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GEOLOGY OF THE BASS RIVER COMPLEX, COBEQUID HIGHLANDS, NOVA SCOTIA

BY

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©

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ABSTRACT

Rocks of the Bass River Complex are interpreted as a multiply deformed sequence comprising four lithotectonically distinct units. The Folly River Schist and Gamble Brook Schist form the supracrustal sequence and are, respectively, heterogeneously deformed submarine meta-volcanic and meta-sedimentary sequences within which primary structures have been recognised. The Great Village River Gneiss represents circa. 900 Ma basement to the supracrustals and is composed of three distinct sequences: the Portapique River Schist, the Bass River Amphibolite, and the Great Village River Orthogneiss.

Two structual domains are identified within the Complex. Rocks of the Great Village River Gneiss define Domain A and bear local evidence of an early deformation peak (DB1) not seen in the Gamble Brook Schist and Folly River Schist of Domain S. Maximum metamorphic conditions reached sillimanite grade during this period. Both domains were deformed by the strong DB2 event which shortened the sequence by extensive tight to isoclinal folding, transposition, and possibly by translation. Granitoid rocks (circa 810 Ma??, Gaudette et al., 1983) were emplaced prior to or during this event.

Subsequently, Late Hadrynian regional folding (DB3) produced east to northeast trending, variably plunging, asymmetric folds that are overturned to the north. Locally a crenulation fabric associated with these folds has been enhanced by chlorite crystallisation. Dioritic intrusions of probable late Precambrian age post-dated the DB3 regional

folding deformation and effectively date the end of pervasive tectonism in Bass River Complex rocks.

In Late Devonian and Early Carboniferous time the Complex was intruded by post- Acadian plutonic rocks and suffered right lateral displacement along the Gamble Lake Fault (DB4a) and Rockland Brook Fault (DB4b). These are the earliest phases of deformation attributed to Glooscap Fault System (King et al., 1975) displacements. Subsequently, a paleo-stress field of uncertain origin (DB4c) acted upon the rocks of the Complex producing a poorly-developed, high angle, conjugate fracture system which crosses the Rockland Brook and Gamble Lake structures but is truncated by the Londonderry Fault.

Early Carboniferous fault movement (DB5) along the Londonderry Fault emplaced the Londonderry Formation in its present location along the south margin of the Bass River Complex, locally overprinting the DB4a (Gamble Lake Fault) fabric. The final deformation to affect these rocks (DB6) involved normal slip along numerous north-trending faults which disrupt Lower Jurassic strata.

Domain A Bass River Complex rocks are correlated with circa 900 Ma gneisses of the Mount Thom Complex, 30 km to the east, and are also similar in age to gneisses in southwest New Brunswick. Relative age-equivalence of Bass River Complex Domain S rocks with the Greenhead Group of southwest New Brunswick is indicated and a similarity of Late Precambrian intrusive histories in the Bass River Complex, the Cape Breton Highlands, and southeast Newfoundland is suggested.

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CHAPTER 1

INTRODUCTION

1:1 General Statement

Rocks of the Bass River Complex form the largest metamorphic terrane of the Cobequid Highlands, north-central Nova Scotia. Their complex structural and metamorphic history has been attributed to either a single Devonian shearing event (e.g., Eisbacher,1967) or to one or more episodes of late Precambrian tectonism followed by Devonian and younger shear events (Donohoe, 1983). Donohoe and Wallace (1980) documented the existence of four lithologically distinct units within the Complex; the Great Village River Gneiss, the Gamble Brook Schist, the Folly River Schist, and the Granite Gneisses. This thesis describes the first detailed study of the entire Bass River Complex subsequent to those studies of Donohoe and Wallace and offers a more detailed analysis of structural, stratigraphic, and metamorphic relationships.

The aim of the thesis project has been to refine the geological interpretation of these rocks by gaining a more detailed understanding of their structural and metamorphic characteristics in order to constrain timing of structural development. This has allowed a comprehensive comparison of these rocks with others of the Cobequids, such as the Mount Thom Complex (Donohoe and Wallace, 1980), and with other similar terranes within the Avalon Zone (Williams, 1979). Studies of this type

are fundamentally important in developing a more complete picture of structural and metamorphic variations in late Precambrian rocks of northern Nova Scotia. The position of the Bass River Complex along the Glooscap Fault, which in this area marks the Avalon Zone-Meguma Zone boundary (Williams, 1979), makes it particularly significant to a regional synthesis.

1:2 Location, Physiography, and Glaciation

The Bass River Complex study area encompasses approximately 200 square kilometres in the Cobequid Highlands of north-central Nova Scotia within a narrow, east-west trending zone adjacent to the southern boundary of the Highlands. It extends from the Debert River area in the east to the Economy River valley, west of Pleasant Hills (Figure 1:2:1).

Physiographically the study area occurs within the "highland complex" of Goldthwait (1924), between the 100 metre and 275 metre contour levels. Stream valleys are deeply incised into the undulating Cretaceous peneplain, remnants of which are defined by the flat topped, pseudo-orthogonal topographic highs. Valleys are typically north-south trending and steep sided, commonly attaining relief of greater than 60 meters from floor to rim. The Cobequid Highlands in general are characterised by deciduous second growth on hill crests and ridges and by thick coniferous cover in stream valleys and at lower elevations.

The history of glaciation in the area is relatively complex with

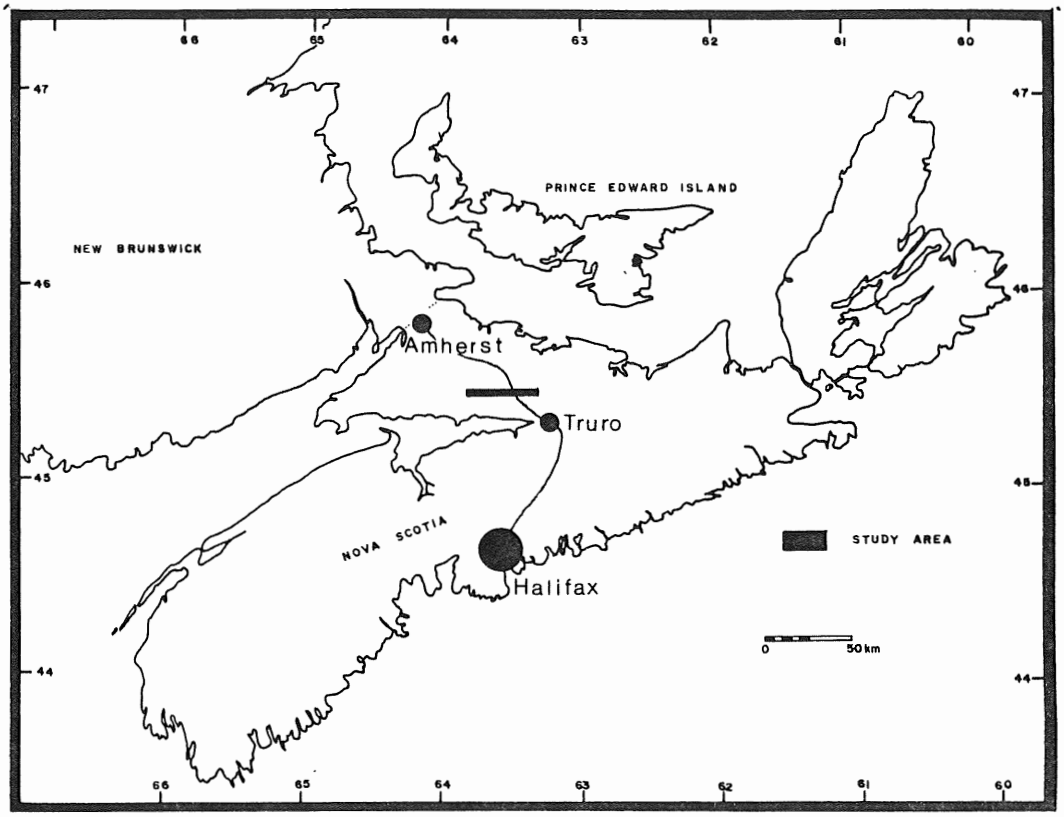
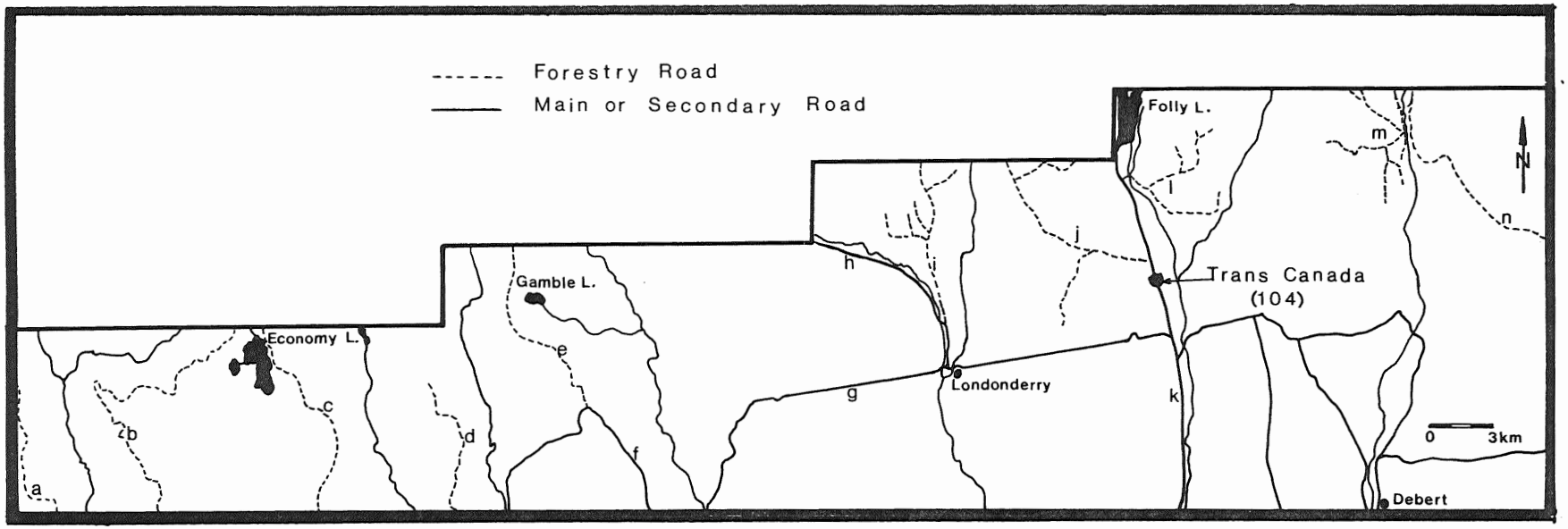


Figure 1:2:1 Location of the Bass River Complex study area, north-central Nova Scotia.

evidence of ice advances from both local and distant centers (Grant,1975; Stea and Fink,1983). Various glacial and post-glacial features occur in the area mapped but study of these beyond recording orientations of bedrock striae was not attempted. Striae measured trend within the range of 345 Az. to 035 Az. but the structures lack obvious indications of movement sense .

1:3 Access

The area is accessible via the Trans Canada Highway (104) in the Folly Lake - Londonderry area and by gravel secondary roads at Higgins Mountain and Sutherland's Lake. Eastward and westward of this central zone the McCallum Settlement - West North River Road and Lynne Road, respectively, are maintained for all-weather traffic. Tracks which are generally passable to trucks and four wheel drive vehicles cross the Highland at Byers Lake - Debert River, Castlereagh- Sugarloaf, and Economy River. An upgraded forestry road from Bass River to Poison Lake provides excellent access to the western area but is limited to three season travel. Private logging road networks mark some portions of the study area providing good access to interstream zones. At present the most extensive are those near Debert River, in the Folly Mountain area, and in the West Bass-Bass River area. An outline of the road systems and their names is presented in Figure 1:3:1.



(a) Economy Lake Road
(b) Newton Lake Road
(c) Economy River Road
(d) West Bass River Maze
(e) Old Castlereagh Road
(f) New Castlereagh Road
(g) Baseline Road
(h) Sutherland Lake Road
(i) Higgins Mountain Road
(j) Folly Mountain Road
(k) Trans Canada Highway
(l) East Folly Mountain Road
(m) Debert River Road
(n) Byers Lake - Debert River Road

Figure 1:3:1 Outline map of main, secondary, and major forestry access roads within the study area.

1:4 Regional Tectonic Framework

Five distinct litho-tectonic zones were identified by Williams (1979) in Newfoundland and incorporated into a general model for Late Precambrian to Late Paleozoic tectonic development of the Northern Appalachian System. Numerous workers have since published refined and altered geologic interpretations (Keppie, 1980; O'Brien et al., 1983) but the general zonation model is still reasonably applicable.

The Bass River Complex occurs along the eastern edge of the Avalon Zone (Figure 1:4:1). In Nova Scotia this eastern boundary is defined by the Glooscap Fault system which separates Avalonian rocks from those of the Meguma Zone. Most workers consider the Avalon Zone to be a micro-continent that was accreted to the northern Appalachians after the Middle Devonian Acadian Orogeny. Precise timing of this emplacement, and the overall significance of transcurrent movement along the orogen are not clear at present but have been considered by a number of workers in recent years (Schenk, 1978; Keppie, 1980; Currie, 1983; O'Brien et al. 1983). Since the Bass River Complex occurs on the eastern margin of the Avalon Zone the results of this thesis place some constraints on possible models for timing and style of emplacement within the Appalachian Orogen.

1:5 Geological Outline of the Cobequid Highlands

The Cobequid Highlands extend across north central Nova Scotia for

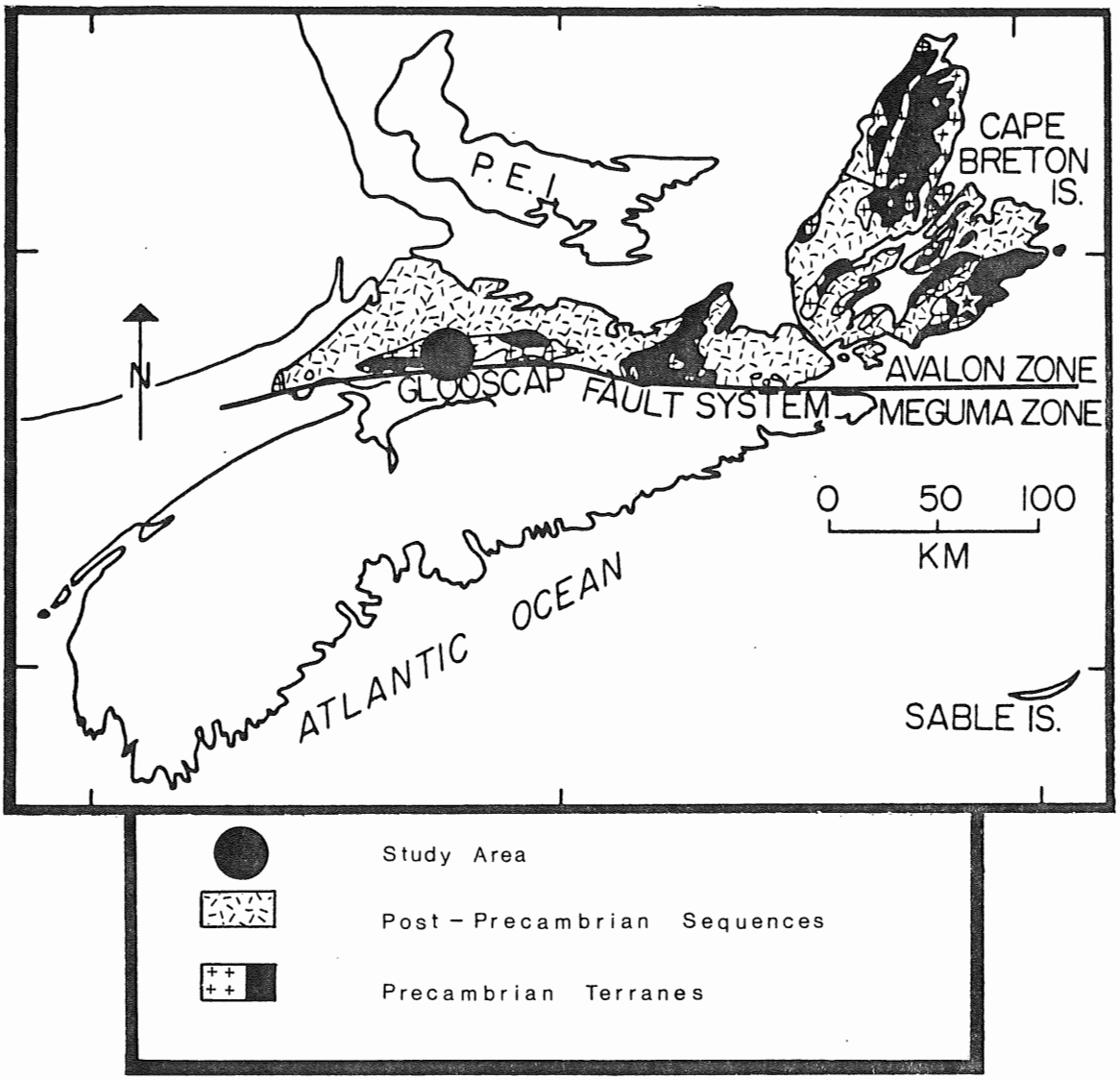


Figure 1:4:1 The Avalon Zone in Nova Scotia (after Graves and Zentilli 1982).

approximately 170 kilometers and are underlain by deformed and undeformed, plutonic, sedimentary, volcanic and metamorphic rocks. To both north and south, sub-parallel Carboniferous sedimentary basins flank the zone and show differing contact relationships with the older rocks of the Highlands. At the surface along the north side Pennsylvanian clastic sedimentary rocks of the Riversdale Group (Kelley, 1966; Poole, 1968) unconformably onlap the Highland, while similar age rocks along the south side (Parrsboro Formation, Donohoe & Wallace, 1980) have been disrupted by movement along splays of the Glooscap Fault system.

Donohoe and Wallace (1980) presented a 1:50,000 scale geology map of the Highlands and described numerous distinct litho-tectonic units ranging in age from Late Precambrian to Early Carboniferous. The distribution and nature of these units are summarised in Figures 1:5:1 and 1:5:2. These figures provide a suitable framework for the following brief outline of Cobequid Highlands regional geology.

The Mount Thom and Bass River Complexes are at least partially correlative metamorphic terranes (Donohoe and Cullen 1983). They have been inferred by Keppie (1981) to form the crystalline basement upon which the Late Hadrynian volcanic-sedimentary sequences of the Jeffers Brook and Warwick Mountain Formations accumulated.

Lower Paleozoic stratigraphic development in the Cobequid Highlands approximates that of the type Avalon Zone of Newfoundland as described by Williams (1979) but distinct variations in total section thicknesses are, of course, evident. No Cambrian sedimentary or volcanic accumula

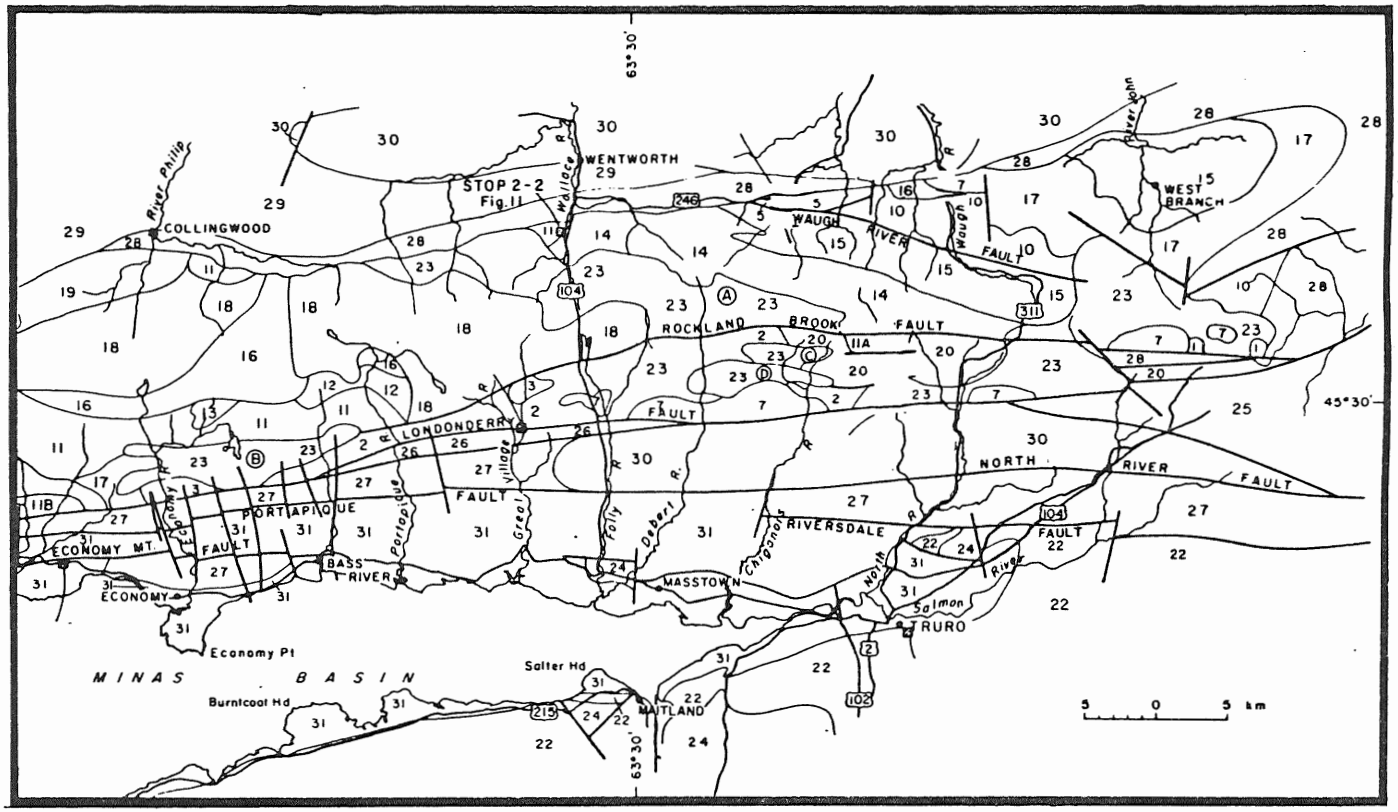


Figure 1:5:1 Regional geology map of the eastern Cobequid Highlands (taken from Donohoe and Wallace 1980) See Figure 1:5:2 for map legend.

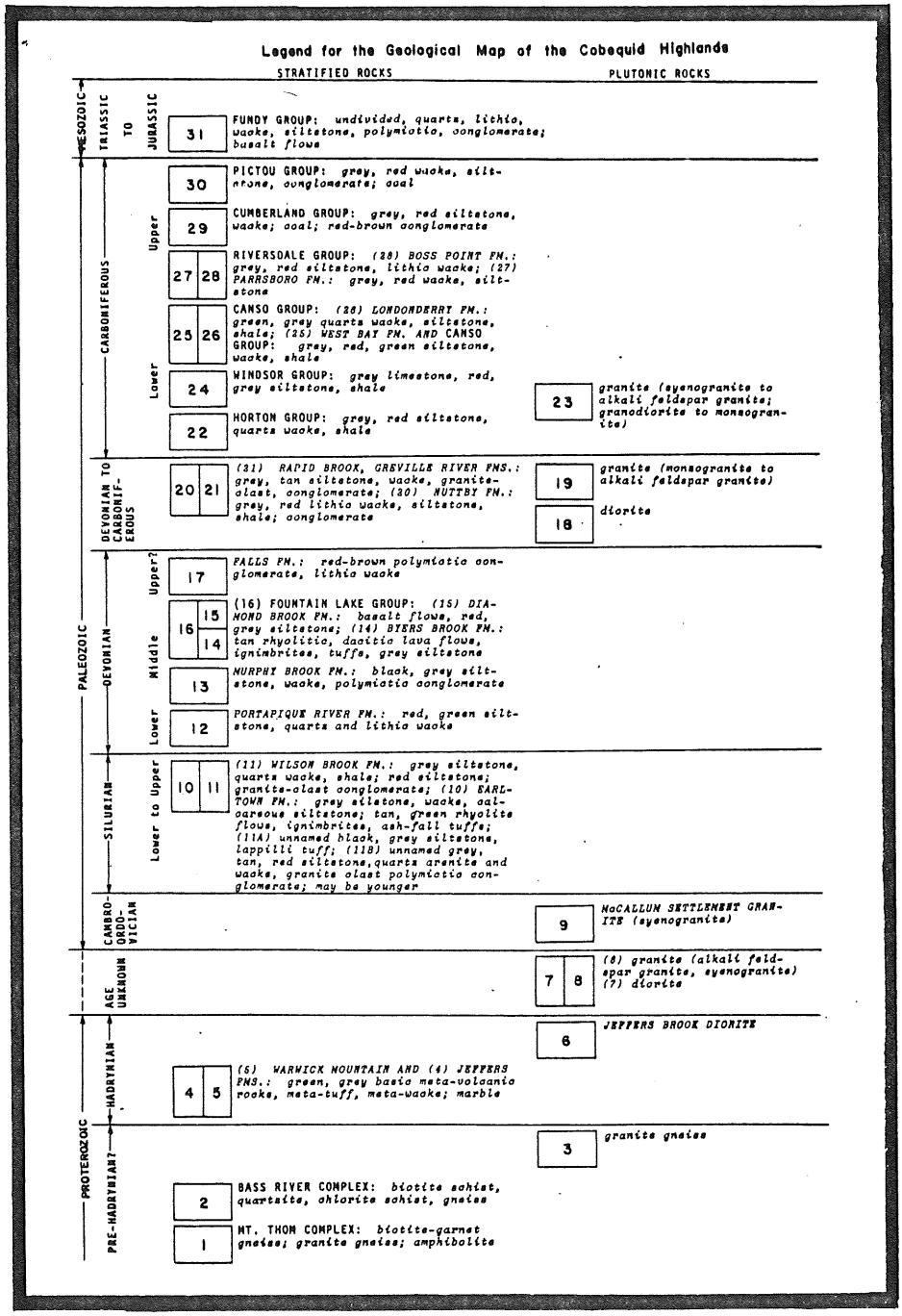


Figure 1:5:2 Stratigraphic outline for rocks occurring in and adjacent to the Cobequid Highlands (taken from Donohoe and Wallace,1980).

tions have been mapped in the Highland although Cambrian rocks were identified 120 kilometers to the east by Benson (1974) and Murphy et al. (1979) in the Antigonish Highlands.

Middle Ordovician to Early Silurian cleaved siltstones, shales, volcanic wackes, and lapilli tuffs characterise the unnamed OS-2 and OS-3 units of Donohoe and Wallace (1980) in the Cobequids and in combination with Early to Late Silurian Wilson Brook and Earltown Formations of the same workers record pre-Lower Devonian sedimentation and restricted volcanism. The Early Mid-Devonian Portapique River Formation and Murphy Brook Formation (Donohoe and Wallace 1980), consisting of siltstones, lithic wacke, polymictic conglomerate and volcanic conglomerate complete the pre-Acadian stratigraphic succession and mark a transition from predominantly marine to terrestrial conditions of sedimentation. These rocks were affected to varying degrees by Acadian deformation (Donohoe, 1980) and intruded by large volumes of Late Middle Devonian to Early Carboniferous diorite and granite.

The Londonderry Formation and Nuttby Formation are the youngest stratified sequences within the Highland. The former occurs as a fault lenticle along the southern boundary of Bass River Complex and the latter unconformably overlies older rocks east of the study area .

1:6 Previous Investigations in the Study Area

Dawson (1873) noted the occurrence of a metamorphic belt in the

Cobequid Highlands south of Folly Lake and interpreted it as a metamorphosed Upper Silurian correlative of the fossiliferous strata near Earltown. Distinction between rock types and units beyond granite, slate, and quartzite was not made. These same rocks were subsequently assigned to the Archean by Honeyman (1881), presumably on the basis of their metamorphic history and absence of fossils.

In an 1885 report Ells distinguished schists and quartzites occurring south of the central intrusive complex from those rocks studied by Dawson (1873) and Honeyman (1881) in the Wentworth Station area north of that complex. Ells used the criterion of similar metamorphic grade as the basis for a tentative correlation with Precambrian rocks encountered in the Caledonian Highlands of southern New Brunswick. His study of rocks now considered part of the Bass River Complex appears to have been concentrated in the Folly River - Debert River area. Although reference is made "blackish hornblendic gneisses" and "talc, chlorite, gneissoid feldspathic schist" encountered along the road from Portapique to Sugarloaf (Castlereagh Road of Figure 1:3:1), no clear indication is given as to inferred age or position relative to other strata of the Highland. He considered the "schistose rocks" of the eastern Highland to overlie the central granite-diorite complex but equivalence of these and the gneissoid rocks near Portapique is not stated directly.

Fletcher's work between 1895 and 1905 resulted in publication of 1:63,360 scale geology maps for most of northern Nova Scotia by 1905 and these provided the first comprehensive outline of Cobequid Highland

geology. Detailed descriptions of these efforts are contained in the Reports of Activities of the Geological Survey of Canada for the years Fletcher mapped in this province.

That portion of the Complex occurring within the Londonderry-Bass River map area was mapped by Weeks (1948) for the Geological Survey of Canada. He interpreted all of those rocks north of the "iron bearing series" (Londonderry Formation of Donohoe and Wallace,1980) and south of the central plutonic complex as belonging to a common, undivided, sedimentary-volcanic unit. Distinction was made between these rocks and those of the Silurian sedimentary sequence in the Portapique River area but uncertainty was expressed in regard to the Silurian age designation for the latter.

An interpretation of faulting along the southern Highland margin offered by Weeks (1948) makes no reference to the boundary faults of the volcanic-sedimentary sequence but does invoke dip slip movement for the Cobequid Fault which marks the Highland boundary with Upper Carboniferous rocks. He implied similar movement along the sub-parallel Portapique and Gerrish Mountain faults to the south.

In 1958 Stevenson reported on the geology of the Truro map area which includes some Bass River Complex rocks in the north. His Silurian age determination for these was consistent with that of Weeks (1948) for their strike-equivalents on the adjoining Londonderry-Bass River map area to the west.

The Geological Survey of Canada began an extensive mapping project

in the Cobequids in 1965 under the direction of Dr. D. J. Kelley and Reports of Activities of the Geological Survey of Canada for 1965 through 1968 outline the course of work. Open file field maps by Kelley are available at the Nova Scotia Department of Mines and Energy Library and data accumulated during field work formed the basis for a Ph.D. dissertation on tectonics of the Cobequid Highlands by Eisbacher (1967). Eisbacher's detailed study included rocks now known as the Bass River Complex and outlined a kinematic model for their deformation history. He concluded that this strongly deformed terrane was the strained equivalent of Silurian and Devonian rocks that occur to the north within the Highlands. Fabric development was attributed to Devonian ductile shear and later brittle faulting.

Most recently Dr. H.V. Donohoe and P.I. Wallace mapped the Highlands for the Nova Scotia Department of Mines and Energy. Various reports and maps have outlined the course of the investigation (Donohoe, 1983; Donohoe and Wallace, 1980; Wallace and Donohoe, 1977; Donohoe and Wallace, in preparation). As a result of this work, the polydeformed nature of the Bass River Complex was recognised and a fourfold lithologic subdivision was introduced. They defined the distinct Gamble Brook Schist metasedimentary unit, the Folly River Schist metavolcanic unit, and a gneissic metaigneous unit termed the Great Village River Gneiss. On the basis of crosscutting relationships and available radiometric age information these rocks were inferred to be at least pre-Early Cambrian in age. Their work outlined areas in which further investigation was necessary and indicated the need for more detailed

information in regard to structural and metamorphic histories. This thesis primarily addresses these aspects of Bass River Complex geology.

1:7 Field Studies

Mapping in the field area began in early June, 1981 and continued until September, 1981, with subsequent short field trips as necessary during the fall of that year and in the summer and fall of 1982.

CHAPTER 2

LITHOLOGY

2:1 General Outline

Donohoe and Wallace (1978) divided the Bass River Complex into four lithological-structural units, the Folly River Schist, the Gamble Brook Schist, the Great Village River Gneiss, and the Granite Gneisses. This scheme proved to be most useful for regional considerations but modification has been necessary to accommodate detailed information collected during this study. Table 2:1:1 illustrates the expanded and revised version of the original four-fold system.

Initial classification of rock types was based on hand specimen descriptions and field associations with modification as necessary based on thin-and/or slab-section observations from nearly two hundred samples. In particular, this led to recognition of a high grade mineral assemblage in rocks that had been grouped initially with the Gamble Brook Schist. As a result, a new subunit termed the Portapique River Schist, was introduced to distinguish these high-grade rocks from others of the Complex.

In the following sections each subunit of the main associations as presented in Table 2:1:1 is described in order of its inferred relative age from oldest to youngest.

BASS RIVER COMPLEX STRATIGRAPHY

Age	Unit Name		Details
Hadrynian or Older	Strongly Foliated Granitoids		circa 810 Ma (Gaudette *et al.*, 1983)
	Gamble Brook Schist	COVER	Metasedimentary sequence
	Folly River Schist		Mafic meta-volcanic sequence
Neo-Helikian or Older	Great Village River Gneiss	BASEMENT	
	(a) Bass River Amphibolite		Hornblende amphibolite with felsic interlayers
	(b) Portapique River Schist		Sillimanite-bearing paragneiss and schist
	(c) Great Village River Orthogneiss		Felsic orthogneiss

TABLE 2:1:1

2:2 Great Village River Gneiss

2:2:1 General Outline

Donohoe and Wallace (1980) define a zone of felsic to intermediate gneisses and amphibolites extending from west of Folly Lake to the Economy River and term it the Great Village River Gneiss, the type section being on that river north of the Londonderry Fault. The present study divides this package of rocks into three new subunits, the Bass River Amphibolite, Great Village River Orthogneiss and the Portapique River Schist, on the basis of lithology and metamorphic grade. Spatial distribution of the subdivisions is illustrated in Plate 1 and Figure 2:2:1 and details of each are presented below.

2:2:2 Bass River Amphibolite

General Outline

This discontinuous zone extends for approximately 10 km westward from Rockland Brook reaching a maximum width along Bass River and probably terminating west of Economy River (Figure 2:2:1). Previous studies (Ells, 1885; Weeks, 1948; Eisbacher, 1967; Donohoe and Wallace, 1980, in preparation) did not distinguish the amphibolites as a mappable unit but sufficient aeromagnetic and geologic data are now available to do so (N.S. Department of Mines and Energy Aeromagnetic Maps 11E5C, 11E5D, Plates 1 and 4, this study).

The section after which the subunit is named occurs on Bass River,

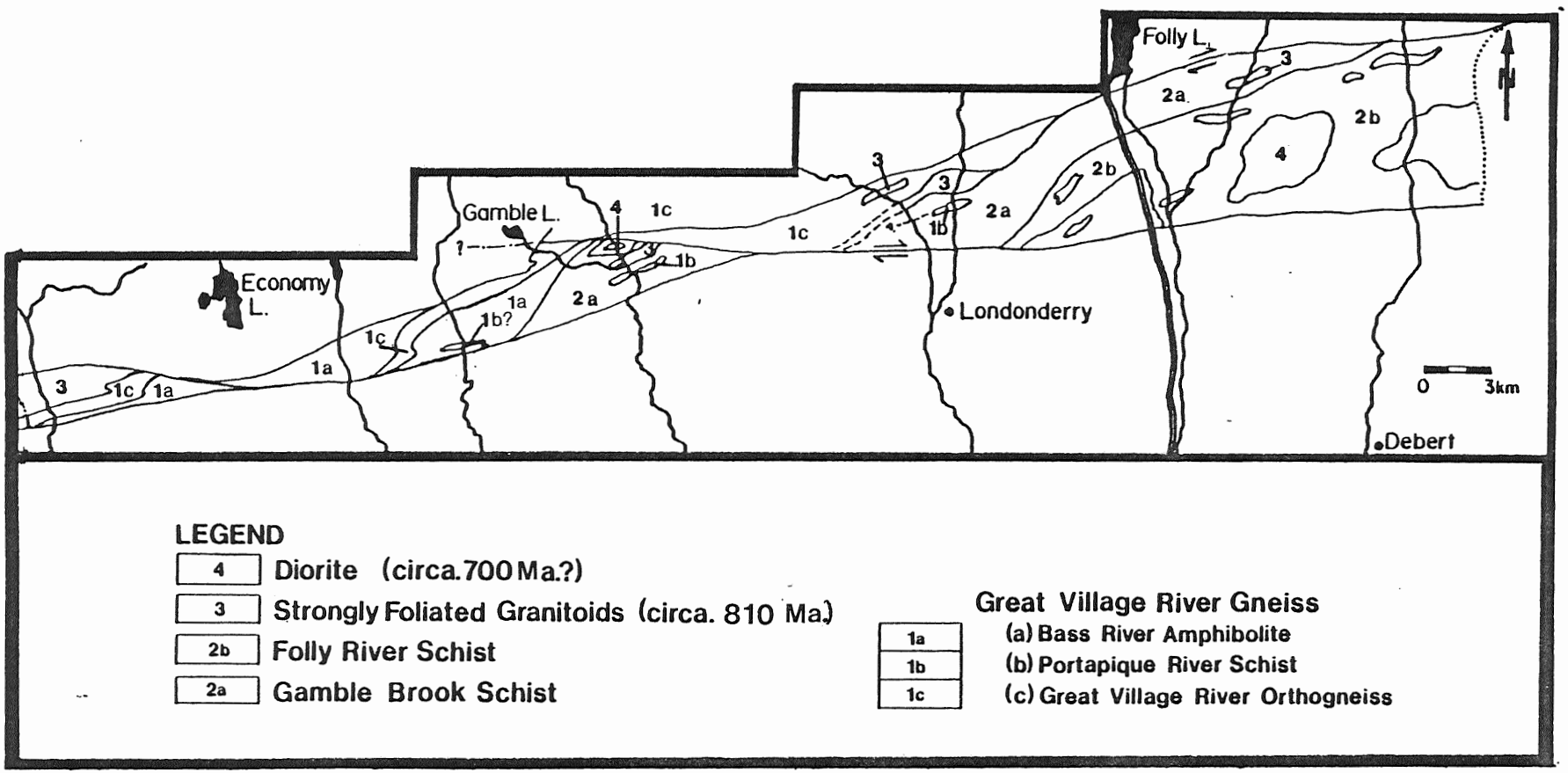


Figure 2:2:1 Outline map of Bass River Complex regional geology. See Plates 1 and 2 for more detailed considerations of lithology and structure.

a short distance north of the northernmost splay of the Londonderry Fault. These rocks can be traced to the next exposed sections westward on West Bass and Economy Rivers by a distinct magnetic anomaly pattern (Plate 4). The northern and southern boundaries in this interstream area are based primarily upon this interpretation of the aeromagnetics, combined with unpublished lithologic information of Kelley (1965 to 1968 openfile base maps available at N.S.D.M.E. library), and the boundary trace shown by Donohoe and Wallace (1980) for their undifferentiated Great Village River Gneiss.

Lithology

The amphibolites are characterised by medium to fine grained, idioblastic, dark green hornblende (30-60%) which defines a strong shape fabric with intergranular altered albite (15-35%), calcic plagioclase (15-35% ; at least in the Higgins Mountain Road area) , epidote (<10%), and biotite (<5%). Feldspars occur as subequant, twinned grains and exhibit varying degrees of sericitic and kaolinitic alteration (15-85%). Plagioclase is generally most severely affected. Rounded inclusions of quartz and feldspar in the coarse hornblende grains are common and chlorite occurs as surface alteration of this mineral and as interstitial, light green, bladed aggregates (Figure 2:2:2). Accessory interstitial opaque minerals (<2%) occur in all sections studied. Although opaque mineralogy was not investigated thoroughly under reflected light, the magnetic signature of the amphibolite implies that magnetite is present in significant amounts. Pyrite and pyrrhotite are not visible commonly in hand specimen but have been observed at several locations on

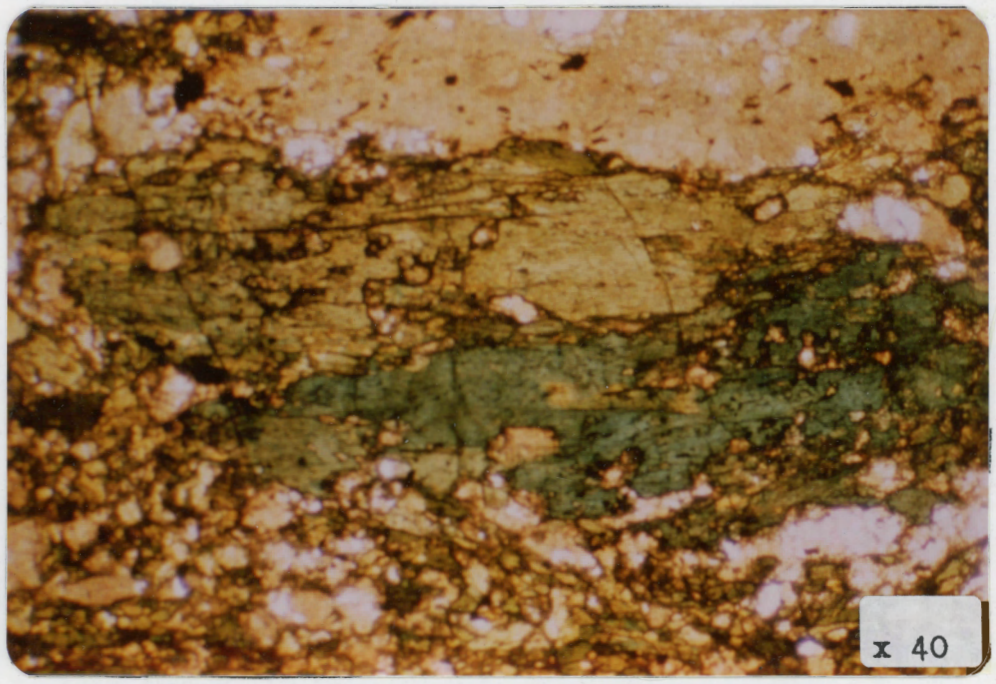


Figure 2:2:2 Chloritisation of hornblende and kaolinisation of plagioclase feldspar in Bass River Amphibolite; note that quartz and feldspar occur as irregular inclusions in amphibole. The matrix is composed of fine grained amphibole, plagioclase, epidote, quartz, and opaques.

various river sections. It is reasonable to suggest therefore, that pyrrhotite content as well as magnetite contributes to the magnetic anomaly associated with these rocks.

Three types of mesoscopic compositional layering have been identified locally:

(1) quartz-dominated segregation layers (2mm to 10cm thick)

quartz > 70%
plagioclase <10%

albite <10%

biotite <5%

epidote <5%

hornblende < 1%

garnet <1%

(2) plagioclase-dominated segregation layers (5mm to 2cm thick)

plagioclase >80%

epidote <5%

sericite <5%

hornblende < 1%

(3) massive hornblende-plagioclase amphibolite (1cm to 15m thick)

hornblende >45%

plagioclase >35%

epidote <10%

biotite <5%

opaques <2%

garnet <<1%

The relative importance of purely metamorphic differentiation processes in developing these layering types cannot be fully evaluated from this study but indications of synmetamorphic tectonism have been noted. Much of the compositional layering seen in these rocks probably developed in close association with a corresponding kinematic culmination (see Chapter 4, section 4:3).

Cullen (1981) investigated hornblende geochemistry from a very restricted area along Higgins Mountain Road and determined that tschermakitic compositions characterise the amphibolite at that location. Plagioclase compositions all fall within the labradorite-bytownite range (An 43 to An 52) and grains showed little compositional variation from rim to core.

The distinction between amphibolite and deformed hornblende diorite is difficult to make in the field and therefore some material within sections of the Bass River Amphibolite may be more aptly termed highly strained diorite. This distinction has been made only where the tectonic fabric is truly non-pervasive, where the equigranular texture of the diorite could be discerned, or where intrusive relationships indicate the diorite to be clearly younger than the amphibolite.

Distribution

The southern contact of the amphibolites is, in most locations, a fault boundary with the Londonderry Formation along the Londonderry Fault (Plate 1, Figure 2:2:1). However, between Portapique and Bass

Rivers thin lenticles of dark brownish grey, Fe-sulphide rich, magnetite bearing, argillacious and psammitic rocks occur within the fault zone. These are structurally isolated from the Bass River Complex but previous to this study had not been recognised as such. At present their age can only be described as pre-dating Early Carboniferous movement along the Londonderry Fault (see Chapter 3, section 3:7).

The nature of the northern boundary varies considerably along the structural strike. On Bass River the amphibolites are interlayered with quartz-feldspar mylonite and fine grained biotite-chlorite-quartz schist to within a short distance of the contact of the Pleasant Hills Pluton (315 + 25 ma; Cormier, 1980). This granite exhibits a grainsize gradation, fining toward the Bass River Complex, but is mylonitised at the contact exposed on the river (Figure 2:2:3). The relationship is therefore interpreted as an intrusive contact which has been subsequently modified by strain along a splay of the Rockland Brook Fault.

On Economy River a tectonic interlayering of amphibolite with rocks of the Great Village River Orthogneiss is characterised by very thin (< 35 cm thick) and somewhat discontinuous bands of amphibolite and orthogneiss. The aeromagnetic trend crosses this section south of the falls (Plate 4), indicating that the interlayering may only mark the transition zone from orthogneiss to more amphibolite underlying the magnetic high.

East of Bass River the southern boundary of the amphibolites continues as the Londonderry Fault but the northern contact is not as

Figure 2:2:3 Mylonitised Pleasant Hills Pluton granite from the contact zone with Bass River Complex rocks on Bass River. Mylonitic layering is well developed although quartz has been extensively recrystallised. The grey porphyroclast is a rounded orthoclase feldspar crystal with quartz-filled micro-cracks at a high angle to the mylonitic layering. (crossed polars)

well defined. Interlayering of these rocks with those of the Great Village River Orthogneiss , Portapique River Schist, and Gamble Brook Schist successively characterises the boundary to the eastward across this area. The amphibolites are inferred to be structurally overlain a short distance west of Portapique River by the Gamble Brook Schist.

2:2:3 Portapique River Schist

Rocks of this subunit outcrop in a narrow, probably discontinuous, band from Rockland Brook to Bass River (Plate I; Figure 2:2:1). In each section they occupy similar structural levels south of Strongly Foliated Granitoid rocks. The widest and best exposed section occurs on Portapique River where biotite - quartz $\pm$ sillimanite $\pm$ garnet $\pm$ staurolite schist and paragneiss outcrop south of the Gamble Brook confluence. Rocks of this unit have not been identified on Great Village River but further sampling and thin section study could result in redefinition of some rocks, now mapped as Gamble Brook Schist associates, as being within the Portapique River Schist. This would depend upon recognition of the high grade metamorphic assemblages.

Lithology

The Portapique River Schist unit is composed of metapelites and paragneisses that occur as discrete interlayers with rocks of the other subunits. Study of available thin sections indicated that these rocks are characterised by a biotite - staurolite $\pm$ garnet $\pm$ sillimanite high grade metamorphic assemblage which is associated with a relict trans

posed layering. Exposures typically show interlayering of light grey to brown, fine grained, quartzitic paragneiss with darker grey, coarser grained , feldspar - quartz - biotite orthogneiss and fine to medium grained biotite - quartz schist. Individual layers of the various rock types range in thickness from less than a few centimetres to 1 metre or more.

The early high grade-layering was overprinted by a later pervasive regional metamorphic foliation. Staurolite, sillimanite, garnet and feldspar porphyroblasts exhibit evidence of post crystallisation fracture, rotation, and alteration relative to the biotite grade overprinting fabric (Figure 2:2:4). Most rocks of the unit have been subjected to shearing strains which generated fabric relationships ranging from transposition to mylonitisation. Truly mylonitic textures are generally confined to discrete zones less than 2 m wide but protomylonitic rocks are more common.

Distribution

As a result of their distinction from other rocks of the Complex by petrographic observations, rocks of the Portapique River Schist are difficult to define in the field. The best exposures are on an isolated section of Portapique River, south of the Gamble Brook confluence, where the previously described interlayering with other Bass River Complex rocks is inferred to occur over about 100m along the river. Section width can only be estimated, of course, since the index minerals were not observed during field mapping. The Rockland Brook and Bass River

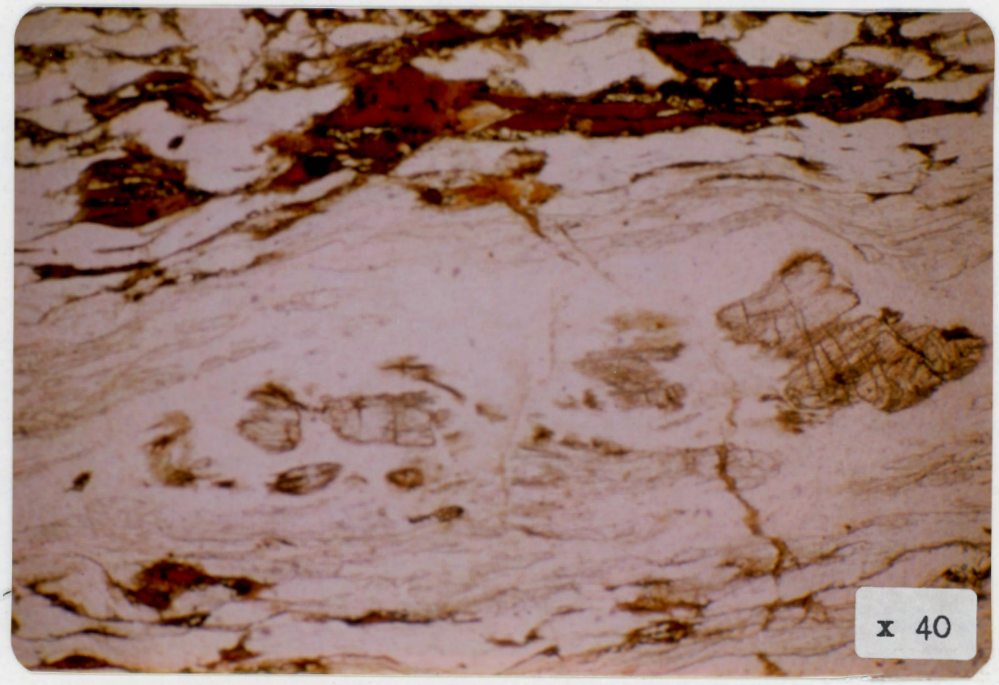


Figure 2:2:4 DB1 generation sillimanite relicts showing extensive replacement by muscovite and evidence of a fracture and rotation history. Sillimanite occurs as crystal fragments and aggregates of fragments within quartz and muscovite rich, discontinuous leucosomes. Biotite, coarse-grained muscovite, and quartz define the DB2 generation overprint. Sample is from Portapique River Schist section on Portapique River. (plane light)

occurrences are similar to that of Portapique River but thinner across strike (Plate 1).

2:2:4 Great Village River Orthogneiss

General Outline

This subunit is characterised by felsic to intermediate, generally protomylonitic or mylonitic gneisses which appear to have had an igneous protolith. All were penetratively deformed and metamorphosed under amphibiolite facies metamorphic conditions. Vestiges of an earlier metamorphic layering have been identified locally, indicating that a significant difference in age and/or deformation-emplacement history distinguishes these rocks from the Strongly Foliated Granitoids. Their spatial distribution on each major river section west of Folly River is considered a reflection of their generally low structural level below the supracrustal rocks of the Folly River Schist and Gamble Brook Schist (Plate 1, Figure 2:2:1).

Lithology

Interlayered, light to medium grey, quartz - plagioclase - biotite + garnet + hornblende gneisses characterise this subunit and generally exhibit a strong tectonic fabric discernible in outcrop as 5 mm to 3 cm discontinuous compositional layering. Pervasive chloritic alteration is typical of the mafic mineral components as is sericitisation, kaolinisation, and saussuritisation of orthoclase and plagioclase feldspars Discontinuously to consistantly banded quartz - plagioclase - biotite +

muscovite + hornblende gneisses are characteristically medium to fine grained and locally exhibit protomylonitic to mylonitic textures (Figure 2:2:5). The dominant foliation is defined by a strong preferred orientation of phyllosilicate phases in combination with porphyroclastic feldspars and recrystallised, quartz- dominated augen structures. The latter produce a characteristic linear fabric on some exposed foliation surfaces.

Generally the early metamorphic fabric is overprinted by the dominant regional foliation . A slight retrogression of the regional mineral phases is exemplified by chloritic surface alteration of hornblendes and it seems highly probable that at least some of the feldspar alteration also developed at this time.

Distribution

The Great Village River Orthogneiss extends in a continuous belt from approximately 3 km west of Folly Lake Quarry as far westward as the Economy River. The widest exposed section is on the Great Village River where these rocks occur for almost 1 km along the stream bed. The northern boundary is marked by the Rockland Brook Fault except along the upper Portapique River where a section of deformed hornblende diorite and a thin slice of deformed granitoid rock occur between the orthogneisses and the fault trace. Diorite appears to have intruded the gneisses at this point but the contact is marked by a strain gradient toward highly sheared porphyroclastic orthogneiss. The foliation within the diorite occurs at a measurable angle to that in the gneisses (110 /55 S : 090/45 S respectively) and is not as well developed. Interlayered foliated granitoid rocks and chloritic, hornblende-bearing

Figure 2:2:5 Mylonitic orthogneiss from the Great Village River Orthogneiss. Recrystallisation of quartz in feldspar porphyroclast pressure shadow and along mylonitic layering has been extensive. Fine-grained muscovite occurs interstitially to quartz and feldspar and parallels the tectonic fabric. (crossed polars)

tectonite occurring between this point and the first upstream outcrop of Silurian Wilson Brook Formation (Donohoe and Wallace 1980) are probably sheared equivalents of the diorite/gneiss contact zone which mark the Rockland Brook Fault. Alternatively, these rocks may represent a thin slice of interlayered Bass River Amphibolite which has been subsequently deformed in the fault zone.

Westward from the Bass River area, where only a thin section of orthogneiss is exposed, its occurrence is difficult to trace due to poor exposure. Occurrences noted by Kelley (Open File maps at NSDM Library) and Donohoe and Wallace (1980) have been interpreted as indicating the existence of a thin but generally continuous band of orthogneiss north of the amphibolite sequence and extending westward to the Economy River. In the east, the inferred structual trend of the zone asymtotically approaches and merges with the Rockland Brook Fault.

A contact with the Gamble Brook Schist is not exposed but is believed to be a transposition-associated shear zone that merges with the southern boundary of the Rockland Brook Strongly Foliated Granitoid body (Plate 1). This conclusion is based on: (1) the local presence of protomylonite along the contact trend west of Slack Lake (Plate 2); (2) the absence of a pre-regional-foliation high grade metamorphic fabric in the Gamble Brook Schist; (3) the consistent intrusive relationship shown by the Strongly Foliated Granitoid rocks with both the Gamble Brook Schist and Great Village River Gneiss.

2:3 Supracrustal Rocks

2:3:1 Folly River Schist

General

The Folly River Schist mafic metavolcanic sequence is confined to the eastern portion of the study area, occurring in a continuous zone from near Londonderry in the west as far east as the Debert River area (Plate 1, Figure 2:3:1). Similar metavolcanic rocks also occur in isolated lenticles along the eastward extensions of both the Rockland Brook and Londonderry faults and have been correlated with the Folly River Schist unit by Donohoe and Wallace (1980) on the basis of lithologic and structural similarities.

The unit is characterised by an abundance of mafic metavolcanic rocks and minor amounts of associated volcaniclastic and chemical sedimentary material. As a result of polyphase deformation, definition of a true stratigraphic sequence within the unit is extremely difficult. The distinctive heterogeneous strain which marks the earliest deformation in these rocks has allowed preservation of primary features such as bedding and volcanic flow structures in certain areas, although in an undoubtedly flattened and/or rotated form. Where younging direction indicators have been noted they show tops to the south on the south limbs of regional folds. This indicates that the volcanic sequence was right-side-up prior to the major folding events that deformed these rocks.

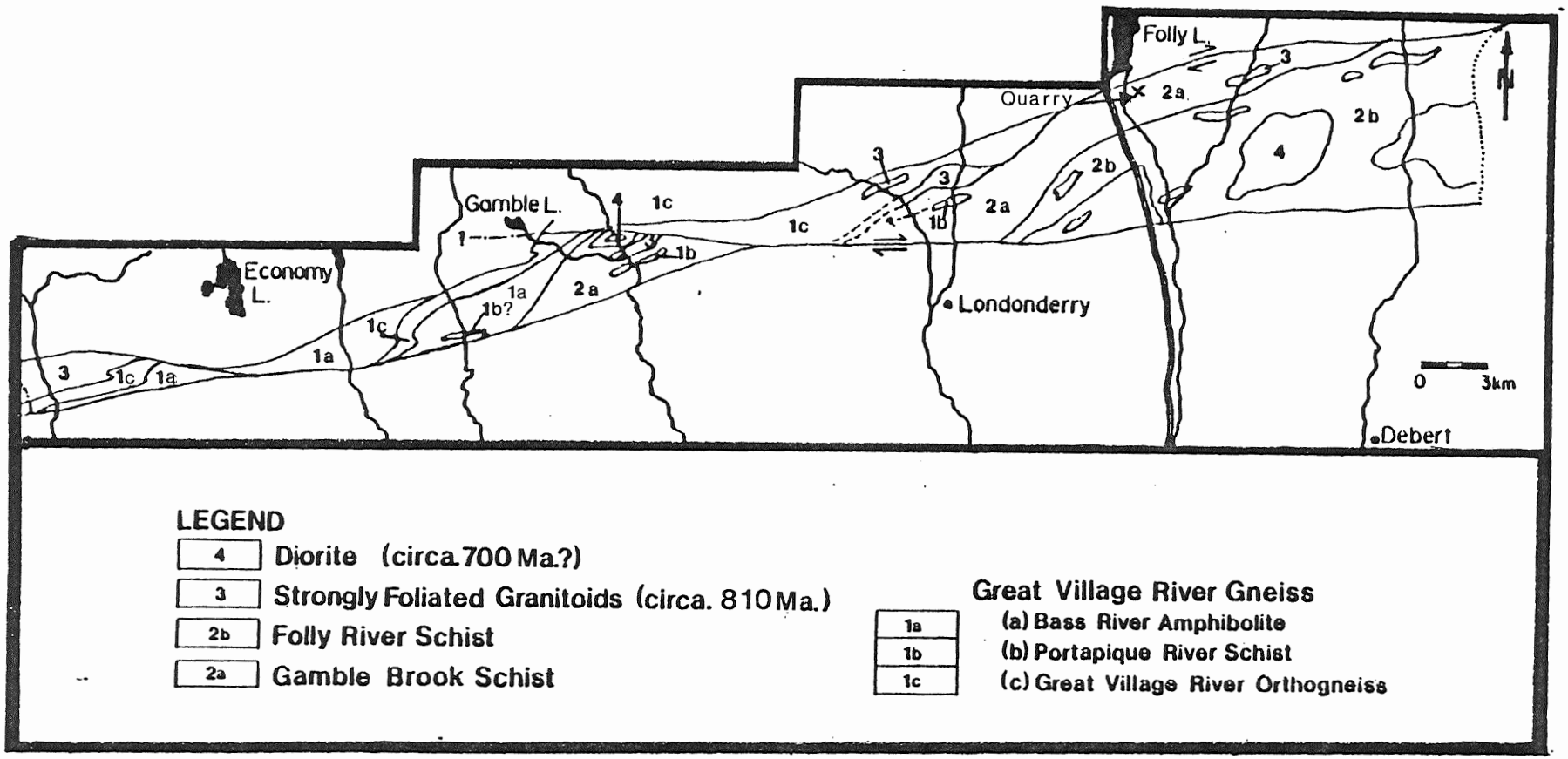


Figure 2:3:1 Outline map of Bass River Complex regional geology. See Plates 1 and 2 for more detailed considerations of lithology and structure.

The following eight lithotypes were mapped in the field and subsequently studied in slab and thin sections: (1) greenschists; (2) mafic volcaniclastic sedimentary rocks; (3) mafic flows and breccias; (4) coarse grained mafic volcanic pyroclastic rocks; (5) rhyolite; (6) mafic volcanic agglomerate; (7) jasperitic iron formation; (8) pyritic schist.

Greenschists

Greenschists are fine grained metavolcanic rocks characterised by a pervasive millimeter scale metamorphic foliation which obliterates most primary volcanic textures. Generally the only suggestion of pre-metamorphic layering visible in outcrop is an intersection lineation visible on some foliation surfaces . This results from intersection of detached and highly attenuated lenticles of sericite, chlorite, quartz, and albite with the metamorphic foliation.

Greenschists have been mapped on the Folly River, along its eastern branch, on East Folly Mountain, and in the Debert River area. Generally these occur in distinctive medium grey-green color outcrops and exhibit a strong parting trend parallel to the metamorphic fabric. Mineralogically, the schists are composed of fine grained chlorite (>25% <40%), albite (<20%), actinolite (>15% <30%), quartz and carbonate (<15%), green hornblende (<10%), biotite (<15%), epidote (<15%), and opaque minerals (<5%). Biotite grains, co-existing with chlorite along the foliation, and knots and fine aggregates of granular epidote are common. Zoisite and allanite locally occur together as zoned crystals

with allanite cores. Compositional variation across the foliation is evident in outcrop by variation in color from light or medium grey-green to dark green over 5 mm or less. In thin section this color change is seen to be attributable to increased sericite, quartz and albite content in the leucocratic layers.

Throughout the metavolcanic sequence greenschists commonly exhibit somewhat ambiguous foliation-parallel contact relationships with more massive, penetratively deformed mafic volcanic or deformed diabasic rocks. This indicates that at least locally greenschists could represent more strongly sheared equivalents of the more massive metavolcanic.

Volcaniclastic Rocks

Zones bearing centimetre-scale layering were identified in rocks of the metavolcanic subunit on the East Folly River. These have been interpreted as volcaniclastic zones within the flow-breccia and agglomerate sequences. The rocks are characterised by generally well developed, alternating, dark grey to black and light grey layers which range in thickness from <5 mm to approximately 15 cm (Figure 2:3:2) and are magnetic. In thin section a direct correlation between the centimetre-scale color changes and mineralogic variation was observed. Although both light and dark layers carry abundant (15-30%) recrystallised quartz, the darker layers are marked by a much higher interstitial opaque mineral component. In addition, they contain more chlorite and biotite than the light quartz-sericite bearing zones. It is unclear whether the systematic variation in opaque mineral concentrations is a

Figure 2:3:2 Fine-grained volcaniclastic rock showing alternation of quartz-rich and chlorite-opaque-rich bands. The outcrop is on East Folly River approximately 20 m north of the powerline (see Plate 1).

relict sedimentary feature but grainsize grading in East Folly River exposures confirms a clastic sedimentary origin for the rocks at this location.

Flows and Breccias

Outcrops in the Debert River area, on a short section of the Upper East Folly River and on East Folly Mountain exhibit moderately to relatively unstrained flow and breccia textures. In the southern part of Debert River area altered breccias, with angular clasts ranging from 5 to 60 cm in section, predominate in the most northerly exposures. Those to the south are more massive, characterised by fracture associated epidote and carbonate. Curvilinear alteration or bleach zones, in association with volcanic breccia and cherty jasperitic iron formation are less common. Unfortunately, the isolated nature of individual outcrops along logging roads prevents study of specific variations in texture even over short distances, and the situation is further complicated by numerous undeformed, crosscutting diabase dikes.

Along the Upper East Folly River section essentially undeformed, spherical structures have been identified that may be the result of quenching (Figure 2:3:3). These exist a short distance from chloritic outcrops exhibiting irregular, closed, alteration rinds of 1.5 to 5 cm width which do not follow the metamorphic foliation (Figure 2:3:4). The spherical structures are interpreted as varioles, similar to those described by Furnes (1975), Jamieson (1977), and Carstens (1967) from submarine lavas and are considered attributable to extremely rapid

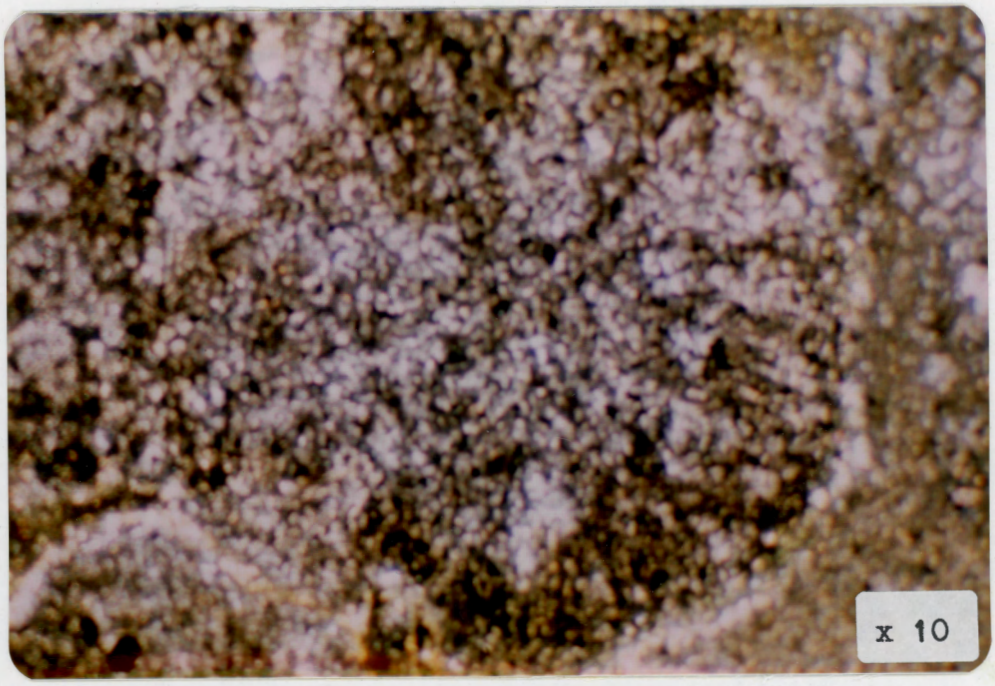


Figure 2:3:3 Quench spherules from a mafic flow unit exposed on the upper portion of East Folly River; note the poorly defined radial structure described by fine grained plagioclase and opaques. Orthogonal slabs of the specimen indicate only slight flattening. (crossed polars)

cooling of lavas by seawater. The alteration rinds illustrated in Figure 2:3:4 are believed to be selvages on irregular pillow structures also indicative of submarine volcanism. At one location on East Folly Mountain moderately well preserved pillows occur in two lowlying outcrops and exhibit markedly bleached and texturally distinct selvages. Rims are characterised by irregularly spaced and radially distributed joints and are demonstratably finer grained than interior zones (Figure 2:3:5). These features are considered strongly indicative of the presence of subaqueous volcanic flows within the supracrustal sequence.

Coarse Pyroclastic Rocks

Coarse grained, fragmental volcanic rocks occur in discontinuous outcrops along the Folly River for approximately 120 meters north of the powerline (Plate 1, Figure 2:3:3). This rock type can be traced westward for at least 175 meters to a point west of the Canadian National Railway track but to the east it is lost beneath overburden on the valley side. The next along-strike outcrops to the east are those of deformed diabase on the logging road south from Folly Lake Quarry.

Hand-specimens and thin-sections from this unit reveal a great range in clast size but a marked consistency of clast composition. Coarse, subangular clasts of chloritic mafic volcanic material and diabase, as large as 25 cm in exposed section, were noted approximately 75 m north of the powerline on the river section. These occur as grain-supported individuals in a finer, ash-clast aggregate with mean clast size between 2.5 and 0.5 mm. At one location on Folly River (80 m north

Figure 2:3:4 Contorted, curvilinear alteration "rinds" in a highly altered mafic volcanic flow unit on East Folly River , north of the powerline (see Plate 1). North is to the left.

Figure 2:3:5 Flattened mafic pillow structures exposed on the East Folly Mountain logging road. Spaced radial fractures in the selvage bleached zones is characteristic. The regional metamorphic foliation dips southward here at a slightly higher angle than the outcrop surface. The hammer handle is oriented approximately east-west.

of the powerline crossing) light-pinkish-grey fragments of rhyolite were noted as well.The presence of rhyolite fragments in the unit may suggest that more felsic volcanic material exists at a lower or laterally equivalent level within the stratigraphic pile. Very fine-grained ash interbeds also occur within the otherwise coarse fragmental material. At the powerline crossing on Folly River grainsize grading over approximately 12 centimeters indicates that strata young southward at this level in the metavolcanic sequence.

Mafic Agglomerate

This distinct rock type occurs in two short outcrop sections on east Folly River and in single exposures on both Folly and Debert rivers. Agglomerates are characterised by amygdaloidal, oblate clasts of medium to fine grained, chloritic metavolcanic material. Clasts range in size on the foliation surface from 10 cm x 3 cm x 1.5 cm to 3 cm x 7 cm x 5 cm and exhibit an extremely consistent elongate shape fabric (Figure 2:3:6). The best exposures are in the East Folly River zone where large ($\leq$ 1.5 m) boulders of agglomerate occur in the stream bed for approximately 70 m. The source of the boulders can be traced to a stratigraphic layer along the west bank near the valley rim. The agglomerate is not seen in outcrop at stream level and its boulders cease to be found immediately to the north in a canyon of deformed diabase.

Amygdules of calcite and chloritic knots as large as 4 mm in cross-section, are common in most clasts. The matrix consists of finer grained, chloritised ash and detritus. Leaching of carbonate and reces-

Figure 2:3:6 Mafic agglomerate boulder from the East Folly River section north of the powerline (see Figure 2:3:1); note the preferred planar orientation of the clasts. Cobble-size clasts rest in a matrix of chloritised ash and coarse-grained volcanic detritus. North is to the left.

sive weathering of some chloritic blebs results in a characteristic pitted clast surface.

Rhyolite

Light purplish to cream colored rhyolite that exhibits locally tuffaceous texture and Liesegang weathering features characterises this subunit. These outcrop in a narrow (< 175 m) east-west belt immediately west of the Shatter Brook - Debert River confluence and along Shatter Brook approximately 150 m from the stream junction (Plate I, Figure 2:3:3). The contact with the mafic metavolcanic sequence west of the Debert River is not exposed but in the Shatter Brook section a steeply south-dipping, east-west-trending mylonite zone marks this boundary. The felsic volcanics carry a distinct tectonic foliation adjacent to the mylonite zone and the mafic volcanic rocks of the hanging wall are pervasively foliated as well.

Grouping of the rhyolites within the metavolcanic unit of the Bass River Complex is based primarily upon the nature of the exposed contact and their structural position south of the Rockland Brook Fault (Plate 1, Figure 2:3:3). It is possible, however, that these rocks represent a structural slice of the Byers Brook Formation, a Devonian-Carboniferous felsic volcanic sequence described by Donohoe and Wallace (1980). The occurrence of these volcanics between the deformed Shatter Brook Pluton and the greenschists, isolated from the Byers Brook Formation volcanic terrane to the north, favors the present grouping within the Folly River Schist. A second line of reasoning to support grouping with the Folly

River Schist lies in the presence of felsic fragments in the previously-described coarse pyroclastic unit. These clasts and the rhyolite may be related to the same volcanic event.

Iron Formation

Cherty, jasperitic iron formation occurs within the metavolcanic unit as detached lenticles parallel to the dominant tectonic foliation. It probably also occurs as relatively thick (to 35 cm at least) stratigraphic units since boulders of this size were located along a logging road in the Debert River area.

Centimetre-scale jasperitic lenses are commonly seen intrafolially within the greenschists on the northern logging roads of the Debert River area and along the lower segment of Shatter Brook. Lenses range in size from discontinuous, 3 to 5 mm thick structures less than 10 cm long, to isolated , 25 cm thick boudins up to 1m in length.

Although most commonly seen within the greenschists, lenses and thin interlayers of iron formation were also detected in bedded, quarzitic metavolcaniclastic or sedimentary rocks on the East Folly River and in phyllites on Shatter Brook (Figure 2:3:7).

A detailed study of this lithotype was not attempted but two types of iron formation were recognised on the basis of field observations and examination of available thin sections:

(1) massive, inconsistently layered jasperite and white quartz $\pm$ fine

Figure 2:3:7 Jasperitic iron formation interbedded with quartz-rich volcaniclastic rocks on East Folly River. This exposure is approximately 150 m upstream from the confluence with Folly River. Well defined, cm-scale bedding is characteristic of the volcaniclastic units and grain-size grading is common. Bedding strikes easterly , dips at a low angle to the south and is right-side up. Photo faces southward.

grained magnetite;

(2) interlayered jasperite, massive specularite, magnetite, carbonate and mafic volcanic material (with tiny, circular section, spherule-like structures characterising the jasperite).

No lithologic association was identified which could be considered exclusively characteristic of either type of iron formation.

The discontinuous and relatively thin nature of the iron formation in the Folly River Schist limits its use as a stratigraphic marker but a large-scale distinction in regard to distribution can be made. Outcrops containing iron formation are confined to 2 northeasterly-trending bands at separate structural levels within the deformed volcanic pile (Figure 2:3:8). The southern zone is defined by only a few outcrops of the quartzitic metasedimentary or metavolcanic material whereas that to the north is characterised by the association of iron formation with chloritic schists or massive deformed volcanics. Both zones are developed along East Folly River but only the northern one is present in the Debert River area.

Pyritic Schists

Strongly foliated sericite-quartz-feldspar protomylonite and schist, containing as much as 70% pyrite, occur as boulders in scree slopes along some of the rivers crossing the metavolcanic sequence. Although outcrop verification was not possible, similarity of fabric and metamorphic grade and the obvious restriction of source area indicate that these are associated with the Bass River Complex metavolcanics.

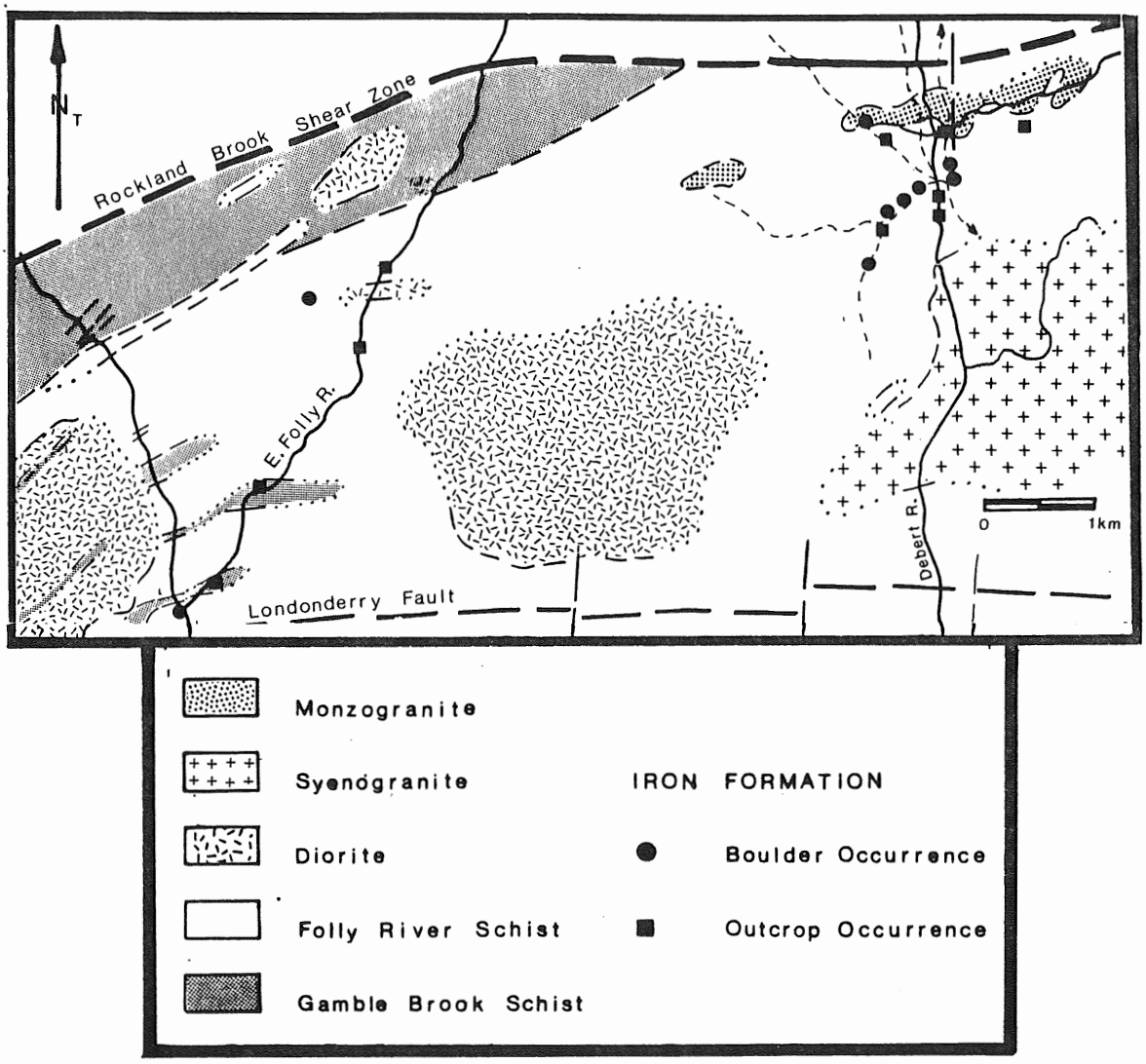


Figure 2:3:8 Distribution of Iron Formation occurrences in the Folly River Schist.

In thin-section the sulphide phase is concentrated along the penetrative metamorphic fabric and as interstitial disseminations between strained and recrystallised quartz, feldspar grains, and sercite (Figure 2:3:9). The highly strained nature of the sample precludes identification of primary features and therefore timing of the mineralisation is unclear. Modification by strain of a syngenetic sulphide concentration within a felsic tuffaceous volcanic or volcaniclastic zone is the favoured interpretation at present but cannot be proven.

2:3:2 Gamble Brook Schist Unit

General

Donohoe and Wallace (1980) recognised a metasedimentary unit composed of interlayered deformed quartzites, quartz - biotite schists, and biotite - chlorite schists, which they termed the Gamble Brook Schist and which Eisbacher (1967) grouped within the "mylonite schist zone" of his study. It occurs north of the Folly River Schist metavolcanic sequence. These rocks extend west-southwesterly from East Folly River to the area between Portapique and Bass Rivers, reaching a maximum width of 1.6 km in the Portapique River - Gamble Lake area. The zone reaches almost the same width in the Great Village River - Higgins Mountain Road area where the quartzites extend northward above the Great Village River Gneiss (Plates 1 and 2).

Based on the findings of this study a re-interpretation of the

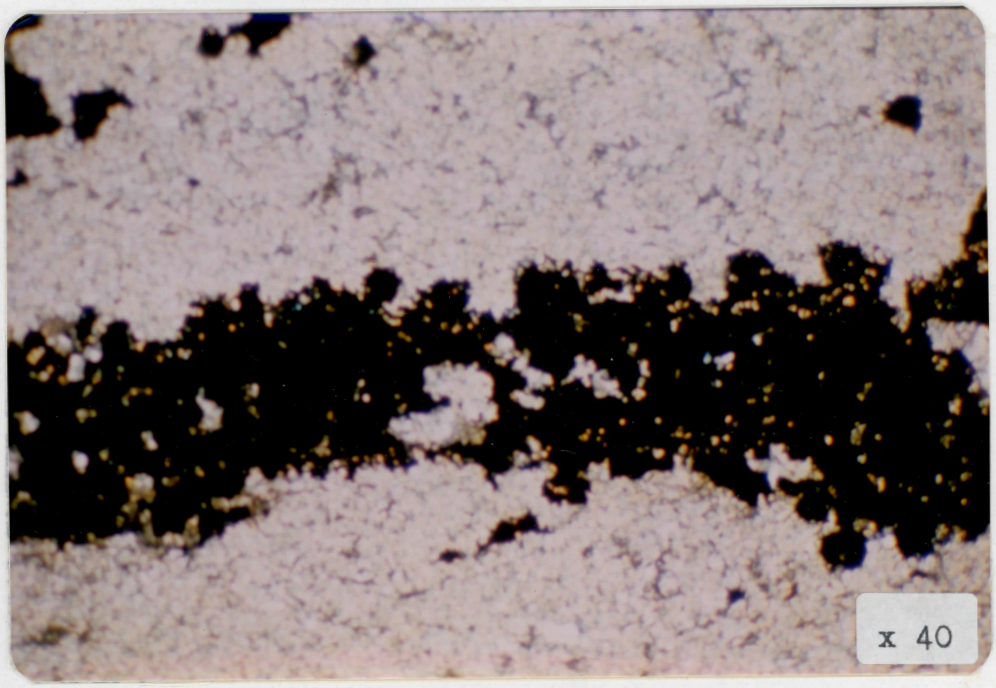


Figure 2:3:9 Sericitic quartzo-feldspathic schist with a pyrite interlayer; note the close association of fine-grained chlorite with the pyrite. Recrystallisation of the sulphide phase is inferred from the fabric overgrowth relationship. The sample was collected from a scree slope on the lower portion of Folly River, north of the East Folly River confluence. (plane light)

Gamble Brook Schist outcrop area was necessary. A discontinuous zone of metasedimentary rocks mapped along the lower reaches of East Folly and Folly Rivers is correlated with the Gamble Brook Schist on the basis of similarities of composition, metamorphic grade, and deformation history. The southern belt is much narrower than that to the north, reaching a maximum width of 250 m on East Folly River. To the west it is intruded by the McElman Brook Pluton and may be tectonically interlayered with rocks of the Great Village River Gneiss. Rocks of the southern belt do not outcrop west of the McElman Brook intrusion.

Lithologies

The Gamble Brook Schist is characterised by light tan to white weathering, medium- to fine-grained quartzite and biotite - garnet quartzite, interlayered with biotite - quartz - chlorite - garnet - schist (Figure 2:3:10). In a few locations thin (< 25 cm thick) zones of diopside- bearing carbonate occur as bedding-parallel interlayers which probably define primary compositional variations. The carbonates weather recessively relative to both schists and quartzites and are characterised by light tan or buff colored, pitted surfaces (Figure 2:3:11). On East Folly River jasperitic iron formation identical to that seen in the Folly River Schist accompanies the tan carbonate in thin (<15 cm thick) layers parallel to the quartz - biotite - chlorite defined metamorphic fabric.

Composition changes and sedimentary structures are moderately well represented in some low strain areas within the unit although in

Figure 2:3:10 Very tightly folded and interlayered, light grey quartzite and dark brown, fine-grained, biotite-chlorite schist of the Gamble Brook Schist. The transposed layering strikes east-northeasterly and dips steeply to the south. The outcrop is on the east side of Folly Lake Quarry (see Figure 2:3:1 for quarry location).

Figure 2:3:11 Massive, fine-grained carbonate layer occurring within light grey quartzites of the Gamble Brook Schist. The outcrop is located at the north end of Folly Lake Quarry (see Figure 2:3:1 for quarry location).

most cases recrystallisation of quartz has obscured original grain sizes A complete compositional and textural series from clean quartzite to biotite - quartz - garnet schist is apparent over the width of the sequence but quartzite and biotite- garnet quartzite predominate in the northern belt. In the south, fine grained quartzite, biotite + chlorite + garnet schist and fine grained chlorite - biotite schist are most common.

Quartzites are characterised by a distinct parting parallel to the penetrative foliation and, as a result of competency and an approximately orthogonal joint set at a high angle to the foliation and layering, most exposures are blocky and marked by "stepped" surfaces. The parting occurs on both metre and centimetre scales depending, probably, upon strength of the protolith and fabric intensity. Metasedimentary rocks with abundant biotite and chlorite are more recessive in weathering style than the psammites and generally show a more pervasive parting trend.

Predominance of clean quartzite in the northern exposures relative to those in the south, combined with the presence of minor amounts of iron formation in the southern belt, may indicate that the southern rocks represent a transition zone between the predominantly metavolcanic Folly River Schist and the clean northern belt of the Gamble Brook Schist. Whether this is a lateral or vertical transition is difficult to assess because transposition by the first deformation to affect these rocks complicates stratigraphic evaluation.

2:4 Intrusive Rocks

2:4:1 Introduction

Plutonic rocks important to this study have been classified according to Streckeisen (1976) and divided into three groups:

(1) those known or inferred to be Precambrian in age;
(2) those known or inferred to be Early Paleozoic (Cambrian) in age;
(3) those known to be Devonian to Carboniferous in age.

Precambrian intrusions of Group 1 include Strongly Foliated Granitoids and nonpervasively deformed diorites, both of which are characterised by intrusive or strain-modified intrusive contacts with host Bass River Complex rocks.

Early Paleozoic syenogranites (Group 2) occur only in the eastern Highlands. The Shatter Brook Pluton (Donohoe and Wallace 1980) is entirely within the Folly River Schist terrane whereas the McCallum Settlement Pluton intrudes the Folly River Schist east of the study area.

Devonian to Carboniferous age intrusions underlie much of the central Cobequid Highlands area and are, in general, in contact with Bass River Complex rocks along the northern Complex boundary. In the Debert River area intrusions of this age occur along the southern boundary as well and also define the eastern termination of most of the metamorphic terrane. In all cases intrusive relationships with the metamorphic rocks, or with rocks known to be intrusive into them, have

been observed. Some contacts, particularly those in the Debert River valley, have been deformed but sufficient crosscutting evidence exists in each case to define the intrusive relationship. A combination of radiometric age and field observations of crosscutting relationships allows good absolute and relative age control for most of these rocks and all were emplaced in Late Devonian to early Carboniferous time (Clarke et al., 1980).

Mapping extended into all of the post-Precambrian plutons but work beyond that necessary to outline contact relationships with Bass River Complex rocks was not done. Those of Precambrian age, Group 1, were more thoroughly investigated. In the following sections rocks of each major group are reviewed with emphasis placed upon those which occur within the confines of the Bass River Complex.

2:4:2 Precambrian Intrusions - Group 1

2:4:2(a) Strongly Foliated Granitoid Rocks

General Outline of Field Relationships

On each major stream section west of Debert River strongly foliated to mylonitic granitoid rocks occur interlayered with those of the meta-sedimentary and metaigneous units. Donohoe and Wallace (1980) describe these as granite gneisses. They range from zones marked by numerous, relatively thin, deformed veins (less than 2 m thick) such as on Bass River, to river-section scale bodies such as that of Economy or East

Folly River.

A strongly porphyroclastic, biotite - garnet grade, mylonitic to protomylonitic fabric characterises most exposures of these granitoids but to date, no pre-existing metamorphic fabric has been identified. Their emplacement prior to, or during, the Complex-wide regional-foliation- related tectonic event is indicated by general parallelism of lithologic boundaries with the dominant fabric in the host rocks and the presence of a pervasive, internal fabric coplanar to that of the host. On the upper Rockland Brook section a syn-deformational emplacement period is indicated by the discordance of granitoid vein boundaries to the host amphibolites' foliation which can be traced across the contact into the vein itself (Figure 2:4:1). The nature of this contact and the presence of multiply interlayered veins generally parallel to the foliation, but crosscutting it at low to moderate angles, indicate a strain-modified intrusive relationship between these rocks and their hosts.

In some areas tectonic shortening post-dated vein emplacement, resulting in close repetition of granitoid veins across the axial planar foliation. Non-parallelism of these veins with the foliation may simply be a function of their initial orientation in the deformation field. However, the variable fabric intensity across the compositional contacts indicates that the granitoids have not been as highly strained and metamorphosed as the host. They therefore may have been emplaced slightly later than the deformation-metamorphism peak with which the

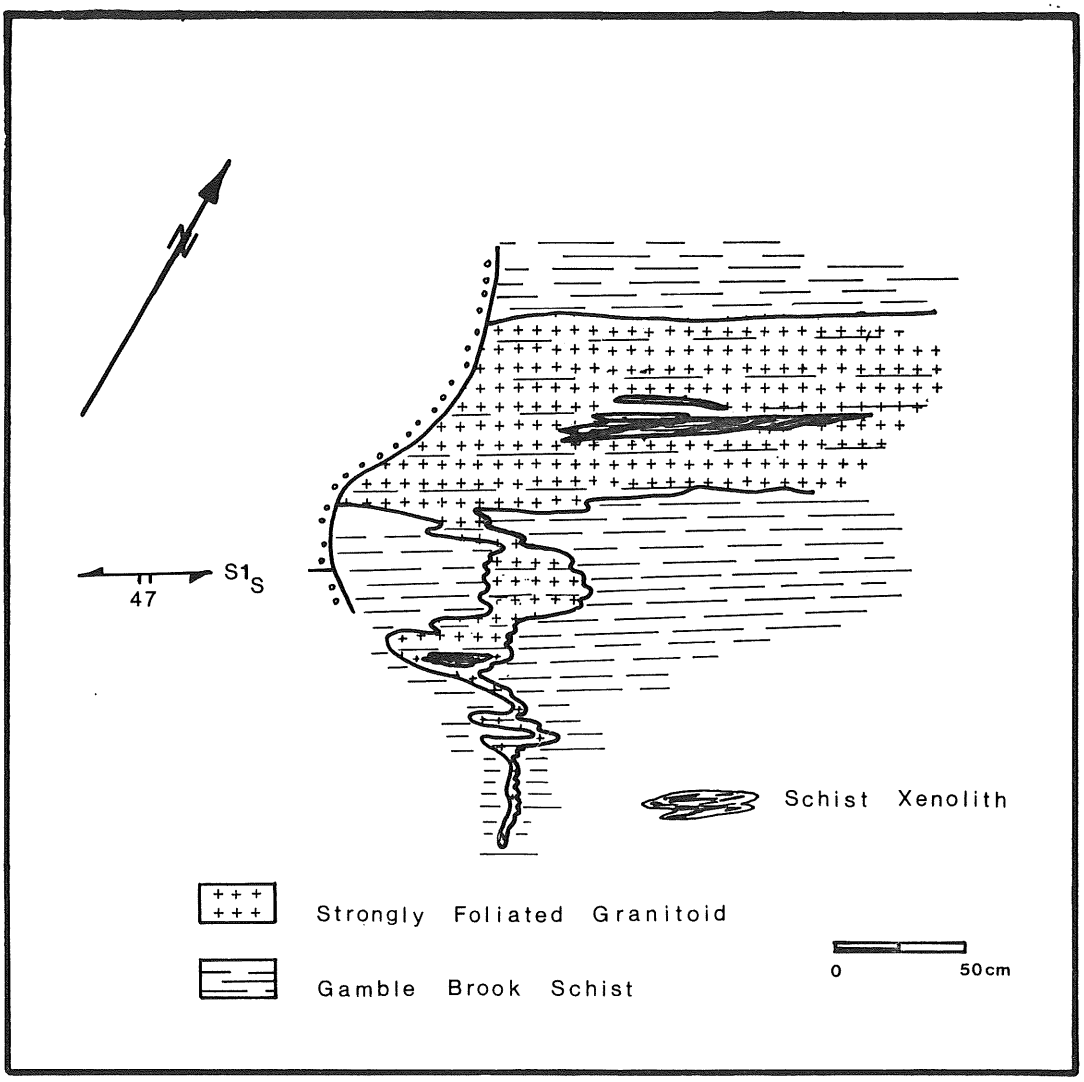


Figure 2:4:1 Host-vein relationship between Strongly Foliated Granitoid intrusive rocks and quartz-biotite schist of the Gamble Brook Schist. The host shows much stronger fabric development than the intrusion but $S1_S$ is clearly continuous across the vein contacts. The granitoid is inferred to have been emplaced during deformation.

regional foliation is associated. No evidence of superimposed shearing is evident at the location of Figure 2:4:1 which is typical of this type of contact relationship.

Lithology

Medium-grained, porphyroclastic, orthoclase - quartz - biotite - muscovite mylonites and protomylonites characterise this group of intrusions. Grain size reduction and rounding of feldspars and quartz has been extensive with both C and S plane fabric components (Berthe et al., 1979) visible in many cases. Segregation of the coarser, porphyroclastic material from finer-grained constituents produces a poor to moderate textural layering visible on both hand-specimen and thin-section scales ($\leq$ 2 mm in thickness). Both pyroxene and hornblende occurred as accessory minerals in the protolith (< 10%) and show varying degrees of grain size reduction by fracturing and fragmentation in a more ductile matrix. Crystallisation of biotite accompanied tectonism for generally idioblastic biotite and muscovite occur within the foliation.

Distribution

The Strongly Foliated Granitoid rocks crop out at twelve different locations within the Complex (Plate 1). As noted previously, they vary in size and nature from zones marked by numerous relatively thin, deformed veins , to large-scale bodies . In the Portapique - Great Village River - Rockland Brook area sufficient geologic control exists to infer direct

correlation of the bodies between rivers but continuous extension of the zone to both east and west seems unlikely . The other zones cannot be directly correlated across interstream areas but their consistent occurrence in two spatially-distinct east-west zones (Plate I), combined with similarities of mineralogy and deformation history, strongly indicates their close association.

Age

Whole-rock radiometric dating of Strongly Foliated Granitoid rocks on Great Village, Economy, and Portapique Rivers (Gaudette et al., 1983, see Figure 2:4:2) showed a large spread in calculated ages. These ranged from circa 810 Ma, for the Portapique River sample, to 370 Ma for that from the Economy River. The Great Village River intrusion was determined to be 690 Ma old but Donohoe and Cullen (1983) suggested that this may represent a reset age. At present their suggestion of resetting by thermal aureole affects of a dibase dike swarm seems somewhat improbable due to a lack of associated anatectic features. The possibility of resetting by localised Paleozoic shearing cannot be eliminated, however. Without benefit of the actual isochrons and more specific details of the sample locations , it seems wise to treat this information with a certain degree of caution. The circa 810 Ma age for the Portapique River intrusion appears to be the most reliable and may place a reasonable minimum age on that body. Considering the similarities of texture, composition, and deformation history existing within this group of intrusions it is suggested that all of the Strongly

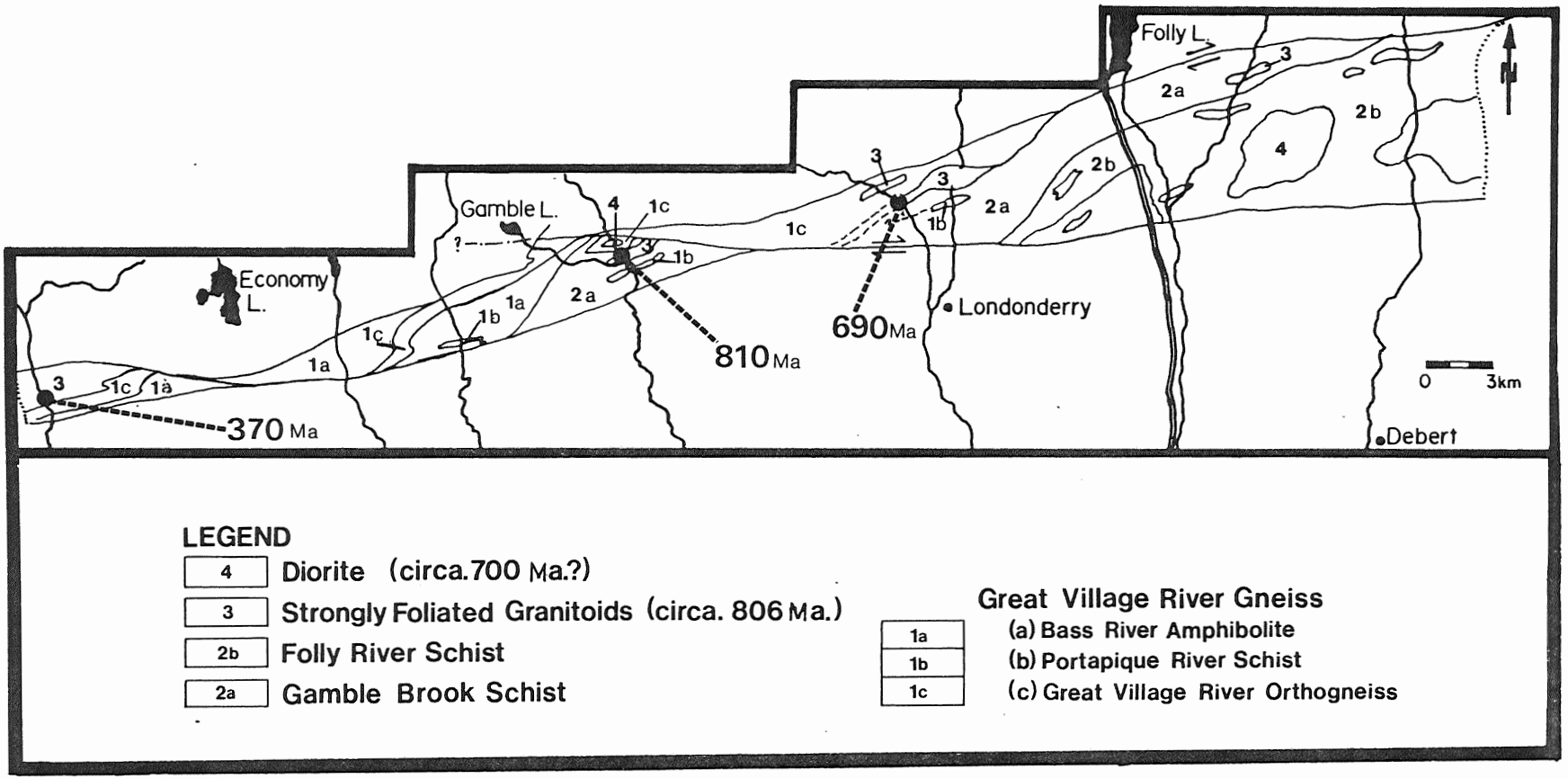


Figure 2:4:3 Location map for radiometric age samples discussed by Gaudette et al. (1983).

Foliated Granitoid bodies are related. If the radiometric age reported by Gaudette et al. (1983) for the Portapique River site is reliable, intrusion must have taken in late-Precambrian time.

2:4:2(b) Diorite Association

General Outline

Numerous distinct diorite bodies have been identified within the Bass River Complex and included in the Precambrian age group. All are essentially undeformed intrusions locally cut by shear zones. Emplacement after development of the regional metamorphic fabric is evident from the presence of foliated metamorphic roof pendants in nonfoliated zones of the intrusions. This is most apparent in the McElman Brook Pluton (see Plate 1). Post-regional-folding intrusion is indicated by truncation of the northeast to east-northeast regional structural strike by the diorites (Plate 2).

The age of these post-regional metamorphism intrusions is difficult to ascertain except in relative terms. Their deformation by Middle Devonian to Early Carboniferous shearing (see chapter 3, sections 3:5 and 3:6) places a minimum constraint on their emplacement age and their position in regard to regional folding suggests a maximum Late Precambrian age (i.e., postdating the last regional folding of circa 810 Ma Strongly Foliated Granitoids)

These diorites may be of the same association as the post-metamorphic Jeffers Brook Pluton granodiorite that intrudes Jeffers-Warwick

Formation (Donohoe & Wallace 1980) greenschist volcanics in the western Cobequids near Parrsboro. This intrusion has been dated as circa 600 Ma old (K/Ar by Wanless et al., 1969; Rb/Sr whole-rock by Cormier, 1980) and the Jeffers-Warwick Formation rocks are believed to be younger than any of those in the Bass River Complex. Jamieson (in press) documented a period of late Precambrian diorite emplacement in the Cape Breton Highlands and Krogh et al. (1983) outline a similar trend in the Avalon Zone of southeastern Newfoundland. In this light it seems reasonable to suggest that similar-aged dioritic igneous activity could have affected other portions of the Avalon Zone, such as the Cobequid Highlands . Final resolution of this age-affinity question will come, of course, only after reliable radiometric dating of the Cobequid diorites.

The largest diorite body inferred to be of late Precambrian age, the McElman Brook Pluton, is described below and all intrusions included in this group are indicated on Plate 1.

McElman Brook Pluton

This intrusion occurs between the Folly River and Rockland Brook (Plate 1) and is exposed primarily along logging roads on Folly Mountain. Short sections also outcrop on Folly River, McElman Brook, and along the Trans Canada Highway. The outline presented in Plate 1 is based on a combination of outcrop distribution patterns and aeromagnetic trends in Plate 5.

The pluton is a generally equigranular, medium grained, hornblende-

plagioclase diorite with less than 10 percent quartz and carries no pervasively developed tectonic fabric. Idioblastic hornblende grains up to 3 mm in length occur in a fine grained matrix of equigranular, intergrown plagioclase and quartz with small amounts of opaque material (< 5%). Amphibole is chloritized preferentially along grain boundaries and cleavage traces (Figures 2:4:3) with sericite commonly replacing plagioclase feldspars. Quartz grains show evidence of strain such as undulose extinction and locally-developed subgrain textures. Locallised shear-related fabrics do occur and are defined by preferred planar and linear orientations of chloritised hornblende and altered feldspar. These small shears are generally less than 1 m in width and are oriented at low to moderate angles to the east-northeast regional structual trend of the Complex. In zones affected by shearing strain gradients can usually be observed from undeformed diorite to a strongly foliated equivalent (Figure 2:4:4).

The existence of a large roof pendant west of Highway 104 (Plate1), combined with the truncation of the structural grain by the diorite's eastern margin, suggests that emplacement of the body took place after the latest regional folding to affect the metamorphic sequences.

2:4:3 Lower Paleozoic Plutonic Rocks - Group 2

Introduction and Field Relationships

Two granitoid plutons occur within or adjacent to the Bass River Complex which show no marked similarities to the Devonian - Carboniferous

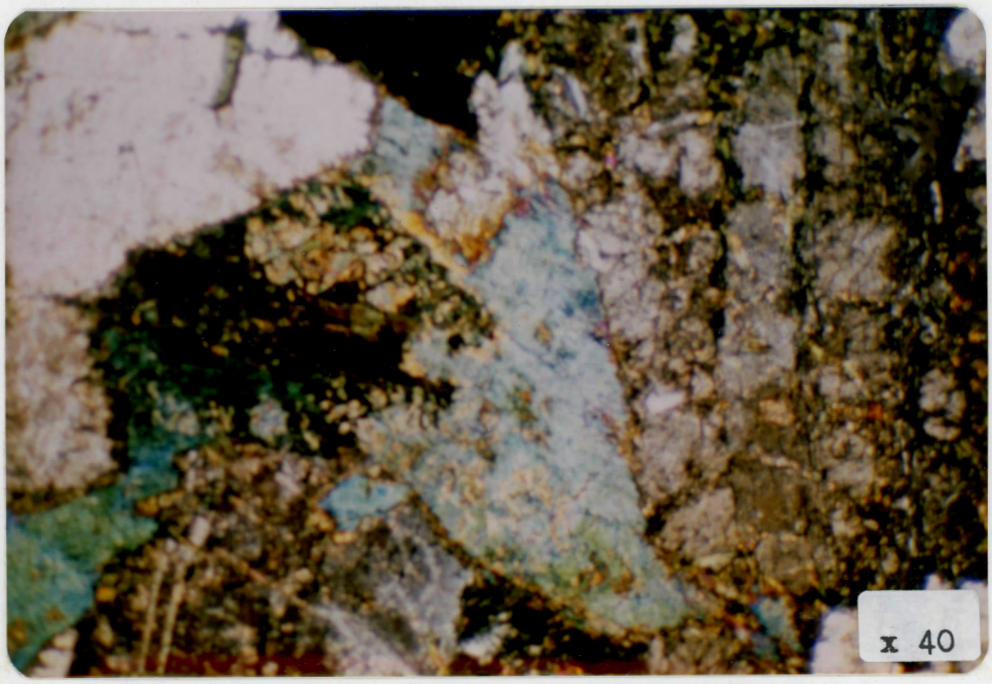


Figure 2:4:3 McElman Brook Pluton diorite; note the quartz inclusions in the hornblende and chloritic alteration, particularly along grain boundaries. Intergrown plagioclase shows sericitic alteration. The sample was collected from an exposure on a logging road approximately 1 km west of the Trans Canada Highway, near McElman Brook. (crossed polars)

Figure 2:4:4 A small shear within the non-pervasively deformed McElman Brook Pluton. The fine, equigranular textures at the top and bottom of the photo are progressively modified toward the shear trace (2.5 cm above the knife blade) where a thin zone of dark grey schist occurs. The outcrop is on a logging road east of Highway 104 near Slack Lake. Shearing is attributed to movement on the Gamble Lake and/or Rockland Brook faults. North is to the left and the foliation dips vertically.

association or to the Precambrian intrusions . Both the Shatter Brook and McCallum Settlement bodies have a syenogranitic composition (50-60% perthitic feldspar; 20-25% quartz; 10-15% Na plagioclase) and are characterised by low mafic mineral content (<10%). Lack of a pervasive tectonic fabric and presence of clearly intrusive contacts with the Folly River Schist are typical.

Age

Cormier (1980) reported a 504 ± 27 Ma Late Cambrian - Lower Ordovician whole-rock radiometric age for the McCallum Settlement Pluton. This determination is helpful in qualifying relative ages of other rocks in the area, especially the Shatter Brook Pluton. Veins crosscutting the regional metamorphic fabric adjacent to the Shatter Brook Pluton are not common but do occur at one location on Shatter Brook, approximately 100m east of the Byers Lake Road. Small veinlets cut the fabric of the host schist and originate in the underlying intrusion. Grainsize change toward the host is not definitive on the Shatter Brook section but an exposure on a logging road west of the Debert River bears foliation - parallel, medium-grained granitic veins which show obvious chilling over 2 cm toward the metavolcanic host.

About 500 m further due west of the Shatter Brook-Debert River confluence, interlayered foliated granite and metavolcanic material outcrop along a logging road. No unambiguous intrusive contact is apparent at this location but shearing is localised to within a few metres of the granite contact and may indicate structural modification

of an originally intrusive relationship.

The Shatter Brook Pluton was emplaced after the deformation that produced the pervasive metamorphic foliation of the Bass River Complex and probably after the subsequent regional folding episode because the western end of the pluton abruptly truncates this structural trend. Texturally and mineralogically this pluton differs from the Strongly Foliated Granitoids (i.e., low mafic mineral content and no pervasive tectono-metamorphic fabric) but bears a strong resemblance to the light pink, mafic-poor McCallum Settlement Granite (circa 504 Ma; Cormier, 1980) which also shows similar crosscutting relationships. Correlation of the two intrusions allows grouping with the Cambrian family of Avalon Zone intrusions discussed in Clarke et al. (1980).

2:4:4 Devonian to Carboniferous Plutons

Granitic and dioritic intrusions underlie large areas of the central and eastern Cobequid Highlands and their characteristics have been outlined in some detail by previous workers (Fletcher, 1905; Weeks, 1948; Eisbacher 1967, Donohoe and Wallace, 1980). Emplacement of the granitic phases typically post-dated diorite intrusion as evidenced by common diorite-granite vein stockwork field relationships (Figure 2:4:5). The Bass River Complex's northern boundary, the Rockland Brook Fault, locally brings rocks of the stockwork zone into fault contact with Bass River Complex metamorphic rocks but clear intrusive relationships have also been noted. Intrusions of Devonian to Carboniferous age

Figure 2:4:4 Devonian-Carboniferous "stockwork" or contact zone rocks exposed along the west side of Highway 104 at the south end of Folly Lake. At this location Hart Lake-Byers Lake Pluton granite has intruded Folly Lake Pluton diorite. The Rockland Brook Fault trace is approximately 175 m to the south.

are essentially undeformed except where cut by major shear zones, such as the Rockland Brook Fault, or by minor shears inferred to be associated with the larger-scale movement zones.

Donohoe and Wallace (1980) described the following Devonian to Carboniferous age intrusions that occur in the eastern Cobequid Highlands adjacent to the Bass River Complex:

(1) Hart Lake-Byers Lake Pluton (granite to syenogranite)
(2) Folly Mountain Pluton (hornblende diorite)
(3) Frog Lake Pluton hornblende diorite to granodiorite)
(4) Pleasant Hills Pluton (alkali feldspar granite).

Table 2:4:1 summarises important contact relationship, composition, and radiometric age information for each pluton. Locations of the intrusions referred to are presented in Plate 1.

DEVONIAN TO CARBONIFEROUS INTRUSIONS

Name of Intrusion	Rock Type	Mineralogy, Texture	Age (Ma)	Details	References
Pleasant Hills	Biotite-hornblende Syenogranite	Medium grained hypidiomorphic: 10-20% albite-oligoclase, 5-10%, biotite, 20-35% quartz, 35-60% K-feldspar, minor hornblende	315 ± 25	Rb-Sr whole rock age	1,3,4
Byers Lake and Hart Lake	Hornblende-biotite Syenogranite	Medium grained hypidiomorphic: 10-15% albite-oligoclase, 5-10% hornblende and biotite, 20-35% quartz, 35-55% alkali feldspar	331 ± 27 331 ± 17	Rb-Sr whole rock age	1,2,3
Debert River	Monzogranite to Granodiorite	Coarse grained hypidiomorphic: 40-55% oligoclase, 5-30% alkali feldspar, 20-30% quartz, ≤15% biotite, hornblende and chlorite	Carboniferous	Cut by Carboniferous Gain Brook Pluton	4
Economy River	Diorite	Medium to fine grained diabasic to equigranular: 55% oligoclase-andesine, 10-45% hornblende, 0-15% clinopyroxine, ≤30% chlorite (after mafics)	Devonian to Carboniferous	Cut by Pleasant Hills Pluton Cuts Silurian strata	4
Folly	Diorite			Cut by Hart Lake-Byers Lake Pluton	4

References: (1) Donohoe (1976) (2) Wallace and Donohoe (1977)
(3) Donohoe and Wallace (1978) (4) Donohoe and Wallace (Cobequid Highlands Memoir; in preparation)

Table is modified from Clarke *et al.* (1980)

TABLE 2:4:1

CHAPTER 3

STRUCTURAL GEOLOGY

3:1 General Introduction

In the first detailed consideration of deformation in rocks of the Bass River area, Eisbacher (1967, 1969) concluded that a single, ductile flow strain event produced the mappable deformation structures (other than north-south fractures) between the Londonderry Fault and the plutonic core of the Highlands. Later brittle fault movements were ascribed to a separate kinematic framework. In contrast, Donohoe and Wallace (1980, in preparation) recognised polyphase deformation in metamorphic rocks that they named the Bass River Complex and presented evidence for three locally-distinct phases. The findings of this study support the general hypothesis offered by Donohoe and Wallace and indicate that local to pervasive, sequential overprinting by six deformation phases occurred between Late Precambrian and Jurassic time.

The Complex has been divided into two structural domains. Domain "A" is composed of the Great Village River Gneiss and locally bears evidence of an early metamorphic fabric not seen in the Domain "S" Folly River Schist and Gamble Brook Schist supracrustal sequences. This basic distinction is interpreted as being the result of a basement - cover relationship existing between Domain A rocks (metamorphic basement) and Domain S rocks (volcanic and sedimentary cover sequences). Field

observations, augmented by thin- and slab-section studies of nearly 200 samples, form the basis for correlations outlined in this study.

Although six distinct events mark the deformation history, evidence of the first is seen only in rocks of Domain A. In the Domain S supracrustal rocks evidence of the second and third events is widespread but no indication of the earlier deformation is present. The remaining deformation signatures take the form of discrete, shear zones, fracture sets, and breccia zones related to either east-west or north-south faulting or to an inferred paleostress configuration of undetermined origin. Each of the six phases is discussed in this chapter with emphasis placed upon the spatial extent of deformation, fabric element associations, and the relative timing of deformation within the Complex and the Highlands.

Table 3:1 outlines terminology relevant to the discussion. For the sake of continuity with the terminology of previous workers (Donohoe, 1983; Donhoe and Wallace 1980, in preparation) a "B" designation (eg., DB1, DB2) has been used to distinguish the Bass River Complex deformation history from that of other rocks in the Highlands. "B" was derived from Bass River Complex. This nomenclature is consistent for consideration of Complex-wide fabrics such as foliations ("S" surfaces). When deformation features of each domain are being considered separately, a subscript system is used that indicates the domainal affinity (eg., DB2 of the regional terminology equates with $D2_A$ of Domain A and with $D1_S$ of Domain S).

CORRELATION OF DEFORMATION PHASES

	DEFORMATION	DOMAIN A Great Village River Gneiss	DOMAIN S Folly River Schist Gamble Brook Schist
Pervasive on Map Scale	DB1	$D1_A$	
	DB2	$D2_A$	$D1_S$
	DB3	$D3_A$	$D2_S$
Fault or Shear Zone Associated	DB4(a)	No distinction made between Structural Domains A and S	
	DB4(b)		
	DB4(c)		
	DB5		
	DB6		

Responsible for regionally pervasive metamorphic foliation S_{B2}*

*Fabric elements follow same subscript scheme as above

TABLE 3:1

3:2 The DB1 Event - $D1_A$ of Structural Domain A

Evidence of DB1 is confined to the highest-grade metamorphic terrane and is found in the earliest fabric identified in the Complex. This event is evident only from rare metamorphic fabric relicts which have not been totally overprinted and reoriented by subsequent dynamothermal events. Micro-scale as well as outcrop examples of the $S1_A$ fabric have been recognised in widely separate areas of Domain A (Figure 3:2:1) and occur in two general associations:

(1) as compositional layers in hinge zones of second generation isoclinal folds;

(2) as foliated metamorphic xenoliths in pre-DB2-peak intrusive rocks.

On Bass River both outcrop- and micro-scale second generation isoclinal folds ($F2_A$) are defined by $S1_A$ quartzo-feldspathic metamorphic layering and have the pervasive metamorphic fabric ($S2_A$) as an axial planar foliation (Figure 3:2:2a,b). The $S1_A$ layering - $S2_A$ relationship is evident in outcrop from serrate $S1_A$ layer boundaries defined by reoriented quartz and feldspar within the $S2_A$ plane. Extensive compositional differentiation by $S2_A$ has not occurred in the majority of recognisable hinges and the earlier fabric is easily discernible. Identical microscale relationships have been observed between the higher grade $S1_A$ and the crosscutting lower grade $S2_A$ axial surface foliation that characterises Portapique River Schist rocks. On the Economy River section strongly attenuated, isoclinal, intrafolial $F2_A$ folds outline a

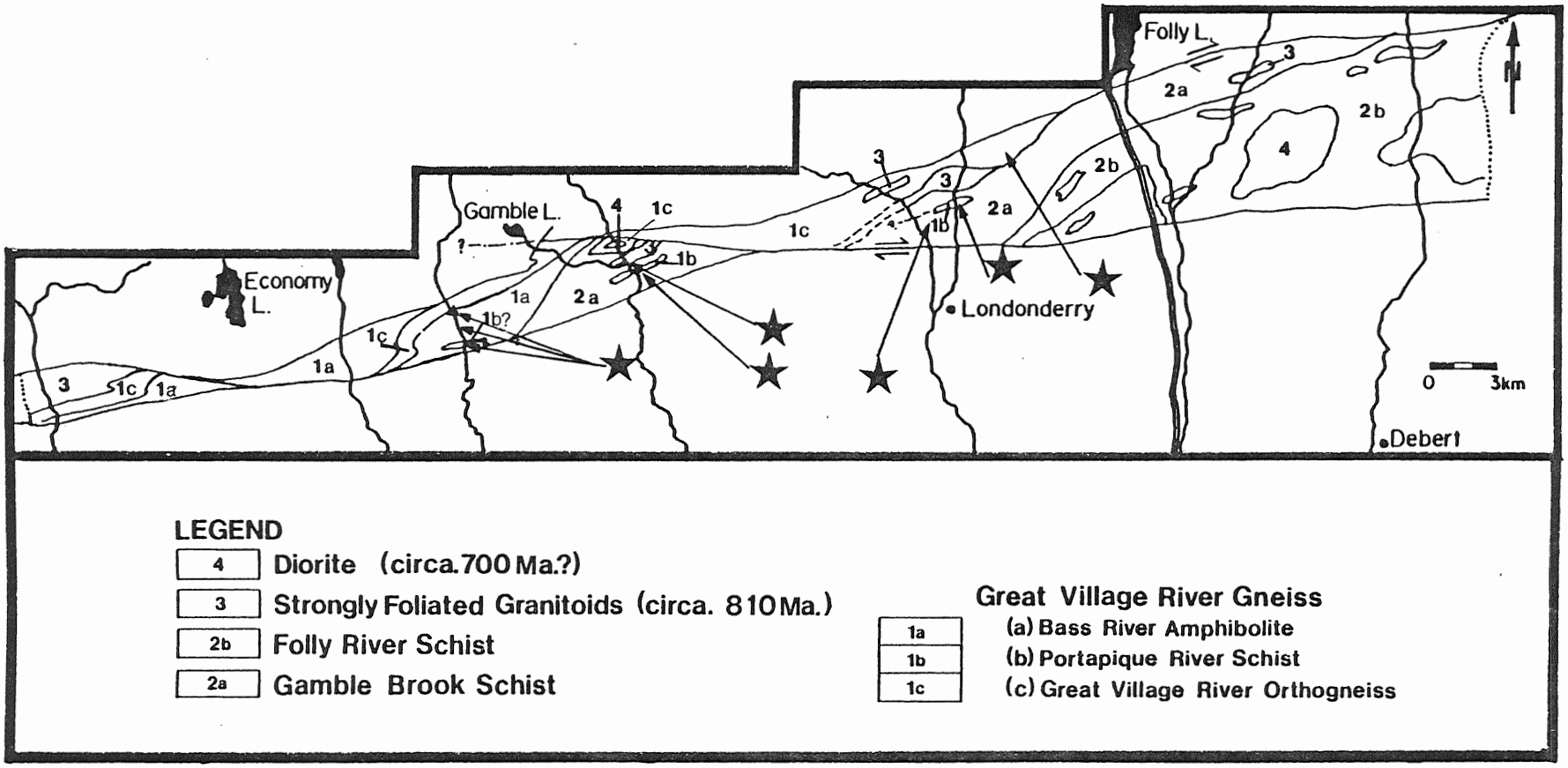


Figure 3:2:1 Location map for DB1 fabric examples (★).

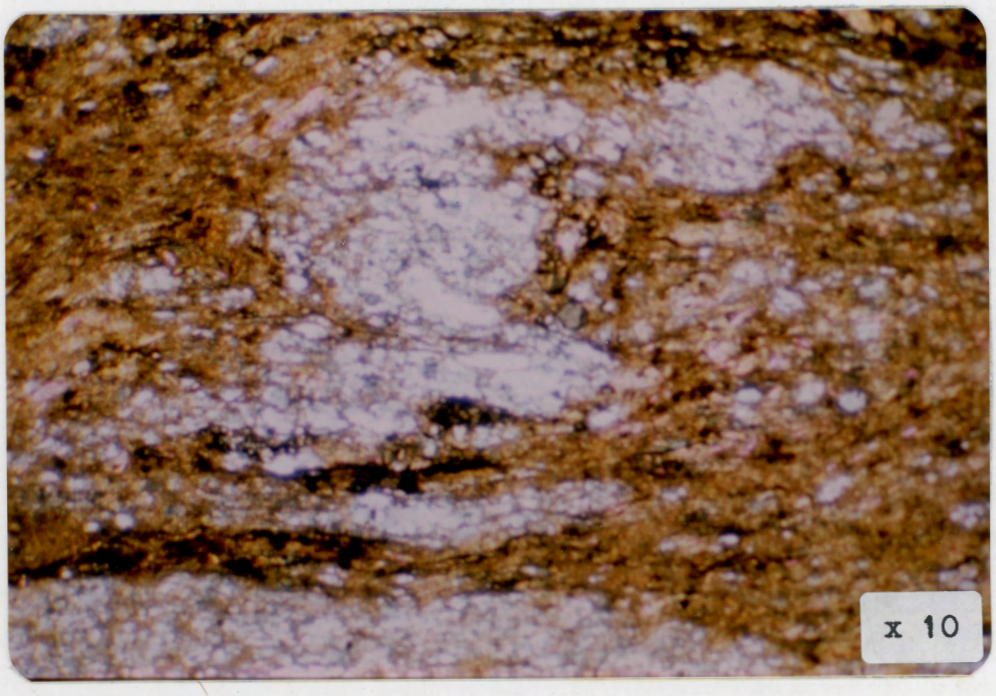


Figure 3:2:2(a) Microscale DB2 generation folds from the Portapique River Schist. Recrystallised quartz and occasional rounded feldspar grains define the early metamorphic layering which has been overprinted by biotite and muscovite of the DB2 axial planar fabric. Note the DB1 generation, inclusion-bearing staurolite grain in the uppermost fold hinge. The sample was collected from the Rockland Brook section. (plane light)

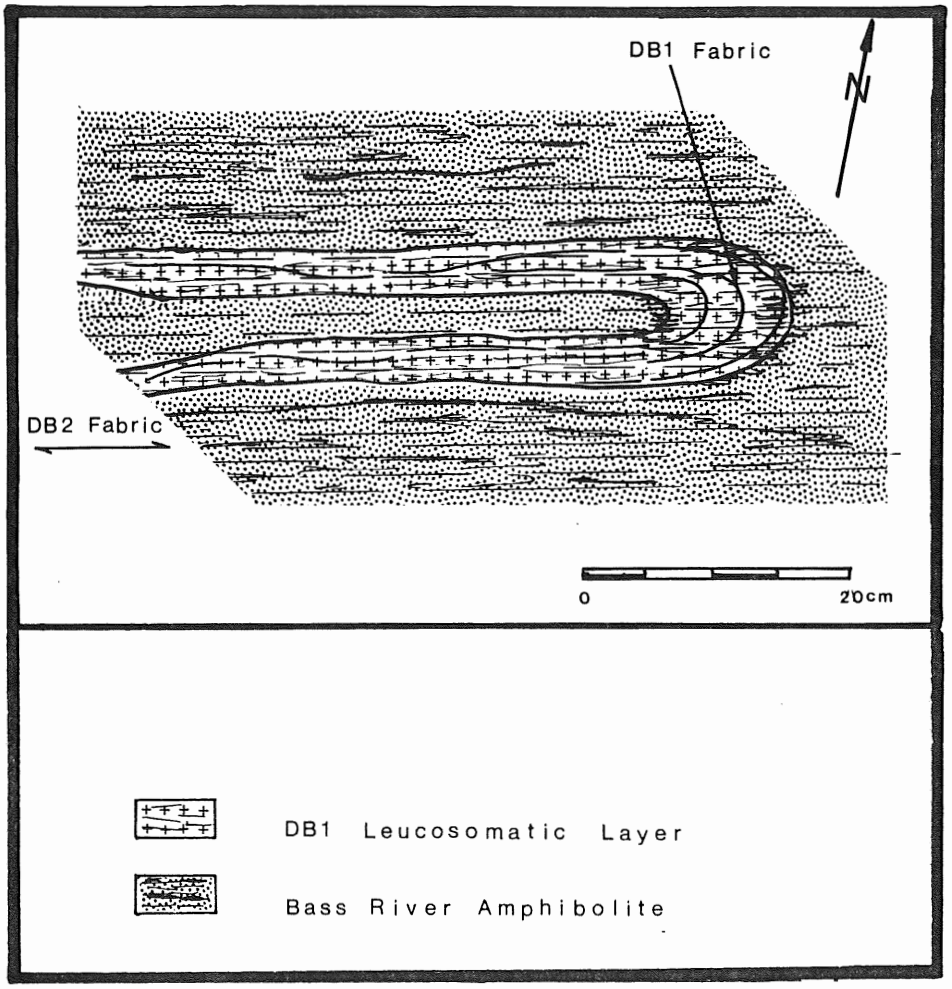


Figure 3:2:2 (b) Sketch of DB1 generation leucosomatic layering defining a rare DB2 generation fold closure. The overprinting fabric penetratively deforms the earlier layering but both fabrics are essentially parallel in most exposures. This example of DB1 layering occurs on Bass River approximately 425 m south of the Rockland Brook Fault.

similar overprinting relationship by having deformed thin, (<20 cm) $S1_A$ - bearing amphibolite layers within quartz-feldspar-biotite-hornblende tectonite of the Great Village River Gneiss. Donohoe (oral communication, 1982) identified at least two other locations at which $S1_A$ - equivalent metamorphic layering is evident.

Fabric relics of the xenolith type are not common but have been identified at two locations within Domain A. Paragneiss and/or amphibolite xenoliths exhibiting the first deformation's fabric ($S1_A$) occur in a pre-DB2-peak granitoid intrusion on Portapique River near the confluence with Gamble Brook. The DB2 fabric is penetratively developed in both xenoliths and hosts, being marked by a crenulation of the earlier DB1 fabric within suitably oriented xenoliths (Figure 3:2:3).

3:3 The DB2 Event - $D2_A$ in Structural Domain A

- $D1_S$ in Structural Domain S

3:3:1 Domain A Structures

The DB2 event developed the strongest tectono-metamorphic fabric seen in the Bass River Complex. Intense shortening and transposition were accompanied by greenschist to amphibolite facies metamorphism and possibly syntectonic emplacement of granitoid bodies. Folds of this generation are visually tight to isoclinal, noncylindrical structures which show extensive limb attenuation and/or hinge detachment within the $S2_A$ foliation (Figure 3:3:1). Amplitudes recorded vary significantly,

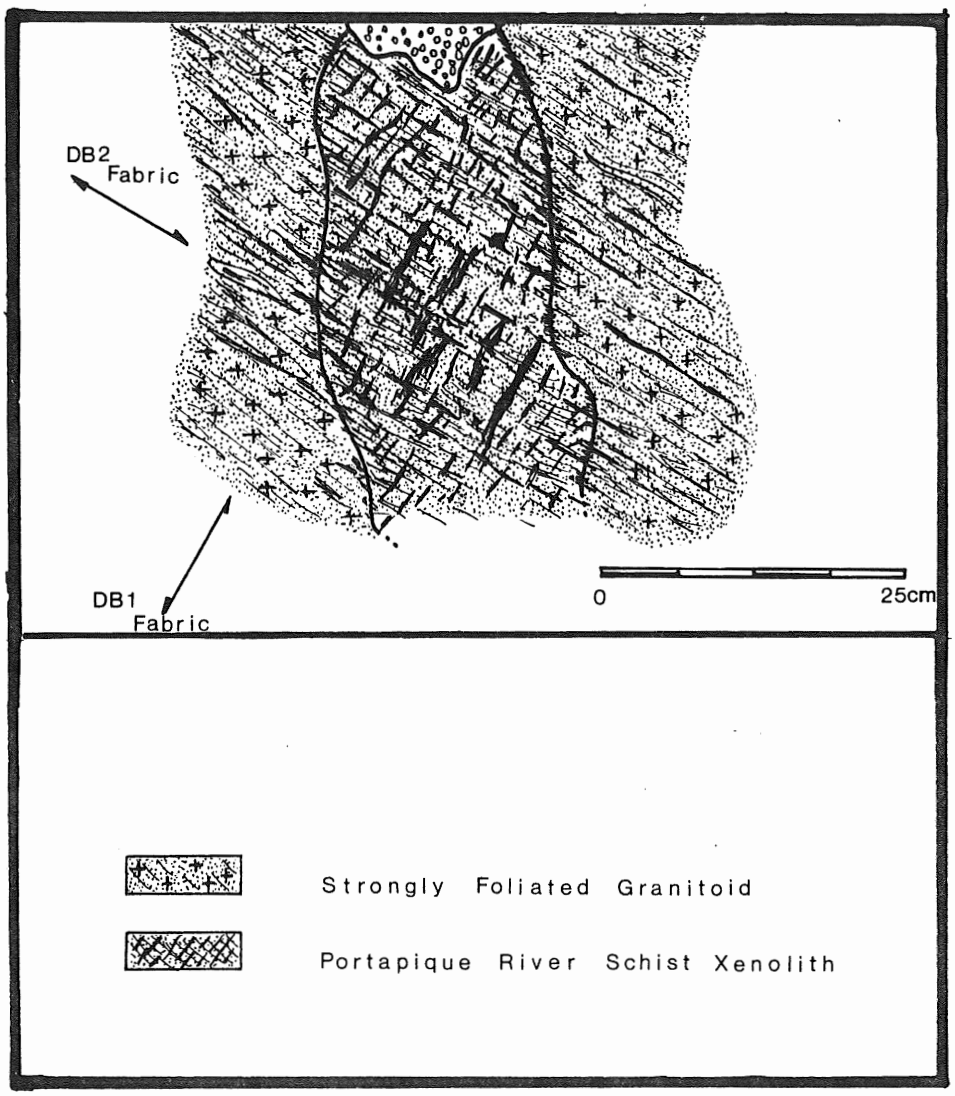


Figure 3:2:3 Sketch of Portapique River Schist xenolith in the Strongly Foliated Granitoid intrusion on Portapique River near the confluence with Gamble Brook. The early DB1 fabric of the xenolith has been overprinted by the regional DB2 foliation.

Figure 3:3:1 A DB2 generation fold from the Great Village River Gneiss. Note how compositional layering in the fine-grained gneiss is attenuated along the steeply south dipping - east trending DB2 overprinting fabric. This exposure is about 100 m north of the Londonderry Fault on Higgins Mountain Road. North is to the left.

but are in every case greater than 5 x wavelength. These fall between the Class 2 and 3 categories of Ramsay (1967, pp. 365-372) (Figure 3:3:2). As noted in Chapter 2, outcrop-scale $F2_A$ folds are not common in Domain A. Where evident, they are relatively small scale structures never exceeding 20 cm in wavelength. Because fold generation is controlled, to a large degree, by viscosity and competency contrasts and layer thicknesses of the deforming materials (Ramberg, 1964, Biot,1964, Hobbs et al. 1976, pp. 200-211) the apparent scarcity of $F2_A$ folds may be a reflection of relative homogeneity within the protolith. However, this scarcity is more likely an observational bias resulting from fold limb reorientations toward the foliation plane.

A strongly developed penetrative metamorphic foliation, ($S2_A$), that is essentially parallel to the axial surface of $F2_A$ structures, is the dominant planar fabric in all rocks of Domain A. It ranges from a pervasive, preferred mineral orientation fabric to a protomylonitic or mylonitic foliation. Mylonitic fabric developed locally during heterogeneous shearing or flattening in discrete zones that could not have been exactly parallel with the axial planar fabric $S2_A$ (Figure 3:3:3). Textures ranging from a quartz - biotite - muscovite grain shape fabric to ultramylonite have been observed in single outcrops but mylonite is commonly associated with distinct contact zones such as between metadiabase (pre- or syn- DB2) and porphyroclastic granitoid rocks. Possibly rheologic contrast between the juxtaposed rock types or fluid flow characteristics played a role in this localisation of the high strain zones.

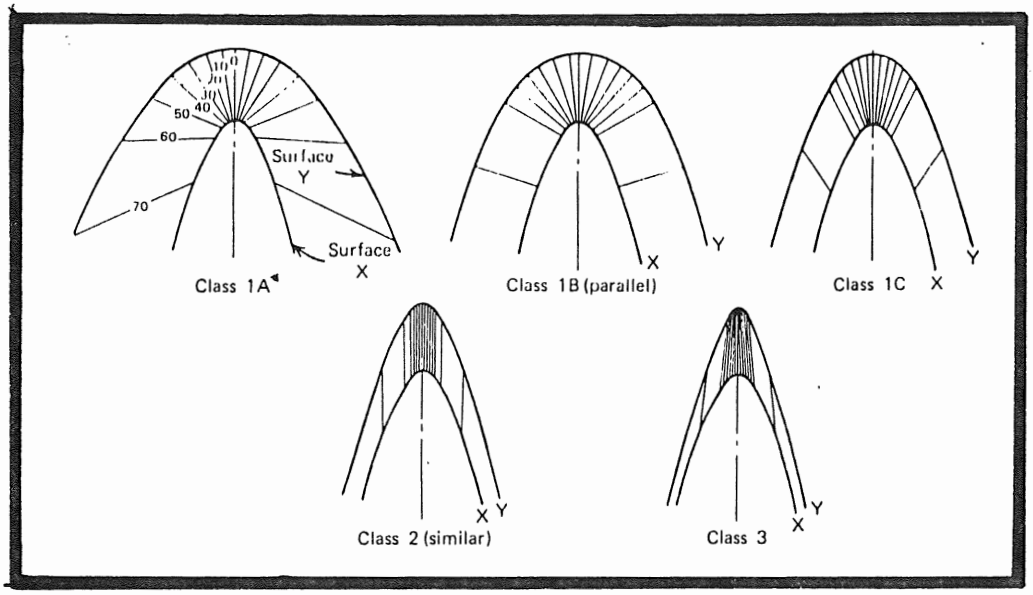


Figure 3:3:2 An outline of the non-genetic fold classification of Ramsay (1967, pp. 365). The lines drawn between x and y surfaces are idealised dip isogons.

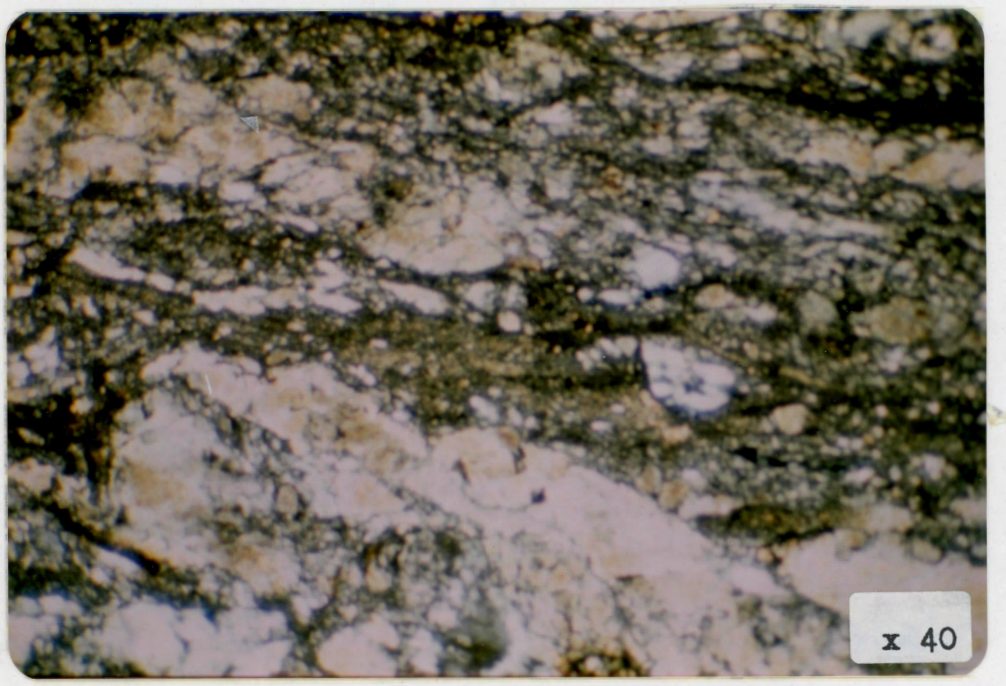


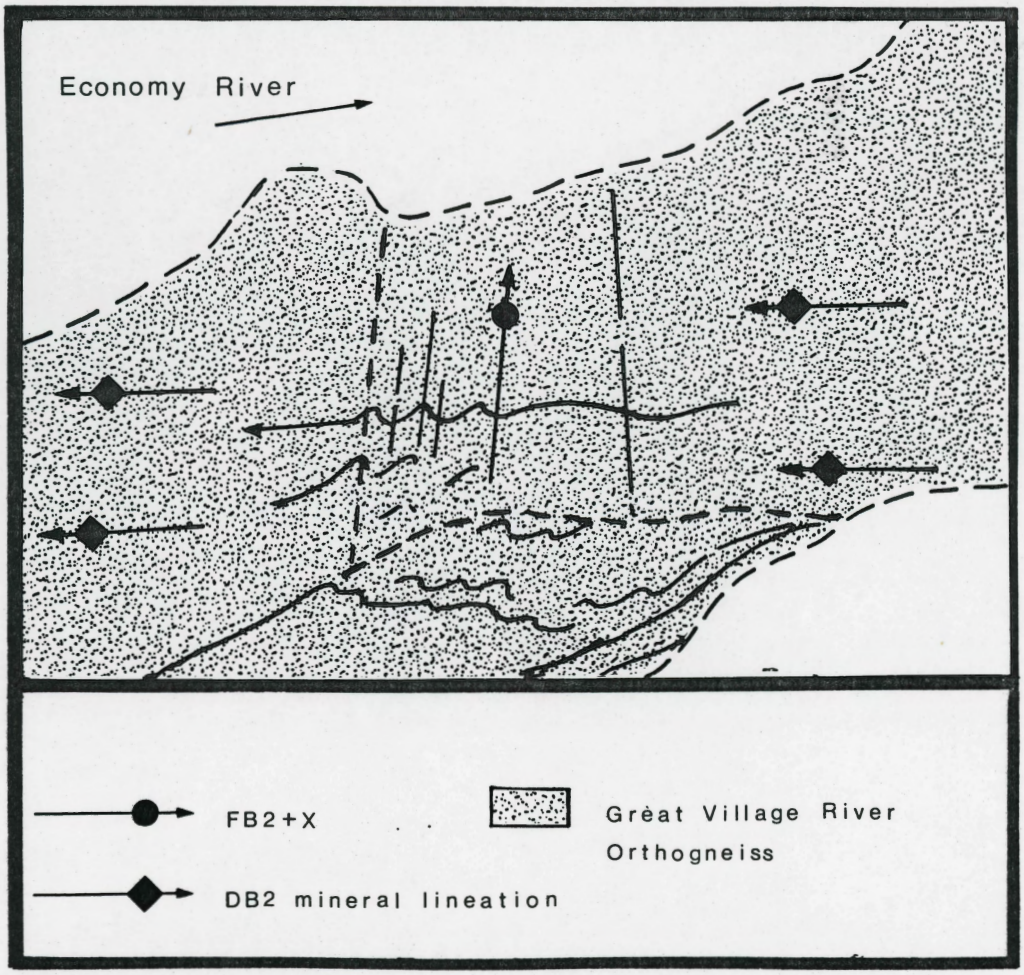
Figure 3:3:3 Syn-tectonic, mylonitic DB2 fabric from Domain A. Discontinuous quartz - feldspar aggregates occur in a matrix of much finer-grained , altered feldspar. Sample was collected from an exposure on the northern portion of the Rockland Brook section. (crossed polars)

Throughout much of Domain A a moderate to strongly developed mineral lineation ($LM2_A$) is developed within $S2_A$. This feature is defined by the long axes of hornblende grains in the amphibolites and by quartz or quartz-feldspar-muscovite augen structures or aggregates in the orthogneisses and granitoids. In some cases elements defining the trend are coaxial with $F2_A$ fold axes. The preferred mineral orientation fabric has a similar angular relationship with neck lines of intra-$S2_A$ pinch-and-swell structures, although these are not common. No pinch-and-swell features with necklines perpendicular to the mineral lineation were observed. The total linear element population plunges at a low to moderate angle down the moderate to steep dip of foliation and although intensity varies considerably it is best developed in the Bass River Amphibolite and in the Great Village River Orthogneiss.

Although not widely observed, a fold generation intermediate between $F2_A$ and $F3_A$, termed $F2_A+X$ was identified in protomylonitic orthogneisses of the Economy River. These structures deform the $LM2_A$ and $S2_A$ fabric components but also appear to be enveloped by the $S2_A$ foliation on the outcrop scale (Figure 3:3:4). Maximum wavelengths of approximately 30 cm, amplitudes of 15 to 30 cm and a periodicity of 3-5 meters are characteristic of these "Z" sense (viewed down-plunge) structures. Since $S2_A$ effectively envelopes these folds they must have developed during DB2 as disharmonic structures after the main folding and metamorphic peaks. Alternatively, they can be interpreted as structures developed during the final phases of local ductile deformation, that had

Figure 3:3:4 A general outcrop view of FB2+X generation folds. The accompanying line sketch clarifies the overprinting relationship. The DB2 generation foliation fabric and its associated mineral lineation have been folded by the FB2+X folds but these structures are ultimately enveloped by the DB2 foliation. Note the non-coaxiality of the fold axial trends. This location, at the falls on Economy River, is the only one at which folds of this generation have been identified. North is to the bottom of the photo.

Figure 3:3:4 cont'd



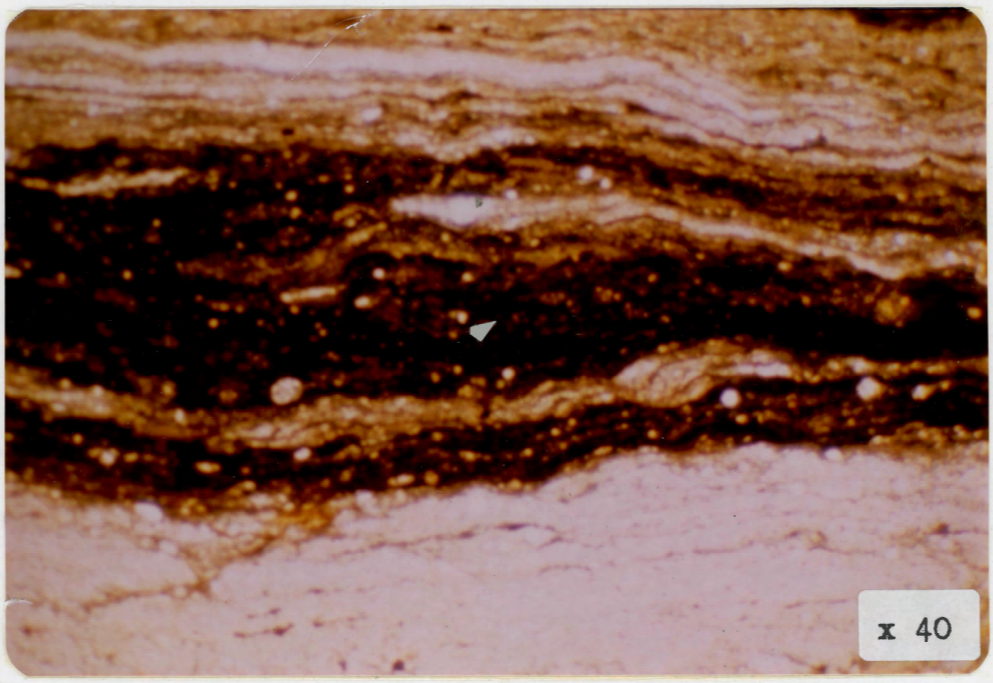


Figure 3:3:5 Mylonitic DB2 fabric in Domain S rocks exposed on Folly River. Note the discontinuous, quartz-dominated flaser structures. The rounded porphyroclasts are predominantly feldspar grains and aggregates in an altered matrix of plagioclase and mafic minerals. (plane light)

$S1_S$ foliation. These folds define a transpositional regime within which primary layering rotated progressively toward the $S1_S$ trend.

Fold styles vary with lithology. Where small- scale $D1_S$ structures developed in metavolcanic greenschists, nearly total obliteration of primary structures has occurred. $F1_S$ folds are evident as detached, isoclinal closures visible only on the centimetre- or micro-scale (Figure 3:3:6). The largest well-defined $F1_S$ structures are exposed in the east wall of Folly Lake Quarry. Transposition features characterising the metasedimentary Gamble Brook Schist are dominated by tight to isoclinal folds that commonly show detached hinges. Wavelengths are on the order of 1 to 5 m where discernible and associated amplitudes are estimated as being in the 5 to 15 m range (Figure 3:3:7). Parasitic folds generally mimic larger scale sense and geometry but are not pervasively distributed. Detachment of one or both limbs of outcrop-scale and microscale $F1_S$ folds is common (Figure 3:3:8a, b, c).

It is somewhat unclear at present whether very large scale reclined to recumbent $F1_S$ folds exist within Structural Domain S. The largest and most open structures of this generation are partially exposed on the East Folly River at the powerline crossing and approximately 620 meters to the south (Plate 1). Both are defined by quartzitic metavolcaniclastic and metasedimentary rocks in which primary layering is well preserved. At the first location local $S1_S$ trends indicate the presence of a relatively large, reclined, structure with a wavelength of at least 20 m. At the second site, bedding-foliation relationships define the north

Figure 3:3:6 Microscale DB2 generation fold in quartzitic metasedimentary rock from an exposure on East Folly River. Quartz, biotite, epidote, and chlorite define the primary compositional layering. Note the abrupt thinning of the uppermost fold limb toward the discrete DB2-related shear plane. $S1_S$ is the axial planar fabric to this fold generation in Domain S and is equivalent to $S2_A$ of Domain A. In a regional context the domainal distinction may be dropped and SB2 used to indicate association with the DB2, Complex-wide deformation.

Figure 3:3:7 A large scale DB2 generation isoclinal fold in a quartzite-dominated portion of the Gamble Brook Schist exposed on the east wall of Folly Lake Quarry. A rusty biotite-rich layer can be traced around the closure, structurally above a thick, grey quartzite layer. The 075 Az. trending fold axial surface is essentially vertical and the structure plunges to the east at about 25°. North is to the left. Note the hammer for scale.

Figure 3:3:8(a) A close-up view of a DB2 generation fold closure zone in Gamble Brook Schist rocks. Bedding is clearly visible as the light brown and white color banding defining the fold. The associated axial planar fabric is apparent as a spaced, healed fracture pattern. The fold axial surface is approximately 065 Az. trending and dips to the south at about 80°. The structure plunges to the east at 28°. North is to the right.

Figure 3:3:8(b) Evidence of progressive rotation of bedding toward the foliation plane. The axial-planar DB2 generation fabric is vertical and east-trending and bedding is defined by the color banding that cuts across the tectonic fabric at a low angle. " Z" sense (viewed down-plunge) minor fold geometry indicates that an antiformal closure lies to the south of this exposure. The minor folds plunge eastward at 35° . North is to the left.

Figure 3:3:8(c) A fold limb high strain zone in which bedding has been rotated into near-parallelism with the tectonic fabric. Fold closure zones are locally recognisable (as in the lower right corner of the photo) but are detached generally. The DB2 enveloping surface which dips shallowly south in Figure 3:3:8(a), is steeply south dipping in this photo and approximates the DB2 generation foliation orientation. North is to the left, and the exposure surface is approximately horizontal.

limb of an asymmetric anticline that is exposed over several tens of metres.

At many localities the $S1_S$ foliation carries a variably developed linear element fabric. Four distinct elements make up this population:

(1) intersection of bedding (SO_S) on $S1_S$ foliation surfaces (referred to as $LI1_S$);
(2) mineral lineation predominantly defined by quartz and feldspar with chlorite, epidote, sericite and biotite (referred to as $LM1_S$);
(3) rodding produced by cm to microscale detached $F2_S$ hinge zones (referred to as $LR1_S$);
(4) pinch-and-swell necking within the $S1_S$ foliation (referred to as $LP1_S$).

The most common of these is (1), the $SO_S/S1_S$ intersection lineation. Exposures on East Folly River south of the powerline (Plate 1) confirm that $LI1_S$ seen on the $S1_S$ foliation plane approximates the axial trend of associated folds in the same outcrops. In rocks not bearing immediately-apparent layering this proved to be a useful indicator of $F1_S$ structures.

The detached closure and composite rodding fabric ($LR1_S$)was infrequently observed, and appears to be restricted to quartz - chlorite - mica schist. On the outcrop scale pervasive 2-3 mm quartz-feldspar rodding exists, each rod representing a detached minor fold or a small-scale, parasitic fold within a centimeter-scale, detached closure. Seg-

regation of mica and quartz aggregates occurred during the $D1_S$ transposition phase and resulted in a pattern of enclosed, hinge-zone relicts. Ductile matrix deformation, in combination with quartz recrystallisation and bedding relict rotation produced the strong composite element fabric seen in the schist outcrops (Figure 3:3:9). However, its composite nature is clear only in sections at a moderately high angle to the $LR1_S$ trend. Samples slabbed perpendicular to the lineation clearly illustrate the detached closures.

A strong $LM2_S$ mineral lineation is restricted to discrete zones of intense deformation (mylonitisation) or in strained pre- or syn- $D1_S$ intrusive rocks such as metadiabases and Strongly Foliated Granitoid rocks. Quartz-feldspar-mica mineral aggregates or hornblende grains generally define the lineation. Relative to Domain A,rocks bearing a strong mineral lineation are far less common in Domain S.

Pinch-and-swell necking in syn- to pre-kinematic diabase dikes and quartz veins are rare in Domain S but in every case the recorded necklines are parallel to the linear fabric developed in other lithologies. The largest scale examples of pinch-and-swell deformation in Domain S occur in the vicinity of the Folly River Schist-Gamble Brook Schist transition on East Folly River (Figure 3:3:10). At this location metadiabase dikes which carry a strong $S1_S$ foliation and a weak lineation are affected. Warping of $S1_S$ into the neck zones has locally resulted in well defined, open folds whose axes are parallel to the necking trends.

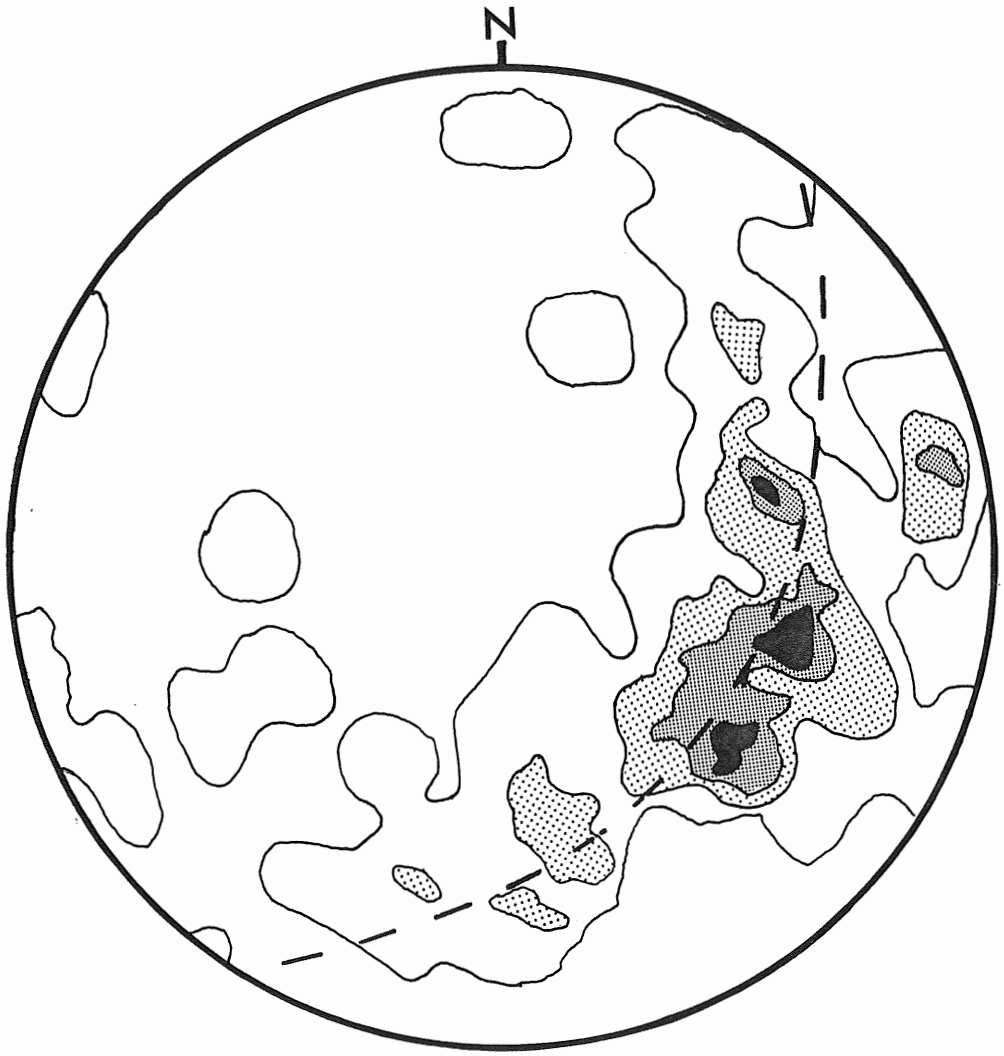
Figure 3:3:9 Composite linear fabric in sericitic, quartzo-feldspathic schist. The surface feature is a combination of quartz-feldspar rodding fabric and the bedding-cleavage intersection. Both elements coincide with axes of detached DB2 generation folds in the same outcrop. The exposure occurs on a logging road west of Highway 104 at Folly Mountain. North is to the top right of the photo.

Figure 3:3:10 Pinch and swell structure in deformed diabase on the upper portion of East Folly River, near the Gamble Brook Schist - Folly River Schist transition. East trending-south dipping biotite - quartz - chlorite - sericite schist hosts the diabase. North is to the left.

The angular relationships between orientation data for various linear element populations are illustrated in Figure 3:3:11a. It is important to note that these data do not solely describe a strong easterly-plunging group such as that presented by Eisbacher (1967; pp. 44 ; see Figure 3:3:11b). A combination of all linear element data in one synoptic plot defines a northeast trending, moderately southeast dipping girdle (Figure 3:3:11c). Its orientation is a function primarily of subsequent $D2_S$ folding but a regional coaxiality of $D1_S$ linear elements is indicated from field observations of coincident features such as fold axes and mineral lineations. The post-deformation distribution pattern is interpreted as representing reorientation of an originally strong southeast-plunging orientation grouping of linear elements that probably plunged to the southeast (Weiss, 1964, pp.482-483; Ramsay, 1967, pp.470-472; Hobbs et al., 1976, pp.192-193). No precise pre-DB2 distribution can be discerned, however, because the effects of similar style folding cannot be realistically removed by orientation analysis of the presently available data. This difficulty arises from the lack of detailed kinematic information pertaining to the superimposed folding strains.

3:3:3 Observations on The Nature of DB2 Deformation

Geometric relationships between outcrop and microscale DB2 fabric elements in both structural domains generally developed during DB2 but reorientation of most features by DB3 deformation must be assumed on



CONTOURS AT 1, 3, 6, AND > 9% PER 1% AREA

≥ 3% ≥ 6% ≥ 9%

N = 72

Figure 3:11 (a) 1 Schmidt equal area plot of all DB^2 generation mineral lineations. Note the north-northeast trending, southeast dipping girdle (lower hemisphere projection).

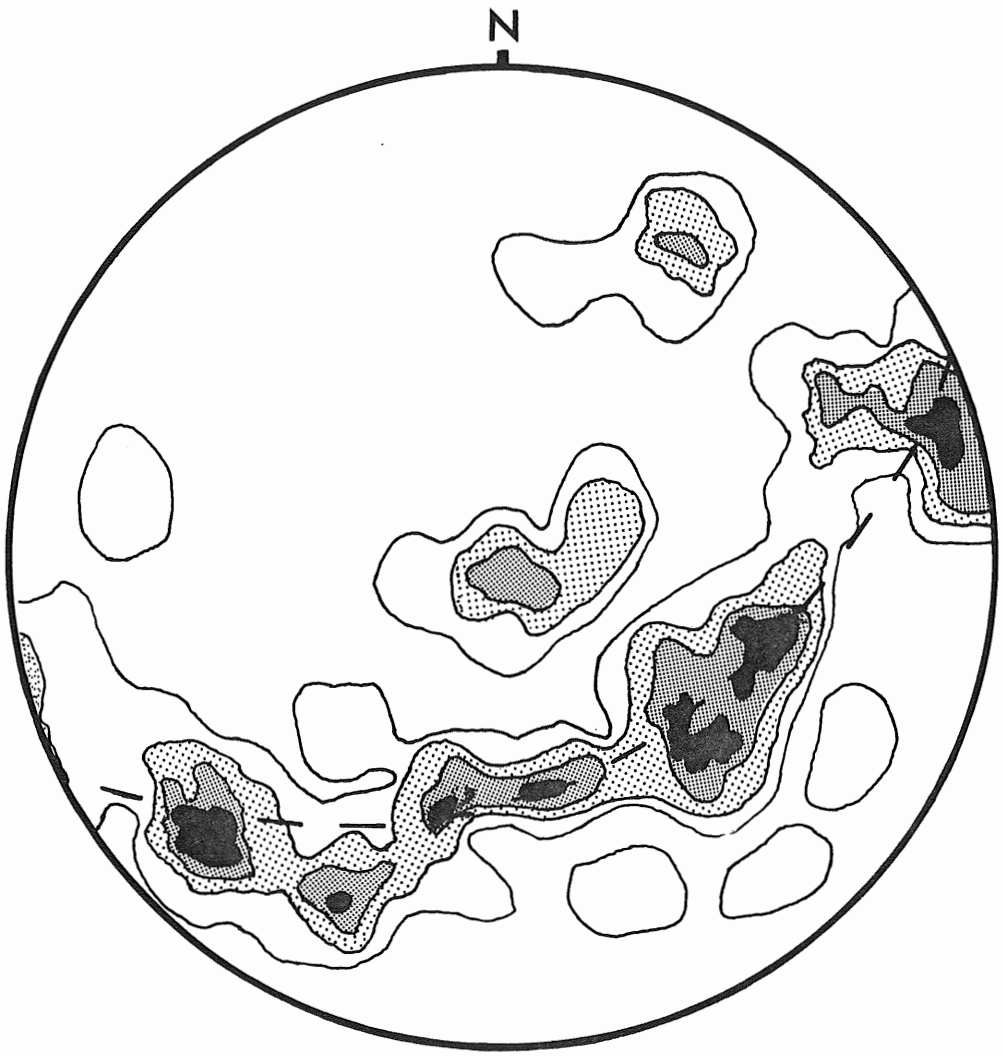


Figure 3:11 (a) 2 Schmidt equal area plot of all DB^2 generation fold axes. Note the dominant northeast trending, southeast dipping girdle (lower hemisphere projection).

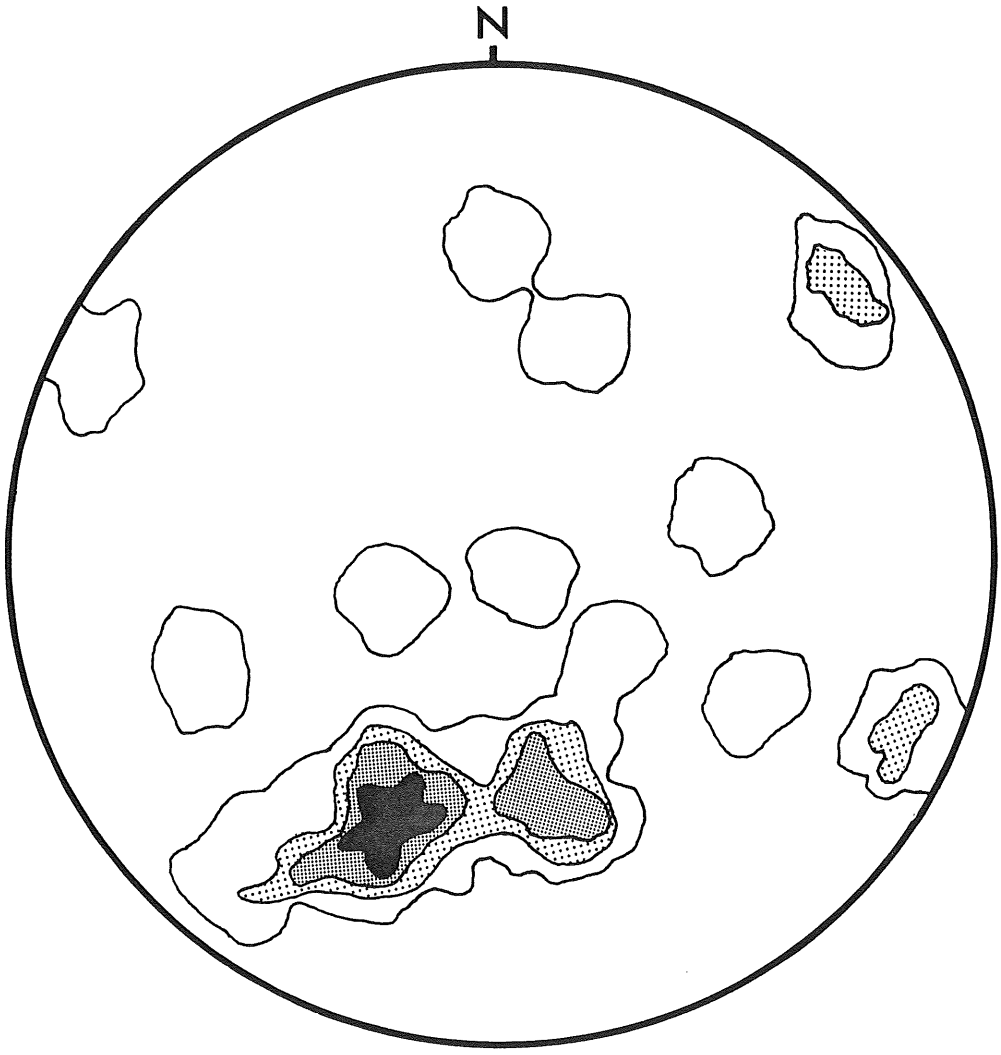
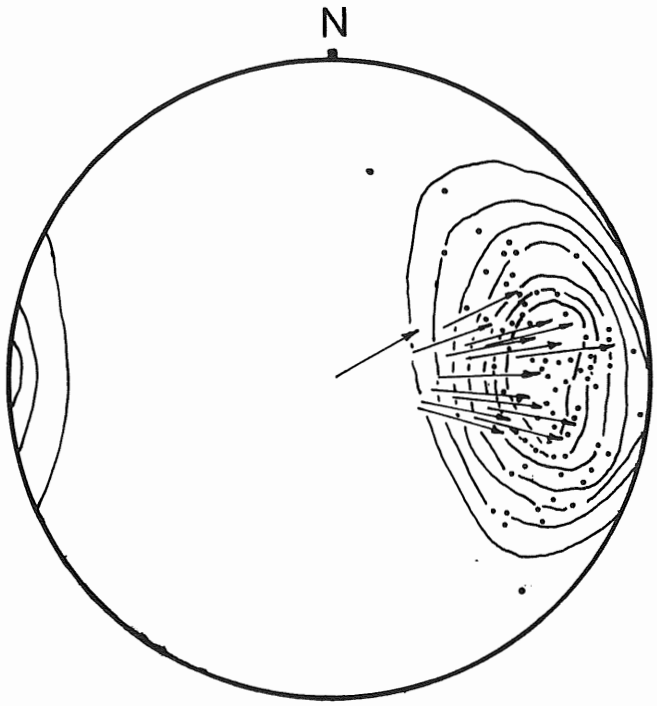


Figure 3:11 (a)3 Schmidt equal area plot of DB^2 generation intersection lineations and pinch and swell necklines. Where present in the same outcrop, these features were always observed to be parallel (lower hemisphere projection).



Folds ——→

Mineral Lineation •

CONTOUR INTERVAL NOT INDICATED

Figure 3:11 (b) Mineral lineation and fold axis data presented by Eisbacher, 1967, p.44). These data do not describe the girdle trend outlined in Figure 3:11 (c) of this study (lower hemisphere projection).

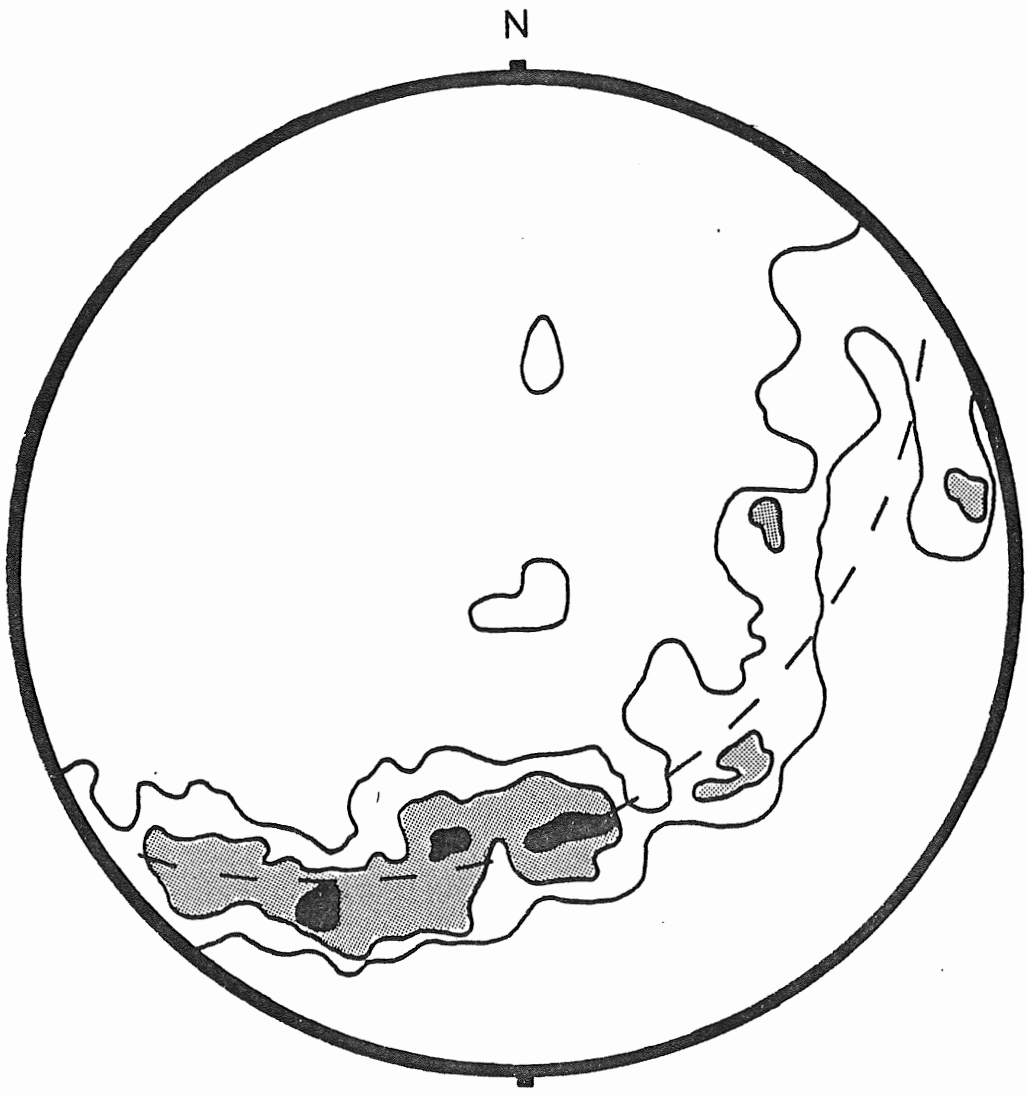


Figure 3:3:11(c) Synoptic Schmidt equal area plot of all Domain S linear elements. Note the northeast trending, south dipping partial girdle.

the kilometre scale. As a result, orientation plots of foliation and linear elements supply only indirect information in regard to the earlier, large-scale DB2 structural geometry. The small scale structures and their interelationships are useful, however, in discerning general deformation trends on a more restricted scale.

Coincidence of mineral lineations and isoclinal fold axes, similar to that noted in both Domain A and S has been described by numerous workers in other areas (Cloos, 1957 , Bryant and Reed, 1969 , Bell, 1978; Berthe et al., 1978; Watkinson and Cobbold, 1981) and related to either the intermediate or maximum axis of the finite strain ellipsoid . Ramsay and Graham (1979) and Cobbold and Quinquis (1980) outline sequential development of such relationships under conditions approximating simple shear and structures generated under such a regime have been documented elsewhere (Minnigh, 1979; Henderson, 1981). The coincidence of linear fabric elements need not be a result of simple shear, however, for Treagus & Treagus (1981) describe similar features resulting from a simple flattening strain. The distinction between straining mechanisms is difficult, if not impossible, to make unless discrete marker horizons can be identified which allow definition of relative displacements and the deformation path.

Pinch-and-swell necks and mineral lineations coincide throughout the Complex (see Figure 3:3:11a) . On Economy River DB2 mylonitic layering in penetratively deformed diabase bears a preferred mineral orientation coinciding with the neck axes. Assuming that the neck axis trend

locally approximates the intermediate finite strain axis (Hobbs et al., 1976, pp. 283-288 ; Schwertdner et al., 1978) and that the lack of chocolate block boudins reflects at least local strain configurations, the coincidence of inferred x and y finite strain axes becomes problematic.

It is unlikely that necking predated the penetrative strain imprint of DB2 because the orientation of the linear features is consistent in both host and diabase dike. Superposition of a local flattening strain in the post-fabric-development stage of the deformation history must have occurred but is difficult to confine relative to other deformations. These may be related to the superimposed DB3 generation folding for some M2 hornblende grains in amphibolites (see Chapter 4, section 4:1) bear evidence of brittle fracture and rotation and occur in association with micro-pinch-and-swell structures along limbs of DB3 minor folds.

3:3:4 Summary of DB2 Deformation

The DB2 history in Domains A and S is essentially the same. Syn-kinematic fabric development and possibly emplacement of granitoid bodies were accompanied by localised ductile shear along zones oriented close to the foliation trend. This resulted in undetermined amounts of translation and tectonic attenuation of the sequence. An intermediate disharmonic fold generation developed which locally deformed DB2 linear and planar fabric elements but was enveloped on the outcrop-scale by the general DB2 foliation trend. All rocks were later affected to varying

degrees by a pervasive, but relatively weak, flattening strain, evident from outcrop and microscale pinch-and-swell structures.

Large scale DB2 orientation analysis is hindered by DB3 overprinting, but the noted heterogeneous strain characteristics of DB2 indicate that transposition - related translation occurred along zones closely approximating the SB2 orientation. This resulted in limb attenuation of FB2 folds at all scales. Large-scale mobilisation along SB2 high strain zones may be indicated by the approximate coincidence of apparent metamorphic isograds with the lithotectonic boundaries. Sillimanite distribution on Portapique River (that is, confined solely to the very thin Portapique River Schist) exemplifies this point.

The pre-DB3 structural configuration was a tightly folded, interleaved package of high and low grade rocks that had been significantly modified by transposition and shearing. At the close of this deformation period the two structural domains were probably separated by significant structural discontinuities. Rocks with vastly differing early tectonic and metamorphic histories had been juxtaposed.

3:4 The DB3 Event - $D3_A$ in Structural Domain A
- $D2_S$ in Structural Domain S

3:4:1 DB3 Structures

This deformation affected all rocks of the Bass River Complex, producing first-order, asymmetric, folds characterised by kilometre-

scale wavelengths and estimated amplitudes of hundreds of meters. Second order folds of similar geometry and tens of metres wavelengths and amplitudes are described by variations in $S2_A$ and $S1_B$ foliation trends on the outcrop-section scale. Third order small-scale structures are sporadic in their occurrence, being generally confined to second order hinge zones. Asymmetry relationships mapped in continuons exposures across second order closures indicate that where sufficient data is available, vergence of minor folds can be relied upon to define larger scale structures. Unfortunately, the sporadic distribution of FB3 minor folds hinders their use in defining the larger-scale structures (Plate 4). Although no pervasive axial planar fabric is associated with these folds, local crenulations of prexisting S planes and microscopic quartz grain shape fabrics commonly mark the axial planar trends.

Style of folding varies systematically on the outcrop scale according to rock type: phyllites and chlorite schists of the metavolcanic sequence have variably plunging, similar style minor folds as do the biotite and biotite - chlorite - quartz - garnet schists of the metasedimentary unit (Figure 3:4:1). More competent quartzites and quartz-biotite psammites bear fewer minor folds, but where they do occur these are small and show less hinge thickening (Figure 3:4:2). The massive volcanic units, which bear only a slight $D1_S$ imprint, are similar to the resistant metasedimentary rocks in that small scale $F2_S$ folds are rare and show lesser amounts of hinge thickening.

Domain A structures are similar to those of Domain S but show a

Figure 3:4:1 Upright, "Z" sense, third order, DB3 generation folds in the Folly River Schist sequence. The pen is parallel to the surface trace of the incipient crenulation fabric ($S2_S$ in Domain S) and the fold axis plunges at 15 into the photo. This exposure occurs about 500 m south of the powerline crossing on East Folly River. North is to the right.

Figure 3:4:2 DB3 generation third order fold in quartzite of the Gamble Brook Schist. The color banding of the bedding defines DB2 generation isoclinal folds which have been refolded by the DB3 structure. This sample was collected from an outcrop along the East Folly Mountain logging road system, south of Moose Pond (see Plate 2).

axial-planar crenulation fabric. Orthogneisses at one location on Rockland Brook have developed an equivalent, spaced, variably differentiated foliation in association with second order folds.

The Folly River section under the powerline exemplifies DB3 folding style in the supracrustal sequence, and the Higgins Mountain Road section provides similar information for the Domain A sequence. Details of each are presented in Figures 3:4:4 and 3:4:3 respectively. The inferred general similarity of folding style throughout both structural domains is illustrated in the Plate 6 structural cross sections from different locations along the strike of the Complex.

3:4:2 DB3 Orientation Trends

DB3 fold axial surface traces in Plate 3 range from east-west to northeast, describing a very gentle "S" between the north and south boundaries of the Complex. Stereographic equal area plots of pre-existing S surface orientations for both domains define differences in fold trends along the length of the study area (Plate 3). The mean northeast flexure is described in a synoptic plot by data distribution along a partial gridle at 060/20°N (Figure 3:4:5).

The elongate stratigraphic units of the Complex parallel the DB3 generation's structural trend and probably developed by approximate coaxiality of large-scale DB2 and DB3 structures. The lack of recognisable interference structures on outcrop or map scales and local coincidence of DB3 and DB2 linear fabric components in (Figure 3:4:6)

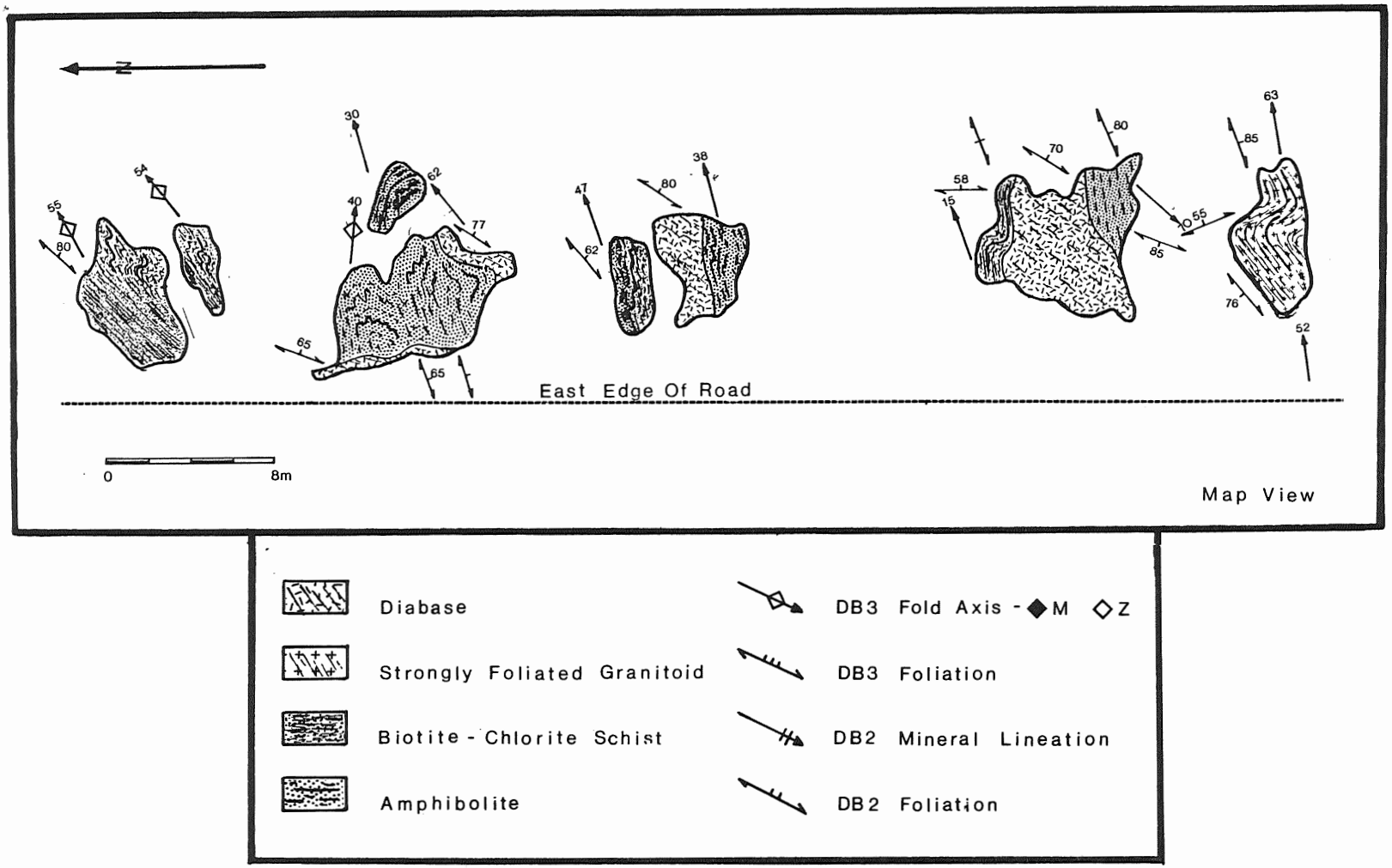


Figure 3:4:3 An outline sketch of second order DB3 folds that are exposed along Higgins Mountain Road.

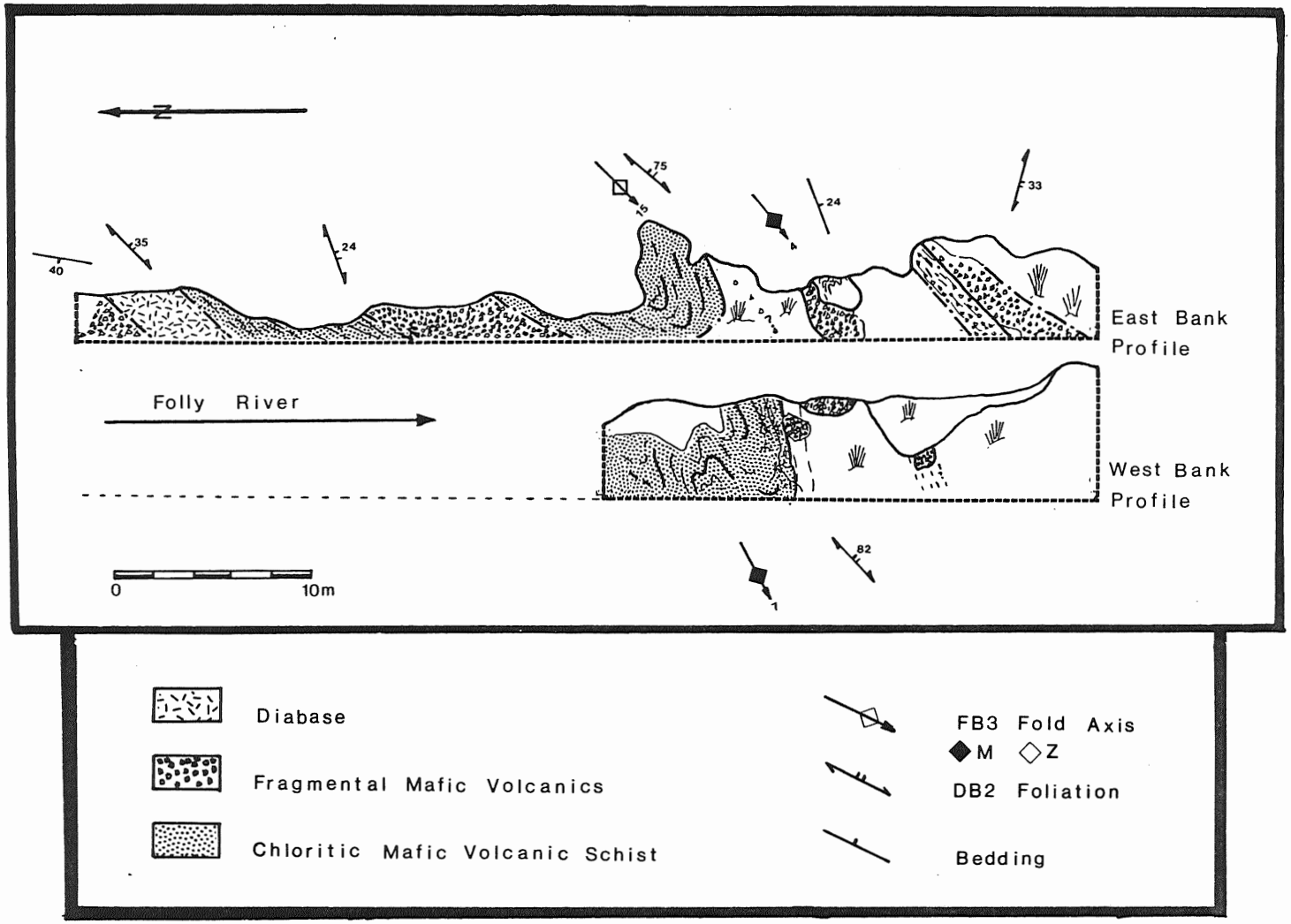


Figure 3:4:4 An outline sketch of second order DB3 folds that occur along Folly River at the powerline crossing.

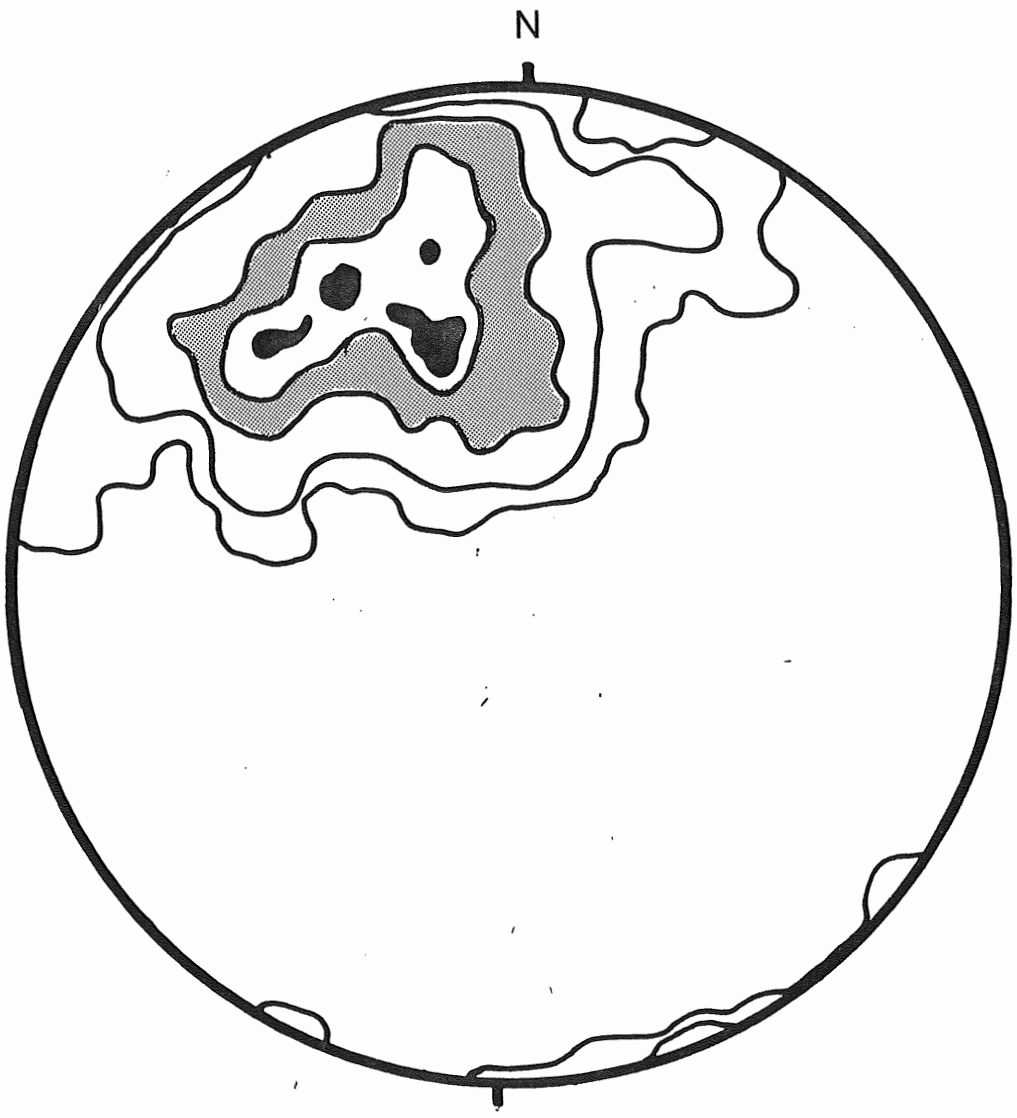


Figure 3:4:5 Schmidt equal area plot of DB2 foliation poles. Note the northwest and north-northeast trending partial girdles. These define the regional DB3 fold trends across the Bass River Complex.

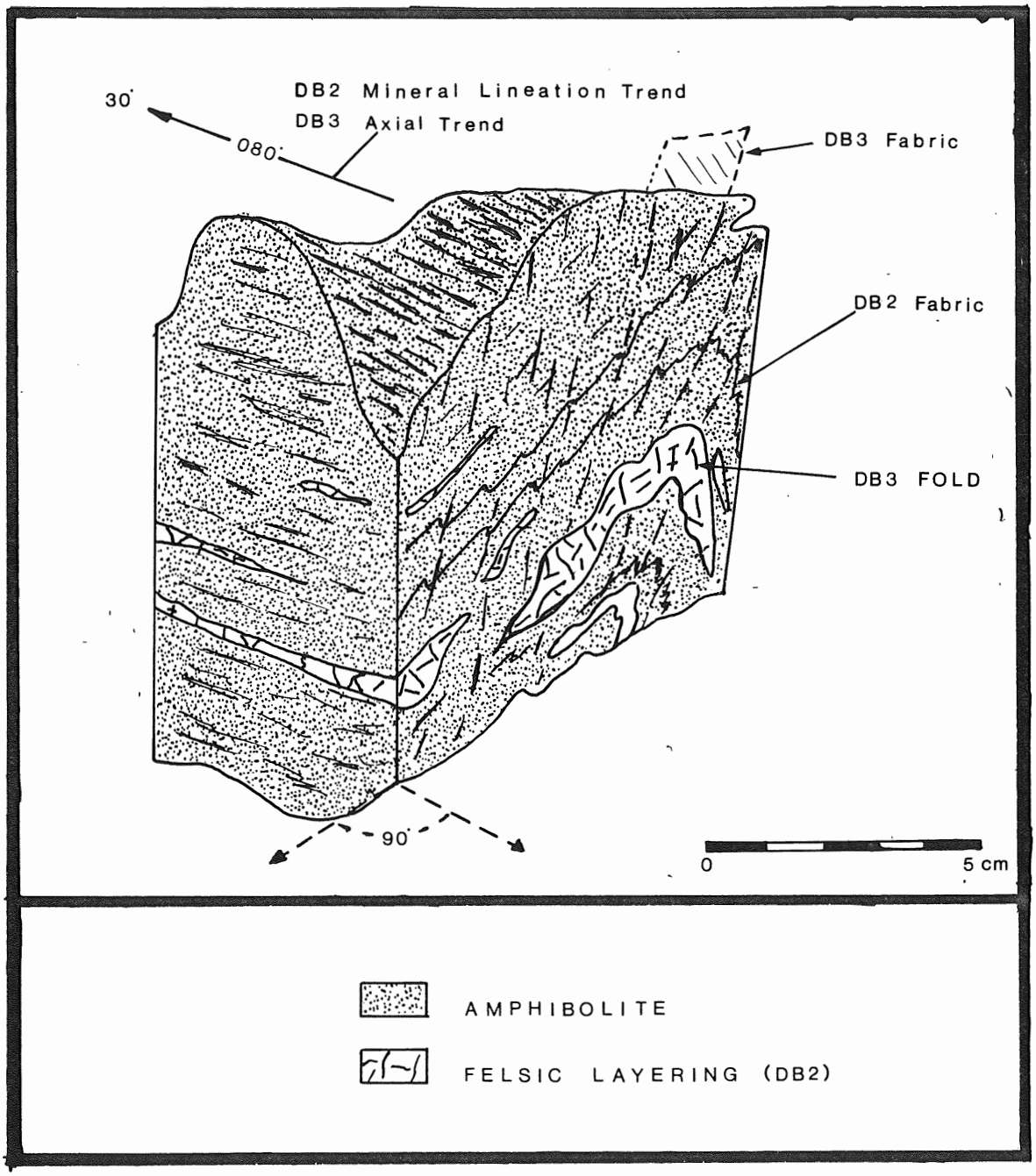


Figure 3:4:6 Sketch of a slabbed hand sample showing relationships between DB2 and DB3 fabric elements in amphibolite from Higgins Mountain Road. This sample was collected from the section depicted in Figure 3:4:3. Note that on this scale linear structures of both generations are essentially coaxial.

supports this conclusion. If these trends had been significantly different, interference patterns would surly have resulted.

Complexity of linear element and fold axis trends developed during the DB3 event increases east of Portapique River (Plate 4) and three broad trend domains are recognisable across the Complex. These are characterised by:

(1) south plunging features;
(2) easterly or westerly plunging features;
(3) a combination of (1) and (2).

In the Rockland Brook - Great Village River area the first group predominates whereas trend domain (3) occurs in the Folly River area and (2) predominates east of Folly River. DB3 axial trend data all fall along a $055^{\circ}/42^{\circ}S$ best-fit partial girdle and define several maxima (Figure 3:4:7). Some minor folds plunge to the southeast at angles between 35° and 70° but the strongest grouping of data defines a southwest-northeast trend. Unimodal trends mark the first two trend domains noted above but those of the third domain outline an interesting, but somewhat diffuse, bimodal pattern separated along the best-fit girdle between 090 Az and 180 Az .

Trend domain 3 data have been interpreted as defining a large-scale fold axial trend variation pattern which culminates updip, within the plane of the best-fit girdle for fold axis data. The transition from west to east through trend domains 1 to 3 therefore marks a systematic

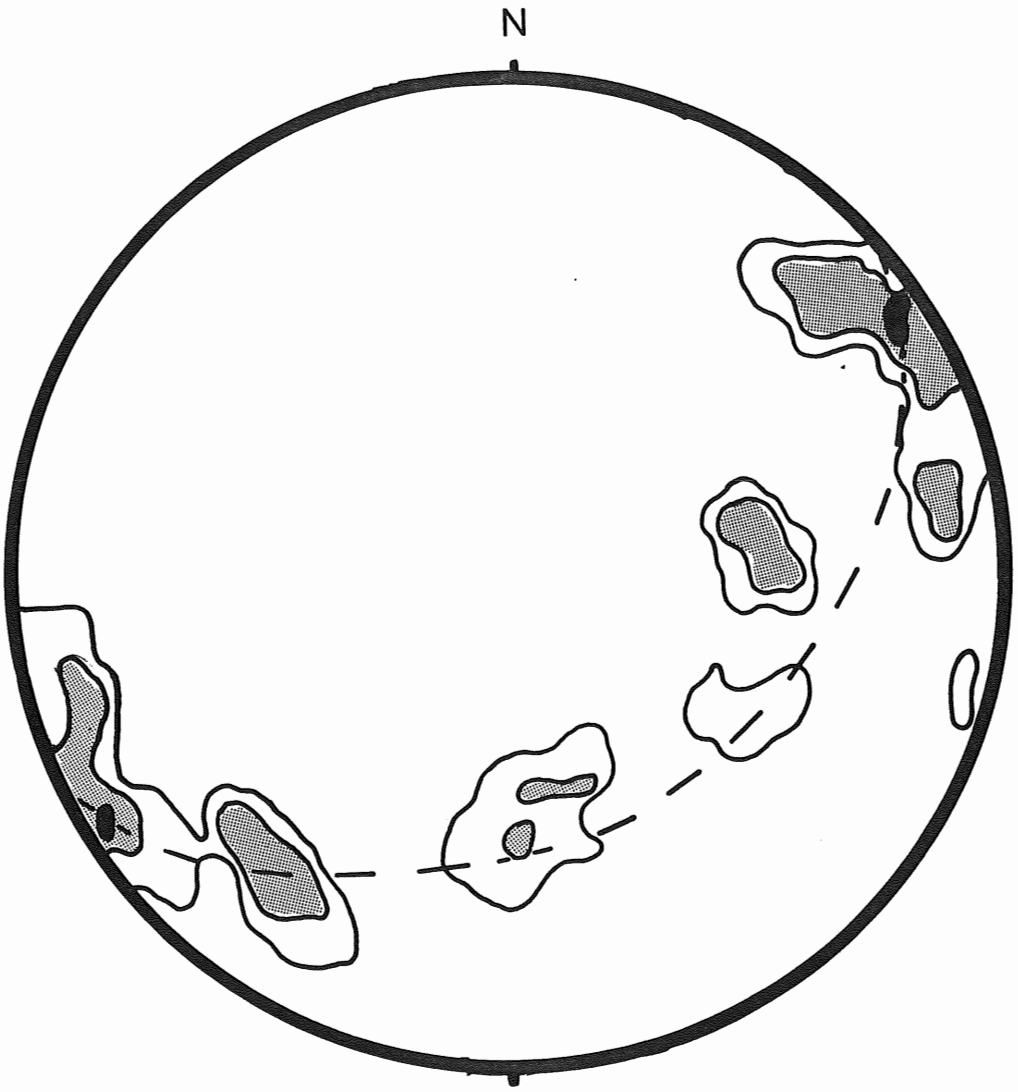


Figure 3:4:7 Schmidt equal area plot of DB3 axial trends. Note that these define a northeast trending, moderately southeast dipping, partial girdle.

change in DB3 fold axis orientations that can be directly related to change in structural level within the Bass River Complex. Consistent orientations mark the structural Domain A rocks and part of the metasedimentary sequence but only rocks of the metavolcanic sequence, highest of the inferred structural sucession, bear evidence of the more diverse fold axis orientations.

3:4:3 Observations On The Nature of The DB3 Deformation

The changes in orientation of DB3 fold axes across the Bass River Complex that are described above are related, obviously, to some deformation process that was either directly related to the DB3 event or one that post-dated that event. The possibilities can be summarised in the following manner:

(1) post-DB3 folding;

(2) post-DB3 penetrative shearing;

(3) markedly heterogeneons DB3 strain across the fold hinge zones.

The first two possibilities must be argued as specific cases. The observed regional consistency in fold trends could only be accomplished by simple rotational movement if the axis of rotation had been nearly perpendicular to the DB3 axial surfaces in both Domains A and S. Penetrative folding about such an axis by mechanisms approximating those of similar folding must be envisioned to explain the observed partial girdle plot (Weiss 1964 pp.482-483; Ramsay,1967 pp.470-472) but no evidence of such a widely developed folding phase has been observed

anywhere in the Bass River Complex. Reorientation by post DB3 shearing should be evident from major discrete shear zones at orientations close to those of the DB3 axial surfaces and therefore crosscutting or overprinting the post-folding orientation patterns of the pre-existing foliation. No such major zones have been identified in association with the folds. The third possibility, that of heterogeneous DB3 strain resulting in variable distribution of fold axes within the mean axial surface trend, seems more tenable.

Ramsay (1967, pp.470-472), and Hobbs et al. (1976, pp. 192-193), describe distribution patterns resulting from similar folding of materials bearing linear fabric elements. They show that originally coaxial linear elements deform as passive marks under ideal conditions. These ultimately describe partial or complete girdle distributions at some angle to the fold axial surface. The angular separation between the girdle defining the linear element orientation and the axial surface increases as a function of the original angular separation between them. It follows that for a deformation superimposed upon a pre-existing linear fabric, which is oriented at a low angle to axial surface of the superimposed phase, distribution of the elements will occur within a plane that's orientation is close to that of the superimposed, generation's axial surface. This appears to be the case in the highest structural levels of the Bass River Complex.

Sanderson (1973) discussed statistically significant trend variations in fold axis populations from numerous localities and determined

that a certain natural variation in axial orientation will be statistically predictable within large-scale structures. In some instances however, the distribution patterns within the mean axial surface trends in rocks of the same type and metamorphic grade were not compatible with the general statistical models. Examination revealed direct correlation between axis trend variations and zones of high strain. Ductile deformation in these zones, associated with the folding strain, accomodated rotation of early formed fold axes toward the bulk extension direction. Axes outside the zones were not rotated as extensively, essentially defining a rotation gradient toward the higher strain zones.

Craw (1982) detected similar distribution patterns in the Otago Schists of the Mount Aspiring area, New Zealand and was able to correlate "deviant" fold axis data with high strain zones. The Figure 3:4:7 girdle of DB3 Bass River Complex data shows a data pattern that could be a result of heterogenous strain during folding. In this regard it is quite similar to the situations described by Craw and Sanderson. A comparison of Bass River Complex DB3 fold axis data with pre-existing linear element trends was made by plotting all of the pre-DB3 data in combination with the FB3 fold information. Presumably, these earlier elements responded passively to the superimposed deformation and originally defined a relatively close spatial grouping. Their best-fit girdle is coincident with that which describes DB3 data alone but the larger data population fills the girdle more completely. This indicates that the pre-DB3 elements deformed in much the same way as those that were generated during DB3 folding (Figure 3:4:8).

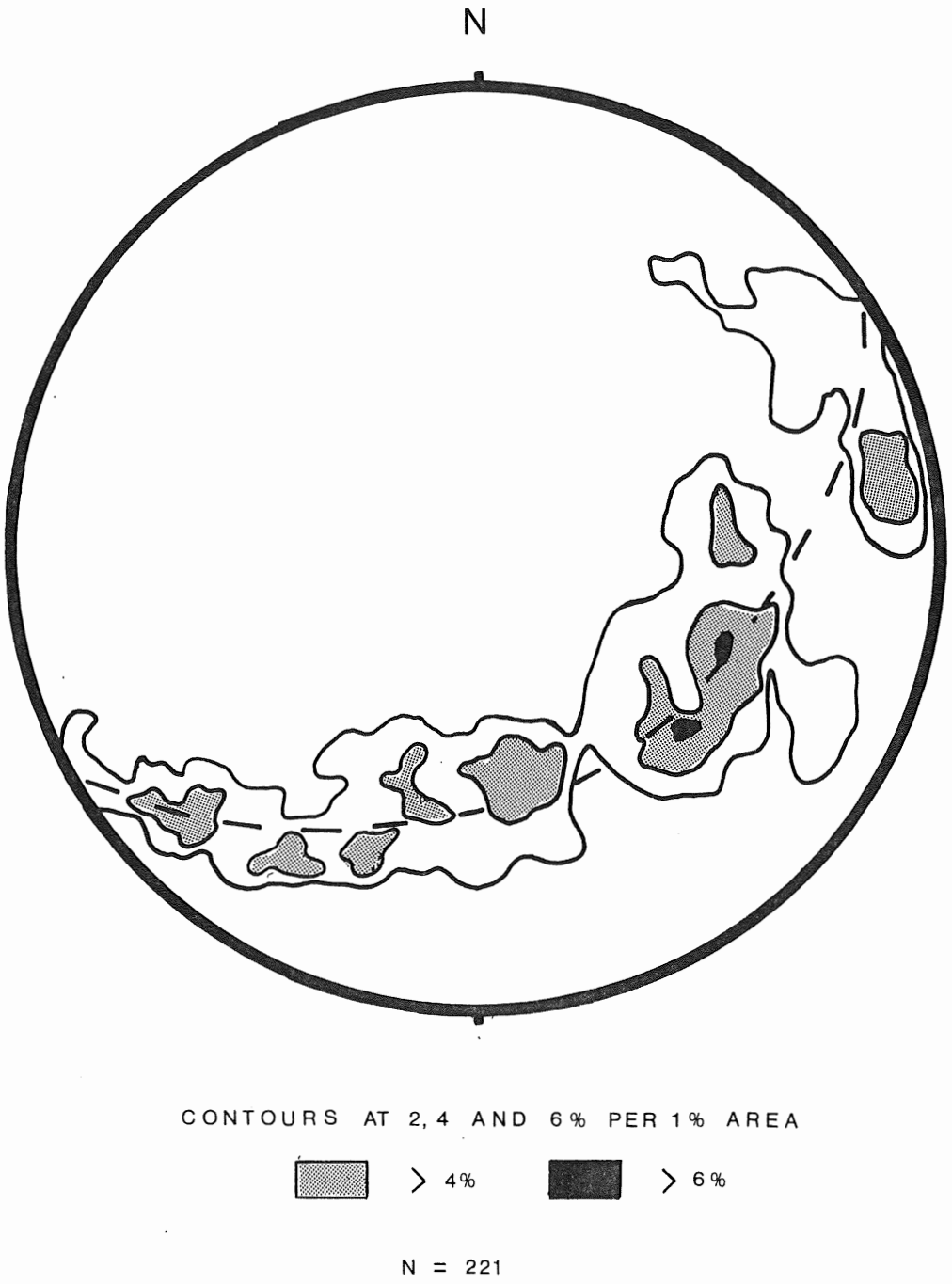


Figure 3:4:8 Synoptic Schmidt equal area plot of DB2 and DB3 linear element data. Note the same northeast trending partial girdle orientation as that of Figure 3:4:7.

The question arises as to which population presently defines the most highly strained (and therefore most extensively rotated) terrane. The axial trend domain west of Rockland Brook may represent a more highly strained equivalent of the eastern configuration. On the other hand, this may mark a later, and distinct, nonpervasive orientation difference between regional finite strain axes across the Complex. As noted above, discrete high strain zones on the outcrop scale cannot fully explain the distribution pattern in either east or west. Reorientation of linear elements must therefore have been accomplished by processes which were heterogeneous on the large scale but essentially homogeneous on the scale of an outcrop. Bulk inhomogeneous shortening or flattening strain could have resulted in the observed trends but the lack of large-scale DB3 shear zones limits applicability of simple shear as the major factor in that portion of the deformation history.

3:5 The DB4a Event - Gamble Lake Fault Movement

A previously undetected east-west trending structure, here termed the Gamble Lake Fault, approximately coincides with the northernmost 4200 gamma magnetic field contour between Portapique and Londonderry on Plate 4. Post-DB3 diorites along the south boundary of the Bass River Complex east of Portapique River locally carry a tectonic fabric attributed to shear along this break. Foliated and lineated diorite fragments occur in the breccia that marks the later Londonderry Fault at Rockland Brook, indicating that for a portion of its length the Gamble

Lake structure was overprinted by brittle DB5 deformation of the Londonderry Fault. The diorite is not pervasively deformed so a correlation of the fabric in the breccia fragments with the DB2 penetrative foliation is not reasonable. The diorites clearly post-date this tectonic event.

The fault trend west of Great Village River is best defined by an aeromagnetic gradient. This distinguishes rocks of the Bass River Complex from the juxtaposed Carboniferous sequence (Plate 5). Magnetic field highs greater than 4200 gammas indicate proximity of the Great Village River Gneiss west of Rockland Brook while signatures less than 4200 gammas generally define the distribution of supracrustal rocks. The higher magnetic signature over rocks of the Gamble Brook Schist on Portapique River probably reflects a relatively thin cover of these rocks over those of Domain A.

A significant gradient in the magnetic field contours occurs between Great Village and Portapique Rivers where a strongly developed east northeast trending anomaly associated with Domain A rocks disappears. From the east the magnetic signature describing the extension of the Bass River Complex from the Great Village River area trends almost perfectly east-west, passing north of this discontinuity and making an angle of approximately 30 with the previously noted trend (Plate 5) that defines the western portion of Domain A. Although the drop in field intensity west of Portapique River may be explained by the presence of overlying metasedimentary cover, the divergence is best explainedby right-lateral displacement along an east-west fault between

Portapique and Londonderry. Apparent strike-slip is estimated at 2 km, assuming an original along-strike continuity of the Strongly Foliated Granitoid rocks on Rockland Brook, Great Village River, and Portapique River. This movement resulted in thinning of the Complex across the regional structural strike between Portapique and Great Village Rivers. To the west of Portapique River the fault approaches the Rockland Brook Fault and ultimately merges with it in the vicinity of Gamble Lake.

Relative timing of movements along the break is difficult to ascertain. The apparent lack of displacement of the eastern Pleasant Hills Pluton (circa 315 Ma.; Donohoe and Wallace (1978)) along the Gamble Lake Fault suggests that major movement predated or accompanied emplacementof that intrusion.

3:6 The DB4b Event - Rockland Brook Fault Movement

3:6:1 General Outline

The Rockland Brook Fault is an east-west-striking, steeply-south-dipping ductile shear zone that forms the northern boundary of the Bass River Complex east of Bass River. Metamorphic rocks of the Complex are juxtaposed against younger igneous and sedimentary rocks occurring on the north (footwall) side according to location along the length of the break. Although it is generally a single well-defined zone, in at least one locality significant splaying of the main shear has taken place, resulting in isolation of relatively undeformed rocks as lenticles in the

movement zone.

3:6:2 Nature of the Shear Zone

On both East Folly River and Higgins Mountain Road rocks of a diorite-granite stockwork define a strong strain gradient over a few tens of meters toward the shear zone. Foliation and mylonitic layering trend generally east-northeast to northeast and dip south at greater than 60°. Progressive elongation of diorite xenoliths was accompanied by development of a mineral lineation parallel to the long axis of xenoliths. The lineation trends southeast on Higgins Mountain Road and east southeast on East Folly River (Figure 3:6:1a,b,c; Plate 4). Adjacent, pre-existing fabric elements in Bass River Complex rocks also appear to have been rotated northeastward, since trends and plunges are parallel to those within the shear. Local isoclinal folding of the igneous protolith is defined by sheared granitic veins of the stockwork-xenolith zone and axial trends of these rare structures appear to be essentially coaxial with the lineations noted above.

The East Folly River exposures are similar to those on Higgins Mountain Road and Rockland Brook, where diorite xenolith-bearing granite stockwork forms the footwall, but metaquartzites form the hanging wall. An intense mineral lineation and surface rodding coincident with small-scale fold axial trends mark the metamorphic sequence for approximately 90 m south of the first outcrops of the plutonic rocks. These folds (FB4b generation of this study) differ from earlier regional types by being open to tight, chevron type structures with centimeter-scale

Figure 3:6:1(a) Undeformed rocks of the granite-diorite contact or "stockwork" zone north of the Rockland Brook Fault.

Figure 3:6:1(b) Foliated granite - diorite contact zone rocks showing a planar preferred orientation of vein and xenolith boundaries. Note the small, isoclinally folded granite vein at the right center of the photo. This oucrop is about 10 m from the shear zone axis on East Folly River.

Figure 3:6:1(c) Mylonitic equivalent of rocks in Figure 3:6:1(b); photo shows rocks that are characteristic of the shear zone axis. Note the strongly-developed mineral lineation and the absence of original igneous texture. The hammer handle is oriented east-west, north is to the bottom in the photo.

wavelengths and amplitudes. A partial progression in folding style has been identified toward the zone axis, since tight to isoclinal structures in the high strain areas could represent more highly deformed equivalents of the more open folds noted above. To the north the limit of penetrative shearing is probably marked by the presence of foliation in granite-diorite stockwork rocks and to the south by the extent of rodding in the metasedimentary rocks of the Bass River Complex..

Shearing along the Rockland Brook Fault could not have been related to the deformation that produced the DB2 linear-planar fabric because tectonite xenoliths of this generation occur in undeformed parts of the Folly Lake Pluton and Hart Lake-Byers Lake Pluton. These Devonian to Carboniferous bodies intrude the latest regional folds of the Bass River Complex and were themselves sheared by movement along the Rockland Brook Fault.

On Rockland Brook the shear zone is complex in that at least 2 distinct, major splays have been identified (Figure 3:6:2). The northernmost occurs within the diorite-granite stockwork terrane and exhibits the complete progression from undeformed to mylontic textures when approached from the north. A similar textural trend presumably exists in the south, but the axis' location at the edge of a waterfall plunge pool prevents close inspection of immediately adjacent rocks. The presence of moderately deformed to undeformed granite-diorite stockwork south of the north splay does, however, allow definition of the general structural scheme. The southern splay marks the actual boundary of Bass

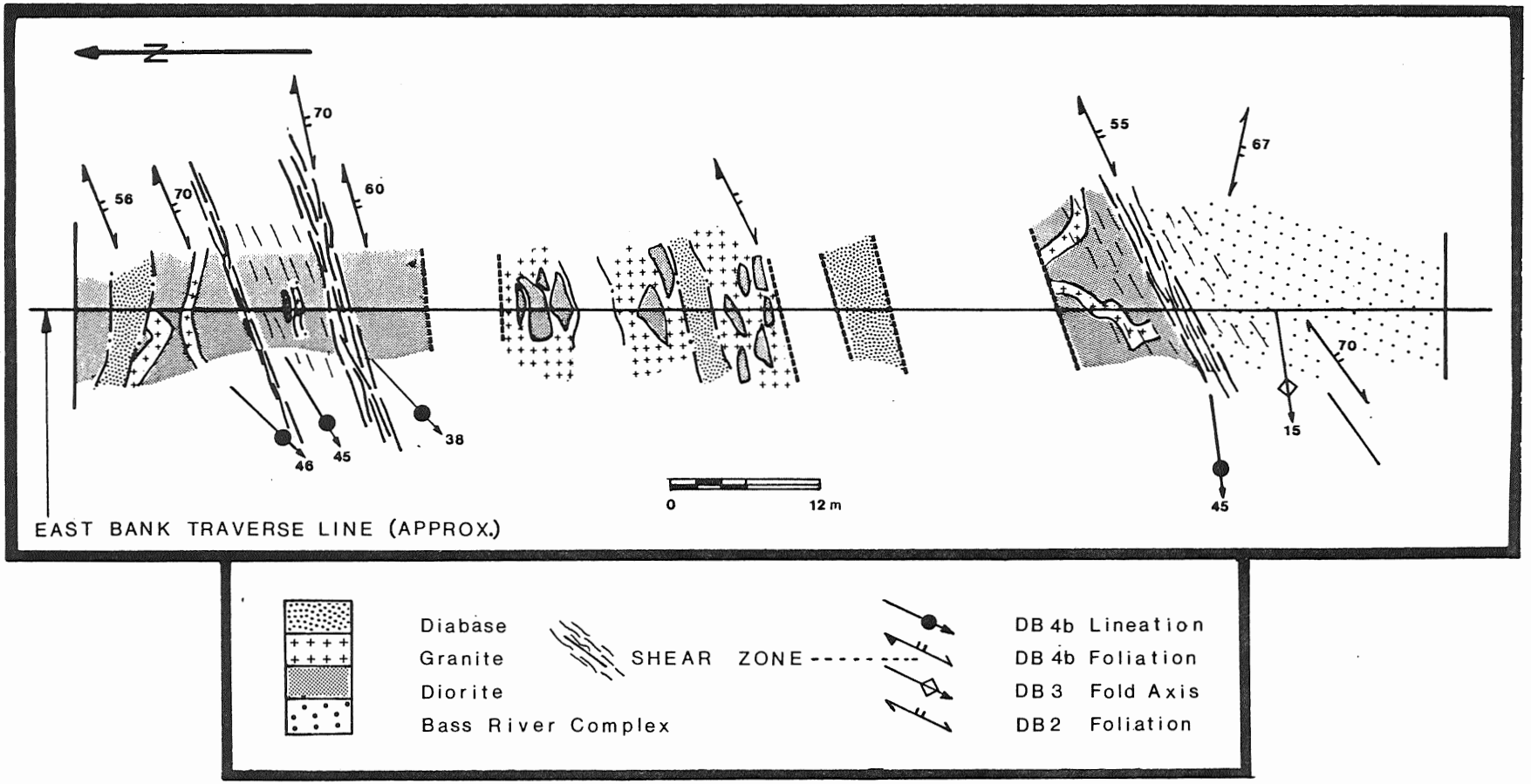


Figure 3:6:2 A sketch map of field relationships at the Rockland Brook Fault on Rockland Brook. Note that the shear zone has two distinct splays at this location separated by undeformed granite-diorite contact zone rocks.

River Complex rocks with the stockwork zone structural lenticle and a strain gradient toward the zone axis is obvious (Figure 3:6:3).

In the Debert River area the Rockland Brook Fault is marked by a 15-20 meter wide mylonite zone of light grey to white quartz and feldspar which exhibits an extremely well developed linear-planar fabric. An inconsistent foliation detectable for approximately 100 m northward from the shear zone axis in the Hart Lake - Byers Lake Pluton locally defines the deformation zone within that otherwise nonpervasively deformed igneous complex. The effect of shearing on the metavolcanic sequence is not clear, for these rocks are not seen in direct association with the highest strain zone along the fault trace.

Donohoe and Wallace (1978, 1980, in preparation) considered the Rockland Brook Fault to be truncated by the Lower Carboniferous (circa 315 Ma) Pleasant Hills Pluton. Major displacements had, no doubt, ceased by this time. However, field evidence obtained in the present study requires revision of this interpretation. The movement zone meets the inferred eastern contact of the pluton at a high angle southeast of Gamble Lake (Plate 2) but a clear truncation of the associated fabric is not seen. A significant east-trending mylonite zone marks the pluton's contact with the Great Village River Gneiss on Bass River and may mark a splay of the Rockland Brook shear west of the Gamble Lake area. Accurate further extension westward is currently not possible but aeromagnetic anomalies can be interpreted as indicating that the Rockland Brook Fault continues west-southwest and ultimately merges with the

Figure 3:6:3 Strain gradient on Rockland Brook at the contact of granite-diorite contact zone rocks with those of the Bass River Complex. Strained diorite xenoliths are visible near the hammer head but not at the left edge of the photo, where these rocks and metamorphic rocks of the Complex have been mylonitised and tectonically interlayered. The photo was taken facing west and mylonitic layering is steeply south dipping.

east-trending Cobequid Fault in the Economy River area.

Ductile shear along the Rockland Brook Fault must have predated the brittle faulting which affected Lower Carboniferous sedimentary rocks and older metamorphic rocks of the Highlands which are juxtaposed across the Londonderry Fault. The presence of sheared zones within the Pleasant Hills Pluton does indicate that Rockland Brook Fault movement did not entirely pre-date emplacement of the granite.

The regional structural trend of the Complex between the northern and southern boundary faults can be used to define relative movement sense along that structure. The regional DB3 fold trends (and therefore the overall structural grain of the Complex) change as they approach the shear zone. Their generally northeast trend swings easterly into near-parallelism with the Rockland Brook Fault, producing an orientation pattern that would be expected adjacent to a right-lateral shear. Dextral movement on this fault is compatible with a similar displacement sense previously determined (chapter 3, section 3:5) for the Gamble Lake Fault and it should be borne in mind that these structures merge in the Gamble Lake area.

Indirect evidence of a differing displacement sense for the Rockland Brook Fault is found in minor shear orientation data from Folly Mountain power line, Debert River area, and Higgins Mountain Road. In these areas numerous discrete, northeast to east-northeast trending, steeply dipping shears up to 1 m in width affect otherwise undeformed Devonian to Carboniferous rocks of the granite-diorite contact zone. A

schematic combination of this minor shear data with the Rockland Brook Fault trend defines at least part of a classic Reidel shear geometry (Figure 3:6:4). The main Rockland Brook Fault trend corresponds to the primary shear orientation and the north-easterly trending associates represent the high-angle reverse-sense secondary structures. These should not be confused with that set of shears that developes at less than 15 degrees to the primary shear. If this interpretation is correct, their orientation supports left lateral movement along the Rockland Brook Fault after emplacement of the Hart Lake - Byers Lake Pluton (circa. 315 ma.; Cormier, 1979). This determination of movement sense is opposite to that indicated by the regional structural trends but is considered to be less reliable. A more detailed study of rocks within and adjacent to the shear zone would be necessary to fully substantiate either argument.

3:7 The DB4c Event

A detailed 1:10,000 air photo interpretation was undertaken to study surface expressions of the major and minor faults of the study area. Azimuths of all linear topographic features described by 5m contouring were recorded on a grid of the study area and classified according to bedrock geology, location and magnitude. The first two parameters were derived from geology and reference-grid overlays and feature magnitude determined using the following criteria:

(1) Magnitude 3 > 1 km in length;

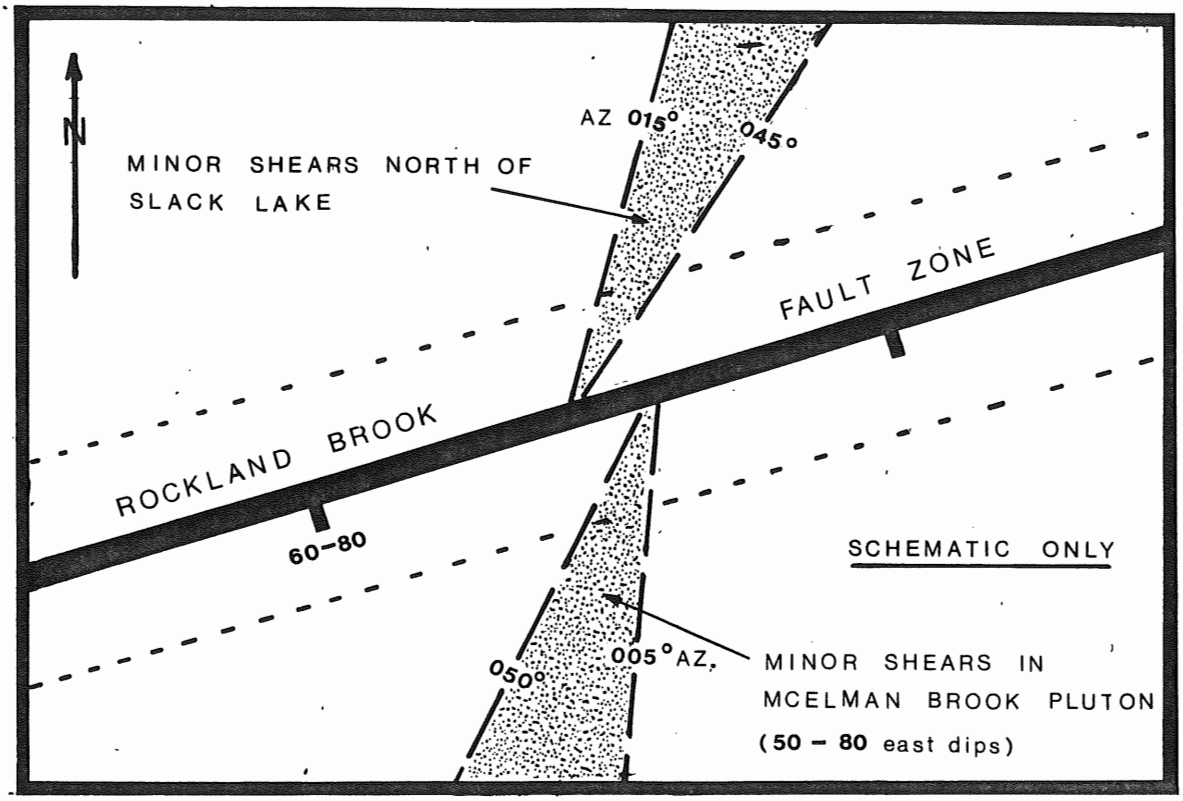


Figure 3:6:4 A comparison of Rockland Brook Fault major and minor shear orientations. The inferred Reidel shear configuration indicates left-lateral movement sense.

(2) Magnitude 2 < 1 km in length > 0.5 km in length;

(3) Magnitude 1 < 0.5 km in length.

Orientation plots for these features are presented in the Appendix .

A first-order categorisation of data was made based upon feature magnitudes. Magnitude 3 data include east-west structures parallel to the Cobequid Fault zone and are the most continuous and pronounced. Magnitude 2 features are generally those trending between 345 Az. and 015 Az. whereas those of magnitude 1 describe a more complex distribution pattern, north of the Cobequid Fault, in both areas. Correlation of overprinting relationships seen in bedrock with those in the photo interpretation data is most clearly evident from the left-lateral apparent displacement of magnitude 3 (070-090 Az.) structures by the 345-015 Az. magnitude 2 trends. This is a reflection of the well documented relationship (Donohoe and Wallace, 1980) between the north-south and east-west trending faults and fractures: the north-trending faults offset the earlier east-west structures and can be shown to have accomodated normal dip-slip movement

Magnitude 1 features can be detected in all rocks of the area studied north of the Londonderry Fault but are best developed east of Portapique River. This trend domain coincides with the Carboniferous and Devonian age intrusions and is probably a reflection of the homogeneity of these rocks relative to the markedly heterogeneous layered rocks of the Bass River Complex. The orientation pattern is interpreted as representing a conjugate joint or fracture set, separated by internal angles

in the 25° to 45° range, which developed in all rocks of the highland that are now north of the Londonderry Fault trace. Fractures were generated after movement on the Gamble Lake and Rockland Brook faults, but prior to activation of the Londonderry Fault. This interpretation of timing is qualified by the apparent absence of the conjugate pattern south of the Londonderry Fault. Changes in the mean conjugate bisector trends occur from west to east (Figure 3:7:1) indicating, possibly, a corresponding change in paleostress trajectories along the length of the map area. The regional significance of this trend may be related to the arcuate trace of the Londonderry Fault but this would be difficult to prove. Their generation has therefore been ascribed to a post-DB4b paleostress configuration, DB4c, of undetermined origin.

3:8 The DB5 Event - Londonderry Fault Movement

The Londonderry Fault forms the southern boundary of the Bass River Complex for almost its entire length, coinciding with, and overprinting, the Gamble Lake Fault east of Great Village River and merging with the Cobequid Fault near Economy River. Foliated diorite, metamorphic rocks of the Bass River Complex, Londonderry Formation sedimentary rocks, and deformed granitoid rocks occur as breccia fragments along the fault trace. These generally reflect the variability of hanging and footwall rocks along strike and down-dip. The width of the transition from the Lower Carboniferous rocks of the hanging wall to metamorphic or igneous rocks of the footwall changes according to location. On Rockland Brook a

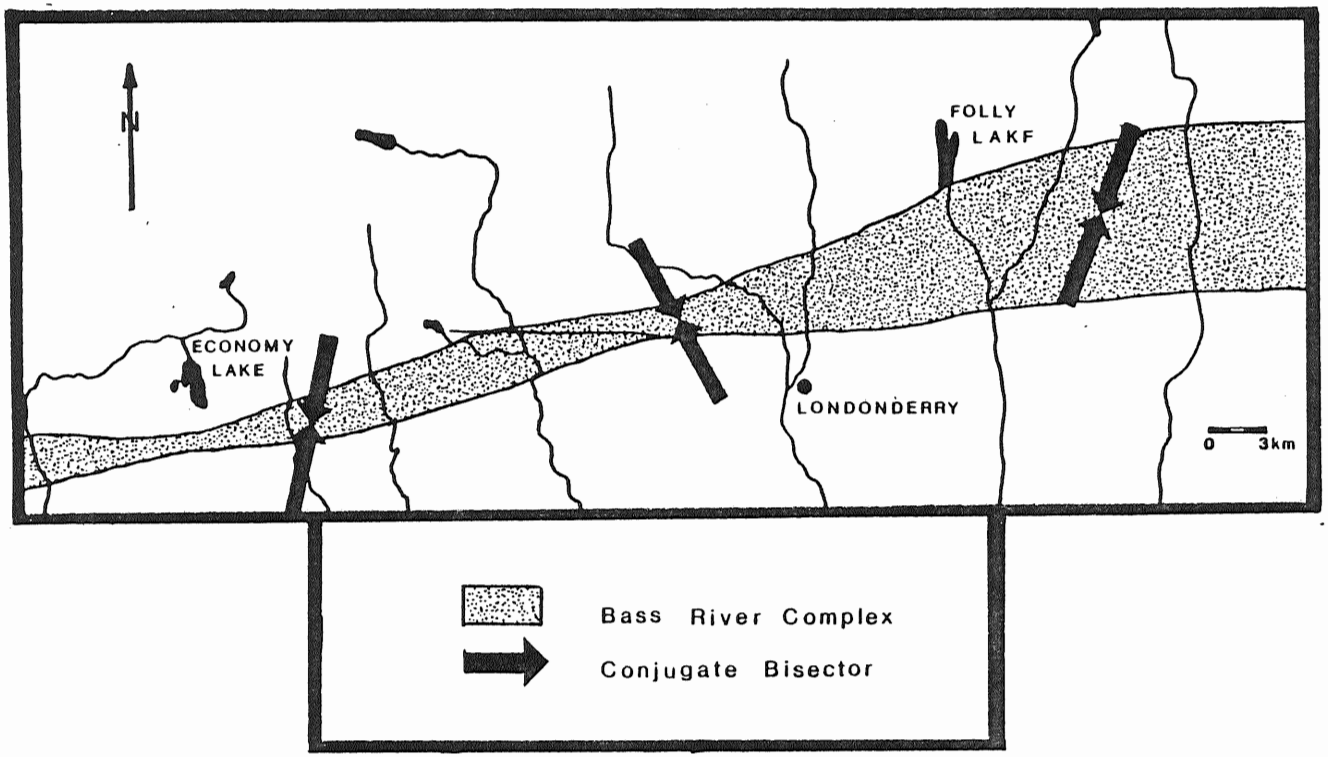


Figure 3:7:1 An outline map of conjugate bisector trends derived from analysis of airphotos.

complex 15 m wide breccia and lenticle zone marks the fault and on Folly River brecciation of both Londonderry Formation and Bass River Complex rocks is apparent over a 10-15m section. Where movement has been distributed over a number of subparallel zones the isolated fault lenticles occasionally bear reclined, minor folds of < 35 cm wavelength whose axes rake at low to moderate angles within fault zone. Donohoe and Wallace (1980) link cleavage development and asymmetric folding of Londonderry Formation rocks south of the fault to movement along the fault.

The Bass River section is unusual in that the trace of the fault is marked not only by brecciation but also by a zone of ductilely deformed carbonate (Figure 3:8:1). The carbonate is considered part of a thin (<12 m true thickness) fault zone lenticle which hosts the Bass River magnetite deposit described by Wright (1975).

3:9 The DB6 Event - North South Fault Movements

Previous workers (Weeks, 1948; Stevenson, 1958; Kelley, 1965-1969; Eisbacher, 1967; Donohoe and Wallace, 1980) have recognised north-south structural breaks within the Cobequid Highlands. Because drainage patterns in the Highlands are generally north-south, river sections in most cases do not cut the fault traces and evidence of the actual break is often obscured by stream bed or valley-side debris. Where they can be documented in outcrop, zones are marked by moderate to intense, nearly vertical fractures and breccia zones. In some cases, such as along Bass

Figure 3:8:1 A zone of ductilely deformed carbonate within the Londonderry Fault on Bass River. The carbonate appears to be associated with a thin structural lenticle that hosts the Bass River Magnetite Deposit described by Wright (1975). The carbonate zone strikes generally east-west and dips to the south. North is to the bottom of the photo.

River, a strong north-south fracture cleavage associated with the main fault can be detected for 15 to 20 m across strike (Figure 3:9:1). Discrete breccia zones have been identified in both the metasedimentary and metavolcanic sequences ranging in width from 25 cm to 2 m and mappable along-strike for up to 100 m. No characterstic hydrothermal alteration or dikes mark these faults but Donohoe (1983) suggests that passage of the fluids which deposited some of the iron in rocks of the Londonderry Formation was controlled locally by north-south brittle structures.

Dip-slip apparent displacements are indicated by the outcrop patterns mapped by Donohoe and Wallace (1980) in the Carboniferous to Jurassic rocks adjacent to the Highland. They cite a Lower Jurassic or younger age for this fault generation on the strength of offsets of the Triassic-Jurassic Scots Bay Formation.

Figure 3:9:1 North-trending fracture cleavage associated with late stage DB6 faulting. This exposure of Londonderry Formation rocks occurs a short distance south of the Bass River Complex on Bass River. North is to the top of the photo.

CHAPTER 4

METAMORPHISM

4:1 General Outline

At least one distinct metamorphic event affecting the terrane now known as the Bass River Complex was recognized by early workers (Ells 1885: Fletcher 1905: Weeks 1948) and Eisbacher (1967, pp. 48) described rocks of his mylonite schist zone as a very low grade metamorphic or sub-metamorphic association. He attributed fabric development to Devonian translation accomplished by laminar flow of the various lithologies. In contrast, Donohoe and Wallace (1980; in preparation) recognised that greenschist and possibly amphibolite facies rocks occur within the Complex. The present study supports the conclusion of Donohoe and Wallace in regard to metamorphic grade and documents relict high grade mineral assemblages in the newly defined Portapique River Schist unit.

Because variation of metamorphism with time is interpreted on the basis of textural relationship of mineral assemblages to tectonic fabrics, metamorphism in the Complex is discussed in terms of the corresponding deformation phases. Table 4:1:1 has been included at this point to clarify salient points and outline related terminology. Section 4:2 deals with mineral assemblage distribution and relative timing of deformation and metamorphism is discussed in section 4:3.

TABLE 4:I:I

SUMMARY OF METAMORPHISM

Metamorphic Grades

Phase	Structural Domain	Low Grade	Medium Grade	High Grade
M1	A Only			
M2	A + S	Domain S / Domain A		
M3	A + S			
M4	S + A ??			
M5	A + S			

Mineralogy

Phase	Structural Domain	Characteristic Metamorphic Mineral Assemblages
M1	A Only	sillimanite, staurolite, garnet, quartz, K feldspar?,
M2	A + S	Domain S – pelitic rocks: biotite, muscovite, quartz, chlorite, garnet, albite, epidote, opaques – mafic rocks: chlorite, actinolite, biotite, quartz, albite, opaques, minor hornblende Domain A – pelitic rocks: biotite, muscovite, garnet, quartz, chlorite, opaques – mafic rocks: hornblende, biotite, garnet, calcic plagioclase, epidote calcite, quartz, opaques
M3	A + S	chlorite associated with DB3 crenulation limbs
M4	S + A ??	chlorite, sericite, diopside, tremolite: contact metamorphic effects documented in Domain S
M5	A + S	chlorite, sericite, kaolinite, quartz: shear zone related retrogression

4:2 Distribution of Metamorphic Mineral Assemblages

The history of metamorphism in Bass River Complex is defined by five phases.termed M1. M2. M3. M4 and M5. Characteristic mineral assemblages attributed to each are presented in Table 4:1:1 with corresponding outlines of relative distributions.

The M1 metamorphic event coincided with the first recognisable deformation (DB1) and resulted in crystallisation of amphibolite facies (maximum of sillimanite grade locally) mineral assemblages that define the DB1 fabric (Table 4:1:1). No FB1 folds have been identified. This early deformation was followed by the DB2 event which affected all rocks of both structural domains and was accompanied by greenschist to lower amphibolite facies M2 metamorphism. The second generation fabric almost totally overprinted SB1 but, as noted in the preceeding chapter. relicts of the early layering have been identified in some rocks and have been used to distinguish Structural Domain A from Structural Domain S (Figure 4:2:1).

Mineral assemblages that define M2 change across the Complex from chlorite - biotite - garnet grade in pelitic supracrustal rocks to amphibolite (calcic plagioclase - hornblende - garnet) grade in mafic rocks of the Great Village River Gneiss. Table 4:1:1 presents in more detail characteristic mineral assemblages for the major lithologic units. DB2 generation folds. of the early metamorphic layering in Domain A and primary layering in Domain S. bear this metamorphic foliation as

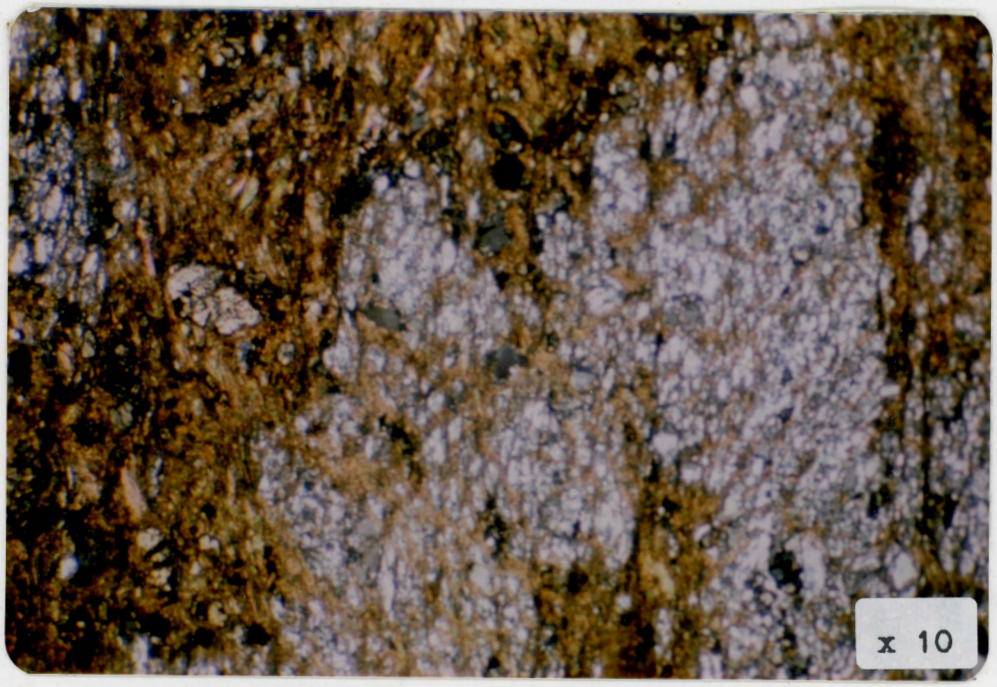


Figure 4:2:1 Biotite-grade transpositional fabric overprinting DB1 generation quartzo-feldspathic layering. Note the rounded garnet and staurolite porphyroclasts in the left portion of the photo. The slight crenulation of the biotites may be related to subsequent DB3 folding. This sample is from Rockland Brook outcrops of the Portapique River Schist. (crossed polars)

an axial planar fabric and also locally show mimetic MB2 metamorphic mineral overgrowths of the pre-existing layering . Post-metamorphic-peak crystallisation of chlorite is associated with DB3 generation crenulations and marks the latest, regional-folding- associated metamorphic imprint, M3, seen in rocks of both structural domains. The occurrence of a well-developed DB3 planar fabric is restricted to relatively tight fold closure zones in the least competent rocks of the Complex but an equivalent quartz grain-shape fabric is often evident in rocks lacking a chlorite-defined cleavage (Figure 4:2:2a,b). M4 contact metamorphism is locally significant and is defined by phyllosilicate crystallisation, restricted calcsilicate crystallisation, and extensive retrogressive kaolinisation and sericitic alteration. All of these are spatially restricted to the immediate contact zones of late Precambrian and younger intrusions. M5 marks retrogressive alteration of rocks that occur within the Rockland Brook and Gamble Lake fault zones. In these zones mafic minerals show extensive chloritisation and feldspar shows sericitisation and kaolinisation. This type of low grade alteration primarily affected rocks of the Bass River Complex and Devonian to Carboniferous intrusions that are cut by the the shear zones.

4:3 Metamorphic Culmination - Strain Relationships

Petrographic study of nearly two hundred thin- and slab-sections, in combination with field observations, allowed definition of distinct fabric-overprinting relationships. This information forms the basis of a

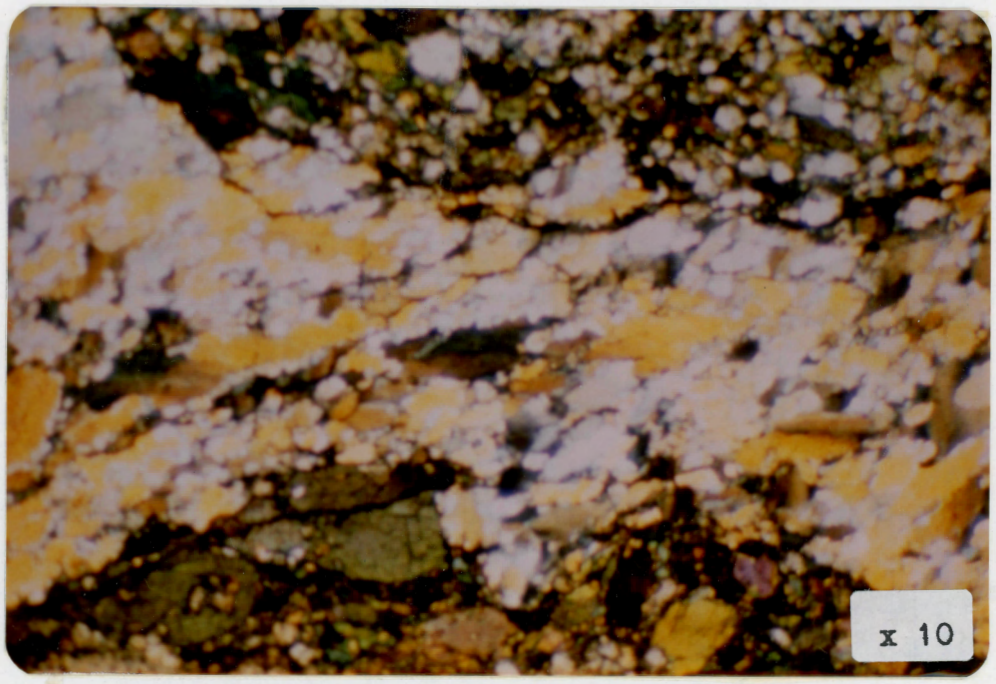


Figure 4:2:2(a) A strongly developed DB3 generation quartz grain-shape fabric in a sample of amphibolite from Higgins Mountain Road. The sample was collected from the outcrop section pictured in Figure 3:4:3. (crossed polars)

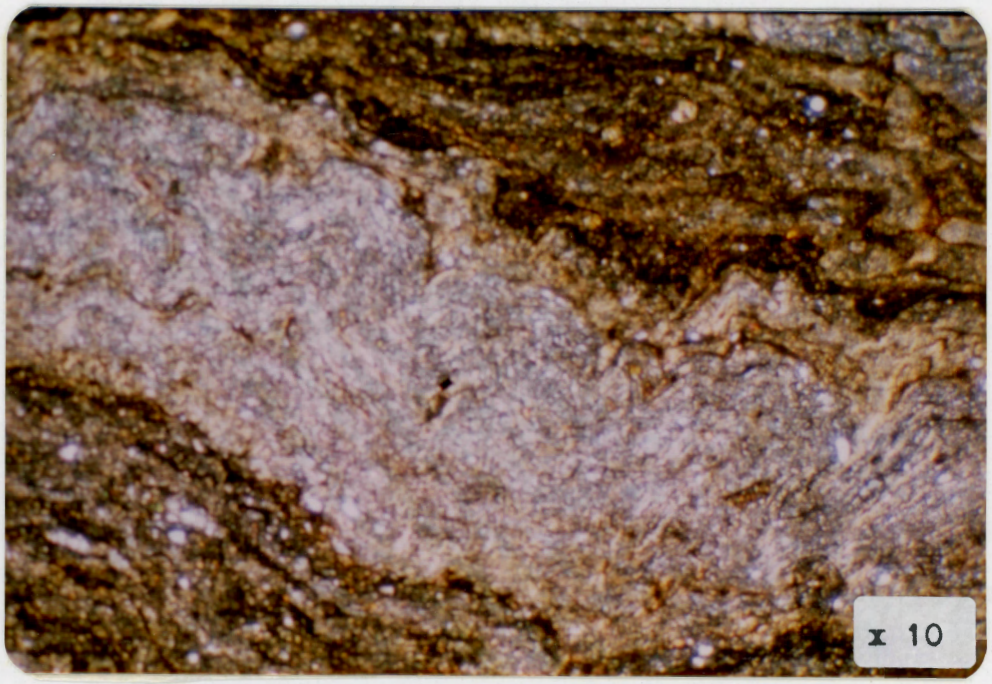


Figure 4:2:2(b) DB3 generation crenulations of DB2 fabric in schist from the Debert River area. Close inspection reveals a weakly-defined chlorite fabric at about 50° to the dominant layering on the right side of the photo. (crossed polars)

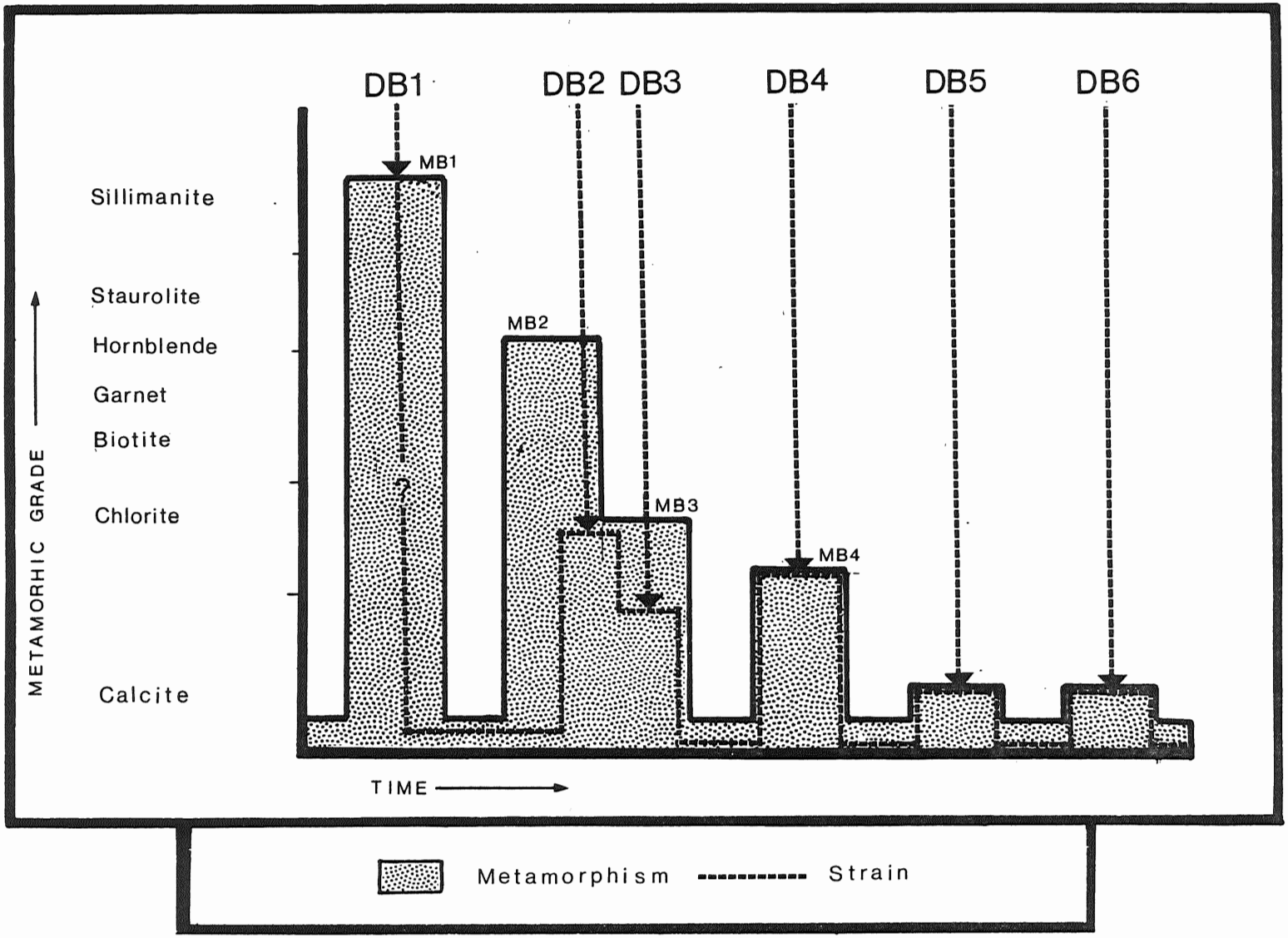
generalised correlation of deformation and metamorphism that is presented in Table 4:3:1. All relationships were determined with consideration of the textural qualifications presented by Spry (1969), Vernon (1978), and Park (1969).

Domain A

The systematic overprinting as described in Table 4:3:1 is self-evident but alternative interpretations are possible. This is most apparent in regard to the high grade M1 index minerals (sillimanite + staurolite + garnet) where arguments for either a syn- or pre-kinimatic crystallisation history can be developed. Both options are considered in this section.

Brittle deformation of all high grade phases is pervasive and warping of the DB2 host fabric into micro-pinch-and-swell neck zones is common between sillimanite crystal fragments (Figure 4:3:1). Fractured M1 staurolites and garnets and some M2 hornblendes bear evidence of rigid body rotation within the same fabric (Figure 4:3:2). Therefore, these may be interpreted as early syn-kinematic M2 phases which were strained later in that same deformation period. The observation that DB2 deformation was characterised by intrafolial zones of heterogeneous high strain is consistent with this interpretation. Grainshape modification of early M2 porphyroblasts that took place under ambient M2 conditions would be difficult to discern from that of pre-kinimatic (ie. M1) porphyroblasts if crystallisation of minerals such as biotite took place both syn- and post-kinematically in DB2.

TABLE 4:3:I


DB1
DB2 DB3
DB4
DB5
DB6
MB1
MB2
MB3
MB4
Sillimanite
Staurolite
Hornblende
Garnet
Biotite
Chlorite
Calcite
METAMORHIC GRADE
TIME
Metamorphism
Strain

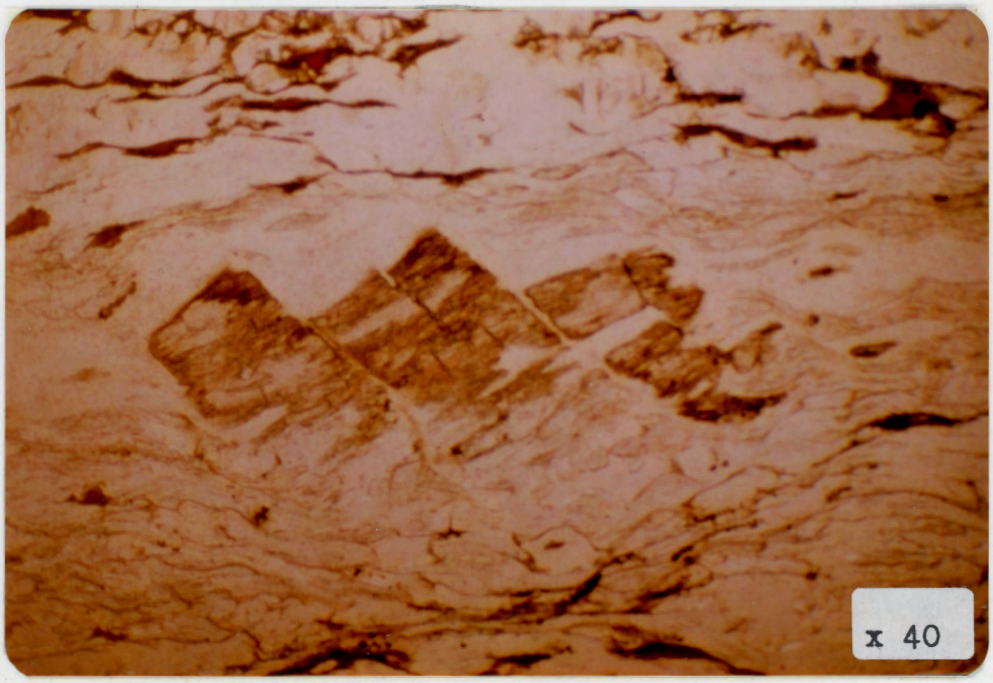


Figure 4:3:1 DB1 generation sillimanite that has been extensively replaced by muscovite. Fracture and rotation of the crystal fragments took place during the DB2 deformation that emplaced the well defined biotite - quartz - muscovite fabric. (plane light)

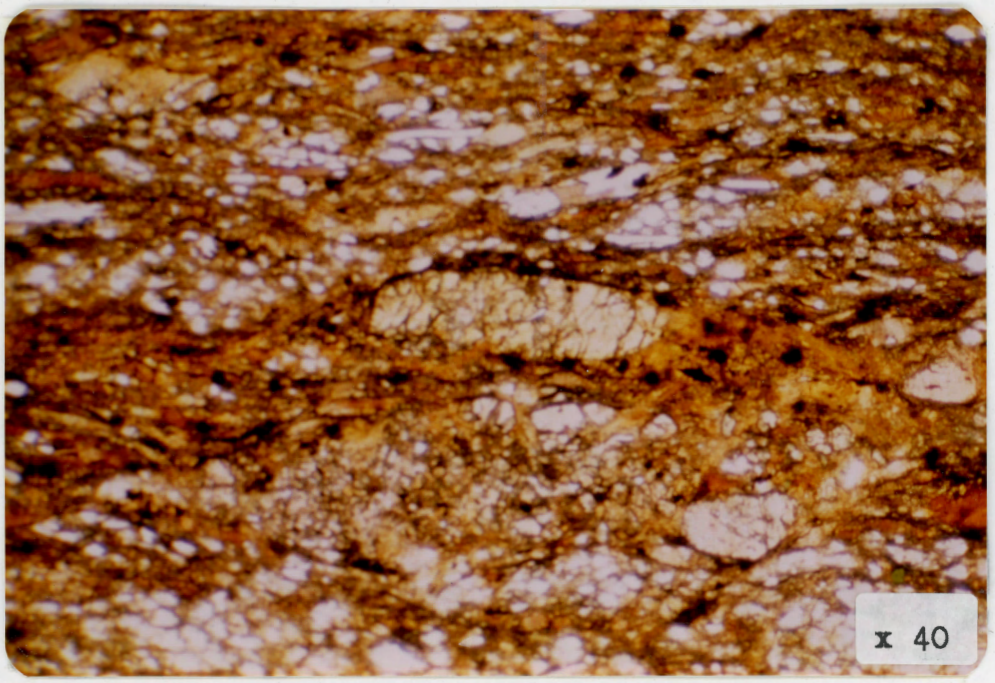


Figure 4:3:2 DB1 generation garnet and staurolite porphyroblasts that were altered, fractured, and rotated during DB2 overprinting. Sample is from the same outcrop section as that of Figure 4:3:1.

Syn-kinimatic crystallisation of high grade minerals can be accommodated within the DB2 framework as noted above, but at least two arguments are apparent that strongly favour their associtions with a distinct M1 metamorphic event that was unrelated to subsequent thermal regimes.

(1) The highest grade rocks are known to occur in relatively narrow, poorly defined zones parallel to the DB2 fabric. This poor definition is a result of tectonic interleaving with lower grade rocks of both structural domains. The strong regional fabric in the lower grade lithologies of structural Domain S is correlated with the overprinting, retrogressive, transposition-related fabric in the higher grade rocks of structural Domain A. It is unlikely that these apparently small, interleaved slabs could have transcended the necessary change in burial depth required by the early M2 crystallisation hypothesis because rapid vertical movement of many kilometres would be required. Given the interleaving relationship with lower grade rocks, this seems improbable.

(2) An independent line of reasoning can be developed based on contact relationships of Strongly Deformed Granitoid rocks with those of both structural domains. The M2 fabric penetratively deforms the granitoid rocks in every exposure but clear intrusive relationships can be defined with rocks of each structural domain. This indicates that prior to DB2 and M2, Domains A and S existed at similar levels in the crust. Anatectic assimilation of country rock is not evident, for xenolith boundaries in both domains are sharp. Anatexis would be ex-

pected adjacent to granitic intrusions that had been emplaced under sillimanite grade T-P conditions. It follows then, that an older high grade metamorphic terrane was in the immediate proximity of the lower grade Domain S sequences prior to their initial deformation and metamorphism. In this light, the interleaving of rocks from the two domains is best explained by infolding of autochthonous or possibly allochthonous cover sequences with older basement material.

Domain S

In Domain S, euhedral garnets of the M2 generation generally overgrow both primary layering and the DB1 fabric (Figure 4:3:3), indicating a locally post-kinimatic metamorphic peak. This does not preclude, however, a history of DB2 and M2 deformation and metamorphism similar to that outlined for Domain A. After formation of the axial planar chlorite- to garnet-grade fabric in the supracrustal rocks, extensive transposition-related shearing took place under ambient metamorphic conditions. Presumably this contributed to the relatively abrupt metavolcanic to meta-sedimentary transition zone. The change from a more homogeneous early DB2 strain distribution, presumably characterised initially by buckling, to one characterised by distinct shear-related heterogeneities produced a corresponding textural spectrum incorporating both static overgrowth and rotational features. Ideally a strain gradient between rotational and non-rotational DB2 phases in Domain S should be observed. Intact rotational garnets have not been observed in Domain S but garnet fragment trains were detected in samples taken from this volcano-sedi-

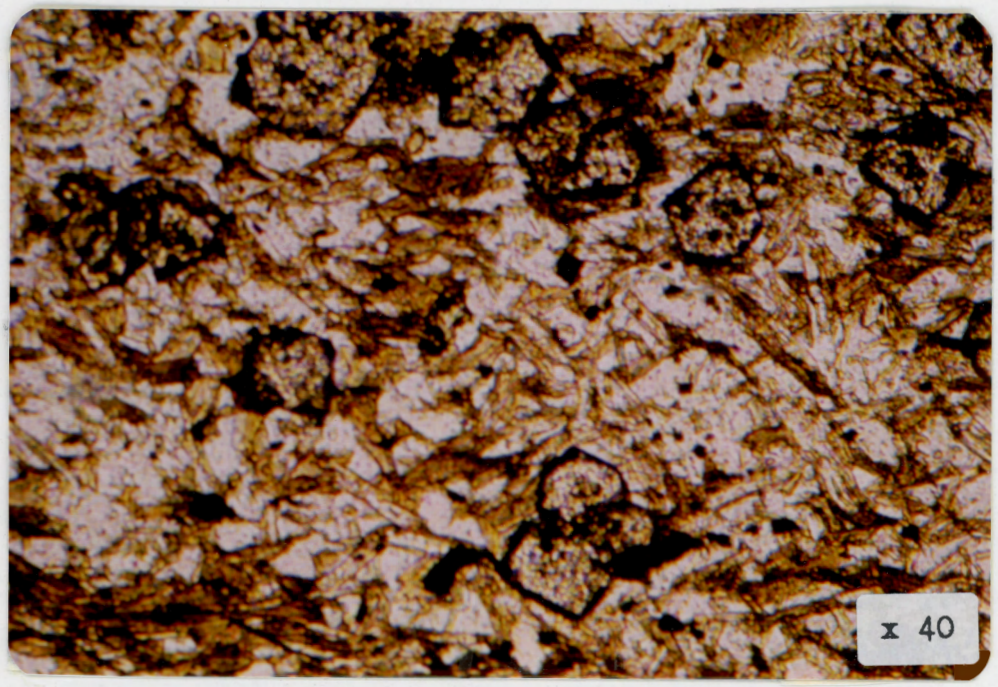


Figure 4:3:3 Euhedral DB2 generation garnets overgrowing the DB2 foliation in a metaquartzite from Folly River. This represents the static growth end member referred to in the text (page 153). See Figure 4:3:4 for the larger scale significance of the overgrowth relationship.

mentary sequence. Figure 4:3:4 shows how M2-peak garnets of the supra-crustal sequence are inferred to have both overgrown and been porphyroclastic within the same generation fabric.

4:4 Post-M2 Metamorphism and Alteration

Local crystallisation of M3 chlorite and flattening of quartz to describe the SB3 axial planar fabric were apparently not regionally significant. Throughout the Complex no post-M2 retrogression is evident except that related to intrusions (M4) or shear zones (M5). The presence of M3 chlorite in association with the late crenulation cleavage, SB3, is interpreted as indicating folding-related crystallisation under falling temperatures following the M2 metamorphic culmination. Regional folding may therefore be considered either the last major increment of strain in a DB2 continuum or a distinct DB3 event closely related in time. The second possibility is favoured considering the sporadic distribution of the DB3 fabric.

Subsequent to M3, local contact metamorphic crystallisation of chlorite (and locally of tremolite and diopside) and shear zone-related retrogression characterise the M4 and M5 imprints, respectively.

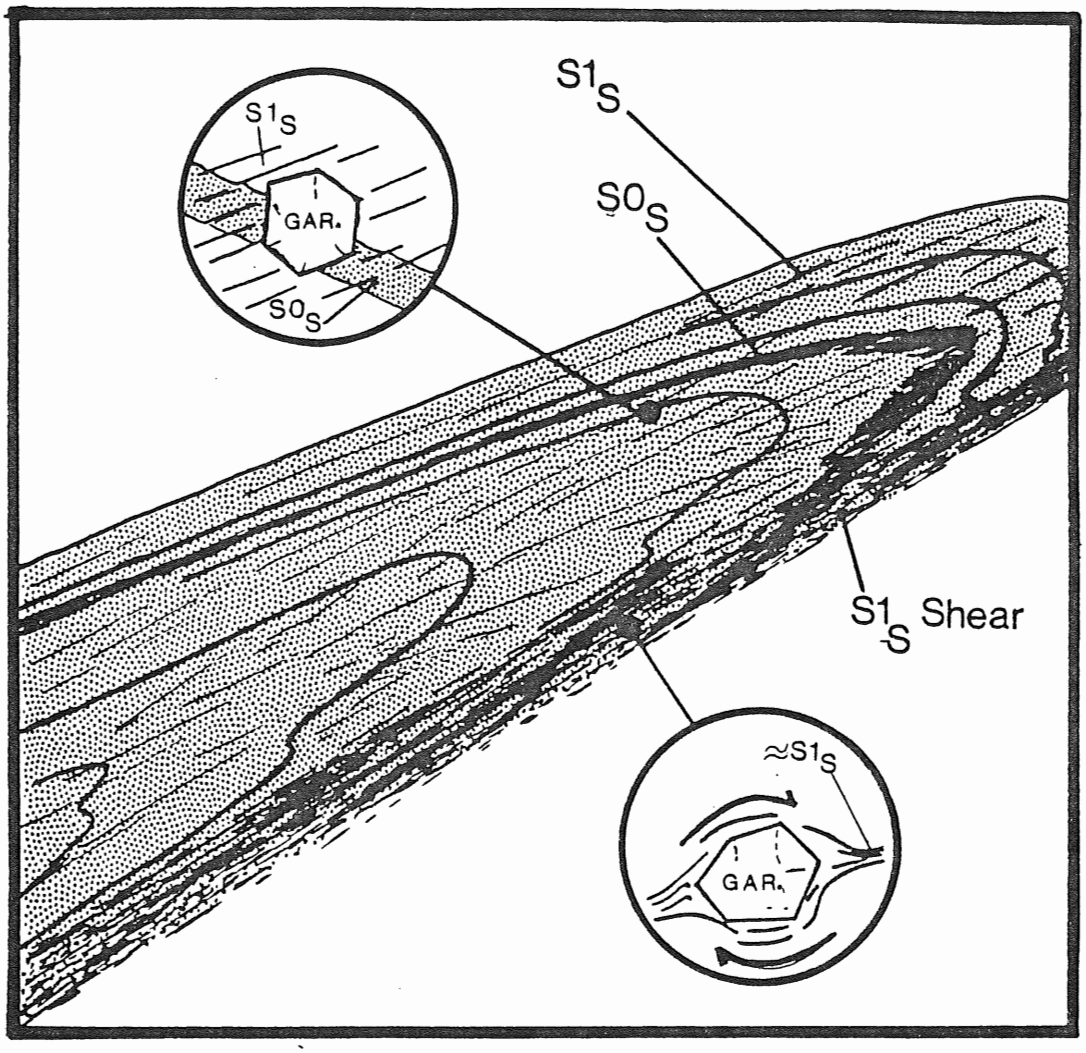


Figure 4:3:4 Idealised relationship between DB2 generation overprinting garnets in Structural Domain S and discrete zones of DB2-related shear strain. Both static growth and rotational histories can be accommodated in this framework. See Figure 3:3:8 (a, b, and c) for the corresponding outcrop - scale structural progression in rocks of the Gamble Brook Schist.

CHAPTER 5

DISCUSSION

5:1 Summary of Structural and Metamorphic Histories

Structural and metamorphic evolution of the Bass River Complex can be described by six distinct deformation episodes, ranging in age from Early Hadrynian or older (DB1) to post-Early Jurassic (DB6).These varied greatly in style and intensity. Little evidence remains of the DB1 metamorphic fabric but quartzo-feldspathic compositonal layering; relict sillimanite indicates that medium pressure upper amphibolite facies metamorphic conditions were reached locally. Distribution of fabric relicts of this generation defines the structural Domain A, or basement terrane. The intense DB2 event deformed Domain A orthogneisses, paragneisses and amphibolites plus the younger Domain S volcanic-sedimentary sequence. The latter is composed of the Folly River Schist and Gamble Brook Schist which bear no DB1 fabric. DB2 mobilisation resulted in intense shortening of the basement-cover package by tight to isoclinal FB2 folds. This may have been marked by syntectonic emplacement of some granitoid bodies. The ultimate DB2 configuration was a transposed structural package showing diachronous metamorphism and strain peaks. Stacking of large structural packages resulted from DB2 transposition and translation along discrete zones of high ductile strain that occur at low angles to the SB2 fabric. The abrupt transitions between large

lithologic units and in metamorphic grade that characterise the Bass River Complex are primarily a result of this generation of tectonism.

Metamorphism was diachronous within the DB2 kinematic framework. Domain A transposition and translation coincided with or slightly post-dated the regional metamorphic culmination, M2. In Domain S a post-M2-culmination strain peak is also evident. Retrogression and penetrative transposition of the DB1-M1 high grade metamorphic fabric of structural Domain A is attributed to this event.

The DB3 folding event deformed the DB2 structural package and generated both regional- and outcrop-scale folds of the pre-existing planar fabrics. Near-parallelism of DB3 axial surfaces and the regional foliation trend of DB2 is notable and the present orientation girdle defined by DB2 linear elements is believed to be a composite of DB2 and superimposed DB3 deformation. The approximate coincidence of these linear element trends in Domains A and S may be a result of orientation control exerted on early formed DB3 elements by DB2 linear structures.

Finite element studies of linear heterogeneities within a deforming matrix support this conclusion.Watkinson and Cobbold (1981) determined that within certain bounds of strain rate and orientation within a given stress field, superimposed fold axes will nucleate on pre-existing linear elements. In natural systems Hanmer (1979) discussed the role of fabric heterogeneities in localising superimposed kink folds in phyllites of the Armorican Shear Zone, and reached a conclusion similar in principle to that noted by Watkinson and Cobbold (1981).

A marked reorganisation of linear elements characterises DB3 strain in the Folly River Schist. This is true primarily on the outcrop scale because regional fold trends appear to be more consistent. Plots of linear element data of the metavolcanic unit are interpreted as indicating that either local heterogeneity in the DB2 trends characterised the volcanic sequence or DB3 deformation was such that reorientation of approximately coaxial DB2 and early DB3 linear elements occurred. The distinction in linear element orientations between the metavolcanic sequence and the rest of the metamorphic terrane indicates that significant differences existed in the forces acting across the Bass River Complex during DB3 deformation. The more competent Great Village River Gneiss and Gamble Brook Schist rocks responded to this deformation with axial trends mimicing the pre-existing linear fabric elements on all scales. These show generally east-southeasterly or west-southwesterly plunges between 12 and 55 degrees. The Folly River Schist rocks are characterised by more diverse linear element orientations. These do, however, fall on the same north-easterly striking, moderately-southwest-dipping girdle as those in other areas of the Complex.

Inferred pre- or syn-kinematic emplacement of the Portapique River Strongly Foliated Granitoid body (circa 800 Ma, Gaudette et al.,1983) places a minimum Hadrynian age on the penetrative DB2 deformation. Another geochronological restriction on this event is supplied by the truncation of later folds by the circa 600 Ma (Cormier, 1980) McCallum Settlement Pluton. At present the DB2 and DB3 events are considered distinct, but both related to the same late Precambrian tectonic event.

DB4a movement along the Gamble Lake Fault could have occurred at approximately the same time as that along the Rockland Brook Fault (DB4b) since these breaks merge west of Gamble Lake, affect post-DB3 diorites, and are marked by discrete ductile deformation zones. The Rockland Brook Fault is believed to have initially displaced the Gamble Lake structure prior to or contemporaneous with emplacement of the Pleasant Hills Pluton (circa 315 ma., Cormier,1980). At least some post-emplacement movement is also indicated but rock type distributions indicate relatively minor offset, if any, occurred during this period. Right lateral apparent displacement is inferred from Strongly Foliated Granitoid outcrop trends on opposite sides of Gamble Lake Fault.

The Hart Lake - Byers Lake Pluton (circa 340 ma.; Cormier, 1978) is cut by the Rockland Brook Fault but little offset ,if any, marks the eastern contact of the Pleasant Hills Pluton at Portapique River. It is therefore suggested that the Rockland Brook and Gamble Lake Faults moved more or less contemporaneously, with the majority of early displacement occurring along the Gamble Lake structure (DB4a) and later movement (D4Bb) occurring along the Rockland Brook Fault. All movement occurred prior to activation of the Londonderry Fault (DB5). Right lateral displacement is indicated from regional structural trends that bend toward the shear zone. Minor shears that may be high-angle Reidel associates of the main shear zone occur in some areas and their orientations relative to the main shear zone can be interpreted as indicating left-lateral movement. A right-lateral movement sense wold have to be

documented on these structures to validify their Reidel association and, therefore, the left-lateral displacement along the Rockland Brook Fault. Insufficient movement-sense data for the minor shears is available at present to fully substantiate the left-lateral hypothesis. The structural trend changes toward the Rockland Brook Fault that are indicative of dextral displacement are considered the most reliable indicators of displacement sense available at present. Future study of micro-structures in highly strained rocks of the zone could provide additional information in this regard.

The D4Bc fracture-joint trend overprints the Rockland Brook Fault and terminates abruptly at the Londonderry Fault. Its development is therefore constrained to Early Carboniferous - Late Devonian time . The regional significance of these breaks in the deformation history of the Highlands is unclear at present. If a determination of paleostress directions from the inferred conjugate orientation relationship is reasonable, NNE-SSW to NNW-SSE directed maximum compressive stresses could have formed these structures.

DB5 Londonderry Fault movement juxtaposed Lower Carboniferous Londonderry Formation rocks against the metamorphic rocks of Bass River Complex. Brittle overprinting of the DB4a Gamble Lake Fault fabric is seen in foliated diorite east of Great Village River. Donohoe and Wallace (in preparation) attribute cleavage development in the Londonderry Formation to this movement.

A general movement trend is apparent from the DB4a to DB5

deformation sequence. A southward stepping-out of activity along east-west faults marked the late Devonian to Jurassic movements between the Cobequid Highlands and the present Minas Basin. Progressively younger rocks are disrupted by the Gamble Lake, Rockland Brook, Londonderry, Cobequid, Portapique, Gerrish Mountain and Economy Mountain faults. The last three structures noted are subparallel faults that occur south of the study area and are not indicated on the accompanying plates.

The last major structural event evident in Bass River Complex rocks (DB6) is defined by high angle, north-south faults which, on the basis of surface displacement trends, stratigraphic juxtaposition in the adjoining sedimentary basin, and interpretation of aeromagnetic anomalies appear to show normal displacement. These offset the east-west fault system of the highland boundary and effectively date as pre-Early Jurassic in age.

5:2 Regional Geologic Framework

The preceeding review of the structural and metamorphic history of the Bass River Complex provides a base from which correlations with other Avalonian terranes of the Atlantic area may be considered. Rast et al. (1976), Rast (1980) and Williams (1979) described the typical lithostructural sequence of the zone noting that although no basement to the supracrustal section is exposed in the Avalon Peninsula of Newfoundland, rocks that are inferred to unconformably underlie these occur in southwest NewBrunswick and in Cape Breton. More recently, Keppie (1980)

described Avalon Zone rocks in Nova Scotia, determining them to be basement to the supracrustals of the Avalon Peninsula. Models combining plate convergence and subsequent transpression have been proposed by Currie (1983) and O'Brien et al.(1983) and a time constraint for arrival of the Avalon Zone within the orogen is presented by Elias and Strong (1982).

The question that must be addressed in light of any regional model is "where do the rocks of the Bass River Complex fit? ". Subaqueous mafic volcanic rocks of the Folly River Schist could be consistent with a convergent boundary - volcanic arc interpretation. Gamble Brook Schist metasedimentary rocks cannot be as easily accommodated because of their very clean quartz sand protoliths. However, they may have accumulated before initiation of volcanism in the paleo-sea. The few southward younging directions recorded from rocks of the Folly River Schist may support this hypothesis. To date, good chronological data defining a lower limit for the Bass River Complex volcanism are not available and therefore regional correlations must be qualified accordingly.

Deformation in part of the Fourchu Group of southeastern Cape Breton Island is described by Helmstadt and Tella (1972), MacDonald (1979), and Donohoe (1980) and distinct variations are apparent between these and their possible Bass River Complex correlatives. The low-grade metamorphic foliation in the Fourchu Group is associated with first phase, upright regional folds which are only weakly overprinted by later deformation. This differs in style and metamorphic grade from the

strong DB2 of the Bass River Complex which records the earliest deformation in the volcanic and sedimentary cover rocks. A Fourchu Group - Folly River Schist correlation seems unlikely.

Reports on recent work in the Cape Breton Highlands (Jamieson, in press ; MacDonald and Smith, 1980; Jamieson et al., 1983; and Doucet, 1983) outline general trends in Avalonian tectonism and intrusion. Jamieson (in press) describes metamorphism and deformation in the Middle River area as pre-dating 750 Ma diorite emplacement. Radiometric ages determined for Strongly Foliated Granitoid rocks of the Bass River Complex (circa 810 Ma.; Olszewski et al., 1983) may place a minimum age on the major DB2 regional deformation but, as noted in chapter 2, the writer has not seen the actual isochron. The question of its reliability therefore could not be addressed at the time of this writing. If the age is valid, it indicates that deformation took place in rocks of the Bass River Complex in the same broad period of time as that indicated by Jamieson (in press) and Doucet (1983) for tectonism in part of the Cape Breton Highlands.

Considering the age of deformation and similarity of intrusion sequence, it is suggested that the inferred late Precambrian diorites of the Cobequids, such as the McElman Brook Pluton, are associated with the same 700 Ma period of intrusion documented in Cape Breton by Jamieson (in press) and indicated by Krogh et al. (1983) as being characteristic of the Avalonian terrane of southeast Newfoundland.

On the basis of lithology and recent whole-rock radiometric age

data, Donohoe and Cullen (1983) have suggested correlation of deformation fabrics seen in Bass River Complex Domain A rocks with those of the Mount Thom Complex further to the east. They note age similarity of the circa 900 Ma. granitic gneiss to the similar age Brookville Gneiss near Saint John and to the Kelly's Mountain Gneiss in Cape Breton. Olszewski et al.(1983) propose a correlation of the oldest gneisses in southwest New Brunswick with those of the Cobequids and consider the Caledonian Highland rocks as basement to the Greenhead Group. This differs significantly from the interpretation of Wardle (1978) who considered the Brookville Gneiss to be either a high-grade metamorphic equivalent of Greenhead Group rocks or part of an older terrane that had been synchronously metamorphosed.

If correlation of Mount Thom Complex rocks with Structural Domain A rocks of the Bass River Complex is valid, the high grade basement upon which the Gamble Brook Schist and Folly River Schist supracrustal sequences were deposited must signicantly predate the circa 810 Ma minimum age for the DB2 tectono-metamorphic event. Domain A correlation with the Mount Thom Complex requires these basement rocks to be at least 900 Ma old. The correlation offered by Gaudette et al. (1983) for the circa 900 Ma. gneisses of the Cobequid and Caledonian highlands strengthens the case for age equivalence of the overlying Greenhead Group of New Brunswick with the supracrustals of the Bass River Complex. The Gamble Brook Schist is lithologically similar to Unit 5 of the Lily Lake Formation (Greenhead Group) as described by Wardle (1978) and these may be depositional equivalents. To date, however, mafic volcanics that

could be equivalents of the Folly River Schist have not been defined within the Greenhead Group. Further study of mafic schists outlined in western Cape Breton Island may show these to be more likely correlatives of the Folly River Schist.

Regional correlations are essential, but far too often characterised by inconsistencies in terms of local geology. It is not the intention of this study to add further to the confusion. Table 5:1 summarises the correlations suggested in the preceeding discussion and combines them with those of other workers . Certainly these correlations must be considered open to further speculation.

TABLE 5:I

REGIONAL CORRELATION

	SW NEW BRUNSWICK	CENTRAL COBEQUIDS	EASTERN COBEQUIDS	CAPE BRETON
COVER	GREENHEAD GROUP	BASS RIVER COMPLEX DOMAIN S >810 ma	NONE EXPOSED	GEORGE RIVER GROUP
	Lily Lake Formation	Folly River Schist / Gamble Brook Schist		mafic schists ?
BASEMENT	BROOKVILLE GNEISS >950 ma	BASS RIVER COMPLEX DOMAIN A (GREAT VILLAGE RIVER GNEISS)	MOUNT THOM COMPLEX >930 ma	KELLY S MOUNTAIN GNEISS

CHAPTER 6

CONCLUSIONS AND RECOMMENDATIONS FOR FURTHER WORK

6:1 Conclusions

This study outlines several new and important points regarding the stratigraphy, deformation history, and regional correlation of Bass River Complex rocks. These can be summarised as follows:

(1) Previously undocumented sillimanite grade mineral assemblages were identified in the Great Village River Gneiss and used to define a new unit, the Portapique River Schist;

(2) Previously undocumented primary sedimentary and volcanic textures were identified, respectively, in the Gamble Brook Schist and Folly River Schist;

(3) The presence of a previously unknown, major, east-west fault, termed the Gamble Lake Fault, is postulated and a right-lateral movement sense inferred from displacement of lithologic boundaries;

(4) A framework for the structural and metamorphic evolution of the Bass River Complex is presented that indicates strongly the existence of a late Proterozoic basement-cover relationship between the Great Village River Gneiss (basement) and the Gamble Brook Schist and Folly River Schist (cover);

(5) Shear-zone-type tectonic features are documented as having developed in both Late Precambrian time and in Late Devonian to Early Carboniferous time. The early shearing is related to infolding and transposition of the basement-cover complex and is not thought to be evidence of early movement along the Glooscap Fault System . The earliest displacements that are considered to be related to the Glooscap Fault System are those that occurred along the Gamble Lake and Rockland Brook faults in Late Devonian to Early Carboniferous time.

(6) Correlation of the Great Village River Gneiss with circa 900 Ma rocks of the Eastern Cobequid Highlands is supported and similarities to late Precambrian intrusive events in Cape Breton, southwest New Brunswick, and southeast Newfoundland are noted. Correlation of the Gamble Brook Schist and Folly River Schist with the Greenhead Group of New Brunswick is supported.

6:2 Recommendations for Further Work

During the course of the investigation several avenues of further study became apparent but could not be pursued within the thesis time frame. The most significant, and possibly most difficult, is further study of the high grade metamorphic rocks of the Portapique River Schist. Because their present map distribution is based only on a few thin sections, further detailed work is warranted.

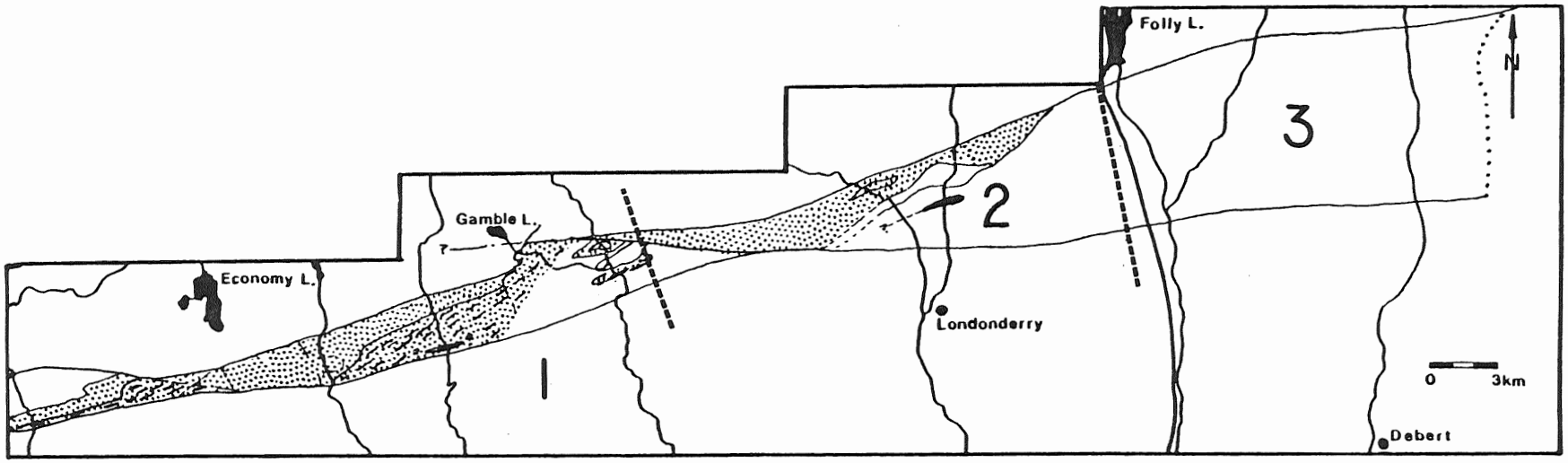
A more complete understanding of intrusion geochronology is

required to constrain more closely some overprinting relationships. This is particularly important for the inferred late Precambrian diorite bodies (unit "4" of Plate 1) and would augment the initial work of Gaudette et al. (1983) on the deformed granitoid rocks.

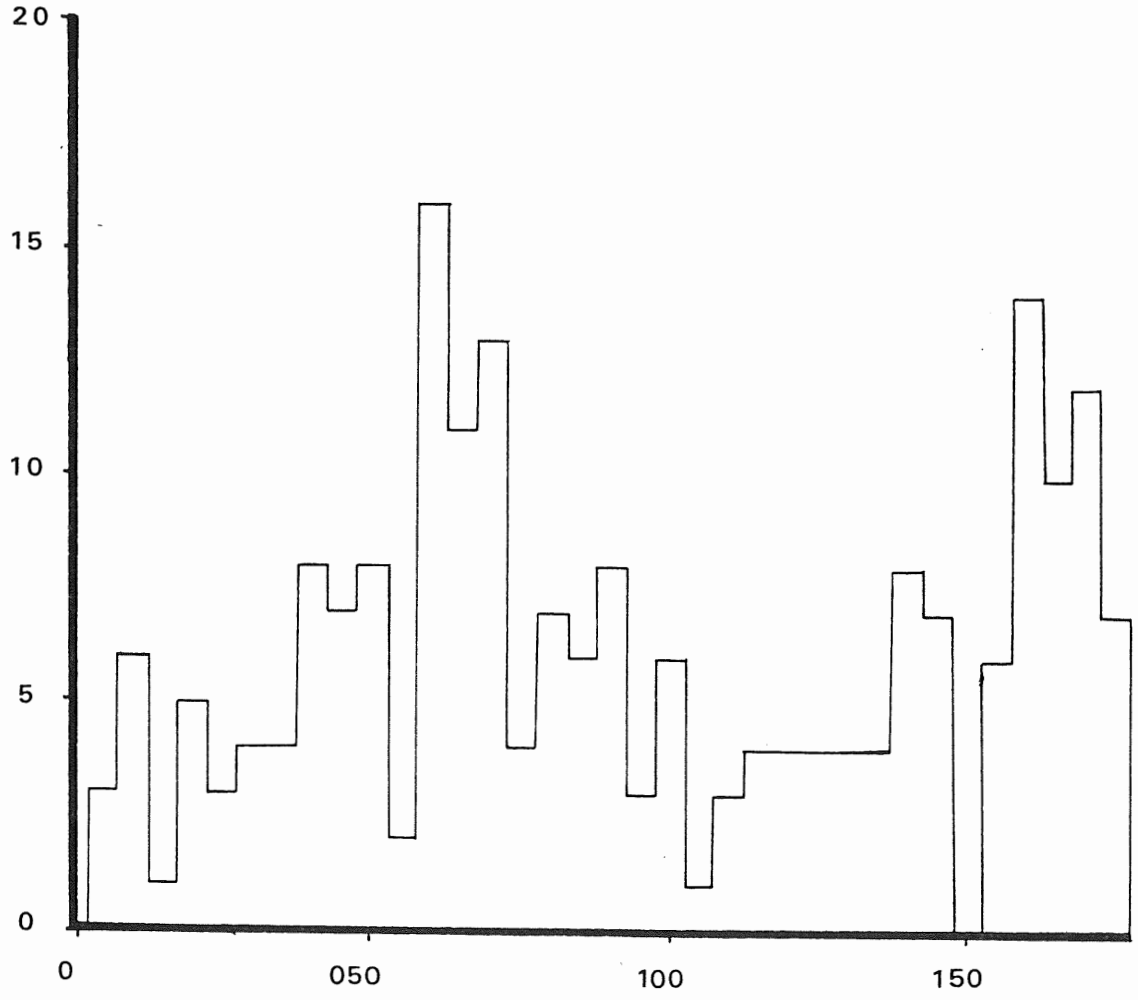
Kinematic analysis of the Bass River Complex would be a beneficial but difficult effort. Domain A rocks , the Strongly Foliated Granitoids and the Gamble Brook Schist of Domain S lend themselves to this type of study as a result of their textural and compositional characteristics. Those of the Folly River Schist would be much more difficult to deal with for exactly the same reason. The results of such a study could shed light on the nature of basement-cover stacking that occurred during DB2. Did this take place within a regional thrusting regime or as a result of oblique translation along a major Precambrian shear zone?

Although a large-scale kinematic analysis would be difficult to accomplish, a more restricted study of rocks that have been affected by Rockland Brook Fault shearing would be valuable. The additional information gained in regard to the movement history of this break would clarify the present relative inconsistency of displacement sense indicators that is noted in Chapter 3, Section 3:6. Was early sinistral movement accommodated along this break?

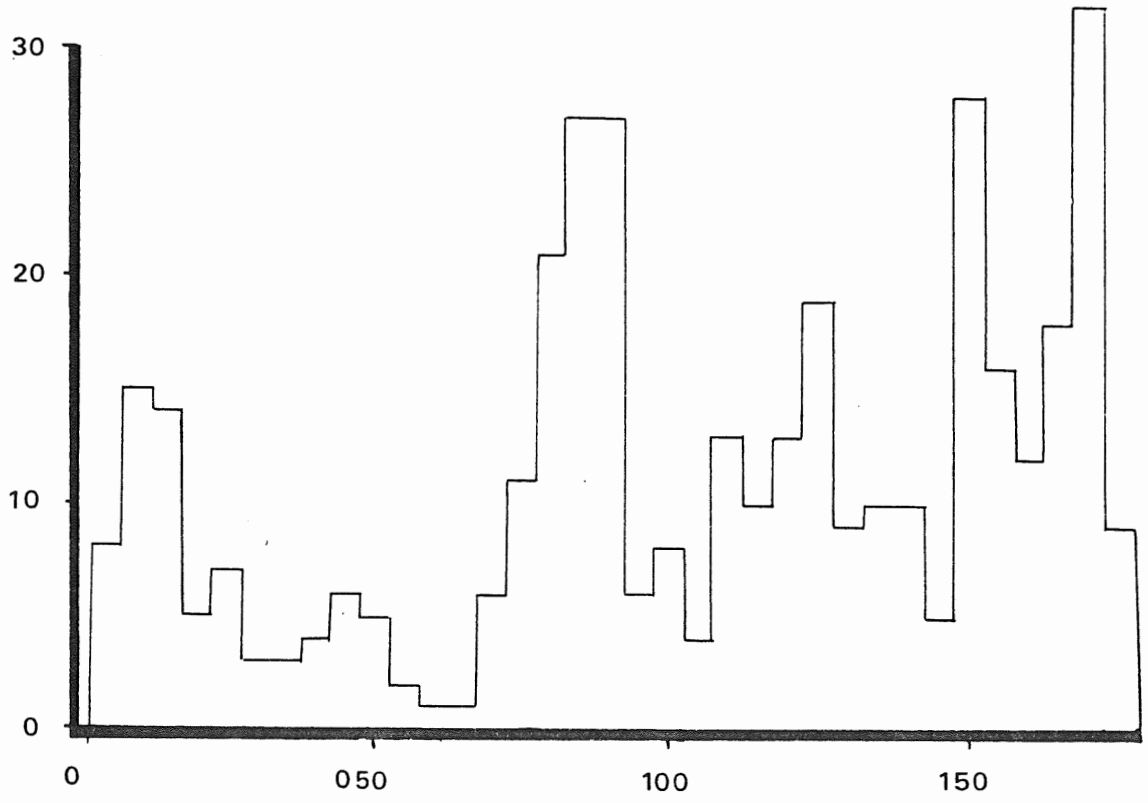
APPENDIX



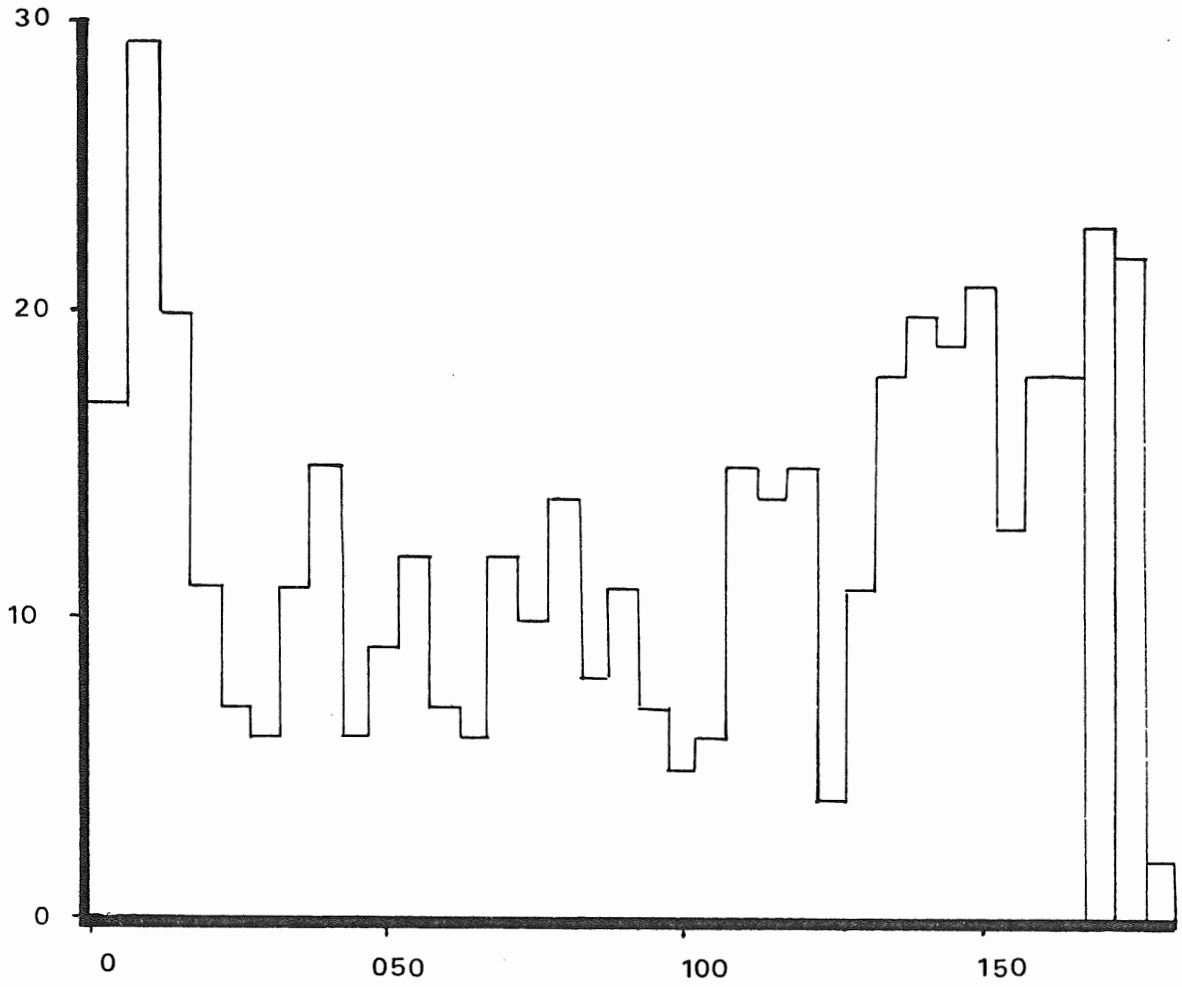
DB4c Orientation Zones



ZONE 1 TRENDS



ZONE 2 TRENDS



ZONE 3 TRENDS

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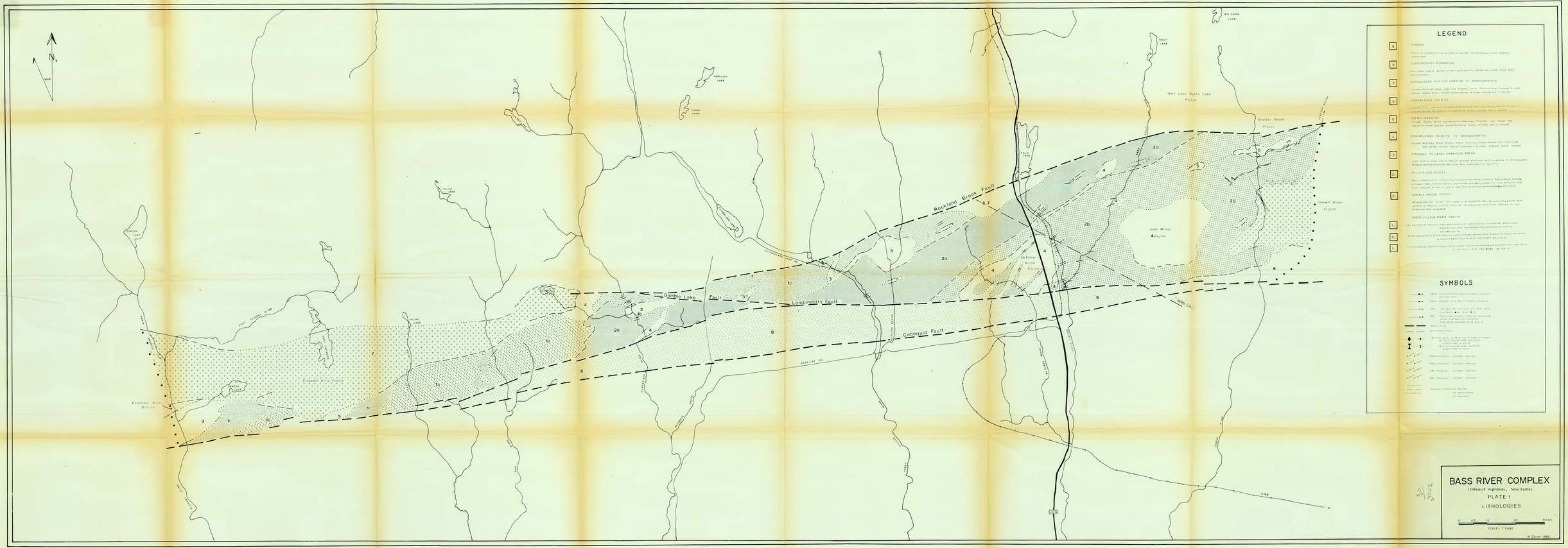

N
Rockland Brook Fault
Gamble Lake Fault
Londonderry Fault
Cobequid Fault
Economy River Diorite
Pleasant Hills Pluton
Hart Lake Byers Lake Pluton
Shatter Brook Pluton
Debert River Pluton
East Mines Pluton
McElmon Brook Pluton
LEGEND
DIABASE
Dykes to subophitic, fine to medium grained hornblende-plagioclase diabase; unfoliated
LONDONDERRY FORMATION
Grey green quartz wacke, siltstone, argillite, minor red purple black shale and volcanics
HORNBLENDE BIOTITE GRANITE TO MONZOGRANITE
Includes Hamilton Byers Lake and Pleasant Hills Plutons; varies texturally from coarse porphyritic to equigranular; weakly foliated and or sheared
HORNBLENDE DIORITE
Includes Folly Lake and Economy River Plutons; dark grey greasy massive to fine gneissic diorite, monzonite to hornblendite, mostly foliated and or sheared
SYENO-GRANITE
Includes Shatter Brook and McElmon Settlement Plutons; light orange red, medium to coarse grained, hypidiomorphic locally foliated and or sheared
HORNBLENDE DIORITE TO GRANODIORITE
Includes McElmon Brook Pluton mostly diorite, locally sheared and chloritized East Mines Pluton mostly granodiorite locally sheared, poorly exposed
STRONGLY FOLIATED GRANITOID ROCKS
Light pink to grey, fine to medium grained granitoids with moderate to strong quartz-feldspar-biotite-hornblende fabric; variably cataclastic to mylonitic
FOLLY RIVER SCHIST
Mafic metaclastic rocks (fine-grained to schistose hornblende-plagioclase-biotite-quartz schist, hornblendite, and foliated diabase); porphyritic equivalents common; minor amounts of felsic schist and chlorite-epidote-hornblende-garnet schist
GAMBLE BROOK SCHIST
Metasedimentary rocks; light brown to white, quartz-feldspar-biotite-garnet schist, generally massive, locally calcareous schist; garnet and mica rich amounts of iron formation and ironstone
GREAT VILLAGE RIVER GNEISS
1a Amphibolite, hornblende-plagioclase-biotite-quartz gneiss commonly calcareous
1b Amphibolite, hornblende-plagioclase-biotite-quartz gneiss commonly calcareous and migmatitic
1c Porcupine Subunit: plagioclase-quartz-clinopyroxene-hornblende gneiss commonly calcareous or migmatitic with interlayered marble and CS
SYMBOLS
Rockland Brook Fault, mineral lineation and fold axes
Gamble Lake Fault mineral lineation
Londonderry Fault mineral lineation and fold axes
Fold axes, mineral lineation, bedding-cleavage intersection lineation
Fold axes indicated by M, F, or S
Mesh Fault
Secondary Fault
Fold axial surface trace (approximate location) (solid) vertical (dash) plunge (arrow) vertical axial surface (solid) vertical axial surface (dash) overturned to north
Foliation: inclined, vertical
Foliation: inclined, vertical
Foliation: inclined, vertical
Foliation: inclined, vertical
Geologic contact (at surface) (?) approximate (?) assumed
BASS RIVER COMPLEX
(Cobequid Highlands, Nova Scotia)
PLATE 1
LITHOLOGIES
0 5.0 10 20 30 km
SCALE: 1:31680
M. Cullen 1983

LEGEND

9 DIABASE
Diabase to subophitic, fine to medium-grained ferromelanocratic dolerite; unaltered.

8 LONGONDERRY FORMATION
Grey to dark quartz wacke, siltstone, feldspathic wacke and purple black shale and argillite.

7 HORNBLENDE BIOTITE GRANITE TO MONZOGRANITE
Include: Mcginnis Brook Pluton, heavy, distinctly foliated and intermixed with Folly Lake Pluton; Hart Lake Byers Lake and Pleasant Hills Plutons; medium to coarse grained, locally porphyritic, commonly massive and in sheeted.

6 HORNBLENDE BIOTITE GRANODIORITE
Includes Folly Lake and Economy River Plutons; dark grey green, medium to fine grained, locally megaporphyritic to subophitic; locally foliated and in sheeted.

5 SYENO-GRANITE
Includes: Spatter Brook and McGinnis Deformed Plutons; light orange red, coarse to fine grained, locally megaporphyritic; locally foliated and in sheeted.

4 HORNBLENDE DIORITE TO GRANODIORITE
Includes McGinnis Brook Pluton; heavy, distinctly foliated and intermixed with Hart Lake Pluton; locally porphyritic; locally sheeted; poorly exposed.

3 STRONGLY FOLIATED GRANITOID ROCKS
Light pink to grey, fine to medium grained granitoids with moderate to strong quartz-feldspar-biotite-hornblende fabric; variably cataclastic to mylonitic.

2 FOLLY RIVER SCHIST
Mainly metapsammitic rocks (muscovite-biotite schist, chlorite-hornblende, biotite, garnet schist, metaconglomerate, and foliated diabase); cataclastic, iron formation and minor amounts of basic schist and chlorite-actinolite-epidote-garnet schist.

2' GAMBLE BROOK SCHIST
Metasedimentary rocks; light brown to white, biotite-chlorite-quartz-feldspar-garnet with lenticular masses; biotite-chlorite, calcite-garnet, pink mineral; amounts of iron formation are subordinate.

1 GREAT VILLAGE RIVER GNEISS
1a. Amphibolite; dubiotite massive, porphyroblastic, and myloniticinterbedded plagioclase porphyroblastic with interbedded felsic and minor ultramafic metavolcanic rocks.
1b Portapique River Schist; dubiotite-quartz-feldspar-garnet schist-hornblendic schist and amphibolite with interbedded felsic and IFR.
1c Orthogneiss; dubiotite, plagioclase-quartz-biotite-muscovite gneiss, commonly cataclastic or mylonitic, with interbedded felsic and IFR.

SYMBOLS

- DBMn Rockland Brook Fault; mineral lineation and fold axes
- DBMn Gamble Lake Fault; mineral lineation
- DB3 Intersection lineation on fold axes
- DB2 Fold axes; B-1 ▲, B-2 ●, B-3 ■
- Fold axes; mineral lineation, trend/rake
- Fold axes; mineral lineation, rake
- Fold axes reformed by B-2 or 3
- Major Fault
- Secondary Fault
- FB3 fold axial surface trace (approximate) (vertical) (anticlinal crest; synform, synclinal trough; synform, overturned to north; antiform, anticlinal trough; antiform, overturned to north)
- DBMn Foliation; inclined, vertical
- DBMn Foliation; inclined, vertical
- DB3 Foliation; inclined, vertical
- DB2 Foliation; inclined, vertical
- Geologic contact (a) defined (b) approximate (c) assumed

BASS RIVER COMPLEX
(Cobequid Highlands, Nova Scotia)

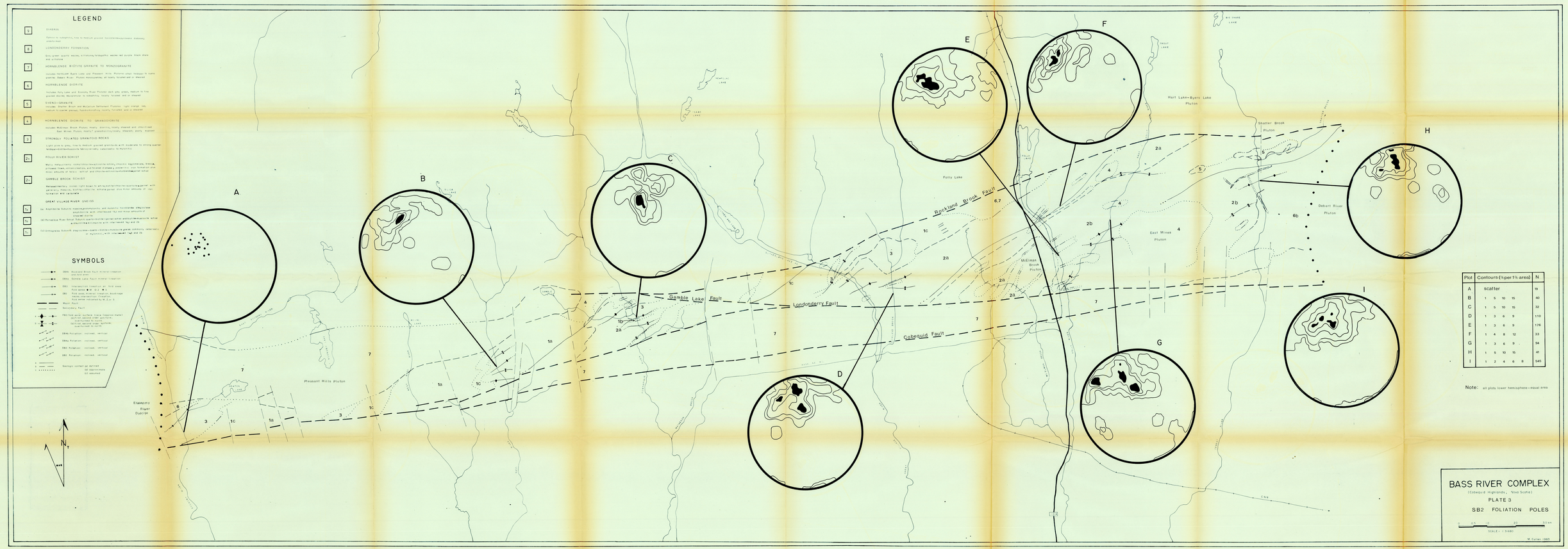
PLATE 2

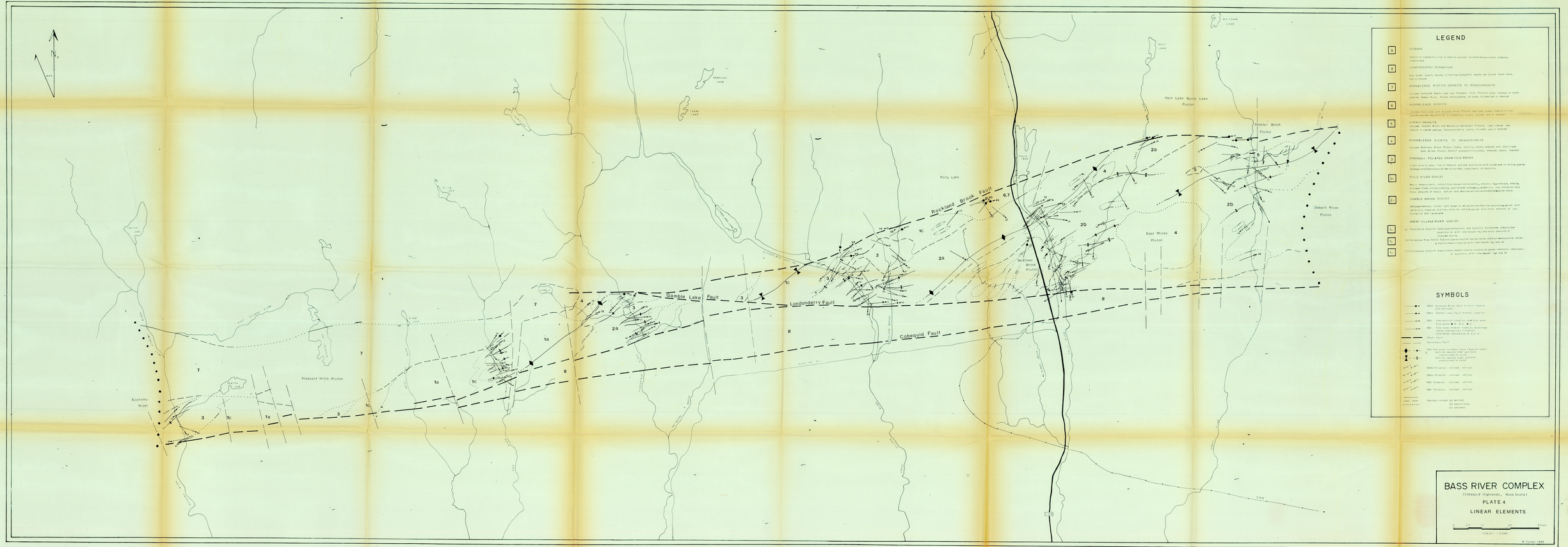
STRUCTURE

0 0.5 0 1.0 2.0 Km

SCALE = 1:31680

M. Currie 1983


LEGEND
SYMBOLS
A
B
C
D
E
F
G
H
I
Rockland Brook Fault
Gamble Lake Fault
Londonderry Fault
Cobequid Fault
Pleasant Hills Pluton
Folly Lake
Hart Lake–Byers Lake Pluton
Shatter Brook Pluton
Debert River Pluton
East Mines Pluton
McElman Brook Pluton
Economy River District
Big Snake Lake
Trout Lake
Plot | Contours (1% per 1% area) | N
A | scatter | 19
B | 1 5 10 15 | 40
C | 1 5 10 15 | 32
D | 1 3 6 9 | 110
E | 1 3 6 9 | 178
F | 1 4 8 12 | 33
G | 1 3 6 9 | 94
H | 1 5 10 15 | 41
I | 1 2 4 6 8 | 545
Note: all plots lower hemisphere–equal area
BASS RIVER COMPLEX
(Cobequid Highlands, Nova Scotia)
PLATE 3
SB2 FOLIATION POLES
SCALE: 1:50,000
M. Culan, 1985


Rocklend Brook Fault
Gamble Lake Fault
Londonderry Fault
Cobequid Fault
Pleasant Hills Pluton
Folly Lake
McElman Brook Pluton
East Mines 4 Pluton
Hart Lake Byers Lake Pluton
Potter Brook Pluton
Debert River Pluton
Economy River
LEGEND
SYMBOLS
BASS RIVER COMPLEX
(Cobequid Highlands, Nova Scotia)
PLATE 4
LINEAR ELEMENTS
SCALE 1:50,000
M. Cullen 1983

Buckland Brook Fault
Gamble Lake Fault
Londonderry Fault
Cobequid Fault
ECONOMY LAKE
SUTHERLAND LAKE
FOLLY LAKE
SYMBOLS
Fault
Gamma Contour
Area Boundary
BASS RIVER COMPLEX
(Cobequid Highlands, Nova Scotia)
PLATE 5
AEROMAGNETIC MAP
SCALE 1:31680
M. Culter 1983

LEGEND

9 DIABASE

Ophitic to subophitic, fine to medium grained hornblende-pyroxene diabase; undeformed

8 LONDONDERRY FORMATION

Grey green quartz wacke, siltstone, feldspathic wacke red purple black shale and siltstone

7 HORNBLENDE BIOTITE GRANITE TO MONZOGRANITE

Includes Hart-Lake Byers Lake and Pleasant Hills Plutons: alkali feldspar to syeno granite: Debert River Pluton: monzogranite; all locally foliated and or sheared

6 HORNBLENDE DIORITE

Includes Folly Lake and Economy River Plutons: dark grey green, medium to fine grained diorite; equigranular to subophitic; locally foliated and or sheared

5 SYENO–GRANITE

Includes Shatter Brook and McCallum Settlement Plutons: light orange red, medium to coarse grained, hypidiomorphic; locally foliated and or sheared

4 HORNBLENDE DIORITE TO GRANODIORITE

Includes McElman Brook Pluton: mostly dioritic, locally sheared and chloritised
East Mines Pluton: mostly? granodioritic; locally sheared; poorly exposed

3 STRONGLY FOLIATED GRANITOID ROCKS

Light pink to grey, fine to medium grained granitoids with moderate to strong quartz-feldspar-biotite-muscovite fabric; variably cataclastic to mylonitic

2b FOLLY RIVER SCHIST

Mafic metavolcanic rocks: chlorite-actinolite schist, chloritic agglomerate, breccia, pillowed flows, volcaniclastics, and foliated diabase; jaspertitic iron formation plus minor amounts of felsic schist and chlorite-actinolite-horblende±garnet schist

2a GAMBLE BROOK SCHIST

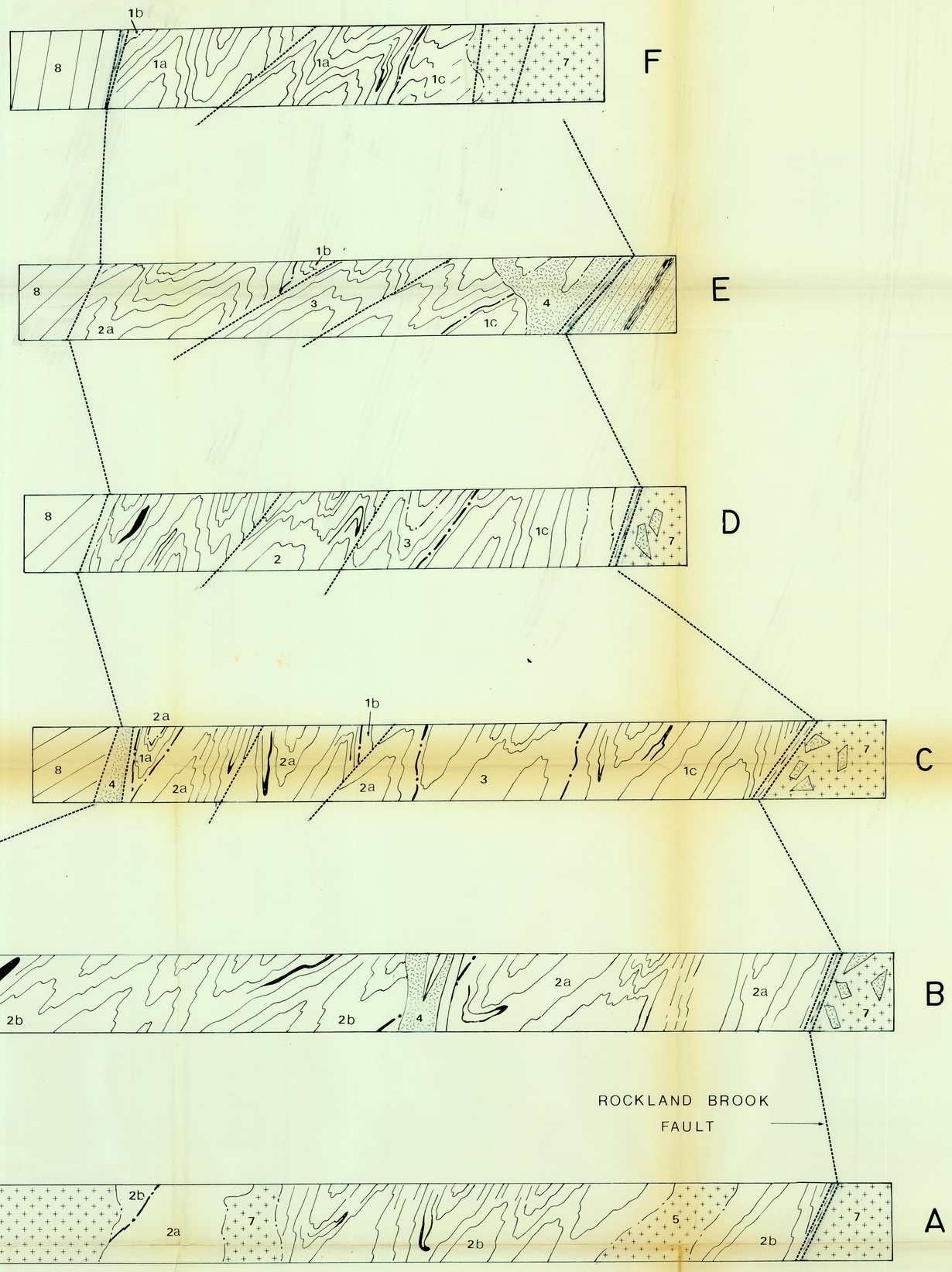
Metasedimentary rocks: light brown to white, biotite-chlorite-quartzite±garnet with generally massive, biotite-chlorite schist±garnet plus minor amounts of iron formation and carbonate

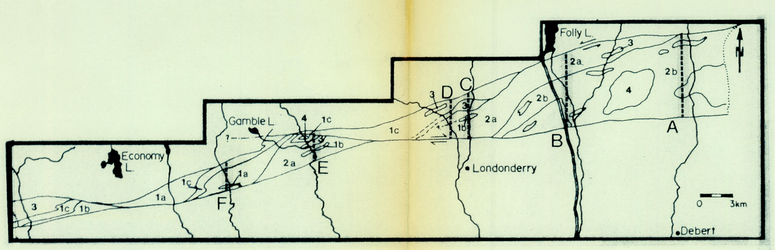
GREAT VILLAGE RIVER GNEISS

1a (a) Amphibolite Subunit: massive, protomylonitic and mylonitic hornblende plagioclase amphibolite with interleaved 1b,c and minor amounts of sheared diorite

1b (b) Portapique River Schist Subunit: quartz-biotite-garnet schist and biotite-muscovite schist ± staurolite± sillimanite with interleaved 1a,c and 2b

1c (c) Orthogneiss Subunit: plagioclase-quartz-biotite-muscovite gneiss commonly cataclastic or mylonitic, with interleaved 1a,b and 2b





Section Locations

BASS RIVER COMPLEX

(Cobequid Highlands, Nova Scotia)

PLATE 6

CROSS SECTIONS

0 1km

M Cullen 1983