CHAPTER 16

MIDDLE MESOPROTEROZOIC (M_2 1350–1200 Ma)

Middle Mesoproterozoic rocks in eastern Labrador represent minor crustal additions, but are important in providing information regarding this relatively quiescent period. Rocks addressed in this chapter are: i) the Fox Harbour bimodal volcanic belt, ii) the Upper North River granitoid intrusion, iii) the Mealy dykes, iv) other mafic intrusive activity, and v) supracrustal rocks at Battle Island.

16.1 FOX HARBOUR BIMODAL VOLCANIC BELTS

16.1.1 DISCOVERY

The Fox Harbour bimodal volcanic rocks are located in the Fox Harbour area in southeast Labrador and trend parallel to the regional structure, extending inland from the coast for a strike length of over 55 km (Figure 16.1). They were discovered by Search Minerals Inc., which has been investigating them for their REE economic potential since 2009, resulting in the locating of numerous REE prospects, and a 9.2-million-tonne indicated resource at the Foxtrot deposit. The belt was not identified during reconnaissance geological mapping by the author (Gower *et al.*, 1987, 1988), although a few occurrences of supracrustal rocks that are now recognized as belonging to the belts were recorded during traversing in the area. The discovery of the volcanic belts was partly the result of unexpected consequences, inasmuch as the author suggested to prospectors (Alterra Resources Inc.) that the region might be worth searching for U mineralization because of its abundance of very large pegmatites, including graphic-textured varieties. Also in mind was an earlier investigation by Meyer and Dean (1988) of a Pb–Cd–W–Cu lake-sediment anomaly 2 km east-southeast of St. Lewis. They reported that a mylonitic rock had anomalous radioactivity (800–1000 counts/second) and a geochemical assay of a sample returned a value of 3.4% Zr, plus anomalous values for Ce, Hf, La, Rb, Th, Sn, Yb, and U. Later prospecting for U revealed several occurrences of anomalous radioactivity, from which the subsequent involvement by Search Minerals Inc. led to the recognition of the Fox Harbour volcanic rocks. The information presented here is summarized from Haley (2014) and Miller

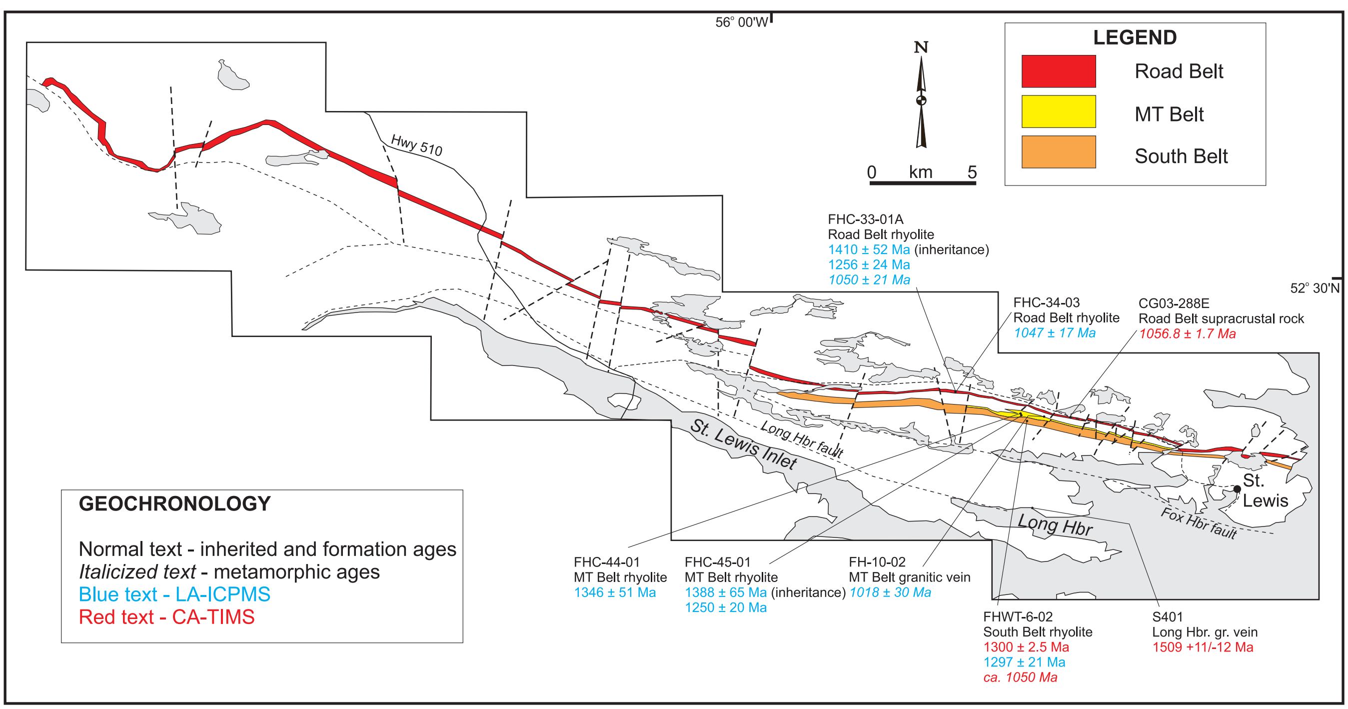


Figure 16.1. *Distribution of supracrustal rocks of the Road Belt, MT Belt and South Belt in the Fox Harbour area identified by Search Minerals Inc. (Miller, 2015).*

(2015), but a huge resource of untapped unpublished (much now non-confidential) assessment report material resides in the Geological Survey of Newfoundland and Labrador archives.

16.1.2 DESCRIPTION

The Fox Harbour volcanic rocks are situated within the southern part of the Lake Melville terrane (as defined by the author), but are in a separate structural entity, termed the Fox Harbour domain based on detailed mapping by Search Minerals Inc., (Miller, 2015). Rock types include mildly peralkaline rhyolite (comendite and pantellerite), mafic volcanic rocks (subalkaline tholeiitic basalt), quartzite, and volcaniclastic/metasedimentary units. The felsic volcanic rocks may include high-level, subvolcanic intrusions. The rocks are extremely deformed, and no textural features remain that might assist protolith identification of the felsic volcanic rocks, which is based on lithogeochemical signatures (Haley, 2014; Miller, 2015). The supracrustal rocks have been divided (from south to north) into three belts named the South Belt, Magnetite (MT) Belt, and Road Belt. These have been interpreted by Haley (2014) as the refolded repetition of a single supracrustal package.

The South Belt has the thickest package of rhyolitic and basaltic rocks, ranging in thickness from 100 to 250 m. Rhyolite (50 to 100 m thick) on the south side is succeeded to the north by a basaltic unit, also 50 to 100 m thick. The basaltic unit is characterized by epidote-rich pods. A 2–5-m-quartzite unit separates the rhyolite and basalt for a 10-km-long segment of their mutual contact. A >1-m-wide quartzite and a fine-grained felsic unit (either a flow or minor intrusion) are present within the basalt layer.

The Magnetite Belt is characterized by interlayered rhyolitic and basaltic units that are variably boudinaged and migmatized. The Magnetite Belt hosts major REE mineralization (Foxtrot deposit) and has been the subject of very detailed study. This has resulted in the identification of seven rhyolite units, although it is noted that they may represent folded/thrust repetitions. The mineralized rhyolite units carry aegerine–augite and Na-rich amphibole, suggesting peralkaline character. Miller (2015) emphasizes that mineralized pantellerite, such as that at the Foxtrot deposit, is very rare and must represent extreme differentiation of peralkaline or near-peralkaline magma in high-level crustal magma chambers close to the source vent. As for the South Belt, the basaltic units have epidote pods, which Haley (2014) considers might be replacing pillows or form alteration pipes (where seen elsewhere by the author, he has favoured interpretation as interpillow material). A thin quartzite unit (0.2–0.5 m thick) is present at the southern margin of the belt, and narrow garnetiferous volcaniclastic/metasedimentary units are found between individual basaltic layers. Concordant boudinaged pegmatites are commonly amazonite-bearing (green, Pb-bearing K-feldspar[2]).

The Road Belt, which is interpreted by Haley (2014) as the other-fold-limb correlative of the Magnetite Belt, shares the same rock types as the Magnetite Belt, but in a much more strongly deformed state. It is the longest of the three belts, having been traced the full 64 km length of the Fox Harbour domain, whereas the other two have been traced for the eastern 30 km.

16.1.3 GEOCHRONOLOGY

Haley (2014) reported results for six samples investigated by LA–ICPMS, one of which was also analyzed by the CA–TIMS method. One sample is from the South Belt; three are from the Magnetite Belt, and two from the Road Belt. An additional sample of the author's, from the Road Belt, has also been independently analyzed by S. Kamo (Jack Satterley laboratory; unpublished result to GNLS).

From the South Belt, a rhyolite (sample FHWT-6-02) gave a concordant age of 1297 ± 21 Ma, based on 11 of 12 zircon grains analyzed by LA–ICPMS. In addition, this sample was analyzed by CA–TIMS and gave an age of 1300 ± 2.5 Ma, based on the weighted average of 5 (of 6) concordant zircon fractions. One fraction showed Pb loss along a discordia line to *ca.* 1050 Ma.

From the Magnetite Belt, one rhyolite sample (FHC-44-01) provided an upper intercept age of 1346 ± 51 Ma, based on 8 discordant zircon grains. A second rhyolite sample (FHC-45-01) gave an age of 1250 ± 20 Ma, based on 15 zircon grains (of 18; 17 on concordia). An earlier date of 1388 ± 65 Ma was interpreted to be inheritance. A third sample (FH-10-02), a granitic vein within basalt, yielded a concordant age of 1018 ± 30 Ma, based on 3 zircon grains. All results from the Magnetite Belt were obtained by LA-ICPMS.

From the Road Belt, one rhyolite sample (FHC-33-01A) proved complicated. Of 12 zircon grains analyzed, 5 define a concordant cluster indicating a date of 1256 ± 24 Ma and 5 a cluster indicating a date of 1050 ± 21 Ma. An outlier grain giving a date of 1410 ± 52 Ma was considered to represent inheritance. A second rhyolite sample (FHC-34-03) was more straightforward, in that 10 zircon grains define a concordia age of 1047 ± 17 Ma. Both samples were investigated by the LA–ICPMS method. The author's sample (CG03-288E; Plate 16.1A; Photomicrograph 16.1A) from

[2] Note that amazonite is also known from other localities in southeast Labrador (Gower, 2010c), and has been reported from the Makkovik area (Wheeler, 1935; King, 1963). *See* also Section 17.3.4.1.

the Road Belt analyzed using CA–TIMS yielded a concordant age of 1056.8 ± 1.7 Ma, based on three zircon analyses.

One other sample in the area was dated by Scott *et al.* (1993) from Long Harbour, and yielded an age of 1509 +11/–12 Ma. Details are given in Section 13.2.4.1.

As concluded by Haley (2014), the 1300 ± 2.5 Ma age is taken as defining the time of formation of the Fox Harbour volcanic package. Departures from this result are the product of Grenvillian metamorphic disturbance, the timing of which is most precisely defined by the 1056.8 ± 1.7 Ma age, but supported by three other *ca.* 1050 Ma dates.

The reader is reminded of syenitic and alkali-feldspar rocks farther northwest in the Lake Melville terrane (Section 10.3.2.4) that are regionally on strike with the Fox Harbour volcanic rocks. In this report they have been included with Labradorian granitoid gneiss, but, as was mentioned then and iterated here, there is no proof that such is the case and they could be correlative with the Fox Harbour volcanic rocks. They do provide a spatial link between the Fox Harbour volcanic rocks and the coeval Upper North River granitoid pluton, which is addressed in the next section.

16.2 UPPER NORTH RIVER GRANITOID INTRUSION (M_2gr, M_2yq)

The Upper North River intrusion (the name first used by Schärer *et al.*, 1986) is situated in the northern part of the Lake Melville terrane (Figure 16.2). A much smaller granitoid body (underlying *ca.* 12 km^2) was distinguished by Eade (1962) within the subsequently defined boundaries of the intrusion (inferred to underlie *ca.* 340 km^2). Eade assigned the rock to his Unit 4 (granite to granodiorite, with some granitic gneiss and minor syenite), which is now known to include a wide range of granitoid rocks of various ages. The Upper North River body was first delineated to approximately its present extent during 1:100 000 mapping by Gower *et al.* (1981, 1982b). No further field investigations have been carried out since that mapping and the indicated boundary of the pluton has remained essentially unchanged since the release of the preliminary 1:100 000-scale map for the region. It should be kept in mind that the

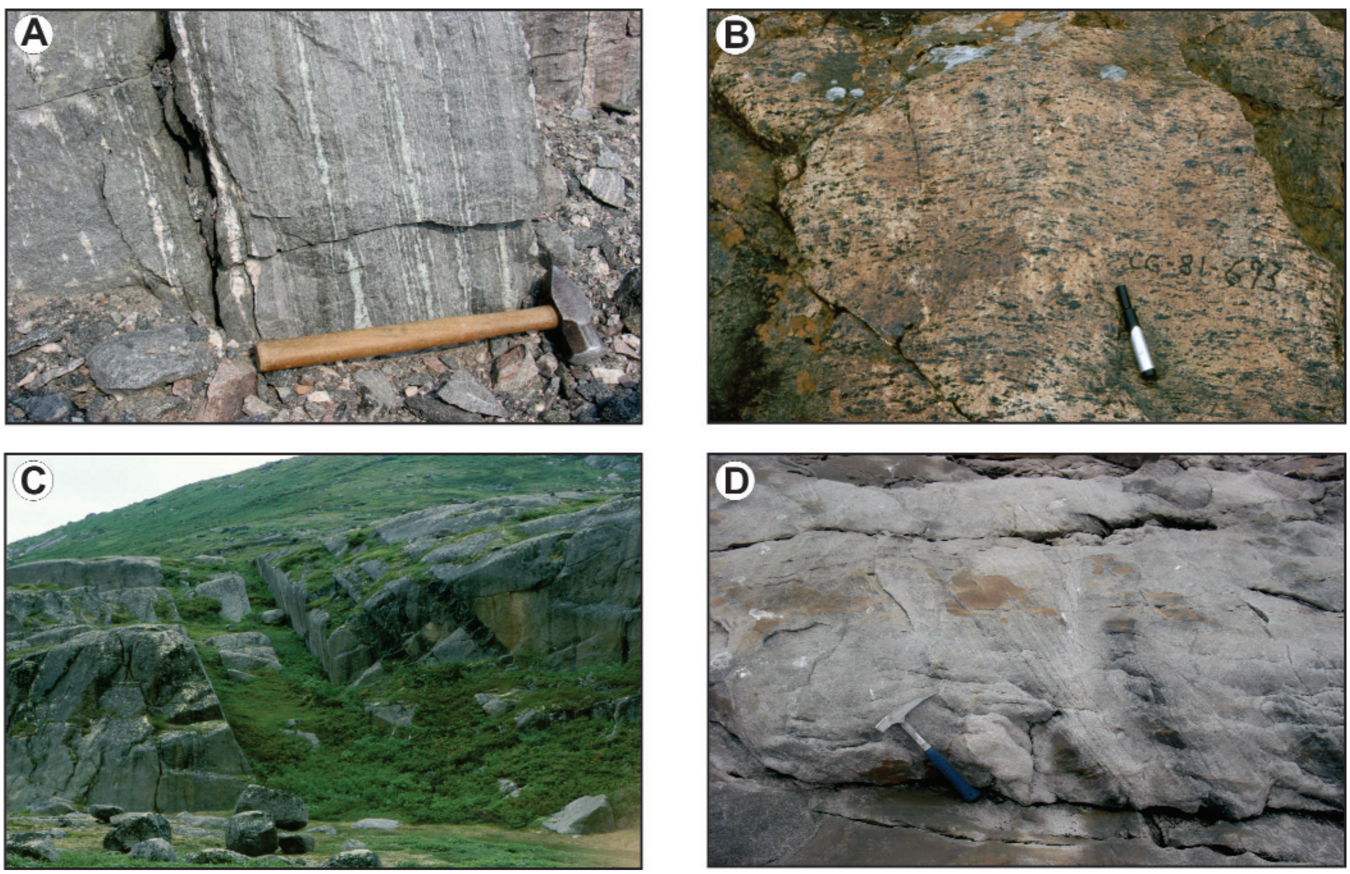
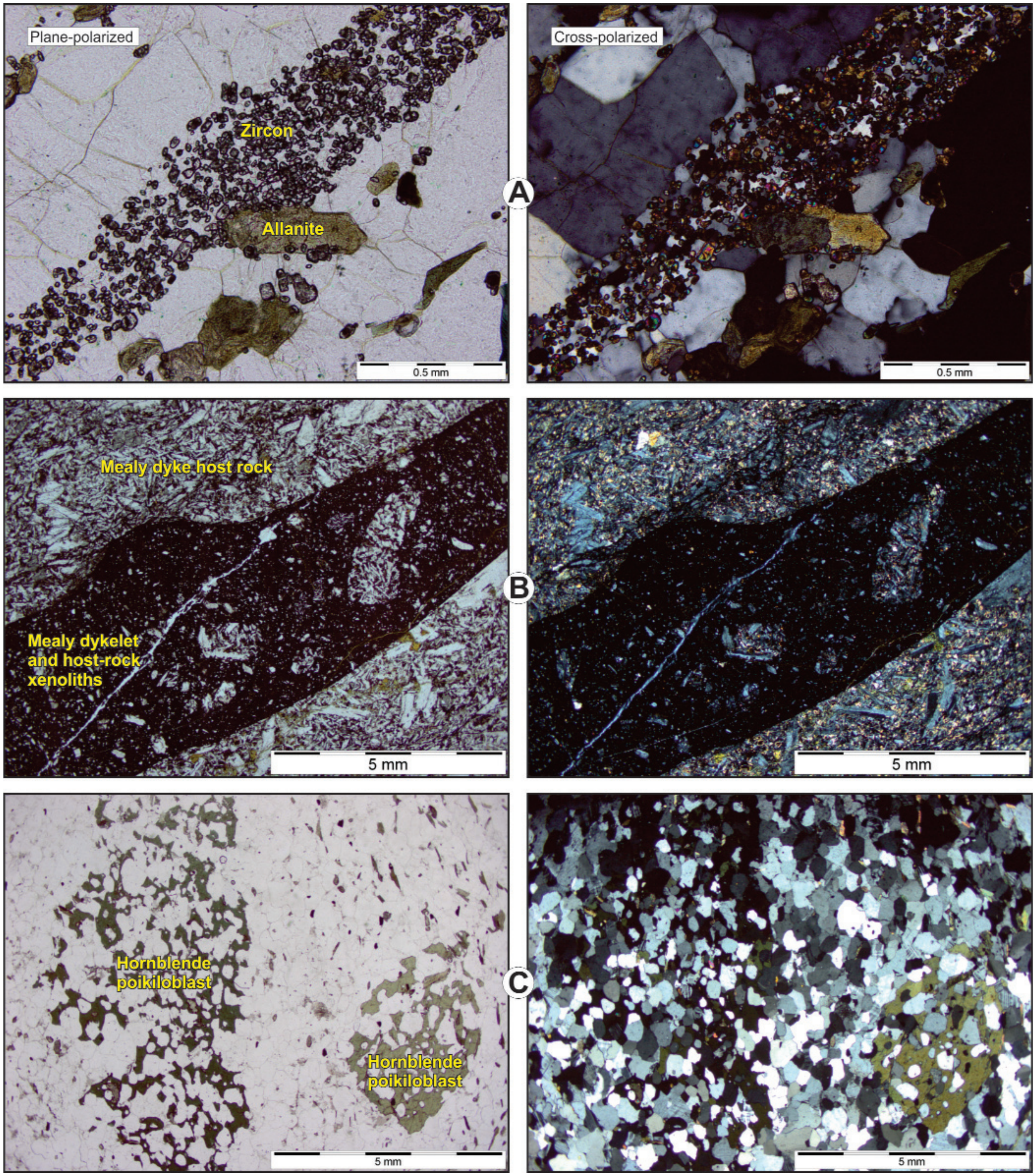


Plate 16.1. *Examples of 1300–1200 Ma Mesoproterozoic units (Fox Harbour volcanic rocks; Upper North River granitoid intrusion; Mealy dykes; and Battle Island supracrustal rocks). A. Fox Harbour supracrustal rock (Road Belt) (CG03-288E), B. Upper North River syenite CG81-693), C. Mealy dykes, north-dipping. At least five dykes present, two of which are in gullies (GF81-167), D. Battle Island crossbedded quartzite (CG07-156).*



Photomicrograph 16.1. *Examples of Middle Mesoproterozoic (1350–1200 Ma) units. A. Hundreds of tiny zircons concentrated in layer. Fox Harbour volcanic belt (CG03-288E). Metamorphic age from this site 1056.8 ± 1.7 Ma, B. Mealy dyke and xenolith-bearing Mealy dykelet within it (VN95-110B), C. Hornblende poikiloblasts in Battle Island psammite (CG07-130C).*

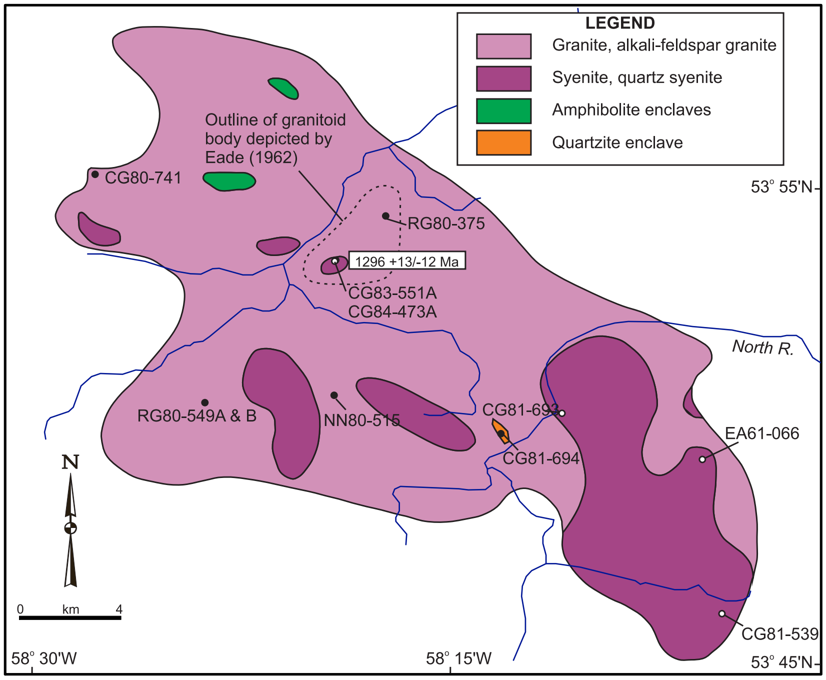


Figure 16.2. *Upper North River granitoid intrusion.*

mapping was done during the early stages of the Eastern Labrador Grenville project when familiarity with the rocks was still being acquired, so it is possible that unrelated rocks have been included. The few post-mapping refinements that have been made are based on more detailed examination of samples collected at that time. The most significant modifications have established that most of the rock present is granite and quartz syenite. Depiction of their separate distributions is shown on the 1:100 000-scale map of Gower (2010a; English River map region).

A sample of quartz syenite from near the centre of the body (CG83-551A; also within Eade's originally defined unit) was determined to have a U–Pb age of 1296 +13/–12 Ma by Schärer *et al.* (1986), based on three near-concordant zircons.

Gower *et al.* (1981) described the body as a pink- to buff-weathering, coarse grained, moderately to strongly foliated, homogeneous granitoid rock containing hornblende as the dominant mafic mineral, but also having biotite and garnet. They noted a dearth of mafic dykes compared to the surrounding units and suggested that it might be younger than them. The description of the intrusion by Gower *et al.* (1982b) is similar, although the range of rock types present was expanded to include quartz monzonite, granite and quartz syenite. A review of field notes provides little further information, except to note the sporadic presence of diabasic and amphibolitic mafic dykes (*cf.* next section), pegmatite and a possible quartzite enclave (CG81-694). A syenitic variant of the intrusion is shown in Plate 16.1B.

Eight thin sections are available, five of which are termed quartz syenite (CG81-539, CG81-693, CG83-551A, EA61-066, NN80-515), two are granite (CG80-741, RG80-549A), and one is alkali-feldspar granite (RG80-549B). All the rocks contain relict igneous and/or metamorphic anhedral plagioclase, K-feldspar and quartz. K-feldspar is present as both microcline and perthite. Myrmekite is also typical. Dark-orange-brown or red-brown biotite is seen in all thin sections, and dark-green (sodic) hornblende is present in all except RG80-549A and RG80-549B. Pale-green metamorphic clinopyroxene is common in CG81-539; relict igneous(?) weakly pleochroic orthopyroxene in EA61-066; and garnet in both of them. Other minerals are igneous and metamorphic opaque oxide, apatite, allanite and zircon (typically fairly large, and zoned/rimmed in part). Secondary minerals are white mica, chlorite, and, in RG80-549A, minor epidote. Thin section EA61-066 is from Eade's collection, although the sample does not come from his originally mapped 12-km^2 body, but *ca.* 16 km to the southeast.

Whole-rock geochemical data are available for 4 samples – two of granite (RG80-549A, RG80-375), and two of quartz syenite (CG81-693, CG83-551). Not surprisingly, the two rock types are each compositionally distinct, but the two samples in each group are similar enough to each other to accept they are related.

16.3 MEALY DYKES (M_2d)

16.3.1 MEALY DYKES IN MEALY MOUNTAINS REGION

16.3.1.1 Introduction

The earliest mention (known to the author) of diabase dykes in the Mealy Mountains region was by Scharon (1952), who merely made passing reference to them in a report focused on mineral exploration. Samples of the dykes were collected, however, and K–Ar age determinations later obtained on them by Gittins (1972; *see* next section). Further observation of diabase dykes intruding anorthosite in the Mealy Mountains was made by Eade (1962) and the first (sketch) map to show their distribution was by Emslie (1976). Emslie also briefly discussed the significance of the dykes, especially in terms of comparing their trend with that of the Harp dykes about 150 km to the north, and speculating that both swarms might be similar in age (in the absence of adequate geochronological control at the time) and related to a rifting event that also included the Seal Lake Group.

The label 'Mealy dykes' was probably never formally introduced. The dykes are mentioned without name by Gower *et al.* (1980, 1981) and Erdmer (1984). The first published appearance of the name was by Park and Emslie

(1983; paleomagnetic study), followed by Emslie *et al.* (1984; complementary petrological investigation). Brief reference to the dykes and limited Nd isotopic data were given by Ashwal *et al.* (1986); an assessment of the significance of magnetic fabric was carried out by Park *et al.* (1988); whole-rock geochemical comparison with other Mesoproterozoic mafic dyke suites in Labrador was made by Gower *et al.* (1990a); and further comparison with the same suites, especially using isotopic data, was made by Emslie *et al.* (1997).

16.3.1.2 Geochronology

The most precise (and reliable) emplacement age available for the Mealy dykes is that of Hamilton and Emslie (1997), who reported a baddeleyite date of 1250 ± 2 Ma (Figure 16.3). Several age determinations preceded that work, however. The earliest results were K–Ar whole-rock dates of 1133 ± 50 Ma and 1088 ± 48 Ma (recalculated) from the Eskimo Paps area by Gittins (1972). An additional K–Ar whole-rock date of 1222 ± 101 Ma from the same area was reported by Fahrig and Loveridge (1981) and K–Ar biotite ages of 955 ± 27 Ma and 964 ± 21 Ma were provided by Emslie *et al.* (1984). An Rb–Sr whole-rock errorchron date of 1380 ± 54 Ma was also obtained by Emslie *et al.* (1984). In addition, Reynolds (1989) obtained Ar–Ar biotite ages of *ca.* 980, *ca.* 1011, *ca.* 1043 and *ca.* 1045 Ma (the latter three each being the average of two separate determinations) and Ar–Ar hornblende ages of *ca.* 1178 Ma (average of two results), *ca.* 1215, *ca.* 1227, *ca.* 1230 and *ca.* 1249 Ma (average of two results). These data include both total-gas and plateau ages. Reynolds's interpretation was that Grenvillian metamorphic effects were sufficient to cause Ar loss from biotite but not amphibole, so the hornblende dates were likely close to the time of emplacement – the latter conclusion subsequently endorsed by the baddeleyite age.

16.3.1.3 Description

To a large extent, the distribution of the Mealy dykes shown in Figure 16.3 relies on information extracted from Emslie's field notes (1975 and 1995). Mapping of the dykes by the author was mostly confined to southern and eastern areas (*cf.* Figure 11.10). This account draws on information from both sources. The Mealy dykes are most abundant in the northeast part of the Mealy Mountains intrusive suite (MMIS), especially in the anorthositic, leuconoritic and leucotroctolitic Etagualet and Kenemich massifs. The dykes are well exposed on the shores of Lake Melville. The dykes also occur in the enveloping monzogranite and in the country-rock pelitic gneiss to the east of the MMIS. The concentration of dykes correlates with the most barren, best-exposed areas, and their apparent distribution may be an artifact of outcrop quality. Even where their ground presence has been established, the dykes cannot be extrapolated reliably along strike using aerial photographs. As noted by Emslie *et al.* (1984), the course of large dykes may be marked by trenches due to plucking of the jointed dyke rocks during glaciation, leaving patches of fine-grained diabase on the walls of the trench as the only evidence of a dyke's former presence. Most of the dykes are too small to have an expression on regional aeromagnetic maps, although east-northeast magnetic trends in places may be an expression of the larger dykes. No high-resolution magnetic data are available for the region in which the Mealy dykes occur. There are obvious east-northeast-trending photolineaments, but it is unclear whether these correlate with Mealy dykes or to faults associated with the Lake Melville rift system (or both).

The dykes tend to occur in closely spaced parallel intrusions (Plate 16.1C), separated by areas where they are sparse or absent. Individual intrusions range in width from a few centimetres to tens of metres (up to 70 m wide reported by Emslie *et al.*, 1984) with concomitant increase in grain size in centres of the thicker dykes. Overall, the trend of the dyke swarm is slightly arcuate, convex-to-the-north, such that the westernmost dykes trend northeast and the easternmost dykes almost due east (Figure 16.3). If thought of as being on the circumference of a circle, the circle would have a radius of about 100 km, centred to the southeast. A stereographic plot illustrates that the predominant dip direction of the dykes is to the north (Figure 16.4; contour peak at 336.9°/75.6°). Assuming that the dykes were emplaced vertically, an explanation for the northwesterly dip of the dykes is that they were tilted backward during Grenvillian thrusting along shallow south-dipping shear surfaces. A few dykes have anomalous strikes (EC75-057, EC75-166, ECD75-069, ECD75-260) at high angles to the prevailing trend, thus appearing to be radial to the circle. At data station ECD75-118 a 15-m-wide dyke trending at 082° is crosscut by a 10-cm-wide dyke trending at 076°. One flat-lying intrusion (CG95-260) is clearly anomalous, but there is little doubt from textural evidence that it belongs to the Mealy dyke swarm. The dyke is about 2 m thick and was observed in two places about 700 m apart, dipping 11° northwest at one and 30° southeast at the other.

The Mealy dykes are grey-, black- or brown-weathering, massive, homogeneous, and have a well-developed subophitic texture. They are mostly medium grained (rarely coarse grained), but grade into fine-grained or aphanitic rocks at chilled margins and in narrow dykes. Some mafic dykes of uncertain affinity initially were subsequently confidently assigned to the Mealy dyke swarm on the basis of their appearance after staining. Despite the rocks showing a wide range in grain size, they are all (except the grain-size extremes) characterized by a distinct 'felted' texture governed by randomly arranged, acicular, quenched plagioclase

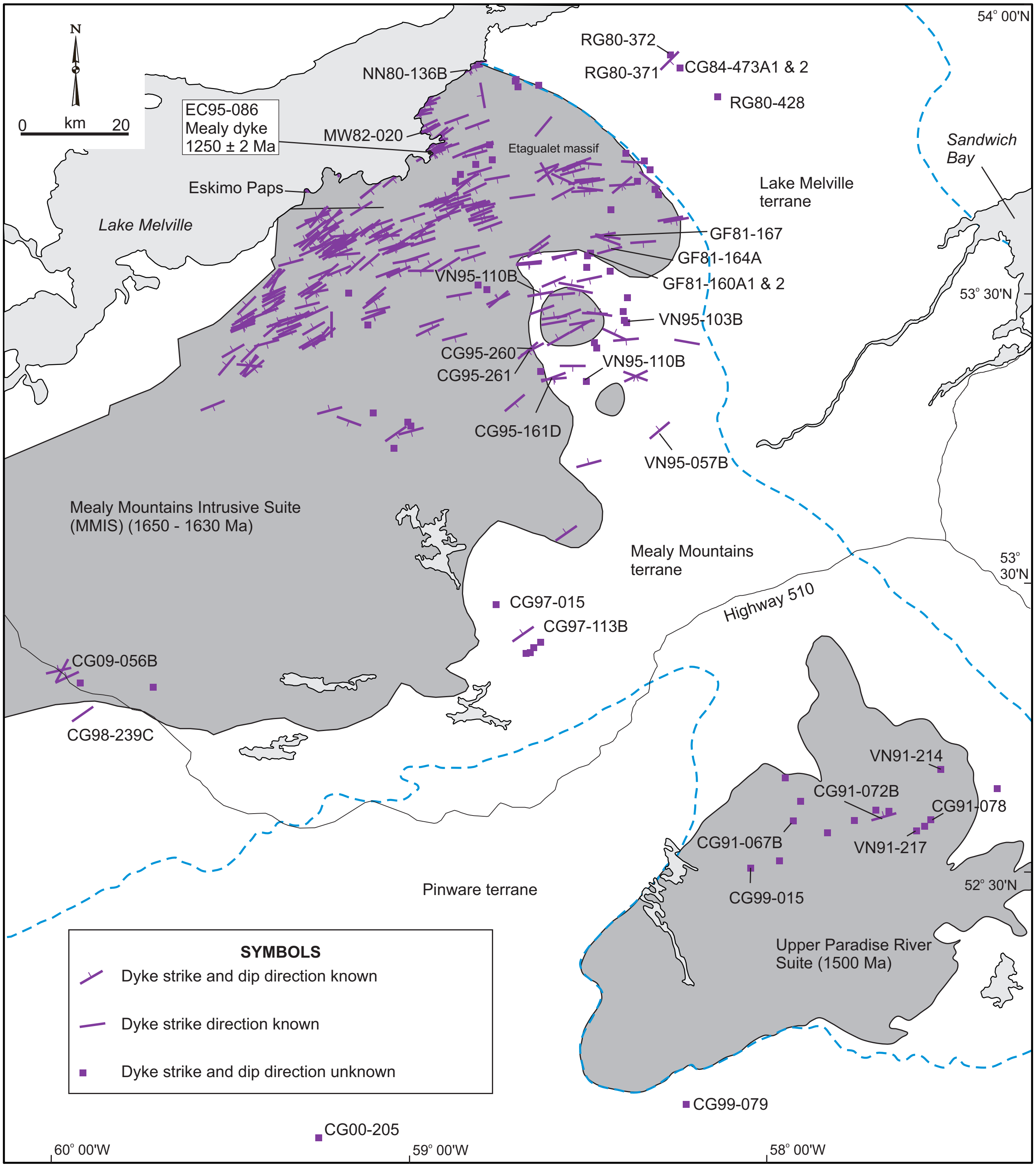


Figure 16.3. *Distribution of Mealy dykes, Upper Paradise River dykes(?) and other potentially correlative minor mafic intrusions.*

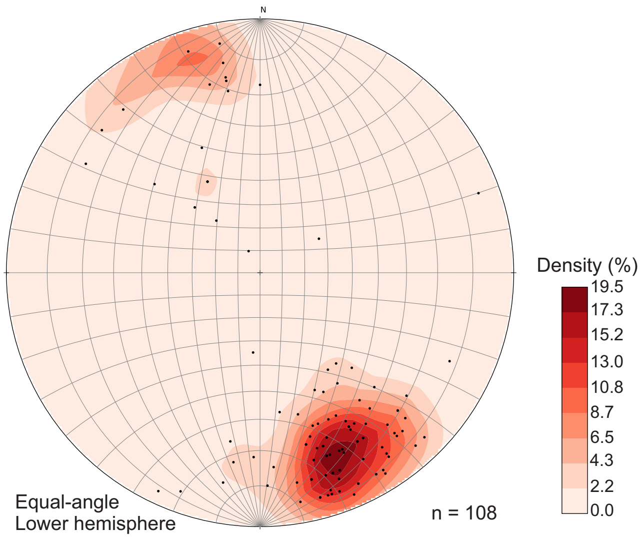


Figure 16.4. *Stereographic plot for the Mealy dykes. Maximum: strike 247°/76°N.*

grains that were clearly early crystallizing. This point is important in assigning mafic dykes in the country-rock gneiss east of the MMIS to the Mealy dyke swarm, especially as the dykes show more metamorphic modification and have more diverse orientations than their western counterparts. In some cases, poor exposure prevents dyke trend being known at all. The more varied strike is attributed to Grenvillian deformation, to which the host gneiss was more responsive than was the MMIS. Note that the Mealy dykes are confined to the Mealy Mountains terrane, but this may be due to preferential preservation. The Mealy Mountains region escaped severe Grenvillian effects (*e.g.*, Emslie *et al.*, 1984), but coeval dykes could have been have been metamorphosed beyond recognition in other terranes, especially the adjacent Lake Melville terrane, which experienced much more severe Grenvillian deformation and metamorphism than the Mealy Mountains terrane.

16.3.1.4 Petrography and Mineral Chemistry

The petrographic characteristics of the Mealy dykes given here are a composite summary based on detailed description and mineral analyses provided by Emslie *et al.* (1984), plus information from thin sections obtained during 1:100 000 mapping (CG95-161D, CG95-261, CG97-015, CG97-113B, CG98-239C, GF81-160A.1, GF81-160A.2, GF81-164A, MW82-020, NN80-136B, NN80-139B, VN95-057B, VN95-103B, VN95-110B (Photomicrograph 16.1b), VN95-143). Two additional thin sections (CG09-003B, CG09-056B), from samples collected along Highway 510 west of longitude 60oW, are also available. The location of CG09-003B, from a quarry immediately south of the causeway across the Churchill River, can be located on Figure 11.9, being the same site as that from which the host rock granite gave a 1641 ± 4 Ma age.

The dykes are typically non-porphyritic, but finer grained dykes may contain microphenocrysts of plagioclase, and, progressively more rarely, clinopyroxene and olivine. Subophitic textures are characteristic. Olivine is anhedral to subhedral, colourless, generally fresh and contains opaque inclusions. Orthopyroxene forms partial rims on olivine. Subophitic to ophitic clinopyroxene is colourless to pale or purplish-brown. Emslie *et al.* (1984) recorded that it is characteristically heavily charged with exsolved magnetite, and that lamellae and blebs of exsolved Ca-poor pyroxene are also common. Plagioclase forms subhedral to euhedral, well-twinned, zoned primary laths. Most show quenched, skeletal habit. Grains are typically heavily clouded with minute opaque inclusions (suggested, but not confirmed, to be hematite by Emslie *et al.*, 1984). Slightly wavy grain margins hint at incipient recrystallization in sample CG98-239C. Red-brown biotite forms veneers around opaque oxides and occurs as an interstitial phase. Apatite forms subhedral to euhedral crystals. Emslie *et al.* (1984) also note narrow rims of brown amphibole on clinopyroxene and very fine-grained interstitial intergrowths of clinopyroxene, plagioclase and possibly K-feldspar. The latter are suggested as representing devitrification texture. Late crystallizing K-feldspar in interstices was also noted by the author (CG97-015 and CG98-239C). Emslie *et al.* (1984) also provide detailed information on the opaque minerals, reporting the presence of separate grains of homogeneous ilmenite and magnetite–ilmenite intergrowths, also noting associated dark-green spinel, pyrrhotite and minor chalcopyrite.

Microprobe mineral data from six Mealy dyke samples were reported by Emslie *et al.* (1984; EC75-002, EC75-016, EC75-029, EC75-037, EC75-040, EC75-176A); these are paleomagnetic sampling sites, which are different from 'regular' Emslie field stations. Refer to Emslie *et al.* (1984) for locations and to, 'Table *EmsliePaleomag*' in the author's database for cross referencing between paleomagnetic and 'regular' Emslie field stations. Mineral data are condensed as follows: olivine $Fo_{47–28}$; orthopyroxene rims on olivine $Ca_{\leq 1.1}$, $Mg_{59–49}$, $Fe_{41–52}$; clinopyroxene $Ca_{44–21}$, $Mg_{43–35}$, $Fe_{21–40}$; plagioclase cores $An_{63–47}$; plagioclase rims $An_{45–30}$; biotite Ca_2, Mg_{45}, Fe_{53}. In addition, sample EC75-040 has groundmass orthopyroxene Ca_1, Mg_{51}, Fe_{48}. Microprobe data were also obtained to allow estimation of bulk compositions of original magnetite–ulvospinel solid solutions and thence to obtain estimates of T and fO_2 at near solidus conditions (T = 980 ± 55°C; $\log_{10} fO_2 = –12.8 \pm 0.7$).

16.3.1.5 Whole-rock Chemistry

Whole-rock geochemical variation diagrams for the Mealy dykes are shown in Figures 16.5 and 16.6. Emslie *et al.* (1984) published whole-rock major- and trace-element data for 15 of 29 samples that they analyzed. Pulverized splits of some of their samples were provided by Emslie to the author, who obtained REE data for Emslie *et al.*'s 15 published samples. The REE data were included in a chondrite-normalized plot by Gower *et al.* (1990a). The eastern Labrador database contains the numerical data, plus whole-rock major- and trace-element data for 9 of the author's own samples, for which extra trace-element data (including REE) are reported. The author's data plot in the same cluster as that for Emslie *et al.*'s results and there is no reason to doubt that all samples are Mealy dykes. The only anomalous data

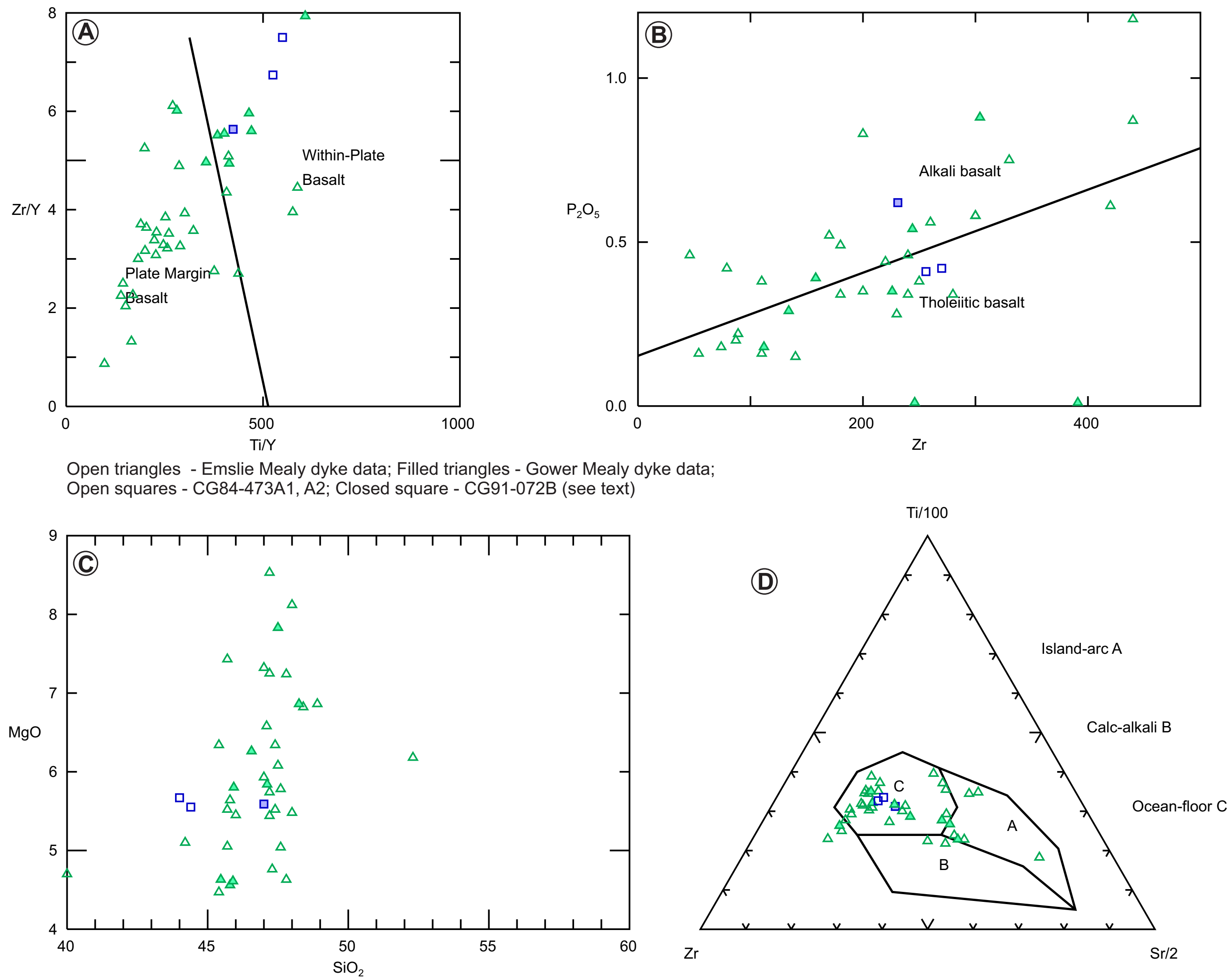


Figure 16.5. *Lithogeochemical characteristics of the Mealy dykes – part 1.*

are from Emslie *et al.*'s suite of rocks in that one of their samples (EC75-044) has abnormally low SiO_2 at 40.0% and one sample (EC74-176A) has abnormally high SiO_2 at 52.3%. The low SiO_2 sample has very high P_2O_5 at 4.76%. This correlates with high CaO and high light REE. The chemistry can be explained by unusually abundant apatite, but the question 'why?' still remains. Perhaps it is not a Mealy dyke. Field notes describe a gabbroic body 50 m wide and having 070° trend. The high SiO_2 sample comes from an area much farther south. The sampled dyke intrudes diorite–monzonite that is associated with pelitic gneiss, so the rock may have been contaminated by its host rock.

Emslie *et al.* (1984) interpreted the whole-rock variations as being consistent with a fairly Fe-rich tholeiitic magma having undergone some fractionation of olivine and calcic plagioclase, either during crystallization of the thicker dykes, or during dyke emplacement as more fractionated magma was supplied from a deeper source. Petrotectonic variation diagrams plotted by the author inconsistently classify the rocks, but allow the designation as tholeiitic, subalkaline transitional to alkaline, and having a within-plate or plate-margin setting (Figures 16.5 and 16.6). In their comparison with other Mesoproterozoic mafic dyke suites in Labrador (namely, the Michael and Shabogamo gabbros, and the Harp and Nain dykes), Gower *et al.* (1990a) considered that the Mealy dykes do not compare closely with them (being enriched in K_2O, P_2O_5, TiO_2 and Zr, and depleted in MgO) and concluded that a likely cause was emplacement into geologically distinct regions that existed long before the various tectonic terranes achieved their final configuration during Grenvillian orogenesis.

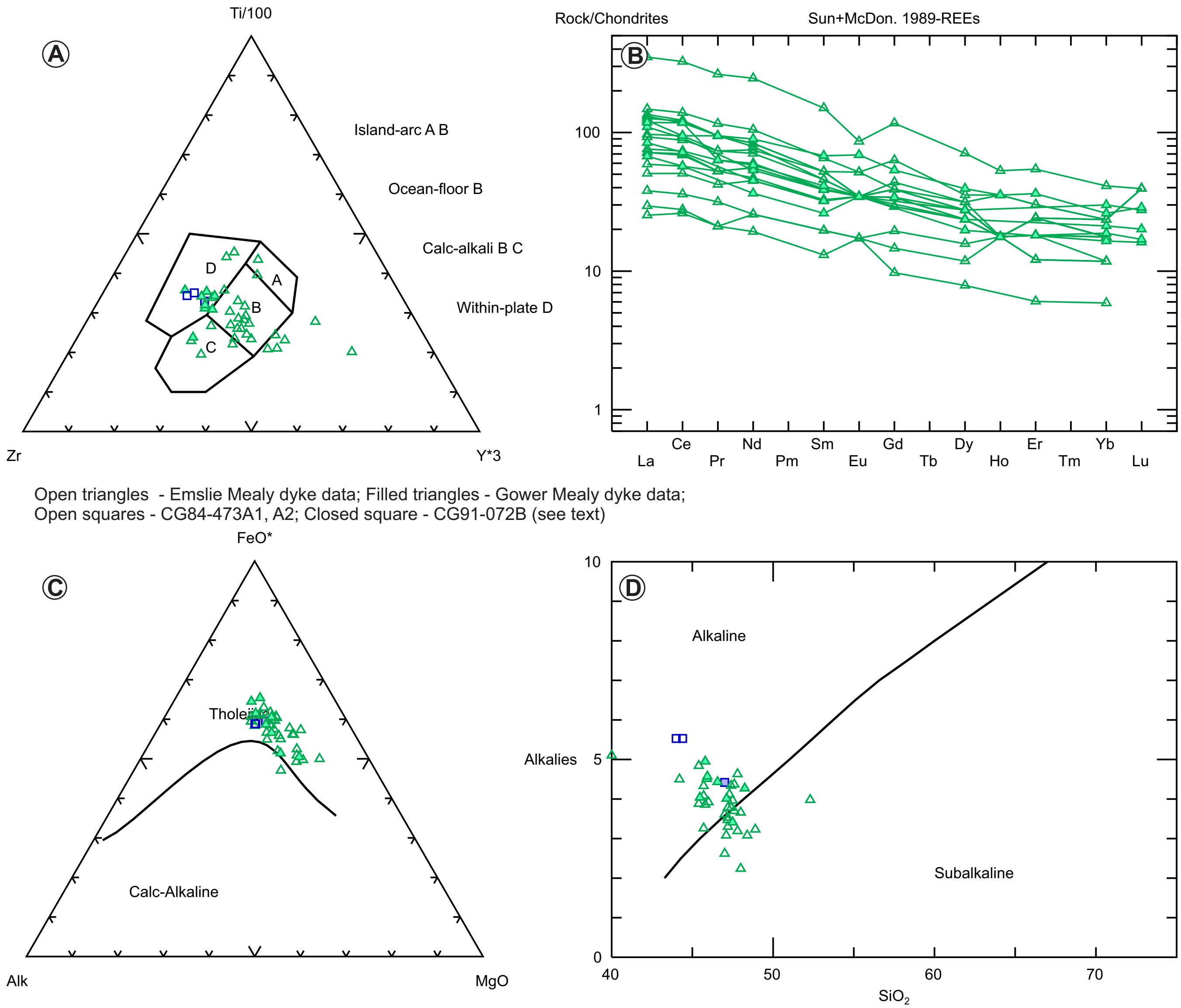


Figure 16.6. *Lithogeochemical characteristics of the Mealy dykes – part 2.*

Additional contrasts with respect to the other suites were made by Emslie *et al.* (1997). For their contribution, precise ages were then also available for all the mafic suites, establishing, in particular, that the 1250 Ma Mealy dykes are significantly younger than the *ca.* 1460–1425 Ma Michael and Shabogamo gabbros. Using averaged data for the various suites, and displaying the results in trace-element plots for REE, extended elements, TiO_2 *vs.* P_2O_5, and TiO_2 *vs.* Nb, Emslie *et al.* (1997) pointed out that the Mealy dykes are the most REE enriched, especially in LREE, compared to the other suites, and that they tend to be relatively richer in other incompatible elements. A negative Nb anomaly is present, but is not so marked as for the other suites. Isotopic data show that the Mealy dykes have less radiogenic Sr than the Michael and Shabogamo gabbros. The Mealy dykes also have a strongly depleted Nd isotopic signature (εNd = +3.8 to +5.6), quite different from the slight to moderately enriched character of all the other suites. Collectively, these data are consistent with fractional crystallization of a less contaminated source. Emslie *et al.* (1997) iterate that the contrasts, with respect to the other suites, suggest that the Mealy Mountains terrane is tectonically exotic relative to its surroundings. Where that exotic location was, prior to Grenvillian thrusting, other than having been somewhere to the southeast, remains conjectural.

16.3.2 PERIPHERAL MEALY DYKES

A few Mealy dykes in peripheral regions are addressed separately, mostly to facilitate better the basis for including them as part of the suite. The dykes are: i) in the northeast, outside the MMIS, ii) in the southeast, also outside the MMIS, and iii) in the southwest, within and outside the MMIS (Figure 16.3).

Northeast of the Mealy Mountains, four Mealy dyke localities are shown on Figure 16.3. Field notes for one locality (RG80-371) record fairly planar diabase dykes having a 045° strike, and diabase dykes of unspecified trend at another (RG80-372). At the other two (CG83-551/CG84-473, RG80-428), the rocks are described as amphibolite dykes. The author has only seen one of these sites. Note that CG83-551 is the same location as CG84-473; the author visited the site in 1983 and collected a sample of the host quartz syenite, then revisited the site in 1984 when the amphibolite dyke was collected and a new data station number used. As the quartz syenite is homogeneous and both the syenite and amphibolite have a common fabric (Grenvillian), the amphibolite is not demonstrably discordant. Its interpretation as a dyke rests on its sheet-like form, albeit somewhat contorted due to subsequent deformation. Both the quartz syenite and the amphibolite were investigated by U–Pb geochronological methods. As noted earlier, the quartz syenite yielded an age of 1296 +13/–12 Ma. The amphibolite yielded a zircon age of *ca.* 1038 Ma, interpreted to date time of Grenvillian metamorphism (Schärer *et al.*, 1986).

Thin sections of the amphibolite (CG84-473A1, CG84-473A2) consist of a polygonal assemblage of plagioclase, green-brown amphibole, pale-green clinopyroxene, red-brown biotite, and opaque oxide, apatite, zircon, and trace allanite.

Two whole-rock geochemical analyses are also available. They are higher in K_2O, Rb, Li, LOI and slightly lower in SiO_2 and CaO compared to typical Mealy dykes (Figures 16.5 and 16.6), which are features that can be related to metamorphism.

In the southeast, two outcrops of fine- to medium-grained mafic rocks were mapped on the Eagle River (Figure 16.3; CG97-015, CG97-113B), each exceeding 10 m in width. Contacts at CG97-015 are not exposed, thus preventing verification that the mafic rock is, in fact, a dyke and, if so, what its orientation might be. It is inferred to strike northeast, however, on the basis of internal joint trends. The localities are justified to be Mealy dykes from stained slabs and thin section evidence, plus a whole-rock geochemical analysis of one of them (CG97-015) that has the same characteristics as other Mealy dykes. A few kilometres to the southeast, on another branch of the Eagle River, four outcrops of black- to brown-weathering, massive, homogeneous, ophitic-textured diabase showing quenched plagioclase features and rare plagioclase phenocrysts (1 by 0.2 cm) were discovered. On the 1:100 000 map for the area, they have been depicted as belonging to a single northeast-trending Mealy dyke, although lack of outcrop away from the northeast-trending river, or any independent evidence (such as magnetic expression), means that the four outcrops could belong to a body having an entirely different shape. The river is coincident with a brittle fault of probable Neoproterozoic age. The rocks are inferred to predate brittle faulting because: i) they are injected by a few quartz-feldspar veins, ii) they are transected by late fractures along which alteration has occurred, and iii) they are heavily jointed. Stained slabs of samples from the four outcrops show textural similarity with other Mealy dykes, but none of these samples was examined in thin section.

In the southwest, three Mealy dyke occurrences are shown, plus one west of longitude 60°W.

Two thin sections are available (CG98-239C and CG09-056B), and whole-rock analytical data were obtained for CG09-056B. Both thin sections and the whole-rock analysis are characteristic of Mealy dykes.

16.4 OTHER MAFIC INTRUSIVE ROCKS

16.4.1 ISOLATED MAFIC ROCK OCCURRENCE AT CG00-205 (on map as M_1d)

Extrapolating 100 km from the south-southwestern outcrop (CG99-015) of the aligned mafic rocks within the Upper Paradise River intrusive suite (Section 15.3; Upper Paradise River mafic dykes(?)), and parallel to their trend, brings one to an isolated outcrop of mafic rock at data station CG00-205 (Figure 16.3). The rock is black-weathering, fine to medium grained, homogeneous, and injected by a few quartz–feldspar veinlets. In the field, the rock was noted as having a diabasic texture and lacking signs of recrystallization.

A thin section prepared from a sample from the outcrop (CG00-205) confirms the distinctiveness of the rock. Plagioclase forms interlocking, randomly oriented, elongate laths to equant, grains showing excellent twinning and no alteration – almost certainly a primary diabasic texture. Both clino- and orthopyroxene are present. Clinopyroxene forms stubby pale to mid-green, pleochroic grains containing very finely exsolved spindles of an unidentified mineral. Orthopyroxene is strongly pleochroic from pink to pale-green and has smudgy-looking orange-brown bars normal to the dominant cleavage (exsolution or biotite alteration?). Patches of apparently mutually isolated or near-isolated orthopyroxene grains are in optical continuity and must be, in fact, poikilitic crystals enclosing plagioclase and opaque grains (but not clinopyroxene). Dull-green, anhedral amphibole is also abundant, forming clusters of anhedral grains and broad coronas around opaque oxide grains. Orange- to red-brown biotite is common, also occurring in clusters. Apatite is ubiquitous. The rock has not undergone severe metamorphism and the overall texture is primary igneous and intrusive. Petrographic data support field assertions that this rock is unlike any other mafic unit in the Grenville Province in eastern Labrador with which the author is familiar. No whole-rock geochemical data are available.

The affiliation of this rock is unknown. Inasmuch as the primary igneous minerals are similar to the Upper Paradise River mafic dykes(?), it could be a correlative. The apparent alignment with the Upper Paradise River occurrences is probably coincidental, given that Grenvillian deformation generated a major reclined fold in the region. Furthermore, CG00-205 is in the Pinware terrane, whereas the Upper Paradise River occurrences are in the Mealy Mountains terrane and that there is evidence that the boundary between them is a major thrust. The rock is sufficiently unlike other mafic rocks in the region to claim a separate origin. It may be related to the Petit Mecatina AMGC suite in eastern Québec.

16.4.2 LOURDES-DU-BLANC-SABLON GABBRONORITE (M_2rg)

The Lourdes-du-Blanc-Sablon gabbronorite is, strictly, outside the area addressed in this report, being in Québec, 7 km west of the Labrador–Québec border at Blanc-Sablon (*see* Figure 22.10). It was, however, included in a study of part of the Pinware terrane by Heaman *et al.* (2004) and the occurrence is relevant to mafic magmatism in the easternmost Grenville Province.

The gabbronorite forms a mega-boudin within well-banded gneissic granitoid rocks at Lourdes-du-Blanc-Sablon. The rock is coarse grained to pegmatitic (Appendix 2, Slab 16.2) and shows compositional and grain-size layering. Igneous crossbedding is present in places. Dykes or sills, exhibiting fine-grained chilled margins and grading to more coarse-grained gabbro inside the layered complex, indicate multiple injections of gabbronorite magma. Pegmatitic gabbro forms metre-sized pockets and veins that intrude the rest of the complex. Granophyre veins and veinlets are locally present. Corona textures are evident, including olivine mantled by orthopyroxene, clinopyroxene mantled by hornblende + garnet, and opaque minerals mantled by biotite and hornblende. Although igneous textures are preserved, most coronas show extensive recrystallization. Some metamorphic hornblende veinlets are present.

The margin of the gabbronorite intrusion is sheared and contains shredded pegmatite and remnants of fine-grained metagabbro/amphibolite. The fine-grained metagabbro or amphibolite is composed mostly of granoblastic hornblende and plagioclase with lesser fine-grained granoblastic orthopyroxene and relict crystals of cataclastic igneous plagioclase. A gradation between the coronitic gabbronorite and fine-grained amphibolite is observed, expressed by grain-size reduction and transformation of coronas into poikiloblastic hornblende–plagioclase–biotite intergrowths. Pegmatites truncate layering in the gabbronorite but are transposed into parallelism with the sheared contact. The contact between the gabbronorite and the country rocks is sharp.

The country rock to the mafic intrusion was derived originally from coarse-grained, strongly foliated, K-feldspar-rich biotite granite. Near the mafic intrusion, pegmatitic and amphibolitic layers in the granitoid rock are boudinaged and parallel to the contact, giving the bulk rock an overall gneissic appearance. Farther away, where deformation is less severe, mafic dykes truncate pegmatite within foliated granitoid rock. If the mafic dykes within the granitoid rock are genetically related to the gabbronorite intrusion, the sequence of events is: i) granitoid emplacement, ii) pegmatite injection, iii) gabbronorite and satellite mafic dyke intrusion, iv) pegmatite emplacement, and v) shearing along contact.

A sample of coarse-grained gabbronorite (97SP-068B) contains baddeleyite, polycrystalline zircon overgrowths (interpreted to be metamorphic) on baddeleyite, and colourless zircon fragments (interpreted to be primary magmatic). Five fractions were selected from this sample and include zircon fragments, irregular platy zircon, polycrystalline zircon and one baddeleyite fraction consisting of brown blades and fragments. Perhaps the most significant analyzed fraction is baddeleyite, which yielded a concordant age of 1248 ± 5 Ma. This date is the best estimate for the emplacement age of the gabbronorite.

Late Elsonian activity, defined by Gower and Krogh (2002) as extending from 1290 to 1230 Ma, was previously unknown in the southeasternmost Grenville Province, so the 1248 ± 5 Ma Lourdes-du-Blanc-Sablon gabbronorite age represented a significant new finding by Heaman *et al.* (2004). Other examples of late Elsonian mafic magmatism in eastern Laurentia are the 1273 ± 1 Ma Harp dykes (Cadman *et al.*, 1993), the 1250 ± 2 Ma Mealy dykes (Hamilton and Emslie, 1997), and gabbro sills in the Seal Lake Group from which ages of 1250 +14/–7 Ma and 1224 +6/–5 Ma have been obtained (Romer *et al.*, 1995). All these rocks are distant (the Mealy dykes being the closest at 200 km to the north-northwest), and no obvious candidates have been recognized in the intervening area. The only other rocks of similar age recognized in the eastern Grenville Province are granites from the Natashquan River area, having emplacement ages of 1245 ± 3 and 1239 ± 3 Ma (Clark and Machado, 1994). These are located about 300 km to the west, with the intervening area being both geologically and geochronologically very poorly known.

16.5 BATTLE ISLAND SUPRACRUSTAL ROCKS

The supracrustal rocks on Battle Island (mentioned by Kranck, 1939; and Christie, 1951; mapped by Gower *et al.*, 1988; and Gower, 2008, 2009) have been subdivided by the author into six lithological units (Figure 16.7). From east to

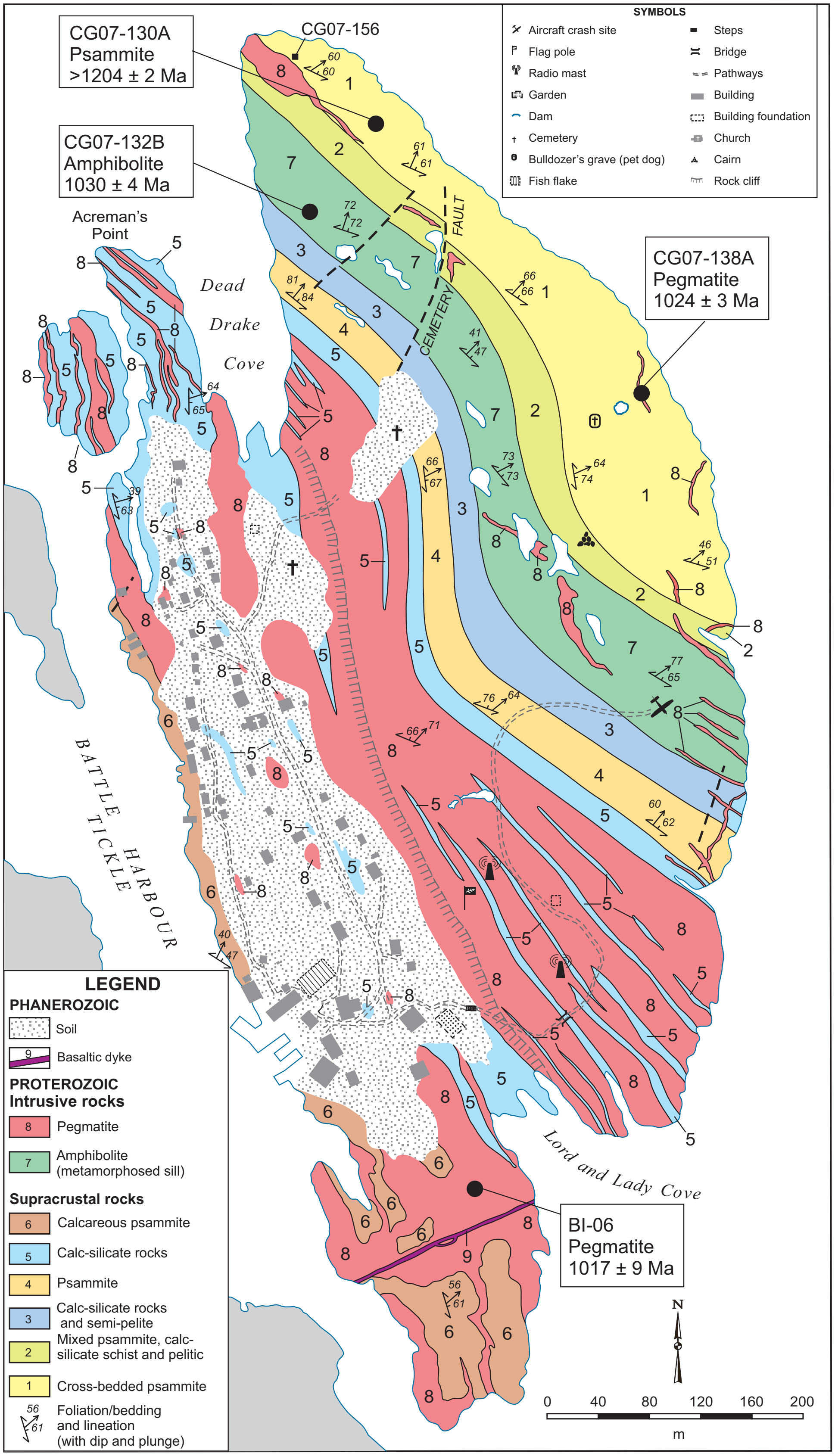


Figure 16.7. *Battle Island geological features and geochronological sites.*

west, these are: i) crossbedded psammite, ii) mixed psammite, calc-silicate schist and pelitic schist, iii) calc-silicate rocks and semi-pelite, vi) psammite, v) calc-silicate rocks, and vi) calcareous psammite.

16.5.1 CROSSBEDDED PSAMMITE (on map as PMss)

The crossbedded psammite has a light-grey, sugary texture on fresh surface, but brownish crust where weathered. The psammite is maroon where hematized. It is a relatively homogeneous, fine to medium grained, and about 100 m thick at its widest part. Its total width is unknown as its eastern boundary is not exposed.

Three features are characteristic of this unit. The first is the presence of bedding in the form of parallel and crossbedding lamination (Plate 16.1D, but *see* Gower, 2009 for additional photographs of the full range of Battle Island supracrustal rocks). The laminae are enriched in biotite and opaque minerals. The crossbedding laminations, first described by Kranck (1939), are confined to particular layers and are commonly about 30 cm thick. Tops are generally to the west, but, as tight folds were seen, some tops to the east may also be present.

A thin section (CG07-130A) that includes two heavy mineral laminations consists mostly of recrystallized quartz, lesser plagioclase and interstitial well-twinned microcline associated with aligned flakes of buff-green biotite. The heavy mineral layers consist mostly of an opaque oxide (*cf.* ilmenomagnetite), but rounded zircon and apatite grains are common. Titanite is also common in these layers, partly mantling the opaque oxide. Minor epidote and chlorite are found throughout. A whole-rock chemical analysis of the thin-sectioned sample has 86.5% SiO_2 (Gower, 2009).

The second feature is a spotted appearance due to hornblende poikiloblasts up to about 1 cm across and surrounded by a haloes depleted in mafic minerals. Commonly, the poikiloblasts are concentrated into particular zones typically extending along original bedding planes. Outside of the amphibole poikiloblasts, the mineral assemblage is similar to that seen in the previously described sample.

The poikiloblasts are dark-green to blue-green and are thoroughly sieved with abundant quartz, plagioclase and microcline. The area around the poikiloblasts is depleted in biotite, which was consumed at the expense of amphibole growth. Other minerals present are an opaque oxide, apatite, zircon and late-stage metamorphic epidote and chlorite. The zircons are small and show a spatial affinity with the hornblende porphyroblasts and may be a metamorphic product themselves (thin sections CG87-489A, CG07-130B, C; Photomicrograph 16.1C).

The third feature is the presence of a few yellowish-green layers, generally less than 1–2 centimetres in width.

In thin section (CG07-130C) these are seen to be quartz-rich compared to the typical psammite described above, although still containing some plagioclase and microcline. The yellow-green colour is due to markedly pleochroic epidote. Other minerals include relict, dark-green to blue-green amphibole, an opaque oxide, titanite (associated with the opaque mineral and epidote), apatite, minor chlorite (after amphibole), and a few small grains of zircon.

16.5.1.1 Geochronology of Crossbedded Psammite

Kamo *et al.* (2011; sample CG07-130A) reported that eight thermally annealed and chemically leached zircon crystals gave a range of ages with the youngest at 1204 ± 2 Ma (2.0% discordant). The youngest age of 1204 ± 2 Ma was interpreted as representing a maximum for deposition of the sedimentary protolith. This is supported by a 1208 ± 3 Ma age for a zircon grain from one of the pegmatites (*see* Section 17.3.4.1, sample CG07-138B). The zircon from the pegmatite, which intrudes the psammite, must be interpreted as inherited because the emplacement of the intrusion is known to postdate 1024 ± 3 Ma.

Despite the near-concordant 1204 Ma grain, it appeared possible that it was on a mixing line with other data between Pinwarian and Grenvillian events. The five youngest data points together define a line with upper and lower intercepts of 1532 ± 20 and 1138 ± 34 Ma (MSWD = 3.0), respectively. Within error, the upper intercept could be Pinwarian, and the lower intercept, although predating Grenvillian orogenesis in eastern Labrador, could be coeval with the age of emplacement of the Gilbert Bay granite (1132 +7/–6 Ma; Gower *et al.*, 1991) or a granitic vein close to the Gilbert Bay pluton having an age of 1113 +6/–5 Ma (Scott *et al.*, 1993) (next chapter).

To investigate further the possible existence of young detrital zircons of igneous origin in this sample, a broader selection of zircon grains from the psammite was imaged and dated using LA–MC–ICPMS (144 analyses from 101 grains; Kamo *et al.*, 2011). These grains were selected from the same concentrate used for TIMS dating. Based on these analyses, several age modes are recognized in this detrital zircon population at ~2.72, 2.05, 1.95, 1.85, 1.80, 1.75, 1.65, 1.60, 1.45, 1.30 and 1.05 Ga. The data obtained are consistent with the conclusion that the youngest concordant TIMS date of 1204 Ma provides a maximum depositional age constraint for the psammite.

A separate study of a sample of Battle Island crossbedded psammite was carried out by Spencer *et al.* (2015; sample CS11-6 – termed Battle Harbour psammite). Dominant age peaks were identified at 1800, 1520, 1250 and 1150 Ma.

16.5.2 MIXED PSAMMITE, CALC-SILICATE SCHIST AND PELITIC SCHIST (on map as PMss)

This, very variable, metasedimentary unit is about 25 m thick. The rocks weather to various hues of grey, brown, green, creamy and white, are mostly medium to coarse grained, and are thinly to thickly bedded. Component rock types are micaceous amphibolite; epidote-rich rock with amphibole porphyroblasts; psammite; schistose semi-pelite; amphibole-rich material with psammitic partings; psammite with amphibole-rich layers or partings; and psammite with lensy, discontinuous layers. Lenses of a white-weathering calc-silicate rock occur up to a few metres in length, but rarely more than 30 cm thick in contact with amphibolite located to the west. Pale-green epidote lenses, typically 5 to 10 cm thick, are common throughout the unit.

Four samples from this unit, representing the main lithological variants, were examined in thin section. The most mafic-mineral-rich rock type is hornblende–biotite schist also containing quartz, plagioclase, microcline, titanite, epidote, apatite, allanite and minor chlorite (CG07-131A). A mesocratic rock (CG07-131C) consists of quartz, plagioclase, phologopitic mica, tremolite, clinozoisite and titanite. A leucocratic variant (CG07-131B) is carbonate-rich calc-silicate rock, containing quartz, plagioclase, K-feldspar, carbonate, clinozoisite, actinolite, opaque mineral(s) and titanite. A white-weathering, hornfelsic-textured calc-silicate rock (CG07-131D) devoid of mafic minerals, occurs as discontinuous lenses adjacent to the amphibolite to the west and consists of quartz, carbonate, and diopside. The protolith for this unit was probably a heterogeneous sequence of muddy calcareous rocks, grading into limestone in places.

The next unit to the west is amphibolite, which is interpreted to be intrusive.

16.5.3 CALC-SILICATE ROCKS AND SEMI-PELITE (on map as PMsc)

West of the amphibolite is a grey-, greenish- and black-weathering, fine to medium grained, very well layered unit displaying alternation of rock types at the centimetre to decimetre scale. It consists mostly of schistose micaceous calc-silicate rocks, but also includes some psammitic material. A distinctive 3-m-wide pale-green rock containing pods of orange-brown garnet (*cf.* andradite) is present immediately west of the contact with the amphibolite.

The four samples of this unit examined in thin section are all calc-silicate rocks, although their mineral assemblages differ. The garnet-bearing rock (CG07-133A) also contains clinozoisite/zoisite, quartz, carbonate, diopside, highly sericitized plagioclase, titanite and minor, possibly secondary, tremolite. Clinozoisite/zoisite is partly symplectic with quartz, which commonly also occurs as inclusions in garnet. Diopside is also a major constituent of sample CG07-133B, where it is associated with plagioclase, phologopitic mica and titanite. A few tremolite blades and chlorite flakes are also present. Sample CG07-133C contains tremolite as a major mineral, along with phlogopitic mica, clinozoisite, plagioclase and quartz. Minor titanite and allanite are present. The fourth sample (CG07-133D) contains plagioclase, tremolite, phologopitic mica, microcline and titanite.

The mineral assemblages in this unit are very similar to those in the mixed unit on the eastern side of the amphibolite. If it is accepted that the amphibolite is intrusive, it is not unreasonable to regard the rocks flanking the eastern and western sides of the amphibolite as being parts of a single unit prior to being separated by emplacement of the amphibolite.

16.5.4 PSAMMITE (on map as PMss)

The contact between the schistose micaceous calc-silicate rocks and the psammite to the west is sharp and regular. The psammite is about 20 m wide, grey-, creamy- or reddish-weathering, fine to medium grained, homogeneous, and thinly to thickly bedded. The contact with calc-silicate rocks farther west is also sharp.

A thin section from this unit (CG07-134) comprises quartz, plagioclase, microcline (all polygonal, equant and clearly thoroughly recrystallized), together with buff-brownish, ragged, short laths of biotite, and small accessory grains of apatite, zircon, titanite and chlorite. The biotite is concentrated at grain interfaces between felsic minerals and is a typical example of the texture described by Gower (2007) in which the phyllosilicate flakes are of similar length to the adjacent felsic minerals, and interpreted as the product of metamorphism of an intergranular mud. The titanite and chlorite are secondary and derived, at least in part, from the retrogression of biotite. Most of the opaque mineral is secondary hematite or limonite, from the breakdown of primary opaque oxide, and perhaps from minor sulphide also.

16.5.5 CALC-SILICATE ROCKS (on map as PMsc)

The calc-silicate unit occupies a wide swath from the northwest tip of Battle Island to the central part of the southeastern shoreline, albeit heavily injected by pegmatite. The rocks are generally light- to dark-green-weathering, medium grained, thinly and well bedded, have variable mineral assemblages, both in mineral composition and their habits of (metamorphic) growth. Typically, the rocks have a very ribbed or pitted surface appearance due to alternating positively and recessively eroding layers and differential weathering of the various minerals present.

Apart from calc-silicate rocks, layers of amphibolite, semipelite and psammite are also present. Rusty-weathering, sulphide-bearing patches are common, locally forming pods up to about 1 m thick and 2 m long. Although rubiginous and have a sulphurous smell on fresh surface, sulphide content is minor.

Four samples were collected near the contact with the psammite to the east and examined in thin section. Two of them (CG07-135A, CG07-135C) and the leucocratic part of a third (CG07-135D) have similar mineral assemblages, although mineral proportions differ. The minerals are plagioclase, carbonate, K-feldspar, diopside, tremolite and titanite. The plagioclase, carbonate and K-feldspar are typically anhedral, show straight grain boundaries and 120° triple junctions, and are obviously recrystallized. Plagioclase is typically very sericitized and poorly twinned. K-feldspar is well-twinned microcline. Diopside is locally altered to tremolite, which, growing discordantly to the prevailing fabric, is clearly posttectonic in places. Titanite is typically dark-brown and anhedral to subhedral. Sample CG07-135B and the melanocratic part of sample CG07-135D consist mostly of a pale-orange mica (*cf.* phologopitic mica), diopside, tremolite, minor plagioclase, with minor oxide opaque mineral, apatite and possibly zircon (very small inclusions showing pleochroic haloes in phologopitic mica). In sample CG07-135B, the diopside occurs as large poikiloblastic grains containing abundant inclusions of phologopitic mica and plagioclase.

Six samples were examined in thin section from the western part of the calc-silicate unit, in the Acreman's Point area (CG87-487B, CG07-136A to CG07-136E). All contain plagioclase, K-feldspar, phlogopitic mica, (an) opaque mineral(s), and titanite. Plagioclase is typically anhedral with straight grain boundaries and 120° triple junctions and heavily sericitized. K-feldspar shows the same habit, and is characteristically well-twinned microcline. The phologopitic mica ranges from pale-orange to reddish-orange and commonly defines a strong fabric. Both oxide and sulphide opaque minerals are present in all five samples except CG07-136E, in which only oxide was observed. Sulphide is especially abundant in CG07-136A and CG07-136B, which were both described as gossanous in the field. The sulphide is thought to be mostly pyrite, but pyrrhotite is suspected in CG07-136A. Titanite varies from anhedral and amoeboid to euhedral. It is abnormally large in sample CG07-136E, where it occurs as grains up to 3 mm long. Other minerals include diopside and minor secondary tremolite (in sample CG07-136E) and tremolite (as a stable phase in CG87-487B, CG07-136A, C and D). Neither diopside nor tremolite is present in CG07-136B, which is more psammitic than the other samples. Graphite was noted in CG87-487B.

Whole-rock analyses were obtained for gossanous calc-silicate samples CG07-136A and CG07-136B. Neither sample is especially anomalous in elements typically of interest to mineral explorationists. The most noteworthy chemical feature of both of them is their high F content.

16.5.6 CALCAREOUS PSAMMITE (on map as PMss/PMsc)

The calcareous psammite, forming the westernmost supracrustal unit on Battle Island, is pale–grey-, green-, or brownish-weathering, fine to medium grained, and thinly to thickly bedded. Bedding is defined by darker, amphibole- or biotite-rich layers. Locally oblique layering is present that could be relict crossbedding. Some layers have a mottled texture, which is attributed to incipient melting. The psammitic rocks are intercalated with calc-silicate layers, locally showing oblique layering that resembles, and may be, crossbedding.

Three samples were examined in thin section from this unit (CG07-137A, B and C). Sample CG07-137A is a calcareous psammite and is distinct in the wide range of minerals present. Twelve minerals were identified in thin section; these are quartz, plagioclase, K-feldspar, diopside, garnet, clinozoisite/epidote, hornblende, tremolite, carbonate, titanite, apatite and leucoxene. The rock is clearly recrystallized, as shown by polygonization of several minerals; most notably quartz, plagioclase and carbonate, which all have straight grain boundaries and 120° triple junctions. K-feldspar is a very minor phase, more obvious in stained slab than thin section. Diopside occurs as pale-green, anhedral poikiloblasts, enclosing quartz and plagioclase, especially. Garnet is also poikiloblastic, enclosing quartz, clinozoisite/epidote, diopside, plagioclase and K-feldspar. The clinozoisite/epidote is colourless and shows its characteristic anomalous birefringence. No formal attempt was made to distinguish between clinozoisite and epidote, but the fact that the diopside is green (in contrast to colourless diopside seen in other Battle Island samples) suggests that the Fe content is higher in this sample than elsewhere, and that the mineral is epidote. Hornblende and tremolite are both secondary, as alteration products from diopside. Of the other minerals; titanite occurs as dark brown euhedral grains; apatite forms small euhedral crystals; and leucoxene is the name applied to granular, amorphous-looking opaque material that is white in reflected light. This is clearly a disequilibrium assemblage.

Samples CG07-137B and C are both psammitic rocks and, like CG07-137A, are mineralogically varied. Both contain quartz, plagioclase and K-feldspar, all of which have polygonal form and show straight grain boundaries. The mafic and accessory minerals differ between the two, however. Sample CG07-137A contains dark-green (sodic?) amphibole, dark-brown titanite, zircon, apatite, biotite, epidote, and chlorite (the latter three secondary after amphibole), whereas CG07-137C lacks amphibole, epidote and zircon.

16.5.7 BATTLE ISLAND SUPRACRUSTAL ROCKS CORRELATIVES? (on map as PMsp)

Quartzite, south of Long Harbour, forms a distinctive unit of grey to white weathering, medium-grained rock and was previously mapped by Christie (1951). The extrapolation of this unit westward on the 1:100 000-scale map (St. Lewis River) from the coast is based on Christie's map, as it was only examined on the shoreline during mapping. It seems likely, as suggested by Christie, that the quartzite south of Long Harbour is the lateral equivalent of quartzite on Battle Island.

Two thin sections (CG87-576, JS87-571) of psammite from this area consist of plagioclase, microcline, quartz, green to orange biotite, muscovite, and opaque oxide, and accessory apatite, titanite, zircon and allanite.

Farther afield, the reader is reminded of the supracrustal rocks on Black Island on the north side of Groswater Bay in the Groswater Bay terrane. These undated rocks are at a lower metamorphic grade than other rocks having a supracrustal protolith in the area and the suggestion was made that they could be comparable in age to the Battle Island supracrustal rocks (Section 7.3.1.4).