CHAPTER 7

LATE PALEOPROTEROZOIC (PRE-LABRADORIAN) SUPRACRUSTAL ROCKS (P_{3A} 1810–1770 Ma)

7.1 INTRODUCTION

Supracrustal rocks interpreted to be late Paleoproterozoic and pre-Labradorian are mostly pelitic and psammitic gneiss, with minor quartzite, calc-silicate rocks and amphibolite. The pelitic rocks are typically sillimanite–garnet–biotite–gneisses, but kyanite occurs with (and instead of) sillimanite in some parts of the region. Cordierite and/or hypersthene are/is common in specific areas, and osumilite, sapphirine, staurolite and relict andalusite are also known. A wide range of pressure–temperature conditions are therefore represented. Psammitic gneiss is difficult to discriminate from orthogneiss in many instances, introducing uncertainty as to true proportions. Quartzite and calc-silicate rocks are not common, but are invaluable in diagnosing a supracrustal protolith for the rocks with which they are associated. The amphibolite is derived, in part, from mafic volcanic rocks. Felsic fragmental or conglomeratic rocks are unknown, except for two small occurrences in the Groswater Bay terrane (Gower, 1996). The distribution of the metasedimentary gneiss is shown in Figure 7.1. A few very minor occurrences of similar rocks occur farther south in the Pinware terrane, which may or may not be correlative; these are addressed in Chapter 13.

7.2 ISOTOPIC DATA

In the following review of isotopic data, the simplifying assumption is made that all the high-grade supracrustal gneisses are broadly coeval hence isotopic data from all terranes can be addressed collectively. While this assumption is permissive from field relationships and encouraged by broad lithological uniformity, it has yet to be rigorously demonstrated.

Determining the age of deposition of supracrustal rocks is much more challenging than dating crystallization/recrystallization events in igneous and metamorphic rocks. Unless there are interlayered volcanic rocks (which there are, at Bull Island – Section 7.3.3.1), the best that can be achieved are maximum and minimum constraints on time of deposition. Even here, reliance may be placed on very indirect (and commonly inconclusive) evidence.

The U–Pb geochronological age constraints are reviewed first, then those offered by other isotopic data. The principal geochronological sites are indicated in Figure 7.1.

7.2.1 U–Pb DATA

7.2.1.1 Maximum Age of Deposition Based on Inherited Zircon

One indirect approach to determining maximum time of deposition utilizes ages of inherited zircons in younger rocks, even if the younger rock has an igneous protolith. Obviously, the time of deposition must postdate the youngest inherited zircon age. The inherited zircons may have come from supracrustal or plutonic protoliths, of course, but if the zircons have a wide spread of ages then it is more likely that the source was supracrustal. The more diverse the inherited-zircon age spectrum, the greater probability that this was the case. Samples that provide a range of inherited ages in eastern Labrador are CG95-341A and GN95-392 (both from the Mealy Mountains terrane), and CG00-154A (Pinware terrane).

Sample CG95-341A is well-banded to mylonitic quartzofeldspathic gneiss of uncertain protolith. It was originally mapped as granodioritic orthogneiss, but the presence of minor muscovite was taken by Gower *et al.* (2008b) as a weak hint that an aluminous metasedimentary component might be present (but no garnet, cordierite or aluminosilicates were seen). Five single zircon grains were collectively interpreted to mean that the rock was intruded at 1800 ± 40 Ma (based on two analyses) and that it contained zircon of detrital origin having $^{207}Pb/^{206}Pb$ ages of 2605 ± 3 Ma (9.2% discordant), 1902 ± 2 Ma and 1894 ± 3 Ma (both less than 1.3% discordant). The conclusion to be drawn is that, if a sedimentary component is present, it must have been deposited between 1894 and 1800 Ma.

Sample GN95-392 has a more colourful interpretational history. It was initially mapped as monzonite belonging to the 1646–1635 Ma Mealy Mountains intrusive suite (Chapter 11), but analysis yielded unexpectedly old zircons having $^{207}Pb/^{206}Pb$ ages between 1739 and 1709 Ma. On revisiting the outcrop and examining it in more detail, it was

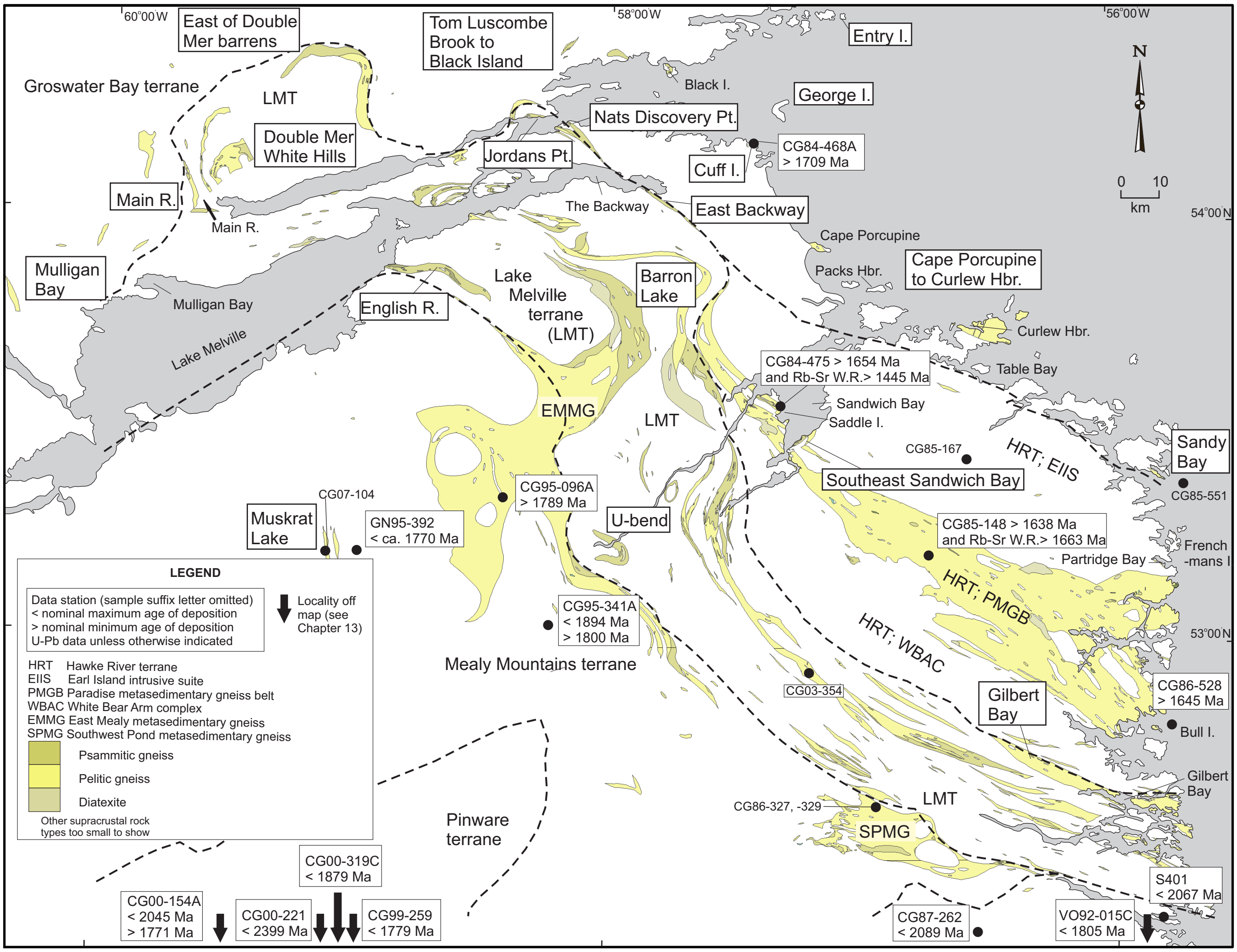


Figure 7.1. *Distribution of late Paleoproterozoic (pre-Labradorian) supracrustal rocks in eastern Labrador, also locating sites offering depositional age constraints from geochronological data. Subareas addressed in text are indicated in boxed text.*

discovered that the monzonite contains numerous enclaves (extremely similar in appearance to the monzonite), introducing the possibility that one of the enclaves had been previously sampled. To test this hypothesis, the same monzonite (on the basis of field and petrographic appearance) was resampled at a locality known to be homogeneous (CG95-154), thus reducing the likelihood of inheritance, if that was the issue. Four of five zircons gave concordant data (the fifth was 1.1% discordant) yielding an age of 1642 ± 4 Ma. Using this age as a lower intercept anchor for regressing the data from sample GN95-392, gives upper intercept ages between *ca.* 1914 and 1770 Ma (Krogh *et al.*, 1996; Gower *et al.*, 2008b). The range of ages suggests a detrital origin, implying sedimentation after *ca.* 1770 Ma.

Sample CG00-154A is a quartzofeldspathic enclave within a moderately foliated, homogeneous K-feldspar megacrystic granodiorite. Seven of ten, single-zircon analyses provide an upper intercept of 1771 ± 4 Ma, when projected from 1043 Ma (the age of the separately analyzed K-feldspar megacrystic granodiorite host). The three other zircons give $^{207}Pb/^{206}Pb$ ages of 2745 ± 2 Ma, 2377 ± 2 Ma and 2045 ± 2 Ma. The results were interpreted by Gower *et al.* (2008b) to mean that the quartzofeldspathic enclave was derived from sediment having a *ca.* 1800 Ma provenance, but incorporating older material.

In addition to the samples above, from which multiple inherited zircon ages were obtained, several samples were analyzed from which only one inherited age was obtained. It is not implied that these necessarily came from a detrital source, although they might have. All are from the Pinware terrane. As obtaining inherited ages was not a primary objective at the time, they should be considered excellent subjects for supplementary investigations of this nature. Samples that fall into this category are as follows:

i) Pegmatitic infill (CG00-319C) in amphibolite, which forms a concordant layer within gneissic granodiorite. In addition to its Grenvillian 1017 ± 3 Ma zircon age of formation, it contains inherited zircons of various ages, including two collinear zircons (anchored at 1017 Ma) that give an age of 1879 ± 10 Ma (Gower *et al.*, 2008b),
ii) Granite to alkali-feldspar granite (CG99-259B) in concordant contact with Labradorian tonalite gneiss. When anchored at 1029 Ma (time of Pb loss in the associated tonalitic gneiss), discordant zircon lies on a line projecting to 1779 ± 18 Ma (Gower *et al.*, 2008b),
iii) A migmatitic quartz monzonite (CG87-262), having a lower intercept age of 1450 +15/–21 Ma that was interpreted to date time of metamorphism, has an upper intercept at 2089 ± 140 Ma (Wasteneys *et al.*, 1997), and
iv) A granitic vein (S401), having a lower intercept age of 1509 +11/–12 Ma that was interpreted to date emplacement, has an upper intercept of 2067 ± 28 Ma (Scott *et al.*, 1993).

Two other samples deserve mention. Both are from the Pinware terrane. Feldspathic quartzite, CG00-221, has a lower-intercept metamorphic age of *ca.* 1020 Ma, which anchors discordant zircon data yielding upper intercept dates of *ca.* 2460 Ma and 2399 ± 7 Ma (Gower *et al.*, 2008b). Quartzite VO92-015C, which also has a lower intercept metamorphic age of 1020 Ma, yields upper intercept ages of 2720 ± 20 Ma, 1878 ± 5 Ma and 1805 ± 10 Ma for discordant zircon results (Wasteneys *et al.*, 1997). Beyond both having a minimum age of *ca.* 1020 Ma as defined by time of metamorphism, the time of deposition is unknown. Both samples may belong to younger supracrustal assemblages, hence are addressed in Chapter 13 (*cf.* Pitts Harbour Group and correlative rocks).

7.2.1.2 Minimum Age of Deposition Based on Field Relationships

Only one locality can be cited that meets the twin demands of: i) having reliable identification of a metasedimentary host rock and an igneous rock intruding it (soon?) after deposition, and ii) having had U–Pb geochronological investigations carried out. At the qualifying site, an enderbitic granulite gneiss (CG95-096A) intrudes pelitic metasedimentary gneiss. It yielded an age of 1789 ± 29 Ma based on three almost-concordant zircon analyses, when anchored by a lower intercept of 1640 ± 2 Ma from metamorphic zircon data in a crosscutting mafic dyke (Gower *et al.*, 2008b).

Another locality that comes close to meeting the criteria is CG84-468A in the Groswater Bay terrane, where a granodioritic rock dated to be 1709 +7/–6 Ma contains enclaves of gneiss (Schärer *et al.*, 1986). It falls short in that, although the gneiss enclave is most likely of supracrustal origin, such has not been conclusively demonstrated.

In field notes, numerous localities elsewhere in eastern Labrador are reported at which gneissic granitoid rocks host metasedimentary gneiss enclaves (*e.g.*, in the 1670 Ma Earl Island intrusive suite). No site-specific geochronological data exist for these, however.

7.2.1.3 Minimum Age of Deposition Based on Time of Metamorphism

Two samples of directly dated sillimanite-bearing pelitic gneiss have yielded metamorphic U–Pb ages. Both are from the Paradise metasedimentary gneiss belt in the Hawke River terrane.

Three zircons from the first dated sample (CG84-475) plot highly discordantly close to 977 ± 2 Ma titanite from the same sample, thus demonstrating extensive Pb loss during Grenvillian metamorphism. The extrapolated upper intercept for the zircon data define an age of 1654 +29/–28 Ma, which was interpreted at the time as either an average age for detrital zircon or due to Labradorian metamorphism (Schärer and Gower, 1988).

Seven near-concordant upper-intercept zircons from the second sample (CG85-148) having an age range from 1647 to 1627 Ma were initially interpreted (with reservations) as detrital, because the zircons displayed a very high range in U content (81–1876 ppm), coupled with a variety of morphologies and degrees of rounding (Gower *et al.*, 1992). Accepting this interpretation introduced conflicts in the then-understood regional geological history, so additional samples from the same outcrop were investigated, namely a microgranite vein (CG85-148B) and a pegmatite considered to be a melt pod (CG84-148C). Results from both samples for five concordant or near-concordant zircons and one monazite overlap within analytical uncertainty, so analyses were regressed together to yield an age of 1638 +11/–3 Ma (Kamo *et al.*, 1996). The data thus required rejection of the 1647–1627 Ma detrital-zircon hypothesis in favour of a mid- to late-Labradorian migmatitic event.

These results obviously offer little by way of constraining the depositional age of the metasedimentary gneisses, although they do provide valuable information regarding Labradorian and Grenvillian metamorphism in the region.

7.2.1.4 Depositional Age Based on Interlayered Volcanic Rocks

One might think that, since dating of interlayered volcanic rocks is the most direct way of determining the age of the associated metasedimentary rocks, any attempt to do so would have been reported first. Such an attempt was made, but depositional ages were not obtained, hence relegation here.

The samples studied come from Bull Island where well-preserved pillowed and unpillowed mafic flows, interlayered sediments, crosscutting mafic dykes and migmatitic leucosome are exposed. This site is within the Paradise metasedimentary gneiss belt. Four near-concordant zircons from the unpillowed flow (CG86-528A) yielded an age of 1645 ± 4 Ma, interpreted to date metamorphism (Kamo *et al.*, 1996).

7.2.1.5 Conclusions Regarding Depositional Age of Supracrustal Gneiss

Despite repeated attempts, the age of deposition of the supracrustal gneisses has not been rigorously determined. Inasmuch as the supracrustal gneisses are intruded by rocks having 1800–1770 Ma ages, the conclusion is forced that they must be earlier. However, if inherited zircons having ages as young as 1770 Ma are detrital, then the supracrustal rocks cannot be older. Given analytical latitude, these two suppositions are not necessarily contradictory. They do, however, imply that sediment deposition and igneous activity likely occurred more-or-less simultaneously. Gower and Krogh (2003), updated by Gower (2012), made comparison with the Ketilidian Mobile Belt in southern Greenland (Figure 5.2), where supracrustal rocks were deposited between 1820 and 1790 Ma and are thought to represent sediments eroded from the emerging 1850–1800 Ma calc-alkaline Julianehåb batholith and its felsic volcanic carapace, and deposited in intra- and fore-arc basins (Chadwick and Garde, 1996; Garde *et al.*, 2002).

Despite the attractive temporal comparison with the Ketilidian Mobile Belt, there may be conflict regarding previously suggested depositional environments. Going beyond eastern Labrador, most interpretations envisage some form of subaqueous environment that was distant from high-energy sources of detritus supply (in contrast to the Ketilidian setting), in keeping with the high proportion of pelitic gneiss, but lack of coarse clastic material. A subaqueous environment is supported by the presence of pillowform mafic volcanic rocks in the Hawke River terrane. Nevertheless, Arima *et al.* (1986) reported high $\delta^{18}O$ values, averaging 12.2 ± 2 per mil SMOW (standard mean ocean water), from pelitic gneiss in the Grenville Province in central Labrador (Figure 5.4) and argued that the protolith was oxidized shale. Leong and Moore (1972) proposed a lateritic origin for rocks from the same region based on mineral assemblages and compositions, especially having hematite as the oxide phase and the refractive index of sapphirine being high, which implies a high Fe^{3+}/Fe^{2+} ratio.

7.2.2 OTHER ISOTOPIC DATA

7.2.2.1 Sm–Nd

Fourteen Sm–Nd isotopic analyses for high-grade metasedimentary rocks are listed in Table 7.1 (Schärer, 1991; Prevec *et al.*, 1990; Hegner *et al.*, 2010; Hewitson, 2010; Moumblow, 2014; R. Creaser, personal communication, 1999). Three of them are the same samples as used for U–Pb analysis (CG84-475, CG95-341A, GN95-392 – *see* previous section), and one sample is from the same outcrop as that sampled for U–Pb analysis (SPM86-09A/CG85-148). All have been examined petrographically and the sedimentary protolith for the pelitic gneiss is not in doubt, and that for the psammitic gneiss is also confidently asserted. The protolith for the remaining samples is more equivocal.

Based on protolith age for the high-grade metasedimentary rocks from U–Pb evidence, an 1800 Ma age for their formation has been adopted, generating T_{DM} = 2.29 to 1.77 Ga and εNd (1.8 Ga) = –2.65 to + 4.37. In a linear regression of εNd (1.8 Ga) *vs.* T_{DM}, only sample GN95-392 falls slightly off the line. The cluster of εNd (1.8 Ga) values around + 4.0 suggests a mantle origin for the source rocks. Inferred is that this source was mixed with homogenous older material, to give the ranges of εNd (1.8 Ga) and T_{DM} observed. Sample CG03-354F (locality indicated on Figure 7.1) contrasts from other pelitic gneisses in its εNd (1.8 Ga) and T_{DM} values, but it is not macroscopically or petrographically distinct. Its εNd (1.8 Ga) and T_{DM} values (–2.65 and 2.29 Ga) can be generated by mixing an Archean source having T_{DM} = 2.70 Ga and εNd = – 8 with a Proterozoic source having T_{DM} = 1.80 Ga and εNd = + 4 in a 55:45 percentage ratio.

Such a conclusion is consistent with the previously suggested juvenile-source comparison made with the Ketilidian Mobile Belt.

7.2.2.2 Rb–Sr

Two Rb–Sr isotopic studies have been carried out on high-grade metasedimentary gneiss, both from the Paradise metasedimentary gneiss belt.

Table 7.1. Sm–Nd isotopic data for metasedimentary gneiss in eastern Labrador

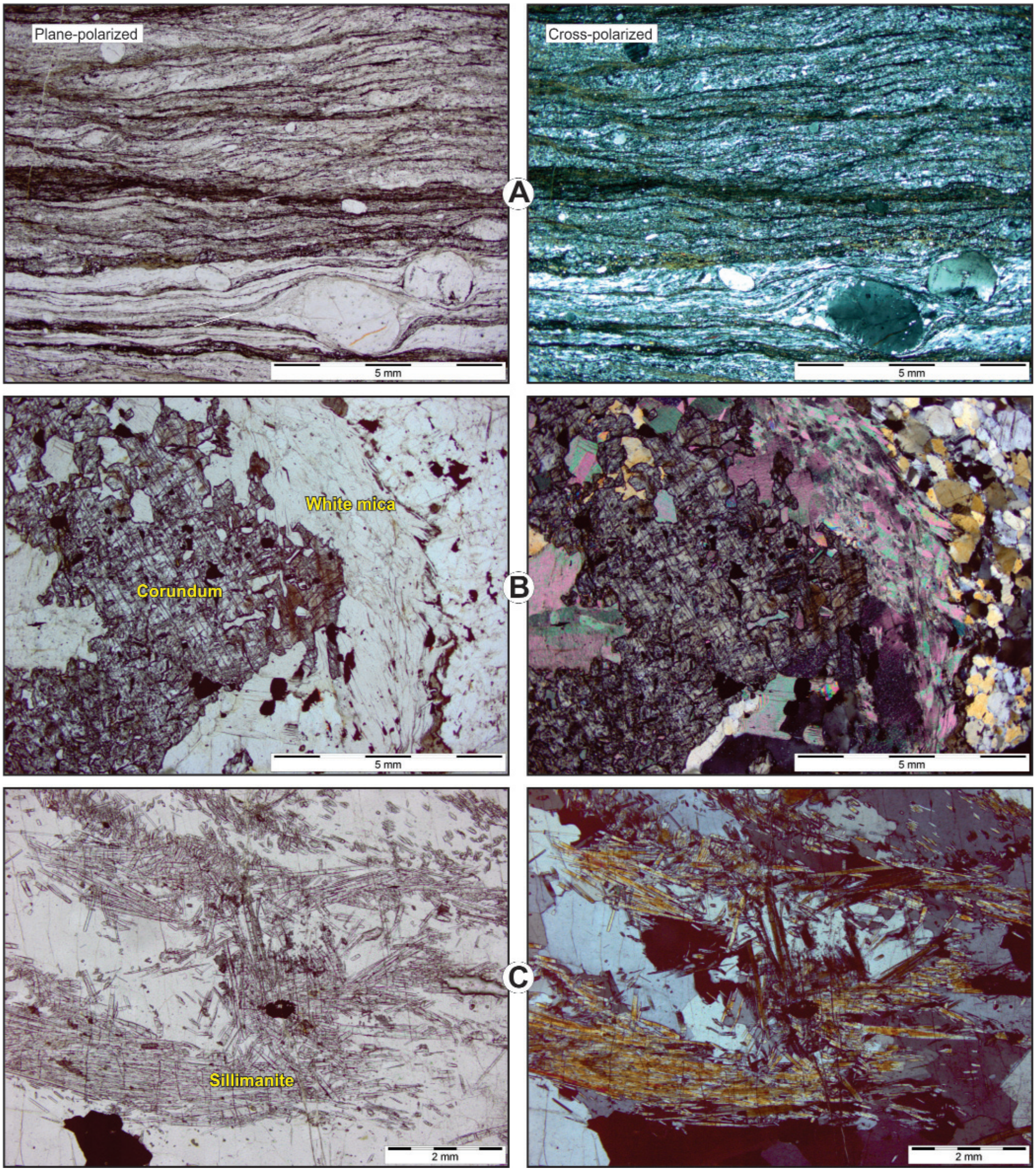
Sample No	Rock Type	Age Basis	Terrane	eNdt	T_{DM} (Ga)	Reference
CG84-475	Pelitic gneiss, biotite–sillimanite–garnet (Saddle Island)	1654 +29/-28 Ma U–Pb age ((Schärer and Gower, 1988; same sample)	Hawke River	0.13	2.07	Scharer (1991)
SPM86-09A	Pelitic gneiss, sillimanite, biotite (Paradise Metased. gn. Belt)	1638 +11/-3 Ma U–Pb metamorphic age (Kamo *et al.*, 1996; same locality)	Hawke River	2.32	1.91	Prevec *et al.* (1990)
SPM86-10B	Pelitic gneiss, musc, garnet, biotite (Paradise Metased. gn. Belt)	Regional correlation with Paradise Arm metasedimentary Gneiss belt	Hawke River	2.62	1.89	Prevec *et al.* (1990)
CG03-354F	Pelitic gneiss, biotite–sillimenit–garnet (Lake Melville terrane)	Regional correlation (same locality as RH-017)	Lake Melville	-2.65	2.29	Moumblow (2014)
RH-017	Psammite/granite, garnet, biot (within Lake Melville terrane granitoids)	Regional correlation (Gower inference)	Lake Melville	4.37	1.77	Hewitson (2010)
RH-020	Pelitic gneiss, sill, garnet, biotite (Gilbert Bay metasedimentary gneiss)	Regional correlation (Gower inference)	Lake Melville	4.27	1.78	Hewitson (2010)
RH-024a	Psammite granite, garnet, biot (within Lake Melville terrane granitoids)	Regional correlation (Gower inference)	Lake Melville	4.13	1.79	Hewitson (2010)
EC75-033	Granulite	Regional correlation as pre-Labradorian	Mealy Mountains	3.93	1.80	Hegner *et al.* (2010)
EC75-101	Granulite	Regional correlation as pre-Labradorian	Mealy Mountains	2.35	1.93	Hegner *et al.* (2010)
EC75-113	Granulite	Regional correlation as pre-Labradorian	Mealy Mountains	0.95	2.20	Hegner *et al.* (2010)
GN95-392	Quartzofeldspathic enclave, orthopyroxene, biotite	1914–1770 Ma U-Pb ages (Gower *et al.*, 2008; same sample)	Mealy Mountains	0.10	2.14	Creaser (unpublished)
RH-014A	Psammite, garnet, biotite (Mealy Mountains terrane)	Regional correlation (Gower inference)	Mealy Mountains	3.97	1.80	Hewitson (2010)
BS20	Pelitic gneiss, biot–sill–gnt (probable Labaradorian supracrustal -Dickin)	1650 Ma regional correlation (could be 1500 Ma)	Pinware	0.27	2.07	Dickin (2000)
CG00-154A	Gneiss enclave, amphibole, biotite (supracrustal?)	1771 ± 4 Ma U-Pb age (Gower *et al.*, 2008; same sample)	Pinware	2.95	1.85	Creaser (unpublished)

The first was by Brooks (1983), who investigated a suite of biotite–sillimanite–garnet–pelitic gneisses near Saddle Island, in the southern part of Sandwich Bay. Six samples were analyzed and define a poor errorchron of 1445 ± 220 Ma (I_{Sr} = 0.7048). The result is too imprecise for rigorous interpretation. At best, the only observation that can be made is that the nominal age is consistent with a late Pinwarian thermal imprint. The samples come from the same area as sample CG84-475 (which yielded a U–Pb metamorphic age of 1654 +29/–28 Ma) and was also analyzed for Sm–Nd isotopic data. Two of the Rb–Sr samples (CG81-148A, C) are from the same site as CG84-475. Calculating individual initial Sr ratios for the six samples using an 1800 Ma time of formation yields I_{Sr} = 0.69057 to 0.70197. Rb–Sr isotopic data for sample CG84-475 were also reported by Schärer (1991), yielding $I_{Sr(t)}$ = 0.70148 (t = 1654 Ma). These values deny any long crustal residence time for the source material.

The second study was carried out by Prevec *et al.* (1990) from localities 55 km to the southeast. Samples of biotite–sillimanite–garnet–pelitic-gneiss were subdivided into three groups; amphibolite facies (4 samples), mylonitized amphibolite facies (five samples from locality CG85-310; Photomicrograph 7.1A) and granulite facies (five samples). Note that two samples (SPM86-09A, 09B) come from locality CG85-148, used for U–Pb and Sm–Nd investigations. Excluding the granulite-facies samples, an age of 1629 ± 90 Ma (I_{Sr} = 0.7043) was determined, whereas including them gave an age of 1663 ± 64 Ma (I_{Sr} = 0.7040). Calculating individual initial Sr ratios for the fourteen samples using an 1800 Ma time of formation yields $I_{Sr(t)}$ = 0.70105 to 0.70371. Prevec *et al.* (1990) concluded that their results dated metamorphism, and pointed out that, as the mylonitized rocks fall on the same linear array as the other samples, Labradorian mylonitization is implied. The Rb–Sr data show the same resetting contrast as the U–Pb data between sites CG84-475 (close to Lake Melville terrane) *vs.* CG85-148 (central Hawke River terrane), implying a severe (Grenvillian) resetting event in the Lake Melville terrane that the interior Hawke River terrane did not experience.

7.2.2.3 K–Ar and Ar–Ar

Two results are available, neither shown on Figure 7.1. One is a K–Ar biotite date of 1175 ± 40 Ma, recalculated from the originally reported 1165 ± 40 Ma for biotite hornblende gneiss (Wanless *et al.*, 1972; sample GSC70-137). The site has been visited by the author who mapped the rock as psammitic gneiss derived from an arkosic metasediment. The age is interpreted in terms of partial Ar loss during Grenvillian metamorphism (Section 22.4). The other is a K–Ar biotite date of 1105 ± 62 Ma (recalculated from an originally reported 1095 ± 62 Ma) for a biotite paragneiss (Lowdon, 1961; sample GSC60-146). The latitude–longitude co-ordinates place it in the Lake Melville terrane about



Photomicrograph 7.1. *Mylonitized pelitic gneiss in Paradise metasedimentary gneiss belt, and pelitic gneiss enclaves in Earl Island intrusive suite, Hawke River terrane. A. Mylonitized pelitic gneiss. One of several samples from this site (CG85-310G) collectively yielding a Labradorian Rb–Sr WR age result, hence denying severe Grenvillian deformation, B. Pelitic gneiss enclave in Earl Island intrusive suite showing part of a corundum poilikoblast wrapped in white mica (margarite?) (CG85-167B), C. Pelitic gneiss enclave in Earl Island intrusive suite showing sillimanite-rich gneiss (CG85-515B).*

35 km south of Sandwich Bay, where high-grade metasedimentary gneiss has been mapped by the author. The age is roughly consistent with the time of Grenvillian metamorphism in this area.

Farther north, Wanless *et al.* (1972) reported a K–Ar biotite date for ‘paragneiss’ in the Groswater Bay terrane (sample GSC70-135, collected by I.R. Stevenson). The rock at this locality was examined by the author and considered to be orthogneiss. K–Ar biotite and hornblende dates were also obtained from another sample submitted for analysis by Stevenson, but proved to be anomalously old and the dates were never published by the Geological Survey of Canada. They are included on Gower’s (2010a) Groswater Bay map region, however. The rock at the site was mapped as paragneiss by Stevenson, but is considered to be syenite to alkali-feldspar syenite by the author.

7.3 FIELD AND PETROGRAPHIC CHARACTERISTICS

Field characteristics of the high-grade metasedimentary gneisses are addressed on a terrane-by-terrane basis, and by specific areas within each terrane. Representative stained slabs are presented in Appendix 2, Slab images 7.1, 7.2, 7.3 and 7.4, which more-or-less correlate with the Groswater Bay, Hawke River, Lake Melville and Mealy Mountains terranes, respectively.

Although the distribution of metamorphic minerals is reviewed for each terrane, the author judged that figures showing their distribution for the whole region are helpful, so such maps (for pelitic gneiss) are included (Figures 7.2 and 7.3).

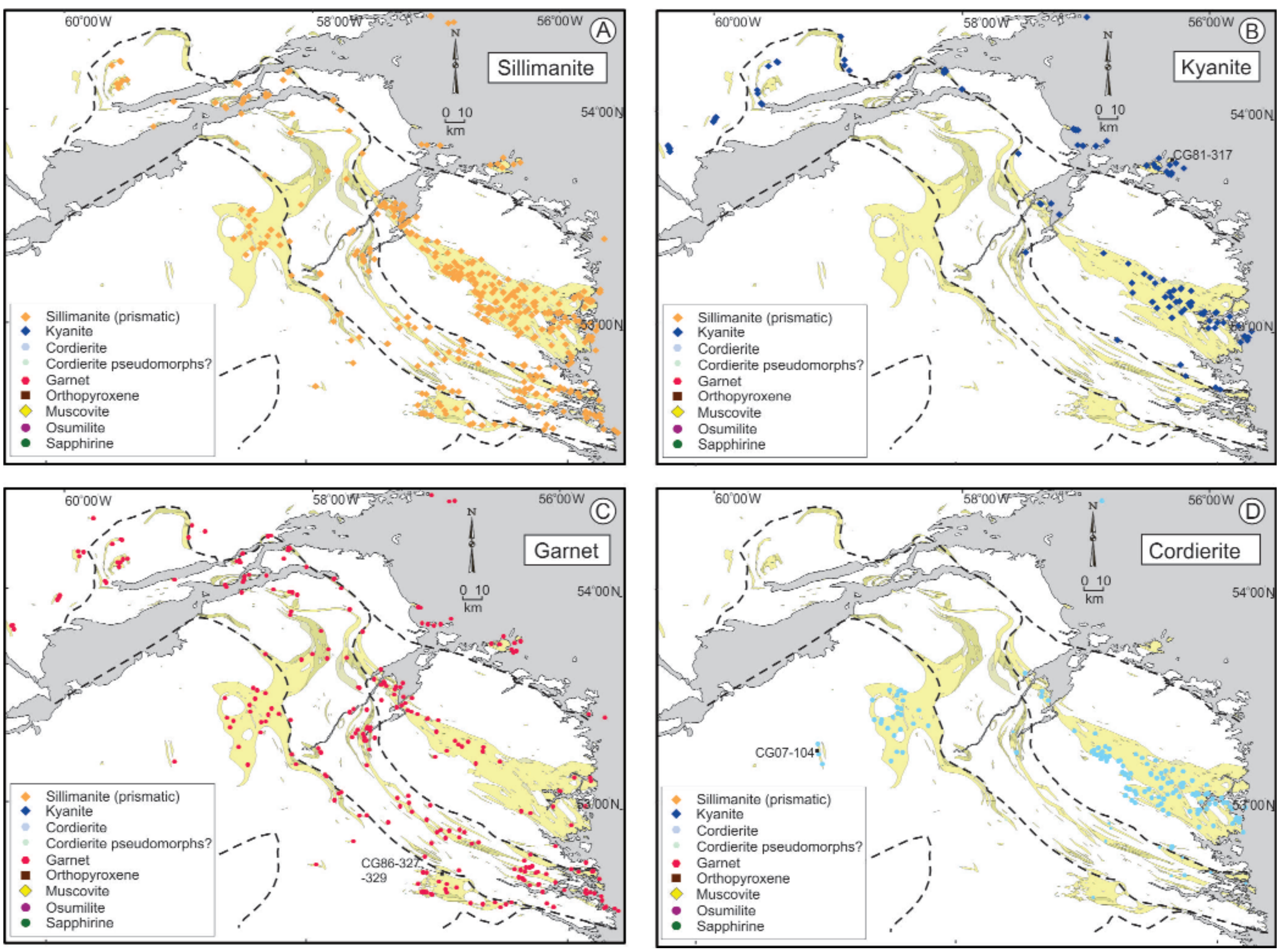


Figure 7.2. *Distribution of selected metamorphic minerals in late Paleoproterozoic pelitic gneiss. A. Sillimanite, B. Kyanite, C. Garnet, D. Cordierite.*

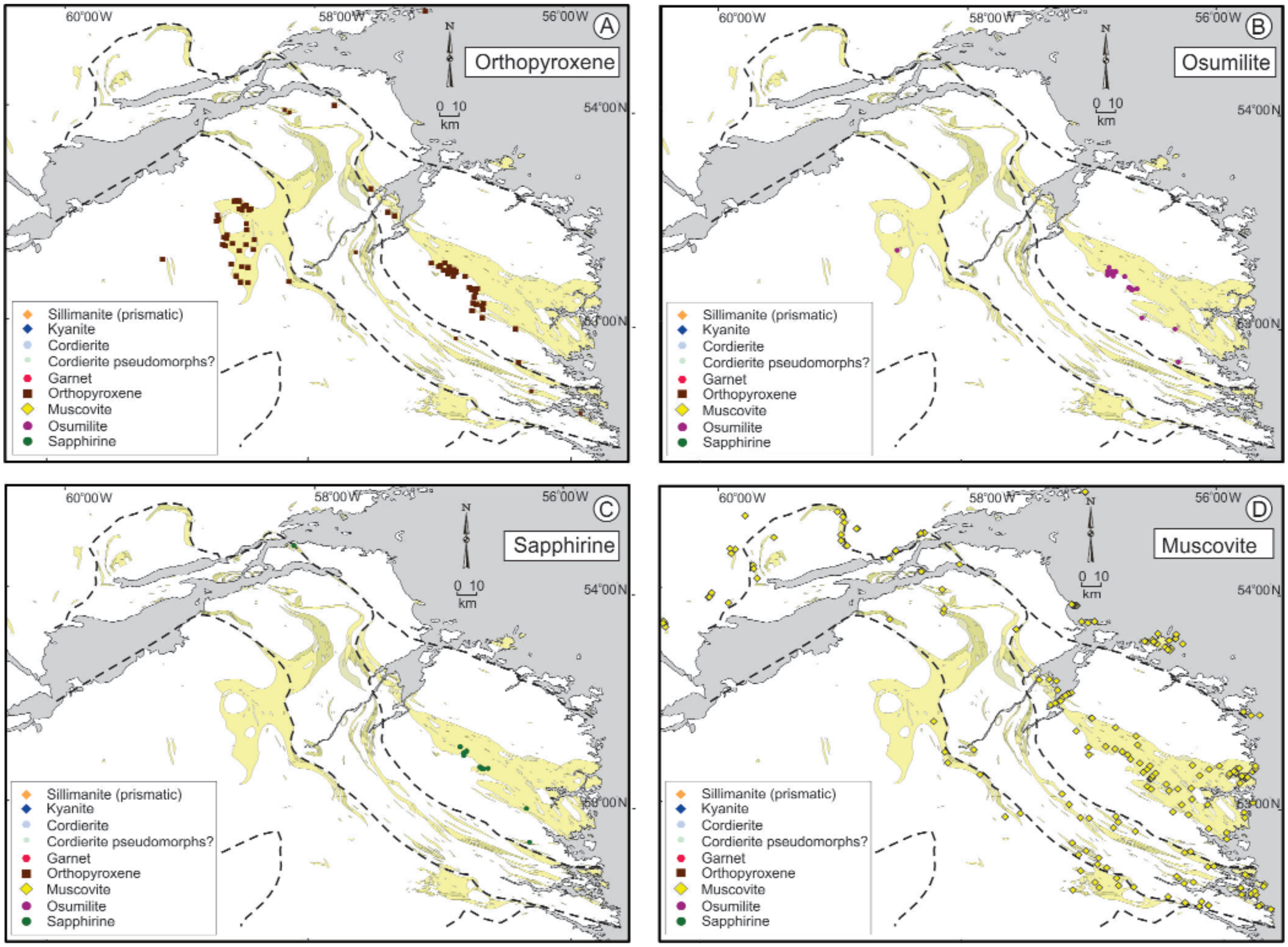


Figure 7.3. *Distribution of selected metamorphic minerals in late Paleoproterozoic pelitic gneiss. A. Orthopyroxene, B. Osumilite, C. Sapphirine, D. Muscovite.*

7.3.1 GROSWATER BAY TERRANE

Field characteristics are described for groups of outcrops of high-grade metasedimentary gneiss from west to east. The pattern of outcrops suggests that these represent generally narrow zones rarely more than a few kilometres wide but several kilometres long, orientated parallel to the prevailing structural trend in the areas in which they occur.

7.3.1.1 Mulligan Bay (P_{3A}sp)

The westernmost clusters of high-grade metasedimentary gneiss in the Groswater Bay terrane are located in two areas, 34 km west of Mulligan Bay and 16 km northwest of Mulligan Bay (Figure 7.4). They were mapped by Erdmer (1983, 1984) as rafts within granitoid orthogneiss, each raft being over 7 km long and about 1 km wide. Erdmer describes them as distinctive, pink to white, granoblastic, finely migmatite-banded, kyanite-bearing quartzofeldspathic biotite gneiss (his Unit Hgd-ky).

A large collection of thin sections (14 sections from 9 of the 11 outcrops included in the unit) establishes without question that all are pelitic gneiss. All contain kyanite, except PE82-087, which is otherwise similar, especially in that it contains muscovite and garnet, as do most of the other thin sections from these two areas. The kyanite occurs as blue-green crystals up to 1 cm long or as recrystallized fine-grained aggregates.

7.3.1.2 Main River (P_{3A}sp, P_{3A}ss)

Metasedimentary gneiss in the Main River area was mapped by Gower (1986). It occurs in two clusters, situated west of Main River and exposed along the shores of Main River itself (Figure 7.4), plus a few minor occurrences scattered elsewhere. It should be noted that Gower *et al.* (2008a) repositioned the Groswater Bay terrane–Lake Melville terrane boundary to coincide with a major thrust located to the west and north of another thrust-defined boundary used in previous interpretations (*e.g.*, Gower, 1986). The 'Main River itself' occurrences are thus transferred to the Lake Melville terrane. Note that it was already identified as a sep-

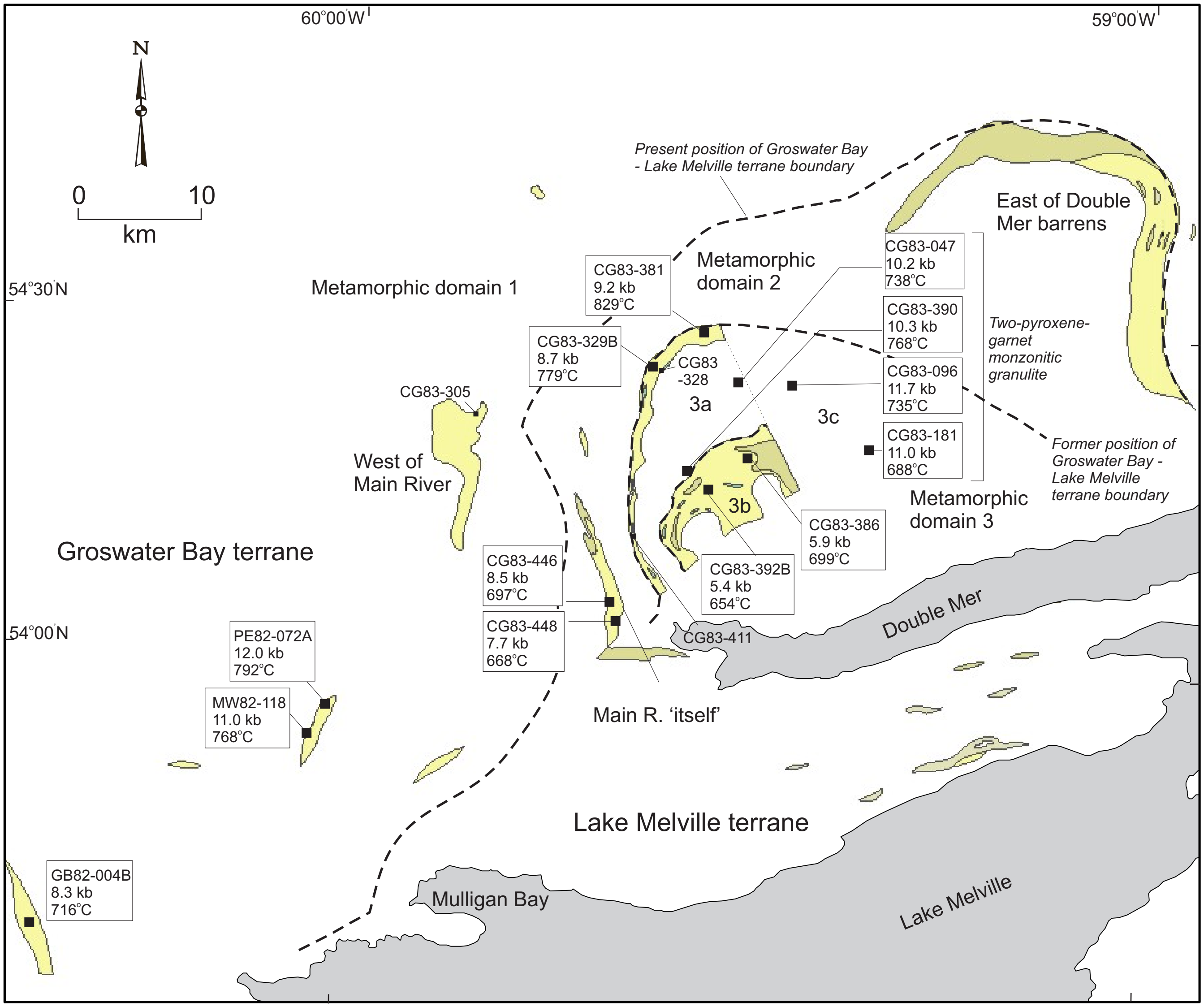


Figure 7.4. *Distribution of late Paleoproterozoic supracrustal rocks north of Lake Melville, also depicting metamorphic domain classification and locating geothermobarometric sites and results obtained from them (*cf. *Gower and Erdmer, 1988). 3a, 3b, and 3c are subdivisions of metamorphic domain 3. Apart from pelitic gneiss results, those four monzonitic granulites are also included.*

arate structural entity named by Gower (1984, 1986) as the Partridge Point domain. Both interpretations have some merit and the issue is more related to conceptual shortcomings in depicting complex structural interfaces as single lines, rather than one or other locus having overriding merit. Distribution of key metamorphic minerals in pelitic gneiss in the area is given in Figures 7.2 and 7.3.

The occurrences 'west of Main River' are grey-weathering, medium-grained, quartzofeldspathic pelitic and semipelitic schists that are interlayered, and gradational into, granitic gneiss. Some anomalously quartz-rich layers are present at CG83-404. Banding, where present, is defined by a melanosome of muscovite + biotite ± garnet, alternating with a coarser grained leucosome of quartz, plagioclase and microcline (Plate 7.1A). Locally the leucosome is deformed into complex ptygmatic folds (Plate 7.1B). The 'Main River itself' localities are pale-pink to black-and-white-weathering, medium grained and schistose to gneissose. They are characterized by a kyanite + muscovite + garnet melanosome in association with a coarser grained plagioclase + quartz + K-feldspar (microcline and/or perthite) leucosome, although not all rocks have a separately developed melanosome and leucosome. Biotitic selvedges and a phlebitic appearance suggest close to *in-situ* partial melting. Kyanite is glassy green and typically found as recrystallized

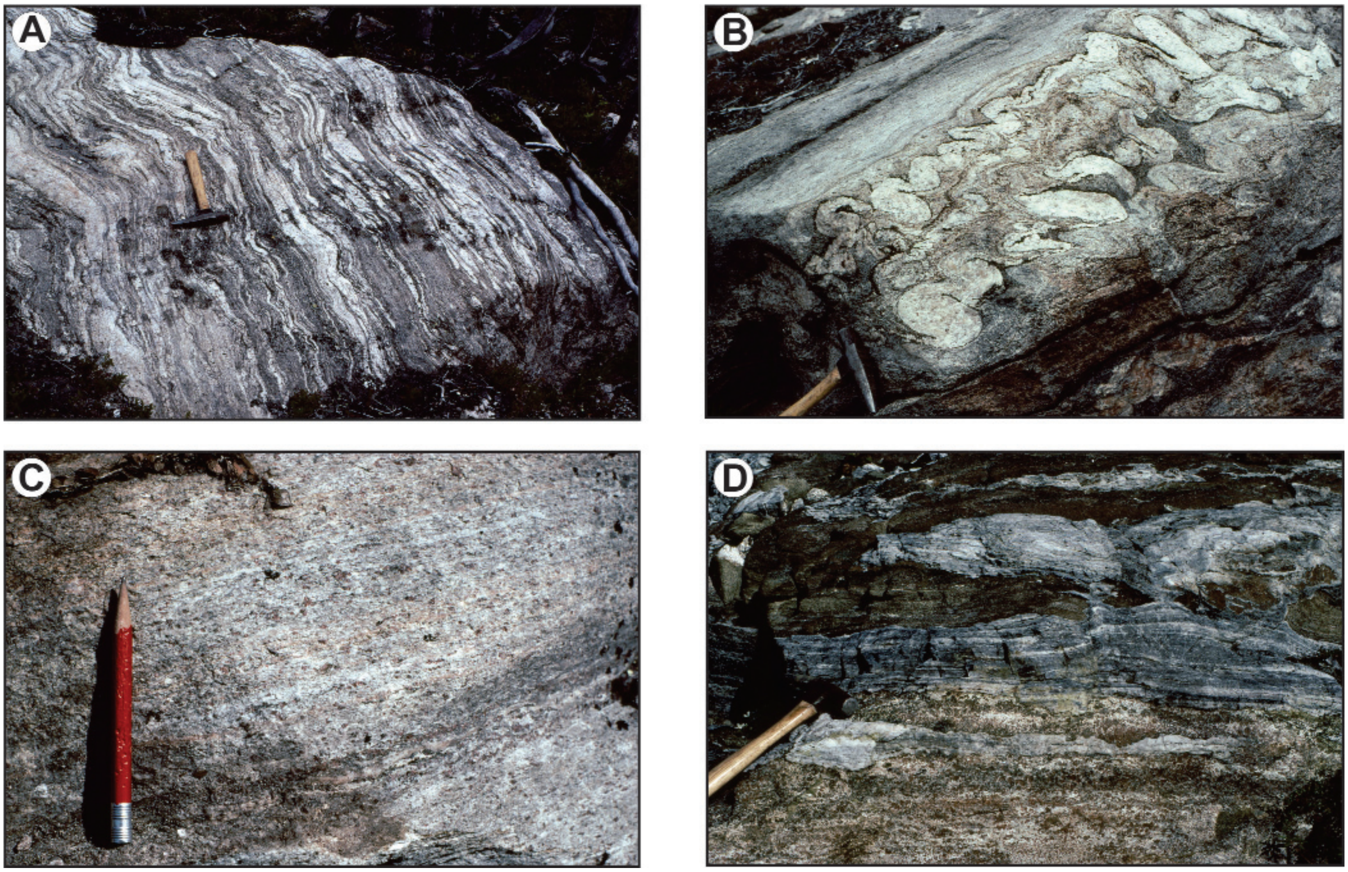


Plate 7.1. *Paleoproterozoic metasedimentary gneiss from the Double Mer region. Images A and B referred to in Section 7.3.1.2 and images C and D in Section 7.3.5.2. A. Muscovite-rich pelitic gneiss with interlayered white granitic leucosome (CG83-305), B. Detail of same locality as in A showing complex ptygmatic folds in leucosome (CG83-305), C. Poorly banded kyanite-bearing pelitic gneiss (CG83-328), D. Interbedded calc-silicate (dark-green) and quartzite (pale-grey) layers (CG83-411).*

clusters. It is only present in the 'Main River itself' occurrences. Outside these two clusters, occurrences are too small and inadequately exposed to assess them fully. Some are ocherous-weathering, due to associated pyrite and other altered sulphides.

In conjunction with mineral assemblages found in associated rock types, Gower (1986) and Gower and Erdmer (1988) assigned the non-kyanite bearing rocks to metamorphic domain 1, and those with kyanite to metamorphic domain 2. Gower (1986) observed that, in domain 2, kyanite preferentially co-exists with either garnet or muscovite, which he took to be evidence of the reaction (Holdaway, 1980):

Muscovite + *Garnet* = *Kyanite* + *Biotite* + *Quartz*

in which either muscovite or garnet survived, depending on which reactant was exhausted first.

7.3.1.3 East of Double Mer Barrens (P_{3A}sp, P_{3A}ss)

The metasedimentary gneiss east of the Double Mer barrens is exposed in a north-trending belt about 20 km long by 2.5 km wide (Figures 7.1 and 7.4). It was outlined by Gower (1986) and links up at its southern end with kyanite-bearing pelitic gneiss earlier mapped by Gower *et al.* (1983a). The north end of the belt is concealed by glacial deposits but, on the basis of a strong positive magnetic anomaly, is interpreted to swing round into a westerly trend. Two data stations were established in the west-trending part, but rocks at both sites may be glacial boulders rather than *in-situ* bedrock. Elsewhere, rock exposures are mixtures of psammitic and pelitic schist and gneiss. Quartz-rich gneiss is present at CG80-571 and CG83-054. Amphibolite is present at CG83-069, in association with muscovite–garnet diatexite. Pegmatite, containing muscovite books up to 3 cm across, intrudes muscovite-rich gneiss at CG83-037.

Most of the rocks contain muscovite, which appears to co-exist stably with kyanite and K-feldspar in some samples and to be retrograde in others. Garnet is seen in 3 of the 9 thin sections, and was recorded in the hand sample of a fourth (*cf.* Figures 7.2 and 7.3).

7.3.1.4 Tom Luscombe Brook to Black Island (P_{3A}sp, P_{3A}ss, P_{3A}sq)

The supracrustal rocks in this group are aligned as a west-northwest-trending discontinuous belt about 30-km long north of Groswater Bay (Figure 7.5). They contrast from the previously addressed metasedimentary gneiss in lacking significant pelitic gneiss, being, instead, mostly psammitic, accompanied by some rocks possibly having a felsic volcanic protolith. They are also at a lower metamorphic grade.

North of Tom Luscombe Brook. The rocks north of Tom Luscombe Brook are described in field notes as feldspathic gneiss, feldspathic quartzite and quartzite. The rocks are white- or rusty-weathering, granoblastic, well foliated and banded, and, in addition to felsic minerals, contain biotite, muscovite, garnet, hornblende, epidote and pyrite. The presence of quartzite and the very muscovite-rich nature of the schist at DB79-214 (examined petrographically) are the principal criteria for designating these rocks as supracrustal.

West of Black Island. The occurrences on the mainland west of Black Island are grey, fine- to medium-grained, banded, quartz-rich, biotite ± hornblende gneiss with a few quartzofeldspathic segregations. Thulite (a rose-coloured, manganiferous epidote group mineral) was recorded at CG79-897.

Four thin sections are available from this area (oddly, from a sampling viewpoint, all from a single outcrop; two were prepared from Stevenson's sample (SG68-058.1 and .2) and two from the author's sample (CG79-637.1 and .2). All have plagioclase, microcline, quartz and green biotite. In addition, SG68-058 has a dark-green (sodic) amphibole and aegerine, which are lacking in CG79-637, having garnet instead. Note that it is from this outcrop that Stevenson collected a sample for K–Ar dating, which yielded the earlier-mentioned 1175 ± 40 Ma K–Ar biotite date.

Black Island. The supracrustal rocks here show considerable variation in both protolith and its subsequent response to metamorphism. In the lowest grade condition displayed, primary sedimentary layering is preserved. The rock at this locality shows pale creamy-grey, medium-

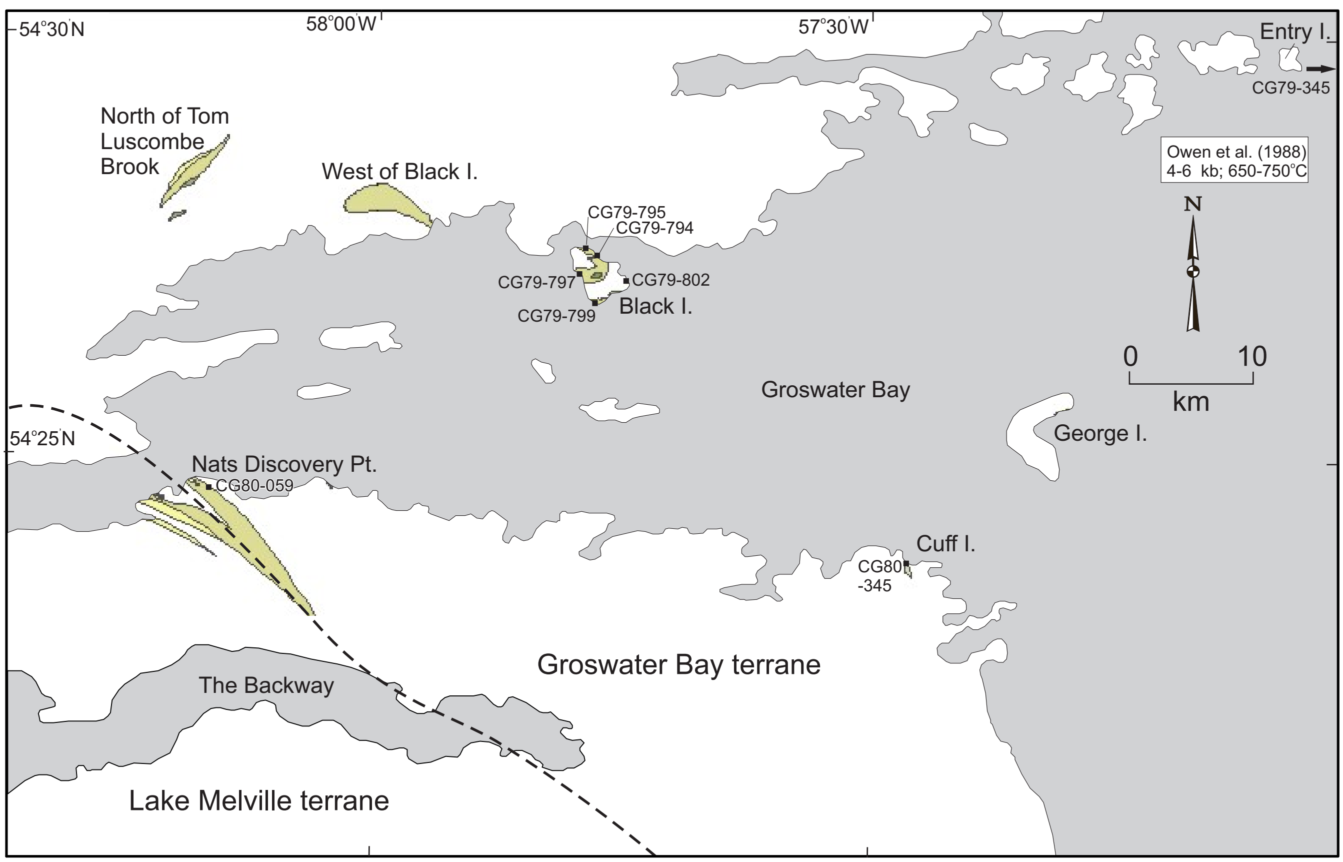


Figure 7.5. *Distribution of late Paleoproterozoic supracrustal rocks flanking Groswater Bay. Some localities too small to be shown.* See *text for discussion regarding possible younger age for supracrustal rocks on Black Island.*

grained clastic material, interstratified with mid-grey, fine-grained material. The interface between the two is marked by a narrow layer of dark-grey, very fine-grained material (Plate 7.2). At other localities on Black Island, the rocks were described as pale grey, buff, or creamy-weathering, continuously and evenly banded, so as to define layers that range from fine laminations to broader composite units roughly 3 m wide.

In several places, the rocks have a spotted appearance due to small hornblende poikiloblasts (Plate 7.3A). These form strings and elongate clusters and are enveloped in creamy white leucosome within the original rock paleosome. Sporadically, the hornblende poikiloblasts are much larger (up to about 4 cm) and the intervening leucosome all pervasive, reducing the paleosome to small remnant patches (Plate 7.3B). Garnet is also a common porphyroblastic mineral in some outcrops on Black Island.

Some of the supracrustal rocks on Black Island show incipient to extensive partial melting. In one example, continuously banded grey psammitic gneiss shows selective partial melting along particular layers. One particular layer, which is made up of multiple much narrower quartzofeldspathic layers, shows eye-catching boudinage (Plate 7.3C). In rocks having this metamorphic state, all vestiges of unequivocal primary stratification are obliterated, resulting in quartzofeldspathic gneiss of uncertain protolith. When initially seen, some of these were termed tonalitic gneiss by the author, although his field notes hedge on whether an igneous or metasedimentary protolith was considered most probable. With the advantage of greater experience, it now seems likely that at least some of the so-called tonalitic gneiss in this area has a supracrustal protolith.

Thulite was also seen on Black Island. It occurs in the cores of yellow-green, epidote-rich pods within the quart-

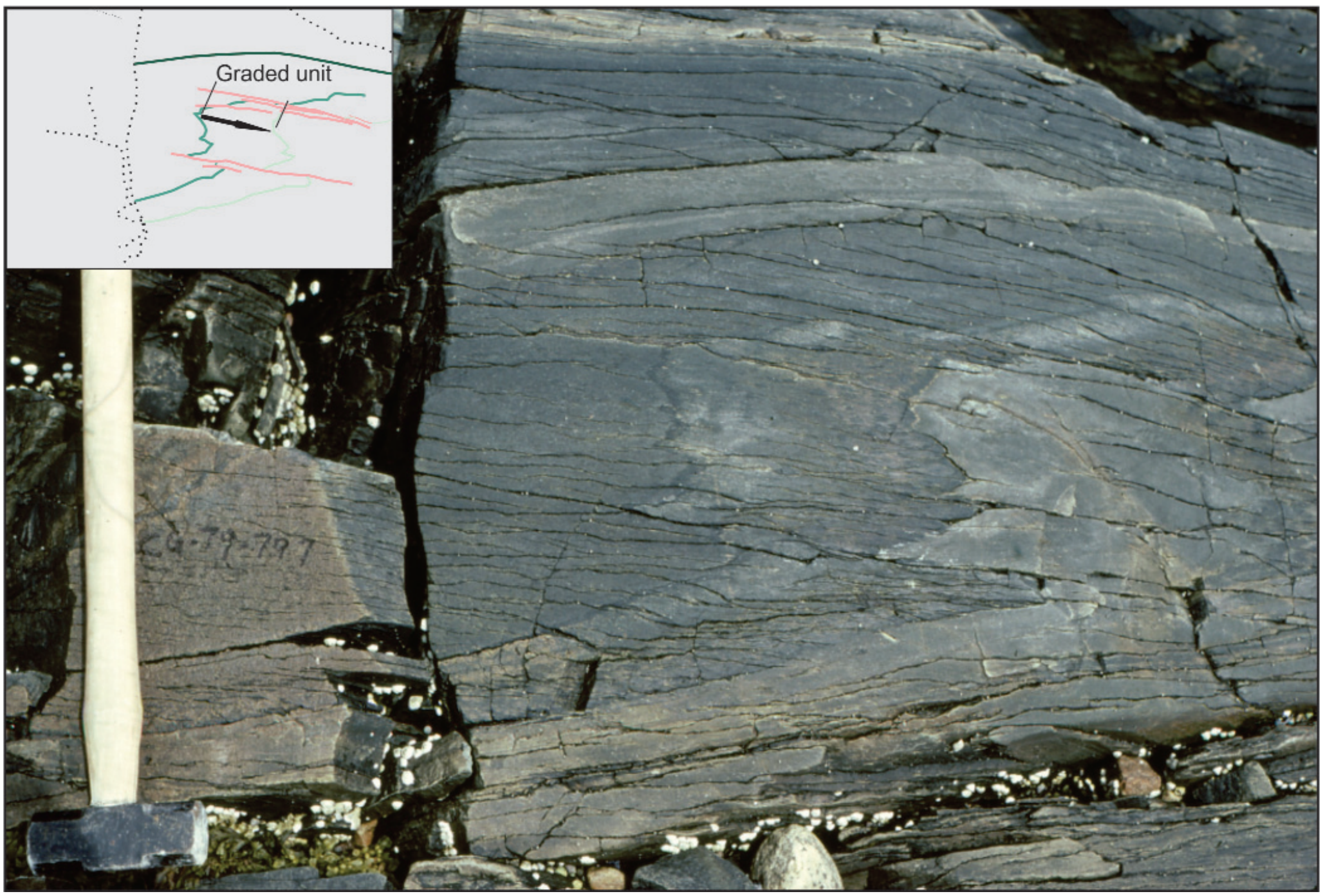


Plate 7.2. *This rock is interpreted as a pelitic–psammitic metasediment in which graded bedding is preserved. The dark-grey, very fine-grained material marks the top of one graded unit that was then overlain by the basal creamy-grey, medium-grained material of the next unit. The sedimentary layers have been folded into an S-fold and show a well-defined axial-planar cleavage. The cleavage has been deflected, or is lacking, in the coarser grained material. The fracture cleavage has acted locally as small faults that have displaced bedding. Inset diagram provides clarification (CG79-797).*

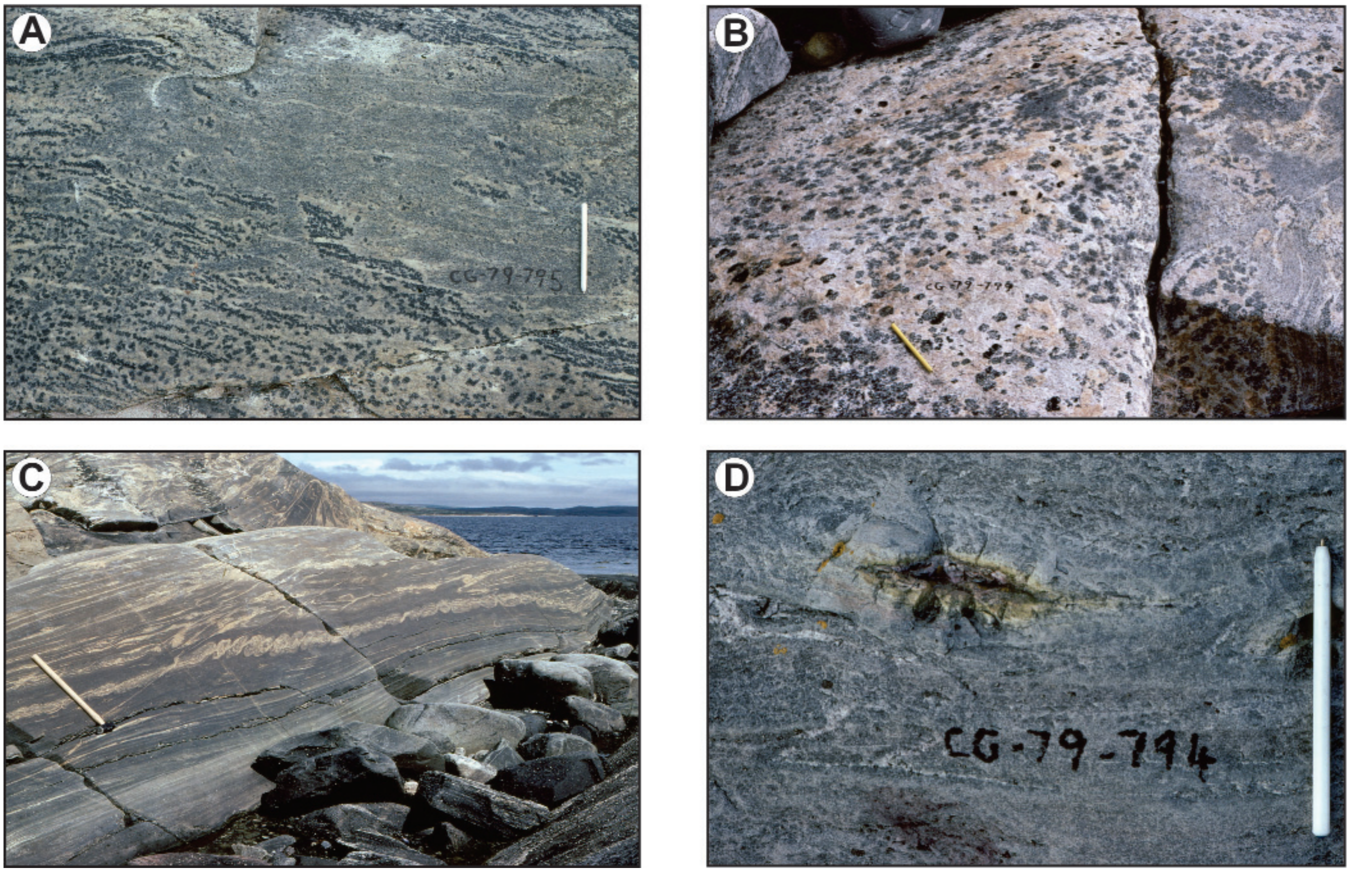


Plate 7.3. *Some features of psammitic gneiss on Black Island, north side of Groswater Bay. The rocks are interpreted as part of the same unit as that shown in Figure 7.2. A. Hornblende poikiloblasts enveloped in creamy-white leucosome with remnant psammitic paleosome preserved (CG79-795), B. More advanced development of hornblende poikiloblasts, with more leucosome and small remnants of paleosome (CG79-799), C. Psammitic(?) gneiss with eye-catching boudinage of leucosome layer (CG79-802), D. Thulite (manganiferous epidote-group mineral) in core of epidote-rich pods in psammitic gneiss (CG79-794).*

zofeldspathic metasedimentary rocks (Plate 7.3D). A poor sample of thulite from data station CG79-794 was identified as an epidote-group mineral by X-ray diffraction (Appendix 2 of Gower, 2010c).

Several small pits related to Cu exploration are present in the centre of Black Island. The earliest description known to the author is by Halet (1946) who noted that they had been dug 40 years earlier. The occurrences have been visited and described by the author (Gower, 2010c).

It was mentioned at the start of this section that these supracrustal rocks are unlike the pelitic gneisses found elsewhere in the Groswater Bay terrane. They were distinguished from the high-grade metasedimentary gneiss by Gower and Owen (1984) and noted to include greywacke and siltstone (no advances have been made since then). Gower and Owen (1984, page 682) wrote that "… it is not clear whether the differences in metamorphic grade are attributable to regional metamorphic variations or because the lower grade rocks postdate the high-grade metamorphism expressed in the pelitic paragneiss. Because regional metamorphic variations would need to be drastic we favour the second explanation". In several respects, a close analogue is the supracrustal rocks at Battle Island (Gower, 2008, 2009; *see* Section 16.3) some 275 km to the south-southeast, but whether or not these rocks might be coeval remains a tantalizing speculation. The supracrustal rocks at Battle Island have a maximum depositional age of *ca.* 1200 Ma (Kamo *et al.*, 2011). Noteworthy comparisons are overall similarity of composition, the preservation of primary sedimentary stratification, and the abundance of hornblende poikiloblasts.

7.3.1.5 Nats Discovery Point (P_{3A}ss?)

The rocks at this locality (CG80-059) are severely mylonitized (Figure 7.5). They are mid-grey-weathering, continuously and evenly banded, and fine- to medium-grained. Narrow amphibolite bands and lenses are present. Although shown on Gower's map (2010a; Rigolet map region) as having a psammitic protolith, this is far from certain, especially given their intense mylonitization. Alternatively, the lack of quartz and K-feldspar suggests a dioritic to mafic protolith.

A thin section contains plagioclase, bronzy biotite, hornblende, garnet, an opaque oxide and sulphide, allanite, epidote and scapolite.

7.3.1.6 Cuff Island (P_{3A}vf/P_{3A}ss)

The rocks on the mainland near Cuff Island (Figure 7.5) were depicted on Gower's (2010a) Groswater Bay map region as psammitic to quartzitic gneiss. This label, however, does not acknowledge the possible existence of conglomeratic or agglomeratic rocks at the site (CG80-345; Plate 7.4A, B). Whereas most of the outcrop comprises mid-grey-weathering, fine-grained, quartz-rich rocks showing a fairly continuous lamination (not mylonitic) that may represent primary bedding, the 'conglomeratic' zones display a wide assortment of angular to rounded fine- to coarse-grained felsic to mafelsic 'clasts'. To the author, the variability in shape, texture and composition of the clasts makes other potential protoliths unattractive (such as being the product of multiple granitoid injection and deformation – the most likely alternative). Recall that the presence of primary sedimentary structures in supracrustal rocks at Black Island (Plate 7.2) demonstrates that preservation of such structures is tenable in the region.

A thin section (CG80-345) of homogeneous material from the outcrop was interpreted to be a meta-arkose. It contains plagioclase, microcline, rounded quartz, buff-orange biotite, an opaque oxide and minor secondary white mica, chlorite, epidote and prehnite.

7.3.1.7 George Island (P_{3A}ss)

An outcrop of white-weathering, well-banded muscovite–biotite–garnet gneiss was mapped on George Island by the author (CG80-701; *cf.* Gower, 2010a; Groswater Bay map region; Figure 7.5). The outcrop was only examined briefly during a short helicopter stop and the areal extent of the rock is not known. Muscovite was recorded to be abundant in outcrop.

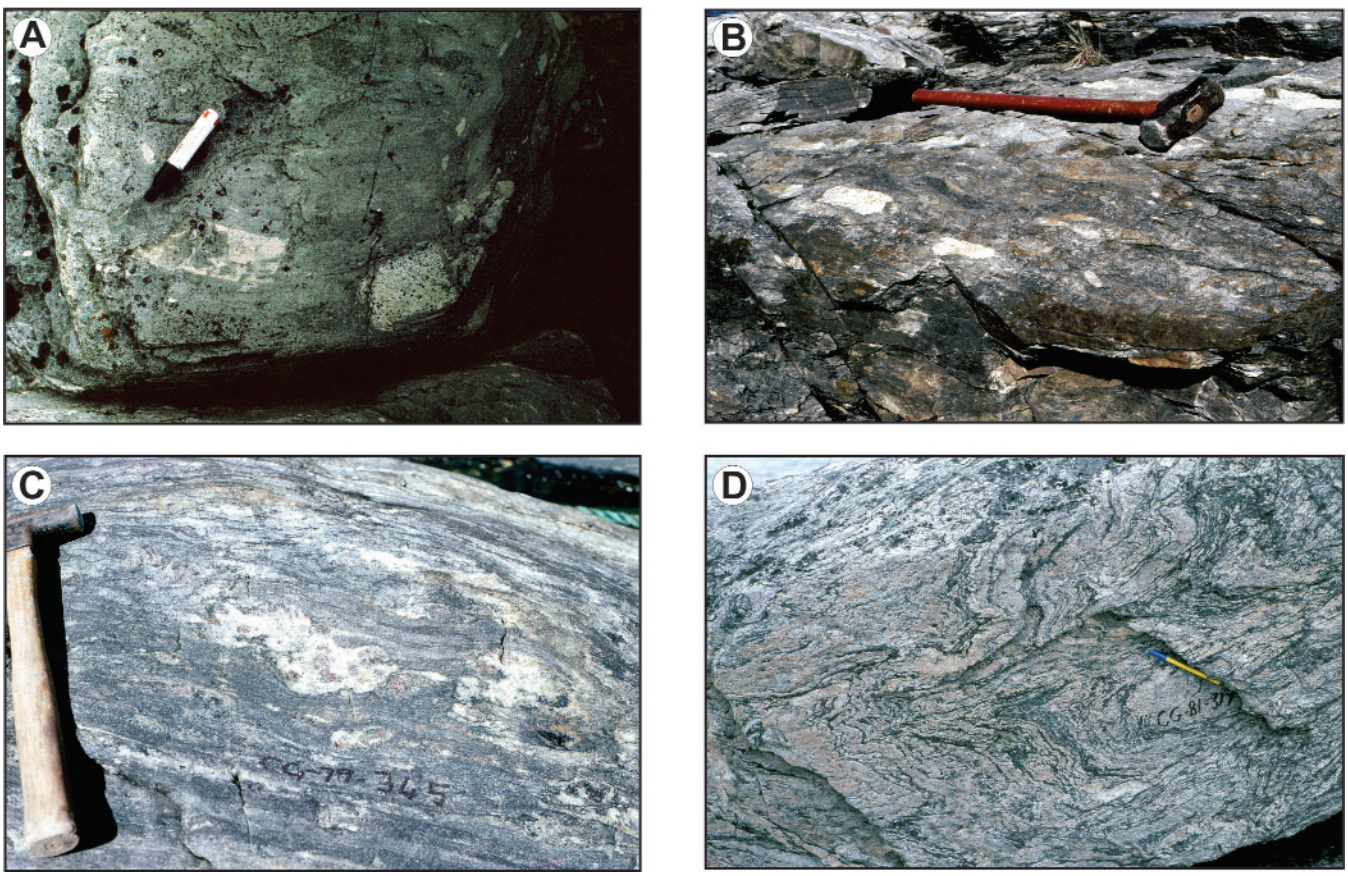


Plate 7.4. *Rocks derived from supracrustal protoliths in the eastern Groswater Bay terrane. A. Rock possibly having a conglomeratic protolith, displaying clasts of fine- and coarse-grained felsic material (CG80-345), B. Same outcrop as in 7.4A, also showing possible conglomeratic/agglomeratic texture (CG80-345), C. Garnet–sillimanite pelitic gneiss of the White Bear Islands granulite complex (CG79-345), D. Typical kyanite-bearing pelitic gneiss from the Curlew Harbour area (CG81-317).*

In thin section, the sample is seen to contain plagioclase, microcline, quartz, muscovite, buff-green biotite, garnet, an opaque oxide, sulphide (*cf.* pyrrhotite), apatite, allanite, secondary epidote and chlorite. The abundance of muscovite makes a pelitic supracrustal protolith most probable.

7.3.1.8 Entry Island and Correlatives (on map as P_{2C}sp)

Stevenson (1970) was the first to recognize metasedimentary gneiss at Entry Island (Figure 7.5). He included the island as part of his Unit 6 (feldspar–quartz–hornblende–biotite granitic gneiss), but, in his field notes, he describes the rock at the site as almost pure quartzite in which relict bedding is preserved.

Of special interest is a thin section from the locality (SG68-181A). On the original Geological Survey of Canada punched-hole data card (signed by S.K. Singh, who presumably did the petrographic examination), the minerals listed are quartz (70%), pyroxene (10%), sillimanite (15%), cordierite (5%), muscovite (trace), garnet (trace). The thin section was later re-examined by J. Bourne, during his study of metamorphism in the eastern Grenville Province (Bourne, 1978). He listed quartz, kyanite, muscovite and an opaque mineral, and included the kyanite locality on the Metamorphic Map of the Canadian Shield (Fraser *et al.*, 1978, Map 1475A). The author has also examined the same thin section, identifying quartz, sillimanite, kyanite, an opaque mineral, rare tourmaline(?) and retrograde muscovite. Orthopyroxene was equivocally identified, but no cordierite or garnet seen. The kyanite occurs as high-relief, bladed crystals and there is no doubt regarding its identification. The author has visited the island, collecting samples of 'muscovite schist' that were subsequently examined petrographically (CG79-963, CG81-763 – the same site despite two different data station labels). Sample CG79-963 contains quartz, microcline, sillimanite and muscovite (not retrograde), whereas sample CG81-763 contains quartz, microcline, plagioclase, muscovite (not retrograde), garnet, an opaque oxide and secondary epidote. No kyanite was seen in either.

In addition to the Entry Island site, high-grade pelitic schist and gneiss were found during 1:100 000-scale mapping at several other sites as rafts within orthogneiss farther east in the Smokey archipelago (*cf.* Gower, 2010a; Groswater Bay map region; Figure 7.2; Plate 7.4C).

Two thin sections (CG79-345, CG79-347) both contain garnet and sillimanite and CG79-345 also contains relict cordierite.

The area was mapped in detail by Owen (1985) and Owen *et al.* (1988), who included the metasedimentary gneiss as part of the (then, newly defined) White Bear Islands granulite complex. Pelitic rocks were subdivided into three subunits, namely 1A – sillimanite-bearing diatexite lacking relict pelitic paleosome (quartz, microcline, plagioclase An_{32}, biotite, sillimanite and garnet), 1B – the kyanite-bearing locality described above, and 1C – cordierite metatexite characterized by well-defined neosome and paleosome (minerals are plagioclase $An_{32–36}$, quartz, antiperthitic K-feldspar, biotite, garnet, cordierite, orthopyroxene, sillimanite, magnetite, ilmenite and green spinel).

Collectively, the status of the rocks as high-grade quartzitic to pelitic gneiss (± kyanite, ± sillimanite, ± garnet, ± cordierite, ± muscovite) can be considered reliably established. Textural relationships between the various mineral and the reactions indicated have been described in detail by Owen (1985), and summarized by Owen *et al.* (1988). A few of the more obvious relationships are as follows.

Garnet encloses biotite ± quartz ± sillimanite and, elsewhere, is replaced by the same minerals, reflecting the reaction:

$$Garnet + K\text{-}feldspar + H_2O = Biotite + Sillimanite + Quartz$$

It is also replaced by cordierite, reflecting the reaction:

$$Garnet + Sillimanite + Biotite + Quartz = Cordierite + K\text{-}feldspar + H_2O$$

The coexistence of sillimanite with K-feldspar indicates the reaction:

$$Muscovite + Quartz = Sillimanite + K\text{-}feldspar + Vapour$$

Given the distribution of kyanite and sillimanite elsewhere in the Groswater Bay terrane (in its southern part), the presence of these aluminosilicates together in the Smokey archipelago is atypical. In view of the elusive Makkovik–Grenville province boundary location in the area, they might well have greater affinity with the Makkovik Province, rather than having close linkage with the high-grade pelitic gneiss farther south, although this would be a unique occurrence of kyanite in the Makkovik Province.

7.3.1.9 Cape Porcupine to Curlew Harbour (P_{3A}sp)

The best exposed and most extensive high-grade metasedimentary gneisses in the Groswater Bay terrane are in the Curlew Harbour area (Figure 7.1). The gneisses were mapped by Gower *et al.* (1982b). They underlie an area roughly 10 km wide and about 20 km long, but similar rocks occur along strike to the west-northwest as far as Cape Porcupine, so the length of the 'belt' (mostly underwater) could extend to about 60 km. The description below also applies to the western occurrences (to include those between Cape Porcupine and Pack's Harbour), except that these are more commonly strongly mylonitic.

The rocks are overwhelmingly pelitic gneiss, but grade into semi-pelitic gneiss locally. Very minor calc-silicate rock was recorded (MC77-235B). The rocks are strongly deformed, and north-verging recumbent folds are evident in several places. Only rarely are the rocks severely mylonitized, but an exception is site VO81-075, which is at the southern margin of the belt. The rocks include both schist and gneiss and are generally thoroughly migmatized, displaying complexly contorted fabrics. Most typical is an irregular, discontinuously banded, schlieric, knotty-looking appearance (Plate 7.4D). The rock comprises pale-grey paleosome, white and/or pink partial melt/neosome, and black restite/melanosome. Grain size varies from fine to medium in the paleosome and restite, to medium or coarse in the neosome. The neosome, which may reach up to 40% of the rock, forms pods, patches and irregular layers and grades into inhomogeneous diatexite. It is typically K-feldspar rich. The melanosome tends to form irregular layers and lenticular pods of black material comprising muscovite, biotite, garnet (mauve and red), magnetite and graphite. The restite also contains kyanite. Curiously, kyanite was only once recorded in the field (CG81-749), but is present in 27 of 31 thin sections – such was the unfamiliarity with the rocks at the time. Felted fibres of sillimanite were recorded at one locality as a retrograded product associated with garnet breakdown.

Concordant layers and pods of amphibolite, locally garnetiferous, are found throughout. The rocks are also discordantly intruded by metamorphosed mafic dykes. Mafic dykes present include the late Proterozoic Long Range dyke swarm (*see* Chapter 18).

The rocks are intruded discordantly by microgranite and pegmatite. The microgranite tends to be very pale-pink-weathering and, rarely, contains euhedral K-feldspar megacrysts. Alaskitic microgranite discordantly intrudes earlier microgranite. The pegmatite is commonly muscovite bearing, and may contain muscovite books up to 10 cm wide and 16 cm in diameter (*cf.* CG81-315, CG81-317, CG81-326). Rarely the pegmatite also carries tourmaline (CG81-316).

The southern flank of the Curlew Harbour metasedimentary gneiss shows a progressive change into finer grained, more homogeneous quartzofeldspathic material lacking the characteristic knotty appearance seen elsewhere. It has less muscovite and more chlorite, less neosome and less alaskitic microgranite, but more discordant K-feldspar-rich veins. The significance of this gradational change is left unexplained.

Thirty-one samples of pelitic gneiss were examined in thin section. All contain plagioclase, K-feldspar (microcline, perthite, or untwinned) and quartz, except for absence of K-feldspar in CG81-758, which is confirmed by a lack of K-feldspar in a stained slab. Muscovite is present in all but two thin sections (CG81-756, GF81-023) and olive-green to orange-brown biotite in all but one (GF81-015). Kyanite occurs as clusters of small, equant, recrystallized grains in 27 of the 31 thin sections (absent from CG81-325, CG81-758, GF81-015, VO81-075A). Its granular nature in these rocks undoubtedly contributed to its lack of recognition in the field. Only in thin sections CG81-312 and CG81-338 does it have its characteristic bladed form. Sillimanite was recorded in nine thin sections, and in all cases it appears to be a derivative mineral, rather than part of the stable high-grade assemblage. In CG81-299A, it occurs in clusters with biotite and some kyanite and may be pseudomorphous after cordierite (if so, this is the only instance in this group of rocks). Garnet was seen in 60% of the thin sections. Other minerals are an opaque oxide, zircon and rare apatite, monazite (possibly present in two thin sections) and allanite. The presence of kyanite apparently stable with K-feldspar (albeit subsequently recrystallized) indicates bathozone 6 conditions, which imply pressures and temperatures exceeding 7 kb and 650°C (Carmichael, 1978; Pattison *et al.*, 2005).

The high-grade metasedimentary gneiss in the Curlew Harbour–Cape Porcupine area has been investigated in detail by van Nostrand (1988), including geothermobarometric analysis (*cf.* Chapter 22). The typical assemblage is noted to be plagioclase + K-feldspar + quartz + biotite + kyanite + garnet granitic veins. In a petrogenetic grid, he concluded that the co-existence of kyanite + K-feldspar resulted from vapour-absent dehydration melting, which, in the K_2O–Al_2O_3–SiO_2–H_2O system, is expressed by the reaction:

Muscovite + Quartz = K-feldspar + Kyanite + Liquid

or, in the more embracing eight-component system K_2O–Al_2O_3–SiO_2–H_2O–Na_2O–MgO–FeO–CaO, as

Muscovite + Quartz + Biotite + Albite
= K-feldspar + Kyanite + Liquid

Assuming the activity of water (aH_2O) = 0.5, then minimum metamorphic conditions of 7 kb and 650°C are implied.

The importance of understanding textural relationships was addressed by van Nostrand (1988), who concluded, in conjunction with mineral chemistry, that garnet and matrix biotite (as opposed to biotite inclusions in garnet) both show consistent compositions and could be accepted as being in equilibrium. Some of the muscovite was inferred to be stable with kyanite, but elsewhere was determined to be retrograde from kyanite.

7.3.1.10 Sandy Bay (P_{3A}sp)

The Sandy Bay occurrences are situated at the southeast end of the Groswater Bay terrane (Figure 7.1) and similar rocks are present on Mark Islands, 4 km east of the mouth of Sandy Bay. The rocks were somewhat cursorily examined during 1:100 000-scale mapping (Gower *et al.*, 1986b; Gower, 2010a; Sand Hill River map region) and the only augmentation of information since then has been from petrographic studies.

The rocks are poorly to well-banded pelitic and semipelitic gneiss having a pink or creamy quartzofeldspathic leucosome and a grey muscovite biotite opaque mineral restite. Garnet is sporadically present in the restite.

Tourmaline was suspected at one outcrop (LC85-079), but was not found in any of the four thin sections examined. Sillimanite was recorded in the field at three localities (CG85-551, LC85-078, LC85-086), but, although thin sections of samples from all three localities were examined, sillimanite was only seen in CG85-551A. Kyanite was suspected at CG85-551, but not verified in a thin section from the locality. Pods and more continuous layers of amphibolite are present and suspected to be the remnants of former mafic dykes.

Felsic minerals seen in thin section are plagioclase, microcline and quartz, except for a lack of K-feldspar in LC85-086. Stained slabs confirm that K-feldspar is lacking in 5 of 8 slabs, suggesting that the rocks did not reach the threshold for the reaction:

Muscovite + Quartz = K-feldspar + Sillimanite + Liquid

Other minerals are muscovite (lacking in psammitic gneiss CG85-551B), biotite (olive-green to orange-brown) and garnet (lacking in LC85-078). Accessory minerals include an opaque oxide, apatite, zircon, monazite, rutile, allanite, sillimanite and epidote (not all minerals in every thin section).

7.3.2 HAWKE RIVER TERRANE; META-SEDIMENTARY REMNANTS IN EARL ISLAND INTRUSIVE SUITE

High-grade metasedimentary gneiss in the Hawke River terrane can be readily subdivided into: i) that within the Earl Island intrusive suite (EIIS), ii) the swath forming the Paradise metasedimentary gneiss belt (PMGB), and iii) tectonic slivers in the White Bear Arm complex (WBAC).

The high-grade metasedimentary gneiss within the Earl Island intrusive suite (the latter emplaced at 1670 Ma – *cf.* Section 10.2) occurs as widely dispersed slivers that vary in size from a metre to several kilometres long and appear to be narrow bodies elongate parallel to the prevailing structural fabric in the host rocks. They are interpreted as xenoliths, although only concordant contacts between the two rock types were seen. The main occurrences are: i) south of Table Bay, ii) in central EIIS, iii) in the Frenchmans Island area, and iv) close to the EIIS's southern boundary (Figures 7.1 and 7.2).

The occurrence south of Table Bay is the least confidently mapped. The rocks were described in the field as grey, well-foliated, migmatitic garnetiferous granodiorite, containing streaky layers of coarse- and fine-grained biotite garnetiferous leucosome, and a few amphibolite boudins. Based partly on stained-slab appearance, the rocks are considered to be psammitic to semi-pelitic gneiss. The rocks are regionally on strike with the pelitic and psammitic gneiss at Sandy Bay (described in the previous section), 60 km to the east-southeast. A thin section of sample VO81-409A contains plagioclase, quartz, orange-brown biotite, apatite and zircon (plus garnet seen in hand sample). Two high-grade pelitic gneiss occurrences were found during mapping of the central EIIS. One is an isolated outcrop, seen in thin section to be straightforward biotite sillimanite pelitic gneiss (CG85-375). At the other, white-mica-rich xenoliths are found within granodioritic gneiss.

The central EIIS occurrences are unusual in that they contain corundum, which was seen in both thin sections from the locality (CG85-167A, CG85-167B; Photomicrograph 7.1B). The corundum forms anhedral poikiloblasts hosting numerous opaque mineral inclusions. The white mica in which it is enveloped is assumed to be margarite as neither rock contains K-feldspar and only trivial amounts of biotite (*i.e.*, low K activity, hence not muscovite). In CG85-167A, other minerals are a very heavily sericitized, relict plagioclase (Ca-rich) and a colourless, unaltered, well-twinned plagioclase (albite). Associated minerals are quartz (with the albite, and separated from corundum by margarite), garnet (grossularite?), an opaque oxide, clinozoisite and zircon. In the other thin section (CG85-167B), minerals associated with corundum are margarite, clinozoisite, an opaque oxide and zircon. In the CaO–Al_2O_2–SiO_2–H2O system, the formation of margarite takes place by the reaction:

Anorthite + Corundum + Fluid = Margarite

Lack of any mineral chemical data, in conjunction with more complex phase relationships when other components are present, particularly Na and Fe (*e.g.*, Rosing *et al.*, 1987), precludes detailed exploration of specific reactions, which undoubtedly involve the other Ca–Al phases present.

The rocks in the Frenchmans Island and Partridge Bay areas are strongly migmatized, muscovite ± sillimanite ± biotite pelitic and semi-pelitic schist and gneiss, locally containing minor garnet (up to 1 cm across at CG84-495). Muscovite-rich schist is interlayered with quartzite layers at CG85-504, whereas very sillimanite-rich knots are found at CG85-515B (Photomicrograph 7.1C). A metamorphosed mafic dyke discordantly intrudes pelitic gneiss at CG85-514. Thin sections examined are CG85-515A, B, SP85-063 and VN85-522.

The occurrences close to the EIIS's southern boundary appear to be aligned parallel to the boundary between the EIIS and the PMGB, and located 3 to 5 km north of it. It is possible that they might all belong to a single zone. The rocks are well-banded, muscovite-bearing pelitic gneiss, schistose in part. Field descriptions emphasize the presence of abundant leucosome or muscovite-rich pegmatitic material in irregular pods and patches, with which biotite ± muscovite ± sillimanite-rich veneers are associated. Minor amphibolite is also present.

Of three thin sections prepared, all contain plagioclase, microcline, quartz, muscovite, buff-green biotite, an opaque oxide, zircon and monazite. Sillimanite was recorded in samples VN85-083 and VN85-281, but is lacking in GM85-051.

7.3.2.1 Southeast Sandwich Bay Assemblage (P_{3A}sp, P_{3A}ss, P_{3A}sq, P_{3A}sc, P_{3A}vm)

The Southeast Sandwich Bay assemblage is not readily assigned to any of the established subdivisions in the Hawke River terrane and is only addressed here for want of a better option. The rocks are situated in the southeast part of Sandwich Bay (Figure 7.1).

The key feature of the assemblage is its structural trend, striking northeast and dipping to the northwest, in contrast to the prevailing northwest regional strike and near-vertical dips. The rocks are also characterized by severe mylonitization, isoclinal folding and a generally highly contorted gneissosity. The author's 1:100 000-scale maps for the area (Gower, 2010a; Sandwich Bay and Paradise River map regions) interpret the assemblage as contained in a southeast-verging thrust slice. The area is poorly understood and merits detailed structural study, which would be logistically easy, given good exposure, and road and shoreline access. Present knowledge is based on mapping by Cherry (1978a, b), Gower *et al.* (1982b) and Gower *et al.* (1985), plus follow-up petrographic studies by the author and an Ar–Ar geochronological result by van Nostrand (1988).

The rocks are a mixture of muscovite- or sillimanite-bearing pelitic gneiss, psammitic gneiss, minor quartzite and calc-silicate rocks, mafic rocks, rare ultramafic pods and granitoid units. The pelitic rocks are thoroughly migmatized, having garnet–biotite–sillimanite–restite and pink quartzofeldspathic leucosome. Some have K-feldspar porphyroblasts up to 2 cm across. Some of the mafic rocks are undoubtedly metamorphosed, crosscutting, diabase mafic dykes, but it also seems likely that rocks derived from a pillowform mafic volcanic protolith are present (VN84-007). Quartzite and calc-silicate rocks are minor and, in the case of the quartzite, possibly derived from a chert protolith. Some muscovite-bearing schistose rocks are also present and may represent retrogressed equivalents of the sillimanite-bearing pelitic gneiss. Garnet retrogression to biotite and plagioclase is also seen and could be another facet of the same process. Fluorite was recorded in muscovite–chlorite shear zones at VN84-048. Also present are quartz diorite, granodiorite and tonalitic rocks, similar to those present in the Earl Island domain. These contain xenoliths of both supracrustal rocks and amphibolite. The rocks are intruded by pegmatite (some muscovite bearing) and aplite, in various deformational states.

The Ar–Ar geochronological result obtained by van Nostrand (1988) alluded to earlier was obtained on hornblende separated from quartz diorite. Ages of 1241 ± 3 Ma (total gas) and 1246 ± 1 Ma (plateau) were obtained. The low-temperature fraction yielded a date of 1080 ± 12 Ma, but as this represented only 2% of the total gas released, the result was discounted as having any significance by van Nostrand.

Five samples of pelitic gneiss were examined petrographically (GF81-107A, GF81-107B, MC77-077A, VN84-007B, VN84-288), four of which are sillimanite bearing. The sillimanite has been folded, indicating at least two deformational events. Given that sillimanite is partially retrograded to muscovite, it seems likely that the pelitic gneiss thin section that lacks sillimanite (VN84-288) may also have contained it prior to retrogression. One of the thin sections is also garnet bearing (VN84-007B). A separate group of thin sections has been termed semi-pelitic schist (MC77-255D, MC77-255E, VN84-046B, VN84-050A, VN84-269, VN84-271). They are all characterized by muscovite- and biotite-bearing mineral assemblages. In addition, VN84-050A contains fibrolitic sillimanite and has graphite, and VN84-271 has garnet. Four samples (MC77-079A, MC77-080A, MC77-255C, VN84-046A) are deemed to be psammitic gneiss, although MC77-080A could, alternatively be a granodioritic plutonic rock. Granitoid plutonic protoliths were also favoured for MC77-082A, VN84-267, VN84-275 and VN84-282. Other rock types include: i) a calc-silicate rock containing orange-brown garnet (*cf.* andradite), pleochroic epidote, bright green clinopyroxene (*cf.* diopside), quartz, blue-green amphibole, minor apatite and an opaque oxide (VN84-007D), ii) a hornblende-dominant ultramafic rock, also having minor quartz, plagioclase, titanite and an opaque mineral (VN84-054), iii) an amphibolite showing hints of a former diabasic texture (VN84-007C), and iv) a plagioclase-dominant rock of uncertain protolith (anorthositic?), containing some biotite and an opaque mineral, and veined by quartz (MC77-254A).

7.3.3 HAWKE RIVER TERRANE; PARADISE METASEDIMENTARY GNEISS BELT

The Paradise metasedimentary gneiss belt (PMGB) is one of the best-established units in eastern Labrador, hence is addressed in more detail than other examples (Figure 7.6). The belt is about 180 km long, 45 km wide in the southeast and tapers out to the northwest, where it is attenuated, transposed, and, ultimately, completely sheared out by thrusts and strike-slip faults. The PMGB was not known as a separate entity prior to mapping at 1:100 000 scale (in stages by Gower *et al.*, 1982b, 1985, 1986b, 1987), although Eade (1962) depicted some small areas of paragneiss that are now included as part of the belt. The name 'Paradise metasedimentary gneiss belt' was introduced by Gower *et al.* (1986b).

The PMGB supracrustal rock types can be subdivided into two groups, namely: i) mafic extrusive rocks accompanied by calc-silicate rocks and quartzite/metachert, and ii) volumetrically dominant pelitic gneiss, with which minor psammitic gneiss and metasedimentary diatexite are associated. The PMGB is also host to a range of small igneous intrusive bodies, which are also reviewed in this section as their affiliation with other units is uncertain.

7.3.3.1 Mafic Extrusive Rocks (P_{3A}vm, P_{3A}am)

Rocks deemed to have a mafic extrusive protolith are mainly found in the extreme southeastern part of the PMGB, which includes the Bull Island geochronology site mentioned in Section 7.2.1.4. More equivocal occurrences are situated farther northwest (*see* later in this section). With one exception, all are in coastal areas. As these rocks are recessive weathering, it seems likely that other instances exist in less-well-exposed inland regions. The various occurrences were discovered during 1:100 000-scale mapping by Gower *et al.* (1985, 1986b, 1987).

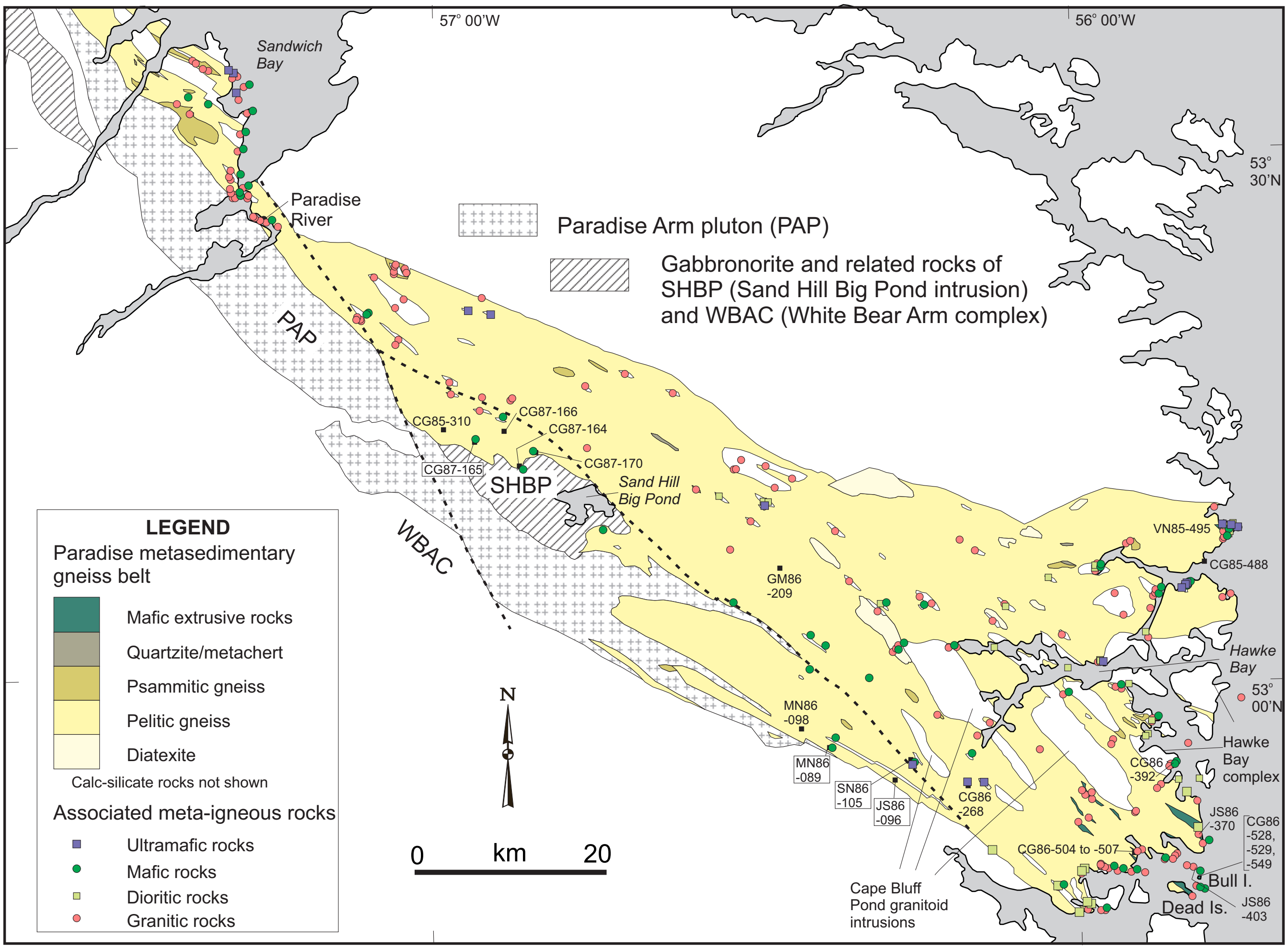


Figure 7.6. *Distribution of supracrustal rock types in Paradise metasedimentary gneiss belt, and associated meta-igneous rocks.*

The rocks are typically black-weathering (locally dark-green), fine-grained amphibolite that have a banded appearance due to compositional variations. The banding is commonly straight, even, and may be quite delicate (Plate 7.5A). It is somewhat discontinuous and lensy in places, but differs from banding resulting from migmatization and/or deformation. The delicate banding is interpreted to be a primary bedding feature, but, in more deformed rocks, it may be related to retaining textural and mineralogical contrasts between melanocratic rims and lighter coloured interiors of pillows in extremely attenuated units. Other features are discordant quartzofeldspathic stringers and quartz veins and a few pegmatites. A possible 60-m-thick mafic sill associated with probable mafic extrusive rocks is present at data station GM85-527.

A key locality is CG86-528 on Bull Island, at which an interlayered sequence of unequivocal pillowed and unpillowed mafic volcanic rocks are superbly exposed (Plate 7.5B). This is the only locality known in the Grenville Province in eastern Labrador where pillowed mafic rocks are known for certain. The locality was first mapped by Gower *et al.* (1987), but the most detailed description, including limited whole-rock geochemical data is by Gower and Swinden (1991). In the pillowed mafic volcanic rocks (Plate 7.5C), individual pillows are generally less than 40 cm in diameter, have thick, darker weathering pillow rinds, and locally have an internal radial structure, which may be related to vesicle patterns. Quartzofeldspathic mobilizate is locally present between the pillows. Thin layers of siliceous, laminated sediment (which could be of felsic volcaniclastic origin) are preserved between unpillowed flows and may be traced across the outcrop as deformed, sinuous bands (Plate 7.5D). The lavas are intruded by mafic dykes (Plate 7.5B).

In addition to an age of 1645 ± 4 Ma (Figure 7.1) from an unpillowed mafic flow, interpreted to date time of metamorphism (Kamo *et al.*, 1996; sample CG86-528A, Section 7.2.1.4), a sample of a crosscutting mafic dyke (CG86-

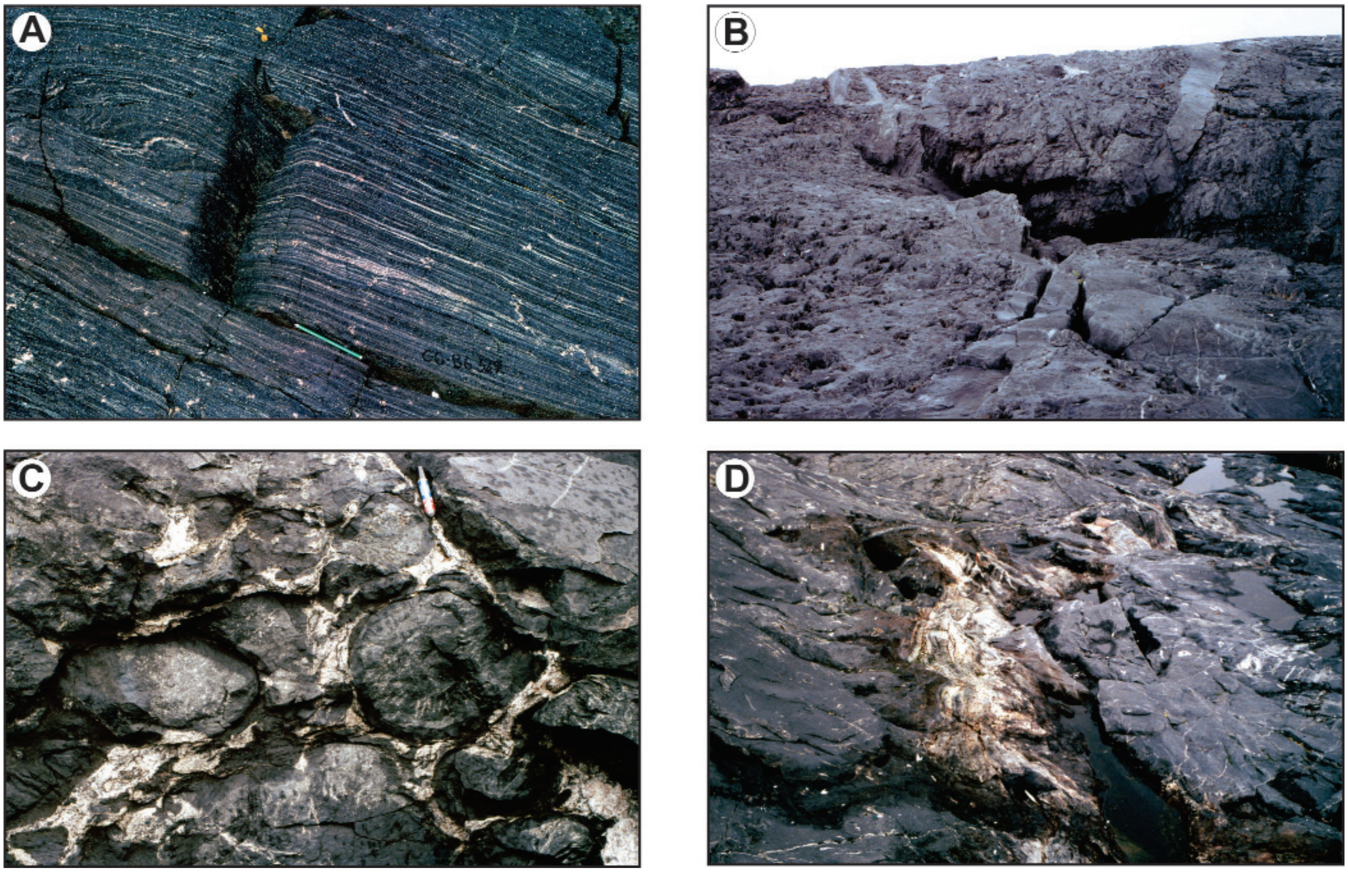


Plate 7.5. *Mafic volcanic rocks from Paradise metasedimentary gneiss belt. A. Delicate banding in mafic rock interpreted to be primary bedding feature (CG86-549), B. General view of outcrop at which pillowed mafic volcanic rocks seen. Note cross-cutting mafic dykes (CG79-528), C. Pillowed mafic volcanic rocks with white interpillow quartzofeldspathic mobilizate. Note dark pillow margins and internal radial structure in pillows, attributed to vesicle patterns (CG79-528), D. Sinuous layer of creamy-white, siliceous laminated sediment between unpillowed flows (CG86-528).*

528C) was also investigated. It yielded an upper-intercept titanite age of 1567 ± 6 Ma, which was regarded as providing a minimum age for the time of dyke emplacement.

In nearby areas, where deformation and/or migmatization have masked the primary nature of the rock, protolith can still be inferred from a characteristic association with other supracrustal rock types. Most diagnostic are the fine-grained, quartz-rich rocks that are interpreted to be metachert. These are white or rusty-weathering, depending on sulphide content (mostly pyrite). Very commonly, the sulphide content is sufficient for the rocks to be mapped as gossans. A few banded quartz-garnet-magnetite rocks are also associated and interpreted as lean banded iron formation.

Almost equally diagnostic as the metachert, is a variety of calc-silicate-rich rocks. These commonly carry grossularite/andradite and/or clinopyroxene, and, in addition to forming fairly continuous layers, also occur in pockets, pods, lenses and stringers. Some of the calcic material is found as interpillow mesostasis. The author has mapped similar material in Archean metamorphosed pillowed lavas in the Yilgarn craton in Western Australia and in the Superior Province in Ontario. The topic of calcareous sediments associated with mafic volcanic rocks in general, including in pillow interstices, has been addressed by Nutman *et al.* (2010) with reference to the Archean Isua supracrustal belt in Greenland.

Also present are psammitic and pelitic schists and gneisses (which may be garnet and/or cordierite bearing), and a few occurrences of rocks that have a lensy, fragmental appearance (possible volcaniclastic rocks).

Potential correlation with occurrences distant from the Bull Island district was considered by Gower and Swinden (1991), who mentioned an outcrop of fine-grained amphibolite on the shore of Sandwich Bay, 4 km northwest of the community of Paradise River, in the northwest part of the PMGB (Figure 7.6). The mafic rocks are characterized by melanocratic seams and pods of calc-silicate minerals. Attention was also drawn by Gower and Swinden (1991) to an additional occurrence of similar rocks that occurs in association with pelitic gneiss on the southeast shore of The Backway (Figure 7.1). This occurrence cannot be demonstrated to be part of the PMGB, although it is regionally, on strike with it and it seems likely that it was once linked. Note that Gower and Swinden (1991) also mention rocks probably having a mafic volcanic protolith associated with pelitic gneiss in the Grenville Province in central/ western Labrador.

Ten thin sections were examined (CG86-394, CG86-528A, CG86-528B, CG86-531B, CG86-549, GM85-527B, GM85-528B, GM85-531B, MN86-377, SN86-353B). All contain polygonal plagioclase, orange-brown biotite, leaf-green hornblende and an opaque oxide. Quartz forms ovoids of recrystallized grains up to 0.5 cm in diameter in CG86-394, with which plagioclase and, in one instance, orthopyroxene are associated. Although the outline of most of the ovoids is a bit ragged, a few have smooth margins. The ovoids are interpreted by the author to represent amygdaloidal infillings. Quartz is also present in four other thin sections as inclusions or as an interstitial mineral. Orthopyroxene is seen in two other thin sections (CG86-549 – weakly pleochroic, SN86-353 – strongly pleochroic) and appears to be part of the stable metamorphic assemblage. Amoeboid garnet hosting numerous opaque oxide inclusions is abundant in CG98-528B, a rock interpreted in the field to be part of the scouriaceous margin of a flow. An opaque sulphide is present in half of the thin sections. Other trace minerals sporadically present are zircon and apatite, and secondary chlorite, carbonate and serpentine. Carbonate forms discordant veinlets in GM85-528B.

On the basis of very limited whole-rock geochemical data (three samples from the Bull Island locality), Gower and Swinden (1991) concluded that the rocks were best interpreted to be variably fractionated transitional tholeiites to mildly alkali basalts, erupted in a within-plate tectonic setting. An extended REE plot showed lack of either positive Th or negative Ta and Nb with respect to La, suggesting that the magmatism was not related to subduction.

7.3.3.2 Calc-silicate Rocks (P_{3A}sc)

Calc-silicate rocks in the PMGB are typically found as part of a heterogeneous package of rocks that includes supracrustal amphibolite, and quartz-rich rocks believed to be derived from banded metachert (*cf.* Plate 7.6A, B). Pelitic and psammitic material may also be associated. On the basis of common listing of the unit (P_{3A}sc) in unit designator strings, one might assume a slightly higher proportion, but calc-silicate rocks tend to comprise only a small part of outcrops. The proportion of calc-silicate rocks in the PMGB is <1%. Typically, they are found as small (<20 cm across) pods or lenses, as layers (usually only a few centimetres thick), and in pockets and patches. The rocks are mostly green-grey or ocherous weathering, fine to coarse grained and massive to moderately foliated. Gower *et al.* (1985) suggested that some of the calc-silicate material could be metamorphosed inter-pillow-lava mesostasis that was originally calcareous mud (Plate 7.6A such an example?).

Nineteen thin sections are available (CG86-127, CG86-505, CG86-528D, GM85-115, GM85-227, GM85-392, GM85-460B, GM85-484C, GM85-484D, GM85-507B, GM85-514A, LC85-010B, LC85-026B, LC85-034, SP85-049A, VN84-038, VN84-125, VN84-131, VN85-495B). The most common minerals are quartz, amphibole (pale-green, leafy-green, blue-green), pale-green clinopyroxene (yellow-green in GM85-460B – hedenbergite?), orange-brown garnet (andradite/grossularite), an opaque oxide, and titanite, although not all the listed minerals are present in all thin sections. Other minerals commonly present are plagioclase (mostly minor), orange-brown biotite, sulphide, apatite, epidote and carbonate. K-feldspar was recorded in one thin section and, interstitially, trace amounts suspected in two others. This brief description hardly does justice to the varied mineral assemblages, retrograde reactions and, possibly, diverse protoliths of these rocks.

7.3.3.3 Quartzite/Metachert (P_{3A}sq)

Quartzite/metachert represents probably <1% of the supracrustal rocks in the PMGB, but, as it is a relatively easy rock to recognize, it is easy to overestimate its abundance. The rocks are white or grey-weathering (locally pink, green, black, or ocherous), fine to coarse grained, thinly to thickly banded (probably reflecting original bedding), and unfoliated to mylonitic (Plate 7.6C). A mutual association of quartzite, calc-silicate rocks, supracrustal amphibolite, and pyritic gossans (Plate 7.6D) is quite common, especially in the Dead Islands area. Quartz-rich rocks are rarely extensive enough to be mapped as the dominant unit at an outcrop, commonly occurring as thin laminae (<1 cm thick), or thicker layers, intercalated with the other rock types listed above. The thickest quartzite layer reported in field notes was estimated to be 40 m thick (GM85-473).

It seems likely that the quartzite was derived from banded chert and lean sulphide-facies banded iron formation, and that the pyritic gossans are the weathered surface expression of the sulphide-bearing varieties. Based on 1:100 000-scale mapping, most of the gossans are within a few kilometres of the coast. As this is where exposure is best, there is no reason to suppose that these easily weathered and eroded rocks are confined to these areas. In interior areas where exposure is good, similar gossanous material exists. Quartzite, interbedded with calc-silicate rocks, correlates with a marked magnetic anomaly 8 km northeast of Sand Hill Big Pond. Quartzite was also recorded at three widely spaced outcrops situated immediately adjacent to the northern margin of the White Bear Arm complex.

Thirteen samples from the PMGB were examined in thin section (CG81-170A, CG81-291, CG86-546A, EA61-049A, GF81-235C, GM85-388C, GM85-388D, JS86-435B, MN86-374, VN85-096A.1, VN85-096A.2, VN85-096B, VN85-096C). The rocks were termed

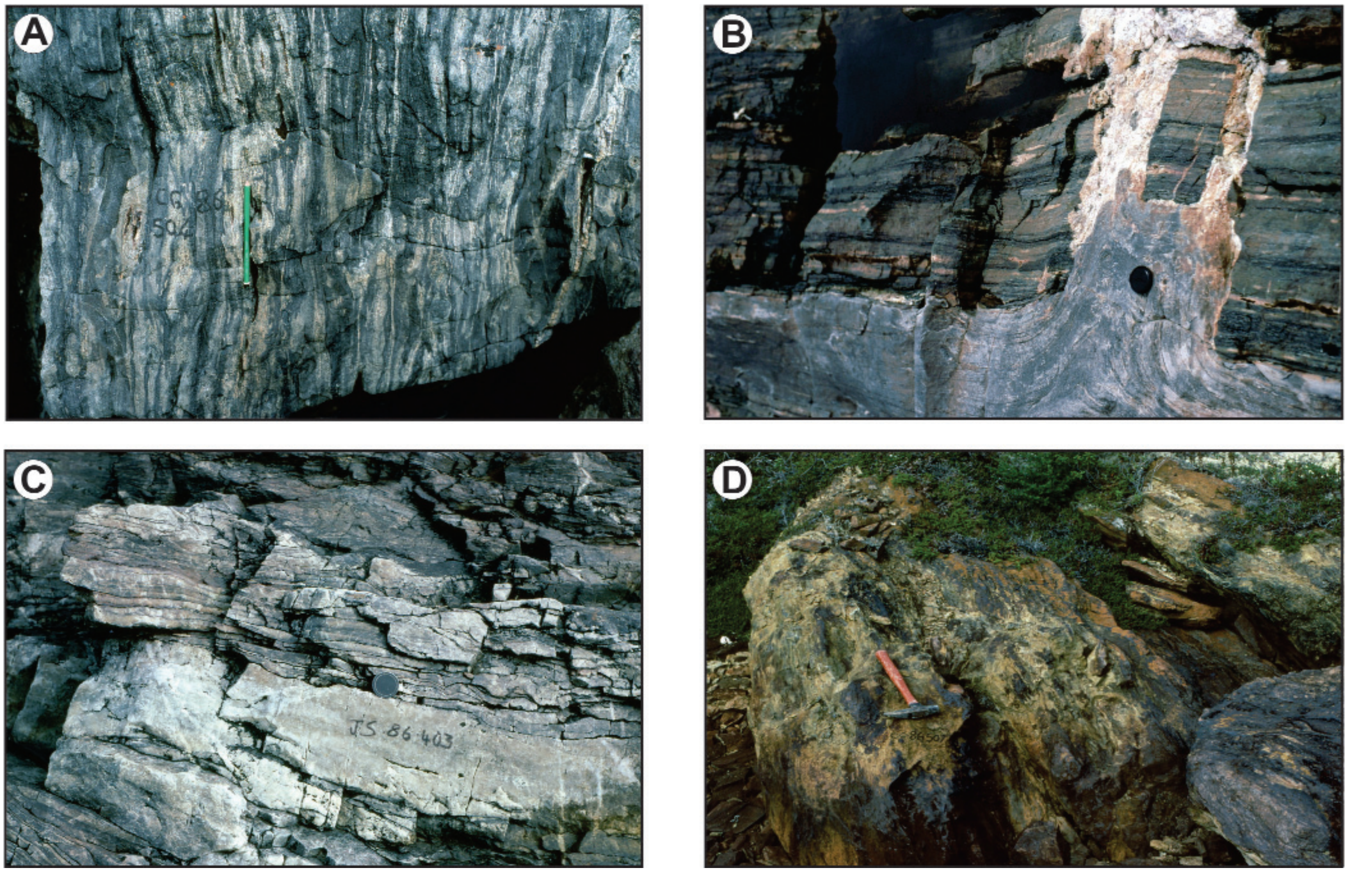


Plate 7.6. *Calc-silicate rocks, quartzite–metachert and sulphide-rich material associated with mafic volcanic rocks in Paradise metasedimentary gneiss belt. A. Irregularly banded mafic–calc-silicate rock associated with mafic volcanic rocks (CG86-504), B. Calc-silicate rocks and quartzite. The quartzite has flowed into boudinaged calc-silicate rock and some pegmatite generated (VN85-495), C. Quartzite (JS86-403), D. Sulphide-rich material associated with calc-silicate rock – quartzite–metachert – mafic volcanic assemblage (CG86-507).*

quartzite or meta-banded iron formation. Except for one sample, all contain quartz. The exception is JS86-435B, which, apart from opaque oxide and sulphide, consists entirely of a colourless, non-pleochroic, polysynthetically twinned amphibole (*cf.* grunerite). A similar amphibole is also present in EA61-049A, but associated with a much more diversified mineral assemblage that includes quartz, orange-brown biotite, garnet and minor apatite. Four other thin sections contain a colourless to very pale-green amphibole suspected to be tremolite/actinolite. In those sections lacking amphibole, the mafic minerals are orange-brown biotite and garnet. Three sections contain minor plagioclase and one has accessory K-feldspar, and may be transitional into psammite. Chlorite, epidote, carbonate, zircon and secondary white mica are sporadically present in trivial amounts.

7.3.3.4 Psammitic Gneiss (P_{3A}ss)

Psammitic gneiss is one of the more difficult rocks to recognize reliably in high-grade metamorphic terrains, especially distinguishing it from granitoid units at an advanced stage of deformation and metamorphism. For this reason, some of the designators involving Unit P_{3A}ss are commonly given alternate granitoid unit labels. Furthermore, psammitic gneiss also grades into similar rocks enriched in calcareous, aluminous or siliceous components (Units P_{3A}sc, P_{3A}sp and P_{3A}sq), so these may also be offered as alternates. Ultimately, choice of designator becomes somewhat arbitrary, and the map user should be aware and accepting of this. In any case, psammitic gneiss is a very minor rock type in the PMGB, probably <2% of supracrustal rocks present.

Rocks most confidently identified as psammitic gneiss are pink, grey, buff, creamy or rusty-weathering, fine to coarse grained and, typically, not well banded/bedded. Layers are commonly only a few centimetres thick, and rarely more than a metre without intervening layers of some other supracrustal rock. They are broadly associated with the pelitic gneiss, but tend to be more intimately associated with quartzite and calc-silicate units. There is no indication that any of these rocks were derived from a felsic volcanic protolith.

Twelve thin sections from the PMGB were identified as psammitic gneiss (CG81-124, GM85-321, GM85-358, GM85-385B, GM85-479, GM85-480, GM85-485A, JS86-113, SN86-359, SP85-052, VN84-132, VN85-509). All contain plagioclase and quartz, and all but three contain K-feldspar (microcline and/or perthite). All contain biotite (green-buff or buff-orange) and seven also have minor muscovite. Two samples associated with calc-silicate supracrustal units have leaf-green hornblende (GM85-385B, JS86-113) and the latter sample also contains clinopyroxene. Garnet is present in CG81-124, GM95-480 and VN84-132. All contain an opaque oxide and sulphide is also present in GM85-385B. Apatite, zircon, monazite, allanite, titanite and epidote (in order of decreasing abundance) are sporadically present. Compositional banding that is not likely migmatitic or mylonitic, and the dispersed nature of mafic and opaque minerals (*cf.* Gower, 2007), are features that assist in protolith identification.

7.3.3.5 Pelitic Gneiss (P_{3A}sp)

Pelitic gneiss in the PMGB is a marvel of metamorphic mineral diversity. Minerals included are muscovite, sillimanite, kyanite, staurolite, cordierite, garnet, orthopyroxene, osumilite, sapphirine and spinel. It is one of the author's great regrets that more time could not be devoted to unravelling the intricacies of the PMGB, as doing so has the potential of shedding much light on pre-Grenvillian metamorphism in the region. Here, the author has focused on first-order distributional and petrographic features of various 'index' minerals and only a very preliminary assessment of their significance. The distribution of samples examined in thin section is shown in Figure 7.7A.

Muscovite bearing. Muscovite-bearing pelitic gneiss is concentrated in two main areas of the PMGB, with scattered occurrences elsewhere (Figure 7.7B). One area is in the northeast part of the belt, north of Hawke Bay; the other is in a diffuse zone on either side of a major northwest-trending, strike-parallel fault that transects the PMGB.

Muscovite-bearing rocks north of Hawke Bay (Plate 7.7A) were commonly mapped as pelitic schist. Some of the samples in this group lack other metamorphic indicator minerals (CG85-488, GM85-493, GM85-510, GM85-516, VN85-338), but a key feature of several samples is the presence of fibrolitic sillimanite (Figure

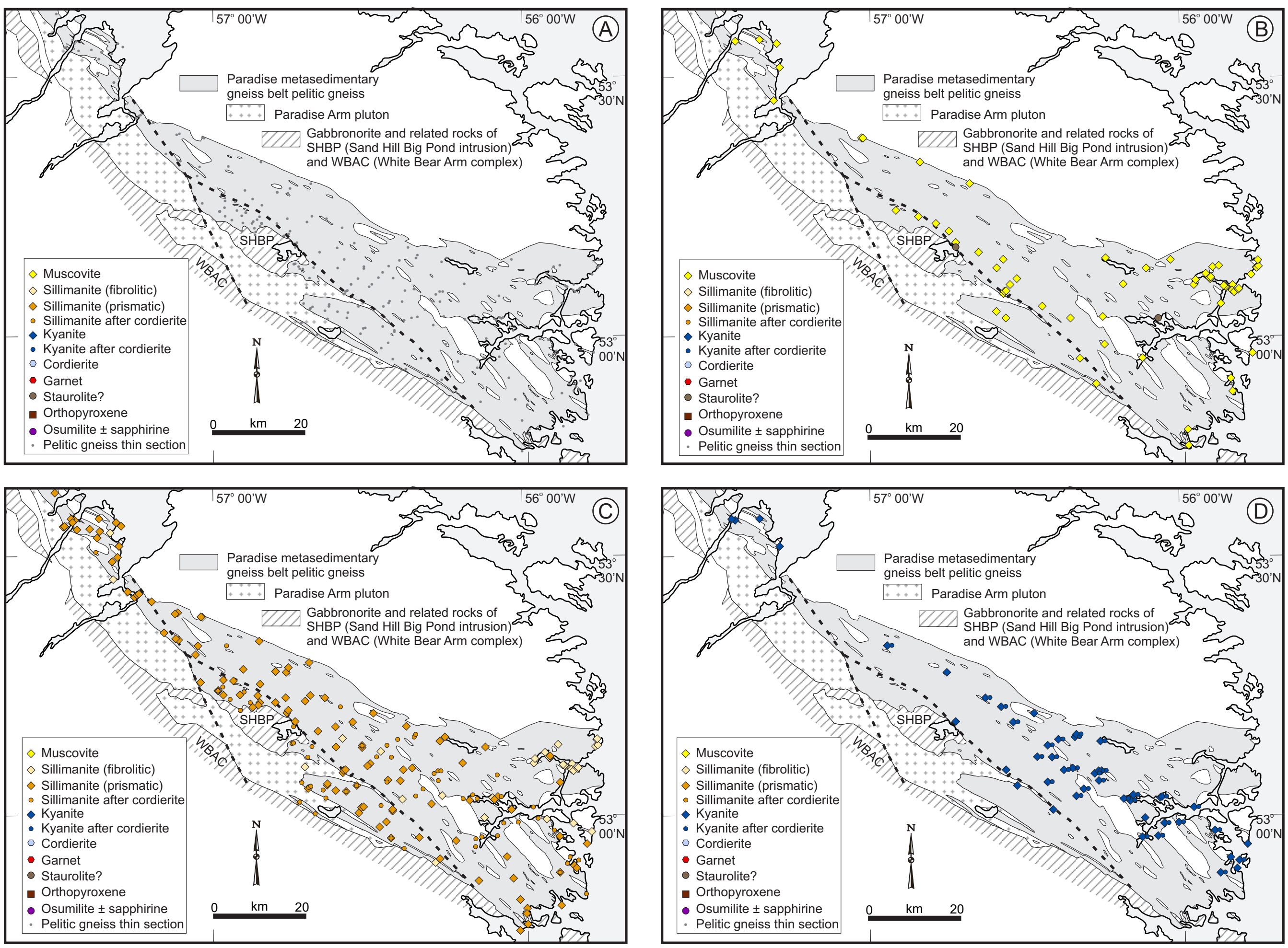


Figure 7.7. *Distribution of thin section locations and metamorphic minerals in Paradise metasedimentary gneiss belt pelitic gneiss. A) Thin section locations, B. Muscovite and staurolite, C. Sillimanite, D. Kyanite.*

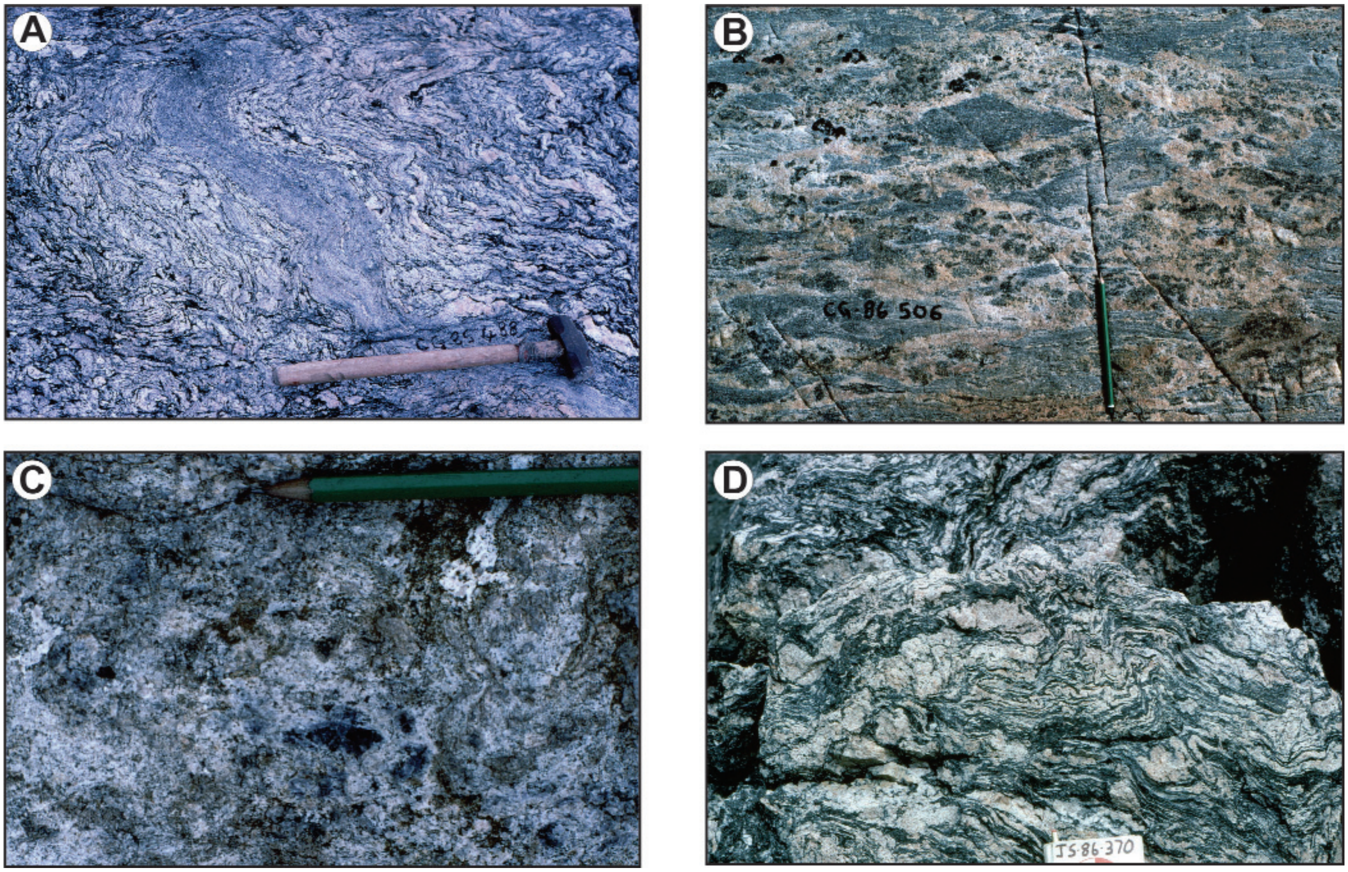


Plate 7.7. *Pelitic gneiss in Paradise metasedimentary gneiss belt (PMGB) – part 1. A. Muscovite-bearing pelitic gneiss from northeast PMGB (no cordierite or prismatic sillimanite). Note grey psammitic layer across the centre (CG85-488), B. Cordierite (dark clusters) within creamy-pink leucosome enveloping remnants of mid-grey semi-pelitic paleosome (CG86-506), C. Large dark-purple cordierite grains (centre and below centre) in leucosome-rich pelitic gneiss (CG86-268), D. Cordierite-bearing pelitic gneiss in which cordierite has been largely pseudomorphed by sillimanite and biotite (in dark layers) (JS86-370).*

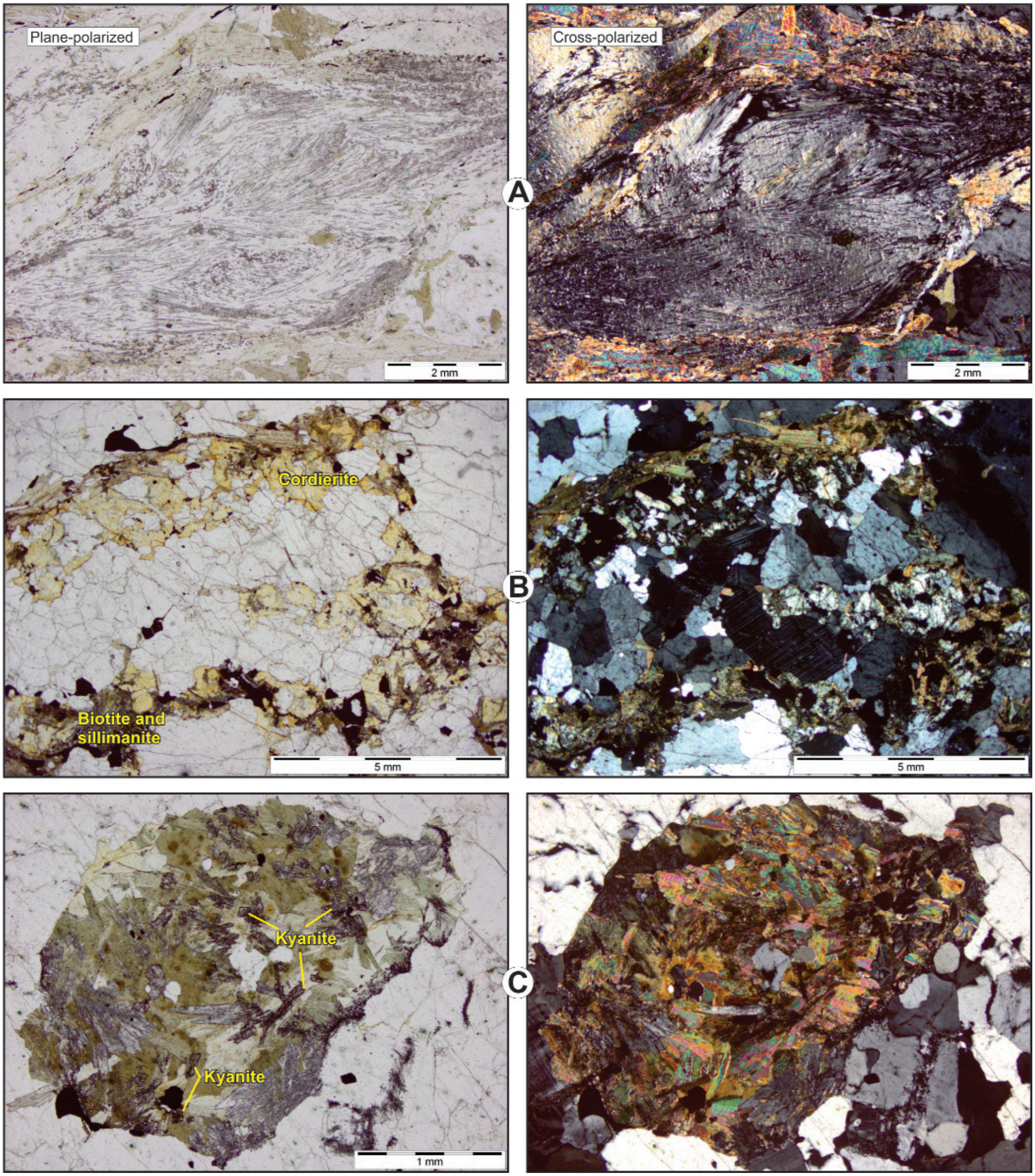
7.7C) in the cores of some flakes of muscovite (CG85-433, CG85-444, GM85-499, GM85-513, GM85-515D, GM85-531A, LC85-024A, LC85-025, LC85-032.1, LC85-032.2, LC85-033, SP85-044). Although petrographic textural evidence is commonly ambiguous, it appears to the author that the relatively low-grade state of these rocks is an original metamorphic feature, rather than them having been regressed from a former high-grade assemblage. The presence of fibrolite, by this interpretation, would be evidence for incipient development of sillimanite, probably according to the reaction:

Muscovite + Quartz = K-feldspar + Sillimanite + H_2O

Most of the muscovite-bearing thin sections flanking the strike-parallel fault also lack other indicator minerals (CG85-218, CG85-258, CG86-150B, GM85-229, JS86-085A, LC85-005B, LC85-007A, LC85-007B, LC85-008, SN86-315, VN85-134A, VN85-237, VN85-251). In three examples, however, sillimanite is seen in the cores of muscovite flakes. In two of the examples, the sillimanite is prismatic (CG85-292, JS86-078A) and it appears to the author that the muscovite is retrograde. In the third sample (JS86-096), sillimanite is fibrolitic, but is atypical in that it is clearly within a stair-step kinematic structure that is related to the fault separating the PMGB from the WBAC (Photomicrograph 7.2A). Overall the muscovite occurrences flanking the fault are interpreted as retrograde.

Staurolite bearing. One anomalously Fe-rich sample (GM85-386; Figure 7.7B) has tentatively identified staurolite. Staurolite is normally not difficult to identify and the uncertainty here is due to the small size of the crystals. Staurolite was also suspected in CG85-132A, near Sand Hill Big Pond.

Sillimanite bearing. Almost all of the pelitic gneisses in the PMGB are sillimanite bearing (Figure 7.7C). Sillimanite that is found in thin sections with kyanite, cordierite, orthopyroxene, osumilite, sapphirine is addressed in sections on those minerals. This section is restricted to pelitic gneiss in which such minerals are lacking (thin sections CG81-148B, CG81-148C, CG81-208, CG81-218B.1, CG81-218B.2, CG81-225, CG85-051, CG85-140, CG85-148A.1, CG85-211, CG85-223, CG85-226, CG85-244, CG85-264, CG85-310D, CG85-310F, CG85-310G, CG85-310H, CG86-131, EA61-059A, GM85-105, GM85-118,



Photomicrograph 7.2. *Features of Paradise metasedimentary gneiss belt muscovite- or cordierite-bearing pelitic gneiss. A. C/S (stair-step) shear-sense indicator (top to right), defined by fibrolitic sillimanite within muscovite (JS86-096), B. Cordierite (yellow) reacting to biotite and sillimanite (GM85-209), C. Cordierite pseudomorphed by biotite, kyanite and sillimanite (CG86-392).*

GM85-137, GM85-140, GM85-200A, GM85-204, GM85-212, GM85-214, GM85-220, GM85-221, GM85-232, GM85-255, GM85-381, GM85-481, GM85-491B, GM85-492, GM85-521, JS86-321, JS86-431, JS86-435A, LC85-003, LC85-004, LC85-009, MC77-086A, MC77-223A, MN86-383B, MN86-386, SN86-103B, SN86-134, SP85-047A, VN84-036, VN84-122, VN84-313, VN84-351, VN85-055, VN85-082, VN85-170, VN85-295, VN85-328).

The characteristic habit for sillimanite is in knots and layers in association with biotite, in which the sillimanite varies from randomly oriented needles to well aligned within the prevailing fabric. Based on many examples where relict cordierite is preserved (*see* later section), it is suspected that the sillimanite–biotite association is the result of these minerals totally replacing cordierite, generally under fabric-generating conditions, such that any evidence of pseudomorphing has been destroyed. A variant of this texture is displayed in a few thin sections (*e.g.*, CG85-244, GM85-137, GM85-221), which show clusters of coarse prismatic sillimanite in the cores of knots and enveloped by biotite–fibrolite symplectite. Another variant comprises felted clusters of tiny acicular crystals at borders of felsic minerals.

In the author's petrographic database sillimanite was recorded as prismatic, fibrolitic or both. There is a very strong correlation between fibrolitic sillimanite and the presence of muscovite. In several instances, sillimanite occurs in the central parts of muscovite flakes. In some cases, it looks like muscovite formed as a retrograde product around sillimanite, whereas, elsewhere, sillimanite seems to have developed from pre-existing muscovite.

Kyanite bearing. Kyanite was confidently identified in 39 thin sections of PMGB pelitic gneiss (Figure 7.7D) (CG81-166, CG81-170B, CG81-203, CG81-218C, CG85-024, CG85-132A, CG86-113A, CG86-122, CG86-249, CG86-355, CG86-441, EA61-719, GM85-126, GM85-209, GM85-231, GM85-246, GM85-250, GM85-268, GM85-275, GM85-276, GM85-278, GM85-281, GM85-289, GM85-291, GM85-293, GM85-345, GM85-363, GM85-386, GM85-387, GM85-445, JS86-100, MN86-344B, SP85-001, SP85-011, VN85-097, VN85-152, VN85-270, VN85-284, VN85-381).

Quite commonly, kyanite occurs in close spatial association with opaque minerals, the latter possibly helping its nucleation as kyanite can accommodate small amounts of Fe^{3+}. Excluding occurrences in the northwest PMGB (*see* next paragraph), the kyanite-bearing samples are confined to a zone in the central part of the PMGB that is about 90 km long and, at its widest, 15 km wide. All of the thin sections also have sillimanite.

The kyanite commonly has a slight brownish hue, which, along with its high relief, inclined extinction, characteristic non-orthogonal cleavages and bladed habit makes its identification fairly definite in most thin sections. In 8 other thin sections (CG81-218A, CG86-251, CG86-392, LC85-005A, SN86-317, VN85-080, VN85-234, VN85-244) identification is less certain, as the grains are quite small and uncommon. These examples tend to be from sites on the fringes of the kyanite-present area. Kyanite in these rocks is derived from the breakdown of cordierite; either directly, or from sillimanite, which itself is derived from cordierite breakdown (GM85-281, SP85-001). A possibly applicable reaction is:

Cordierite + K-feldspar + H_2O *=*
Sillimanite/kyanite + Biotite + Quartz

In the northwest part of the PMGB, cordierite is lacking in kyanite-bearing thin sections, and it appears that kyanite + K-feldspar co-exist (CG81-166, CG81-170B, CG81-203, CG81-218C). If the effects of dextral displacement across the northwest-trending fault are removed, then these occurrences are aligned with the higher-grade pelitic gneisses along the south side of the PMGB.

Cordierite bearing. Cordierite is present is most parts of the PMGB, except in the northeast corner (Figure 7.8A). This section only considers cordierite that does not occur in association with orthopyroxene, garnet, or osumilite, which are all addressed in later sections (Plates 7.7B–D and 7.8A).

Attention is confined here to cordierite that was the paramount mafic mineral in a former high-grade mineral assemblage, but has now been partially or completely pseudomorphed by sillimanite and phlogopitic mica. Thirty-six thin sections meet this criterion (CG85-147, CG85-174, CG85-255, CG85-310B, CG85-310C, CG85-310E, CG85-323, CG86-143, CG86-396A, CG86-400, EA61-047C, GM85-185, GM85-222, GM85-353, GM85-370, JS86-104, JS86-370, MN86-098, MN86-113, MN86-310, MN86-337, SN86-078, SN86-081, SN86-110B, SN86-115, SN86-312, VN84-028B, VN84-126, VN85-235, VN85-243A, VN85-250, VN85-252, VN85-253, VN85-289, VN85-394, VN85-425).

Although cordierite can be troublesome to identify in thin section, in these samples it is not a problem. It is euhedral to anhedral, colourless or yellow (Photomicrograph 7.2B) (EA61-047C, GM85-209, GM85-370, VN84-126), commonly displays yellow pleochroic inclusions, and, locally, has well-developed sector twinning. Its most characteristic feature is a ubiquitous replacement by pale orange-brown phlogopitic mica (rarely, chlorite), sillimanite and/or kyanite (Photomicrograph 7.2C). The sillimanite occurs as randomly oriented felted masses, or it displays radiating fronds. It ranges from fibrolitic to prismatic. Mostly, the alteration to sillimanite and biotite was static and former euhedral outlines of the original cordierite crystal can be easily seen. Locally, rocks show evidence of mylonitization that postdates the formation of sillimanite and biotite (*e.g.*, CG85-310A-E, VN85-394). In such rocks, the former presence of cordierite is demonstrated by knots of sillimanite and biotite that, commonly, still retain residual kernels of cordierite.

The reaction that seems likely to apply to the retrogression is:

Cordierite + K-feldspar + H_2O *= Sillimanite/*
kyanite + Biotite + Quartz

Garnet bearing. Garnet was identified in 34 thin sections (plus two hand samples for which the derivative thin sections lack garnet) as follows: CG04-285B, CG81-148, CG81-155, CG81-198, CG81-223, CG81-286, CG85-127, CG85-130B, CG85-144, CG85-151, CG85-479, CG86-265, CG86-362B, CG86-537A, EA61-046, EA61-059B, GF81-126, GF81-131, GF81-235A, GF81-236 – hand sample only, GM85-148, GM85-384B, GM85-484A, GM85-484B, GM85-494, MC77-193A, MC77-194B, MN86-089C, VN84-033B, VN84-035B, VN84-102, VN84-139 – hand sample only, VN84-150, VN84-151, VN84-332, VN85-144).

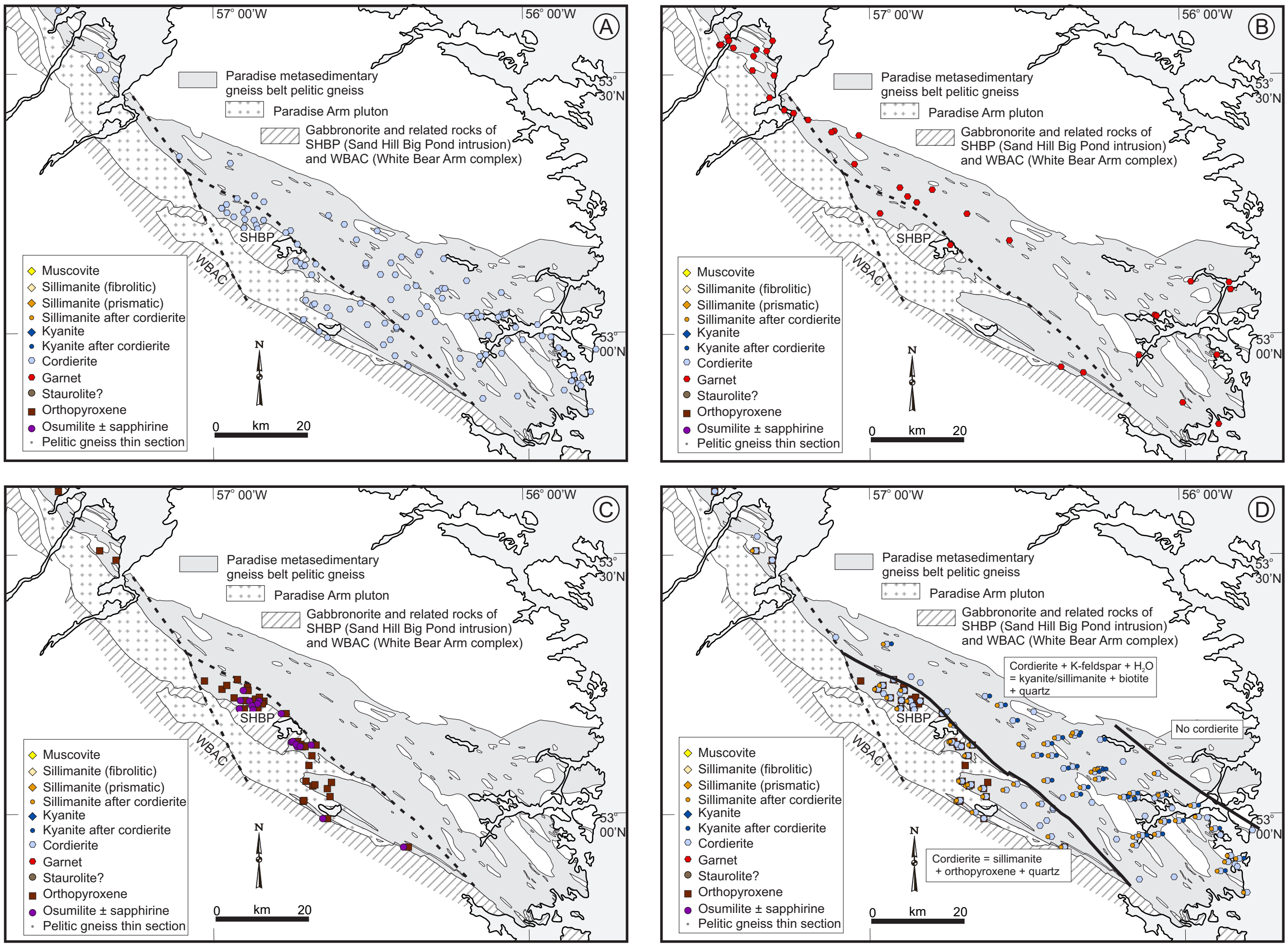


Figure 7.8. *Distribution of metamorphic minerals in Paradise metasedimentary gneiss belt pelitic gneiss and simplified metamorphic domain interpretation. A. Cordierite, B. Garnet, C. Orthopyroxene, osumilite and sapphirine, D. Metamorphic domain interpretation.*

Most of the garnet-bearing samples are in the northwestern part of the PMGB (Figure 7.8B), with more sporadic occurrences in the southeastern area.

Garnet ranges from euhedral to anhedral and typically hosts quartz, an opaque oxide and biotite inclusions, although the opaque mineral may be graphite in a few instances. A possible reaction to explain the quartz, opaque oxide and biotite inclusions is:

Quartz + Magnetite + Biotite + Muscovite
= Garnet + K-feldspar + H_2O

All samples also have sillimanite, except two in which garnet is severely retrograded. Four samples also contain kyanite. No inclusions of sillimanite or kyanite were seen in garnet in any garnet-bearing sample. In MC77-194B, textural evidence suggests garnet and cordierite are reactants according to the reaction:

Garnet + Cordierite = Sillimanite + Biotite + Quartz

Two samples, less than 5 km apart and very close to the contact between the PMGB and the WBAC (CG86-265, MN86-089C), show clear evidence of garnet being pseudomorphed by biotite, quartz and plagioclase. Note that MN86-089C is also the site of sample MN86-089A, which is suspected to contain pseudomorphed osumilite (*see* later in this section).

The concentration of garnet in the northwestern part of the PMGB (along with other evidence; *e.g.*, kyanite + K-feldspar), indicates that the rocks in this area were subjected to higher pressures. The occurrences in the southeast area may be related to more Fe-rich whole-rock compositions.

Orthopyroxene bearing. Orthopyroxene-bearing pelitic gneiss is confined to a fairly narrow zone on the south side of the PMGB (Figure 7.8C; Plate 7.8B). If the effects of dextral displacement along a northwest-trending fault and the effects of wedging apart by the Paradise Arm pluton are both removed, the distribution of the orthopyroxene forms a simple northwest-trending, narrow, lensoid pattern centred on the SHBP. The lens is about 80 km long and up to 10 km wide, tapering to points at both ends. The lensoid zone is bounded on its northeast side by a fault.

Orthopyroxene was recorded in 21 thin sections, 9 of which are additional to those included in the study by Arima and Gower (1991).

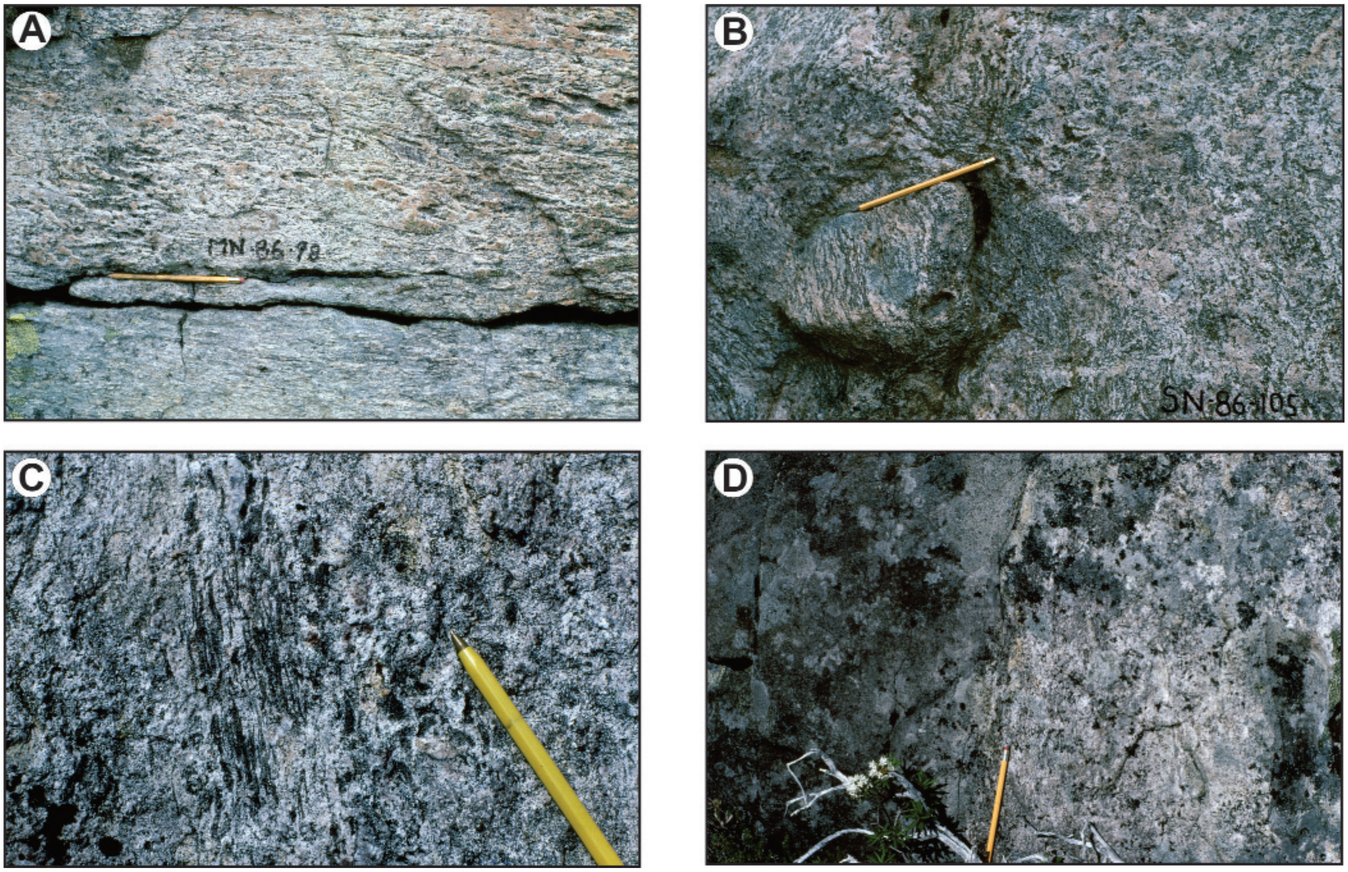


Plate 7.8. *Pelitic gneiss in Paradise metasedimentary gneiss belt (PMGB) – part 2. A. Fairly homogeneous cordierite-bearing pelitic gneiss, depleted in hydrous mafic minerals (MN86-098), B. Pelitic gneiss near the orthopyroxene-in isograd (SN86-105), C. Osumilite- (orthopyroxene + cordierite) bearing pelitic gneiss (CG87-166), D. Contact between PMGB osumilite-bearing pelitic gneiss (right) and fine-grained Sand Hill Big Pond gabbronorite (left) (CG87-164).*

The thin sections are CG81-285, CG85-122, CG85-198, CG85-203, CG85-242, CG85-243, CG85-245, CG85-246, CG85-290, CG85-310A, CG87-165, GM85-143, GM85-194, GM85-196, GM85-201, MC77-194A, VN84-028A, VN84-201, VN85-139, VN85-165 and VN85-167. From petrographic evidence (with six exceptions), it is clear that orthopyroxene, along with sillimanite and quartz, are the products of cordierite breakdown as follows:

Cordierite = Orthopyroxene + Sillimanite + Quartz

The exceptions are CG85-122, CG85-198, CG85-203, VN85-165 and VN85-167, in which orthopyroxene forms an independent phase separate from the cordierite pseudomorphs. These sites are close to the osumilite-in isograd (*see* next section) and near the core of the lensoid distribution. They are therefore regarded as the highest grade, orthopyroxene-bearing rocks that lack osumilite. From experimental data, the above reaction (Mg end member, written in the direction of increasing pressure) takes place at 800–900°C and 11.0–11.5 kb, and has a very shallow P/T slope (Motoyoshi *et al.*, 1993), hence making for a useful geobarometer.

Associated minerals are plagioclase, K-feldspar (mostly perthite; some microcline), quartz, pale orange-brown (phlogopitic) mica, an opaque oxide, and accessory apatite, zircon, monazite, spinel and corundum. Not all of the accessory minerals are present in every thin section, and none of the thin sections carry garnet.

Osumilite and Sapphirine bearing. Osumilite (Figure 7.8C) is a rare potassium ferromagnesian aluminous silicate having the formula $(K,Na)(Mg,Fe)_2(Al,Fe,Mg)_3(Si,Al)_{12}O_{30}$. It is an important indicator mineral in the high-temperature subfacies of granulite-facies metamorphism. Only two osumilite occurrences are known in Canada, the other also in Labrador, in the Nain area (Berg and Wheeler, 1976). Both occurrences are related to granulite-facies contact aureoles of deep-seated AMCG-related intrusions. In the case addressed here, osumilite is related to the emplacement of the Sand Hill Big Pond gabbronorite–leucogabbronorite–anorthosite body (SHBP) into the PMGB. Sapphirine is a ferromagnesian aluminous silicate having the formula $(Mg,Fe,Al)_8(Si,Al)_6O_{20}$, and is also an important indicator mineral in granulite facies terranes.

The pelitic gneisses adjacent to the SHBP (Plate 7.8C, 7.8D) were first recognized as distinct by Gower *et al.* (1986b, page 105) who reported that "Adjacent to the Sand Hill Big Pond gabbronorite, the metasedimentary gneiss has a buff, grey, or brown-weathering appearance, and lacks the well-banded leucosome–restite contrasts that occur throughout the remainder of the area." In the 'Metamorphism' section of the same report, it was noted that these rocks were depleted in hydrous mafic silicates, lacked garnet, but contained orthopyroxene, sapphirine and abundant magnetite, and the conclusion reached that the rocks indicate a high-pressure and high-temperature aureole adjacent to the SHBP intrusion. The presence of osumilite was not mentioned by Gower *et al.* (although was suspected at the time – partly as a result of the author having seen osumilite-bearing rocks adjacent to the Rogaland AMCG suite in southern Norway the previous year). The presence of osumilite was established by M. Arima (Arima and Gower, 1987). After more detailed sampling in 1987, a petrogenetic study involving whole-rock and extensive microprobe mineral geochemical analysis was carried out by M. Arima, culminating in a comprehensive journal publication in 1991 (Arima and Gower, 1991), from which the following information has been extracted.

The Arima and Gower (1991) study included 21 osumilite-bearing thin sections, to which 6 additional thin sections are added here. Two of the 'new' sites are remote from the SHBP and receive separate mention at the end of this section. The osumilite-bearing (and very closely related) thin sections are CG85-121B, CG85-177, CG85-183A, CG85-183B, CG85-183C, CG85-208, CG86-095, CG87-164B, CG87-166, CG87-167A.1, CG87-167A.2, CG87-168, CG87-169, CG87-170A, CG87-170B, CG87-170C, CG87-170D, CG87-171A, MN86-089A, VN85-164A, VN85-176, VN85-185A, VN85-188, VN85-207, VN85-208, VN85-209 and VN85-211.

The osumilite-bearing rocks, although lacking obvious compositional banding may have a very finely laminated to lensy or crudely layered appearance. From a distance they look massive or have a nodular/knotty aspect. The osumilite and sapphirine occurrences are limited to a 1- to 3-km-wide zone adjacent to the SHBP intrusion, hence allowing an 'osumilite-in' isograd to be defined. Osumilite occurs as idioblastic grains up to 5 mm in diameter.

In thin section, osumilite commonly shows extremely fine-grained symplectic intergrowths around grain margins and along cleavages (*e.g.*, CG87-167A.2; Photomicrograph 7.3A, B). In places, the symplectite completely replaces osumilite. The symplectic intergrowths consist of K-feldspar, orthopyroxene, cordierite and quartz, suggesting the reaction:

K-feldspar + Orthopyroxene + Cordierite + Quartz = Osumilite

This reaction is considered by Arima and Gower (1991) as defining the 'osumilite-in' isograd, whereas the thin section evidence for the reverse-direction reaction is taken to represent the formation of secondary products during retrogressive cooling.

Sapphirine occurs as discrete grains in silica-deficient zones of the rocks, where spinel, cordierite, sillimanite, corundum, hematite and/or magnetite are relatively abundant. Note that sapphirine occurrences are restricted to the quartz-absent zones in the gneisses, suggesting that the assemblage sapphirine + quartz was not stable under the peak metamorphic conditions (*i.e.*, reaction of sillimanite + orthopyroxene = sapphirine + quartz was not operative – in contrast to central Labrador). Orthopyroxene and sillimanite intergrowths are locally developed around Fe-oxide grains and between cordierite and sapphirine grains, indicating the reaction:

Sapphirine + Cordierite + Fe-oxides
= Orthopyroxene + Sillimanite

Discrete grains of sapphirine associated with cordierite are locally mantled by sillimanite, indicating the continuous reaction:

Sapphirine(1) + Cordierite = Sapphirine(2) + Sillimanite

Orthopyroxene is present as porphyroblasts or as fine-grained crystals mantling Fe oxides, prismatic sillimanite, and phlogopite, or in various symplectic reactions. The rocks lack garnet and micas, except for the sporadic occurrence of large platy grains of phlogopite mantled by orthopyroxene, which is interpreted to indicate the reaction:

Phlogopite + Quartz = Orthopyroxene + K-feldspar + Vapour

The assemblage sapphirine + cordierite + sillimanite, is constrained to lower temperatures and pressures than the reaction:

Cordierite + Sillimanite = Sapphirine + Quartz

and higher temperatures than the reaction:

Enstatite + Sillimanite = Cordierite + Sapphirine

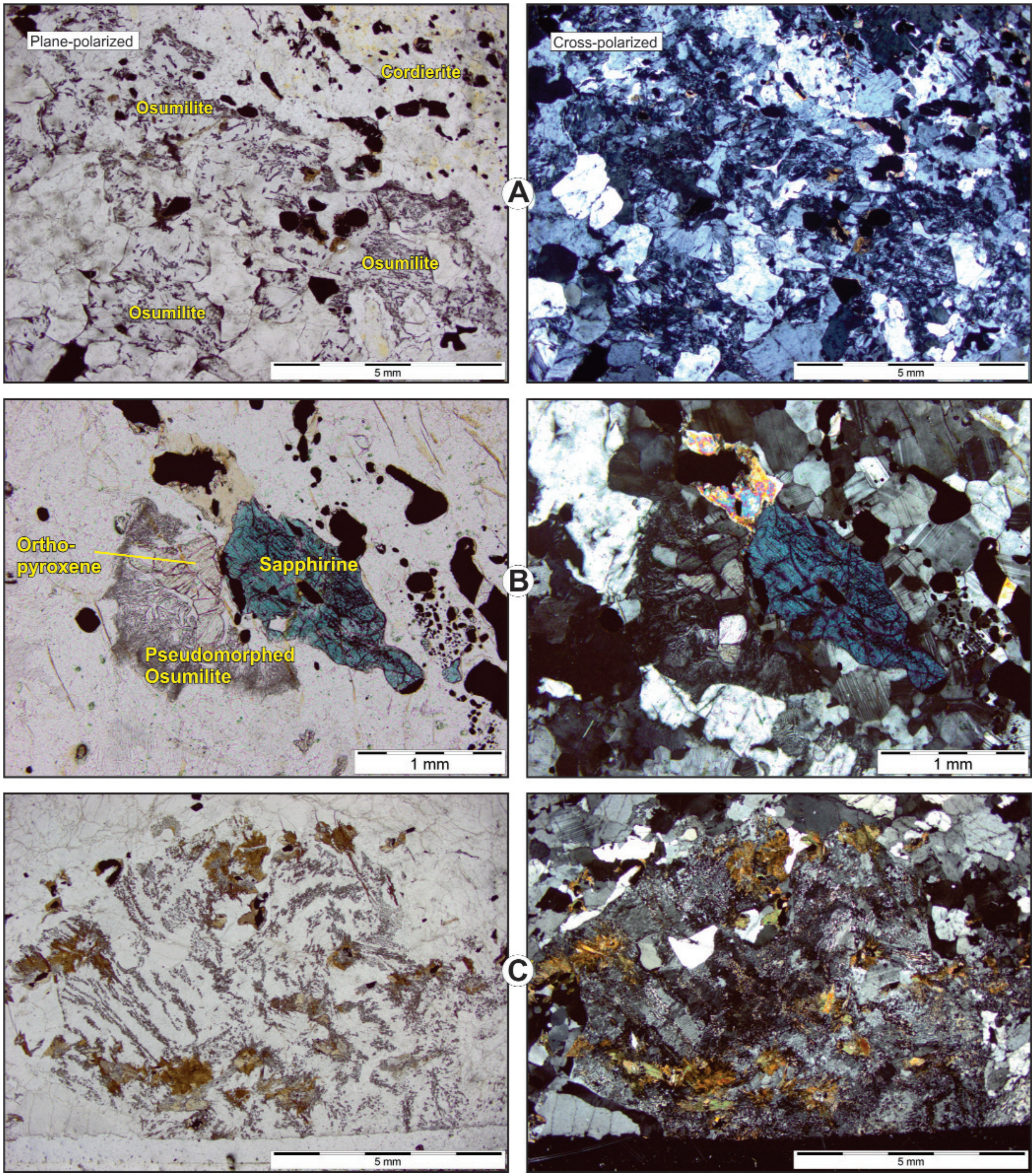
and higher pressures than the reaction:

Cordierite + Corundum = Sapphirine + Sillimanite

Arima and Gower (1991) concluded that the phase assemblages and petrographic features in the osumilite-bearing and associate metasedimentary gneisses, combined with P–T grids in the MAS, KMAS and KMAS–H_2O–CO_2 (M–MgO, A–Al_2O_3, S–SiO_2, K–K_2O) systems suggest that the peak metamorphic conditions were in the vicinity of an invariant point located at 825–950°C and 6.7–7.5 kb. Motoyoshi *et al.* (1993) reported, from an experimental study, that osumilite breaks down to cordierite + enstatite + K-feldspar + quartz at P = 11–12 kb and T = 950–1100°C. This reaction is isobaric, thus limiting the upper stability of osumilite (roughly, the base of a normal-thickness crust). M. Arima (personal communication, 2015), using recent data, concluded that pressures were probably less than 8 kb, and temperatures exceeded 850°C (for recent research *cf.* Kelsey and Hand, 2015).

Arima and Gower (1991) also delivered whole-rock chemical data for both osumilite-bearing and related osumilite-free gneisses and mineral compositional data for osumilite, sapphirine, cordierite, sillimanite, corundum, orthopyroxene, micas, spinel, feldspar and opaque minerals, the details of which are not addressed here.

Two other possible occurrences of osumilite are at sites CG86-095 and MN86-089, 17 and 33 km southeast of the SHBP intrusion, respectively (Figure 7.8C). Neither is near the SHBP intrusion, but they are adjacent to the (probably faulted) contact between the PGMB and the White Bear Arm complex. In both cases only pseudomorphs remain. The characteristic alteration to symplectic intergrowths of orthopyroxene, K-feldspar, cordierite and quartz is very convincing in MN86-089A (Photomicrograph 7.3C), but some-



Photomicrograph 7.3. *Features of Paradise metasedimentary gneiss belt osumilite-bearing pelitic gneiss. A. Osumilite partially altered to K-feldspar, orthopyroxene, cordierite and quartz (CG87-165A.2), B. Sapphirine–orthopyroxene and pseudomorphed osumilite in pelitic gneiss (CG87-170D), C. Probable pseudomorph of osumilite altered to K-feldspar, orthopyroxene, cordierite and quartz (MN86-089A).*

what less so in CG86-095. Site MN86-089 is intruded by several closely spaced, parallel, (now metamorphosed) mafic dykes that may have augmented high-temperature, anhydrous conditions conducive to localized osumilite formation.

Interpretation of regional distribution of index minerals in PMGB pelitic gneiss. The following is a brief analysis of the above observations; a thorough investigation of these rocks is still required.

Three metamorphic domains are defined (Figure 7.8D). These trend obliquely to the overall strike of the PMGB, but are parallel to the orientation of the granitoid intrusions within it, especially those in the southeastern area.

In the northeast corner (north of Hawke Bay, which is located on Figure 7.6), the diagnostic feature is that the rocks lack cordierite, or any evidence of its former presence. The rocks are characterized by quartz–feldspar–biotite–muscovite–opaque mineral assemblages. Fibrolitic sillimanite found in some muscovite flakes is interpreted as indicating a prograde reaction.

The core of the PMGB is occupied by quartz + feldspar ± cordierite ± garnet ± opaque mineral prograde assemblages, in which cordierite is partially replaced by sillimanite and/or kyanite, together with biotite and quartz. The distribution of kyanite (*vs.* sillimanite) may be controlled more by lower temperatures rather than higher pressures.

The southern flank of the PMGB is occupied by a similar prograde mineral assemblage, but replacement of cordierite occurred under anhydrous, higher temperature conditions such that orthopyroxene developed and kyanite is lacking. The osumilite-bearing rocks reflect the most anhydrous and highest temperature conditions achieved.

Geochronological data indicate that the PMGB was not severely affected by Grenvillian metamorphism (except at its northwest end, where it runs into the Lake Melville terrane) so it seems most likely that the high-grade retrograde reactions are Labradorian and postdate the emplacement of the SHBP and WBAC (as implied from osumilite contact metamorphism), but predate the emplacement of the Paradise Arm pluton. As these two emplacement events have overlapping ages (within uncertainly limits) the high-grade retrogression must have occurred at *ca.* 1640 Ma.

7.3.3.6 Diatexite (P_{3A}sx)

Diatextite is found in all parts of the PMGB (Figure 7.6) and ranges from small segments of outcrops to forming entire hilltops. The term, as used here, refers to the product of advanced melting of pelitic gneiss. The resultant end-member rock type is a white-weathering (locally buff, creamy, or pink), texturally variable granite, but there is a complete gradation from pelitic gneiss, through inhomogeneous diatexite to homogeneous diatexite. The textural variability is partly the result of grain-size variation (from fine grained to pegmatitic) that is commonly very patchy, showing localized and irregularly shaped areas of a particular textural type. It is also partly the result of darker coloured restite material having variable sizes and shapes, ranging from rounded, blebby clots to drawn-out wispy rafts, schlieren and skialiths. The white-weathering part is the neosome, composed of quartz, plagioclase and K-feldspar, and the darker coloured part (melanosome) may contain one or more of sillimanite, garnet, biotite, muscovite, cordierite, opaque oxides/sulphides and graphite.

The five samples examined in thin section (CG81-288, CG85-136A, CG85-148C, GM85-190, LC85-029) do not provide additional insights (at least, to the author). They contain the above-listed minerals, plus apatite, zircon and monazite. GM85-190 lacks K-feldspar and contains garnet. The others are sillimanite-bearing.

7.3.3.7 Ultramafic Rocks (on map as P_{3C}um)

Ultramafic rocks in the PMGB are not common, having been located at only 17 of 1400 data stations established in the belt, and, even where they were recorded, they are typically minor. Although sparse, they are found in all parts of the PMGB, and lack any obvious pattern in their distribution (Figure 7.6). In some cases, the ultramafic rocks are represented as pods less than 1 m wide, and even the larger examples are only tens of metres across. The rocks are dark-grey-, dark-green- or black-weathering, generally fine to medium grained (the central parts of some bodies may be coarse grained), unfoliated to strongly deformed, and homogenous. No evidence of discordance against their host rocks was found. Sharp contacts with pelitic gneiss were recorded at MC77-096. The rocks are not migmatized, but have experienced severe deformation, as demonstrated by tight folding (*e.g.*, MC77-096). In some cases, they show indistinct mineralogical layering (CG86-268, GM85-488, GM85-527). At GM85-488 and GM85-527, the ultramafic rocks are associated with gabbroic rocks and are interpreted to be part of small ultramafic–mafic layered intrusions. It is the author's preferred interpretation that they are sills and dykes emplaced into the PMGB, and most likely emplaced coevally with the White Bear Arm complex (hence their designation as Unit P_{3C}um in the database).

Ten samples were examined petrographically (CG85-029, CG86-268A, GM85-039, GM85-488A, GM85-527A, MC77-094B, MC77-096B, SN86-104C, SP85-049B, VN85-258). The most common mineral assemblage comprises pale-green amphibole (70–90% of the rock), with strongly pleochroic orthopyroxene and an opaque oxide, and little else (CG85-029, CG86-268A, MC77-096B, VN85-258). The 'little else' includes minor accessory plagioclase, pale-orange-brown, phlogopitic biotite, hercynite (relatively abundant in VN85-258), apatite, or sulphide. Olivine (with associated hercynite) is preserved in GM85-039, and, despite the numerous microshears transecting it, is only slightly serpentinized. Minor pale-brown, relict igneous clinopyroxene forms the cores of spectacular pseudomorphs in GM85-488A and colourless, metamorphic pyroxene is present in MC77-094B. Secondary minerals are titanite, rutile, carbonate and serpentine (all uncommon).

7.3.3.8 Mafic Plutonic Rocks and Mafic Dykes (on map as P_{3B}am, P_{3C}am, P_{3C}rg, P_{3C}d)

Excluding the mafic extrusive rocks already addressed, mafic rocks in the PMGB include small plutonic bodies and dykes (Figure 7.6).

Small mafic plutonic bodies. These comprise medium- to coarse-grained metagabbro, metagabbronorite and metanorite, locally grading into leucocratic equivalents. Most commonly, the rocks are massive or weakly foliated, homogeneous, non-migmatitic and show ophitic or sub-ophitic texture. Orthopyroxene, up to 1 cm diameter, was recorded at VN84-119. Generally, they lack coronitic textures, but coronas were recorded at three sites (*see* next paragraph). In some places (*e.g.*, GM85-488 to GM85-490) the gabbroid rocks are associated with ultramafites and, collectively, probably represent small layered intrusions. At GM85-488, pyroxenite enclaves are found in gabbro. One poorly understood ‘body’ included here occurs at the north-west end of the belt. Only two outcrops were seen. One was referred to in the field as granodiorite and the other as anorthositic gabbro. Both are non-migmatitic. Samples from the two outcrops look similar in stained slab (justifying their grouping). They are neither granodiorite nor anorthositic gabbro, but something in between – hence designation as monzogabbro. No petrographic data are available for either.

In general, the relatively undeformed state and lack of migmatization in the gabbroid rocks are taken as reasons for assigning them as mid- to late-Labradorian. The gabbroid rocks are intruded by mafic dykes, which must be therefore be later, hence designation as P_{3C}d (*e.g.*, VN85-134).

Twenty three samples were examined in thin section (CG85-124, CG85-149A, CG85-149B, CG85-310J, CG85-394, CG86-113B, CG86-150A, GM85-388A, GM85-388B, GM85-488B, GM85-519A, JS86-366, LC85-024B, MC77-224A, MN86-091, MN86-097, SN86-104A, VN85-134B, VN85-134C, VN85-243D, VN85-249, VN85-287B, VN85-504B). Three of them (MN86-091, VN85-249, VN85-287B) contain olivine, or its pseudomorphed equivalent, and show poorly developed double coronas. The coronites compare most closely to the lowest grade Type 1 coronas seen in the White Bear Arm complex (Section 11.4.4.11). In VN85-249, the orthopyroxene forms fern-like dendritic symplectic intergrowths with plagioclase. All three samples occur in the same area, 30–35 km south-east of Sand Hill Big Pond. Fifteen of the samples contain orthopyroxene, which is weakly to strongly pleochroic. Clinopyroxene is present in fifteen thin sections. All samples contain plagioclase; five contain interstitial quartz; most have orange-brown to red-brown biotite, pale-green to green-brown hornblende, an opaque oxide, and apatite. Other minerals sporadically present are pyrite, hercynite (in the coronitic samples) and scapolite (in SN86-104A). Five of the samples have textures that allow them to be called mafic granulite (CG85-124, CG85-149B, CG85-310J, MC77-224A, SN86-104A), and six of them (CG85-394, GM85-388A, GM85-388B, LC85-024B, MW84-097, VN85-504B) are best termed amphibolite.

Mafic dykes. Rocks probably derived from mafic dykes are scattered throughout the PMGB and include amphibolite gneiss and agmatitic/migmatized amphibolite, and unmigmatized amphibolite. The rocks are generally black- or brown-weathering, fine to medium grained, and have fabrics that are concordant with their pelitic or psammitic associated rocks. They occur as pods, blocks and more continuous layers that show boudinage in places. One example contains large (up to 2 cm long) plagioclase crystals in a fine-grained mafic matrix (MW84-097). Some are intruded by minor granitoid veins. Unlike the rocks considered to have a mafic extrusive protolith described earlier, these lack internal banding and are not part of a characteristic supracrustal association. Localities having rocks confidently asserted to be mafic dykes include CG86-525, CG86-528C (*cf.* Section 7.3.3.1), CG86-529 (6 dykes recorded), CG86-548, JS86-103, JS86-370, MN86-089 (two dykes) and VN85-287.

Seven thin sections examined display a wide range in mineral assemblage and fabric. Minerals present are plagioclase, quartz, orange-brown biotite, blue-green amphibole and an opaque oxide (not all minerals present in all thin sections), plus sporadic opaque sulphide, apatite, titanite, allanite and epidote. Fabrics vary from unfoliated (but recrystallized) to mylonitic. The plagioclase phenocrysts in MW84-097 have been recrystallized to polygonal aggregates.

Mafic unit assignment. The mafic rocks in the PMGB belong to several different groups, as follows: i) those thought to be extrusive and addressed earlier are given the designators P_{3A}vm or P_{3A}am on the 1:100 000-scale geological maps, ii) medium- to coarse-grained mafic intrusive rocks have been assigned the designator P_{3C}rg, iii) P_{3B}am and P_{3C}am units are considered to represent amphibolitized ‘early’ migmatized minor mafic intrusions and ‘late’ unmigmatized minor mafic intrusions, respectively, despite lack of evidence (*e.g.*, discordance) that they are dykes, and iv) where a minor mafic intrusion is confidently identified as a dyke it is given the designator P_{3C}d. Remember, although their designation in the PMGB may be speculative and no assurance can be given that a particular mafic body has been correctly assigned to its chosen unit, all these units are unequivocally established somewhere in eastern Labrador.

7.3.3.9 Dioritic Rocks (on map as P_{3B}dr and P_{3C}dr)

Dioritic rocks in the PMGB are brown- or grey-weathering, medium to coarse grained, massive to strongly foliated and homogeneous to migmatized. Occurrences are minor and scattered throughout the PMGB, although are mostly in the northeastern part of the PMGB (Figure 7.6). The rocks tend to be associated with mafic bodies and some have enclaves of fine-grained mafic rocks.

In thin section (GM85-350, GM85-356, GM85-448A, GM85-496, GM85-522, GM85-523A, VN85-265) the rocks are seen to consist of plagioclase; interstitial, poorly twinned or untwinned K-feldspar; quartz; buff-orange biotite; relict ragged hornblende; garnet (only in thin section GM85-356); an opaque oxide and apatite (not all minerals present in all thin sections). Other minerals sporadically present are an opaque sulphide, zircon, allanite, epidote and minor chlorite. Pyroxene is absent.

Given the field association with amphibolites and their similar mineral assemblages (except for the presence of K-feldspar and the minor-phase status of hornblende in the diorite), it is plausible that they are genetically linked – perhaps being the product of hybridization between metasedimentary gneiss and mafic intrusive rocks.

7.3.3.10 Granitoid Rocks (on map as Various P_{3B}, P_{3C} Units)

Granitoid rocks within the PMGB can be subdivided into four groups. Excluded here are the larger plutonic bodies of the Hawke Bay complex and the Cape Bluff Pond granitoid intrusions (Figure 7.6), which are addressed later (Sections 12.3.2 and 12.3.3).

The first group comprises pale-pink to grey, fine- to medium-grained, granodioritic gneiss that is locally gradational into dioritic or tonalitic gneiss. The rocks tend to be fairly homogeneous, but may have associated amphibolite lenses, microgranite layers and pegmatitic patches. In the northwest part of the PMGB, garnet is sporadically present, at one site up to 3 cm in diameter (VN84-143). In a few instances, field notes suggest that these rocks may have a psammitic protolith, rather than being igneous plutonic. Alternatively, the gneisses might be linked with the dioritic to granodioritic rocks of the Earl Island intrusive suite.

The second group consists of white or pink, medium- to coarse-grained or pegmatitic granitic material. It is not as inhomogeneous as the earlier described metasedimentary diatexite, but it could have a similar origin.

The third group is massive to weakly foliated K-feldspar megacrystic (or seriate textured) granitoid rock. Megacrysts are typically less than 2 cm, but up to 4 cm diameter were recorded. The rock is not migmatitic, but it carries a few pelitic gneiss enclaves (*e.g.*, VN84-104) or amphibolite pods. The occurrences may be satellite intrusions of the Paradise Arm pluton or the Cape Bluff Pond granitoid intrusions, as all the rocks are very similar.

The fourth group comprises a few small dykes (up to 40 cm wide at GM85-516) of pink syenitic rock.

The mineral assemblages of these groups, seen in 31 thin sections, include plagioclase, K-feldspar (microcline and relict perthite), quartz, buff-orange biotite, hornblende (in three samples – CG81-513, GM85-528, R61-108a), garnet (also in three samples – CG81-513, SN86-104B, VN84-337), an opaque oxide, apatite, zircon, monazite (six samples), allanite and chlorite. Rare titanite, rutile, carbonate and epidote were also noted. Not all minerals are present in every sample. The thin sections are CG81-513, CG85-049, CG85-148B, CG85-434, CG86-150C, CG86-531A, CG86-578, CG86-674, EA61-054, EA61-401, GM85-274, GM85-286, GM85-295, GM85-491A, GM85-495, GM85-508, GM85-516, GM85-519C, GM85-528C, GM85-529, JS86-429, JS86-430, MC77-093A, MC77-095A, SN86-104B, SN86-111, SP85-002,SP85-025, SP85-027B, VN84-337C, VN85-150, VN85-157 and VN85-329.

7.3.4 HAWKE RIVER TERRANE; METASEDIMENTARY GNEISS IN WHITE BEAR ARM COMPLEX

The slivers of metasedimentary gneiss addressed here are all situated in the southeast part of the White Bear Arm complex (WBAC). The larger occurrences are depicted on Figure 7.1. They were first mapped as having a metasedimentary protolith by Gower *et al.* (1987), but some data stations of Eade (1962) and Wardle (1976), in the same units but at which granitoid rocks were recorded, have been reinterpreted by the author as supracrustal where justified by his own data.

The rocks have not been well mapped or well described but are probably tectonic slivers less than 1 km wide and several kilometres long. Gower *et al.* (1987) suggested that the metasedimentary material might be more abundant than depicted on their Figure 2, but the only additional metasedimentary gneiss outcrops found by the author during later examination of roadside outcrops along the branch road to Charlottetown are strike extensions of previously mapped slivers. Their intercalation with amphibolitic gneissic rocks of the WBAC and pervasive mylonitization are the main reasons for regarding them to be thrust-bound units.

Most of the rocks are typical pink- and black-weathering (some grey or buff), fine- to coarse-grained pelitic gneiss like that found in the PMGB. Very minor calc-silicate rock, psammitic gneiss, quartzite, banded amphibolite and ocherous gossanous material are associated. Some of the rocks are schistose, especially those in muscovite-bearing outcrops near the northern fringe of the WBAC.

Given their tectonized setting, it is perhaps not surprising that they display varied mineral assemblages. Nine thin sections of pelitic gneiss were examined (CG04-001C, CG86-274, MN86-084, MN86-148, MN86-347, SN86-187, SN86-188, SN86-189, SN86-190A). All contain plagioclase, quartz, biotite (olive-green to orange-brown), an opaque oxide and prismatic sillimanite (fibrolitic in

MN86-347). Minor kyanite is present in SN86-187. K-feldspar is present in all except MN86-347 and SN86-187. Garnet is seen in two thin sections (CG04-001C, MN86-347). Partially pseudomorphed cordierite (to skeletal sillimanite and biotite) is present in three closely clustered samples (SN86-187, SN86-188, SN86-189) and suspected to have been formerly present in MN86-148. Strongly mylonitized sample SN86-190A differs in containing, what is suspected to be, pseudomorphed osumilite with orthopyroxene and sapphirine. Other minerals include accessory zircon and monazite and secondary white mica and chlorite.

In addition to the pelitic gneiss samples, two associated calc-silicate rocks were examined in thin section (CG86-453, SN86-380). Both contain plagioclase, clinopyroxene (bright apple-green – *cf.* diopside/hedenbergite), orange-brown garnet (grossularite/andradite), quartz, titanite and epidote. In addition, CG86-453 contains high-relief scapolite (*cf.* meionite) and apatite, and SN86-380 contains an opaque oxide and carbonate. The presence of meionite indicates high-temperature and high CO_2/H_2O conditions. Epidote isolates garnet from plagioclase, consistent with being the product of the retrograde reaction (written as Ca end-member compositions):

$$Grossularite + Anorthite + H_2O = Zoisite + Quartz$$

7.3.5 LAKE MELVILLE TERRANE

A general review of metasedimentary gneiss in the Lake Melville terrane is made by rock type throughout the terrane, followed by focus on selected individual areas. The reader is referred to Section 7.2 for pertinent age-constraint data.

7.3.5.1 General Review

Pelitic gneiss. Pelitic gneiss in the Lake Melville terrane (Figure 7.9; Plate 7.9A) weathers pink and black, grey, creamy, brown, or rusty orange-yellow. The brown/orange/yellow colours are due to sulphide oxidation (including sporadic Cu sulphides), which seems to be more prevalent in the Lake Melville terrane than adjacent areas (except in coastal parts of the Paradise metasedimentary gneiss belt). The rocks are typically well banded and strongly deformed to ultramylonite, and may also have a marked lin-

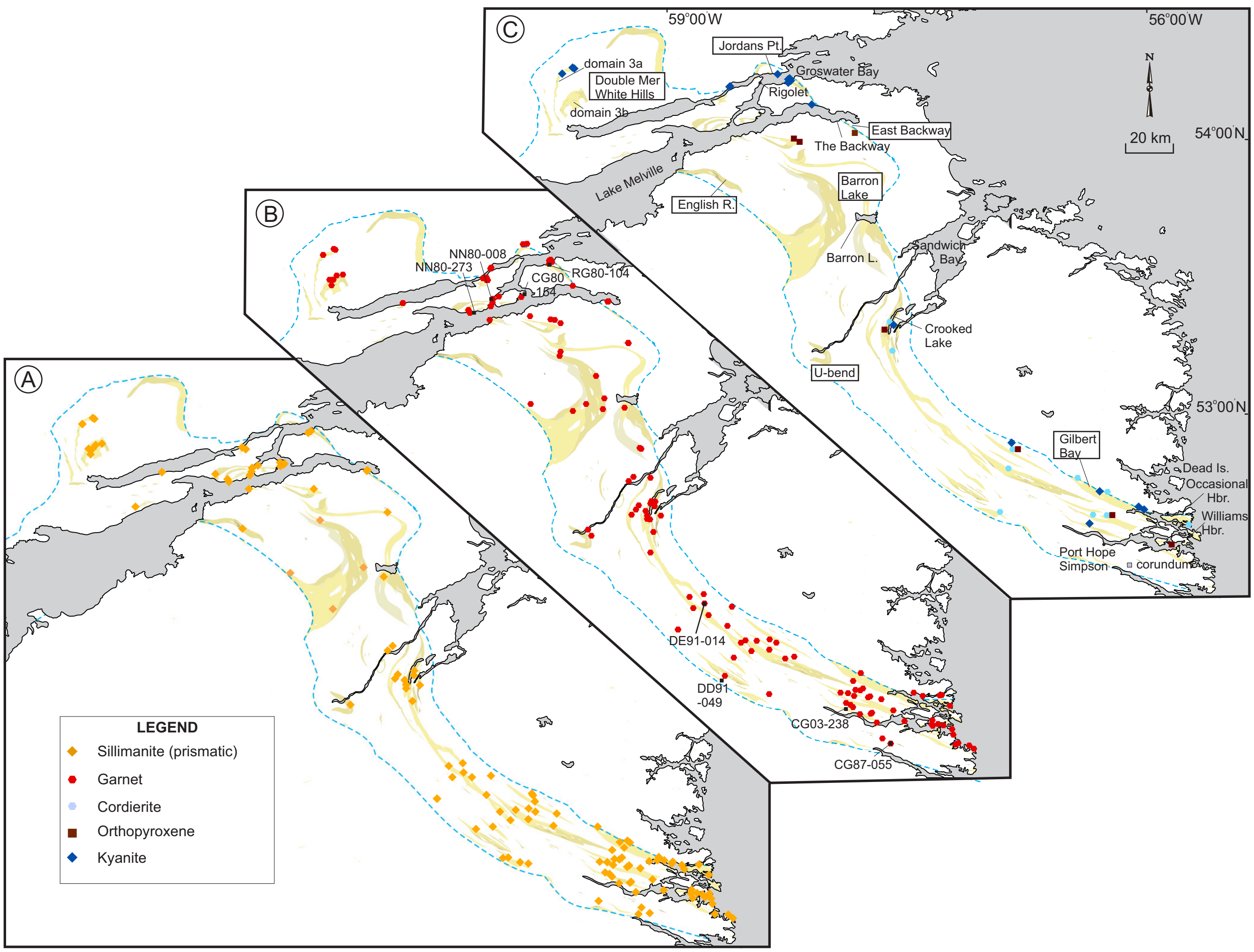


Figure 7.9. *Distribution of selected metamorphic minerals in late Paleoproterozoic pelitic gneiss in the Lake Melville terrane. A. Sillimanite, B. Garnet, C. Cordierite, orthopyroxene and kyanite.*

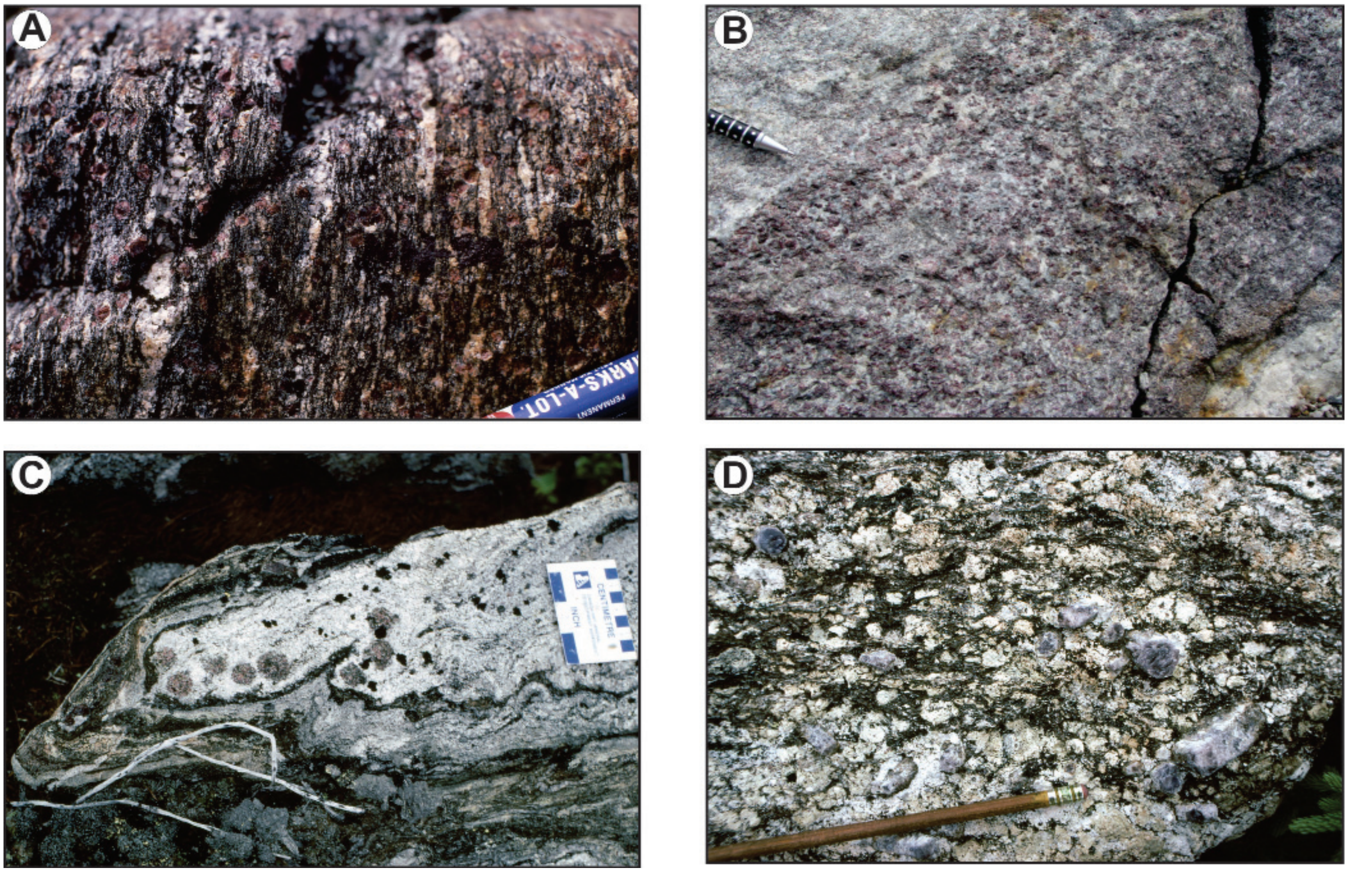


Plate 7.9. *Garnetiferous and corundum-bearing pelitic gneiss in Lake Melville terrane. A. Typical garnet–sillimanite–biotite pelitic gneiss (NN80-008), B. Very abundant garnet, showing pink colour characteristic of pelitic gneiss (CG03-238), C. Partially retrograded garnet (to biotite and plagioclase) in leucosome of pelitic gneiss (DD91-049), D. Lilac corundum (*cf.* sapphire) in pelitic gneiss (CG87-055).*

eation. They are commonly schistose. Compositional banding is expressed by veneers or schlieren of biotite-rich material that, sporadically, include sillimanite, garnet and, in places, muscovite, associated with white or pink, lensy, poddy leucosome layers. The leucosome may also be garnet bearing. A common feature is a K-feldspar porphyroblastic appearance that leads to difficulties in distinguishing the pelitic gneiss from the associated K-feldspar megacrystic granitoid rocks. The rocks are host to amphibolitic layers and lenses and common microgranite and pegmatite dykes, which may be garnet-, muscovite- or pyrite-bearing. Quartz veins are also found in places.

Stable high-grade mineral assemblages in pelitic gneiss of the Lake Melville terrane overwhelmingly involve combinations of garnet, sillimanite and biotite (along with plagioclase, K-feldspar, quartz and accessory minerals). The distribution of sillimanite and garnet is shown in Figures 7.9A and B. Repeated mention is made in field notes that garnet is mauve, lilac or rose coloured (Plate 7.9B), although dark-red garnet is also present. The cause of the colour has not been determined. One possibility is that the garnet has a higher pyrope to almandine component. Pure pyrope garnet is colourless, so more pyrope relative to almandine results in colour dilution from the deep-red characteristic of almandine. Garnet is extremely abundant in some places (*e.g.*, CG86-020, CG03-226), forming up to 80% of some layers (CG03-238). It is also quite large locally – up to 3 cm in diameter at NN80-571 and 4 cm in diameter at CG80-151. It may also be extensively or entirely retrograded to plagioclase-dominated pseudomorphs (*e.g.*, DD91-049, VN91-453; Plate 7.9C). Sillimanite was identified at most pelitic gneiss outcrops. It also occurs as inclusions in garnet in the northwest part of the Lake Melville terrane (PE82-235.1, PE82-235.2).

Cordierite and orthopyroxene are not part of stable mineral assemblages in pelitic gneiss in the Lake Melville terrane, being found only as relict minerals (Figure 7.9C). Kyanite is also a relict mineral south of Lake Melville and The Backway in the Lake Melville terrane. Cordierite occurs as relict primary grains or is completely pseudomorphed. Both modes of occurrences are restricted to the southern half of the Lake Melville terrane, from Crooked Lake to Williams Harbour.

Cordierite seen in a thin section from one of the Crooked Lake occurrences (VN84-452B) has ragged, relict form and is surrounded by garnet, biotite, plagioclase, K-feldspar and quartz. No aluminosilicate is seen in the immediate vicinity, but sillimanite is present elsewhere in the thin section. The reaction:

Cordierite + Biotite = Garnet + K-feldspar + H_2O

may apply. Cordierite in the other Crooked Lake occurrence (CG84-103) occurs as small pinitized grains in a thin section also containing garnet. The two minerals are not in contact and their mutual stability cannot be judged petrographically. Both thin sections contain common graphite.

Farther southeast, north of Port Hope Simpson, cordierite is seen as a relict phase associated with orthopyroxene, biotite, sillimanite and minor garnet in thin section JS86-282. A reaction that may be applicable is:

Cordierite + Orthopyroxene + Garnet = Biotite + Sillimanite

If garnet is a reactant, then it has been almost entirely consumed as very little is present.

For the remaining cordierite-bearing samples, including samples from the Gilbert Bay metasedimentary gneiss (CG04-024E, CG86-174, CG86-195B, CG86-387A, JS86-472, MN86-205), the former presence of cordierite is suggested by clusters of sillimanite and biotite that have the overall shape of typical cordierite grains. It seems very likely that cordierite was once present in these rocks.

Two possible occurrences of orthopyroxene south of the middle part of The Backway (RG80-410, RG80-459) are both serpentinous pseudomorphs. A third example (SG68-108) from the same area (south of the east end of The Backway) does have stable, unaltered orthopyroxene as part of a high grade mineral assemblage that includes sillimanite and garnet, but there is a possibility that this occurrence should be correlated with the northwest end of the Paradise metasedimentary gneiss belt, where other samples have similar mineral assemblages, rather than the Lake Melville terrane. Orthopyroxene in the next occurrence to the south (CG84-082), in the Crooked Lake area, is relict and largely altered to serpentine. Progressing southeast, the next two occurrences (CG86-174, JS86-282) both show biotite–quartz symplectite enveloping relict orthopyroxene, providing strong evidence for the reaction:

Orthopyroxene + K-feldspar + H_2O *= Biotite + Quartz*

Note that sample CG86-174 is also suspected to have former cordierite, and that relict cordierite is present in JS86-282. The cordierite and orthopyroxene breakdown reactions could therefore be coupled. The final thin section (SN86-419) is a hybrid rock (half pelitic gneiss and half mafic granulite) and the presence of orthopyroxene in the pelitic part could be explained as due to contamination. One other orthopyroxene-bearing pelitic gneiss sample in the area is from a tectonic sliver of pelitic gneiss within the White Bear Arm complex (SN86-190A),, which is addressed as part of the Hawke River terrane metasedimentary gneisses.

Kyanite, in a sample from the Crooked Lake area (EA61-411C), is uncertainly identified because of the poor quality of the thin section. Farther to the southeast, kyanite is relict at all localities where it was found (CG86-302, CG86-387, CG86-491, JS86-159, SN86-015B), forming irregular grains in the cores of clusters of sillimanite, to which it is partially transformed.

Despite cordierite, orthopyroxene and kyanite all being relict minerals, attention is drawn to their distribution, in that they are concentrated along the northeast side of the Lake Melville terrane and found mostly at its southeast end. Their adjacency to the Hawke River terrane is in keeping with similar mineral assemblages in pelitic gneiss in the Paradise metasedimentary gneiss belt, on the other side of the White Bear Arm complex, and all part of the close linkage between the PMGB and the WBAC. Their preservation at the southeast end of the Lake Melville terrane is suspected to be due to a lower grade Grenvillian metamorphic overprint (but still fairly high grade) than present throughout the remainder of the Lake Melville terrane.

Other minerals. Muscovite, in many cases, is demonstrably not part of the stable high-grade mineral assemblage (being a discordant posttectonic phase or forming an obvious alteration product), but such cannot be determined in all instances.

The thin sections in which muscovite appears to be stable are all from the southwestern side of the Lake Melville terrane (CG86-649, CG87-519B, DD91-049, JS87-049, VN91-466A).

Corundum occurs at CG87-055, southeast of Port Hope Simpson (Figure 7.9C), as large lilac-coloured crystals several centimetres long and over 1 cm wide (Plate 7.9D). The site has been commercially evaluated for its sapphire potential, details of which were summarized by Gower (2010c, and GSNL Mineral Occurrence 013A/08/Gem001).

An unidentified blue-green mineral, suspected to be apatite, occurs at CG80-152. Staurolite, osumilite and sapphirine have not been found in Lake Melville terrane pelitic gneiss.

Diatexite. Diatexite is typically associated with pelitic gneiss in the Lake Melville terrane as a subsidiary rock type. It is usually white-weathering, has a coarse grained, irregular texture, may be well banded and commonly contains schlieric biotite-rich layers. It discordantly intrudes pelitic gneiss locally and may contain rafts of it (*e.g.*, CG80-427). Mauve or red garnet is common (garnet up to 3 cm in diameter at CG80-531) and, sporadically, outcrop surfaces display canary-yellow staining that is associated with high radioactivity (not found in other terranes). White leucogranite sheets mentioned by Hanmer and Scott (1990), which contain pinhead garnet and are spatially associated with pelitic metasedimentary gneiss, are grouped as diatexite here.

Most of the thin sections examined (CG86-197, LG80-015, MC77-107B, MW82-107, NN80-034A, PE82-031, PE82-177, PE82-179, PE82-249, RG80-384, VN91-058B) have over 95% felsic minerals, namely quartz, K-feldspar (mostly microcline; some perthite) and plagioclase, but proportions differ. For example, MC77-107B probably has less than 5% K-feldspar. All, except two samples, have minor biotite (orange-brown) and five of the samples have subhedral to euhedral garnet with quartz inclusions, and most likely the product of incongruent melting. Accessory minerals sporadically present include an oxide opaque mineral (sulphide in VN91-058B), zircon (in most samples), and monazite (rare). Secondary minerals are rutile, white mica, chlorite and rare epidote, allanite and prehnite.

Psammitic gneiss. Psammitic gneiss is dark-grey, brown, buff, or ocherous-weathering. It tends to be fairly well banded, but may be finely laminated or have a crude flaggy appearance. Although layering is largely the product of metamorphism, compositional and grain size contrasts between individual layers, not readily attributable to partial melting, may well reflect original primary bedding (*e.g.*, CG83-486, CG84-025). Psammitic gneiss is easily confused with granitoid orthogneiss and clues to protolith commonly rely on interlayering with pelitic gneiss, calc-silicate rocks or quartzite. The rocks are generally garnetiferous, garnet being reported to be 5 x 8 cm at NN84-021.

Sixty thin sections from the Lake Melville terrane metasedimentary gneiss are deemed to have a psammitic protolith (Plate 7.10A), although an alternative granitoid protolith is possible for some of them (CC87-055, CG04-054B, CG80-151A, CG80-154D, CG80-156A, CG80-426, CG80-587, CG84-008, CG84-025, CG84-029, CG84-030, CG84-181, CG84-346B, CG86-072, CG86-082, CG86-231, CG86-310, CG87-154, DE91-109B, EA61-008, EA61-039, EA61-405B, EA61-410, JS86-169, JS86-178, JS86-195, JS86-257, JS86-492, JS87-197, MC77-129A, MN86-231, NN80-258A, NN80-273A, NN80-549A, NN84-008, NN84-028, NN84-220D, NN84-400, NN84-441, PE82-234A, RG80-001, RG80-474B, RG80-532, RH-017, SN86-005, SN86-168, SN86-183, SN86-243, VN84-450B, VN84-473, VN84-474, VN87-134B, VN91-048, VN91-061, VN91-064, VN91-069A, VN91-094, VN91-108A, VN91-111, VN91-161A). Of the felsic minerals, all sections contain plagioclase and quartz, and 75% have K-feldspar (microcline, perthite, or untwinned/exsolved). In many of the samples, the abnormally high quartz content (compared to most granitoid rocks) is one criterion in assigning their protolith to be psammitic. Muscovite is present in 15% of the samples and biotite in all of them. As in the pelitic gneisses, biotite is dominantly red-, buff- or orange-brown. A subset of the samples (about 15%) contains hornblende ± clinopyroxene. These are interpreted as calcareous psammite. Orthopyroxene is present in three samples (CG04-054B, NN84-441, RG80-532), one of which also contains hornblende and clinopyroxene (NN84-441). Garnet is present in 60% of the thin sections and appears to be a stable, prograde phase. It is commonly poikiloblastic (*e.g.*, CG80-151, CG80-426, CG86-072, CG86-231, CG86-310, MN86-231, NN80-258), containing inclusions of quartz, an opaque oxide, plagioclase and, less commonly, biotite. A narrow rim of plagioclase separates garnet from quartz in SN86-005. The garnets show little or no sign of deformation, suggesting that they are syn- to late-Grenvillian. The opaque mineral is mostly oxide, but sulphide is also present in about 15% of the samples. Apatite, zircon, monazite, allanite (in order of abundance) are variously present and secondary minerals include

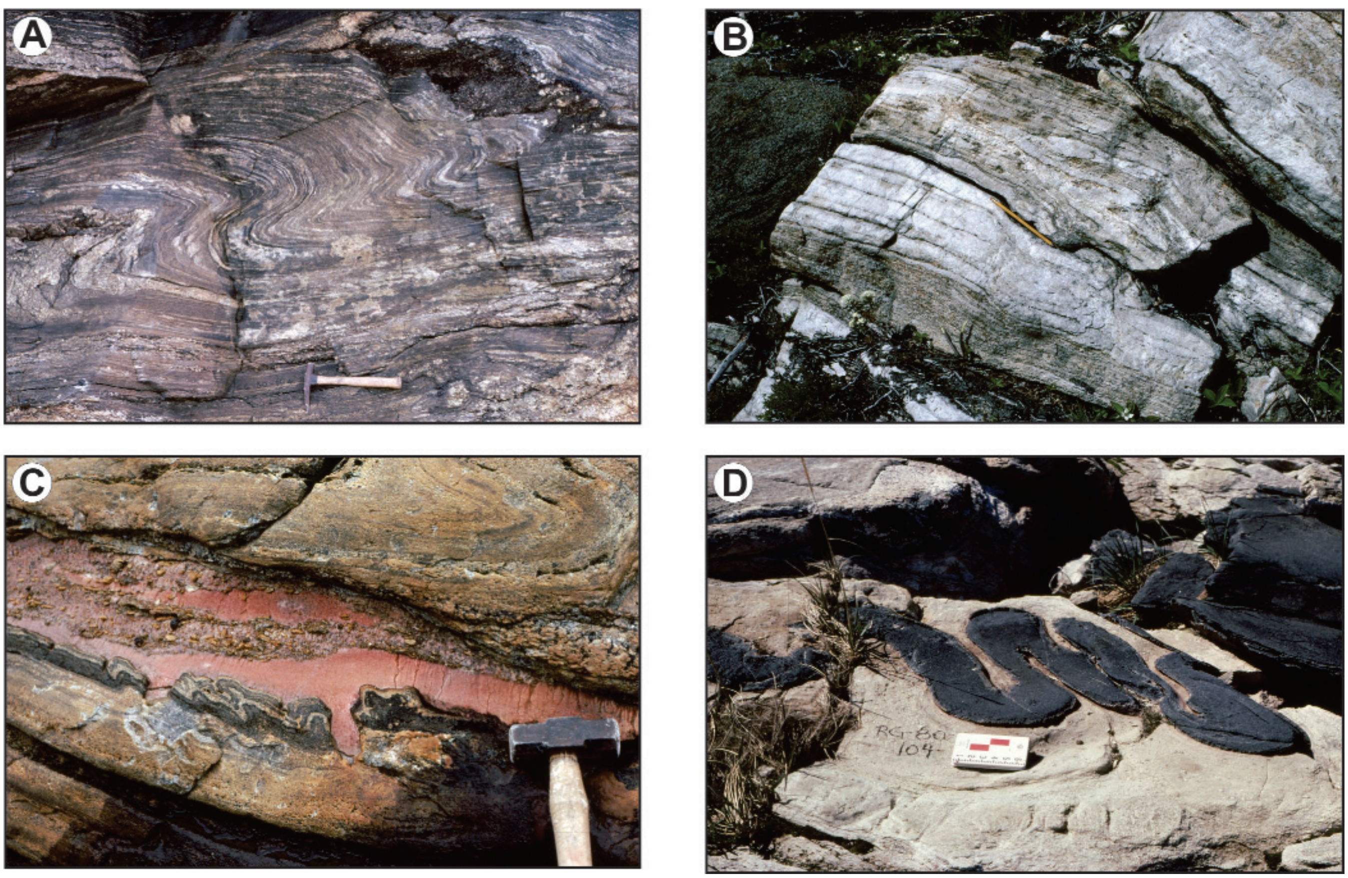


Plate 7.10. *Psammitic gneiss, quartzite and marble in Lake Melville terrane. A. Psammitic gneiss with some semi-pelitic layers (CG80-154), B. Quartzite (CG87-055), C. Pink marble with psammitic and pelitic gneiss. Note protrusion of marble into boudinaged dark layer underneath (NN80-273), D. White marble intruded by now-buckled mafic dyke (RG80-104).*

titanite, white mica, chlorite and epidote. The accessory minerals and mafic silicates commonly show a 'dispersed' pattern that the author has come to associate with quartzofeldspathic supracrustal rocks (especially well seen in EA61-039 and EA61-405B).

Quartzite. Quartzite is white- or grey-weathering, and found in association with pyritic pelitic or psammitic gneiss or calc-silicate rocks (Plate 7.10B). It is massive to thickly bedded, having quartzite layers separated by biotite- or feldspar-rich horizons. Layering at 1 to 3 cm is most typical, but the rocks may be finely laminated or massive. The total thickness of a given quartzite layer is normally less than 2 m, but a 10-m-thick quartzite layer was recorded at VN87-289. Garnet, graphite and magnetite are commonly associated minerals.

None of the samples examined in thin section (CG03-354E, CG80-170B, CG80-423B, CG83-487, EA61-425A, JS86-163, LG80-014A, MC77-114A, MC77-143A, NN80-265, NN80-274B, NN80-592, NN84-475, RG80-472, VN84-440) are pure meta-orthoquartzite, although the quartz content is too high for any rock name, other than quartzite, to apply. Further subdivision is possible on the basis of those containing significant feldspar, mica, Ca-bearing minerals, or Fe-rich minerals being gradational into psammitic gneiss, pelitic gneiss, calcareous quartzite, or cherty iron formation, respectively. Samples of some mineralogical interest are: i) CG83-487, which contains sillimanite needles in garnet cores, ii) EA61-425A, which contains polysynthetically twinned amphibole (*cf.* grunerite), and iii) LG80-014 and MC77-114, both of which contain substantial graphite. Graphite in Mesoproterozoic metaquartzite in the New Jersey Highlands is suggested by Volkert *et al.* (2000), on the basis of field, petrographic, and carbon and sulphur isotopic data, possibly to have been derived from algal mats. No carbon or sulphur isotopic data are available from eastern Labrador for any graphitic metasedimentary rocks, so biotic *vs.* abiotic origin remains unevaluated.

Calc-silicate rocks. Most calc-silicate rocks in the Lake Melville terrane are relatively minor, occurring as layers rarely more than a few ten's of centimetres thick. Quite commonly, the rocks occur as green- or grey-weathering pods, lenses or boudins within pelitic or psammitic gneiss. At CG80-183 and CG80-186, calc-silicate rocks form agmatized breccia zones of elongate pods within a semi-pelitic envelope. Apart from hornblende and plagioclase, most of the occurrences have diopside and carbonate as major phases. Pink marble (Plate 7.10C) occurs as 10- to 20-cm-thick layers interlayered with pelitic and psammitic gneiss, into which they discordantly protrude (NN80-273, NN80-274). White marble intruded by a mafic dyke occurs at RG80-104 (Plate 7.10D).

An extensive thin section collection of calc-silicate rocks is available from the Lake Melville terrane metasedimentary gneiss (CG03-227D, CG80-167C, CG80-170A, CG80-172, CG80-186, CG84-105, CG86-025, CG86-279B, CG87-142A, DE91-014C, MC77-130B, MW84-035A, NN80-009A, NN80-009B, NN80-009C, NN80-013, NN80-100A, NN80-273B, NN80-273D, NN80-274A, NN80-274C, NN80-571, NN80-572, SN86-011A, SN86-431, SN86-435, VN87-391, VN87-393). The rocks are texturally heterogeneous and represent a wide variety of mineral assemblages. The detailed study necessary to establish all viable stable parageneses has yet to be attempted. Minerals present in about half of the thin sections (but not all the same ones) are plagioclase, K-feldspar, quartz, pale-orange-brown mica (phlogopitic), tremolite/actinolite, an opaque oxide and titanite. Diopside is present is all but three sections, being absent in CG03-227D, NN80-009A and VN87-373. Sample NN80-009A contains forsterite instead, and the other two are retrograded to amphibole-bearing assemblages. Sample CG80-172 also contains forsterite and is on strike with NN80-009, being about 10 km farther east, both located in the Henrietta Island area, south of Rigolet. Bright orange-brown garnet (grossularite/andradite) is present in DE91-014C, NN80-273D (Photomicrograph 7.4A) and NN80-274, and is associated with vividly green diopside and scapolite. Paler coloured orange-brown garnet and paler green diopside are present in MC77-130B (which also contains scapolite), and colourless garnet is present in NN80-571, SN86-011 and VN87-393 (relict scapolite in NN80-571). Scapolite is present in almost half of the thin sections. Other minerals sporadically present include sulphide, apatite, zircon and secondary allanite, epidote, white mica and chlorite. Wollastonite and vesuvianite are additional phases that might be anticipated, but, if either is present, it has escaped recognition. Sample NN80-273B is of particular interest in that it is a marble containing small, very rounded, ovoid (detrital?) grains of quartz, plagioclase, K-feldspar and diopside in a calcite matrix that makes up about 90–95% of the rock (Photomicrograph 7.4B).

Scapolite and quartz have reacted to give grossularite in DE91-014C (Photomicrograph 7.4C), possibly reflecting the reactions (*e.g.*, Mathavan and Fernando, 2001):

Anorthite + Calcite = Scapolite

and

*Grossularite + 2*CO_2 *= Anorthite + Calcite + Quartz*

Other reactions for which there is some petrographic evidence (several thin sections) are:

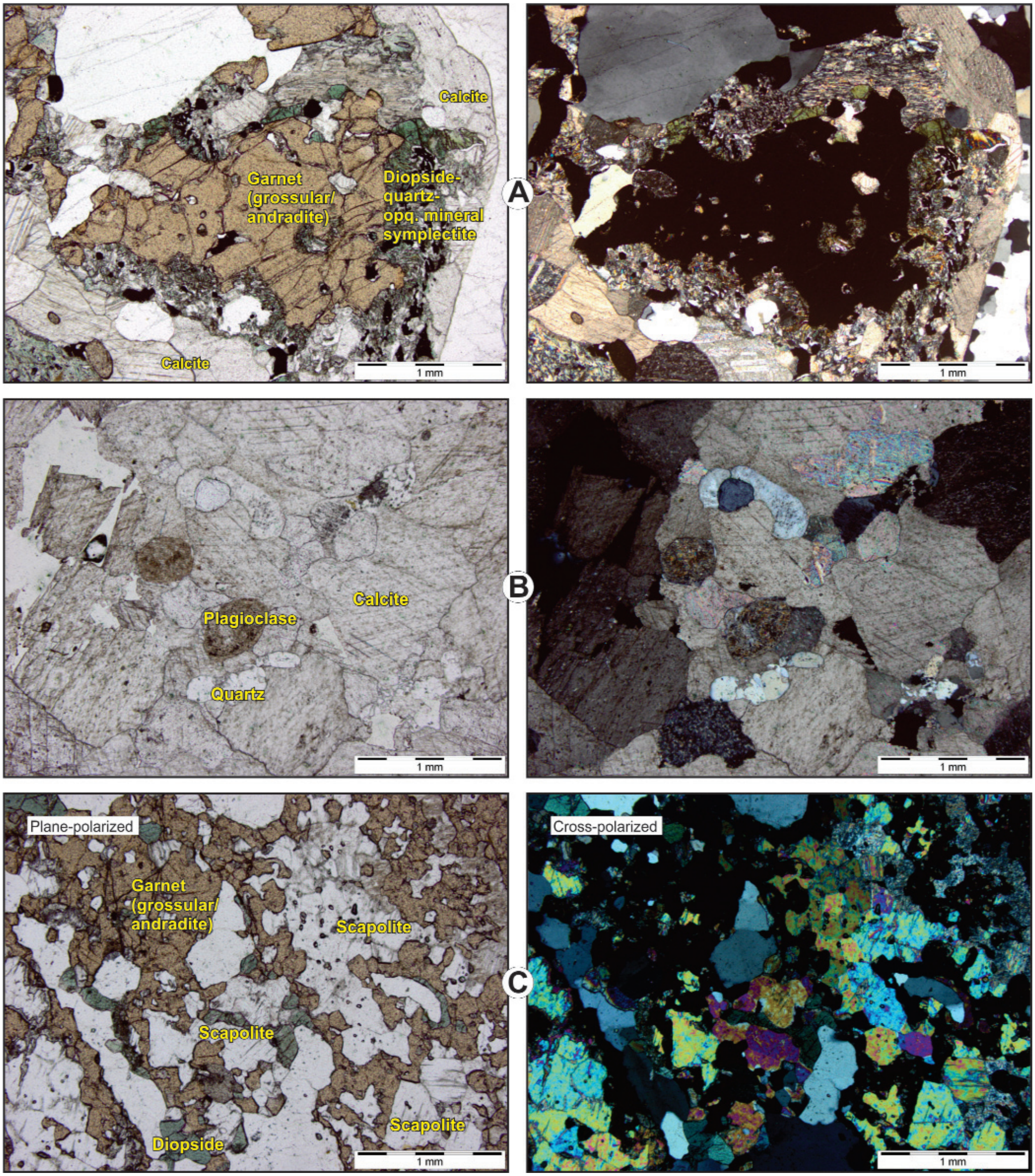
Diopside + Anorthite + CO_2 *=Amphibole + Calcite + Quartz*

plus an unquantified reaction which appears to have generated a diopside, quartz, ± an opaque mineral symplectite between grossularite/andradite and calcite (NN80-273D).

Mafic rocks. These include amphibolite, mafic granulite and metamorphosed mafic dykes within Lake Melville terrane metasedimentary gneiss. Amphibolite occurs as concordant layers and lenses that most probably represent the remnants of former mafic dykes. Metamorphosed mafic dykes that discordantly intrude metasedimentary gneiss occur at CG04-001, CG86-053 and CG86-711. Mafic granulite is a subsidiary rock, having rather sporadic distribution.

Three mafic granulites (CG86-309, NN84-444A, NN84-476A), and one retrogressed granulite (VN91-264C), were examined in thin section. All contain plagioclase, orthopyroxene, biotite (orange-brown), an opaque oxide and apatite. The unretrogressed granulite samples also contain clinopyroxene, garnet and hornblende (no hornblende in NN84-444A). Other minerals present in the retrogressed sample VN91-264C are interstitial quartz (also in CG86-309), hornblende, sulphide, chlorite and epidote. Sample CG86-309 is of particular interest in that garnet is surrounded by veneers of plagioclase, and clinopyroxene is rimmed by polygonal orthopyroxene. Locally, juxtaposed garnet and clinopyroxene are separated by veneers of orthopyroxene, plagioclase and quartz. These mineral relationships suggest the reaction (written retrogressively):

Garnet + Clinopyroxene + Quartz =
Plagioclase + Orthopyroxene



Photomicrograph 7.4. *Lake Melville terrane calc-silicate rocks. A. Calc-silicate rock showing diopside–quartz–opaque mineral symplectite between garnet (grossularite/andradite) and calcite (NN80-273D), B. Impure marble containing minor amounts of felsic minerals (NN80-273B), C. Calc-silicate rock dominated by garnet, diopside and scapolite (DE91-014C).*

Granitoid rocks. Granitoid rocks, including dioritic rocks, tend to form separate, mappable units in the Lake Melville terrane, albeit tectonically interleaved with metasedimentary gneiss. Rock types include granite, syenite, monzonite and K-feldspar megacrystic rocks. The K-feldspar megacrystic units are of genetic relevance to the metasedimentary gneiss, inasmuch as they appear to be gradational into it in places (*e.g.*, CG86-734, CG86-741).

Thin sections examined include two dioritic rocks (CG80-152, VN91-463A), one quartz monzonite (VN91-173), one alkali-feldspar syenite/quartz syenite (EA61-726B), two tonalitic to granodioritic rocks (DD91-043, VN91-264A.1), and the remainder granite (CG80-594, DD91-039, MC77-113A, MH86-009C, MN86-235B, NN80-033, NN80-100B, NN80-273C, NN84-022, NN84-213, PE82-125, PE82-165, PE82-166, VN91-264A.2). Several of the samples could have pelitic or psammitic origin, especially PE82-165 and PE82-166 (pelitic?), and CG80-152, CG80-594 and VN91-463A (psammitic?). Minerals present include plagioclase, K-feldspar, quartz, biotite (green-buff to orange-brown), hornblende (5 samples), garnet (9 samples), orthopyroxene (only in NN84-213), an opaque mineral, sulphide (4 samples), apatite, zircon, and less commonly monazite, allanite, and secondary white mica and chlorite. Not all minerals are present in every thin section.

The following sections address specific areas of supracrustal rocks within the Lake Melville terrane.

7.3.5.2 Double Mer White Hills Metasedimentary Gneiss (P_{3A}sp, P_{3A}sc, P_{3A}ss)

The Double Mer White Hills metasedimentary gneiss is situated north of Double Mer forming a thrust-soled structural outlier (Figures 7.4 and 7.9). It has been generally been assigned to the Lake Melville terrane, mainly on the basis of having a similar high-grade metamorphic character. On the other hand, if comparable rock types are emphasized instead, then the rocks are equally in keeping with the Mealy Mountains terrane.

The presence of metasedimentary gneiss in the area was recognized by Stevenson (1970) and mapped at 1:100 000 scale by Gower (1984, 1986). It occurs as two gently south-dipping sheets, alternating with monzonitic granulite, which the author interprets to be due to structural interleaving during thrusting. The metasedimentary package consists of kyanite–garnet ± sillimanite pelitic gneiss, associated with calc-silicate rocks and minor quartzite (Plate 7.1C, D). The gneisses tend to have an evenly, regularly banded, somewhat flaggy appearance. This is related more to strong mylonitization rather than migmatization. A well-developed stretching lineation is also present. In pelitic gneiss, kyanite, sillimanite and mauve garnet are easily recognizable in the field. The proportion of calc-silicate rocks is subjectively judged to be higher than elsewhere in metasedimentary gneiss in eastern Labrador, albeit still very subordinate. It tends to be much darker-weathering than the pelitic gneiss and commonly has alternating green/grey, red, or (less commonly) white layers, reflecting concentration of diopside, garnet and carbonate, respectively. It may contain quartzofeldspathic leucocratic partings or quartz stringers. The associated quartzite is either white- or grey-weathering. It contains mauve garnet and pyritic layers. At CG83-392, multiple pyritic bands are present, up to 1 m thick and guesstimated to contain up to 30% sulphide.

Gower and Erdmer (1988) assigned the structurally lower sheet of metasedimentary gneiss to metamorphic domain 3a (zone 3.1 of Gower, 1986; thin sections CG83-052, CG83-329B, CG83-329C, CG83-381) and the structurally upper sheet to metamorphic domain 3b (zone 3.2 of Gower, 1986; thin sections CG83-352B, CG83-386, CG83-389, CG83-391, CG83-392B) (Figure 7.4). The principal difference between the two is that kyanite is the dominant Al_2SiO_5 polymorph in domain 3a, whereas sillimanite is the exclusive Al_2SiO_5 polymorph in domain 3b. In both domains, associated minerals are plagioclase, perthitic K-feldspar, quartz, orange-brown biotite, mauve garnet, an opaque oxide, spinel and zircon. K-feldspar and an Al_2SiO_5 polymorph are lacking in sample CG83-329C. Muscovite and apatite are lacking in all samples, in contrast to their presence in similar, but lower-grade, pelitic gneiss to the north and west.

Despite pelitic gneiss being the dominant metasedimentary rock type, the author's collection of calc-silicate thin-sections is larger (CG83-051, CG83-196, CG83-204, CG83-329A, CG83-341, CG83-342, CG83-347B, CG83-348, CG83-352A, CG83-354, CG83-362, CG83-365, CG83-380, CG83-387, CG83-392A.1, CG83-392A.2, CG83-402, CG83-411A, SG68-229B). The only mineral in common to all of them is plagioclase, but minerals found in most are hornblende, clinopyroxene, garnet, an opaque oxide and apatite. Other common minerals are orange-brown biotite and scapolite. Sporadically present are orthopyroxene, sulphide, titanite, zircon, rutile, allanite, white mica, chlorite and epidote, some of which are secondary. Gower (1986) made the following observations regarding changes with increasing metamorphic grade in the calc-silicate rocks: i) amphibole changes from pale-green to olive-green/brown, ii) titanite and epidote are eliminated, iii) garnet and scapolite appear, and iv) titanite and garnet seem to be incompatible.

Two thin-sectioned samples cannot be grouped as either pelitic gneiss or calc-silicate rocks. One is composed almost entirely of quartz, apart from a sulphide opaque mineral and minor apatite (CG83-411B). The other contains mostly garnet, lesser quartz and an opaque sulphide mineral, plus trace plagioclase, apatite, biotite (inclusion in garnet) and secondary carbonate (CG83-392C). Both samples are interpreted by the author as having a metachert or lean banded iron formation protolith.

7.3.5.3 Jordans Point Metasedimentary Gneiss (P_{3A}ss, P_{3A}sc)

Rocks referred to as the Jordans Point metasedimentary gneiss (name introduced here) are situated at the Lake Melville–Groswater Bay terrane boundary near Rigolet (Figure 7.1). The rocks were included as part of Stevenson's (1970) granitic gneiss unit and were remapped at 1:100 000 scale by Gower *et al.* (1981). Gower *et al.*'s mapping in the area was done at an early stage of the eastern Labrador project and, in hindsight, rock identification suffered from inexperience with these high-grade, intensely deformed units.

Follow-up petrographic work by the author has alleviated some of the field rock-identification deficiencies, but further investigation is needed.

A step in the right direction was made by Corrigan *et al.* (2000), who carried out a geochronological study in the area, addressing timing of deformation and metamorphism (as opposed to time of deposition, which was reviewed in Section 7.2). Three of ten samples in their study are relevant to the Jordans Point metasedimentary gneiss. Sample 6 is from a pre- to syn-thrusting pegmatite from the north side of Groswater Bay that intrudes mylonitized rocks of uncertain protolith (but within the Jordans Point metasedimentary gneiss package). The pegmatite shows comparable strain to that in the host rock. The pegmatite yielded an upper intercept Labradorian age of 1619 +16/–14 Ma and a lower intercept of 1120 +141/–126 Ma. The age is based on a regression that included two nearly concordant, single zircon analyses and five single-grain monazite analyses. Sample 8, only 170 m farther east, is mylonitic kyanite-, garnet-bearing pelitic gneiss. Kyanite blades are parallel to the extension lineation and were interpreted by Corrigan *et al.* (2000) to be syn-thrusting. Two single monazite grains yielded ages of 1612 ± 5 Ma and 1610 ± 2 Ma. Sample 9, pelitic gneiss comparable to sample 8 but from the south side of Groswater Bay and at a higher structural level, yielded two single-grain, slightly reversely discordant, $^{207}Pb/^{206}Pb$ monazite Grenvillian ages of 1047 ± 4 Ma and 1046 ± 4 Ma. In a companion quantitative geothermobarometric study, Tulk (1996) determined that P–T conditions reached 11–13 kb and 830–860°C in the area. These results thus provide evidence for thrusting and granulite-facies metamorphism during relatively late stage Labradorian orogenesis and another high-grade event during Grenvillian tectonism.

Apart from the above-mentioned pegmatite and pelitic gneiss, a wide range of rock types is present and includes: i) muscovite biotite schists interlayered with pegmatite, ii) white-weathering muscovite garnet diatexite, iii) psammitic gneiss (some informally termed metagreywacke in the petrographic database), iv) quartz-rich psammitic gneiss grading into quartzite, v) rocks suggested to have a volcaniclastic protolith in field notes, but perhaps the product of extreme deformation of some other protolith, vi) marble (RG80-104, Plate 7.10D), vii) hornblende- and epidote-rich rocks that may represent retrograded calc-silicate rocks, viii) extremely garnetiferous rocks (commonly also quartz rich) of uncertain protolith having garnet up to 2 cm across, showing plagioclase-rich retrograde rims, and forming to about 25% of the rock (CG80-008, CG80-028) (possibly meta-banded iron formation), ix) banded mafic rocks suggested to have a volcanic protolith (RG80-469), but having banding that may be the product of severe deformation, x) intercalated K-feldspar megacrystic rocks, and xi) garnet–pyroxene–hornblende, granulite-facies mafic rocks forming narrow layers or lensoid, boudinaged form that are probably former mafic dykes, sills, or (originally) larger mafic intrusions.

The rocks are commonly continuously and evenly banded, or show lensy fabrics, due to intense mylonitization (straight gneiss). Also present are rocks that have a flaggy or fissile appearance and a well-developed fracture cleavage. It seems likely that these are the product of multiple deformational events. Added to these fabrics are discordant and concordant pegmatites, probably of various ages, and products of late-stage, brittle deformation (brecciation and hematitic, chloritic, saussuritic, or silicic alteration) related to extensional faulting in the late Neoproterozoic Lake Melville rift system.

Twenty-six thin sections are available from the Jordans Point metasedimentary gneiss, and represent a good spectrum of the rock types present. Six thin sections of kyanite-bearing pelitic gneiss from four samples (RG80-019.1, RG80-019.2, RG80-025, RG80-027A, RG80-464.1, RG80-464.2) have plagioclase–K-feldspar–quartz–biotite–garnet–kyanite assemblages, although plagioclase is very scarce in RG80-464.1. An important additional mineral is sillimanite (RG80-019.2, RG80-464.1 and .2), which appears to be stable with kyanite. Minor opaque oxide(s) and apatite are sporadically present, as is secondary white mica and chlorite. RG80-464.1 also contains minor corundum. The co-existence of kyanite and K-feldspar (except perhaps RG80-464.1) means these rocks can be assigned to bathozone 6 (pressures exceeding 7 kb). One of the two pelitic gneiss samples of Corrigan *et al.* (2000) enlarges the known extent of kyanite-bearing pelitic gneiss along strike to the northwest of the author's sample sites.

Six thin sections of muscovite schist were prepared from five samples all within a 2.3 km strike length (CG79-656, CG79-657A.1, CG79-657A.2, CG79-657B, CG80-214, CG80-215). All samples contain plagioclase, quartz and strongly aligned muscovite, and all but one have an opaque oxide. K-feldspar, zircon, apatite, monazite are present in some thin sections, and sporadically present secondary minerals are chlorite, white mica, allanite and epidote. Relict garnet was seen in one thin section, and recorded in hand sample in two others. Clearly these rocks have lower-grade mineral assemblages than the above-described kyanite-bearing pelitic gneiss, but it is not evident to the author whether the muscovite schists are retrograded equivalents or original metamorphic assemblages. They are located in a subsidiary, northwest segment of the unit and could be separated from the kyanite-bearing rocks by major thrusts.

Rocks deemed to be of psammitic origin examined in thin section include CG80-001, RG80-035, RG80-299, RG80-300, RG80-466, RG80-467B and RG80-468. The rocks contain (apart from sporadic exceptions of specific minerals) plagioclase, K-feldspar, quartz, orange-brown biotite, hornblende, garnet, an opaque oxide, apatite and zircon, with secondary white mica and chlorite. Features favouring a psammitic protolith are: i) higher than typical quartz content, compared to most granitoid rocks, ii) common to abundant garnet, and iii) field association with other, more diagnostic metasedimentary gneisses. The presence of amphibole is taken to imply a calcareous component in the sediment. Interpretation of a (calcareous) psammitic protolith is, of course, far from conclusive. That rocks of calcareous psammitic origin are present, however, is supported by associated thin-sectioned samples of calc-silicate rocks (CG80-

014C, CG80-028A, CG80-028B). The key minerals present are clinopyroxene (CG80-014C), orange-brown garnet and scapolite (the latter two minerals in all three samples). In addition, a small occurrence of nearly pure marble (RG80-104) is associated, consisting mostly of calcite, with K-feldspar, quartz, clinopyroxene, garnet, apatite (all minor).

Other rocks present examined in thin section include quartzite, having mostly quartz with minor plagioclase, K-feldspar, red-brown biotite, common garnet, an opaque oxide, sulphide and apatite (RG80-027C), and two mafic to intermediate clinopyroxene–orthopyroxene–garnet granulite samples (RG80-467C, RG80-469).

7.3.5.4 English River Metasedimentary Belt (P_{3A}sq, P_{3A}ss, P_{3A}sp)

This group of rocks comprises a 20-km-long by 2-km-wide belt, extending inland in an east-southeast direction from Lake Melville (Figure 7.9). Metasedimentary gneiss close to Lake Melville was first recorded by Emslie (field notes, 1975), but the presently interpreted distribution of metasedimentary rocks in the whole belt is based on the mapping of Gower *et al.* (1981). The name 'English River metasedimentary belt' is newly introduced here.

From field descriptions, the essential feature of the rocks is that they are unusually quartz-rich compared to most metasedimentary gneiss in the Lake Melville terrane, reaching near-pure orthoquartzite in a few places. Magnetite and, less commonly pyrite or garnet, are sporadic accessory minerals. Generally, however, the rocks are referred to in field notes as 'dirty' quartzite and described as interlayered with feldspar + biotite + quartz ± amphibole psammitic rocks and(or?) granodioritic gneiss. Biotite-rich layers are found in places, but, with one exception (NN80-613), the rocks lack an aluminosilicate mineral. Muscovite is recorded as disseminated in quartzite and found in associated white-weathering pegmatite and quartz pods. Some of the pegmatite is discordant. Amphibolite layers, garnetiferous locally, are common and interpreted in field notes to be remnants of mafic dykes.

Of the 17 thin sections from the belt, only three are deemed, convincingly, to be quartzite. Two of them (NN8-141A, NN80-141B) lack feldspar and, apart from quartz, contain only muscovite, minor biotite and sparse secondary hematite and chlorite. The third sample NN80-144A, differs in having minor plagioclase and K-feldspar and traces of secondary minerals, but is too quartz-rich not to be quartzite. Other associated rocks examined in thin section accepted as having a metasedimentary protolith are diopside-bearing calc-silicate rock (NN80-463) and pelitic gneiss characterized by fibrolitic sillimanite and orange-brown biotite (NN80-613). The remainder of the rocks in the belt lack features providing unequivocal protolith identification. A psammitic protolith seems most probable for those having both muscovite and a 'clastic-looking' texture (NN80-144B, NN80-145A, NN80-442, NN80-445), whereas for the remainder, apart from field association, either a granitoid or psammitic protolith seem equally viable (NN80-140, NN80-436, NN80-440, NN80-443, NN80-472A, NN80-614, NN80-616). A mela-amphibolite to ultramafic rock (NN80-451) consisting largely of hornblende, with interstitial plagioclase and quartz, buff-orange biotite, and apatite is probably an intrusive rock. Note that garnet was not seen in any thin section from the belt, in keeping with its sparse mention in field notes.

7.3.5.5 East Backway Metasedimentary Gneiss Sliver (P_{3A}ss, P_{3A}sp, P_{3A}vm)

The East Backway metasedimentary sliver is situated at the boundary between the Groswater Bay and Lake Melville terranes at the southeast end of The Backway. It might seem hardly justifiable to single it out, as it is only about 15 km long and less than 1 km wide. The reason for doing so, however, is to highlight a rather diverse package of supracrustal rocks that includes sillimanite–garnet pelitic schist, quartzite, psammitic gneiss, calc-silicate rocks, possible intermediate volcaniclastic rocks and mafic volcanic rocks (which, despite severe deformation, retain hints that they were once pillowed). Some of the rocks are very pyritic and also host minor Cu mineralization (*cf.* Beavan, 1980). This assemblage is intruded by granodioritic rocks, gabbro/diabase and pegmatite. Knowledge of the rocks is based on two clusters of outcrops mapped by Gower *et al.* (1981). One cluster is on the south shore of The Backway and the other inland about 12 km to the southeast. Based on their regional on-strike alignment and correlation with a strong linear positive magnetic anomaly (Figure 5.8), the two groups of outcrops are considered to be related. The most similar packages of rocks elsewhere in eastern Labrador are situated in the southeast Sandwich Bay and Dead Islands areas in the Paradise metasedimentary gneiss belt, with which the East Backway metasedimentary sliver could be an attenuated northwest extension.

Thin sections are available from samples representing both clusters of outcrops. From the shoreline of The Backway, thin sections of sillimanite-garnet pelitic schist and gneiss were investigated (RG80-113, RG80-114A.1, RG80-114A.2) and one of garnet-bearing psammitic gneiss (RG80-114B). One fine-grained amphibolite of undetermined probable protolith (RG80-114C) was also examined. Two thin sections from the inland cluster are a granulite-facies calc-silicate rock (CG80-758A) and quartzite/psammitic gneiss, also having a granulite-facies mineral assemblage.

7.3.5.6 Barron Lake Metasedimentary Gneiss Package (P_{3A}sx, P_{3A}ss, P_{3A}sp)

The Barron Lake metasedimentary gneiss package is one of the least confidently defined assemblages in the Lake Melville terrane. The area is very poorly exposed and few of the rocks have characteristics that allow certain identification. Rocks having non-supracrustal protoliths are also present and it is questionable whether or not metasedimentary gneisses are even dominant. Information here is based on mapping by Gower *et al.* (1982b) and subsequent follow-up studies by the author.

The rocks are a mixture of pelitic gneiss, metasedimentary diatexite and psammitic gneisses. Calc-silicate rock and quartzite are very rare (but are present at RG80-440, RG80-442). That pelitic gneiss is unequivocally present is indicated by garnet- and sillimanite-bearing rocks at data stations GF81-141, RG80-441 and RG80-442, and by abundant graphite at GF81-141 and CG81-488. Garnet at these sites is the mauve/lilac colour characteristic of high-grade pelitic gneiss. Associated diatexite is white- or ocherous-weathering, has variable grain size and, generally, only diffuse banding. Both inhomogeneous and homogeneous types are present. One strong indication of the genetic affiliation of the diatexite with pelitic gneiss is that it also carries mauve/lilac garnet. The rocks considered to be psammitic gneiss are less certainly identified. The presence of garnet and foxy red-brown biotite are helpful indicators, but not diagnostic. Some so-called psammitic gneiss could have a granitoid plutonic protolith. Mafic rocks are a minor rock type.

All the rocks deemed to have supracrustal protoliths (of varied nature) examined in thin section carry plagioclase, K-feldspar (microcline or perthite), quartz, red-brown biotite, and an opaque oxide (CG81-476, CG81-479, CG81-488D, CG81-488E, GF81-141, RG80-530). Apatite and zircon are present in most thin sections. Minor sulphide and monazite are sporadically found, and secondary minerals include white mica, chlorite and serpentine. Garnet is present in four thin sections, graphite in three, pseudomorphed orthopyroxene in three, and sillimanite in one (GF81-141, but reported also to be present in CG79-476 and CG81-479 by van Nostrand (1988)). The identification of orthopyroxene is uncertain as no primary material remains and the pseudomorphs are not well preserved. Its (former) presence is consistent, however, with orthopyroxene-bearing pelitic gneiss mineral assemblages found in the adjacent, northwest end of the Paradise metasedimentary gneiss belt.

Five other rocks from the belt seen in thin section are two granitic to granodioritic rocks (CG81-493), a monzonite (GF81-216B), an intermediate granulite (CG81-515) and a mafic granulite (GF81-246A). The monzonite lacks quartz, the intermediate granulite lacks biotite, and both granulites contain amphibole, clinopyroxene, orthopyroxene and garnet.

7.3.5.7 Gilbert Bay Metasedimentary Gneiss (P_{3A}sp, P_{3A}ss, P_{3A}sx)

The Gilbert Bay metasedimentary gneiss forms a belt of northwest-trending rocks about 60 km long and 3 km wide. Although grouped here as part of the Lake Melville terrane, this belt of rocks is less tectonized than most of the metasedimentary gneiss slivers elsewhere in the terrane. The rocks are bounded to the north by the White Bear Arm complex, and to the south by granitoid units. Knowledge of the metasedimentary belt is based on mapping by Eade (1962), Wardle (1976) and Gower *et al.* (1987). The informal name 'Gilbert Bay metasedimentary gneiss' is newly introduced here.

The gneisses are mostly grey, pink and black, rusty or brown-weathering, well-banded, fine- to coarse-grained, migmatized pelitic rocks. They are strongly deformed and commonly intensely mylonitized, in places being recognizable only by incorporated lenses of less-deformed rock. Muscovite- or muscovite–chlorite schists are common in the Gilbert Bay metasedimentary rocks, the schistosity having been imposed on pre-existing migmatite. Field notes repeatedly refer to the rocks as being porphyroblastic and easily confused with associated K-feldspar megacrystic rocks. There is clearly a close spatial relationship between the two rock types. Where contacts between the two rock types are not mylonitized, they are transitional. Gower *et al.* (1987) suggested that the K-feldspar megacrystic rocks were derived from the pelitic gneiss by porphyroblastesis of the gneiss and its eventual homogenization.

Pelitic gneiss is associated with minor psammitic gneiss, calc-silicate gneiss, pyritic quartzite and pyritic gossans. Pyritic gossans are up to 40 m wide (JS86-462) and are especially abundant south of Occasional Harbour. Other associated rocks are banded amphibolite and ultramafic pods and layers. The pelitic gneiss is discordantly intruded by metamorphosed mafic dykes (*e.g.*, JS86-473, MN86-187), deformed quartzofeldspathic veins, two generations of granitic dykes (one group of which can be linked to the Gilbert Bay pluton, *see* Section 17.1.1.1), and pegmatite (some of which is muscovite bearing). Late stage, cataclastic shear zones comprising sulphide- or hematite-rich material are seen locally, and green chloritic schist is also be present. The impression gained is that the metasedimentary gneiss is similar overall to that in the PMGB, but the overall package differs in having common minor granitoid intrusions and low-grade superimposed metamorphism. The differences can be attributed to more pervasive syn- and late-Grenvillian effects.

Twenty-two thin sections of Gilbert Bay pelitic gneiss are available (CG04-024E, CG86-382, CG86-387A, CG86-460, CG86-463, CG86-491, EA61-392, JS86-159, JS86-345, JS86-472, JS86-474, JS86-478, MN86-187A, MN86-196, MN86-205, MN86-258, MN86-263, SN86-303, SN86-305, SN86-393, SN86-394B, SN86-396). A feature common to almost all thin sections is the presence of a very strong mylonitic fabric. All contain plagioclase, quartz, biotite (olive-green to orange-brown). Most have K-feldspar, an opaque oxide, zircon and sillimanite. Seen in less than half to a third of the thin sections are muscovite and monazite. Sporadically present are garnet (CG86-491, JS86-472, MN86-263, SN86-303, SN86-393, SN86-394B, SN86-396), kyanite (JS86-159, CG86-387A, CG86-491) and cordierite (CG04-024E, CG86-387A, JS86-472, MN86-205). Garnet is found mostly at the southeast end of the belt. Sillimanite is mostly prismatic, but is fibrolitic in a few thin sections. Recall that kyanite is rare in the Lake Melville terrane and is mostly found on its northeast side (as applies to these instances). Cordierite is mostly pseudomorphed by sillimanite and biotite. It is suspected to have been present in other thin sections, but now obliterated by pseudomorphing and deformation.

Five non-pelitic rocks from the Gilbert Bay metasedimentary gneiss belt were also examined in thin section. They do not have much in common, being: i) orthopyroxene- and garnet-bearing sulphide/silicate-facies iron formation (CG04-074), ii) gabbronorite (CG86-210), iii) metagabbro (JS86-473B), iv) amphibolite (MN86-187B), and v) diorite (JS86-477).

7.3.5.8 U-bend Metasedimentary Gneiss (P_{3A}ss, P_{3A}sx, P_{3A}sp)

The U-bend metasedimentary gneiss is shown on 1:100 000-scale maps for the area (Paradise River and Southeast Mealy Mountains maps) as a southeast-closing folded unit having an unfolded total strike length of 20 km and a width of 2 km (Figure 7.9). This depiction is substantially influenced by interpretation of aeromagnetic patterns. Outcrop is poor in the area and many of the rocks do not have obvious protoliths, so representation of the unit must be considered approximate at best. It is based on field observations by Cherry (1978a, b), Gower *et al.* (1985) and Gower and van Nostrand (1996).

The rocks were mostly described in the field as granitic to granodioritic gneiss. Some features, however, suggest a clastic supracrustal origin, as follows: i) sillimanite-bearing pelitic gneiss and white-weathering diatexite are commonly associated rock types, ii) the rocks are repeatedly described as highly garnetiferous, and, locally, the garnets have the characteristic mauve colour of pelitic gneiss (*e.g.*, NN84-044), and iii) field notes commonly suggest that the rocks could (alternatively) be psammitic gneiss. One key locality is CG84-435. Here, the outcrop is dominated by metasedimentary diatexite containing abundant garnet with sillimanite-bearing, biotite-rich schistose bands. The diatexite here contains numerous calc-silicate enclaves. Elsewhere in the unit, amphibolite enclaves are common.

Three pelitic gneiss samples (CG84-435B, MC77-158A, NN84-062), two psammitic gneiss samples (CG95-050, NN84-044) and one calc-silicate rock (CG84-435A) were examined petrographically. Apart from the calc-silicate rock, all have plagioclase, K-feldspar, quartz, orange-brown biotite, garnet and opaque oxide and zircon. Pale-green (actinolitic?) amphibole and strongly pleochroic orthopyroxene are additional non-secondary minerals in CG95-050, and sillimanite is present in MC77-158A and NN84-062 (also seen in outcrop at CG84-435). Thin section CG84-435B shows rims of K-feldspar around garnet and may indicate the reaction:

Biotite + Sillimanite + Quartz + Plagioclase
= Garnet + K-feldspar + Melt

in which all the sillimanite has been consumed.

The mineral assemblage in the calc-silicate rock comprises highly sericitized plagioclase; pale-green poikiloblastic amphibole; moderately pleochroic, high-relief orthopyroxene; orange-brown biotite and quartz. Despite field association and name applied, this is not a typical calc-silicate rock, so it may have a different protolith (mafic igneous rock?).

7.3.6 MEALY MOUNTAINS TERRANE

7.3.6.1 East Mealy Metasedimentary Gneiss (P_{3A}sp, P_{3A}ss, P_{3A}sx)

As depicted on the 1:100 000-scale maps of Gower (2010a) and Figure 7.1, half of the East Mealy metasedimentary gneiss (new informal name; EMMG abbreviation) is within the Mealy Mountains terrane and the other half in the Lake Melville terrane. The distribution is based on mapping by Gower *et al.* (1981, 1982b), Gower and van Nostrand (1996) and van Nostrand (1992). The extent of the metasedimentary gneiss is open to re-interpretation as exposure is very poor in the region (although it improves as the MMIS is approached). It is equally possible that the metasedimentary gneiss in the Lake Melville terrane is not linked to that in the Mealy Mountains terrane at all. The uncertainty is partly due to a total lack of outcrop in the vicinity of the narrow neck between the two parts, and partly to difficulty in interpreting magnetic data. Both northwest and southeast of the narrow neck, well-defined magnetic lows correlate with zones of severe deformation on the ground. In the vicinity of the neck, however, a magnetic low is lacking and there is hint of a northeast-trending magnetic fabric instead (Figure 5.8). As the Lake Melville terrane–Mealy Mountains terrane boundary is interpreted elsewhere in eastern Labrador to be a crustal-scale fault on either side of which considerable displacement may have occurred, the continuity implied by linking the two parts of the East Mealy metasedimentary gneiss is in conflict with the structural interpretation. The current, problematic, interpreted distribution of units is not likely to be much improved by additional field work. High-resolution magnetic data would be helpful. The description below centres on the western part of the East Mealy metasedimentary gneiss, but strays into the northeast tongue (in the Lake Melville terrane), and the southeast tongue (at the boundary between the Mealy Mountains and Lake Melville terranes).

The East Mealy metasedimentary gneiss is, overwhelmingly, pelitic, with which metasedimentary diatexite is commonly associated. Other rock types present having supracrustal protoliths are psammitic gneiss, calc-silicate rocks and quartzite. Minor mafic rocks (amphibolite to granulite facies) and small granitoid bodies are also present.

The key feature of the pelitic gneiss is its change in appearance and mineral assemblage from east to west as the Mealy Mountains intrusive suite is approached (Figure 7.10). The changes are related to increasing metamorphic grade along with accompanying dehydration. The progressive changes are subdivided into four groups, namely: i) sillimanite ± garnet, lacking cordierite, ii) sillimanite ± garnet, with cordierite, iii) cordierite + orthopyroxene, lacking sillimanite, and iv) orthopyroxene, lacking cordierite and sillimanite.

Pelitic Group 1 (sillimanite ± garnet; lacking cordierite). Metasedimentary gneiss in the eastern part of the East Mealy metasedimentary gneiss is mainly well-banded (locally mylonitic) pelitic gneiss consisting of black or grey, 1–5-

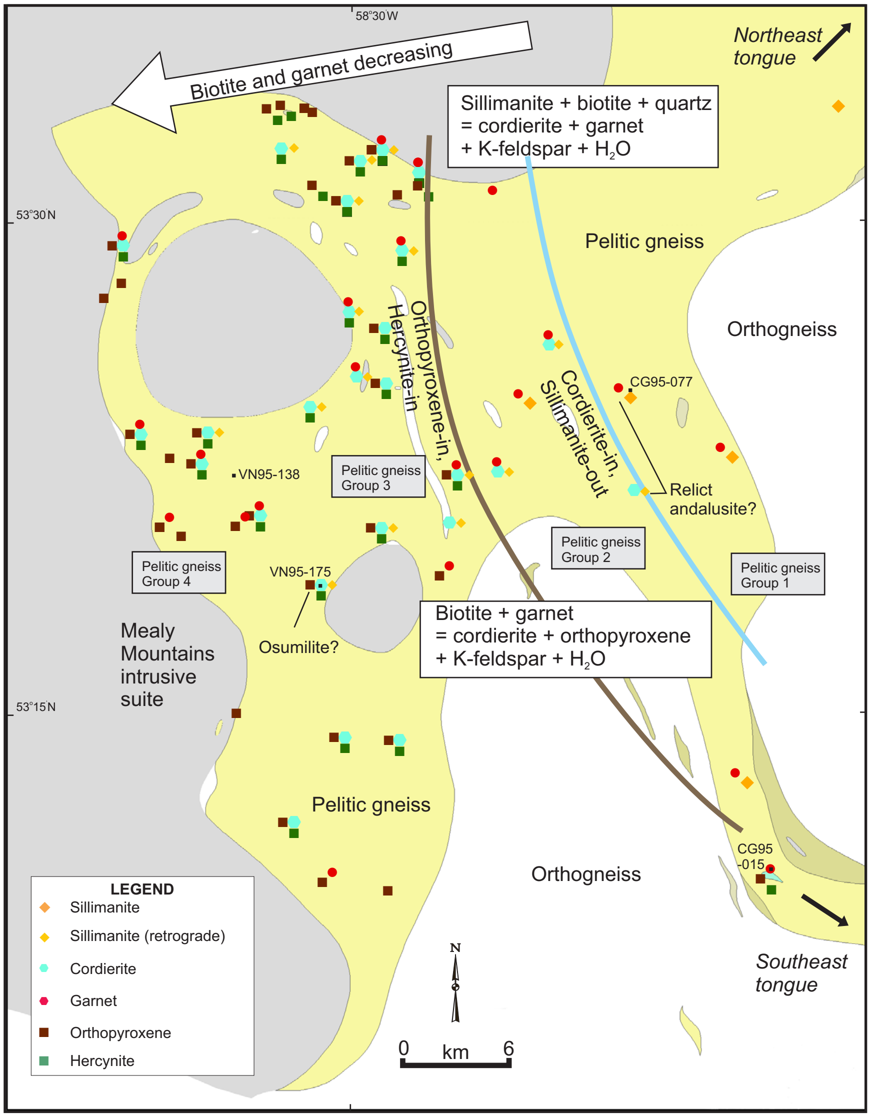


Figure 7.10. *Progressive metamorphism in pelitic East Mealy metasedimentary gneiss.*

mm-wide, schlieric sillimanite + garnet + biotite ± muscovite melanosome and 10–50-cm-wide irregular layers of creamy-white or pink K-feldspar + plagioclase + quartz ± garnet leucosome. Garnet (mauve to burgundy) is very abundant, in some layers making up to 30% of the rock. It forms grains commonly about 1 cm across, but locally up to 6 cm in diameter. Typically, it is partially retrograded and also occurs as elliptical aggregates elongated in the plane of the gneissosity. Minor quartzite, rare calc-silicate gneiss and medium-grained amphibolitic pods and layers are also present.

The most common mineral assemblage includes garnet-, sillimanite- and biotite-bearing assemblages (CG81-567A, CG84-224B, CG84-226, CG84-338, CG91-040B, CG95-077, EA61-070, RG80-436, VN84-482, VN91-439, VN95-040, VN95-069, VN95-087, VO81-518). Quartz is anhedral, polygonal to interlobate, and shows grain-size reduction in places. Plagioclase is anhedral, poor to moderately twinned, and lightly to moderately sericitized. Much of the K-feldspar is well-twinned microcline, but some untwinned material and bead/stringlet perthite is present is places. Biotite forms aligned flakes showing a range of colours, including foxy red-brown, buff orange-brown and olive-green. Garnet is anhedral and shows rounded or irregular grain boundaries. Inclusions of quartz and, less commonly, plagioclase are present. Fluid inclusions are present in garnet in CG95-077 (Photomicrograph 7.5A). Sillimanite occurs as prismatic crystals, generally associated with biotite. Ubiquitous accessory minerals are zircon and an opaque mineral. The latter is mostly oxide (in part hematized or limonitized), but graphite is locally present. Minor corundum is present in CG84-224B and VN84-482. Muscovite occurs in RG80-436, VN84-482 and VN91-439, but it is equivocal whether it belongs to the stable metamorphic assemblage or is retrograde. It may be that conditions for its breakdown (through dehydration reactions) were exceeded, but perhaps only marginally, hence either allowing some muscovite to be preserved, or, perhaps more likely, more readily facilitating its reappearance (*e.g.*, if fluids had not been expunged). Muscovite may have been eliminated by the reaction, or re-introduced as part of the same reaction in reverse:

Muscovite + Quartz = Sillimanite + K-feldspar + H_2O

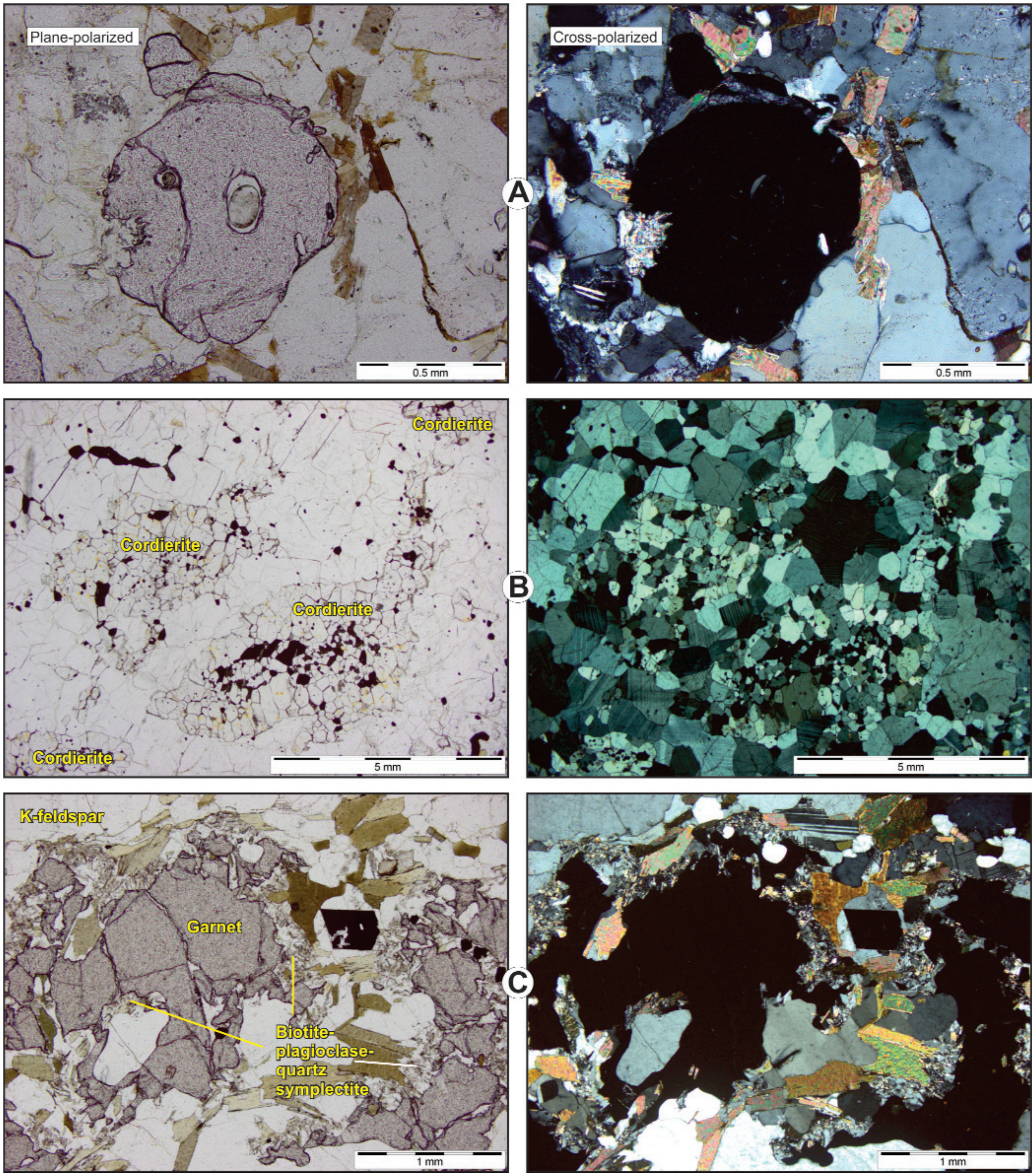
Similarly, garnet may have formed (and subsequently retrograded, *e.g.*, RG80-436), according to the reaction:

Sillimanite + Biotite + Quartz = Garnet + K-feldspar + H_2O

In sample CG81-567A, which comes from a site adjacent to MMIS in the English River fault zone, small grains of polygonal plagioclase surround sillimanite sheaves, separating it from K-feldspar, suggesting a reaction involving sillimanite and K-feldspar as reactants and plagioclase as a product, but it is not obvious from the thin section as to what the complete reaction might be.

Several thin sections, possibly transitional to more psammitic compositions, carry garnet, but no sillimanite (CG81-578, CG81-583, CG81-663, CG81-684, DE91-147A, GF81-308, VN95-014A, the last one listed only having garnet seen in outcrop). Two thin sections carry sillimanite, but no garnet (CG81-622, DD91-167). DD91-167 is distinctive in having minor hercynite in the cores of sillimanite sheaves.

Pelitic Group 2 (sillimanite ± garnet; with cordierite). Progressing west, pink-weathering pelitic gneiss is less common, displaying grey or rubiginous shades instead. Banding, plus leucosome–melanosome contrasts, also becomes less distinct, being replaced by heterogeneously textured diatexite showing marked variation in grain size in irregular patches, coupled with more swirly, convoluted and discontinuous gneissosity. The rocks are characterized by



Photomicrograph 7.5. *East Mealy metasedimentary gneiss. A. Fluid inclusions in garnet in pelitic gneiss (CG95-077), B. Polygonized cordierite in pelitic gneiss (VN95-175), C. Psammitic gneiss showing garnet (+K-feldspar) breakdown to biotite + plagioclase + quartz (CG95-015).*

cordierite, which is easily recognized as blue–mauve glassy crystals, locally up to 3 cm across but more typically less than 1 cm. Biotite decreases westward.

Cordierite-bearing pelitic gneiss in the East Mealy metasedimentary gneiss was examined in 13 thin sections (CG81-657, CG95-080, CG95-082, CG95-095A, CG95-111, CG95-126, CG95-146A, GF81-147.1, GF81-147.2, GF81-157, VN91-432, VN95-078A, VN95-097). All samples contain undulose quartz; moderately to well-twinned, anhedral, lightly sericitized plagioclase; poorly twinned, homogeneous, or finely perthitic K-feldspar, orange-brown to red-brown biotite, an oxide opaque mineral, and zircon. Most of them contain garnet and cordierite in stable association, but also show cordierite breakdown to sillimanite (mostly fibrolitic) and biotite. Hercynite is present in some samples. Sample VN91-432 is distinct, in that it contains a stable association of cordierite, prismatic sillimanite and garnet, but it is also remote from the other cordierite-bearing localities, so it may belong to a separate pressure–temperature regime. Two localities lacking evidence of breakdown of cordierite to sillimanite and biotite are CG81-657 and GF81-147, both of which are close together and near the contact with the Mealy Mountains intrusive suite. Thin section CG95-082 is of particular interest in that it contains relict andalusite, enveloped in an elliptical mosaic (strain shadow?) of quartz, secondary muscovite and carbonate. The andalusite is clearly a metastable relict, rather than part of the equilibrium assemblage, but it is of interest in that it suggests a relatively low-pressure prograde path of metamorphism. It is also the only confidently identified occurrence of andalusite known in eastern Labrador (although a very tentatively identified second example may occur in cordierite-lacking sample CG95-077, which is located 5 km to the north).

Of the numerous cordierite-generating reactions that are possible, those in accord with the observed westward decrease of biotite, sillimanite and quartz and increase in K-feldspar and garnet (and, conversely, their retrogressive reactions) include the following:

Sillimanite + Biotite + Quartz = Cordierite + K-feldspar + H_2O

Sillimanite + Biotite + Quartz = Cordierite
+ Garnet + K-feldspar + H_2O

Pelitic Group 3 (cordierite + orthopyroxene; lacking prograde sillimanite). Farther west still, banding in the metasedimentary gneisses is very diffuse and indistinct, being defined by wispy, dark-grey- or black-weathering, generally fine-grained, granulitic-textured, melanocratic schlieren/lenses within a more-or-less homogeneous monzonitic- to granitic-looking host (Plate 7.11A). Many of these rocks superficially resemble granitoid rocks of the MMIS to the west, and have a similar magnetic signature. The fine-grained nature of the granulitic melanosome makes mineral identification difficult in the field, although quartz, garnet, orthopyroxene, plagioclase and a magnetic opaque mineral can normally be confidently identified. Cordierite was strongly suspected in the melanocratic schlieren as they commonly have a dark blue-grey hue, but the mineral could only rarely be positively recognized in these fine-grained aggregates. Quartz-rich and rubiginous-weathering layers occur locally and are interpreted to be related to quartzite and sulphide-rich mudstone protoliths, respectively.

Pelitic gneiss containing both cordierite and orthopyroxene was examined in thin sections CG81-659, CG95-084, CG95-092, CG95-121, CG95-257, CG95-290, CG95-293, CG95-326B, CG95-335, GF81-145A, GF81-145B.1, GF81-145B.2, GF81-151.1, GF81-151.2, VN95-101, VN95-109, VN95-136, VN95-140 and VN95-175 (Photomicrograph 7.5B). Three additional thin sections are grouped here because they occur in the same area, although they lack orthopyroxene (CG95-126, CG95-146A, VN95-097). In these rocks, the orthopyroxene (undoubtedly hypersthene) is dark-coloured and strongly pleochroic. It forms equant, high-relief grains, commonly sheathed in retrograde bastite. The intensity of pleochroism increases from east to west, and in the westernmost samples, shows bright hues of sea -green, brick-red and yellow. Cordierite forms both single grains and polygonized aggregates and is easily recognizable because of its yellow pleochroic haloes around inclusions, well-developed sector twinning, a distinctive pattern of microfractures, and pinnitic alteration. In VN95-175, cordierite has a subtle vermiform texture that is very reminiscent of osumilite. Both the orthopyroxene and cordierite are locally altered to green/orange serpentinous products.

The rocks all contain plagioclase, K-feldspar, quartz, biotite, an opaque mineral, zircon and hercynite. Plagioclase typically shows polygonal form, is moderate to well twinned and only very lightly sericitized; K-feldspar occurs as very fine stringlet and bead perthite, such that the grains appear homogeneous in thin section except under high magnifications. Only rarely is there any inversion to microcline. Quartz is anhedral, and typically shows granoblastic, interlobate texture. Biotite is orange-brown to red-brown, as part of the high-grade metamorphic assemblage and as a retrograde mineral at cordierite grain boundaries. Hercynite is very common, being found in every thin section examined in this group. Garnet normally has anhedral form and contains a variety of inclusions, such as cordierite, quartz, biotite, orthopyroxene and opaque oxide. Note that the rocks lack an Al_2SiO_5 polymorph, except sillimanite as a retrograde mineral associated with biotite as breakdown products from cordierite.

The elimination of sillimanite, decrease in biotite and garnet, appearance of orthopyroxene and increase in cordierite and hercynite can be accounted for by potentially applicable reactions as follows:

Garnet + Sillimanite = Cordierite + Spinel + Quartz

Biotite + Garnet = Cordierite + Orthopyroxene
+ K-feldspar + H_2O*,*

Biotite + Quartz = Orthopyroxene + K-feldspar + H_20

Although all samples contain biotite, it is a very minor phase in some of them and, along with garnet (which looks relict, at least in part), may be left over from the above reactions. Furthermore, the cordierite- and orthopyroxene-rich rocks may not represent original protolith compositions, but represent the residue from melting, along the lines of the reaction:

Biotite + Garnet + Quartz + Albite + H_2O
= Cordierite + Orthopyroxene + Melt

Farther south, similar cordierite- and hypersthene-bearing metasedimentary gneiss generally lacks garnet. Four samples were examined in thin section (CG95-290, CG95-293, CG95-326B, VN95-175). In most respects (except for the absence of garnet), these samples are similar to the cordierite- and hypersthene-bearing metasedimentary gneisses described above. One notable difference is the presence of corundum in CG95-293 and VN95-175, the latter sample also lacking quartz. In many samples, but especially obvious in VN95-175, it is clear that cordierite aggregates are the result of polygonization of larger grains commonly over 0.5 cm in diameter. An opaque oxide and hercynite are concentrated along cordierite grain boundaries, especially at triple points, and seemingly were products developed during the polygonization process.

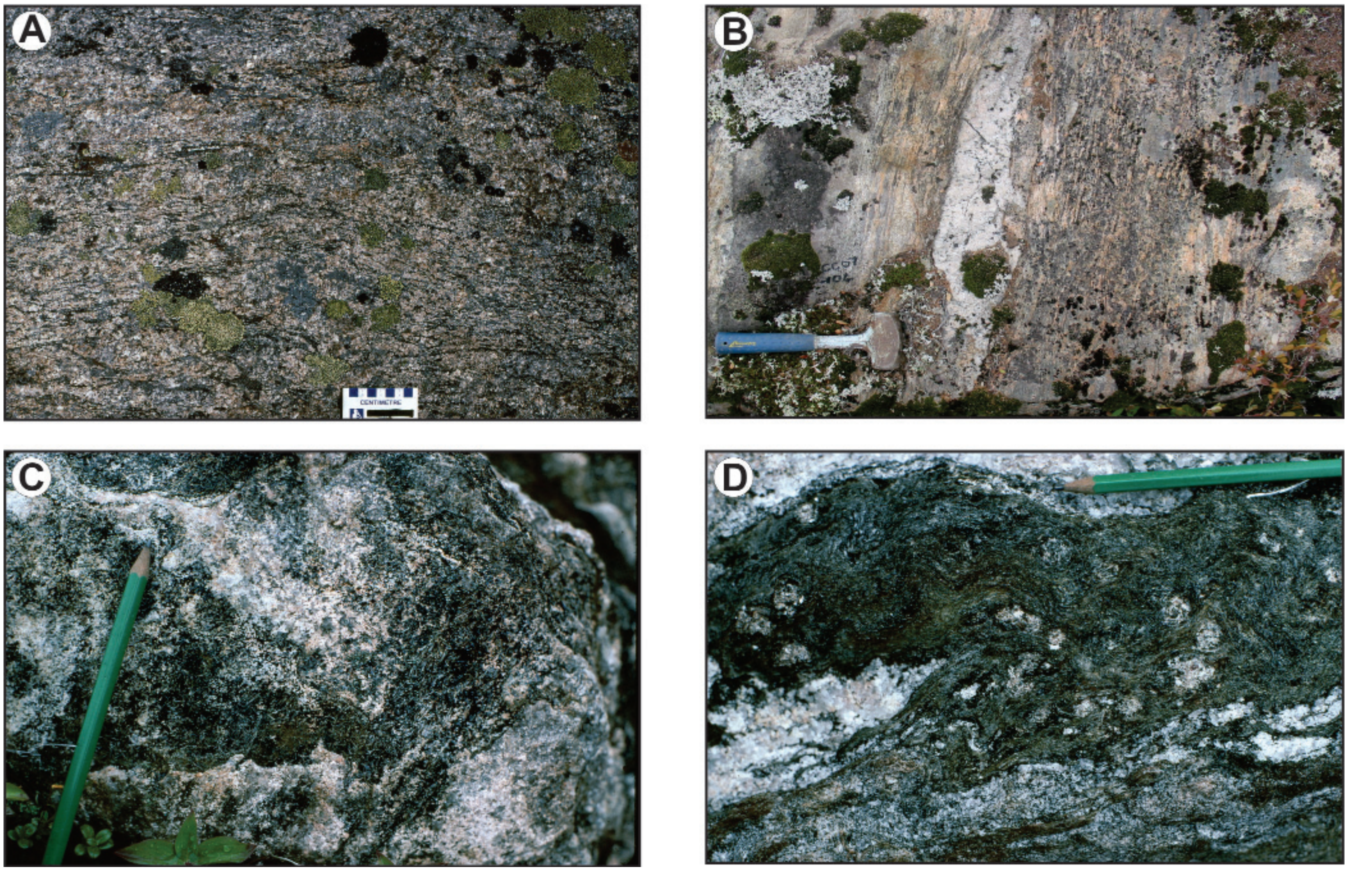


Plate 7.11. *Pelitic gneiss in Mealy Mountains terrane. A. Fairly homogeneous cordierite–orthopyroxene–spinel ± garnet pelitic 'gneiss'; East Mealy metasedimentary gneiss (VN95-138), B. Cordierite-bearing pelitic gneiss. Muskrat Lake metasedimentary gneiss (CG07-104), C. Pelitic gneiss showing very coarse sillimanite needles. Southwest Pond metasedimentary gneiss (CG86-327), D. White retrograde garnet (to plagioclase and biotite) in melanosome of pelitic gneiss (CG86-329). Compare/contrast with that in Plate 7.9C. Southwest Pond metasedimentary gneiss.*

Pelitic Group 4 (orthopyroxene; lacking cordierite). No clear field distinction is evident between the cordierite-absent rocks (this group) and the cordierite-present rocks (previous group); although tendencies exist in this group toward less biotite and quartz, finer grain size and perhaps more compositional variation. The cordierite-absent rocks are confined to a zone within 5 km of the Mealy Mountains intrusive suite.

In addition to lacking cordierite, the rocks also lack any Al_2SiO_5 polymorph. A large group of samples was examined in thin section (CG81-658, CG81-660, CG81-661, CG81-662, CG95-160, CG95-161A, CG95-161B, CG95-161C, CG95-163B, CG95-172A, CG95-210, CG95-212, CG95-303, CG95-324, CG95-337, EA61-002, EA61-339, GF81-159, GF81-160B, GF81-161A.1, GF81-161A.2, GF81-161B.1, GF81-162A, VN95-015, VN95-130A, VN95-134).

Orthopyroxene (hypersthene) forms anhedral, equant, somewhat poikiloblastic grains, commonly partially retrogressed to bastite. The prevailing characteristic of the hypersthene in this zone is its intense pleochroism, only matched in immediately adjacent rocks to the east. Other minerals forming part of the stable metamorphic assemblage are plagioclase, perthitic K-feldspar, quartz, an opaque mineral and ± biotite. Plagioclase is anhedral, poor to well twinned, lightly sericitized and commonly antiperthitic. K-feldspar is very finely perthitic, but appears almost homogeneous at low magnifications. Quartz occurs as granoblastic polygonal to interlobate grains and is present in all samples except CG95-210.

Biotite occurs in several forms, much of it occurring as red-brown to locally orange-brown, aligned flakes that are part of the high-grade assemblage, but retrograde biotite, of various colours and containing vermiform quartz inclusions, is a breakdown product of hypersthene. Thin veneers of secondary biotite marginal to opaque minerals are also seen locally. Biotite is absent in CG95-160, which vies with CG95-285A as being the site in the metasedimentary gneiss sampled closest to the Mealy Mountains intrusive suite contact. Hercynite is also present in a few samples but it is not as abundant or as ubiquitous as in the orthopyroxene-present/cordierite-present rocks. In sample CG95-324, a biotite + quartz symplectite separates garnet from orthopyroxene, suggesting the retrograde reaction:

Garnet + Orthopyroxene + K-feldspar +
H_2O = Biotite + Quartz

A similar texture is seen in EA61-002, and, to a very minor extent, in VN95-015.

Garnet occurs in CG95-161A, CG95-172B, CG95-324, EA61-002, EA61-339, VN95-109 and VN95-116, and tends to form spongy, anhedral grains engorged with quartz inclusions and surrounded by a quartz-rich envelope. Thin section CG95-161A has symplectic garnet and quartz surrounded by plagioclase, K-feldspar, quartz and biotite. Garnet–quartz symplectite, up to 3 cm across, was also recorded in the field at CG95-110. It is not part of the stable metamorphic assemblage, and may be a decompression product. In VN95-116, garnet forms coronal aggregates around both orthopyroxene and an opaque oxide. The only other noteworthy point is an elliptical inclusion of cordierite within garnet at VN95-109.

The presence of orthopyroxene, hercynite and apatite in VN95-015 is anomalous for the region where it is found. The rock was described in field notes as part of a leucosome within sillimanite + garnet + biotite + muscovite pelitic gneisses.

Generalized interpretation of pelitic gneiss. The sequence of prograde mineralogical changes from east to west toward the Mealy Mountains intrusive suite (MMIS) was reported in preliminary form by Gower and van Nostrand (1996). The present analysis is based on more petrographic data than was available at that time, but the same general result applies. Gower and van Nostrand (1996) concluded that there was a genetic link between prograde metamorphism and the MMIS, but that the linkage is not necessarily direct, as there does not appear to be a close relationship between the distribution of mineral assemblages and the eastern boundary of the MMIS. The east-to-west mineralogical changes in the pelitic gneiss reflect progressively higher temperatures and increasingly anhydrous conditions. To summarize, the main (east to west) changes are: i) incoming of cordierite, then orthopyroxene and hercynitic spinel, ii) the elimination of sillimanite, except that due to retrograde alteration of cordierite, and iii) depletion of garnet and biotite, although one or the other of the latter persists up to the contact with the MMIS. Reactions that might be responsible for these changes were given earlier. They only provide partial explanation and do not take into account all phases, or the reality of a more complex multi-component system than the mentioned reactions imply. Also, the issue of melting and what role it might have played have been largely ignored.

Diatexite. Diatexite in the East Mealy metasedimentary gneiss is only depicted as an extensive mappable rock type in the northeast tongue (Figure 7.1). Its distribution here must be regarded as speculative as outcrop is sparse. Mostly diatexite is integral to, but a subsidiary part of, the pelitic gneiss unit representing a once-melted component of it. It is typically white-weathering, and has a nebulitic, inhomogeneous appearance. Rafts of pelitic gneiss are common within it.

Two thin sections were examined from the EMMG that were specifically termed diatexite. CG95-172B is intimately associated with Group 4 pelitic gneiss, but lacks orthopyroxene. Instead, it comprises about 98% felsic minerals plus an opaque oxide with narrow fringes of garnet, biotite or white mica. The other thin section, RG80-432, comes from the mappable unit in the northeast part of the belt and carries substantial orange-brown biotite and garnet. It could equally well be classified as Group 1 pelitic gneiss.

Psammitic gneiss. Most psammitic gneiss in the East Mealy metasedimentary gneiss occurs in the northeast tongue, with additional minor occurrences in the southeast tongue (Figure 7.1). The rocks in the northeast tongue were mapped during the early stages of the project when rock identification was most tentative and it was unclear to mappers whether the rocks were psammitic gneiss or granitoid orthogneiss, so some interpretational hindsight has been used (*e.g.*, data stations VO81-504 to VO81-517, at which the rocks were merely termed quartzofeldspathic gneiss). The rocks are creamy or pale-grey, discontinuously banded, and schlieric textured. Mylonitic psammitic gneiss occurs in the southeast tongue (CG84-228). Garnet is prevalent throughout the gneiss, forming garnet-rich melanosome layers, in which garnet may constitute up to 40% of the rock (*e.g.*, VN95-012, CG95-015). Garnet may also be dispersed throughout the leucosome. Some rocks were described in the field as unusually quartz rich (VO81-520, VO81-521).

Four samples were examined petrographically (CG81-523, CG91-047, CG95-015, RG80-534). They contain quartz, well-twinned plagioclase, untwinned K-feldspar, olive-green to red-brown biotite, garnet (not in CG81-523), an opaque oxide, allanite, zircon and retrograde chlorite (after garnet and biotite). The most interesting petrographic feature is the breakdown/formation of garnet and K-feldspar to/from plagioclase, biotite and quartz, seen in CG91-047 and CG95-015 (Photomicrograph 7.5C), both of which are from the southeast tongue of the EMMG. A reaction that might apply is:

$$\textit{Garnet} + \textit{K-feldspar} + Al_2O_3 + H_2O$$
$$= \textit{Biotite} + \textit{Plagioclase} + \textit{Quartz}$$

Quartzite. Quartzite is rare in the EMMG and generally present only as thin layers within pelitic or psammitic gneiss. Some quartz-rich rocks are also associated with sulphide occurrences and may have been derived from metachert. The only notable occurrence of a quartz-rich rock interpreted to be of clastic origin, is at data station CG81-553, in the northwest tongue of the belt. Here, the rock was described in the field as being well-bedded with layers from 2 mm to several centimetres thick, and termed, from a thin section, muscovite-bearing arkosic quartzite. This occurrence is close to the English River metasedimentary gneiss, which is dominated by quartzite.

Calc-silicate rocks? Only two occurrences of possible calc-silicate rock have been recorded from the EMMG, both in the southeast tongue. At CG84-227, a massive layer of grey-green rock about 10 m wide was seen, and, at VN95-016, pods of black to dark-green rock were recorded.

Thin sectioned sample CG84-227 has an entirely metamorphic mineral assemblage consisting of plagioclase, pale-green amphibole, pale-green clinopyroxene, weakly pleochroic orthopyroxene, and minor opaque oxide and sulphide. A thin section from VN95-016 contains minor undulose quartz, minor well-twinned, anhedral plagioclase, colourless high-relief clinopyroxene, pale-orange biotite, pale-green amphibole, and a sulphide opaque mineral (pyrrhotite?). In both cases, an alternative interpretation is that both instances were derived from ultramafic intrusive protoliths, noting that there are several other examples of ultramafic rocks in the southeast tongue (*cf.* Section 9.3.2).

One other rock requiring mention is CG81-586A, which is located in the northern part of the EMMG. It was described in the field as having an evenly and continuously banded appearance.

In thin section, it is seen to consist of plagioclase, clinopyroxene and blue-green amphibole. The author's preferred interpretation is that it has an anorthositic/leucogabbronoritic protolith, but a calc-silicate supracrustal origin remains a possibility.

Granitoid rocks. Granite, granodiorite, monzonite, monzodiorite and their gneiss equivalents were recorded in field notes throughout the EMMG. Many have been demonstrated from petrographic evidence to have a pelitic/diatexitic protolith, and have been reassigned accordingly. This section, therefore, addresses the remainder, for which protolith remains equivocal. The rocks are mostly grey to creamy (some rusty), fine- to coarse-grained gneisses displaying various fabrics from well banded or obviously mylonitic, through discontinuously banded, to schlieric or nebulitic. Biotite-, amphibole-, or garnet-rich mafic schlieren are characteristic. Garnet is also found in the leucosome, where it is locally large and poikiloblastic. A few have a K-feldspar seriate to megacrystic aspect. Concordant and discordant pegmatite and microgranite are seen sporadically, but are not common. One rock that may have been erroneously included in this group is a massive hornblende quartz syenite at data station CG81-653.

Thin sections examined are CG80-817, CG81-568, CG81-577, CG81-656, GF81-149, GF81-161B.2, GF81-162B, GF81-302, GF81-315 and VO81-504. With the exception of GF81-149 to GF81-162B (from adjacent to the MMIS), all are from the northern part of the EMMG. They comprise plagioclase, K-feldspar (microcline and perthite), quartz, orange-brown biotite, and zircon, plus sporadic posttectonic muscovite (CG81-568), hornblende (CG80-817, GF81-149, GF81-315), orthopyroxene (GF81-149), garnet (CG81-656, GF81-302, GF81-315, VO81-504), and opaque oxide and sulphide, apatite, and secondary minerals. Tenuous evidence in favour of an igneous origin includes prevalence of myrmekite (not seen in the pelitic gneiss), relict 'igneous-looking' textures, and clustering of mafic and accessory minerals.

Mafic rocks. Mafic rocks associated with the East Mealy metasedimentary gneiss are either amphibolite or mafic granulite. These rocks are not a major component of the EMMG, being found most commonly as pods and layers within pelitic and psammitic gneiss. They are interpreted to be remnants of fairly small mafic intrusions. A few examples are discordant mafic dykes (CG95-096, CG95-113, VN95-014).

Petrographic data from a small, but diverse group of samples serve to demonstrate some of the group's variability. Sample CG84-224A, from the southeast tongue, is the lowest grade, in keeping with it being the farthest from the MMIS. It consists of plagioclase, amphibole (blue-green hornblende and actinolite), biotite, an opaque oxide and apatite. Two mafic dyke samples (CG95-096B, VN95-014B) are both orthopyroxene-bearing amphibolite/mela-amphibolite.

Mafic dyke CG95-096B gave a near-concordant weighted average $^{207}Pb/^{206}Pb$ age of 1640 ± 2 Ma, interpreted to date time of metamorphism (Gower *et al.*, 2008b).

The most anomalous rock is CG95-285A, which occurs at the boundary between the MMIS and the EMMG and is intermediate rather than mafic. Strongly pleochroic orthopyroxene allies it with the Group 4 pelitic gneiss, but it also contains minor clinopyroxene, which is atypical of the pelitic gneiss but, conversely, links it to the monzonitic rocks of the MMIS. The rock is the closest seen to being transitional between the two units.

Enderbitic granulite. In a few places within the metasedimentary gneisses, leucocratic, quartz-, plagioclase- and pyroxene-bearing rocks were recorded, for which the name enderbitic granulite was used by Gower and van Nostrand (1996). They depicted three small bodies, aligned along a northerly trend, parallel to the regional structural grain of the area. Integrating this mapping with information for other localities extracted from Eade's field notes and petrographic thin sections, led Gower (2010a; Southeast Mealy Mountains map region) to re-interpret the occurrences as possibly originally belonging to a single tabular unit, now preserved as boudinaged remnants. The southern body is best exposed, consisting of fairly homogeneous melanocratic mafic rocks in the east, grading into more heterogeneous leucocratic variants farther west (CG95-097). Metamorphosed mafic dykes, some of which are clearly discordant to the gneissosity in the host rocks, are also present in the vicinity. Stained slabs of the rocks show both grain-size and textural variation. The finer grained examples have a granoblastic texture and are the most homogeneous, but one sample (CG95-117), from the central body, has, what appears to be, a medium- to coarse-grained, relict-ophitic texture. The author is still at somewhat of a loss regarding the protolith of this unit, but favours the notion that it might represent remnants of a metamorphosed, layered leucogabbronorite intrusion.

A sample of enderbitic granulite gneiss (CG95-096A; *see* Figures 7.1 and 8.1) yielded an age of 1789 ± 29 Ma based on three almost-concordant zircon analyses, when anchored by a lower intercept age of 1640 ± 2 Ma from metamorphic zircon in a crosscutting mafic dyke. The Nd–Sm isotopic data for the same sample are T_{DM} = 2195 Ma and εNd (1.79 Ga) = –0.08.

Eight thin sections of the unit are available: from the south, CG95-096A, CG95-099 and CG95-102; from the centre, CG95-116 and CG97-117; and from the north, EA61-327 and EA61-327A. Another outcrop, 6 km northwest of EA61-327, may have affinities with this group of rocks (CG95-108). Minerals present include minor granoblastic, polygonal quartz (absent from CG95-108, CG95-117), moderate- to well-twinned plagioclase, red-brown or orange-brown to olive-green biotite, weakly to moderately pleochroic orthopyroxene (pseudomorphed by biotite + amphibole + quartz + opaque minerals in CG95-116), an opaque oxide, apatite (not seen in CG95-116), and zircon (in CG95-116, EA95-327A). K-feldspar is present in CG95-116, clinopyroxene in CG95-117, garnet in EA61-327A (and in outcrop at CG95-116), and amphibole is part of the stable assemblage in EA61-327 and forms a retrograde mineral in CG95-116. Sample CG95-116, containing K-feldspar, abundant hydrous mafic silicate minerals and only orthopyroxene pseudomorphs, is probably the most (metasomatically) modified.

The broad list of minerals, with its additions and exceptions, serves to emphasize the variability of these rocks – perhaps even to the point of questioning whether they should be grouped. Sample CG95-117, consisting of quartz–plagioclase–orthopyroxene–clinopyroxene–apatite–opaque minerals and having a relict ophitic texture, is probably closest to having an igneous protolith appearance. Sample CG95-116, containing K-feldspar, abundant hydrous mafic silicate minerals and only orthopyroxene pseudomorphs, is probably the most (metasomatically) modified. Zoning in plagioclase in CG95-096A and EA61-327A also supports the suggestion that these rocks are derived from an igneous protolith.

7.3.6.2 Muskrat Lake Metasedimentary Gneiss Slivers (P_{3A}sp)

The Muskrat Lake metasedimentary slivers occur as two, north-trending, probably thrust-bound wedges within the Mealy Mountains intrusive suite (Figure 7.1). Their recognition was an awkward process. The Muskrat Lake area was first mapped by Emslie (1975 field notes). He refers to 40% paragneiss inclusions within pyroxene quartz monzonite at EC75-014, streaky diorite–monzonite to quartz diorite at EC75-037, and 'garnet present in streaks and blobs of contaminated quartz monzonite' at EC83-217. His assistant (G. Dunning) mentions biotite metasediment inclusions and 'blue quartz stringers' at ECD75-030, and blue quartz at ECD75-014 (blue quartz being misidentified cordierite? – author). All sites are within the presently defined area of metasedimentary gneiss. No formal geological map was ever produced from this project and there is no indication that any coherent metasedimentary gneiss unit was suspected. The area was mapped at 1:100 000 scale by Nunn and van Nostrand (1996b). A traverse by van Nostrand identified rusty- to grey-brown-weathering, medium-grained, garnet–cordierite–biotite-bearing pelitic gneiss associated with pink- to white-weathering diatexite in two separate segments of his traverse (TN95-084 to -086 and TN95-092 to -094). Follow-up petrographic work was never carried out and, unfortunately, samples from the project were discarded, so no confirmation of van Nostrand's mapping was possible without revisiting the area. The author did this in 2007. Very strongly deformed to mylonitic cordierite-garnet pelitic gneiss was found over an 11-km-long strike distance (CG07-086, CG07-101 to -109; Plate 7.11B). Cordierite occurs as recrystallized masses concentrated into elongate, irregular dark-purple lenses and schlieren. The data now form the basis for interpreting the western sliver, also including van Nostrand's original observations. Interpretation of the eastern sliver relies largely on Emslie's and van Nostrand's field notes, plus two inconclusive data stations of the author (CG07-113, CG07-114). Currently, the delineation of metasedimentary gneiss in the Muskrat Lake area must be regarded as only approximate.

That the rocks are cordierite-bearing pelitic gneiss is confirmed by three thin sections (CG07-086B, CG07-104B, CG07-109), all of which contain plagioclase, perthitic K-feldspar, quartz, an opaque oxide, and cordierite. The cordierite is thoroughly polygonized and closely associated with an opaque oxide. It is partially pseudomorphed by a feathery felsic myrmekite-like symplectite in CG07-086B, which is reminiscent of alteration of osumilite. Orange-brown biotite is present in two thin sections, but absent from CG07-086B, which contains intensely pleochroic orthopyroxene instead. Garnet is present in CG07-109. Monazite and dark green spinel are accessory minerals. No Al_2SiO_5 polymorph is present. These rocks have comparable mineral assemblages to those in the western part of the East Mealy metasedimentary gneiss, with which they must surely be genetically related.

Three other samples were examined in thin section. All are mafic to intermediate granulite-facies rocks. One is from an outcrop 1.4 km west of the western sliver and is a two-pyroxene mafic granulite (CG07-085). The other two are from within the western sliver and were termed enderbite–pyroxene quartz diorite (CG07-101) and two-pyroxene mafic granulite (CG07-107). Both the pelitic gneiss and the intermediate to mafic granulites appear to have experienced similar high-grade metamorphism, assumed to have occurred during a single thermal event.

7.3.6.3 Southwest Pond Metasedimentary Gneiss (P_{3A}sp, P_{3A}ss)

The Southwest Pond metasedimentary gneiss is a wedge-shaped area of mostly pelitic gneiss, about 16 km wide in the west and tapering out to the east over a strike length of about 50 km (Figure 7.1). The broad western end of the wedge interfingers with K-feldspar megacrystic and non-megacrystic granitoid rocks. The northern and southern boundaries of the belt are most likely strike-slip faults. An ovoid body of K-feldspar megacrystic rocks is present in the central-eastern part of the belt, although its border with the enveloping pelitic gneiss is probably not as simple as depicted on Gower's (2010a) Port Hope Simpson map region. The belt is also intruded by the Southwest Pond late- to post-Grenvillian (962 ± 3 Ma) granite pluton, the simple ovoid shape of which is confidently represented. The belt was mapped during three separate 1:100 000-scale projects (Gower *et al.*, 1987, 1988; van Nostrand, 1992). The name 'Southwest Pond metasedimentary gneiss' is informally introduced here.

The gneiss is white-, pink-, grey-, buff- or brown-weathering, fine to coarse grained and migmatitic, being characterized by schlieric and nebulitic fabrics. It varies from being relatively homogeneous in semi-pelitic compositions, to well-banded where entirely pelitic. The pelitic rocks, in particular, have well-defined, white or pink, quartzofeldspathic neosome and grey or black garnet-sillimanite-biotite restite. Both the sillimanite and garnet may be quite large – sillimanite reaching 1.5 cm long (*e.g.*, CG86-337; Plate 7.11C) and garnet up to 4 cm in diameter (*e.g.*, VN91-329). Muscovite is common throughout the belt. Some white-weathering, irregularly shaped patches of diatexite are associated. In places, the rocks have a porphyroclastic aspect and are spatially associated with K-feldspar megacrystic granitoid rocks. Associated supracrustal rocks include psammitic gneiss (fairly common), quartzite and calc-silicate rocks (both rare). Some sulphide-rich (mostly pyrite) rocks are intercalated with the pelitic gneiss. Other rock types include pods of amphibolite or gabbroic rocks, which are the boudinaged remnants of mafic dykes or larger mafic intrusions. Also present are concordant quartzofeldspathic veins, discordant microgranite and pegmatite dykes (the latter commonly muscovite bearing) and rare quartz veins.

Forty-five thin sections of pelitic or psammitic gneiss are available (CG03-341C, CG86-332, CG86-338, CG86-343, CG86-622, CG86-624, CG87-270, CG87-354, DD91-005A, DD91-005B, DD91-006, DD91-097, DD91-101B, DD91-101C, DD91-153, DE91-038A, DE91-038B, DE91-062, DE91-120, DE91-130A, EA61-376, EA61-379, JS86-302, JS86-305A, JS87-001, JS87-012, MN86-289, SN86-269, SN86-273, SN86-283, VN87-006, VN87-041, VN91-030, VN91-223B, VN91-265A, VN91-266A, VN91-283, VN91-293A, VN91-296, VN91-301, VN91-323B, VN91-336A, VN91-370A, VN91-370B, VN91-388). All contain plagioclase, quartz and biotite (olive green to orange-brown); most contain K-feldspar, an opaque oxide and zircon. Two-thirds have garnet, half have sillimanite, about one-third have muscovite, apatite (unusually abundant in CG86-343), and/or monazite, and secondary chlorite. Relict cordierite was doubtfully identified in two thin sections as a possible relict phase (SN86-273, VN91-323B), but certainly needs verification.

Garnet commonly shows evidence of breakdown to biotite, plagioclase and quartz (Plate 7.11D), with muscovite as a sporadic product (CG86-338, CG86-343, DE91-130A, SN86-283, VN87-041, VN91-223B, VN91-265A, VN91-266A, VN91-323B). The reaction may be written as:

$$\textit{Garnet} + \textit{K-feldspar} + Al_2O_3 + H_2O$$
$$= \textit{Biotite} + \textit{Plagioclase} + \textit{Quartz}$$

Garnet is also replaced by sillimanite + biotite + quartz (DE91-062, DE91-120, CG86-622, CG87-354, VN91-301, VN91-370A). Typically, embayed relicts of garnet are surrounded by symplectic intergrowths of green biotite, sillimanite and quartz ± muscovite ± plagioclase, as previously noted by van Nostrand (1992). A similar reaction may apply, namely:

$$\textit{Garnet} + \textit{K-feldspar} + \textit{Melt} = \textit{Biotite}$$
$$+ \textit{Sillimanite} + \textit{Quartz} + \textit{Albite}$$

Muscovite is clearly a late-stage phase in some samples, being either a breakdown product of sillimanite (*e.g.*, CG86-343), or forming large, posttectonic poikiloblastic flakes (CG86-624, DD91-153).

Two mafic rocks associated with the Southwest Pond metasedimentary gneiss were also examined in thin section. One is a granoblastic garnet amphibolite, retaining textural hints of a former gabbroic protolith (CG86-333). The other is a garnet-orthopyroxene-hornblende granoblastic mafic granulite (JS87-003). Beyond inferring that both are mafic plutonic rocks, little is known about them.