OPHILITE EMPLACEMENT ALONG THE BAIE VERTE-BROMPTON LINE AT GLOVER ISLAND, WESTERN NEWFOUNDLAND

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OPHIOLITE EMPLACEMENT ALONG THE BAIE VERTE - BROMPTON LINE AT GLOVER ISLAND, WESTERN NEWFOUNDLAND

by

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Cobble Cove

ABSTRACT

The Glover Island Area of western Newfoundland spans the Humber-Dunnage Zone Boundary at the northern end of the Appalachian Orogen. Here a deformed ophiolitic pluton separates schists and gneisses (west) from mafic volcanic rocks (east). To the west, quartzo-feldspathic gneisses of the Cobble Cove Gneiss are structurally overlain by the Keystone Schist. To the east, the ophiolitic Grand Lake Complex is structurally juxtaposed against the Keystone Schist and unconformably overlain by volcanic rocks of the Glover Group. Farther east the Glover Group is unconformably overlain by the Corner Pond Formation and to the south, volcanic rocks and dykes of the Otter Neck Group are unconformably overlain by the Red Point Formation.

The Cobble Cove Gneiss consists of granodioritic gneiss and metasomatized mafic dykes. It is followed eastward by metamorphosed greywackes, quartzites, conglomerates and mafic volcanics of the Keystone Schist. Intruded by mafic dykes and trondhjemite, cumulate ultramafic and mafic rocks of the Grand Lake Complex are structurally juxtaposed against the Keystone Schist. The Glover Group unconformably overlies the Grand Lake Complex and is divided into a lower metasedimentary unit termed the Kettle Pond Formation which contains abundant clasts derived from the underlying ophiolite, and an upper metavolcanic unit termed the Tuckamore Formation. Pillow basalts and mafic tuffs of the Tuckamore Formation are similar to low-K tholeiites. The Otter Neck Group consists of tholeiitic sheeted dykes and aquagene tuffs intruded by porphyries geochemically comparable to the Tuckamore Formation. It is unconformably overlain by metasediments of the Red Point Formation which are correlated with the Corner Pond Formation. Arenig conodants have been

recovered from a newly discovered fossil locality within the Corner Pond Formation.

The Cobble Cove Gneiss is interpreted as Grenvillian basement overlain by a Paleozoic metaclastic cover represented by the Keystone Schist. The Grand Lake Complex and Otter Neck Group are interpreted to represent Iapetus Ocean crust obducted during Taconic Orogenesis. Exhumation during ophiolite obduction led to the formation of unconformities upon which the Glover Group, Red Point Formation and Corner Pond Formation were deposited. Similarities in lithology and structural sequence between Glover Island and Baie Verte indicate that the Baie Verte-Brompton Line is continuous between these two localities.

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Chapter 1

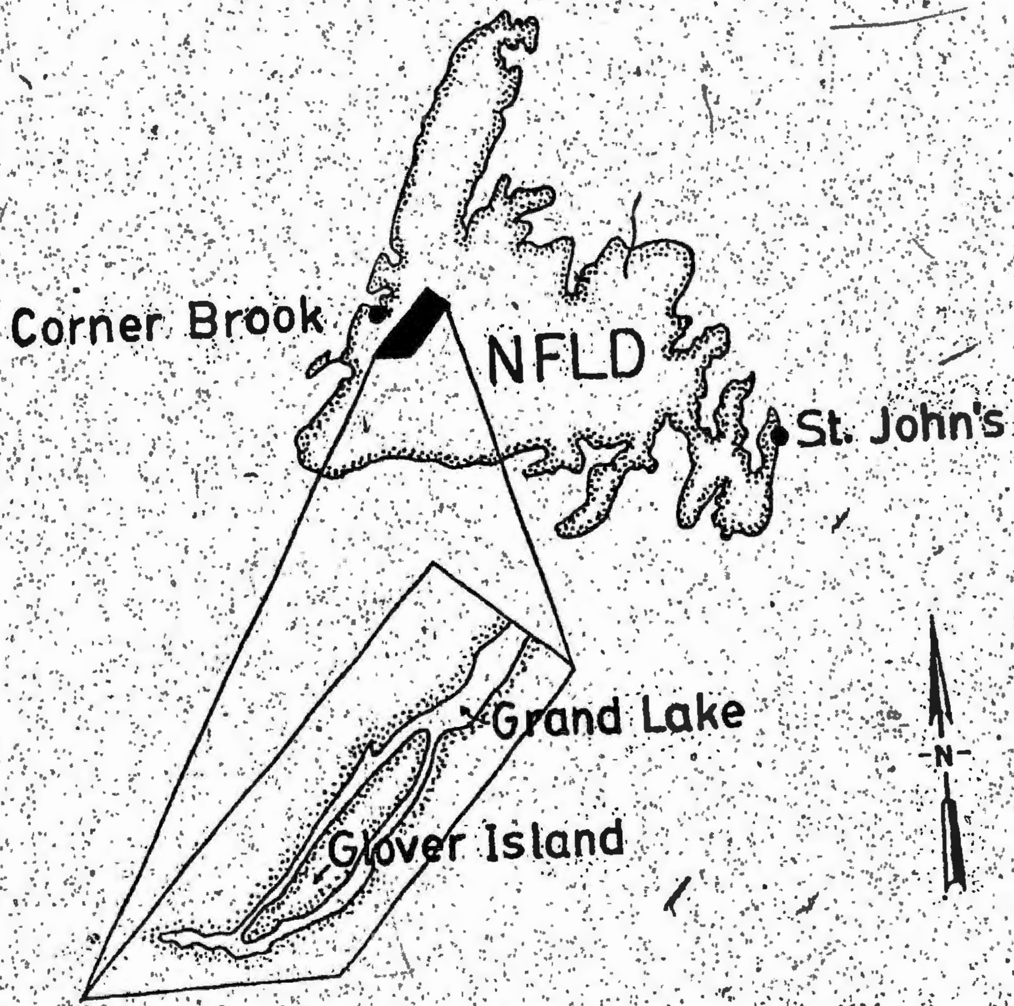
INTRODUCTION

The Glover Island Area is uniquely located along the west flank of the northern Appalachians as it spans an important structural junction between Grenvillian basement gneisses and metaclastic rocks to the west and an ophiolite complex overlain by a thick volcanic sequence to the east. Grenvillian basement gneisses and metaclastic rocks form the eastern part of the Humber Zone (Williams, 1979), the westernmost zonal subdivision of the Appalachian Orogen, while ophiolitic rocks and volcanics form part of the more easterly Dunnage Zone. These two zones evolved separately as continental margin and ocean, respectively, until they became structurally juxtaposed along the Baie Verte-Brompton Line (Williams and St. Julien, 1982). The Glover Island Area represents one of the few areas in Newfoundland (and indeed in the Appalachian Orogen) where this structural junction can be observed.

LOCATION AND ACCESS

Located in western Newfoundland, the Glover Island Area consists of roughly 200 km^2 in the southern half of Grand Lake, Newfoundland's largest inland body of water (Fig. 1.1). Glover Island is a large, 40 km long island located in the southern half of Grand Lake after which the

Figure 1.1 Location of study area.


Corner Brook
NFLD
St. John's
Grand Lake
Glover Island
N
KM
0
100
200
300

study area is named.

The Southern half of Grand Lake is accessible at the extreme west end of Grand Lake at "Camp 33" (NTS map sheet 12 B/9E) and at Northern Harbour via a 15 km woods road from the town of South Brook (NTS map sheet 12 A/13). This road is owned by Bowaters of Newfoundland, Incorporated, and company permits are required.

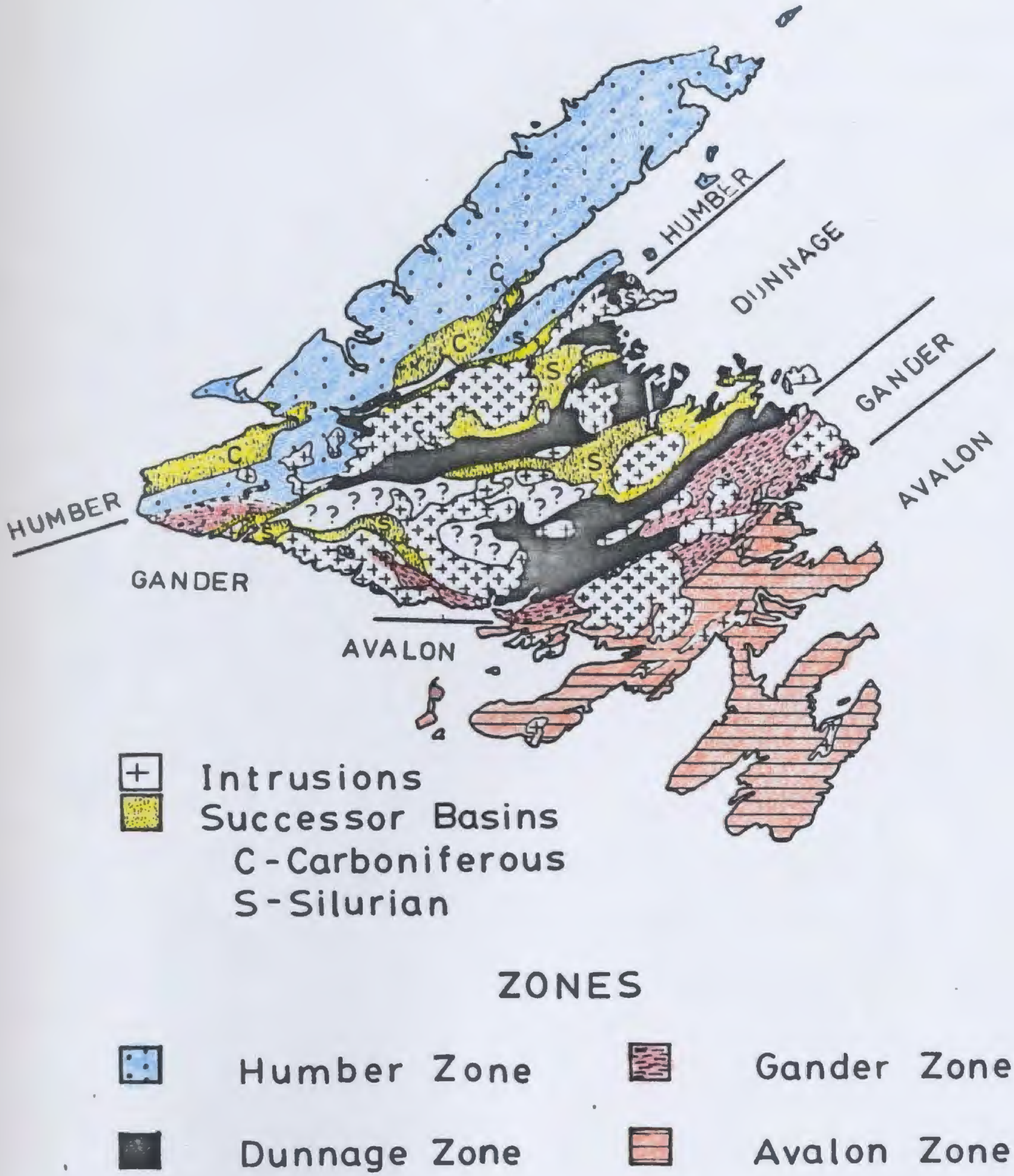
Temporary camps were set up by boat on Glover Island and on the mainland adjacent to the island to gain easy access to the island and the surrounding area. Interior portions of Glover Island and Corner Pond were mapped from fly camps established by float plane and helicopter.

GEOLOGIC SETTING

The Glover Island Area represents part of the Humber and Dunnage Zones of the five-fold zonal subdivision of the Appalachian Orogen proposed by Williams (1978, 1979). In this zonal subdivision, the Appalachian Orogen is divided from west to east into the Humber, Dunnage, Gander, Avalon and Meguma Zones (Fig. 1.2).

The Humber Zone is the most external zone, bounded to the west by undeformed platformal rocks of the North American craton and to the east by more outboard, accreted elements of the Orogen. It consists of Grenvillian basement overlain by a thin cover of clastic rocks and a Cambrian to Middle Ordovician carbonate sequence. Farther east

Figure 1.2 Tectonic Lithofacies map of Newfoundland after Williams (1978).


HUMBER
DUNNAGE
GANDER
AVALON
HUMBER
GANDER
AVALON
Intrusions
Successor Basins
C-Carboniferous
S-Silurian
ZONES
Humber Zone
Dunnage Zone
Gander Zone
Avalon Zone

retrograded Grenvillian Gneisses occur together with metaclastic rocks. A variety of westerly transported allochthonous rocks including ophiolite also occur within the Humber Zone.

Based upon facies reconstructions and comparisons with modern day examples, the Humber Zone is interpreted as the ancient continental margin of eastern North America formed during the opening of the Iapetus Ocean (Stevens, 1970; Williams, 1975).

The Dunnage Zone consists of a complex Ordovician and Silurian volcanic terrane that separates the Humber and Gander Zones. In Newfoundland the Dunnage Zone is underlain by an ophiolitic basement strongly influencing its interpretation as a volcanic arc built upon Iapetus, oceanic crust.

The Gander Zone consists of a thick sequence of metaclastic rocks intruded by abundant megacrystic granite bounded to the west by the Gander River ultramafic belt and to the east by the Dover Fault. It is interpreted to represent the eastern margin of the Iapetus Ocean, or an accreted terrane of unknown significance.

The Avalon and Meguma Zones occur east of the Gander Zone and represent the most outboard elements of the Appalachian Orogen. The Avalon Zone consists of a volcanic terrane of Late Precambrian age overlain by a Cambrian platformal sequence. The Meguma Zone consists of a thick sequence of undeformed, early Paleozoic clastic rocks recognized only in Nova Scotia.

The Avalon and Meguma Zones both escaped the widespread Ordovician deformation and metamorphism that characterizes more westerly zones. This suggests these zones originated in an environment unrelated to the Iapetus cycle and were accreted to the orogen at a later stage.

Williams's zonal subdivisions highlight one of the key structural junctions in the Northern Appalachians: the Humber-Dunnage Zone boundary. For over 1000 kilometres this boundary is marked by discontinuous fragments of ophiolite that separate strikingly contrasting rock groups. Arenaceous metaclastic rocks of the Humber Zone occur west of the Humber-Dunnage Zone boundary. East of this boundary are dominantly mafic volcanic rocks of the Dunnage Zone. Along the boundary, fragments of ophiolite occur locally. Although generally poorly preserved, some of these ophiolite fragments contain a recognizable ophiolite stratigraphy. A unique feature of these ophiolite occurrences is that they are locally overlain by coarse conglomerates containing abundant ophiolitic detritus.

Williams and St. Julien (1978, 1982) recognized that the Humber-Dunnage Zone boundary was a fundamental, Taconic structure related to the destruction of the continental margin of eastern North America and traced it through Newfoundland and Quebec. They named it the Baie Verte-Brompton

Line and postulated that it extended from Baie Verte, where its position is well known, through Glover Island, based upon a single occurrence of ultramafic rock, to the Cape Ray Fault in southeastern Newfoundland and from there across the Gaspe Peninsula to Brompton Lake, Quebec.

PURPOSE AND SCOPE

The recognition of ophiolitic rocks at Glover Island and the probable continuation of the Baie Verte-Brompton Line presented an ideal opportunity to study the relationships between ophiolitic rocks and rocks of the Humber and Dunnage Zones across the Baie Verte-Brompton Line. This structural junction represents the interface between the ancient continental margin of eastern North America and the Iapetus Ocean and by studying it much can be learned about the destructional history of this part of the orogen. Considerable work has been done on a similar part of the Baie Verte-Brompton Line along strike to the northeast at Baie Verte. Here conflicting interpretations of local relationships have precipitated a storm of controversy. However by a comparison between these two areas some insight into the regional continuity of structures and major rock groups can be gained.

The data base for this study was collected over a period of 4 years while a full time graduate student at Memorial University. Three seasons of 12, 10 and 6 week intervals were spent in the field mapping the central part of Glover Island in detail and reconnaissance mapping of the rest of the Glover Island Area. This work is reported in a 1:50,000 scale geologic map of the Glover Island Area included in the jacket of this thesis (Plate 1) and in the text.

Data collected in the field were supplemented by petrographic analyses and geochemistry. 400 thin sections were examined in detail, and as much of the inferred structural and metamorphic history of rocks in the Glover Island Area is based upon microstructures, this petrographic data base is of considerable importance and is emphasized in chapters on stratigraphy, structure and metamorphism. 37 samples were analyzed for major and trace elements by G. Andrews (M.U.N.) and D. Press (M.U.N.). Major elements were determined by standard wet chemical methods and trace elements by X-ray fluorescence. These data were used to elucidate the affinities and probable origins of volcanic rocks on Glover Island and to investigate the metasomatism of mafic rocks.

Carbonate samples were also tested for microfossils by the Geological Survey of Canada. However extensive deformation and metamorphism restricted the number of samples that could be reasonably analyzed.

Minerals were identified by several different techniques in the course of this study including optical examination, universal stage measurements, staining techniques, microprobe analysis and X-ray diffraction. Details of sample preparation and methods for the major and trace element analyses and microprobe analyses are found in Appendices A and B.

PHYSIOGRAPHY

The Grand Lake Area of western Newfoundland represents a classic example of a dissected peneplain. Gently rolling terrain characterizes the regions west and east of Grand Lake. Elevations average around 1500 feet (450 m) above sea level with isolated monadnocks locally reaching 2000 feet (600 m). Twenhofel and MacClintock (1940) have described this terrain as a partial peneplain. Grand Lake is situated in a deeply incised valley superimposed upon this moderately level topography.

The transition between peneplain and valley is abrupt and locally spectacular. Along the west side of Glover Island, level, mature woodlands end at a sheer cliff that drops precipitously into Grand Lake 1000 feet (300 m) below.

Twenhofel and MacClintock hypothesize this partial peneplain in the Grand Lake Area developed as a preglaciation, fluvial peneplain which they term the High Valley Peneplain. The High Valley Peneplain postdates an older, more completely developed peneplain preserved best along the west coast of Newfoundland termed the Long Range Peneplain.

The Grand Lake Valley is glacial in origin as indicated by its oversteepened sides and numerous waterfalls and constitutes a series of lowlands extending from Georges Lake to White Bay. A number of high-angle, possibly strike-slip faults occur in this lowland area and may have influenced the pattern of subsequent glaciation.

PREVIOUS WORK

Early workers in the Grand Lake Area were largely concerned with the coal-bearing Carboniferous rocks of northern Grand Lake (e.g., Murray and Howley, 1881). Systematic mapping of the crystalline rocks which underlie the southern half of Grand Lake was initiated by the Geological Survey of Canada. Working for the G.S.C., Riley mapped the one degree sheet: Red Indian Lake - west half, which was later published by the G.S.C. at a scale of 1:253,440 (Riley, 1957).

Riley recognized schistose metasediments, volcanics and plutonic mafic rocks on Glover Island. He named the volcanics the Glover Formation and interpreted ophiolitic plutonic rocks as intrusions. He also mapped Carboniferous rocks and delineated a large granite pluton at the north end of Glover Island.

In 1970 P. Dimmel discovered a rich graptolite locality at Corner Pond while working for the Noranda Exploration Company, Limited (G.S.C. fossil locality 87420). The fossil locality consists of black shales containing abundant graptolite remains identified by R.B. Rickards

(Cambridge) and described in Dean (1976). This graptolite fauna is assigned to the Didymograptus nitidus zone.

No additional work in the area was undertaken until Williams became involved in tracing ophiolitic rocks across Newfoundland. During the 1977 field season, Williams concentrated on the Glover Island Area and demonstrated that the diorite plutons mapped as intrusions by Riley were structurally sited along the interface between volcanic rocks of the Glover Formation and schists and gneisses he correlated with the Fleur de Lys Supergroup (Church, 1969). The results of this work are summarized in Williams and St. Julian (1978).

The present study was begun in 1978 as part of a doctoral program at M.U.N. Initial results are summarized in various G.S.C. publications during 1978 and 1980 (Knapp et al., 1979; Knapp, 1980a) and in several abstracts (Knapp, 1979, 1980b, 1981). In these reports, the major rock groups and structures of the Glover Island Area have been described in a preliminary fashion. A more complete treatment has been reserved for this thesis.

ACKNOWLEDGEMENTS

I am grateful for the support and encouragement of a great number of people during my tenure at Memorial University. In particular I wish to thank my advisor, Dr. Williams, for his continued support and interest and for faithfully reading seemingly endless first drafts of various papers, reports and maps. Numerous discussions with T. Calon, J. Hibbard, E. Stander, D. Kennedy, H. Longerich, S. Godfrey, G. Dunning, R. Jameison and many others have all contributed to this thesis and I wish to thank them all. The capable field assistance of Martin Blanchard of Shallop Cove and Greg Tobin of Corner Brook is gratefully acknowledged as well.

Financial support was provided by a Memorial University of Newfoundland Fellowship to the author and field support was provided by the Geological Survey of Canada through Dr. Williams. I gratefully acknowledge this assistance.

Chapter 2

GENERAL GEOLOGY

The Glover Island Area encompasses portions of both the Dunnage and Humber Zones as it spans the Baie Verte-Brompton Line (Plate I). The Dunnage Zone is primarily represented by greenschist facies, mafic and silicic, extrusive rocks that unconformably overlie an ophiolitic basement of mafic and ultramafic rocks. In places younger coarse metaclastic rocks unconformably overlie this crystalline basement. Amphibolite grade schists and gneisses of the Humber Zone bound the western margin of these mafic rocks and record a complex structural history. These rocks are briefly described below and summarized in the accompanying Table of Formations (Table 2.1).

Strongly foliated gneisses containing metasomatized mafic dykes represent a fragment of Grenvillian basement within schists and gneisses of the Humber Zone. This gneissic unit is termed the Cobble Cove Gneiss.

The Cobble Cove Gneiss is overlain by a sequence of semipelitic, psammitic and calcareous schists and amphibolite termed the Keystone Schist. The protolith of these rocks is interpreted to have been quartzites, conglomerates, shales and limestone interlayered with mafic volcanic rocks. Metamorphic grade ranges from greenschist to amphibolite. The age of the Keystone Schist is inferred to be Cambrian

Table 2.1 Table of Formations for the Glover Island Area. (a) Glover Island and the area immediately east of Grand Lake except for the southern end of Glover Island; (b) Southern end of Glover Island.

Table 2.1a

<table>
<tr><th>AGE</th><th colspan="2">FORMATION</th><th>LITHOLOGY</th></tr>
<tr><td rowspan="6">ORDOVICIAN</td><td colspan="2">Corner Pond Formation 200m</td><td>Tan metasandstone and metaconglomerate; grey, black and green slate, red silicic tuff, green and purple pillow lava (fossiliferous)</td></tr>
<tr><td colspan="3">Unconformity</td></tr>
<tr><td rowspan="2">Glover Group</td><td>Tuckamore Formation 4000m</td><td>Pillow lava, mafic and silicic tuff, breccia, gabbro and argillite metamorphosed to the greenschist facies.</td></tr>
<tr><td>Kettle Pond Formation 600 m</td><td>Thinly layered quartz-sericite schists and polymictic metaconglomerate</td></tr>
<tr><td colspan="3">Unconformity</td></tr>
<tr><td colspan="2">Grand Lake Complex 1500 m</td><td>Ophiolite, serpentinized ultramafic rocks, talc-carbonate schists, metagabbro, metatrondhjemite, metabasalt dykes, and greenschists</td></tr>
<tr><td colspan="4">Thrust Fault</td></tr>
<tr><td>ORDOVICIAN and CAMBRIAN</td><td colspan="2">Keystone Schist 700m</td><td>Semipelitic and psammitic schist and gneiss, amphibolite, marble and metaconglomerate</td></tr>
<tr><td colspan="4">Thrust Fault</td></tr>
<tr><td>PRECAMBRIAN</td><td colspan="2">Cobble Cove Gneiss 250m</td><td>Quartzo-feldspathic microcline gneiss and biotite-microcline mafic schist layers</td></tr>
</table>

Table 2.1b

AGE	FORMATION	LITHOLOGY
ORDOVICIAN	Red Point Formation 200 m	Green and tan metasandstone and metaconglomerate, purple phyllite and metagreywacke
	Unconformity	
	Otter Neck Group Thickness unknown	Metagabbro, fine-grained mafic and silicic metaporphyry and medium-grained metatrondhjemite
		Intrusive Contact
		Ophiolite; pillow lava, pillow breccia, aquagene tuff and sheeted dykes metamorphosed to the greenschist facies

to Ordovician.

The Keystone Schist is overlain abruptly by a zone of intensely deformed serpentinite marking a major fault termed the Grand Lake Complex Thrust Fault. Structurally overlying the Keystone Schist along this major thrust fault is the Grand Lake Complex.

The Grand Lake Complex is a kilometre-thick, plutonic massif of ophiolitic affinity consisting of serpentinite and metaclinopyroxenite at the base and hornblende gabbro at the top. All units of the massif are intruded by mafic dykes while the upper part of the massif is also intruded by trondhjemite. Volcanic rocks are absent in the Grand Lake Complex due to erosion along an unconformity located along the upper contact of the massif.

The Grand Lake Complex is overlain by the Glover Group which is divided into two formations: The Kettle Pond Formation and the Tuckamore Formation. The Kettle Pond Formation occurs at the base of the Glover Group and unconformably overlies the Grand Lake Complex. It is primarily sedimentary in origin and contains numerous clasts which closely resemble rocks of the underlying ophiolite. Coarse conglomerate beds locally occur within the Kettle Pond Formation.

The Tuckamore Formation gradationally overlies the Kettle Pond Formation. Abundant silicic tuffs mark the contact and are quite abundant in the Tuckamore Formation. Mafic pillow lavas, mafic and silicic lapilli tuffs and

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agglomerates make up most of the Tuckamore Formation. Intrusive rocks are restricted to small dykes and sills of cogenetic gabbro and several later granitic intrusions.

At the south end of Glover Island a unit of pillow lava, sheeted dykes, pillow breccia and aquagene tuff known as the Otter Neck Group outcrops in an isolated, fault-bounded block. This unit contrasts with the Tuckamore Formation in its lack of pyroclastic rocks and abundance of intrusions. Based upon the occurrence of sheeted dykes the Otter Neck Group is interpreted to be ophiolitic.

Intrusions within the Otter Neck Group consist of trondhjemites, gabbros, and mafic and silicic porphyries. These intrusions are younger than the Otter Neck Group and based upon petrography and chemistry are interpreted to be related to the Tuckamore Formation.

The Red Point Formation unconformably overlies a trondhjemite that intrudes the Otter Neck Group on Glover Island. It consists of purple phyllites and metagreywacke. Its unconformable position indicates it is younger than the Otter Neck Group.

Rocks similar to the Red Point Formation at Corner Pond unconformably overlie the Tuckamore Formation. Here pillow lavas and silicic lapilli tuffs occur together with conglomerates and red and black slates to form the Corner Pond Formation. This unit is correlated with the Red Point Formation. Conodont fragments recovered from a limestone bed within the Corner Pond Formation are Early Ordovician in age. Graptolites from a nearby but isolated outcrop of black slate are also Early Ordovician in age and are interpreted to also be part of the Corner Pond Formation.

The Cabot Fault forms a major structural boundary along the west side of Grand Lake that truncates the Glover Group, Grand Lake Complex, Keystone Schist, Cobble Cove Gneiss, Otter Neck Group and Red Point Formation. Throughout most of its extent in the Glover Island Area the Cabot Fault is concealed by Grand Lake. Where it is exposed it consists of a 20 metre thick vertical zone of extensive mylonitization and brecciation. Numerous upright folds in Carboniferous rocks located along the shoreline of Grand Lake in the northern part of the study area (Plate I) may also be related to this fault.

Because this major, high-angle fault truncates all rock units and structures on Glover Island it is interpreted to be younger and therefore unrelated to the formation of the Baie Verte - Brompton Line. It is not discussed in greater detail in this thesis for this reason. However it is clearly of major significance to the subsequent history of this area and indeed western Newfoundland.

Thus the Glover Island Area is characterized by a large number of contrasting rock types disposed in a relatively small area. Structural boundaries and unconformities are common and far outnumber gradational, conformable contacts. Metamorphism ranges between greenschist and amphibolite facies and has obscured many primary features and structures. However a study of these rock units has revealed a detailed geologic history.

Chapter 3

GRENVILLIAN BASEMENT - The Cobble Cove Gneiss

The Cobble Cove Gneiss is named after a small inlet located at the mouth of Keystone Brook named by the author (Plate I). It is applied to a unit of quartzo-feldspathic gneiss that is overlain by the Keystone Schist. The base of the Cobble Cove Gneiss is not exposed.

Excellent outcrops of the Cobble Cove Gneiss occur along the shoreline of Glover Island for about 1 km north of Cobble Cove and inland for about 250 metres. Exposure south of Cobble Cove is limited to a few small outcrops in the woods. The type locality and best exposures of the Cobble Cove Gneiss occur along Keystone Brook. A small, isolated sliver of Cobble Cove Gneiss is also recognized along a thrust fault within the Keystone Schist.

Thickness of the Cobble Cove Gneisss cannot be determined as the base of the unit is not exposed. Maximum outcrop ~~width is~~ about 250 metres and a moderate dip implies a structural thickness of about 175 m.

LITHOLOGY AND PETROLOGY

The Cobble Cove Gneiss consists of foliated, quartzo-feldspathic gneisses and thin layers of mafic schist described separately below.

Quartzo-feldspathic gneisses

Quartzo-feldspathic rocks of the Cobble Cove Gneiss are buff-coloured, foliated gneisses streaked with discontinuous, 1-5 X 0.5 cm, mafic laminae composed of concentrations of fine-grained (0.1 mm) mafic minerals (Fig. 3.1). The parallel alignment of these mafic laminae defines a foliation recognized throughout the Cobble Cove Gneiss.

Quartzo-feldspathic rocks of the Cobble Cove Gneiss consist primarily of equidimensional, granoblastic albite, microcline, quartz and randomly oriented laths of biotite (Fig. 3.2). Albite is most abundant followed by microcline, quartz and biotite.

Albite is generally equidimensional and ranges from .5 to 1.0 mm. Self boundaries are curved to cuspate and irregular. Albite twinning is widespread and although locally continuous, generally these twins intersect or are wedge-shaped and probably developed during deformation. Albite locally contains fine-grained sericite, epidote and calcite because of secondary alteration.

Microcline is widespread in the Cobble Cove Gneiss and is the only potassium feldspar recognized. Grains are xenomorphic, 0.5-1.0 mm and cuspate in contact with quartz and plagioclase. Microcline is everywhere twinned and shows a distinctive cross-hatched or tartan pattern. Microprobe analysis yields a composition or Or=96, Ab=4, An=0. No exsolution effects are present.

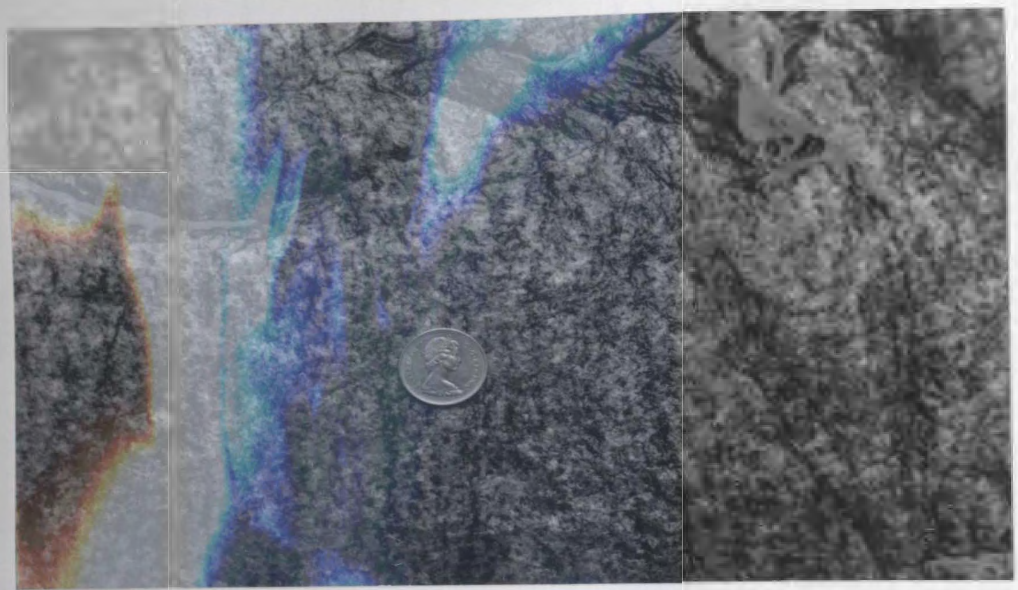


Figure 3.1 Quartzo-feldspathic Cobble Cove Gneiss displaying prominent foliation.

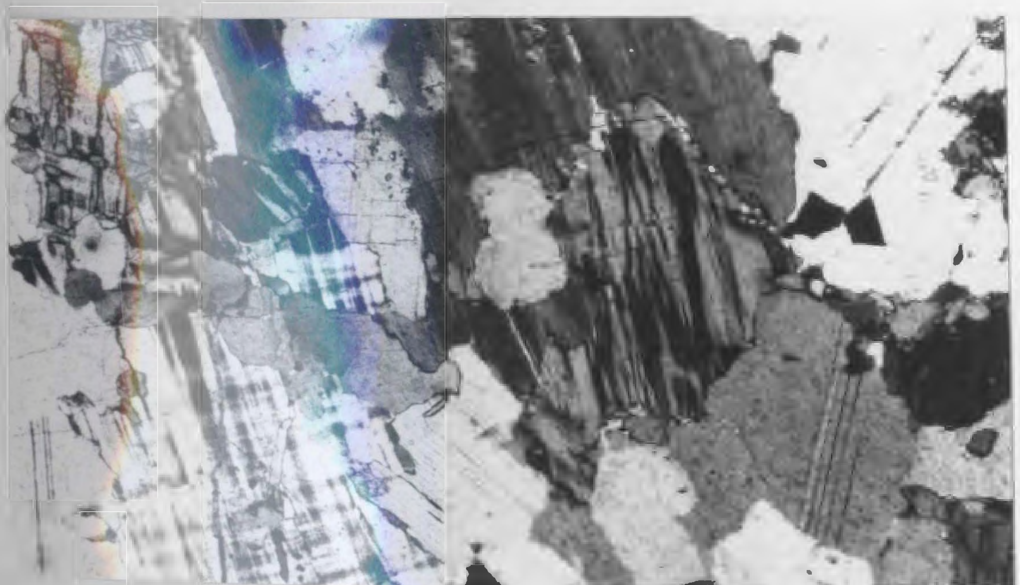


Figure 3.2 Photomicrograph of quartzo-feldspathic Cobble Cove Gneiss. Abbreviations: mi, microcline; pl, plagioclase; q, quartz. 15X.

Quartz occurs as 0.2-0.5 mm, equidimensional grains with curved grain boundaries in contact with plagioclase and microcline. The curvature of these interphase grain boundaries is always concave away from the center of the quartz grain. Self boundaries of quartz are nearly straight or very slightly curved. Quartz aggregates form polygonal microstructures with numerous, nearly 120 degree, triple junctions. Quartz grains are free of inclusions and optical strain effects such as undulatory extinction and deformation lamellae.

The foliation in quartzo-feldspathic rocks of the Cobble Cove Gneiss is formed by concentrations of biotite, opaques, sphene, epidote, zircon, garnet and porphyroclasts of deep-brown to reddish-brown, pleochroic allanite and sphene.

Biotite is a pleochroic green colour (X = light yellow green; Y = Z = dark green) and occurs as subidiomorphic and xenomorphic, randomly oriented, equant grains. The random orientation of the biotite indicates it is post-tectonic with respect to formation of the foliation.

Sphene is an abundant mineral in these mafic laminae and generally forms equant, anhedral grains and rims on iron-titanium oxides.

Zircon is idiomorphic or very slightly rounded in shape and quite coarse (1 mm). Very thin rims surrounding an idiomorphic core are locally observed.

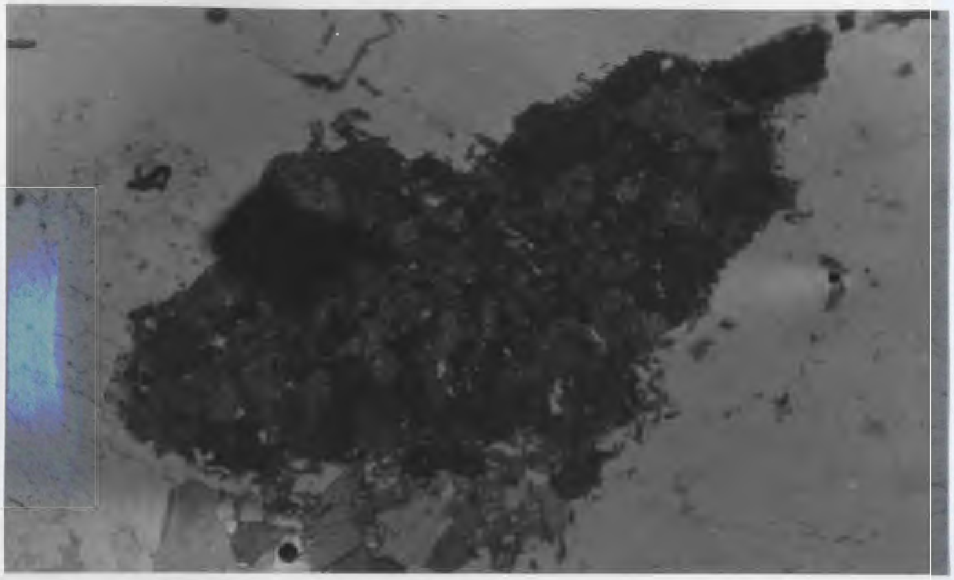
Epidote is anhedral in shape and pleochroic yellow in colour. Locally garnet occurs as web-like or skeletal, anhedral grains.

Distinctive porphyroclasts of a deep-reddish-brown pleochroic allanite containing up to 20 weight percent light rare earth elements (microprobe analysis) also occur in these mafic laminae (Fig. 3.3). These rounded, elongate porphyroclasts consist of an aggregate of allanite and sphene. The allanite forms an interlocking granoblastic aggregate of stumpy, equant grains. Numerous 120 degree triple junctions are present in these allanite aggregates indicating a close approach to microstructural equilibrium. Equant grains of sphene compose up to 15 modal percent of these porphyroclasts. Porphyroclast margins are invariably rimmed by idiomorphic, pleochroic, yellow grains of epidote.

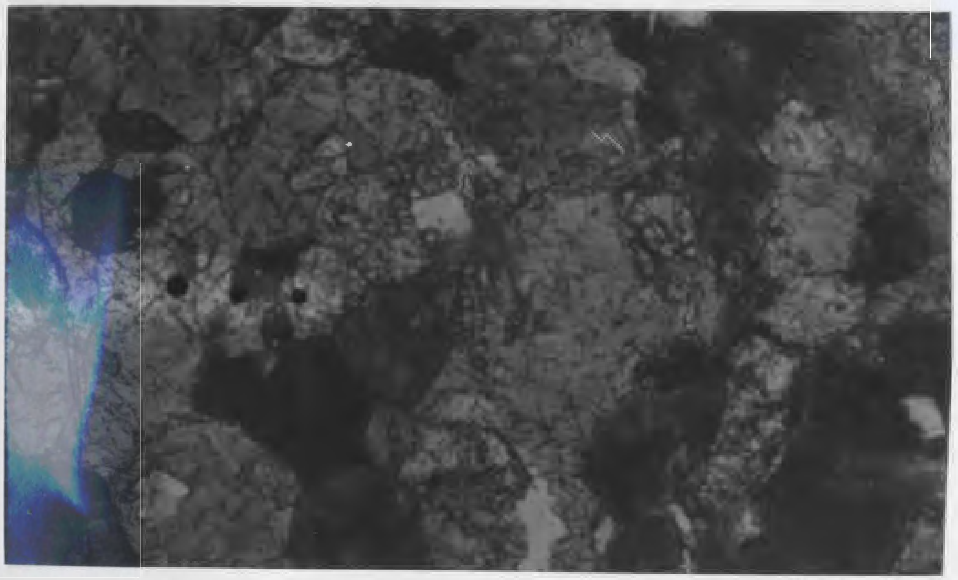
Allanite-sphene porphyroclasts represent an older, relict microstructure as evidenced by their occurrence as porphyroclasts, distinctive mineralogy and the ubiquitous epidote rim that separates the porphyroclast from the groundmass of the rock. The present distribution of these porphyroclasts as augen in the foliation is clearly the result of deformation.

<u>Modal analyses of quartzo-feldspathic gneisses</u>

The composition of quartzo-feldspathic rocks of the Cobble Cove Gneiss has been calculated by modal analysis of nine representative samples. 750 points were counted



a.



b.

Figure 3.3 Photomicrograph of an allanite-sphene porphyroclast in the Cobble Cove Gneiss. (a) entire porphyroclast, X15; (b) Close-up of texture between allanite and sphene within the porphyroclast, X50. Abbreviations: sp, sphene; al, allanite.

for each analysis. The results are shown in Table 3.1 and graphically displayed in Figure 3.4. In Figure 3.4, the modal proportion of quartz (Q), plagioclase (P), and potassium feldspar (K) is plotted. Albite is not plotted with potassium feldspar as it is not the result of exsolution. Appropriate fields are shown labelled after Streckheisen (1967, 1976).

Based upon this system the Cobble Cove Gneiss ranges between granite, granodiorite, tonalite, quartz monzodiorite and monzodiorite in composition. The bulk of the analyses fall into the granodiorite and quartz monzodiorite fields and the area defined by these analyses is centered in the granodiorite field.

Mafic schists

A minor proportion of the Cobble Cove Gneiss consists of mafic rocks which occur as subparallel layers generally less than a metre in thickness. Although these mafic layers contrast in composition with quartzo-feldspathic rocks, their ubiquity in these rocks and absence in the surrounding area indicates they are an integral part of the Cobble Cove Gneiss.

Mafic rocks consist of dark-green to black, medium-grained (1-2 mm), biotite-microcline schists that locally contain distinctive 1-3 mm porphyroblasts of microcline. They occur as 10-100 cm thick layers within quartzo-

Table 3.1 Modal analyses of quartzo-feldspathic rocks of the Cobble Cove Gneiss based upon 750 points per analysis in percent.

	1	2	3	4	5	6	7	8	9
microcline	23.37	15.92	14.46	17.60	24.44	30.15	5.95	.00	25.65
plagioclase	46.43[2]	60.75	73.78	72.45	48.45	35.93	43.88	61.03	57.74
quartz	23.29	8.38	6.22	1.70	26.31	32.97	47.08	32.67	7.06
biotite[1]	3.30	9.36	3.78	4.49	.50	.00	.00	4.78	8.67
muscovite	.00	.00	.00	.00	tr.	tr.	3.09	.00	.00
calcite	2.77	.85	1.22	.97	.00	.81	.00	1.28	.25
opaques	.61	2.92	.13	1.33	.00	.13	.00	.00	.25
hornblende	.08	.36	tr.	.00	.00	.00	.00	.00	tr.
epidote	.15	1.46	.40	.00	.00	.00	.00	.00	.12
zircon	tr.	tr.	tr.	.68	tr.	.00	.00	.23	.25
apatite	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.	tr.
sphene	.00	tr.	tr.	.78	.00	.00	.00	.00	.00
garnet	.00	tr.	.00	.00	.00	.00	.00	.00	.00

[1] biotite is locally altered to chlorite [2] albite (microprobe analysis)

1 Granodioritic, biotite orthogneiss; Biotite is partially retrograded to chlorite; Epidote forms overgrowths on relict hornblende.

2 Quartz monzodioritic, garnet-biotite orthogneiss; Epidote forms overgrowths on relict hornblende; Sphene occurs as overgrowths on opaques; A large (1 mm) poikiloblastic garnet is present locally.

3 Quartz monzodioritic biotite orthogneiss; Biotite is olive-green; Epidote forms overgrowths on unresolvable mineral, probably hornblende.

4 Monzodioritic, biotite orthogneiss; Biotite is highly altered to chlorite; Zircon is present as large, xenomorphic grains and as small, idiomorphic grains.

5 Leucocratic, granodioritic, biotite orthogneiss; Biotite is locally altered to chlorite.

6 Leucocratic, granitic orthogneiss; Mafic minerals are absent.

7 Leucocratic, granodioritic, muscovite orthogneiss; Muscovite is secondary.

8 Tonalitic, biotite orthogneiss; Biotite is oxidized and altered to chlorite; Zircons are small and idiomorphic.

9 Quartz monzodioritic, biotite orthogneiss; Biotite is olive-green and locally altered to chlorite; Zircon is coarse (0.1 mm) and unaltered; unresolvable clots are inferred to be altered hornblende.

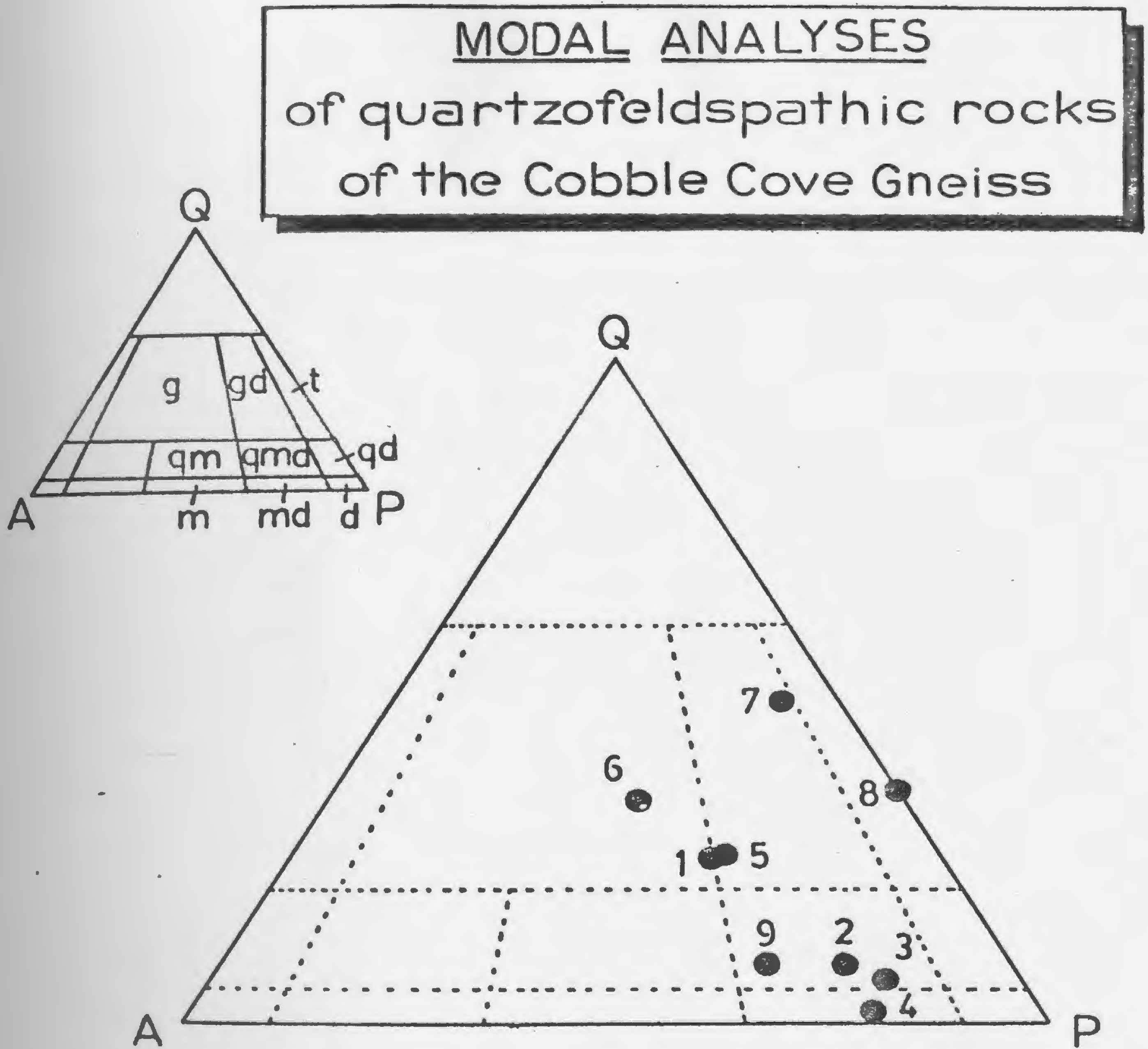


Figure 3.4 Modal analyses of quartzo-feldspathic rocks of the Cobble Cove Gneiss based upon 750 points per section. Fields shown are those of Streckheisen (1967, 1976): g, granite; gd, granodiorite; t, tonalite; qm, quartz monzonite; qmd, quartz monzodiorite; qd, quartz diorite; m, monzonite; md, monzodiorite; d, diorite; A, alkali feldspar; Q, quartz, P, plagioclase. Numbered points refer to analyses given in Table 3.1.

feldspathic rocks of the Cobble Cove Gneiss (Fig. 3.5). Contacts with the quartzo-feldspathic rocks are abrupt. At the type locality of the unit in Keystone Brook, a single 50 cm thick layer can be traced continuously for 16 metres.

Biotite-microcline schist layers parallel the strong gneissosity in quartzo-feldspathic rocks of the Cobble Cove Gneiss or intersect at a very low angle. They contain a weak schistosity.

Biotite-microcline schists consist of biotite, microcline, plagioclase, calcite, quartz, muscovite, epidote, chlorite, sphene and apatite (Fig. 3.6). Visually estimated modal analyses of three of these mafic schists are given in Table 3.2.

Mafic schists are composed primarily of pleochroic, olive-green biotite (X = light yellow green; Y = Z = dark olive green). It is generally subidioblastic and elongate. No preferred orientation is apparent.

Microcline is the next most abundant mineral in these mafic schists. Locally it is porphyroblastic (1-3 mm) and in most cases exhibits cross-hatched or tartan twinning. Exsolution effects are absent but porphyroblasts contain numerous inclusions of epidote.

Epidote occurs as inclusions within microcline porphyroblasts but is absent in the groundmass. It is generally idiomorphic or subidiomorphic and is pleochroic yellow-green

(Z = yellow green; Y = Z - colourless to very pale yellow green). Optically it is negative and has a 2V of 72 degrees.

Chemical analyses of mafic schists

Three chemical analyses of mafic schists, determined by Atomic Absorption Spectrometry, are given in Table 3.3 (see Appendix for methods and precision). Based upon these analyses it is concluded that the composition of these schists is very similar to that of a basalt except for anomalously high K_2O. However K_2O is notoriously mobile during metamorphism suggesting this compositional similarity may be significant.

CONTACT RELATIONS

The contact between the Cobble Cove Gneiss and the Keystone Schist is exposed in Keystone Brook and along the shoreline of Glover Island northeast of Cobble Cove. In Keystone Brook this contact is interpreted to be a fault. Here microcline-porphyroblastic, semipelites of the Keystone Schist are in sharp contact with a high-strain facies of the Cobble Cove Gneiss. The tan weathering, high-strain zone is approximately 10 m wide and consists of fine-grained, quartzo-feldspathic gneiss and highly deformed mafic schist which form a matrix to rounded, tectonic clasts of undeformed gneiss and vein quartz. Faulting along the contact between the Cobble Cove Gneiss and the

Figure 3.5 Mafic schist layer within quartzo-feldspathic rocks of the Cobble Cove Gneiss in Keystone Brook.

Figure 3.6 Photomicrograph of mafic schist from the Cobble Cove Gneiss. Abbreviations: mi, microcline; bio, biotite; epi, epidote. X20.

Table 3.2 Visually estimated modal analyses of mafic rocks in the Cobble Cove Gneiss.

	1	2	3
biotite	40	53	56
microcline	27	25	25
plagioclase	27	2	0
sphene	2	tr.[1]	tr.[1]
quartz	0	0	3
epidote	5	2[2]	1[2]
chlorite	0	0	tr.
calcite	7	10	10
opaques	tr.	3	3
muscovite	tr.	5	2
apatite	tr.	tr.	0

[1] cores of opaque material only

[2] as inclusions within microcline only

1 - Dark-green to black, medium-grained (1 mm), biotite-microcline schist; Biotite is pleochroic green; Epidote is light-green and occurs in the groundmass and as inclusions in microcline.

2 - Dark-green to black, medium-grained (1 mm), porphyroblastic, biotite-microcline schist; Microcline forms 1-2 mm porphyroblasts; Biotite is pleochroic green as in 1; Epidote is a light-green colour and is present as inclusions within microcline porphyroblasts; Sphene occurs locally in the cores of opaques and as inclusions within microcline.

3 - Same as 2 above.

Table 3.3 Chemical analyses of samples 1, 2 and 3 of Table 3.2 compared with the average tholeiitic basalt and alkali basalt computed by Manson (1967). See appendices for methods and precision.

	1	2	3	4	5	6
SiO_2	43.6	44.8	42.9	46.1	50.7	46.6
TiO_2	1.84	1.72	1.83	1.90	1.6	2.5
Al_2O_3	14.1	14.2	14.0	14.9	16.1	15.7
$Fe_2O_3^t$	14.68	15.78	14.66	15.86	11.4	12.0
MnO	.28	.47	.42	.41	.17	.16
MgO	5.54	4.79	4.61	5.25	6.1	7.0
CaO	5.05	3.82	5.92	5.21	9.8	10.0
Na_2O	1.09	.85	.78	.96	2.5	3.2
K_2O	7.38	8.12	8.06	8.28	.7	1.4
P_2O_5	.09	.07	.05	.07	.22	.50
L.O.I.[1]	4.60	4.67	6.12	1.00	.7	1.0

[t] total iron as ferric iron

[1] Loss On Ignition

1, 2, 3 - Chemical analyses of mafic schists 1, 2 and 3 of Table 3.2, respectively.

4 - Average of analyses 1, 2 and 3 normalized to 100 percent with 1.0 percent Loss On Ignition.

5 - Average of 897 tholeiitic basalts from Manson (1967).

6 - Average of 661 alkali basalts from Manson (1967).

Keystone Schist is interpreted to have formed this high strain facies.

The contact between these rocks is also exposed along the shoreline of Glover Island northeast of Keystone Brook. Unfortunately small intrusions of pink, massive, undeformed granite and granitic pegmatite obscure contact relationships at this locality.

A tectonic sliver of Cobble Cove Gneiss also occurs along a structural surface separating the Upper and Lower Members of the Keystone Schist in Keystone Brook. Contacts with the Keystone Schist are abrupt. The Cobble Cove Gneiss is slabby in appearance due to structural overprinting. This fault sliver outcrops only in Keystone Brook and is thus of limited size and extent.

STRUCTURE AND METAMORPHISM

Rocks of the Cobble Cove Gneiss contain structures related to two deformational events that have been overprinted by a static, post-tectonic metamorphism. D1 structures consist of a well developed foliation in quartzo-feldspathic rocks while D2 structures consist of a weak foliation in mafic rocks.

S1 in quartzo-feldspathic rocks is defined by discontinuous mafic laminae that are constant in orientation and intensity throughout the Cobble Cove Gneiss.

S2 consists of a weak grain shape preferred orientation of plagioclase recognized only in mafic schists.

It is very difficult to directly assign relative ages to S1 and S2 because they are subparallel, occur in contrasting lithologies and are not directly observed overprinting one another. The basis for the assigned relative age of these structures is instead, the origin of the mafic schist layers. These layers are interpreted to be mafic dykes intrusive into the quartzo-feldspathic gneisses based upon cross-cutting relationships, field appearance and chemistry. These mafic schist layers locally truncate S1 at a low angle indicating dyke intrusion occurred after D1. The faint foliation observed in these mafic schists must therefore post-date dyke intrusion.

A static, post-tectonic metamorphic event followed D2 and recrystallized quartzo-feldspathic rocks forming aggregates of polygonal, strain-free quartz (Fig. 3.7) and decussate biotite. Porphyroblasts of biotite and microcline formed in mafic schists.

INTERPRETATION AND AGE

Quartzo-feldspathic gneisses of the Cobble Cove Gneiss are interpreted to represent orthogneisses based upon their modal composition and homogeneity. The granodioritic modal composition of these quartzo-feldspathic gneisses is typical of a granodiorite pluton, a common lithology in orogenic terranes. The overall homogeneity in composition and

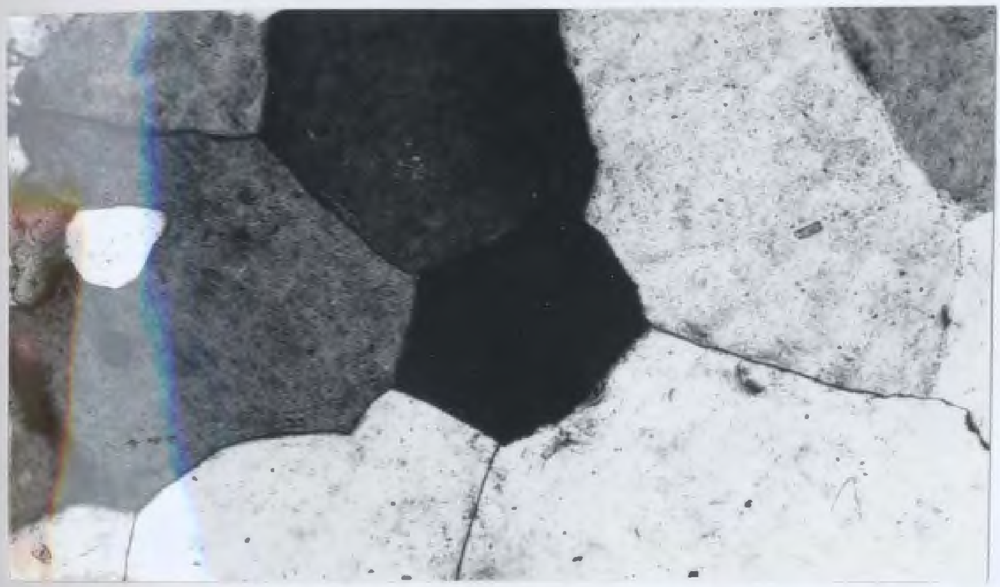


Figure 3.7 Photomicrograph of quartz microstructure within the Cobble Cove Gneiss. Note the lack of undulatory extinction and 120 degree triple junctions displayed by the quartz. X20.

field appearance also support this interpretation.

Mafic schists strongly resemble intrusive dykes or sills based upon cross-cutting relationships and appearance. They are also very similar to basalt in chemistry except for anomalously high values of potash (Table 3.3). However this discrepancy is easily accounted for if potash metasomatism is considered. Potash is a highly mobile element during metamorphism. Mafic schists of the Cobble Cove Gneiss underwent a static metamorphic event during which porphyroblast growth of biotite and microcline was extreme. These minerals both contain significant potash. It is concluded that this metamorphic event was accompanied by the influx of a potash-rich fluid that resulted in the growth of microcline and biotite and the elimination of the original mineralogy. Epidote may represent a relict mineral preserved from this original mineralogy. Epidote occurs solely as inclusions within microcline where it is isolated from metasomatic fluids. Epidote inclusions would thus be preserved while groundmass epidote would have been destroyed. The occurrence of epidote inclusions thus supports a metasomatic origin for these mafic schists. Epidote is a common mineral in greenschists and epidote amphibolites, a likely protolith composition of these mafic schists.

Highly deformed and metasomatized dyke-intruded orthogneisses contrast strongly with the less deformed overlying metaclastic rocks of the Keystone Schist. In particular the pervasive, well developed foliation in the Cobble Cove Gneiss is not represented in these less intensely deformed rocks in which primary sedimentary structures can still locally be recognized. This contrast implies a basement-cover relationship.

In western Newfoundland it is commonplace to observe metasediments overlying a deformed, dyke-intruded basement complex. Examples occur at the Strait of Belle Isle (Williams and Stevens, 1969), Indian Head (Riley, 1962) on the Baie Verte Peninsula (DeWit, 1974) and south of Grand Lake (Kennedy, 1980; Knapp et al., 1979). Thus based upon the regional geology of western Newfoundland it is concluded that the Cobble Cove Gneiss is basement to the Keystone Schist and Grenvillian in age. The pervasive S1 foliation is also interpreted to be Grenvillian in age as high temperatures and strains are required to produce such a well developed foliation. Such conditions are typical of the Grenville Structural Province.

Post-Grenvillian structures are recognized only in mafic schists of the Cobble Cove Gneiss as granodioritic gneiss is a relatively insensitive recorder of deformation events.

Chapter 4

LOWER PALEOZOIC METACLASTIC ROCKS - The Keystone Schist

Polydeformed and metamorphosed clastic rocks that structurally overlie the Cobble Cove Gneiss and are structurally overlain by the Grand Lake Complex (Plate I; Fig. 4.1) are named the Keystone Schist after Keystone Brook. The Keystone Schist is subdivided into two members designated the Upper Member and the Lower Member. These two units are separated by a thrust sliver of Cobble Cove Gneiss in Keystone Brook. Shoreline exposures and exposures in Keystone Brook are excellent. Keystone Brook is designated as the type locality for this unit.

Structural thickness of the Keystone Schist is 700 metres. The true thickness of this unit is unknown as the stratigraphic top and base of the section are missing and the rocks are complicated by faulting and isoclinal folding.

TYPE SECTION

Keystone Brook runs approximately normal to the strike of the Keystone Schist and exposes stream-polished outcrops throughout its extent. This superb type section is described below and summarized in Figure 4.2.

The Lower Member of the Keystone Schist is in abrupt contact with the Cobble Cove Gneiss in Keystone Brook. A 10 m thick zone of highly deformed gneiss occurs at the top

Figure 4.1 Geologic map of the central part of Glover Island.

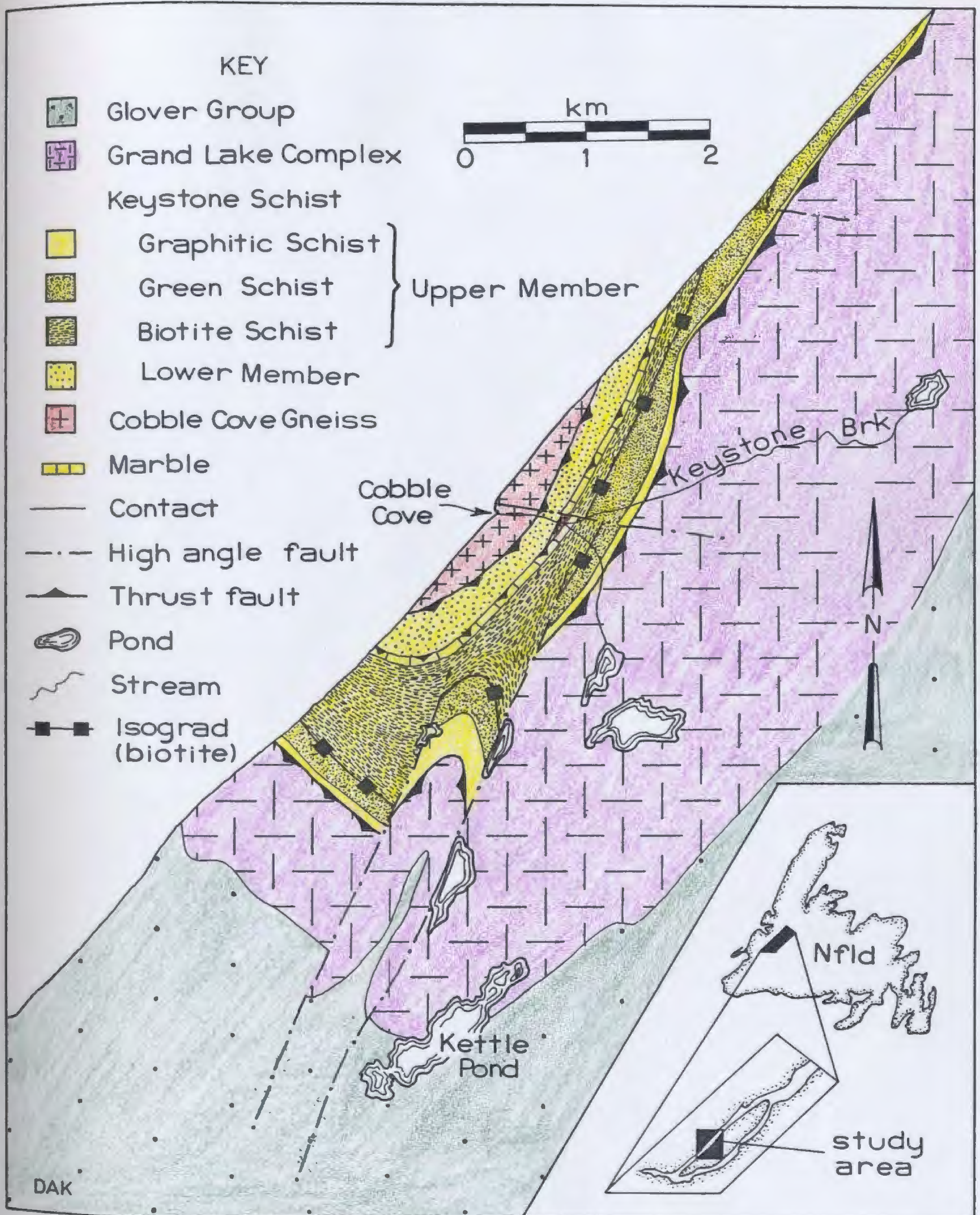
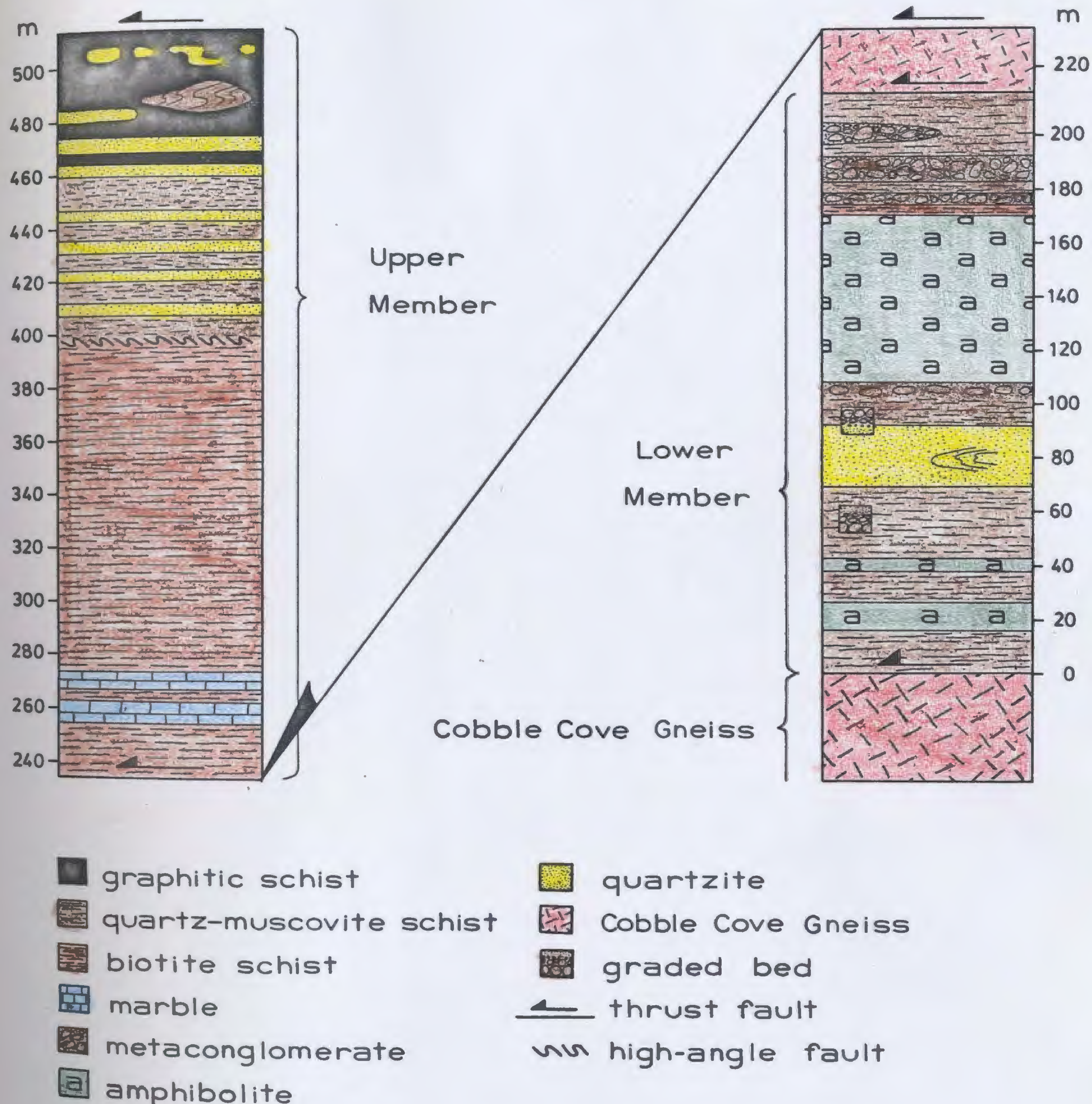

KEY
Glover Group
Grand Lake Complex
Keystone Schist
Graphitic Schist
Green Schist
Biotite Schist
Upper Member
Lower Member
Cobble Cove Gneiss
Marble
Contact
High angle fault
Thrust fault
Pond
Stream
Isograd (biotite)
km
0
1
2
Cobble Cove
Keystone Brk
N
Kettle Pond
Nfld
study area
DAK

Figure 4.2 Schematic section through the Keystone Schist in its type locality at Keystone Brook.



m
500
480
460
440
420
400
380
360
340
320
300
280
260
240
Upper Member
Lower Member
Cobble Cove Gneiss
m
220
200
180
160
140
120
100
80
60
40
20
0
graphitic schist
quartz-muscovite schist
biotite schist
marble
metaconglomerate
amphibolite
quartzite
Cobble Cove Gneiss
graded bed
thrust fault
high-angle fault

of the Cobble Cove Gneiss. The contact is interpreted to be a thrust fault and the tectonized gneisses, the result of dislocation along it.

Highly foliated, microcline-biotite schists with 2 mm porphyroblasts of microcline occur at the contact and are interlayered with quartz-muscovite schists and thin 1-2 m concordant amphibolites upstream of the contact.

Above the amphibolites a second schistosity becomes macroscopically recognizable in the hinge region of a major, late (F2) fold. Here early and late cleavages are conspicuous as they intersect at right angles. Elsewhere the two are indistinguishable in the field as they are parallel.

Bedding is also quite distinct in the hinge region of this fold and graded beds can be locally recognized. From facing directions of these graded beds the major fold is observed to be an upright, isoclinal syncline plunging gently to the south as discussed more completely in a later chapter.

Rocks in the hinge area of this major syncline consist of medium to fine-grained (1-2 mm) white-weathering, arkosic quartzites. The second cleavage is most obvious in these quartzites where it forms a differentiated crenulation cleavage.

Upstream the facing direction of graded beds changes from east to west across the hinge of this major fold and semipelitic schists outcrop. Interlayered with these schists is a 60 m thick amphibolite that consists of dark-green hornblende grains, 1-3 mm in length, in a plagioclase matrix. At its stratigraphic base (east) are biotite-epidote schists with porphyroblasts of biotite up to 4 mm in diameter. The top of this thick amphibolite layer (west) consists of a quartz-muscovite schist containing flattened clasts of quartzite up to 20 cm in length (Fig. 4.3). This thick unit of amphibolite is inferred to be continuous with the much smaller layers of amphibolite at the same stratigraphic position on the west limb of this major fold.

Upstream from this thick unit of amphibolite, quartz-muscovite schists are locally interbedded with 10 cm to 1 m thick pebble conglomerate horizons (Fig. 4.4). These are discontinuous, ungraded, poorly sorted, clast-supported conglomerates containing abundant rounded quartz and plagioclase clasts up to 2 cm in diameter.

These conglomerate rocks are succeeded by a 50 m thick section of Cobble Cove Gneiss emplaced along a thrust fault within the Keystone Schist which separates the Lower Member from the Upper Member.

This gneiss is in contact with rusty-weathering calcareous schists and white marbles (Fig. 4.5) on its upstream side which grade into a thick section (240 m) of homogeneous, black to dark-grey biotite schists.

Biotite schists are abruptly truncated by a minor, east-trending fault that cuts across Keystone Brook (Fig. 4.1). Upstream of this fault are sub-biotite grade, green and tan,

Figure 4.3 Flattened clasts of quartz at the top of the thickest amphibolite layer in the Keystone Schist in Keystone Brook.

Figure 4.4 Pebble metaconglomerate from the Lower Member of the Keystone Schist in Keystone Brook. Note stretched quartz clasts.

Figure 4.5 Interbedded calcareous schist and marble at the base of the Upper Member of the Keystone Schist in Keystone Brook.

quartz-sericite schists and quartzites. A significant amount of section including the biotite isograd is missing across this small fault. However, a continuous section across the biotite isograd occurs at this stratigraphic position along the shoreline of Glover Island north of Keystone Brook.

Quartz-sericite schists are locally interbedded with thin (2-5 cm) orange-weathering beds of plagioclase and carbonate, and, at the very top of the Keystone Schist, with graphitic schists. Graphitic schists are very highly deformed and locally contain boudinaged beds of arkosic metasandstone (Fig. 4.6). One 2 m, equant pod of orange-weathering carbonate may have been derived from an ultramafic rock. However it cannot be determined if this pod represents a block in black shale or simply a sliver of altered ultramafic rock emplaced along a minor fault.

Black schists are abruptly juxtaposed against greenschists of the Grand Lake Complex and mark the top of the Keystone Schist.

Although the above descriptions refer specifically to the Keystone Brook section, these lithologies can be traced throughout the extent of the Keystone Schist as shown in Figure 4.1.

a.

b.

Figure 4.6 Graphitic schist at the top of the Keystone Schist. (a) Thin layers of graphitic pelite alternating with thin layers of psammite. (b) Large block of psammite in a black, pelitic matrix. Block is part of a disrupted bed.

LITHOLOGY AND PETROLOGY

The Lower Member of the Keystone Schist consists of microcline schists, conglomeratic schists, biotite schists, amphibolite and quartzite. The Upper Member consists of biotite schist, quartz-sericite schists, marble and graphitic schists. The distribution of these lithologies is given in Plate I and Figure 4.1, modal analyses estimated from thin sections are presented in Table 4.1 and petrologic descriptions are given below.

Lower Member of the Keystone Schist

Microcline schists

Microcline schists are abundant in the Lower Member of the Keystone Schist. They consist of medium-grained, quartz-muscovite-plagioclase-biotite schists with 2 mm, equant porphyroblasts of microcline (Fig. 4.7 and 4.8). Quartz inclusions within porphyroblasts define straight inclusion trails except in rim areas where these inclusion trails abruptly curve into parallelism with the second foliation. Based upon this microstructure these microcline porphyroblasts are interpreted to be pre- and syntectonic with the second foliation.

Conglomeratic schists

Conglomeratic schists consist of flattened clasts of quartz and equant clasts of plagioclase up to 1 cm in length

Table 4.1 Visually estimated modal analyses of rocks of the Keystone Schist.

	1	2	3	4	5	6	7
garnet	0	5	0	0	0	0	0
biotite	5	10	15	35	0	0	0
muscovite	5	25	25	15	20	25	25
quartz	65	60	30	33	40	45	60
plagioclase	25	tr.	30	15	35	30	10
chlorite	tr.	tr.	0	0	5	0	0
epidote	tr.	0	0	2	0	0	0
calcite	tr.	0	tr.	tr.	0	tr.	5
apatite	tr.	0	0	0	0	tr.	tr.
sphene	0	0	0	0	0	0	0
zircon	0	0	tr.	tr.	0	tr.	0
tourmaline	0	0	0	0	0	0	tr.
opaques	tr.	tr.	tr.	tr.	tr.	tr.	tr.

1 Biotite schist - Lower Member
2 Garnet-biotite schist - Lower Member
3 Biotite schist - Upper Member
4 Biotite schist - Upper Member
5 Chlorite phyllite - Upper Member
6 Green phyllite - Upper Member
7 Graphitic phyllite - Upper Member

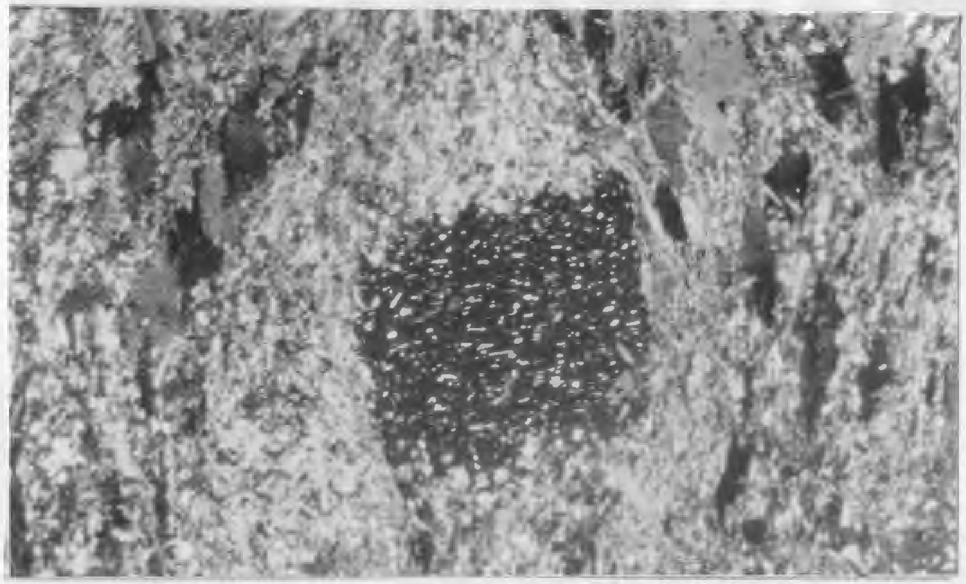


Figure 4.7   Microcline schist from the Lower Member of the Keystone Schist. Microcline porphyroblast contains straight inclusion trails of quartz that define S1. The dominant foliation in the matrix of the schist is S2. X10.

Figure 4.8   Microcline schist in the Keystone Schist. Note millimetre-sized porphyroblasts of microcline.

in a fine-grained, granoblastic matrix of quartz and muscovite. Coarse quartz grains locally contain abundant rutile needles.

Biotite schists

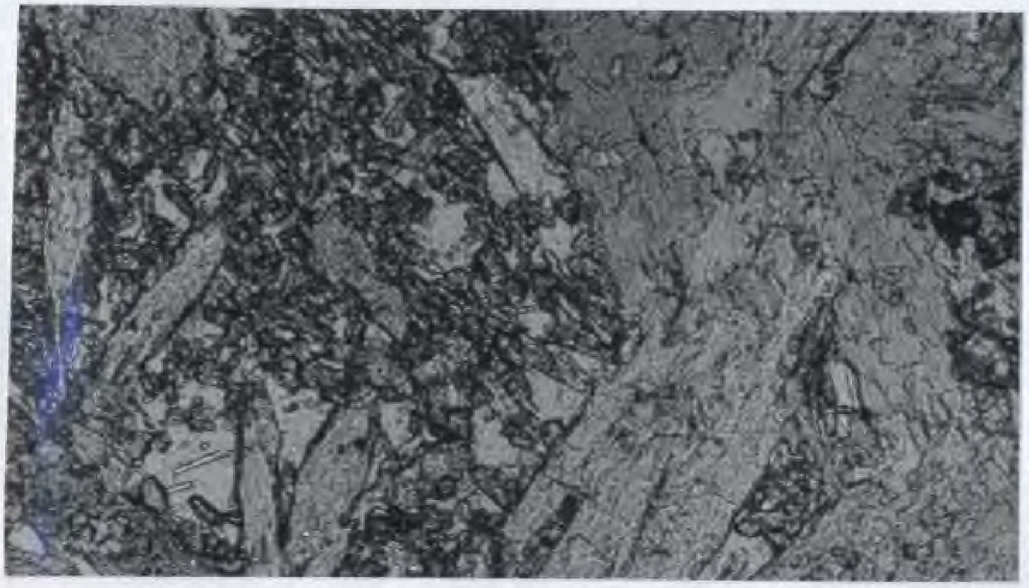
Fine-grained, light-green biotite schists occur at the base of the amphibolite in Keystone Brook. These schists consist of fine-grained epidote, quartz, muscovite and opaques overgrown by 2-4 mm porphyroblasts of olive-green, pleochroic, randomly-oriented biotite. This rock is interpreted to be a metamorphosed mafic sediment or tuff.

Amphibolite

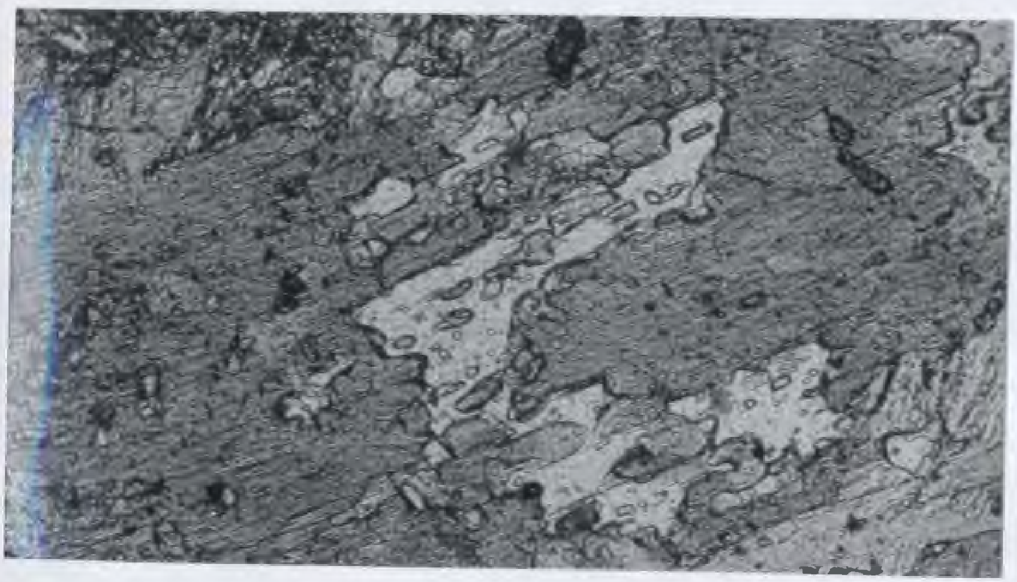
Amphibolites in the Lower Member consist of plagioclase, hornblende and epidote (Fig. 4.9). The hornblende is pleochroic blue-green (Z = blue green; Y = light yellow green; X = light brownish yellow), has a 2V of 76° and a $Z \wedge C$ of 17°. It is intimately intergrown with quartz in a symplectite microstructure. Plagioclase is oligoclase (An 21; microprobe analysis) and contains abundant idiomorphic inclusions of fine-grained epidote. Olive-green biotite is present in some examples.

Quartzites

Quartzites consist of a granoblastic, polygonal aggregate of quartz, plagioclase and muscovite. The



a.



b.

Figure 4.9 Amphibolite within the Keystone Schist. (a) Hornblende-oligoclase-epidote microstructure, X20. (b) Symplectite texture within hornblende, X50.

plagioclase is interpreted to represent detrital grains recrystallized by subsequent metamorphism.

Upper Member of the Keystone Schist

Quartz-sericite schists

Quartz-sericite schists consist of fine-grained, quartz-muscovite-plagioclase-chlorite schists. Quartz and plagioclase grains, 1-3 mm in diameter, occur in a fine-grained, schistose quartz-plagioclase-sericite matrix. Chlorite generally forms 1-2 mm clots or blebs. Traces of brown biotite are locally present in the cores of these chloritic patches suggesting they represent retrograded biotite grains. Coarse quartz and plagioclase grains are interpreted to be detrital. Tourmaline, apatite and zircon grains are also recognized. Tourmaline is metamorphic in origin but apatite and zircon may be detrital.

Graphitic schist

Graphitic schists occur at the very top of the Upper Member. They are thinly layered and composed of millimetre-scale alternating quartz-plagioclase and chlorite-muscovite layers.

Quartz-plagioclase layers consist of variably-sized grains of detrital plagioclase in a fine-grained quartz matrix. These alternate with millimetre-scale muscovite layers containing abundant, fine-grained dusty inclusions of graphite.

Other minerals recognized in the graphitic schists include tourmaline, apatite and biotite. Tourmaline is quite widespread and consists of strongly pleochroic brown and green, idiomorphic elongate crystals which occur in both quartz-feldspar and micaceous layers. A single, large rounded detrital grain of apatite has also been recognized.

Biotite schists

Biotite schists in the Upper Member consist of 1-3 mm, randomly oriented, brown, pleochroic porphyroblasts of biotite set in a fine-grained, schistose matrix of muscovite and quartz. These are clearly metasediments as bedding is locally preserved.

Marble

Marble units within the Upper Member consist of equant, polygonal grains of calcite. Muscovite is also present in more schistose marble beds.

STRUCTURE OF THE KEYSTONE SCHIST

Four deformational events are recognized in the Keystone Schist: D1, D2, D3 and D4. Structures relating to these deformational events are described below. A geometric analysis of these structures follows this section.

D1 Structures

A well-developed, penetrative, regional schistosity is recognized throughout the Keystone Schist and is inferred to have formed during D1. No D1 folds or other structures are recognized due to transposition and high strain related to D2. It is therefore not known for certain if this schistosity formed during a deformational event. However as most regional schistosities are associated with folds (e.g., Hobbs et al., 1976) this schistosity is also inferred to be associated with folding and therefore to have formed during a deformational event.

Although difficult to recognize in the field due to transposition, S1 is easily identified in thin section as it is ubiquitously overprinted by post-tectonic porphyroblasts of biotite, microcline, garnet and epidote.

S1 is defined by the parallel alignment of muscovite and chlorite. Quartz and plagioclase are anhedral and generally form grain boundaries with muscovite or chlorite. Epidote and tourmaline form idiomorphic or subidiomorphic elongate prisms parallel to S1 and are interpreted to have formed synkinematic with S1.

D2 Structures

Structures related to D2 deformation consist of a penetrative differentiated crenulation cleavage, S2; a well-developed pressure shadow lineation, L2; and major and minor isoclinal folds, F2.

S2 forms the dominant, planar, fabric element observed in the field (Fig. 4.10). It consists of a differentiated crenulation cleavage formed by transposition of S1. S2 is most apparent in psammitic rocks in the hinge regions of major F2 folds. Here S2 is defined by alternating mica-rich and quartz-rich laminae. Quartz-rich laminae are 1-2 mm in width and are inferred to have formed by the transferral of quartz from mica-rich domains to quartz-rich domains during cleavage formation as described by Hobbs et al., (1976; p. 243) and Gray (1977).

In limb regions of major F2 folds, S2 is subparallel with S1 and difficult to identify in the field where it appears as a single schistosity. However it is easily identified in thin section in these areas because of its unique microstructure. Abundant porphyroblasts of biotite, microcline, garnet and epidote formed in the interkinematic period between D1 and D2. These porphyroblasts are overprinted and deformed by S2 but post-tectonically overprint S1. S2 is thus easily identified by a consideration of porphyroblast-cleavage relationships.

S2 structures are generally associated with a conspicuous pressure shadow lineation, L2. This lineation is best developed around deformed porphyroblasts that formed during the interkinematic period separating D1 and D2. Pressure shadows are composed of quartz and muscovite and appear on

Figure 4.10 Semipelitic schist of the Lower Member of the Keystone Schist in Keystone Brook. Thin laminae is an S2 crenulation cleavage. Diffuse banding may represent bedding.

S2 foliation surfaces as small ridges extending away from the host porphyroblast. This lineation is most obvious where the host porphyroblast is microcline.

Minor folds associated with D2 deformation are extremely rare and are only recognized in Keystone Brook and along the shoreline of Glover Island south of Cobble Cove. In Keystone Brook, fine pelitic layers in a psammite are folded into an isoclinal minor fold with a well-developed, axial-planar, S2 crenulation cleavage. Biotites, 1-2 mm in size, are truncated by this crenulation cleavage. This minor fold is strongly similar in profile (Fig. 4.11).

Minor folds affect a black to dark-grey semipelite of the Upper Member along the shoreline of Glover Island south of Cobble Cove. The minor folds are defined by pelitic layers within a semipelite and are extremely difficult to recognize due to almost complete transposition of bedding. Hinge regions of minor folds involving pelite layers consist of thin (1-10 mm) layers of pelite alternating with thin layers of more quartz-rich material. These layers are in sharp contact with one another and are strictly parallel with S2. The gross distribution of these pelite layers defines a zone oriented at high angles to S2 which is interpreted to reflect the original orientation of bedding.

Lack of other recognizable F2 folds in the Keystone Schist is attributed to the intense transposition that accompanies their formation. F2 folds in Keystone Brook occur in the hinge region of a major F2 fold. Elsewhere, recognizable F2 folds are limited to areas of exceptional outcrop exposure where strongly contrasting lithologies are folded.

D3 Structures

Structures related to D3 deformation consist of minor and major folds, a heterogeneously developed crenulation cleavage and abundant intersection lineations between S3 and bedding or S2. D3 structures are heterogeneously developed on a regional scale. They are more abundant and more intensely developed in the upper parts of the Keystone Schist and in basal parts of the overlying Glover Group.

Tight to isoclinal asymmetric folds are the most obvious D3 structure (Fig. 4.12). These folds are abundant in the upper parts of the Keystone Schist and basal parts of the Glover Group where they locally have amplitudes of 3-4 m. These folds characteristically have curvilinear fold axes and similar profiles in homogeneous lithologies. In heterogeneous lithologies the more rigid beds form buckle folds with little or no hinge thickening while the more ductile material forms nearly ideal, similar folds that demonstrate extreme hinge thickening and limb thinning.

Folds of D3 age are generally associated with a well-developed, axial planar, S3 crenulation cleavage. This crenulation cleavage is defined by rotated S1 muscovite

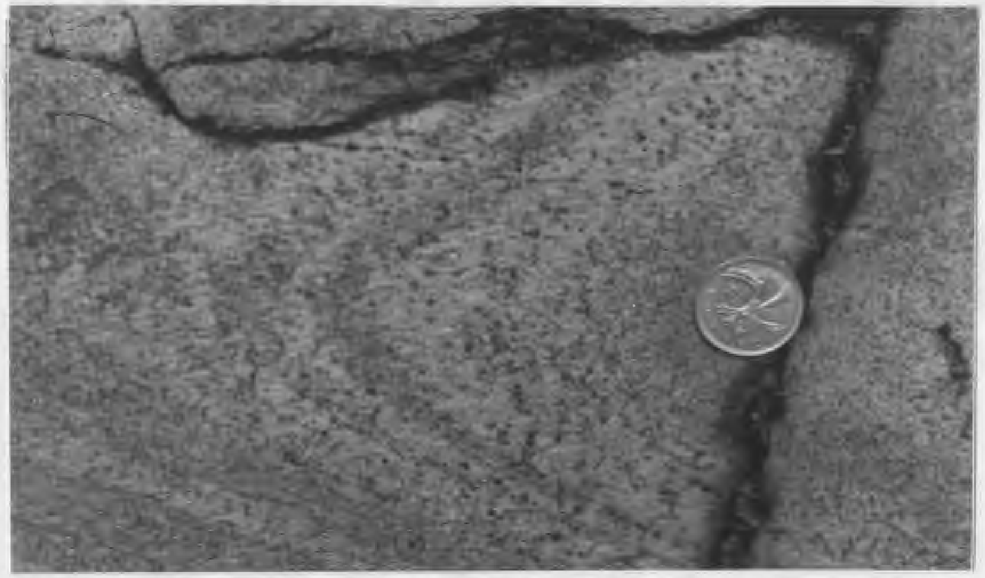


Figure 4.11 F2 fold within the Keystone Schist. Photo is taken from hinge area of major F2 fold in Keystone Brook.

Figure 4.12 F3 folds within the Keystone Schist. Photo is taken approximately 3 km southwest of Bluff Head along the shoreline of Glover Island.

and is locally differentiated. Some muscovite growth occurred during D3 but this is generally very minor. Chlorite growth is widespread during D3 however.

Abundant lineations formed furing D3. These are invariably intersection lineations formed by the intersection of S3 with S2 and are generally expressed as tiny crenulations on S2 foliation surfaces.

D4 Structures

The distribution of the various lithologic units on Glover Island is largely dominated by a major, south-plunging north-south trending anticline termed the Glover Anticline. The deformational event that formed this major anticline is designated D4. The Cobble Cove Gneiss and Keystone Schist are exposed in the core of the Glover Anticline testifying to the scale of this structure.

Correlation of minor structures with a major structure the size of the Glover Anticline is difficult. However a heterogeneously-developed crenulation cleavage overprints S3 and is axial planar to the Glover Anticline. This crenulation cleavage maintains its orientation across the Glover Anticline and is accordingly inferred to be of D4 age. No other structures are correlated with D4.

GEOMETRIC ANALYSIS

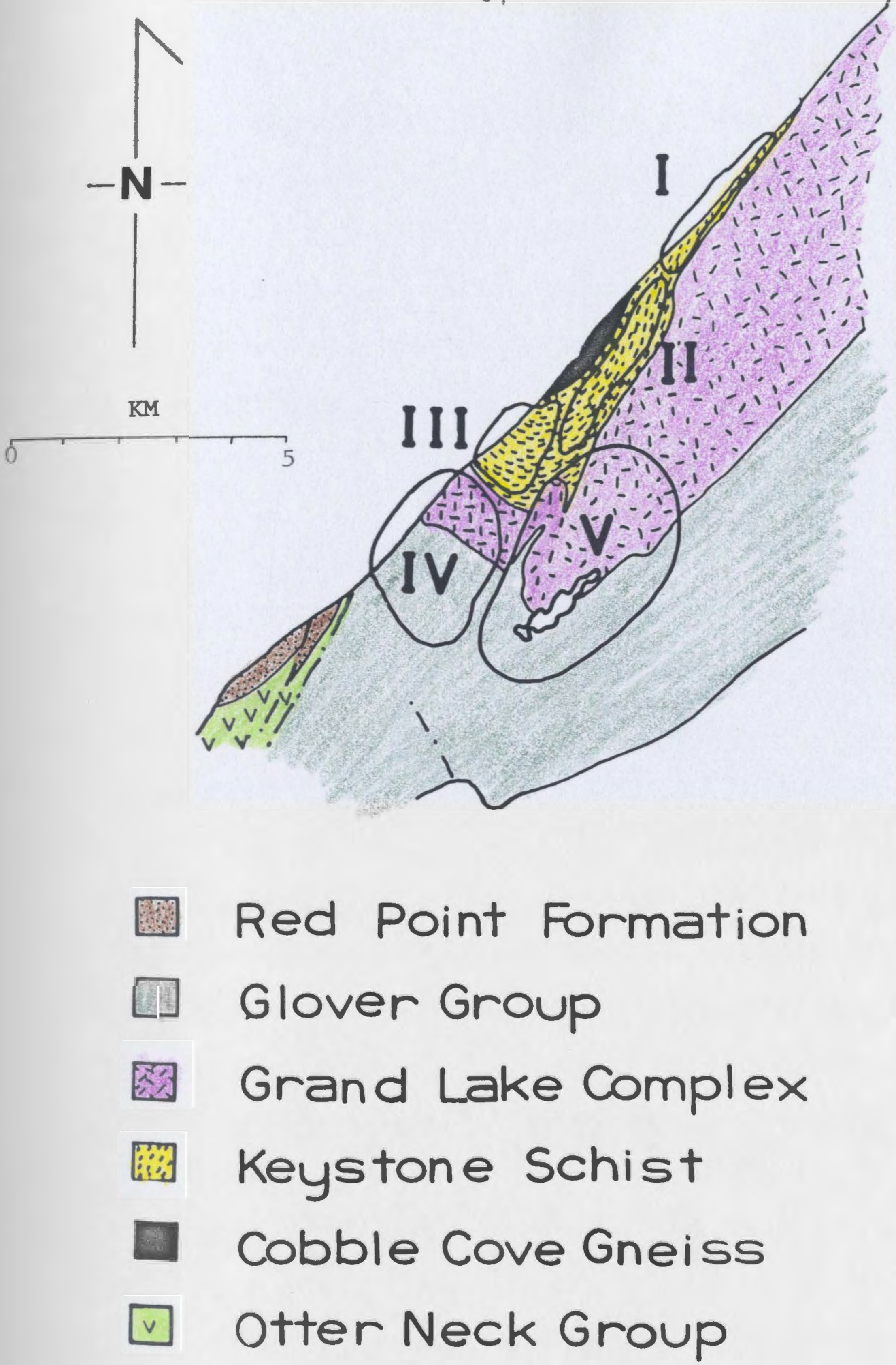
Geometric analysis is a method by which the orientation and distribution of different fabric elements are established for an area. The method is purely descriptive and seeks only to generate an internally consistent geometry of fabric elements.

A geometric analysis of the Keystone Schist is presented in this section to more fully characterize the structural sequence that has been presented. Methods used are largely those outlined by Turner and Weiss (1963). Where little structural data is available no geometric analysis is presented.

Subareas

For the purpose of this and subsequent geometric analyses the Glover Island Area has been divided into a number of subareas symmetrically located about the hinge line of the Glover Anticline to eliminate the effects of this folding event. These various subareas are shown in Figure 4.13 and are labelled I-V. Subareas I-III are located symmetrically about the Glover Anticline within the Keystone Schist. Subareas IV and V are also located symmetrically about the Glover Anticline within the Glover Group.

Figure 4.13 Structural subareas in the Glover Island Area used in the geometric analysis described in the text. Individual subareas are designated by Roman Numerals.


N
KM
0
5
I
II
III
IV
V
Red Point Formation
Glover Group
Grand Lake Complex
Keystone Schist
Cobble Cove Gneiss
Otter Neck Group

S1 Structures

The oldest recognized fabric element in the Keystone Schist consists of an S1 schistosity that has been largely transposed by S2. This transposition is so complete that the orientation of S1 is nearly impossible to determine in the field. As a result, no geometric analysis of D1 structures was attempted.

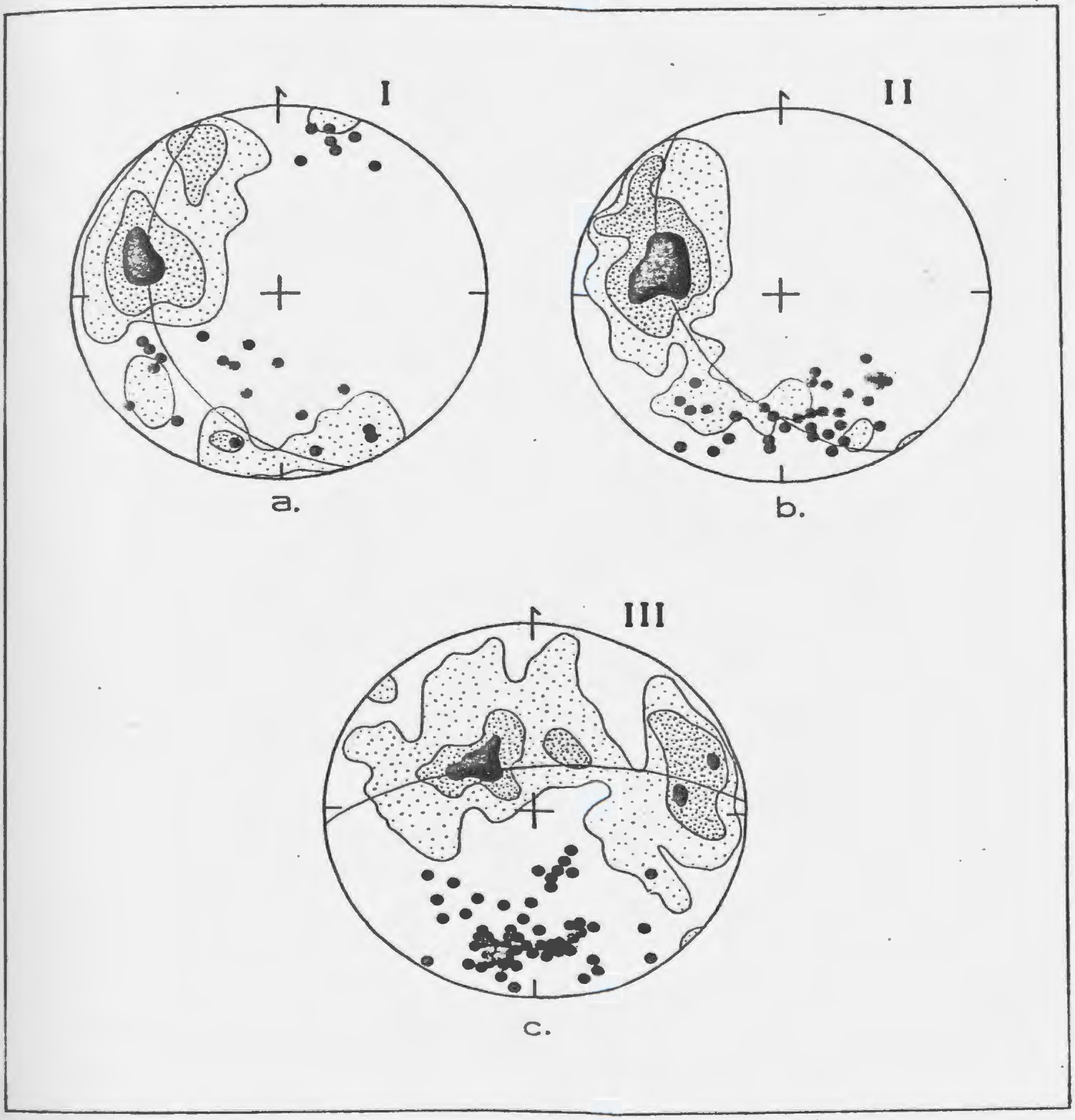
S2 Structures

Poles to S2 and the cogenetic pressure shadow lineation L2 are plotted in equal-area, lower hemisphere, stereographic projection in Figure 4.14 for the three subareas within the Keystone Schist. These subareas group D2 data from different parts of the Glover Anticline. In each of these subareas, poles to S2 form partial, great circle girdles and L2 forms a diffuse, south-plunging, single, point maximum.

The northernmost subareas, I and II, are located on the eastern limb of the Glover Anticline and are essentially removed from the effects of folding related to the Glover Anticline. In these subareas, poles to S2 form a single, great circle girdle with a single point maximum. This great circle girdle reflects extensive folding of S2 by F3 folds about north-south, upright axial planes.

The cogenetic L2 lineation forms a point maximum in subareas I and II. An older lineation folded by a younger folding event will describe a great circle or a small circle

Figure 4.14 Poles to S2 surfaces (contoured) and L2 lineations (solid dots) for subareas I, II, and III of the Keystone Schist plotted in equal area, lower hemisphere projections. Great circles are visually estimated, best-fit, great circles for poles to S2. Contour intervals in percent per one percent area, the number of points contoured and the number of lineations plotted are as follows: (a) .6,3.0,12.0; S2=166; L2=21; (b) .9,4.5,8.2; S2=110; L2=34; (c) .7,3.6,5.7; S2=139; L2=63.


I
II
III
a.
b.
c.

girdle unless folding is coaxial. However girdle development in subareas I and II is weak and L2 lineations are therefore not significantly rotated by D3.

In subarea III a pronounced change in morphology and orientation of poles to S2 is observed while L2 remains unaffected as before. Subarea III is located in the hinge area of the Glover Anticline and the great circle girdle described by poles to S2 is the result of D4 folding.

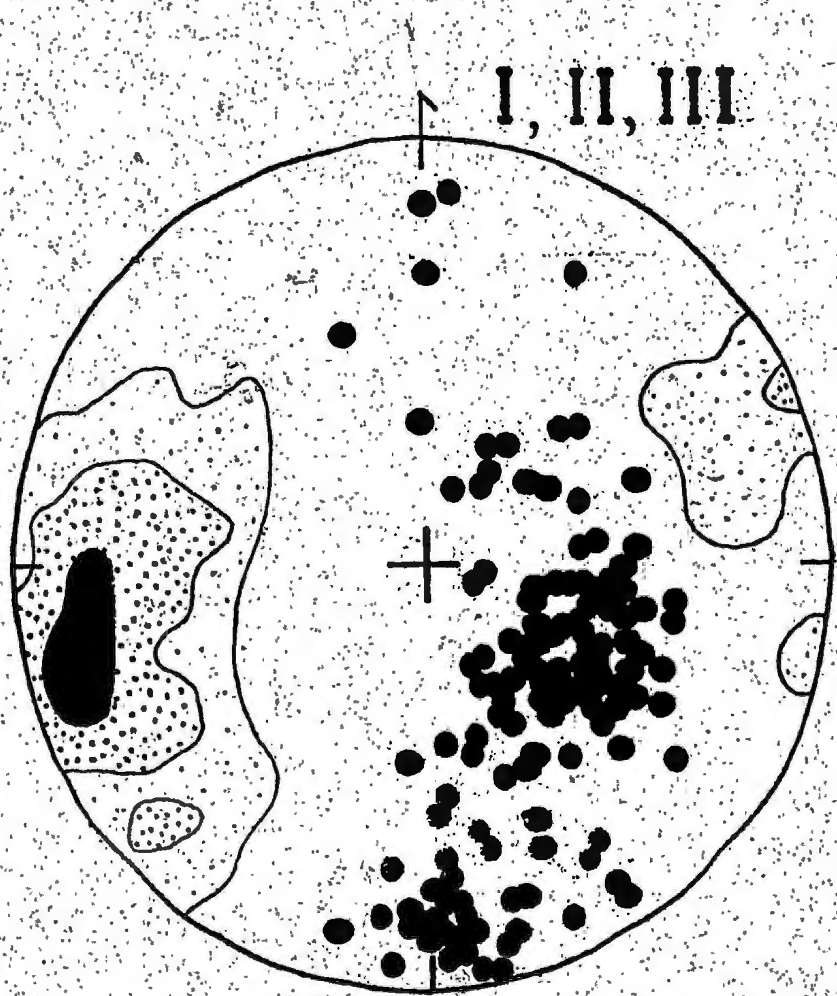
The diffuse nature of the great circle girdle defined by poles to S3 surfaces in subarea III is interpreted to be due to D4 folding of an S2 surface that has already been complicated by D3 folding.

L2 lineation data in subarea III forms a single point maximum indicating coaxial folding as in the case of L2 data in subareas I and II. Thus L2 is coaxial with respect to L3 and L4 and as a result is remarkably consistent throughout the study area.

D3 Structures

Poles to S3 surfaces, F3 fold axes and intersection lineations between S3 and S2 are plotted in equal-area, lower hemisphere projection in Figure 4.15 for subareas I, II and III. Poles to S3 surfaces form a single point maximum indicating that these rocks have not been significantly refolded after D3. The effects of D4 deformation are not apparent because very little D3 data

Figure 4.15 Poles to S3 surfaces (contoured) and L3 lineations (solid dots) for Subareas I, II, and III plotted in equal area, lower hemisphere projections.
Contour intervals in percent per one percent area, the number of points contoured and the number of lineations plotted are: .9,4.3,12.9; S3=116; L3=129.


I, II, III

was collected in the hinge area of the Glover Anticline in these three subareas. The data were taken predominantly from the east limb of the Glover Anticline in which D4 folding is essentially absent.

L3 intersection lineations from subareas I, II and III form a single, great circle girdle within the S3 cleavage plane as shown in Figure 3.15. This great circle distribution of L3 lineations cannot be the result of refolding as poles to S3 form a single point maximum and are therefore not folded. Two alternative possibilities exist that will produce this observed geometry. One is to refold an already folded surface and the other is to physically rotate these lineations towards the axis of maximum elongation of the finite strain ellipsoid by an increase in strain. Lineations form passive material lines and will rotate towards X with an increase in strain. At very high strains, such lineations will lie subparallel to X. It is not possible to determine which of these possibilities is correct in this situation but as F3 certainly does fold an already folded surface, the first possibility seems most logical.

METAMORPHISM OF THE KEYSTONE SCHIST

All recognized structural events in the Keystone Schist are associated with syntectonic metamorphic events. In addition, a major metamorphic event corresponding to the peak of metamorphism occurred during the first interkinematic period.

D1 Metamorphism

The S1 schistosity, defined by muscovite, chlorite and elongate prisms of epidote, formed during syntectonic metamorphism related to the first deformational event.

A gradient in temperature may have existed during this metamorphic event as coarser grain sizes occur at the base of the Keystone Schist compared to its top. However this may also reflect coarsening during later overprinting events.

Interkinematic metamorphism

Randomly-oriented, 1-2 mm porphyroblasts of biotite, garnet, microcline and epidote post-tectonically overgrow S1 and define an interkinematic metamorphic event which corresponds to the peak of metamorphism in the Keystone Schist.

Biotite is the most abundant porphyroblast and consists of dark green to olive green subidiomorphic laths which randomly overgrow S1 in the Lower Member of the Keystone Schist. An identical microstructure is observed at the base of the Upper Member but the biotite is a deep brown colour. Idiomorphic inclusions of epidote and dusty opaques occur within these biotite porphyroblasts and define straight inclusion trails that are continuous with S1 in the matrix of the rock.

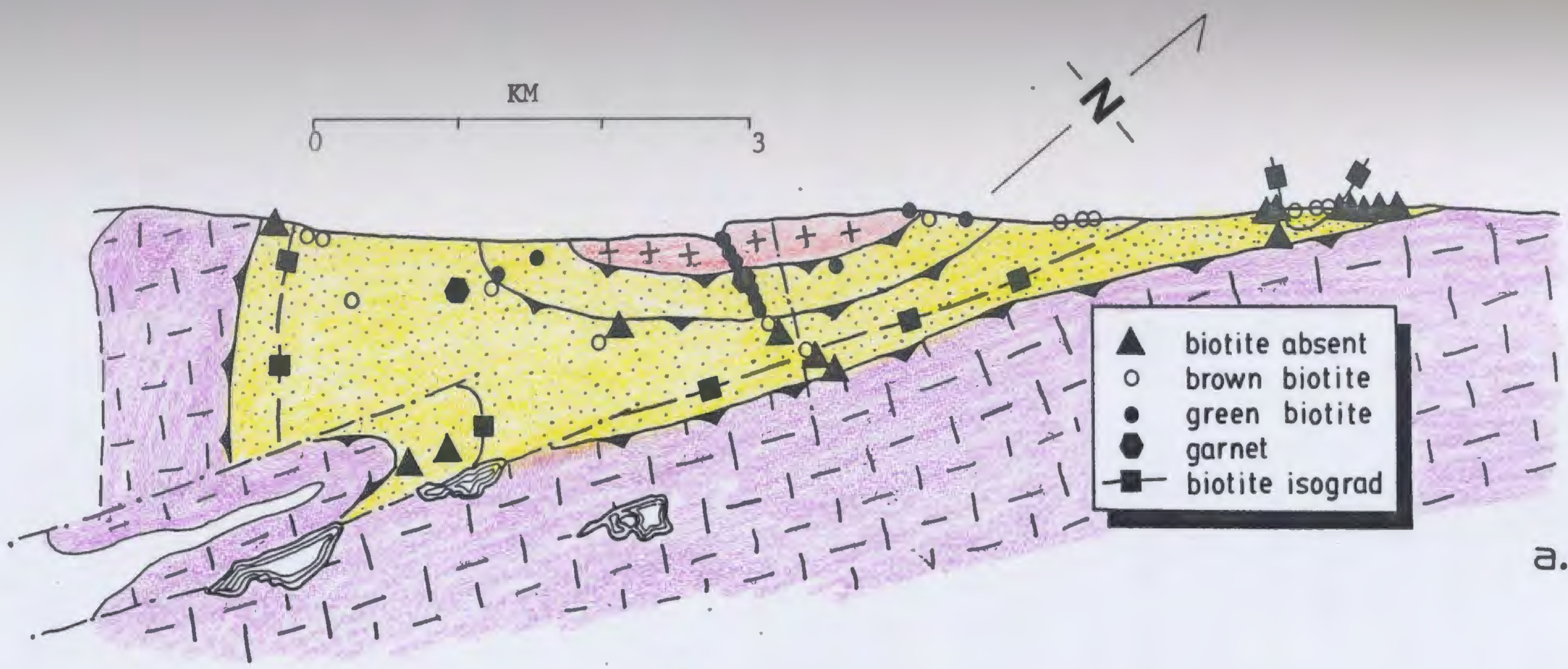
Equant, 1-3 mm, microcline porphyroblasts occur in the Lower Member and contain abundant quartz inclusions. These inclusions define straight inclusion trails in porphyroblast cores which abruptly curve into parallelism with S2 in porphyroblast rim areas (Fig. 4.7). S1 is transposed by S2 in these rocks.

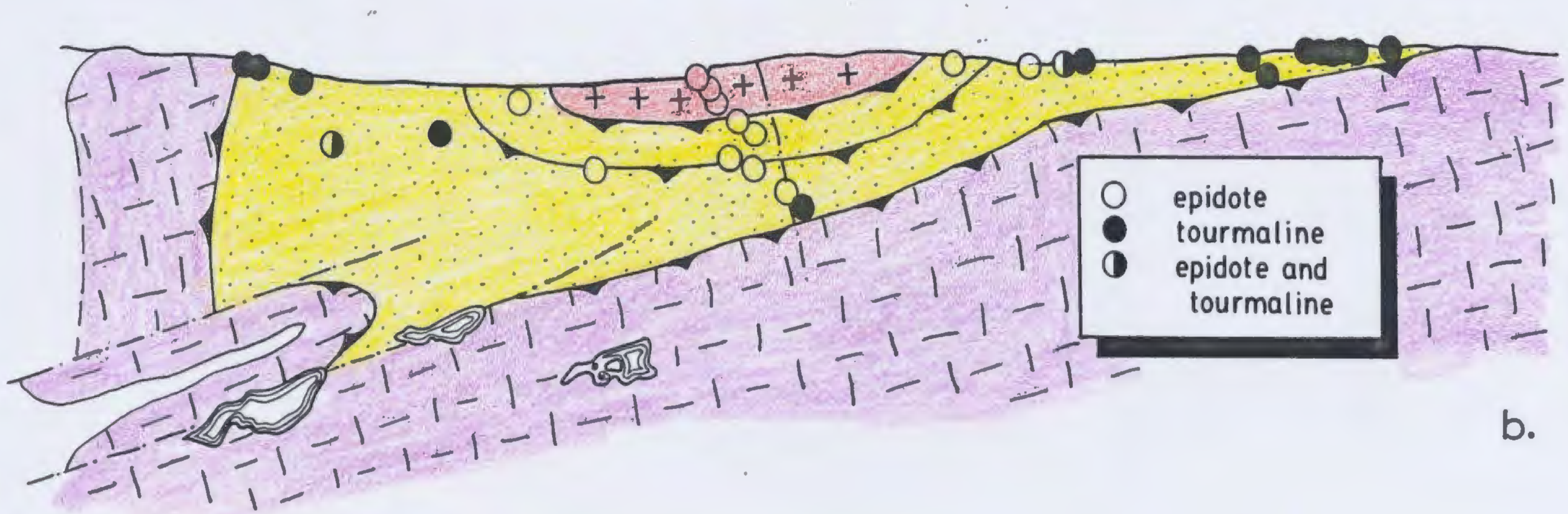
Garnet and epidote also occur as porphyroblasts in the Lower Member and locally contain straight inclusion trails. As in biotite and microcline porphyroblasts, these inclusion trails are continuous with S2 in the matrix of the rock where S1 has been transposed by S2.

The ubiquity of straight inclusion trails and the random orientation of porphyroblasts indicate that this metamorphic event was post-tectonic with respect to D1. However the abrupt rotation of inclusion trails into parallelism with S2 in porphyroblast rim areas indicates porphyroblast growth continued and overlapped with D2 deformation.

A strong gradient in metamorphic conditions existed during this interkinematic metamorphic event (Fig. 4.16). The growth of garnet and biotite in semipelitic rocks and hornblende and oligoclase in mafic rocks indicates the lower amphibolite facies was reached in the Lower Member. However a steady decrease in metamorphic grade occurs in the Upper Member from amphibolite to the lower greenschist facies. The biotite isograd occurs within the Upper Member. East of the biotite isograd, large clots of chlorite occur

Figure 4.16 Metamorphic map and distribution of metamorphic minerals in the Keystone Schist and Cobble Cove Gneiss. (a) Biotite isograd and occurrences of biotite and garnet. (b) Distribution of epidote and tourmaline. Ornamentation: crosses: Cobble Cove Gneiss; stippled pattern: Keystone Schist; dashed pattern: Grand Lake Complex.


KM
0
3
N
biotite absent
brown biotite
green biotite
garnet
biotite isograd
a.


epidote
tourmaline
epidote and tourmaline
b.

which are interpreted to be retrograded biotites. Farther east these chlorite clots disappear and no porphyroblasts are recognized. The disappearance of chlorite clots is interpreted to mark the original interkinematic biotite isograd. Retrograde metamorphism during D2 displaced the biotite isograd west to its observed position.

The progressive decrease in metamorphic grade upwards in the Keystone Schist is accompanied by a parallel decrease in grain size. Rocks at the top of the Upper Member are fine-grained, quartz-sericite schists transitional to phyllites. The base of the Keystone Schist contains coarse-grained, porphyroblastic schists. Intermediate rocks are transitional between these two extremes and support the conclusion that a strong gradient existed during this interkinematic metamorphism.

Microstructures in mafic rocks also suggest that this interkinematic metamorphic event involved potash metasomatism. Mafic rocks occur at the base and within the central part of the Lower Member. At the base of the Lower Member mafic rocks are in contact with the Cobble Cove Gneiss northeast of Cobble Cove. In the central part of the Lower Member they form a continuous, conformable layer of amphibolite folded by a major D2 fold. These two contrasting amphibolites are described below.

Mafic rocks at the base of the Lower Member consist primarily of plagioclase, actinolite, biotite, epidote, sphene and minor microcline. Biotite is green (X = light yellow green; Y = Z = dark olive green) and coexists with a nearly colourless, subidiomorphic actinolite (microprobe analysis). Idiomorphic, pleochroic pale yellow epidote is also present. Microcline occurs in these gneisses as well but is rare.

Actinolite is partially replaced by quartz yielding a pseudosymplectite texture in which rounded actinolite fragments occur in a quartz matrix. The quartz forms irregular patches partially controlled by the strong amphibole cleavage of the actinolite. This microstructure is interpreted to be a replacement texture.

Mafic rocks in the central part of the Lower Member are epidote-bearing amphibolites. They consist of hornblende, plagioclase and clinozoisite. Hornblende (microprobe analysis) is subidiomorphic and strongly pleochroic (Z = blue green; Y = light yellow green; X = light brownish yellow). It occurs together with xenomorphic oligoclase (An 21, microprobe analysis) which is invariably cluttered with numerous tiny, idiomorphic clinozoisite prisms. Hornblende is partially replaced by quartz yielding a pseudosymplectite texture as in the amphiboles described above.

These two amphibolites contrast strongly in mineralogy. Amphibole is dark green hornblende in amphibolites from within the middle of the Lower Member. In amphibolites at the base of the Lower Member, the amphibole is actinolite.

Biotite is abundant in this leucocratic amphibolite as well.

This contrast between these two rocks is attributed to potash metasomatism. The influx of potassic fluids during metamorphism resulted in the development of biotite at the expense of amphibole in the leucocratic amphibolites. The source of the potassic fluids is inferred to be the Cobble Cove Gneiss. Amphibolites from within the Keystone Schist were remote from this source and hence escaped similar metasomatic effects.

D2 Metamorphism

D2 syntectonic metamorphism resulted in the formation of a well developed, locally differentiated, S2 crenulation cleavage. Porphyroblasts that formed during the interkinematic period between D1 and D2 are deformed by S2 and are associated with elongate pressure shadows of quartz.

Microcline porphyroblast growth continued from pre-D2 to syn-D2 as evidenced by quartz inclusion trails (Fig. 4.7). Microcline porphyroblasts contain straight inclusion trails in core regions. Near rim areas these inclusion trails abruptly rotate into parallelism with S2 in the matrix of the rock. Rocks in which this microstructure is observed were collected from the hinge of a major F2 fold. In this hinge region, S1 has been transposed and now parallels S2. These porphyroblasts clearly record the interkinematic growth of microcline based upon straight inclusion trails in their cores. Rotation of S1 into parallelism with S2 during D2 is recorded by the abrupt rotation of inclusion trails into parallelism with S2 in porphyroblast rims. This microstructure indicates microcline growth continued through D2 and links these two metamorphic events.

Microstructural evidence also demonstrates that albite formed during D2. Graphitic, planar traces, highly crenulated by D2 deformation are locally entirely overgrown by helicitic albite. This recrystallization is inferred to have occurred immediately after D2 deformation.

Additional mineral growth during D2 is restricted to retrogression of mafic minerals. Biotite and garnet are both partly replaced by chlorite along S2 cleavages. This retrograde metamorphism is accordingly inferred to be of D2 age.

D3 and D4 Metamorphic events

Microstructures related to S3 and S4 crenulation cleavages are similar and discussed together. Each consists of a crenulation cleavage defined by tiny crenulations and microfolds of the preexisting schistosity. Minor muscovite, chlorite and quartz recrystallization occurred during these deformations and defines these syntectonic metamorphic events. Retrogression of earlier minerals such as biotite also occurred during D3 and D4. However these events are, in general, associated with little or no mineral growth and where mineral growth did occur it is restricted to cleavage planes.

Metamorphic conditions

The D1 metamorphic event is defined by muscovite-chlorite-quartz and therefore took place within the middle greenschist facies. The first interkinematic metamorphism locally reached the lower amphibolite facies in basal parts of the Keystone Schist based upon the assemblage: hornblende-oligoclase, but decreases upwards to middle greenschist assemblages of albite-chlorite-sericite at the top of the section. D2 metamorphism overlapped with the interkinematic metamorphism but only reached the upper greenschist facies, as biotite did not form during D2 metamorphism. Later events took place in the middle or lower greenschist facies based upon the recrystallization of quartz, chlorite and muscovite.

Thus the metamorphic history of the Keystone Schist involves primarily middle to lower greenschist facies conditions and spans four deformational events punctuated by a single, static interkinematic metamorphic event which locally reached lower amphibolite facies conditions.

CORRELATION AND AGE

The Keystone Schist bears a striking similarity to metaclastic rocks found west of Grand Lake that extend as far north as the Baie Verte Peninsula. Polymictic conglomerates, marble, amphibolite and semipelitic and psammitic schists have been described from the Fleur de Lys Supergroup of the Baie Verte Peninsula that closely resemble the Keystone Schist (e.g., Church, 1969; Kennedy, 1971; DeWit, 1974; ~~Bursnall and DeWit, 1975)~~. Rocks immediately west of Grand Lake consist of metaclastic rocks and marble that also resemble the Keystone Schist. Kennedy (1980) described a conglomerate horizon from west of Grand Lake comparable to conglomerates of the Keystone Schist. The structural and metamorphic history of these rocks has been complex and polyphase, and also comparable to that observed in the Keystone Schist. Based upon these many similarities, the Keystone Schist is correlated with metaclastic rocks west of Grand Lake.

Metaclastic rocks of western Newfoundland have not been directly dated as yet due to their unfossiliferous nature and high metamorphic grades. Rocks interpreted to be correlatives occur farther west however in lower units of the Humber Arm Supergroup (Stevens, 1970; Williams, 1973, 1975). These allochthonous rocks have escaped the high metamorphic conditions suffered by their more easterly equivalents and a few fossil localities have been discovered.

Coarse clastic rocks of the Summerside and Irishtown Formations occur at the base of the Humber Arm Supergroup. The Summerside Formation is undated but conglomerates containing fossiliferous Lower Cambrian limestone clasts do occur at the top of the Irishtown Formation. These coarse clastic rocks are overlain by interbedded shale, limestone and limestone breccias of the Cooks Brook Formation which ranges from Middle Cambrian to Lower Ordovician in age. The uppermost unit in the Humber Arm Supergroup is the Middle Arm Point Formation consisting of Middle Cambrian to Lower Ordovician shales. Based upon correlation of the Keystone Schist with coarse, clastic rocks at the base of the Humber Arm Supergroup, the Keystone Schist is inferred to be Lower Cambrian or somewhat older in age.

SIGNIFICANCE

The Keystone Schist consists of Lower Paleozoic metaclastic rocks that overlie a unit of Grenvillian basement gneiss, the Cobble Cove Gneiss. This basement-cover relationship, although structurally discontinuous, is well known in western Newfoundland where autochthonous Grenvillian basement of the North American continent is overlain by coarse metaclastic rocks. Based upon this regional relationship, the Keystone Schist is interpreted to represent part of this basal clastic unit. This interpretation further refines the inferred age of the Keystone Schist as Cambrian or Upper Precambrian.

Chapter 5

LOWER ORDOVICIAN OPHIOLITE SUITE:

The Grand Lake Complex

Mafic and ultramafic plutonic rocks that structurally overlie the Keystone Schist to the west and are nonconformably overlain by the Glover Group to the east are named the Grand Lake Complex. The Grand Lake Complex consists of a basal greenschist overlain by talc-carbonate schists, serpentinized peridotite, cumulate-layered gabbro, massive gabbro and an upper greenschist. Mafic dykes intrude all units of the complex while trondhjemite intrudes only the upper part of the complex.

The Grand Lake Complex is folded by the broad, open, south-plunging Glover Anticline. Both eastern and western limbs of this major fold intersect Grand Lake along the northwestern shoreline of Glover Island. The southern limit of the Grand Lake Complex occurs at the hinge of this fold which is located at Kettle Pond.

Massive mafic plutonic rocks of the Grand Lake Complex occur exclusively in the central part of the complex while greenschists containing large blocks of gneiss enclose these massive rocks. Traced to the south the thickness of these massive gneisses decreases until the basal greenschist and upper greenschist are nearly in contact.

The Grand Lake Complex is well exposed along the shoreline of Glover Island where the eastern and western fold limbs plunge beneath Grand Lake. It is also well exposed immediately south of Bluff Head and north of Kettle Pond where rugged, higher elevations occur. Intervening areas are mainly tree covered with the exception of Keystone Brook which exposes the basal part of the Grand Lake Complex.

Thickness of the Grand Lake Complex is approximately 1.5 km but it thins to 1 km in the hinge region and western limb of the Glover Anticline. As the base of the complex is a thrust fault and the top a nonconformity, this thickness is minimum.

STRATIGRAPHY

Keystone Brook section

In its type section at Keystone Brook the Grand Lake Complex consists of a basal greenschist overlain by talc-carbonate schists, serpentinized peridotite, layered gabbro and massive gabbro.

Abruptly overlying graphitic phyllites of the Keystone Schist in Keystone Brook is a 30 m thick unit of greenschist. It consists of alternating bright-green, purple and blue, millimetre-scale, fine-grained layers and metre-thick, fine-grained, massive greenschists. The pronounced layering exhibited by this rock is quite distinctive and not recognized elsewhere on Glover Island, although similar rocks are associated with ultramafic rocks at Little Sandy Point on the west side of Grand Lake (Plate I).

Talc-carbonate schists abruptly overlie the basal greenschist. These schists are orange or tan weathering and are strongly schistose although more massive rocks occur locally. The dominant cleavage is an anastomosing, phacoidal cleavage which overprints an earlier schistosity. Black lenses of serpentine, up to several metres in length locally occur within orange-weathering talc-carbonate rocks and represent layers that have been structurally disrupted by the late cleavage. Large boudins of serpentine are also locally present within these schists. The early schistosity wraps around these large boudins indicating they formed during or before this cleavage. Talc-carbonate schists are about 20 metres thick in Keystone Brook.

Talc-carbonate schists are abruptly juxtaposed against massive, serpentinized peridotites. This sharp contact is interpreted to mark the location of a small, high-angle fault along which part of the Grand Lake Complex is missing. From comparison with other sections, this missing part probably consists of serpentinite mélange.

Massive, serpentinized peridotites are well exposed for several hundred metres upstream of this contact. These white to orange weathering rocks locally contain abundant, 2-3 mm, relict clinopyroxenes which stand out prominently on weathered surfaces. Clinopyroxenes are locally concentrated into bands which define a diffuse gneissosity.

The serpentinized peridotites are overlain by mafic and leucocratic metagabbros. The transition is gradational as mafic gabbros occur first followed by more leucocratic gabbros but the transition is rather abrupt and occurs over a rather short distance (50 m). Outcrop exposure is poor at this point in Keystone Brook and rapidly ends shortly upstream such that much of the details of this transition are greatly obscured.

Metre-thick, east-facing, graded layers locally occur within the gabbros. However these rocks have been overprinted by a foliation and most graded contacts are extremely difficult to recognize. Much better graded layers occur along the shoreline of Grand Lake and at the top of Bluff Head.

Metabasalt dykes locally intrude metaperidotite and metagabbro in Keystone Brook. These dykes all strike to the north and are vertical or very steeply dipping. They range up to 1 m in thickness and have sharp chilled margins. Plagioclase phenocrysts are locally present and concentrated in the central part of the dykes.

<u>Shoreline section at Bluff Head</u>

The Grand Lake Complex forms an imposing headland over 1000 feet in relief at Bluff Head composed primarily of massive gabbro (Plate I).

Towards the southwest, patches of east-facing, layered gabbros occur within the complex. These consist of layers up to 1 m in thickness composed of clinopyroxene and hornblende at the base grading upward into leucocratic gabbro at the top. A regular sequence composed of fifteen to twenty graded layers can be observed at two localities along the shoreline of Grand Lake southwest of Bluff Head. These layers can be traced for about 15-20 m upwards before they are masked by fractures and lichen-cover.

Farther along the shoreline southwest of Bluff Head, ultramafic rocks are covered by talus derived from the overlying gabbros. A single outcrop extends above this talus slope and marks the base of the Grand Lake Complex along the shoreline. This outcrop consists of massive, orange-weathering, talc-carbonate schists.

<u>Halfway Brook Section</u>

Halfway Brook is a small stream located about halfway between Cobble Cove and Bluff Head which exposes a very important section through the base of the Grand Lake Complex.

Green phyllites of the Keystone Schist are well exposed at the mouth of Halfway Brook where it enters Grand Lake. Upstream, black phyllites and metasandstone beds outcrop and are abruptly overlain by orange-weathering, talc-carbonate schists. These rocks are generally strongly cleaved but are locally massive. Several 2 m thick fault zones separate talc-carbonate schists from a 20 m thick zone of serpentinite mélange.

Halfway Brook descends sharply through an open, meadow-like clearing halfway up the hillside of Glover Island. This clearing is underlain by scaly, finely comminuted, highly sheared, serpentinite schist containing blocks of massive serpentinized peridotite (Fig. 5.1). The sheared matrix material is poorly consolidated and can be scooped-out by hand. More resistant clasts of massive serpentinized peridotite stand out conspicuously on the hillside where this matrix material has been eroded away. The clasts are rounded, slickensided, and range in size from several centimetres up to over 4 metres in diameter. The exposure is typical of serpentinite mélange.

Farther up Halfway Brook, fine-grained serpentine matrix material disappears and the rock consists of a massive, serpentinized peridotite similar to that found in Keystone Brook. These rocks are overlain by massive gabbro that outcrops on top of Glover Island above Halfway Brook.

Area west of Kettle Pond

The area immediately west of Kettle Pond is structurally complex as rocks in this area are involved in a major D3 fold that folds the Grand Lake Complex into a gently south-plunging, upright syncline associated with a pronounced S3 crenulation cleavage which has overprinted bedding and earlier cleavages.

The eastern limb of the syncline consists of three separate lithologic units. The basal unit is an orange-weathering talc-carbonate schist with a strong S3 cleavage. It is overlain by a thin (10-50 m) unit of serpentine containing a strong S3 schistosity defined by tiny flakes of green mica (fuchsite?) and by the parallel alignment of serpentine minerals. No relict clinopyroxene grains are recognized in contrast to other ultramafic rocks recognized on Glover Island.

A strongly foliated greenschist containing lenticular, discontinuous pods of trondhjemite and gabbro overlies the serpentine and composes the upper greenschist unit.

The talc-carbonate rocks and thin unit of serpentine are considered to be equivalent to ultramafic rocks at the base of the complex east of Kettle Pond and upper green-schists are considered to be equivalent to greenschists at the top of the Grand Lake Complex east of Kettle Pond. The thick section of plutonic rocks that forms the bulk of the Grand Lake Complex east of Kettle Pond appears to be missing. The overall shape of the complex is thus one of a large lenticular pod, terminating at Kettle Pond, and enclosed in a sheath of highly strained greenschist and talc-carbonate rocks.

Shoreline section south of Cobble Cove

The Grand Lake Complex is discontinuously exposed along the shoreline of Glover Island south of Cobble Cove. Here graphitic pelites are abruptly overlain by magnesite-dolomite-green mica (fuchsite?) schists containing a well-developed S2 schistosity complexly folded by large scale, upright, F3 folds that trend subparallel to the shoreline. Metagabbros, greenschists and serpentinites are complexly interlayered with these carbonate rocks. A 100 m section of overburden containing large blocks of talc-carbonate schist separates these rocks from massive gabbro intimately associated with abundant trondhjemite and metabasalt.

A north-trending, high angle fault separates the gabbro from highly sheared greenschists containing large boudinaged pods (2 m) and 30 cm thick layers of orange-weathering, magnesite-dolomite marble which grade into similar, strongly-sheared greenstones containing abundant blocks and pods of highly deformed gabbro up to 2 m across. The foliation in the greenschists is an anastomosing, strongly curviplanar schistosity that sweeps smoothly around the more resistant blocks of gabbro. The upper contact of this greenschist consists of a quartzose greywacke of the overlying Glover Group.

The shoreline section described above consists of talc-carbonate rocks virtually in contact with upper greenschists. These two lithologies are separated by a thin unit of comingled gabbros, metabasalts and trondhjemites. As in the area west of Kettle Pond this section is comparable to the eastern section but the thick plutonic massif is missing and replaced by a cryptic thin zone of various intrusive rocks.

LITHOLOGY AND PETROLOGY

The Grand Lake Complex consists of a basal greenschist, talc-carbonate schist, ultramafic rocks, metagabbros, metabasalts and metatrondhjemites. These rocks are described below.

Basal greenschist

A thin (40 m) unit of greenschist occurs at the base of the Grand Lake Complex in Keystone Brook. It has not been recognized elsewhere and its absence may be due to structural omission.

The greenschist consists of bright green, purple and blue, millimetre-scale, fine-grained, schistose laminae and metre-thick, fine-grained, massive greenschists. Thinly laminated parts are conspicuous in the field due to purple and blue layers alternating with bright green layers.

Bright green layers of the basal greenschist consist primarily of chlorite and epidote. Purple and blue layers consist of plagioclase, carbonate and opaques. The opaques occur as fine, dust-like particles which give a distinctive bluish colour to these thin layers in hand specimen. Plagioclase grains range up to .2 mm in size and are equidimensional but not well rounded in shape. Carbonate generally occurs as a matrix to these plagioclase grains. Epidote is abundant as very small grains (.001 mm) concentrated mainly in chloritic layers. Tourmaline occurs as green, pleochroic, idiomorphic prisms but is minor in abundance.

<u>Talc-carbonate schists</u>

Talc-carbonate schists are a distinctive orange or tan colour derived from the weathering of iron-magnesium carbonate minerals. The rocks vary from massive to strongly schistose. In Keystone Brook talc-carbonate rocks contain a strong late fabric (D3) which forms an anastomosing, crenulation cleavage overprinting all earlier structures. In contrast, talc-carbonate schists that locally outcrop along the shoreline of Glover Island several kilometres southwest of Bluff Head are massive in appearance.

Talc-carbonate schists consist of talc, dolomite, magnesite, serpentine and opaques in varying proportions. In thin section, equidimensional carbonate grains up to 2 mm in diameter are set in a fine-grained, talc groundmass. Minor serpentine locally occurs within this talc groundmass and opaques occur as widely disseminated coarse grains

(.001 mm), as fine, dust-like grains rimming carbonate grains, and as inclusions within carbonate grains. The carbonate grains are generally equidimensional in shape and have xenomorphic shapes. Minor, small, idiomorphic carbonate rhombs locally occur in the groundmass, however. The larger carbonate grains are generally composed of a core region clouded with opaque inclusions and a rim region free of opaques, suggesting two distinct episodes of carbonate growth. Carbonate minerals consist of magnesite and dolomite, based upon microprobe analysis.

Ultramafic rocks

Ultramafic rocks consist of numerous, dark-green, clinopyroxene grains set in a fine-grained, orange-weathering serpentine matrix. Relict, igneous clinopyroxenes are generally 1-3 mm in diameter. Coarser clinopyroxenes occur up to 2 cm in diameter, but these may be dykes. Where clinopyroxene is rare or absent, the serpentinite is white weathering although all colour variations between white and orange occur. Eighty percent or more of the ultramafic rocks east of Kettle Pond contain relict clinopyroxene.

Clinopyroxene grains locally define a crude gneissosity in these serpentinites formed by alternating, diffuse layers of serpentine and clinopyroxene-rich serpentine (Fig. 5.2). No dimensional or lattice-preferred orientation of clinopyroxene is apparent.

Figure 5.1   Serpentinite mélange at the base of the Grand Lake Complex in Halfway Brook. Large, rounded block is 1 m in diameter and is embedded in a finely-comminuted, scaly matrix of serpentinite.

Figure 5.2   Weathered surface of clinopyroxene-serpentinite in Keystone Brook. Dark clots are relict, igneous clinopyroxene grains set in a groundmass of serpentine. Banding in rock defines S2.

Ultramafic rocks of the Grand Lake Complex consist of clinopyroxene, serpentine, tremolite and an opaque mineral inferred to be magnetite.

Clinopyroxene grains are equidimensional in shape and average about 2-5 mm in diameter although grains up to 3 cm in diameter occur locally. Clinopyroxenes have a 2V of 5^{o} and a $Z \wedge C$ of 39^{o}. They contain prominent but thin exsolution lamellae of orthopyroxene (?) parallel to (100) and numerous stacking faults parallel to (010) (Fig. 5.3). Selected analyses are given in Table 5.1 and are shown plotted in the pyroxene quadrilateral in Figure 5.4. These pyroxenes define a small compositional area centered at wo:en:fs = 48:7:45 and are generally diopsidic in composition.

Serpentine ranges from 5 to 10 modal percent in these ultramafic rocks and consists of fine-grained, decussate aggregates. The serpentine polymorph was identified as antigorite by X-ray diffraction (J. Vahtra, M.U.N.).

Tremolite is abundant in most of these ultramafic rocks. It consists of discrete subidiomorphic laths and epitaxial overgrowths on pyroxene. It is colourless and has a $Z \wedge C$ of about 15 degrees. Its composition is given in Table 5.2 as determined by electron microprobe analysis. It is tremolite in composition.

Table 5.1 Clinopyroxene analyses from cumulate clinopyroxenites of the Grand Lake Complex determined by microprobe analysis. Recalculation basis = 6 oxygens.

	1	2	3	4
SiO_2	52.95	52.45	52.46	51.89
TiO_2	.08	.18	.15	.11
Al_2O_3	1.16	1.19	1.80	2.20
FeO^T	4.01	4.11	3.95	4.44
MnO	.11	.12	.09	.15
MgO	15.55	15.76	15.86	15.72
Cao	25.67	25.23	25.32	23.74
Na_2O	.00	.01	.00	.04
K_2O	.00	.04	.00	.00
Tot	99.61	99.15	98.75	98.41
Si	1.957	1.948	1.922	1.932
Al	.050	.051	.079	.096
Ti	.002	.004	.003	.002
Fe^{++}	.124	.127	.123	.138
Mg	.857	.872	.882	.875
Mn	.003	.003	.002	.004
Ca	1.017	1.004	1.013	.950
Na	.000	.001	.000	.002
K	.000	.001	.000	.000
Tot	4.011	4.011	4.026	4.006

Figure 5.3 Photomicrograph of relict, igneous clinopyroxene from a clinopyroxene serpentinite showing prominent exsolution lamellae parallel to (001) and an (010) cleavage. X50.

Figure 5.4 Clinopyroxenes from clinopyroxenites and cumulate gabbros of the Grand Lake Complex plotted in the pyroxene quadrilateral. Compositions are determined by electron microprobe analyses. Ruled field is of cumulate clinopyroxenes from the Bay of Islands Complex after Malpas (1976). Solid lines represent the tholeiitic Skaergaard pyroxene trend (Wager and Brown, 1967) and the dashed line represents the mildly alkaline, Shiant Isles Sill trend from Gibb (1973).

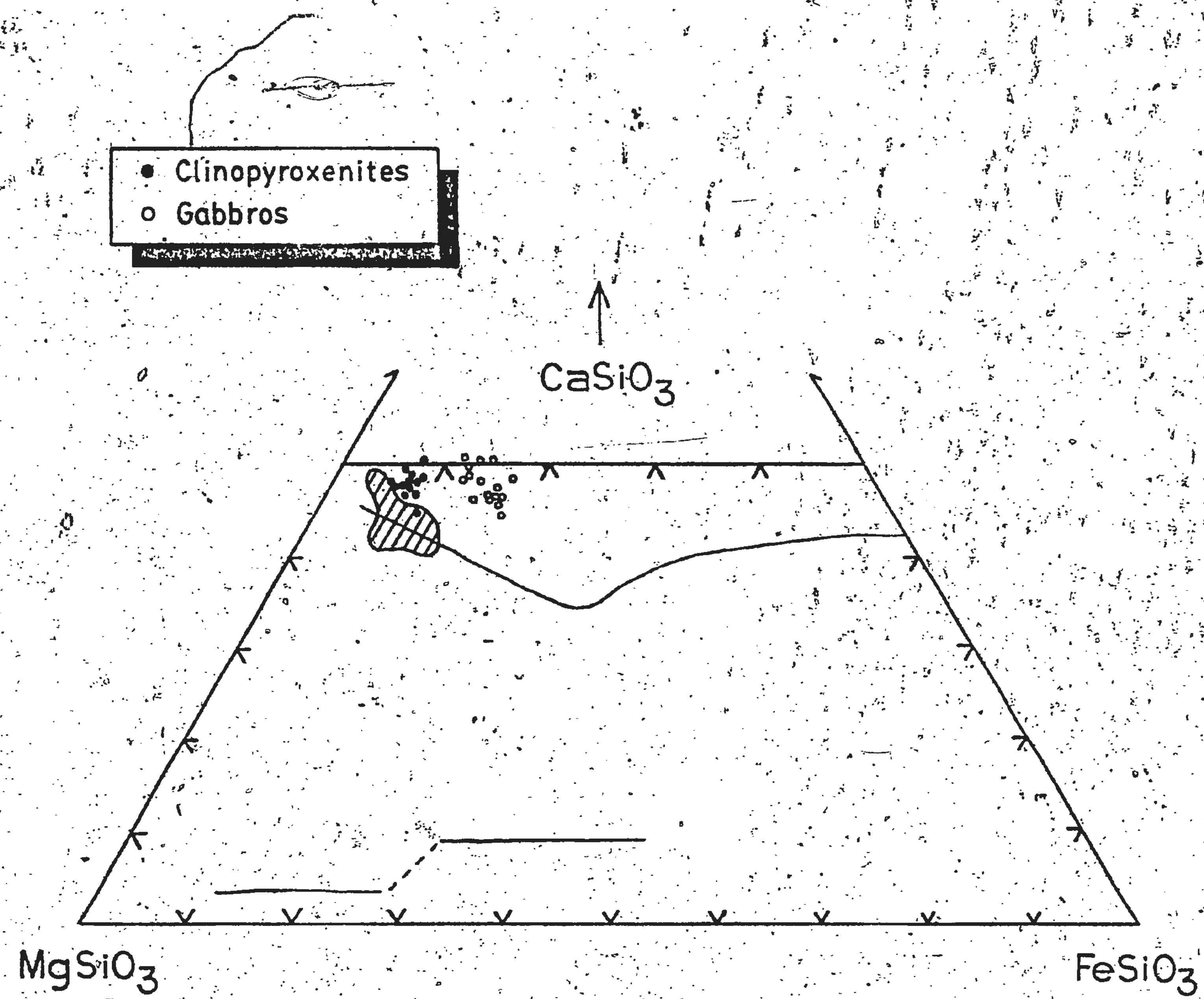

Clinopyroxenites
Gabbros
CaSiO3
MgSiO3
FeSiO3

Table 5.2 Tremolite analyses from clinopyroxenites of the Grand Lake Complex determined by microprobe analysis. Recalculation basis = 23 oxygens.

	1		2	
SiO_2	56.32		55.35	
TiO_2	.00		.13	
Al_2O_3	.73		2.66	
Cr_2O_3	.25		.27	
FeO^T	5.59		7.20	
MnO	.14		.14	
MgO	20.59		21.04	
CaO	12.43		11.81	
Na_2O	.05		.54	
K_2O	.01		.01	
Tot	96.16		99.21	
Si	7.935	8.00	7.635	8.000
Al	.065		.365	
Al	.055		.064	
Ti	.000		.012	
Cr	.024	5.077	.028	5.274
Fe^{++}	.657		.830	
Mg	4.325		4.324	
Mn	.016		.016	
Ca	1.873		1.745	
Na	.012	1.885	.142	1.887
K	.000		.000	
Tot	14.967		15.164	

Tremolite occurs primarily as a replacement of clinopyroxene. It generally forms along grain boundaries presumably due to nucleation difficulties and forms rational grain boundaries with the clinopyroxene. The result is a strong host control on tremolite orientation and a microstructure comparable to metallurgical Widmanstätten textures (e.g., Chadwick, 1974). Locally a different type of replacement occurs in which tremolite replaces clinopyroxene along a single lattice direction forming a uralitized clinopyroxene but these are minor in abundance.

Serpentine is locally associated with anhedral opaque minerals inferred to be magnetite as magnetite is commonly formed during the serpentinization process.

Two distinct textural types of ultramafic rocks are recognized in the Grand Lake Complex depending upon serpentine content. Where serpentine is minor or absent, equant clinopyroxene grains form a polygonal, granoblastic, monomineralic rock interpreted to be of cumulate origin. Wager et al. (1960) termed such rocks adcumulates and hypothesized they formed by the growth of unzoned cumulate crystals which gradually forced out all intercumulus liquid resulting in a rock composed entirely of interlocking cumulate crystals.

Aggregates of magnetite and serpentine locally occur in these rocks as small equant inclusions within clinopyroxene grains and as interstitial material. Serpentine-magnetite aggregates within clinopyroxenes are interpreted to represent metamorphosed inclusions of either olivine or orthopyroxene within clinopyroxenes. Interstitial serpentine-magnetite aggregates are molded about clinopyroxene grains as if they represented a metamorphosed intercumulus mineral or a poikilitic cumulus mineral such as orthopyroxene. However it is equally likely these aggregates represent metamorphosed olivine or orthopyroxene grains which have lost their original grain shapes during metamorphism and deformation and have deformed and flowed around the more rigid clinopyroxene grains to yield this microstructure. Grains trapped as inclusions within clinopyroxene grains retained their original grain shapes as they were protected by their host.

A second distinct microstructure present in ultramafic rocks of the Grand Lake Complex consists of 20 to 100 modal percent serpentine and equant, relict clinopyroxene grains. In these rocks, grains of clinopyroxene are entirely surrounded by a decussate, serpentine matrix.

These two microstructural types are clearly extreme cases of a continuous sequence. All intermediate combinations occur, locally within the same thin section. The origin of the serpentine is interpreted to be the same in each case. Orthopyroxene or olivine cumulate grains are interpreted to have formed a proportion of the modal mineralogy in these rocks. During subsequent metamorphism these minerals were converted to serpentine-magnetite aggregates which flowed around the more rigid clinopyroxene grains during accompanying deformation.

Metagabbro

The Grand Lake Complex gabbros consist of tan-weathering, medium-grained (1-4 mm), hornblende-clinopyroxene-clinozoisite-albite rocks. Mafic gabbros and graded, cumulate-layered gabbros occur in the central and lower parts of the Grand Lake Complex. In the upper parts of the Grand Lake Complex the gabbro is massive and in most places leucocratic. Locally, coarse (1-2 cm), hornblende-gabbro pegmatite occurs in small, irregular patches. Trondhjemite locally intrudes the gabbro in the upper part of the Grand Lake Complex. No trondhjemite is recognized in the central and lower parts of the Grand Lake Complex. Dykes and small stocks of basalt occur throughout the Grand Lake Complex gabbros.

At the base of the Grand Lake Complex gabbro section, graded layers consisting of clinopyroxenite at the base and gabbro at the top are abundant (Fig. 5.5). All layers with recognizable grading face to the east. These layers are locally obliterated by a strong gneissosity.

Mafic gabbros consist primarily of clinopyroxene, hornblende and clinozoisite with minor albite,

Figure 5.5 Cumulate gabbros in Keystone Brook structurally overprinted and intruded by a mafic dyke. Deformation has largely obscured facing directions at this locality.

sphene and ilmenite.

Clinopyroxene forms equant grains, 1-5 mm in diameter similar to the clinopyroxene found in ultramafic rocks. Exsolution lamellae of orthopyroxene (?) parallel to (100) are conspicuous and stacking faults parallel to (010) are recognized. In composition these clinopyroxenes are salites, similar in Ca to those in ultramafic rocks but more iron-rich based upon microprobe analysis (Table 5.3).

Plagioclase is ubiquitously pseudomorphed by aggregates of clinozoisite and minor albite. Clinozoisite was identified by X-ray diffraction (J. Vahtra, M.U.N.).

A conspicuous mineral within the metagabbros is a light green hornblende (X = medium olive green; Y = very light olive green; Z = dark olive green) with a 2V = 42 and $Z \wedge C = 15$. Its composition was determined by microprobe analysis. It is a ferroan pargasite based upon the classification of Leake (1980). Selected analyses are given in Table 5.4.

The primary shape of plagioclase grains is well preserved in these gabbros as they are pseudomorphed by aggregates of clinozoisite and minor albite. The clinozoisite consists of an aggregate of anhedral grains. In hand specimen it is white and the gabbro appears unaltered. In thin section these grains are found to consist almost entirely of clinozoisite rather than plagioclase.

Table 5.3 Clinopyroxene analyses from hornblende-clinopyroxene cumulate layers of the Grand Lake Complex determined by microprobe analysis. Recalculation basis = 6 oxygens.

	1	2	3
SiO_2	48.32	45.66	49.62
TiO_2	.94	.83	.66
Al_2O_3	5.31	4.74	4.20
FeO^T	8.58	8.32	8.03
MnO	.26	.22	.23
MgO	11.77	12.23	12.91
CaO	23.28	23.03	23.68
Na_2O	.23	.21	.14
K_2O	.01	.01	.01
Tot	98.82	95.31	99.48
Si	1.838	1.811	1.869
Al	.237	.221	.189
Ti	.026	.024	.018
Fe^{++}	.272	.275	.252
Mg	.667	.722	.724
Mn	.008	.007	.006
Ca	.949	.977	.955
Na	.017	.015	.010
K	.000	.000	.000
Tot	4.017	4.053	4.020

Table 5.4 Pargasite analyses from hornblende-clinopyroxene cumulate layers of the Grand Lake Complex determined by microprobe analysis. Recalculation basis = 23 oxygens.

SiO_2	41.41		41.04	
TiO_2	1.66		1.65	
Al_2O_3	13.33		13.45	
FeO^T	13.71		13.62	
MnO	.20		.16	
MgO	12.17		11.63	
CaO	12.20		12.36	
Na_2O	2.46		2.51	
K_2O	.07		.07	
Tot	97.21		96.52	
Si	6.175	8.000	6.171	8.000
Al	1.825		1.829	
Al	5.6		.555	
Ti	.184		.185	
Fe^{++}	1.707	5.133	1.712	5.079
Mg	2.705		2.609	
Mn	.021		.018	
Ca	1.947		1.991	
Na	.709	2.669	.728	2.728
K	.013		.009	
Tot	15.802		15.807	

Pseudomorphs of iron-titanium oxides are present throughout the gabbro. These occur as lamellae of ilmenite exsolved along octahedral planes of the primary iron oxide mineral phase which is altered to fine-grained granular sphene and albite. These occur as anhedral grains in the gabbros and also as inclusions within pargasite.

Metatrondhjemite

A small pluton of trondhjemite intrudes gabbros of the Grand Lake Complex 3 km northeast of Kettle Pond. Other small (less than 5 m) bodies of trondhjemite occur within gabbro and within the upper greenschist at the top of the Grand Lake Complex.

Trondhjemites are medium-grained (2-4 mm) and are white-weathering with quartz grains forming conspicuous, positive-relief, irregularities on weathered outcrop surfaces. On fresh surfaces the trondhjemite is light green.

Trondhjemite consists of plagioclase, quartz and chloritized mafic minerals. Mafic minerals are entirely replaced by chlorite which composes up to 10 percent of the rock. Quartz is abundant ranging between 30 and 50 model percent. The remainder of the rock consists of anhedral or subidiomorphic plagioclase. No alkali feldspar is present based upon petrography and staining with sodium cobaltinitrate.

Metabasalt

Mafic dykes up to 1 m in thickness are abundant in the Grand Lake Complex northeast of Kettle Pond. Dykes strike to the north and are steeply dipping. Most are aphyric but locally they contain about 5 modal percent plagioclase phenocrysts in their central part. Dyke margins are generally aphyric and chilled while the central part of the dyke grades into ophitic microgabbro.

Stocks of massive microgabbro intrude the Grand Lake Complex north of Kettle Pond. These massive black rocks range between microgabbro and gabbro in texture and locally contain euhedral igneous amphiboles intermediate between pargasite and kaersutite in composition.

Metabasalts of the Grand Lake Complex generally contain well-preserved relict, igneous minerals. Three petrographic types are recognized: a titanaugite basalt, a clinopyroxene basalt and a pargasite basalt. These metabasalts are described below.

Dykes sampled from Keystone Brook and Bluff Head (e.g., DK-208, 2082) contain relict igneous clinopyroxenes rimmed by actinolite. The clinopyroxenes are augitic in composition (wo:en:fs: = 42:43:15) based upon microprobe analysis and occur within interstices formed by laths of albitized plagioclase to define a relict, intergranular texture (Fig. 5.6). Pseudomorphs of fine-grained granular sphene after euhedral, iron-titanium oxides are abundant.

A single dyke of basalt from the bluff north of Kettle Pond (DK-2322) contrasts with these intergranular basalts as it consists of coarse (1 mm) subidiomorphic plates of clinopyroxene. It is light purple and faintly pleochroic suggesting it may be titanaugite.

A third, very distinct petrographic type that also occurs north of Kettle Pond (DK-232) consists of a ferroan pargasite microgabbro (Fig. 5.7). It consists of euhedral, deep-reddish-brown pleochroic pargasite in a matrix of anhedral albitized plagioclase. The pargasite is iron-rich and titaniferous and resembles kaersutite in composition, but contrasts with most kaersutites in a greater iron content (Table 5.5). Comparable ferroan pargasites are termed barkevikites by Deer et al. (1966), but this name is poorly defined (c.f., Leake, 1980). Green actinolite forms epitaxial overgrowths on these pargasites and is clearly of metamorphic origin.

Upper greenschist

Upper greenschists of the Grand Lake Complex consist of strongly foliated, green schistose rocks locally containing discontinuous pods of trondhjemite and tonalite up to several metres in size. The foliation wraps smoothly around these pods.

Figure 5.6 Photomicrograph of intergranular metabasalt dyke that intrudes gabbros of the Grand Lake Complex. Abbreviations: cpx, clinopyroxene; ab, albite with tiny inclusions of clinozoisite. X15.

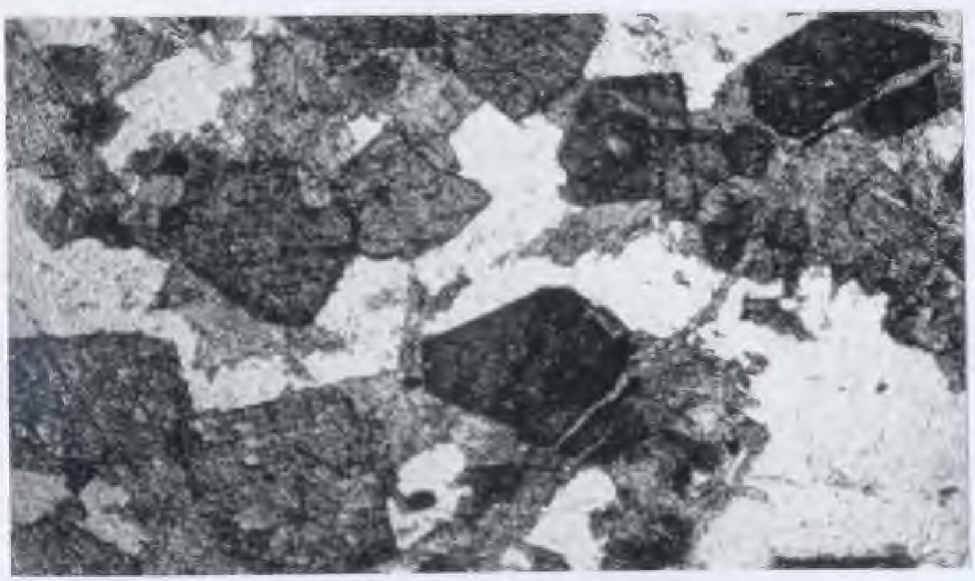


Figure 5.7 Photomicrograph of barkevikitic microgabbro that intrudes ultramafic rocks of the Grand Lake Complex just north of Kettle Pond. Abbreviations: bk, ferroan pargasite (barkevikite); ab, albite; act, actinolite. X15.

Table 5.5 Analyses of a ferroan pargasite (barkevikite) which occurs as a primary igneous phase in a microgabbro (DK-232) that intrudes ultramafic rocks of the Grand Lake Complex north of Kettle Pond. Recalculation basis = 23 oxygens.

	1	2	3
SiO_2	39.96	40.49	42.03
TiO_2	3.73	3.49	3.27
Al_2O_3	11.37	11.34	9.94
FeO^t	20.74	20.29	21.73
MnO	.30	.32	.38
MgO	7.74	7.88	7.32
CaO	11.21	11.34	10.63
Na_2O	3.62	3.41	3.20
K_2O	.24	.22	.25
Tot	98.98	98.83	98.73

Fe^{+3}	min	int	max	min	int	max	min	int	max
Si	6.105	6.100	6.095	6.170	6.167	6.164	6.417	6.405	6.392
Al^{IV}	1.895	1.900	1.905	1.830	1.833	1.836	1.583	1.595	1.608
Al_{VI}	.149	.142	.136	.204	.200	.196	.205	.190	.173
Ti	.426	.426	.425	.397	.397	.397	.372	.371	.371
Fe^{+3}	.000	.037	.074	.000	.023	.046	.000	.088	.176
Fe^{+2}	2.649	2.610	2.571	2.584	2.560	2.535	2.772	2.680	2.586
Mg	1.761	1.760	1.758	1.788	1.787	1.786	1.663	1.659	1.656
Mn	.036	.036	.036	.040	.040	.040	.044	.044	.044
Ca	1.833	1.831	1.830	1.851	1.850	1.849	1.738	1.735	1.735
Na	1.071	1.070	1.069	1.006	1.005	1.005	.945	.943	.943
K	.044	.044	.044	.040	.040	.040	.044	.044	.044

The upper greenschist consists of chlorite, epidote, plagioclase, quartz and calcite in varying proportions. The strong foliation is defined primarily by chlorite. Quartz is fine-grained due to deformation. No earlier relict minerals or textures are recognizable in this unit.

GEOCHEMISTRY

Eight rocks from the Grand Lake Complex were analyzed for major and trace elements. These included four metabasalts that intrude gabbroic rocks, two gabbros, a cumulate leucogabbro and a metaperidotite. These analyses are given in Table 5:6 and are discussed below. Methods and precision are discussed in the Appendix and the effects of alteration during metamorphism are discussed in Chapter 6. Sample locations are given in Plate II.

DK-2189

DK-2189 is a metaperidotite containing approximately 70 modal percent clinopyroxene and 30 percent serpentine. Cumulate textures are represented by the relict, igneous clinopyroxenes.

The effect of clinopyroxene accumulation is demonstrated by high calcium, magnesium and chromium. Nickel is rather low suggesting serpentine is a replacement of orthopyroxene rather than olivine. This observation can also be supported by major element modelling.

Table 5.6 Major and trace element analyses of rocks of the Grand Lake Complex.

	1	2	3	4	5	6	7	8
	2322	2082	208	232	283	206	2185	2189
SiO_2	45.8	46.8	47.9	47.9	48.6	46.2	40.6	48.3
TiO_2	1.65	2.20	2.79	2.21	.24	.26	.80	.16
Al_2O_{3T}	16.4	13.9	13.1	13.4	17.0	15.4	25.2	4.15
Fe_2O_3	10.43	12.88	14.77	14.27	4.93	11.36	8.56	7.86
MnO	.17	.21	.21	.24	.09	.20	.14	.14
MgO	8.13	6.74	5.85	4.83	9.84	9.68	3.44	20.00
CaO	10.59	10.23	7.85	9.58	14.31	12.19	14.28	16.58
Na_2O	3.18	2.56	3.67	4.22	2.52	1.63	1.43	.13
K_2O	.06	.04	.07	.18	.07	.01	.21	.01
P_2O_5	.19	.11	.25	.27	.31	.18	.17	.03
L.O.I.	2.99	2.58	2.20	1.36	2.67	2.81	3.43	2.96
Tot	99.59	98.25	98.64	98.38	100.58	99.92	98.26	100.32
Nb	2	0	8	9	0	0	0	0
Zr	133	166	247	303	8	0	0	0
Y	38	54	82	89	8	9	4	1
Sr	193	125	107	139	112	91	262	4
U	0	1	3	0	1	1	0	0
Rb	0	0	1	0	1	0	1	0
Th	1	4	7	3	2	2	2	2
Pb	0	3	4	6	2	2	0	6
Ga	15	13	13	17	9	7	17	0
Zn	59	75	100	103	18	48	66	23
Cu	15	5	16	0	64	0	0	86
Ni	64	39	33	57	141	40	0	216
Cr	217	130	19	51	882	11	0	2192
V	228	379	428	331	120	313	169	272
La	4	4	22	27	0	0	0	0
Ce	4	5	10	12	4	1	0	6
Ba	20	8	10	13	11	0	14	4

1 subophitic, titaniferous-augite-bearing metabasalt
2 intergranular augite-bearing metabasalt
3 intergranular augite-bearing metabasalt
4 ferroan, titaniferous pargasite-bearing microgabbro
5 cumulate layered metagabbro
6 cumulate layered metagabbro
7 cumulate layered meta-leucogabbro
8 metaperidotite

The composition of the clinopyroxene in this rock (determined by microprobe analysis) can be combined with different proportions of a hypothetical orthopyroxene (wo:en:fs = 4:84:12) and a hypothetical olivine (fo = 86) to duplicate the observed analysis. The resulting best fit proportions should indicate if the serpentine is a replacement of olivine or orthopyroxene.

A solution to this problem can be obtained by means of a linear least squares calculation in which the sum of the squared deviations between the calculated analysis and the observed analysis are minimized. The results of this calculation are given in Table 5.7 and it is clear that the sum of the squared deviations is minimized for a rock composed of 72 percent clinopyroxene, 24 percent orthopyroxene and 4 percent olivine. This suggests the protolith of the clinopyroxene-serpentinite was a websterite. However this calculation assumes isochemical conditions. Clearly the serpentinization process is not an isochemical process by virtue of the water which must be added to the system for the reaction to proceed. Yet, the agreement between this calculation and the observed abundances of nickel and chromium suggest it is valid.

DK-206, 283

DK-206 and 283 are coarse-grained, hornblende-clinozoisite metagabbros. The hornblende is an olive pargasite of

Table 5.7 Parameters used in least squares mixing calculation. By utilizing the method of least squares, the mineral compositions given in 1, 2 and 3 are manipulated to yield a single chemical analysis, 5, that most closely approximates the observed analysis 4. The resulting proportions are clinopyroxene:olivine:orthopyroxene = 72:4:24. This suggests that serpentine is a replacement of orthopyroxene rather than olivine assuming isochemical conditions.

	1	2	3	4	5
SiO_2	49.74	39.58	55.45	49.62	50.70
TiO_2	.07	.00	.00	.16	.05
Al_2O_3	3.24	.00	2.41	4.26	2.91
$Fe_2O_3^T$	6.93	14.73	8.76	8.08	7.68
MnO	.16	.00	.00	.14	.12
MgO	16.26	45.69	31.25	20.55	21.03
CaO	23.40	.00	2.12	17.03	17.36
Na_2O	.19	.00	.00	.13	.14

1 Clinopyroxene composition (Wo:En:Fs = 46:44:10)

2 Olivine composition (Fo = 86)

3 Orthopyroxene composition (Wo:En:Fs = 4:84:12)

4 Chemical analysis of clinopyroxene serpentinite recalculated to 100 percent anhydrous (serpentine: clinopyroxene = 70:30)

5 Calculated chemical analysis (Cpx:Ol:Opx = 72:4:24)

metamorphic origin. Aggregates of clinozoisite pseudomorph plagioclase grains.

Both gabbros are similar in composition except for higher iron and vanadium in DK-206 perhaps reflecting the accumulation of an iron oxide mineral. Both have rather high calcium and magnesium values around 10.

DK-2185

This rock is a leucogabbro from the top of a cumulate gabbro. It consists primarily of clinozoisite and albite after plagioclase.

Its plagioclase-rich nature is demonstrated by high calcium, aluminium and strontium. The low silica content suggests a calcic plagioclase. Recalculation of this analysis to mole proportion and normalizing such that aluminium and silica total four, yields 1.09 calcium and .2 sodium. This is very close to the stoichiometry of anorthite supporting the above observation and suggesting the protolith of this rock was an anorthositic gabbro.

DK-2082, 208

DK-2082 and 208 are intergranular metabasalts containing equant grains of augite bounded by laths of albitized plagioclase. Composition of the augite is wo:en:fs = 42:43:15 based upon microprobe analysis. Augites are extensively overgrown and partially replaced by green, pleochroic hornblende.

These two basalts are strongly differentiated as both contain high iron, titanium, and vanadium and low aluminium and magnesium. In contrast, DK-208 is strongly depleted in chromium and enriched in the light rare earth elements.

DK-2322

DK-2322 contains subophitic, purplish augites that may be titaniferous, overgrown by actinolite and chlorite. Plagioclase is pseudomorphed by albite and iron-titanium oxides by granular sphene.

DK-2322 is a typical basalt in composition and contrasts with DK-208 and DK-2082 in that it has lower iron, titanium and vanadium and higher aluminium, chromium and nickel. It may represent a less fractionated composition cogenetic with these other basalts.

DK-232

DK-232 is a ferroan, titaniferous pargasite-bearing microgabbro. Pargasites are euhedral to subhedral and primary in origin. They are rimmed by epitaxial overgrowths of green, pleochroic actinolite. Pargasite grains are set in a matrix of equant, albitized plagioclase grains.

DK-232 is highly enriched in the incompatible trace elements, titanium, iron and vanadium, and strongly depleted in aluminium and magnesium. Nickel and chromium values are about equal and light rare earth elements are high. This basalt may represent a strongly differentiated rock as the major and trace element abundances are consistent with typical differentiation trends.

Graphical representation

Basaltic rocks of the Grand Lake Complex are plotted in various standard diagrams in Figures 5.8 and 5.9. A complete discussion of the rationale behind the use of these diagrams is given in the next chapter.

$Zr-P_2O_5$ Diagram (Fig. 5.8)

All analyses from the Grand Lake Complex plot within the sub-alkaline field in this diagram. Clearly no alkaline rocks are present in the Grand Lake Complex.

SiO_2-FeO^T/MgO Diagram (Fig. 5.9)

Basalts show a range of iron/magnesium values in this diagram at roughly the same silica content. A trend away from the calc-alkaline field boundary to strongly tholeiitic compositions is apparent. It is not known if these basalts are all related however.

Ni-Cr Diagram (Fig. 5.9)

Two compositional groups are observed in the nickel/chromium diagram with DK-2082 and DK-2322 clustering

Figure 5.8 P_2O_5 - Zr diagram for rocks of the Grand Lake Complex. Alkaline field boundary is from Winchester and Floyd (1976). Numbered data points refer to Table 5.6

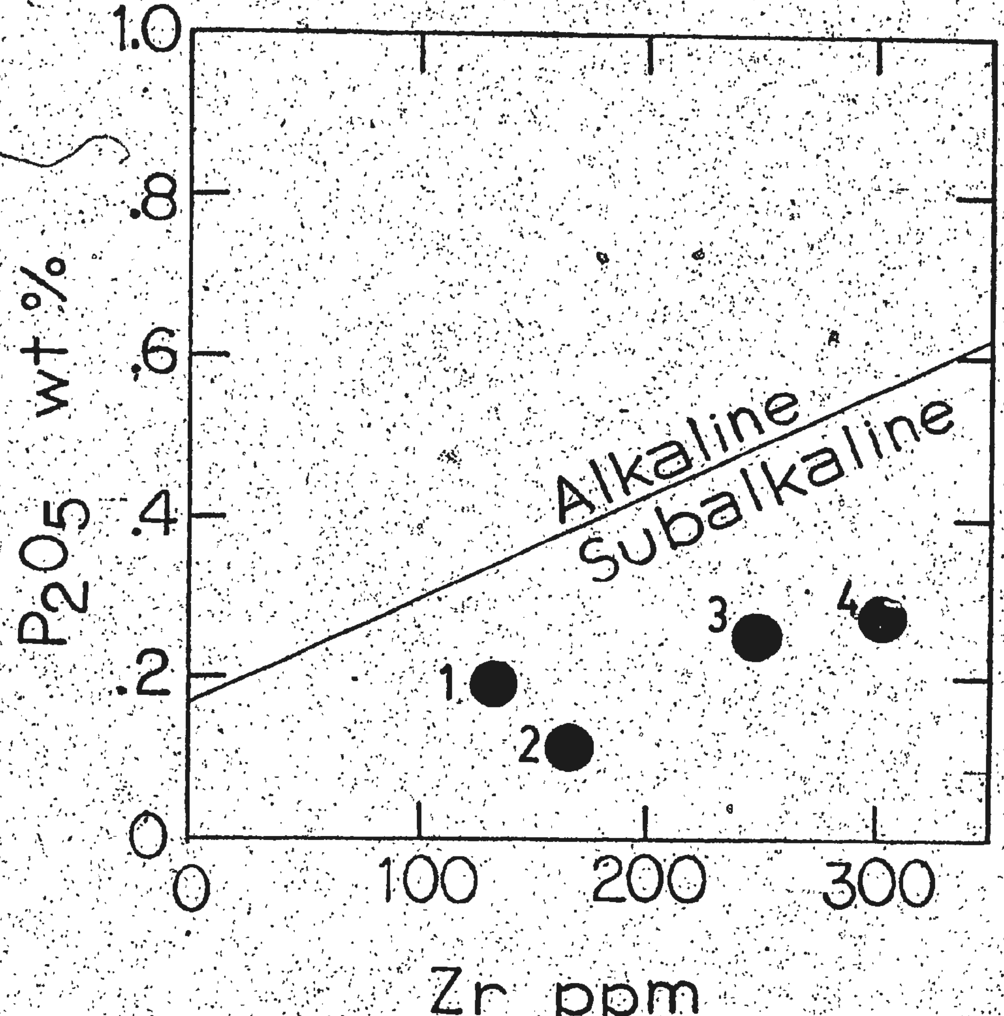
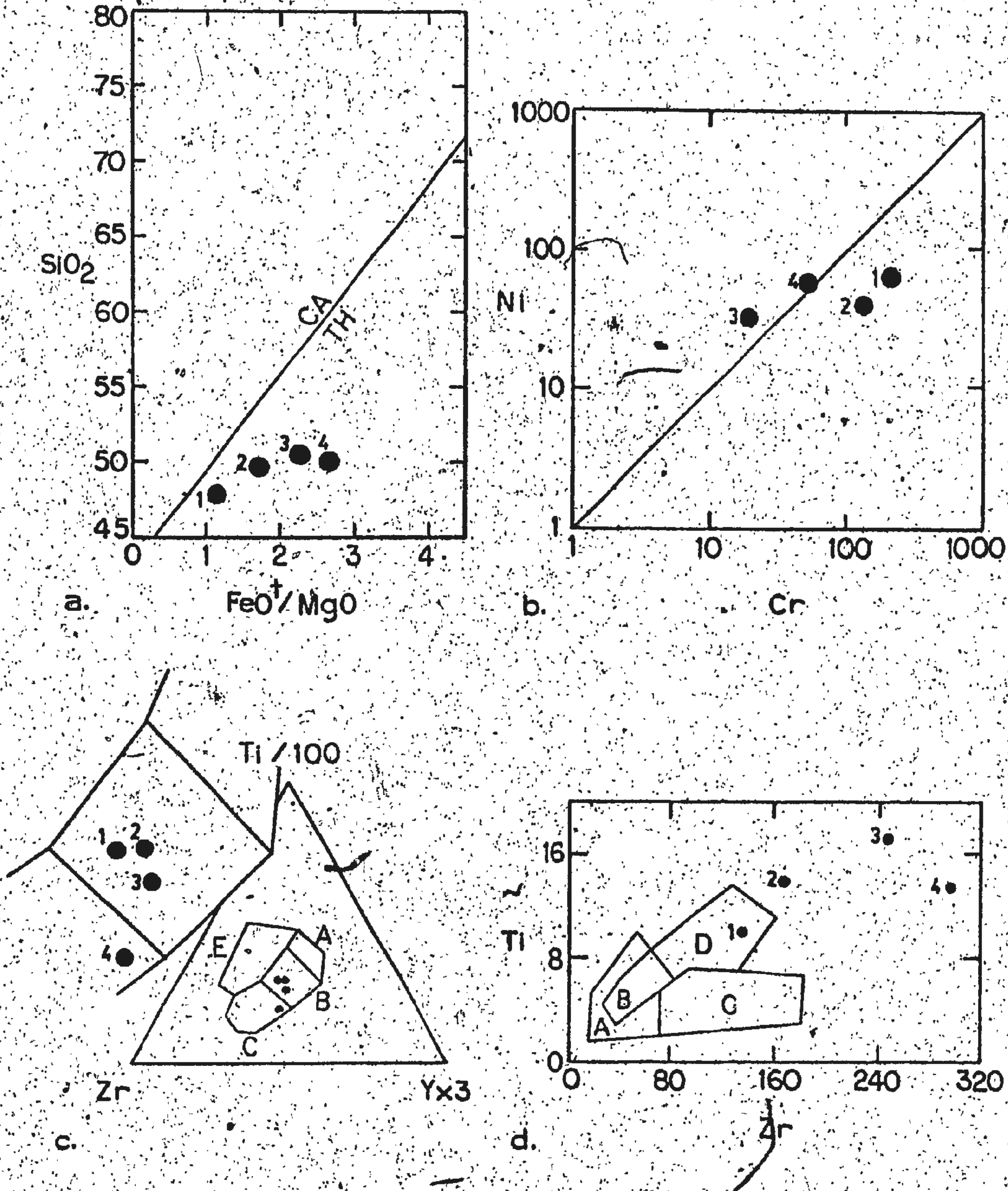

1.0
.8
.6
.4
.2
0
0
100
200
300
P2O5 wt %
Zr ppm
Alkaline
Subalkaline
1
2
3
4

Figure 5.9. Metabasalts of the Grand Lake Complex plotted in various standard diagrams. Numbered points refer to analyses given in Table 5.6. Figure a is after Miyashiro (1974). Figures c and d are after Pearce and Cann (1971, 1973).


80
75
70
65
60
55
50
45
SiO2
CA
TH
0
1
2
3
4
FeO†/MgO
a.
1000
100
10
1
Ni
1
10
100
1000
Cr
b.
Ti / 100
E
A
B
C
Zr
Yx3
c.
16
8
0
Ti
D
C
B
A
0
80
160
240
320
Zr
d.

together at high chromium, low nickel values.

Ti/100-Yx3-Zr Diagram (Fig. 5.9).

A similar relationship is observed in this diagram with DK-2082 and DK-2322 plotting close to one another while DK-208 and DK-232 plot at lower titanium values.

All basalts plot in the field of overlap in this diagram except DK-232 which plots in the calc-alkaline basalt field.

Zr-Ri Diagram (Fig. 5.9)

DK-2322 plots within the ocean floor basalt field in this diagram. However all the other basalt analyses fall outside of the field of normal basalts reflecting their highly differentiated nature.

Analysis of graphical presentation

DK-2322 and DK-2082 plot close to one another in all of the diagrams described above. On this basis, it is likely that they are related to one another.

DK-232 and DK-208 are rather unusual in that they are highly differentiated and enriched in the incompatible trace elements, iron and vanadium. DK-232 plots away from DK-208 in diagrams involving titanium. In a highly differentiated rock such as this it is possible that titanium was behaving as a compatible element thereby introducing the scatter into this diagram.

Based upon extreme iron enrichment and iron/magnesium ratio, DK-232 and DK-202 are tholeiites. It is unlikely they represent ocean floor basalts however, as primary pargasite is rare in such a setting.

DK-2322 and DK-2082 contrast with these other basalts as they are less iron enriched and plot closer to the calc-alkaline field in the silica-iron/magnesium diagram. However DK-2082 is enriched in iron suggesting a tholeiitic origin. An ocean floor origin for these basalts is not unreasonable.

Thus basaltic rocks of the Grand Lake Complex can be divided into two groups based upon major and trace element geochemistry. One group consists of iron-enriched tholeiites locally containing primary ferroan pargasite. This group is enriched in the light rare earths and depleted in chromium. The second group is less iron-enriched, has high chromium values and low abundances of rare earth elements.

Based upon the limited data available, it is concluded that two groups of basalts occur within the Grand Lake Complex. However it cannot be established that these groups are genetically separate. It is possible they simply represent a single magma which has been extensively fractionated. The two groups defined above may simply represent this single magma captured at contrasting fractionation stages.

STRUCTURE AND METAMORPHISM

Structures related to four deformational events are recognized in the Grand Lake Complex. A hornblende foliation formed in mafic gabbros during D1 and clinopyroxene was replaced by hornblende. An antigorite foliation formed during D2 in ultramafic rocks. These were metasomatically altered to carbonate and talc during the second interkinematic period at the base of the complex. Talc-carbonate schists were foliated and a pressure shadow lineation formed during D3. D4 was associated with the formation of many upright minor folds and a crenulation cleavage. These structures are described below. No geometric analysis is presented as the quantity of structural data collected does not warrant it.

D1 Structures

A weak dimensional preferred orientation of pargasite in mafic gabbro or hornblendite layers defines S1. The individual grains of pargasite that compose these nearly monomineralic rocks are unusually coarse (7 mm) and display a well-developed equilibrium metamorphic microstructure involving straight grain boundaries and abundant 120 degree triple junctions. Locally, irregular, anhedral fragments of salitic clinopyroxene (microprobe analysis) occur as inclusions in interstitial grains in these amphibole layers. These salites are clearly out of microstructural equilibrium and are interpreted to represent relict igneous clinopyroxenes.

Based upon this microstructure D1 is inferred to have been associated with a high temperature syntectonic metamorphic event. The widespread development of pargasite indicates that temperatures reached amphibolite facies conditions.

Clinopyroxenes of the Grand Lake Complex are plotted in the pyroxene quadrilateral in Figure 5.4. Compared with other pyroxenes from tholeiitic and alkaline rocks, these clinopyroxenes are unusually calcic. This compositional difference is attributed to re-equilibration during metamorphism. The Grand Lake Complex pyroxenes are thus interpreted to represent relict igneous pyroxenes that have re-equilibrated to diopsidic compositions during D1 syntectonic metamorphism.

D2 Structures

S2 is defined by diffuse, alternating antigorite and clinopyroxene layers (Fig. 5.2). The foliation is best observed in hand specimen or on weathered outcrop surfaces. Antigorite generally forms a fine-grained decussate aggregate surrounding grains of relict, igneous clinopyroxene. Epitaxial overgrowths of diopside on clinopyroxene are locally recognized.

Based upon the occurrence of antigorite and diopside, syntectonic D2 metamorphism is inferred to have reached upper greenschist facies conditions.

Interkinematic metasomatism

During the second interkinematic period, antigorite schists at the base of the complex were metasomatized to carbonate-talc assemblages. The resulting schists are massive with decussate talc replacing antigorite.

D3 Structures

Structures related to D3 deformation consist of a foliation and a cogenetic pressure-shadow lineation. These structures are recognized in talc-carbonate schists at the base of the complex where they are overprinted and locally transposed by D4 structures.

The most conspicuous structure developed during D3 consists of a heterogeneously developed foliation in talc-carbonate rocks defined primarily by talc. Foliation development is very heterogeneous. Locally talc-carbonate rocks are massive. Elsewhere these rocks contain a well developed foliation defined by fine-grained talc and coarse talc pressure shadows formed around carbonate grains. Heterogeneity of fabric development is so extensive that these two extremes may be observed in the same thin section.

A strong pressure shadow lineation also formed during D3. This lineation is observable only in thin section where it is very conspicuous on sections cut parallel to the lineation. Relict magnesite grains are clouded with fine-grained opaque material and epitaxially overgrown by optically distinctive rims of inclusion-free carbonate. Overgrowths are restricted to a single direction contained within the S2 schistosity parallel to the lineation.

Pressure shadows of coarse-grained talc also define this lineation as they are coarse-grained in comparison with groundmass talc and are oriented with their basal cleavage at a high angle to the carbonate grain-matrix interface. Talc pressure shadows are strictly parallel with the epitaxial carbonate overgrowths and clearly relate to the same tectonic-metamorphic episode.

Syntectonic metamorphism occurred within the greenschist facies based upon the mineral assemblage magnesite-talc.

D4 Structures

D4 structures consist of numerous upright open folds that affect the margins of the Grand Lake Complex and are associated with a heterogeneously developed crenulation cleavage.

Syntectonic metamorphism is inferred to have reached medial or lower greenschist facies conditions.

INTERPRETATION

The Grand Lake Complex is interpreted to represent a structurally isolated and metamorphosed layered basic pluton. The overall sequence observed in the Grand Lake Complex consists of: ultramafic rocks at the base rich in clinopyroxene and perhaps orthopyroxene, succeeded by graded, cumulate layers and massive leucocratic gabbro. This is a familiar sequence in layered basic intrusions supporting this interpretation.

Clinopyroxene-bearing, cumulate-layered sequences comparable to the protolith of the Grand Lake Complex are a conspicuous component of the ophiolite suite (e.g., Coleman, 1977). The idealized ophiolite suite is composed of a basal metamorphic tectonite composed primarily of harzburgite that is overlain by cumulate rocks. The base of the cumulate section generally consists of cumulate dunite and minor chromite layers which grade upward into wherlite (olivine-clinopyroxene rocks). Olivine gradually decreases in abundance upward in the cumulate sequence and plagioclase becomes a cumulate phase forming troctolites or clinopyroxene gabbros depending upon the abundance of olivine. Massive leucocratic gabbros and minor trondhjemite generally compose the uppermost part of the cumulate section. The zone separating ultramafic cumulates and massive gabbro is referred to as the transitional or critical zone as peridotite and gabbro are interlayered throughout this interval.

Individual ophiolites commonly display considerable variation in the thickness and sequence of different cumulate layers. The Vourinos Ophiolite Complex of northern Greece has one of the most extensive cumulate sequences recognized to date (Jackson et al., 1975; Moores, 1969). This 1500 m thick cumulate section contains numerous, discrete, cyclic units each of which represents part of a single cumulate sequence.

A similar sequence of cumulate rocks occurs in the Bay of Islands Complex (Malpas, 1976). The basal cumulate section of the Bay of Islands Complex consists of dunite with minor spinel and clinopyroxenite. It is overlain by a transition zone in which plagioclase, clinopyroxene and olivine are present in varying proportions. These rocks are overlain by a thick sequence of massive gabbro. Orthopyroxene is rare in the cumulate section as in the Vourinos Ophiolite Complex where orthopyroxene occurs only locally in the plagioclase cumulates.

The Betts Cove Ophiolite (Kidd et al., 1978) contains a well-exposed ultramafic cumulate section composed primarily of harzburgite overlain by a thin transition zone. The transition zone is in turn overlain by a thin zone of gabbro and a well-developed sheeted dyke complex. The ultramafic cumulate sequence in the Betts Cove Ophiolite contrasts with the Bay of Islands and Vourinos Ophiolite complexes in that it contains a significant proportion of orthopyroxene demonstrating the variety of cumulate minerals found in ophiolite cumulate sequences.

Based upon comparison with these well documented ophiolites, the Grand Lake Complex is interpreted to represent a fragment of disrupted ophiolite. The lower, ultramafic portion of the Grand Lake Complex which consists of metamorphosed cumulate websterite and clinopyroxenite is interpreted to represent part of an ophiolite ultramafic cumulate sequence. Overlying layered gabbros are interpreted to represent the transition zone overlain by massive gabbro and minor trondhjemite. The top of the massive gabbro section and the base of the ultramafic cumulate section are both absent due to structural omission or erosion.

Many on-land ophiolite complexes contain a dynamothermal aureole at their base composed of pyroxene and garnet amphibolites, amphibolites and greenschists derived from a volcanic protolith (e.g., Williams and Smyth, 1973; Jameison, 1980; Clague et al., 1981). This aureole is inferred to have formed at the base of the ophiolite complex during initial obduction and then to have been transported as an integral part of the ophiolite to its final position.

Greenschists at the base of the Grand Lake Complex in Keystone Brook occupy the same structural position in which aureole rocks occur in other ophiolitic complexes. They contrast sharply with underlying graphitic pelites and semipelites as they are of probable volcaniclastic origin. Later metamorphism has retrograded any earlier mineral assemblages so that, unfortunately, no high temperature relict mineral phases are preserved. The fine, millimetre-scale laminae present in this greenschist may represent a relict metamorphic layering formed during aureole formation. Based upon these similarities the basal greenschist of the Grand Lake Complex is interpreted to represent a sliver of retrograded metamorphic aureole.

AGE

Many of the ophiolite complexes present in western Newfoundland, or their metamorphic aureoles, have been dated radiometrically. All yield Lower Ordovician ages (Mattinson, 1975, 1976; Dallmeyer and Williams, 1975; Archibald and Farrar, 1976; Dallmeyer, 1977). The Grand Lake Complex is interpreted to have a similar Lower Ordovician age.

SIGNIFICANCE

Rocks interpreted to be ophiolitic outcrop along the structural boundary separating schists and gneisses from volcanic rocks on the Baie Verte Peninsula of northwest Newfoundland known as the Baie Verte - Brompton Line

(Williams and St. Julien, 1978; Kidd, 1977; Williams et al., 1977). The Grand Lake Complex is sited along a similar structural interface separating schists and gneisses from volcanic rocks. The interpretation of the Grand Lake Complex as ophiolite therefore reinforces the interpretation of this structural surface as the Baie Verte-Brompton Line and documents its continuity from Baie Verte to Glover Island.

Chapter 6

LOWER ORDOVICIAN VOLCANIC ROCKS: The Glover Group

The Glover Formation was first defined by Riley (1957) to include virtually all volcanics and metasediments east of the Cabot Fault except the Keystone Schist and Cobble Cove Gneiss. As this includes a variety of contrasting lithologies the Glover Formation is redefined to exclude the Red Point Formation, Otter Neck Group and Corner Pond Formation and is elevated to Group status. Two formations are defined in the Glover Group: the Kettle Pond Formation at the base and the Tuckamore Formation which composes the rest of the Group. The Glover Group overlies the Grand Lake Complex along a nonconformity and is unconformably overlain by the Corner Pond Formation.

THE KETTLE POND FORMATION

The Kettle Pond Formation outcrops along the shoreline of Glover Island south of Cobble Cove (Plate I). It extends inland from this point to where it is folded by the Glover Anticline together with the Grand Lake Complex. From the hinge area of this major fold, the Kettle Pond Formation continues north to within about 2 km of Bluff Head where it ends. Its truncation is interpreted as stratigraphic overlap rather than structural.

High strains and complex folding characterize the Kettle Pond Formation throughout its extent and make an accurate estimate of thickness impossible. Outcrop width of the steeply dipping Kettle Pond Formation ranges between 300 and 600 metres.

Lithology of the Kettle Pond Formation

The Kettle Pond Formation is well exposed at its type locality along the southeastern shoreline of Kettle Pond. Here it consists of a fine-grained (.001 mm), thinly-laminated (1-2 mm), light tan to brown-weathering, phyllitic or schistose, arenaceous metasediment containing sparse clasts of basalt, gabbro and trondhjemite. The laminated appearance of this fine-grained rock is quite distinctive in the field. More micaceous layers generally weather-out such that the more psammitic layers form millimetre-scale ridges. Chevron folds are conspicuous in these rocks reflecting the strong control this lamination has on the deformational style. It is not known to what degree deformation has contributed to the formation of these laminae.

A clast-supported polymictic conglomerate containing clasts up to 20 cm in diameter is also exposed at the type locality along the southeastern shoreline of Kettle Pond. The matrix of this conglomerate consists of a green, volcaniclastic metasediment. Similar conglomerates have not been recognized elsewhere in the Kettle Pond Formation. Clast composition is the same as that for the rest of the formation.

Polymictic conglomerates that occur at the very base of the Kettle Pond Formation overlie greenschists and ultramafic rocks of the Grand Lake Complex. The actual contact is not exposed. They are overlain by thinly laminated, arenaceous schists that compose the remaining part of the Kettle Pond Formation. The contact between these two units is gradational as observed in discontinuous but closely spaced outcrops on the hillside along the southeast shoreline of Kettle Pond. The conglomerate is interpreted to represent a coarser facies of the Kettle Pond Formation.

Petrology of the Kettle Pond Formation

A sample of thinly laminated schist from the Kettle Pond Formation containing a 4 cm long flattened jasper clast consists of alternating, 2-3 mm, quartz-plagioclase and chlorite-sericite layers. Quartz-plagioclase layers are extremely fine-grained and are composed almost exclusively of extremely fine-grained, irregular grains of quartz. Angular, coarser grains of plagioclase are evenly scattered throughout these siliceous laminae together with variable amounts of sericite and chlorite.

Arenaceous laminae are separated by chlorite-sericite layers consisting of decussate aggregates of chlorite intimately associated with wispy flakes of sericite.

A large (3 mm) aggregate of polygonized, 0.5 mm quartz grains with straight grain boundaries, 120 degree triple junctions, and minor undulose extinction is interpreted to represent a detrital grain of quartz that has undergone deformation and annealing recrystallization. Pressure shadows of quartz and calcite form tails parallel to the schistosity on either side of this detrital quartz grain.

Lithology and petrology of clasts of the Kettle Pond Formation

Sparse, rounded clasts up to 10 cm in diameter occur throughout the Kettle Pond Formation (Fig. 6.1). As these clasts are of key importance in interpretation of the Kettle Pond Formation, they are discussed in detail below.

A total of 33 rounded clasts were collected from the Kettle Pond Formation during routine field work. Care was taken to ensure that these were clasts and not boudinaged beds. Only those clasts which were clearly detrital were collected. Undoubtedly this has biased the sample towards lithologies which tend to remain undeformed during metamorphism, such as gabbros and trondhjemites. Clasts of argillite, phyllite and schist would deform along with the matrix such that they would be unrecognizable. Offsetting this bias, however, is the fact that during field work all well-exposed outcrops of the Kettle Pond Formation were closely inspected for any clasts of schist or gneiss. None were found in spite of this intense scrutiny suggesting their scarcity is not entirely due to structural obliteration or sampling bias. Quartz clasts are abundant in the Kettle Pond Formation but were not collected as they are not particularly informative.

All clasts collected from the Kettle Pond Formation are igneous in origin except for clasts of vein quartz and a single clast of orthoquartzite (Table 6.1). Igneous clasts consist of gabbro, leucogabbro, basalt, trondhjemite, myrmekitic trondhjemite, porphyritic microtrondhjemite, rhyodacite and rhyolite. These various clast types are described below.

Trondhjemite clasts are abundant at the base of the Kettle Pond Formation along the southeast shoreline of Kettle Pond and for several kilometres to the north. These well-rounded clasts range up to 10 cm but are generally less than 5 cm in diameter. They are quite distinctive in the field as they weather white and are coarser-grained than the matrix material. Medium grained (1-4 mm) trondhjemite, myrmekitic trondhjemite and porphyritic microtrondhjemite are recognized.

a.

b.

Figure 6.1 Clasts within the Kettle Pond Formation of the Glover Group. (a) Rounded trondhjemite clasts. (b) Flattened chloritic clasts.

Table 6.1 Clast types recognized in the Kettle Pond Formation.

Clast Type	Number
Medium-grained trondhjemite	8
Myrmekitic trondhjemite	5
Porphyritic microtrondhjemite	3
Aphyric Rhyolite	1
Quartz-plagioclase phyric rhyolite	3
Hematized rhyolite	1
Quartz-epidote rock	2
Basalt	5
Gabbro	2
Leucogabbro	2
Orthoquartzite	1
TOTAL:	33

Trondhjemite clasts consist of medium-grained (1 mm) quartz and plagioclase in an allotriomorphic-granular texture. Alkali feldspar is absent based upon petrography and staining (Sodium Cobaltinitrate). Mafic minerals are also absent. Small amounts of calcite, mica and epidote occur due to sericitization.

Myrmekitic trondhjemite clasts are identical to trondhjemite clasts except for the conspicuous occurrence of vermicular quartz in plagioclase (myrmekite). These rocks are composed of medium-grained (1 mm), quartz and plagioclase, and are free of mafic minerals. Quartz ranges between 30 and 40 modal percent.

Porphyritic microtrondhjemite forms a third, distinct, clast type. These rocks consist of plagioclase-quartz phyric, micro-trondhjemite and leucocratic microdiorite. The groundmass consists of fine-grained laths of plagioclase and interstitial quartz. With a slight decrease in groundmass grain size these rocks would grade into dacites and andesites, or with a slight coarsening of the groundmass, into trondhjemite and leucocratic diorite. Mafic minerals are limited to minor eipdote and opaques.

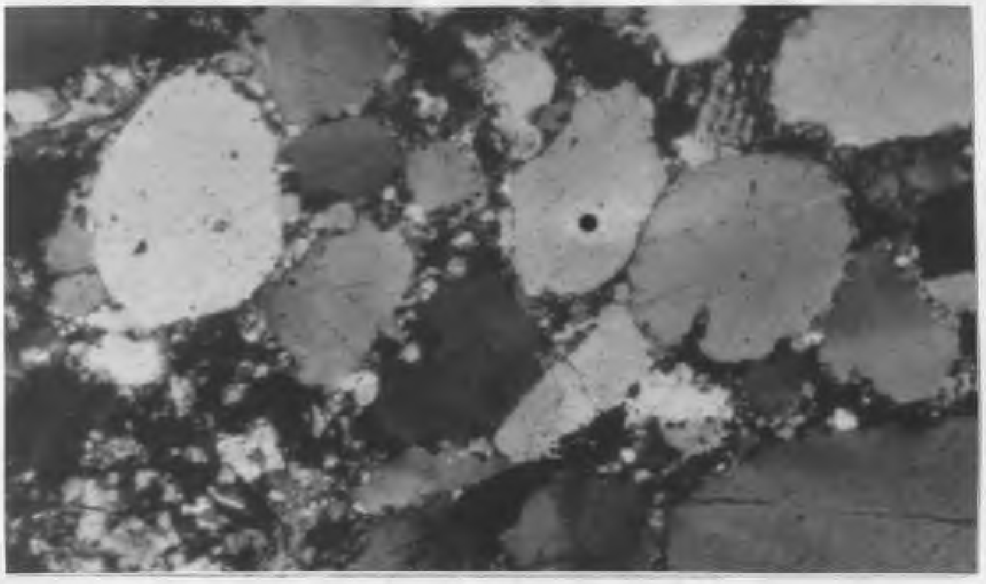
Silicic volcanic clasts are the most abundant clast type in the Kettle Pond Formation aside from trondhjemitic clasts. These silicic volcanic clasts include quartz-plagioclase phyric rhyolite, aphyric rhyolite, hematized, rhyolitic, lapilli tuff, jasper and quartz-epidote rocks.

All are interpreted to represent silicic volcanic rocks in varying states of alteration and metamorphism.

Mafic clasts of the Kettle Pond Formation consist of metabasalt and metagabbro. Metabasalt clasts are the more abundant of these two types and consist of actinolite, chlorite, epidote and albite. Except for a general idea of grain size, primary igneous textures are absent.

On the other hand, primary igneous textures are well preserved in gabbroic rocks as the plagioclase in these rocks is invariably replaced by epidote and clinozoisite. The epidote or clinozoisite preserves the original outline of the plagioclase during metamorphism. Of the four gabbroic clasts recovered from the Kettle Pond Formation, two are leucogabbros and are quite coarse in grain size (5-10 mm) reminiscent of anorthositic leucogabbros from ophiolite cumulate sequences. These distinctive clast types are illustrated in Figure 6.2.

The only clast recovered from the Kettle Pond Formation that is not igneous in origin is a single clast of mature orthoquartzite shown in Figure 6.2. It consists of a well-sorted, well-rounded, fine-grained orthoquartzite. A well-developed mortar texture has formed in response to deformation and metamorphism.



a.

b.

Figure 6.2 Clast types present within the Kettle Pond Formation of the Glover Group. (a) Photomicrograph of a mortar-textured quartzite clast; X25. (b) Coarse-grained gabbro and leucogabbro clasts drawn to scale.

THE TUCKAMORE FORMATION

Volcanic rocks overlying the Kettle Pond Formation are named the Tuckamore Formation (Plate I). Type area is designated as the south end of Kettle Pond.

No accurate estimate of thickness can be made for this unit due to structural complexity largely in the form of high-angle faults. A minimum thickness of 3-4 km seems likely based upon local continuous sections and regional extent.

Lithology and petrology of the Tuckamore Formation

Much of the Tuckamore Formation consists of greenschists and massive greenstone. However relict igneous textures and volcanic structures are locally well preserved and in most cases allow the protolith to be identified.

The Tuckamore Formation consists of a heterogeneous sequence of metamorphosed mafic tuffs, pillow lavas, silicic tuffs, volcaniclastic sediments and coarse volcanic breccias. All units are intruded by cogenetic sills of basalt and microgabbro and dykes or stocks of coarse gabbro and tonalite. These lithologies occur repeatedly within the section and form lenticular, discontinuous map units further complicated by abundant high-angle faults.

The base of the Tuckamore Formation consists of quartz-sericite schists and greenschists. At the north end of Glover Island these are overlain by pillow basalts intruded by massive, cogenetic sills. At the south end of Glover Island quartz-sericite schists are overlain by well-bedded sequences of greenschist, minor quartz-sericite schists and lenticular units of pillow basalt. Along the eastern side of the Otter Neck Group are well-layered greenschists and pillow basalt. Locally along the south shore of Glover Island these greenschists are interrupted by a coarse, polymictic, volcanic breccia containing distinctive jasper clasts. The very southern tip of the island consists of well-bedded greenschists, pillow basalts and minor quartz-sericite schists.

Rocks south and east of Grand Lake have not been mapped. However Red Indian Brook exposes over 6 km of homogeneous pillow basalts, and rocks east of the Corner Pond Formation consist of homogeneous pillow basalt as well. The eastern shoreline of Grand Lake consists of pillow lava, massive sills and amphibolite hornfels. These different lithologies are described below.

Tuffs and volcaniclastic sediments

Greenschists are fine-grained, well-bedded, light, medium and dark-green in colour. Locally plagioclase-bearing layers are present. These contain abundant, irregular, white, plagioclase pseudomorphs that constitute up to 50 modal percent of the rock. Thickness of individual beds varies from less than a centimetre up to a metre.

Thin (1 m or less) layers of quartz-sericite schist locally occur interbedded with these greenschists. Flattened quartz grains and equant plagioclase crystals up to 5 mm in length are a conspicuous feature. They occur in an extremely fine-grained, quartz-plagioclase matrix.

Quartz-sericite schists locally exhibit a distinctive, fragmental appearance. These fragmental, silicic rocks are particularly well exposed along the shoreline of Glover Island north of Bluff Head and on a well exposed hilltop in the interior of Glover Island located exactly 4 km ESE of Cobble Cove. Flattened clasts of extremely fine-grained, quartz-sericite schist, 1-10 cm in length, occur in a silicic matrix at these localities and are interbedded with massive quartz-sericite schists such as those described above.

Well-bedded greenschists are interpreted to represent mafic tuffs, volcaniclastic sediments and water-lain tuffs. Plagioclase-rich beds are probably crystal tuffs and quartz-sericite schists are interpreted to represent silicic crystal tuffs with phenocrysts of quartz and plagioclase. Fragmental quartz-sericite schists probably represent agglomerates or reworked silicic tuffs.

A rock that is clearly of tuffaceous origin occurs 1.5 km due north of South Point. This rock, DK-193, consists of flattened, irregular, chloritic fragments, 2-10 mm in length in a plagioclase-sericite matrix. These chloritic fragments have irregular, shard-like shapes and concentric rings that parellel their margins. This rock is interpreted to be a lapilli tuff.

Rocks interpreted to be mafic tuffs or fine-grained, volcaniclastic sediments consist of light, medium and dark green, alternating layers of chlorite, epidote, plagioclase and quartz and much unresolvable, fine-grained material.

Plagioclase crystal tuffs occur interbedded with these fine-grained rocks at the south end of Glover Island. Sample DK-80 is interpreted to be one of these plagioclase tuffs. It is a light-green, medium to fine-grained (1 mm or less), greenschist, containing approximately 20 modal percent anhedral, equant, plagioclase crystals. It occurs as a 1-2 m thick layer interbedded with fine-grained mafic tuffs.

Relict plagioclase crystals are completely pseudomorphed by aggregates of clinozoisite, albite and muscovite but preserve their original morphology. The groundmass consists of a microcrystalline aggregate of albite, chlorite, epidote, colourless actinolite and sphene surrounding microphenocrysts of strongly zoned amphibole. Amphibole microphenocrysts consist of discrete grains with pleochroic, dark-brown cores and blue-green rims. The brown portions of these amphiboles consist of titaniferous pargasite (Table 6.2) and green areas of actinolite based upon microprobe analysis. Relict plagioclase,

Table 6.2 Composition of primary amphiboles from tuffs of the Tuckamore Formation determined by microprobe analysis. See appendices for methods and precision. Recalculation basis = 23 oxygens.

	1	2
SiO_2	41.93	41.08
TiO_2	3.32	3.10
Al_2O_3	11.71	12.91
FeO^t	12.37	16.01
MnO	.13	.24
MgO	14.33	10.49
CaO	10.92	11.47
Na_2O	3.04	2.59
K_2O	.26	.25
Total	98.02	98.20

	Min	Int	Max	Min	Int	Max
Si	6.171	6.114	6.058	6.143	6.113	6.084
Al^{IV}	1.829	1.886	1.942	1.857	1.887	1.916
Al^{VI}	.200	.124	.050	.415	.374	.334
Ti	.367	.364	.360	.346	.344	.343
Fe^{+3}	.000	.426	.844	.000	.222	.441
Fe^{+2}	1.520	1.080	.648	2.000	1.768	1.540
Mg	3.144	3.115	3.086	2.334	2.323	2.312
Mn	.013	.013	.013	.026	.026	.026
Ca	1.718	1.702	1.686	1.834	1.825	1.816
Na	.868	.860	.852	.746	.742	.739
K	.047	.047	.046	.043	.043	.042
Total	15.876	15.731	15.584	15.749	15.673	15.598

1. Titaniferous pargasite from DK-109
2. Titaniferous pargasite from mafic tuff (DK-80)

and this pargasitic amphibole, are interpreted to represent mineral fragments in a crystal tuff. Actinolite is interpreted to be of metamorphic origin.

The mafic lapilli tuff located due north of South Point (DK-193) is a distinctive rock consisting of 2-10 mm long, flattened, irregular, shard-like, chloritic fragments in a plagioclase-sericite matrix (Fig. 6.3). These fragments are all concentrically zoned with each individual zone strictly parallel to the fragment margin. A consistent zonation occurs in each chloritic fragment and is interpreted to be in part a primary devitrification feature modified to an unknown extent by later metamorphism.

Microlites of euhedral plagioclase are locally present within these chloritic fragments and are rimmed by a massive, dark-brown unresolvable material and feathery sericitic patches.

The matrix which encloses these chloritic fragments consists of fine-grained, plagioclase, epidote and sericite. Coarse-grained patches of plagioclase and epidote occur locally in the goundmass and may represent metamorphic recrystallization.

Quartz-sericite schists are abundant at the base of the Tuckamore Formation and are interpreted to represent crystal tuffs. They consist of uniformly fine-grained (less than .01 mm), quartz, plagioclase, epidote, biotite and sericite. Quartz and plagioclase phenocrysts are

Figure 6.3 Lapilli tuff within the Tuckamore Formation; X20.

generally present but aphyric varieties are also well represented.

Plagioclase phenocrysts are 1-5 mm in size, equant and euhedral. Composite phenocrysts are abundant and although some broken phenocrysts are recognized, most are well formed.

Quartz is an abundant phenocryst phase forming many large (1-5 mm), equant phenocrysts. These phenocrysts are invariably well rounded and locally embayed indicating some resorption has taken place typical of most silicic volcanic rocks.

Biotite is locally present but is very minor in amount and strongly retrograded to chlorite. It is interpreted to have been the primary mafic phase.

Fine-grained epidote, chlorite and sericite are clearly secondary in origin and occur in all the silicic rocks.

Volcanic breccia

A distinctive lithology within the Tuckamore Formation outcrops 5 km southwest of South Point along the southeast shoreline of Glover Island. This lithology consists of a polymictic volcanic breccia with a metasedimentary and possibly metavolcanic matrix. It is exposed along the shoreline at this locality, is well exposed on the bluff directly above this locality, and for a short distance inland along strike both on Glover Island and on the mainland southeast of Grand Lake. It is apparently of local significance only as similar rocks do not occur elsewhere.

The volcanic breccia consists of clast-supported and matrix-supported rounded, semi-rounded and angular clasts of basalt, rhyolite, argillite, dacite, and bright red jasper (Fig. 6.4). These occur in unsorted layers rich in clasts and in other layers containing sparse, rounded clasts. A crude layering or bedding is present locally. The matrix to this breccia or conglomerate consists of a green, epidote-chlorite-actinolite-plagioclase-quartz schist. Locally it is of obvious sedimentary orgin but elsewhere may be tuffaceous or volcanic in origin.

The volcanic breccia is interpreted to have been deposited proximal to an active volcanic source. Breccia beds containing coarse, clast-supported angular clasts are interpreted to be debris flows. Where the matrix material is of possible tuffaceous origin these debris flows may represent lahars.

A wide variety of clast types are represented in this volcanic breccia and are described below.

A massive, bright-green breccia containing abundant, flattened volcanic clasts, 2 mm to 2 cm in length was collected from the bluff overlooking the south end of Glover Island (DK-106). All clasts are of extrusive igneous rocks of intermediate or silicic composition.

Dominating the clast assemblage is a distinctive, intermediate, extrusive rock locally containing hornblende phenocrysts. The groundmass of this rock is cryptocrystalline and largely unresolvable. Based upon relief and birefringence it is interpreted to represent a fine-grained intergrowth of epidote, quartz and albite. From its cryptocrystalline nature and structureless appearance this epidote-rich material is interpreted to represent a devitrified glassy groundmass and the rock a hornblende porphyry of intermediate composition.

One large fragment of hornblende porphyry consists of a cryptocrystalline, irregular, epidote-rich rim surrounding a chloritic core. The reason for this zonation is unknown but may relate to devitrification or metamorphism of the chilled margin and core of a pyroclastic fragment.

Fine-grained silicic rock fragments are an abundant component of this breccia and are generally composed of quartz and variable amounts of opaques. The fragments are sharply angular and may represent chert clasts.

The most distinctive component of this breccia consists of irregular, splashy-shaped jasper fragments that occur throughout the breccia (Fig. 6.4). These consist of fine-grained quartzites with variolitic textures defined by opaques and hematite inclusions. They are clearly of pyroclastic origin based upon their splashy, shard-like shapes.

The matrix material of this breccia varies between chlorite-quartz-calcite schists to quartz-plagioclase-actinolite-chlorite schists. Due to the fine grain size of this matrix material it is difficult to determine its origin but it is likely that it is locally tuffaceous

Pillow basalts

Pillow basalts occur interbedded with greenschists interpreted to be tuffs and volcaniclastic sediments throughout the Glover Island Area. Locally thick sequences occur composed almost entirely of pillow basalt such as along the northwestern side of Glover Island opposite Northern Harbour. Elsewhere pillow basalt, in some cases only metres thick, occurs within sequences of tuffs.

Pillow basalts consist of light green, massive greenstone (Fig. 6.5). They are generally poor in amygdules and plagioclase phenocrysts, and have thin (1 cm) pillow rims. Interpillow material is lithically identical to that making up the pillow itself. With increasing deformation, these units of pillow lava are rapidly converted to massive greenstone as the pillows themselves contrast little with the interpillow material and pillow rims are thin and poorly developed. Pillows range between 20 cm and 1 m in diameter. Facing directions cannot be determined in most places.

Figure 6.4 Volcanic breccia within the Tuckamore Formation of the Glover Group. Note large jasper clast.

Figure 6.5 Pillow basalt of the Tuckamore Formation of the Glover Group. Photograph was taken along the shoreline of Grand Lake opposite South Point.

One kilometre due north of South Point a small stream cuts through a thick sequence of pillow basalt. Pillow basalt is exposed for 400-500 metres in this stream and consists of an extremely porphyritic pillow basalt. Plagioclase composes the only recognized phenocryst phase. Plagioclase phyric pillow basalts also occur in a stream located 5.5 km northeast of South Point. This stream cuts across almost 1 km of plagioclase phyric, light green pillow basalts and may represent the same unit of pillow basalt. Elsewhere pillow basalts contain very few plagioclase phenocrysts.

Profusely amygdular pillow basalts occur at South Point. These pillow basalts are extremely rich in calcite amygdules which range up to 50 percent in the well developed rim areas. A distinctly greater abundance of amygdules occurs in the top of these pillows which is easily identified from pillow morphology. The extreme vesicularity of these pillow basalts suggests the original magma contained an abundance of volatiles or was erupted in extremely shallow water. The facing direction of these well-formed pillow basalts is to the east.

A feeder dyke within these pillow lavas displays an unusual petrographic texture with rounded plagioclase phenocrysts in an equigranular, fine-grained matrix. This may represent a fluidized dyke which transported these volatile-rich magmas. No other fluidized dykes or anomalously amygdular basalts are recognized in the Tuckamore Formation and are apparently of local significance only.

Well-formed pillows at South Point contain up to 40 modal percent amygdules filled with calcite, minor quartz and epidote. The groundmass consists of subradiating feathery laths of intimately intergrown plagioclase and an unresolvable mafic mineral that is probably actinolite after clinopyroxene. This feathery intergrowth forms a conspicuous variolitic texture.

Pillow basalts along the shoreline adjacent to South Point also contain calcite-filled amygdules and display variolitic textures. Plagioclase microphenocrysts are crudely aligned imparting a weak trachytic texture to these rocks as well.

Elsewhere pillow basalts are composed of a fine-grained aggregate of actinolite, albite, epidote, chlorite and sphene. No relict igneous textures are recognized in these massive greenstones.

Sills

Mafic sills interpreted to be cogenetic with the pillow basalts occur adjacent to South Point and along the shoreline northeast of South Point. They occur as thick tabular sheets of light green basalt that roughly parallel bedding where the latter can be identified. Sill margins are generally amygdular in contrast with the pillow basalts which are generally poor in amygdules. Locally sill margins are coarsely amygdular, and where the carbonate vesicle filling is weathered out the rock appears scoriacious.

Argillites

Argillites are locally present interlayered with quartz-sericite schists at the base of the Tuckamore Formation, at South Point and along the shoreline opposite South Point. All argillites have been metamorphosed to slates in the Tuckamore Formation.

Several thin (5-10 m) graphitic black slate horizons occur interbedded with quartz-sericite schists at the base of the Tuckamore Formation. Black slates are thinly bedded, strongly cleaved and richly pyritiferous in thin (1-4 mm) sedimentary layers and tiny stockworks where the sulphide has moved into fractures. Adjacent silicic tuffs contain abundant disseminated pyrite and minor chalcopyrite as well.

Purple, green and grey, fine-grained, thinly-bedded slates occur at South Point and adjacent to South Point on the southeastern shoreline of Grand Lake. They are discontinuous and of local importance only.

A single, purple slate from South Point has been examined in thin section. It consists of a fine-grained interlocking aggregate of quartz, sericite and opaques and extremely fine-grained, unresolvable minerals. Bedding is clearly represented by distinctive clastic horizons composed of small rounded fragments similar to the matrix of the rock but coarser grained.

Mafic intrusions

The Tuckamore Formation is intruded by numerous sills, dykes and stocks of microgabbro and gabbro. These rocks are distinctive in the field because of their massive, coarse-grained appearance. Microgabbro consists of medium-green, millimetre-sized, equant amphibole crystals in a white or very pale green matrix. Each amphibole grain is entirely surrounded by a lighter, plagioclase matrix yielding a conspicuous, spotted appearance that contrasts strongly with the well-layered tuffs and finer-grained pillow lavas that it intrudes. Coarse-grained gabbros are in every way comparable to microgabbro except that individual amphibole grains range up to 3 mm in size.

A coarse-grained gabbro consists of 1-2 mm laths of randomly-oriented plagioclase pseudomorphs which enclose interstitial grains of green, pleochroic actinolite. Pseudomorphs of iron-titanium oxides composed solely of fine-grained granular sphene up to 1 mm in diameter occur sporadically throughout the section.

The protolith of this coarse-grained rock is inferred to have been an iron-titanium oxide gabbro with an intergranular texture. Mafic minerals may have been clinopyroxene or amphibole.

Geochemistry of the Tuckamore Formation

The geochemistry of the Tuckamore Formation was studied to more completely characterize this important unit and to infer its tectonic setting. Twenty-four samples from scattered locations within the Tuckamore Formation were collected and analyzed for major and trace elements. The sample locations are given in Plate II.

Major elements were determined by G. Andrews (M.U.N.) using atomic absorption spectrometry with the exception of phosphorus which was determined by colourimetry. Total iron was analyzed by atomic absorption spectrometry and is expressed as ferric iron in all chemical analyses. Ferrous iron was not determined.

The abundances of seventeen trace elements were determined by X-ray fluorescence using standard pressed pellets. Sample preparation, accuracy and precision for both trace elements and major elements is discussed in the appendix.

All analyses were normalized to 100 percent anhydrous with total iron expressed as ferrous iron before they were plotted in the various diagrams presented in this chapter.

Alteration

All rocks within the Glover Group have been metamorphosed to the greenschist facies. Primary plagioclase is ubiquitously replaced by albite and clinozoisite and mafic minerals are replaced by chlorite and actinolite. Where relict, igneous pyroxene and amphibole are recognized they are overgrown by metamorphic actinolite. Epidote occurs in the groundmass of these rocks and calcite-filled amygdules occur in some samples. There can be no question that on a local scale elemental mobility has been widespread in all samples as virtually all primary igneous minerals have recrystallized.

Under certain conditions, major elements, trace elements, and rare earth elements may show pronounced mobility (e.g., Dostal et al., 1980; Strong et al., 1979). However under normal conditions of weathering, alteration and greenschist facies metamorphism, Zr, Ti, Nb and Y have been shown to be relatively immobile (Cann, 1970; Smith and Smith, 1976). These elements are accordingly emphasized in the following analysis.

Major elements may or may not have behaved in a mobile fashion during metamorphism. It is reasonably clear that the alkalis have been mobilized. Basalts and rhyolites are all depleted in potash and enriched in soda as a cursory examination of the analyses presented in this section will attest. They therefore correspond to spilite and keratophyre in chemistry and petrology.

Other major elements and trace elements generally reflect primary abundances. This contention is supported by the strong correlation observed between petrography, major element chemistry, mobile trace elements and immobile trace elements. For example, rocks containing relict igneous amphiboles are invariably enriched in the light rare earths. Rocks high in aluminium and calcium invariably contain appreciable modal plagioclase and abundant strontium, consistent with plagioclase accumulation. Rocks high in zirconium are also enriched in iron, titanium and vanadium and depleted in calcium, magnesium and aluminium consistent with differentiation trends observed in unaltered basaltic rocks. Several rock samples contain anomalously high values of magnesium and chromium consistent with the accumulation of a mafic phase. One of these samples (DK-109) is also enriched in the light rare earths and contains abundant relict, igneous titaniferous pargasite indicating the mafic phase is amphibole. Thus major elements, trace elements, and light rare earth elements all show strong qualitative agreement between relict, igneous modal mineralogy, immobile trace elements and each other and are therefore interpreted to reflect primary compositions.

Description of chemical analyses

A total of thirteen samples were analyzed from the Tuckamore Formation. Sample locations are given in Plate II and analyses are listed in Table 6.3. Most of the analyses are from tuffaceous rocks. DK-124, 14, 80, 496, and 532 are all feldspar crystal tuffs and quartz-feldspar silicic crystal tuffs. DK-91 and 173 are pillow basalts and compare quite closely with mafic crystal tuffs. DK-109 is interpreted as a dyke and NH-3 is a massive greenstone of unknown origin. DK-191 and NH-191 are rhyolite sills. These analyses are briefly described below.

DK-109

This fine-grained, massive green rock is extremely interesting petrographically as it consists of euhedral prisms of titaniferous pargasite (microprobe analysis) and anhedral plagioclase. The relict igneous pargasite grains are ubiquitously overgrown by epitaxial rims of actinolite (microprobe analysis) and the plagioclase is an albite clouded with secondary, tiny inclusions of sericite and epidote. It occurs as a métre-thick layer within volcaniclastic metasediments and is interpreted as a sill.

DK-109 is unusual in that is has very low aluminium and incompatible trace element abundances but very high iron, magnesium, nickel, chromium and light rare earth element abundances. Coupled with the modal abundance of pargasite,

Table 6.3 Major element and trace element chemistry of rocks of the Tuckamore Formation.

	1	2	3	4	5	6	7	8	9	10	11
	109	124	14	91	80	173	NH-3	496	532	191NH	191
SiO_2	47.9	44.5	45.6	45.6	45.2	48.7	47.7	75.4	77.7	73.9	69.5
SiO_2	.83	1.24	1.22	1.28	1.51	1.55	1.72	.19	.14	.14	.42
Al_2O_3	10.6	18.3	16.8	17.8	17.4	14.0	15.2	11.8	11.7	14.0	15.8
$Fe_2O_3{}^t$	11.08	9.37	9.42	9.24	9.81	10.57	12.65	2.45	1.63	.80	2.32
MnO	.21	.14	.14	.16	.15	.16	.20	.05	.01	.03	.07
MgO	14.15	8.54	8.80	6.42	6.98	6.33	5.84	1.70	1.12	.08	1.22
CaO	8.80	11.70	10.56	10.38	11.63	9.94	8.69	.06	.06	.92	1.05
Na_2O	2.74	1.85	2.43	3.31	2.62	3.29	3.97	3.36	3.87	4.13	5.43
K_2O	.35	.04	.36	.14	.16	.17	.25	1.73	1.09	4.09	2.86
P_2O_5	.16	.07	.11	.12	.32	.17	.18	.02	.01	.00	.21
L.O.I.	3.15	3.51	3.36	3.66	3.12	3.41	1.94	1.86	1.37	1.48	1.79
Total	99.97	99.26	98.80	98.11	98.90	98.29	98.33	98.62	98.70	99.57	100.67
Nb	0	0	1	0	0	3	1	2	2	27	22
Zr	63	80	89	93	108	117	137	135	134	52	193
Y	18	29	28	32	34	40	54	55	57	14	16
Sr	103	333	224	318	219	165	227	65	141	51	359
Ga	9	10	12	14	11	12	23	16	15	27	23
Ni	322	100	136	62	69	55	42	32	22	24	26
Cr	809	229	271	222	217	244	20	0	1	0	22
V	194	204	214	229	245	316	347	26	27	8	50
La	39	8	3	5	10	7	28	74	66	0	32
Ce	19	2	5	0	5	9	0	10	6	61	131

1 Pargasite basalt
2,3 Plagioclase crystal tuff
4 Plagioclase-phyric basalt
5 Plagioclase crystal tuff containing primary pargasite
6 Pillow basalt
7 Massive greenstone
8,9,10,11 Silicic tuffs

these data are interpreted as indicating amphibole accumulation.

DK-124, 14, 91, 80

These rocks consist of light-green greenstones containing abundant phenocrysts of plagioclase replaced by clinozoisite-albite aggregates. All occur as thin (1-2 m) layers within sequences of mafic tuffs and volcaniclastic sediments and are interpreted to be feldspar crystal tuffs. DK-80 contains relict, titaniferous pargasite.

The effects of plagioclase accumulation are readily observed in these rocks as all are high in aluminium, calcium and strontium.

DK-173

DK-173 consists of a light-green, pillow basalt with minor calcite amygdules. In thin section it consists of weakly-aligned, plagioclase microphenocrysts in a brown, unresolvable groundmass containing randomly-oriented plagioclase microlites.

DK-173 closely resembles an average tholeiitic basalt in its composition.

NH-3

NH-3 is a massive, dark green greenstone from the north end of Glover Island. It contrasts with other basalts in the Tuckamore Formation in that it is somewhat more differentiated. Iron is unusually high, magnesium is low and incompatible trace elements, vanadium and titanium are all high, typical of an iron-enriched, differentiated, tholeiitic basalt. Chromium is subordinate to nickel in NH-3 as well, in contrast with other rocks of the Tuckamore Formation.

DK-496, 532

Two samples of quartz-feldspar phyric, silicic tuff collected from the base of the Tuckamore Formation were analyzed. They have silica contents around 75 weight percent indicating rhyolitic compositions but have low potash and high soda indicating they have undergone metasomatism to keratophyre.

DK-191, NH-191

Two foliated, rhyolitic sills were sampled from the Tuckamore Formation. These rocks are tan to pink and contain alkali feldspar phenocrysts in addition to plagioclase and quartz phenocrysts. Silica content is around 70 weight percent. These rhyolites contrast with the rhyolitic tuffs described above in that they contain alkali feldspar phenocrysts and have up to 4 weight percent potash.

Graphical representation

These chemical data are interpreted using several standard diagrams to establish variations and differences among the rocks analyzed and, if possible, to indicate a tectonic environment (Fig. 6.6 and 6.7).

The silica, iron-magnesium ratio plot used here is taken from Miyashiro (1974). Rocks from a single differentiating magma will yield a single, linear array in this figure. If rocks from different magmas with contrasting degrees of iron-enrichment are plotted, they will form separate groups at the same silica content. This diagram is used here to discriminate between different magmas.

The nickel-chromium ratio is largely a function of source composition and degree of fractionation of olivine, spinel and clinopyroxene. Contrasting nickel-chromium ratios indicate a different source or a different olivine, spinel or clinopyroxene fractionation history. Contrasting nickel-chromium ratios thus indicate different magmas and can be used to establish magmatic affinities.

Immobile trace elements have been used by Pearce and Cann (1971, 1973) to discriminate between basalts formed in different tectonic settings. The field of within plate basalts can be discriminated from other basalts by using a Ti/100-Yx3-Zr ternary diagram. Overlap in the non-within plate field can be eliminated to some extent by using a Zr-Ti plot although this diagram is very sensitive to fractionation and crystal accumulation as it uses absolute abundances rather than ratios. These diagrams are presented and interpreted below.

Subalkaline nature of the Tuckamore Formation

All rocks of the Tuckamore Formation are subalkaline in composition based upon their trace element abundances. Niobium is very low or below detection in all analyzed basalts yielding a very high Nb/Y ratio. This ratio is tyically quite low in alkalic rocks (Pearce and Cann, 1973) clearly requiring the Tuckamore Formation to be subalkaline in composition.

Alkalic rocks are also generally quite high in phosphorus. Phosphorus is very low in rocks of the Tuckamore Formation in most cases further substantiating the subalkaline nature of these rocks (Fig. 6.6). DK-80 is an exception as it appears to be transitional. Based upon Nb, Y and Ti, however, DK-80 is also interpreted to be subalkaline.

SiO_2 - FeO^T/MgO Diagram

Porphyritic rocks of the Tuckamore Formation define an elongate field parallel to the boundary between calc-alkaline and tholeiitic fields in this diagram. DK-109 plots out of this field due to amphibole accumulation.

In contrast to porphyritic rocks, NH-3 plots well within the field of tholeiitic basalts. It is clearly

Figure 6.6 P_2O_5 - Zr Diagram for rocks of the Tuckamore Formation. Alkaline field boundary is from Winchester and Floyd (1976). Numbered data points refer to Table 6.3.

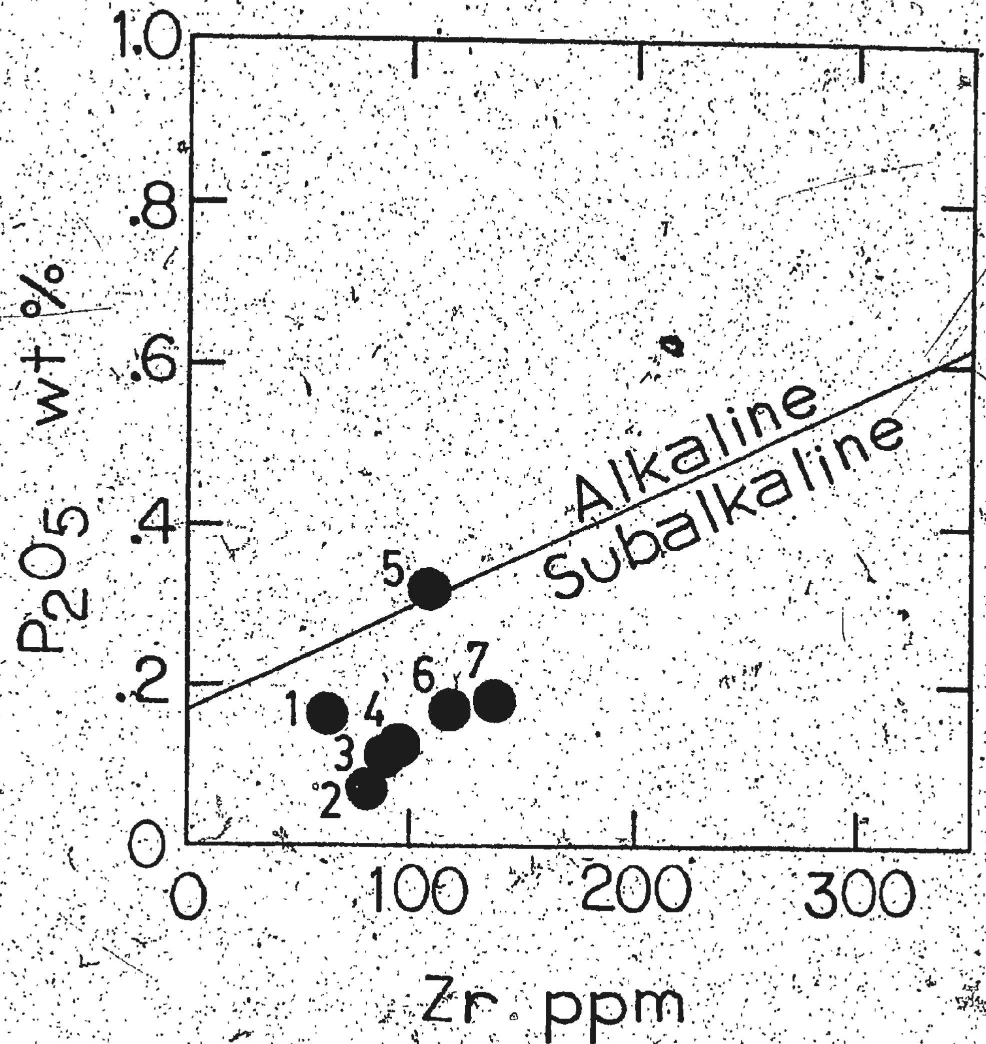
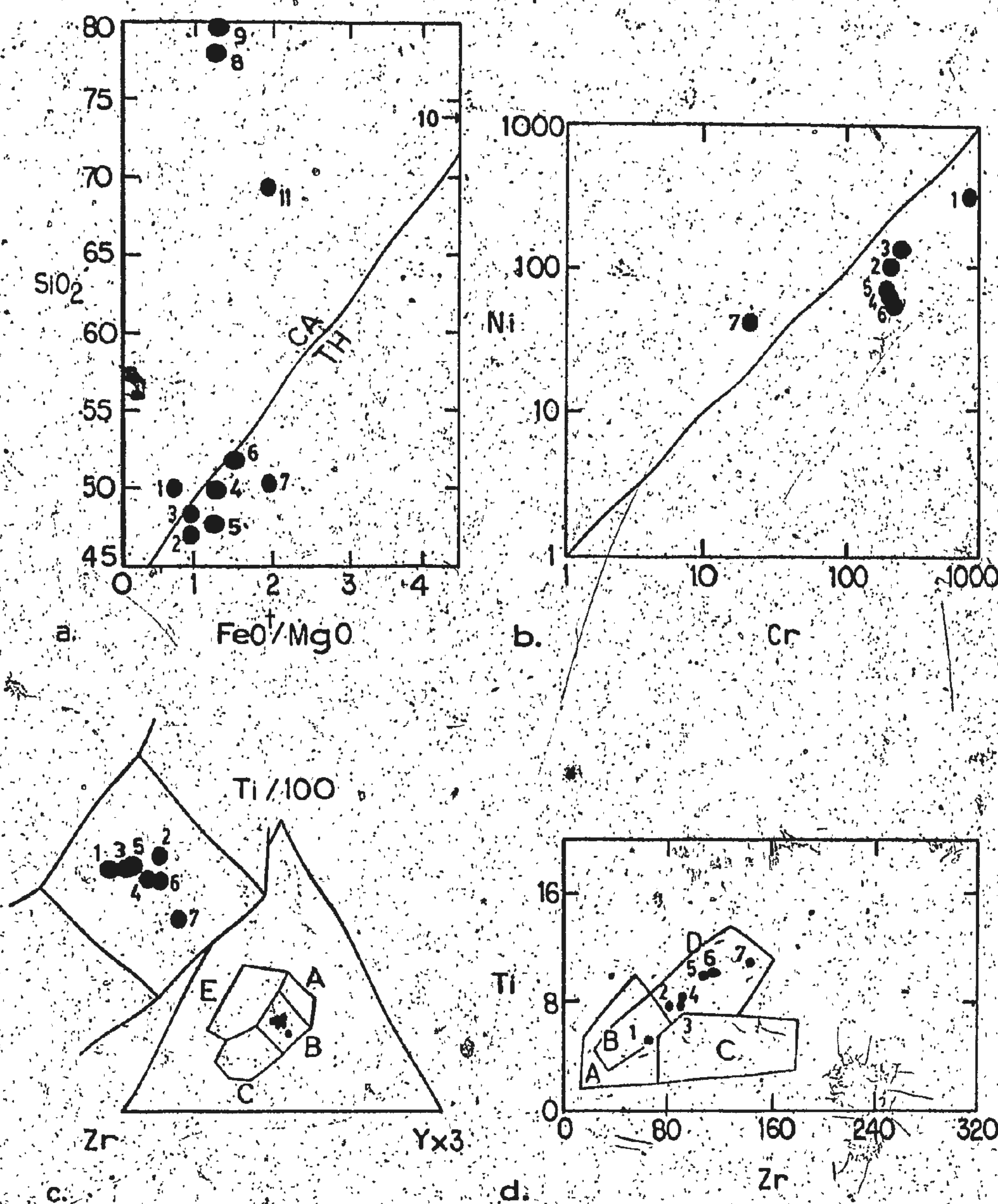

1.0
.8
.6
.4
.2
0
0
100
200
300
P_2O_5 wt %
Zr ppm
Alkaline
Subalkaline
1
2
3
4
5
6
7

Figure 6.7 Chemistry of the Tuckamore Formation of the Glover Group. Numbered points refer to analyses given in Table 6.3. Figure a is after Miyashiro (1974) and Figures c and d are after Pearce and Cann (1971, 1973).


SiO2
FeO†/MgO
CA
TH
a.
Ni
Cr
b.
Ti/100
Zr
Y×3
A
B
C
E
c.
Ti
Zr
A
B
C
D
d.

more iron-enriched than porphyritic rocks at the same silica content and thus appears to represent a separate, unrelated magma.

Ni/Cr Diagram

The nickel-chromium diagram also suggests that porphyritic rocks of the Tuckamore Formation are related as they form a single, well-defined compositional field. DK-109 plots outside of this field as expected due to its cumulate nature.

NH-3 also plots well out of this compositional area as in the silica-iron magnesium ratio diagram indicating that it is unrelated to porphyritic rocks of the Tuckamore Formation.

Ti/100-Yx3-Zr Diagram

Rocks of the Tuckamore Formation are plotted in a Ti/100-Yx3-Zr diagram in Figure 6.7 where two, distinct, compositions are defined. Porphyritic rocks of the Tuckamore Formation form one distinct field and NH-3 defines the second.

It is important to note that DK-109 falls into the field of porphyritic basalts in this ternary diagram but plots outside of this field in other diagrams. This is the behavior expected of a cumulate rock in which incompatible trace elements have been diluted by the accumulation of a phenocryst phase. This effect is subtracted in plots involving ratios of incompatible trace elements where the true affinity of such a rock can be detected.

NH-3 plots well out of the field of porphyritic rocks consistent with other diagrams.

All rocks plot outside of the field of within plate basalts in this diagram.

Ti-Zr Diagram

In a zirconium-titanium diagram, basalts of the Tuckamore Formation form a single linear trend with porphyritic rocks defining a single compositional space within the ocean floor basalt field. The dilution of incompatible trace elements by accumulation of amphibole phenocrysts is clearly observed in DK-109 which plots closer to the origin in the field of overlap.

The more evolved nature of NH-3 is clearly brought out in this diagram as NH-3 plots in the high zirconium, high titanium part of the ocean-floor basalt field.

Summary of geochemistry

Analyzed basalts from the Tuckamore Formation define three, distinct compositional fields related to two magmas. Porphyritic basalts are all clearly related as they form a single cluster of points in all diagrams. DK-109 falls outside of this field in many diagrams because of its cumulate nature but in ternary plots involving incompatible trace elements it falls within the field of porphyritic basalts with which it is clearly related. Titaniferous, relict, igneous pargasite occurs in DK-109 and DK-80 establishing a further, petrologic link between these two rocks.

In the silica-iron/magnesium diagram these porphyritic basalts fall along the boundary between calc-alkaline and tholeiitic fields. In trace element plots they fall in the ocean floor basalt field but on a trend through the low-K tholeiite field. They are therefore similar in many respects to either low-K tholeiites or ocean floor basalts. However the occurence of a titaniferous pargasite and rhyolitic rocks, together with these basalts is very unusual in an ocean floor setting, and on that basis they are tentatively interpreted to be low-K tholeiites.

NH-3 consistently falls out of the field of porphyritic basalts in all diagrams and on this basis is considered to be derived from a separate magma. It is more iron-enriched at a similar silica content than are the porphyritic rocks and is thus much more tholeiitic.

NH-3 consists of a massive, basaltic greenstone. It may represent a later dyke or stock intrusive into the Tuckamore Formation as it is clearly unrelated to the surrounding rocks based upon its iron-enriched, highly fractionated tholeiitic chemistry.

STRUCTURE OF THE GLOVER GROUP

Structures relating to three deformations are recognized in the Glover Group. A penetrative foliation and cogenetic stretching lineation formed during the first deformation. The second and third deformations are defined by abundant minor folds and axial planar, heterogeneously developed, crenulation cleavages.

D1 Structures

S1 consists of a well developed foliation recognized throughout the Glover Group. It is most apparent in quartz-rich lithologies such as silicic tuffs where it is defined by millimetre-scale, fine-grained layers of quartzite and sericite. Layers are in sharp contact with one another and continuous on the scale of the outcrop. The strict conformity between these layers and S1 suggests they are structural in origin.

A conspicuous, cogenetic, stretching lineation (L1) is invariably associated with S1 and is particularly well developed in quartz-rich lithologies. L1 is defined by thin (5 mm) ridges or mullions of quartz on cleavage surfaces that are continuous on the scale of the outcrop.

L1 is particularly well developed in deformed, silicic volcanic rocks at the base of the Tuckamore Formation that outcrop along the shoreline of Glover Island halfway between Red Point and Cobble Cove. These silicic rocks generally contain numerous phenocrysts of plagioclase and quartz. During deformation, quartz becomes highly strained while plagioclase phenocrysts remain comparatively undeformed. Pressure shadows of quartz invariably form around these rigid plagioclase grains which are locally brittlely broken and pulled apart. Pressure shadows are extremely elongate and define a single extension direction. They are continuous on the scale of a thin section and may be continuous on the scale of an outcrop.

These relationships indicate that L1 is a stretching lineation that defines the direction of maximum elongation of the finite strain ellipsoid. From the exceptional length of these pressure shadows D1 is inferred to have been a high strain event.

In greenschists, S1 is defined by the parallel orientation of chlorite, sericite and where present, actinolite. L1 is generally absent as it is restricted to quartz-rich lithologies.

D2 Structures

D2 structures consist of minor folds and a crenulation cleavage. F2 folds occur throughout the Kettle Pond Formation and base of the Tuckamore Formation. They are small, ranging up to several metres in amplitude, moderately to strongly asymmetric and tight to open. Orientation and plunge varies considerably because of refolding.

F2 folds are restricted to the Kettle Pond Formation and base of the Tuckamore Formation where slabby, well-layered lithologies occur. Higher in the Tuckamore Formation where more massive lithologies predominate, minor folding is rare.

A heterogeneous crenulation cleavage formed axial planar to F2 folds during D2. S2 is well developed in highly folded areas and poorly developed in unfolded, massive rocks.

D3 Structures

The recognition of D3 structures is rather difficult, because to positively identify a D3 structure it must clearly overprint D1 and D2 structures. The number of D3 structures recognized is, as a result, rather small. However a widely spaced crenulation cleavage and open, chevron folds are recognized which overprint D1 and D2 structures.

S3 consists of an upright, heterogeneously developed crenulation cleavage defined by tiny folds and crenulations of S1 or S2.

Minor F3 folds are open, chevron, upright folds with steeply plunging axes. Amplitudes are less than 1 m. These folds range in size down to microfolds transition to S3.

GEOMETRIC ANALYSIS OF THE GLOVER GROUP

D1 Structures

Poles to S1 surfaces and L1 lineations are shown in lower hemisphere, equal-area projections in Figure 6.8 for subareas IV and V (see Chapter 4 for discussion of subareas). Poles to S1 surfaces form diffuse, great circle girdle distributions in each subarea and L1 lineations form single point maxima.

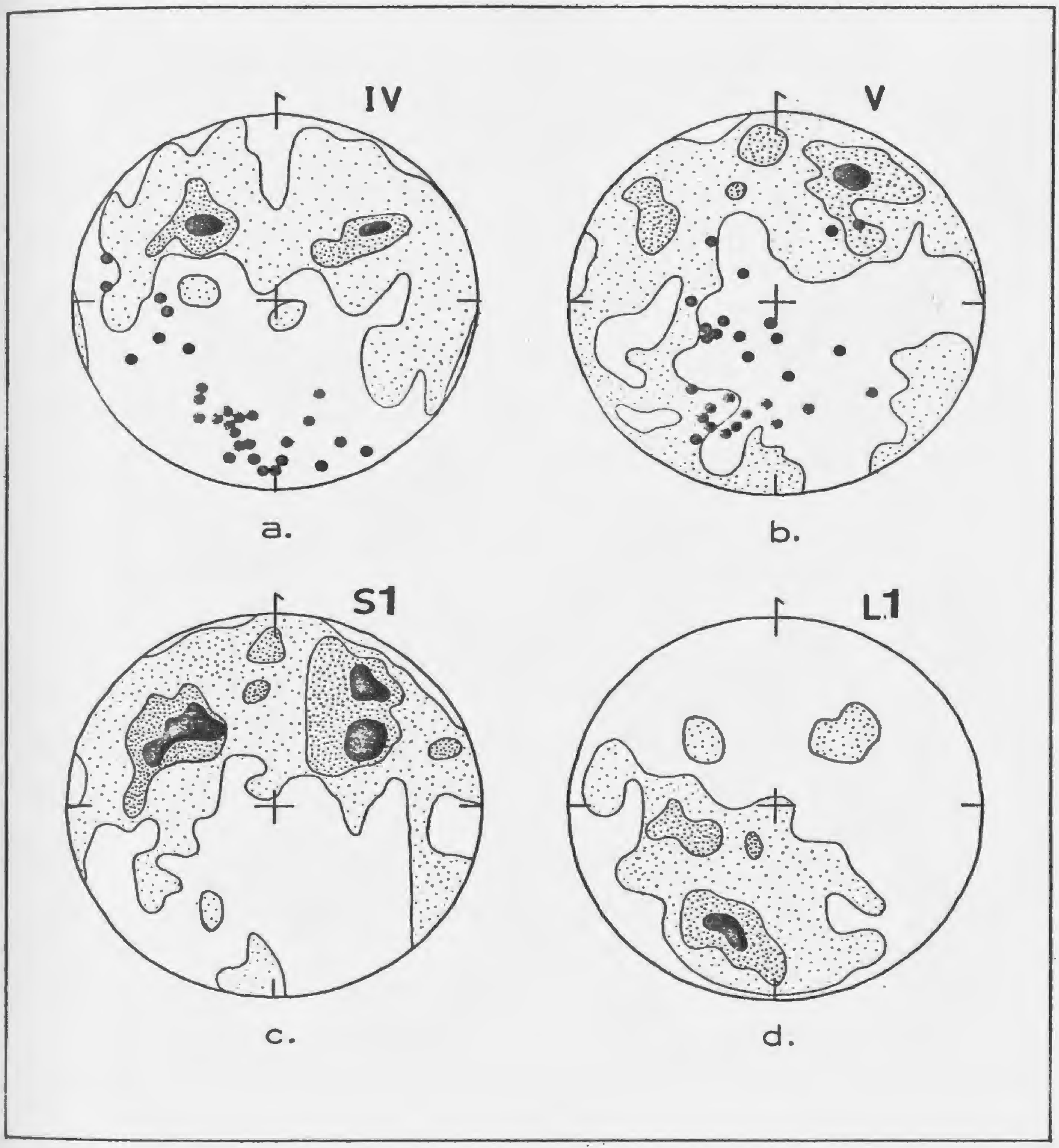
In subarea IV, poles to S1 surfaces form a diffuse but well-defined great circle girdle. This girdle is largely the result of D3 folding related to the Glover Anticline. Its diffuse nature is inferred to result from limited girdle development during D2 refolded during D3.

L1 lineations in subarea IV form a single point maximum plunging to the south. No spread of this maximum occurs due to F3 folding because L1 and L3 are coaxial.

Poles to S1 surfaces form a diffuse great circle girdle distribution in subarea V. This distribution is interpreted to be the result of F2 folding. Major and minor F2 folds deform poles to S1 into a great circle girdle that plunges to the south.

L1 data in subarea V plunges moderately to the southwest and defines a single point maximum. This distribution of points indicates the dominance of D2 strain in this subarea as D3 tends to displace L1 in a great circle girdle pattern.

Figure 6.8 D1 data from the Glover Group plotted in equal area, lower hemisphere projections. (a) Poles to S1 surfaces (contoured) and L1 lineations (solid dots) from Subarea IV; (b) Poles to S1 surfaces (contoured) and L1 lineations (solid dots) from Subarea V; (c) Poles to S1 surfaces from Subareas IV and V, combined; (d) L1 data from Subareas IV and V, combined. 29 lineations are plotted in a and 28 lineations in b. Contour intervals for the contoured data in percent per one percent area and the number of points contoured for each figure are as follows: (a) .7,3.6,7.2; S2=139; (b) .6,3.2,6.5; S2-154; (c) .3,1.7,3.0; S2=293; (d) 1.7, 5.1,15.5; L2=58.


IV
V
a.
b.
S1
L1
c.
d.

D2 Structures

In subareas IV and V, D2 structures are affected by D3 deformation related to the Glover Anticline. This deformation is expressed by the gradual rotation of poles to S2 surfaces from west to east along a poorly defined great circle girdle inclined steeply to the north (Fig. 6.9). This great circle girdle is poorly defined because it is largely composed of two point maxima, one related to the east limb and the other related to the west limb of the Glover Anticline. Data in the hinge region of the Glover Anticline is scanty reflecting extremely poor outcrop exposure.

L2 lineations form a point maximum to the southwest in subarea V and a great circle girdle in subarea IV (Fig. 6.9). Combined, this data describes a single, diffuse, great circle distribution. This great circle distribution is attributed to the rotation of L2 lineations around the Glover Anticline.

METAMORPHISM OF THE GLOVER GROUP

A major syntectonic metamorphic event during D1 recrystallized rocks of the Glover Group. Quartzo-feldspathic rocks recrystallized to fine-grained quartz, plagioclase, sericite and epidote. Mafic rocks recrystallized to chlorite-calcite-quartz-plagioclase greenschists at the base of the Tuckamore Formation and actinolite-

Figure 6.9 Poles to S2 surfaces (contoured) and L2 lineations (solid dots) for Subareas IV and V of the Glover Group plotted in equal-area, lower hemisphere, projections. Contour intervals in percent per one percent area, the number of points contoured and the number of lineations plotted are: (a) 1.3,5.4,12.1; S2=74; L2=91; (b) 1.5,4.7,7.9; S2=63; L2=81.

epidote-plagioclase greenschists higher in the Glover Group. As this contrast in mineralogy is considered significant, these two rock types are described below more completely.

Actinolite-bearing rocks of the Glover Group are composed of actinolite, chlorite, epidote, albite, sphene, clinozoisite, quartz and muscovite. Actinolite is generally colourless or very pale green. It forms fibrous or acicular aggregates together with chlorite which replace mafic, igneous minerals. Otherwise they occur in parallel aggregates together with chlorite to define S1. Epidote occurs as idiomorphic equant prisms in the groundmass of the rock. Clinozoisite forms tiny idiomorphic prisms in albite which generally replace primary subidiomorphic plagioclase or forms xenomorphic interstitial grains in the groundmass. Calcite is totally absent or present in very minute amounts in these rocks. Quartz is rare or absent.

Mineral assemblages lacking actinolite consist of chlorite-calcite-epidote-albite-quartz-muscovite and chlorite-calcite-albite-quartz-muscovite. No relict grains of actinolite are observed. Epidote is present as irregular grains in a chloritic matrix and quartz is abundant. Muscovite is also abundant in these rocks and shows a strong negative correlation with actinolite and strong positive correlation with calcite and chlorite, overall. Blackened, corroded granular aggregates represent degraded sphene and occur in all mineral assemblages lacking actinolite.

Epidote is present in these rocks in variable amounts or is entirely absent. It is optically clean but irregular in shape. If a stable phase, it has certainly re-equilibrated with the groundmass and been partially resorbed.

A number of sections examined contain mineral assemblages that are intermediate between these two end-member types. However these are rare and may represent disequilibrium conditions.

Carbonate metasomatism of mafic rocks is a well documented phenomenon and many examples have been given in the literature (e.g., Hynes, 1981; Barron and Barron, 1976). By analogy, the sudden change in mineralogy of mafic rocks at the base of the Glover Group is interpreted to be due to carbonate metasomatism.

The source of carbon dioxide required for this metasomatic alteration probably originated from the Grand Lake Complex. Ultramafic rocks are commonly metasomatized by carbonate-rich fluids during metamorphism and mineral assemblages such as dolomite and magnesite in the Grand Lake Complex support this contention. Based upon mineral assemblages and microstructures in the Glover Group, the activity of CO_2 increases towards the Grand Lake Complex and therefore points to the complex as the source for the carbonate that metasomatized the Glover Group. Origin of the carbonate needed to originally alter the Grand Lake Complex is unknown, but may have formed during decarbonation reactions of the underlying rocks.

Mineral growth during D2 and D3 deformations is restricted to minor chlorite and sericite along crenulation cleavages and retrogression of relict minerals.

Based upon the mineral assemblage: actinolite-epidote-albite, D1 is inferred to have occurred during greenschist facies conditions. D2 and D3 are inferred to have taken place during lower grade conditions based upon the growth of chlorite and sericite and retrogression of other minerals.

CONTACT RELATIONS OF THE GLOVER GROUP

The contact between the Kettle Pond Formation and the underlying Grand Lake Complex is exposed along the shoreline of Glover Island and at Kettle Pond. Along the shoreline of Glover Island south of Cobble Cove a quartzose schist containing quartz grains up to 5 mm in diameter in a fine-grained, quartz-sericite matrix marks the base of the Kettle Pond Formation. No compositional layering or other internal structure is recognized in this unit.

Although the protolith of this distinctive lithology is difficult to determine due to recrystallization, it is clear that the coarse grains of quartz are relict grains within a pelitic or semipelitic matrix. This psammitic schist is accordingly interpreted to represent a quartzose metagreywacke containing detrital grains of quartz. The coarse size of these quartz grains suggests a plutonic source terrane.

The contact is also exposed 1 km north of Kettle Pond. Here foliated greenschists of the Grand Lake Complex are in sharp contact with quartz-sericite schists of the Kettle Pond Formation containing abundant clasts of trondhjemite.

Throughout the Kettle Pond Formation, clasts are recognized that greatly resemble various lithologies of the underlying Grand Lake Complex. Based upon this sedimentological link and the abrupt nature of the contact it is interpreted to represent an unconformity. The conglomerate located at the base of the Kettle Pond Formation along the east side of Kettle Pond marks the unconformity as well.

The Kettle Pond Formation changes gradually from south to north to a more mafic composition until it becomes indistinguishable from mafic rocks of the overlying Tuckamore Formation. This gradual change is interpreted to be due to a facies change resulting in a stratigraphic pinch-out of the Kettle Pond Formation just south of Bluff Head. At Bluff Head, the Tuckamore Formation overlies the Grand Lake Complex.

INTERPRETATION

The Glover Group is interpreted to represent a thick sequence of volcanic rocks including pillow basalts and mafic and silicic pyroclastic rocks built upon a basement of ophiolite. The basal part of the volcanic pile contains an abundance of silicic tuffs perhaps accounting for the abundance of quartz in the Kettle Pond Formation. Silicic volcanism may have contributed material to the Kettle Pond Formation in the form of reworked silicic tuffs. This fine-grained silicic matrix material contrasts with coarse, rounded clasts derived from the underlying ophiolite emphasizing two source terrains.

The Tuckamore Formation is interpreted to have formed in a volcanic island setting. Major and trace elements indicate the dominant mafic rock is basaltic and subalkaline in composition. Based upon the iron/magnesium ratio these rocks compare favorably with low-K tholeiites although they overlap considerably with the ocean floor basalts in terms of composition and various trace element ratios. However these rocks also contain primary titaniferous amphibole and significant silicic volcanics which are unusual in ocean floor environments. It is concluded an island arc origin for rocks of the Tuckamore Formation is more likely.

AGE

The Lower Ordovician Corner Pond Formation unconformably overlies the Glover Group at Corner Pond thereby providing a minimum age of Lower Ordovician for the Glover Group. As underlying rocks of the Grand Lake Complex are inferred to be Lower Ordovician, the Glover Group must fit into a rather restricted time frame between the Grand Lake Complex and Corner Pond Formation. It is interpreted to be Lower Ordovician in age.

SIGNIFICANCE

The Glover Group is very important to understanding the evolution of the Appalachian Orogen in western Newfoundland as it links volcanic island volcanism and ophiolite emplacement. An unconformity formed on top of the Grand Lake Complex eroding the upper part of the ophiolite sequence. Subsidence coupled with silicic volcanism formed a cover of silicic greywacke containing large rounded clasts of the underlying ophiolite. Continued subsidence was followed by silicic and mafic volcanism to form the Tuckamore Formation. The volcanism that formed the Glover Group must have closely followed exhumation and subsidence of the underlying Grand Lake Complex.

Chapter 7

LOWER ORDOVICIAN OPHIOLITIC VOLCANICS: The Otter Neck Group

The Otter Neck Group is named after Otter Neck, located at the south end of Glover Island. It is applied to a fault-bounded block of mafic volcanic rocks that obliquely transects the southern end of Glover Island and extends southeast onto the mainland (Plate I).

A high-angle fault on Glover Island divides the unit into a northern block and a southern block. The northern block consists largely of massive basalt and microgabbro interpreted to be dykes, but contains a thin unit of aquagene tuff along its southeastern side. It is unconformably overlain along the northwestern shoreline of Glover Island by the Red Point Formation. The unconformity is developed on a trondhjemite that intrudes the Otter Neck Group.

The southern block consists of dykes of basalt and microgabbro along the northwestern shore of Glover Island These dykes locally alternate with screens of pillow basalts and are intruded by a number of silicic porphyries.

The Otter Neck Group is also well exposed on the southern shore of Grand Lake. Here dykes of basalt and microgabbro occur along its eastern and western margins while pillow basalt, tuffaceous rocks and coarse pillow breccias occur within the central part. A trondhjemite intrudes the Otter Neck Group along this shoreline and produces a thin aureole of hornfels.

Many intrusions of silicic porphyry and trondhjemite intrude the Otter Neck Group. These are grouped together and discussed at a separate point in this chapter.

Thickness of the Otter Neck Group is unknown. Outcrop width ranges from 1 km to 5 km.

LITHOLOGY AND PETROLOGY

The Otter Neck Group consists primarily of mafic extrusive rocks and dykes intruded by a younger suite of silicic rocks. These two groups of rocks are discussed separately below.

Dykes and volcanic rocks of the Otter Neck Group

Dykes

Rocks interpreted to compose a sheeted dyke complex are exposed along the northwestern shore of Glover Island where dark, blue-black, massive fine-grained diabase forms much of the shoreline. Close inspection locally reveals fine-grained, dark-black, chilled margins within these basaltic rocks. Chilled margins are upright and strike roughly north and south. Dykes are thick and range from a minimum of about 1 m up to at least several metres. Chilled margins are vague and their recognition is greatly dependent upon the quality of the outcrop surface. Chilled margins are most conspicuous on surfaces that are smooth, slightly weathered and uncomplicated by fractures.

Most of these dykes are basaltic in composition. However at the western margin of the Otter Neck Group along the northwestern shoreline of Glover Island, several 2 m thick, light grey-green, fine-grained, aphyric dykes are present. These dykes are siliceous and andesitic or dacitic in composition. They parallel the basaltic dykes and are interpreted to be an integral part of the dyke complex. However the possibility that they are related to nearby silicic porphyries cannot be dismissed.

Mafic dykes consist of massive, fine-grained, greenstone composed of albitized plagioclase and actinolite in a matrix of fine-grained actinolite, albite, chlorite and epidote.

Silicic dykes consist of subhedral laths of albitized plagioclase and patches of chlorite and epidote in a groundmass of myrmekitic quartz and albite. This texture is similar to granophyres but may be metamorphic in origin rather than primary. In composition these dykes are micro-trondhjemites. With a decrease in grain size they would grade into dacites or andesites.

Pillow lavas

Pillow lavas of the Otter Neck Group outcrop about

30 m along the northwestern shoreline of Glover Island in its western part and in its central part along the south-eastern shoreline of Grand Lake.

Pillows on Glover Island range between 1 m to 3 m in length and are well formed. No pillow rim is developed and the pillows are a dark green or black.

No amygdules are found in any of these pillow lavas. However irregular quartz-epidote clots locally occur in pillow rim areas.

Interpillow material is minor and generally identical to the pillows themselves. However locally on Glover Island and on the mainland black and grey chert occurs as interpillow material. Chert on Glover Island contains microscopic, spheroidal shapes defined by concentrations of dusty opaques. These may have formed around radiolarian skeletal material.

Pillow lavas consist of microphenocrysts of albitized plagioclase in a very fine-grained, dark-brown variolitic textured groundmass. The variolitic texture is well developed in these pillow lavas and is defined by intimate, radial intergrowths of a brown mineral and albitized plagioclase. The brown mineral is probably altered clinopyroxene.

Locally pillows contain blebs of quartzite. These blebs have rounded margins and are locally spherical. Elsewhere they are highly irregular. Many have cores of epidote and all are composed of polygonized quartz grains.

These structures resemble amygdules but are more irregular in shape, size and distribution. They strongly resemble quartz segregations found in glassy pillow rim areas of unaltered pillows, however, and are interpreted to represent a similar structure.

Pillow breccia and aquagene tuffs.

Rocks interpreted to be pillow breccias and aquagene tuffs outcrop in the central part of the Otter Neck Group along the southeastern shoreline of Grand Lake.

Angular, pale-green fragments up to 4 cm in size occur along this shoreline in a dark, blue-black basaltic matrix. These fragments are interpreted to represent broken pillow basalts and these units, pillow breccias.

Fine-grained greenschists occur in the Otter Neck Group at one location along the northwest shoreline of Glover Island and along the eastern margin of this unit on Glover Island. Similar rocks also occur in the central part of the Otter Neck Group along the southeastern shoreline of Grand Lake.

The greenschists locally contain black, irregular, basaltic fragments that may represent broken pillow lavas. They also contain abundant spheroids that are commonly zoned or hollow and have various irregular shapes. The origin of these spheroids is unknown. They resemble amygdules but contrast in their nonspherical shape and irregular sizes. They may be varioles. Greenschists are interpreted to represent aquagene tuffs as described by Carlisle (1963) as they occur in close association with pillow breccias.

Most of the rocks along the southeastern shoreline of Grand Lake are hornfelsed by a trondhjemite intrusion that outcrops locally in the central part of the Otter Neck Group along the shoreline. Virtually all these rocks are overgrown by decussate actinolite aggregates, as a result of intrusions. However some structures are preserved as described below.

Spheroids are abundant in these rocks and are found in aquagene tuffs and in the matrix to pillow breccias. Some of these spheroids are perfectly spheroidal but many others are irregular or oblong in shape. All are composed of quartz and epidote in varying proportions.

Irregular, dark-green fragments also occur in these aquagene tuffs. In thin section these mafic patches are found to consist entirely of chlorite. These chloritic patches may represent glassy fragments from hyaloclastites that have devitrified to chlorite during later metamorphism.

Mafic gabbro

Mafic gabbro locally cuts other rocks of the Otter Neck Group along the southeastern shoreline of Glover Island. These mafic gabbros are coarse-grained (1-5 mm) and very distinctive in the field. One gabbro located along the eastern margin of the Otter Neck Group is associated with a pronounced, positive magnetic anomaly.

Mafic gabbros are composed of laths of plagioclase and subhedral actinolite grains up to 5 mm in length with interstitial microcrystalline actinolite, albite and chlorite.

Coarse (1 mm), iron-titanium oxides are an important component of these gabbros (Fig. 7.1). These are replaced by fine-grained, granular sphene but retain the original, euhedral shape of the primary oxide phase which is probably titanomagnetite. Thin plates of ilmenite occur within these pseudomorphs where they have exsolved along octahedral planes.

Leucocratic gabbro

Leucocratic gabbro outcrops along the northwestern shoreline of Glover Island and along the southeastern shoreline of Grand Lake. It consists of 2 mm long, dark-green, amphibole grains set in a white, plagioclase matrix. Subhedral plagioclase grains are locally present but most are anhedral.

Actinolite grains up to several millimetres in length are subhedral and pleochroic green. Simple twins are locally recognized. These may represent relict, igneous amphiboles that have re-equilibrated to actinolitic

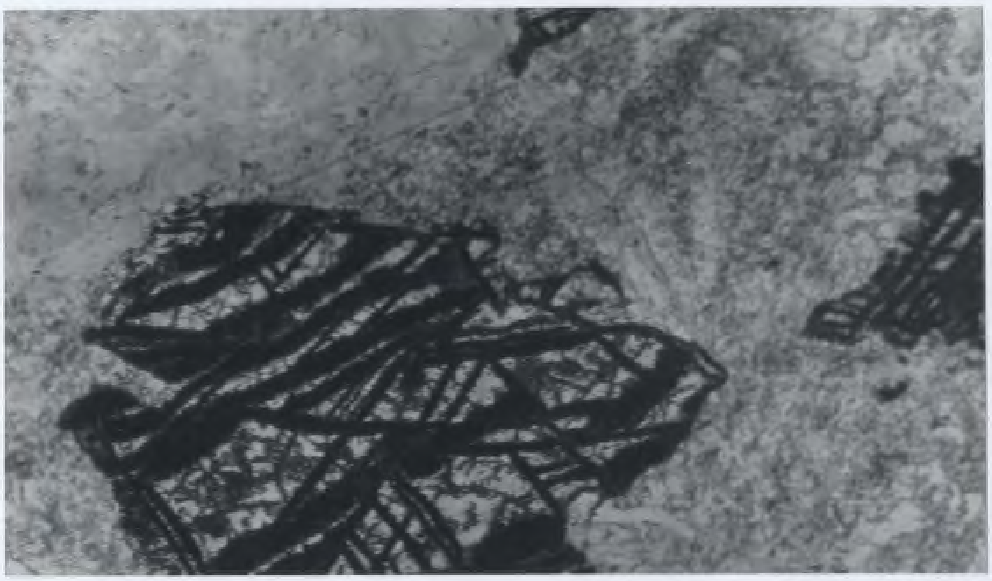


Figure 7.1 Photomicrograph of mafic gabbro within the Otter Neck Group. Relict iron-titanium oxides are defined by thin plates of ilmenite exsolved along octahedral planes within the original grain which has now been largely replaced; X10.

compositions. Plagioclase is invariably replaced by clinozoisite and albite.

DK-101 contains appreciable quartz as small blebs of polygonized grains and as myrmekite. It is likely this sample is of a metadiorite. Quartz is rare or absent in other samples.

Late dykes

Massive, aphyric, olive-weathering dykes are recognized which intrude basalt dykes of the Otter Neck Group along the northwest shoreline of Glover Island and along the southeast shoreline of Grand Lake. These dykes strike north, dip steeply and are clearly younger than dykes of the Otter Neck Group.

The olive-weathering dyke located along the north-western shoreline of Glover Island is composed of milli-metre-scale ophitic plates of a light purplish, weakly pleochroic clinopyroxene and subhedral laths of albitized plagioclase. Granular sphene pseudomorphs, after iron-titanium oxides, are scattered throughout the rock. The purplish colour of this clinopyroxene is distinctive and suggests it may be titaniferous.

A second dyke occurs along the southeast shoreline of Grand Lake and is interpreted also as a late intrusion. However all primary minerals in this dyke have been replaced by greenschist facies metamorphic minerals.

Intrusions within the Otter Neck Group

Many small intrusions cut the Otter Neck Group. These rocks are recognized only in this group and are described below.

Silicic porphyry

A number of small stocks or dykes of plagioclase-quartz phyric, silicic porphyry intrude the Otter Neck Group along the northwestern shoreline of Glover Island. These porphyries are irregular, generally less than 10 m across and contain abundant mafic xenoliths and hornblende xenocrysts.

In thin section they consist of equant phenocrysts of plagioclase and quartz up to 5 mm in size in an extremely fine-grained, quartz-feldspar groundmass (Fig. 7.2). Plagioclase phenocrysts are invariably replaced by epidote and albite. The original plagioclase phenocrysts were subhedral and generally composite. Quartz phenocrysts are abundant, equant and generally rounded in shape. Resorption features are present in places.

Locally these porphyritic rocks are congested with xenoliths consisting of fragments of hornblende, diorite, hornblendite and single grains of hornblende, epidote, allanite and apatite. These occur together with euhedral plagioclase phenocrysts and rounded, resorbed quartz phenocrysts.

a.

b.

Figure 7.2 Silicic porphyry intrusion within the Otter Neck Group. (a) Euhedral plagioclase phenocrysts; X10. (b) A partially resorbed quartz phenocryst; X15.

Trondhjemite

A small unit of trondhjemite intrudes the Otter Neck Group at its northernmost extent. This trondhjemite is associated with a conspicuous intrusion breccia at its base in which angular blocks of basalt up to a metre in size occur in a medium-grained trondhjemite matrix. It is unconformably overlain by the Red Point Formation.

The trondhjemite is medium-grained (2 mm) and white-weathering. Fine-grained, sugary-textured, aplitic dykes cut the trondhjemite locally and are also truncated by the overlying unconformity.

A second trondhjemite occurs along the southern shoreline of Grand Lake within the Otter Neck Group and others are known to be present several kilometres inland from this point (P. Dean, personal communication). A thin metamorphic aureole is associated with this intrusion.

The trondhjemite consists of 1-4 mm, equant, grains of plagioclase, quartz and fine-grained chlorite, epidote and sericite. Equant, subhedral or anhedral grains of plagioclase and anhedral grains of quartz make up most of the rock with aggregates of chlorite and epidote ranging up to 10 modal percent. Chlorite and epidote are secondary and replace the primary mafic mineral phase. Plagioclase is invariably clouded with inclusions of clinozoisite and sericite but has relatively clear rims. Quartz is weakly to strongly undulose and has sutured grain boundaries.

Mafic porphyry

A plagioclase-phyric mafic dyke intrudes the Otter Neck Group along the northwestern shoreline of Glover Island. This porphyry is quite distinct due to its abundant plagioclase phenocrysts. It outcrops in close proximity to silicic porphyries with which it may be related.

Plagioclase pseudomorphs of clinozoisite and albite dominate the texture of this mafic porphyry. These occur in a microcrystalline groundmass composed of actinolite, albite and chlorite.

GEOCHEMISTRY OF THE OTTER NECK GROUP

Ten samples from the Otter Neck Group were analyzed. One of these is a pillow basalt (DK-741). The rest are of various intrusive rocks. These include mafic and silicic porphyry (DK-85, 69, 1352, 89, 86), leucocratic gabbro (DK-1351, 101) and mafic gabbro (DK-742, 76). The chemistry of these rocks is described below and given in Table 7.1. Sample locations are given in Plate II.

Descriptions of chemical analyses

DK-743

DK-743 is an aphyric, fine-grained, black rock. It is composed of microphenocrysts of albitized plagioclase in a fine-grained, variolitic-textured groundmass of

Table 7.1 Major element and trace element chemistry of rocks of the Otter Neck Group.

	1	2	3	4	5	6	7	8	9	10
	743	1351	101	742	76	85	69	1352	89	86
SiO_2	54.4	51.5	50.9	50.2	48.3	45.2	74.2	73.6	71.7	64.5
TiO_2	1.89	.42	.88	2.17	2.22	.98	.35	.33	.25	.38
Al_2O_3	13.3	14.9	15.3	13.3	14.2	21.3	12.7	13.2	13.7	15.4
$Fe_2O^t_3$	13.23	10.56	9.36	15.14	14.34	7.51	3.07	3.23	2.69	3.60
MnO	.16	.16	.17	.24	.20	.13	.05	.04	.04	.07
MgO	3.77	7.48	6.67	4.44	5.89	6.18	1.06	.70	.60	2.24
CaO	5.30	9.75	9.31	8.00	9.38	11.40	2.90	3.35	3.50	4.51
Na_2O	5.06	2.22	2.73	2.74	3.56	2.37	4.67	4.69	4.79	5.48
K_2O	.11	.21	.22	.68	.18	.11	.03	.10	.13	.54
P_2O_5	.17	.06	.07	.24	.36	.01	.03	.04	.02	.17
L.O.I.	1.00	2.72	2.44	1.49	2.11	3.20	1.18	1.32	1.03	1.65
Total	98.39	99.98	98.05	98.64	100.74	99.39	100.24	100.60	98.45	98.54
Nb	4	0	0	3	2	0	1	1	1	1
Zr	174	29	79	155	117	67	117	139	146	104
Y	73	17	28	66	52	23	38	47	49	12
Sr	120	243	210	129	136	316	106	206	188	773
Ga	11	9	14	16	14	13	16	18	15	10
Ni	8	40	53	19	46	73	19	16	19	14
Cr	0	321	89	0	0	197	52	0	0	58
V	402	230	270	401	508	176	54	30	30	81
La	18	0	2	13	9	4	47	55	53	47
Ce	15	1	0	9	2	6	3	4	3	20

1 Variolitic pillow lava
2,3 Massive leucogabbro
4,5 Hornblende gabbro
6 Basalt porphyry
7,8,9, Rhyolite Porphyry
10 Contaminated Rhyolite Porphyry

albite and green, pleochroic amphibole. The variolitic texture is defined by radiating, intimate intergrowths of albite and amphibole and is very well developed in this rock.

In composition this rock is a basaltic andesite showing pronounced iron enrichment towards an icelanditic composition. Total iron is high and aluminium and magnesium low, consistent with a tholeiitic differentiation trend. It is enriched in the light rare earths and strongly depleted in chromium.

DK-101, 1351

These rocks consist of coarse-grained (2-3 mm) massive, leucogabbro or diorite composed of equant aggregates of albite and epidote pseudomorphic after plagioclase and actinolite. DK-101 contains quartz and myrmekite.

DK-1351 has unusually low abundances of incompatible trace elements and titanium. It may represent a primitive composition or may be diluted somewhat by clinopyroxene accumulation as it is marginally higher in chromium, calcium and magnesium.

DK-742, 76

DK-742 and DK-76 are coarse-grained (5 mm), black mafic gabbros. They consist of albitized laths of plagioclase and subhedral actinolite grains up to 5 mm in length.

Fine-grained granular sphene aggregates replace iron-titanium oxides.

Both are chemically similar gabbros with high iron, titanium, incompatible trace elements and vanadium, typical of a strong, iron-enrichment tholeiitic trend. Chromium is strongly depleted in these gabbros.

DK-85

DK-85 is a light green, strongly porphyritic metabasalt with abundant pseudomorphs after plagioclase. It intrudes the sheeted dyke complex along the northwestern shore of Glover Island.

In composition, DK-85 reflects the modal abundance of plagioclase phenocrysts in having high aluminium, calcium and strontium. Incompatible trace elements are likewise low also suggesting some plagioclase accumulation.

DK-69, 1352, 89, 86

These rocks are all leucocratic, quartz-plagioclase phyric, silicic porphyries that cut the Otter Neck Group along the northwestern shoreline of Glover Island. Silica is over 70 weight percent in these rocks and soda about 5 weight percent. Potash is absent and may reflect metasomatic alteration to a keratophyric composition.

DK-86 is anomalous in that it is dacitic in composition. However this rock is choked with amphibole xenocrysts and xenoliths and is clearly contaminated. It is assumed to have originally been more silicic in composition as are the other porphyries.

Graphical representation

Rocks of the Otter Neck Group are plotted in various standard diagrams in Figure 7.3 and 7.4. The use of these diagrams is discussed in Chapter 6 and the results summarized below.

P_2O_5 - Zr Diagram. (Fig. 713)

Mafic rocks of the Otter Neck Group are plotted in a phosphorus-zirconium diagram in Figure 7.3. DK-1351, 101 and 85 plot well within the subalkaline field and clearly show no alkaline affinities. DK-743 and 742 plot closer to the field boundary and DK-85 plots within the alkaline field.

SiO_2 - FeO^T / MgO Diagram (Fig. 7.4)

Geochemical data for the Otter Neck Group shows a wide scatter in this figure. DK-76, 742 and 743 all plot well within the tholeiitic field and encompass a range of silica values. The mafic porphyry, DK-86, also plots within the tholeiitic field but more towards the calc-alkaline portion of this field. Leucocratic gabbros plot well within the calc-alkaline field as do the

Figure 7.3 P_2O_5 - Zr Diagram for rocks of the Otter Neck Group. Alkaline field boundary is from Winchester and Floyd (1976). Numbered data points refer to Table 7.1.

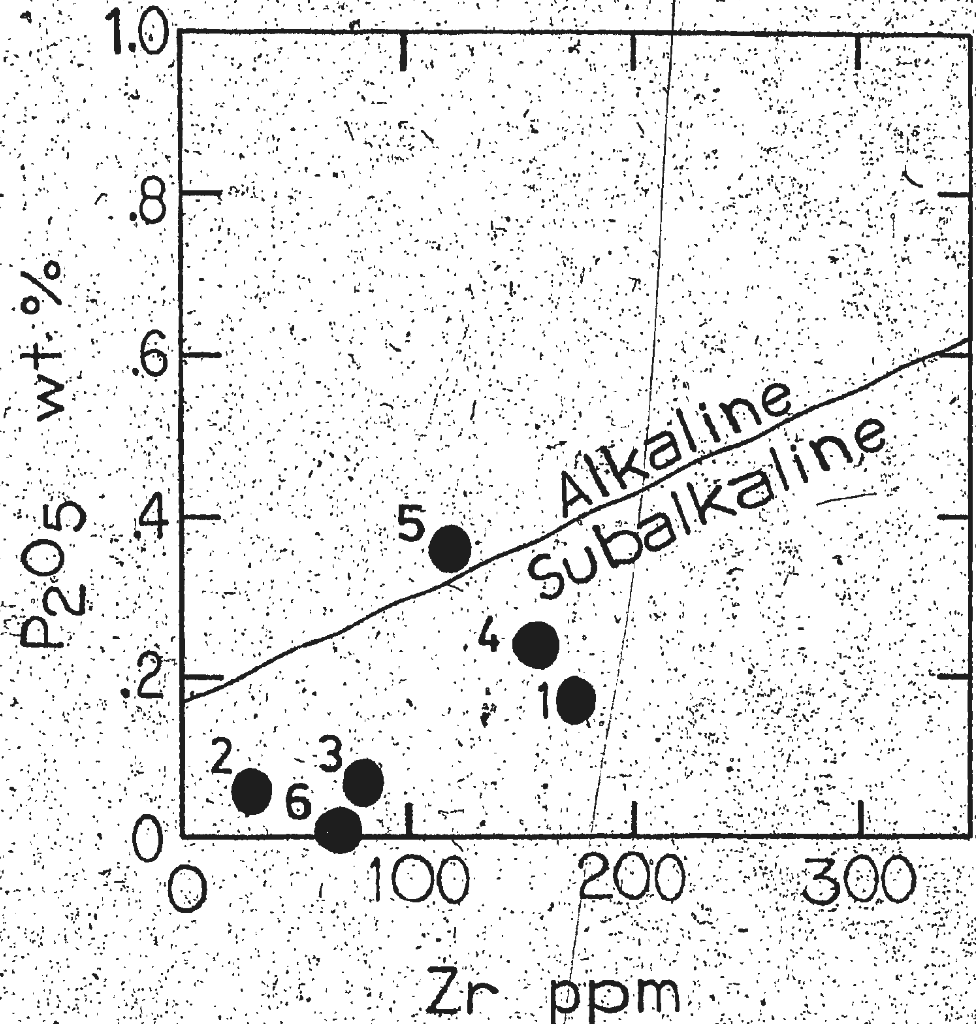
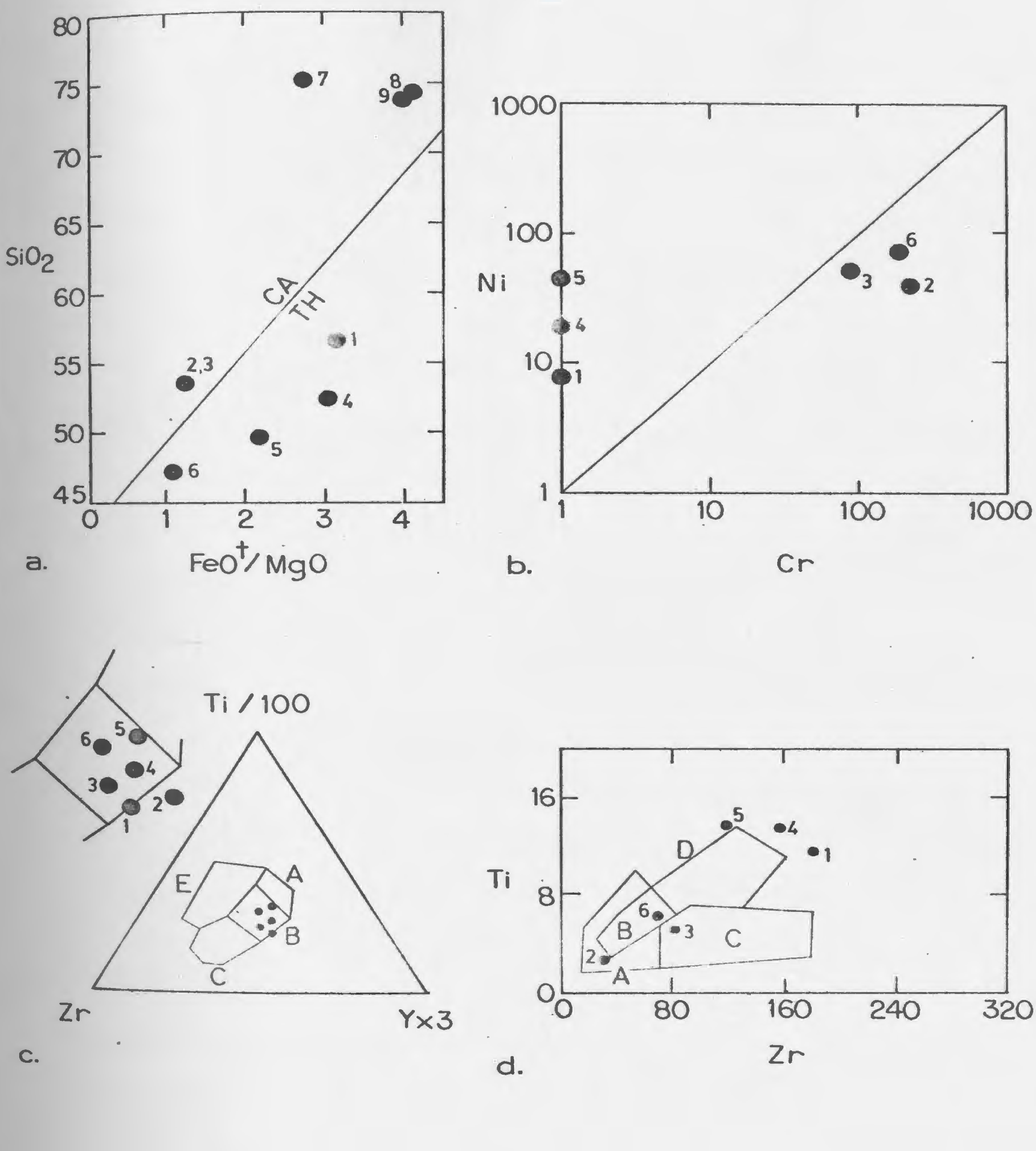

1.0
.8
.6
.4
.2
0
P_2O_5 wt.%
0
100
200
300
Zr ppm
Alkaline
Subalkaline
1
2
3
4
5
6

Figure 7.4 Chemistry of the Otter Neck Group. Numbered points refer to analyses given in Table 7.1. Figure a is after Miyashiro (1974) and Figures c and d are after Pearce and Cann (1971, 1973). Analysis 1 is of a basaltic andesite (DK-743) and is included in Figures c and d for comparison purposes only as the field boundaries in these diagrams do not apply to andesitic rocks.


SiO2
FeO^t/MgO
CA
TH
a.
Ni
Cr
b.
Ti / 100
Zr
Y×3
c.
Ti
Zr
d.

silicic porphyries.

Ni/Cr Diagram (Fig. 7.4)

In this diagram, rocks of tholeiitic affinity are clearly separated from those of calc-alkaline affinity defined by the silica - iron/magnesium diagram described above.

Ti/100 - Yx3 - Zr Diagram (Fig. 7.4)

Analyses are widely scattered in this diagram. It is clear that tholeiitic rocks have similar zirconium-yttrium ratios but a range of titanium values. DK-135 plots well away from the other leucogabbro, DK-101, which plots closer to DK-85. This anomaly may reflect the very low abundances of zirconium and yttrium in DK-1351. These low abundances greatly increase the relative error and may account for the discrepancy between DK-101 and DK-1351 in this diagram which otherwise generally plot quite close to one another.

All analyses plot within the ocean floor basalt field, however it is important to note that DK-743 is an andesite and cannot be plotted in this figure which pertains to basalts only.

Zr - Ti Diagram (Fig. 7.4)

As in the ternary trace element diagram, tholeiitic rocks cluster together at high zirconium and titanium values in this diagram. DK-85 and DK-101 also cluster together and are separated from DK-1351 by a substantial gap.

Mafic gabbros are strongly enriched in incompatible trace elements and plot outside the field of normal basalts. DK-743 also plots outside the field of normal basalts, but because it is andesitic these fields do not apply to it. The leucocratic gabbros and mafic porphyry plot within the low-K tholeiite field, calc-alkaline basalt field and within the field of overlap.

Interpretation of geochemistry

The Otter Neck Group consists of pillow basalts and dykes intruded by a variety of gabbros and porphyries. A single analysis of pillow lava is strongly tholeiitic basaltic andesite.

Intrusive rocks can be divided into two major groups based upon major and trace element chemistry. Mafic gabbros are strongly iron-enriched tholeiites that resemble the pillow basalt in many respects. It is possible they are comagmatic but no conclusion can be made based upon this scanty information.

A second major group of rocks that intrude the Otter Neck Group consists of leucocratic gabbros, basaltic porphyry and silicic porphyry. These rocks may be related as all show calc-alkaline affinities based upon the degree of iron enrichment, low incompatible trace element abundances and similar nickel-chromium ratios.

Mafic gabbros all plot close to the ocean floor basalt field but out of the field of normal basalts. Because they intrude pillow lavas their age is not well known. They may be comagmatic with the pillow lavas or they may be much younger.

Other intrusive rocks in the Otter Neck Group have degrees of iron-enrichment typical of calc-alkaline basalts or low-K tholeiites based upon the silica - iron/magnesium diagram. Trace element abundances are also rather low and fall in low-K tholeiite and calc-alkaline fields on standard diagrams. Based upon this data it is concluded that these rocks may be arc-related and strongly contrast with the analysis of pillow lava they intrude. Mafic gabbros are also highly tholeiitic and may or may not be related.

STRUCTURE AND METAMORPHISM

The Otter Neck Group is structurally simple as a single foliation is recognized which defines D1. S1 is very heterogeneous. In massive units of greenstone no foliation is apparent whereas in more tuffaceous lithologies S1 is well developed. No F1 folds are recognized.

Synkinematic metamorphism during D1 reached middle greenschist facies conditions defined by actinolite, epidote, albite, clinozoisite, sphene and chlorite, typical of the greenschist facies. Pseudomorphs are common with plagioclase laths and pyroxene replaced by albite and actinolite, respectively.

CONTACT RELATIONS

The Otter Neck Group is faulted against the Tuckamore Formation along its western and eastern margins. These faults are vertical and abrupt with little or no fault zone breccia developed. Folding and contorted beds occur along the fault zones and are inferred to have formed during dislocation. The direction and amount of movement along these faults is unknown.

The contact between the Otter Neck Group and the Red Point Formation is an unconformity developed on a trondhjemite that intrudes the Otter Neck Group. Clasts of trondhjemite occur in the overlying Red Point Formation along the contact.

INTERPRETATION

The Otter Neck Group consists of pillow lavas locally rimmed by black radiolarian cherts, pillow breccias and aquagene tuffs. Aquagene tuffs contain spheroids interpreted to represent varioles. Amygdules are not present.

No evidence of pyroclastic rocks is observed and silicic tuffs are absent.

The environment of formation of this sequence of rocks is inferred to be a deep water, oceanic environment. Lack of amygdules suggest deep water volcanism (Moore, 1965) and radiolarian cherts, pillow breccias and aquagene tuffs with no pyroclastic rocks indicates that volcanism was wholly submarine. Sheeted dykes locally occur within this unit and are interpreted to be feeder dykes to pillow lavas.

Pillow lavas are strongly tholeiitic and highly fractionated based upon a single analysis. Similar rocks are known to occur in an oceanic setting (Bryan et al., 1976) thus supporting an oceanic origin for the Otter Neck Group.

The occurrence of sheeted dykes in the Otter Neck Group together with evidence of a deep water, oceanic origin suggests the Otter Neck Group may represent the upper part of an ophiolite sequence. Sheeted dykes, aquagene tuffs and radiolarian chert are typical of many ophiolites. The Otter Neck Group is accordingly interpreted to be ophiolitic.

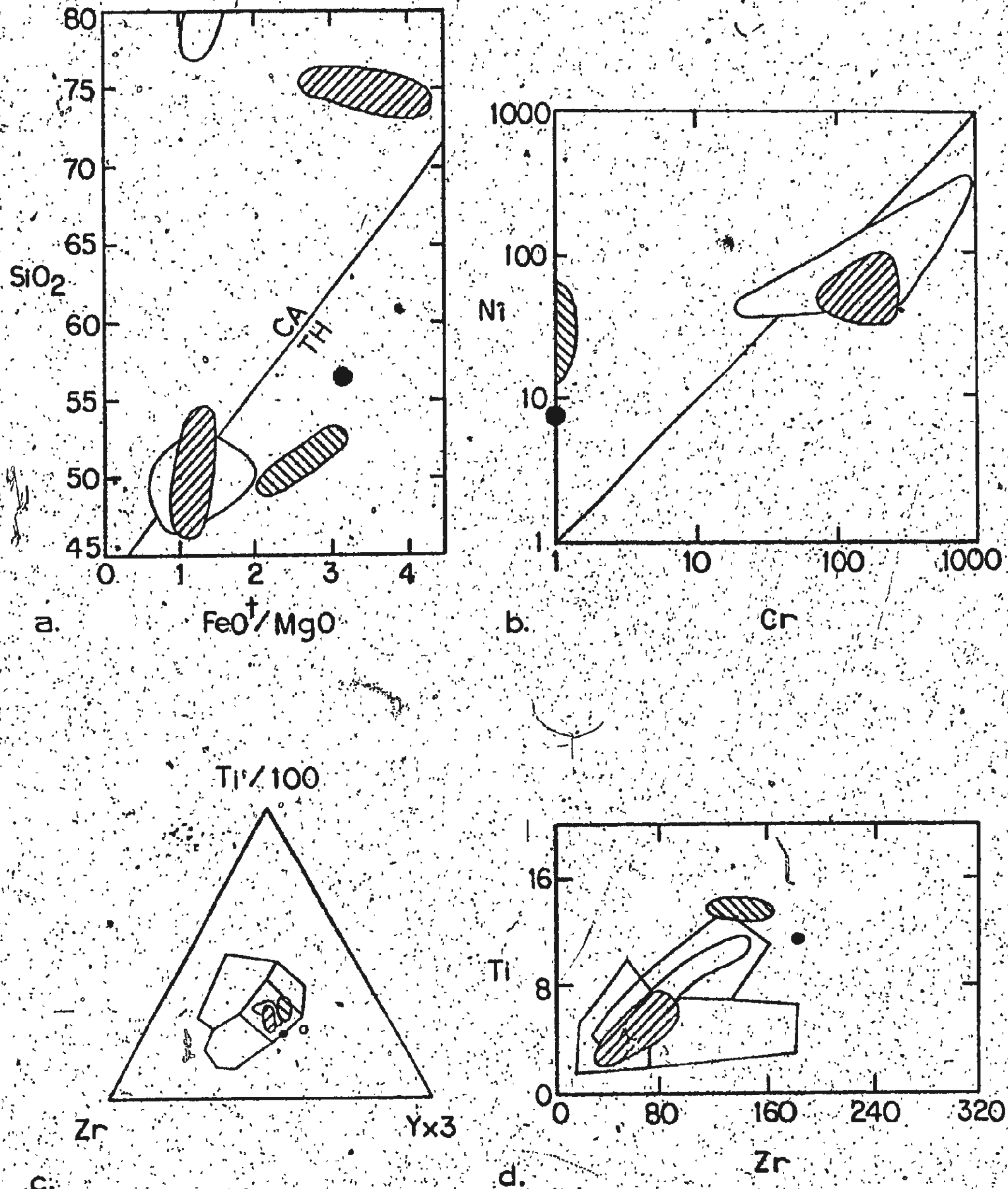
Intrusive rocks of the Otter Neck Group form two separate compositional groups. Mafic gabbros are strongly tholeiitic and iron enriched and may be comagmatic with the pillow lavas as they are chemically similar. Mafic and silicic porphyries and leucogabbro contrast with these rocks as they are less iron-enriched and have calc-alkaline affinities. It is concluded they are unrelated to tholeiitic rocks of the Otter Neck Group and relate to a younger episode of calc-alkaline volcanism.

CORRELATION

Non-tholeiitic intrusive rocks of the Otter Neck Group are correlated with volcanic rocks of the Tuckamore Formation based upon strong chemical and petrographic similarities. The composition of these rocks is compared in Figure 7.5 where it is clear that a strong positive compositional correlation exists indicating the two may be related in some way. Petrographic types are also comparable. Silicic tuffs of the Tuckamore Formation are petrographically identical to silicic porphyries of the Otter Neck Group. A major difference between the two groups is that the Tuckamore Formation is exclusively extrusive. This discrepancy is attributed to having sampled the same volcanic pile at different structural levels and therefore does not weaken the proposed correlation. Intrusions within the Otter Neck Group are interpreted to represent the source of an overlying volcanic pile represented by the Tuckamore Formation.

Mafic gabbro intrusions of the Otter Neck Group are compositionally similar to analysis NH-3 of the Tuckamore Formation. These may represent a single suite of highly

Figure 7.5 Chemistry of the Otter Neck Group and Tuckamore Formation compared. Unornamented field encloses analyses of the Tuckamore Formation while the ruled fields enclose analyses of the Otter Neck Group. The solid dot corresponds to a basaltic andesite pillow lava (DK-743) from the Otter Neck Group. It is included in Figures c and d for comparison purposes only as these diagrams do not apply to andesitic rocks. The open circle in Figure c is an anomalous point from the Tuckamore Formation.



80
75
70
65
60
55
50
45
SiO2
CA
TH
0
1
2
3
4
a.
FeO†/MgO
1000
100
10
1
Ni
1
10
100
1000
b.
Cr
Ti/100
Zr
Yx3
c.
16
8
0
Ti
0
80
160
240
320
Zr
d.

fractionated mafic gabbros.

SIGNIFICANCE

The correlation proposed above requires the Otter Neck Group together with its younger calc-alkaline intrusions to form an ophiolitic basement to a thick pile of extrusive volcanic rocks represented by the Tuckamore Formation. The Glover Group is known to overlie the Grand Lake Complex along its western margin thus requiring the Grand Lake Complex and Otter Neck Group to occupy the same stratigraphic position. This geometry supports the interpretation of the Otter Neck Group as ophiolitic as it occurs in the familiar position of basement to the Glover Group.

It also implies a relationship between the Grand Lake Complex and the Otter Neck Group. Yet these two ophiolites contrast with one another. The Grand Lake Complex is highly deformed and has been partially eroded during exhumation. The Otter Neck Group is only weakly deformed and has clearly not suffered the same exhumation history.

This contrast is known elsewhere along the Baie Verte - Brompton Line. Ophiolites along the Line are invariably highly deformed and associated with coarse clastic rocks. Those off the Line are generally more complete and well preserved. A classic example occurs at Baie Verte where the Advocate Complex, a highly deformed ophiolite marks the Line and only kilometres away the Point Rousse Complex is a nearly intact ophiolite. Farther east there is a second well preserved ophiolite, the Betts Cove Ophiolite. These "off Line" ophiolite complexes are in each case remarkably well preserved while those "on Line" are highly disrupted, deformed and associated with coarse clastic rocks. Accordingly the Grand Lake Complex and Otter Neck Group are interpreted as examples of "on Line" and "off Line" ophiolites, respectively. They serve to strengthen observations relating to the Baie Verte - Brompton Line elsewhere and further confirm the conclusion that the Baie Verte - Brompton Line is present on Glover Island.

AGE

The Otter Neck Group is interpreted as Lower Ordovician based upon the observation that other ophiolites in western Newfoundland are Lower Ordovician as summarized in Chapter 5. Also based upon correlation of the Red Point Formation with the Lower Ordovician Corner Pond Formation, the Otter Neck Group can be no younger than Lower Ordovician.

Chapter 8

PARALLOCHTHONOUS ORDOVICIAN CLASTIC ROCKS:

The Red Point and Corner Pond Formations

The Red Point and Corner Pond Formations consist of coarse metaclastic rocks unconformably overlying trondhjemite of the Otter Neck Group and mafic volcanic rocks of the Glover Group, respectively. Graded beds, coarse conglomerates and minor volcanic rocks are common to both. Each occurs in a single, isolated, synclinal core, the Red Point Formation at the southern end of Glover Island and the Corner Pond Formation at Corner Pond. They are interpreted to have once formed a single, continuous unit.

THE RED POINT FORMATION

The Red Point Formation is named after a small promontory located along the northwest shoreline of Glover Island at its southern end named Red Point by the author (Plate I). This promontory is also designated as the type locality.

The Red Point Formation unconformably overlies trondhjemite that cuts the Glover Group. Its areal extent is restricted to the core of a north-plunging syncline which is bounded to the northwest by Grand Lake (Plate I; Fig. 8.1). Its total areal extent is thus limited to a narrow strip along the northwest shoreline of Glover Island about 2 km long and

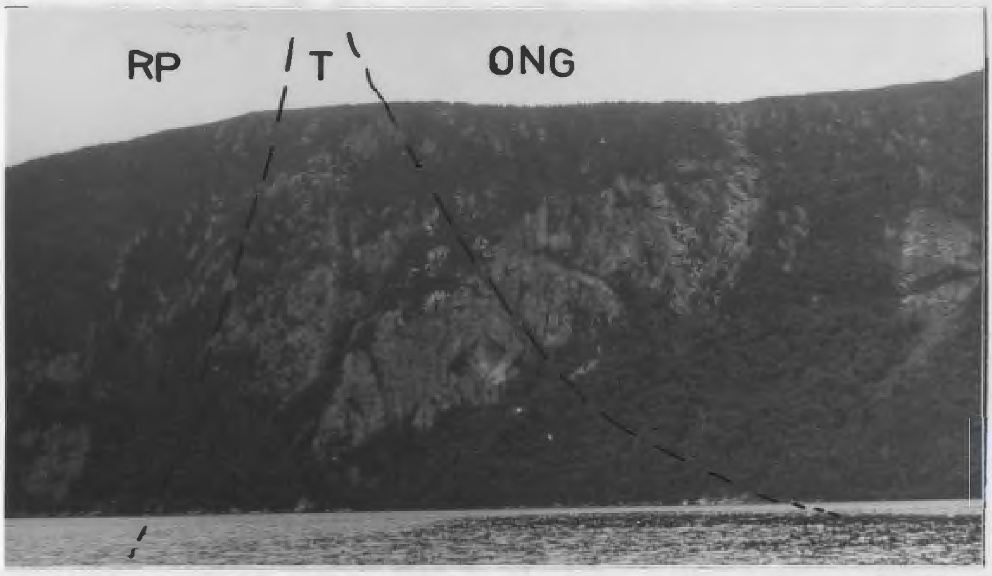


Figure 8.1 Photograph of the hillside and shoreline immediately southwest of Red Point which is located at the extreme left margin of this photograph. Abbreviations are as follows: RP, Red Point Formation; T, Tonalite, ONG, Otter Neck Group.

.5 km wide. Preservation of the Red Point Formation is undoubtedly due to the fact that it occupies the core of a syncline. The rest of the formation has been removed by erosion.

Thickness of the Red Point Formation is impossible to determine as its top is not exposed. It has a minimum thickness of about 200 m in the section north of Red Point.

Lithology

The Red Point Formation consists of purple phyllite, metagreywacke, grey sandstone and metaconglomerate. Distinctive in outcrop, the purple phyllite consists of sericite, quartz and plagioclase with abundant disseminated small grains of hematite giving the phyllite its purple colour. The metagreywacke consists of millimetre-sized quartz and plagioclase in a purple phyllitic matrix. Sandstones and conglomerates are also abundant in this formation. These lack the purple phyllitic matrix and weather tan to white depending upon the relative proportion of detrital, plagioclase-bearing rock fragments. They are all light green on freshly broken surfaces.

Graded beds are locally well preserved. At Red Point, spectacular, metre-thick graded beds occur along the shoreline of Glover Island (Fig. 8.2). Undoubtedly the preservation of these graded beds is due to their structural position in the hinge region of a major fold. Shear strain is minimized in such regions thus preserving the sedimentary structures.

The bases of the graded beds at Red Point consist of white-weathering, green, pebble metaconglomerate containing a variety of quartz and feldspar-bearing clasts. The conglomerates grade upward into a coarse greywacke with a purple phyllitic matrix and at the top of the greywacke into purple phyllite. Individual beds range between .3 to 1 m in thickness and contain spectacular load casts at their base. All graded beds are right side up at Red Point.

Boulders of Red Point conglomerate contain well-preserved cross-bedding accentuated by thin, purple, phyllitic laminae at Red Point (Fig. 8.3). Cross-bedding was not observed in outcrop, however.

An unusual lithology interpreted to be a tuff is locally exposed in the stream valley along the high angle fault which transects the Red Point Formation. It consists of a grey, fine-grained greywacke with distinctive elongate blebs of dark-green chlorite. The flattened ellipsoidal blebs are elongate down-dip 2-6 cm and are 1-2 cm wide in the plane of the schistosity. They are interpreted to be deformed, devitrified, fragments of pyroclastic material and the rock, a reworked tuff or agglomerate. This is the only evidence of volcanism recognized in the Red Point Formation.

Figure 8.2 Photograph of the Red Point Formation at Red Point. Bedding is gently inclined to the left. Cleavage (S1) is steeply inclined to the left. Light coloured layers are feldspar-rich conglomerates at the base of graded beds. The thickest light coloured layer is approximately 1 m thick for scale.

Figure 8.3 Arkosic metasandstone block at Red Point. Cross bedding is defined by thin, purple phyllitic laminae.

Petrology

Greywackes in the Red Point Formation are composed of quartz, sericite, muscovite, epidote, sphene, plagioclase and opaques in varying proportions. Quartz, epidote, sphene, muscovite and rock fragments are all detrital in origin. Quartz ranges up to 5 mm in diameter and rock fragments up to 3 cm. Epidote, sphene and muscovite grains are generally .5 mm or smaller in diameter. All are anhedral and rounded or angular in shape. The strong contrast in size and the irregular, angular and locally rounded shapes of quartz, epidote, sphene and muscovite attest to their origin as detrital grains. This interpretation is supported by the common occurrence of rock fragments of similar size together with these detrital grains.

Structure

Structures in the Red Point Formation consist of an axial planar, single regional schistosity and a cogenetic stretching lineation. S1 of the Red Point Formation consists of a penetrative regional schistosity defined by sericite, flattened relict grains of quartz and quartz pressure shadows formed around quartz grains. The schistosity is primarily defined by sericite but contributing to it are coarse (1-4 mm) grains of detrital quartz flattened within the schistosity which contain prominent pressure shadows of quartz.

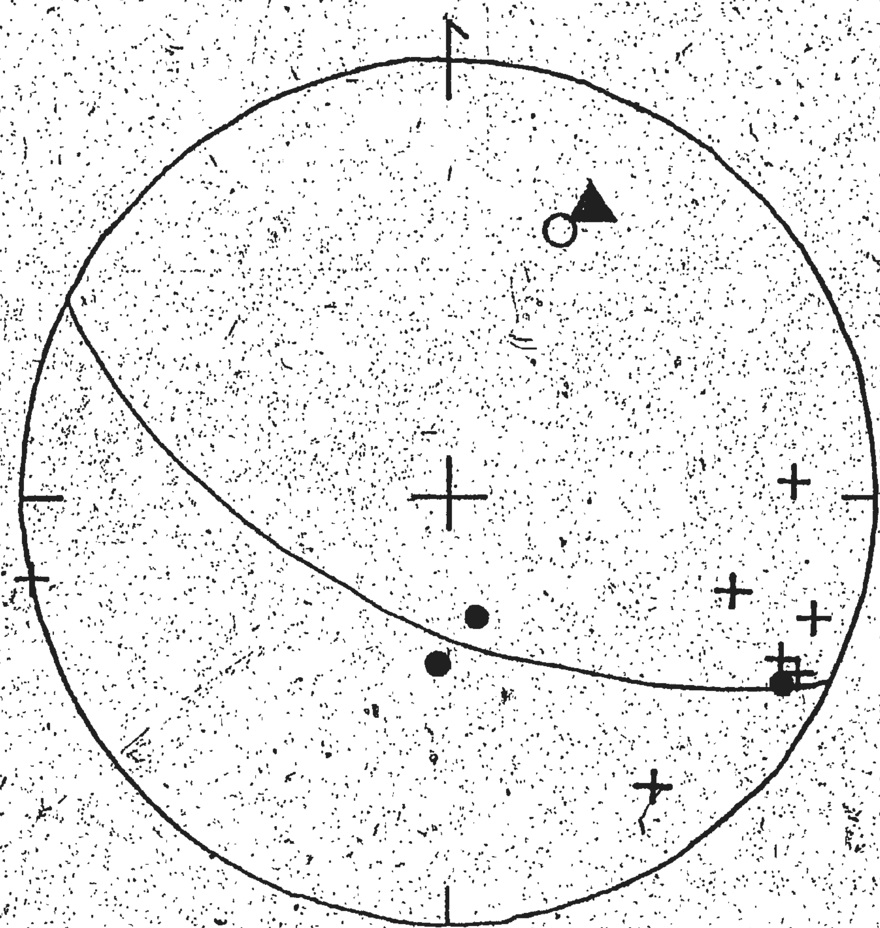
The Red Point Formation is folded into a major, open, north-plunging syncline which largely controls the areal distribution of the Red Point Formation. This major fold is interpreted to be F1 in age as S1 is geometrically axial planar to it. The hinge of this fold is exposed at Red Point and the eastern limb is exposed along the shoreline of Glover Island.

Fold axes related to this major fold plunge to the north. They reflect the pre-D1 orientation of bedding relative to S1 and do not have kinematic significance. Chlorite blebs in a green greywacke are interpreted to represent reworked pyroclastic material. These chlorite blebs are highly flattened and elongated down dip. This down dip stretching lineation defines the direction of maximum elongation during D1 in rocks of the Red Point Formation. Lack of coincidence between this lineation and F1 fold axes reflects the low strains associated with D1.

Structural geometry

Poles to S1, L1 and poles to bedding are shown plotted in equal-area, lower hemisphere projections in Figure 8.4. Poles to bedding define a great circle girdle inclined to the south striking roughly northwest. Poles to S1 form a diffuse point maximum inclined gently to the southwest.

Figure 8.4 Poles to bedding (solid dots) and poles to S1 (crosses) of the Red Point Formation plotted in equal-area, lower hemisphere projections. The great circle represents a visually estimated, best-fit, great circle girdle to poles to bedding. The pole to this great circle is shown as a solid triangle and an F1 fold axis is plotted as an open circle.



The great circle girdle defined by poles to bedding in Figure 8.4 is interpreted to be the result of F1 folding. S1 forms a point maximum indicating this area is homogeneous with respect to S1 orientation and has not been significantly deformed by later deformational events.

F1 fold axes and a downdip stretching lineation are oriented at high angles to one another. This indicates that bedding and S1 intersect in a line oriented at a high angle to X of the finite strain ellipsoid for D1. That the fold axes are not rotated towards X suggests D1 was not a high strain event.

Metamorphism

D1 is inferred to have occurred during lower greenschist facies conditions based upon the synkinematic growth of sericite and chlorite which define S1.

Contact relations

The base of the Red Point Formation is exposed on the slope above Red Point where it unconformably overlies a medium-grained, massive, plagioclase-quartz-chlorite trondhjemite locally intruded by comagmatic dykes of microtrondhjemite (Fig. 8.1). Chlorite is secondary after biotite or amphibole and plagioclase is ubiquitously sericitized. The trondhjemite is tabular or sheetlike in shape and vertical in orientation. It is roughly 300 m in thickness at its southern end and thins to the north presumably due to truncation along the base of the Red Point Formation.

The trondhjemite is massive in appearance throughout most of its areal extent. Within 2 m of the Red Point Formation contact, the trondhjemite is massive but contains hematite finely disseminated along cracks and grain boundaries. The hematite is undoubtedly derived from the overlying hematitic, purple metasediments of the Red Point Formation. About 1 m from the contact the typically massive trondhjemite is schistose. The contact between the schistose trondhjemite and overlying purple phyllitic metasediments is abrupt and represents the actual unconformity. The schistose margin of the trondhjemite is certainly due, in part, to a strain gradient between the schistose metasediments and the massive trondhjemite. Hematite staining of the massive trondhjemite may be related to fluid migration at the erosion surface during its post-exhumation history.

Elsewhere the Red Point Formation directly overlies massive volcanics of the Glover Group. At the very northernmost limit of the Red Point Formation this contact plunges beneath Grand Lake. The actual contact is not exposed but graded beds of purple shale and green sandstone occur 2 m from massive volcanics. Without the distinctive intervening trondhjemite the unconformity resembles a stratigraphically conformable contact.

Correlation and age

Based upon lithic similarity the Red Point Formation is correlated with the Corner Pond Formation. Both constitute clastic sequences unconformably overlying volcanic basement with red slate beds at their base. Both are preserved in the cores of tight synclines interpreted to have once formed a single continuous unit.

The Red Point Formation is inferred to be of Lower Ordovician age based upon correlation with the fossil-bearing Corner Pond Formation.

THE CORNER POND FORMATION

The Corner Pond Formation is named after Corner Pond which is located approximately 7 km southeast of Grand Lake (Plate I). The name is a local name and has not yet been formalized.

The Corner Pond Formation outcrops in the core of a major, northeast-trending, gently-plunging syncline which extends for at least 7 km along strike and probably farther. This syncline is bounded to the east by a high-angle, reverse fault which locally overturns bedding. The western limb unconformably overlies pillow basalts of the Glover Group. Preservation of this unit is attributed largely to its structural position in a syncline. Its equivalents have been eroded elsewhere except at Red Point where similar rocks in another synclinal keel outcrop.

Type section of this formation is designated as the stream outlet in the northwest corner of Corner Pond. Over a kilometre long section through the formation is exposed at this locality, including the important basal part of the formation. Superb exposure of the upper part of the unit is present in a west-trending stream located 2 km north of Corner Pond.

Thickness of the Corner Pond Formation is difficult to estimate due to folding and because the top of the section is not exposed. Minimum thickness is estimated as 700 m based upon the section exposed at the type locality.

Stratigraphy and lithology

As the Corner Pond Formation has a complex internal stratigraphy, detailed descriptions of three well-exposed localities are given below: the type locality in Corner Pond Brook; the stream flowing into Corner Pond at its northeast corner; and a larger stream located 2 km north of Corner Pond.

Type section

The outlet to Corner Pond trends across bedding at Corner Pond and for several kilometres downstream. Exposure in the stream bed is good for over a kilometre including the critical lower part of the Corner Pond Formation, and it is designated the type locality.

The base of the Corner Pond Formation at its type locality consists of bright red and reddish-purple slates and phyllites unconformably overlying massive volcanics of the Glover Group. Red slates contain a single strong cleavage and no recognizable primary sedimentological features. They are overlain by purple pillow lavas with red interpillow material identical to the underlying red slates.

Overlying these red phyllites and purple pillow lavas are massive silicic volcanic rocks and tan sandstones. Bright red patches locally occur within these volcanic rocks due to concentrations of hematite.

These silicic volcanic rocks are overlain by massive, tan sandstones. Locally thin, argillaceous layers are present but these are generally subordinate.

Proximal to Corner Pond, undeformed tan rhyolite dykes and small sills intrude these metasediments. A crude banding is present in the intrusive rocks defined by concentrations of pink spheroidal varioles (Fig. 8.5). Layering of varioles is interpreted to have formed by devitrification about discrete nuclei which have been concentrated in layers by flow of the magma analogous to

Figure 8.5 Photomicrograph of spherulitic, silicic dyke that intrudes the Corner Pond Formation; X25.

flow banding in extrusive silicic rocks.

Northeast Corner Pond section

A similar suite of rock types outcrops at the opposite end of Corner Pond. Here a small stream drains into the northeast corner of Corner Pond. An Arenig graptolite fossil locality is located along the lower part of this stream. Northward the stream is situated along a fault zone. Dark green pillow basalts outcrop along the east bank while a variety of lithologies typical of the lower part of the Corner Pond Formation outcrop along the down-thrown west bank.

The stratigraphic position of different lithologies exposed in this stream cannot be determined due to the complicating effect of the fault zone along which they outcrop. However red, silicic volcanic rocks are well exposed in this stream making it a key locality.

Upstream of the black graptolitic shales described earlier are several minor outcrops of polymictic conglomerate cut by basaltic dykes. Farther upstream, light green pillow basalts are well exposed. These contain feldspar phenocrysts pseudomorphed by light-green aggregates of sericite. Rounded basalt clasts in red-shale matrix conglomerate are locally present in isolated blocks along the fault zone.

The west side of the stream consists of sharp, ragged outcrops and small cliffs of red silicic volcanics. These locally contain hematite-rich patches which weather to a brilliant red. Hematite is abundant along fracture surfaces and in clots within the volcanics. The red silicic volcanics are fragmental in origin and are interpreted to be lapilli tuffs.

Upstream from these red volcanics the stream veers to the east away from the fault zone. Plagioclase-phyric basaltic pillow lavas outcrop for several hundred metres in the stream at this point. These pillow basalts are part of the underlying Glover Group.

Red silicic volcanics are well exposed in this stream but are evidently not abundant elsewhere. They are overlain by basalts and a volcanic breccia to the west. The breccia has a red, hematitic matrix and may be related to red slate matrix conglomerates in the stream. Above these volcanic rocks, exposure is very poor and limited to an isolated outcrop of tan sandstone.

Stream section 2 Km North of Corner Pond

An east-trending stream located 2 km north of Corner Pond cuts across the entire Corner Pond Formation and has excellent outcrop exposure. Unfortunately the lower part of the formation is not exposed. The red slate unit at the base of the Corner Pond Formation is continuous, however, based upon abundant red slate float in a small stream that crosses the basal contact.

Rocks exposed in this stream consist predominantly of thinly bedded (4-10 cm) flysch alternating with thick (19 m) conglomerate horizons. The conglomerate beds are highly resistant and form ridges and numerous large erratic bounders. The flysch portion of the formation tends to form low, swampy areas. In areas of incomplete exposure virtually all outcrops and all loose float is conglomerate and it is tempting to overestimate the abundance of this lithology. In areas of more complete exposure such as in the stream, conglomerate forms a minor proportion (10-25 percent) of the formation.

An excellent exposure of conglomerate occurs in this stream 1 km east of where it crosses the base of the unit. Here a single, steeply-dipping, conglomerate bed 10 m thick outcrops in contact with thinly-bedded flysch. The conglomerate bed and the flysch beds are graded and face to the northwest.

The base of the conglomerate bed consists of 1-6 cm clasts in a matrix of 1-3 mm angular quartz and feldspar grains (Fig. 8.6). Locally clasts up to 10 cm in length are present. Most clasts are angular or poorly rounded. The small feldspar grains are sharply angular.

Clast types consist primarily of silicic, igneous extrusive rocks and cherts. The silicic igneous rocks

a.

b.

Figure 8.6 Conglomerate within the Corner Pond Formation. (a) Silicic volcanic clasts within a thick conglomerate bed. (b) Same conglomerate bed with an elongate, angular clast of chert.

consist of tan, green and pink porphyritic rhyolites and micrograníites. The chert clasts are grey, green and red and are generally sharp and angular.

The conglomerate bed grades upward into a homogeneous arkosic sandstone with 1 mm grains of quartz and feldspar. No break in sedimentation is recognized in this single bed although a faint colour banding is present at the top.

Elsewhere in this stream, flysch is abundant. It consists of thin beds, 4-10 cm in thickness, that are invariably graded. Plagioclase and quartz greywacke forms the base of these beds which grade upward into cherty black, grey or green slates. Load casts are locally conspicuous.

Argillaceous beds throughout the Corner Pond Formation are invariably rich in silica and have the luster of typical cherts. However it is clear these siliceous beds are turbidites and the silicification is therefore interpreted as secondary.

An outcrop of coarse conglomerate located near this stream (symbol F in Plate I) consists of unsorted, rounded and angular clasts of various igneous and sedimentary rocks up to 20 cm in size. These clasts are the coarsest recognized in the Corner Pond Formation and unusual in that they include carbonate clasts. A fossiliferous metre thick bed of limestone also occurs at this locality.

Structure

Structures present within the Corner Pond Formation include a penetrative regional schistosity (S1) axial planar to broad, open folds (F1) locally refolded near faults.

A penetrative, regional schistosity or slaty cleavage is recognized throughout the Corner Pond Formation. In conglomerate beds it is cryptic but in more argillaceous lithologies it is well developed. This cleavage is designated S1. It is defined by fine-grained chlorite and sericite and has no linear elements associated with it.

The overall distribution of rock units within the Corner Pond Formation indicates it is folded into a broad, open syncline. Red slates or phyllites and minor volcanic rocks occur at the base of the Corner Pond Formation and locally along the eastern margin indicating the overall configuration of the Corner Pond Formation is synclinal. This synclinal geometry is further complicated by faulting along its eastern margin.

A major, open, upright, north-plunging syncline is recognized in the central west half of this synclinal belt. S1 is axial planar to this syncline indicating these folds formed during S1.

In the eastern part of this syncline, cleavage and bedding are overturned to the west and inclined to the east. This folding event is not associated with any axial planar fabric. The close proximity of the major high angle fault which bounds the eastern extent of the Corner Pond Formation suggests this late folding may be fault related.

Structural geometry

Fabric elements recognized in the Corner Pond Formation consist of bedding and a single cleavage.

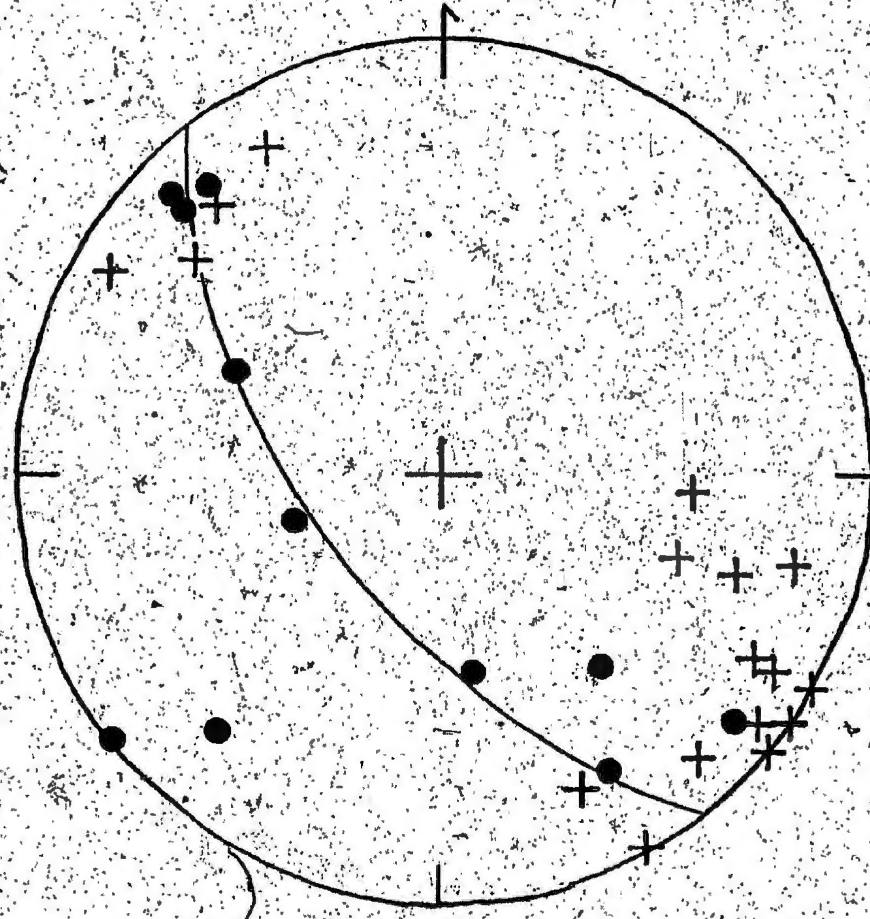
Poles to bedding and cleavage are plotted in equal-area, lower hemisphere projections in Figure 8.7 for rocks of the Corner Pond Formation. Bedding defines a single great circle girdle inclined steeply to the southwest and striking northwest. Poles to cleavage form a single horizontal point maximum.

The orientation of fabric elements in the Corner Pond Formation is interpreted to be the result of a single deformational event during which the single cleavage formed. This is supported by the orientation of cleavage axial planar to minor folds and from the data shown plotted in Figure 8.7. Bedding is displaced in a great circle distribution while cleavage forms a single point maximum consistent with a single, syncleavage folding event.

Metamorphism

Mafic rocks in the Corner Pond Formation are altered to calcite-chlorite-sericite and S1 in argillites is

Figure 8.7 Poles to bedding (solid dots) and poles to S1 (crosses) of the Corner Pond Formation plotted in equal area, lower hemisphere projections. Great circle girdle represents visually estimated best fit great circle girdle to poles to bedding.



defined by very fine-grained chlorite and sericite. These minerals are inferred to have formed syntectonically with D1.

Based upon the observed mineral assemblages of calcite, chlorite and sericite, metamorphism is inferred to have reached low greenschist facies conditions.

Contact relations

The Corner Pond Formation unconformably overlies mafic volcanic rocks of the Glover Group along its western margin. The contact is exposed at the type locality in Corner Pond Brook. Here red slates of the Corner Pond Formation are in sharp contact with green, massive volcanic rocks of the Glover Group. The contact is exposed in the stream bed and can be precisely located. It is interpreted to represent an unconformity based upon its abruptness and the lack of interbedding between pillow basalts and red slates.

Paleontology and age

The age of the Corner Pond Formation is based upon a fossil locality in a stream along the faulted eastern margin of the formation and from the interior part of the formation.

A metre-thick, limestone bed outcrops in the central part of the Corner Pond Formation north of Corner Pond

(Plate I). It clearly represents a bed and not a large limestone clast as it contains porphyritic clasts similar to nearby conglomerates and is laterally continuous.

During routine field work several samples of this limestone bed were collected. Partial digestion of this sample revealed it contained an abundance of macrofossils including brachiopods, bryozoans, ostracodes and corals (T. Bolton, G.S.C.). G. Nowlan processed the remaining sample together with additional material collected during 1981 and recovered the following forms:

Drepanodus sp. s.

Periodon flabellum (Lindstrom)

Protopanderodus rectus (Lindstrom)

Scolopodus n. sp. cf. *S. gracilis* (Ethington and Clark)

Teridontus ? sp.

The Colour Alteration Index (Epstein et al., 1977) of these specimens is 5.5-6.

Nowlan interprets the age of this sample to be mid to late Arenig.

A second fossil locality occurs in a stream located at the northeastern corner of Corner Pond. This fossil locality was discovered by Peter Dimmel in 1970 working with Noranda. Specimens collected by Dimmel and also by

F.D. Anderson (G.S.C.), at the same time were identified by R.B. Rickards of Cambridge. Forms identified by Rickards were reported in Dean (1976) and are listed below:

Tetragraptus fruticosis (Hall)

Tetragraptus pendens (Elles)

Tetragraptus reclinatus (Elles and Wood)

Loganograptus logani (Hall)

Sigmagraptus sp.

Didymograptus extensus (Hall)

Didymograptus constrictus (T.S. Hall)

Didymograptus similis (Hall)

Williams also collected at this same locality during the 1977 field season. D. Skevington (M.U.N.) identified two additional species in this collection which are given below:

Phyllograptus anna (Hall)

Didymograptus protobifidus (Elles)

This rich graptolite fauna is assigned to the *Didymograptus nitidus* zone by all workers indicating a mid Arenig age.

In spite of this extensive collecting, the geologic setting of this fossil locality has never been fully described. Because of the obvious importance of this fossil locality, the author visited it in 1980. The following descriptions are based upon this work.

Two isolated outcrops of fossiliferous black shale intruded by a tan dyke of rhyolite occur in a small stream that drains into the northeastern corner of Corner Pond (Fig. 8.8). Upstream this small brook follows a high angle fault that separates metaclastic rocks and metavolcanics of the Corner Pond Formation from green, pillow basalts. This fault projects west of these isolated outcrops of fossiliferous black shale which have been traditionally considered to be an integral part of the pillow basalt terrane. However as these fossiliferous shale outcrops are isolated and occur along a fault zone separating contrasting rock groups it is uncertain to which rock group they belong.

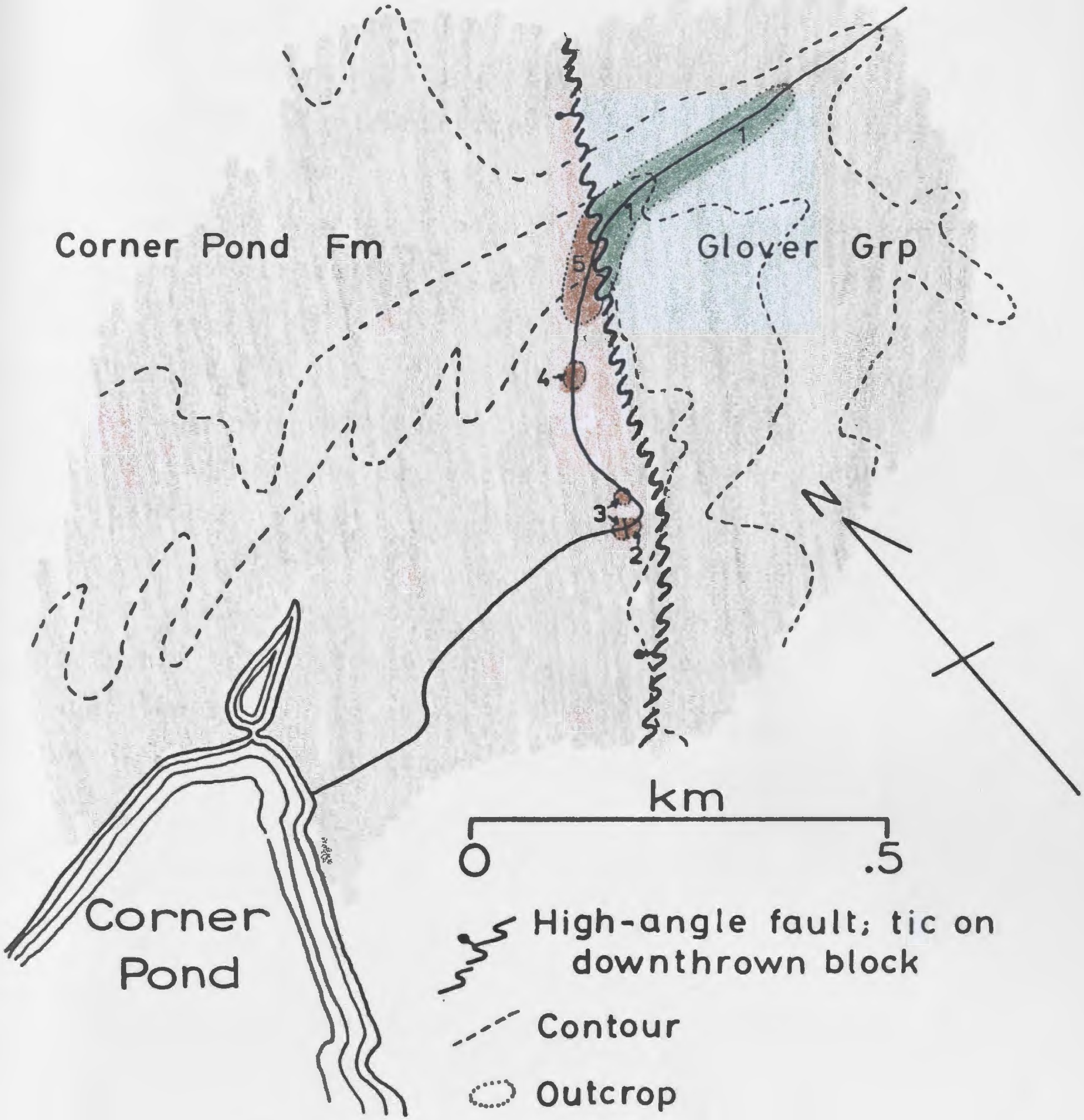
No black shales are recognized in volcanic rocks of the Glover Group in the Corner Pond area despite excellent exposure. In contrast, black shales are a common lithology in the Corner Pond Formation. Based upon lithic similarity and identical ages, these graptolitic black shales are inferred to represent part of the Corner Pond Formation and not the Glover Group as has been previously assumed.

The Corner Pond Formation is interpreted to be Mid to Late Arenig based upon these fossil localities.

Correlation

The Corner Pond Formation is correlated with the Red Point Formation based upon lithology and stratigraphic

Figure 8.8 Geologic map of the northeast corner of Corner Pond showing details of the Arenig graptolite fossil locality described in the text. Key: 1, pillow basalts; 2, aplite; 3, fossiliferous black shale; 4, conglomerate; 5, silicic volcanics, pillow lava and red shale matrix breccia.


Corner Pond Fm
Glover Grp
1
1
5
4
3
2
N
km
0
.5
Corner
Pond
High-angle fault; tic on
downthrown block
Contour
Outcrop

position. Both units unconformably overlie mafic volcanic rocks and consist of red slates or phyllites overlain by a sequence of coarse clastic rocks and minor volcanic rocks. They are interpreted to have once been continuous and are presently preserved because of their structural positions in synclinal keels.

Significance

The Corner Pond and Red Point Formations document the Lower Ordovician uplift and erosion of the volcanic basement terrane upon which they are deposited. Emplacement of ophiolitic rocks is known to have continued long after the Lower Ordovician indicating these clastic rocks formed during the emplacement process. They are therefore parallochthonous in origin and presumably formed during an episode of uplift related to the emplacement process.

Chapter 9

STRUCTURAL ANALYSIS

The structural sequence recognized in each major lithic group in the Glover Island Area has been presented in the preceding chapters. Each individual structural sequence is based upon cross-cutting relationships involving minor structures and the correlation of style groups. To derive a structural history applicable to the entire Glover Island Area, these individual structural sequences must be integrated into a single, common history by the correlation of structures between each different lithic group. The validity of the resulting structural history is wholly dependent upon the strength of these individual structural correlations.

In this chapter the structural sequences presented for each lithic group in the Glover Island Area are reviewed and correlated to yield a single structural history. The structural correlations upon which this history is based are discussed at length and inferred ages are assigned to individual deformations where possible. A discussion of the tectonic significance of these deformations is given in the following chapter.

CORRELATION CRITERIA

Structural correlation forms the basis for combining structural sequences from several isolated lithic groups into a single structural history. A number of criteria are used in the correlation procedure analogous to those used in establishing a structural sequence in a single lithic group. These criteria are: known age; overprinting; position in a structural-metamorphic sequence; style; and orientation, in order of decreasing reliability.

Absolute age is a reliable means of correlating individual structures if the structures can be dated and if other structures of approximately the same age are absent.

In more complex situations, major folds involving more than one lithic group or cleavages overprinting the contact between two lithic groups offer a more reliable means of correlating structures between lithic groups.

Position in a structural-metamorphic sequence may be a good criterion if the structural-metamorphic sequence is well established and distinctive. If not, this criterion may prove ambiguous.

Style and orientation are generally poor criteria to use in structural correlation as they may vary extensively over short distances. Style and orientation are most useful over short distances where strongly contrasting structures or orientations are involved. They are generally used when more reliable data is unavailable.

These criteria are analogous to those used in establishing a structural history in a single unit where outcrop exposure is incomplete. Position in a structural-metamorphic sequence, style and orientation are used in exactly the same manner. Absolute age could be used in either situation in precisely the same manner as well, if it could be determined with any degree of accuracy. However overprinting is not used in exactly the same manner as it applies to contact relationships.

In correlating structures between lithic groups based upon overprinting criteria, contact relationships are emphasized. If a major structure such as a fold involves two or more lithic groups, the correlation of that structure between these lithic groups is implicit. Likewise if a structure overprints the contact between two lithic groups, correlation of that structure between the two lithic groups is implicit. Such contact relations are critical as they establish unambiguous structural correlations between major lithic groups. These unambiguous correlation points serve to construct a structural-metamorphic sequence that can then be used as the basis for additional correlations. This procedure is followed in establishing a structural history for the Glover Island Area.

INDIVIDUAL STRUCTURAL SEQUENCES

The structural history of each lithic group in the Glover Island Area has been presented in the preceding chapters. These structural histories are briefly summarized below and in Table 9.1 so that the following correlations may be more easily understood.

Cobble Cove Gneiss

Quartzo-feldspathic rocks of the Cobble Cove Gneiss contain a prominent, penetrative, S1 foliation defined by thin discontinuous laminae of biotite gneiss in a microcline gneiss host. No preferred orientation of mica exists in these biotite gneisses or in the host gneisses which have been statically recrystallized during the second interkinematic period. Evidence of a second deformational event is only observed in mafic schists.

Mafic schists of the Cobble Cove Gneiss contain a faint S2 foliation defined by a dimensional preferred orientation of plagioclase that has been extensively overprinted by decussate biotite and porphyroblastic microcline. The growth of biotite and microcline is inferred to have occurred during the same interkinematic interval during which recrystallization of quartzo-feldspathic gneisses took place.

The mafic schists are interpreted to have originated as basaltic dykes that intruded quartzo-feldspathic rocks

Table 9.1 Summary of structural sequences recognized in individual lithic groups in the Glover Island Area.
(with no attempt to correlate structural sequence among areas)

	COBBLE-COVE GNEISS	KEYSTONE SCHIST	GRAND LAKE COMPLEX	GLOVER GROUP	OTTER NECK GROUP	RED POINT AND CORNER POND FORMATIONS
D4		S4 - poorly developed crenulation cleavage F4 - minor, open folds transitional to crenulations	S4 - crenulation cleavage F4 - minor, upright, tight, asymmetric folds in margins of complex			
D3	Faulting - extensive tectonic breccia formed along fault	S3 - crenulation cleavage defined by chlorite and sericite F3 - tight, asymmetric folds with axial planar S3 crenulation cleavage	S3 - heterogeneous foliation defined by talc and magnesite L3 - pressure shadow lineation defined by talc and epitaxial rims on magnesite grains			
I2	Quartz is recrystallized and decussate biotite and porphyroblastic microcline forms	Helicitic albite overgrows S2 crenulation cleavage	Ultramafic rocks are metasomatically altered to talc and magnesite			
D2	S2 - faint banding defined by plagioclase recognized only in mafic layers interpreted as metasomatized dykes	S2 - Intensely developed locally differentiated crenulation cleavage F2 - Rare similar folds with S2 axial planar foliation L2 - pressure shadow lineation of quartz and muscovite	S2 - foliation defined by antigorite banding; recognized only in ultramafic rocks	S2 - crenulation cleavage defined by chlorite and muscovite		
I1		Static metamorphism and growth of biotite, microcline and garnet porphyroblasts				
D1	S1 - foliation defined by biotite-rich streaks in a microcline gneiss	S1 - penetrative schistosity defined by muscovite and chlorite	S1 - foliation defined by grain shape preferred orientation of hornblende inferred to have formed during synkinematic metamorphism of clinopyroxene	S1 - penetrative schistosity defined by chlorite and muscovite L1 - pressure shadow lineation defined by quartz	S1 - Penetrative schistosity defined by chlorite muscovite and actinolite	S1 - penetrative schistosity defined by sericite and chlorite F1 - minor, tight to open upright folds with an axial planar S1 schistosity

of the Cobble Cove Gneiss after D1 deformation based upon geochemistry and field appearance. Metasomatism to biotite schists is inferred to have occurred during the second interkinematic period.

Keystone Schist

Structures related to four deformational events are recognized in the Keystone Schist. The oldest structure, S1, consists of a penetrative schistosity defined by muscovite and chlorite.

Abundant, randomly oriented porphyroblasts of biotite, microcline and garnet overgrow S1 and define a static metamorphic event that occurred during the first interkinematic period.

An intense S2 crenulation cleavage developed axial planar to rare, tight to isoclinal, minor F2 folds during the second deformational event. S2 is locally differentiated where it involves psammitic lithologies and contains a well-developed pressure shadow lineation.

Abundant, tight, upright minor folds formed during the third deformational event. An S3 cleavage associated with minor chlorite and sericite formed axial planar to these minor folds.

A heterogeneously developed, crenulation cleavage formed during the fourth deformational event and is not associated with any recognizable mineral growth.

Grand Lake Complex

Structures related to four deformational events are recognized in the Grand Lake Complex. The first structure consists of a hornblende foliation in mafic metagabbros defined by a weak dimensional preferred orientation of hornblende. The Hornblende is inferred to have formed as a synkinematic replacement of clinopyroxene under amphibolite facies conditions. Relict igneous clinopyroxene grains equilibrated to diopsidic compositions during D1.

Evidence of D2 deformation is restricted to a poorly developed S2 foliation recognized only in ultramafic rocks. This foliation is defined by diffuse bands of antigorite in a clinopyroxene-antigorite gneiss.

During the second interkinematic period, ultramafic rocks at the base of the Grand Lake Complex were metasomatically altered to talc-carbonate schists.

Talc-carbonate schists were heterogeneously foliated during the third deformational event. The S3 foliation consists of a talc schistosity and bands of carbonate. A pressure shadow lineation also formed during D3 and is defined by fringes of talc and epitaxial rims on carbonate grains.

Abundant, upright, asymmetric minor folds formed at the margins of the Grand Lake Complex during the fourth deformational event. A crenulation cleavage associated with minor chlorite and sericite growth formed axial planar to these folds.

Glover Group

Structures related to two deformational events are recognized in the Glover Group. The oldest structure consists of a penetrative schistosity defined by chlorite and sericite and a pressure shadow lineation of quartz. These formed during the first deformational event and are restricted to the base of the Glover Group.

Abundant, tight, asymmetric minor folds formed during the second deformational event. A crenulation cleavage associated with minor chlorite growth formed axial planar to these minor folds.

Otter Neck Group

A single foliation is recognized in the Otter Neck Group. This foliation consists of a penetrative schistosity defined by chlorite, actinolite, epidote and sericite. It is best developed in tuffaceous rocks and weak to absent in massive dykes and pillow lavas.

Red Point Formation

The Red Point Formation contains a single penetrative schistosity defined by sericite and minor chlorite oriented axial planar to a major, open syncline.

Corner Pond Formation

The Corner Pond Formation contains a single penetrative schistosity defined by chlorite and sericite oriented axial planar to open, upright minor folds.

MAJOR STRUCTURES

Major structures in the Glover Island Area play a key role in correlating deformational sequences between different units as they involve more than one unit. Major structures in the Glover Island Area are described below.

Three major structures, the Keystone Syncline, the Kettle Pond Fold Pair and the Glover Anticline are recognized in the Glover Island Area. Abundant thrust faults and high angle faults also occur in the study area and as these are regional in extent and involve many different lithic groups they are also considered major structures and are discussed below.

The Keystone Syncline

The Keystone Syncline is a major F2 fold located within the Lower Member of the Keystone Schist. Its hinge area is well exposed in Keystone Brook and is described below.

In the central part of the Lower Member, a psammite outcrops between two units of amphibolite (Fig. 4.2). In this psammite, S2 forms a differentiated crenulation cleavage oriented at a high angle to bedding which parallels S1. Bedding-cleavage (S2) relationships in this hinge region indicate a synformal fold hinge is present. The psammite is inferred to occupy the core of this fold and the bounding units of amphibolite are inferred to represent a single unit that has been folded into a synform. On either side of the hinge region of this fold, bedding and cleavage (S2) are parallel on the scale of the outcrop and are interpreted to have been rotated into parallelism during D2 deformation.

The structural contrast between hinge and limb regions of the Keystone Syncline is very pronounced and is a characteristic feature of this structure. Limb regions appear simple. They are composed of a single, macroscopic schistosity that parallels the compositional layering on the scale of the outcrop. Original bedding is cryptic in limb regions and is replaced by homogeneous compositional layers bounded by sharp, structural (S2) surfaces.

In hinge regions, however, bedding is easily recognized with grading locally conspicuous. Bedding parallels S1 in the hinge area and S2 is oriented at a high angle to both bedding and S1. As a result, the hinge region of the Keystone Syncline has a very distinctive field appearance and is easily recognized.

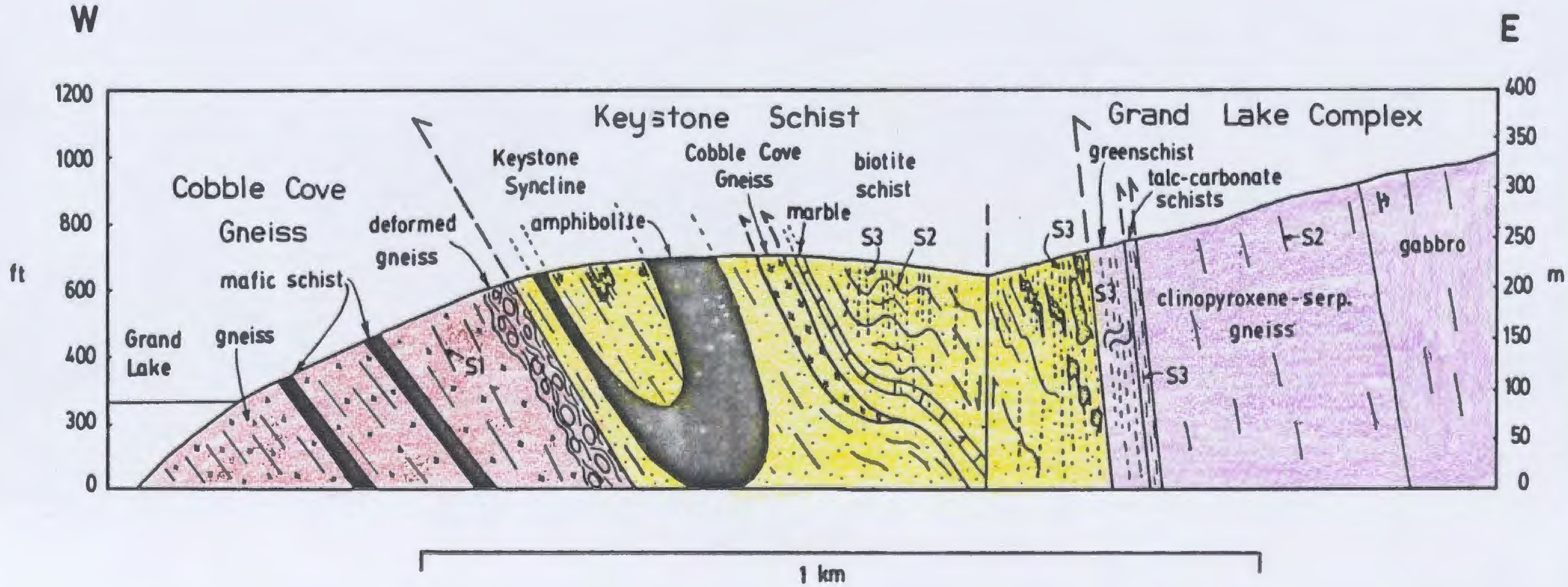
Grading is apparent in many beds in the hinge region of the Keystone Syncline. Many of these are well preserved and allow facing directions to be determined. These beds consistently face toward the axial surface of the synform indicating it is synclinal and west facing.

A profile section through the Keystone Syncline in Keystone Brook is presented in Figure 9.1. The syncline is upright, isoclinal, overturned to the west, and west-facing. Because the anticlinal partner of this syncline is absent, the vergence of the structure cannot be determined. If bedding was right-side-up prior to D2 deformation, it would be west-verging.

The regional extent of the Keystone Syncline has been determined using bedding-cleavage (S2) relationships and facing directions from graded beds. These minor structures have been used to locate the axial trace of the Keystone Syncline in Figure 9.2. The synclinal trace extends from the shoreline of Glover Island, inland to Keystone Brook, and back to the shoreline of Glover Island, parallel to major lithic boundaries.

Facing directions reverse across a ductile shear zone or fault located between the Upper and Lower Members of the Keystone Schist but bedding-cleavage (S2) intersections remain the same. Evidently one limb of the Keystone Syncline is sheared-out along this surface. A sliver of Gobble Cove Gneiss is exposed along this structural break.

Figure 9.1 Schematic structural cross-section taken through Keystone Brook. No vertical exaggeration.


W
E
1200
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800
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400
300
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ft
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150
100
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0
m
Keystone Schist
Grand Lake Complex
Cobble Cove
Gneiss
deformed
gneiss
Keystone
Syncline
amphibolite
Cobble Cove
Gneiss
marble
biotite
schist
S3
S2
S3
greenschist
talc-carbonate
schists
S2
gabbro
mafic schist
Grand
Lake
gneiss
S1
clinopyroxene-serp.
gneiss
S3
S3
1 km

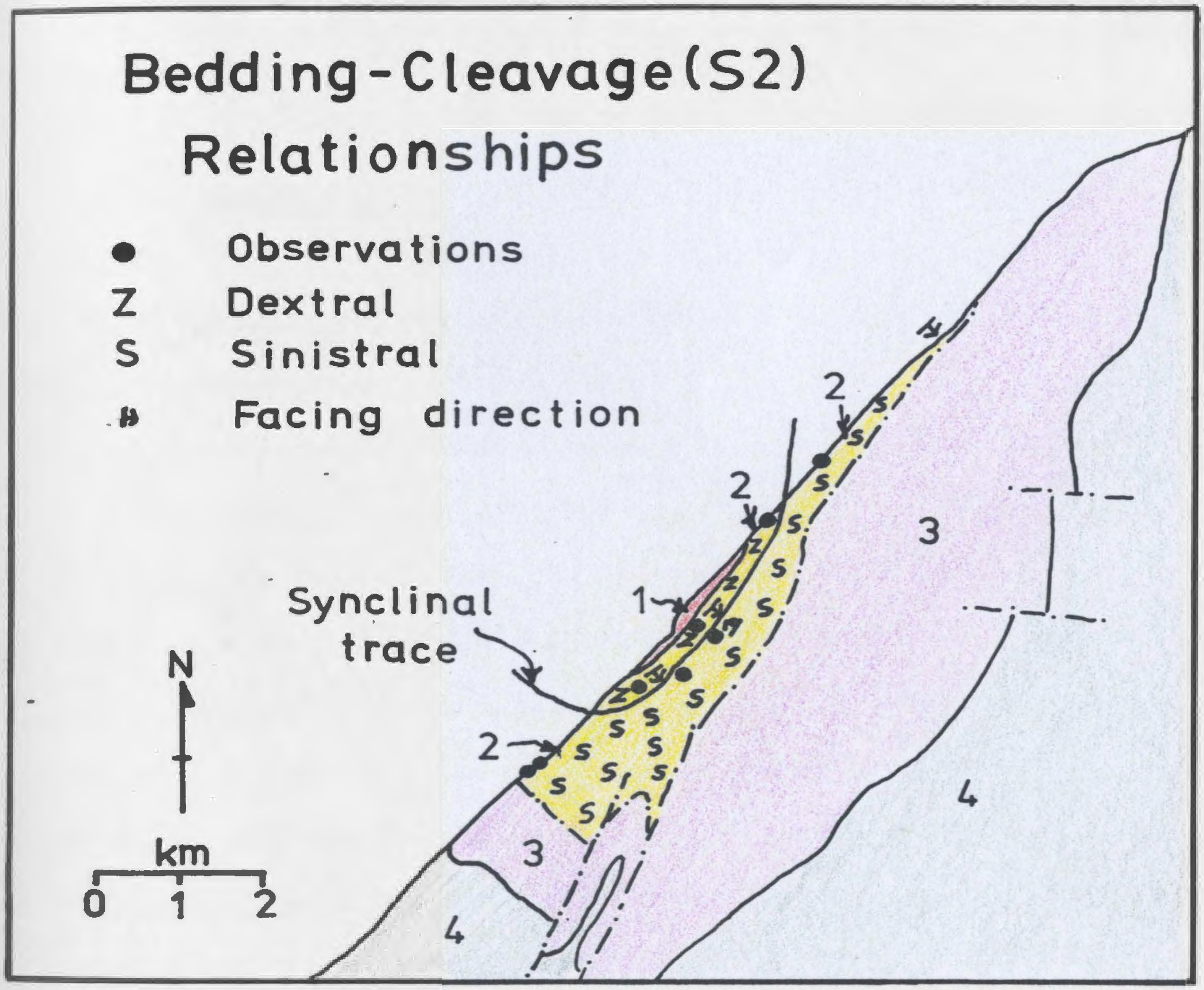


Figure 9.2 Axial trace of the Keystone Syncline based upon minor structures. 1, Cobble Cove Gneiss; 2, Keystone Schist; 3, Grand Lake Complex; 4, Glover Group.

The Kettle Pond Fold Pair

The Kettle Pond Fold Pair consists of a sequence of north-trending, south-plunging anticlines and synclines located in the Kettle Pond area of Glover Island. A major anticlinal hinge referred to as the Kettle Pond Anticline is located at Kettle Pond and a tight syncline referred to as the Kettle Pond Syncline is located about one kilometre to the west. These major structures are described separately below.

At Kettle Pond the Grand Lake Complex is folded into a major, south-plunging, upright anticline designated the Kettle Pond Anticline. The Kettle Pond Formation is folded into two smaller anticlines and a syncline related to this major anticline. Numerous minor folds associated with these major structures occur throughout the Kettle Pond Formation at the southwest end of Kettle Pond.

Separated from this anticline to the west by a north-northeast-trending, high-angle fault is a major syncline known as the Kettle Pond Syncline. Displacement across this fault is minor and the east limb of the syncline is interpreted to have once been continuous with the west limb of the anticline prior to faulting. The east limb of the Kettle Pond Syncline is unusual in a number of features.

The surface trace of the east limb of the Kettle Pond Syncline underlies a narrow topographic ridge which parallels the high angle fault between the Kettle Pond Syncline and Anticline. This ridge also parallels an intense foliation developed axial planar to the Kettle Pond Syncline. The ridge is composed of a thin, sinuous, upright unit of serpentine of the Grand Lake Complex and greenschist and quartz-sericite schists of the Kettle Pond Formation. As the ridge is traversed from south to north, quartz-sericite schists, greenschists and finally, serpentine is encountered while the foliation remains unchanged and oriented parallel to the axial plane of the syncline and parallel to the ridge.

The intensity with which this axial planar foliation is developed is highly unusual. In the Kettle Pond Anticline and elsewhere, foliation development is restricted to a weak crenulation cleavage in the hinge regions of minor folds. Foliation development in the east limb of the Kettle Pond Syncline is far more intense and consists of a pervasive, locally differentiated, crenulation cleavage. It is clear this limb has undergone a greater degree of strain than the rest of the Kettle Pond Fold Pair.

The west limb of the Kettle Pond Syncline is less well exposed and less intensely foliated. Quartz-sericite schists of the Kettle Pond Formation are exposed along the west limb of the Kettle Pond Syncline and show a pronounced rotation of structures from axial planar to the Kettle Pond Syncline to a more east-west orientation before truncation by a second high-angle fault. This rotation of structures in the west limb of the Kettle Pond Syncline is the result of later folding related to the Glover Anticline.

A cross-section through the Kettle Pond Fold Pair is given in Figure 9.3. The axial planar foliation related to these structures is vertical except where folded by the Glover Anticline in the west limb of the Kettle Pond Syncline. The resulting structures are upright and tight but not isoclinal.

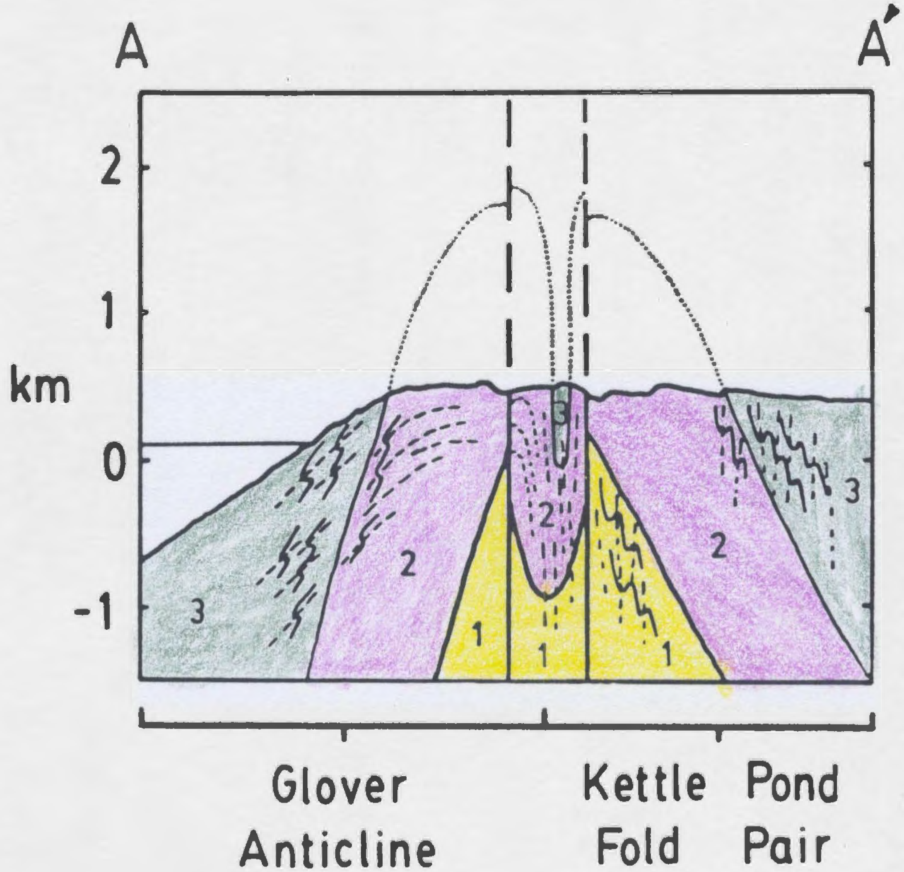
The cross-section also demonstrates that the Kettle Pond Fold Pair is strongly asymmetric. The east limb of the Kettle Pond Syncline represents the short limb of this highly asymmetric structure accounting for the intensity of foliation development it exhibits. The Kettle Pond Fold Pair verges west based upon this profile section and is west facing.

The Kettle Pond Fold Pair is of considerable importance to structural correlation in that it involves, and thus links, the Keystone Schist, Grand Lake Complex and Glover Group. These structures are easily correlated between each lithic group based upon overprinting criteria and thus form an unambiguous correlation point.

The Glover Anticline

The Glover Anticline is a major, upright, north-trending

Figure 9.3 Schematic structural cross-section through the Glover Anticline and the Kettle Pond Fold Pair. Key: 1, Keystone Schist; 2, Grand Lake Complex; 3, Glover Group; Short solid lines: S2 in the Keystone Schist, S3 in the Grand Lake Complex, and S1 in the Glover Group; Short dashed lines: S3 in the Keystone Schist, S4 in the Grand Lake Complex and S2 in the Glover Group.


A
A'
2
1
km
0
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1
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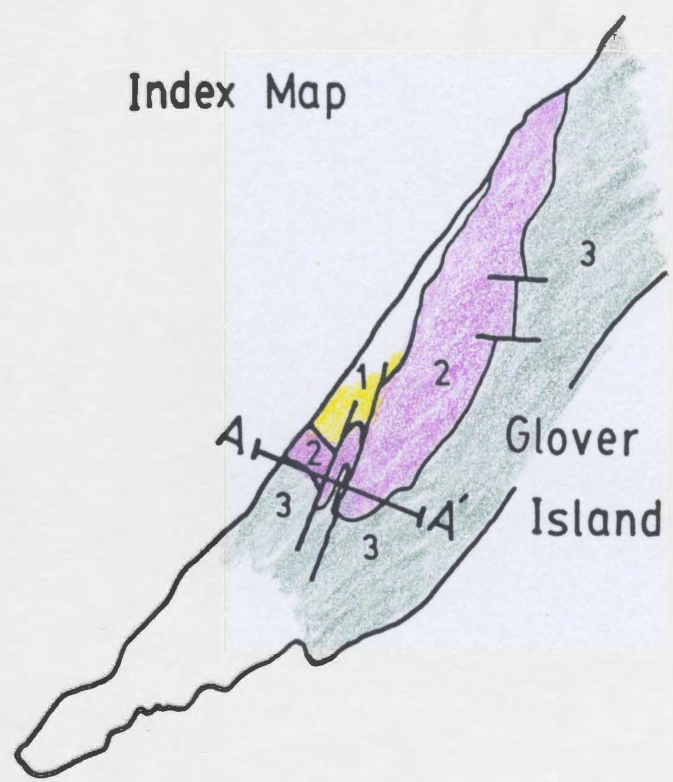
Glover
Anticline
Kettle
Fold
Pond
Pair


Index Map
3
2
1
A
2
3
3
A'
Glover
Island

south-plunging anticline that transversely intersects the central part of Glover Island. The Cobble Cove Gneiss, Keystone Schist and Grand Lake Complex occupy the core of this structure. A profile section through the Glover Anticline is given in Figure 9.3. The eastern part of this profile is dominated by the Kettle Pond Fold Pair up to the westernmost high-angle fault. Throughout this part of the profile, minor structures related to the Kettle Pond Fold Pair such as minor folds and an axial planar foliation remain constant in orientation. Towards the west these structures are gradually rotated to an east-west orientation parallel with structures in the western half of the profile section. This rotation is the result of folding by the Glover Anticline.

The change in orientation of structures from the west to the east limb of the Glover Anticline across the western fault is very abrupt and appears anomalous in terms of the scale of the structure. Evidently part of the hinge region of the Glover Anticline is missing along this high-angle fault thus accounting for the sharp changes in orientation observed.

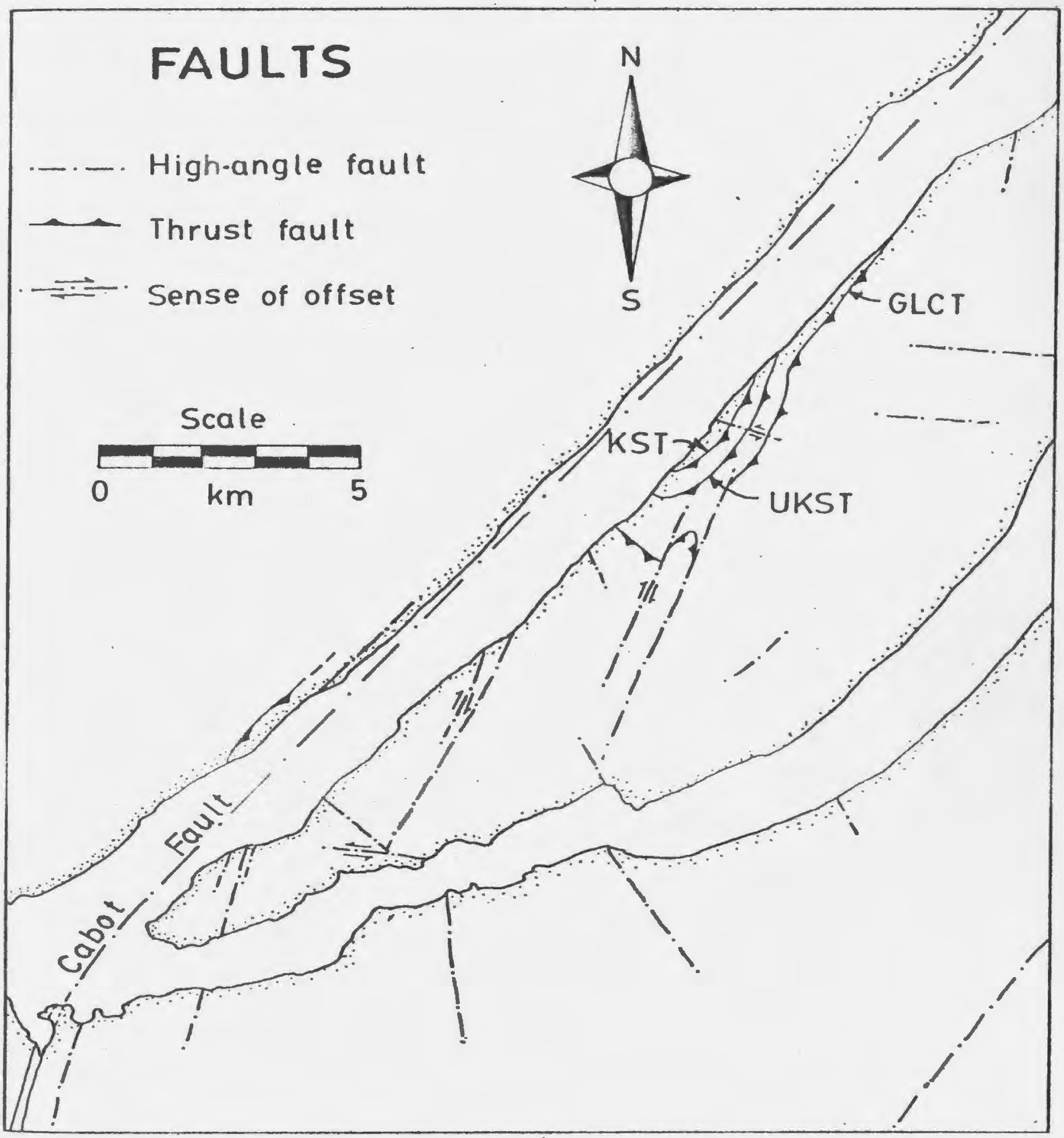
The Glover Anticline is of limited use with regards to structural correlation between different rock groups because minor structures associated with it are rare, yet form the basis for determination of the structural sequences in individual units. However age relationships can be inferred by observing the orientation of minor structures on a larger scale. In the Keystone Schist, the Glover Anticline rotates D2 and D3 structures and is therefore inferred to be a D4 structure. In the Grand Lake Complex D1, D2, D3 and D4 structures are rotated such that the Glover Anticline is inferred to be a D5 structure. Likewise D1 and D2 structures of the Glover Group are rotated by the Glover Anticline which is accordingly inferred to represent a D3 structure.

FAULTS

Faults in the Glover Island Area are divided into high-angle faults and thrust faults. Because many of these faults have been refolded, their present orientation cannot be used to distinguish between thrust faults and high-angle faults. Faults are determined to be refolded thrust faults based upon the orientation of bedding or compositional layering in the upper and lower plates of the fault. Where the fault surface and layering in both upper and lower plates is parallel for significant distances, the fault is inferred to be a thrust fault.

Based upon this criteria, faults in the Glover Island Area have been divided into high-angle faults and thrust faults (Fig. 9.4). Two sets of high angle faults are recognized and four major thrust surfaces. These major structures are described below.

Figure 9.4 Distribution of high-angle faults and thrust faults in the Glover Island Area. GLCT, Grand Lake Complex Thrust Fault; UKST, Upper Keystone Schist Thrust Fault; KST, Keystone Schist Thrust Fault.



FAULTS
High-angle fault
Thrust fault
Sense of offset
N
S
Scale
0
km
5
GLCT
KST
UKST
Cabot
Fault

Thrust faults

Four major thrust faults are recognized in the Glover Island Area. Three of these occur on Glover Island and one along the shoreline of Grand Lake opposite the southern tip of Glover Island (Fig. 9.4). Each thrust is referred to by the name of the lithic group that forms the upper plate of the thrust. This convention concisely identifies a thrust without introducing unnecessary and confusing names.

Grand Lake Complex Thrust Fault

The Grand Lake Complex Thrust Fault is located at the base of the Grand Lake Complex and separates these rocks from the underlying Keystone Schist. Bedding and compositional layering in both Grand Lake Complex and Keystone Schist parallel this fault within the limits of geologic measurement. At the fault contact, black carbonaceous phyllites of the Keystone Schist are in contact with orange weathering, talc-carbonate schists or basal greenschists of the Grand Lake Complex. The black phyllite is estimated to be less than 20 m in thickness and is present at the thrust surface everywhere on Glover Island emphasizing the parallel orientation of thrust fault and major lithic boundaries.

Age of formation of this thrust surface must clearly predate the Glover Anticline and the Kettle Pond Fold

Pair as it is involved in these major structures. Additional information concerning the age of this structure must be based upon the correlation of structures in the upper and lower plate of the thrust. This topic will be discussed in detail at a later point.

Upper Keystone Schist Thrust Fault

The contact between the Upper and Lower Members of the Keystone Schist is interpreted to be a thrust fault based upon the disruption of the Keystone Syncline and a sliver of Cobble Cove Gneiss that occurs along this contact. The thrust fault is interpreted to represent a D2 or slightly post-D2 fault of minor displacement. Its original orientation is unknown and to be consistent with the criteria used in this thesis it is inferred to be a thrust.

Keystone Schist Thrust Fault

The basal contact of the Keystone Schist consists of a poorly exposed fault in contact with the Cobble Cove Gneiss. Its orientation closely parallels bedding in the Keystone Schist and foliation in the Cobble Cove Gneiss and it is accordingly inferred to be a thrust fault. A wide (10 m) brecciated zone occurs in the Cobble Cove Gneiss adjacent to the fault.

Unnamed Ophiolite Thrust Fault

Unnamed ophiolitic rocks outcrop on the western shore of Grand Lake opposite the south tip of Glover Island. These rocks are separated from semipelitic schists to the west by a fault oriented parallel to the foliation in these schists. This surface is inferred to be a thrust fault and may represent the continuation of the Grand Lake Complex Thrust Fault on the west side of the Cabot Fault.

High-angle faults

High-angle faults are abundant in the study area and consist of a north-trending set of vertical, dextral faults and an east-trending set of faults of variable dip and shear sense (Fig. 9.4).

North-trending faults strike north-northeast in the central part of Glover Island and due north at the southern end of Grand Lake. This change may represent a gradual change in orientation or a separate generation of faults. The possible significance of this change in orientation if it relates to a single deformational event is that these minor faults roughly parallel the Cabot Fault. The Cabot Fault, although hidden beneath Grand Lake, must strike roughly north-northeast adjacent to Glover Island, while at the south end of Grand Lake it strikes more northerly. Thus, the observed orientation of minor faults roughly parallels the Cabot Fault. These dextral, high-angle faults may therefore be related to the Cabot Fault.

A second set of high-angle faults disrupts the rock units on Glover Island but this set trends easterly. These have strike separations that vary between a kilometre and several hundred metres. These faults are interpreted to be late, normal faults.

STRUCTURAL CORRELATIONS

The structural sequences of eight different lithic groups or structural domains have been summarized above (Table 9.1). In this section these different structures are correlated based upon the criteria discussed at the beginning of this chapter together with the additional information provided by major structures. The correlations to be discussed are given in Table 9.2.

Correlations based upon major structures

The Kettle Pond Fold Pair folds the Keystone Schist, the Grand Lake Complex and the Glover Group. It has abundant minor structures associated with it. This fold pair thus yields a known correlation point between these three lithic groups based upon overprinting criteria. D3 of the Keystone Schist, D4 of the Grand Lake Complex and D2 of the basal Glover Group are therefore correlated as shown in Table 9.2

Table 9.2 Correlation chart of structures in the Glover Island Area.

	Cobble Cove Gneiss	Keystone Schist	Grand Lake Cmplx	Glover Group	Otter Neck Group	Red Point Fm	Corner Pond Fm
DVI		D4[1]	D5[1]	D3[1]			
DV		D3[2]	D4[2]	D2[2]			
DIV		D2[3]	D3[3]	D1[3]	D1	D1	D1
INTIII	M	M	M				
DIII	D2[4]	D1[4]	D2[4]				
DII			D1[5]				
DI	D1[5]						

Correlation Basis: [1]The Glover Anticline; [2]The Kettle Pond Fold Pair; [3]Structural sequence, style and orientation; [4]Structural sequence; [5]Assumed age.

M – Static metamorphic event

Correlations based upon structural sequence, style and orientation

The correlation of structures related to the Kettle Pond Fold Pair and Glover Anticline is unambiguous because it is based upon overprinting criteria. This yields a fixed point in terms of structural correlation and allows additional, tentative correlations to be made by subtraction. Subtraction is used here to refer to the correlation of those structures that predate the known, correlated structure by one deformation. Thus D2 of the Keystone Schist, D3 of the Grand Lake Complex and D1 of the basal Glover Group are correlated based upon this concept as they all predate the Kettle Pond Fold Pair by a single deformation. However this correlation criteria is ambiguous and other criteria must be used to support it.

Style is a correlation criteria which must be used with caution when used alone (Williams, 1970). When used with other criteria it can be useful, however. Structures formed in the Keystone Schist during the D2 deformational event consist of a crenulation cleavage and a conspicuous pressure shadow lineation. This pressure shadow lineation is a distinctive feature of D2 deformation. A pressure shadow lineation also formed during D3 deformation of the basal Grand Lake Complex. The formation of a pressure shadow lineation during each of these deformational events strongly supports their correlation. This correlation basis is strengthened by the observation that only a single deformational event is associated with a pressure shadow lineation in each individual lithic group and thus constitutes a unique style attribute.

The orientation of these structures is difficult to compare directly because of the complicating effect of the Glover Anticline. However if structural data from the Keystone Schist, Grand Lake Complex and basal Glover Group is compared from the same limb of the Glover Anticline a striking similarity is observed (Fig. 9.5). Poles to the foliation and the pressure shadow lineation both demonstrate a strong similarity in orientation adding support to the validity of this correlation.

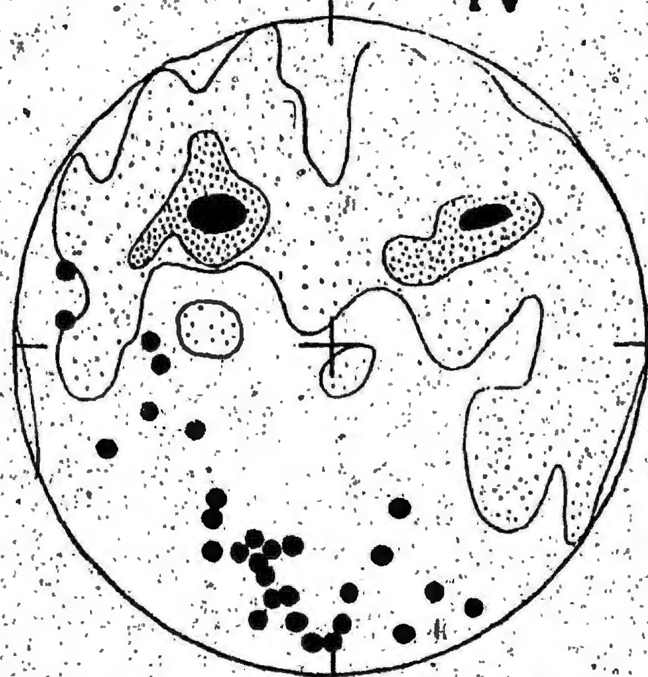
The proposed correlation of D2 of the Keystone Schist, D3 of the Grand Lake Complex and D1 of the basal Glover Group is therefore accepted as this correlation is supported by structural sequence, style and orientation. The correlated deformational event directly precedes a deformational event of known relative age based upon major structures. It involves the formation of a foliation and a pressure shadow lineation in each unit which constitutes a distinctive style group and shares a common orientation of both foliation and lineation where measurements can be directly compared.

This correlation is of great importance as it places constraints on the age of the Grand Lake Complex Thrust

Figure 9.5 Equal area, lower hemisphere projections of D1 structures of the Glover Group from Subarea IV (a) and D2 structures of the Keystone Schist from Subarea III (b). The similarity in orientation supports the proposed correlation of these structures as discussed in the text. Poles to the foliation are contoured and lineations are represented as dots. Contour intervals in percent per one percent area, the number of points contoured and the number of lineations plotted are as follows: (a) .7,3.6,7.2; S1=139; L1=29; (b) .7,3.6, 5.7; S2=139; L2=63.

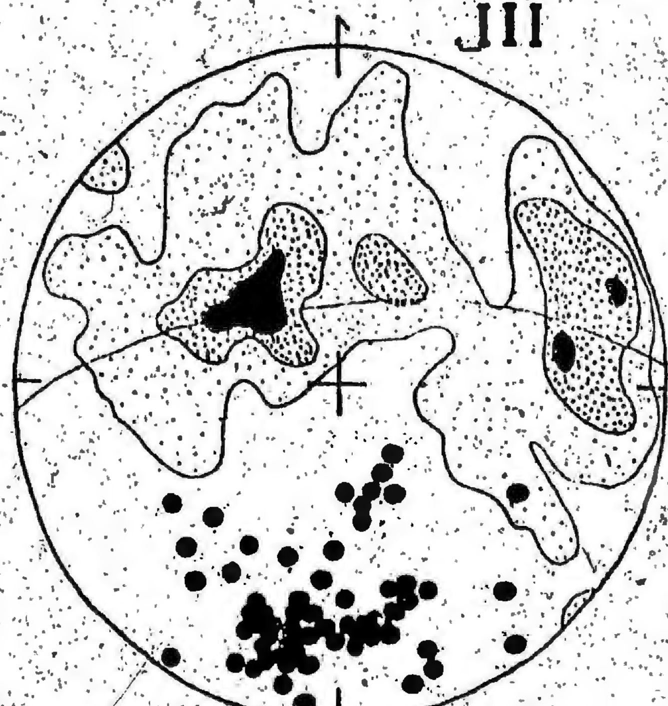
IV

a.



III

b.



Fault. This thrust fault separates the Keystone Schist from the Grand Lake Complex. The correlation of structures across this thrust requires major movement on the thrust to predate or be synchronous with the correlated structure. Thus major dislocation on the Grand Lake Thrust Fault must predate or be synchronous with D2 of the Keystone Schist, D3 of the Grand Lake Complex and D1 of the basal Glover Group. This topic is treated more fully in the next chapter.

In the same manner D1 of the Keystone Schist and D2 of the Grand Lake Complex may be correlated by subtraction. The correlation of these structures is far more speculative due to lack of data and must be considered tentative.

Correlation of structures based upon age

Where the age of a structure is known or can be inferred from the age of the unit in which it is found, this may form a reliable means of structural correlation if other structures of about the same age are absent.

The Cobble Cove Gneiss is inferred to be Precambrian based upon field appearance. Its well developed S1 foliation is accordingly assigned a Precambrian age. As no other rock groups in the Glover Island Area are Precambrian in age, no other deformational events correlate with D1.

D1 of the Cobble Cove Gneiss is followed by an episode of dyke intrusion, foliation development during D2 and static metamorphism. Based upon structural sequence and style the D2 deformational event is correlated with D1 of the Keystone Schist and the static metamorphism with the first interkinematic period in the Keystone Schist which also involved a static metamorphic event.

The first deformational event recognized in the Grand Lake Complex involved the synkinematic alteration of clinopyroxene to hornblende under amphibolite facies conditions. The resulting hornblende grains are extremely coarse (up to 1 cm) and exhibit a weak, dimensional preferred orientation that defines S1. Comparable coarse-grained microstructures are wholly unknown in any of the other rock groups in the Glover Island Area. D1 of the Grand Lake Complex is also unique in terms of structural sequence. Only D1 of the Cobble Cove Gneiss occurs at the same structural position but cannot be correlated because the Grand Lake Complex is inferred to be Lower Ordovician in age. Thus D1 of the Grand Lake Complex correlates with no other structures in the Glover Island Area.

RELATIVE AGES OF DEFORMATIONAL EVENTS

Individual deformational events can be assigned relative and absolute ages in a formal manner now that they have been correlated with one another following the criteria discussed at the beginning of this chapter. Relative ages are given in Table 9.2. Where an age is applicable to the entire study area and not simply one lithic group it is given in Roman Numerals.

The first deformational event recognized in the study area is D1 of the Cobble Cove Gneiss. This deformational event is inferred to be Grenvillian in age and is designated DI.

The second deformational event corresponds to D1 of the Grand Lake Complex which predates all other remaining structures but must be no older than Lower Ordovician. This deformational event is designated DII.

The third deformational event involves D2 of the Cobble Cove Gneiss, D1 of the Keystone Schist and D2 of the Grand Lake Complex and is designated DIII.

Static metamorphism during the second interkinematic period in the Cobble Cove Gneiss, the first interkinematic period in the Keystone Schist and the second interkinematic period in the Grand Lake Complex defines the third interkinematic period of the regional structural sequence.

D2 of the Keystone Schist, D3 of the Grand Lake Complex, and D1 of the Glover Group, Otter Neck Group, Red Point Formation and Corner Pond Formation are all equated with the fourth regional deformation event which is designated DIV.

D3 of the Keystone Schist, D4 of the Grand Lake Complex and D2 of the basal Glover Group correspond with the fifth regional deformational event which is designated DV.

D4 of the Keystone Schist, D5 of the Grand Lake Complex and D3 of the basal Glover Group are included within the DVI regional deformational event.

OROGENIC EPISODE

It is customary to assign a deformational event of known age to a particular orogeny. Crystalline Precambrian rocks along the west flank of the Appalachian Orogen are generally found to have undergone an episode of metamorphism and deformation dated at about 1 Ga. termed the Grenville Orogeny. DI deformation is accordingly inferred to have taken place during the Grenville Orogeny.

The DII deformational event contrasts with later events in terms of microstructures and its occurrence only in the Grand Lake Complex. As the Grand Lake Complex is interpreted to be ophiolitic it is likely this deformational event took place in an oceanic setting far removed from a continental margin. Intense deformation occurs along oceanic fracture zones and may represent a possibility. It may also represent initial displacement of the Grand Lake Complex.

The DIII, DIV and DV deformational events are interpreted to represent different phases of the same orogeny as they are geometrically comparable and linked in time based upon metamorphic microstructures as discussed more fully in the following chapter. They also represent the dominant orogeny in the study area. In western Newfoundland the dominant orogeny affecting Paleozoic rocks is the Taconic Orogeny. Based on this reasoning these structures are accordingly inferred to have formed during the Taconic Orogeny.

The DVI deformational event is inferred to have taken place during the Acadian Orogeny as it post-dates the peak of metamorphism and contrasts with Taconic structures in style and scale.

Chapter 10

TECTONIC SYNTHESIS AND CONCLUSIONS

A detailed geologic history of the Glover Island Area is presented below summarizing the relationships described in preceding chapters. This history is applicable to only a tiny part of the Appalachian Orogen but carries important implications for other parts of the orogen inferred to have evolved in a similar manner. One such area is the Baie Verte Peninsula of northwest Newfoundland. A comparison between these two areas is accordingly presented and significant similarities and differences are discussed.

Examining these relationships on a still larger scale, a plate tectonic model is presented based on the geologic history of the Glover Island Area, but drawing upon data from all of western Newfoundland.

GEOLOGIC HISTORY OF THE GLOVER ISLAND AREA

The geologic history of the Glover Island Area is presented below with important or controversial aspects discussed. This history is summarized in Table 10.1 and the reader is encouraged to refer to this table in the following discussion.

Table 10.1. Geologic history of the Glover Island Area.

	STRUCTURES	METAMORPHISM	ROCK UNIT
	Faults: north-trending, high-angle faults affect all map units Faults: east-trending, high-angle faults affect all map units		
DVI	Crenulation Cleavage: sporadically developed Folds: The Glover Anticline, a major, upright, south-plunging structure forms	Lower Greenschist	Inferred to affect all map units in the central Glover Island area
DV	Crenulation Cleavage: upright; most intense where associated with DV minor folds Folds: abundant, tight to open, upright; A major asymmetric DV fold faces and verges west	Medial to lower Greenschist facies	Recognized in the Keystone Schist, Grand Lake Complex, and Glover Group
DIV	Crenulation Cleavage: intensely developed, locally differentiated; oriented subparallel to bedding Lineation: conspicuous pressure shadow lineation Folds: minor folds are rare and similar in profile; The hinge of a major gently south-plunging west-facing syncline is recognized in the Keystone Schist	Greenschist facies mineral growth is restricted to sericite, chlorite and microcline	Recognized in the Keystone Schist, Grand Lake Complex, and Glover Group
		Static metamorphic event ranges between greenschist facies at top and amphibolite facies at base of the Keystone Schist; Abundant porphyroblasts of biotite, microcline and garnet occur in pelitic rocks and tremolite in ultramafic rocks; Locally ultramafic rocks are altered talc-carbonate schists	The Grand Lake Complex is uplifted and eroded and the Glover Group is deposited; The Red Point and Corner Pond Formations are deposited during a second cycle of uplift and erosion
DIII	Schistosity: defined by muscovite and chlorite in semipelitic rocks and antigorite in ultramafic rocks; parallels bedding in semipelites	Medial greenschist facies; muscovite, chlorite and epidote occur in semipelites and antigorite occurs in ultramafic rocks	Recognized in Keystone Schist and Grand Lake Complex
DII	Foliation: defined by a weak, dimensional preferred orientation of hornblende	Coarse polygonal hornblende replaces clinopyroxene and relict clinopyroxene equilibrated to diopsidic compositions during amphibolite facies metamorphism	DII structures are recognized only in the Grand Lake Complex
			The Grand Lake Complex forms as part of the Iapetus Ocean; The protolith of the Keystone Schist is deposited along the western margin of the Iapetus Ocean Basaltic dykes intrude the Cobble Cove Gneiss; May be related to initial formation of the Iapetus Ocean
DI	Gneissosity: defined by biotitic streaks	Uncertain due to overprinting by later metamorphism	DI structures are recognized only in the Cobble Cove Gneiss
			Granodioritic protolith of the Cobble Cove Gneiss forms

Grenvillian basement

The Cobble Cove Gneiss is interpreted to represent a fragment of Grenvillian basement. The protolith of the Cobble Cove Gneiss is inferred to have been a granodiorite that was deformed and foliated during the Grenville Orogeny.

The inferred Precambrian age of the Cobble Cove Gneiss is based solely on field appearance. Strongly foliated orthogneisses are conspicuously absent in the Paleozoic schists and gneisses west of Grand Lake yet are abundant in areas of Precambrian age such as south of the west spur to Grand Lake. Thus the Cobble Cove Gneiss is assigned a Precambrian age. As the strong foliation in the Cobble Cove Gneiss represents an integral part of its field appearance it is also inferred to be Precambrian in age. Precambrian inliers along the west flank of the Appalachians all exhibit Grenvillian deformation and by analogy this strong foliation is also inferred to be Grenville in age.

Following the Grenville Orogeny, the Cobble Cove Gneiss was intruded by abundant basaltic dykes. These dykes are cross-cutting, have sharp contacts with surrounding gneisses and in some cases can be traced along strike for several tens of metres. They are inferred to be Lower Paleozoic or Upper Precambrian in age and to have intruded Grenvillian basement during rifting related to the opening of the Iapetus Ocean.

Mafic dykes are now composed of biotite and microcline and thus contrast strongly with the expected mineralogy of a typical metamorphosed basaltic dyke which is amphibole and plagioclase. The unusual mineralogy is interpreted to be the result of potash metasomatism. The source for the potash is inferred to be the surrounding granodioritic gneisses.

In many parts of western Newfoundland, Precambrian rocks are intruded by basaltic dykes. Where dated these basaltic dykes are Late Precambrian or Lower Cambrian in age (Pringle, et al., 1971; Stukas and Reynolds, 1974) and are inferred to have formed during rifting related to the initial formation of the Iapetus Ocean. By analogy, metasomatized basaltic dykes in the Cobble Cove Gneiss are interpreted to also have formed at this time. Metasomatism occurred during the Taconic Orogeny.

The Cobble Cove Gneiss is interpreted to represent a basement to the Keystone Schist by analogy with Precambrian rocks elsewhere in the Appalachians. However it is unknown whether the Cobble Cove Gneiss represents a highly allochthonous fragment of basement interleaved with metasediments or if it represents the exposed tip of an autochthonous basement terrane wholly underlying Glover Island. High grade rocks of the Grenville Structural Province underlie large parts of the terrane southwest of Grand Lake. These rocks are interrupted to be northeast by volcanic rocks of the Otter Neck and Glover Groups. The transition is extremely abrupt and presumably faulted. However metasediments such as those west of Grand Lake are minor or lacking. This is significant in that metasediments are also very thin on Glover Island. Metasediments on Glover Island are represented by the Keystone Schist which is less than a kilometre in thickness. It thus appears that in the Glover Island Area and in regions to the south, metasediments are minor or lacking while basement is extensive. This suggests that the Cobble Cove Gneiss does not represent a highly allochthonous sliver, but a basement complex continuous at depth. However the area south of Grand Lake is not well known at present and may prove to be much more complex.

Formation of the Keystone Schist, Grand Lake Complex and Otter Neck Group

The Keystone Schist is inferred to have formed as a clastic wedge along the continental margin of North America. The protolith of the Keystone Schist includes graded beds, arkosic pebble conglomerate, arkosic sandstones and limestone consistent with this interpretation.

The Grand Lake Complex and Otter Neck Groups are inferred to have formed as an integral part of the Iapetus Ocean or as part of smaller marginal basins bounding the Iapetus Ocean.

The Grand Lake Complex consists of a polymetamorphosed, layered, basic pluton consisting of cumulate ultramafic and mafic rocks. Its protolith is inferred to have been composed of websterites and clinopyroxenites at the base and pyroxene gabbros at the top separated by a narrow transition zone dominated by cumulate layered gabbros.

Ophiolites contain basic cumulate sections and are commonly located in orogenic zones. Ophiolitic rocks are conspicuous along the Baie Verte-Brompton Line. As the Grand Lake Complex is a basic layered pluton along the Baie Verte-Brompton Line it is interpreted to be ophiolitic.

The Otter Neck Group consists of pillow lavas, pillow breccias, aquagene tuffs and locally, sheeted dykes intruded by abundant small stocks and dykes of silicic and mafic porphyry. These intrusions of silicic and mafic porphyry are geochemically and petrologically identical to extrusive silicic tuffs and porphyritic basalts of the Glover Group which unconformably overlies the Grand Lake Complex. This suggests the Otter Neck Group forms a basement to the Glover Group.

The Otter Neck Group is interpreted to be ophiolitic based upon the occurrence of sheeted dykes and evidence that the Otter Neck Group forms a basement to the Glover Group. Throughout Notre Dame Bay, volcanic rocks and a variety of pyroclastic rocks comformably overlie an ophiolitic basement. The Glover Group consists of a comparable group of volcanic and pyroclastic rocks. Thus evidence that the Otter Neck Group forms a basement to the Glover Group supports its interpretation as ophiolite.

DII Deformation

DII is defined by a weak dimensional preferred orientation in hornblendites of the Grand Lake Complex. It is associated with an amphibolite facies, synkinematic metamorphic event.

Pyroxene was extensively replaced by an olive-green hornblende during DII. The hornblende is coarse and locally forms monomineralic layers presumably replacing clinopyroxenite layers. In these monomineralic layers the hornblende is coarse, polygonal and exhibits numerous triple junctions.

These microstructures all reflect a close approach to microstructural equilibrium most easily attained with hornblende at elevated temperatures corresponding to the amphibolite facies. Clinopyroxene also re-equilibrated during this metamorphism. The composition of clinopyroxene from clinopyroxenites and clinopyroxene-hornblende rocks was determined by microprobe analysis and plotted in the pyroxene quadrilateral. These clinopyroxene compositions define a crude alkalic trend. However a similar trend would result if subalkaline compositions were re-equilibrated to diopsidic compositions during metamorphism. This last interpretation is considered to be the most probable interpretation of these calcic clinopyroxene compositions. The composition of these clinopyroxenes cannot be compared with those in other ophiolites because of this metamorphic overprinting and any such comparison would be meaningless.

DII is only recognized in the Grand Lake Complex and is unique because of the coarse microstructures and high metamorphic grades inferred to have accompanied it. It is interpreted to have originated in an oceanic environment. Similar metamorphic rocks have been dredged and drilled from the modern oceans and are thus known to form in an oceanic environment at spreading ridges, along transform faults or during initial displacement of the ophiolite.

Karson and Dewey (1978) have suggested that ophiolitic rocks of the Bay of Islands area include a transform fault and attribute pronounced structural contrasts within these rocks as the result of this transform fault. If so, a likely cause for this deformational event is activity along a transform fault.

Alternatively, great strains are involved in initial obduction of the ophiolite. Initial obduction generally occurs in an oceanic environment and involves new ocean crust that has not had time to cool and is still at high temperatures. Initial obduction is thus a likely cause of DII deformation and is the interpretation that is favoured.

The Taconic Orogeny

Three deformational events are inferred to have occurred during the Taconic Orogeny in the Glover Island Area, DIII, DIV and DV. A static metamorphic event corresponding to the peak of metamorphism occurred during the interkinematic period separating DIII and DIV.

DIII is interpreted to signal the beginning of the Taconic Orogeny in the Glover Island Area. This deformational event is defined by a penetrative schistosity recognized in the Keystone Schist. Muscovite, chlorite and epidote define this schistosity and also the medial greenschist facies conditions under which it is inferred to have developed.

No folds or other structures are recognized associated with the DIII deformational event. Thus the possibility exists that the SIII penetrative schistosity is not associated with a deformational event but formed by the metamorphic overprinting of a sedimentary fabric. However early penetrative schistosities are recognized in many orogenic areas where they are commonly associated with folding and deformation. For this reason the penetrative schistosity within the Keystone Schist is inferred to have developed during a deformational event. The absence of recognized DIII folds is attributed to the lack of exposure and the intensity of later overprinting structures rather than to a true absence of such structures. Nevertheless this interpretation must clearly be understood to represent an assumption.

An antigorite foliation is recognized in ultramafic rocks of the Grand Lake Complex at the same position in the structural sequence occupied by DIII of the Keystone Schist and these two structures are correlated on this basis. This correlation corresponds to the age of juxtaposition of the Keystone Schist with the Grand Lake Complex or the age of ophiolite emplacement at the continental margin. If these structures are correlative as is hypothesized here, ophiolitic emplacement on the continental margin must have occurred before or during DIII deformation. Unfortunately this correlation is weak as it is made solely on the basis of structural sequence and not bracketed by more reliable correlative events.

Support for this correlation is present in the metamorphic history of the Grand Lake Complex, however. The base of the Grand Lake Complex consists of ultramafic rocks. Proximal to the thrust at the base of the Grand Lake Complex, referred to here as the Grand Lake Complex Thrust Fault, ultramafic rocks are altered to talc-carbonate schists. This metasomatic event is inferred to have occurred during the third interkinematic period.

It is a reliable, relative age as it is known to predate DIV based upon overprinting relationships. The importance of this metasomatic event is that its distribution at the base of the Grand Lake complex is clearly related to the Grand Lake Complex Thrust Fault. The thrust fault must therefore have developed before metasomatism of the ultramafic rocks, which took place before DIV based upon overprinting relationships. The age of inception of the Grand Lake Complex Thrust Fault must predate DIV based upon this reasoning and is likely to be synchronous with DIII deformation. DIII structures are therefore interpreted to be related to ophiolite emplacement.

A major, static, metamorphic event occurred during the interkinematic period just before the fourth deformational event. This metamorphic event is well represented in the Cobble Cove Gneiss, Keystone Schist and Grand Lake Complex.

Quartzo-feldspathic rocks are annealed and mafic rocks metasomatized to biotite schists during this event in the Cobble Cove Gneiss. The gneisses contain abundant quartz which is nearly strain free and forms polygonal, coarse grains with abundant 120 degree triple junctions. Albite and microcline did not recrystallize as they do not show comparable microstructures and comprise large, irregular grains deeply embayed by recrystallized quartz.

The Cobble Cove Gneiss is inferred to have recrystallized during this static, metamorphic event under upper greenschist facies conditions based upon the occurrence of albite.

In the Keystone Schist, this static metamorphic event is well documented because it is associated with the growth of abundant porphyroblasts of biotite, microcline and garnet. These post-tectonically overgrow the SIII schistosity and contain straight inclusion trails in porphyroblast core regions where inclusion trails can be recognized.

A gradient in metamorphism is inferred to have existed during this metamorphic event. At the base of the Keystone Schist, biotite and microcline porphyroblasts are coarse and abundant. Higher in the Keystone Schist porphyroblasts gradually decrease and finally disappear. This change is accompanied by the gradual loss of garnet and epidote and the appearance of tourmaline towards the top of the Keystone Schist. A strong decrease in grain size accompanies this change in mineralogy. Grain size changes can result from variations in impurity content, structural overprinting or kinetic variations. However the variation observed in the Keystone Schist is coupled with consistent mineralogical changes that indicate this variation is due to a true metamorphic gradient.

Metasomatism of ultramafic rocks at the base of the Grand Lake Complex is also inferred to have occurred during this static matemorphic event. Metasomatism required the influx of substantial carbonate-rich fluids to proceed. It is likely the source for this fluid was the surrounding schists. Decarbonation reactions in these schists would have released substantial carbonate-rich fluids easily accounting for that required to metasomatize ultramafic rocks along the margin of the Grand Lake Complex.

A major episode of uplift and erosion occurred between DIII and DIV. DIII and older structures are present in the Keystone Schist and Grand Lake Complex but are absent in the unconformably overlying Glover Group. The Glover Group is interpreted to be younger than DIII based upon this relationship. The Corner Pond Formation is younger than the Glover Group as it unconformably overlies it. However it is older than DIV as it shares DIV structures with the Glover Group. By correlation, the Red Point Formation is also pre-DIV in age. Thus between DIII and DIV, erosion of the Grand Lake Complex occurred followed by deposition of the Glover Group then erosion of the Glover Group and Otter Neck Group and deposition of the Red Point and Corner Pond Formations, respectively.

The peak of metamorphism and uplift appear to have occurred at the same time suggesting the two processes are linked in some way. Loading of the continental margin is a likely explanation. During tectonic loading of the continental margin the burial of basal portions of the continental margin sedimentary clastic wedge resulted in metamorphism while crustal thickening simultaneously caused isostatic uplift. Thus metamorphism and uplift are linked in a rational manner.

The second phase of the Taconic Orogeny is represented by the fourth deformational event, DIV. Structures formed during this deformational event consist of a crenulation cleavage, pressure shadow lineation and both major and minor folds.

The most conspicuous structure to form during DIV is a crenulation cleavage which is differentiated in psammitic lithologies. It forms the dominant foliation in the Keystone Schist. A cogenetic pressure shadow lineation formed with this crenulation cleavage. It plunges gently to the south throughout the Keystone Schist. This uniform orientation is the result of coaxiality of this pressure shadow lineation with the Glover Anticline. This coaxiality is due to coincidence and has no genetic implication as these structures are clearly of contrasting ages.

Minor folds of DIV age are rare in the Keystone Schist but a major syncline of DIV age has been recognized. This major fold, referred to as the Keystone Syncline, plunges gently south and faces west based upon graded beds preserved in its hinge region. Its fold axis parallels the pressure shadow lineation developed during DIV and its axial plane parallels the SIV, crenulation cleavage.

DIV structures do not show clear evidence of westerly transport although the evidence is equivocal. For example, the Keystone Syncline is west-facing but because its anticlinal partner is absent, possibly sheared-out along the Upper Keystone Schist Thrust Fault, its vergence is unknown. If the assumption is made that bedding was not overturned prior to DIV deformation then it must be west-verging. However an older deformational event is recognized in the Keystone Schist (DIII), such that this assumption cannot be made. The vergence of the Keystone Syncline and its sense of tectonic transport is unknown.

The DIV pressure shadow lineation and the fold axis of the Keystone Syncline are interpreted to have kinematic significance. The pressure shadow lineation is interpreted to have the same significance as does a stretching lineation in that it is interpreted to define X, the axis of maximum elongation of the DIV strain ellipsoid. Pebble conglomerates of the Keystone Schist contain stretched quartz pebbles (Fig. 4.4) that are oriented roughly parallel to this pressure shadow lineation. Pull-apart structures of DIV age involving relict plagioclase phenocrysts are also recognized in the Glover Group. These relationships support interpretation of this pressure shadow lineation defining the axis of maximum elongation for DIV strain. Although other interpretations have been offered for pressure shadow lineations in mylonites (Lister and Price, 1978), in view of the above data these alternative interpretations are not considered applicable.

The axis of the Keystone Syncline is oriented parallel to the pressure shadow lineation. A similar geometry between fold axis and the axis of maximum elongation of finite strain has been reported in many areas where high strains are observed. Although a number of different explanations have been offered to account for this geometry (see review in Hobbs et al., 1976, p. 283), the viewpoint taken by most recent workers is that it represents the progressive rotation of the fold axis into near parallelism with the stretching axis as the result of high strains (e.g., Johnson, 1965; Bryant and Reed, 1969). Thus both fold axis and pressure shadow lineation are interpreted to define the stretching axis of DIV strain. They have kinematic significance as this axis is inferred to represent the direction of tectonic transport during DIV deformation.

Fold axes and the DIV pressure shadow lineation are gently inclined to the south signifying that DIV structures are not related to ophiolite emplacement, a conclusion already reached on the basis of age of the Grand Lake

Thrust Fault.

DIV deformation and static metamorphism during the third interkinematic period are interpreted to be closely linked in time. This interpretation is based upon inclusion trails in microcline porphyroblasts from the Keystone Schist. In most porphyroblasts these inclusion trails are perfectly straight throughout the porphyroblast and are continuous with the matrix of the rock. In some porphyroblasts from rocks of the Keystone Syncline these inclusion trails begin to sharply curve in rim areas and continue to curve until they become continuous with SIV of the matrix. This gradual curvature is interpreted to represent the progressive rotation of the SIII schistosity in the matrix into parallelism with the SIV crenulation cleavage. Microcline growth therefore occurred first as post-tectonic growth and then as synkinematic growth during DIV. No optical discontinuity exists at the boundary between these two growth episodes and on that basis the interkinematic, static metamorphic event and DIV deformation are interpreted to have occurred penecontemporaneously during the same tectono-metamorphic pulse.

The last deformational event related to the Taconic Orogeny is DV. It consists of a crenulation cleavage and abundant minor and some major folds. The crenulation cleavage is upright and best developed in areas of intense DV folding. Minor tight to open, upright folds formed during DV and a major fold pair developed near Kettle Pond referred to as the Kettle Pond Fold Pair. It consists of a major, west-facing, west-verging, asymmetric fold. The short limb of this major fold has an intensely developed crenulation cleavage.

DV occurred during the Taconic Orogeny based upon its similarity to DIV in terms of style and orientation of structures. Both involve west-facing major structures associated with an axial planar crenulation cleavage.

DVI is inferred to have taken place during the Acadian Orogeny as it contrasts with older structures. Major open upright folds formed during DVI which lack the dominant west-facing direction of Taconic structures. The scale of the structures is entirely different as well. The Glover Anticline, a DVI structure, is a very large-scale structure in contrast to the smaller-scale structures typical of the Taconic Orogeny

Kinematics of the Taconic and Acadian Orogenies

The Taconic Orogeny is defined by three deformational events in the Grand Lake Area, DIII, DIV, and DV. DIII and DIV are separated by a thermal peak and uplift signified by the development of several unconformities. DIII is interpreted to relate to ophiolite transport and the gradual thickening of the continental margin by the stacking of thrust slices. This represents a progressive simple shear deformation in which the direction of tectonic transport is roughly horizontal towards the continent. Basal parts of this thickened crust became metamorphosed while upper parts were uplifted and eroded. Thus metamorphism and erosion occurred penecontemporaneously. Collision of outboard elements of the orogen with the continent along an east dipping subduction zone and their aborted obduction introduced an element of pure shear into the deformation resulting in locally aberrant directions of tectonic transport such as the north-south direction shown by DIV structures. DIV and DV structures share the same style and west facing geometries reflecting similar movement patterns as they represent different phases in the same overall deformational event.

Acadian structures contrast as they are upright and large in scale. They represent an episode of crustal thickening unrelated to Taconic westward movement.

Late faults

Numerous high angle faults dissected the Glover Island Area after the Acadian Orogeny. These faults can be divided into a north-trending set and an east-trending set.

North-trending faults are older than east-trending faults based upon cross-cutting relationships at the southern end of Glover Island where a north-trending fault is offset by an east-trending fault (Plate I). These north-trending faults may be related to the Cabot Fault as they trend into the Cabot Fault at a low angle but do not offset it. Their orientation may simply be coincidental, however. Since the Cabot Fault cuts Carboniferous rocks these structures must be Carboniferous or younger in age.

East-trending faults are interpreted to be normal faults of minor offset. Their age is uncertain.

REGIONAL CORRELATIONS AND IMPLICATIONS

The geological history summarized above is strictly applicable to only the Glover Island Area. However it carries important implications for other parts of the Appalachians inferred to have developed in the same manner. One such area is the Baie Verte Peninsula of northwest Newfoundland. This area contains many rock units and rock associations comparable to those on Glover Island and may once have been continuous with rock units on Glover Island. A comparison between the two areas is clearly demanded.

The geology of the Baie Verte Peninsula has been described in a number of publications (e.g., Church, 1969; Bursnall and DeWit, 1975; Williams et al., 1977; Kidd, 1977; Williams, 1977; DeGrace et al., 1976) and the reader is referred to these publications for familiarity with this geology. In the following section it is assumed the reader is familiar with the geology of the Baie Verte Peninsula.

Geology of the Baie Verte Peninsula

The western half of the Baie Verte Peninsula is underlain by the Fleur de Lys Supergroup which consists of complexly deformed and metamorphosed metasediments and amphibolite. Basement rocks and pods of eclogite are locally recognized.

The Fleur de Lys Supergroup is bounded to the east by disrupted fragments of ophiolite of the Advocate Complex. Along the contact the Fleur de Lys Supergroup consists of greenschists and ophiolitic mélanges termed the Birchy Complex.

The Advocate Complex is overlain by megabreccias, conglomerates and silicic and mafic volcanic rocks of the Flatwater Group. These coarse clastic rocks contain abundant fragments and blocks of ophiolitic rocks and rare clasts of polydeformed schists.

The eastern half of the Baie Verte Peninsula is underlain by a group of Ordovician volcanic rocks intruded by Ordovician and Silurian plutonic rocks. Locally these Ordovician rocks are unconformably overlain by Silurian sediments and mafic and silicic subaerial, volcanic rocks.

Ordovician volcanic rocks of the Baie Verte Peninsula are interpreted to overlie ophiolite because well preserved ophiolite complexes are locally exposed in this area including the Point Rousse Complex and the Betts Cove Complex.

Comparison between Glover Island and the Baie Verte Peninsula

The most obvious similarity between the geology of Glover Island and the Baie Verte Peninsula is the presence of a major structural junction between schists and gneisses (west) and ophiolitic rocks and volcanics (east). This major break is referred to as the Baie Verte-Brompton Line based on the belief that it is continuous between Baie Verte and Brompton Lake, Quebec.

This major structural break is marked by disrupted ophiolitic rocks at Glover Island and on the Baie Verte Peninsula. Thus the Grand Lake Complex and Advocate Complex are located in structurally analogous positions.

Ophiolitic rocks are followed eastward by coarse clastic rocks and volcanics at both localities. Thus the Flatwater Group is located in an analogous position as is the Glover Group. Both consist of coarse clastic rocks at the base, overlain by volcanic or volcaniclastic rocks. Both silicic and mafic volcanic rocks are recognized.

The Keystone Schist and Cobble Cove Gneiss are located west of this volcanic terrane in a position structurally analogous to the Fleur de Lys Supergroup on the Baie Verte Peninsula. Both consist largely of semi-pelitic and psammitic schist and minor amphibolite and are thus strongly similar.

Ophiolite complexes occur in the eastern part of the Baie Verte Peninsula. These consist of the Point Rousse Complex and Betts Cove Ophiolite Complex, and differ from ophiolitic rocks of the Advocate Complex in that they are less deformed and are conformably overlain by volcanic sequences. The Otter Neck Group is interpreted to be an ophiolite complex that forms a basement to the Glover Group. This ophiolite complex contrasts with the Grand Lake Complex in that it is comparatively undeformed and is therefore analogous to these other ophiolite complexes on the Baie Verte Peninsula.

From these relationships it is clear that rocks of the Baie Verte Peninsula bear a striking similarity in lithology and distribution as do those in the Glover Island Area and that it is likely the two areas evolved in a similar manner. The geologic history developed at Glover Island should therefore be in part, applicable to both areas.

The Baie Verte - Brompton Line

The Baie Verte-Brompton Line is the surface expression of a major structure that separates schists and gneisses from ophiolite and volcanic rocks. The Baie Verte-Brompton Line, defined on the Baie Verte Peninsula, clearly outcrops on Glover Island based upon the widespread similarity between the two areas. At each place it is marked by ophiolite followed by coarse clastic rocks and volcanic rocks to the east. These similarities indicate the Baie Verte-Brompton Line is certainly continuous to Glover Island and can be defined by the same criteria everywhere it outcrops.

The dogged persistence of a number of distinctive characteristics over vast distances makes the Baie Verte-Brompton Line unusual. It is vitally important to examine these characteristics at this point as these must be reflected in the geologic history of the Glover Island Area. First and foremost, the Line is invariably marked by ophiolite. These ophiolitic rocks along the Line are always severely deformed, fragmented and discontinuous in sharp contrast to the far less deformed, nearby ophiolites of the Point Rousse, Betts Cove and Otter Neck Groups.

On the Baie Verte Peninsula the Birchy Complex bounds these fragments of ophiolite to the west. The Birchy Complex is a highly deformed unit containing a variety of lithologies including dismembered ophiolite and is interpreted to include an ophiolitic mélange formed by ophiolite transport. (Williams, 1977). The Birchy Complex marks the line on the Baie Verte Peninsula with great reliability but is not well represented on Glover Island and is not a continuous aspect of the Line.

Megabreccias and conglomerates lie above ophiolites along the Line wherever exposed while similar rock units are absent farther east of the Line. These distinctive coarse sediments are clearly a key aspect of the Line.

East of these coarse sediments is a vast, volcanic-plutonic terrane in which stratigraphy, deformation and sedimentation are typical of western parts of Notre Dame Bay.

A strong structural contrast occurs along the Line on the Baie Verte Peninsula. The Fleur de Lys Supergroup has undergone a long and complex structural history unrepresented in the nearby Flatwater Group. It is controversial whether ophiolitic rocks along the Line have undergone the same structural history as has the Fleur de Lys Supergroup but ophiolitic fragments in the Birchy Complex clearly have. This structural contrast is an important aspect of the Baie Verte-Brompton Line throughout its extent.

Ophiolitic mélange along the west flank of the Line is attributed to ophiolite transport by Williams (1977). This mélange, the Coachman's Melange of the Birchy Complex, then underwent the full sequence of Fleur de Lys deformations and metamorphisms. This aspect of the Line is therefore the result of ophiolite transport prior to deformation and metamorphism.

Ophiolites along the Line are also related to ophiolite transport and sharply delineate the Baie Verte-Brompton Line. Based upon relationships at Grand Lake, these ophiolites then underwent deformation and metamorphism after emplacement.

The Glover Group and Flatwater Group are structurally simpler than rocks to the west. As the Glover Group is inferred to unconformably overlie the Grand Lake Complex based on the nature of sediments along the contact, this structural contrast is inferred to be due to an unconformity. A similar argument is equally applicable to the Flatwater Group. Thus the Baie Verte-Brompton Line marks a conspicuous unconformity.

Several of the criteria that make the Baie Verte-Brompton Line so distinctive are clearly related to ophiolite transport while others are related to an episode of erosion and deposition. Yet these criteria unfailingly pinpoint the Line with equal precision. The Baie Verte-Brompton Line must in some way link the processes of ophiolite obduction, transport and emplacement with uplift, erosion and deposition to be so faithfully defined by such contrasting criteria.

A solution to this enigma is suggested by the concept of parallochthonous and parautochthonous rock units. These units are characterized by having been deposited during the movement of a thrust sheet. In this situation the thrust sheet is represented by the ophiolite as it is emplaced or thrust westwards above a thickened wedge of continental margin sediments above an east-dipping subduction zone. Detritus derived from the allochthon is deposited in front of the advancing allochthon on the autochthon to form a parautochthonous unit. Detritus derived from the allochthon and deposited on the allochthon forms a parallochthonous rock unit.

Stevens (1970) first described parautochthonous rock units in western Newfoundland. The Blow-Me-Down Brook Formation is a parautochthonous rock unit containing abundant ophiolite detritus. It was deposited in front of the advancing allochthon and eventually over-ridden and incorporated into the allochthon.

Parallochthonous rocks have recently been described from the Bay of Islands Ophiolite Complex on North Arm Mountain. Here Llanvirn breccias, sandstones and argillites of the Crabbe Brook Group (Casey and Kidd, 1981) unconformably overlie ophiolite. These sediments formed during ophiolite transport and are therefore parallochthonous.

Megabreccias and coarse clastic rocks located along the Baie Verte-Brompton Line are interpreted to be parallochthonous as well accounting for the close association of erosional features with those related to ophiolite transport. The emplacement of ophiolite along the continental margin is evidently accompanied by abundant erosional features.

This conclusion allows one to take a closer look at the ophiolite emplacement process. Exhumation would appear to be the rule rather than the exception in the emplacement process based upon the vast regional extent of the Baie Verte-Brompton Line. This exhumation is most logically the result of the underthrusting of increasing amounts of sediments beneath the ophiolite as the continental margin began to be subducted. Thus the ophiolite was thrust upwards to be eroded as is happening at Timor in Indonesia today (Hamilton, 1978).

Other ophiolites located just east of the Baie Verte-Brompton Line have not been eroded based upon their gradational contacts with overlying volcanic rocks. The transition between exhumed and intact ophiolites must be very rapid and appears to occur at the point at which the Baie Verte-Brompton Line outcrops. Thus although the Baie Verte-Brompton Line is said to simply be the surface expression of a complex surface it is far more significant than that because it is also the point at which erosion of ophiolite commences. The present position of the Line is not simply an accident of later deformational events but is the point at which the over-riding ophiolite meets the thick wedge of continental metasediments and steepens to go over the imbricated sediment wedge. At this point the ophiolite breaks the surface and becomes eroded thus accounting for the ubiquitous relationship that conglomerates occur above ophiolites along the Line and to the west, but not to the east of the Line. The Line thus represents a fossil point of ophiolite exhumation.

This process of uplift, erosion, and deposition, may occur several times. The resulting parallochthonous sediments will become younger each time it occurs, recycling older sediments. This contrast in age indicates that although all parallochthonous rocks are related to one another by having evolved in a similar manner, they may be of different ages. For example the Crabbe Brook Group is Llanvirn while the Corner Pond Formation is Arenig. Yet both are interpreted to be parallochthonous.

Thus the ophiolite emplacement process offers a satisfactory solution accounting for the common association of emplacement features and erosional features along the Baie Verte-Brompton Line. The Glover Group, Flatwater Group, Corner Pond Formation and Red Point Formation are considered to be at least in part, parallochthonous based on these relationships.

PLATE TECTONIC RECONSTRUCTION

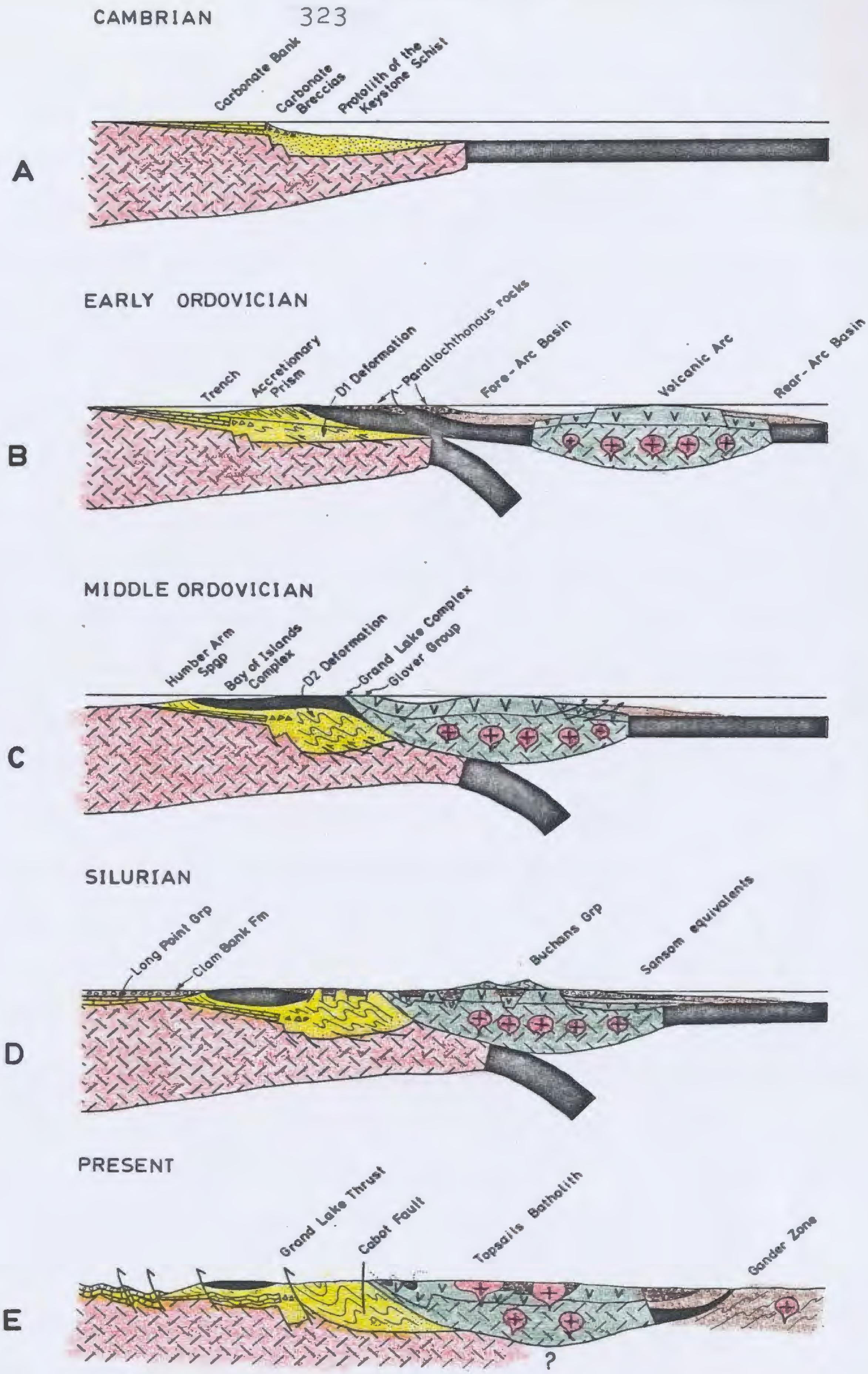
Plate tectonics has provided a comprehensive framework in which to view an orogen. In the following section a plate tectonic model is presented which satisfactorily accounts for many of the observed relationships in western Newfoundland and places the geologic history developed for the Glover Island Area into a plate tectonic framework. This model is schematically illustrated in Figure 10.1 and is discussed below.

Late Cambrian (Fig. 10.1a)

The Proto-Atlantic or Iapetus Ocean formed during the Late Precambrian by continental rifting and sea-floor spreading. This ancient ocean was bounded to the west by the North American continent along a passive, Atlantic-type, continental margin. The trailing edge of this continent was composed of Grenvillian, sialic crust that was thinned and intruded by abundant mafic dykes during the initial rifting episode. A thick section of coarse clastic rocks formed a sediment wedge along the continental margin at this time as well.

Initial rifting was followed by gradual subsidence and the development of a carbonate bank along the continental shelf. Oceanwards this carbonate bank was succeeded by coarse breccias, and still farther oceanwards, by pelagic sedimentation.

Figure 10.1 Evolution of the Glover Island Area. See discussion in text.


CAMBRIAN
Carbonate Bank
Carbonate Breccias
Protolith of the Keystone Schist
A
EARLY ORDOVICIAN
Trench
Accretionary Prism
D1 Deformation
Parallochthonous rocks
Fore-Arc Basin
Volcanic Arc
Rear-Arc Basin
B
MIDDLE ORDOVICIAN
Humber Arm Spgp
Bay of Islands Complex
D2 Deformation
Grand Lake Complex
Glover Group
C
SILURIAN
Long Point Grp
Clam Bank Fm
Buchans Grp
Sansom equivalents
D
PRESENT
Grand Lake Thrust
Cabot Fault
Topsails Batholith
Gander Zone
E
?

The Cobble Cove Gneiss is interpreted to represent part of this thinned, sialic, Grenvillian crust. It contains an intensely developed foliation that may have formed during the Grenville Orogeny. The abundant mafic dykes that intrude the Cobble Cove Gneiss may be related to rifting.

The Lower Member of the Keystone Schist consists of graded beds, pebble conglomerates and arkosic quartzites interlayered with metavolcanic sills or flows. These rocks are interpreted to have been deposited along the continental margin during initial rifting.

Early Ordovician (Fig. 10.1b)

During the Late Cambrian or Early Ordovician the polarity of plate motion changed and a convergent plate margin formed. Eastward subduction along this convergent plate margin resulted in the formation of a volcanic arc floored by ophiolite.

The age of subduction and obduction is based upon the Late Arenig age of the syntectonic Blow-Me-Down Brook Formation (Stevens, 1970). Chromite in this unit requires obduction to have begun prior to Late Arenig. The Corner Pond and Red Point Formations may represent parallochthonous equivalents of the Blow-Me-Down Brook Formation.

The Grand Lake Complex is interpreted to represent a part of the basement to this oceanic arc. As this arc evolved, it developed a root of intermediate composition formed by the intrusion of gabbro and trondhjemitic granitoid rocks.

<u>Middle Ordovician</u> (Fig. 10.1b, c)

Although a great many complex, multi-stage models of ophiolite obduction have been proposed for western Newfoundland, the most logical method of ophiolite obduction is to gradually subduct the thin, leading edge of the continental margin. The aborted subduction of continental crust will result in the emplacement of the fore-arc region of a volcanic arc onto the continental margin. As the fore-arc region of many island arcs are floored by ophiolite (e.g., Seely and Dickenson, 1979; Hamilton, 1978), this represents a mechanically reasonable way of obducting large, thick sheets of dense oceanic crust onto the continental margin.

The obduction of ophiolitic rocks of the Humber Arm Allochthon is accordingly interpreted to have taken place by the partial subduction of the continental margin together with its carbonate bank cover sequence beneath the fore-arc region of the volcanic arc (Malpas and Stevens, 1977). Clastic rocks of the Humber Arm Allocthon are interpreted to represent an accretionary prism related to this subduction zone. Rocks in this accretionary wedge represent a sampling of the continental margin as it was gradually subducted. However, this sampling was confined to upper structural levels as the carbonate bank and clastic rocks east of the Humber Arm Allochthon are interpreted to be part of the autochthon. Evidently they were carried partway down into the subduction zone without being incorporated into the accretionary wedge. Autochthon and allochthon were apparently separated by a master thrust surface at this time. The DIII deformational event is correlated with this underplating episode. As the obducting oceanic lithosphere encountered the wedge of sediments flanking the continental margin it was thrust upwards and eroded locally forming coarse breccias. Arc-related volcanism was continuous throughout exhumation and the erosional surface was quickly mantled by volcanic material to form the Glover Group.

Loading of the continental margin by thrust slices of continental margin clastic material resulted in a widespread, tectonometamorphic event of classic Barrovian regional metamorphic type. The peak of metamorphism in the Glover Island Area corresponds to this loading event.

The DIV deformational event is correlated with the collision of the arc system with the continental margin. Ophiolite emplacement related structures and thrust surfaces were deformed and locally reactivated by this deformational event.

Subduction related volcanism ceased during decoupling as the axis of magmatic activity was displaced oceanward relative to its former position. The cessation of volcanism in the volcanic arc is marked by a graptolitic black shale of Caradocian age (Dean, 1978). The age of final emplacement of allochthons in western Newfoundland coincides with this cessation of volcanism based upon the Caradocian age of the neoauthochthonous Long Point Group (Bergstrom et al., 1974) that unconformably overlies the allochthon on the Port Au Port Peninsula (Rodgers, 1965). This linkage between volcanism and tectonism in western Newfoundland strongly supports the contention that both are subduction related.

<u>Silurian</u> (Fig. 10.1d)

During Silurian, the thickened orogen began to isostatically uplift and a widespread unconformity developed. Farther east, thick clastic sequences, derived from the uplifted terrane to the west were deposited in the rear-arc or retro-arc basin. Sedimentation in the uplifted area was probably confined to fault-bounded, successor basins.

<u>Subsequent history</u> (Fig. 10.1e)

Broad, open folding occurred during the Acadian Orogeny locally associated with the formation of an axial planar foliation. The DVI deformational event is inferred to have formed at this time. The Glover Anticline typifies the structures formed during the Acadian Orogeny as it is upright, broad and open and large in scale. High-angle faulting locally modified this complex terrane to produce the present day cross-section as shown in Figure 10.1e.

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APPENDIX A

MAJOR AND TRACE ELEMENT CHEMICAL ANALYSES

Chemical analyses and trace element abundances determined in this study are listed in tables at appropriate locations in this thesis. Sample locations for analyzed rocks are given in Plate II. Sample preparation and determinative methods are described below.

Major elements

Only relatively unaltered and undeformed samples were selected for analysis. These were carefully trimmed to eliminate veins or segregations of quartz, calcite or other secondary minerals. The sample was then washed, dried and broken into small chips. A disc mill was used to powder the chips.

Chemical analyses were done by G. Andrews (M.U.N.). The sample was first heated in a ceramic crucible to determine the loss on ignition then dissolved in concentrated HF. Aluminium, silica, titanium, iron, calcium, sodium, magnesium, potassium and manganese were determined by atomic absorption spectroscopy. Phosphorus was determined by photometric methods.

Analyses are given in this thesis with total iron expressed as ferric iron. These were recalculated to 100 percent anhydrous and total iron was recast as ferrous iron before they were plotted in various standard diagrams.

Trace elements

Samples were formed into standard pressed pellets for trace element analysis. Each sample was combined with a binding agent, placed in a shaker until mixed and then heated.

Trace element abundances were determined by x-ray flourescence by D. Press (M.U.N.). Calibration was done using known standards.

APPENDIX B

MICROPROBE ANALYSIS

Microprobe analyses determined in this study are given in tables at appropriate locations within this thesis. Sample locations are given in Plate II.

Analyses were made on carbon-coated, polished thin sections using a fully automated Jeol electron microprobe supervised by H. Longerich (M.U.N.). An accelerating voltage of 15 kv was used and measurements were made for 10 seconds per analysis. The data was corrected using the scheme of Bence and Albee (1968) and the correction factors of Albee and Ray (1970). Calibration of the microprobe was done using known standards.

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PLATE I 52

GEOLOGIC MAP OF THE 50°
GLOVER ISLAND AREA

1982 48°

Douglas A. Knapp
Memorial University of Newfoundland 46°

Scale 1:50,000

0 1 2 3 4 5 6 7 8 9 10 km

EXPLANATION

Geologic contact (known, inferred)

High-angle fault (known, inferred)

Thrust fault; teeth on upper plate (known, inferred)

Bedding; tops known (inclined, vertical, overturned)

Bedding; tops unknown (inclined, vertical)

Cleavage; generation shown by tic (inclined, vertical)

Igneous layering; tops known (inclined, vertical, overturned)

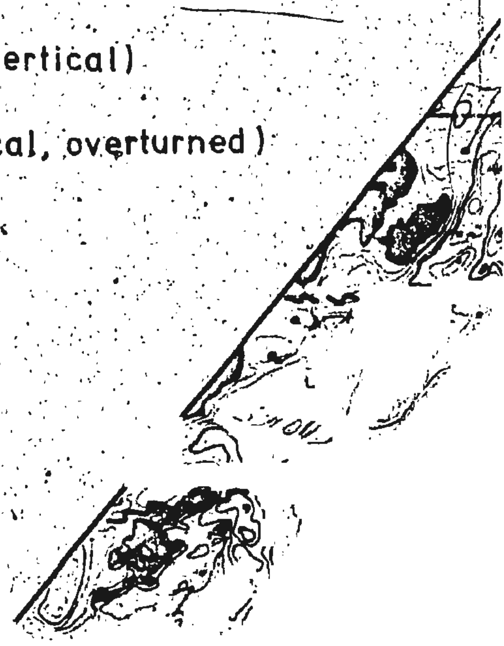
Synform (upright, overturned)

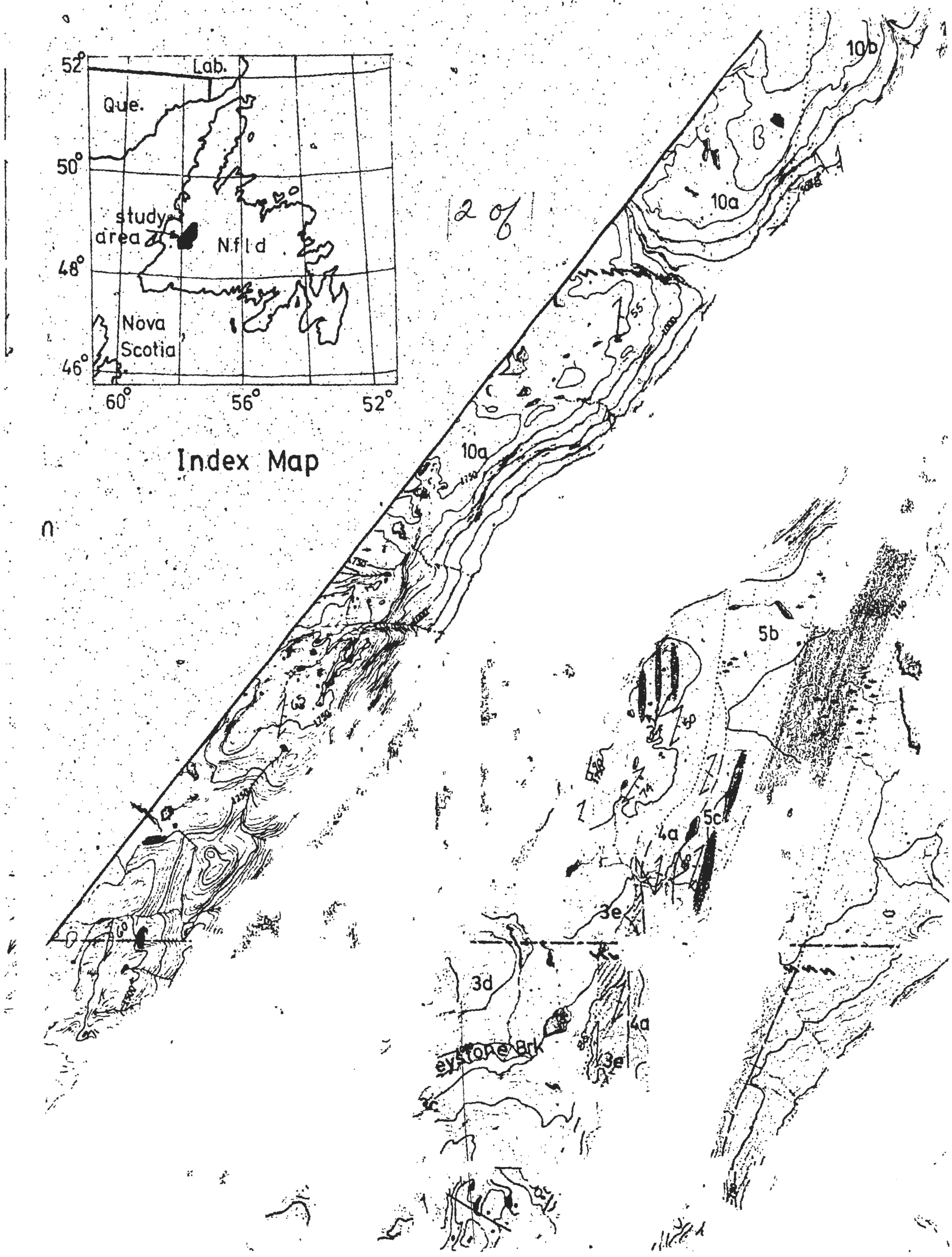
Antiform (upright, overturned)

F Fossil locality

CFZ Cabot Fault Zone

Biotite isograd




52°
Lab.
Que.
50°
study area
Nfld
48°
Nova Scotia
46°
60°
56°
52°
Index Map
2 of
10b
10a
10a
5b
5c
4a
3e
3d
4a
3e
eystone Brk

E X P L

WEST SHORE OF GRAND LAKE

CARBONIFEROUS

[13] ANGUILLE GROUP: Grey siltstone, sandstone, conglomerate and impure limestone; black carbonaceous mudstone.

DEVONIAN AND SILURIAN

ORDOVICIAN

3 g

[] Ophiolitic rocks; , serpentine and talc-carbonate schist; , greenschist; , massive metabasalt.

ORDOVICIAN AND CAMBRIAN

[10] Psammitic and semipelitic schist and gneiss; 10a, albite schist and gneiss; 10b, quartz-rich psammite and minor amphibolite.

PRECAMBRIAN

[9]
met
pur

sil
ph

7
op
br
dy

L A N A T I O N

SOUTH END OF GLOVER ISLAND

GLOVER ISLAND AND CORNER POND

44

9 RED POINT FORMATION: Green and tan
ietasandstone and metaconglomerate;
urple phyllite and metagreywacke.

Intrusive rocks; 8a, metagabbro; 8b,
ilicic metaporphyry; 8c, mafic metapor-
hyry; 8d, metatrondhjemite.

OTTER NECK GROUP: Greenschist facies
hiolitic rocks; [illegible], pillow lava; 7b, pillow
eccia and aquagene tuff; 7c, sheeted
ykes.

12 Intrusive granitoid rocks.

6 CORNER POND FORMATION: Tan metasandstone and metaconglomerate; grey, black and green slate; red silicic tuff; green and purple pillow lava.

GLOVER GROUP (4-5)

5 TUCKAMORE FORMATION: Greenschist facies, mafic and silicic volcanics; 5a, pillow lava; 5b, mafic tuff and volcaniclastic sediment; [illegible], silicic tuff; 5d, breccia; 5e, purple and black argillite; 5f, gabbro.

4 KETTLE POND FORMATION: Quartz-sericite schist and metaconglomerate; 4a, thinly layered quartz-sericite schist; [illegible], polymictic metaconglomerate.

3 GRAND LAKE COMPLEX: Ophiolitic rocks; 3a, basal greenschist; [illegible], serpentinized ultramafic rocks; 3c, talc-carbonate alteration of 3b; 3d, metagabbro; ~~3e~~, metatrondhjemite; [illegible], metabasalt dykes; 3g, greenschist.

2 KEYSTONE SCHIST: Semipelitic and psammitic schist and gneiss; 2a, microcline schist, quartz-muscovite schist, metaconglomerate and amphibolite; 2b, biotite schist, quartz-sericite schist and carbonaceous schist.

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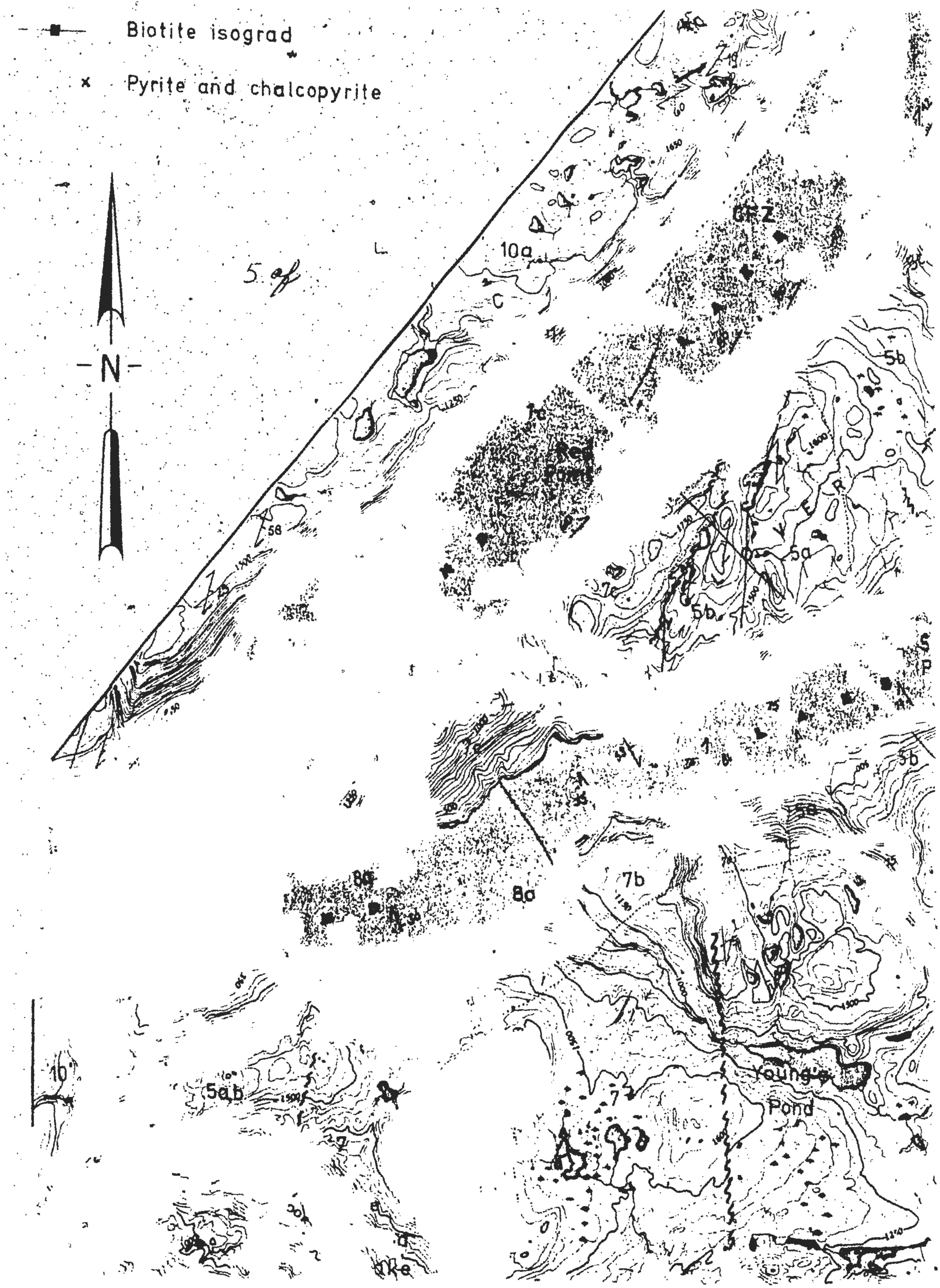

Biotite isograd
Pyrite and chalcopyrite
N
5 of
10a
GFZ
5b
5a
7c
8a
7b
Young's
Pond
5ab
10


I S L A N D
5c
5b
5b
5a
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6
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6 of
1000
500
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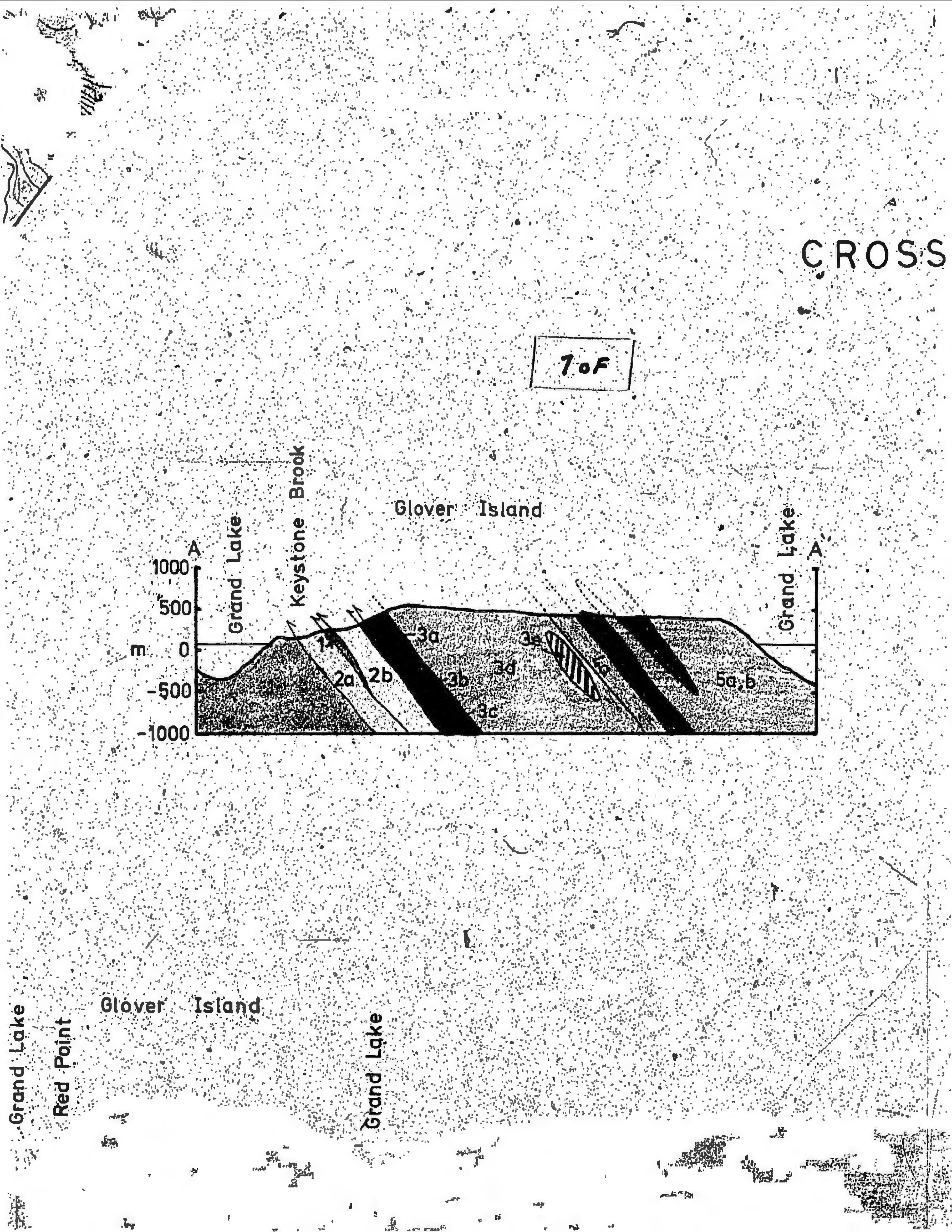
-500
m
10a


CROSS
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Glover Island
Grand Lake
Keystone Brook
Grand Lake
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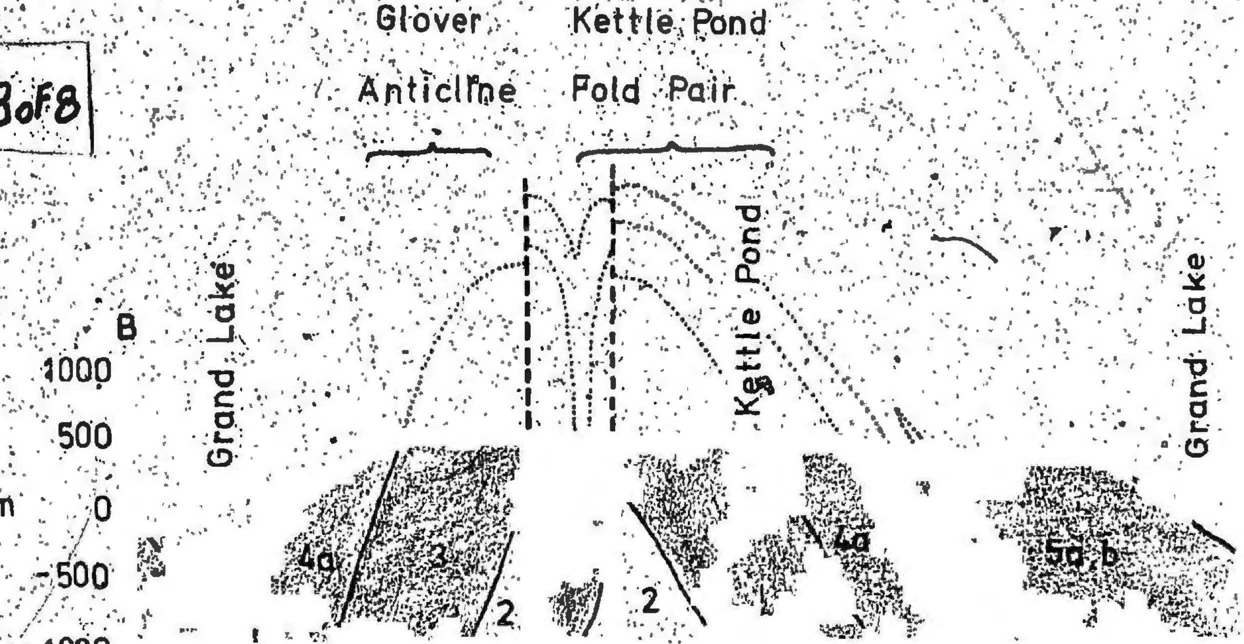
1000
500
m 0
-500
-1000
1
2a
2b
3a
3b
3c
3d
3e
4a
5a,b
Glover Island
Grand Lake
Red Point
Grand Lake

1 COBBLE COVE GNEISS: Quartzo-dspathic microcline gneiss and biotite-microcline mafic schist layers.

SS' - SECTIONS

8 of 8



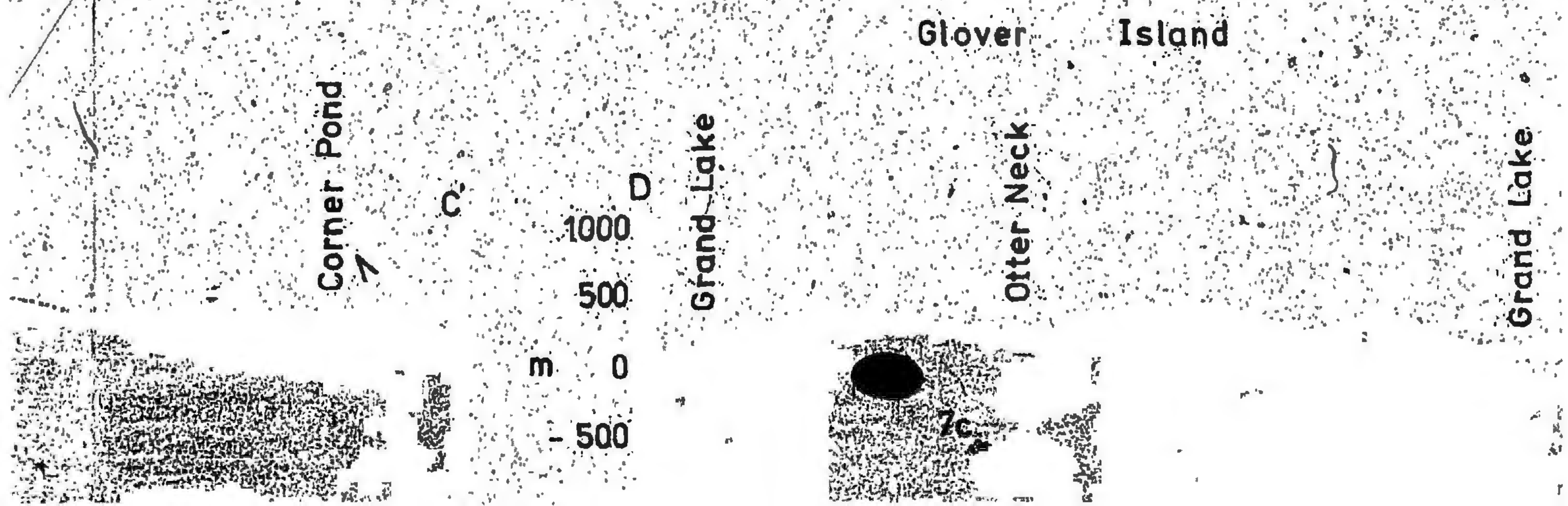


PLATE II

SAMPLE LOCATION MAP OF THE GLOVER ISLAND AREA

1982

1 of

Douglas A. Knapp
Memorial University of Newfoundland

Scale 1:50,000

0 1 2 3 4 5 6 7 8 9 10 km

EXPLANATION

Geologic contact

High-angle fault

Thrust fault

CFZ Cabot Fault Zone

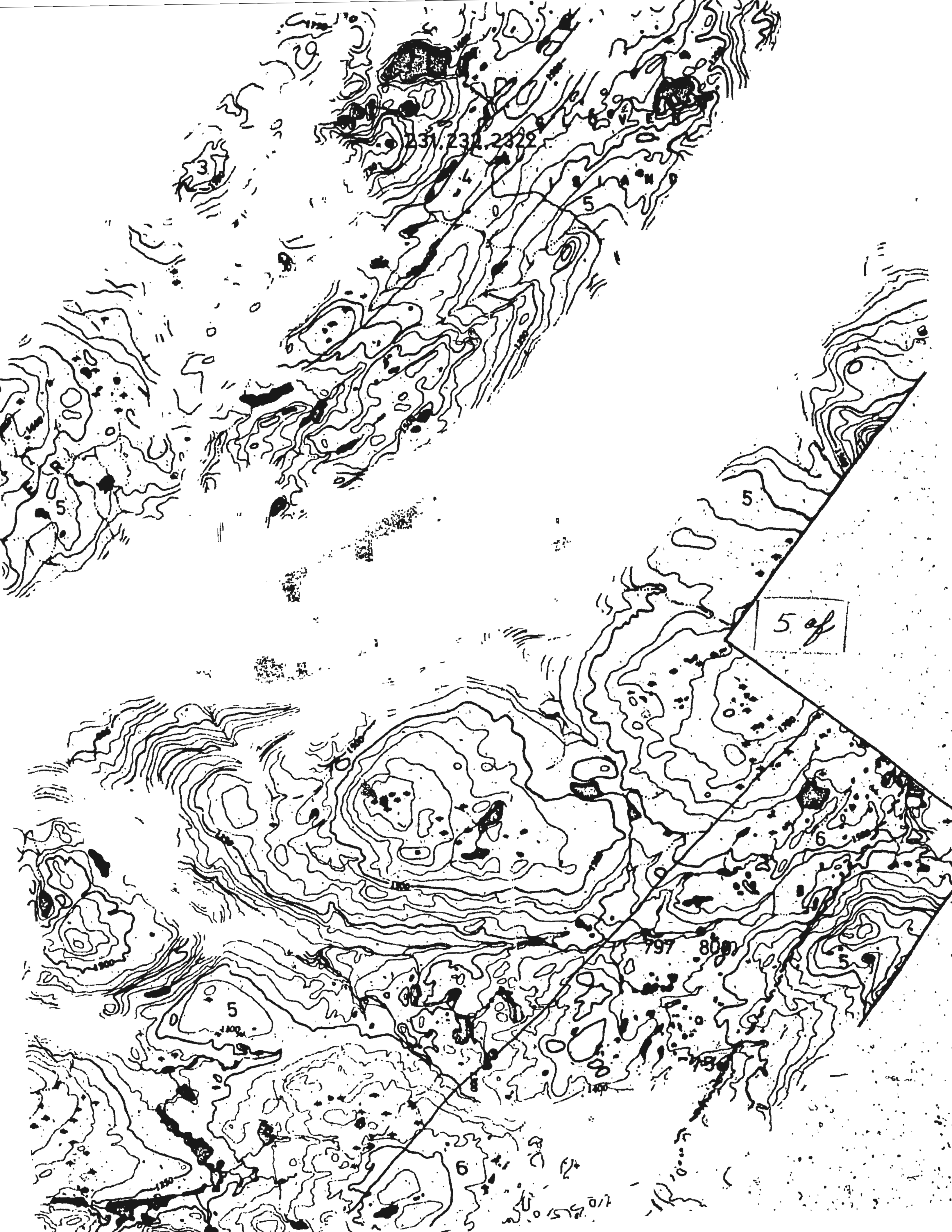
1, 2, etc. Geologic units, see Plate I

345 Sample locations

2 of


13
3
10
5
GLOYER
12
2082
208
5
3
206
3
5
2
2185, 2189, 21810
5
12
3

44



231,232,2322
3
4
5
5 of
6
797
800


5
4
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797
801
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6