GEOLOGICAL SURVEY OF CANADA
BULLETIN 556

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2001

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Catalogue No. M42-556E
ISBN 0-660-18554-7

Available in Canada from
Geological Survey of Canada offices:

601 Booth Street
Ottawa, Ontario K1A 0E8

3303-33rd Street N.W.
Calgary, Alberta T2L 2A7

101-605 Robson Street
Vancouver, B.C. V6B 5J3

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Price subject to change without notice

Cover illustration

Layered granite of the Late Devonian Hart Lake–Byers Lake pluton containing a xenolith of hornblende gabbro, which attributed to the Folly Lake Pluton. Photograph by B. Murphy. GSC 2000-016

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Original manuscript submitted: 1999-09
Final version approved for publication: 2000-03

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GEOLOGY OF THE EASTERN COBEQUID HIGHLANDS, NOVA SCOTIA

Abstract

The eastern Cobequid Highlands, part of the Avalon terrane in mainland Nova Scotia, include Late Proterozoic to Late Paleozoic rocks. The Rockland Brook Fault divides the highlands into two blocks. To the south, Late Proterozoic rocks include granitoid gneiss and amphibolite of the Mount Thom Complex, and the Bass River Complex which includes the 734 ± 2 Ma Economy River Gneiss, platformal metasedimentary rocks of uncertain age, and the ca. 610–580 Ma Great Village River Gneiss. Coeval supracrustal rocks consist of continental tholeiitic basaltic flows and dykes and interbedded turbidite sequences. North of the Rockland Brook Fault, similarly aged arc-related volcanic and sedimentary rocks occur. Taken together, Late Neoproterozoic rocks in the eastern Cobequid Highlands may expose an oblique section across a Neoproterozoic magmatic arc. Penetrative fabrics are interpreted to reflect ductile shear-zone deformation associated with sinistral intra-arc transtension and basin development, followed by dextral transpression resulting in deformation of the basin.

Late Ordovician to Early Devonian rocks comprise shallow marine sedimentary and volcanic rocks and probably correlate with the Arisaig Group to the northeast. Devonian–Carboniferous rocks comprise continental sedimentary rocks and intra-continental bimodal volcanic rocks (Fountain Lake Group) and coeval bimodal plutonic rocks (emplacement age ca. 360 Ma). Locally intense dextral transpression between the Visean and Namurian resulted in a positive flower structure, local isoclinal folds, and may be related to dextral movement on the Avalon–Meguma terrane boundary.

Résumé

Les roches des hautes terres de Cobequid orientales, qui font partie du terrane d'Avalon en Nouvelle-Écosse, datent du Protérozoïque tardif au Paléozoïque tardif. La faille de Rockland Brook divise les hautes terres en deux blocs. Au sud, les roches du Protérozoïque tardif incluent des gneiss granitoïdes et des amphibolites du Complexe de Mount Thom ainsi que le Complexe de Bass River qui comprend le Gneiss d'Economy River (734 ± 2 Ma), des roches métasédimentaires de plate-forme d'âge incertain et le Gneiss de Great Village River (610–580 Ma environ). Les roches supracrustales contemporaines comprennent des dykes et des coulées de basalte tholéiitique d'origine continentale et des séquences turbiditiques interstratifiées. Au nord de la faille de Rockland Brook reposent des roches sédimentaires et volcaniques d'arc d'âge semblable. L'ensemble des roches du Néoprotérozoïque tardif des hautes terres de Cobequid orientales peut présenter une coupe oblique à travers un arc magmatique néoprotérozoïque. Les fabriques pénétratives sont interprétées comme étant le résultat d'une déformation de zone de cisaillement ductile associée à une transtension senestre intra-arc et à la formation d'un bassin, qui aurait été suivie par une transpression dextre ayant causé la déformation du bassin.

Les roches de l'Ordovicien tardif au Dévonien précoce incluent des roches volcaniques et des roches sédimentaires déposées dans un milieu marin peu profond et que l'on pourrait corréler au Groupe d'Arisaig au nord-est. Les roches dévoniennes–carbonifères comprennent des roches sédimentaires continentales et des roches volcaniques bimodales intracontinentales (Groupe de Fountain Lake) et des roches plutoniques bimodales contemporaines (mises en place vers 360 Ma). Une transpression dextre, localement intense, survenue entre le Viséen et le Namurien a créé une structure de faille bivergente positive et des plis isoclinaux locaux; elle pourrait être associée à un mouvement dextre sur la limite des terranes d'Avalon et de Meguma.

SUMMARY

This report describes and interprets the geology of the Neoproterozoic to Early Carboniferous rocks of the eastern Cobequid Highlands, Nova Scotia. The highlands are bound to the south by the Cobequid Fault, and to the north by unconformably overlying middle to Late Carboniferous rocks. The Cobequid Highlands comprise part of the Avalon terrane which extends discontinuously along the southeastern flank of the Appalachian Orogen, outcropping in isolated fault-bounded blocks from the Avalon Peninsula type area in southeastern Newfoundland to southern New England. This terrane is characterized by Neoproterozoic arc-related rocks overlain by Early Paleozoic successions with fauna of Gondwanan affinity.

The eastern Cobequid Highlands is divided by the Rockland Brook Fault into two fault blocks with contrasting pre-Namurian geology. To the south, Neoproterozoic rocks include poorly exposed granitoid gneiss and amphibolite of the Mount Thom Complex, whose age and tectonic significance remain uncertain, the Bass River Complex (Gamble Brook Formation, Economy River Gneiss, Folly River Formation, Great Village River Gneiss), and granitoid plutons (Frog Lake, Debert River, and McCallum Settlement plutons). The Economy River Gneiss is dated at 734 ± 2 Ma which provides a minimum age for the platformal Gamble Brook Formation which it intrudes. It is exposed only near Economy River Falls. This gneiss has an arc-related chemistry and is interpreted as a local representative of an early phase of Avalonian magmatism prior to the main arc phase. The gneiss is intruded by amphibolite dykes of uncertain age. The Great Village River Gneiss consists of strongly deformed granitoid rocks and amphibolite. It was previously interpreted to represent Neoproterozoic basement, however U-Pb zircon ages of 610–580 Ma indicate that it is a coeval, deformed, plutonic correlative of both less deformed plutons and low-grade arc-related Folly River Formation.

The contact between the Gamble Brook Formation and the Great Village River Gneiss is exposed on several stream sections and is a ductile shear zone. Syntectonic granitic intrusions during deformation are dated at ca. 605 Ma and constrain the age of ductile shear. The contact between the Folly River and Gamble Brook formations is interpreted as an unconformity, because fabrics in the Gamble Brook Formation are cut by a dyke that defines the contact between them. The Folly River Formation consists of continental tholeiitic basalt and coeval, locally sheeted mafic dykes. The Dalhousie Mountain Formation occurs in the easternmost highlands and is lithologically isolated. As a result, its depositional age is uncertain. It is intruded by ca. 610 Ma granitoid pluton, suggesting a Neoproterozoic age. Its geochemistry is indistinguishable from the Folly River Formation, suggesting it may be correlative.

SOMMAIRE

Le présent rapport contient une description et une interprétation de la géologie des roches du Néoprotérozoïque au Carbonifère précoce des hautes terres de Cobequid orientales en Nouvelle-Écosse. Les hautes terres sont limitées au sud par la faille de Cobequid, et au nord par des roches du Carbonifère moyen à tardif qui les recouvrent en discordance. Les hautes terres de Cobequid comportent une partie du terrane d'Avalon qui longe de façon discontinue le flanc sud-est de l'orogène appalachien, affleurant dans des blocs isolés limités par des failles depuis la région type de la presqu'île d'Avalon dans le sud-est de Terre-Neuve jusque dans le sud de la Nouvelle-Angleterre. Ce terrane est caractérisé par des roches d'arc du Néoprotérozoïque sur lesquelles reposent des successions du Paléozoïque précoce marquées par une faune d'affinité gondwanienne.

La faille de Rockland Brook devise les hautes terres de Cobequid orientales en deux blocs qui contrastent par leur géologie pré-namurienne. Dans le sud, les roches néoprotérozoïques comportent des gneiss granitoïdes et des amphibolites mal exposés du Complexe de Mount Thom dont l'âge et l'importance tectonique n'ont pas été précisés, le Complexe de Bass River (Formation de Gamble Brook, Gneiss d'Economy River, Formation de Folly River, Gneiss de Great Village River), et des plutons granitoïdes (plutons de Frog Lake, de Debert River et de McCallum Settlement). Le Gneiss d'Economy River a été daté à 734 ± 2 Ma, ce qui permet d'attribuer un âge minimal aux roches de plate-forme de la Formation de Gamble Brook qu'il recoupe. Il n'affleure que près des chutes de la rivière Economy. D'après sa composition chimique, ce gneiss est lié à un arc et serait la représentation locale d'une phase précoce de magmatisme avalonien, antérieure à la principale phase d'arc. Des dykes d'amphibolite d'âge incertain le recoupent. Le Gneiss de Great Village River se compose d'amphibolites et de roches granitoïdes fortement déformées. Selon une interprétation antérieure, il représenterait le socle néoprotérozoïque; cependant, des datations U-Pb sur zircon donnent un âge de 610–580 Ma et il s'agirait donc d'une roche plutonique contemporaine déformée en corrélation avec des plutons moins déformés et la Formation de Folly River faiblement métamorphisée et liée à un arc.

Le contact entre la Formation de Gamble Brook et le Gneiss de Great Village River est exposé dans plusieurs coupes fluviatiles et correspond à une zone de cisaillement ductile. Des intrusions granitiques syntectoniques contemporaines de la déformation remontent à 605 Ma environ, ce qui permet de dater le cisaillement ductile. Le contact entre la Formation de Folly River et la Formation de Gamble Brook est interprété comme une discordance étant donné que des fabriques dans la Formation de Gamble Brook sont recoupées par un dyke qui définit le contact. La Formation de Folly River se compose de basalte tholéiitique continental et de dykes mafiques contemporains localement feuilletés. La Formation de Dalhousie Mountain se rencontre dans l'extrême est des hautes terres et se distingue par sa lithologie. C'est pourquoi elle n'a pas pu être datée avec certitude. Comme un pluton granitoïde de 610 Ma environ la recoupe, on peut supposer un âge néoprotérozoïque. Sa composition géochimique est tout à fait semblable à celle de la Formation de Folly River, d'où une corrélation possible entre les deux formations.

North of the Rockland Brook Fault, the Neoproterozoic rocks are assigned to the Jeffers Group and Warwick Mountain Formation. Uranium-lead age dating of the Jeffers Group in the western Cobequid Highlands yields an age of 628 ± 3 Ma. The depositional age of the Warwick Mountain Formation is less certain. On the basis of similar lithology, geochemistry, and tectonic setting, the Jeffers Group and Warwick Mountain Formation are thought to be broadly correlated with the Dalhousie Mountain and Folly River formations to the south of the Rockland Brook Fault and equivalent rocks in the Antigonish Highlands.

The eastern Cobequid Highlands appear to expose an oblique section across a Neoproterozoic magmatic arc that formed and deformed between 630 and 580 Ma. Kinematic analysis indicates that the shear-zone contact between the Gamble Brook Formation and the Great Village River Gneiss has normal and sinistral components. This data is interpreted to reflect an episode of intra-arc transtension which heralds the formation of a pull-apart basin into which the overlying Folly River Formation was deposited. Kinematic analysis of structures in the Folly River Formation indicate shear-related deformation with reverse and dextral components and is interpreted to represent dextral intra-arc transpression and inversion of the basin. Penetrative fabrics in Neoproterozoic gneiss, syntectonic granite, and volcano-sedimentary successions are penecontemporaneous, not sequential, and represent fabrics formed at different structural levels during protracted deformation associated with arc-related tectonothermal activity.

Cambrian to Early Ordovician rocks are not exposed in the eastern Cobequid Highlands. North of the Rockland Brook Fault, Late Ordovician to Early Devonian rocks comprise shallow marine fossiliferous siliciclastic sedimentary and minor volcanic rocks (Wilson Brook, Murphy Brook formations) containing Rhenish–Bohemian faunal assemblages, similar to the Arisaig Group of the Antigonish Highlands. South of the fault, rocks of similar age are poorly exposed in fault slices, and their relationship to coeval strata north of the fault is not understood.

Devonian–Carboniferous rocks north of the Rockland Brook Fault comprise continental sedimentary and intracontinental bimodal volcanic rocks (Fountain Lake Group). Voluminous bimodal plutonic rocks (emplacement age ca. 360 Ma) on both sides of the fault may be cogenetic with the volcanic sequence. Deformation may have accompanied their emplacement. Ductile shear deformation on and adjacent to the Rockland Brook Fault also occurred between the Visean and Namurian. Associated fabrics are heterogeneous and locally intense and exhibit dextral kinematic indicators. Deformation produced a positive flower structure, local isoclinal folds, bimodal plutonism; reactivation of the Rockland Brook Fault may be related to movement on the Cobequid–Chedabucto (Minas) fault zone during dextral motion along the Avalon–Meguma terrane boundary.

Au nord de la faille de Rockland Brook, les roches néoprotérozoïques sont attribuées au Groupe de Jeffers et à la Formation de Warwick Mountain. La datation U-Pb du Groupe de Jeffers dans les hautes terres de Cobequid occidentales donne un âge de 628 ± 3 Ma. L'âge du dépôt de la Formation de Warwick Mountain est moins certain. En se basant sur le fait que leur lithologie, leur composition géochimique et leur cadre tectonique sont sembables, il existerait une corrélation générale entre le Groupe de Jeffers et la Formation de Warwick Mountain et les formations de Dalhousie Mountain et de Folly River au sud de la faille de Rockland Brook et des roches équivalentes dans les hautes terres d'Antigonish.

Les hautes terres de Cobequid orientales semblent présenter une coupe oblique à travers un arc magmatique néoprotérozoïque dont la formation et la déformation se situent entre 630 et 580 Ma. Une analyse cinématique indique que le contact (zone de cisaillement) entre la Formation de Gamble Brook et le Gneiss de Great Village River a des composantes normale et senestre que l'on peut interpréter comme témoignant d'un épisode de transtension intra-arc ayant abouti à la création d'un bassin d'extension dans lequel se sont déposées les roches de la Formation de Folly River sus-jacente. L'analyse cinématique des structures dans la Formation de Folly River indique qu'il s'est produit une déformation liée à un cisaillement et ayant des composantes inverse et dextre qui représenterait une transpression dextre intra-arc et une inversion du bassin. Les fabriques pénétratives observées dans le gneiss néoprotérozoïque, le granite syntectonique et les successions volcano-sédimentaires sont pénécontemporaines et non séquentielles et représentent des fabriques formées à différents niveaux structuraux lors d'une déformation prolongée associée à une activité tectonothermique d'arc.

Les roches du Cambrien à l'Ordovicien précoce n'affleurent pas dans les hautes terres de Cobequid orientales. Au nord de la faille de Rockland Brook, les roches de l'Ordovicien tardif au Dévonien précoce comprennent des roches sédimentaires silicoclastiques fossilifères déposées dans un milieu marin peu profond et des roches volcaniques accessoires (formations de Wilson Brook et de Murphy Brook) renfermant des associations faunistiques rhénanes–bohémiennes, semblables à celles du Groupe d'Arisaig dans les hautes terres d'Antigonish. Au sud de la faille, les roches d'âge équivalent sont mal exposées dans des écailles et les liens qu'elles auraient avec les strates contemporaines au nord de la faille n'ont pas été élucidés.

Les roches dévoniennes–carbonifères situées au nord de la faille de Rockland Brook sont des roches sédimentaires continentales et des roches volcaniques bimodales intracontinentales (Groupe de Fountain Lake). Les vastes roches plutoniques bimodales (mise en place vers 360 Ma) de part et d'autre de la faille et la séquence volcanique pourraient être cogénétiques. Les roches plutoniques ont pu être déformées pendant leur mise en place. La déformation par cisaillement ductile le long de la faille de Rockland Brook et près de celle-ci a également eu lieu entre le Viséen et le Namurien. Les fabriques associées sont hétérogènes et localement intenses et présentent des indications de mouvement dextre. La déformation a produit une structure de faille bivergente positive, des plis isoclinaux locaux et un plutonisme bimodal. La réactivation de la faille de Rockland Brook pourrait être associée à un mouvement survenu sur la zone de faille de Cobequid–Chedabucto (Minas), pendant qu'a eu lieu le mouvement dextre le long de la limite entre les terranes d'Avalon et de Meguma.

INTRODUCTION

Regional geological setting

The eastern Cobequid Highlands (Fig. 1) form a complexly faulted block of Late Proterozoic and minor Paleozoic rocks, bounded to the south by the Cobequid–Chedabucto (Minas) fault system, which separates the highlands from the Meguma terrane, and to the north by unconformably overlying middle to late Carboniferous rocks (Donohoe and Wallace, 1982a, b, c, d).

The Cobequid Highlands lie within the Avalon composite terrane of the Appalachian Orogen (Williams and Hatcher, 1983; Keppie, 1985; Fig. 2), which extends discontinuously from Newfoundland to southern New England. This terrane is characterized by Late Precambrian (Neoproterozoic III, ca. 630–585 Ma) arc-related volcanic-sedimentary sequences and cogenetic plutons, ca. 565–550 Ma bimodal volcanic and continental clastic rocks deposited in rift and/or wrench basins, and Early Paleozoic platformal rocks with fauna that suggest that Avalon was separated from cratonic North America until the Silurian (Boucot, 1975; Williams, 1979; Rast and Skehan, 1983; Keppie, 1985). In Nova Scotia, these Early Paleozoic platformal rocks include a Cambrian platformal overstep sequence that contains Acado–Baltic (Avalonian) faunas (Theokritoff, 1979; Keppie, 1985; Murphy and Nance, 1989; Keppie et al., 1991), and Late Silurian–Early Devonian sedimentary rocks that contain Rhenish–Bohemian faunas (Keppie et al., 1991). Middle Proterozoic or older basement rocks have nowhere been unequivocally identified within the Avalon composite terrane. Most workers agree that the Late Proterozoic rocks of the Avalon composite terrane developed along the periphery of Gondwanaland, but the precise location is uncertain.

Diversity of the late Precambrian sequences has been interpreted to represent variations in tectonic setting within one terrane or several distinct terranes (cf. Keppie, 1985; Krogh et al., 1988). In the type area of the Avalon composite terrane in Newfoundland the rocks have been variously interpreted as an ensialic rift zone (Strong et al., 1978), a series of rift-related intracratonic troughs and small ocean basins (O'Brien et al., 1983), and a basin-and-range type environment (Krogh et al., 1988). The succession in New Brunswick has been interpreted as both rift-related (Giles and Ruitenberg, 1977) and ensialic volcanic arc (Rast et al., 1976; Nance, 1987; Barr and White, 1989; Currie and Eby, 1990). The late Precambrian rocks of Nova Scotia have been interpreted as rifted volcanic-arc sequences underlain by continental crust (Keppie et al., 1979; Keppie, 1982; Murphy and Keppie, 1987; Barr and White, 1989; Pe-Piper and Piper, 1989, Murphy et al., 1990; Dostal et al., 1990). Although a wide variety of tectonic environments have been suggested for the Late Precambrian history of the Avalon composite terrane in Atlantic Canada, a wealth of recent geochemical data supports an ensialic arc-related environment (Barr and White, 1989, 1990; Pe-Piper and Piper, 1989, Murphy et al., 1990; Dostal et al., 1990; Currie and Eby, 1990).

Variations in Precambrian geology contrast with the Cambro-Ordovician stratigraphy which may be correlated across the entire Avalon composite terrane (Landing, 1991), and has been interpreted as an overstep sequence (Keppie and Murphy, 1988). The various late Proterozoic sequences were therefore probably assembled by latest Precambrian time (Neoproterozoic III of Cowie and Bassett (1989)) to form the Avalon composite terrane (Keppie, 1985; Keppie et al., 1991). These sequences were polydeformed and intruded by a suite of dioritic, granodioritic, and granitic plutons yielding latest Precambrian to Cambrian radiometric ages (Barr et al., 1988; Keppie et al., 1990). The mid- to late Paleozoic history of the Avalon composite terrane is thought to be related to rifting from Gondwanaland and accretion to North America, but the precise timing of these events is controversial. In the Late Paleozoic, significant telescoping and dispersion of Avalonian terranes occurred during translations relative to cratonic North America.

Minor mineralization has been found throughout the eastern Cobequid Highlands (Donohoe and Wallace, 1982a, b, c, d; Chatterjee, 1984) comprising 1) sandstone-hosted Cu-U mineralization in Late Carboniferous sedimentary rocks; 2) sandstone-hosted Fe mineralization in Late Carboniferous sedimentary rocks; 3) fault- and joint-related Fe-Mn-Cu mineralization adjacent to the Cobequid Fault, mainly within Carboniferous sedimentary rocks, and 4) epithermal U-Au-Th-F-Cu-Ag-Sn-Mo-Fe mineralization in Devonian–Carboniferous bimodal volcanic and intrusive complexes, probably related to rhyolite domes (Chatterjee, 1984). Mapping of the eastern Cobequid Highlands therefore provides both a guide to mineral exploration in a poorly understood area and a contribution to a much larger scientific problem.

Previous work

Initial reconnaissance mapping (Honeyman, 1874; Ells, 1885; Fletcher, 1905a, b, c, d) considered the rocks now known as the Bass River Complex to be pre-Early Silurian and possibly as old as Archean, although Fletcher subdivided them into Devonian and "igneous rocks of unknown age" and Weeks (1948) grouped these rocks into the "Cobequid Complex", which he interpreted as Silurian. The Carboniferous cover rocks were studied in detail by Bell (1927, 1948) with particular emphasis on the flanks of the Cobequid Highlands. Kelley (1964, 1966, 1968) provided a series of reports and a preliminary map of the Cobequid Highlands at a scale of 1:50 000. Eisbacher (1969) studied the strain history along the Cobequid Fault, to which he attributed Late Paleozoic dextral fault motion.

The entire Cobequid Highlands were mapped by H.V. Donohoe, Jr. and P.I. Wallace at 1:50 000 scale (Donohoe and Wallace, 1982a, b, c, d), and a series of reports and field guides published by these authors (Donohoe and Wallace, 1985, and references therein) defined many of the stratigraphic names used in the present report, either formally or informally (Table 1; *see also* Williams et al., 1985). Donohoe and Wallace (1982a, b, c, d) considered the oldest rocks in the Cobequid Highlands to be the Mount Thom and Bass River complexes, interpreted as basement gneiss

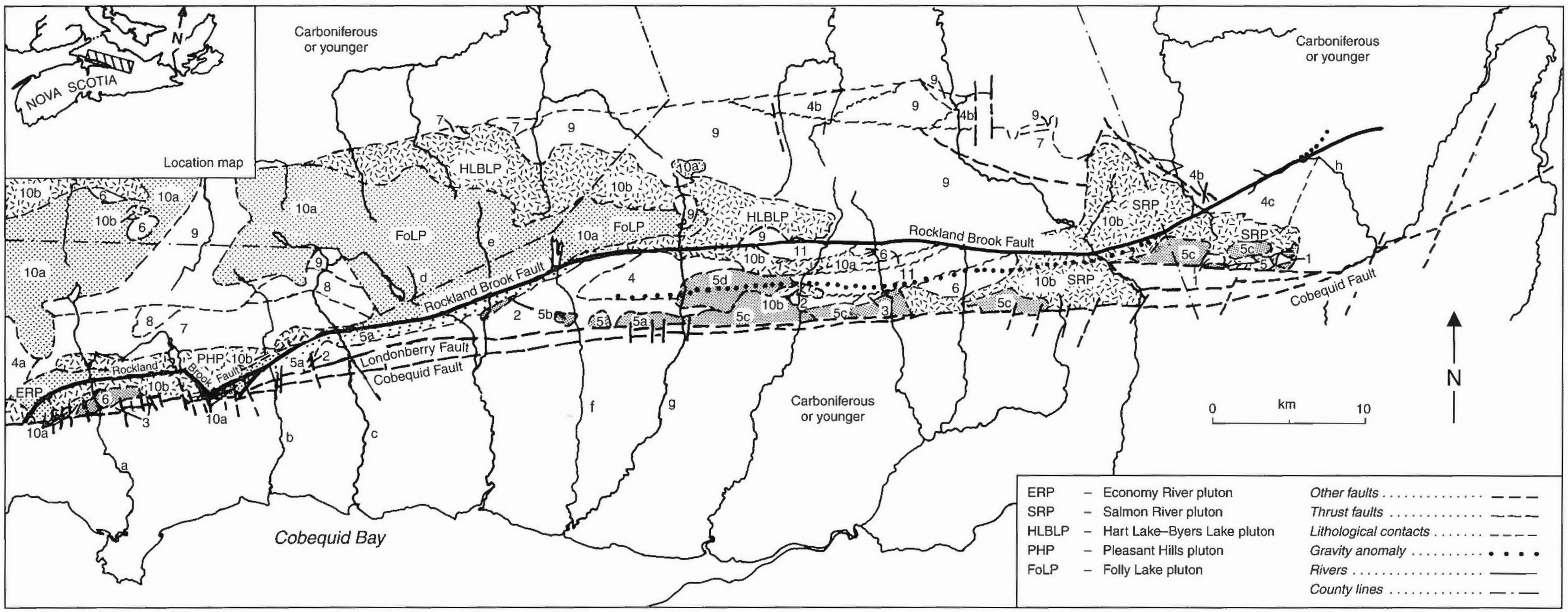


Figure 1. *Summary geological map of the eastern Cobequid Highlands.* See also *Table 1 for key to numbered map units.*

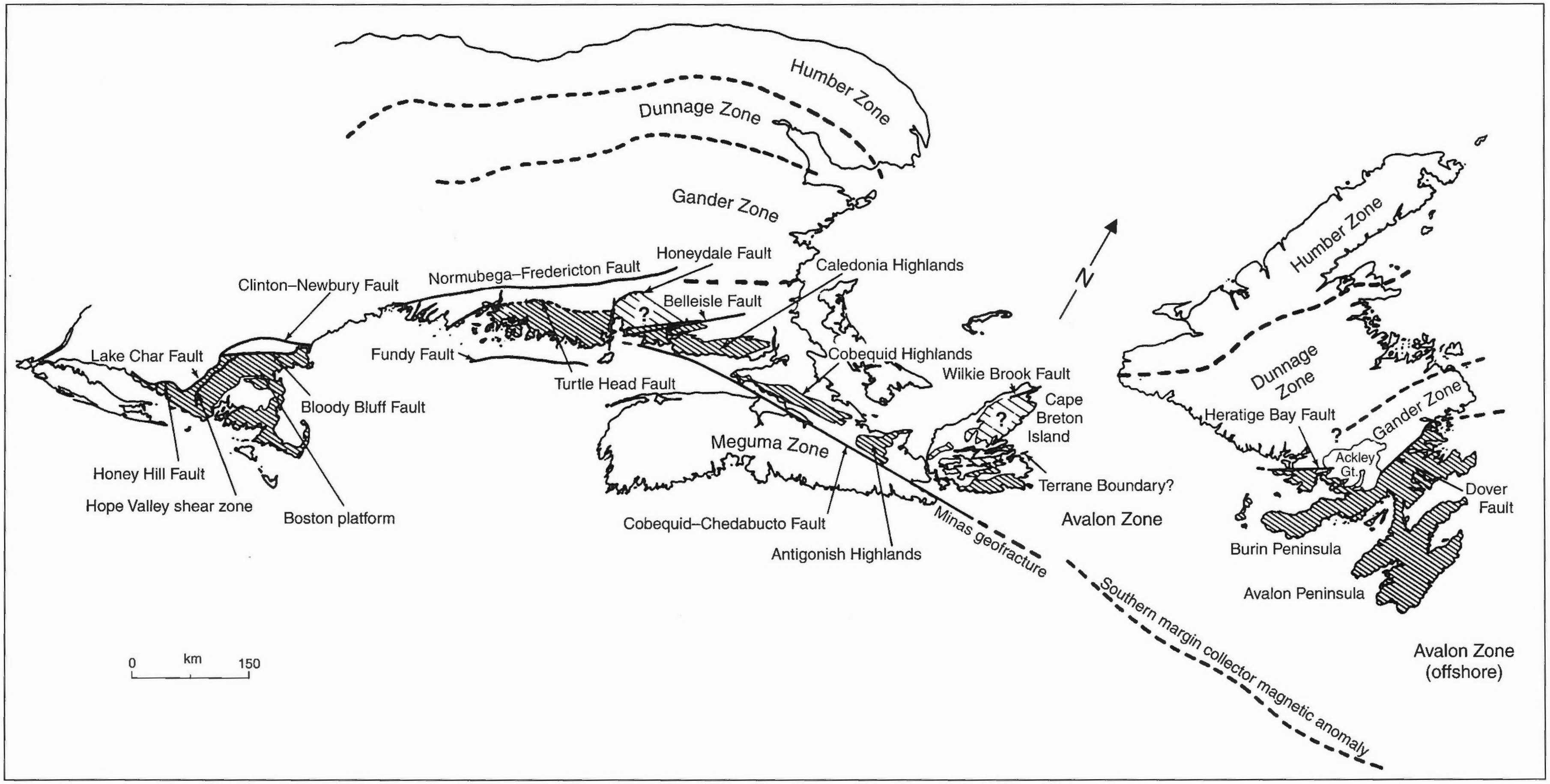


Figure 2*. Location of the Cobequid Highlands within the Avalon composite terrane* (modified from *Zen, 1983, Fig. 1).*

Table 1. Stratigraphy of Precambrian to Lower Carboniferous rocks of the Cobequid Highlands (*modified from* Donohoe and Wallace, 1982a, b, c, d; Murphy et al., 1988).

North		South	
Unit		**Unit**	
Carboniferous **Early Carboniferous**			
		11	Nuttby Formation: quartzite, siltstone, minor conglomerate
Devonian **Late Devonian**			
10	Plutons 10a gabbro diorite 10b granite	10	Plutons 10a gabbro diorite 10b granite
9	Fountain Lake Group 9a basalt 9b rhyolite 9c sedimentary rocks		
Early–Mid-Devonian			
8	Portapique and Murphy Brook formations: conglomerate, sandstone, siltstone		
Ordovician–Silurian			
7	Wilson Brook Formation: sandstone, siltstone, minor volcanic rocks	6	Undivided sandstone, siltstone
Neoproterozoic III			
4	4a Jeffers Group 4b Warwick Mountain Formation 4c Dalhousie Mountain Formation: mafic and felsic volcanic rocks interlayered with greywacke	5	Late Proterozoic plutons 5a Great Village River Gneiss 5b Granite gneiss 5c Appinitic gabbro 5d Granite
		4	Folly River Formation: mafic volcanic rocks, interlayered greywacke
		3	Economy River Gneiss Deformed granite
		2	Gamble Brook Formation 2a lower: quartzite, pelitic schist, carbonate 2b upper: schist, phyllite, minor quartzite
		1	Mount Thom Complex

unconformably overlain by medium-grade metasedimentary and metavolcanic rocks (which they considered to be Helikian or Hadrynian), that had been deformed and metamorphosed prior to deposition of the late Precambrian Jeffers and Warwick Mountain formations. Donohoe and Wallace (1982a, b, c, d) divided the Paleozoic rocks into late Ordovician to late Silurian siltstone, wacke, and minor volcanic rocks of the Wilson Brook Formation; rhyolitic flows and volcanic wacke of the middle to late Silurian Earltown Formation; overlain by middle to upper Devonian redbeds, and thick, bimodal, volcanic rocks (Fountain Lake Group) and redbeds. Cullen (1984) interpreted the Bass River Complex as a polydeformed basement-supracrustal sequence, with the Great Village River Gneiss unconformably overlain by the Gamble Brook Schist and Folly River Schist. Pe-Piper and Piper (1987) upgraded the Jeffers Formation of the western highlands to group status and divided it into formations, with basal bimodal volcanic rocks overlain by turbiditic rocks.

Obtaining an accurate determination of the age of plutonic rocks posed a major problem in early work on the Cobequid Highlands. A summary of previous work, mainly using Rb-Sr and K-Ar isotopic systems, is given in Table 2. These data suggested that ages of plutonism concentrated into two groupings, ca. 660–570 Ma (i.e. Late Proterozoic, Neoproterozoic III) and ca. 360–310 (Devonian to Carboniferous), but the data were not sufficiently precise to determine whether magmatism spanned the entire duration of each grouping or whether magmatism was concentrated in discrete pulses within each grouping.

Table 2. Geochronological compilation showing references and sources of age data (*see* Murphy et al., 2000) from the eastern and western Cobequid Highlands. Locations given in Figure 1.

Location number	Unit and/or rock type	Age (Ma)	Technique	Reference
1	Mount Thom Complex; orthogneiss	934 ± 82 (0.7113)	Rb-Sr (W)	Gaudette et al., 1984
2	Economy River Gneiss; orthogneiss	734 ± 2	U-Pb (Z)	Doig et al., 1989, 1993
3	Great Village River gneiss	626 ± 22 (0.7057)	Rb-Sr (W)	Gaudette et al., 1984
4	Great Village River gneiss	600 ± 5 580 ± 5 589 ± 5	U-Pb (Z)	Doig et al., 1991a, b
5	Rockland Brook granite-gneiss	652 ± 99 (0.7051) 642 ± 15 (0.7053)	Rb-Sr (W)	Gaudette et al., 1984
6	"Portapique River" granite gneiss	*839 (0.7003) *587 (0.7046) *350 (0.7088)	Rb-Sr (W)	Gaudette et al., 1984
7	Great Village River gneiss; deformed granodiorite	605 ± 5	U-Pb (Z)	Doig et al., 1989, 1991a, b
8	Great Village River gneiss	425 ± 17	K-Ar (m)	Wanless et al., 1972; Keppie and Smith, 1978
9	Great Village River gneiss	398 ± 17	K-Ar (b)	Wanless et al., 1972; Keppie and Smith, 1978
10	Great Village River gneiss; sheared syenogranite	370 ± 4	U-Pb (Z)	Doig et al., 1989
11	Economy River gneiss; granite intrusion	363 ± 12	Rb-Sr (W)	Gaudette et al., 1984
12	McCallum Settlement pluton; granite	575 ± 22 (0.7065)	Rb-Sr (W)	Gaudette et al., 1984
13	McCallum Settlement pluton; granite	575 ± 5	U-Pb (Z)	Doig et al., 1991a, b
14	Debert River pluton; granite	612 ± 4	U-Pb (Z)	Doig et al., 1989, 1991a, b.
15	Debert River pluton; granite	596 ± 70 (0.7059)	Rb-Sr (W)	Donohoe et al., 1986; Donohoe and Wallace, 1985
16	Jeffers Brook pluton; diorite	616 ± 28 626 ± 28	K-Ar (h)	Wanless et al., 1973; Keppie and Smith, 1978
17	Jeffers Brook pluton; diorite	607 ± 3	$^{40}Ar/^{39}Ar$ (h)	Keppie et al., 1990
18	Jeffers Brook pluton; diorite	605 ± 4	$^{40}Ar/^{39}Ar$ (h)	Keppie et al., 1990
19	Jeffers Brook pluton; diorite	544 ± 22 564 ± 22 585 ± 23	K-Ar (b)	Wanless et al., 1973; Keppie and Smith, 1978
20	Jeffers Brook pluton; diorite	541 ± 25 (0.7060)	Rb-Sr (b)	Cormier, 1979
21	Frog Lake pluton; diorite	622 ± 3	$^{40}Ar/^{39}Ar$ (h)	Keppie et al., 1990
22	Hart Lake–Byers Lake pluton; granite	348 ± 9	Rb-Sr (W)	Donohoe et al., 1986
23	Hart Lake–Byers Lake pluton; granite	361 ± 2	U-Pb (Z)	Doig et al., 1996
24	Hart Lake–Byers Lake pluton; granite	331 ± 17	Rb-Sr (W)	Cormier, 1979
25	Pleasant Hills pluton; granite	358 ± 2	U-Pb (Z)	Doig et al., 1996
26	Pleasant Hills pluton; granite	315 ± 25 (0.7082)	Rb-Sr (W/k/q-p)	Cormier, 1980
27	Fault zone, Grenville River	287 ± 34	K-Ar (W)	Wanless et al., 1966
Western Cobequid Highlands				
28	Davidson Brook pluton; granite	409 ± 87	Rb-Sr (W)	Pe-Piper et al., 1989
29	Gilbert-Wyvern pluton (east) granite	*400	Rb-Sr (W/b)	Pe-Piper et al., 1989
30	Gilbert-Wyvern pluton (west) granite	*362 *353	Rb-Sr (W/k/b)	Pe-Piper et al., 1989
31	Fountain Lake Group; (old) volcanic rocks	**387 ± 2 (0.7075)	Rb-Sr (W)	Cormier, 1982; Pe-Piper et al., 1989
32	Fountain Lake Group; (young) volcanic rocks	**341 ± 4 (0.7075)	Rb-Sr (W)	Cormier, 1982; Pe-Piper et al., 1989
33	North River pluton; granite	**356 ± 17	Rb-Sr (W/k)	Pe-Piper et al., 1989
34	Cape Chignecto pluton; granite, diorite	349 ± 12 (0.7052)	Rb-Sr (W/k)	Pe-Piper et al., 1989
35	Cape Chignecto pluton; granite	339 ± 22	Rb-Sr (W)	Cormier, 1979
36	Cape Chignecto pluton; deformed granite-diorite	329 ± 9 327 ± 11 303 ± 11	K-Ar (b)	Waldron et al., 1989
37	West Moose River pluton; gabbro-diorite	342 ± 5 (0.7084)	Rb-Sr (W/k)	Pe-Piper et al., 1989
38	Hanna Farm pluton; granite	338 ± 17 (0.7069)	Rb-Sr (W/k)	Pe-Piper et al., 1989

(Sample types: b - biotite; h - hornblende; k - K-feldspar; m - muscovite; q-p - quartz-plagioclase; W - whole rock; Z - zircon)

* - Rb-Sr isochron not formed (minimum ages); ** - Rb-Sr errorchron; Numbers in parentheses are Rb-Sr initial ratios

A lineament analysis using Seasat, TM (Thematic mapper), stream geochemical, and mineral occurrence data (Harris et al., 1987) identified predominantly north- and east-trending faults, subordinate north-northeast- and north-northwest-trending faults, and circular fracture patterns attributed to buried plutons. Our mapping shows that these lineaments are fracture sets with little displacement that correspond to deeply cut river valleys. Rogers et al. (1986) and Bonham-Carter et al. (1987) integrated bedrock geology, topographic, drainage, and geochemical data to conclude that Carboniferous sedimentary units have potential for Au-Hg-Pb-Zn mineralization and that Devonian–Carboniferous volcanic rocks and granitic bodies have exploration potential near their contacts.

Piper et al. (1993) used magnetic data as guide to re-interpret the position of geological boundaries in areas of limited outcrop, and showed that the total magnetic field is high beneath the eastern Cobequid Highlands as far west as the Folly Lake pluton, suggesting the presence of subsurface gabbro-diorite plutons in this region. The high magnetic anomaly extends west to the latitude of Parrsboro in two subparallel belts, corresponding to the belts of plutons along the southern and northern margins of the highlands. A belt with a lower magnetic signature extends westwards from the Folly Lake pluton to Parrsboro, corresponding to Neoproterozoic, Silurian, and Devono-Carboniferous sedimentary rocks. The magnetic field is also lower west of Parrsboro, except for a few small anomalies between the Cobequid and Kirkhill faults, some of which correspond to known diorite outcrops in the Cape Chignecto pluton. Two linear zones of positive anomalies 8 km and 15 km north of the Cobequid Fault and west of Parrsboro correspond approximately to the Spicer Cove and Sand Cove faults of Ryan et al. (1990). On a smaller scale, strong linear trends in the total-magnetic-field and first-derivative maps can be used to extend mapped faults, and to hypothesize others.

Present work

This report was prepared under contracts to J.B. Murphy and commenced in the summer of 1987 as part of the Canada–Nova Scotia Mineral Development Agreement 1984–1989. The report presents a description and interpretation of the field, structural, and laboratory studies of pre-Carboniferous rocks of the eastern Cobequid Highlands based upon mapping at a scale of 1:10 000. Geochemical work in 1987–1989 was subcontracted to G. Pe-Piper. A geological map was compiled from results of this study, from open file maps of Pe-Piper and co-workers, and from Donohoe and Wallace (1982a, b, c, d), and represents knowledge of the geology in 1990, but the text has been selectively updated to include work carried out since 1991. The map, GSC Open File 3703 (Murphy et al., 2000), is available separately and may prove helpful to the reader in understanding the geology of the study area. Geochemical and geochronological work were partially funded by N.S.E.R.C. (Natural Sciences and Engineering Research Council of Canada), the Geological Survey of Canada, and the Canada–Nova Scotia Cooperative Agreement, 1990–1992.

Acknowledgments

We thank Fred Chandler and Joe Whalen for organizational support; Sandra Barr, Fred Chandler, Randy Cormier, Ken Currie, Wes Gibbons, Jana Horak, Duncan Keppie, Brent Miller, Bill Poole, and Chris White for informative discussions on the outcrop; Kim Bellefontaine, Kathy Minehan, Rick Morell, Denise Turner, and Mary Tramble for their excellent field assistance; Marie Gillis for excellent secretarial skills; and Kathryn Baker for technical assistance. Special thanks to Sid Taylor, Rob Raeside, and Sandra Barr for very constructive scientific editing. We are especially indebted to Ken Currie of the Geological Survey of Canada for a thorough final edit which greatly improved the organization and clarity of the manuscript. Funding for this project, provided by the Geological Survey of Canada, under the Canada–Nova Scotia 1984–1989 Mineral Development Agreement, N.S.E.R.C. operating grants to J.B. Murphy and G. Pe-Piper, and Ohio University grants to R.D. Nance are gratefully acknowledged. We acknowledge, with gratitude, the contribution of the work of Howard Donohoe, Peter Wallace, and Mike Cullen to the understanding of the geology of the Cobequid Highlands. They laid a very important foundation which greatly facilitated our investigations.

PRECAMBRIAN ROCKS

Introduction

Precambrian rocks of the Cobequid Highlands vary in lithology, structural style, and grade of metamorphism according to their position relative to the Rockland Brook Fault (Fig. 1; *see also* Murphy et al., 2000). South of the fault, the Mount Thom Complex and the structurally lower part of the Bass River Complex consist of highly deformed medium- to low-grade metamorphic rocks of predominantly igneous protolith. The upper part of the Bass River Complex consists of medium- to low-grade platformal sedimentary rocks (Gamble Brook Formation) unconformably overlain by a low-grade volcanic-sedimentary sequence (Folly River Formation). All of these units are intruded by late Precambrian plutonic rocks ranging in composition from gabbro to granite, some of which have appinitic tendencies.

Adjacent to the eastern boundary of the highlands, the Dalhousie Mountain Formation consists of low-grade interbedded volcanic and sedimentary rocks which lack a regional penetrative fabric. North of the Rockland Brook Fault, Precambrian supracrustal rocks are assigned to the Warwick Mountain Formation and Jeffers Group. The Folly River, Dalhousie Mountain, and Warwick Mountain formations and Jeffers Group are lithologically similar and are probably correlative.

Mount Thom Complex

The Mount Thom Complex (Fig. 1), originally defined by Donohoe (1975) with a type section along the Glen road (NTS E/10, Pictou county), is bounded to the north by younger granitic and gabbroic intrusions, and to the south by an east-trending fault. Until recently, outcrop of the complex consisted of poor exposures of granitoid orthogneiss and amphibolite along logging roads (Donohoe and Wallace, 1985). A new quarry opened up in the area of Mount Thom Complex exposes a wide range of deformed metamorphic rocks with a predominant quartzofeldspathic orthogneiss containing a foliation defined by biotite and muscovite and thought to be related to deformation and greenschist- to amphibolite-facies metamorphism. This main foliation is folded into isoclinal folds with an axial planar foliation defined by garnet- and biotite-rich layers, which is in turn deformed by folds and an associated crenulation cleavage (Donohoe, 1976; Donohoe and Cullen, 1983). Subordinate amphibolite, characterized by elongate hornblende (45–65%), and minor biotite-rich (20–30%), garnetiferous paragneiss with abundant rounded accessory detrital zircon grains also occurs. Contacts between these lithologies are all

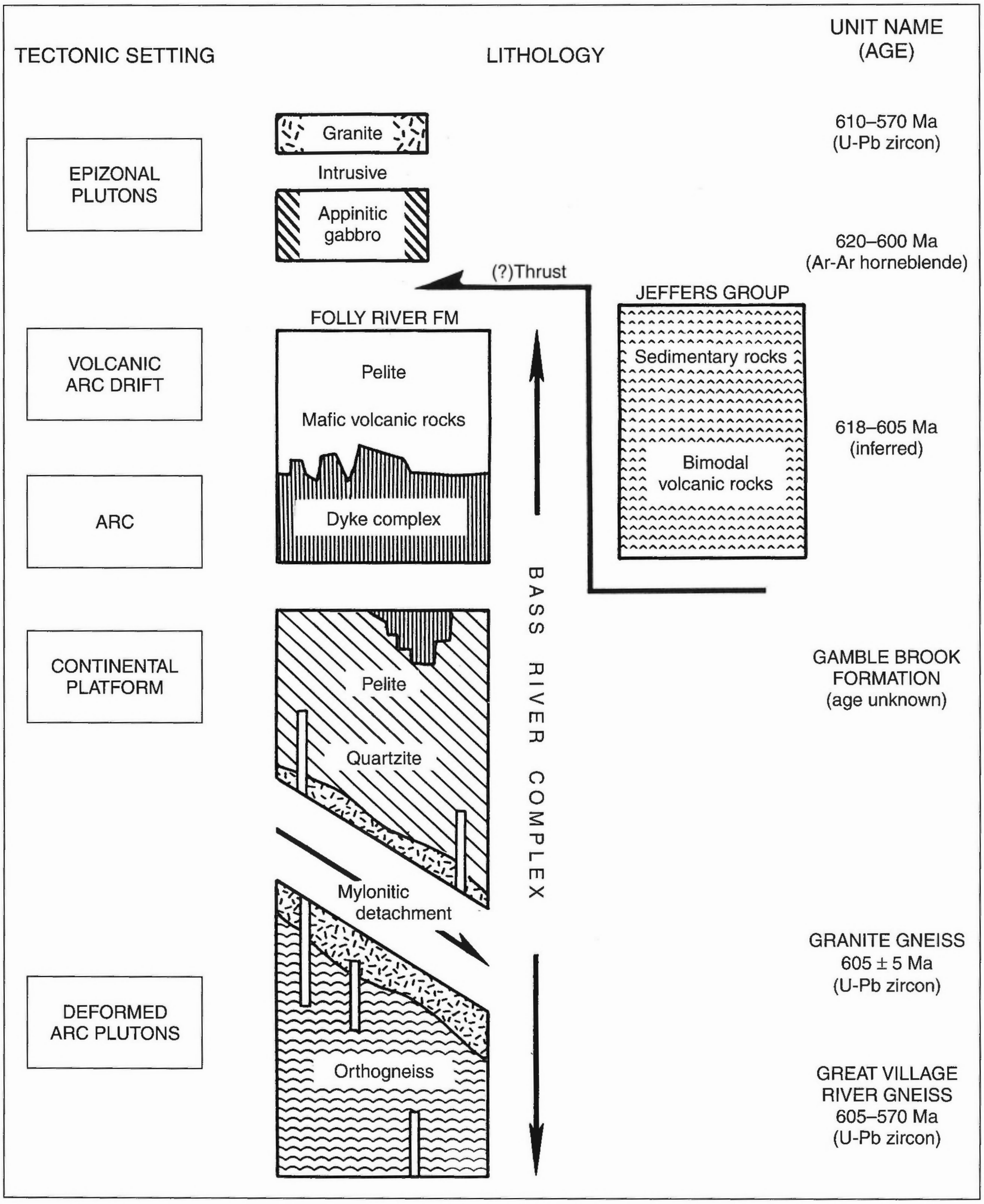


Figure 3. Simplified interpretative tectonostratigraphic column for Late Precambrian to Early Paleozoic rocks of the eastern Cobequid Highlands.

transposed, so that original relationships are unknown. Postdeformation granite, aplite, and mafic dykes of unknown intrusive age crosscut the complex. Other than these minor intrusive relationships, no relationships of the Mount Thom Complex to other units are known.

Orthogneiss samples of the Mount Thom Complex yielded a poor Rb/Sr whole-rock age of 934 ± 82 Ma and a $^{87}Sr/^{86}Sr$ initial ratio of 0.711 (Gaudette et al., 1984), interpreted by Donohoe and Wallace (1985) as the age of main foliation (D_1 fabric). Samples from the main quarry are currently being analyzed for U-Pb (zircon) dating.

Bass River Complex

The Bass River Complex, as defined and described by Donohoe (1975, 1983), consisted of a) the Great Village River Gneiss (orthogneiss, paragneiss, amphibolite, and schist), b) the structurally juxtaposed Gamble Brook Schist (greenschistfacies platformal orthoquartzite, pelitic schist, and phyllite), and c) the Folly River Schist (greenschist-facies mafic volcanic rocks and turbidite sequences) which probably overlies the Gamble Brook rocks unconformably (Murphy et al., 1988). Our work, however, shows that units b and c can be defined in stratigraphic terms as the Gamble Brook and Folly River formations, and that unit a is composite, consisting of an older Economy River Gneiss (ca. 734 Ma), a younger Great Village River orthogneiss (600–570 Ma), and mylonitized granitic gneiss. Despite the presence of stratigraphic units within it, we retain the name Bass River Complex for the whole assemblage, since it has thus far proved impossible to assign all exposures of the complex to chronologically ordered units (Fig. 3).

Gamble Brook Formation

The metasedimentary Gamble Brook Formation structurally overlies the Great Village River Gneiss and separates the Great Village River Gneiss from the Folly River Formation. The contact between the Gamble Brook Formation and the Great Village River Gneiss is interpreted as a ductile shear zone (Murphy et al., 1988) characterized by mylonitization, development of C-S fabrics, local tectonic interleaving of the Great Village River Gneiss and Gamble Brook metasedimentary rocks, synkinematic intrusion of mylonitic granite, and the development of several generations of cogenetic, small-scale isoclinal folds. These features all die out away from this shear zone. The upper contact with the overlying Folly River Formation is interpreted as an unconformity (Murphy et al., 1988). The Gamble Brook Formation is excellently exposed on Rockland Brook, Portapique River, and Great Village River as well as in a quarry east of the old Trans-Canada Highway near Folly Mountain.

The Gamble Brook Formation consists of two distinct units. A lower part contains abundant tan to white orthoquartzite and arkosic quartzite with interlayered biotite-muscovite-garnet psammitic schist and minor carbonate beds or lenses. Mylonitic fabric and a well developed mineral lineation occur adjacent to the contact with the Great Village River Gneiss. In these localities, bedding is transposed into the metamorphic foliation. The intensity of the metamorphic fabric decreases away from the contact and bedding can be recognized in some instances (e.g. the quarry south of Folly Lake). The upper part of the Gamble Brook Formation is distinctly more pelitic and is dominated by biotite-muscovite and biotite-garnet schist and phyllite, in addition to quartzite and psammitic schist.

A minimum depositional age for the Gamble Brook Formation is provided by the ca. 605 Ma synkinematic granitic rocks which intruded the formation adjacent to the contact with the Great Village River Gneiss. If xenoliths within the ca. 734 Ma Economy River Gneiss belong to the Gamble Brook Formation, then the antiquity of the formation relative to typical Avalonian rocks can be demonstrated.

The depositional setting of the Gamble Brook Formation has been interpreted as a shallow marine platformal environment (Donohoe and Wallace, 1985; Nance and Murphy, 1990) and the formation may be broadly correlative with other platformal successions of the Avalon composite terrane such as the Green Head Group of southern New Brunswick (Wardle, 1978); however, owing to the lack of control on depositional ages, regional correlations of lithologically similar sequences should be viewed with caution (compare for example Barr and Raeside, 1989, 1990; Keppie 1990; Barr and White, 1990). Although the platformal lithologies suggest presence of a continental basement, the age and nature of this basement is unknown. Uranium-lead geochronological data and field relationships preclude the possibility that the Economy River Gneiss or Great Village River Gneiss are the basement upon which the Gamble Brook metasedimentary rocks were deposited.

Economy River Gneiss

The Economy River Gneiss, previously considered to be part of the Great Village River Gneiss (Donohoe and Wallace, 1982a, b, c, d), outcrops on Economy River (Fig. 4, 5), bounded on the south by the Cobequid Fault and on the north by a granitic gneiss body (Great Village River Gneiss). Although the contact between the gneiss bodies is a ductile shear zone, the granitic gneiss locally intruded the Economy River Gneiss, which was also intruded by deformed amphibolite dykes and a deformed syenogranite dated at ca. 360 ± 5 Ma (Doig et al., 1989) implying that deformation of the syenogranite, as well as part of the deformation of the Economy River Gneiss, was late Paleozoic.

The Economy River Gneiss consists predominantly of quartz, plagioclase, and hornblende with local retrograde alteration to sericite, epidote, and chlorite. This amphibolite-facies assemblage is similar to that in the crosscutting amphibolite dykes. Both rock types contain a variably developed mylonitic fabric with a strong subhorizontal to moderately inclined mineral lineation defined by elongate quartz grains in the orthogneiss and hornblende in the amphibolite. Pronounced obliquity of the margins of the amphibolite dykes to the mylonitic fabric suggests that deformation and intrusion of the dyke may have been penecontemporaneous. The mylonitic fabric is clearly overprinted by deformation that affected the ca. 360 Ma syenogranite (Fig. 6A), attributed to

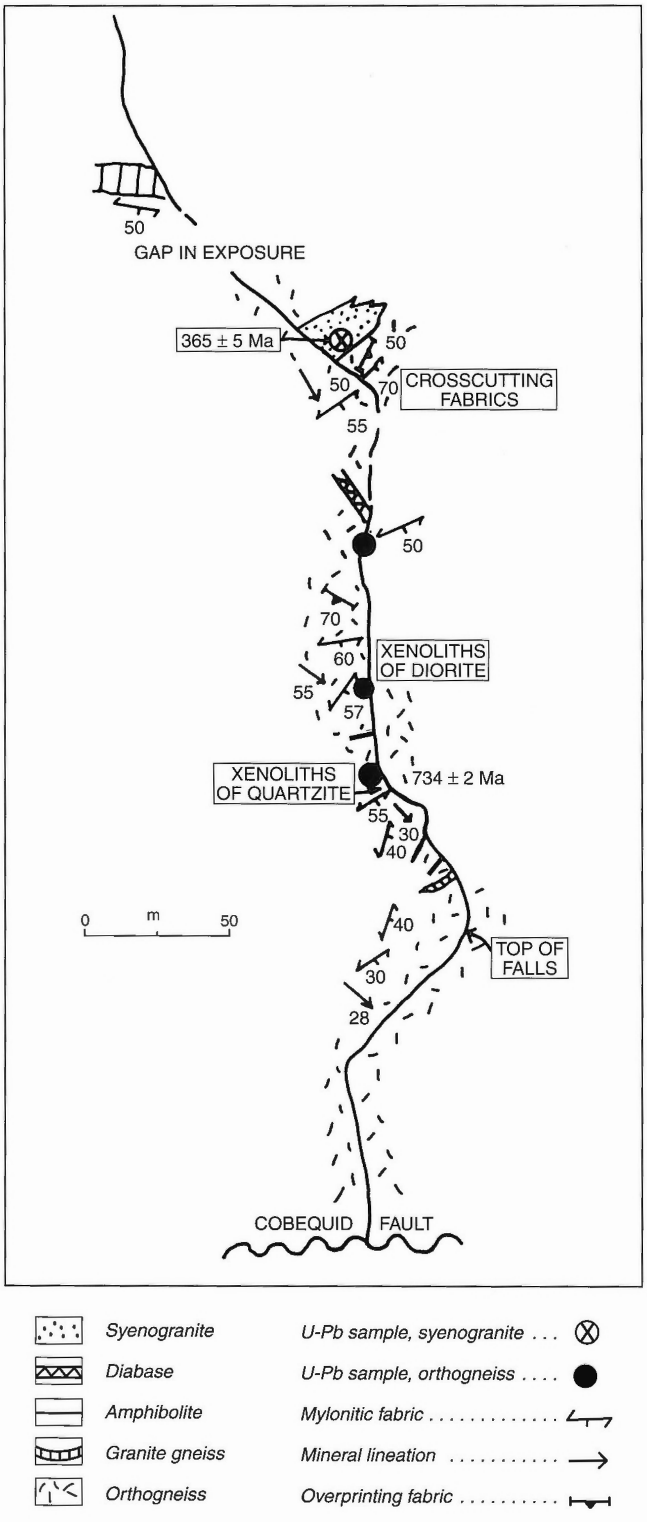


Figure 4. *Detailed map of the Economy River Gneiss on Economy River. For location of Economy River* see *Figure 1* (see also *Murphy et al., 2000).*

Late Paleozoic movement on the Cobequid fault system. This relationship demonstrates the polyphase nature of shear in the eastern Cobequid Highlands. The orthogneiss contains blocks of orthoquartzite (Fig. 6B) interpreted as xenoliths of the Gamble Brook Formation (Murphy et al., 1988; Doig et al., 1991b). This relationship may provide a minimum age for the deposition of the Gamble Brook Formation.

The geochemistry of the Economy River Gneiss is summarized in Table 3 (*see* below) which shows it to be predominantly intermediate to felsic in composition with SiO_2 contents in the 59–65 weight per cent range (volatile free). The gneiss is metaluminous and plots as tonalite in the normative An-Ab-Or diagram. The Fe_2O_3/MgO ratio varies from 1.5 to 2.6, typical of calc-alkaline rocks of intermediate SiO_2 composition. The gneiss is characterized by moderately enriched large-ion-lithophile elements such as K, Rb, Ba, and Th, as well as Ce and Sm relative to high-field-strength elements such as Nb, Zr, and Y (Fig. 7A), a trace-element signature typical of intermediate to felsic magmas generated in a volcanic-arc environment (Pearce et al., 1984). Rare-earth-element profiles (Fig. 7B), although not diagnostic in themselves, are consistent with the interpretation of the trace-element signature. These profiles display moderate LREE enrichment with $(La/Sm)_n$ varying from 6.8 to 2.9 and $(La/Yb)_n$ varying from 29.0 to 5.7.

The gneiss was sampled for geochronology at three localities on the Economy River from 300–600 m north of the Cobequid Fault (Fig. 5). Table 4 lists the isotopic data. All zircon fractions lie within error of a discordia (Fig. 5B) with an upper intercept of 734 ± 2 Ma and a lower intercept of -47 ± 115 Ma, indicating that there are no inherited components and/or that metamorphism is negligible.

The Economy River Gneiss is interpreted to represent part of an early or pre-Avalonian arc-related tectonothermal event, which together with the platformal Gamble Brook Formation form part of the basement to the late Precambrian stratigraphy of the Avalon composite terrane. The presence in the gneiss of xenoliths of the Gamble Brook Formation provides evidence for the antiquity of these platformal rocks. The ca. 734 Ma age of the Economy River Gneiss may provide a minimum age for similar, possibly correlative platformal units such as the George River Group of Cape Breton Island (Milligan, 1970). These data indicate that the basement to the platformal sequences has yet to be identified, contrary to arguments expressed in many Avalonian syntheses. This stresses the importance of evaluating basement-cover relationships in areas of the Avalon composite terrane where precise geochronological data are not available.

Folly River Formation

The Folly River Formation (unit 3, Fig. 1) comprises mafic flows (Fig. 8), hyaloclastites and tuffs, volcanogenic sedimentary rocks and minor jasperitic ironstone (Folly River Schist of Donohoe and Wallace (1982a, b, c, d); Cullen, 1984) together with interlayered chloritic metasedimentary rocks (interpreted as distal turbidite sequences), and abundant mafic dykes. The thickness of the formation is difficult to estimate due to structural complexities, but is probably

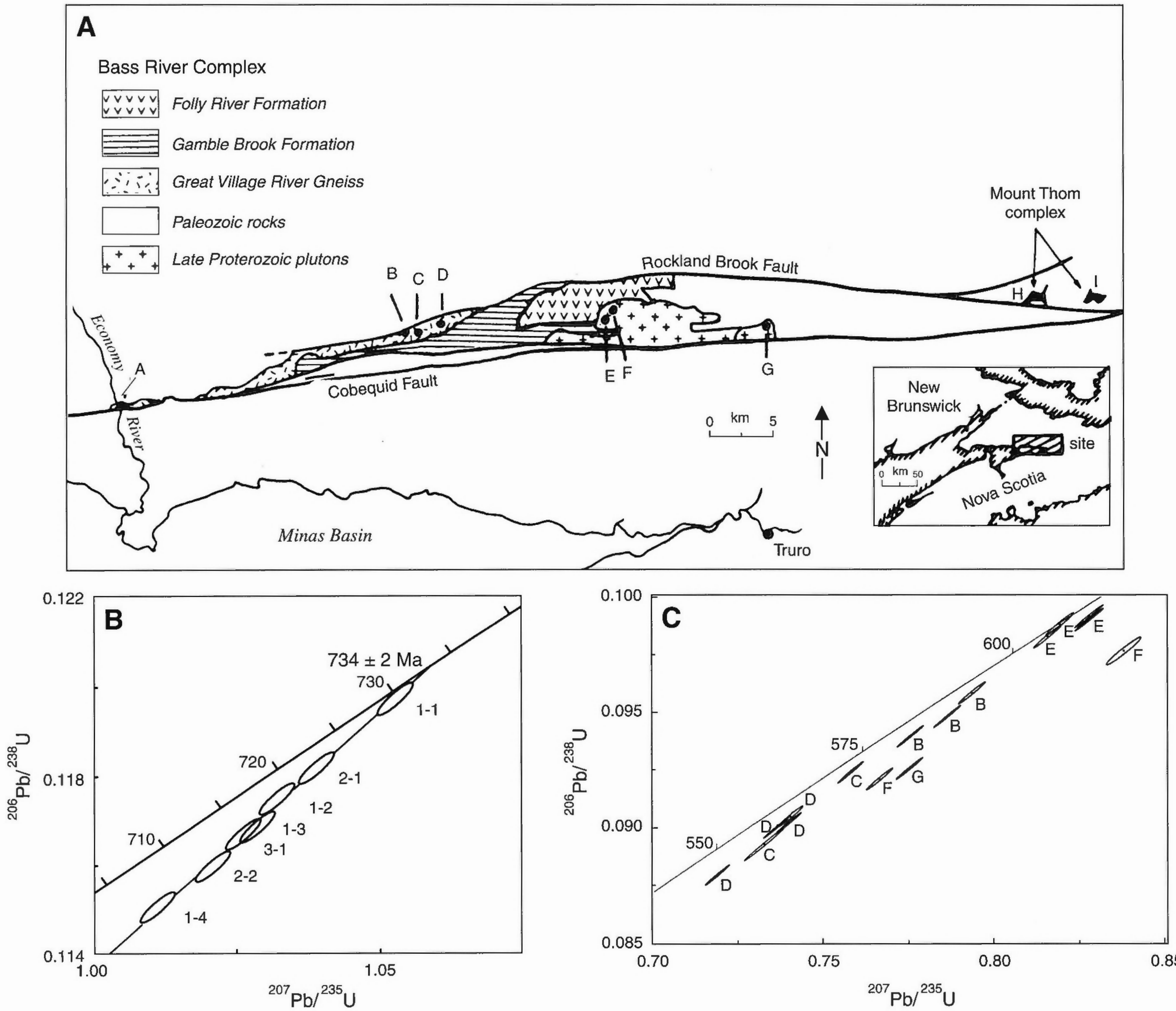
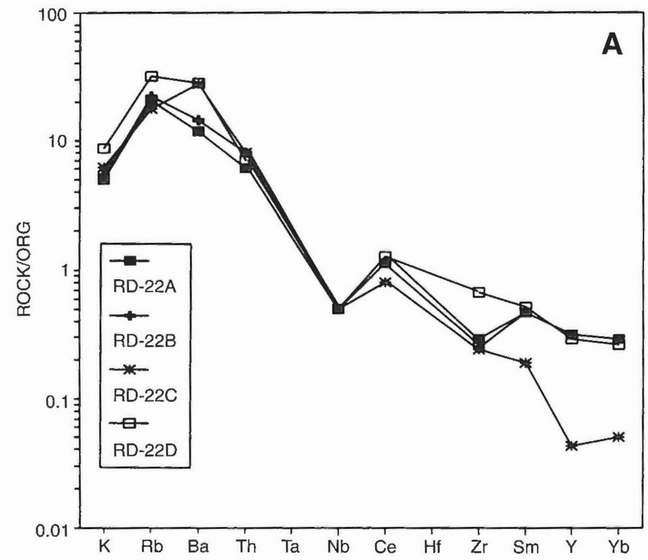


Figure 5. *Results of U-Pb zircon dating, eastern Cobequid Highlands.* ***A)*** *Location map of samples from Economy River Gneiss (A, unit 2, Fig. 1), mylonitic granite intruding along the contact between the Gamble Brook Formation and the Great Village River orthogneiss (B, unit 5b, Fig. 1), amphibolite in the Great Village Gneiss (C, unit 5a, Fig. 1), Great Village River orthogneiss (D, unit 5a, Fig. 1), and the Debert River (E) and McCallum Settlement (F) plutons. "H" and "I" are samples from the Mount Thom Complex. These samples are currently being dated.* ***B)*** *Uranium-lead concordia diagram for the Economy River Gneiss (sample A). Data is given in Table 4.* ***C)*** *Uranium-lead concordia diagram for samples B, C, D, E, F. Data is given in Table 4. Two additional points for McCallum Settlement occur below the base of the figure. Variation in* $^{207}Pb/^{206}Pb$ *dates of zircon fractions from the same sample is attributed to a small, but significant, inherited component. Both the gneissic rocks and the undeformed plutons yield similar ages, in the 610–580 Ma range.*

Figure 6.

A) Photograph showing overprint of earlier amphibolite-grade mylonitic fabric by later Paleozoic low-grade deformation. This demonstrates the polyphase nature of shearing in the eastern Cobequid Highlands. Photograph by B. Murphy. GSC 2000-015A. ***B)*** *Block of quartzite, attributed to the Gamble Brook Formation, within the Economy River Gneiss. If so, this demonstrates the antiquity of the Gamble Brook Formation. Photograph by B. Murphy. GSC 2000-015B*



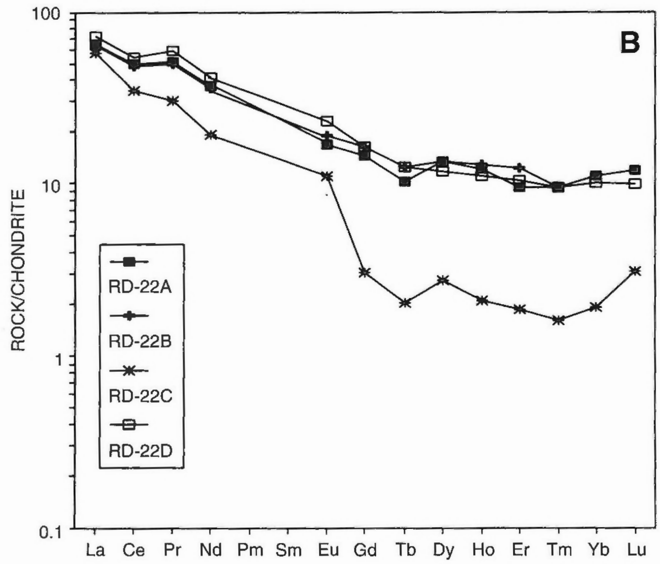


Figure 7. A) *Composition of samples from the Economy River Gneiss normalized to ocean-ridge granite (ORG; fields* after *Pearce et al., 1984).* ***B)*** *Rare-earth element plot for samples from the Economy River orthogneiss normalized to chondrite* (after *Sun, 1982).*

Table 3. Major, trace, and rare-earth element geochemistry of the Economy River Gneiss.

Sample	RD-22A	RD-22B	RD-22C	RD-22D	RD-31
SiO_2 (%)	58.57	60.27	64.04	60.35	61.92
Al_2O_3(%)	16.74	16.90	15.94	18.20	14.34
Fe_2O_3 (%)	7.60	6.66	4.50	5.04	8.98
MgO (%)	2.78	2.38	3.07	2.38	3.49
CaO (%)	5.29	4.83	3.43	3.86	1.61
Na_2O (%)	3.48	3.66	3.83	3.92	1.96
K_2O (%)	2.00	2.18	2.49	3.49	3.05
TiO_2 (%)	1.09	0.78	0.74	0.58	1.47
MnO (%)	0.09	0.07	0.07	0.06	0.25
P_2O_5 (%)	0.35	0.26	0.12	0.17	0.19
LOI (%)	1.30	1.10	0.90	1.10	1.90
Total (%)	99.29	99.09	99.13	99.15	99.16
Ba (ppm)	587	714	1376	1400	470
Rb (ppm)	81	89	71	125	153
Sr (ppm)	453	398	585	348	178
Y (ppm)	22	22	3	20	26
Zr (ppm)	97	83	82	229	280
Nb (ppm)	<5	<5	<5	<5	17
Pb (ppm)	<10	<10	<10	<10	20
Ga (ppm)	17	18	17	19	20
Zn (ppm)	44	39	84	29	202
Cu (ppm)	73	42	25	63	48
Ni (ppm)	7	12	36	10	40
V (ppm)	96	99	107	82	194
Cr (ppm)	29	39	82	36	109
La (ppm)	20.5	20.0	18.1	22.7	45.1
Ce (ppm)	39.8	39.3	27.9	44.1	90.6
Pr (ppm)	5.1	4.9	3.0	5.9	11.6
Nd (ppm)	21.8	21.3	11.5	24.7	46.5
Sm (ppm)	4.2	4.2	1.7	4.7	9.0
Eu (ppm)	1.22	1.36	0.80	1.66	1.92
Gd (ppm)	3.7	4.2	0.8	4.2	7.5
Tb (ppm)	0.5	0.6	0.1	0.6	1.1
Dy (ppm)	4.3	4.3	0.9	3.8	5
Ho (ppm)	0.87	0.92	0.15	0.79	1.00
Er (ppm)	2.0	2.6	0.4	2.2	2.8
Tm (ppm)	0.3	0.3	<0.1	0.3	0.3
Yb (ppm)	2.3	2.3	0.4	2.1	2.2
Lu (ppm)	0.38	0.38	0.10	0.32	0.35
Th (ppm)	6	8	8	7	11
U (ppm)	2	1	2	21	1

Note: LOI = loss on ignition.

about 600 m. The contact between the Gamble Brook Formation and the overlying Folly River Formation, exposed on East Folly River and on a tributary of East Folly River, is interpreted as an unconformity. In the latter locality, intense mylonitic fabric in the Gamble Brook Formation is truncated by a mafic dyke assigned to the Folly River Formation. In both localities, the contact marks a major lithological break between the quartzite-dominated metasedimentary rocks to the north (Gamble Brook Formation) and the mafic volcanic-dyke complex to the south (Folly River Formation). An unconformity is also suggested by contrasting structural styles in the Gamble Brook and Folly River formations (Nance and Murphy, 1990). The top of the Folly River Formation is not exposed. Excellent sections of this formation may be seen on Folly River, east Folly River, and Debert River, with the most complete section exposed on Debert River.

Mafic volcanic rocks and dykes vary locally from massive to schistose along sharp to gradational contacts. In the schistose rocks, primary igneous minerals and textures are largely replaced by a fabric of chlorite, sericite, epidote, and more rarely, actinolite and biotite. In the Debert River section, lavas, consisting of plagioclase augite, actinolite, hornblende, Fe-Ti-oxide minerals, chlorite, epidote, rare quartz, and biotite, show a variety of igneous textures, including ophitic to subophitic, pilotaxitic, porphyritic, and vesicular. Flows are a few decimetres to metres in thickness. Flow tops, commonly highly vesicular, locally show pillowing and the development of hyaloclastite (Cullen, 1984). Hyaloclastite breccia, volcanogenic sedimentary rocks, crystal and lithic tuffs, and rare agglomerate beds also occur. The freshest lavas contain relict primary augite (about 25%) present as fine to coarse anhedral to subhedral grains with various degrees of alteration to actinolite and dusty opaque minerals. Plagioclase (about 15%) occurs as fine- to medium-grained laths (An_{48} to An_{58}) showing partial to complete albitization and alteration mostly to actinolite and chlorite. Actinolite (about 20–30%) forms aggregates and medium to coarse grains in lenticular pockets with chlorite and epidote and is also present as scattered grains throughout the matrix. Minor biotite is spatially associated with actinolite. Chlorite (about 15–25%) forms fibrous aggregates subparallel to the fabric and commonly contains inclusions of epidote. Epidote (about 14–20%) forms fine to coarse anhedral grains mostly in association with actinolite and chlorite. Opaque minerals (about 15–20%) occur as anhedral crystals displaying igneous textures, and as dusty secondary anhedral patches which may form a discontinuous and locally irregular foliation in the rock. Quartz and calcite, where present, form very fine-grained patches and may occur in veins along with tiny epidote crystals. They are very minor constituents of the rock.

Table 4. Uranium-lead data and locations of samples. **A)** Uranium-lead data of the Great Village River Gneiss, mylonitic granitic rocks, and Debert River pluton.

Zircon fraction[a]	Weight (mg)	U (ppm)	Pb_{rad} (ppm)	$^{206}Pb/^{204}Pb$[b]	$^{208}Pb/^{206}Pb$[c]	$^{206}Pb/^{238}U$[c]	$^{207}Pb/^{235}U$[c]	$^{207}Pb/^{206}Pb$[c]	$^{207}Pb/^{206}Pb$ age (Ma)
Orthogneiss, Great Village River road									
B-1 N0, 25 x 100	0.380	446	43.3	1092	0.131	0.09480	0.7865	0.06017	610
B-2 M0, 50 x 125	0.284	313	31.1	1470	0.147	0.09581	0.7937	0.06008	606
B-3 M2, 70 x 150	0.441	479	46.4	1601	0.140	0.09394	0.7756	0.05988	599
Amphibolite, Great Village River									
C-1 M1, 70 x 200	0.250	899	94.3	1222	0.304	0.08927	0.7328	0.05954	587
C-2 N4, 70 x 200	0.090	984	107.1	2853	0.308	0.09240	0.7582	0.05951	586
Orthogneiss, Rockland Brook									
D-1 N0, 150 x 250	0.240	241	23.7	2327	0.203	0.09020	0.7400	0.05950	585
D-2 M0, 75 x 150	0.143	286	27.0	905	0.186	0.08794	0.7191	0.05931	579
D-3 M1, 75 x 150	0.223	277	26.8	1761	0.190	0.08999	0.7363	0.05934	580
D-4 M2, 75 x 150	0.103	350	33.7	1419	0.177	0.09048	0.7405	0.05936	580
Debert River pluton									
E-1 N0, 125 x 300	0.177	414	42.2	1141	0.139	0.09880	0.8194	0.06015	609
E-2 N0, 25 x 150	0.102	228	23.4	1209	0.155	0.09826	0.8258	0.06021	611
E-3 N0, 75 x 200	0.323	212	21.8	1848	0.147	0.09900	0.8280	0.06066	627
E-4 N0, 60 x 150	0.160	251	25.9	1054	0.151	0.09914	0.8279	0.06057	624
F-1 N0, 70 x 200	0.073	418	42.1	1113	0.138	0.09763	0.8380	0.06225	683
F-2 M2, 80 x 250	0.111	186	17.9	1228	0.152	0.09210	0.7666	0.06037	617
McCallum Settlement pluton									
G-1 N0, 50 x 150	0.241	357	34.4	1022	0.148	0.09256	0.7754	0.06076	631
G-2 M0, 50 x 150	0.094	588	51.3	873	0.137	0.08484	0.6972	0.05960	589
G-3 M1, 50 x 150	0.157	649	51.9	1043	0.124	0.07853	0.6426	0.05935	580

[a] N and M are nonmagnetic and magnetic fractions, respectively and the numerals 0, 1, 2, and 4 are degrees tilt on a Frantz separator; 25 x 100 indicates zircon dimensions in micrometres (μm).
[b] Atomic ratios corrected for fractionation and spike.
[c] Atomic ratios corrected for fractionation, spike, blanks, and common Pb from the model of Stacey and Kramers (1975).

Table 4. B) Uranium-lead data of the Economy River Gneiss.

Zircon fraction[a] (mg)	Weight (ppm)	U (ppm)	Pb_{rad}	$^{206}Pb/^{204}Pb$[b]	$^{208}Pb/^{206}Pb$[c]	$^{206}Pb/^{238}U$[c]	$^{207}Pb/^{235}U$[c]	$^{207}Pb/^{206}Pb$[c]	$^{207}Pb/^{206}Pb$ age (Ma)
1-1 N0, Ab, 50 x 100	0.210	360	50.3	1.362	0.2904	0.11974	1.0527	0.06377	734
1-2 N0, 60 x 125	0.263	386	51.8	1.414	0.2634	0.11750	1.0320	0.06371	732
1-3 N0, 35 x 70	0.236	429	57.5	1.110	0.2671	0.11687	1.0286	0.06383	736
1-4 M4, 35 x 70	0.148	437	56.8	1.282	0.2475	0.11506	1.0112	0.06374	733
2-1 N0, 125 x 250	0.287	611	78.6	1.220	0.2009	0.11819	1.0389	0.06375	733
2-2 M0, 100 x 200	0.146	843	106.0	1.339	0.1958	0.11598	1.0208	0.06383	736
3-1 N0, 80 x 200	0.299	408	52.7	2.453	0.2223	0.11671	1.0261	0.06377	734

[a] N and M are nonmagnetic and magnetic fractions, respectively, and the numerals 0 and 4 are degrees tilt on the Franz separator; 50 x 100, etc. are zircon dimensions in micrometres (μm). All grains were clear, euhedral and pale yellow to brown. Only fraction 1-4 contained small black inclusions. Ab, abraded.
[b] Atomic ratios corrected for fractionation and spike.
[c] Atomic ratios corrected for fractionation, spike, blanks, and common Pb from the model of Stacey and Kramers (1975).

Table 4. C) Summary of U-Pb age dates and locations.

Sample	Unit and lithology	Locality	NTS, grid reference	Age (Ma)
A	Economy River Gneiss; orthogneiss	Economy River	11 E/5, 280329	734 ± 2
B	Mylonitic granite	Road adjacent to Great Village River	11 E/5, 510385	605 ± 5
C	Great Village River Gneiss; amphibolite	Great Village River	11 E/5, 507387	589 ± 5
D	Great Village River Gneiss; orthogneiss	Rockland Brook	11 E/5, 530397	580 ± 5
E	Debert River pluton; granite	Southern contact Derbert River	11 E/6, 653394	609 ± 4
F	Debert River pluton; granite	Northern contact of pluton	11 E/6, 662413	612 ± 4
G	McCallum Settlement pluton; granite	Tributary of North River	11 E/6, 790402	575 ± 5
H	Mount Thom Complex; amphibolite	Road west of Mount Ephraim	11 E/6, 042425	Indeterminate
I	Mount Thom Complex; gneiss	Mount Thom	11 E/7, 996425	Indeterminate

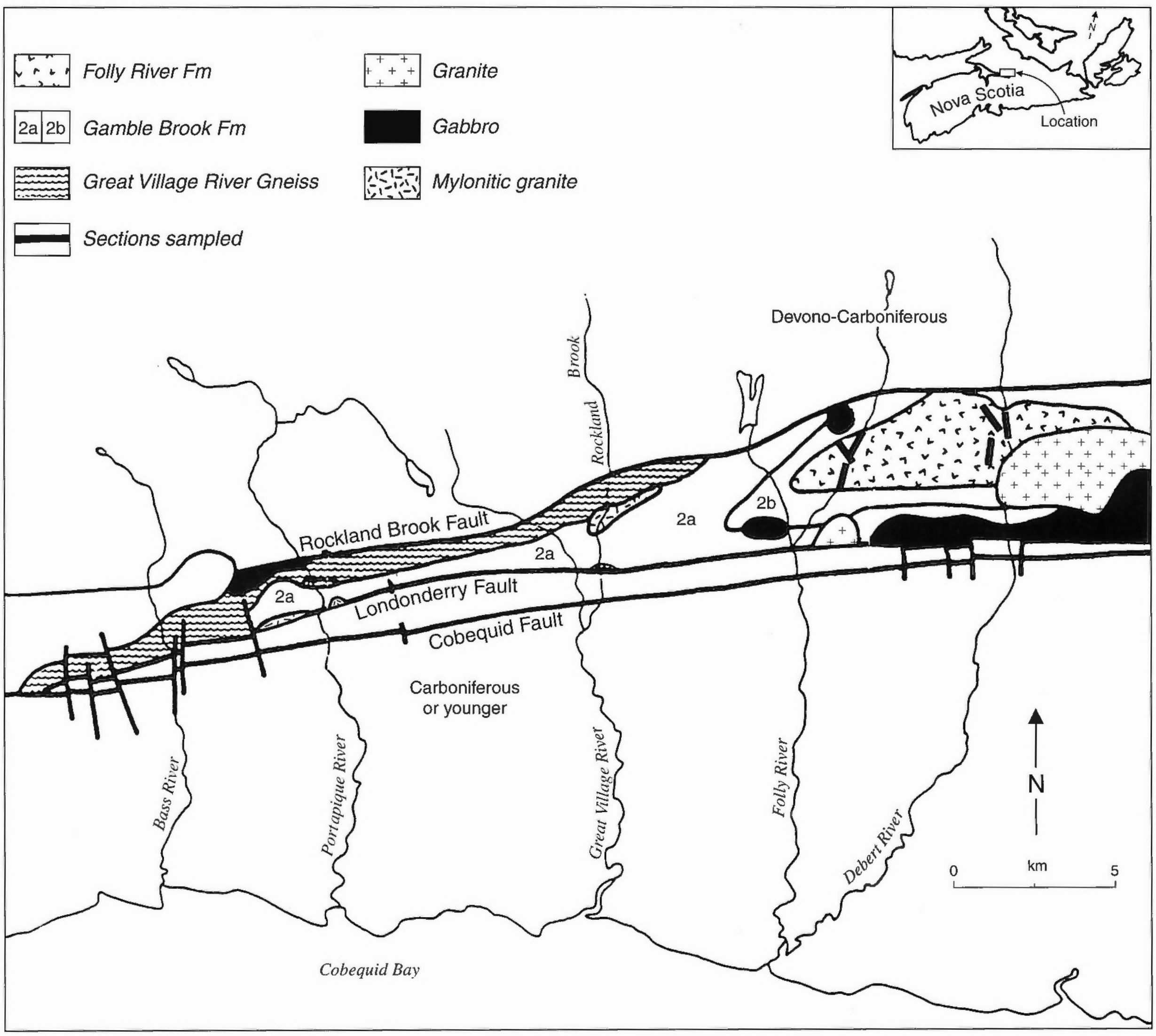


Figure 8. Geological map of the Bass River Complex showing locations of the sections from which the Folly River Formation was sampled.

Table 5. A) Location of analyzed samples, Folly River Formation. *See* Figure 8 for sections sampled.

Sample number	Latitude	Longitude	Location
10-4-11	45°31′N	63°27′W	Debert River
10-4-5	45°32′N	63°27′W	Debert River
10-4-1	45°32′N	63°27′W	Debert River
6-3-2	45°32′N	63°27′W	Debert River
10-4-3	45°32′N	63°27′W	Debert River
10-4-9	45°31′N	63°27′W	Debert River
10-4-2	45°32′N	63°27′W	Debert River
10-4-10	45°31′N	63°27′W	Debert River
6-2-2	45°31′N	63°27′W	Debert River
10-4-7	45°31′N	63°27′W	Debert River
10-4-8	45°31′N	63°27′W	Debert River
10-5-1	45°31′N	63°27′W	Debert River
10-4-6	45°31′N	63°27′W	Debert River
9-6-6	45°31′N	63°30′W	East Folly River
9-5-5	45°31′N	63°30′W	East Folly River
9-5-9	45°31′N	63°30′W	East Folly River
9-5-8	45°31′N	63°30′W	East Folly River
9-5-1	45°31′N	63°30′W	East Folly River
5-2-2	45°30′N	63°36′W	Rockland Brook
20-1-6	45°30′N	63°36′W	Rockland Brook
21-1-6	45°30′N	63°38′W	Great Village River
13-3-1	45°28′N	63°48′W	Gamble Brook
12-6-1	45°30′N	63°38′W	Great Village River
20-1-3	45°30′N	63°37′W	Rockland Brook
2-5-2	45°31′N	63°32′W	Folly River
24-3-4	45°31′N	63°30′W	East Folly River
22-1-2	45°32′N	63°30′W	Trib. East Folly River
22-1-1	45°32′N	63°30′W	Trib. East Folly River

Table 5. B) Geochemistry of the mafic flows and dykes of the Folly River Formation.

Analysis	1	2	3	4	5	6	7	8	9	10	11	12	13
Sample number	10-4-1	10-4-5	10-4-1	6-2-3	10-4-3	10-4-9	10-4-2	10-4-10	6-2-2	10-4-7	10-4-8	10-5-1	10-4-6
Major elements (wt %)													
SiO_2	46.65	47.29	47.47	47.52	47.62	47.75	47.95	48.11	48.78	49.03	49.51	50.00	51.52
TiO_2	2.24	2.27	2.13	3.41	2.18	2.29	2.56	2.21	2.73	2.77	1.76	2.56	2.97
Al_2O_3	12.68	12.78	12.65	12.51	13.31	12.64	12.94	12.46	12.70	12.56	13.13	13.48	13.93
$Fe_2O_3^t$	14.71	14.76	15.07	15.00	13.16	15.15	14.42	14.36	14.64	14.50	11.13	13.05	12.83
MnO	0.22	0.26	0.25	0.23	0.23	0.24	0.29	0.23	0.38	0.33	0.23	0.24	0.23
MgO	6.45	6.77	7.02	5.22	6.55	6.30	6.16	6.54	5.45	5.28	7.21	5.03	3.66
CaO	10.94	9.88	10.00	10.07	9.25	9.49	10.37	10.17	9.51	9.10	9.89	11.68	5.33
Na_2O	2.45	2.01	2.50	2.55	3.03	2.71	2.79	2.55	2.58	2.16	3.29	0.61	2.96
K_2O	0.13	0.08	0.13	0.87	0.42	0.15	0.77	0.16	0.80	1.10	0.17	0.12	2.51
P_2O_5	0.19	0.19	0.17	0.49	0.19	0.19	0.31	0.18	0.38	0.37	0.12	0.16	0.87
LOI	2.60	2.40	2.00	0.60	3.30	2.30	0.20	1.90	1.10	1.20	3.10	3.80	2.10
Total	99.26	98.69	99.39	98.47	99.24	99.21	98.76	98.87	99.05	98.40	99.54	100.73	98.91
Trace elements (ppm)													
Ba	29	35	60	258	103	89	216	43	262	280	89	24	929
Rb	1	0	0	19	10	2	21	0	18	27	2	3	50
Sr	190	85	208	261	204	233	253	248	260	232	157	290	443
Y	37	40	37	38	31	41	35	36	37	37	27	30	42
Zr	147	151	133	240	140	153	107	144	218	222	110	187	449
Nb	7	4	5	16	7	3	12	8	14	13	5	13	26
Th	0	9	0	5	0	0	0	12	0	4	0	0	0
Pb	0	0	0	3	0	0	0	0	0	3	0	0	0
Ga	17	21	25	29	15	25	21	8	22	23	14	30	9
Zn	119	130	132	127	111	134	173	199	150	129	90	107	144
Cu	75	82	62	116	60	86	157	81	134	138	68	37	39
Ni	56	49	53	44	62	54	59	55	52	55	69	92	24
V	467	484	468	400	410	491	397	454	375	397	369	237	252
Cr	102	91	87	56	110	83	149	111	87	113	219	169	16
La	n.d.	n.d.	5.70	20.90	n.d.	n.d.	n.d.	n.d.	n.d.	19.70	n.d.	n.d.	43.20
Ce	n.d.	n.d.	17.00	52.00	n.d.	n.d.	n.d.	n.d.	n.d.	47.00	n.d.	n.d.	96.00
Nd	n.d.	n.d.	14.00	27.00	n.d.	n.d.	n.d.	n.d.	n.d.	27.00	n.d.	n.d.	49.00
Sm	n.d.	n.d.	4.88	7.63	n.d.	n.d.	n.d.	n.d.	n.d.	7.10	n.d.	n.d.	11.00
Eu	n.d.	n.d.	1.70	2.46	n.d.	n.d.	n.d.	n.d.	n.d.	2.18	n.d.	n.d.	3.10
Tb	n.d.	n.d.	1.30	1.40	n.d.	n.d.	n.d.	n.d.	n.d.	1.40	n.d.	n.d.	1.50
Yb	n.d.	n.d.	4.64	4.09	n.d.	n.d.	n.d.	n.d.	n.d.	4.10	n.d.	n.d.	4.59
Lu	n.d.	n.d.	0.67	0.59	n.d.	n.d.	n.d.	n.d.	n.d.	0.60	n.d.	n.d.	0.68

Notes: Sample locations are listed in Appendix A.
Analyses 1 to 13: flows and dykes, Debert River area; 14 to 18: flows interlayered with turbiditic rocks; 19 to 24: dykes cutting Great Village River Gneiss; 25 to 26: dykes cutting the Gamble Brook Formation; 27 to 29: dykes cutting the Folly River Formation lavas. Major elements and Ba-Cr trace elements were analyzed by X-ray fluorescence at St. Mary's University on a Philips PW1400 sequential spectrometer using a Rh-anode X-ray tube. International standards with recommended values from Abbey (1983) as well as in house standards were used for calibration. Analytical precision, as determined from replicate analyses is generally better than 2%, except MgO, Na_2O, and Nb which are better than 5%, and Th which is better than 10%.
Loss on ignition (LOI) was determined by treating the sample for 1.5 h at 1050°C in an electric furnace. The rare-earth element concentrations were determined by neutron activation analysis at McMaster University.
n.d.: no data

Interbedded chert and siliciclastic sedimentary rocks are generally thinly plane-laminated, thin-bedded, medium- to fine-grained turbidite, with rare grading and crossbedding. The thinly bedded character of these turbidite sequences and the general absence of slump structures suggest either a proximal levee, back levee, or distal outer fan environment (Murphy et al., 1990).

Abundant dykes within the Folly River Formation may be isolated, or form continuous zones of sheeted dykes several hundred metres in length. The dykes petrographically and geochemically resemble the lavas, but exhibit chilled margins up to 3 cm in width, and vary from fine-grained porphyritic margins to coarse-grained ophitic to subophitic cores. The cores of the dykes are medium to coarse grained and contain plagioclase, and clinopyroxene relics (about 10%) in ophitic to subophitic intergrowth in addition to secondary phases such as actinolite, opaque minerals, biotite, clinopyroxene, and locally, hornblende. Quartz, chlorite, and epidote occur as vein minerals. Plagioclase (about 40–50%) is present as medium to large laths that show alteration to sericite. Actinolite (about 25–40%) is present in the groundmass as coarse laths, and as an alteration of clinopyroxene, leaving only clinopyroxene relics behind. Opaque minerals (about 10–20%) are present throughout the rock as an alteration product together with actinolite and as fine to medium anhedral to subhedral grains. Biotite (about 5%) is present as tiny flakes spatially associated with actinolite and opaque minerals, as stringers, and may form as an alteration product of clinopyroxene. Similar dykes cut the Gamble Brook Formation and the high-grade gneiss, and may have been feeders to the lavas. However these dykes geochemically resemble Carboniferous dykes, for example those analyzed by Pe-Piper (1991) from the North River pluton and by Pass (1993) from the Squally Point volcanic rocks). Unpublished lead isotope data suggests that some dykes previously regarded as equivalent to the Folly River Formation are probably Carboniferous, implying that correlations of dykes based on geochemical data should be viewed with caution.

Thirty samples of lavas and dykes from the Folly River Formation were analyzed in order to determine their geochemical affinities. The locations of the analyzed mafic rocks are given in Table 5A and their whole-rock geochemistry are listed in Table 5B. Relics of igneous minerals in both lavas and dykes were analyzed by electron microprobe using the method of Clarke (1976). Our data indicate that the original plagioclase was predominantly labradorite. Of the amphiboles, igneous hornblende with a variable TiO_2 content (0.9–2.8%) is dominant. The rare biotite flakes are iron-rich

Table 5B (cont.)

Analysis	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29
Sample number	9-6-6	9-5-5	9-5-9	9-5-8	9-5-1	5-2-2	20-1-6	21-1-6	13-3-1	12-6-1	20-1-3	2-2-3	2-5-2	24-3-4	22-1-2	22-1-1
Major elements (wt %)																
SiO_2	45.20	48.03	48.83	49.40	49.73	44.73	46.43	47.76	48.79	48.88	48.91	45.88	47.96	48.34	48.84	48.90
TiO_2	2.86	2.14	2.06	1.93	1.96	2.59	3.86	2.89	1.48	2.21	1.57	2.68	2.76	1.32	2.10	2.14
Al_2O_3	15.31	12.47	12.93	13.02	13.12	14.38	13.00	13.54	14.47	14.01	14.70	14.06	13.15	15.35	16.06	16.11
$Fe_2O_3^t$	14.20	15.15	13.24	13.84	13.00	15.82	15.37	14.11	11.69	12.89	11.79	16.75	13.70	11.12	11.46	11.84
MnO	0.31	0.29	0.24	0.22	0.22	0.28	0.42	0.36	0.19	0.35	0.17	0.32	0.26	0.17	0.34	0.35
MgO	6.67	7.30	6.70	5.97	6.43	6.51	5.37	5.26	7.10	6.41	6.37	5.66	5.92	7.28	6.22	5.94
CaO	8.15	9.62	10.68	10.23	10.40	8.29	8.55	8.76	11.40	9.13	11.68	8.17	10.35	10.54	8.44	8.35
Na_2O	2.27	2.64	2.85	2.98	3.62	2.42	2.88	2.57	2.16	2.84	2.20	3.07	3.05	2.79	2.84	2.20
K_2O	1.59	0.17	0.14	0.18	0.20	1.85	1.25	1.23	0.60	1.15	0.61	1.19	0.98	0.29	1.24	1.21
P_2O_5	0.40	0.17	0.17	0.15	0.17	0.23	0.79	0.42	0.12	0.32	0.14	0.21	0.37	0.11	0.33	0.33
LOI	2.20	0.90	0.90	0.60	0.50	1.40	0.70	1.20	0.80	0.70	1.00	0.50	0.40	1.70	1.10	1.10
Total	99.16	98.88	98.74	98.52	99.35	98.50	98.62	98.10	98.80	98.89	99.14	98.49	98.90	99.01	98.97	98.47
Trace elements (ppm)																
Ba	286	23	27	35	93	472	474	460	162	281	285	408	247	86	278	245
Rb	63	0	2	2	1	61	31	34	15	34	21	30	26	0	44	41
Sr	301	95	120	148	102	214	341	317	235	308	211	315	340	186	526	524
Y	32	32	30	30	35	39	40	41	19	35	22	31	30	28	38	39
Zr	201	132	131	123	143	162	289	213	92	245	109	155	183	90	191	188
Nb	11	5	4	4	5	12	25	18	7	16	8	10	14	4	22	21
Th	0	0	0	0	0	0	0	0	0	0	0	0	0	0	2	0
Pb	0	0	0	0	0	0	0	0	0	28	0	0	0	0	0	0
Ga	21	17	22	21	16	24	24	23	20	22	22	26	22	17	22	22
Zn	166	137	119	107	127	128	227	192	138	242	100	143	111	98	141	133
Cu	56	91	71	88	62	223	75	70	123	77	141	78	98	62	75	40
Ni	92	45	45	35	61	68	41	40	101	61	93	73	46	81	76	77
V	282	450	410	409	413	537	398	412	362	290	368	487	386	278	254	241
Cr	45	58	59	61	107	96	67	94	228	147	218	125	134	295	24	32
La	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	29.50	n.d.	n.d.	n.d.	7.90	n.d.	n.d.	n.d.	n.d.	n.d.
Ce	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	65.00	n.d.	n.d.	n.d.	18.00	n.d.	n.d.	n.d.	n.d.	n.d.
Nd	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	39.00	n.d.	n.d.	n.d.	11.00	n.d.	n.d.	n.d.	n.d.	n.d.
Sm	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	9.09	n.d.	n.d.	n.d.	3.42	n.d.	n.d.	n.d.	n.d.	n.d.
Eu	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	2.91	n.d.	n.d.	n.d.	1.11	n.d.	n.d.	n.d.	n.d.	n.d.
Tb	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	1.40	n.d.	n.d.	n.d.	0.70	n.d.	n.d.	n.d.	n.d.	n.d.
Yb	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	4.20	n.d.	n.d.	n.d.	2.27	n.d.	n.d.	n.d.	n.d.	n.d.
Lu	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.59	n.d.	n.d.	n.d.	0.32	n.d.	n.d.	n.d.	n.d.	n.d.

Table 6. Average electron microprobe analyses of clinopyroxene samples from the mafic rocks of the Folly River Formation.

Sample	LAVAS		DYKES	
	6-2-3	6-2-2	10-4-7	2.5-2
SiO_2 wt %	48.94(1.13)	49.72(1.12)	51.01(0.89)	47/56(0/99)
TiO_2 wt %	1.76(0.43)	1.35(0.39)	1.07(0.26)	1.29(0.34)
Al_2O_3 wt %	4.21(0.77)	3.36(1.09)	3.44(0.76)	3.54(0.97)
FeO* wt %	10.75(1.25)	12.17(2.76)	9.45(0.77)	9.32(1.62)
MnO wt %	0.26(0.04)	0.32(0.10)	0.24(0.03)	0.24(0.05)
MgO wt %	13.46(1.09)	13.44(1.27)	15.14(0.56)	14.37(1.15)
CaO wt %	20.07(0.73)	19.20(1.01)	19.46(0.73)	20.33(0.58)
Na_2O wt %	0.46(0.07)	0.39(0.06)	0.38(0.03)	0.44(0.07)
Cr_2O_3 wt %	0.14(0.10)	0.10(0.12)	0.28(0.10)	0.17(0.13)
Number of analyses	11	19	19	12
Standard deviations are given in parentheses; FeO* = total Fe recalculated as FeO.				

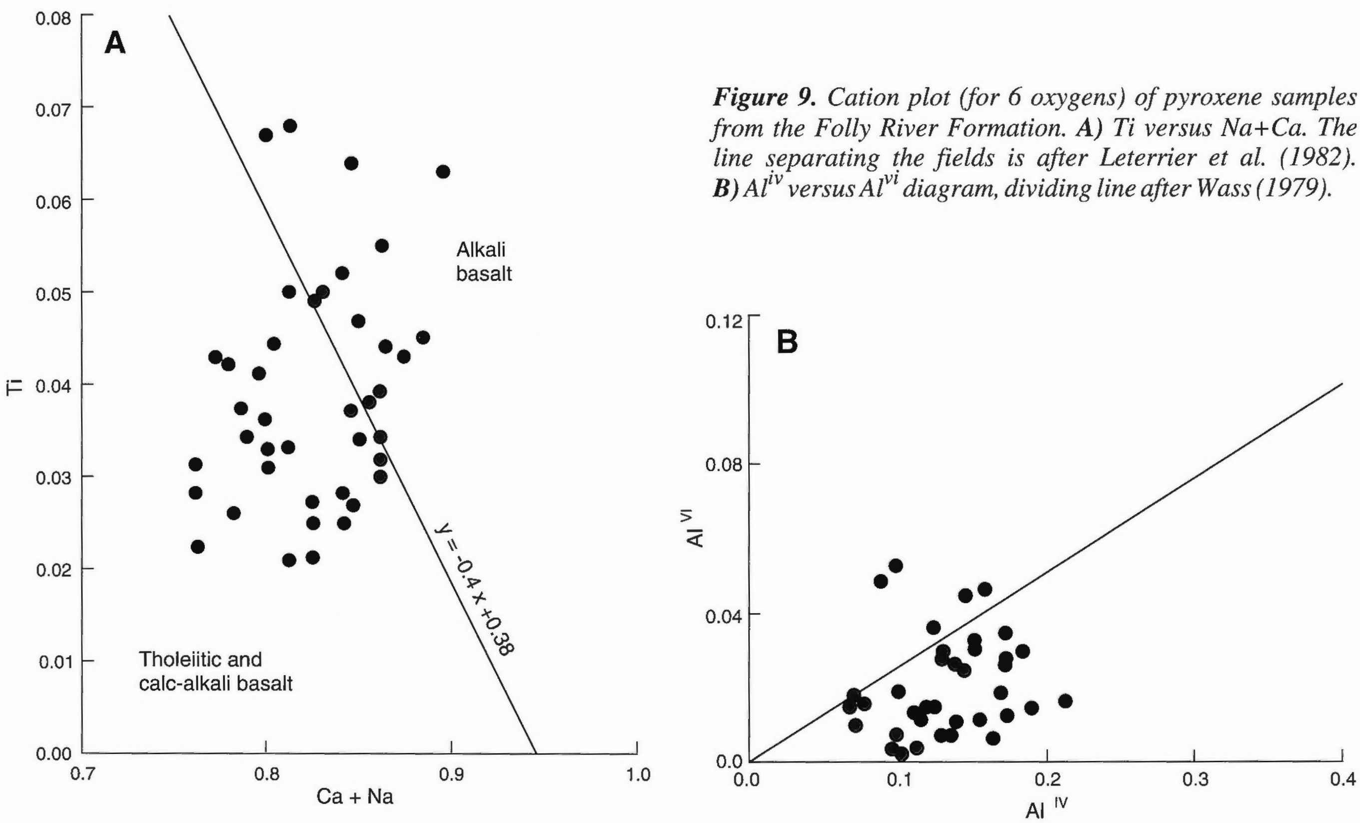
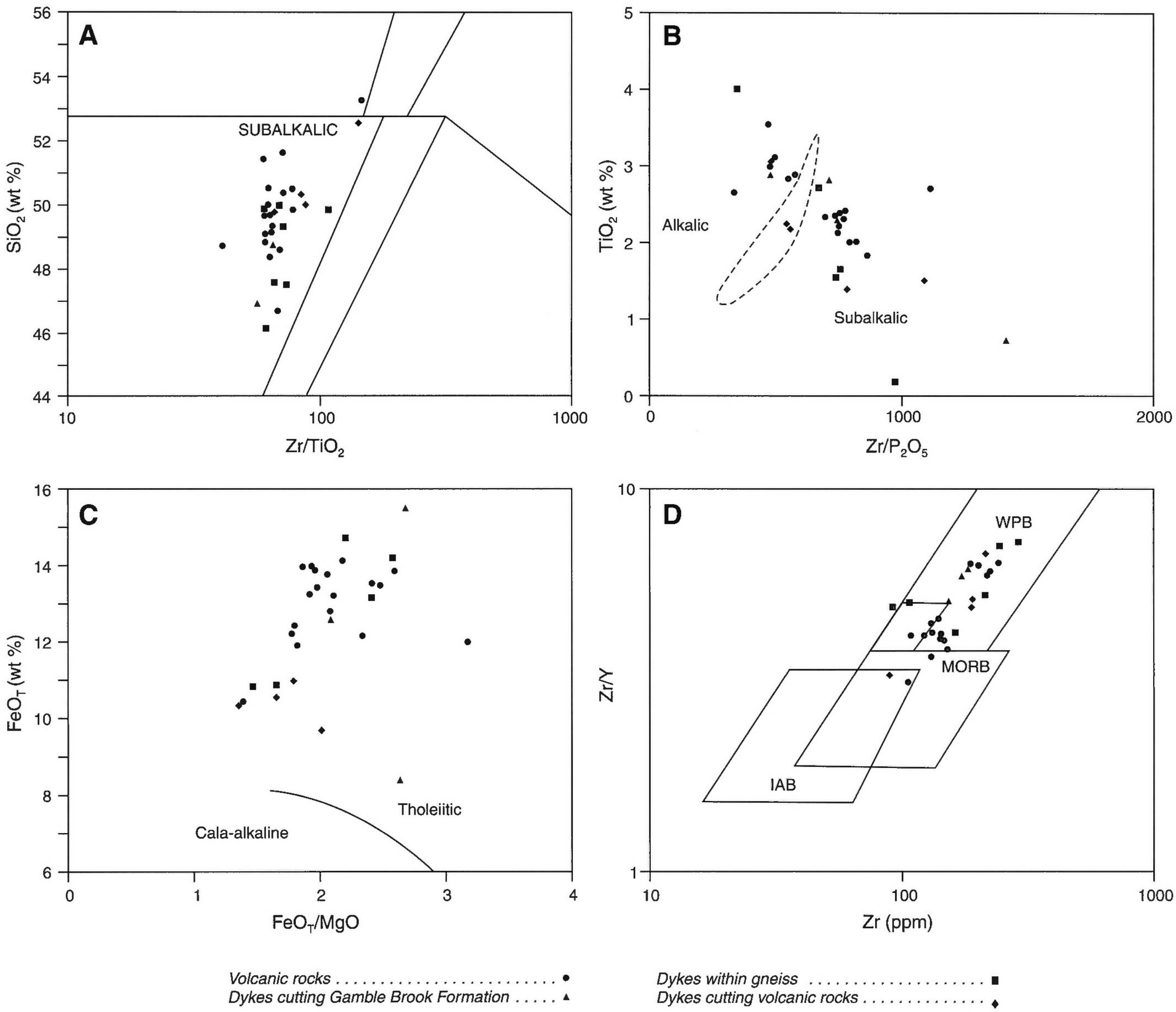


***Figure 9.** Cation plot (for 6 oxygens) of pyroxene samples from the Folly River Formation. **A)** Ti versus Na+Ca. The line separating the fields is after Leterrier et al. (1982). **B)** Al^{iv} versus Al^{vi} diagram, dividing line after Wass (1979).*

biotite (FeO_T about 24%) with high TiO_2 (about 3%). Average analyses of clinopyroxene crystals from representative samples are given in Table 6 and are typical of titanaugite. On a plot of Ti versus Ca+Na (Fig. 9A), most individual clinopyroxene analyses plot in the field for subalkalic basalt of Leterrier et al. (1982), although a substantial proportion (one third) of the analyses plot in the alkali-basalt field. This may indicate a transitional character for these mafic rocks. A plot of Al^{vi} versus Al^{iv} for the same analyses (Fig. 9B) indicates in general a low-pressure origin.

Whole-rock geochemical analyses show typical magmatic trends as exemplified by the positive correlations between Fe, Ti, and P. The SiO_2 ranges from 46 to 52% (Table 5B) and together with the high Fe and Mg confirm the basaltic and mafic character of the magmatism in the Folly River Formation. According to SiO_2 versus Zr/TiO_2 and TiO_2 versus Zr/P_2O_5 discrimination diagrams (Fig. 10A, B, *after* Floyd and Winchester, 1975; Winchester and Floyd, 1977), the basalt samples are subalkalic and according to FeO_T versus FeO_T/MgO the basaltic rocks are of tholeiitic nature (Fig. 10C, *after* Miyashiro, 1974). The apparent alkalic character of some samples (Fig. 10B) is attributed to Ti enrichment during crystal fractionation of a primary subalkalic magma. This Ti enrichment is typical of fractionated tholeiitic suites (Miyashiro 1974). The high FeO_T and wide range in FeO_T/MgO for rocks with similar SiO_2 is typical of differentiated tholeiitic suites as is the positive



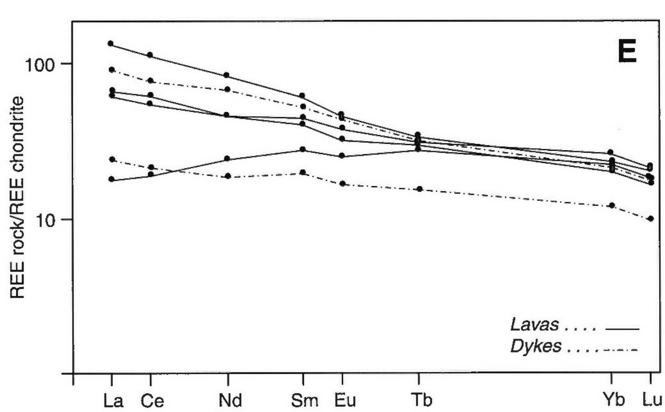


Figure 10. *Petrogenetic affinity of mafic rocks of the Folly River Formation;* ***A)*** *SiO_2 versus Zr/TiO_2 fields* after *Winchester and Floyd, 1977,* ***B)*** *TiO_2 versus Zr/P_2O_5 fields* after *Floyd and Winchester, 1975,* ***C)*** *FeO_T versus FeO_T/MgO fields* after *Miyashiro, 1974,* ***D)*** *Zr/Y versus Zr fields* after *Pearce and Norry, 1979. IAB = island-arc basalt; MORB = mid-ocean ridge basalt; WPB = within-plate basalt.* ***E)*** *REE plot.*

correlation between Fe, Ti, P, Zr, and V. The REE distribution (Fig. 10E) displays moderate LREE enrichment typical of continental tholeiitic rocks (Basaltic Volcanism Study Project, 1981). The within-plate tectonic setting as shown by the Zr/Y versus Zr (Fig. 10D, *after* Pearce and Norry, 1979) diagram is consistent with a continental tholeiitic magmatic affinity. Thus the geochemical character indicates that the rocks are intracontinental tholeiitic rocks.

The rocks display a wide range in Ni and Cr suggesting that olivine and clinopyroxene represent fractionating phases. However wide variations in Zr/La (9–25), Zr/Y (3.5–11, Fig. 10D), and Zr/Nb (8–50) ratios indicate that crystal-liquid fractionation cannot account for all the trends observed and require derivation from a heterogeneous or enriched source (e.g. Pearce and Norry, 1979; Le Roex et al., 1983). These features are common in continental tholeiite and are attributed to the derivation of parental magmas from a heterogeneous upper mantle source (e.g. Erlank, 1986). Murphy (1988) postulated that the mantle source beneath the Antigonish Highlands may have been contaminated by dehydration of a subducting slab and it is possible that this contamination also contributed to the geochemical heterogeneities discussed above.

Field relationships indicate that the Folly River Formation unconformably overlies the Gamble Brook Formation, and probably postdates the ca. 605 ± 5 Ma deformation within the Gamble Brook Formation (Nance and Murphy, 1990). A minimum age for the Folly River Formation is given by the ca. 610 ± 5 Ma age of the Debert River pluton (U-Pb on zircon, Doig et al. (1991a, b)) which postkinematically intruded the formation. These data imply that the age of deposition and deformation of the Folly River Formation is tightly constrained to ca. 610–605 Ma. The close temporal relationship between deposition, deformation, and intrusion supports the interpretation that the formation was deposited in a strike-slip tectonic regime (Nance and Murphy, 1988, 1990; Pe-Piper and Murphy, 1989). The stratigraphy, geochemistry, and age of the Folly River Formation suggest that it may be a correlative of the Jeffers Group and Warwick Mountain Formation.

Mylonitic granite gneiss

Heterogeneously deformed granite bodies ranging in size from veins and dykes to narrow plutons (unit 5b; Fig. 1) occur along the ductile shear zone between the Gamble Brook Formation and the Great Village River Gneiss. These deformed granite bodies, excellently exposed on Rockland Brook, Great Village River, and Portapique River, consist of sericitized porphyroclasts of orthoclase, plagioclase, and quartz with rare biotite and muscovite, and contain a mylonitic foliation defined by ribbon quartz, biotite, and muscovite. Foliation surfaces display a strong linear fabric. These bodies are concordant with the ductile shear zone on a regional scale but locally, crosscutting, welded contacts with both Gamble Brook Formation and Great Village River Gneiss are preserved. The intensity of mylonitization in these granite gneiss bodies varies considerably over short distances. Larger bodies commonly crosscut the principal fabrics in the host rocks, yet show well developed C-S fabrics and asymmetric augen that are coplanar with the stronger fabric of their host rock. These relationships, coupled with available geochronological data, support the contention of Cullen (1984) that emplacement of the granite was broadly synkinematic with respect to deformation of gneiss and host rocks.

Previous geochronology (Table 2) yielded ages ranging from ca. 650–600 Ma for these rocks. A light grey hornblende and biotite-bearing mylonitic granite collected on a road outcrop adjacent to Great Village River is homogeneously deformed both internally and with respect to its host rock and its contacts with the host are oblique to the shear zone suggesting that it was intruded during the waning stages of movement along the shear zone. Gapais (1989) attributed this style of deformation to shearing during crystallization and cooling of the magma. The age of the deformed granite should therefore give a minimum age for the development of the shear zone.

There is a small but nonsystematic inherited component, such that the three data points roughly point to a zero-age lower intercept. Uranium-lead zircon analyses (Doig et al., 1991a; sample B, Fig. 5A, C, Table 4) indicate that these deformed granitic rocks were emplaced at 605 ± 5 Ma. A similar U-Pb age was obtained for deformed granite in the same shear zone near Rockland Brook (J.D. Keppie, pers. comm., 1992). These ages imply a genetic relationship with the slightly younger Great Village River Gneiss.

Great Village River Gneiss

The generally east-trending Great Village River Gneiss, well exposed in Portapique River, Great Village River, and Rockland Brook, is truncated to the north by the Rockland Brook Fault and to the south by a ductile shear zone separating it from the structurally overlying Gamble Brook Formation. The gneiss (unit 5a, Fig. 1) consists of massive to layered quartz-plagioclase, hornblende amphibolite (Bass River Amphibolite of Cullen (1984)), hornblende-bearing granitoid orthogneiss (Great Village River Orthogneiss of Cullen (1984)), and biotite±garnet-rich, psammitic paragneiss (Donohoe and Wallace, 1985). Cullen (1984) included the pelitic "Portapique River Schist" which locally contains sillimanite and staurolite (Cullen, 1984); however, on the basis of lithology and spatial association, we consider this unit to be Gamble Brook Formation that was contact metamorphosed by adjacent mylonitic granite along the shear zone separating the Great Village River Gneiss and the Gamble Brook Formation. In some localities, amphibolite dykes crosscut fabrics in the gneiss, but in others contacts are transposed by a strong mylonitic fabric. All lithologies of the Great Village River Gneiss display several generations of small-scale folds all of which fold the dominant metamorphic foliation.

Two samples from the Great Village River Gneiss were dated by the U-Pb zircon method, namely a foliated amphibolite (C, Fig. 5A, the only one of three samples which yielded zircon), and a granodioritic orthogneiss (D, Fig. 5A). The amphibolite sample had few zircons, and most were highly coloured crystals in the zero-degree magnetic fraction. A second sample of 90 μg was picked indiscriminately from the more magnetic fractions. Both gave essentially the same $^{207}Pb/^{206}Pb$ dates of 586 Ma and 587 Ma (Table 4A, Fig. 5C) so that inheritance is unlikely, as is expected from a rock with a likely mafic to intermediate igneous origin. Three of the different magnetic fractions from the orthogneiss, a medium grey hornblende-bearing rock, gave identical $^{207}Pb/^{206}Pb$ dates of about 580 Ma (Table 3A, Fig. 5C) so that this is almost certainly the age of crystallization. The inherited component of the other fraction is small, and equivalent to 5 Ma. The important point is that the age of the orthogneiss is no more than 580 Ma. Uranium-lead (zircon) analyses therefore indicate that orthogneiss and amphibolite of the Great Village River Gneiss were emplaced at ca. 590–580 Ma. The age of the host paragneiss is unknown.

Jeffers Group

The Jeffers Group forms an east-trending belt north of the Rockland Brook Fault which outcrops within the map area in the headwaters of East River of Five Islands and River Philip. On the basis of more detailed studies to the west, both southern and northern contacts of the belt are faulted, generally against the Silurian Wilson Brook Formation (Pe-Piper and Piper, 1987, 1989). Many of the unfossiliferous clastic rocks assigned by Donohoe and Wallace (1982a, b, c, d) to the Wilson Brook Formation, including those on East River of Five Islands, were reassigned to the Jeffers Group, on the basis of conformable relationships with typical Jeffers Group lithologies (Pe-Piper and Piper, 1987).

The Jeffers Group consists of a basal Gilbert Hills Formation (basalt, basaltic andesite, rhyodacite, and rhyolite), passing gradationally upward into felsic volcanogenic turbidite of the Humming Brook Formation, overlain in turn by a thick sequence of turbiditic lithic arkose (Cranberry Lake Formation). Outcrops in the headwaters of East River of Five Islands belong to the Cranberry Lake Formation, whereas those in the headwaters of River Philip belong to the Gilbert Hills Formation and possibly the Humming Brook Formation. A granodiorite similar to the Jeffers Brook pluton (Pe-Piper, 1987) also occurs in the latter area. The Jeffers Group is characterized by a penetrative flat-lying cleavage of uncertain age.

Geochemistry of the Jeffers Group is important to understanding the tectonic setting of volcanic successions in the Cobequid Highlands. Pe-Piper and Piper (1987) have shown that volcanic rocks predominantly occur within the Gilbert Brook Formation of the Jeffers Group which is subdivided into two facies, the Lakeland facies of voluminous felsic flows and tuffs with interlayered high Ti-rich basalt flows and associated dykes, and the Harrington River facies of low-Ti calc-alkalic basaltic andesite to rhyodacite. Pe-Piper and Piper (1989) deduced magmatism to be related to rifting within a volcanic arc, with the two facies related by fractional crystallization.

The SiO_2 versus Zr/TiO_2 diagram (Fig. 11A) clearly displays the subalkaline traits of Jeffers Group magmatism and the range in chemistry from basalt to rhyolite. The FeO_T versus FeO_T/MgO and Ti/100- Zr–Y.3 (Fig. 11B, C) diagrams show that the mafic rocks can be subdivided into two distinct chemical types, one with relatively high FeO_T and high TiO_2 plots in the within-plate field, whereas the other has lower FeO_T and TiO_2 and plots in the calc-alkaline field. The Jeffers Group has three distinct magma types: Fe-Ti-rich within-plate tholeiitic basalt, calc-alkalic basaltic andesite, and felsic volcanic rocks with arc signatures, remarkably similar to the chemical signature of the Keppoch Formation in the Antigonish Highlands. The geochemistry of the Fe-Ti-rich mafic rocks is indistinguishable from that of the Folly River Formation (Fig. 9, 11).

Volcanic rocks of the Cobequid Highlands can be precisely correlated with the better exposed and less fault-dissected Georgeville Group of the Antigonish Highlands. In the southern Antigonish Highlands, the Keppoch Formation of interlayered continental tholeiite, calc-alkalic basaltic andesite, and felsic volcanic rocks with volcanic arc affinities is conformably to unconformably overlain by thick turbidite and minor continental tholeiite in the central highlands (Clydesdale Formation) which are probably laterally equivalent to calc-alkalic basaltic andesite of the Chisholm Brook Formation in the northern Antigonish Highlands. The entire Georgeville Group has been interpreted to represent limited ensialic rifting within a volcanic-arc setting (Murphy and Keppie, 1987; Murphy et al., 1990). The environment envisaged is comparable to the narrow ensialic intra-arc basin in the southern Andes (Saunders and Tarney, 1984) where volcanic rocks with both calc-alkalic and tholeiitic affinities developed.

The Jeffers Group (Pe-Piper and Piper, 1989) and Dalhousie Mountain Formation (Murphy et al., 1988) lithologically and geochemically resemble the Keppoch Formation, whereas the Folly River Formation resembles the Clydesdale Formation of the Antigonish Highlands. Detailed correlations between the Antigonish and Cobequid highlands therefore suggest that late Precambrian volcanism in the Cobequid Highlands also represents limited rifting within a volcanic-arc setting.

A minimum age for the Jeffers Group is provided by the age of the Jeffers Brook pluton, interpreted to intrude deformed Jeffers Group (Donohoe and Wallace, 1985; Pe-Piper and Piper, 1987), although all known contacts of the Jeffers Brook pluton are faulted. However xenoliths within the pluton include metasedimentary rocks, mafic igneous rocks, and rhyolite similar to the Jeffers Group. The pluton gave $^{40}Ar/^{39}Ar$ dates on hornblende of 607.1 ± 3.4 Ma and 604.5 ± 4.4 Ma (Keppie et al., 1990). The Jeffers Group strongly resembles, and is correlated with, the late Precambrian rocks of the Keppoch Formation in the Antigonish Highlands which has a well constrained ca. 615–610 Ma depositional age (Murphy et al., 1991). The relationship of these rocks with Precambrian rocks to the south of Rockland Brook Fault is not clear. Correlation of the Jeffers Group with the Folly River, Warwick Mountain, and Dalhousie Mountain formations is based on lithological comparison with each other and with the rocks of the Antigonish Highlands and is consistent with available geochronological data, but no definitive field relationships are exposed.

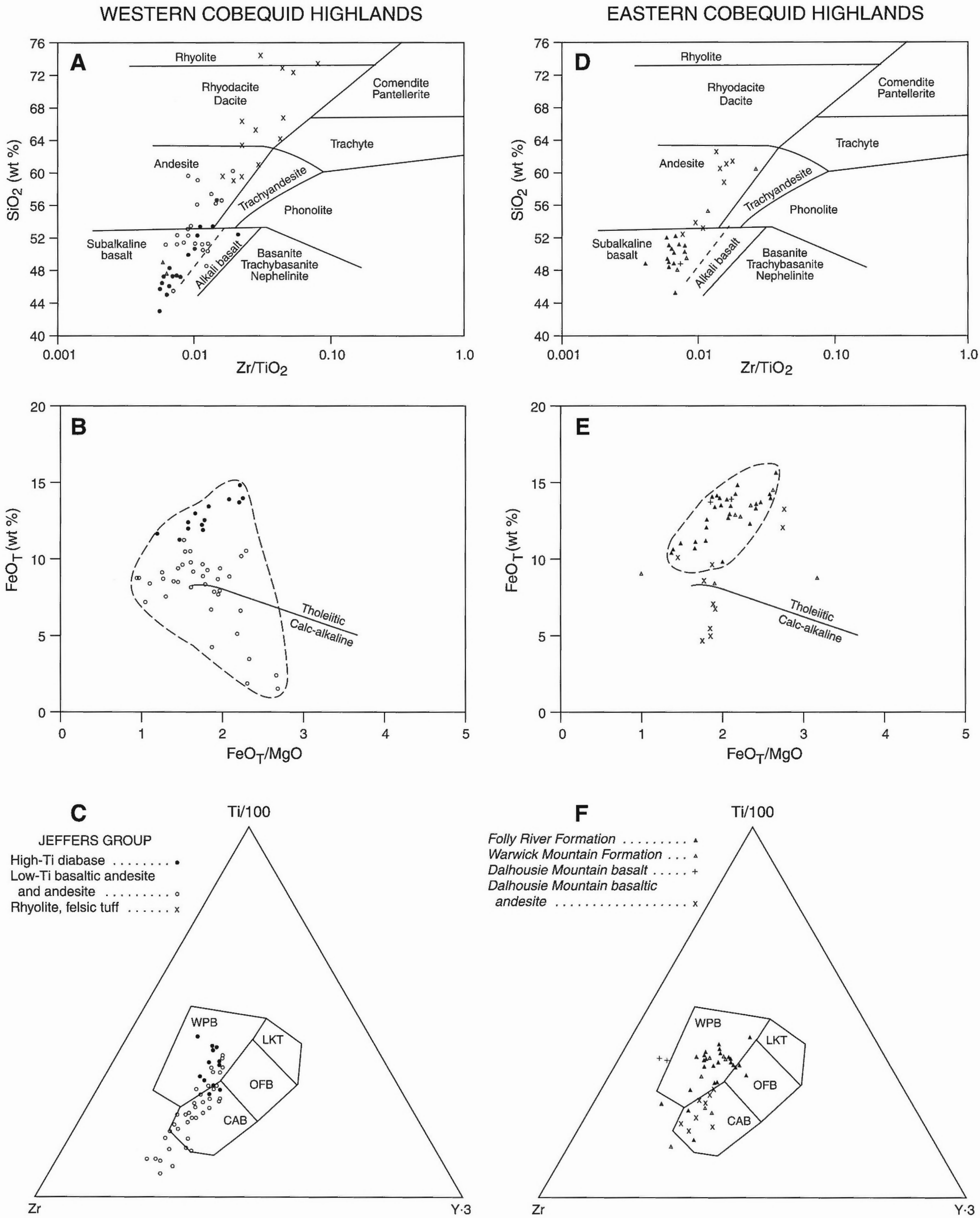


Figure 11. *Geochemical character of the Jeffers Group and Folly River, Dalhousie Mountain, and Warwick Mountain formations. Outlying data points are thought to be altered (data* after *Pe-Piper and Piper, 1989).* ***A), D)*** *SiO_2 versus Zr/TiO_2 (data* after *Floyd and Winchester, 1975);* ***B), E)*** *FeO_T/MgO versus FeO_T (boundary* after *Miyashiro, 1974);* ***C), F)*** *Ti-Y-Zr (fields* after *Pearce and Cann, 1973), WPB = within-plate basalt, CAB = calc-alkalic basalt, MORB = mid-ocean ridge basalt, LKT = low-K tholeiite.*

Warwick Mountain Formation

The Warwick Mountain Formation, defined by Donohoe and Wallace (1980, 1982a, b, c, d) with a type section along French River, east of Warwick Mountain (NTS 11 E/11), is also well exposed along Miller Brook (NTS 11 E/11). Its structural thickness is about 400 m, but its stratigraphic thickness is unknown. The formation is unconformably overlain to the north by the Late Carboniferous Boss Point Formation and is in faulted contact to the east and west with Silurian and Devonian–Carboniferous rocks. The southern contact is a thrust, where the Warwick Mountain Formation structurally overlies the Devonian–Carboniferous Fountain Lake Group.

The Warwick Mountain Formation consists of interlayered felsic and mafic volcanic rocks, and turbiditic mudstone to lithic arkose. The formation characteristically displays a penetrative flat-lying cleavage. The lithology and structure and strongly resemble those of the Jeffers Group (Donohoe and Wallace, 1982a, b, c, d). The age of the Warwick Mountain Formation is unknown. On the basis of field relationships it could be as young as Carboniferous, but a late Precambrian age is favoured because of lithological and geochemical similarities with the Jeffers Group (Fig. 11D, E, F).

Dalhousie Mountain Formation (new)

The Dalhousie Mountain Formation, defined here as a sequence of interlayered felsic to mafic volcanic rocks and abundant tuffaceous turbidite, occurs only in the easternmost Cobequid Highlands, bounded to the north by the Rockland Brook Fault, to the south by the Late Devonian Salmon River pluton (which intruded it), and to the east by the unconformably overlying Late Carboniferous Boss Point Formation. Neither the base nor the top of this succession is exposed. The formation is well exposed in tributaries of Eight Mile Brook which drain from Dalhousie Mountain.

The Dalhousie Mountain Formation consists of felsic ignimbrites, tuffs, and flows with a significant proportion of andesitic tuffaceous material, and basalt and basaltic andesite distinguished from one another by the absence or presence of plagioclase phenocrysts. The volcanic rocks are characteristically massive and lack a regional penetrative fabric. Interlayered sedimentary rocks are thin, fine-grained, pale to dark green turbiditic mudstone units displaying plane lamination, local grading, and crossbedding. Individual units attain a thickness of up to 2 m. They consist of subrounded quartz, albite, abundant sericite and chlorite, and minor pyrite. Sericite and chlorite define a weak cleavage, generally subparallel to bedding. The formation was deposited in a submarine environment (turbidite sequences) probably in a low-energy environment, as suggested by the fine, planar laminations, possibly at the distal end of a submarine fan (Walker, 1984). The presence of a weakly developed cleavage, defined by sericite and chlorite, indicates moderate shortening and lower greenschist-facies metamorphism.

Donohoe and Wallace (1982a, b, c, d) included these rocks in the Silurian Earltown Formation on the basis of the abundance and style of the interlayered sedimentary rocks which they correlated with the Silurian Arisaig Group of the Antigonish Highlands, and contrasted with known late Precambrian successions. However, the thin, plane-laminated bedding, the chloritic argillaceous composition, and the weak bedding-parallel cleavage in the mudstone are typical of sedimentary rocks of the late Precambrian Keppoch Formation in the Antigonish Highlands and contrast with unmetamorphosed, uncleaved, arenaceous, very fossiliferous Silurian sedimentary rocks. Murphy et al. (1988) therefore reassigned these rocks to the Late Precambrian and informally termed them the Dalhousie Mountain volcanic unit. This assignment is supported by recent geochronological data from plutons that intruded these rocks, and which were emplaced at 605 ± 5 Ma (R. Doig, pers. comm. (1993), U-Pb on zircon). Given the better age control, the Dalhousie Mountain volcanic unit is elevated to formation status.

Precambrian plutons

Precambrian plutonic rocks not part of the Bass River Complex include the appinitic diorite Frog Lake pluton (Hubley, 1987), and the granodioritic-granitic Debert River and McCallum Settlement plutons. Other small plutons are correlated with these plutons on the basis of lithological comparison. In general, Late Precambrian granitic rocks are medium- to coarse-grained, equigranular with subhedral granitic texture. Major minerals are quartz, plagioclase (An_{30-45}), orthoclase or microcline, with minor biotite (commonly replaced by chlorite), opaque oxides, apatite, and accessory allanite and zircon. Sericite and epidote occur as secondary minerals. The granodiorite bodies are medium-grained equigranular rocks and consist of plagioclase, biotite, hornblende, quartz, and minor perthite, apatite, titanite, opaque oxide minerals, and actinolite. Representative modal analyses are given in Table 7.

Frog Lake pluton

The Frog Lake pluton forms an elongate body of about 20 km^2 and occurs about 5 km north of Debert (Donohoe and Wallace, 1982a, b, c, d). The Londonderry and Cobequid faults merge to form a ductile shear zone near its southern boundary. The pluton is best exposed on Debert River, Chiganois River, and in a quarry and isolated outcrops along logging roads north of Belmont that head towards the dam on Debert River. The pluton intruded the Gamble Brook Formation, contains xenoliths and/or pendants of Gamble Brook lithologies, and was intruded by the Debert River pluton. The Frog Lake pluton consists of two predominant suites, a predominantly dioritic mafic suite and a granodioritic-granite suite. The mafic suite (Hubley, 1987) is dominated by plagioclase and amphibole which constitute about 95 modal per cent of the rock, with interstitial biotite. Most specimens are equigranular and medium grained. Locally, the pluton exhibits appinitic tendencies and contains highly variable textures including coarse-grained hornblende pegmatite. Plagioclase is oligoclase-andesine. Quartz and alkali feldspar occur locally. This diorite strongly resembles the Late Precambrian Jeffers Brook diorite in the western Cobequid Highlands.

Table 7. Modal mineralogy of representative sample from Neoproterozoic granitoid plutons.

	Debert River pluton						Frog Lake pluton					McCallum Settlement pluton				
Sample number	**3104**	**4521**	**4513**	**3185**	**4524**	**4547**	**2812**	**2943**	**3330**	**2815**	**2941**	**3118**	**3136**	**3171**	**4090**	**4091**
Quartz	30.5	27.1	12.1	33.6	32.3	24.8	18.7	32.9	17.1	31.8	33.3	35.9	40.0	30.5	31.2	33.7
K–feldspar	7.3	16.2	–	36.6	25.2	57.6	17.1	9	1.2	27.0	38.4	35.4	20.0	34.3	17.71	26.0
Plagioclase	53.6	48.0	40.4	26.5	32.8	16.3	54.0	48.4	59.7	35.0	26.6	25.6	34.5	29.5	43.5	35.7
Hornblende	–	1.4	43.3	–	4.8	–	–	–	5.6	–	–	–	–	–	–	–
Biotite	5.1	2.3	–	1.7	1.3	–	9.2	8.1	15.0	2.3	0.8	1.9	3.5	3.2	0.7	–
Opaque minerals	0.5	1.4	1.3	0.6	1.6	0.6	–	0.7	–	0.2	0.3	–	1.0	0.5	0.1	0.9
Apatite	0.2	–	–	–	–	–	–	0.1	0.3	–	–	–	0.1	–	0.1	–
Chlorite	0.5	2.2	0.4	0.5	0.8	0.1	–	0.4	0.4	–	0.3	0.5	0.5	1.0	5.8	2.7
Epidote	2.4	1.2	2.5	0.5	1.0	0.5	1.0	0.5	0.8	3.7	0.4	0.5	0.3	1.0	0.7	1.0
Titanite	–	–	–	0.1	–	0.1	–	–	–	–	–	–	–	–	–	–
Accessory minerals	–	0.2	–	–	0.2	0.3	–	–	–	–	–	–	0.2	–	0.2	–
Rock type	Grano-diorite	Grano-diorite	Tonalite	Granite	Granite	Granite	Grano-diorite	Grano-diorite	Tonalite	Granite	Granite	Granite	Granite	Granite	Grano-diorite	Granite

– = no data

Granitoid rocks of the Frog Lake pluton occur as stocks, veins, and dykes which occur only within the pluton, where they intruded the mafic suite. Contacts between mafic rocks and granitoid rocks are sharp with no chilled margins. No evidence of mixing was observed. The granitoid rocks are medium- to coarse-grained, reddish-pink to whitish-red with subhedral grains of pink plagioclase and white alkali feldspar as the dominant visible constituents. Minor amounts of biotite (up to 5%) and quartz can be distinguished in hand specimen and accessory amounts of apatite and opaque minerals are interstitially distributed. Secondary minerals include chlorite, epidote, and sericite. Granodiorite, distinguished from granite by presence of significant hornblende and biotite, and minor tonalite, consisting of 20% (biotite+hornblende), 60% plagioclase, and approximately 15% quartz with accessory titanite and apatite also occur. In thin section, the plagioclase (An>30) forms subhedral laths that have been severely altered to sericite and muscovite. Twin lamellae display kinking. Primary quartz (large, anhedral, and strained) is distinct from secondary quartz (fine-grained, rhombohedral, and filling grain boundaries); interstitial alkali feldspar displays perthitic texture. Biotite occurs as subhedral laths, partially replaced by chlorite. Subhedral apatite grains form inclusions within biotite.

Geochronology (^{40}Ar-^{39}Ar hornblende) of the Frog Lake pluton yielded an age of 622.1 ± 3.3 Ma (Keppie et al., 1990).

Debert River pluton

The Debert River pluton, as originally mapped by Donohoe and Wallace (1982a, b, c, d), included a Late Precambrian part (to the south) and a Devonian–Carboniferous part (to the northeast) which intruded the Nuttby Formation (Murphy et al., 1988). The term Debert River pluton is retained for the Late Precambrian rocks.

Two varieties of granitic rocks yielding indistinguishable U-Pb ages occur within the Debert River pluton. A medium- to coarse-grained equigranular red granite with few visible mafic constituents outcrops on the upper Chiganois River and near the northern contact along Debert River. The granite is sheared near the contact of numerous diabase dykes. The granitic rocks are subhedral, medium-grained, and consist of plagioclase, alkali feldspar, quartz, minor amounts of biotite, accessory minerals such as titanite, and opaque minerals; and secondary muscovite, epidote, and chlorite. A greenish-grey, equigranular granodiorite containing varying amounts of mafic minerals occurs near the contact between the Frog Lake and Debert River plutons and is best exposed along Debert River where xenoliths of diorite (attributed to the Frog Lake pluton) and mylonitic quartzite (attributed to the Gamble Brook Formation) occur. This equigranular, medium- to coarse-grained granodiorite consists of interlocking anhedral grains of alkali feldspar, plagioclase, and quartz with interstitial biotite, muscovite, and opaque minerals. Apatite grains appear as inclusions within the biotite grains and secondary chlorite and epidote are common. Albite twinning in the plagioclase has been obliterated by alteration and perthitic textures in the alkali feldspar have a waxy appearance.

Uranium-lead (zircon) data from the two distinct lithologies gave ages of 609 ± 4 Ma and 611 ± 4 Ma respectively (Table 4A, Fig. 5C). Zircon fractions nearly all contain an inherited component, but one fraction (E-1) consisting of the largest crystals is concordant at 609 Ma. The fraction of clear, thin needles of zircon (E-2) is nearly concordant and yields a date that is only slightly greater (611 Ma). The Rb-Sr date is 596 ± 70 Ma (Donohoe et al., 1986).

McCallum Settlement pluton

The McCallum Settlement pluton consists mainly of granite with some aplite and rare diorite, all well exposed on the West Branch North River. At its northern contact, the pluton post-tectonically intruded the Folly River Formation. The western and eastern contacts are not exposed. The southern contact is the Cobequid Fault. The pluton is intruded by diabase dykes of unknown age. The granite is coarse-grained with perthitic feldspar in which plagioclase abundance varies inversely with modal K-feldspar. Minor biotite and muscovite occur with accessory opaque minerals, rutile, titanite, and secondary epidote, chlorite, and sericite. Aplite displays micrographic texture with plagioclase highly altered to sericite. Quartz is highly variable in abundance and occurs as isolated anhedral crystals and within K-feldspar crystals. Diorite is medium grained, and contains sutured quartz, saussuritized plagioclase, minor nonperthitic K-feldspar, and abundant chloritized brown biotite (containing epidote, zircon, apatite, and opaque minerals as inclusions). Laths of actinolite are sparsely distributed as a secondary phase throughout the samples.

The age of the McCallum Settlement pluton is somewhat problematic because the $^{207}Pb/^{206}Pb$ dates (Table 2) are all different and the data are the most discordant of all the samples analyzed. The maximum age of emplacement is therefore interpreted to be the youngest $^{207}Pb/^{206}Pb$ date of 580 Ma, which is very similar to the 575 ± 22 Ma Rb-Sr date of Gaudette et al. (1984).

Geochemistry

A representative suite of samples of fine-grained granitoid rocks from the Debert River and Frog Lake plutons have been analyzed for major and selected trace elements (Table 8).

These rocks are peraluminous to metaluminous and most samples contain significant normative corundum. Relative to Devonian–Carboniferous granitic rocks (*see* Table 11 *in* "Paleozoic rocks" section) of the eastern Cobequid Highlands, they contain lower average concentrations of SiO_2, K_2O, Rb, Rb/Sr, Y, Nb, and Th and higher average concentrations of TiO_2, Al_2O_3, Fe_2O_3, MgO, CaO, P_2O_5, Sr, Ba, Zn, V, and Cr. These distinctions may be useful in combination with petrography in mapping areas for which geochronological data may not be available. The relatively high Sr and low Rb appear to be particularly distinctive.

Most of the data display magmatic trends (Fig. 12) typical of volcanic-arc granitic rocks (Fig. 13). The geochemical data are consistent with late Precambrian arc-related granitoid plutonism typical of the Avalon Terrane (e.g. Krogh et al., 1988). Recent age data indicate that this magmatism within the terrane ranges from 630 Ma to 570 Ma, but it is uncertain whether magmatism was continuous or episodic. However,

Table 8. Geochemistry of selected granitoid rocks from Neoproterozoic plutons.

	Debert River pluton							Frog Lake pluton		
Sample number	**3171**	**3104**	**2943**	**4521**	**4524**	**4525**	**4533**	**4496**	**4490**	**4491**
SiO_2 (wt %)	71.10	64.79	67.90	66.80	73.58	65.99	69.08	67.39	70.75	74.47
TiO_2 (wt %)	0.36	0.85	0.61	0.68	0.36	0.64	0.49	0.43	0.35	0.24
Al_2O_3 (wt %)	14.90	16.65	15.31	15.27	13.55	16.02	15.00	15.93	14.97	13.22
Fe_2O_3(t) (wt %)	3.15	5.37	4.05	5.00	2.77	4.54	3.57	3.91	2.99	1.64
MnO (wt %)	0.08	0.09	0.12	0.09	0.08	0.15	0.08	0.08	0.08	0.04
MgO (wt %)	0.94	2.16	1.46	1.66	1.19	1.55	1.39	1.66	1.29	1.08
CaO (wt %)	1.91	2.54	2.34	1.71	1.83	3.35	1.92	2.71	1.67	0.71
Na_2O (wt %)	3.81	4.11	4.29	3.92	3.83	4.34	4.04	3.77	3.98	4.18
K_2O (wt %)	3.04	2.01	2.94	3.41	2.28	2.10	2.99	2.58	2.62	4.17
P_2O_5 (wt %)	0.10	0.09	0.16	0.15	0.07	0.18	0.12	0.16	0.08	0.07
LOI (wt %)	0.80	1.30	0.60	1.20	0.60	0.60	1.10	1.10	0.80	0.50
Total	100.27	99.96	99.79	99.89	100.14	99.46	99.78	99.72	99.58	100.32
	813	321	1169	1388	785	987	949	800	565	655
Ba (ppm)										
Rb (ppm)	76	72	85	103	68	58	73	97	91	123
Sr (ppm)	259	273	298	319	264	336	319	419	324	157
Y (ppm)	19	50	32	23	21	43	29	12	22	23
Zr (ppm)	192	255	299	339	206	312	261	188	159	150
Nb (ppm)	12	17	21	13	11	18	13	12	27	20
Th (ppm)	<10	<10	<10	<10	<10	<10	<10	<10	<10	34
Pb (ppm)	11	12	<10	<10	<10	<10	<10	<10	32	69
Ga (ppm)	22	24	22	17	15	21	17	16	17	13
Zn (ppm)	39	56	58	44	36	64	53	43	84	72
Cu (ppm)	<5	<5	13	<5	<5	<5	<5	<5	<5	<5
Ni (ppm)	9	26	18	8	<10	<10	9	8	<10	8
V (ppm)	25	61	39	38	23	35	29	34	25	13
Cr (ppm)	3	13	<5	18	49	28	24	30	39	26
La (ppm)	49.7	43.8	38.6	–	–	–	–	–	–	–
Ce (ppm)	106.0	74.0	66.0	–	–	–	–	–	–	–
Nd (ppm)	38.0	29.0	29.0	–	–	–	–	–	–	–
Sm (ppm)	6.2	7.0	7.5	–	–	–	–	–	–	–
Eu (ppm)	1.55	1.24	1.06	–	–	–	–	–	–	–
Yb (ppm)	2.32	5.95	3.07	–	–	–	–	–	–	–
Lu (ppm)	0.35	0.96	0.57	–	–	–	–	–	–	–

LOI = loss on ignition
– = no data

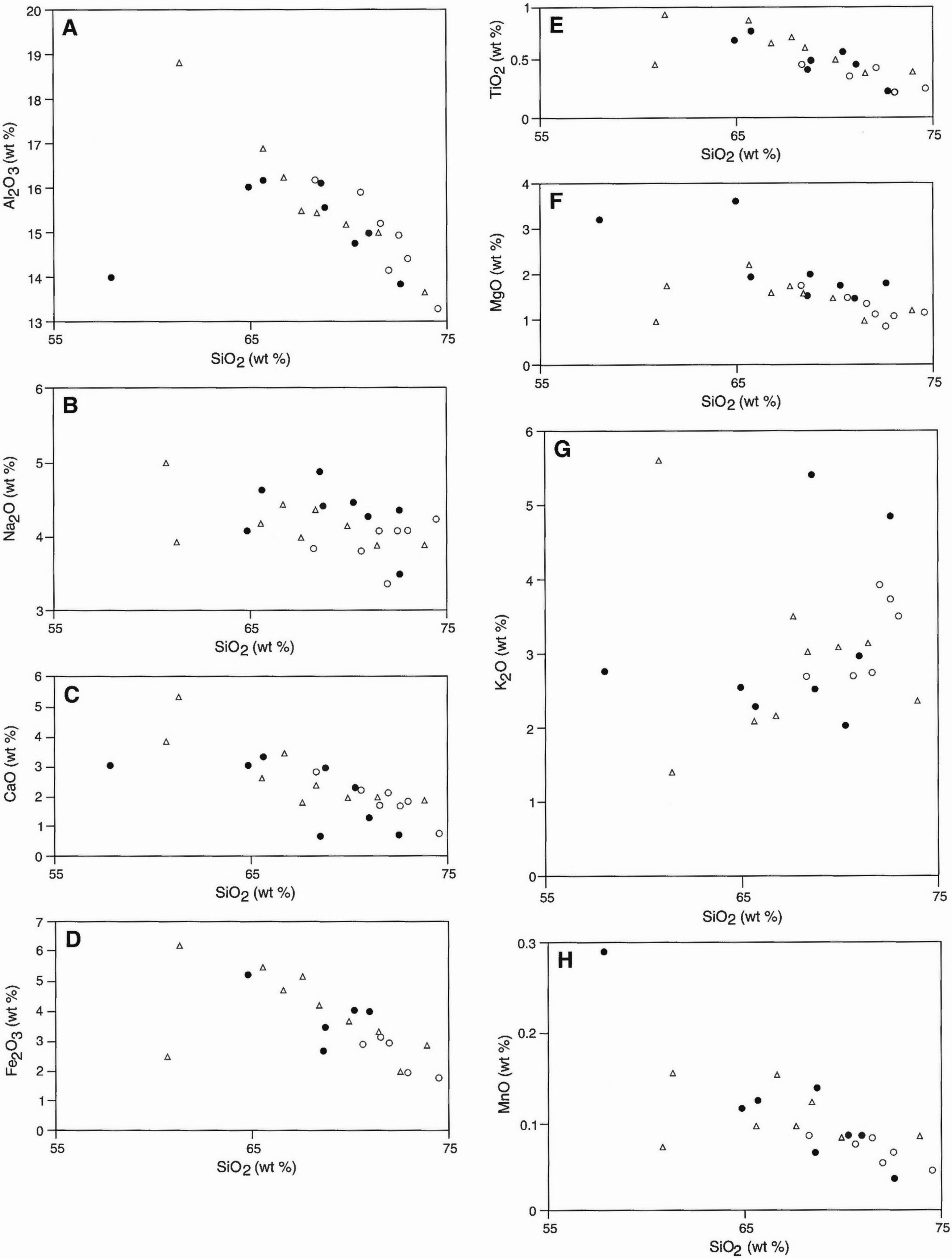
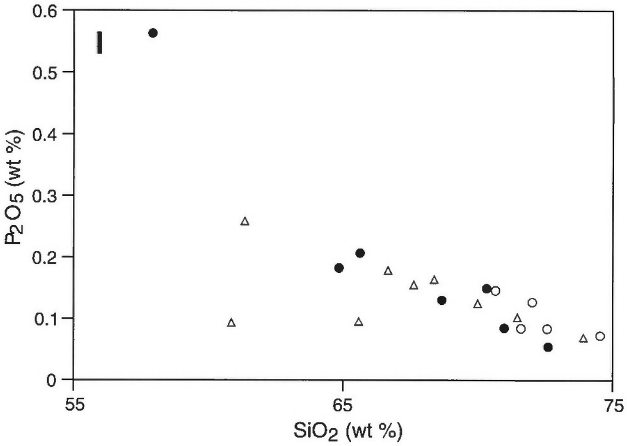
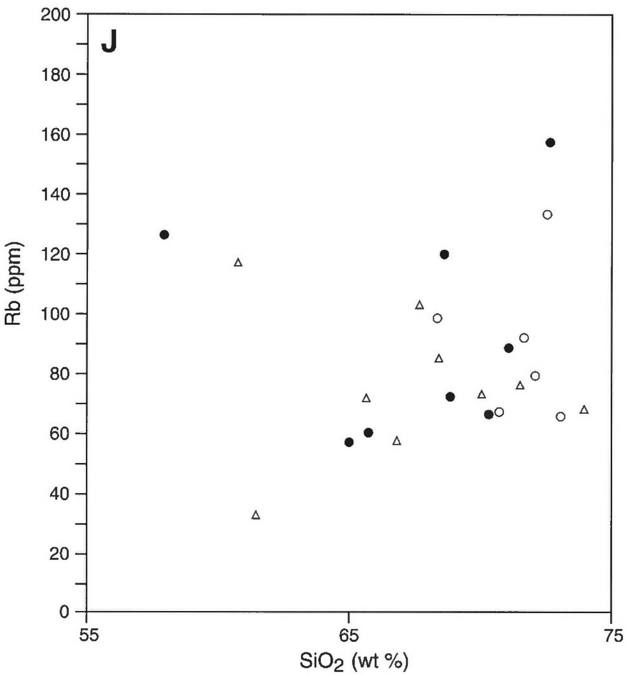
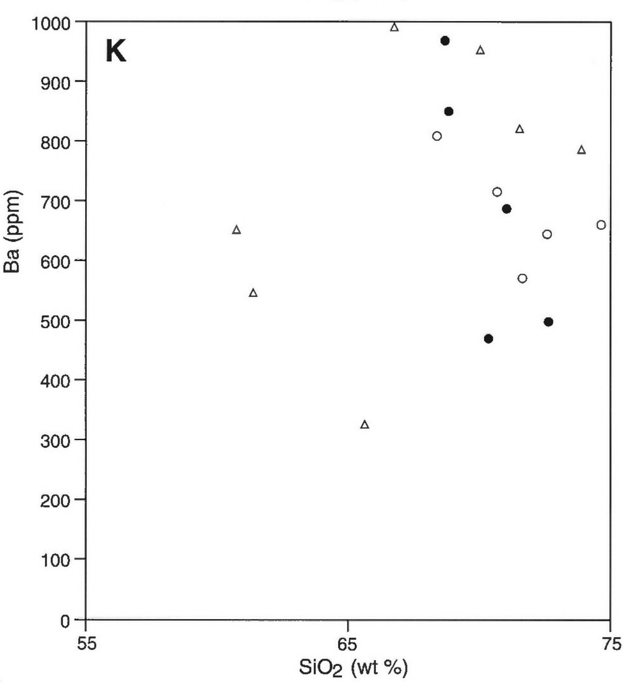
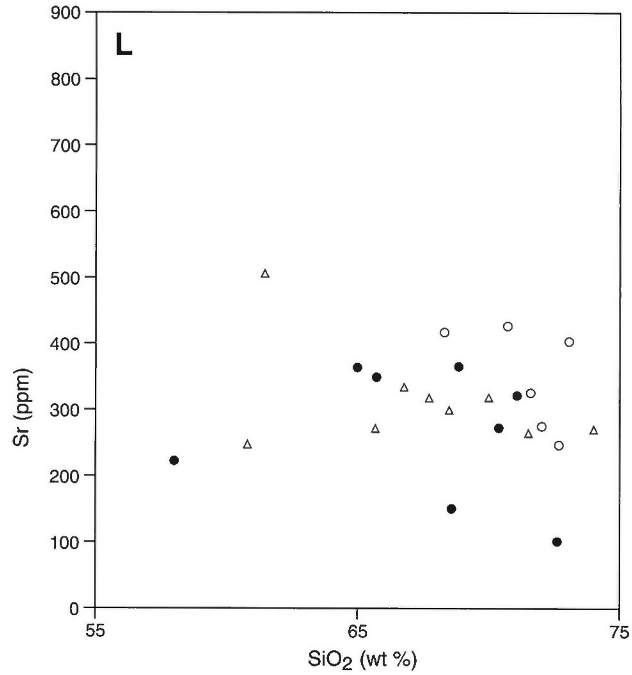


Figure 12. *Major and trace element variations with SiO_2 for the Neoproterozoic granitoid rocks.*









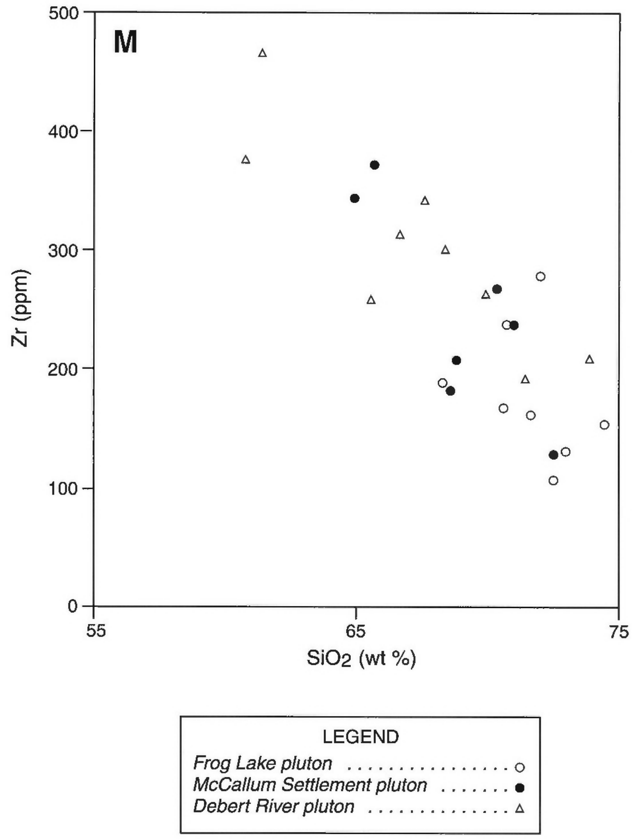


Figure 12 *(cont.).*

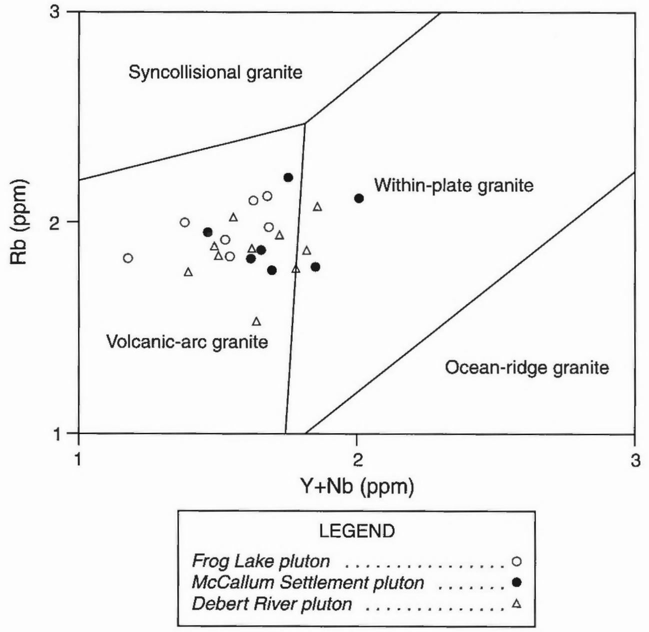


Figure 13. *Plot of Rb versus Y+Nb for the Neoproterozoic granitoid rocks. Discrimination fields* after *Pearce et al. (1984).*

the range in ages is similar to that of the Great Village River Gneiss. The late Precambrian plutons may be less deformed and metamorphosed equivalents of the Great Village River Gneiss. The chemistry of the plutons lends support to regional tectonic interpretations of late Precambrian arc-related magmatism in the Avalon Terrane of Nova Scotia (e.g. Pe-Piper and Piper, 1989; Murphy et al., 1990; Keppie et al., 1991). Furthermore, it demonstrates the Avalonian affinity of the host rocks intruded by these plutons, including the entire Bass River Complex.

PALEOZOIC ROCKS

Stratigraphy

The Cambrian–early Ordovician overstep sequence diagnostic of the Avalon composite terrane is not present in the Cobequid Highlands. Late Ordovician to early Devonian rocks are sparsely represented in thrust slices south of Rockland Brook Fault (unit 6, Fig. 1), and are more widespread north of the fault (Wilson Brook Formation). Undivided Ordovician–early Devonian strata south of the fault consist of fossiliferous grey and green siltstone and shale with an Ordovician or Silurian (Donohoe and Wallace, 1982a, b, c, d) Rhenish–Bohemian fauna (Boucot, 1975). Detailed comparison with rocks of a similar age to the north of the Rockland Brook Fault is hindered by poor exposure.

The Wilson Brook Formation, defined by Donohoe and Wallace (1980, 1982a, b, c, d) with a type section on Portapique River north and south of its confluence with Wilson Brook, has a measured thickness of 1850 m (Donohoe and Wallace, 1985). On Portapique River the northern boundary of the Wilson Brook Formation is conformable with the Early–Middle Devonian Portapique River Formation. Differences in structural style suggest that the unit is overlain unconformably by the mid-Devonian to early Carboniferous Fountain Lake Group, although the actual contacts appear to be faulted. The Wilson Brook Formation is dominated by green, grey, and black micaceous siltstone, wacke, and minor tuffs. Thin coquina beds are locally present in the upper part of the formation (Donohoe and Wallace, 1980). In general the sequence coarsens upward and deposition is thought to have taken place in a storm-dominated, shallow-marine environment (Donohoe and Wallace, 1985). Faunas range from Llandovery to Early Gedinnian. These rocks resemble in lithology, fauna, and depositional environment the Llandovery to Early Gedinnian Arisaig Group, also deposited in a storm-dominated, shallow-marine environment (Cant, 1980).

The "Earltown Volcanics" as mapped by Donohoe and Wallace (1982a, b, c, d) is now believed to be essentially absent. Most rocks formerly assigned to the Earltown Formation have been reassigned to the Late Precambrian Dalhousie Mountain Formation, whereas an area of volcanic rocks near Wyvern, tentatively assigned a Silurian age by Murphy et al. (1988), is now considered to be part of the Devonian–Carboniferous Fountain Lake Group. Although the structural complexity is such that the presence of Silurian volcanic rocks cannot be excluded without detailed geochronological studies, the name Earltown Formation is abandoned and all Silurian sedimentary rocks north of Rockland Brook Fault are assigned to the Wilson Brook Formation.

The Portapique River Formation which conformably overlies the Wilson Brook Formation on Portapique River, consists of dark grey, green, and red clastic rocks. No fossils have been found to date. Based on field relationships, the age of the formation is bracketed between the underlying Wilson Brook Formation (Gedinnian) and overlying Emsian–Tournaisian volcanic rocks. The relationship of the Wilson Brook and Portapique River formations to the Murphy Brook Formation (unit 8, Fig. 1) is not understood. The Murphy Brook Formation consists of ca. 200 m of interbedded red and grey massive conglomerate beds which contain abundant clasts of rhyolite and granite and grey to dark grey siltstone with plant fossils of Emsian to Eifelian age (Forbes et al., 1979). The Portapique River and Murphy Brook formations may be correlative with the Early Devonian Knoydart Formation in the Antigonish Highlands.

The contact between the Fountain Lake Group (unit 9, Fig. 1) and the Murphy Brook Formation is not exposed, although differences in structural style suggest an unconformity. The Fountain Lake Group, which outcrops extensively north of the Rockland Brook Fault, consists of a thick succession of interlayered felsic ignimbrites and flows, mafic volcanic rocks, and minor interbedded clastic sedimentary rocks with abundant associated dykes and sills (Chatterjee, 1984). The Byers Brook Formation consists of felsic volcanic rocks, predominantly bedded airfall tuff (locally fine agglomerate) and volcaniclastic sandstone, which achieve stratigraphic thicknesses of several hundred metres in upper Bulmer Brook, Arsenic Brook, and lower Bulmer Brook, and basalt flows locally interbedded with sedimentary rocks. The rocks are steeply dipping, locally overturned, but are not cleaved. Minor rhyolite and basalt associated with the sequence may be either lavas or fine-grained hypabyssal rocks. Rather rubbly weathering and amygdales probably indicate lava. Rarely, spherulitic rhyolite is found. The Diamond Brook Formation (principally basalt) contains Emsian–Eifelian and Tournaisian spores (Donohoe and Wallace, 1982a, b, c, d) whereas felsic volcanic rocks yielded a Rb/Sr isochron of ca. 387 Ma (Pe-Piper et al., 1989). All of the rock units of the Fountain Lake Group of the eastern Cobequid Highlands are cut by gabbro and granite, whereas in the western Cobequid Highlands, the Fountain Lake Group volcanic rocks appear broadly synchronous with intrusion of plutons and no intrusive or unconformable relationships have been found between plutons and the Fountain Lake Group.

Our mapping of the Fountain Lake Group suggests that many of the contacts mapped by Donohoe and Wallace (1982a, b, c, d), and used as the basis of their chronology (Donohoe and Wallace, 1985), are tectonic. It is not possible to demonstrate an unconformable relationship between the Fountain Lake Group and the Murphy Brook Formation north of Economy Lake. The regional setting makes a fault contact more likely. The supposed intrusive contact of the Folly Lake pluton with the Fountain Lake Group in the upper Portapique River is a thrust fault. Donohoe and Wallace (1982a, b, c, d) suggested that the Fountain Lake Group is predominantly northward facing, but this was not confirmed by our work, and clear evidence of southward-facing units was found in several areas, indicating that the Diamond Brook Formation underlies (structurally or stratigraphically) the Byers Brook Formation. Way-up indicators, including basaltic flow tops and scours in siltstone, are commonly to the south (Piper, 1994).

Devonian–Carboniferous strata south of Rockland Brook Fault are represented by the Nuttby Formation which contains early Tournaisian spores (Donohoe and Wallace, 1982a, b, c, d). The Nuttby Formation, first defined by Donohoe (1975) as the "Nuttby Succession" and redefined as the Nuttby Formation by Donohoe and Wallace (1980), appears only in the Nuttby area where it has a thickness of approximately 850 m (Donohoe and Wallace, 1980). The type section outcrops on North River, north and south of Nuttby (NTS 11 E/11) and the formation is also well exposed on West Branch North River. The formation contains grey fluviatile quartzite and siltstone, minor polymictic conglomerate, and dark grey siltstone, and shale. Relationships with older or younger stratigraphic units are not exposed. The northern boundary of the Nuttby Formation is a fault.

North of the Rockland Brook Fault, an undated unit at the northern margin of the Pleasant Hills plutons may be correlative to the Nuttby Formation (Piper, 1994). Purplish and green siltstone, sandstone, and conglomerate with granite clasts similar to the pluton, outcrop along the Economy River. Rhyolite and vein quartz fragments also occur. The occurrence of hornfels and crosscutting granite veins suggest that this unit is essentially synchronous with the Pleasant Hills pluton, and hence early Carboniferous. In West Branch Economy River, this unit appears to be thrust over the Wilson Brook Formation.

Late Devonian plutons

Late Paleozoic plutonism predominantly occurs north of the Rockland Brook Fault, which forms the southern contact for many of the Late Devonian plutons in the eastern highlands (Fig. 1). These Late Devonian plutons are bimodal consisting predominantly of gabbro (unit 10a, Fig. 1) and granite (unit 10b, Fig. 1). Where intermediate compositions occur, they appear to be mixtures between mafic and felsic end members, and almost all plutons display some evidence of magma mixing. Gabbroic to dioritic bodies include the Folly Lake and Economy River plutons, whereas granitic plutons (terminology of Streckeisen (1976)) include Pleasant Hills, Gilbert Mountain, Hart Lake–Byers Lake, and Salmon River. Uranium-lead data on zircons from selected granitic plutons indicate an emplacement age of ca. 360 Ma (Table 9, Fig. 14; Doig et al., 1996). On the basis of field evidence for magma mixing, the emplacement age of the gabbroic and dioritic bodies is assumed also to be ca. 360 Ma.

The Pleasant Hills and Hart Lake–Byers Lake plutons yielded ages of 358–362 Ma, 19–46 Ma greater than previously reported Rb-Sr whole rock isochron ages (Cormier, 1979; Donohoe et al., 1986; Pe-Piper et al., 1989). Zircon from different localities in the Pleasant Hills pluton consistently contain 5–7 ppm common Pb, presumably in small, dark inclusions that occur in all grains.

Folly Lake and Economy River plutons

The Folly Lake pluton intruded folded Early Devonian strata and is intruded by, and mixed with, granite of the Hart Lake–Byers Lake pluton. The pluton is bounded to the south by the Rockland Brook Fault and to the north by the Hart Lake–Byers Lake pluton and unconformably overlying Late Carboniferous sedimentary rocks. To the southwest of the pluton, gabbro is thrust over Fountain Lake Group.

The Folly Lake pluton ranges in composition from gabbro to diorite, although much of the diorite may be due to mixing with the Hart Lake–Byers Lake pluton. Hybridization along lobate boundaries and inclusions of partly digested diorite or gabbro in the granite suggest injection of granite into still partly molten mafic magma. The Economy River pluton consists mainly of gabbro and diorite.

Subophitic textured gabbro contains large crystals of purplish-brown augite (45–60%), overgrown and pseudomorphed by olive-green hornblende which itself is overgrown by, and unmixed to greenish and bluish actinolite and stubby laths of oligoclase. Blue actinolite is commonly associated with titanite. Deep brown, strongly absorbing biotite occurs as poikilitic tabular grains in fine-grained aggregates with actinolite, or as overgrowths on hornblende. Accessory minerals include strongly prismatic apatite, subhedral zircon, and penninite. Magnetite occurs as large anhedral to subhedral grains which commonly have needle-like exsolution of ilmenite. Pyrite, subhedral to anhedral, commonly occurs in clinopyroxene and ilmenite occurs as anhedral to subhedral (prismatic) large crystals included in clinopyroxene or actinolite. Small anhedral grains of

Table 9. Uranium-lead data for Devonian–Carboniferous plutonic rocks of the Cobequid Highlands.

Zircon fraction*	Weight (mg)	U (ppm)	Pb_{rad} (ppm)	$^{206}Pb/^{204}Pb$†	$^{208}Pb/^{206}Pb$‡	$^{206}Pb/^{238}U$‡	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$‡	$^{207}Pb/^{206}Pb$ age (Ma)
Cape Chignecto pluton									
2M, 50 x 125 Ab	0.047	280	17.3	2617	0.225	0.05603	0.4148	0.05368	358
2M, 75 x 200	0.136	431	24.9	1376	0.214	0.05311	0.3933	0.05372	359
3M, 50 x 125	0.131	327	19.6	1821	0.210	0.05508	0.4077	0.05368	358
Pleasant Hills pluton									
0M, 75 x 200	0.166	1014	57.1	710	0.161	0.05395	0.4003	0.05381	363
1M, 75 x 200	0.194	1320	75.2	844	0.180	0.05368	0.3975	0.05371	359
2M-1, 125 x 250	0.271	1144	63.3	659	0.151	0.05338	0.3959	0.05379	362
2M-2, 100 x 225	0.272	1324	75.4	637	0.167	0.05425	0.4022	0.05378	362
Hart Lake–Byers Lake pluton									
0NM, 75 x 200 Ab	0.094	1219	72.3	1234	0.184	0.05571	0.4136	0.05384	365
0NM, 75 x 200	0.464	1587	89.1	4121	0.159	0.05385	0.3991	0.05376	361
0M, 124 x 250	0.397	1438	79.6	4405	0.161	0.05297	0.3922	0.05374	360
1M, 125 x 250	0.398	1798	98.8	2124	0.162	0.05257	0.2902	0.05384	364
West Moose River pluton: zircon with an inherited component									
6NM, 40 x 100	0.021	283	16.9	1115	0.198	0.05543	0.4235	0.05542	429
6M-1, 75 x 200	0.100	1176	66.7	3566	0.144	0.05506	0.4087	0.05383	364
6M-2, 50 x 125	0.019	592	33.2	1268	0.137	0.05482	0.4088	0.05408	375

*N and M are nonmagnetic and magnetic fractions, respectively, and the numerals are degrees tilt on the Frantz LB-1 separator; 75 x 250, etc. are zircon dimensions in micrometres (μm); Ab, Abraded.

†Atomic ratios corrected for fractionation and spike.

‡Atomic ratios corrected for fractionation, spike, blanks, and common Pb from the model of Stacey and Kramers (1975).

chalcopyrite commonly occur in actinolite. Most of the opaque minerals show a vermicular-type intergrowth with the silicate minerals.

Porphyritic and equigranular diorite and tonalite intruded gabbro, and xenoliths of porphyritic diorite have been found in equigranular diorite, thereby yielding a sequence of intrusion. Apart from the porphyritic nature of plagioclase, the two rocks are very similar (color index 28–45), consisting of a granitic to subophitic-like intergrowth of plagioclase (prismatic and commonly zoned), hornblende (olive-green, subhedral with actinolite rims), biotite (dark brown equant laths), and tabular, lamellar, or skeletal opaque minerals. Minor amounts of quartz or alkali-feldspar form anhedral interstitial grains. Accessory minerals include apatite, epidote, zircon, and titanite.

The Folly Lake pluton is cut by late numerous fine-grained granitic dykes and sills, locally pegmatitic, and rare porphyritic mafic dykes. These alkali feldspar-rich granitic rocks exhibit a characteristic granophyric or granitic texture with acicular hornblende crystals. Alkali feldspar forms intergrowths of orthoclase-perthite, subhedral orthoclase grains, and rims on plagioclase crystals. Quartz may be graphically intergrown with feldspar (commonly perthite). Plagioclase (albite, An<3) occurs only in perthite in some samples, but in others reaches 30% modal in well zoned subhedral crystals. Olive-green actinolitic hornblende forms ragged elongate crystals containing inclusions of feldspar, quartz, and/or zircon. Green biotite occurs in clusters, with a radiating habit in some cases. Brown biotite is present in subhedral isolated crystals. Accessory minerals include apatite, magnetite with exsolution lamellae of ilmenite, zircon, subhedral titanite, zoned tourmaline, rutile, and large subhedral, translucent, red crystals of hematite. Alteration of hornblende to epidote, plagioclase to clay minerals, orthoclase and perthite to muscovite, and biotite to chlorite is common. Epidote also occurs in secondary veins.

Hart Lake–Byers Lake and Salmon River plutons

The Hart Lake–Byers Lake and Salmon River plutons are similar in field appearance and are discussed together. The Hart–Lake Byers Lake pluton is a composite pluton that lies on the northern margin of the Folly Lake pluton (Fig. 1) and is over 30 km in length. The northwestern edge of this pluton is onlapped by the Westphalian A Boss Point Formation. The southeastern part of the pluton is cut by the Rockland Brook Fault, producing a ductile shear zone. On the northern and eastern margins, the pluton contacts Devonian–Carboniferous volcanic rocks of the Byers Brook Formation (Fountain Lake Group). A fine-grained porphyritic or equigranular granophyric marginal unit of the pluton hundreds of metres wide suggests an intrusive relationship. A similar fine-grained margin occurs where the granite contacts with the Folly Lake pluton, and granitic dykes cut the gabbroic pluton. The Salmon River pluton is very poorly exposed in separate outcrops along logging roads (Fig. 1) and on White River and a tributary of West Branch River John. Its inferred surface area is considerably less than that implied by Donohoe and Wallace (1982a, b, c, d), because age dating revealed that supposedly correlative rocks south of the Rockland Brook Fault are Neoproterozoic.

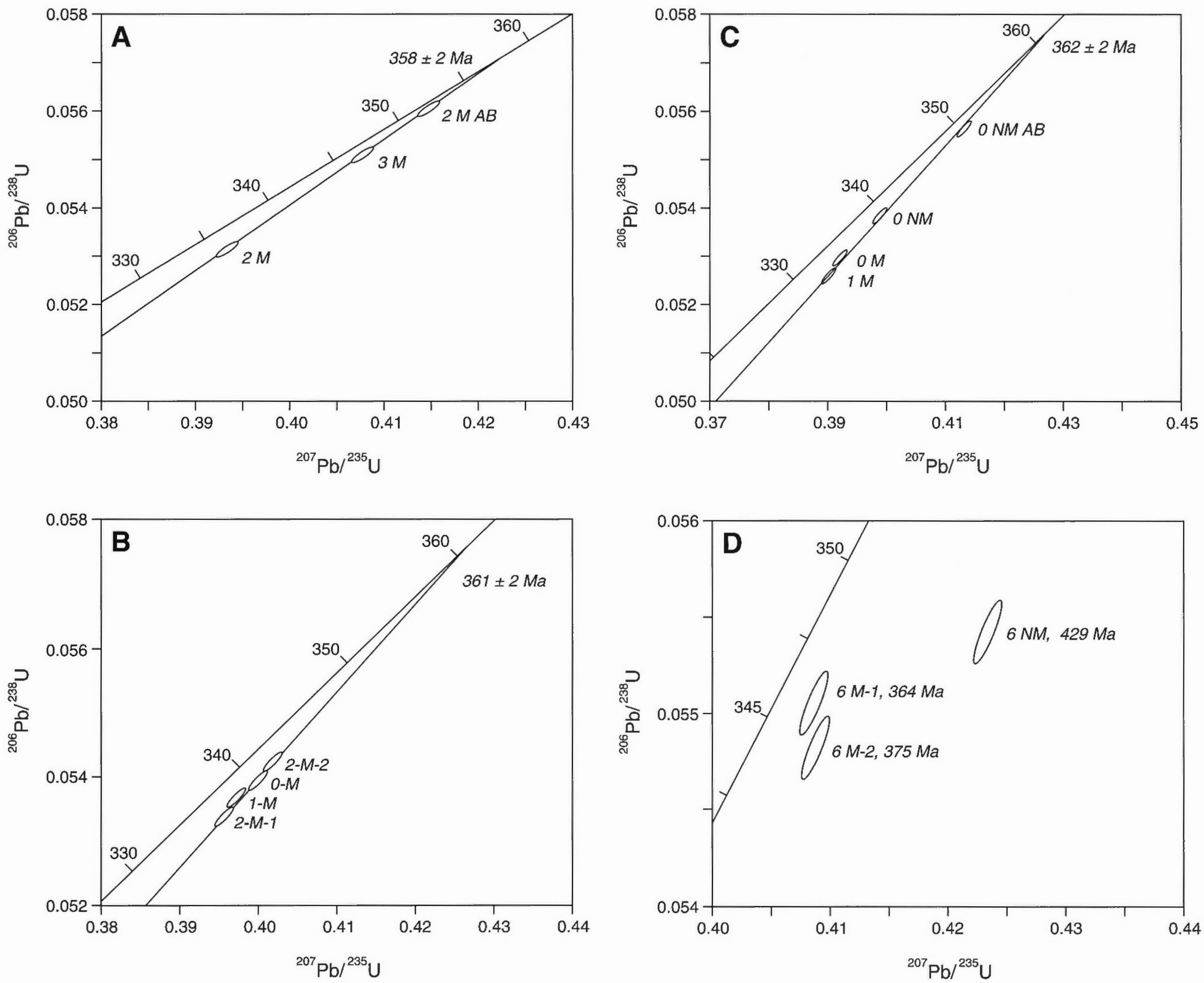


__Figure 14.__ Uranium-lead concordia diagram for zircon from the __A)__ Cape Chignecto, __B)__ Pleasant Hills, __C)__ Hart Lake–Byers Lake, and __D)__ West Moose River plutons. Variations in $^{207}Pb/^{206}Pb$ dates of zircon fractions from the same sample is attributed to a small, but significant, inherited component.

The main lithology of both plutons is a coarse-grained, quartz-rich, locally graphic, alkali-feldspar leucogranite (Table 10) cut by pegmatitic veins, many of which are mineralized (especially in the Folly River–Debert River area) with magnetite, chalcopyrite, pyrite, tourmaline, and (?)siderite. Between Folly and Debert rivers a small inlier of Folly Lake gabbro is cut by coarse leucocratic granite with chilled margins. In this same area, xenoliths of flow-banded rhyolite occur in the granite.

Feldspar crystals occur almost exclusively in perthitic intergrowths of orthoclase and albite, although some discrete subhedral albite laths have been observed. Point counts (in the Hart Lake–Byers Lake pluton only) yielded 23–40% K-feldspar and 23–30% plagioclase. Plagioclase (An<3) in both perthitic and individual crystals exhibits albite and pericline twinning. Quartz occurs in large polycrystalline masses which may have cuneiform outline. The individual grains are fairly coarse, anhedral, and show undulose extinction and subgrain boundaries. Mafic minerals occur in polymineralic clusters. Biotite is present in ragged brownish to greenish grains. Many of these are altered to chlorite (penninite). Subhedral to anhedral amphibole is commonly an olive-green hornblende, although some samples contain blue to grey arfvedsonite that is partly altered to a yellow-brown allanite or orange-red titanite. Polycrystalline patches of ilmenite are associated with arfvedsonite and hornblende. Crystals of magnetite with hematite rims are common as inclusions in feldspars. Magnetite also occurs as inclusions in arfvedsonite and commonly shows exsolution of ilmenite and hematite. Zircon is an abundant accessory mineral, occurring as subhedral inclusions in chlorite, hornblende, or opaque minerals, or as isolated grains. Rare titanite forms large anhedral grains. Fluorite is rarely found as interstitial grains within clots of ferromagnesian minerals.

Table 10. Modal compositions (in per cent) of representative samples from Late Devonian Pleasant Hills and Hart Lake–Byers Lake plutons.

	Hart Lake–Byers Lake pluton												
	Pleasant Hills pluton					Marginal granite bodies		Coarse granite bodies			Leucocratic granite		
Sample	**25-5-1**	**25-5-6**	**25-5-8**	**26-13-1**	**26-13-2**	**35-5-1**	**35-5-2**	**29-8-7**	**35-7-1**	**36-7-1**	**36-5-1**	**36-6-2**	**44-1-1**
Plagioclase	–	–	–	–	3.1[2]	1.8[2]	–	–	–	–	–	–	–
Albite[1]	20.8	22.6	12.7	27.7	26.0	20.9	26.7	30.1	23.6[3]	28.1	22.2[3]	18.5	28.6
Quartz	25.3	24.6	32.2	25.3	26.5	32.2	24.8	35.0	46.8	29.4	31.3	43.1	33.1
K-feldspar	52.4	50.7	52.9	41.7	43.4	39.5	43.0	31.6	23.7	40.2	44.5	37.4	36.2
Hornblende	–	–	–	–	–	2.2	1.6	0.6	0.6	–	0.3	–	–
Biotite	–	–	–	–	0.5	0.3	–	0.2	–	1.0	0.3	–	0.5
Opaque minerals	1.5	2.1	2.1	4.8	3.2	0.4	1.1	0.7	0.8	0.5	1.5	0.8	0.8
Chlorite	–	–	–	–	–	1.0	0.6	1.7	–	0.9	–	–	–
Epidote	–	–	–	0.3	0.3	0.1	0.2	–	–	–	–	0.2	–
Zircon	–	–	–	0.3	–	–	0.2	–	–	–	–	–	–
Red alteration mineral	–	–	–	–	–	0.3	–	0.1	1.0	–	–	–	0.2
Fluorite	–	–	–	–	–	–	–	–	0.1	–	–	–	–
Arfvedsonite	–	–	–	–	–	–	–	–	3.5	–	–	–	–
Total	100.0	100.0	99.9	100.1	99.9	100.0	100.0	100.0	100.1	100.1	100.1	100.0	100.0
Color index	1.5	2.1	2.1	5.1	4.0	4.3	3.7	3.2	6.0	2.3	2.0	1.0	1.9
IUGS Q-P-A (normalized to 100)													
Plagioclase	–	–	–	–	–	3.2	1.9	–	–	–	–	–	–
Quartz	25.7	25.1	32.9	26.7	27.6	33.6	25.8	36.2	49.7	30.1	31.9	43.6	33.8
Alkali–feldspar[4]	74.3	74.9	67.1	73.3	45.2	63.1	72.3	63.8	50.3	60.9	68.1	56.4	66.2
Rock name	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite	Alkali-feldspar granite

Notes: 1. Plagioclase in perthite structure and as individual crystals
2. Plagioclase refers to individual plagioclase grains of which composition is unknown
3. Feldspars analyzed by electron microprobe
4. Albite is counted as alkali-feldspar in IUGS calculations

– = below detection

Granitic dykes intruding mafic rocks within the Hart Lake–Byers Lake pluton are very leucocratic light pink, graphic or glomeroporphyritic textured, with phenocrysts of quartz and perthite in an allotriomorphic groundmass. Mineralogy resembles that of the main phases (Table 10). Zoned, six-sided crystals of yellow-brown tourmaline have been found in one sample. A number of unusual rock types have also been noted, including a porphyritic banded rhyolite.

Diabase dykes cutting both plutons are aphanitic and approximately 3 m wide on average, with sharp planar chilled margins. They show subophitic, equigranular texture of large subhedral to anhedral crystals of pinkish augite (20–25%) and elongate or stubby twinned, oscillatory zoned laths of twinned plagioclase (50–55%). Olive green to very dark brown, strongly absorbing biotite forms 10–15%. Polyhedral clumps of extremely fine-grained green minerals (?chlorite, ?amphibole, ?antigorite) (10–15%) may be an alteration of olivine. Opaque minerals occur as tabular, tetragonal or very large anhedral, skeletal grains. Apatite may be present as an accessory mineral occurring in fairly large elongate grains. A purplish flaky mineral, fairly abundant (0.2%) in one sample, was tentatively identified as perovskite.

Pleasant Hills pluton

The Pleasant Hills pluton, a 17 km long body, is a composite polyphase pluton with granitoid rocks in the north, felsic subvolcanic rocks in the east, and diorite in the south. The pluton is cut by numerous diabase dykes and a few gabbroic bodies. At the northern boundary, granitoid rocks intruded synplutonic sedimentary rocks, whereas the southern boundary is generally a ductile shear zone. Shear zones within the complex impart an intense mylonitic fabric. Miller et al. (1989) demonstrated that penetrative deformation along the Rockland Brook Fault affected the pluton and that granite lithologies on opposite sides of the fault are significantly different.

Lithologies include very coarse- to fine-grained granodiorite and granite, locally with rapakivi texture, rhyolite, gabbro, diabase, diorite, and hybrid rocks. Fine-grained granite from the northwestern margin of the pluton consists of equigranular varieties (group a), and porphyritic varieties (group b) which are comparatively quartz-rich and contain lesser amounts of zircon and opaque minerals than group a. In both groups the groundmass is microcrystalline and usually granophyric.

In general, the granitic rocks are alkali-feldspar granite and display perthitic intergrowths, although embayed quartz and plagioclase (An_3) are also present. The groundmass displays myrmekitic, micrographic, and granitic textures. Clusters of very fine-grained, magnetite crystals, randomly dispersed, comprise 2–3% of the rock. Minor subhedral pyrite and secondary hematite occur. Biotite is olive green, although minor brown biotite (less than 1%) occurs in fine-grained clusters. Accessory titanite and zircon are subhedral to euhedral and commonly associated with ferromagnesian mineral clusters. Acicular rutile may occur as inclusions within quartz. Anhedral fluorite has been found as interstitial grains to the felsic minerals. Muscovite occurs as vein-like or pervasive alteration of the groundmass and phenocrysts, and chlorite occurs as alteration of biotite.

Gilbert Mountain pluton

The Gilbert Mountain pluton as mapped by Donohoe and Wallace (1982a, b, c, d) consists of granite, quartz diorite to tonalite, gabbro, and diabase. The granitic rocks, pink, white and black, medium- to coarse-grained equigranular rocks, consist mainly of quartz and K-feldspar with lesser amounts of plagioclase, hornblende, and biotite. Some rocks are porphyritic with quartz and K-feldspar phenocrysts. Hornblende and biotite, variably altered to actinolite and chlorite respectively, occur in clots associated with zircon, epidote, titanite, and chlorite. Fractures are filled with epidote and actinolite. Quartz diorite, predominantly equigranular rocks with "salt-and-pepper" texture, consist of anhedral quartz, plagioclase, hornblende, biotite, and, in some samples, phenocrysts of K-feldspar which may be partially altered to albite. Accessory minerals include titanite, apatite, zircon, and opaque minerals. One sample contains a fragment of granophyre about 0.6 mm that is composed of altered plagioclase, quartz, opaque minerals, biotite, muscovite, hornblende, and calcite. In general, the rocks are moderately altered and deformed, as shown by sericitized plagioclase, chloritized hornblende and biotite, and strained and grain-reduced quartz.

Geochemical data

Twenty-two samples from the Pleasant Hills and Hart Lake–Byers Lake plutons and granitic pods in the Folly Lake pluton were selected for chemical analysis. Major-element and trace-element chemistry for these granitic rocks is presented in Table 11. Note the relatively low concentrations of Al_2O_3, MgO, and CaO and high alkali content. Normative quartz, anorthosite, and orthoclase contents and the molar ratio $(Na_2O+K_2O)/Al_2O_3$ (Table 11) suggest alkali affinities, although the analyses commonly give very small amounts of corundum (0–1.6% average 0.7) in the CIPW norms. The Hart Lake–Byers Lake pluton is generally comparable to alkali granite whereas some of the Pleasant Hills granitic rocks fall in their field of subalkali granite (Fig. 15). Geochemical data indicate that group a and group b of the Pleasant Hills pluton are distinguished by SiO_2 content, but both exhibit decrease of Al_2O_3, TiO_2, Fe_2O_3T, Na_2O, CaO, MnO, Ba, Zr; similar MgO, P_2O_5, and Sr; and increase of K_2O and Rb with increasing SiO_2 (Fig. 16).

Fractionation trend plots for these two groups (Fig. 17) indicate fractionation of K-feldspar (Ba vs. Sr and Ba vs. Rb) and biotite (Ba vs. Sr). Thus, the overall trend seen in the

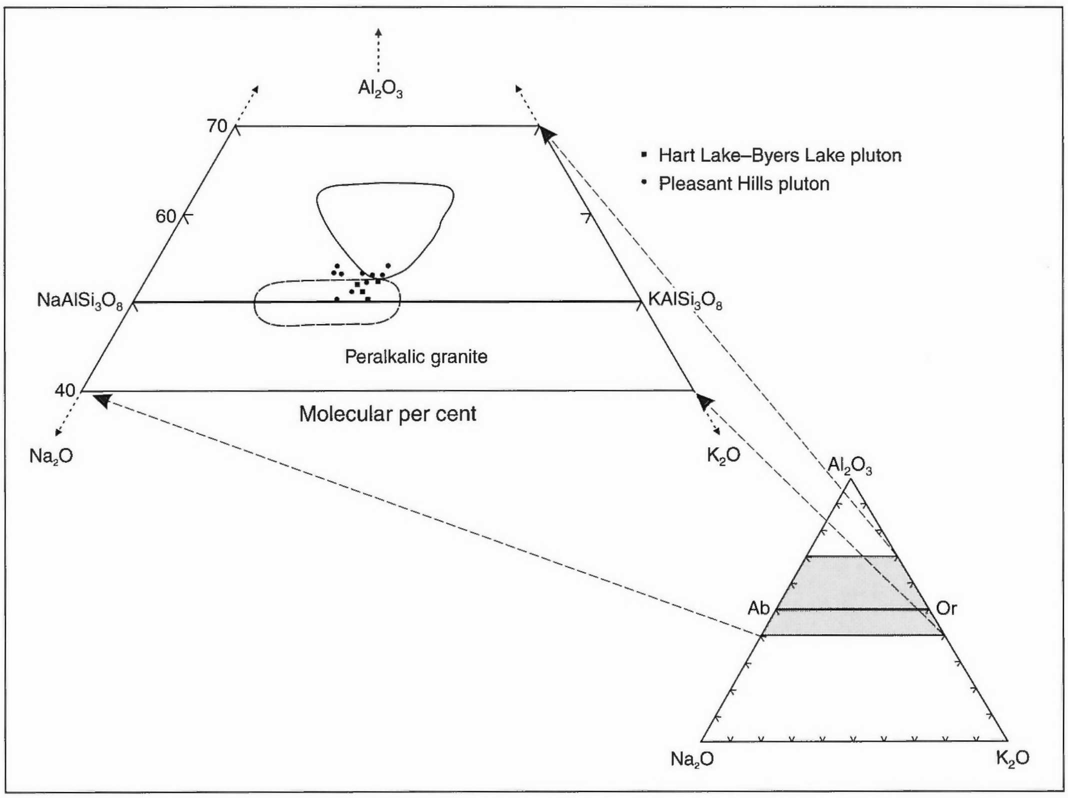


Figure 15. Molecular proportion of Na_2O, K_2O, and Al_2O_3 for selected samples from the Carboniferous granite bodies of the eastern Cobequid Highlands compared with alkali (dashed-line field) and subalkali granitic rocks (solid-line field) from New England (field after *Hermes et al., 1978). Note because some analyses are very similar, in this and subsequent figures, for clarity not all analyses have been plotted.*

Pleasant Hills granite could be due to a fractionation controlled mainly by K-feldspar±biotite with crystallization (saturation) of accessory minerals, such as zircon, titanite, and apatite. Differentiation trend plots for the Hart Lake–Byers Lake pluton granite (Fig. 17) also indicate K-feldspar and biotite fractionation and are most similar to those for group b of the Pleasant Hills pluton, although the extent of differentiation to higher SiO_2 values is more pronounced. Both the Pleasant Hills and the Hart Lake–Byers Lake granitic rocks are less aluminous and richer in Ti than typical collision granitic rocks (Harris et al., 1986), but are persistently richer in K_2O than Na_2O, a feature typical of collision granitic rocks. The trace-element compositions of these granitic rocks are typical of within-plate granitic rocks (Fig. 18), consistently showing compositions of Rb, Nb, and Y similar to those described by Pearce et al. (1984). These granitic rocks do not show characteristics of postcollisional granite (Harris et al., 1986), such as depletion in Nb, Zr, Y, Ce, and Hf.

Table 11. Geochemistry of the Pleasant Hills and Hart Lake–Byers Lake Late Devonian granitoid plutons.

Sample number	26-13 -3	26-13 -6	26-13 -5	26-13 -1	26-13 -4	26-13 -2	25-5 8	25-5 -6	25-5 -5	25-5 -2
SiO_2	70.36	70.37	71.05	71.31	71.46	71.94	73.10	73.20	73.76	74.34
TiO_2	0.27	0.40	0.43	0.39	0.27	0.28	0.20	0.24	0.20	0.18
$A1_2O_3$	13.27	13.94	14.08	13.80	13.34	13.78	13.02	13.52	13.05	13.02
Fe_2O_3(t)	3.21	3.64	3.27	3.74	3.19	3.20	1.86	2.11	1.72	1.89
MnO	0.05	0.05	0.05	0.05	0.05	0.06	0.02	0.01	0.02	0.02
MgO	0.88	0.95	1.00	0.95	0.85	0.73	0.85	0.79	0.86	0.75
CaO	0.86	0.79	0.80	0.78	0.78	0.85	0.17	0.07	0.18	0.50
Na_2O	4.56	4.65	4.37	4.35	4.09	4.90	4.00	3.80	3.84	3.63
K_2O	4.97	4.33	4.34	4.51	4.87	5.12	5.25	5.17	5.21	5.27
P_2O_5	0.04	0.07	0.07	0.06	0.03	0.04	0.03	0.04	0.04	0.03
LOI	0.60	0.30	0.20	0.10	0.50	0.00	0.30	0.20	0.30	0.30
Total	99.07	99.49	99.66	100.04	99.43	100.90	98.80	99.15	99.18	99.93
CIPW norms (wt %)										
Qz	20.47	21.52	23.95	23.27	24.21	19.37	27.35	28.98	29.10	29.76
Or	23.93	25.89	25.87	26.77	29.19	30.08	31.56	30.94	31.19	31.32
Ab	39.31	39.82	37.30	36.97	35.10	41.42	34.43	32.57	32.92	30.89
An	1.08	3.64	3.70	3.64	3.71	0.48	0.72	0.19	0.70	2.37
Hy	5.92	7.90	7.34	8.03	7.07	5.14	4.98	5.14	4.75	4.75
Ilm	0.52	0.77	0.82	0.74	0.52	0.53	0.39	0.46	0.38	0.34
Ap	0.09	0.16	0.16	0.14	0.07	0.09	0.07	0.09	0.09	0.07
Cor	–	0.29	0.86	0.44	–	–	0.51	1.63	0.85	0.48
Trace elements (ppm)										
Ba	597	578	647	630	583	627	255	438	223	281
Rb	200	164	176	180	197	203	210	198	230	188
Sr	47	45	68	54	38	42	35	31	28	59
Y	87	73	64	72	82	83	64	71	78	89
Zr	577	498	409	464	566	554	256	308	241	257
Nb	36	28	22	27	35	32	19	28	28	30
Th	12	15	12	12	14	13	13	14	17	16
Pb	10	4	2	3	16	12	27	29	28	18
Ga	19	17	20	20	23	21	20	20	17	22
Zn	44	28	40	31	41	49	40	19	33	16
Cu	7	6	7	7	3	6	7	7	4	7
Ni	8	3	7	3	5	5	7	7	5	4
V	0	16	22	14	0	3	6	4	2	6
Cr	28	39	38	26	30	24	37	39	23	28
K/Rb	210	221	206	208	208	208	211	219	190	234
Rb/Sr	4.3	3.6	2.6	3.3	5.2	4.8	6.0	6.4	8.2	3.2
Sr/Ba	0.08	0.08	0.11	0.09	0.07	0.07	0.14	0.07	0.13	0.21
Mol $\frac{Na_2O+K_2O}{Al_2O_3}$	0.87	0.89	0.84	0.87	0.90	0.99	0.99	0.88	0.92	0.90

Major elements and Ba-Cr trace elements were analyzed by X-ray fluorescence at St. Mary's University on a Philips PW1400 sequential spectrometer using a Rh-anode X-ray tube. International standards with recommended values from Abbey (1983) as well as in house standards were used for calibration. Analytical precision, as determined from replicate analyses is generally better than 2%, except MgO, Na_2O, and Nb which are better than 5%, and Th which is better than 10%.
Loss on ignition (LOI) was determined by treating the sample for 1.5 h at 1050°C in an electric furnace.

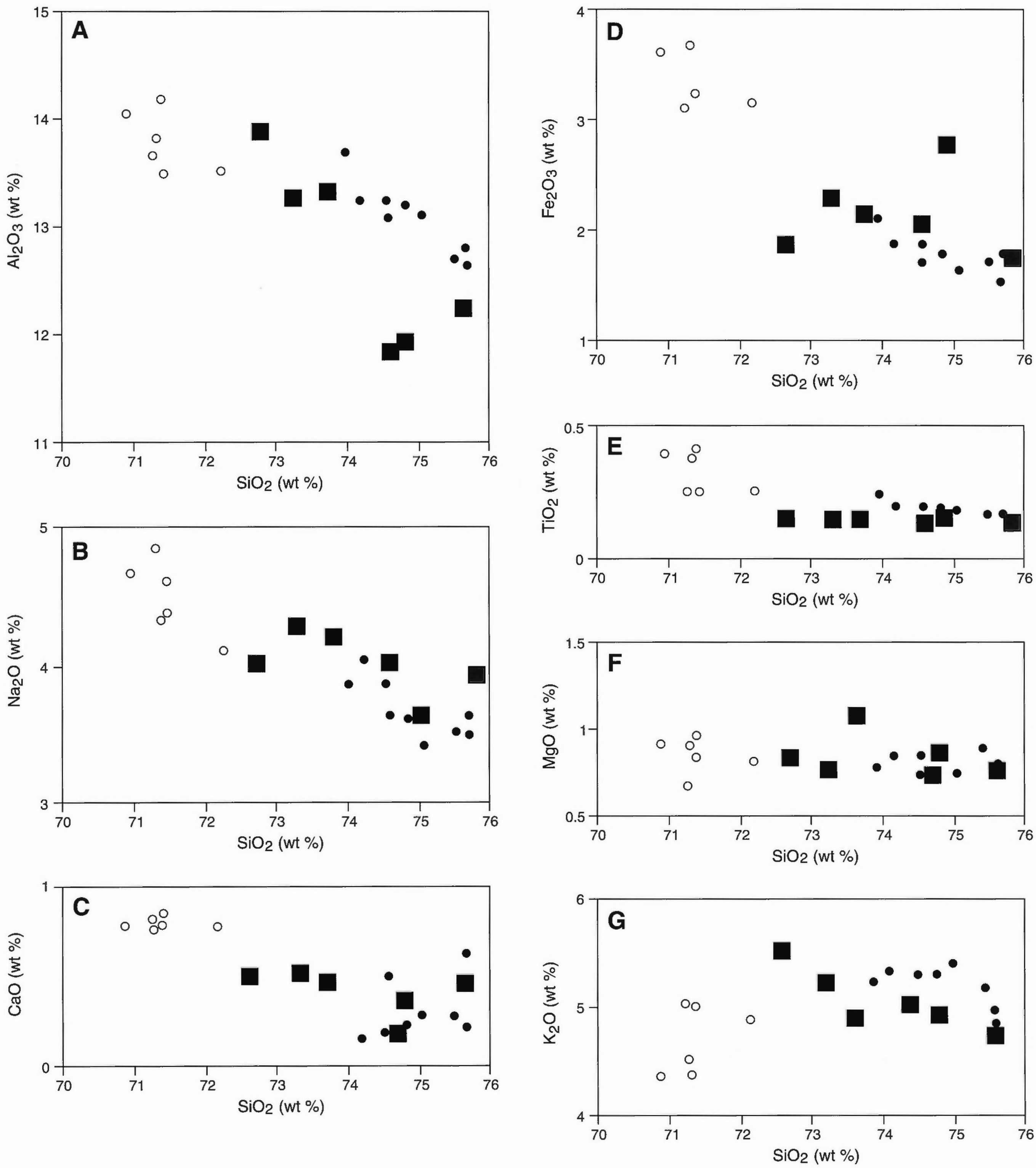


Figure 16. *Major-element and trace-element variations with SiO_2 for the Hart Lake–Byers Lake and Pleasant Hills granitic rocks.*

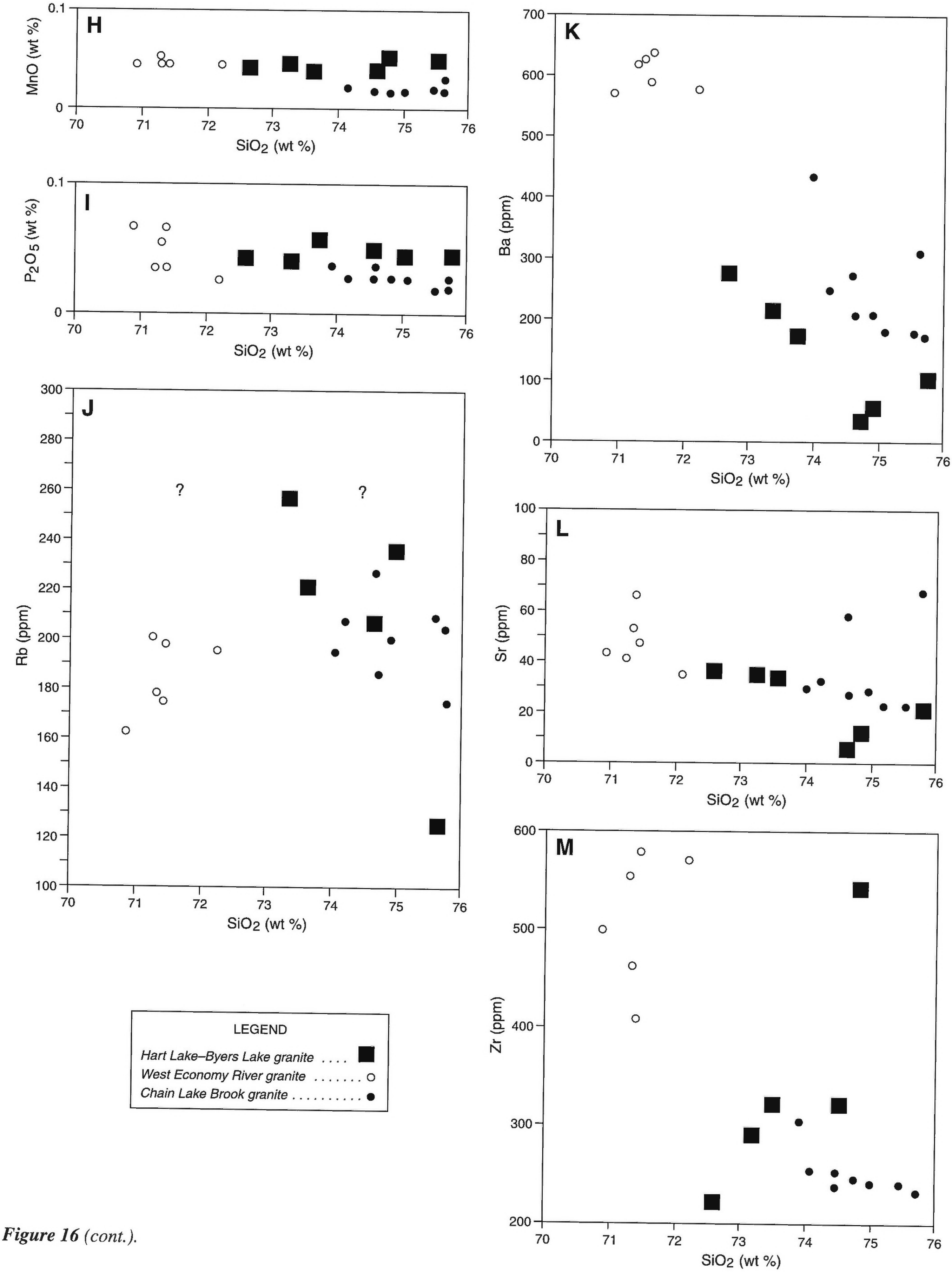


***Figure 16** (cont.).*

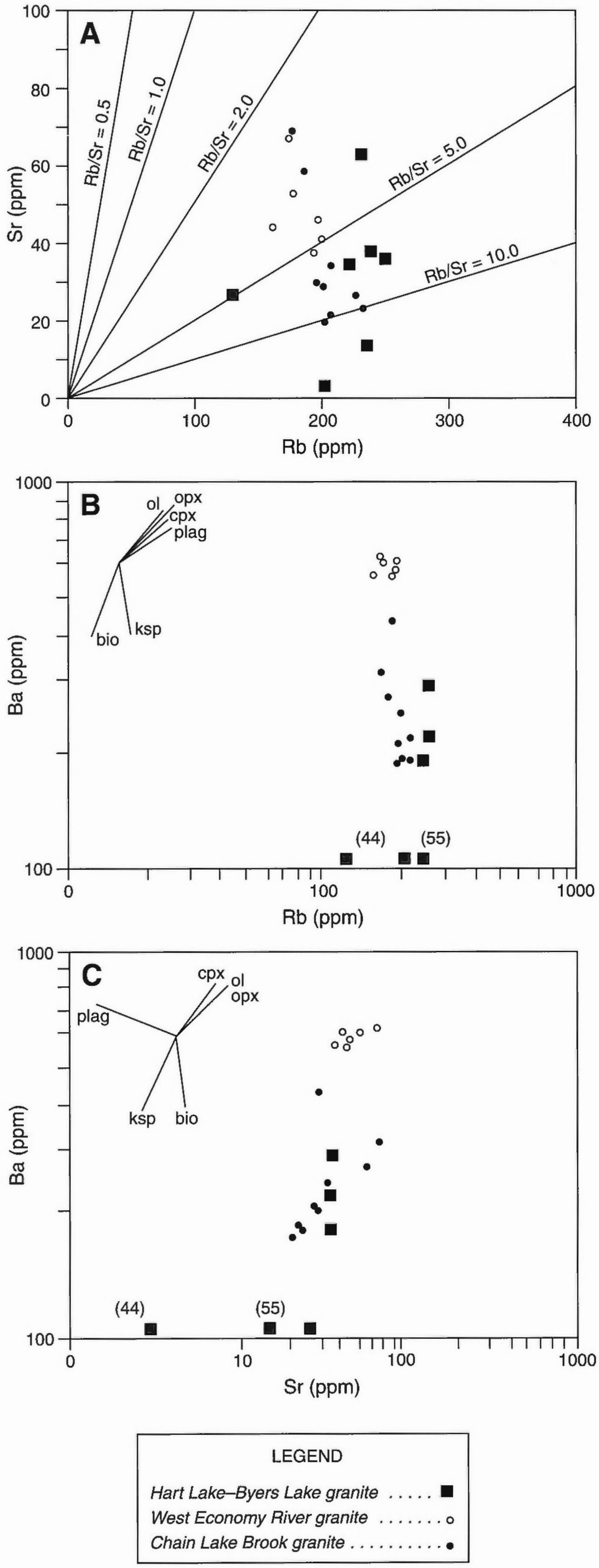


Figure 17. ***A)*** *Sr versus Rb,* ***B)*** *Ba versus Rb, and* ***C)*** *Ba versus Sr for the Hart Lake–Byers Lake and Pleasant Hills granitic rocks. The mineral vectors are taken* from *Tindle and Pearce (1981); ol = olivine; opx = orthopyroxene; cpx = clinopyroxene; plag = plagioclase; bio = biotite; ksp = K-feldspar. Numbers in parentheses show Ba abundances outside the compositional range of the figure.*

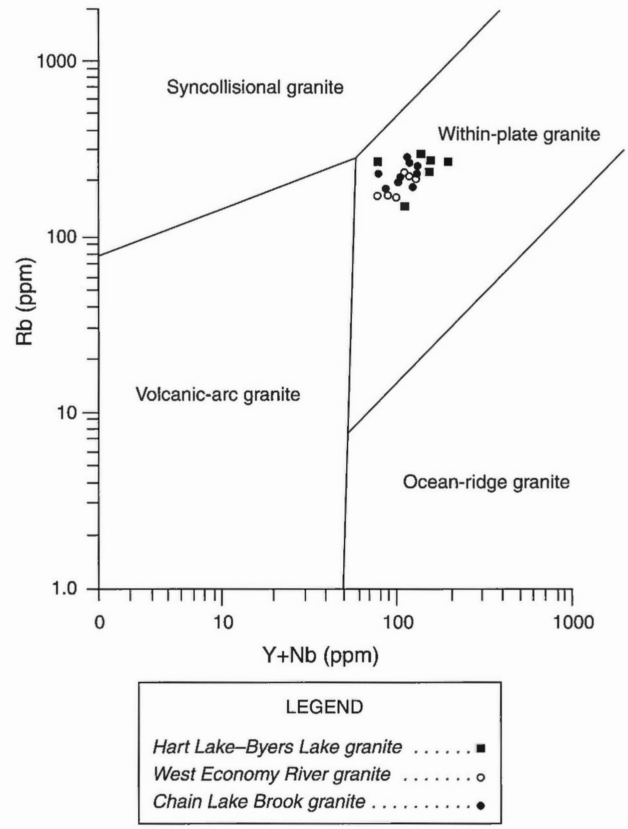


Figure 18. Plot of Rb versus Y+Nb for Hart Lake–Byers Lake and Pleasant Hills granitic rocks. Discrimination fields after *Pearce et al. (1984).*

STRUCTURAL GEOLOGY

Precambrian deformation

Bass River Complex

Three distinct phases of deformation can be distinguished in the Bass River Complex (Nance and Murphy (1990) here designated D_{1a}, D_{1b}, and D_{1c}. Development of metamorphic fabric D_{1a} in the Great Village River orthogneiss and amphibolite is interpreted as having accompanied igneous emplacement under amphibolite-facies conditions, dated at ca. 587–580 Ma. Development of mylonitic fabric (D_{1b}) accompanied emplacement of deformed granitic rocks into active ductile shear zones along the contact between the Gamble Brook Formation and the Great Village River Gneiss, dated at about 605 Ma. D_{1b} mylonitic overprinting of the metamorphic fabric in the orthogneiss indicates that intrusion and metamorphism of the Great Village River Gneiss must have commenced prior to ca. 600 Ma, implying that plutonic rocks of the Great Village River Gneiss vary in age by at least 20 Ma. Folds and locally penetrative fabric (D_{1c}) in the Folly River Formation predate the postkinematic Debert River pluton (ca. 610 Ma). The Folly River Formation has been correlated with part of the lithologically similar Georgeville Group in the Antigonish Highlands (Murphy et al., 1992) that contains detrital zircons with ages of ca. 613 Ma (U-Pb, Keppie and Krogh, 1990). If this correlation is valid, deposition and deformation of the Folly River Formation are constrained to ca. 615–610 Ma.

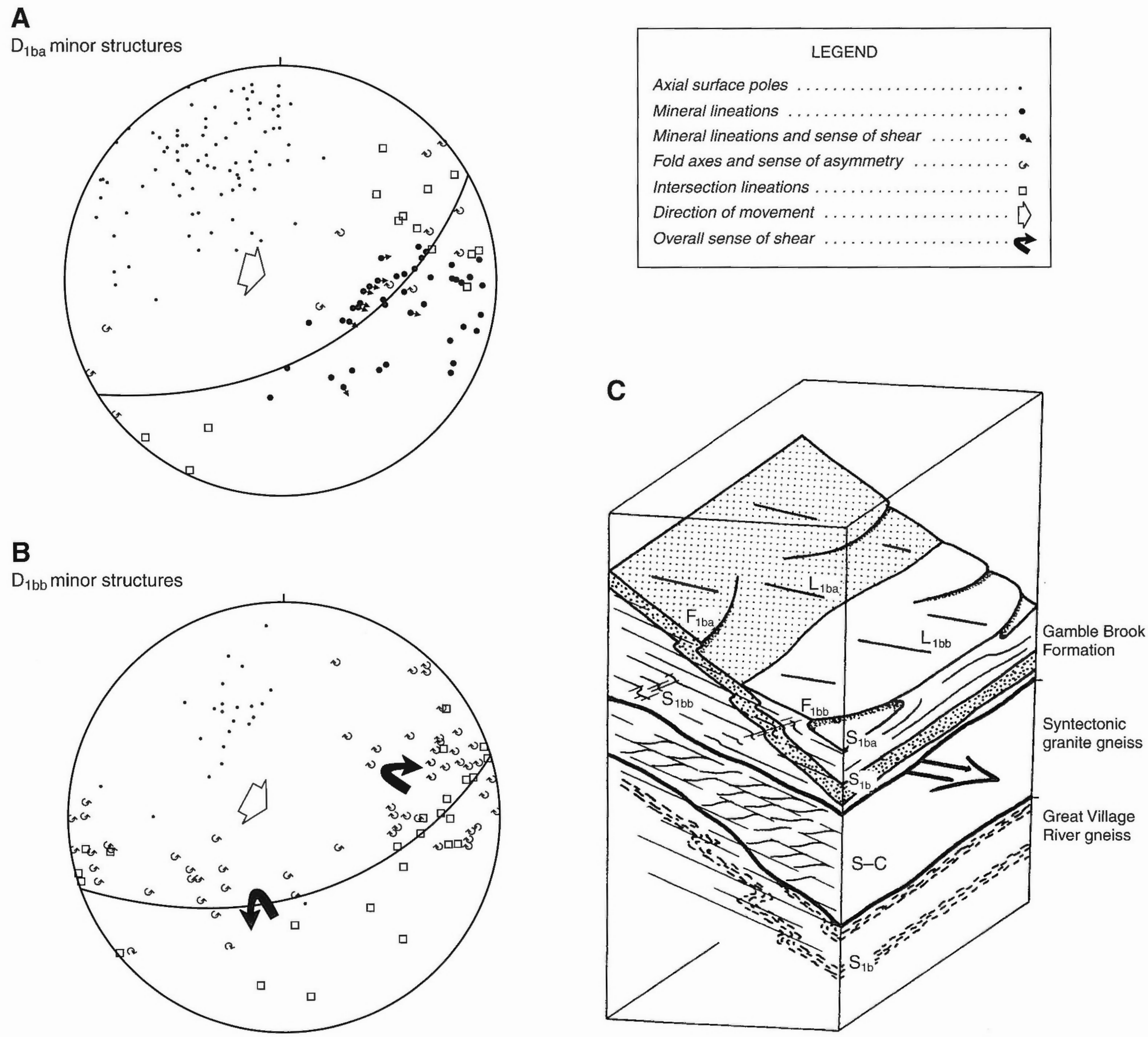


Figure 19. *Structural data and interpretation of D_1 structural elements within the Bass River Complex. The data imply oblique slip with normal and shear components toward the east-southeast within the ductile shear zone.* ***A)*** *Synoptic equal-area stereographic projection of D_{1ba} structural data from the ductile shear zone contact between the Great Village River Gneiss and the Gamble Brook Formation.* ***B)*** *Synoptic equal-area stereographic projection of D_{1bb} structural data from the ductile shear zone contact between the Great Village River Gneiss and the Gamble Brook Formation. Large curved arrows show overall sense of fold asymmetry.* ***C)*** *Conceptual block diagram of D_{1b} relations at the ductile shear zone contact between the Gamble Brook Formation (hanging wall) and Great Village River Gneiss (footwall).*

D_{1a} *structures*

Evidence of D_{1a} deformation is confined to areas where the earliest recognizable folds (to which the D_{1b} mylonitic fabric is axial planar) deform a pre-existing S_{1a} amphibolite-facies foliation defined by hornblende-plagioclase compositional banding in amphibolite and by quartzofeldspathic layering in the gneiss. (In the Gamble Brook Formation, folds to which the mylonitic fabric is axial planar, deformed only the bedding.) D_{1a} folds and linear fabrics have not been observed.

D_{1b} *structures*

D_{1b}, the principal deformation fabric throughout much of the Bass River Complex, formed during protracted progressive deformation of a ductile shear zone between the Great Village River Gneiss and the Gamble Brook Formation, producing strong L-S tectonite fabrics under greenschist- to amphibolite-facies conditions. Two generations of D_{1b} structures and associated fabrics can be recognized which are interpreted as essentially coeval progressive phases of deformation. D_{1ba} fabrics are produced within the ductile shear zone, and D_{1bb} fabrics deform these fabrics, but produce only local fabric development of their own. The structural data associated with each phase is summarized in Figure 19A, B. In the Great Village River Gneiss, the dominant planar fabric (S_{1ba}) ranges from amphibolite-facies metamorphic foliation to mylonitic schistosity developed under heterogeneous ductile shear. The fabric dips southeast at moderate to steep angles and is axial planar to small, tight to isoclinal, doubly plunging folds (locally sheath folds) that deformed the earlier S_{1a} fabric (Fig. 19A). The L_{1ba} lineation, defined by preferred orientation of hornblende and quartzofeldspathic augen, records transport direction during deformation. Kinematic indicators (S-C fabrics, asymmetric augen, fold asymmetry) suggest oblique slip, with normal and sinistral movement towards the present southeast (Fig. 19A).

In mylonitic granitic gneiss, the heterogeneous, L-S tectonite fabric is coplanar and colinear with the D_{1ba} fabric of the amphibolite, but F_{1ba} folds have not been observed within the main body of the rock unit. However, stringers of the granitic gneiss that are discordant to, but contain, the S_{1ba} fabric, frequently define F_{1ba} folds within the Great River Village Gneiss. Some dykes and veins of granite gneiss that cut the S_{1ba} fabric at high angles are only mildly influenced by the D_{1ba} deformation, supporting the conclusion of Cullen (1984) that the granite gneiss bodies are largely synkinematic and that their ca. 605 Ma emplacement age dates development of the D_{1ba} fabric. Within the ductile shear zone, where the granite gneiss intruded both the Great River Village Gneiss and the Gamble Brook Formation, kinematic indicators suggest oblique-slip transport towards the present east-southeast (Fig. 19A, B).

The Gamble Brook Formation contains a heterogeneous, L-S tectonite fabric coplanar, colinear, and continuous with the D_{1ba} fabric in both the Great Village River Gneiss and mylonitic granite gneiss. The fabric is more intense within the ductile shear zone, where it formed largely under greenschist-facies conditions, but the reported presence of staurolite and sillimanite (Cullen, 1984; R.A. Jamieson, pers. comm., 1990), imply a local increase in metamorphic grade to amphibolite facies. In quartzite, an intense S_{1ba} mylonitic fabric is defined by quartz ribbons, the elongation of which also defines a strong lineation (Fig. 19A). The features and movement sense of this fabric are identical to those in the Great Village River Gneiss. The intensity of the fabric decreases with distance from the Gamble Brook Formation–Great Village River Gneiss contact and grades into muscovite-chlorite or muscovite-biotite schistosity in more pelitic lithologies and massive quartzite.

Within the ductile shear zone, fabrics are deformed to produce strongly asymmetric folds, that locally are sheath folds, and are coaxial, coplanar, and spatially associated with F_{1bb}. They differ in style from normal F_{1bb} structures in that they fold S_{1ba} and only locally develop axial planar fabrics. They plunge at varying angles to the northeast, east, south, and southeast and, when plotted stereographically (Fig. 19B), define a partial great circle girdle that is broadly coplanar with the mean orientations of S_{1bb} axial surfaces and a locally developed axial-planar crenulation cleavage. The kinematics and structural geometry mimic those of D_{1ba} and the two are considered phases of a single progressive deformation involving heterogeneous ductile shear with continuous, oblique movement having left-lateral and normal components of slip. The style of deformation in the Gamble Brook Formation probably represents sinistral transtension (Nance and Murphy, 1990).

D_{1c} *structures*

The Folly River Formation is affected by D_{1c} deformation, of which two phases (D_{1ca} and D_{1cb}) can be recognized. D_{1ca} produces the main fabrics and D_{1cb} locally deforms these fabrics. D_{1ca} produced a low-grade planar schistosity (S_{1c}) that dips to the southeast at moderate to gentle angles, a mineral lineation (L_{1c}) that plunges gently to the east-southeast, and an S_{1c} bedding (S_0) intersection lineation parallel to the axes of associated F_{1c} folds which plunges gently to moderately to the southeast (Fig. 20A). S_{1c} is defined by chlorite, actinolite, epidote, and opaque minerals in the mafic volcanic rocks and a bedding-subparallel, muscovite-chlorite cleavage in interlayered pelitic rocks. The L_{1c} mineral lineation, defined by elongation of metamorphic minerals, lies within S_{1c}. Planar to tight, asymmetric folds (F_{1c}) plunge southwest, through south to east, varying in plunge on an outcrop scale. They may be sheath structures, but this has yet to be demonstrated. The sense of movement implied by the distribution of F_{1c} fold symmetry groups is everywhere consistent with northwest-directed shear parallel to the L_{1c} mineral lineation, and with rare C-S fabrics and asymmetric epidote augen in metavolcanic rocks. Closely associated folds deform S_{1c} (Fig. 20B) and possess a weakly developed, spaced, axial-planar fracture cleavage and a rare axis-parallel intersection lineation. However, they are broadly coaxial and coplanar with those of F_{1c} and suggest a similar sense of shear (Fig. 20B). Hence they are attributed to a progression of the same tectonic episode. The overall style of deformation of the Folly River Formation is consistent with dextral transpression (Nance and Murphy, 1990; Fig. 20).

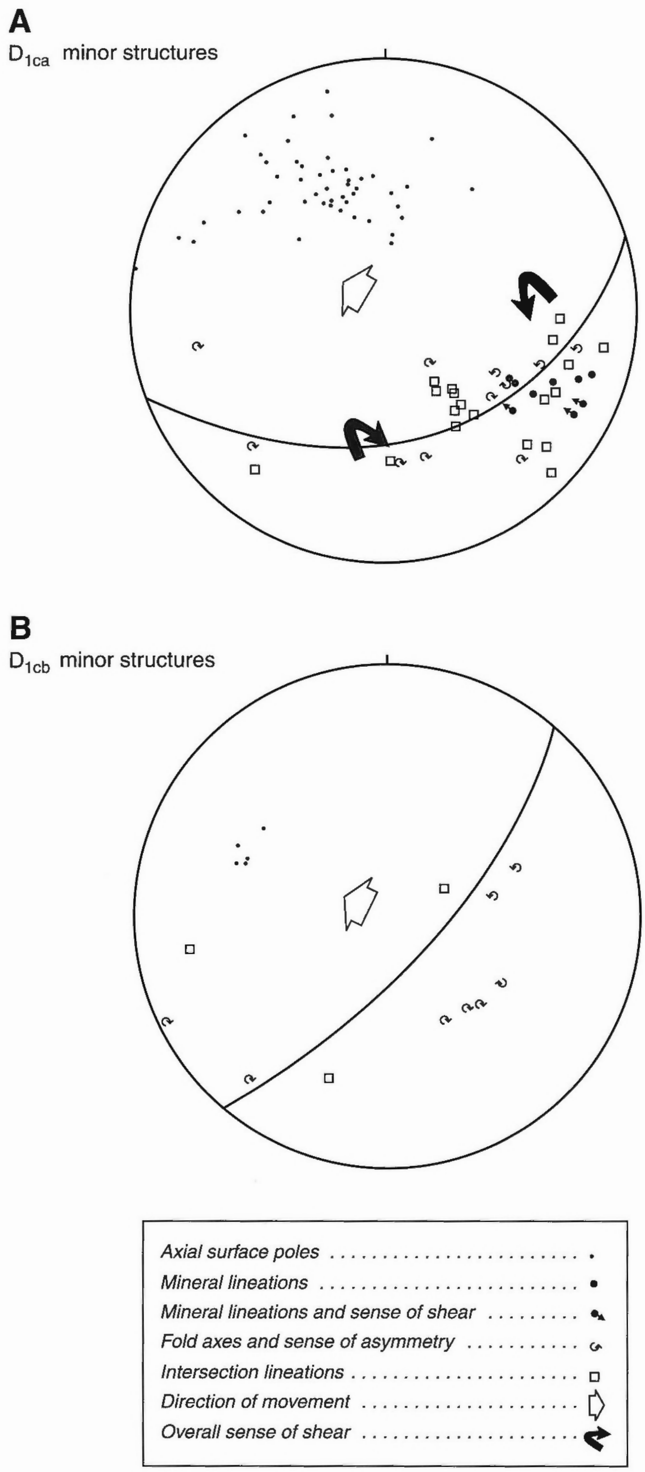


Figure 20. *Synoptic equal-area stereographic projection of **A)** D_{1ca} and **B)** D_{1cb} structural data from the Folly River Formation. Large arrows indicate the overall sense of asymmetry of F_{1c} folds. The sense of shear indicators imply oblique slip with reverse and dextral components toward the west-northwest.*

Jeffers Group

The Jeffers Group is characterized by subhorizontal penetrative cleavage, which on Jeffers Brook (west of the map area), predates intrusion of the ca. 605 Ma Jeffers Brook diorite. Along Jeffers Brook, southward dipping cleavage appears to crosscut earlier subhorizontal folding and forms wide zones (tens of metres) of intense foliation, commonly with steeply dipping kink bands, which strike subparallel to the cleavage. A shallow, south-plunging, chlorite lineation is associated with subparallel calcite veining. Mesoscopic kinematic indicators are lacking but microstructures suggest a northerly direction of movement. Elsewhere around the diorite, kink bands formed by offset of east-trending joints and metre-scale folding occur adjacent to an east-trending Carboniferous fault, as well as a shallow, north-dipping zone of brittle reverse faulting. The Jeffers Group west of the map area therefore displays predominantly Late Precambrian deformation possibly overprinted by Carboniferous deformation (Pe-Piper and Piper, 1989). On East River, the structural style is similar, with shallow-dipping, north-northeast-striking, penetrative cleavage defined by sericite and chlorite.

Based on data collected west of the mapped area, the pervasive, shallowly dipping cleavage is attributed to late Precambrian thrusting. Localized zones of intense foliation are common indicators of thrust movement (Bosworth, 1984; Nickelson, 1986). Either cleavage duplexes or anastomosing shear zones may be defined by a change in cleavage dip or a discrete zone of more intense cleavage, both of which were observed in the Jeffers Group. Bedding-parallel veins are either the result of bedding slip movement during compressional thickening (Fitches et al., 1986) or overthrusting (McClay, 1992), in which veins form as a result of extension and fluid overpressures. Kink and/or crenulation bands are common features in thrust belts and have been attributed to ramp formation (Beutner et al., 1988), the sense of which depends on their orientation and sense of kinking. The fracturing associated with cleavage kinking suggests ductile behaviour as evident from warping and/or flexing of the cleavage. The cleavage probably formed within a thrust belt, wherein the cleavage developed in successive thrust sheets close to the level of décollement (Mitra and Elliot, 1980; Mitra and Yonkee, 1985), as suggested by a relatively ductile environment, lack of folding and brittle structures (breccia and extensional veining), and the occurrence of crenulation zones and a stretching lineation (*see* Fig. 9, Ramsay, 1981).

Alternative models for the development of subhorizontal cleavage include a genetic association with diapirs (Dixon, 1975), core complexes (e.g. Davis, 1980), or listric faults that curve into deeper, ductile zones of subhorizontal deformation (Wernicke, 1985). Formation of a synmetamorphic, subhorizontal cleavage and mineral lineation may occur during irrotational extension by emplacement of a pluton at depth (e.g. Reymer and Oertel, 1985). Evidence that might support this type of model for the Jeffers Group include the crosscutting character of the S_2 cleavage to earlier folding and the apparent lack of mesoscale kinematic indicators.

Warwick Mountain and Dalhousie Mountain formations

The age of deformation in the Warwick Mountain and Dalhousie Mountain formations is not well constrained, but deformation of the Dalhousie Mountain Formation predates the ca. 605 Ma age (R. Doig, pers. comm., 1993) of a granite that postkinematically intruded it. The age of deformation of the Warwick Mountain Formation is thought to be similar because of its lithological and structural similarity to the Jeffers Group, also characterized by gently dipping bedding and a penetrative, bedding-parallel cleavage.

The Dalhousie Mountain Formation, although similar lithologically to the Jeffers Group and Warwick Mountain Formation, is distinguished by the lack of a regional penetrative cleavage. In the mudstone and finer grained tuffs, a weak cleavage is present, generally subparallel to bedding. The Dalhousie Mountain Formation appears correlative with the Keppoch Formation in the Antigonish Highlands. The general lack of cleavage in the Keppoch Formation has been attributed to its behavior as a rigid mass that maintained internal cohesion during deformation (Murphy et al., 1991).

Paleozoic deformation

Paleozoic deformation in the eastern Cobequid Highlands is predominantly Carboniferous. South of the Rockland Brook Fault, thrusts and isoclinal folds postdate the Nuttby Formation and Fountain Lake Group (syn- or post-Namurian; Fig. 1). A regional syncline whose axial trace has a east-northeast trend has a profound influence on the outcrop pattern of the Bass River Complex. This fold trace can be mapped into the early Carboniferous lithologies to the east, where it forms one of a set of en echelon east-northeast-trending folds that have a dextral orientation with respect to the Rockland Brook Fault. These folds are attributed to dextral motion on the Rockland Brook Fault some time in the Carboniferous.

North of the Rockland Brook Fault, intense deformation is confined to ductile shear zones and associated minor thrusts (Waldron et al., 1989), but evidence of regional deformation is given by an angular unconformity between the Fountain Lake Group and Westphalian rocks (Donohoe and Wallace, 1982a, b, c, d), and by the occurrence of north-northwest-dipping thrusts and reverse faults. The best example of thrusting north of the Rockland Brook Fault occurs on Miller Brook where the Warwick Mountain Formation is thrust upon the Fountain Lake Group.

Rockland Brook Fault

A detailed study of the Rockland Brook Fault was undertaken as an example of Devonian–Carboniferous deformation processes (Miller et al., 1989, 1995) and the most important results from these studies are summarized below. The Rockland Brook Fault, approximately 80 km in length, forms a major, Late Paleozoic strike-slip structure up to 1 km wide in the central Cobequid Highlands of Nova Scotia (Fig. 21). The fault is parallel and adjacent to the Cobequid Fault (Donohoe and Wallace, 1980, 1985), a part of the east-trending Cobequid–Chedabucto system (Minas Geofracture, Keppie (1982)) which separates the

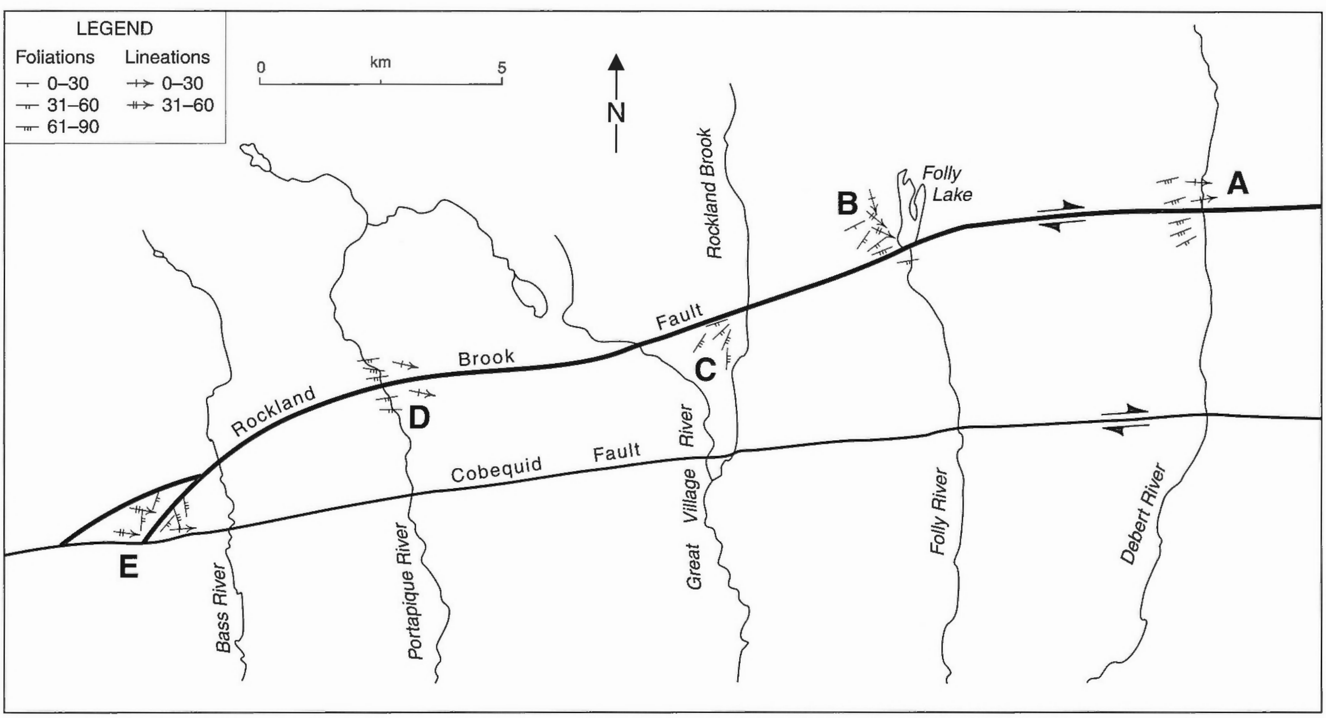


*Figure 21. Summary map emphasizing the central portion of the Rockland Brook Fault (*see also *Fig. 1; Murphy et al., 2000) and the orientations of structural data along its surface trace. Stereoplots of regions A, B, C, D, and E are shown in Figure 22.*

Avalon and Meguma terranes (Williams and Hatcher, 1983). A major portion of the movement history of the Rockland Brook Fault is related to (and probably a product of) movement on the Cobequid Fault, which has accommodated major displacements (e.g. Eisbacher, 1969; Keppie, 1982; Mawer and White, 1987); however, the fault has been interpreted to have an earlier history of movement (termed the "ancestral Rockland Brook Fault", Miller et al. (1995)) that formed Precambrian terrane boundary within the Avalon composite terrane that separated the proposed Bass and Cobequid terranes (Keppie, 1985).

Donohoe and Wallace (1980, 1982a, b, c, d, 1985) described the fault as a broad zone of mylonitization in the central Cobequid Highlands, which they attributed to predominantly strike-slip ductile shear of mid-Carboniferous age. This zone followed but did not define the northern margin of the Bass River Complex and generally separated this polydeformed Precambrian basement-cover sequence to the south from undeformed Devonian–Carboniferous plutons and volcanic rocks and Silurian volcanogenic sedimentary rocks to the north (Fig. 1). At its eastern extremity, where the zone was considered to merge with the Cobequid Fault, they positioned the fault south of the Mount Thom Complex with which parts of the Bass River Complex have been correlated (Donohoe and Cullen, 1983). Such a path would imply a minimum dextral displacement of about 40 km between the two complexes. This segment does not mark the course of a ductile shear zone, but that of a brittle fault. Murphy et al. (1988) proposed a different extension of the Rockland Brook Fault, which Yeo (1985) and Yeo and Ruixiang (1987) found to terminate within the Late Namurian to Early Westphalian Riversdale Group east of the Salmon River pluton (Fig. 1). Both traces are here considered to define paths relevant to the Rockland Brook Fault, the former marking the trace of its Late Paleozoic (brittle) reactivation, and the latter the path of the ancestral Rockland Brook Fault. The Salmon River pluton stitches late Precambrian motion, and possibly also the main Late Paleozoic ductile phase of movement, and hence shows only late-stage brittle faulting. At its western extremity, Donohoe and Wallace (1982a, b, c, d) considered the fault to be truncated ("stitched") by the Pleasant Hills granite pluton (Fig. 1), but our observations show that the fault lies close to the southern margin of the pluton, deforms it, and merges with the Cobequid Fault, only to re-emerge further west to enclose the ca. 734 Ma Economy River Gneiss.

The eastern end of Rockland Brook Fault splays into two segments. The southern splay links Rockland Brook Fault to the Cobequid Fault and separates Carboniferous sedimentary rocks to the south from Precambrian and Carboniferous crystalline rocks to the north. The northern splay links a previously unnamed, northeast-trending fault near Six Mile Brook (Fig. 1) with Rockland Brook Fault (Murphy et al., 1988) and places basement rocks of the Mount Thom and Bass River complexes within the same (southern) fault block. Furthermore, the path separates deformed, late Proterozoic volcanic and volcanogenic sedimentary rocks to the northwest (Fig. 1) that are similar to those of the western Cobequid Highlands (Jeffers Group) from undeformed, late Proterozoic volcanic and volcanogenic sedimentary rocks to the southeast (Dalhousie Mountain Formation, unit 4c, Fig. 1; *see* Fig. 25) that are similar to those of the Antigonish Highlands to the east (Murphy et al., 1992). The change in trend was detected from a linear gravity anomaly which broadly coincides with the Rockland Brook Fault and bends in the area of the Salmon River pluton (Fig. 1). The path could not be unequivocally demonstrated on the basis of field evidence due to a lack of exposure, but other nearby major fault systems such as the Cobequid–Hollow fault system (Eisbacher, 1969; Yeo and Ruixiang, 1987) make a similar bend.

In straight segments of the Rockland Brook Fault, the effects of deformation are most obvious in Late Devonian plutons, which are otherwise undeformed. Mylonitic fabric defines a steep, parallel-walled ductile shear zone with alternating quartz-feldspar-rich and chlorite-epidote-rich bands containing a subhorizontal stretching lineation defined by preferred orientation of quartz, minor feldspar, rare mica, and flattened quartzofeldspathic augen, and, in more mafic phases, hornblende. However, major departures from this simple pattern occur in curved sections of the fault near the Pleasant Hills and Folly River plutons (Fig. 1).

In the Debert River area (subarea A, Fig. 22), where the fault is straight, mylonitic foliation trends 240°–270° and dips steeply to the south or southeast at angles greater than 60°. The associated mineral lineation plunges gently east or southeast at angles no greater than 30° (Fig. 21, 22). The orientation of these fabric elements indicates fault movement was primarily strike-slip with only a minor component of northwest-vergent oblique slip. In the vicinity of Folly Lake (subarea B, Fig. 22), fabric south of the fault trends 045° to 075° and dips steeply southeast. North of the fault, orientations progressively change from a steeply southeast-dipping 030° trend at the fault centre to a gently dipping 060° trend at its northern margin (Fig. 21). The fault zone has a trace concave to the south, and indicates development of a half-flower structure by dextral transpression within a restraining bend. West of Rockland Brook (subarea C, Fig. 22) clockwise rotation of fabric elements within the Great Village River Gneiss as the fault zone is approached indicates dextral shear (Fig. 22C). In Portapique River (subarea D, Fig. 22) where the fault trace is straight, axial surfaces and hinge lines of associated folds in the gneiss to mylonitic fabric in the Pleasant Hills pluton strike broadly east-west, dip steeply south, and contain a gently east-plunging (less than 30°) mineral lineation (Fig. 21, 22). West of Bass River (subarea E, Fig. 22) Rockland Brook Fault takes a sharp southwesterly swing to eventually merge with the Cobequid Fault (Fig. 21). This bend was sharp enough for dextral shear to produce low-angle northwest-directed thrusting so that mylonitic fabric trends between 010° and 020°, dips at moderate to gentle angles, and contains a down-dip mineral lineation that plunges southeast at 36°–50° (Fig. 22E).

The Rockland Brook Fault is one of the few northern Appalachian faults to show both east-west and northeast-southwest segments and the structures associated with the two orientations differ considerably. The east-west section from Economy River to the Salmon River pluton is a boundary separating polydeformed Precambrian rocks from Devonian–Carboniferous intrusive and sedimentary rocks.

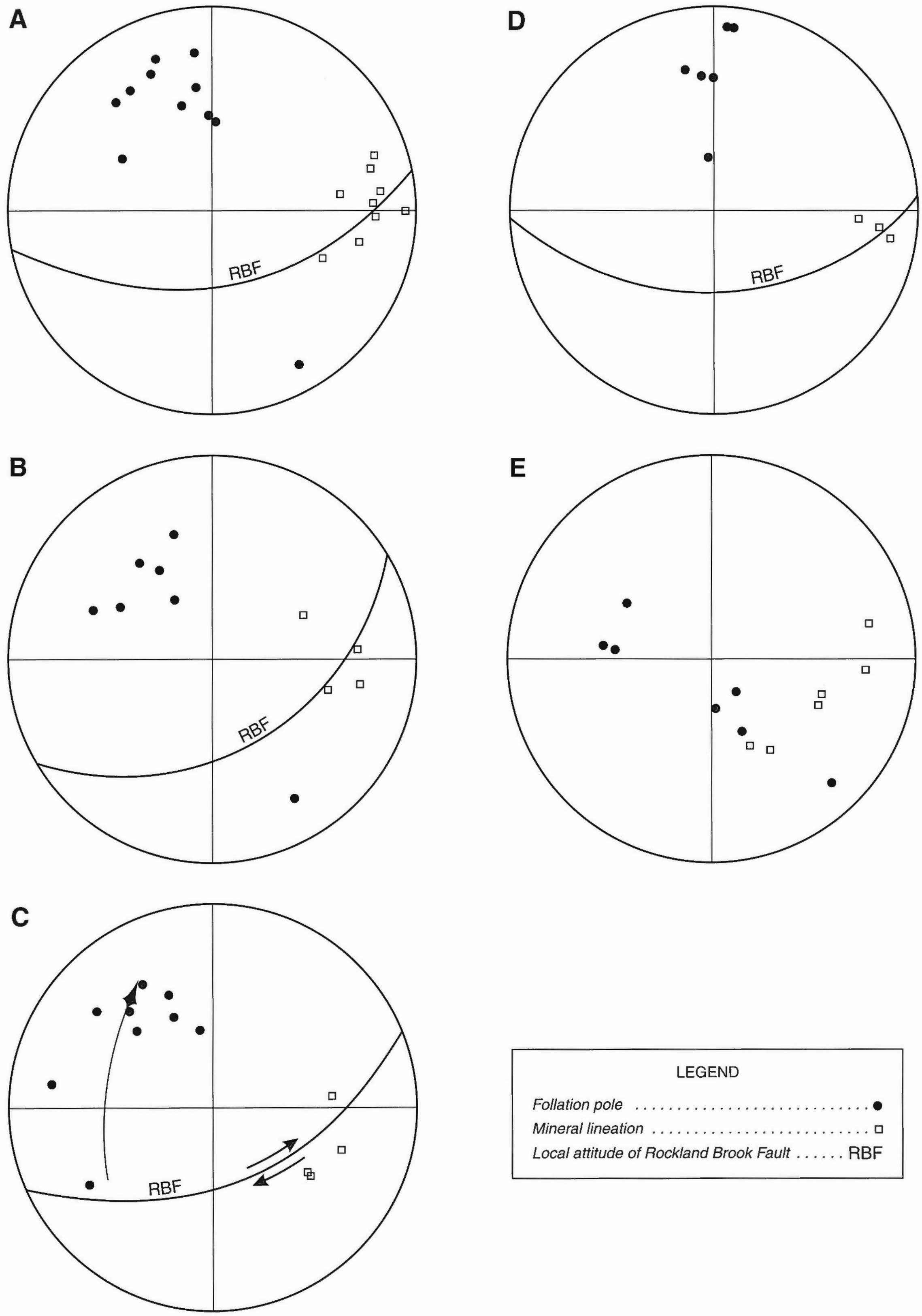


Figure 22. *Equal-area stereoplots showing representative structural data from sub-areas A through E on the east-west segment of the Rockland Brook Fault (Miller et al., 1995).*

The fault zone centre is marked by 10–25 m of ultramylonite and deformation directly attributable to the fault, and extends up to a hundred metres from the centre in the previously undeformed rocks north of the fault zone. Deformation may extend for similar distances south of the fault but it is difficult to positively identify fabrics because of the pre-existing, polyphase structural history. Minor fold vergence, S-C fabrics, and asymmetric augen (sigma and delta structures of Passchier and Simpson (1986)) indicate dextral movement, although the indicators of opposite sense of shear are common. In most cases these sinistral indicators can be explained by antithetic shearing, local folding, or complex ductile flow within an overall dextral regime. In most localities quartz is ductilely deformed whereas the feldspar crystals remained relatively rigid, suggesting temperatures between 350°C and 600°C (Tullis and Yund, 1977). In subareas A and D (Fig. 21, 22) both minerals are ductilely deformed.

Along the northeast-trending segment of the fault within the Salmon River pluton, faulting is largely brittle and offsets along individual fractures minimal. Locally, sedimentary bedding can be seen at a high angle to the fault trace, suggesting strike-slip displacement along this section of the fault was less significant than on the east-west portion. Significant post-Carboniferous deformation in the Six Mile Brook area is precluded by tilted, but otherwise undeformed beds of the Late Namurian "Millsville Conglomerate" (Gillis, 1964) that overstep the predominant fracture trend in the Dalhousie Mountain Formation. Large boulders in this fanglomerate resemble the Salmon River and other Late Devonian plutons of the Cobequid Highlands which are strongly deformed along the east-west segment of the Rockland Brook Fault. Hence Late Paleozoic dextral movement along this segment of the fault appears been minimal. However, the fault separates a Late Precambrian volcanic-sedimentary succession with a pervasive flat-lying Precambrian cleavage (Pe-Piper and Piper, 1987) to the northwest, from a lithologically similar but undeformed Late Precambrian sequence, and broadly corresponds to the northern margin of a linear positive Bouguer anomaly (Keppie, 1982) that suggests the basement is upthrown on its southeastern side. The northeast-trending segment of the fault therefore may be a site of a significant late Precambrian southeast-dipping thrust (ancestral Rockland Brook Fault) which separates the relatively undeformed Dalhousie Mountain Formation in the hanging wall from underlying cleaved volcanic rocks of the Warwick Mountain Formation, or, more regionally, where relatively undeformed Late Precambrian rocks in the Antigonish Highlands overthrust intensely deformed Precambrian rocks of the Cobequid Highlands.

Along the east-west portion of the fault, the ancestral Rockland Brook Fault may have provided a structural weakness that accommodated later displacements during the emplacement of the Meguma terrane. This movement rotated the fault's ancestral northeast trend into an east-west orientation. A similar scenario was implied by Keppie (1982) for the development of the Cobequid–Hollow fault system. Further movement is thought to be responsible for the development links to the Cobequid Fault such as those proposed by Donohoe and Wallace (1982a, b, c, d). These links appear to be late-stage brittle faults and may imply that the Salmon River pluton stitches Late Paleozoic ductile shear on the Rockland Brook Fault. Cessation of major movement on the

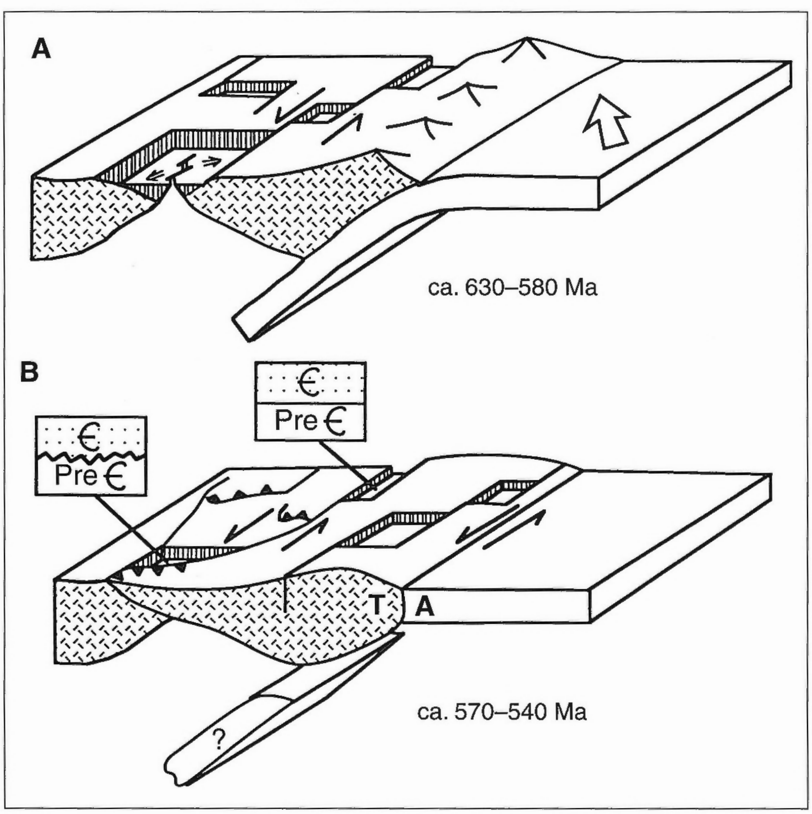


Figure 23.

*Regional tectonic environment for the Avalon composite terrane in the Late Proterozoic showing, **A)** 630–580 Ma coeval subduction, arc magmatism dominated by volcanic and plutonic rocks and intra-arc rifts which contain thick turbidite successions in addition to volcanic rocks, and **B)** termination of subduction and development of a transform margin.*

Rockland Brook Fault is of uncertain timing. As noted above, right-stepping en echelon folds affect late Visean to early Namurian rocks between the Cobequid and Londonderry faults implying movement on the Cobequid Fault and, hence, related movement on the Rockland Brook Fault continued until at least the early Namurian. Rare mafic dykes that cut the mylonitic fabric on Bass River are likely Triassic.

SUMMARY AND SYNTHESIS

Basement rocks

The eastern Cobequid Highlands lies within the Avalon composite terrane, which was affected by a global-scale series of Late Proterozoic orogenic events generally referred to as "Pan-African". Late Precambrian reconstructions (e.g. Bond et al., 1984; Taylor and Strachan, 1990; Dalziel, 1991; Hoffman, 1991) and paleomagnetic data (Johnson and Van der Voo, 1986) suggest that the arc-related activity that typifies the Avalon composite terrane (Fig. 23) occurred on the periphery of Gondwana (e.g. Fig. 24).

The age and tectonic history of basement rocks of the Avalon composite terrane are critical to palinspastic reconstruction of the terrane because they help to identify the continental margin along which Avalonian rocks were developed and constrain the accretionary history of the Appalachian Orogen (Williams, 1979; Barr and Raeside, 1989; Keppie et al., 1991).

Uranium-lead dating in the Cobequid Highlands, together with recent data from New Brunswick have fundamentally changed our understanding of Avalonian tectonostratigraphy in the Maritime provinces. The data show that late Proterozoic gneiss units and platformal sedimentary rocks represented by the Bass River and Mount Thom complexes in the Cobequid Highlands and the Brookville Gneiss in New Brunswick, formerly thought to underlie the volcanic-sedimentary supracrustal rocks, are in fact coeval with them. They yield ca. 600 Ma protolith ages and are similar in age to the volcanic-sedimentary sequences (Bevier et al., 1990; Doig et al., 1991a, b). Hence, their genesis should be considered part of the ca. 630–570 Ma orogenic cycle.

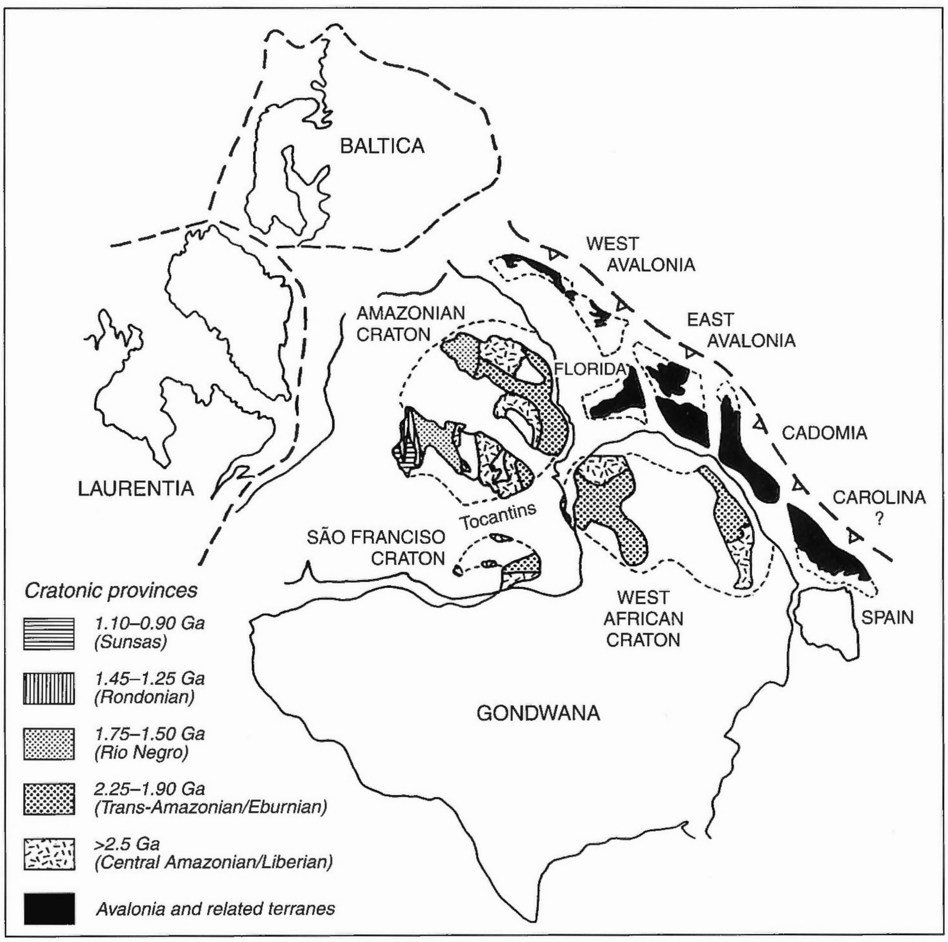


Figure 24. *Examples of Late Proterozoic reconstructions showing a possible configuration of a Late Proterozoic supercontinent,* (after *Murphy and Nance (1989),* modified from *Dalziel (1991)), showing the proposed position of the Avalonian belt* (see *Van der Voo, 1988; Murphy and Nance, 1989).*

The oldest confirmed protolith age for rocks in the eastern Cobequid Highlands is 734 Ma for the Economy River Gneiss. Similar ages between 820–660 Ma occur in other parts of the Avalon composite terrane of North America, the probably correlative Cadomian belt of western Europe, and Pan-African belts in Africa, Arabia, and South America. In North America these ages include the ca. 730 Ma Fishbrook Gneiss of Massachusetts (Olszewski, 1980), the ca. 676 volcanic rocks of the Stirling block, southeastern Cape Breton Island (Barr et al., 1990), the 763 Ma Burin mafic volcanic rocks (Krogh et al., 1988) (which have been correlated with the ca. 784 ± 10 Ma (Clauer, 1976) Bou Azzer ophiolite in Morocco), the 683 Ma Connaigre Bay Group rhyolite (Swinden and Hunt, 1991) of southeastern Newfoundland, and the ca. 686 Ma Roti granite of southwestern Newfoundland (O'Brien et al., 1990). In western Europe, similar ages include the ca. 700 Ma orthogneiss of the Rosslare complex, Ireland (R. Doig, unpub. data, 1990), the ca. 677 Ma deformed granite of the Malvern plutonic complex (Tucker and Pharoah, 1991), the ca. 702 Ma granitoid rocks of the Stanner–Hunter Complex (Patchett et al., 1980) and the ca. 700 Ma quartz diorite of Guernsey (Dallmeyer et al., 1991). Granitic rocks dated at 729 ± 8 Ma occur in the eastern Hoggar of northwest Africa and are interpreted as basement to ca. 600 Ma strata (Caby, 1987). Recent U-Pb and Nb-Sm geochronological data from the west African trans-Saharan orogenic belt of Mali (Caby et al., 1989) indicate important arc development at both ca. 730 Ma and ca. 635 Ma. Igneous ages of ca. 820–660 Ma are also common in the Arabian Shield (Pallister et al., 1988) where they represent a long-lived tectonothermal event (early Pan-African, Kroner et al. (1990)) which preceded voluminous, relatively undeformed ca. 630–600 Ma Cordilleran-type volcanic-sedimentary successions (Stern et al., 1984). Evidence of a ca. 800–700 Ma tectonothermal event is preserved in South America in the Rio Preto fold belt (Davison and Santos, 1989), Tocantins Province (Pimental and Fuck, 1992), and Montiqueira–Ribera belt (Bernasconi, 1987) to the east of the Amazonian Craton, which may be a continuation of the Trans-Saharan orogenic belt of West Africa. Late Proterozoic events are not well represented in cratonic North America, implying that the Avalon composite terrane lay far away from the craton at this time.

Constraints on the Late Proterozoic paleogeography of the Avalon composite terrane in Nova Scotia is provided by a U-Pb study of detrital zircons (Keppie and Krogh, 1990) from clastic rocks of the Georgeville Group, Antigonish Highlands which are correlated with the Folly River Formation of the eastern Cobequid Highlands (Murphy et al., 1992). These data suggest that the Avalon composite terrane was built on basement of Middle–Late Proterozoic (Grenvillian), Archean, and Early Proterozoic age. Grenvillian ages are not well documented in West Africa orogens, but are present in the southern Amazonian Craton of South America (Teixeira et al., 1989) which is positioned adjacent to the Late Proterozoic West African Craton on most reconstructions of the late Precambrian (e.g. Bond et al., 1984; Dalziel, 1991; Hoffman, 1991). These data suggest that the Avalon composite terrane of mainland Nova Scotia has the strongest genetic links with the "Pan-African" orogens of eastern South America.

In reconstructions of the late Precambrian (e.g. Bond et al., 1984; Dalziel, 1991), the Arabian Shield and the Avalon composite terrane are both positioned at the northern extremity of a supercontinent (Fig. 23) where a transform plate boundary may have evolved from a convergent plate boundary in the late Precambrian (Murphy and Nance, 1989, 1991). The Arabian rocks have been interpreted as a collage of previously independent exotic terranes that were accreted by late Precambrian convergence and strike-slip activity, analogous to the Cordillera of western North America (e.g. Kroner, 1985). If so, the ca. 820–660 Ma rocks of the Arabian Shield and Avalon composite terrane may both represent vestiges of a widespread peri-Gondawanan subduction-related event. Late Precambrian strike-slip activity may be responsible for terrane dispersal in some areas (hence, the fragmentary evidence of the pre-630 Ma tectonothermal event in the Avalon composite terrane) and amalgamation in others (such as the collage of terranes preserved in the Arabian Shield).

The age and nature of basement to the platformal Gamble Brook Formation, which is intruded by the 734 Ma Economy River Gneiss, remains unknown. The Gamble Brook Formation has also been correlated with the Polletts Cove and McMillan Flowage formations and George River Group of Cape Breton Island, which are all dominated by thick platformal sedimentary successions (Keppie et al., 1991). However Barr and Raeside (1986, 1989) suggest that the Cape Breton Highlands are exotic with respect to the Avalon composite terrane. If this is the case, the above correlations are invalid. The poorly exposed Mount Thom Complex could be basement to the Gamble Brook Formation, but its age and relationship to other Late Proterozoic rocks of the Cobequid Highlands remain unclear.

"Avalonian" rocks

Age data, geochemical data, and field relationships indicate that both the Great Village River Gneiss and low-grade volcanic-sedimentary sequences on both sides of the Rockland Brook Fault are part of a complex ca. 630–580 Ma history of magmatism, deformation, and metamorphism in the Cobequid Highlands. The geochemistry of the Jeffers Group and Folly River Formation (e.g. Pe-Piper and Murphy, 1989; Pe-Piper and Piper, 1989) and cogenetic plutons (e.g. Pe-Piper, 1988) indicate that magmatism was arc related, and stratigraphy (Pe-Piper and Piper, 1987) is consistent with development of a rifted arc. The 590–580 Ma dates for the Great Village River orthogneiss suggest that it is an integral part of the Avalonian orogenic cycle. Plutonism of a similar age includes the 605 Ma mylonitic granite (location B of Table 2) which occurs within the ductile shear zone separating the Great Village River Gneiss and Gamble Brook Formation, the undeformed 610 Ma Debert River pluton, and the 575 Ma McCallum Settlement pluton. We interpret these plutons to be shallow equivalents of the Great Village River orthogneiss.

The absence of a ca. 560 Ma tectonothermal event and basin development in the Cobequid Highlands (and indeed the Antigonish Highlands) is problematic, given their widespread occurrence in most portions of the Avalon composite terrane in Atlantic Canada and New England. We speculate that during this period of wrench-related regional tectonothermal activity, that local faults were less active than elsewhere and did not provide the conduits for the ascent of magma to the surface or the generation of intracontinental sedimentary basins.

There is an apparent conflict between the age data for the Precambrian rocks and their field relationships. Two samples of the Great Village River Gneiss yield ca. 590–580 Ma ages. Yet fabrics in this gneiss are cut by mylonitic granitic rocks which yield older (ca. 605 Ma) ages. Epizonal plutons range from ca. 620–575 Ma. Taken together, arc magmatism appears to have spanned 40 Ma during which time three distinct phases of deformation took place. Although these phases can be separated on the basis of field relationships (Nance and Murphy, 1990), age data indicate that these deformations were broadly cogenetic and related to deformation of arc plutons and associated supracrustal rocks.

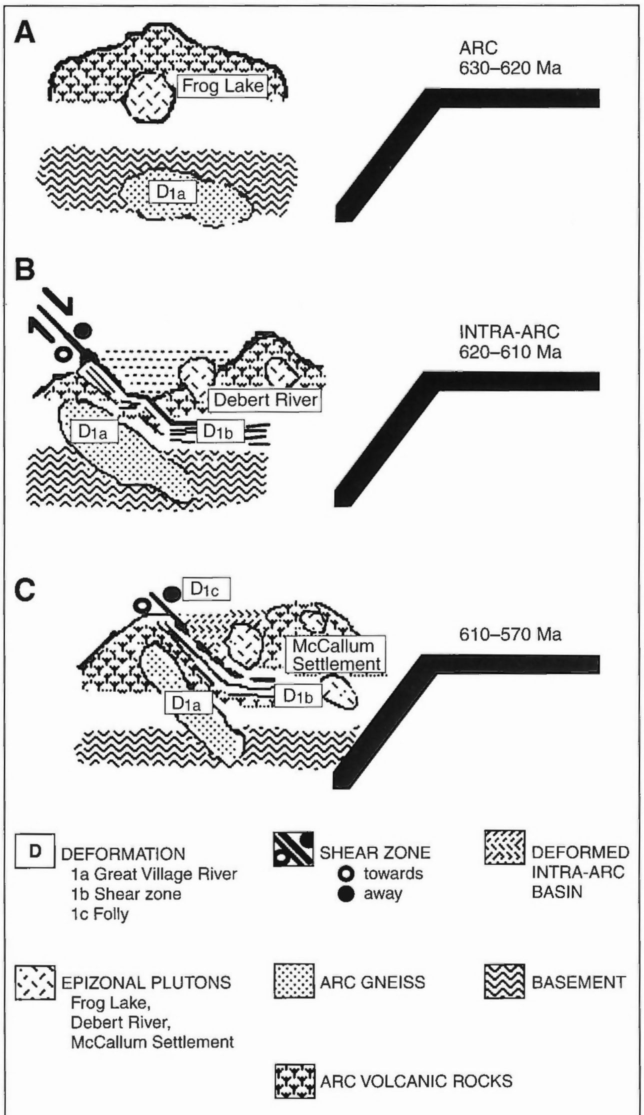


Figure 25. *Interpretative schematic cross-sections showing the sequential, but essentially penecontemporaneous, deformational events affecting the Late Precambrian rocks of the Cobequid Highlands.*

Clearly more U-Pb data is needed from localities where crosscutting relationships occur. A model attempting to reconcile the available age data and field relationships is outlined in Figure 25. The Great Village River Gneiss is inferred to comprise the plutonic equivalents of all epizonal plutonic and volcanic units, and is therefore assumed to range from 620–580 Ma.

The intrusion of the Great Village River Gneiss is interpreted to have accompanied the development of D_{1a} fabric at deeper levels, whereas emplacement of 605 Ma deformed granite into active ductile shear zones accompanied the development of D_{1b} mylonitic fabric, and at relatively shallow levels, the 609 Ma Debert River pluton is post-tectonic with respect to the D_{1c} deformation in the Folly River Formation, which unconformably overlies mylonitic fabric within Gamble Brook Formation.

The development of fabric related to each deformation must thus have occurred over a considerable period of time. It is also clear that they are essentially coeval. Geochronological data indicate that intrusion and metamorphism (D_{1a}) persisted to at least 580 Ma, whereas ductile shear and associated D_{1b} mylonitic deformation commenced along the Great Village River Gneiss–Gamble Brook Formation contact prior to 600 Ma. Since the development of D_{1a} and D_{1b} overlapped temporally in some localities, D_{1b} mylonitic fabric overprinted the D_{1a} metamorphic fabric. The lack of deformation within some epizonal Late Precambrian plutons can be attributed to emplacement at some distance from active shear zones and at relatively shallow depths.

An amphibolite sample from the Great Village River Gneiss, chemically indistinguishable from the Folly River Formation (Doig et al., 1991a), yielded an age of 589 Ma (Table 2). The association of penecontemporaneous amphibolite with volcanic-arc orthogneiss suggests an environment of intra-arc extension, a tectonic setting similar to of other late Precambrian low-grade volcanic-sedimentary sequences in the Antigonish and Cobequid highlands (Pe-Piper and Piper, 1989; Murphy et al., 1990). The presence of a southward-dipping normal and sinistral ductile shear zone (Nance and Murphy, 1990) with higher grade rocks to the north is thought to be the result of sinistral transtension opening a rift within which Folly River Formation was deposited (Fig. 25).

Schematic cross-sections of late Precambrian events in the eastern Cobequid Highlands (Fig. 25) sections show: A) 630–620 Ma arc magmatism with emplacement of volcanic rocks and plutons (e.g. Frog Lake) at shallow levels, and deformed gneiss with metamorphic fabric (D_{1a}) of the Great Village River Gneiss at deeper levels; B) 620–610 Ma development of an intra-arc rift into which the Folly River Formation was deposited with the Jeffers Group on the flanks and shear zones with sinistral and normal sense producing D_{1b} mylonitic fabric in slightly older arc and arc-rift magmatic rocks; C) 610–570 Ma continued plutonism producing younger components of the Great Village River Gneiss at depth, continued shearing along the Great Village River Gneiss–Gamble

Brook Formation contact and intensifying the mylonitic (D_{1b}) fabric, and intrusion of epizonal calc-alkaline plutons (e.g. Debert River, McCallum Settlement), and deformation producing D_{1c} fabric.

In southern New Brunswick, Barr and White (1989) prefer to separate the ca. 600 Ma Brookville Gneiss, a possible correlative of the Great Village River Gneiss, as a terrane distinct from the Avalon, rather than being a deeper portion of the Avalonian magmatic event. However, in the Cobequid Highlands, the field relationships between the Great Village River Gneiss and typical Avalonian epizonal rocks suggests that both of these units formed an important part of the Neoproterozoic tectonothermal activity in the Avalon composite terrane.

With the exception of the strong strike-slip component of motion, many of the features of the D_{1b} tectonic episode (such as the strong L-S tectonite fabric, the mylonitic décollement that separates a plutonic infrastructure from a metasedimentary suprastructure, and polyphase folding during protracted deformation), broadly parallel those of Cordilleran metamorphic core complexes (e.g. Davis, 1980). Many of the features of the Folly River Formation (such as the dominance of continental tholeiitic basalt, the presence of a dyke complex, hyaloclastite and jasperitic ironstone in the volcanic assemblage, and the close association of the volcanic rocks with pelitic rocks that resemble distal turbidite) suggest an extensional tectonic environment, attributed to an episode of local crustal extension that may have culminated in the transtensional tectonics of D_1.

Except for their sense of shear, structures in the Folly River Formation closely resemble those of the Gamble Brook Formation in form and orientation. As a result, Donohoe and Wallace (1980, 1985) and Cullen (1984) attributed deformation of the Folly River Formation to a second phase of deformation and equated its principal fabric to that of the Gamble Brook Formation. Kinematic contrasts and style distinctions between D_{1b} and D_{1c} make a direct correlation between these fabrics unlikely. Moreover, a mafic dyke, interpreted as a feeder to the volcanic rocks, cuts the D_{1b} fabric in the Gamble Brook Formation. S_{1c} appears to have resulted from a deformational event that affected only high structural levels of the Folly River Formation. The reversal of tectonic vectors implied by the D_{1c} kinematics could be attributed to the transpressional closure of this short-lived basin, or alternatively to kinematic decoupling at various structural levels. In summary, the kinematic analysis of the Bass River Complex is consistent with late Precambrian development and destruction of a small basin of limited time span. The limited extent of rifting is indicated by the continental affinity of the mafic volcanic rocks of the Folly River Formation.

These events in the Bass River Complex may be indicative of more regional tectonic events. The Antigonish and Cobequid highlands have been interpreted as a rift-related sequence within a volcanic arc possibly associated with the late Precambrian calc-alkalic Fourchu Group (Keppie et al., 1979; Murphy et al., 1990; Dostal et al., 1990) in southeastern Cape Breton Island of Nova Scotia. The Folly River Formation with its turbidite and Fe-Ti-rich continental tholeiitic volcanic rocks is similar to part of the upper sequence in the Antigonish Highlands (Clydesdale Formation) and may

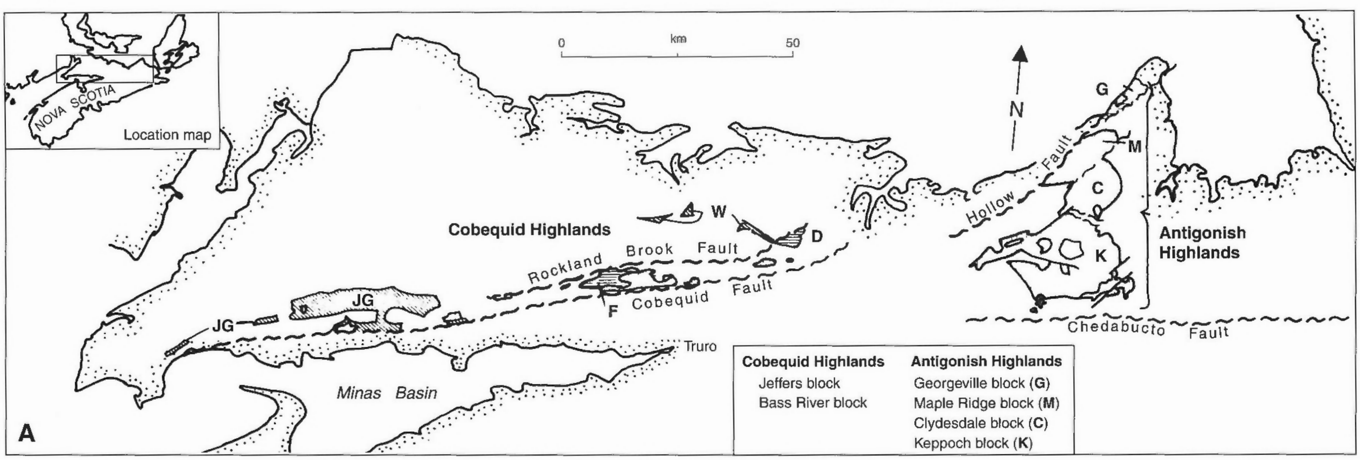


***Figure 26. A)** General geological map of the Precambrian rocks of northern mainland Nova Scotia (*modified from *Williams, 1979; Donohoe and Wallace, 1982; Pe-Piper and Piper, 1987; Murphy et al., 1991). In the Cobequid Highlands, the Jeffers block occurs north of the Rockland Brook Fault and consists of volcanic rocks of the Jeffers Group (JG) and the Warwick Mountain Formation (W). The Bass River block is south of the Rockland Brook Fault and consists of the Folly River Formation (F) and the Dalhousie Mountain Formation (D). **B)** Schematic stratigraphy of the late Pecambrian rocks in the Antigonish and Cobequid highlands of northern mainland Nova Scotia. Only the names of formations with volcanic rocks are given. Stratigraphic details are given by Murphy and Keppie (1987), Pe-Piper (1987), and Pe-Piper and Piper (1987, 1989). Correlation within the Antigonish Highlands are based on recognition of large-scale fining- and coarsening-up cycles (Murphy and Keppie, 1987).*

therefore also be genetically related to the volcanic-arc rift. The lack of true oceanic volcanic rocks in these basins indicates that the basins had limited width and did not involve the creation of significant oceanic crust.

Correlations with the Antigonish Highlands

Mapping in the Antigonish (Murphy et al., 1992) and Cobequid Highlands allow litho-stratigraphic correlation, one of the few examples of terranes within the Avalon composite terrane where such correlations can be demonstrated. These correlations imply considerable Paleozoic dispersion.

The Antigonish Highlands comprises four fault blocks (Fig. 26). In the southernmost Keppoch block, the Keppoch Formation consists of at least 3000 m of subaerial to submarine, interlayered felsic volcanic rocks, basaltic andesite, and interlayered turbidite towards the top of the formation. The basaltic rocks are within-plate continental tholeiite, the basaltic andesite are calc-alkalic, the felsic volcanic rocks have volcanic-arc affinities (Fig. 27) and were probably generated by anatexis of the crust (Murphy et al., 1990). In the central Clydesdale block the sequence is dominated by proximal and distal turbiditic sequences with thin interlayered mafic volcanic rocks which have within-plate, continental tholeiitic affinities (Fig. 27). The Maple Ridge block to the north has a thick turbidite succession. In the northernmost Georgeville block, calc-alkalic basaltic andesite and interlayered marble are overlain by distal turbidite and channel-fill conglomerate. Murphy and Keppie (1987) proposed the correlation shown in Figure 26 between formations of the Georgeville Group based on similarity of large-scale upward-coarsening and upward-fining trends in different fault blocks. There was a significant source of volcanic detritus to the north of the Antigonish Basin, whereas on the south side of the basin, sedimentary rocks interfingered with thick

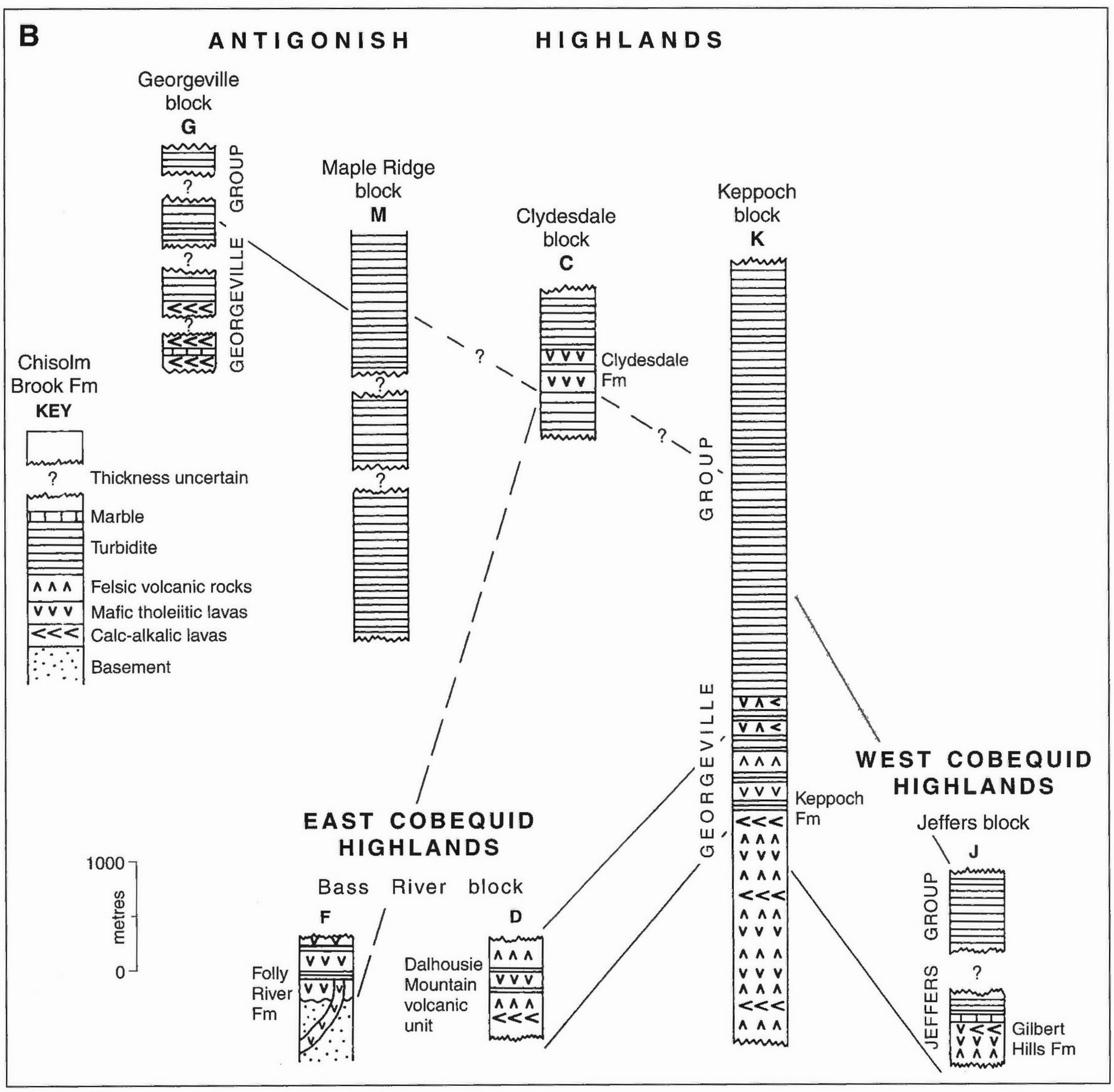


Figure 26 (*cont.*).

synchronous volcanic rocks. A lower age limit for deposition of the Georgeville Group is given by the 613 ± 5 Ma Pb-Pb age for detrital zircons in the Livingstone Cove Formation (Keppie and Krogh, 1990). Deformation of the Georgeville Group is dated as late Precambrian because structures and fabrics are truncated by Late Precambrian intrusive rocks and cleaved fragments derived from the Georgeville Group occur in unconformably overlying Cambro-Ordovician rocks (Murphy and Keppie, 1987). These Late Precambrian plutons include hornblende gabbro to diorite which yielded $^{40}Ar/^{39}Ar$ (hornblende) plateau ages ranging from 620 Ma to 600 Ma (Keppie et al., 1990). These data indicate that the Georgeville Group was deposited, deformed, and intruded in a narrow ca. 615–610 Ma time interval. Other granitoid rocks, are less precisely dated and have yielded ages of 535 ± 13 Ma (Rb-Sr whole rock isochron, R.F. Cormier, pers. comm. (1982)), and 604 ± 14 Ma (K-Ar age on muscovite, K. Wanless, pers. comm. (1980)).

The Keppoch Formation in the southern Antigonish Highlands and the Gilbert Hills Formation in the Cobequid Highlands are both northerly facing sequences consisting of interlayered basalt, basaltic andesite, and felsic volcanic rocks with minor volcanogenic sedimentary rocks, overlain by thick turbidite. The Warwick Mountain Formation and the Dalhousie Mountain Formation include similar volcanic rocks with interbedded turbidite. Such sequences are rare in the Avalon terrane; they are not reported in southern New Brunswick or in Cape Breton Island. These sequences are therefore correlated, as shown in Figure 26. In the Antigonish Highlands, the Clydesdale Formation was sediment starved, but is thought to correlate with the upper part of the thick sequence of turbidite in the Maple Ridge block to the north (Fig. 26). In the Cobequid Highlands, the Folly River Formation is geochemically and lithologically similar to the Clydesdale Formation; however, the lack of proximal turbidite in the Folly River Formation suggests a more distal environment of deposition. No basement is seen to the Clydesdale Formation in the Clydesdale block, but the geochemistry of the volcanic rocks indicates the presence of continental crust (Murphy et al., 1991), possibly equivalent to the Gamble Brook Formation beneath the Folly River Formation. The Folly River Formation lies within the same fault block as the

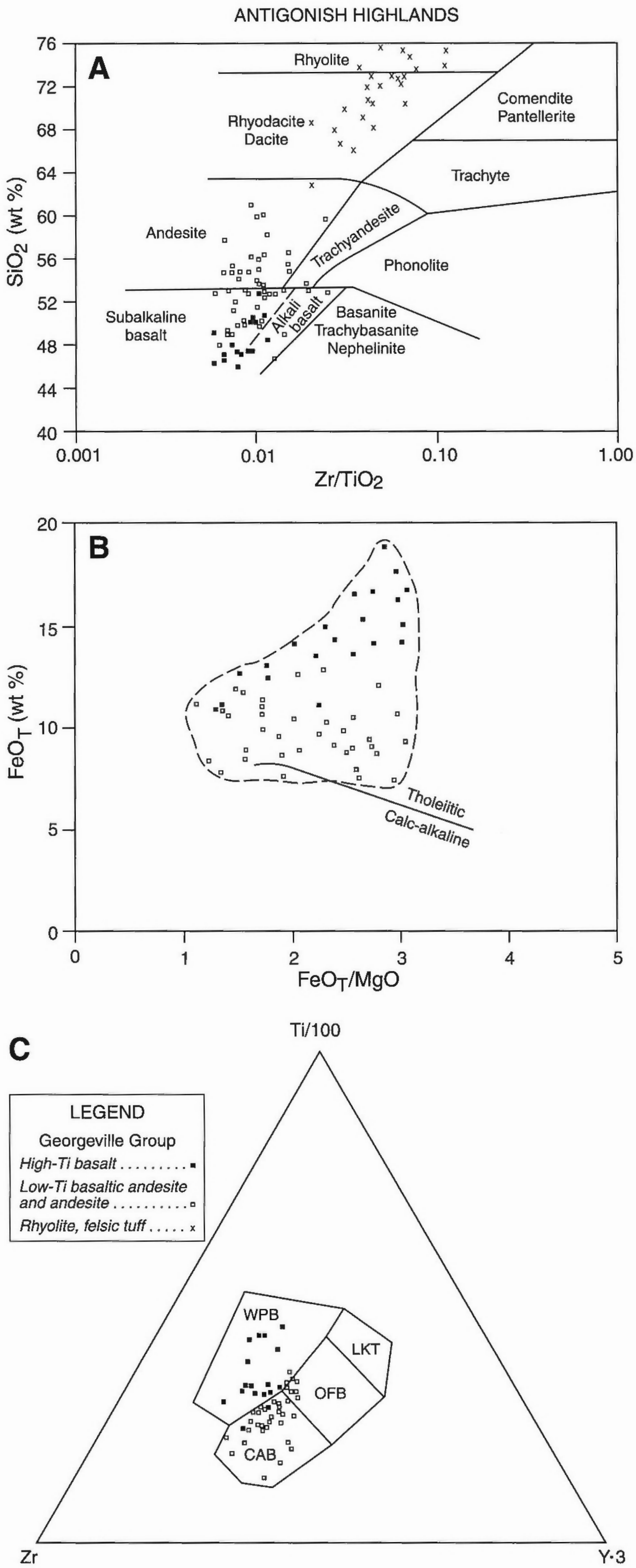


*Figure 27. Plots of selected elements from Neoproterozoic volcanic rocks of the Antigonish Highlands to illustrate their geochemical character. Open symbols indicate calc-alkalic basaltic andesite; solid symbols for within-plate continental tholeiitic basalt. For further details see Murphy et al. (1991). **A)** SiO_2 versus Zr/TiO_2 (fields after Floyd and Winchester, 1975) **B)** FeO_T versus FeO_T/MgO for mafic rocks (tholeiitic–calc-alkalic boundary after Miyashiro, 1974) **C)** Ti/100-Y-Zr-3 (tectonic environments after Pearce and Cann, 1973). WPB = within-plate basalt, CAB = calc-alkalic basalt, MORB = mid-ocean ridge basalt, LKT = low-K tholeiitic rocks.*

Dalhousie Mountain Formation, suggesting a facies transition between them similar to that proposed between the Keppoch and Clydesdale blocks (Murphy and Keppie, 1987). The extrusion of basaltic magmas in the Clydesdale and the Folly River formations is probably broadly coeval with the intrusion of dykes of similar composition in the Gilbert Hills Formation and Keppoch Formation.

If these correlations between the Cobequid Highlands and Antigonish Highlands are correct, then any simple reconstruction (Fig. 28) requires the Keppoch Formation, Dalhousie Mountain Formation, and Gilbert Hills Formation to be deposited on the same side of a subsiding basin, with the Clydesdale and Folly River formations in a more distal part of the basin. The Cobequid Highlands could lie either north or south of the Antigonish Highlands, as shown in Figure 28. If they lay south of the Antigonish Highlands, then dextral motion on the east-west Cobequid Fault followed by dextral motion on northeast-trending faults such as the Hollow and Rockland Brook faults (e.g. Keppie, 1982; Yeo and Ruixiang, 1987) and/or northerly directed thrusting (e.g. Nance, 1986) could achieve the present configuration (Fig. 28A). On the other hand, if the Cobequid blocks initially lay north of the Antigonish Highlands, then major sinistral motion on the Rockland Brook and Hollow faults (and their continuation in the Cobequid Fault) would be required, followed by minor dextral motions (Fig. 28).

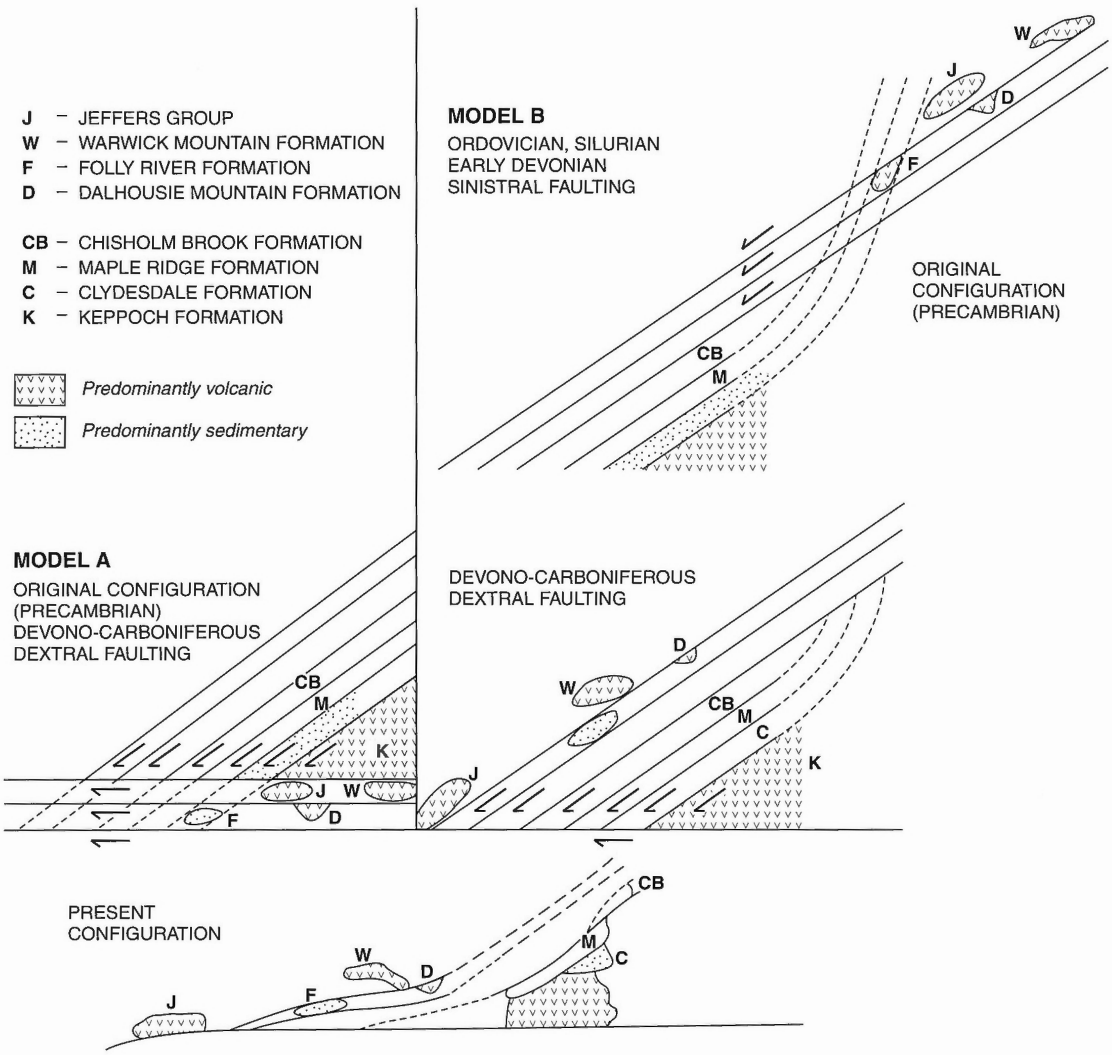


Figure 28. Possible schematic palinspastic reconstructions of the late Precambrian volcanic rocks of northern mainland Nova Scotia, showing possible dispersion resulting from movement on the Cobequid, Hollow, and Rockland Brook faults. Model A shows Cobequid Highlands rocks originally lying to southwest of Antigonish Highlands rocks and emplaced principally by dextral slip on the Minas geofracture, followed by dextral slip on the Rockland Brook and Hollow faults. Model B shows Cobequid Highlands rocks originally lying to northeast of Antigonish Highlands rocks and emplaced principally by sinistral slip along the Rockland Brook and Hollow faults.

Dextral movements on the Cobequid and Hollow faults occurred during Devonian–Carboniferous time (Keppie and Dallmeyer, 1987; Mawer and White, 1987; Yeo and Ruixiang, 1987), consistent with model A shown in Figure 28; however, an earlier period of sinistral shear would be consistent with model B. The east-west faults bounding the blocks in the Cobequid Highlands swing north of the Antigonish Highlands, suggesting that they originated to the north of the present Antigonish Highlands. Keppie and Murphy (1988) proposed sinistral movements on the Hollow Fault during the Ordovician and Silurian, consistent with sinistral accretion of Avalon to Laurentia. Transpressive deformation in the late Precambrian was both sinistral and dextral on northeast-trending faults (Nance and Murphy, 1988; Murphy et al., 1991). Thus, it is unclear which model is correct. Regardless of the original configuration, the presence of continental crust beneath the late Precambrian basin suggests limited rifting. The mixed volcanic arc-rift affinity of the magmatism and the presence of abundant volcaniclastic sedimentary rocks indicate rifting of a magmatic arc. Murphy and Keppie (1987) and Nance and Murphy (1988) proposed that this rifting was produced by strike-slip movements suggesting that the rift basin represents a pull-apart basin. Keppie (1982, 1985) and Nance (1987) suggested that oblique subduction and translation along faults in the volcanic arc were associated with late Precambrian orogeny. In general, these models result in the opening and closing of strike-slip basins, and explain the heterogeneous intensity of late Precambrian deformation in the Avalon composite terrane. For example in Newfoundland, late Precambrian deformation is relatively mild and only a disconformity separates late Precambrian and Cambrian sequences (Williams, 1979; O'Brien et al., 1983, 1990).

If the above correlations are valid, then the D_{1b} sinistral deformation in the Bass River Complex may be genetically associated with the development of a volcanic-arc rift in the Antigonish Highlands and basin closure may also have been accomplished by strike-slip faulting. All late Precambrian sequences are post-tectonically intruded by appinitic and granitic plutons and in the Antigonish Highlands the sequence is unconformably overlain by Early Cambrian continental to shallow marine sedimentary and volcanic rocks (Murphy et al., 1991) indicating that this basin closed in the late Precambrian.

Paleozoic rocks

The extent and quality of outcrop of Early to Middle Paleozoic rocks is very poor in the Cobequid Highlands. They occur only in narrow fault-bounded slices immediately south of the Rockland Brook Fault. Devono-Carboniferous rocks consist of voluminous plutonic and volcanic rocks. Available data indicate that the bulk of this activity took place at ca. 360 Ma, although further age dating is necessary to confirm this. Available radiometric ages imply that the plutonism is penecontemporaneous with the younger phases of plutonism within the Meguma terrane (*see* Pe-Piper et al., 1989).

Plutonism in the eastern Cobequid Highlands is typically bimodal and within plate. Mafic magmatic rocks resemble continental tholeiite. Felsic magmas resemble within-plate and A types. There appears to have been synchronous intrusion of mafic and felsic magmas throughout the entire length of the highlands. Textures, xenoliths, and close association of plutonic and volcanic rocks suggest that the plutons crystallized at a high structural level, probably in part intruding extrusive counterparts of the Fountain Lake Group.

The abundance of Late Devonian magmatism in the Cobequid Highlands compared with other parts of the Avalon composite terrane suggests a local structural control. The ca. 360 Ma ages for the plutonic rocks are similar in age to the $^{40}Ar/^{39}Ar$ plateau ages from muscovite crystals in shear zones adjacent to the Avalon–Meguma terrane boundary, attributed to ca. 360 Ma dextral motion along the terrane boundary. The great abundance of tholeiitic mafic products suggests a relationship with extension, presumably Late Devonian–Carboniferous extension of the Magdalen Basin, which resulted in a thick underplated gabbroic crustal layer (Marillier and Verhoef, 1989). The A-type geochemical affinities of the granitoid plutons and rhyolite (Pe-Piper, 1991) probably result from melting of relatively dry lower crust. Several lines of evidence suggest that the Cobequid–Chedabucto fault system was probably the main path for magma ascent. It is larger than any of the others in terms of its present offset and width of the deformation zone. Both outcrop studies and aeromagnetic data (Piper et al., 1993) indicate that mafic magma emplacement in the western Cobequid Highlands was concentrated along this fault zone and the Kirkhill Fault a few kilometres to the north. Two phases of deformation accompanying magma emplacement are recognized in the plutons of the western Cobequid Highlands. Early magmas intruded ductile rocks during left-lateral oblique thrust movements. A second stage of right-lateral oblique-slip faulting accommodated increasing uplift of the plutons. At this time coarse granite was emplaced in the most elevated region through a process of roof lifting. Crosscutting, late-stage porphyries, granitic clasts in marginal basins cut by granitic dykes, and superposition of brittle on ductile structures all indicate rapid uplift of the plutons. Pluton emplacement was not the result of extension in releasing bends during transcurrent shear. Rather, flower-structure, high-angle faults acted as magma conduits and space was created by wall-rock deformation and hanging-wall expansion on thrust faults at depth, and by roof lifting as the plutons were uplifted to higher structural levels (Pe-Piper and Koukouvelas, 1994).

Regional considerations indicate that emplacement and evolution of Devonian–Carboniferous igneous rocks of the Avalon composite terrane of Nova Scotia was strongly influenced by motion on major faults associated with dextral emplacement of the Meguma terrane (Dallmeyer and Keppie, 1987; Keppie and Dallmeyer, 1987) which produced penecontemporaneous pull-apart basins, thrusts, uplifts, and depressions that strongly influenced sedimentation and magmatism (Ryan et al., 1987).

Structural analysis of the eastern Cobequid Highlands suggest another important phase of deformation may have occurred later in the Carboniferous. This deformation produced east-northeast-trending, en echelon folds and may also be associated with dextral shear on east-west Visean–Namurian faults. This deformation may be a local representative of regional episodic dextral shear during the Late Carboniferous on major faults adjacent to the Avalon–Meguma terrane boundary. Yeo and Ruixiang (1987) attributed the origin of the Stellarton Graben, immediately to the east of the Cobequid Highlands, to Late Carboniferous development of a local pull-apart basin associated with dextral shear. In southern New Brunswick, Nance (1986) interpreted Westphalian thrusting as associated with dextral transpression on the Cobequid–Chedabucto fault system.

Mineral potential

Mineral exploration in the Cobequid Highlands has been active periodically. The most recent burst of activity occurred in the early 1980s prior to the moratorium on uranium exploration. Known mineral occurrences for the entire Cobequid Highlands were compiled by Donohoe and Wallace (1982a, b, c, d; *see also* Murphy et al., 2000). Base-metal and uranium exploration has been active in the Late Devonian–Carboniferous basins surrounding the Cobequid Highlands and is described by Ryan et al. (1987, 1990). On the southern flank of the highlands, fault- and joint-related Fe-Mn-Cu mineralization occurs adjacent to the Cobequid Fault, mainly within Carboniferous sedimentary rocks. A deposit of this type in the Londonderry area was mined in the late nineteenth century.

One of the goals of this report is to provide a regional geological framework as a guide to mineral exploration. In that context, it is important to note that the Late Proterozoic rocks of the Great Village River Gneiss, epizonal plutons and volcanic supracrustal rocks preserve an oblique section across a volcanic arc. Typically, this can result in a Cu-Au-Mo-Hg association (Mitchell and Garson, 1981). The Folly River Formation demonstrates that there was a rift component to the arc, a setting similar to Kuroko-type mineralization. To date, only a few minor occurrences have been found.

Most recent mineral exploration has focused on the voluminous Late Paleozoic bimodal plutonic and volcanic rocks. This report indicates that the bulk of this magmatism is probably Devonian and was emplaced in an intracontinental rift environment. These environments typically have Sn-W-Mo-F-U-Ag-Nb mineralization potential (Mitchell and Garson, 1981). In that context, epithermal U-Au-Th-F-Cu-Ag-Sn-Mo-Fe mineralization in Devonian–Carboniferous bimodal volcanic and intrusive complexes (Donohoe and Wallace, 1982a, b, c, d; *see* Murphy et al., 2000) were interpreted by Chatterjee (1984) to be related to rhyolite domes. In addition, the GIS studies of Rogers et al. (1986) suggest that the both the plutonic and volcanic rocks have considerable exploration promise for these metals adjacent to their contacts.

CONCLUSIONS

Some of the major conclusions of the present work are summarized as follows.

1. Formational status is assigned to the Late Proterozoic (Neoproterozoic III) Gamble Brook and Folly River successions because they can be mapped by lithotype and stratigraphic position.
2. The distribution of lithologies varies considerably across the Rockland Brook Fault which defines a boundary between the northern and southern fault blocks.
3. The oldest dated rocks are the ca. 734 Ma gneiss on Economy River.
4. The Great Village River Gneiss is penecontemporaneous with Late Proterozoic arc-related sequences, rather than its basement.
5. A ductile shear zone separates Gamble Brook Formation from Great Village River Gneiss.
6. The Folly River Formation rests unconformably on Gamble Brook Formation.
7. The Folly River Formation correlates with other Late Precambrian volcanic-sedimentary sequences such as the Jeffers Group in the Cobequid Highlands and the Georgeville Group in the Antigonish Highlands.
8. Most of the "middle Silurian Earltown volcanic rocks" of Donohoe and Wallace (1982a, b, c, d) belong to a late Precambrian sequence, described here as the Dalhousie Mountain Formation. The name Earltown Formation is therefore abandoned.
9. Much outcrop previously assigned to the early Tournaisian Nuttby Formation is reassigned to an unnamed late Ordovician–Silurian formation.
10. A major north-northeast-trending synclinal Carboniferous fold occurs in the in the southern highlands.
11. Devonian–Carboniferous thrusting and reverse faulting related to dextral-strike-slip movement are pervasive in the northern highlands.

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Appendix
Analytical Techniques

CHEMICAL ANALYSES

Major elements and fourteen trace elements were analyzed by X-ray fluorescence at Saint Mary's University on a Phillips PW1400 sequential spectrometer using a Rh-anode X-ray tube. International standards with recommended values from Abbey (1983) as well as in house standards were used for calibration. Analytical precision, as determined from replicate analyses, is generally better than 2%, except MgO, Na_2O, and Nb which are better than 5% and Th which is better than 10%. Loss on ignition (LOI) was determined by treating the sample for 1.5 hours at 1050°C in an electric furnace. The rare-earth element concentrations were determined either by neutron activation analysis at McMaster University or by ICPMS at Memorial University. The analytical error of the determinations using either technique is about 10%.

URANIUM-LEAD ISOTOPIC ANALYSES

Uranium-lead analysis on zircons were processed by a method similar to that of Krogh (1973), but using a 0.3 mL resin volume and a mixed ^{205}Pb-^{233}U-^{235}U spike. Isotope ratios were measured on a VG-Sector mass spectrometer in the UQAM-McGill laboratory at the University of Quebec. The blank for the entire analytical procedure ranged from 7–30 pg Pb. Errors are calculated at the 95% confidence level and include measurement error, confidence in the fractionation factors, error in the U/Pb ratio of the spike, and the effect of the common Pb correction. Unless stated otherwise, all zircon fractions consist of clear, euhedral crystals interpreted to be of magmatic origin.

^{40}Ar-^{39}Ar ANALYSES (by Dave Dallmeyer, University of Georgia)

^{40}Ar/^{39}Ar incremental-release dating (Dalrymple and Lanphere, 1971) techniques were followed. Pure mineral concentrates (>99%) were prepared using heavy liquid and magnetic separation techniques. Concentrates were wrapped in aluminum-foil packets, encapsulated in sealed quartz vials, and irradiated (total neutron dose of approximately 4 x 1018 nV) for 40 hours at 1000 kW in the central thimble position of the United States Geological Survey TRIGA reactor in Denver, Colorado. Samples were incrementally heated. Each heating step was maintained for 30 minutes Temperatures were controlled with an infrared sensing thermometer and are accurate to ± 25°C with internal monitoring to ±10°C. The crucible was cooled to room temperature between heating steps, and the evolved gas was purified with hot Cu-CuO and Al-Zr getters. A "plateau" is defined if the ages recorded by two or more contiguous gas fractions that together constitute more than 50% of the total gas released from a sample and are mutually similar within a ±1% intralaboratory uncertainty. Analyses of the MMhb-1 monitor indicate that apparent K/Ca ratios may be calculated through the relationship 0.518 (±0.005) x (^{39}Ar/^{37}Ar) corrected.