THESIS

THE EVOLUTION OF THE BAIE VERTE LINEAMENT, BURLINGTON PENINSULA,
NEWFOUNDLAND

Submitted for the degree of Ph.D.

by

William Spencer Francis Kidd
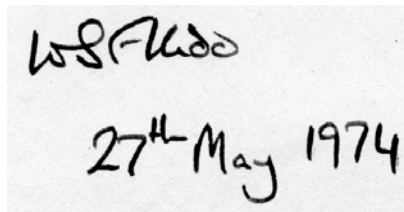
of

Darwin College, Cambridge

1974

DECLARATION

I declare that the contents of this thesis have not been submitted
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submitted for any other degree,etc.
The contents of this thesis are my own work, done without collaboration
except where specifically stated.

Signed,

27ᵗʰ May 1974

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"Below its beautiful surface, the entire region is a badly confused
chunk of Mother Earth . . ."

Tourist guide to the Burlington Peninsula.

Frontispiece: Flatwater Pond, looking southeast towards Skippens Ridge.

Summary

An area measuring about 5 km by 30 km, of the relatively well-exposed
central part of the Baie Verte Lineament, has been mapped in detail. Two
major rock divisions are recognised by their differing structural and
metamorphic histories. A terrain of psammitic, semipelitic and mafic schists
borders the western side of the Lineament. These rocks (part of the Fleur
de Lys Supergroup) are in upper greenschist to epidote-amphibolite metamorphic
facies, and are affected by a polyphase deformation sequence. They
are intruded by post-kinematic granite and tonalite. A large, generally
unfoliated, granodiorite (Burlington Granodiorite) borders the eastern side
of the Lineament. Outside the map area, it is seen to be either a pre- or
early syn-kinematic intrusion into Fleur de Lys schists.
The Baie Verte Lineament consists, in the mapped area, of two adjacent
narrow-parallel belts; to the west the Baie Verte Group of uncertain (?Arenigian)
age, and to the east the early Devonian Mic Mac Lake Group. Large
lensoid "alpine-type" (ophiolitic) ultramafic bodies are situated along the
tectonic contact between the Fleur de Lys schists and the Baie Verte Group.
A narrow discontinuous strip containing bodies of ophiolitic gabbro with
parallel diabase dykes is found at the western side of the Baie Verte Group
adjoining the tectonic contact. The internal structures and textures of the
ophiolitic rocks are described. The Baie Verte Group consists mainly of mafic
pillow lava and mafic volcaniclastic sediments, subvertical to moderately
west-dipping, and facing east, for which a stratigraphy is defined. At or

near the base, a conglomerate, resting on ophiolite gabbro megabreccia,
contains clasts mainly referable to the mafic parts of an ophiolite complex,
with some clasts identical to the Burlington Granodiorite, and rare clasts
referable to previously deformed silicic volcanics and related sediments of
the Fleur de Lys Supergroup found to the east of the map area. Redeposited
carbonate rocks are very occasionally found near this horizon. The Mic Mac
Lake Group rests with spectacular unconformity, including a significant amount
of palaeotopographic relief, on the Burlington Granodiorite. It also contains
a similar erosion surface within the subaerial sequence of silicic volcanics,
mafic and trachyte lava flows, and mostly conglomeratic sediments. A stratigraphy
is defined for the Mic Mac Lake Group. The sequence dips west at
moderate angles and, with the significant exception of a narrow strip at the
western side, faces west. The Baie Verte and Mic Mac Lake Groups share a
single steep cleavage and low greenschist facies metamorphic grade.
An autochthonous contact between Baie Verte Group and east-facing Mic
Mac Lake Group in the southern part of the area indicates that the Mic Mac
Lake Group is in a highly-disrupted syncline, and is unconformable on the
Baie Verte Group. To the north an attenuated Mic Mac Lake Group is overthrust
by the Baie Verte Group, and must eventually be cut out to the north of the
map area where the Baie Verte Group directly overthrusts Burlington Granodiorite.
It is shown that the deformation of the Baie Verte and Mic Mac Lake Groups,
and the tectonic emplacement of the ophiolitic ultramafic bodies into their
present position, was wholly later than all the regional polyphase deformation
and metamorphism of the Fleur de Lys Supergroup, and that this later
deformation is almost wholly localised in the belt of less complexely deformed
rock.

Joint mapping with J.M. Bird and J.F. Dewey at the northeast end of
the Baie Verte Lineament demonstrated the presence of all members of a full
ophiolite suite, overlain by mafic volcaniclastic sediments and a thick pile
of pillow lava. The internal igneous-relationships of the ophiolite complex
are described. These rocks are disposed in three thrust sheets (two being
inverted), which are overthrust eastward above previously deformed and metamorphosed
Fleur de Lys schists.

One model has been proposed that coherently interprets the Palaeozoic
evolution of the central and western Newfoundland Appalachians in terms of
present-day continental margin / island arc sedimentation, magmatism, and
tectonics. The results of this mapping are interpreted within the framework of
this model. The Fleur de Lys schists and Burlington Granodiorite represent
rifted continental margin sediments, and a volcanic arc built on and intruding
them prior to polyphase deformation and metamorphism. The Baie Verte
Group is interpreted as the remains of the oceanic crust and mantle floor and
mafic volcanic fill of one of several small marginal (inter-arc) basins that
developed in the northwestern Newfoundland area in lower Ordovician times.
It is interpreted to be in the place where the basin formed relative to the
eastern and western Fleur de Lys blocks that border it. The Mic Mac Lake
Group is a proximal section to an early Devonian calc-alkaline cordilleran
caldera complex which may have been related either to subduction or to continental
collision processes. It was laid down over an eroded surface of
Burlington Granodiorite and over little to undeformed mafic rocks filling
the Baie Verte marginal basin. Both were then deformed during the Acadian
continental collision orogeny by contraction of the Baie Verte basin, with
medium to high angle eastward overthrusting.

Notes for the reader.

Thin sections and corresponding rock samples used for this study

are deposited in the Harker collection of the Department of Mineralogy and

Petrology, University of Cambridge. (Catalogue no. 33, p. 251-253, 255-

283; serial numbers 110403-110456, and 110476-110990).

The magnification quoted in captions to photomicrographs is the

true scale of the print to the thin section. The prints are about 85 X 57

mm., and their longest side therefore represents about 8.5, 3.4, 2.1 and

0.85 mm. for magnifications of X 10, X 25, X 40 and X 100 respectively.

The camera lens cap seen in photographs of outcrop surfaces is 45 mm across.

Pagination of thesis of W.S.F. Kidd

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CHAPTER 1 INTRODUCTION

This thesis is concerned with a long narrow area of Lower Palaeozoic rocks, possessing complex

tectonic relationships, that form part of the northwestern side of the northern Appalachian

'orogenic belt' in Newfoundland. The map area is in the Burlington Peninsula, the triangular

piece of land adjoining the southeast end of the Long Range Peninsula (Fig. 1.1).

(i) Physical features of the Burlington Peninsula

The Trans-Canada Highway skirts the southern end of the peninsula, and the Baie Verte Road

runs 64 km NNE from it to the town of Baie Verte. Subsidiary dirt roads branch off it at

Flatwater Pond and at Baie Verte, leading to settlements ('outports') on the east, west and

northeastern coasts of the peninsula (Fig. 1.2), and another road leads north from Baie Verte to

the asbestos mine, and to two settlements at the northern end of the peninsula.

The topography of the peninsula is, in general, relatively subdued, except near the rocky cliffed

coast, which possesses some small fjords; the land is almost always 500 to 700 feet (150 to 220

m.) high within a mile of the coast all round the peninsula. The eastern part of the peninsula has

a summit level rarely exceeding 800 feet (250 m.). In the northwestern and west central parts of

the peninsula, the land gradually rises to a summit level of about 1100 feet (340 m.). In the

southern part of the peninsula, summits are commonly about 1300 feet (400 m.), and the highest

point in the peninsula (1575 ft. 475 m. A.S.L.) occurs in the southeastern part. Although the

topography inland is generally subdued, a few isolated escarpments occur on some whaleback

ridges, which are essentially large *roches moutonnées*.

The drainage of the peninsula is fairly typical of the immature, disrupted type that occurs where

recently glaciated hard bedrock is at the surface; numerous small ponds and bogs occupy hollows

between low rock ridges

Fig. 1.1. Location of the Burlington Peninsula.
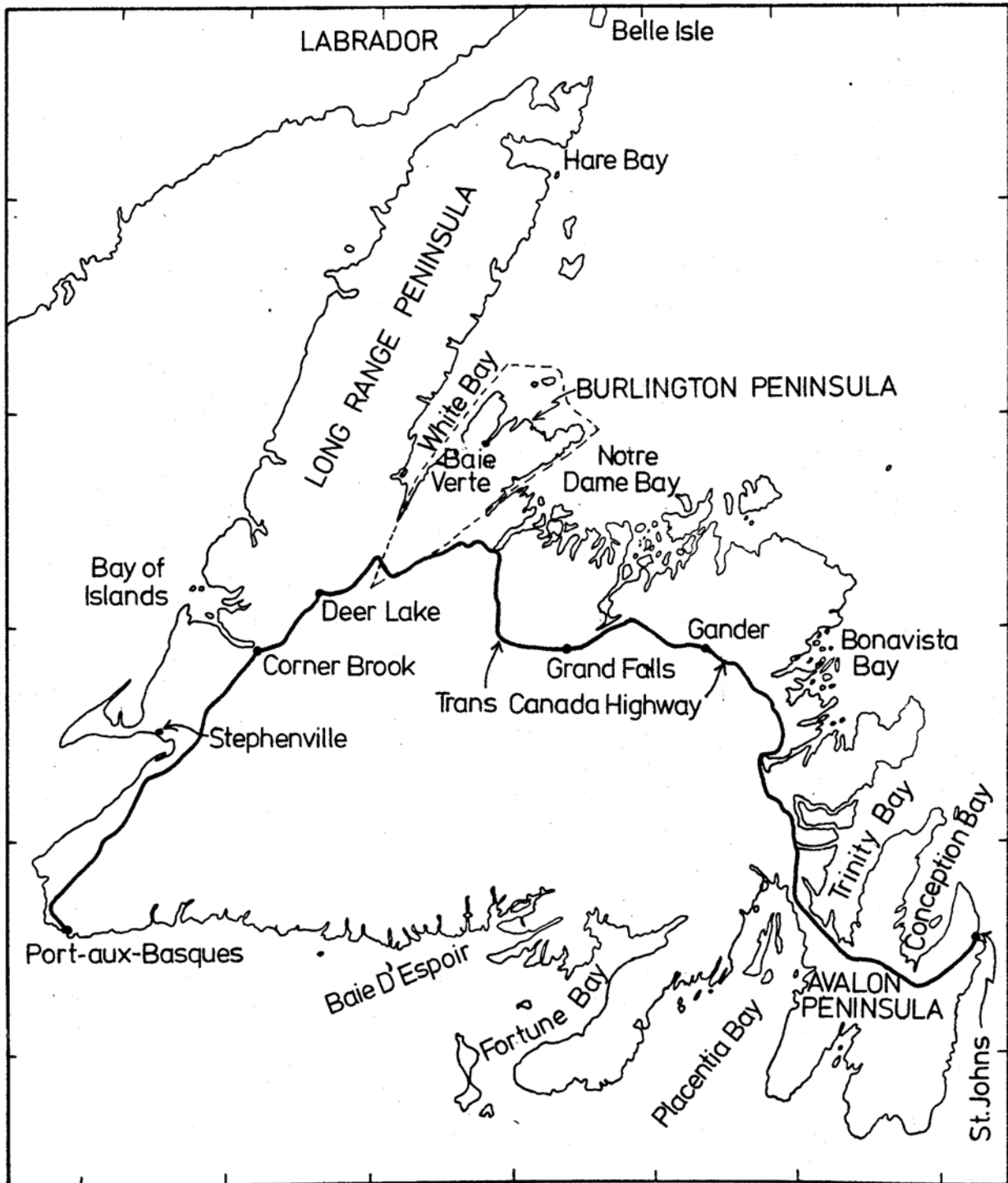
Fig. 1.2. Physiography of the Burlington Peninsula.
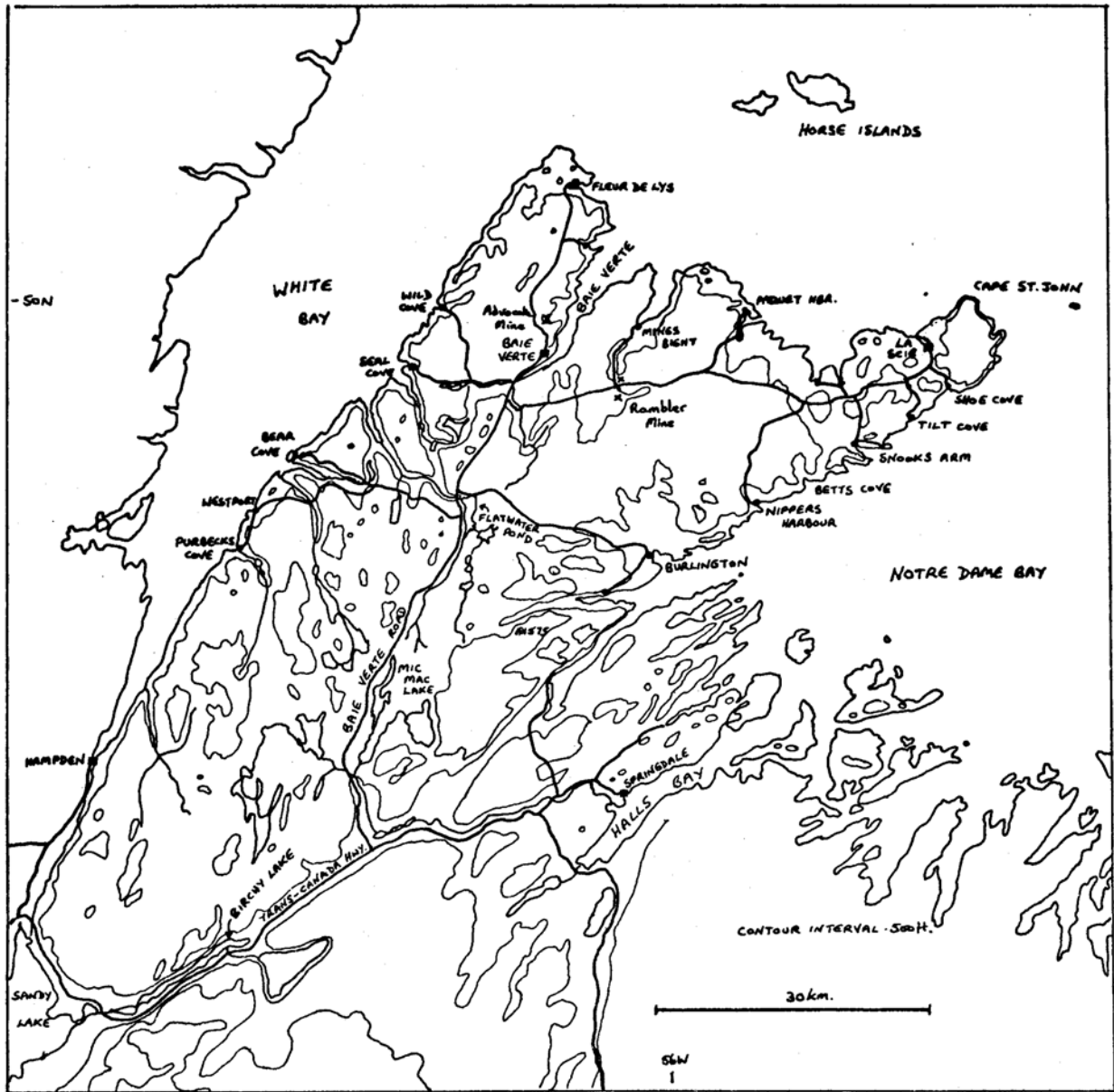
and larger streams are far from their base levels. Some large lakes occur, and two of these, Mic Mac Lake and Flatwater Pond (Fig. 1.2), occupy the central subdued valley that follows the outcrop of the Baie Verte Lineament.

(ii) Superficial deposits and physical features of the map area

The area mapped inland extends in a strip averaging 5 km wide from the centre of Mic Mac Lake to beyond the northern end of Flatwater Pond, and its boundaries and size are described in a later section.

The latest Pleistocene glaciation has left very little till in the area; a thin, very patchy veneer occurs, mainly in the boggy hollows between ridges. Glacial erratics are scattered over the whole area; almost all are locally derived, and very few exceed two metres in diameter. A centre of ice dispersal is thought to have been on the highland to the south of the peninsula, at least in the last glacial episode (Neale and Nash, 1963). Indicators of glacial flow direction seen in the map area, such as striae and *roches moutonnées*, show that the flow in this episode followed closely the existing topography. Thus, the main ice flow was up the Baie Verte Lineament to the NNE, and distributary flow occurred down valleys leading to small fjords of the west and east coasts, where it joined two other main glacial flows down White Bay and Green Bay. The only significant deposit of drift in the map area occurs at the south end of Flatwater Pond.

Most of the Burlington Peninsula is thickly forested, with the exception of rocky 'barrens' of silicic igneous rock in its southeastern and northeastern parts.

In 1959, a forest fire started at Mic Mac Lake in the southern part of the peninsula, and burnt a swath NNE about 30 km long averaging 5 km wide, essentially following the broad subdued valley of the Baie Verte Lineament. Bowaters salvaged the dead timber, and in doing so incidentally made the area very much easier to traverse; the few remaining uncleared patches display

a cautionary waist-high tangle of fallen dead trees to the passer-by.

During the logging operation, many dirt roads were constructed, and this made access to most of the area exceedingly easy, although the bridges and culverts are rapidly deteriorating. Far more outcrop is presently available in the burnt area than before the fire, and it is extremely clean, being similar to wavewashed coastal exposures. The contrast with the sparse, lichen and moss covered, tree root-entwined outcrops in the forested areas could hardly be more extreme. The burnt area controlled, to a large extent, the area that was eventually mapped.

The frontispiece shows typical aspects of the physiography of the map area. Relatively good outcrop occurs in places along the power transmission line; in roadcuts on the Baie Verte Road and subsidiary roads; on the northern shores of Flatwater Pond; and in the area burnt by the forest fire, seen in the frontispiece where abundant white outcrops of Burlington Granodiorite occur on and beyond the farther (southeastern) shore of Flatwater Pond. Outcrop is relatively poor to non-existent in forested areas.

Fig. 1.3. Geological Sketch map of the Burlington Peninsula.
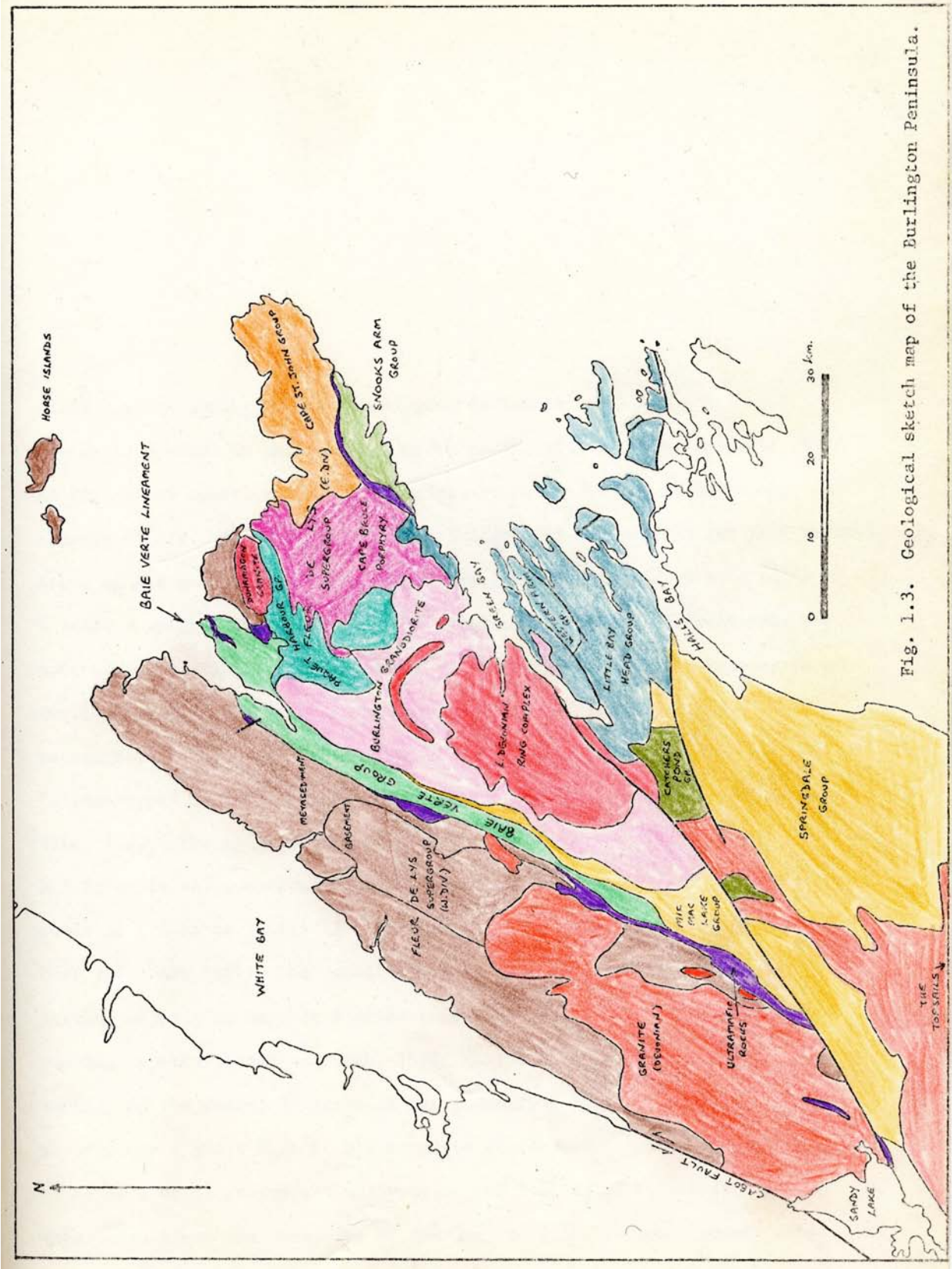
Fig. 1.3. Geological sketch map of the Burlington Peninsula.

(iii) Outline geology of the Burlington Peninsula

Early studies in the Burlington Peninsula, of small and separated areas adjoining the coast, are well summarised by Neale (1957) and Neale and Kennedy (1967). One of these studies (Watson, 1947) involved the Baie Verte .Group on the north coast of the Peninsula, and is referred to more fully in a later chapter. The first map of a substantial part of the peninsula was made by Baird (1951). In the course of compiling exploration company data for inclusion in the first geological map of Newfoundland, Baird (1954) recognised the presence of the Baie Verte Lineament from the line of lensoid 'alpine-type' ultramafic bodies running down the length of the peninsula (Fig. 1.3). The whole Burlington Peninsula was systematically mapped by E.R.W. Neale and coworkers from 1956 to 1961, and mostly published on a scale of 1 inch to 1 mile (1:63;360), (Neale 1958, 1959a, b, Neale, Nash and Innes 1960); the results of this work are contained together with further mapping on maps at 4 miles to the inch (1:253,440) with accompanying reports (Neale and Nash, 1963, Williams 1962). All subsequent mapping in the peninsula rests on the foundation laid by Neale, and all workers owe a great debt to his accurate field-work. Since the publication of Neale's maps, geological activity in the Burlington Peninsula has increased, especially since the relevance of the area to plate tectonic models was

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Table 1. Pre-Carboniferous geological events of the Burlington Peninsula.
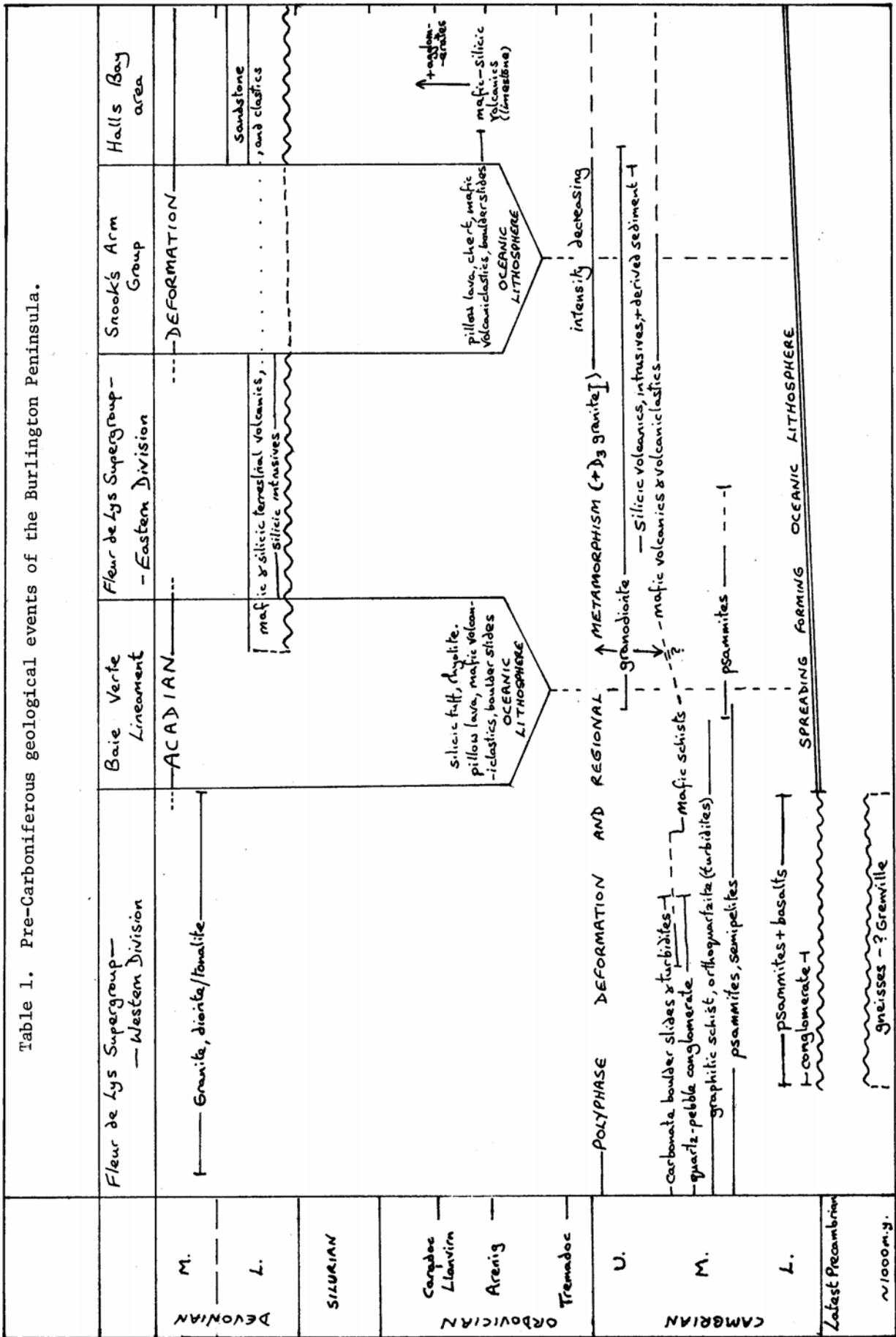
realised (Dewey, 1969, Bird and Dewey, 1970).

The Burlington Peninsula contains four main rock associations; in order of relative age these are, (1) a terrain of regionally metamorphosed and complexly-deformed rocks (Fleur de Lys Supergroup), which includes large silicic early intrusions (Burlington Grandiorite, and Cape Brulé Porphyry); (2) two belts of less-complexly deformed, low metamorphic grade, submarine mafic volcanic rocks, associated with ophiolite complex lithologies of which the most prominent are the linear belts of ultramafic bodies; (3) subaerial silicic and mafic volcanic rocks with coarse terrestrial sediments, and a silicic intrusive complex which was the immediate source of the silicic volcanics; and (4) a large granite-diorite intrusive complex. The ages of the rocks and the events that have affected them range from about latest Pre-Cambrian to Devonian; they are shown in more detail in Table 1 and discussed below.

Perhaps the most striking gross feature of the geological map of the peninsula is the Baie Verte Lineament, the narrow belt of ultramafic rock bodies adjoined to the east by mafic volcanic rocks, which divides the peninsula into two halves. The western part of the peninsula is underlain by a thick sequence mainly of psammitic, semipelitic, and mafic metasedimentary schists, the western division of the Fleur de Lys Supergroup (Church 1969), together with some remobilised basement gneisses of probable Grenville (~1000 m.y.) age (M. J. de Wit, in preparation). The presence of a gneissic basement indicates that most of the Fleur de Lys schists in this part of the peninsula were deposited on continental-type lithosphere, but this is not to say that it was necessarily connected to the continental lithosphere of the North American shield at the time the Fleur de Lys sediments were laid down, or that at this time the crust was of the typical thickness of continental crust. All these rocks in the western part of the peninsula have suffered

polyphase deformation and accompanying regional metamorphism, generally reaching epidote-amphibolite facies. Aluminosilicate minerals occur in a few separated localities, but are not common. The overall structural trend parallels the western coast of the peninsula, and the dominant foliation is steep, except near the northern end of the peninsula, where it is more flat-lying. Similar psammitic and mafic schists, deformed mafic metavolcanics and silicic metavolcanic and metaplutonic rocks occur in the northeastern part of the peninsula, and form the eastern division of the Fleur de Lys Supergroup (Church 1969). As the silicic metaplutonic rocks (Cape Brulé Porphyry) contain abundant and large inclusions of ophiolite complex lithologies (Neale 1957, l959a) it is likely that these rocks were deposited mainly on oceanic lithosphere (Dewey and Bird, 1971). The intensity of deformation and grade of metamorphism of the Fleur de Lys rocks along the northeast coast of the eastern part of the peninsula is about the same as in the western division, but southwards across the eastern part of the peninsula the metamorphic grade drops to greenschist facies, the overall intensity of the deformation is reduced, and it changes its character from a bulk to a zonal type of deformation. The structural trend in the eastern division of the Fleur de Lys Supergroup is about east-west, and generally dips gently north, which contrasts with the western division of the Fleur de Lys Supergroup. There is no direct evidence for the age of the Fleur de Lys metasediments and metavolcanics. Very general regional lithologic correlation, and regional tectonicistratigraphic evidence, suggest that their age is latest Pre-Cambrian, Cambrian and perhaps lowest Ordovician.

A large proportion of the eastern half of the peninsula is occupied by the Burlington Granodiorite (Baird 1951). The relationship of this plutonic complex to the Fleur de Lys Supergroup is somewhat uncertain, but

several observations by the writer suggest that it is either a pre- or early syn-kinematic intrusion into the schists of the Fleur de Lys Supergroup, so it is probably of late Cambrian to earliest Ordovician age. That this age is an upper limit is suggested by the presence of clasts of this fairly distinctive granodiorite in a conglomerate in the Baie Verte Group (Neale and Kennedy, 1967), and by the occurrence of debris possibly derived from the granodiorite in the Arenig Snooks Arm Group (Church, 1969).

The Baie Verte Lineament, which runs for about 90 km NNE up the whole length of the peninsula, is most clearly defined by the string of large lensoid 'alpine-type' ultramafic bodies, which in most cases abut the Fleur de Lys schists to the west. Along the lineament, except in the south, these are adjoined to the east by the low greenschist facies, simply deformed, steeply dipping submarine mafic volcanic and mafic volcaniclastic rocks of the Baie Verte Group. The internal structure of the rocks parallels the lineament. The ultramafic bodies, together with some of the other rocks of the Baie Verte Group, are parts of a variably dismembered ophiolite complex. A large fault, the Baie Verte Road Fault, has been postulated along the western side of the Lineament, but its sense and amount of displacement were not known (Neale and Kennedy, 1967). The age of the Baie Verte Group and the associated ophiolite complex (oceanic crust and upper mantle) is not known; the lithologically similar Snooks Arm Group is of lower Ordovician (Arenig) age.

The Baie Verte Group is adjoined to the east along the central and southern half of the Lineament by the Mic Mac Lake Group. This rests unconformably on the Burlington Granodiorite to the east (Neale and Nash, 1963), and consists of ignimbrites and rhyolites, mafic and minor trachyte lavas, and coarse terrestrial sediments. Neale, Nash and Innes (1960)

described the relationship of this Group to the Baie Verte Group as a fault obscured by rhyolite sill intrusions, but in contrast Neale and Kennedy (1967) interpreted the Mic Mac Lake Group as conformably underlying the Baie Verte Group. The Baie Verte and Mic Mac Lake Groups appear to share the same simple deformation and incipient low greenschist facies metamorphism (Neale and Kennedy, 1967). The depositional age of the Mic Mac Lake Group is probably early Devonian, as two independent determinations of whole rock Rb-Sr isochrons both give an age of about 395 m.y. (R. K. Wanless, in Neale and Kennedy, 1967), (I. R. Pringle, pers. comm.). In the southeast of the peninsula, a large elliptical complex contains silicic volcanics, hypabyssal silicic intrusives and syenite intrusions. It represents the roots of an eroded (multiple) caldera complex, the source for the silicic volcanics of the Mic Mac Lake Group (Neale and Nash, 1963; Neale and Kennedy, 1967; Neale, in prep.). A half ring dyke occurs to the north of this complex. Most of the southwestern part of the peninsula is occupied by granite and diorite that is post-kinematic with respect to the Fleur de Lys deformation; a K-Ar biotite age of 358 m.y. obtained from the granite (Neale and Nash, 1963) may not represent the emplacement age.

The Snooks Arm Group (Neale, 1957) occurs on the eastern side of the peninsula, and like the Baie Verte Group, contains alpine-type ultramafic rocks that abut rocks of the Fleur de Lys Supergroup to the northeast with fault contact. The bulk of the Group consists of three horizons of submarine mafic volcanics, with two horizons of mafic volcaniclastics, chert and banded pelitic sediment. Poorly preserved graptolites from a black pelite horizon indicate an Arenig age (Snelgrove, 1931). The lower horizon of pillow lava forms the top of the ophiolite complex that lies at the base of the Group; in one place the full ophiolite complex sequence is intact.

The Snooks Arm Group was therefore based on oceanic lithosphere of probable Arenig age (Upadhyay, Dewey and Neale, 1971). The rocks of the Snooks Arm Group are steeply dipping, but are not as strongly-deformed as the rocks of the Baie Verte Group.

The Burlington Peninsula is bounded on the west by the White Bay-Hampden-Grand Lake fault, part of the Cabot palaeotransform fault system of Carboniferous age (Wilson, 1962). From regional geological considerations, it is this author's opinion that the motion on this fault is left lateral, and that the amount of displacement is probably of the order of 100-200 km. To the south and east, the peninsula is bounded by a small transcurrent fault, the Green Bay-Birchy Lake fault, with a dextral offset of 20 km (Dewey and Bird, 1971). At the north coast of the peninsula, structures and rock groups disappear under the Atlantic ocean, and little is known about the local offshore continuation. Fleur de Lys Supergroup metasediments reappear briefly on the Horse Islands, 20 km offshore (Fig. 1.3). The available aeromagnetic maps show that the expression of the Baie Verte Lineament in terms of highly magnetic ultramafic bodies stops abruptly at the northern end of the peninsula. The maps also show that the structural trend in the western Fleur de Lys block curves quickly around from NNE-SSW to E-W immediately beyond the northern end of the peninsula.

(iv) Problems investigated; location and limits of map area.

The following problems were investigated during this project and are discussed in this thesis:

1. The determination of the relationships of the Baie Verte Group to the Fleur de Lys Supergroup, and to the Burlington Grandiorite, with special emphasis on the nature and development of the contacts between them.

2.   The investigation of the nature and relationships of the 'alpine-type' ultramafic rocks, and their association with the contact between the Baie Verte Group and the western Fleur de Lys Supergroup. This problem became generalised into an investigation of the ophiolite complex rocks when the other members of the suite were recognised within the Baie Verte Lineament.

3.   The determination of the stratigraphic relationship between the Baie Verte and Mic Mac Lake Groups, and the investigation of their structural relationships.

4.   The definition of a stratigraphy in the Baie Verte and Mic Mac Lake Groups, and the documentation of the palaeotopographic relationships in the latter.

During the course of this project, the rapid development of plate tectonics and its corollaries, and their application to the Newfoundland Appalachians, provided an added stimulus to the investigation of these problems. In particular, the resulting actualistic hypotheses helped considerably in both investigation and synthesis.

This thesis is concerned with the results of detailed mapping; relatively little was known (in definitive terms) about the contents and their relationships of all the six major rock units in the area when it was chosen for investigation. The author considered that a study of the chemistry of the volcanic rocks would be premature, until their detailed distribution and visible properties were known. Also, whole rock analyses of altered volcanic rocks are known to be of dubious significance; this point is discussed later. Thus the techniques used consist of detailed lithologic and structural field mapping, assisted by the examination of about 600 thin sections. In the investigation of the problems listed previously, the main effort in this

Fig. 1.4. Location of detailed map areas.
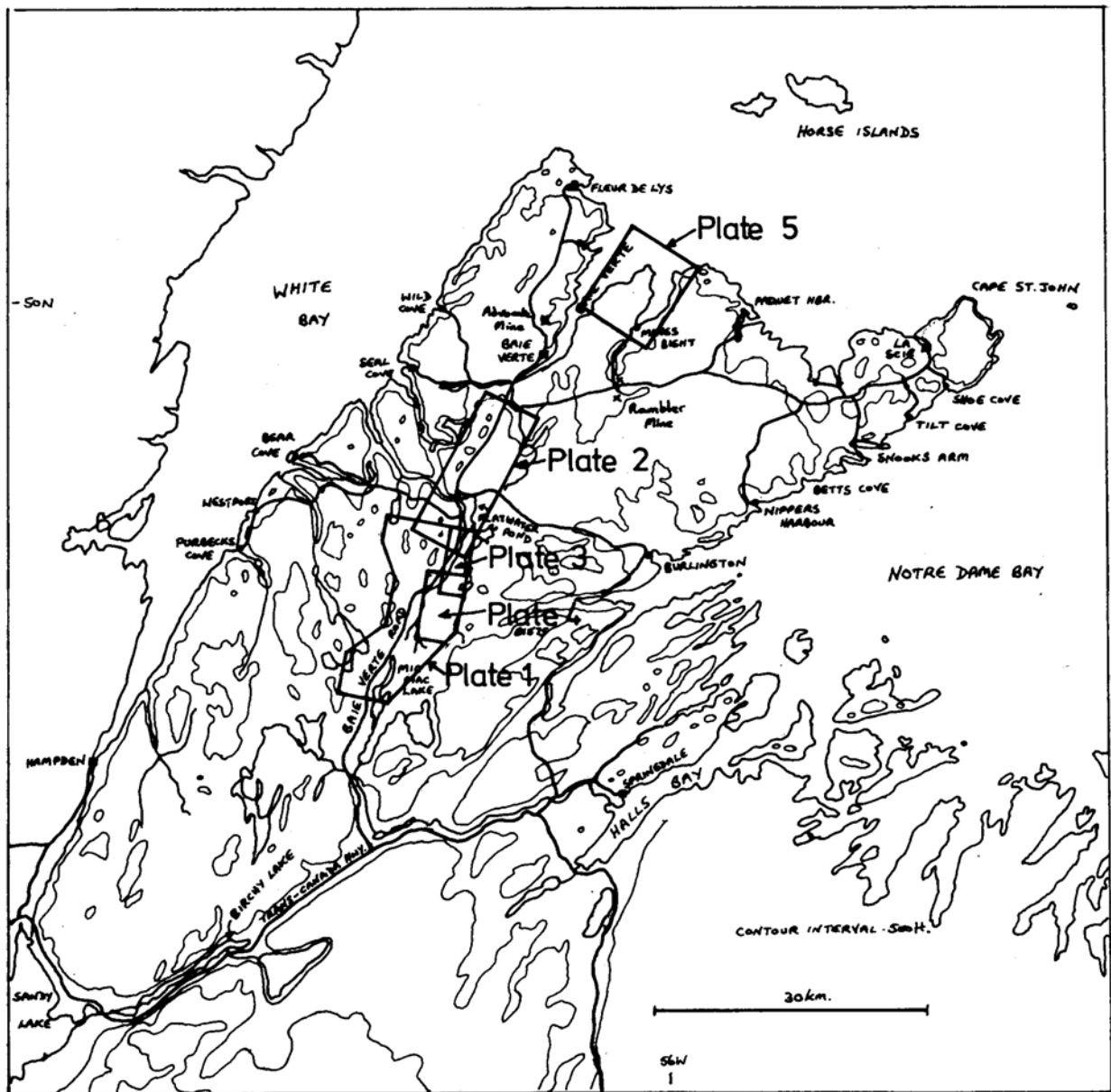
project was directed toward the Baie Verte and the Mic Mac Lake Groups. While a large amount of effort had to be expended in mapping the Fleur de Lys schists, they were investigated from the point of view of using them to help solve the problems relating to the Baie Verte Group, and its relationship with them, not as an end in themselves. Similar comments apply to the Burlington Granodiorite.

The detailed mapping is presented as a series of outcrop maps (Plates 1-5), whose locations are shown on Fig. 1.4. The inland area, where most of the time was spent, has an area of approximately 150 km^2. The total length of the area of detailed mapping shown on Plates 1 and 2 is 36 km., and the width varies from almost nothing along the Baie Verte Road in the north, to a maximum of about 8 km. Reconnaissance traverses were made in a few places up to 5 km west and 12 km south of the southern end of Plate 1, and the results from these traverses are included on an overall summary geological map (Plate 6). The large scale maps of part of the Mic Mac Lake Group (Plates 3 and 4) cover an area of about 13 km^2, which is 11 km long and between 0.4 and 2 km wide. The northern map (Plate 3) has been mapped solely by this author; the southern map (Plate 4), comprising the bulk of this area of very detailed mapping, was first mapped under the author's guidance by his assistant in the 1970 season, P. A. Randall. I have re-mapped critical areas, comprising approximately 30% of the total area of this map. Further acknowledgement of Peter Randall's contribution is made in the appropriate chapter. The large scale of this mapping was necessary to obtain the detail for the stratigraphy, local palaetopography, and some structural problems in this well-exposed area of the Mic Mac Lake Group; the results are incorporated into Plate 1.

The boundaries of the inland area, especially the northern and southern

boundaries, were greatly influenced by the extent of the burnt area (Fig. 1.4).

The eastern and western boundaries of the map area are delineated by geological boundaries, beyond which it was not necessary to map for the purposes of this project, except in the southwest of the map area. The western boundary is a sharply defined tectonic contact between two groups in the Fleur de Lys Supergroup, and is described in Chapter 3. The eastern boundary is the unconformity of the Mic Mac Lake Group on the Burlington Granodiorite.

The well-exposed coastal section, and adjoining inland area of the Baie Verte Lineament was all originally part of J. T. Bursnall's map area, but in 1970 he decided that he would not have time to include the Ming's Bight Peninsula in his project. Thus, in 1971 this author spent a month mapping the coast in this area jointly with J. M. Bird and J. F. Dewey. With their agreement, some of the results from this mapping are presented here in a separate chapter. The mapping is displayed on Plate 5 and is incorporated in the summary map (Plate 6). Selected cross sections of the Baie Verte Lineament together with stratigraphic columns of the Baie Verte Group are presented as Plate 7.

(v)   Mapping method and construction of base maps

The outcrop maps are essentially partially-controlled air-photomosaics; the scale varies over these maps, and they must not be used for accurate measurements of distances or angles, especially large distances, and angles over large distances. However, the important point is that the outcrop maps are, in this compiler's opinion, sufficiently accurate for geological purposes, and specifically for the local and overall relative positions of outcrops. Anyone taking the outcrop maps into the field should be able to find a specific outcrop quickly and without difficulty; this would not have been the case if the enlarged topographic maps had been used as a base. Distances between reference points and particular outcrops that are quoted in the text

were measured directly from the outcrop maps. They are therefore probably not the true distance

on the ground, especially for larger distances. They are intended to define an outcrop location

accurately on the map, and the outcrop can then be found in the field from more local features.

CHAPTER 2. REGIONAL SETTING

(i) Summary of Newfoundland geology

Rocks forming the Burlington Peninsula are part of the western side of the 'central mobile belt' of the Newfoundland Appalachian orogenic belt (Williams 1964). This central mobile belt (Fig. 2.1) is defined by the presence of what, for want of a better term, might still be called an 'eugeosynclinal assemblage'; that is thick sequences of immature, rapidly deposited clastic sediments and volcanics, intruded by much granite, with belts of regionally metamorphosed rock, and showing intense and complex deformation. These characteristics can be generalised to the previous existence of a zone of vertical and lateral crustal instability and periodic high heat flow. This central mobile belt contrasts with the little-deformed to undeformed stable shallow water marine sediment accumulations, mainly of Cambrian and Ordovician age, that comprise the cover of the western and eastern platforms of the Newfoundland Appalachians. These two platforms are adjoined by regionally metamorphosed clastic wedges marginal to the central mobile belt (Fig. 2.1) (Williams 1964). The psammites and semi-pelites of the Fleur de Lys Supergroup form a part of the western clastic wedge. The basement of the western platform is composed mainly of gneisses and granites of Grenville age, deformed and metamorphosed about 1000 m.y. ago. It is overlain unconformably by arkoses and orthoquartzites derived from the craton. Minor amounts of basalts are preserved in places at the base, and the feeder dyke swarm for the basalts cuts the Grenville rocks, and parallels the margin of the mobile belt. A carbonate bank accumulation overlies the clastic sediments. The sediments on the western platform are a wedge-shaped accumulation, thickening toward the mobile belt (Williams 1969), and thus form a miogeoclinal wedge (Dietz and Holden 1967). There is slight evidence that the

Fig 2.1. Sketch map of the major elements of Newfoundland geology.
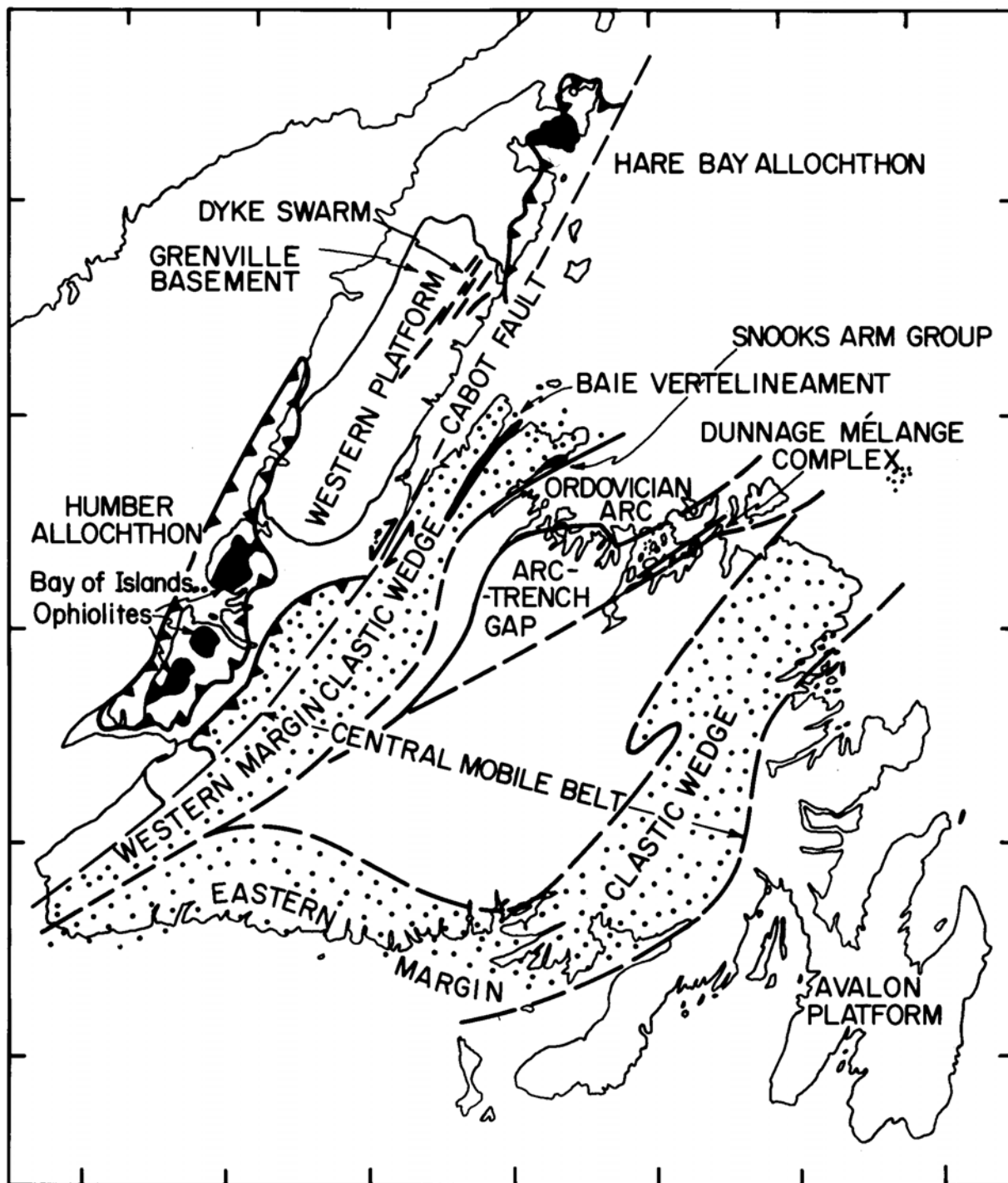
sequence of facies from clastic rocks through mixed clastic and carbonate to carbonate is diachronous, becoming slightly younger to the west (Bird and Dewey, 1970).

Two large composite klippen of the off-shelf, deeper water equivalents of the platform sediments, consisting of carbonate bank edge breccias, more distal carbonate turbidites, shales, and quartz-rich turbidites, were emplaced from the east onto the western platform in middle Ordovician time (Rodgers and Neale, 1963, Stevens 1970). They were overridden by large klippen of oceanic crust (ophiolite complexes) (Stevens 1970), still wholly structurally intact in the Bay of Islands area (Williams, Malpas, and Comeau, 1972). The arrival of these allochthonous slices on the platform was heralded by a westward retreat of the carbonate edge, and the klippen moved into a temporary basin (exogeosyncline) accumulating black shale, and turbidites derived from the direction of the advancing allochthon. Minor calcareous sandy shallow marine sediments of Upper Ordovician and Lower Devonian age overlie the allochthons unconformably.

It has been suggested that the Baie Verte Lineament may be the 'root zone', or telescoped line of derivation for the allochthonous ophiolite complexes on the western platform (Neale and Kennedy, 1967), or that it may be a downfaulted slice of a once far more extensive allochthon derived from farther east (Church and Stevens, 1971). However, an alternative 'root zone' for the allochthonous ophiolites on the western platform has been suggested by Lock (1972) and Dewey and Bird (1971), to be to the west of the Burlington Peninsula, along White Bay, and south through a thrust complex east of the Humber allochthon (Fig. 2.1). This thrust complex brings regionally metamorphosed rocks greatly resembling the Fleur de Lys Supergroup over autochthonous sediments of the western platform. A sliver

of ultramafic rock occurs on one of the thrusts (Williams, 1967). Palinspastic facies

reconstructions, seismic refraction data, and plate tectonic models favour this zone of telescoping

as the site of derivation of the sedimentary and ophiolite allochthons (Dewey and Bird, 1971), in

preference to derivation from the Baie Verte Lineament, or further east.

The eastern, or Avalon platform consists of little deformed, mainly clastic shallow marine rocks

of Cambrian and early Ordovician age, that unconformably overlie a thick sequence of gently

folded late PreCambrian (c. 600 m.y.) sediments and volcanics. These latter rocks form the bulk

of the bedrock on the Avalon peninsula, and an underlying basement is not known. The basement

is assumed to be continental, partly because of the gently- deformed nature of the rocks, but this

need not be wholly the case. The relationship of these rocks to the metamorphosed marginal

clastic wedge on the eastern side of the central mobile belt, which includes some gneissic base-

ment rocks (Kennedy and McGonigal, 1972), is not known. In contrast, the psammites and semi-

pelites of the Fleur de Lys Supergroup on the western side of the mobile belt are thought to be

the very much thicker equivalents, in gross terms, of the arenaceous early Cambrian, and perhaps

late PreCambrian sediments at the base of the succession on the western platform. Rare and thin

carbonate-bearing rocks are intercalated with psammites, semipelites and quartz-bearing pebbly

psammites near the inferred top of the preserved Fleur de Lys succession in the western

Burlington Peninsula (M. J. de Wit, in preparation). They are probably the equivalent of the

transition between the clastic and carbonate sediments on the western platform, of about middle

Cambrian age. They also greatly resemble parts of the unmetamorphosed Summerside Formation

in the allochthonous sediments on the western platform (Stevens, 1970).

In the western part of the Newfoundland Appalachians, there is evidence for three periods of major compressive deformation. First the polyphase deformation of the Fleur de Lys Supergroup, accompanied by regional metamorphism, was post middle-Cambrian and pre-Arenig, because debris derived from the terrain is found in the Snooks Arm Group (Dewey and Bird, 1971). Secondly, the expulsion of the klippen onto the western platform is well dated by sediment provenance and an unconformity on the allochthonous rocks as middle Ordovician (Llanvirn-Caradoc) (Stevens, 1970, Williams, 1971). It is the equivalent of the Taconic event further south, and has been termed the Humberian in Newfoundland (Bird and Dewey, 1970). Thirdly, the folding deformation that affects most of the younger rocks in the central mobile belt is the Acadian event. This is dated in Newfoundland as between lower Devonian and lower Carboniferous, and is probably middle Devonian (Rodgers, 1970). The graben formation, transcurrent faulting, and gentle folding of the Carboniferous rocks in the White Bay-St. Georges Bay belt are best considered separately from the main development of the Northern Appalachians, but the effect of the transcurrent faulting on the relative positions of the earlier rocks cannot be ignored.

Except for the Acadian event, there is no obvious correlation between events and rock groups of the eastern part of the central mobile belt and the eastern platform, and their western counterparts. The eastern boundary of the western part of the central mobile belt can be conveniently placed at the Dunnage Melange (Fig. 2.1). This is thought to represent a deposit formed in the back wall of an oceanic trench, in this case with subduction leading under to the northwest (Kay, in press, Bird and Dewey, 1970). The age of formation of the Dunnage complex is not well controlled, as it is isolated in a separate fault block. It is probably Ordovician, but may be

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Table 2. Summary of the main pre-Carboniferous geological events in Newfoundland.

Table 2. Summary of main pre-Carboniferous geological events in Newfoundland.

modified after Bird and Dewey (1970)

AVALON PLATFORM

- Fe-oolitic sst, shale
- minor basalt stuff
- sst., shale, pink lst.
- quartzite
- FOLDING
- shallow clastics
- shales
- turbidites
- sub-alkalic
- basalt-rhyolite
- gneisses age?

CENTRAL MOBILE BELT (thick sequences)

'CLASTIC WEDGE'

ACADIAN DEFORMATION

- bimodal basalt-rhyolite and terrestrial-shallow marine sediments
- shallow marine sediments
- flysh
- slide conglom.
- vclastics
- andesitic, silicic arc volcanics & intrusives
- mafic volcanics/vclastics
- OPHIOLITE GENERATION
- POLYPHASE DEFORMATION
- METAMORPHISM
- silicic arc vulcanism
- mafic
- carbonate - carbonate turbidites - shales & shales
- basalts, clastics
- gneisses - Grenville?
- SPREADING FORMING OCEANIC LITHOSPHERE
- DUNNAGE MÉLANGE COMPLEX
- flysh
- POLYPHASE DEFORMATION METAMORPHISM & GRANITES
- gneisses age?

WESTERN PLATFORM (thin sequence thickening to E)

'CLASTIC WEDGE'

- calc. sst. ?
- calc. sst.
- KLIPPEN EMPLACED
- flysh
- black shale
- Carbonate bank
- carbonate breccia
- clastics
- gradation
- arkoses, quartzites
- basalts, clastics - RIFTING
- Grenville gneisses

m.y.		
350	D	
400	S	
450	O	
500		
550	€	
600		
	pC	

older or younger. Middle Cambrian trilobites are found in one of the large blocks within the melange (Kay and Eldredge, 1968). Rocks in the central mobile belt to the northwest of the Dunnage Complex and southeast of the Burlington Peninsula are mostly volcanic, plutonic and volcaniclastic rocks, with sediments derived from their erosion. Their age ranges from (?)Cambrian to lower Devonian, and their relationships are complex and in most cases still rather poorly known. There is no indication of a sialic basement in this area; indeed all indications suggest that oceanic lithosphere of various ages originally underlay the whole zone (Dewey and Bird, 1971). These rocks will be described briefly, where appropriate, in the last chapter. These events are summarised in Table 2.

(ii) Geotectonic model.

The brief summary of the regional geology set out above is intended to be just an introductory outline. If the reader wishes a fuller summary of the geology of the Island of Newfoundland, Williams (1969) and Rodgers (1970) give data summaries, but many critical field relationships have been established since they were written. Many of these are described in Williams and others (1972) and Dewey and Bird (1971).

The fact that the British Caledonides are the disrupted continuation of the Newfoundland Appalachians is generally accepted, given the evidence for plate tectonics (Morgan, 1968), and the spreading history of the Atlantic (Pitman and Talwani, 1972). There is no longer any point to using the similarities on either side to contribute to this argument in view of the far more direct and convincing evidence from the ocean. Rather, it is more productive to use the similarities and differences to help document the evolution of the Appalachian-Caledonian orogenic belt. The overall correlation between gross tectonic elements, stratigraphic assemblages, and timing of events between the Newfoundland Appalachians and the British Caledonides

is striking (Dewey and Kay, 1968; Dewey, 1969). However, it is probably at best misleading, and at worst futile, to attempt to draw very detailed correlations, especially stratigraphical ones, across the 750 km of present continental shelf that separates the two coastlines on a well known reconstruction (Bullard, Everett and Smith, 1965), considering the changes that occur over a similar distance southward from Newfoundland along the Appalachians, or northward from Scotland into the Scandinavian Caledonides.

The geological data from Newfoundland and the Appalachians and Caledonides in general, are now interpreted as a fragmentary record of the opening and closure of an ocean (Wilson 1966; Bird and Dewey, 1970; Dewey and Bird, 1971). At present the maximum former width of the Northern Appalachian-Caledonian ocean is uncertain. Lithologic palaeolatitude indicators, (glaciation in N. Africa, carbonate reefs in N. America), and scarce reliable palaeomagnetic data show that an ocean 6000 to 9000 km wide existed between North Africa and eastern North America in the upper Ordovician (Fairbridge, 1971, McElhinny and Opdyke, 1972). However, the position of the Avalon Platform is uncertain, and it is not known whether it was ever detached on its own plate, or attached to SE Britain and the Scandinavian-Russian platform, or attached to Africa-South America (McKerrow and Ziegler, 1972). Palaeomagnetic evidence for the Lower Palaeozoic is scarce and the data points of widely scattered ages, and therefore reconstructions are at present liable to large errors.

From such evidence as is known, it appears that spreading may have occurred in the New York-Newfoundland-British sector of the Appalachian-Caledonian ocean during most of the Cambrian (c. 70-100 m.y.) without any subduction on either margin (Dewey, 1969). At a modest rate of spreading (say 1 cm/yr half rate), this would lead to an ocean 2000 km wide at the end of this time. Perhaps this can be regarded as a minimum figure for the

maximum width of this sector of the Appalachian-Caledonian ocean, as it coincides with a palaeomagnetic reconstruction estimate (Briden, Smith and Drewry, 1972). Thus the geology of the Appalachian-Caledonian orogenic belt is a fragmentary record of events generated by plate interaction during the opening and closure of a substantial and major ocean.

The geology also shows that the western and eastern platforms of Newfoundland were continental margins from the earliest Cambrian to Devonian times; that the marginal clastic wedges of the central mobile belt consist of metamorphosed outer continental margin sediments resting partly on continental-type crust and partly on oceanic crust; and that the remainder of the central mobile belt consists of oceanic crust, island arc rocks, and the sediments derived from erosion of island arcs and cordillera. The Baie Verte Lineament is situated very near the change from continental-type to oceanic crust in the western marginal clastic wedge (Fig. 2.1). This thesis is, therefore, concerned with a part of the history of the western side of the Appalachian-Caledonian ocean as recorded in Newfoundland. This western side has been interpreted in terms of plate tectonic processes by Bird and Dewey (1970), Dewey and Bird (1971). It is not a purpose of this thesis, and neither is it necessary to construct from scratch another synthesis of the geology in terms of plate tectonic processes. However, the data gathered in the map area and set out in this thesis are interpreted in terms of, and related to, processes engendered by plate interactions occurring during the ocean opening and closing cycle. In particular the observations and inferences are pertinent to the evolution of inter-arc basins, and the generation of oceanic crust. The interpretations of the map area, together with their implications for other areas, and some interpretations of other areas, are set out in the final chapter.

CHAPTER 3. ROCK GROUPS BORDERING THE BAIE VERTE LINEAMENT

A. Fleur de Lys Supergroup (Western Division).

(1) Introduction

These regionally metamorphosed rocks are on the western side of the map area, and continue outside it along the full length of the Baie Verte Lineament and to the west coast of the Burlington Peninsula on White Bay (Fig. 1.3). The rocks have been complexly and severely deformed in a 'polyphase' deformation sequence consisting of two successive penetrative schistosities followed by a strain-slip type schistosity, all with accompanying folds. They are also more or less thoroughly reconstituted by accompanying regional metamorphism reaching biotite or garnet grade in most of the map area. The overall amount of exposure of these rocks in the map area is not sufficient to enable the early large scale fold structures to be defined with confidence, although it is sufficient to show the broad distribution of the lithological units and define the later large fold structures. As mentioned in the first chapter, these rocks were mapped to provide evidence on their relationship to the Baie Verte Group, and thus on the evolution of the Baie Verte Lineament. If it appeared that the information potentially available from investigation of a particular aspect of these rocks would not assist materially in the solution of problems relating to the Baie Verte Lineament, then the matter was not pursued in great detail. Thus, some problems, in particular the metamorphic recrystallisation sequence and its relations to the structural sequence, have not been studied in detail. The detailed outcrop mapping of these rocks is shown on Plates 1 and 2, and the geology is summarised on Plate 6. The descriptive parts, of this thesis are intended to be read with the outcrop maps to hand, and repeated references

to them are not made.

The units in the mapped area fall into two sequences, each containing two major units. However, no conclusive facing evidence has been detected in these rocks in the map area. In view of the potential tectonic disruption and complexity of folding, let alone original facies changes, that can occur in this kind of terrain, it is not wholly certain how far the relationships of the units mapped represent a stratigraphy. Therefore, the units are described in what is thought to be a likely order of relative age, mainly on the basis of evidence from other parts of the Burlington Peninsula. This evidence derives mainly from the work of M. J. deWit (in preparation), J. T. Bursnall (in preparation), and Kennedy (1971), and is quoted where appropriate. For same reason, all but one of the lithologic units are given an informal nomenclature, so that they can eventually be assigned to stratigraphy erected on better evidence elsewhere, and in the meantime a proliferation of names is avoided.

(ii) Lithologic units

(a) Western sequence

(1) Psammite - amphibolite unit (Psammite with amphibolite boudins)

This unit consists mainly of homogeneous white-weathering, grey psammitic schist, with mica-rich folia spaced at large and irregular intervals. Uncommon bands of semipelitic schist up to about 10 cm thick also occur. The psammite consists of a granular annealed aggregate of quartz, plagioclase, muscovite and biotite, with minor epidote, apatite and sphene, and rare allanite. The micas are usually strongly aligned, but are coarse-grained and relatively strain free. The green amphibolite, which occurs as plump 'semiductile'-type boudins a few centimetres to at least thirty metres across,

is an abundant and distinctive constituent of this unit; almost all outcrops contain at least one boudin. The proportion of amphibolite is tentatively estimated to be 5% to 10% of the unit, although locally it can be much greater. Most of the amphibolite is essentially non-foliated, being a fine to medium grained granular mass of amphibole and subordinate plagioclase, with biotite, quartz, epidote, apatite, sphene and minor magnetite. It may contain numerous plagioclase porphyroblasts, or (rarely) porphyroblastic amphibole, or no porphyroblasts. In the south of the map area, west of Mic Mac Lake, some of the amphibolite is richly garnetiferous, and displays coarse poikiloblastic plagioclase and hornblende. No genuine eclogitic rocks, have been found, even with only relict pyroxene, except as float derived from southwest of the map area. An area of typical outcrops of this unit occurs just over a kilometre WNW of Kidney Pond, and another just east of Marty's Pond. Minor folds of early phases are not seen in this unit, except in the westernmost outcrop marked on the Bear Cove Road (Plate 2), where the lithology is atypical, and best described as a quartzo-feldspathic gneiss with amphibolite boudins. There are three possible origins for the psammite-amphibolite unit; (1) variably reconstituted gneisses forming the original basement to part of the Fleur de Lys metasediments; (2) Psammitic and pelitic sediments of pre-Fleur de Lys age; that is sediments metamorphosed and incorporated in the basement gneisses prior to the Fleur de Lys sedimentation; (3) sediments of lower part of the Fleur de Lys Supergroup, of which the most similar unit is the Southern Arm Formation of the Seal Cove Group (deWit, thesis 1972; in preparation). It is quite likely that all three possibilities are present, there is no evidence to separate them in this map area. The outcrop of quartzo-feldspathic gneiss on the Bear Cove Road mentioned above is assigned to the remobilised basement gneisses by M. J. deWit (pers. comm.). Very little of the unit mapped has sufficient semipelitic and pelitic material in

well-defined bands to suggest an obvious bedding, and as such it does not resemble the more obviously sedimentary parts of the Southern Arm Formation shown to the author by M. J. deWit. The unit in the map area is probably too thoroughly reconstituted to have retained other sedimentary structures and none were seen. Whatever the nature of the psammitic component, the amphibolite is almost certainly derived from basaltic dykes, sills and perhaps flows. M. J. deWit (pers. comm.) has found evidence of intrusive relationships in his mapping of these bodies elsewhere in the peninsula, and the boudins show that the original bodies were relatively brittle and tabular in shape. DeWit finds that most is probably of the same age as the deposition of the psammitic sediments of the lower Fleur de Lys succession in which they are found in abundance, because they do not occur above a certain point in the succession in his map area. In my map area, the amphibolite is too completely recrystallised and strongly-deformed to retain any relict textures or clear intrusive /extrusive relationships in the small outcrops available.

(2) Leucocratic quartzo-feldspathic gneiss

This occurs in two small separate areas at the western extremity of the map area (Plate 1). The southern area, west of the north end of Mic Mac Lake, consists of very homogeneous medium-grained granular white-weathering grey rock, with very rare, thin and diffuse micaceous laminae. A thin section shows an even, fine to medium-grained, granular aggregate of quartz and plagioclase with some microcline (perhaps metasomatic) and minor biotite, muscovite, apatite and epidote. As the more micaceous laminae are spaced on the order of 50 cm apart, the outcrops have a coarse slabby appearance, and this slabbing is almost all that can be seen of the main regional schistosity. The area of this unit that was mapped had been burnt, and the

outcrops are clean, fairly numerous, and some quite large. There is a conspicuous absence of amphibolite boudins, and the rock exposed is extraordinarily homogeneous. The contact with the psammite-amphibolite unit to the southeast is not exposed, and there are only two outcrops of the psammite near it, but they define a segment quite closely. The outcrops of gneiss nearest to the contact have a strong, steeply-plunging, lineation formed by extreme elongation of the mineral grains and it appears to be near a pure L-fabric. This lineation dies out away from the contact, and thus implies a zone of strong localised deformation along the contact, perhaps caused by a significant ductility contrast between the two units during deformation. The origin of this area of this unit is problematic, as it cannot be shown from the outcrops mapped that the pre-Fleur de Lys tectonic fabrics occur in it. While it may be part of the basement gneiss complex, it could alternatively perhaps be a thoroughly reconstituted fine-grained granite, intruded during early Fleur de Lys sedimentation. The former is perhaps favoured by the fact that the rock shows no trace of relict granitic textures, even well away from the contact, while the absence of amphibolite boudins perhaps favours the latter.

The second small area of this unit that was encountered is at the southeast corner of Gull Pond (Plate 1). Here the rock is similar to that described above, but has a slightly greater content of micas, and has better-defined and more common micaceous laminae, spaced at around 15 cm interals, and the main foliation is more clearly defined as a result. This rock is thus more like a common gneiss, although it is still relatively fine-grained. It is along strike from some intensely-banded basement gneisses mapped by M. J. de Wit 4 km to the north on the Bear Cove Road. Only a few small outcrops were examined in the thickly-forested terrain, but no amphibolite was seen in them.

(3) Pegmatite

Coarse to extremely coarse-grained plagioclase-microcline-quartz-muscovite-biotite-epidote-garnet pegmatite occurring as concordant veins is common in the psammite-amphibolite unit from Marty's Pond northnortheastward for 4 km. These veins range from a centimetre upward to at least 10 metres wide, always show pinch-and-swell structure, and have not been seen other than sensibly parallel to the main schistosity. While the large crystals of feldspar and quartz seem to be anhedral, they are never severely crushed, but the margins of the veins often have a crude foliation parallel to the main (second) schistosity in the host schist. This suggests that the pegmatites formed during, or before, the development of the second (main) Fleur de Lys schistosity, but their relationship to minor third or second phase folds has not been seen, as these are not well-developed in the area where the pegmatities occur. Minor development of similar veins not more than a few centimetres wide occurs in the quartzo-feldspathic gneiss NNE of Marty's Pond, and a few tiny lensoid sweats a centimetre or two wide were seen in the semipelitic schist west of Red Cliff Pond but these may not belong to the same event. No pegmatite occurs in the psammite-amphibolite WNW of Kidney Pond or in that mapped on the Bear Cove Road. However, a zone of pegmatite veins occurs in rocks similar to the psammite-amphibolite in M. J. deWit's map area, on the Bear Cove Road about 5 km west of Red Cliff Pond. A discrete zone containing pegmatite may be present between this occurrence and the one by Marty's Pond. The only intrusion in the peninsula known to be of similar age to the pegmatites is the Dunamagon granite, situated 60 km NE of Marty's Pond near the north coast of the peninsula (Fig. 1.3), which was intruded synchronously with the third deformation in the Fleur de Lys schists adjacent to it and has associated zones of abundant pegmatite very

similar to those in the map area (Baird 1951). Otherwise no suggestions as to the origin of the pegmatites can be made, and no evidence has been seen to link them with the Celebes Pond and Wild Cove Pond granites (q.v.), which were intruded after Fleur de Lys deformation, or with the Burlington Granodiorite, which appears from published descriptions to lack significant associated pegmatites (Baird 1951; Neale and Nash 1963). A small area of the Wild Cove Pond granite, marked on published maps northeast of Marty's Pond (Neale and others, 1960; Neale and Nash, 1963), is due to confusion of whole outcrops of the pegmatite with the granite, from which the pegmatite is distinguished by the lack of large microcline phenocrysts, and the presence of muscovite and garnet.

(b) Rattling Brook Group (metasediments)

The Rattling Brook Group was erected by Watson (1947) to cover semi- pelitic and mafic schists occurring to the west and northwest of the town of Baie Verte. Subsequent work by Neale, summarised in Neale and Nash (1963), confirmed that these rocks are continuous southward into my map area. Further subdivisions of the Rattling Brook Group made by J. T. Bursnall and M. J. deWit, and possible correlations, are discussed in a later section.

(1) Semipelitic schist unit

In most of this area, this is a rather homogeneous and uniform rock, being a typical white-weathering, grey quartz-plagioclase-muscovite-(biotite) schist. It is always well-foliated with typical metamorphic segregation and transposition banding of thin folia of micas alternating with polygonised quartz-plagioclase folia. Biotite is generally brown or orange and either subordinate to muscovite or absent. Apatite and magnetite are usually present as minor constituents and tourmaline, epidote, sphene and allanite may also occur. Garnet porphyroblasts may be abundant, although they rarely exceed a centimetre in diameter, and are usually much smaller. The presence or

absence of garnets in this semipelitic lithology are used to map a garnet isograd. Magnetite is found abundantly speckled though the schist in some outcrops, but is usually a minor constituent. Plagioclase may be present as porphyroblasts up to a centimetre diameter, but is not very common in this form. Minor chlorite, retrograde after garnet and biotite, occurs in some thin sections. A distinctive feature of the rock unit is the almost ubiquitous presence of lensoid quartz veins concordant with the composite schistosity. They commonly show pinch-and-swell structure, and sometimes occur as disrupted isoclinal fold hinges with the main (second) schistosity axial planar. The veins are generally a centimetre or less wide, though a few larger ones have been found. Few traces of bedding remain, except where beds of strongly contrasting lithologies occur. Possible bedding in more homogeneous semipelitic schist can be seen 2 km north of Kidney Pond [1]*, where diffuse, somewhat more psammitic, layers up to 1 cm thick alternate with diffuse, relatively more pelitic, layers of similar thickness. If the greater part of the unit originally had this thin and poorly-defined bedding, it is perhaps clearer why the deformation and metamorphism have obliterated it elsewhere. In outcrops in the area west and northwest of Slink Pond (e.g. [2]), well-defined psammitic beds up to 20 cm thick alternate with fairly sharply-defined pelitic to semipelitic beds up to a few centimetres thick, although the latter are often strongly disrupted by the deformation. In this area, outcrops between about 100 metres west of the Roadside slide zone and 300 metres east of the western belt of Birchy Schist Formation are mostly of this kind of lithology and its western boundary with homogeneous semipelitic schist can be mapped as far south as the first cross-fault. Lack of outcrop prevents the eastern boundary, and any extension of either boundary south of

*Numbers in square brackets refer to specific outcrops marked on Plates 1 and 2 in the same way.

the cross-fault, from being mapped with any certainty. This type of lithology is not seen in the wide belt of the semipelitic schist that extends north from Kidney Pond, although thin (cm) beds of psammite are seen in a few outcrops. It may be possible to map subtle variations in quartz content in the whole semipelitic schist unit, but this has not been attempted here. Outcrops noted with relatively, abundant garnet or magnetite did not make any mappable sub-units. In contrast, a facies of the semipelitic schist containing a noticeable but minor graphitic component is always observed, when it occurs, in outcrops within about 200 metres of the contact with the Birchy Schist Formation. However, this interval does not always contain a graphitic component, and this is probably due to sedimentary rather than tectonic causes. Some bands of mafic schist, usually not more than a few centimetres thick, and identical to that which forms the bulk of the Birchy Schist Formation, are also often present within about 200 metres of the contact with that Formation, and more rarely elsewhere. However, they are also not necessarily found in any particular traverse towards the contact.

The magnetite content of this semipelitic unit, and also of some parts of the Birchy Schist Formation, is large relative to that of the western psammite-amphibolite and other units. This allows a fairly accurate location of the boundary between these two sequences of rocks on the aeromagnetic maps, from Kidney Pond northwards in this map area, and elsewhere in the north and west of the Burlington Peninsula.

Typical outcrops of this unit occur 0.5 km NNE of the southeastern corner of Tom's Pond. This terrain has suffered severe and complex deformation, and a determination of a thickness for this unit is difficult because of possible undetected fold repetition and tectonic sliding. A very tentative estimate of the minimum present thickness is 800 metres (the width of the section along

the Bear Cove Road and Middle Arm Brook). The real thickness may be much greater, as the only exposed structurally lower contact of the unit in this area seems to be a tectonic junction with the western sequence. Much greater cross-strike widths of this kind of rock occur to the north outside this map area. The upper contact of this unit and possible correlations with other areas are discussed in sections below.

<center>(2) Birchy Schist Formation</center>

This formation was erected by Fuller (1941) in the Fleur de Lys village area and has been retained by later workers (Kennedy 1971, Bursnall, in prep.). There is an essentially continuous belt of this formation from the type area into this map area (Neale 1958, 1959b; Neale and Nash, 1963). It is composed mainly of varieties of mafic schist.

In the map area, the mafic schist is medium to dark green, medium-grained and well-foliated, with some yellow epidosite, and some white calc-silicate segregations. The main constituents are actinolite and plagioclase, or hornblende and plagioclase at higher metamorphic grade. Minor constituents are epidote, clinozoisite, magnetite, sphene, apatite, quartz, calcite and rare chlorite. However, at lower metamorphic grade, mainly in the small area mapped north of Red Cliff Pond, chlorite becomes a major constituent, and may be the only mafic mineral present. Around Kidney Pond occasional bands contain green biotite instead of amphibole. The mafic schist contains little magnetite, except in chlorite grade rocks in the northern part of the map area along the Baie Verte Road, where small magnetite octahedra are abundant in some outcrops.

There is a strong foliation on a millimetre scale in the outcrops due to the schistosity and uncommon segregations parallel to it, and there is also a banding on the scale of a few centimetres to a metre or so, which is bedding. This banding results from contrasts between bands with small plagioclase

porphyroblasts, those with none, bands containing a large quantity of epidote, etc. In an outcrop 0.4 km NW of Kidney Pond [3], these bands are concordant with a bed that is either an agglomerate or volcanic conglomerate, containing mafic clasts mostly from 1 to 30 cm across. Only one other occurrence of agglomerate has been found in this unit in the map area, 0.15 km NW of the Burlington Road junction [4].

An outcrop 0.3 km NW of the junction of the Camp 164 Road and the Baie Verte Road [5], contains a pre-kinematic crosscutting dyke about 10 cm wide of fine-grained amphibolite in semipelitic schist. The dyke is not foliated except for its fine-grained margins. In thin section, some ragged pseudomorphs of plagioclase phenocrysts are seen. This is probably a feeder dyke associated with the Birchy Schist Formation vulcanism, but as it occurs within interbanded semipelitic and mafic schist it cannot prove the way up of the succession. No corresponding massive amphibolites that might have been lavas or sills have been found in outcrops with other mafic schist lithologies. However, a whole outcrop of massive fine-grained little-foliated amphibolite very similar to the dyke occurs 1.9 km SW of the inlet to Kidney Pond. This may have been a lava or sill, but the outcrop is isolated and its relationship to other lithologies is not known. Similar massive amphibolite occurs within the agglomeratic horizon 0.4 km NW of Kidney Pond [3], and is here either a few large clasts, or the boudinaged remnants of a dyke about a metre wide.

At three localities, metachert bands are found. The best of these is 1.6 km WSW of the southern end of Slink Pond [6]. The metachert is a finely- banded purplish, pale-weathering rock in beds 1 to 10 cm thick. A fine- grained quartz mosaic contains abundant small zoned porphyroblasts of garnet (? spessartitic) and magnetite, some epidote and minor actinolite. Also found in the Formation, well-displayed at the metachert locality

mentioned, are beds of psammite and quartzitic psammite, 1 to 30 cm thick, with thinner pelitic

and semipelitic interbeds. The psammite beds are exceptionally up to 2 metres thick, but are a

minor component of the Formation and are almost wholly-confined to the western belt of

outcrop. However, around Tom's Pond, NE of Jack's Pond, and in two outcrops on Red Cliff

Pond, a mappable member of psammite with pelitic and semipelitic interbeds occurs, which is

structurally above much of the mafic schist of the Birchy Schist Formation in this area. It is

included within the Formation, because a few thin bands of mafic schist occur within the

member and more mafic schist appears to occur conformably above it in the area west of Tom's

Pond. The pelitic interbeds in this unit can be very mica-rich and may contain abundant

magnetite. No sedimentary structures apart from bedding were seen in this unit. Psammite beds

are usually between 5 and 20 cm thick, and the more pelitic interbeds between 1 and 10 cm thick.

The thickness of the Birchy Schist Formation in this area is uncertain, as the structural top is not

seen, and there are complications with tectonic sliding, both detected and possibly undetected

and also perhaps misinterpreted. The maximum thickness preserved in the map area is about 500

metres, but in most of the area the preserved thickness is 300 metres or less. The psammite

member mapped is up to 100 metres thick; this is included in the estimates for the whole

Formation. Typical outcrops of both the psammite member and the mafic schists of the

Formation may be found from the southwest shore of Tom's Pond for 0.5 km to the southwest.

The structurally lower contact of the Birchy Schist Formation with the semipelitic schist

unit is either an abrupt lithologic change across a sharp contact, or is a zone usually about 150

metres wide in which the two lithologies are interbanded on all scales from a centimetre to 20

metres thick. The interbanded type is mostly confined to the westernmost contact of the two

units, although this western contact is sharp south of the Celebes Pond

Granite. It also occurs in the noses of two second-phase folds just north of Kidney Pond, and one of these zones is a rather irregular shape for the interbanding to be solely due to minor folds. In addition the whole exposed width of the Rattling Brook Group opposite the northern half of Mic Mac Lake is an .area where the two units cannot be separated, and this is not entirely due to the poor outcrop in this area. Although some of the interbanding in all these cases, especially of the thick units, is probably due to isoclinal folding, it seems likely that much is primary, especially where bands a few centimetres thick are concerned. There is no sign of any major tectonic slide along this contact in the case of either the interbanded or sharp of contact. From the distribution of the two units, and the structural interpretation of this distribution, I conclude that there are not two separate contacts represented. In other words there is not more than one unit either of the Birchy Schist or of the semipelitic schist in this area. Therefore, it seems likely that the variable nature of the contact is due to minor facies change.

Northward from the Bear Cove Road, a lithology intermediate between the normal actinolite-chiorite mafic schist and the quartz-plagioclase muscovite semipelitic schist is commonly developed in the interbanded zone. Its variable composition probably indicates original sedimentary mixing of the two lithologies. Also found in the interbanded unit, in outcrops 0.8 km northwest [7] and north [8] of the north end of Marty's Pond, are thin continuous beds of rather pure metaquartzite. These are commonly 2 to 5 cm thick, and are traceable for strike lengths of at least 20 metres. Such beds of quartzite in this intimate association with semipelitic and mafic schists are most likely to have been deposited as turbidites.

3.14

A comparison of the relative state of deformation of the small amphibolite dyke [5], and the agglomeratic horizon [3], with the adjoining mafic schists indicates that all of the mafic schist was medium and fine-grained mafic volcaniclastic rock when deposited. The common occurrence of mafic schist beds only a few centimetres thick confirms this contention. The relative lack of quartz in the mafic schists, and the lack of any associated silicic volcaniclastic rocks perhaps suggest that the mafic volcaniclastics were derived from a basaltic parent.

The associated psammitic, semipelitic and pelitic metasediments were originally dirty sandstones, siltstones and shales. Their association in a relatively thick monotonous succession with mafic volcaniclastics, chert, graphitic siltstone and shale, and probable orthoquartzite turbidites indicates that they are very much more likely to be turbidites and associated shales, rather than shallow water deposits.

(3) Correlation

Stratigraphic columns have been set up for the Rattling Brook Group in the Fleur de Lys village area by Kennedy (1971), and in the adjacent area west of Baie Verte by J. T. Bursnall (pers. comm.). The stratigraphy of the White Bay Group, a broadly correlative sequence on the west coast of the Burlington Peninsula, has been set up by deWit (thesis, in prep.) (Fig. 3.1). While the Birchy Schist Formation in the map area can be clearly correlated from the type section near Coachman's Cove, other correlations are problematic. A thick sequence of semipelitic schists structurally underlies the Birchy Schist in the Coachman's Cove area, and while Kennedy (1971) states that there is a tectonic junction between the two sequences, Bursnall (pers. comm.) considers the contact to be essentially conformable. Kennedy named the semipelitic schists the 'Harbour sequence', and divided it into four formations. The semipelitic schist unit in this map area probably corresponds
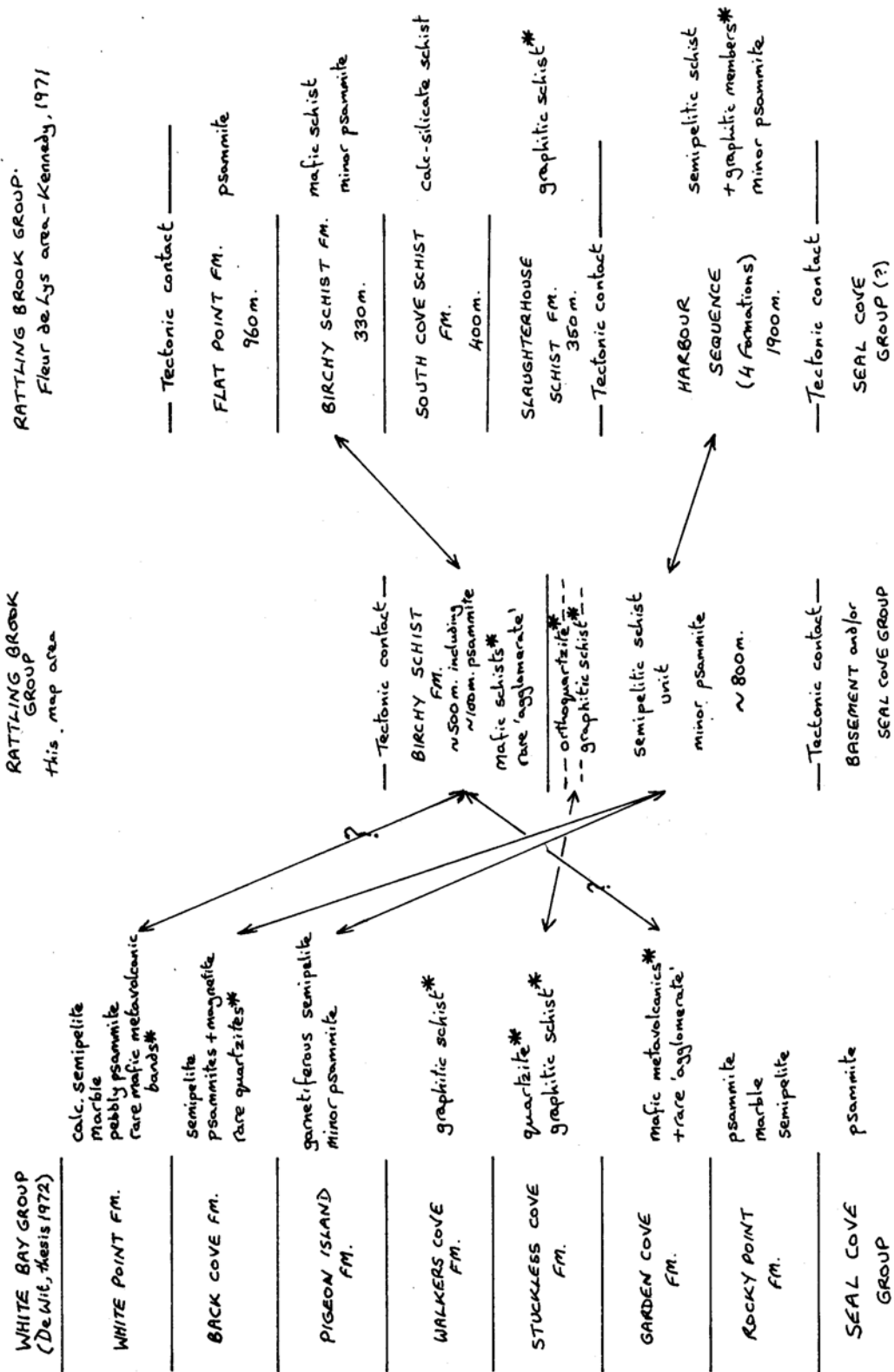
Fig. 3.1. Tectono-stratigraphic columns for possible correlations of the Rattling Brook Group.

Fig. 3.1. Tectono-stratigraphic columns for possible correlations of the Rattling Brook Group.

with some part of the Harbour sequence, and it is clear in my area that there is a conformable

contact between this and the Birchy Schist Formation, and that two formations put below the

Birchy Schist Formation in the Coachman's Cove area by Kennedy (1971) are missing. If the

stratigraphic facing evidence in the Coachman's Cove area is correct, it suggests that the

succession in this map area is stratigraphically upright. The psammite member recognised in this

map area is possibly an equivalent of the bottom of the Flat Point Formation, but is more likely

to be a thicker than normal development of psammites found intercalated in the Birchy Schist

Formation in the Coachman's Cove area (Kennedy, 1971). Correlation with the White Bay

Group is more problematic (Fig. 3.1). In that group, (deWit, thesis) orthoquartzite turbidites

occur mainly in the graphitic schists of the Stuckless Cove Formation, but are also found in the

Back Cove Formation. A prominent graphitic schist unit, the Walkers Cove Formation, is above

the Stuckless Cove Formation. However, the most prominent mafic metavolcanic horizon, the

Garden Cove Formation, is below the graphitic Stuckless Cove Formation. Most of the

semipelitic schists in this succession are, however, above this graphitic horizon. In addition

minor mafic metavolcanic horizons are also found in the White Point Formation near the top of

the succession as preserved. Therefore, although a general similarity exists between some

lithologies in the White Bay Group and the Rattling Brook. Group in this map area, no specific

correlations can be made at this time, other than the probable broad equivalence of the two

Groups.

(c) Meta-granodiorite

Apophyses of a white-weathering, whitish-grey, medium to coarse- grained foliated

granitic rock and finer grained aplitic equivalents intrude the Rattling Brook Group

metasediments in two main areas adjacent to the contact with the Baie Verte Group. These are

from Kidney Pond for 1.4

km NNE and 4 km NNE of Kidney Pond. Two other widely separated minor occurrences are in outcrops 0.25 km SW of the outlet of Slink Pond [10] and 0.8 km NW of the north end of Marty's Pond [7]. The area just north of Kidney Pond consists of a narrow mappable zone adjoining the contact with the Baie Verte Group (Plate 1). Minor occurrences of apophyses, indicated on Plate 1, are commonly found up to 0.4 km west of this main occurrence, and one occurs 0.7 km west of it. These apophyses range from a few centimetres to about a metre wide, and are strongly-foliated with the regional Fleur de Lys schistosity, and are mostly concordant. The more northerly main occurrence contains outcrops with only weakly foliated rock, that shows relict plutonic textures. Aplitic veins are present cutting coarse grained granular rock, that when unfoliated consists of 60% andesine, 30% quartz, some of which is poikilitic, and interstitial large muscovite flakes, with accessory epidote, apatite, sphene and calcite. When the rock becomes foliated, and the original grains augened by a developing muscovite schistosity, the plagioclase is converted to albite, and a proportion of the albite grains contain myriads of aligned fine-grained muscovite inclusions.

Although most occurrences of this lithology contain the main second schistosity, in one locality it can be shown that an aplitic apophysis contains the first schistosity. The intrusion of this granodiorite is therefore synchronous with, or predates, this first regional deformation. This outcrop [9] is just east of the power transmission line almost 1.4 km NNE of the outlet of Kidney Pond.

These apophyses of meta-granodiorite occur along a strike length of 15 km on the very eastern edge of the western division of the Fleur de Lys Supergroup terrain. The eastern division of the Fleur de Lys Supergroup to the east of the Baie Verte Lineament, is replaced in the map area by the

Burlington Granodiorite (Fig. 1.3). I suggest that these apophyses of meta-granodiorite were originally part of the western contact of that body. The nature of the main part of the Burlington Granodiorite is discussed later in this chapter.

(iii) Deformation; minor structures

(a) Introductory comments

The whole of the western division of the Fleur de Lys Supergroup is a regionally metamorphosed terrain and, as is usual in such terrains, the accompanying deformation is intense and has a polyphase nature. The result is a sequence of folds and their attendant schistosities and lineations, where each succeeding phase affects the previous ones to a greater or lesser extent, depending on the local intensity of deformation and local amount of metamorphic recrystallisation during a particular phase. Thus, complex refolded structures and composite fabrics can be present. While the sequence of folds and fabrics seen in any one outcrop may appear, and can be described in terms of, a set of discrete deformational events, this does not necessarily mean that the overall regional deformation was a series of discrete events, or that apparently similar sequences of deformational events from widely separated areas are necessarily correlative, or if correlative, were each geologically simultaneous. The nature of any particular deformational event may change as an area is traversed, but it is perhaps reasonable to assume, at least at first,that any changes in intensity of deformation, orientation of structures, and general style of folds and fabrics in a given lithology, are likely to occur relatively gradually. Thus correlation of deformational event sequences between closely-spaced outcrops (up to 100 metres apart) is usually an easier and more certain process than correlation between outcrops more than a kilometre apart. However, it is not necessarily a simple process, as in a particular region an event may not have occurred, or be weakly repre-

sented, or alternatively have greatly increased in intensity. Also, the evidence for the earlier events may be lacking due to its obliteration in most outcrops by later events. In addition, local narrow zones of intense deformation occur (tectonic slides). However, my approach to this problem is that a discrete structural event sequence adequately describes what can be seen in outcrop, and can with care be correlated from outcrop to outcrop. Although it may be misleading to think of the kinematic development of the large scale regional structure in terms of discrete events, this aspect of the problem is not a concern of this thesis. Rather, this approach enables the data to be satisfactorily condensed, and is in a suitable form for use in the solution of problems relating to the Baie Verte Lineament. The comments above derive in large part from many helpful discussions with J. T. Bursnall and M. J. deWit, for which I am grateful, although the views expressed are my own.

The terminology and abbreviations used to describe the deformational sequence are those in common use, where events and structures are given a subscript numeral, the oldest being 1, then 2, etc.; and the structures in question are abbreviated to a capital letter, where the deformational events are D_1, D_2, D_3, etc.; folds and their axes F_1, etc.; schistosity/cleavage S_1, etc.; linear structures L_1, etc.; and where F_1, S_1, L_1, are the product of the same event, not the earliest event of that type that happens to be seen in a particular outcrop.

In this map area, there are three phases of deformation that are regionally important. The main regional schistosity is the most widely recognisable structure, which together with the subparallel lithologic units is responsible for the dominantly NNE-SSW grain of the terrain occupied by the western division of the Fleur de Lys Supergroup (Neale and Nash, 1963).

(b) First Deformation (D_1)

Evidence for the regional development of this event is less abundant than for the later events. Rare small-scale interference fold structures, where F_1 is refolded by F_2, are found. In all except one of these examples, the F_1 folds are nearly coaxial with F_2 folds and the interference structures are of type 3 of Ramsay (1967). The original orientation of the axial surfaces of the first folds cannot be stated with confidence because of the small number and small scale of the examples seen. On the basis of the inadequate sample, it would appear to have been subparallel to the now steep axial surfaces of the second folds. Folds deformed by nearly coplanar and coaxial folds are merely accentuated and no characteristic interference pattern is generated (Ramsay, 1967). The large-scale lithologic distribution pattern does not contradict the suggestion that this may be the case in this area. It may be objected that the small number of minor F_1 folds seen might be due to sediment-slumping phenomena, not to compressive deformation. None of the minor F_1 folds seen had the character of slump folds and there is other more widespread evidence of the first deformation.

In several widely scattered localities along the full length of the map area, clear evidence is seen in thin sections of the transposed relicts of the primary schistosity (S_1) between mica folia defining the second schistosity (S_2) in the semipelitic schists. This can also be seen megascopically in the cores of minor second folds (F_2), where the first schistosity (S_1) is dominant, and the second schistosity (S_2) is reduced from its normal penetrative nature on the limbs of the folds to a strain-slip type. It might be postulated that the first schistosity in the semipelitic schist is merely a mimetic metamorphic recrystallisation of a shaly parting (J. T. Bursnall, pers. comm.), and this cannot be disproved for the semipelitic schist. However, in the mafic schist of the Birchy Schist Formation, the penetrative primary schistosity (S_1) can be followed round the hinges of almost all

minor second folds. In mafic volcaniclastic rocks there is little chance that much, if any, of this foliation is due to mimetic overgrowth of a 'sedimentary' foliation. Indeed, the schistosity on the limbs of most of the minor second folds in mafic schist is indistinguishable from that around the hinge, suggesting that the first schistosity (S_1) is responsible for most of the tectonic foliation seen in the mafic schist, although the second deformation may have accentuated the schistosity on the limbs of the isoclinal second folds.

Minor first folds are isoclinal and extremely attenuated, which may be due partly to modification during the second deformation. Their original orientation cannot be determined, but they are now nearly coaxial and coplanar with the steep, generally gently southward-plunging second folds. Lineations belonging to the first deformation have not been identified. The first schistosity (S_1), when detected, is parallel to any bedding that can be recognised with the exception of the outcrops on the Baie Verte Road between 3.8 and 6 kilometres north of the Burlington Road junction. Here, the first schistosity consistently dips to the east less steeply than the bedding, indicating an F_1 antiform to the east. The significance of this observation is not known, as the hypothetical fold has been removed by the tectonic boundary with the Baie Verte Lineament, and may have been of relatively small scale. Also, M. J. deWit (pers. comm.) has suggested that the schistosity crosscutting the bedding is not S_1 but a later schistosity; the author disagrees with this interpretation as both schistosities have been seen folded by what seem to be F_3 folds. The original nature of the first schistosity over the whole area has almost certainly been greatly modified by the deformation and both dynamic and static metamorphic recrystallisation subsequent to its formation. Examples of minor first folds folded by second

Fig. 3.2. Minor folds in the Rattling Brook Group.
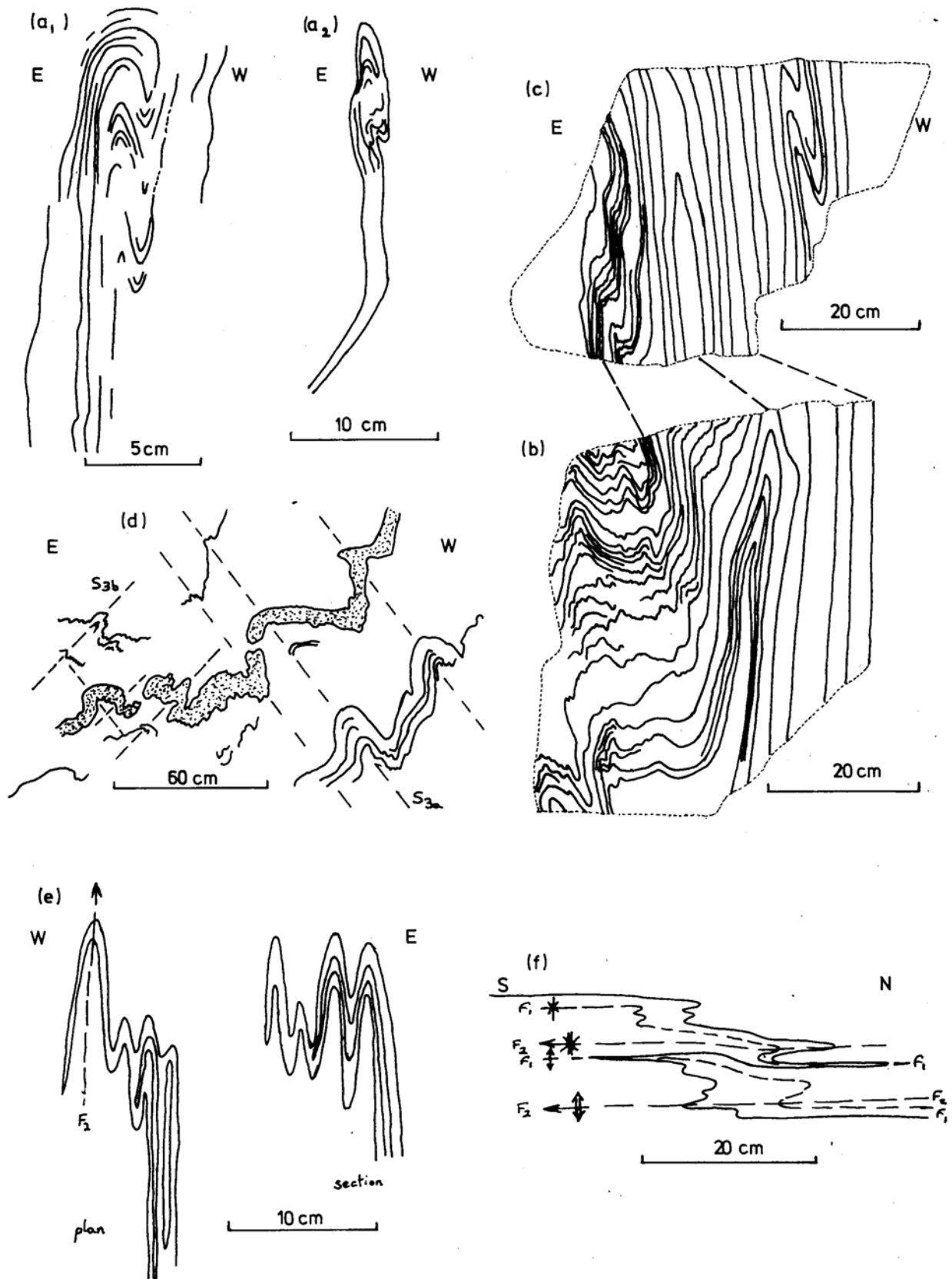
Captions for Fig. 3.2.

(a) Two examples of F_3 refolded by F_2: (a_1) section; (a_2) plan of a separate example; mafic schist, Birchy Schist Fm., Kidney Pond [11].

(b) F_2 isocline folding S_1 schistosity; F_3 fold-pair on left refolds the F_2 fold. mafic schist, Birchy Schist Formation [12]. Sectional view perpendicular to fold axes.

(c) plan view of same part of outcrop [12] as in (b), showing southward plunging F_2 and F_3 minor folds.

(d) conjugate S_3 strain-slip schistosity and elastica-type F_3 folds mafic schist, Birchy Schist Formation, Kidney Pond [11].

(e) plan and section of F_1 minor fold refolded by F_2 minor folds. psammitic schist in semipelitic schist unit, 1 km NW of S. end of Slink Pond.

(f) plan view of type-3 F_1/F_2 interference structure; psammitic schist, semipelitic schist unit, NW of Slink Pond [2].

3.2 a-d are sketches from photographs; e and f are from field sketches.

folds are shown in Fig. 3.2.

(c) Second deformation (D$_2$)

Minor second folds are fairly common and are tight to isoclinal with steep axial surfaces. They often plunge gently southward at 20 to 30 degrees but this is by no means the rule. Mineral lineation (L$_2$) is extremely uncommon, but occurs in a few scattered localities formed by amphibole in the mafic schists and micas in the semipelitic schists and, where present, is coaxial with the local second fold axes. The second schistosity is penetrative in the semipelitic and psammitic schists but apparently less so in the mafic schists. However, in most of the area, the second and first schistosities are so nearly parallel that S$_2$ is usually partly formed by transposition of S$_1$. Where S$_1$ is the dominant fabric, as in much of the mafic schist, it is essentially parallel to the S$_2$ orientation due to the isoclinal second folds. Thus the 'main schistosity' seen in this area of the western Fleur de Lys terrain is a composite fabric made of varying contributions from both S$_1$ and S$_2$. The second schistosity is in practical terms parallel to bedding over the whole area. Examples of second folds are shown in Figure 3.2. Connected pairs of minor second folds giving major fold vergence data are not commonly seen, and this technique was not used for interpretation of the major structure.

(d) Third Deformation (D$_3$)

In the area bounded to the north and south by east-west lines through Red Cliff Pond and Kidney Pond, minor third folds are common. These are open to close folds of similar type with rather angular hinge zones and straight limbs. Their axial surfaces are subvertical, and axes generally plunge southward at 15 to 30 degrees, essentially coaxial with second folds in the area around Kidney Pond. Their axial surfaces and the third strain-slip

schistosity are almost always oriented so that they dip less steeply to the west than the main schistosity. If a sufficiently broad view is taken for any particular scale, the F_3 folds in this area are all asymmetric step folds rising to the west with the steep limb the long limb. Fairly symmetric F_3 folds occur on the gently-dipping short limbs of the larger asymmetric folds. A typical example of a third fold is shown in Fig. 3.2. The third 'schistosity' is, at its most intense, a strain-slip schistosity but it is not developed at all in much of the area. Even where well-developed in the semipelitic schist, very little recrystallisation of micas has taken place on the discrete schistosity planes. The common third lineation, either a crenulation or an intersection lineation, is always parallel to local fold axes.

In the area to the south of the Celebes Pond Granite, megascopic third folds are extremely rare, except for large-scale very open examples in the quartzo-feldspathic gneiss. However, an L_3 crenulation lineation is very common in this area. In the outcrops mapped to the north of the Bear Cove Road, little is seen of D_3 structures, except occasional steeply plunging open to close asymmetric minor folds usually without an axial planar fabric. Most of these are probably correlative with D_3 in the south of the area and they have been marked as D_3 on the map (Plate 2).

A few peculiar developments of the D_3 deformation were seen, especially in the psammite-amphibolite unit, where shallow to moderately dipping conjugate strain-slip 'schistosities' intersecting in L_3 may be found in place of the normal single subvertical S_3. This phenomenon is also seen in an outcrop of Birchy Schist on Kidney Pond [1] where it has locally generated elastica-type folds (Fig. 3.2).

(e) Late Deformations (D_4)

Small scale kink bands are very rare in the area of the Fleur de Lys

Schists that was mapped. Those few that were seen have not been marked on the outcrop maps. All other deformational structures later than D_3 are only found in particular small areas, and all appear to be related to the later deformation of the rocks in the Baie Verte Lineament rather than the deformation of the Fleur de Lys Supergroup with its attendant regional metamorphism. Small scale structures are found in two areas, one of which is from Camp 26 Pond southwest to the Old Camp 32 Road. Here, a fine crenulation and associated strain-slip type "fracture cleavage" is found in some parts of some outcrops, and is associated with a single large very open sinistral kink fold that seems to be connected with the shape of the southwest side of the Flatwater ultramafic body. The strain-slip fabric is steeply dipping and strikes about SW-NE. The plunge of the crenulation is variable, depending on the local attitude of the main schistosity.

The other occurrence of a late strain-slip cleavage occurs within a few outcrops located near the tectonic junction with the Baie Verte Group between 1.2 and 3.2 km. NNE of the outlet of Kidney Pond. The best example is seen in an outcrop 3.1 km. NNE of the outlet of Kidney Pond [12] where the steep "fracture cleavage" striking NE-SW is associated with a few small open vertically-plunging folds with sinistral asymmetry. The few separated occurrences of this structure would not merit comment if they were well within the Fleur de Lys terrain, but in their position adjacent to the junction with the Baie Verte Group they have significance, as they appear to be directly correlative with a strain-slip fabric in the same orientation in the Baie Verte Group. This will be referred to in the chapter on deformation in the Baie Verte Lineament.

Although both these strain-slip fabrics have been marked as D_4 on the map (Plate 1), they are probably not the same event; as they are not seen together in one outcrop, there is no way to find out with certainty.

(iv) Deformation; major structures

(a) Tectonic Slides

1. Introduction

Before the evidence for the major folds in the Fleur de Lys of the map area is discussed, major syn-kinematic tectonic dislocations present in this terrain must be described because their location is a part of the evidence used for the interpretation of the major fold structures.

The term tectonic slide (or slide for short) was first used by Bailey (1938) to describe a folded fault or décollement plane. In particular, the term was used for such tectonic junctions as the base of the Iltay succession, and for other major breaks with the Moine-Dairadian terrain that were developed during the compressive deformation. The term has been generalised from Bailey's original use (see, for example, Fleuty, 1964), and is now used for any major tectonic junction or boundary that developed during a particular episode of compressive deformation that results in development of cleavage/schistosity (with or without folding) of more than very local significance. The tectonic slides are parallel to the cleavage/schistosity, and therefore the definition does not include crosscutting transcurrent faults, although a slide may also be a transcurrent fault. Thus a tectonic slide may or may not be folded, depending on whether large scale folding followed its development (see, for example, Dewey, 1967).

A slide may develop where an element in a deforming terrain suffers much greater stretching and/or flattening than adjacent elements. The difference in the deformation may be expressed in narrow zones of intense tectonism (slides) bounding such elements, instead of being distributed across much broader gradational zones. Examples are found where extreme attenuation occurs along the limb of an isoclinal fold in a particular susceptible horizon or along the junction of two units of greatly differing

ductility, for example, a basement-cover interface. A slide zone may also be found within a deformed terrain where no obvious difference in character or amount of deformation, or in original ductility, can be detected from structural evidence seen in outcrop. In this case, it may be that the evidence is insufficient, or that the structural 'cause' of the slide is not exposed. It is theoretically possible for a slide to exist that has no displacement across it, being merely a zone of intense attenuation and flattening. However, most major slide zones usually have significant displacement on them, but they may die out in any direction. Slide zones are usually characterised by a distinct lithology, depending on the parent rock. A very intense foliation and fine grain-size are characteristic of some; these lithologies are often referred to as 'slide facies' and are recrystallised mylonites.

The term slide is a useful one for the following reasons. First, it makes clear that the structure is a major dislocation developed during a compressive foliation-producing deformation. Secondly, the lithology developed by the intense deformation along it is usually distinctive and often of mappable extent (at the scale used here). Thirdly, a slide has particular and useful geometric properties, being essentially parallel to the axial plane of folds that form in the same phase of deformation. Fourthly, in complexly deformed terrains, the original attitude or the sense of displacement of the slide is often not ascertainable. The term slide is noncommital and preferable to designating a particular kind of fault, and the characteristics listed above distinguish it from a common fault produced by brittle failure, and also from such terms as 'shear zone', used in a loose sense (not the precisely defined term of Ramsay and Graham, 1970), although they are often slides in the sense defined. Slides, particularly those with

originally steep attitudes, are usually developed only in intensely-deformed terrains and they can be of any scale; minor examples may be seen disrupting folds in outcrop. These small-scale examples are paper-thin planes rather than wider zones although they have the same geometrical properties as the major slides. Examples of them will be cited in the chapter on deformation in the Baie Verte Group. Large-scale examples are structurally important in the Fleur de Lys terrain, and are now described.

(2) Roadside Slide

This is informally named the Roadside Slide because it is exposed in several roadcuts and quarries on the western side of the Baie Verte Road, from the north end of Mic Mac Lake to where the stream from Celebes Pond crosses the road. Further exposures that indicate proximity to the slide occur immediately north of Tom's Pond and Kidney Pond. The best exposures of the slide occur in roadside quarries on the west side of the Baie Verte Road; one is 0.9 km NW of the north end of Slink Pond [13], the other 0.8 km south from the north end of Mic Mac Lake [14]. The slide zone is composed of strongly-foliated semipelitic schist, with subordinate graphitic semipelitic schist and mafic schist mixed in a zone about 100 metres wide. In outcrop [14], some parts of the zone have a gneissic appearance with regularly-spaced highly-segregated folia of quartz-plagioclase, and mica, including a large proportion of orange biotite relative to muscovite. In other outcrops on the margin of the slide zone, bands of monomineralic orange- biotite schist a few centimetres thick occur, best seen in the roadcut 0.6 km WNW of the south end of Slink Pond [15]. However, the most striking, consistent, and diagnostic feature of this slide zone is the presence of abundant small lenticular bodies up to 1 metre long of tremolite-fuchsite rock. These bodies are usually strongly-foliated, but some are statically recrystallised with coarse sheaves and rosettes of acicular tremolite.

These tectonic lozenges are seen in outcrop [14] folded by F_2 minor folds. Also minor F_2 folds of the intense foliation elsewhere in this and other outcrops indicate that the development of the slide zone is of D_1 age, and the emplacement of the meta-ultramafic tremolite-fuchsite lenses is of this age or earlier. Also seen in outcrop [14] is a large (30 x 5 metres) tectonic lozenge of metagabbro, and some of the mafic schist in this outcrop, especially the leucocratic varieties, has been derived from this metagabbro.

Almost all the tremolite-fuchsite lenses are apparently contained within the intensely-deformed zone of the slide from Mic Mac Lake to the Celebes Pond Granite, but in some of the occurrences north of Kidney Pond they are in schists that do not have the intense foliation of the slide zone. However, on the structural interpretation chosen, they are very close below the slide zone, and probably have been tectonically introduced from it, perhaps later in the deformation sequence. They are not found randomly within the Birchy Schist Formation, as has been previously implied (Kennedy, 1971); their association with this D_1 slide is quite clear.

The tremolite-fuchsite lenses derive from an ultramafic, chromite-bearing parent rock. This may have been either the ultramafic basal part of an ophiolite complex (oceanic uppermost mantle), or a picritic sill. I favour the ophiolite parentage because of the association with the gabbro, and because of their occurrence on a major tectonic dislocation over a strike length of 11 km.

The amount and sense of displacement and original orientation of this major D_1 slide zone is uncertain. Because of the width of the deformed zone and the presence of probable ophiolite-complex-derived slivers on it, it is my view that the displacement was westward on a gently dipping thrust plane, and of considerable magnitude. The westward displacement is favoured because

continental-type crust occurs under the Fleur de Lys metasediments a few kilometres to the west (deWit, in prep.) and oceanic crust is known under the eastern division of the Fleur de Lys Supergroup (Dewey and Bird, 1971).

(3) Noseeum boundary slide

The lithologic boundary between the Western sequence and the Rattling Brook Group is always extremely abrupt. An outcrop exposing the contact zone is not present and the smallest gap across the contact is 8 metres between two outcrops located 1.2 km WNW of the northern end of Mic Mac Lake [16]. Here, the psammite closest to the gap is somewhat finer-grained and well-foliated than that away from the contact. The semipelitic schist contains thin bands of mafic schist that appear rather more strongly-foliated than usual. On the outcrop of semipelitic schist in this locality, there is a very angular boulder about a metre across, consisting of flinty psammite with amphibolite bands about 5 centimetres thick, both very fine-grained and intensely-foliated. This rock is a typical sample of the rock type developed on slides involving psammitic rock. The amphibolite is probably a very-deformed version of the normal plump unfoliated amphibolite boudins found in the psammite-amphibolite unit to the west. The angularity of the boulder indicates that it has not travelled very far. The indicated local ice flow direction is along the contact, and the contact zone is the most likely position of a slide from which to derive this boulder.

In some other outcrops near the contact, the psammite appears to be rather more deformed than usual and the amphibolite boudins are more elongate)and quite markedly so in a few outcrops. This indicates greater than normal local flattening. However, most outcrops near the contact, especially those of semipelitic schist, are highly uninformative. The proposed slide is probably D_1 in age, although this cannot be proved in the absence of

outcrop, and it may have had significant D_2 movement on it as well. No sign of any ultramafic tectonic lenses were seen in outcrops near the contact. The displacement and original orientation of this slide are not known.

M. J. deWit (pers. comm.) has mapped most of the width of the semipelitic schist along the Bear Cove Road and Middle Arm Brook as a large slide zone, and while this may be so, to me it does not differ materially from the rest of the semipelitic schist in the map area, and I prefer to map it as the same lithologic unit, differentiated sharply from the psammitic rocks to the west.

The width of the semipelitic schist unit adjoining the contact varies from about 500 metres on the Bear Cove Road to nothing west of northern Mic Mac Lake. Also, near Gull Pond, the quartzo-feldspathic gneiss abuts the contact instead of the psammite-amphibolite unit.

It is admitted that the evidence for a tectonic slide along the contact is weak. However, the alternative stratigraphic hypotheses of a conformable or unconformable contact are less attractive. Both require the original absence of most and locally all of the 5 km thick Seal Cove Group that is present between basement and Rattling Brook Group equivalent on White Bay. In the author's view, the evidence favours a tectonic slide contact.

(b) Major fold structures

(1) Introduction

In the Fleur de Lys terrain mapped, nearly all the available outcrop has been examined in the area south of a line running east through the small pond north of Tom's Pond and thence northeast to the Baie Verte-Fleur de Lys contact. Thus, remaining uncertainties of interpretation are inevitable in this area. North of the line, outcrops are available in some areas that were not examined. Airphoto interpretation has been used to

Fig. 3.3. Sketch map of proposed early structures ($D_1 + D_2$) of the Rattling Brook Group in the map area, with D_3 and later structures removed.
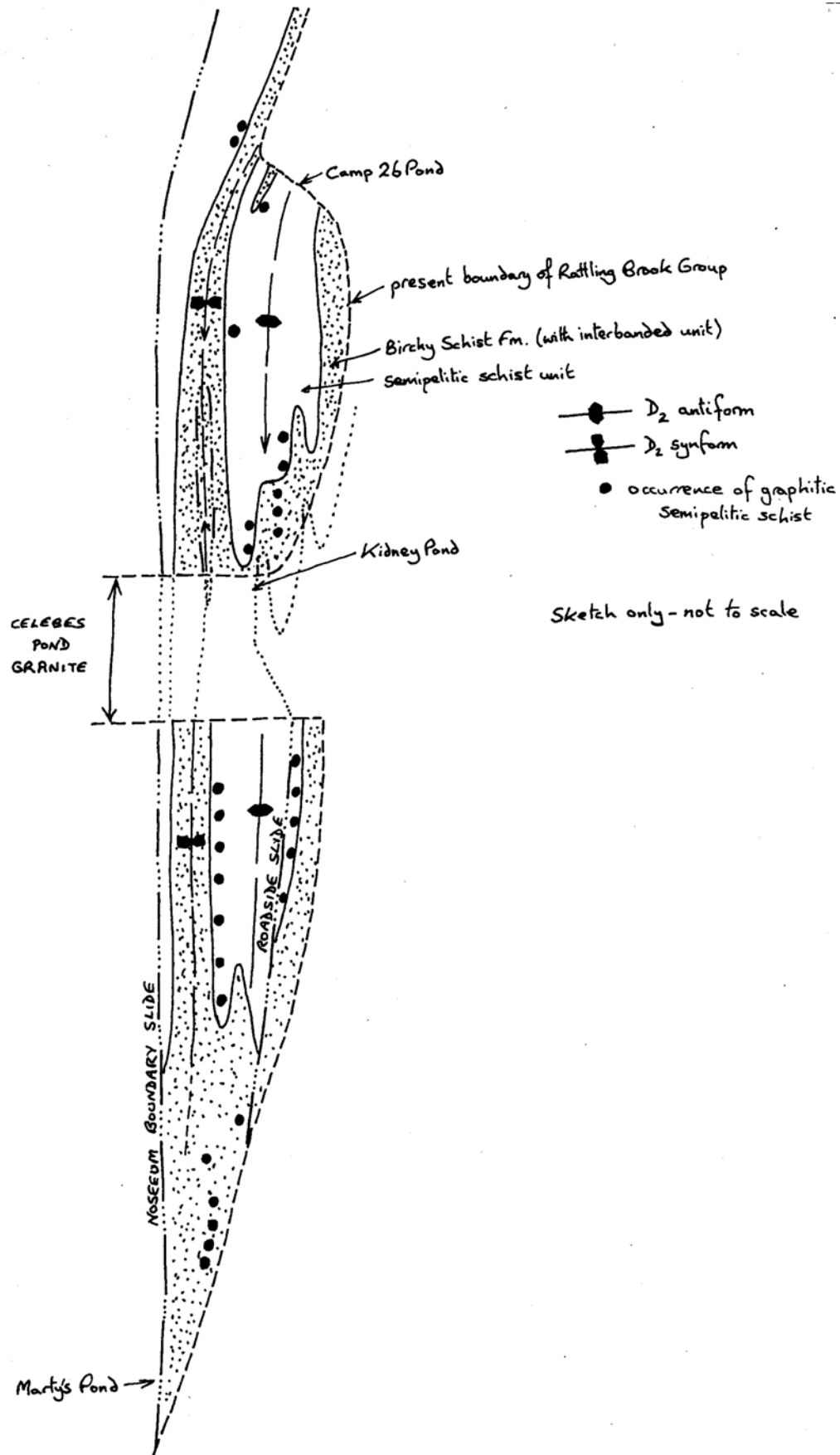
supplement the interpretation in all areas. In particular, in the strip of Birchy Schist and

psammite marked from northeast of Tom's Pond to east of Jack's Pond, a rather speculative

airphoto interpretation results from combining the airphoto data with geological data at either

end of the strip. There is some outcrop in this strip, but it was not examined. The medium- scale

F_2 antiform-synform pair marked 2.4 km NNE of Kidney Pond outlet is based on a form

observed on the air photographs as there is no outcrop in the critical area. In addition, the exact

position of the boundary between the Western sequence and the Rattling Brook Group is poorly

defined by mapped outcrops from the Bear Cove Road to the area northwest of Kidney Pond.

Except for a single traverse that confirmed the extent of the semipelitic schist unit, its position is

based on airphoto interpretation and extrapolation of the structure known around Kidney Pond,

supplemented by data from the aero-magnetic maps.

(2) Early folds (D_1 and D_2)

In the area just north of Kidney Pond, a large-scale fold closure, accompanied by several

smaller parasitic folds, is defined by the mapped lithologic distribution. This large fold is

interpreted as a second (F_2) fold because it seems to fold the D_1 Roadside slide, and as an

antiform because the plunge of minor F_2 folds in this area is mostly southward. The axial-trace of

this fold continues northward to Camp 26 Pond (Fig. 3.3). The graphitic facies of the semipelitic

schist unit occurs in places near the contact with the Birchy Schist Formation on both limbs of

this major fold, and also on the Bear Cove Road in the same relative position but on the western

side of the belt of Birchy Schist Formation that forms the western limb of this F_2 antiform (Fig.

3.3). This indicates that all the semi-pelitic schist in this area is probably the same unit,

structurally below the Birchy Schist Formation, and that there is a major early fold in this

western belt of the latter formation (Figs. 3.3 and 3.4). The roadside slide is present in this belt along the eastern side of Tom's Pond, but it has not been detected on the old Camp 32 Road or further north, which suggests that the fold closure within this belt of Birchy Schist is a major syn- formal second (F_2) fold.

The area south of the Celebes Pond Granite, although offset, appears to be equivalent to the area north of the Granite at a different (lower?) structural level. After correlating the Noseeum boundary slide across the Granite, the four main lithologic belts seem to equate with those to the north, with a postulated F_2 synform in the western belt of Birchy Schist, and an F_2 antiform in the main belt of semipelitic schist. The occurrence, and absence in other positions, of the D_1 Roadside Slide is compatible with this interpretation (Fig. 3.4).

However, two observations may imply that a major first fold closure (F_1) is present. The first is the continuity of the belt of Birchy Schist Formation from Jack's Pond at least as far as the northern end of the area mapped, which may indicate that its position is controlled by a structure larger and more fundamental than a major F_2 synform. The continuity could be accounted for if the plunge of the F_2 synform changes to northward in the area north of Jack's Pond. The little data on the plunge of F_2 minor folds in this area suggests that this might be the case (Plates 1, 2). Secondly, in the area opposite the northern half of Mic Mac Lake, the well-defined separate belts of Birchy Schist and semipelitic schist found to the north cannot be traced, and the two lithologies become closely interbanded and unseparable on the scale of Plate 1. In the north of this area this may be partly due to the poor outcrop, but in the south the merging of the two units may indicate the presence of a major first fold closure (F_1), rather than a facies change. However, the interpretation of the major

Fig. 3.4. Sketch cross-sections of the Fleur de Lys Supergroup in the map area (a) near Kidney

Pond; (b) west of Slink Pond
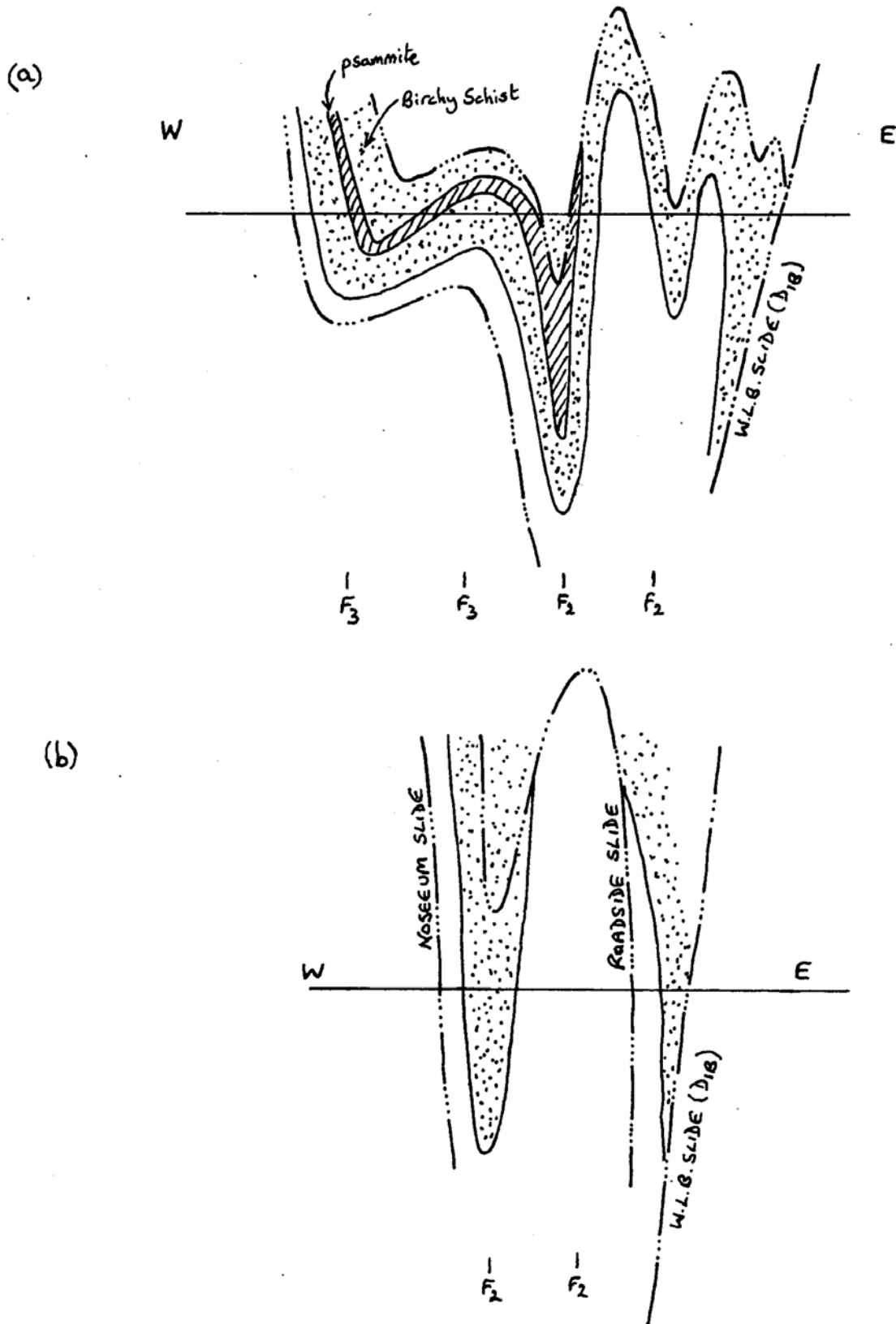
Section (a) is about 3 km. long; (b) about 2 km. long.

structure presented (Figs. 3.3, 3.4) is the simplest that is consistent with the inadequate data and does not involve major F_1 folds, but it depends critically on two particular interpretations. First, it depends on the correct identification of the outcrop of the D_1 Roadside slide (that is both where it is present and where it is absent), especially north of the Celebes Pond Granite. For instance, if it crosses the Old Camp 32 Road and continues to Red Cliff Pond, the western belt of Birchy Schist may contain a major first fold closure (F_1), instead of a second fold closure. Secondly, the hypothesis that both belts of semipelitic schist are the same unit could be incorrect. It is therefore possible that the interpretation of the large-scale early structures may need revising if more data are collected.

(3) Third folds (D_3)

Large-scale third folds are present only in a strip from the area of Gull Pond to the area around Kidney Pond. Axial traces of these folds, which essentially combine into one huge asymmetric antiform-synform pair, are summarised on Fig. 3.5, and the folds illustrated in a cross section on Fig. 3.4(a). They are open asymmetric step folds verging up to a hypothetical antiforin to the west, identical to their minor equivalents. The large folds plunge gently southward, except for the area between 4.4 and 6.6 kilometres north to northnorthwest of Kidney Pond. Here the plunge reverses to gently northward, at least partly causing the great extent of the semipelitic schist in this general area. Within the strip occupied by these large-scale folds, small-scale F_3 folds are abundant. Elsewhere in the area (except east of Tom's Pond) minor F_3 folds are either less common or absent; this is in contrast to the L_3 crenulation lineation which is widespread, at least south of the Bear Cove Road. The deformation is the last of the regionally significant deformations of the Fleur de Lys schists in the map area. It is important to note that these large-scale F_3 folds and their axial-

Fig. 3.5. sketch map of late structures ($D_3 + D_4$) in the Fleur de lys Supergroup of the map area.

local D_4 kink plane trace

S_2 orientation

(a) Kidney Pond – N. Flatwater Pond area

NOSEEUM BOUNDARY SLIDE

Sketch maps – not to scale

F_3 axial trace

(b) Marty's Pond area

MARTY'S POND FAULT

NOSEEUM BOUNDARY SLIDE

Birchy Schist Fm.
(includes interbanded unit)
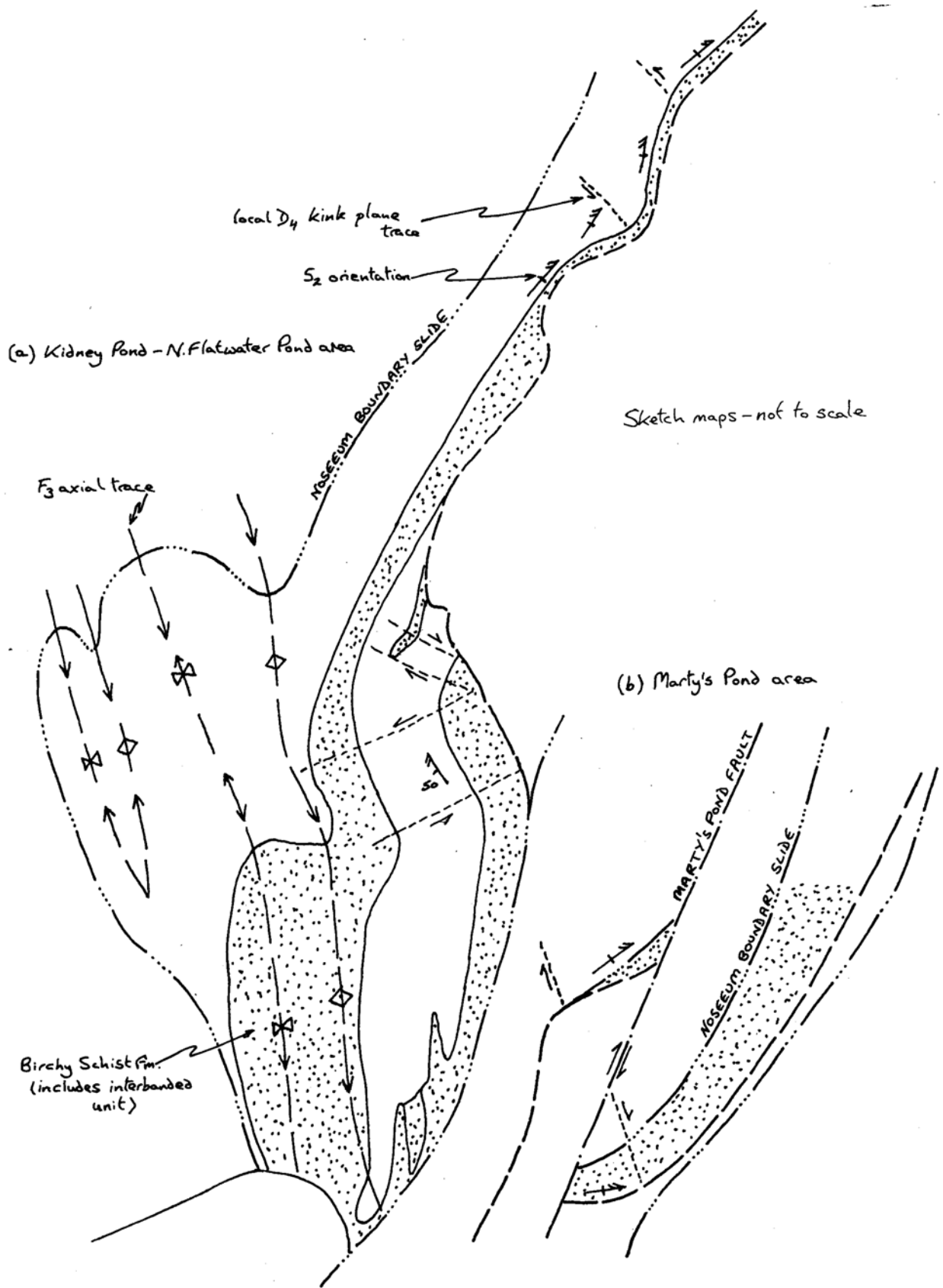
traces are truncated by both the Fleur de Lys-Baie Verte tectonic boundary, and within the Fleur de Lys terrain by the post-kinematic Celebes Pond Granite, whose relationship is described later.

(4) Later deformations ('D$_4$')

These consist of three separate very open kink folds, all local and apparently connected with the shape of parts of the western sides of the two large ultramafic bodies in the map area. A large very open sinistral kink fold defined by the attitude of the main schistosity occurs locally near the northern end of the Flatwater ultramafic body (Fig. 3.5a). Here the main foliation in the schists remains subvertical through the fold, indicating that its axes are subvertical. No associated minor structures have been detected. The second kink fold is a sinistral structure in the area of the southern end of the Flatwater ultramafic body (Fig. 3.5a), and is modified by a smaller dextral kink fold to the north, making an incomplete box fold. The sporadic occurrences of associated minor structures have already been described. In the main limb of this kink fold the main foliation in the schists strikes NNW and dips on average about 50 degrees to the southwest. Since the foliation is subvertical on the unkinked limbs, the kinkfold axes plunge moderately southward. The aeromagnetic map shows that the anomaly due to the Flatwater ultramafic body extends westward in this area under all the southern part of Camp 26 Pond, where Fleur de Lys schists would outcrop. The ultramafic body probably expands at shallow depth in this area, and the kink fold is an accommodation to this shape.

The other late fold is found at the southern end of the map area, and is now in two pieces, south and north of Marty's Pond, disrupted by a large dextral fault with an apparent horizontal displacement of 3.2 kilometres (Fig. 3.5b). Only one diffuse hinge area of this dextral fold is seen, because the other is in the Wild Cove Pond granite-diorite complex.

Fig. 3.6. Lower-hemisphere stereographic projection showing orientation change of S_2 and L_3 around D_4 fold in Marty's Pond area.
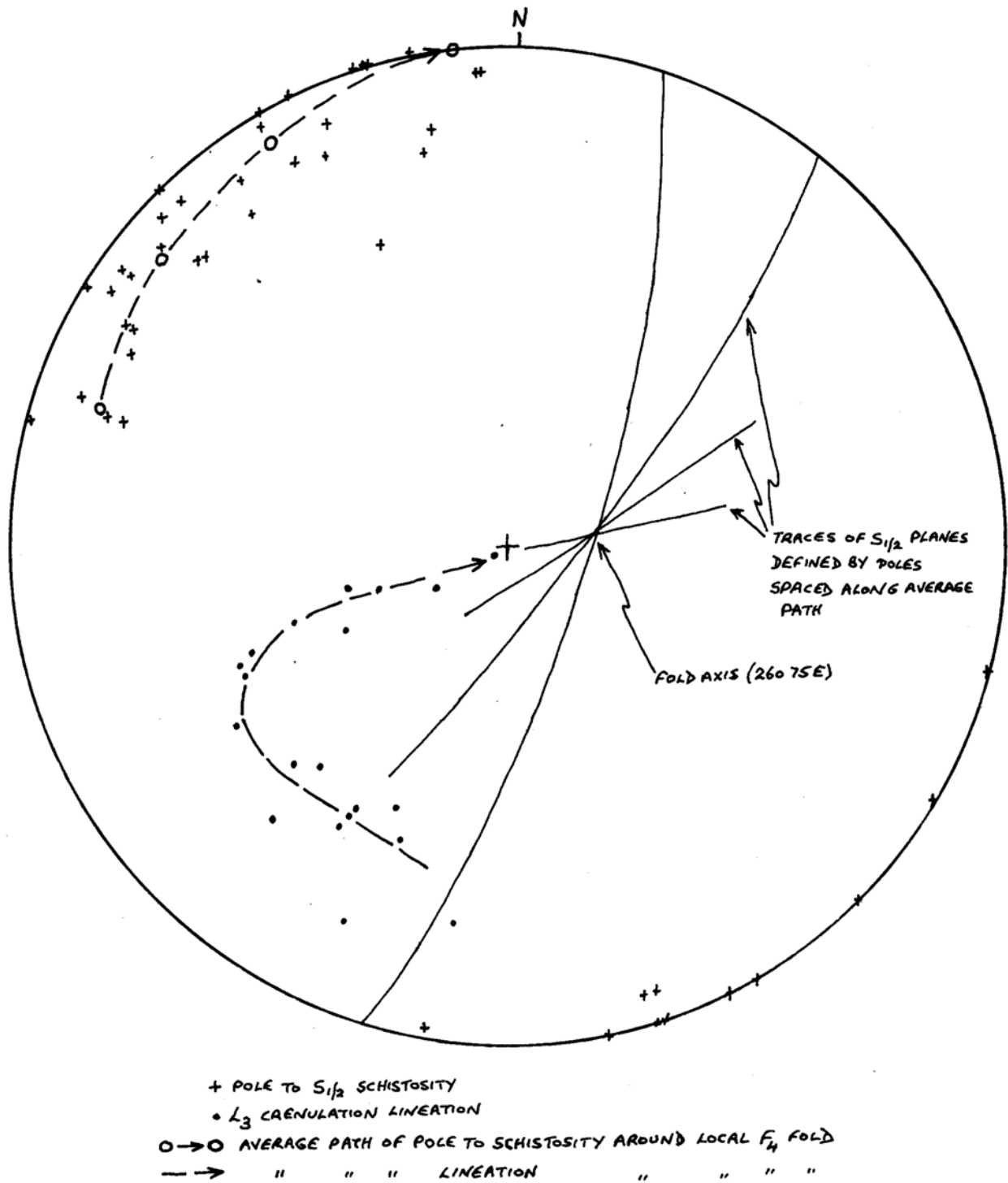
The fold and its disruption by the large fault is best defined by the Noseeum boundary slide. The combined effect of the fold and fault provides the explanation for the U-shaped northern end of the Mic Mac ultramafic body, which previously was ascribed to regional D_3 Fleur de Lys folding (Neale and Kennedy, 1967). It is clearly not due to an F_3 fold, as the L_3 lineation together with the main S_2 schistosity is folded by it. The orientation of the S_2 schistosity and the L_3 lineation around this fold and its displaced continuation are summarised on a stereogram (Fig. 3.6). The same data on Plate 1 show that the S_2 schistosity dips steeply east to south all round the fold, while the L_3 lineation steepens to subvertical in the east-west limb, far more than would be expected around a normal cylindrical fold. This may indicate that the fold is conical (Ramsay, 1967), but the average S_2 planes intersect in a normal single fold axial direction on the stereogram. Therefore large-scale simple shear deformation has probably occurred on the short limb, with downthrow on the west side of a north-south plane. It is suspected that the large dextral fault has an element of vertical displacement in this sense, because the Fleur de Lys terrain to the east appears to be a lower structural level than to the north of Kidney Pond. The fault does not have the correct orientation (but has the right displacement sense) to be a kink plane for the late fold. If displacement on the fault is removed, the fold becomes an accommodation structure to the shape of the ultramafic body, as in the other two examples. The Wild Cove Pond granite-diorite complex was almost certainly in place before the tectonic emplacement of the ultramafic body, and the formation of this fold. Several NW-SE lineaments are visible around this south-east end of Wild Cove Pond, and it is proposed that these are dextral faults accommodating the massive intrusive rocks to the fold. The main part of Wild Cove Pond (see Plate 6) is aligned parallel to these lineaments, so the pond may occupy a (glacially scoured)

shatter belt due to many parallel faults. These three late folds are most probably due to deformation in the Baie Verte Lineament and the accompanying emplacement of the ultramafic bodies. This view is amplified in a later chapter.

(5) Faults

Some small late faults mostly belonging to a conjugate wrench set have been marked, on outcrop evidence combined with lineaments seen on air photographs. Some more lineaments of this nature are present, but if they represent faults, the displacement is not sufficient to be shown by the outcrops available. Most appear to have formed before the development of the Baie Verte-Fleur de Lys tectonic contact.

(v) Metamorphism

The Fleur de Lys schists have a complex metamorphic history; the observations set down here are not exhaustive, and there is scope for further study of the metamorphism in this map area.

The rocks seen in thin section are all well-crystallised and have a 'clean' appearance; in general little retrograde metamorphic mineralisation is present. Most of the non-porphyroblastic constituents have grown at the latest during the formation of the second schistosity although they have probably been annealed and undergone static grain growth later. In the semipelitic and psammitic schists a few large muscovite flakes are commonly seen overgrowing the aligned S_2 fabric. Growth of minerals on S_3 is either negligible or confined to minor, partly transposed muscovite. No recrystallisation definitely attributable to D_3 has been seen in the mafic schists.

Plagioclase porphyroblasts are often weakly augened in S_2. Inclusion trails are not common, and most of those seen are uninformative. Some indicate post-S_1, pre-S_2 growth. Some thin sections show garnet porphyroblasts

weakly augened in S_2 and containing S_1 inclusion trails. In most thin sections, garnets are not obviously augened in S_2 but an unequivocal relationship to has not been seen.

A garnet isograd has been mapped, mainly from outcrop observation, but supplemented by information from a few thin sections. This isograd is mostly subparallel to the regional strike (the main composite foliation), except east of Tom's Pond, where it cuts directly across it. This relationship may indicate that the garnet growth in this area, and perhaps in the rest of the area, is mainly post-D_2. However, garnets in two thin sections from the area east of Tom's Pond are weakly augened in S_2. It may be that these grew during the time the S_2 schistosity was forming at a late stage in the D_2 folding process, or that the S_2 schistosity has been enhanced during D_3.

The bulk of the rocks in the mapped area are in either biotite or garnet grade (upper greenschist and epidote-amphibolite facies assemblages, (Turner, 1968). A small area adjacent to the Baie Verte Lineament north of Middle Arm Brook is in chlorite grade, but insufficient mapping was done here to define a biotite isograd. Two areas elsewhere contain assemblages of higher metamorphic grade. The first is the metamorphic aureole of the northern margin of the Celebes Pond Granite, and this contains rocks in amphibolite facies. The mafic schists contain hornblende and andesine, but no garnet. The semipelitic and psammitic schists contain oligoclase, but garnet is no more abundant than elsewhere inside the garnet isograd. However, andalusite is common, and clearly overgrows F_3 microfolds. Plagioclase porphyroblasts in the schists within the andalusite isograd marked on Plate 1 may be partially to wholly replaced by microline with myrmekite spots. A few post-D_3 zoned porphyroblasts of sodic plagioclase occur in pelitic bands in the psammite member of the Birchy Schist Formation. In some of these pelitic bands, some andalusite porphyroblasts appear to have

been replaced by fibrous sillimanite but others are retrogressed to a felted micaceous aggregate. Where the latter occur, garnets are partially retrogressed to chlorite and tourmaline is common, suggesting some local pneumatolytic alteration. A more widespread alteration in this area is the association of garnet and clumps of biotite, which may indicate a retrograde reaction. The andalusite isograd has been mapped mainly from thin section data, as it is not easy to identify positively in outcrop in the coarse grained schists, except where it is abundant in mica-rich pelitic schist. However, all samples that were suspected in the field to contain andalusite were found to do so on thin section examination. A few samples found to contain only a relatively small amount of andalusite were not detected in the field. The cross-strike deviation of the garnet isograd east from Tom's Pond appears on the map as if it may be due to the Celebes Pond Granite. This is probably not so, as syn-S_2 garnet occurs north of the outlet of Kidney Pond, as the granite and its andalusite isograd were intruded and formed after D_3, and as the garnet isograd is cut by the south-eastern margin of the granite.

The other area where high-grade metamorphic assemblages have been found is around Marty's Pond. Garnet-amphibolite occurs in outcrops of the psammite-amphibolite unit east of the centre, at the north end, and 2 km NE of the north end of Marty's Pond. However, garnet is not seen in mafic schist of the Rattling Brook Group just east of these localities. These coarse poikiloblastic garnet amphibolites were probably formed during the regional metamorphic event rather than the later contact metamorphic event due to the Wild Cove Pond granite-diorite complex. In addition, outcrops of semipelitic schist in the Rattling Brook Group near the junction of the Camp 164 Road with the Baie Verte Road (e.g. [5]) contain post-D_3 andalusite, but there is insufficient data in this area to draw an isograd. In two thin sections andalusite is seen replacing staurolite; in one case, the latter was tentatively

identified in the field. One of the specimens comes from an outcrop only 70 metres from the contact with the Mic Mac ultramafic body and the Baie Verte Group. The age of the staurolite is post-D_2, but its relationship to D_3 was not seen. It is probably due to the regional metamorphism; the andalusite is post-D_3 and due to the contact metamorphism of the Wild Cove Pond Complex. In roadcuts, on the eastern side of the Camp 164 Road 1.2 km from the north end of Marty's Pond, psammite of the psammite-amphibolite unit contains abundant well-formed post-D_3 zoned microcline-perthite porphyroblasts up to 10 cm long, with marginal myrmekite spots. Their cores are microperthite, and the rims are microcline. They are accompanied by oscillatory zoned plagioclase porphyroblasts up to a centimetre across. This is an example of the rock known as 'dent-du-cheval'. It represents local potash metasomatism from the nearby Wild Cove Pond Complex.

Lastly, minor but significant retrograde metamorphic effects are seen near the tectonic contact between the Baie Verte Group and the Fleur de Lys Supergroup schists. In an outcrop on Kidney Pond [13], some F_3 microfolds in amphibolite facies schist, which were fortuitously thin sectioned, have been tightened by buckling subsequent to their formation, and prehnite sheafs have grown in the crescentic voids between some of the folia around the fold hinges, together with minor spots of (?)pumpellyite. An outcrop next to the contact 1.5 km NNE of Kidney Pond outlet has coarse-spaced remobilisation along some of the S_2 planes, and chlorite has grown in the S_2 planes so reactivated. Where the local D_4 fracture cleavage occurs [12] 3.1 km NNE of Kidney Pond outlet, chlorite is present, perhaps as a retrograde metamorphic effect. Lastly, in a roadcut 3.7 km north along the Baie Verte Road from the Burlington Road Junction, an outcrop of mafic schist is severely retrograded and structurally sliced. This effect is confined to the eastern half of the outcrop (about 1 metre width). These occurrences

are respectively 400, 20, 80 and less than 10 metres from the tectonic contact with the Baie Verte

Group. Further north along the Baie Verte Road from the last occurrence, phyllitic muscovite

sheen and porphyroblastic muscovite occur on irregularly rumpled S_2 schistosity surfaces in

chloritic semi-pelitic schist. Thin sections show that this muscovite is part of the S_2 schistosity,

which is here relatively fine-grained. The rocks here are of particularly low original metamorphic

grade, and the muscovite is probably not a retrograde metamorphic effect, as previously

suggested (Neale and Kennedy, 1967). Four specimens were taken from between 3 and 6 metres

away from the tectonic contact with the Flatwater ultramafic body, and show no trace of

retrograde mineral growth or structural modification. It is concluded that very little retrogressive

metamorphism is present near the tectonic contact with the Baie Verte Group, and most is

confined to a zone within a few metres of the contact.

Observations and inferences pertinent to the evolution of the Baie Verte Lineament that

have been found in the western division of the Fleur deLys Supergroup are summarised at the

end of this chapter.

B. Post-Kinematic Granite and Diorite.

(1) Introduction

Large areas of the central and southern terrain of the western division of the Fleur de Lys

Supergroup are occupied by a distinctive porphyritic biotite granite (Fig. 1.3). Published

descriptions of this rock unit are brief (Neale and others, 1960; Neale and Nash, 1963), and state

that the rock is somewhat gneissic in some places, and that this often grades into 'hybrid gneiss'

at its contacts with the Fleur de Lys schists. This intrusive body covers a large area and only a

very small part has been examined in the course of this mapping.

(ii) Celebes Pond Granite

This intrusion is separate from the main one to the southwest, and typical outcrops of the granite are seen near the small pond that has been named here for the purpose of naming the intrusion. This porphyritic biotite granite contains pink anhedral to subhedral phenocrysts of microcline microperthite commonly up to 3 cm long. Subordinate zoned subhedral plagioclase and interstitial quartz occur with biotite and accessory magnetite, epidote, sphene, apatite, and zircon. Some large (0.5 cm) subhedral zoned allanite crystals rimmed by epidote are also found. In precise terms, the rock is an adamellite, not a granite. Aplite veins are common, but xenoliths are very rare, and pegmatitic features were not seen. The northern contact northwest of Kidney Pond is expressed only by small apophyses up to a metre wide of pink granite and aplite in the Fleur de Lys schists. These follow joint planes, indicating a stoping method of 'passive' emplacement, and were definitely intruded after D_3 deformation. The contact proper is not exposed but is inferred from the sudden disappearance of outcrop. The granite has caused contact metamorphism, partially overprinting the Fleur de Lys regional metamorphic assemblages. This is most clearly shown around this northern contact by the occurrence of post D_3 andalusite in the pelitic and semipelitic schists. The andalusite isograd mapped (Plate 1) is broadly parallel to the northern margin of the granite, but diverges from it to the west, suggesting that the margin is less-steeply dipping in this direction. The actual contact of the granite along its southeastern side is also unexposed. In contrast to the area north of Kidney Pond, the garnet isograd is not concordant, and appears to be cut by the contact. Garnet growth was probably not promoted by the contact metamorphism, and the apparent concordance of the contact and the garnet isograd north of Kidney Pond is probably fortuitous. However, andalusite was not detected in the field in the semipelitic schists

near its southeastern contact. No thin sections were cut from the rocks in this area, and andalusite is difficult to recognise in outcrop in these schists, but it is not thought to be present in significant quantities (see p. 3.34). The postulated dextral fault along the contact does not affect this problem. A possible explanation for the apparent absence of an aureole in this area is that it is very narrow. Almost all the few outcrops of granite nearest this southeast contact are quartz-vein breccias and the biotite is chioritised in several outcrops near the contact. It is thought that this is evidence that indicates a fault along this contact, although quartz-vein breccia also occurs in one outcrop elsewhere.

It is important to note that the zone of apophyses and the andalusite isograd adjacent to the northern contact of the granite are cut by the tectonic contact with the Baie Verte Group.

(iii) Wild Cove Pond Complex.

This name is proposed for the intrusion of granite and related rocks between Wild Cove Pond and the large fault running NW out of Black Lake (Plate 6). It forms an elliptical body, cut off from the main area of granite to the south by the fault, which appears as a 'neck' in the boundary (Fig. 1.3). Only a small amount of this body was seen, partly in reconnaissance traverses outside the area of Plate 1.

Southwest of Marty's Pond (Plate 1), porphyritic biotite granite outcrops, and is identical to that in the Celebes Pond Granite. Northward there is an area lacking outcrop, and then around the southeast end of the main part of Wild Cove Pond, diorite and quartz diorite (tonalite), (collectively termed 'diorite' hereafter), are the main rock type found. In this area and elsewhere in the complex the 'diorite' has a variable composition, ranging from hornblende diorite through biotite-hornblende diorite and hornblende-biotite-quartz diorite (tonalite) to biotite-quartz diorite (tonalite). A

small proportion has sufficient quartz to be called granodiorite, although it has an atypically large proportion of mafic minerals, and is not particularly coarse-grained. Plagioclase may be granular, but is more commonly in subhedral laths, sometimes much coarser-grained than the rest of the constituents. It may be weakly or strongly zoned, and the composition is usually in the range. An_{45} to An_{18}*. Besides green hornblende and olive-green biotite, some or all of the accessories epidote, magnetite/ilmenite, apatite and sphene may be abundant, and zircon and allanite may occur. Quartz ranges from absent to 30% of the rock, and may be poikilitic. The variation from 0 to 30% can occur over a metre, perhaps indicating that incomplete mixing of different melts has occurred. Minor microcline microperthite may occur in some of the quartz diorites.

Around the southeastern end of the main part of Wild Cove Pond, and near the contact with the psammitic schists of the psammite-amphibolite unit, complex mixture rocks of both dioritic and 'granitic' phases are found together with large 'schlieren-type', rather 'plastic'-looking xenoliths of the psammite. The 'granitic' phase in these rocks may be diorite or psammite overgrown by porphyroblastic K-feldspar and sodic plagioclase, rather than granite itself. Veins of granite and aplite are present in many outcrops of diorite, and beyond into the psammite-amphibolite unit, which in this area is often affected by growth of many large microcline microperthite porphyroblasts, previously described. Relationships of xenoliths and veins observed in this area show clearly that both the diorite and the granite were intruded after D_3 of the Fleur de Lys polyphase deformation. They also show that the granite is later than the diorite.

* Specific plagioclase compositions of igneous rocks are not quoted elsewhere because this is the only igneous rock unit in the map area where it is not totally albitised and/or saussuritised in the thin sections examined.

The variable composition of the 'diorite' alone is not restricted to this small area, or to the vicinity of the contact, but seems general in the areas examined. A woods road runs into the centre of the intrusion, around the southern extension of Wild Cove Pond, and northwest for about 4 km along the southern side of the ridge on the southwest side of Wild Cove Pond. In several places in the western 3 km, where the outcrops are almost all of 'diorite', there are intrusion 'breccias' of diorite in quartz-diorite. The diorite xenoliths are rounded and have slightly diffuse margins, indicating intrusion of the quartz-diorite into still very hot diorite. From the western contact of the Mic Mac ultramafic body, there are few outcrops along the first 4 km of this woods road, but those seen are porphyritic biotite granite, some of which shows flow alignment of the microcline phenocrysts. Within the 'diorite' beyond this, at about 4.5 and 6 km from the ultramafic contact, a granular muscovite-biotite microgranite occurs, but its relationship to the 'diorite' was not observed. In the more westerly of these two localities, many large angular boulders probably not far from outcrop show single shell orbicules in the microgranite. These have a blackish core about 2 cm across and a pink rim about 0.5 cm wide. The cores contain a slightly greater amount of micas than the matrix micro- granite. The rim contains a slight excess of microcline and lacks micas relative to the matrix microgranite, but no textural change occurs in either part of the orbicules.

Outcrops in two areas near the contact show evidence that the diorite was, at least in part, forcefully emplaced in contrast to the passive stoping emplacement of the granite. On the central eastern shore of Black Lake (Plate 6), exposures of diorite and quartz-diorite farthest from the contact with the Fleur de Lys schists are isotropic, but the outcrops nearest the contact (which is not exposed) have one variably-developed

steep foliation. In these, bands of leuco-quartz diorite alternate with dark diorite in a migmatitic rock. In thin section they are seen to be rocks deformed and foliated while still hot. Large zoned plagioclase laths in the leuco-quartz diorite bands are kinked, slightly broken and augened by biotite, although not all are affected, and some appear to overgrow the foliation. The quartz is strained and granulated, and some internal sutured boundaries have developed. These rocks are deformed agmatites, or intrusion breccias. All gradations are seen in outcrop in this area between these foliated rocks with aligned mineral fabrics, but still containing recognisable plutonic features, and isotropic rocks, although undeformed agmatite is not seen. The deformation represents movement in the still plastic 'diorite' near the contact during emplacement. Very similar deformed diorite-quartz diorite agmatite occurs in outcrops near the contact with Fleur de Lys schists on the Camp 164 Road, 3.6 km north of the north end of Marty's Pond, and beside the stream at Camp 164. The regional map (Neale and Nash, 1963) shows that foliation near the margin of this intrusive complex parallels its elliptical outline and also that the main foliation in the Fleur de Lys schists to the north of it is aligned subparallel to the contact instead of the usual NNE-SSW regional strike. A circular aeromagnetic anomaly is also associated with this elliptical intrusion. I suggest that it is a separate sub-intrusion from the main mass to the south, and it was,at least in part, forcefully emplaced with a zone of movement on or near its contact.

Just before Camp 164 is reached an angular boulder of porphyritic granite 3 metres high occurs by the road. The K-feldspar phenocrysts are aligned and the biotite forms a parallel foliation defining weak augen round the subhedral phenocrysts; this could be due to a magmatic flowage phenomenon, perhpas related to motion on the contact. This boulder displays, on its northern face, a piece of Fleur de Lys amphibolite elongate across the foliation.

3.44

I interpret this as a xenolith, but M. J. deWit (pers. comm.) interprets this, and the foliation in the deformed agmatites, to mean that the intrusive complex was involved in the Fleur de Lys deformation. I disagree with this interpretation because both granite and 'diorite' elsewhere are seen to crosscut D_3 Fleur de Lys structures.

Insufficient thin sections have been cut from rocks in the area around Marty's Pond to define an andalusite isograd. The post-D_3 andalusite in semipelitic schist near the junction of the Camp 164 and Baie Verte Roads is probably due to contact metamorphism by the Wild Cove Pond complex, but it is not known whether it is due to an unexposed offshoot underneath this occurrence, or to the main intrusion as exposed when displacement on the Marty's Pond Fault is removed. Near this locality a post-D_3 dyke 8 metres wide of chalky-weathering grey micro-spherulitic rhyolite is found [17]. A microporphyritic rhyolite also occurs in the northernmost outcrop of the northeast prong of the Mic Mac ultramafic body [18], where the grey flinty brecciated rock is tectonically abutted by sheared and foliated talc- carbonate rock. The relationship of the rhyolite to the Fleur de Lys schists is not exposed. Both these rhyolite dykes could be related to the Wild Cove Pond Complex. Also in this general area, in an outcrop just east of Marty's Pond [19], and in outcrops of the Roadside Slide 0.8 km south of the north end of Mic Mac Lake [14], there are post-D_3 mafic dykes 10 cm to a metre wide. They contain resorbed xenocrysts of quartz, saussuritised zoned plagioclase, and pink K-feldspar up to 3 cm long, which were probably derived from the porphyritic biotite granite. The dykes appear to be andesites, perhaps related to the diorites of the Wild Cove Pond Complex. They contain calcite-filled amygdales, which are abundant in outcrop [19], indicating that their structural level at the time of intrusion was probably

not more than a kilometre or so below the surface (Moore, 1965).

In the Wild Cove Pond-Mic Mac Lake area no granite or diorite apophyses are found in any outcrops of the ultramafic body or in the Baie Verte Group, and there is a clear and abrupt drop in metamorphic grade across the tectonic contacts between the Fleur de Lys schists, and the ultramafic body and the Baie Verte Group.

The possible age range for the granite-diorite complex is therefore between the cessation of regional deformation in the Fleur de Lys schists, and the Acadian deformation of the Baie Verte Group; that is between lowest Ordovician and mid-Devonian. A K-Ar biotite age of 358 m.y. from granite on Wild Cove Pond (Neale and others, 1960) is very similar to the other ages in the western Fleur de Lys terrain; a K-Ar muscovite age of 355 m.y. from schist in the northeastern part of this terrain (Neale and Kennedy, 1967) and K-Ar dates of 368 and 353 m.y. respectively from the post kinematic granite and lamprophyre dykes on the Grey Islands (Kennedy, Williams, and Smyth, 1973). They could represent uplift ages corresponding with the Acadian deformation of the Baie Verte Lineament. Samples were taken from outcrops of porphyritic biotite granite on the Trans-Canada Highway near Sandy Lake, ground up, and given to I. R. Pringle with a view to obtaining an Rb-Sr isochron, but so far they have not been processed. Therefore there is a possible age range of about 140 m.y. for this intrusive complex, from lower Ordovician to mid-Devonian. However, as Neale and Nash (1963) pointed out, the granite strongly resembles that cutting upper Silurian- lowest Devonian rocks near Sops Arm, to the west across White Bay, and is therefore most likely to be Devonian in age.

C. Burlington Granodiorite.

This rock unit, named by Baird (1951), occupies a large proportion

of the eastern half of the Burlington Peninsula (Fig. 1.3), and occurs along the full length of the eastern boundary of the inland map area. This rock unit is distinctive and varies little in appearance over most of its extent. In the tiny proportion mapped it is usually a chalky-weathering hornblende-biotite granodiorite with a consistent medium to coarse grain size. Subhedral plagioclase is the dominant constituent, and is wholly albitised and/or saussuritised. Some used to be zoned, and it occasionally occurs in glomeroporphyritic clots. Quartz may be granular or poikilitic; minor microperthitic K-feldspar (now altered) occurs in some places, occasionally in sufficient quantities for the original rock to approach adamellite. Green hornblende may be euhedral, but most commonly is not, and may be in excess of or subordinate to green biotite. Muscovite occurs in a few places. Accessory primary epidote, sphene and ilmenite are common, apatite is usually present, and occasional zircons are seen.

Besides the pervasive alteration of plagioclase and K-feldspar, hornblende may be altered to actinolite or, in extreme cases, to a fine-grained chlorite-calcite-magnetite assemblage. Biotite is often partially to wholly chloritised. This pervasive retrogression from the original igneous assemblage is found in most areas of the whole intrusion (Neale and Nash, 1963), and probably occurred during the Fleur de Lys regional metamorphic event.

In the map area, pink aplite veins are common, and include an example of a porphyritic aplite. Xenoliths and pegmatite are absent. Round autoliths of fine-grained granodiorite are seen, sometimes with pink microperthite-rich rims, resembling orbicules. This fine-grained phase is more extensive in one locality (by the Mic Mac Lake Group unconformity, 3.6 km south of the southern end of Flatwater Pond) where it is invaded by veins of coarse-grained granodiorite.

The granodiorite is overlain unconformably by the subaerial volcanic and clastic sequence of the Mic Mac Lake Group (Neale and Nash, 1963), the unconformity dipping west at a medium angle. A cleaved and slightly reddened, highly-altered derivative of the granodiorite occurs locally under this unconformity by the Camp 166 Road. It is interpreted as a palaeoregolith, but is is not common and the unconformity is in most places clean and sharply defined. Northnortheast of Flatwater Pond, the granodiorite has been involved in deformation relating to that in the Baie Verte Lineament, and here is highly altered and cleaved over extensive areas near the contact with the rocks of the Lineament.

The age of the Burlington Granodiorite relative to the Fleur de Lys Supergroup deformation sequence is in doubt. This author examined briefly the eastern contact with the mafic schists of the eastern division of the Fleur de Lys Supergroup on the La Scie Road (Plate 6, Fig. 1.2, Fig. 1.3). This contact is complex and cryptic in detail, due to the presence of meta-gabbroic rock of uncertain affinities, but the essential feature is that the granodiorite contains in places the main schistosity and accompanying strong stretching lineation seen in the adjacent Fleur de Lys schists. However, it is not known if this tectonic fabric is D_1 or D_2 in this area. Thus the minimum age of intrusion of the granodiorite is synchronous with, or prior to D_2 in this area. At this locality, there are no clearcut apophyses of the granodiorite in the mafic schist, and no obvious hornfels or other metamorphic aureole. It is possible that the contact here is a tectonic slide. The syn- or pre-D_1 granodiorite and aplite apophyses that are found in the Fleur de Lys schists adjoining the western side of the Baie Verte Lineament, and which may have been part of the Burlington Granodiorite, also apparently lack a hornfels or metamorphic aureole. The reason

for this is not known. It cannot be proved whether the tectonic fabrics in this area correlate or not

with those near the contact on the La Scie Road. Therefore the sum of evidence for the relative

age of the Burlington Granodiorite is that it was intruded during or prior to the early (D_1 and D_2)

deformation in the Fleur de Lys Supergroup. The author also briefly examined exposures around

Burlington Village (Figs. 1.2, 1.3), where the Burlington Granodiorite invades mostly unfoliated

greenstones and metadolerites of the Nippers Harbour Group (part of the Fleur de Lys

Supergroup). The only deformation observed here in either rock type were local steep foliated

zones. Since these may be due to later deformation, there is no information here on the age of

intrusion relative to the Fleur de Lys deformation sequence. A preliminary Rb-Sr mineral

isochron from samples of Burlington Granodiorite gave an age of 444 ± 20 m.y. (I. R. Pringle,

pers. comm.). This age relies almost entirely on the biotite, as the range of Rb/Sr ratios in the

other minerals and whole rock is very small. As the granodiorite was intruded during or before

the early Fleur de Lys deformation sequence, which is pre- Arenigian (see below), this Rb-Sr

date is probably too young. Pebbles of the Burlington Granodiorite occur in a unit in the Baie

Verte Group (Neale and Kennedy, 1967, and described in the next chapter). The age of the Baie

Verte Group is not known from internal evidence, but it may perhaps be a correlative of the

Arenigian Snooks Arm Group. The latter contains granule-size clasts of 'granodioritic material'

in greywacke (Church, 1969), possibly derived from the Burlington Granodiorite. Thus the

Granodiorite is likely to have been intruded prior to the Arenig. The presence of deformed Fleur

de Lys Supergroup clasts in the Snooks Arm Group (Dewey and Bird, 1971) shows that Fleur de

Lys deformation is also pre-Arenigian.

The lowermost ignimbrite cooling unit of the Mic Mac Lake Group south of the Camp

166 Road (Plate 4) contains rare pebble-size inclusions of

altered chromiferous cumulate gabbro and altered chromiferous ultramafic rock, probably

derived from an ophiolite complex. These inclusions are not seen in any other units higher in the

succession. The unconformity on the granodiorite underneath this ignimbrite is only about 1.6

km from the western margin of the elliptical ring complex that was the caldera source for the

ignimbrites (Neale and Nash, 1963) (Fig. 1.3). No similar members of an ophiolite complex are

known within or adjacent to the granodiorite as now exposed, and therefore it is possible that

these inclusions indicate that the granodiorite, at least in some part of the area now occupied by

the ring complex, was a sheet intruded above lower oceanic crust and mantle, or that large

inclusions of these lithologies were present at depth in the granodiorite when the ring complex

was formed.

D. Summary of Major Events.

The chronology of the major events and relationships in the rock groups bordering the

Baie Verte Lineament in and adjacent to the map area, are summarised below. The age ranges

given are estimates from local and regional evidence and vary in reliability. Some are not

generally accepted.

1a.   Thick psammitic sediments with basalt flows (Seal Cove Group) laid down on

         Grenville(?) gneissic basement, and both intruded by basalt dykes and sills. Age in the

         range late pre-Cambrian to lowest Cambrian.

1b.   Deposition of semipelitic and mafic volcaniclastic sediments (Rattling Brook Group),

         including minor psammitic, agglomeratic, and graphitic sediments, and rare

         orthoquartzite turbidites and chert. An unknown amount of the original stratigraphy is

         missing above and below this sequence in the map area. Age range: some part of the

         Cambrian.

2. Polyphase deformation (D_{1-3}) of Fleur de Lys Supergroup sediments and basement with accompanying regional metamorphism. Age in the range upper Cambrian(?) to Arenig.

    (a) Intrusion of granodiorite apophyses during or before D_1. Intrusion of Burlington Granodiorite during or before main deformation (D_2) in Rambler Mine area.

    (b) D_1; minor isoclinal folds, penetrative schistosity; large westward(?) translational movements on D_1 tectonic slide involving transport of ophiolite lithologies.

    (c) Local intrusion of pegmatite veins during or before D_2.

    (d) D_2; major isoclinal folds, steep penetrative and transposed composite schistosity; local remobilisation of tremolite-fuchsite pods, ? some renewed movement on slides.

    (e) ? Culmination of regional metamorphism, leaving Fleur de Lys schists in most of the map area in biotite or garnet grade, with local chlorite grade to the north, and local staurolite grade in the south. Later retrograde metamorphic effects are generally insignificant.

    (f) D_3; major asymmetric gently plunging folds, crenulation lineation, steep weak strain-slip schistosity/cleavage. Dunamagon granite (and pegmatite?) intruded synchronously with D_3 in Mings Bight area.

    (g) Small conjugate wrench faults.

3. Pebbles of Burlington Granodiorite eroded and deposited in part of the Baie Verte Group. Age (?) Arenig.

4. Celebes Pond Granite and Wild Cove Pond granite-diorite complex intruded, generating an andalusite-bearing aureole that partially overprints Fleur de Lys regional assemblages. Celebes Pond Granite and aureole crosscuts large D_3 folds. Possible age range from lowest

Ordovician to mid-Devonian; on regional similarities probably Devonian.

5.   Mic Mac Lake Group unconformably laid down over Burlington Granodiorite. Age

lowest Devonian (see next chapter).

6.   Tectonic contact developed between ultramafic bodies and Fleur de Lys schists. Local

accommodation kink folds develop near the contacts; dextral faulting on one kink plane

through Marty's Pond develops U-shape of northern Mic Mac ultramafic body. Tectonic

contact between Baie Verte Group and Fleur de Lys schists develops, and cuts aureole of

Celebes Pond Granite. Minor retrograde metamorphism occurs locally in the Fleur de Lys

schists within about 100 metres of the contact. Minor development of strain-slip cleavage

correlative with one in the Baie Verte Group also occurs very locally in the same position.

Age probably mid-Devonian.

This data leaves open the possibility that the single steep cleavage in the Baie Verte

Group is equivalent to D_3 (or even D_2) in the Fleur de Lys schists bordering the western side of

the Baie Verte Lineament. To investigate this possibility it is necessary to find out when the

tectonic contact between the Baie Verte Group and the Fleur de Lys schists was developed as

now preserved, especially where it cuts the aureole of the Celebes Pond Granite in the vicinity of

Kidney Pond. This problem is discussed in the next chapter, after description of the rock units

and their relationships that occur within the Baie Verte Lineament.

CHAPTER 4. ROCK GROUPS OF THE BAIE VERTE LINEAMENT

A. Baie Verte Group - Ophiolite Plutonic Rocks

(i) Introduction

The Baie Verte Group was defined originally as the 'Baie Verte Formation' by Watson (1943, 1947), who examined exposures in the coastal and adjoining areas around Baie Verte and Mings Bight at the northern end of the Lineament (Figs. 1.2, 1.3, and Plate 6). Baird (1951) called these rocks the Baie Verte Group, and all subsequent workers have followed this usage.

Watson described the association of greenstones and concordant bodies of gabbroic and ultramafic rock, recognising 'pseudostratification' in the plutonic rocks, and illustrating cumulate textures from them. The plutonic rocks were described as intrusive, although there is no evidence of aureoles or clear intrusive relationships, and no such evidence has been cited since his work. A large area of mafic volcanic rocks that belong to the Fleur de Lys Supergroup (eastern division) were included in the Baie Verte Group as a more highly metamorphosed equivalent. This mistake was not rectified until Church (1969) recognized that there is a great difference between the structural histories of the two units. Baird (1954) first recognised the existence of the Baie Verte Lineament from the string of large ultramafic bodies running down the centre of the peninsula. The Lineament defined by these and the adjoining mafic rocks contains all the Baie Verte Group as redefined (Fig. 1.3). Neale (1958, 1959b) emphasised the spatial relationship between the ultramafic and gabbroic plutonic rocks in the Baie Verte Lineament, and suggested a genetic relationship between them. Neale and Kennedy (1967) recognised that there is a major but narrow tectonic junction, which they termed the Baie Verte Road Fault, between the western division of the Fleur

de Lys schists (with complex deformation and relatively high metamorphic grade) and the adjoining Baie Verte Group (with a single cleavage and low metamorphic grade) along the inland part of the Lineament. They also reported that clasts of Burlington Granodiorite occur in a conglomerate unit in the Baie Verte Group, and in addition they demonstrated the identity of the minor deformation structures in the Baie Verte and Mic Mac Lake Groups. Church (1969) reported the presence of detrital chromite in the Baie Verte Group, but the locality was not specified. The presence of pebbles of Burlington Granodiorite is the only internal evidence for the relative age of the Baie Verte Group. The lithologic correlation with the Snooks Arm Group (of Arenigian age) is the only justifiable suggestion for the actual age of the Baie Verte Group, so this age is still in doubt.

The Baie Verte Group, as redefined, consists mainly of mafic pillow lava, mostly fine-grained (but with some conglomeratic) mafic volcaniclastic rocks, and dolerite sills. Minor green and maroon chert, green, grey, black and maroon slate and argillite, conglomerate, marble and pink silicic tuff also occur. In addition, the Baie Verte Group is here expanded to include the ultramafic and gabbroic plutonic rocks and sheeted diabase dyke complex of the variably disrupted ophiolite complex (sensu stricto — Geotimes, 1972) that occur within and on the margins of the Lineament. This procedure is justified because 1) this ophiolite complex is shown to have formed the floor to the bulk of the mafic volcanics and sediments of the Baie Verte Group and is an integral part of the stratigraphy; 2) the pillow lavas of the ophiolite complex were originally included in the Baie Verte Group; 3) this procedure has been successfully applied to the nearby ophiolite complex underlying the Snooks Arm Group (Upadhyay and others, 1971). It is admitted that the status of some of the ultramafic bodies in the Lineament

is in doubt, because there is an older set of ophiolite ultramafic bodies that occur in the Fleur de Lys terrain and are involved in its polyphase deformation. However, some are definitely an integral part of the assemblage in the Baie Verte Lineament, and in time those now of doubtful status will be assigned to one or other category. The pillow lavas and mafic volcaniclastic sediments of the Baie Verte Group are aptly described as 'greenstones'. They dip subvertically to steeply west, and are variably deformed in the single steep penetrative cleavage found in the Group. All the rocks in the Lineament are weakly metamorphosed in lowest greenschist facies. Thus the coarser grained plutonic rocks almost always retain their igneous textures, but only clinopyroxene or hornblende may remain from the original mineralogy; plagioclase is ubiquitously altered. Such rocks should technically be called 'metagabbro' , 'metadolerite', etc., but as their original affinity is almost always apparent, the terms gabbro, dolerite, etc. , are used for convenience, regardless of the variable alteration. The originally finer grained volcanic rocks are all pervasively altered, though not necessarily deformed, so they are given the noncommittal nomenclature of mafic lava, mafic volcaniclastics, etc. The Baie Verte cleavage is near coplanar to all the three near coplanar schistosities in the adjoining Fleur de Lys schists, and it might be correlated with any one of them. As any such correlation is not obvious, and may not exist, the structural abbreviations for the cleavage, folds and crenulations in the rocks of the Baie Verte Lineament are kept strictly separate from those for the Fleur de Lys sequence. The one regional cleavage in the rocks of the Baie Verte Lineament is termed S_{1B}, the deformation responsible. D_{1B}, and the accompanying folds F_{1B}. The abbreviations for local secondary structures (crenulations) are defined later, when needed.

The Baie Verte Group in the Mings Bight area is described separately in a later chapter.

4.4

The term ophiolite suite has often been misused since it was proposed by Steinmann (1926) for the common <u>association</u> of ultramafic rocks, gabbro, pillow lava and chert found in Mesozoic and Tertiary mountain belts. Dietz (1963) proposed that this association is derived from the oceanic crust and upper mantle. The description of the sequence and relationships of the rocks found in undisrupted ophiolite complexes, and shown to be unlikely to be anything other than that forming the oceanic crust and upper mantle (Moores and Vine, 1971, Dewey and Bird, 1971), has led to reexamination of the relationships of the ophiolite association wherever it occurs. A recent conference (Geotimes, 1972) agreed to a definition of the term ophiolite complex. The complete undisrupted sequence ranges from depleted Mg-rich non-cumulate ultramafic rocks at the base, passing upward through cumulate ultramafic rocks to at least partly cumulate gabbro, and then through gabbro cut by parallel diabase dykes into a 100% sheeted diabase dyke complex that feeds pillow lava. Sediments cap the pillow lava, and they are usually cherts or other pelagic sediments. The conference also agreed that the term ophiolite complex, when suitably qualified, can be used for the more common occurrence of these lithologies in belts of tectonically disrupted, and even metamorphosed, fragments. The essential character is the association of all the lithologies. If a block of 100% sheeted diabase dykes is found, this is fairly diagnostic by itself, as the sea-floor spreading process is the only one known to produce more than small cross-strike widths of 100% dykes. However, it seems that sheeted dykes are either very difficult to recognise or that they are rarely preserved in disrupted ophiolitic associations, as they are only described from a few well preserved ophiolite complexes. The non-cumulate ultramafic rocks are also diagnostic, in this author's opinion, as they lack cumulate

textures, and have a depleted, Mg-rich, composition distinct from the cumulate ultramafic rocks of lopoliths. small Alaskan-type zoned complexes, and small intrusive bodies (e.g. Skaergaard). This fact was first documented by Hess (1938). In the past, it was often assumed that the plutonic rocks, especially the ultramafic rocks, were necessarily intrusive in their present position. Most ultramafic and gabbroic members of ophiolite suites are no more intrusive in their present position than an outlier of granite on a thrust plane, or isolated in a fault block. However, it is true that the faults bounding ophiolite plutonic rocks are difficult to recognise, and that the tectonic slivers may appear superficially concordant and in place.

The Baie Verte Lineament contains a number of large ultramafic bodies, most but not all of which occur on its western side (Fig. 1.3), and also a number of smaller bodies, and derivatives of ultramafic rock. These rocks were shown to be of the depleted Mg-rich 'alpine' (or ophiolitic) type by Watson (1943). In the inland map area, the whole of one and part of another of these large bodies of ultramafic rock were mapped, as well as several other small occurrences. Small, but significant bodies of cumulate ultramafic rocks, ophiolitic gabbro with diabase dykes, and sheeted diabase dykes are also found in the inland map area. Mafic pillow lava is an abundant component of the Baie Verte Group in the inland map area, but it is not part of the disrupted ophiolite complex, as it occurs conformably above a significant thickness of sediments.

(ii) Depleted non-cumulate ultramafic rocks

(a) Distribution

Most of the data on the nature of the ultramafic rocks of the inland map area were gathered from outcrops in the Flatwater ultramafic body (Plate 2). Some additional information was derived from the part of the Mic Mac ultra-

-mafic body that was mapped (Plate 1). Other minor bodies just north of Kidney Pond and at the northern end of the map area did not contribute additional data. Highly-altered derivatives of ultramafic rock that are more widely distributed along the contact between lavas and sediments of the Baie Verte Group and the western Fleur de Lys schists are described in a later section.

The large ultramafic bodies (allowing for the fault modification of the Mic Mac body) have the typical lensoid, or 'orange pip' outline of 'alpine- type' bodies. The smaller ones tend to be longer relative to their width, and a transition between the two is seen in the northern part of the Flatwater body. Disruptive tectonic effects only occur very near the margins of the large bodies; almost all of their interior is unaffected by this marginal deformation and retains undeformed the original megascopic plutonic textures. Serpentinisation is more widespread, and is most complete in a zone around the outer part of the large bodies. Some of the original microscopic plutonic textures are progressively obliterated as serpentinisation proceeds, but some textures are preserved pseudomorphed by different serpentine habits. The smaller ultramafic bodies are completely serpentinised.

(b) Primary textures and mineralogy

Ninety-five percent or more of the ultramafic bodies mapped consist of magnesian harzburgite (saxonite); the remaining 5% or less is magnesian dunite which occurs in bands with sharp margins against the harzburgite. Chromitite bands are very rare and small.

The harzburgite in outcrop is orange-brown weathering with conspicuous ovoid orthopyroxene crystals up to 10 mm (commonly about 2 mm) across spotting the surface, and comprising between 5 and 30% (commonly about 20%) of the rock. The dunite has a smooth featureless orange-brown weathering surface, and occurs as bands commonly between a centimetre and a metre wide,

though a very few small outcrops consisting wholly of dunite probably indicate the presence of wider bands.

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

In addition to the fairly common occurrence of chromite seams one grain thick in dunite bands, very uncommon chromitite bands are also found. The examples seen, on Middle Arm Brook [20] and in two other localities 1.2 [21] and 1.3 km north of the Burlington Road junction, are bands 1 to 2 cm thick, not pods, although one band in outcrop [20] has parts where semi-ductile boudins tend towards the typical 'podiform' chromite (Fig. 4.1a). A chromite lens in the outcrop 1.4 km north of the Burlington Road junction, reported as 3 metres long by 70 cm wide (Neale, 1958) is no longer exposed. Chromite grains are typically between 0.05 and 0.8 mm on the edge of the bands, but they rapidly become amalgamated within the band into a solid mass of chromite, suggesting considerable grain growth during and perhaps shortly after formation. In specimens from outcrop [20] a gradational sequence can be seen from examples of seams of euhedral chromite one grain thick to examples of thicker bands, with amalgamation between grains becoming more pronounced in thicker examples.

Fig. 4.1 Banding in non-cumulate ultramafic rocks.
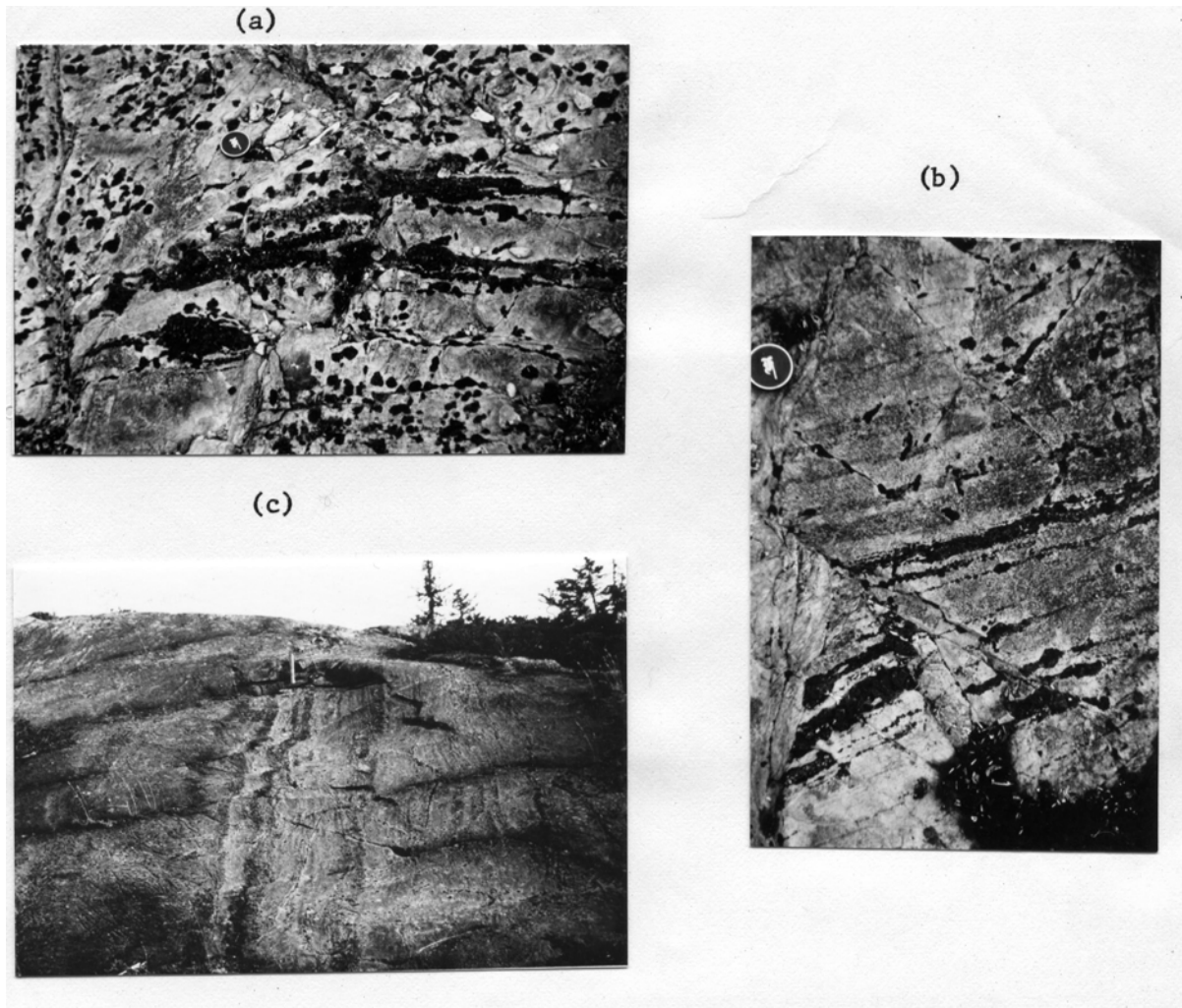
(a) Chromitite bands with semi-ductile boudins; Middle Arm Brook [20].

(b) Clinopyroxenite bands (grey) and chromitite bands (black) in dunite; Middle Arm Brook [20].

(c) Orthopyroxenite bands in harzburgite [22].

4.8

Small tension cracks (ac) are common in the chromite bands, and although most seem to have developed during serpentinisation, some in the bands in outcrop [20] have olivine in them, indicating brittle deformation of the chromite band during the high temperature plutonic processes generating the ultramafic rocks. Clinopyroxene (diopside or diopsidic augite) is seen in a few thin sections as lensoid exsolution lamellae in bastite. In one outcrop [20] clinopyroxenite bands 1 to 5 cm thick are found paralleling adjacent chromitite bands, all in dunite (Fig. 4.1b). Clearly-defined orthopyroxene-rich bands are also seen only in one outcrop [22], SSE. of Red Cliff Pond, where they are up to 10 cm wide (Fig. 4.1c). Otherwise, gradual variations in orthopyroxene content of the harzburgite are difficult to map objectively. This content ranges from about 5 to 30% on thin section estimates. However, it is most commonly in the range 15 to 20%, and there is a definite break in the range of compositions of the lithologies between pyroxene-poor harzburgite and dunite.

Pegmatitic harzburgite is seen in outcrops 0.8, 0.9 and 1.8 km [23] south of the outlet of Flatwater Pond along the power transmission line. Grey bastite pseudomorphs after enstatite up to 5 cm, most commonly 2 cm across, set in serpentinised white weathering rock, are found in irregular patches, and in the southern outcrop [23] in lensoid bands with diffuse margins in white weathering normal serpentinised harzburgite. In the northern outcrops some of the orthopyroxenes are strongly kinked.

Also in outcrop [23] a probable xenolith (strictly autolith) of cumulate lherzolite and harzburgite is found. This occurs about half way along the eastern edge of the outcrop and is isolated near a 'corner' made by three orthogonal joint faces. The reasons for calling this unusual rock a xenolith is that the cumulate rock, which appears to form a band on its isolated corner,

is not traceable into the main body of the outcrop, and is therefore a lump apparently contained within non-cumulate rock. The exposed part is about 30 cm across, and displays large green subhedrai poikilitic chrome diopside crystals up to 5 cm long, that enclose slightly rounded and resorbed olivine crystals. These occur in a layer where they comprise up to 30% of the rock. The adjacent harzburgite also seems to be a cumulate rock from thin section examination, but it was not possible to differentiate this from the non-cumulate harzburgite in outcrop. Thus the boundary of the xenolith away from the layer containing diopside is difficult to define. In thin section, the cumulus phases are olivine and (now serpentinised) orthopyroxene in the ratio about 4 1, with an accessory opaque ore mineral, possibly altered chromite. Very rare small clinopyroxene crystals are enclosed in cumulus orthopyroxene. Postcumulus processes were growth of poikilitic chrome diopside, probable slight resorbtion of orthopyroxene, and slight resorbtion of olivine where in contact with diopside, but slight overgrowth elsewhere. Olivine grains are 1 to 12 mm, usually 6-8 mm across, orthopyroxene are 0.4 to 4 mm, usually 1 to 2 mm across, and the chromite (?) 0.1 to 3 mm, typically 0.2 to 0.5 mm across. Olivine composition is in the range Fo_{80-90}, probably about Fo_{85}. The diopside may have orthopyroxene lamellae parallel to [010]. This rock is fairly typical of cumulate rocks described from the transition zone between ultramafic and gabbroic rocks in ophiolite complexes (see Wilson, 1959). However, the range of cumulate grain sizes is rather large when compared with cumulates described from layered complexes (e.g. Jackson, 1961) and the postcumulus relationships of the olivine seem to be equivocal.

The dunite and other bands, and a high temperature foliation, are sub-parallel over most of the interior of the wide part of the Flatwater body, (Plate 2), generally trending between NNE-SSW to NE-SW and dipping subvertically.

4.10

In the part of the Mic Mac body mapped, these fabrics are not so consistently aligned.

Within both the Flatwater and Mic Mac bodies, a high temperature foliation is variably developed, generally in local zones. It is most intensely developed in some dunite bands, (but dunite is not necessarily foliated) and the foliation is widespread, though not as intense, in harzburgite. A millimetre scale foliation is seen on the weathered surface of rocks in which selective serpentinisation has occurred, but it is not seen on the weathered surface of very fresh rocks. On freshly broken surfaces, it is generally not apparent unless the rock is nearly or wholly serpentinised. Thus observation of this foliation in outcrop is wholly dependent on the amount and type of serpentinisation that has occurred. However, as outcrops of rock that are too fresh for the foliation to show on the weathered surface are very rare, the observations made in the areas mapped are thought to be representative.

The most intense foliation was found in a specimen from outcrop [24] at the northwestern end of the Mic Mac body, and in a dunite band in an outcrop just south of Middle Arm Brook [25]. In these examples, a very fine grained mylonitised dunite contains a few extremely elongate bastites (after enstatite). However, a few ovoid, undeformed bastites also occur, apparently unaffected by the intense deformation, and they are presumed to have overgrown the fabric after the deformation. The olivine ($Fo_{90\pm}$) occurs as tiny equant grains down to 0.01 mm, and generally not larger than 0.5 mm, most commonly about 0.1 to 0.2 mm across. They show a high degree of crystallographic orientation, with one or two statistically preferred extinction directions in any thin section, generally at a low angle (~15°) to the foliation. In addition, they show a high proportion of grains with either

high, or medium, or low maximum birefringence in each of the three sections cut orthogonal to the foliation (and lineation where present). Individual grains do not generally have extremely elongate shapes, although many have a small elongation in the foliation. In these strongly deformed rocks, individual strings or chains of olivine grains parallel to the foliation very commonly have simultaneous, or nearly simultaneous extinction and very nearly or exactly the same birefringence; that is more so than surrounding grains. This suggests either derivation by deformation of a single pre-existing larger grain, or nucleation from a single source. The former explanation is preferred. The olivine has suffered a certain amount of grain growth subsequent to the deformation. The grains are annealed, and there is a definite lower limit to the grain size, suggesting that finer grains have been absorbed. A few grains have amalgamated, leading to optically continuous grains elongate along the foliation and of larger dimensions than normal. Most examples of foliated rock have a flattening fabric (S-fabric). In all except two examples of these strongly foliated rocks, the accessory chromite is equant and often sub- to euhedral, and has perhaps recrystallised after the deformation, although this need not be the case. However, in samples of dunite from outcrop [20] on Middle Arm Brook, some of the chromites are deformed in the foliation. The grain size of the olivine in this rock is commonly in the range 0.1 to 1 mm, somewhat larger than in the previous examples. However, both smaller and larger grains do occur, in one case to 8 x 2 mm elongate in the foliation. The olivine shows preferred extinction directions, birefringence, and elongate 'grain trains', as in the previous examples. Some of the chromite grains in this rock are flattened and brittly boudinaged in the foliation (Fig. 4.2a). Olivine is present in the gaps between the parts of the boudinaged grains, in some cases optically continuous with an

Fig. 4.2. Boudinaged chromite grains in dunite with intense sub-magmatic high-temperature foliation.
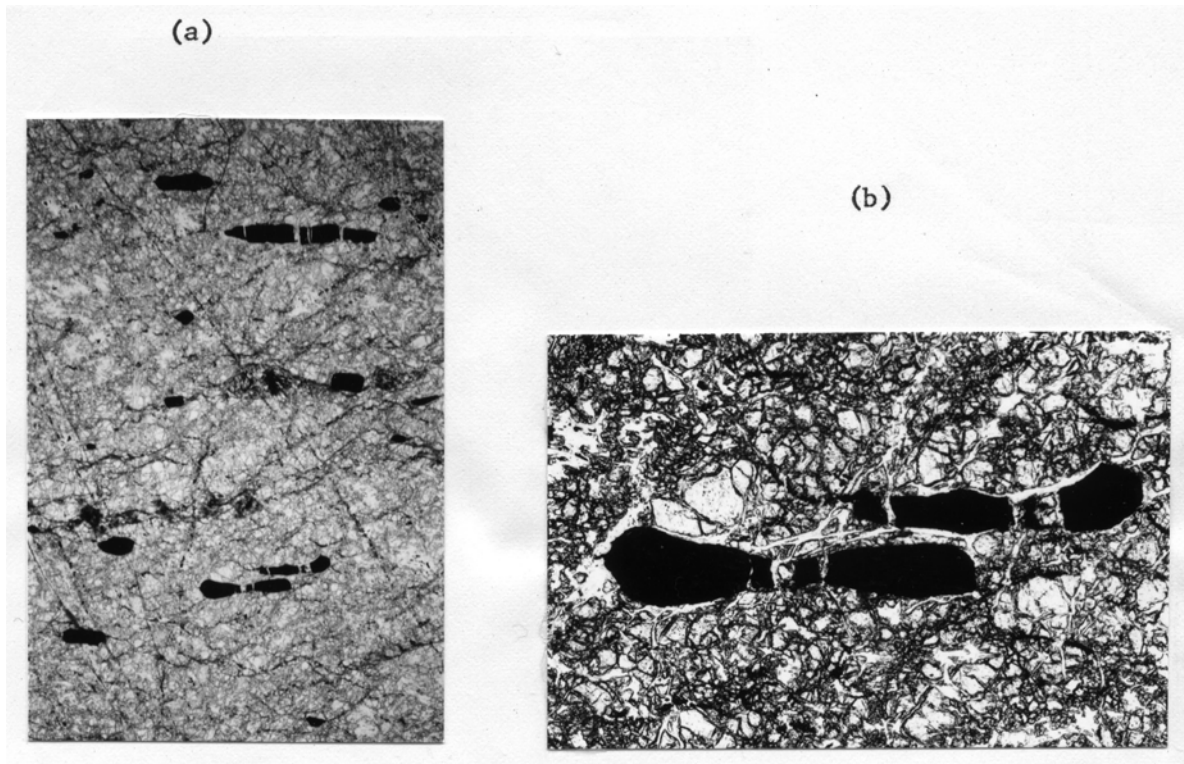
(a)   General view (X 10). Middle Arm Brook [20].

(b)   Detail—olivine in boudinage gaps (X 40)

olivine grain outside the boudinage gap (Fig. 4.2b). This shows clearly that the tectonic fabric in this rock was formed at high temperature, in general terms at the same time that the ultramafic rocks were being formed. Some of the chromite grains in this rock are euhedral, perhaps indicating recrystallisation after deformation. Also the olivine is partly annealed. The chromite bands in this rock show both semi-ductile pinch-and-swell structure and brittle tension cracks with olivine in them. The clinopyroxenite bands also occur near the chromite bands. They contain both large, often kinked, ovoid crystals of diopside or diopsidic augite 1 to 2 mm across, set in a finer granular matrix of the same pyroxene. averaging 0.2 mm across. The contact of the clinopyroxene band with the dunite is sharp, and the bands show weak pinch-and-swell structure. These bands, therefore, also show the effect of deformation, granulation and some annealing recrystallisation.

Very highly deformed rocks such as these are not common. Rocks showing grain sizes intermediate between these and the normal common harzburgite are also very uncommon. The two samples examined show that the bastites are of normal grain size, although often elongate and kinked, while the olivine is finer grained than in the normal rocks. Rocks of normal grain size that show a high temperature foliation on the weathered surface usually have strongly deformed elongate (flattened) and kinked bastites, and while the olivine is also often kinked and strained, it shows no particular preferred shape orientation. Not all the bastites in any sample are deformed; some are normal ovoid equant crystals. Even in rocks that do not show a megascopic foliation, it is common for some of the bastites to be kinked, and occasionally elongate. Olivine in these apparently undeformed rocks is more often than not kinked, twinned, and slightly strained. In rocks of normal grain size, whether they possess a megascopic foliation or not, the

olivine quite often shows a preponderance of high or low maximum birefringence, and one or two preferred extinction directions. This is usually apparent only when the thin section is examined at very low power. Thus the coarse- grained rocks also show evidence of some high temperature deformation.

Olivine is very easily deformed at high temperature, as illustrated by a small slump in a cumulate dunite band from the Stiliwater complex that shows kinked, twinned, and slightly strained olivine (Jackson, 1961).

The deformed bastites in the rocks of the map area indicate somewhat stronger deformation in quite a substantial proportion of the ultramafic rocks. Further, it is clear that all the non-cumulate ultramafic rocks have undergone some high temperature deformation, and the amount of recrystallisation that accompanied this is not known. Therefore it is not clear how much the textures of the apparently least-deformed rocks reflect their original plutonic textures. Although some of these rocks show ophitic relationships between some of the olivine grains, the typical xenomorphic granular texture shows no evidence of cumulate origin.

The zones of megascopically foliated rock range in width from a few centimetres upward to 0.5 km. However, all the examples of intense foliation are zones not more than a few metres wide. The zone 0.5 km wide contains megascopically foliated but mostly coarse grained rock. It has been marked on the map of the Flatwater ultramafic body (Plate 2), and data from an unpublished map belonging to Advocate Mines Ltd. has been used to define the southernmost part of its eastern boundary.

In all but one example, the foliation is parallel to lithologic banding in the same outcrop, or subparallel to that in adjacent outcrops. The fairly consistent orientation of foliation and banding over the main part of the Platwater body is parallel to the orientation of a transition zone between

Fig. 4.3. Refracted high-temperature foliation in dunite band.
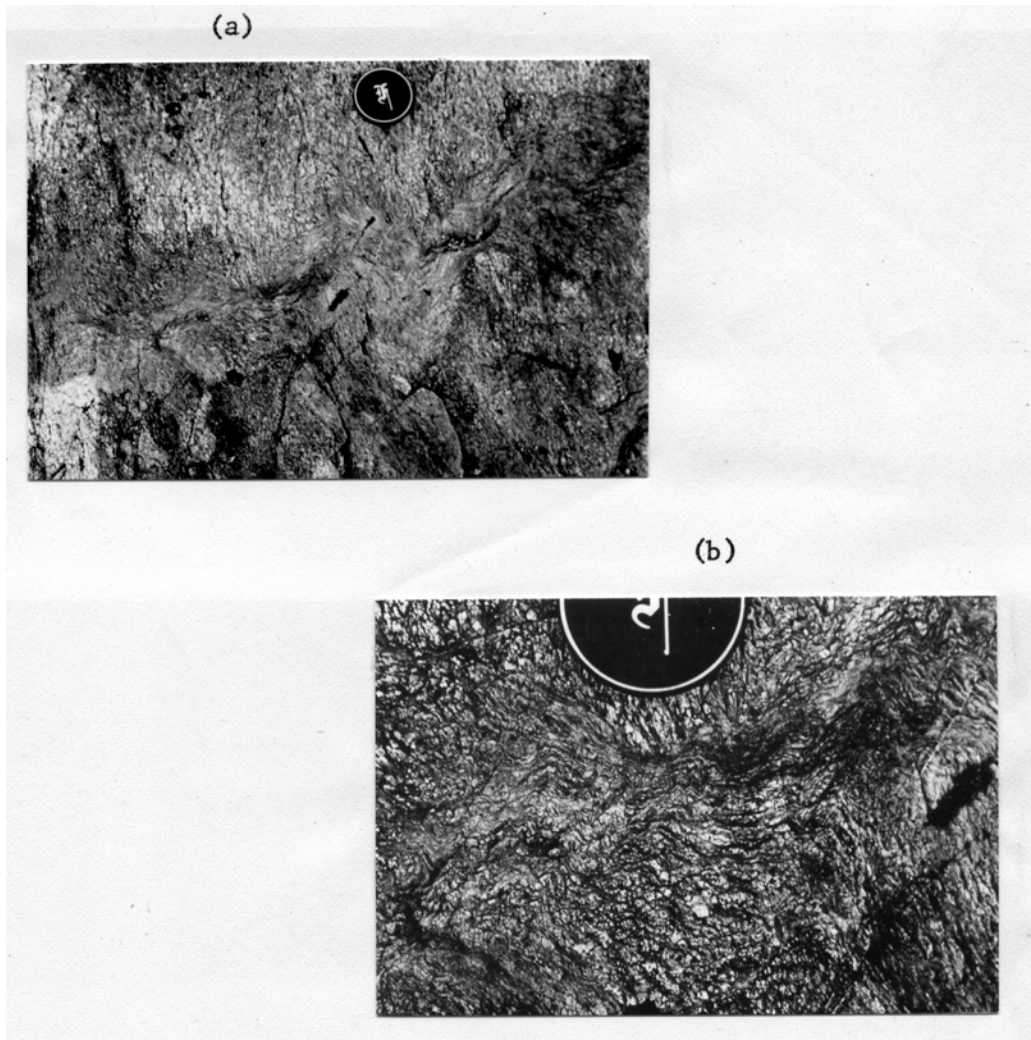
(a) General view

(b) Detail, showing crenulated high-temperature foliation.

4.14

ultramafic and gabbroic rock that occurs at the southeast end of the Flatwater ultramafic body. The cumulate rocks in this transition zone must have been originally horizontal, and therefore it is most probable that banding and foliation in the non-cumulate ultramafic rocks were also originally sub-horizontal.

One small example of weakly folded foliation and foliation oblique to a band is seen in outcrop [22]. At the north end of this outcrop the high-temperature foliation parallels orthopyroxene-rich bands. However, in the centre of the outcrop, this foliation enters a dunite band that is locally at a high angle to it. The foliation is abruptly refracted and crosses the band at a small angle and resumes its former orientation on leaving it. However, the band and the subparallel foliation within it are moderately crenulated, and the same foliation outside the band is axial planar to this crenulation (Fig. 4.3). This example indicates folding of a dunite band where locally slightly oblique to the original foliation, probably followed by some simple shear along the dunite band, followed by mild crenulation resulting from further flattening on the same foliation outside the band. This minor example does not justify the term 'polyphase deformation', although it conveniently illustrates how apparently polyphase deformation can result from continuous, but locally inhomogenous, deformation. Otherwise, no more than one high temperature foliation is present in the ultramafic bodies mapped.

(c) Serpentinisation and other alterations

In this thesis, a clear distinction is made between serpentinised ultramafic rock, and serpentinite. The former term is used for rock that may be totally serpentinised, but is undeformed and can be assigned to its original composition from its remaining pseudomorphic textures. The term serpentinite is strictly reserved and precisely used for rock that has been deformed and/or

recrystallised, and is no longer attributable to its original igneous composition. Serpentine is used exclusively as a mineral name. Alteration products of ultramafic rock containing more than 50% Mg-rich carbonate are not called ultramafic rocks, but are referred by their mineral constituents.

Rocks from the non-cumulate ultramafic bodies mapped range from about 10% to wholly serpentinised, but only in a very few exposures are they less than 50% serpentinised. In the main, the exposures are either chalky-white to grey weathering, or orange-brown weathering. The weathering colour reflects the general amount of serpentinisation, and white/grey weathering rocks are always 95% or more and, usually 100% serpentinised. The transition between white and orange weathering areas is nearly always a narrow zone, less than 100 metres wide, in which the rocks are orange weathering with flecks of white or *vice versa*. This zone has been marked by a dotted line in the Flatwater body (Plate 2). In the northwestern prong of the Mic Mac body, white weathering rocks are confined to the northeasternmost two or three outcrops, so it is inappropriate to draw such a boundary in this area, even though a few of the outcrops west and southwest of Marty's Pond have mixed white and orange transition weathering and are 90% serpentinised. All the outcrops in the northeast prong of the Mic Mac body, and those around the bottom of the 'U' south of Marty's Pond, are white weathering. Within the areas of fresher, orange-brown weathering rocks in both bodies, highly serpentinised rock occurs along joint planes, and the alteration is usually not more than a few centimetres wide. When these joints from part of the surface of an outcrop, they have a characteristic gouged and cracked appearance, referred to informally as 'rhinohide'.

In the main part of the Flatwater body (Plate 2) it is significant that

the width of the zone of white-weathering 95%+ serpentinised rock is very such narrower on the western side than on the eastern side. The northern mapped part of the Mic Mac body shows the same asymmetric distribution, but with even stronger contrast, when the displacement on the Marty's Pond fault is removed. This asymmetric distribution of serpentinisation shows that the water content of the Baie Verte Group was much greater than that of the Fleur de Lys schists when the ultramafic bodies were emplaced into their present position. Also it is possible that the relative widths of the western highly serpentinised zones are not unconnected with the higher metamorphic grade (garnet, andalusite, ± staurolite) of the Fleur de Lys schists adjoining the wider part of the Mic Mac body compared with that (biotite-chlorite) in the schists adjoining the Flatwater body.

Uncommon alterations visible in outcrop include red smooth-weathering rock, that is always found in the white and transitional weathering zones of the Flatwater body near Middle Arm Brook, and in the thin ultramafic body at the northern end of the inland map area (Plate 2). This red-weathering rock is wholly serpentinised, and is distinct from the surrounding rock only because it contains a relative abundance (~ 30%) of Mg-rich carbonate. In a small elongate area between 0.5 km north and south of Middle Arm Brook 0.15 km west of where the Baie Verte Road crosses it, this rock occurs in sharply defined subparallel veins between 1 and 20 cm thick. Some of these have shallowly diverging apophysae, and some include fragments of the white/grey weathering host rock, and thus they greatly resemble dykes. However, in one locality a thin chromite band is seen crossing a red vein at right angles. This observation, and the thin section data show that these red weathering veins are not dykes, but an alteration effect.

In thin sections of white and orange-weathering normally serpentinised rocks, the serpentine replacing olivine is almost always a randomly oriented

rather coarse-grained form-antigorite identical to that illustrated by Francis (1956). Two samples out of about fifty sectioned show mesh-serpentine. Enstatite is always replaced by an ultrafine-grained fibrous serpentine, where the fibres are coaxial with the intersection of the two pyroxene cleavages. This type of pseudomorph is conventionally termed bastite. The initial replacement is a nephritic mineral, brown in transmitted light and a characteristic porcellanous green in white weathering rocks in outcrop. This is then replaced in the most serpentinised rocks by a colourless fibrous serpentine mineral with normal grey birefringence. Chromite almost always has some alteration, but the relative amount of alteration does not correspond with the degree to which the host rock is altered. The alteration starts from a thin skin of opaque material around the rim and bordering cracks in the chromite grain, whose fresh interior may be anywhere from light brown through burgundy to nearly opaque reddish-brown in colour. It progresses by the patina advancing inwards into each piece of each grain until the whole is opaque. In reflected light the opaque phase shows a bluish cast that the dusty black reflecting chromite does not. It is possible that the opaque alteration is magnetite. No chrome-bearing alteration mineral has been positively identified around partly or wholly altered chromite grains in serpentinised ultramafic rock or serpentinite. However, in some thin sections a 'halo' effect is seen in the serpentine around chromite; under crossed polars a ring of abnormal blue birefringence occurs a short distance away from the edge of the chromite grain, separated by an isotropic ring from anomalous brown birefringent colourless fibrous serpentine adjoining the chromite grain, with the fibres arranged radially. This is perhaps due to diffusion of some Cr away from the grain, but it seems likely that even the wholly altered opaque grains still retain an appreciable proportion of their original Cr content. Very fine grained magnetite dust is an ubiquitous

product of serpentinisation, and tends to be concentrated along some of the original olivine

grain boundaries, and in the bastites. Scattered crystals of Mg-rich carbonate, are found in

some thin sections of the highly serpentinised white-weathering rocks, and a small amount of

brucite has also been identified in some samples. Talc has not been seen in the little to

undeformed serpentinised ultramafic rocks.

A sample of brown-weathering harzburgite showing blackish orthopyroxene was

collected 0.4 km SW of where the Bale Verte Road crosses Middle Arm Brook [36]. It is or

normal coarse grain size, and contains about 10% brown bastite pseudomorphs after

orthopyroxene. The olivine is only about 5% serpentinised. The bastites are partly to wholly

replaced by olivine. In any one bastite, several grains of olivine with different optic

orientations are present and tend to have relatively long straight grain boundaries parallel to the

bastite fibre direction and short straight boundaries across the fibre direction. The composition

of the replacement olivine is not detectably different ($Fo_{90\pm}$) from that outside the bastite in the

sample and in the rest of the ultramafic body. Many of the other samples examined in thin

section from the area between Middle Arm Brook and the Bear Cove Road show a colourless

to brown acicular mineral, with cracks crossing the elongate grains, in some of the bastites. It

has a maximum second order yellow birefringence, but has mostly oblique extinction and is

probably cummingtonite. When fine-grained, this is very difficult to distinguish from possible

fine-grained olivine. Therefore, while the one sample described above definitely has olivine

replacing a nephritic bastite, it is not certain how widespread this process is. The examples of

(?) cummingtonite in bastite appear to form a sequence, seen in different bastites, grading by

decreasing fibre size into the brown nephritic fibrous bastite mineral which could also be

micro-crystalline cummingtonite.

4.19

It is observed that in all except one outcrop in the inland map area, orthopyroxene (enstatite) in the non-cumulate ultramafic rocks is completely altered to nephritic bastite. This is wholly independent of the amount olivine is serpentinised. It is the colourless, grey birefringent, lizardite bastite replacement of nephritic bastite that is correlated with highly serpentinised, white weathering rocks. The fact that enstatite is replaced while olivine is relatively unaffected is unusual because the majority of descriptions known to the author speak of orthopyroxene rather than olivine resisting serpentinisation. The clear evidence of olivine overgrowth of bastite in the one sample indicates a temperature locally greater than 400+°C and enstatite is not stable in the presence of water vapour <u>above</u> about 850°C, whereas olivine is stable under the same conditions (Deer and others, 1966). Therefore, this evidence from the ultramafic rocks of the Flatwater and Mic Mac bodies seems to indicate access of water to the upper oceanic mantle while it was still at high temperature close to the spreading ridge that generated it. This could happen by a fault or faults penetrating the upper oceanic mantle. These would most probably be transform faults as it is unlikely that the rift faults forming the axial valley of spreading ridges cut any more than the upper part of the oceanic crust. Alternatively, the faults could be block faults formed outside the axial valley, such as those with diapiric serpentinite on them, described from the 45°N area just west of the mid-Atlantic Ridge (Aumento, Longcarevic, and Ross, 1971).

Within 100 metres, and usually less than 20 metres of the external contacts of the ultramafic bodies, and within a few metres of internal faults, there is a progressive tectonic disruption of the serpentinised ultramafic rock. The style of deformation is distinctive and characteristic of ultramafic rocks, and is mainly due to the extreme ductility contrast between serpentinised ultramafic rock and derived serpentinite. The progressive

4.20

deformation follows a sequence that starts with the production of 'shear polyhedra' (Jahns, 1967). Blocks of the serpentinised ultramafic rock defined by joint and conjugate shear planes start to move with respect to each other, and serpentine recrystallises in a fibrous form (slip-fibre) subparallel to the planes. As movement continues, the serpentine on the slip planes is deformed, granulated and recrystallised, and some granulation of the surfaces of the blocks of the ultramafic rock may occur, and the finely granulated material is recrystallised and incorporated into the serpentinite matrix. Further deformation leads to flattening and breakdown of the blocks of serpentinised ultramafic rock by formation of more brittle extension cracks and movement on newly-developed conjugate shear planes. This flattening deformation (with or without a preferred extension direction) leads to a progressive rotation of slip planes towards the plane of flattening and the recrystallisation of serpentine minerals to form a foliation in this plane. A component of simple shear deformation is usually inferred to be important in these deformed zones, and may well be dominant, although a pure flattening deformation will produce similar results. Under simple shear, an open-ended rotational stress is applied to the brittle blocks of ultramafic rock, that will probably result in their dissolution being faster than under a pure flattening stress, where the rotational stress diminishes as the long axis of the block approaches the plane of flattening. Further deformation produces a rock with a larger volume of foliated serpentinite matrix and a smaller volume and smaller fragments of ultramafic rock. The matrix serpentinite itself is deformed by simple shear, and becomes separated into thin scaly plates of glassy surfaced serpentinite (commonly referred to as fish-scale serpentinite). Remaining small (mostly pebble-size) lumps of serpentinised ultramafic rock are more rounded, and are surrounded by envelopes

of glassy serpentinite, forming objects shaped like mango seeds (phacoids). The end member of the deformation process is reached when all the brittle lumps of serpentinised ultramafic rock have been destroyed, and the rock is a schistose serpentinite. This is only found in a few localities in the map area, and then only over widths of about a metre or two in some outcrops at the external contacts of the ultramafic bodies. Good examples of parts of the progressive deformation process can be observed in a roadcut just south of Middle Arm Brook [26] (shear polyhedra), in an outcrop 0.9 km N of the Burlington Road Junction [27] (progressive from little deformed rock through to fish-scale serpentinite), and in outcrops on the western contact of the Flatwater body on Middle Arm Brook [28] (Schistose and fish-scale serpentinite). This progressive deformation sequence does not involve any diapiric behaviour of serpentinite which leads to a different rock type, described in the chapter on the Mings Bight area. Rocks from locality [28] show an intense foliation on the weathered surface. In thin section most specimens show a statically-recrystallised, randomly-oriented, aggregate of flaky form-antigorite and the foliation is only shown by parallel streaks and trails of magnetite dust and elongate trails and lensoid veins of Mg-rich carbonate crystals. One thin section shows an aligned flaky antigorite fabric partly recrystallised to a randomly oriented flaky antigorite fabric. In several sections, occasional partly to wholly-altered chromite grains display augen defined by magnetite dust trails. Most of the rock in this locality contains small pebble-size lumps of undeformed but thoroughly-recrystallised, serpentinised, ultramafic rock. These are very difficult to see in these particular outcrops, perhaps due to the recrystallisation. The foliation in this locality is commonly cut by microshear surfaces at a low angle to the foliation. There are also a few small-scale asymmetric open folds of the foliation.

4.22

The deformation here and in other localities on the external contacts of the ultramafic bodies are discussed with reference to the tectonic history of the area in a later section.

In some areas showing mild to moderate development of shear polyhedra deformation, veins of cross-fibre asbestos (chrysotile) generally not more than 5 mm wide are found. These are only abundant in an area 0.25 km west of the Burlington Road Junction, and this has been trenched and drilled by the Advocate mining syndicate and found to lack economic potential. In the quarry just west of the Baie Verte Road opposite the Burlington Road Junction, veins of bright yellow fibrous 'picrolite' (supposedly antigorite) are abundant. Around 0.5 km along Middle Arm Brook from the Baie Verte Road, veins of porcellanous green 'nephrite' up to 1 cm wide occur. The chrysotile, 'picrolite', and 'nephrite' veins all show growth of parallel fibres of a serpentine mineral in the direction of opening of the vein. They also show a microbanding paralleling the margins of the vein, indicative of many individual increments of opening and fibre growth. The reason why chrysotile grows in one area but 'picrolite' or 'nephrite' in another is not known. However, it is clear that the growth is controlled by the rate and direction of opening, and 'cross-fibre' (where the fibre is at a high angle to the walls of the vein) is distinct from 'slip-fibre' (where it is at a low angle to the walls) only in the direction of opening relative to the orientation of the vein, but the fine silky chrysotile asbestos that is economically important does not occur as slip fibre, and the reason for this is not known. In one outcrop 0.3 km west of the Burlington Road Junction two parallel bands about 10 cm thick and about a metre apart are defined by myriads of thin (~2 mm) *en echelon* tension gashes filled by chrysotile asbestos, and each tension gash has smaller subsidiary tension gashes. It is possible that this example reflects a control by pre-existing igneous banding over

the formation of the tension gash veins. Elsewhere a consistent vein orientation or direction of opening was not seen over an area larger than a square metre within any outcrop.

Parts of two internal fault zones show Mg-Fe carbonate-quartz alteration of the ultramafic rock. Both are near the eastern contact of the Flatwater body, one south and one north of Middle Arm Brook. The southern example is better exposed. Serpentinised ultramafic rock becomes rather abruptly and fully replaced by an 80% carbonate-20% quartz assemblage about 60 metres north of the fault, but only within 10 metres south of it in an exposure gap. The replacement of the serpentinised ultramafic rock is selective, and quartz tends to be concentrated in and near the sites of the bastites, so that an imperfect pseudomorphic texture results. The rough, rusty-weathering rock begins to become foliated with elongation and flattening of the quartz and carbonate crystals within a metre or so north of the fault zone. First the weak foliation dips gently north, and then quickly turns over through horizontal to dip steeply south to vertical and becomes very intense just after passing through the horizontal. This behaviour is exactly that of a shear zone (*sensu stricto*) Ramsay and Graham (1970). Other carbonate-rich alteration products of ultramafic rock occur along the eastern margin of the Flatwater body and elsewhere on the Baie Verte-Fleur de Lys tectonic contact, but as they are closely connected with the development of that tectonic contact, they are discussed in a later section.

The source of the CO_2 in the Mg-Fe carbonate found in small quantities throughout the more serpentinised parts of the ultramafic bodies, and forming most of the rock in some tectonic movement zones, is an apparent problem. Roedder (1965) has shown that non-depleted mantle-derived ultramafic nodules in Hawaiian basalts contain significant numbers of CO_2-filled fluid inclusions.

4.24

The author therefore used Roedder's crushing stage method to determine whether gas-bearing

inclusions occur in the depleted upper oceanic mantle-derived ultramafic rocks of the map

area. Small pieces of the freshest specimens were crushed and the 0.1 to 0.25 mm size fraction

put in oil ($\mu = 1.64$) under a thick cover glass made from a piece of microscope slide.

Fragments showing tiny inclusions (0.01 to 0.001 mm or less, usually not more than 0.005

mm), rather smaller than Roedder's examples, were crushed by pressing on the cover glass

with a metal spike. A small proportion of the inclusions yielded gas bubbles that were quickly

resorbed into the oil. The quick absorption into the oil is quite strong, although not conclusive

evidence that the gas evolved was CO_2 (Roedder, 1965). Most of the apparent inclusions did

not contain gas; indeed some became separated from their host grain and are clearly solid.

Their refractive index is much less than olivine, and the oil ($\mu = 1.64$). They may perhaps be

glass, as Roedder (1965) mentions glass inclusions in the minerals of the Hawaiian ultramafic

nodules. They are definitely not serpentine or other alteration minerals, because in thin section

planar arrays of inclusions commonly observed in olivine are interrupted, and continue in the

olivine on the other side of serpentinised cracks. That is, the inclusions are destroyed by

serpentinisation. The inclusions almost always occur in planar and gently curving surfaces

running through part of an olivine grain. The inclusions are often spherical to ellipsoidal, and

sometimes are cigar-shaped or more irregular bodies. The planar arrays are likely to mark

crystallographic edge dislocations, and occasional linear arrays of cigar-shaped inclusions may

perhaps mark screw dislocations. Two-phase fluid inclusions were not seen either in thin

section or in the crushed material. However, many inclusions show a single dark speck on their

boundary. Some of these are tiny grains of chromite, but most are only just resolvable at

X1000, so their nature is uncertain.

They do not show Brownian motion as gas bubbles in two-phase inclusions do. Thus, it is probable that some CO_2 filled fluid inclusions occur in the depleted non-cumulate ultramafic rocks, and are the source of the CO_2 responsible for the ubiquitous carbonate alteration. The concentration of this type of alteration on zones of tectonic dislocation is readily accounted for by channeling of the gas being evolved from a large volume of the rock, during serpentinisation and tectonic transport of the ultramafic bodies.

Gas evolution from the ultramafic rocks is also indicated by the occurrence of tuffisite bodies within them. In the inland map area, these are found 0.2 km north of Marty's Pond in the northwest prong of the Mic Mac body. Well rounded clasts of ultramafic rock up to about 20 cm across are set in a foliated serpentine matrix and, in outcrop, only ultramafic clasts were seen. The foliation in the matrix is probably of the same age as the disruptive tectonic fabric on the margins, formed during the tectonic emplacement of the ultramafic body. The main area of tuffisite is fairly equidimensional, about 80 metres across, and could well be a section through a pipe. It also has two steeply-dipping dykes running out of it to the north (Plate 1). In this area several other small dykes and presumed dykes of tuffisite are found, ranging from less than a metre to a few metres wide. Some have a matrix that is not foliated. The matrix of all the tuffisites contains patches of talc and Mg-carbonate, but it is presumed to be mainly a serpentinised and recrystallised derivative of finely comminuted ultramafic rock. This is what is seen in thin section of abundant glacially transported boulders of ultramafic tuffisite found in the area 2 to 2.4 km ESE of the outlet of Black Lake (Plate 6). These boulders are most probably derived from the ultramafic bodies to the SSW. Most of the extremely well-rounded but not necessarily equidimensional clasts up to about 10 cm across are wholly serpentinised

harzburgite, but occasionally are foliated serpentinite. A few clasts show both a very well

rounded surface and a freshly-broken angular surface cutting it. One clast in thin section is

seen in the process of being split apart. Clasts range down to 0.05 mm across, and those

smaller than 1 mm are mostly parts of single serpentinised olivine or bastite crystals. These

smaller fragments may be quite angular. The matrix is fine-grained randomly-oriented

serpentine and subordinate talc with some patches of Mg-rich carbonate. The serpentinised

clasts often show mesh-texture serpentine, and the centres of the mesh are often selectively

replaced by talc. One small clast of a pre-existing fine-grained tuffisite was found in thin

section, and in addition one 0.8 mm clast of interlocking quartz grains, and one 1.2 mm clast of

spherulitic clinozoisite were observed. The quartz could be derived from many alternative

sources; the clinozoisite from associated gabbros or mafic volcanics of the Baie Verte Group.

These two 'foreign' clasts suggest, however, that the tuffisites were generated during tectonic

emplacement of the ultramafic rocks into their present position. The subvertical attitude of the

tuffisite dykes north of Marty's Pond also suggest that they were formed with the ultramafic

body in or near its present attitude, but the foliation in the matrix of some of the tuffisites here

indicates that these at least were formed before the deformation and emplacement ceased

completely.

(iii) Cumulate ultramafic rocks (Transition zone)

In well-preserved ophiolite complexes, these rocks form a zone of transition between

non-cumulate depleted ultramafic rocks and cumulate gabbros. In the inland map area, only

one occurrence of cumulate ultramafic rocks is still preserved in this position, and it is very

poorly exposed. Most other occurrences are in tectonically bounded slices, one is a large

boulder in Baie

4.27

Verte Group sediments, and some are glacial erratics.

At the southern end of the Flatwater ultramafic body (Plate 1 or 2) a group of four outcrops [29] of white-weathering ultramafic rock were examined in haste at the end of the last field season. Some of the outcrop is serpentinised non-cumulate (?) harzburgite, but a large proportion of the outcrop is thought to be serpentinised dunite. In addition, there are abundant bands and veins usually between 1 and 10 cm wide of serpentinised pegmatitic orthopyroxenite. These mostly trend SW-NE and dip steeply, concordant with the orientation of banding in the non-cumulate rocks to the north, and with the inferred attitude of the transition zone. However, some apophyses ramify off these bands, and wide veins in other orientations are quite common. Many of the pegmatite veins are slightly affected by tectonic movements, and anastomosing fractures and incipient augen of clumps of bastite grains are common. The bastites are often kinked, fractured, and variably granulated. The subhedral to anhedral pseudomorphs after orthopyroxene range up to 23 x 15 cm across in two dimensions, though they are most commonly between 0.5 and 1 cm across. Some chromite grains from 0.01 to 0.8 mm across are also present in the pegmatite. It is not known whether the host serpentinised dunite is cumulate. The reason for placing these outcrops in the transition zone is that abundant pyroxene pegmatites are generally restricted to this interval in ophiolite complexes (Dewey and Bird, 1971).

Between these outcrops, and outcrops of gabbro 100 metres to the east, there is no exposure, but in this interval there are several angular glacially-transported boulders about a metre across, that consist of cumulate lherzolite and coarse-grained cumulate gabbro (the particular lithologies are not seen as erratics elsewhere). The orange-weathering lherzolite boulder is almost certainly derived from this transition zone marked at the south end of the Flatwater body, because there is certainly nowhere else it can

be derived from in the direction of glacial flow before Jacob's Lake, 21 km to the SSW.

The cumulate lherzolite contains diopsidic clinopyroxene. Both olivine and orthopyroxene are wholly serpentinised, but the pseudomorphic replacement does not destroy all the critical textures. The proportions of clinopyroxene: orthopyroxene: olivine were about 45 : 35 : 20, and accessory magnetite (? altered chromite) is also present. All four components were cumulus phases; post-cumulus processes included overgrowth of both pyroxenes and resorbtion of olivine. Both pyroxenes occur as sieve-like poikilitic crystals up to 1 cm across, but most are smaller more equant grains. Grain sizes of pyroxenes and olivine are usually between 0.4 and 4 mm; magnetite(?) grains are usually between 0.05 and 0.4 mm across.

Very similar cumulate ultramafic rocks are found as small glacial erratics in the ditches of the Baie Verte Road 7.5 km north of the Burlington Road Junction. The place of origin of these distinctive mottled dark green rocks is less certain, as it is possible that they have been brought from some distance to make the road foundation, but it is more usual practice to use very local material. They are of cumulate lherzolite, with cumulus olivine and minor chromite, and sieve-like poikilitic post-cumulus orthopyroxene and diopsidic clinopyroxene. Typical olivine grain size is about 0.5 to 3.0 mm and the poikilitic pyroxenes are up to at least 1 cm across. Olivine and orthopyroxene are again both totally serpentinised, and large rosettes of spherulitic talc are abundant in one specimen.

Several angular erratics up to 50 cm across were found at the south end of the pond opposite the Burlington Road Junction. These are all of a distinctive dark green rock studded with ovoid whitish-gold clinopyroxenes that stand out on the weathered surface. The ovoid diopside grains commonly

between 1 and 4 mm across comprise about 30% of the rock and together with a few small chromite grains are set in a matrix of mesh-serpentine liberally overgrown by acicular tremolite and some chlorite. This rock was a probably cumulate wehrlite, but the alteration has destroyed too much of the original textures to define the cumulus and post-cumulus phases with confidence. This group of erratics are almost certainly derived from the transition zone at the southern end of the Flatwater body.

Clinopyroxenite forms the whole of two small lensoid tectonic bodies on a Baie Verte slide zone, 0.7 km south along the power line, and 0.6 km north along the Baie Verte Road, from the Burlington Road Junction. Apart from a thin skin of sheared serpentinite a metre or so wide, both these bodies consist entirely of undeformed silver-grey weathering granular clinopyroxenite. Xenomorphic granular augite(?) from 0.25 to 3 mm, commonly 1 to 2 mm across forms more than 95% of the rock. The remainder is small interstitial patches of bastite serpentine, and several neighbouring patches commonly have exactly the same fibre orientation. This shows that poikilitic orthopyroxene was present. No ore mineral grains occur. This rock was formed by clinopyroxene as the single cumulus phase, followed by postcumulus overgrowth of the same composition (no zoning is present) accompanied by interstitial crystallisation of small amounts of postcumulus orthopyroxene.

In outcrop [30], 2.1 km N of the Burlington Road Junction on the power line, very similar but slightly granulated clinopyroxenite containing small (1-10 cm) Ca-metasomatite ('rodingite'- *sensu lato*) inclusions occurs as a large boulder in the Baie Verte Group mafic volcaniclastic sediments.

All these occurrences of clinopyroxenite are most probably derived from a cumulate ultramafic transition zone.

Thus, the evidence from the inland map area is sufficient to show that rocks typical of ophiolite complex transition zones are present in the Baie

Verte Lineament, but no evidence is available on the original relationships between these rocks and the non-cumulate ultramafic rocks, or between them and the diabase dykes in the gabbros above. The thickness of the transition zone marked at the southern end of the Flatwater ultramafic body is about 300 metres, but only a maximum of 100 unexposed metres thickness may be banded cumulate ultramafic rocks, and it is not known if this is structurally intact. The remainder of this transition zone is serpentinised dunite and harzburgite that may or may not be cumulate, and is included in the transition zone only because of its abundant orthopyroxene pegmatite content. A little more data on transition zone rocks is available from the Mings Bight area, described in the following chapter.

(iv) Ophiolite complex gabbros

(a) Introduction

A generally thick sequence of both layered and homogenous gabbro form a significant component of all well-preserved ophiolite complexes (Dewey and Bird 1971). The layered rocks, most of which are originally cumulate, mainly occur in the lower part of the gabbro immediately above the ultramafic transition zone, and the lowest part of the layered gabbro usually has some olivine- and orthopyroxene-bearing ultramafic cumulate layers. The homogenous gabbro is mostly a feldspathic, somewhat leucocratic variety and generally forms a majority of the whole. Some may be cumulate but it has not been demonstrated to consist wholly, or even dominantly of cumulate rocks in any well-preserved ophiolite complex. Some parts of the gabbro may be cut by parallel diabase dykes, and these increase rapidly over a short distance upward into the sheeted diabase dyke complex above the gabbro. It is of critical importance for models of ophiolite complex formation at oceanic spreading ridges to determine whether diabase dykes commonly cut cumulate gabbros.

In the inland map area, ophiolite complex gabbros are mostly found in places other than their original position in the ophiolite complex sequence. At the southern end of the Flatwater body (Plates 1 or 2) a very poorly exposed triangular area west of the Western Lineament Boundary slide (W.L.B. slide for short — a Baie Verte, D_B slide zone) interpreted as gabbro is probably more or less in its original position with respect to the adjacent transition zone, at least when compared with all other occurrences in the inland map area. All these areas are found to the east of the W.L.B. slide. They are divided into four categories, based on their type of occurrence.

1) In the south (Plate 1), a fairly continuous strip adjoining the W.L.B. slide from Mic Mac Lake to 1.5 km NNE of Kidney Pond, and up to 200 metres wide, displays outcrops of various ophiolitic gabbroic rocks. For reasons discussed in a later section, this strip, which is everywhere directly overlain by the ophiolite debris-bearing Kidney Pond Conglomerate Formation, is thought to be composed of closely-packed gigantic blocks (sedimentary as opposed to tectonic blocks). This is only relevant to the present discussion in that internal igneous textural and structural relationships cannot be correlated between adjacent outcrops, nor can they be related precisely to their position in the original ophiolite complex.

2) Discontinuous areas of gabbroic rocks also adjoin the east side of the W.L.B. slide from the Old Camp 32 Road northward (Plate 2). They are separated by the slide from the gabbro at the southern end of the Flatwater body. Although all this second group may, like the first group, be deposits of gigantic sedimentary blocks, or single enormous sedimentary blocks, there is no proof that each not a relatively intact, tectonically bounded and derived, body. They are all separated from the Kidney Pond Conglomerate Formation by a variable thickness of mafic volcaniclastic and other sediments, in contrast to the first group. However,

because these bodies may be collections of disoriented blocks, the relationships seen are treated in the same way as for the first group.

3) Near the northeast corner of Flatwater Pond (Plate 2) one ultramafic and one gabbroic body are thought to be large sedimentary blocks in mafic volcaniclastic sediments. Some more pieces of ophiolite gabbro too small to show on the map are found at the same horizon in outcrops on the northeastern shore of the pond, and they are definitely boulders in mafic sediments. No internal igneous relationships are seen in these blocks that are not better displayed elsewhere.

4) The tectonic and/or sedimentary relationships of outcrops of gabbro adjoining the Baie Verte Road around 7.5 km north of the Burlington Road Junction are not known. They may be boulders in the Baie Verte Group sediments, but insufficient outcrop is available to solve the problems of this very complex small area. However, 9.5 km north of the Burlington Road Junction, a roadcut shows ophiolite gabbro that is definitely a large tectonic lozenge on a Baie Verte D_{1B} slide zone.

The ophiolite complex-derived gabbros are distinct from dolerite and minor gabbro sills in the Baie Verte Group mafic volcanics and volcaniclastic sediments. The criteria used to distinguish them are textural and, to a lesser extent, compositional. The sills are almost all dolerites (grain size much less than ophiolite gabbros), had ophitic to poikilitic (not cumulate-type seive-poikilitic) pyroxene and subhedral to euhedral long plagioclase laths, always contained some accessory ilmenite, and are nearly always meso- to melano-cratic. The ophiolite gabbros are mostly rather leucocratic (feldspar-rich) , and rarely contain much, or any ilmenite or other ore mineral. They generally had xenomorphic granular pyroxene and equant subhedral to anhedral plagioclase. There are a few outcrops of gabbro that are difficult to place in one or other category, and these are described in a later section.

(b) Cumulate rocks

Clearly cumulate gabbros are found in outcrops in only three small areas in the inland map area.

On the west side of the Baie Verte Road 7.5 km north of the Burlington Road Junction, an outcrop [31] of ophiolite gabbro has uncertain external relationships. A band of granular greenish-gold clinopyroxenite about 30 cm wide is contained in the outcrop and, although there has been some tectonic movement on both margins, it is most probably a band within the gabbro. The xenomorphic granular unzoned diopside forms 99%+ of the rock, and grains range from 0.1 to 1.6, commonly 1.0 mm across. There are also 1% or less of small interstitial pseudomorphs after orthopyroxene, and the rock is definitely cumulate. It is almost the same as that occurring in larger isolated bodies, described in the previous section and illustrates how the transition from ultramafic cumulate to gabbroic cumulate rocks involves the interbanding of components from both units. The gabbro in this outcrop is mostly atypical with an intense high-temperature foliation and banding, but it still contains some relict sieve-poikilitic diopside grains, demonstrating its cumulate origin.

At the southern end of the Flatwater body, the gabbro outcrops east of the transition zone are cumulate rocks but they are not layered or banded. In these, as in all gabbros and all other rocks in the Baie Verte Group, plagioclase is totally albitised and may be more altered to fine-grained calc-silicate aggregates. However, in the gabbros, the original shapes of the grains are usually retained unless the rock has been affected by (low-temperature) destructive tectonic processes. Three thin sections of gabbros from the outcrops near the south Flatwater transition zone were examined, and all were two-pyroxene cumulate gabbros. One contains a few (~1%) pseudomorphs of cumulate olivine but no ore minerals are present

in any of the specimens. The proportions of clinopyroxene: orthopyroxene: plagioclase in each the three specimens is about 55:15:30, 45:20:35, 35:15:50. Grain sizes range from 0.1 to 2.5 mm and are most commonly between 1 and 2 mm. Occasional grains of either pyroxene occur up to 8 mm across. Plagioclase is in subhedral stubby prisms but is anhedral in the plagioclase-rich specimen. The clinopyroxene is a diopside or diopsidic augite. Orthopyroxene is wholly altered and pseudomorphed by bastite or chlorite and tremolite. Both pyroxenes are mostly in anhedral prisms, though some is interstitial and some orthopyroxene especially is also ophitic to sub-poikilitic. The feldspar-rich specimen displays some glomeroporphyritic clumping of both plagioclase and clinopyroxene. To summarize, these three specimens all have somewhat variable proportions of both pyroxenes and plagioclase as cumulus phases, with a very minor amount of cumulus olivine in addition in one specimen. Post-cumulus processes appear to involve overgrowth of clinopyroxene and especially orthopyroxene with either minor overgrowth or resorbtion of plagioclase. These outcrops of cumulate gabbro are each cut by one or two parallel diabase dykes.

The third place where cumulate rocks are found is in two outcrops on the shore, one on each side of the narrows at the northwest corner of Flatwater Pond. Both outcrops are cut by rather irregular dykes and small bodies of diabase. On the southern side of the narrows, two specimens were found to be cumulate. One is very similar to the two relatively plagioclase-poor specimens described in the previous paragraph, except that orthopyroxene is rare. The other specimen shows sieve-poikilitic and interstitial green hornblende instead of a pyroxene, is plagioclase-rich and also contains some accessory cumulus ilmenite. On the small peninsula forming the northern side of the narrows, fine-grained granular gabbro,

which is not apparently cumulate, contains lensoid pods of cumulate anorthosite up to about 30 cm wide and several metres long. One thin section of anorthosite shows 100% cumulate (albitised) plagioclase between 0.4 and 3.0, commonly 1.5 mm across. Another section has a small amount (~1%) of interstitial serpentine and tremolite, probably after orthopyroxene. In outcrop, slickensided shear surfaces within the anorthosite show scattered flecks of fuchsite, indicating the presence of a little chromite, but this was not detected in thin section. These anorthosite pods are probably remnants of cumulate layers. It is thought that the enveloping fine-grained granular gabbro may have been recrystallised during and after deformation in a plastic state at high temperature shortly after its original crystallisation, as a cumulate, accompanying the anorthosite. The cumulate clinopyroxenite in an outcrop of foliated gabbro that was described at the beginning of this section also illustrates the same phenomenon. This will be discussed after the other gabbros are described in the next section. Although other parts of the various areas of ophiolite gabbro nay be cumulate, they have not been positively identified as such. This is partly due to the ubiquitous alteration but most cumulate two-pyroxene gabbros are difficult to recognise even when unaltered (Davies, 1971).

In the inland map area, cumulate gabbros typical of ophiolite sequences are present but uncommon. From the examples found, little can be concluded about their relationships to one another, or about their place in the gabbro unit of the ophiolite complex that they were derived from.

(c) Other ophiolite gabbroic rocks

Most of the ophiolite gabbro in the inland map area is not provably of cumulate origin, and besides the difficulty in recognising the cumulate nature of many fresh gabbros, retrograde mineral alterations and overgrowth

obscure the textures. Also, some of the gabbro has been affected by a high-temperature (sub-magmatic) gneissic foliation (described later) that has changed the original textures.

The bulk of the non-cumulate gabbros are granular relatively feldspathic rocks, commonly containing 50% or more plagioclase. A few were two-pyroxene gabbros before alteration, but most cannot be shown to have been other than clinopyroxene gabbros. Most have xenomorphic granular textures, with typical grain size between 1 and 2 mm. There is a conspicuous absence of ilmenite (or its sphene pseudomorphs) in most rocks, although it is seen in a few examples, most of which are very mafic, pyroxene-rich gabbro. All the gabbro and related rocks in the occurrences south of the Old Camp 32 Road is wholly-altered, with pale actinolite after clinopyroxene and albitised plagioclase. More extreme, generally texture-destructive, alteration may involve large amounts of clinozoisite and rarely zoisite and chlorite. North of the Old Camp 32 Road, clinopyroxene is often unaltered, although plagioclase is always albitised or otherwise replaced. Primary hornblende occurs in a few places in the gabbro in both areas, and is unaltered in every case except one.

Minor amounts of other related lithologies are found within the 'normal' gabbro. Anorthosite bands from 0.5 to 4 cm wide occurring in outcrop [32] at the south end of Kidney Pond were probably of cumulate origin, but they have been subsequently affected by the high temperature gneissic foliation. Similar bands are seen in the next but one outcrop to the south, and in outcrop [33] west of the northern part of Mic Mac Lake. Mafic gabbro, containing 15% or less plagioclase and minor occurrences of altered, originally clinopyroxene-rich, picrite are best displayed in parts of outcrops up to 1.6 km south of Kidney Pond. Examples of mafic gabbro and picrite are

occasionally found elsewhere, but they form a very small proportion of the total exposure of ophiolite gabbro. The mafic gabbro may form bands probably with gradational boundaries of the order of tens of metres thick in the 'normal' gabbro but a clear relationship between them has not been seen. The picritic rocks form bands about 30 cm wide in 'normal' and mafic gabbro, but they are parallel to local diabase dykes, and may themselves be dykes. The texture and grain size of both mafic gabbros and picrites are very similar to 'normal' gabbro, except that they usually contain significant accessory sphene after ilmenite. It is not clear whether some or all of these gabbros and picrites are cumulates. Scattered occurrences of pegmatitic gabbro are seen in most of the areas of ophiolite gabbro. They are either in irregular patches up to about 10 cm across or in rather irregular veins a few centimetres wide, both having rapidly gradational contacts with 'normal' gabbro. Grains within them typically range up to 1 cm across. Most cut the high temperature gneissic foliation in the gabbro, but some are affected by the foliation (e.g. in [32]). Gabbro pegmatites are more abundant and far better displayed in the Mings Bight area, described in the next chapter.

Trondhjemitic rocks are found in the gabbro in a few localities. On the west side of the Baie Verte Road, 9.5 km north of the Burlington Road Junction, 'normal' altered ophiolite gabbro contains a few crosscutting veins or dykes from 1 to 5 cm across. The thicker (5 cm) veins are grey and, in thin section, show a fine-grained (0.05-0.5 mm) granular aggregate of quartz and slightly sericitised, albitised, plagioclase, in the approximate proportions 2:3. A few of the larger plagioclase grains are subhedral stubby phenocrysts, but most are anhedral. The larger plagioclase grains have a very narrow zone of clear original albite on their rims. Accessory

apatite and ilmenite grains around 0.05 mm across and a few patches of chlorite and Al-epidote of similar size are present. The narrower veins (1-2 cm) in this locality are pinkish buff. A thin section shows 70% euhedral rhombic to lath-shaped saussauritised piagioclase with narrow clear albite rims, 15% interstitial quartz, 15% green hornblende, and accessory ilmenite and apatite. The plagioclase is not aligned, ranges from 0.1 to 4, commonly about 0.8 mm long. Quartz grains have irregular shapes with grains from <0.1 to 1.2 mm across. Hornblende phenocrysts from 0.1 to 1 mm across are generally interstitial to the plagioclase. There is a slight decrease in average plagioclase grain size towards the margin of the vein, but no genuine chilled margin is present. Both these rocks are trondhjemites. A large outcrop consisting of rock almost identical to but slightly coarser grained than the second type is found on the northern shore of Flatwater Pond 0.3 km west of the stream at the northeastern corner. It also contains pink aplite veins. It has been marked as ophiolite gabbro on the map and is thought to be a large boulder within the mafic volcaniclastic sediments, though it could be part of a sill.

On the eastern side of the Baie Verte Road, 0.8 km south of Middle Arm Brook, an outcrop of white granophyre is just within the gabbro unit but its relationship to the gabbro is not seen. A granular fine-grained (<0.1 to 0.5 mm) aggregate of quartz and albite forms half the rock, and most of the remainder consists of patches of well-formed granophyric quartz and feldspar. About 5% is composed of clots up to 1 mm across of acicular fine-grained stilpnomelane. Rare specks of accessory ilmenite also occur.

Small (~30 cm) pods of 'albite granite' (trondhjemite) are found in 'normal' ophiolite gabbro 0.6 km directly south of the Burlington Road Junction. They also occur in the two outcrops on both sides of the narrows

at this corner of Flatwater Pond. Most of the boudin-like pods exposed do not show a clear

relationship to the gabbro, or to the cumulate anorthosite. However, one pod in the outcrop on

the northern side of the narrows has a fine-grained grey margin, that may perhaps be due to

original chilling. A thin section of a typical sample shows granular albite and subordinate and

somewhat interstitial quartz, both typically between 0.1 and 1 mm across, with minor

hornblende of similar grain size. Accessory apatite is present.

(d) High-temperature gneissic foliation and hot mylonites

In addition to the isotropic granular gabbros, a proportion of the outcrops of ophiolite

gabbro possess a foliation best described as gneissic. This foliation is defined by flattening of

the pyroxene and plagioclase grains and, in the generally plagioclase-rich gabbro, this is mostly

shown in outcrop by parallel alignment of the elongate sections through flattened mafic

mineral aggregates. A preferred elongation lineation has not been seen. This foliation is

independent of and far more widespread than the rare examples of lithological banding, but

when banded rock is also foliated, as in outcrop [32], the foliation is coplanar with the banding.

Similar gneissic fabrics in both ultramafic and gabbroic parts of other ophiolite complexes

have been termed 'flow layering' (Thayer 1963). This term was proposed with the belief that

the banding was produced by the deformation, which is not the case. It is therefore

inappropriate and avoided in this thesis. In the map area, the foliation varies in intensity and all

gradations are present between the most foliated and unfoliated rock. In the case of unfoliated

and weakly-foliated gabbro, the change takes place gradually over several metres. Most of the

foliated gabbro is fairly weakly foliated. Its original mineralogy appears to have been

clinopyroxene-plagioclase. There is no evidence of replacement of pyroxene by hornblende

during the foliating process in rocks with the foliation here described as gneissic,

except in one locality where the foliation is extremely intense and in a relatively narrow zone, described later in this section. As the plagioclase is now ubiquitously altered, any possible slight change there may have been in its composition during foliation is not detectable. In thin section, the flattened mafic mineral blebs consist of aggregates of granular fairly equant grains (of pale actinolite after clinopyroxene) that are evidence for the breakdown of a large original grain into several smaller ones during the production of the gneissic foliation. Fine-grained granulation products are not present as far as can be seen in the altered rocks available. Plagioclase shows a similar granulation and polygonisation to the mafic mineral clots. Typical grain sizes in moderately foliated rock are 0.1 to 0.5 mm. Pegmatitic gabbro may be affected by the foliation but most commonly it postdates it. No banding is later than the foliation where the rocks are foliated. This gneissic foliation always predates diabase dykes.

Some of the gabbro in outcrop [33], west of the northern part of Mic Mac Lake, is a 'flaser-gabbro', where the gneissic foliation defined by the mafic clots is wispy and wavy. All of the flattening fabric and some of the wavy nature of the mafic aggregates is due to granulation and flattening during high temperature deformation, but some later microshear planes are also responsible for some of the wavy shapes. All gradations between original large (~ 4 mm) plagioclase grains with incipient kinking and granulation to fully-polygonised aggregates can be seen in this rock. The mafic aggregates are now almost wholly altered but one retaining clinopyroxene in its centre and several retaining some orange-brown amphibole and/or green hornblende were seen. Adjacent fine-grained polygonal granulitic-textured gabbro contains only strongly coloured green-brown hornblende. This polygonal granulitic fine-grained (0.01 to 1 mm, commonly 0.1 to 0.5 mm)

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gabbro contains a few narrow (2 mm to 1 cm) anorthosite bands, which were probably cumulate in origin, parallel to the gneissic flaser foliation in gabbro a metre away. No traces of outlines of larger mafic grains were seen in this granulitic gabbro. Very similar fine-grained granulitic gabbro, containing a brown-orange, possibly titaniferous, amphibole, is associated with the boudined cumulate anorthosite pods on the north side of the narrows of Flatwater Pond. However, a loose block (not a glacial erratic) found next to an outcrop of gabbro west of the Baie Verte Road and north of the northwestern corner of Flatwater Pond shows the same type of fine-grained granulitic gabbro, containing green hornblende, in sharp contact with hornblende gabbro of normal (1-2 mm) grain size. This is almost certainly an original magmatic contact, and is either a hot dyke or sill contact, or a cumulate layer contact. Therefore, it is not clear whether the polygonal granulitic relatively fine-grained texture of the other examples is due to strong high-temperature deformation and thorough annealing, or to normal magmatic crystallisation. Examples of fine-grained granulitic gabbro have been seen in a few other localities, but they do not yield additional information. The lithology is not connnon, but it always seems to contain hornblende or an orange-brown possibly titaniferous amphibole, never clinopyroxene.

The most intense gneissic high temperature foliation seen in gabbro in the inland map area is found in an outcrop on the west side of the Bale Verte Road 7.5 km north of the Burlington Road Junction [31]. Here, the intense foliation parallels both poorly-defined more mafic and more feldspathic bands less than 1 cmwide and well-defined coplanar bands up to 5 cm wide overprinting the former, and containing either amphibole or pyroxene. In thin section, granular altered plagloclase (~ 0.1 mm) with somewhat irregular weakly-sutured boundaries is found both in the bands that contain bright orange amphibole, and those containing diopside. Rare diopside grains occur

in the amphibole bands and *vice versa,* but the segregation is very marked. The change from amphibole to pyroxene across the banding is abrupt. The amphibole grains are all anhedral lozenges weakly elongate in a section across the foliation, and typically 0.1 to 0.5 mm across. Diopside grains are mostly about 0.2 to 2 mm across and markedly elongate in a section across the foliation. A few diopsides are larger (to 4 mm) and some of these show relict sieve-poikilitic cumulus texture with included plagioclase. The rock also contains elongate sphene in the diopside bands, a few of which are very large (up to 6 x 2 mm). Small sphene grains also occur sparingly in the amphibole bands, but small apatite grains are only present in the diopside bands. Fine-grained magnetite(?) is found in small amounts throughout the rock. It may be after chromite because a fuchsite-green colour pervades the plagioclase areas in hand specimen, although fuchsite was not identified in thin section. The orange amphibole is probably titaniferous although the darkest pleochroism it displays is a bright orange, insufficiently dark for it to be kaersutite. Both the rock type and the strong foliation are unusual but the coexistence of both amphibole and pyroxene in it is fairly clear evidence that the gneissic foliation was formed at high temperature. Not all the gabbro in this outcrop is foliated. The relationship of the foliated rock to the cumulate clinopyroxenite in the northern part of the outcrop is not known. Although the relationship between the intensely-foliated gabbro and little to undeformed gabbro is not exposed, the foliated rock must form a zone a metre or two wide within the little deformed gabbro. It represents a transitional example between the broad zones of (weakly) gneissically-foliated gabbro, and gabbro with narrow zones of very intense foliation also developed at high temperature, described below.

Fig. 4.4 Ophiolite gabbro textures.
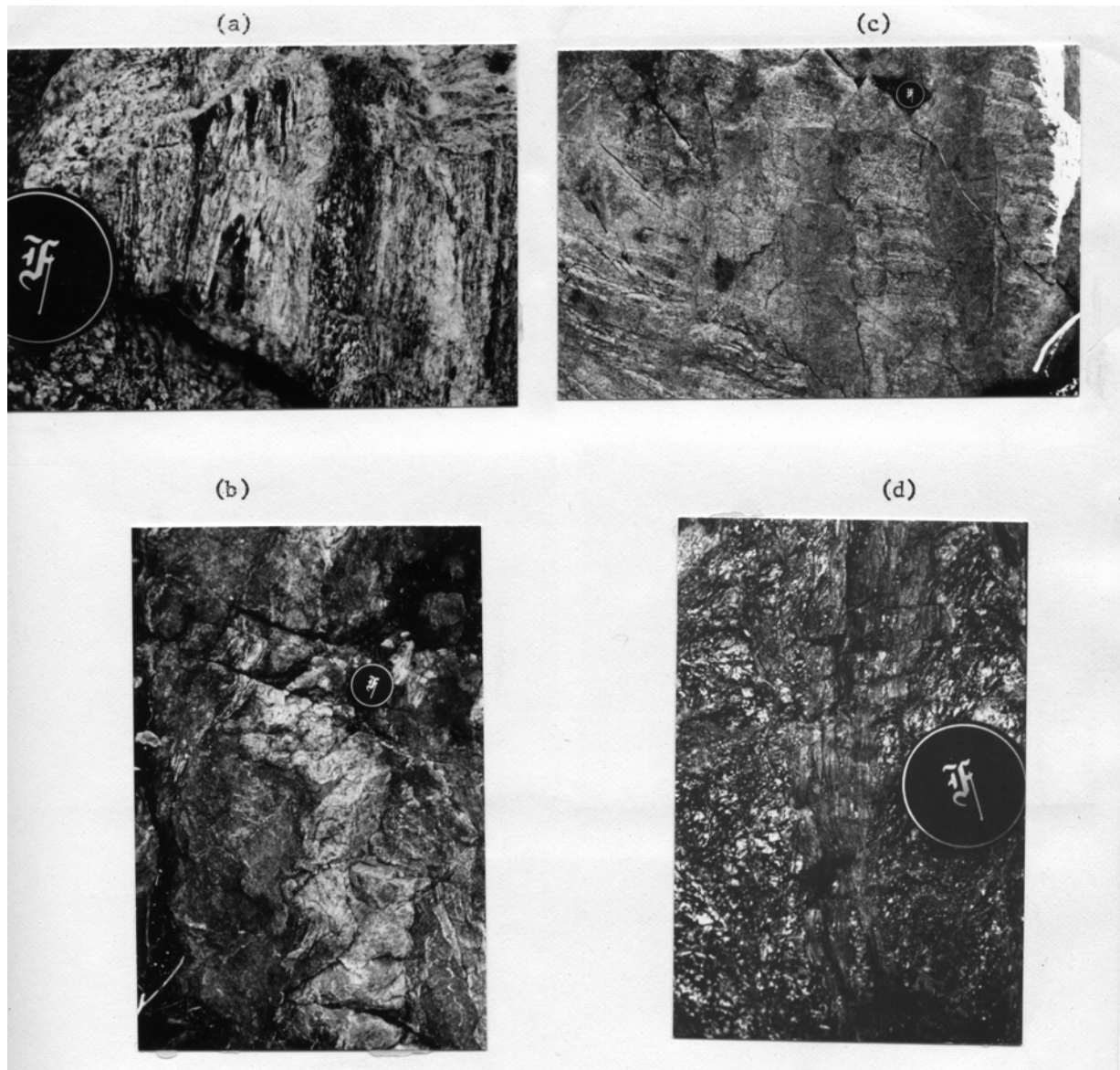
(a) hot mylonitised gabbro; intrafolial folds of mafic clots. Kidney Pond [34].

(b) plastically folded diabase dyke in gabbro; Kidney Pond [34].

(c) diabase dykes truncated along resealed plane; Kidney Pond, 60 m NE of [32].

(d) hot mylonite zone involving diabase; N. Mic Mac Lake, [33].

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In outcrop [34], just south of Kidney Pond, mafic and normal gabbro are cut by narrow hot mylonite zones. These are typically 1 to 10 cm wide and show intense deformation within the zones with a rapid gradation outward across their margins into essentially undeformed gabbros. The zones and their foliation are parallel to the weak gneissic high temperature foliation in the adjacent outcrop [32], but this fact may not be significant as they may be in separate blocks. The hot mylonite zones show very strongly flattened and sometimes intrafolially-folded mafic clots (Fig. 4.4a). In thin section, the rocks are a very fine-grained granular aggregate (0.01 mm or less to 0.1 mm, mostly less than 0.05 mm) of albitised plagioclase and green hornblende. Both form trains and narrow long lenses of fairly equant grains that define the foliation. Occasional larger ovoid green hornblende grains (~ 1 mm across) are found augened in the foliation. In terms of the deformation, this is just a much more extreme development of the high-temperature gneissic fabric but its localisation in sharply-defined narrow zones and the development of hornblende in the zones, while the host gabbro retained clinopyroxene, are distinguishing factors. In this outcrop, a diabase dyke is seen plastically folded and 'back-veined' by ordinary non-foliated mafic gabbro (Fig. 4.4b). This is not apparently directly connected with any of the hot mylonite zones. Plastic remobilisation of gabbro after brittle failure and dyke intrusion raises an apparent problem. However, this is the only example seen, and may be explicable in terms of relative strain rates or perhaps by local reheating of the gabbro (by nearby intrusion of more gabbro ?). Nearby, in the outcrop about 60 metres NE of outcrop [32], two closely-spaced parallel diabase dykes are truncated at a high angle by a plane in the gabbro that has very little expression on the weathered surface, and that appears to be a dislocation resealed at high temperature (Fig. 4.4c). It is not a (low temperature) brittle failure plane, or a hot mylonite zone. This is the only other example

seen in the area of 'hot' deformation of diabase dykes. Other diabase dykes cut the hot

mylonite zones in [34].

Some parts of some of the hot mylonite zones in [34] consist partly of homogenous or

foliated very fine-grained grey-green weathering rock that quite strongly resembles the diabase

dykes, but is, at least in part, the product of extreme mylonitisation and granulation of the

gabbro. In [34] these are rather obscured in thin section by alteration mineral growth, but

identical features are seen in one other outcrop [33] south of the north end of Mic Mac Lake

where their relationships in outcrop and in thin section are better displayed.

These 'diabase' hot mylonite zones are abundant in [33], where they cut narrow

anorthosite bands in granulitic fine-grained hornblende gabbro, and the foliation in the flaser

gabbro. Their relationship to diabase dykes has not been seen. These zones are grey-

weathering, and have a flinty conchoidally-fractured fresh surface. Some parts show a lensoid

foliation, produced by streaks of slightly different shades of grey-green, and also by lensoid

inclusions both of partly granulated host gabbro and of pre-existing diabasic hot mylonite. The

zones are tabular and range from less than a centimetre up to at least 10 cm wide. In the

northern part of outcrop [33] they occur in such abundance that locally they comprise up to

30% of the rock. It is, however, more significant that the hot mylonite zones run in several

highly oblique orientations, outline rectangular blocks of gabbro and that some zones cut

others (Fig. 4.4d). These last observations show that these zones must be hot rather than cold

mylonites. In thin section the flinty grey portions are brown irresolvable material that is

identical to devitrified basaltic glass seen elsewhere in the chilled margin of a diabase dyke.

This material contains fragments of plagioclase, clinopyroxene

and green and orange hornblende grains derived by granulation from the host gabbro, besides the larger lensoid partly granulated fragments of host gabbro, and pre-existing foliated hot mylonite. It is not clear if the zones were formed by intrusion of basaltic magma into granulation zones with further deformation of the resultant mixture, or whether they consist entirely of milled and perhaps melted derivatives of the gabbro. It seems most likely that basaltic magma was involved, as some of the narrowest zones consist entirely of granulated material, whose matrix does not resemble the flinty diabasic matrix of the wider zones.

Some other zones less than a centimetre wide consist of flinty diabasic material that could only have its present extent along such a thin zone with slightly diffuse margins by continued movement across the zone during and after intrusion of the magma.

The weak gneissic high-temperature foliation described from outcrop [32] parallel to originally cumulate anorthosite bands and found fairly commonly in the exposures of ophiolite gabbro is the only fabric that is found in gabbros in the Mings Bight area. Published descriptions, of which this author is aware, of the gneissic foliation found in gabbros of other well-preserved ophiolite complexes do not include foliations like the hot mylonites described above. There appears to be a sequence in this map area from the weak gneissic foliation affecting wide zones, through the more intense foliation in zones a metre or so wide, to the hot mylonitic zones in outcrop [34] affecting gabbro over widths of a few centimetres. The granulation on zones in [33] may represent an extension of this sequence but the possible involvement of basalt magma in these zones is not a direct part of this sequence. There seems also to be a correlation between this sequence and the replacement sequence clinopyroxene to hornblende or an

orange-brown amphibole. The former sequence seems likely to be due to decreasing temperature and the latter may be also, although it is more likely to reflect the relative ease of access of water to the deforming zones.

The uncommon narrow foliated zones that appear (but probably are not) unique to the ophiolite gabbros of the map area may be accounted for by the hypothesis that they reflect processes occurring along a transform fault, close to a spreading ridge within the active portion of the transform. Several lines of evidence that appear to support this hypothesis are described in some of the subsequent parts of this chapter.

(e) Tuffisites

A large (50 X 20 metre) boulder of ophiolite gabbro is contained in two outcrops [35] in the Kidney Pond Conglomerate Formation, 1.1 km NE of the north end of Mic Mac Lake. The gabbro in this boulder is atypical and the block is clearly within sediment. Although gas-brecciated gabbro is found elsewhere within the gabbro unit underneath the conglomerate, this block is described because it is by far the best exposure of gas-brecciated ophiolite gabbro. The gabbro in this block mostly has a coarse doleritic texture with clear albitised plagioclase in squat laths (0.2 to 1 mm) with intergranular altered clinopyroxene (0.1 to 0.6 mm) and accessory magnetite (0.1 mm). It is broken into angular to subangular blocks up to 10 cm across, separated by white-weathering veins up to a few centimetres wide. The veins greatly resemble albitised anorthosite but consist entirely of a fine-grained granular aggregate of clinozoisite. All sizes of angular fragments down to pieces of the minerals constituting the doleritic gabbro are found in the veins. A very few small fragments are altered granular mafic gabbro and clinopyroxenite. In outcrop, a small proportion of the clasts are

diabase and one example of a larger block of vein brecciated diabase was seen. Most diabase clasts are more or less isolated and a chilled contact between diabase and gabbro is seen in one of these. Presumably the diabase is derived from dykes and shows that the clasts in the breccia are mostly very disjunct despite the apparent similarity of most of the gabbro, the angularity of the clasts and the narrowness of the veins. A clast 1 mm across of fresh brown chromite, seen in a vein in thin section, confirms the disjunct nature of the clasts. It is very likely to have been derived from the ultramafic part of the ophiolite complex (upper oceanic mantle). Although it is not understood why the veins now contain clinozoisite, it seems probable that this breccia is a gas-breccia (tuffisite). The chromite grain suggests that the gas responsible in this particular case was CO_2 originally contained in the ultramafic rocks, demonstrated in a previous section. Very similar tuffisite is common in outcrops within the gabbro exposed from 0.6 km north of [35] to Mic Mac Lake, but is rare or absent elsewhere. A diabase dyke in [33] is brecciated by very similar white veins confined to the dyke. This is probably also a gas breccia. Diabase in outcrops around the northwest corner of Flatwater Pond contains abundant clinozoisite veins in some places and these may be due to the same process. One dyke in the sheeted diabase dyke complex at the northern end of the map area contains a brecciated zone but the matrix here contains small clasts of finer-grained diabase and is composed of comminuted diabase debris rather than clinozoisite. This example is certainly a tuffisite vein.

Although the tuffisites previously described in the ultramafic rocks are thought, from their present steep attitude, to have been generated during Bale Verte deformation, a tuffisite in ultramafic rock in the Mings Bight area is shown in the next chapter to have been developed on an ocean

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floor fault. This and the gabbro tuffisite containing chromite in outcrops [35] are thought to be additional evidence that ocean floor fault activity is represented in the textures seen in the ophiolite gabbro.

(f) Serpentinite inclusions

Several outcrops in the gabbro unit contain small slivers of shear polyhedra serpentinite. These are (Plate 1) outcrop [33] and (Plate 2) the small strip just south of the Bear Cove Road, in the larger area north and south of the northwest corner of Flatwater Pond and in the area of sheeted diabase dykes at the northern end of the map area. In the one example seen in each of the first two localities, the shear polyhedra serpentinite is about 20 cm wide in presumed small faults. The examples from the other two localities are wider, up to 20 metres across and all have been indicated on the map. In the cases of these larger examples, it is not clear whether they have been introduced on faults or on Baie Verte tectonic slides (as appears to be the case for the two pyroxenite bodies north of Flatwater Pond) or whether some or all were introduced as large boulders. Judging by the ubiquitous shear polyhedra deformation of even the larger bodies, when compared with the essentially undeformed adjacent gabbros, it seems unlikely that they were introduced as boulders, unless they were already shear polyhedra serpentinite before they became boulders.

(g) Metasomatism

Within the gabbro bodies, just north of Kidney Pond and just north of the northwest corner of Flatwater Pond, very small and rare patches not more than 20 cm across are affected by calcium metasomatism (Coleman, 1967). This is not apparently related to the inclusions of shear polyhedra serpentinite. Gabbro affected by calcium metasomatism becomes a fine-grained more or less homogenous and featureless white rock, which grades into normal gabbro outside the metasomatised patch. One example

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retained some green clots of pale actinolite and chlorite within the metasomatic white rock. The examples of metasomatite seen in thin section are mostly composed of zoisite and subordinate clinozoisite with varying amounts of pyrophyllite(?) and minor chlorite and calcite. One white vein about 3 mm across in a sample from the thin strip of gabbro just south of the Bear Cove Road consists of coarse prehnite. This may or may not be due to alteration prior to the Baie Verte Lineament deformation. Apart from these rare and minor examples of Ca-metasomatism, all other alteration mineralogy and assemblages in the gabbros are entirely compatible with the low greenschist fades metamorphism associated with the Bale Verte Lineament deformation described later.

(v) Diabase dykes

(a) Dykes in gabbro

Parallel diabase dykes are common in all of the areas of ophiolite gabbro in the inland map area. These are grey-green weathering fine-grained hornogenous rocks. Almost all the dykes are between 10 and 60 cm wide, and most are between 20 and 40 cm wide. Obviously chilled devitrified glassy margins are not often seen. Fine-grained margins are common, although they are not easily observed in outcrop. Most dykes are non-porphyritic but a minority contain some (perhaps 5%) small (2-3 mm) fairly equant plagioclase phenocrysts. One thin section contained a small (0.5 mm) clinopyroxene phenocryst and a few small pseudomorphs probably after orthopyroxene phenocrysts. Similar orthopyroxene pseudomorphs were seen in a thin section containing the devitrified glassy chilled margin of another dyke. When the diabase retains its original texture (where it has not been destroyed by alteration mineral overgrowth) all diabase examined has a very characteristic and diagnostic intergranular texture (Fig. 4.5). The plagioclase laths (0.1 to 0.3 mm long) are randomly oriented and homogenously

Fig. 4.5. Diabase dyke texture; ophiolite gabbro textures.
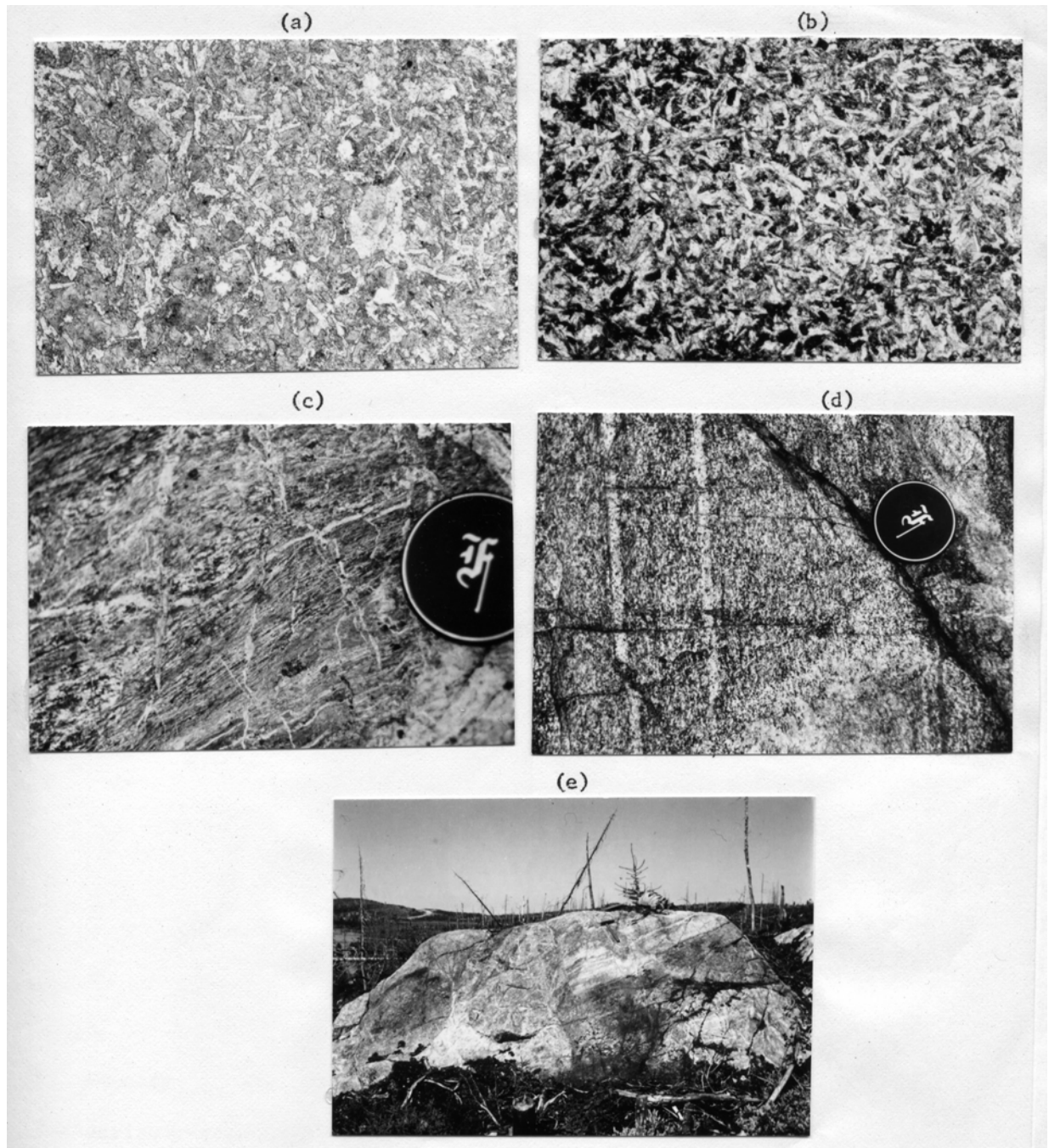
(a) Dyke in gabbro; Kidney Pond [34] (x 25)

(b) Dyke from sheeted dyke complex; 9.2 km N from Burlington Road Junction (x 25).

(c) Hot mylonitised gabbro; Kidney Pond [34].

(d) Normal weak gneissic foliation in gabbro with anorthosite bands. Kidney Pond [32].

(e) Gabbro with anorthosite bands as (d), cut by diabase dykes [32], Kidney Pond.

distributed. Their usual length/breadth ratio and the typical triangular interstices occupied by clinopyroxene that are defined by the abutment of three laths are absolutely characteristic of these ophiolite complex diabase dykes. This texture is quite distinct from dolerite (diabase) sills in the Baie Verte Group volcanics and sediments, described later. All diabase dykes in the map area are wholly altered. In those showing the original texture, the most common alteration has clinopyroxene wholly replaced by pale actinolite or chlorite and plagioclase wholly replaced by albite. In one example zoisite/clinozoisite entirely replaces the plagioclase, instead of albite. Minor and very small sphene aggregates represent a very minor amount of original ilmenite, and occasional small pyrite grains are present.

Diabase dykes often occur at intervals of around 2 to 3 metres in much of the ophiolite gabbro in the map area, although they are often more widely spaced and are not present in a minority of the outcrops. Around the northwestern corner of Flatwater Pond, especially in the two gabbro outcrops on either side of the narrows, diabase is found in more irregularly bounded masses, as well as in dykes. The reason for the irregular distribution in this area is not known. Diabase dykes cut cumulate gabbro at the southern end of the Flatwater body. The other specific examples of cumulate rocks described above are all presumed, although none were seen, to be cut by diabase dykes that occur in the same outcrops. Apart from the plastically folded dyke and the two dykes cut by a resealed plane in the gabbro, all other dykes cut banding, pegmatites and the widespread weak gneissic foliation, and they clearly fill brittle tension cracks in the gabbro. The relationship of diabase dykes to the other more zonal high temperature deformation fabrics is equivocal; some cut the zones, while others may be

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affected or involved in them. Their relationship to the trondhjemitic rocks is not known.

(ii)   Sheeted diabase dyke complex

At the northern end of the inland map area (Plate 2), there is a very poorly exposed area consisting, as far as the available outcrop shows, of more than 99% parallel diabase dykes. One screen of gabbro about 5 cm wide was seen between two dykes but, otherwise, the exposures consist entirely of diabase dykes. Poorly developed fine-grained margins are seen on one or both margins of the dykes, that are mostly about 20 cm wide, a width comparable to those in the gabbro. The texture and grain size range are essentially identical to the dykes in the gabbro. Again, it is not known whether this area of sheeted dykes, which is in an identical structural position to the gabbro bodies, is a single tectonically derived slice, a single 'boulder' or a multiple collection of boulders. Identical diabase, although not seen to consist of dykes, appears to be a large lensoid tectonic(?) inclusion in the thin body of ultramafic rock adjoining the east side of the area of sheeted dykes.

Both diabase dykes in gabbro and sheeted dyke complex occur in far larger, better exposed, and structurally intact bodies in the Mings Bight area described in the next chapter.

(vi) Summary

The variably-serpentinised Mg-rich ultramafic rocks of the map area are mostly harzburgite with minor dunite in small bands. Rare chromitite, orthopyroxenite and clinopyroxenite bands are also present. These rocks are typical of ophiolite complexes and, by inference, were formed as upper oceanic mantle. A large minority of the rocks show a high-temperature foliation parallel to banding that was probably sub-horizontal when formed.

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If so, about 2.4 km thickness of these rocks are preserved in the main part of the Flatwater body. CO_2-bearing microinclusions in the olivine are probably the source for the common carbonate mineral grains, seen in highly serpentinised rocks, and Mg-carbonate rock found on certain tectonic movement zones. This source is also thought to be responsible for small tuffisite pipes and dykes that cut the Mic Mac body; the particular examples seen probably formed late during tectonic emplacement. Disruptive tectonic fabrics are confined to a thin zone around the margins and to small faults within the ultramafic bodies. Serpentinisation is most complete in a zone subparallel to the outlines of the bodies, wider on the eastern compared with the western side. Early selective alteration of enstatite, rather than olivine, in the fresher rocks indicates ingress of water when temperatures exceeded 850°C.

Cumulate ultramafic rocks are inferred to form a thin transition zone to gabbro at the southeastern corner of the Flatwater body. Other occurrences of such rocks are inferred to be derived from such a transition zone. The ophiolite complex gabbro found in the inland map area, with the exception of that at the southern end of the Flatwater body, is not in its original position in an ophiolite complex sequence. The gabbro in the southern part of the area is a collection of closely-packed enormous boulders. In the northern part of the area, some are tectonically derived slices on slide zones, a few are boulders and most are probably one or the other of these alternatives. The bulk of the gabbro is a homogenous feldspathic (leucocratic) clinopyroxene gabbro lacking ore minerals. Some hornblende gabbro, minor mafic gabbro, and rare clinopyroxenite, picrite and anorthosite bands are found. A few samples of the gabbro are cumulate, including some hornblende gabbro and disrupted anorthosite bands but most is not provably cumulate

in its present condition. A large minority of the gabbro has a weak gneissic high-temperature foliation, parallel to rare probably cumulate banding, with accompanying granular recrystallisation. Uncommon narrow bands, with more intense high-temperature foliation and finer-grained recrystallisation, are hot mylonite zones. Recrystallisation to hornblende or a Ti-rich amphibole occurred on these narrow zones while the surrounding little-deformed gabbro retained clinopyroxene. There appears to be a sequence from more common zones many metres wide, with weak foliation and diffuse boundaries, to rare narrow zones a few centimetres wide with sharp boundaries and intense foliation. This sequence is probably a function of decreasing temperature, but also reflects access of water into the narrower zones. These high-temperature foliations in the gabbros are directly comparable with those in the depleted ultramafic rocks. The entry of water vapour at high temperatures is indicated both in the gabbros and in the ultramafic rocks; it probably reflects faulting of the ophiolite complex (oceanic crust and upper mantle) shortly after its formation at a spreading ridge axis, probably by a transform fault or faults.

It is suggested that uncommon tuffisite breccias of gabbro were generated by release of CO_2 contained in the microinclusions in ultramafic rocks. In order to release sufficient CO_2, introduction of water and local serpentinisation of underlying ultramafic rocks is probably needed. Therefore the CO_2 was probably mixed with water vapour. If so, this is further tentative evidence of faulting of the oceanic crust. This faulting must be prior to or, more likely, coeval with the deposition of the gabbro (including the tuffisite) as large boulders in and below Baie Verte Group sediments and volcanics, which are described next.

Parallel diabase dykes are found cutting most of the exposures of

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ophiolite gabbro. Only two examples of dykes are slightly affected by high-temperature deformation. Cumulate gabbros are cut by some diabase dykes. One small area of sheeted diabase dyke complex is found in the inland map area but it is tectonically isolated from other ophiolite complex lithologies.

Pillow lava belonging to the ophiolite complex has not been identified in the inland map area. All pillow lava in the Baie Verte Group occurs conformably above sediments containing ophiolite gabbro and diabase clasts, and therefore is not part of the ophiolite complex.

CHAPTER 4B. BAIE VERTE GROUP VOLCANICS AND SEDIMENTS

(i) Introduction

In the inland map area, the Baie Verte Group mostly consists of volcanics and sediments that occupy a linear belt from 1 to 3 km across (Plate 6). All these rocks are submarine, and almost all are mafic lavas and mafic volcaniclastics. They form a homoclinal sequence, strike parallel to the Lineament, and dip subvertically at the western side changing gradually to moderate westerly dips at the eastern side. They are affected by a single penetrative cleavage that is everywhere essentially coplanar with bedding. The amount of flattening and subordinate predominantly subvertical elongation deformation associated with this cleavage is variable; least in the south, between Mic Mac Lake and Kidney Pond, becoming greater to the north. All quoted thicknesses are uncorrected for the deformation. The amount of deformation not usually measurable but, in general, thicknesses are estimated to be reduced from their original values by at least 50%.

The stratigraphy proposed here is the first subdivision of the Baie Verte Group attempted in this area. The North American Stratigraphic Code (1961) Is followed, but some qualifying statements about the necessarily cut and dried nature of the stratigraphic definitions are needed because of some properties of the lithological assemblage, the deformation and the outcrop in the area.

1) Stratigraphic subdivision of a relatively strongly-deformed suite of rocks like the Baie Verte Group can only be made with the proviso that undected tectonic slides nearly coplanar with bedding may be present within the apparent stratigraphy. These may remove or duplicate significant thicknesses of the original sequence. This may be undetectable in a sequence like the Baie Verte Group which mostly consists of a very few repeated lithologies.

Tectonic slides with large displacements may occupy very narrow zones merely a few metres across. If they are not exposed, it is very unlikely that their effects can be differentiated from lensing of units and facies changes. In the particular case of Baie Verte Group, in the southern part of the area between Mic Mac Lake and Kidney Pond, it is thought that there are no significant tectonic slides present within the sequence, but this cannot be proved in the absence of at least one section with near 100% exposure.

2) A stratigraphy set up for a particular section in a sequence of rocks like the Baie Verte Group, which consists of repetitions of a very few related lithologies, is much more likely to be less precisely correlatable or even not correlatable along strike than in a more varied sequence. For example, the main pillow lava unit in the Baie Verte Group separates two broadly similar units of mafic volcaniclastic rocks. The pillow lava unit both thins and is tectonically cut out to the north. If it disappears in the relatively poorly exposed area (not mapped) to the north, it will be impossible to separate the two mafic volcaniclastic units.

3) Most of the proposed units rely on the appearance or disappearance of a particular lithology to define their boundaries. In areas of poor exposure, especially where the units are involved in lensing or intertonging and facies change, a boundary may depend on one or two outcrops of the diagnostic lithology out of several times this number of outcrops of undiagnostic lithologies. Where there is also uncertainty in identification of the particular diagnostic lithology, in areas that have small outcrops of strongly-deformed rock, this problem is compounded. In the Baie Verte Group, the lower boundary of the main pillow lava unit west and north of Flatwater Pond is uncertain for both these reasons.

4) The relatively poor overall exposure and small outcrops typical of

the Baie Verte Group in this area mean that it is usually impossible to define a 'linear' type section that exposes anywhere near all of any particular Formation, or one that is much superior in total exposure to any other. Actual exposures of the top and bottom of most of the units are rare. Where they occur, they are not usually found where it is most convenient to designate a representative, relatively well-exposed type section. For the Baie Verte Group in this area, it is my opinion that it is more useful in most cases to designate a broad swath across any unit where the exposure is relatively good as a type area. This will result in a larger proportion of the total thickness of a unit being exposed in the type 'section'. The stratigraphic units in the area were not defined by mapping a type section and extrapolating the defined units. They only became clearly apparent when a large proportion of the total area had been mapped. This shows why type section (*sensu stricto*) definition is inappropriate for this area of the Baie Verte Group. Theoretically, definition of a type 'swath' with finite dimensions along strike could lead to squabbles over the precise position of a bed within the Formation. Such arguments against this procedure in the particular case of the Baie Verte Group in this area are trivial, compared with the need to establish a representative section that is of practical use, as long as the section or swath is precisely and clearly defined.

5) Geographic names are very scarce in the area. Several have been proposed by the author merely to name stratigraphic units. They may or may not prove acceptable to the Canadian Committee on Geographic Names. Names of stratigraphic units also may or may not be acceptable when submitted to the Committee on Stratigraphic Nomenclature before they are formally proposed.

Mineral assemblages throughout the volcanics and sediments are those

characteristic of the lowest greenschist facies and are described later. However, this means that in almost all cases, the primary mineralogy of the volcanics, and of matrix and clasts in the sediments has been destroyed. Primary microscopic textures in the volcanic rocks are often wholly destroyed. Recrystallisation of the fine-grained matrix of the sediments obliterates its original texture; rarely-observed grading through silty to argillaceous layers shows that a gross resemblance of original and present grain size remains, at least in some places. In the less deformed clastic sediments, clastic grains from coarse silt to fine sand size upward are pseudomorphically replaced and preserve some of their original relationships. All but a tiny proportion of the sediments are mafic volcaniclastics; that is clastic rocks ranging from the very finest silts to conglomerates derived and deposited in unspecified ways from either or both of basaltic and andesitic sources. The volcaniclastics appear poorly-sorted in most cases. Clearly graded beds are rare, although recrystallisation of the fine-grained matrix may have obscured more widespread poorly-developed grading. Bottom (sole) structures are almost non-existent, although the type of outcrop is not advantageous for their detection. In this thesis, the volcaniclastic rocks are described as silty, sandy, gritty or conglomeratic, referring to the coarsest clasts in the beds, using the usual definitions for the sizes of clasts (Wentworth scale). The generally poor sorting means that a large proportion of the clastic material will have been finer grained than the size specified. In the thin sections examined, the clastic mineral grains in the volcaniclastics are mostly albitised plagioclase. Few beds contain visible clastic mafic mineral grains, and volcaniclastic rocks can be assumed to contain only altered plagioclase clastic grains unless it is stated otherwise. Clastic quartz is essentially absent from the mafic volcaniclastic

rocks in the inland map area. A few grains are seen in very rare instances. All detected

occurrences of clastic quartz and clasts containing quartz are specifically mentioned in the

appropriate section. Argillaceous rocks are a minor component of the sediments. Weakly-

cleaved, usually banded argillaceous rocks are called argillites or cherty argillites; more strongly

cleaved, usually thicker and homogenous argillaceous rocks are called slate. Flinty siliceous

argillite and chert are a minor component of the bedded sediments, but are common as interstitial

fillings between pillows within flows. Dolerite and minor gabbro are common as sills throughout

the sequence. They are described separately from the stratigraphy, although they are probably

intimately related to the mafic vulcanism. Stratigraphic thicknesses quoted include these sills, as

there is no objective way to remove them, in the absence of total exposure. The maximum

proportion occupied by sills is estimated to be about 25% of the total thickness (the section east

from Kidney Pond). Elsewhere, the proportion appears to be 15% or less.

The stratigraphic subdivisions of the Baie Verte Group of the map area are set out first as

summaries of definitions and essential data. Detailed descriptions, of various aspects of the

rocks, together with discussion and interpretations are reserved to following sections. Rocks

mapped beyond 6 km north of the Burlington Road Junction cannot be correlated with this

stratigraphy until the intervening ground is mapped. They are therefore described separately.

(ii) Stratigraphy

(a) Boudin Pond Ophiolite Gabbro Megabreccia Formation.

A small pond just southwest of Slink Pond (Plate 1) was named for the purpose of

naming the Formation. The Formation consists of closely-packed enormous boulders derived

entirely from the gabbro layer of an ophiolite

Fig. 4B.1. Type section of the Boudin Pond Ophiolite Gabbro Megabreccia Formation.
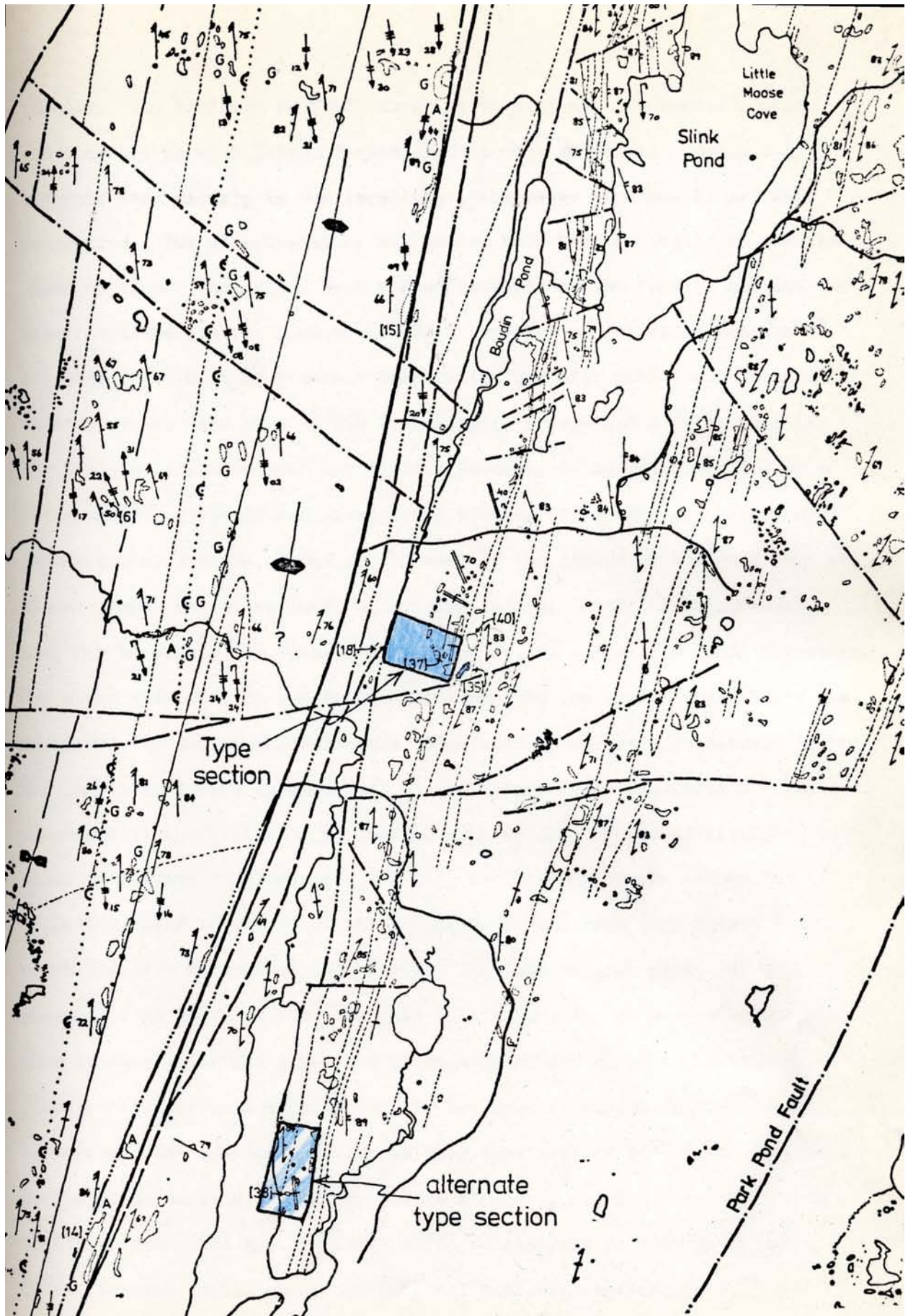
complex. The boulders probably range up to at least 100 metres across and some may be much larger. Bedded argillaceous sediments between them are only seen clearly in two localities, the range of sizes is not well controlled. The lithologies of the gabbro have been described in the previous section. They range from clinopyroxenite to anorthosite and include gas-brecciated gabbro, gabbro with high temperature foliation and minor trondhjemite. Most is somewhat leucocratic granular gabbro cut by parallel diabase dykes. The base of the Formation is everywhere a tectonic slide, the Western Lineament Boundary Slide. However, it is possible that three outcrops of a strongly but irregularly cleaved green 'gritty' rock with abundant quartz veins, found to the west of the gabbro at the north end of Mic Mac Lake, may be sediment underlying the Formation. Alternatively, they may be within the Formation, or they may be a tectonic slide lithology, but as no thin section was seen, this cannot be resolved. From Kidney Pond southward the top of the Formation is defined by the contact between gabbro and pebble to cobble conglomerate of the Kidney Pond Conglomerate Formation. Occasionally the gabbro is in contact with up to 2 metres of argillite and slate underlying conglomerate, and this is included in the Kidney Pond Formation. The thickness of the Formation in the area from Kidney Pond southward is a minimum of 200 metres. The type section (Fig. 4B.l) is chosen 1.1 km north of the north end of Mic Mac Lake to emphasize the very rare evidence that the gabbro is in megaboulders. It is not chosen to include typical or well exposed gabbro, nor does it expose the top of the Formation. The critical outcrop in this type section [37] shows about 1.3 metres of normally striking, vertically dipping green and minor grey banded argillite with fine-grained sandy mafic volcaniclastic beds up to 10 cm thick, between gabbro which has vertical contacts conformable with the sediments. The top 10 cm of the sediment consists of pebble-size clasts

of green argillite with a few pebbles of gabbro and diabase in a green argillite matrix. A crack in the overlying gabbro subparallel with bedding is seen filled with the pebbly argillite but soft sediment deformation of the other sediment beds is not apparent. A few metres to the east in this outcrop some similar sediment is seen, but its relationship to the gabbro is not well-exposed. These sediment occurrences are about 50 metres below the top of the formation. It is possible that this outcrop will be covered or ruined by activities connected with a farming experiment just started in this immediate area. If so, an alternate type section could be designated 0.55 km south of the north end of Mic Mac Lake (Fig. 4B.1). An outcrop in this section [38] is the only other in the area that shows unequivocal sediment among gabbros. About 2.5 metres of green and grey argillite and slate occur, and some of the grey slate contains black cherty argillite clasts. Otherwise, the sediments are not affected by soft sediment deformation. The contacts of the sediment with the gabbro are not well-exposed. Elsewhere, there is no other direct evidence that the gabbros are in large blocks, except for one outcrop where a thin lens of material could be deformed pebble conglomerate and another outcrop where the gabbro was suspected to be in disjunct blocks because of interrupted banding. Above the type section designated, a 50 x 20 metre section through a gabbro megaboulder is exposed [35] in the overlying Kidney Pond Conglomerate Formation. This fortuitous occurrence clearly shows that very large gabbro boulders were deposited at this time.

The ophiolite gabbro occurring around 1.3 km north of the outlet of Kidney Pond probably belongs to the Formation, but the shear polyhedra ultramafic rock may be a tectonic sliver introduced along the Western Lineament Boundary slide. It is not clear whether ophiolite gabbro areas

to the north of the Old Camp 32 Road belong to this Formation or are tectonically-introduced slices. A few centimetres of sandy mafic volcaniclastic rock, seen within gabbro on the point on the south side of the narrows of northwestern Flatwater Pond, could be in place, or may have been tectonically introduced. It is possible that each area is an enormous single 'clast'. If they are part of the Boudin Pond Formation, its top contact may also be defined by contact of gabbro with green and grey slaty argillite and sandy and silty mafic volcaniclastics of the Teardrop Pond Formation, which underlies the Kidney Pond Conglomerate Formation in the area north of Kidney Pond. The existence of at least one large boulder of ophiolite-derived cumulate clinopyroxenite, a large boulder of marble, and slaty argillites with argillite pebbles, within the Teardrop Pond Formation, perhaps suggests that the Boudin Pond Formation may be partially or wholly a facies equivalent of the Teardrop Pond Formation but it is not possible to prove this. The nature of abrupt contacts strongly oblique to the normal strike, in particular those between the ophiolite gabbro and the Teardrop Pond Formation 1.4 km N of the outlet of Kidney Pond, and 0.8 km N of the Burlington Road Junction, is not known. If they are faults they predate deposition of the Kidney Pond Conglomerate Formation. It is possible that they are normal sedimentary contacts on the sides of single enormous boulders, or heaps of smaller boulders, but it is not possible to state the order of deposition. If the gabbro bodies north of the Old Camp 32 Road are part of the Formation, the sheared ultramafic lenses within them may be either clasts or slivers introduced on faults prior to deposition as enormous blocks.

(b) Teardrop Pond Formation.

This Formation is named after a small pond 1.4 km N of the Burlington

Fig. 4B.2. Type section of the Teardrop Pond Formation.
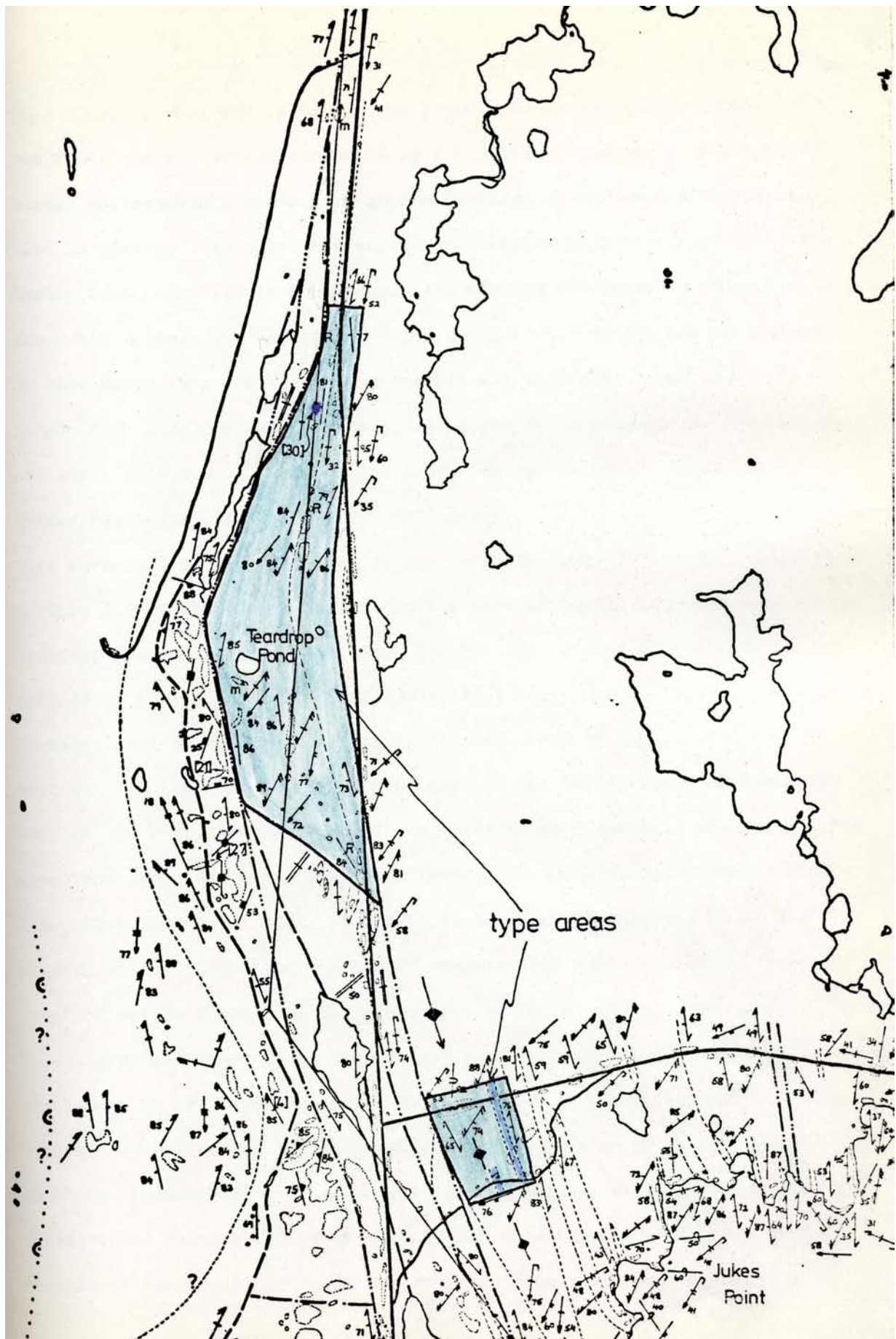
Road Junction that was named for the purpose of naming the Formation. The lower contact of the Formation is a tectonic boundary; it may be a normal conformable contact with gabbro boulders of the Boudin Pond Formation in places. Its upper contact is conformable with conglomerate of the Kidney Pond Conglomerate Formation. Its minimum thickness is estimated as about 550 metres, but undetected folds and tectonic slides may be present. In some ways, this Formation is a dustbin for sediments found under the Kidney Pond Conglomerate Formation, and correlation between the four separate areas in which it is found is poor. It may be partially or wholly a facies equivalent of the Boudin Pond Formation. A 'linear' type section for this Formation is impossible to define if it is to be of any use. Therefore a swath 1.6 km wide north of the Burlington Road Junction in the area around Teardrop Pond and the section along the Burlington Road (Fig. 4B.2) is defined as a type section. This includes at least one tectonic slide, and an anticlinal fold closure is interpreted to occur in the Burlington Road section. In the type 'section' northwest of the Burlington Road Junction, most of the rocks are sandy and silty mafic volcaniclastics, and subordinate green and grey, and minor black and maroon, banded argillite, cherty argillite chert and slate. Beds of mafic volcaniclastics range up to at least several metres thick, and the banded argillaceous beds form units up to at least 10 metres thick. On the power line southeast of Teardrop Pond, black, grey and green slate with fairly abundant pebble-size slaty argillite clasts are at least 10 metres thick. A boulder of clinopyroxenite at least 5 metres across occurs in silty volcaniclastic rock on the power line 0.8 km NE of Teardrop Pond. The underlying silty volcaniclastic rock is mainly resedimented pebbles, although this is not at all obvious in outcrop. The section on the Burlington Road is composed mainly of sandy mafic volcaniclastics

with minor banded green cherty argillite and black slate. Some of the thinnest volcaniclastic beds are well-graded but this can only be seen clearly in thin section. A minority of the clasts are mafic minerals (now actinolite) in this section of rocks. The contact with the Kidney Pond Conglomerate Formation is exposed in an outcrop between the Burlington Road and an abandoned section of the Old Baie Verte Road to the south. The polymictic black slate matrix conglomerate overlies rather nondescript mafic volcaniclastic rocks that may either be resedimented and conglomeratic, or else include some pillow breccia. Marble occurs in three localities in the Teardrop Pond Formation. The outcrop immediately south of Teardrop Pond contains about 10 metres of microbrecciated homogenous grey marble that is almost certainly a large boulder, overlain by a metre of calcareous black slate, within homogenous sandy mafic volcaniclastic rock. About 2 metres of banded calcarenite marble with black slate interbeds is found under the power line 3.0 km N of the Burlington Road Junction. Several beds of banded calcarenite marble are found 0.18 km south of the point on the south side of the narrows on the western shore of Flatwater Pond. These beds are in black and grey slate, and most are 7-10 cm thick. Two beds also contain lenses of pebbly calcirudite marble and are about 60 cm thick. This tectonically isolated section of the Teardrop Pond Formation consists of sandy and silty mafic volcaniclastic beds and minor grey and green slate and argillite to the north west of the marble, including another possible outcrop of calcirudite marble. Similar rocks form the first part of the section to the southeast of the marble, and are succeeded by a (relatively) thick section of black slate with occasional green chert laminae. The strip of Teardrop Pond Formation running from north of the Bear Cove Road to 1.5 km north of Kidney Pond is composed of nondescript mafic volcaniclastic rocks, probably mainly highly-deformed sandy mafic volcaniclastics, pumiceous conglomeratic

mafic volcaniclastics and disrupted green cherty argillite laminae. Precise identification of the lithologies in this strip was in most cases impossible. These rocks conformably underlie a very thin development of the Kidney Pond Conglomerate Formation.

(c) Kidney Pond Conglomerate Formation.

This Formation is apparently continuous over a strike length of 24 km, while having a maximum thickness of 60 metres. In the stretch between 1.6 km N of the outlet of Kidney Pond and its next crossing of the Baie Verte Road near Flatwater Pond, the only two exposures of the Formation show about 1 metre and 40 cm total thickness. Elsewhere the thickness is mostly between 20 and 50 metres. The Formation consists almost entirely of polymictic conglomerate. Most of the clasts are between about 2 mm and 15 cm across, and boulders up to about 3 metres across are not uncommon. Clasts larger than 3 metres across are very rare but the largest boulder seen has an exposed section of 50 x 20 metres. Conglomerate with clasts supported in a homogenous black slaty matrix is found over the whole strike length mapped. The clasts are flattened in the cleavage, the amount of flattening depending on the local intensity of deformation and the relative ductility of the different clasts. Most of the rest of the Formation consists of conglomerate with a green sandy or silty matrix with clasts forming a self-supporting framework. This always occurs underneath black slaty matrix conglomerate, and is only found southward from 1.3 km north of Kidney Pond outlet. It is also not present opposite the northern part of Kidney Pond, and in the outcrops immediately south of Kidney Pond. Clasts in this conglomerate are not deformed as much as those in black slaty matrix conglomerate. The clast assemblage in the conglomerates is dominated by ophiolite complex-derived gabbroic lithologies, with subordinate diabase, mafic volcanic rock, and

Fig. 4B.3. Type section of the Kidney Pond Conglomerate Formation.
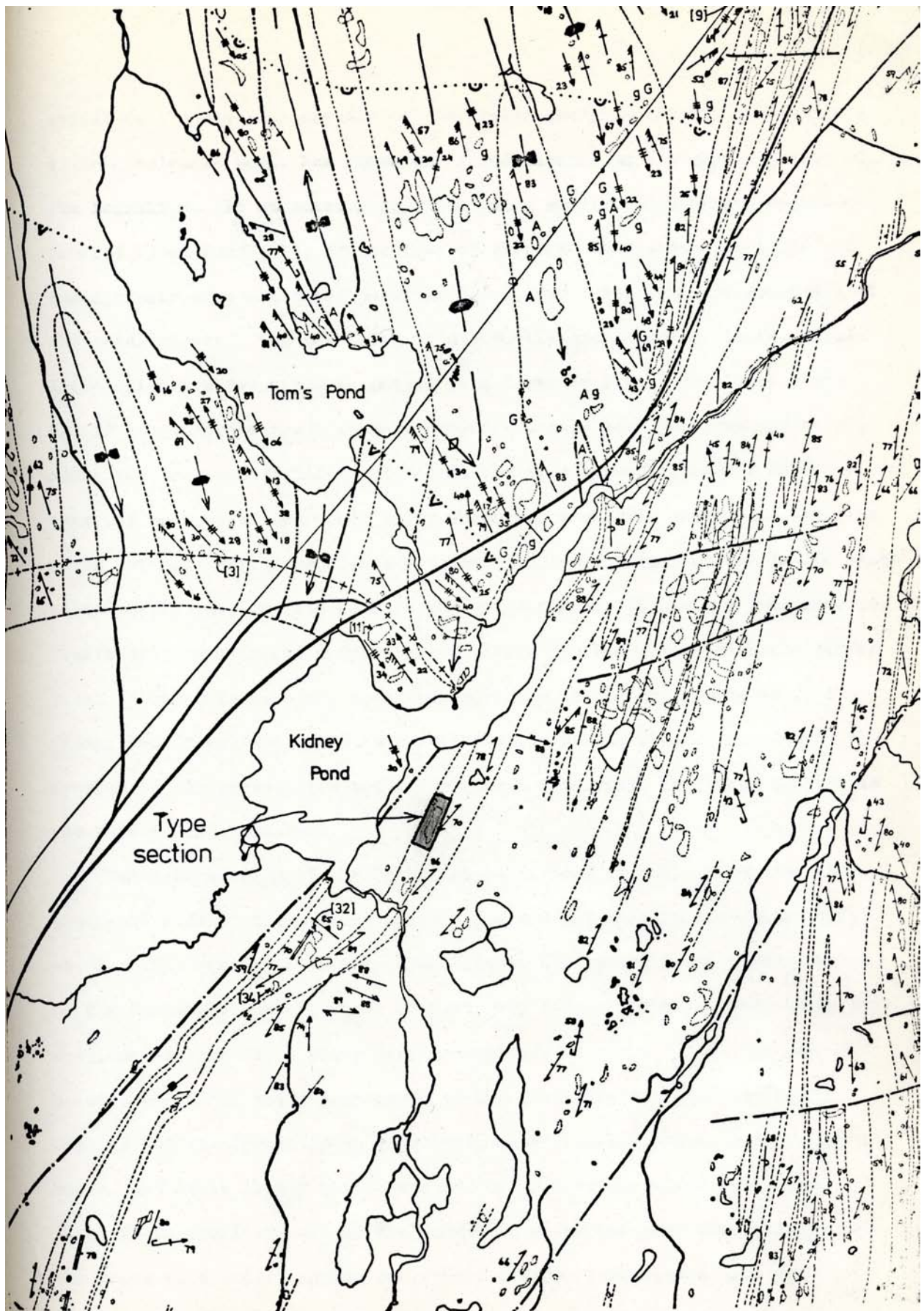
argillite. A minor proportion of the clasts are granodiorite and foliated silicic volcanic rock, but these are found throughout the mapped extent of the Formation. In an outcrop of green sandy matrix conglomerate opposite central Slink Pond [39], proportions of gabbroic to diabase and mafic volcanic clasts were estimated as about 2:1, and these comprise about 95% of the total clasts. The remaining 5% are mostly granodiorite clasts. Clast lithologies are described in detail in a later section. The black slate matrix conglomerate tends to have proportionately less gabbroic and more mafic volcanic and argillite clasts in the area around Flatwater Pond and this may be true of the whole unit, which has a uniform appearance over its whole extent. Argillite clasts are generally much more abundant in the black slate matrix conglomerate than in the green sandy/silty matrix conglomerate. Clasts of serpentinite, serpentinised olivine/orthopyroxene rich ultramafic rock, or chromite have not been seen anywhere in the conglomerates of the Kidney Pond Formation. All rusty ultramafic-looking clasts were found to be altered clinopyroxenite and mafic gabbro (picrite), identical to that in the Boudin Pond Formation.

The type section of the Formation is defined as the outcrops at the southeast side of Kidney Pond between 1.0 and 1.2 km from the outlet (Fig. 4B.3). This section does not expose either the upper or the lower contact of the Formation, but it shows in part very little deformed green sandy matrix conglomerate and black slaty matrix conglomerate above this. The contact between these two major components of the Formation is never completely exposed but it appears to be relatively sharp, taking place in less than a metre, and it is likely that these two units were deposited separately. Black slaty argillite clasts are common in a few outcrops near the base of the green sandy/silty matrix conglomerate from Slink Pond southward. In outcrops southward from 1.6 km N of Kidney Pond outlet, the top of the

black slaty matrix conglomerate unit almost always shows an upward gradation into clast-poor and often into clast-free black slate. This is not seen north of this point, where clasts persist to the top of this unit and the Formation. Just north of the north end of Mic Mac Lake, one outcrop shows sandy mafic volcaniclastic pebbles and cobbles in black slate near the top of the Formation. Apart from this example, mafic volcaniclastic clasts have not been seen in the Formation. The exposure of the Formation just south of the Bear Cove Road shows two beds of black slaty matrix conglomerate each about 15 cm. thick separated by about 10 cm of fine-grained sandy mafic volcaniclastic rock and minor green argillite. At the southern end of Slink Pond, the usual clast-poor top of the black slaty matrix conglomerate is overlain by a few metres. of sandy mafic volcaniclastic and banded green cherty argillite, that are in turn overlain by 1.6 metres of pebbly black slate-matrix conglomerate. Interbeds of mafic volcaniclastic rocks are also seen within the black slaty clast-poor top of the Formation in the roadcut 1.4 km N of Kidney Pond, and possibly on the north shore of Flatwater Pond, but have not been seen elsewhere. It is likely that the black slaty matrix conglomerate was deposited in several events in at least some places. Evidence of slumping is seen in an outcrop on the northern shore of Flatwater Pond and in an outcrop 1.2 km N of the north end of Mic Mac Lake where large bedded clasts of black slate with silty and sandy greywacke display slump folds.

The contacts of the Kidney Pond Conglomerate Formation are conformable everywhere except on the western limb of the fold pair around the northwestern corner of Flatwater Pond. The lower contact with the Teardrop Pond Formation is defined by the incoming of conglomerate. The lower contact with the Boudin Pond Formation is defined by the contact between either gabbro and conglomerate or gabbro and argillaceous sediment. In some places,

up to 2 metres of banded argillite, or slate, or slate containing black argillite pebbles may underlie conglomerate and be in contact with gabbro. These are included in the Kidney Pond Formation. The upper contact of the Formation with the Jukes Point Formation is defined by the contact of the uppermost black slaty matrix conglomerate or black slate over this conglomerate with mafic volcaniclastic rocks or banded green argillite. In places, between Kidney Pond and Mic Mac Lake, the Formation has been shown on the map as in contact with the Slink Pond Pillow Lava Formation because there is no outcrop evidence, in these places, for the thin development of the Jukes Point Formation. However, pillow lava is never seen in direct contact with the black slate or black slaty matrix conglomerate. In these stretches where the Jukes Point Formation is not marked, there is probably always at least a few metres of volcaniclastic and argillaceous sediment that would be ascribed to the Jukes Point Formation if it were seen in outcrop.

The conglomerate of the Kidney Pond Formation is one of only two distinctive and extensively traceable marker units in the Baie Verte Group of the map area. As such, it plays a disproportionately large role in the definition and tracing of the stratigraphy. The conglomerate with quartzite and quartz-pebble-conglomerate boulders, found on either side of the Baie Verte Road 7.8 km N of the Burlington Road Junction, may be approximately equivalent to the Kidney Pond Conglomerate Formation. It is described with the detailed description of some aspects of the conglomerate of the Kidney Pond Formation following description of the remainder of the Baie Verte Group stratigraphy.

(d) Jukes Point Formation

This Formation is named after a feature on the northern shoreline of Flatwater Pond within the type section. This feature was named for this purpose after the first person to make and record geological observations

Fig. 4B.4. Type section of the Jukes Point Formation.
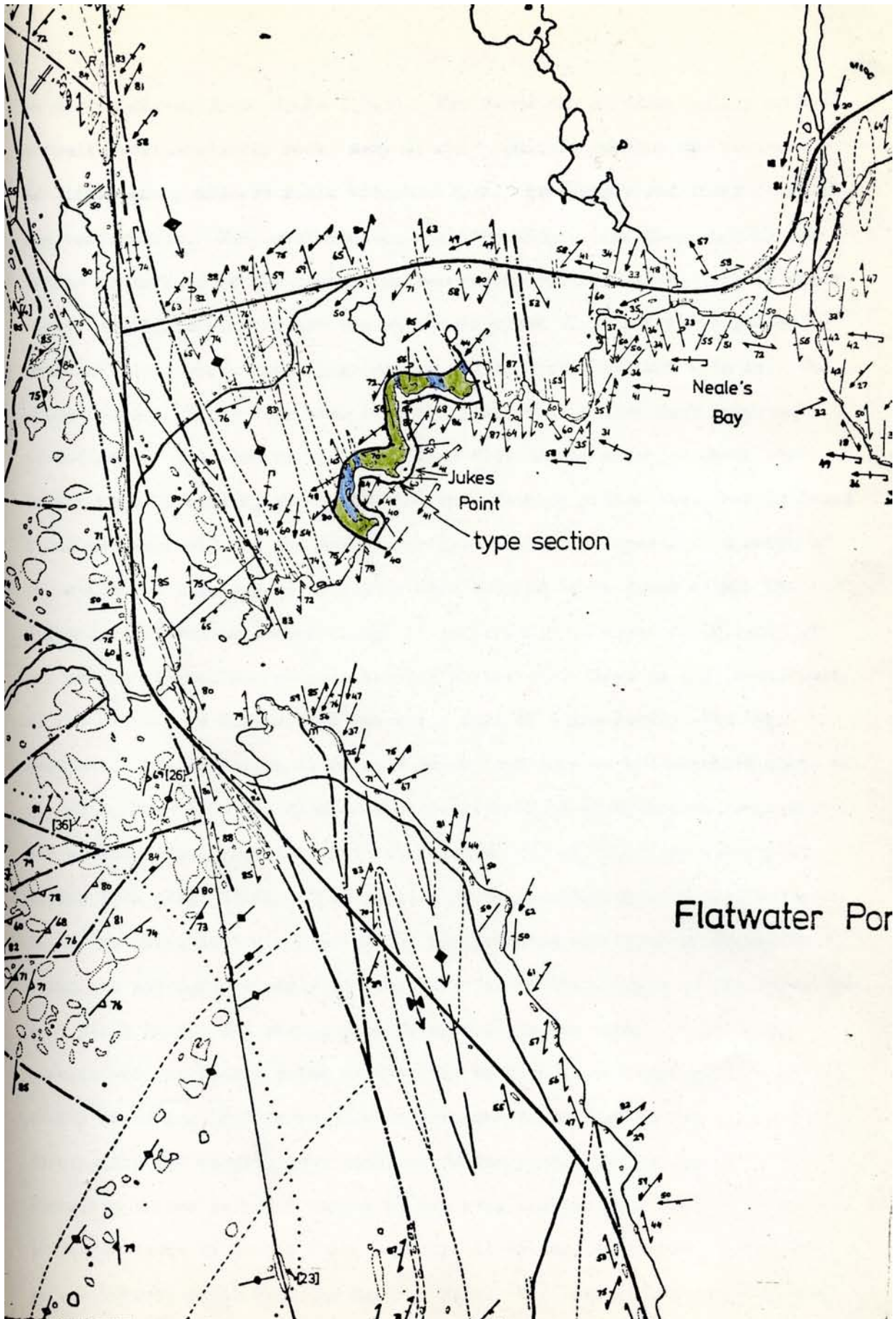
in Newfoundland, J. B. Jukes (1843). The Formation consists mainly of beds of mafic volcaniclastic rock, many of which contain pebbles and boulders of lithic and pumiceous mafic volcanic rock. Both upper and lower contacts are conformable. The lower contact with the Kidney Pond Conglomerate Formation is defined by the contact between mafic volcaniclastic rock or green cherty argillite with either the uppermost black slaty matrix conglomerate or black slate above this conglomerate and in direct contact with it. The upper contact of the Formation with the Slink Pond Pillow Lava Formation is defined by the base of the first lava flow in the sequence above the Kidney Pond Formation, except for the thin lens of pillow lava that is found along the west side of the Baie Verte Road between 1.1 and 1.7 km north of the outlet of Kidney Pond (which is here defined as included within the Jukes Point Formation, even though it is probably an exact correlative of one of the plagioclase phenocryst-rich pillow lava flows in the lowest part of the Slink Pond Pillow Lava Formation west of Slink Pond). The type section of the Formation is defined as the outcrops on the northern shore of Flatwater Pond from the exposed contact with black slaty matrix conglomerate of the Kidney Pond Formation eastward as far as the first outcrop of pillow lava (Fig. 4B.4). This section is representative of the lithologies in the thicker, northern part of the Formation as far south as Kidney Pond, which are all more strongly deformed than in the thin strip of the Formation past Slink Pond. The strong elongation deformation shown by the conglomeratic volcaniclastic rocks in the type section is not typical of the rest of the Formation, but it results in sections orthogonal to the elongation deformation not showing much relative deformation. The thickness of the Formation varies from 430 metres in the type section to a minimum of about 10 metres south of Kidney Pond, although it is usually between 15 and 25 metres between Slink Pond and Mic Mac Lake. The Formation north of Kidney

Pond is mostly a time equivalent of the lower part of the Slink Pond Pillow Lava Formation in the Slink Pond area but the lithic and pumiceous conglomeratic mafic volcaniclastics are not pillow breccias, or otherwise directly related to the pillow lavas of that Formation.

In the type section, about 60 metres of sandy and silty mafic volcaniclastics in beds from a centimetre to a few metres thick, with minor thin (cm-mm) beds of green cherty argillite forming the tops of uncommon thin graded beds, occur below the first conglomeratic bed. The next 270 metres thickness contain abundant beds of conglomeratic mafic volcaniclastics. The beds are typically 20 cm to 2 metres thick and are sometimes graded, although grading is not often seen elsewhere in the Formation. Lithic clasts up to 30 cm across are mostly supported by the sandy to gritty mafic volcaniclastic matrix. Interbeds of sandy to silty mafic volcaniclastic rock without large clasts are commonly between about 10 and 20 cm thick. Sole structures have not been seen and all the bedding in this Formation appears to be planar. Clasts are essentially all wholly-recrystallised mafic volcanic rocks, often moderately amygdaloidal with calcite fillings. Less than five white siliceous clasts not more than 3 cm across are of trachyte. In the section of this Formation on the western shore of Flatwater Pond, some apparently sandy mafic volcaniclastic beds have a characteristic laminated appearance due to slight colour banding. This lamination is lenticular, and is thought to represent exceedingly flattened mafic pumiceous clasts, because where lithic clasts occur in the same bed, as in the outcrop on the shore of Flatwater Pond 0.24 km NNE of the Bear Cove Road Junction, they are not very flattened. The upper 100 metres of the Formation in the type section does not contain many beds with conglomeratic clasts and consists mostly of sandy mafic volcaniclastic rock.

Most of the crystal fragments in the mafic volcaniclastics are altered plagioclase, except for a few beds in the lower part of the Formation, exposed on the northern and western shore of Flatwater Pond, that contain a significant amount of actinolitised mafic crystals as well as altered plagioclase. The outcrop by the waterfall at the outlet of Kidney Pond, often visited on field trips as it is next to the road, is part of the lower portion of the Jukes Point Formation. It shows beds of sandy to silty mafic volcaniclastics with some green, white-weathering chert beds up to 1 cm thick. It is thus more typical of the thin southern part of the Jukes Point Formation. Beds containing conglomeratic clasts occur some way to the northeast of this outcrop. On the south side of the waterfall, poorly exposed at the western side of the outcrop, is a strongly cleaved and kinked grey iron-stained gritty rock. It contains grit-sized clastic quartz and albitised plagioclase in equal amount and also contains much (recrystallised) calcite. This quartz-rich greywacke-type rock has not been seen elsewhere in the Jukes Point Formation.

Between Slink Pond and Mic Mac Lake, this Formation consists of: 1) relatively little-deformed sandy and silty mafic volcaniclastic beds from less than a centimetre up to about 1 metre thick; 2) thin beds of cherty argillite and chert a few millimetres thick; and 3) beds between about 50 cm and 2 metres thick of sandy mafic volcaniclastics with matrix supported pebbles and boulders up to 30 cm, commonly 10 cm or less across, of any one, two or all of sandy mafic volcaniclastic rock, banded argillite and cherty argillite, and mafic volcanic rock. The mafic volcanic clasts are often amygdaloidal, and occasionally porphyritic (plagioclase), in contrast with the mafic volcanic clasts in the underlying Kidney Pond Formation, which are neither porphyritic or amygdaloidal. Cherty argillite clasts are always equidimensional and somewhat rounded and sometimes plastically

deformed, indicating that the sediment was soft when eroded, except for one example where brittle flat flakes of cherty argillite were seen. A very few clasts of doleritic and ophiolite gabbroic rock were seen, but granodiorite or silicic volcanic clasts are not present. Identical beds are found between the first few pillow lava flows of the overlying Slink Pond Formation up to about 70 metres above its lower contact, but only in this same area from 2.6 km south of Kidney Pond outlet to Mic Mac Lake. The mafic volcanic clasts tend to be fairly well-rounded. They are not pieces of pillows or pillow breccia. These rocks are little deformed and relatively well-exposed. The Jukes Point Formation in this area strongly resembles the well-exposed sediments directly overlying ophiolite complex pillow lavas on the western side of Mings Bight (Plate 6).

(e) Slink Pond Pillow Lava Formation.

This Formation is named after the large slim pond north of Mic Mac Lake. It is mostly composed of pillow lava, with minor very vesicular massive lava, and rare pillow agglomerate. A minor proportion (~5-10%) consists of sandy and silty mafic volcaniclastic beds with rare beds and units of banded green, and very rare grey-black, cherty argillite, forming sediment units mostly less than 10 metres thick between some flows. Green white-weathering chert is only found as interstitial fillings between pillows but very rare banded maroon chert is found in beds. The base of the Formation is defined by the base of the first pillow lava flow above the Kidney Pond Formation, with the one exception defined in the previous section. The top of the Formation is defined in general terms by a clear change from a sequence consisting mainly of pillow lava to one consisting mainly of mafic volcaniclastic rocks. In the area west and north of Flatwater Pond this is defined as the top of the uppermost pillow lava in the sequence. In the area southward from 1 km north of the Camp 166 Road,

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Fig. 4B.5. Type section of the Slink Pond Pillow Lava Formation.

White
Bay
Waters

Little
Moose
Cove

Slink
Pond

boundary of
type area
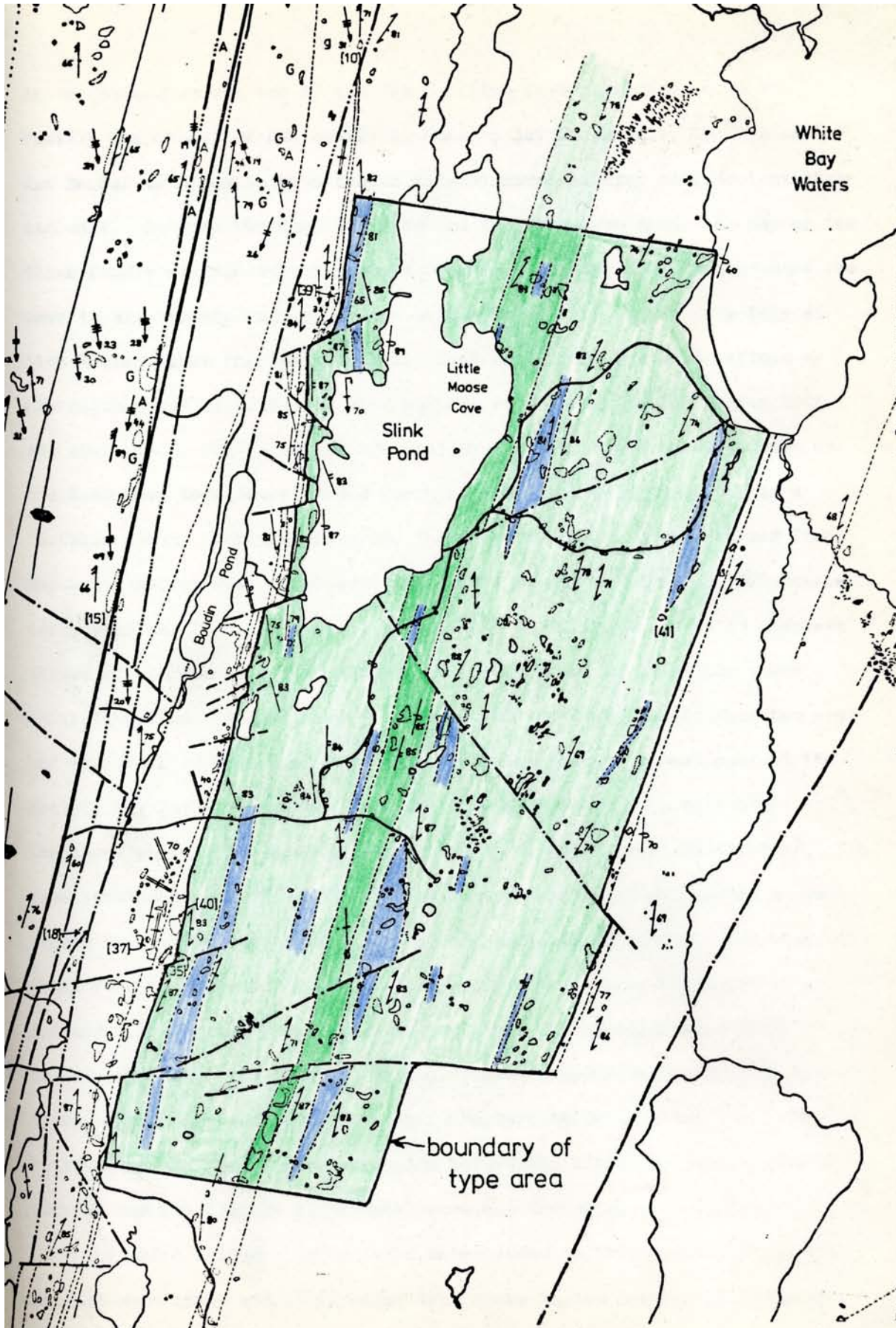
it is defined as the top of the first pillow lava flow underneath the Prairie Hat Member of the overlying Neale's Bay Formation. The Prairie Hat Member is a distinctive coarse clinopyroxene-bearing mafic volcaniclastic unit. Between the Camp 166 Road and the Bear Cove Road, the top of the Slink Pond Formation is very poorly defined, and west of Flatwater Pond the base is also poorly defined by the available outcrop, which is mainly of lithologies other than pillow lava. Both boundaries are well-defined on the north shore of Flatwater Pond but not so well on the Burlington Road. All contacts of the Formation are conformable except the upper contact by the Bear Cove Road Junction and north of Flatwater Pond, where it is a tectonic slide. The stratigraphic thickness of the Formation varies from about 260 metres west of Flatwater Pond, to a maximum of about 1600 metres north of Slink Pond. The lower part of the Formation where it is thickest around Slink Pond is a time equivalent of the upper parts of the Jukes Point Formation. A small part of the top of the Formation in the area west and southwest of Park Pond is a time equivalent of the lowest part of the Neale's Bay Formation on the Camp 166 Road and more may be equivalent to the lower part of the Neale's Bay Formation on northern Flatwater Pond. The type section of the Slink Pond Pillow Lava Formation is defined as a swath extending 3.0 km along strike, running perpendicular to strike from west of Slink Pond to White Bay Waters, with northern and southern boundaries 1.3 and 1.7 km respectively from the southern end of Slink Pond (Fig. 4B.5). It is not possible to define a useful 'linear' type section for this Formation. Approximate correlation within the type swath is given by the marker horizons of the Kidney Pond Formation below, the Little Moose Cove Member within, and the Prairie Hat Member above the Formation.

The mafic volcaniclastic rocks interbedded in this Formation are all unremarkable sandy and silty beds, like those in the underlying and overlying

Formations, except for the conglomeratic beds near the base, south of 2.6 km south of Kidney Pond outlet. The banded green, and rare grey, argillites are almost all found in the lower 1/3 of the Formation. One bed 30 cm thick of cleaved buff silicic volcaniclastic rock was seen in an outcrop 0.8 km NNE of the north end of Slink Pond, about 160 metres above the base of the Formation. A strongly-cleaved rusty quartz-rich greywacke type bed not more than 30 cm thick is exposed immediately on the eastern side of a coarse gabbro sill 0.6 km east of Kidney Pond. It resembles very closely the quartz-rich bed in the Jukes' Point Formation at the outlet of Kidney Pond. Two adjacent thin (~ 5 cm) beds of strongly cleaved, almost totally recrystallised quartz-rich sediment are found in an outcrop of this Formation on the north shore of Flatwater Pond. They also contain much calcite, and also possibly clastic muscovite.

Most of the pillow lava in the Formation is a very pale green, with pale yellow epidotic rims on the pillows typically about a centimetre thick. Pillow breccia/agglomerate is very uncommon. Pillows are typically 30 cm to 1 metre long, but in some flows are mostly 1-2 metres long. Flow thicknesses are not well controlled by the available data, but are probably mostly between 3 and 50 metres, and might be as much as 100 metres in rare cases. The pale green pillow lava is sometimes mildly amygdaloidal, with calcite and/or clinozoisite fillings. It very often has abundant white weathering green chert in the interstices between the pillows. Massive lava is relatively rare, and all that seen is pale green and highly amygdaloidal. It forms thin units 1 to 2 metres thick, that are seen in some cases to be at the base of mostly pillowed flows, but seem to be sharply defined and separate from the pillowed part. It is suggested that these highly vesicular massive lavas are material from the same extrusive event as each overlying pillowed flow, but that it was discretely intruded along

the base of the innnediately previously formed pillows, and may perhaps have fed pillow formation at the flow front beyond. In the area around the Camp 166 Road, the microscopic texture of the lava within the pillows is wholly obliterated by recrystallisation. However, in the least deformed area of the pillow lava from about 1.8 km north of the north end of White Bay Waters southward, the original texture is preserved. It consists of a very fine-grained randomly-oriented felted aggregate of albitised plagioclase needles, with very pale green interstitial actinolite at least partly replacing clinopyroxene. Generally, pale green pillow lava only occasionally contains a minor amount of altered plagioclase phenocrysts, but the first three or so flows in the Formation in the area west and south of Slink Pond are exceedingly porphyritic, and the rock is crowded with equant to squat plagioclase phenocrysts up to 1 cm across. Dark green pillow lava forms a minority of flows in the Formation, perhaps about 20% in the type section. It has a distinctive bluish tinge on the weathered surface, and usually has orange-coloured epidotic rims to the pillows, that contain either or possibly both of altered vermicular chlorite and stilpnomelane. The dark green pillow lava is sometimes amydaloidal, occasionally with large (~1 cm) vesicles filled with pink clinozoisite. A few flows of dark green pillow lava have a relatively minor amount of small (< 5 mm) lath-shaped (altered) plagioclase phenocrysts. Pillow agglomerate (dark green) occurs in places, but is not common. This pillow lava is clearly associated with beds of maroon banded chert up to 20 cm thick, that are not found among the pale green pillow lava. Some maroon chert and, in places, green, white-weathering chert are occasionally interstitial to the dark green pillows. In thin section, the dark green lava that is least deformed, east of Slink Pond, also shows a pseudomorphed original microscopic texture. Like the pale green pillow lava, it also has a randomly oriented aggregate of plagioclase

laths, but these are much larger and less elongate than in the pale green lava. The interstitial spaces are filled with a green, as opposed to a very pale green, actinolite accompanied by some chlorite, which is not seen in the pale green lava. Some (partially altered) ilmenite grains also occur, not seen in the pale green lava. No other visible differences have been detected between the two types. The differences suggest higher Ti and Fe and, perhaps, lower Mg contents in the dark green lava.

The lowest occurrence of this dark pillow lava forms a mappable member wholly composed of dark green pillow lava flows, termed (by definition informally) the Little Moose Cove Member, after a bay in Slink Pond named for the purpose. Between Slink Pond and Mic Mac Lake, this Member is between 180 and 50 metres thick. It is not continuously traceable northward due to lack of outcrop, but is present (about 140 metres thick) east of Kidney Pond, where the pillows are quite strongly deformed. It is not certain whether it disappears abruptly just before Flatwater Brook because this lithology was not known when the Brook section was examined, and the outcrop in the Brook is exceedingly poor, being mostly underwater, frost shattered, and covered with black algal slime.

In the upper half of the Slink Pond Formation, south of the Camp 166 Road some flows of dark green pillow lava are interbedded with more abundant flows of pale green pillow lava. They cannot be shown on Plate 1, mainly because there is insufficient outcrop to trace each flow, except for a unit composed of several dark green flows at the top of the Formation from 0.6 to 2 km south of the Camp 166 Road.

A very few outcrops of pillow lava are a medium green, and cannot be put in either pale or dark green categories. These few outcrops are found

Fig. 4B.6. Type section of the Neale's Bay Formation.
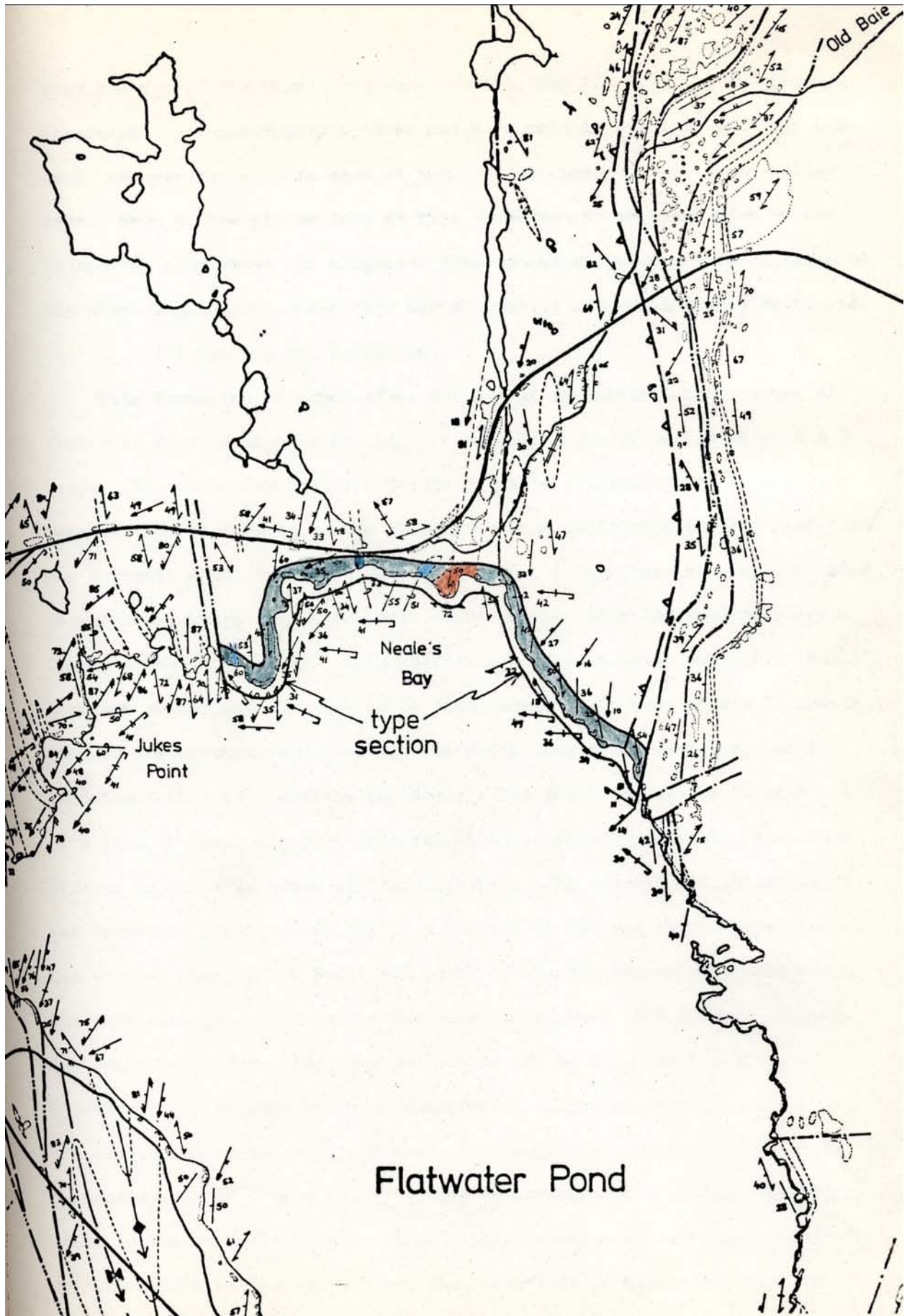
near the top of the Formation, south of the Camp 166 Road. Clinopyroxene phenocrysts (or pseudomorphs) have not been seen in outcrop, and only one small example was seen in each of two thin sections of pale green pillow lava. None of the pillow lava in this Formation or anywhere else in the inland map area shows the slightest development of varioles, in contrast to the Mings Bight area, where they are abundantly and consistently developed.

(f) Neale's Bay Formation.

This Formation is named after the bay at the northeastern corner of Flatwater Pond, named for the purpose of naming the Formation after E.R.W. Neale. The Formation mostly consists of mafic volcaniclastic rocks. The type section is defined as the outcrops on the shore of Flatwater Pond from the tectonic slide contact with the uppermost pillow lava of the Slink Pond Formation eastward to the tectonic slide contact with deformed Burlington Granodiorite (Fig. 4B.6). This section includes at least one other (small?) tectonic slide, and also the large late normal fault that enters Flatwater Pond at its northeastern corner. The fault downthrows westward, but it is not possible to correlate the Neale's Bay Formation across it with the data from the area mapped. Assuming no major fold or tectonic slide repetitions in the area north of Flatwater Pond, the approximate thickness of the Formation preserved in the type section is 850 metres and substituting the section east of the Fault and north of the Burlington Road, the maximum thickness preserved in the area mapped is about 1100 metres. As with the Jukes Point Formation, the section of the Neale's Bay Formation on Flatwater Pond is both thicker and somewhat different from the section in the area south of the Camp 166 Road. Although the type section has tectonic boundaries, it is chosen in preference to sections with conformable boundaries in the south because it shows a thick section containing lithologies not seen south of Flatwater Pond, and because it probably contains rocks

younger than any preserved in the south. However, the Formation is described first from the southern area, and the conformable contacts are defined.

The lower contact of the Formation is conformable with the Slink Pond Pillow Lava Formation south from the area west of Flatwater Pond. It is defined by the top of the uppermost pillow lava in the area west of Flatwater Pond but, in the area south of Flatwater Pond, it is defined as the top of the uppermost pillow lava that overlies less than about 20 metres of mafic volcaniclastic rocks. This change in definition is necessary as scarce pillow lava intertongues with more abundant mafic volcaniclastics in the badly-exposed area west of Flatwater Pond and because minor discontinuous pillow lava flows are found in the southern part of the Neale's Bay Formation (marked on Plate 1). In the southern part of the area, the change from mainly pillow lava to mainly volcaniclastic sediments is sharp, and quite clear in the field. In the unmapped area to the north of Flatwater Pond, it will probably be necessary to use the 'uppermost pillow lava' definition because of relatively poor exposure and suspected eventual lensing out of the Slink Pond Pillow Lava.

In the area between White Bay Waters and Park Pond, the Formation is defined to be conformably overlain by another unit consisting mainly of pillow lava, the White Bay Waters Pillow Lava Formation. This latter Formation is probably wholly a time equivalent of part of the section of the Neale's Bay Formation on Flatwater Pond, but is defined as a separate Formation to point up the relationship of the Baie Verte Group to the Mic Mac Lake Group. Thus the conformable top contact of the Neale's Bay Formation, where the White Bay Waters Formation is present, is defined as the base of the first pillow lava that is succeded mainly by pillow lava, with only thin interbeds of mafic volcaniclastics. Elsewhere in the area between

Fig. 4B.7. Reference section for the southern part of the Neale's Bay Formation.
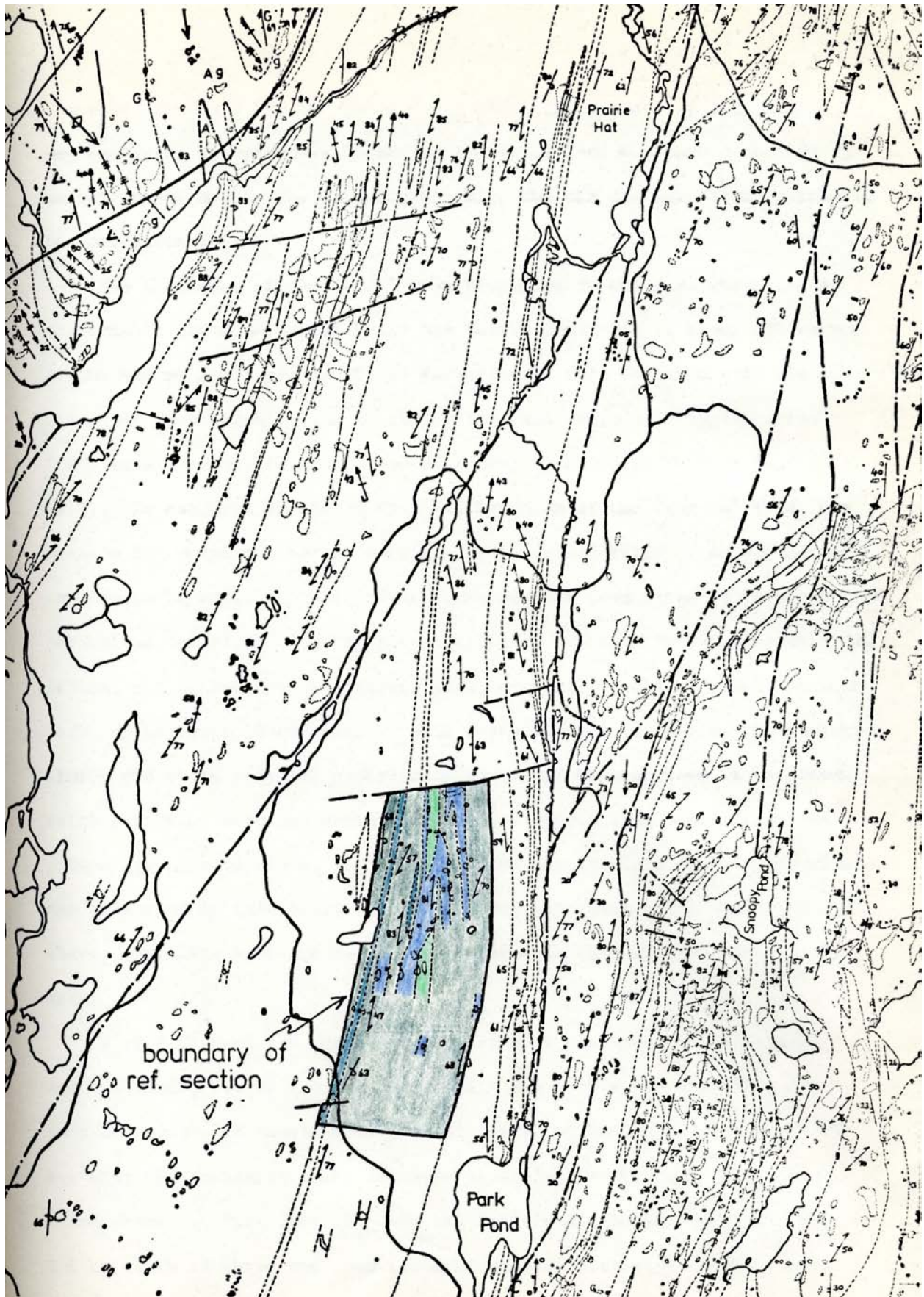
Flatwater Pond and Mic Mac Lake, the present top contact of the Neale's Bay Formation is variously formed by the large late westward downthrowing normal fault, and by the unconformity with the Mic Mac Lake Group overlying it, described later.

The thickness of the Neale's Bay Formation in the area where it is conformably overlain.by the White Bay Waters Formation is about 300 metres. As the nature and contacts of the Formation in this southern area are so different from those to the north, an informal secondary 'type-section' (reference section) is designated as a swath just north of Park Pond (Fig. 4B.7). In the area southward from 2.2 km north of the Camp 166 Road, the Neale's Bay Formation mostly consists of sandy and silty, with subordinate conglomeratic, mafic volcaniclastic rocks in beds from around a centimetre to several metres thick. The conglomeratic beds contain lithic, amygdaloidal lithic, and probably very deformed pumiceous clasts, all of mafic volcanic rock, up to about 10 cm across. Most examples seem to have matrix-supported clasts and these beds are essentially identical to some beds in the Jukes Point Formation west and north of Flatwater Pond. Some pillow lava and pillow agglomerate flows, and possibly some massive lava flows occur within the Formation in this southern area. Minor thin (mm) green argillite and cherty argillite beds are found between some of the mafic volcaniclastic beds.

A very distinctive unit occurs at or near the base of the Formation, and was traced 9.6 km from south of White Bay Waters to just north of the Camp 166 Road. It consists of the only mafic volcaniclastic rock in the map area that contains abundant mafic minerals, in this case fresh green clinopyroxene. This unit is named the Prairie Hat Member. A small hill 0.6 km south of where the Camp 166 Road crosses Flatwater Brook is named for the purpose of naming this unit. There appear to be two units of

Fig. 4B.8. Thin sections from the Prairie Hat Member I.
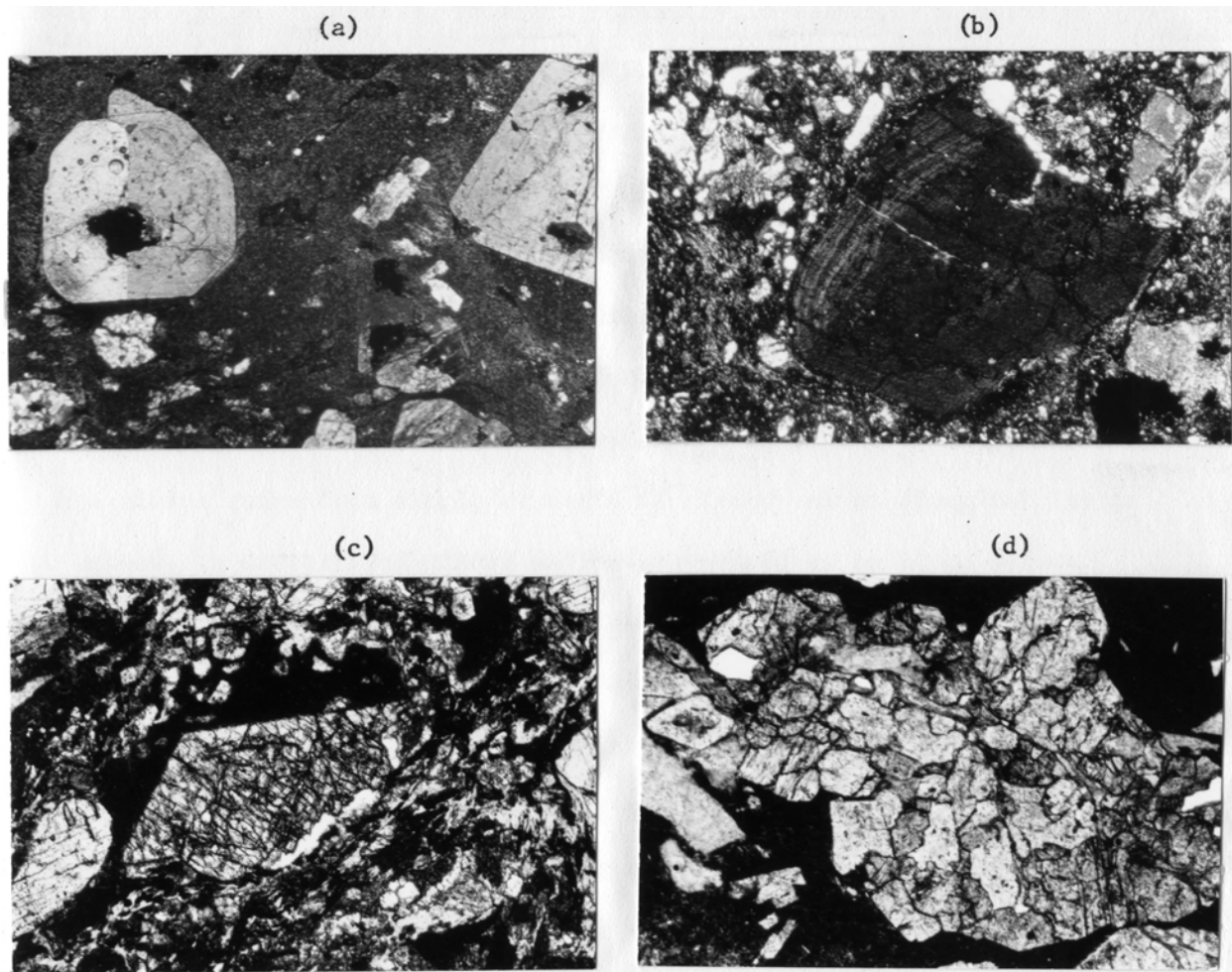
(a) Devitrified glassy clast containing large euhedral clinopyroxene, small euhedral plagioclase, and albite-filled amygdales (X 10).

(b) Broken clinopyroxene crystal with oscillatory zoned margin (X25).

(c) Broken clinopyroxene crystal with dark devitrified glass only on euhedral margin (X 25).

(d) Possibly cumulate clinopyroxene-plagioclase aggregate in devitrified glassy clast (X 25).

this lithology, one directly, or nearly directly, overlying the Slink Pond Pillow Lava Formation from south of White Bay Waters to the Prairie Hat, the other appearing north of Park Pond and continuing to the north as the lower one dies out. As it is wished to name only the clinopyroxene-bearing rocks and as it is not permitted to call two units exactly the same name, the two units are designated the Prairie Hat I and Prairie Hat II Members, respectively. The thickness of these Members ranges from 0 to a maximum of 25 metres. The rock in these Members is gritty to conglomeratic and the clasts range from single crystals of clinopyroxene (?augite), up to 1 cm across, to devitrified glassy mafic lava clasts up to 20 cm across containing abundant identical clinopyroxene phenocrysts. The rock also contains some very flattened recrystallised pumiceous clasts without phenocrysts and very rare clasts of green cherty argillite and fine-grained doleritic rock. Albitised plagioclase crystals occur as clasts and as phenocrysts in the devitrified glassy mafic clasts. These also contain fragments of coarse-grained often cumulate-looking intergrowths of clinopyroxene grains, and of clinopyroxene and plagioclase grains (gabbro). Many of the clinopyroxene phenocrysts are marginally zoned, and some have multiple oscillatory zoned borders (Fig. 4B.8). Most of the clinopyroxene crystals are either euhedral, or are broken pieces of euhedral crystals. The latter have devitrified mafic glass on their euhedral outside faces, but not on their angular broken faces, in rock containing the devitrified glass. The devitrified glassy clasts are often moderately amygdaloidal, the vesicles being filled with albite intergrowths. The rock as a whole is poorly sorted, with all sizes of clast from sand size upward in any rock. In a few places, the pebbles and cobbles are abundant enough to appear to form a self-supporting framework, but most often, the pebbly and larger clasts are matrix supported. Beds range from a few centimetres to about 5 metres thick, and some are very

roughly graded. The beds with the largest and greatest abundance of clasts are mostly restricted to the southern part of the area, opposite and south of Park Pond. North of Park Pond, pebbles are less common and most of the pyroxene grains are individual clasts. These coarse clinopyroxene-bearing volcaniclastics may be derived from andesite or from ankaramitic basalt. It is not clear whether some or all were directly deposited from either a subaerial or submarine volcanic eruption or from viscous mudflows or turbidites originating around a submarine vent, or whether they have been subaerially eroded.

The unique character and restricted stratigraphic range of this lithology suggests a correlation with 1-2 metres of a similar lithology in the type section on the northern shore of Flatwater Pond (Plate 2), described next.

Several beds 5-20 cm thick in an outcrop 0.42 km west of the northeast corner of Flatwater Pond contain abundant mafic crystal clasts, as well as albitised plagioclase and small shreds of recrystallised flattened mafic pumiceous clasts. The mafic mineral, dark green in outcrop, is a bright orange amphibole (?Ti-rich). Thus the similarity with the Prairie Hat Members I and II is not exact but this occurrence is in about the expected position in the sequence and therefore is tentatively correlated with them.

Most of the Neale's Bay Formation in the type section consists of undistinguished sandy and silty mafic volcaniclastic rocks, in beds commonly from less than a centimetre to a few tens of centimetres thick, but often not obviously bedded, perhaps due to strong deformation and recrystallisation. No pillow lava occurs in this section, and green argillite laminae between mafic volcaniclastic beds are extremely rare. Some of the apparently thinly laminated silty to sandy mafic volcaniclastics may be very deformed mafic volcaniclastic rocks with many flattened pumiceous clasts but no lithic mafic volcanic clasts were seen in outcrop in this section. Apart from the amphibole

clasts in the beds referred to the Prairie Hat Member, almost all other thin sections and outcrops showed only albitised and otherwise altered plagioclase crystal clasts. A few samples from relatively near the beds with amphibole clasts showed a minor amount of actinolitised mafic crystal clasts.

Although no lithic mafic volcanic clasts were detected in this section, small pebbles to enormous boulders of ophiolite complex-derived plutonic rocks are found, mostly within a limited horizon on the eastern shore of Flatwater Pond, 0.44 km south of the northeast corner. Here, clasts of somewhat cleaved leucocratic granular gabbro, from 1 cm pebbles to a boulder about 5 metres long, are found scattered through about 60 metres thickness of sandy mafic volcaniclastics. The gabbro consists of equant anhedral altered plagioclase and actinolitised clinopyroxene, with significant partly-altered accessory ilmenite. Apparently a little lower in the sequence, a section through an enormous block 50 metres wide (or possibly a collection of smaller blocks) of similar leucocratic gabbro is seen on an abandoned section of the Old Baie Verte Road, 0.48 km NE of the northeast corner of Flatwater Pond (Plate 2). This gabbro is cut by parallel diabase dykes. A thin section of slightly-cleaved gabbro shows equant anhedral altered plagioclase and similar actinolitised pyroxene. No ore mineral is present, although a few specks of fuchsite (after chromite) were seen on the specimen. The texture suggests that the rock was cumulate. Part of the next outcrop to the east is composed of similar gabbro with diabase dykes; this is probably also a smaller boulder within mafic volcaniclastics. Elsewhere, similar leucocratic gabbro is found in two localities near the Eastern Lineament Boundary Slide, 0.4 km S and 1.6 km NE of the Burlington Road. In the former locality, the probable boulder is small (< 2 m) but shows probable cumulate texture in thin section and flecks of fuchsite in outcrop. In the latter locality, the gabbro clasts are exceedingly deformed to strongly-cleaved

'garbenschiefer'-type rock, very close to the Boundary slide. Also on the abandoned Old Baie Verte Road, 0.24 km from the northeast corner of Flatwater Pond, several outcrops of shear polyhedra serpentinite with minor fish-scale serpentinite occur below the large gabbro block and are separated from it by mafic volcaniclastics (Plate 2). The outcrop nearest the contact with mafic volcaniclastics shows somewhat talcose fish-scale serpentinite with the foliation dipping moderately westward, concordant with the cleavage in the mafic volcaniclastics in nearby outcrops. There is no evidence for a slide zone on strike with this body in the outcrops on Flatwater Pond or the Burlington Road and therefore it is not thought to be a sliver tectonically emplaced on such a zone. It cannot be a tectonic sliver on the large, high angle late normal fault adjacent to it because of the moderately dipping concordant nature of the cleaved margin. It is also not a diapiric serpentinite. Therefore, it is thought to be a huge boulder deposited, like the gabbro above it, within the mafic volcaniclastic sediments. It could consist of several smaller blocks; no evidence was seen to support or refute this possibility. This is the only large occurrence of ultramafic rock in the inland map area where there is some evidence for its being in a sedimentary block.

Immediately west of the first promontory west of the northeast corner of Flatwater Pond, an outcrop of trondhjemite with pink aplite veins is found (marked on Plate 2 as ophiolite gabbro). The contacts with the mafic volcàniclastic sediments are not seen. It could be a sill but it is not seen immediately along strike on the Burlington Road. It is possible that it is another large boulder within the sediments. Identical rock was seen as thin veins in ophiolite gabbro with diabase dykes elsewhere (described previously). If this is a boulder, it is about 20 metres across. Below this horizon, a few pebbles of leucocratic gabbro and diabase rock, 5 to 10 cm long, are

found immediately adjacent to the amphibole-bearing beds referred to the Prairie Hat Member and are the lowest occurrence of ophiolite-complex derived plutonic clasts above the Kidney Pond Formation in this section.

About 60 metres above the supposed Prairie Hat Member, in the type section, the first bed of cleaved pink silicic volcaniclastic rock is found. Beds of this lithology from 2 cm up to perhaps as much as 80 metres, but not provably more than a few metres thick, are found occasionally and form perhaps up to 5% of the total section to the east of the first bed. A thick unit consisting almost entirely of this lithology, with a few thin mafic volcaniclastic interbeds, is mappable at the scale of, and is marked on Plate 2. North of the Burlington Road, beds of this lithology are very common near the Eastern Lineament Boundary Slide, and locally form at least 50% of the section. This lithology is distinctive and easily recognised, although the thinnest beds are greyish, rather than pink. All thin sections show a recrystallised very fine-grained granular quartzo-feldspathic matrix with very fine-grained white mica defining a well developed cleavage. Clear sub- to euhedral albite crystals typically 0.5 to 2 mm across are common (~20%) in all of this lithology, and they are mildly kinked, brittly shear-fractured, and augened by the cleavage but are relatively undeformed compared to the matrix. The fact that they are clear, rather than even weakly saussauritised, in all thin sections seen, suggests that their original composition was near albite. Their state of deformation compared with the matrix shows that the rock was fragmental. The euhedral nature of the albite crystals and the lack of lithologies indicating mixing of this silicic fragmental rock with the mafic volcaniclastics perhaps suggest that these beds were deposited directly as tuff. The lithology is not found south of the last outcrop in the thick unit shown on the map by the Flatwater Pond Provincial Park in the detailed map area. Beds of a

very similar lithology are found in the section along Black Brook (Plate 6) just south of the sharp

bend in the stream east of Black Brook Bridge, but these may not be correlative with the Neale's

Bay Formation. The incoming of this lithology could be used to delineate the stratigraphy when

the area north from the Burlington Road is mapped.

North of the Burlington Road near the Eastern Lineament Boundary Slide, massive

chalky-buff weathering grey rhyolitic rocks are found (Plate 2). In the western part of this strip,

the rock is fairly severely cataclastically deformed by an attempt to form a rude cleavage. The

massive undeformed rock on the eastern side shows just an ultrafine-grained granular quartzo-

feldspathic aggregate with occasional ovoid quartz grains up to 0.2 mm across that may be very

resorbed phenocrysts. This rock may be either a flow or a sill; it is possible that it is related in a

general way to the pink silicic tuffs.

About a metre of siliceous grey slate is found in one outcrop adjoining the contact with

the Mic Mac Lake Group 1.72 km north of the bridge where the Camp 166 Road crosses

Flatwater Brook. This is only mentioned because it may indicate a correlation with a thick

sequence of similar grey siliceous slate found in reconnaissance immediately east of Black

Brook Bridge, underlying the mafic volcaniclastics containing pink silicic tuff beds.

(g) White Bay Waters Formation.

In the area between east of White Bay Waters and Park Pond, a section consisting mostly

of pillow lava overlies the mafic volcaniclastics of the Neale's Bay Formation. This is named the

White Bay Waters Formation as there is no nameable geographic feature within the area

occupied by the rocks concerned. The Formation is poorly exposed, so the type section is defined

as a swath covering the group of outcrops opposite the southern end

Fig. 4B.9. Type section of the White Bay Waters Pillow Lava Formation.
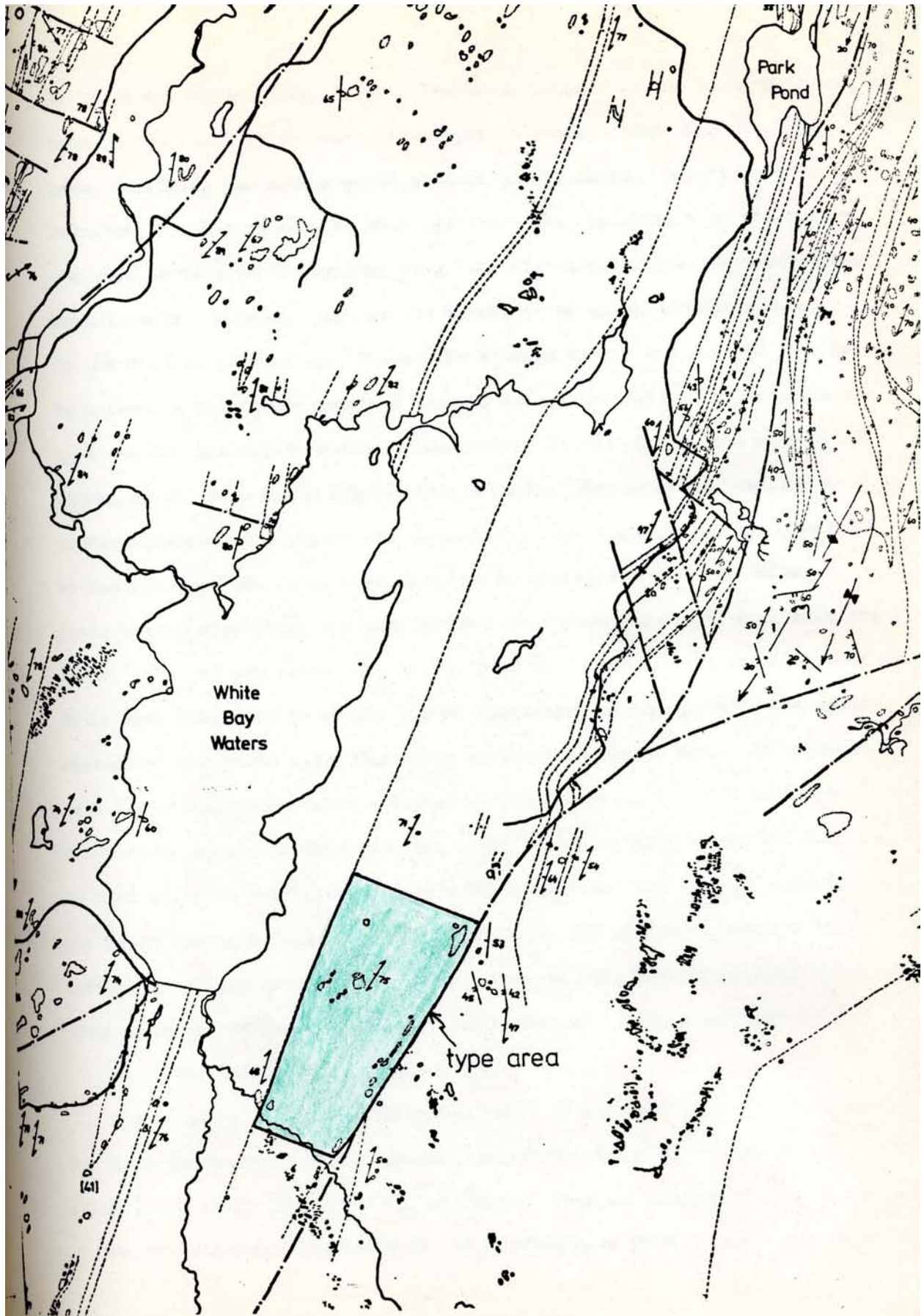
of White Bay Waters (Fig. 4B.9). The lower contact of the Formation, the base of the first pillow lava in a sequence mostly consisting of pillow lava, overlying the mainly mafic volcaniclastic section of the Neale's Bay Formation, is not exposed in this section. The top contact in the area of the type section is interpreted as a late high-angle westward-downthrowing normal fault. However, just to the north, it is an unconformity overlain by the Mic Mac Lake Group. Due to the limited extent and outcrop of this Formation, a relatively complete section is not available. Outcrops in the type section are mainly medium-green rather cleaved pillow lava with minor interbeds of sandy mafic volcaniclastic rocks. To the north, massive lava or fine-grained dolerite occurs, especially near the unconformity overlain by the Mic Mac Lake Group, and sandy mafic volcaniclastic beds, minor green banded argillite, and one thin bed of red banded cherty argillite are found. The maximum preserved thickness of this Formation is about 430 metres. It is very likely to be wholly a time-equivalent of some part of the upper section of the Neale's Bay Formation north of Flatwater Pond. It is defined as a Formation, rather than a member within the Neale's Bay Formation because it is rather thicker than the small lenses of pillow lava below it in this southern part of the Neale's Bay Formation, and also to emphasize the relationship between the Baie Verte Group and the unconformably overlying part of the Mic Mac Lake Group. However, the White Bay Waters Formation could be reduced to Member status within the Neale's Bay Formation.

(h) Dolerite and gabbro sills

Mafic sills are common throughout most of the Baie Verte Group in the map area and most are medium-grained dolerite. Sills range from about 10 cm to at least 120 metres in thickness. They are estimated to form 25% of the stratigraphic thickness in the section east from Kidney Pond;

elsewhere they form no more than 15% of any section. In general, there are relatively fewer sills in the Neale's Bay Formation and fewer and thinner sills in the north Flatwater Pond-Burlington Road section compared with the area south of the Camp 166 Road. In other words, there is a rough correlation with the amount of pillow lava in any section, and there are relatively fewer sills above the main pillow lava horizon, the Slink Pond Formation.

Exposures of sill margins show a zone of decreasing grain size a few centimetres wide adjoining a chilled margin a few millimetres wide. Most dolerite is non-porphyritic, and has randomly oriented albitised plagioclase laths and ophitic actinolitised pyroxene, with significant accessory altered ilmenite. Rare sills are highly porphyritic dolerite with rather equant altered plagioclase phenocrysts up to 1 cm across. These resemble very closely the highly porphyritic pillow lava flows at the base of the Slink Pond Formation around Slink Pond, but most of the porphyritic sills are intruded above this horizon and do not have equivalent flows. Sills of dolerite with altered mafic phenocrysts (~4 mm across) are even less common, and equivalent flows have not been seen. It is possible that a few massive lava flows have been mistaken for sills, but in general the distinction is clear, mainly based on the coarser grain size, less cleaved nature, and lack of amygdales in the sills. Very rare examples of coarse-grained gabbro sills are found. A relatively thick coarse-grained gabbro sill is situated about 0.6 km east of the eastern shore of Kidney Pond. The gabbro appears to fine slightly, but is not chilled, in the one contact seen with mafic volcaniclastics. Several outcrops of this sill contain one or two thin subparallel fine-grained dolerite sills, between 5 and 20 cm thick, that are strongly chilled against the gabbro. In this respect, and also in the unique leucocratic nature of this particular gabbro, it greatly

resembles (boulders of) ophiolite gabbro with diabase dykes. However, the texture of the gabbro is wholly unlike ophiolite gabbro, as it consists of randomly oriented fat laths of sub- to euhedral plagioclase, interstitial to ophitic partly-actinolitised clinopyroxene, and abundant large jagged ilmenite grains.

The distribution and nature of the sills suggests that they are comagmatic with the pillow lavas. None have been seen that post-date the partial to total low-greenschist-facies alteration of the enclosing sediments and lavas.

(iii) Baie Verte Group of other parts of the inland map area.

(a) North of the Burlington Road

Between 3.4 and 9.8 km north of the Burlington Road Junction, rocks were mapped (Plate 2) that cannot yet be correlated accurately with the stratigraphy defined above. They are mainly rather homogenous, not obviously bedded, sandy to silty mafic volcaniclastic rocks, with very rare green argillite partings. They are all strongly cleaved with the exception of the northernmost roadcut mapped on the Baie Verte Road. Two beds of cream weathering ultrafine-grained silicic tuff about 20 cm and 1 m thick were seen in a quarry next to the road 9.2 km N of the Burlington Road Junction. At 3.64 km and from 6.2 to 7.5 km north of the Burlington Road Junction, a light olive-grey-green pillow lava horizon is found immediately overlain by sediments that are mainly sandy mafic volcaniclastics, but also include a 1-2 metre thick black slate horizon. The northernmost roadcut mapped on the Baie Verte Road contains a lens (channel?) of conglomeratic mafic volcaniclastic rock about a metre thick, which contains only mafic volcanic clasts up to 25 cm across. This outcrop also contains several beds seen in thin section to contain about equal amounts of fairly euhedral actinolitised clinopyroxene and albitised plagioclase crystals, and a few aggregates of

such clinopyroxene crystals. A few other sandy mafic volcaniclastic beds in this northern area seen in thin section contain albitised plagioclase with rare flattened and recrystallised pumice fragments. Some contain significant quantities of actinolitised mafic crystal fragments.

From 7.0 km N of the Burlington Road Junction, the mafic volcaniclastics contain dolerite and gabbro in some outcrops at specific horizons, assuming that bedding is essentially parallel to cleavage as it is in the area to the south. Some of these gabbro and dolerite occurrences may be sills, but much is leucocratic gabbro strongly resembling ophiolite gabbro and, in addition, some of this contains tabular diabase bodies 30 to 50 cm wide, parallel to each other and to bedding adjacent to the gabbro. Although the example cited in the previous section shows that diabase sills may intrude larger gabbro sills in the area, it is possible that many of these gabbro outcrops are of boulders or larger slabs of ophiolite gabbro, sometimes with diabase dykes, deposited at particular horizons. The gabbro in roadcuts on the western side of the Burlington Road at 8.6 km north of the Burlington Road Junction is in particular likely to be ophiolite gabbro in a large slab or boulders, because it shows a few anorthosite bands and possible cumulate texture in thin section. Unlike the gabbro body 9.5 km N of the Burlington Road Junction, which is a tectonic sliver on a tectonic slide, this gabbro is definitely not so located, and there is no evidence to suggest that any other of these gabbros are on such slide zones. These gabbro localities have been marked as lenticular and continuous horizons on Plate 2, but if they are horizons containing boulders, the gabbro will probably be discontinuous.

In this northern area, 7.8 km north of the Burlington Road Junction, outcrops of deformed boulder conglomerate are found on either side of the

Baie Verte Road. It consists of pebbles and boulders originally from about 1 to 50 cm across, mostly of green quartzite and quartz-pebble conglomerate, with minor brown siltstone, and rare reddish jaspery chert, ultramafic, and vein quartz clasts. The matrix is dark green chlorite, probably derived from an ultramafic matrix, and it coats all the clasts. Two very poor outcrops of sheared serpentinite within the outcrop area of the conglomerate may represent a very large boulder, or a tectonically introduced sliver. Most clasts are strongly deformed, and are now ellipsoids and more irregular shapes. A little deformed vein quartz clast is well-rounded and moderately spherical, but the irregular shapes of some of the deformed clasts suggest that many were irregularly shaped and probably angular. Weathered surfaces perpendicular to the subvertical maximum elongation axis show that the clasts probably formed a self-supporting framework, although this may not have been true for the whole unit. The clasts and their derivation are discussed in detail in a following section. The contacts of this conglomerate are not exposed, and its relationships to, and the status of, the adjoining ophiolite gabbro and ultramafic outcrops are not certain. The shape of the conglomerate body on the map is like a channel-section, and it may be in a channel. If so, there is a possibility that the strip of ultramafic rock running north from this point is composed of large boulders.

The Kidney Pond Conglomerate Formation is the only unit in the Baie Verte Group to the south that contains significant clastic quartz, until the Black Brook section, described next. As the cleavage in the Baie Verte Group of the inland map area is essentially parallel to bedding, and as it parallels the western margin of the Lineament, it is quite possible that this quartz-rich boulder conglomerate is the lateral equivalent of the Kidney Pond Conglomerate Formation. If so, the mafic volcaniclastics above it may be equivalent to part of the Jukes Point Formation, and those below

it to the south, to part of the Teardrop Pond Formation. It is not clear where this conglomerate horizon might continue to the north, and it is therefore difficult to assign the mafic volcaniclastics here to either unit.

(b) Mic Mac Lake — Black Brook area.

Roadcuts along the Baie Verte Road opposite the southern part of Mic Mac Lake (Plate 6) are all of pillow lava, and are probably part of the Slink Pond Formation.

The section along Black Brook and the woods road leading to the eastern side of Pittman's Pond (Plate 6) cannot at present be correlated with the stratigraphy set out previously, and it may be above much or all of the defined sequence. At least 30 metres entirely of thin bedded and laminated maroon argillite 1.1 km west of Black Brook Bridge, is succeeded to the east by a section of strongly to intensely cleaved green sandy mafic volcaniclastics and very minor green and grey argillite, which continue to Black Brook Bridge. It contains two gritty beds a metre or so thick with about equal amounts of albitised plagioclase and quartz grains. Some of the quartz is blue, and some shows abundant resorbtion cavities. A minor amount of lithic fragments of both red and pale rhyolite occur, some containing quartz and feldspar phenocrysts. On the woods road leading to Pittman's Pond, some outcrops on strike with the eastern part of this section in Black Brook show a few metres of pebbly slide conglomerates containing flaky brittle cherty argillite clasts, and subordinate mafic lava clasts supported in a silty mafic volcaniclastic matrix. One outcrop also contains rhyolite and granodiorite pebbles up to a few centimetres across. The granodiorite has reddish-pink K-feldspars, and the rock consists of granular quartz, albitised plagioclase in subhedral crystals and subordinate altered K-feldspar. A few tens of metres of pillow lava with interstitial chert overlies these pebbly conglomerates. It is not known whether either of these lithologies

occur on the Black Brook section, because the part on strike was not examined. Immediately on the east side of Black Brook Bridge, a section at least 80 metres thick, almost entirely composed of siliceous grey slate, succeeds the strongly-cleaved mafic volcaniclastics. There is no sign of any fault crossing the section near Black Brook Bridge, as previously mapped (Neale and Nash, 1963). The grey slate is also exposed in the roadcut just south of the bridge, where it contains a reddish cleaved silicic tuff band about 20 cm thick. South of the sharp bend in Black Brook, reconnaissance of outcrops as far as 0.5 km from the bridge revealed cleaved sandy mafic volcaniclastics with some beds about 20-50 cm thick of buff to pink cleaved silicic tuff. It is possible that these may be correlative with those in the upper part of the Neale's Bay Formation north of Flatwater Pond, at least in the most general way.

A road cut on the south side of the Baie Verte Road 1.1 km from Black Brook Bridge exposes greyish-buff to brown non-porphyritic flinty eutaxitic ignimbrite. It is cut by very narrow (~1-2 cm) devitrified glassy mafic dykes, and contains a few pebbles of Burlington Granodiorite. This rock is wholly unlike the ignimbrites in the Mic Mac Lake Group on Mic Mac Lake, and further north, which are a bright maroon colour, and mostly are highly porphyritic. It is possible that it, and other outcrops like it further south along the Baie Verte Road, and to the west of this part of the Baie Verte Road (H. R. Peters, pers. comm.) are not, as previously mapped (Neale and Nash, 1963), part of the Mic Mac Lake Group, but are part of a sequence originally conformably overlying the mafic rocks of the Baie Verte Group and representing a continuation of the same volcanism. They might be assigned to the Baie Verte Group, or to a separate new Group, if this turns out to be the case when the area is mapped in detail.

4B.39

(iv) Clast types in the Baie Verte Group.

(a) Mafic and sediment clasts

1. Ophiolitic Gabbro and Diabase

Identifiable clasts of ophiolite gabbro with or without parallel diabase dykes, range from small pebbles up to enormous blocks of the order of 100 metres across, and perhaps larger. The large blocks occur in the Boudin Pond Megabreccia Formation, within and perhaps below the Teardrop Pond Formation, occasionally within the Kidney Pond Conglomerate Formation, and locally within the Neale's Bay Formation northeast of Flatwater Pond. The smaller boulders and pebbles are mostly found within the Kidney Pond Conglomerate and a few are found in the lowest part of the Jukes Point Formation near Slink Pond, and in the Neale's Bay Formation on northeast Flatwater Pond. In the Kidney Pond Formation, a few small clasts were seen that contained the contact of a diabase dyke with gabbro (Fig. 4B.10a). The lithologies of the ophiolite gabbroic and related rocks have been described in detail in previous sections, except for pebbles of a greasy-green white-weathering anorthosite quite commonly found in the Kidney Pond Formation. It consists of anhedral granular albitised plagioclase, with a few grains of actinolitised clinopyroxene and significant accessory subhedral sphene in large grains. It is emphasised that all lithologies can be matched with the internally undeformed gabbros in the Mings Bight ophiolite complex, and other well preserved ophiolite complexes. It is also strongly emphasised that the gabbro in the clasts shows no trace of any penetrative deformation apart from the high temperature (sub-magmatic) foliation that is cut by the diabase dykes, (and is found in all well preserved ophiolite complexes) and in places a destructive cataclastic proto-foliation and associated alteration produced during the deformation of the surrounding

Fig. 4B.10. Conglomerates.

(a) Diabase dike-gabbro contact in pebble, Kidney Pond Formation [39].

(b₁)   (b₂)
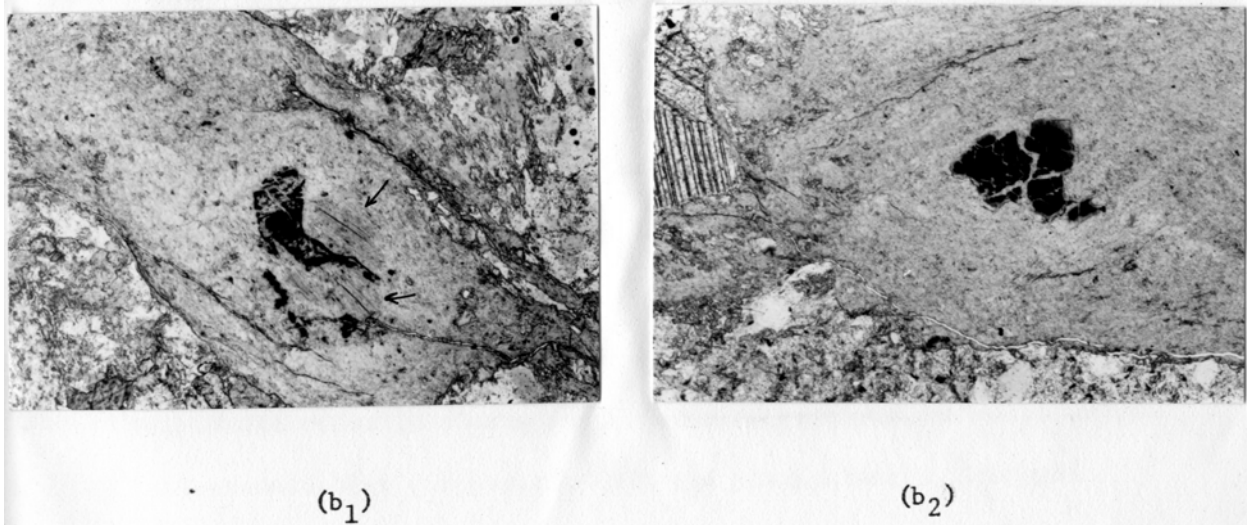
(b) Chloritic ultramafic pebbles from quartz-rich conglomerate 7.8 km N of the Burlington Road

Junction (X 40)

(b$_1$) bastite pseudomorphs (arrowed); chromite with skeletal magnetite alteration overgrowth.

(b$_2$) fractured chromite.

and overlying sediments and volcanics of the Baie Verte Group. Therefore the clasts came from an ophiolite complex that had not been penetratively deformed by a regional compressive deformation.

2. Ophiolitic depleted ultramafic rock.

A probable large block of this lithology is found at the northeastern corner of Flatwater Pond, and possibly in large blocks in the area immediately around a point on the Baie Verte Road 7.6 km N of the Burlington Road Junction. Pebbles of altered ultramafic rock up to about 1 cm across are found in places in the quartz-rich boulder conglomerate in the latter locality. These are now composed of dark green chlorite, contain fractured brown chromite grains, and occasionally show chloritised bastite pseudomorphs (Fig. 4B.10b)

3. Mafic volcanic rock.

Pebbles to boulders up to 30 cm across are found in the Kidney Pond Conglomerate Formation, the Jukes Point Formation, the lowest 70 metres of the Slink Pond Formation south of Kidney Pond, and in the Neale's Bay Formation south of Flatwater Pond. Clasts of mafic volcanic rock in the Kidney Pond Conglomerate are almost all non-vesicular/amygdaloidal and non-porphyritic and consist of abundant fine-grained (albitised) plagioclase needles and interstitial chlorite and actinolite, with a random basaltic to trachytic texture. None contain any trace of a pre-depositional tectonic foliation. Mafic volcanic clasts in the other Formations are often amygdaloidal, and occasionally porphyritic (altered plagioclase). The mafic volcanic clasts in the Kidney Pond Conglomerate may be derived from pillow lavas of the ophiolite complex represented by the gabbro and diabase clasts, but the mafic volcanic clasts in the other Formations closely resemble the lavas of the Slink Pond Formation. Flattened and recrystallised probably originally pumiceous mafic clasts are found, and are probably very common,

in the Neales Bay Formation, and in the Jukes Point Formation north of Kidney Pond. Devitrified glassy vesicular mafic volcanic clasts with clinopyroxene and altered plagicclase phenocrysts are found in the Prairie Hat Members of the Neale's Bay Formation. All mafic volcanic clasts could have been derived from either basaltic or andesitic parent material.

4. Argillaceous and cherty clasts.

Grey to black argillite clasts are common in the Kidney Pond Formation, especially in conglomerate with slaty matrix, and were most probably derived from slightly more lithified sediment in the same place that the soft slaty matrix originated. Green and grey pebbly argillites in the Boudin Pond Formation at least partly reflect soft sediment deformation under blocks of gabbro as they were emplaced. Pebbly argillites in the Teardrop Pond Formation, and green banded argillite clasts showing soft sediment deformation and rare brittle flakes of green chert, in the Jukes Point and lowest southern Slink Pond Formation, the Prairie Hat Member and outcrops on the Pittman's Pond woods road are also locally derived, as the same sediment is interbedded with the conglomeratic beds containing these clasts.

5. Mafic volcaniclastic clasts.

These have been found in the Teardrop Pond Formation at the base of the large clinopyroxenite boulder, and in conglomeratic beds in the Jukes Point and lowest Slink Pond Formations south of Kidney Pond. The occurrence in the Teardrop Pond Formation is clearly due to soft sediment deformation during emplacement of the boulder overlying it. In all occurrences these clasts are essentially identical with the nearby bedded mafic volcaniclastics and must be locally derived. In the case of the Teardrop Pond Formation occurrence, the clasts are identical to the matrix, and close packed, and it is very difficult to detect them. It is quite possible that similar rocks are more widespread, and were not identified as clast-bearing.

6. Crystal fragments in volcaniclastic rocks.

The overwhelming majority of beds of green mafic volcaniclastic rocks now contain only altered plagioclase crystal clasts, with minor and variable amounts of small altered ilmenite fragments in some cases. The plagioclase is locally abundant and up to 5 mm across in some beds in the Jukes Point Formation, but mostly they are 2 mm down to about 0.05 mm across, and only visible in thin section. Occasionally in little deformed beds, clasts as small as 0.01 mm survive. They may be euhedral, but are most commonly equant anhedral abraded grains. The green colour of the beds, due to recrystallised fine-grained actinolite and/or chlorite in the matrix, perhaps suggests that there was originally a fine-grained mafic crystal fragment component in most or all beds. Beds that now contain a significant proportion of actinolitised mafic crystal fragments are relatively rare, found in 1) The Teardrop Pond Formation on the Burlington Road; 2) the lower Jukes' Point Formation on western and northern Flatwater Pond; and 3) in the northernmost mapped outcrop on the Burlington Road (?Jukes' Point Formation) and other outcrops in this area. The only other beds known to contain large quantities of mafic crystals are the Prairie Hat Members south of Flatwater Pond (clinopyroxene) and the presumed equivalent on northern Flatwater Pond (orange amphibole). Probable originally sodic plagioclase is found in the pink silicic tuff beds in the Neale's Bay Formation north from the southern end of Flatwater Pond. Staining of about 20 selected thin sections of mafic volcaniclastic rock revealed no K-feldspar, and it is thought to be wholly absent from the volcaniclastic rocks. It is emphasised that quartz crystal fragments are lacking in almost all volcaniclastic beds. A few grains form a very small proportion of the total content of a few samples. Clastic grains of magnetite have not been seen All opaque grains are altered to leucoxene or sphene

and were originally ilmenite.

(b) Exotic clasts

1. Granodiorite

Clastsof chalky-weathering granodiorite are restricted to the Kidney Pond Formation, but are

found throughout the strike length mapped from Mic Mac Lake to the Burlington Road. Most

clasts are less than 15 cm across, but a few are found up to about 1 metre across. They appear to

be fairly well-rounded in outcrops where there is relatively little deformation. They usually form

5% or less of the clast assemblage, although a few outcrops near Mic Mac Lake show local

concentrations of up to 10% of granodiorite clasts. This is the only clast type that has been

mentioned by previous workers (Church, 1969; Neale and Kennedy, 1967), and they have

suggested that it was derived from the Burlington Granodiorite. The clasts are megascopically

extremely similar to the Burlington Granodiorite, which has a uniform and characteristic

appearance over most of its present extent. This resemblance can be demonstrated in the field by

an outcrop of relatively little deformed green sandy matrix conglomerate opposite the centre of

Slink Pond [39], where it happens that a 1 metre glacial erratic of Burlington Granodiorite rests

on the northern part of the outcrop. Granodiorite clasts in the conglomerate can be directly

compared with it. The reasons why the clasts resemble the Burlington Granodiorite are: 1) the

grain sizes of the minerals are very similar in both; 2) the proportion of mafic to felsic minerals

is approximately the same in both, although the clasts as a whole perhaps tend to have a slightly

lesser proportion of mafic minerals; 3) the altered plagioclase is subhedral and tends to be

slightly glomeroporphyritic in both; 4) the altered K-feldspar is generally very subordinate to

absent in both; and 5) the quartz tends to have a slight grey-blue tinge in the Burlington

Granodiorite, and this is seen in the clasts, although some clasts have a very pronounced bluish

or even violet colour in the quartz that the

author has not seen in the Burlington Granodiorite. There are three slight differences seen in thin section between the clasts and the Burlington Granodiorite sectioned by the author, which consists of 7 thin sections from a tiny fraction of its total area. First the clasts tend to have a somewhat larger proportion of quartz, some of which is in large grains, whereas it is mostly interstitial in the Granodiorite. Secondly, the clasts seen in thin section all contain green hornblende, but only two contain chlorite after biotite, and only one more than a trace amount, while the Granodiorite seen has fairly even proportions of green hornblende and variably chloritised green biotite. Thirdly, only one clast was seen to contain a large well formed sphene grain, and the other clasts contain altered ilmenite grains, whereas well-formed sphene is common in the Granodiorite seen. These differences are not thought to be too significant in view of the other very close resemblances, and the author therefore endorses the suggestion that these clasts are derived from the Burlington Granodiorite. The clasts resemble it far more than they resemble granodiorite specimens the author has seen that were collected from the Topsails area (to the south of the Burlington Peninsula — Fig. 1.3) and that may be of the same age as the Burlington Granodiorite (Neale and Nash, 1963). The author has not seen thin sections of granodiorites probably of the same age, found pre- or syn-kinematically intruding highly deformed mafic rocks, in the Coastal Slice of the Bay of Islands allochthon, and in the Twillingate area. A few specimens and outcrops of these rocks seen by the author do not match the clasts as closely as the Burlington Granodiorite.

The granodiorite with reddish-pink K-feldspars found in pebble conglomerate on the Pittman's Pond woods road is probably derived from the Burlington Granodiorite, although the resemblance is not as close as it is for the

clasts in the Kidney Pond Formation.

2. Silicic volcanic clasts

These pale buff clasts form only a very minor proportion (< 1%) of the assemblage in the Kidney Pond Conglomerate, but are found over its whole mapped extent from Mic Mac Lake to the Burlington Road. They are not found outside the Kidney Pond Formation. The clasts contain sparse phenocrysts up to 2 mm across, mainly of quartz with minor (probably originally K-) feldspar, set in an ultrafine-grained granular quartzo-feldspathic groundmass, which contains a minor amount of a pale chloritic or micaceous mineral. The latter defines a foliation that augens around the phenocrysts, and is axial-planar to some folds of granular quartz veins. The phenocrysts are strained, and some are slightly sheared and broken, but are little affected by the foliation compared with the matrix. The foliation is a cleavage, not a flow-banding effect, and the deformation contrast between phenocrysts and matrix suggests the rock was a tuff or other volcaniclastic rock prior to deformation and recrystallisation of the matrix. A few of the quartz phenocrysts have bottle-shaped resorbtion cavities, and most are anhedral, fairly equant grains. Most of the feldspar phenocrysts are subhedral prisms. These foliated silicic volcanic clasts tend to have tabular shapes controlled by their foliation, and thus to lie flat in the bedding orientation in the conglomerate, which is parallel to the Bale Verte D_1 slaty cleavage. However, the foliation in the clasts is pre-depositional. The largest clast of this type, seen on the northern shore of Flatwater Pond, is a 1 x 1 x 0.15 metre slab. The only possible source of foliated silicic volcanics at the time of deposition was the Cape St. John Group, which is the highest preserved unit in the Fleur deLys Supergroup (eastern division), and these clasts are interpreted here as being derived from this source.

In the Black Brook area, described previously, a few non-foliated

rhyolite clasts were found in pebbly slide conglomerate and gritty 'greywacke', accompanied in the 'greywacke' by blue quartz with resorbtion cavities. The presence of blue quartz and the granodiorite pebbles suggests that the rhyolite was derived from the Cape St. John Group, although it could have been derived from rhyolite equivalent to silicic tuffs seen higher in the section on Black Brook.

3. Granophyre

Two clasts both about 2 mm across were seen in thin sections of two specimens collected from the Kidney Pond Conglomerate on the north shore of Flatwater Pond. These were probably derived from rocks related to the Cape St. John Group, because a fragment of granophyre was also seen in one clast of foliated silicic volcanic rock.

4. Fleur de Lys schist clast.

In the western part of the southernmost outcrop of the Kidney Pond Conglomerate Formation, west of northern Mic Mac Lake, there is a large boulder of crenulated schistose rock. It is not particularly well-exposed; a section 1 x 2 metres can be seen, and the whole boulder may be twice as large. The boulder consists of cream-weathering dark grey-green foliated rock that could be termed either a phyllite or a fine-grained schist. In thin section the rock consists of thin pelitic laminae, wholly consisting of recrystallised muscovite, interbanded with slightly thicker beds of 'dirty siltstone'. The latter consist of a recrystallised ultrafine-grained turbid brown aggregate that contains some silt to fine sand size clastic quartz grains and a few clastic albite grains. The schistosity defined by the muscovite in the pelitic layers is seen in places oblique to the layering (bedding), and axial planar to tight (plastic) microfolds of this bedding (Fig. 4B.11a). The thin bedding and generally subparallel schistosity and subparallel quartz veins are folded in places by open to close crenulation

Fig. 4B.11. Fleur de Lys schist clast from Kidney Pond Formation.
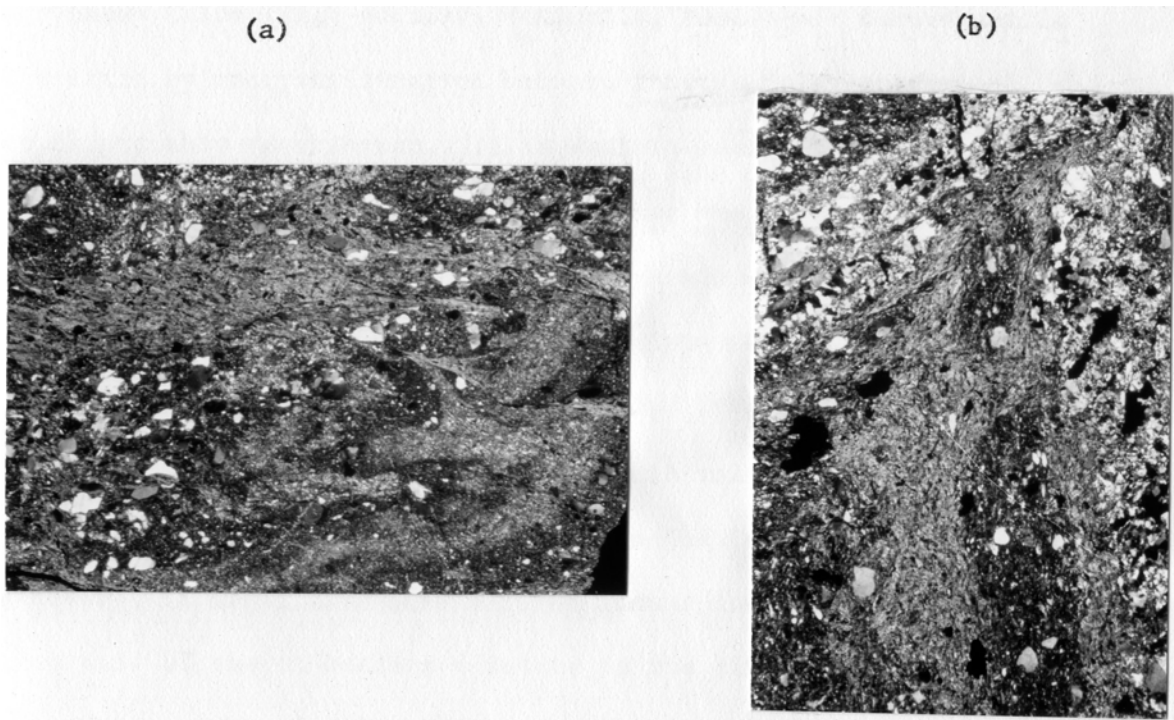
(a) 'F$_1$' folds of bedding, clastic quartz and albite, pelitic lamina with crosscutting schistosity (X 10)

(b) 'F$_2$' folds of bedding; brittly boudinaged silty layer (X 10)

folds (Fig. 4B.11b). The 'silty' layers are brittly boudinaged on the limbs of these folds (Fig. 4B.11b), indicating that their turbid matrix became brittle by recrystallisation between the two fold generations, and that it is probably quartz-rich. It is certain that the schistosity in this clast is pre-depositional, and very likely that the crenulation is also pre-depositional. This low-grade partially-recrystallised fine-grained schist was fairly certainly derived from a high level in the Fleur de Lys terrain. The ductility change in the matrix of the silty layers may have happened because they consisted mostly of glassy silicic volcanic detritus If so, it is likely that this clast was derived from the Cape St. John Group. Alternatively, it might have been derived from a low grade equivalent of any other unit of quartz-bearing clastics in the Fleur de Lys Supergroup Although source rocks of suitably low metamorphic grade are now only preserved in the central and southern parts of the Cape St. John Group, low-grade rocks were probably more extensive when this boulder was eroded. This clast is important because it provides definite evidence that the Fleur de Lys terrain had been at least partly deformed and metamorphosed before deposition of the Baie Verte Group volcanics and sediments, and also that they were deposited close to the Fleur de Lys terrain.

5. Quartz-pebble conglomerate, quartzite, siltstone, vein quartz.

These clast types comprise almost all of the boulder conglomerate found on either side of the Baie Verte Road around 7.8 km north of the Burlington Road Junction, described in outline previously. The greenish quartzite clasts show subangular to subrounded grains 0.04 to 2 mm of clastic quartz with a few albite grains in a very fine-grained matrix of recrystallised quartz (and albite?) with very fine-grained pale green mica and variable amounts

of calcite.

The siltstone clasts are all less than a few centimetres across, and show variable amounts of silt-grade subangular quartz, or none, in a brown turbid ultrafine-grained irresolvable matrix that is probably quartz rich as the clasts are not (relatively) strongly-deformed. They resemble the silty parts of the schistose boulder described above.

Thin sections of parts of quartz-pebble conglomerate boulders show a self-supporting framework of well rounded clasts, mostly of quartz, with sizes mostly between 0.2 and 8 mm ranging from 0.02 to 20 mm. The following clast-types were seen: single crystals of unstrained quartz, strained quartz, quartz containing muscovite flakes, or biotite flakes, or an epidote grain, quartz with resorbtion cavities, albite, K-feldspar; multiple grain clasts of coarse grained (plutonic) quartz-albite, quartz-K-feldspar, and quartz-albite-K-feldspar (granite); vein quartz aggregates, sutured quartz aggregates, and brown turbid irresolvable fine-grained clasts to angular quartz siltstone with this as matrix (identical to the separate siltstone clasts in the boulder conglomerate outside). In hand specimen, K-feldspar is often an olive green (amazonite?), quartz is clear or bright blue, and in the largest boulder seen there is also rose quartz, and siltstone is brown, green or red. The matrix of the quartz-pebble conglomerate is wholly re-crystallised and very fine grained, and the rock originally may have been bimodal. It consists of very fine-grained granular quartz (and minor albite?), and very fine-grained pale greenish mica either randomly oriented or slightly aligned. The matrix of some boulders also contains abundant euhedral pyrite.

No bedding has been seen in any clast of quartzite or quartz-pebble conglomerate, nor has any clearly defined pre-depositional deformation fabric been detected. However, rock very similar to the quartz-pebble conglomerate, including blue quartz, occurs in the Cape St. John Group. A

Fig. 4B.12. Quartz-pebble conglomerate; clast from Baie Verte Group compared with Cape St. John Group sediment.
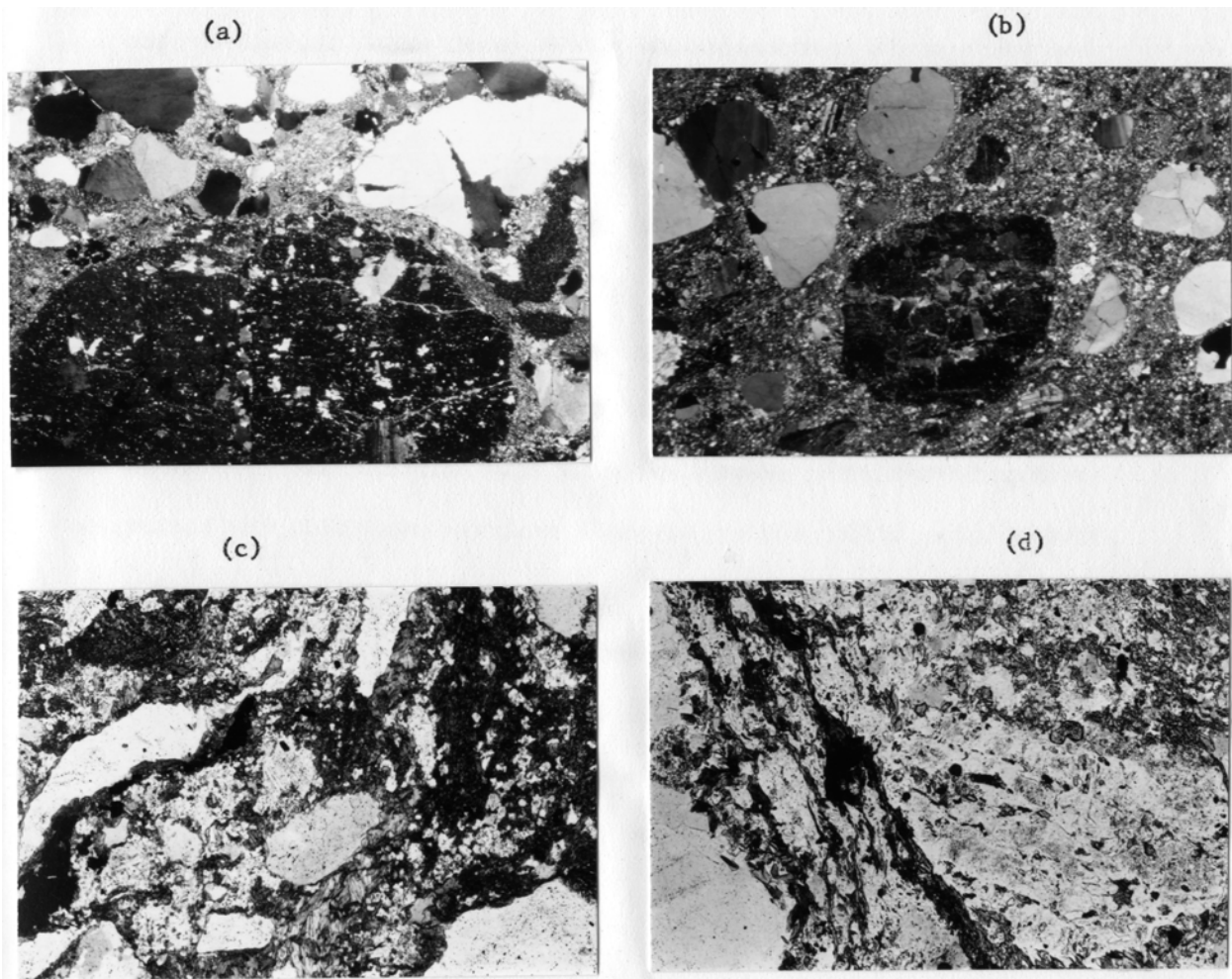
(a) Baie Verte Group clast; rounded clasts of ±K-feldspar, quartz and albite (X 10).

(b) Cape St. John Group sediment; rounded grains of K-feldspar, quartz and albite (X 10).

(c) Baie Verte Group clast; matrix of quartz, albite, pale green mica (X 40)

(d) Cape St. John Group sediment; matrix of quartz, albite, pale green mica (X 40).

sample of such weakly foliated 'quartz-feldspar porphyry', taken from an outcrop in the Cape St. John Group that contains cobbles of rhyolite and suspected bedding, was thin sectioned. It shows a quite astounding resem blance to the quartz-pebble conglomerate clasts, not only in the types and form of the pebbles, but also in the grain size, composition and fabric of the very fine-grained matrix (Fig. 4B.l2). A sample of quartz-pebble conglomerate with blue quartz from the Mings Bight Group on the eastern side of Mings Bight, and one of pebble conglomerate contains quartz from the Pacquet Harbour Group near Rambler Mine were also sectioned. Both have far more coarse-grained and schistose matrices than the quartz pebble conglomerate clasts, which is an allowable factor, but their clast assemblages do not resemble that of the quartz pebble conglomerate boulders so closely as the sample from the Cape St. John Group. It is, therefore, suggested that the quartz-pebble conglomerate, quartzite, siltstone, and vein quartz clasts in this quartz-rich boulder conglomerate were derived from the Cape St. John Group.

6. Jasper

Pale maroon jaspery chert clasts form a few percent of the assemblage in the quartz-rich boulder conglomerate. They are not larger than 10 cm long, and are apparently strongly flattened, although most of this may be due to their original shape. They are mostly composed of fine-grained recrystallised quartz that contains pyrite, and also a few grains of (detrital) chromite. These clasts are most probably from an ophiolite-complex association, as are the small ultramafic clasts also found in this conglomerate. They might be derived from old ophiolite complex inclusions in the Cape St. John Group, or from the younger ophiolite complex associated with the Baie Verte Group.

7. Trachyte

Not more than six pebbles between 2 and 5 cm across of a white siliceous rock resembling quartzite were seen in conglomeratic mafic volcaniclastic beds of the Jukes Point Formation on Flatwater Pond and the Burlington Road. One was sectioned and is an undeformed trachyte with euhedral sodic plagioclase phenocrysts. It was probably derived from a volcanic centre that was providing the mafic volcaniclastic debris in the same rocks.

8. Marble

The 10 m boulder of grey marble found at Teardrop Pond is composed of wholly recrystallised calcite, and is microbrecciated with calcite veins. The pebbles in the calcirudite lenses found on the northwest shore of Flatwater Pond are similarly wholly recrystallised and microbrecciated. The banded and laminated grey marble found with the calcirudite lenses, and also under the power line km NE of Teardrop Pond, is also totally recrystallised, but retains a grain size contrast between the different bedding laminae. It almost certainly was turbidite deposited calcarenite and calcisiltite. These recrystallised limestone occurrences are all at more or less the same horizon, and are suggested to have been derived from a small, perhaps short-lived, reef surrounding either a volcano, or islands consisting either of ophiolite complex lithologies, or of Burlington Granodiorite/Cape St. John Group lithologies, such as are also found as boulders in this part of the Baie Verte Group.

9. Slumped black slate and quartz-rich greywacke

At the west side of the outcrop 1.2 km north of the north end of Mic Mac Lake [40], a slump-folded raft of bedded gritty graded greywacke, black slate, and green cherty argillite is found within black slaty matrix conglomerate

of the Kidney Pond Formation. This raft is at least 5 x 17 metres in exposed section. The greywacke is quartz-rich, and is the only clast of this type seen in the area. The grit-sized clasts in the greywacke are a sorted and fine-grained version of the clast assemblage in the surrounding conglomerate. That is, the subrounded to rounded quartz is dominant, accompanied by subordinate actinolite (after clinopyroxene), saussauritised plagioclase, hornblende, and altered ilmenite. Lithic clasts consist of granular quartz, quartz-plagioclase aggregates, silicic volcanic clasts with corroded quartz phenocrysts, small pieces of granular gabbro, mafic lava, and diabase, and chips of green and grey to black argillite. Thus pre-depositional sorting has increased the quartz-bearing granodiorite and silicic volcanic components at the expense of the mafic (gabbro, diabase, lava) components, relative to the conglomerate of the Kidney Pond Formation. This greywacke therefore provides clear evidence that all the non-sedimentary clast-types in the Kidney Pond Formation were exposed subaerially. It also shows that subaerially-eroded material was sorted in shallow water before its emplacement, first as a turbidite into an area of black and green mud deposition, and secondly within this large raft incorporated by slumping into the conglomerate.

10. Clastic quartz.

Quartz-bearing pebbles and boulders are only found in the Kidney Pond Formation, in its presumed equivalent to the north, and in one bed in the unassigned sequence on the Pittmans Pond woods road. Individual clastic quartz grains are abundant in the quartz-rich greywacke in the slump 'raft' in the Kidney Pond Formation, described above, and commonly form a small proportion of the matrix of the conglomerates in that Formation. A few quartz grains are found in a few mafic volcaniclastic beds but apart from these, only six beds containing a significant quantity of clastic quartz grains

(or presumed to have done so) have been detected within the sequence above the Kidney Pond Formation. These have been described previously. One is in the Jukes Point Formation, three are in the middle of the Slink Pond Formation, and two are in the unassigned sequence west of Black Brook Bridge. The beds are not obviously graded, and range from 5 cm to perhaps 2 metres thick. All have a large proportion of calcite in the recrystallised matrix, but this is probably not due to carbonate detritus. Some quartz grains in the beds in the Black Brook section have resorbtion cavities, and some are blue quartz, and were probably derived from the Cape St. John Group, as were some of the grains in the Kidney Pond Formation. Erosion of silicic volcanic centres that were coeval with the volcanics preserved in the Baie Verte Group may have also provided the quartz in some of the other beds.

(c) Clast provenance — summary and inferences

Sedimentary structures that give a clear provenance direction have only been seen in one outcrop, 0.72 km north of the south end of Slink Pond. Two separate examples of small asymmetric slump folds in thin silty volcaniclastic beds interbedded with cherty argillite laminae were seen 1 metre above the top of the Kidney Pond Formation and both indicated movement more or less directly up dip. This is equivalent to movement from the east towards the west when the vertical beds are restored to horizontal by a strike rotation. If a set of two consecutive rotations is used, the first may deviate considerably from a strike rotation without affecting the conclusion of a general east to west transport direction.

The granodiorite, silicic volcanic, and quartz-rich sediment clasts are suggested to be derived from the Burlington Granodiorite and the Cape St. John Group. These two rock units are now only found to the east of the Baie Verte Lineament, in the eastern Burlington Peninsula, and locally in the immediately adjoining part of the Halls Bay Zone to the southeast (Fig. 1.3).

It is possible to construct models (by tectonic removal of some Burlington Granodiorite sideways or downwards, and erosion of the Cape St. John Group) that result in the occurrence of both lithologies on the western side of the Baie Verte Group during its deposition (assuming it to be autochthonous in the broadest sense with respect to eastern and western Fleur de Lys terrains, a point justified later). However, given the slim evidence from the slump folds of a palaeoslope from east to west, this is not necessary at present, and it is proposed that the granodiorite and Cape St. John clasts, and by inference all the others associated with them, were derived from an easterly direction.

This terrain therefore contained subaerial exposures of Burlington Granodiorite, deformed Cape St. John Group, and an undeformed ophiolite complex down to the lower gabbro (containing clinopyroxenite layers), and in the north some depleted ultramafic rocks, at the time of the Kidney Pond Formation. There must also have been a large fault scarp consisting of ophiolite complex gabbro, and locally of depleted ultramafic rock, probably located not more than a few kilometres away from the present section, to provide the large blocks in the Boudin Pond, Teardrop Pond, Kidney Pond and Neale's Bay Formations. A local, probably ephemeral, carbonate reef was present in this source area and because of the size of the large carbonate boulder found was probably on top of the fault scarp. However, during most of the time of deposition of the Baie Verte Group, this source area must have consisted mainly or entirely of a volcano or volcanoes, consisting mainly of feldspar-phyric mafic volcanics, and with local rhyolitic centres mostly during the later part of the time represented by the section preserved.

    (v) Sedimentary structures; transport and deposition of sediments.

Sedimentary structures in the mafic volcaniciastic and subordinate argillaceous and cherty sediments do not show any evidence of reworking by wave or current action. Essentially, all bedding seen is planar with very rare exceptions. Graded bedding is uncommon but is the most frequent structure seen in the mafic volcaniclastics and is mainly found in the conglomeratic beds of the Jukes Point Formation. Small slump folds are found in three localities in thin-bedded fine-grained mafic volcaniclastic and argillaceous sediments but it is only possible to observe the transport direction in the one locality mentioned above. Very small washouts (channeling) a few millimetres deep are occasionally seen in argillite laminae. It is possible that some or all of these sections through small flute or groove casts, but this was not determinable. On the northern shore of Flatwater Pond, just above the amphibole-bearing sediments, three examples of beds with overturned flow casts at the base (probably deformed load casts or flute casts) were seen but the exposure did not permit determination of their orientation. In laminated green argillite, in the westernmost roadcut on the Burlington Road, occasional examples of small-scale truncated laminations resembling low-angle cross-bedding are not sedimentary structures, but are due to small-scale tectonic slides disrupting minor isoclinal folds (described in detail later). Slump folding is seen in two localities in the Kidney Pond Conglomerate Formation. One locality is in the outcrop near Mic Mac Lake, described previously, where the slump-folded quartz-rich graded greywacke and black slate is a large raft in the conglomerate. The other locality is an outcrop within the Formation on the northern shore of Flatwater Pond. Black and grey slate and silty slate with sparse and small (< 1 cm) pebbles, mostly of granodioritic, silicic volcanic and argillite derivation, displays slump folds of its rather cryptic thin bedding. This bedded slump-folded

material is also probably a large raft within the conglomerate. The direction of slump transport could not be interpreted from either of these two occurrences.

Argillite pebbles supported in a slaty argillaceous matrix, seen under a gabbro block in the type section of the Boudin Pond Formation, are clearly locally derived. They were probably generated by the emplacement of the overlying gabbro block, as a micro-melange similar to that found on a much larger scale under allochthonous gravity slide sheets like the Bay of Islands Allochthon in western Newfoundland. The silty mafic volcaniclastic pebbles in identical mafic volcaniclastic rock, found under the clinopyroxenite boulder in the Teardrop Pond Formation, represent the same situation. Other occurrences of pebbly argillites (without directly associated boulders) are similarly due to mobilisation and variable resuspension of unlithified sediment, and breakup and incorporation of underlying or adjacent partially-lithified sediment as clasts. This process can result from passage of a large boulder over the sediment, or from any other type or combination of types of mechanical disturbance of a marginally stable accumulation of sediment (by earthquake, eruption, abnormal sediment dumping, slumping, or storm activity). A marginally stable sediment accumulation requires a (relatively) steep slope on the edge of a basin, and this may be produced by an active fault scarp, or an active volcanic centre, or maintenance or progradation of a sedimented slope originally formed (but not necessarily maintained) by faulting. Depending on the degree of resuspension achieved, the downslope sediment movement may range from a coherent slump, through a viscous mudflow (with or without large clasts), to a turbidity current. The viscous mudflows (fluxoturbidites) especially are likely to incorporate some of the sediment they override, as matrix and/or clasts. These may

include sediments beyond the foot of the steep slope that would otherwise have been stable. Matrix-supported pebbles to boulders of argillaceous, cherty, and sandy mafic volcaniclastic sediment that are essentially identical to adjacent bedded sediments, are found in the Teardrop Pond, Kidney Pond, Jukes' Point, lowest Slink Pond and southern Neales' Bay Formations. They were therefore derived by mechanical disturbance of partially lithified sediments on a steep slope, and probably also by disturbance of similar sediments beyond the foot of the steep slope due to the passage of the resulting viscous mudflows and turbidites. Graded pebbly to bouldery conglomeratic beds in the Jukes' Point Formation indicate considerable resuspension of the matrix and turbidite-type emplacement. Adjacent beds, containing large matrix supported clasts, indicate less resuspension and a more viscous mudflow-type (fluxoturbidite) emplacement mechanism. The two slump rafts in the black slaty matrix conglomerate of the Kidney Pond Formation show that it may have started partly or wholly as a slump, and perhaps changed to a mudflow as the black slate became remobilised into the present homogenous matrix. The green sandy matrix conglomerate in the Kidney Pond Formation has a matrix composed of small clasts and crystal fragments that are mainly referable to the ophiolitic clasts, with a few quartz grains referable to the quartz-bearing clasts. As this conglomerate appears to have a self-supporting clast framework and is not apparently graded, the method of its deposition is not clear. The lower part of a steep submarine slope is characterised by channels containing pebbly to conglomeratic viscous mudflow (fluxoturbidite) deposits (Stanley and Unrug, 1972). The outcrop in the inland map area is not adequate to determine whether most of these matrix-supported conglomerates are in channels. However, the black slaty matrix conglomerate of the Kidney Pond Formation (clearly a viscous mudf low) is very extensive and continuous along strike, although it varies

somewhat in thickness. It cannot be positively determined whether it is composed of several

units separate along strike, but it is unlikely to consist of many channels because of its

homogeneity, although the thickness variations may indicate very wide shallow channels. The

meagre slump-fold evidence suggests that the section now exposed at the surface is

approximately perpendicular to its emplacement direction. Therefore the continuity of this unit is

evidence that this section is not part of the channeled lower part of the steep slope at the edge of

the basin, but is part of the gentler slope beyond. However, it cannot be far beyond the bottom of

the steep slope because the large (10 m +) blocks of gabbro, ultramafic rock, and marble, in this

and other Formations, which represent submarine rockfalls, are very unlikely to be more than a

few kilometres from their fault-scarp source. Pebble-size lithic clasts in the least deformed

exposures of the Kidney Pond Formation are mostly fairly well-rounded, although some are very

angular. The evidence from the quartz-rich greywacke in the slump raft in the Kidney Pond

Formation shows that all lithic clast-types were available for subaerial erosion, and it seems

likely that all lithic clasts in this Formation were eroded subaerially (on land or in shoreline

cliffs). Similarly most of the lithic mafic volcanic clasts in little deformed exposures of

conglomeratic mafic volcaniclastic beds in the Jukes Point Formation are also fairly well

rounded. It is probable that these lithic clasts were also subaerially eroded, although it is possible

that some rounded mafic volcanic clasts could have been produced during explosive eruptions. It

Is not clear whether the viscous mudflows and turbidites that contain these lithic clasts that have

passed through shallow water, were triggered by the clasts being dumped onto the

unconsolidated matrix sediment or by other processes sometime after the clasts were deposited

on the sediment that became the matrix. Clasts of the order of 1 to 2 metres across,

specifically of Burlington Granodiorite and Cape St. John Group lithologies, could have been eroded subaerially as long as the river slopes were continuously steep to the shoreline. It is perhaps more probable that these larger boulders were eroded from shoreline cliffs. However, in either case, it is necessary that the width of the shelf from shore line to the top of the slope into the basin be narrow. This probably was achieved or accentuated by a submarine canyon-head occurring very close to the shoreline, and it is most likely that the subaerially shoreline eroded lithic clasts were transported into the basin down a submarine canyon.

Beds of mafic volcaniclastics usually contain equant abraded (altered) plagioclase crystal fragments, and less commonly have fairly euhedral laths. Similarly, mafic crystals in the rare beds seen to contain them may be abraded, or non-abraded euhedral grains. Probable small pumiceous clasts are now too deformed and recrystallised in all examples seen to comment on their original shape. Many of the volcaniclastic beds in the area north of the Camp 166 Road, in the eastern part of the Baie Verte Group, and north of 5 km north of the Burlington Road, are so deformed and recrystallised that only the largest or no crystal fragments remain. However, the dominance of abraded grains elsewhere suggests that most of the material has passed through a shallow water environment before deposition. The relatively rare examples of euhedral crystals may indicate deposition directly from explosive submarine or subaerial eruption, or indirectly from redeposition of fragmental material deposited around a submarine vent, or merely a short residence time of subaerial fragmental material in shallow water.

Most of the crystal fragments are therefore abraded grains, derived in a subaerial or shallow shoreline environment. It cannot be determined what proportion was derived from lava versus fragmental volcanic sources.

It is not certain if the clastic grains are derived from the same volcanic pile as the lavas and sills

preserved in the sequence, although the great preponderance of plagioclase phenocrysts in these

is evidence that there is a close connection with the source of the mafic volcaniclastic sediments.

The characteristics of most beds of sandy and silty mafic volcaniclastics are those of 'grain

flows' as summarised by Blatt and others (1972), except that many of the beds are thin although

beds thicker than 2 metres would be difficult to detect in most outcrops), and 'dish structure' has

not been seen. The description of ungraded, internally structureless, sharply bounded beds

lacking sole marks and traction structures, and containing 'outsize' clasts in a sandy matrix

perfectly characterises most of the conglomeratic mafic volcaniclastic beds, and without the

outsize clasts, is a good description of almost all the other sandy and silty mafic volcaniclastic

beds. However, they were not 'relatively well sorted'. The rare well graded sandy and

conglomeratic mafic volcaniclastic beds were probably deposited as turbidites. The combined

evidence from the sediments therefore suggests that most if not all of the lithic and crystal clasts

have passed through a shallow water environment. The evidence suggests that most were

deposited by 'grain-flow' , and viscous mudflow mechanisms, with minor turbidite, slump and

rockfall contributions, and therefore represent a very 'proximal' environment close to the base of

a steep, basin-bounding slope. The enormous boulders emplaced by rockfall, and their

accumulation in the Boudin Pond Formation, which is probably a submarine scree, show that this

slope was mostly a fault scarp, at least for the inland map area.

The abundant interstitial green chert in the pillow lavas is mostly structureless and moulded to

the pillows. Occasionally, bedding laminations remain in the chert fragments, well displayed in

an outcrop 1.05 km E of the south end of Slink Pond [41]. Here, the laminations in the

chert fragments are plastically folded, but also truncated at the edges of the fragments. It is clear that all the chert interstitial to the pillows is semi-lithified bedded sediment picked up and incorporated by each advancing lava flow. This process can be seen frozen in progress in two outcrops of dark green pillow lava 0.4 km SE of the south end of Slink Pond [42]. In the northern outcrop a laminated maroon chert bed between two pillow lava flows is seen being squeezed up in the gaps between the pillows resting on it. In the adjacent outcrop to the southwest, two large lenses (~2-3 metres long) of a maroon laminated chert bed are seen incorporated into a dark green flow. Although in these two examples the sediment was perhaps more lithified than the common interstitial green chert seen elsewhere, and the pillow lava flows further from source and perhaps less effective in disrupting the sediment, they illustrate the kind of process involved. Whether the source of the pillow lavas was a central vent or a fissure system, it must have built up above its surroundings. If it remained below wave base, sediments apart from chert or argillite would not be deposited on it between eruptions, as long as fragmental volcanic rocks directly deposited from explosive eruptions were not commonly available, as is thought to be the case. This model accounts for the occurrence of abundant chert (but not other sediment) interstitial in the pillow lavas, but not for the rarity of argillaceous as opposed to sandy and silty volcaniclastic bedded sediments within the pillow lavas In the exposed section. Laminated green chert beds are only seen below the pillow lava in the Jukes Point Formation south of Kidney Pond, and in sediment interbeds in the lowest 100 metres of the southern part of the Slink Pond Formation. Laminated green argillite beds are restricted to the lower third or less of the Slink Pond Formation. Very rare beds of maroon chert associated with the dark green pillow lava are essentially the only bedded argillaceous

sediment seen within the upper part of the Slink Pond Formation. Unless green chert and argillite was originally deposited within the pillow lava section and has been preferentially picked up by the pillow lavas, its absence in most of the section is unexplained. The pillow lava pile of the Slink Pond Formation does seem to have had an effect on the distribution of conglomeratic mafic volcaniclastics. They appear just above the Kidney Pond Formation in the area south from Kidney Pond and are not seen above 100 metres above the top of the Kidney Pond Formation in this area. In the north Flatwater Pond section the first 60 metres above the Kidney Pond Formation do not contain conglomeratic beds, while the next 270 metres contains abundant beds. It is likely that these viscous 'mudflow' deposits were deflected from their original path to the deposition site in the south of the preserved section to a path leading to the northern site by the construction of the pillow lava pile.

(vi) Facing evidence and major structure

Reliable facing (way-up) evidence is scarce in the Baie Verte Group volcanics and sediments. Bedding and cleavage are everywhere coplanar within the limits of any outcrop or thin section observation, except in the hinges of the less than 10 minor Baie Verte D_1 isoclinal fold closures seen in the whole area. This criterion, therefore, is not useful. Facing criteria that were found and that are regarded as wholly reliable are restricted to the following: 1) Wash-outs and truncations of argillaceous laminae (small erosional channels a few millimetres deep, and one example of a truncated slumped layer); 2) One example of a chert bed incipiently disrupted by the overlying pillow lava flow; 3) Three examples of overturned flow casts; 4) Vesicles/amygdales concentrated consistently on one side of the pillows in any outcrop; and 5) Sequences like those reported by Carlisle (1963), involving a sequence in a lava flow of several of the units massive

lava—large pillows—small pillows—pillow breccia/agglomerate—volcaniclastic/ argillaceous sediment. Graded bedding is regarded as a probably reliable criterion, because it is consistent with reliable criteria in the few cases that they can be compared. Very few clear examples of grading have been seen, and used on the map, and they do not critically affect the interpretation of the major structure. Inverse grading is known to occur in 'viscous-mudflows' ('fluxoturbidite', 'grain flow') deposits (Blatt and others, 1973), and in directly deposited tuffs, but has not been seen in the map area.

The shapes of pillows in pillow lavas is often quoted as a facing criterion. In the author's experience, this is an unreliable criterion even in undeformed rocks. It has not been used in the mapping, even though the required pillow shapes were present in a few outcrops, and were consistent with reliable facing indications nearby. The rocks in the area are all deformed, even though a large proportion of the pillow lavas do not show a cleavage, leading to the spurious impression that the pillow shapes are little changed. For example, very strongly deformed pillows in the outcrops on northern Flatwater Pond have typical sections measuring about 1 metre by 4 cm in the plane perpendicular to the maximum elongation direction, but show only the very weakest cleavage. Not only is pillow shape an unreliable facing criterion, but it is also totally unacceptable as a measure of bedding orientation. In undeformed pillow lavas it is vaguely and poorly defined. In deformed lavas the long axis of any pillow section is controlled by the plane of flattening (the cleavage), not bedding, and may be at a high angle to bedding. Bedding and facing measured from pillow lavas are often used in desperation because of the rarity of reliable indicators. In the author's view, it is preferable to use a restricted amount of information known to be reliable, rather than to mix in unreliable data, and therefore to make the reliable data dubious and unidentifiable unless

the whole set is rechecked. Two cautionary tales known to the author illustrate the inadvisability of using bedding or facing criteria from pillow lavas. A section mostly consisting of strongly deformed pillow lavas in the area north of Springdale to the southeast of the Burlington Peninsula was mapped using pillow shapes as a facing criterion, and a series of folds interpreted on this basis (Neale and Nash, 1963). Remapping of this ground using only reliable criteria for facing showed that it consists of a southward facing homocline. (J. R. DeGrace, pers. comm.).

An isolated exposure of pillow lava in the Bay of Exploits area, central Newfoundland, was viewed as one of the stops on a field trip during the Gander Conference in 1967. Several of the participants were overheard separately voicing opinions on the bedding orientation and facing direction, and (inevitably!) these opinions on both items were completely contradictory.

On the basis of the facing evidence and the distribution of the stratigraphic units, the major structure of the Baie Verte Group in the inland map area between Mic Mac Lake and Flatwater Pond is an eastward facing homocline. Subvertical bedding in the west changes more or less gradually to moderately-dipping overturned bedding in the east and overall strikes parallel to the margins of the Lineament. Without the facing evidence, a plausible case can be made for a major fold closure whose axial-trace would run in the centre of the Slink Pond Formation. However, the facing evidence in the area between the Camp 166 Road and Mic Mac Lake is sufficiently common and well-distributed to preclude the existence of any major fold repetition in the overall sequence. In particular, none of the synclinal fold axial-traces shown in the Baie Verte Group on Fig. 8 .1 of Neale and Kennedy (1967) exist, including the one between Slink Pond and White Bay Waters, based on an observation of west-facing plotted where

this author found reliable evidence for eastward facing. The lithological distribution in the area around the northwestern corner of Flatwater Pond, and observations of beds facing both east and west in the western outcrop on the Burlington Road, are interpreted to be due to a medium scale parasitic Baie Verte fold pair modified by tectonic sliding (Plate 2). It is possible that similar fold pairs of this scale or smaller may be present in the Teardrop Pond Formation north of the Burlington Road, and the Neale's Bay Formation east of the large late normal fault that enters Flatwater Pond at its northeastern corner. Facing evidence is lacking in these areas, and stratigraphic control is not firm, although the distribution of maroon argillite and marble in the former, and pink silicic tuff in the latter, do not suggest that any folds are present.

Minor structures in the Baie Verte Group are described and discussed after description of the Mic Mac Lake Group because they are partially shared with that Group and are better classified as minor structures associated with the Baie Verte Lineament as a whole.

(vii) Attempts to date the Baie Verte Group

No direct evidence for the age of the Baie Verte Group is presently available. The clasts of Burlington Granodiorite and Cape St. John Group suggest deposition was younger than most or all of the Fleur de Lys deformation and metamorphism. The structural evidence (discussed later) shows that the Baie Verte deformation is wholly separate and later than the Fleur de Lys deformation. This mapping shows (discussed later) the Baie Verte Group to be older than the Mic Mac Lake Group, which is dated by two independent Rb-Sr isochrons as lower Devonian. The Snooks Arm Group, in the eastern part of the Burlington Peninsula, is dated by graptolites as Arenigian (Sneigrove 1931). It resembles the Baie Verte Group both lithologically, and in the nature of its stratigraphic sequence of pillow lava and

mafic volcaniclastic units resting on an ophiolite complex base. It also contains some clasts of Fleur de Lys-derived rock deformed prior to deposition (Dewey and Bird, 1971). No other sequences in Western Newfoundland resemble the Baie Verte Group nearly as closely, and it is likely on this basis and on another argument set out in the final chapter, that the Baie Verte Group is also Arenigian in age.

During the mapping in the inland map area, all occurrences of grey and black argillite and slate were investigated for fossils insofar as time and the particular exposures permitted (i.e. usually about 10-15 minutes, but occasionally up to 2 hours, were spent in looking). No trace of anything potentially identifiable as a graptolite (let alone a specific identification) was seen, although a few scraps of material seen in one or two localities might be shredded and deformed small pieces of graptolite. Many exposures of black slate are rather strongly deformed, and are perhaps unpromising material for preservation of graptolites. All the more promising less-deformed exposures are typical glacially smoothed 'whalebacks', and in these it is not possible to extract more than a few small scraps of the rock. It is quite possible that a search aided by a drill and dynamite might be productive.

The author dissolved (in HF) a specimen of clast-poor black slate from the Kidney Pond Formation on the northern shore of Flatwater Pond. The residue contained only structureless black flecks, and so study of other specimens was not attempted. However, it is possible that a systematic search of the least-deformed black/grey argillite might produce some resistant microfossils. A pound or so of the calcirudite and banded marble found on the shore of northwest Flatwater Pond was gently dissolved but no conodonts were found. Lars Fahreus of Memorial University was kind enough to dissolve and examine several pounds of the massive microbrecciated

marble boulder found at Teardrop Pond, but he found no conodonts in it (pers. comm., 1973). The buff aphyric silicic volcanics south of Black Brook may be part of the Baie Verte Group, and they might produce an Rb-Sr isochron. Samples were collected for this purpose, but so far have not been processed. The only other possibility for dating the Baie Verte Group lies In obtaining lead ages from zircons in the trondhjemitic rocks in the ophiolite gabbro. It is hoped to pursue this in the near future.

     (viii) Metamorphism

Rocks in the Baie Verte Group in the inland map area (and elsewhere) contain mineral assemblages of the lower greenschist facies, as summarised by Turner (1968). Retrogression of igneous minerals in the mafic rocks is nearly complete. Plagioclase is universally altered to albite, or to other minerals, but clinopyroxene survives in part, or wholly, in some places. Igneous hornblende is not visibly affected in most cases. Large ilmenite grains are only slightly to partly altered in a few places, but most ilmenite is wholly altered.

The survival of clinopyroxene seems to depend partly on its grain size, and therefore in most cases on the grain size of the rock containing it. However, it also seems to depend on the position of the rock within the Baie Verte Lineament. Fig. 4B.13 is a sketch map showing where clinopyroxene has been detected in thin section, and a rough indication of its grain size from the type of rock containing it. In the area south of the Camp 166 Road, the lower half of the succession, including the ophiolite gabbro, contains no relict clinopyroxene. Some coarse grained pyroxene survives in the upper half of the succession in this area, and three examples of fine-grained clinopyroxene are known, two very close to the top of the succession.

Fig. 4B.13. Sketch map showing distribution of relict clinopyroxene detected in the Baie Verte Group.

Flatwater Pond

Kidney Pond →

Occurences in :

G - gabbro sill
D - dolerite sill
L - lava
P - volcaniclastic rock
X - clinopyroxenite
O - ophiolite gabbro
H - harzburgite/dunite

Sketch map at 1 : 126,720
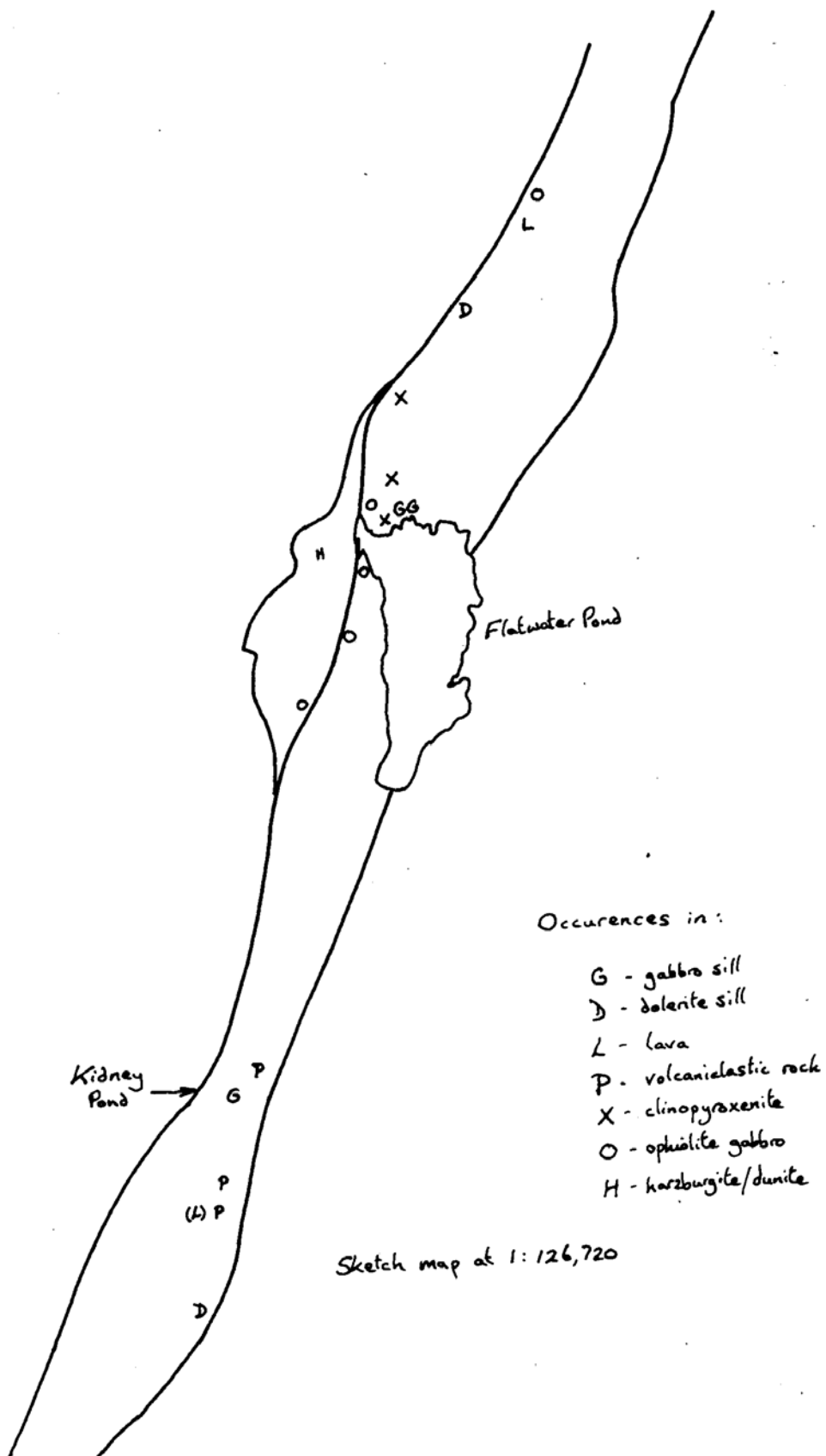
In the other example, fine-grained clinopyroxene partly survives in very little deformed light green pillow lava about two thirds of the way up the Slink Pond Formation. The survival is therefore very patchy, and sample distribution is only adequate in a strip east from Kidney Pond to be fairly sure that the data approximates the actual distribution of clinopyroxene. In the area northward from the southern end of Flatwater Pond, the surviving clinopyroxene is restricted to the lower half of the sequence. Sample distribution is adequate to define this along the northern Flatwater Pond and Burlington Road sections, but there are insufficient occurrences of the coarser dolerite sills in the eastern half of the sequence to be sure of the situation. However, boulders of ophiolite gabbro in the eastern part of the section do not contain any relict clinopyroxene, while they do at the western side.

As deformation is relatively weak and, if anything, stronger in the upper half of the sequence in the southern part of the area, it is possible that the relict clinopyroxene distribution relates to burial alteration. In the northern part of the area, deformation is generally relatively more intense than in the south, but it is not apparent (and not possible to measure) that it is more intense in the eastern relative to the western part of the northern section. Therefore it is a fact, albeit unexplained, that dolerite sills and ophiolite gabbro in the western part of the north area tend to retain clinopyroxene in places while those in the south do not, despite the stronger deformation in the northern area. There is no detected change in metamorphic mineral assemblages between the two areas.

In the metamorphic alteration of igneous minerals, clinopyroxene in the coarser rocks (dolerite and gabbro) is altered wholly or mainly to a very pale or colourless actinolite. In diabase dykes, mafic lava, and grains in mafic volcaniclastics, chlorite may be the dominant alteration

mineral, although pale actinolite usually dominates in pale green pillow lava. Plagloclase is everywhere altered. It may range from fairly clear albite to albite crowded with microinclusions of either a sericitic mineral (? pyrophyllite or paragonite) and/or zoisite/clinozoisite ('saussaurite'). Plagioclase is alternatively sometimes altered to a turbid brown fine to coarse grained aggregate almost wholly composed of zoisite and clinozoisite. Ilmenite is most often wholly altered, either to a turbid irresolvable brown aggregate (leucoxene), or more commonly to a slightly coarser grained version of this that is detectably composed of very fine-grained sphene. Some large ilmenite grains may be nearly unaltered, or have an unaltered core surrounded by a rim of sphene, but are most commonly more or less totally replaced by relatively coarse grained sphene. In most examples of dolerite and gabbro sills, and of ophiolite gabbro and diabase, minerals are pseudomorphically replaced with little or no overgrowth of the original texture. In most examples of mafic lava and the matrix of the volcaniclastic sediments, metamorphic minerals randomly overgrow and destroy the original textures.

Generalised mineral assemblages found in the various rock types are set out below. No systematic change in the distribution of the assemblages has been found in the map area, apart from the distribution of relict clinopyroxene.

1. Mafic

(a) gabbro, dolerite, little deformed lava

Actinolite-albite ± sphene ± clinozoisite/Al-epidote ± calcite ± chlorite ± pyrite ± sericite (± ? quartz in groundmass). Occasional zoisite, Fe-epidote; rare oxidised vermicular chlorite, stilpnomelane. Occasional relict ilmenite, hornblende, brown/orange amphibole, rare clinopyroxene, apatite.

(b) Cleaved lava, mafic volcaniclastics, slide facies schist

Chlorite-albite ± actinolite ± calcite ± Al-epidote/clinozoisite ±sphene ± pyrite ± sericite (± ? quartz in groundmass). Occasional Fe-epidote, zoisite, rare oxidised vermicular chlorite, stilpnomelane (?), ankerite, Fe-bearing magnesite, fuchsite. Rare relict clincpyroxene, orange amphibole, ilmenite, apatite.

    2. Mg-rich Ultramafic

        (a) serpentinised only: antigorite - lizardite and/or nephrite - magnetite ±

         magnesite ± brucite. (relict olivine, chromite, clinopyroxene, v. rare enstatite)

        (b) other alterations:

            (i) Fe-bearing magnesite - quartz -fuchsite ± pyrite ± magnetite.

            (relict chromite).

            (ii) Talc – Fe-bearing magnesite – antigorite

            (iii) Chlorite - (relict chromite)

        (c) tuffisite: Talc - lizardite

    3. Calcareous

Calcite ± (pyrite ± ?dolomite ± chlorite ± white mica ± carbonaceous matter) (± rare clastic quartz, albite, ilmenite)

    4. Siliceous

        (a) Chert

            (i) maroon: Quartz - haematite ± magnetite ± pyrite

            (± epidote ± sericite ± chlorite ± ?rutile) (± relict chromite).

            (ii) green: Quartz - chlorite(?) ± actinolite ± clinozoisite/Al-epidote.

        (b) Tuff: Albite - quartz - white mica - pyrite ± calcite

        (c) Sediment: Quartz - albite - calcite - chlorite ± white mica ± pyrite.

In the mafic rocks, the actinolite is most commonly nearly colourless, and ranges from colourless to a medium pale green. The epidote mineral is

usually either an aluminous epidote or clinozoisite, and Fe-rich epidote is uncommon. Chlorite is usually more common in mafic volcaniclastic rocks than in lavas, dolerites and gabbros, where actinolite is the dominant mafic mineral. Stilpnomelane is found in deformed dark green pillow lava, and a brown-orange mineral that resembles it, but which is probably oxidised vermicular chlorite, is also seen in these rocks, and in a few other localities. Prehnite is found in a few veins in ophiolite gabbro, but neither prehnite nor pumpellyite have been seen in the volcanics and sediments. A few examples of crudely spherulitic yellowish calc-silicate mineral aggregates in volcaniclastic sediment thought at first to be prehnite are probably Fe-rich epidote. A maroon chert from near the Prairie Hat consists of wholly recrystallised quartz containing laminae rich in magnetite octahedra. Minerals in argillite and slate are usually irresolvable and are, presumably, mainly chlorite and sericite. Occasional veins containing quartz seen in mafic volcaniclastic rocks perhaps indicate that it is present together with albite in the very fine-grained recrystallised groundmass of these rocks.

Much of the mineral growth in the rocks is either replacement of igneous grains, or is very fine grained, so it is difficult to define the time of growth of most of the metamorphic minerals relative to the deformation. However, no conclusive evidence has been seen for substantial growth of minerals before or after the deformation. The best evidence comes from veins, that may contain one or more of albite, quartz, epidote, chlorite and calcite. Some are seen deformed in the cleavage, indicating formation prior to or synchronous with the cleavage, some are later than the cleavage, and one was seen cutting the cleavage but folded by a local secondary crenulation. In cleaved rocks, most chlorite defines the cleavage, and therefore grew during its formation. Actinolite is somewhat aligned in rocks containing a 'longrain' (preferred elongation direction) in the main cleavage,

and therefore also grew during the deformation. Some of the larger clinozoisite and Al-epidote crystals are probably post-cleavage as they are not augened. Most but not all pyrite grains do not have strain shadows, and are also probably post-cleavage. Otherwise, most minerals as now seen probably crystallised during the deformation, shortly before, during, or shortly after formation of the one regional cleavage. However, this does not rule out the possibility of significant, but fine-grained alteration having occurred previously during burial. Conditions during low greenschist-facies metamorphism (Turner, 1968), involve a fairly restricted temperature range (250°—350°C, probably not more than 300° C), but a large and only slightly temperature-dependent pressure range, from about 2kB upward. The depth of burial during and after deformation, estimated from the regional geology, almost certainly did not exceed 9 km, equivalent to a pressure of about 3 kB, and was probably less.

CHAPTER 4C. MIC MAC LAKE GROUP

(i) Introduction

A sequence of moderately west-dipping, mostly west-facing subaerial silicic and mafic volcanics with coarse clastic sediments adjoins the eastern side of the belt of mafic rocks of the Baie Verte Group southward from a few kilometres north of Flatwater Pond. (Plate 6). It forms a belt that is nowhere wider than 2.6 km. This sequence rests unconformably on the Burlington Granodiorite (Neale and Nash, 1963). The relationship between it and the Baie Verte Group is rather cryptic, and has been interpreted as a fault obscured by rhyolite intrusions (Neale and Nash, 1963), or conformable with the Baie Verte Group overlying the contact (Neale and Kennedy, 1967). A different interpretation is made on the basis of the data collected during this project, that this sequence unconformably overlies the Baie Verte Group along part of the contact. In other parts, the contact is a tectonic junction.

This sequence has never been formally and separately named. Formally named sequences containing large amounts of silicic volcanics are found nearby, in the Burlington Peninsula (Cape. St. John Group), and in the area to the southeast of the Peninsula (Springdale Group). Prior to the recognition that the Cape St. John Group was at least partly (Church, 1969), and probably wholly (Dewey and Bird, 1971) involved in the Fleur de Lys metamorphism and deformation, and is much older than the Springdale Group, these two groups were, at least inferentially, correlated This sequence that adjoins the Baie Verte Group was correlated with both (Neale and others, 1960, Neale and Nash, 1963, Neale and Kennedy, 1967, Church, 1969), and in one instance (Neale, 1958b) a small part was placed in the Cape St. John Group. This sequence in the map area was informally termed the Mic Mac Lake sequence by Neale and Kennedy (1967) It is clearly not part of the Cape St John Group. It is

certainly correlative in broad lithological terms with the Springdale Group, but it could be somewhat older or younger. Indeed, Neale and Nash (1963) placed this Mic Mac Lake sequence in a different map unit than the Springdale Group. To clarify the situation, it is proposed to name the sequence formally the Mic Mac Lake Group. The alternative, to assign it to the Springdale Group, is rejected for the reasons set out below.

In the Burlington Peninsula and the region adjoining it (Fig. 1.3), there are three main areas occupied by similar subaerial sequences containing much red porphyritic silicic volcanics, mafic lavas and immature clastic sediments (Neale and Nash, 1963). These are the strip from the Trans-Canada Highway to Flatwater Pond occupied by the Mic Mac Lake sequence, the strip south of the Trans-Canada Highway extending west from opposite the end of the Baie Verte Lineament, and the large area to the southeast of the Burlington Peninsula occupied by the Springdale Group. The second of these areas was included in the same map unit as the Mic Mac Lake sequence by Neale and Nash (1963). However, the Mic Mac Lake sequence is separated from the other two areas by the Birchy Lake-Green Bay Fault, that possesses about 20 km dextral displacement (Dewey and Bird, 1971). There is also an elliptical area just to the east of the Mic Mac Lake sequence that consists of silicic volcanics contained in hypabyssal silicic porphyry and syenite intrusions, and there is a half ring dyke of silicic porphyry to the north of it. This is proposed (Neale and Nash, 1963) to be the eroded roots of a caldera complex, and the source for the silicic volcanics of the Mic Mac Lake sequence, and probably for some of the silicic volcanics in the Springdale Group.

Thus, the Mic Mac Lake sequence 1) is on a separate tectonic block from the other two areas; 2) is much closer to the source of the silicic volcanics; 3) contains much coarse conglomerate relative to the other areas;

4) rests unconformably on the Burlington Granodiorite while the basal relationships of the other two areas are poorly known; 5) is preserved in a different structural environment and is generally speaking more strongly deformed than the rocks in the other two areas. Also it will probably be very difficult or impossible to correlate the Mic Mac Lake sequence with the units in the Springdale Group because 1) the lower parts of the Springdale Group (that contain the mafic and silicic volcanics) are poorly exposed, 2) rapid and drastic facies changes and erosional lensing of units are known to occur in the Mic Mac Lake sequence, and 3) there is almost no outcrop of rock that is definitely part of the Mic Mac Lake sequence in the southern 13 km of strike length of the sequence south from Mic Mac Lake, the area that would be most critical for correlation. If it proves unacceptable to give the Mic Mac Lake sequence Group status, it could be made a sub-Group of the Springdale Group, or alternatively the Springdale Group could be raised to Supergroup status. However, it is thought to be justifiable to give the Mic Mac Lake sequence Group status. It is also thought that this will facilitate discussion of the regional geology with the least prejudice to any further correlations that may be made.

The Mic Mac Lake Group has been dated as lower Devonian by two independent Rb-Sr whole rock isochrons. These gave 393 ± 23 m.y. (R. K. Wanless, pers. comm. to Neale and Kennedy, 1967), and 395 ± 5 m.y. (I. R. Pringle, pers. comm.). The Silurian-Devonian boundary is not very accurately dated, and could be between about 395 and 430 m.y. (Lambert, 1971). Therefore the Mic Mac Lake Group may be anywhere from lowest to upper lower Devonian in age.

The Mic Mac Lake Group is well exposed in the burnt area, from a little north of the Burlington Road south to the area around Park Pond. Elsewhere

the exposure is poor, and inadequate to define such detailed map units and relationships found in the well exposed area.

P. A. Randall, who assisted the author during part of the 1970 field season, first mapped the area from the Camp 166 Road for 6 km south, approximately to an east-west line through the middle of White Bay Waters. The author has checked and remapped parts of about 30% of this area, including all parts critical to the palaeotopographic and overall structural interpretation. Although this resulted in a few minor differences, the author found Peter Randall's mapping to be accurate and of high quality. The Mic Mac Lake Group between the south end of Flatwater Pond and 6 km south of Park Pond was mapped on enlarged air photographs of 1:4800 scale, and is presented at this scale in Plates 3 and 4. Plate 3 is the author's mapping between the Camp 166 Road and Flatwater Pond; Plate 4 is Peter Randall's map with corrections and additions by the author, and it contains a compilation diagram showing which parts the author has examined. The author gratefully acknowledges Peter Randall's contribution, and it should be noted that the larger part of this area would not otherwise have been mapped in detail unless another part of the author's overall map area had been left unmapped. The mapping done at 1:4800 scale has been reduced to the scale of, and is presented, slightly generalised, on Plate 1.

The Mic Mac Lake Group in the area from Flatwater Pond to south of Park Pond consists almost wholly of a homoclinal west-facing sequence. In general, dips in the outcrops furthest from the Baie Verte Group are gently west at 20-30 degrees. Westward, dips increase to moderately, and locally steeply, west. One cleavage is seen in suitable lithologies, and may be at a high angle to bedding, although it is often near coplanar. The rocks are

variably deformed, and in general terms the outcrops furthest from the Baie Verte Group are less deformed than most. However, the lithologies vary greatly in ductility, and in most places this is more significant as a control over deformation and cleavage development. Three main generalised groups of lithologies are present in this area of the Mic Mac Lake Group, namely silicic volcanic rocks, mafic lava flows, and immature, mainly conglomeratic, clastic sediments. In general terms, these are grouped into two associations, the mafic lavas and clastic sediments being interbedded and mostly separated from the silicic volcanics. The relationships in the very detailed map area are complicated by original depositional, and especially erosional, lensing of units, and also by local complex palaeotopography. In a few places structural complexity allied with these factors makes small areas uninterpretable in the absence of 100% outcrop.

(ii) Stratigraphy

(a) General Distribution

Most of the Mic Mac Lake Group is a homocline facing west from the unconformity on the Burlington Granodiorite. However, a narrow discontinuous strip of the Mic Mac Lake Group is found from 1.8 km north of the Camp 166 Road southward between the Baie Verte Group and the large high-angle fault that runs from Flatwater Pond through Park Pond to Mic Mac Lake (the Park Pond Fault). This narrow strip is not part of the main west-facing homocline, and it will be shown later that it overlies the Baie Verte Group unconformably and faces east.

The main homoclinal sequence of the Mic Mac Lake Group is divided into two sequences by a major internal erosion surface. These two sequences are defined below as Formations. The lower sequence is almost wholly restricted to the area south of the Camp 166 Road, where it overlies the unconformity on the Burlington Granodiorite. It consists almost wholly of silicic volcanic

rocks with minor conglomerate. The major internal erosion surface at the top of this sequence is overlain by the upper sequence, whose lower part consists of mafic lava flows and conglomerate. Further up interbedded porphyritic silicic volcanic rocks occur in places, and then there is a rapid upward change to wholly porphyritic silicic volcanic rocks. The major erosion surface between the two sequences is not traceable northward of where the mafic lava and conglomerate of the upper sequence disappear 1.0 km south of the Camp 166 Road. This disappearance is due mainly to unmeasurable relative contributions from both nondeposition and erosion during deposition, and is also affected by the structural complexity in this area. If it is not everywhere faulted out, the erosion surface must lie within the section consisting wholly of silicic volcanic rocks north of the point where it is lost. It most probably lies (and in one place has been tentatively identified) fairly close to the unconformity on Burlington Granodiorite in the area just north of the Camp 166 Road. North of a point 1.8 km N of the Camp 166 Road, a sequence consisting almost wholly of mafic lava and conglomerate rests on the unconformity with the Burlington Granodiorite, and appears to underlie the upper silicic volcanics, although a stratigraphic contact between the two is not exposed. It is most likely that this is the same unit as that overlying the internal erosion surface to the south.

Therefore the unconformity on Burlington Granodiorite under this northern mafic lava/conglomerate sequence is the same as the erosion surface within the Group to the south. The unconformity on the Burlington Granodiorite southward from the disappearance of this northern mafic lava/conglomerate sequence is therefore a different, slightly older erosion surface.

(b) Armageddon Formation (lower sequence)

The lower sequence in the Mic Mac Lake Group lies between the unconformity

Fig. 4C.1. Type section of the Armageddon Formation.
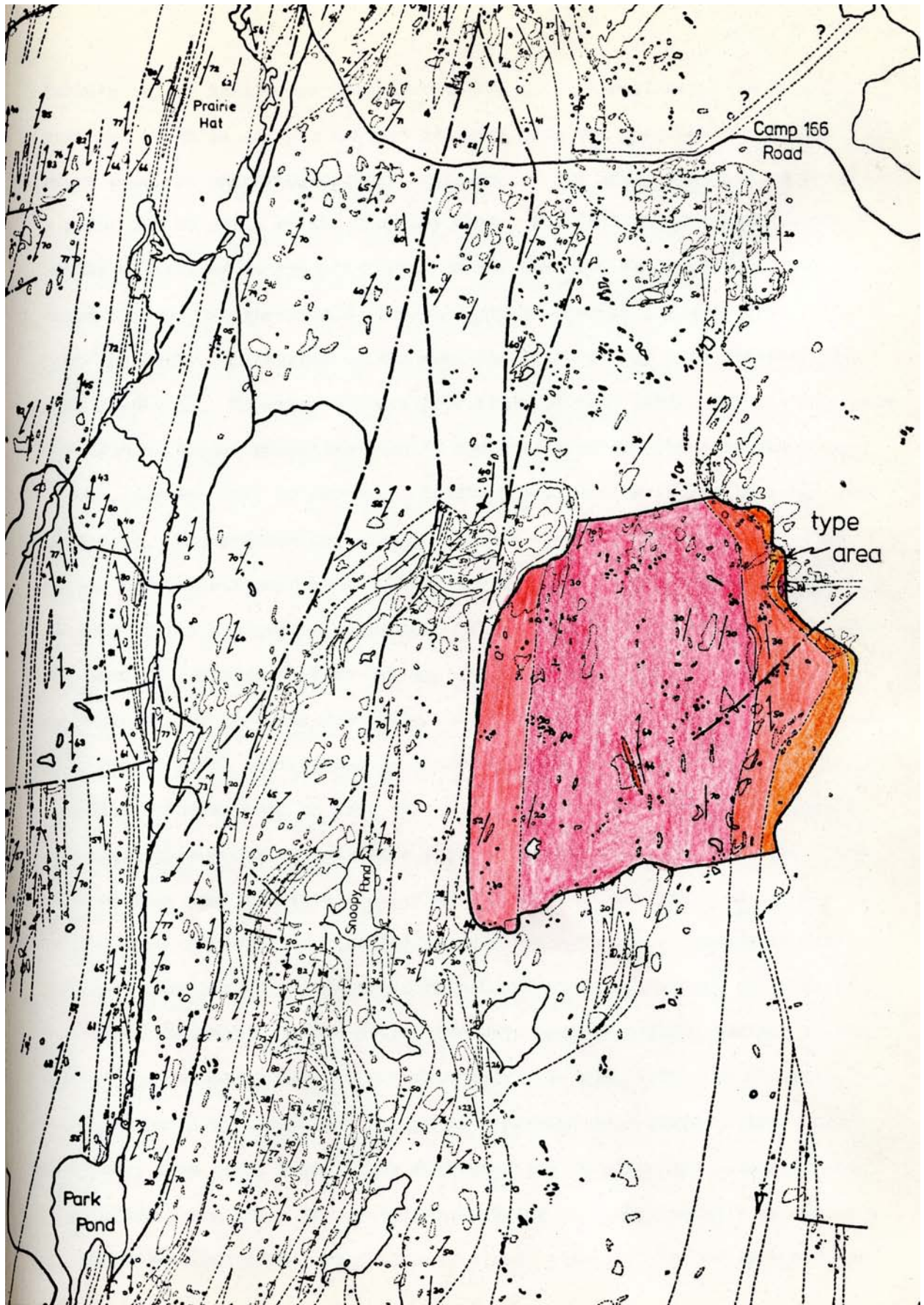
on the Burlington Granodiorite and the internal major erosion surface, which is clearly defined by mafic lava and conglomerate of the upper sequence overlying silicic volcanics of the lower sequence south of a point 1.0 km south of the Camp 166 Road. The lower sequence is almost wholly restricted to the area south of the Camp 166 Road, and has been removed by penecontemporaneous erosion north of a point 1.8 km north of the Camp 166 Road. It extends to the southern limit of the very detailed map area (Plate 4). Exposure is very poor south of this limit, but the available map (E.R.W. Neale, unpublished data) suggests that the lower sequence continues, perhaps with an erosional interruption, southward past Mic Mac Lake (Plate 6). No nameable topographic feature occurs within the area of detailed mapping occupied by the lower sequence. The author therefore chooses to call it the Armageddon Formation. The type section is chosen as a swath 1.1 km wide perpendicular to average strike, whose northern boundary is 1.0 km south of the Camp 166 Road (Fig. 4C.1). Most of this Formation consists of a single thick unit of maroon flow-banded and massive non-porphyritic rhyolite, that is seen to have thin interbeds of other lithologies (one of conglomerate and one of eutaxitic ignimbrite) in only two localities. The base of this unit consists locally of rhyolite autobreccia. In the area of the type section, the rhyolite is overlain by a unit of red quartz-K-feldspar porphyry eutaxitic ignimbrite, essentially identical to those in the upper sequence. To the south, the major palaeo-erosion surface at the top of the Formation has cut out this unit. To the north, the quartz-K-feldspar porphyry ignimbrite cannot be separated from those in the upper sequence; some of it immediately overlying the rhyolite may belong in the Armageddon Formation. The stratigraphy below the rhyolite unit is complex, due to palaeotopographic and palaeoerosional effects. In the type section,

a unit of non-porphyritic, mostly scarlet red eutaxitic ignimbrite occurs immediately below the rhyolite. In the north, near the Camp 166 Road, part of this unit is eroded and replaced by boulder conglomerate. A lens of similar non-porphyritic eutaxitic ignimbrite is found within the boulder conglomerate that lies below the main horizon in this area. In the type section one main unit of a very distinctive pink quartz-K-feldspar-plagioclase porphyry eutaxititic ignimbrite occurs below the main non-porphyritic eutaxitic ignimbrite unit. Erosional remnants of this pink porphyry and its buff cleaved unwelded base are found to the north, including a lens within conglomerate just south of the Camp 166 Road. Below it in the type section conglomerate is found in places resting on the unconformity with the Burlington Granodiorite. A tiny thin remnant of a second lower unit of pink porphyry eutaxitic igniinbrite is found in one outcrop below this conglomerate. To the south of the type section, the main unit of pink porphyry ignimbrite is traceable to the southern end of the detailed map area (Plate 4). It is always near or on that unconformity, except in the poorly exposed area from 2.8 to 4.0 km south of the Camp 166 Road. There, a unit of maroon flow-banded non-porphyritic rhyolite, a second unit of very similar pink porphyry ignimbrite, and some conglomerate underlies it. One rather dubious outcrop of mafic lava at the base of this sequence is the only possible occurrence of a mafic lava flow in the Armageddon Formation. The conglomerates in the lower part of the Formation do contain a few pebbles of mafic lava, although most of the clasts are granodiorite and non-porphyritic maroon rhyolite, with rare quartz-feldspar porphyry. A purplish-red non-porphyritic rhyolite containing sparse streaky thin axiolitically devitrified lenses has been distinguished from the maroon flow-banded rhyolite. It is found on and to the north of the Camp 166 Road, and is interpreted to have been extruded from a

dyke reaching the surface at the Camp 166 Road, forming a tholoid above it, and a flow to the north. The top of this rhyolite at a point 0.4 km north of the Camp 166 Road is interpreted to be the major erosion surface between the Armageddon Formation and the upper sequence, but elsewhere some of the porphyritic ignimbrite above it may belong to the Armageddon Formation. The thickness of the Armageddon Formation in the type section is a maximum of 700 metres. The maximum thickness preserved is about 750 metres.

(c) Snoopy Pond Formation (upper sequence)

This Formation overlies the erosion surface cut into the Armageddon Formation south of the Camp 166 Road, and overlies the same erosion surface where it becomes the unconformity on the Burlington Granodiorite 1.8 km N of the Camp 166 Road. The upper contact of the Formation is everywhere one of two kinds of tectonic boundary, and the stratigraphic top contact is defined as not seen. Thus, this Formation includes all of the Mic Mac Lake Group in the map area as far west as the Park Pond Fault, with the exception of the Armageddon Formation as defined above. In general terms, the lithological sequence in the Snoopy Pond Formation south of the Camp 166 Road starts with mafic lavas and conglomerates. Upward, a few limited red quartz-K-feldspar porphyry eutaxitic ignimbrites are interbedded with more mafic lavas and conglomerates. These conglomerates pass southward with decreasing clast size into a sequence mostly of sandstone. Near the top of the preserved sequence just north of Park Pond, there is an abrupt upward change to a sequence wholly of silicic volcanics. These are highly deformed near Park Pond, but northward they consist wholly of red quartz-K-feldspar porphyry eutaxitic ignimbrites, a large proportion of which are separated from the rest of the Snoopy Pond Formation by a large high-angle fault. In this separate fault block, north of the Camp 166 Road, several non-porphyritic maroon rhyolite sills also occur.

Fig. 4C.2. Type section of the lower part of the Snoopy Pond Formation.
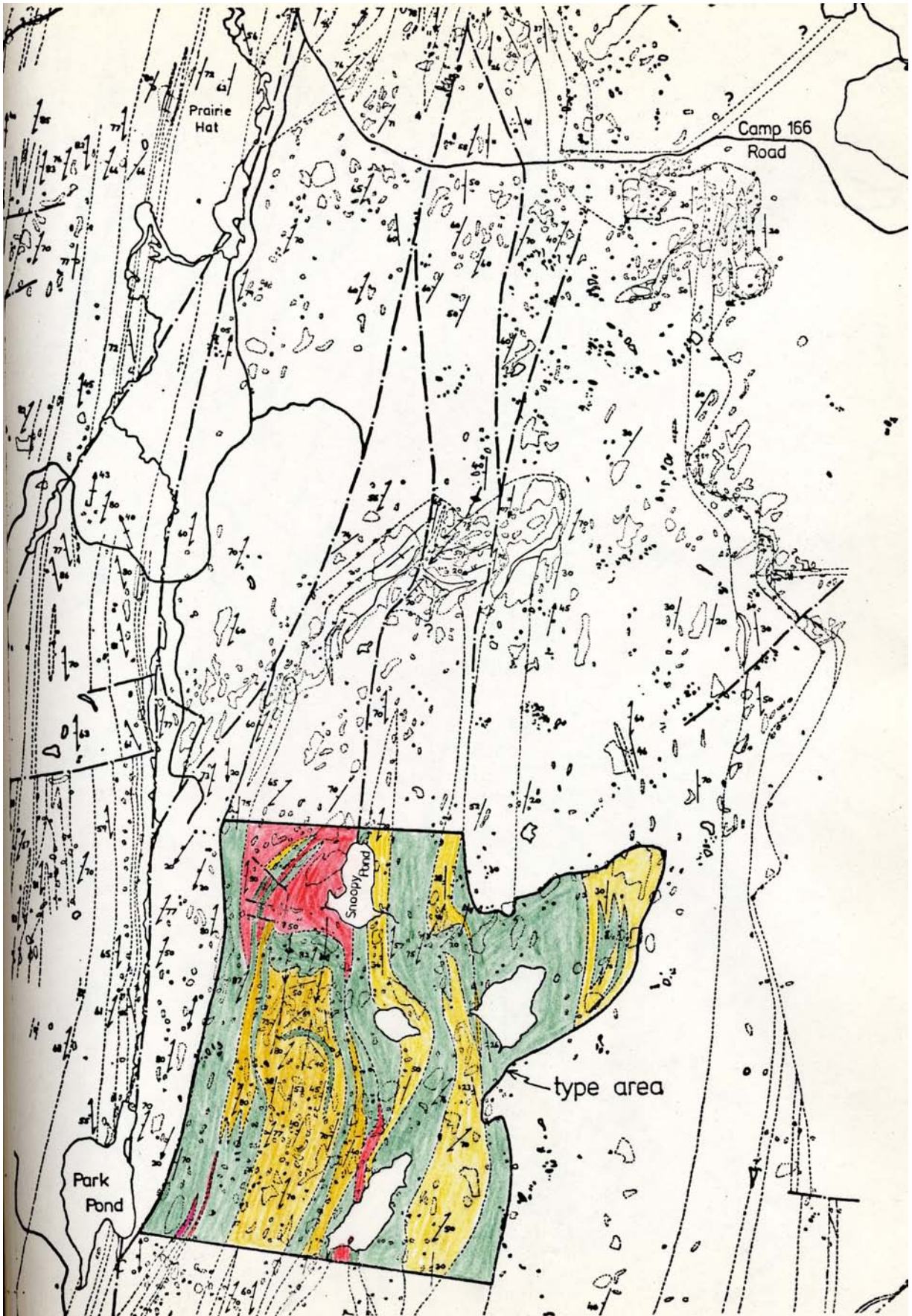
Fig. 4C.3. Type section of the upper part of the Snoopy Pond Formation.
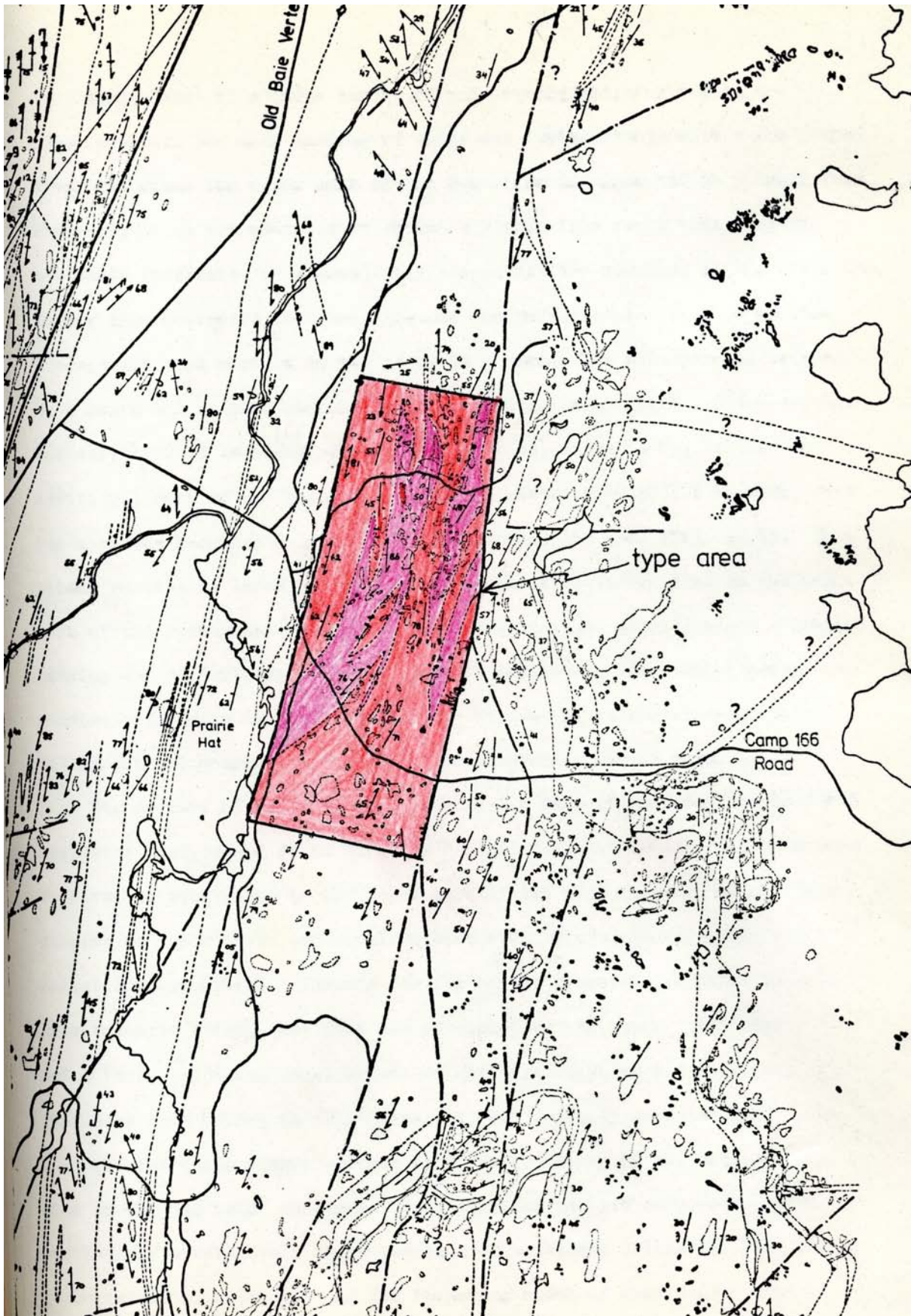
4C.10

Since there is a large amount of palaeotopographic relief on the lower contact, and much lensing of units and facies change within the Formation, and since the upper part of the Formation is separated by a fault from the rest, it is not possible to define a single type swath that includes the whole thickness, or a completely representative section, of the Formation. Rather than designate the whole area of the Formation as its type section, a composite type section in two parts is chosen. The main part is defined as a swath 1.2 km wide approximately perpendicular to strike, whose northern boundary is 2.15 km south of the Camp 166 Road, shown in Fig. 4C.2. An additional part of the type section is defined as a swath 1.4 km wide, with its northern boundary 1.16 km north of the Camp 166 Road (Fig. 4C.3). This second section is added to include the porphyritic ignimbrites in the upper part of the Formation. However, there is an unknown stratigraphic thickness missing due to faulting between these two sections. This rather messy compromise definition is the best that can be done given the stratigraphic and structural complexities, and the uncertainties due to them.

The section of the Formation between 1.8 km N of the Camp 166 Road and Flatwater Pond, where it is directly unconformable on the Burlington Granodiorite, is equivalent to the lower part of the type section, mainly consisting of mafic lavas and conglomerates with very minor red porphyritic eutaxitic ignimbrites. Towards the top of the preserved sequence in this area trachyte flows occur that are not found to the south. Sandstone, mafic lava, and minor conglomerate on the eastern shore and northeast of Flatwater Pond belong to this Formation, but are only correlatable in the most general terms with the lower part of it. South of the detailed map area on Mic Mac Lake, the mafic lavas interbedded with conglomerate are overlain by a relatively thick sequence of sandstone (Plate 1). This seems to correlate with the part of the Formation south of Park Pond. However,

the pillow lava and chert that outcrops on islands just to the east of the mafic lavas and conglomerate maybe an inlier of Baie Verte Group, and therefore the mafic rocks to the east of the pillow lavas are of doubtful status at present. The thickness of the Snoopy Pond Formation is about 1150 metres in the main (lower) part of the type section, with about an additional 470 metres in the separate upper part of the type section.

(d) Western correlative of the Armageddon Formation

Between the Baie Verte Group and the Park Pond Fault, there is an intermittent narrow strip of rocks, belonging to the Mic Mac Lake Group, from the waterfall on Flatwater Brook, 1.8 km north of the Camp 166 Road, south to opposite the south end of White Bay Waters. The rocks in this strip are not any more deformed than typical examples in the main area of the Mic Mac Lake Group east of the Park Pond fault. At least one unit of mafic lava is found in all parts of this strip, and one unit of a distinctive pale pink porphyritic eutaxitic ignimbrite is found in all parts except the southernmost small exposure. In the northernmost part, by the waterfall on Flatwater Brook, there is also a narrow non-porphyritic rhyolite sill in contact with the Baie Verte Group, with a thin very haematised conglomerate and then a thick maroon quartz-K-feldspar porphyry sill to the east before the mafic lava and pink porphyritic ignimbrite. In the part on either side of the Camp 166 Road, a few metres of maroon siltstone and pink sandstone locally replaces the pink porphyritic ignimbrite. From 1 to 1.6 km north of Park Pond a thin mostly non-porphyritic, but in part porphyritic rhyolite sill is found locally in contact with the Baie Verte Group. Nearer to Park Pond there are two units of pink porphyritic ignimbrite separated by a mafic lava unit. It is not clear to which unit the single pink porphyritic ignimbrite to the south of Park Pond is correlative. The thickest preserved part of this sequence (about 300 metres) is found south of Park Pond, and

contains several units of mafic lava and conglomerate, a very local non-porphyritic eutaxitic ignimbrite, and a non-porphyritic maroon to purple rhyolite intruded on its eastern side by a pink non-porphyritic spherulitic rhyolite that is vesicular in places.

It will be shown later that the rocks in this strip are inverted, and overlie the Baie Verte Group unconformably. Some of the lithologies enable the thin sequence preserved in this western strip to be correlated with the lower part of the Armageddon Formation in the main part of the Mic Mac Lake Group. Although non-porphyritic sills are found locally in the Snoopy Pond Formation, all extrusive non-porphyritic silicic volcanics in the main part of the Mic Mac Lake Group are in the Armageddon Formation. The fortuitous preservation of a small piece of non-porphyritic eutaxitic ignimbrite directly below the extrusive rhyolite in the section south of Park Pond happens to correlate perfectly with the same contact in the Armageddon Formation, but this is probably not significant beyond the fact that both rocks are non-porphyritic. A much stronger correlation comes from the pink porphyritic ignimbrites, found in this western strip and near the base of the Armageddon Formation. These are identical in appearance in both areas, and absolutely distinct from any other porphyritic ignimbrites in the Mic Mac Lake Group, for the following reasons. 1) The pale pink colour contrasts with the scarlet red to maroon or purple of the main parts of all other porphyritic ignimbrites. 2) They are slightly to moderately porus-weathering, indicating incomplete closure of the pore space by sintering and vapour phase crystallisation, and later cementation by calcite. This is only seen elsewhere in part of the non-porphyritic ignimbrites in the Armageddon Formation, and very rarely in a thin development of buff ignimbrite near the base of two porphyritic ignimbrites in the Snoopy Pond Formation. 3) They contain almost

no flattened pumice clasts (fiamme), while other porphyritic ignimbrites almost always have an abundance of them. 4) Rare small pebble-size angular inclusions of red chert are found locally in both areas in these rocks, but have not been seen elsewhere. 5) Very rare fuchsite-green (chrome-bearing) highly altered serpentinite clasts are found in places in these rocks in both areas, but have not been found elsewhere. These are mostly between a millimetre and a centimetre across. 6) These are the only ignimbrites in the map area that possess a well-preserved devitrified welded glass shard texture in thin section. In all others examined this texture has been wholly destroyed. 7) The main pink ignimbrite unit in the Armageddon Formation and the unit in the western strip south of Park Pond are the only ignimbrites in the map area that contain altered plagioclase phenocrysts as well as quartz and K-feldspar phenocrysts. The pink ignimbrite in the western strip near the Camp 166 Road and near the waterfall on Flatwater Brook does not contain plagioclase phenocrysts, and it may be that only one of the two units in the western sequence contains them. It is not known whether the lower, more restricted, pink ignimbrite in the Armageddon Formation contains plagioclase phenocrysts. The rocks containing plagioclase phenocrysts have a very much better preserved glass shard texture than those that do not contain them. Since single ignimbrite units are often deposited over extensive areas, and are the most likely lithology in the western strip to form a marker horizon, it is proposed that this western strip is equivalent to the lower part of the Armageddon Formation. The conglomerates in the western strip south of Park Pond are compatable with those at the base of the Armageddon Formation, both containing mostly Burlington Granodiorite and non-porphyritic maroon rhyolite clasts, with very few of quartz-feldspar porphyry. In the western strip, mafic lava pebbles are much more common than at the base of the Armageddon Formation. Mafic lava flows are also common in the western

strip, but only one probable outcrop is found in the Armageddon Formation. It is probable that the mafic lava flows originated from dykes cutting the Burlington Granodiorite, and they are not found (E.R.W. Neale, pers. comm.) in the elliptical ring complex source of the silicic volcanics that is a short distance to the east of the base of the Armageddon Formation. Therefore, this difference.between the western strip and the lower part of the Armageddon Formation is not evidence against the correlation.

(iii) Lithologies

(a) Mafic lavas

The mafic lavas in the Mic Mac Lake Group of the map area are all basically massive lava flows. All examples in the main sequence, except for one dubious outcrop, are in the Snoopy Pond Formation, but they are common in the thin western strip correlative with the Armageddon Formation, west of the Park Pond Fault. Almost all mafic lava is green, purplish-green or purplish weathering, even in well-cleaved examples, and this contrasts with, and is the major distinction from the pale-cream weathering mafic rocks of the Baie Verte Group. Just south of Park Pond a few outcrops of strongly cleaved mafic lava are cream-weathering, like the Baie Verte Group rocks. These are the only examples seen where a mistake could be made, and they are not part of the Baie Verte Group because they are interbedded with other Mic Mac Lake Group lithologies, and because they have a vesicular base with included pink sandstone like most other Mic Mac mafic lavas. The purplish cast of many flows, especially in their vesicular bases and tops, is due to haematite, but red strongly haematised mafic lava is very rare. Most of the mafic lava is not porphyriric, but it is often possible to see the 'groundmass' altered plagioclase laths on the weathered surface of outcrops without using a hand lens. They range up to 1 cm long, but are

usually not more than 5 mm long. Rocks in which they are not visible are sometimes found to be wholly recrystallised, although they were probably fine-grained before alteration. In the few mafic lavas sectioned, clinopyroxene (augite?) interstitial to the altered (albitised or saussauritised) plagioclase laths is often partly preserved, even in strongly haematised rock. Small magnetite grains are usually abundant. The texture of the lavas is usually a random 'basaltic' aggregate of the plagioclase laths, but occasionally a weakly trachytic preferred orientation is seen. The few examples of porphyritic flows contain very large altered plagioclase phenocrysts up to 3 cm long.

The top and bottom of most flows are highly vesicular (amygdaloidal) with calcite and/or quartz fillings. In addition to their more haematised nature, the top and bottom metre or two of most flows have a large amount of included pink sandstone. That at the base was picked up by the lava during flow, and in many cases forms a 'pillow roll mud', where small ovoid lumps of vesicular mafic lava are detached from the main flow and are floating in a pink sandstone matrix (Fig. 4C.4a). The tops of flows with included sandstone are often difficult to distinguish from the base. The reason for this is not known, and it is presumed that this sandstone in the tops of flows is filling gaps between blocks. The sandstone is usually a flinty, baked rock. Units of mafic lava flows range up to 180 metres thick and each probably represents one relatively rapid sequence of eruptions. Individual flows are found down to a few metres thick, but a maximum thickness is not well defined. In one outcrop a banding between about 10 cm and a metre thick is formed by variations in haematite purplish colour, amount of vesicles, and changes from basaltic to weakly trachytic texture. This was probably caused by advances of discrete tongues of lava during the emplacement of a single flow. In general, the textures and structures of the lavas

indicate that they were relatively fluid and mobile.

Pillow lavas were reported within the Mic Mac Lake Group outcropping on islands in Mic Mac Lake (Neale and Nash, 1963). These contain interstitial green white-weathering chert, occasionally with a pinkish tinge, a sediment type wholly foreign to the Mic Mac Lake Group. It is quite possible that these pillow lavas are an upthrust inlier of Bale Verte Group unconformably overlain by the mafic lava, conglomerate, and sandstone to the west. Most of the mafic rocks to the northeast of the pillow lavas are massive green lavas without haematitic alteration, so some of these could also be Baie Verte Group rocks. As all these rocks are isolated on small islands within Mic Mac Lake, it may not be possible to decide by field methods whether Baie Verte Group rocks are represented here.

No mafic sills have been definitely identified within the Mic Mac Lake Group. Almost all examples of mafic rock are definitely flows, but some of the uppermost mafic rocks in the area from just east of to 1.4 km NNE of Park Pond may be sills.

Mafic dykes 10 cm to several metres wide are found occasionally cutting the Burlington Granodiorite just below the unconformity with the Mic Mac Lake Group. Most were found between 2.8 km north of the Camp 166 Road and just north of the Burlington Road, and in an area east of the southern part of Mac Mac Lake. They are often purplish and haematised to a few metres below the unconformity, and some are moderately vesicular (amygdaloidal). Although these dykes are almost certainly feeders to the mafic lavas in the Mic Mac Lake Group, they have not been seen cutting any of the rocks in the Group. In the very detailed map area, all but two were seen under the mafic lava and conglomerate sequence of the Snoopy Pond Formation north of the Camp 166 Road, and it is possible that most of these dykes fed mafic lavas in the Armageddon Formation and were eroded before the sequence now seen

overlying them was deposited. The southern and smaller of the two areas mapped as possible

Baie Verte Group within the Burlington Granodiorite east of Mic Mac Lake (Neale and Nash,

1963) seems to be an area with a number of Mic Mac mafic dykes. One of these has a coarse-

grained dolerite core containing saussauritised euhedral plagioclase laths poikilitically included

with a few ovoid probable olivine pseudomorphs within clinopyroxene (augite), and accessory

leucoxene after ilmenite. Dykes in this area mostly strike about east-west; in the area south of

Flatwater Pond most strike southwest-northeast; in both areas they are subvertical. The author

also found similar mafic dykes cutting Burlington Granodiorite on the eastern part of the

Burlington Road northnortheast of the silicic ring complex. No mafic dykes are present within

the elliptical silicic ring complex that was the source for the silicic volcanics (E.R.W. Neale,

pers. comm.). The three widely separated areas where mafic dykes have been found suggest that

the dykes may occur in a ring-like zone surrounding the ring complex. It is therefore possible

that many of the feeder dykes for the mafic lavas in the Mic Mac Lake Group are further to the

west than the Section now exposed, and therefore still buried. This situation is indicated by the

difference in mafic lava content between the strip west of the Park Pond Fault and the lower

Armageddon Formation, if the correlation is correct.

(b) Trachyte flows

Trachyte flows are found only in the upper part of the Snoopy Pond Formation in the area

extending 3.6 km south from Flatwater Pond. In the southern part of this area the trachytes are

within highly deformed rocks, and although they themselves are not penetratively deformed, they

are relatively altered. North of the Camp 163 Road, a much fresher trachyte flow is found

between two little deformed mafic lava flows. The trachytes

consist of a flinty, dark purplish, slightly mottled rock, discoloured to pale buff near thin quartz veins. Near the base of the northern flow, very small (~1 mm) vesicles occur (now calcite-filled amygdales). In thin section, the strongly aligned mat of feldspar laths contains abundant fine-grained magnetite, scattered phenocrysts of K-feldspar, a few larger magnetite grains, a few small apatite grains, and rare phenocrysts of sphene. Areas ('pools') of granular quartz are common. The map (Plate 3) shows that this northern single flow rests on an irregular, probably erosional, surface. The flow ranges from 11 to 39 metres thick. One flow in the southern area is a maximum of 48 metres thick. The source of the trachytes is probably represented now by the syenite intrusions mapped by Neale and Nash (1963) in the ring complex to the east.

(c) Silicic volcanic rocks

1. Introduction

Most of the silicic volcanic rocks in the Mic Mac Lake Group are ignimbrites. The author is indebted to B. E. Lock for a lucid introduction to and discussions on the properties of ignimbrites. The statements that follow on some features and terminology of ignimbrites in general stem primarily from an extended review that he wrote and included in his thesis (Lock, 1969).

Ignimbrites are the deposits of the glowing hot basal avalanche of nudes ardentes eruptions. The glowing avalanche consists of a turbulent suspension of hot glass shards, hot pumice fragments, phenocrysts if present in the parent magma, and sometimes a minor amount of clastic debris picked up during its travel. After the avalanche has come to rest, the resulting deposit of still very hot glass shards and pumice fragments may weld (sinter) by collapse and compaction if the glass is still sufficiently hot and plastic. This results in flattened glass shards and pumice fragments, and the latter then are coplanar discoid objects with lensoid cross-sections (termed fiamme),

and they are usually the commonest indication that an ancient rhyolite is an ignimbrite.

Ignimbrite units are often found to consist of the deposits of several hot avalanches, and/or sub-units of the same avalanche that travelled down separate valleys. The deposit of each avalanche is termed a flow-unit. As long as each flow-unit is deposited soon (not more than a few days) after the previous one, the deposit consisting of several separate flow units will cool as, or almost as, a single unit. It will be separated from earlier and later ignimbrite deposits that cooled independently by unwelded material at the top and sometimes at the base, or by interbeds of other lithologies. These units that may consist of one or several flow-units, but that cooled as one deposit are termed cooling units. The degree of welding in a vertical section through a cooling unit may be complex, but is often most complete in the central part, decreasing slowly towards the top, and rapidly near the base, where chilled, unwelded, material is found. However, the base may be thoroughly welded, and the unconsolidated top is easily removed by penecontemporaneous erosion. During and after welding, the evolution of hot gases in the cooling unit generates fumaroles, but more significantly, leads to vapour phase crystallisation in the remaining pore space, and also promotes devitrification. This process has an equal, if not more important influence than the welding on the final mechanical properties of the rock, and little to unwelded material, especially at the base of the cooling unit, may be consolidated to a very tough strong rock by this process (Lock, 1971). It is very difficult to separate the effects of welding and vapour phase crystallisation in old devitrified rocks where the glass-shard texture has been destroyed. The evolution of hot gases after deposition also often forms 'gas-explosion nuclei' (Beavon, Fitch, and Rast, fide Dewey, 1963).

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These are essentially amygdales, but unlike those in basalt, they form suddenly in very viscous rock, and often contain angular splinters of the host rock, and deform the microtexture in the host rock, to prove it. In the Mic Mac rocks they are filled by fairly coarse crystals of feldspar and form white speckles and patches in the red rocks. They may be spherical or of irregular shape, and are usually from about a millimetre to a centimetre, but occasionally to ten centimetres across, and are most often found within the fiamme (flattened pumice clasts) of the most flinty ignimbrites, and in flow banded rhyolite autobreccia. Ross and Smith (1961), in an excellent summary, state that fiamme are often axiolitically devitrified, and further that axiolitic devitrification is diagnostic of ignimbrites. Axiolitic devitrification texture involves the inward growth of a beard of very fine-grained acicular silica and/or feldspar lining the margins of the discoid fiamme, with a relatively coarse-grained anhedral mosaic of quartz and/or feldspar in the central portion. This texture is not seen in the fiamme of ignimbrites in the Mic Mac Lake Group, but it is seen in objects that could be very deformed vesicles in rhyolite sills and a rhyolite lava flow, and is therefore not diagnostic of ignimbrites.

Ignimbrite units that were exceptionally hot and/or composed of relatively less viscous glass than usual may undergo partial to complete secondary flowage after extreme welding. This secondary movement leads eventually to a rock almost indistinguishable from a rhyolite lava flow, and is termed a rheoignimbrite (Rittman, 1962). Rocks partially remobilised with slight to moderate modification of the normal welded shards and flattened pumice are termed parataxitic ignimbrites (Beavon, Fitch, and Rast, 1961). Normal ignimbrites without secondary mobilisation are termed eutaxitic ignimbrites, or eutaxites, after the welded glass shard texture, termed eutaxitic texture.

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All silicic volcanic rocks in the Mic Mac Lake Group are wholly devitrified. Most contain much haematite, and their colours are very intense hues of scarlet, crimson, maroon and purple. The rhyolite flows and sills, and the matrix of the ignimbrites consist of an almost irresolvable homogenous quartzo-feldspathic and haematite groundmass. Fiamme are very slightly more coarsely crystalline than the groundmass, but always consist of a mosaic of equant grains. They always contain a larger proportion of haematite than the matrix. The ignimbrite cooling units have sporadically developed originally little to unwelded and little vapour phase crystallised tops and bottoms. These are now cleaved, usually purple, rocks with pale buff diffuse reduced patches in them. Their sporadic occurrence at the base of cooling units is due to variable welding and vapour phase crystallisation. The original unconsolidated tops would have been very susceptible to erosion, and this is probably the reason why they are not consistently present. Where the sequence consists wholly of ignimbrite, it is usually not possible to tell the difference between the top of one flow and the base of the next. Although the change from massive flinty rock to cleaved rock is usually fairly abrupt, a facies intermediate between the flinty, well-sintered and vapour phase crystallised central (and lower) parts, and the cleaved originally unconsolidated margins of the cooling units is well developed in some places. It either forms a near-basal, and less commonly near-top, facies, or the whole of a cooling unit. It is distinguished by its porus-weathering, as opposed to flinty-weathering nature, and is not cleaved, although it is seen in places to grade rapidly into well-cleaved marginal material. The porus-weathering nature indicates incomplete closure of the pore space by welding and vapour phase recrystallisation, and later cementation by calcite, that weathers out. It is very

possible that some flinty-weathering rocks were of this type, but were cemented by silica instead. If so, it would be very difficult to detect.

Besides the distinction between ignimbrites, and rhyolite flows and sills, there is another easily mappable distinction, between non-porphyritic and porphyritic rocks. All rhyolite lava flows, all of which are in the Armageddon Formation (lower sequence) and all but one sill are non-porphyritic. Most of the eutaxitic ignimbrites in the Armageddon Formation are non-porphyritic; all in the Snoopy Pond Formation (upper sequence) are porphyritic. Non-porphyritic rocks are not wholly devoid of phenocrysts, but usually contain a very few K-feldspars, similar in size to those in porphyritic rocks. The porphyritic rocks contain a fairly equal proportion of mostly equant quartz and K-feldspar phenocrysts, of a uniform size (typically 0.5 to 2 mm), and uniform proportion of the rock (20-30%) throughout the whole sequence. Most phenocrysts are broken and/or resorbed, with K-feldspar more commonly retaining some euhedral faces than quartz. The K-feldspar now appears to be a cryptoperthite or a microperthite, and shows Carlsbad and Baveno twins. The perthite is sometimes a diffuse patch type, but more commonly is a type that mimics chessboard albite. It is possible that some of the feldspar is a chessboard albite, but this has not been positively identified. No mafic mineral grains are preserved. A few ovoid semi-opaque haematitic objects may be highly altered pseudomorphs after a mafic mineral.

2. Ignimbrites

2.1 Non-porphyritic

Non-porphyritic ignimbrites are restricted to the Armageddon Formation. The eutaxitic ignimbrites are all found below the main thick rhyolite unit, with the one thin exception within that unit. They show a thin (~1m or less) cleaved purple non-welded and non-vapour phase crystallised base and

Fig. 4C.4. Megascopic textures of mafic lava and eutaxitic ignimbrites.
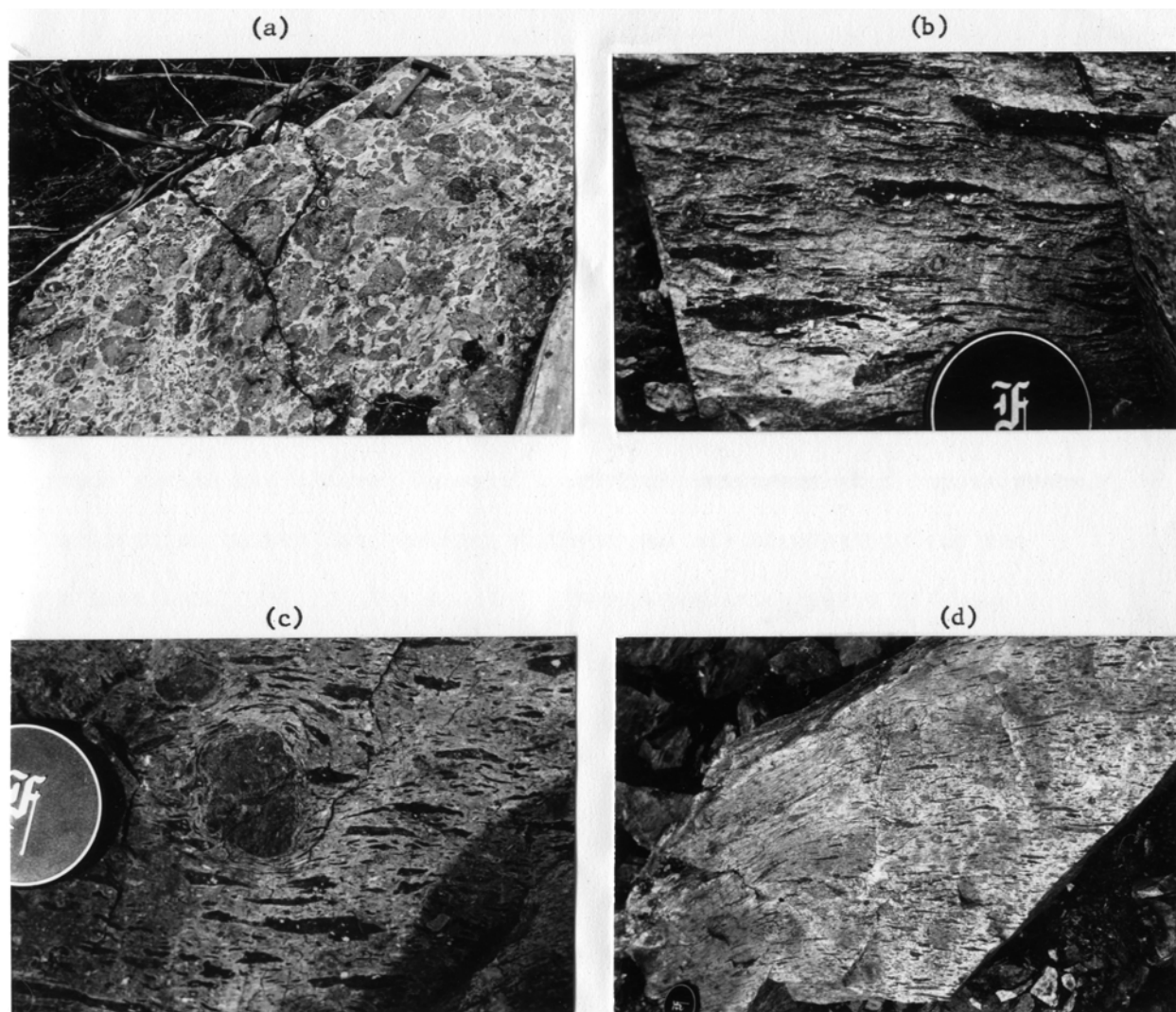
(a) Mafic lava basal pillow roll 'mud' – lava in pink sandstone.

(b) Eutaxitic ignimbrite flattened pumice texture – fiamme.

(c) Fiamme differentially flattened around rhyolite pebble.

(d) Larger-scale differential flattening of fiamme.

top in a few places. A large proportion of these eutaxites is porus-weathering, compared to the porphyritic ignimbrites. The porus-weathering rocks are pale buff, or bright pink, or pinkish purple, with dark brown or red or black fiamme. The gradation from the cleaved marginal lithology into the porus-weathering uncleaved material takes place in a zone less than a metre wide (outcrop < 1 >*, 1.1 km south of the Camp 166 Road). Within this zone slightly diffuse lumps of the non-cleaved porus-weathering eutaxite are found within the cleaved material, and this indicates that vapour phase crystallisation rather than welding differences are responsible for the change (see Lock, 1971). The scarlet flinty weathering parts of these ignimbrites appear to form the central part of a cooling unit 1.1 km south of the Camp 166 Road. However, this material also replaces the porus weathering material where the cooling unit thickens to the south. As there is little or no difference in the amount of flattening of the abundant fiamme between the porus and flinty weathering types, it is suggested that vapour phase crystallisation is also mostly responsible for the difference between the two lithologies.

The porus-weathering non-porphyritic eutaxitic ignimbrites show the megascopic features of eutaxitic ignimbrites better than any other ignimbrites in the map area, and (incidentally) more clearly than any other ignimbrites the author has seen described. Typical eutaxitic fiamme texture in these ignimbrites is shown in Fig. 4C.4b. The fiamme are seen differentially flattened around pebbles of rhyolite or granodiorite that are commonly found within the ignimbrites, picked up by the glowing avalanche during its travel (Fig. 4C.4c). One example of larger scale differential

* Localities marked on Plates 3 and 4 in the same way.

Fig. 4C.5. Eutaxitic ignimbrite textures.
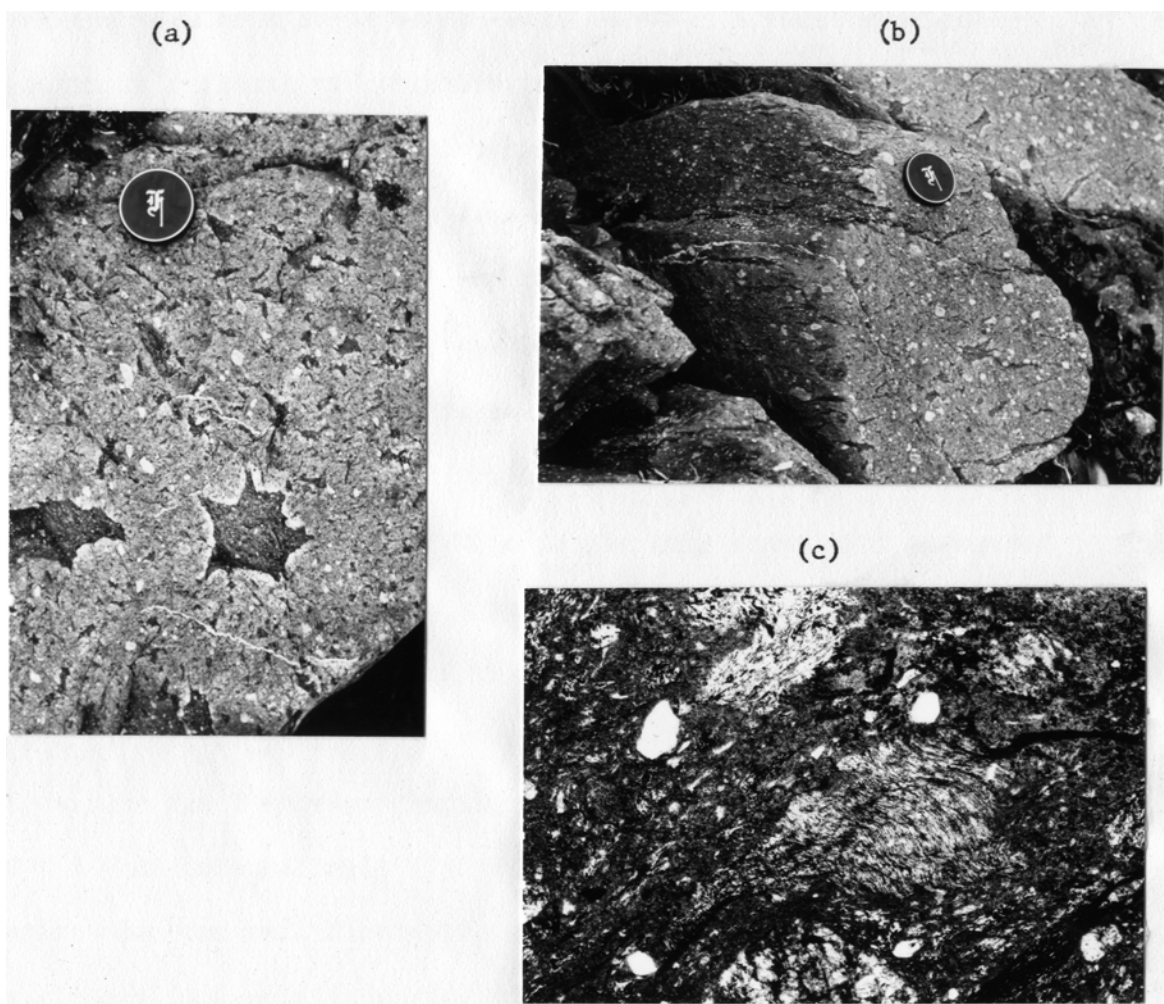
(a) Large unflattened but elongate fiamme.

(b) View of outcrop showing sections parallel and perpendicular to elongated fiamme.

(c) Flattened vesicle traces in fiamme (X 25).

flattening was seen in a loose block (Fig. 4C.4d). A small erosional channel cut into a flinty red eutaxite, exposed in an outcrop 0.9 kin south of the Camp 166 Road < 2 >, is filled by a bright pink porus-weathering eutaxite with black fiamme. These fiammes are little flattened (Fig. 4C.5a), and are elongated down the axis of the channel (Fig. 4C.5b). The steep plunge of the lineation compared with the shallow local dips suggests that the channel floor had a local steep slope in the section exposed, and the ignimbrite oozed over this (like a waterfall) after deposition, elongating the pumice rather than flattening it. This is the only example of elongated fiamme found in the area. In thin sections of the porus-weathering eutaxites, the fiamme preserve traces of the internal flattened vesicles (Fig. 4C.5c), but the matrix shards have been almost totally destroyed by devitrification. In the flinty-weathering eutaxites 1.5 km south of the Camp 166 Road < 3 > an internal unit (?major flow unit) that is purple rather than scarlet also has well displayed internal flow-units a few metres thick. The base of each flow-unit is marked by a layer about 5-10 cm thick containing granodiorite pebbles (Fig. 4C.6a). This is the only unit in the map area where this is clearly displayed, although included pebbles are common in many other ignimbrites, and are probably localised, but not so sharply, at the base of flow units.

Some of the thick non-porphyritic flow-banded rhyolite that forms a large proportion of the Armageddon Formation is possibly rheoignimbrite. A description of this is found in the section below describing the rhyolite.

2.2. Ignimbrites, Porphyritic.

All porphyritic silicic volcanic rocks in the map area are thought to be extrusive and also to be ignimbrites, with two minor exceptions, described later. All except 3 porphyritic ignimbrite cooling units are in the Snoopy Pond Formation (upper sequence). The two units near the base of

Fig. 4C.6. Eutaxitic ignimbrite textures
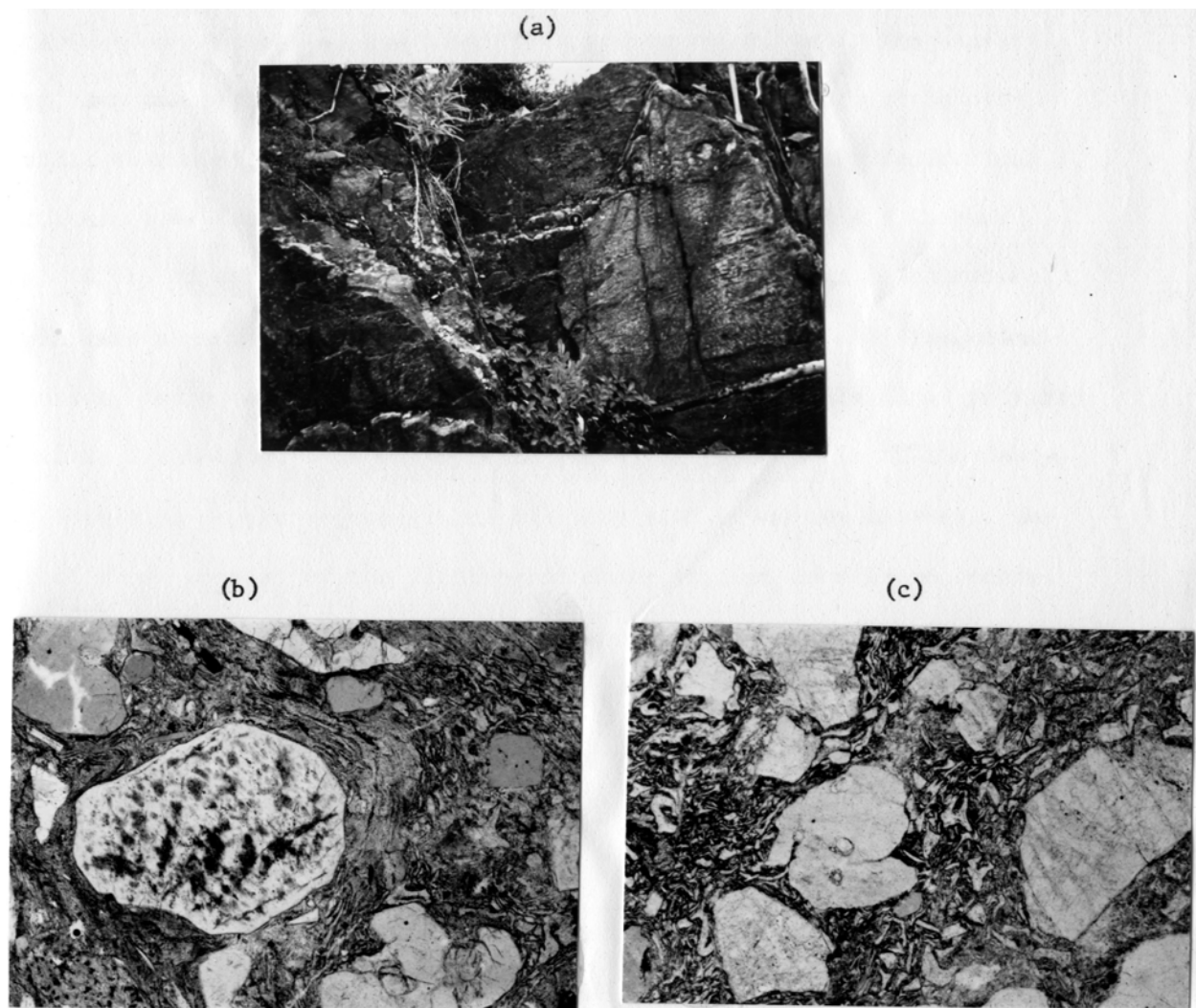
(a) Flow unit boundary – layer of white granodiorite pebbles.

(b and c) Pink porphyritic eutaxitic ignimbrite – eutaxitic welded shard texture (X 25).

the Armageddon Formation, and their presumed correlatives in the western strip, are the pink

porus-weathering quartz-K-feldspar-plagioclase porphyritic eutaxites, described previously. They

are the only ignimbrites in the map area that have preserved the eutaxitic welded shard texture

(Fig. 4C.6b, c) and also (in part) the only ones that contain (albitised?) plagioclase phenocrysts.

The third unit is at the top of the Armageddon Formation, and is a normal red to reddish-orange

quartz-K-feldspar porphyry eutaxitic ignimbrite. The porphyritic eutaxitic ignimbrites in the

Snoopy Pond Formation (upper sequence) are all very similar to one another. The bulk of these

consist of the flinty-weathering scarlet to crimson central parts of the cooling units. Near the base

of two cooling units, a fairly discrete layer up to 20 metres thick is found, consisting of a non-

cleaved porus-weathering pale buff eutaxite with dark brown fiamme, and they are probably flow

units. The cleaved purple, originally little consolidated material, forming the tops and bases of

the cooling units is far more common. It often contains diffuse-margined off-white patches where

the haematite has been reduced during deformation. The cleaved marginal facies contains sericite

forming the cleavage. Occasional isolated pseudomorphs of glass shards are seen in it, but

otherwise shard textures in the porphyritic ignimbrites have been destroyed. A vague intimation

of welded shard texture is defined by dusty haematite (magnetite?) in one sample of the buff

porus-weathering rock. In another sample of this rock, perlitic cracks are outlined by the opaque

dust in the fiamme but not in the matrix.

Included pebbles (and occasionally cobbles) of rhyolite and granodiorite are common in

the porphyritic ignimbrites, and the former are easily distinguished from the fiamme. In the

marginal cleaved facies and the rare buff porus-weathering near-basal facies, the brown fiamme

are clearly flattened pumice clasts, and in many cases the traces of the flattened

Fig. 4C.7. Parataxitic textures; resorbtion cavities.
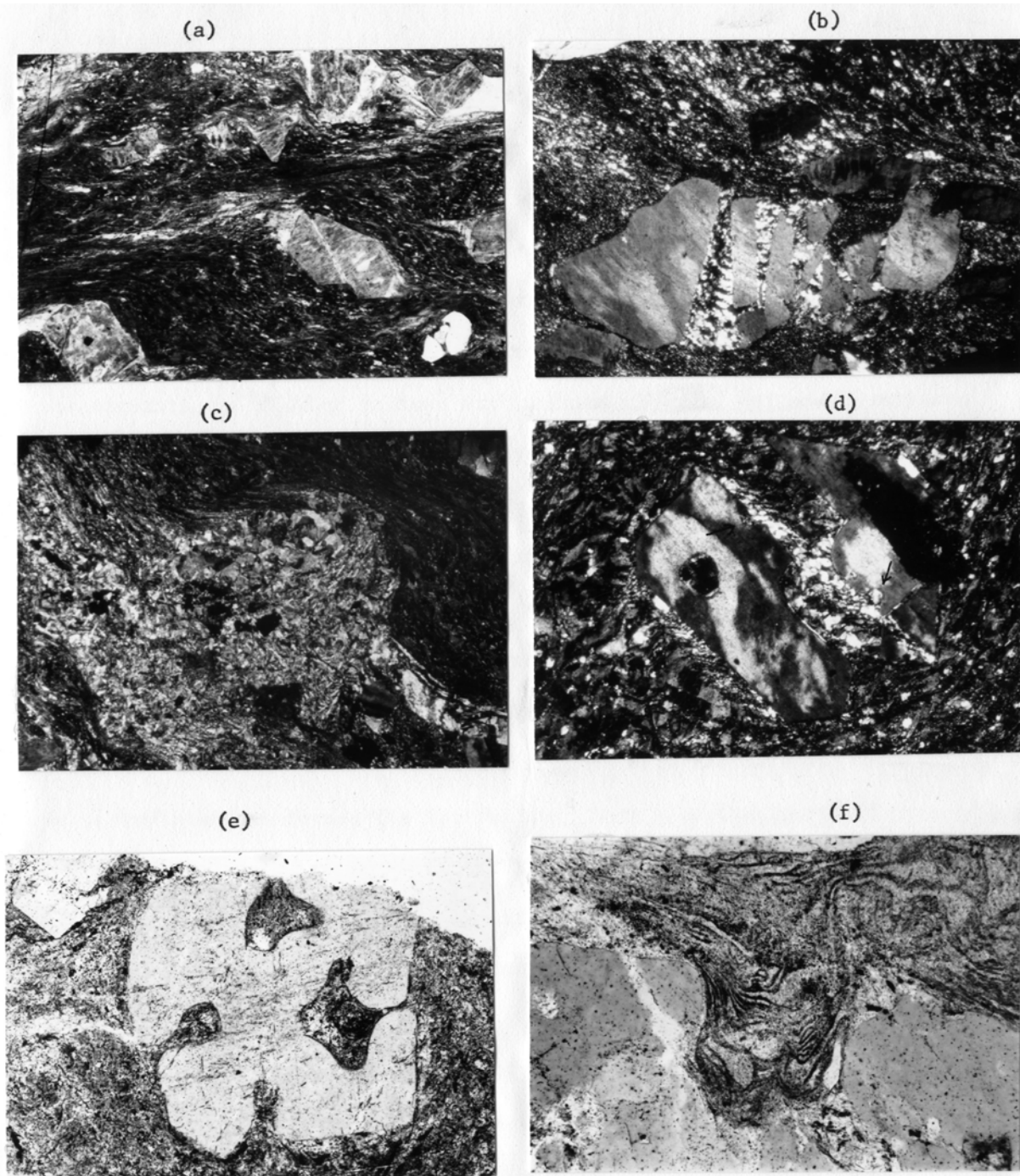
(a) viscous flow texture in a 'fiamme' (X 10).   (b) pull-apart of K-feldspar phenocryst (X 10)

(c) feldspar-filled gas-explosion nucleus deforming flowage texture (X 10).

(d) pulled-apart K-feldspar; small resorbtion cavities on broken surface (X 10).

(e) Quartz phenocryst with resorbtion cavities (X 40).

(f) welded glass shards packed into outer part of resorbtion cavity (X 100).

vesicles can be seen within them. However, in the main flinty-weathering part of the ignimbrites, the fiamme may not all be flattened pumice clasts. In many outcrops, the fiamme are not all, or even mostly, coplanar discoid objects. They are generally compact, and range from a somewhat flattened to an equidimensional shape, but the latter are not included pebbles. In thin section (Fig. 4C.7a), and sometimes in outcrop, clear evidence of viscous shearing and flowage is seen within these objects, defined by streaks alternately rich and poor in haematite dust, and K-feldspar phenocrysts are pulled apart by it (Fig. 4C.7b). The pull-apart gaps are filled by vapour phase quartz and feldspar. In addition, small gas explosion nuclei filled by white vapour-phase feldspar are very common within these rocks, and mostly occur within the fiamme, and deform the flowage structure (Fig. 4C.7c). It is suggested that many of these equidimensional viscously-deformed objects are unvesiculated plastic rhyolite, erupted with the material that became vesiculated and formed the ignimbrite. They show that parts of some of the cooling units underwent a little flowage after welding, and are therefore partly tending towards parataxitic ignimbrites, as opposed to the clearly eutaxitic, merely welded material nearer the margins of the cooling units.

A thin section of an ignimbrite with plastically deformed 'fiamme' showed one K-feldspar phenocryst that could be approximately fitted back together. Small resorbtion cavities are seen on the pulled-apart faces (Fig. 4C.7d), showing that at least some resorbtion of phenocrysts takes place after deposition of the ignimbrites. This is also indicated by the delicate and breakable form of some resorbed phenocrysts (Fig. 4C.7c). Resorbtion after eruption is probably indicated by the occurrence of glass shards packed into the outer part of resorbtion cavities in phenocrysts in the pink eutaxite in the Armageddon Formation (Fig. 4C.7f).

4C.27

Although the main red flinty-weathering parts of the ignimbrites are not cleaved and apparently not deformed, in some places quite strongly cleaved and flattened included pebbles of granodiorite are found within them, showing that in places the rock has been significantly internally deformed. The phenocrysts are always strained, but the flinty rocks do not show any other evidence of deformation. In most places, however, included granodiorite pebbles are undeformed. Fiamme orientation may not therefore be a reliable measure of bedding orientation, besides its inherent inaccuracy when compared with sediments. In the cleaved marginal facies fiamme are sometimes seen, and are in places oblique to the cleavage. The presence of discoid collapsed pumice clasts in this cleaved marginal material shows that some of it compacted and presumably welded somewhat upon deposition, and is further evidence that welding is not the dominant process affecting the mechanical properties of the rock. An outcrop of red flinty ignimbrite on the south side of the Camp 166 Road < 4 > a few metres below cleaved marginal facies, shows a polygonal (tetragonal) joint set oriented with the intersection perpendicular to bedding. Some joints are filled with narrow veins of sandstone, and others with concentrations of magnetite. These are polygonal cooling joints filled by fumarolic magnetite and small sedimentary dykes. No other examples of these have been found. The lowermost porus eutaxitic ignimbrite just south of the Camp 166 Road is mostly within conglomerate, but also occurs in an originally subvertical cleft between granodiorite and a large granodiorite block that has moved slightly, 0.48 km south of the Camp 166 Road < 5 >. A conglomerate mostly containing granodiorite pebbles is exposed directly overlying the ignimbrite in the cleft. It is cemented by botryoidal (presumably originally colloform) silica, and the feldspar in the pebbles is kaolinised. A few similar silica veins cut the granodiorite forming the side of the cleft. This is the only evidence

of fumarolic activity known in the area.

Due to patchy development and penecontemporaneous erosion of the cleaved bases and tops of the ignimbrites in the Snoopy Pond Formation, and because they are less likely to outcrop than the flinty central parts, individual cooling units cannot usually be precisely defined or traced along strike for any distance. This is also complicated by the rhyolite sills present north of the Camp 166 Road. Such cooling units in both Formations that can be defined range from 15 to 120 metres (and perhaps as much as 200 metres) present thickness. Internal flow units are seen about 3 metres thick, and probably range up to a thickness equal or comparable to the cooling units.

3. Rhyolite flows

The major part of the Armageddon Formation consists of a thick non-porphyritic rhyolite unit. It is all crimson to maroon in colour, and while much is flow banded with a millimetre-scale colour banding, some is homogenous. Flow folding of the flow banding is common, but is irregularly oriented. The whole unit would appear to be one flow, were it not for two thin (1-2m) interbeds, and others may be unexposed. The thickness of the whole unit is about 440 metres. At the base, 0.25 km and from 1.1 to 1.8 km south of the Camp 166 Road, a rhyolite autobreccia up to 15 metres thick is developed. Flow banded and homogenous rhyolite clasts from small pebbles to boulders a metre across are contained in a scarlet rhyolite matrix. The clast/matrix ratio is large. The clasts may be angular brittle fragments or partly to wholly plastically deformed. Where most clasts are deformed, the rock strongly resembles the flinty eutaxitic and slightly parataxitic ignimbrites. The autobreccia locally overlies the unconsolidated (now cleaved) top of a porus eutaxitic ignimbrite 0.25 km south of the Camp 166 Road < 6 >. Ovoid lumps of plastic rhyolite have

been loaded into the unconsolidated material, and some of the latter has been picked up and incorporated into the autobreccia. Also in this locality, some rhyolite blocks contain ellipsoidal gas explosion nuclei up to 10 cm across (Fig. 4C.8a). These deform the flow banding, contain angular splinters of the rhyolite, and are filled by a white feldspar aggregate. The contact of the autobreccia with the overlying rhyolite is rather cryptic, and takes place over less than a metre. In places brittle tongues of flow banded rhyolite are frozen in the process of being pushed down into the autobreccia, and deforming plastic clasts in the autobreccia around them. Autobreccia is not found within the rhyolite unit.

In the outcrop immediately overlying the interbed of porus eutaxitic ignimbrite, 1.8 km south of the Camp 166 Road, the flinty maroon rhyolite shows a lensoid banding. This is flow folded in places, and some layers show a strong flow lineation. The lenses are moderately to closely packed, are typically less than a centimetre thick and several tens of centimetres long, and usually have white feldspar speckles like that in gas explosion nuclei concentrated along their centres. These could be very strongly deformed pumice clasts. In addition layers parallel to bedding up to 1 metre thick, and also irregular patches, show a ramifying network of rather thin and diffuse white vapour phase feldspar veins and veinlets. These appear to be zones of strong gas evolution and incipient brecciation formed immediately after deposition. These characters suggest that at least this part of the rhyolite unit may be a rheoignimbrite. Near the base of the overlying porphyritic eutaxite and above this outcrop, an outcrop of flinty non-porphyritic eutaxitic ignimbrite may be the upper less remobilised part of this unit, but neither this nor the possible rheoignimbrite are traced along strike. It is possible that they occupy a fairly large undetected valley cut in the underlying flow banded rhyolite. The large content of

Fig. 4C.8. Rhyolite flow and sill textures; Lahar
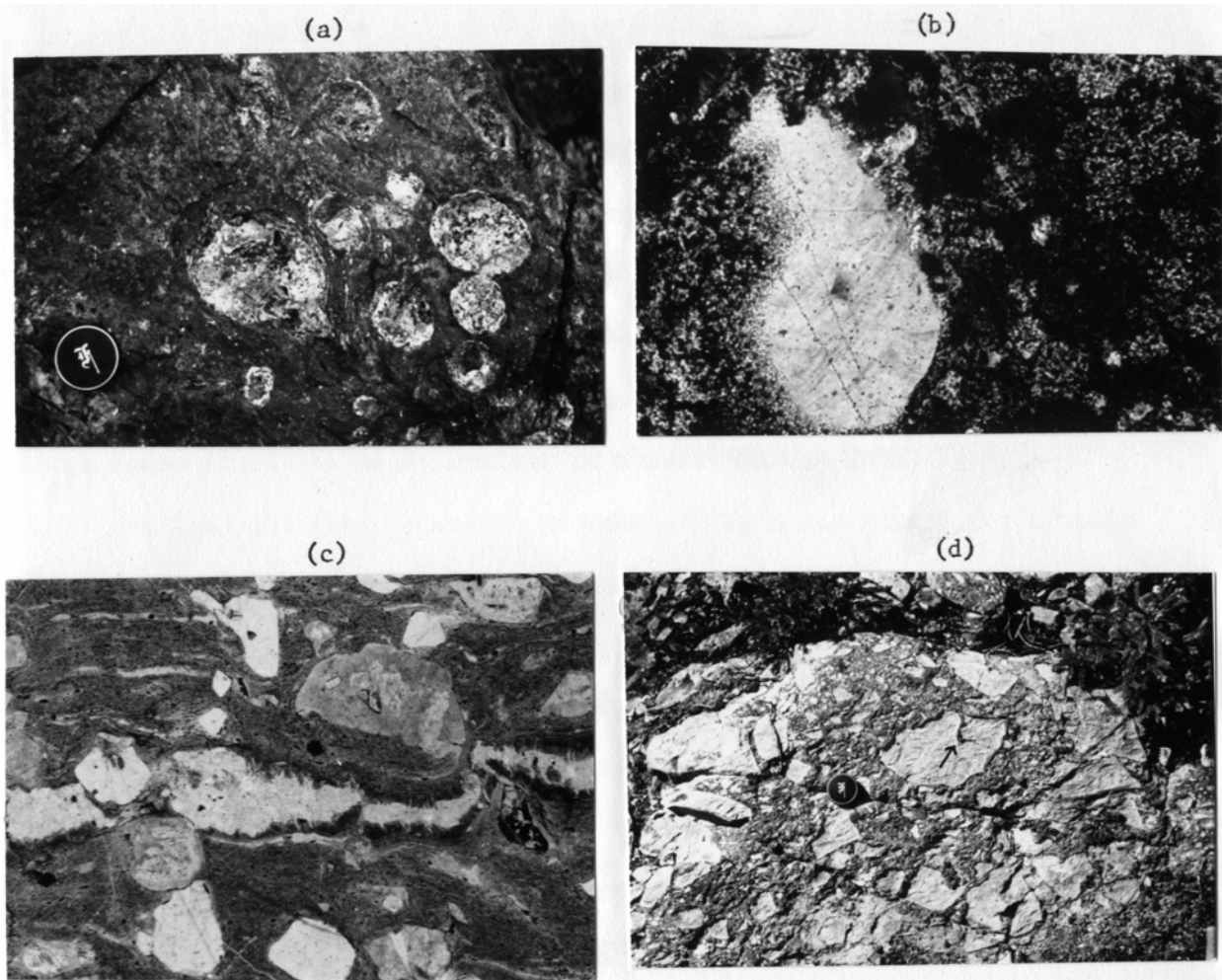
(a) Large gas-explosion nuclei in rhyolite autobreccia.

(b) Micropoikilitic devitrification ('snowflake') texture – rhyolite sill (X 100).

(c) Axiolitically devitrified objects in porphyritic rhyolite (?)xenolith (X 10).

(d) Lahar; mafic lava clasts near valley wall. Note incipient pull-apart (arrowed).

white vapour phase feldspar in this outcrop is not found in the remainder of the rhyolite unit, except locally in the basal autobreccia.

A peculiar dark pink to dark maroon non-porphyritic rhyolite extends from the Camp 166 Road northward along the unconformity on Burlington Granodiorite. In places it possesses a cleaved purple base up to a metre thick, that is seen to penetrate a few centimetres into cracks in the underlying granodiorite. Its upper contact is exposed in one locality < 7 >, north of the Camp 166 Road, where it is overlain by a red flinty porphyritic ignimbrite containing included pebbles within about a metre of the irregular contact. This exposed contact is probably erosional, and may be the major erosion surface between the Armageddon and Snoopy Pond Formations. The relationships of this rhyolite on and south of the Camp 166 Road are partly unclear, but it appears from the map that this rhyolite is extrusive, and forms a flow to the north and a tholoid on the Camp 166 Road. It overlies a discontinuous thin conglomerate covering the Burlington Granodiorite in the area of complex detailed palaeotopography on the south side of the Camp 166 Road below the tholoid. A small amount of rhyolite breccia in the easternmost exposure of this unit may be diatremic (Neale and Kennedy, 1967), and it is interpreted (Plates 1, 4) that this rhyolite was extruded from an (unexposed) dyke that reached the surface here. Internally, this rhyolite is flinty and homogenous, apart from sparse axiolitically devitrified discoid dark red porus-weathering objects, typically 1-3 mm thick and 5-20 cm long. In one instance they are very elongate, but otherwise these objects are parallel to local bedding orientation. It is possible that they are highly deformed and altered pumice clasts, and that the rock is a parataxitic ignimbrite, but the geometry of the unit does not favour this interpretation, and the rock and the axiolitic streaks are identical to those in definite rhyolite sills in the Snoopy Pond Formation. A possible interpretation of

the axiolitic streaks is that they are deformed vesicles formed during flow of the rhyolite and filled after deformation by vapour phase activity. A few small rhyolite pebbles are found in the rock, and they indent the axiolitic streaks when the two are seen in contact. The significance of the cleaved base is not known, and the status of this rock unit is somewhat uncertain.

4. Intrusive rhyolites

4.1 Non-porphyritic

Non-porphyritic rhyolite sills are abundant in the porphyry ignimbrites up to 1.2 km north of the Camp 166 Road. They are a maroon to dark purplish colour, and are either homogenous, or more commonly contain abundant very attenuated axiolitically devitrified discoid streaky objects, typically 1 mm thick and 10 cm or more long. These parallel lensoid objects are rarely flow-folded and very rarely lineated. They are very similar to the objects found in the rhyolite flow described previously. These rhyolites have variable contact relationships. The contact may be sharp, with flinty non-porphyritic rhyolite adjoining either flinty or cleaved porphyry ignimbrite. In other places, the flinty rhyolite with axiolitic streaky banding contains abundant quartz and K-feldspar phenocrysts in the nearest 1 to 3 metres adjoining the top or bottom contact, usually with the cleaved marginal porphyry, or where it is interpreted to have been. The non-porphyritic rhyolite is interpreted to have picked up the phenocrysts (and shards) from unconsolidated marginal parts of the porphyritic ignimbrites. In a few places, the contact shows a thin (~30 cm) development of a cleaved non-porphyritic maroon (with diffuse-margined off-white patches) marginal facies to the rhyolite, that does not seem to have a systematic distribution. Its cause is not known. The contact relationships therefore do not help to distinguish these rocks as sills, except where the top contact

has entrained phenocrysts from the overlying ignimbrites,which is the case, in places, along the upper three main units. The map distribution shows that the uppermost unit is a sill and that the one below it probably is a sill, The outcrop to the east of the large fault is interpreted as part of a sill from its similarity to the other rocks, and the unexposed dyke feeding it is interpreted from a double lineament and the offset of a buff porus-weathering porphyritic ignimbrite.

On the unconformity on Burlington Granodiorite, 1 .3 km south of the Camp 166 Road a small tholoid of homogenous dark maroon non-porphyritic rhyolite is fed by a dyke in the Granodiorite. The pink porphyry eutaxite is severely brecciated in the zone overlying the tholoid. In the first few metres above the contact the breccia blocks are separated by rhyolite veins up to a few centimetres wide. Above this the veins are filled with sandstone, presumably washed into empty spaces from the surface.

North of this, 1.0 km from the Camp 166 Road, a diatremic dyke about 15 cm wide cuts the Burlington Granodiorite just below the unconformity. Angular pebble and smaller size comminuted fragments of Granodiorite are contained in a homogenous crimson matrix, which probably consisted of glass shards. This is probably a gas-breccia dyke as the clast/matrix ratio is large and there are some very thin veins of the crimson matrix.

Elsewhere, a tongue of a sandy-weathering maroon spherulitic rhyolite sill is found just east of Park Pond. A very similar pink sandy-weathering spherulitic rhyolite intrudes maroon rhyolite interpreted to be extrusive in the western strip south of Park Pond. The pink spherulitic rhyolite contains in places empty vesicles about a centimetre across lined with small quartz prisms. Also in the western strip 1.0 to 1.6 km north of Park Pond a maroon flinty rhyolite sill is intruded between the Baie Verte Group and a mafic lava of the Mic Mac Lake Group. This sill is patchily

porphyritic in the outcrop 1.2 km north of Park Pond that exposes the contact with the Baie Verte Group. It is interpreted to have picked up unconsolidated phenocrysts and glass from the margin of the pink porphyritic ignimbrite that adjoins the contact to the south, but which is not present in this outcrop. A thin steely dark maroon flinty non-porphyritic rhyolite sill 2 to 3 metres thick is also found on the exposed contact with the Baie Verte Group in the western strip 1.4 km north of the bridge on the Camp 166 Road, on the south side of Flatwater Brook. It has a relatively coarse-grained micropoikilitically devitrified 'snowflake texture' in thin section (Fig. 4C.8b), as also does the crimson quartz-K-feldspar porphyry sill to the east of it in this locality. Anderson (1970) proposed that this texture was diagnostic of ancient ignimbrites. This is clearly not the case, and only these two sills and the one north of Park Pond possess this texture. Apart from the spherulitic sills, all other silicic rocks in the map area have an ultrafine-grained almost irresolvable mosaic groundmass. The maroon homogenous rhyolite in the Mic Mac sequence at the northern end of the map area northeast of Flatwater Pond may be either a flow, or more probably a sill.

<div align="center">4.2. Intrusive rhyolites, Porphyritic</div>

None of the quartz-K-feldspar porphyry in the Snoopy Pond Formation is thought to be intrusive. Although not all outcrops show fiamme, sufficient of them do to rule out any significant intrusive bodies. The crimson quartz feldspar porphyry in the western sequence on Flatwater Brook is thought to be intrusive because of its coarse-grained, micro-poikilitically devitrified matrix, and because an internal intrusive contact is seen in it in the outcrop on the north side of the bend in the Brook. This consists of a band, with a sharp western contact, about 3 cm thick that is phenocryst-free. It is not a chilled margin, but is due to a boundary layer process that

excludes particles in a flowing liquid from the immediate vicinity of the container wall (as in

blood flowing in a vein). An outcrop 1.08 km north of the Camp 166 Road < 8 >, near the

unconformity on Burlington Granodiorite, shows the only red quartz-K-feldspar porphyry in the

area that contains axiolitically devitrified very long thin lenses (Fig. 4C.8c) identical to those in

the non-porphyritic sills. It is interpreted to be a xenolith, perhaps carried up from the ring

complex to the east, in the non-porphyritic dyke. This porphyry, and the porphyry sill on

Flatwater Brook, both contain some granophyre beards on K-feldspar phenocrysts. These have

also been seen in one sample of purple quartz-K-feldspar porphyry ignimbrite, that might be

intrusive as it contains no fiamme and the outcrop is isolated, 0.36 km east of the waterfall on

Flatwater Brook.

(d) Sediments

1. Conglomerates

Most of the sediments in the Mic Mac Lake Group are conglomerates. They range from

pebble to boulder conglomerate with clasts usually not more than 1 metre across, but occasional

larger clasts up to 8 metres across are found. The clasts can all be matched with local lithologies.

That is, the clast assemblages consist of variable proportions of Burlington Granodiorite, pink

aplite from the granodiorite, purplish mafic lava, rare epidosite pebbles from the mafic lava,

homogenous and flow-banded non-porphyritic maroon rhyolite, red quartz-feldspar porphyry

with or without ignimbrite fiamme, and rare non-porphyritic eutaxitic ignimbrite. One clast of

maroon flow-banded rhyolite agglomerate is similar to the auto-breccia, but is not precisely

matchable within the Mic Mac Lake Group, although it obviously belongs to the same silicic

volcanic suite. Two or three clasts of a purple fine-grained rock with K-feldspar phenocrysts up

to 1 cm across were seen. This lithology is apparently common as a

marginal facies to the syenite intrusions in the ring complex to the east (E.R.W. Neale, pers. comm.). It should be noted that no foliated metamorphic rock clasts are present.

Clasts are almost always fairly well rounded and moderately equidimensional, but the latter is probably controlled by the joints in the parent rock rather than abrasion, and the former to penecontemporaneous spherical exfoliation weathering as well as abrasion. Granodiorite clasts in particular are sometimes tabular with rounded edges, and lie flat in the bedding. No imbrication has been detected. Clasts always form a self-supporting framework. Bedding is relatively uncommonly seen within the thick mapped units of conglomerate, but interbedded units of coarse sandstone (granule-size clasts) and sandstone 1 to 2 metres thick are seen in places. These show that the boulder conglomerate occurs in individual beds at least 4 metres thick, and thicker beds are probably present, especially in the bottom of the larger abrupt palaeotopographic depressions. Pebble conglomerates occur in proportionately thinner beds, down to one pebble thickness. The clast size and proportions of clast types present are usually fairly constant within most of any one of the mapped conglomerate units. The only consistent exception to this is that there is usually an overall fining upward in the upper 10 to 20 metres of the thicker map units, which consists mainly of pebble conglomerate, coarse sandstone and some sandstone. In two cases, in the first two thick units of conglomerate 0.8 km south of the Camp 163 Road, there is a coarsening upward over the first 15 to 20 metres at the base of the units. These thick conglomerate units range up to 160 metres thick, and are commonly between 50 and 70 metres thick. The matrix of the conglomerates is mostly a granule to coarse sand size mixture of material from the Granodiorite, phenocrysts from the

porphyries, and lithic grains of the silicic and mafic volcanics. In general terms the conglomerates are rather poorly sorted, but there usually seems to be a restricted size range for a majority of the clasts in a bed, and thus a distinction between clasts and matrix. The first conglomerate unit above the unconformity on Granodiorite usually has a pink sandstone matrix. This is seen in all the conglomerate at the base of the Armageddon Formation. It is also found in the lower 30 metres or so of both thick conglomerate units from 1.0 km south of the Camp 163 Road north to Flatwater Pond, but only where each directly overlies the Granodiorite.

Most of the mapped lower contacts of the conglomerate units are to some degree erosional. Although there are variations in the proportions and types of the locally derived clasts present both within and between units, the only criterion that produces mappable units of any size or continuity is the presence or absence of significant quantities of Granodiorite clasts. It is probable that the lower two extensive conglomerate units in the two areas of the Snoopy Pond Formation are equivalent, with those in the north containing abundant granodiorite clasts, and those in the south almost none. Although a strict correlation is not possible, in particular the thick mafic lava units filling the palaeovalleys just under the lower of the two extensive conglomerate units in both areas may be equivalent, and also the first occurrence of porphyritic ignimbrite in both areas may be approximately equivalent. If so, it illustrates the very local derivation and deposition of the conglomerate clasts, with very different clast supply to two areas only 3 km apart. In the southern area of Snoopy Pond Formation, the thick conglomerate unit, above the lower two, contains abundant granodiorite clasts at the northern end. In this unit the general proportion of granodiorite clasts drops southward, accompanied by a reduction in general clast size, and eventually by passage sideways into a

sequence mainly of coarse sandstone and sandstone. Other variations in clast proportions also reflect very local and temporary availability of particular sizes and mixtures of the different clast types. For example, the conglomerates at the base of the Armageddon Formation contain very few mafic lava clasts, while those in its presumed equivalent south of Park Pond in the western strip contain a substantial proportion. Mafic lava clasts form a significant but variable proportion of almost all the conglomerates of the Snoopy Pond Formation. Similarly, non-porphyritic rhyolite clasts are very much more abundant than porphyry clasts in the Armageddon Formation and its western equivalent, reflecting their relative abundance as volcanics in the Formation. In the Snoopy Pond Formation, conglomerate locally may contain an abundance of rhyolite compared with porphyry clasts, and in other places the reverse. In a few places, the clast content of the conglomerates is closely related to the immediately underlying rock. In particular, the boulder conglomerate lying on the major erosion surface at the base of the Snoopy Pond Formation 1.2 km south of the Camp 166 Road is probably a fossil scree on the southern side of the palaeo-valley. It contains mainly porphyry boulders where it lies on the porphyritic ignimbrite, and both rhyolite and porphyry clasts below the contact with rhyolite. The only known example of a very angular sedimentary breccia is found 0.2 km east of Snoopy Pond <9>, where the angular pebble-size clasts are mostly of maroon rhyolite. There is a coarsening upward sequence from sandstone into the breccia in this locality. The pebbles are locally imbricated, and indicate transport from the northeast.

2. Lahar

Within the thick mafic lava unit that fills the palaeo-canyon 0.3 km south of the Camp 163 Road, a smaller palaeo-canyon has been cut, and is filled by a rock that is superficially a boulder conglomerate with very

few granodiorite clasts. It has a dark maroon haematite mud and silty quartz matrix. The northern

originally subvertical side of the valley occupied by this deposit is exposed, and shows that it is a

volcanic mudflow, or lahar. The first two metres of the conglomeratic rock measured sideways

from the valley wall of massive mafic lava consists almost entirely of pebble to cobble-size

clasts of vesicular and massive mafic lava. Within 1 metre of the wall, the clasts are highly

angular, and a few are seen caught in the process of being split apart by veins of the maroon

matrix (Fig. 4C.8d), which also penetrates the wall in places. This evidence is particularly

diagnostic of a viscous mudflow. The mafic clasts and the mafic lava forming the wall are a pale

green, discoloured relative to mafic lava further away, presumably due to alteration by the hot

water in the mudflow. Between about 2 and 3 metres from the wall, a rapid transition takes place

from almost wholly mafic lava clasts to a mixture of much larger and more rounded clasts

commonly up to 1 m and occasionally up to 2 m across, of mafic lava, rhyolite, and porphyry.

The silicic clasts are commonly tabular and lie flat in the bedding orientation. The clast/matrix

ratio is rather high for a mudflow, being at least 4:1. The southern outcrop shows a poorly

exposed contact with Granodiorite. There are very few clasts of granodiorite in the lahar, even at

the contact, and veins of the matrix are not seen to penetrate the underlying Granodiorite. It is

possible, but unlikely, that similar lahars have been mistaken for normal conglomerate

elsewhere.

3. Sandstone and coarse sandstone

Arenaceous sediments are uncommon relative to conglomerate in the mapped part of the

Mic Mac Lake Group. Bedded units of pink arkosic coarse sandstone (granule size 2-4 mm

clasts) and sandstone 1 to 3 metres thick

are found in places within the conglomerate units. Sandstone is more common in the upper half of the thick conglomerate unit southsouthwest of Snoopy Pond, which passes southward into a thick unit consisting mostly of sandstone and coarse sandstone with minor pebble conglomerate. A thick pink sandstone unit is also found overlying a sequence of alternating mafic lava and conglomerate 2.4 km south of Park Pond and on the southern part of Mic Mac Lake, and is most probably the same unit. Other mappable units of sandstone in the very detailed map area are only found near the base of the Snoopy Pond Formation (Plate 4), east of Snoopy Pond, and in the northern part of the Formation north and south of the Camp 163 Road (Plate 3). In addition, sandstone forms most of the sedimentary part of the thin development of the Mic Mac Lake Group northeast of Flatwater Pond.

The most common sedimentary structure in the sandstones is flat bedding, and these flat beds are often well graded. A current lineation has not been seen. Low-angle planar cross-bedding is uncommon. Approximately 15 outcrops provided some indication of the current direction, and currents from all directions except southeast are indicated, with a predominance from north and south that is probably due to the bias of the homoclinal outcrop. Although the sediments are thought to be derived from the general direction of the ring complex to the east, this direction is least commonly indicated. In most of the 15 outcrops, the direction indicated by several foresets is restricted to a particular quadrant, but in one, on the pond eastsoutheast of Snoopy Pond, omnidirectional currents are indicated. Rare ripple cross-lamination and small ripple marks were seen in only 3 localities in the sandstone near the unconformity north and south of the Camp 163 Road. Their orientation indicated currents from the north, and perhaps from the south, only. Small channels are commonly seen in the sandstone. The

sandstone beds are usually between 0.5 and 10 cm thick, and the graded beds may have a thin siltstone top up to 0.5 cm thick. One example of mudcracks has been seen in the latter. Two exposures contain units up to 2 m thick of graded crimson siltstone and shale. Beds in these are between 0.5 and 5 mm thick, and are mostly graded beds. They also show small channels, rafted mudflakes, and rare syn-sedimentary micro-faults. The clastic grains in the siltstone are mainly angular to sub-angular quartz, with minor feldspar and magnetite, and much haematite in the shaly parts. Clastic grains in the sandstones consist of angular to sub-angular quartz, with about an equal proportion of K-feldspar, albitised plagioclase, and magnetite. Minor quantities of rhyolite fragments, muscovite, biotite, sphene and epidote occur. Heavy mineral laminae are found occasionally, and in abundance in one outcrop. These contain mainly magnetite and sphene, with lesser epidote, zircon, and apatite, and rare hornblende and rutile. As in the conglomerates, all clastic material is therefore referable to the granodiorite and the volcanic rocks. Sandstone grain size is generally 1.5 mm or less, although beds with granule-size grains (2-4 mm) are common. The beds are overall rather poorly sorted, although the common grading produces better sorting within each bed. Some beds contain a few pebbles between 5 mm and 5 cm across 'floating' within the bed. Sandstones in the area northeast of Flatwater Pond are mostly finer grained (0.4 to 0.02, av. 0.1 mm) and better sorted, and with a higher proportion of quartz, than those to the south, although the grains are still subangular. These sandstones are a paler pink or a greenish-white weathering colour. The matrix of all sandstones examined in thin section is mostly calcite with some sericite.

Although sandstone is uncommon relative to conglomerate, the ubiquitous sandstone included in the bases and tops of the mafic lavas suggests

that rather more was transported through the section exposed than is now preserved in it.

4. Regolith

The unconformity of the Mic Mac Lake Group on the Burlington Granodiorite is exposed in many places in the burnt area. In all cases the unconformity is sharply defined, and in almost all cases the Granodiorite is massive and no more altered than elsewhere directly below it. However, for 0.6 km along the unconformity from the Camp 166 Road, and at 1.3 km north of the Camp 166 Road, the Granodiorite is cleaved for up to 60 metres thickness below the unconformity, and also contains small diffuse slightly reddened patches. It is suggested that this is a regolith of palaeo-weathered rock. At 0.24 km north of the Camp 166 Road a sedimentary sandstone dyke is seen in this regolith. Just below it is a rock that appears to be conglomerate with 100% granodiorite clasts and a chalky granodiorite derived matrix, and it is interpreted as spheroidally palaeo-weathered rock. The only other sandstone dyke seen in the Granodiorite is found at the north end of the remnant piece of regolith 1.3 km north of the Camp 166 Road.

Cleaved Granodiorite is also found a few metres under the unconformity in an outcrop on the north side of the Camp 163 Road, but the Granodiorite immediately under the unconformity at this locality is massive. It is suggested that this cleaved Granodiorite is connected with a small eastward dipping high angle thrust fault that has faulted out the unconformity just south of the Camp 163 Road. This is the only place known where it is faulted by a subparallel fault. Most of the Granodiorite north of the Burlington Road is cleaved and involved in eastward thrusting. However it is least deformed immediately under the unconformity with the Mic Mac Lake Group lithologies, and therefore it is suggested that the cleavage is mostly due to the deformation, although it could be influenced by a regolith in places.

(iv) Sedimentary environment

Welded ignimbrites require a subaerial environment, and it is therefore most likely that whole of the mapped part of the Mic Mac Lake Group was deposited subaerially. The massive mafic lavas, and the haematised nature of all the rocks are consistent with this environment, as are the sediments and their internal structures, the erosional nature of many contacts, and the steep palaeotopography preserved on the major erosion surfaces. The dominance of large clasts, the locally derived clast lithologies, and the immature nature of the sandstones show that the sediments were deposited close to their source.

The upward coarsening seen at the base of two and through the whole of another conglomerate units is probably due to the progradation of small alluvial fans. The more common fining upward at the top of thick conglomerate units could be due to the retreat of the head of alluvial fans, or perhaps to the erosional exhaustion of local sources after an increment of uplift. The alternation of mafic lava units and such conglomerate units could indicate a connection between such uplift and eruptions of mafic lava. Most of the conglomerates were probably deposited at the head of alluvial fans as sieve deposits (Bull, 1972). In these the conglomeratic debris is dumped as a self-supporting framework at the head of the fan, while the water and finer debris accompanying the larger clasts travel further through the natural soakaway formed by the framework of large clasts. The interstices are filled by finer material of later supplies of similar sediment. The thick conglomeratic unit south of Snoopy Pond passes southward with generally decreasing maximum grain size into a sequence mainly of sandstone. This is very probably a longditudinal section through a small alluvial fan. On a smaller scale, the individual graded sandstone beds, and graded pebble conglomerate to coarse sandstone beds, are probably due

to transportation and deposition by flash floods, with the finer material deposited later in each event as the current velocity decreased. Some sandstone beds, especially the graded beds,.and those containing 'floating' pebbles, are probably sandflows (grain flows or debris flows), that are commonly produced by flash floods on alluvial fans (Allen, 1970). The variable current directions indicated by the cross-bedding, and the rarity of indications of currents from the source area to the east and northeast is not understood. It is possible that some could be anti-dune cross-bedding, but this was not investigated. The thick sandstone unit found in three areas south of Park Pond could be more distal alluvial fan deposits or perhaps braided stream deposits. It has not been determined from the sedimentary structures whether the latter environment is represented. The local scree conglomerate/breccia on the side of the smaller canyon on the internal major erosion surface has been described above. Some of the conglomerate at the bottom of the larger canyon on this erosion surface is also probably a scree deposit. However, scree deposits have not been detected on the unconformity on Granodiorite, although they may be present.

(v) Palaeotopography and palaeogeography

Abrupt local palaeotopographic features of considerable original relief are preserved on the two major erosion surfaces of the Mic Mac Lake Group. As preservation of such features is unusual in older rocks, and as the map pattern might be confusing, the more striking features are described below.

Two palaeo-canyons are preserved on the internal major erosion surface south of the Camp 166 Road (Plates 1, 4). The larger southern canyon was originally about 300 metres deep, and this is the largest amount of palaeo-topographic relief of a very local nature that is preserved in the area. At least part of the conglomerate filling the bottom is probably fossil scree. The canyon can only be viewed on the map from a rather restricted

Fig. 4C.9. Palaeotopographic profiles and sketch reconstruction of local palaeotopography (descriptions and locations in text).
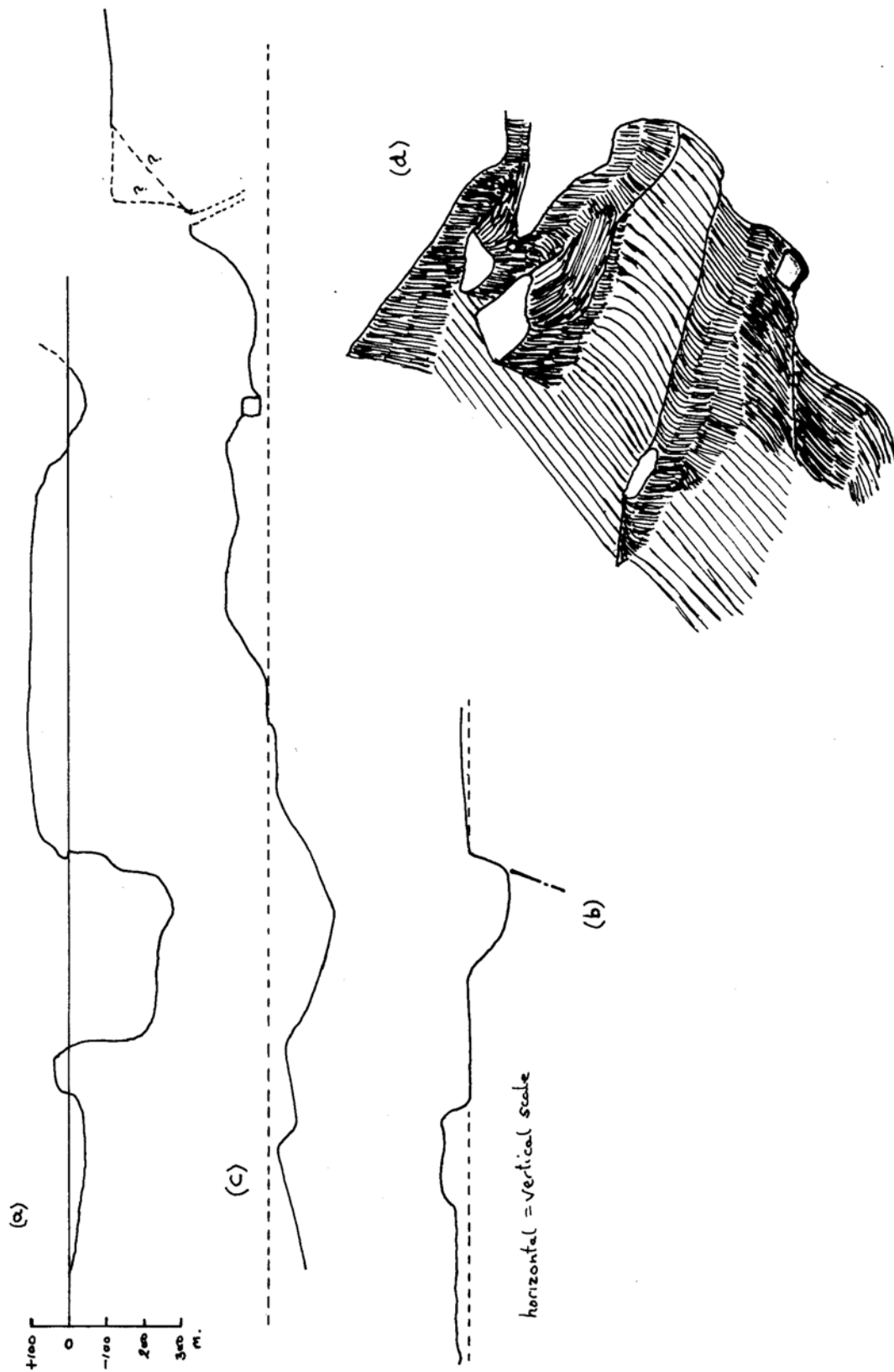
sector looking southwest if its profile (Fig. 4C.9a) is not to have an overhanging wall, and therefore the trend of the canyon axis is probably in this direction in (and near) the exposed section. A section through a small flat-topped steep-sided hill is preserved in the rhyolite at the top of the south wall of the canyon. On the north side, the protruberance at the base of the porphyritic ignimbrite is not due to movement on the contact with rhyolite below, and is probably a pillar rather than an overhang, because its exposed lower contact with mafic lava is not deformed and is subvertical. The base of the porphyritic ignimbrite intersects the underlying rhyolite when projected across the canyon. The units above the canyon are not faulted, and therefore a fault penecontemporaneous with deposition is a possible cause. However, even the poor outcrop below the canyon is sufficient to limit any such displacement to a barely adequate amount, and further, no suitable lineament is seen. Alternatively, it is possible that the overall shape of the canyon, and thus the relative position of the two sides, has been changed by bulk deformation. Although the rocks forming and filling the canyon are not visibly deformed, this is not conclusive evidence against this possibility. The smaller northern canyon on the internal erosion surface has had its northern side severely affected by complex deformation; the southern side has a probable fossil scree banked against it, as described above. This southern side shows a marked step at the contact of the rhyolite with the overlying porphyritic ignimbrite, and there is a less well-defined step on the erosion surface higher up the porphyritic ignimbrite. This stepped topography is characteristic of canyons cut in ignimbrite terrains, controlled by the cooling unit, and sometimes the flow unit, boundaries. It is possible that the upper step in this palaeo-canyon is controlled by a flow unit boundary.

4C.45

The unconformity of the Snoopy Pond Formation on the Burlington Granodiorite defines a small asymmetric palaeo-valley originally 60 metres deep 0.8 km north of the Camp 163 Road. South of the Camp 163 Road, a larger palaeo-canyon originally about 110 metres deep has a very straight, originally steep, northern wall (well controlled by the available outcrop). It is interpreted as. a fossil fault scarp because a very obvious lineament is seen in the granodiorite directly below it and in the same orientation. Several mafic dykes cutting the Granodiorite just north of this wall are in nearly the same orientation, and possibly fed the mafic lavas filling the canyon. This canyon also has to be viewed on the map from a restricted position looking southwest to avoid an overhanging wall in its original profile (Fig. 4C.9b). The smaller valley cut in the mafic lava filling the canyon, now filled by the lahar, is clearly due to renewed downcutting of the same watercourse. The remainder of the palaeotopography along this segment of the unconformity overlain by Snoopy Pond Formation is relatively subdued and small scale, except 1.2 km south of the Camp 163 Road, where conglomerate and mafic lava are banked against a section through a small ridge originally about 60 metres high. It is probably a ridge because of the change in the stratigraphy of the lower units across it, particularly the occurrence of sandstone to the south.

The other area where local and abrupt palaeotopography is well developed and controlled by exposure is at the base of the Armageddon Formation south for 2.3 km from the Camp 166 Road (Fig. 4C.10). The southern part of this segment is relatively simple, and shows a section through a relatively broad valley (profile in Fig. 4C.9c) that is blanketed by the non-porphyritic eutaxitic ignimbrite. The distribution of the pink porphyritic ignimbrite on the steeper northern side of this valley may indicate the presence of several penecontemporaneous faults, but no suitable

Fig. 4C.10. View of part of the Armageddon Formation.

Fig. 4C.10a. View north across the type section of the Armageddon Formation towards the Camp

166 Road. Red outcrops are rhyolite and ignimbrite. White outcrops are

conglomerate or Granodiorite.
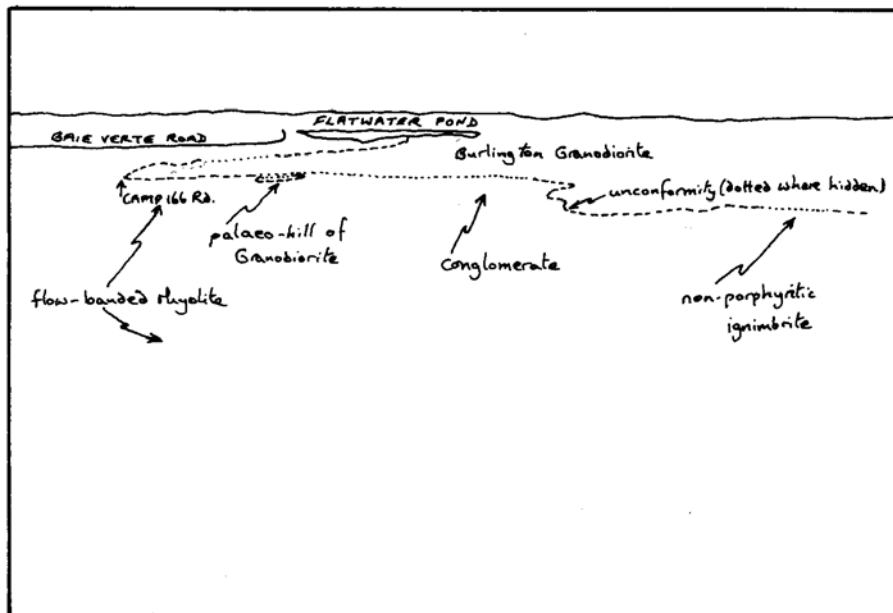
10b. Sketch showing unconformity and geographic features in 10a.

lineaments are present in the Granodiorite. The distribution does not require faulting, and could

be attained by erosion of most of a thick ignimbrite sheet.

A complex outcrop pattern is formed by intricate palaeotopography in the northern part of

this segment, in the small area up to 0.44 km south of the Camp 166 Road. The

palaeotopography here consists basically of a valley with an originally steep to vertical northern

wall about 300 metres high, and a less steep and less high southern slope. However, the oblique

section that is the present land surface intersects several small hills, and a ridge, of Granodiorite

within the valley. Present-day topography is insignificant as a factor in the outcrop pattern here,

or elsewhere. At the bottom of the valley, a rectangular block of Granodiorite about 100 metres

across is surrounded by conglomerate. The narrow strip on its south side is an originally vertical

cleft about a metre wide and at least 3 metres deep between the block and Granodiorite in place,

and contains the ignimbrite overlain by fumarolically cemented and altered pebble conglomerate.

The narrow strip of conglomerate at the eastern side is interpreted as an originally vertical cleft

filled by boulders. This rectangular block of Granodiorite is therefore interpreted as a slab which

became detached and slid a few metres downslope so as to open both clefts. The conglomerate

that fills the lower part of the valley has consistently large clasts. Its upper part overlies the

eroded lower portion of the second eutaxitic ignimbrite horizon, and to the south abuts the

remainder at a steep erosional contact. The ignimbrite overlying the conglomerate has a complex

outcrop pattern because of the local palaeotopography. It abuts the north wall of the valley, and

laps around the eastern half of a steep sided hillock of Granodiorite to the west, and its extent in

this area shows that conglomerate and Granodiorite underlie it at no great depth.

Further south the same ignimbrite extends in a finger and laps around the eastern side of an isolated inlier of Granodiorite. This inlier is also a steep-sided hillock, and the finger-like outcrop of the ignimbrite indicates an underlying ridge of Granodiorite running towards it. This ridge is also seen as a more subdued feature under the conglomerate at the bottom of the valley. This ridge also has a southwesterly trend. The ridge, and the valley to the north defined by the top contact of the eutaxitic ignimbrite, are blanketed by the flow-banded rhyolite, whose locally autobrecciated base thickens toward the axis of this valley. The north wall of the lower part of the valley defined by the Granodiorite unconformity becomes a ridge and ends in two small hillocks. This ridge separates the eutaxitic ignimbrite from the streaky rhyolite that covers these two hillocks, and the other one next to them. The streaky rhyolite is interpreted to have been extruded from a dyke that reached the surface on the north side of these hillocks and the ridge leading to them. This small ridge is traceable eastward to where the upper contact of conglomerate abuts the wall of the valley, because small outlying smears of conglomerate are found on the Granodiorite a few metres to the north of the main contact. A sketch of the palaeotopography in this small area is given in Fig. 4C.9d.

The lowest units preserved in the Armageddon Formation are found in the area 4 km south of the Camp 166 Road. They occupy a broad, rather poorly defined valley in the erosion surface, that was probably in part fault-controlled. In particular it is likely that the large southwest-trending fault bounding the southern side (or rather its precursor) was active during deposition of these lower units because they are not found immediately to the south of the fault.

Additionally, abrupt palaeotopography is partly responsible for the complex relationships just southwest and west of Snoopy Pond. Structural

complexity in the same area hampers the decipherment of the original relationships, but the conglomerate and mafic lava to the south probably rested originally against a hill formed of the red porphyry ignimbrite. Similarly, the units above the northern valley in the internal erosion surface are far too structurally complex to interpret their palaeotopographic relations, which are clearly important in forming their present pattern.

On a larger scale, the Armageddon Formation is almost wholly restricted to the area south of the Camp 166 Road, presumably mainly because of palaeo-topographic control. One heap of large angular boulders of purple porus-weathering non-porphyritic eutaxitic ignimbrite was found 0.3 km north of the Camp 163 Road and may be outcrop, and if so may be an erosional relict of the Armageddon Formation. If this is the case, the present unconformity in the segment north of the Camp 166 Road may not be at a very different level than it was when the Armageddon Formation was being deposited.

The second larger scale feature is the absence of the mafic lava and conglomerate of the lower Snoopy Pond Formation in the segment from 1.0 km south to 1.8 km north of the Camp 166 Road. Although the internal major erosion surface has not been positively identified in this area, it is probably close to the unconformity on the Granodiorite north of the Camp 166 Road, and becomes that unconformity 1.8 km north of the Camp 166 Road. It is probable that this central segment was the site of a positive topographic feature relative to north and south during the deposition of the lower part of the Snoopy Pond Formation, but presumably was not so during the deposition of the porphyritic ignimbrites of the upper part of that Formation. However, it is not understood why the units of the southern part of the Snoopy Pond Formation converge on the internal erosion surface at the bottom of the Formation at its northern detected limit 10 km south of the Camp 166 Road. There is no evidence that the sequence wholly of

porphyritic igniinbrites is unconformable on the upper mafic lavas in the Snoopy Pond Formation. A possible explanation is that there is a pene-contemporaneous fault (or faults) whose trace is nearly parallel to strike, either outcropping or below the exposed section. The strike and near strike faults interpreted in this general area may be such faults, or connected them. If so, the potential complexities in the relationships of the rock units that may have resulted are probably impossible to interpret. It is probable that this central segment was in part structurally controlled as the mafic lava and conglomerate of the lower part of the Snoopy Pond Formation reappear to the north across a southwest-trending fault, which well have had penecontemporaneous movement on it. Also it may not be coincidence that this central segment is where 1) almost all the rhyolite sills in the main sequence are found, 2) two dykes feeding rhyolite are interpreted to occur, and 3) the ring complex boundary (Plate 6) has a protruberance, while it has a salient under the thicker more complete sequence to the south.

The third major feature in the map area is the thickest preserved part of the main sequence adjoining the large southwest-trending fault at the southern end of the very detailed map area (Plate 4). It is quite possible that there is a large valley cut in the internal erosion surface at the base of the Snoopy Pond Formation next to the fault, but there is no exposure in this area. The Snoopy Pond Formation adjacent to the Fault is composed almost wholly of sediment, with a thick upper unit of sandstone. Although the exposure is poor, it is probable that the mafic lava units to the north are eroded out in this section next to the fault, and they appear to be truncated at approximately the same distance from the fault. It seems more likely that the mafic lavas would be truncated at the edge of the course of a persistent braided stream, rather than on the distal part

of an alluvial fan that is the alternative interpretation of this section, especially since they are well preserved on the more proximal (coarser) fan deposits to the north. Thus, this sediment sequence adjacent to the large fault may mark the course of a persistent braided stream, although it has not been shown that the sandstones are such braided stream deposits. It is presumed that the reason for the location of this thickest section of both Formations here is penecontemporaneous faulting on the precursor to the large southwest-trending fault. This could have controlled the location of a braided stream by erosion of a gorge along its trace further towards the source area.

Thus a dominant southwest-northeast trend is shown by the orientation of the palaeotopographic features, faults with probable penecontemporaneous movement and the mafic dykes in the Granodiorite. This is also the dominant fracture orientation in the Burlington Granodiorite, visible on the air photographs from the Camp 166 Road to the north end of the map area, and this is probably the controlling factor. Southwest-trending faults and parallel lineaments that are probably faults are common in the ring complex to the east (Neale and others, 1959). In the map area these faults are not radial, but are almost tangential to the outline of the ring complex.

There is a remarkable correspondence between the positions of the two valleys in the unconformity on Granodiorite south of the Camp 166 Road, and the two in the internal erosion surface, when the map is viewed looking down the axis of the valleys towards the southwest. It is probable that this indicates persistance of faults controlling the position of the valleys. However, this does not necessarily mean that such faults cut the section now preserved, but only that they were active further toward the nearby source area to the east.

(vi) Facing evidence and major structure

There are abundant examples of reliable facing indicators in the main sequence of the Mic Mac Lake Group. These include channeling, ripple marks, cross-bedding, erosional contacts, mafic lava picking up underlying sediment, and cleavage/bedding intersections in sandstones and occasionally with fiamme in ignimbrite cleaved margins. Graded bedding is always normal in the many instances it can be checked with the other facing indicators.

On the basis of these criteria, and the non-repetition of stratigraphy, the main part of the Mic Mac Lake Group in the map area as far west as the Park Pond Fault is a westward-facing homocline. Church (unpublished manuscript map) suggested that there was a major synclinal fold closure in the southern part of the Snoopy Pond Formation. Its axial trace was placed in the centre of the thick conglomerate unit that starts just southwest of Snoopy Pond. It is true that there is a fairly large synclinal fold at the northern end of this conglomerate unit, as shown by the boundaries of the map units and abundant facing indicators (Plate 4). However, it seems that there is a complementary anticlinal closure just to the west, that is difficult to recognise because of palaeotopographic and erosional relationships. This fold pair degenerates rapidly to the south, where abundant indications of bedding and facing from sandstone beds in the thick conglomerate unit show that there is a flexure in the wholly westward-facing bedding corresponding to the locally pronounced fold to the north (Plate 4). Smaller fold pairs exist in a few other places, including the complex area just above the northern valley on the internal erosion surface (also complicated by secondary deformation); just east of Snoopy Pond; 2.6 km north of the Camp 166 Road; and in the sandstone south of Park Pond. There is a suggestion that all these fold pairs result from differential deformation near palaeotopographic irregularities, and all appear to be

strongly generative and of very local extent in both directions along their axial planes. It is possible that there is a major synclinal fold closure in the thick sandstone unit on the western shore of Mic Mac Lake, but there is insufficient exposure and facing evidence to prove this. Several of the more westerly outcrops consist wholly of overturned beds.

The discontinuous strip of Mic Mac Lake Group to the west of the Park Pond Fault was interpreted above to be correlative with the lower Armageddon Formation. It is also interpreted (described later) to be unconformable on the Baie Verte Group, and to face east. Outcrop evidence for east-facing in this strip is found in three outcrops south of Park Pond where mafic lava is seen picking up and baking sandstone to the west of it in two places, and is seen to have an included cobble of pink porphyritic ignimbrite in another outcrop where the lava is to the east of the ignimbrite. The only other outcrop where reliable facing evidence is seen in this strip is in crimson siltstone and slate just on the east side of the Camp 166 Road (Plate 3 or 4). This rock is microjointed and it is difficult to obtain a specimen, but four specimens thin sectioned all showed west-facing evidence, from channeling, grading and cleavage/bedding intersection. The outcrop is very small, the specimens all came from a small part of it, and there is about 2 metres unexposed between it and the mafic lava to the west. It is suggested that there is a small fold in this locality, and that the specimens come from its west-facing limb. It would be possible to dig out the unexposed interval to check. It is thought that this is not sufficient evidence to negate the interpretation that this western strip faces east and is equivalent to the lower part of the Armageddon Formation.

This western strip is therefore a small remnant of the western limb of a large syncline or synclinorium, whose eastern limb is represented by

the main west-facing homoclinal part of the Mic Mac Lake Group. The Park Pond Fault is westward-downthrowing and is partly responsible for the modification of this synclinal major structure. However, there are highly deformed rocks adjoining the eastern side of the Park Pond Fault (but not structurally caused by it) from Park Pond northward for 1.3 km and from Flatwater Pond southward for 3.6 km. They are due to a major D_{1B} tectonic slide zone that is interpreted to have previously modified the synclinal structure by thrusting of the western part up over the eastern part.

(vii) Metamorphism

All the rocks except the mafic lavas are silicic and are relatively uninformative as to the metamorphic state of the sequence. In the main part of the map area between Flatwater Pond and Mic Mac Lake, haematite is abundant, although it is locally reduced in small diffuse patches near faults, quartz veins, and in the more deformed cleaved margins of the ignimbrites. The mafic lavas in this main area often partially retain clinopyroxene, although they are haematised and the plagioclase is wholly albitised. The wholly recrystallised mafic lavas consist of assemblages containing some or all of: Chlorite-calcite-haematite-albite-Al-epidote or Fe-epidote-sphene-sericite-actinolite-quartz. This assemblage is wholly compatable with those in the Baie Verte Group, of low greenschist facies. However, the general grain size in the main part of the Mic Mac Lake Group is somewhat finer than in Baie Verte Group rocks. On the eastern shore of Flatwater Pond, rhyolite cobbles in conglomerate become progressively more bleached from the rim inwards (reduction of haematite) towards the north. Rhyolite clasts are absent northeast of the Pond, but sandstone also becomes a very pale pink to white-weathering rock with little or no haematite northeast of the pond, and contains quite coarse-grained greenish white mica where most deformed. Mafic lava in

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this area is wholly recrystallised, with the same assemblage listed above, but without haematite,

and of a grain size more comparable with rocks in the adjacent Baie Verte Group. These changes

are attributable to the greater deformation of the Mic Mac Lake Group rocks in this area.

CHAPTER 4D. STRUCTURES OF THE BAIE VERTE LINEAMENT

(i) Main deformation and cleavage (D_{1B})

The Baie Verte Group in the map area is affected by one regional cleavage (S_{1B}) that is parallel to bedding within the limits of observation, except in the hinges of extremely rare and small-scale isoclinal folds (F_{1B}). It dips subvertically at the western side of the Lineament and changes more or less gradually to moderate westerly dips at the eastern side. It is traceable on the finest scale across the exposed contacts between the Baie Verte and Mic Mac Lake Groups, both where the contact is stratigraphic in the south, and where it is tectonic northeast of Flatwater Pond. The single cleavage seen in both Groups is, without question, the same, as proposed by Neale and Kennedy (1967). In the Mic Mac Lake Group, the cleavage is usually slightly oblique to, and steeper than, bedding, where the relationship can be accurately defined (usually only in sandstone). The development of the cleavage is variable, depending primarily on lithologies, so that in the Baie Verte Group, mafic volcaniclastics are usually well cleaved, pillow lava little cleaved, and dolerite sills and ophiolite gabbro not cleaved. In the Mic Mac Lake Group, unconsolidated margins to ignimbrites are well-cleaved, mafic lavas are only cleaved where deformation was relatively intense, sandstones are moderately cleaved to uncleaved, and the main flinty parts of ignimbrites and rhyolites, and conglomerates, are uncleaved. The cleavage is also found locally in Burlington Granodiorite adjoining the Mic Mac Lake Group, where regolith is under the unconformity, and in the area where the Granodiorite has been involved in tectonic slides northeast of Flatwater Pond.

A peculiar development of the cleavage is sometimes seen locally in dolerite sills and ophiolite gabbro. It consists of usually narrow zones

Fig. 4D.1. Miscellaneous structures.
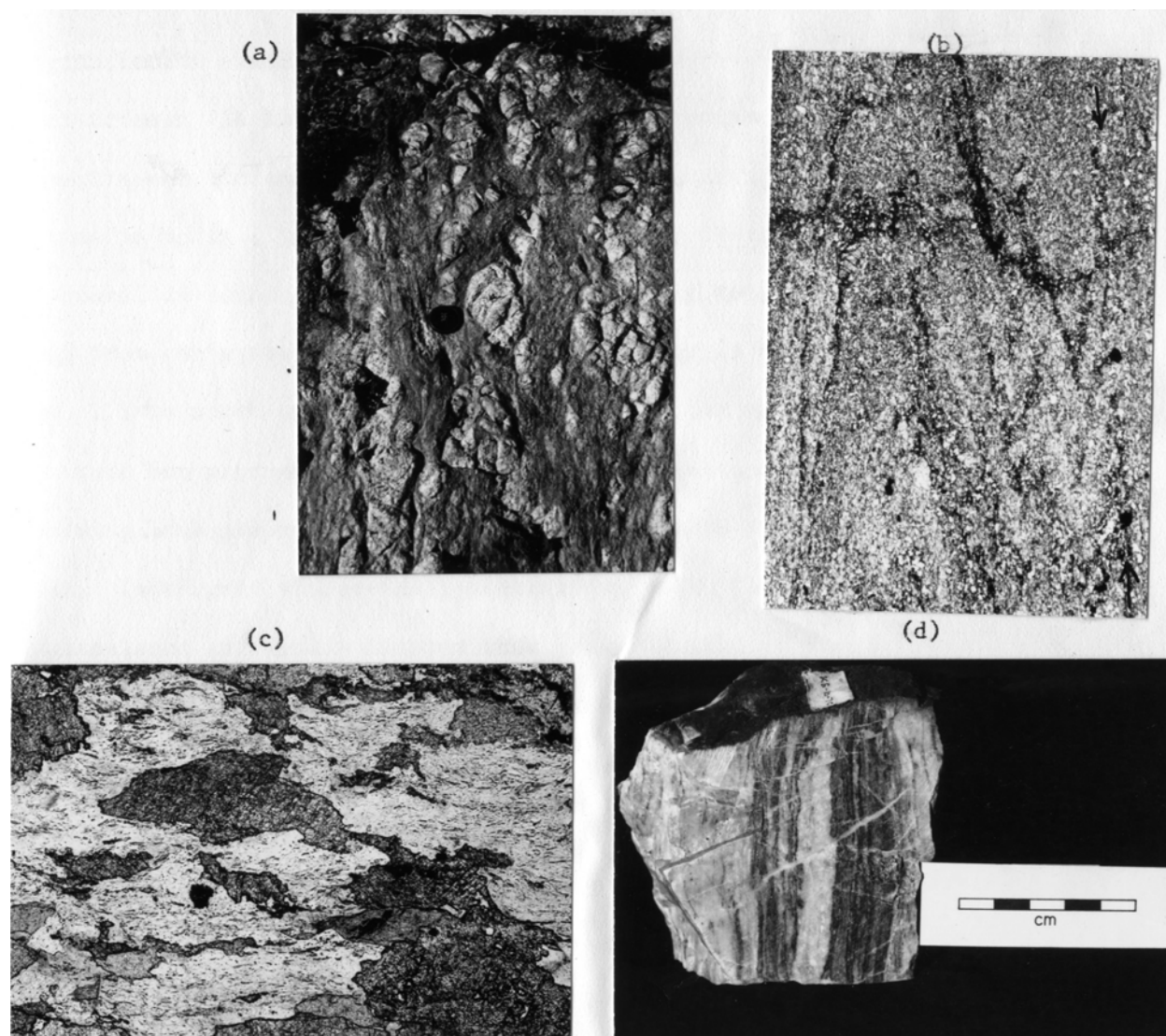
(a) Tectonic pseudoconglomerate; gabbro in Boudin Pond Formation

(b) F_{1B} folds in banded marble; tectonic slide near right of frame (arrowed). (Thin section X10)

(c) Virginite; S_{1B} foliation of elongate magnesite grains and fuchsite inclusion trails in quartz.

   (Thin section X40)

(d) Virginite; S_{1B} foliation defined by magnetite dust foliae.

of granulation outlining phacoids of undeformed rock between about 1 and 10 cm across. It is very similar to the shear polyhedra structure seen in serpentinised ultramafic rock, which, on the eastern margins of the large ultramafic bodies, is also a product of the Baie Verte D_{1B} deformation. The structure, in both the gabbro and the ultramafic rocks, is presumed to develop from conjugate fractures. In one locality in the Boudin Pond Formation, 1.1 km north of the outlet of Slink Pond, an extreme development of this structure has produced a classic example of a tectonic pseudo-conglomerate, involving both gabbro and the diabase dykes, which are seen to be still in place. Undeformed ellipsoidal phacoids are seen 'floating' in wide zones of granulated and weakly cleaved rock (Fig. 4D.1a).

The coplanar to sub-coplanar orientation of the schistosities in the Fleur de Lys schists to the cleavage in the Baie Verte Group adjacent to them across the narrow tectonic contact zone cannot be used to prove equivalence between the cleavage and one or other of the schistosities. However, the orientation of the S_{1B} cleavage near the boundaries of the Baie Verte Group is everywhere essentially parallel to the local orientation of the nearby boundary. Overall the general strike of the cleavage parallels the Lineament.

In the northern and especially the northwestern part of the Baie Verte Group, north from Flatwater Pond, there is often a subvertically pitching 'longrain', or stretching fibre (L_{1Ba}), on cleavage surfaces. This is only seen occasionally, and weakly, developed south of the north shore of Flatwater Pond. It is formed by a subparallel alignment of fibrous minerals (actinolite) in the cleavage surface, and is seen megascopically as a very small-scale and often not obvious hackly fibre on cleavage planes. On the north shore of Flatwater Pond, maf ic volcanic clasts in the Jukes Point Formation are strongly elongated colinear with the longrain, and elsewhere

deformed clasts are seen with a less pronounced preferred elongation direction also colinear with the longrain. The longrain therefore reflects a preferred elongation direction in the plane of the cleavage, but it does not necessarily mean that the deformation was a constrictional as opposed to a flattening deformation. The finite strain ellipsoid resulting from any deformation can be characterised by a shape parameter, $K = a—1/b—1$, where $a = z/y$, and $b = y/x$, and z, y, x are the long, intermediate, and short principal axes of the ellipsoid (Flinn, 1962). When $K = 1$ the deformation is plane strain; $K > 1$ is constriction; and $K < 1$ is flattening. In the Baie Verte Group mapped, only one small area shows strongly elongated (constricted) clasts. This is in the Jukes Point Formation on the north shore of Flatwater Pond and on the Burlington Road. Measurements of the axial ratios of deformed clasts in one specimen of conglomeratic mafic volcaniclastic rock from the Burlington Road gave a value for K of about 3. Measurements on argillite and mafic volcanic clasts in a specimen of black slate matrix conglomerate from the Kidney Pond Formation on the Burlington Road gave a K value of about 1. Elsewhere in the map area, qualitative observations of deformed clasts and pillows suggest the K value is 1 or less, so the deformation mostly resulted in plane strain or flattening strain, with a usually subvertical maximum elongation direction. The amount of deformation obviously varies according to the lithology, and also rocks south of Flatwater Pond are generally less deformed than those to the north. The argillaceous clasts in the measured specimen from the Kidney Pond Formation give values of 75 to 90% for the amount of flattening, and some pillows on the north shore of Flatwater Pond also show this extreme flattening. However, in the area south of Flatwater Pond, flattening probably does not exceed 50%. It may be noted that the high values for the flattening are rather larger than those recorded for rocks in the paratectonic British Caledonides (Dewey, 1969).

5

Fig. 4D.2. Baie Verte Lineament minor structures.

(a) F_{1B} fold closure; fine-grained mafic volcaniclastics near Camp 166 Road.

(b) F_{1B} fold closure refolded by F_{2M}; sandstone near Burlington Road.

(c) F_{2B} folds; mafic volcaniclastics, Jukes Point Fm., N. shore Flatwater Pond.
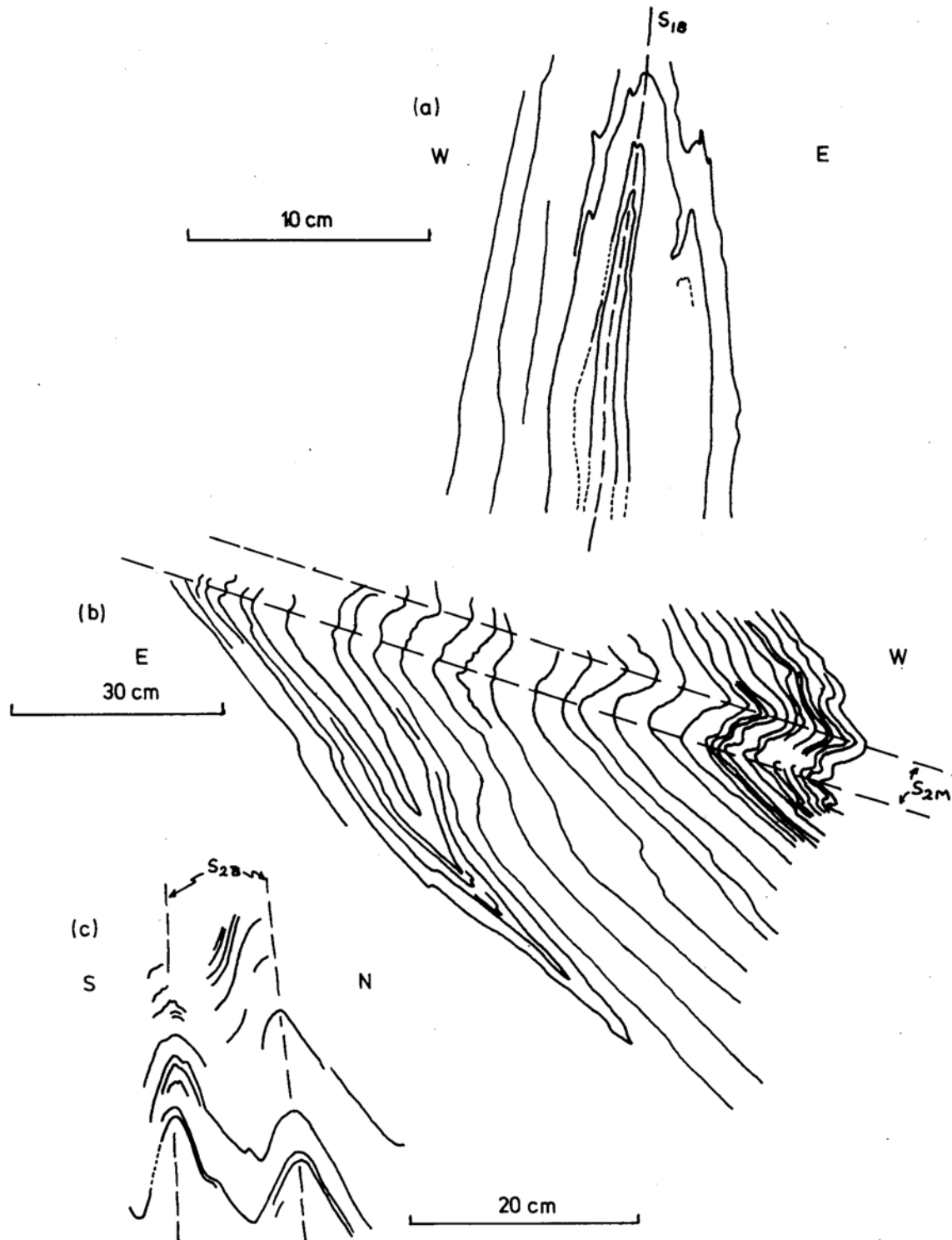
Minor folds associated with the S_{1B} cleavage are rare, seen in less than 10 outcrops in the Baie Verte Group mapped. A relatively open parasitic fold pair with a wavelength of about 5 metres is seen west of Slink Pond in the outcrop just south of [39]. Otherwise all are isoclinal and of hand specimen scale or a little larger. An example is shown in Fig. 4D.2a. The axes of the F_{1B} folds in the banded marble 3.0 km north of the Burlington Road Junction are colinear with the longrain, which is anomalously shallowly plunging in this segment. Similarly, a minor F_{1B} fold east of Kidney Pond plunges at about the same angle as weak longrain in a nearby outcrop. However, the fold pair west of Slink Pond is shallowly plunging while the weak elongation of clasts in nearby mafic volcaniclastic conglomerate is steeply plunging. It is possible that fold axes have been rotated towards the elongation direction in areas of stronger deformation (Dewey and McManus, 1964), but the rarity of minor folds does not permit a definite conclusion.

The longrain mostly pitches steeply, and in one segment moderately, southward on the cleavage in the strip along the Baie Verte Road north of Flatwater Pond. Eastward along the shore of Flatwater Pond, it changes, and usually pitches steeply northward. South of the Camp 166 Road the few observations show a change southward from moderate to steep northerly pitch. These changes reflect slight local adjustments to the dominant subvertical major extension direction. The frequency of development of the longrain is clearly correlated with the change from less to more deformed rocks north across Flatwater Pond.

F_{1B} folds are uncommon in the Mic Mac Lake Group, and the longrain has not been seen. In the area south of Flatwater Pond, F_{1B} folds are very impersistent (generative) relatively open parasitic fold pairs with a wavelength up to 200 metres. Northeast of Flatwater Pond, minor F_{1B} are quite common in sandstone just north and south of the Burlington Road.

These are tight, nearly isoclinal folds, reflecting the more intense deformation in this area (Fig. 4D.2b). In the area south of Flatwater Pond, most F_{1B} folds plunge shallowly to moderately southward, but the development and plunge of the folds are thought to be mostly controlled by ductility contrasts across, and the geometry of contacts with local palaeo-topographic relief. Northeast of Flatwater Pond, most minor F_{1B} folds plunge shallowly southward.

In the Baie Verte Group, the only larger scale F_{1B} folds are a parasitic fold pair interpreted in the area of northwest Flatwater Pond. In the axial region of the anticline of this pair on the Burlington Road, minor isoclinal F_{1B} folds on a centimetre scale in laminated green argillite are seen cut by minor tectonic slide planes. The slide planes are so small and narrow that they hardly differ from normal cleavage surfaces, and the displaced parts of the folds make an apparent truncated cross-lamination. This structure is also seen in banded marble in the outcrop 3.0 km north of the Burlington Road Junction, and an example seen in thin section (Fig. 4D.1b), shows an isoclinal fold closure cleanly sliced off on the plane of cleavage.

(ii) D_{1B} tectonic slide lithologies

Major tectonic slides that formed during the D_{1B} deformation are found on the western boundary, within, and north from Flatwater Pond, on the eastern boundary of the Baie Verte Lineament. Tectonic slides are well-localised zones of high strain, formed during a compressive deformation, and essentially coplanar with the foliation produced during such deformation (see Chapter 3 (iv) (a)), and containing an intensely developed version of the foliation. Although it is preferable to confirm their presence by the offset and/or removal of stratigraphic units, they are readily recognised on the basis of their distinct, strongly foliated lithologies in outcrop.

4D.6

The Western Lineament Boundary Slide, equivalent in part to the Baie Verte Road Fault of Neale and Kennedy (1967), separates Fleur de Lys schists from the volcanics and sediments of the Baie Verte Group, and also runs along the eastern margin of the two large ultramafic bodies. In places, this slide zone contains up to 80 metres, but usually less than 40 metres width of a rusty-weathering, ultramafic-derived, Fe-bearing magnesite-quartz-fuchsite rock, known locally as 'virginite'. The proportions of the three minerals are typically 95%, 5% and a trace, respectively, and accessory magnetite dust and pyrite are usually present. It usually contains one strong foliation, defined by the strongly-flattened shape of magnesite grains (Fig. 4D.lc), fuchsite clots, magnetite dust trails (Fig. 4D.ld), and fuchsite inclusion trails in quartz grains (Fig. 4D.lc). In places it is statically recrystallised, which obscures the foliation, but remnants can usually be detected, especially the fuchsite inclusion trails in quartz grains. This one foliation is coplanar with the Baie Verte D_l cleavage, and is thought to be its equivalent in this carbonate rich rock. Lithologies on the slide zones within the Baie Verte Group are characterised by an intense platy cleavage, and are wholly distinct from any of the normal volcaniclastic and argillaceous sediments. They are often rusty-weathering, due to abundant disseminated Fe-bearing carbonate mineralisation. These slide zones outcrop just west of Flatwater Pond Provincial Park, in two places on the north shore of Flatwater Pond, and on the Baie Verte Road 0.3, 1.0, 2.0, and 9.2 km north of the Burlington Road Junction. They are typically up to a few metres wide, but in the last locality mentioned the zone is about 50 metres wide. Rocks in a zone adjoining the narrow sharply-defined zone of platy cleavage are often relatively more intensely cleaved than elsewhere. In the first of the localities mentioned previously, and in an outcrop (which has now been destroyed) of the same slide, just south of the Bear Cove Road .Junction, specks of fuchsite

are seen. This fuchsite, and the more widespread carbonate mineralisation, which is restricted to the slide zones, show the affinity of the slide zones within the Baie Verte Group to the Western Lineament Boundary Slide. The CO_2 and the Cr are in the latter certainly, and in the former probably, derived from the ultramafic rocks during the D_{1B} deformation.

The Eastern Lineament Boundary Slide, found on the eastern shore and northeast of Flatwater Pond between Baie Verte Group and either Mic Mac Lake Group or Burlington Granodiorite, is not more than 20 metres wide. It is partially exposed on the promontary in the central part of the eastern shore of Flatwater Pond, where sandy mafic volcaniclastic rock rapidly becomes very intensely cleaved and rusty-weathering (?carbonate-bearing) within about 10 metres of the contact with Mic Mac mafic lava. It is almost fully exposed just south of the Burlington Road, where the slide lithology consists of wholly recrystallised mafic volcaniclastic rock containing abundant long, thin (1 cm or less) schileren derived from pink silicic tuff. It does not have a platy cleavage, because.the rock contains little chlorite, but the pale schlieren define a strong foliation, and the rock is clearly intensely and disruptively deformed. The adjacent Granodiorite is moderately to strongly cleaved, and retrograded. A tectonic slide contact between a thrust slice of cleaved Granodiorite and underlying Mic Mac Lake Group is exposed 1.1 km north of the Burlington Road at the western side of the isolated lens-shaped area of Mic Mac Lake Group. All that is seen at the contact is 8 cm thickness of platy highly-cleaved material between cleaved Granodiorite above, and moderately cleaved Mic Mac mafic lava below. Most of the platy slide lithology is derived from the Granodiorite. This example clearly illustrates the hazards involved in defining stratigraphy in moderately to strongly deformed terrains, because it is estimated that

there has been at least 500 metres of thrust displacement on this thin slide zone.

A major tectonic slide zone is exposed in two separate segments at the western side of the main sequence of the Mic Mac Lake Group south of Flatwater Pond. Both segments abut the eastern side of the Park Pond Fault and its major splay, which are responsible for the disappearance of the slide zone, between the two exposed segments, and south of Park Pond. The southern segment, which extends for 1.8 km north from Park Pond, consists almost wholly of extremely strongly deformed derivatives of silicic volcanic rocks. Their original haematite has been wholly reduced, and the rocks are mostly white to pale buff quartz-albite-sericite 'schists'. They appear, at least in part, to pass abruptly along strike into normal flinty red ignimbrites but, even if they were originally unconsolidated lateral equivalents, they are now far more intensely deformed than the unconsolidated cleaved marginal material of the normal ignimbrites elsewhere. Therefore, in contrast to the interpretation of Neale and Kennedy (1967), these rocks are interpreted here as due to deformation on a major tectonic slide zone within the Mic Mac Lake Group. Similarly, the other segment of intensely deformed rocks, which runs for 3.6 km south from Flatwater Pond, is interpreted as the same major slide zone. Most of the rock in this zone is derived from pebble to cobble conglomerate, with minor amounts derived from mafic lava. A small proportion may have been derived from weakly consolidated silicic volcanics, but as the trachyte flows within this zone are internally little deformed, it is unlikely that any is derived from flinty silicic volcanics. The trachytes have been severely sliced by dislocations subparallel to strike, and the eastern boundary of this zone of highly deformed rocks cuts across stratigraphic units in this same area 2.9 km south of the Camp 163 Road (Plate 1). The amount of flattening defined by the

Fig. 4D.3. Baie Verte Lineament minor structures.
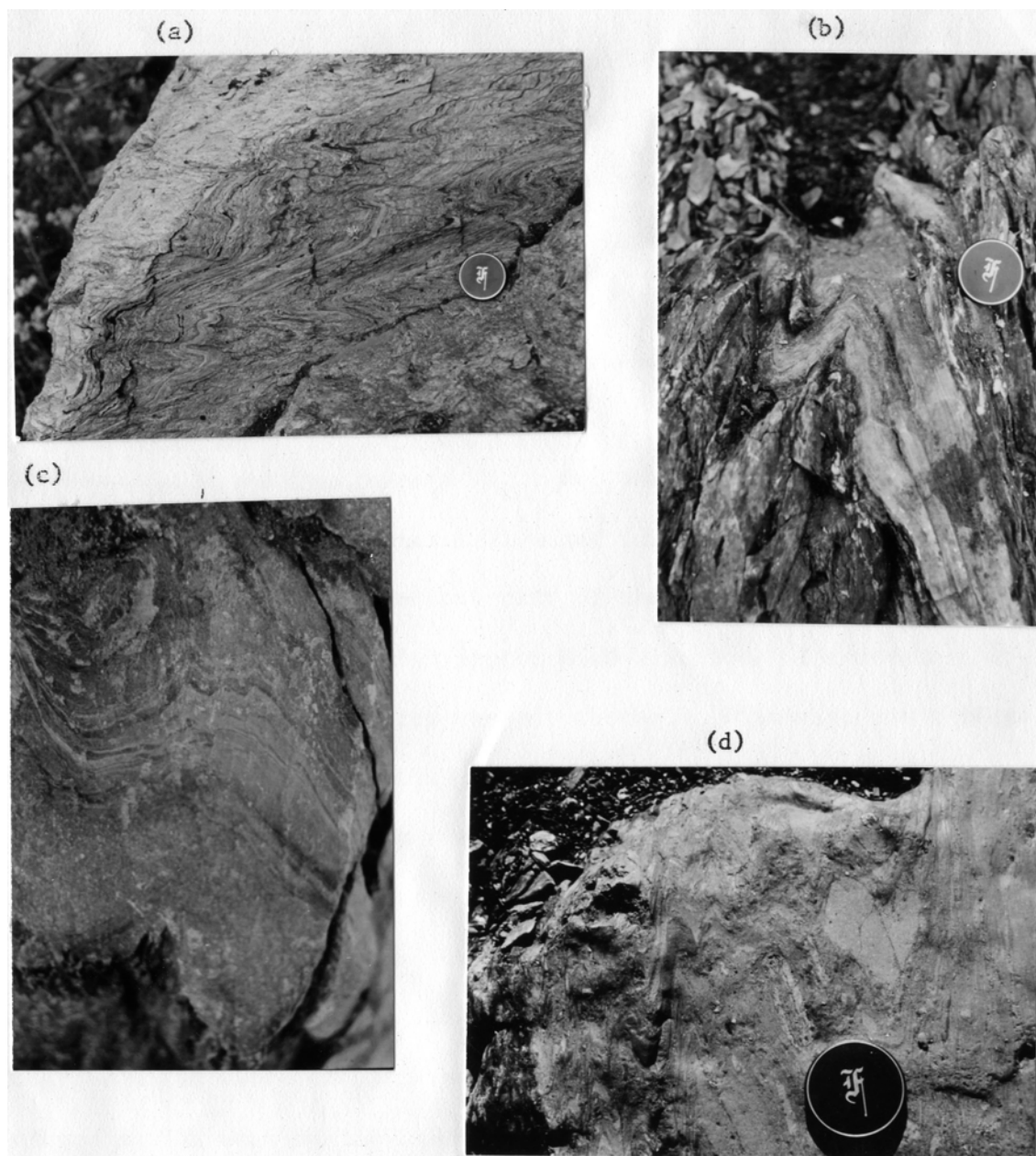
(a) F_{2M} minor folds; (W—E section) 1.0 km. S of Camp 163 Road.

(b) F_{2B} minor folds; banded marble, NW shore Flatwater Pond.

(c) F_{2B} minor folds; mafic volcaniclastics and argillite, W. Burlington Road.

(photo about 25cm. across).

(d) F_{2B} minor folds deforming clasts in Kidney Pond Fm., N shore Flatwater Pond.

4D.9

conglomerate clasts in these rocks is estimated to be commonly as much as 90%. The clasts

become hard to recognise when more strained than this, but some of the rock is almost certainly

derived from conglomerate.that has suffered flattening strains greater than 90% shortening (Fig.

4D.3a).

(iii) Sub-regional secondary deformations

(a) Secondary deformation in the western half of the Baie Verte Group (D_{2B})

A subvertical strain-slip cleavage (S_{2B}) and associated usually steeply plunging open to

close asymmetric sinistral minor folds (F_{2B}) of bedding and the S_{1B} cleavage are found in the

western part of the Baie Verte Group from Kidney Pond to 1.2 km north of the Burlington Road

Junction. The strain-slip cleavage is consistently oriented with respect to the S_{1B} cleavage and

strikes from 10 to 60, usually about 30 degrees more northeasterly than S_{1B}. The minor F_{2B} folds

are therefore all southwesterly plunging, and asymmetric, verging toward a hypothetical antiform

to the south. Typical examples of these folds are illustrated in Figs. 4D.3b, c; and 2c. This

deformation is restricted to within 0.7 km east of the Western Lineament Boundary Slide, except

along the north shore of Flatwater Pond and the Burlington Road, where it is developed as far

east as the lower contact of the Slink Pond Formation. The structure is sporadically developed,

and is only common on the western and northern shores, and north of Flatwater Pond. One

development of a possible conjugate member is seen in an outcrop of the Kidney Pond

Formation just south of Kidney Pond. Two examples are seen between Kidney Pond and Slink

Pond, but otherwise the structure is restricted to the segment stated above. The minor folds are

occasionally developed without much, or any, associated strain-slip cleavage. The strain-slip

cleavage may be coarse-spaced (~1 cm) but is usually more finely spaced (~1 mm). F_{2B} fold

plunge

Fig. 4D.4. Northwest Flatwater Pond, showing F_{2B} fold pair.
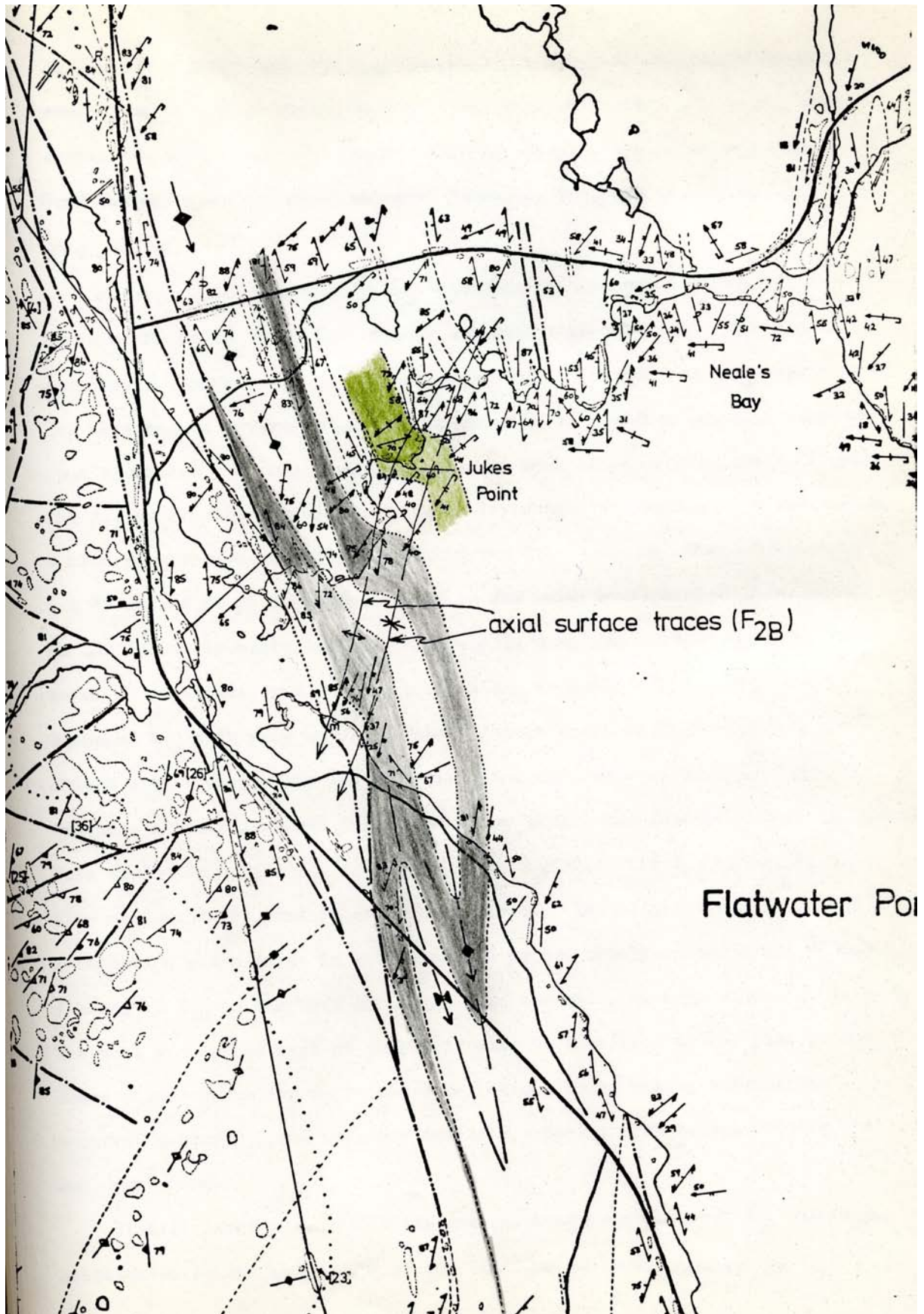
is always colinear with the L_{1Ba} longrain and maximum elongation direction within the limits of discrimination, and it is possible that the D_{2B} deformation was 'keyed' by this earlier fabric. Clasts in part of the Kidney Pond Formation on the north shore of Flatwater Pond are folded by F_{2B} (Fig. 4D.3d).

A larger scale asymmetric F_{2B} fold pair is developed on the northern shore of Flatwater Pond (Fig. 4D.4), and this fold pair also crosses the northwestern shore. On the short limb of this fold pair, the F_{2B} minor folds are very common and are relatively symmetric, or have the opposite vergence than elsewhere. In two localities, one in each hinge zone of the fold pair, the S_{2B} strain-slip cleavage is abnormally strongly developed and approaches a true penetrative cleavage, obscuring the S_{1B} cleavage. These localities are 80 metres NW of Jukes Point, and on the point southwest of Jukes Point. In the latter locality, the S_{2B} cleavage is very intense and platy in a zone about 15 cm across, and this is a small D_{2B} tectonic slide. This large S-shaped F_{2B} fold pair mapped on the northern shore of Flatwater Pond conflicts with a sketch map and interpretation published by Church (1969), defining a later Z-shaped fold in the same area. His interpretation has resulted from confusion of the S_{2B} strain-slip cleavage with the S_{1B} cleavage, and also of some later kink folds with F_{2B} folds. The slight variations in S_{2B} orientation which occur in this area can be adequately accommodated by some fanning of S_{2B} in the fold cores, as seen elsewhere in some minor F_{2B} folds. The most pronounced part of this fold pair is developed in the part of the Jukes Point Formation where the local strong constriction deformation occurred during D_1, and this may indicate a genetic connection between the two structures.

Similar large scale F_{2B} folds are interpreted from local S_{1B} cleavage orientation in the area south of the Old Camp 32 Road. Between the Old

4D.11

Camp 32 Road and opposite the Camp 166 Road, F_{2B} minor folds and the S_{2B} strain-slip fabric are seen cutting and folding the foliation (S_{1B}) in two separate outcrops of virginite on the Western Lineament Boundary Slide. In the same segment, three separated outcrops of Birchy Schist adjacent to the W.L.B. Slide contain a very local D_4 fracture cleavage and asymmetric minor folds of the same geometry and orientation as S_{2B} and F_{2B}. They are interpreted as due to the same deformation. This is best displayed in [12], where Fleur de Lys F_3 folds are clearly folded by the local F_4 folds.

A secondary cleavage very close to the orientation of S_{1B} was occasionally suspected to be present in some outcrops in the small area mapped beyond 6.5 km north of the Burlington Road Junction. However, a thin section of one of the clearer examples showed that the structure is produced by discrete, relatively coarse-spaced, somewhat anastomosing S_{1B} planes rich in chlorite. Thus, no convincing or consistent examples of any secondary cleavage or folds were seen in this area.

(b) Secondary deformation in the Mic Mac Lake Group and eastern part of the Baie Verte Group (D_{2M})

A set of secondary open to close asymmetric minor folds (F_{2M}) of the S_{1B} cleavage and bedding, with an associated sub-horizontal strain-slip cleavage (S_{2M}) is seen in suitably foliated lithologies in the main sequence of the Mic Mac Lake Group from Park Pond to the northern limit of mapping, 2 km north of the Burlington Road. It is also seen in cleaved Granodiorite northeast from Flatwater Pond. South of Flatwater Pond, it is not seen west of the Park Pond Fault, except in one outcrop of the Baie Verte Group on Flatwater Brook, 0.35 km north of the Camp 166 Road. The structure is common in the Baie Verte Group along and to the north of the northeast shore of Flatwater Pond as far west as the Park Pond Fault. The sub-horizontal

strain-slip cleavage generally dips gently southwest, west, or northwest, always at a shallower

angle than S_{1B} and bedding. The folds are therefore asymmetric and 'monocline-like' stepping

down westward with short steep limbs, and plunging gently north or south. Most examples of F_{2M}

minor folds seen in outcrop are small-scale, and have short limbs from a centimetre to a few tens

of centimetres long. These small-scale folds are often parasitic folds on the short limbs of larger,

more open 'monoclinal' step-folds, whose short limbs range from a few metres to a few tens of

metres long. Examples of minor F_{2M} folds are illustrated in Figs. 4D.3a; 2b. F_{2M} folds large

enough to affect the map pattern significantly have not been positively identified, and all large

scale open flexures in the Mic Mac Lake Group appear to be due to later deformation. These

asymmetric F_{2M} folds represent minor net movement of the moderately dipping material at the

eastern side of the Lineament westward toward the main steeply inclined part. This is significant,

as it will be shown later that the movement of material at the eastern side of the Lineament

during the D_{1B} deformation was eastward.

There are a few examples of possible small-scale F_{2M} minor folds west of the Park Pond

Fault, on the northern shore of Flatwater Pond to the western side of Neale's Bay and north of

the Provincial Park (Plate 2). Almost all these examples are folds without associated strain-slip

cleavage. Small-scale possible examples of F_{2B} minor folds, also without associated strain-slip

cleavage, are found at the western side of Neale's Bay and northwest of the Provincial Park. The

structures marked as F_{2B}, S_{2B} on :he western shore of Flatwater Pond north from opposite the

Bear Cove Road junction are definitely D_{2B} structures. However, the other examples mentioned

were identified as F_{2M} or F_{2B} on the basis of their vergence and the pitch of their axes in S_{1B}. They

have been marked on the map as either F_{2B} or

Fig. 4D.5. Block diagram showing typical orientations and vergences of D_{1B}, D_{2B} and D_{2M} structures.
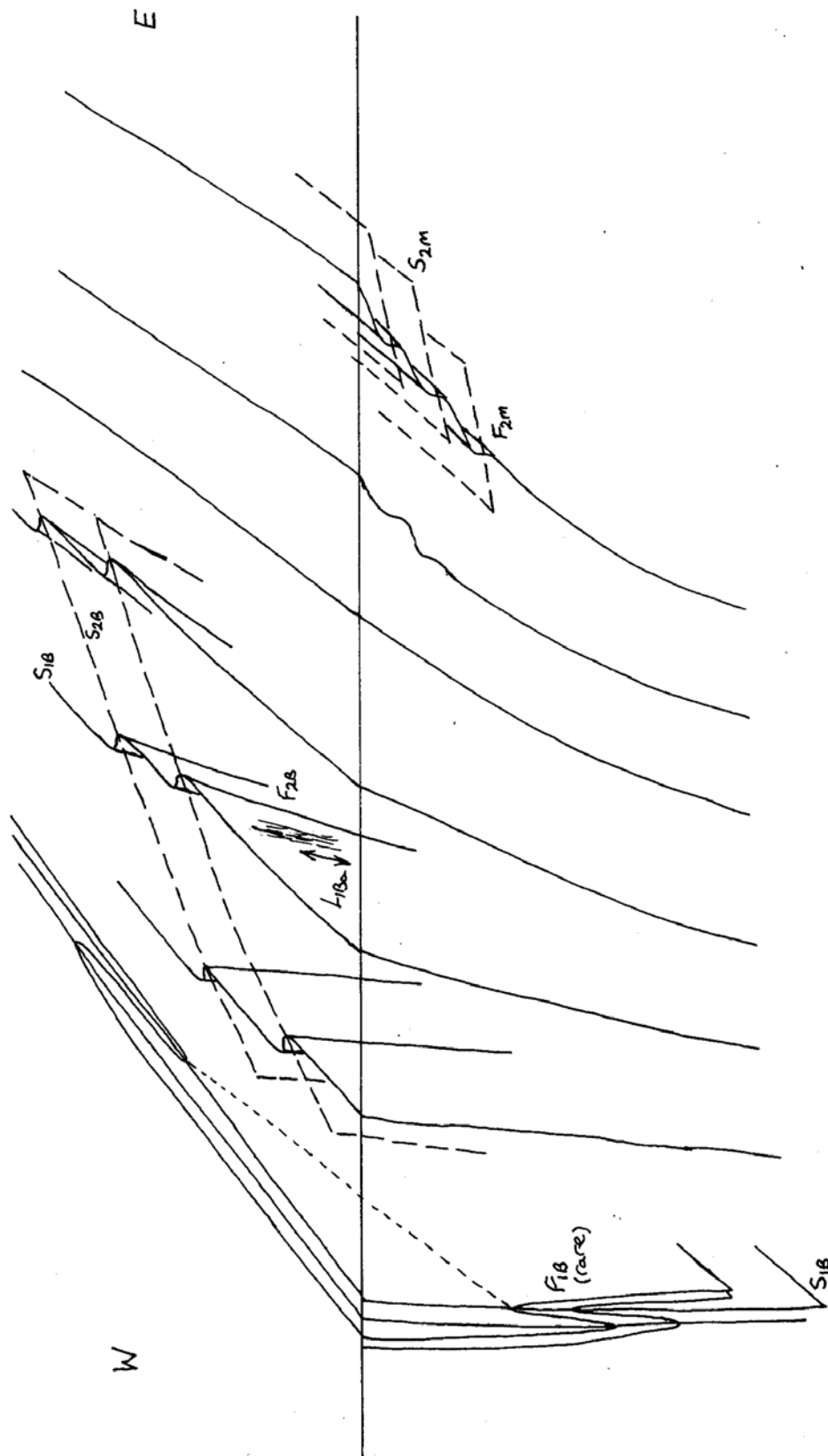
F_{2M}, but they may be isolated minor folds and crenulations not due to D_{2B} or D_{2M}. In the area of the point at the western side of Neale's Bay, a few of these folds, marked as F_{2B} on.the map, have a moderate southerly pitch in S_1, between the normal steep and shallow pitch of F_{2B} and F_{2M} respectively. These may be due to the steeply-dipping S_{2B} intersecting moderately dipping S_{1B} or later reorientation by large shallowly plunging flexures of S_{1B}, or they may be an intermediate development indicating that D_{2B} and D_{2M} were coeval. The evidence is insufficient to prove the latter hypothesis.

A sketch block diagram showing the typical orientation and style of the primary (D_{1B}) and secondary (D_{2B}, D_{2M}) structures of the Baie Verte Lineament is shown as Fig. 4D.5.

(iv) Late local deformations

(a) Kink bands and large-scale gentle to open flexures

Small-scale kink bands with angular hinges are seen occasionally in well-cleaved rocks of the Baie Verte Lineament. They were not identified as belonging to well-defined sets, apart from vertically plunging examples, representing minor axial shortening along the Lineament, found in places in the area north of the Burlington Road Junction and on the northwestern shore of Flatwater Pond.

A large open dextral steeply-plunging kink-fold pair is defined by the orientation of S_{1B} and the boundary of the thick pink silicic tuff unit at the northeast corner of Flatwater Pond. Although the obvious displacement on the subvertical Park Pond Fault is a westward downthrow, it is possible that there is some lateral displacement on it, and that this kink fold is connected with such displacement. Another similar large dextral steeply-plunging kink-fold pair is defined by the S_{1B} cleavage orientation just north of Slink

Pond, at the southwestern end of the large splay from the Park Pond Fault.

Several other large open asymmetric dextral kink-fold pairs which are apparently steeply plunging, are defined by the map units in several places in the Mic Mac Lake Group. These are just north, and 1-2 km south of the Burlington Road, and 1.2 km south of the Camp 166 Road in the Snoopy Pond Formation. As their short limbs are mostly steeply-dipping, and as they appear to reorient minor F_{2M} folds, it is likely that they post-date D_{2M}. However, as S_{2M} is somewhat variable in attitude, they might pre-date D_{2M}, or some or all could be large F_{2M} folds. Similar flexures in the lower part of the Snoopy Pond Formation east of Park Pond may be the same type of structure, or could be F_{2M} or F_{1B} fold pairs. A large more gentle and extensive dextral kink-fold pair is defined by the W.L.B. Slide and Kidney Pond Formation just south of Kidney Pond, and the Baie Verte Group-Mic Mac Lake Group contact south of Park Pond, and by a few measurements of S_{1B} between these two areas. Faults interpreted to cut the contact southwest of Park Pond are in the orientation of the kink-planes of this fold pair, but have the wrong displacement sense. Smaller kink folds that affect the western strip of the Mic Mac Lake Group in the same area also have the opposite sense to the larger flexure.

The dip of the S_{1B} cleavage, and bedding, define large asymmetric shallowly plunging gentle flexures in places in the Baie Verte Group. These form fold pairs with a less steeply-dipping short limb, and are seen in the section across Neale's Bay west of the Park Pond Fault, and just north and west of Flatwater Pond Provincial Park. The relationship of these flexures to D_{2B} and is not known, and they may be earlier or later than either event.

4D.15

(b) Minor crenulations of uncertain affinity

Micro-crenulations on S_{1B} cleavage surfaces that are not attributable to D_{2B} are fairly common in the area beyond 6 km north of the Burlington Road Junction. They do not appear to form sets with consistent orientations. Unattributable micro-crenulations are seen occasionally elsewhere.

(c) High strain-rate minor dextral folds

A few small open to close vertically plunging asymmetric dextral folds of S_{1B} are exposed in the two roadside quarries in the black slaty matrix conglomerate of the Kidney Pond Formation 1.4 km northeast of Kidney Pond. The larger open folds have a short limb not more than a metre long. The smaller and tighter fold pairs are rather irregularly formed, and have angular hinges that have broken the S_{1B} cleavage, indicative of formation at a relatively high strain-rate. These examples are not more than 80 metres from the W.L.B. Slide. Identical hand-specimen scale folds with broken foliation across their hinges are seen in one other outcrop. This is a roadcut of retrograded Birchy Schist, not more than 10 metres from the W.L.B. Slide 3.7 km north of the Burlington Road Junction. The dextral Marty's Pond Fault joins the W.L.B. Slide at Kidney Pond. It is suggested below that much of the displacement on the Marty's Pond Fault predated the development of the foliated virginite and the W.L.B. Slide, which formed late in the D_{1B} deformation. However, it is possible that these high-strain rate dextral folds in two localities next to the W.L.B. Slide are due to a small amount of dextral displacement (perhaps just one slip episode) on the Marty's Pond Fault and along the W.L.B. Slide north of Kidney Pond. A breccia of foliated virginite and quartz from veins, cemented by statically recrystallised virginite, is seen in one outcrop west of the Baie Verte Road 7.7 km north of the Burlington Road Junction. This, and the common static recrystallisation of the virginite, could be connected with minor dextral displacement along the

4D.16

W.L.B.Slide north of Kidney Pond.

(v) Faults

A conjugate set of small strike-slip faults, symmetrically-oriented with respect to the main compression direction defined by the normal to the S_{1B} cleavage, are present in the Baie Verte Lineament. Because their displacement is always small, they can only be detected where narrow marker horizons are well-exposed. This is the reason for their apparent abundance between Kidney Pond and Mic Mac Lake at the western side of the Lineament. Some have the opposite sense of displacement to a normal conjugate shear set. All probably cut the W.L.B. Slide, although many have such a small displacement that this is not detectable. They have not been extrapolated across the W.L.B. Slide on the map. One appears to have been generated from a similar pre-existing fault in the Fleur de Lys schists, which has more displacement in the latter than in the Baie Verte Group.

Similar small steep faults cutting the Mic Mac Lake Group oblique to strike are mostly oriented southwest-northeast, and appear to be controlled by a dominant fracture orientation in the Burlington Granodiorite. Most, as discussed previously, appear to possess some, probably dip-slip, displacement penecontemporaneous with deposition of the Mic Mac Lake Group. Displacement on some after D_{1B} deformation may have been strike-, dip-, or oblique-slip. The major strike fault in the area is the Park Pond Fault. The obvious displacement on it is a westward downthrow that is estimated to exceed 1000 metres. However, the large steeply-plunging dextral kink folds described previously, one of which is clearly associated with a splay of the Park Pond Fault, suggest that it may also possess some dextral strike-slip movement. The splays of this fault within the Mic Mac Lake Group between Park

Pond and the Camp 163 Road may have been controlled in part by faults penecontemporaneous with deposition of the Mic Mac Lake Group. The major splay that rejoins the Park Pond Fault 1 km north of Park Pond appears to downthrow eastward, defining a lenticular horst between it and the Park Pond Fault.

A thrust fault post-dating D_{1B} is marked (Plate 2) just west of the D_{1B} tectonic slide contact of the Baie Verte and Mic Mac Lake Groups northeast from Flatwater Pond. A marked lineament is used to interpret its presence, together with indications of incipient fault brecciation in an outcrop 1.7 km north of the Burlington Road.

(vi) Quartz-vein breccia

Abundant quartz-vein breccia post-dating D_{2M} is found in the Baie Verte Group north of the Burlington Road east of the Park Pond Fault. It cuts both mafic volcaniclastic rock, pink silicic tuff and the grey rhyolite. Minor amounts of quartz-vein breccia are found in the cleaved Burlington Granodiorite to the east. The reason for the existence and localisation of this quartz-vein breccia were not discovered.

(vii) Dolerite on the Park Pond Fault

Fine-grained non-cleaved spilitised dolerite containing fresh clinopyroxene forms an outcrop directly on the trace of the Park Pond Fault at the north end of Park Pond. Two similar outcrops are found adjacent to the fault, 0.8 and 1.4 km south of Park Pond. The contacts of this rock are not seen, but as they are not deformed, they are interpreted as a dyke intruded along the fault, rather than tectonic slivers. If they are a dyke, it may be of any age younger than the deformation of the Baie Verte Lineament.

CHAPTER 4E. MAJOR STRUCTURES OF THE BAIE VERTE LINEAMENT

(i) Major contacts

(a) Fleur de Lys schists — large ultramafic bodies.

The western contacts of the two large ultramafic bodies in the map area are treated separately from their eastern contacts because it appears that the former are slightly older than the latter.

These western contacts are tectonic; shear polyhedra to schistose serpentinite is developed adjacent to them in a zone usually not more than a few tens of metres wide. A relatively small width of 95%+ serpentinised ultramafic rock adjoins the western contacts compared with the eastern contacts of the two bodies. This is interpreted to mean that the Fleur de Lys schists were already dewatered and metamorphosed when the western tectonic contacts developed. The facts that the western contact of the northeast prong of the Mic Mac body crosscuts the garnet isograd and that the Flatwater body crosscuts a large F_2 antiform when garnet of syn- or pre-S_2 age is seen in the Fleur de Lys schists, support this inference. The western contacts of both bodies, although they are nearly coplanar with adjacent Fleur de Lys foliation in any one locality, on a large scale clearly crosscut the lithologic units and all three regional schistosities in the Fleur de Lys terrain. It is inferred that the Mic Mac body has a tectonic contact with the Wild Cove Pond Complex. The contact must crosscut the andalusite isograd due to this post-D_3 intrusive complex, and no contact metamorphic effects or apophysae from the complex are seen in the ultramafic body.

Although the contacts crosscut the regional Fleur de Lys foliations, local large-scale steeply-plunging gentle kink-fold pairs of post-D_3 age are found locally within the Fleur de Lys schists, and they are correlated with local changes in orientation of the contacts.. These local D_4 kink-folds

have been described previously, and are found in the area around Camp 26 Pond, 1 km north of

Middle Arm Brook, and north and south of Marty's Pond disrupted by the Marty's Pond Fault.

These folds appear to represent local accommodation of the more ductile Fleur de Lys schists to

'indentation' by the 'orange-pip'-shaped less ductile ultramafic bodies.

The eastern tectonic contact of the ultramafic bodies is thought to be younger than the

western contacts because two faults that cut the latter either are cut by or have a much

diminished throw across the former. These faults are the one crossing the southern end of the

Flatwater body, and the Marty's Pond Fault, respectively.

A complete section across the western contact of either ultramafic body is not exposed.

The two places where there is least gap between Fleur de Lys schists and ultramafic rock are on

Middle Arm Brook and 0.93 km due north of the Burlington Road Junction. In the first locality

there is a 40-metre gap between normal chlorite/biotite grade semipelitic Fleur de Lys schists and

schistose serpentinite. About 40 metres of schistose and fish-scale serpentinite adjoin a rapid

transition to undeformed serpentinised ultramafic rock to the east. The foliated serpentinite in

this locality is abnormal, in that it is mostly statically recrystallised and contains a relative

abundance of carbonate (~ 20%). The foliation seen in outcrop is mostly due to the weathering

out of lenticular carbonate veins, which formed subparallel to, and later than the one original

subvertical foliation. In places poorly-developed coarsely-spaced slip planes cut the foliation at a

low angle, occasionally accompanied by small impersistent crenulations. A few larger

asymmetric crenulation folds of the same type, with a wavelength up to a metre and verging up

toward a hypothetical antiform to the west, are seen in the outcrop to the south of the Brook.

These resemble F_3 folds in the Fleur de Lys, but no such F_3 folds are seen in the Fleur de Lys

outcrop to the west. There is no reason to assume, given the large-scale evidence, that these minor crenulations are connected with anything other than local late movement on the tectonic contact. They are not seen anywhere else in foliated serpentinite on the western contact. The other well-exposed locality shows platy fish-scale to schistose serpentinite with one vertical foliation 8 metres from normal chlorite/biotite grade interbanded mafic and sernipelitic schist. A few metres of schistose/fish-scale serpentinite grades rapidly eastward to a few metres of shear polyhedra serpentinite, and in turn to massive serpentinised ultramaf Ic rock. This exposure is typical, although better-exposed, of all other localities near the western contact of the ultramafic bodies. In this locality, the semipelitic and mafic schists, show no structural or metamorphic retrogression effects in outcrop or thin section attributable to the tectonic contact, which is probably not more than 5 metres away from the nearest exposed rock. In other places along the western contact of either body, the shear polyhedra zone is usually as narrow as in this locality, but south of Red Cliff Pond it is up to 100 metres wide, although the polyhedra are mostly large and with little interstitial serpentinite. North of Marty's Pond, the zone of white-weathering 95%+ serpentinised rock is exceptionally narrow, and the tectonic contact zone must also be very narrow, as it is essentially not exposed.

Thus, the evidence is all consistent with the development of the western tectonic contact of the large ultramafic bodies after the Fleur de Lys schists had been deformed, metamorphosed and intruded by the post-D_3 Wild Cove Pond Granite/Diorite Complex. The schistose/shear polyhedra serpentinite developed on the contact is compatable with either or both of shearing and flattening deformation. The large kink-folds in the Fleur

de Lys schists are indicative of strong flattening and moulding of the more ductile schists against the less ductile ultramafic bodies. These probably formed after most or all of the displacement on the contact had been accomplished. The eastern contact of the ultramafic bodies cuts faults which cut the western contact, and therefore formed slightly later. This eastern tectonic contact, described next, is the same structure as that separating Fleur de Lys schists from Baie Verte Group sediments and volcanics where the large ultramafic bodies are absent. Therefore, although the large ultramafic bodies belong stratigraphically to the Baie Verte Group, they can be regarded, after formation of their western contact,as structurally part of the western Fleur de Lys block in relation to the volcanics and sediments of the Baie Verte Lineament.

(b) Fleur de Lys schists and large ultramafic bodies — Baie Verte Group; Western Lineament Boundary Slide.

The linear contact between the Fleur de Lys schists, and volcanics and sediments of the Baie Verte Group, where the large ultramafic bodies are absent, is a major and abrupt subvertical structural and metamorphic discontinuity. The Baie Verte Group volcanics and sediments, with one cleavage coplanar with bedding and the contact, adjoin Fleur de Lys schists with three regional foliations which on a large scale are truncated by the contact, as is the garnet isograd within them. On this basis, the contact is either a major tectonic contact or an unconformity. No evidence supports the presence of an unconformity, and all evidence seen indicates that it is a major tectonic junction.

In general, where the large ultramafic bodies are absent, no obvious structural or metamorphic changes are seen within either Fleur de Lys schists or Bale Verte Group sediments and volcanics close to the narrow

contact zone. Some minor retrograde metamorphic effects seen in Fleur de Lys schists mostly within a few tens of metres of the contact have been described previously, and are probably attributable to the development of the contact. It is possible that Baie Verte Group mafic volcaniclastic sediments are more highly deformed than elsewhere in the zone west of the Kidney Pond Formation between 2.4 km south of the Old Camp 32 Road and 0.4 km north of the Bear Cove Road. Elsewhere, there is no obvious change in deformation of Baie Verte Group rocks adjacent to the contact. In the south, there is an abrupt change in metamorphic grade across the contact. North of the Burlington Road, the Fleur de Lys schists adjacent to the contact are of lower (chlorite) grade, containing a medium-green actinolite. The adjacent Baie Verte rocks are of the same grade as in the south, containing a weaklyicoloured to colourless actinolite, and there is therefore still a difference, although less marked, in the metamorphic assemblages across the contact. However, the major differences that apply all along the contact are the grain size and degree of recrystallisation, and the structural histories, of the rocks on either side. The Fleur de Lys schists always contain at least two marked penetrative schistosities, and are totally and coarsely recrystallised. The Baie Verte rocks contain a single cleavage, and are only partially recrystallised with a much finer grain size. Therefore, in the northern part of the area there is still a marked and abrupt structural and metamorphic change across the contact, and there is nowhere any difficulty in distinguishing Fleur de Lys mafic schists from Baie Verte mafic volcanics and volcaniclastic sediments.

Long lenticular bodies of virginite (magnesite-quartz-fuchsite rock, described previously) up to 80 metres wide, but not usually more than 40 metres wide, are found along parts of the contact, where the large ultra-

-mafic bodies are absent, from 2.6 km south of Kidney Pond north to the Old Camp 32 Road (Plate 1) and from 5.7 to 8.3 km north of the Burlington Road Junction (Plate 2). This distinctive rock-type is always found exactly on the contact between Fleur de Lys schists and Baie Verte Group volcanics and sediments. It is also found, in the same long lenses, on the eastern margin of the Flatwater ultramafic body from south of the Bear Cove Road to just north of Middle Arm Brook. In the area just north and south of Middle Arm Brook there is a narrow strip of the ultramafic body to the east of the virginite. The particular elements, and their proportions, in the virginite indicate that it is derived mostly or entirely from ultramafic rock, but it is not due to *in situ* alteration of serpentinite, because sharply-defined phacoids of serpentinite are occasionally seen within it. The way it was formed is not known, but the large CO_2 content may be explained by channeling of the CO_2, derived from microinclusions in a large volume of serpentinising ultramafic rock, along a major tectonic discontinuity. The localisation of the virginite is fairly clear evidence that the contact that is now seen between Fleur de Lys schists and Baie Verte Group sediments and volcanics continues along the eastern margin of the Flatwater ultramafic body, and was developed later than the western contact of that body for the reason stated in the previous section. The contact is also interpreted to continue along the eastern contact of the Mic Mac ultramafic body because of an outcrop of virginite mapped by E.R.W. Neale (unpublished manuscript map) on the eastern side of Jacob's Lake.

Roadcuts through the virginite within the ultramafic body just south of Middle Arm Brook show that it is bordered by a zone of foliated carbonate-serpentine-talc rock about a metre wide, which is in turn adjoined by a narrow zone of schistose/fish-scale serpentinite, which passes outward through shear polyhedra serpentinite to massive ultramafic rock. This

demonstrates that the virginite marks a major tectonic discontinuity. The virginite terminates just to the north of Middle Arm Brook, and the tectonic discontinuity continues northward within the ultramafic body as a poorly-exposed zone of foliated carbonate-serpentine-talc rock, bordered by shear polyhedra serpentinite. Such obvious effects of a tectonic discontinuity are not seen where Fleur de Lys schists directly adjoin Baie Verte Group volcanics and sediments. Two outcrops of Baie Verte Group rocks within a few metres of the virginite, and one outcrop of Birchy Schist in a similar position show local carbonate metasomatism. Otherwise, rocks exposed as close as 3 metres to the virginite do not show any obvious structural or compositional modification.

The virginite possesses a single foliation, described previously, which is subvertical and coplanar with the contact and the adjacent Baie Verte cleavage (S_{1B}). On a small scale, it is also coplanar with local adjacent Fleur de Lys schistosities, but on a large scale this western boundary of the Baie Verte Group volcanics and sediments, and the virginite lenses on it, crosscut the orientation of all three regional schistosities in the Fleur de Lys schists. The foliation in the virginite is therefore interpreted as the equivalent of the Baie Verte S_{1B} cleavage. The contact zone containing the virginite lenses is interpreted as a D_{1B} tectonic slide, here named the Western Lineament Boundary Slide. The foliation in the virginite probably represents the last part of the significant deformation it underwent, because the carbonate-rich rock is relatively susceptible to recrystallisation, shown by the statically-recrystallised fabric that is commonly seen. There must have been a significant amount of thrust movement of the ultramafic rocks on the western side of the W.L.B. Slide over the Baie Verte Group volcanics and sediments, because it can be shown in

the Mings Bight area that the identical ultramafic rocks form the lower part of an ophiolite complex base to the volcanics and sediments. However, the foliation in the virginite is probably a flattening fabric formed at the same time as the Baie Verte S_{1B} cleavage, after most or all of the translational movement and simple shear deformation on the contact had taken place, because the virginite is probably relatively easily recrystallised. A large (3m) angular erratic block of virginite adjacent to the power line 0.4 km south of Middle Arm Brook contains a strong lineation of magnesite and altered chromite grains in the foliation. Such a strong lineation was not seen in outcrop, although a weak subvertical preferred elongation lineation is seen in a few outcrops. This lineation is interpreted to be L_{1Ba}. The one foliation in the virginite is cut and folded by S_{2B} and F_{2B} in the outcrops up to 1.7 km south of the Old Camp 32 Road, and in this same segment S_{2B} and minor F_{2B} folds are also seen as local F_4 and S_4 structures in the Birchy Schist adjacent to the contact (best shown in [12]). Minor D_{2B} structures are also seen cutting the foliation in the virginite outcrops north and south of Kidney Pond. The one foliation in the virginite therefore formed prior to D_{2B}, and in view of the large-scale relationships along the contact, it is unlikely that the foliation is due to anything other than D_{1B}. Later structures in the virginite are confined to abundant and ubiquitous quartz veins, mostly from a millimetre to a few centimetres wide. The narrower veins form a conjugate shear set with respect to the foliation, and the larger veins an ac tension fracture set. Three large quartz veins each several metres wide were also found. The one example of a breccia of foliated virginite, described previously, is probably due to the minor dextral strike-slip motion inferred on the segment north from Kidney Pond, connected with the last movement on the Marty's Pond Fault. Sub-horizontal slickensides on some outcrops of virginite, especially around

6.2 km north of the Burlington Road Junction, are also probably due to the same minor adjustments. However, there is no evidence that this inferred late minor dextral strike-slip movement has modified the position or attitude of the Western Lineament Boundary Slide. The virginite between Kidney Pond and Slink Pond shows that it was formed prior to this late movement and is essentially not affected by it. There is also no evidence of more substantial strike-slip motion on the Western Lineament Boundary Slide, and this narrow zone does not compare with the relatively wide zones of brecciation seen along major strike-slip faults, as for example in the Sops Arm area to the west of White Bay.

Virginite is shown as present along approximately 40% of the mapped length of the W.L.B. Slide. It may or may not be present elsewhere, but it is replaced by a zone of foliated carbonate-serpentine-talc rock north of Middle Arm Brook within the ultramafic body, and talc-carbonate schist with a single foliation is seen on the W.L.B. Slide at the northern limit of mapping. On the woods road 3.1 km north of the Burlington Road Junction, poorly-exposed strongly-cleaved mafic volcaniclastic and argillaceous sediments are seen 10 metres from normal Birchy Schist, and although there is no evidence that ultramafic-derived material is present on the contact, it is likely that a thin smear of such material is present and unexposed. Roadcuts just to the north of this locality expose very strongly deformed pillow lava overlain by black slate. The next roadcut to the north of the pillow lava, about 10 metres across strike, consists of chlorite-grade mafic and semipelitic Fleur de Lys schists. The easternmost metre of the schist is severely retrograded and discoloured, and somewhat brecciated, and the outcrop contains the high strain-rate minor dextral folds described previously. These two localities are where the smallest gap is seen between Baie Verte sediments and volcanics and Fleur de Lys

schists. As the Baie Verte S_{1B} cleavage is coplanar with the contact, they are consistent with the hypothesis that it is primarily a D_{1B} slide.

The D_{1B} Western Lineament Boundary Slide crosscuts on a large scale all three regional foliations and folds in the Fleur de Lys schists. It is interpreted to have formed relatively slightly later than the western tectonic contacts of the two large ultramafic bodies, which also crosscut Fleur de Lys D_3 structures. The disrupted D_4 kink fold associated with the Mic Mac body appears to have deformed the post-D_3 Wild Cove Pond Complex by a number of small southeast-northwest dextral faults, and the western tectonic contact of the Mic Mac ultramafic body is interpreted to cut the Complex, its contact with Fleur de Lys schists, and its andalusite isograd. Therefore, both the western tectonic contacts of the ultramafic bodies, and the D_{1B} Western Lineament Boundary Slide, and the Baie Verte D_{1B} deformation all post-date not only all the D_{1-3} regional Fleur de Lys deformations, but also the post-D_3 intrusion of the Wild Cove Pond Granite-Diorite Complex and the Celebes Pond Granite. Although the Marty's Pond Fault joins the W.L.B. Slide at Kidney Pond, the strike-slip displacement transmitted onto the latter is probably not more than 1 km and may be negligible. If this displacement is removed, the W.L.B. Slide still truncates the major D_3 fold pair in the Fleur de Lys, and the andalusite isograd due to the Celebes Pond Granite.

(c) Baie Verte Group — Mic Mac Lake Group stratigraphic contact.

Between the Park Pond Fault and the Baie Verte Group, the narrow discontinuous strip belonging to the Mic Mac Lake Group has been shown to resemble, in several lithological aspects, the lower part of the Armageddon Formation at the base of the main sequence of the Mic Mac Lake Group.

Fig. 4E.1. Sketches of stratigraphic contacts between Baie Verte and Mic Mac Lake Groups.

[descriptions and locations given in text]
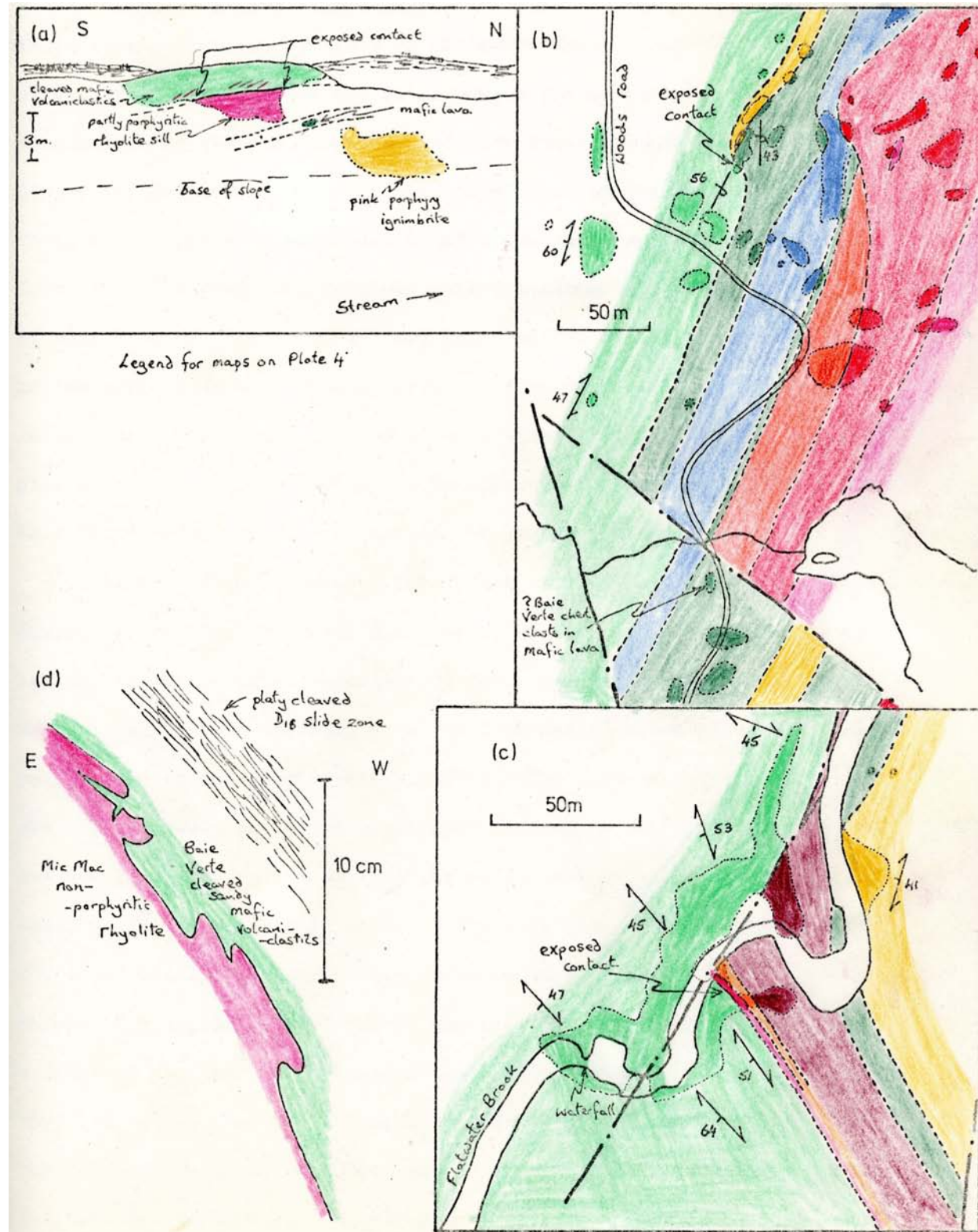
Facing evidence in this strip, with one possible exception, suggests that the beds face eastward and are overturned. The stratigraphy is very constant along the strip, with a pink porphyritic ignimbrite east of a mafic lava unit from the falls on Flatwater Brook to Park Pond. A second pink ignimbrite underlies the mafic lava north of Park Pond, and this is presumed to be the one seen south of Park Pond. In contrast, the stratigraphy in the Baie Verte Group, and in particular the White Bay Waters Formation, appears to be truncated at a low angle against the contact. The contact is fully-exposed in three outcrops along this strip, 0.8 km south of Park Pond, 1.2 km north of Park Pond, and on the sides of Flatwater Brook in the gorge just below the waterfall 1.4 km north of the Camp 166 Road. In all three, the contact appears to be stratigraphic, and therefore this part of the Mic Mac Lake Group is interpreted to be unconformable on the Baie Verte Group, an interpretation not previously suggested.

A sketch of the outcrop 1.2 km north of Park Pond is given in Fig. 4E.1a. At the foot is about 3 metres of weakly cleaved pink porphyritic ignimbrite, structurally overlain by about a metre of vesicular purple mafic lava, in turn overlain by about 3 metres of a non-cleaved patchily porphyritic to non-porphyritic rhyolite. The contacts of the mafic lava are not exposed. The structural upper contact of the rhyolite is sharp and not tectonic, and is overlain by well-cleaved cream-weathering green sandy mafic volcaniclastic rocks of the Baie Verte Group. Weakly-cleaved and occasionally brittly-boudinaged veins up to 5 cm wide of the patchily porphyritic rhyolite are found in the mafic volcaniclastic rock within a metre of the contact. Although they are not seen to join the main rhyolite below, they are almost certainly apophyses of the rhyolite below the contact, which is therefore probably a sill. The rhyolite veins are not tectonic inclusions. A narrow (~1 cm) protruberance about 10 cm long

of the Baie Verte rock is seen extending from the contact into the rhyolite at a shallow angle, consistent with an intrusive contact. The rhyolite is flinty right to the contact and in the veins, and it has a micropoikilitic 'snowflake' devitrification texture, all more consistent with it being a sill rather than a flow. This contact is therefore definitely stratigraphic, but the facing direction, and conformable as opposed to unconformable nature, have to be found elsewhere and in larger-scale evidence.

Another exposed contact is found 0.8 km south of Park Pond 40 metres north of the woods road, shown in a sketch map in Fig. 4E.1b. To the north, about 1.5 metres of the pink porphyritic ignimbrite is stratigraphically overlain by about a metre of sandstone, and then by a purplish mafic lava. Towards the woods road, the ignimbrite and sandstone are cut out, and the vesicular base of the Mic Mac mafic lava is seen in contact with about 30 cm exposed thickness of a weakly cleaved homogenous cream-weathering Baie Verte Group rock that is probably a lava. The contact shows no sign of tectonic modification. There is a thin (mm) cream-yellow streak along it, due to ultra-fine-grained calc-silicate mineral alteration of the underlying Baie Verte rock. The disappearance of the pink ignimbrite and sandstone is interpreted as partly due to non-deposition against a small ridge in the unconformable Baie Verte Group surface (Fig. 4E.1b), and partly to pene-contemporaneous erosion.

To the south of this exposure, the lowermost (westernmost) conglomerate has a chloritic matrix, in contrast to the quartzo-feldspathic sandy to granular matrix of all other conglomerates in the Mic Mac Lake Group. It is suggested that this chloritic matrix is derived from the unconformably underlying Baie Verte Group. None of the clasts in this conglomerate, which are mostly cobbles of Granodiorite and non-porphyritic rhyolite, were definitely identified as derived from the Baie Verte Group. An outcrop

of Mic Mac mafic lava, on the western side of the woods road just south of the small stream that has washed out the road, contains a few angular fragments of white-weathering green chert, which are suggested to have been derived from the Baie Verte Group.

On the north side of the gorge below the waterfall on Flatwater Brook, 1.4 km north of the Camp 166 Road, the contact between Baie Verte and Mic Mac Lake Groups is faulted. A classic granulated fault gouge zone about 30 cm wide is seen between well-cleaved mafic volcaniclastics and a crimson flinty quartz-feldspar porphyry sill. On the south side of the gorge (Fig. 4E.1c), there is about 3 metres of phacoidally-cleaved, intensely-haematised pebble conglomerate to the west of the crimson porphyry sill, and then about 2 metres of a steely maroon flinty aphanitic rhyolite, interpreted to be a sill. This rhyolite has an exposed contact with cleaved to strongly-cleaved cream-weathering green sandy/silty mafic volcaniclastics of the Baie Verte Group. There is a cream streak 2-3 mm thick, composed of ultra-fine-grained calc-silicate minerals, on the contact. Less than a centimetre away from the contact, there is a zone about 10 cm wide in the Baie Verte rocks of intense platy cleavage, which is interpreted as a D_{1B} tectonic slide that may have significant displacement across it. However, the slide leaves a thin skin of Baie Verte Group rock in autochthonous contact with the rhyolite. On the south side of the small ridge beside the stream, the contact of the aphanitic rhyolite and Baie Verte is irregular in detail, with small rounded 'flow-cast'-like protrusions of the rhyolite into the thin skin of mafic volcaniclastics to the east of the platy cleaved slide zone. They are not minor folds because they 'verge' both ways along a short length of the broadly planar surface (Fig. 4E.1d). Besides the micropoikilitic devitrification texture of the rhyolite, also seen in the crimson porphyry

sill to the east, the flinty nature of the rhyolite across the whole of its small thickness, and the small irregularities on its contact, are thought to be evidence that it is a sill. The pebble conglomerate to the east contains a few green clasts, but these are not positively identifiable, even in thin section, as derived from the Baie Verte Group. A few very fine-grained white siliceous pebbles might be Baie Verte chert, or devitrified silicic volcanics of Mic Mac Lake Group affinities. This exposure therefore also shows a stratigraphic contact. The apparently small slide zone next to the contact represents a transition towards the situation in the area north of Flatwater Pond where the contact is a major tectonic slide zone, described below.

Baie Verte rocks within a few metres of the contact contain rare narrow (~1 mm) veins of red haematitic material, in all cases post-dating the S_{1B} cleavage, in three localities, on the south side of the gorge on Flatwater Brook, and 1.6 km north and 2.3 km south of Park Pond. In the first locality the veins locally form a small zone of net breccia. In the third locality, one vein is about 5 mm wide, and is seen in thin section to consist of microcrystalline quartz and haematite. These are not sedimentary dykes because they post-date the S_{1B} cleavage. It is interpreted that they formed by deposition from circulating groundwater after the rocks reached their present attitude.

Therefore there is a stratigraphic contact between the Baie Verte Group and the Mic Mac Lake Group west of the Park Pond Fault, from the gorge on Flatwater Brook southward to 2.2km south of Park Pond, and probably as far as opposite the south end of White Bay Waters. The large-scale relationships show that this contact is a very slightly angular unconformity, and although the contact is not stratigraphic north of Flatwater Pond,

the greater thickness of Neale's Bay Formation preserved there shows that the unconformity must have been slightly angular.

(d) Baie Verte Group — Mic Mac Lake Group/Burlington Granodiorite tectonic contact; Eastern Lineament Boundary Slide.

In the area north of the southern part of Flatwater Pond, the contact between the Baie Verte Group and the Mic Mac Lake Group is in basic terms a D_{1B} tectonic slide, named here the Eastern Lineament Boundary Slide. On the central eastern shore of Flatwater Pond, and north of a point 0.6 km north of the Burlington Road, the E.L.B. Slide occurs directly between Baie Verte Group and a thin sequence of the Mic Mac Lake Group. In this area east and northeast of Flatwater Pond, the Mic Mac Lake Group rocks are exposed in many places in stratigraphically unconformable contact with the Burlington Granodiorite, and the unconformity is nowhere known to be the site of a tectonic slide or strike-fault. Between the northeast shore of Flatwater Pond and just north of the Burlington Road, the E.L.B. Slide separates Baie Verte Group from a cleaved to strongly cleaved tectonic slice of Burlington Granodiorite, that in turn is in D_{1B} tectonic slide contact with Mic Mac Lake Group. In the outcrop along the northeast shore of Flatwater Pond, the situation is more complex. D_{1B} tectonic slides result in the repetition of the sequence Burlington Granodiorite-Mic Mac Lake Group at least three times, and two small cross-faults repeat the sequence a fourth time. As the Mic Mac Lake Group preserved here consists almost wholly of sandstone, which resembles the remobilised Granodiorite when both are strongly cleaved, it is initially difficult to perceive the situation on this shore section. North of the Burlington Road, a third D_{1B} tectonic slide appears within the widening slice of cleaved Granodiorite and isolates two narrow lenticular areas of west-facing Mic Mac Lake Group on its eastern side. These are still in autochthonous unconformable

Fig. 4E.2. Tectonically remobilised Burlington Granodiorite.

contact with the cleaved Granodiorite to the east. All the D_{1B} tectonic slides in this area, including the E.L.B. Slide, dip moderately westward and have eastward overthrust displacement.

The cleaved, tectonically remobilised Granodiorite in this area can be clearly identified because abundant aplite veins and occasional strongly-cleaved mafic dykes can be traced for significant distances. It does not resemble the local Mic Mac sandstones because it is unbedded, and more heterogeneous, with coarse remnant grains of quartz in the retrograded and deformed fine-grained chlorite-sericite-albite matrix derived from the feldspar and mafic minerals. The sandstones are homogeneous fine-grained, thinly bedded, and relatively quartz-rich. Much of the cleaved Granodiorite is not strongly deformed, and the plutonic texture can still be recognised. However, in places, it becomes highly-deformed, and a characteristic "migmatitic"-looking fabric develops, consisting of long (~1 metre) narrow (a few centimetres) 'long augen' of little deformed pink granodiorite contained in a strongly cleaved chloritic granulated matrix of similar width (Fig 4E.2). The formation of this structure is not understood. Occasionally, an incipient 'gneissic' segregation banding, formed by parallel lenticular areas rich in chlorite up to a centimetre wide, is seen in the more homogenous cleaved granodiorite.

The thin strip of Mic Mac Lake Group preserved in this area was traced 2.3 km northward from the Burlington Road. Outcrop is poor north of this limit. However, it is clear that the Mic Mac Lake Group must eventually be cut out by the Eastern Lineament Boundary Slide, because it is not present on or north of the La Scie Road, or in a drill hole that passed through the contact somewhere near the La Scie Road. This drill core showed (E.R.W. Neale, pers. comm.) a narrow zone of platy cleaved material derived from the Granodiorite on the contact with Baie Verte Group, and it

is presumed (the core has been lost) that this was part of the E.L.B. Slide.

The E.L.B. Slide is the major surface of tectonic movement at the eastern side of the Baie Verte Lineament. The D_{1B} slides between remobilised slices of Granodiorite and the Mic Mac Lake Group are subsidiary, sympathetic structures. The E.L.B. Slide juxtaposes east-facing Baie Verte Group against west-facing Mic Mac Lake Group in this area on the eastern side and north of Flatwater Pond, and has therefore removed the core of a major synclinal structure. To the south of Flatwater Pond, a major D_{1B} slide zone has been identified, preserved in two segments adjacent to the eastern side of the Park Pond Fault. Allowing for westward downthrow on the Park Pond Fault, this slide zone also occurs between a westward-facing and an eastward-facing sequence, in this case both of Mic Mac Lake Group. It is therefore interpreted as the southern extension of the Eastern Lineament Boundary Slide. It has been removed and concealed by the Park Pond Fault south of Park Pond, and between the two exposed segments.

(ii) Overall structure of the Baie Verte Lineament.

The overall structure of the Baie Verte Lineament in the map area is a highly modified syncline (Plate 7). This is also true for the Mings Bight area, described below. In the map area, the Baie Verte Group, and the thin remnant of eastward-facing Mic Mac Lake Group unconformable on it, form the eastward-facing, homoclinal, vertical to overturned, western limb of this structure. The main part of the Mic Mac Lake Group, unconformable on the Burlington Granodiorite, forms the westward-facing, homoclinal, moderately-dipping, eastern limb of the structure. This syncline has been modified by major westward downthrow on the Park Pond Fault, and previously by major eastward thrusting during the D_{1B} deformation on the Eastern Lineament Boundary Slide. The latter probably formed on the axial-surface of

of the major synclinal structure, because it separates east-facing from west-facing rocks north from Flatwater Pond, and probably did so, before the Park Pond Fault developed, in the region south of Flatwater Pond.

The interpretation of the development of the D_{1B} structure at the western side of the Lineament depends on the recognition that the sediments and intercalated volcanics of the Baie Verte Group were deposited on an ophiolite complex (oceanic crust and mantle). Direct evidence of this is exposed in the Mings Bight area, described below. For the purposes of the structure, it does not matter whether the ophiolite complex and Baie Verte sediments and volcanics are broadly *in situ*, or allochthonous with respect to the eastern and western Fleur de Lys terrains. However, the evidence summarised in the final chapter all indicates *in situ* development of the ophiolite complex. The large ultramafic bodies (and local overlying transition zone and gabbro) west of the Western Lineament Boundary Slide are therefore the dismembered relics of the upper oceanic mantle (and crust) that originally underlay the sediments and intercalated volcanics of Baie Verte Group. Their juxtaposition against those sediments across the W.L.B. Slide shows that, together with the western Fleur de Lys block, they have moved upward relative to the sediments on that major D_{1B} slide zone. Although there was probably also relative movement of the western Fleur de Lys block upward on the western tectonic contacts of the large ultramafic bodies, this would have to have happened before development of the W.L.B. Slide. The moulding of the Fleur de Lys schists to the shape of the western contacts of the ultramafic bodies suggests strong flattening across the contact after any such relative movement that there may have been. Therefore, the major D_{1B} structures of the Baie Verte Lineament primarily reflect major horizontal shortening. These structures are the major tight synclinal. fold, the shortening across the moderately to steeply inclined

cleavage, the eastward thrusting on the Eastern Lineament Boundary Slide, and the flattening of the Fleur de Lys schists around the western contacts of the large ultramafic bodies. The inferred subvertical upward displacements of western relative to eastern blocks at the western side of the Lineament are consistent with the overall horizontal shortening together with a general sense of overthrusting from west to east. Structures at the western side of the Lineament are subvertical, and change to moderately west-dipping at the eastern side, and therefore the relative movement of more westerly to more easterly structural units was mostly upwards at the western side of the Lineament, while it had more of an overthrust component at the eastern side. The preferred elongation direction (L_{1Ba}) in the S_{1B} cleavage also reflects this upward and eastward direction of transport of material. The deformation and the structures associated with it are almost wholly localised within the Baie Verte Lineament. Although there are local effects in the Fleur de Lys schists and Burlington Granodiorite immediately adjacent to the Lineament, the large horizontal shortening and upward and eastward overthrusting essentially do not affect the western and eastern Fleur de Lys terrains. The deformation can therefore be viewed in terms of the relative approach of these two bordering blocks as rigid bodies, deforming the Baie Verte Group and Mic Mac Lake Group between them, and with the western Fleur de Lys block attempting to override the eastern Fleur de Lys/Burlington Granodiorite block.

The later deformations on the eastern side of the Lineament indicate precisely the opposite kinematics to the D_{1B} deformation. D_{2M} folds reflect relative westward motion of material back towards the axis of the Lineament and the major westward downthrow of the Park Pond Fault is also opposite to the sense of relative movement during D_{1B} deformation. These two structures are thought to indicate stress relaxation soon after the major horizontal

shortening during D_{1B}. The significance of the minor D_{2B} structures found at the western side of the Lineament is not known, although they represent minor and local horizontal shortening along an approximately north-south axis.

CHAPTER 4F. CHEMISTRY OF VOLCANIC ROCKS OF THE BAIE VERTE LINEAMENT.

No systematic study of the chemistry of the various groups of mafic and silicic volcanic rocks in the Baie Verte and Mic Mac Lake Groups was made, for the following reasons. First, the number and distribution of the groups and subgroups of the volcanic rocks, the location of least-deformed examples and the thickest and best-exposed sections were unknown prior to this mapping. The effort may be largely wasted if samples are taken and analysed in ignorance of these data. Secondly, it has been demonstrated that analyses of the major elements in weakly metamorphosed mafic volcanics lead at best to uncertain conclusions about their original composition, and in most cases probably reflect no more than the present composition of the particular samples chosen (Smith, 1968). Noble (1965) demonstrated that ignimbrites are highly likely to suffer selective leaching, especially of Na, by penecontemporaneous weathering and groundwater circulation. In addition, it is clear that the Mic Mac Lake Group ignimbrites and rhyolites suffered severe oxidation, and therefore analyses are not likely to be very informative on the original composition of their magma.

Pearce and Cann (1973) developed a method for discrimination of basalts of the four types: oceanic crust, low-K arc tholeiite, calc-alkaline arc, and hot spot, based on the trace elements Ti, Zr, Y, Sr, Nb. Euan Nisbet very kindly analysed 8 selected powdered samples from the Baie Verte Lineament for these elements. The samples selected (Table 4F.1) are two of ophiolite diabase dykes, one from gabbro (1), and one from sheeted dyke complex (2) in the inland area, one ophiolite pillow from Mings Bight (3), two of light green pillow lava from the Slink Pond Formation (4, 5), one of dark green pillow lava from the little Moose Cove Member (6), one of mafic lava from the Mic Mac Lake Group (7), and one of a vesicular Mic Mac

Table 4F.1. Partial trace element analyses

Analyses by E. Nisbet

Sample	Ti	Rb	Sr	Y	Zr .	Nb	(ppm.)
1	1850	0.5(?)	181.9	11.1	13.1	1.7	Ophiolite dyke
2	1900	4.2	154.4	11.7	14.3	1.8	Ophiolite dyke
3	2100	4.3	97.5	9.8	18.2	2.1	Ophiolite pillow
4	8300	1.0	186.6	26.8	108.5	5.0	Slink Pond Fm. lt. green pillow
5	8750	2.8	190.1	25.5	109.3	5.2	Slink Pond Fm. lt. green pillow
6	16000	1.4	196.1	35.3	224.1	12.3	Slink Pond Fm. dk. green pillow
7	8750	72.2	363.9	26.3	192.2	14.1	Mic Mac Lake Group lava
8	5750	46.3	402.9	21.4	138.4	14.8	Mic Mac Lake Group dyke

mafic dyke (8) in the Burlington Granodiorite. Although there are a small number of samples, there is a clear grouping of and discrimination between ophiolite diabase/lava, light green Slink Pond Fm. pillow lava, and Mic Mac Lake Group mafic lava/dykes. However, the ophiolite lavas fall into the low-K tholeiite arc basalt field on all Pearce and Cann's discrimination diagrams, while the light green Slink Pond Fm. lavas, known to be interbedded with sediments, fall into their oceanic crust tholeiitic basalt fields. The dark green pillow lava in the Slink Pond Formation falls into the hot spot field. The Mic Mac Lake Group mafic lava falls into the calc-alkaline arc basalt field, compatable with their association with ignimbrites. If these groupings are maintained when further samples are obtained, these elements may be used to discriminate tectonically isolated lavas of uncertain affinity in the Baie Verte Lineament. However, the results suggest that Pearce and Cann's criteria for discrimination of oceanic, arc, and hot spot tholeiites are not universally applicable, at least in weakly metamorphosed rocks. Smith (1968) showed that Ti varies from 0.6 to 1.3% in just the 'unaltered'-looking parts of a single outcrop of a weakly metamorphosed mafic lava flow, and thus the method used by Pierce and Cann may not be applicable to such rocks.

CHAPTER 5. BAIE VERTE GROUP IN THE MINGS BIGHT AREA

(i) Introduction

Three weeks of the 1971 field season was spent mapping in detail the coastal exposures

of the Mings Bight peninsula jointly with J. M. Bird and J. F. Dewey. The data collected are

shown on Plate 5. The geology of the inland part of the peninsula was compiled by using the

detailed coastal data and air photo interpretation to modify the maps of Neale (1958, 1959b) and

Watson (1947). The inland geology is therefore a very preliminary compilation, although there is

good control from both coastal sections on either side of the northern part of the peninsula. This

map is generalised and incorporated into Plate 6. A short summary of the geology follows.

(ii) Overall structure and stratigraphy

The rocks in the Mings Bight area are disposed in three major thrust sheets, dipping

moderately northwest, and consisting in large part of various units of the Mings Bight Ophiolite

Complex. The lower thrust sheet is upright, consisting of non-cumulate harzburgite and minor

dunite to the south and east of Mings Bight, overlain to the west by a rather disrupted transition

zone, and gabbro. The other two major thrust sheets are inverted. Although they are modified by

internal faults, the inverted sequence of ophiolite complex stratigraphy can be seen on the map

(Plate 5). The sequence in the two sheets runs from non-cumulate harzburgite and minor dunite

(in the large body underlying Baie Verte; and at Red Point) through layered cumulate ultramafic

'transition zone' (not seen in the upper sheet; Western Point, south of Red Point), gabbro and

gabbro with a few coplanar diabase dykes (Point Rousse; Western Point, south of Red Point),

gabbro with dykes, increasing to 100% sheeted diabase dyke complex (south of Point Rousse;

north of Big Head), then diabase dykes cutting pillow lava (not

seen in the upper sheet; Upper Green Cove), to pillow lava (Eastern Point; Big Head, Upper Green Cove). The diabase dykes, gabbros and ultramafic rocks are not cleaved or penetratively deformed, except very locally near thrust faults and D_{1B} slide zones. Pillow lavas on Big Head are hardly deformed and those in Upper Green Cove are weakly deformed, but the section south of Eastern Point is strongly-cleaved and highly-deformed.

Just south of Big Head, a 30 cm bed of slump-folded maroon chert has an overturned but stratigraphic contact with the highly variolitic ophiolite complex pillow lava, and this contact is the top of the Mings Bight Ophiolite Complex. An overturned section about 300 metres thick of mostly very weakly cleaved mafic volcaniclastic sediments stratigraphically overlies the chert. It, in turn, is stratigraphically overlain by an overturned sequence about 1000 metres thick consisting mostly of very strongly deformed and highly cleaved variolitic pillow lava. This is interpreted to be in thrust contact with the lower upright thrust slice, although the contact is modified by a late high-angle fault in the coastal section.

(iii) Mings Bight Ophiolite Complex

The ultramafic rocks are mostly non-cumulate harzburgites identical to those forming the large ultramafic bodies in the inland map area. Minor thin dunite bands are present. The contact between these and cumulate ultramafic rocks has not been seen structurally intact in the coastal sections. Transition zone layered cumulate ultramafic rocks are wholly serpentinised and consist mostly of harzburgite with sieve-poikilitic orthopyroxene, and dunite, with minor clinopyroxenite and gabbro layers. Very coarse-grained clinopyroxenite pegmatite layers and veins are present. Large (10 m) autoliths of non-cumulate(?) harzburgite are found in the layered transition zone. The gabbro unit consists mostly of rather feldspathic

gabbro, with minor clinopyroxenite, anorthosite, and mafic gabbro bands in the banded cumulate parts. A weak gneissic fabric coplanar with the cumulate banding is common in both banded and non-banded gabbro. Gabbro pegmatite veins are very common, and almost all post-date the gneissic fabric. Parallel diabase dykes, commonly 20-50 cm wide, cut the gabbro, including the cumulate gabbro, and always post-date both gneissic fabric and the pegmatite veins. This sequence of events is applicable to most of the gabbro unit although, locally, more complex relationships between crystallisation and the genissic fabric are seen. For example, angular autoliths of gabbro with a gneissic fabric are found in homogenous gabbro, or in banded gabbro with a gneissic fabric; homogenous gabbro is seen with a crosscutting intrusive contact against banded gabbro with a gneissic fabric; and a dyke of banded anorthosite and gabbro with a gneissic fabric coplanar with the banding and dyke walls crosscuts, at right angles, banded cumulate gabbro with a gneissic fabric coplanar with its banding. The transition from gabbro to 100% sheeted diabase dyke complex is exposed south of Point Rousse. It takes place through about 200 to 300 metres original 'stratigraphic' (vertical) thickness, and is characterised by screens a few tens of metres wide of dykes (of similar width as in the gabbro) separating screens of gabbro of similar width (with few dykes). The gabbro screens narrow upward at the expense of the dyke screens until they disappear and the rock consists wholly of sheeted diabase dykes. The sheeted dyke complex possesses a sheet jointing near coplanar with the dykes. Gabbro with diabase dykes also has a preferred sheet jointing in places, especially south of Red Point. The transition from the dyke complex up into ophiolite pillow lava is not well displayed, but can be seen in Upper Green Cove, where screens of pillow lava a few metres wide are found between screens of dykes. The pillow lava screens widen and become more common upwards. The Mings Bight Ophiolite Complex is

therefore composed of a sequence of lithologies which possess physical relationships that are readily explicable in terms of formation of the ophiolite complex as oceanic crust and upper mantle at a spreading ridge (Dewey and Bird, 1971).
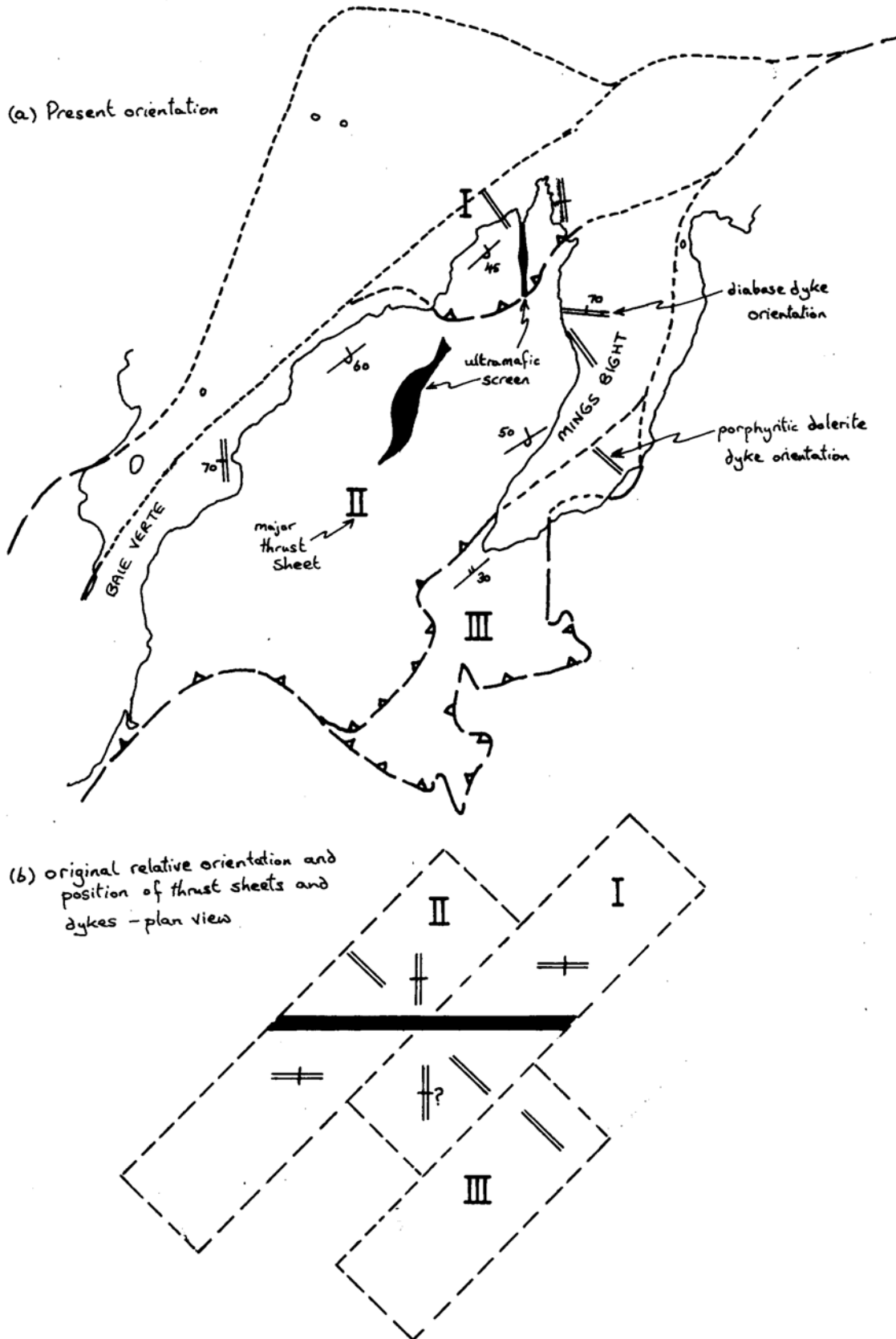
(iv) Sediments and volcanics overlying the ophiolite complex

The 300 metres of sediments conformably overlying the ophiolite pillow lavas consist mostly of weakly-cleaved mafic volcaniclastic sediments with minor thin interbeds of yellow-green argillite and cherty argillite. The mafic volcaniclastics are mostly sandy non-graded grain-flow deposits, with subordinate graded beds probably deposited as turbidites. Beds are typically between 1 cm and 1 metre thick. Non-graded conglomeratic beds, with clasts supported by the mafic sandy matrix, form a small proportion of the section. These beds are typically from 30 cm to 2 m thick; clasts are usually not more than 30 cm across, and typically are 5-10 cm across. The conglomeratic beds near the bottom of the section only contain very angular flakes of laminated green chert and cherty argillite, and uncommon clasts of sandy mafic volcaniclastic rock. Stratigraphically further up the section, these beds also contain somewhat rounded mafic lava clasts, and rare doleritic and gabbro clasts. One particularly coarse-grained deposit is exposed in a channel about 2 metres deep. Slump folds in thin argillite beds and overturned flow casts at the base of sandy volcaniclastic beds are occasionally seen. After a rotation back to horizontal about strike, these indicate a palaeoslope from the east/northeast. A conglomeratic bed with mafic lava clasts containing abundant black actinolitised pyroxene phenocrysts typically 5 mm across, and a graded bed of such pyroxenes, are seen near the top of this section of sediment. The sediments in this section overlying the ophiolite complex strongly resemble those of the Jukes Point Formation south of Kidney Pond in the inland map area and represent the same base-of-slope

Fig. 5.1. Present and original orientations of dykes and the ultramafic screen in the Mings Bight Ophiolite Complex.

5.5

(a) Present orientation

diabase dyke orientation

ultramafic screen

porphyritic dolerite dyke orientation

MINGS BIGHT

BAIE VERTE

major thrust sheet

(b) original relative orientation and position of thrust sheets and dykes – plan view

environment. The pyroxene-bearing bed, on the other hand, resembles the Prairie Hat Members of the Neale's Bay Formation.

The thick section of pillow lava overlying these sediments is mostly exceedingly highly deformed, and resembles homogenous sandy mafic volcaniclastic sediments. The pillows are, however, highly variolitic, and the deformed varioles can be recognised defining the very deformed pillows in much of the section. In two places, sequences from pillow lava to pillow agglomerate show that the section is inverted. Near Mings Bight Wharf, two thick sills of porphyritic dolerite with squat altered plagioclase phenocrysts are seen intruding the section. Narrow non-porphyritic dolerite sills are seen occasionally in both sediments and lavas.

(v) Dyke orientation and penecontemporaneous faulting in the ophiolite complex.

Wide dykes (10-30 m) of porphyritic dolerite, identical to that forming the thick sills in the upper pillow lava, cut gabbro in both inverted thrust sheets, and harzburgite in the lower upright thrust sheet (Plate 5, Fig. 5.1). These dykes have a very consistent orientation, being subvertical and within about 10° of a mean northwest-southeast strike. This is thought to be evidence that there has not been significant rotation of the thrust sheets relative to one another about a vertical axis. Diabase dykes, in the gabbro of the upper thrust sheet from Point Rousse to Red Point, and in the middle thrust sheet in Upper Green Cove, strike north-south and are subvertical. Those in the middle thrust sheet from Red Point to Big Head strike east-west and dip steeply north, orthogonal to the other areas (Fig. 5.1). After the inverted thrust sheets are rotated back to horizontal about strike, these two dyke orientation sets are oriented east-west, and north-south, respectively. As the dykes must have been originally near vertical, the rotation is closely controlled, and a strike rotation must be very near the

real axis of rotation for the deformation. Therefore, as there is evidence that the thrust sheets have not rotated about a vertical axis relative to one another, these two mutually perpendicular diabase dyke orientations are proposed to represent two such spreading directions. In both the upper and middle thrust sheets, there is a narrow strip of ultramafic rock, oriented north-south, separating an eastern and western part of each thrust sheet. In the upper thrust sheet, this runs into Devils Cove. The ultramafic rock is a knotty diapiric serpentinite and its contacts with gabbro to the west and sheeted dykes to the east are subvertical. The exposure on the western side of Devils Cove contains the remaining half of a tuffisite pipe, which becomes subvertical when the inverted thrust sheet is restored to horizontal. The western contact of knotty serpentinite with gabbro is intruded by five or so 10-20 cm wide diabase dykes, and one cuts off a septa of the serpentinite. Therefore this steep sheet of knotty diapiric serpentinite, which separates gabbro from sheeted dykes, is interpreted as a fault formed in the oceanic crust and mantle during its generation. It is presumed to continue, perhaps modified by thrusting, in the middle thrust sheet, and therefore to separate presently east-west dykes in the eastern half from presently north-south dykes in the western half. It is possible that the presently north-south set of diabase dykes formed after the presently east-west set. Therefore this fault occupied by diapiric serpentinite is probably a normal fault associated with the change of spreading direction, although it may have generated from a small transform fault associated with the spreading direction represented by the presently east-west dyke set.

Two other anomalous relationships seen in the area may indicate faulting penecontemporaneous with the generation of the ophiolite complex. At Red Point, there are intrusion breccias of ultramafic autoliths in ultramafic rock, cut by a few north-south diabase dykes. This may have resulted from

Fig. 5.2. Duck Island gabbro.

modification of oceanic upper mantle when it passed the end of a ridge segment across a small

transform fault. On Duck Island (Plate 6), mafic gabbro possessing an intense gneissic foliation

is intruded by non-foliated leuco-gabbro (Fig. 5.2), which in turn is cut by a few north-south

diabase dykes. The intense foliation in the mafic gabbro is not comparable to the weak gneissic

fabric seen in the other gabbro of the Mings Bight Ophiolite Complex or in other described

ophiolite complex gabbros. It is comparable, however, with the intense foliation and hot

mylonite zones seen locally in gabbro in the inland map area. It clearly developed during

formation of the ophiolite complex, and it is suggested that its formation was also connected with

processes on a transform fault.

(vi) Structure

The three main thrust sheets are bounded by thrust zones that contain the Baie Verte S_{1B}

cleavage; they are therefore also tectonic slides. The thrust at the base of the lower upright sheet

involves sympathetic splay thrusts and schuppen of Fleur de Lys schists; these and the main

thrust clearly crosscut and post-date all the regional Fleur de Lys deformation structures, contrary

to the statements of Kennedy and Phillips (1971). The major structure defined by the two

inverted thrust sheets overlying an upright sheet is a disrupted syncline with a D_{1B} tectonic slide

in its axial zone. This, and the southeastward to eastward thrusting on moderately northwest to

west dipping D_{1B} slide planes is wholly consistent with the major structure in the inland map

area. S_{1B} cleavage is everywhere parallel to bedding in the Mings Bight area, where the

relationship can be defined. The cleaved pillow lavas, north of Mings Bight Wharf, and south of

Eastern Point, show a strong longrain and colinear preferred elongation direction of the pillows

that pitches near 90° in the cleavage. The metamorphic assemblage in the pillow lavas and

sediments is the same as in the inland map area. Undeformed diabase dykes in the

ophiolite complex are wholly altered, but gabbros may retain all their clinopyroxene. Plagioclase in the gabbros is wholly albitised or saussauritised. A very large, steeply-inclined Baie Verte D_{1B} tectonic slide zone is exposed on the northwest coast of the Mings Peninsula at Whalesback, and an exposure close to this slide zone is formed by the Deer Cove Islands. This slide zone truncates the moderately dipping middle and upper thrust sheets, and the moderately dipping S_{1B} cleavage in the pillow lavas near Eastern Point and the sheeted dykes in Lower Green Cove. Therefore the cleavage in this slide zone probably belongs to a late stage of the D_{1B} deformation. It may indicate that the steep inclination of bedding and cleavage in the main part of the Baie Verte Lineament is due to an event late in the D_{1B} deformation that steepened originally moderately-inclined structures by continued convergence across the Lineament. A sheet of gabbro overthrusts cleaved pillow lava on a very shallowly northward dipping plane at Eastern Point, and the thrust crosscuts the moderately dipping cleavage in the pillow lava. This local south to southeastward thrusting post-dates the S_{1B} cleavage in the rock underlying it, but almost certainly predates the large D_{1B} slide zone that is located just offshore, and therefore also indicates that there are several parts to the D_{1B} deformation.

Orthogonal (north-south and east-west) normal faults downthrowing consistently westward or northward towards the axis of the Lineament are found throughout the Mings Bight area. These therefore represent the same late stage reversal of tectonic transport found in the inland map area, and are probably correlative with the Park Pond Fault.

Kennedy and Phillips (1971) alleged that there were two exposures of an unconformity of other rocks on ultramafic rocks in the Mings Peninsula. One of these is the half tuffisite pipe in Devils Cove, mentioned above. The

other, in Hammer Cove, is partly an incipient phacoidal tectonic breccia, partly autoliths of altered ultramafic rock in altered ultramafic rock (alleged conglomerate), and partly altered (talcose) cumulate dunite and chromitite (alleged serpentine sediment). Even if these were unconformities, which they are not, the first would post-date Baie Verte deformation, and the second shows a cleavage-bedding relationship indicating that the alleged sediments face downwards towards the unconformity they are supposed to lie on.

CHAPTER 6. SYNTHESIS — EVOLUTION OF THE BAIE VERTE LINEAMENT

"…and please don't be offended at my views, they're really only offered to amuse."

— The Wife of Bath's Prologue; Canterbury Tales.

(i) Fleur de Lys Supergroup; Burlington Granodiorite

The metasediments of the White Bay, Rattling Brook and Mings Bight Groups are interpreted

(Dewey and Bird, 1971) as continental slope-rise sediments. The mafic volcanics and

volcaniclastics of the Pacquet Harbour Group and the overlying silicic volcanics and

subaerial/shallow-water sediments of the Cape St. John Group are interpreted as an island arc

subsequently built on the outer part of the continental rise, prior to deformation and

metamorphism. In the western Fleur de Lys terrain, psammitic and psephitic metasediments and

amphibolites of the Seal Cove Group, which overlie a quartzo-feldspathic gneissic basement also

riddled with amphibolites (deWit, thesis, 1972) are interpreted as initial rifting 'graben facies'

clastics and basalts overlying continent-derived basement. These are correlated (deWit, op.cit.)

with the Grenville basement and Labrador dyke swarm, and locally overlying basalt and

conglomerate of the Western Platform. The basalt and dyke swarm are thought to be related to

the initial rifting that formed the northern Appalachian/Caledonian ocean (Bird and Dewey,

1970; Strong and Williams, 1972). The youngest preserved units in the western Fleur de Lys

terrain are in the White Bay Group which consists of pelitic to partly pebbly quartzo-feldspathic

schists with interbedded resedimented carbonate beds. These are probably no younger than

Middle Cambrian, the age of the upward change from sandstone/quartzite to carbonate

deposition on the Western Platform. The status of the mafic volcanics of

the Garden Cove Formation in the White Bay Group (deWit, op. cit.), and of the mafic

volcaniclastics of the Birchy Schist Formation, is uncertain, and they may not be correlative.

Either might represent the initial rifting, an intraplate volcanic event, or be the distal equivalent

of the mafic rocks of the Paquet Harbour/Cape St. John island arc. That arc was constructed on

oceanic lithosphere (Dewey and Bird, 1971), seen as giant inclusions of gabbro and ultramafic

rock in the Cape Brulé Porphyry, a subvolcanic silicic intrusive complex in the Cape St. .John

Group. It is also seen as areas of sheeted dykes, in the Nippers Harbour Group, and in the Little

Bay Head Group in the Halls Bay area (Dewey and Bird, 1971). These two groups consist

mainly of pillow lava and mafic volcaniclastics, and probably consist partly of older oceanic

crust and partly of overlying equivalents to the Pacquet Harbour mafic arc rocks. They are

mostly strongly and complexly deformed, but of low metamorphic grade, and indicate a waning

of the Fleur de Lys deformation and metamorphism southeastward. The Burlington Granodiorite,

if it is pre-kinematic to the Fleur de Lys deformation, is probably the plutonic equivalent of

intermediate volcanic rocks in the Pacquet Harbour/Cape St. John arc. The Dunamagon granite is

syn-kinematic with the deformation and metamorphism and later than the island arc. The major

part of the western Fleur de Lys terrain is underlain by continental-derived basement. However,

the D_1 Roadside slide, and a probable equivalent structure in the central northern part of the

western Fleur de Lys terrain, carry meta-ultramafic and gabbroic tectonic slivers, probably

derived from underlying oceanic crust and mantle. Therefore an interface between continental

and oceanic lithosphere lay under the metasediments seen in the eastern part of the western Fleur

de Lys terrain. A possible reconstructed cross section for the rocks older than Fleur de Lys
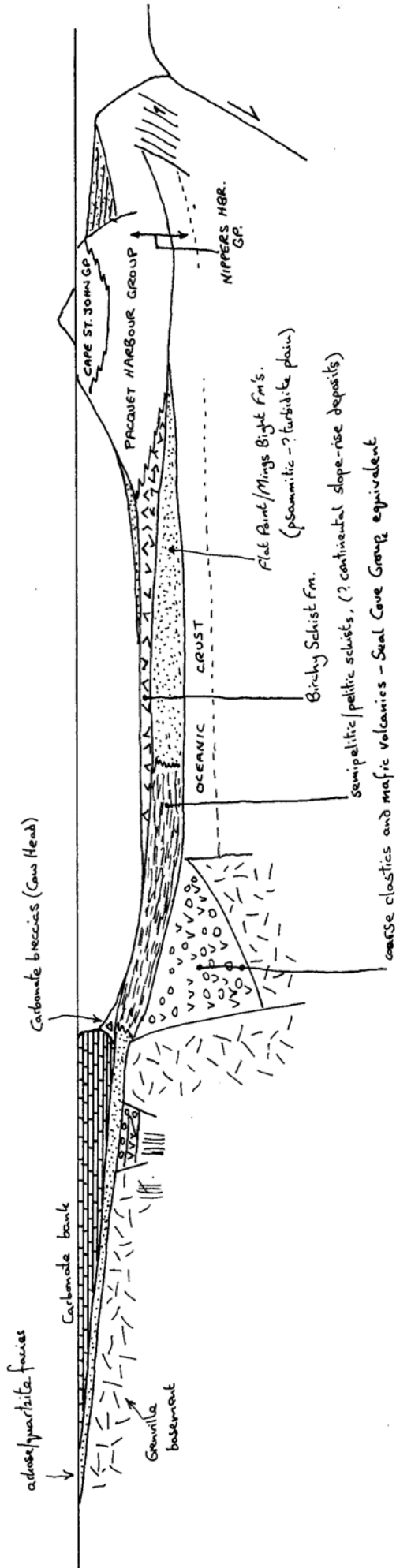
deformation and metamorphism, and

Fig. 6.1. Schematic cross section of the western margin of the Newfoundland Appalachian ocean prior to Fleur de Lys deformation and metamorphism.

their relation to the Western Platform, is shown in Fig. 6.1. It should be noted that the 'graben facies' clastics in the Burlington Peninsula are involved in Fleur de Lys deformation and metamorphism, while similar rocks in the Hare Bay allochthon were little to unaffected (Smyth, 1971), and carbonate turbidites and black shale, deposited on the continental rise, and now found in the.Humber and Hare Bay allochthons were also unaffected. This shows that the metamorphic front originally cut across the sedimentary facies belts, and that disjunct parts have been juxtaposed across the Carboniferous strike-slip Cabot Fault.

The Fleur de Lys deformation and metamorphism is dated as pre-Arenigian, because deformed clasts of Cape St. John Group rocks are found in the Snooks Arm Group (Dewey and Bird, 1971). It is also indirectly dated as pre-Tremadocian. Undeformed allochthonous pillow lavas in the Hare Bay area are of Tremadocian age (Kay, 1969), and are presumably part of the allochthonous Bay of Islands/White Hills ophiolite complexes. The Bay of Islands ophiolites are internally undeformed and have sheeted dykes in a northwest-southeast orientation; dolerite dykes in this orientation cut penetratively foliated amphibolite, gabbro and granodiorite forming the allochthonous Little Port slice to the west of the Bay of Islands ophiolite slice (Williams, 1973). The amphibolite and meta-gabbro, intruded by granodiorite and then penetratively deformed, are presumed to be a sample of an equivalent to the Fleur de Lys terrain, and the mafic rocks might have been old oceanic crust. This evidence also shows that the Bay of Islands ophiolites formed within the previously deformed and metamorphosed terrain.

The available evidence suggests that the building of the Pacquet Harbour/Cape St. John arc, and the subsequent Fleur de Lys deformation and metamorphism, are events that took place between Middle Cambrian and Tremadocian, and are probably of late Cambrian age. These two events are the earliest

evidence of subduction in this sector of the western margin of the northern Appalachian ocean.

(ii) Baie Verte Group

    (a) General relations

The mafic volcaniclastic sediments and overlying pillow lavas in Mings Bight conformably overlie the internally undeformed, little-disrupted ophiolite complex basement. The conglomeratic and sandy mafic volcaniclastic and argillaceous sediments here very closely resemble those in the southern Jukes Point Formation, and the Kidney Pond Conglomerate and Boudin Pond Formation underneath the latter are therefore interpreted also to have been conformably deposited on an ophiolite complex basement. Clasts of deformed Fleur de Lys rock in the Kidney Pond Conglomerate show that it was deposited after Fleur de Lys deformation and metamorphism. Large (2 m) clasts of Burlington Granodiorite indicate that the section exposed was not deposited more than a few tens of kilometres from that source, and some sedimentary structures show that the clasts were derived and transported from the east with respect to their present position.

If the ophiolite complex base to the Bale Verte Group is *in situ* with respect to the eastern and western Fleur de Lys blocks, then it must have formed later than the Fleur de Lys metamorphism and deformation, because the ophiolite complex is not penetratively deformed, is not significantly metamorphosed, and is not overlain by any significant arenaceous quartzo-feldspathic sediment. The Fleur de Lys terrain, however, is penetratively deformed, is in at least epidote-amphibolite fades, and contains large quantities of arenaceous quartzo-feldspathic metasediments, on both sides of the Baie Verte Lineament. If the ophiolite complex base was a downfaulted piece of an allochthonous slice of old oceanic crust and mantle derived from the southeast of the Burlington Peninsula, then the Baie Verte Group

sediments would not be expected to overlie it conformably and evidence of the emplacement of such an allochthonous slice, such as underlying melanges, should be present, and none is seen. Furthermore the sense of tectonic transport in the eastern Fleur de Lys terrain is southeastward, and structures in the Little Bay Head terrain are steep, also with a southeastward sense of overthrusting (J. deGrace, pers. comm.).

It might be proposed that the eastern Fleur de Lys block was moved in along a strike-slip fault after the hypothetical allochthon of the ophiolite complex base had been emplaced and before the deposition of the Kidney Pond Formation containing Granodiorite boulders. However, the meta-granodiorite apophyses found along 15 km of the eastern edge of the western Fleur de Lys terrain are proposed to be the original western intrusive contact of the Burlington Granodiorite. Also, the aeromagnetic expression of the large ultramafic bodies does not continue beyond the northern end of the peninsula and the magnetics indicate that the structures in the Rattling Brook Group curve around without interuption across the end of the large body under Baie Verte, and continue eastward. These data suggest that there has not been significant net strike-slip displacement along the site of the Lineament.

The Boudin Pond Formation might be termed a melange, but in the author's view this is not the best term to describe it. The megaboulders are only of ophiolite gabbro and they appear to be closely packed, whereas a polymictic deposit with a larger amount of wholly soft-sediment-deformed matrix would be expected in a melange derived from an advancing allochthon, judging by examples seen in the Humber Allochthon. The two examples of sediment seen between the gabbro blocks are bedded, and conformable with the sediments overlying the Formation. Only small-scale sediment disruption is seen in discrete beds within the intercalated sediment, and in one case clearly

relates to the deposition of a single gabbro block. Therefore, it is thought that this deposit was accumulated block by block, not be mass movement of boulders and matrix as in an olistostrome deposit. The conglomerate of the overlying Kidney Pond Formation could be described as an olistostrome, but there is no indication that it was derived from a moving allochthon, which would have to have contained some Burlington Granodiorite and some Cape St. John Group besides the ophiolite complex. Both these Formations are interpreted to have conformably overlain an ophiolite complex base, and they are conformably overlain by a large thickness (now 2½ km, before deformation 4-5 km) of sediment and intercalated volcanics, not an allochthon. Furthermore, large blocks of ophiolite gabbro are found near the preserved top of this sequence. This fact is not easily reconcilable with the hypothesis of an allochthonous ophiolite complex, because the sediments must be *in situ* relative to the Burlington Granodiorite, and the hypothetical allochthonous ophiolite complex should have stopped moving so that part could be supposedly downfaulted and buried by the Baie Verte Group sediments and volcanics.

The Bay of Islands ophiolites were expelled in middle Ordovician time from a zone of telescoping and thrusting that now runs along western White Bay and south through Deer Lake (Dewey and Bird, 1971). Allowing sufficient strike-slip movement on the Cabot Fault, it might be alleged that it originally continued southward along the Baie Verte Lineament. However, there is no evidence of compressional deformation of the ophiolite complex or the conformably overlying sediments and volcanics in the Baie Verte Lineament until after deposition of the early Devonian Mic Mac Lake Group. There is also no evidence of westward tectonic transport in the Lineament of any ophiolite complex rocks or volcanics and sediments that are definitely part of the Baie Verte Group. There is therefore no reason to postulate a source for.the Bay of Islands ophiolites from the Baie Verte Lineament, or from

further east, or to postulate any other allochthonous ophiolite sheet. These hypotheses involving

allochthons are all somewhat forced, and the data and inferences are all more compatible with

the development of oceanic lithosphere along the present site of the Baie Verte Lineament

between eastern and western Fleur de Lys blocks, after their deformation and regional

metamorphism had taken place. This is the same situation inferred for the development of the

Arenigian Snooks Arm Group in its present site and the Tremadocian Bay of Islands ophiolite

complex in its original site, for reasons stated previously. Lower and middle Ordovician (Arenig-

Caradoc) mafic to silicic volcanics, and shallow water clastics and limestones, that are

interpreted as island arc volcanics and associated sediments, are found in the Halls Bay area to

the southeast of the Burlington Peninsula (Catchers Pond, Cutwell, and upper Western Arm

Groups). Rocks of equivalent age in a zone adjoining these to the southeast, consisting of sandy

volcaniclastics, slates and minor limestone and mafic lavas (Horne and Helwig, 1969) are

interpreted as the arc-trench gap. The Dunnage melange, next to the southeast, is interpreted as

the rear trench wall deposit marking the site of the northwest-dipping subduction zone

responsible for the arc vulcanism (Bird and Dewey, 1970).

Formation of oceanic lithosphere by spreading within and just behind island arcs (above active

subduction zones) has been proposed for the present Marianas (Karig, 1971), Tonga-Kermadec

(Lau-Havre trough) (Sclater and others, 1972), New Hebrides (Karig and Mammerickx, 1972),

Scotia (Barker, 1972), and Andaman (Rodolfo, 1969) island arcs. Appropriate spreading

magnetic anomalies have been shown to exist behind the Scotia arc and in the Lau-Havre trough.

In the other areas, truncated morphology of the 'remnant' arc split off the active arc by the

proposed spreading, and high heat flow and a

Fig. 6.2. Sketch map of the New Hebrides island arc (from Karig and Mammerickx, 1972).
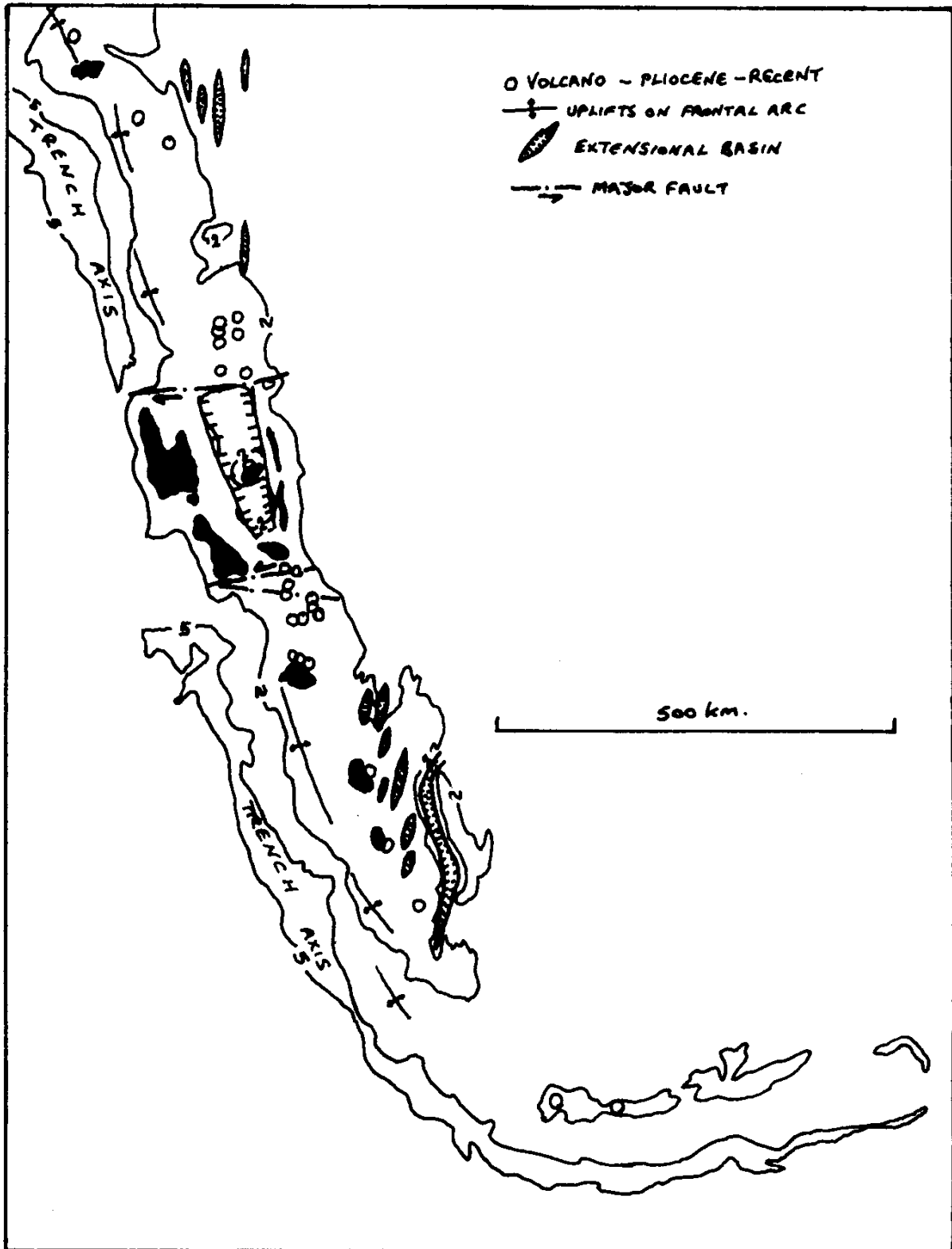
relative lack of sediment in the 'inter-arc' basins given the present supply, are fairly persuasive evidence that the basins are being formed by active sea-floor spreading behind the arc, and as a consequence of subduction, not as part of the primary plate boundary system.

Given this mechanism, the features of the Baie Verte Group and its relationship to the Fleur de Lys terrain may be readily explained. A basin formed by splitting apart the basement (Fleur de Lys Supergroup, etc.) to the early Ordovician island arc, and spreading formed oceanic crust and mantle as its floor. The basin filled mainly with mafic volcaniclastic rocks, mostly eroded subaerially from mafic volcanics and volcaniclastics erupted by the arc to the southeast. Mafic volcanics intercalated with the sediments were probably erupted from central volcanoes or fissures built in the basin, although some may have spilled in from the arc to the east. The fact that a thick section of mafic volcanics and volcaniclastics, rather than pelagic sediments, conformably overlies any ophiolite complex is thought to be strong evidence in itself that the ophiolite complex was part of the floor of an inter-arc basin rather than the main ocean.

The Baie Verte Lineament is about 100 km long, and is not thought to have a northward extension for the reason stated above. It might have originally extended via Birchy Lake and Sandy Lake to the Glover Island Formation at the southern end of Grand Lake, a total length of 230 km. The section in the Mings Bight area is generously estimated to represent a width of oceanic crust not exceeding 30 km when reconstructed. There are no indications of later subduction in the Lineament, and although some width of the original oceanic crust may have been lost by under- or over-thrusting, there seems no good reason to suppose that the basin ever exceeded about 30 km in width. Most present-day active inter-arc basins have widths of 200 km or more, and lengths commonly around 1000 km. However, the New

Hebrides arc (Karig and Mammerickx, 1972) possesses at least 15 small *en echelon* inter-arc basins oblique to the trend of the arc, ranging from 20 to 130 km long and 5 to 30 km wide, and one larger basin of 200 x 60 km (Fig. 6.2). These are proposed to be ideal analogues for the Baie Verte and Snooks Arm basins. The Bay of Islands ophiolites also formed the floor of another early Ordovician inter-arc basin, but this was probably a large structure similar to the large present-day examples because the length of its telescoped source line is at least 500 km, and the ophiolites were obducted as a slab at least 25 km wide, probably indicating closure of its basin by subduction (Dewey and Bird, 1971).

(b) Detailed aspects

The two orthogonal diabase dyke orientation sets in the ophiolite complex in the Mings Bight area are interpreted as due to an originally east-west spreading direction, succeeded by an originally north-south spreading direction. There are several independent indications of major contemporaneous faulting of the ophiolite complex basement to the east of the section now exposed in the Baie Verte Lineament. It is interpreted that water gained access to the ultramafic rocks while they were still at high temperature, because enstatite is pervasively altered while olivine is not. Reintruded ultramafic rocks at Red Point may be due to activity where a spreading ridge axis abuts previously formed lithosphere across a transform fault. The large blocks of ophiolite gabbro accumulated in the Boudin Pond Formation are interpreted to be a submarine scree, made by large blocks spalling off a major fault scarp. The gabbro blocks contain local zones with intense high-temperature gneissic foliation and associated hornblende, hot mylonite zones involving basalt magma, and much tuffisite. As these features are not seen in intact ophiolite gabbro in the Mings Bight area, it is suggested that their development is connected with the fault zone, and two

of these features also suggest entry of water into the ophiolite complex. The intense local high-temperature gneissic foliation developed during formation of the ophiolite complex because it is seen intruded by non-foliated leucogabbro on Duck Island. Small serpentinite screens in at least one block of gabbro in the Boudin Pond Formation also indicate faulting in the zone from which the blocks were derived. Coplanar diabase dykes in the gabbro blocks are mostly oriented so that they lie flat in the bedding orientation. This is to be expected, because intact ophiolite gabbro containing diabase dykes (in Mings Bight) shows a sheet jointing coplanar with the dykes, and blocks spalled off a scarp of such rock would tend to have their longest dimensions in the plane of the sheet jointing. It may also indicate that the dykes and sheet jointing were coplanar with the fault scarp, and if so, reinforces the suggestion that north-south spreading followed east-west spreading. These data and inferences can all be combined by the suggestion that the major fault zone was at least in part a transform fault subparallel to the Baie Verte Lineament and connected with the second, north-south spreading direction. This is not to say that more than a few kilometres of strike-slip displacement accumulated across it. This fault zone also formed the eastern margin of the basin, and must have differed from a normal transform fault because several kilometres of vertical displacement are required across it to obtain subaerial exposures of ophiolite gabbro for erosion of the clasts now in the Kidney Pond Conglomerate. It is probable that this proposed fault zone involved slivers of the western side of the eastern Fleur de Lys block, forming ephemeral islands from which the Burlington Granodiorite and Cape St. John clasts were eroded. The marble boulder and rare resedimented carbonate beds suggest that local and ephemeral carbonate reefs formed on some of the proposed islands of ophiolite complex at the top of the main fault scarp. The lower three Formations (Boudin

Pond, Teardrop Pond and Kidney Pond Fms.) in the inland map area all contain large ophiolite-complex derived boulders as evidence of active faulting at that time. Above these, there are no large boulders or ophiolite complex-derived clasts until they reappear in the middle of the northern mapped part of the Neale's Bay Formation. It is suggested that the Fault became quiescent during Kidney Pond Formation times, and that renewed movement occurred briefly later. If this interpretation, and that of east-west preceding north-south spreading, are correct, then most or all of the spreading that formed the basin was complete by the time of deposition of the Kidney Pond Formation. If Ti contents of ophiolite basalts reflect spreading rates (Nisbet and Pearce, 1973), and if they are applicable to inter-arc basins, the very low Ti contents of the Baie Verte ophiolite basalt (0.3%) mean they were formed at a very low spreading half-rate, not covered by Nisbet and Pearce's data, but probably 0.5 cm/yr or less. The most common dyke width in the Baie Verte ophiolites is 20-50 cm, systematically smaller than that of Troodos (1-2 m), which is estimated by Nisbet and Pearce (1973) to have formed at a spreading rate of around 1 cm/yr. This may also reflect a slow spreading rate for the Baie Verte ophiolites. At 0.5 cm/yr half rate, a 30 km wide basin would take 3 m.y. to open, and in view of the uncertainties in the original width and the spreading rate, this might be taken as an average estimate for the time taken to form the Baie Verte basin.

The Baie Verte basin was mostly filled by mafic volcaniclastic rocks, and mafic pillow lavas. The mafic volcaniclastics appear to have been plagioclase-rich, and mainly to have been eroded at shorelines or subaerially in the volcanic arc to the east, and emplaced mainly by a grain flow mechanism. The mafic lavas may contain some plagioclase phenocrysts, but mafic phenocrysts are very rare. A relatively minor amount of albite phenocryst-bearing

Fig. 6.3. Schematic map for the western margin of the Newfoundland Appalachian ocean during the lower Ordovician (based on Fig. 6.2).
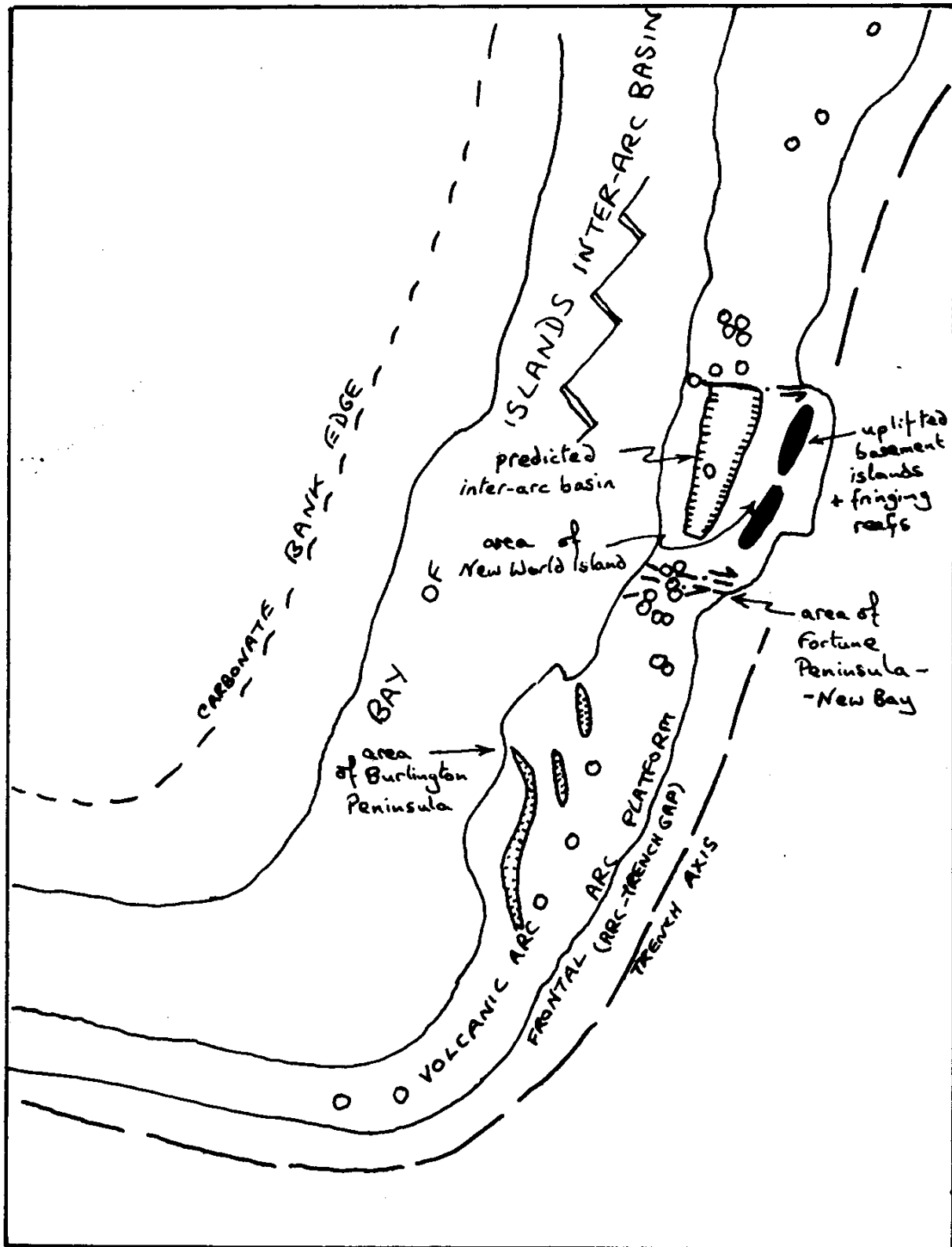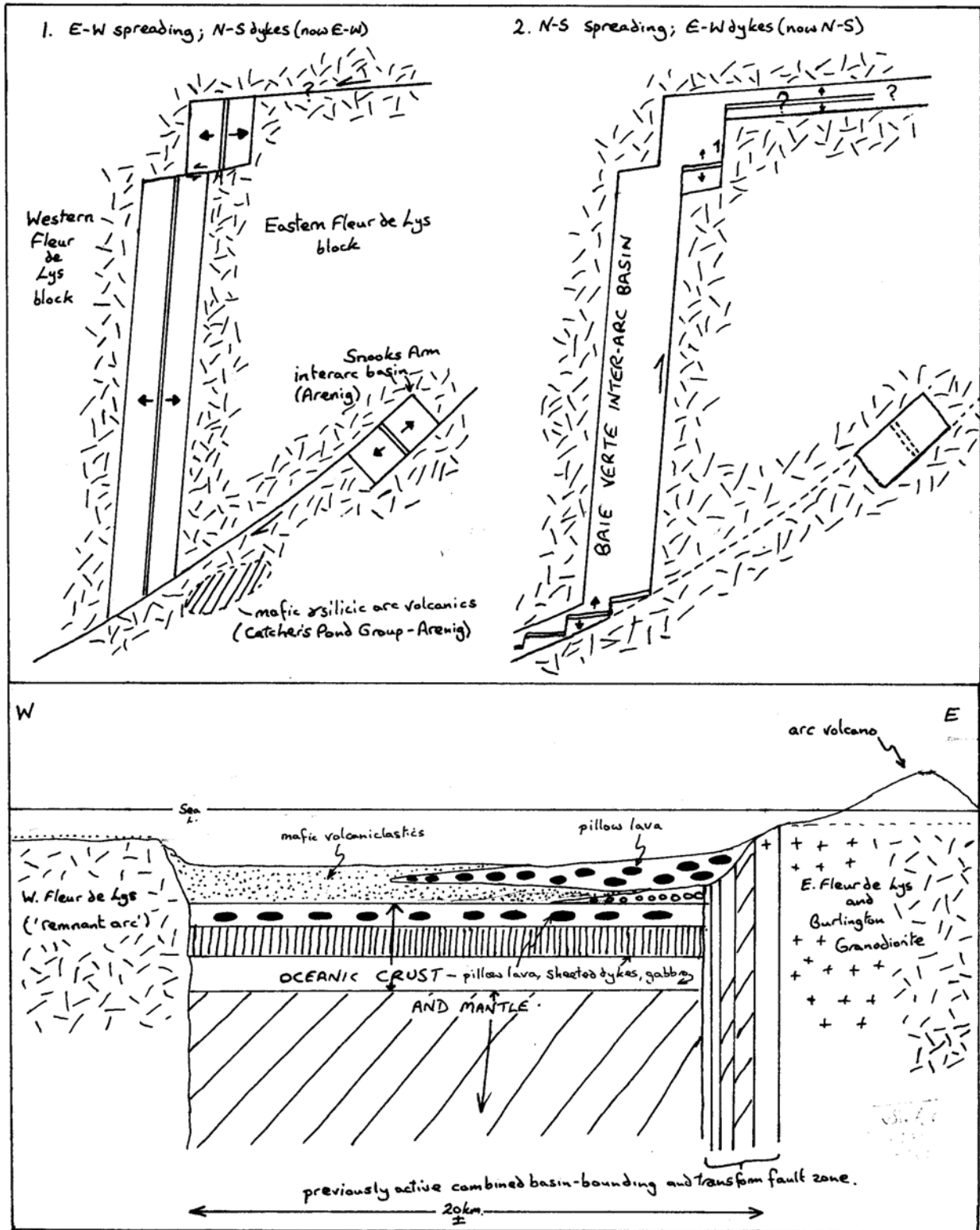
Fig. 6.4. Sketch maps of proposed two-stage spreading history of the Baie Verte inter-arc basin;

schematic cross section during eruption of Slink Pond pillow lavas.

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pink silicic tuff is found in the upper part of the northern Neale's Bay Formation. There is a remarkable correspondence between this record and the Pliocene-Recent vulcanism in the New Hebrides arc. That vulcanism mainly consists of feldsparphyric basalts and basaltic andesites, with minor andesite lavas and fragmental deposits and rare dacite pumice (Mitchell and Warden, 1971). The pillow lavas of the Slink Pond Formation represent a partial section across an outer lobe of a volcano, which was probably sited on the eastern boundary fault, or within the Baie Verte basin. The largest, central, interarc basin in the New Hebrides has an arc volcano within it. It has erupted feldsparphyric basalt, but currently erupts ankaramitic basalt (Mitchell and Warden, 1971). The latter may be very similar material to that contained in the Prairie Hat Members of the Neale's Bay Formation.

Thus, not only does the present New Hebrides arc provide the best modern analogues for the Baie Verte interarc basin, but also some gross characteristics of the accompanying arc vulcanism are apparently very similar. The map of the New Hebrides, when mirrored, is proposed as an ideal generalised map for western Newfoundland in the early Ordovician, if the carbonate bank platform edge and the larger Bay of Islands interarc basin are inserted, (Fig. 6.3). Sketch maps showing the proposed spreading history and possible extensions of the Baie Verte interarc basin, and a schematic cross-section at the time of eruption of the Slink Pond Pillow Lava Formation, are shown in Fig. 6.4.

There is no known geological record between Ordovician and early Devonian in the Burlington Peninsula. There is no evidence that the sediments filling the Baie Verte basin were deformed prior to the deposition of the Mic Mac Lake Group. The slightly angular unconformity between the Mic Mac Lake and Baie Verte Group contrasts with the rugged palaeotopography on the

unconformity between Mic Mac Lake Group and Burlington Granodiorite. This contrast suggests that the slight tilting and differential uplift of the Baie Verte Group before deposition of the Mic Mac may be entirely due to domal uplift associated with the development of the caldera complex source of the Mic Mac silicic volcanics.

(iii) Mic Mac Lake Group

Quartz-feldspar porphyry and syenite was intruded into the Burlington Granodiorite in early Devonian time, and formed a ring complex probably composed of at least two successive overlapping calderas, and provided the silicic volcanic rocks of the Mic Mac Lake Group. The initial erosion of the associated domal uplift produced a rugged topography near the ring complex. Mafic lavas were erupted from dykes, cutting the Granodiorite and probably located in a ring-like zone surrounding the silicic ring complex. Mainly coarse (conglomeratic) immature, locally-derived clastic material was mostly deposited in alluvial fans by flash floods. Two eruptive episodes separated by a period of strong erosion, and perhaps renewed doming, are preserved in the map area. The volcanics are bimodal, and both eruptive sequences start with mafic lavas and change upward to rhyolitic volcanics, without any intermediate rocks. The rhyolites are mainly ignimbrites, with a thick unit of rhyolite flows in the older sequence (Armageddon Formation). The ignimbrites and rhyolites in the older sequence are mainly non-porphyritic; those in the younger sequence are all porphyritic, although a few non-porphyritic rhyolite sills are found among them. Rhyolite sills found on the unconformity between Baie Verte and Mic Mac Lake Groups suggest that there was some intrusion of silicic magma along the pre-existing major fault zone between Baie Verte Group and Burlington Granodiorite, because these sills are not seen in the equivalent position to the east. Abrupt palaeotopographic features on both major erosion surfaces include sections

Fig. 6.5. Schematic cross-section of the Baie Verte interarc basin after deposition of the Mic Mac Lake Group.
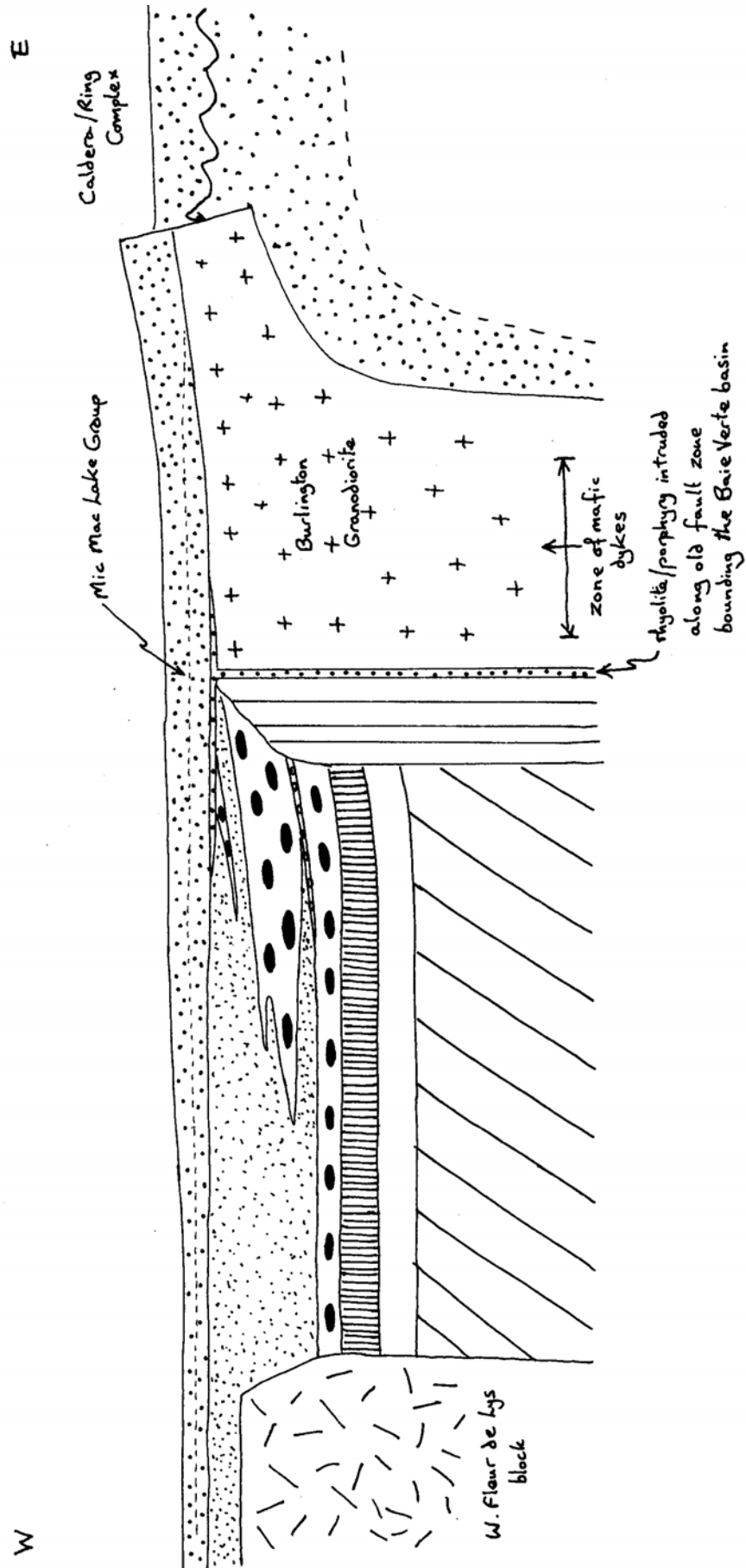
across Devonian canyons as much as 300 metres deep. The larger palaeo-topographic features appear to have been structurally controlled by faults formed along a preferred fracture orientation in the underlying Burlington Granodiorite. The maximum thickness of the present section in the map area is about 2000 metres. The order of eruption, and the large volume of silicic relative to mafic volcanics, perhaps indicates that the rhyolites are from crust melted by the mafic magma. This cordilleran-type volcanic episode perhaps indicates renewal or acceleration of subduction on the northwestern side of the northern Appalachian ocean during the late Silurian-early Devonian, because there are no known arc/cordilleran volcanics on this margin between Llandovery and very late Silurian ages (Williams, 1967). A schematic restored cross-section of the Baie Verte basin after deposition of the Mic Mac Lake Group is shown as Fig. 6.5.

(iv) Deformation of the Baie Verte Lineament

The Baie Verte and overlying Mic Mac Lake Groups were deformed, and the present Baie Verte Lineament produced, due to the relative approach and attempted collision between the western Fleur de Lys and eastern Fleur de Lys/Burlington Granodiorite rigid blocks that bordered the Baie Verte interarc basin. The structure is steep and asymmetric, with attempted overthrusting of the western block over the eastern block (Fig. 6.6). There is no indication of lithosphere subduction during this convergence, and the basin seems to have been initially buckled into a large syncline. However, a large proportion of the upper mantle below the oceanic crust flooring the basin must have been overridden by the western Fleur de Lys block, and it is likely that a narrow strip of the western Fleur de Lys block that was immediately adjacent to the western side of the Baie Verte basin was overridden and is not now exposed. Progressive convergence tightened the syncline, overturning its western limb followed by eastward overthrusting

Fig. 6.6. Interpretative and simplified cross-sections of the present Baie Verte Lineament.
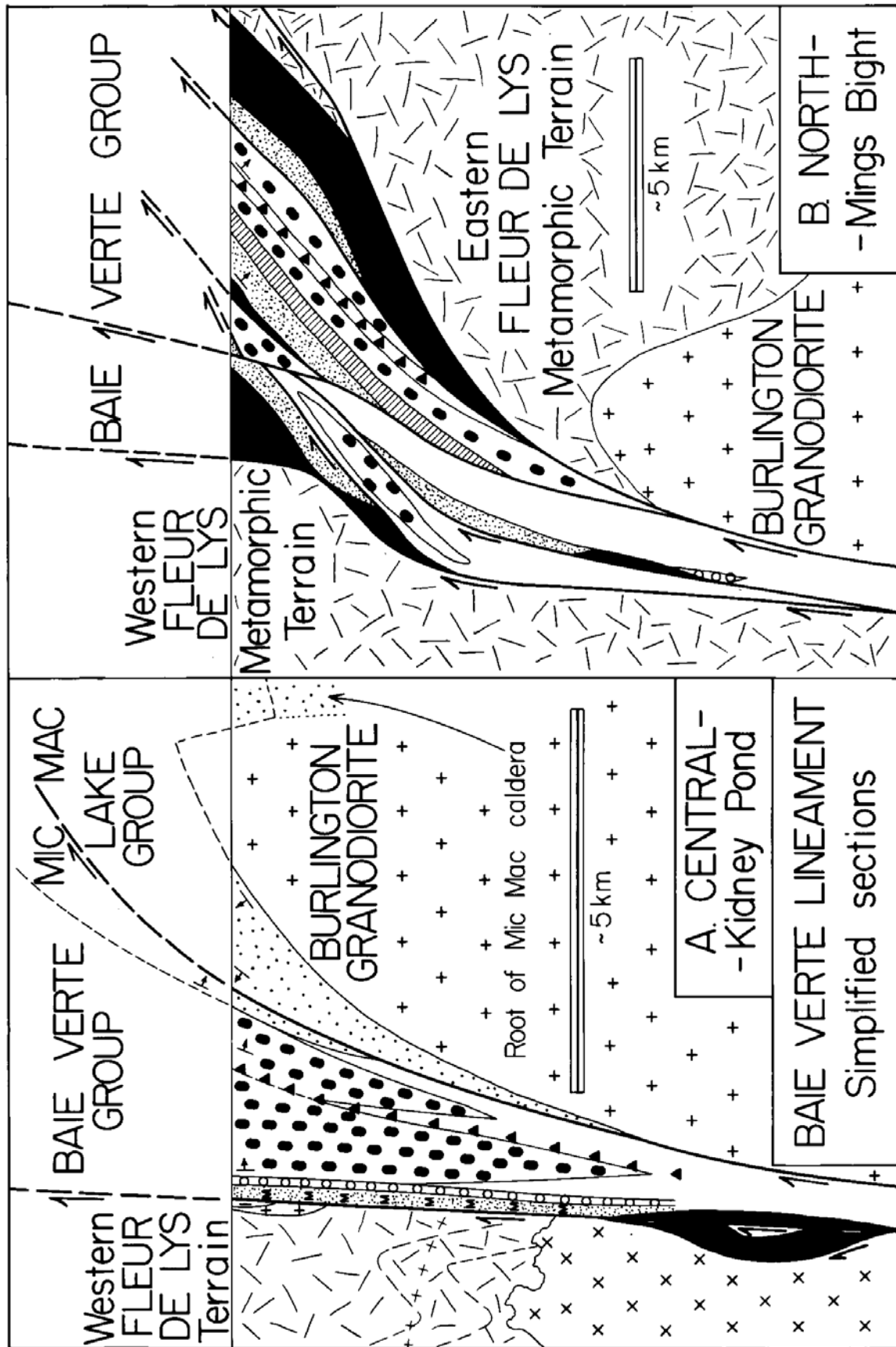
on a moderately inclined major tectonic slide along its axial surface. In the Mings Bight area, the

upright lower limb of the syncline was thrust over the eastern Fleur de Lys schists, and also the

overturned upper limb was segmented by a major eastward moving thrust. During this time, the

western Fleur de Lys schists developed their tectonic contact with the ultramafic rocks that

underlay the oceanic crust of the basin floor. The western Fleur de Lys block together with the

indented ultramafic bodies then converged on the Baie Verte Group sediments and volcanics,

probably steepening them from a moderately dipping to a subvertical attitude in the western part

of the Lineament, and forming the Western Lineament Boundary Slide. During the early part of

this process, a sliver of the western Fleur de Lys schists together with part of the Mic Mac

ultramafic body was moved sideways and upward on the Marty's Pond Fault, presumably

because of a temporary local stress difference due to the shapes of the impinging rigid blocks

along this part of the zone of convergence. The present sideways motion of the Turkish plate

between the converging Arabian and Eurasian forelands (McKenzie, 1972) seems an appropriate,

although larger scale, analogy to this situation. After convergence across the Baie Verte

Lineament had ceased, there was a minor reversal of the tectonic transport direction, with

relative uplift of the Burlington Granodiorite/eastern Fleur de Lys block. This was perhaps due to

isostatic readjustment of the underthrust block and may also have been connected with erosion of

the material thrust over the eastern block.

The Wild Cove Pond Granite/Diorite complex, and Celebes Pond Granite were intruded post-

kinematically into the Fleur de Lys schists prior to the deformation in the Baie Verte Lineament.

These and other similar granites in the Fleur de Lys terrain, and one sample of Fleur de Lys

schist, all give K-Ar ages of about 370-350 m.y. There is sufficient uncertainty

in the geological time-scale (Lambert, 1971), and error in the particular age determinations for these to be middle rather than late Devonian ages. The vesicular mafic dykes associated with the Wild Cove Pond Complex suggests that the present land surface was not deeper than about 2 km when they were intruded, and therefore that the 358 m.y. age from the Wild Cove Pond Complex is the age of intrusion. However, if during the Baie Verte deformation the western Fleur de Lys block did not move a lesser distance upward than the ultramafic bodies that they now adjoin, the present surface of the western Fleur de Lys block would have been not less than 7 km deep just before the deformation. This is below the level for argon retention in micas, even with a normal geothermal gradient (Dewey and Pankhurst, 1970). Therefore it is possible that the ~360 m.y. ages are cooling ages due to uplift, and date the Baie Verte deformation. That deformation is, therefore, probably Acadian, the middle Devonian event due to major continental collision and suturing across the site of the northern Appalachian ocean (Bird and Dewey, 1970). Subsequent geological events recorded in the map area are restricted to a possible dyke of unknown age on the Park Pond Fault, and the Pleistocene to Recent glacial and post-glacial features.

(v) Regional implications

The closest modern analogue of the early Ordovician island arc recorded in the northwest Newfoundland Appalachians is proposed to be the Plio-Pleistocene New Hebrides arc. The largest (200 x 60 km) interarc basin in the. present New Hebrides (Fig. 6.2) is bounded at its ends by transcurrent faults that cut both the arc and the frontal arc platform and arc-trench gap. In the sector of the arc defined by these two transcurrent fault zones, the trench is not present as a topographic feature, and the arc-trench gap

is occupied by recently uplifted islands of the present basement that lie along strike of the trench

to the north and south. It appears that the frontal arc platform has been pushed out over the

trench (and uplifted) in this sector, perhaps by the spreading in the inter-arc basin behind it.

Therefore the frontal arc platform in this sector has a very narrow width for arc-trench gap

sediment accumulation, and it is shallow compared to the arc outside this sector. In the Notre

Dame Bay area in Newfoundland, the early to middle Ordovician rocks interpreted as arc and

arc-trench gap show a very similar relationship. The boundary between these two facies is

marked by the Lobster Cove-Lukes Arm fault, presumed to be the main frontal arc fault for the

Ordovician arc. This, and the two facies belts, are offset dextrally by a series of close-spaced

cross-faults in the area of the Fortune Peninsula (Fig. 6.3). To the west of these faults, the

sediments interpreted as representing the arc-trench gap occupy a wide zone and are thicker than

their equivalents to the east of the cross faults, which occupy a narrow zone. In addition, some

shallow water limestone is found to the east, but not to the west of the cross-faults in the section

interpreted as arc-trench gap. Therefore, it is predicted that the remains of a relatively large early

to middle Ordovician interarc basin underlie the waters of Notre Dame Bay east of the line

defined by the cross-faults (Fig. 6.3). This may be detectable from aeromagnetic maps if

remnants of the ophiolite complex ultramafic upper mantle of this proposed basin currently

outcrop on the sea bed.

There is a good correlation between the time of development of the Arenigian Murrisk interarc

basin in Ireland (Dewey, 1963, 1967) and the Tremadocian to Arenig interarc basins in

Newfoundland, which were all formed on the northwestern side of the northern Appalachian-

Caledonian ocean (Bird and others, 1971). The Murrisk basin was, like the Baie Verte, also

deformed during the continental collision across the site of the northern Appalachian-Caledonian

ocean, although the somewhat earlier late Silurian-early Devonian age of this deformation in

Ireland shows that the collision was diachronous. The major structure of the Murrisk basin also

resembles that of the Baie Verte Lineament, containing a major synclinal structure, with steep

dips on the north side, and moderate dips on the south side, and with associated southward

thrusts. However, the Baie Verte fold is much tighter, and the thrusting more pronounced.

Another structural resemblance between the two basins is that the inferred ophiolite complex

pillow lavas of the Lough Nafooey Group, on the south side of the Murrisk basin, were uplifted

on penecontemporaneous faults relative to the main part of the basin (Dewey and others, 1970).

This is an equivalent structure to the large scarp formed of part of the ophiolite complex floor

inferred to have been at the southeast margin of the Baie Verte basin.

If the Taconic and Sillery Allochthons of the northern Appalachians (Dewey and Bird, 1970)

were formed by the closure of one or more larger interarc basins, as was the Humber Arm

Allochthon in Newfoundland (Dewey and Bird, 1971), then interarc basins of varying sizes were

formed along at least 2500 km of the northwestern side of the Appalachian-Caledonian ocean in

early Ordovician times. Interarc basins form over the zone of high heat flow and anomalously

thin lithosphere that occurs just behind the volcanic chain in island arcs and volcanic cordillera

(Karig, 1971b). Armstrong and Dick (1973) have suggested that large overthrust sheets may

originate from this zone of high heat flow and thin lithosphere if it is placed in compression,

because the thin lithosphere is more easily broken than elsewhere. Thus, this zone may be a site

of compression or extension, depending on the intraplate stresses. It is therefore suggested that

the regional metamorphism and deformation of the Fleur de Lys Supergroup, the

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succeding development of the interarc basins, and the later obduction of the Bay of Islands

ophiolite slab, are varied results of the same fundamental process.

The geological record preserved in the Burlington Peninsula records parts of the history of a

small part of the margin of the Appalachian-Caledonian ocean (Bird and Dewey, 1970) during its

opening and closing cycle (Wilson, 1968). It happens that the Baie Verte Lineament also records

the history of the opening and closing of an ocean, albeit a small one.

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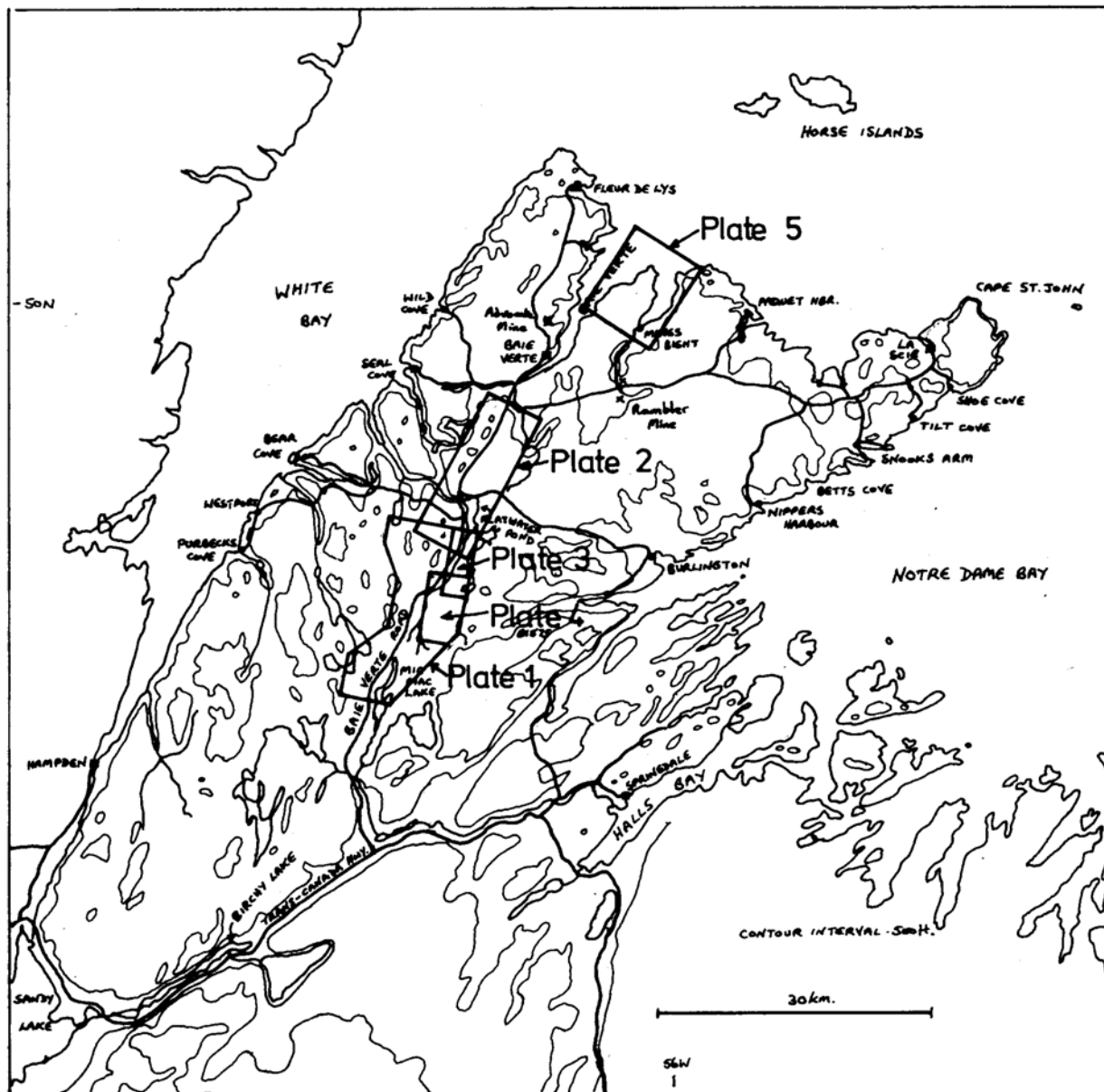
Location index map and list of Map Plates (all in pocket)

Plate 1. Geology of the Baie Verte Lineament; Mic Mac Lake — Flatwater Pond area. 1:15,840.

Plate 2. Geology of the Baie Verte Lineament; Flatwater Pond — La Scie Road area. 1:15,840.

Plate 3. Geology of the Mic Mac Lake Group; Flatwater Pond— Camp 166 Road area. 1:4,800

Plate 4. Geology of the Mic Mac Lake Group; Camp 166 Road — Park Pond area. 1:4,800.

Plate 5. Geology of the Baie Verte Lineament; Mings Bight area. 1:15,840.

Plate 6. Geology of the Western Burlington Peninsula and the Baie Verte Lineament. 1:63,360.

Plate 7. Cross sections of the Baie Verte Lineament: stratigraphic columns for the Baie Verte Group.

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