with Murray's findings. Schuchert and Dunbar (1934), in their memoir on western Newfoundland, correlated the carbonates and the overlying sandstones and shales of the Canada Bay area with the Cambro-Ordovician carbonates and the overlying clastic Humber Arm series of western Newfoundland. They thought that all of the sediments of western Newfoundland comprised a single sequence of Cambrian to upper middle Ordovician age and this idea stood until 1963.

Cooper (1937) mapped the Hare Bay area on a scale of 1" to one mile and published the first comprehensive geological report on part of the thesis area. Cooper recognized two contrasting sequences and described a lower division consisting of limestone and an upper division characterized by sandstone and lava flows. Cooper divided the rocks into formations and his terminology has been adopted by later workers. Although Cooper failed to recognize the allochthonous nature of the rocks in the area he did realize the importance of thrust faults. He placed the White Hills Peridotite Sheet and the underlying amphibolite and greenschists in a separate thrust sheet or "decke" and estimated a displacement of eleven miles for this sheet on the Hare Bay thrust. He attributed the amphibolites and greenschists to contact metamorphism by the peridotites and suggested that another ultramafic mass lay seaward east of Fishot Island to account for similar metavolcanic rocks exposed there.

In 1939 Betz mapped the Canada Bay area and recognized Precambrian gneisses unconformably overlain by a Cambro-Ordovician sequence which he subdivided into 9 formations. He correlated the upper four formations with Cooper's Ordovician sequence in Hare Bay. Betz recognized a number of thrust faults and recumbent folds in the area but placed the entire sequence in a normal stratigraphic succession.

The first revolution in thought on the geology of western Newfoundland occurred in 1963 when Rodgers and Neale, following suggestions by Johnson (1941) and Kay (1945), rejected the theory of a single conformable sequence for the rocks of the Humber Arm and Hare Bay areas. They recognized that fossils collected in the Humber Arm Series showed the upper clastic sequence to be in part contemporaneous with the lower carbonate sequence. These anomalous relationships were previously explained by the clastic sequence being local facies variations in the carbonate sequence. Rodgers and Neale, realizing the similarities of the geology and tectonic setting of west Newfoundland with the Taconic Klippe of New York State, proposed that the upper clastic sequence constituted a separate but contemporaneous sequence which was deposited to the east of its present location. In Middle Ordovician times the clastic sequence and associated ultramafic rocks were emplaced as klippen on top of the carbonate sequence, presumably by gravity sliding.

Tuke (1968) mapped the area north of Hare Bay to investigate this possibility. He confirmed Rodgers' and Neale's prediction of two superimposed sequences of different lithologies but of similar age. Tuke delineated three separate slices and showed that the White Hills Peridotite Sheet occupied the highest slice.

The second revolution in the interpretation of the geology occurred with the advent of plate tectonics when Stevens (1970) and Church and Stevens (1970, 1971) interpreted the Bay of Islands Complex and the White Hills Peridotite Sheet as oceanic crust and mantle. Stevens (1970) viewed the sediments in the allochthon and in the upper part of the autochthon as flysch deposits related to ocean birth and destruction. Stevens (1968) previously had recognized that the metamorphic rocks at the base of the ophiolite slice had suffered polyphase deformation prior to emplacement. He interpreted the metamorphic rocks as previously deformed low grade schists that were "overprinted by thermal metamorphism bordering (the) ultramafics" (1968, p. 8). This view was later subscribed to by Church and Stevens (1971) and Church (1972), who considered the aureole to include at most only the amphibolites.

Smyth (1971),in a preliminary report on the present work, outlined the various tectonic slices of the allochthon from Hare Bay to Croque and recognized the importance of pre- and post-emplacement deformations in that area. He concurred with Stevens (1968) that the greenschists and amphibolites of the ophiolite slice had suffered polyphase deformation prior to emplacement but showed that the growth of metamorphic minerals accompanied the deformations rather than overgrowing previously deformed rocks. He concluded that if the heat was supplied by the ultramafic plutons then the ultramafic rocks must have been emplaced before or during the first and second deformation.

The Hare Bay Allochthon has featured in most of the recent plate tectonic models for the development of the Newfoundland Appalachians, most notably by Bird and Dewey (1970) and Dewey and Bird (1971). However, their interpretations of this area are based on previously published reports and no new rock relationships were added, although they put the area in sharp perspective.

Recent work north of Hare Bay by Williams, Smyth and Stevens (1973) outlined four major tectonic slices in the allochthon and their order of structural stacking has been determined for the first time. A map of the entire Hare Bay Allochthon by Williams and Smyth, incorporating their work north of Hare Bay with the results of this study is in preparation.

Purpose and Techniques

The present study was undertaken to determine the stratigraphy and structure of the little known area south of Hare Bay and to map for the first time the intervening 40 km between Cooper's and Betz's map areas. Of special importance was the determination of the number of tectonic slices that comprise the allochthon in that area and to determine their order of structural stacking.

Three field seasons were spent mapping the area on a scale of 1:50,000 and locally on a scale of 1:12,500 approximately, using air photographs supplied by the Department of Energy, Mines and Resources, Ottawa. The principles of small scale structural mapping as expounded by Wilson (1961) and the concept of facing directions of folds (Shackleton, 1958) were used throughout. Special attention was paid to the nature of the boundaries between the various allochthonous slices and the mélange zones. Using the principles of polyphase deformation, the presence of three major phases of deformation have been recognized. By determining whether the structures produced are truncated by or cut across the mélange zones it has been possible to date the deformations as having occurred pre-, syn- or post- the emplacement of the allochthon.

The field work was supplemented by a detailed petrologic study of thin sections. Growth phases of metamorphic minerals in relation to the phases of deformation have been determined using the textural methods of Zwart (1960) and Spry (1968). Mineral identification was by normal optical methods using the data of Kerr (1959) and of Deer, Howie and Zussman (1966). Plagioclase composition was determined by the Michel--Lévy method using a four axes universal stage.

Reconnaissance work north of Hare Bay and correlations with the work of Williams, Smyth and Stevens (1973) in that area allows discussion of the Hare Bay Allochthon in its entirety.

Acknowledgements

The work was supervised by Dr. M.J. Kennedy to whom special thanks are due for many suggestions and helpful criticisms. Dr. H. Williams is thanked for his interest in the project, for field visits and for reading the manuscript. I am grateful to Mr. R.K. Stevens for proposing the problem and for many useful discussions on the geology of western Newfoundland. My colleagues at Memorial University are thanked for valuable discussions and help, especially B. Marten, Dr. D.F. Strong, Dr. E.R.W. Neale, Dr. L. Fahraeus, S.P. Colman-Sadd, and P. Brown.

British Newfoundland Exploration Limited provided logistic support in the field and also supported D. Martin, B. White and W. Hoddinott who acted as excellent field assistants. The technical staff of the Geology Department, Memorial University, are all thanked for assisting in the production of this work.

Tenure of a Memorial University Fellowship and a N.R.C. Post-Graduate Scholarship is gratefully acknowledged.

Lastly, I wish to thank Miss Ruby Hudson for her encouragement throughout all phases of this work and for typing.

CHAPTER II

THE AUTOCHTHON

Introduction

The autochthonous sequence consists of four units namely, basement gneisses of Precambrian age; a basal clastic-volcanic unit of Eocambrian age; a middle carbonate unit of early Cambrian to early Ordovician age; and an upper flysch unit of Middle Ordovician age. The formations which comprise these units are listed below in Table 1.

UNIT	FORMATION
UPPER FLYSCH UNIT	Goose Tickle Formation Sugarloaf Schists
MIDDLE CARBONATE UNIT	Table Head Formation St. George Formation Treytown Pond Formation * Cloud Rapids Formation * Forteau Formation Devils Cove Formation
BASAL CLASTIC-VOLCANIC UNIT	Bradore Formation Lighthouse Cove Formation
BASEMENT GNEISSES	Long Range Complex Acid and Basic Gneisses

TABLE 1: Table of Formations in the Autochthonous Sequence (* Denotes formations that do not outcrop in the map area).

Basement gneisses outcrop in the south of the area where they form the northernmost exposures of a large inlier of Grenville gneisses named the Long Range Complex (Clifford and Baird, 1962). Just to the north of this inlier, infaulted into cover rocks, is a small block of lithologically distinct, but probably related gneisses, referred to as the Acid and Basic Gneisses.

The basal clastic-volcanic unit unconformably overlies the Long Range Complex around its northern periphery. It consists of conglomerate, basaltic lava, and sandstone. A small area of the clastic-volcanic unit occurs east of its main outcrop belt faulted into the upper flysch unit. The middle carbonate unit consists of a clear succession of limestone and shale, dolostone and rubbly-weathered limestone from lower Cambrian to lower Ordovician age.

The upper flysch unit conformably overlies the carbonate unit and marks an abrupt cessation of carbonate deposition. The flysch was easterly derived and contains detritus that can be matched with the allochthonous rocks that now structurally overlie it. The flysch unit outcrops in a north-south trending belt that parallels the western margin of the allochthon. It is also exposed in Whites Arm Window, an area of autochthonous rocks exposed through the allochthon cover following Acadian (Devonian) folding. In the extreme south of the area the flysch appears to grade into lithologically distinct semi-pelitic schists, referred to as the Sugarloaf Schists. The Sugarloaf Schists may be an equivalent of the upper flysch unit that was intensely deformed by the emplacement of the allochthon.

The allochthon was studied primarily from a structural viewpoint to determine the deformation effects of allochthon emplacement on it and to understand the effects and extent of the post-emplacement Acadian orogeny. However, a number of important stratigraphic relationships and problems were encountered, namely:

1. The occurrence of a faulted block of basement gneisses of uncertain affinities within autochthonous cover rocks.
2. The occurrence of rock stratigraphic units within the faulted block of the basal clastic-volcanic unit not present in the main outcrop belt of this unit only 2.4 km to the west.
3. A decrease in grain size and sand content of the upper flysch unit from north to south.
4. The discovery of autochthonous rocks in Whites Arm Window, clear evidence of the importance of post-emplacement folding.
5. The enigmatic occurrence of semi-pelitic schists below allochthonous rocks in the south of the area.

BASEMENT GNEISSES

The gneisses of the Long Range Complex (Clifford and Baird, 1962) define the southern limit of the map area where they are unconformably overlain by the Lighthouse Cove and Bradore Formations.

Sixty meters north of the Bradore unconformity with the Long Range Complex at Seans Cove in White Bay, and north of a major normal fault, is a small area of lithologically distinct Acid and Basic Gneisses. The top of the Acid and Basic Gneisses is a tectonic contact with cover rocks, named the Sugarloaf Schists. The stratigraphic relations of the Sugarloaf Schists to the overlying allochthonous rocks is uncertain and hence the relationships of the Acid and Basic Gneisses to the allochthon and also to the fault separated Long Range Complex are problematical. However, the Acid and Basic Gneisses are considered to be of Grenvillian age and to be related to the Long Range Complex, as:

1. They possess a gneissic foliation that predates the first fabric recognized in the cover rocks.
2. They are intruded by a post-tectonic granite that is similar to a late granite in the Long Range Complex at Seans Cove.
3. They are spatially related to the Long Range Complex.

LONG RANGE COMPLEX

The Long Range Complex consists of coarse crystalline acid gneiss and minor basic gneiss. Acid gneiss is excellently exposed at Seans Cove in White Bay and around the north shore of Otter Cove (Fig. 3). The rock is pink, massive to well foliated, quartz-plagioclase-K-feldspar-biotite gneiss. At Seans Cove the gneisses are intruded post-tectonically by a pink microgranite (Plate 1, fig. a). Basic gneiss is rare and occurs as foliated pre-kinematic sills or dykes in the acid gneiss and also as larger meta-gabbro bodies. Meta-gabbro is exposed in Canada Bay northwest of Gouffre Island where it is unconformably overlain by Bradore sandstones. The meta-gabbro is retrogressed to a chlorite-sericite-hematite rock.

Along the west shore of Canada Bay the gneisses are cut post-tectonically by mafic dykes.

ACID AND BASIC GNEISSES

Acid and basic gneisses outcrop for approximately 610 m along the shore of White Bay, northeast of a major north-south trending fault, the Wild Cove Fault of Betz (1937) (See Fig. 3, section E-F). The gneisses are poorly exposed inland and have been traced only for 305 m northwards. Foliation in the gneisses dips moderately to the northeast and their upper contact is a late minor fault which separates them from a major tectonic slide (Fleuty, 1964 b), named the Sugarloaf Slide, which consists in part of mylonitized gneiss and granite. The gneisses are cut by post-tectonic granite dykes that occur in great profusion eastwards towards the structural top of the gneisses. Rare post-tectonic mafic dykes cut the gneisses and the granite dykes.

Lithology

The gneisses consist of light grey acid gneiss with minor interlayered hornblende gneiss. The layers of hornblende gneiss are from 0.6 m - 3.6 m thick and are the thickest and most abundant to the west near the structural base of the gneisses.

The gneisses possess a moderate to strong banding. The banding is best expressed in the acid gneiss by alternating quartzo-feldspathic and mafic bands from 1-4 cm thick and the mafic bands possess a concordant biotite foliation. The hornblende gneiss is only locally banded and consists of weakly defined hornblende and plagioclase bands from 1-2 cm thick. In most places the hornblende gneiss possesses a strong foliation that is concordant to the banding. Rare blocks of acid gneiss occur as inclusions in the hornblende gneiss and the foliation is concordant with both.

Fabric

The gneissose foliation and the post-tectonic granite dykes are cut by a gently dipping schistosity that intensifies eastwards towards the structural top of the gneisses.

Petrography

The acid gneiss is composed of plagioclase (73%), quartz (20%), biotite (7%), with minor hornblende, epidote and sphene and accessory apatite, zircon and opaques.

Plagioclase (An_{32}) is the dominant mineral and occurs as idiomorphic to xenomorphic crystals that are partially to completely sericitized. Small rounded quartz inclusions are common but form no definite internal fabric. Quartz occurs as xenomorphic crystals and elongate aggregates that parallel the foliation. Brown biotite occurs as laths up to 2 mm long that lie in the plane of the foliation. Hornblende, associated with biotite, is present in small amounts in subidiomorphic crystals. Epidote occurs as small xenomorphic crystal aggregates and also as an alteration product of the plagioclase.

The amphibolite gneiss is composed of hornblende (60%,) plagioclase (20-40%), with minor quartz, epidote, sphene, apatite and opaques. Hornblende forms subidiomorphic crystals 1-2 mm across that are aligned parallel to the foliation. Plagioclase forms twinned and untwinned xenomorphic crystals 1-2 mm across that are generally highly altered. Quartz is locally present in major amounts and occurs as small irregular crystals.

Where the gneiss has been affected by the later deformation it has suffered extensive retrogressive metamorphism. The feldspar is commonly completely sericitized and saussuritized and locally the sericite is aligned and defines the superimposed fabric (S_1). Biotite flakes are kinked and broken and partially chloritized and hornblende is altered to chlorite. In highly deformed examples the acid gneiss consists of isolated quartz and chloritized biotite crystals set in a sericite groundmass and the rock resembles a deformed sediment. However, the large, kinked biotite flakes are distinctive.

BASAL CLASTIC-VOLCANIC UNIT

The basal clastic-volcanic unit consists of a thin, locally developed conglomerate, overlain by basalts of the Lighthouse Cove Formation, in turn followed by arkoses of the Bradore Formation. The basal conglomerate has been referred to as the Bateau Formation on Belle Isle (Williams and Stevens, 1969), but in the map area it is thin and of limited extent and hence is not given formational status but is included in the Lighthouse Cove Formation.

LIGHTHOUSE COVE FORMATION

The basalts and the underlying and intercalated coarse clastics which unconformably overlie the Long-Range Complex are assigned to the Lighthouse Cove Formation. The formation outcrops at the south shore of Otter Cove and in a fault bounded block west of Burnt Point. The name Lighthouse Cove Formation was first introduced by Williams and Stevens (1969) for volcanic rocks that overlie basement gneiss on Belle Isle and was later extended by Strong and Williams (1972) to include volcanic rocks that occupy similar stratigraphic levels at southeastern Labrador and Canada Bay. The main outcrop of the Lighthouse Cove Formation in Canada Bay at Cloud Mountain was not examined.

At Otter Cove, 9 m of lava with top unexposed outcrops in two small separate areas. The lava either rests directly on the Long Range Complex or else is underlain by up to 3 m of coarse sandstone and pebble-conglomerate. The southernmost exposure consists of two flows separated by a 1 m thick bed of pebble conglomerate containing rusty weathered quartzite clasts. Locally, where lavas directly overlie clastics, the basal 1 - 1.5 m of the flow contains vesicular pillows (Plate 1, fig. b; Plate 2, fig. a). Towards the top hexagonal cooling joints are developed (Plate 2, fig. b) that are well exposed on wave washed surfaces.

West of Burnt Point a high angle thrust block of the basal clastic-volcanic unit is exposed. It consists of a porphyritic microgranite intrusion at the base, structurally overlain by a plutonic boulder conglomerate (6 m thick), and followed in turn by pillow lava and Bradore sandstone. The contacts where exposed are thrusts. The conglomerate has moderate to well rounded clasts from cobble to boulder size (Plate 3, fig. a) that consist of pink granite gneiss, amphibolite, and quartzite. The clasts are grain supported and the interstices are infilled with a fine schistose mudstone.

The volcanic rocks of the Lighthouse Cove Formation are light to dark green and generally altered. In section the lava is either microporphyritic with amygdules or porphyritic without amygdules. Phenocrysts of plagioclase (An_{48}) and rare quartz are present. The fine grained groundmass of augite, plagioclase and iron oxide is generally altered to fibrous actinolite, chlorite, epidote and hydrated iron oxide. Interstitial patches of chloritized volcanic glass are common and small patches and veins of secondary albite and quartz occur locally. The amygdules are infilled with quartz, chlorite, epidote and natrolite and some have a thin rim of iron oxide.

The lavas are of tholeiitic composition and were once thought to be chemically similar to the mafic dykes which cut the basement gneisses (Strong and Williams,1972). However, further analyses now contradict this correlation (D.F. Strong, pers. comm.).

BRADORE FORMATION (CLOUD MOUNTAIN FORMATION OF BETZ)

Betz named the sandstone unit that unconformably overlies the Long Range Complex and presumably overlies the Lighthouse Cove Formation, the Cloud Mountain Formation, and he correlated it lithologically with the Bradore Formation (Schuchert and Dunbar, 1934) of southeast Labrador. The name Bradore Formation is now widely used and is favoured here.

The sandstones crop out along the north shore of Otter Cove (Betz's type section) and southwards along the west shore of Wild Cove (Fig. 3). From Wild Cove the beds continue inland and reappear on the shore of White Bay at Seans Cove. A small area of Bradore sandstone occurs in a faulted block west of Burnt Point where it structurally overlies the Lighthouse Cove Formation. Unconformable relations with the Long Range Complex are exposed at Otter Cove, Fly Point, northwest of Wild Cove and Seans Cove (Fig. 3). The sandstones are conformably overlain by limestones of the Devils Cove Formation at Otter Cove and at Wild Cove.

Betz (1939) measured 585 ft. (178 m) of strata at the type section, however, this study shows the formation to be 64 m thick. Two thrust faults near Dieppe Point repeat the top of the succession and may account for Betz's figure. West of Wild Cove 60 m of strata were estimated. This compares with a thickness of 220 to 230 ft. (67 to 70 m) at southwest Labrador and of 300 to 400 ft. (91 to 121 m) at Belle Isle (Williams and Stevens, 1969).

The Bradore Formation consists of a locally developed basal breccia followed by grey to white, dark to light purple, sub-arkoses, green sandstones and siltstones with local pebble conglomerate. Towards the top of the section, black, hematite-stained, sandstones with thin hematite pebble conglomerate predominate.

The basal breccia is best exposed at Seans Cove in White Bay where it is up to 0.5 m thick and consists of angular blocks of granite gneiss and undeformed microgranite set in a green sericite matrix (Plate 3, fig. b). West of Wild Cove the basal beds consist of 18 m of fine-grained, green, subarkose that does not appear on the coast and cannot be followed along strike. It appears that the basal beds represent infilling of a local depression.

The succeeding beds onlap onto unweathered basement gneisses and are well-exposed on the coast. They consist of a thin pebble conglomerate bed containing clasts of quartz, quartzite and gneiss, followed by 36 m of alternating beds of grey to green to purple, fine to coarse sandstones with rare pebble conglomerates (Plate 4, fig. a). The upper 12 m of the section consists of black, medium-grained, hematitic sandstones with minor interbedded grey sandstones. Rounded to subangular hematite grains form up to 25% of the detritus and the quartz grains commonly have a black iron coating. An 8 cm thick pebble conglomerate with hematite clasts up to 2 cm in diameter occurs at the base of the iron rich sandstone sequence (Plate 4, fig. b). The clasts are weakly magnetic suggesting the presence of some remnant magnetite.

The type section at Otter Cove consists of about 48 m of medium-grained purple cross-bedded sandstones overlain by 15 m of black hematitic sandstones. West of Fly Point, near the top of the exposed Bradore sequence, an 8 cm bed of purple crystalline limestone is interbedded with medium-grained purple to black sandstones. The limestone bed is overlain by a thin breccia horizon containing clasts up to 2 cm in diameter of limestone, sandstone, hematitic sandstone and granite gneiss set in a matrix of medium-grained sandstone.

The black hematitic sandstones are distinctive and are useful marker horizons. West of Burnt Point, volcanic rocks of the Lighthouse

Cove Formation are structurally overlain by 8 m of sandstones and arkoses and are followed in turn by a 1 m thick bed of hematitic sandstone and a 2.5 m thick bed of white crystalline limestone. The limestone is highly sheared and is the site of a major thrust fault. The stratigraphy within the fault block conforms to that of the Lighthouse Cove-Bradore Formation succession and the hematitic sandstone beds allow precise correlation.

Mega ripples with crests striking about 65^{o} occur near the base of the sequence and large scale tabular and trough cross-bedding and graded bedding is common throughout the formation (Plate 5, fig. a). South of Fly Point, the orientation of the cross-bedding indicates a paleocurrent flow from the southwest.

Petrographically the sandstones plot in the arkose, sub-arkose and orthoquartzite field as defined by PettiJohn (1957) (See Fig. 8). The individual grains vary in any one section from sub-angular to well-rounded with a moderate to high sphericity index. The feldspar consists of both potassic and plagioclase varieties and commonly a single section contains both fresh and highly altered feldspar grains (Plate 5, fig. b). This is diagnostic of tectonic arkose (Folk, 1968) and indicates a source of detritus from fresh bedrock and from older sediments. Detrital magnetite, now almost completely altered to hematite, is the only abundant heavy mineral with accessory amounts of zircon. The hematite grains are commonly well rounded and nearly spherical (Plate 6, fig. a).

The absence of volcanic detritus in the Bradore Formation, and the sharp contact between the sandstones and the lavas of the Lighthouse Cove Formation on Belle Isle (Williams and Stevens, 1969) suggests a conformable relationship between the two formations and hence the lavas cannot be considered as a source for the hematite detritus. Hematite ore in gneisses west of Cloud Mountains (Howley 1918, p. 513) might constitute the source of the iron detritus.

MIDDLE CARBONATE UNIT

The middle carbonate unit is well exposed across Canada Bay where it dips steeply eastwards away from the lower clastic-volcanic unit. The succession is repeated by numerous high-angle thrust faults and the complete unit is not exposed in the map area. Northwards the dip of the beds becomes gentle, thrust faults are rare or absent, and only the upper part of the unit, the Table Head Formation, is exposed in the western part of the map area. The Table Head Formation is also exposed in the central part of Whites Arm Window through the allochthon cover.

The complete middle carbonate unit consists of six formations as defined by Betz (1939), and a table of the formations showing Betz's terminology and the present usage is given in Table 2.

Age	Betz (1939)	Present Usage	Lithology
Lower Middle Ordovician	Bide Arm Fm.	Table Head Fm.	Limestone, minor dolomite and shale.
Middle Cambrian to Early Ordovician	Chimney Arm Fm.	St. George Fm.	Pink dolomite and limestone.
Middle Cambrian	Treytown Pond Fm.	Not present in map area.	Grey limestone, black limestone with chert.
Early Middle Cambrian	Cloud Rapids Fm.	Not present in map area.	Blue-black limestone, shale, minor orthoquartzite.
Lower Cambrian	Forteau Fm.	Forteau Fm.	Grey shale, sandy limestone.
Lower Cambrian	Devils Cove Fm.	Devils Cove Fm.	Purple and white limestone with thin, purple shale.

TABLE 2. Relationship of Present Stratigraphic Terminology to that of Betz (1939).

DEVILS COVE FORMATION

Thin bedded purple and white limestones of the Devils Cove Formation (Betz, 1939) overlie the Bradore Formation with sharp contact along the west shore of Canada Bay. The contact appears to be conformable and marks a sudden change from sand to carbonate deposition.

The type section is at the northeast shore of Otter Cove where approximately 14 m of limestone, with top unexposed, intervenes between the Bradore Formation and the Forteau Formation. At the west shore of Wild Cove approximately 8 m of purple limestone and thin minor shale, with top unexposed, conformably overlies the Bradore Formation (Plate 6, fig. b). At the southeast side of Wild Cove similar limestone occurs in fault contact with the Goose Tickle Formation. There the limestones are sheared and recrystallized and form a mylonite up to 0.6 m from the fault (Plate 7, fig. a). A similar relationship is seen west of Burnt Point where a 2.5 m thick zone of sheared limestone overlies the Bradore Formation and is in fault contact with the Goose Tickle Formation.

The formation is fossiliferous, and poorly preserved gastropods were collected from pink limestones west of Dieppe Point. At Wild Cove the pink limestones contain micro-fragments of algae filaments and brachipod shells (Plate 7, fig. b). Betz collected a variety of gastropods and trilobites from localities north of the map area that indicate a lower Cambrian age.

FORTEAU FORMATION

The grey shales and sandy limestones that overlie the Devils Cove Formation along the northwest shore of Canada Bay are assigned to the Forteau Formation (Schuchert and Dunbar, 1934; Betz, 1939). In the map area the Forteau Formation outcrops at Fly Point and along the west shore of Chimney Bay. South of Canada Bay the formation is cut out by thrust faults.

The base of the formation was not observed. At Dieppe Point the lowermost beds follow the Devils Cove Formation across a small gap in exposure and consist of thin interbeds of grey limestone, mudstone and shale. These are followed by blue, sandy limestone and interbedded grey mudstone. Approximately 108 m of strata with top unexposed outcrops east of Dieppe Point and Betz estimated the formation to be 213 m thick at Castor Cove north of the map area.

The formation contains a variety of brachiopods, trilobites, gastropods and cephalopods that indicate a lower Cambrian age (Betz 1939).

ST. GEORGE FORMATION

Introduction

Limestones and shales that disconformably overlie the Forteau Formation to the northwest of Canada Bay were assigned to the Cloud Rapids Formation and the Treytown Pond Formation by Betz (1939). These two formations are absent from the map area due to thrust faulting.

The succeeding formation, the St. George Formation (Kindle and Whittington, 1965) consists of shallow water dolostones and minor limestones and was named the Chimney Arm Formation by Betz (1939). The contact with the Treytown Pond Formation is not exposed in the Canada Bay area (Betz, 1939). The St. George Formation underlies the western half of the Bide Arm Peninsula and strikes northwards to Hare Bay where equivalent rocks were named the Southern Arm Limestone by Cooper (1937).

Description

In the map area the lowermost beds of the St. George Formation outcrop along the east shore of Chimney Arm. At the east side of Wild Cove the formation is thrusted over the Goose Tickle Formation. The St. George Formation is conformably overlain by limestones of the Table Head Formation and the contact is exposed at Handy Harbour and east of Wild Cove. The contact is defined by the first appearance of limestone above the last chert horizon in the sequence. 460 m of strata with base unexposed outcrop across the southern end of Bide Head Peninsula and Betz estimated the total thickness of the formation to be 1800 ft. (548 m).

The lowermost beds are exposed along the east shore of Chimney Bay and consist of 49 m of thin bedded black hackly limestones. These are followed by about 430 m of alternating dolomites and sandy dolomites in well defined beds 0.5 - 1 m thick. The dolomites weather to white, buff or brown but are grey to pink on fresh surfaces. Thin chert beds and nodules occur towards the top of the sequence near Handy Harbour. The uppermost chert horizon is overlain by a few metres of dolomite followed by black and grey limestones with minor interbedded dolomite. The limestone beds mark a distinct change in lithology from the thick monotonous dolomites and are assigned to the overlying Table Head Formation.

Large scale cross bedding is rarely developed in the sandy dolomites. Mega-ripples, striking east-west, and mud cracks occur in the lower half of the sequence. Algal mounds measuring 4.5 cms long and 18 cms high occur in grey dolomite beds approximately 60 m above the black limestone member.

The formation is only locally fossiliferous. Betz collected gastropods, brachiopods and trilobites from localities at Chimney Arm and east of Wild Cove that indicate an early Ordovician Age.

TABLE HEAD FORMATION

The uppermost formation of the middle carbonate unit consists of rubbly-weathering limestone and shale and records a change in sedimentation from the underlying dolostones of the St. George Formation. The limestones were named the Bide Arm Formation (Betz, 1939) in the Canada Bay area and the Hare Island limestone (Cooper, 1937) in the Hare Bay area and were correlated by both workers with the Table Head Formation (Schuchert and Dunbar, 1934; Whittington & Kindle, 1963) which outcrops extensively in western Newfoundland. The name Table Head Formation is used here on account of its widespread application.

No complete section of the Table Head Formation is exposed in the Canada Bay area where the formation is cut by thrust faults. Northwards towards Big Springs Inlet in Hare Bay the formation is less disturbed but the base of the formation was not mapped. The upper part of the formation is also exposed in Whites Arm Window (Fig. 3). The thickness of the formation is estimated to be about 490 m at Canada Bay but estimation is complicated by incomplete sections and the development of recumbent folds there.

The Table Head Formation records a gradual change in lithology from base to top and three informal members are recognized. The basal member is about 30 m thick and consists of black to blue to white crystalline limestone with interbedded minor dolomite beds and is exposed at Bide Head and east of Wild Cove. The middle member is approximately 300 m thick and consists of a well bedded sequence of black to grey to white limestone and marble in beds from 0.5 to 1 m thick. At Canada Bay it includes rare dolomite beds. The upper member consists of approximately 180 m of thin bedded black hackly limestone with minor thin chert beds and black shale and is best exposed at Big Springs Inlet and in Whites Arm River. At Canada Bay the shale is more abundant and is interbedded with thin limy siltstone. Intraformational breccia with clasts from 2 cm to 15 cm across occurs in beds from 0.5 to 1 m thick near the top of the formation at the north shore of Coles Pond.

The intensity of deformation and of recrystallization of the Table Head Formation increases from north to south and in the Canada Bay area most of the middle unit consists of marbles. The upper unit is generally cleaved and bedding is unrecognizable except where interbeds of chert are present (Plate 8, fig. a).

The Table Head Formation is fossiliferous but preservation of the fossils is poor. Gastropods were collected from the middle unit 1.6 km north of Bide Head on the west shore of Bide Arm. The upper black limestone unit is abundantly fossiliferous and brachiopods, trilobites, gastropods, cephalopods and crinoid stems were collected from locations at Big and Little Springs Inlet, from the river gorge 0.8 km west of the head of Whites Arm, and at the southwest side of Coles Pond. Fahraeus (1970) recovered conodonts from the top of the black limestone member at the west side of Little Springs Inlet that indicate an Arenig-Llanvirn age.

UPPER FLYSCH UNIT

The upper flysch unit consists of northeasterly derived greywackes and shales. The greywackes record a general decrease in grain size, sand content and bed thickness from north to south across the area. The flysch marks the cessation of carbonate deposition that existed in the area from lower Cambrian to middle Ordovician times. The flysch was named the Goose Tickle slate at Hare Bay (Cooper, 1937) and the Englee Formation at Canada Bay (Betz, 1939). The name Goose Tickle Formation (Tuke, 1968) is used here for the entire area as the rocks are best exposed and contacts more easily defined in the Hare Bay area.

GOOSE TICKLE FORMATION

The Goose Tickle Formation forms a continuous narrow belt that parallels the western margin of the allochthon from Big Springs Inlet to Canada Bay and is also exposed below the allochthon in Whites Arm Window. South of Canada Bay the formation is repeated by thrust faults and outcrops in four north-south trending belts, and the most easterly one parallels and dips under the allochthon. South of Canada Harbour the greywackes and slates appear to grade into semi-pelitic and calc-silicate schists referred to as the Sugarloaf Schists (new name), which are well exposed around Sugarloaf Cove in White Bay. The Sugarloaf Schists may be a highly deformed and metamorphosed equivalent of the Goose Tickle Formation. However, because of their special structural setting and lithological differences, and as other correlations are possible, the schists are described separately at the end of this section.

The top of the Goose Tickle Formation is defined by a tectonic contact with the allochthon. In the north of the area the contact is marked by a few tens of meters of black shaley mélange but, at Canada Harbour, mélange is absent and the contact is a sharp thrust. At Big Springs Inlet the upper part of the formation is tectonically disrupted towards the contact with the mélange and the formation boundary is drawn at the first appearance of exotic material in the mélange. A complete section of the Goose Tickle Formation is exposed at Big Springs Inlet where it is 258 m thick. This compares with approximately 1000 ft. (305 m) at the type section at the head of Hare Bay (Stevens 1970).

Lithology

At Hare Bay, the Goose Tickle Formation consists of medium to fine-grained greywacke beds from 0.3 to 1 m thick with interbedded black shales up to 0.3 m thick. The sand content, grain size and bed thickness decrease southwards and in the Canada Bay area the formation consists dominantly of dark grey slates to brown siltstones with minor fine-grained greywacke beds towards the top of the section.

A conglomerate lens, 2 m long and 0.5 m thick, is interbedded with greywackes and slates 0.4 km west of the allochthon contact on the south side of Whites Arm. The conglomerate contains small oval shaped [illegible]les of sandstone and rare diabase clasts (Plate 8, fig. b). A bed of coarse sandstone outcrops near the top of the formation at the western shore of Whites Arm that contains clasts up to 1 dm across of sandstone, green, brown, and black shale, and rare diabase.

The greywackes at Big and Little Springs Inlet exhibit a variety of sedimentary structures, namely, graded bedding, cross bedding, ripple-drift lamination, convolute lamination, load casts and flute casts (see Plate 9, figs. a and b). The latter indicate a northeasterly transport of sediment (see Fig. 9). Typically the greywackes show a basal graded interval and an upper cross or parallel laminated division (Plate 10).

Fabric

In the north of the area the formation is openly folded and is cut by a single penetrative cleavage. Southwards the intensity and complexity of deformation increases and at Canada Bay the formation consists of slates and phyllites with two cleavages.

Petrography

The greywackes in the north of the area have the following modal composition: quartz (40-60%), feldspar (2-6%), 'rock fragments' (6-18%), and matrix (25-30%). At Canada Bay the greywackes consist of quartz (50%), matrix (40-60%), rare feldspar, 'rock fragments', and heavy minerals (less than 2%). The quartz is of vein and plutonic origin and the grains vary from subangular to subrounded. However, perfectly rounded and nearly spherical grains are sparse. Plagioclase (up to An_{30}) is the commonest feldspar although a few grains of microcline were identified. 'Rock fragments' include diabase, felsite, myrmekite, granophyre, chert, limestone, ooliths, sandstone, shale and chlorite (See Plates 11 and 12). Chlorite grains are present in most sections and locally contain inclusions of hematite and chromite (Plate 12, fig. b). The diabase clasts are cryptocrystalline and contain twinned and untwinned plagioclase (An_{30}), rare quartz, interstitial chlorite, and rare leucoxene. The diabase clasts are petrographically similar to the allochthonous Maiden Point volcanics (see Chapter 3). The felsite, myrmekite and granophyre detritus indicates a granite source. The coarse sandstone at the head of Whites Arm is petrographically similar to coarse conglomerate present in the formation north of Hare Bay (Tuke, 1968; Stevens, 1970), that contain detritus derived mainly from the allochthonous Northwest Arm Formation (Williams et al., 1973).

Heavy minerals include chromite, hematite, zircon, and rare tourmaline. The chromite is translucent and reddish and is distinctive. In the greywackes north of Hare Bay chromite occurs within grains of serpentine (R.K. Stevens, pers. comm.) indicating derivation from an ultramafic source (Stevens, 1970). No chromite was seen in the greywackes at Canada Bay.

Fossils and Age

Poorly preserved brachiopods were found in siltstone bands within the outlier that outcrops west of White Hump Pond and rare graptolite fragments were recovered from outcrops at the western shore of Coles Pond and at Little Springs Inlet in Hare Bay. However, the best preserved and most abundant graptolites in the formation occur north of the map area at Pistolet Bay (Tuke, 1968) and at the western shore of Hare Bay (Erdtmann, 1971). The assemblages collected from these two areas indicate a lower Middle Ordovician age (Llanvirn) for the formation.

SUGARLOAF SCHISTS

3.2 km (2 miles) southwards along strike from Canada Harbour the easternmost belt of the Goose Tickle Formation apparently grades into semi-pelitic schists, informally referred to as the Sugarloaf Schists. Exposure along this belt is poor and the contact was nowhere observed. The Sugarloaf Schists are well exposed on the coast around Sugarloaf Cove where they overlie, across a major tectonic contact, the Acid and Basic Gneisses. This contact is marked by approximately 53 m of mylonitized gneiss, granite and cover rocks and is referred to as the Sugarloaf Slide Zone. At the north side of Sugarloaf Cove the schists are overlain by greywackes of the allochthonous Maiden Point Formation. The contact is sharp and appears conformable although the greywackes are less deformed than the schists. To the west the schists are separated from the Bradore Formation by the Wild Cove Fault and to the northwest they apparently overlie the Table Head Formation although the contact is not exposed and may be tectonic.

The Sugarloaf Schists are especially enigmatic as they occur within an area of major thrusting and it is uncertain whether they are allochthonous and underlie the Maiden Point Formation or whether they are autochthonous and represent a highly deformed part of the upper flysch unit.

Most lines of reasoning support the autochthonous nature of the schists, namely:

1. They occupy a similar stratigraphic and structural position to the autochthonous Goose Tickle Formation. For example, (a) they are overlain along their eastern contact by the allochthonous Maiden Point Formation: and (b) they overlie the Table Head Formation.

2. Nowhere else in the Hare Bay Allochthon is the Maiden Point Formation stratigraphically underlain by semi-pelitic schists.

3. Nowhere does the Maiden Point Formation rest on any other part of the autochthon other than the Goose Tickle Formation.

However, arguing against these points is the lack of structural features at the Sugarloaf Schist-Maiden Point Formation contact that would suggest a structurally significant contact, such as the emplacement thrust of the allochthon. However, at the east side of Canada Harbour the basal thrust of the allochthon is also a sharp, apparently insignificant contact.

In this thesis the Sugarloaf Schists are considered to be part of the autochthonous sequence and to be deformed and metamorphosed equivalents of the Goose Tickle Formation. Therefore, the underlying Acid and Basic Gneisses must also be autochthonous and the slide contact between them must represent a major tectonic break, similar to the thrusted contact between the basal clastic-volcanic unit and the Goose Tickle Formation exposed west of Burnt Point.

Lithology

The Sugarloaf Schists are typical low grade regionally metamorphosed schists with two sets of cleavages, but have been affected by post-tectonic thermal metamorphism producing extensive hornfels. This late metamorphism forms the southern margin of a large area of thermal metamorphism that also affected the structurally overlying Maiden Point Formation as far north as Canada Head and is described in Chapter VII.

The pelites and semi-pelites are fine grained, thin bedded, dark purple to dark grey schists. White weathered calc-silicate bands, 2-3 cm thick, are interbanded with the schists in varying amounts throughout the formation (Plate 13, fig. a). The calc-silicate bands are conspicuous, and diagnostic of the schists in inland exposures. Marble beds up to 0.3 m thick (Plate 13, fig. b) are interbedded with semi-pelites near the contact with the Maiden Point Formation. The marble beds are severely deformed and are commonly broken up into boudins on the fold limbs. Graded, gritty, psammites with interbanded basic schists, occur below banded semi-pelitic and calc-silicate schists in a cliff face near the northwest corner of Sugarloaf Cove. The psammite-basic schist horizon is about 5 m thick and appears to occupy the nose of a southwest facing recumbent fold. The psammites occur in graded beds up to 0.3 m thick and contain pebbles of blue quartz.

Fabric

The dominant fabric element of the Sugarloaf Schists is a penetrative S_1 schistosity that is refolded by tight, inclined F_2 folds with an associated S_2 crenulation cleavage. Minor F_3 folds locally refold the early structures and appear to be of purely local significance.

Petrography

The semi-pelites and pelites are fine grained biotite-muscovite-quartz-oligoclase schists with minor amounts of epidote, zoisite, and accessory apatite, zircon, tourmaline and ore.

Oriented flakes of light brown, biotite and muscovite define the S_1 fabric. Oligoclase occurs as untwinned cloudy, xenomorphic crystals up to 0.8 mm across. Quartz occurs in single grains and in quartzite bands up to 3 mm thick. Chlorite after biotite, and muscovite laths are interleaved with quartz at the margins of the quartzite bands and here the quartz has a strong preferred shape orientation parallel to S_1. The interior of the bands is free of inclusions and here the quartz has recrystallized to a granoblastic polygonal network with grains from 0.5 - 0.8 mm across (see Plate 14, fig. a).

Post-tectonic porphyroblasts include biotite, epidote and cordierite and these overgrow and include the regional fabrics.

The calc-silicate bands are composed of quartz and carbonate with subsidiary oligoclase and epidote. Post-tectonic porphyroblastic growth of diopside and tremolite-actinolite is extensive. Carbonate forms from 0-30% of the calc-silicates and occurs as granular crystals 0.2-0.5 m across associated with bands of quartz. Diopside occurs as large xenomorphic to subidiomorphic crystals up to 8 mm across, but more commonly as crystals 2-3 mm across. Almost colourless tremolite-actinolite occurs at the margins of the diopside-rich bands as small laths up to 0.8 mm long.

The marble beds are coarse grained, granoblastic and weakly foliated. They are composed almost entirely of calcite with a few grains of quartz and apatite. The calcite crystals are from 0.5-6 mm across and exhibit a polygonal texture. Post-tectonic diopside is rare and occurs as subidiomorphic to xenomorphic crystals, up to 3 mm long scattered throughout the rock.

The basic schists are composed dominantly of biotite (30-50%) and actinolite (30%) with subsidiary oligoclase (10-15%), quartz (10-15%), epidote (3%) and accessory magnetite octahedra. Small brown biotite laths up to 0.5 mm long define the S_1 schistosity that is overgrown by large unoriented biotite and tremolite-actinolite crystals up to 1.5 mm long. The biotite is partially chloritized. Oligoclase occurs as untwinned, cloudy xenomorphic crystals up to 0.8 mm long. In some sections the feldspar is completely sericitized. Quartz forms up to 2% of the rocks and occurs as stringers of recrystallized quartz up to 0.5 mm long. Dark brown epidote occurs as small xenomorphic crystals in the groundmass and accessory minerals include magnetite, octahedra and disseminated chalcopyrite. The basic schists are interbedded with gritty psammite and appear to be of sedimentary origin.

Discussion

Although at first glance the Goose Tickle Formation appears lithologically distinct from the Sugarloaf Schists the differences can be largely attributed to deformation and to the hornfelsing of the schists. The lithologies of two rock units are compared below in Table 3.

GOOSE TICKLE FORMATION	SUGARLOAF SCHISTS
Greywackes	Psammite and semi-pelitic schists
Slates	Pelitic Schists
Limy Siltstones	Calc-silicate Schist
Limestone	Marble

TABLE 3: A Comparison of the Goose Tickle Formation and the Sugarloaf Schists.

This lithological and stratigraphical correlation of the two units appears to be well-founded and it is concluded that the Sugarloaf Schists are part of the Goose Tickle Formation.

SUGARLOAF SLIDE

A zone of intensely deformed granite and metasediment 53 m thick, named the Sugarloaf Slide, intervenes between the Acid and Basic Gneisses and the Sugarloaf Schists. Although it has not been possible to demonstrate that the slide was formed in close connection with folding, and hence does not fall within Fleuty's (1964b) definition of a slide, the term is used as the rocks exhibit typical textures and fabric of a slide.

The slide outcrops in the cliffs 300 m southwest of Sugarloaf Cove and consists of phyllonitized granite (30 m) overlain by tectonic schists (23 m). The phyllonitized granite is separated from the underlying Acid and Basic Gneisses by a late minor fault in which a 4 m thick wedge of gritty psammite and interbedded dark green quartz-biotite schist intervenes. The psammite contains pebbles of blue quartz (90%) in a sericite matrix. The quartz-biotite schist contains in addition, minor albite and microcline and accessory zircon.

Phyllonitized Granite

The phyllonitized granite (Knopf, 1931) is an intensely foliated coarse grained rock that consists of augen and lenticles of quartz and K-feldspar, up to 1 cm across with intervening layers up to 2 mm thick of fine-grained phyllonite (Plate 14, fig. b). The intensity of cataclasis varies from place to place within the slide so that lenses of schistose granite can be traced over a few feet into a phyllonite. The phyllonite fabric is locally folded about minor F_2 folds (Plate 15, fig. a) indicating that the sliding was a pre-D_2 event. The contact with the overlying tectonic schists is sharp and is exposed in a small stream gulley in the cliffs. The contact is parallel to the tectonic fabric in both units and is considered to be a slide of the same age as the major slide.

Tectonic Schists

The tectonic schists consist of an intensely foliated, light green to white, rusty weathered, fine grained calc-silicate rock with minor interbanded purple semi-pelitic schists (Plate 15, fig. b). The calc-silicate is non-schistose and consists of alternating discontinuous foliae, 2-5 mm thick of white, epidotic-carbonate foliae and light green, quartzo-feldspathic foliae. The semi-pelites are fine-grained, purple schists with thin, 1-2 mm, quartzo-feldspathic bands, and occur as discontinuous lenticles from 2 mm to 0.6 m thick in the calc-silicates.

The tectonic schists have been affected by post-tectonic thermal metamorphism with extensive porphyroblastic growth of diopside and tremolite-actinolite and minor sphene and idocrase in the calc-silicates foliae and of biotite and tremolite-actinolite in the semi-pelitic foliae.

The petrography of the tectonic schists is similar to the calc-silicates and semi-pelites included in the overlying Sugarloaf Schists and a description is not repeated here.

The banding of the tectonic schists is subparallel to S_2 but is folded around minor F_2 folds indicating that it is of D_1 age.

Discussion

The phyllonitized granite is a highly deformed granite and is probably related to the post-tectonic granites that intrude the underlying basement gneisses. The tectonic schists are petrographically similar to parts of the Sugarloaf Schists and likely represent highly deformed parts of that unit.

The Sugarloaf Slide is interpreted as a zone of intense shearing associated with the juxtaposition of the Sugarloaf Schists against basement gneisses.

RECONSTRUCTION OF THE DEPOSITIONAL ENVIRONMENTS OF THE AUTOCHTHON

The depositional environments of the autochthonous rocks are discussed in terms of a plate tectonic model whereby the rocks are considered to have formed a stable shelf area bordering a continental margin. It has long been known that the carbonate rocks found all along the western margin of the Appalachians record a vast carbonate bank (Cloud and Barnes, 1948; Rogers and Neale, 1963; Rodgers, 1969; Stevens, 1970) and that the rocks conform to Pettijohn's (1957) orthoquartzite-carbonate association. Pettijohn considered this association to have formed in a tidal environment bordering a shallow shelf sea and this work essentially confirms these views.

Discussion

The basal-clastic volcanic unit records a westerly transgression onto the Precambrian Long Range Complex. The basal plutonic boulder conglomerate seen only near Burnt Point, contains well rounded clasts that are grain supported and likely formed in a high energy, shallow water environment. The overlying volcanic rocks are pillowed near Burnt Point indicating sub-aqueous conditions, but 6.4 km to the north-west at Otter Cove and nearby Cloud Mountain the volcanics exhibit columnar jointing indicating a subaeriel environment.

The overlying Bradore sandstones rest on unweathered basement gneisses. The sandstones are compositionally mature with a low clay content and well rounded clasts. The unweathered nature of the basement suggests that the maturity of the sandstones was produced, not by the long period of weathering, but by prolonged transport and abrasion. Such an environment is found in an intertidal area and the sedimentary structures, the textural and compositional maturity, and the association with overlying carbonates conform to this model. The Bradore unconformity with fresh basement gneiss therefore probably represents marine erosion during initial transgression.

Williams and Stevens (1969) suggested a fluvial environment for the Bradore sandstones on Belle Isle. However such an environment is unlikely in the Canada Bay area as the sandstones do not show a fining upwards sequence or the poor sorting in grain size that is typical of a fluvial environment (Allen, 1965).

The sharp contact between the Bradore sandstones and the Devils Cove limestones and shales attests to an abrupt cutting off of the terrigenous supply. The limestone clasts in the breccia contained in the uppermost part of the Bradore Formation west of Fly Point suggest that carbonate deposition had been established nearby prior to the end of Bradore times. Gneiss clasts are also contained in the breccia indicating that the basement was also exposed at that time.

The Devils Cove Formation contains a shelly fauna that suggests a shallow marine environment. The overlying formations (the Forteau, Cloud Rapids and Traytown Pond) contain mostly limestones but also shales, sandy limestones and rare orthoquartzites that indicate occasional influx of terrigenous material into a shallow marine basin.

The Cloud Rapids and Traytown Pond limestones have not been recognized on the western side of the Great Northern Peninsula. There, a sequence of quartzites with minor limestone and shale named the Hawke Bay Formation (Schuchert and Dunbar, 1934) overlies the Forteau Formation. Betz (1939) attributed the absence of the Hawke Bay orthoquartzites in the Canada Bay area to erosion or non-deposition, a view later subscribed to by Williams (1964). However, the absence can also be explained by a lateral facies change from west to east. The Hawke Bay Formation, with mudcracks, flat pebble conglomerates, cross-bedded orthoquartzites and cryptozoan fossils, is typical of a near shore tidal deposit, whilst the limestones to the east are representative of a shallow marine environment. This suggests a westerly transgression of facies.

There was a temporary deepening of the basin of deposition at the beginning of St. George Formation times (upper Cambrian) indicated by deposition of black rubbly limestones. The succeeding dolomites with algal mounds, shelly fauna, mudcracks and cross bedding, were probably deposited in a shallow sea. The absence of terrigenous material indicates complete isolation from any detrital source. The St. George Formation was recently interpreted to be a slightly subtidal deposit (Swett and Smit, 1972).

At the end of St. George times (lower Ordovician) subsidence finally outpaced deposition and there was a deepening of the basin of deposition witnessed by deposition of black hackly limestones and shales of the Table Head Formation. Stevens (1970) interpreted the Table Head Formation to represent a bank edge and basin slope deposit. The intraformational conglomerate near the top of the sequence at Coles Pond suggests instability probably associated with subsidence. Further deepening of the basin in upper Table Head times resulted in cessation of lime deposition and formation of black pyritiferous shale.

In Llanvirn times there was a sudden influx of a northeasterly derived flysch (the Goose Tickle Formation) which heralds the arrival of the allochthon. The greywackes exhibit sedimentary structures typical of turbidites and the shales contain graptolites indicating deposition in deep water. The greywackes in the north of the area are medium grained, graded, show the lower three intervals of Bouma's (1962) ideal turbidite, and occur in beds of moderate thickness, indicating deposition from distal turbidites (Walker, 1967). In the south of the area the beds are thinner and finer grained indicating deposition from a more distal current. Thick, coarse conglomerate beds occur near the top of the formation north of Hare Bay (Tuke, 1968; Williams et al., 1973) indicating deposition from a proximal turbidite (Stevens, 1970).

As the flysch was derived from the east-northeast and contains detritus from the allochthon which later overran it, it is obviously related to the emplacement of the allochthon (Stevens, 1970). Using Reading's (1972) recent classification of flysch the Goose Tickle Formation can be considered as "Mediterranean Flysch".

CHAPTER III

THE ALLOCHTHON

The rocks of the Hare Bay Allochthon belong to a number of separate thrust slices. Each thrust slice is defined by different stratigraphy, different structural histories and by distinct upper and lower contacts that are generally marked by black shaley mélange zones.

South of Hare Bay four distinct and separate thrust slices comprise the allochthon. These are named from structurally lowest to highest, the Maiden Point Slice, the Croque Head Slice, the St. Julien Island Slice and the White Hills Slice. There is no continuous structural succession containing all four slices and relationships had to be built up from observations throughout the area. The St. Julien Island Slice is locally missing but reversals in the stacking order are unknown.

The rock units of each slice are bounded tectonically and separate formations within a slice generally have tectonic contacts. The allochthonous rocks cannot, therefore, represent a continuous sedimentary sequence and hence the stratigraphy of each slice is described separately.

The allochthon tectonically overlies the autochthonous sediments of the middle Ordovician Goose Tickle Formation, which gives a lower age limit for emplacement. Unfortunately no fossils were found in any of the allochthonous units in the map area and their ages are conjectural. The allochthonous rocks have suffered at least two phases of deformation and a slaty cleavage is ubiquituous in the lower slices and a schistosity is developed in the lower part of the White Hills Slice.

The structural stacking order of the slices and the stratigraphy of each slice is shown schematically in Fig. 10.

MAIDEN POINT SLICE

The Maiden Point Slice is areally the most extensive and structurally the lowest slice of the allochthon south of Hare Bay. It consists entirely of rocks assigned to the Maiden Point Formation which is composed mostly of coarse greywackes, slates and mafic volcanic rocks. The Maiden Point Formation has so far defied all attempts at subdivision on account of complex internal structure and lack of continuous marker horizons.

The Maiden Point Slice everywhere overlies the autochthonous Goose Tickle Formation except south of Canada Harbour where it overlies the Sugarloaf Schists. In the north of the area the contact is generally marked by black shaley mélange and contact relations are well exposed along the south shore of Hare Bay at Big Springs Inlet, Indre Point and Little Cormorandier Island (Fig. 3). The base of the slice is exposed again in Whites Arm Window and basal contacts are defined at Whites Arm. At Canada Bay mélange is absent and the contact is a sharp thrust with the underlying Goose Tickle Formation.

MAIDEN POINT FORMATION

Introduction

The Maiden Point Formation (Tuke, 1968; Cooper, 1937) in the map area consists predominantly of coarse greywackes and slates. Thick pebble conglomerates are locally interbedded with the greywackes in the north of the area. Mafic volcanic rocks, informally referred to as the Maiden Point volcanics, occur at or near the structural base of the formation in the central parts of the area to the east and west of Whites Arm Window, and at Little Cormorandier Island in Hare Bay. A conglomerate unit is associated with the volcanic rocks at Croque Harbour.

Two small areas of ultramafic rocks were found west of Grandois that appear to be related to the volcanic rocks. Thin marble wedges outcrop at the base of the Maiden Point Slice along the eastern margin of Whites Arm Window and are included as part of the formation.

Cooper (1937) defined the excellently exposed section along the south shore of Hare Bay as the type section. However, this section only includes greywackes and slates, estimated to be in the order of 1,200 - 1,800 m (4,000 - 6,000 ft) thick. An accurate estimation of the thickness cannot be made on account of numerous late faults, complex structure, and lack of continuous marker horizons. Betz (1939) estimated a similar order of thickness for identical greywackes and slates exposed at Canada Bay. Betz named these rocks the Canada Head Formation, but as they are clearly correlatable with the Maiden Point Formation to the north and as they occur in the same structural slice, the name Maiden Point Formation is applied here to all the rocks of this slice.

Small areas of mélange locally separate continuous sections of the slice at Deep Bay and at Little Canada Harbour. They are not continuous, do not separate slices, and their possible origins are discussed in Chapter IV.

Greywackes and Slates

Monotonous greywackes and slates comprise approximately 90% of the formation in the map area. The greywacke is typically a dark grey, massive, medium to coarse grained rock that weathers light grey to buff. The greywackes form graded beds from 0.5 to 3 m thick with 1 m being the average thickness (Plate 16, fig. a). Generally two intervals can be recognized in a sandstone bed. A lower well-graded interval 8-20 cms thick, containing small pebbles to coarse sand and a thick upper interval of coarse to medium grade sands with poor to absent grading (Plate 16, fig. b). Black to grey but occasionally red, green and turquoise blue slates are interbedded with the greywackes. The slates vary in thickness from a few centimeters to 9 metres, but are generally 0.3 to 0.6 meters thick. The slates are not everywhere developed and they only comprise about 20% of the greywacke-slate lithology. Internally the slates are thinly-laminated and rarely they show small-scale cross bedding.

The greywackes are coarsest grained and thickest bedded in the northwest at Hare Bay where they are associated with pebble conglomerates. They are finest in the southeast at Cat Cove where the lower coarse-graded intervals are poorly developed and the upper intervals commonly consist of fine-grained pelites 0.6 to 1 m thick. Betz (1939) incorrectly identified the pelites as volcanic rocks. But, they are obviously sediments as they pass downwards into recognizable fine-grained greywackes, and in thin section contain sparse grains of clastic zircon. The fine greywackes and pelites are generally featureless apart from numerous thin (2-5 m) quartz veins that are usually arranged parallel to bedding. The veins are folded by inclined F_2 folds and are considered to be of D_1 age. Late thermal metamorphism (see chapter VII) in the Cat Cove area has altered the pelites to a purple hornfels.

Graded bedding is nearly always present and is expressed by a decrease in the number and size of the coarse grains from the base upwards. The finer grains are present throughout both intervals. Cross-bedding, conspicuous by its rarity, is restricted to the top of beds, and generally forms in those parts of the sequence where the slates are absent. It is usually found in thin lenses of buff to brown weathered greywackes that have a slightly calcareous matrix (Plate 17, fig. a). The lenses are up to 0.6 m thick and 3.6 m long, and generally contain only single set cross beds. Similar greywacke with a calcareous rich matrix forms large spheroids up to 0.3 by 0.6 m at the top of graded beds south of Cobblers Cove.

Soft sediment deformation structures are extremely scarce and consist of loadcasts and slumps. Load casts were only seen north of Conche and are best developed north of Pilier Bay. They form at the base of greywacke beds that overlie shales and vary from large bulbous loads up to 0.6 m wide to lobate casts up to 30 cms long and 15 cms wide. (Plate 17, fig. b). The lobate casts overlap regularly towards the southeast, suggesting that their formation was current controlled.

Slump structures were only seen in two exposures. On Spring Island hook and roll up (Ksiaziewicz, 1958) structures (Plate 18, fig. a) occur at the top of a bed in a lens of cross-bedded greywacke. South of Mare Cove thin laminae of alternating medium and very fine sandstones are disturbed producing tabular and rolled up discontinuous lenticles of medium grained greywacke (Plate 18, fig. b). The lenticles are commonly 7-10 cms long and are confined by the primary bedding planes which are not affected. Their formation appears to have been by slumping or sliding within beds.

Petrography:

The greywackes are composed dominantly of quartz (up to 60%), matrix (up to 74%), and plagioclase (up to 24%), with minor amounts of metaquartzite, shale flakes, K-feldspar, chloritized biotite, muscovite, zircon, apatite, tourmaline, and iron ore. When grouped into the three fundamental end members (Fig. 11) they plot in the greywacke to sub-greywacke field.

The quartz is of plutonic (85%), vein (10%) and metamorphic (5%) origin. Zircon inclusions were seen in four grains of plutonic quartz, and chlorite is a common inclusion in the vein quartz. The feldspar consists almost entirely of twinned and untwinned plagioclase (An 0-15), and only a few grains of microcline and orthoclase were positively identified. The feldspar, in particular the untwinned variety, is partially altered to sericite and chlorite. Detrital biotite, now almost completely chloritized is sparse, but muscovite flakes are present in most sections. The accessory minerals are small (0.5 mm) and sparse and pyrite is the only accessory to form over 1% of the total rock in one section.

Extreme variations in the grain size of the quartz and feldspar grains are common and most sections contain grains from granule to medium sand size. The quartz grains are generally larger than the feldspars in any one section. The matrix consists of a fine-grained (less than 0.05 mm) aggregate of sericite, chlorite, quartz and feldspar and is partially replaced by carbonate.

Most grains are subrounded to angular (0.3 - 0.5 mm), however some sections contain a small number of highly rounded and near spherical quartz grains (Plate 19, fig. a). The greywackes are classed as poorly sorted and texturally immature on account of their wide variations in grain size and their high matrix content.

Quartz Pebble Conglomerate

Quartz pebble conglomerates, (marked by C in Fig. 3) occur throughout the Maiden Point Formation but are most common in the north of the area, especially along the south shore of Hare Bay where 13 occurrences from 1 to 6 m thick are exposed.

They consist of subrounded pebbles up to 2.5 cm in diameter of blue, milky, and black, quartz; metaquartzite, chert, plagioclase feldspar, microcline and sandstone. Black mudstone flakes are frequently associated with the conglomerates and form from 0-50% of the rock (Plate 19, fig. b). The flakes generally show a rough alignment of their long axis parallel to bedding. Rarely the mudstones form angular slabs up to 0.6 m long (Plate 20, fig. a and b). The mudstone clasts are interpreted to represent contemporaneous channelling and erosion of intercalated shales.

Generally no internal grading exists within the conglomerates which have sharp tops and bottoms. The conglomerates could not be traced along strike for more than 30 m and are thought to represent coarse channel infills within the greywackes. Not surprisingly they have a similar petrology to the greywackes. One section examined contains a clast of medium-grained hematitic sandstone (Plate 21, fig. a) identical to the sandstones at the top of the Bradore Formation.

Maiden Point Volcanics

Altered agglomerates, tuffs, and basaltic pillow lavas are interbedded with the greywackes of the Maiden Point Formation in the north of the area. They form a number of discontinuous horizons up to 150 m thick and 8 km (5 miles) long and are located at or near the structural base of the Maiden Point Slice. They are exposed north of Coles Pond, west of Whites Arm Pond, from Croque Harbour to Whites Arm and on Little Cormorandier Island (See Figs. 3 and 4). Each horizon consists of a number of flows separated by flow brecciated lava or agglomerate, or more rarely, by tuff, and in one locality north of Croque by a 0.6 m thick limestone bed.

The lavas are green to grey to purple, amygdaloidal and generally well-cleaved. They vary from fine-grained, altered, greenstones at the margins of flows, to medium grained less altered rock in the interior. On Little Cormorandier Island the volcanic rocks rest directly on a mélange zone, and are conformably overlain by greywackes and slates. There the volcanic rocks consist of a massive flow followed by tuff and pillow lava. The pillows vary from 8 to 45 cms in diameter with limestone filled interstices (Plate 21, fig. b). At northeast Croque subrounded inclusions of dark grey, medium-grained basalt up to 8 cm across are included in typical green fine-grained basalt (Plate 22, fig. a).

Petrography:

The medium-grained basalt has a subophitic texture consisting of cloudy albite phenocrysts up to 1.5 mm long, subidiomorphic clinopyroxene grains up to 1 mm long and interstitial pools of chlorite. Generally, however, the rock is heavily altered with complete sericitization of the feldspar and growth of fibrous amphibole, chlorite and epidote. Typically the fine-grained basalts show intersertal texture with the irregular spaces between small (0.8 m) albite laths occupied by chlorite epidote, calcite and leucoxene. In some cases the albite laths are completely sericitized or replaced by calcite. Rare stilpnomelane laths statically overgrow and post-date the main alteration minerals (Plate 22, fig. b). The basalts at the outer margins of the horizons are fine grained, aphanitic, extremely altered and well cleaved. The cleavage, defined by aligned chlorite and fibrous amphibole tends to form augen around large grains of epidote and cuts across locally developed calcite veinlets. At least some of the alteration, therefore, pre-dates the regional deformation. Amygdules, up to 3 mm across are infilled with clear twinned albite, chlorite and epidote.

Geochemistry:

Major element analyses were carried out on 16 specimens of the Maiden Point volcanics by atomic absorption spectrometry. The results are summarized in Fig. 12 and 13 where they are compared with data from two different groups of basalts, the Lighthouse Cove Formation and the Lushs Bight Group. The Lighthouse Cove basalts are Eocambrain continental tholeiites (Strong and Williams, 1972) from the mio-geosynclinal Western Platform, and the Lushs Bight basalts on the other hand are Cambrian (?) oceanic tholeiites (Smitheringale, 1972) from the Central Mobile Belt.

The alkalies-silica diagram (Fig. 12) shows both the Maiden Point Volcanics and the Lighthouse Cove Formation to be transitional between tholeiitic and alkali basalts, whilst the Lushs Bight basalts are clearly tholeiitic. However, calculation of the CIPW norms for the Lighthouse Cove Formation showed in fact these basalts to be tholeiitic (Strong & Williams, 1972).

Variations of SiO_2, MgO and CaO with Na_2O (Fig. 13) show no detectable differences between the three groups. However, the TiO_2 variation diagram revealed a significant difference. Both the Maiden Point volcanics and the Lighthouse Cove Formation contain similar proportions of TiO_2 that are significantly greater than that of the Lushs Bight Group. Ti, according to Cann (1970), is one of the few elements whose abundance is not affected by secondary alteration, and it must be remembered that the Maiden Point volcanics show petrographic indications of extensive alteration.

The Maiden Point volcanics thus show definite chemical similarities to the Lighthouse Cove Formation and none to the Lushs Bight Group. They are therefore considered to be tholeiites that formed in a similar environment of continental rifting, distension and separation. The field associations of both the Lighthouse Cove Formation and the Maiden Point volcanics with coarse clastics further emphasizes their similarities.

Ultramafic Rocks

Two small areas of serpentinized ultramafic rocks occur within the Maiden Point Slice near Grandois. One occurrence is 0.8 km west of Grandois, at the foot of a prominent ridge that marks the contact between volcanics on the ridge and greywackes on the lower ground. Contact relations are not exposed. The second occurrence is within Whites Arm Window, 2.4 km west of Grandois. There a 60 m long hummock of brown weathered ultramafic rock appears to overlie and be surrounded by slates and greywackes of the Goose Tickle Formation. The ultramafic rock is cleaved and folded by regional deformations and is completely serpentinized. The ultramafic rock may represent an erosional remnant of the once continuous Maiden Point Slice (See Fig. 3, section C-D). Nearby other remnants of volcanics and greywackes of the Maiden Point Slice occur on topographically higher ground.

The ultramafic rocks show a spatial relationship with the Maiden Point volcanics and are considered to be related. Alternatively, the ultramafic block exposed in Whites Arm Window might represent a 'knocker' that was derived from the White Hills Peridotite Sheet during emplacement and was later overridden by the Maiden Point Slice. Occurrences of large ultramafic blocks below the lower sedimentary slices have recently been reported from the Humber Arm Allochthon (Stevens and Williams, 1973).

The possibility of intrusions being associated with the volcanic rocks raises the question that some of the coarsest parts of the Maiden Point volcanics may also be intrusive, for example, the mafic rocks at northeast Croque and north of Coles Pond. However, the field evidence necessary to prove this was not found.

Conglomerate

A 7 m thick conglomerate unit occurs near the base of the Maiden Point volcanics at the north west arm of Croque Harbour. The unit outcrops on both sides of the arm but could not be traced inland. The stratigraphy, constructed from both exposures is shown diagrammatically in Fig. 14.

The lower contact of the conglomerate unit is not exposed on the south side of the arm, where it is underlain across a small gap in exposure by the autochthonous Goose Tickle Formation. On the north side a basalt flow underlies the conglomerate. The upper contact of the conglomerate unit is exposed on the south side of the arm where cross-bedded sandstones are overlain by another basalt flow (Plate 23, fig. a). The conglomerate unit appears to be contained within the Maiden Point volcanics and is therefore included as part of the Maiden Point Slice.

The conglomerate consists of boulders and cobbles of basalt (80%), marble (20%), and rare sandstone. The basalt clasts are from 8 cms to 30 cms in diameters, well rounded and oval-shaped (Plate 23, fig. b). Some show concentric chilled margins and may have been pillows. In thin section they are completely altered and consist of calcite, chlorite, and relict plagioclase laths. The marble clasts are up to 0.6 m long and

10 cms thick and are stretched and flattened in the plane of the cleavage which is subparallel to the bedding. The marbles are pure carbonates with only a few sparse quartz grains. The conglomerate matrix and the overlying cross-bedded sandstones are petrographically similar and consist of medium-grained sand with a calcareous rich cement.

Marble Wedges

Thin marble wedges up to 10 m thick intervene between the volcanics at the base of the Maiden Point Slice and the autochthonous Goose Tickle Formation at 1.6 km south of Grandois Pond and south of Croque. A similar marble bed 0.6 m thick occurs between two lava flows north of Croque at Forest Hill. Contact relations of the basal wedges were not seen, but they are considered to be part of the Maiden Point Formation as there are no known limestones at the top of the Goose Tickle Formation, and as they are similar to the obviously allochthonous bed at Forest Hill.

The marbles are fine-grained, white to buff, but at Croque are purple where they are interbedded with greywacke beds. Thin laminae, less than 3 mm thick, of green chloritic slate characterize the marbles and exhibit the bedding and the tectonic fabric. The slates consist of chlorite, sericite, and stilpnomelane, with a few grains of quartz and plagioclase. Late, static tourmaline up to 0.55 mm long is sparsely developed in the shales.

Age and Provenance

The age of the Maiden Point Formation is uncertain but several lines of evidence suggest a lower Cambrian-Eo-Cambrian age. The presence of blue quartz and metaquartzite detritus indicates a source from a metamorphic crystalline terrane and a westerly derivation from the Grenville Province has previously been suggested (Stevens, 1970). If the Grenville was the source, then the sediment must have been eroded prior to deposition of the autochthonous carbonate cover rocks in lower Cambrian times. This suggests that the formation is coeval with the lower Cambrian Lighthouse Cove-Bradore Formations and the presence of a hematitic sandstone clast and the close chemical similarity between the basalts of the Lighthouse Cove and Maiden Point Formations tend to confirm this view.

Fabric

The Maiden Point Formation suffered two deformational episodes that varied in intensity from place to place so that the rocks exhibit marked variations in fabric. Generally, the rocks in the northwest of the area are less deformed and here the greywackes retain their original clastic structure, the volcanic rocks are freshest, and a single weak cleavage is developed. Southeastwards the volcanic rocks are extensively altered and well cleaved, and locally the greywackes give place to semi-schists (Turner, 1938; Spry, 1969) in which the clastic structure has been partially obliterated, the grain size is reduced, and a composite tectonic fabric is developed. Semi-schists occur in restricted areas within the Maiden Point Slice at Cow Bay in Hare Bay, and in the Canada Bay-Cat Cove area.

Discussion

The greywackes of the Maiden Point Formation are compositionally mature but texturally immature. The sedimentary features are typical of proximal turbidites (Walker, 1967) and the general southeastwards-fining and decrease in bed thickness suggest sediment transport from the northwest. The associated volcanic rocks are tholeiitic in composition which suggests formation in an environment of distension.

CROQUE HEAD SLICE

The Croque Head Slice overlies the Maiden Point Slice and outcrops in a narrow strip up to 0.8 km wide along the coast from St. Juliens to the Croque Head Peninsula. The basal contact is sharp and is defined by a well exposed mélange zone at St. Juliens, Irish Bay and Rets Point (see Figs. 3 and 5).

The Croque Head Slice consists almost entirely of greywackes and slates of the Maiden Point Formation with one thin flow or sill of mafic volcanic rock exposed south of Wild Cove. However, the greywackes are finer grained, generally thinner bedded, and exhibit a more complex internal structure than those in the Maiden Point Slice. In most places the greywackes have been converted to semi-schists.

The petrography of the greywackes and volcanic rock is similar to those in the Maiden Point Slice and a description is not repeated here.

ST. JULIEN ISLAND SLICE

The St. Julien Slice overlies the Croque Head Slice and is areally the least extensive slice of the allochthon. It outcrops only on St. Julien Island and on nearby Black Island (Fig. 3 and 5). The basal contact with the Croque Head Slice is marked by a mélange zone that is best exposed at the south end of St. Julien Island. The top of the slice is not exposed.

The St. Julien Island Slice contains two formations, juxtaposed by an early high angle fault, the Irish limestone to the west, and the St. Julien Island Formation to the east. These rock units were previously assigned to the basal part of the uppermost White Hills Slice (Williams et al., 1973). However, they do not outcrop elsewhere in the Hare Bay Allochthon and are atypical of the lithologies of the White Hills Slice. Consequently they are considered to comprise a distinct and separate slice that intervenes between the Croque Head Slice and the overlying White Hills Slice.

IRISH LIMESTONE

Sandy limestone that outcrops along the western side of St. Julien Island was named the Irish Limestone (Smyth, 1971). The formation overlies a mélange zone which separates it from the underlying Croque Head Slice (Fig. 5). The formation occupies the western limb of a faulted northeast plunging syncline so that the base of the slice is exposed along the west and south sides of the island. The formation consists of three units and each tectonically overlaps onto the mélange. The stratigraphy and thickness of each unit is summarized in Fig. 15.

Lithology

The basal unit consists of quartzites with minor interbedded greenschist. The quartzites are sheared and brecciated, cut by numerous thin quartz and carbonate veins and characteristically have brown, carbonate stained, weathered surfaces. The greenschists are up to 1 m thick and are best exposed at the southern tip of the island where they overlie the mélange. They possess a fine schistose lamination of alternating dark and light green bands composed of varying amounts of chlorite and sericite. This schistosity is folded by late upright, post-emplacement folds that form the synclinal structure of the formation.

The contact between the quartzite unit and the overlying siliceous limestones and slates is gradational. The basal limestones are thinly bedded (5-8 cm) and weather brown. They are interbedded with minor thin green slates and fine-grained green sandstones. Higher up the succession, these limestones gradually lose their brown colour and weather grey, they become thicker bedded, and the green slates and sandstones are absent. Individual beds are from 15 cms to 30 cms thick and have sharply defined bottoms and indistinct top surfaces with transition to black, thin laminated, shales (Plate 24, fig. a). Graded bedding is developed but is difficult to recognize on account of the small grain size (0.2 - 0.5 mm).

The limestone is locally extensively brecciated within a few meters of the contact with the mélange zone and in two localities the breccia passes into the mélange. Both the mélange and the breccia are folded by the upright F_2 folds indicating that their formation was a pre-D_2 event. (See Chapter IV).

Petrography

The limestones are impure and contain up to 30% terrigenous material that consists of quartz (70%), feldspar (30%) and rare grains of muscovite, tourmaline, zircon, and pyrite. The quartz grains are of the plutonic type with predominantly straight to slightly undulose extinction. Feldspar consists of fresh, and strongly altered, albite grains in approximately equal proportions. The plagioclase alteration occurred after deposition and generally includes sericitization and calcitization. The terrigenous material floats in a matrix of fine grained recrystallized calcite with some authigenic pyrite.

The clastic grains are of medium sand size at the base of beds and grade into fine sands. Texturally the limestone is immature and moderately sorted. Quartz and feldspar grains show a variety of rounding stages but are mainly subangular.

The sandy limestones show clastic textures and are considered to have formed from carbonate rich turbidity currents. The thin beds and the fine grain size suggest deposition from distal turbidites.

Age and Provenance

A search for conodonts was unsuccessful and the age of the formation can only be inferred. The only known source of carbonate detritus is from the autochthonous platform to the west, and the admixture of carbonate and sand detritus suggests that the formation may be equivalent to the middle Cambrian impure limestones and shales of the Forteau, Cloud Rapids and Treytown Pond Formations of the autochthon.

ST. JULIEN ISLAND FORMATION

Introduction

The polymictic cobble conglomerate that outcrops on the western half of St. Julien Island and on Black Island (Fig. 5) was named the St. Julien Island Formation (Smyth, 1971). The formation is in high angle fault contact with the Irish Limestone to the west. Neither top nor bottom of the formation is exposed, and a minimum thickness of 60 m is estimated from exposures on St. Julien Island. The conglomerate is locally cut by small pre-tectonic gabbro intrusions at Black Island and at the east and north sides of St. Julien Island. The petrology of the gabbro is described in Chapter V.

Lithology

Cobble conglomerate with less common boulder and pebble clasts, is the dominant lithology. It is interbedded with minor amounts of red to green to purple coarse greywackes (Plate 24, fig. b). Massive bedding is typical, grading is poorly developed except where interbeds of greywackes occur. Trough cross-bedding (Plate 25, fig. a and b) with the lower bounding surface downcutting 1.8 m is excellently exposed in interbedded coarse greywackes and cobble conglomerates at one locality on the east side of the island. The trough axis runs NNE-SSW but a current direction was not evident.

Clasts constitute approximately 60% of the rock and they are isolated and scattered throughout the matrix (Plate 26, fig. a). The matrix varies from red to purple to green and consists of the same components as the clasts dispersed in a fine-grained sericite-chlorite groundmass. The matrix fragments vary from granule to coarse sand size and are subrounded to angular. The interbedded greywacke beds are similar in composition and texture to the matrix.

The conglomerate is very poorly sorted with clasts ranging from pebble to boulder size. Generally the larger clasts are moderately to well rounded whilst the finer particles are more angular. Clast sphericity is moderate but is difficult to estimate on account of later tectonic flattening. Some of the red greywacke clasts are recognizable as portions of broken up beds and form intraformational monomictic conglomerate.

The St. Julien Island conglomerate is texturally and compositionally immature as it contains a variety of unstable (for example, granite and shale) and stable (for example, quartz, jasper) materials, as well as containing a high matrix content. The dispersed texture, high matrix content, and graded bedding, indicate deposition from a highly turbulent flow such as a turbidity current.

Petrography of the Clasts

The petrography and approximate abundance of the clasts is as follows:

Quartz	30%	Dark grey, red and green shale.
Red Sandstone	40%	Mafic Volcanic Rock.
Acid Volcanic Rock	15%	Sericitic Phyllite
Granite	5%	Mafic Tuff
Jasper	2%	Gneiss Epidozite Quartz-chlorite schist

Rare, Less Than 1%.

Quartz: All clasts sectioned are of vein quartz variety.

Red Sandstone: Orthoquartzite with rare clasts of plagioclase, microcline and orthoclase. Clasts are sub-spherical and of medium sand grain size. Grains are closely packed and the sericite-carbonate-hematite matrix forms less than 10% of the rock.

Acid Volcanic Rock: The acid volcanic clasts contain embayed and rounded phenocrysts, up to 4 mm across, of quartz and andesine set in a fine-grained felsitic matrix. Relict garnet now altered to epidote was seen in one section. The matrix includes interstitial pools of chlorite. The rock is probably a dacite.

Granite: Two varieties of microgranite clasts are present, a leucogranite and a red hematite stained granite. The leucogranite consists of quartz (40%), albite (40%), orthoclase (10%), minor chlorite and epidote and accessory iron ore. Myrmekitic texture is characteristically developed (Plate 26, fig. b). This rock can be classed as a microgranodiorite.

The red stained granite contains approximately 40% alkali feldspar and hematite forms from 1-3% of the rock. The rock is an alkali-microgranite.

Mafic Volcanic Rock: The mafic volcanic clasts contain phenocrysts (up to 3 mm long) of cloudy plagioclase (now An_{40}) in a highly altered groundmass composed mainly of epidote with minor chlorite and calcite. The rock is probably an altered basalt.

Gneiss: One clast of a thinly banded quartzitic gneiss was found. The foliation is defined by alternating quartz rich and mafic rich bands 1-3 mm thick. (Plate 27, fig. a). The mafic bands contain relict euhedral garnet (Plate 27, fig. b), now altered to sericite and chlorite, hematite, epidote and apatite. The garnet crystals statically overgrow and postdate the foliation. Unfoliated epidote rich areas up to 4 mm across are contained within the gneissose foliae.

Sericitic Phyllite: Clasts composed almost entirely of pale sericite laths are present in minor amounts. The clasts show a strong schistosity that is coincident with the fabric in the surrounding matrix. The original rock was probably a feldspar pegmatite, and the sericitization appears to be a post-depositional event.

Quartz-Chlorite-Schist: A single clast of dark green, quartz chlorite schist was found. The quartz shows a strong dimensional, predepositional, preferred orientation that is partially obliterated by recrystallization. Chlorite forms approximately 15% of the rock and occurs as scattered flakes that show no preferred orientation. The original rock was probably a deformed quartz vein.

Epidozite: Light green, fine grained epidozite clasts are rare. They contain an intergranular texture of epidote and quartz with minor carbonate. Their origin is problematical but they resemble the small epidote rich areas in the gneiss clast described above.

Jasper: Jasper clasts are conspicuous and form clasts of cobble and boulder size. R.K. Stevens (pers. comm.) reports radiolaria from the jasper, but none was seen in my samples.

Fabric

The St. Julien Island Formation is strongly deformed. The matrix and the interbedded greywackes are converted to foliated quartz-sericite-chlorite semi-schists (Spry, 1969). This foliation forms augen around the clasts which are flattened and elongated in the plane of the foliation. The amount of flattening depends largely on the composition of the clasts, and this is discussed further in chapter VI.

Age and Provenance

Clast provenance must have been from a volcano-plutonic terrane that included areas of sedimentary rocks. No such terrane existed in the stable platform to the west and hence a source must be sought from within the eugeosyncline to the east.

The Cape St. John Group (Baird, 1951) which forms the upper part of the pre-Arenig Fleur de Lys Supergroup (Church, 1969) on the eastern side of the Burlington Peninsula, consists of rhyolite, trachyte, andesite, tuff, sandstone, quartz boulder conglomerate and chert and is the only known likely source. The presence of pre-tectonic gabbro in both rock units is a further correlative. The source of the single gneiss and rare schist clasts are problematical and suggest derivation either from a gneissic terrane or a reworked conglomerate.

The age of the formation is uncertain but is likely to be of similar or slightly younger age than the Cape St. John Group, which is considered to be of late Cambrian age (Dewey and Bird, 1971).

WHITE HILLS SLICE

The White Hills Slice is the highest slice of the allochthon. It consists of a lower unit of schistose metavolcanic and minor sedimentary rocks referred to as the Goose Cove Formation, (Tuke, 1968; Smyth, 1971) overlain by ultramafic rocks named the White Hills Peridotite Sheet (Cooper, 1937). In the map area only rocks of the Goose Cove Formation are exposed. However north of Hare Bay the complete slice is excellently exposed in the White Hills, from whence it takes its name.

In the map area, the White Hills Slice overlies the Croque Head Slice south of Croque Head (Fig. 3), and a narrow mélange zone marks the contact. The base of the White Hills Slice is also well exposed all along the western side of Fishot Island where it overlies mélange. Greywackes of the Maiden Point Formation underlie the mélange at the southwest end of the island (Fig. 4) but it is not known if they belong to the Maiden Point or Croque Head, Slice.

GOOSE COVE FORMATION

Introduction

The Goose Cove Formation (Tuke, 1968; Smyth, 1971) consists of a polydeformed and metamorphosed sequence of mafic pillow lava, tuff, agglomerate, and gabbro with thin interbands of greywacke and limestone. The type section is north of Hare Bay where the formation underlies and forms a metamorphic aureole to the White Hills Peridotite Sheet. The metamorphic grade of the formation decreases with distance downward from the base of the peridotite sheet and the resultant lithologies grade outwards from foliated pyroxene bearing amphibolites directly below the ultramafics, through schistose amphibolites to schistose greenschists. The transition from amphibolite to greenschist occurs over a few tens of meters and is marked by a decrease in grain-size of macroscopic hornblende and an accompanying increase in development of chlorite characteristic of the underlying greenschists. The transition is sufficiently distinct that it can be mapped and an amphibolite member and a greenschist member distinguished. In the map area the base of the White Hills Slice occurs at greenschist level on Fishot Island but transgresses upwards to amphibolite level south of Croque Head. It is clear then that the base of the slice bears no genetic relationship to the aureole.

Cooper (1937) originally restricted the term Goose Cove Schist to the greenschists. However, it is clear the greenschists and overlying amphibolites were derived from similar protoliths and that the distinction between the two is purely a function of metamorphic grade. Consequently, the term Goose Cove Formation (Tuke, 1968; Smyth, 1971) has been used to include all of the metamorphosed supracrustal rocks of the aureole. Map units of the formation defined on grade of metamorphism are informally referred to as members.

Distribution and Thickness

The formation underlies most of the Fishot Islands and a small area south of Croque Head. A cross section from the structural base of the greenschist member through to the amphibolite member is excellently exposed across the south end of Fishot Island where it is repeated twice by a late fault. At Fishot the formation is approximately 120 m thick with top unexposed. The lower 80 m is assigned to the greenschist member, but any estimate of the thickness is complicated by the internal polyphase folding that is characteristic of the formation.

Greenschist Member

Lithology:

Medium-grained, light grey to red, greywackes up to 10 m thick form the base of the greenschist member on Fishot Island. It outcrops in a narrow strip along the western edge of the island and structurally overlies the basal mélange of the slice. At the base, the beds are about 1 m thick, poorly graded and are interbedded with red slate. At the top, the beds are thinner and are interbedded over a few meters with purple and green agglomerate and tuff. Single greywacke beds up to 1 m thick occur interbedded with tuff and lava on Pigeon and Great Verdon Island (Fig. 4). The greywackes petrographically and lithologically resemble the semi-schists (Turner, 1938; Spry, 1969) of the Croque Head Slice with which they were correlated (Smyth, 1971).

Agglomerates with mafic volcanic clasts up to 10 cms across occur near the base of the sequence south of the western entrance to Fishot Harbour (Plate 28, fig. a). Recognizable pillows are rare in the thesis area but are common north of Hare Bay. The lavas are fine-grained, light to dark green, and commonly contain feldspar phenocrysts, up to 1.5 cm across in the centres of pillows (Plate 28, fig. b). Thinly-foliated, light to dark green, tuff is the dominant lithology of the member. Bedding in the tuff is preserved in the less deformed parts of the sequence. Thin limestone beds up to 0.3 m thick are interbedded with tuffs at the east side of Landing Cove and on the west side of Great Verdon Island.

Fabric:

The intensity of deformation is zoned within the greenschist member and the rocks vary from semi-schists to intensely foliated schists. In the semi-schist areas, the greywackes retain clastic textures, pillows are moderately flattened, and feldspar phenocrysts are flattened and elongated with XYZ ratio of 1:2:3.

Most of the member, however, has been converted to fine-grained, well foliated schist. The greywackes are converted to fine-grained psammites and semi-pelites with a fine muscovite schistosity. The volcanic rocks are altered to chlorite-actinolite-schists. Feldspar phenocrysts are almost completely flattened into thin bands with XYZ ratio of 1:4:10.

With increasing deformation in narrow restricted zones the rocks pass into intensely foliated streaky schists. These schist zones were referred to as tectonic slides by the author (1971). The slide zones are up to 6 m thick and appear to be conformably bounded by the surrounding schists into which they grade. The best example is at the southwest tip of Fishot Island where a zone of intensely foliated purple quartzite occurs interlayered between mafic tuffs. It consists of discontinuous quartzitic foliae 1-2 mm thick separated by streaks of muscovite and magnetite. The banding is folded by inclined F_2 folds and is clearly of D_1 age.

The S_1 schistosity is the dominant fabric element of the greenschist member. It is folded by tight F_2 folds that produced a strain slip S_2 schistosity. Small open folds, kink bands, and shear zones of D_3 age affect the earlier structures. The shear zones are from 0.5 to 2 m wide and consist of angular schist fragments set in a brown weathered quartz-carbonate matrix.

Petrography:

<u>Tuff</u> - The tuffs consist of thin alternating actinolite rich and calcite rich bands. The actinolite is subidiomorphic, 0.1 - 0.25 mm in grain size, and shows a strong preferred orientation that defines the S_1 schistosity. It occurs with scattered granular and prismatic epidote crystals up to 0.5 mm long, that have a weakly developed preferred orientation. The S_1 fabric forms augen around some of the single granular crystals suggesting that part of the alteration preceded the first deformation. In some sections, chlorite and granular epidote also form distinct, but discontinuous bands. The calcite rich bands contain, in addition, minor plagioclase (An_{6-10}), epidote, and quartz. Most of the plagioclase is twinned according to albite and albite-pericline laws, but the grains are almost completely saussuritized.

<u>Porphyritic Pillow Lava</u> - The feldspar phenocrysts are completely altered to a light-coloured sericite-carbonate-epidote assemblage in the interior with a thin outer rim of dark brown, epidotic-carbonate. A weak schistosity in the groundmass, defined by tremolite, forms augen

around the phenocrysts (see Plate 28, fig. b). The groundmass consists of plagioclase laths, tremolite, epidote and accessory iron ore. The tremolite is subidiomorphic, 0.3 - 1 mm in grain size, and shows a weak preferred orientation. The plagioclase (An_{12}) occurs as small, twinned, laths up to 1 mm long that are only partially saussuritized. Epidote occurs as scattered granular grains up to 0.2 mm across and the iron ore is oxidized to leucoxene.

Psammites and Semi-Pelites - The psammites contain sub-angular grains, from 0.5 - 3 mm in diameter, of quartz, quartzite, K-feldspar and plagioclase set in a fine-grained sub-schistose sericitic matrix. In the semi-pelites muscovite and chlorite laths, up to 1 mm long, define a strong L-S tectonite fabric. Quartz occurs in stringers and lenticles, 0.5 to 2 mm thick. In the thinner bands the quartz shows a strong preferred shape orientation but this gives way to a granoblastic polygonal texture in the thicker bands. Magnetite is a common accessory and occurs as small rods (0.5 mm long) aligned parallel to the schistosity, which is a composite S_2 fabric.

Slide Zone - The intensely folded quartzitic schist zone at the southwest tip of Fishot Island consists of a granoblastic polygonal network of quartz crystals, 0.05 mm across, with a high proportion of magnetite (20%) and minor sericite, chlorite, and apatite. Magnetite, muscovite and minor garnet occur concentrated in thin interbands up to 1.5 mm thick. Laths of sericite and chlorite, 0.03 - 0.05 mm long, grow along and across the quartz grain boundaries and show a strong preferred orientation that defines the S_2 schistosity. The garnet is idiomorphic, 0.03 - 0.05 mm in diameter and dirty brown. The S_2 fabric forms augen around the garnet indicating that it predates this fabric. Magnetite forms small octahedra up to 1 mm across that also pre-date the S_2 fabric. Muscovite microporphyroblasts up to 0.5 mm across, overgrow and post-date the S_2 fabric.

Small discontinuous quartz veins are folded by the F_2 folds and consist of crystals, up to 1.5 mm long, that have sutured grain boundaries and a weak preferred orientation parallel to S_2.

Amphibolite Member

The amphibolite member is derived from mafic volcanic rocks, tuff, minor gabbro, and minor thin interbeds of limestone and pelite. This is a similar protolith to that of the greenschist member except for the lack of psammite and semi-pelite and the presence of gabbro.

The amphibolite member structurally overlies the greenschist member and is exposed in a 400 m wide strip along the east side of Fishot and Northeast Islands. The narrow transition zone which marks the contact is well exposed at the east side of Landing Cove and at Easter Tickle. Another occurrence of amphibolites at Fishot Island is in a narrow, discontinuous strip at the west side of Landing Cove (Fig. 4). Here the amphibolite member is only 15 m thick, with top cut out by a late fault that repeats the greenschist succession. South of Croque Head the amphibolite member rests directly on a mélange that separates it from the Croque Head Slice.

Lithology and Fabric:

Volcanic rocks and tuffs, converted to hornblende schists, are the dominant lithology of the member. No primary volcanic features are preserved and the rock is a dark grey to black, generally massive, well foliated schist. However, it is distinctly banded where derived from bedded tuffs (Plate 29, fig. a). The schists are mostly medium to fine-grained with hornblende crystals up to 1 mm long and have a strong platy L-S tectonite fabric. In places, especially near the exposed top of the section, the rock is coarser grained with hornblende crystals up to 5 mm across.

Pelites, converted to dark grey, garnet-biotite schists, occur as thin, rare, interbands from 5 to 25 cm thick in the hornblende schists. The garnet porphyroblasts are generally 1-3 mm in diameter but at one locality on the north side of Easter Tickle they occur up to 12 mm in diameter. Hornblende is generally only poorly developed within the pelites and biotite best defines the schistosity (Plate 29, fig. b).

Impure limestone beds up to 0.6 m thick are conspicuous, but rare, in the amphibolite member. They are exposed near the top of the member at Easter Tickle where they are interlayered and complexly folded with hornblende schists (Plate 30, fig.a). The limestones are converted to white to grey to bluish grey, thinly banded, foliated, marbles that contain rare porphyroblasts of garnet.

The narrow strip of amphibolite member exposed on the west side of Landing Cove includes a lens of metagabbro up to 8 m thick. The gabbro structurally overlies the hornblende schists with sharp contact to the west and to the east is cut out by a fault that repeats the greenschist member.

The gabbro has a distinct compositional banding composed of light, feldspar rich and dark, amphibole rich bands from 2 to 15 cms thick (Plate 30, fig. b). The banding is thought to be a primary igneous layering A strong S_1 tectonic fabric, defined by actinolite crystals up to 3 mm long is developed parallel or subparallel to the primary gabbro layering. The schistosity is folded by tight F_2 folds that have an associated strain-slip fabric. Small areas of meta-gabbro were also discovered in reconnaissance work near the top of the amphibolite member at the southwest corner of the western White Hills, north of Hare Bay (Fig. 2).

Petrography:

Hornblende Schist - The hornblende schists are composed dominantly of hornblende (75 - 90%) and plagioclase (15%) with minor epidote and quartz and accessory sphene and magnetite. The hornblende is strongly pleochroic in X, light brown; Y, olive green; and Z, emerald green. It is subidiomorphic and shows a moderate to strong preferred orientation. The hornblende commonly contains small rounded inclusions of plagioclase and more rarely of magnetite. The hornblende is commonly altered along thin cracks to penninite.

Plagioclase (An_{33-40}), generally heavily saussuritized, occurs as scattered xenomorphic grains, 0.1 - 0.3 mm across, between hornblende crystals. Albite and albite-pericline twinning is preserved in the less altered varieties.

Pelite - The pelites are composed dominantly of plagioclase 50%, biotite 25%, with minor quartz 5-15%, garnet 5% and hornblende 0-10%; and accessory magnetite, epidote and apatite.

Plagioclase (An_{33}) occurs as porphyroblasts up to ~2 mm across that contain rare inclusions of biotite and rounded quartz. Most of the porphyroblasts are untwinned but the finer grained plagioclase in the groundmass commonly shows albite twinning.

Brown biotite occurs as small laths 0.2-0.4 mm long. They show a strong preferred orientation that forms augen around the plagioclase and garnet porphyroblasts. Quartz is present as small grains 0.1 - 0.2 mm long and as stringers up to 7 mm long. The quartz shows undulose extinction and a weakly developed dimensional orientation. In the stringers the quartz is partially polygonized with sutured grain boundaries.

Garnet is ubiquitous in the meta-pelite bands and generally forms well developed porphyroblasts from 0.5 - 3 mm in diameter. They contain inclusion trails of magnetite and quartz and, more rarely, of biotite, that are arranged in circular and straight forms.

Hornblende is poorly developed or absent in the pelites. It occurs as subidiomorphic crystals up to 1 mm long, scattered throughout the rock, and in discrete, 2 - 3 mm thick hornblende rich bands. The hornblende shows a strong preferred orientation, interpreted as S_1, that is folded by F_2 folds with associated further hornblende growth.

Magnetite occurs as small rods and equant grains up to 0.1 mm long. The rods are aligned in S_1 and define straight inclusion trails within the garnet porphyroblasts.

Idiomorphic apatite, 0 - 0.5 ~~mm~~ across, and xenomorphic epidote, up to 0.5 mm across, occur in accessory amounts. Local chloritization and sericitization partially alter the pelites.

Marble - The marble bands are composed of diopside, carbonate, sericite, epidotic-carbonate and minor quartz and garnet. The diopside is colourless to light green and occurs as subidiomorphic crystals in monomineralic masses, 5-7 mm wide. The diopside is partially altered, especially at the margins of the bands, to colourless tremolite.

Garnet is rare and occurs as poorly developed, pale pink, porphyroblasts up to 2 mm across. The groundmass consists of sericite, brown epidotic-carbonate, and minor chlorite, that show a strong preferred orientation and define the S_1 foliation of the rock.

Gabbro - The meta-gabbro is composed of pale green amphibole (40 - 80%), plagioclase, and accessory sphene. The amphibole is pleochroic in X, colourless; Y, light to olive green; Z, pale green, has an extinction angle Z: c of 18^{o}, and is probably actinolite. The actinolite is subidiomorphic, up to 2 mm in grain size, and shows a strong preferred orientation that defines the S_1 schistosity. The plagioclase is completely saussuritized to a brown epidotic-carbonate mixture, and the hornblende shows partial alteration to chlorite.

Contact Zone

The contact between the Goose Cove Formation and the overlying White Hills Peridotite Sheet is not exposed in thesis area but is believed to lie "beneath the sea not far east of Fishot", (Cooper, 1937).

A high magnetic anomaly in this general area (Fig. 6) suggestive of an ultramafic body, adds to Cooper's postulation.

The contact below the White Hills Peridotite Sheet, north of Hare Bay was briefly studied on four separate traverses. The lithology and fabric of the Goose Cove Formation in the White Hills is similar to that on the Fishot Islands and a greenschist and amphibolite member can be mapped. However, a petrographical subdivision of the amphibolite member is possible and a "contact zone" is defined on the basis of a different metamorphic mineral assemblage. The "contact zone" occurs directly below the White Hills Peridotite Sheet and is characterized by the presence of brown hornblende and the appearance of augite. The zone varies in thickness from 10 to 27 m and the transition outwards from brown hornblende to green hornblende, typical of the amphibolite member, occurs over a few meters. The actual contact between the peridotite and schist was nowhere observed but can usually be defined over an exposure gap of a meter or two.

Lithology and Fabric:

Dark hornblende schists and meta-gabbro are the lithologies of the contact zone. The hornblende schists are texturally similar to those in the amphibolite member apart from a slightly coarser grain size. Hornblende crystals from 2 - 4 mm long define a strong L-S tectonite fabric. The meta-gabbros possess a strong foliation defined by segregation of amphibole and feldspar into discontinuous layers 1 - 3 mm thick. Within a few meters from the contact the schists possess a strong schistosity in which knots and augen of hornblende or hornblende and augite stand out in a light, feldspar rich matrix that wraps around them (Plate 31, fig. a).

The foliation of the schists in the contact zone is parallel to the contact with the White Hills Peridotite Sheet and also to the axial plane of the F_2 recumbent folds well developed in the underlying amphibolites. No recumbent folds were seen in the contact zone but the fabric is considered to be a composite S_2 schistosity.

Petrography:

The contact zone is characterized by the coexistence of brown hornblende (35-40%) and augite (20%), and the presence of plagioclase (An_{40}). The augite is colourless or pale green and non-pleochroic. In the metagabbro it occurs in the amphibole rich bands and in the hornblende schists it is scattered throughout the rock and exists side by side with hornblende. The grains are subidiomorphic, up to 2 mm long, but locally form porphyroblasts up to 4 mm across.

Hornblende occurs as discrete grains and as partial replacements and inclusions in pyroxene crystals. The hornblende is pleochroic in X, pale brown; Y, Z brown to reddish brown, but in one section of metagabbro the pleochroic scheme is X, colourless; Y, Z light brown. In the strongly banded metagabbro, hornblende forms eye shaped porphyroblasts 6 mm long that are augened by the S_2 schistosity.

Plagioclase forms xenomorphic grains up to 2 mm across that are commonly partially, or completely altered, to saussurite. A fine polysynthetic twinning is present in most grains. In the feldspar rich bands in the metagabbro the plagioclase forms a granoblastic type texture with a slight preferred dimensional orientation. The grains are from 0.1 - 0.3 mm long and show undulose extinction.

Apatite is a common acessory in the ferro-magnesian rich bands in the metagabbro. It forms xenomorphic grains 0.3 mm across and often occurs as inclusions in hornblende. Sphene and iron ore are less commonly present.

Most sections of the contact zone show some degree of retrogressive metamorphism. The first stage, a marginal alteration of pyroxene to brown hornblende, is only poorly developed. The second stage affected both pyroxene and hornblende alike and they are retrogressively altered to a yellow, fibrous amphibole or more commonly to colourless to light green, tremolite-actinolite. In more extreme cases chlorite is also developed. Feldspar alteration also accompanied this alteration stage, which post-dates the S_2 foliation.

Thin prehnite veins cross-cut and clearly post-date all of the previous metamorphic events.

Age and Provenance

The age and provenance of the rocks of the Goose Cove Formation is uncertain. The greywackes at the base of the formation on Fishot Island are similar to parts of the Maiden Point Formation and must have been derived from the Grenville basement to the west. The volcanic rocks may represent a volcanic rich facies of the Maiden Point Formation similar to the Maiden Point north of Hare Bay around St. Lunaire (Fig. 2) (Williams et al., 1973).

The Goose Cove Formation has been correlated (Smyth, 1971) with part of the eastern sequence (Kennedy, 1971) of the pre-Arenig Fleur de Lys Supergroup. All of these tenuous lines of evidence suggest a Cambrian age for the Goose Cove Formation.

CHAPTER IV

MÉLANGE ZONES

Mélange zones consisting of a chaotic mass of blocks of various shapes, sizes and attitudes set in a black, or sometimes black and green shaley matrix (see Plate 31, fig. b) occur at the soles of the various thrust slices in the north of the area but are absent in the south of the area at Canada Bay where the contact below the Maiden Point Slice is a sharp thrust. The location of the various mélange zones is shown diagrammatically in Fig. 16.

Basal Mélange to the Maiden Point Slice

A mélange separates the Maiden Point Slice from the autochthonous Goose Tickle Formation in the north of the area (see Fig. 3) but is absent in the Canada Bay area.

At Big Springs Inlet the Goose Tickle Formation grades upwards over a distance of about 9 m from unbroken beds through broken up beds composed of detached and rotated blocks and overlain by mélange consisting of blocks derived from both the Maiden Point and Goose Tickle Formation. Similarly the base of the Maiden Point Slice is broken up and injected with shale and shows a gradational contact with the mélange. The mélange at Whites Arm also has gradational upper and lower contacts.

Two mélange zones occur within the Maiden Point Slice at Deep Bay in Hare Bay, and at Little Canada Harbour. The mélange zones are not continuous, and do not define separate slices. Three possible explanations may account for their presence.

I. They represent the basal mélange of the Maiden Point Slice exposed in eroded anticlinal cores.

Structural evidence to prove that these areas are major anticlines was not found. The presence of basal mélange at Little Canada Harbour would be anomalous as only 1.6 km to the west at Canada Harbour there is no mélange at the base of the Maiden Point Slice.

II. They are tectonically-disturbed, brittle, parts of the Maiden Point Formation.

Both of these mélange zones contain in addition to the usual black shaley matrix and blocks of Maiden Point greywacke, blocks of brown weathered limy siltstone. The Little Canada Harbour occurrence also contains a pre-tectonic dyke or block of diabase. Limy siltstone was not found outside the mélange zones and would appear to be an exotic lithology to the Maiden Point Formation, which would negate this possibility.

III. They represent early, discontinuous, zones of imbrication within the Maiden Point Slice that contain mélange.

This possibility circumvents the lack of structural evidence necessary for the first hypothesis and also might account for the presence of the limy siltstone blocks. Such mélange zones may not have formed at the base of the slice or may have been removed by emplacement, which would account for their absence from the present leading edge of the slice at Canada Harbour and Englee.

Basal Mélange to the Croque Head Slice

Basal mélange to the Croque Head Slice is well exposed at the localities listed in Fig. 16. It varies from 12 to 45 m in thickness and attains maximum development at St. Juliens. The lower contact is gradational into broken beds of greywacke but the upper contact is sharply defined (Plate 32, fig. a).

The mélange consists of clasts of Maiden Point greywacke, brown weathered limy siltstone and a single clast of altered diabase occurs south of Cobblers Cove. Murray (1861) noted chalcopyrite and siderite mineralization in a quartz veined sandstone at St. Juliens. The sandstone occurs as a block 18 m long in the mélange and is of no economic significance.

Basal Mélange to the St. Julien Island Slice

The basal mélange to the St. Julien Island Slice outcrops along the west and south side of the island (See Fig. 5). It overlies the Croque Head Slice with sharp contact,and the upper contact with the Irish Limestone varies from sharp to gradational over a few meters. Locally the basal limestone beds are brecciated and clasts of limestone and brown stained quartzite are incorporated in the uppermost part of the mélange. The majority of the clasts, however, are of Maiden Point greywacke.

Basal Mélange to the White Hills Slice

The basal mélange to the White Hills Slice is excellently exposed across a 60 m wide wave cut platform that parallels the western margin of Fishot Island. The lower contact of the mélange is a normal fault with the Maiden Point Formation which is exposed on a small island at the southwest tip of Fishot (Fig. 4). The upper contact with the Goose Cove Formation is sharp and is defined in several localities. The mélange is approximately 45 m thick with base unexposed.

The clasts consist almost entirely of greywacke and minor quartz-pebble conglomerate derived from the Maiden Point Formation. A single altered microgabbro boulder, similar to the coarser parts of the Maiden Point volcanics was found. A pre-tectonic diabase block or dyke, now cut up into small imbricate slices(Plate 32, fig. b) occurs in the mélange near the contact with the Maiden Point Formation. Post-tectonic dykes cut the mélange and the mélange-White Hills Slice contact in several places (Plate 33, fig. a).

Lithology and Thickness

The various mélange zones vary in thickness from a few meters to over 40 meters at Fishot Island. They are all lithologically similar, in that they contain isolated clasts (10-30%) in a shaly matrix. Practically all of the clasts consist of Maiden Point greywackes and it should be noted that rocks of the Maiden Point Formation form at least one of the bounding formations to all of the mélange zones. Other clasts are rare and consist of brown weathered limy siltstone, microgabbro and diabase. Limy siltstone was not seen in any of the bounding allochthonous or autochthonous units in the map area and appears to be an indigenous lithology of the mélange zones.

The matrix consists of black shale that in most cases has been converted to a slate or phyllite (Plate 33, fig. b). At Fishot Island, apple green slates are locally intermixed with black slate and at Irish Bay light green siltstone is incorporated in the matrix. Secondary nodules and euhedral crystals of pyrite are common in the black slates.

The clasts show a variation in size from pebbles to boulders to blocks up to 30 m long (Plate 34, fig. a). However, boulders from 0.3 to 0.6 m across are the most common size. There is no internal grading and clasts of all sizes occur chaotically intermixed. The boulder and pebble size clasts vary from angular to rounded and are generally oval shaped (Plate 34, fig. b). The larger blocks tend to be angular and tabular shaped and lie with their long axes parallel to the first cleavage (S_1) in the rock. The brown weathered limy siltstone clasts commonly form slabs up to 3.6 m long and up to 1 m thick. The slabs are composed of thin beds up to 15 cms thick that show internal grading, lamination and cross bedding. At St. Julien's siltstone slabs can be traced into broken-up, angular, boulders.

Fabric

In most places the mélange possesses an early tectonic fabric (S_1), that is arranged parallel to its upper and lower contacts, and is folded by open upright folds (F_2) with an associated weak crenulation cleavage (S_2) (Plate 35, fig. a). The intensity of the early fabric (S_1) varies from west to east across the area and is most intense below the White Hills Slice at Fishot Island. However, the later folds (F_2) are also tightest in the east and they may have accentuated the S_1 fabric (See Chapter VI)

At Big Springs Inlet the matrix possesses a fissility that forms augen around the clasts. Eastward, at Whites Arm in the same mélange zone below the Maiden Point Slice, the S_1 fabric is defined by flattened clasts and a slaty cleavage in the matrix. The S_1 fabric is intense at Fishot Island where quartz pebbles are completely flattened into thin veinlets. At Fishot Island and Little Canada Harbour minor flat lying F_1 folds are locally associated with this fabric (Plate 35, fig. b).

Discussion

The origin of mélange zones is especially enigmatic. Bruckner (1966), impressed by the absence of primary deformational structures in mélanges in the Humber Arm Allochthon, concluded that they must have had an original chaotic structure before they were overridden and deformed by a thrust slice. He proposed a regolith produced by mass wasting as the cause of the original chaos.

Stevens (1970) concluded that there are two types of mélange, namely, a sedimentary olistostrome developed in front of the allochthon, and a tectonic mélange developed between the thrust slices, but gave no criteria by which this distinction is made. The only apparent difference between the two types is in the presence of a tectonic fabric in the tectonic melange, which appears to be related to overriding by the thrust slices. Both types probably had a similar origin. Hsü (1968) concluded that it is impossible to distinguish between a deformed olistostrome and a tectonic mélange.

Any theory on the origin of mélange between the allochthonous thrust slices must explain the following points:

1. the source of the indigenous black shale that forms up to 80% of the melange.
2. the source of the clasts.
3. the general uniformity of all the melange zones regardless of structural position.

The tectonic base of a slice does not always occur at the same stratigraphic level along its length and breadth, for example the White Hills Slice. It follows, therefore, that the base of a slice bears no genetic relationship to the mélange it overlies. Rather, a mélange represents an independent lithology over which a slice traversed during emplacement.

There are two formations in the Hare Bay area that may have been a source of the black shale and may have been lain in an area that was traversed by the thrust slices during emplacement:

1. The uppermost part of the autochthon, the Goose Tickle Formation.

2. The black shales of the allochthonous Northwest Arm Formation (Cooper, 1937) that constitute the basal slice of the complete Hare Bay Allochthon (Williams et al., 1973). The Northwest Arm Formation is exposed north of the map area at the head of Hare Bay (Fig. 2) where it is structurally overlain by both the Maiden Point Slice and the White Hills Slice.

The Goose Tickle Formation is an unlikely source as it does not contain limy siltstone or green shale and also it underlies the entire Hare Bay Allochthon, even where mélange is absent, for example, below the Maiden Point Slice at Canada Bay.

The Northwest Arm Formation, on the other hand, contains black shale, minor green shale and buff weathered limy siltstone (Williams et al., 1973). The formation crops out close to the northern extremity of the map area at Hare Bay (Fig. 2) where mélange is best developed, but does not occur southwards where mélange is absent. The development of mélange appears, therefore, to be closely linked to the spatial distribution of the Northwest Arm Formation.

It is suggested that equivalents of the Northwest Arm Formation are the source of the indigenous lithologies of the mélange zones. The exotic blocks of greywacke, microgabbro, and diabase were derived from the thrust slices during emplacement. They were either shed by gravity sliding or were tectonically incorporated from the advancing slices. The structural evidence (see Chapter VI) suggests that the mélange zones formed early in the tectonic evolution of the allochthon and this is discussed further in Chapter VIII.

CHAPTER V

INTRUSIVE IGNEOUS ROCKS

PRE-EMPLACEMENT INTRUSIONS

AUTOCHTHON

MICROGRANITE AND GRANITE

Pink microgranite (senso lato) post-tectonically intrudes the gneisses of the Long Range Complex at Seans Cove (see Plate 1, fig. a) and both gneiss and granite are unconformably overlain by Bradore sandstones. The granite is therefore of post-Grenville and of pre-Cambrian age. A similar pink microgranite also occurs in thrust contact with the Bradore sandstones (Plate 36, fig. a) and nearby it intrudes a structurally overlying dark grey, medium grained, highly altered, schistose granodioritic rock. The granodiorite may represent either an earlier more calcic intrusion or a highly retrogressed gneiss.

To the northeast post-tectonic pink, granite dykes and pegmatites cut the Acid and Basic Gneisses and a highly deformed pink granite constitutes the lower part of the structurally overlying Sugarloaf Slide Zone.

All of these pink granites show similar relationships. namely, either intrusive into, or thrust over, basement gneisses. The granites are deformed, apparently in relation to the extensive thrusting developed in this area. The microgranite thrust over the Bradore sandstones contains thin (1 - 2 mm), closely spaced (2 - 20 mm) anastomosing dark grey crush zones (Plate 36, fig. b) that give the rock a brecciated appearance. The granite dykes in the Acid and Basic Gneisses are folded, cleaved and boudinized (Plate 37, fig. a). The granite in the Sugarloaf Slide Zone is the most intensely deformed and is converted to augen-schist and phyllonite.

Petrography

The microgranite is composed of a fine grained (0.1 - 0.5 mm) microcrystalline, granular, mosaic of quartz and K-feldspar with rare microphenocrysts up to 2 mm across of quartz, plagioclase and K-feldspar. Graphic intergrowth of quartz and microcline is sometimes developed. Accessories include opaques and muscovite.

The granite dykes in the Acid and Basic Gneisses consist of plagioclase (An_{5-15}) 55 , quartz 35 and K-feldspar 10 . Mafic minerals (1) consist of biotite, commonly chloritized, and muscovite, and accessories include sphene and zircon. In the granites of the Sugarloaf Slide Zone, K-feldspar and plagioclase (An_{5-15}) are in approximate equal proportions. The plagioclase is commonly sericitized in the more deformed parts of the slide.

The composition of the granites ranges from alkali granite to granodiorite and although the various bodies are separated from each other by major thrusts they appear to be post-tectonic intrusives of the same age.

PORPHYRITIC MICROGRANITE

A porphyritic microgranite (senso lato) occurs at the base of the narrow thrust block of the lower clastic-volcanic unit of the autochthon west of Burnt Point in Canada Bay (see Fig. 3). A major thrust separates the base of the porphyry from the underlying Goose Tickle Formation, and the upper contact is also a thrust that separates it from either conglomerate or volcanic rocks of the Lighthouse Cove Formation. The porphyry is strongly deformed (see Plate 37, fig. b) and the deformation is probably related to the thrusting.

Two varieties of porphyry, in fault contact with each other, are present:

(a) pink porphyritic alkali-microgranite (UPPER).

(b) black to green, porphyritic microgranodiorite (LOWER).

The dark green variety consists of phenocrysts of oligoclase (60%), and quartz (30%) in a groundmass of biotite, quartz and plagioclase. Oligoclase forms subidiomorphic phenocrysts up to 6 mm across. The dark green porphyry is highly deformed near the contact with the basal thrust. There the plagioclase phenocrysts are broken and kinked and commonly completely altered to sericite and minor epidote. The quartz is strained and broken down to sub-grains that show a dimensional preferred orientation and the biotite is chloritized.

A dark green, quartz-sericite-chlorite-epidote schist outcrops in the cliffs structurally above the dark green porphyry and may represent a more highly deformed part of the porphyry.

The pink variety structurally overlies the green porphyry and can be traced for 0.8 km south of the coast. It consists of xenomorphic phenocrysts of quartz and K-feldspar up to 5 mm across, in a fine-grained

(0.1 - 0.5 mm) schistose groundmass of quartz, K-feldspar and aligned sericite. The quartz phenocrysts show undulose extinction with sutured sub-grain boundaries.

The pink porphyritic microgranite near Burnt Point has certain petrographic similarities ~~with~~ the microgranite at Seans Cove and both occur at a similar structural level, namely in thrust contact with the autochthonous basal clastic-volcanic unit, and both are probably related.

LATE PRECAMBRIAN TO EARLY CAMBRIAN DIABASE DYKES

Post-tectonic diabase dykes intrude the Long Range Complex at Canada Bay. They have been described by Strong and Williams (1972) and are not included in this study.

MIDDLE ORDOVICIAN (?) MAFIC DYKES AND SILLS

Rare pre-tectonic dykes and sills cut the Table Head and the Goose Tickle Formations at Canada Harbour, Englee, Whites Arm and southwest of Indre Point in Hare Bay. They are deformed by the middle Ordovician emplacement deformation (D_E), which puts an upper age limit on the intrusions. At Canada Bay the dykes are commonly broken up into boudins (Plate 38, fig. a).

The intrusions vary from fine-grained, altered, greenish, diabase dykes, up to 0.6 m thick, at Canada Bay and at the northwest side of Whites Arm, to a 9 m wide gabbro at Indre Point. The outer margin of the gabbro is coarsely crystalline with plagioclase laths up to 2.5 cm long.

In thin section it consists of interlocking laths of albite (80%) with interstitial chlorite (15%), accessory opaques and late sheaves of brown stilpnomelane. Cooper (1937) incorrectly identified the rock as extrusive as he interpreted the chilled margin as an ash band.

ALLOCHTHON

ST. JULIEN ISLAND GABBRO

Small areas of pre-tectonic gabbro intrude the St. Julien Island Formation on St. Julien Island and at Black Island (Fig. 5). The gabbro bodies have fine grained chilled margins, and medium-grained central parts that locally include coarse porphyritic patches. The chilled margins are up to 1.8 m across and have been converted to dark greenschists by the deformations that affected the St. Julien Island Formation. The microgabbro shows an xenomorphic granular texture with average grain diameter between 1.5 and 2.5 mm. The igneous texture is well-preserved but, locally, the deformation has produced a sub-schistose fabric in the rock. The porphyritic patches are up to 1.2 m wide and contain subidiomorphic plagioclase and augite crystals up to 3 cm long that are fractured and kinked by the deformation.

The greenschist areas consist predominantly of chlorite with minor carbonate, albite, epidote and magnetite. These minerals define a weak schistosity (S_1) that is folded about the upright F_2 folds. The gabbro is composed of plagioclase (An_{10-15}) and augite, in approximate equal amounts, and minor hornblende and magnetite. The plagioclase is unzoned, twinned according to both albite and pericline laws and is commonly extensively saussuritized. Augite forms subidiomorphic crystals with irregular small inclusions and discontinuous rims of brown hornblende. Both the hornblende and augite are kinked and folded together and both show secondary alteration to fibrous tremolite. Magnetite, commonly with rims of hornblende, is intergrown with augite in skeletal forms.

Both the plagioclase (An_{10-15}) and hornblende appear to be primary and their unusual occurrence can be explained by a high water content of the ~~original~~ magma (D.F. Strong, pers. comm.).

POST-EMPLACEMENT INTRUSIONS

DIABASE DYKES

Post-tectonic intrusions with respect to the regional Acadian deformation (D_A) are sparse and consist only of dykes. Approximately forty dykes were noted in the area which cut all of the slices of the allochthon, the mélange zones, and the upper part of the autochthon. At Sugarloaf Head a late dyke in the Maiden Point Formation cross cuts the Acadian structures but is itself folded and cleaved by the late local folding developed in this area.

The dykes generally trend in a northeast direction and range in thickness from 0.6 to 6 m. The smaller dykes are fine-grained but the wider dykes have medium to coarse-grained central parts. The texture varies from holocrystalline to porphyritic to subophitic. The dykes are slightly to moderately altered but igneous textures are everywhere preserved. Less altered varieties consist of plagioclase An_{40} (80%), augite (20%), and accessory quartz, sphene and opaques. The altered dykes consist of albite sometimes completely sericitized, fibrous amphibole and chlorite, probably after pyroxene, and variable combinations of secondary sericite, actinolite, epidote and carbonate.

Major element analyses were made on 21 dykes and all show broadly similar compositions (Fig. 17). Their compositions are similar to the pre-tectonic Maiden Point Volcanics interpreted to have formed in an environment of continental distension (see Chapter 111).The post-tectonic dykes may also have formed in a similar environment.

CHAPTER VI

STRUCTURE AND DEFORMATION

The different tectonic slices of the Hare Bay Allochthon not only contrast in lithology but in the intensity and number of phases of deformation they suffered, for example, the White Hills Slice contains poly-deformed amphibolites whilst the Maiden Point Slice contains mildly-deformed basalts. The earliest deformations in each slice either pre-dated or formed contemporaneously with emplacement. After middle Ordovician emplacement, the entire area was affected by a major period of folding. These folds affect the underlying autochthon, the mélange zones, the allochthon-autochthon contact, and refold the earlier folds and fabrics in the allochthonous rocks. This post-emplacement deformation has been assigned to the Acadian (Devonian) orogeny (Smyth, 1971).

By careful examination of the contacts between the various slices and their basal mélange zones it has been possible to demonstrate that the pre-Acadian deformations were produced either during emplacement or else prior to emplacement. Syn-emplacement structures are characteristic of the Maiden Point, Croque Head, St. Julien Island, Slices and also of the mélange zones and the upper part of the autochthon. Pre-emplacement structures are truncated by the mélange zones and are developed only in the upper slice, the White Hills Slice.

The pre-, syn- and post- emplacement deformations are labelled D_P, D_E, and D_A respectively. As the style and intensity of development of structures related to these phases of deformation varies from slice to slice, each deformation event is described separately for each slice. The structural evolution of the various slices of the allochthon is summarized in Fig. 18.

PRE-EMPLACEMENT DEFORMATION

Introduction

Pre-emplacement deformations could only be proven in the Goose Cove Formation, which comprises the basal part of the White Hills Slice. The evidence for pre-emplacement deformations in these rocks is:

(1) Psammites with a composite second phase schistosity (S_2) are truncated by black and green shales of a mélange along the south-west side of Fishot Island.

(2) Schistose amphibolites overlie unmetamorphosed mélange west of Lock's Cove, north of Hare Bay.

(3) Boulders of polydeformed Goose Cove schists are incorporated in mélange along the north shore of Hare Bay (Williams et al., 1972).

(4) Schistose amphibolites overlie much less deformed mélange south of Croque Head.

Two phases of pre-emplacement deformation have been recognized and are labelled D_{P1} and D_{P2}.

D_{P1}

The first deformation in the Goose Cove Formation produced a strong schistosity (S_1) and minor tectonic slides (Fleuty, 1964 b). No megascopic or microscopic folds were observed related to this deformation.

The S_1 fabric is well preserved in the metavolcanic rocks of the greenschist member and is defined by a preferred orientation of chlorite, tremolite-actinolite, and flattened plagioclase phenocrysts. In the semi-pelites and psammites the S_1 muscovite-chlorite fabric is crenulated by the S_2 schistosity (Plate 38, fig. b) and on the fold limbs S_1 is transposed into the S_2 planes.

S_1 in the amphibolites is defined by aligned hornblende, biotite, and rods of opaques. It is strongly overprinted by the S_2 fabric and S_1 is only recognizable in the hinges of the F_2 folds between the S_2 planes, and as an included fabric in post-D_{P1} garnet porphyroblasts (Plate 39, fig. a).

In the slide zone at the southwest tip of the island, S_1 consists of a strong, streaky, foliation.

D_{P2}

The second deformation (D_{P2}) in the Goose Cove Formation produced inclined to recumbent folds (see Plate 39, fig. b) with an associated axial plane crenulation cleavage in the greenschist member and a schistosity in the amphibolite member. F_2 folds are common in the greenschists but are difficult to recognize in the overlying monotonous dark amphibolites except where interbedded contrasting lithologies occur, (for example, marble beds north and south of Easter Tickle)(see Plate 30, fig. b).

The folds are close to sub-isoclinal (Fleuty, 1964a) (interlimb angle 5-35^{o}), similar folds that trend northeast to north and plunge gently northeastwards. The present dip of the axial plane of the folds is controlled by the post-emplacement (D_A) structures. The folds are asymmetrical and exhibit a constant sense of vergence that indicates an antiform upwards to the west. Facing directions were determined from the psammites at the base of the formation where the folds face upwards to the west on the S_2 cleavage.

S_2 in the greenschists consists of a closely spaced (1-3 mm), kink-style, crenulation cleavage with little or no associated metamorphic growth except in the semi-pelites where S_2 is defined by chlorite and sericite. In the amphibolites S_2 becomes the dominant fabric element by complete transposition of S_1 into the S_2 planes and accompanying growth of aligned biotite and hornblende. The D_{P2} structures in the amphibolites are separated from the S_1 fabric by a period of static growth of garnet and plagioclase porphyroblasts.

Conclusions

The pre-emplacement deformation falls into two distinct phases separated by a period of static mineral growth. The first phase is a flattening type of deformation characterized by a strong S_1 schistosity that is most intense in the slide zone. The second phase was a period of recumbent folding with concomitant aligned mineral growth in the amphibolites.

D_E SYN-EMPLACEMENT DEFORMATION

The major elements of syn-emplacement deformation (D_E) are the detachment thrusts that bottom and define the various tectonic slices of the allochthon. These basal thrusts are generally marked by mélange zones which form distinct map units and have been described in Chapter IV. Internal deformation within the Maiden Point, Croque Head, and St. Julien Island Slices produced a variety of structures that are generally concordant with the basal thrusts. The emplacement deformation also affected the mélange zones and the upper part of the allochthon. At Canada Bay, where mélange is absent, an imbricate structure is developed in the autochthonous rocks. All of these deformation structures are believed to have accompanied slice emplacement (D_E) and can be demonstrated to have preceded the post-emplacement Acadian event (D_A).

AUTOCHTHON

Introduction

The D_E deformation produced the first tectonic structures recognized in the autochthon, namely, a widespread slaty cleavage in its uppermost part, and minor recumbent folds and an imbricate structure in the Canada Bay area.

Imbricate Structure

A number of imbricate thrust faults are developed in the autochthonous rocks in the Canada Bay area below the Maiden Point Slice. The imbricate structure is best exposed along the south shore of Canada

Bay where five thrusts repeat the autochthonous succession (See Fig. 3, section E-F). Southwards, the thrusts appear to be truncated by the Wild Cove Fault and only two thrusts appear at the coast of White Bay. Minor thrusts are exposed across the north shore of Canada Bay at Englee Island, Baard Island, Bide Head, and Dieppe Point.

The uppermost and most powerful thrust of the imbricate structure is, of course, the emplacement thrust of the Maiden Point Slice. Within the autochthon, the thrust with the largest displacement occurs 0.4 km west of Burnt Point and superposes the Lighthouse Cove Formation (Cambrian) on the Goose Tickle Formation (Middle Ordovician). A highly sheared porphyry occurs at the base of this thrust sheet (See Plate 37, fig. b). An escarpment and a strong lineation on the air photographs mark the trace of this thrust southwards where it is truncated by the Wild Cove Fault. Above this thrust, 91 m to the east, a second major thrust is developed, which brings the Goose Tickle Formation over the Devils Cove Formation. The Devils Cove Formation and the directly underlying Bradore Formation are mylonitized up to 6 m below the fault (Plate 40, fig. a). A possible continuation of this thrust is exposed south of Sugarloaf Cove in White Bay where the Sugarloaf Schists are thrust over the Acid and Basic Gneiss. The Acid and Basic Gneiss probably forms a thrust block of basement gneiss within the Sugarloaf Schists. The basal contact is cut out by the Wild Cove Fault but the upper contact is well-exposed and is marked by a 53 m thick zone of mylonitized rocks referred to as the Sugarloaf Slide Zone. The tectonic fabric of the Slide Zone is described later.

The displacements on the other thrusts in the autochthon are much smaller; for example the St. George Formation rests on Goose Tickle Formation east of Wild Cove; the Table Head Formation rests on the Goose Tickle Formation west of Canada Harbour, at Baard Island, and at Englee Island (the Englee Island Thrust of Betz, 1939); and the St. George Formation is superposed on the Table Head Formation at Bide Head (the Handy Harbour thrust of Betz, 1939) and at the west side of Handy Harbour. The thrust planes dip moderately eastwards but are locally folded by the D_A Acadian event. The rocks at the base of the minor thrusts are locally sheared and generally marked by thin (0.6 m) zones of mylonitized rock. Deformed pebbles in Bradore sandstones at the east shore of Wild Cove and west of Burnt Point record a northwesterly stretching lineation that probably reflects the direction of movement of the D_E minor thrusts. Locally, minor recumbent folds are developed below the thrust planes and are described in the next section.

Minor Structures

Recumbent to inclined D_E folds (F_1) are common in the autochthon at Canada Bay but in the north of the area are rare and were only seen at the south shore of Whites Arm and at the west end of Coles Pond. At Canada Bay the folds are most common below thrust faults, and thrusting and folding was a related event.

The F_1 folds are best developed in slates and minor limestones of the Goose Tickle Formation, at the west side of Baard Island

(Plate 40, fig. b) and at Englee Island, below overthrust sheets of the Table Head Formation. Recumbent folds are also developed at the west side of the Bide Head in grey limestone and minor dolostone of the Table Head Formation (Plate 41, fig. 2) below an overthrust sheet of the St. George Formation.

These F_1 folds are subisoclinal, east-trending, gently-plunging, similar, folds. At Bide Head, the folds have an amplitude of 1.5 - 3 m with highly-attenuated limbs and thickened hinge zones. The more rigid dolostone beds are commonly drawn apart to form isolated, oval-shaped, boudins whose axes are parallel to the F_1 folds.

Other occurrences of F_1 folds that are not obviously related to thrusting occur in the Table Head Formation at Burnt Point, in the basal black limestones of the St. George Formation along the east side of Chimney Arm, and in the Forteau Formation at Fly Point. The folds in the Forteau Formation are close, moderately-inclined, similar, folds that face up to the northwest, and are more open (interlimb angle 50^0) and of smaller amplitude than those developed in the overlying formations (Plate 41, fig. b).

In the north of the area at the south shore of Whites Arm, a single recumbent fold (F_1) was seen in the Goose Tickle Formation. This fold is folded about upright D_A folds that are locally downward-facing (Shackleton, 1959), further indicating the presence of syn-emplacement recumbent folds in this area.

Other minor structures include boudins, quartz rods and quartz veins. Rare, pre-tectonic, diabase dykes intruded into the Table Head

Formation at the east side of Canada Harbour, Baard Island, and at Englee Harbour are disrupted into boudins with rectangular shaped cross sections (see Plate 38, fig. a). The boudins lie in the plane of the D_E S_1 cleavage and their long axes are aligned parallel to the F_1 fold axes. Veins, rods and boudins of quartz and calcite are locally developed in the Goose Tickle Formation at Canada Bay. The S_1 cleavage locally forms augen around the veins that probably formed synchronously with F_E. These minor structures are all folded by the upright D_A folds.

Fabric

The D_E deformation is represented in most places in the autochthon by a flat lying slaty cleavage (S_1) that is an axial plane fabric to the rare F_1 recumbent folds. S_1 is most intense and most extensively developed in the Goose Tickle Formation and the Sugarloaf Schists and locally near the thrust faults of the imbricate structure. S_1 shows a general decrease in intensity with depth and distance away from the allochthon contact. At Canada Bay the development of S_1 extends down to the basal Bradore Formation but in Whites Arm Window no sign of S_1 was found in the Table Head Formation, and in this area it appears that the effects of D_E die out rapidly with depth. West of the present leading edge of the allochthon at Big Springs Inlet, S_1 is weak and only locally developed in the Goose Tickle Formation.

In the slates of the Goose Tickle Formation S_1 is defined by aligned sericite, minor chlorite and trains of opaques. In the greywackes S_1 consists of anastomosing films of sericite that are locally arranged parallel and at Canada Head, east of Englee and at the south end of Whites Arm Window, the micas are associated with a dimensional preferred orientation of the detrital material (Plate 42, figs. a and b). However, in all of these localities, the D_A, F_2 folds are tight and may have caused a morphological change in the S_1 fabric.

A syn-emplacement (S_1) schistosity is well developed in the Sugarloaf Schists and is defined by aligned biotite in the pelites and semi-pelites, by tremolite in the calc-silicates, and by biotite and tremolite-actinolite in the mafic schists. The S_1 fabric is preserved in the hinge regions of the F_2 folds between the S_2 crenulation cleavage planes. However, on the limbs of these folds S_1 is commonly transposed into the S_2 planes.

In the underlying Sugarloaf Slide Zone the S_1 fabric is most intense. It is represented in the phyllonitized gneiss and granite by zones of granulation that are in most places schistose. These zones bound, and form augen, around lenses and bands of quartzo-feldspathic material that are up to 2.5 cm thick (Plate 43, fig. a). With increasing deformation the zones of granulation are thicker, the schistose matrix finer-grained, and the quartzo-feldspathic bands are broken down into small, augen shaped, lenticles. The following stages of cataclasis and resultant rock types, that are of course gradational into each other have been distinguished in thin section.

Stage 1 - Development of thin (1 mm) irregular zones of granulation with marginal crushing of the original minerals and accompanying growth of sericite (Plate 43, fig. b). Quartz crystals become strained and broken, and develop sutured grain boundaries with a weak alignment of their long axes. Feldspars are strained, kinked and broken (Plate 44, fig. a).

Stage 2 - The zones of granulation are wider (2 mm) and arranged parallel (Plate 44, fig. b). Fine-grained sericite and chlorite in these zones give a sub-schistose fabric to the rock. The schistosity forms augen around broken and strained feldspar and elongate "polygonized" quartz porphyroblasts.

Stage 3 - Extreme granulation producing a fine grained phyllonite with relict porphyblasts of aligned quartz and feldspar arranged in lenticles. The fine-grained groundmass consists of a schistose aggregate of quartz, feldspar, sericite and minor chlorite (Plate 45, figs. a and b).

In the tectonic schists, D_E produced a fine streaky foliation (see Plate 14, fig. a). A biotite schistosity (S_1) is developed in the pelitic foliae which is preserved between the S_2 cleavage planes in the hinges of the D_A (F_2) folds, but is completely transposed into the S_2 planes on these fold limbs. D_A (F_2) folds are poorly developed in the calc-silicate foliae and the S_1 fabric is in most places transposed into the S_2 planes.

MAIDEN POINT SLICE

Introduction

The D_E deformation in the Maiden Point Slice produced a weak to strong slaty cleavage with rare associated recumbent folds.

Minor Structures

Emplacement (D_E) folds in the Maiden Point Slice were only observed in the Croque Harbour area at 2 km south of Cobbler's Cove, 3.2 km north of Pilier Bay and at the north shore of the northwest arm of Croque Harbour. Locally, in the Maiden Point Formation, especially south of Canada Head, F_1 recumbent folds are inferred from the downward-facing directions (Shackleton, 1958) of the later D_A (Acadian) F_2 folds. North and south of Croque Harbour the F_1 folds are close to open (interlimb angle 60^0 - 100^0), symmetrical, similar, recumbent folds that trend east-west and face up to the north on the S_1 cleavage. These folds appear to occupy the nose of a north facing recumbent fold that is of small amplitude as most of the F_2 folds are upward facing. The fold at the north shore of Croque Harbour occurs in an isolated exposure of medium to fine-grained greywackes and trends northeast.

Major recumbent F_1 folds are inferred from Canada Head to Cat Cove on the basis of downward-facing F_2 folds developed in belts up to 1 km wide. North of Canada Bay downward-facing F_2 folds are less common and of limited areal extent. This implies that no major F_1 recumbent folds are developed in the north of the area.

Fabric

Elsewhere within the Maiden Point Slice the D_E deformation produced a fabric (S_1) that is an axial plane fabric to the recumbent F_1 folds and hence is generally arranged close to the bedding. The intensity of development of S_1 decreases northeastwards, being very intense at Canada Head and east of Englee and weak to absent west of Maiden Point in Hare Bay.

The contrast in the development of S_1 between pelites and greywackes is marked. In the pelites a slaty cleavage, defined by a preferred orientation of sericite, is generally developed. In the greywackes the S_1 fabric varies with increasing deformation from:

(1) Anastomosing trains of sericite that meander around unstrained detrital grains, to

(2) parallel films of sericite that form augen around detrital grains and define definite cleavage planes (see Plate 19, fig. a), to

(3) aligned films of sericite associated with a dimensional preferred orientation of the clasts (see Plate 46, fig.a).

Examples of the last stage are rare and are only developed where the later D_A (F_2) folds have strongly over printed the S_1 cleavage, for example, at Canada Head, Englee, and Cow Bay in Hare Bay. Williams (1972) noted a similar relationship in greywackes that had suffered two periods of deformation and concluded that the development of preferred orientation of detrital material was accomplished by the F_2 folding which caused a morphological change of the S_1 fabric.

An S_1 slaty cleavage defined by chlorite and rods of opaques is also present at the margins of the Maiden Point volcanics north of Croque. It is crenulated and commonly transposed by the later D_A deformation.

CROQUE HEAD SLICE

Fabric

A penetrative S_1 slaty cleavage is developed in the greywackes and slates of the Maiden Point Formation that comprise the Croque Head Slice. Here the D_A F_2 folds are strongly superimposed on the S_1 fabric which is crenulated and transposed by S_2. A slaty cleavage S_1 is also developed in the underlying mélange that is folded by the F_2 folds. For this reason the S_1 fabric in both rock units is assigned to D_E deformation.

No associated F_1 folds were seen in the Croque Head Slice and all the F_2 folds are upward facing. S_1 is everywhere developed within the slice and is in most places more intense than in the underlying Maiden Point Slice. S_1 is locally defined by a dimensional preferred orientation of the detrital material in addition to the usual aligned sericite films.

ST. JULIEN ISLAND SLICE

Fabric

The D_E deformation produced a slaty cleavage S_1 in the St. Julien Island Slice that is folded by the upright D_A F_2 folds. S_1 is best developed in the interbedded tuff at the base of the Irish Limestone at the southeast tip of the island where a slaty cleavage, defined by chlorite, sericite, and rods of opaques, is crenulated by the D_A F_2 folds (Plate 46, fig. b).

In the Irish Limestone S_1 is weak and consists of closely spaced, parallel, dark, lines that are composed of rods of opaques and fine-grained sericite.

In the St. Julien Island Formation an S_1 slaty cleavage is locally preserved between the D_A (S_2) cleavage planes in the conglomerate matrix and the interbedded medium grained sandstones. However, in most places the external transposition of S_1 into S_2 is complete and S_1 cannot be identified.

Faults

A steeply dipping, northeast trending fault juxtaposes the Irish Limestones to the east and the St. Julien Island Formation to the west. The fault is exposed near the south end of the island where it is marked by a 0.6 m thick zone of vein quartz. The D_A upright cleavage is developed in the fault zone thus the fault was a pre D_A event, probably related to the D_E deformation.

DISCUSSION

The emplacement deformation resulted in a number of distinct but probably related events:

(1) the formation of the major detachment thrust of the various allochthonous slices;

(2) the production of mélange zones;

(3) the internal deformation of the slices, deformation of the mélange zones, deformation of the autochthon, and the production of the imbricate structure in the autochthon at Canada Bay.

The facing directions of the recumbent folds in the Maiden Point Slice and in the autochthon, the direction of overthrusting in the Canada Bay area, and the stretching lineations locally developed below the minor thrusts, all combine to suggest a northwesterly to northerly direction of allochthon emplacement.

D_A POST-EMPLACEMENT DEFORMATION

After the D_E deformation and the emplacement of the allochthon in middle Ordovician times, the entire area was affected by another major deformation event termed D_A. Folds related to D_A trend northeast and produce the regional structure of the area. They fold the allochthon-autochthon contact (evident in map pattern at Coles Pond and in Whites Arm Window) and the contacts between the various slices. The earlier tectonite elements related to D_P and D_E are crenulated and partially transposed by the D_A effects.

AUTOCHTHON

Minor Structures

In the Goose Tickle Formation west of the Maiden Point Slice at Big Springs Inlet the D_A folds are open (interlimb angle 90°), westerly inclined, northeast-plunging, flexural-slip, folds (Plate 47, fig. a) with an associated fracture to slaty cleavage that is in most places the first cleavage in these rocks. The amplitude of the folds decreases westwards where open anticlines and synclines are developed. Eastwards, into the Maiden Point Slice the intensity of D_A deformation increases and in the autochthonous rocks exposed through Whites Arm Window, the D_A folds are tighter (interlimb angle 60-80^{o}) and more upright than folds of the same age developed only 4.8 km to the west at Big Springs Inlet. At Whites Arm the associated axial plane fabric is crenulation S_2 cleavage in the mélange zone, in the Goose Tickle Formation, but in the Table Head Formation is a first phase slaty cleavage.

Southwards along the western margin of the Maiden Point Slice, D_A folds are well exposed at Coles Pond where the allochthon-autochthon boundary is folded around northeast-plunging open folds. In the Canada Bay area D_A folds are common, especially across the south shore of Canada Bay. The D_A folds are well-developed in the Goose Tickle Formation east of Englee and west of Canada Harbour (Plate 47, fig. b) and are upright to moderately-inclined, northeast-trending, similar, folds. They fold the D_E (S_1) slaty cleavage, and a superimposed S_2 crenulation cleavage is developed making these folds second phase structures. At Burnt Point a D_E fold in the Table Head Formation is refolded by an upright D_A (F_2) structure (Plate 48, fig. a). Elsewhere in the Table Head Formation D_A fold closures are rare and of small amplitude (0.5 - 1 m).

In the Sugarloaf Schists D_A folds are common. Here the folds are tighter (interlimb angle 20-40^{o}) and more inclined than the D_A folds at Canada Bay. The folds are asymmetrical with a sense of vergence that indicates an antiform to the west (see Plate 48, fig. b). Facing directions are rare and were only determined from the gritty psammite that outcrops midway across the cove where the D_A folds face up to the southwest on the S_2 cleavage.

The D_A folds in the Sugarloaf Schists are clearly second phase structures (F_2) since the earlier schistosity (S_1) is visibly preserved between the S_2 cleavage planes (see Plate 13, fig. a). Marble beds are drawn apart on the limbs of the F_2 folds into boudins and the boudinage axis and fold axis are parallel and trend northeast.

In the Sugarloaf Slide Zone the D_A folds are tighter and of smaller amplitude (0.3 m) and less common than in the overlying schists. The tectonic fabric (S_1) of the slide zone is generally arranged subparallel to S_2 and is locally indistinguishable from it. In the phyllonitized gneiss small kink-style F_2 folds are developed (see Plate 15, fig. a).

MAIDEN POINT SLICE

The D_A deformation produced upright to inclined folds in the Maiden Point Slice. In places, especially in the northeast of the area at Hare Bay these are first phase folds but elsewhere the folds fold an earlier slaty cleavage (the emplacement (D_E) fabric) and are second phase folds (F_2). Northeast trending minor D_A folds are especially well exposed along the south shore of Hare Bay where they are open to close (interlimb angle 60-80°) similar folds (Plate 49, fig. a). The folds generally plunge moderately northeastwards but reversals due to fold culminations do occur. The folds are steeply inclined and locally have slightly overturned western limbs, for example at the western shore of Whites Arm Pond. The persistent sense of vergence shows an antiform to the west but major fold closures could not be determined on account of lack of marker horizons and extensive late faulting. At Hare Bay the Maiden Point Formation faces upwards on the D_A cleavage even when this cleavage is demonstrably a second phase crenulation cleavage.

At the north shore of Whites Arm the D_A folds visibly fold the mélange and overlying Maiden Point sandstones (Plate 49, fig. b). Southwards around Croque the D_A folds refold the D_E recumbent F_1 folds and an S_2 crenulation cleavage is developed in the slates and a fracture cleavage in the sandstones. From Conche to Englee the D_A folds are second phase structures (F_2) and fold the D_E slaty cleavage. Downward facing D_A folds occur 3 km north of Hillier Harbour but elsewhere the D_A folds are generally upward facing. However, south of Canada Head, downward-facing D_A folds are areally extensive indicating the presence of major D_E recumbent folds. At Canada Head the D_A folds are steeply-inclined to the west (dip of axial plane 60-74^o) but southwards the folds become moderately to gently inclined but still maintain northeast trends.

Fabric

The nature of the D_A fabric is dependent on lithology and on the presence or absence of the earlier D_E fabric. Where D_E produced a penetrative slaty cleavage (S_1) the D_A cleavage is a crenulation cleavage (S_2) (see Plate 50, fig. a). The S_1 fabric is preserved at the fold hinges between the S_2 planes but on the fold limbs S_1 is completely transposed into S_2. It is common to find that the D_A fabric is a S_2 crenulation cleavage in the slate horizons but is a closely spaced fracture cleavage or a weak slaty cleavage in the greywackes and volcanic rocks where the D_E fabric is absent. Locally in the greywackes where D_E fabric was a penetrative cleavage, this cleavage has been modified by the D_A folds and now consists of a dimensional preferred orientation of the detrital material that is crenulated by the D_A fabric (S_2), for example, at Cow Point, east of Englee and Canada Head.

CROQUE HEAD SLICE

D_A folds are well-developed in the Croque Head Slice. They are upright, close (interlimb angle 60-65^o), similar, folds and fold a strong S_1 slaty cleavage and are thus second phase structures (Plate 50, fig. b). Fold culminations are common south of Croque Head and the folds plunge gently to moderately northeast and southwest. The folds show a constant sense of vergence that indicates an anticline to the west and are upward facing on the S_2 cleavage. However west of Black Island the F_2 folds are locally downward facing. The axial plane of the F_2 folds is upright or dips steeply to the east and north of Rets Point dips 80^o to the west.

The contact between the Croque Head Slice and its basal mélange is exposed at Rets Point and in several exposures along the east side of Irish Bay and at the north side of Cobblers Cove. At all localities the contact is folded by the upright D_A folds (See Plate 32, fig. a).

Fabric

The D_E S_1 cleavage is a penetrative fabric in the greywackes and slates of the Croque Head Slice. Consequently the D_A crenulation cleavage S_2 is well - developed. Commonly there is a dimensional preferred orientation of the detrital material in the S_1 plane (Plate 51, fig. a) and locally the deformation was sufficiently intense to destroy the clastic texture in the greywackes to convert the rocks to semi-schists.

ST. JULIEN ISLAND SLICE

Minor Structures

Minor upright D_A folds are common in the Irish Limestone but are rare in the St. Julien Island Formation. The contact between the St. Julien Island Slice and its basal mélange is folded around the south end of the island by a northeast plunging (F_2) syncline. In the Irish Limestone the D_A folds are best exposed at the northwest side of the island where they are close to open (interlimb angle 60-75^0), upright, similar, folds (see Plate 24, fig. a). The folds are upward facing on the S_2 cleavage.

In the St. Julien Island Formation on Black Island the D_A folds are upward-facing and asymmetrical indicating an anticline to the west (see Plate 51, fig. b). On St. Julien Island the D_A folds are nearly symmetrical and are also upward-facing.

Fabric

In the Irish Limestone the S_2 fabric is generally a closely spaced fracture cleavage, but is a crenulation cleavage in the interbedded greenschists.

A striking feature of the St. Julien Island conglomerate is the strong preferred orientation of the long axis of the clasts in the plane of the D_A (S_2) cleavage (Plate 52, fig. a). The cleavage forms augen around the clasts and, in some cases, cleaves them. The clasts record a variety of shapes that appear to be dependent primarily on lithology. The granite, quartz, and rhyolite, clasts are deformed the least and are oval-shaped with an average XYZ ratio of 1:1.5:1.8. The sandstone and shale clasts are much more deformed and are disc-shaped with an average XYZ ratio of 1:3:5.2. Some of the sandstone and shale fragments are so flattened that they no longer resemble clastic pebbles. This may, however, be partially due to initial shapes, whereby some of the clasts were originally angular slabs.

Rare quartzite clasts lie athwart the cleavage and show no signs of flattening. The deformation of the conglomerate appears to have been accompanied first by rotation, whereby the long axes of the clasts lie in the S_2 plane, and second by flattening. The amount to which the various clasts deformed appear to have been controlled by the ductility contrast between clast and matrix.

WHITE HILLS SLICE

Minor Structures

Minor, open, northeast-trending, D_A folds are developed in the Goose Cove Formation exposed on Fishot Island and south of Croque Head. These folds refold the pre-emplacement (D_P) recumbent folds that are second phase structures, so that the D_A folds in the Goose Cove Formation are third phase structures (F_3).

The contact between the Goose Cove Formation and its basal mélange is folded into a steeply dipping position by the D_A folds (see Plate 52, fig. b). Open, upright similar folds of small amplitude (1 - 2 m) are developed across the Fishot Islands. The folds generally plunge moderately northeastwards but on the north shore of Easter Tickle they plunge southwest. Refolded recumbent F_2 folds in the greenschists are exposed at the north and south side of the western entrance to Fishot Harbour (Plate 53, fig. a) and in the amphibolites at the north and south shores of Easter Tickle in interbedded marbles (see Plate 53, fig. b and Plate 30, fig. a). The F_3 folds show a constant sense of vergence that indicates an antiform up to the west.

Northeast trending shear zones, from 0.3 to 1.2 m wide, are locally developed at the Fishot Islands. The F_3 folds are especially well developed in the shear zones and are tighter and of greater amplitude than the F_3 folds in the surrounding schists. The variation in style is sharp and occurs at the shear zone margins (Plate 54). Brown stained carbonate veins that surround brecciated angular particles of schist are extensively developed in the shear zones.

The small area of the Goose Cove Formation exposed south of Croque Head is extensively brecciated. The brecciation post-dates the S_2 schistosity and is folded by open upright F_3 folds. The formation of breccia may have been produced during emplacement of the allochthon.

Fabric

A weak crenulation cleavage (S_3) that is axial planar to the F_3 folds is locally developed. In thin section single amphibole crystals are kinked but there is no associated progressive metamorphic growth (See Chapter VII).

MAJOR D_A STRUCTURES

In the east and south of the area the minor D_A folds are symmetrical, upright to slightly inclined, but become asymmetrical, moderately inclined, and more open in the west. The style of the major folds mirror that of the minor folds.

A major northeast-trending, gently-plunging, D_A anticline occurs in the autochthon at Big and Little Springs Inlet. The fold is inclined towards the west and its western limb is faulted (see Fig. 3, section A-B). The fold repeats the Maiden Point Slice on its western limb at Indre Point. In the autochthon south of Canada Bay, the major anticlines and synclines are more upright and of smaller wavelength. At Burnt Point, the Table Head Formation is exposed in the crest of a major D_A anticline.

It has not been possible to outline the major folds within the Maiden Point Slice on account of the lack of marker horizons. However, the base of the slice is an easily mapped contact and delineates a major anticline exposed across Whites Arm Window. This major fold is asymmetrical with a slightly overturned western limb and a moderately dipping eastern limb (see Fig. 3, section C-D).

The minor D_A folds in the Croque Head, St. Julien Island, and Fishot Islands slices all show a constant sense of vergence that indicates these slices lie on the western limb of an upward facing synform. The base of the St. Julien Island Slice is folded around the south end of the island clearly indicating the northeast plunging, synclinal form of the major fold.

CONCLUSIONS

The D_A deformation affected the entire area and was a major period of folding. The deformation is most intense in the east of the area where the folds are upright, nearly symmetrical, similar folds. Westwards the intensity of deformation decreases and the folds become more open, asymmetrical and inclined to overturned towards the west and locally, at Hare Bay, are associated with high angle reverse faults. This variation in style and decrease in intensity of deformation westwards accords with that expected from a section across an orogen from its central region to the platformal margin and reflects a variation in the attitude of the principal deforming stresses (Wilson, 1961).

LATE MINOR STRUCTURES

Late minor structures that post-date the D_A deformation are sporadically developed in the eastern parts of the area and consist of gentle folds and kink bands.

Northeast-trending minor folds are locally developed in the Sugarloaf Schists and Maiden Point Formation around Sugarloaf Cove. The folds are open to gentle (interlimb angle 100-125^0), upright, and are developed on the inclined limbs of the D_A folds. At Sugarloaf Point a 1.2 m thick diabase dyke, that intrudes the Maiden Point formation post-tectonically with respect to D_A, is folded by late upright minor folds, that are weak to absent in the bounding greywackes. An associated fracture cleavage is developed in the dyke but elsewhere no fabric was seen associated with the minor folds. Late northeast and south-southeast trending minor folds are developed in the St. Julien Island Formation that warp the D_A cleavage.

Reverse kink bands (Dewey, 1965) are locally developed in the northeast of the area. They show a variety of trends but most are steeply-dipping. The kink bands show no apparent spatial relationship to the minor folds of the area and are considered to represent a discrete late phases of deformation.

LATE FAULTS

Numerous high angle northeast trending faults cut the allochthonous rocks and post-date the D_A folding.

The Wild Cove Fault (Betz, 1939) is the only major late fracture in the area and appears to be related to the horsting of the Long Range Complex. At Wild Cove the fault throws the Long Range Complex and the overlying Bradore and the Devils Cove Formations against the Goose Tickle Formation. There, the fault consists of two branches, and the eastern branch is exposed at the east side of the cove where it is marked by 0.6 m of steeply dipping mylonitized marble (see Plate 7, fig. a). Southwards towards White Bay the fault cuts out the autochthonous sequence that is exposed on the coast of Canada Bay between Wild Cove and Burnt Point. At White Bay the fault juxtaposes the thrusted block of Acid and Basic Gneiss to the east, against microgranite and the underlying Bradore Formation and Long Range Complex to the west. The fault plane is marked by an inaccessible, sub-vertical gully.

The uplift of the Long Range Complex likely occurred over a long period of time. The evidence from the Wild Cove Fault suggests final movements occurred in post-Acadian (Devonian) times.

CHAPTER VII

METAMORPHISM

The lithological and structural contrast between the White Hills Slice and the lower, dominantly sedimentary, slices is further emphasized by their different metamorphic histories. The Goose Cove Formation records pre-emplacement metamorphism up to pyroxene-amphibolite grade, whilst the rocks of the lower slices record syn-emplacement, low-greenschist, metamorphism. Little or no progressive metamorphism accompanied the post-emplacement Acadian deformation (D_A).

The relationship between the growth of metamorphic minerals and the various structural events (D_P, D_E, D_A) has been elucidated using microtextural methods. Microtextures present in the rocks have been used as "time markers" in determining the growth history of the metamorphic minerals.

PRE-EMPLACEMENT METAMORPHISM

Metamorphism related to the pre-emplacement deformations (D_{P1}, D_{P2}) only affected the basal part of the White Hills Slice - the Goose Cove Formation. In the greenschist member of this formation the main metamorphic growth was syn-tectonic with D_{P1} movements whilst in the amphibolite member it was pre- to syn-tectonic with respect to the D_{P2} deformation. The metamorphic growth history of the formation is summarized in Fig. 19.

D_{P1} Syntectonic Growth

Syn-tectonic growth with respect to D_{P1} was non-porphyroblastic. In the greenschist member it led to the development of fine-grained tremolite-actinolite, chlorite and epidote, defining the S_1 schistosity (see Plate 38, fig. b) and accompanying growth of plagioclase (An_{8-12}). In the psammite and semi-pelite S_1 is defined by muscovite.

In the amphibolite member the S_1 schistosity is defined by a dimensional parallelism of fine to medium-grained hornblende. Syn-tectonic biotite occurs as oriented inclusions in garnet and plagioclase porphyroblasts (Plate 55, fig. a) that are post-kinematic with respect to D_{P1}. Garnets rarely have small S-shaped inclusion trails in the cores of porphyroblasts indicating growth may have initiated syn-tectonically with D_{P1}. Magnetite is developed in small rods oriented in the S_1 planes.

D_{P1} Post-tectonic Growth

Post-tectonic crystallization with respect to the first deformation is of minor importance in the greenschists but is extensive in the amphibolites. In the greenschists it is marked by rare, randomly oriented crystals of tremolite-actinolite and epidote that overgrew the S_1 fabric. Small post-tectonic garnets grew in the slide zone near the base of the formation.

In the amphibolites D_{P1} post-tectonic crystallization is marked by porphyroblastic growth of garnet and plagioclase (An_{36-40}). Some of the garnets overgrew and include the S_1 fabric (Plate 55, fig. b) and one crystal with an included helicitic fold of opaques was noted (Plate 56, fig. a). The aligned opaques either represent bedding or else an early tectonic fabric that pre-dated the D_{P1} event. However, no other evidence was found to substantiate a pre-D_{P1} event in the Goose Cove Formation on Fishot Islands. Commonly the inclusions in the garnets are arranged in a number of concentric rims that mimic the hexagonal shape of the porphyroblasts (Plate 56, fig. b). This inclusion pattern is believed to be the result of rapid post-tectonic growth that engulfed impurities concentrated at the advancing grain boundaries.

The plagioclase porphyroblasts are generally inclusion free but, in a few places, they contain both oriented and unoriented biotite flakes.

D_{P2} Syntectonic Growth

D_{P2} syn-tectonic growth is restricted in the greenschists to the psammite and semi-pelitic rocks where oriented chlorite and muscovite define the S_2 schistosity. Towards the contact with the amphibolite member minor syntectonic growth of tremolite-actinolite and epidote is developed. In the amphibolites a strong dimensional parallelism of biotite in the pelitic rocks and of hornblende in mafic rocks defines the S_2 schistosity. The S_2 schistosity forms augen around the garnet and plagioclase porphyroblasts (see Plate 56, fig. b) and, in rare cases, garnet shows narrow overgrowth zones which contain curved inclusion trails in the D_{P2} syn-tectonic portions. Quartz shows almost complete regrowth in dimensionally oriented crystals.

The schists in the contact zone exhibit a strong to moderate S_2 fabric, defined by aligned pyroxene and deep brown hornblende, and plagioclase (An_{40}) crystals,(Plate 57, fig. a). There is no preserved remnant of the S_1 fabric.

D_{P2} Post-Tectonic Growth

Post-tectonic crystallization effects with respect to D_{P2} deformation are weak and are characterized by annealing of feldspar in the contact zone, annealing of quartz throughout the formation and by restricted growth of unoriented muscovite in the psammites. The muscovite overgrows the S_2 fabric and represents static growth after the D_{P2} deformation. In the contact zone, rare unoriented hornblende grew across the S_2 schistosity. Sphene overgrows S_2 in both the greenschist and amphibolite members.

Late post-tectonic sericitization and saussuritization of feldspar, chloritization of biotite, and retrogression of hornblende and pyroxene is widespread but not strongly developed except locally in the contact zone. Part of this retrograde metamorphism may have accompanied the post-emplacement deformation (D_A) and is described later.

Discussion

The Goose Cove Formation records an increase in metamorphic grade, in grain size, and in the development of new minerals toward the contact with the ultramafic sheet during D_p metamorphism. Whilst temperature is a dominant and obvious control, the role of strain is also important as witnessed by the strong tectonic fabric of the schists. Deformation and heating have combined to produce the high grade dynamothermal aureole.

Increase in the grade of metamorphism from the structural base to the top of the formation is witnessed by an increase in the An content of plagioclase from An_8 to An_{40}, a change from tremolite-actinolite through green hornblende to brown hornblende, and by the appearance of augite, and these changes are summarized diagrammatically in Fig.20.

The metamorphic assemblage of the basic schists in the greenschist member, namely, chlorite-albite-epidote-muscovite-tremolite-actinolite, is characteristic of the greenschist facies of regional metamorphism.

The typical assemblages of the amphibolite member are: green hornblende—plagioclase (An_{36-38}) for mafic rocks, and quartz-biotite-garnet-plagioclase (An_{36-38}) for pelitic rocks. These are typical assemblages of the amphibolite facies of regional metamorphism.

The assemblage hornblende-plagioclase (An_{38-40})-augite of the contact zone is absent from the hornblende-hornfels or pyroxene-hornfels facies but occurs in the transition between the amphibolite and the granulite facies. It should be emphasized here that the rocks of the contact zone do not show any hornfels textures and the term hornfels is inapplicable.

D_E SYN-EMPLACEMENT METAMORPHISM

Low-grade regional metamorphism accompanied the emplacement deformation (D_E) and affected the lower sedimentary slices, the intervening mélange zones and the upper part of the autochthon (see Plate 57, fig. b).

Most of the rocks have reached chlorite grade and typical assemblages are:

I. Pelitic rocks and greywacke matrix:
sericite-chlorite-quartz.

II. Mafic rocks (Volcanics and Tuffs):
chlorite-sericite-epidote-opaques-quartz.

In the Sugarloaf Schists pale brown biotite is locally developed and the assemblage consists of sericite-biotite-epidote-quartz.

D_A POST-EMPLACEMENT METAMORPHISM

Little or no metamorphic growth of new minerals accompanied the post-emplacement D_A deformation, which was mainly a folding event.

In the Goose Cove Formation shearing, with associated cold working of the schists is locally developed. Quartz recrystallization accompanied D_A since quartz is not strained in the hinge zones of the D_A folds. In the amphibolite member local kinking of hornblende crystals is developed. Since the D_A (S_3) crenulation cleavage is only locally developed it is difficult to determine how much of the retrograde metamorphism of the Goose Cove Formation is related to this deformation.

In the lower sedimentary slices and in the autochthonous rocks the D_A deformation caused recrystallization of quartz but only in the east of the area in the St. Julien Island Formation was there minor growth of sericite along the D_A (S_2) crenulation cleavage planes.

POST D_A THERMAL METAMORPHISM

A late local period of low to medium grade thermal metamorphism affected a restricted area of allochthonous and autochthonous rocks from Canada Head southwards to Sugarloaf Cove. It produced a variety of porphyroblasts that overgrew and include the D_A crenulation cleavage (S_2) and clearly post-date the D_A deformation.

The grade of metamorphism increases southwards from Canada Head and within the Maiden Point Formation chlorite and biotite zones can be recognized. The biotite zone continues south into the Sugarloaf Schists and the Sugarloaf Slide Zone. The Acid and Basic Gneisses also record growth of unoriented micas but the growth phase cannot be accurately dated as these rocks are unaffected by the D_A deformation.

Zone of Chlorite

Chlorite porphyroblasts are developed locally in the Miaden Point Formation from Twillingate Cove north to Canada Head. Poor sampling from inland exposures does not permit accurate delineation of the zone boundary inland but it appears to run northwest to Wild Cove Pond.

Chlorite is best developed on a small island at the north entrance to Twillingate Cove where ovoid porphyroblasts up to 2 mm long form a knotted texture in slates. The ovoids have their long exes parallel to the D_A crenulation cleavage (S_2) but their margins clearly overgrow this fabric (Plate 58, fig. a). Locally the chlorite has nucleated about quartz and the ovoids have quartz rich cores. At Canada Head, unoriented chloritoid crystals are associated with the chlorite.

Zone of Biotite

Biotite porphyroblasts are developed in the Maiden Point Formation south of Twillingate Cove, in the Sugarloaf Formation, and in the

Sugarloaf Slide Zone, and define the zone of biotite. The various rock types record a variety of assemblages and they will be described separately.

Maiden Point Formation:

The Maiden Point greywackes attain a darker colour south of Twillingate Cove due to hornfelsing. Some of the fine-grained beds exhibit a basic igneous appearance and these may be what Betz recognized as volcanic rocks (Betz, 1939, p. 22). Randomly oriented red-brown biotite flakes are developed in the greywacke matrix and in the slate horizons. Biotite is associated with quartz, plagioclase, muscovite and locally with cordierite or andalusite or both. Cordierite and andalusite are best developed in slates northeast of Cat Cove. Cordierite forms oval, spongy, poikiloblastic crystals that exhibit nodular texture in places (Plate 58, fig. b). Andalusite is less common than cordierite and forms radiating laths up to 2 cm long (Plate 59, fig. a). Plagioclase is cloudy, rarely twinned, and has a composition of An_7.

Sugarloaf Schists:

Hornfelsing of the semi-pelites produced abundant biotite and minor cordierite, andalusite, tourmaline, tremolite-actinolite and epidote. The biotite is dominantly randomly oriented (Plate 58, fig. a) but is occasionally mimetic after the D_A (S_2) crenulation cleavage. Locally, biotite porphyroblasts are kinked which may be the result of the late minor folding developed in this area. Recrystallization of the quartzite bands has produced an unstrained polygonal mosaic, however, in some of the D_A hinge zones a dimensional preferred orientation of quartz parallel to D_E (S_1) is preserved. Cordierite was only seen at the western corner of Sugarloaf Cove where it forms inclusion packed ovoids.

The calc-silicates are characterized by growth of diopside, tremolite-actinolite, and epidote. Diopside forms large poikiloblastic crystals that tend to be arranged with their long axis parallel to D_A (S_2) fabric but commonly overgrows S_2. Small tremolite-actinolite flakes surround large diopside porphyroblasts and may be the product of degradation. In the thin calc-silicate bands included in the semi-pelites, tremolite-actinolite is developed instead of diopside and there is no evidence to suggest that it is secondary. Epidote forms thin veinlets and small aggregates of grains (see Plate 60, fig. a) and may be the result of retrograde metamorphism. Carbonate has recrystallized to form a sub-polygonal mosaic and quartz is generally unstrained although locally it still shows a dimensional preferred orientation parallel to D_E (S_1). Sphene was also a result of this event.

In the marble beds rare poikiloblastic diopside crystals are developed that contain inclusions of carbonate. The carbonate has recrystallized to a sub-polygonal mosaic.

In the mafic schists porphyroblasts of biotite and smaller grains of actinolite and epidote are developed. These minerals mimic and overgrow the composite D_A (S_2) fabric (Plate 60, fig. b).

Sugarloaf Slide Zone:

Hornfelsing of the tectonic schists of the Sugarloaf Slide Zone resulted in growth of diopside, epidote, and sphene in the calc-silicate bands and of biotite and tremolite-actinolite in the semi-pelitic bands. Diopside forms poikiloblastic crystals that mimic the D_A (S_2) fabric but locally overgrow it (Plate 61, fig. a). The quartzo-feldspathic bands have recrystallized to a sub-polygonal, generally unstrained, network. In the semi-pelites small flakes of red-brown biotite clearly overgrow the D_A (S_2) crenulation cleavage. Diopside, and/or tremolite-actinolite are developed in the quartzo-feldspathic bands and overgrow the S_2 fabric. In one section idocrase was recognized.

Hornfelsing effects on the phyllonitized gneiss and granite are slight and the fine-grained sericitic groundmass shows only minor growth of biotite, epidote and chlorite. The biotite mimics the phyllonite fabric and may in fact be relict micas unaffected by the D_E deformation. The underlying Acid and Basic Gneisses show no affects of thermal metamorphism.

Discussion

The assemblage of the chlorite zone accords to the albite-epidote-hornfels facies or to low-greenschist facies. The biotite zone represents higher grade of metamorphism and the occurrence of cordierite, andalusite and diopside accord with the lowest part of the hornblende-hornfels facies (Turner, 1968). This late thermal metamorphic event was possibly produced by an unexposed granite intrusion. Late Acadian (Devonian) granites are common in the Central Mobile Belt to the east but are rare in the Western Platform. The only known occurrences are at the eastern margin of the Western Platform at Sops Arm at the head of White Bay. There, late Acadian granites intrude Silurian sediments (Lock, 1972).

CHAPTER VIII

CORRELATIONS, SOURCE AREA AND TECTONIC EVOLUTION OF THE HARE BAY ALLOCHTHON

In this chapter the allochthonous rocks are first correlated with the northern part of the Hare Bay Allochthon and within this framework the stratigraphic evolution of the allochthonous rocks is discussed. The Hare Bay Allochthon is then correlated with the other allochthons found along the western margin of the northern Appalachians. A possible source area of the Hare Bay Allochthon is proposed and, finally, the tectonic evolution of the allochthon and autochthon is discussed in terms of an evolving continental margin.

CORRELATIONS WITH THE NORTHERN PART OF THE HARE BAY ALLOCHTHON

The Maiden Point Slice and the White Hills Slice continue north of Hare Bay and underlie most of the allochthon in that area (Fig. 2). However, in addition, two areally small slices, the Northwest Arm Slice (Tuke, 1968; Williams et al., 1973) and the Cape Onion Slice (Williams et al., 1973) outcrop near the western margin of the allochthon in the north. The occurrence of slices common to both areas allows direct correlations and establishment of the structural stacking order of the slices for the entire allochthon for the first time. (See Fig. 21). A brief description of the two slices represented only in the north, and of the White Hills Peridotite Sheet, is included, as it is essential that the allochthon be viewed as a whole when discussing its depositional environments, source area and tectonic evolution.

Northwest Arm Slice

The Northwest Arm Slice is structurally the lowest slice of the allochthon and outcrops around its western extremity at Hare Bay and Pistolet Bay (Fig. 2). It directly overlies the autochthonous Goose Tickle Formation and this formation is brecciated at the contact.

The Northwest Arm Slice consists of a chaotic and rubbly sequence of black and in places, green shale, buff-weathered limy siltstone, grey sandstone, green chert, white to grey limestone and limestone breccia (the Northwest Arm Formation of Cooper, 1937). The beds are everywhere broken up into angular to subrounded pebbles, boulders, and slabs up to 9 m long that are set in a shaley matrix. The slabs and boulders have their long axes oriented parallel to the bedding and thus the attitude of bedding can be discerned. Rarely, the shales attain a thickness of 60 m without any clastic interbeds, and here the formation lacks the rubbly appearance. Limy siltstone is the most common and distinctive lithology after shale. Flat lying, discontinuous, isoclinal, slumped folds are developed but cannot be traced on account of the discontinuity of the beds. The formation contains early Ordovician (Tremadocian) graptolites (Tuke, 1968) and the black shale-graptolite association suggests deposition in deep water.

Cape Onion Slice

The Cape Onion Slice intervenes between the Maiden Point Slice and the White Hills Slice at Cape Onion Peninsula and possibly at Lock's Cove. The slice consists of basaltic pillow lava, agglomerate and minor slate that are informally referred to as the Cape Onion Volcanics (Williams et al., 1973). The slates contain Tremadocian graptolites (Williams, 1971), crustaceans and brachiopods (Erdtmann, 1971), and are therefore coeval with the Northwest Arm Formation. However, these formations must have been deposited in widely separated areas as the Northwest Arm contains no volcanic rocks.

White Hills Slice

The White Hills Slice outcrops extensively north of Hare Bay and consists of a sheet of ultramafic rocks, referred to as the White Hills Peridotite Sheet (Cooper, 1937), underlain by a metamorphic aureole referred to as the Goose Cove Formation. This structural unit comprises the highest slice of the allochthon. The ultramafic rocks occupy the highest ground in the eastern and western White Hills and are at least 300 m thick with top not seen. A high magnetic anomaly occurs seaward of the Goose Cove Formation on Fishot Island which suggests the presence of submerged ultramafic rocks (see Fig. 6).

White Hills Peridotite Sheet:

(This description is based on reconnaissance work by the author).

The ultramafic rocks consist of banded harzburgite and lherzolite, minor dunite and thin garnet-amphibole layers that are restricted to the base. The contact with the Goose Cove Formation was nowhere observed and is generally marked by a narrow (0.6 - 1.2 m) gap in exposure. The ultramafic rocks show a distinct compositional banding which is a function of the relative amounts of olivine and pyroxene. Numerous thin (2-15 cms) orthopyroxene bands that are resistant to weathering are most distinct.

The rocks contain a strong L-S tectonic fabric that is axial planar to rare recumbent folds (Plate 61, fig. b) so that in most cases the tectonic fabric is arranged parallel to the primary mineralogical bands. The tectonic fabric is most intensely developed around the periphery of the sheet and the intensity of its development appears to reflect proximity to its base.

The orthopyroxenite bands are of two generations. The first pre-date the tectonic fabric and are parallel to the other primary bands. The second are post-tectonic and cross-cut the banding and the tectonic fabric (Plate 62, fig. a).

A cold cataclastic deformation is locally developed at the base of the sheet. It is best observed in the vicinity of Daniels Lookout, northeast of Irelands Bight, where the earlier tectonic fabric is brecciated and refolded by minor discontinuous flat lying folds.

Petrography - Garnet amphibole layers were only seen at the base of a small outlier of the peridotites east of the Eastern White Hills. They consist of phenocrysts of olivine and enstatite that are augened by fibrous secondary amphibole and small colourless garnets. The olivine is partially serpentinized and enstatite has recrystallized to a granular texture and shows partial alteration to amphibole at the margins of phenocrysts. The amphibole is colourless, biaxial positive, and, optically, resembles cummingtonite. It exhibits a strong preferred orientation that defines the tectonic fabric.

Dunite is only present in significant proportions in the Eastern White Hills. It occurs near the base of the sheet along its northwestern and eastern margins. It consists of almost 100% olivine with local grains of enstatite and accessory chrome-spinel. All of the samples sectioned show approximately 30% serpentinization, with the serpentine forming a mesh texture.

Harzburgite and lherzolite are the major phases present. The harzburgite weathers to a medium yellow brown and is contrasted with dark weathering lherzolite. In both types enstatite occurs as long laths up to 5 mm long and 0.8 mm across that are arranged parallel to the tectonic foliation (Plate 62, fig. b). Olivine has recrystallized and shows a rough dimensional orientation that tends to form augen around the enstatite phenocrysts. Augite forms up to 20% of the lherzolites and is present as equant grains up to 1 mm across. Chrome-spinel is always present in accessory amounts in both rock types. It tends to be concentrated into stringers from 1 - 2 mm long arranged parallel to the foliation.

The orthopyroxenite bands contain up to 90% pyroxene, (approximately 60% enstatite, 30% augite), 10% olivine, and accessory chrome-spinel. Large enstatite crystals stand out on weathered surfaces and form both laths and equant grains.

Age - The White Hills Peridotite Sheet is petrographically and lithologically similar to the lower part of the Bay of Islands Igneous Complex (Cooper, 1936; Smith, 1958), which forms the highest slice of the Humber Arm Allochthon (Stevens, 1970). The Bay of Islands complex consists of a complete ophiolite suite (Williams and Malpas, 1972) and, as in the case of the White Hills Peridotite Sheet, the peridotites overlie a metamorphic aureole that is an integral part of the highest slice. The White Hills Peridotite sheet is correlated with the basal ultramafic part of the Bay of Islands complex and can therefore be viewed as a partial ophiolite.

The age of the White Hills Peridotite Sheet is uncertain. Stevens (1970) considered the Tremadocian Cape Onion Volcanics to form the roof of the White Hills Ophiolite suite. However, it is now known that the volcanic rocks constitute a separate slice that is overlain by amphibolites of the White Hills Slice (Williams et al., 1973). Thus the stratigraphic relationship of the volcanic rocks to the peridotites is uncertain.

Ophiolites have recently been recognized on the Burlington Peninsula at Baie Verte (Bird et al., 1971), Mings Bight (Dewey and Bird, 1971) and Snooks Arm (Upadhyay et al., 1971). The latter is of lower Ordovician age (Arenig), (Snelgrove, 1931) and a similar or slightly older age for the White Hills Peridotite Sheet has been suggested (Stevens et al., 1969).

A characteristic feature of many ophiolites is that a relatively short time elapsed between extrusion of the basalt cap and thrust emplacement (Reinhardt, 1969; Moores and Vine, 1971; Smith, 1971; Milsom, 1973). If the Cape Onion Volcanics are taken as the top of the White Hills Ophiolite then the time interval between formation of the basalts and thrust emplacement is in the order of 30 to 50 m.y. A similar time interval elapsed between formation and thrusting of the Papuan Ultramafic Belt (Davis and Smith, 1971).

Discussion

The complete Hare Bay Allochthon consists of six distinct thrust slices (Fig. 21) that are, in most places, separated from each other by mélange zones. There is no continuous vertical succession showing all six slices and the stacking order was built up from relationships throughout the allochthon. The relative position of the Cape Onion Slice to the St. Julien Island Slice is not seen. Both rest on slices of the Maiden Point Formation (either the Maiden Point or Croque Head Slice) and the Cape Onion Slice is overlain by the White Hills Slice. It will be argued on paleogeographic grounds that the St. Julien Island Slice is the lower as shown in Fig. 21 (c). The succession of slices therefore, from lowest to highest, is as follows: Northwest Arm, Maiden Point, Croque Head, St. Julien Island, Cape Onion and White Hills.

STRATIGRAPHIC RECONSTRUCTION OF THE ENTIRE HARE BAY ALLOCHTHON

Introduction

All of the sediments in the Hare Bay Allochthon are considered to have been deposited in relatively deep water by turbidity currents and, with the exception of the St. Julien Island conglomerates, were derived from the west from the Grenville basement or from the overlying shallow water clastics and carbonates. The St. Julien Island conglomerates were derived from a different source, most likely from an island arc, within the geosynclinal depositional environment. The volcanic rocks of the higher slices (Cape Onion Volcanics and the Goose Cove Formation) formed well within the geosyncline, generally out of reach of clastic detritus. The White Hills Peridotite Sheet consists of mantle peridotite and must have formed in an oceanic domain that lay farther to the east.

These more or less contemporaneous rocks units are now found in a vertical sequence with tectonic contacts and cannot represent a normal stratigraphic succession. A key to understanding the stratigraphy comes from the constant stacking order of the slices, their contrasting facies, and contrasting pre-emplacement structural style. All combine to suggest that the slices originated in separate areas and that the higher slices formed farther from the continental margin and are the farthest travelled. A simple palinspastic restoration of the slices can thus be made by placing them in lateral order away from a continental margin (see Fig. 21, d).

This interpretation of the allochthnous rocks is valid only if it explains the facies variations, depositional environments, relative ages of the slices, and their pre and syn-emplacement structural and metamorphic development.

Discussion

In the scheme proposed the Northwest Arm Slice being the lowest tectonic slice of the allochthon must have formed closest to the continental margin. It is coeval with the authochthonous St. George Formation but yet contains sparse carbonate detritus. This suggests either deposition distal from the shelf or the presence of a physical barrier at the shelf edge that prevented eastward spreading of lime detritus.

The Maiden Point Slice consists of mafic volcanic rocks near the base overlain by a thick monotonous greywacke and pebble conglomerate sequence (Maiden Point Formation) derived from the Grenville basement to the west. The Maiden Point Formation must have been deposited prior to the Tremadocian Northwest Arm Formation, which, on palinspastic restoration, lay closer to the continental platform but contains no coarse clastic detritus.

The Croque Head Slice contains finer grained and thinner bedded greywackes than those of the Maiden Point Slice, which suggests that they represent a more distal (eastwards?) facies of the Maiden Point Formation.

If the palinspastic restoration is correct, the rocks of the St. Julien Island Slice formed east of the Maiden Point Formation.

The thin quartzite at the base of the Irish Limestone may represent the top of the Maiden Point Formation and mark the transition from eastward spreading sand to carbonate detritus. The Irish Limestone contains a high sand content and is lithologically distinct from the shaley Northwest Arm Formation and hence was probably deposited prior to the establishment of the black shale environment that lay closer to the continental margin. On this reasoning the Irish Limestone is of pre-Tremadocian age.

The St. Julien Island Formation contains volcano-plutonic detritus that was derived from within the geosyncline. This represents the first major change in the direction of sediment transport, which was related to a major tectonic event. No such detritus was found in any of the other formations, which suggests that the St. Julien Island conglomerate may be younger than them and may be of post Tremadocian age.

The Cape Onion Volcanics were considered (Stevens, 1970) to form the upper part of the White Hills Ophiolite but it is now known that they constitute a separate slice that is overridden by the White Hills Slice. If the stacking order of the slices reflects their geographic locations prior to slice emplacement, the Cape Onion Volcanics should have formed closer to the continent than the rocks of the White Hills Slice. However, there is no continent-derived detritus in the Cape Onion Volcanics as in the Goose Cove Formation of the highest slice. It is possible that the Cape Onion Volcanics are oceanic volcanics and formed the basaltic carapace of the White Hills

Ophiolite Suite, but during transport were dislodged and later overridden by the dismembered ophiolite. An alternative possibility is that the volcanic rocks formed in an island arc environment that lay between the continent and the site of ophiolite obduction. Chemical analyses of the Cape Onion Volcanics should provide valuable information on their origin.

The White Hills Slice is the highest slice of the allochthon and therefore should have originated farthest from the stable platform. The upper part of the slice consists of a partial ophiolite (the White Hills Peridotite Sheet) and the interpretation of ophiolite suites as slabs of on land oceanic-crust and mantle is now almost universally accepted (Bailey et al., 1970; Church, 1972; Coleman, 1971; Davies and Smith, 1971; Dewey and Bird, 1970, 1971; Moores, 1970; Moores and MacGregor, 1972; Reinhart, 1969; Stevens, 1970; Smith, 1971; Upadhyay et al., 1971; Zimmerman, 1972). This implies that an oceanic domain lay east of the stable platform and of the depositional sites of the lower structural slices.

The lower part of the White Hills Slice, the Goose Cove Formation, is welded to the base of the peridotite sheet. The intensity of deformation and grade of metamorphism decreases away from the peridotite contact and is therefore probably related to obduction and earliest transport of 'hot' ophiolite over the supracrustal rocks. The supracrustal rocks, the greywackes, agglomerates, pillow lavas and marbles, must have lain west of the site of ophiolite obduction. The association of greywackes interbedded with mafic volcanic rocks is atypical of oceanic crust, which suggests that the ophiolite obducted onto sediments and volcanics that formed at or near the margin of the proto-ocean.

Figure 22,a summarizes the approximate stratigraphic ranges of the rocks of each slice. The only definite lithostratigraphic correlations that can be made are between the Maiden Point and Croque Head Slices.

CORRELATION OF THE HARE BAY ALLOCHTHON WITH THE "TACONIC" ALLOCHTHONS OF THE NORTHERN APPALACHIANS.

The Hare Bay Allochthon can be correlated directly with the other "Taconic" allochthons that occur in a similar tectonic position along the western margin of the Appalachian orogen (Fig. 7). Each allochthon is characterized by a pile of tectonic slices with the lower slices consisting dominantly of sedimentary rocks and the upper slices consisting of varied assemblages of metasedimentary metavolcanic and plutonic rocks. Restored stratigraphic sections for each allochthon showing possible lithostratigraphic correlations are given in Fig.22 .

<u>Humber Arm Allochthon</u>

The Humber Arm Allochthon consists of five major tectonic slices (Williams et al., 1972; Williams, in press). The lower slice is a composite structural unit of a number of undefined slices that contain a clastic sedimentary succession of Cambrian to lower Ordovician age referred to as the Humber Arm Supergroup. Stevens (1970) subdivided the Humber Arm Supergroup into a basal quartzo-feldspathic flysch division (Summerside, Meadows and Irishtown Formations); a middle carbonate flysch division (Cooks Brook and Middle Arm Point Formations), and an upper quartzo-feldspathic flysch division (Blow Me Down Brook Formation). The basal flysch division is correlated with the Maiden Point Formation, the carbonate flysch is probably equivalent to the Irish limestone and Northwest Arm Formation and the upper flysch appears to be a fine-grained equivalent of the St. Julien Island Formation. The upper flysch was also derived from a volcano-plutonic source within the geosyncline but contains, in addition, ophiolite detritus (Stevens, 1970). Locally intervening between the clastic sedimentary rocks and the highest slice in the Humber Arm Allochthon are three slices of variably deformed volcano-plutonic rocks (Williams et al., 1972) that are not represented in the Hare Bay Allochthon. The highest slice consists of a complete ophiolite suite referred to as the Bay of Islands Complex (Smith, 1958; Williams and Malpas, 1972) with a basal metamorphic aureole of supracrustal rocks that are lithologically, petrographically, and structurally, similar to the Goose Cove Formation.

Taconic Allochthon of New York

The classic Taconic Allochthon of New England and New York contains at least six tectonic slices (Zen, 1967; 1972a). The lower five slices comprise a clastic sedimentary sequence of Cambrian to middle Ordovician age referred to as "the low Taconic Sequence" (Zen, 1967). This sequence is directly comparable to the sedimentary successions in the Newfoundland Allochthons. In particular, the basal Rensselaer Formation (Balk, 1953) can be correlated with the Maiden Point Formation. The highest slice of the Taconic Allochthon consists of black, grey and green phyllites and schists - the "High Taconic Sequence" (Zen, 1967), which contrasts with the slabs of ophiolite found in the highest slices of the Newfoundland Allochthons. Zen (1972b) recognized a sequence of tectonic events in the Taconic Allochthon that are comparable to these described from the Hare Bay area. The first event resulted in the formation of gravity slides. The second event resulted in recumbent folds and low grade regional metamorphism and accompanied or immediately followed slice emplacement. These early folds were later refolded by upright to inclined folds assigned to the Acadian. However, a notable difference is the absence of pre-emplacement structures in the Taconic Allochthon that are characteristic of the higher slices of the Newfoundland Allochthons.

Quebec City and Gaspe Allochthons

The Quebec City and Gaspe Allochthons are imperfectly known. It appears that they consist of lower, dominantly sedimentary slices (Sillery Group and Quebec Group) with the higher slices containing metasediments, meta-volcanics (Bennet Schists and Schickshock Group), and ophiolite suites referred to as the Thetford Mines Ophiolite (St. Julien, 1973; Laurent, 1973) and the Mount Albert Pluton (MacGregor, 1962). The Mount Albert Pluton has a metamorphic aureole of metavolcanic rocks and minor metasediments welded to its base. K-Ar ages from metamorphic mica in the aureole gives an early Ordovician age (MacGregor, 1962). In a recent paper Carrara and Fyson (1973) doubt the allochthonous nature of these rocks in the Gaspe area, but the clear analogy between the Mount

Albert Pluton and its surrounding metamorphic aureole to the high slices of the Newfoundland allochthons is striking and is compelling evidence in favor of a transported origin for these rocks at least.

Summary

In summary, the Hare Bay Allochthon shows striking lithostratigraphic and structural similarities to the other allochthons. This suggests that similar tectonic processes controlled their deposition, formation and subsequent westward transport in giant thrust sheets.

SOURCE AREA OF THE HARE BAY ALLOCHTHON

The proposed reconstruction of the Hare Bay Allochthon suggests that the area of deposition was at least 100 km wide and that it lay between the stable continental margin to the west and an oceanic domain to the east. Northwesterly to northerly transport of the allochthon is indicated by the north to northwest facing D_E recumbent folds in the autochthon at Canada Bay and in the Maiden Point Slice near Croque. Such a reconstruction would place the source area approximately 30 - 48 km northeast of the Burlington Peninsula - Notre Dame Bay area (see Fig. 1). However, the source must have lain much further to the north if a post-emplacement sinistral displacement is assumed to have taken place between the Western Platform and the Burlington Peninsula along the Cabot Fault (Williams et al., 1970).

The gross similarities between the transported sedimentary and volcanic rocks with those of the Fleur de Lys Supergroup on the Burlington Peninsula has long been known (Rodgers and Neale, 1963; Tuke, 1968; Stevens, 1970; Bird and Dewey, 1970; Smyth, 1971). Of prime importance in locating the source area is the establishment of the source of the ophiolites of the White Hills Slice. For if this can be ascertained, then the source of the sedimentary and volcanic rocks of the lower slices can be sought to the west of the oceanic domain.

Ophiolites outcrop on the Burlington Peninsula at Baie Verte, Mings Bight and at Snooks Arm in Notre Dame Bay. The Notre Dame Bay area is now known to be largely underlain by lower Paleozoic oceanic crust, referred to as the Snooks Arm Group and the Lushs Bight Group (Upadhyay et al., 1971; Strong, 1972; Smitheringale, 1972).

The Snooks Arm Ophiolite is thrust against the Cape St. John Group (Baird, 1951) which forms the upper part of the Fleur de Lys Supergroup. The top of the Snooks Arm Group contains abundant angular detritus derived from the Cape St. John Group (H.D. Upadhyay, pers. comm.) which implies that the ophiolite formed close to its present position at the eastern margin of the Fleur de Lys Supergroup. It is unlikely, therefore, that the Snooks Arm Ophiolite can represent the trailing edge of a large overthrust slab of oceanic crust and mantle, which according to Church and Stevens (1971) includes the Baie Verte Ophiolite, the White Hills Peridotite Sheet and the Bay of Islands Complex to the west. In addition, the Cape St. John Group probably represents an upper Cambrian to lower Ordovician Island Arc (Bird and Dewey, 1970) which means that if the allochthonous ophiolites originated in Notre Dame Bay then they must have passed over this arc. Obduction of ophiolite across an island arc is not known to have happened anywhere else in the geological past and is considered unlikely.

The Baie Verte Lineament consists of a line of ophiolites and must be considered a possible root zone for the White Hills Peridotite Sheet. The lineament is of fundamental importance and marks the junction between the contrasting western and eastern divisions of the Fleur de Lys Supergroup (Church, 1969). The western division overlies gneissic basement (de Wit, pers. comm.) and consists of a basal conglomerate, overlain by a thick coarse greywacke unit with syn-sedimentary mafic igneous rocks, followed by a thinner, more varied clastic sequence with carbonate breccias towards the top in the west and mafic volcanic rocks in the east. The eastern sequence, on the other hand, consists of an oceanic foundation (Dewey and Bird, 1971; Kennedy, 1973a) referred to as the Nippers Harbour Group, that is overlain by intermediate and acid volcanic rocks of the Cape St. John island arc (Bird and Dewey, 1970). Extensive pre-tectonic acid plutonic rocks referred to as the Burlington Granodiorite and the Cape Brule Porphyry intrude the eastern sequence and are interpreted to be related to the island arc.

The age of the Baie Verte ophiolite and its relationship to the Fleur de Lys Supergroup is uncertain and controversial. Kennedy (1973,a) distinguished two ages of ophiolite in this belt on structural criteria. The older, highly deformed, ophiolite, according to Kennedy formed in situ and was deformed with the Fleur de Lys Supergroup in pre-Ordovician times, and the younger, mildly deformed, ophiolite was thrust in from the Notre Dame Bay area after the Fleur de Lys deformation. This must have involved thrusting up considerable gradients over a deformed belt and is considered unlikely. De Wit (pers. comm.) reports that the deformation of the Fleur de Lys Supergroup and the Baie Verte Ophiolite belt occurred at the same time and that the difference in the intensity of deformation between the two groups is the result of different structural levels. If so, then the Baie Verte Ophiolite Belt can be viewed as remnants of a single ophiolite suite that may have formed in situ. This suggests the former presence of an oceanic domain along the Baie Verte lineament, which must be considered a most likely source area of the White Hills Peridotite Sheet.

Bird and Dewey (1970) considered the White Bay Zone to be the source of all the allochthonous rocks. However, this area is considered an unlikely source of ophiolite as westerly derived coarse carbonate breccias occur on both sides of White Bay in the lower Paleozoic Coney Arm Group to the west (Lock, 1972) and in the Fleur de Lys sequence to the east. These breccias can be correlated (de Wit , pers. comm.) and preclude the presence of a 100 km wide basin of deposition. In addition the recognition of Grenville (?) gneissic basement east of White Bay (de Wit, pers. comm.) raises the possibility that gneissic basement may underlie the entire White Bay zone.

In this thesis a northeasterly continuation of the Baie Verte Zone is considered to be a likely source of the White Hills Peridotite Sheet. The source area of the lower sedimentary slices must have lain to the west of the ocean on strike with the western division of the Fleur de Lys Supergroup, and a lithostratigraphic comparison between these rock groups is given in Table 4.

Rock Units of the Lower Slices of the Hare Bay Allochthon.	Western Division of the Fleur de Lys Supergroup[1].
TREMADOCIAN Northwest Arm Formation Black shales, minor limestones, limy siltstones.	PRE-LOWER ORDOVICIAN Graphitic Schists Marble breccias
PRE-ORDOVICIAN Maiden Point Formation Greywackes, slates, quartz pebble conglomerates, mafic volcanic flows, and minor ultramafic intrusions.	Thick psammitic schists, quartz pebble greywackes, with syn-sedimentary mafic igneous intrusions, minor ultramafic intrusions and a basal conglomerate.

[1]Description after Church (1969); Kennedy (1973a), and de Wit (pers. comm.).

Table 4. Comparison of rock units of the lower allochthonous slices with the western division of the Fleur de Lys Supergroup.

TECTONIC EVOLUTION OF THE HARE BAY ALLOCHTHON

The evolution of the Hare Bay Allochthon is described in terms of plate tectonics and is related to the birth and destruction of a continental margin, a proto-ocean and a marginal ocean basin from late Precambrian to upper Paleozoic times. The model draws heavily from present day analogues of developing continental margins, for example the present day Atlantic, and of the destruction of oceanic crust and mantle by subduction and obduction, for example, at Troodos, Papua and Oman.

Any model for the evolution of the Hare Bay Allochthon must take into account the following:

1. The metamorphic evidence from the aureole to the White Hills Peridotite Sheet suggests obduction of 'hot' ophiolite.
2. The possibility that the Tremadocian Cape Onion Volcanics represent the detached basaltic carapace of the White Hills Ophiolite.
3. Evidence from other on land ophiolites suggests that some ophiolites form only 30-50 my prior to thrusting.
4. The sedimentological evidence from the St. Julien Island conglomerates suggests derivation from an island arc.
5. The most satisfactory method of ophiolite obduction appears to be the result of an abortive subduction of a continental margin. For western Newfoundland, this requires a lower Ordovician, easterly dipping, Benioff Zone.

The model proposed here is an attempt to explain the geological facts derived from the allochthonous rocks. By necessity, it involves the tectonics of the Burlington Peninsula, however, it is not intended as a final or an authorative view on this controversial area.

Birth of the Proto-Atlantic Ocean

After the Grenvillian orogeny (900 my) North America and Europe formed a single plate. In late Precambrian times this plate began to split by a slow prolonged phase of crustal extension with the formation of a series of graben and the intrusion of dyke swarms into the gneisses of the Long Range Complex. The dykes give K-Ar ages of 805 ± 35 my (Pringle et al., 1971) and are chemically distinct (D.F. Strong, pers. comm.) from the overlying flows of the Lighthouse Cove Formation that they were once thought to feed (Strong and Williams, 1971).

The age of the final split, the initiation of the proto-Atlantic ocean, and the formation of an Atlantic-type continental margin in western Newfoundland, are all uncertain. The Eo-Cambrian (?) flood basalts of the Lighthouse Cove Formation have been compared with the Paleocene flood basalts of northwest Scotland (Dewey and Bird, 1971) and may mark initiation of sea floor spreading. Kennedy (1973,b) however, relates the 805 my date from the diabase dykes in the Long Range Complex to formation of the proto-Atlantic ocean, and relates the Lighthouse Cove basalts to formation of an early Cambrian (?) marginal ocean basin at Baie Verte. If the transported ophiolites formed in the Baie Verte marginal ocean then an Eo-Cambrian age for their formation appears to be too old. For this reason the Lighthouse Cove basalts, the chemically similar Maiden Point Volcanics, and the mafic igneous rocks in the basal parts of the western division of the Fleur dy Lys Supergroup, are all believed to relate to initiation of axial plate accretion in the proto-

Atlantic ocean in Eo-Cambrian times. At the same time a thick, monotonous sequence of Atlantic type (Reading, 1972) quartzo-feldspathic flysch spread eastwards from the Grenville basement into a westerly transgressive sea and was deposited on the continental rise,(for example, the basal psammites of the Fleur de Lys and possibly also the Maiden Point greywackes.)

In the proto-Atlantic ocean, oceanic crust and mantle was being generated at a spreading ridge probably in a manner described by Dewey and Bird (1971). It is unlikely that the allochthonous ophiolites formed at that time (late Precambrian) in the main ocean because of the resultant long time interval between their formation and middle Ordovician thrusting (about 130 my).

Establishment of the Carbonate Bank

By middle Cambrian times carbonate deposition had been established along the eastern margin of the Long Range initially represented by the Forteau, Cloud Rapids and Traytown Pond Formations at Canada Bay. This eventually cut off the westward supply of coarse terrigenous detritus to the continental rise area and instead, carbonate flysch and carbonate breccias spread eastwards. The allochthonous Irish Limestone, Cow Head Breccias, Cooks Brook Formations formed at this time. In the western division of the Fleur de Lys Supergroup, a varied clastic succession that includes greywackes, shales and carbonate breccias was deposited.

By Tremadoc times the Hare Bay Allochthon source area no longer received abundant coarse carbonate flysch possibly due to the formation of a barrier at shelf edge. Instead the black graptolitic shales and rare carbonate flysch beds of the Northwest Arm Formation accumulated.

Development of a Westerly Dipping Subduction Zone, Production of an Island Arc and a Marginal Ocean Basin.

In late Cambrian times (?) a westerly dipping subduction zone initiated in central Newfoundland east of the Burlington Peninsula (see also Bird and Dewey, 1970, fig. 8B; Kennedy, 1973a). The resultant island arc, the Cape St. John Group, built up on oceanic crust (Dewey and Bird, 1971) and these rocks constitute the eastern division of the Fleur de Lys Supergroup. Behind the arc a marginal ocean basin initiated in the Baie Verte area to the east of the already established continental rise prism (Fleur de Lys psammites). Oceanic crust and mantle was generated in the marginal basin and is probably represented by the Baie Verte ophiolite, the Mings Bight ophiolite, the White Hills Peridotite Sheet and the Bay of Islands Complex. At the western margin of this ocean mafic volcanic rocks, tuffs, and agglomerates, with an occasional influx of coarse greywackes from the west, were accumulating and represent the Goose Cove Formation.

The regional tectonic fabric present in the lower part of the White Hills Peridotite Sheet is interpreted as mantle fabric produced either by syn-tectonic recrystallization (Ave 'Lallemant and Carter, 1970) or by plastic deformation (Nicolas et al., 1971). The fabric is locally associated with recumbent folds and both folds and fabric are cut by late orthopyroxite dykes suggesting that the deformation preceded final solidification of the mantle.

Accumulating gravity (Milsom, 1973) and stratigraphic data (Kairg, 1972) on other on-land ophiolite suites favors a history of formation in marginal basins at the rear of island arcs. Interpretation of ophiolites as on-land marginal basin crust satisfactorily explains their large gravity fields (Milsom, 1973), their great thickness, their close age relationships between formation and thrusting, and the requirement of 'hot' ophiolite for aureole formation.

Upadhyay (pers. comm.) also favors an origin of the Snooks Arm ophiolite in a marginal ocean basin. However, it is uncertain whether it formed in relation to the same subduction zone as the Baie Verte ophiolite or if it resulted from later eastward subduction of the Baie Verte ocean.

Methods of emplacement of slabs of oceanic crust and mantle over continental margins are poorly understood. However, many workers favor emplacement by means of a collision between a continental margin and a subduction zone (Temple and Zimmerman, 1969; Zimmerman, 1972; Davies,1971; Moores, 1970; Dewey and Bird, 1971; Moores and MacGregor, 1972), whereby the leading edge of the oceanic plate is thrust over the partially subducted continental margin. To cause obduction in this manner in western Newfoundland, there must have been a reversal in arc polarity (Karig, 1972) whereby a new subduction zone initiated along the rear edge of the Cape St. John Island Arc. Reversals in arc polarity have been described from Papua (Karig, 1972) but the process is poorly understood and the cause unknown.

Subduction of the Baie Verte marginal ocean basin eventually resulted in the collision of the continental margin with the active arc system in lower Ordovician times. This collision is believed to mark the initiation of the Fleur de Lys deformation and the first pre-emplacement deformation in the allochthonous rocks.

In the case of the Hare Bay Allochthon the slab of ophiolite (White Hills Peridotite Sheet) was thrust over previously undeformed supracrustal rocks (Goose Cove Formation) that lay at the edge of the continental margin. The leading edge of this overthrust slab is unlikely to contain any sign of island arc igneous activity as volcanism within 30 km of a trench is virtually unknown, even where the Benioff zone dips steeply.

The first regional deformation and metamorphism (D_{P1}) of the Goose Cove Formation probably relates to initial ophiolite displacement. A period of static growth marked by garnet and plagioclase porphyroblasts followed. Further movements, possibly related to final displacement, produced the D_{P2} deformation and metamorphism. The associated westward facing recumbent folds imply westerly displacement of the ophiolite assuming there has been no rotation of the slice during subsequent transport. Kennedy (1973a) also noted a similar facing direction of recumbent folds in the western division of the Fleur de Lys Supergroup.

The history of the aureole rocks records an increase in the intensity of deformation and grade of metamorphism towards the contact with the ophiolite sheet. The metamorphic grade increases over approximately 300 m from low greenschist at the base up to pyroxene-amphibolite at the top and likewise the intensity of deformation increases from a crenulation cleavage to a mylonite. The aureole is viewed, therefore, as a dynamothermal aureole related to obduction of hot ophiolite onto a continental margin. Metamorphism of the basal aureole to the Mount Albert ophiolite occurred in early Ordovician times (MacGregor, 1962) and a similar age of metamorphism is likely for the Goose Cove Formation.

The partially subducted continental crust, which originally "plugged" the subduction zone, later rose isostatically under the overthrust ophiolite sheet. It caused a large structural high that eventually resulted in detachment of the leading edge of the ophiolite from the now inactive island arc to the east. From this upwelling highland, island arc type (Reading, 1972) flysch spread westwards in lower Ordovician times to form the St. Julien Island conglomerates and the Blow me Down Brook Formation, and the latter contains both island arc and ophiolite detritus (Stevens, 1970).

Formation of Gravity Slides, Emplacement of the Allochthon and the Taconic Orogeny.

In lower to middle Ordovician times the structural high became unstable. The ophiolite slice detached at a new structural level below its welded metamorphic aureole so that this structural couple comprised a distinct slice that moved westwards by gravity sliding. Detachment of the lower structural slices occurred at this time either by peel thrusting (Stevens, 1970) below the ophiolite slice or by gravity sliding. The present stacking order of the slices relates to their position of initial detachment whereby the higher slices originated farther from the emplacement site and are the farthest travelled.

The emplacement of the allochthon was heralded by deposition of the middle Ordovician Goose Tickle Flysh in the deepening autochthonous basin. This northeasterly derived flysch contains detritus that can be matched with most of the slices of the allochthon (Stevens, 1970).

The lowermost, Northwest Arm, Slice contains slumped and rubbly beds and was emplaced in a semi-consolidated condition. Conglomerate beds near the top of the Goose Tickle Formation contain clasts derived almost entirely from the Northwest Arm Formation indicating active erosion of this slice during transport.

The Maiden Point, Croque Head and St. Julien Island Slices display rare syn-emplacement recumbent folds but a widely developed slaty cleavage, indicating emplacement and deformation in a well indurated state. Low greenschist regional metamorphism accompanied the emplacement deformation. The syn-emplacement (D_E) recumbent folds in the Maiden Point Slice face up to the north and north-west implying transport in that direction.

During emplacement, the slices traversed across a basin of unconsolidated black shales and limy siltstones that probably represent an easterly facies of the Northwest Arm Formation. Blocks of shale and siltstone, together with blocks shed from the advancing slices, were tectonically incorporated into the imbricate thrust zones between the slices to form mélange zones. The mélange zones formed early in the evolution and transport of the slices as they show similar syn-emplacement deformation and metamorphism as the lower slices. However, the deformation and metamorphism of the Goose Cove Formation of the White Hills Slice clearly preceded mélange formation as blocks of greenschist occur in the basal mélange to that slice.

Final emplacement of the allochthon occurred in middle Ordovician times (Rodgers, 1965; Stevens, 1970). The directly underlying autochthonous rocks are, in most places, unaffected by the emplacement deformation, apart from minor igneous intrusions. At Canada Bay, where mélange is absent and the allochthon contact is a hard thrust, stress transmittance into the underlying rocks was possible, and as a result (D_E) northwest-facing recumbent folds and imbricate structure were developed.

The emplacement deformation (D_E) is correlated with the middle Ordovician "Taconic Orogeny" or Humberian orogeny of Bird and Dewey (1970). Effects of this orogeny are not solely restricted to areas of allochthonous rocks but have also been recognized in the lower Paleozoic Coney Arm Group at the head of White Bay (Lock, 1972) and appear to be of regional extent along the western margin of the Appalachians (Rodgers, 1967; Poole, 1967; Zen, 1968).

Final Closure of the Proto-Atlantic Ocean and the Acadian Orogeny

After the Baie Verte subduction zone closed in early Ordovician times a new subduction zone developed to the east in the

Notre Dame Bay area. Its direction of dip is controversial but preliminary geochemical evidence suggests it also dipped to the east (D.F. Strong, pers. comm.). Final closure of the proto-ocean in southwest Newfoundland (P. Brown, pers. comm.) locked the system in Devonian times and resulted in the Acadian Orogeny.

The Acadian deformation produced open upright folds and a crenulation cleavage that overprint the earlier pre- and syn-emplacement structures. The intensity of the Acadian deformation dies out rapidly west of the allochthon-autochthon contact. In the thesis area, little or no metamorphism accompanied the Acadian deformation. The zone of late thermal metamorphism in the south of the area post-dates the Acadian structures and is probably related to a late Acadian granite.

REFERENCES

Allen, J.R.L. (1965) - A review of the origin and characteristics of recent alluvial sediments; Sedimentology, v. 5, p. 91-191.

Ave'Lallemant, H., and Carter, N.L. (1970) - Syntectonic recrystallization of olivine and modes of flow in the upper mantle; Geol. Soc. American Bull., v. 81, p. 2203-2220.

Bailey, E.H., Blake, M.C., Jr., and Jones, D.L. (1970) - On-land Mesozoic oceanic crust in California Coast Ranges; U.S. Geol. Survey Prof. Paper 700-C, p. C70 - C81.

Baird, D.M., (1951) - The Geology of Burlington Peninsula, Newfoundland; Canada Geol. Survey Paper 51-21, p. 70.

Balk, Robert (1953) - Structure of greywacke areas and Taconic Range, east of Troy, New York; Geol. Soc. America Bull., v. 64, p. 811-864.

Betz, F. Jr. (1939) - Geology and Mineral Deposits of the Canada Bay area, Northern Newfoundland; Newfoundland Geol. Surv. Bull. No. 16, 53 p,

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

Bird, J.M., Dewey, J.F., and Kidd, W.S. (1971) - Proto-Atlantic Oceanic Crust and Mantle: Appalachian/Caledonian Ophiolites; Nature, Phys. Sci., v. 231, p. 28-31.

Bouma, A.H. (1962) - Sedimentology of some flysch deposits. A graphic approach to facies interpretation; Elsevier Pub. Co., Amsterdam, 168 p.

Bruckner, W.D. (1966) - Stratigraphy and Structure of West Central Newfoundland; in Poole, W.H. (editor); Guidebook, Geology of Parts of Atlantic Provinces; Ann. Meeting., Geol. Assoc. Can., Min. Assoc. Can., p. 137-151.

Cann, J.R. (1970) - Rubidium, Sr, Y, Zr, and Nb in some ocean floor basaltic rocks; Earth Planet. Sci. Lett., v. 10, p. 7-11.

Carrara, A. and Fyson, W.K. (1973) - Taconic and Acadian folds in northern and western Gaspe Peninsula, Quebec; Can. J. Earth Sci., vol. 10, p. 498-509.

Church, W.R. (1969) - Metamorphic rocks of the Burlington Peninsula and adjoining areas of Newfoundland, and their bearing on continental drift in North Atlantic; in Kay, M. (editor), North Atlantic - Geology and Continental Drift; Am. Assoc. Petrol. Geol. Mem. 12, p. 212-233.

Church, W.R., (1972) - Ophiolite: Its definition, origin as oceanic crust and mode of emplacement in orogenic belts, with special reference to the Appalachians; Pub. of the Earth Physics Branch - Dept. of Energy, Mines and Resources (Canada), v. 42, no. 3, p. 71-85.

Church, W.R. and Stevens, R.K. (1970) - Mantle Peridotite and early Paleozoic ophiolite complexes of the Newfoundland Appalachians (abstr.); Program and Abstracts, Internation. Symposium on Mechanical Properties and Processes in the Mantle, June 24 - July 3, Flagstaff, Arizona.

Church, W.R. and Stevens, R.K. (1971) - Early Paleozoic ophiolite complexes of the Newfoundland Appalachians as mantle-oceanic crust sequences; Journal Geophys. Res., v. 76, p. 1460-1466.

Clifford, P.M. and Baird, D.M. (1962) - Great Northern Peninsula of Newfoundland - Grenville Inlier; Canadian Inst. Mining and Met. Bull. No. 599, p. 150-157; 267-277.

Cloud, P.E., Jr. and Barnes, V.E. (1948) - The Ellenburger group of central Texas; Univ. Texas (Bur. Econ. Geology) Pub. No. 4621, 473 p.

Coleman, R.G. (1971) - Plate tectonic emplacement of upper mantle peridotites along continental edges; Jour. Geophys. Research, v. 76, p. 1212 - 1222.

Cooper, J.R. (1936) - Geology of the southern half of the Bay of Islands igneous complex; Newfoundland Dept. Nat. Res., Geol. Sec. Bull. No. 4.

Cooper, J.R. (1937) - Geology and Mineral Deposits of the Hare Bay area; Nfld. Dept. Nat. Res., Geol. Sec. Bull., no. 9, p. 36.

Davies, H.L. (1971) - Peridotite-gabbro-basalt complex in eastern Papua: An overthrust plate of oceanic mantle and crust; Australian Bur. Mineral Resources, Geology and Geophysics Bull. 128, 48 p.

Davies, H.L., and Smith, I.E. (1971) - Geology of Eastern Papua; Geol. Soc. America Bull., v. 82, p. 3299-3312.

Deer, W.A., Howie, R.A., and Zussman, J. (1966) - An introduction to the rock forming minerals; Longmans, London, 528 p.

Dewey, J.F. (1965) - Nature and origin of kink-bands; Tectonophysics, v. 6, p. 459-494.

Dewey, J.F. and Bird, J.M. (1970) - Mountain belts and the new global tectonics; Jour. Geophys. Res., v. 75, p. 2625-2647.

Dewey, J.F., and Bird, J.M. (1971) - Origin and emplacement of the ophiolite suite: Appalachian ophiolites in Newfoundland; Jour. Geophys. Res., v. 76, p. 3179-3206.

Dietz, R.S. (1963) - Collapsing continental rises: an actualistic concept of geosynclines and mountain building; J. Geol., v. 71, p. 314-333.

Erdtmann, B.-D. (1971) - Ordovician graptolite zones of western Newfoundland in relation to paleogeography of the north Atlantic; Geol. Soc. America Bull., v. 82, p. 1509-1529.

Fahraeus, L.E. (1970) - Conodont-based correlations of Lower and Middle Ordovician strata in western Newfoundland; Geol. Soc. America Bull., v. 81, p. 2061-2076.

Folk, R.L. (1968) - Petrology of sedimentary rocks; Hemphill's, Texas, 170 p.

Fleuty, M.J. (1964a) - The description of folds; Proc. Geologists Assoc., v. 75, part 4, p. 461.

Fleuty, M.J. (1964b) - Tectonic slides; Geol. Mag., v. 101, pp. 452-456.

Howley, J.P. (1917) - Report upon a geological exploration in the District of White Bay, 1902, 18 p. Geological Survey of Newfoundland Reports,1882-1906, St. John's.

Hsu, K.J. (1968) - Principles of mélanges and their bearing on the Franciscan-Knoxville paradox; Geol. Soc. America Bull., v. 79, p. 1063-1074.

Johnson, Helgi (1941) - Paleozoic lowlands of western Newfoundland; New York Acad. Sci. Trans., ser. 2, v. 3, p. 141-145.

Karig, D.E. (1972) - Remnant arcs; Geol. Soc. America Bull., v. 83, p. 1057-1068.

Kay, M. (1945) - Paleogeographic and palinspastic maps; Bull. Am. Assoc. Petroleum Geol., v. 29, p. 426-450.

Kay, M. (1967) - Stratigraphy and structure of northeastern Newfoundland: bearing on drift in North Atlantic; Bull. Am. Assoc. Petroleum Geol., v. 51, p. 579-600.

Kennedy, M.J. (1971) - Structure and stratigraphy of the Fleur de Lys Supergroup, Newfoundland; Geol. Assoc. Canada Proc. v. 24, no. 1, p. 59-71.

Kennedy, M.J. (1973a) - Pre-Ordovician polyphase structure in the Burlington Peninsula of the Newfoundland Appalachians; Nature, Phy. Sci., v. 241, p. 114-116.

Kennedy, M.J. (1973b) - The relationship between Pre-Ordovician polyphase deformation and ophiolite obduction in the Newfoundland Appalachian system; (abstr.); Geol. Soc. America, N.E. Sec. Meeting Program, Allentown, Penn. (March), p. 183.

Kerr, P.F. (1959) - Optical Mineralogy; McGraw-Hill, London, 442 p.

Kindle, C.H. and Whittington, H.B., (1965) - New Cambrian and Ordovician fossil localities in western Newfoundland; Geol. Soc. America Bull., v. 76, p. 683-688.

Knopf, E.B. (1931) - Retrogressive Metamorphism and phyllonitization; Am. J. Sci., v. 21, p. 1-27.

Krumbein, W.C. and Sloss, L.L. (1963) - Stratigraphy and Sedimentation; Freeman, London, 660 p.

Ksiaziewicz, M. (1958) - Submarine slumping in the Carpathian Flysch; Ann. soc. geol. Pologne, v. 28, p. 123-150.

Laurent, Roger, (1973) - The Thetford-Mines ophiolite, Paleozoic "flake" of oceanic lithosphere in the Northern Appalachians of Quebec (abstr.); Geol. Soc. America, N.E. Sec. Meeting Program, Allentown, Penn. (March), p. 188.

Lock, B.E. (1972) - Lower Paleozoic history of a critical area; Eastern margin of the St. Lawrence Platform in White Bay, Newfoundland, Canada; Internat. Geol. Cong., 24th Sess. Sec. 6, p. 310-324.

Macdonald, G.A. and Katsura, T. (1964) - Chemical composition of Hawaiian lavas; J. Petrol., v. 5, p. 82-133.

MacGregor, I.D. (1962) - Geology, petrology and geochemistry of the Mount Albert and associated ultramafic bodies of Central Gaspe, Quebec; Unpublished M.Sc. thesis, Queen's University, Kingston, Ontario.

Milsom, J. (1973) - Papuan ultramafic belt: Gravity anomalies and the emplacement of ophiolites; Geol. Soc. America Bull., v. 84, p. 2243-2258.

Moores, E.M., (1970) - Ultramafics and orogeny, with models of the U.S. Cordillera and the Tethys; Nature, v. 228, p. 837-842.

Moores, E.M., and Vine, F. (1971) - Troodos Massif, Cyprus, and other ophiolites as oceanic crust: evaluation and implications; Royal Soc. London Philos. Trans., Ser. A, v. 268, p. 443 - 466.

Moores, E.M. and MacGregor, I.D. (1972) - Types of Alpine ultramafic rocks and their implications for fossil plate interactions; in Shagam, R., (ed.); Studies in Earth and Space Sciences; Geol. Soc. America Memoir 132, p. 209-223.

Murray, A.M. and Howley, J.P. (1881) - Geological Survey of Newfoundland from 1864-1880; London, Stanford, 536 p.

Neale, E.R.W., and Kennedy, M.J. (1967) - Relationships of the Fleur de Lys group to younger groups of the Burlington Peninsula, Newfoundland; in Geology of the Atlantic Region; Neale, E.R.W. and Williams, H. (editors); Geol. Assoc. Canada, Spec. Paper No. 4, p. 139-169.

Nicholas, A., Bouchez, J.L., Boudier, F., and Mercier, J.C. (1971) - Textures, Structures and Fabrics due to solid state flow in some European lherzolites; Tectonophysics, v. 12, p. 55-86.

Pettijohn, F.J. (1957) - Sedimentary rocks; Harper and Row, New York, 718 p.

Poole, W.H. (1967) - Tectonic evolution of Appalachian region of Canada; Geol. Assoc. Canada Spec. Paper 4, p. 9-51.

Pringle, I.R., Miller, J.A. and Warrell, D.M. (1971) - Radiometric age determinations from the Long Range Mountains, Newfoundland; Can. J. Earth Sci., v. 8, p. 1325-1330.

Reading, H.G. (1972) - Global tectonics and the genesis of flysch successions; Internat. Geol. Cong., 24th Sess. Sec. 6, p. 59-66.

Reinhardt, B.M. (1969) - On the genesis and emplacement of ophiolites in the Oman Mountains geosyncline; Schweiz, Mineral Petrog. Mitt., v. 49, p. 1-30.

Rodgers, J. (1965) - Long Point and Clam Bank Formations, western Newfoundland; Proc. Geol. Assoc. Canada, v. 16, p. 83-94.

Rodgers, J. (1967) - Chronology of tectonic movements in the Appalachian region of eastern North America; Am. Jour. Sci., v. 265, p. 408-427.

Rodgers, J. (1969) - The eastern edge of the North American continent during the Cambrian and early Ordovician; in Zen, E-an, White, W.S., Hadley, J.B., and Thompson, J.B. Jr., (eds.), Studies of Appalachian geology, Northern and Maritime: New York, Interscience Publ., p. 129-140.

Rodgers, J., and Neale, E.R.W. (1963) - Possible "Taconic" klippen in western Newfoundland; Am. J. Sci., v. 261, p. 713-730.

Schuchert, C., and Dunbar, C.O. (1934) - Stratigraphy of western Newfoundland; Geol. Soc. Amer., Mem. 1, 123 p.

Shackleton, R.M. (1958) - Downward-facing structures of the Highland border; Quart. J. Geol. Soc., v. 113, p. 361-388.

Smith, A.G. (1971) - Alpine deformation and the oceanic areas of the Tethys, Mediterranean, and Atlantic; Geol. Soc. America Bull., v. 82, p. 2039-2070.

Smith, C.H., (1958) - Bay of Islands Igneous Complex, western Newfoundland; Geol. Surv. Canada, Mem. 290, 132 p.

Smitheringale, W.G. (1972) - Low potash Lushs Bight tholeiites: ancient oceanic crust in Newfoundland?; Can. Jour. E. Sci., v. 9, p. 574-588.

Smyth, W.R., (1971) - Stratigraphy and structure of part of the Hare Bay Allochthon, Newfoundland; Geol. Assoc. Canada Proc. (A Newfoundland Decade), v. 24, no. 1, p. 47-57.

Snelgrove, A.K. (1931) - Geology and ore deposits of Betts Cove - Tilt Cove area, Newfoundland; Bull. Can. Inst. Mining, Met., No. 228, p. 477-519.

Spry, Alan (1969) - Metamorphic textures; Pergamon Press, London, 350 p.

Stevens, R.K., (1968) - Taconic klippen of western Newfoundland; Geol. Surv. Canada, Paper 68-1, Part A, p. 8-10.

Stevens, R.K. (1970) - Cambro-Ordovician flysch sedimentation and tectonics in west Newfoundland and their possible bearing on a proto-Atlantic ocean; Geol. Assoc. Canada, Spec. Paper No. 7, p. 165-177.

Stevens, R.K. and Church, W.R., and St.-Julien, P.,(1969) - Age of ultramafic rocks in the Northwestern Appalachians; Geol. Soc. Amer., Abstr. for 1969, pt. 7, p. 215-216.

Stevens, R.K. and Williams, H. (1973) - The emplacement of the Humber Arm Allochthon, Western Newfoundland; (abstr.); Geol. Soc. America, N.E. Sec. Mtg. Program, Allentown, Penn. (March), p.222.

Strong, D.F. (1972) - Sheeted diabases of central Newfoundland: New evidence for Ordovician sea-floor spreading; Nature, v. 235, p. 102-104.

Strong, D.F. and Williams, H., (1972) - Early Paleozoic flood basalts of northwestern Newfoundland: their petrology and tectonic significance; Geol. Assoc. Canada Proc., v. 24, no. 2, p. 43-54.

St. Julien, Pierre (1973) - Structural setting of the Thetford-Minis ophiolite complex (abstr.); Geol. Soc. America, N.E. Sec. Meeting Program, Allentown, Penn., March, p. 215.

Swett, K., Smit, D.E. (1972) - Cambro-Ordovician shelf sedimentation of western Newfoundland, northwest Scotland and central east Greenland; Internat. Geol. Cong., 24th Sess. Sec. 6, p. 33-41.

Temple, P.G., and Zimmerman, J. (1969) - Tectonic significance of alpine ophiolites in Greece and Turkey; Geol. Soc. America, Abstracts with Programs for 1969, Pt. 7 (Ann. Mtg.), p. 221-222.

Tuke, M.J. (1968) - Autochthonous and Allochthonous rocks in the Pistolet Bay area in northernmost Newfoundland; Can. J. Earth Sci., v. 5, p. 501-513.

Turner, F.J. (1938) - Progressive regional metamorphism in southern New Zealand; Geol. Mag., v. 75, p. 160-174.

Upadhyay, H.D., Dewey, J.F., and Neale, E.R.W., (1971) - The Betts Cove ophiolite complex, Newfoundland: Appalachian oceanic crust and mantle; Geol. Assoc. Canada Proc. (A Newfoundland Decade), v. 24, p. 27-34.

Walker, R.G. (1967) - Turbidite sedimentary structures and their relationship to proximal and distal depositional environments; J. Sed. Pet., v. 37, p. 25-43.

Whittington, H.B., and Kindle, C.H., (1963) - Middle Ordovician Table Head Formation, western Newfoundland; Geol. Soc. America Bull., v. 74, p. 745-758.

Williams, Harold (1964) - The Appalachians in northeastern Newfoundland - a two-sided symmetrical system; Am. J. Sci., v. 262, p. 1137-1159.

Williams, Harold (1971) - Mafic-ultramafic complexes in western Newfoundland Appalachians and the evidence for their transportation: a review and interim report; Geol. Assoc. Canada Proc. (A Newfoundland Decade), v. 24, p. 9-25.

Williams, Harold (in press) - Bay of Islands map-area, Newfoundland; Geol. Surv. Canada, Paper and final map.

Williams, H., and Stevens, R.K. (1969) - Geology of Bell Isle - northern extremity of the deformed Appalachian miogeosynclinal belt; Can. J. Earth Sci., v. 6, p. 1145-1157.

Williams, H., Kennedy, M.J. and Neale, E.R.W. (1970) - The Hermitage Flexure, the Cabot Fault, and the disappearance of the Newfoundland Central Mobile Belt; Geol. Soc. of America Bull., v. 81, p. 1563-1568.

Williams, Harold, and Malpas, John (1972) - Sheeted dikes and brecciated dike rocks within transported igneous complexes, Bay of Islands, western Newfoundland; Can. J. Earth Sci., v. 9, p. 1216-1229.

Williams, Harold, Malpas, John, and Comeau, R., (1972) - Bay of Islands Map-area, Newfoundland;Geol. Surv. Canada, Paper 72-1, Part A, p. 14-17.

Williams, Harold, Smyth, W.R. and Stevens, R.K., (1973) - Hare Bay Allochthon, Northern Newfoundland; Geol. Surv. Canada, Paper 73-1, Part A, p. 8-14.

Williams, P.F. (1972) - Development of metamorphic layering and cleavage in low grade metamorphic rocks at Bermagui, Australia; Am. J. Sci., v. 272, p. 1-47.

Wilson, G.B., (1961) - Tectonic significance of small scale structures and their importance in the field; Annal. Soc. geol. Belgium, 74, p. 423-546.

Zen, E-an (1967) - Time and space relationships of the Taconic Allochthon and Autochthon; Geol. Soc. America Spec. Paper 97, 107 p.

Zen, E-an (1968) - Nature of the Ordovician orogeny in the Taconic area; in Zen, E-an, White, W.S., Hadley, J.B., and Thompson, J.B. Jr., (eds.) Studies of Appalachian geology, Northern and Maritime; New York, Interscience Publ., p. 129-140.

Zen, E-an (1972a) - The taconide Zone and the Taconic orogeny in the west part of the northern Appalachian orogen; Geol. Soc. America Spec. Paper 135, 72 p.

Zen, E-an (1972b) - Some revisions in the interpretation of the Taconic Allochthon in West-Central Vermont; Geol. Soc. America Bull., v. 83, p. 2573-2588.

Zimmerman, J., (1972) - Emplacement of the Vourinos Ophiolite Complex, Northern Greece; in Shagam, R., (ed.), Studies in Earth and Space Sciences; Geol. Soc. America Mem. 132, p. 225-239.

Zwart, H.F., (1960) - Relations between folding and metamorphism in the central Pyrenees and their chronological succession; Geologie Minjnb., v. 39, p. 163-180.

THE STRATIGRAPHY AND STRUCTURE OF THE SOUTHERN PART OF THE HARE BAY ALLOCHTHON, N.W. NEWFOUNDLAND

VOLUME 2

W. R. SMYTH

THE STRATIGRAPHY AND STRUCTURE OF THE SOUTHERN PART OF THE HARE BAY ALLOCHTHON, N.W. NEWFOUNDLAND

Volume 2

by

© W. R. SMYTH

MEMORIAL UNIVERSITY OF NEWFOUNDLAND

July, 1973

CONTENTS

VOLUME 2

PLATES

FIGURES

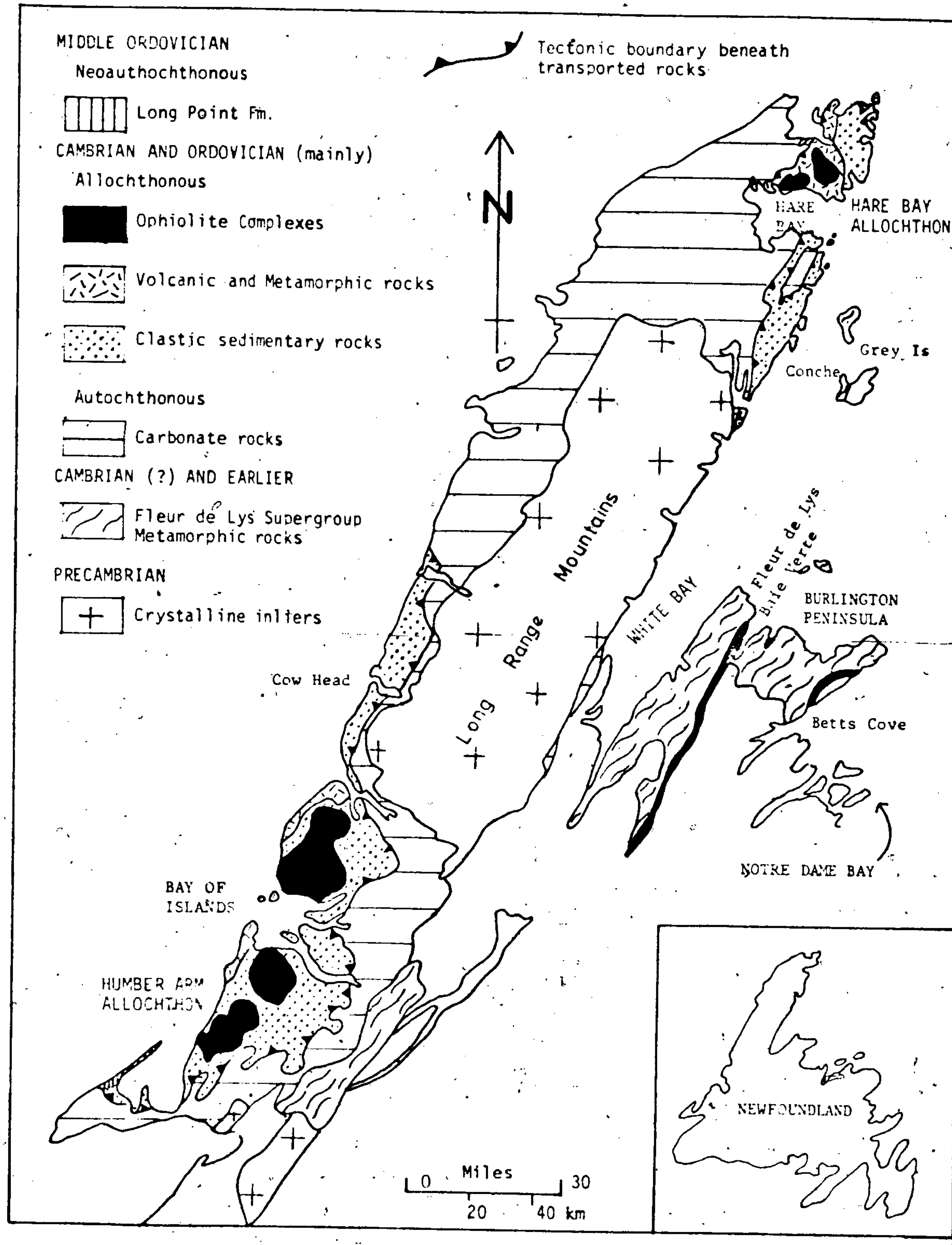


Figure I. Major tectonic elements of western Newfoundland

PLATE 1

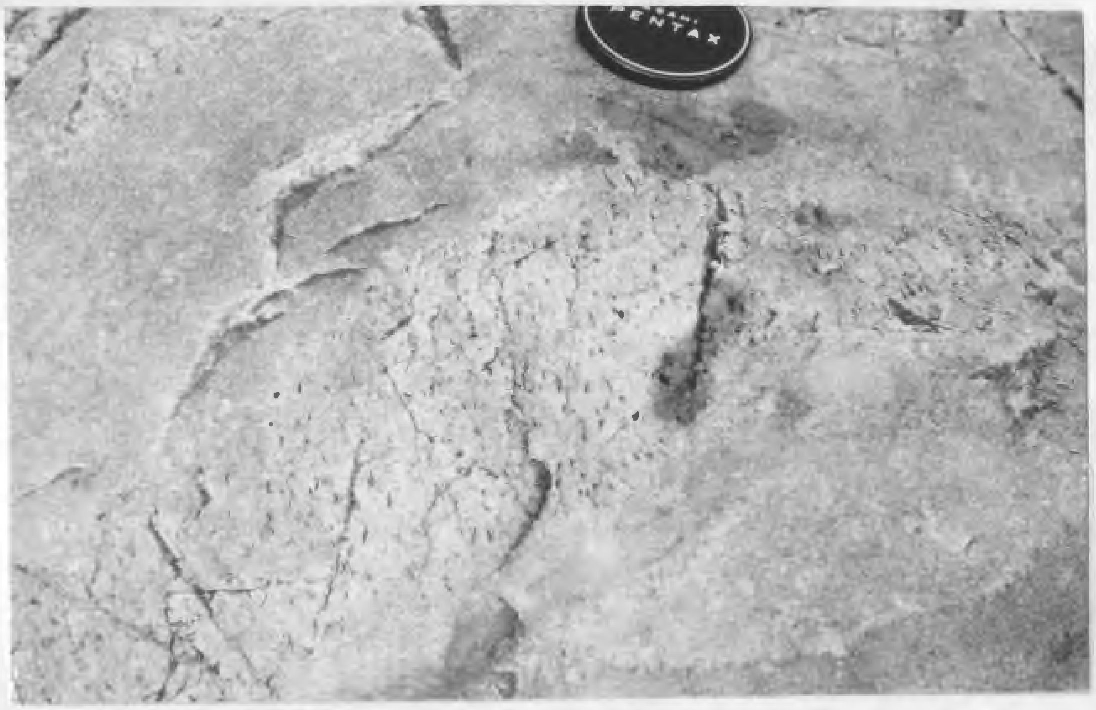
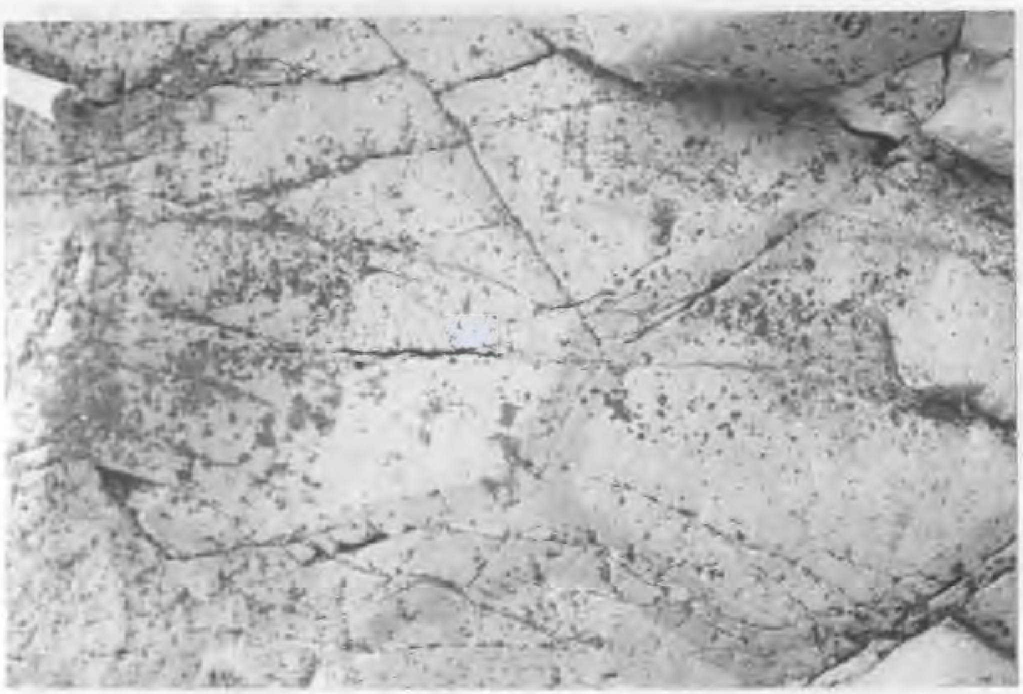


Fig. a: Inclusion of foliated gneiss of the Long Range Complex in microgranite. Seans Cove, White Bay.

b: Pillow basalt of the Lighthouse Cove Formation overlying thin basal conglomerate. Conglomerate unconformably overlies basement gneisses. South side of Otter Cove, Canada Bay.

PLATE 2

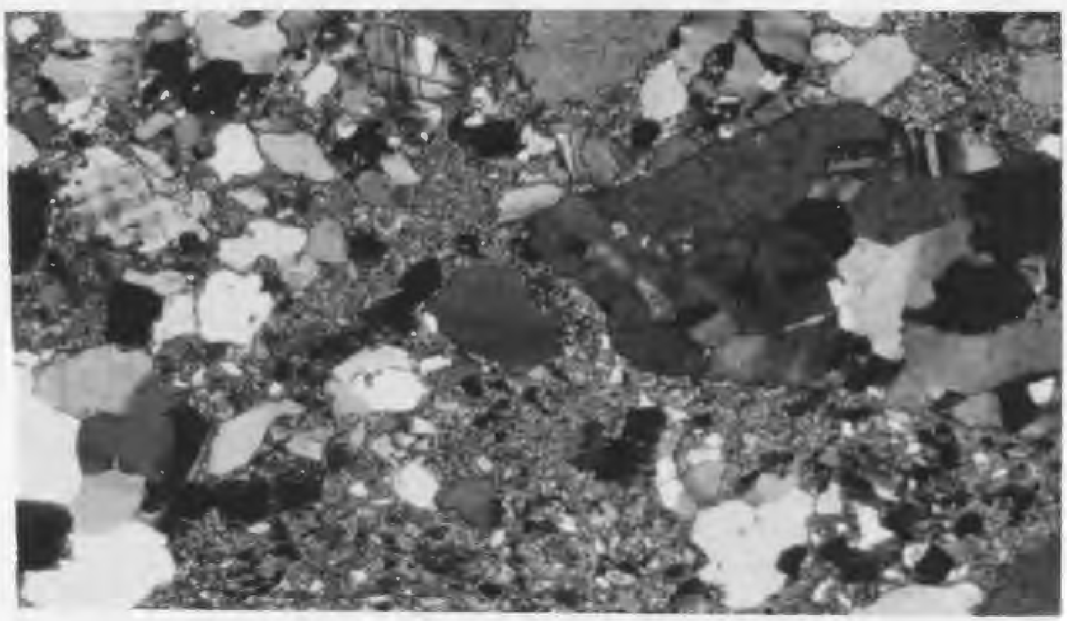
Fig. a: Pillow basalt of the Lighthouse Cove Formation. South side of Otter Cove.



b: Hexagonal cooling joints in basalts of the Lighthouse Cove Formation. Top of flow, south side of Otter Cove.

PLATE 3

Fig. a: Plutonic boulder conglomerate from the Lighthouse Cove Formation. West of Burnt Point, Canada Bay.



1 mm

b: Microgranite clasts in Bradore Formation. Note poor sorting and high matrix content. From Bradore unconformity with Long Range Complex, Seans Cove, White Bay.

PLATE 4

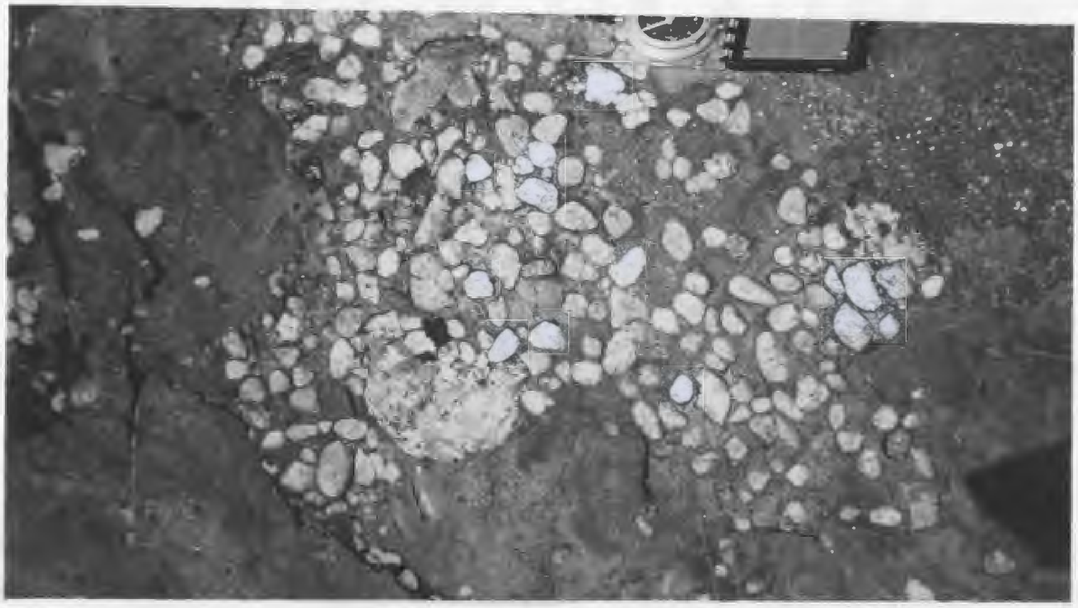
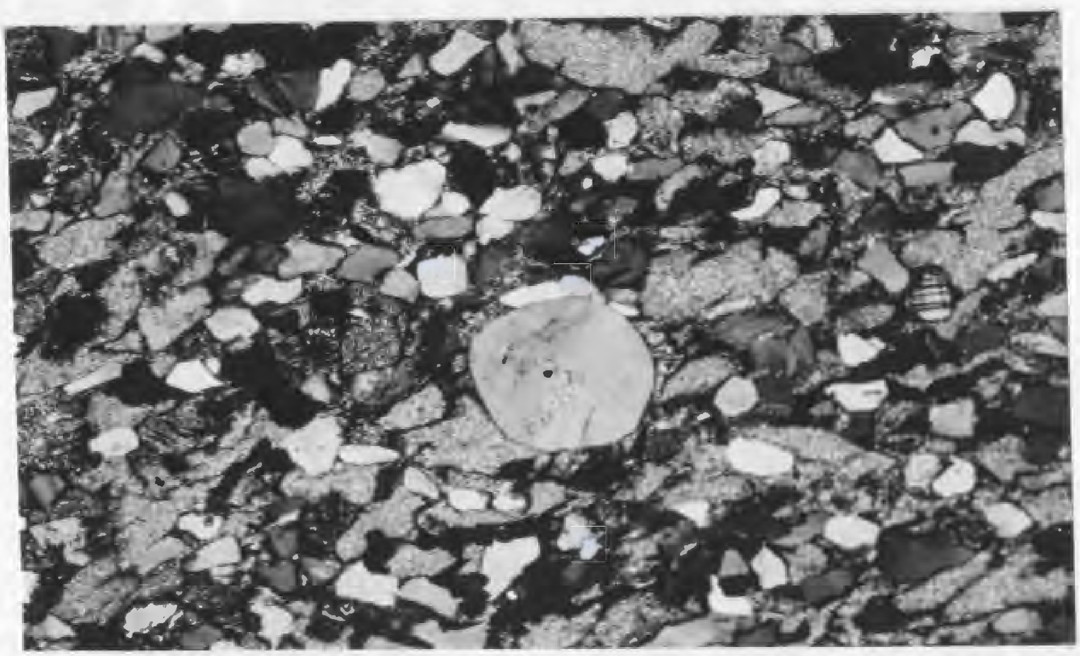


Fig. a: Pebble conglomerate, Bradore Formation. Northwest Shore of Wild Cove.

b: Pebble conglomerate consisting of hematite clasts and sparse quartz grains, Bradore Formation. Near top of formation, west side of Wild Cove.

PLATE 5

Fig. a: Trough cross-bedding in the Bradore Formation. Looking northeast, south of Fly Point, Canada Bay.



1mm

b: Photomicrograph of sandstone from the Bradore Formation. Note variety of roundness and sphericity. Altered feldspar grains are cloudy with indistinct grain boundaries. West shore of Wild Cove. Plane polarized light.

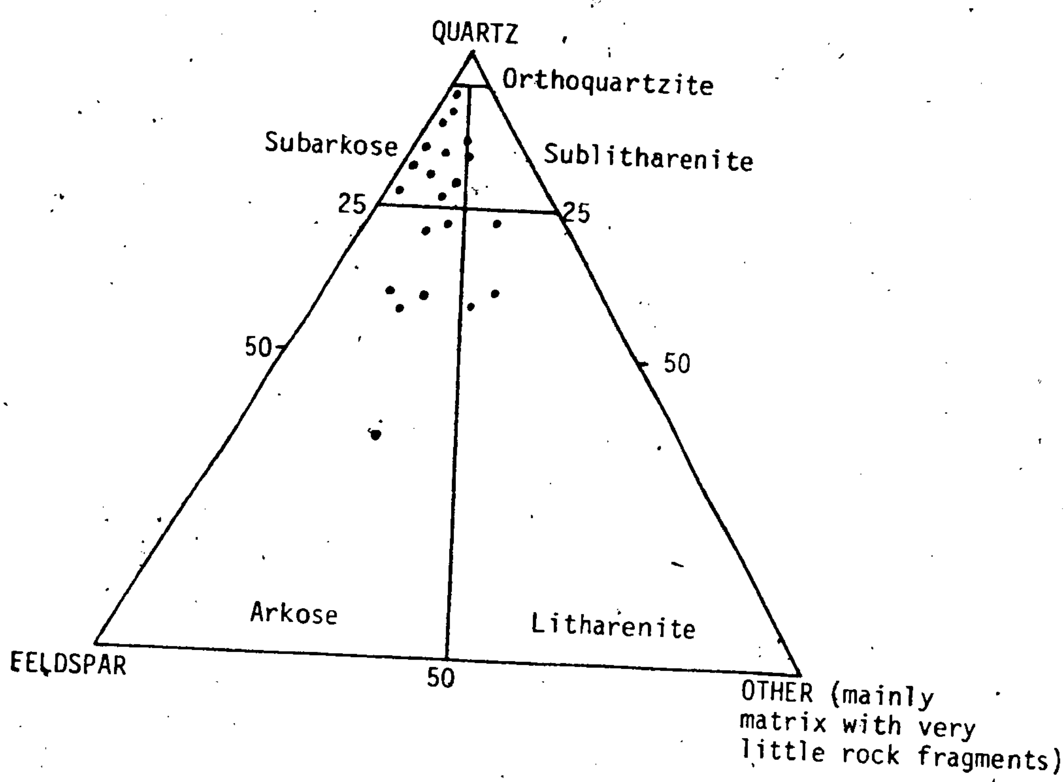


Fig 8: Compositional diagram of the Bradore Formation

PLATE 6

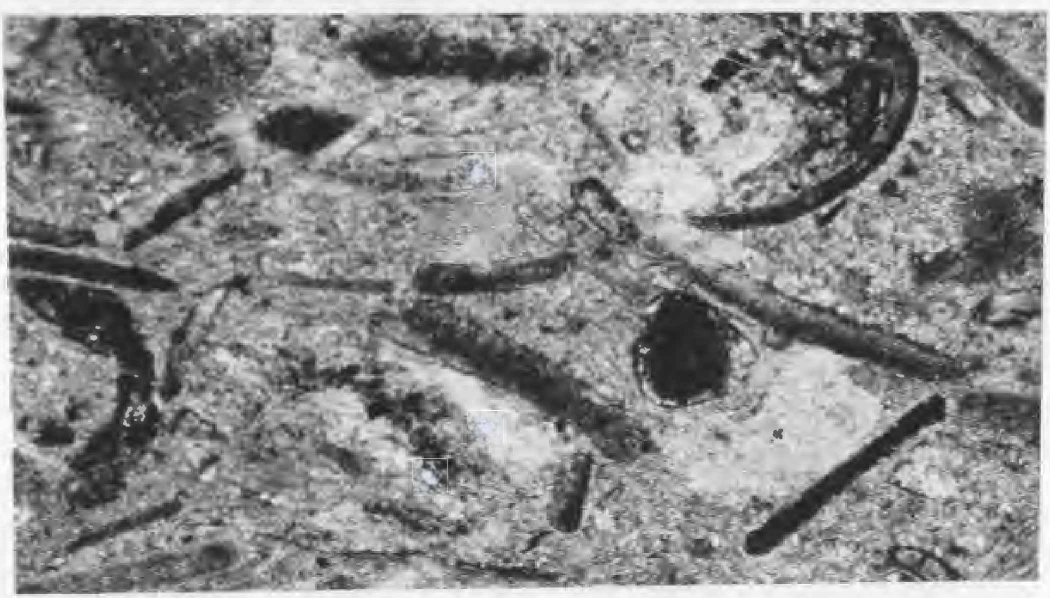
0.5 mm

Fig. a: Detrital hematite in orthoquartzite from the Bradore Formation. Note well rounded and sub-spherical clasts. Wild Cove. Partially polarized light.

b: Contact between Bradore Formation and the overlying Devils Cove Formation. Note sharp contact between black hematitic sandstones below, and lighter, variegated limestones and shales above. West shore of Wild Cove, Canada Bay.

PLATE 7

Fig. a: Faulted contact between Devils Cove Formation and Goose Tickle Formation. Contact is marked by 0.6 m of light coloured, mylonitized marble. Note hematitic sandstones of Bradore Formation (left) underlying limestones of Devils Cove Formation looking north, east side of Wild Cove.



0.5 mm

b: Photomicrograph of fossils in limestones from the Devils Cove Formation. Algae filaments are straight and brachiopod shell is curved. West shore of Wild Cove.

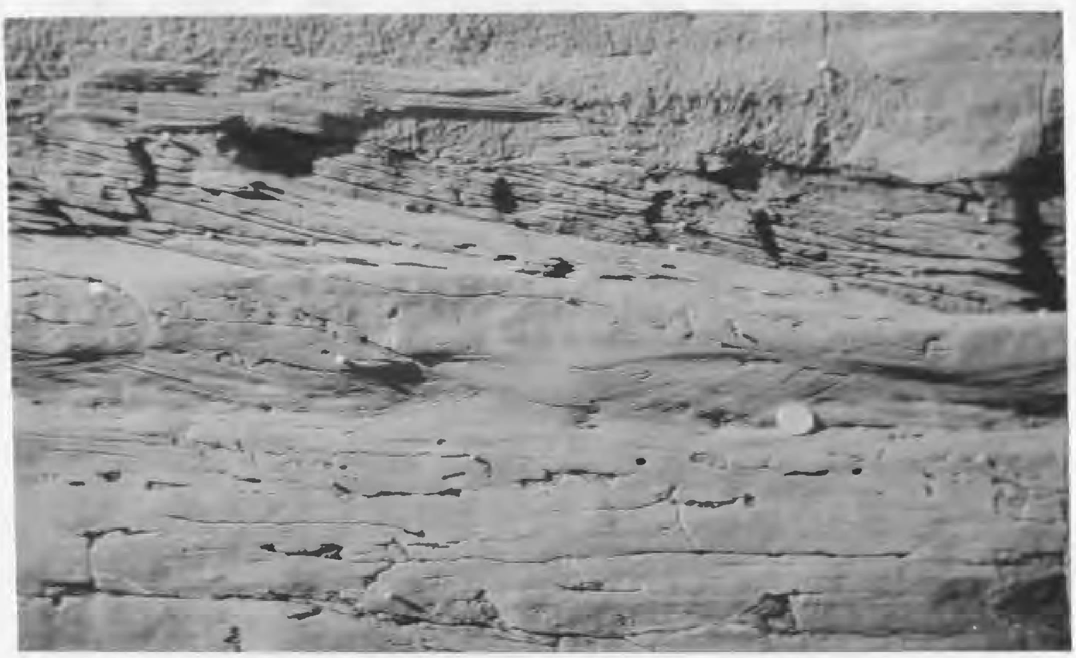
PLATE 8

Fig. a: Chert bed from upper unit of Table Head Formation. Note steeply dipping cleavage refracted in chert bed. Whites Arm Pond.

b: Conglomerate from Goose Tickle Formation. Note flattened clast. Southwest shore of Whites Arm.

PLATE 9

Fig. a: Flute casts, Goose Tickle Formation. From greywacke block in mélange, north end of MacGrey Island, Hare Bay. Note steeply dipping fracture cleavage.



b: Ripple drift lamination, Goose Tickle Formation. East side of MacGrey Island, Hare Bay.

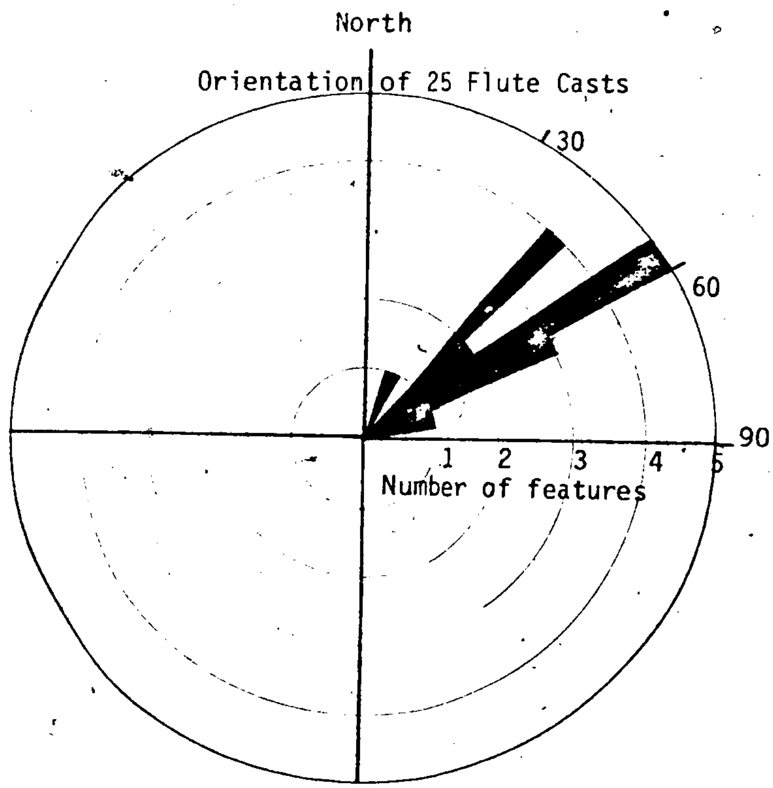
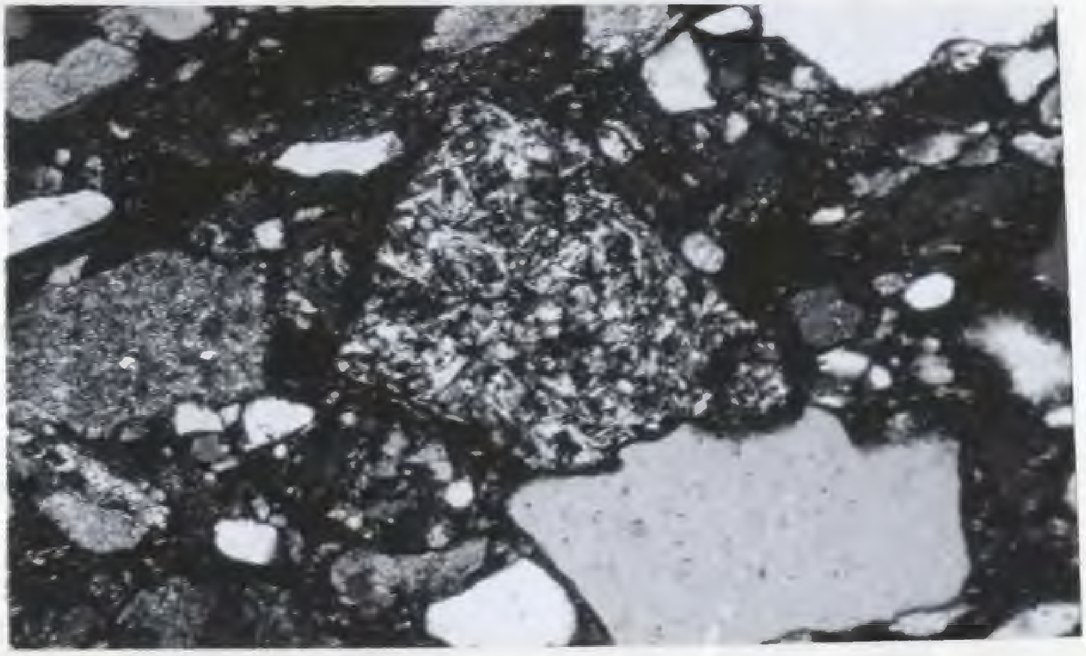


Fig 9: Directional features of the Goose Tickle Formation, Big and Little Springs Inlets, Hare Bay

PLATE 10

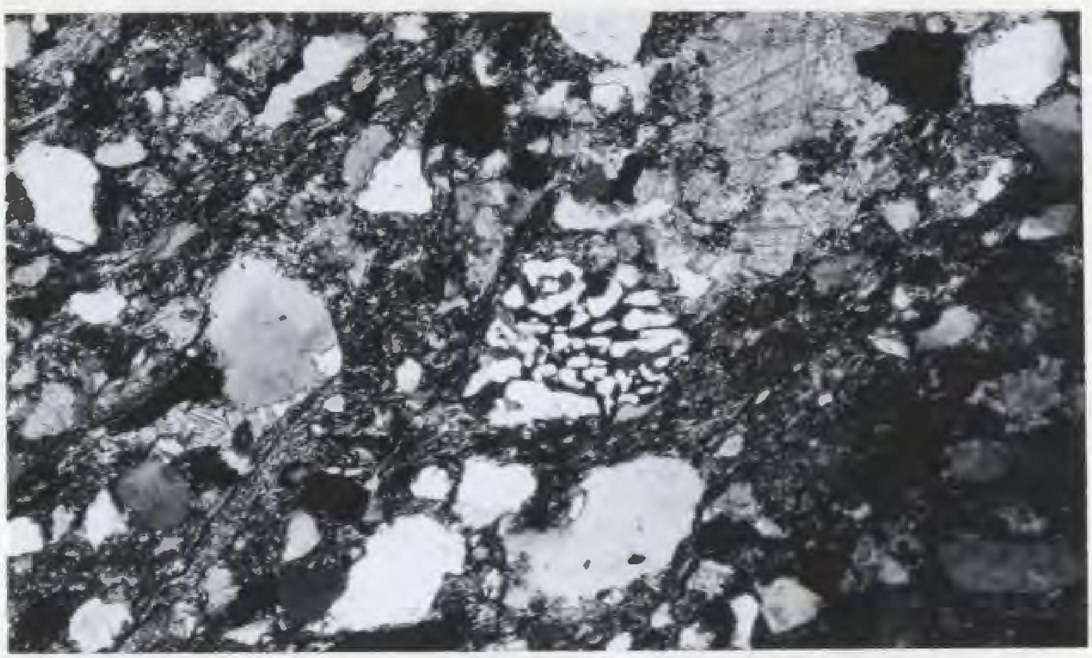
Greywacke bed, Goose Tickle Formation. Note graded lower interval and upper cross-laminated interval. East shore of Big Spring Inlet, Hare Bay.

PLATE 11



0.5 mm

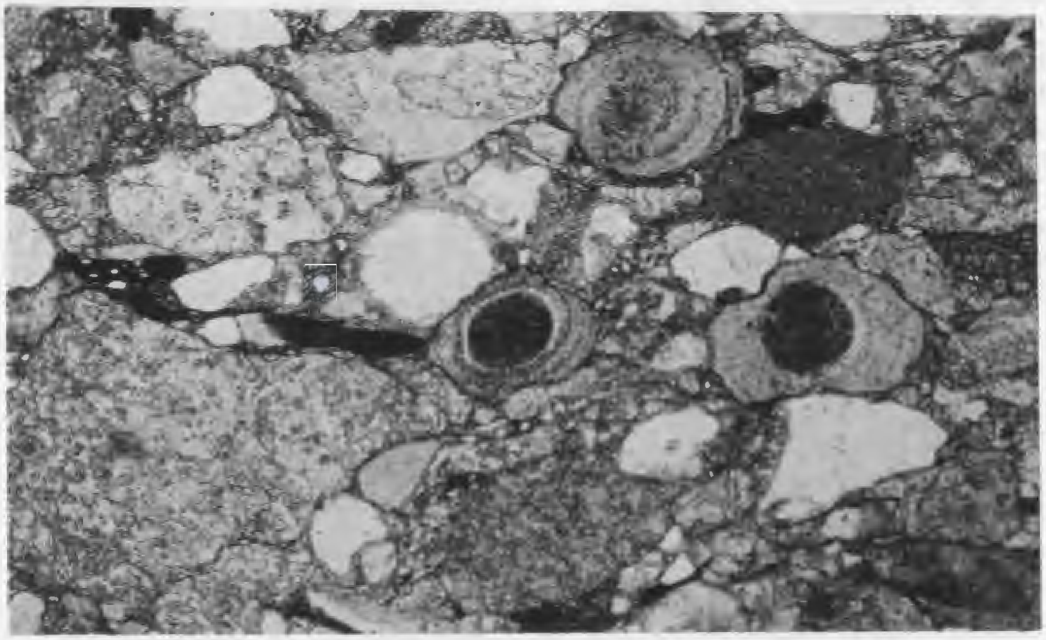
Fig. a: Diabase clast in greywacke from Goose Tickle Formation. Note angular shapes, poor sorting and high matrix content. Northwest shore of Whites Arm. Crossed nicols.



0.5 mm

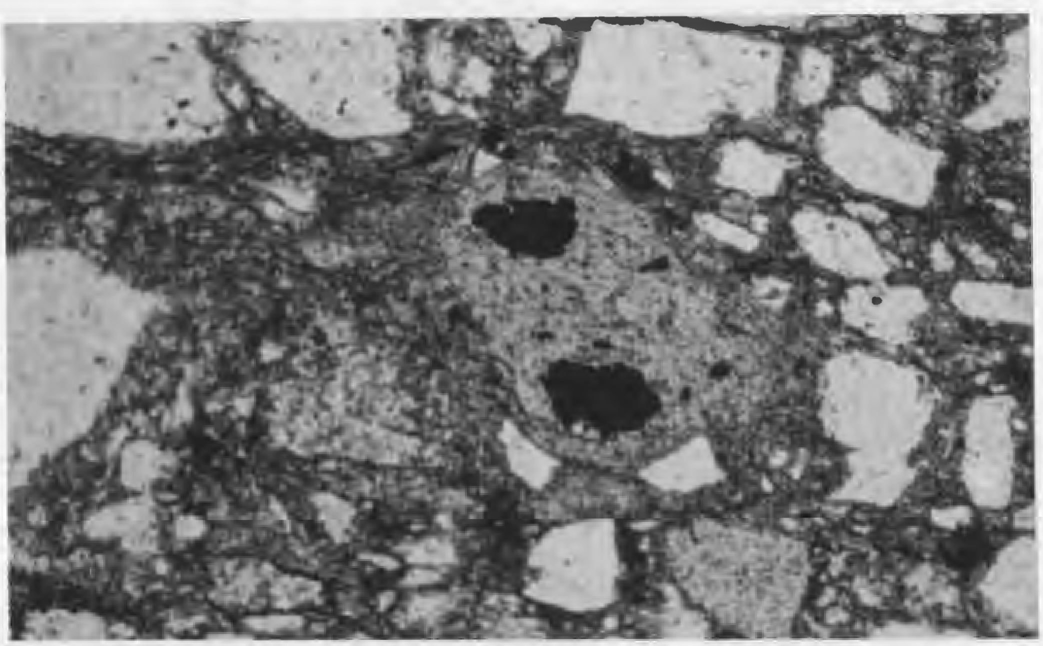
b: Granophyre clast in greywacke from Goose Tickle Formation. Note development of slaty cleavage. East shore of Big Spring Inlet. Crossed nicols.

PLATE 12



0.5 mm

Fig. a: Oolite clasts in greywacke from Goose Tickle Formation. Note partially broken concentric structure of the ooliths. East shore of Big Spring Inlet. Partially polarized light.



0.5 mm

b: Chlorite clast containing chromite, in greywacke from the Goose Tickle Formation. East shore of Big Spring Inlet. Obliquely crossed nicols

PLATE 13

Fig. a: Interbanded calc-silicate and pelitic schists, from Sugarloaf Schists. Calc-silicate bands are white. Note inclined folds and widely spaced strain-slip (S_2) cleavage. South side of Sugarloaf Cove, looking north.

b: Folded marble beds in pelitic and semi-pelitic schists, Sugarloaf Schists. Sugarloaf Cove.

PLATE 14

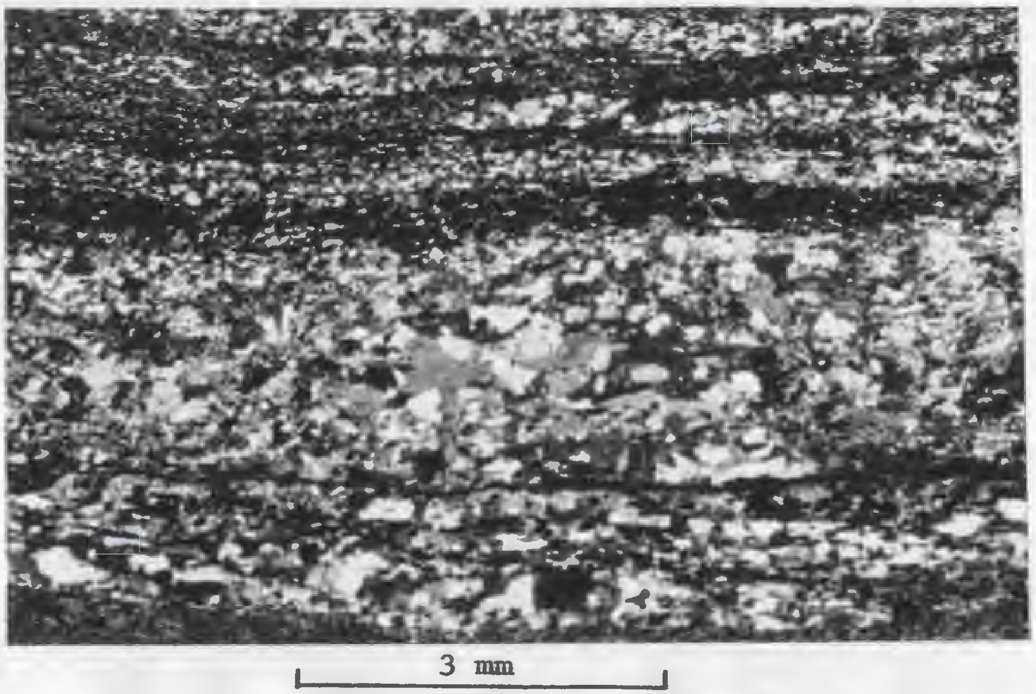
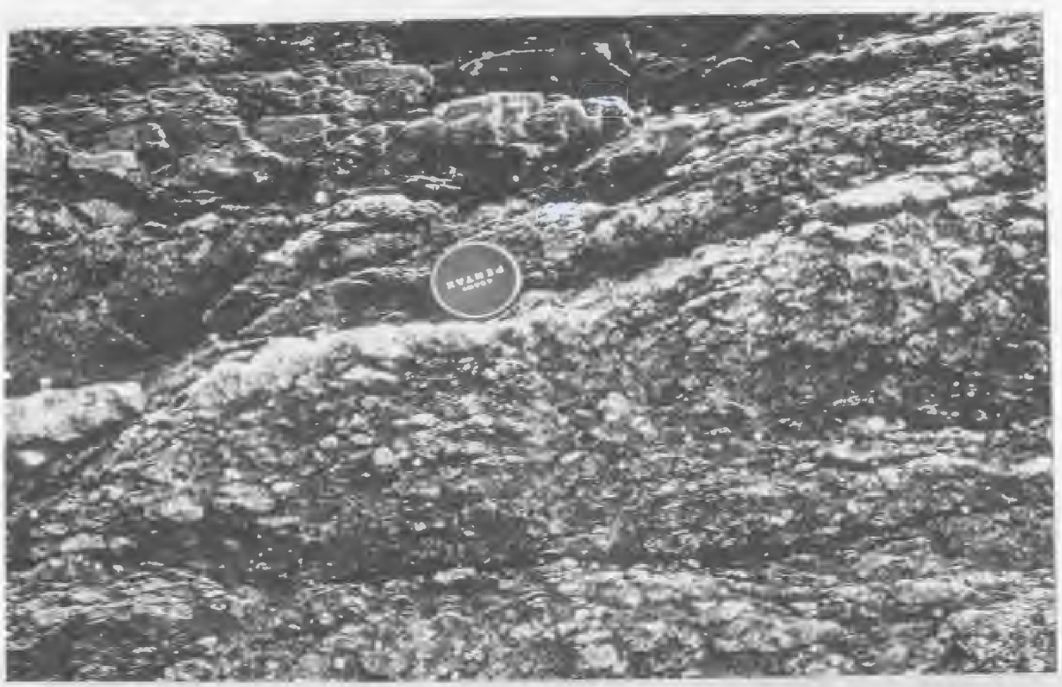


Fig. a: Photomicrograph of a quartzite band from the Sugarloaf Schists. Note the granoblastic texture in center of band and preserved schistoxity (S_1) at the margins where micas are interleaved with quartz. Sugarloaf Cove. Crossed nicols.



b: Highly deformed, schistose granite, Sugarloaf Slide Zone. The schistosity (S_1) is defined by thin zones of phyllonite that form augen around relict phenocrysts of feldspar and quartz. Southwest of Sugarloaf Cove.

PLATE 15

Fig. a: Kink style F_2 folds in phyllonitized granite, Sugarloaf Slide Zone. The phyllonite fabric (S_1) goes around the folds. Southwest of Sugarloaf Cove.

b: Foliated tectonic schists Sugarloaf Slide Zone. Pelitic foliae are dark and calc-silicate foliae light. Southwest of Sugarloaf Cove.

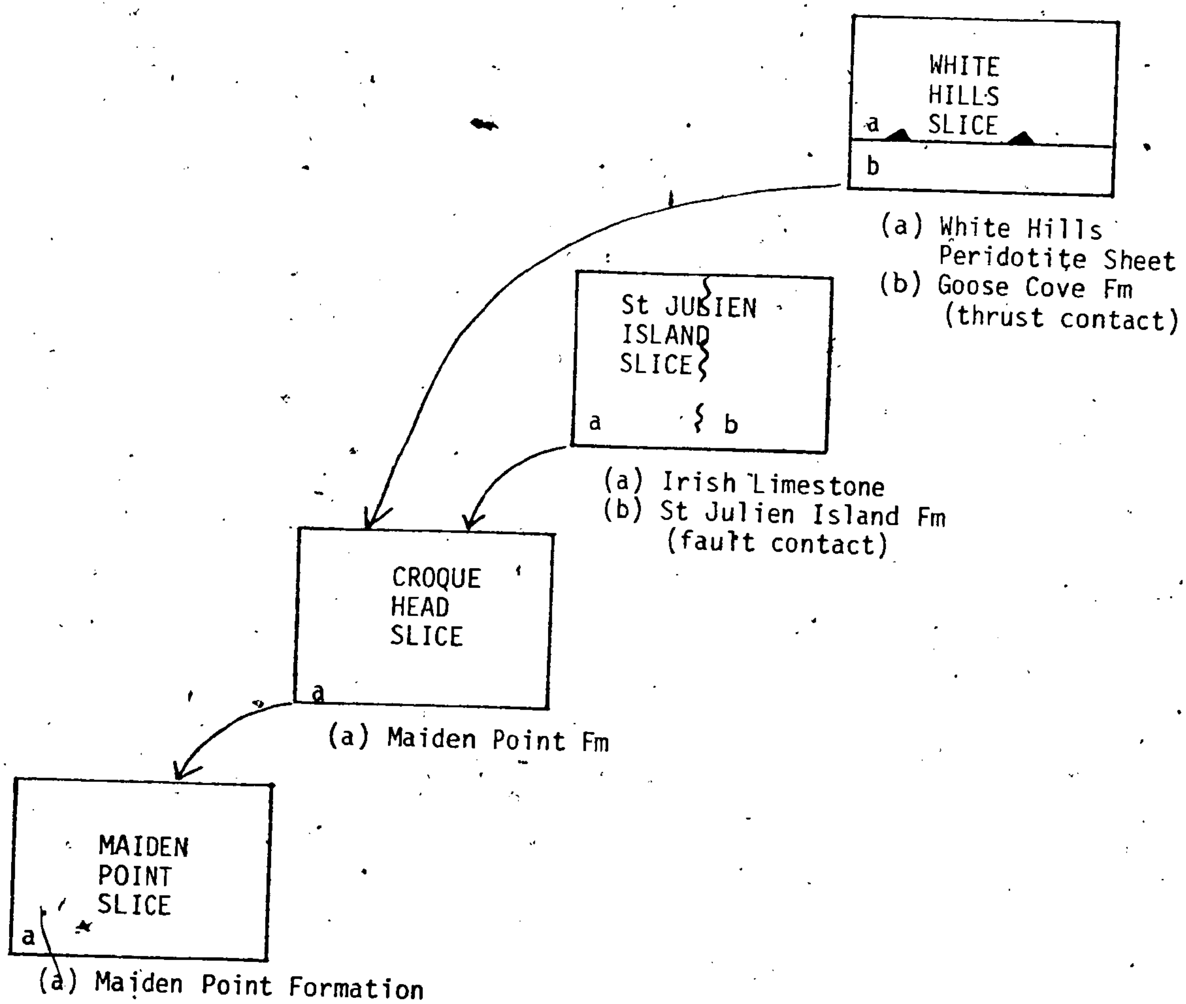


Fig 10: Schematic representation of the structural stacking order of the slices of the southern part of the Hare Bay Allochthon. The formations of each slice are listed. Arrows indicate observed structural relations between the slices.

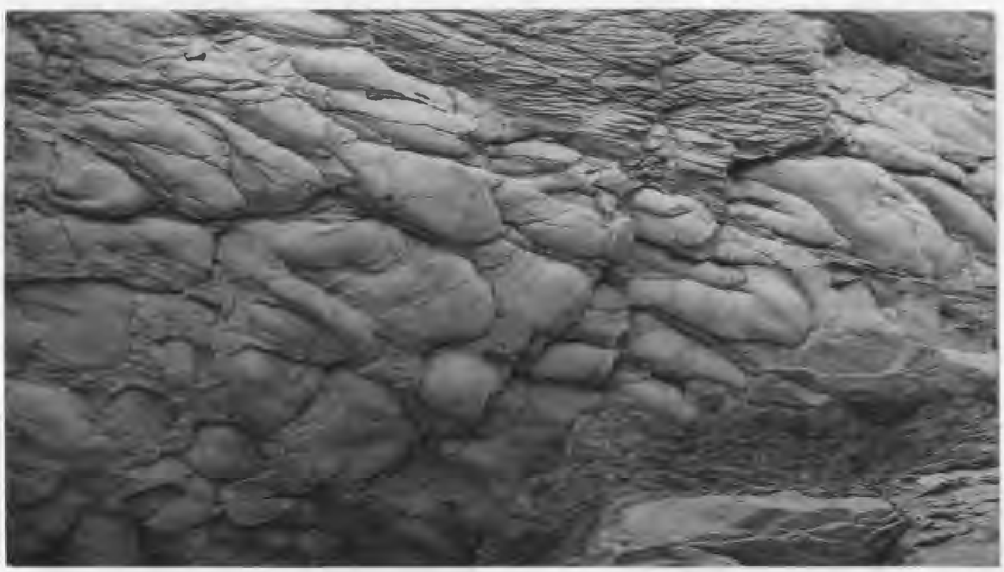
PLATE 16

Fig. a: Graded greywackes from Maiden Point Formation. The beds are slightly overturned. Looking northeast, Spring Island, Hare Bay.

b: Graded greywacke from Maiden Point Formation. Note the lower, coarse graded interval downcutting into shale and the thick, upper, ungraded interval. Spring Island, Hare Bay.

PLATE 17

Fig. a: Cross-bedded greywacke, Maiden Point Formation. Looking north, south of Whites Arm.



b: Load casts from Maiden Point Formation. The casts are approximately 30 cm long. Note regular direction of overlap. Looking north, north of Pillier Bay.

PLATE 18

Fig. a: Hook and roll up structure, Maiden Point Formation. Soft sediment deformation in greywackes. Spring Island.

b: Slumped bedding in greywackes, Maiden Point Formation. Mare Cove.

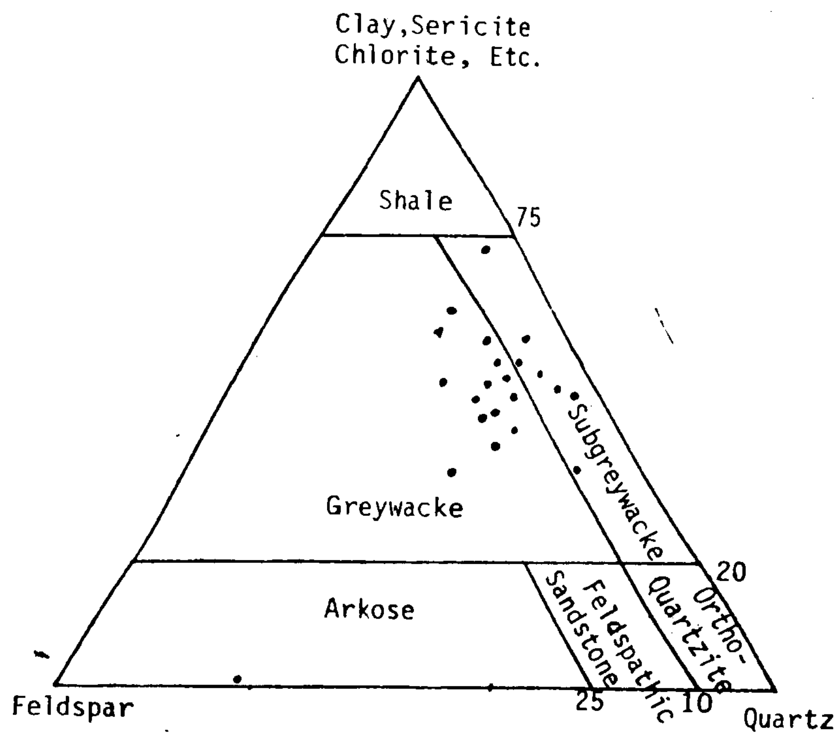


Fig 11: Compositional diagram of the Maiden Point Formation
(Classification after Pettijohn, 1944, in Krumbein & Sloss, 1963)

PLATE 19

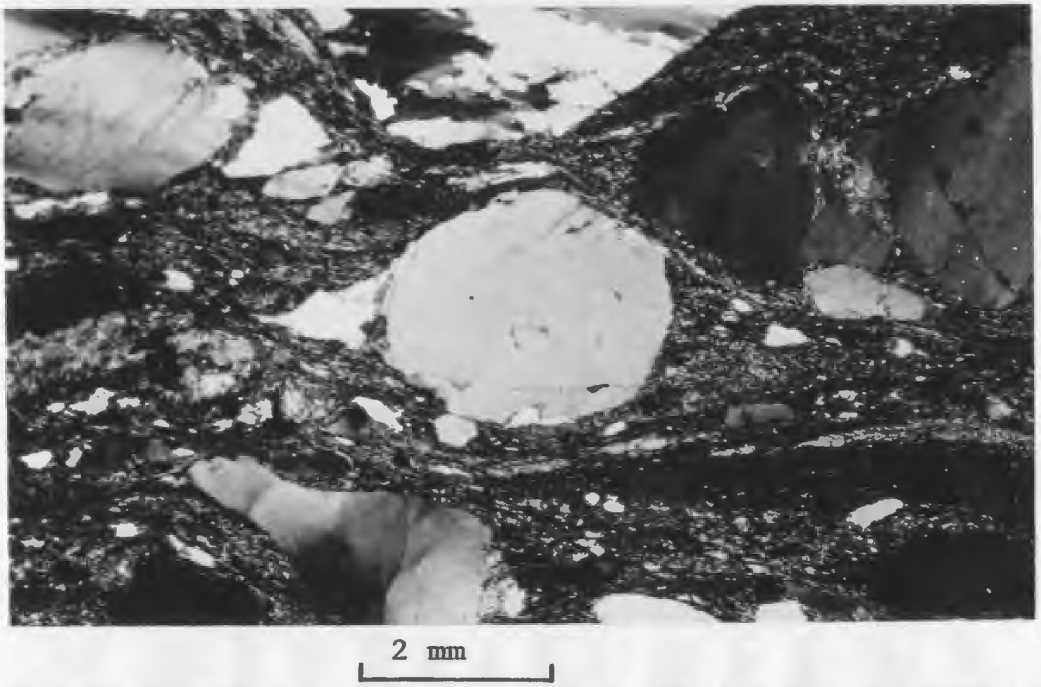
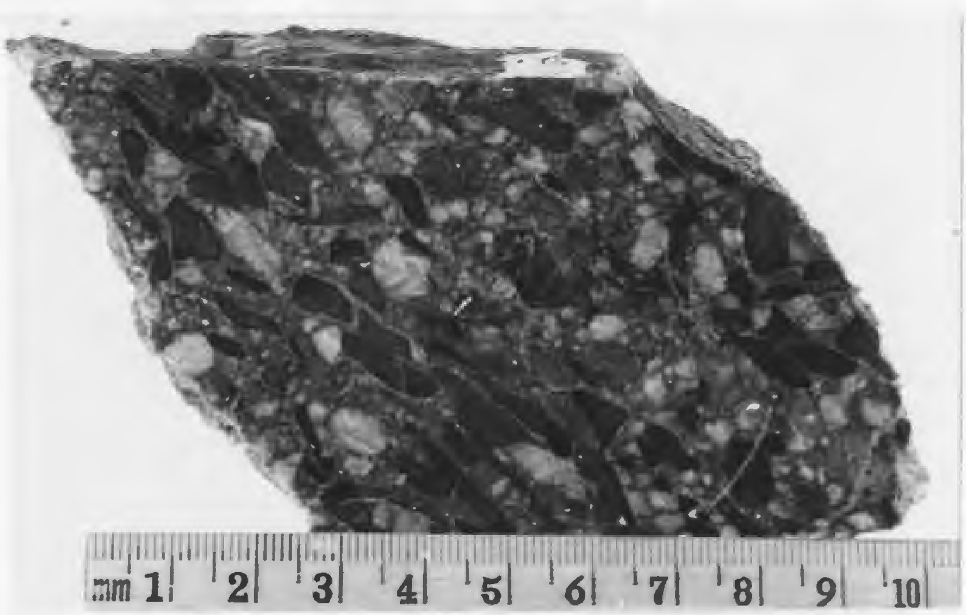


Fig. a: Photomicrograph of greywacke from Maiden Point Formation. Note spherical quartz grain in centre of section, high matrix content, poor sorting and slaty cleavage. Cow Bay, Hare Bay.



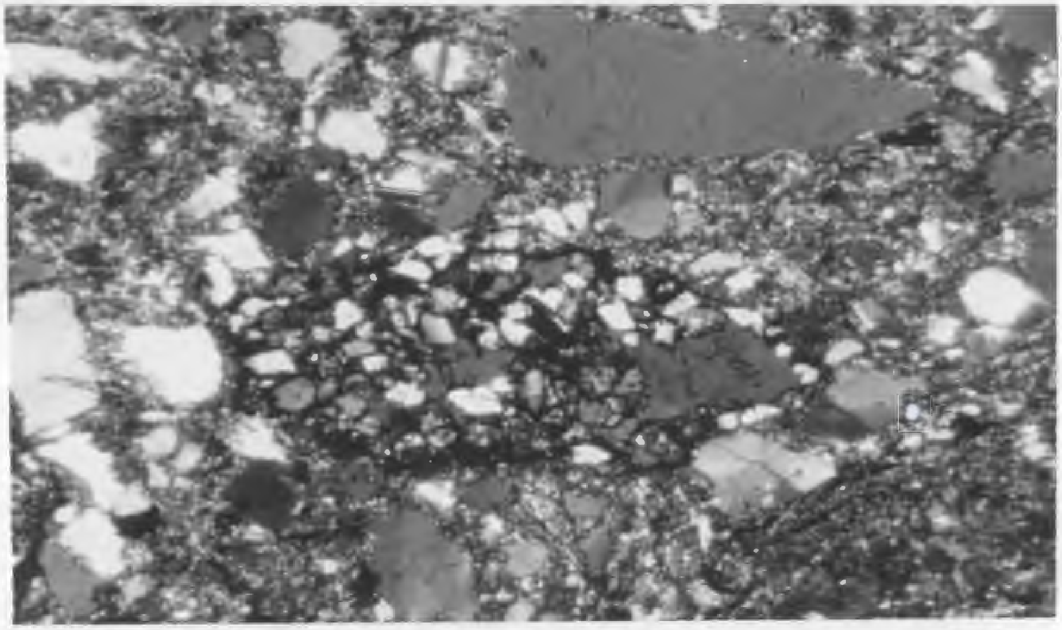
b: Quartz pebble conglomerate from the Maiden Point Formation. Note rough alignment of black mudstone flakes, and poor sorting of quartz grains.

PLATE 20

Fig. a: Angular mudstone clasts in Maiden Point Formation. Note chaotic arrangement of slabs in medium-grained greywacke. Spring Island, Hare Bay.

b: Mudstone slabs in quartz pebble conglomerate, Maiden Point Formation. 1 km east of Big Spring Inlet.

PLATE 21



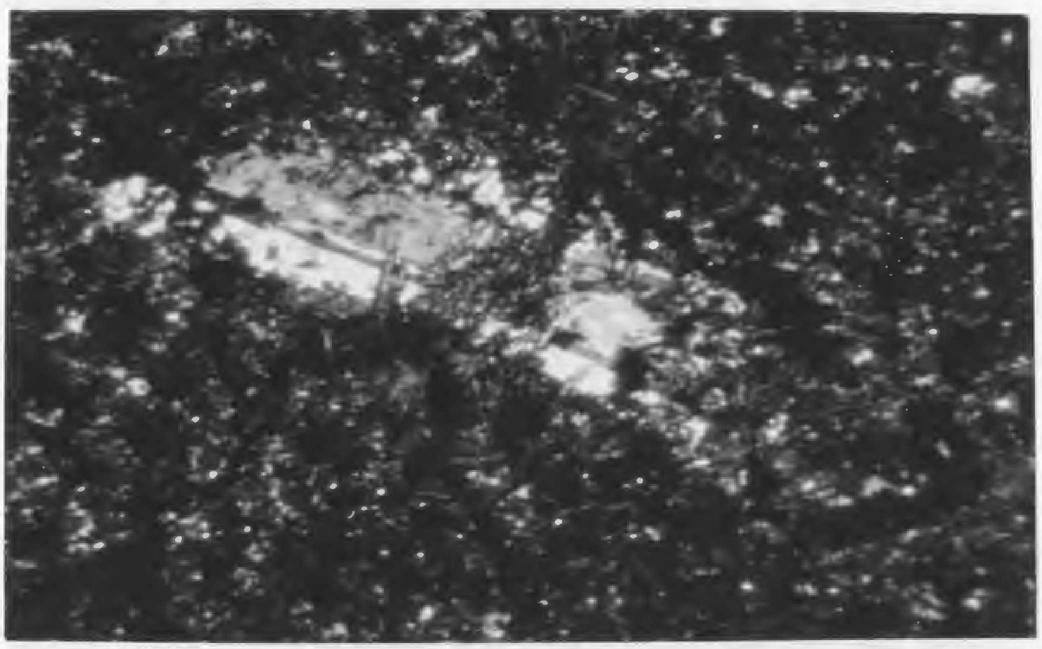
1 mm

Fig. a: Photomicrograph of quartz-pebble conglomerate Maiden Point Formation. Note hematitic sandstone clast in center of section. Conche Road.

b: Pillow lava from Maiden Point Volcanics. Note limestone filled interstices. Little Cormorandier Island, Hare Bay.

PLATE 22

Fig. a: Inclusions of dark, medium-grained basalt in Maiden Point Volcanics. Shore of northwest arm, Croque Harbour.



0.5 mm

b: Post-tectonic stipnomelane overgrowing plagioclase phenocryst, Maiden Point Volcanics. Note fine-grained, altered groundmass. West of Grandois.

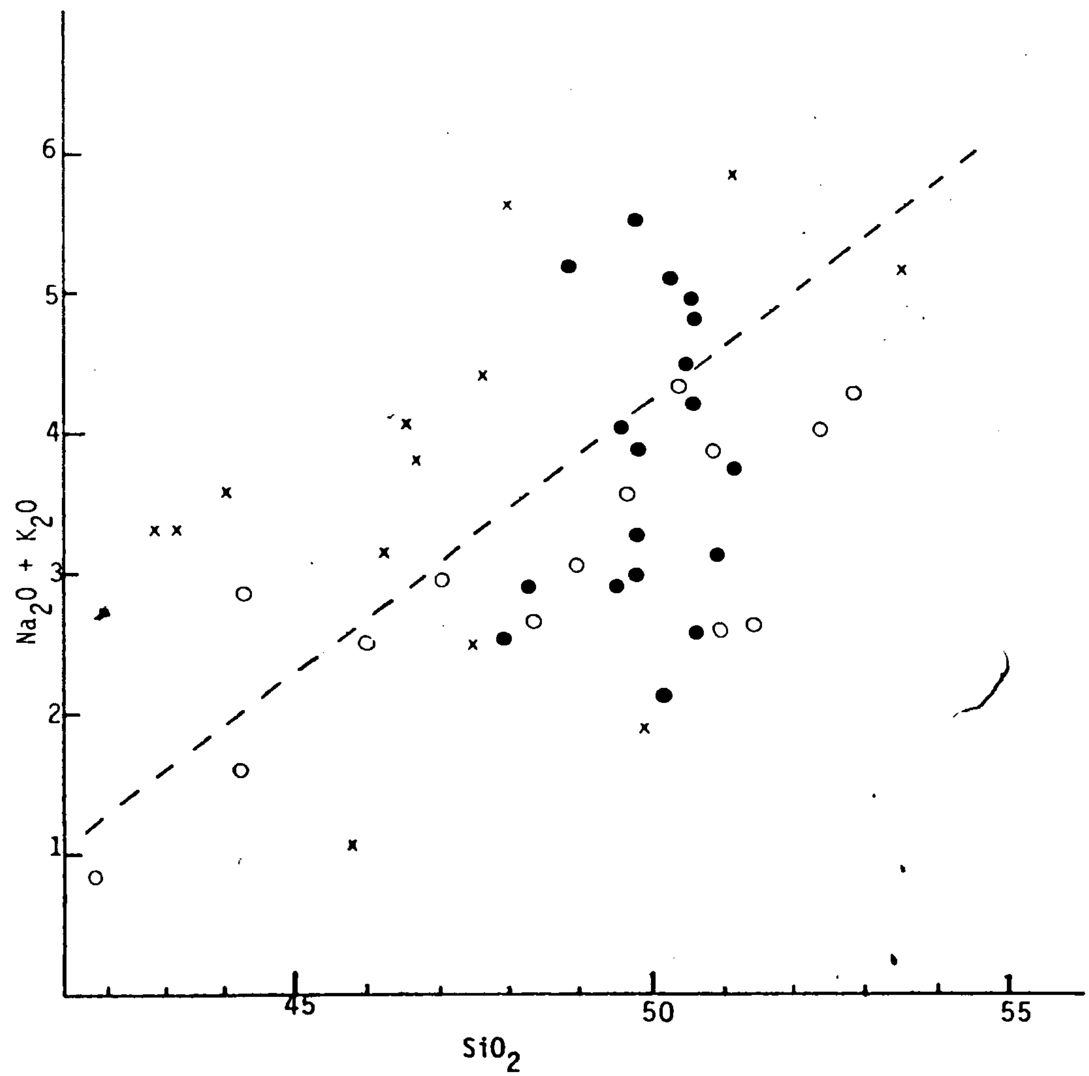


Fig 12: Variation of total alkalis and silica in the Maiden Point Volcanics Lighthouse Cove Formation and Lushs Bight Group. Crosses = Maiden Point Volcanics; dots = Lighthouse Cove flows and dykes; circles = Lushs Bight flows. The dashed line is that separating tholeiitic and alkali basalts in Hawaii (after Macdonald and Katsura, 1964). Data in weight percent. Data for the Lighthouse Cove Formation donated by D.F. Strong and for the Lushs Bight Group donated by D.F. Strong and W.G. Smitheringale, Memorial University.

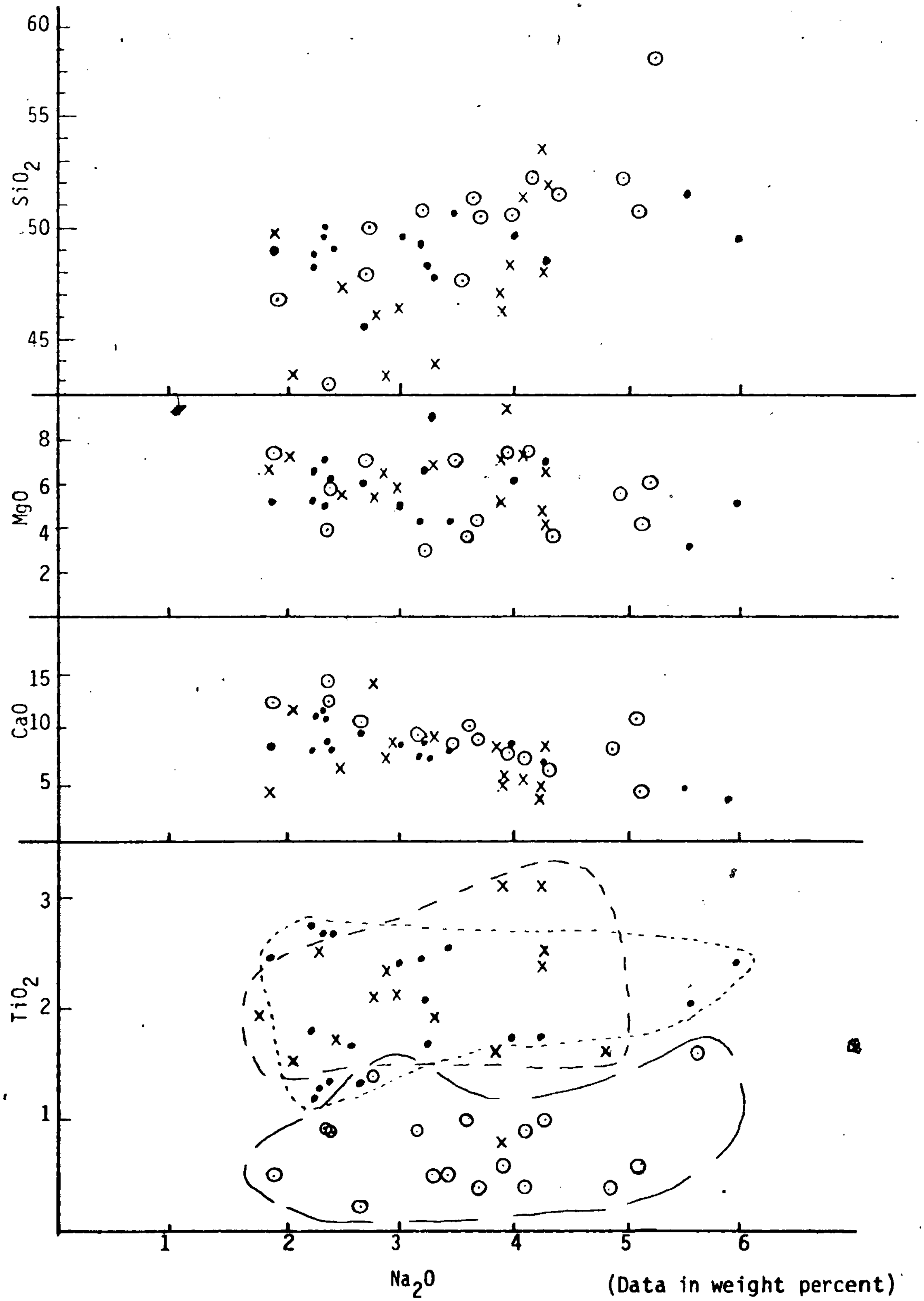


Fig 13: Variation of SiO_2, MgO, Cao and TiO_2 with Na_2O in lavas of the Maiden Point Formation, Lighthouse Cove Formation, and the Lushs Bight Group. Crosses = Maiden Point Volcanics; dots = Lighthouse Cove Formation; circles = Lushs Bight Group.

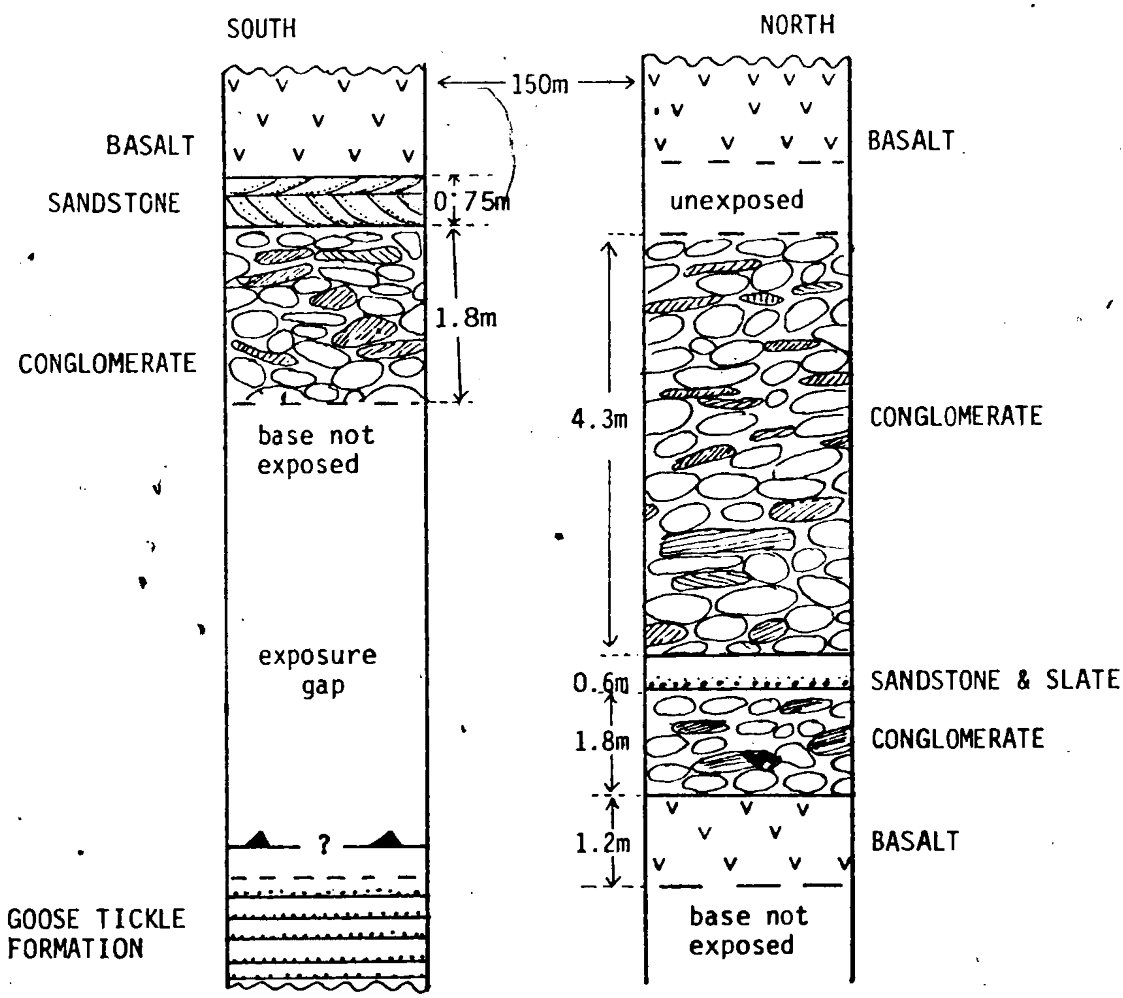


Fig 14: Schematic representation of the stratigraphic relations of the conglomerate near the base of the Maiden Point Formation, northwest arm, Croque Harbour.
Basalt clasts = clear; limestone clasts = hatched.

PLATE 23

Fig. a: Basalt flow overlying sandstone, from the Maiden Point Formation. Basalt (top right of photograph) is underlain by cross-bedded sandstone and conglomerate. South side, northwest arm, Croque Harbour.

b: Conglomerate, Maiden Point Formation. Note dark grey, subspherical, basalt clasts and light grey, slab-shaped, marble clasts. North side, northwest arm, Croque Harbour.

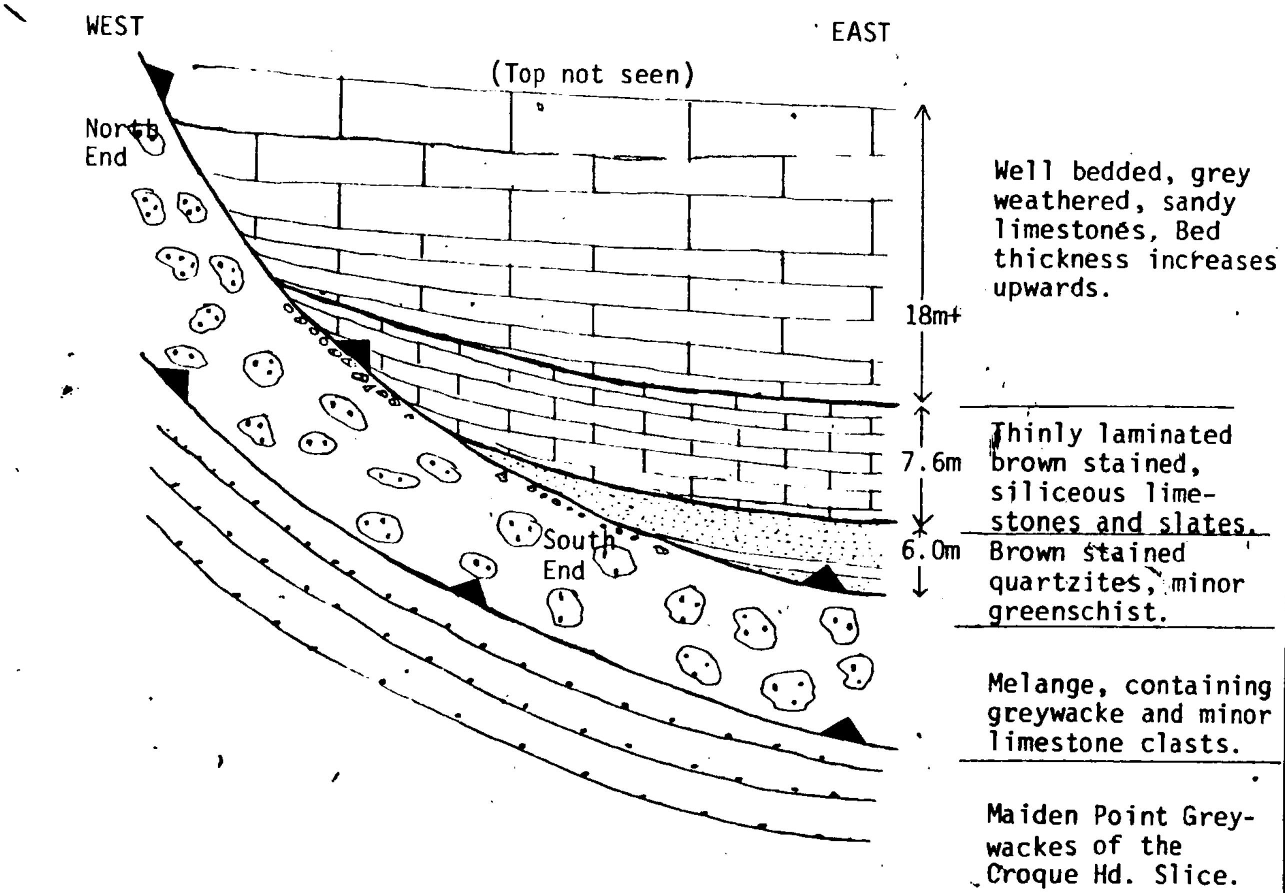


Fig 15: Diagrammatic stratigraphic cross section of the Irish Limestone, west side of St. Julien Island.

PLATE 24

Fig. a: Poorly graded, sandy limestone, Irish Limestone. The beds are upward facing on the steeply dipping cleavage (S_2). Looking northeast, northwest side of St. Julien Island.

b: Bedded conglomerate and greywacke, St. Julien Island Formation. Note poor sorting, dispersed texture and high matrix content of conglomerate. Looking north, south side of St. Julien Island.

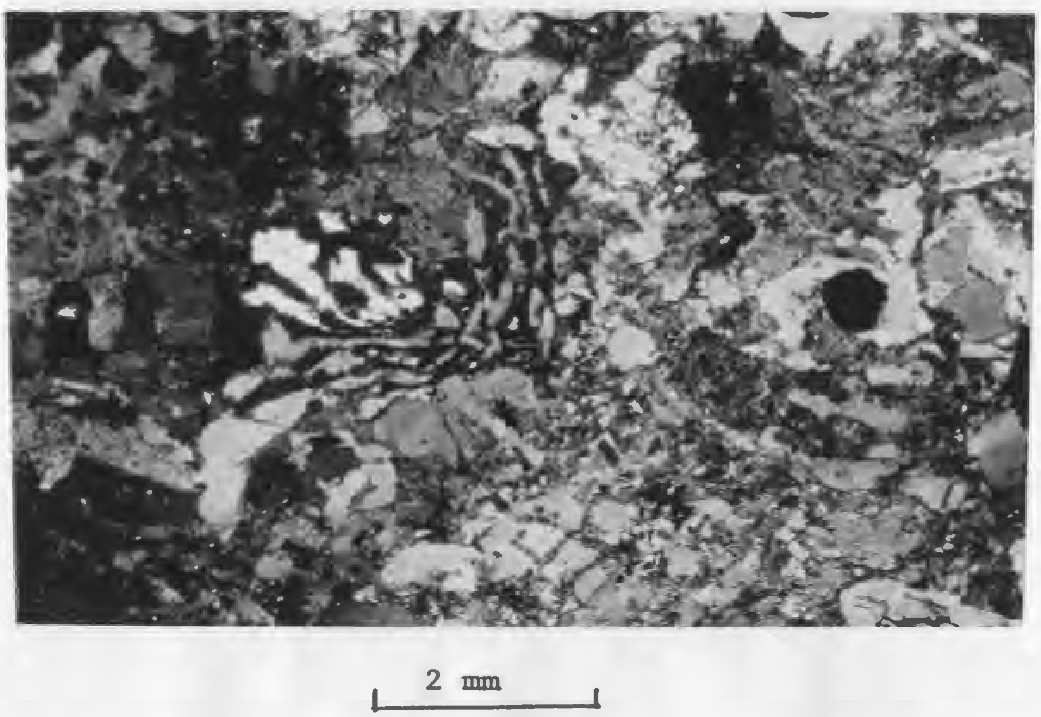
PLATE 25

Fig. a: Large scale trough cross-bedding, St. Julien Island Formation. Note the downcutting nature of bed outlined. Looking south, west side of St. Julien Island.

b: Trough cross-bedding, St. Julien Island Formation. Close up of greywacke bed in Fig. a, above.

PLATE 26

Fig. a: Conglomerate, St. Julien Island Formation. Note dispersed texture and alignment of clasts in (S_2) cleavage plane.



b: Myrmekitic texture in granite clast, St. Julien Island Formation. St. Julien Island. Crossed nicols.

PLATE 27

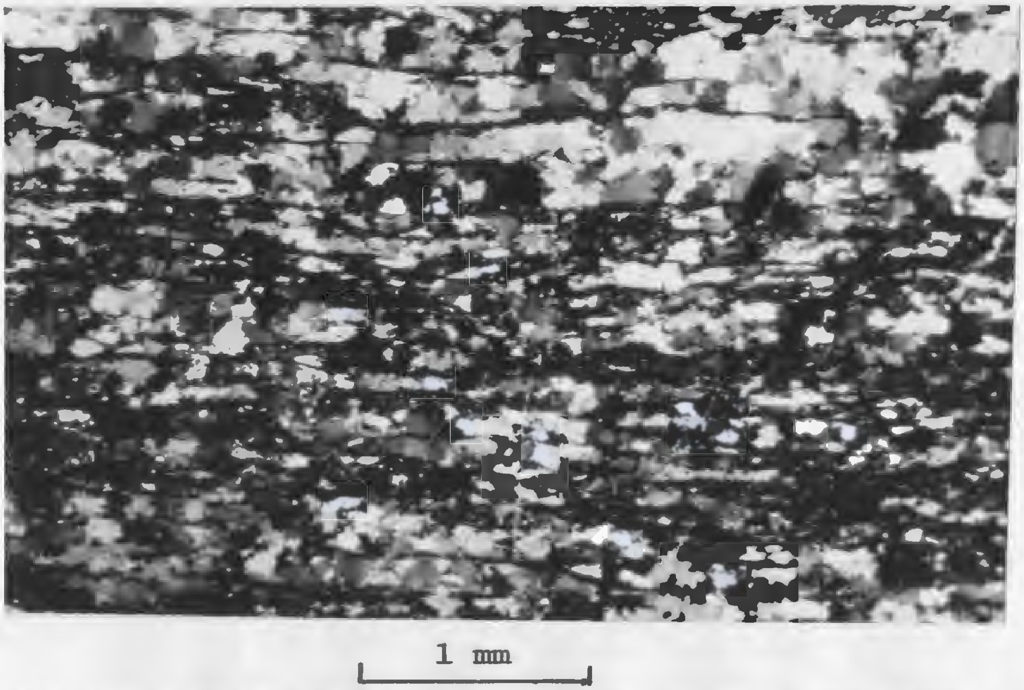
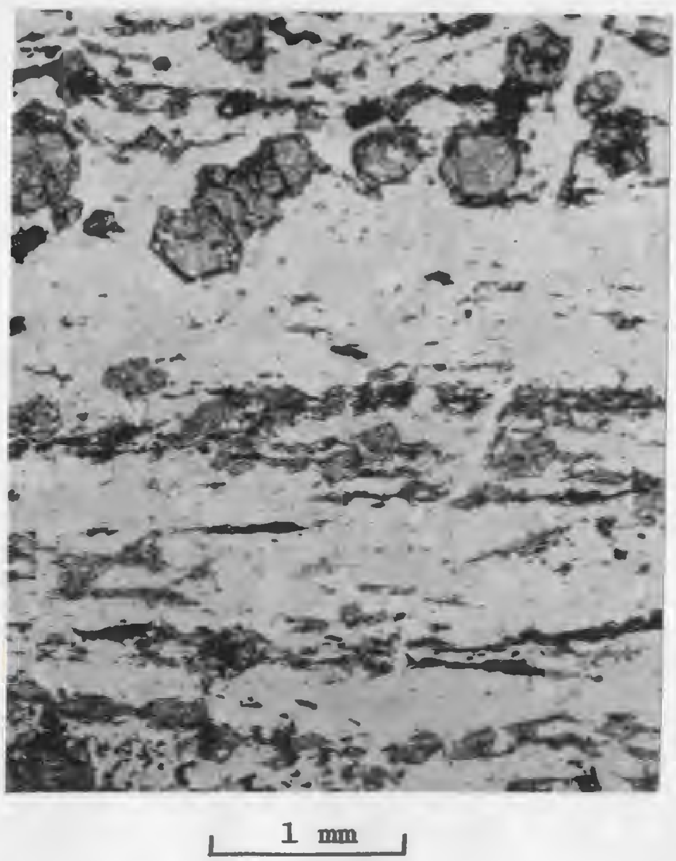


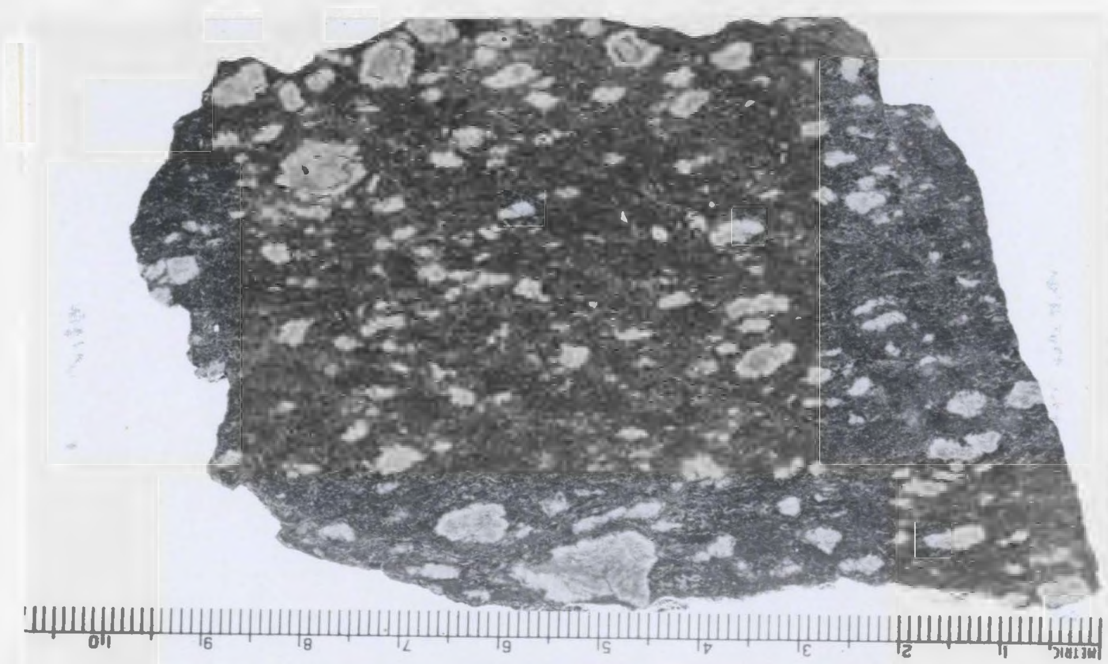
Fig. a: Foliated gneiss clast, St. Julien Island Formation. Note alternating quartz and mafic (dark) foliae. St. Julien Island. Crossed nicols.



b: Relict euhedral garnets in gneiss clast, St. Julien Island Formation. Note characteristic hexagonal shapes of garnets. St. Julien Island. Plane polarized light.

PLATE 28

Fig. a: Agglomerate, Greenschist Member, Goose Cove Formation. Note rounded clasts and schistose matrix that forms augen around them. Shore, south side of western entrance to Fishot Harbour.



b: Porphyritic pillow lava, Greenschist Member, Goose Cove Formation. Note schistose matrix which forms augen around feldspar phenocrysts. Pigeon Island.

PLATE 29

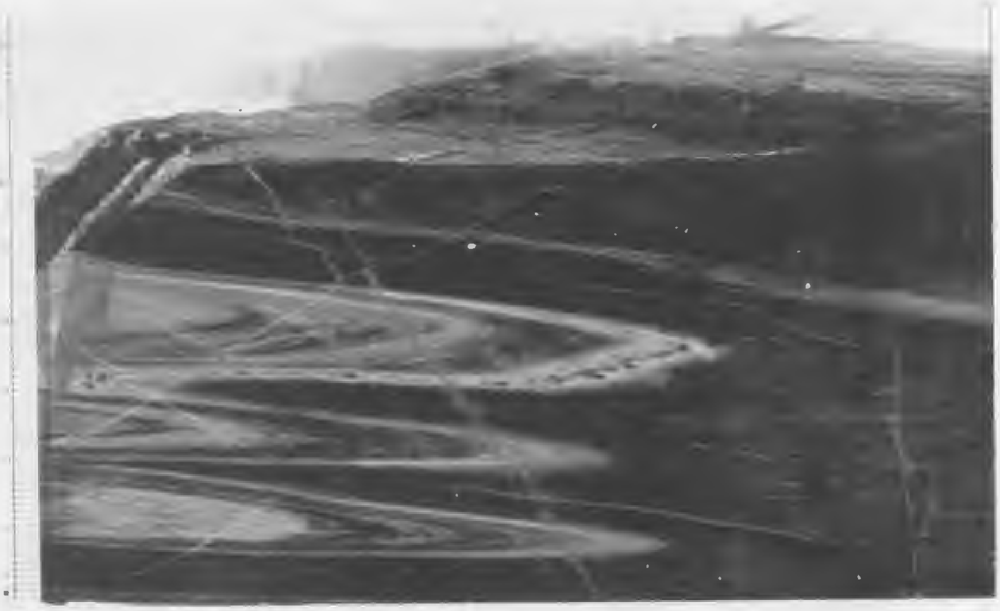
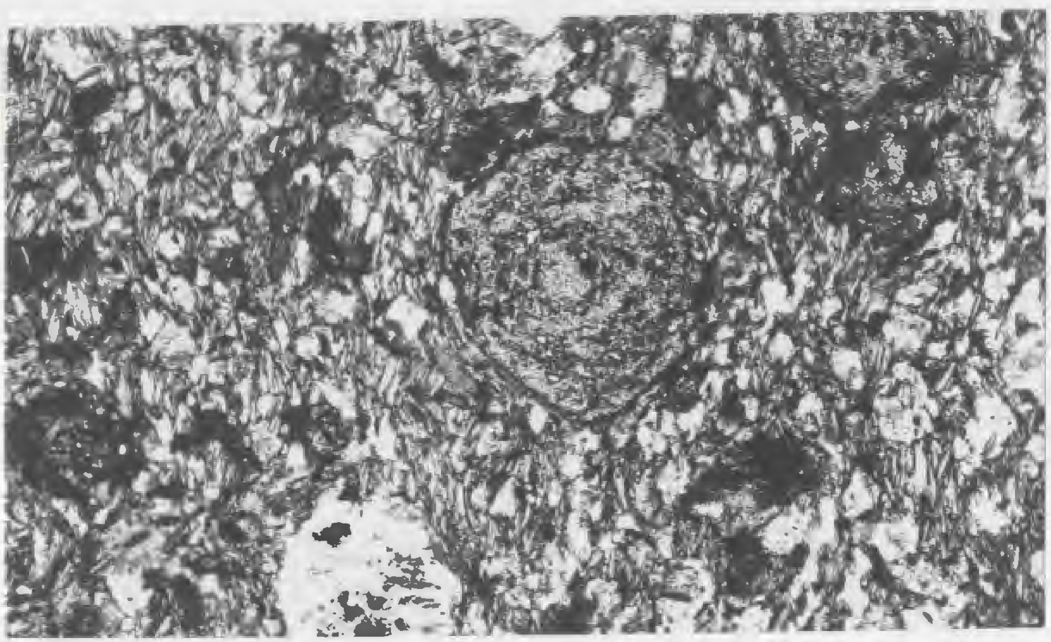


Fig. a: Banded hornblende schist, Amphibolite Member, Goose Cove Formation. The schists are derived from tuffs. The isoclinal folds are F_2 structures. Landing Cove, Fishot Island.



0.5 mm

b: Photomicrograph of pelite, Amphibolite Member, Goose Cove Formation. Note biotite schistosity (S_2) which forms augen around garnet (dark) and plagioclase (light) porphyroblasts. Landing Cove, Fishot Island. Obliquely crossed nicols.

PLATE 30

Fig. a: Marble bands in hornblende schist, Amphibolite Member, Goose Cove Formation. Note how well the marbles (white) display the structure in an otherwise homogeneous looking rock. Looking north, north side of Easter Tickle, Northeast Island.

Fig. b: Metagabbro, Amphibolite Member, Goose Cove Formation. The mineral banding is interpreted as primary igneous layering. East side of Landing Cove, Fishot Island.

PLATE 31

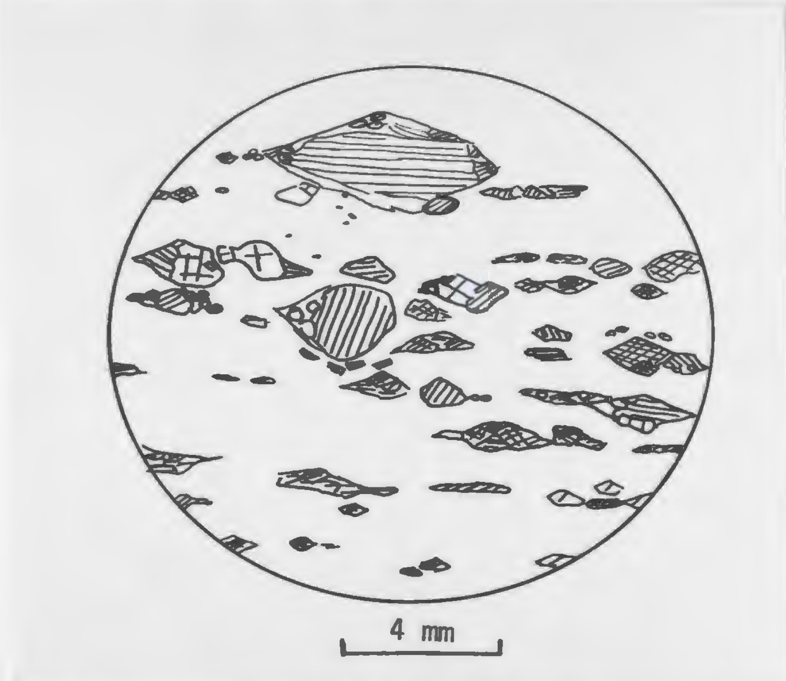


Fig. a: Line drawing of Pyroxene-bearing Amphibolites, Contact Zone, Goose Cove Formation. Note augen shaped hornblende (closely spaced cleavage) and pyroxene (widely spaced cleavage) porphyroblasts showing (S_2) fabric. Western White Hills.

b: Greywacke blocks in Mélange. Note variety of shapes and sizes of blocks. Looking northeast, west side of Fishot Island.

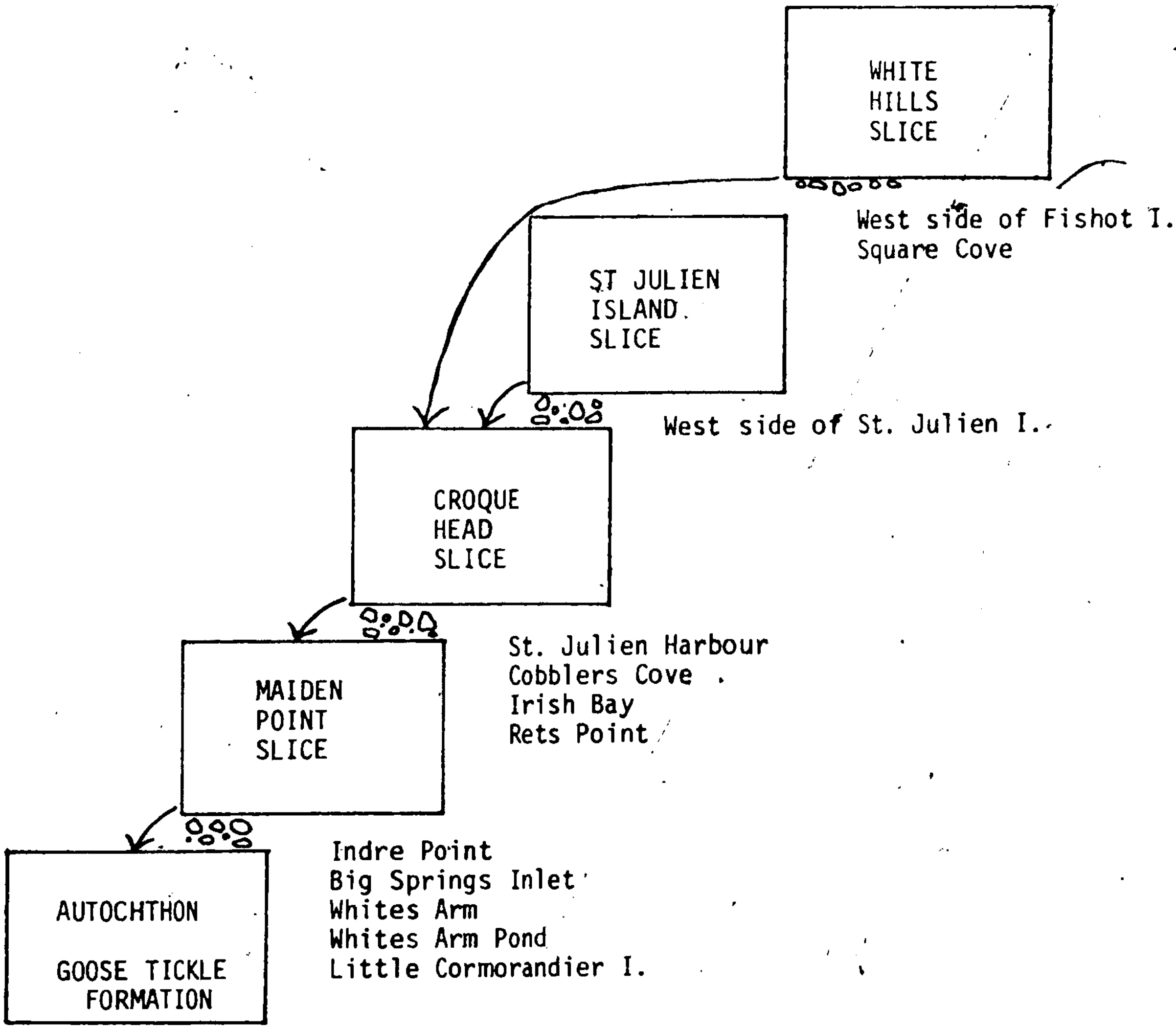
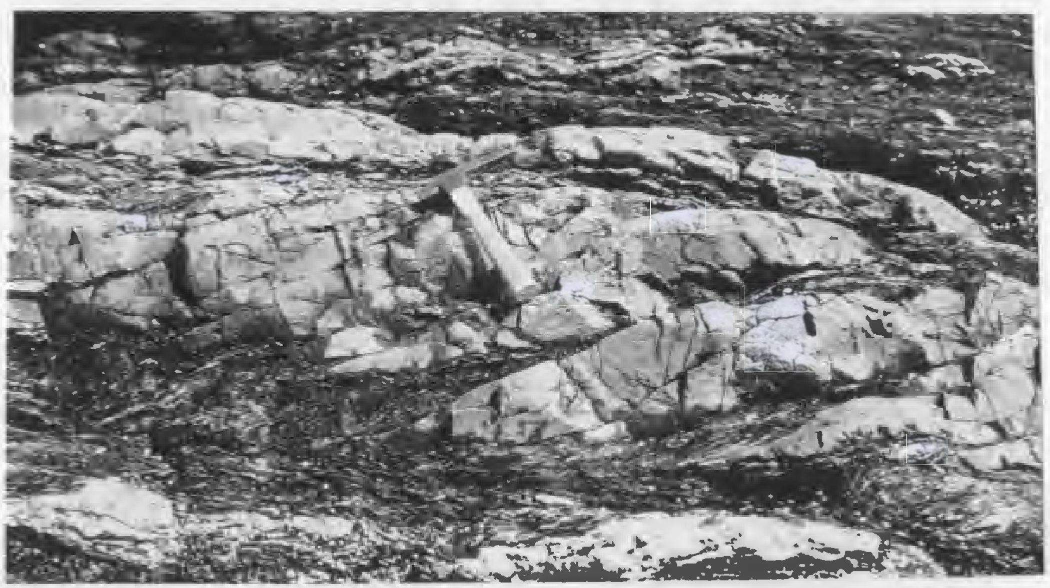


Fig 16: Schematic representation of the structural stacking order of the slices listing the locations of the various melange zones between the slices. Arrows indicate observed structural relations. Localities shown in Fig 3.

PLATE 32

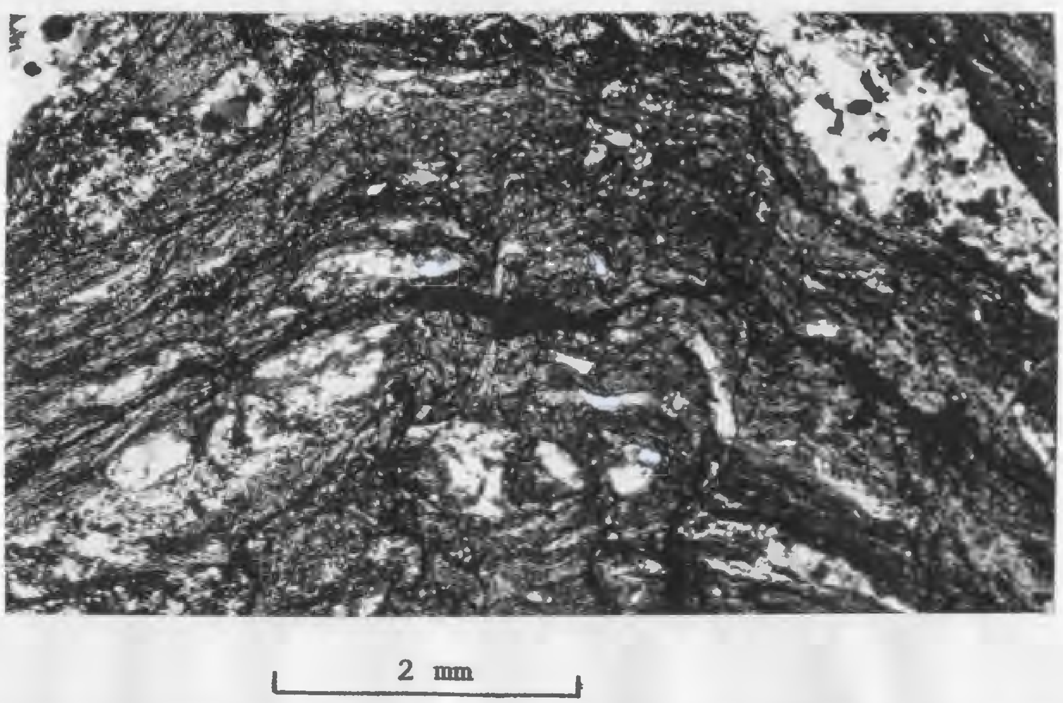
Fig. a: Folded contact between greywackes of the Croque Head slice and Mélange. Note the sharp contact and the upright post-emplacement folds. Looking south, east side of Irish Bay.



b: Pre-tectonic diabase dyke(?) in basal mélange to White Hills Slice. Note that dyke has been cut up into small imbricate slices. Looking east, southwest side of Fishot Island.

PLATE 33

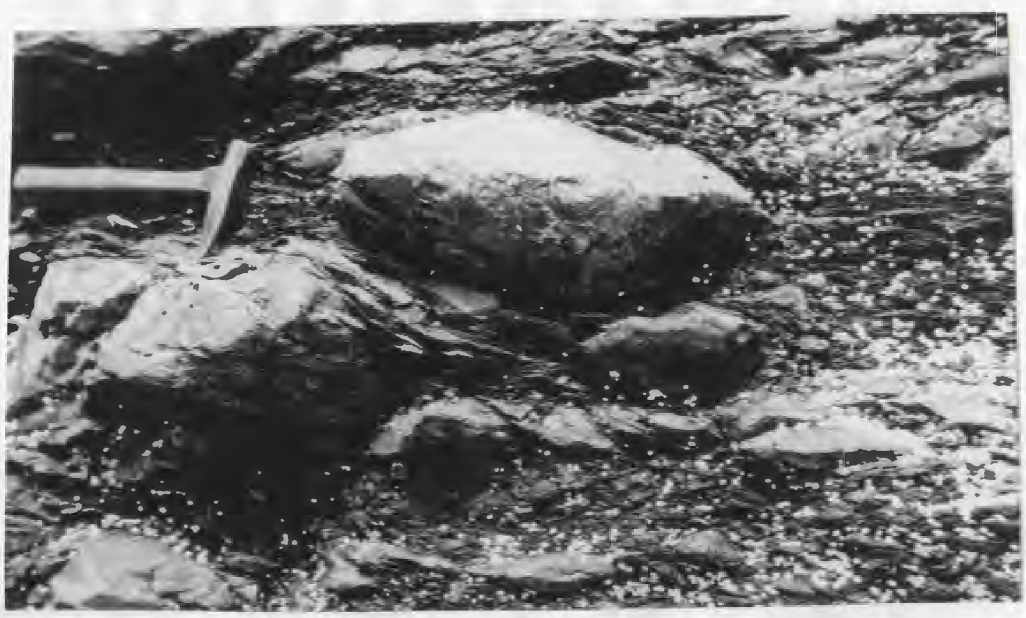
Fig. a: Post-tectonic dyke cutting the basal mélange to the Fishot Island Slide. Looking north, southwest side of Fishot Island.



b: Photomicrograph of black slate matrix of mélange. The schistose matrix is folded by upright, post-emplacement folds.

PLATE 34

Fig. a: Mélange. Note the unsorted texture with clasts ranging from pebbles to blocks. North shore of Little Canada Harbour.



b: Oval shaped boulders in Mélange. Note how black slate matrix forms augen around the blocks. West side of Fishot Island.

PLATE 35

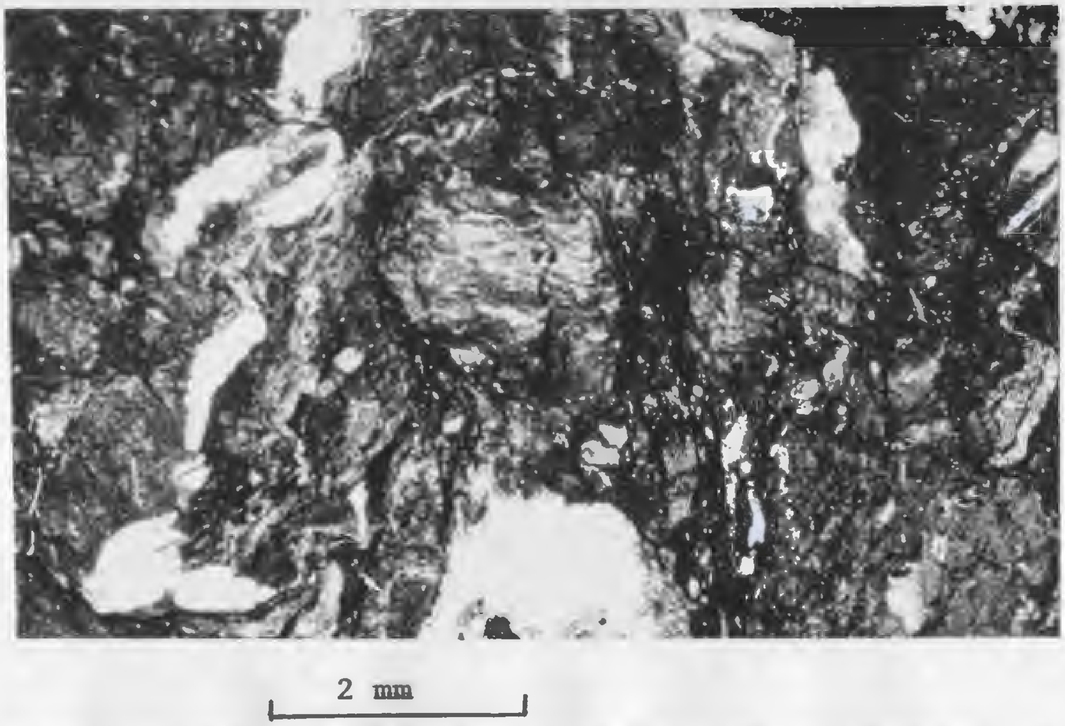
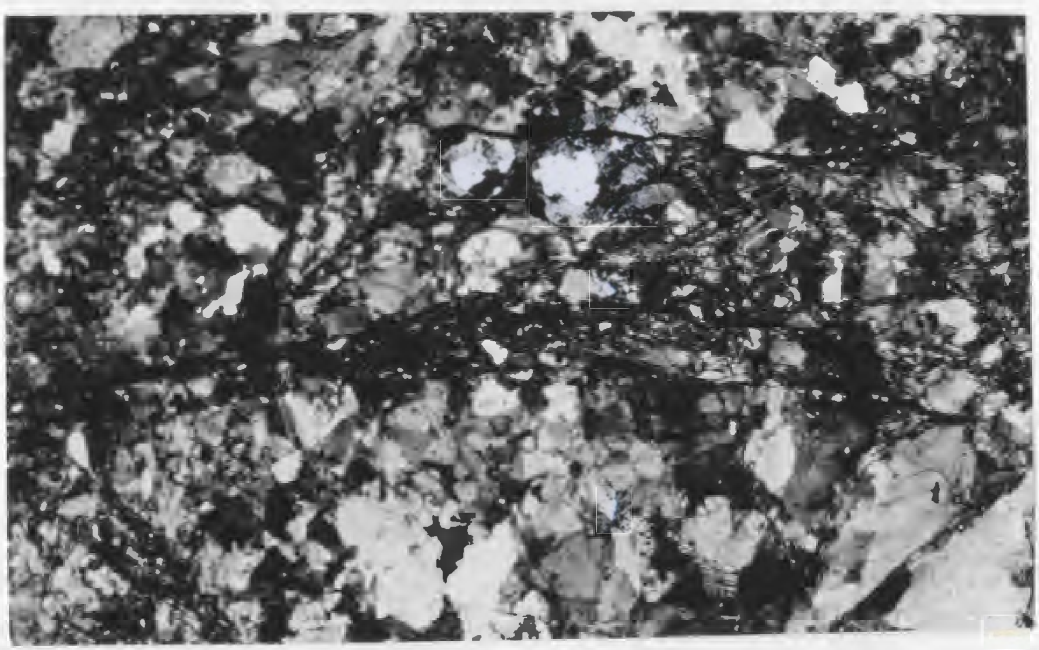


Fig. a: Photomicrograph of Mélange. Note how the emplacement schistosity (S_1) is cut by an upright crenulation (S_2) cleavage. Basal melange to the Croque Head Slice, Cobblers Cove.

b: Recumbent fold in mélange. Note how the F_1 fold is cut by a moderately dipping fracture cleavage (S_2). Looking north, north shore of Little Canada Harbour.

PLATE 36

Fig. a: Thrusted microgranite resting on Bradore Formation. The Bradore sandstones unconformable overlie gneisses of the Long Range Complex. Seans Cove, White Bay.



b: Photomicrograph of deformed micrgranite. Note thin anastomising fracture zones (dark) containing fine-grained, crushed micrgranite. Seans Cove.

PLATE 37

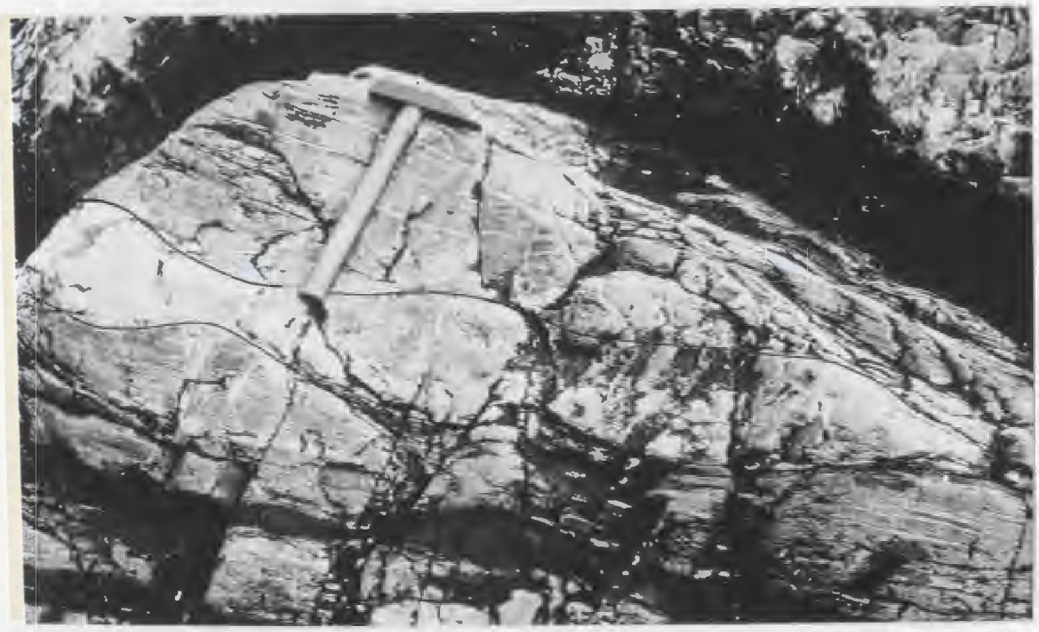


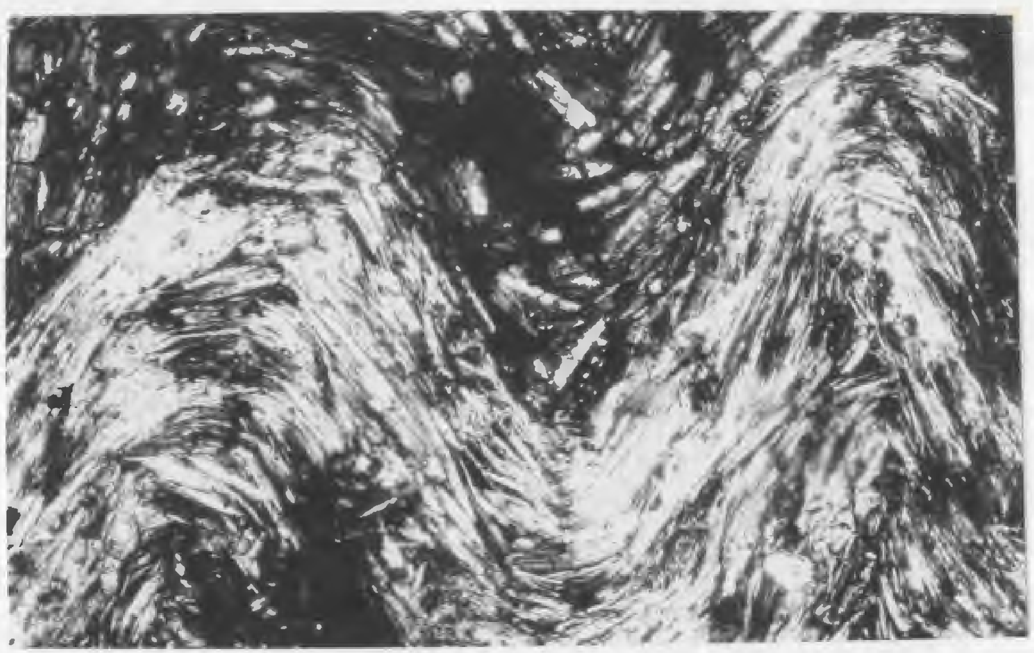
Fig. a: Pinch and swell structure in granite dyke in Acid and Basic Gneisses. Granite dyke outlined. Note foliated gneisses and unfoliated granite. North of Seans Cove.

0.5 mm

b: Photomicrograph of deformed porphyry. Note schistose groundmass and strained relict quartz phenocrysts. Base of faulted block, west of Burnt Point, Canada Bay.

PLATE 38

Fig. a: Deformed diabase dyke from Table Head Formation. Note rectangular shaped cross section of boudins. East shore of Canada Harbour.



0.5 mm

b: Pelite from Greenschist Member, Goose Cove Formation. Note S_1 schistosity defined by muscovite and chlorite, crenulated by small scale F_2 folds. Northwest side of Fishot Island.

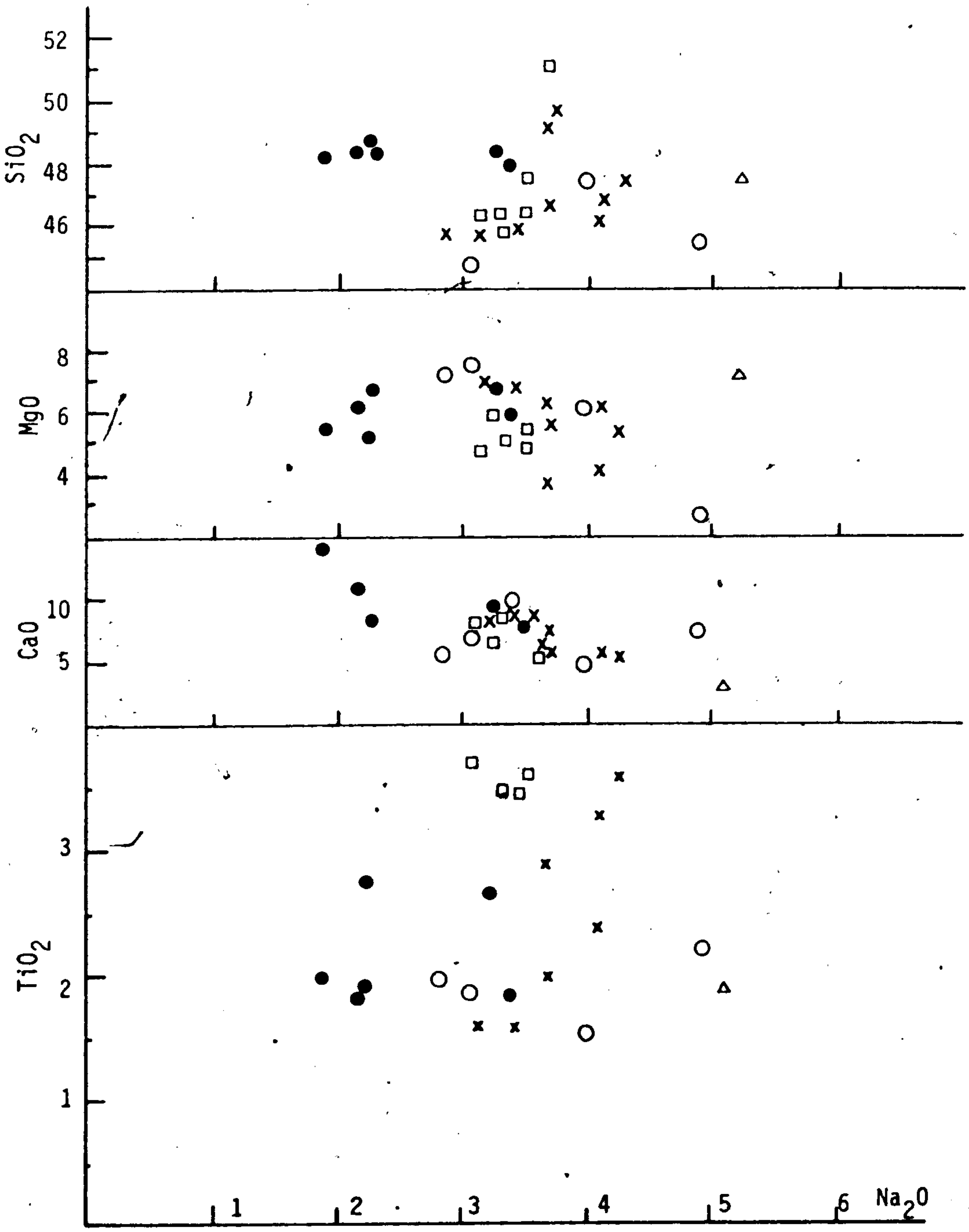


Fig 17: Variation of TiO_2, CaO, MgO and SiO_2 with Na_2O in post-Acadian dykes in allochthonous and autochthonous rocks and in the Lighthouse Cove Formation. Crosses = Dykes in the Maiden Point Slice; squares = dykes in the White Hills Slice; Triangle = dyke in the St. Julien Island Slice; circles = dykes in the autochthonous Goose Tickle Formation; dots= dykes in the Lighthouse Cove Formation. Data in weight percent. Lighthouse Cove data from Strong and Williams (1972).

Fig. 18: SUMMARY OF THE STRUCTURAL HISTORY OF THE ALLOCHTHONOUS & AUTOCHTHONOUS ROCKS SOUTH OF HARE BAY.

DEFORMATION EVENT	AUTOCHTHON	MAIDEN POINT SLICE	CROQUE HEAD SLICE	ST. JULIEN ISLAND SLICE	WHITE HILLS SLICE (Goose Cove Formation only)
POST-EMPLACEMENT Minor Open Folds		Minor open folds and kink bands	Minor open folds	Minor open folds and kink bands	Kink bands
DA (ACADIAN)	S_1 slaty cleavage and open, upright flexural slip folds at Hare Bay. Open westerly inclined similar folds at Canada Bay. Crenulation S_2 cleavage.	Open to close upright to moderately inclined (F_2) similar folds with associated fracture to crenulation cleavage.	Open to close upright similar F_2 folds, with associated crenulation cleavage (S_2)	Open upright similar F_2 folds, with associated slaty to crenulation cleavage (S_2)	Open upright F_3 folds and local shear zones. Weak crenulation S_3 cleavage
SYN-EMPLACEMENT DE (TACONIC?)	Slaty cleavage at Canada Bay and White's Arm Window that dies out northwestwards. Schistosity in Sugarloaf schists. Imbricate thrusts and local recumbent folds at Canada Bay.	Strong to weak to absent slaty cleavage (S_1). Rare recumbent similar folds (F_1).	Penetrative S_1 slaty cleavage. No folds seen.	Penetrative (S_1) slaty cleavage. Fault. No folds seen.	Minor faults
PRE-EMPLACEMENT DP $\frac{D_{P2}}{D_{P1}}$	--	--	--	--	DP_2 Recumbent to inclined, isoclinal to close F_2 similar folds. S_2 crenulation cleavage to schistosity. DP_1 Penetrative S_1 schistosity. Tectonic slides. No folds seen.

PLATE 39

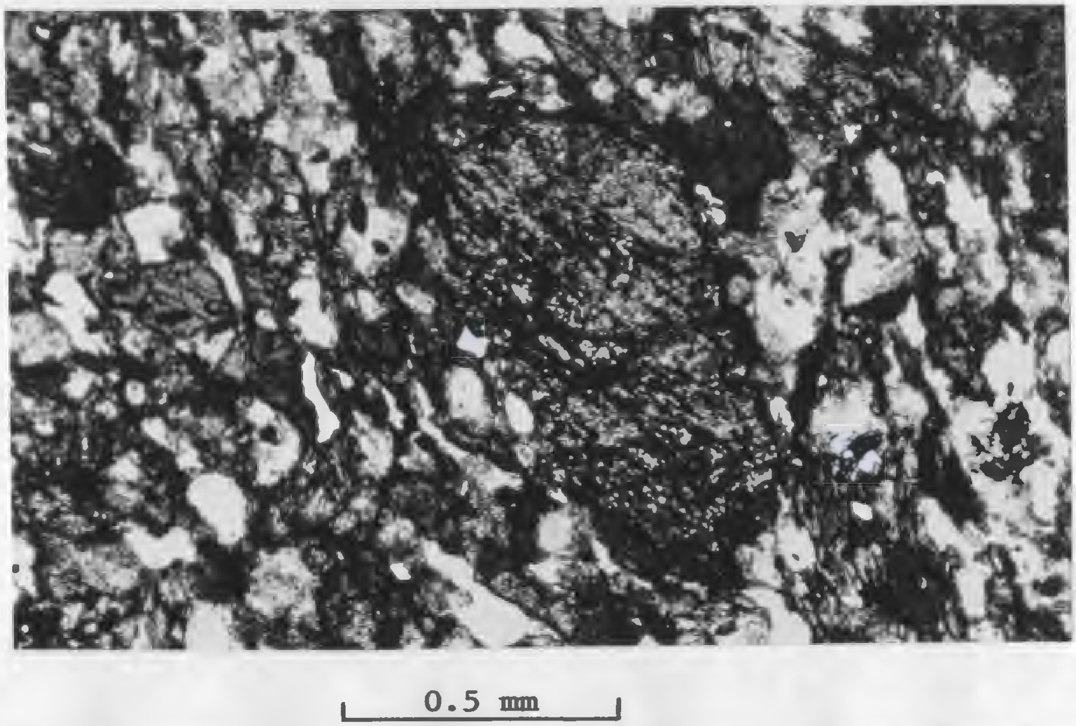
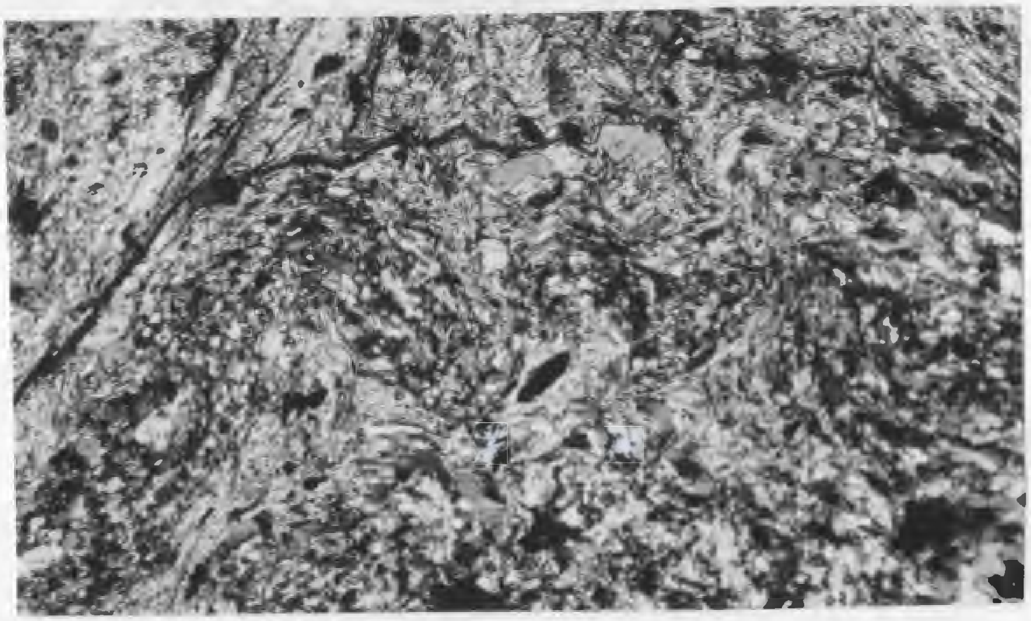


Fig. a: Garnet porphyroblast, Amphibolite Member, Goose Cove Formation. Note straight inclusion trails of elongate ores defining S_1 schistosity. East shore of Landing Cove, Fishot Island.

b: Recumbent fold, Amphibolite Member, Goose Cove Formation. Fold is developed in impure marbles. Looking north, south side of Easter Tickle, Fishot Island.

PLATE 40



1 mm

Fig. a: Granulated sandstone, Bradore Formation. The S_1 fabric is folded about F_2 folds. Near top of faulted block, west of Burnt Point, Canada Bay.

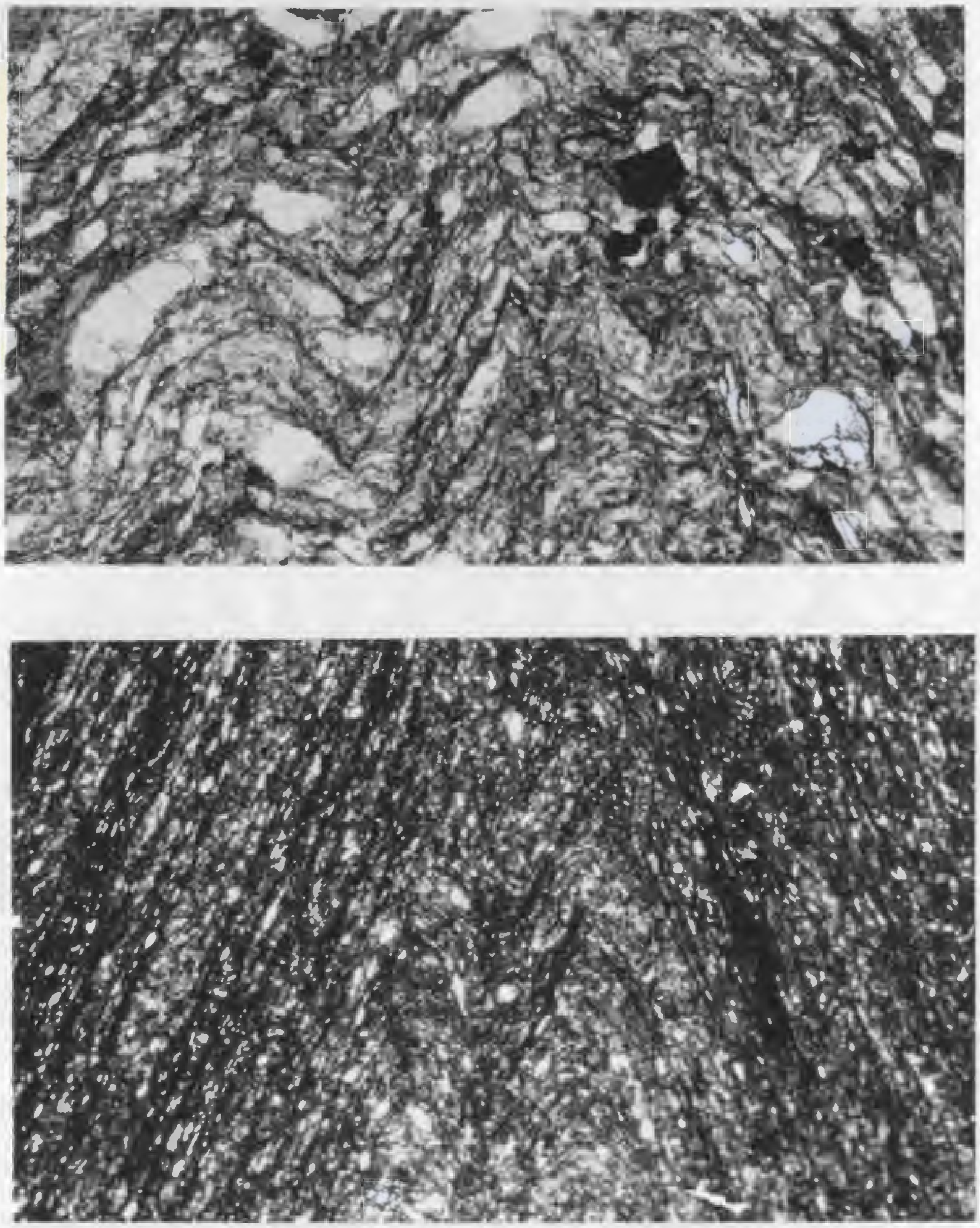
b: Emplacement recumbent folds from the Goose Tickle Formation. Looking east, west side of Baard Island, Canada Bay.

PLATE 41

Fig. a: Emplacement recumbent folds from Table Head Formation below the Handy Harbour Thrust. Note weakly developed, moderately dipping fracture cleavage (S_2). Looking east, west side of Bide Head.

b: Emplacement fold in Forteau Formation. Looking northeast, west of Fly Point.

PLATE 42



1 mm

(a)

2 mm

(b)

Fig. a and b: Post emplacement folds (F_2) in greywackes from the Goose Tickle Formation. Note dimensional preferred orientation of detrital material defining the emplacement S_1 fabric. Obliquely crossed nicols. (a) East of Englee; (b) South shore of Whites Arm.

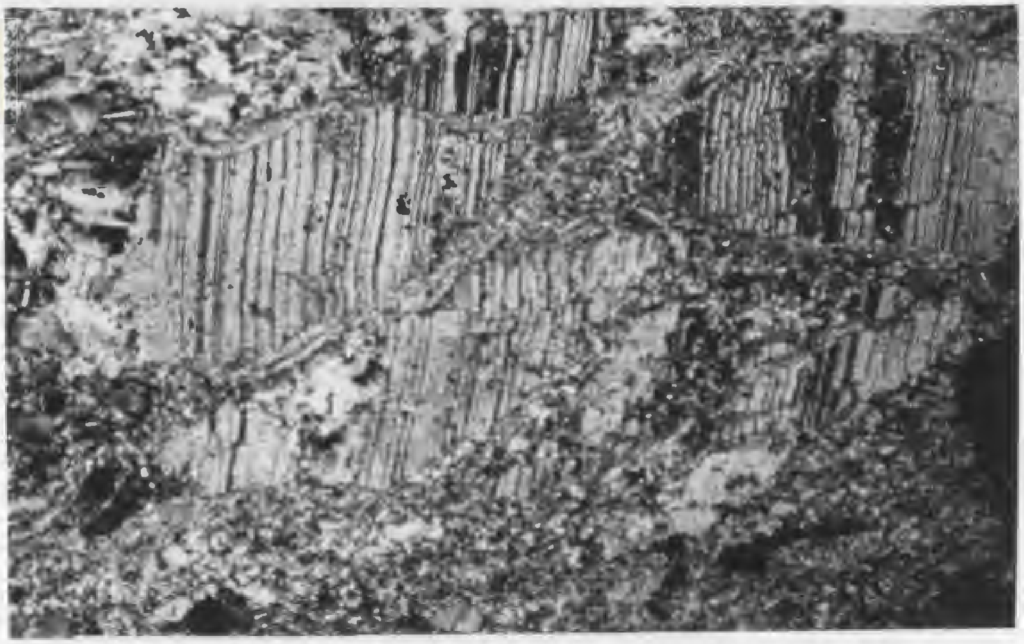
PLATE 43

Fig. a: Phyllonitized granite, Sugarloaf Slide Zone. Note lenticles of quartz and feldspar bounded by narrow zones of finely crushed, schistose matrix. South of Sugarloaf Cove.

2 mm

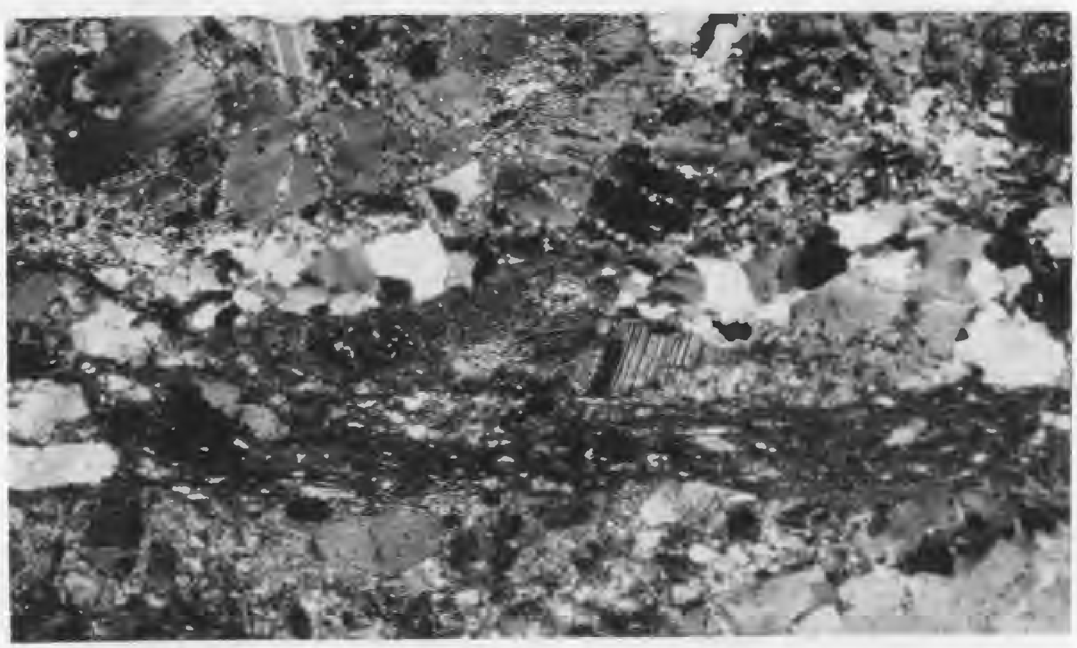
b: Photomicrograph of deformed granite, Sugarloaf Slide Zone. Note narrow irregular crush zones. Shore south of Sugarloaf Cove. Crossed nicols.

PLATE 44



1 mm

Fig. a: Photomicrograph of kinked and broken plagioclase, Sugarloaf Slide Zone. Shore south of Sugarloaf Cove. Crossed nicols.



2 mm

b: Photomicrograph of Phyllonitized granite, Sugarloaf Slide Zone. Note parallel zones of granulated and schistose natrix. Crossed nicols.

PLATE 45

2 mm

Fig. a: Phyllonitized granite, Sugarloaf Slide Zone. Note highly granulated matrix with well developed schistosity that forms augen around aligned quartz and feldspar lenticles. Crossed nicols.

1 mm

b: Relict Plagioclase Porphyroclast enclosed in highly crushed and granulated matrix. Sugarloaf Slide Zone. Shore, south of Sugarloaf Cove. Crossed nicols.

PLATE 46

2 mm

Fig. a: Emplacement cleavage in greywacke from Maiden Point Slice. Note alignment of detrital material in S_1 plane that is folded by the post-emplacement F_2 fold. East side of Canada Head. Crossed nicols.

2 mm

b: Folded greenschist, Irish Limestone. Note S_1 slaty cleavage defined by aligned opagues, chlorite and sericite that is folded by open F_2 folds. South end of St. Julien Island. Obliquely crossed nicols.

PLATE 47

Fig. a: Folded greywackes and slates of the Goose Tickle Formation. Note asymmetry of folds. Looking northeast, east side of MacGrey Island. At the point in the background the greywackes are overlain by melange.

b: Acadian (F_2) folds in slates of the Goose Tickle Formation. Looking north, west of Canada Harbour.

PLATE 48

Fig. a: Refolded emplacement fold (F_1) in the Table Head Formation. Facing directions uncertain. Note steeply dipping S_2 cleavage. Burnt Point South shore of Canada Bay.

b: Folded semi-pelitic schists, Sugarloaf Schists. The folds refold the emplacement S_1 schistosity. Looking north, south of Sugarloaf Cove.

PLATE 49

Fig. a: Acadian folds in greywackes and slates of the Maiden Point Slice. The fold is upward facing on the cleavage that is the first fabric in the rock. Looking northeast, east of Big Spring Inlet.

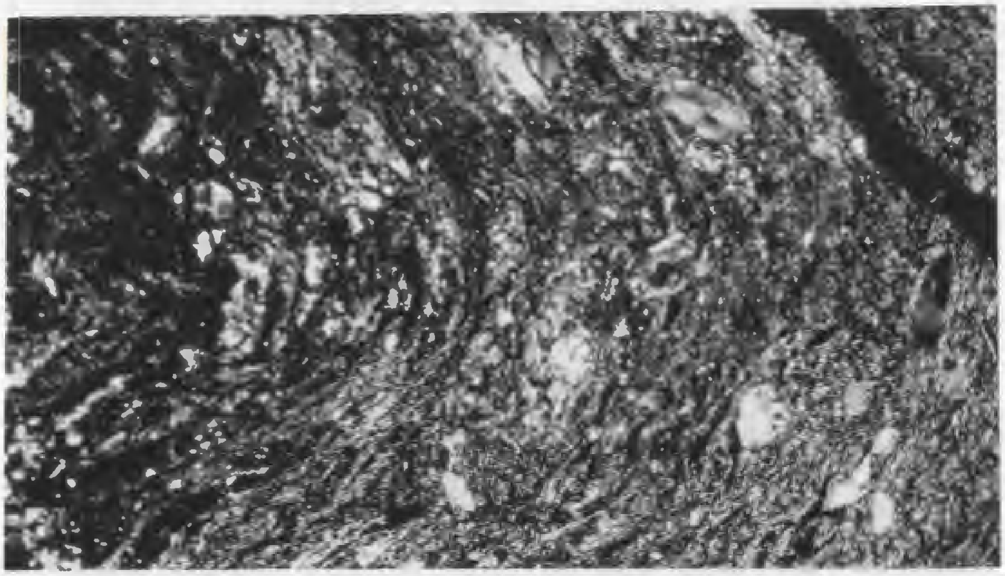
b: Acadian folds in mélange. Note the alignment of the blocks in the S_1 cleavage plane and the development of a crenulation (S_2) cleavage. North shore of Whites Arm.

PLATE 50

Fig. a: Relationship between S_1 and S_2 in greywackes from Maiden Point Slice. The F_1 folds face westwards on S_1 and the F_2 folds face down on S_2. Shore, east of Englee.

b: F_2 folds in greywackes from the Croque Head Slice. The S_2 fabric is a crenulation cleavage.

PLATE 51



2 mm

Fig. a: Semi-schists from Croque Head Slice. The emplacement S_1 cleavage is defined by aligned sericite and chlorite and by a strong dimensional preferred orientation of the detrital material. The clastic texture is partially destroyed. Crossed nicols.

b: Acadian (F_2) fold in St. Julien Island Formation. The fold faces up on the S_2 cleavage. Black Island.

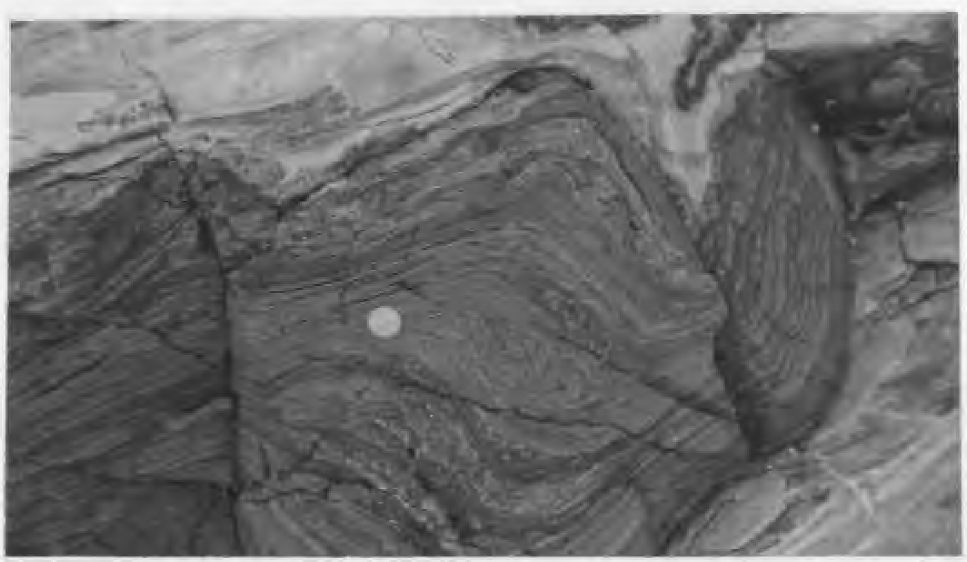
PLATE 52

Fig. a: Alignment of clasts in conglomerate, St. Julien Island Formation. Note the subspherical shape of the quartzite clasts (white) and the elongate shape of the sandstone and shale clasts. St. Julien Island.

b: Steeply folded contact between Croque Head Slice below and White Hills Slice above. The gully is underlain by mélange separating the slices. Looking northeast, south of Square Cove.

PLATE 53

Fig. a: Acadian folds (F_3) in Greenschist Member, Goose Cove Formation. Looking northeast, Fishot Island.



b: Refolded fold in marbles, Amphibolite Member, Goose Cove Formation. South side of Easter Tickle, Fishot Island.

PLATE 54

Shear Zone in greenschists, Goose Cove Formation. Note drag of S_2 fabric into left hand side of shear zone. Zone is marked by brown stained, carbonate veins. West side of Landing Cove, Fishot Island.

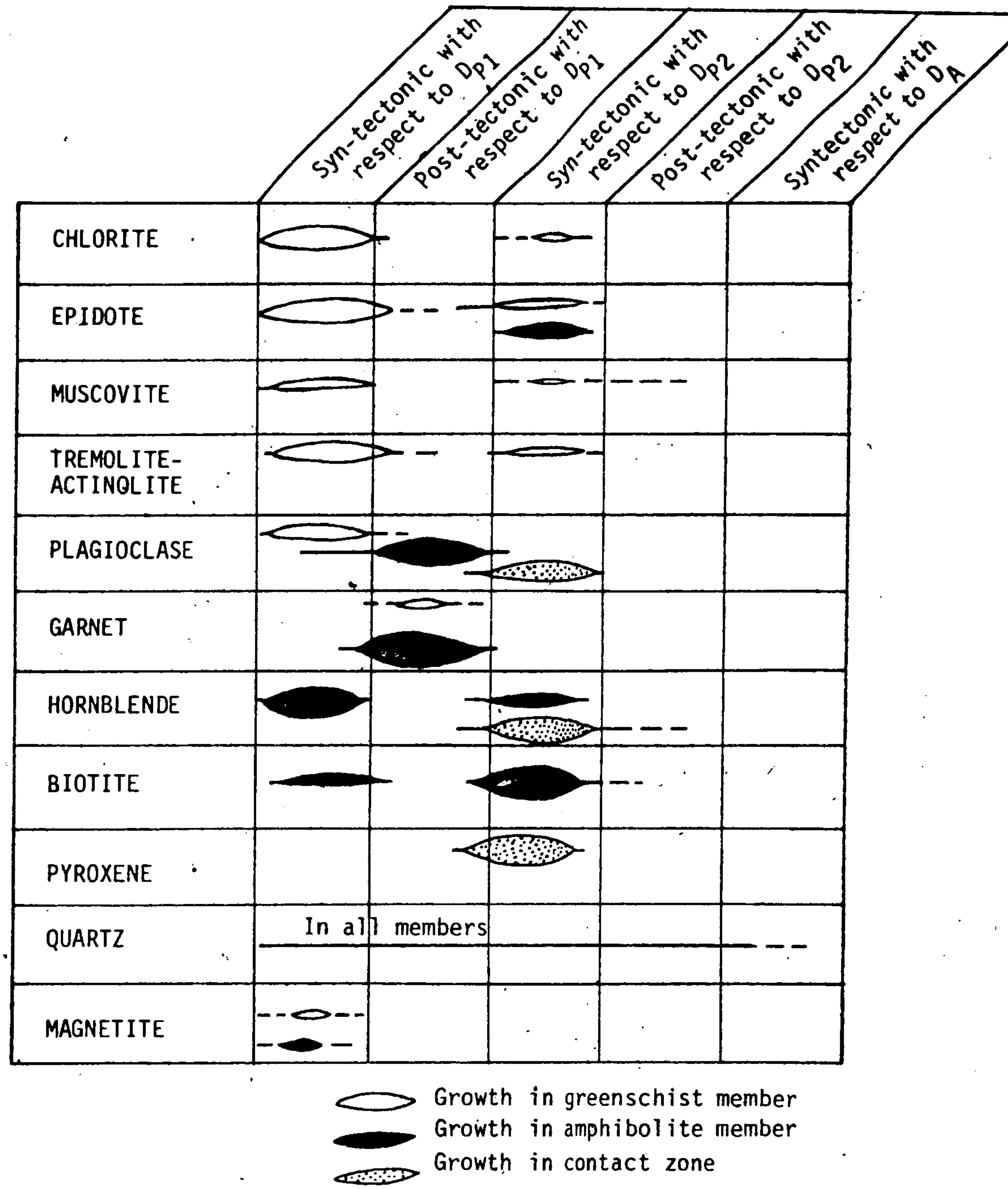


Fig I9 : The relationship between mineral growth and deformation in the Goose Cove Formation.

PLATE 55

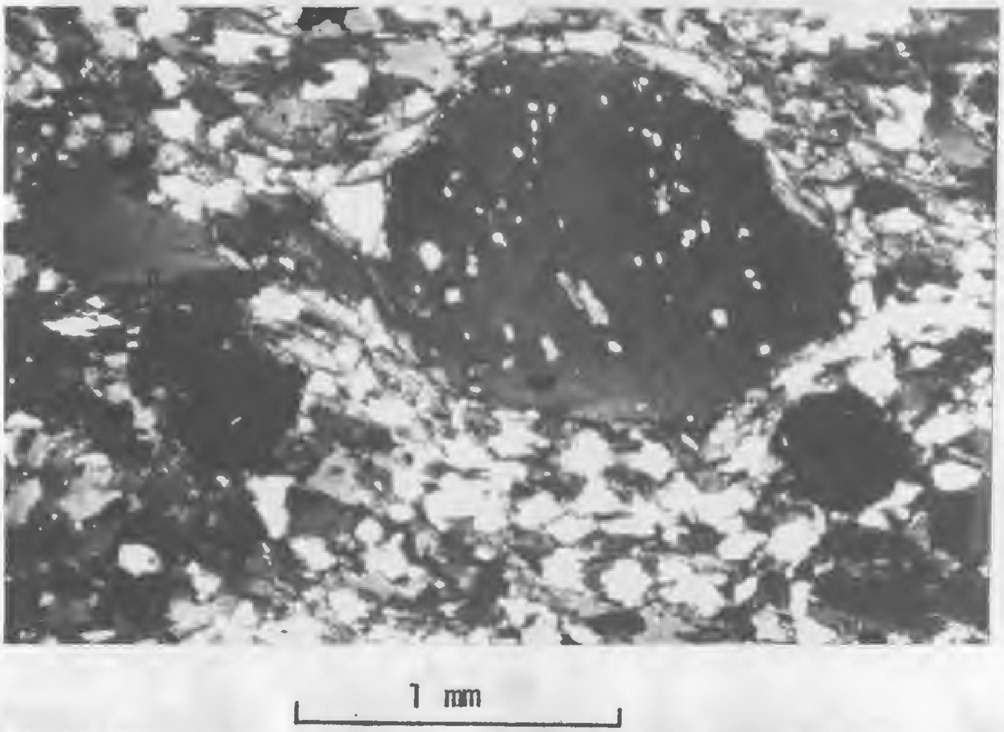
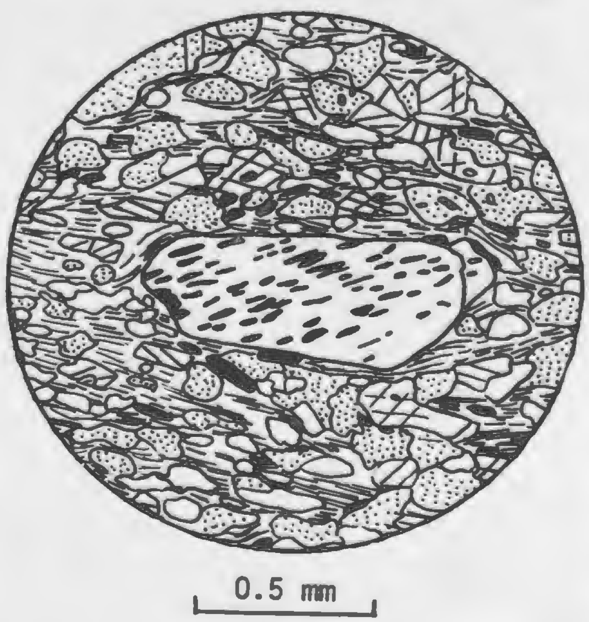
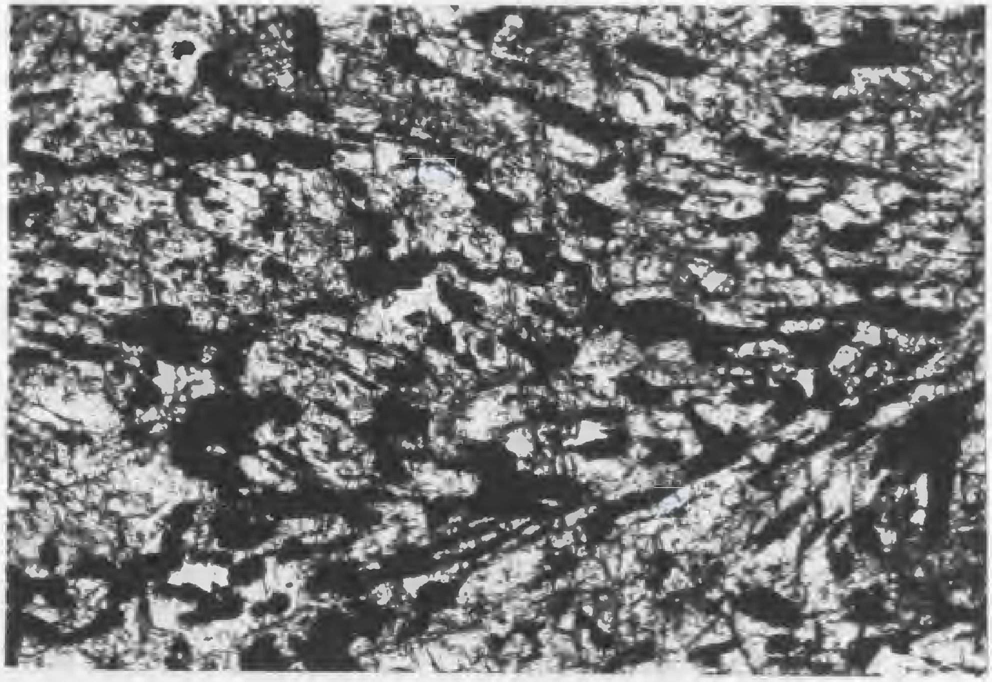


Fig. a: Plagioclase porphyroblast, Amphibolite Member, Goose Cove Formation. Note biotite and quartz inclusions. Crossed nicols.



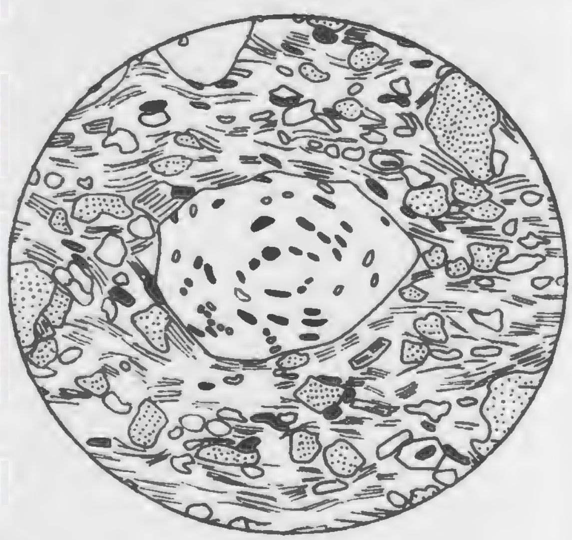
b: Garnet porphyroblast with straight inclusion trails. This is a line drawing of Plate 39, fig. a. The garnet clearly grew after formation of the S_1 fabric. Goose Cove Formation. Fishot Island. Garnet heavy outline and clear; plagioclase stippled; quartz clear; biotite hatched; horblende closely spaced cleavages.

PLATE 56



0.3 mm

Fig. a: Photomicrograph of helicitic garnet porphyroblast. The folded aligned opaques suggest the presence of two periods of deformation prior to garnet growth. Amphibolite Member, East side of Landing Cove, Fishot Island. Plane polarized light.



0.5 mm

b: Garnet porphyroblast, Amphibolite Member, Goose Cove Formation. Note circular inclusion trails. Line drawing from a photograph. Fishot Island. Symbols as in Plate 55, fig. b.

PLATE 57

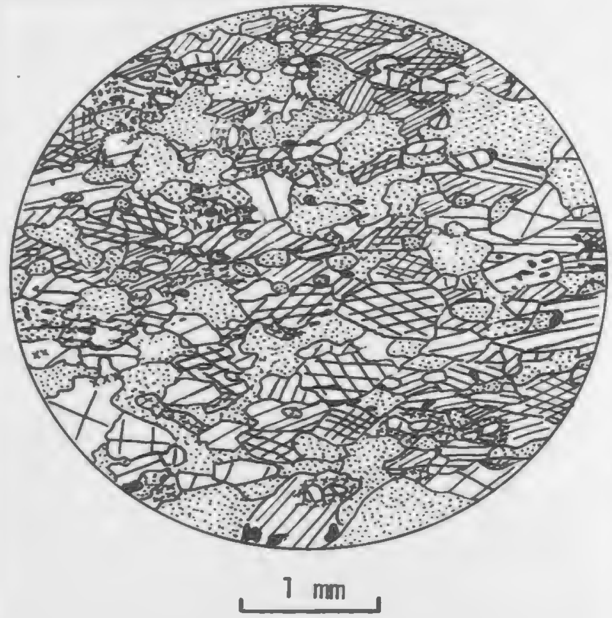
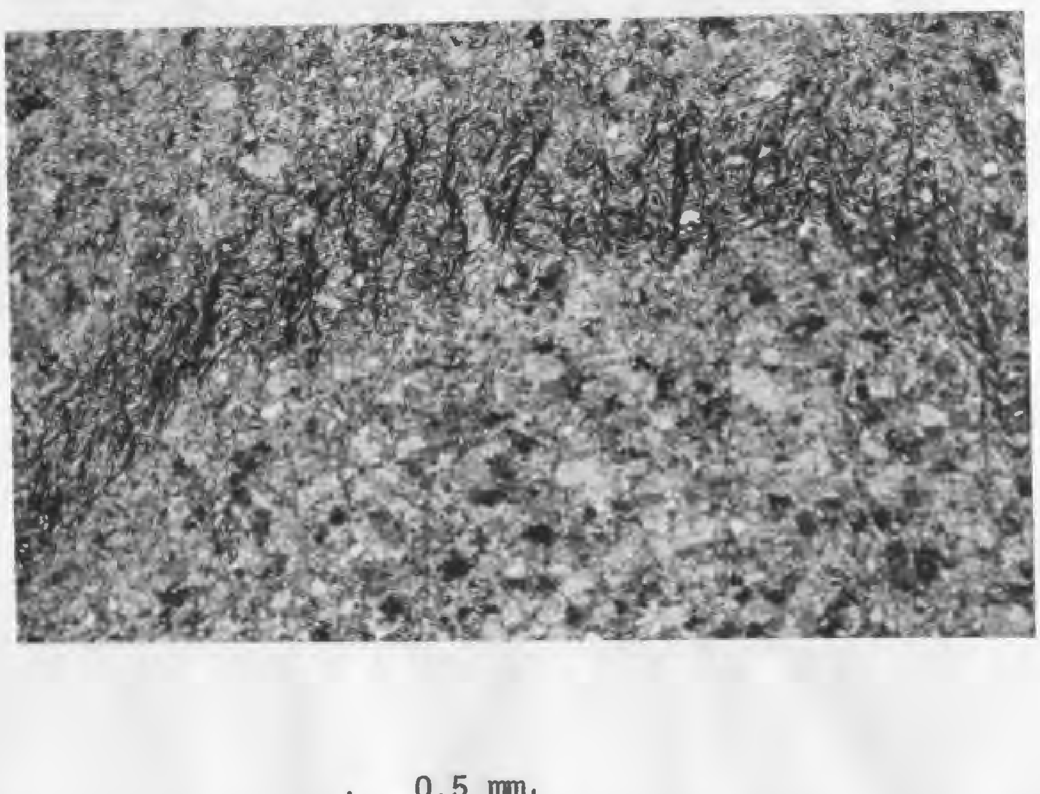


Fig. a: Pyroxene bearing amphibolite, Contact Zone, Goose Cove Formation. Line drawing from a photograph. Pyroxene widely spaced cleavages; other symbols as in Plate 55, fig. b.

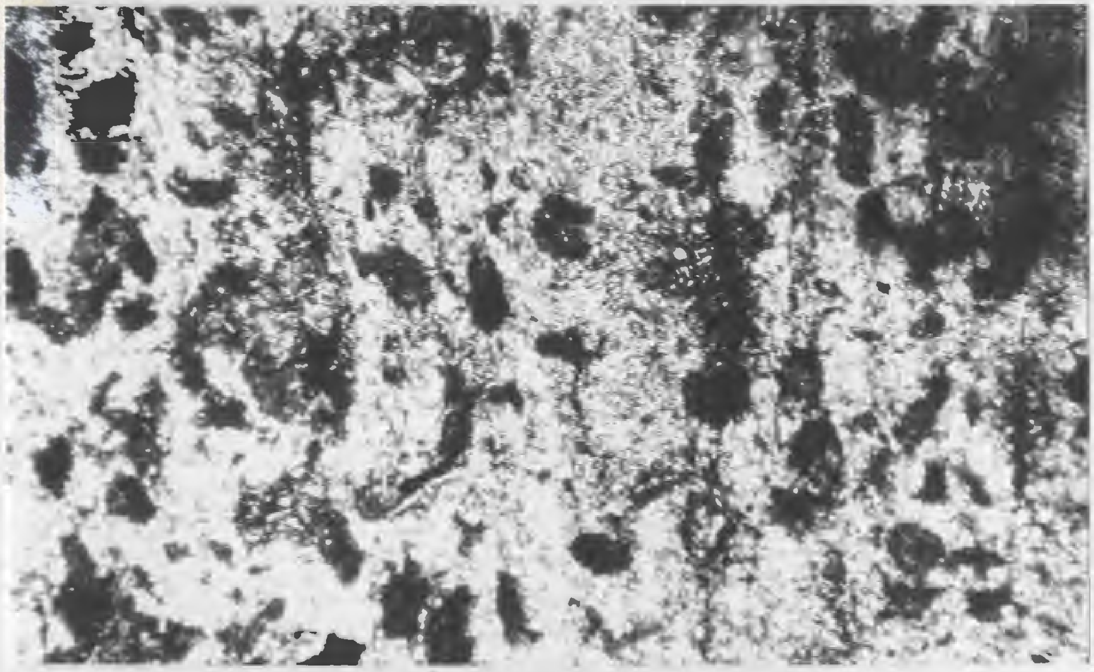


b: Emplacement slaty cleavage, Table Head Formation. S_1, defined by chlorite and seritite, is developed in a pelite band. Note development of S_2 crenulation cleavage. South of Englee. Crossed nicols.

	Metamorphic minerals		
WHITE HILLS PERIDOTITE SHEET			
Thrust	Plagioclase	Amphibole	Pyroxene
GOOSE COVE FORMATION: Contact Zone	An_{40}	Brown Hornblende	Augite
GOOSE COVE FORMATION: Amphibolite Member	An_{36-40}	Green Hornblende	—
GOOSE COVE FORMATION: Greenschist Member	An_{8-10}	Tremolite-Actinolite	—

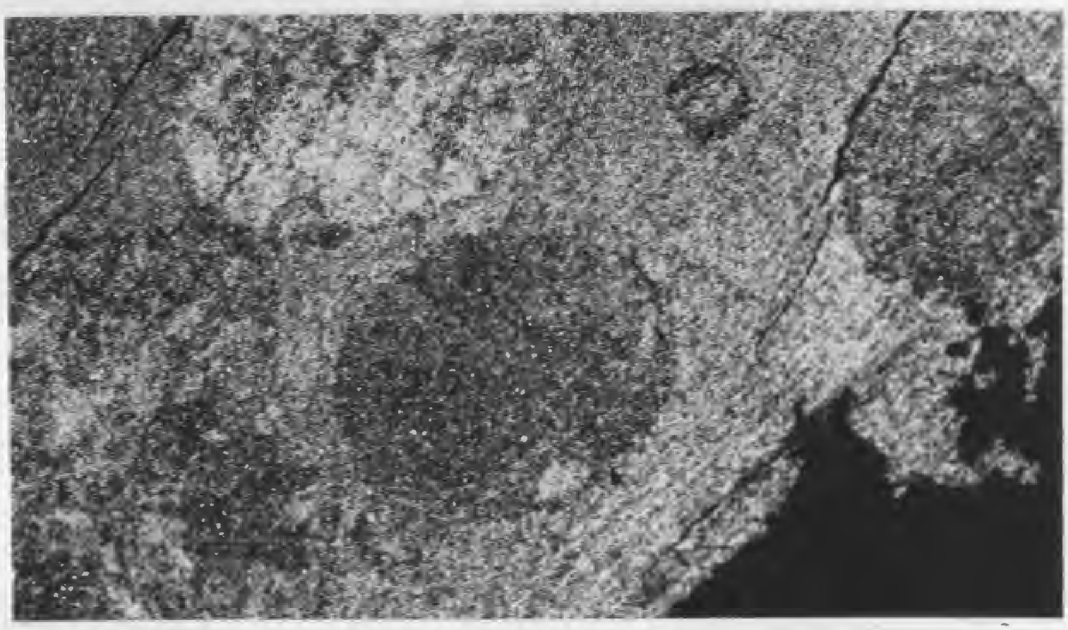
Fig 20: Summary of growth of major minerals in the Goose Cove Formation.

PLATE 58



1 mm

Fig. a: Post-tectonic chlorite porphyroblasts in semi-pelite, Maiden Point Formation. The chlorites (dark) overgrew and include the D_A (S_2) crenulation cleavage. Northeast side of Twillingate. Cove. Crossed nicols.



1 mm

b: Cordierite porphyroblasts in hornfels. Maiden Point Formation. Note inclusion packed prophyroblasts. Shore, Northeast of Cat Cove. Crossed Nicols.

PLATE 59

1 mm

Fig. a: Post-tectonic andalusite overgrowing S_2 crenulation cleavage, Maiden Point Formation. Shore, Northeast of Cat Cove. Crossed Nicols.

0.1 mm

b: Post-tectonic biotite, Sugarloaf Schists. Note biotite overgrowing the S_2 fabric. Plane polarized light.

PLATE 60

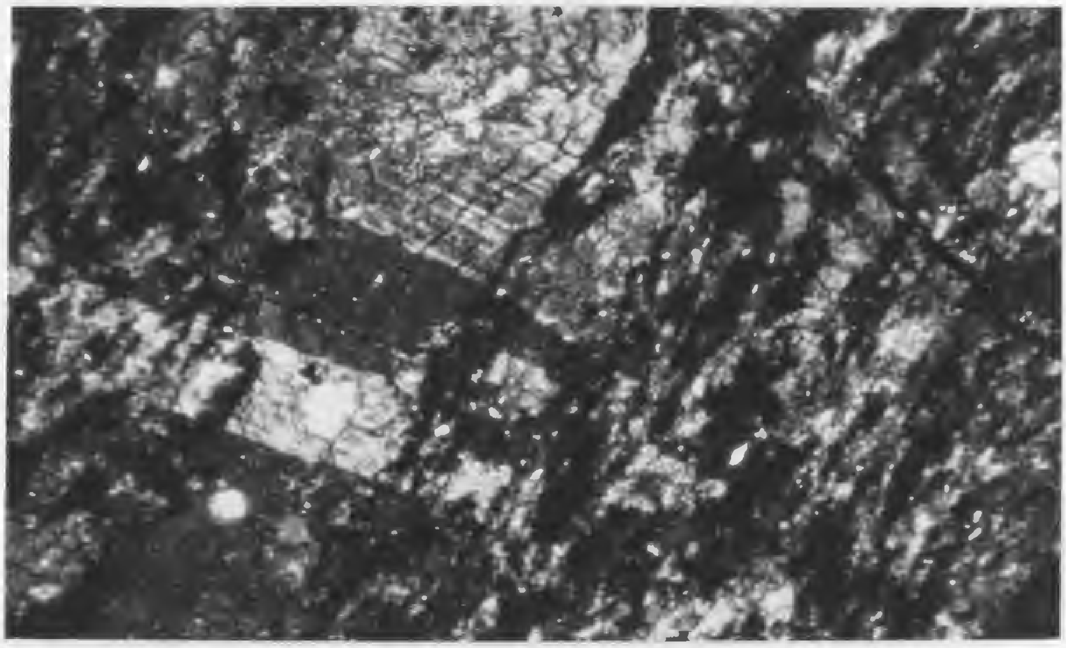
0.1 mm

Fig. a: Post-tectonic epidote porphyroblast, Sugarloaf Schists. Note included fabric. Northeast side of Sugarloaf Cove. Crossed nicols.

0.3 mm

b: Post-tectonic actinolite, Sugarloaf Schists. South side of Sugarloaf Cove. Obliquely crossed nicols.

PLATE 61



0.5 mm

Fig. a: Polysynthetic twinning in post-tectonic diopside, Sugarloaf Slide Zone. Note included fabric. Shore, South of Sugarloaf Cove. Crossed nicols.

b: Recumbent fold in White Hills Peridotites. The dark bands are orthopyroxene rich. South side, Western White Hills.

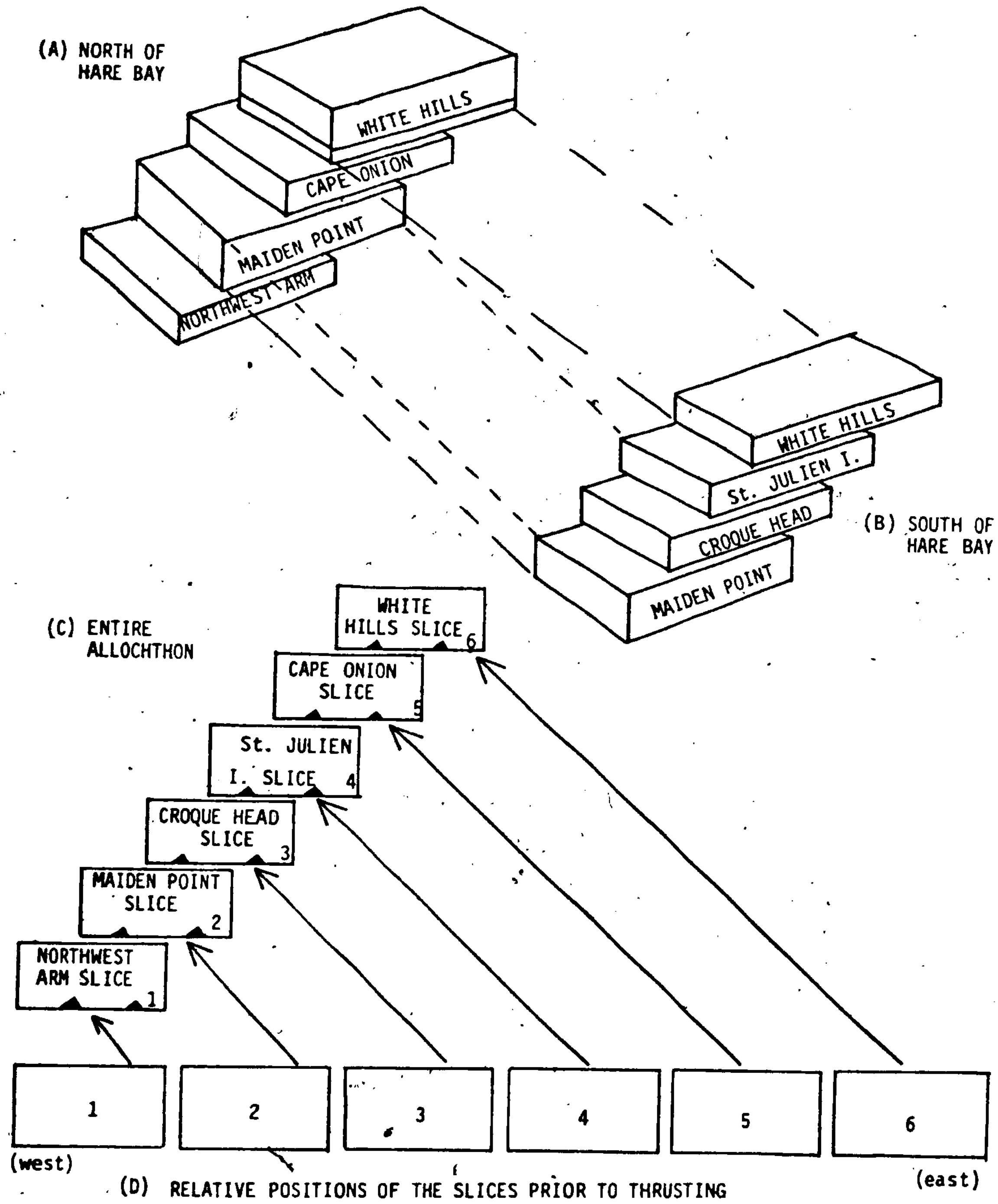
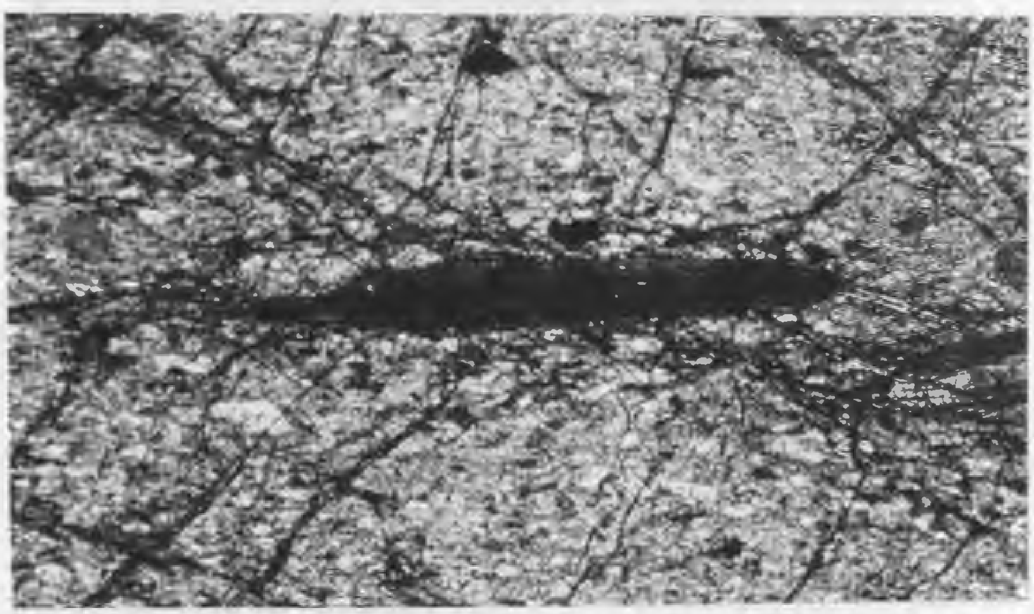


Fig 21: Palinspastic reconstruction of the thrust slices shown schematically.
(A) Structural stacking order of the slices north of Hare Bay.
(B) Structural stacking order of the slices south of Hare Bay.
(C) Structural stacking order of the slices of the entire Hare Bay Allochthon.
(D) Palinspastic reconstruction of the slices to the relative positions they had before emplacement, using the assumption that the higher slices originated farther from the continental margin in the west.

PLATE 62

Fig. a: Two ages of orthopyroxene bands, White Hills Peridotite Sheet. Western White Hills.



0.5 mm

b: Photomicrograph of harzburgite, White Hills Peridotite Sheet. Note enstatite augen. Southeast side, Eastern White Hills. Crossed nicols.

1 of

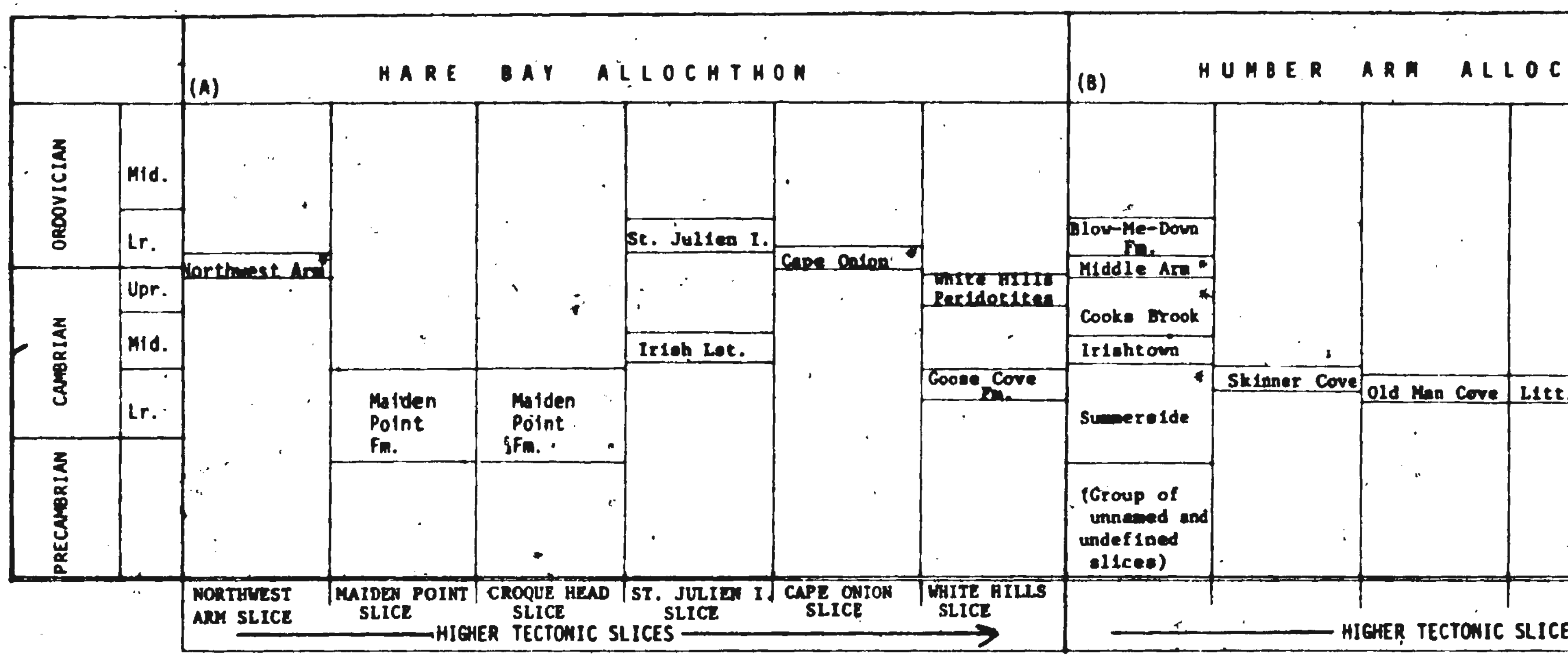


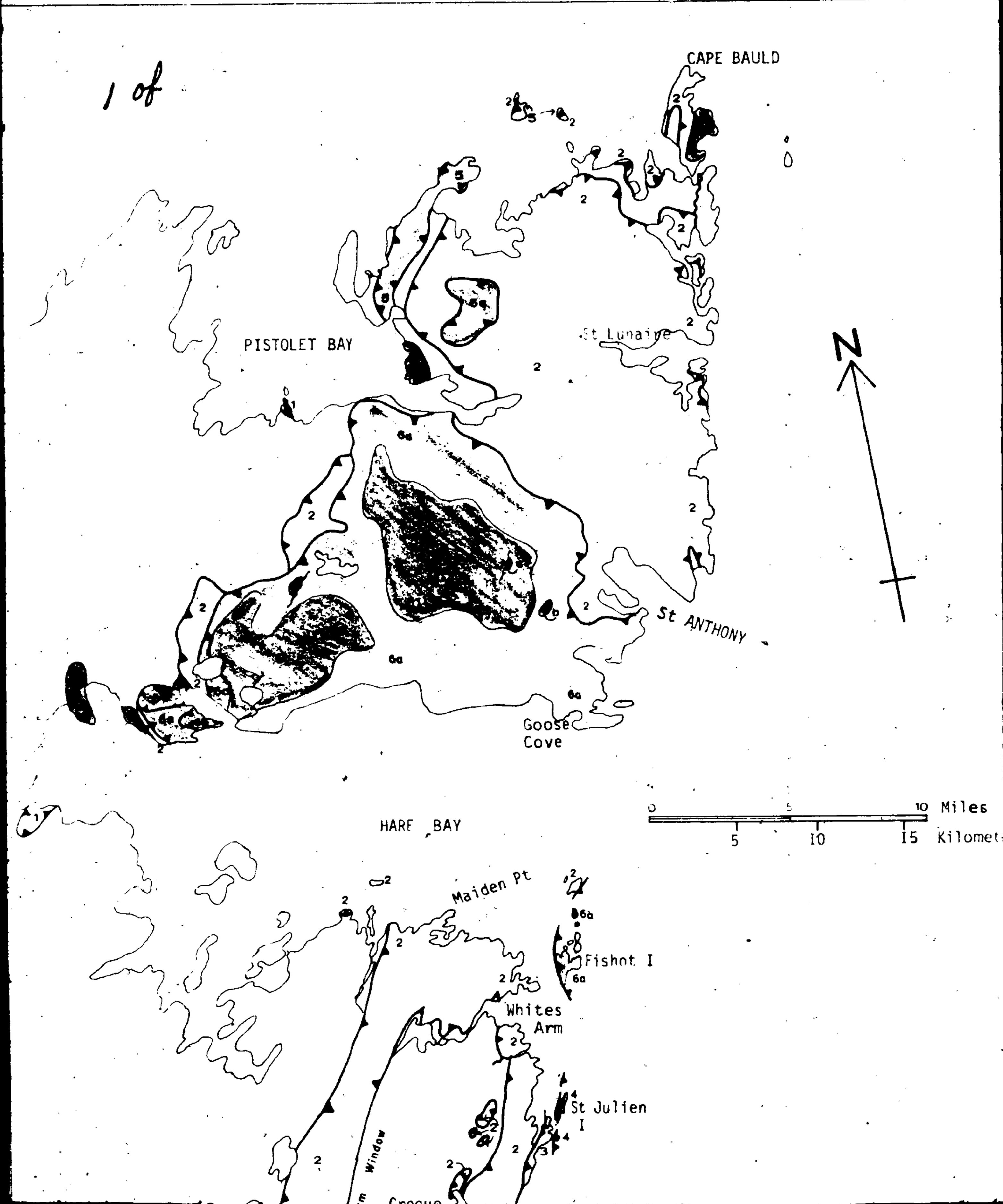
FIG. 22 : CORRELATION CHART SHOWING STRATIGRAPHIC SECTIONS OF THE 'TACONIC' ALLOCHTHONS FOUND ALONG THE WESTERN MARGIN OF THE NORTHERN APPALACHIANS.

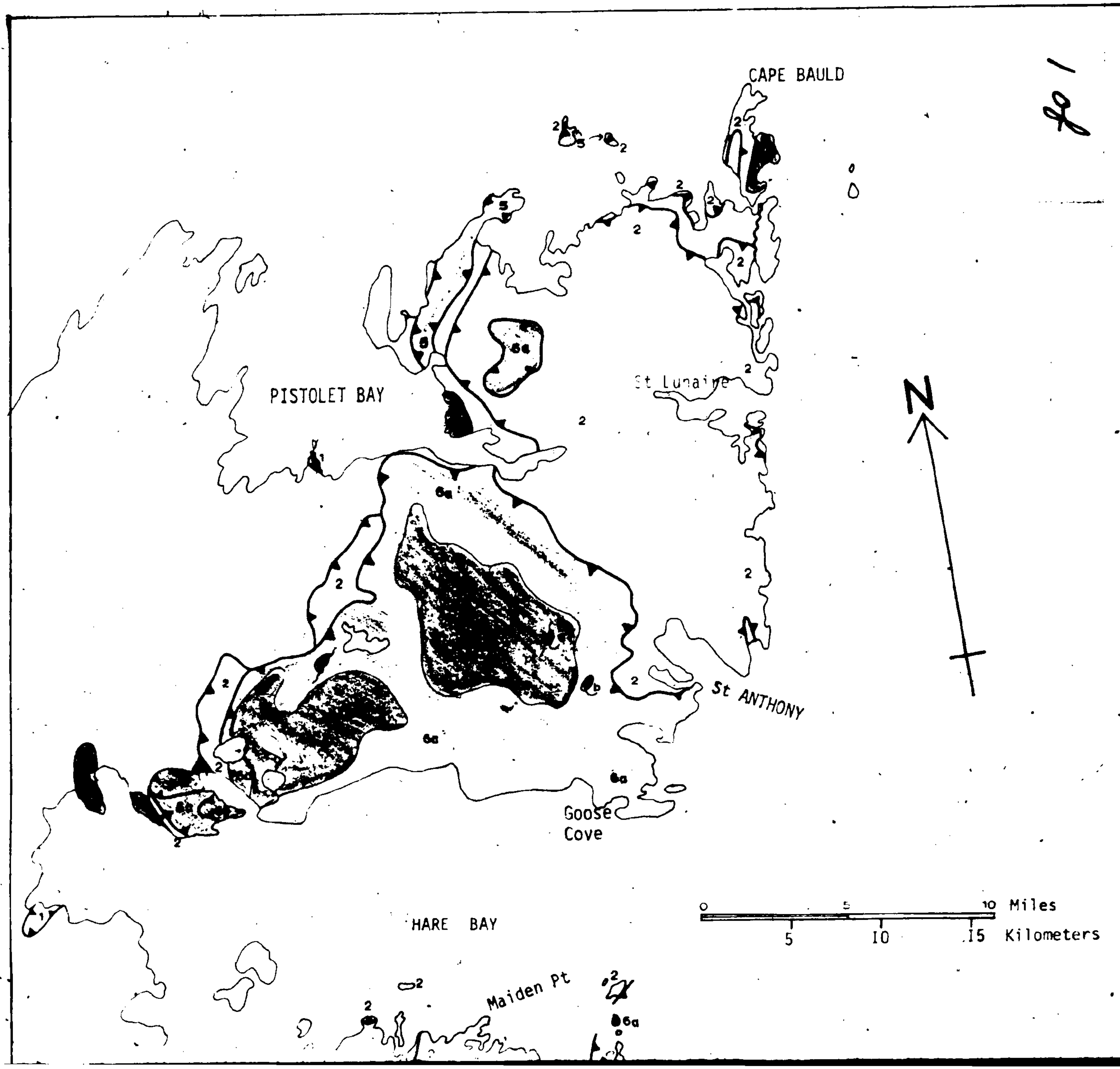
2 of 2

	(B) HUMBER ARM ALLOCHTHON					TACONIC ALLOCHTHON (C)		(D) QUEBEC CITY ALLOCHTHON		(E) GASPE ALLOCHTHON	
White Hills Peridotites Goose Cove Fm.	Blow-Me-Down Fm. Middle Arm Cooks Brook Irishtown Summerside (Group of unnamed and undefined slices)	Skinner Cove	Old Man Cove	Little Port	Bay of Islands Complex Aureole	Pawlet Poultney Hatch Hill West Castleton Bull Rensselaer (contained in 5 separate slices)	Greylock Bellowspipe Berkshire	Levis Fm. Sillery Gp. (Group of undefined slices)	Thetford Ophiolite	Quebec Group (Group of undefined slices)	Mt. Albert Pluton Shickshock
WHITE HILLS SLICE →	——— HIGHER TECTONIC SLICES ———→					LOW TACONICS	HIGH TACONICS	LOWER SLICES	UPPER SLICE	LOWER SLICES	UPPER SLICE

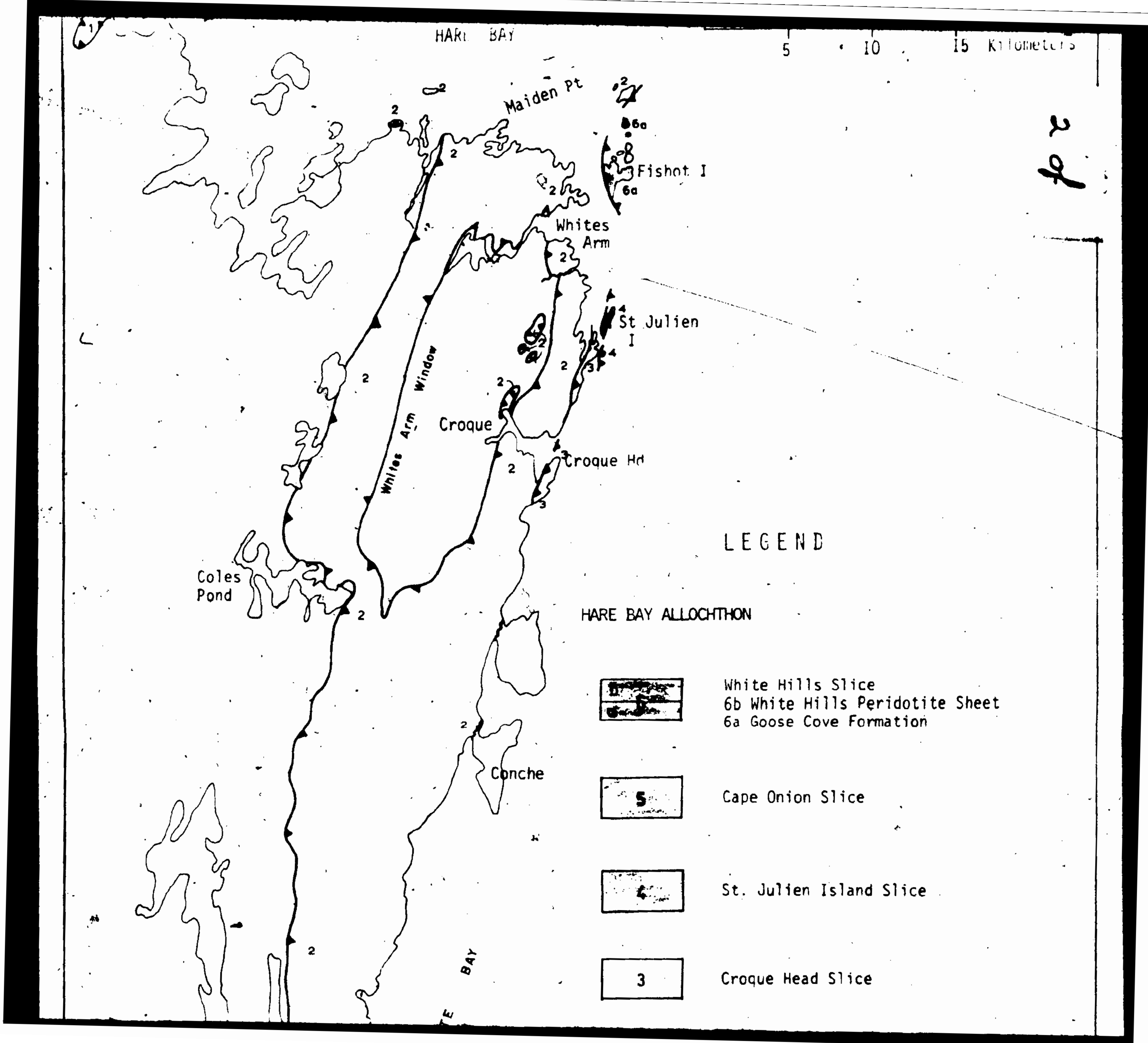
'TACONIC' ALLOCHTHONS FOUND ALONG THE WESTERN MARGIN OF

1 of


CAPE BAULD
PISTOLET BAY
St Lunaire
St ANTHONY
Goose Cove
HARE BAY
0 5 10 Miles
5 10 15 Kilomet
N
Maiden Pt
Fishot I
Whites Arm
St Julien I
Window
Croque
2
5
6a
1
3
4



CAPE BAULD
PISTOLET BAY
St Lunaire
St ANTHONY
Goose Cove
HARE BAY
Maiden Pt
N
0 5 10 Miles
5 10 15 Kilometers
2
5
6a
6b
1


HARE BAY
5 10 15 Kilometers
Maiden Pt
Fishot I
Whites Arm
St Julien I
Whites Arm Window
Croque
Croque Hd
Coles Pond
Conche
BAY
LEGEND
HARE BAY ALLOCHTHON
White Hills Slice
6b White Hills Peridotite Sheet
6a Goose Cove Formation
5
Cape Onion Slice
4
St. Julien Island Slice
3
Croque Head Slice

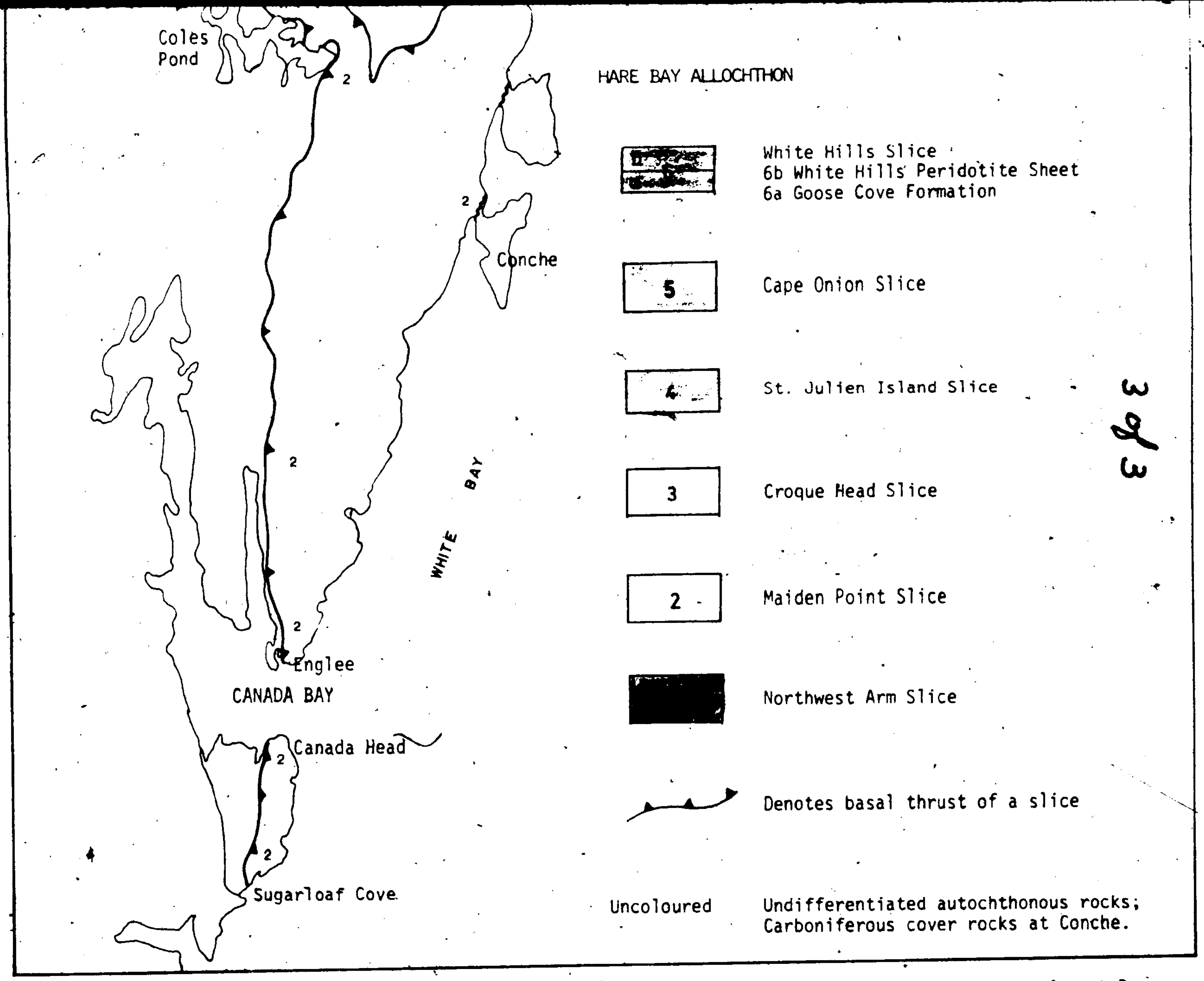
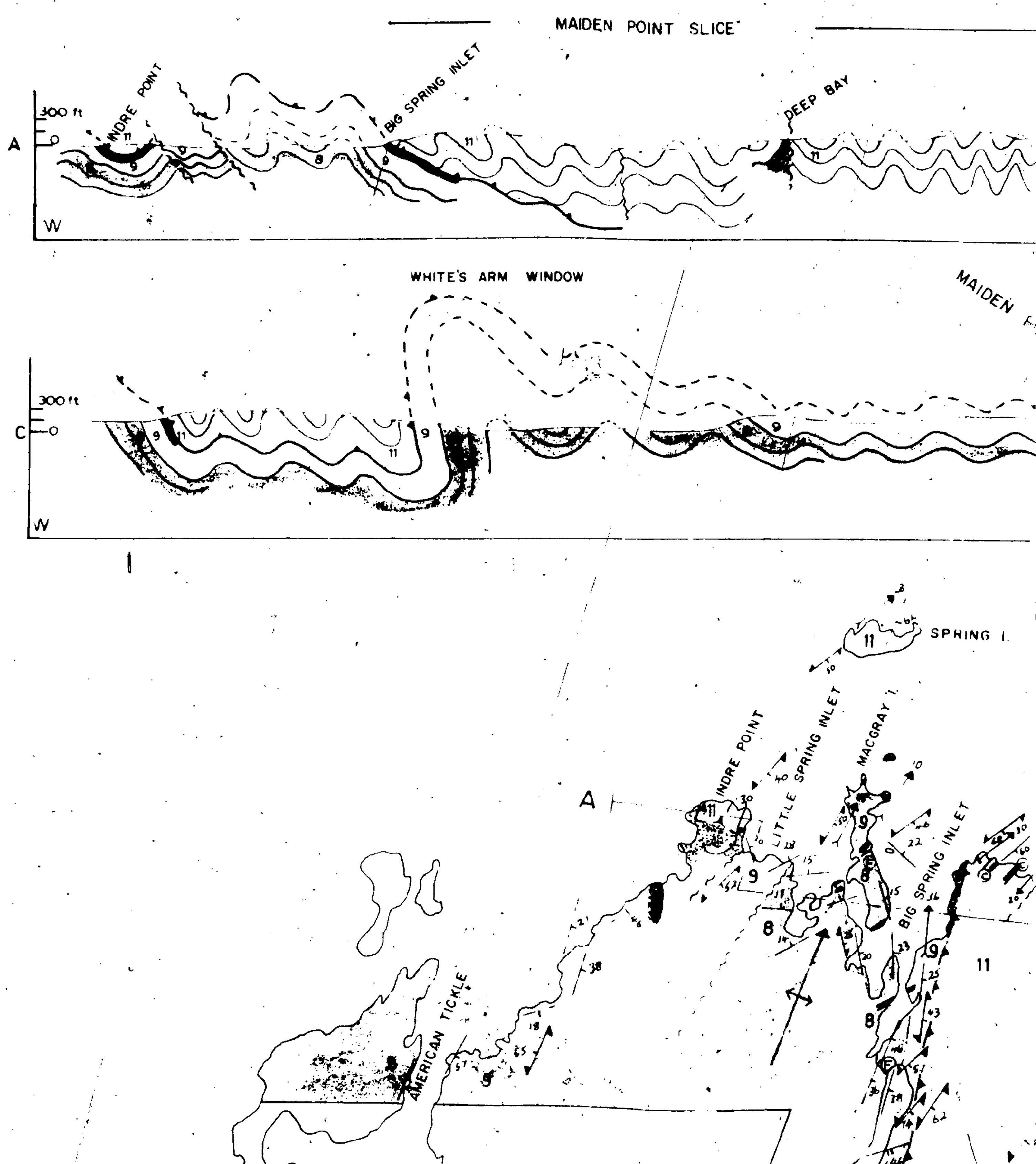
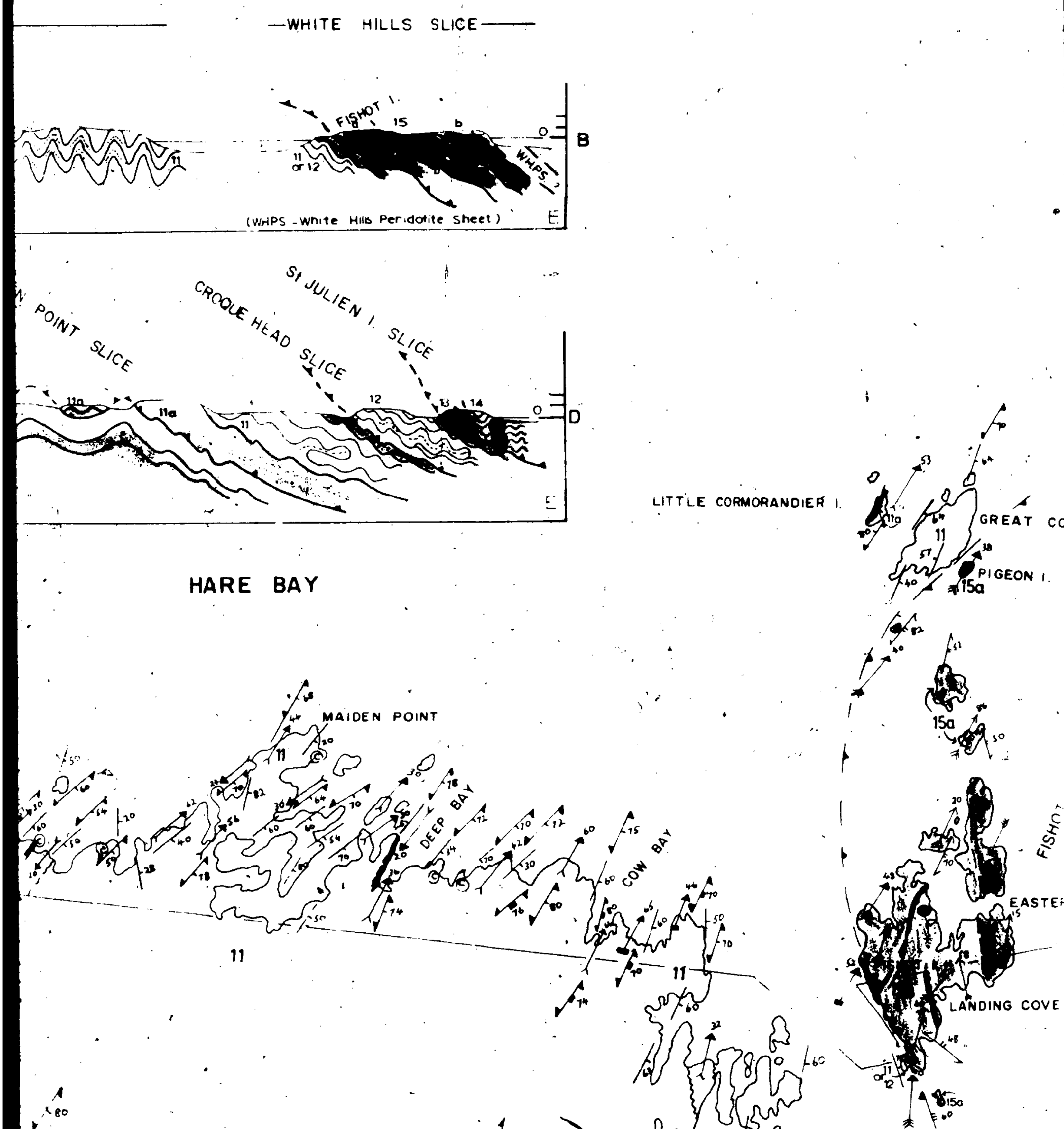


FIG. 2: GEOLOGIC MAP SHOWING THE DISTRIBUTION OF SLICES IN THE HARE BAY ALLOCHTHON. Geology north of Hare Bay after Williams, Smyth & Stevens (1973).

1 of

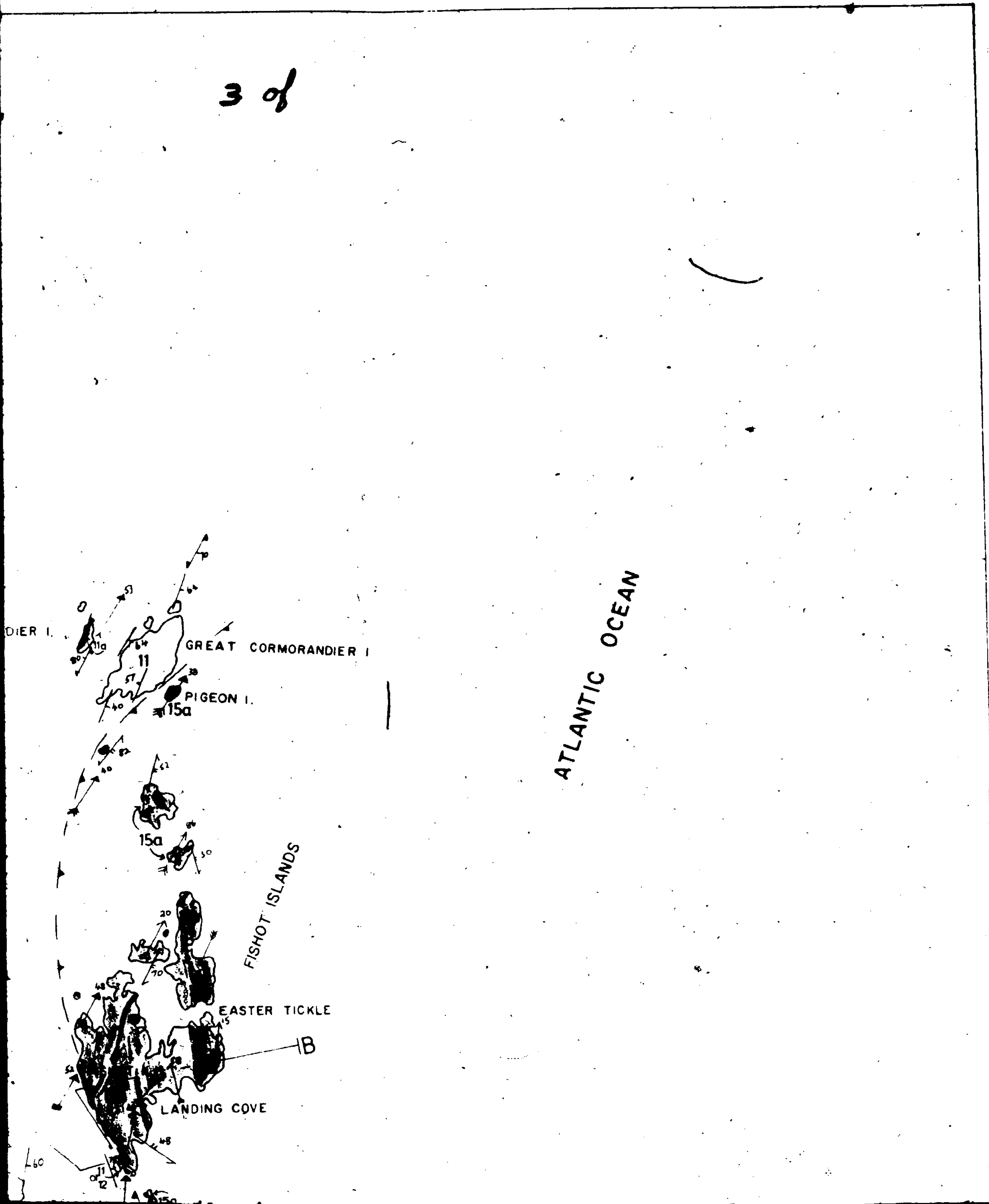

MAIDEN POINT SLICE
INDRE POINT
BIG SPRING INLET
DEEP BAY
300 ft
A
0
W
WHITE'S ARM WINDOW
MAIDEN
300 ft
C
0
W
SPRING I.
INDRE POINT
LITTLE SPRING INLET
MACGRAY I.
BIG SPRING INLET
AMERICAN TICKLE
A

2 of

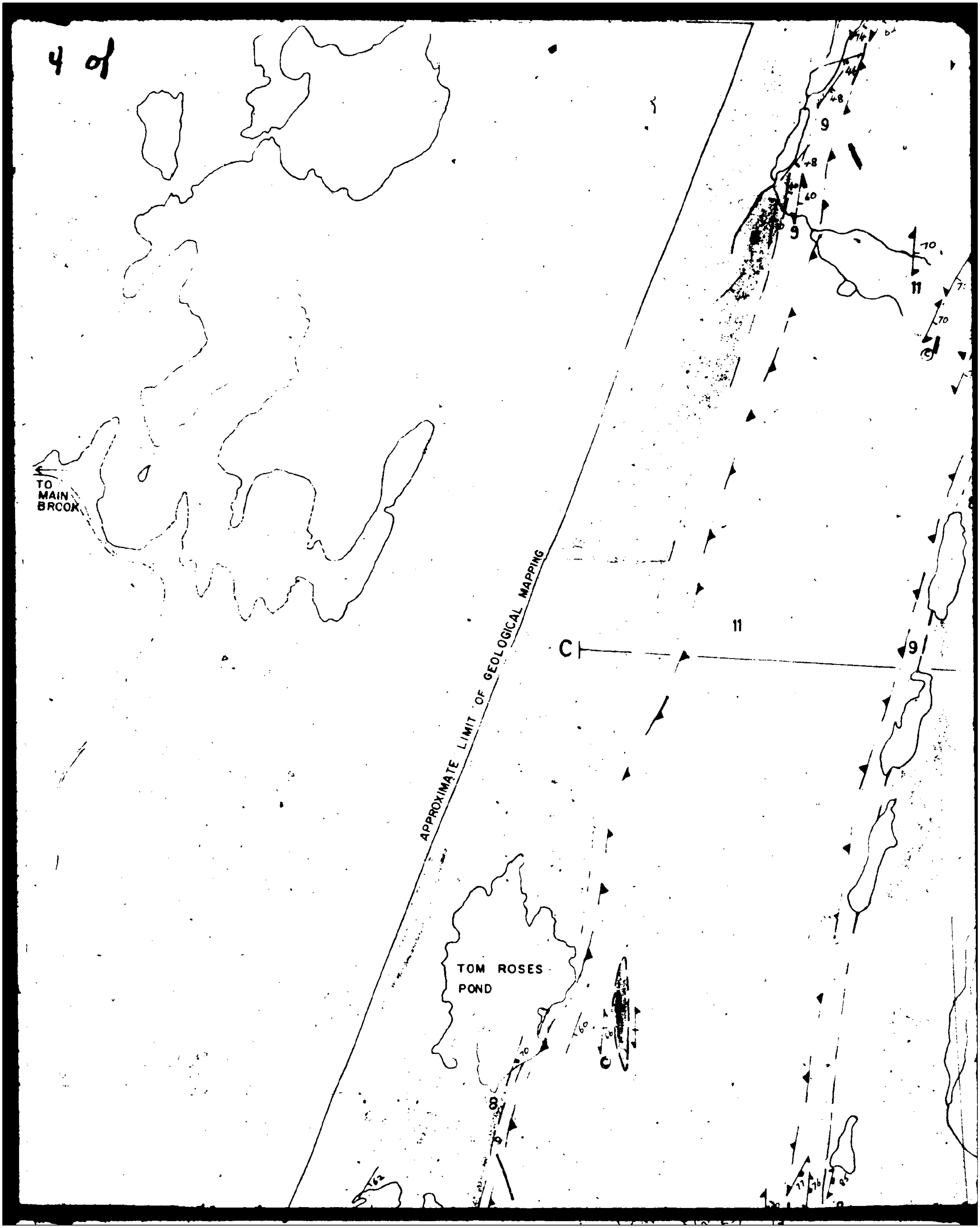


WHITE HILLS SLICE
FISHOT I.
15
b
11 or 12
11
WHPS
0
B
(WHPS -White Hills Peridotite Sheet)
POINT SLICE
CROQUE HEAD SLICE
St JULIEN I. SLICE
11a
12
13
14
0
D
LITTLE CORMORANDIER I.
GREAT CO
PIGEON I.
15a
HARE BAY
MAIDEN POINT
DEEP BAY
COW BAY
FISHOT
EASTE
LANDING COVE
11 or 12

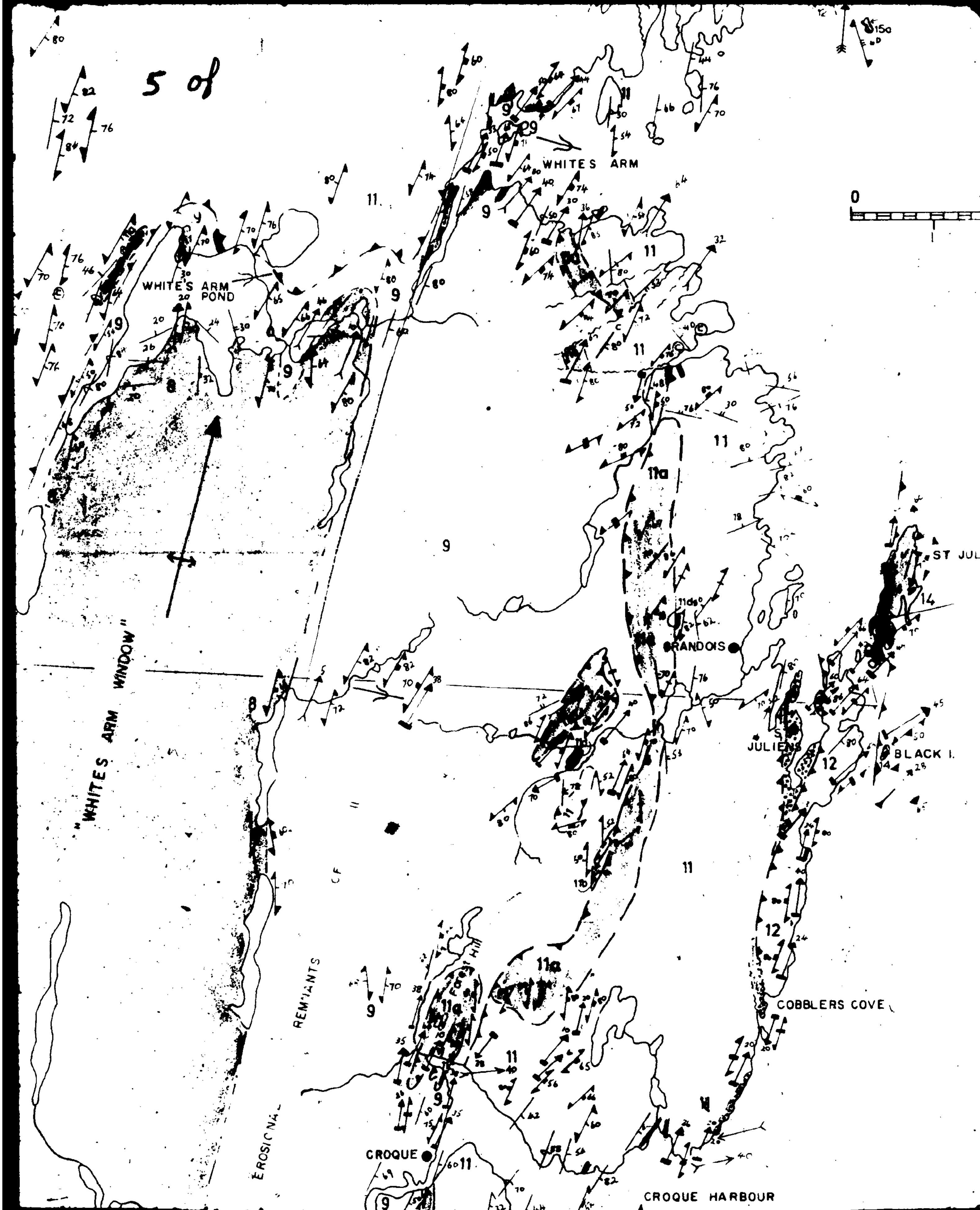
3 of

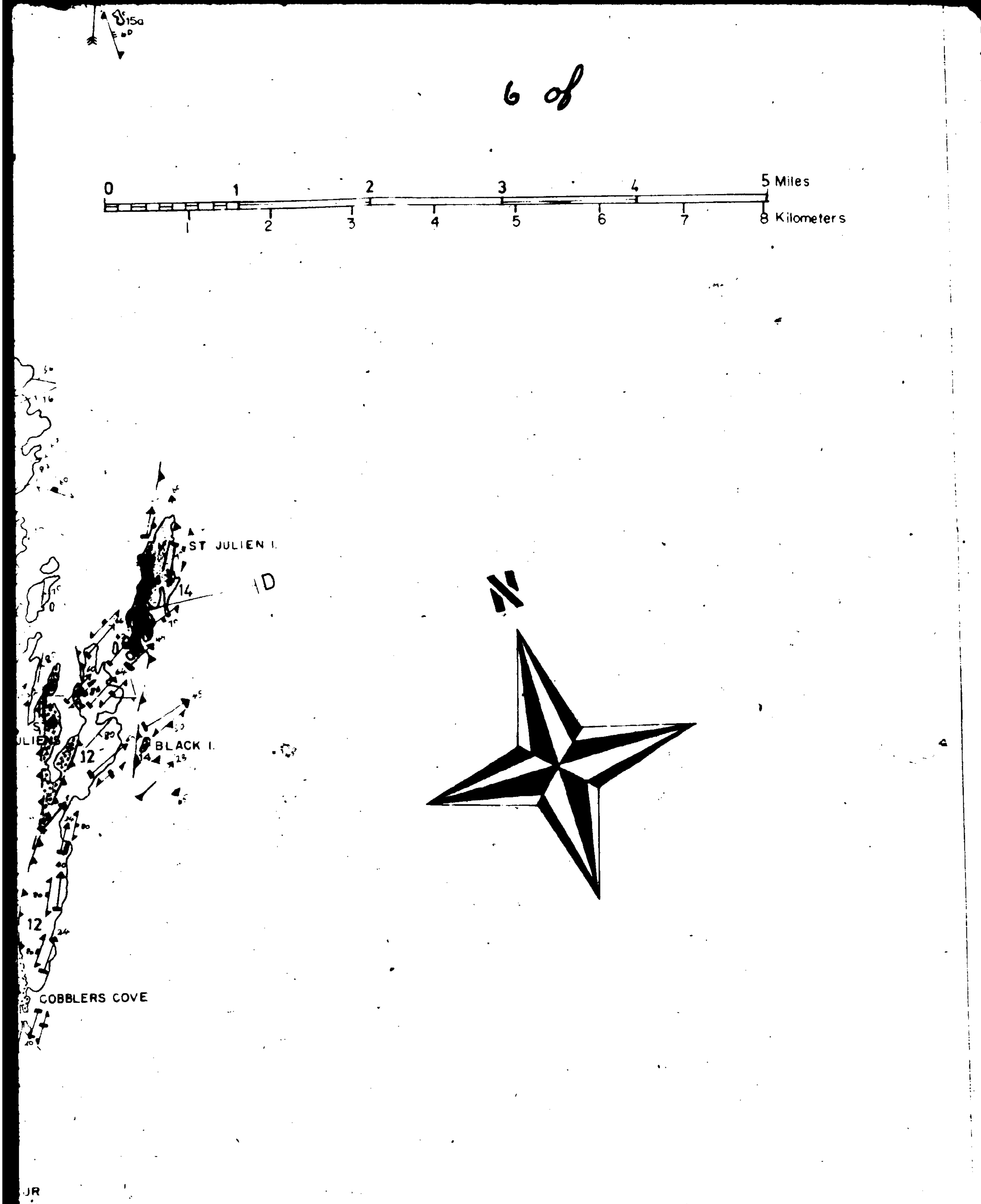

DIER I.
GREAT CORMORANDIER I
PIGEON I.
11
11a
15a
ATLANTIC OCEAN
FISHOT ISLANDS
EASTER TICKLE
B
LANDING COVE

4 of

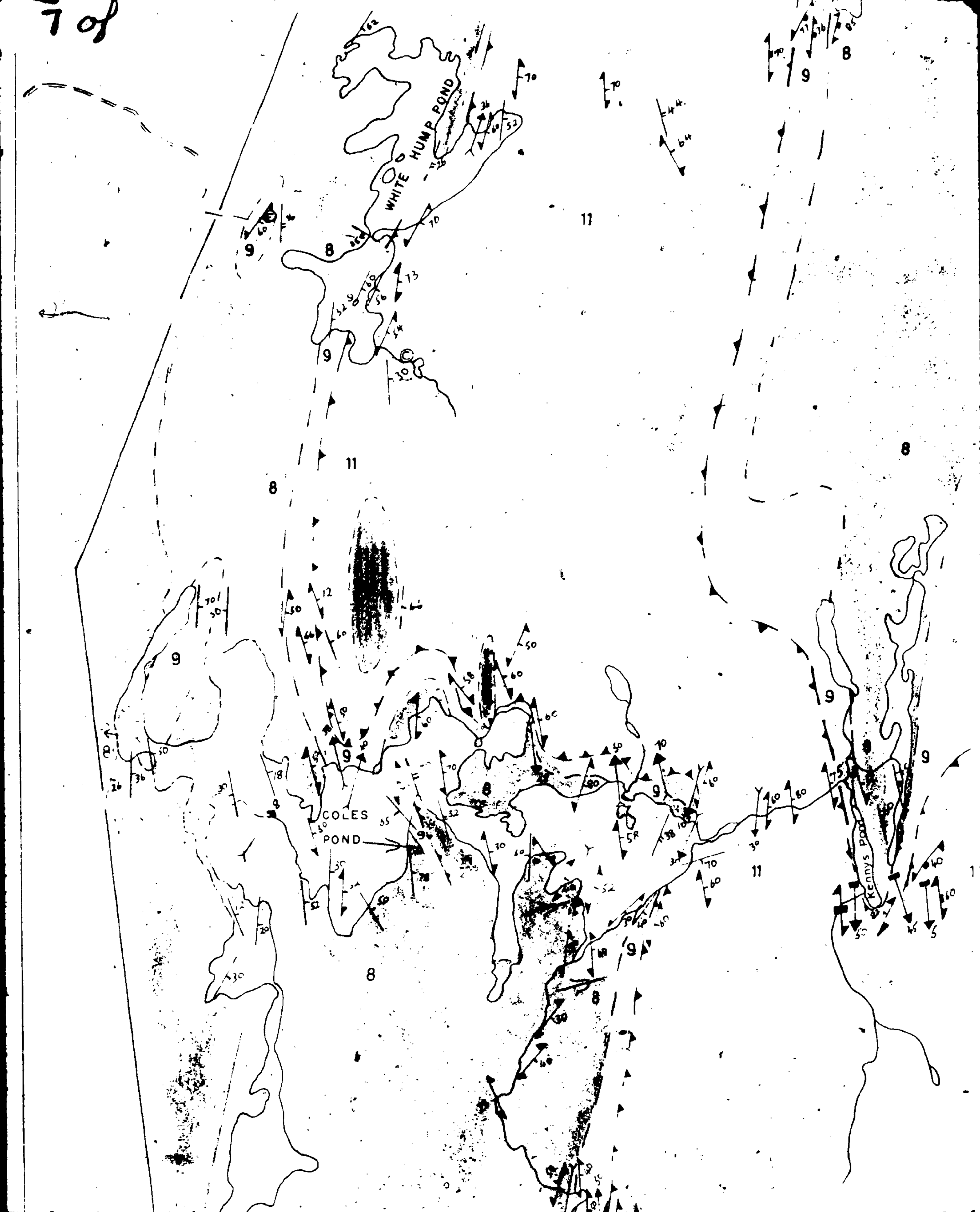


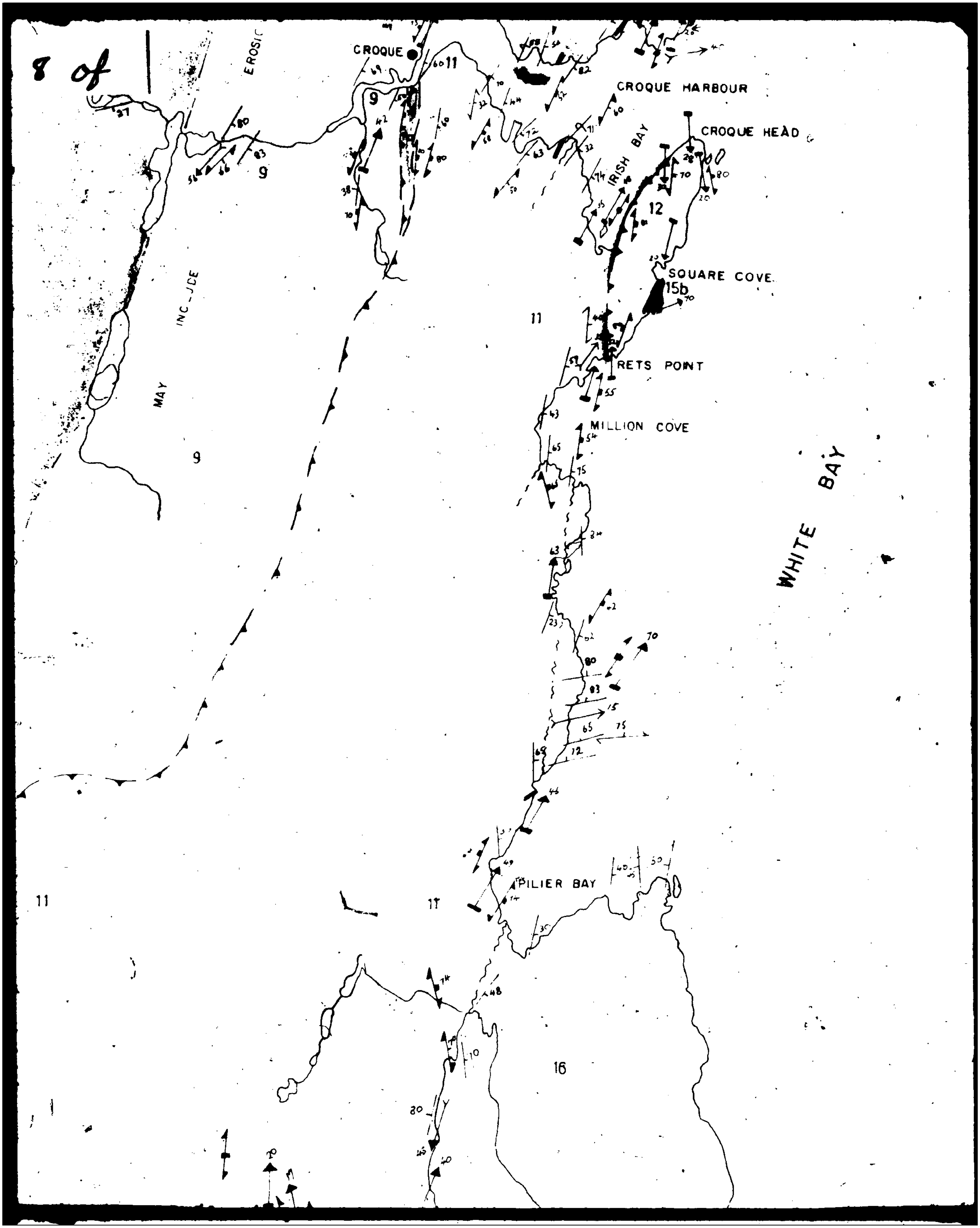
TO MAIN BROOK
APPROXIMATE LIMIT OF GEOLOGICAL MAPPING
TOM ROSES POND
C
11
9
8


WHITES ARM
WHITE'S ARM POND
"WHITES ARM WINDOW"
EROSIONAL
REMNANTS
ST JUL
RANDOIS
JULIENS
BLACK I.
COBBLERS COVE
CROQUE
CROQUE HARBOUR
11a
12
14

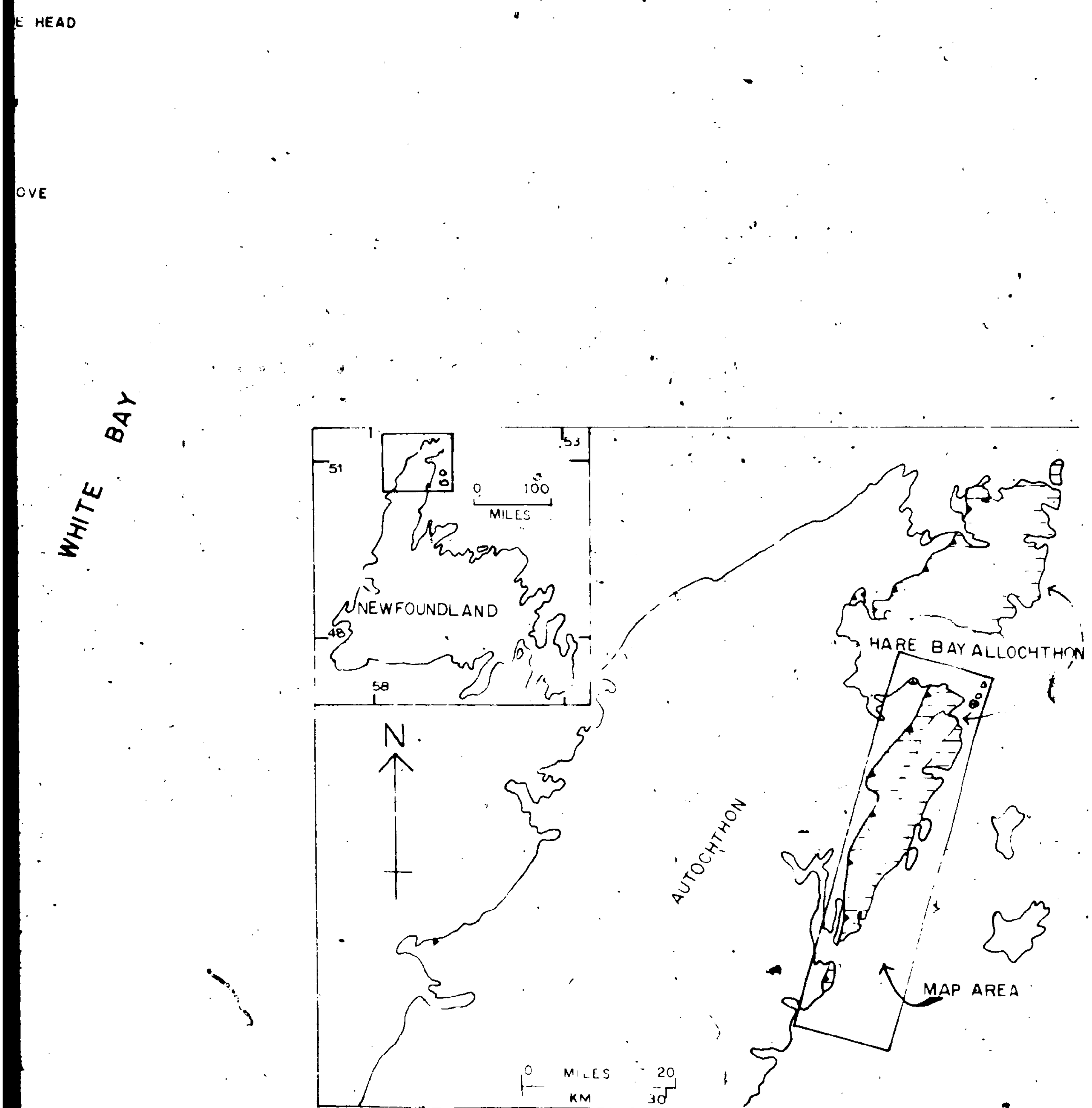

0 1 2 3 4 5 Miles
1 2 3 4 5 6 7 8 Kilometers
ST JULIEN I.
14
BLACK I.
12
COBBLERS COVE
N
JR

7 of



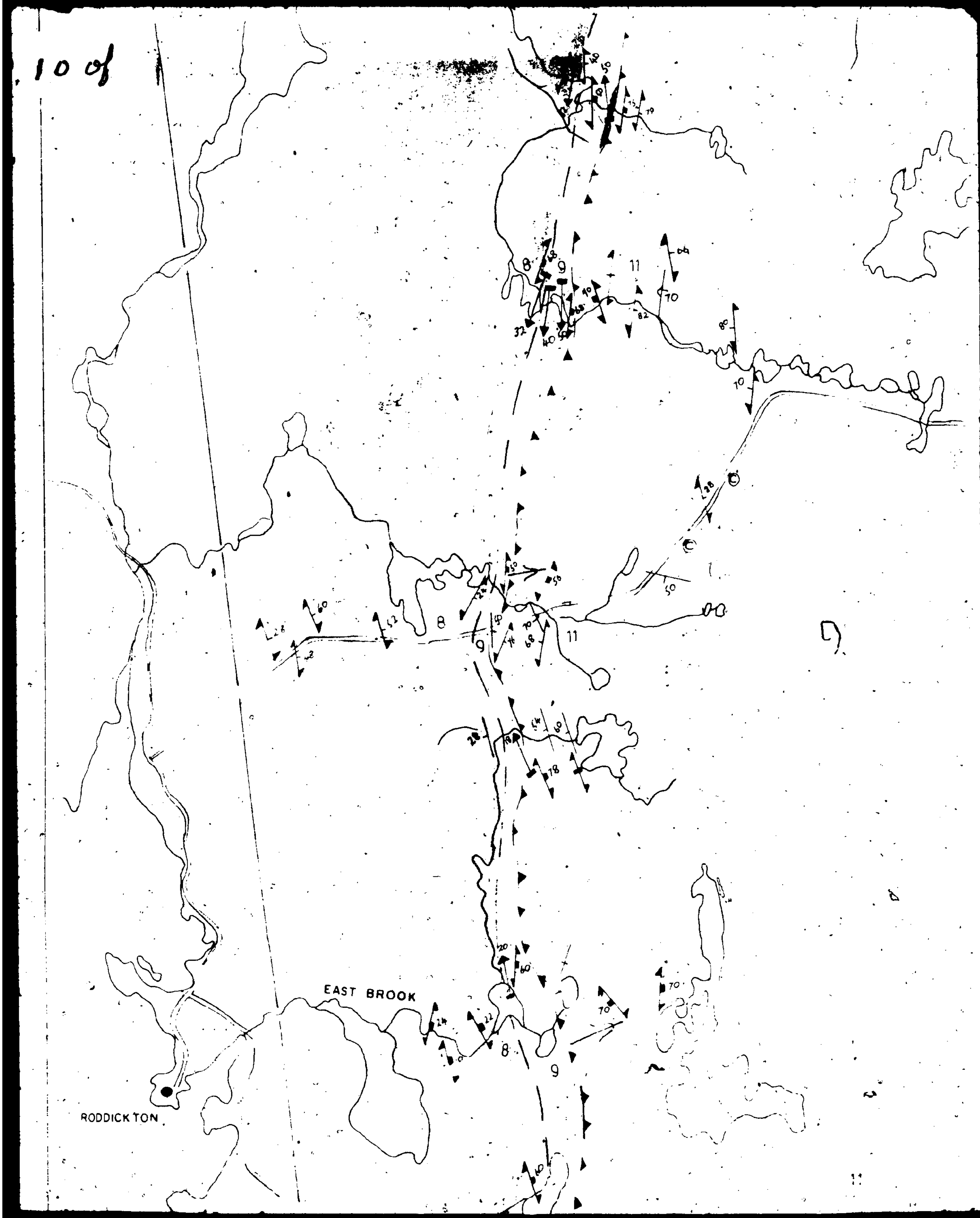

8 of
CROQUE
CROQUE HARBOUR
CROQUE HEAD
IRISH BAY
SQUARE COVE
RETS POINT
MILLION COVE
WHITE BAY
PILIER BAY
EROSIC
INC_JDE
MAY
9
11
12
15b
16

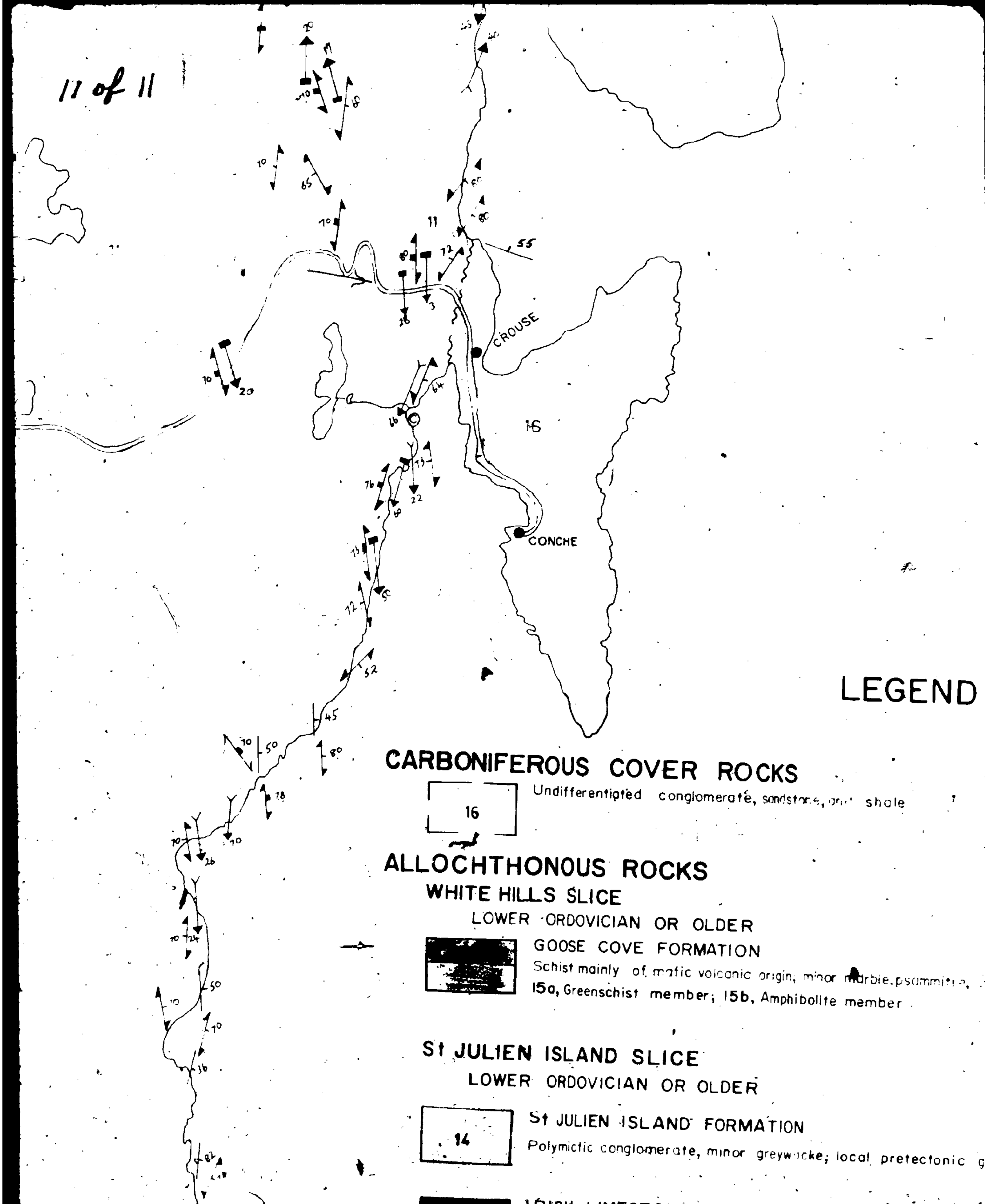
9 of



INDEX MAP

10 of




CROUSE
CONCHE
16
LEGEND
CARBONIFEROUS COVER ROCKS
16
Undifferentiated conglomerate, sandstone, and shale
ALLOCHTHONOUS ROCKS
WHITE HILLS SLICE
LOWER ORDOVICIAN OR OLDER
GOOSE COVE FORMATION
Schist mainly of mafic volcanic origin; minor marble, psammite,
15a, Greenschist member; 15b, Amphibolite member
St JULIEN ISLAND SLICE
LOWER ORDOVICIAN OR OLDER
14
St JULIEN ISLAND FORMATION
Polymictic conglomerate, minor greywacke; local pretectonic
IRISH LIMESTONE

12 of

LEGEND

KS

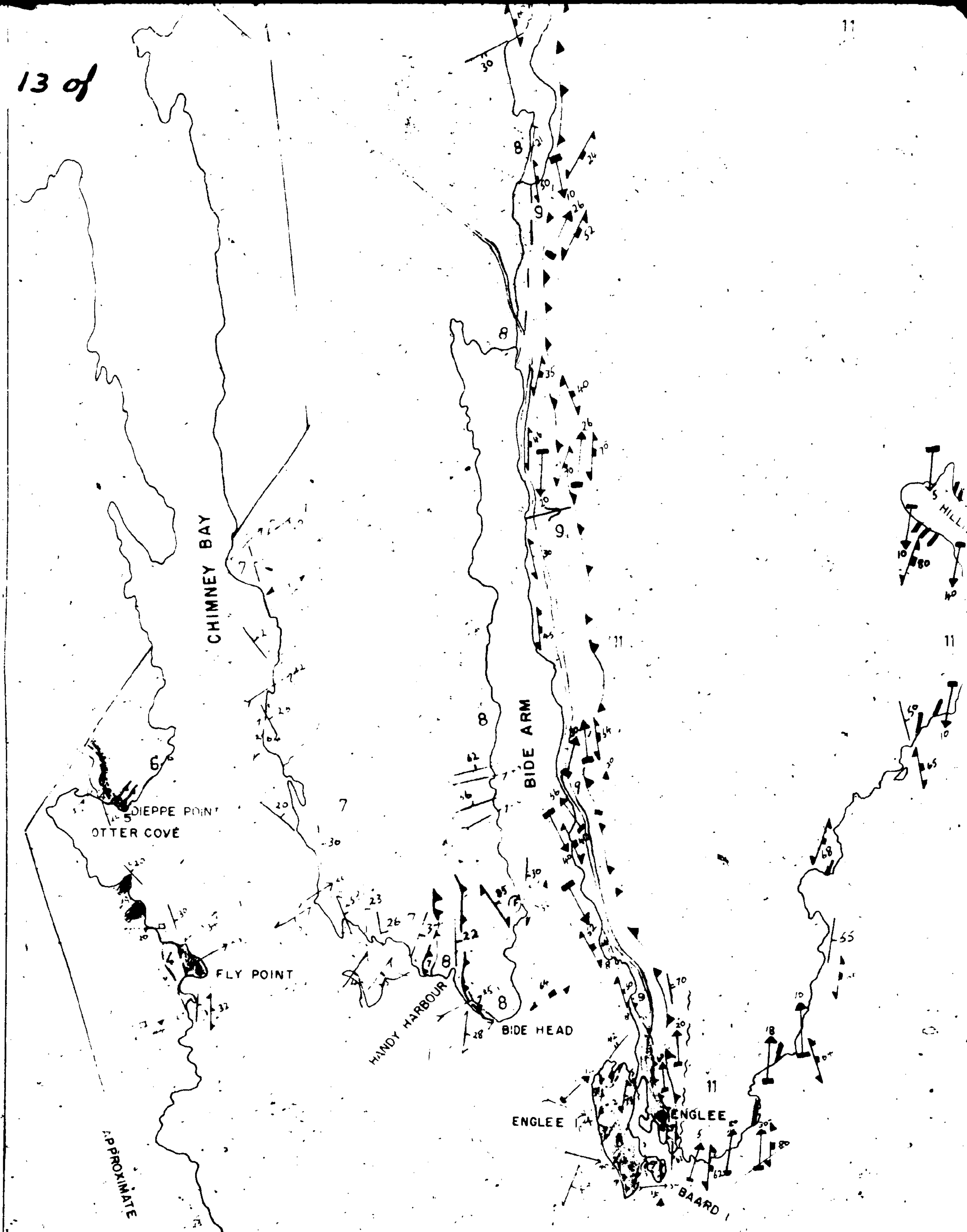
n'stone, and shale

...n; minor marble, psammitic, and meta-gabbro

nphibolite member

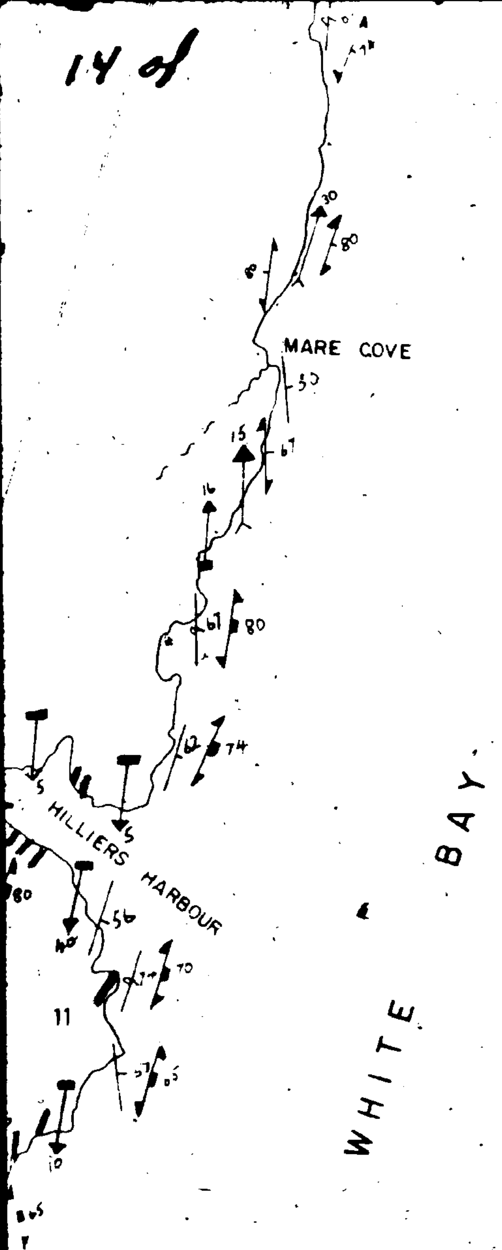
TION

ywacke; local pretectonic gabbro



CHIMNEY BAY
BIDE ARM
DIEPPE POINT
OTTER COVE
FLY POINT
HANDY HARBOUR
BIDE HEAD
ENGLEE
ENGLEE I.
BAARD I.
HILL
APPROXIMATE

14 of



... greywacke, local pretecto...

IRISH LIMESTONE
Sandy limestone; local quartzite and greenschist

CROQUE HEAD SLICE

LOWER CAMBRIAN OR OLDER

12 — **MAIDEN POINT FORMATION**
Greywacke and minor slate

MAIDEN POINT SLICE

LOWER CAMBRIAN OR OLDER

11 — **MAIDEN POINT FORMATION**
Coarse greywacke, green grey and black slate, a local pebb...
11a, massive basalt; **11b**, grey and white marble; **11c**, conglom...

AUTOCHTHONOUS ROCKS

MIDDLE ORDOVICIAN

9 — **GOOSE TICKLE FORMATION**
Grey to dark grey greywacke and shale; local limestone

8 — **TABLE HEAD FORMATION**
Grey to black hackly limestone and minor shale

UPPER CAMBRIAN AND LOWER ORDOVICIAN

7 — **St GEORGE FORMATION**
Grey, buff and pink dolomite

LOWER CAMBRIAN

6 — **FORTEAU FORMATION**
Grey shale and sandy limestone

5 — **DEVILS COVE FORMATION**
Purple and white limestone

icke; local pretectonic gabbro

nd greenschist 15 of

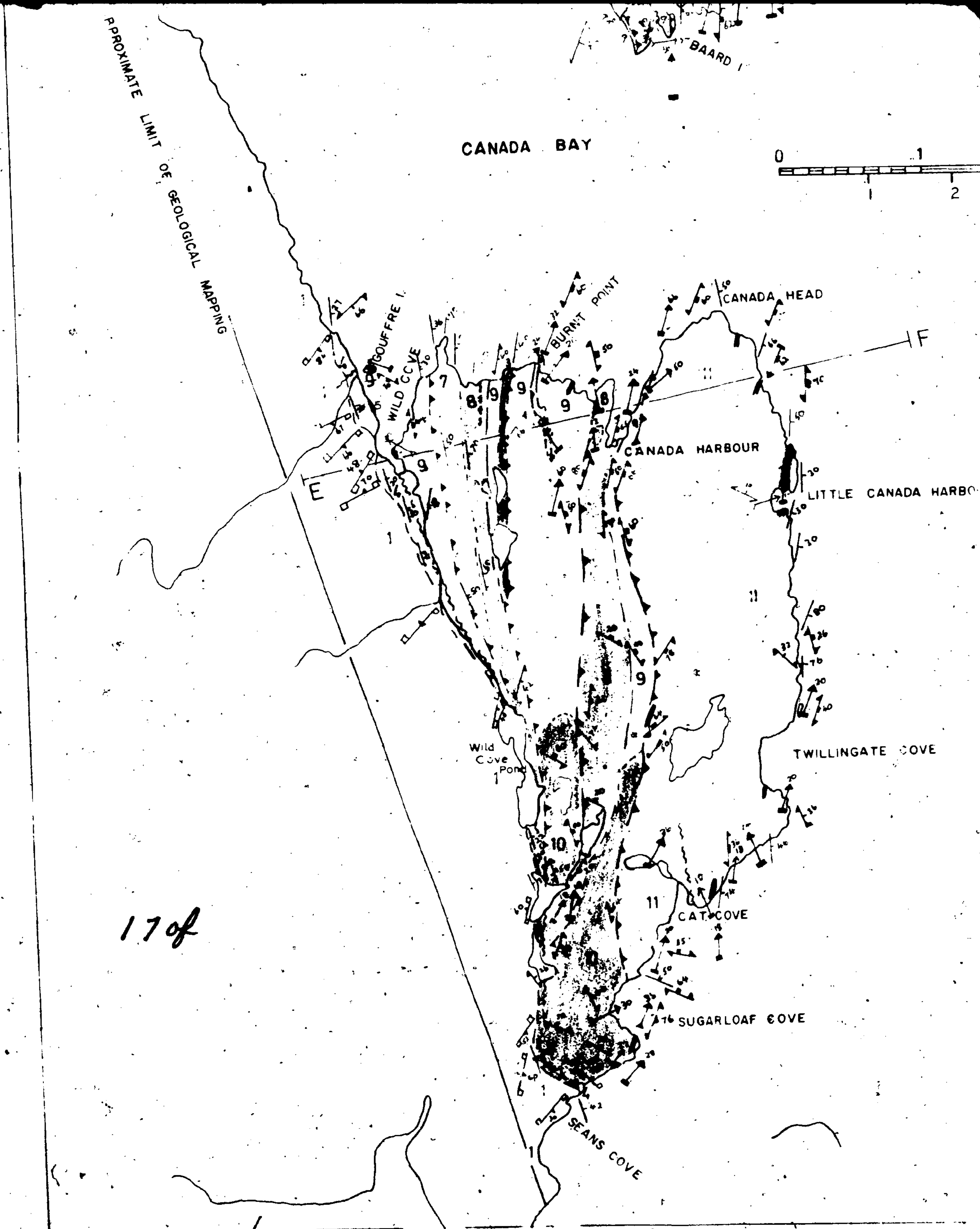
lack slate, c local pebble conglomerate;
hite marble; IIc, conglomerate; IId, serpentinized ultramafic rock

N
hale;

SUGARLOAF SCHISTS
Semi-pelitic, pelitic, and calc-silicate schist;
local marble; probably equivalent to IO'

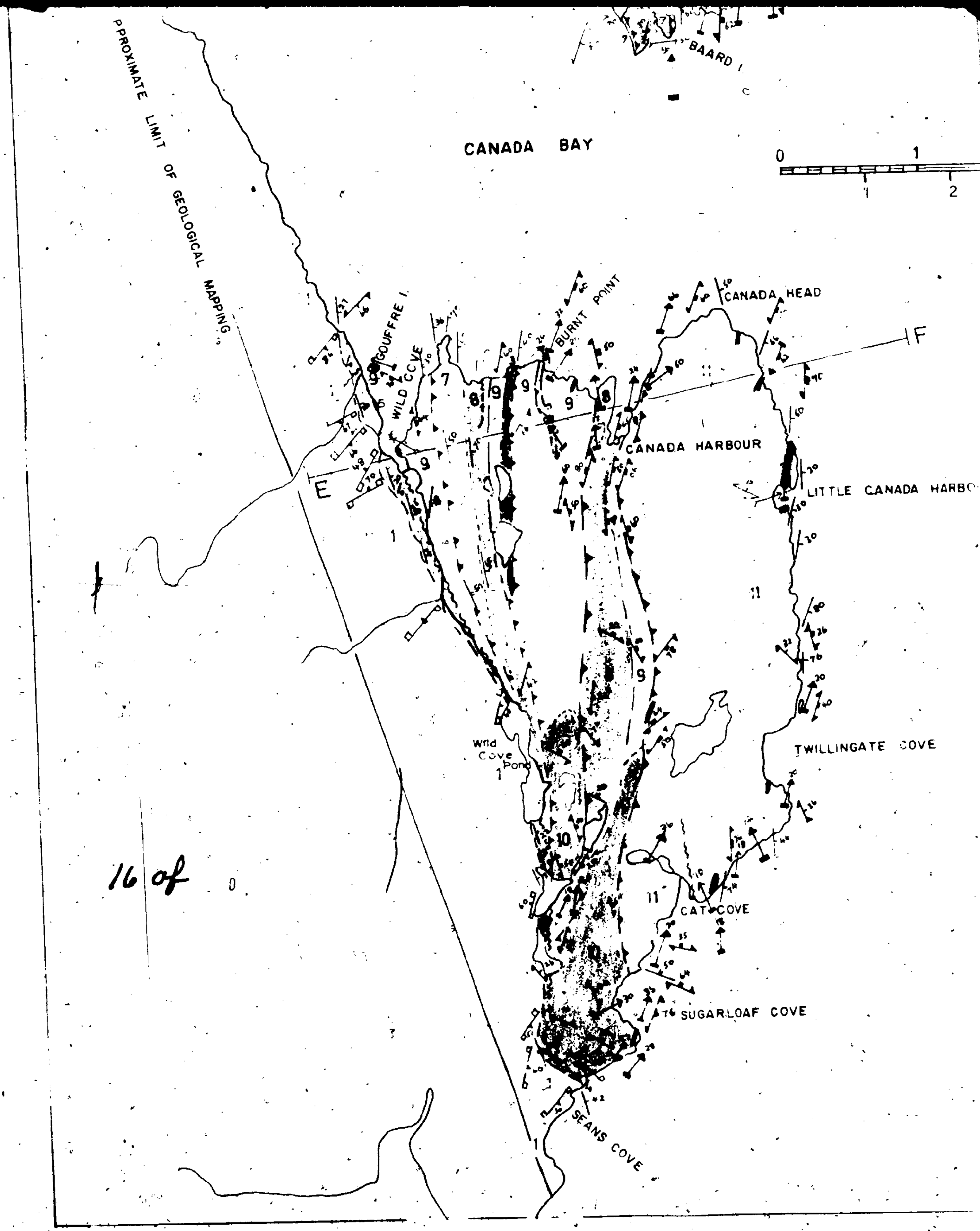
d minor shale

DOVICIAN



APPROXIMATE LIMIT OF GEOLOGICAL MAPPING
CANADA BAY
BAARD I.
GOUFFRE I.
WILD COVE
BURNT POINT
CANADA HEAD
CANADA HARBOUR
LITTLE CANADA HARBO
TWILLINGATE COVE
Wild Cove Pond
CAT COVE
SUGARLOAF COVE
SEANS COVE
E
F
0
1
2

17of



BAARD I.
CANADA BAY
0
1
1
2
APPROXIMATE LIMIT OF GEOLOGICAL MAPPING
GOUFFRE I.
WILD COVE
BURNT POINT
CANADA HEAD
F
E
CANADA HARBOUR
LITTLE CANADA HARBO
TWILLINGATE COVE
Wild Cove Pond
CAT COVE
SUGARLOAF COVE
SEANS COVE

16 of

1 2 3 Miles

2 3 4 Kilometers

4F

ADA HARBOUR

17 of

OVE

4	**BRADORE FORMATION** Arkose, hematitic sandstone, and minor shale

LOWER CAMBRIAN OR OLDER

	LIGHTHOUSE COVE FORMATION Plutonic boulder conglomerate; basalt

PRECAMBRIAN

1	**LONG RANGE COMPLEX** Foliated granite gneiss; foliated meta-gabbro; minor post-tectonic microgranite

SYMBOLS

Melange with black and green shale matrix and mainly sandstone(11) blocks

Dyke or sill

Geological boundary (defined, approximate, assumed)

Fault (defined, approximate, assumed)

Thrust (defined, approximate, assumed)

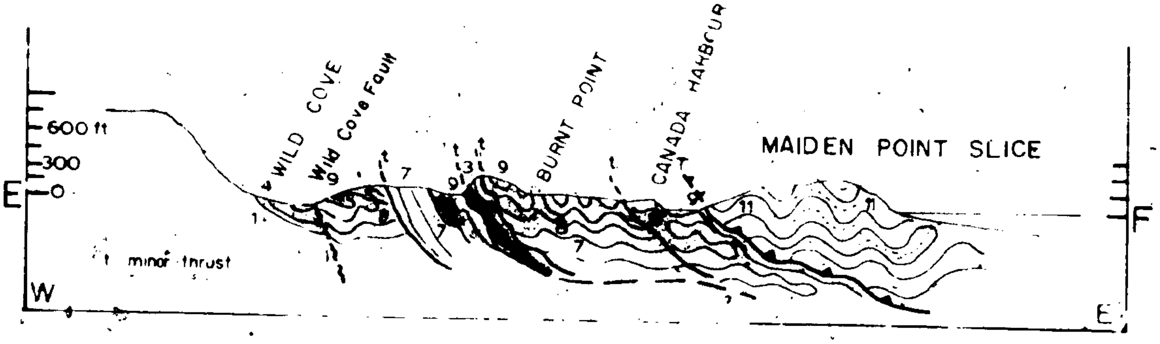
Bedding (tops known, unknown)

Gneissosity (units 1 and 2 only)

Fossil locality

Glacial striae

Road; cart track, trail



shale

ON

abbro; 2. ACID AND BASIC GNEISSES
Foliated acid gneiss; amphibolite; post-tectonic granite dykes; probably equivalent to I

ctrix and mainly

mate, assumed)

PRE-EMPLACEMENT STRUCTURES (White Hills Slice only)

Second schistosity, S_2

Plunge of minor F_2 folds

SYN-EMPLACEMENT STRUCTURES (Lower slices and autochthonous rocks)

Cleavage

Plunge of minor F_1 folds

POST-EMPLACEMENT STRUCTURES (All rocks)

Cleavage

Plunge of minor folds

First or second phase structures in lower slices and autochthonous rocks, third phase structures in White Hills Slice

Plunge of major anticline

Plunge of major syncline

LATE MINOR STRUCTURES

Cleavage

Fold or kink band

18 of

GEOLOGY

SOUTHERN PART OF THE HARE BAY ALLOCHTHON

N.W. NEWFOUNDLAND

SCALE 1:50,000

GEOLOGY BY W.R. SMYTH

DRAWN BY W.R.S

FIG 3

·· shale

ON

·abbro; } 2

ACID AND BASIC GNEISSES

Foliated acid gneiss; amphibolite; post-tectonic granite dykes; probably equivalent to I

·atrix and mainly

·mate, assumed)

PRE-EMPLACEMENT STRUCTURES (White Hills Slice only)

Second schistosity, S_2

Plunge of minor F_2 folds

SYN-EMPLACEMENT STRUCTURES (Lower slices and autochthonous rocks)

Cleavage

Plunge of minor F_1 folds

POST-EMPLACEMENT STRUCTURES (All rocks)

Cleavage } First or second phase structures in lower slices and autochthonous rocks, third phase structures in White Hills Slice

Plunge of minor folds

Plunge of major anticline

Plunge of major syncline

LATE MINOR STRUCTURES

Cleavage

Fold or kink band

GEOLOGY

SOUTHERN PART OF THE HARE BAY ALLOCHTHON

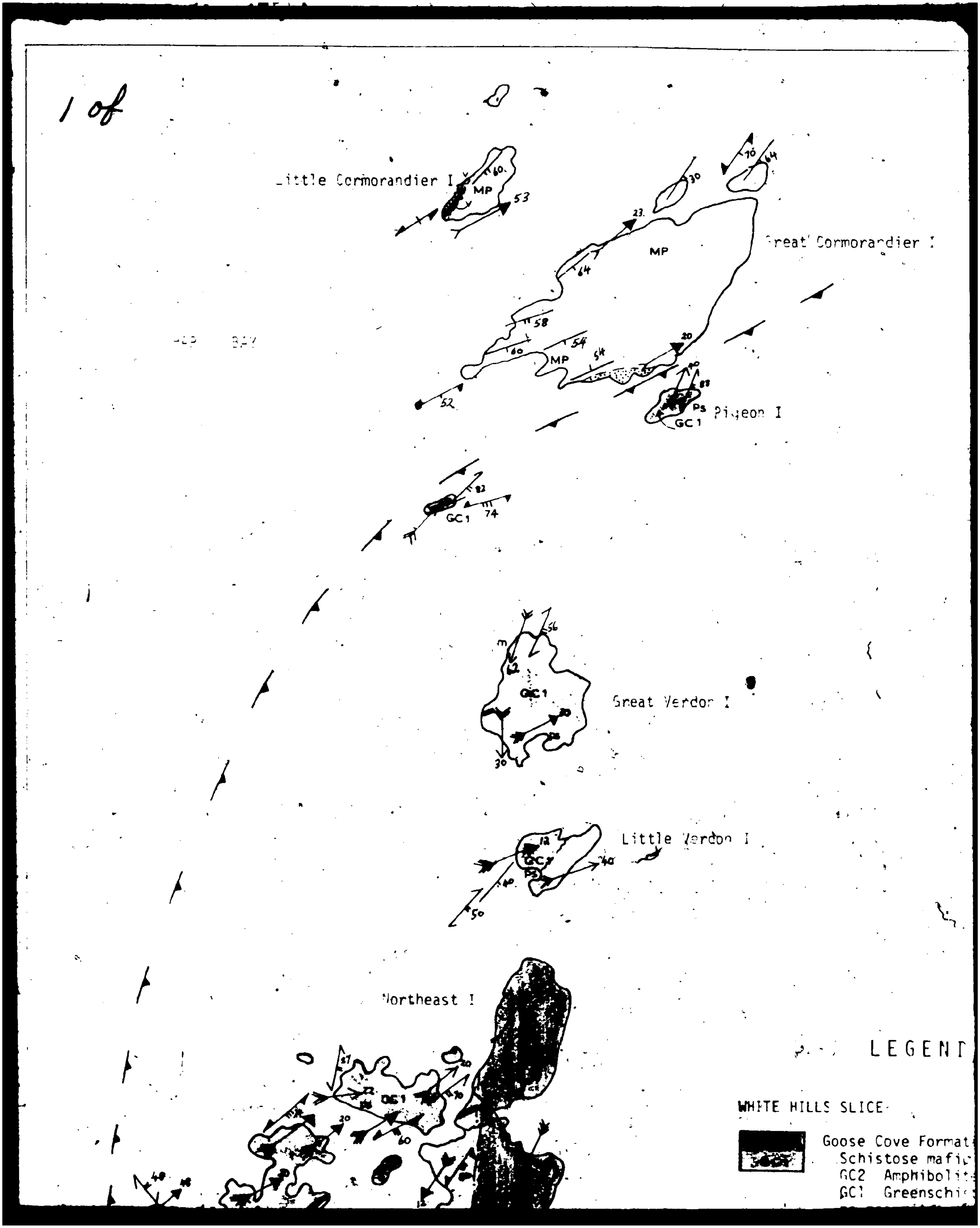
N.W. NEWFOUNDLAND

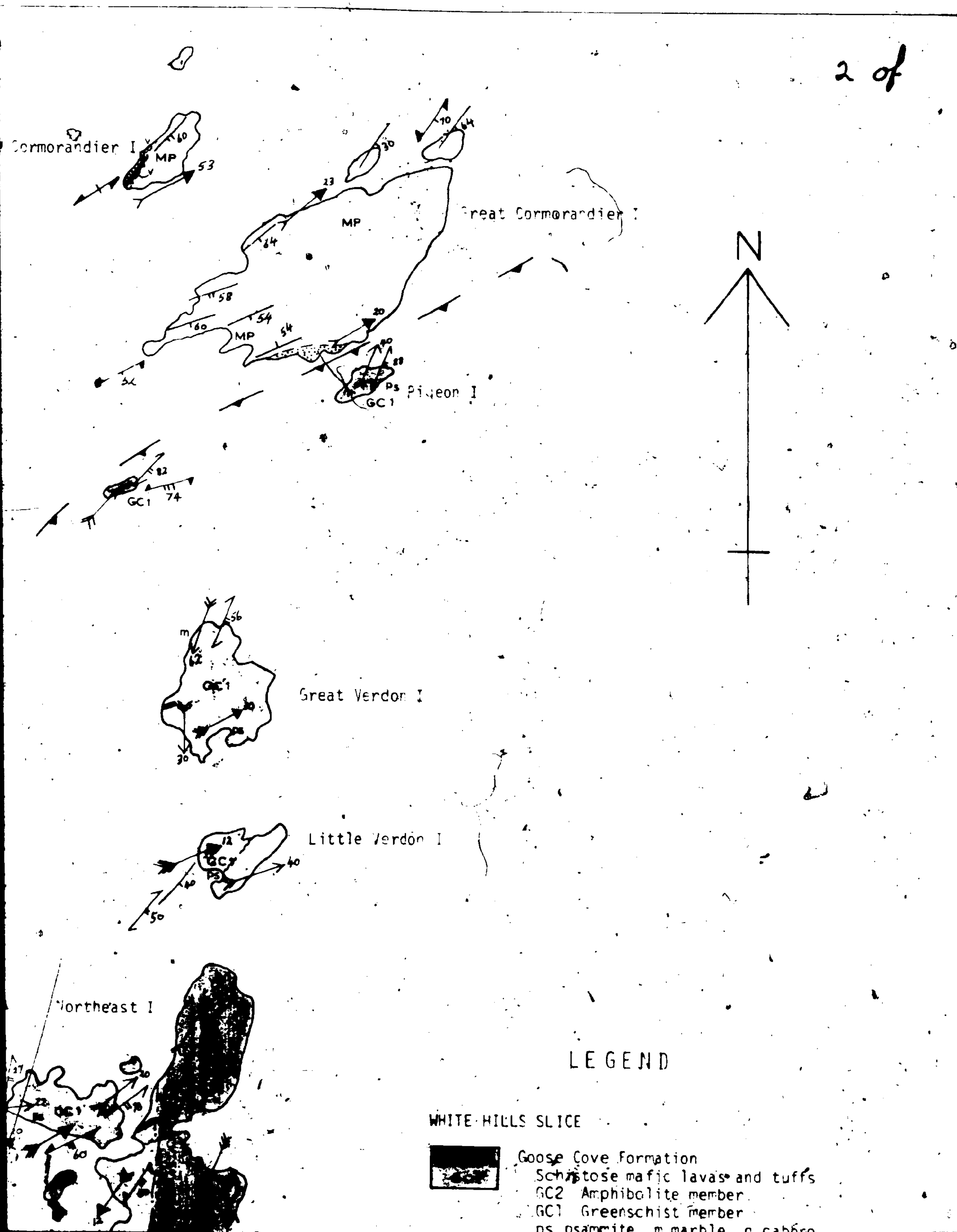
SCALE 1:50,000

GEOLOGY BY W.R. SMYTH

DRAWN BY WRS

FIG 3


1 of
Little Cormorandier I
MP
Great Cormorandier I
MP
MP
Pigeon I
Ps
GC1
GC1
Great Verdon I
GC1
Little Verdon I
GC2
Ps
Northeast I
GC1
LEGEND
WHITE HILLS SLICE
Goose Cove Format
Schistose mafi
GC2 Amphiboli
GC1 Greenschi


Cormorandier I
MP
Great Cormorandier I
MP
MP
N
Pigeon I
Ps
GC 1
GC 1
Great Verdon I
GC 1
Little Verdon I
GC 2
Ps
Northeast I
GC 1
LEGEND
WHITE HILLS SLICE
Goose Cove Formation
Schistose mafic lavas and tuffs
GC2 Amphibolite member
GC1 Greenschist member
ps psammite, m marble, g gabbro

3 of

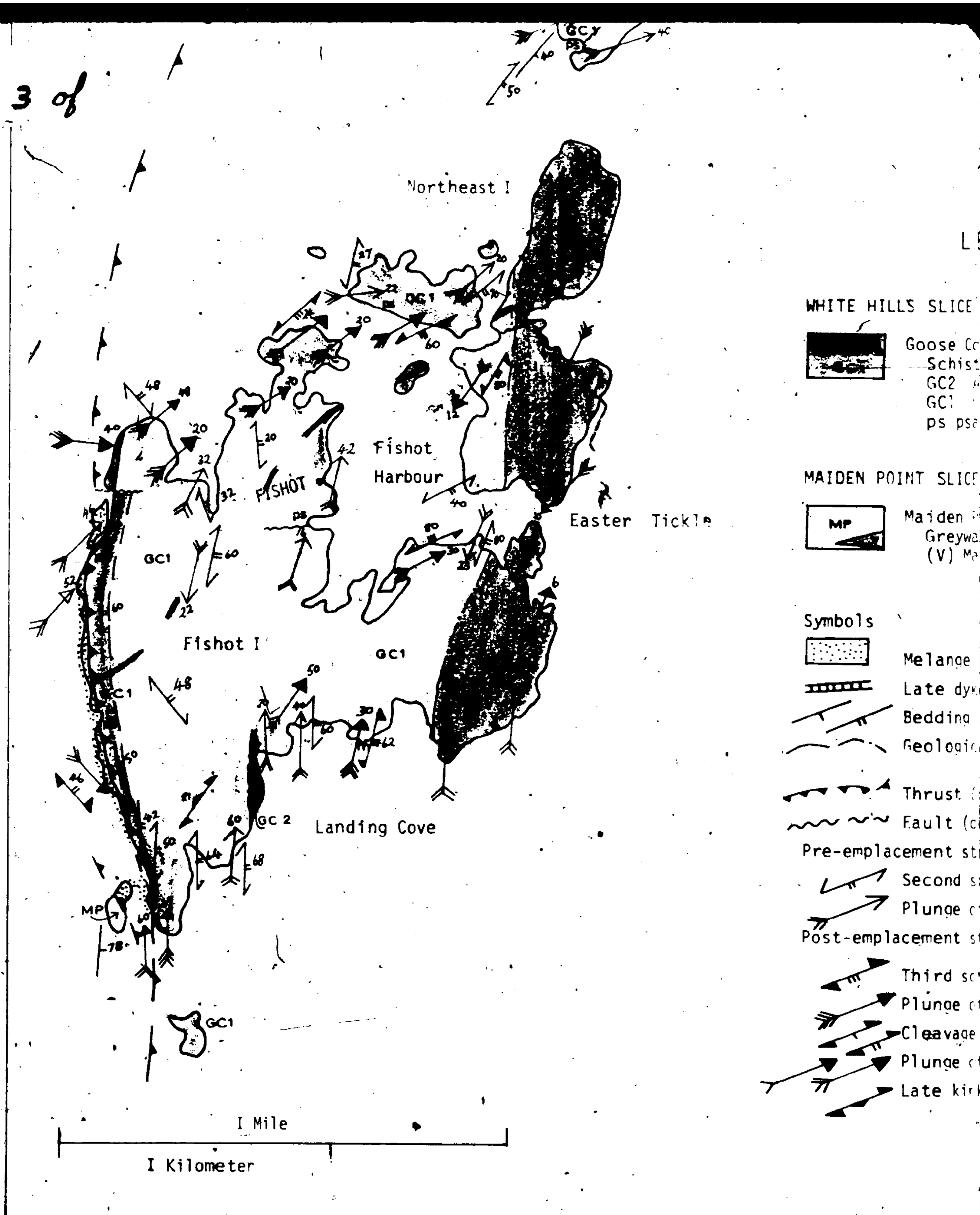


Figure 4: Geologic Map of the Fishot Islands

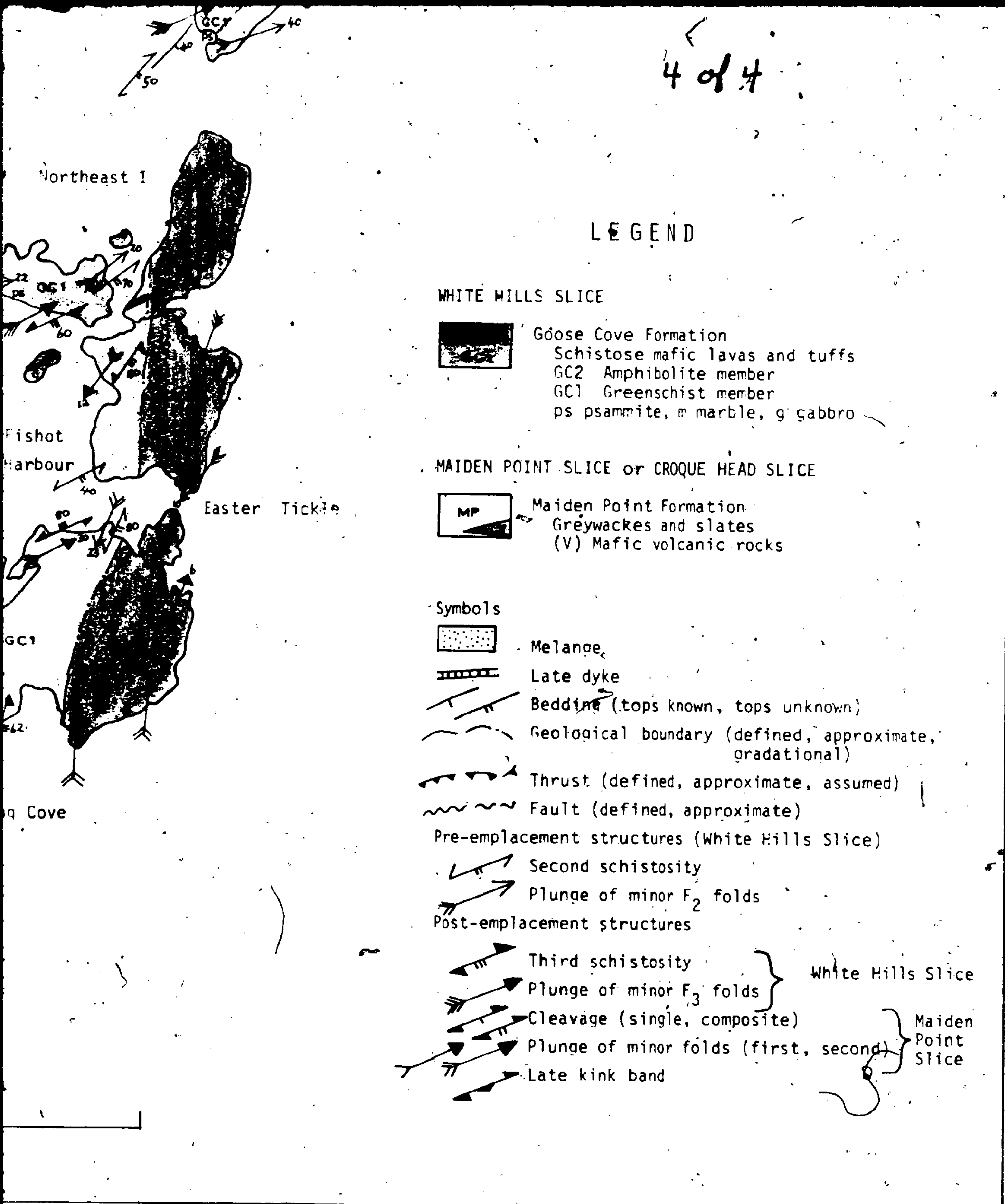


Figure 4: Geologic Map of the Fishot Islands

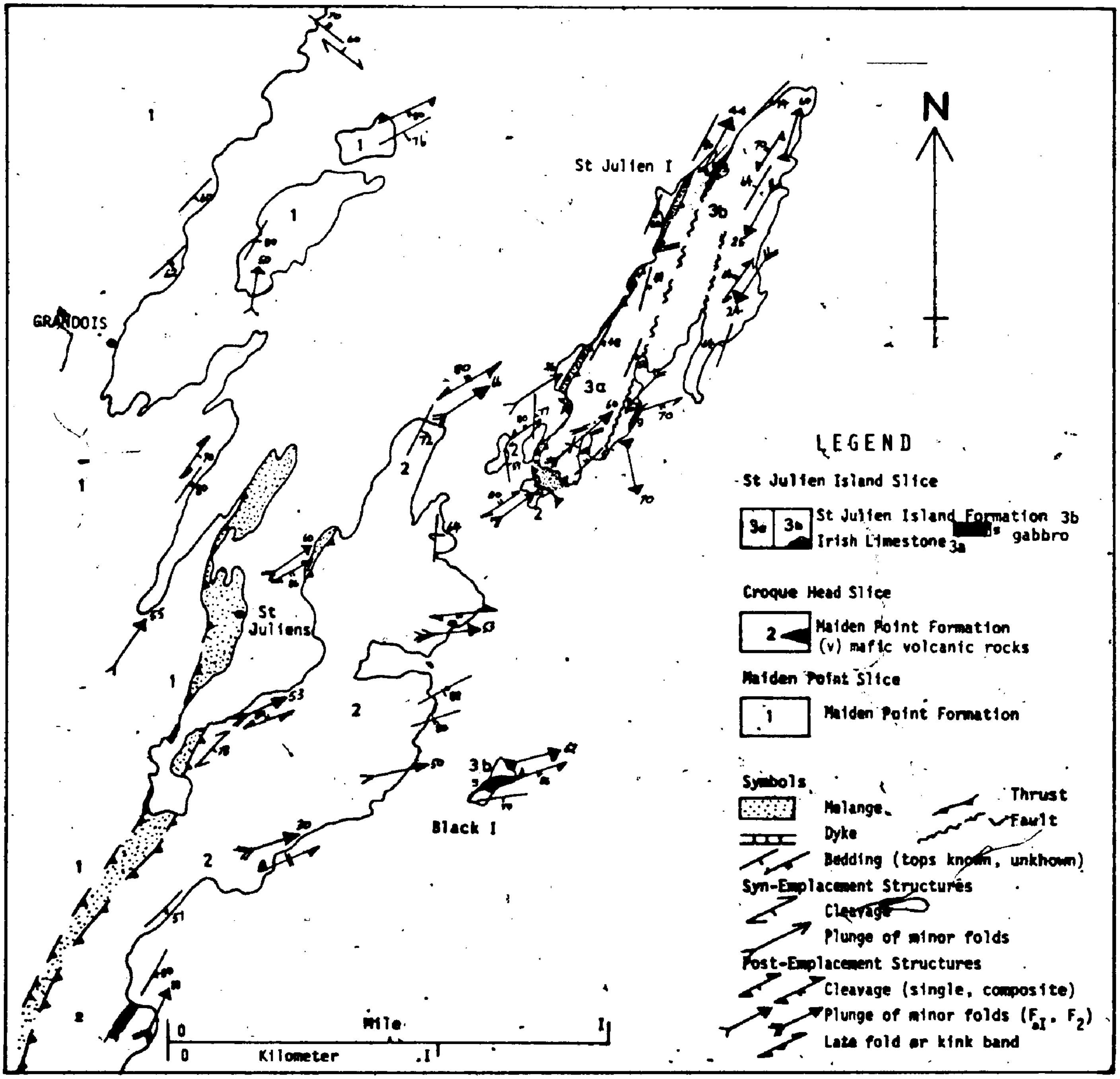
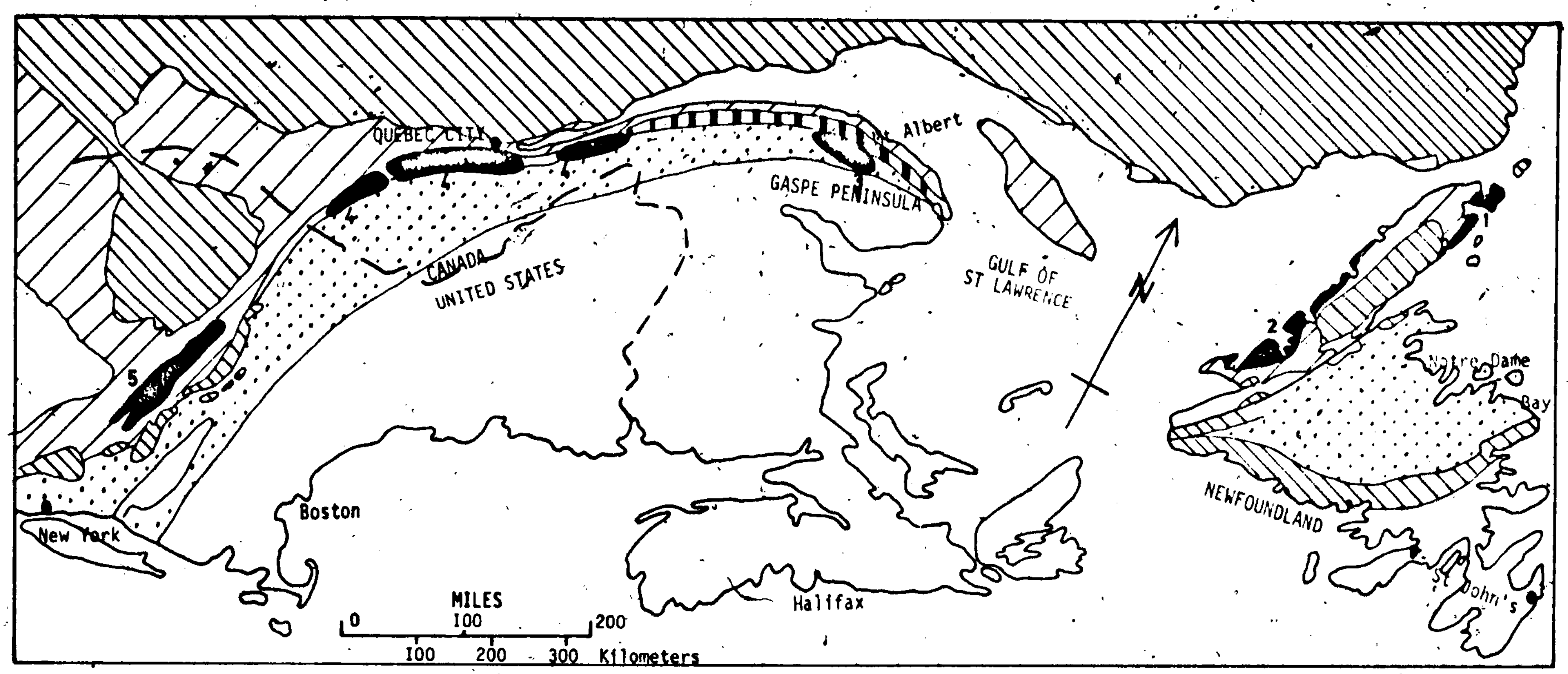


FIG. 5: GEOLOGIC MAP OF THE ST. JULIEN ISLAND SLICE.

Fig. 6 Aeromagnetic map of the Hare Bay Area. (Geological Survey of Canada Map 7366G,1970)



LOWER and MIDDLE ORDOVICIAN

Miogeosynclinal carbonate rocks

Allochthonous rocks
1. Hare Bay Allochthon
2. Humber Arm Allochthon
3. Gaspe Allochthon
4. Quebec City Allochthon
5. Taconic Allochthon

Mainly allochthonous rocks

Eugeosynclinal rocks, including ophiolites in Newfoundland.

PRECAMBRIAN

Crystalline Basement rocks

PRECAMBRIAN and PALEOZOIC

Undifferentiated

Fig 7: Generalized geologic map showing the distribution of the 'Taconic Allochthons' in the northern Appalachians
Map modified after Zen (1972) and Williams et al. (1972).