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EUROGRANITES 2010

FIELD EXCURSION GUIDEBOOK

NOVA SCOTIA

Glaciated Surface of the South Mountain Batholith at Peggys Cove, Nova Scotia

Editor: D. Barrie Clarke

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SAFETY CONSIDERATIONS

Many outcrops are along roadsides. Use extreme caution near or on the travelling surface.

Use care when hammering, both for yourself and others. Eye protection is recommended.

Some outcrops are steep and potentially metastable. Use hard hats and common sense in these places.

Some outcrops are near sea-level. Beware slippery rocks and large rogue waves.

Bears, coyotes, and moose can be very dangerous, especially in Cape Breton Island. Do not venture out alone.

Ticks can be a problem also. Avoid low bushes and tall grass. Check for your skin for ticks at the end of each day.

Stow belongings under the seat in the bus, and remain seated while the bus is moving.

A first aid kit is available on the bus.

The emergency telephone number in Nova Scotia is: 911

PREFACE: OVERVIEW OF PERI-GONDWANAN TERRANES IN NOVA SCOTIA

J. Brendan Murphy, Department of Earth Sciences, St. Francis Xavier University (bmurphy@stfx.ca)

The Appalachian Orogen extends more than 3000 km from Newfoundland to Alabama along the eastern margin of North America (Williams 1979; Hibbard et al. 2006, 2007). Pre-Mesozoic reconstructions imply genetic linkages and former continuity with the Caledonide and Variscan orogens of western Europe (Fig. 1). The Appalachian-Caledonide-Variscan orogen records the formation and subsequent destruction of Paleozoic oceans such as the Iapetus and Rheic oceans (e.g. Williams 1979; Keppie 1985; van Staal et al. 1998) culminating in the formation of Pangea (Murphy and Nance 2008). There is general consensus on major tectonic events that gave rise to the Appalachian orogen. These events include: (i) creation of the Iapetus Ocean between ca. 600-550 Ma resulting in passive margin development along eastern Laurentia (Cawood et al. 2001); (ii) destruction of this passive margin beginning in the Late Cambrian and continuing into the middle Ordovician (Taconic orogeny; van Staal et al. 1998, 2009); (iii) accretion of Gondwanan-derived terranes (collectively known as peri-Gondwanan terranes) sometime between the Ordovician and Devonian (Keppie 1985; van Staal et al. 1998; Barr et al. 2002; Murphy et al. 2004, 2006); followed by (iv) collision between Laurussia and Gondwana, one of the major events in the amalgamation of Pangea, which in North America is known as the Alleghanian orogeny (e.g. Williams 1979), but in western Europe is known as the Variscan orogeny.

Events associated with the destruction of the Laurentian passive margin are assigned to the Taconic orogeny, and coeval deformation and metamorphism along portions of the Gondwanan margin are assigned to the Penobscot orogeny. These events include ophiolite obduction, arc-continent and microcontinent collision events and are broadly coeval with the collision between Laurentia and Baltica to form Laurussia (Scandian orogeny).

The most enduring controversies in the Appalachian orogen are related to various interpretations on the evolution of the peri-Gondwanan terranes. We will see three representatives of these terranes on this trip: (i) the Meguma terrane (Schenk 1971 1997), which is only definitively exposed in southern Nova Scotia, and is separated by the Cobequid-Chedabucto Fault Zone (also known as the Minas Fault Zone) from (ii) the Avalon terrane (or Avalonia) which everyone agrees underlies all of northern mainland Nova Scotia (i.e. the Cobequid and Antigonish Highlands) as well as southeastern Cape Breton Island, and (iii) Ganderia, which is thought to underlie much of Cape Breton Island (e.g. Barr et al. 2002). Avalonia and Ganderia can be identified throughout Atlantic Canada (Hibbard 2007) and have probable correlatives in northwestern Europe. More controversially, correlatives for the Meguma terrane have been proposed for NW Wales (Waldron et al. 2010) and for the South Portuguese Zone of Iberia (Martinez-Catalan et al. 1997).

The Meguma terrane is underlain by a ca. 10 km thick succession of Cambrian (possibly Ediacaran) to Ordovician metaturbiditic rocks of the Meguma Supergroup (White 2008; Waldron et al. 2009) that are unconformably overlain by a mainly Silurian to Early Devonian succession of bimodal volcanic and shallow marine to continental clastic rocks (e.g. Schenk 1997). These rocks were deformed and metamorphosed beginning at ca. 400 Ma (Reynolds and Muecke 1978; Hicks et al. 1999), and were intruded by late syntectonic Late Devonian (ca. 380-372 Ma) granitoids (Kontak et al. 2003, 2004). The basement to the Meguma terrane is not exposed, but Sm-Nd isotopic data from Early Silurian and Devonian crustally-derived felsic rocks indicated derivation from crustal sources with depleted mantle model ages (T_{DM}) between 0.9 and 1.9 Ga (Keppie et al. 1997; Clarke et al. 1997; MacDonald et al. 2002). As the Early Silurian volcanic rocks (White Rock Formation) predate all metamorphic and structural events recorded in the Meguma terrane, these data may be representative of the Sm-Nd isotopic composition of the Meguma basement during, and immediately prior to the deposition of the Meguma Supergroup.

Avalonia in mainland Nova Scotia and southeastern Cape Breton Island is characterized by Neoproterozoic (635-570 Ma) arc-related sequences (Murphy and Nance 2002), unconformably overlain by Cambrian-Early Ordovician platformal successions with minor, localized volcanic rocks (Landing 1996); these are followed by ca. 460 Ma bimodal volcanic rocks (Hamilton and Murphy 2004) that are overlain by Early Silurian-Early Devonian predominantly siliciclastic rocks (Boucot et al. 1974). The basement to the Avalon terrane is not definitively exposed, but Sm-Nd isotopic data from Neoproterozoic and Paleozoic crustally-derived felsic rocks mostly indicate derivation from crustal sources with depleted mantle model ages (T_{DM}) between 0.95 and 1.2 Ga (e.g. Barr and Hegner 1992; Murphy and Nance 2002).

Ganderia (summarized from van Staal et al. 2009) is interpreted as a Late Neoproterozoic-Early Cambrian arc sequence built upon the Amazonian margin (van Staal et al. 1998). Its leading edge was an active margin into the Cambrian (Penobscot arc, Rogers et al. 2006) with middle Cambrian rift-related magmatism reflecting separation from Gondwana in a back-arc environment (White et al. 1994; Schultz et al. 2007), attributed to slab roll-back (van Staal et al. 1998, 2009). At that time, the active leading edge of Ganderia was an arc (Penobscot arc) and the trailing edge was a passive margin (van Staal 1994). During the Ordovician, Ganderia had a complex evolution involving the creation and destruction of back-arc basins, culminating with accretion to Laurentia (the Salinic orogeny, see below) beginning at about 455-450 Ma (van Staal 1994; Zagorevski et al. 2007), and continuing until ca. 430-423 Ma (van Staal et al. 2009).

There is a general consensus that the peri-Gondwanan terranes in Atlantic Canada originated along the northern margin of Gondwana (Amazonia-West Africa) in the Neoproterozoic, separated from that margin in the Paleozoic, and accreted to Laurentia-Baltica later in the Paleozoic (Fig. 1). There is also a general consensus that these terranes have different Paleozoic histories, and the oldest rocks that can be definitively mapped across terrane boundaries are latest Devonian-earliest Carboniferous in age. The main controversies include: (i) their relative original positions along the Gondwanan margin, and (ii) whether they separated from that margin as a large continental ribbon terrane (and so were fellow travellers during the Paleozoic, e.g. Murphy et al. 2004, 2006) or whether they separated at different times and travelled as independent terranes (e.g. van Staal et al. 1998, 2009). In the former scenario, Ganderia is the leading edge and Meguma is the trailing edge of a large continental ribbon that accreted to Laurentia-Baltica as one large terrane beginning in the Silurian. Differences in tectonic histories are attributed to their location along either the leading edge (Ganderia) or trailing edge (Meguma) of the continental ribbon. In the latter scenario, the peri-Gondwanan terranes travelled on different plates and accreted at different time to Laurentia-Baltica and their sequential accretion is responsible for the Salinic (Ganderia), Acadian (Avalonia) and Neoacadian (Meguma) orogenies (van Staal et al. 2009).

One of the principal difficulties in evaluating the possible existence of separate terranes is that the traditional characteristics of suture zones that occur during terrane accretion (e.g. ophiolites, mélanges) are not preserved, implying either that convergence highly oblique, or that the suture zones have been destroyed (or rendered cryptic) by post-accretionary strike-slip movement.

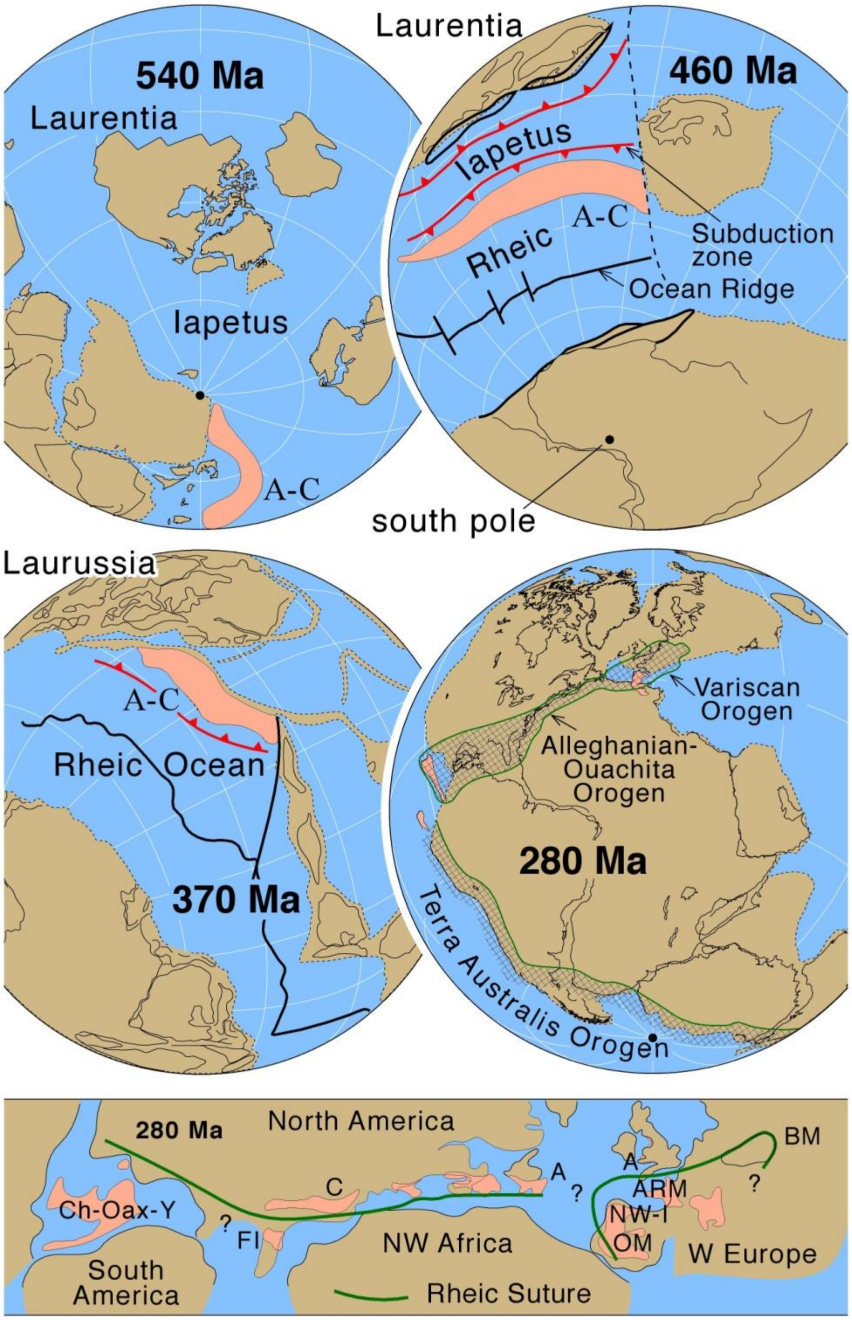

Figure 1. Generalized scheme (from Murphy et al. 2006) showing the migration of the peri-Gondwanan terranes (A-C pink) from the northern margin of Gondwana, their paleogeographic relationship with the Iapetus and Rheic Oceans, and ther accretion to Laurentia-Baltica. Note that other reconstructions (e.g. van Staal et al. 2009) show Ganderia, Avalonia and Meguma as independent terranes.

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CHAPTER 1: THE MEGUMA LITHOTECTONIC TERRANE

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Dedication to Paul Schenk, the Father of the Meguma Terrane

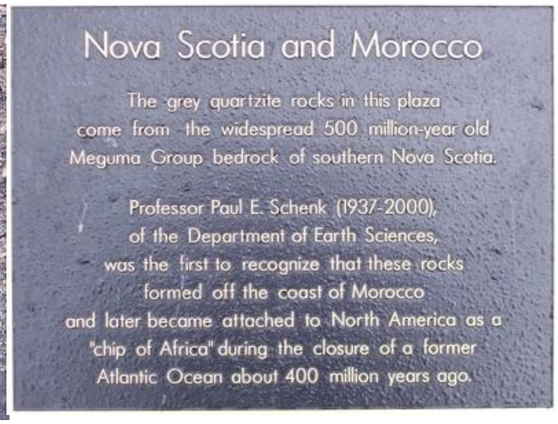

Figure 1-1. Paul Schenk was the first person to connect the Meguma Terrane to Gondwana. A plaque commemorating his outstanding scientific insight stands outside the Killam Library at Dalhousie University. If he were here to greet you today, he would say, "Bienvenue à Nouveau Maroc".

GENERAL INTRODUCTION

We are going to spend the whole day in and around Halifax, observing an eclectic mix of intriguing physical and chemical processes in the South Mountain Batholith and its Meguma Supergroup country rocks. Each locality provides an opportunity to discuss a different topic, and should generate some lively discussions. Geographical constraints interfere with the logical and systematic presentation of the various topics. The principal recurring themes, however, are emplacement, contamination, and internal structures.

A SHORT PRIMER ON THE MEGUMA LITHOTECTONIC TERRANE

Terrane Dimensions and Boundaries

- the terrestrial extent of the Meguma Terrane is ~100 km NS by ~450 km EW (Fig. 1-2), but if the offshore extent is included, the dimensions are closer to ~ 400 km NS by >1000 km EW – that such a large terrane has no known equivalents elsewhere in eastern North America is intriguing, although Waldron et al. (2010) have linked Meguma to the Harlech Dome in Wales
- Meguma's only contact with another Appalachian terrane occurs along the Cobequid-Chedabucto fault zone, a transpressional dextral offset against the Avalon Terrane resulting from oblique collision of Meguma with Avalon/Laurentia
- as determined from xenoliths in lamprophyre dykes, Meguma is underlain by granulite facies meta-igneous and metasedimentary rocks with Nd model ages (~600 Ma) similar to some ages in Avalon

- Meguma Terrane is dominated by just two lithologies: a thick package of turbiditic metasedimentary rocks (Meguma Supergroup) intruded by peraluminous granites (South Mountain Batholith and its satellites)

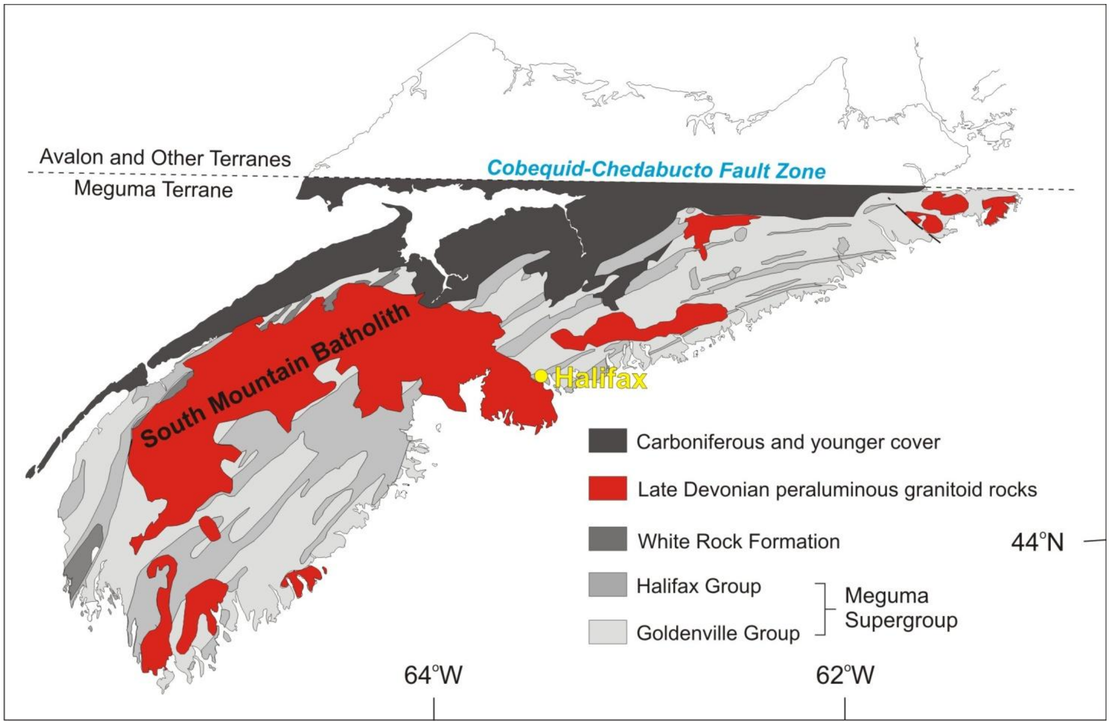

Figure 1-2. Location of the Meguma lithotectonic terrane.

Meguma Supergroup

- Meguma metasedimentary rocks constitute a thick (≥ 10 km)(White and Barr 2010), sparsely fossiliferous, Cambro-Ordovician turbidite succession (Meguma Supergroup – the lower Goldenville Group dominated by psammites and the upper Halifax Group dominated by pelites), with provenance offshore to the southeast, so Schenk (1971) inferred that these sedimentary rocks had been deposited on the passive margin of Gondwana
- metasedimentary rocks were deformed into tight folds and metamorphosed to greenschist to amphibolite facies during the "Neo-Acadian(?)" Orogeny at ~400 Ma (if the Acadian Orogeny represents the collision of Avalon with Laurentia, this deformation of the Meguma Supergroup is later; Murphy et al. 2007)

Meguma Granitoid Rocks

- Meguma granitoid rocks consist of 20-25 late Devonian (~380 Ma) post-tectonic peraluminous granite plutons, including the South Mountain Batholith (SMB), the largest (~100,000 km^3) batholith in the Appalachians (MacDonald 2001)
- the SMB and its satellite plutons cross-cut regional Meguma structures at high angles, the granitic rocks contain no regional solid-state ductile or brittle deformation structures, and late pegmatites and aplites are "isotropic", thus granite emplacement is post-tectonic in the Meguma Terrane (there will be small reward to anyone who finds evidence of syn-tectonic deformation in the SMB)

- the main source rocks of the granite magmas were **not** the Meguma Supergroup (Eberz et al. 1991; Dostal and Chatterjee 2010), and heat for the generation of granite magmas was some combination of crustal thickening when Meguma was thrust over Avalon and underplating/intraplating of arc-related mafic magmas (Late Devonian Mafic Intrusions - LDMIs, including the inaptly-named Weekend Dykes) (Fig. 1-3)
- some LDMIs are synplutonic, but all LDMIs and all granites have uniform ~380 Ma U-Pb and Re-Os ages, and ages of 380 Ma or less by $^{40}Ar/^{39}Ar$ and Rb-Sr (Kontak et al., in prep.)
- the granitic rocks are all peraluminous, and the common mineralogical expressions of this high A/CNK are biotite, muscovite, cordierite, andalusite, and garnet

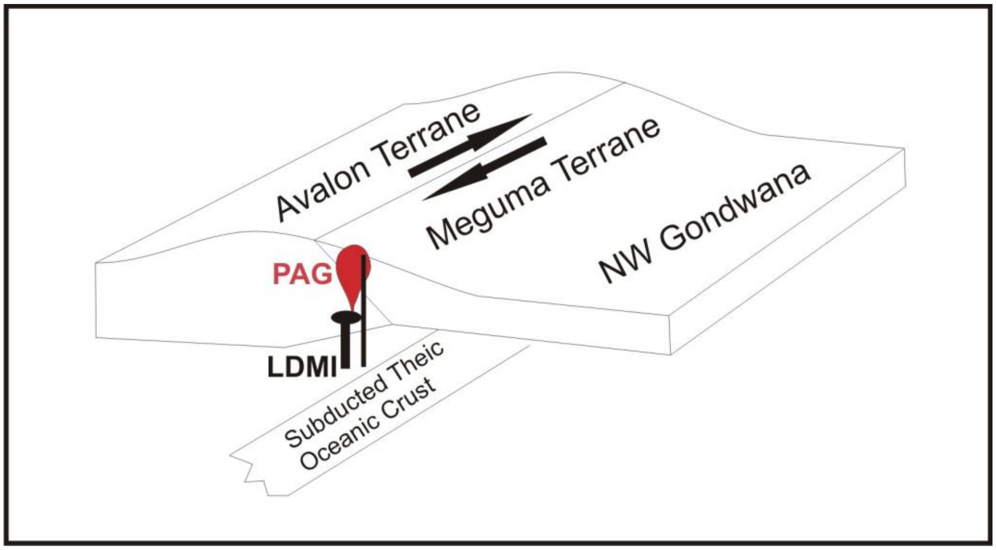

Figure 1-3. The essential lithological and tectonic elements of the Meguma Terrane. Gondwana underwent dextral transpressive collision with Avalon, leading to coeval production of subcontinental mantle-derived Late Devonian Mafic Intrusions (LDMI) and lower Avalon crust-derived Peraluminous Granites (PAG), both of which intruded the Meguma Terrane. Granite magmas were the combined products of overthrusting and underplating.

Post-Granite Events

- after emplacement of the granites at 380 Ma, this part of the orogen was uplifted rapidly, and within 20 my of their intrusion at depths of 10-12 km, the granite plutons were exposed and covered with coarse fluvial clastics of the Early Carboniferous Horton Group, followed by marine carbonates and evaporates of the Windsor Group, all belonging to the Maritimes Basin
- the Maritimes Basin formed near the St. Lawrence Promontory in a broad zone of transtension created by the offset of an orogen-scale dextral strike-slip fault, and eventually accumulated up to 12 km of sediments
- final docking of the Meguma Terrane at its present position against Avalonia occurred at about 320 Ma
- the last major tectonic event in this region was the rifting associated with the opening of the Atlantic Ocean that produced the 200 Ma North Mountain basalts and Shelburne Dyke (possibly related to the Iberian Messejana dyke: Dunn et al. 1998)

In summary, Meguma is most likely a “chip of Africa” left behind in North America when the modern Atlantic Ocean formed. The major puzzle for Meguma-ologists is to determine the original position of this “chip” at the time of deposition of the turbidites. Paul Schenk (1971) was quick to apply the then new plate tectonic hypothesis to determining northwest Africa as the source of the Meguma turbidites, but subsequently others have advanced other potential provenances (Fig. 1-4, Table 1-1).

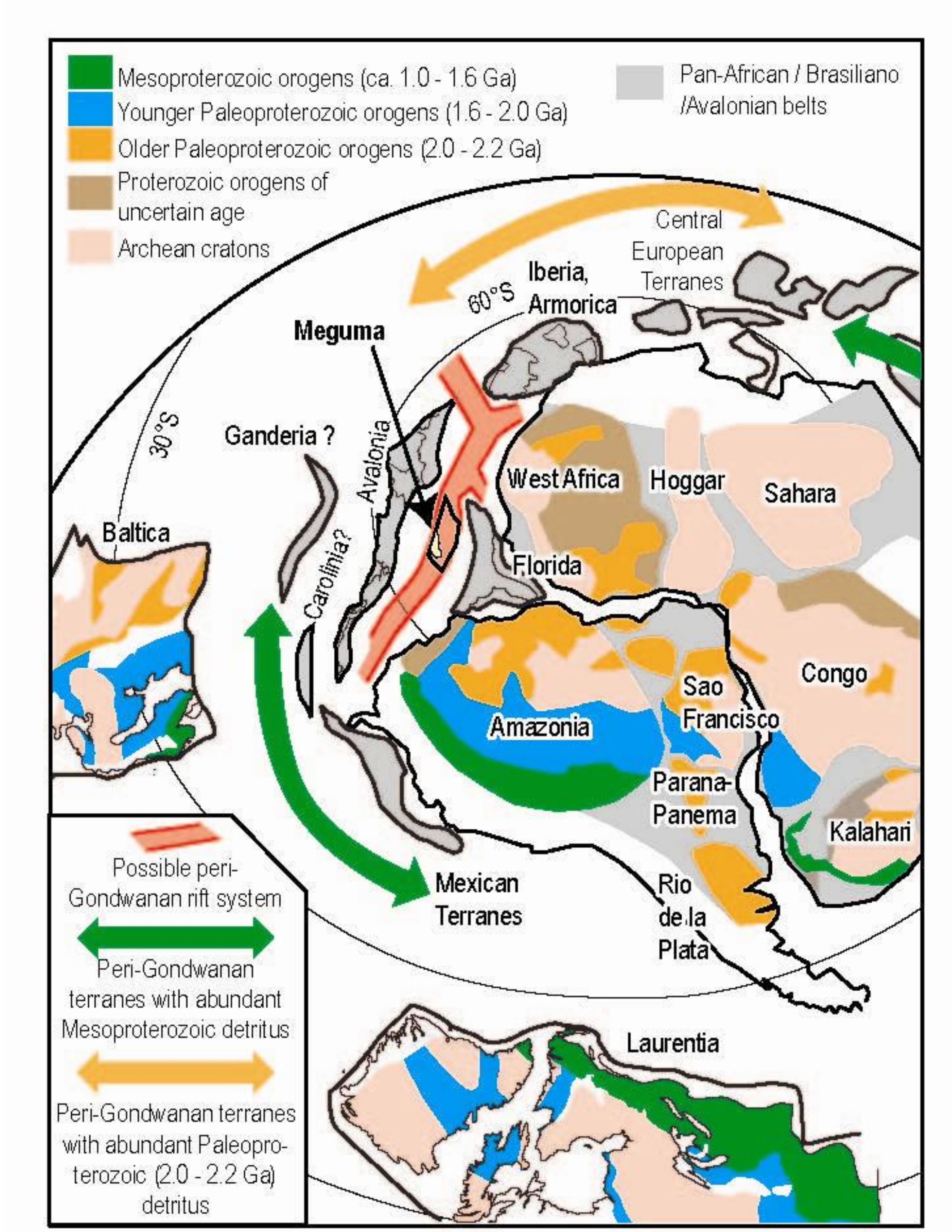

Figure 1-4. A Cambrian reconstruction showing a hypothetical peri-Gondwanan rift in which sediments of the Meguma Lithotectonic Terrane may have accumulated. (Figure courtesy of John Waldron.)

Table 1-1. Possible sources/correlations of the Meguma lithotectonic terrane.

	Similarities	Differences	Comment	References
Amazonia	stratigraphic, deformational, paleomagnetic	no peraluminous granites??	possible fit from palinspastic reconstructions	Keppie (1977); Nance et al. (2008)
Morocco / West Africa	Cambro-Ordovician tubidite sequence cut by peraluminous granites; similar detrital zircon populations	peraluminous granites may be too young in Morocco to correlate with Meguma Terrane granites	best fit from palinspastic reconstructions; granites might be diachronous products of the same collisional event	Schenk 1971, 1997); Clarke and Halliday (1985); Richard and Clarke (1989); Krogh and Keppie (1990)
Iberia	??	peraluminous granites are too young in Iberia (ca. 310 Ma?) to correlate with Meguma Terrane granites	evidence is indirect in the sense that zircon populations connect SW Iberia with the West African Craton (as does the Meguma)	de Albuquerque (1977); Richard and Clarke (1989); Dunn et al. (1998); Pereira et al. (2008)
Wales	similar statigraphies; similar thicknesses; identical depositional ages; Identical zircon source age populations	more fossiliferous than the Meguma, and absolutely no granites	may be just another fragment of a dispersed peri-Gondwanan terrane once proximal to Meguma	Waldron et al. (2009, 2010)

Table 1-2. General locations and rationales for STOPs on Day 1.

STOP	LOCATION	TOPIC
1-1	100 m up the hill from the intersection of Larry Utek Blvd. and the Bedford Hwy.	tectonics: origin and significance of the Meguma Supergroup
1-2	at the La-Z-Boy Furniture Gallery in Bayers Lake - private property – permission must be obtained in advance	mode of emplacement and contamination: chemical effects
1-2a	Hwy. 101 level bridge over railway cut	as at STOP 1-2
1-3	near super mailboxes in Portuguese Cove on Hwy. 349 – private property – permission must be obtained in advance	contamination: why there are no elephants' graveyards
1-4	go to gate at end of Duncans Cove Road of Hwy. 349 and walk down to shore staying off private land	internal structures: the origin of linear schlieren and layering
1-5	take Bald Rock Road off Hwy. 349 and go right to the end of the unpaved section – private property – permission must be obtained in advance	contamination: mafic porphyry supercontamination
1-6	take the Prospect Road off Hwy. 333, stop at the small bridge into the village, and walk northwest along the coast	internal structures: the origin of ring schlieren and ladder dykes
1-7	all signs lead to Peggys Cove!	contamination: chemical and structural synthesis
1-8	Hwy. 103 at top of the hill just west of Exit 5	contamination: peritectic cordierite
1-9	Hwy. 103 at intersection with Bowater logging road 2.6 km west of STOP 1-8	pressure quenching: magmatic cordierite

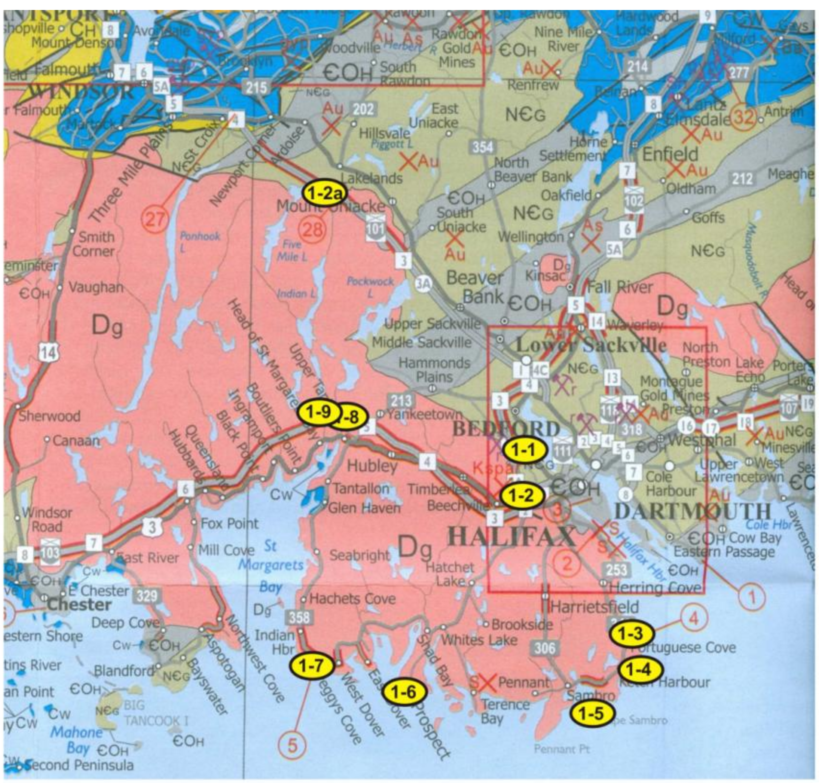

Figure 1-5. Locations of STOPs for Day 1. (Geological Highway Map by permission of Atlantic Geoscience Society)

STOP 1-1 – LARRY UTEK BLVD./ÉCOLE BEAUBASSIN: MEGUMA TURBIDITES (44.699036°N 63.663796°W)

At this location, we can see metre-scale turbidites of the Meguma Supergroup showing variously developed Bouma sequences (Fig. 1-6). The thicknesses of the individual depositional units and the sand/mud ratios both vary. This locality is in the (lower) Goldenville Group where sand/mud > 1, whereas the (upper) Halifax Group has sand/mud < 1. The thickness of this rather uniform material is ≥ 10 km and, as such, it provides an envelope of essentially constant mineral assemblage for the intrusion of the peraluminous granites. Because the host and intrusion have a nearly identical mineral assemblages, detection and quantification of contamination is difficult.

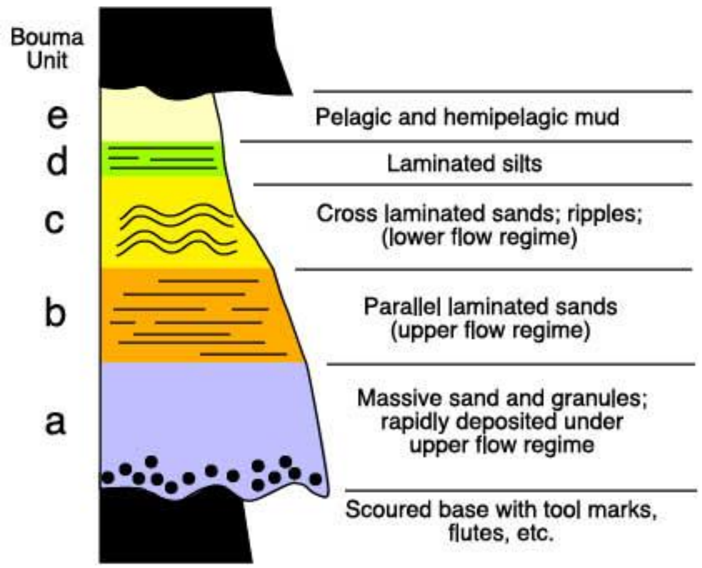

Figure 1-6. Bouma sequence in turbidites.
(http://faculty.gg.uwyo.edu/heller/Sed%20Strat%20Class/SedStrat%208/bouma.jpg)

At this distance of approximately 3 km from the contact with the South Mountain Batholith, the psammitic beds contain contain euhedral crystals of pyrite, and the pelitic tops of these depositional units contain microporphyroblasts of cordierite (Fig. 1-7). Toward the contact, pyrite desulphurizes to become pyrrhotite and the cordierite porphyroblasts grow to become ~5 millimetres in diameter. The bottom line is that the Meguma Supergroup is thick (≥ 10 km), uniform in composition, and has a provenance to the southeast.

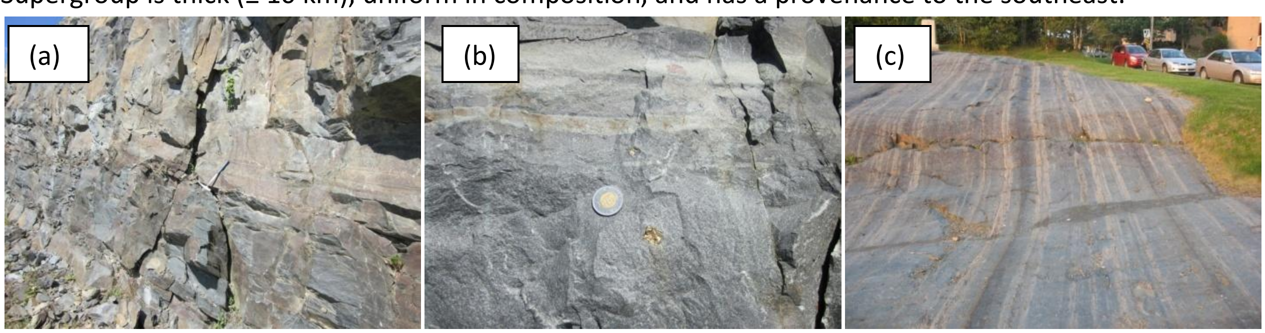

Figure 1-7. (a) Typical package of Meguma turbidites belonging to the psammite-dominated Goldenville Group. (b) Detail of a psammitic bed with euhedral pyrite. (c) Typical package of pelite-dominated turbidite, in Titus Smith Park in Fairview, belonging to the Halifax Group.

STOP 1-2 – OTTER LAKE COURT: THE MEGUMA-SMB CONTACT (Private Property) (44.646725°N 63.669387°W)

At this location, we can see the contact zone between the rusty sulphide-rich rocks of the Goldenville-Halifax Transition Zone and the South Mountain Batholith (Fig. 1-8). Here, the "contact" between the Meguma Supergroup and the South Mountain Batholith (SMB) is approximately 200 m wide. (Is the contact at the outermost occurrence of granite, or is it at the innermost occurrence of country rock? Ideally, it should be at the surface marking the mutual limits of contiguous country rock on one side and granite on the other side, but without 100% exposure, it is not possible to precisely define that surface here.) Clearly, though, large slabs of Meguma country rock had been spalling into the SMB and, at this locality, the mode of emplacement of the ca. 100,000 km^3 of granite magma was stoping. Ice-shelf disintegration is a good two-dimensional analogue (Fig. 1-9). We can continue to discuss the processes of stoping/spalling and disintegration at the next locality.

Figure 1-8. (a) Part of the Meguma-SMB contact zone at Otter Lake Court. More granite to the left; more Meguma to the right. (b) A small, and rare, Meguma xenolith in the marginal granodiorite.

Figure 1-9. Negative of the Larsen ice-shelf collapse, an analogue for stoping in magma chambers (black is solid snow and ice; white is liquid water). Largest "xenolith" is ~20 km long. Note the large roof pendant on the right. Modified from http://earthobservatory.nasa.gov/Features/WorldOfChange/larsenb.php

STOP 1-2a – HWY 101: THE MEGUMA-SMB CONTACT (44.892651°N 63.883835°W)

Time constraints will almost certainly preclude a visit to this locality. The outcrop shows the same spalling of country rock slabs into the SMB as at STOP 1-2, but shows more evidence of disintegration and contamination. All granites are contaminated. The only questions are: by what, and by how much? Because the Meguma Supergroup is so thick, the granite magmas have seen little else, so the "what" is well constrained. Furthermore, the mineral assemblages of the Meguma and the South Mountain Batholith (SMB) are nearly identical (qz+Ksp+plag+bt+musc+ap+mnz+zirc+ilm), but the textures and compositions of the phases are significantly different. If xenocrysts of mineral X enter the granite magma, they attempt to equilibrate chemically and texturally until the differences are eliminated. At that point, magmatic and xenocrystic grains of X are indistinguishable, and assimilation is complete. As long as we can see evidence of arrested reactions, we are able to determine that contamination has taken place. We have systematically investigated what happens to Meguma silicates, oxides, sulphides, and phosphates as they entered the SMB magma. At this locality, evidence of contamination includes abundant garnets, sulphides, and xenoliths with reaction rims in the marginal facies of the SMB granite (Fig. 1-10). The Meguma psammites are rather unreactive, but the pelites have undergone extensive partial melting to produce high modal abundances of garnets that are almost certainly paraxenocrystic in origin. On the other hand, the high concentration of biotite is almost certain a combination of magmatic and Ostwald ripened xenocrystic grains. And the high concentration of sulphide is essentially orthoxenocrystic.

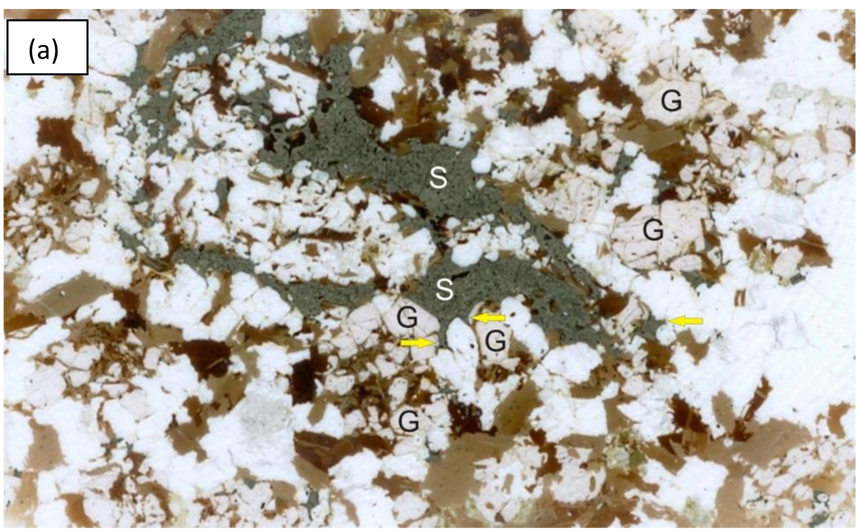

Figure 1-10. (a) Granodiorite contaminated by Meguma pelite resulting in abundant paraxenocrystic garnet (G) and melted orthoxenocrystic sulphides (S). Arrows show immiscible molten sulphide wrapped around silicates. Image width 3 cm. (b) Meguma xenolith with a necklace of Ostwald-ripened biotite xenocrysts on its margin.

So, at this location at the contact, we have a complex mixture of magmatic, orthoxenocrystic, and paraxenocrystic grains that is, nevertheless, relatively easy to deconvolve where the reactions were incomplete, but what about farther into the batholith where the reactions are complete, and where the degree of assimilation is high? The only clue we have is from Sr-Nd systematics which show that the interior of the SMB is more Meguma-like than the margin, even though the physical evidence of contamination is less. It is, after all, the job of assimilation to destroy all physical/textural evidence of contamination.

STOP 1-3 – PORTUGUESE COVE: DISINTEGRATING ELEPHANT (Private Property)(44.522944°N 63.532645°W)

Xenolithic fragments enclosed in granite batholiths indicate incorporation of material from the walls and roof of the intrusion. The actual process of detachment from contiguous country rock is probably one of *thermal stress fracturing*, i.e., the heating, expansion, and spalling of the country rock in direct contact with the hot granite magma. Once a block has spalled off to become a “free-swimming” or sinking xenolith, the process repeats itself, and the granite magma expands outwards, or moves upwards, by this process known as *stoping*. Some geologists believe that stoping is not important in the ascent of granite magmas, because nobody ever finds floors of batholiths strewn with large sunken xenoliths, hypothetical zones known as *elephants’ graveyards*. In simple terms, they say, “No elephants’graveyards, therefore no stoping”. Clarke et al. (1998) say that no elephants’ graveyards exist because the elephants disintegrated, and that disintegration facilitated assimilation.

At Portuguese Cove is a large xenolith (Clarke et al. 1998). It was probably stoped from the roof of the batholith and became stuck in the granite mush as it fell (the arresting of descent is a problem in itself!). What is remarkable about this xenolith is the large number of fractures in it (~375 over a distance of ~ 25 m), filled with granite (Fig. 1-11). If this xenolith had been in a “magma-static” pressure situation, why would the melt penetrate the xenolith – what driving force would there be? On the other hand, if the xenolith were undergoing thermal stress fracturing as it descended into hotter magma, it would crack and inexorably draw the viscous magma into the vacuum of the cracks. All the fractures in this xenolith suggest that it was fragmenting, and the xenolithic shrapnel scattered around this elephant suggest that the disintegration may have been explosive (rapid fracture propagation).

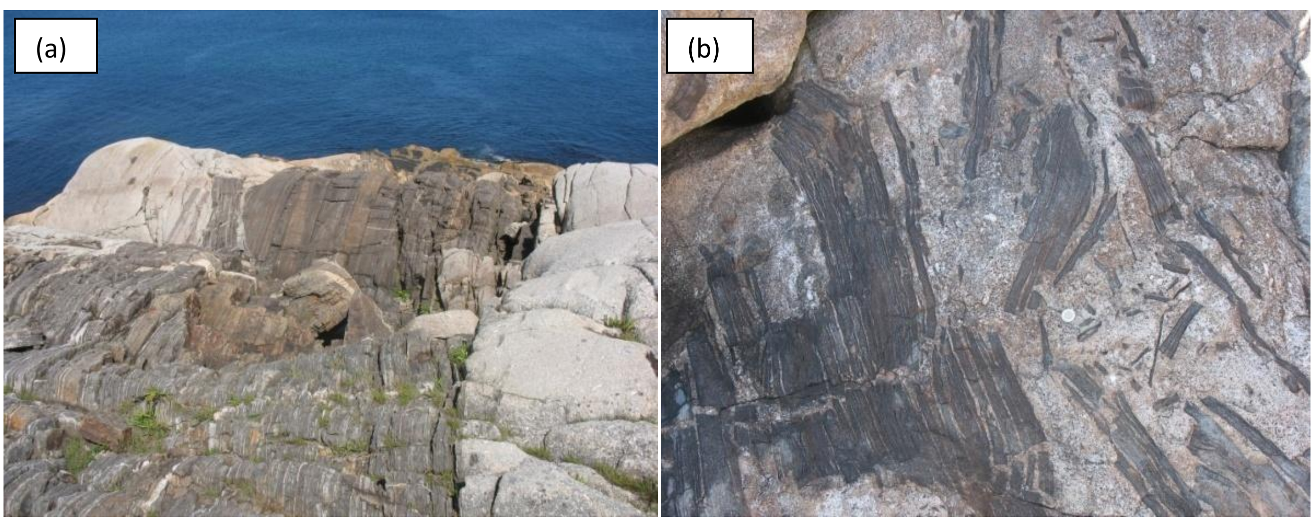

Figure 1-11. (a) Part of the large xenolith at Portuguese Cove with the abundant granite micro-dykes visible in the foreground. (b) Xenolithic shrapnel spalling/exploding off the main xenolith.

At this location, make sure you see:

- two types of external bounding surfaces (new straight ones, and old curved ones)
- depletion zone and K-feldspar concentrations adjacent to the southwest corner of the xenolith
- tapered, high-aspect ratio, internal fractures filled with granite (miniature aplite-pegmatite systems a few millimetres wide)

- xenolithic shrapnel as evidence of the explosive disintegration of the xenoliths
- good flow foliation and curious clusters of coarse constituents south of the elephant

So, if xenolithic elephants disintegrate on descent, there never will be any elephants' graveyards. Instead, all one will ever find is a few elephant body parts floating around in the granite (Fig. 1-12). Their small volumes and relatively large surface areas make them ripe for reaction with, and assimilation by, the magma.

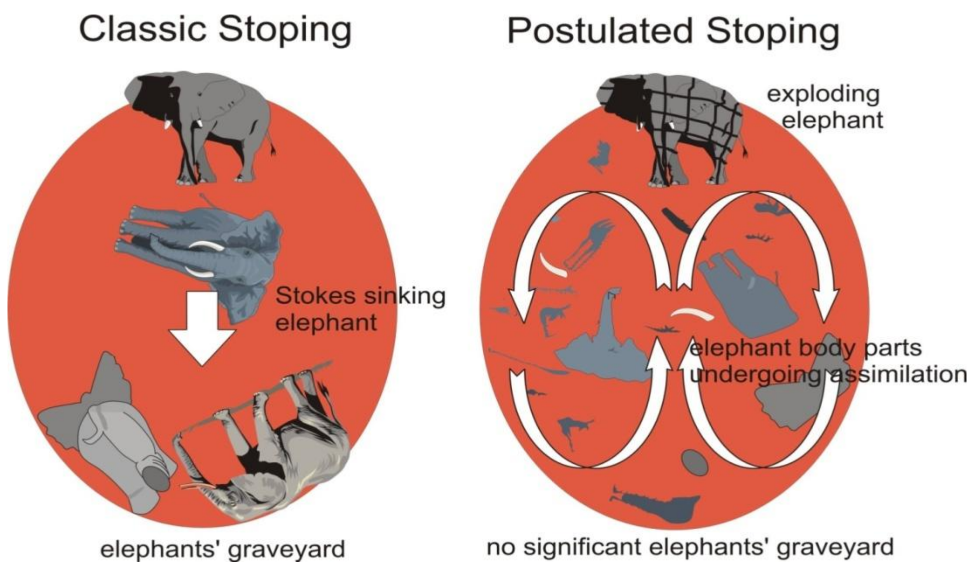

Figure 1-12. A comparison of classic stoping to produce elephants' graveyards, and stoping + disintegration to promote physical and chemical assimilation.

STOP 1-4 – CHEBUCTO HEAD: LINEAR SCHLIEREN AND LAYERING (44.500806°N 63.520071°W)

Layering, graded bedding, cross-bedding, scour-and-fill structures, and slumping are features normally associated with sedimentary rocks such as the Meguma Supergroup. For early workers, it was difficult to convince others that such structures could form in gabbroic magmas (10^3 times more viscous than water) to produce the Skaergaard intrusion. Another three orders of magnitude in viscosity may appear to stand in the way of such structures forming in granites, but they do form nevertheless. At Chebucto Head, we are going to look at a remarkable 7-metre thick layered granite sequence (Clarke and Clarke 1998).

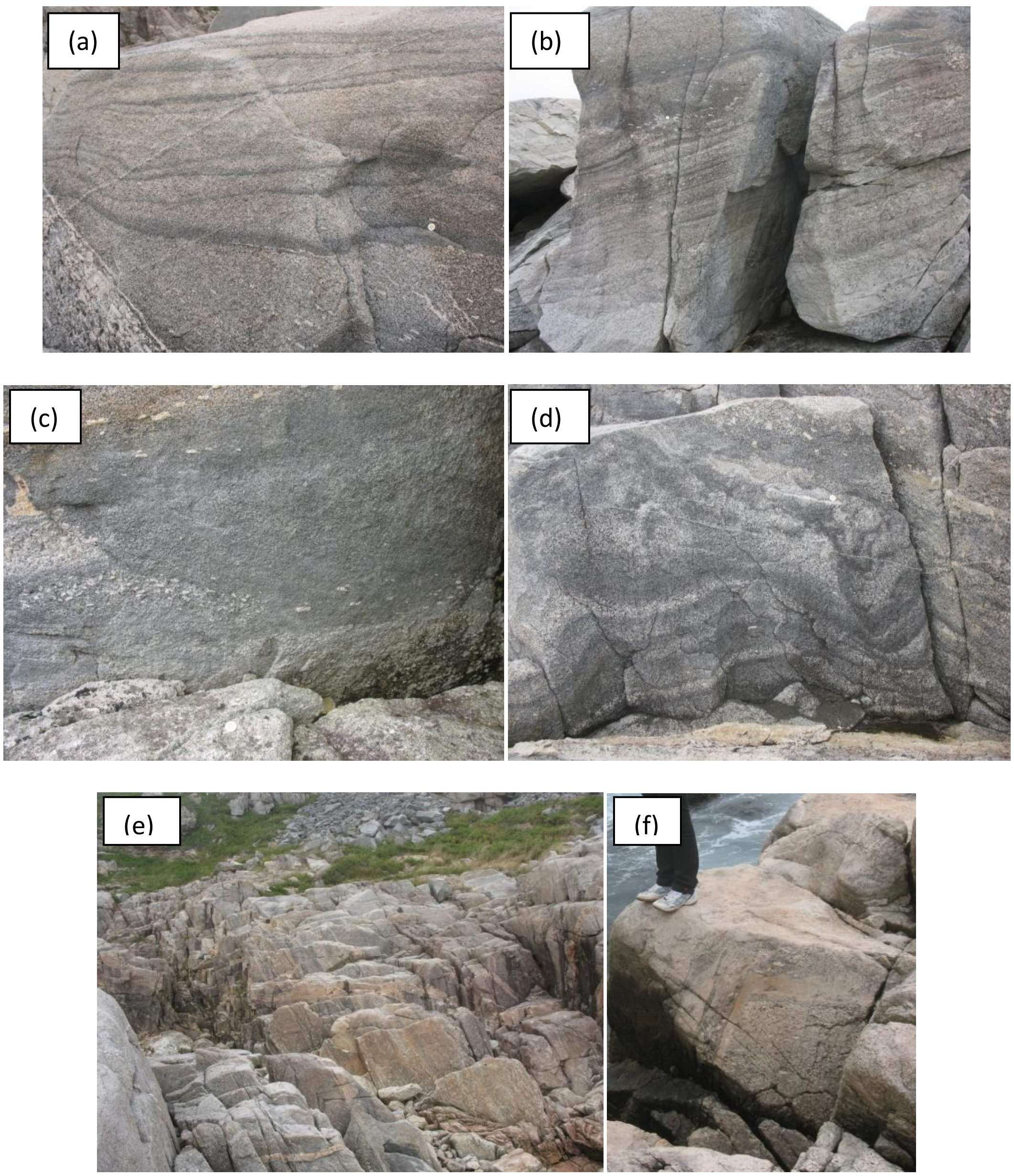

Figure 1-13. Layering features at Chebucto Head. (a) Typical layers with fine mafic bottoms grading into coarse felsic tops. (b) Sequence of layers cut by a channel. (c) Cross-bedding. (d) Typical soft-sediment slump structure. (e) Vertical composite feeder dyke. (f) Right-angle transition from vertical dyke to first horizontal layer.

We will move systematically from the bottom of the layered sequence to the top, looking at features shown in Figure 1-13:

- the razor sharp lower contact of the layered sequence
- scour-and-fill structures
- K-feldspar foliation without lineation
- slump structures
- inverse graded bedding (fine on bottom, coarse on top)
- melanocratic biotite bases grading up to leucocratic quartz-feldspar tops of individual layers
- sharp contacts between layers, but more gradational variation within layers
- bifurcating layers
- cross-bedding
- diffuse top to layered sequence
- diapir from layers penetrating the overlying granodiorite
- log-jam dyke, its termination at the layers, and its connection to the lowermost layer

Then we will try to understand how this layered sequence formed by a process of multiple injection from the log-jam dyke, each new layer being injected above the former ones (Fig. 1-14).

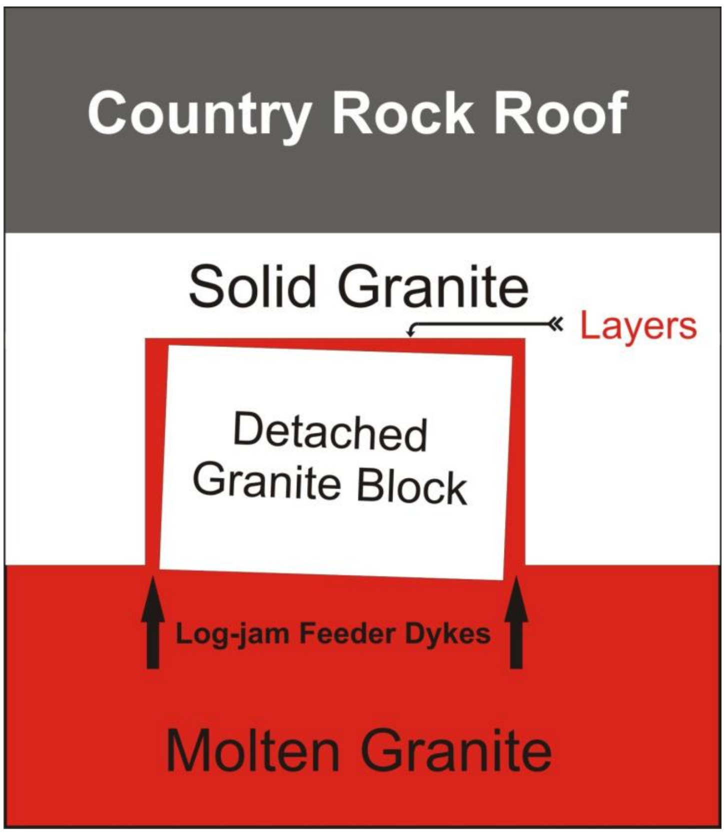

Figure 1-14. A framework model for the formation of the layered granites at Chebucto Head. Detached pieces of the solid granite roof rattle downward, each incremental descent episode drawing magma up the bounding fractures to form a new layer on top of the descending block. Gravity causes settling and graded bedding. Strong currents cause scour-and-fill structures. Jostling causes slumping. The feeders rise only as high as the layers.

STOP 1-5 – SAMBRO HEAD: MAFIC PORPHYRY (Private Property) (44.458081°N 63.580211°W)

One of the most enigmatic facies of the SMB is the so-called mafic porphyries. They constitute less than 0.1% of the volume of the batholith and consist of a relatively melanocratic granitoid rock (hence "mafic") with prominent phenocrysts of alkali feldspar (hence "porphyry"). The mafic porphyries variously contain high modal abundances of Meguma xenoliths, biotite, garnet, cordierite, andalusite, and sulphides. They may also contain plagioclase-mantled alkali feldspars, acicular apatites, and quartz with biotite reaction rims. The mafic porphyries seem to have the characteristics of highly contaminated rocks caught in various stages of chemical disequilibrium. To further darken the plot, we previously thought that only the small satellite intrusions of the SMB contained any evidence of symplutonic intrusions of LDMIs, but at his locality, we suspect that an LDMI is present. If so, perhaps the mafic porphyries are highly contaminated because of the additional heat supplied by the LDMI and that some amount of three-component (felsic magma, mafic magma, Meguma country rocks) mixing is responsible for their unusual textures and compositions (Figs. 1-15 to 1-18).

Figure 1-15. Mafic porphyries at Sambro Head. (a) Typical appearance showing twinned and plagioclase-mantled alkali feldspars. (b) Abundant xenoliths in a mafic porphyry, including the partially melted remnants of a large pelitic xenolith (bottom-left quadrant; Mike MacDonald photo). (c) Detail of corroded K-spar and quartz with biotite rim.

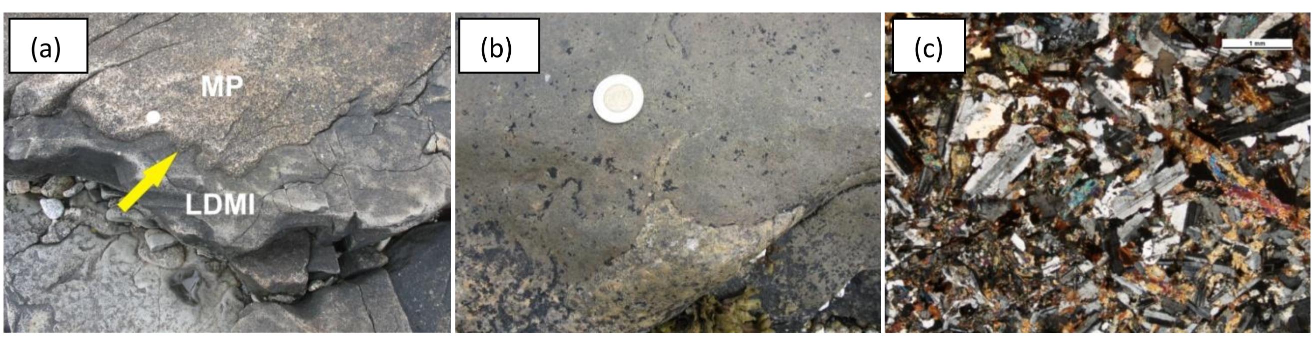

Figure 1-16. Mafic porphyry at Sambro Head. (a) Sharp but lobate contact (arrow) between mafic porphyry (MP) and synplutonic gabbroic dyke (LDMI). (b) Detail of lobate MP-LDMI contact(MP above, LDMI below) .
(c) Photomicrograph of LDMI.

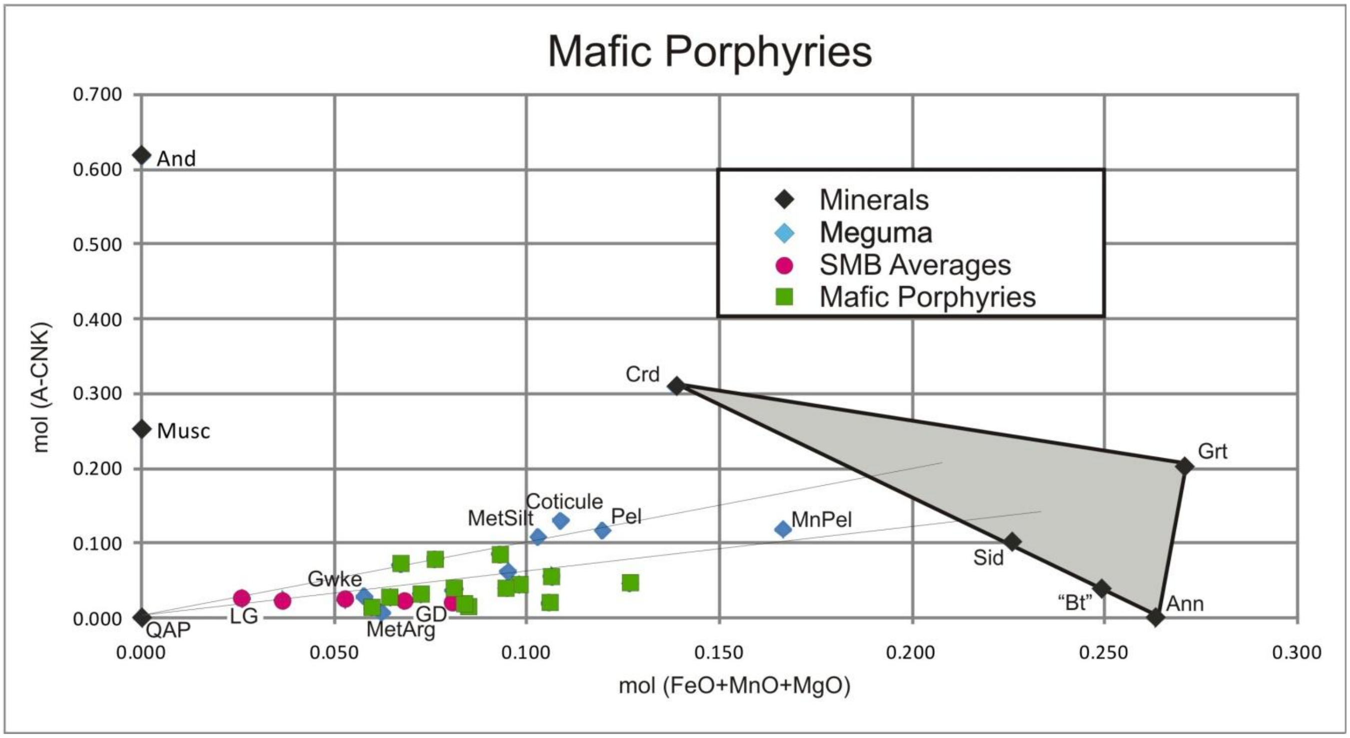

Figure 1-17. Origin of the mafic porphyries. The Meguma rocks can be considered as binary mixtures of QAP + (Bt+Grt+Crd). If they partially melt, they produce a QAP melt fraction and Bt-Grt-Crd refractory/peritectic solids. The QAP melt mixes with the main SMB magma, and the Bt-Grt-Crd solids become dispersed in the magma. Most mafic porphyries plot to the mafic (FeO+MnO+MgO-rich) side of the main SMB evolution trend from biotite granodiorite (GD) to leucogranite (LG).

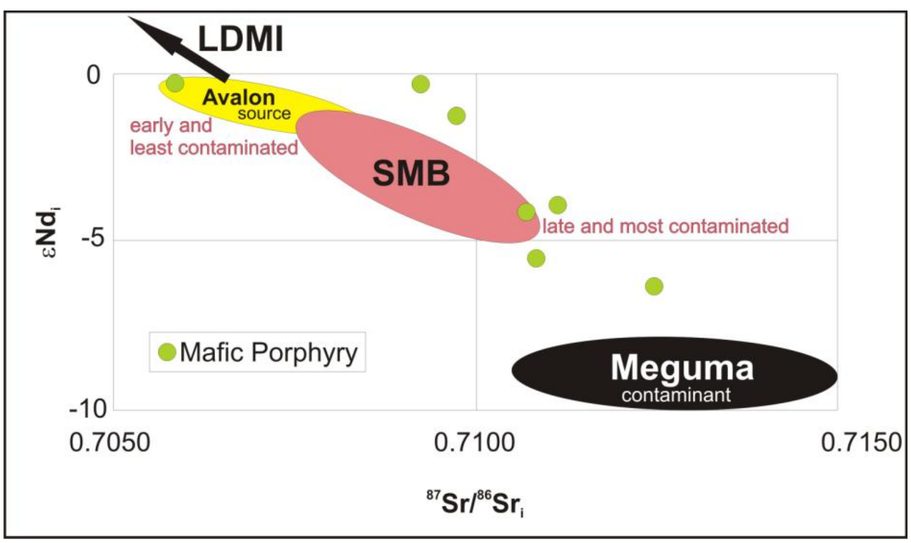

Figure 1-18. Preliminary Sr-Nd isotopic data from several mafic porphyry localities in the SMB show possible influences of contamination by both Meguma and LDMIs. Avalon basement is the putative source of the SMB, but the trend of the SMB towards the Meguma may indicate contamination by Meguma, or Meguma material in the source, especially in the light of some εNd values reported for the Meguma by Waldron et al. (2009).

STOP 1-6 – PROSPECT: RING SCHLIEREN, LADDER DYKES, AND BUBBLE TRAINS (44.475612°N 63.806164°W)

Background

Schlieren are dark streaks in granites with many possible origins. We have already seen one case at STOP 1-4. As we walk out to this locality, we will note the roughly constant contact-parallel orientation of the flow-foliated K-feldspar phenocrysts (dents-de-cheval) and absence of schlieren. But, in a relatively small area (~2500 m^2), some process has clearly disrupted this regular regional pattern – the regular alignment of the Kspar and biotite disappears, and a somewhat irregular development of biotite schlieren appears.

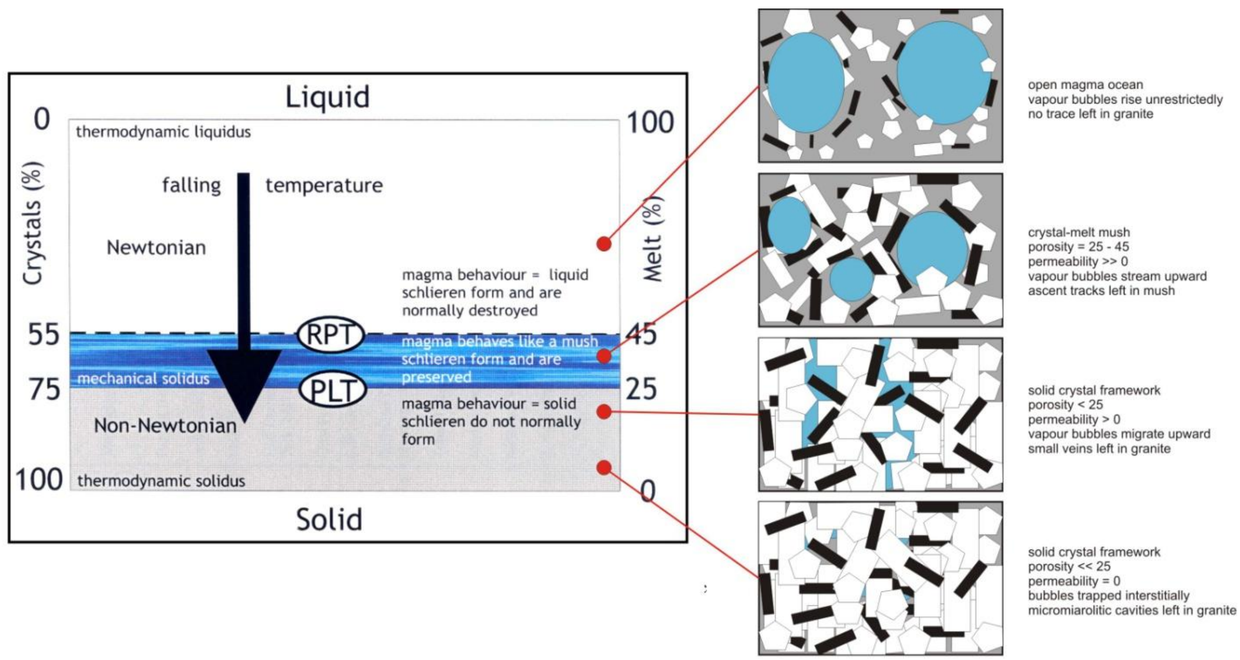

Figure 1-19. Suprasolidus deformation domains.

As granite magmas cool and crystallize, they gradually change from ~0% crystals (ϕ = 0) to 100% crystals (ϕ = 100) (Fig. 1-19). Following Vigneresse and Berg (2000, unpub), at ϕ < 55, crystals are suspended in the melt and the magma behaves as a liquid – any structures that form in the fluid magma at this stage will probably be destroyed. At ϕ = 55, crystals must interfere with one another and the magma reaches the rigidity percolation threshold (RPT). At 55 < ϕ < 75, the crystals interact to the extent that the magmatic mush can flow, but it develops increasing rigidity up to ϕ = 75, which is called the particle locking threshold (PLT). Any structures that form in the crystal-liquid mush at this stage can be locked in permanently. At ϕ > 75, the crystals are completely locked together, and cannot move relative to one another by viscous flow, therefore flow foliations cannot form beyond this limit.

The bottom line is that granite magmas have only a narrow window (55 < ϕ < 75) in which to *form and preserve* any flow structures. At this outcrop, the irregular schlieren and ring schlieren post-date the regional contact-parallel flow foliation, and must be the last structures to have formed in this pluton.

Observations

Table 1-3. Comparison of the regional granite and the area containing ring schlieren.

	Regional Pattern	Local Disruption
Kspar phenocrysts	dispersed flow foliated	clustered random orientations
Biotite schlieren	absent	prominent

At this locality, make sure that you see:

- relatively sharp discontinuity between the regional flow foliation and the chaotic zone
- random orientation of schlieren, size sorting of minerals, and clusters of K-feldspar
- the squashed-spider-shaped schlieren patterns that we call 'arocknids'– circular arocknids, and randomly oriented elliptical arocknids, present major problems for those who believe that the SMB is syntectonic

Table 1-4 below shows the observations we typically make on each ring schlieren structure.

Table 1-4. Typical field measurements on ring schlieren.

Name	**PR2**	**PC4**
GPS Coordinates (NAD 1927)	0435825 E 4924785 N	0426816 E 4927510 N
Long Axis Orientation	-	305°
Dimensions	2m x 2 m	6 m x 5 m
Aspect Ratio	1 : 1	1.2 : 1
Granite Host	biotite monzogranite	coarse-grained leucomonzogranite
Regional Foliation	strong (290° – 345°)	weak to moderate
No. of Nested Ellipses: Observed (Possible)	2	3 (10)
Inner Ellipse	yes	yes
Schlieren Grade	prominent inner bands variable outer bands	prominent to faint inner bands prominent to faint outer bands
Schlieren Thickness	< 1 cm to < 4 cm	< 1 cm to < 3 cm
Terminations	abrupt, gradual, diffuse	abrupt, gradual, diffuse
Cross-cutting	yes	yes
Xenoliths	outside structure	outside structure
Horizontal / Vertical Exposures	horizontal	horizontal / vertical
Dist. to Meguma Contact	5 km	1 km

And, to generalize from a large number of observations, ring schlieren have the following properties (Fig. 1-20):

(i) entire structure has an elliptical shape with an average aspect ratio of ~ 1.5
(ii) entire structure is eccentric and nested, with an average of 5 rings
(iii) where cross-cutting relations are clear, inner rings cut outer rings
(iv) gradations of biotite concentrations normally decrease inward
(v) grain-size sorting, where the mafic bands are finer grained than the felsic bands
(vi) parallel grain orientation (foliation), especially biotite in the schlieren
(vii) considering the large size of the SMB, all ring schlieren are close to the contact
(viii) where three-dimensional observations are possible, the two-dimensional ring schlieren appear to be three-dimensional ca. vertical cylindrical schlieren

Whatever the process responsible for these ring schlieren, it must be capable of producing vertically oriented, cylindrical structures, with many pulses decreasing in size and age from the outside, and capable of producing both modal and grain-size variations in the structure. They must also be late magmatic because they disrupt the flow foliation and are not cut by any other feature. A satisfactory model for the origin of these ring schlieren must account for all of these observations.

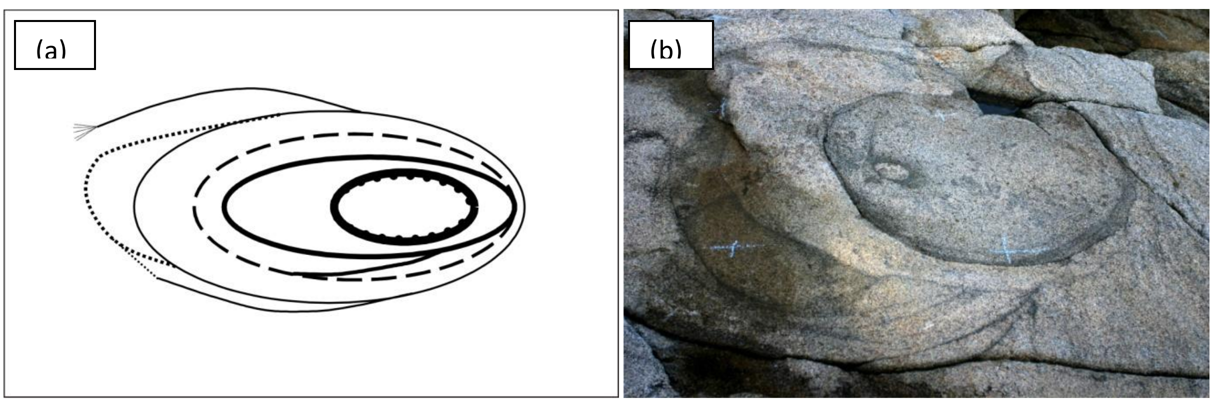

Figure 1-20. (a) Generalized sketch of nested ring schlieren. (b) Typical outcrop pattern (at Aspotogan, not here!). Blue chalk marks a 1 m square grid.

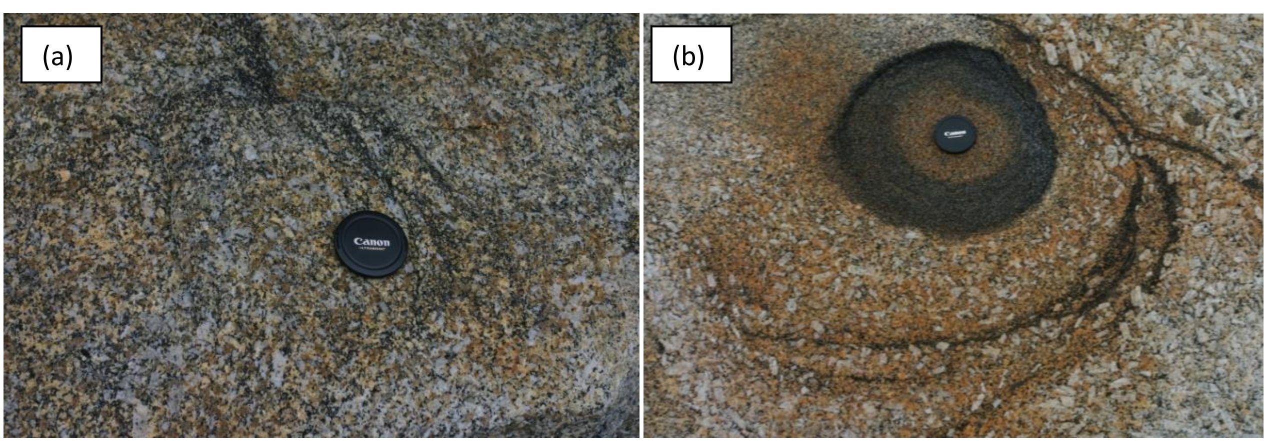

Figure 1-21. (a) Faint arocknid (ring schlieren). (b) More prominent arocknid (ring schlieren).
Photographs by James Sykes.

Discussion

Many locations in the SMB show ring schlieren cutting the regional flow foliation, but at this locality nested ring schlieren and other complex (roll-over?) structures occur in a larger area of disruption of the regional flow foliation. What process(es) can account for these features?

First, what caused the disruption of the regional flow foliation? The agent could be some sort of convective overturn in the 55 > ϕ > 75 mush, caused by differences in density or temperature (Weinberg et al. 2001), but we do not believe that thermal inhomogeneities on this small a scale existed in the SMB. Alternatively, the local disruption might be the fossil pathway of a large xenolith (remember the elephant at STOP 1-3?) that fell through this locality. Such a process might account for both the sharp truncation, and the destruction, of the regional flow foliation. It might also account for the complex roll-over vortex structures. But the more regular ring schlieren are unlikely to have formed in this way. To create these polygenetic structures requires a repetitive process. If falling xenoliths were responsible, one xenolith after another would have to fall through the same path (unlikely), and there is no evidence of spalled debris as they disintegrated during descent. Instead, for the ring schlieren, we favour sequence of rising bubbles (bubble trains) rising from a degassing magma through a 55 > ϕ > 75 mush to produce them (McCuish 2001). Stationary bubble trains can produce ring schlieren; migrating bubble trains should be able to produce ladder dykes (Fig. 1-22).

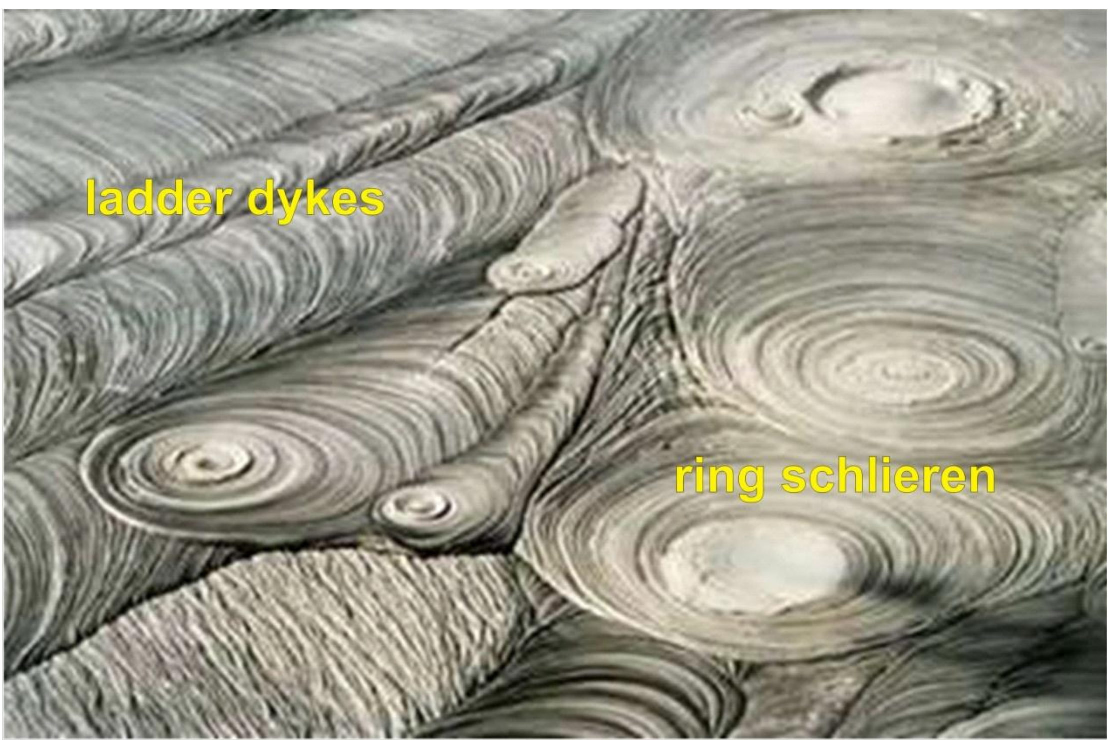

Fig. 1-22. Bubbling mud at Wai-O-Tapu Mud Pool, Taupo, New Zealand. Stationary bubble trains in mud pits produce remarkably good analogues of ring schlieren in granites. Migrating bubble trains advance as ring schlieren and leave ladder dykes in their wakes. Unfortunately no scale was possible for this photograph, but the large rings are roughly metre in diameter. Photograph used with permission of:
http://www.totaltravel.co.nz/travel/north-island/north-island/rotorua/photos/l28_mud

The ring schlieren also speak to the matter of timing of emplacement of the SMB relative to the neo-Acadian (~400 Ma) orogeny. The long axes of the elliptical ring schlieren align with the contact-parallel flow foliation, suggesting that flow may have had a role in the deformation of the rings (Fig. 1-23). If the elliptical shapes of the ring schlieren represent syntectonic deformation of formerly circular structures, the long axes should align parallel to the axial planes of the isoclinal folds in the Meguma Supergroup (NE-SW). That they do not provides further evidence that the SMB is post-tectonic.

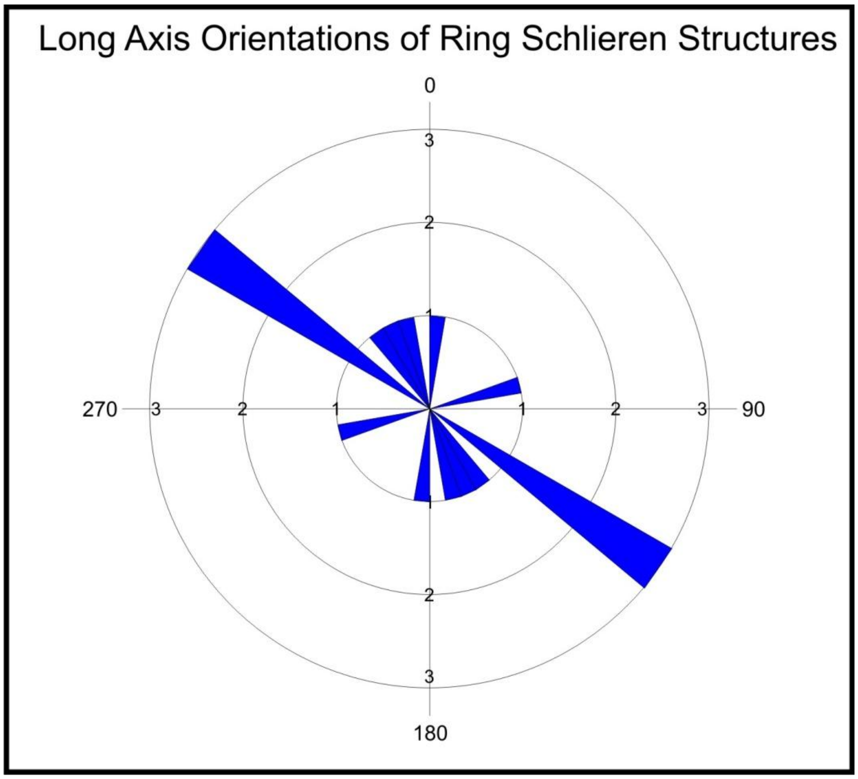

Figure 1-23. Rosette diagram of long-axis orientations of ring schlieren structures in the SMB (Sykes 2006).

As for the good sorting of mineralogical constituents at this locality, the best explanation is some type of shear flow in a vortex or bubble train. If one part of the magma is flowing faster than another, a velocity gradient exists in the melt. If a sheared liquid contains large blocky solids (e.g., Kspars), those solids will migrate to zones of minimum shear stress and may form clusters (cf. slurry pipes). If the sheared liquids contain small anisotropic flaky solids (e.g., biotite), those flaky solids may also become concentrated by shear flow and develop strong shape-preferred orientations, which we now see as schlieren. Overall, the modal proportion of minerals in the locally disrupted area seems to be about the same as that in the regional area, suggesting that there has only been a local sorting of mineralogical constituents, largely on the basis of their sizes and shapes.

STOP 1-7 – PEGGYS COVE: MORE TOURISTIC THAN GEOLOGIC (44.491692°N 63.918945°W)

DO NOT WALK ON WET ROCKS (unless it is raining), LEST YOU BE SWEPT AWAY BY A LARGE WAVE!!

With its quaint fishing harbour, iconic lighthouse, and abundance of gift shops, Peggys Cove is more of a tourist destination than a geological destination, although excellent panels discussing the geology do exist here. But even the tourists like to wander around on the rocks, and we can do the same to reinforce what we have already seen elsewhere today. Features of note are:

- prominent contact-parallel flow foliations, a continuation of what we saw at STOP 1-6
- abundant xenoliths in advanced states of assimilation
- late undulating aplites (are they folded?) and pegmatites (some mineralized with tourmaline)
- two prominent join sets oriented NW and NE
- roche moutonnée-style glacial erosion
- glacial striations

A good reference to the geology at this locality is Kontak et al. (2002).

Figure 1-23. The Peggys Cove lighthouse standing on peraluminous granites of the South Mountain Batholith.

STOP 1-8 – HWY 103: CORDIERITE-RICH FACIES OF THE SMB (44.703139°N 63.876063°W)

Introduction

High modal abundances of AFM minerals (Bt, Crd, Grt, And) occur at various localities in the SMB. The question is: are they the result of magmatic crystallization and concentration by some physical process such as flow or gravity, or are they the result of contamination, either as orthoxenocrysts or paraxenocrysts? This problem is the perennial chicken or egg conundrum, namely: is this rock now rich in AFM minerals because the original magma had high A/CNK, or does this rock now have high A/CNK because the magma was rich in (foreign) AFM minerals? What all these rocks with high modal abundances of AFM minerals have in common is a high concentration of alumina, and the richest source of high alumina is the pelites of the Halifax Group. Granites containing these high concentrations of AFM minerals also tend to contain one or more of the tell-tale high modal abundances of xenoliths (e.g., STOP 1-5 mafic porphyries), high modal abundances of sulphide minerals (e.g., STOP 1-2a biotite granodiorites and STOP 1-5 mafic porphyries), and low εNd, all suggestive of the role of the country rocks.

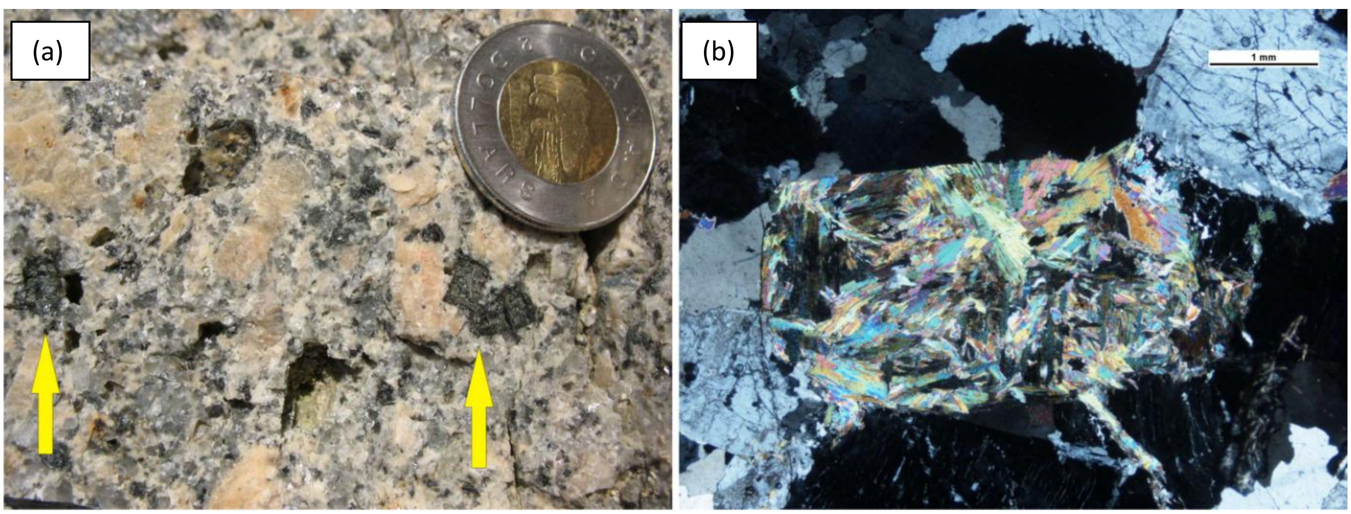

Figure 1-24. (a) Blocky cordierite at STOP 1-8. Coin diameter = 2.8 cm. (b)Photomicrograph showing that the cordierite is nearly 100% pinitized (Mus+Chl), perhaps by the same fluids that formed the fluorite veins.

At this locality, are the abundant cordierite grains (Fig. 1-24) orthoxenocrysts from the Meguma, the result of primary magmatic crystallization and accumulation, or are they paraxenocrysts created in incongruent melting reactions involving pelitic Meguma country rocks? All three types of origin are possible, and only a careful examination of the textures and compositions of the phases can potentially reveal the true origin, or perhaps origins, because in any single sample, it is possible that the phase in question may have more than one origin.

Discussion

Are these cordierites orthoxenocrysts? The cordierites at this locality are certainly not unmodified orthoxenocrysts. Their grain sizes, grain shapes, and lack of inclusions makes these cordierites highly unlikely to be porphyroblasts from country rocks in the contact aureole of the SMB. Of course, it is possible that these grains are modified orthoxenocrysts so that they have come into textural and chemical equilibrium with the SMB magma (Table 3 in Clarke 2007).

Are these cordierites magmatic? The cordierites at this locality could be magmatic. They have euhedral grain shapes and are free of inclusions as they might be if they had crystallized from a magma (Fig. 1-24). Meguma terrane granites are all peraluminous because they have inherited these characteristic from their source region, and their A/CNK ratios increased during closed-system fractional crystallization, open-system contamination, and evolution of fluids (Clarke et al. 2004). For the record, Acosta-Vigil et al. (2003) measured the A/CNK in H_2O-saturated synthetic granite melts in equilibrium with AFM minerals at 700^{o}C and 200 MPa:

Mineral	**A/CNK Necessary to Saturate the Melt**
andalusite	1.16
muscovite	1.27
cordierite	1.20
tourmaline	1.26

For comparison, the SMB magmas had A/CNK = 1.10-1.25, and so a magmatic origin for these phases is certainly possible, at least in terms of bulk rock A/CNK values.

But, if these cordierites are magmatic, then some concerns arise:

- why would a closed-system magma that had been crystallizing biotite from the outset, a process that decreases the FeO+MgO concentration <u>and</u> the A/CNK of the melt, abruptly begin to crystallize yet another AFM phase?
- to claim that magmatically crystallized grains become concentrated in isotopically contaminated rocks (Erdmann et al. 2009, Fig. 8) presents a challenge to the logic of chemistry and space
- the cordierites here are texturally distinct from the generally smaller, non-glomeroporphyritic, twinned magmatic grains in pressure-quenched aplitic sheets (STOP 1-9)

Are these cordierites paraxenocrysts? The third alternative is that the cordierites at this locality are the solid products of an incongruent melting of Meguma pelite, and several lines of evidence point in that direction:

- high modal proportions of AFM minerals (Crd, Grt, And) are **local** phenomena in the SMB, **not universal** phases such as biotite and muscovite – many of these local concentrations of AFM minerals are closely associated with the Meguma contact, suggesting that the country rock may have a role to play in their formation, e.g., biotite as Ostwald-ripened orthoxenocrysts (Fig. 1-10b), garnet as paraxenocrysts forming as the result of incongruent melting reactions, and cordierite forming by the same process as garnet (our experience at STOP 1-2a shows that pelites react readily to produce a biotite- and garnet-rich granodiorite)
- before Highway 103 was widened three years ago, we also saw some garnet with biotite reaction rims at this location – perhaps some keen observer will find more garnets today, suggesting that these are contaminated rocks, similar to those at STOP 1-2a
- this problem is like a classic Greek syllogism involving high modal AFM and Sr-Nd isotope correlation, namely: the cordierite-rich areas are isotopically contaminated, the putative Meguma contaminant has high A/CNK, ergo the AFM mineral (Crd) is related to contamination
- the spatial association of cordierite at the margins of xenoliths (Fig. 1-25) suggests Meguma involvement

- these large biotite and cordierite grains on the margins of xenoliths cannot be quench products against a cold xenoliths because the xenoliths are already rounded (reacted) and cannot have represented thermal anomalies, and because quenching should produce many nuclei and small grain sizes– instead, the textures suggest that the xenoliths have melted and they were releasing their Ostwald-ripened (grain-coarsened) biotites and peritectic cordierites to the surrounding SMB melt
- with their sizes, shapes, aggregation, and oscillatory zoning (Erdmann et al. 2004), these cordierite grains are texturally anomalous – the common glomeroporphyritic clumping and oscillatory zoning are consistent with growth in an incongruent melting reaction where the zoning reflects alternating influence of the SMB melt and the partial melt
- some cordierites have reaction rims of andalusite + biotite on them, suggesting that they were not in equilibrium with the SMB melt
- rusty spots are developing on this outcrop suggesting an anomalous concentration of sulphides in these cordierite-rich rocks

So, the combination of coarsened xenocrystic biotites and growth of peritectic cordierite and garnet can make foreign material look cognate. Thus high concentrations of Bt, Crd, and Grt are likely from contamination.

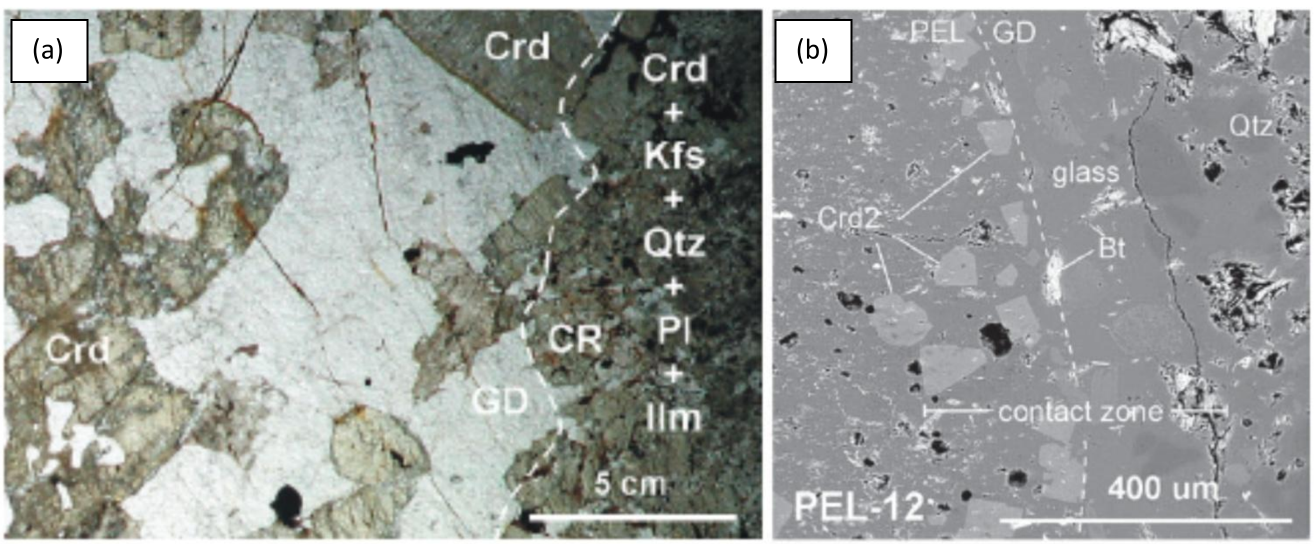

Figure 1-25. The formation of cordierite in nature and in experiment. (a) Meguma xenolith on right side of photomicrograph growing large blocky peritectic cordierite "porphyroblasts" on its margin that drift away as paraxenocrysts into the granite on the left side. (b) Pelitic fragment on left side of photomicrograph growing large peritectic cordierite "porphyroblasts" on its margin that will eventually drift away as paraxenocrysts into the granite magma on the right. (Modified interpretation from Erdmann et al. 2007)

Conclusion

Erdmann et al. (2009) have made the case for a magmatic origin for cordierite in the SMB, whereas García-Moreno et al. (2007) and Ugidos et al. (2009) have made the case for a contamination/assimilation origin for cordierite in Iberian granites. I regard the high modal concentration of cordierite at this location as the last vestige of an assimilated slab of Halifax Group pelite (like the ice shelf analogue in Fig. 1-9), just as the high concentration of garnet (and biotite) at STOP 1-2a was the last vestige of assimilation of pelite. We shouldn't worry that one case is on the decimetre scale and the other case is on a kilometre scale, because Horne (1993) shows Meguma roof pendants on this scale in the SMB. Here at STOP 1-8, we are several kilometres horizontally from any known contact , but we may be looking at the remnants of a roof pendant where nothing is left *physically* of the pelite except for a high concentration of cordierite. The pelite, like the Cheshire cat in "Alice in Wonderland", disappeared leaving nothing but its *chemical* smile, except in this case that smile has some nice cordierite teeth.

Once we are able to make the critical field-petrographic-geochemical link between one AFM mineral (garnet) and pelitic contamination, then we are bound to give consideration to the idea that the same kind of link may exist for both other AFM minerals (biotite and cordierite). Indeed, contamination must become the default interpretation, and we certainly do see high modal proportions of biotite and cordierite associated with disintegrating Meguma xenoliths. The difficult part comes when you can no longer see the xenolith, just the remaining refractory AFM phase. And that is the nub of this petrogenetic problem.

STOP 1-9 – HWY 103: MAGMATIC CORDIERITE IN APLITE (44.710941°N 63.899981°W)

The SMB contains many small intrusions, including aplite-pegmatite dykes. These bodies have essentially the same bulk chemical compositions as the coarser grained rocks of the batholiths, but they commonly have significant modal abundances of andalusite, cordierite, and/or garnet, phases normally absent in their host monzogranites. Stripping of alkalies from the melt into a water-rich pegmatitic fluid can cause a concomitant dramatic increase in the A/CNK of the remaining melt. The AFM phase that crystallizes from such pressure-quenched aplites depends on a complex function of A/CNK, total Fe+Mg+Mn, temperature, and pressure.

At this locality is a subhorizontal aplitic sheet. It contains abundant cordierite (Fig. 1-26) that is texturally completely different from that at nearby STOP 1-8. These quenchy-looking cordierites are almost certainly magmatic in origin. Some euhedral cordierites are also present in this outcrop (Fig. 1-27).

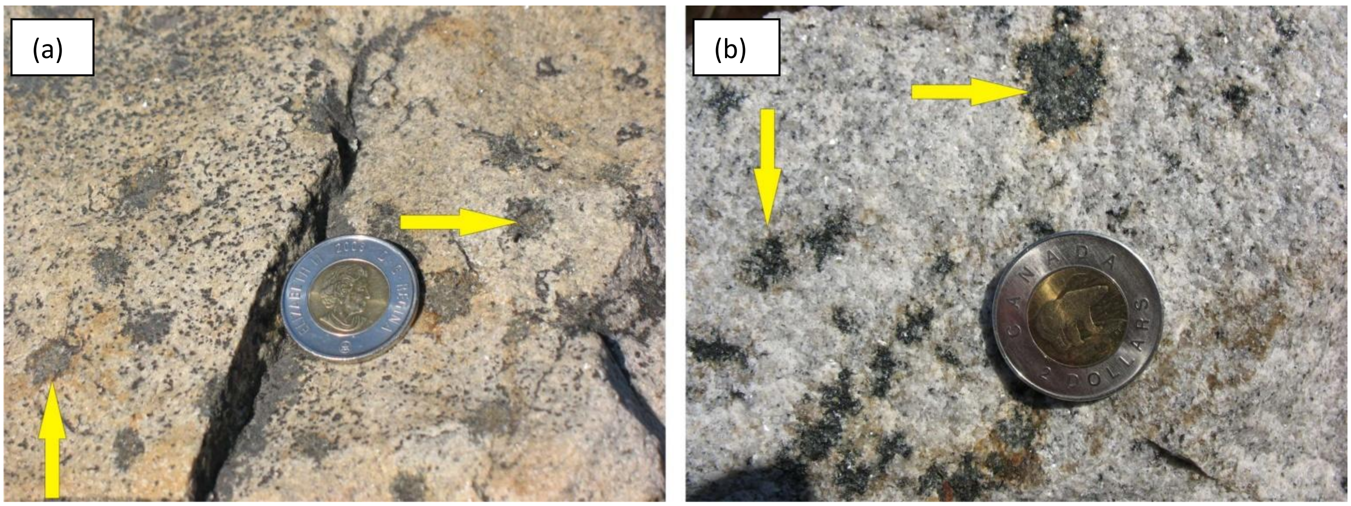

Figure 1-26. (a,b) Cordierite splotches in the aplite at STOP 1-9.

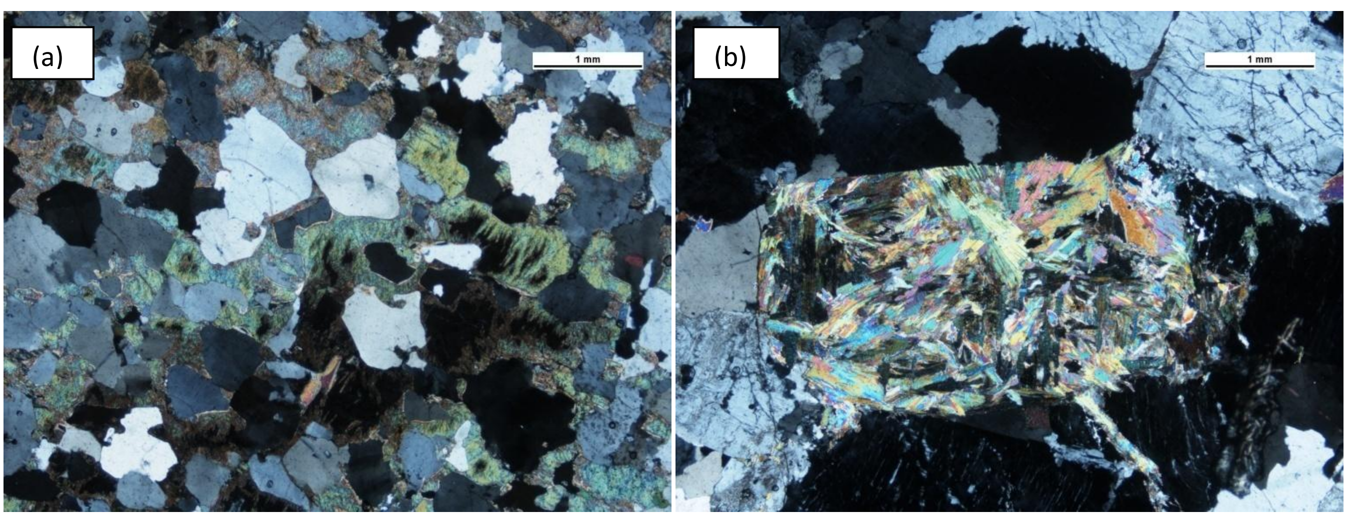

Figure 1-27.(a) Section through large splotchy pinitized poikilitic cordierite of Fig. 1-26.The only inclusions are quartz, suggesting that this mineral was the only phase on the liquidus when cordierite crystallized. (b) This aplite also contains some euhedral cordierite grains that may be of the same peritectic origin as at STOP 1-8.

MEGUMA TERRANE SUMMARY

What we have seen today is generally representative of the Meguma lithotectonic terrane, namely Cambrian-Ordovician proximal and distal turbidites cut by peraluminous granites. Elsewhere in the Meguma Terrane, the regional grade of metamorphism of the metasedimentary rocks may vary up to amphibolites facies, and the composition of the granitoid rocks may vary from tonalite to alkali-feldspar granite, but the similarities outweigh the differences.

Paul Schenk led the way in searching for the source of the Meguma turbidites and postulated that they were deposited on the passive margin of Gondwana. With a ~200-km-wide stretch of the Meguma terrane buried on the Scotian Shelf, it is difficult to make convincing stratigraphic correlations with the source region or correlatives, be they Amazonia, Africa, Iberia, or even Walesia. Nevertheless, it is clear that the Meguma terrane was the last terrane to dock with what is now North America, and it did so along a dextral transpressive fault, the Cobequid-Chedabucto Fault system during the “neo-Acadian” Orogeny about 400 Ma ago.

Many peraluminous granite plutons, including the South Mountain Batholith, intruded this terrane about 20 my after the deformation of the Meguma Supergroup. The SMB is a post-tectonic batholith with a variety of internal structures, including layering and ring schlieren, that provide evidence of magma chamber processes, including its emplacement by stoping. The SMB has also been extensively contaminated by psammitic to pelitic rocks of the Meguma Supergroup providing abundant textural and chemical evidence of the assimilation processes.

ACKNOWLEDGEMENTS

Many thanks to Jim Jordan, Manager of La-Z-Boy Furniture Galleries in Bayers Lake Industrial Park (STOP 1-2), Jim and Frances Wolfe (STOP 1-3), and Ernie and Rachelle Eisler (STOP 1-5) for so generously granting us access to their properties. So much of what I know the Meguma Terrane and its granites has been learned with, or from, my valued colleagues in the various universities of Nova Scotia, my many students from 1970 to 2010, and those who provided world-class technical and analytical support. In addition, I am grateful to John Waldron for his helpful review of the section on the Meguma Supergroup. Ultimately, none of the knowledge we have gleaned from the rocks would have been possible without the unbroken financial support of NSERC over the decades.

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CHAPTER 2: THE COBEQUID HIGHLANDS

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OVERVIEW OF THE AVALON TERRANE

The tectonostratigraphic zones or terranes of the Appalachians (e.g., Williams 1979) record a protracted Paleozoic accretionary history involving major oblique collisional events. Those events marked the sequential docking to Laurentia of mostly peri-Gondwanan outboard terranes and microcontinents, represented by the Gander, Avalon, and Meguma terranes, and the closure of the Iapetus and Rheic oceans. The Iapetus Ocean opened by rifting of Rodinia at the end of the Neoproterozoic and separated Gondwana from Laurentia. The Rheic Ocean had opened when Avalonia rifted from Gondwana during the Early to Middle Ordovician, and it became the host to several small continental terranes, including Meguma. Final closure of Iapetus took place in the early Silurian with the accretion of Ganderia to Laurentia. Convergence of Ganderia and Avalonia took place in the late Silurian–early Devonian, and the accretion of Meguma is recorded by mid-Devonian deformation. Continuing oblique convergence and strike-slip motion along boundaries between the previously assembled terranes is recorded in diverse and protracted tectonism during the Late Devonian to Permian, with the Carboniferous arrival of Gondwana and the formation of the Pangaean supercontinent (van Staal, 2005).

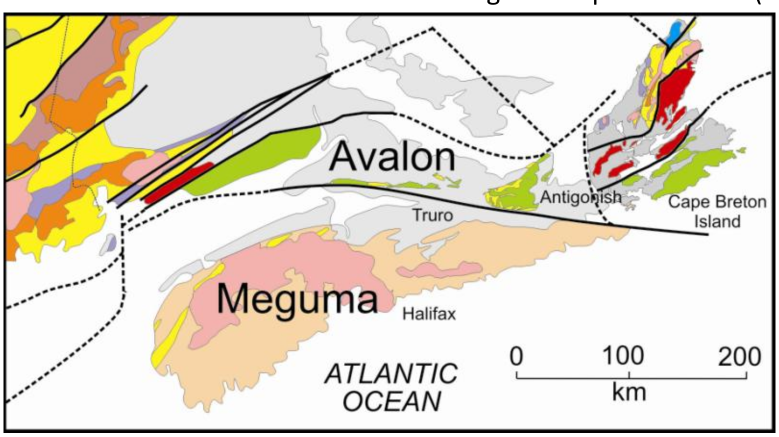

Figure 2-1. General map of the distribution of the Avalon (sensu stricto) terrane (light green colour) in Nova Scotia and New Brunswick. (From S. Barr, pers. comm. 2005).

The exposed rocks of the Avalon Zone mainly comprise a package of Neoproterozoic, dominantly juvenile, arc-related volcanosedimentary successions and associated plutonic rocks that experienced a complicated and long-lived Neoproterozoic tectonic history. These rocks are overlain by a distinctive Cambrian–Ordovician shale-rich platformal sedimentary sequence that locally includes rift-related volcanic rocks. The Caledonia terrane of southern New Brunswick and similar rocks in the Cobequid and Antigonish Highlands and southeastern Cape Breton Island have been termed Avalon terrane *sensu stricto*, to distinguish them from the Composite Avalon Terrane that also includes various inboard terranes and blocks of uncertain tectonostratigraphic affinity, both in southern New Brunswick and Cape Breton Island. Within the Avalon Zone *sensu stricto*, several tectonic blocks (or terranes) with rather different Neoproterozoic tectonic histories have been distinguished.

From **STOP 2-1**, we can see the Cobequid-Chedabucto fault line and behind it the Avalon terrane of the Cobequid Highlands. This fault was part of an active NE-trending dextral strike slip fault system (the Cobequid Shear Zone) in the late Devonian and early Carboniferous. It was reactivated as part of a series of E-W trending dextral strike-slip faults in the late Carboniferous Alleghenian orogeny, resulting in clockwise rotation of the Cobequid Highlands horst. The Cobequid fault was a predominantly dip-slip, basin-margin fault in the Triassic (Withjack et al. 1995) and took up dextral shear deformation in the mid Cretaceous (Pe-Piper and Piper 2004).

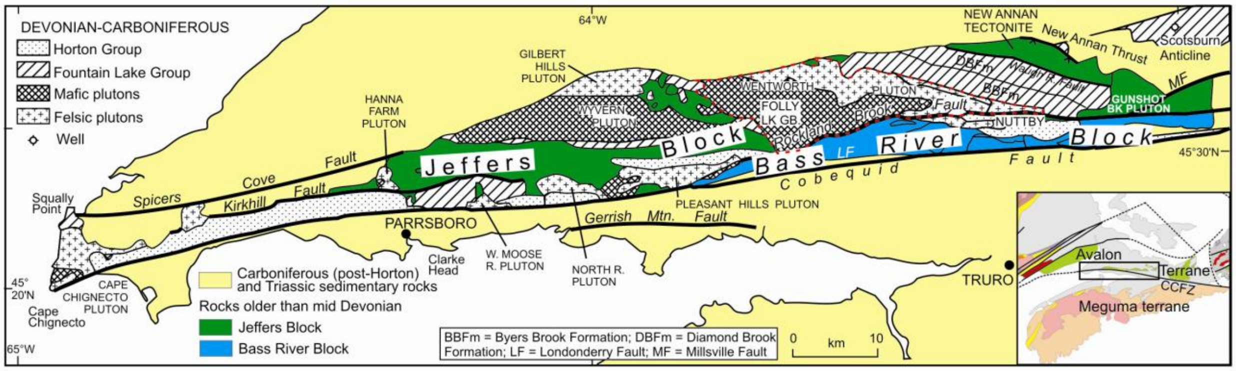

Figure 2-1. General geological map of the Cobequid Highlands, showing the distribution of latest Devonian–earliest Carboniferous plutons, volcanic rocks, and Horton Group sedimentary rocks. (Modified from Pe-Piper and Piper 2003).

INTRODUCTION TO THE COBEQUID HIGHLANDS

The Cobequid Highlands of Nova Scotia (Pe-Piper and Piper, 2003; Fig. 2-2) constitute a horst, originally developed in the latest Devonian, exposing Neoproterozoic Avalon terrane rocks (Fig. 2-3) on the northern side of the Cobequid Fault (as viewed from **STOP 2-1**). The Neoproterozoic rocks consist of arc volcanic and plutonic rocks and associated sedimentary rocks in two major blocks. The geology of the Jeffers block in the west and north is similar to that of the Antigonish Highlands (**Day 3**), with volcanic rocks and turbidites intruded by granodioritic and granitic plutons. The Bass River block, in the southeast, will be seen in **STOPS 2-6 to 2-9**; it shows evidence of shear-zone emplacement of a suite of plutonic rocks ranging from gabbro to granodiorite to granite into tectonically juxtaposed ocean floor volcanic and shelf sedimentary rocks. Not included in either block is the Economy River Gneiss, a geochemically distinct granodioritic orthogneiss that has yielded a U-Pb zircon age of 734 Ma (Doig et al. 1993).

The Cobequid Highlands are dominated by E-W trending faults of the Late Paleozoic Cobequid Shear Zone, developed during oblique convergence between the Avalon and Meguma terranes, that facilitated the emplacement of gabbro and granite plutons and their extrusive equivalents (Fountain Lake Group) in the latest Devonian - earliest Carboniferous (**STOPS 2-2 to 2-5**). The most important fault, the NE-trending dextral Rockland Brook Fault, marks the boundary between the Bass River and Jeffers blocks. Lower Paleozoic sedimentary rocks, including Silurian to early Devonian shelf sandstone and siltstone, were highly dismembered by this faulting associated with terrane convergence. The youngest stratified rocks within the Cobequid Highlands comprise fluvial and lacustrine sedimentary rocks of the latest Devonian - Tournaisian Horton Group, which accumulated in fault-bound basins.

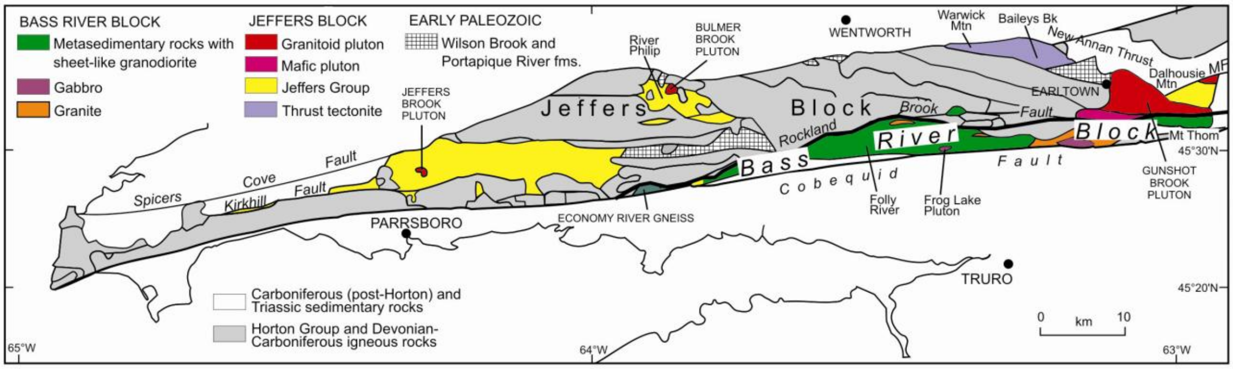

Figure 2-2. Map of the Cobequid Highlands showing distribution of Avalon terrane plutonic rocks. (Modified from Pe-Piper and Piper 2003).

NEOPROTEROZOIC PLUTONISM OF THE BASS RIVER BLOCK

The Bass River block (Nance and Murphy 1990) consists of a series of fault slices of Neoproterozoic shelf siliciclastic sedimentary rocks (Gamble Brook Formation: **STOP 2-6**), tectonically juxtaposed with ocean-floor basalt (**STOP 2-8**) and a thin cover of pelagic sedimentary rocks and fine-grained turbidites (Folly River Formation: Pe-Piper and Murphy 1989), and intruded by a series of plutonic rocks. The Gamble Brook Formation contains detrital zircon grains as young as 1183 Ma (Keppie et al. 1998), thus placing a maximum age on the formation. Individual plutons are difficult to define because of the tectonic dismemberment of the Bass River block, but three principal plutonic rock types are distinguished (Pe-Piper et al. 1996a): the Frog Lake gabbro/diorite unit (**STOP 2-6**), the Debert River granodiorite unit (**STOP 2-7**) and the McCallum Settlement granite unit (**STOP 2-9**). The first two units occur principally as sheets tens to hundreds of metres thick within the Gamble Brook Formation, with structures indicating emplacement during shearing (Pe-Piper et al. 1996a, 2010). Geochronology by $^{40}Ar/^{39}Ar$ on hornblende (Keppie et al. 1990) and U-Pb on zircon (Doig et al. 1991) showed that the Frog Lake gabbro/diorite is about 622 Ma and the Debert River granodiorite 609 - 605 Ma. Poor quality U-Pb and Rb-Sr ages of ca. 575 Ma (Doig et al. 1991) have been obtained from the McCallum Settlement granite unit, which cuts the Debert River granodiorite.

In the Frog Lake pluton, low-Ti hornblende gabbros have trace element abundances similar to subduction-related low-K mafic rocks, including some enrichment in large-ion lithophile elements and marked relative depletion in Nb and Y. High-Ti hornblende gabbros and pyroxene-mica gabbro show more alkaline characteristics, with higher amounts of Nb, Y, P_2O_5, and high-field-strength elements. Tonalite and granite veins within the Frog Lake pluton are geochemically similar to volcanic arc granite (Pe-Piper et al. 2010). Pe-Piper et al. (1996a) interpreted the sequence in the Bass River block as having formed by tectonic accretion at a convergent plate margin, with the plutonic rocks emplaced along shear zones during oblique convergence. These shear zones also provided pathways for rapid rise of water-rich mafic magma to the Frog Lake pluton. Murphy (2002) showed from sedimentary geochemistry that the Gamble Brook Formation formed in a rifted arc environment.

We will not visit the Jeffers block on this field trip. The Jeffers block consists principally of the Jeffers Group, a series of volcanic and volcaniclastic sedimentary rocks at least several hundred metres thick (Pe-Piper and Piper 1989). Its age is Late Neoproterozoic based on a U-Pb age on zircon of 630±2 Ma from rhyolite (Murphy et al. 1997). Cross-cutting plutons are of granite (Gunshot Brook) and granodiorite (Bulmer Brook, Jeffers Brook); the latter was dated by $^{40}Ar/^{39}Ar$ on hornblende at 605-607 Ma (Keppie et al. 1990).

Comparisons of the Avalonian tectonic and igneous events (Fig. 2-4) can be made between the Bass River and Jeffers blocks of the Cobequid Highlands and the Caledonia terrane of southern New Brunswick, the Georgeville block of the Antigonish Highlands, and the Mira terrane of southern Cape Breton island. In most of these areas, there is a long and complex history of Late Neoproterozoic igneous activity. The oldest rocks are the 734 Ma Economy River orthogneiss, a kilometre-wide sliver along the Cobequid Fault. No volcanic rocks are known from the Cobequid Highlands of similar age to the 680 Ma Stirling Group of the Mira terrane. The oldest volcanic rocks are the 630 Ma Jeffers Group, which are a little older than the ca. 620 Ma volcanism of the Mira terrane (East Bay Hills, Coxheath and Pringle Mountain groups) and the 615-610 Ma Keppoch Formation in the Antigonish Highlands, but may be of similar age to the Broad River Group of the Caledonia terrane. In the Bass River block, the 622 Ma Frog Lake gabbro is of similar age to widespread dioritic to granodioritic plutons in the Caledonia and Mira terranes. Similar plutonism in the Antigonish Highlands is regarded as being a little younger (618-611 Ma) and the 609-605 Ma Debert River granodiorite of the Bass River block is younger still. The 605-607 Ma plutons of the Jeffers block are of similar age to the Debert River granodiorite. Metamorphism (590-600 Ma) and intrusions (580 Ma) in the "Great Village River Gneiss" and the McCallum Settlement granite (ca. 575 Ma) are a little older than the Fourchu and Main à Dieu volcanics and associated plutons of the Mira terrane (ca. 575-560 Ma) and the Coldbrook Group volcanics and the younger plutons of the Caledonia terrane (560–550 Ma). The closest regional similarities are between the Jeffers block of the Cobequid Highlands and the Georgeville block of the Antigonish Highlands. The Bass River block, which lacks volcanic rocks related to plutonism, appears to be at a deeper crustal level than either of these blocks or the Caledonia and Mira terranes. From its present geometry, it might have been the most inboard of these peri-Gondwanan terranes.

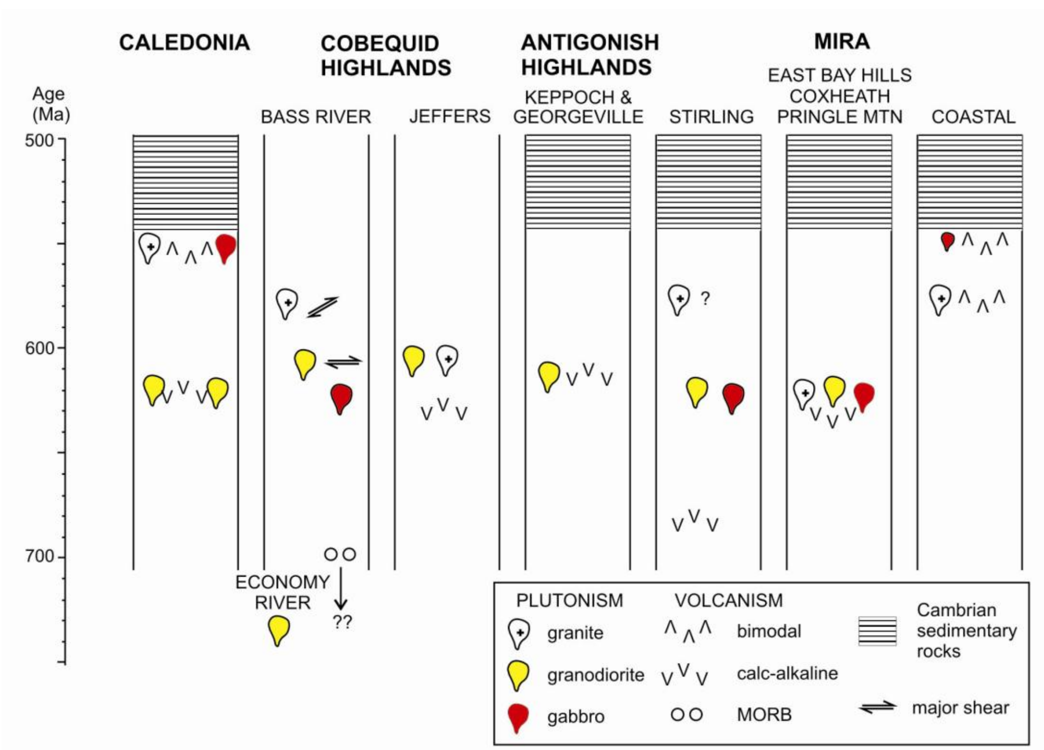

Figure 2-3. Chronology of igneous activity in different parts of the Avalon Terrane (sensu stricto) in New Brunswick and Nova Scotia. (Modified from Pe-Piper and Piper 2003)(cf.Fig. 4-4 for Mira update).

LATE DEVONIAN–EARLY CARBONIFEROUS IGNEOUS ROCKS OF THE COBEQUID HIGHLANDS

Plutonic rocks

Much of the Cobequid Highlands is underlain by plutonic rocks of latest Devonian to earliest Carboniferous age. The main plutonic rock types are granite and gabbro/diorite. In places, the felsic and mafic rocks appear to have been emplaced as co-existing magmas (Pe-Piper et al. 1996b) or as sheet-like alternations. Hybrid granitic/mafic rocks show mingling but incomplete mixing, whereas granodiorite/tonalite has been interpreted to result from complete mixing of the two magmas (Pe-Piper et al. 1996b). Large gabbro/diorite plutons contain variable amounts of granite in irregular patches and dykes. Regional magnetic and gravity data suggest that plutonic rocks extend south of the Cobequid Fault. At Clark Head, near Parrsboro, blocks of Devonian-Carboniferous gabbro-granulite outcrop in a fault breccia (Gibbons et al. 1996).

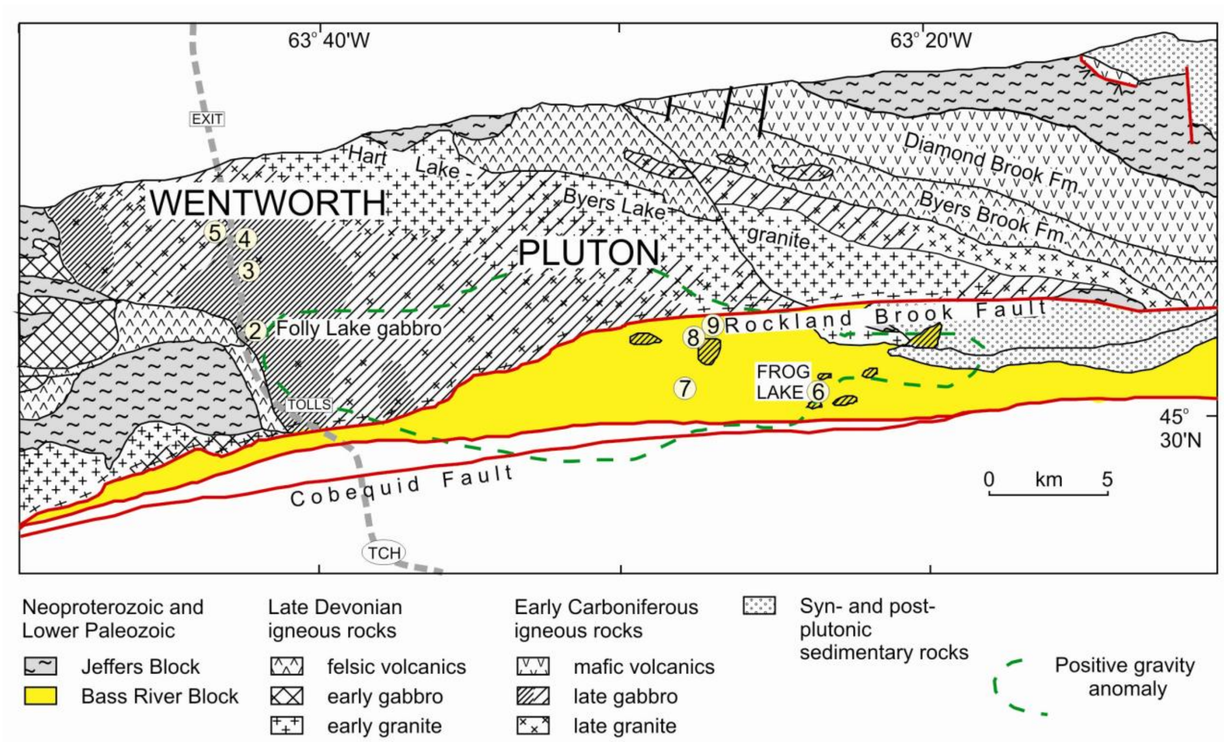

Figure 2-4. Map of the Wentworth Pluton showing STOPs 2-2 to 2-9. (Modified from Koukouvelas et al. 2002).

The three phases of pluton emplacement are:

(1) The main phase of plutonism in the Cobequid Highlands includes granite plutons of the western Cobequid Highlands and their associated gabbro (Koukouvelas et al. 1996), the principally mafic Wyvern Pluton, and the granite in the northern part of the Wentworth Pluton (the Hart Lake - Byers Lake pluton of Donohoe and Wallace 1982; Fig. 2-5). These units have all yielded ages of 362 to 358 Ma by U-Pb on zircon (Doig et al. 1996) or by $^{40}Ar/^{39}Ar$ on amphibole (Pe-Piper et al. 2004) (Fig. 2-6). The emplacement of these plutonic rocks thus straddles the Devonian-Carboniferous boundary. The granitic Gilbert Hills pluton in the north central Cobequid Highlands is probably of similar age.

(2) The Whirley Brook intrusion, a complex body made up of sheets of fine-grained granite and granitic porphyry, located between the northeastern Wentworth pluton and the Byers Brook Formation (Figs. 2-5, 2-7). Irregular enclaves and sheets of diabase occur within this fine-grained granite, with textures suggesting that the

mafic magma was immiscible (cf. Pe-Piper et al. 1996b). The Whirley Brook intrusion is marked by a strong magnetic anomaly (Piper et al. 1993), suggesting the presence of voluminous mafic rocks at shallow depth. The granitic rocks have a high Zr content, similar to distinctive high-Zr rhyolite found only at the extreme top of the Byers Brook Formation (Piper et al. 1999), with an inferred age of about 355 Ma (earliest Carboniferous).

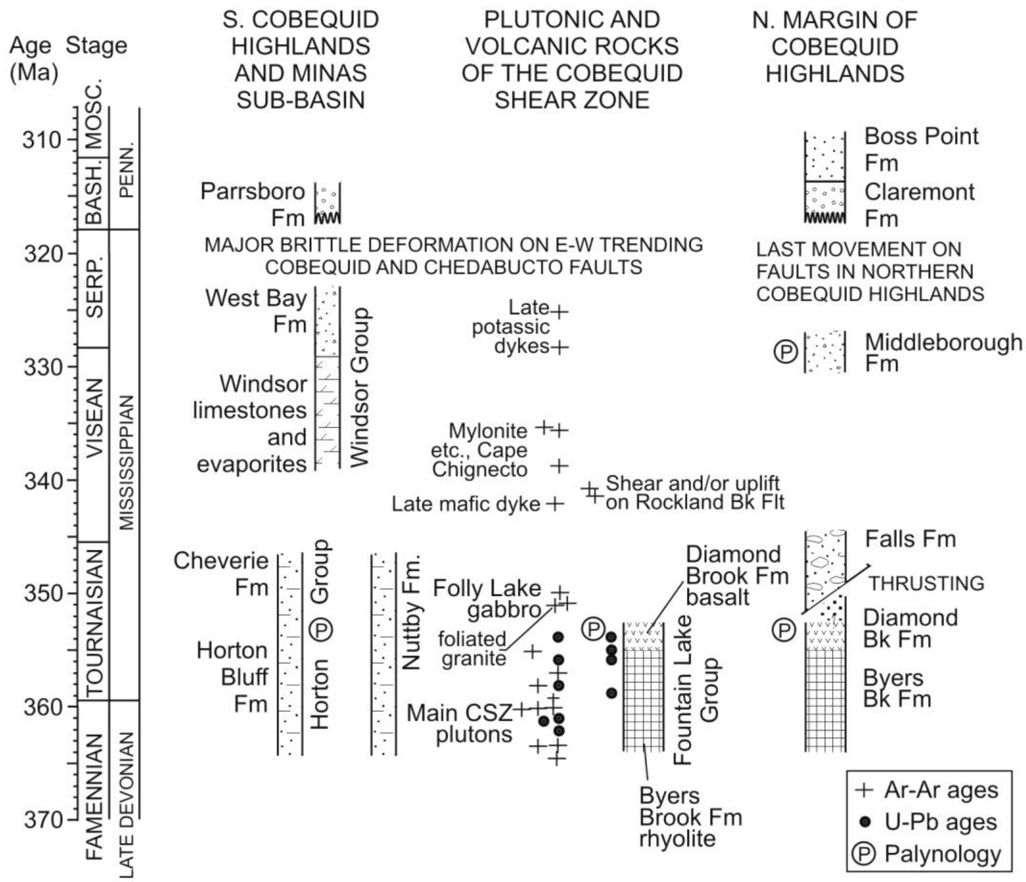

Figure 2-6. Summary of chronologic control for the Upper Devonian–Lower Carboniferous rocks of the Cobequid Highlands. (Modified from Pe-Piper et al. 2004).

(3) Gabbro/diorite of the southwestern Wentworth Pluton (Koukouvelas et al. 2002), termed the Folly Lake pluton by Donohoe and Wallace (1982) (Fig. 2-5). These rocks consistently yield $^{40}Ar/^{39}Ar$ ages on amphibole and biotite of 350-354 Ma (Nearing et al. 1996), dating from the early Carboniferous and synchronous with the extrusion of the voluminous basalt of the Diamond Brook Formation (Figs. 2-6, 2-7). Much of the granite of the Wentworth pluton within a few kilometres of the Folly Lake gabbro either cross-cuts the gabbro or shows textures indicating the co-existence of immiscible mafic and felsic magmas (Pe-Piper et al. 1996b, Koukouvelas and Pe-Piper 1995, 1996). Koukouvelas et al. (2002) argued on the basis of textural, geochemical and radiometric data that these transition zone granites were the consequence of melting of “early” Hart Lake - Byers Lake granite by the Folly Lake gabbro/diorite.

Minor late units of the plutons consist principally of dykes of porphyritic rhyolite and diabase. Some felsic dykes have yielded U-Pb zircon ages of 356 Ma (Dunning et al. 2002); some mafic dykes appear to be as young as 328 Ma (Nearing et al. 1996). Abundant gabbro and diabase dykes occur south of Squally Point cutting Fountain Lake Group (Piper et al. 1993). South of the Rockland Brook Fault, gabbro sills make up perhaps 50% of the section in the Nuttby Formation and intrusive pods of gabbro are common in the Neoproterozoic rocks westward to Folly River. The age of these abundant mafic intrusions is not well constrained, but they probably correlate with basalt in the upper Horton Group of Prince Edward Island (Pe-Piper and Piper 1998b; Dunning et al. 2002), or with early Viséan gabbro dated at 338 Ma at St Peters in southern Cape Breton Island (Barr et al. 1994).

Volcanic rocks

The Fountain Lake Group appears to be the extrusive equivalent of the late Devonian–early Carboniferous plutons. In the eastern Cobequid Highlands, it is divisible into two formations (Figs. 2-5, 2-6, 2-7). The underlying Byers Brook Formation consists of felsic pyroclastic rocks, lesser felsic and mafic flows, and minor interbedded sedimentary rocks, in total several kilometres thick (Piper et al. 1999). A rhyolite flow near the top of the formation has yielded a U-Pb zircon age of 358 Ma (Dunning et al. 2002). The conformably overlying 1.5 km thick Diamond Brook Formation (Dessureau et al. 2000) consists of basalt flows lesser thicknesses of interbedded sedimentary rocks and rhyolite in the lower part of the formation and fluvial conglomerate and sandstone at the top of the formation. Siltstone in the middle of the basaltic facies of the Diamond Brook Formation are Late Famennian (Martel et al. 1993) and Mid Tournaisian (early Tn3) spores were identified in the upper part of the formation in the Scotsburn #1 well (Utting et al. 1989). These ages are consistent with a U-Pb age of 354 Ma on a rhyolite in the lower part of the formation (Dunning et al. 2002). In the western Cobequid Highlands, the Fountain Lake Group consists of alternating rhyolite and basalt and is less than 0.8 km thick. U-Pb zircon ages of 356 and 355 Ma have been obtained from rhyolite (Dunning et al. 2002).

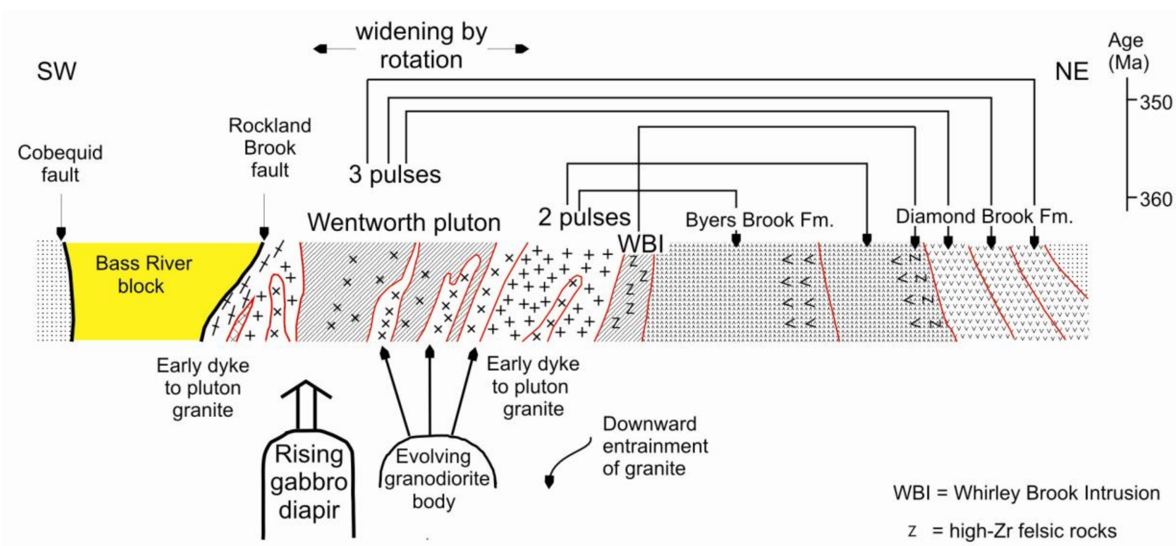

Figure 2-5. Summary cartoon of age relationships between the Wentworth Pluton and the Fountain Lake Group. (Modified from Koukouvelas et al. 2002).

Mineralogy and geochemistry of the Late Paleozoic igneous rocks

The basalt of the Diamond Brook Formation geochemically resembles continental flood basalts, with iron-enrichment and other geochemical features indicating derivation from plume-related melting in the asthenosphere, modified by passage through sub-continental lithospheric upper mantle (Dessureau et al. 2000). Gabbro intrusions of the Cobequid Shear Zone are of olivine tholeiite composition and appear geochemically related to the basalt. The granite plutons have the geochemical character of A-type granite (Pe-Piper et al. 1991). Nd and Pb isotope data show that all the Cobequid Shear Zone granites appear derived from melting of crust compositionally similar to Neoproterozoic granodiorite (Pe-Piper and Piper 1998b).

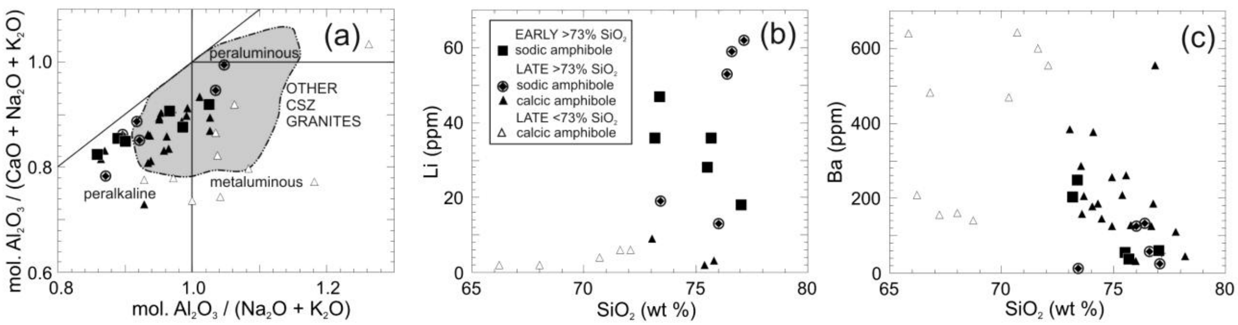

Figure 2-6. Selected geochemical features of the Wentworth Pluton granites, showing contrasts between early and late granites and those with sodic or calcic amphiboles. In (a) the grey tone shows other granites of the Cobequid Shear Zone. (Modified from Pe-Piper 2007).

The Wentworth pluton has higher alkalinity and lower aluminum saturation index than other granites of the Cobequid Shear Zone (Fig. 2-8) and is richer in F than the other granites (Koukouvelas et al. 2002). All these features imply anhydrous melting of lower crust. The presence of the voluminous Folly Lake gabbro with its associated gravity anomaly and the unusually thick coeval basalt and felsic pyroclastic rocks near the Wentworth pluton both suggest the presence of a major heat source in this area, which resulted in more anhydrous melting here than elsewhere in the Cobequid Shear Zone. Along most of the Cobequid Shear Zone, the latest Devonian partial melting of lower crust to form granitic magma under more hydrous conditions was synchronous with development of Horton Group basins and was probably triggered by ingress of voluminous water from lakes through crustal-scale faults of the shear zone (Pe-Piper et al. 1998). Rhyolites of the Fountain Lake Group are geochemically similar to the granites (Pe-Piper et al. 1991; Piper et al. 1999).

The peralkaline character of the Wentworth pluton is reflected in its mineralogical composition, notably the presence of sodic amphiboles (Pe-Piper, 2007) and minerals rich in REE, Y and Nb. Magmatic sodic amphiboles are arfvedsonite, ferrorichterite and katophorite. Ferrowinchite and riebeckite developed by subsolidus oxidation reactions with loss of fluorine. Some granites have calcic amphiboles. The presence of sodic or calcic amphibole in Wentworth pluton granite shows no systematic difference with either host-rock geochemistry or distance from the gabbroic heat source, except that rocks with sodic amphibole have high Li. The first REE mineral to form is chevkinite, followed by allanite. Later action of fluids caused the alteration and partial leaching of the older REE-bearing magmatic minerals (e.g. allanite, chevkinite) and the precipitation of a younger generation of Y-HREE-Th-Nb-Ta-Ti-rich minerals (thorite, hingganite, samarskite, and aeschynite).

Structure of the late Paleozoic Cobequid Shear Zone

The Cobequid Shear Zone, exposed throughout the Cobequid Highlands horst, consists of late Paleozoic, crustal scale, south-dipping, strike-slip faults developed during amalgamation of the Avalon and Meguma terranes (Donohoe and Wallace 1982, 1985; Miller et al. 1995; Durling 1996; Murphy and Keppie 1998). These faults were active during latest Devonian to earliest Carboniferous pluton intrusion and are cut by the younger Cobequid Fault, which experienced major late Carboniferous right-lateral strike-slip motion under brittle conditions (Murphy et al. 1999) and several kilometres of Triassic dip-slip motion (Withjack et al. 1995). The Rockland Brook fault is in part a reactivated Neoproterozoic shear zone between the Bass River and Jeffers blocks (Pe-Piper et al. 2002) that involved amphibolite facies metamorphism of mafic volcanic rocks as late as 589 Ma. This Neoproterozoic lineament was reactivated in the central Cobequid Highlands, but not in the extreme eastern Cobequid Highlands, where late Paleozoic motion was taken up on the Cobequid and Millsville faults (Fig. 2-2) (Pe-Piper et al. 2002). The Cobequid Fault may in part follow another Neoproterozoic lineament between the Bass River block and the Economy River gneiss.

Well exposed latest Devonian - earliest Carboniferous plutons occur along two dextral fault strands and an en echelon transition zone between them in the southern part of the Cobequid Shear Zone. The Kirkhill Fault strand is a fault 60 km long, now truncated at its western end by Triassic basin margin faults. On its northern side is the large Cape Chignecto pluton in the west and a series of smaller plutons to the east. The 60-km-long Rockland Brook Fault strand (Miller et al. 1995) merges at its western end with the younger Cobequid Fault and in its central portion dips about 45^{o}S (Durling 1996). Almost all the Rockland Brook Fault shows intense magmatic activity, with the large Wentworth pluton north of the central part of the fault (Koukouvelas et al. 2002) and the smaller Pleasant Hills pluton at its western end (Pe-Piper et al. 1998). Between these two prominent faults, in the transition zone, the deformation was accommodated by several faults including two shear zones within the western Pleasant Hills Pluton (Pe-Piper et al. 1998), a fault along the southern margin of the North River Pluton, and the fault along the southern margin of the West Moose River Pluton. Significant ductile slip took place on all these faults in the latest Devonian - earliest Carboniferous, associated with NW-vergent thrusting and some wrench faulting (Koukouvelas et al. 1996), and space for pluton emplacement was created by lateral translation of blocks within the shear zones (Pe-Piper et al. 1998; Koukouvelas et al. 2002) and by thrusting and wall-rock compression (Waldron et al. 1989; Piper et al. 1996a; Koukouvelas et al. 1996). Along the major fault zones, highly deformed earlier emplaced plutonic rocks underwent strong solid-state deformation and are cut by progressively less deformed younger plutonic rocks, demonstrating that the faults were persistent conduits for magma (Pe-Piper et al. 1998; Koukouvelas et al. 2002). The distribution of gabbro, the presence of more deformed rocks cut by less deformed rocks, and the wide mylonite zone indicate that the Rockland Brook fault was the master fault of a series of fault conduits for magma that were active during pluton emplacement (Koukouvelas et al. 2002). Most plutons occur north of the Rockland Brook fault, but the Gain Brook granitic pluton and an un-named granitic body south of Earltown intrude the Bass River block (Pe-Piper et al. 2002).

The Rockland Brook - Kirkhill fault system marks the boundary between thick sandstones and mudstones of the Nuttby Formation to the south and thick volcanic rocks of the Fountain Lake Group to the north. Both biostratigraphic and radiometric data (Dunning et al. 2002) indicate that the Nuttby Formation and the Fountain Lake Group are essentially of the same age - late Famennian to middle Tournaisian (ca. 362 to 354 Ma) (Fig. 2-6). The lack of volcanic rocks in the Nuttby Formation suggests that major strike slip motion took place along the Cobequid Shear Zone after the end of voluminous volcanism in the mid Tournaisian (ca. 354 Ma) (cf. Hibbard and Waldron, 2009).

North-vergent thrusting (the New Annan Thrust) of late Tournaisian age (ca. 350 Ma) is recognised in the northeastern Cobequid Highlands (Fig. 2-2) (Piper et al. 1996b; Piper and Pe-Piper 2001), postdating the thrusting associated with pluton emplacement (Fig. 2-6). Possibly synchronous thrusting occurred north of Cape Chignecto (Waldron et al. 1989), where late dykes occupy thrust fault planes (Piper et al. 1993). Mylonite and secondary biotite from the Kirkhill fault trend in the Cape Chignecto pluton date from the early Viséan (336-339 Ma; Nearing et al. 1996) (Fig. 2-6). Mid-Namurian deformation is indicated by the unconformity of the Parrsboro Formation over the West Bay Formation south of the Cobequid Fault (Donohoe and Wallace 1985).

At the southern margin of the Cobequid Highlands, widespread brittle strike-slip movement took place on E-W trending faults in the late Carboniferous (Murphy et al. 1995, 1999), continuous with strike-slip in southern New Brunswick and the Stellarton Basin.

ROAD LOG DAY 2

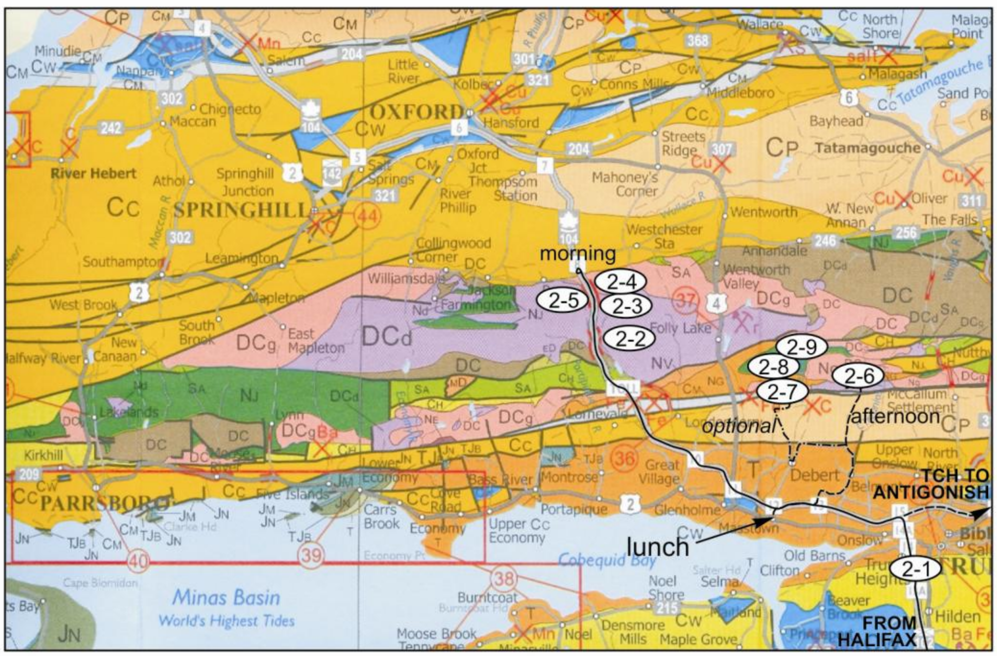

Figure 2-7. Day 2 stops in the Cobequid Highlands. For more detail, see Fig. 2-5.

Depart from Lord Nelson hotel and drive northwards on Highway 102 towards Truro. The first part of the drive takes us across Meguma Supergroup metasediments, but north of Exit 9 at Milford the route passes across the Carboniferous lowlands of central Nova Scotia. The Shubenacadie River, which we cross just north of Exit 10, is tidal across the entire width of the Carboniferous lowlands.

STOP 2-1. Exit right at Exit 13 to Truro Heights and park in the area where the exit ramp forks. Observe view of the Cobequid Highlands (Avalon Terrane) in the distance, bounded by the Cobequid-Chedabucto fault. In the middle ground are the early Carboniferous Kennetcook–St Mary's basins and the feather edge of the Triassic Minas basin, obscuring the tectonic contact of the Avalon and Meguma terranes.

Immediately rejoin Highway 102 northbound on the on-ramp across Truro Heights Road. Drive northwards over the Salmon River. Take the left lane to join the Trans-Canada Highway towards Amherst.

Drive west and north on the Trans-Canada Highway over Cobequid Pass. Note the deformed Lower Carboniferous rocks as the road ascends across the Cobequid Fault. At the toll station, pay toll and reset odometer.
0.0 km Toll booth
2.4–3.5 km. The outcrops on the right are in Fountain Lake Group basalt and rhyolite that are the extrusive equivalent of the Wentworth Pluton.

Figure 2-8. Field photographs of STOPs 2-2 and 2-4 showing relationships between granite and gabbro.

STOP 2-2. Odometer 4.5 km. (45.533767°N 63.710233°W) Old quarry in the Wentworth Pluton on gentle right bend before bridge over highway. Pull off onto the very wide shoulder just beyond the 68 km sign. ***Take care when pulling onto the highway again. Do not attempt to cross the highway. Hard hats should be worn if examining the rock face.*** This outcrop provides a general overview of the western part of the Wentworth

Pluton. The principal rock is the Folly Lake gabbro that has been intruded by subvertical bodies of granite. The whole outcrop shows evidence of both early ductile and later brittle deformation, probably of early and late Carboniferous age respectively. This stop, well clear of the busy highway, provides an opportunity to present an overview of the regional geology: STOPs 2-3 to 2-5 show detailed features more clearly.
The question at this stop is what is the large scale relationship between gabbro and granite? The next three stops will provide better opportunities to examine small-scale structures in granites and gabbros.

STOP 2-3. Odometer 7.4 km. (45.560550°N 63.712233°W) Road cut in Wentworth Pluton. Pull well off onto the narrow shoulder just before the 65 km sign. ***Take care when pulling onto the highway again. Do not attempt to cross the highway. Hard hats should be worn if examining the rock face.***
Questions at this outcrop are: (1) how many different types of gabbro and granite are present; and (2) what rheological properties of these difference types can be inferred from the outcrop?

STOP 2-4. Odometer 8.6 km. (45.570117°N 63.711067°W) Road cut in Wentworth Pluton at hillcrest just past the 64 km sign. Pull well off onto the narrow shoulder in the middle of the outcrop. ***Take care when pulling onto the highway again. Do not attempt to cross the highway. Hard hats should be worn if examining the rock face.***
At this outcrop, there is evidence for at least two styles of emplacement of granite and two types of gabbro distinguished by grain size and by rheological relationships with granite sheets. Is there a pattern of earlier rocks showing emplacement in vertical sheets and later rocks showing emplacement in horizontal sheets? Do the subhorizontal granite sheets mimic the emplacement of larger-scale sills in subvolcanic complexes?

Drive north on the highway to the Collingwood–Wentworth road overpass
Odometer 13.2 km. At Exit 8, take exit ramp to right to the Collingwood–Wentworth road. Turn left onto bridge over highway and re-enter highway to drive south. Drive to just before hillcrest.

STOP 2-5. Odometer 19.0 km. (45.571767°N 63.712450°W) Road cuts in Wentworth Pluton on grade just before hillcrest. Pull off onto the wide shoulder towards the south end of the outcrop, after the 63 km signpost. ***Take care when pulling onto the highway again. Do not attempt to cross the highway.***
At the southern part of the outcrop, examine the gabbro sheets and interpret their relationship to granite. At the northern part of the outcrop, examine the varied relationships between granite sheets and gabbro, including the distribution of gabbro enclaves in granite. Can we make a synthesis of the intrusion sequence and conditions seen at STOPs 2-2 to 2-5?
Continue driving south on Trans-Canada Highway. Pay toll again. Drive Exit 12 and take the ramp to Masstown. Drive south and cross old Highway 4 and enter the parking lot for the Masstown Market for lunch. After lunch, return to the Trans-Canada Highway eastbound and drive a few km east to Exit 13 to Debert. Take ramp and turn left on McElmon Road towards Debert. Pass Tim Hortons. At the Y junction near the water tower, reset odometer.
Note that immediately west of here is the the Debert Paleo-Indian site, dating from before the late glacial Younger Dryas time about 12 000 years BP. Thousands of stone tools and rare living floors have been excavated. The Paleo-Indians occupied the site at a time when the Cobequid Highlands were still glaciated.

0.0 km. Y junction. Take road right (eastwards) towards Belmont
2.3 km. Turn left to Belmont. Drive north through Belmont.
6.9 km. Follow paved road to left at the junction with Graham Road.
7.8 km. Paved road turns sharp left. Yield to oncoming traffic and continue straight up the unpaved Staples Brook Road.
14.1 km. Turn left at Y junction
14.9 km. Turn into the main Quarry at Frog Lake, STOP 2-6.

STOP 2-6. Odometer 14.9 km. The Frog Lake Quarry has been unexpectedly re-activated in the summer of 2010. ***Follow the instructions of the field trip leaders and do not enter the active area of the quarry floor where trucks are driving around. Beware of any potentially unstable freshly excavated surfaces.***

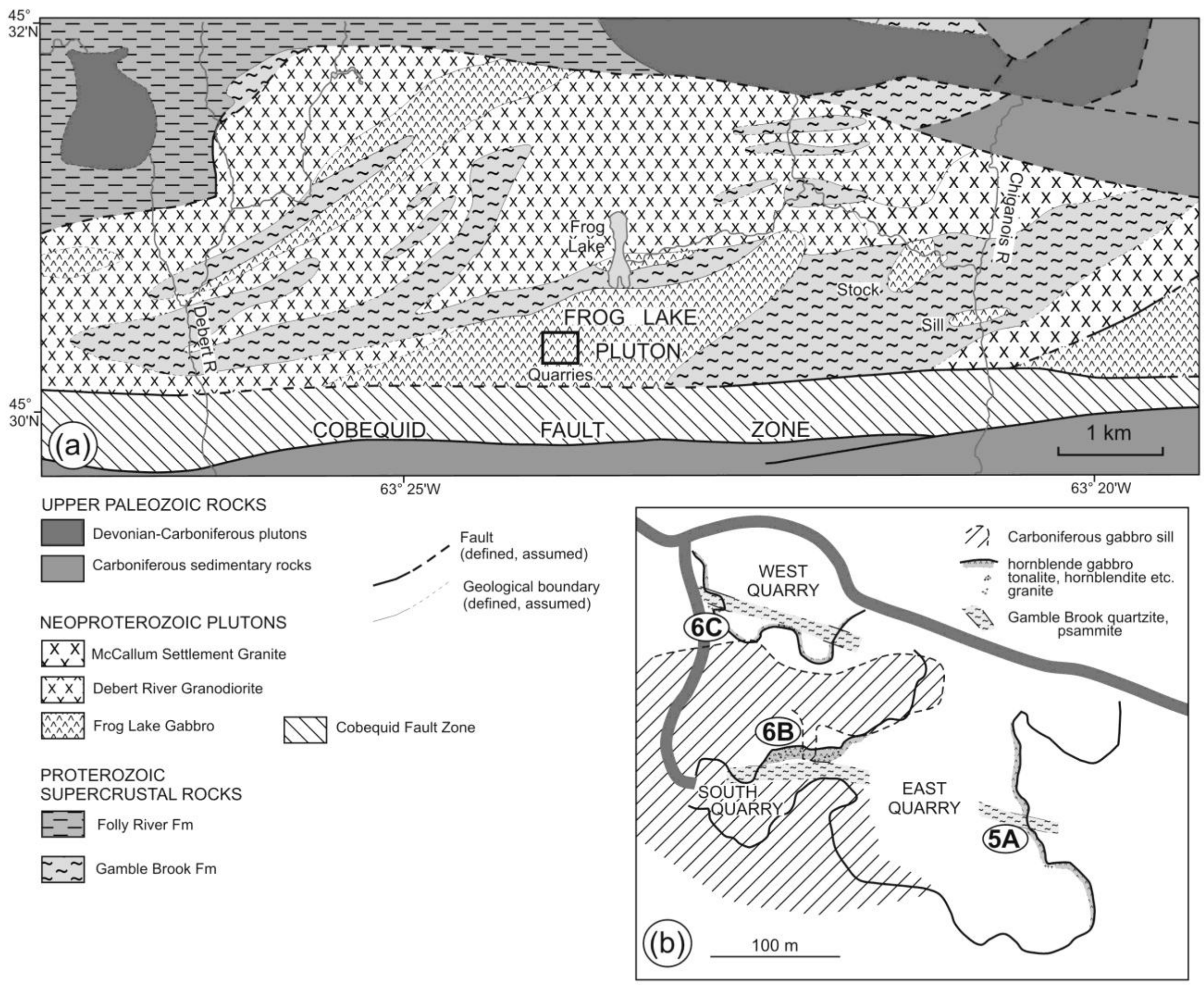

Figure 2-9. Geological maps of (a) the environs of the Frog Lake Quarry and (b) the quarries in 2009. (Modified from Pe-Piper et al. 2010).

The late Neoproterozoic Frog Lake pluton consists predominantly of hornblende gabbro. It shows petrographic similarities to water-rich mafic intrusions known as appinites. The Frog Lake pluton has been dated at 622 ± 3 Ma by $^{39}Ar/^{40}Ar$ on hornblende (Keppie et al. 1990). The country rock of the early Neoproterozoic Gamble Brook Formation experienced late Neoproterozoic deformation and upper greenschist-grade metamorphism and includes biotite-garnet quartzite and biotite-quartz-chlorite-garnet schists (Cullen 1984; Murphy et al. 2001). The main hornblende gabbro was intruded between screens of this metasedimentary country rock that is strongly hornfelsed. Elsewhere in the Bass River block, plutonic rocks can be shown to have been emplaced under conditions of regional dextral shear. The contacts appear to have been pathways for magma of gabbroic, tonalite-granodioritic and granitic composition that carried enclaves of gabbroic lithologies. Some of these magmas had a high volatile content, resulting in abundance of hydrous mineral phases, pegmatites, and diffuse felsic segregations. These varied rocks in the contact zones experienced progressive shear resulting in syn-magmatic deformation. The Frog Lake pluton appears to be at a deeper structural level than the Greendale appinite complex of the northern Antigonish Highlands that will be seen on Day 3. The Greendale appinitic hornblende gabbro intruded greenschist to sub-greenschist grade Neoproterozoic Georgeville Group rocks under conditions of dextral shear and show widespread evidence for syn-magmatic deformation and dyking.

Figure 2-10. Panorama of STOP 2-6A in the Frog Lake quarries, showing alternating hornblende gabbro and sheets of Gamble Brook Formation country rock.

STOP 2-6A. East side of Northeast Quarry. (45.504433°N 63.395367°W) This continuous outcrop shows the Frog Lake hornblende gabbro intruding highly deformed country rock of the Gamble Brook Formation. Thin gabbro sheets are generally highly deformed, larger gabbro bodies are generally less deformed. Both appear compositionally inhomogenous. A muscovite granite body is present near the south end of the main wall. Is there evidence in these outcrops for a sequence of intrusion? What was the relationship of the granitic to the gabbroic magmas?

STOP 2-6B. This is located in the northeast part of the South Quarry. It may be destroyed by renewed quarry operations by the time we visit. It is a zone rich in mafic enclaves and felsic sheets, many of which are deformed, at the southern margin of a gabbro body against country rock.

STOP 2-6C. West side of Northwest Quarry. This is an alternate to STOP 2-6B. It is another contact of gabbro with country rock, showing a particular abundance of mafic enclaves.

Then drive back down the unpaved Staples Brook road. If STOPs 2-7 to 2-9 are to be skipped, then retrace the route through Belmont and past the Tim Hortons back to the TransCanada Highway.

If STOPs 2-7 to 2-9 are to visited, then on reaching the paved road, turn right and drive west and then follow the road to the south to Debert. In Debert, turn right at the T junction and cross the Debert River bridge. Immediately west of the bridge, turn right onto the Upper Debert River Road. Reset odometer.

0.0 km. South end of Upper Debert River Road.
5.4 km. Turn right at T-junction.
6.1 km. Turn left onto woods road.
7.8 km. Bear right at Y junction.

STOP 2-7. Odometer 9.9 km. (45.508283°N 63.460233°W) Debert River granodiorite (Neoproterozoic) and gabbro stock (Carboniferous). This is the first of three stops that illustrate the principal lithologies in the Bass River Block that were not seen at STOP 2-6.

12.7 km. Crossing woods road.

STOP 2-8 Odometer 13.1 km, just before top of hill. (45.532267°N 63.452750°W) Folly River Formation (Neoproterozoic) metabasalt and metagabbro. Is there any field evidence for pillow basalt?

STOP 2-9. Odometer 13.7 km. (45.536800°N 63.455833°W) McCallum Settlement granite (Neoproterozoic) and gabbro dykes (Carboniferous) at Y junction on woods road.

Turn around and retrace route through Debert and past Tim Hortons to the Trans-Canada Highway (35 minutes).

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CHAPTER 3: THE ANTIGONISH HIGHLANDS

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GENERAL OVERVIEW

Like the Cobequid Highlands, the Antigonish Highlands is part of the Avalon terrane. The Cobequid and Antigonish highlands were probably contiguous prior to the Late Carboniferous opening of the Stellarton Basin, which lies between them. We will be visiting the coastal block where three formations of the Georgeville Group occur as well as two spectacular plutonic bodies, known as the Greendale Complex and the Georgeville Granite (**STOPs 3-1 and 3-2**). Time permitting, we will also examine the coastal sections that preserve an excellent record of the Ordovician volcanism that occurred after Avalonia separated from northern Gondwana, when it was probably a New Zealand-style microcontinent.

NEOPROTEROZOIC ROCKS

The Antigonish Highlands is divided into several fault blocks, each block predominantly underlain by Late Neoproterozoic rocks of the ***Georgeville Group***, which record the progressive development of a sedimentary basin within an arc regime (Fig. 3-1; Murphy *et al*. 1990). Geochronological data indicate that the ***Georgeville Group*** was deposited between ca. 620 and 608 Ma. Its lowest stratigraphic units are dominated by calc-alkalic basalt and basaltic andesite, dacite, and rhyolite (dated at 618 Ma $\pm$ 2 Ma; Murphy *et al.*, 1997), all with arc geochemical characteristics. Some interbedded basalts are geochemically distinct, having continental tholeiitic affinities, and are interpreted to reflect rifting and basin formation within the arc. These rocks are locally interbedded with tuff, mudstone, and limestone. In the area that we will visit, these rocks are represented by the Chisholm Brook Formation. The upper units are overwhelming dominated by volcanogenic turbidite deposits, which include matrix-supported conglomerates, greywackes, and mudstones, as well as subordinate tholeiitic mafic volcanic rocks.

The youngest detrital zircon from the turbidites yields 613 ± 5 Ma (single zircons, TIMS, Keppie *et al*. 1998) and provides a maximum depositional age for the formation. In general, the dominant detrital zircon population ranges from 0.61-0.62 Ga, and other populations include ca. 1.1-1.2 Ga, 1.5 Ga, 1.8-2.0 Ga and 2.6 Ga, which may reflect derivation from the Amazonian craton of Gondwana (Keppie et al. 1998).

The Georgeville Group was deformed by F_1 tight to isoclinal recumbent folds with a weak axial planar cleavage developed under lower greenschist facies conditions associated with dextral motion along NE-trending faults followed by two phases of upright folding (Murphy et al., 1991). Intrusion of the "appinitic" mafic to intermediate ***Greendale Complex*** occurred during the later stages of deformation (Murphy and Hynes 1990; Murphy *et al*. 2001).

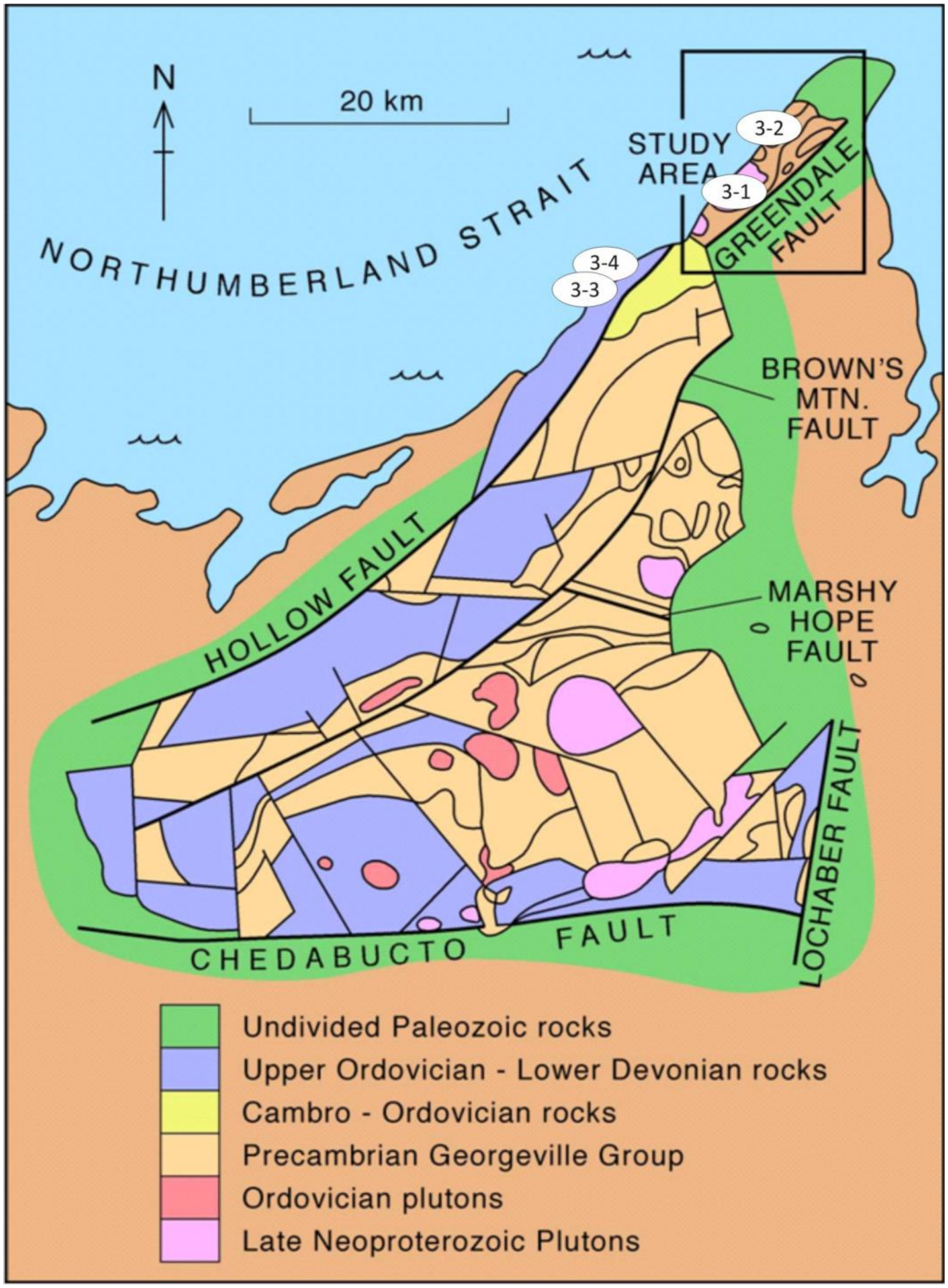


Fig 3-1. Summary geology map (modified from Murphy et al., 1991) of the Antigonish Highlands showing the locations of STOPs 3-1 to 3-4. The inset marked "Study Area" shows the location of Fig. 3-2 (STOPs 3-1 and 3-2). Parts of the southern highlands are after Escarraga (2010) who determined that several plutons, previously thought to be Devono-Carboniferous in age (in Murphy et al., 1991) are either Neoproterozoic or Ordovician in age. Current mapping by Nova Scotia Dept. of Natural Resources in the southern highlands will result in further modifications of map boundaries.

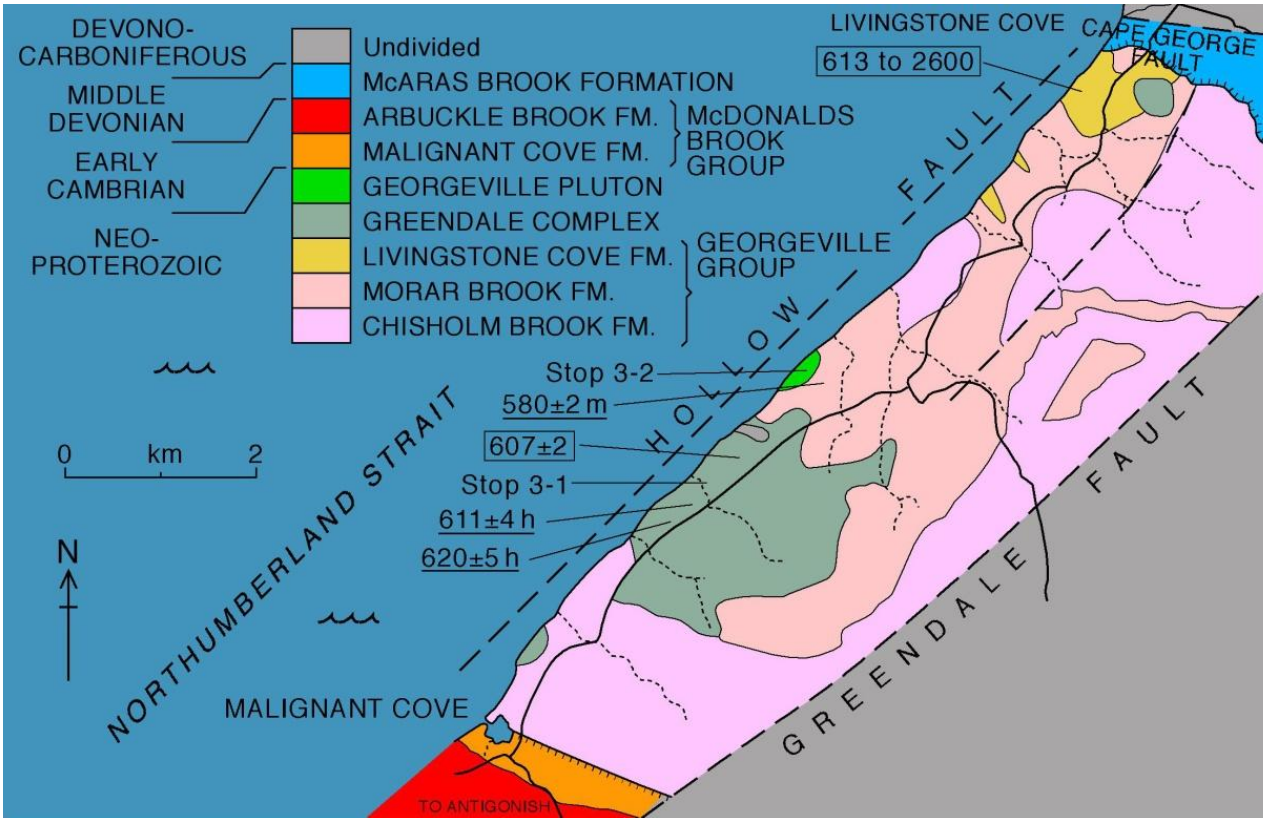


Fig. 3-2. Geological map of the northern Antigonish Highlands showing the locations of the Greendale Complex and Georgeville Pluton. m = ^{40}Ar/^{39}Ar muscovite; h = ^{40}Ar/^{39}Ar hornblende.

The Greendale Complex is remarkably heterogeneous on all scales, and ranges in composition from ultramafic to felsic rocks. Lithologically, it is dominated by hornblende gabbro, which is locally pegmatitic. The felsic rocks intruded late in the evolution of the complex. There are excellent field examples of mingling and mixing on all scales. The Greendale Complex (Fig. 3-2) has yielded ages of 607 $\pm$ 2 Ma (U-Pb, titanite, Murphy *et al.* 1997) and 611 $\pm$ 4 Ma (^{40}Ar/^{39}Ar, hornblende, Keppie *et al*. 1990). These age data suggest a limited time gap between deposition, deformation, and intrusion of the Georgeville Group, consistent with interpretations of synorogenic emplacement in a strike-slip basin within an ensialic volcanic arc regime.

The Georgeville Group is post-tectonically intruded by the ***Georgeville Granite***, which is a small (ca. 1.5 km across) epizonal body that is excellently exposed along the shoreline of the Northumberland Strait (Figs. 3-2, 3-3, 3-4) to the northwest of the Greendale Complex. The pluton consists of a central stock predominantly consisting of leucocratic, medium- to coarse-grained alkali feldspar granite and pegmatite intruded by steep to moderately dipping aplite and pegmatite dykes. Intrusive contacts with the Georgeville Group host rocks are sharply defined and the host rocks show development of spotted hornfels which overprints regional tectonic fabrics. Many, but not all, geochemical and mineralogical features resemble A-type, within plate granites. ^{40}Ar/^{39}Ar (muscovite) data yielded a plateau age of 579.8 $\pm$ 2.2 Ma, interpreted by Murphy *et al*. (1998) interpreted as the age of intrusion.

Fig. 3-3. Left: Field photos of Georgeville Granite (light) and mudstone host rock (dark).Right: an example of a pegmatitic phase.

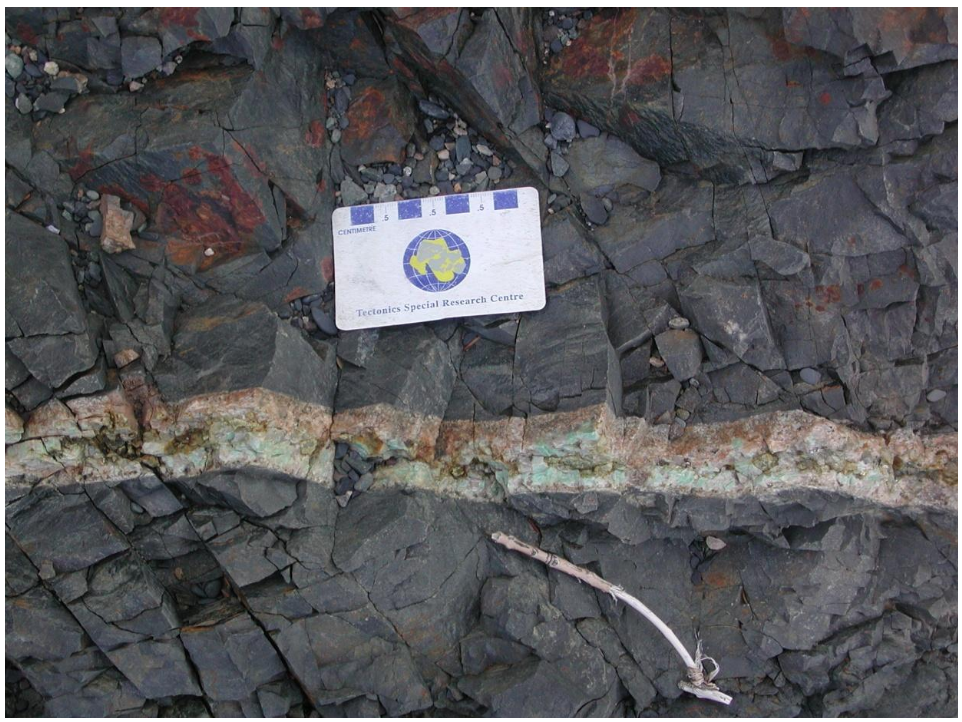


Fig. 3-4. Amazonite-bearing pegmatite dated by $^{40}Ar/^{39}Ar$ on muscovite at 579.8 $\pm$ 2.2 Ma.

CAMBRIAN-EARLY ORDOVICIAN ROCKS

Neoproterozoic rocks of the Georgeville Group are unconformably overlain by Cambrian-Early Ordovician rocks (Fig. 3-1). Unfortunately, we will not have time to see these on the trip. Cambrian-Early Ordovician rocks are divided into two groups that are related by lateral facies variations: the predominantly sedimentary (continental-shallow marine) Iron Brook Group and the predominantly volcanic (bimodal, rift-related) McDonalds Brook Group (Keppie and Murphy 1988). The Iron Brook Group contains limestones that have yielded a late Early Cambrian typically Avalonian fauna of trilobites and paraconodonts (Landing and Murphy 1991).

These rocks were affected by greenschist-facies metamorphism, and were also deformed by recumbent F1 folds with a penetrative axial plane cleavage defined by chlorite, muscovite and biotite, and accompanying thrusts, and then refolded by F2 folds. The deformation is early to mid Ordovician in age, and predates deposition of the Dunn Point Formation and Arisaig Group. The underlying Georgeville Group rocks are virtually unaffected by this deformation, indicating that Cambrian-Early Ordovician sequences are separated from Neoproterozoic sequences by a décollement surface that lies close to, or at, the contact between them.

MID ORDOVICIAN-EARLY DEVONIAN ROCKS

Time permitting, we will examine the Ordovician (ca. 460 Ma) bimodal volcanic sequence of the ***Dunn Point Formation,*** which lies at the base of a Mid Ordovician-Early Devonian succession (Fig. 3-1) and the conformably overlying (ca. 454 Ma) McGillivray Brook Formation, which is dominated by flow-banded felsic volcanic rocks and lahars. These rocks unconformably overlie Cambrian to Early Ordovician and Neoproterozoic sequences and are overlain by a 1900 m thick conformable sequence of fossiliferous earliest Silurian-early Devonian shallow marine to continental siliciclastic rocks of the Arisaig Group.

Recent U-Pb zircon (TIMS) data from a rhyolite yielded a concordant age of 460.0 ± 3.4 Ma for the Dunn Point Formation (Hamilton and Murphy, 2004), indicating a 15-20 million year gap between the Dunn Point volcanism and deposition of the basal formation of the Arisaig Group. Paleomagnetic data indicate that the Dunn Point volcanic rocks lay at a paleolatitude of 41°S ± 8° (Van der Voo and Johnson, 1985), i.e. 2000 km north of Gondwana, and 1700-2000 km south of the Laurentian margin at that time, with a latitudinal component of the convergence rate between Avalonia and Laurentia from 460 to 440 Ma of about 5.5 cm/yr (see also Johnson and Van der Voo 1990). Lying this distance off the Laurentian margin at 460 Ma, Avalonia was the key tectonic element defining the southern margin of the Iapetus Ocean and the northern margin of the Rheic Ocean (e.g. Torsvik *et al*. 1996). The Arenig-Llandeilo northward drift of Avalonia relative to Gondwana is consistent with the presence of an ocean ridge system between Avalonia and Gondwana, implying that the trailing (southern) edge of Avalonia was a passive margin. Thus, the tectonic environment of the Dunn Point Formation is interpreted in terms of an ensialic microplate floored by Avalonian crust, bordered to the north by the Iapetus Ocean and to the south by the Rheic Ocean.

The Dunn Point and McGillivray Brook Formations (DPF and MBF) both resemble A-type silica-rich magmas, and the MBF in particular has high concentrations of Zr (745-1965 ppm), Y (65-213 ppm), Nb (57 to 185 ppm), and high Ga/Al (Murphy and Hamilton, submitted). Several MBF samples exhibit strong LREE depletion, consistent with fractionation of LREE-bearing accessory phases. The εNd_t values for MBF (t = 455 Ma) range from + 1.5 to + 3.9, and overlap with DPF rhyolites, which range from +2.9 to +3.7. Depleted mantle model ages for MBF and DPF samples unaffected by accessory phase fractionation are between 0.9 and 1.2 Ga and are similar to Neoproterozoic and Cambrian crustally-derived felsic rocks. Taken together the data are consistent with derivation from ca. 0.9 to 1.2 Ga lower crust. The Dunn Point and McGillivray Brook felsic rocks were probably erupted in a local extensional environment. Regional considerations, however, suggest that this extensional environment occurred within an ensialic arc, analogous to the modern Taupo Volcanic Zone in northern New Zealand.

In contrast to the Dunn Point and McGillivray Brook Formations, geochemical and isotopic studies of the Arisaig Group sedimentary rocks indicate they are not derived from the underlying, juvenile Avalonian basement (Murphy *et al.* 1996a, 2004), but these data are consistent with derivation from ancient continental crust. Reconstructions show that Avalonia was adjacent to Baltica by 440 Ma and Laurentia by 415 Ma.

ROAD LOG

The outcrops we visit on Day 3 are located along the shore of the Northumberland Strait, north of Antigonish (Fig. 3-1). They may be reached by driving north along Nova Scotia highway 245 from Antigonish to Malignant Cove, which is located at the junction with Nova Scotia Highway 337. To reach STOP 3-1, proceed 5 km east on Highway 337 to a road on the north side of the road leading to the Georgeville Quarry (Fig. 3-2). This road has a metal bar-type gate at its entrance. Park by the gate and walk down the road to the shore. STOP 3-1 is located ca. 700 m southwest of the entrance to the shore.

To reach STOP 3-2, resume driving northeast along Nova Scotia Highway 337 about 3 km and turn left on an unpaved road that goes straight towards the shoreline.

To reach STOP 3-3, return to Malignant Cove along Nova Scotia Highway 337 (Fig. 3-1-2). From here, drive southeast along Highway 245 to Arisaig (Fig. 3-1), where you take the turn down to Arisaig Pier for STOP 3-3 (Fig. 3-1). STOP 3-4 located along the coastline ca. 500 m northeast of STOP 3-3.

STOP DESCRIPTIONS

STOP 3-1 (45.816181°N 62.050308°W to 45.821181°N 62.042519°W)

Neoproterozoic Greendale Complex and on the shore north from the Georgeville Quarry, Northumberland Strait (Fig. 3-2).

This stop demonstrates the remarkable heterogeneity of Greendale Complex on all scales. The complex ranges in composition from ultramafic to felsic rocks (Fig. 3-5). There are more variations in textures than can possibly be described briefly in print, and there will be a contest (prize of a beer) to anyone who can put her/his two feet on the same rock at the same time! Field trip leaders are ineligible to receive this award. Although semi-circular in map pattern, the Complex essentially consists of many dykes of variable composition and orientation, giving the appearance of vertical layering. According to Murphy and Hynes (1990), this layering was developed during coeval strike-slip movement on the adjacent Hollow and Greendale faults. There are excellent examples of mingling and mixing on all scales. Lithologically, it is dominated by hornblende gabbro, which is locally pegmatitic. The felsic rocks were intruded late in the evolution of the complex. Locally, they bear an intriguing resemblance to the Georgeville Granite (STOP 3-2), and we will discuss the possibility of a genetic relationship.

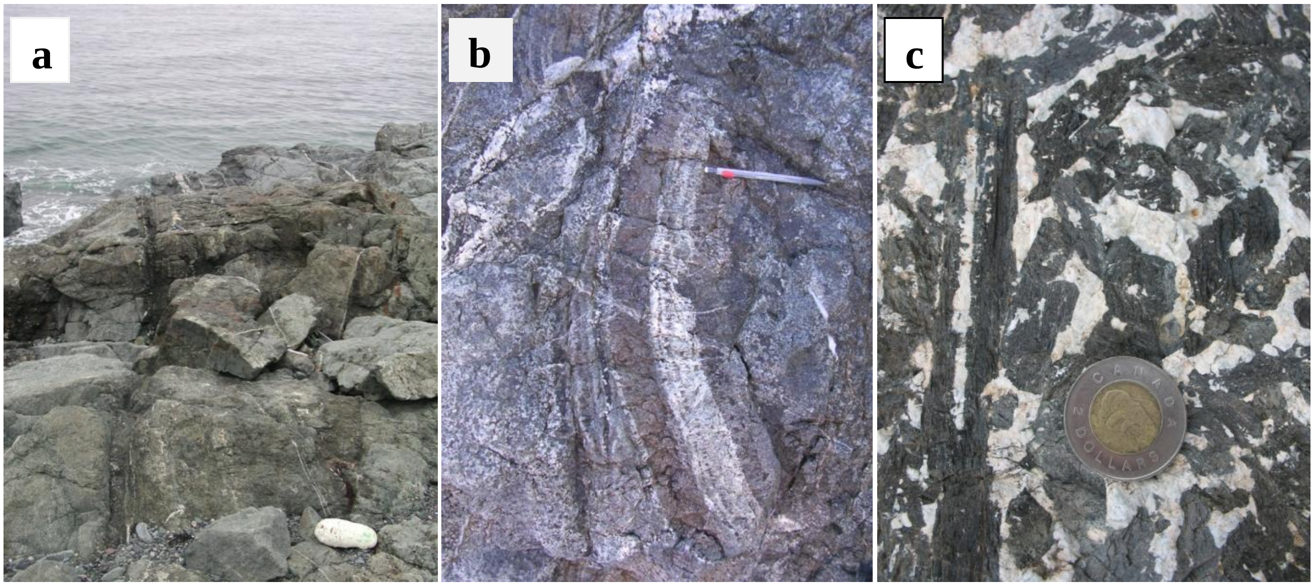


Figure 3-5. Features of the Greendale Complex. a,b Vertical layering, suggesting dilation. c : Large hollow hornblende crystals in a pegmatitic facies of the complex.

We do not understand everything we see, and there is ample scope for many discussions. Depending on time and tide, we also intend to see examples of the following features:

A: Sharp contacts between two dominant phases of the syn- to post- tectonic Greendale Complex; i.e. plagioclase-rich and hornblende-rich gabbro. The plagioclase crystals typically enclose enclaves of matrix and so is a late crystallizing mineral, probably of metasomatic origin. Igneous hornblendes from the Greendale Complex have been dated at 611 ± 4 Ma and 620 ± 5 Ma ($^{40}Ar/^{39}Ar$, hornblende, Keppie *et al.* 1990). Our preferred age for the intrusion is 607 ± 2 Ma (U-Pb, titanite, Murphy *et al*. 1997).
B Basalt xenoliths in the gabbro, with evidence of increasing contamination towards the contact.
The basalt consists of albite, epidote, prehnite and actinolite. The amphibole increases in size towards the contact with the gabbro. Coarse pegmatites with sinuous contacts and a microgranite dyke represent a late felsic stage.
C A fault slice containing metasomatized marble and interbedded impure serpentinized marble and basalt. These rocks probably comprise part of the wall-rock (Chisholm Brook Formation).
D Intrusive contact between late pegmatitic felsic phase of the intrusive body and marble and tuff of the country rocks. The pegmatite contains large crystals of amphibole (up to 8 cm), in places enclosing plagioclase cores, in a matrix of orthoclase, quartz and plagioclase. An aplitic dyke cuts the pegmatite. Host rocks show evidence of contact metamorphism.
E Contact between two formations of Georgeville Group host rock: the Chisholm Brook and Morar Brook Formations. Interbedded tuff and agglomerate of Chisholm Brook Formation grade conformably up into black, thinly laminated mudstone of the Morar Brook Formation. The bedding is overturned.
F Morar Brook Formation: slump folds. Although this fold is probably related to deformation, large-scale isoclinal folds affect the volcanic rocks and, therefore, have a hard-rock component.

G Leucocratic diorite intrusive sheet containing xenoliths of country rock. Both contain large stage metasomatic plagioclase porphyroblasts. The eastern contact of the sheet is broadly concordant. Load casts in the mudstone/siltstone host rock indicate overturned bedding.
H Morar Brook Formation: interbedded mudstone, siltstone, marble and chert. Note the minor intrusive, which alternates from a sill to a dyke.
J Morar Brook Formation: An example of a northeast-southwest, upright, F2 fold.

STOP 3-2 (45.827644°N, 62.033839°W)
The Georgeville Granite, pegmatites and metasomatized aureole.
The Georgeville Granite is dominated by leucocratic, medium- to coarse-grained alkali feldspar granite and pegmatite intruded by steep to moderately dipping aplite and pegmatite dykes. Intrusive contacts with the Georgeville Group host rocks are sharply defined and the host rocks show development of spotted hornfels which overprints regional tectonic fabrics.
The granite consists of medium-grained quartz, microcline perthite and albite (An_{0-1}) with subordinate mafic phyllosilicate minerals and a wide range of accessory minerals including zircon, titanite, rutile, euxenite, cassiterite and pyrite (Murphy *et al*. 1998). Small pegmatitic pods (generally ca. 0.4 m in diameter) occur throughout the exposed granite, and consist mainly of quartz and microcline ($\pm$ amazonite). The granite is characterized by high SiO_2 (between 71-80%), Th, Nb, Y and Zr, low CaO, TiO_2, MgO, FeO, MnO and most notably, extreme LREE depletion. Many, but not all, geochemical and mineralogical features resemble A-type, within-plate granites. However, textural and mineral chemical evidence suggests that the whole-rock trace-element and isotopic systematics were influenced by subsolidus aqueous fluids (Anderson et al., 2008; Dalby *et al*., 2010; Fig. 3-6 to Fig. 3-9; Table 3-1). $^{40}Ar/^{39}Ar$ (muscovite) data yielded a plateau age of 579.8 + 2.2 Ma. Murphy *et al*. (1998) interpreted this as the age of intrusion, suggesting intrusion about 25-30 million years after the Greendale Complex and after arc-related magmatism had ceased in the Antigonish Highlands. We were never particularly happy with this conclusion, and we are keeping an open mind as to whether the Georgeville Granite may be related to the Greendale Complex. We will discuss the unique challenges of dating of the Georgeville Granite. All suggestions are welcome!

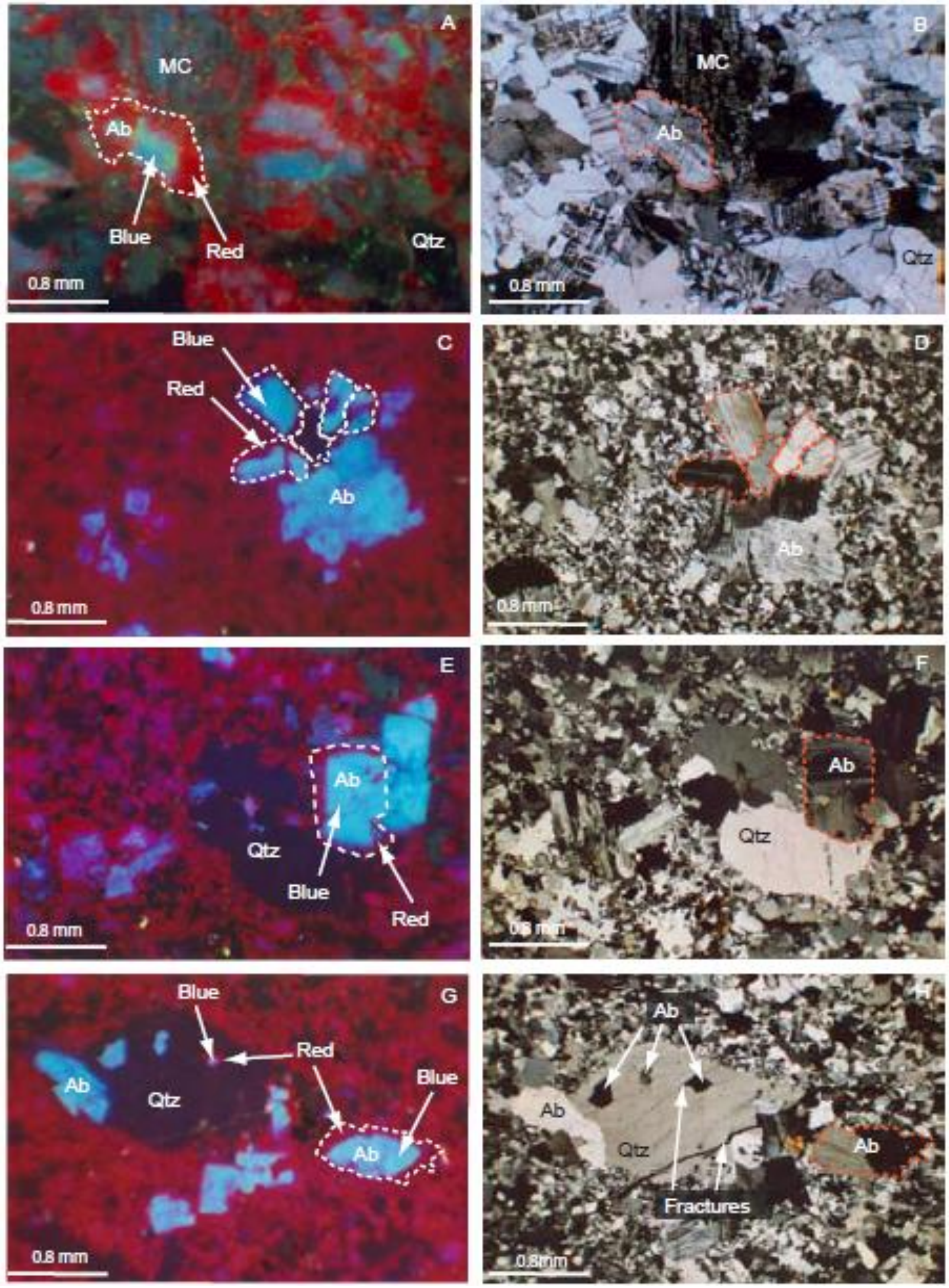
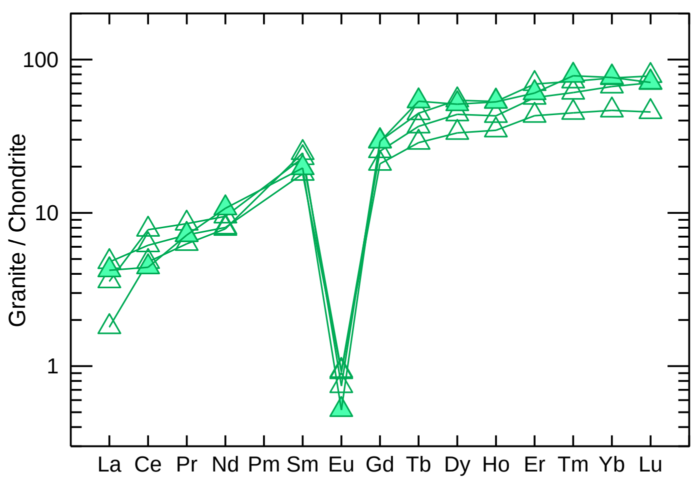


Fig. 3-6 Examples of individual albite grains in the Georgeville granite (some outlined in dashed lines for clarity). On the left are CL images, and on the right are XP (crossed polar) images of the same areas. The CL textures do not correspond to any intragrain features observed with the petrographic microscope.

Fig. 3-7 Veins of aplite cross cutting ultra-coarse quartz + microcline pegmatite near the contact of the Georgeville granite.



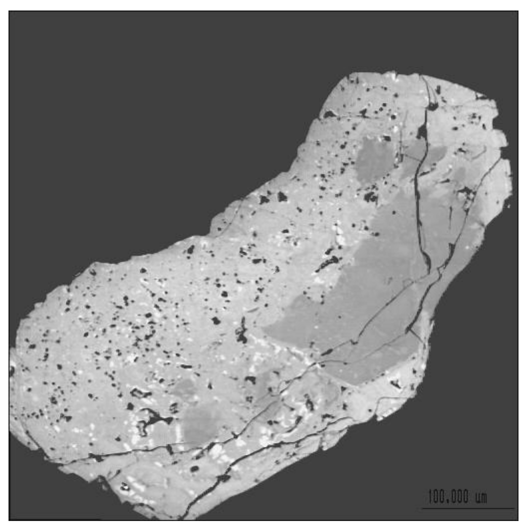


Fig. 3-8 (Left) Chondrite-normalized REE patterns for the chloritized (shaded triangles) and nonchloritized granite (open triangles) from Georgeville. Normalization values are from Sun & McDonough (1989).
Figure 3-9. (Right) BSE image of a zircon grain in the Georgeville granite. Note the micropores and bright inclusions of thorite indicating dissolution and reprecipitation of zircon by late fluids.

Table 3-1. Chemical composition of the Georgeville granite.

Sample	Al – 4[a]	Al-7[a]	Al-10[a]	Chloritized[b] Granite
SiO_2 wt.%	78.07	76.94	77.18	72.12
TiO_2	0.02	0.02	0.05	0.04
Al_2O_3	12.27	12.45	12.51	11.90
Fe_2O_3	0.77	0.68	1.13	5.64
MnO	0.02	0.01	0.02	0.09
MgO	nd	nd	0.13	3.94
CaO	nd	nd	nd	0.12
Na_2O	4.55	4.35	4.27	0.01
K_2O	3.87	4.35	3.98	2.20
P_2O_5	0.01	0.01	0.02	0.02
H_2O	nd	nd	nd	4.08
CO_2	nd	nd	nd	0.03
LOI	0.1	0.1	<0.1	nd
Total	99.68	98.91	99.29	100.20
Ba (ppm)	22	21	36	88
Ga	38	40	37	35
Nb	93	79	67	105
Ni	5	5	5	<10
Rb	326	402	324	130
Sc	0.95	78	1.33	<10
Sr	5	5	16	<10
V	5	5	9	<10
Y	105	99	79	118
Zn	12	144	17	288
Zr	227	187	104	229
La	0.43	0.85	1.13	1.0
Ce	2.93	4.75	3.76	2.7
Pr	0.6	0.81	0.68	0.68
Nd	3.67	4.43	3.75	5.4
Sm	3.76	3.44	2.75	3.4
Eu	0.04	0.05	0.05	0.03
Gd	6.04	5.19	4.3	5.5
Tb	1.67	1.37	1.07	1.5
Dy	13.78	11.16	8.45	13
Ho	3.01	2.43	1.96	2.8
Er	11.47	9.33	7.12	9.7
Tm	1.84	1.55	1.15	1.5
Yb	12.89	11.34	7.92	13
Lu	1.99	1.79	1.16	1.8

[a] trace element analysis by ICP-MS [b] trace element analysis by ICP-OES

STOP 3-2A: Intrusive contact between post-tectonic alkali feldspar granite "alaskite" stock of the Georgeville Granite and mudstones of the Morar Brook Formation showing contact metamorphism. The alaskite consists of quartz, albite and microcline with minor zinnwaldite, chlorite and amazonite. $^{40}Ar/^{39}Ar$ (muscovite) data yields a plateau age of 579.8 + 2.2 Ma (Murphy *et al.* 1998).

STOP 3-2B: Morar Brook Formation: An example of an easterly facing, isoclinal F1 fold with a poorly developed fracture cleavage.

STOP 3-2C: Coarse pegmatite associated with the alaskite.

STOP 3-2D: Thin veins of pegmatitic granite intruding the mudstones are undeformed indicating their post-tectonic age.

STOP 3-2E (Optional): Chisholm Brook Formation: tuff, basalt flows and mudstone intruded by numerous co-magmatic mafic dykes. Typical mineralogy is albite, epidote, actinolite and chlorite. These rocks probably underlie the Morar Brook Formation.

STOP 3-3 (45.762222°N 62.171028°W)

Middle Ordovician Dunn Point Formation and Earliest Silurian Beechill Cove Formation at Arisaig

In the Arisaig area (Fig. 3-10; Fig 3-11), the succession starts with 90-210 m of subaerial, bimodal, (tholeiitic-alkalic), within-plate, rift or anorogenic basalts, rhyolites, and ignimbrites and interbedded laterites of the Dunn Point and McGillivray Brook Formations. These formations are unconformably overlain by the Early Silurian-Early Devonian Arisaig Group, with basal units consisting of shallow marine conglomerate, sandstone and siltstone of the Early Llandovery Beechill Cove Formation that grade upwards into Mid-Late Llandovery black shales, muddy siltstone, tuff and arenaceous limestone of the Ross Brook Formation. The Dunn Point rhyolite has yielded a middle Ordovician age of 460 ± 2 Ma (Hamilton and Murphy, 2004). The depositional age of the Arisaig Group is constrained by abundant fossils (Boucot *et al*., 1974).

STOP 3-3A

Contact between Beechill Cove and Dunn Point Formations as described by Boucot *et al*. (1974). Here, conglomerate, assigned to the base of the Beechill Cove Formation by Boucot *et al*. (1974), lies disconformably upon rhyolite capped by lenses of laterite of the Dunn Point Formation. We have some problems with this relationship and we will discuss them at the outcrop. Most of the pebbles are rhyolite and ignimbrite, presumably derived from the McGillivray Brook Formation. Some jasper pebbles are also present. The rhyolite exhibits well-developed flow banding, and consists of quartz and feldspar phenocrysts set in a devitrified glass matrix, with variable degrees of alteration of albite, sericite and hematite.

STOP 3-3B

Dunn Point Formation: Basalt flows interbedded with laterite horizons and overlain by rhyolite flows. The basalts are generally vesicular, particularly towards their tops, and grade upwards from spheroidally weathered basalt, through a rubbly laterite/basalt mixture into a blocky indurated red laterite, capped by small irregular lenses of bedded laterite. This sequence is interpreted as an undisturbed red lateritic soil profile (Boucot et al., 1974), with the bedded laterite having been deposited by wind. In places, the basalt flows have incorporated pieces of the underlying laterite in their basal portions. The intimate association of vesicles with laterite trains suggests that the laterites were soft and wet at the time of extrusion. The basalts are generally aphyric and consist of plagioclase, epidote, chlorite, calcite, opaques, and rare clinopyroxene relics. Vesicles are filled with quartz, chlorite, and prehnite. The basalts are tholeiitic with some alkaline tendencies and were extruded during an anorogenic or rifting period within a continental plate (Murphy *et al*., 2008). *Caution:* This outcrop contains visible bombed droppings from seagulls; remember, the past is the key to the present!

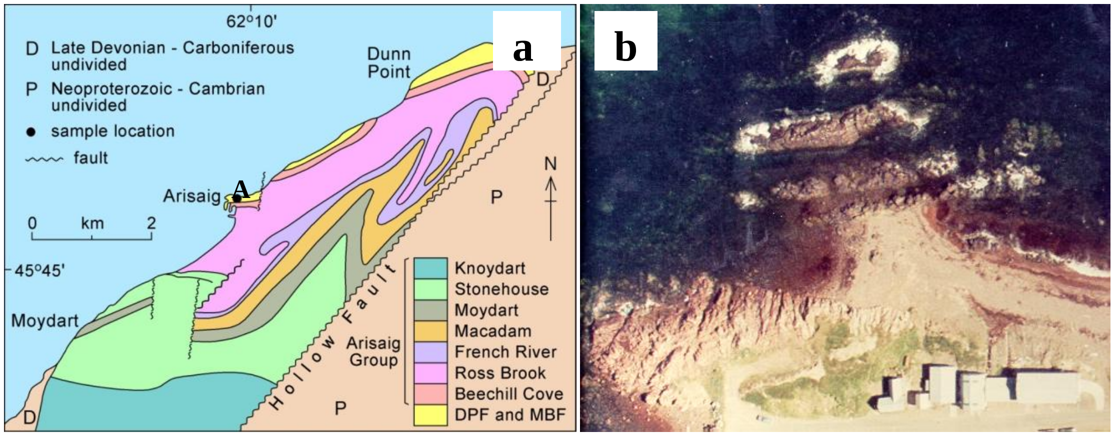


Fig. 3-10. a: Geological map(modified from Boucot et al., 1974) showing the distribution of the ca. 460-454 Ma Dunn Point and McGillivray Book Formations (DPF and MBF) and the overlying Arisaig Group. A and B refer to STOPs 3-3A and 3-3B. b: Air photo of typical Dunn Point Formation rocks showing individual mafic flows (upper) overlain by felsic flows and ignimbrite.

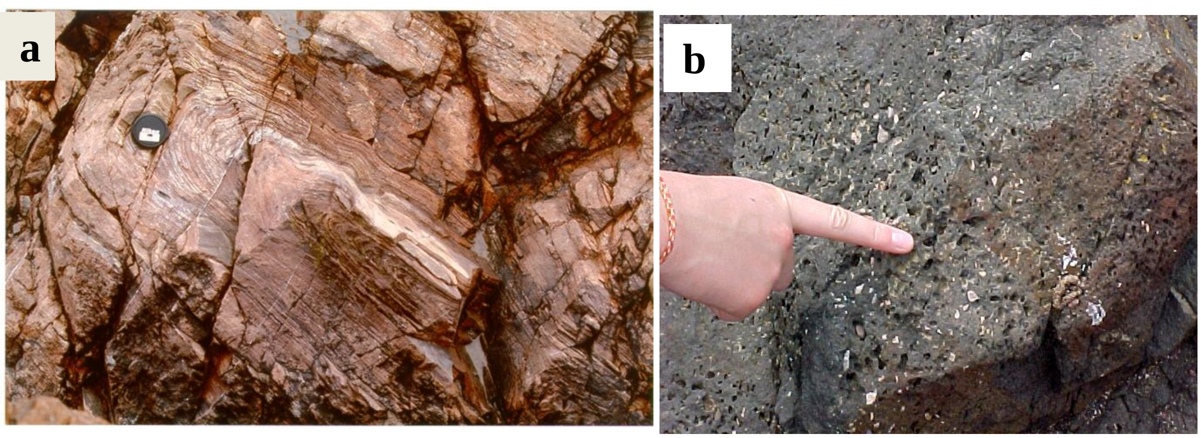


Fig. 3-11. Field photos of (left) flow-banded ignimbrite and (right) amygdaloidal basalt, Dunn Point Formation.

STOP 3-3C

Dunn Point Formation: Basalt and rhyolite flows. The rhyolite flows show a variety of features, such as flow banding and auto- brecciation.

STOP 3-4 (45.765469°N 62.160844°W)

Middle Ordovician McGillivray Brook Formation

The McGillivray Brook Formation (dated at ca. 454+0.7 Ma; Murphy et al., submitted; U-Pb, zircon, TIMS) conformably overlies the Dunn Point Formation and unconformably underlies the Arisaig Group.

In contrast to the bimodal composition of the Dunn Point and Bears Brook Formations, the McGillivray Brook Formation is dominated by felsic rocks, including ignimbrites, as well as subordinate rhyolite flows, lahars and lateritic mudflow deposits that were deposited in a subaerial environment and are described in detail in Boucot *et al*., (1974). The thickness of the McGillivray Brook Formation varies from 60-130 m. The ignimbrites are the dominant lithology and are pale pink to orange in colour and exhibit well preserved flow banding. They consist of compact eutaxitic welded tuffs, and contain euhedral quartz with characteristic “bottle-neck embayments” (Rast, 1963), fragments of devitrified pumice, ash, minor laterite and basalt, in a matrix dominated by devitrified triangular to platy glass shards, granular hematite and clay minerals.

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CHAPTER 4: GRANITES AND TERRANES IN CAPE BRETON ISLAND

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OVERVIEW

Cape Breton Island is a granite-lover's paradise – at least 60 different granitoid plutons, most of them composite, have been named, ranging in age from Mesoproterozoic to Late Devonian. They formed in a variety of tectonic settings, are compositionally varied, and display diversity in isotopic characteristics (e.g., Barr 1990; Barr and Hegner 1992; Ayuso et al. 1996; Potter et al. 2008). This diversity reflects the complex geology and geological evolution of the island. Cape Breton Island is made up of four lithotectonic divisions named (from south to north) the Mira, Bras d'Or, and Aspy terranes and the Blair River inlier (Fig. 4-1). All of these terranes were juxtaposed by the late Devonian because Carboniferous clastic and carbonate rocks correlate across the island (Fig. 4-2), although faulting (including strike-slip faults, thrust faults, and extensional detachment faults) continued during the Late Paleozoic and into the Mesozoic.

On this trip we will be focusing on pre-Carboniferous geology with strong emphasis on contrasts among granitoid rocks in each of the four lithotectonic divisions of the island. Differences in age and petrological character of granitoid rocks played a major role in the recognition of these divisions (e.g., Barr and Raeside 1986, 1989; Raeside and Barr 1992; Barr 1990). In some cases (e.g., Blair River inlier vs. Aspy terrane) the differences are glaring; in other cases (e.g., Bras d'Or and Mira terranes) the contrasts are much more subtle. The schematic stratigraphic columns for each of the lithotectonic divisions in Figure 4-2 should help you to keep track of the differences.

Until the mid-1970s, the granites of Cape Breton Island were assumed to be Devonian, like those in southern Nova Scotia. The pioneering Rb-Sr dating by Randy Cormier of St. Francis Xavier University gave strong evidence that such was not the case, and granitoid petrological characteristics alone gave strong hints (e.g., Barr et al. 1982). However, U-Pb dating (beginning in the late 1980s) was required in order to establish clear patterns (e.g., Dunning et al. 1990; Barr et al. 1990). Williams (1978) had included all of Cape Breton Island in the Avalon Zone on his classic map of the Appalachian orogen because that map was made before much was known about the geology of the island. Geological mapping, petrological studies, and geochronology subsequently revealed the geological complexity, the demonstration of which is one of the goals of this field trip. It is now widely recognized that Cape Breton Island preserves a compressed cross-section of the Appalachian orogen from Avalonia in the southeast to Grenvillian basement in the northwest, a fortunate circumstance that has been attributed to promontory-promontory collision between elements of Laurentia and Gondwana (Lin et al. 1994).

A summary of the geology of Cape Breton Island is presented here as background for the trip. The details on which the interpretations are based are contained in the cited references (although even those have been kept to a minimum). Day 4 of the field trip focuses on the Mira terrane, Day 5 on Bras d'Or terrane, and Day 6 on Aspy terrane and the Blair River inlier.

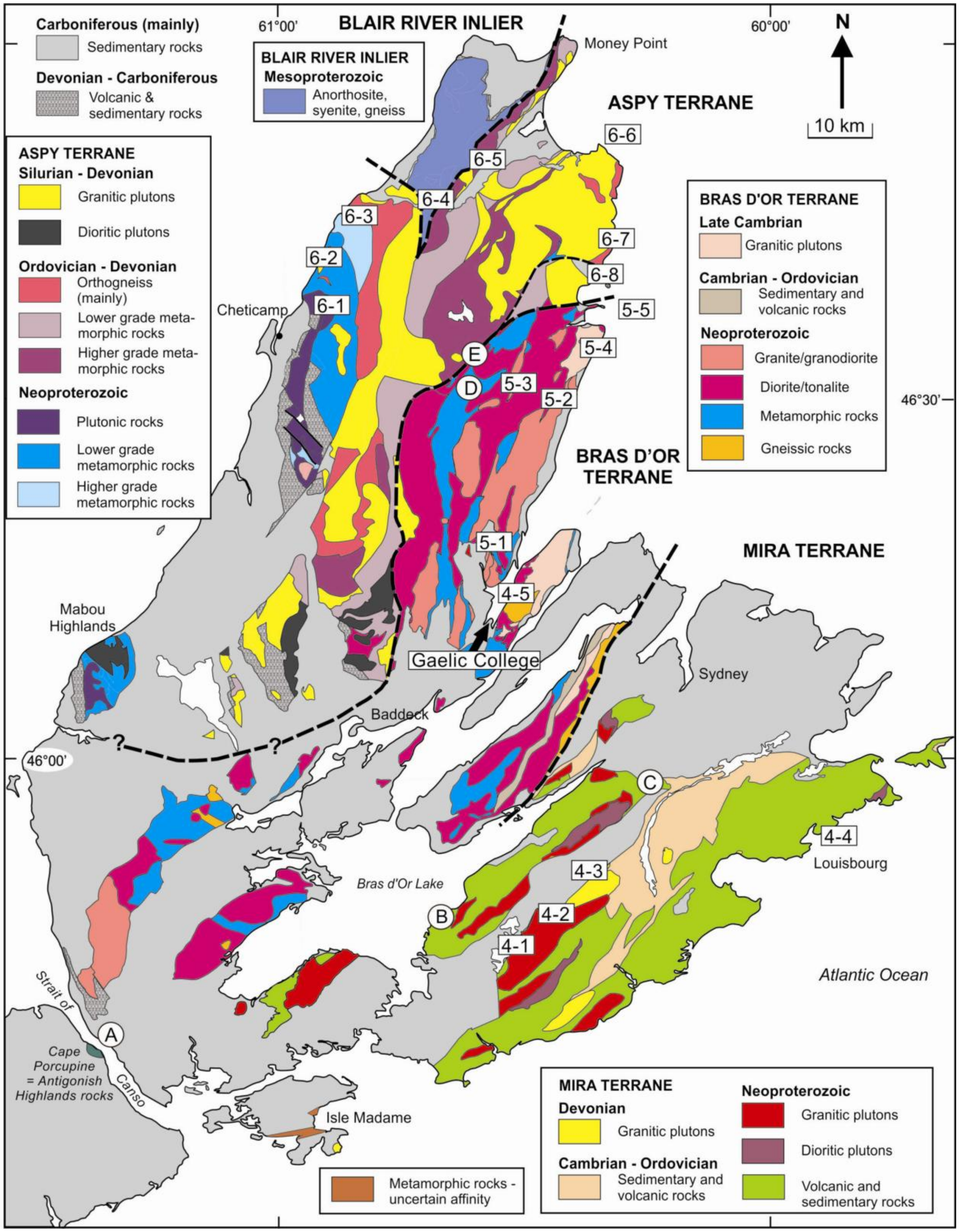

Figure 4-1. Simplified geological map of Cape Breton Island showing the major geological components of Mira, Bras d'Or, and Aspy terranes and the Blair River inlier and field trip stops for days 4, 5, and 6. Points of interest A, B, C, D, and E are described in the text.

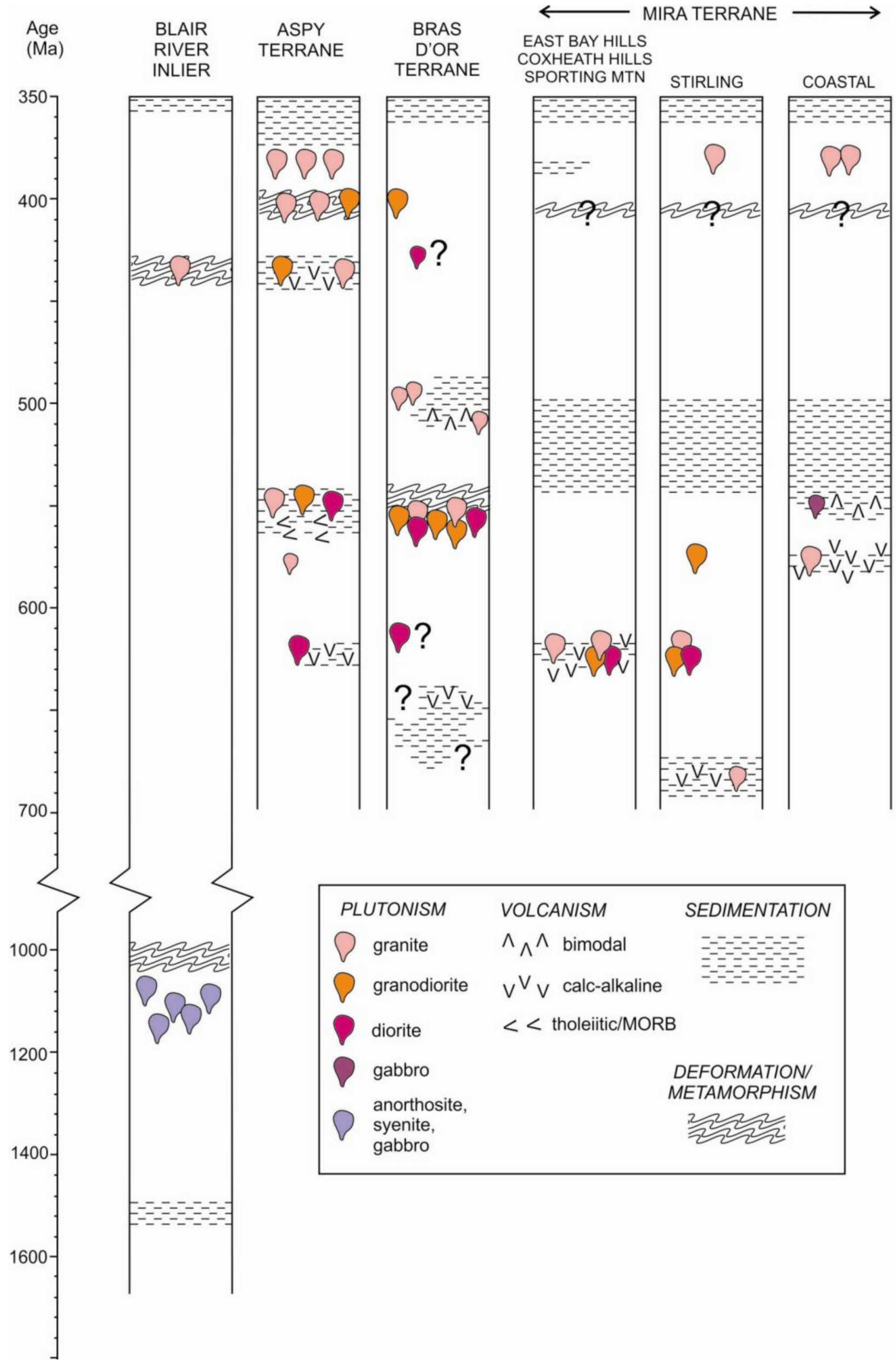

Figure 4-2. Schematic tectonostratigraphic columns summarizing the geological features of the Mira, Bras d'Or and Aspy terranes and the Blair River inlier. In Mira terrane, separate columns are shown for comparison among the Coastal, Stirling, and Coxheath Hill-East Bay Hills-Sporting Mountain belts.

MIRA TERRANE

Mira terrane is the only part of Cape Breton Island which belongs to Avalonia (Hibbard et al. 2006). In the northern Appalachian orogen, Avalonian rocks occur in southeastern New England (USA), the Caledonian Highlands of southern New Brunswick, the Cobequid and Antigonish highlands of northern mainland Nova Scotia, Mira terrane in Cape Breton Island, and the Avalon platform of eastern Newfoundland (Fig. 4-3). Some authors refer to these areas collectively as West Avalonia, to distinguish them from the components of East Avalonia in the UK and elsewhere in Europe. In this guidebook, the term is used to refer only to the northern Appalachian orogen. The characteristic rocks of Avalonia are Middle to Late Neoproterozoic volcanic, sedimentary, and plutonic rocks, although specific ages vary from area to area. Most Avalonian areas also include overlying Cambrian to Lower Ordovician clastic sedimentary units.

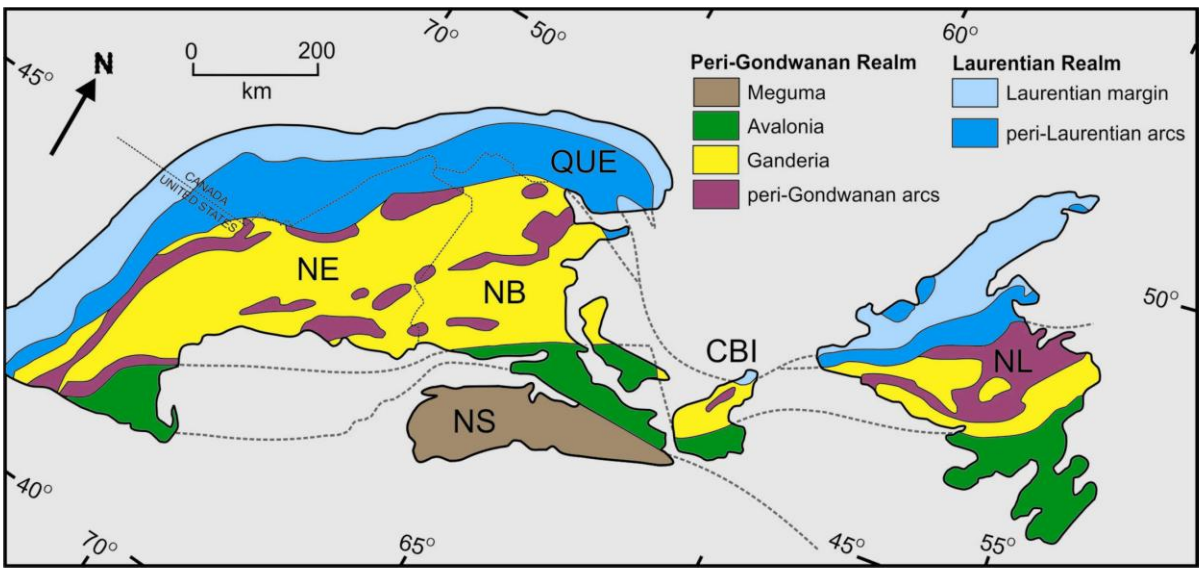

Figure 4-3. Divisions of the northern Appalachian orogen after Hibbard et al. (2006). Abbreviations: CBI, Cape Breton Island; NE, New England states of the USA; NL, Newfoundland, NS, Nova Scotia; NB, New Brunswick; Que, Quebec.

In Mira terrane, Neoproterozoic rocks form three belts of mainly different ages (Figs. 4-2, 4-4): ca. 680 Ma (Stirling), ca. 620 Ma (Sporting Mountain, East Bay Hills, and Coxheath Hills), and 575-550 Ma (Coastal). All three belts are dominated by mafic to felsic volcanic and volcaniclastic rocks and varying abundances of inter-stratified epiclastic and clastic sedimentary rocks. Barr (1993) and Macdonald and Barr (1993) interpreted the Stirling belt to represent an intra-arc or back-arc basin (Fig. 4-5). In contrast, the ca. 620 Ma mainly volcanic, volcaniclastic, and plutonic rocks of the Coxheath Hills, Sporting Mountain, and East Bay Hills belts have petrochemical features typical of high-K calc-alkalic suites formed at continental margin subduction zones (Barr 1993; Barr et al. 1996). However, the presence of plutonic rocks of this age in the Stirling belt indicates that these two belts were juxtaposed by that time (Fig. 4-5). These composite dioritic to granitic ca. 620 Ma plutons (Fig. 4-6) are the most extensive plutons in the Mira terrane.

The ca. 575 Ma mainly tuffaceous volcanic rocks of the Coastal belt (Fourchu Group) appear to be transitional between calc-alkalic and tholeiitic chemical affinity. They are inferred to represent magmas derived early in the development of a ca. 575 Ma northwest-dipping (present coordinates) subduction zone (Fig. 4-5). High-level plutonic rocks are only a minor component. The other major component of the Coastal belt, the Main-à-Dieu Group, contains lava flows, tuffs, debris flows, and fine-grained epiclastic rocks interpreted to have been deposited in intra-arc basins developed adjacent to the stratovolcanoes represented by the tuffs and flows of the Fourchu Group (Barr 1993; Barr et al. 1996). The Main-à-Dieu Group is overlain by mainly clastic marine sedimentary rocks of Cambrian to Early Ordovician age (Barr et al. 1996).

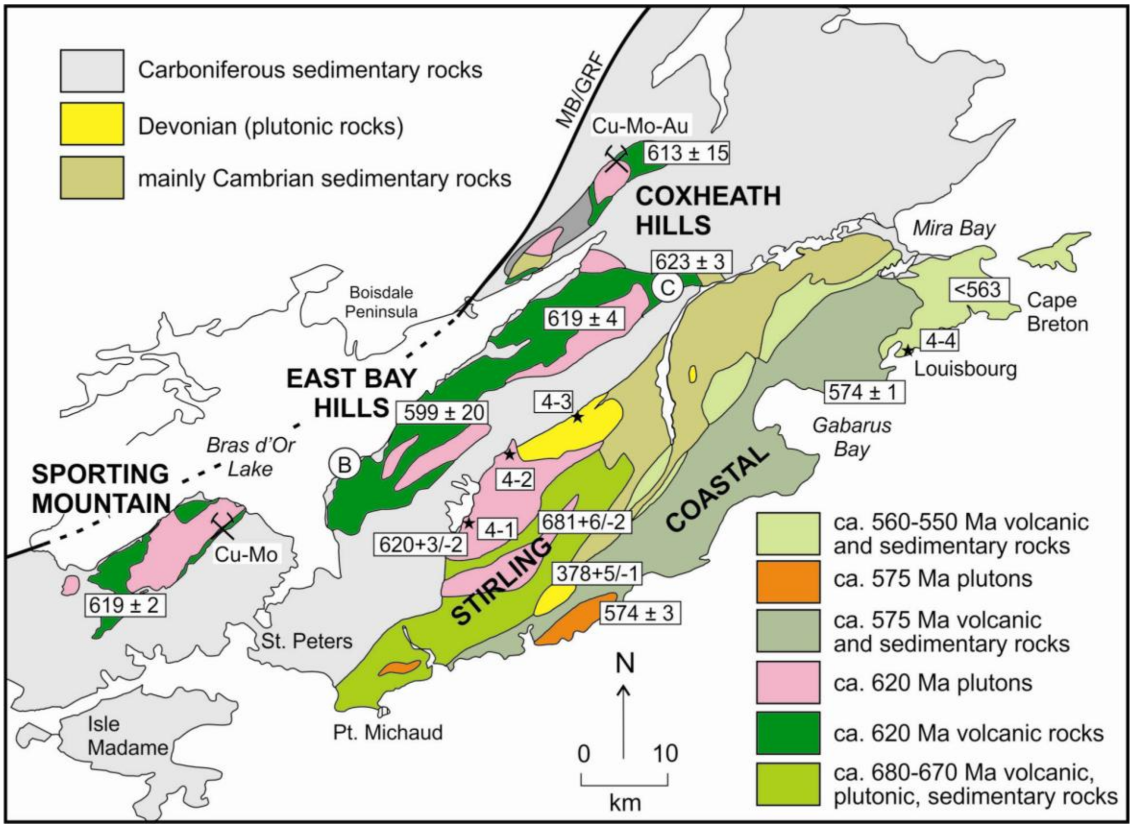

Figure 4-4. Simplified geological map of the Mira terrane (after Barr et al. 1996) showing lithotectonic belts described in the text, U-Pb (zircon) ages in Ma, and locations of stops (stars) for Day 4. Sites B and C are points of interest described in the text.MB/GRF is the McIntosh Brook- George River fault.

Devonian plutons are also present in the Mira terrane. They are shallow intrusions with associated porphyry-type, greisen-hosted, and vein-hosted Cu-Mo-Pb-Ag-Bi mineralization (Barr and Macdonald 1992). Petrological characteristics suggest that they are subduction-related plutons, but no other evidence of a Devonian magmatic arc occurs in southern Cape Breton Island. The apparent arc-signatures could reflect the nature of their source rocks in the roots of a Neoproterozoic arcs, or they could be associated with a more outboard subduction zone relating to juxtaposition of Gondwana with Meguma (e.g., Moran et al. 2007).

The boundary between Mira terrane and Bras d'Or terrane to the north is a "cryptic suture", buried beneath Carboniferous sedimentary rocks or located under water in Bras d'Or Lake (Figs. 4-1, 4-4). On maps, it is rather arbitrarily placed at conveniently located Carboniferous faults through the Boisdale Peninsula. The presence of clasts derived from both Mira terrane and Bras d'Or terrane units in a Middle Devonian conglomerate (McAdams Lake Formation) south of this fault shows that the two areas were in proximity by that time (White and Barr 1998). Magnetic and gravity models across the boundary suggest that the Mira terrane has been thrust under Bras d'Or terrane at the boundary (King 2002).

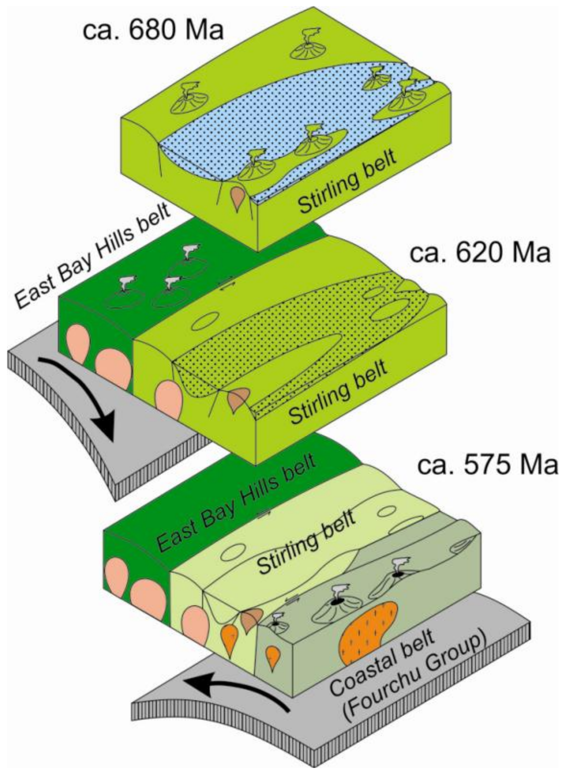

Figure 4-5. A possible tectonic scenario for the development of the volcanic-sedimentary-plutonic belts of Mira terrane (after Macdonald and Barr 1993).

Figure 4-6. A selection of cut slabs from the ca. 620 Ma plutons in Mira terrane, ranging in composition from diorite (upper left) to syenogranite (lower left). Slabs have been etched and stained to emphasize K-feldspar (yellow), plagioclase (white), and quartz (grey). Mafic minerals are dark grey/black.

BRAS D'OR TERRANE

Bras d'Or terrane, like Mira, is dominated by Neoproterozoic and Cambrian rocks but they differ from those of Mira terrane in composition and age, and hence Bras d'Or is interpreted to be part of Ganderia, not Avalonia (Barr et al. 1998; Hibbard et al. 2006). Rocks similar to those in Bras d'Or terrane also have been recognized elsewhere in Ganderia (e.g., O'Brien et al. 1991; White and Barr 1996; Rogers et al. 2006) but they are best exposed in Bras d'Or. Bras d'Or terrane contains fault-bounded blocks of Neoproterozoic low-pressure amphibolite-facies gneiss and much more extensive belts of greenschist-facies (and in places higher grade) quartzite, marble, meta-greywacke, and minor volcanic rocks (Raeside and Barr 1990, 1992). The relationship between the two suites of metamorphic rocks is unknown but in the equivalent Brookville terrane of southern New Brunswick, U-Pb dating has indicated that the gneissic rocks may be younger than the lower grade rocks (Bevier et al. 1990) and we have hints of that in Bras d'Or also (White et al. 2003). In any case, the lithological assemblages and metamorphic histories are so different that it is unlikely that they are equivalent units with different metamorphic histories (Raeside and Barr 1990; White and Barr 1996).

Both metamorphic suites are intruded by a large volume of Late Neoproterozoic (mainly ca. 560-553 Ma) subduction zone-related dioritic, tonalitic, granodioritic, and granitic plutons (Figs. 4-1, 4-7, 4-8). Plutonic rocks are especially abundant in the eastern Cape Breton Highlands, where several of the plutons contain high-alumina hornblende and magmatic epidote, indicative of crystallization at pressures of over 800 MPa (25 km depth) (Farrow and Barr 1989). These rocks are interpreted to represent the deep levels of an Andean-type continental margin subduction zone, whereas plutons (and in places co-magmatic volcanic rocks) in the southern part of the terrane represent higher level parts of the same subduction zone igneous assemblage (Fig. 4-9). Post-orogenic Late Cambrian granitic plutons are also present, and Middle Cambrian to early Ordovician volcanic and sedimentary rocks are preserved in a down-faulted block known as the Bourinot belt in the Boisdale Peninsula. Despite some similarity to the Cambrian sequence on Mira terrane, the Bourinot belt is firmly linked to Bras d'Or terrane and hence Ganderia. Similar Cambrian-Ordovician rocks also occur in Ganderia in southern New Brunswick (Fyffe et al. 2009).

Barr et al. (1998) proposed that Neoproterozoic rocks of the Bras d'Or terrane and its equivalents exposed in southern New Brunswick and locally in central Newfoundland represent the "basement" on which Paleozoic rocks were deposited, and these Paleozoic rocks (e.g., the Gander Group in Newfoundland), which dominate Ganderia elsewhere, were assumed to have been derived from the Bras d'Or terrane (Fig. 4-9).

A major mylonitic high-strain zone known as the Eastern Highlands shear zone (Lin, 1995, 2001) separates Bras d'Or terrane from Aspy terrane to the north. This boundary has a long and complex history, and the original relationship between Bras d'Or and Aspy terranes was likely as basement and cover (Chen et al. 1995). Bras d'Or terrane appears to have been thrust to the northwest over Aspy terrane (Fig. 4-9), and much of the original terrane is missing - the part of Bras d'Or terrane now adjacent to Aspy was unaffected by and hence probably far away during the Silurian-Devonian events which are so prominently recorded in Aspy terrane. These mid-Paleozoic events are not generally recorded in Bras d'Or terrane rocks, except near the Eastern Highlands shear zone, where $^{40}Ar/^{39}Ar$ dating revealed overprinting in Neoproterozoic rocks by younger thermal events associated with Bras d'Or-Aspy terrane collision (Reynolds et al. 1989).

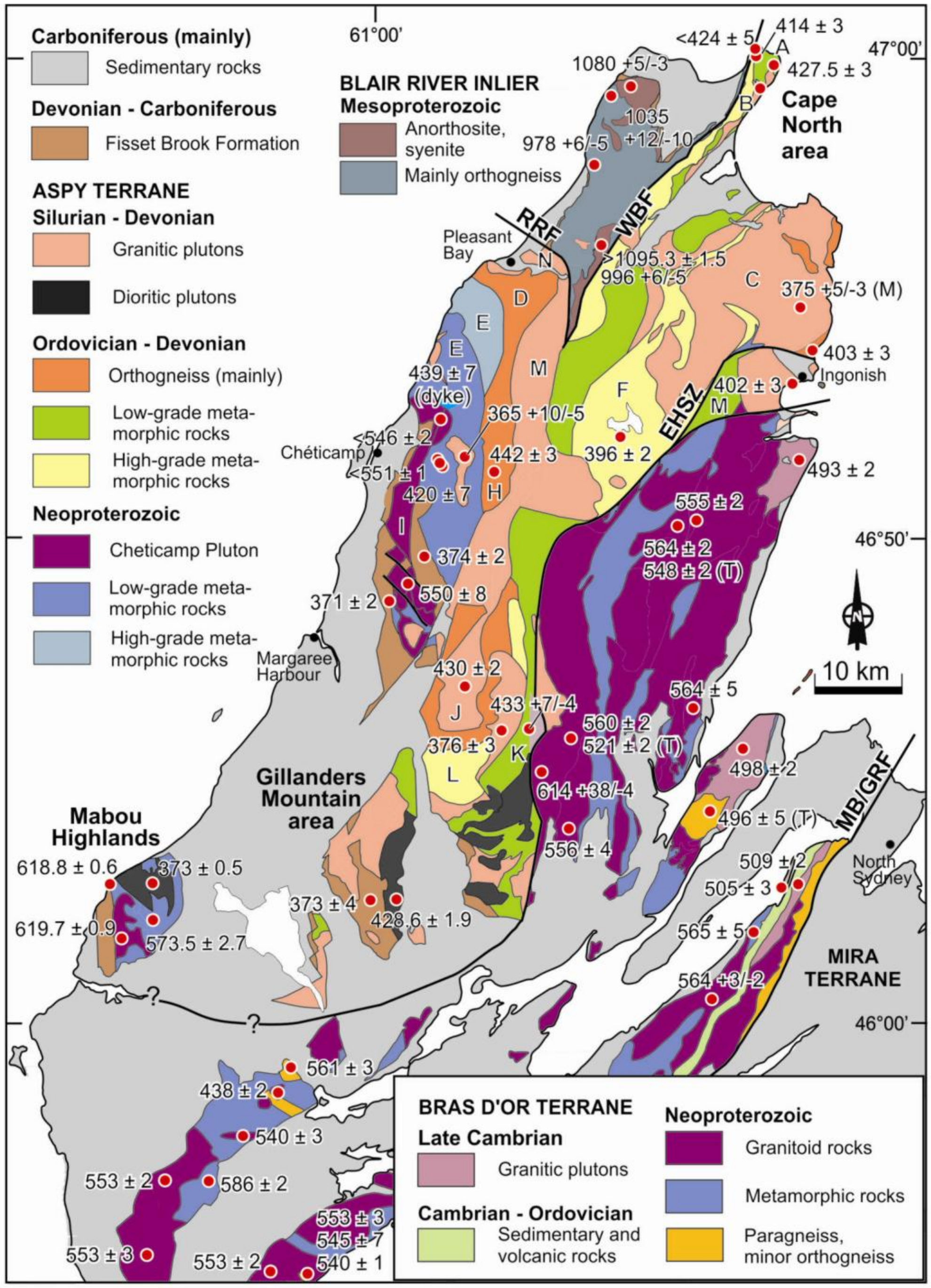

Figure 4-7. Compilation of U-Pb age data (in Ma) for the Bras d'Or and Aspy terranes and the Blair River inlier (modified from Lin et al. 2007). U-Pb ages are for zircon unless otherwise indicated: M, monazite; T, titanite. Lettered units (more or less from north to south) are: A, Money Point Group; B, Cape North Group; C, Black Brook Granitic Suite; D, Pleasant Bay Complex; E, Jumping Brook Metamorphic Suite; F, Cheticamp Lake Gneiss; G, Clyburn Brook Formation; H, Belle Cote Orthogneiss; I, Cheticamp Pluton; J, Taylors Barren Granite; K, Sarach Brook Metamorphic Suite; L, North Branch Baddeck River leucotonalite; M, Margaree Pluton; N, Andrews Mountain Granite. Abbreviations: RRF, Red River fault; WBF, Wilkie Brook fault; EHSZ, Eastern Highlands shear zone; MB/GRF, MacIntosh Brook/Georges River fault.

Figure 4-8. A selection of cut slabs from the ca. 560 Ma plutons in Bras d'Or terrane, ranging in composition from diorite (upper left) to syenogranite (lower right). Slabs have been etched and stained to emphasize K-feldspar (yellow), plagioclase (white), and quartz (grey). Mafic minerals are dark grey/black.

ASPY TERRANE

Aspy terrane (Figs. 4-1, 4-7) contains low- to high-grade metavolcanic and metasedimentary rocks and large areas of orthogneiss and less abundant paragneiss, all metamorphosed in the late Silurian - early Devonian (ca. 420 - 400 Ma) (Dunning et al. 1990; Reynolds et al. 1989; Barr et al. 1998; Price et al. 1999; Horne et al. 2003; Lin et al. 2007). The protolith ages for these metamorphic rocks are somewhat uncertain, but they appear to include both Neoproterozoic and Ordovician-Silurian components (Lin et al. 2007). Rocks with ages of ca. 620 and 550 Ma occur in the western part of the terrane (Mabou and Cheticamp areas) and have difficult-to-demonstrate relationships with the younger mid-Paleozoic metamorphic rocks (and associated plutons) that dominate in the rest of the terrane, but are likely to constitute their basement. The younger metamorphic rocks include metavolcanic, metasedimentary, and gneissic rocks of Ordovician and Silurian (ca. 450-430 Ma) age. They were involved in high-pressure amphibolite-facies metamorphism in the late Silurian - early Devonian (ca. 400 - 420 Ma). They are subdivided into map units separated by plutons, faults, or Carboniferous rocks and assigned local names because of the difficulty of making correlations throughout the area (e.g., Barr and Jamieson 1991; Barr et al. 1992; Lin et al. 2007). Metaconglomerate in the eastern part of the Aspy terrane yielded detrital zircon ages including ca. 495 Ma grains that suggest an original depositional link with Bras d'Or terrane which contains plutons of that age (Lin 1993; Chen et al. 1995).

Metamorphic units in the Aspy terrane are intruded by abundant plutons with igneous crystallization ages of ca. 442 Ma, 430 Ma, 400 Ma, and 375-365 Ma (Dunning et al. 1990; Horne et al. 2003). In simplistic terms, the plutons can be said to include all of "I-type, S-type, and A-type" (Barr 1990). The oldest plutons are orthogneissic/foliated dioritic, tonalitic, and granitic plutons with ages of ca. 440-425 Ma and petrological features consistent with formation in a subduction zone setting (Price et al. 1999). Together with volcanic rocks

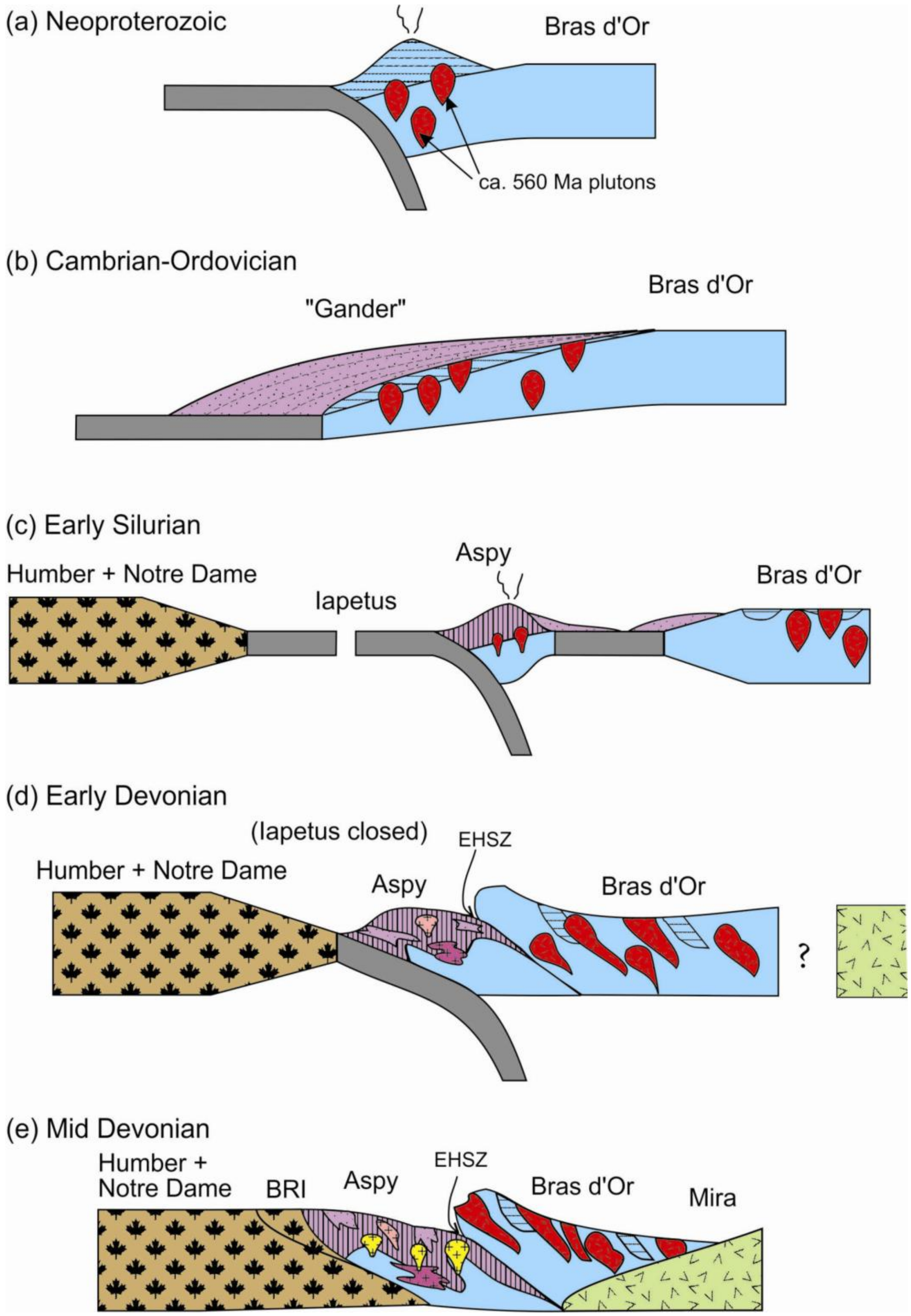

Figure 4-9. A possible tectonic model for the development of Bras d'Or and Aspy terranes after Barr et al. (1998). Abbreviation: EHSZ, Eastern Highlands shear zone.

of similar ages, they may have formed in a Japan-type setting offshore from Bras d'Or terrane and on Bras d'Or terrane crust (Fig. 4-9). Younger ca. 400 Ma plutons mainly occur within splays of the Eastern Highlands shear zone and may have formed in conjunction with the early stages of juxtaposition with Bras d'Or terrane. They also have volcanic-arc characteristics (Barr 1990). The most pervasive units in Aspy terrane are "S-type" granodioritic to granitic plutons and associated pegmatite with a minimum age of about 375 Ma, based on a U-Pb (monazite) age of 375+5/-3 Ma from the Black Brook Granitic Suite (Fig. 4-7). However, based on the fact that they are older than the Bothan Brook Granite dated at 376 ± 3 Ma (U-Pb zircon; Horne et al. 2003) and that they are an intimate part of the Cheticamp Lake Gneiss where they have yielded a U-Pb age of 396 ± 2 Ma (Fig. 4-7), their age is probably closer to 400 Ma than 375 Ma. These plutons have "syn-collisional" petrological characteristics, and may have formed as a result of increased compression between Aspy and Bras d'Or terranes during docking of Mira terrane in the mid-Devonian (Fig. 4-9).

Metamorphic and $^{40}Ar/^{39}Ar$ studies indicate that the Aspy terrane cooled quickly from ca. 600°C through 400°C between 386 Ma and 370 Ma, and by ca. 375 Ma, bimodal volcanic rocks and nonmarine sedimentary successions were forming in rift basins, mainly in Aspy terrane but also locally in Bras d'Or terrane. The bimodal volcanic rocks have within-plate characteristics (e.g., Barr et al. 1995; Barr and Peterson 1998), as do the related granitic plutons. These granitic plutons include the megacrystic Margaree Pluton with its distinctive plagioclase-mantled K-feldspar crystals (Fig. 4-10), as well as widespread coarse-grained equigranular granitic plutons.

Such plutons are absent from Bras d'Or terrane, except in the easternmost highlands where the Eastern Highlands shear zone splays; one splay (the main one?) appears to be stitched by the Black Brook Granitic Suite (Yaowanoiyothin and Barr 1991), whereas the 402 Ma Cameron Brook Granodiorite was emplaced in the southern splay and may also be a stitching pluton (Dunning et al. 1990).

The abundant plutonic components that characterize Aspy terrane also typify other parts of Ganderia. Van Staal et al. (2009) interpreted an equivalent range of granitoid rocks in New Brunswick and central Newfoundland mainly to slab break-off (Silurian) and flat-slab subduction (Devonian), and similar models may apply (in a simplistic way at least) to plutons in Aspy terrane.

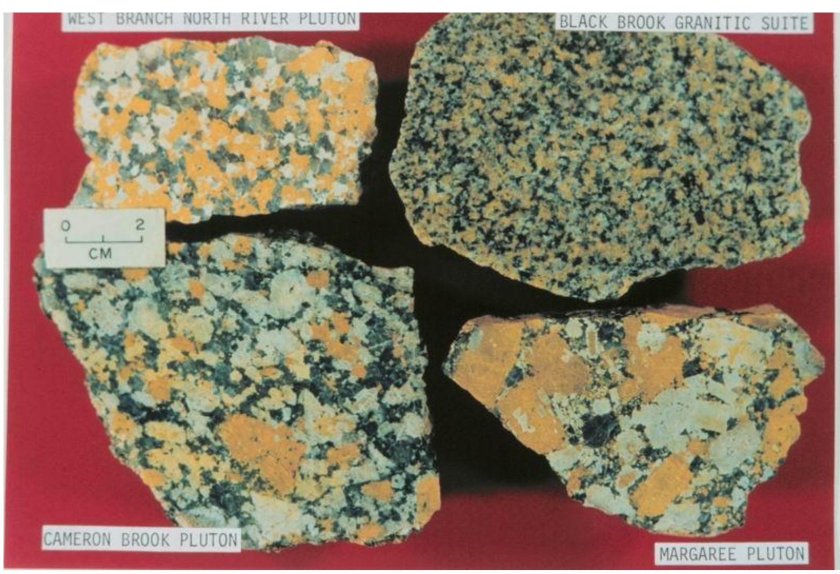

Figure 4-10. A selection of cut slabs from the ca. 402 Ma (Cameron Brook) and 375 Ma plutons in Aspy terrane. Slabs have been etched and stained to emphasize K-feldspar (yellow), plagioclase (white), and quartz (grey).

On the north, Aspy terrane is separated from the adjacent Blair River inlier by the steep Red River and Wilkie Brook fault zones (Fig. 4-1, 4-7) which form a geometrically impossible-looking “V-shape”, with the junction of the “V” cut and thus obscured by Devonian granitic plutons (Margaree and Andrews Mountain-type plutons). Such plutons “stitch” the bounding faults of the Blair River inlier, and show that juxtaposition with Aspy was completed by ca. 375 Ma (Barr et al. 1996). The configuration of the bounding faults at depth is unknown. No units equivalent to the peri-Laurentian arcs and back-arcs of western Newfoundland and Quebec (Fig. 4-3) have yet been identified in Cape Breton Island.

BLAIR RIVER INLIER

The Blair River inlier forms the northwestern part of Cape Breton Island. It consists mainly of composite orthogneissic units, intruded by less deformed plutons of varied compositions including anorthosite, gabbro, syenite, and granite (Fig. 4-11). Miller *et al.* (1996) and Miller and Barr (2000) reported U-Pb dates confirming that major units in the inlier are of Mesoproterozoic age, including the Sailor Brook Gneiss (>1217 Ma), Lowland Brook Syenite (1080 +5/-3 Ma), Red River Anorthosite Suite (>1095Ma), and Otter Brook Gneiss (978 +6/-5 Ma). They also showed that high-grade metamorphism of the Sailor Brook Gneiss occurred at 1035 +12/-10 Ma, and that the Red River Anorthosite Suite was metamorphosed at 996+6/-5 Ma.

Figure 4-11. A selection of cut slabs from the Blair River inlier. Upper and lower left: orthogneiss; upper right, syenite; lower middle, anorthosite; lower right, ferrogabbro. Mafic minerals are darker grey/black.

Paleozoic igneous activity in Blair River inlier is demonstrated by the 435 +7/-3 Ma age of the Sammys Barren Granite (Fig. 4-7). In addition, Paleozoic amphibolite-facies metamorphism is reflected in ca. 425 Ma titanite ages from the Proterozoic units, and subsequent cooling through hornblende, muscovite, and phlogopite $^{40}Ar/^{39}Ar$, and rutile U-Pb, closure temperatures lasted until ca. 410 Ma (Barr et al. 1996; Miller *et al.*, 1996). Although similar in timing to events in adjacent Aspy terrane, these events are unlikely to reflect direct interaction of the Blair River inlier with the Ganderian Aspy terrane, but instead with peri-Laurentian elements incised by subsequent tectonic events. A later, probably Devonian, greenschist-facies overprint is most intense near chlorite-grade shear zones and brittle fault zones, and likely reflects juxtaposition with Aspy terrane.
The Mesoproterozoic units of the Blair River inlier are distinct in both age and composition from rocks in other parts of Cape Breton Island and in northern Appalachian outboard terranes in general. They contain rock types and ages similar to those characteristic of the Grenville Province of Laurentia and similar to those in other Grenvillian basement inliers in the Appalachian orogen, such as the Steel Mountain, Indian Head, and Long

Range inliers in western Newfoundland (e.g., Heaman et al. 2002). Thus, the Blair River inlier is interpreted to be an exposure of Laurentian Grenvillian basement that was deformed, metamorphosed, and intruded by granite during Appalachian orogenic events (Miller *et al.*, 1996; Barr *et al*. 1996, 1998).

The faults that bound the Blair River inlier are marked by steep mylonite zones mainly in Blair River lithologies; contacts between these rocks and Aspy terrane rocks are mainly brittle faults, although Aspy terrane rocks are also locally mylonitic, especially near the bounding faults. Offshore seismic profiles (e.g., Loncarevic et al. 1989) are not clear about the relationship of the Blair River inlier to Grenvillian basement in the Gulf of St. Lawrence, but it is likely that the Blair River inlier is a "flake" that has been detached from its Grenvillian roots in the St. Lawrence promontory (Lin et al. 1994).

DAY 4: MIRA TERRANE (AVALONIA)

Summary of Itinerary (Figs. 4-1, 4-4)
Drive from Antigonish to Cape Breton Island, crossing the Canso Causeway en route. (approximate driving time to STOP 4-1 = 2 hours).
STOP 4-1: Typical (and dated) ca. 620 Ma granodiorite of the Mira terrane (to compare with much younger Bras d'Or terrane granodiorite to be seen on Day 5).
STOP 4-2: A petrological treat – mixing/mingling textures in ca. 620 Ma diorite/quartz monzodiorite (Chisholm Brook Plutonic Suite)
STOP 4-3: Example of Devonian high-level granite in Mira – a beautiful rock!
STOP 4-4. Typical Coastal belt volcaniclastic rocks and a scenic introduction to Louisbourg (if the weather is nice!). [**+ LUNCH**] *[Note: National Historic Site – no hammers or sample collection]*
Cultural stop - visit to the Fortress of Louisbourg

<u>STOP Descriptions</u>

Point of interest A (45.647210^{o}N 61.413635^{o}W) The Canso Causeway opened in 1955. It was constructed from blocks of ca. 610 Ma granite (and cross-cutting mafic dykes of unknown age) from Cape Porcupine, the big cliff located southwest of the causeway. This rock is still quarried for use as aggregate, and mainly shipped to the USA. We now know that the Cape Porcupine rocks are a fragment of the Antigonish Highlands. No correlation is possible between any pre-late Devonian crystalline rock units in Cape Breton Island and mainland Nova Scotia; therefore, the Strait of Canso marks a major geological structure.

Point of interest B: East Bay Hills Group - Johnstown road cut (45.814505^{o}N 60.682830^{o}W)
(An alternative route to STOP 4-1 for those not travelling by bus)
Road cuts along Highway 4 expose dacitic and rhyolitic crystal and lithic lapilli tuffs typical of the ca. 620 Ma East Bay Hills Group in the East Bay Hills, Coxheath Group in the Coxheath Hills belt, and Pringle Mountain Group in the Sporting Mountain belt. Chemically these rocks have volcanic-arc signatures, and may have formed in an Andean- or Japan-style subduction zone. From Irish Cove, a narrow unpaved road crosses the East Bay Hills to Loch Lomond, passing outcrops of the Irish Cove Pluton, which is typical of the ca. 620 Ma plutonic rocks that are co-magmatic with the volcanic rocks of the East Bay Hills Group at B. These high-level intrusions consist mostly of fine- to medium-grained inequigranular monzogranite to granodiorite, characterized by subhedral to

euhedral plagioclase in a finer grained assemblage of K-feldspar, quartz, biotite, and hornblende. They are intimately associated with the volcanic rocks as small plutons, sills, and dykes. They are mineralogically and chemically similar to the Chisholm Brook granodiorite at STOP 4-1.

STOP 4-1: Chisholm Brook granodiorite at Loch Lomond (45.755461°N 60.569064°W)
The Chisholm Brook granodiorite is part of the large Chisholm Brook Plutonic Suite. The plutonic suite also includes diorite and quartz monzodiorite (STOP 4-2) but granodiorite such as that at this stop is the most abundant rock type, at least at the current level of erosion. A sample from this location yielded a U-Pb (zircon) crystallization age of 620 +3/-2 Ma (Barr et al. 1990). Similar $^{40}Ar/^{39}Ar$ (hornblende) cooling ages from other ca. 620 Ma plutons in Mira terrane (Keppie et al. 1990) are consistent with high-level emplacement and rapid cooling of these plutons. The granodiorite ranges from medium- to coarse-grained, and consists of equigranular plagioclase, quartz, alkali feldspar, biotite, and hornblende. Like the East Bay Hills Group and associated plutons, the Chisholm Brook Plutonic Suite has volcanic-arc chemical signatures, and likely formed in an Andean- or Japan-style subduction zone (Fig. 4-5). The Chisholm Brook Plutonic Suite intruded ca. 680 Ma volcanic and sedimentary rocks of the Stirling belt which we will not be seeing on this trip because of time constraints.

STOP 4-2: Chisholm Brook quartz monzodiorite (45.811975°N 60.524355°W)
Quartz monzodiorite, a relatively minor component of the ca. 620 Ma Chisholm Brook Plutonic Suite, is well exposed here in a small quarry, where is it is mingled and mixed with more mafic "dioritic" rocks (Fig. 4-12). The resulting inhomogeneous rocks have varied textures and relative proportions of minerals, but a characteristic feature is needle-like hornblende crystals that reach several centimetres in length. These types of petrological details have received almost no study in Cape Breton Island, so if anyone is interested, you are welcome to collect samples.

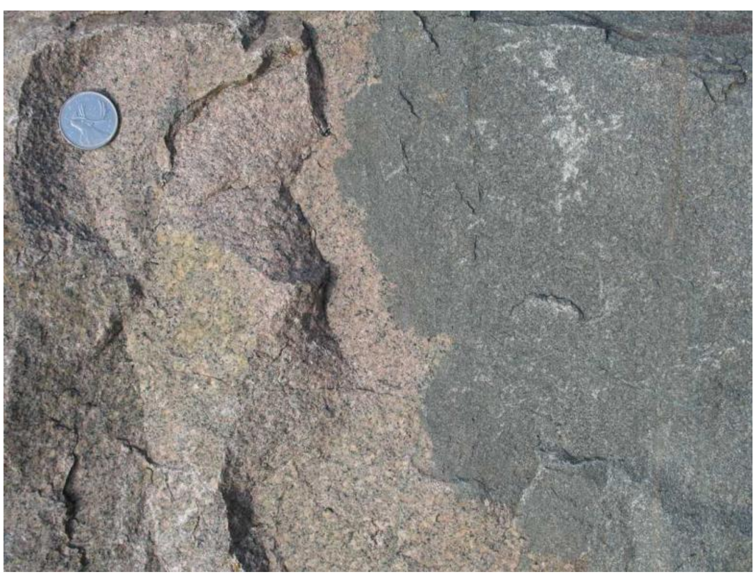

Figure 4-12. Outcrop photograph of quartz monzodiorite and diorite of the Chisholm Brook Plutonic Suite at STOP 4-2.

STOP 4-3: Salmon River rhyolite porphyry (45.858886°N 60.361463°W)
In addition to the Neoproterozoic plutonic suites, the Mira terrane also contains Devonian plutons, including the Salmon River rhyolite porphyry at this stop. This location is the site of the former Yava mine, active in the 1980s, a stratabound lead deposit hosted by Upper Carboniferous sandstone of the Silver Mine Formation. The sandstone nonconformably overlies the rhyolite porphyry (or would you call it a fine-grained poprhyritic granite?), which has well developed exfoliation joints parallel to the contact. Sangster and Vaillancourt (1990) interpreted the lead in the sandstone to have been derived from the rhyolite porphyry and/or the ca. 620 Ma Chisholm Brook Plutonic Suite.

Field observations (not visible at this location) show that the Salmon River rhyolite porphyry intruded both the Chisholm Brook Plutonic Suite and overlying Cambrian coarse clastic rocks of the Kelvin Glen Group. Locally concordant intrusive contacts and the presence of sills of rhyolite porphyry in the Kelvin Glen Group suggest that it is laccolithic or sheet-like in form (McMullin 1984). No contact metamorphic effects are obvious, but Neoproterozoic detrital muscovite in the Kelvin Glen Group have undergone Silurian or younger partial argon loss (P. H. Reynolds, 1991, unpublished data). U-Pb dating of the rhyolite porphyry was attempted but the sample yielded only sparse xenocrystic zircon of Archean age (Doig et al. 1990).

The pluton is relatively homogeneous over its entire area, although the groundmass grain size varies from very fine (as at STOP 4-3) to medium. Phenocrysts include sanidine, quartz, plagioclase, and rarely altered muscovite (possibly xenocrystic in origin?). The groundmass is granular and consists of feldspar, quartz, biotite (mostly altered to chlorite), and abundant sericite. Feldspar phenocrysts are partly altered to sericite and saussurite but it is a good teaching sample for sanidine identification (Fig. 4-13).

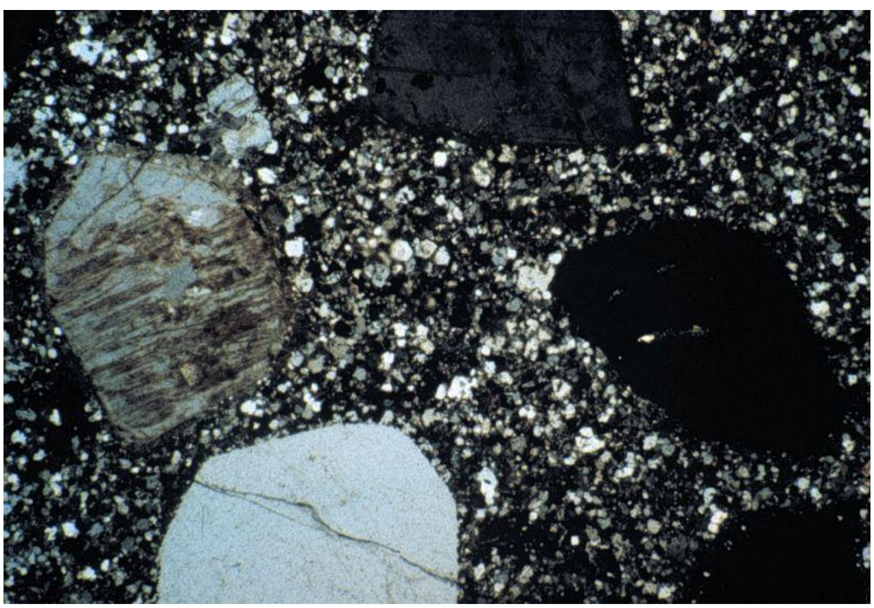

Figure 4-13. Photomicrograph of rhyolite porphyry from the Salmon River pluton (STOP 4-3). Phenocrysts are quartz, plagioclase, and sanidine. Field of view is 4 mm wide. Crossed polars.

Point of interest C: East Bay Hills Group – Morley Road quarry. (45.991060°N 60.264840°W)
Spectacular columnar-jointed rhyolite and felsic tuff of the East Bay Hills Group are exposed in a large quarry on Morley Road. Although early attempts to date the rhyolite here and elsewhere in the East Bay Hills Group initially yielded somewhat problematic results with large errors, perseverance resulted in a better age of 623 ± 3 Ma (Bevier et al. 1993; Barr et al. 1996), consistent with co-magmatic pluton ages (Fig. 4-4).

STOP 4-4. Main-à-Dieu Group, Louisbourg Lighthouse (45.906478°N 59.957268°W)
The Main-à-Dieu Group (age ca. 560-542 Ma) is the youngest of the Neoproterozoic components of the Mira terrane, and underlies the Cambrian rocks of the Mira Group. It consists of volcanic rocks, including basalt and rhyolite flows, varied (and commonly spectacular) lithic lapilli tuff, laminated ash tuff, and red clastic sedimentary sequences. The exposures around the lighthouse provide an opportunity to examine a variety of volcaniclastic rocks typical of the Main-à-Dieu Group. If the weather is clear, you also will have a scenic view of the Fortress of Louisbourg.

Lunch at STOP 4-4.

Cultural stop - visit to Fortress of Louisbourg. We will have a guided tour of the Fortress, and you have an introductory brochure in your registration package.

STOP 4-5: St. Anns look-off (time and weather permitting) (46.245716°N 60.514786°W)
From the look-off above St. Anns you can see the eastern Highlands as far as Cape Smokey, providing an overview of the route for Day 5 (Fig. 4-1). This part of the Highlands is composed mostly of granitoid rocks, with quartzite and other metasedimentary rocks of the McMillan Flowage Formation making up the prominent ridge to the northwest in this view. The lower land is underlain by sandstone, limestone, and gypsum of the Carboniferous Horton and Windsor groups.

At the look-off, you are standing on cordierite-bearing gneiss (Jamieson 1984) of Kellys Mountain, also part of the Bras d'Or terrane. These low-pressure gneissic rocks are characteristic of the Bras d'Or terrane. The gneiss is intruded by the late Cambrian (ca. 498 Ma) Kellys Mountain Granite, as well as by diorite and intermediate granitoid rocks with a probable age of ca. 560 Ma (based on comparisons to similar rock types elsewhere in the terrane).

Proceed to the Gaelic College in St. Anns. (46.212751°N 60.604496°W)

DAY 5: BRAS D'OR TERRANE (GANDERIA)

Summary of Itinerary (Fig. 4-1)
9:00 – 10:30 am: cultural stop – Alexander Graham Bell Museum, Baddeck.

STOP 5-1: Typical 560 Ma granodiorite of the Bras d'Or terrane (compare to Mira)

Lunch: Plaster Picnic Park, Cabot Trail.

STOP 5-2: Typical 560 Ma granite of the Bras d'Or terrane (compare to Mira).
STOP 5-3: Ingonish River Tonalite with magmatic epidote and high-Al hornblende.
STOP 5-4: Late Cambrian Cape Smokey Granite.
STOP 5-5: Hiking on Middle Head (560 Ma diorite, granite, and magma mingling).

STOP Descriptions

STOP 5-1: Indian Brook Granodiorite (46.348293°N 60.538072°W)
The Indian Brook Granodiorite is the largest of the dioritic to granitic plutons which form much of the southeastern part of Bras d'Or terrane. Petrological studies (e.g., Grecco and Barr 1999) indicate that these plutons constitute a co-magmatic suite formed in an Andean-type continental margin subduction zone, with magma evolution controlled by feldspar and amphibole fractionation.

The granodiorite at this location consists of medium-grained plagioclase, perthitic microcline, quartz, hornblende, and biotite, with abundant accessory titanite, zircon, apatite, and magnetite. It displays the intense red (hematite) staining that is typical of the high-level plutons in Bras d'Or terrane, and led early mappers in the area to term them "syenite" or "syenodiorite". Here, we are here near the top of the pluton, because in this area it intruded related volcanic rocks of the Price Point Formation.

U-Pb dating of titanite in the granodiorite has yielded an age of 564 ± 4 Ma, interpreted to be the approximate crystallization age of the granodiorite; the U-Pb zircon age is slightly older and less precise as a result of inheritance (Dunning et al. 1990). Before reliable geochronology shed light on the actual situation, these plutons were assumed to be similar in age to those that we saw in the Mira terrane, but they are about 50-70 million years younger than the similar-looking plutons of Mira terrane such as those of the Chisholm Brook Plutonic Suite that we saw at STOP 4-1. Numerous U-Pb (zircon) ages throughout Bras d'Or terrane show that this Andean-type magmatism occurred from 575 Ma to 540 Ma, with a peak around 560 Ma (Fig. 4-7).

Lunch: Plaster Picnic Park, Cabot Trail.

STOP 5-2: Wreck Cove Road - Birch Plain Granite (46.527842°N 60.431344°W)
The Birch Plain Granite is medium-grained biotite granite, of probable Late Neoproterozoic age (based on petrographic features and relations to adjacent dated units – it forms xenoliths in the Indian Brook Granodiorite - but the pluton itself has not been dated). It is texturally and mineralogically distinct from the younger (ca. 495 Ma) Cape Smokey Granite (STOP 5-4), and is inferred to be part of the same plutonic suite as the Indian Brook Granodiorite (Grecco and Barr 1999). Mineralogy in the granite includes plagioclase, perthitic microcline, quartz, and biotite (no amphibole). Allanite and zircon are abundant accessory phases, together with titanite, apatite, and magnetite. It locally displays weak foliation, but whether it is the result of flow or tectonism is unclear. It is cut by mafic dykes of uncertain age (but likely Devonian – see STOP 5-5) which have sharp contacts and well developed chilled margins.

STOP 5-3: Wreck Cove Road - Ingonish River Tonalite (46.551135°N 60.512494°W)
The Ingonish River Tonalite is a large, elongate pluton that intruded dioritic rocks of similar age on both its eastern and western margins. The tonalite is medium to coarse grained, and consists mainly of plagioclase (andesine), hornblende, biotite, and quartz. Green epidote of inferred igneous origin is a distinctive component (Fig. 4-17), visible even in hand specimen. It indicates that the pluton crystallized at a depth of ca. 25 km (Farrow and Barr 1992). The key petrographic features of magmatic epidote, such as zoning and symplectic intergrowths with plagioclase (e.g., Zen and Hammarstrom 1984), are well displayed in thin section. Foliation, although typically present, may not be obvious in coarse-grained samples. U-Pb dating of zircon has yielded an age of 555 ± 2 Ma, which overlaps the $^{40}Ar/^{39}Ar$ age of 554 ± 5 Ma from hornblende. The similarity in crystallization and hornblende cooling ages indicates rapid uplift, perhaps explaining the preservation of magmatic epidote.

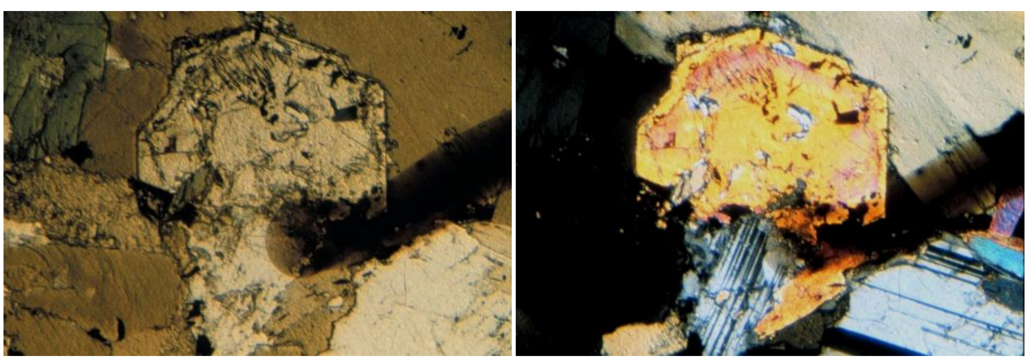

Figure 4-14. Photomicrographs of magmatic epidote in the Ingonish River Tonalite (STOP 5-3). Photograph on left is in plane polarized light; photograph on right is with crossed polars. Field of view is 4 mm wide.

Points of interest D (46.544835°N 60.616492°W) **& E** (46.579129°N 60.632325°W): Owing to time constraints, we will not be visiting the host rocks (D) for the Indian Brook, Birch Plain, and Ingonish River plutons. They include quartzite, pelitic rocks, and marble, of late Neoproterozoic and older age, which were regionally metamorphosed at the same time as the intrusion of the plutons. These rocks are typical of the continental margin sedimentary sequences of the Bras d'Or terrane. We also will not be visiting the Eastern Highlands shear zone (E), the boundary between Aspy and Bras d'Or terranes. It is a high-strain zone up to over 2 km in width, in which rocks were converted to mylonite and subsequently recrystallized in the greenschist facies to blastomylonite (Lin 1995). The age of at least some of the later movement on this shear zone is Devonian, constrained by the fact that it affected (slightly) the ca. 375 Ma Black Brook Granitic Suite of the Aspy terrane, and reset $^{40}Ar/^{39}Ar$ ages in the ca. 560 Ma Kathy Road Diorite of the Bras d'Or terrane to 415 Ma.

STOP 5-4: Cape Smokey Granite (at the picnic park, summit of Cape Smokey) (46.587714°N 60.381551°W)
Cape Smokey is underlain by pink leucocratic biotite granite of late Cambrian age (493 ± 2 Ma; Dunning et al. 1990). The granite is equigranular and unfoliated (post-tectonic), and consists of approximately equal amounts of plagioclase, orthoclase, and quartz. Biotite forms only 1-2% and is mainly altered to chlorite. Similar granite of similar age (498 ± 2 Ma; Dunning et al. 1990) forms much of Kellys Mountain where it intrudes Bras d'Or gneiss. The tectonic setting for these large plutons is enigmatic. They have volcanic-arc chemical signatures (Barr 1990). The only constraints on tectonic activity in Bras d'Or terrane at that time is in the Bourinot belt, where bimodal volcanic rocks and a syenogranite pluton have ages of 505 ± 3 Ma and 509 ± 2 Ma, respectively (White et al. 1994). These rocks have within-plate characteristics and have been linked to separation of Ganderia from Gondwana. Perhaps the Kellys Mountain and Cape Smokey plutons were related to that event also.

STOP 5-5: Middle Head - Wreck Cove Diorite, Cape Smokey Granite, and mingled dyke (46.655787°N 60.373196°W)
We will finish the day with a short hike along Middle Head where we can see the varied dioritic rocks of the ca. 560 Ma Wreck Cove Diorite intruded by ca. 493 Ma Cape Smokey Granite. This area was described in classic papers by Wiebe (1973, 1974). He interpreted the eastern part of Middle Head as a layered and differentiated diorite pluton with layers and inclusions of hornblende gabbro, and inferred that the gabbroic magma was injected in pipe-like bodies into a chamber of actively fractionating dioritic magma. We will not have time to investigate all of the features described by Wiebe (1973, 1974) but will focus on a granitic dyke mingled with locally vesicular gabbroic rocks (Fig. 4-18). Classic mingling textures are well displayed, as documented by Wiebe (1973). The ages of the gabbroic and granitic material are uncertain but my "best guess" is that they are Devonian, the age of rhyolite that forms Ingonish Island located 3 km to the north in Ingonish Bay.

Figure 4-15. Photograph of the magma mingling features in a granitic/gabbroic dyke of probable Devonian age on Middle Head (photograph courtesy of Barrie Clarke).

Return to the Gaelic College.

DAY 6: ASPY TERRANE (GANDERIA)AND BLAIR RIVER INLIER (LAURENTIA)

Summary of Itinerary (Fig. 4-1)
Scenic drive (clockwise on the Cabot Trail). Coffee stop in Cheticamp (approximate arrival time 9:45 am).

Important note: Most of the following STOPs are in the Cape Breton Highlands National Park. No hammering or sample collecting is allowed.

STOP 6-1: Ca. 550 Ma granite of Aspy terrane (Cheticamp Pluton)
STOP 6-2: Metasedimentary rocks of the Jumping Brook Metamorphic Suite (Aspy terrane) + Scenic Look-off
STOP 6-3: Pleasant Bay Complex + Scenic Look-off
STOP 6-4: Red River Anorthosite Suite – Blair River inlier
STOP 6-5: Aspy Fault Scenic Look-off
STOP 6-6: White Point (ca. 375 Ma White Point pluton of the Black Brook Granitic Suite)
STOP 6-7: Mixed margins - Black Brook Granitic Suite and "Neils Harbour Gneiss", Lakies Head
STOP 6-8: Origin of megacrystic granodiorite, Cameron Brook Granodiorite, Kings Point

<u>STOP Descriptions</u>

STOP 6-1: Cheticamp Pluton at Le Grand Falaise (46.675004°N 60.953092°W)
[Note: It is **not** safe to clamber up the rubble at the base of this cliff.]

As seen from the Grand Falaise Picnic Site, the bottom of the cliff consists mostly of dark green to black, shattered and strongly altered basalt (Devonian Fisset Brook Formation) with a prominent layer of brick-red sedimentary rocks. These red sedimentary rocks are associated with rocks of the Fisset Brook Formation elsewhere, and without them it would have been difficult to identify the basalt as belonging to that formation, as Silurian mafic volcanic rocks also occur in Aspy terrane. The upper part of the cliff is granite of the ca. 550 Ma Cheticamp Pluton (Jamieson et al. 1986). The contact between the Fisset Brook Formation below and the granite above is a thrust fault marked by a zone of intense shattering and alteration. The movement of fluids through the fractured rock deposited veins of calcite and gypsum near the thrust fault. The granite is cut by mafic and felsic dykes; one of the felsic dykes in the cliff yielded an age of 439 ± 7 Ma (Currie et al. 1982) and hence they are probably related to Silurian volcanic rocks elsewhere in the Aspy terrane.

This area is in the northern part of the Cheticamp Pluton, which consists mostly of medium-grained biotite granodiorite gradational to tonalite and monzogranite. The major minerals are quartz, plagioclase, microcline, and biotite, with accessory titanite and opaque minerals. The southern part of the pluton consists of muscovite-biotite monzogranite to granodiorite. An early petrological study (Barr et al. 1986) suggested that these units, which were not observed in contact, are cogenetic and related by crystal fractionation involving mafic minerals, plagioclase, and titanite.

STOP 6-2: Metasedimentary rocks of the Jumping Brook Suite (46.735923°N 60.922690°W)
Many of rocks that comprise the Aspy terrane are seen here in the ornamental wall at the Cap Rouge look-off, which is a safer place to stop and see them than on the ascent up French Mountain! Also, from this location, you can get a good view (weather permitting!) of the western Cape Breton Highlands, including the lowlands underlain by Carboniferous rocks and the highlands composed of metamorphic and granitic rocks of the Aspy terrane.

From this area to the east and uphill, metamorphic grade in the Jumping Brook Metamorphic Suite increases systematically from chlorite through biotite, garnet, staurolite, and sillimanite. Little outcrop occurs at the top of the highlands plateau (elevation ca, 450 m) and the relationship between the Jumping Brook Metamorphic Suite and overall higher grade Pleasant Bay Complex (STOP 6-3) is uncertain (Jamieson et al. 1987, 1989).

STOP 6-3: Pleasant Bay Complex + Scenic Look-off (46.789861°N 60.846676°W)
On a clear day, the descent from MacKenzie Mountain (elevation 335 m) provides scenic views across Aspy terrane to the east, underlain in this area mainly by plutonic rocks ranging in age from ca. 442 Ma (or older) orthogneiss (Pleasant Bay Complex) to ca. 365 Ma megacrystic granite of the Margaree Pluton, but also including ca. 375 Ma granite (Pleasant Bay Granite) and schist of the Jumping Brook Metamorphic Suite. Far down in the valley, the MacKenzie River provides access to the interior of Aspy terrane.

Views far to the north show Andrews Mountain, composed of red Devonian granite (see pillars supporting the benches and display panels for nice examples), and across the Red River mylonite zone north of Andrews Mountain, the rugged shoreline of the Grenvillian Blair River inlier. The Lowland area around Pleasant Bay is underlain by Carboniferous sedimentary rocks. The water to the west is the Gulf of St. Lawrence.

The roadside outcrop opposite the look-off is typical of the Pleasant Bay Complex. It includes paragneiss (foliation trending north and nearly vertical) cut by varied biotite and muscovite-biotite granodiorite and granite. These distinctive medium-grained peraluminous granitoid rocks are characteristic of Aspy terrane, and you will see them again at STOPS 6-6 and 6-7. Dykes of granitic pegmatite and associated aplite are also characteristic components of Aspy terrane and are well displayed in this outcrop.

Probable Lunch Stop: MacIntosh Brook picnic park, Grand Anse River valley.

STOP 6-4: Red River Anorthosite Suite - North Mountain (46.808488°N 60.694025°W)
Gneissic and plutonic rocks of the Blair River inlier form most of the roadside outcrops as we ascend North Mountain (elevation 445 m) but stopping anywhere during the ascent is both dangerous and illegal. In any case, the rocks are a mess - they are all strongly deformed and altered because of proximity to the margins of the inlier, and in fact, no outcrops that really do justice to the Blair River inlier occur along the Cabot Trail. Outcrops in the ditch on the south side of the highway near the summit of North Mountain are the best available.

These ditch outcrops are massive anorthosite of the Red River Anorthosite Suite. Farther to the northeast, this anorthosite body has yielded a U-Pb (zircon) minimum igneous crystallization age of 1095 Ma, and a metamorphic age of ca. 996 Ma (Miller et al. 1996; Miller and Barr 2000). The massive anorthosite in this outcrop is typical of the core of the suite, and elsewhere it grades outward into gabbroic rocks. The anorthosite consists mostly of plagioclase of andesine-labradorite composition.

STOP 6-5: Aspy Fault Look-off - MacGregor Brook (46.811444°N 60.642244°W)
As we began our descent from North Mountain, we cross a few outcrops of mylonitic rocks of the Wilkie Brook shear zone and then quickly are back into Aspy terrane. The roadside section exposes sheared and locally mylonitic metasedimentary rocks of the Ordovician-Silurian Cape North Group, cut by abundant Devonian granitic sheets of the "Andrews Mountain type" (age ca. 375 Ma).

From the look-off you can see the extent of the Carboniferous lowlands of the Aspy Valley. The prominent Aspy Fault follows the steep valley to the northeast toward Newfoundland (where it connects with the Cabot or Long Range Fault) and to the southwest toward Margaree where we crossed it earlier today en route to Margaree. Although it is topographically prominent, the Aspy Fault is a Carboniferous feature within Aspy terrane and has only a few kilometres of displacement.

STOP 6-6: White Point (Black Brook Granitic Suite) (46.883822°N 60.351539°W)
White Point is in the northern, more granitic lobe of the Devonian Black Brook Granitic Suite, a large body of biotite granite and muscovite-biotite granite that yielded an age of 375 +5/-3 Ma (U-Pb, monazite; Dunning et al. 1990). The White Point granite has not been dated. It consists of medium-grained monzogranite, containing plagioclase, quartz, microcline, biotite, and muscovite, with accessory (< 1%) apatite, zircon, monazite, and ilmenite. Myrmekitic intergrowths of quartz and plagioclase are common, and form bulbous shapes at the margin of plagioclase grains adjacent to microcline (Fig. 4-16).

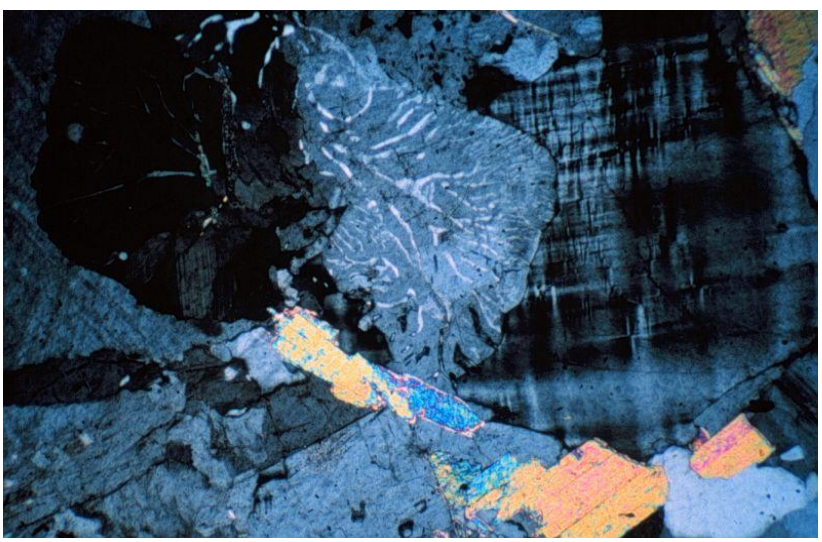

Figure 4-16. Photomicrograph of granite from the Black Brook Plutonic Suite displaying myrmekitic texture. Photograph is with crossed polars. Field of view is 4 mm wide.

If the weather is good, we should walk out to White Point, past the grave of the "Unknown Sailor", for a spectacular view to the north of the Cape North peninsula and the scarp face of the Aspy fault, as well as excellent outcrops of the White Point granite, its many enclaves, and associated pegmatite. The northernmost outcrops are dominantly in the host gneissic rocks, part of the Cheticamp Lake Gneiss, and the foliation trends east-west and dips steeply north; this area is probably at the northern margin of the pluton. If the weather is uncooperative, a good selection of typical rock types is also easily accessible in the breakwater at the wharf.

As noted at STOP 6-3, these peraluminous granites and granodiorites are a major component of the Aspy terrane but are completely absent from Bras d'Or terrane. Examine the gneissic enclaves and screens in order to compare them to those we will see at STOP 6-7.

STOP 6-7: Black Brook Granitic Suite and Neils Harbour Gneiss, Lakies Head (46.734957°N 60.332146°W)
The shoreline outcrops at Lakies Head are a mixture of biotite granite and cross-cutting granitic, aplitic, and pegmatitic dykes, all part of the Black Brook Granitic Suite. Also present are xenolithic blocks composed of biotite-rich pelitic material and megacrystic granodiorite. Collectively these lithologies have been termed the "Neils Harbour Gneiss". The megacrystic orthogneiss yielded a U-Pb (zircon) age of 403 ± 3 Ma, the same as the age of the megacrystic Cameron Brook Granodiorite (402 ± 3 Ma) which we will see at STOP 6-8. It is likely that the megacrystic component of the Neils Harbour Gneiss is the Cameron Brook Granodiorite. However, what is the origin of the biotite-rich xenoliths? Are they remnants of the host rock of both the Black Brook Granitic Suite and the Cameron Brook Granodiorite?

Based mainly on an observed change in the nature of the xenoliths from the type seen here in this outcrop to gneissic lithologies clearly derived from Aspy terrane units (e.g., at STOP 6-6), Yaowanoiyothin and Barr (1991) drew the Aspy/Bras d'Or boundary as a "cryptic suture" through the Black Brook Granitic Suite, and interpreted the granitic suite to be syntectonic, emplaced during movement along the Eastern Highlands shear zone. The terrane boundary then stepped to the south with several branches, as shown in Fig. 4-1. In contrast to the western and eastern margins of the Black Brook Granitic Suite where dykes associated with the pluton are numerous (e.g. STOP 6-7), the southeastern margin is sharp and no dykes derived from the Black Brook Granitic Suite occur in adjacent Bras d'Or terrane units and no xenoliths from those units (with the exception of the Cameron Brook Granodiorite) were recognized in the Black Brook Granitic Suite. These observations suggest that the Bras d'Or terrane units now juxtaposed with the Black Brook Granitic Suite along the Eastern Highlands shear zone were not in close proximity during intrusion of the suite.

STOP 6-8: Cameron Brook Granodiorite, Kings Point (46.685543°N 60.383143°W)
The Cameron Brook Granodiorite was intruded into the boundary between the Bras d'Or and Aspy terranes. It intruded rocks assigned to the McMillan Flowage Formation on the southwest, and the Ingonish River tonalite on the south. A branch of the Eastern Highland shear zone separates the pluton from the Black Brook Granitic Suite to the north. As noted at STOP 6-7, the age of the Cameron Brook Granodiorite is 402 ± 3 Ma (U-Pb zircon; Dunning et al. 1990). The granodiorite consists of coarse-grained variably megacrystic granodiorite gradational to monzogranite. Megacrysts include both plagioclase and perthitic alkali feldspar, and some display plagioclase-mantled K-feldspar textures (Fig. 4-17). Megacryst density varies from clustered to sparse – is this a result of flow? Biotite is the main mafic mineral, with subordinate hornblende. Accessory apatite, titanite, zircon, and magnetite are present. Weak to strong foliation ofuncertain origin is defined by alignment of megacrysts and biotite. Although similar to the megacrystic Margaree Granite (Fig. 4-10), the Cameron Brook Granodiorite is at least 25 million years older. The tectonic setting for this unique pluton remains enigmatic.

Figure 4-17. Photograph of the megacrystic Cameron Brook Granodiorite at Kings Point (STOP 6-8) showing plagioclase-mantled K-feldspar crystals (photograph courtesy of Barrie Clarke).

Return to the Gaelic College and St. Anns for dinner and the Eurogranite Mini-Ceilidh.

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