CHAPTER 18

NEOPROTEROZOIC

Neoproterozoic rocks in eastern Labrador include: i) Double Mer Formation and probably correlative rocks, including clastic dykes, and the lower Labrador Group, and ii) the Long Range dykes and tectonically related quartz and carbonate veins. Images of representative stained slabs of the Double Mer Formation and clastic dykes are displayed in Appendix 2, Slab images 18.1; of the lower Labrador Group in Appendix 2, Slab images 18.2; of a Long Range dyke (White Bear River estuary dyke) in Appendix 2, Slab images 18.3; and of quartz veins and related fault breccia in Appendix 2, Slab images 18.4.

18.1 SUPRACRUSTAL ROCKS

18.1.1 DOUBLE MER FORMATION (N*Dm*)

18.1.1.1 Historical Background

Outcrops of Double Mer Formation on the north shore of Lake Melville were first reported by Low (1896). The strata were subsequently named Double Mer sandstone by Kindle (1924), who outlined their distribution north of Lake Melville and north of Double Mer. Kindle selected a 45 m thickness of sandstone, 13 km east of the head of Double Mer, as the type section, preferring this locality to a less well-exposed section at Mulligan Bay, on the north shore of Lake Melville, previously described by Low. Kindle (1924, page 56) described the rock as a 'firmly-cemented, dull red, arkosic sandstone of coarse texture throughout, with numerous small pebbles in some layers'. The distribution of parts of the Double Mer sandstone is shown on maps of Kranck (1953), Christie *et al.* (1953), Eade (1962) and Stevenson (1970).

The modification of the name to Double Mer Formation was made by Stevenson (1967a, b), who correlated similar strata first mapped by him in the Churchill and Kenamu river valleys with already known occurrences north of Lake Melville and north of Double Mer. Systematic 1:100 000-scale mapping in eastern Labrador has led to refinements in the interpretation of areas underlain by Double Mer Formation (Bailey, 1980; Gower *et al.*, 1981, 1982b; Erdmer 1983, 1984; Gower 1984; Gower *et al.*, 1986a), and to inferences concerning areas covered by surficial deposits but most likely underlain by Double Mer Formation around the western end of Lake Melville (Wardle and Ash, 1986; Wardle *et al.*, 2000b).

The only areas in which substantial thicknesses of Double Mer Formation are exposed are in the Double Mer half graben and in the northern part of the Lake Melville graben. Much of the information presented below is summarized from Gower *et al.* (1986a).

18.1.1.2 Double Mer Half Graben

In the Double Mer half graben (Figure 18.1), the Double Mer Formation comprises subarkosic and arkosic sandstone, conglomerate, siltstone and shale.

The subarkosic and arkosic sandstone is maroon, red or brown, and shows medium-scale cross-stratification in sets ranging up to 50 cm in thickness (Plate 18.1A). Crossbedding is overwhelmingly planar and 10–40 cm thick. Trough crossbedding (10–50 cm) predominates at locality CG80-689. Scour surfaces are common, and beds of planar lamination and thin stratification up to 30 cm also occur.

Conglomerate is most abundant near the eastern end of the half graben (Plate 18.1B), but is also found intermittently adjacent to the Double Mer fault that defines its northern side (Stevenson, 1970). The conglomerate is clast-supported, and has a sandy to pebbly matrix made up mostly of quartz and feldspar. The rock clasts are rounded to angular, quartzo-feldspathic and mafic, intrusive and metamorphic. Minor vein quartz is also found. The source was a high-grade metamorphic, granitic, and gabbroic plutonic region, such as is present in the adjacent Grenville Province. The conglomerate was interpreted by Gower *et al.* (1986a) as fanglomerate deposited adjacent to the northern margin of the basin, with immature arkosic sandstone in more distal areas to the southwest.

The interbedded siltstone and shale are red-brown or maroon, and mostly found in the central and western parts of the half graben, although a minor component elsewhere. Bedding attitudes suggest broad, open folding of the strata. Aeromagnetic data indicates a basement high at the west end of Double Mer. This coincides with a structural culmination, but whether the cause is post-depositional folding, or due to original erosional irregularities, is unknown.

Arkose from the Double Mer half graben has been well examined in thin section (CG80-689, CG83-208, CG83-210, CG80-211, CG80-219, CG80-221, CG80-223, SG68-230). Minerals present are cloud-

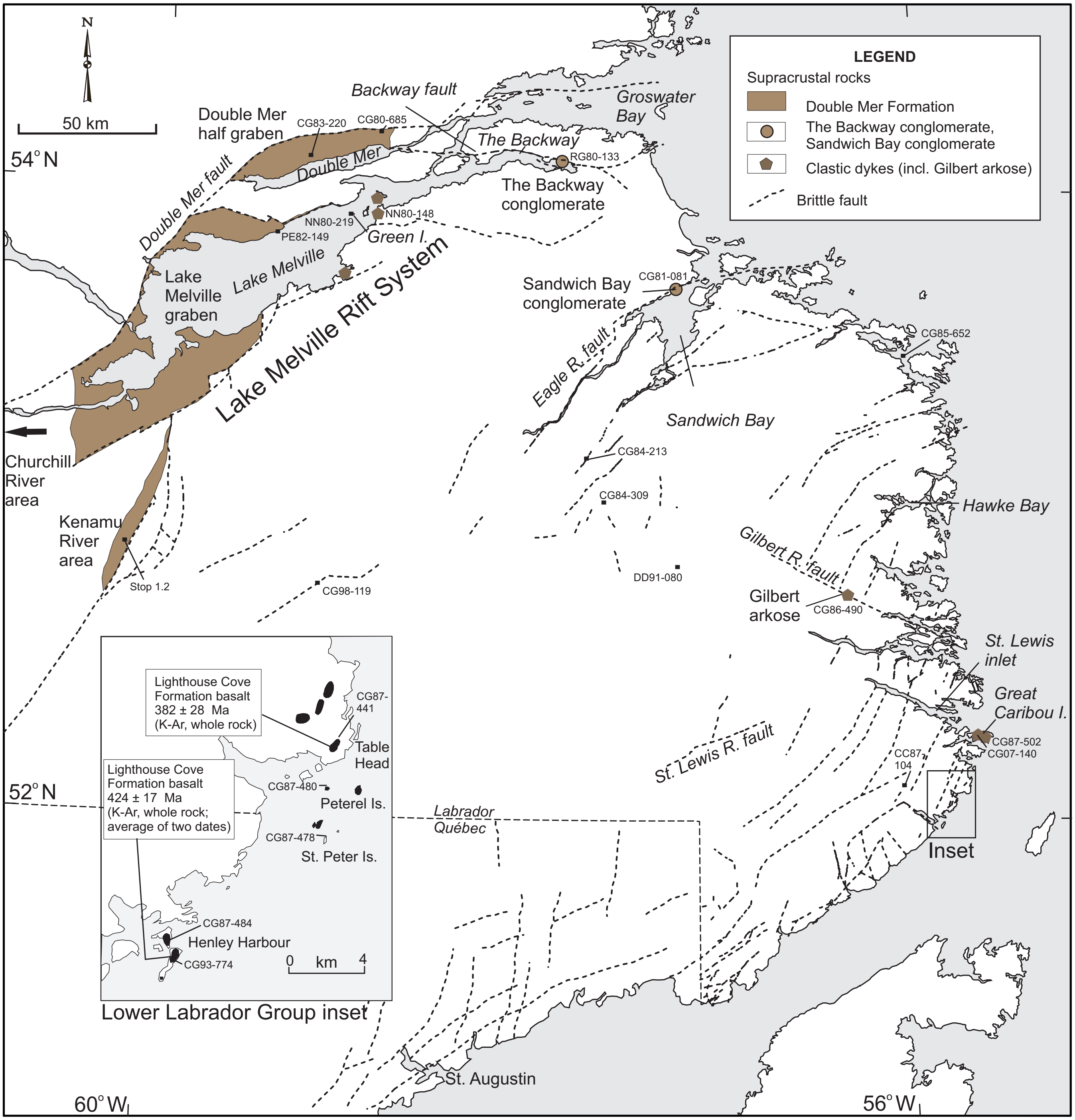


Figure 18.1. *Double Mer Formation, inferred correlative rocks including clastic dykes, and related faults. Inset shows distribution of lower Labrador Group (Bateau and Lighthouse Cove formations).*

ed plagioclase, microcline, perthitic K-feldspar, quartz, muscovite, chloritic patches (after mafic silicates), opaque minerals, biotite, apatite, garnet, and zircon. The opaque minerals, garnet and zircon define heavy mineral laminations, but opaque minerals are also disseminated. Rock fragments, mostly composite quartz-feldspar grains derived from metamorphic–plutonic rocks, are sporadically present. The arkose is mostly grain-supported, having a matrix of quartz and feldspar bound by pervasive hematite cement. The clasts are moderately supported and the best-sorted rocks are found interleaved with shale at the west end of the Double Mer half graben. Modal analyses for all of the arkose examined in thin section were reported by Gower *et al.* (1986a), and are summarized as follows: quartz 21.1–43.8%; K-feldspar 23.3–32.0%; plagioclase 16.5–40.3%; opaque minerals 1.6–10.0%; white mica 0.0–1.1%; chlorite 0.0–7.6%; other minerals (garnet, zircon and apatite) 0.0–0.6%; rock fragments 0.0–2.1%; matrix 3.3–18.4%.

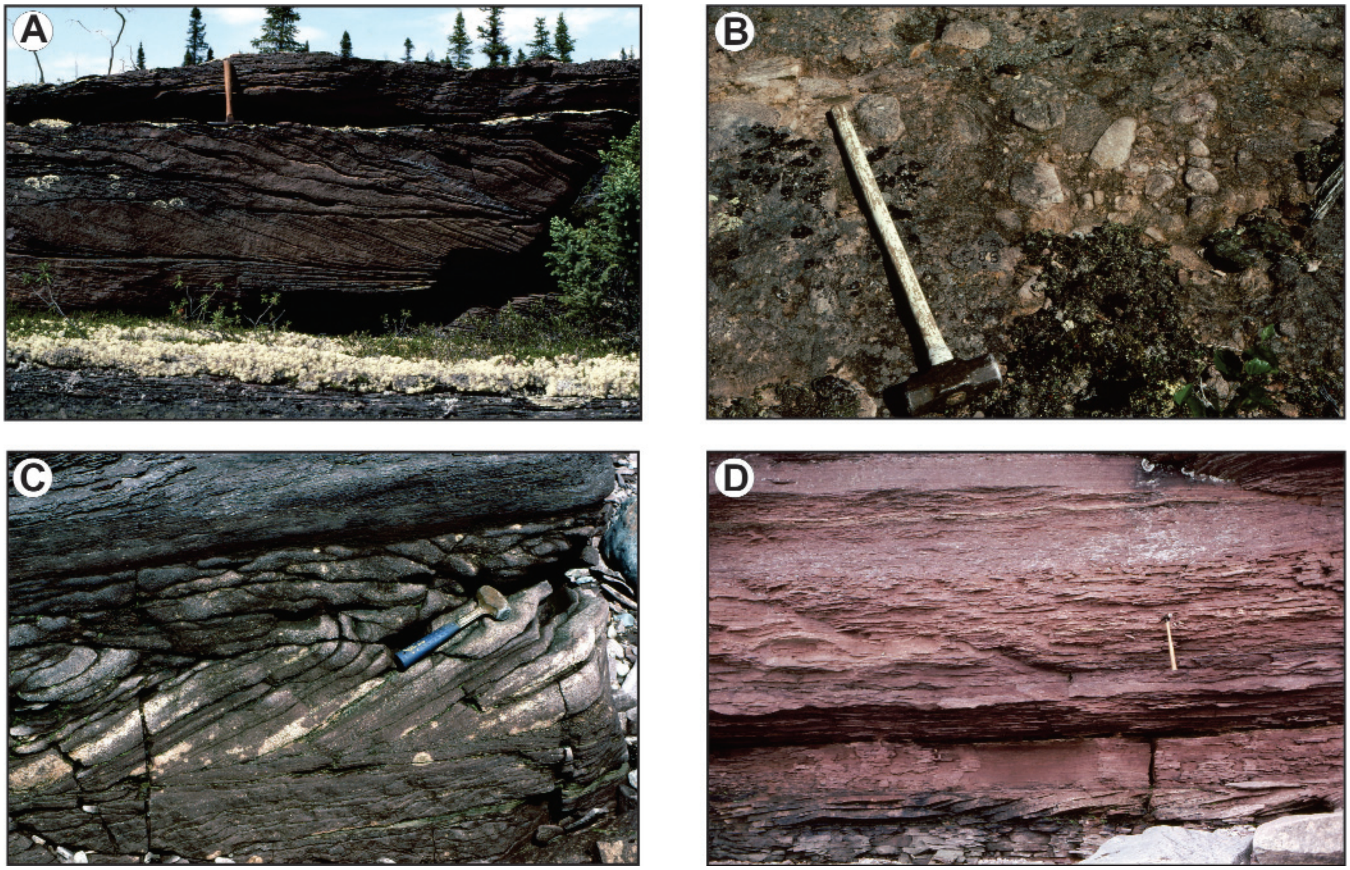


Plate 18.1. *Double Mer Formation in the Double Mer half graben and the Lake Melville graben. A. Cross-stratification in arenite in Double Mer Formation, Double Mer half graben (CG83-220), B. Conglomerate in Double Mer Formation, Double Mer half graben (CG80-685), C. Cross-stratification in Double Mer Formation, Lake Melville graben (PE82-149), D. Siltstone and arkose in the Double Mer Formation, Lake Melville graben (NN80-219).*

18.1.1.3 Lake Melville Graben

The Double Mer Formation in the Lake Melville graben (Figure 18.1) comprises dull-red to reddish-brown, clast-supported, polymictic conglomerate, subarkosic and arkosic sandstone, pebbly sandstone having a hematitic, arkosic, or silt-rich matrix, and minor siltstone and shale (Plate 18.1C). Crossbedding in the sandstone suggests a north-flowing depositional current. Bedding attitudes outline a gently west-dipping homoclinal sequence, except in the most easterly outcrops, where dips are north to northeast (either because of open folding or the effects of post-depositional faulting). Bedding and crossbed attitudes are regular over a large area hence the dip is interpreted to be structural rather than depositional (Erdmer, 1984). Excluding the possibility of repetition due to faulting, this implies a total depositional thickness exceeding 5 km, which is roughly consistent with the 3–4 km thickness interpreted by Grant (1975) for strata of the Double Mer Formation inferred to underlie parts of Lake Melville. The easternmost end of the Lake Melville graben is constrained by an outcrop of Double Mer Formation on Green Island (Eade, 1962; Gower *et al.*, 1986a) (Figure 18.1; Plate 18.1D).

Two samples of arkosic rocks from the Lake Melville graben were examined in thin section (PE82-122, PE82-130). They contain the same mineral assemblages as those samples from the Double Mer half graben. Modal analyses for both were reported by Gower *et al.* (1986a), and are summarized as follows: quartz 35.1–46.6%; K-feldspar 23.1–26.8%; plagioclase 16.8–18.6%; opaque minerals 2.1–2.3%; white mica 0.1%; chlorite 1.6–2.1%; rock fragments 1.0–1.1%; matrix 12.8–15.8%.

18.1.1.4 The Lake Melville Rift System

The term Lake Melville rift system was coined by Gower *et al.* (1986a) to embrace the Lake Melville graben, the Double Mer half graben, and extrapolations of the rift both to the east, (to include The Backway conglomerate), and to the southwest (to include correlative strata in the Churchill and Kenamu river valleys). The fault-bounded basin configuration had, however, been recognized much earlier (Kindle, 1924; Kranck, 1947, 1953). Kranck (1947) termed it the Lake Melville depression.

The Double Mer fault defines the northern paleo-edge of the rift system (Stevenson, 1970). This fault can be extended eastward into the centre of Groswater Bay and southwestward to the north side of the lower Churchill River valley, a total distance of about 270 km. The fault is marked by a broad zone of brecciated, sheared, and mylonitized rocks, and low-grade alteration, up to 1 km wide. Gower *et al.* (1986a) interpreted the mylonite to be earlier (Grenvillian), as it occurs as blocks of breccia pervasively altered to greenschist-facies assemblages and enveloped in low-grade shear zones. The somewhat arcuate form of the faults (convex to the north) may well have been influenced by the location of earlier ductile, north-northwest-verging thrust zones. Stevenson (1970) recorded that conglomerate close to the fault contact has been invaded by quartz veins and has associated hematite-coated fracture planes. Quartz veins within Double Mer Formation conglomerate have also been noted by the author (*e.g.*, CG83-152). Their presence is taken to indicate that faulting and deposition of conglomerate were broadly synchronous.

Between the Mealy Mountains and Lake Melville, the southern side of the rift system is defined by an escarpment that represents a series of linked faults. That such faults also underlie Lake Melville has been interpreted by Grant (1975) from bathymetric and seismic data. A modest modification to the southern boundary of the Lake Melville rift system was made by Gower and Erdmer (2010; data station CG09-035), resulting in the depiction of a 28-km^2, rhomb-shaped, fault-bounded block of Mealy Mountains intrusive suite in an area formerly shown to be underlain by Double Mer Formation.

Adopting these bounding faults implies the rift basin to be 40–50 km wide, which is comparable with other rift valleys of the world. Central horsts are also typical of rift systems, and such is also the case for the Lake Melville rift system, separating the Lake Melville graben and Double Mer half graben, although it has an east–west trend, rather than being parallel to the margins of the rift system.

In a broader context, the Lake Melville rift system is part of the St Lawrence rift system, which also includes the Ottawa–Bonnechere and Saguenay grabens. These were initiated during rifting that led to the opening of Iapetus Ocean, although reactivated in later times (Kumarapeli and Saull, 1966; Kumarapeli, 1985; O'Brien and van der Pluijm, 2012). Noting that alkaline intrusions are an integral part of rift-related setting including the St. Lawrence rift system, Gower *et al.* (1986a) speculated that quartz monzonite at Mokami Hill, together with some other weakly strained intrusions nearby, were potential representatives of analogous magmatism associated with the Lake Melville rift system. Gower and Kamo (1997) reported an age of 1417 ± 5 Ma for the Mokami Hill quartz monzonite, demonstrating that such is not the case.

18.1.1.5 Depositional Environment

No detailed analysis of depositional environment has been carried out on the Double Mer Formation, but some observations have been made by Stevenson (1970), Erdmer (1984) and Gower *et al.* (1986a). On the basis of crossbedding orientations, all three publications infer north-to-northwest sediment transport. For the Double Mer half graben, both Stevenson (1970) and Gower *et al.* (1986a) envisaged a north-sloping graben floor coupled with a faulted-bounded northern paleo-edge as a model that allowed reconciliation between the southwestward fining of conglomerate and northwest current flow. A braided fluvial environment was preferred by Gower *et al.*, but a shallow-marine or lacustrine setting was also considered plausible.

18.1.1.6 Age of the Double Mer Formation

Low (1896) correlated the Double Mer Formation with the Bradore Formation (Early Cambrian) in the Strait of Belle Isle region. This view was echoed by Kindle (1924), Kranck (1947) and Stevenson (1970), although, as observed by Gower *et al.* (1986a), this interpretation had not gained any increased validity for having been repeated several times. In any case, correlation with the Bateau Formation would be a better choice.

The only unequivocal constraint on the age of the Double Mer Formation is that it must postdate Grenvillian uplift (*ca.* 930 Ma, Gower *et al.,* 2008a, b), because it is unmetamorphosed and contains clasts of Grenvillian basement. Detrital geochronological data are available for two samples (Spencer *et al.*, 2015). Sample CS11-1 yielded age peaks at 1500, 1230, 1060 and 1000 Ma. Sample CS11-3 yielded age peaks at 1650, 1500, 1230, 1100 and 980 Ma. No post-Grenvillian material was found. In addition, samples of Double Mer sandstone from five localities were processed for palynomorphs, but no recognizable microfossils were found (Erdmer, 1984).

Gower *et al.* (1986a) offered a strain-ellipsoid structural model as an indirect approach for inferring the age of the Double Mer Formation. Their model integrated the overall configuration and orientation of the Lake Melville rift system with various features within it. The reader is referred to the paper for details, but key elements of the model were that the direction of principal extension of the inferred strain ellipsoid is normal to the Long Range mafic dykes in southeast Labrador, and both graben and dyke emplacement were suggested to have been more-or-less synchronous. At that time, the age of the mafic dykes was only imprecisely known, but was subsequently determined to be 615 Ma by Kamo *et al.* (1989) and Kamo and Gower (1994) (*cf.* Section 18.2 on Long Range dykes).

Murthy *et al.* (1992) carried out paleomagnetic studies on samples of Double Mer Formation. Five sites yielded a mean (tilt-corrected) direction of D = 110.8°, I = 50.1°, which was interpreted to be primary. The direction is similar to that obtained from the Long Range dykes, reported in the same publication (D = 124.8°, I = 55.5°). Although this study would not be considered rigorous by current standards, at the time it was offered (and the offer is still open), it was the first quantitative indication for the age of the Double Mer Formation.

Accepting that the Double Mer Formation, Long Range dykes, and the Lake Melville rift system are integral to the St. Lawrence rift system, then linkage between them is indirectly strengthened by the dating results of O'Brien and van der Pluijm (2012). In this study, the Ar geochronological data obtained from pseudotachylytes from the Montmorency fault, near Québec City (which defines the northern margin of the St. Lawrence rift system in that area), yielded ages of 613.3 and 614.2 Ma. These are analytically identical to the U–Pb baddeleyite/zircon ages from the Long Range dykes obtained by Kamo *et al.* (1989) and Kamo and Gower (1994).

18.1.2 DOUBLE MER FORMATION CORRELATIVES

Apart from the outcrops in the two grabens reviewed in the previous section, there are several small, isolated occurrences of similar rocks that are, most obviously, correlated with the Double Mer Formation. The first two, strictly, are outside the region covered by this report (*cf.* Figure 18.1).

18.1.2.1 Churchill River

Outside the area addressed in this report, two outcrops, 6.5 km apart, of conglomerate and sandstone on the Churchill River were described by Stevenson (1967a) and correlated with the Double Mer Formation. The rocks were described as northwest-dipping, thick-bedded, red conglomerate and sandstone, gradational into layers of maroon shale. The two outcrops were re-examined by Wardle *et al.* (1988). Expanding on Stevenson's description, they noted that the conglomerate clasts consist almost exclusively of quartzite set in a matrix of coarse pink arkose, locally cemented by carbonate. The quartzite clasts are white to pale-green, highly rounded and very fine grained.

Stevenson (1967a) interpreted easterly paleocurrent flow, whereas Wardle *et al.* (1988) reported it to be northwest. Given the limited data, the best that can be said is that it is broadly consistent with that obtained for the Double Mer Formation farther east.

18.1.2.2 Kenamu River

Occurrences of flat-lying to gently dipping conglomerate, arkose and shale are exposed in the banks of the Kenamu River (west of the area addressed in this report, but included in the author's database). They were first mapped by Stevenson (1967a, b) and correlated with the Double Mer Formation. The outcrops were re-examined by James and Lawler (1999). During construction of the Trans-Labrador Highway, additional exposures were created in two large quarries (Plate 18.2A), which were visited by the author in 2009 (CG09-058, CG09-059). In the quarries, the arkose is mauve-, brown-, dull red-, maroon-, or purplish-orange-weathering and pebbly conglomerate layers contain subrounded to angular clasts, generally less than 1 cm in diameter, consisting of quartz, K-feldspar, and granitoid rock. Crossbedding is present, but no paleocurrent measurements have been made.

The quarries were revisited by the author in 2012. The quarry faces are degrading fairly rapidly and will not last long.

18.1.2.3 The Backway (N*Dm*)

The Backway conglomerate is exposed in 2-m-high cliffs (*ca.* 100–200 m in length) at the edge of a small promontory on the north side of The Backway, about 6 km from its eastern end (Figure 18.1). It was discovered during 1:100 000-scale mapping by Gower *et al.* (1981) and is the only known exposure in the area (RG80-133). It consists of clast-supported conglomerate that includes boulders of metagabbro, anorthosite, augen granodiorite, amphibolite, and quartzite (Plate 18.2B). Similar basement rock types occur in the surrounding area.

The outcrop occurs south of a major east-trending fault, termed The Backway fault by Gower *et al.* (1986a). This fault defines the north side of The Backway and can be inferred to continue east of The Backway for about 100 km. The fault is marked by a 0.75-km-wide zone of low-grade alteration and brecciation.

As noted by Gower *et al.* (1986a), outcrops of basement gneiss south of the fault and west of the conglomerate are topographically higher than the conglomerate, implying that the conglomerate is a remnant preserved in a pre-existing basement hollow.

The matrix of the conglomerate and a norite clast were examined in thin section. The matrix (RG80-133A) consists of orthopyroxene,

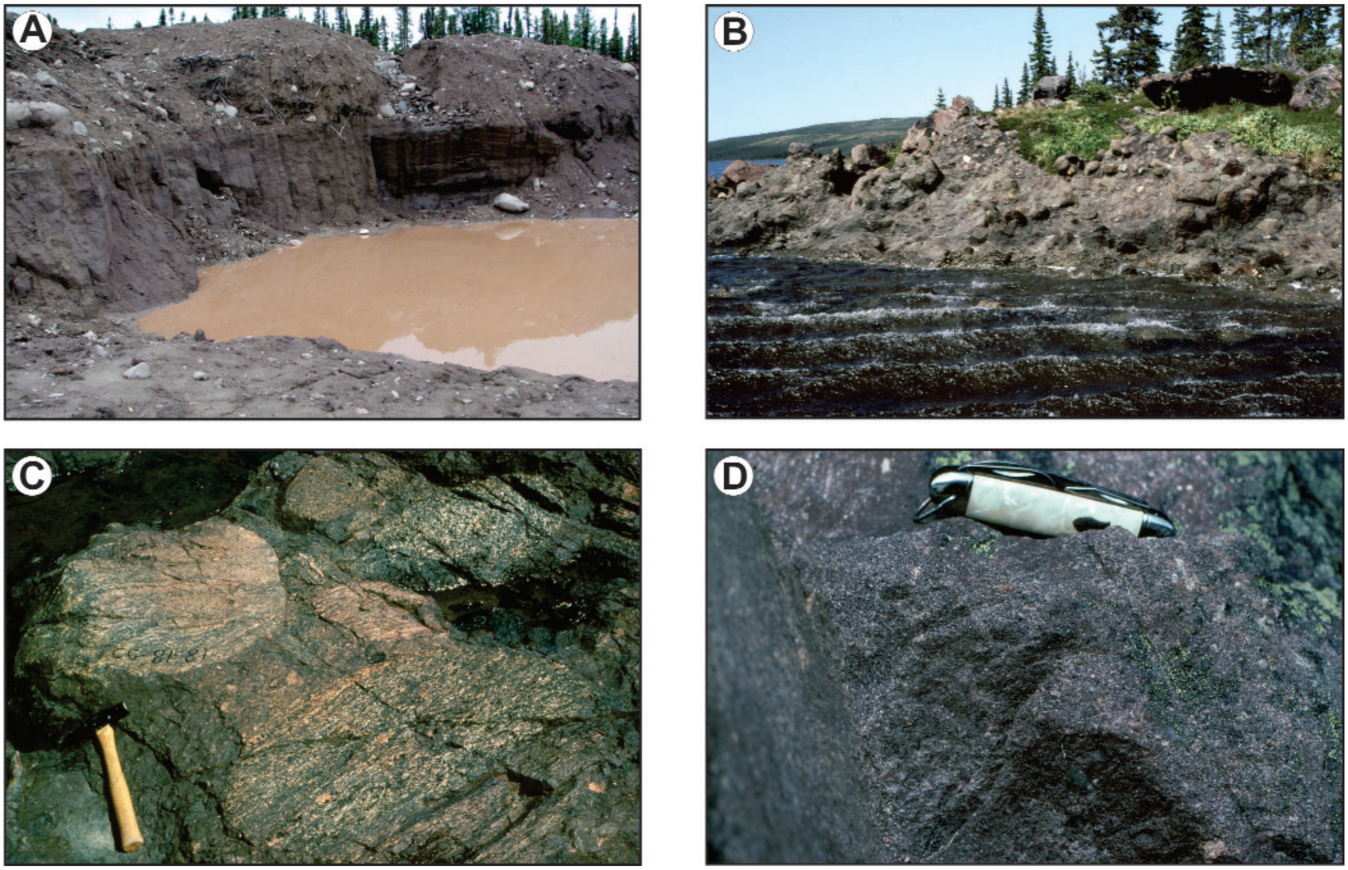


Plate 18.2. *Double Mer Formation correlative occurrences. A. Double Mer Formation arkose correlative, Kenamu River area (Gower, 2012; field trip guide, Stop 1.2), B. Double Mer Formation conglomerate correlative, The Backway (RG80-133), C. Double Mer Formation conglomerate correlative, northwest side of Sandwich Bay (CG81-081), D. Gilbert arkose. Possible Double Mer Formation correlative (CG86-490).*

> plagioclase, quartz, opaque minerals, clinopyroxene, epidote, chlorite, and amphibole. It was clearly derived from a mafic intrusive source. The norite clast (RG80-133B) provides a plausible example of that source, consisting of severely sericitized plagioclase, altered orthopyroxene, aggregates of blue-green amphibole, relict clinopyroxene, titanite and opaque minerals.

18.1.2.4 Sandwich Bay Conglomerate (N*Sb*)

The Sandwich Bay conglomerate is situated on the northwest shore of Sandwich Bay, about 17 km west-southwest of Cartwright (Figure 18.1). It is only known at one locality (CG81-081), which was discovered during 1:100 000 mapping in 1981 (Gower *et al.*, 1882b). Mention of the Sandwich Bay conglomerate was made by Gower *et al.* (1986a) and Gower (1988) in articles addressing the Double Mer Formation and the Lake Melville rift system.

The outcrop has an area of about 100 m^2 (guestimate from memory) on the shoreline and is mostly covered at high tide. The conglomerate consists of rounded to subrounded boulders of gneiss, granite, and gabbro up to 1 m in diameter in a pebbly matrix of similar material (Plate 18.2C). Clasts in the conglomerate are similar to basement rocks to the northwest.

The occurrence is confined on its northwest side by a fault parallel to, and essentially defining, the northwest side of Sandwich Bay. It was termed the Eagle River fault by Gower *et al.* (1986a), and is a zone of breeciation and low-grade alteration that can be traced for more than 40 km southwest of Sandwich Bay. Despite the spatial linkage with the Eagle River fault, the rounded nature of the boulders denies accepting the conglomerate to be simply talus at the base of the fault scarp.

18.1.2.5 Gilbert River Arkose (N*Gi*)

Only one outcrop is known (CG86-490), which is on the north shore of Gilbert River about 800 m west of Gilbert Lake (Figure 18.1). The name 'Gilbert arkose' is used here. It is a modification from 'Gilbert conglomerate', which was the name applied by Piloski (1955), who first discovered it and termed it a coarse grit or pebble conglomerate. The term ‘Gilbert conglomerate’ was also used in descriptions by Eade (1962), Bradley (1966), Gower *et al.* (1986a, 1987) and Gower (1988).

The rock is maroon-weathering, homogeneous arkosic sandstone to pebbly grit. The pebbles and grains are rounded to subangular, generally less than 0.3 cm in diameter and consist mostly of quartz and minor feldspar (Plate 18.2D). Bradley (1966) also noted the presence of granitic gneiss fragments. The arkose and conglomerate are confined to a 5-m-wide, parallel-sided, vertical zone that is parallel to the Gilbert River fault. The country rock is a mylonitized K-feldspar-megacrystic granitoid unit, but has been described previously as paragneiss (Piloski, 1955), or flow-banded rhyolite (Bradley, 1966). Field relationships have also been interpreted in various ways. Eade (1962) regarded the arkose as unconformably overlying mylonite, but notes that the distinct bedding is steep to vertical, and suggested that preservation of the sediment was due to block downfaulting. Eade’s comments are based in the field notes of one of his field assistants (data station R61-100). The site is located on the aerial photograph 300 m from the conglomerate described by Bradley and that seen by Gower *et al.* (1987). Most likely, the discrepancy is due to inaccurate plotting, but the possibility remains that more than one occurrence exists. Bradley (1966), who has made the most detailed examination, interpreted the outcrop as a clastic dyke bounded on its south side by a minor fault. Bradley also observed that the arkose and the surrounding country rock are crosscut by clastic dykes less than 5 cm wide. From observations made by the author in 1986 and during later visits, Bradley’s interpretation is to be preferred.

A thin section of the Gilbert arkose (CG86-490A) comprises rounded to subangular quartz grains, rounded grains of microcline, sparse subrounded plagioclase, opaque material (mostly as hematite cement, but also some larger clastic magnetite grains), and a few small, ragged flakes of biotite within composite quartz-feldspar clasts. The texture is unequivocally that of clast-supported arenaceous sediment. The only sign of any superimposed deformation takes the form of narrow crush zones transecting the sample.

The age of the Gilbert arkose is unknown, beyond that it must postdate Grenvillian orogenesis. It has traditionally been considered to be correlative with the Double Mer Formation in the Lake Melville rift system and/or the Bateau Formation in southeasternmost Labrador. Both of these are considered late Neoproterozoic, related to rifting prior to the opening of Iapetus Ocean. Alternative possibilities are: i) that the Gilbert arkose formed earlier, at the same time as the 974 Ma Gilbert Bay alkalic mafic dykes; or ii) later, at the same time as the Sandwich Bay dykes, Charlottetown Road dykes and/or Battle Harbour dykes. The Gilbert arkose is spatially associated with both the Gilbert Bay and Charlottetown Road mafic dykes, and also with the Gilbert River fault, which was clearly active during the waning stages of Grenvillian orogenesis. Given that the Gilbert Bay and Charlottetown Road dykes were both emplaced at a shallow level, it is not inconceivable that sand could have been transported from the surface down the same system of fissures as those utilized by the ascending mafic dykes. It would be interesting to find a site where sand and mafic magma met. Such is by no means improbable as the closest known Charlottetown Road dyke is only 2.5 km from the Gilbert arkose locality.

18.1.2.6 Clastic Dykes on the Shore of Lake Melville (N*c*)

Four clastic dykes have been discovered on the shores of Lake Melville, all close to its eastern end. One occurrence was mentioned by Emslie in Etagualet Bay (field notes; Emslie, 1975; EC75-235), Emslie recorded a north-north-west-trending, 6–8”-wide dyke having a filling of dark-brown sandstone with angular inclusions of anorthosite and diabase. He also notes a smaller (2”-wide) dyke parallel to the large dyke. Emslie (1976) mentions discovery of two clastic dykes, but it is unclear whether he is including the smaller dyke at this site, or had an additional locality in mind. Evans (1951) noted a small outcrop of conglomerate on the south shore of Etagaulet Bay containing pebbles of anorthosite, quartz and chert and having anorthosite on either side of it. The location is not specified, but the descriptive location is consistent with that of Emslie’s site, and it seems probable, despite apparent clast differences, that it is the same site.

The three localities found during 1:100 000-scale mapping are all outside the area mapped by Emslie (NN80-148, NN80-149 and NN80-350) and were first reported by Gower *et al.* (1981). The narrowest dyke (NN80-350) is about 10 cm wide and the widest about 1 m (NN80-148) (Plate 18.3A). Trends of 166° and 175° were recorded, which are similar to that for Emslie’s locality. The material in the clastic dykes consists mostly of angular clasts of granitoid rocks within brown or mauve sandstone. Clasts probably exceed enveloping sandstone in volume.

One clastic dyke was examined in thin section (NN80-148). It contains a wide variety of angular, unsorted quartz, plagioclase, K-feldspar and opaque-mineral fragments and composite granitoid rock clasts associated with minor sericite and cemented by hematite.

It is noteworthy that at both outcrop- and thin-section-scale the clasts are angular and closely reflect the surrounding host rock, and thus have a nearby source. Emslie (1976) also pointed out that as the dykes are at lake level and if the obvious correlation with the Double Mer Formation is made, then the north and south shores of Lake Melville must have been at about the same relative elevations at the time of deposition of the Double Mer Formation, unless the

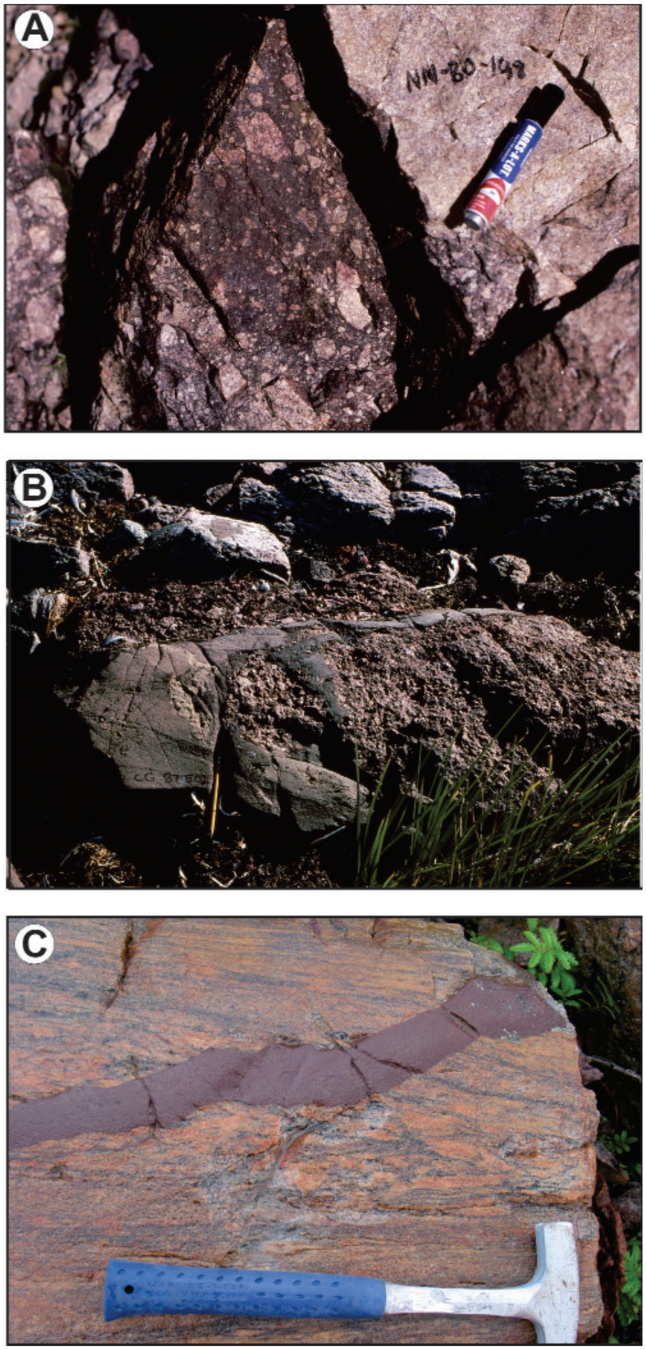


Plate 18.3. *Clastic dykes; assumed correlative with Double Mer Formation. A. Clastic dyke comprising breccia fragments in arkose matrix (NN80-148), B. Clastic dyke breccia with later clastic dyke arkose (CG87-502), C. Clastic dyke arkose (CG07-140).*

dykes were formerly at 0.5–1.0 km depth. He adds that, if the northwest-facing escarpment of the Mealy Mountains is a fault (which has been its consistent interpretation), then movement on that fault must have preceded deposition of the Double Mer Formation.

18.1.2.7 Clastic Dykes on Great Caribou Island (N*c*)

Two clastic dykes were discovered by the author on Great Caribou Island. One was reported by Gower *et al.* (1988); the second one was discovered in 2007, for which information is published here for the first time. The two dykes have trends of 087° and 110°, respectively, which are similar to their common west-northwest orientation, in alignment with St. Lewis Inlet.

The dyke discovered by Gower *et al.* (1988) is 30–50 cm wide and is made up of two parts (Plate 18.3B). The older part is a conglomerate/breccia containing fragments of K-feldspar and quartzofeldspathic rocks. This is cut by a maroon sandstone dyke containing fragments of the earlier breccia in a matrix of carbonate, hematite, quartz and feldspar.

In thin section (CG87-502A), the rock is seen to comprise angular granite clasts, and angular crystal-fragment clasts of plagioclase, microcline, quartz and apatite, all in a matrix of secondary opaque oxides (hematite and limonite), carbonate and chlorite.

The dyke discovered in 2007 is 1.1 km to the west-northwest of the original discovery. The dyke is 4 cm wide and consists of dark-brown to maroon, fine- to medium-grained, homogeneous sandstone (Plate 18.3C).

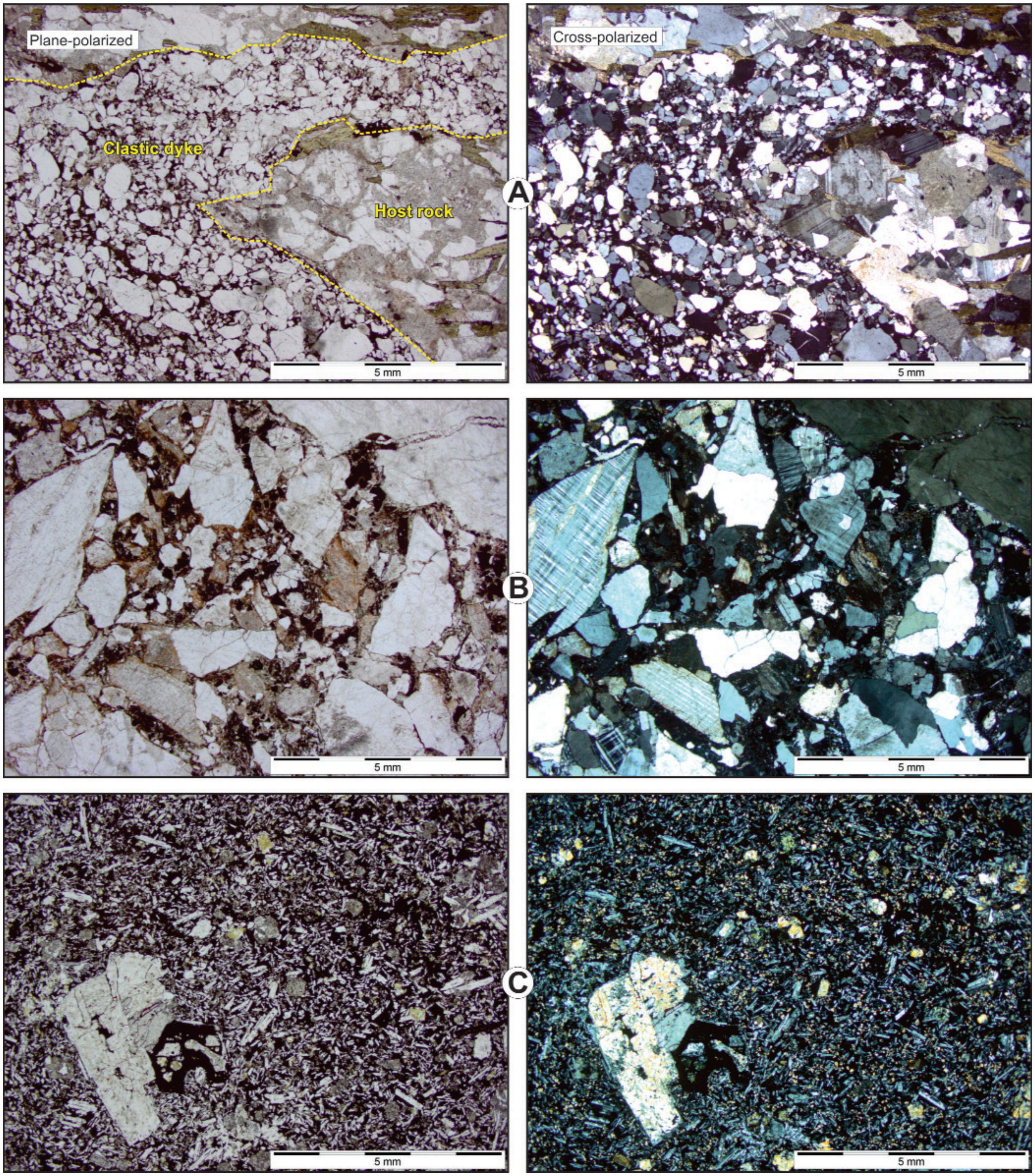
In thin section (CG07-140), the sandstone is seen to contain angular to subrounded quartz, plagioclase and microcline in a matrix of hematite and carbonate (Photomicrograph 18.1A).

18.1.2.8 Clastic Dykes at St. Augustin

Although, strictly, outside the area of this report, clastic sandstone dykes reported by Davies (1968) are worth mentioning. Davies described the clastic dykes being up to 3-cm wide, yellowish-grey, sugary, medium-grained, quartz-cemented sandstone. Given the dykes are presently at sea level, he makes a similar argument to that of Emslie (*see* above) regarding present-day topography closely reflecting that which existed during the Neoproterozoic.

18.1.3 LOWER LABRADOR GROUP

The Labrador Group, as defined by Williams and Stevens (1969), was divided by Knight (in Gower *et al.*, 2001) into lower and upper parts. The lower Labrador Group comprises the Bateau and Lighthouse Cove formations and is interpreted to be Neoproterozoic. The upper part comprises the Bateau and Forteau formations in southeast Labrador and the Hawke Bay Formation in Newfoundland. It is latest Early Cambrian to Middle Cambrian. The upper Labrador Group, by this usage, conforms to the original definition of the Labrador Series by Schuchert and Dunbar (1934).



Photomicrograph 18.1. *Neoproterozoic supracrustal rocks. A. Clastic dyke in granite on Great Caribou Island (CG07-140), B. Regolith at base of Lighthouse Cove basalt on St. Peters Islands (CG87-478D), C. Lighthouse Cove Formation basalt showing plagioclase and opaque mineral phenocrysts and skeletal plagioclase in the groundmass. Henley Harbour (CG87-441B).*

18.1.3.1 Bateau Formation (NC*Ba*)

Orange-, red- and maroon-weathering conglomerate, pebbly sandstone and arkose that unconformably overlies Grenvillian basement and underlies columnar jointed basalt in southeast Labrador is correlated with the Bateau Formation of Williams and Stevens (1969). These rocks are found on islands west of Peterel Island, at Table Head and at outliers northwest of Table Head (Figure 18.1, inset). The present distribution of the Bateau Formation appears to be restricted to the west and south by northeast trending and northwest-trending faults, respectively. That the Bateau Formation was deposited in a fault-bound basin (or basins) is supported by lateral lithological and thickness variations. At Table Head and the outliers to the northwest, talus from the overlying basalt prevents accurate measurement of section thickness. The maximum thickness is probably about 20 m. On the most southerly of four small islands forming the Peterel Islands, an almost complete 5 m section of Bateau red arkose and conglomerate is exposed. In general, there appears to be a southeasterly thinning of the unit.

The arkose includes both parallel lamination and cross-bedded layers, which have master bedding planes, up to 30 cm apart, defined by colour, grain size variation and heavy mineral laminae. The conglomeratic layers contain driekanter-shaped pebbles of quartz, granite and pegmatite derived from the underlying quartzofeldspathic basement (Plate 18.4A, B).

In thin section (CG87-441A, CG87-480A, VN87-305B), the arkose is seen to consist of angular crystal fragments of plagioclase, K-feldspar (microcline and perthite), and quartz of various sizes, grading from medium-grained to fine-grained groundmass material. Accessory zircon is present. Opaque oxide grains, also angular, occur in zones. Secondary oxides, chlorite, titanite, and carbonate are common, partially in the form of interstitial cement. The arkose lacks sorting and, clearly, the material is not far travelled. The fragments are contained within carbonate–hematite cement.

18.1.3.2 Regolith on Southernmost St. Peter Islands (Included with NC*Lc*)

On the most southerly of the St. Peter Islands, the stratigraphic position of the Bateau Formation is represented by a 2–30-cm-thick breccia layer derived from a basement regolith and intermixed with scoriaceous material from the overlying Lighthouse Cove Formation basalt flow (Plate 18.4C, D).

The basal breccia (thin section CG87-478D: Photomicrograph 18.1B) comprises angular clasts of plagioclase, K-feldspar, quartz and opaque oxides in a fine-grained, non-descript matrix composed of epidote, titanite, chlorite, secondary Fe oxides (hematite/limonite), carbonate, and fine-grained quartz and feldspar. A sample taken from the interface between the breccia and the overlying basalt (thin section CG87-478E) contains one 1-cm-long ovoid patch of very fine-grained basalt in a coarse-grained matrix comprising plagioclase, microcline, quartz and a bottle-green mineral thought to be an Fe-rich amphibole. The coarse-grained quartzofeldspathic matrix is intermixed with fine-grained material (in which any mineral is hard to identify with certainty using a petrographic microscope), which is suspected to be a mixture of feldspar, quartz and phyllosilicates. The coarse- and fine-grained components may represent sand and mud incorporated into the base of the basalt flow.

18.1.3.3 Lighthouse Cove Formation (NC*Lc*)

Basalt overlying the Bateau Formation in southeast Labrador is correlated with the Lighthouse Cove Formation in Newfoundland (Williams and Stevens, 1969; Strong and Williams, 1972). The Lighthouse Cove Formation is spectacularly exposed in the Henley Harbour area, where it forms flat-topped hills, one of which is known locally as the Devil's Dining Table. Previous early mention of the basalt at Henley Harbour was made by Douglas (1953), Eade (1962) and Clifford (1965). The Lighthouse Cove Formation is also found about 12 km to the northeast on the St. Peter Islands, and 16 km north-northeast at Table Head and other nearby mainland outliers.

At Henley Harbour, Bostock (1983) describes the formation as consisting of a single flow about 20 m thick (Clifford estimated 28 m) divisible into a lower columnar part and an upper hackly fractured part, resting on crystalline basement (Plate 18.4E, F). At St. Peter Islands, the basalt is approximately 15 m thick and, similarly to that present at Henley Harbour, can be also be divided into a lower columnar jointed part and an upper, hackly fractured part. At Table Head, the basalt is about 10 m thick. The uppermost parts of the basalt have been removed by erosion in all areas, so the original thickness is unknown. No later rocks overlie the basalt.

The basalts are brown- to blue-grey-weathering, locally with green or maroon hues and are plagioclase or pyroxene phyric and amygdaloidal in places. The amygdules are pink near the base of the flow, but black farther up. No pegmatite or mafic dykes were seen intruding either the Bateau or Lighthouse Cove formations.

Wanless *et al.* (1965) reported a K–Ar whole-rock age of 382 ± 28 Ma for the Lighthouse Cove Formation basalt from Table Head and Wanless *et al.* (1974) reported K–Ar whole-rock ages of 419 ± 17 and 429 ± 17 Ma for Lighthouse Cove Formation basalt from Henley Harbour. The dates from Henley Harbour are analytically identical to two K–Ar whole-rock dates of 435 ± 17 Ma and 421 ± 16 Ma from basalt flows northwest of Horse Chops Ridge area at the northern end of the Long Range Inlier in Newfoundland also reported by Wanless *et al.* (1974) (all ages recalculated). It has been assumed that the Henley Harbour flows are correlative with basalt at Canada Bay. As

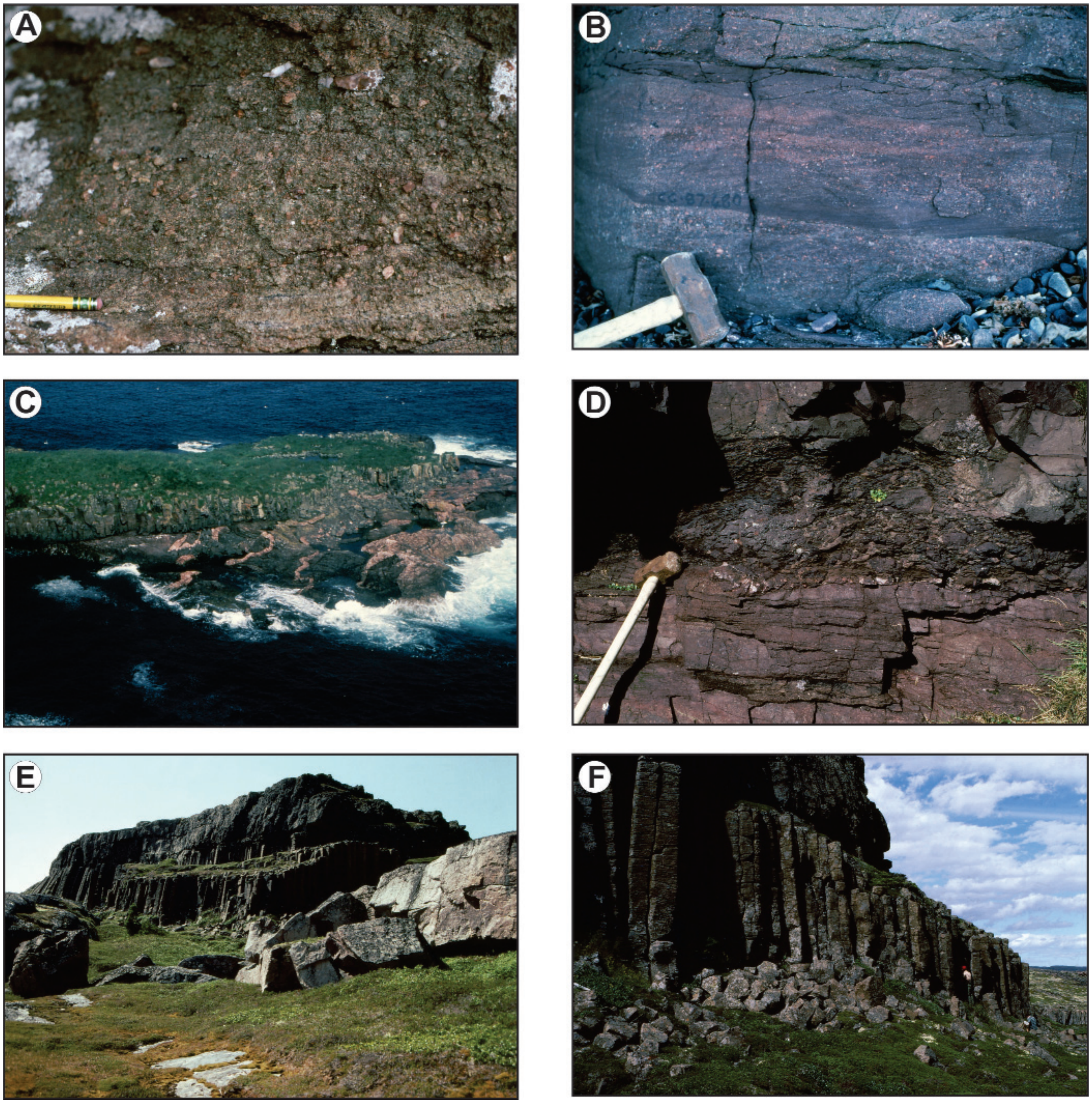


Plate 18.4. *Lower Labrador Group (Bateau and Lighthouse Cove formations). A. Bateau Formation; pebbly conglomerate (CG87-441), B. Bateau Formation; cross-stratified arkose and pebbly conglomerate (CG87-480), C. Basement gneiss, intruded by pegmatite dykes. Unconformably overlain by Lighthouse Cove Formation (CG87-478), D. Regolith between altered basement gneiss and Lighthouse Cove Formation basalt (CG87-478), E. Lighthouse Cove Formation showing lower columnar basalt and upper hackly basalt (CG93-774), F. Lighthouse Cove Formation columnar basalt (CG87-484).*

the latter are overlain by fossiliferous Cambrian strata, it is implied that the Henley Harbour K–Ar ages are too young, leading Wanless *et al.* (1974) to interpret the dates as reflecting Ordovician uplift (Bostock, 1983). If correlation with the basalt at Canada Bay is denied, then the K–Ar ages could be taken to mean a younger extrusion age for the Lighthouse Cove Formation.

Seven samples of Lighthouse Cove basalt were examined in thin section (CG87-441B – Photomicrograph 18.1C, CG87-442B, CG87-478F, CG87-480B, CG87-484, EA61-397 from locality R61-104, VN87-305A). The thin sections show a wide range in colour (black, green, grey, and orange), and grain size (glassy to fine/medium grained). Sample CG87-478F is amygdaloidal. Plagioclase forms euhedral, well-twinned laths exhibiting skeletal/quench textures and commonly having sericitic cores. Clinopyroxene occurs as subhedral, colourless, high-relief grains. Both plagioclase and clinopyroxene range in size from small phenocrysts to very fine-grained crystals in the groundmass. Opaque oxides (mainly secondary hematite and limonite) appear to form much of the groundmass in the finer grained rocks. Other identifiable minerals include granular titanite, rutile, carbonate, chlorite and possibly tremolite–actinolite. A tiny, bright green mineral (pumpellyite?) is present in VN87-305A. Bostock (1983) also examined the basalt at Henley Harbour in thin section (BK71-082.A1, BK71-082.A2), and reported a similar mineral assemblage.

A petrochemical study of the Lighthouse Cove basalt at Henley Harbour was carried out by Greenough and Owen (2002). Like Bostock (1983), they concluded that only one flow was present. They addressed the origin of metre-scale layering evident in outcrop and argued that it formed during emplacement of the basaltic flow and was the result of convective redistribution of evolved residual liquids from a single parental magma.

18.2 LONG RANGE DYKES, QUARTZ VEINS AND CARBONATE VEINS

18.2.1 LONG RANGE DYKES (Nd)

The term ‘Long Range dykes’ was first applied by Bostock (1983) to a swarm of northeast-trending mafic dykes that intrude Grenvillian rocks of the Long Range Inlier in northwest Newfoundland. Although Fahrig and West (1986) introduced the term Trunmore Bay dykes for analogous dykes in southeast Labrador their name did not achieve further usage. Adoption of the name ‘Long Range’ to the dykes in southeast Labrador dates from Murthy *et al.* (1992) who considered that a separate name was not justified.

Note that some dykes having uncertain kindred with the Long Range suite are also addressed in this section (Open Bay, Beaver Brook, Double Island and Truck Island). They are included here because: i) at some previous stage of this project they have been assigned as Long Range dykes, ii) the case to exclude them is not definitive, and iii) they do not obviously belong to any other recognized group of dykes in southeast Labrador.

The dykes are addressed below in corridors. In each corridor, more than one major dyke may be present and, even if one main dyke dominates, minor satellite dykes may also be present (Figure 18.2).

18.2.1.1 White Bear River Estuary Dyke Corridor

This is the westernmost, longest and largest dyke of the Long Range swarm. On land, the dyke extends from St. Augustin, on the north shore of the Gulf of St. Lawrence in Québec, northward to the southern end of Porcupine Strand in Labrador, a distance of 340 km (Figure 18.2). The southern end of the dyke was mapped by Davies (1968) at St. Augustin and is indicated farther north in Québec, to near the Labrador border, by Avramtchev (1983a), who, curiously, failed to include the part that Davies mapped. The northern end of the dyke can be traced 20 km offshore using regional aeromagnetic data, to the limit of data coverage (*e.g.*, Oakey and Dehler, 2004). Patterns apparent in Goggle Earth seabed imagery (Figure 18.2, inset) suggest that the dyke continues at least another 30 km to the north-northeast, for a total length of 390 km.

For this dyke, Kamo *et al.* (1989) obtained a U–Pb near-concordant zircon and baddeleyite age of 615 ± 2 Ma with a long extrapolation to an imprecise lower intercept of 255 +112/–88 Ma from sample CG84-476 collected at the White Bear River estuary (Figure 18.2). This location is also one of Murthy *et al.*’s (1992; Dyke 1) paleomagnetic sites.

The dyke mostly has a 020° trend, except at its southern end, where it is 010°, but in a few places there is divergence. In the White Bear River estuary area, the dyke has a 045° trend. This area is very close to the margin of the Sandwich Bay graben, so the possibility exists that rotation due to faulting may have occurred, but it could be due to bridging between *en échelon* dyke segments. The latter explanation is likely to apply to a segment about 50 km south-southwest of the White Bear River estuary, where the dyke orientation, as determined from aerial photographs is *ca.* 040° for 2.5 km.

The width of the dyke is not well known, as both contacts with the rocks intruded are rarely exposed. At its southern end, Davies (1968) reported the dyke to be 1/8 mile wide (*ca.* 200 m). Topographic features suggest a 300 m width in places and, where the dyke crosses the Trans-Labrador Highway, it has a width of about 400 m. It is probable that the dyke is a series of lensoid, right-stepping, *en échelon* segments and that its width varies considerably according to location within a segment.

Undeformed gabbro (now known to be part of the dyke) was first recorded in the White Bear River estuary area in Labrador by Cherry (1978a), although is not shown on his 1:100 000-scale map (Cherry, 1978b). The gabbro was grouped by Cherry with many other gabbroic rocks that are now known to be unrelated. The White Bear River occurrences were examined by Gower in 1981 (CG81-279, CG81-280, CG81-281) and at a locality farther north (CG81-252). The dyke may have been seen earlier during Eade’s (1962) mapping at an outcrop 14 km to the north-northeast of White Bear River estuary. The rock at that site (R61-146) is reported, by one of Eade’s assistants, to be a

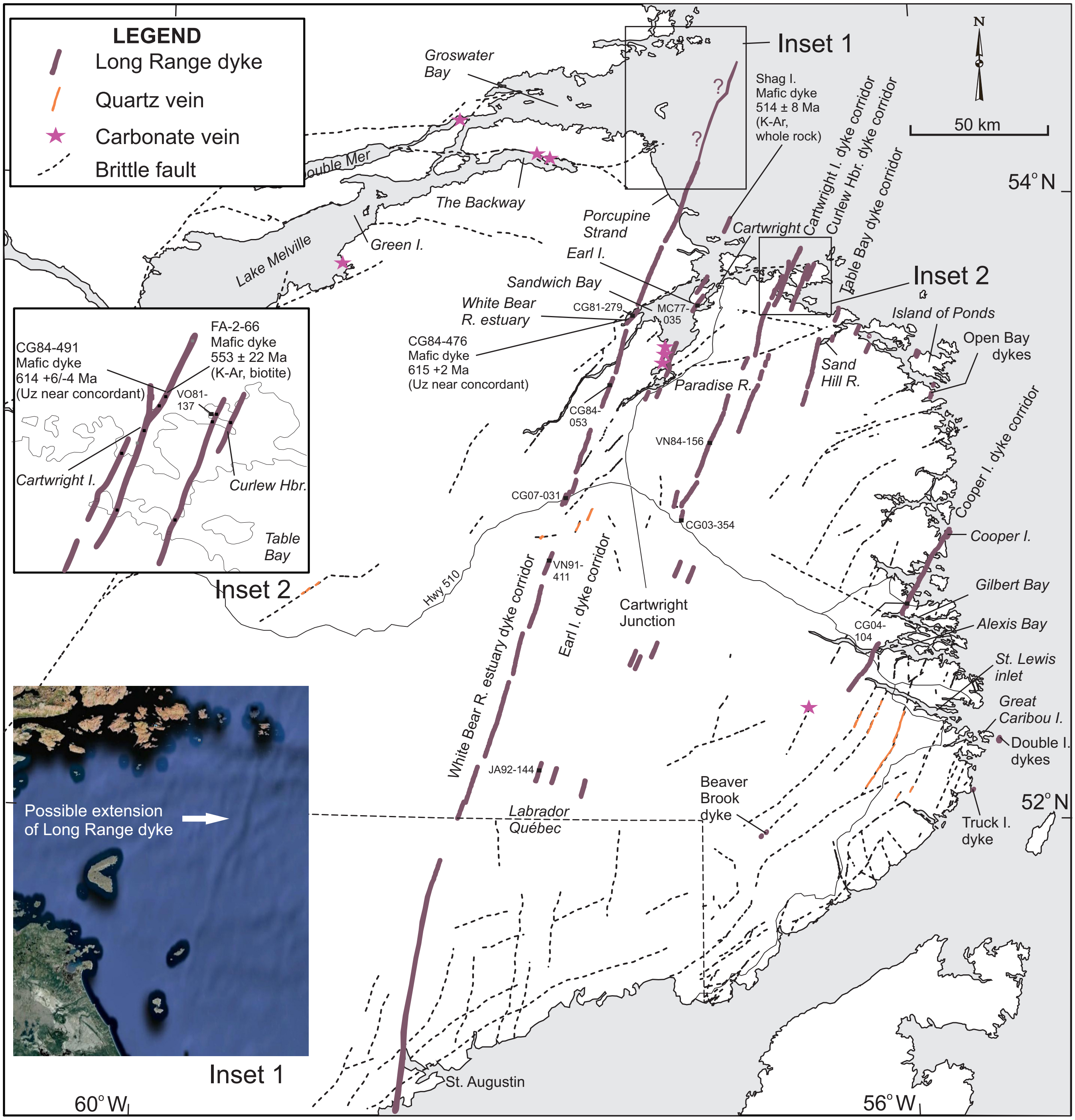


Figure 18.2. *Long Range dykes, quartz veins and carbonate veins.*

massive, coarse-grained gabbro, but this description could also apply to the Labradorian mafic plutonic rocks in that area. Interestingly, field notes for the site also mention a 5-ft-wide dyke-like body running through the outcrop that is suggested to be a conglomerate. It is described as containing quartz-feldspar fragments, rounded 'football-shaped remnants of gabbro' and rounded pieces and bands of white quartz. The author has not visited the outcrop but guesses that it might be a minor, xenolithic dyke within the major dyke – similar to that seen in one of the Long Range dykes at Curlew Harbour (VO81-131 – *cf.* Section 18.2.1.4).

Explicit inclusion of the dyke as part of the Long Range swarm was made by Gower *et al.* (1985), following mapping in the Paradise River map region, which allowed the dyke to be defined an additional 50 km to the south. Even farther south, in the Alexis River map region, other exposures of the same Long Range dyke were reported by van Nostrand (1992). No exposures of the dyke were found in either the Eagle River or Upper St. Paul River map regions (map regions outlined on Figure 17.10), although an obvious photo lineament in the Alexis River map region also extends a few kilometres into the Eagle River map region, and diffuse lineaments hint at where the dyke is located farther south. In the Upper St. Paul River map region, photo lineaments defining the dyke's location are only obvious in the southern part of NTS map area13B/08. Aeromagnetic data do not show any clear indication of the dyke in either map region. Gower (2000) noted that, at one location in the central part of NTS map area 13B/08, a hill of metagabbro is divided into two parts by a deep north-northeast-trending, 200–300-m-wide gully having straight, vertical, parallel-sided margins. He interpreted this to be the site of the now preferentially eroded Long Range dyke. The width is consistent with that recorded for the dyke to the north and south.

In addition to the coarse-grained main dyke (Plate 18.5A), minor, fine-grained diabase satellite dykes were found in several places. At White Bear River estuary, a 30-cm-wide dyke (CG84-476E) having a 025° trend intrudes the main dyke (Plate 18.5B). It is clarified that this minor dyke has Long Range chemical affinity, in contrast to the later Sandwich Bay dykes that also intrude the main dyke along the same stretch of shoreline, but at a *ca.* 100° trend. About 60 km to the south-southwest, two parallel small dykes, of unspecified width, in a single outcrop (CG84-238) were noted. Where the dyke crosses the Trans-Labrador Highway (CG07-031), a 2-m-wide satellite dyke was recorded, and about 25 km farther south-southwest, two satellite plagioclase-phyric dykes, each 1 m wide, were discovered at one locality (CG91-014) and a 25-cm-wide plagioclase-phyric dyke at another (VN91-411; Plate 18.5C). Satellite mafic dykes are also reported by Davies (1968) in the St. Augustin area.

The southern end of the dyke, at St. Augustin, is about 35 km north-northeast of the Baie des Moutons syenitic intrusion (Davies, 1968), and, by offshore extrapolation,

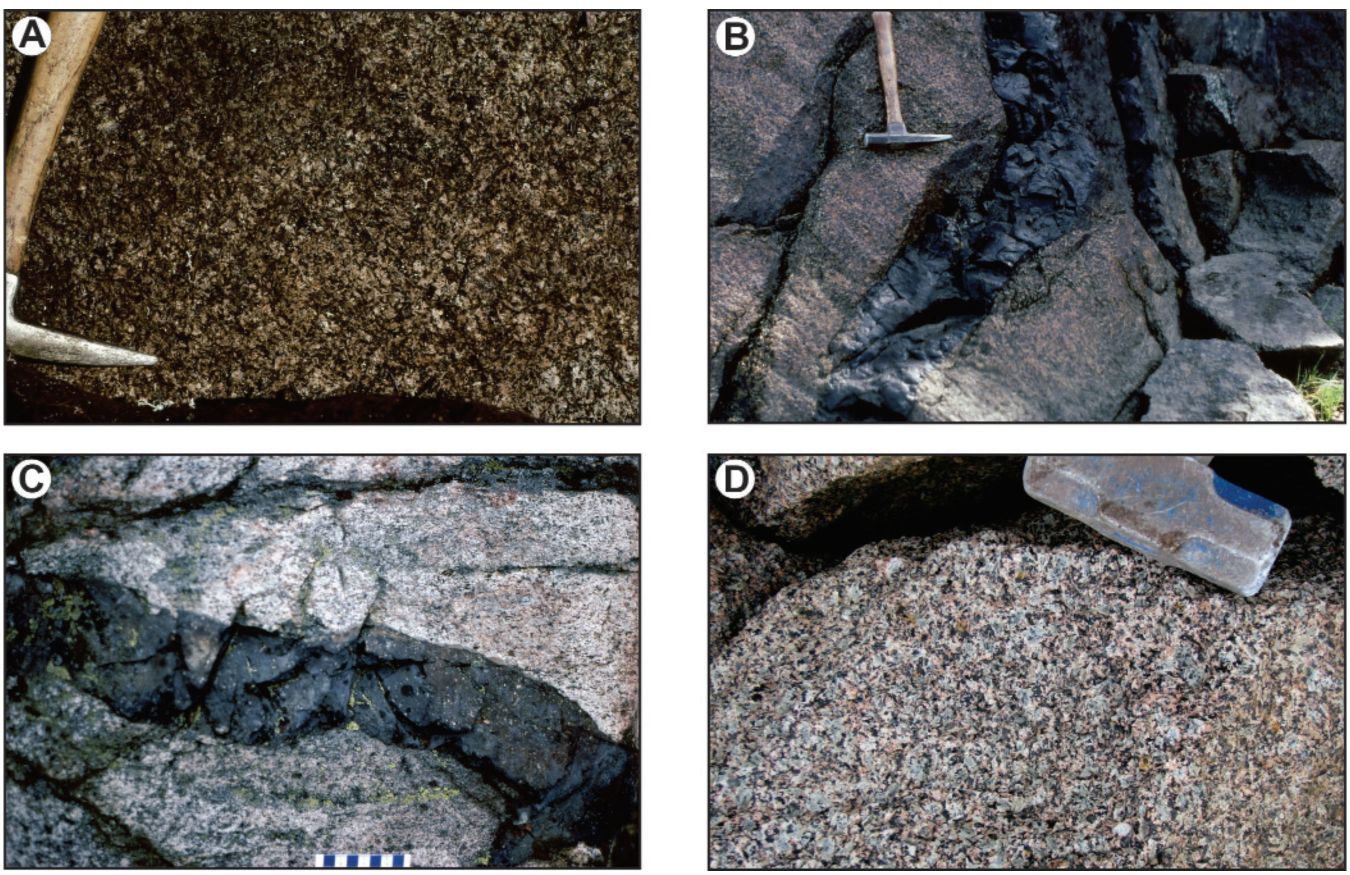


Plate 18.5. *White Bear River estuary Long Range dyke; lithological character and minor associated dykes. A. Typical coarse-grained texture of main dyke (CG84-053), B. Minor subsidiary dyke within main coarse-grained dyke (CG81-279/CG84-476E), C. Minor satellite dyke intruding granitoid rocks (VN91-411), D. Monzogabbro fractionate of dyke (CG07-031).*

could pass through it. Davies (1968) did not find intrusive relationships between the two on offshore islands. Dates reported for the Baie des Moutons intrusion include 631 ± 40 Ma (Rb–Sr, biotite; Davies, 1968), 590 ± 25 Ma and 602 ± 20 Ma (both K–Ar biotite, recalculated from the values reported by Davies, 1968), and 583 ± 2 Ma (Ar–Ar plateau age from overlapping hornblende and biotite spectra; McCausland *et al.*, 2011). Davies (1968) also reported an Rb–Sr biotite date of 470 ± 55 Ma for the dyke.

A remarkable feature of the dyke is that, from the coast of Labrador to 30 km south of the Trans-Labrador Highway, the dyke shows a gradational change in composition from gabbro in the north through K-rich gabbro, then monzogabbro and, in the southernmost exposure in Labrador, to syenite. It should be noted that the southernmost exposure is at a 300-m higher elevation than the coastal outcrops in the north. The gabbro is brown- or black-weathering, massive, coarse grained, and displays well-developed ophitic textures. The monzogabbroic to syenitic rock is pinkish-weathering, massive, coarse grained, and has textural similarity with its more mafic counterpart (Plate 18.5D).

Progressing farther south, the dyke is not seen again until south of the Labrador–Québec border. At St. Augustin, Davies (1968) recorded both gabbroic and syenitic variations and lithological gradations between the two. He noted that the syenite, in places, occurs as irregular veins and patches having gradational contacts with the enveloping gabbro and interpreted the syenite to be residual magmatic liquid. A similar relationship can be seen where the dyke crosses the Trans-Labrador Highway (CG07-031). The east margin of the dyke is a rubbly looking, sulphide-bearing inhomogeneous gabbro to leucogabbro that grades westward over a few metres into a pink-weathering, massive coarse-grained, monzogabbroic to syenitic rock, which extends westward for about 400 m (Plate 18.5D). The west margin of the dyke is not exposed (Gower, 2012).

Both the main dyke and three of the satellite dykes have been examined petrographically.

Thin sections from the main dyke collectively represent much of the exposed length of the dyke in Labrador. Two thin sections of gabbro examined from the White Bear River estuary area (CG81-279A, CG84-476A), the latter also being the sample for which a 615 ± 2 Ma emplacement age was determined by Kamo *et al.* (1989), plus one gabbro from 25 km farther south-southwest (CG84-053) are all petrographically similar. They contain primary, subhedral, well-twinned plagioclase; relict or pseudomorphed olivine; primary, anhedral, colourless to brown clinopyroxene; primary, red-brown biotite; primary, relict or secondary hornblende and/or tremolite–actinolite; accessory opaque oxide, apatite and titanite; zircon and baddeleyite. Secondary minerals are sericite (after plagioclase) and a serpentinous mineral (after olivine). Two thin sections of monzogabbro (CG07-031F, CG07-031G from the Trans-Labrador Highway, and a syenogabbro – CG91-014C, from a locality 30 km farther south) contain similar minerals, although in different proportions. The greatest difference is the presence of primary, poorly twinned, frond-like K-feldspar infilling interstices created by earlier crystallizing minerals. In thin section CG07-031G, K-feldspar shares the interstices with ovoid patches of chlorite, pumpellyite and devitrified glass that clearly represent the last residual liquid to crystallize (Photomicrograph 18.2A). This sample also contains an unidentified colourless high-relief mineral. Note that the stained slab for sample CG84-203, for which no thin section is available, lies geographically and compositionally between gabbro to the north and monzogabbro to the south.

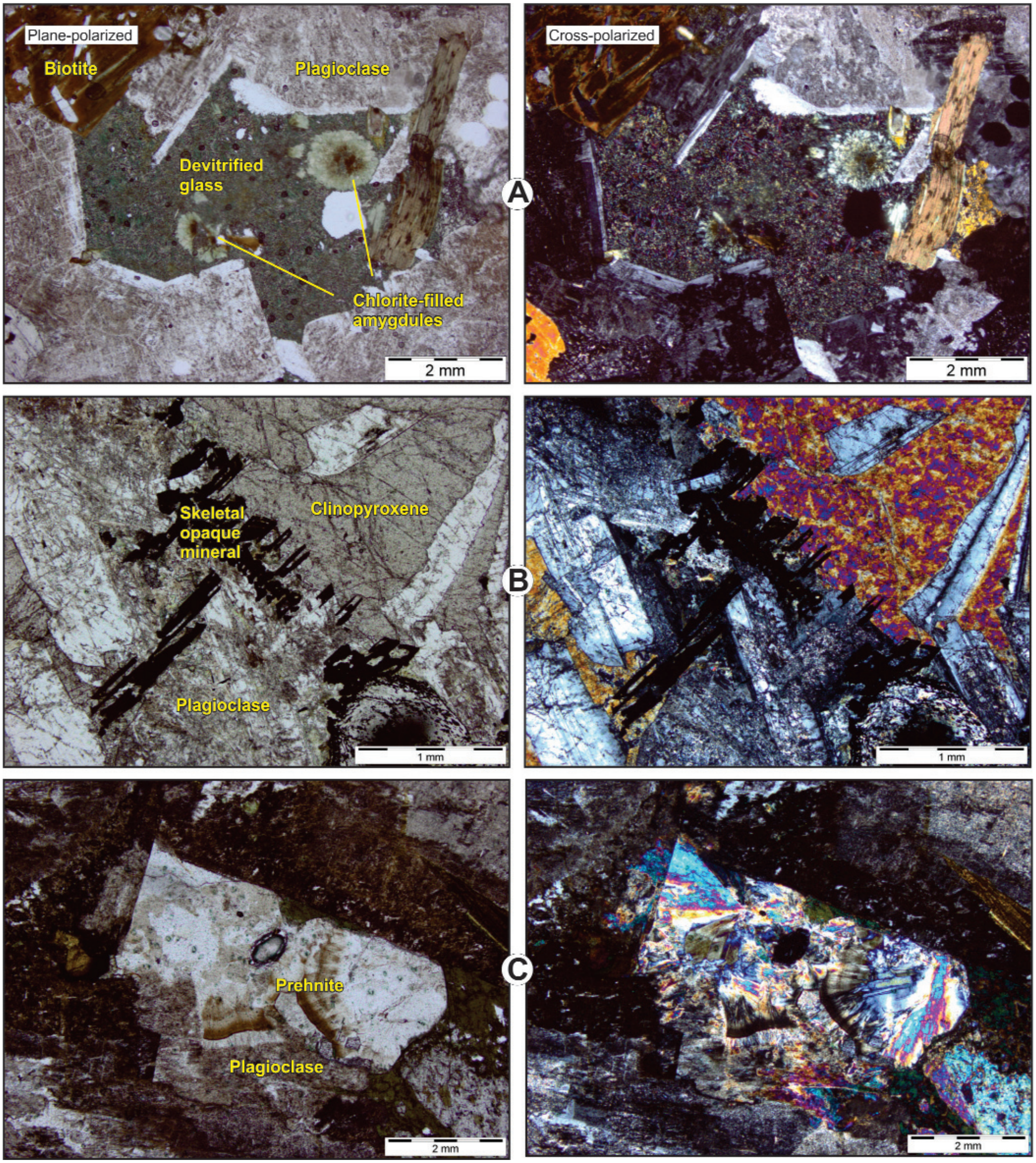
The satellite dykes vary from fine grained to very fine grained. All three thin sections (CG84-476E, CG91-014B, VN91-441B) contain euhedral plagioclase phenocrysts. In the matrix, quenched plagioclase, secondary tremolite/actinolite, biotite, and skeletal opaque minerals can be identified in all three sections. Altered clinopyroxene and olivine were equivocally recorded, and secondary chlorite and sericite noted. All three are amygdaloidal with chlorite fillings.

Despite the coarse-grained nature of the main dyke, the presence of quenched minerals and amygdules in the satellite dyke and the quenched residual interstitial liquid in the main dyke suggest emplacement at a high level, perhaps by transportation of a largely crystalline mush from deeper levels, although the ophitic texture displayed in the gabbro and an analogous texture in the monzo/syeno-gabbro seem at odds with such a mechanism.

18.2.1.2 Earl Island Dyke Corridor

The confirmed extent of the Earl Island dyke corridor is from Shag Island in the north (22 km north of Cartwright) to Paradise River in the south. At two roadcuts south of Paradise River, rocks were tentatively identified as belonging to the Long Range swarm, but no dyke margins were seen. The samples (CG04-195, CG04-200A) have not been followed-up either petrographically or geochemically; so it is probably wise to regard them cautiously. Although the outcrops collectively define a north-northeast trend, individual segments deviate and widths are variable. The Earl Island dyke corridor is equivalent to Dyke 2 of Murthy *et al.* (1992).

In the north, the dyke on Shag Island is 20 to 30 m thick and trends at 028°. Grasty *et al.* (1969), who were the first to map any part of this dyke, reported a K–Ar whole-rock age of 514 ± 8 Ma (recalculated) for it, describing it as fresh, coarse-grained olivine gabbro. Farther south, the major occurrence on Earl Island is on the southern shore where the dyke trends due north and has a width of about 30 m. Two other occurrences on the north side of Earl Island are minor satellite dykes, one having a reported width of 1.3 m. Another minor dyke has been recorded on Diver Island immediately to the north. All four occurrences were first located by Cherry (field notes), which remain the only source of information for one of the dykes (MC77-043).



Photomicrograph 18.2. *Long Range major dykes. A. White Bear River estuary corridor dyke showing devitrified glass and chlorite-filled amygdules (CG07-031G), B. Cartwright Island corridor dyke showing skeletal opaque mineral (JA92-144), C. Cartwright Island corridor dyke showing prehnite infilling between plagioclase crystals (VN84-156).*

From the southern side of Earl Island, Cherry (1978a; MC77-035) described the main dyke as undeformed diabase about 30 m wide. He also mentions that the main dyke is intruded by fine-grained aphanitic, plagioclase-phyric mafic dykes having the same orientation as the main dyke (Plate 18.6A). In the aphanitic dyke, the phenocrysts are concentrated in the core of the dyke. Although not mentioned by Cherry, two other features are evident from his photograph. The first is that the main dyke is texturally inhomogeneous, having dark, fine-grained wispy parts that also carry small plagioclase phenocrysts. These wispy parts show either sharp or gradational contacts with the coarse-grained gabbro and represent local quenching of the crystallizing gabbro crystal mush and incorporation of the chilled material (somewhat akin to scraping cooled material off the side of a saucepan). The second feature is that the separate aphanitic plagioclase-phyric dyke has slightly cuspate contacts against its gabbroic host, indicating emplacement before total consolidation of its host.

The most easily accessible segment in the Earl Island corridor is at Paradise River, in a roadcut at the north end of the community, where the full width and both contacts of a 70-m-wide, 015°-trending dyke are exposed. From Paradise River, the dyke can be easily traced on aerial photographs for 3.5 km to the north where it intersects the shore of Sandwich Bay. Between the two shoreline intersections, the dyke shows an uncharacteristic (for the Long Range swarm) sinistral dog-leg change in trend, locally adopting a 345° direction for 1.5 km. The dyke is brown- or black-weathering, massive, medium to coarse grained (but not as coarse grained as the White Bear River estuary dyke), and has an ophitic to slightly plagioclase-phyric texture.

Six thin sections (CG87-656, CG84-477, CG84-482.1, CG84-482.2, HN81-023, HN81-035) are available. Primary essential minerals are well twinned, euhedral to subhedral plagioclase, subhedral olivine (relict or pseudomorphed in three thin sections); anhedral, mauve-brown clinopyroxene, and red-brown to orange-brown biotite. Primary accessory phases are oxide minerals and apatite, and significant secondary minerals include tremolite/actinolite, chlorite, carbonate and white mica. In contrast to the White Bear River estuary dyke, olivine is common and K-feldspar lacking.

18.2.1.3 Cartwright Island Dyke Corridor

This corridor takes its name from Cartwright Island where the dyke was first recognized and photographed in 1939 from the air by Tanner (1944). The dyke continues north of Cartwright Island through Long Island and Tinker Island. Vague patterns from Google Earth seabed imagery suggest that it might continue north for another 70 km. The dyke is shown on the maps of both Christie (1951) and Kranck (1953), who also locate some of the other Long Range dykes in the vicinity. Mapping of the area was carried out by Kranck in 1949 and by Christie in 1950. Interestingly, although both reports were published by the Geological Survey of Canada, neither author seems to have much awareness of the other's investigations. A distinction between older/metamorphosed- and younger/unmetamorphosed- mafic rocks was made by Taylor (1951), a student of Kranck's, who completed a M.Sc. thesis on samples of gabbro and diabase from the area. The dyke is also shown on maps compiled by Brinex (Piloski, 1955) and by Eade (1962). The Cartwright Island corridor includes Dykes 3 and 4 of Murthy *et al.* (1992).

A K–Ar biotite age of 553 ± 22 Ma for the dyke was reported by Wanless *et al.* (1970), from a sample collected by W.F. Fahrig. Note that the co-ordinates originally reported by Wanless *et al.* (1970) place the dating site about 20 km to the west. There is no Long Range dyke at that location and the published co-ordinates were confirmed to be erroneous by W. Loveridge (personal communication to the author, 1985). A near-concordant U–Pb baddeleyite and zircon age of 614 +6/–4 Ma was reported by Kamo and Gower (1994) – data station CG84-491; Figure 18.2, inset 2; *cf.* also Plate 18.6B).

On the map of Gower (2010a; Table Bay map region), the dyke on Cartwright Island is indicated as dividing into two branches. This configuration is not firmly established. Other possibilities are that two dykes cross, or two parallel dykes exist in close proximity. Two dykes can be traced across the adjacent mainland south of Cartwright Island, and are exposed in the headwater areas of Table Bay. The eastern dyke, in the headwaters area of Table Bay (CG85-020), is intruded by a 0.6-m-wide subsidiary dyke trending at 160°. The two dykes can be traced a few kilometres south of Table Bay, and the westernmost dyke for a further 30 km south. The westernmost dyke is indicated as being offset dextrally by a northeast fault on the map of Gower (2010a), but this is not a relationship that has been rigorously established. The position of the westernmost dyke at its southern end is partly defined using high-resolution industry-obtained magnetic data. The trend of the dykes in the corridor varies between 0° and 30° and widths range from less than 1 m up to a guesstimated width of 160 m.

In a quarry 25 km southeast of Cartwright Junction (CG03-354), where the dyke is projected to cross the Trans-Labrador Highway, four satellite mafic dykes are exposed having widths of 1.0, 0.5, 0.5 and 0.3 m (Gower, 2012). The dykes contain plagioclase phenocrysts up to about 2 cm long and quenched plagioclase can be seen in the matrix in hand samples. The dykes are also amygdaloidal and have marked chilled margins. Three of the dykes are shown in Plate 18.6C.

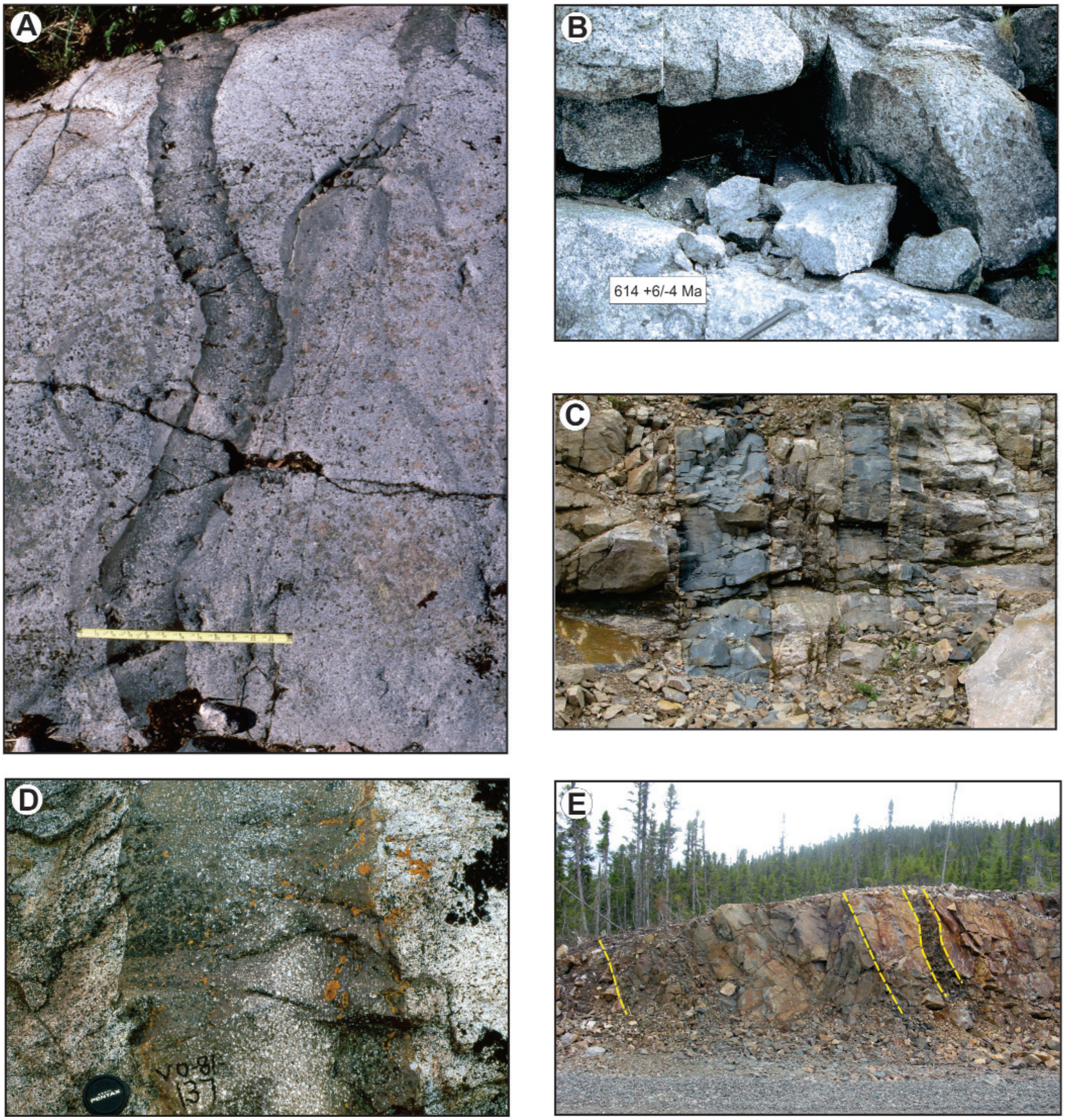


Plate 18.6. *Long Range dykes (Earl Island, Cartwright Island, Curlew Harbour and Cooper Island dykes). A. Inhomogeneous Earl Island Long Range dyke intruded by subsidiary dyke.* See *text for details (MC77-035), B. Cartwright Island Long Range dyke; U–Pb dated sample (CG84-491), C. Cartwright Island Long Range satellite dykes (CG03-354), D. Curlew Harbour Long Range satellite dyke (VO81-137). Compare with Plate 18.6A, 40 km to southwest, E. Cooper Island Long Range satellite dykes (CG04-104).*

South of the Trans-Labrador Highway, nine outcrops of Long Range dykes were found during 1:100 000-scale mapping (van Nostrand, 1992: Gower *et al.*, 1993). Apart from one riverside outcrop exposing a 10-m-wide dyke (VN92-161), very little field information was gathered at any of the outcrops, beyond noting the gabbro to be massive and unmetamorphosed. Samples were collected from all but one of them and stained slabs confirm a Long Range affiliation for most of them (based on their distinctive texture). There are two localities where some doubt remains: i) CG87-630, which is much finer grained and texturally distinct (no thin section or chemistry available), and ii) JA92-124, which is a very fine grained, K-rich, 1.5-m-wide mafic dyke. A thin section and a whole-rock analysis confirm that this rock is different, but its unmetamorphosed state, 020° trend, and alignment with the Cartwright Island dyke support a Long Range affinity. Davies (1968) notes that some trachyte dykes are associated with the Long Range dyke at St. Augustin; perhaps this is an example in Labrador.

The dykes are rusty-brown, black-, creamy- or even white-weathering, massive, generally coarse grained, unrecrystallized, and have an ophitic texture.

Eighteen thin sections are available for samples from the Cartwright Island corridor. Most of these are gabbro, but three are minor mafic dykes (CG85-020B, CG03-354H, JA92-124). As for the White Bear River estuary dyke, a major distinction exists between those in the north and those from the south, especially in K-feldspar content.

Samples of gabbro collected from north of the Paradise Arm pluton lack K-feldspar (CG84-487, CG84-488, CG84-491A, CG84-491C, CG84-491D, CG85-020A, GM85-037, VO81-102A). Their mineral assemblage is primary, subhedral to euhedral, moderately to strongly zoned plagioclase; primary, relict olivine; primary, anhedral brown clinopyroxene; primary, red-brown biotite; and primary, anhedral, green hornblende. Accessory primary minerals are opaque oxide (and, less commonly, sulphide) and apatite. Secondary minerals are tremolite/actinolite, white mica, and, sporadically, carbonate, chlorite, and serpentine.

Gabbro from within and south of the Paradise Arm pluton (DD91-081, DE91-044, DE91-046B, JA92-027, JA92-144, VN84-156, VN91-168B, VN92-161A) have the same primary minerals, but are more leucocratic and contain primary K-feldspar. In the case of DD91-081, K-feldspar is sufficiently abundant for the rock to be termed monzogabbro. Accessory and secondary minerals are mostly similar, except that apatite and the opaque mineral are skeletal in some samples (Photomicrograph 18.2B), and prehnite infilling a late-stage cavity was noted in one sample (VN84-156; Photomicrograph 18.2C), in a texture very similar to that seen in sample CG07-031G from the White Bear River estuary dyke. Crystallization of clinopyroxene and plagioclase, without disruption of the skeletal opaque mineral in JA92-144, implies very passive magmatic conditions.

The minor dykes all have features that characterize them individually. The minor dyke at Table Bay (CG85-020B) is atypical (for a Long Range dyke in southeast Labrador) in having a mottled texture in hand sample. It is also plagioclase-phyric and has oikocrysts of biotite. Its whole-rock composition (very similar to VN92-161A) and paleomagnetic signature indicate that it belongs to the Long Range suite, but perhaps is not genetically closely associated with the major Long Range dyke that it intrudes. The minor dyke from the quarry adjacent to the Trans-Labrador Highway (CG03-354H) is noteworthy in that both plagioclase and mafic minerals are mostly altered to secondary minerals. It contains abundant skeletal opaque oxides and chlorite-filled amydules. The southern example (JA92-124) is characterized by phenocrysts of a skeletal opaque mineral, skeletal plagioclase and chloritic interstitial material, which has completely pseudomorphed clinopyroxene.

18.2.1.4 Curlew Harbour Dyke Corridor

The Curlew Harbour dyke corridor comprises two major dykes and several minor dykes. The western dyke can be traced for about 60 km south from Curlew Harbour, and is well exposed on both sides of Table Bay, 11 km south of Curlew Harbour. The eastern dyke has only been traced for about 6 km in the vicinity of Curlew Harbour. The dykes trend between 20° and 25° and range in width from less than 1 m to a guestimated width of 120 m. Segments of the dykes are shown on the maps of Christie (1951) and Brinex (Piloski, 1955), but not on those of Kranck (1953) and Eade (1962). No geochronological data are available for the dykes in this corridor. The western dyke in this corridor is Dyke 5 of Murthy *et al.* (1992).

As for those dykes farther west, both major and satellite dykes are present. The major dykes comprise brown-weathering, coarse-grained, massive, unrecrystallized gabbro.

Four thin sections of gabbro are available from the Curlew Harbour area (CG81-393, CG84-489, CG84-490A, VO81-131A). The primary mineral assemblage is euhedral plagioclase; relict olivine; anhedral, brown clinopyroxene; red-brown biotite; hornblende (in two samples); accessory oxide and apatite, with secondary white mica and serpentine.

None of these contain K-feldspar, but, consistent with the pattern seen farther west, a stained slab of a sample collected 50 km south of Curlew Harbour does contain some interstitial K-feldspar (CG85-034 – no thin section or geochemical analysis available).

Minor mafic dykes are recorded from three data stations, of which VO81-131 is of particular interest. Here, a 1- to 2.5-m-wide dyke, having a 025° trend, intrudes coarse-grained Long Range gabbro. The minor dyke contains abundant, pink or creamy, 1- to 15-cm-long xenoliths of quartzofeldspathic country rocks (stained slab VO81-131B and outcrop photos VO81-131.1, VO81-131.2). Xenoliths in Long Range dykes in southeast Labrador are very rare, but another example of a xenolith-rich minor mafic dyke may exist at data station R61-146 (White Bear River estuary dyke; *cf.* Section 18.2.1.1). A subsidiary dyke at VO81-137 has plagioclase phenocrysts concentrated in its core, and is very similar in character to the dyke exposed on the southern side of Earl Island, although the two dykes are 40 km apart (compare Plate 18.6D and Plate 18.6A).

Three thin sections of the minor dykes are available (VO81-131D, VO81-137B, and VO81-138D). All contain plagioclase phenocrysts, skeletal in part. The groundmass mineral assemblage is plagioclase, olivine, clinopyroxene, opaque oxide, ± biotite and apatite. Thin section VO81-137B captures a chilled margin contact against coarse-grained Long Range gabbro.

18.2.1.5 Table Bay Dyke Corridor

The Table Bay dyke corridor is perhaps the least well known. A major dyke is exposed on the coast on the south shore of Table Bay (VO81-580). Its immediate extrapolation inland has not been clearly identified, although satellite imagery and aerial photographs do hint as to where it might be. In any case, most likely it links up with a major dyke that is exposed on Sand Hill River, 15 km south-southwest of the coast exposure. From its most northerly exposure on Sand Hill River, the dyke can be traced clearly on satellite aerial photographic imagery for 22 km farther southwest, trending at 015° and up to 160 m wide. No geochronological data are available for this dyke and it was not included in the study of Murthy *et al.* (1992).

Field notes for the outcrop on the south shore of Table Bay (not seen by the author) merely mention the presence of coarse-grained diabase, but there is no doubt, from the sample collected (VO81-580D), that it is from a major Long Range dyke. The stained slab shows it to contain common interstitial K-feldspar (but not abundant enough to be termed monzogabbro). In addition, field notes record the presence of several fine-grained, bronze-weathering, 0.3-m-wide dykes trending at 355°.

The major dyke exposed on Sand Hill River is brown- or creamy-weathering, coarse-grained, massive, unrecrystallized gabbro. At another outcrop on Sand Hill River (VO81-316), a fine-grained mafic satellite dyke was recorded, having a trend of 356°.

The primary essential minerals in a thin section of coarse-grained gabbro (CG85-170) are euhedral plagioclase; relict olivine; anhedral, brown clinopyroxene; hornblende, red-brown biotite and K-feldspar. Accessory minerals are an opaque oxide, sulphide, and apatite. Secondary white mica is also present. One minor dyke was also examined in thin section (VO81-580C). It is plagioclase-phyric and has skeletal plagioclase in the groundmass. Identified mafic minerals are pseudomorphed olivine, altered clinopyroxene, and amphibole.

K-feldspar is also present in the only other coarse-grained gabbro sample collected from this Long Range dyke (CG85-169).

18.2.1.6 Open Bay Dykes

It needs to be stated at the outset that considerable doubt exists as to the true affinity of the Open Bay dykes (Figure 18.2), although they were interpreted to belong to the Long Range swarm by Murthy *et al.* (1992; Dyke 6), and gave a comparable paleomagnetic signature to that seen in some of the other Long Range dykes in eastern Labrador.

Two dykes are present at the locality (CG85-576). One is 2 m wide and the other 0.6 m wide. Both are black-weathering, fine-grained diabase dykes having a typical ‘Long Range’ trend between 010° and 026°. No coarse-grained Long Range gabbro dykes are associated. Another similar dyke was found, on strike, 4.3 km to the north-northeast (GM85-614).

Despite trend and paleomagnetic signature, the dykes differ from typical Long Range dykes in that they are much more altered and also have been injected by quartz–epidote veins. The dyke at GM85-614 has been injected by a granitic vein.

Two thin sections are available; one is amygdaloidal (CG85-576A) and one is plagioclase-phyric (CG85-576B). In both, plagioclase is euhedral, relict primary and severely altered. No olivine is present in either, but one (CG85-576A) contains anhedral clinopyroxene pseudomorphs. Both oxide and sulphide opaque minerals are present and relict orange-brown biotite in CG86-576A. Secondary minerals include chlorite and carbonate.

Whole-rock chemical analyses are available for both rocks examined in thin section and show compositional differences compared to unequivocal Long Range dykes. Perhaps the claim could be made that the compositional differences are due to alteration, but the author’s conclusion is that they are not Long Range dykes; at least not ‘normal’ ones (although their alternate affinity remains to be determined).

18.2.1.7 Cooper Island Dyke Corridor

The Cooper Island dyke is addressed in three parts, from north to south. Clarification regarding two outcrops at Beaver Brook, which previously was considered to be a southern extrapolation of the same dyke (but is no longer), is also included.

Cooper Island to Gilbert Bay. This part of the Cooper Island dyke was first mapped by Wardle (1976; 1977), and has been traced for 33 km from Cooper Island in the north, to Gilbert Bay in the south. The name ‘Cooper Island dyke’ is newly applied here. No geochronological data are available for the dyke and it was not included in the paleomagnetic study of Murthy *et al.* (1992). It was sampled later by Murthy *et al.* at two localities (CG86-209, JS86-001). Paleomagnetic samples collected at JS86-001 were processed; the data is included in the Paleomagnetic table of the author's database.

The dyke was described as 200 m wide, slightly altered, and locally columnar jointed by Wardle (1977). At St. Michaels Bay, it was estimated it to be 100 m wide by the author (CG86-209). The dyke is brown-, orange-brown- or dark-green-weathering, massive, homogeneous, even-textured, medium- to coarse-grained, fresh to mildly altered olivine gabbro (or finer grained compositional equivalents). It lacks deformational fabrics, except at St. Michaels Bay where it has been transected by southeast-trending, chlorite-filled shears. The dyke has clearly defined chilled margins truncating foliation or gneissosity in the surrounding rocks. It generally lacks phenocrysts, apart from some larger-than-groundmass plagioclase in quench-textured chilled margins. Wardle's field notes, however, noted plagioclase phenocrysts 1 to 2 cm across in a 0.5–0.75 cm groundmass at RW75-401.

On the road between Charlottetown and Pinsent's Arm, two unmetamorphosed, amygdaloidal dykes are exposed (CG04-104). One is about 8 m wide, trending at 010°, and the other about 1 m wide, trending at 015° (Plate 18.6E). The 8-m-wide dyke shows altered, pyrite-enriched, 30-cm-wide borders, which have been recognized as a feature of the Long Range dykes elsewhere (Gower, 2010c).

A small dyke (about 2 m wide, and having a trend of 015°) mapped on the north side of Gilbert Bay (CG86-387) was included by Gower *et al.* (1987) as an attenuated southward extension of the Cooper Island dyke. Despite it being aligned with the Cooper Island dyke and having the same trend, its correlation was initially doubted because the intrusion was unusually narrow, compared to the rest of the dyke. The narrowness of the unequivocal Long Range dykes exposed on the road to the north implies that the dyke tapers southward.

Of the five thin sections prepared from this segment of the Long Range dyke, only one is massive, medium- to coarse-grained gabbro (CG86-209), which, nevertheless, is the typical rock type seen in the field. Its primary mineral assemblage consists of euhedral, moderately to strongly zoned plagioclase; anhedral, brown clinopyroxene; relict olivine; orange or red-brown biotite; pale-green amphibole; skeletal opaque oxide and sulphide; and skeletal apatite. Interstitial K-feldspar is present, as it is in other samples of gabbro collected from this part of the dyke (stained, but not thin sectioned samples from CG86-608, MN86-375, SN86-318C).

One of the other four thin sections is from a chilled margin (CG98-446), and the remainder are from small dykes (CG86-387B, CG04-104C, CG04-104D). They are too fine grained to distinguish their complete mineral assemblage, but primary, euhedral to subhedral, well-twinned plagioclase is evident, forming small phenocrysts in a quenched matrix. It is clearly skeletal in CG86-387B. Clinopyroxene and serpentinized olivine also form small phenocrysts. In the matrix, chlorite, carbonate, biotite, skeletal opaque oxides can be identified, in addition to plagioclase, clinopyroxene and, possibly, olivine. Much of the matrix in CG86-446 is probably devitrified glass.

Gower *et al.* (1987) mentioned that two dykes mapped on the south side of St. Michaels Bay (JS86-380, and MN86-349/RW75-256) might be Long Range satellite dykes. Following petrographic studies, these are both now excluded; they are metamorphosed and either Gilbert Bay or Labradorian dykes. The assigning of another locality (actually three closely spaced data stations; SN86-207, SN86-208, SN86-219) to the Long Range swarm on the preliminary 1:100 000-scale map for the Port Hope Simpson region is also discredited. Although the rocks superficially resemble the coarser grained Long Range gabbro, petrographic study of SN86-207 is persuasive that the rocks are unrelated, although their affinity remains unclear (provisionally now assigned as M_{3A}mn – Section 17.1.2). Another example of a rock that superficially resembles a Long Range dyke in the area was collected at JS86-388. A Long Range dyke was recorded in the field at the locality, which is entirely reasonable, given confirmed outcrops on strike to the north and south. A stained slab of the sample shows that the gabbro is coronitic, which Long Range dykes never are. The sample is undoubtedly from Labradorian gabbronoritic host rocks to the Long Range dyke, rather than the dyke itself.

Gilbert Bay to Alexis Bay. The Cooper Island dyke was not located between Gilbert Bay and Alexis River, but it might be very narrow and escaped notice. Its potential location is not evident from either aerial photographs or aeromagnetic data. A supposed Long Range dyke depicted on the south shore of western Gilbert Bay by Gower *et al.* (1987; their Figure 3) relied on field notes that mention a 'rusty-brown weathering, medium-grained, massive gabbro' at the locality (MN86-316). On the basis of field description, which noted pegmatitic veins through the gabbro, coupled with the textural appearance of a stained slab, this outcrop now has been reassigned as Labradorian. The unit designator utilized (P_{3C}rg/Nd?) leaves open both options, however. If there is a 2.5-km-dextral offset along the Gilbert River fault (as proposed by Gower *et al.*, 1987), this is certainly where the Long Range dyke would be expected. Re-examination of the outcrop is needed.

Alexis Bay to St. Lewis Inlet. A Long Range dyke, regionally on strike with the Cooper Island dyke, was mapped on the south side of Alexis River, east of Port Hope Simpson by Gower *et al.* (1987). The dyke was not seen on the north side of Alexis Bay. The dyke is brown-weathering, massive, coarse-grained gabbro, estimated to be about 50 m wide at sites JS86-001/CG86-784. It has been traced 13 km to the St. Lewis River, where it is about 10 m wide (Gower *et al.*, 1988). North of St. Lewis River, it is interpreted to be offset dextrally across a small fault.

Five thin sections are available for this segment of the Long Range dyke (CG86-784, JS86-001, JS86-005, JS87-425, VN87-057; the first two are from the same locality). The rocks are coarse-grained diabase/gabbro. Primary minerals are plagioclase, clinopyroxene, lesser olivine, and accessory phases. Plagioclase is euhedral to subhedral, and moderately to strongly zoned. Clinopyroxene is pale-brown, subhedral to anhedral. Olivine occurs as subhedral to rounded, serpentinized pseudomorphs. Both skeletal opaque Fe-Ti oxide and sulphide opaque minerals are present, although oxide dominates. Other minerals include red-brown biotite (commonly mantling opaque minerals), very acicular and locally skeletal apatite, and zircon. Interstitial and symplectically intergrown quartz and K-feldspar are also present. Feldspars and melanocratic silicates both show extensive alteration to greenschist-facies and possibly sub-greenschist-facies minerals, which include colourless to pale-green amphibole, chlorite, secondary titanite, prehnite, and equivocally identified pumpellyite.

Beaver Brook. Two outcrops of mafic rock on Upper Beaver Brook (Pinware map region) and previously recorded on the map of Bostock (1983), were interpreted by Gower *et al.* (1994) as exposures of a Long Range dyke, and they were depicted as such on the 1:100 000-scale map for the region. This interpretation is based on it having a typical north-northeast Long Range dyke trend and being aligned with the Cooper Island Long Range dyke. On the other hand, there is little evidence from aeromagnetic or topographic lineaments that the Cooper Island dyke continues southwest of the St. Lewis River. An alternative possibility is that the dyke is older, possibly early posttectonic with respect to Grenvillian orogenesis. A sample from an outcrop on Beaver Brook examined in thin section (VN93-718A) is more altered than is typical for Long Range dykes (*cf.* Section 17.4.3.1).

18.2.1.8 Double Island (Southeast of Battle Island) and Truck Island

Several unmetamorphosed or slightly metamorphosed, planar, parallel, near-vertical dykes between 0.5 and 5 m wide intrude gneiss on Double Island, southeast of Battle Island (CG87-661 to CG87-664). A 3-m-wide, texturally similar, vertical dyke (CG87-665) occurs on Truck Island, 20 km to the south-southwest. The dyke at locality CG87-662 is deformed into open folds around a fold axis having an azimuth/plunge of 050°/42° and segmented by apparent sinistral faults having a strike/dip attitude of 260°/60°. An anomalous shallowly eastward-dipping (measured to be 355°/30° and 010°/35°) amygdaloidal dyke (CG87-561), about 10–15 m thick, is also present on Double Island. Another amygdaloidal dyke occurs on the shoreline 7.6 km southwest of Truck Island (JS87-398B).

Despite being reviewed here and given the unit designator Nd by Gower (2010a), there are good reasons to doubt that these dykes belong to the Long Range suite, although they have a typical Long Range trend – that is to say, between 005° and 030°. As this is also the orientation of a line linking Double Island to Truck Island, it is surmised that the dykes at the two localities belong to the same suite. These dykes differ considerably in habit compared to Long Range dykes located elsewhere in southeast Labrador, especially in being small, closely spaced and deformed. There is no *a priori* reason to exclude small dykes from the Long Range swarm, however because: i) along with major dykes in southeast Labrador, sporadic minor dykes are present, and ii) the Long Range dykes intruding the Long Range Inlier are also smaller, more numerous, and more deformed and metamorphosed compared to those in southeast Labrador (and confirmed by geochronological data to be temporally correlative). An important consideration is that evidence from supracrustal rocks indicates that the present coastline also coincides with a change in late Neoproterozoic tectonic setting from stable crust in the west, to much less stable immediately offshore.

The dykes are petrographically distinct from other Long Range dykes in southeast Labrador, although forming a coherent group themselves (CG87-561A.1, CG87-561A.2, CG87-661, CG87-662, CG87-663, CG87-664, CG87-665). Especially characteristic is hollow-tube, quenched euhedral plagioclase. Clinopyroxene, in ophitic relationship to plagioclase, occurs as brown, anhedral grains. The only other essential mineral is ragged, skeletal opaque oxide. Orange-green interstitial material, including some biotite, chlorite and carbonate, might represent devitrified glass. Samples CG87-561A.1 and CG87-561A.2, from the same shallowly dipping dyke, have rounded amygdules up to 1 mm in diameter. Lack of olivine and lack of distinct zoning in plagioclase are major differences compared to well-established Long Range dykes in southeast Labrador.

18.2.1.9 Distribution of K-feldspar in Long Range Dykes

The variation from gabbro to monzogabbro, and even syenogabbro, has been noted above for various dykes. Less clear from the above descriptions is the remarkably systematic pattern in the distribution of K-feldspar shown collectively for all Long Range dykes in southeast Labrador (excluding those rejected above as probably not belonging to the suite). This pattern is illustrated in Figure 18.3, which shows that, northwest of a 060°-trending line passing south of Paradise River, the Long Range dykes lack K-feldspar, whereas, southeast of the line, K-feldspar is present. The instances in Figure 18.3 where the presence of K-feldspar has been indicated as uncertain mostly apply to fine-grained dykes in which staining of slabs for K-feldspar identification was less effective.

Although the empirical pattern is compelling, the author is at a loss to propose a cause. The dividing line is

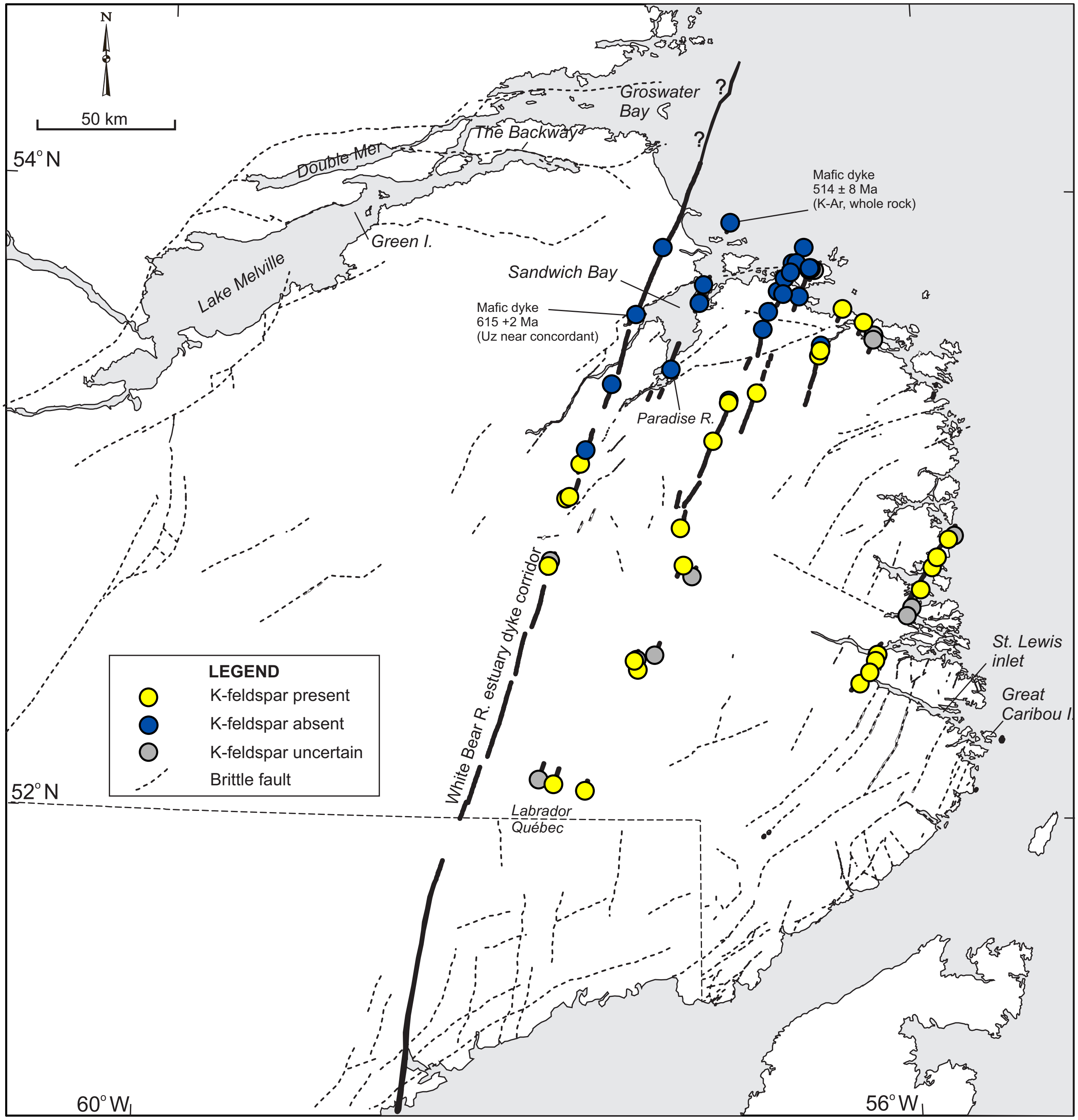


Figure 18.3. *Distribution of K-feldspar in Long Range dykes.*

neither related to Grenvillian or earlier structures, nor has apparent ties to Bouguer anomalies or other geophysical parameters. It is, however, parallel to the Double Mer and Sandwich Bay grabens, which, like the Long Range dykes, are products of Iapetan rifting. It would seem likely that either the deep crust or underlying mantle is involved. A 15–20-km-thick high-velocity lower crustal (HVLC) wedge has been interpreted to exist under southeast Labrador by Funck *et al.* (2001), and is attributed to underplating during Iapetan rifting. The K-feldspar dividing line identified here has no obvious spatial relationship with the outline of the HVLC wedge depicted by Funck *et al.* (2001), but the extent of the wedge was acknowledged by them to be very uncertain.

18.2.1.10 Lithogeochemistry of Long Range Dykes

Some lithogeochemical features of the Long Range dykes are shown in Figure 18.4. The samples show considerable scatter, which is unsurprising given that all have been lumped together, regardless of which part of the dyke they are from, for example, coarse-grained centres of thick dykes, or from fine-grained satellite dykes.

The most compelling feature to the author is that the Double Island–Truck Island corridor samples have a less fractionated character and are clearly distinct from the rest of the Long Range dyke samples from eastern Labrador, thus reinforcing comments made in Section 18.2.1.8 regarding the doubtful affiliation of these rocks.

The analyzed samples are depicted according to the dyke corridor in which they occur, the red end of the spectrum being the westernmost-dyke samples and *vice versa* for the east.

Apart from some anomalous Curlew Harbour and Table Bay–Open Bay high-Ti samples, there is some indication of compositional commonality within samples from specific dyke corridors and, despite the considerable scatter of points, a suggestion that the western dykes are more fractionated than those in the east.

18.2.2 QUARTZ VEINS (Nq)

In addition to the giant quartz veins described below, minor quartz veins were recorded throughout the region during 1:100 000-scale mapping. It is probable that many were formed at the same time as the giant quartz veins, but in the absence of any means to demonstrate this, the veins have simply been designated as "q" on the 1:100 000-scale maps of Gower (2010a). The major occurrences are indicated on Figure 18.2.

18.2.2.1 St. Lewis River Area

Three, large, parallel quartz veins (up to 400 m wide and, discontinuously, up to 20 km long) transect the area southwest of St. Lewis Inlet (Figure 18.2). They were only briefly examined during the 1:100 000-scale mapping by Gower *et al.* (1988), but have been investigated in more

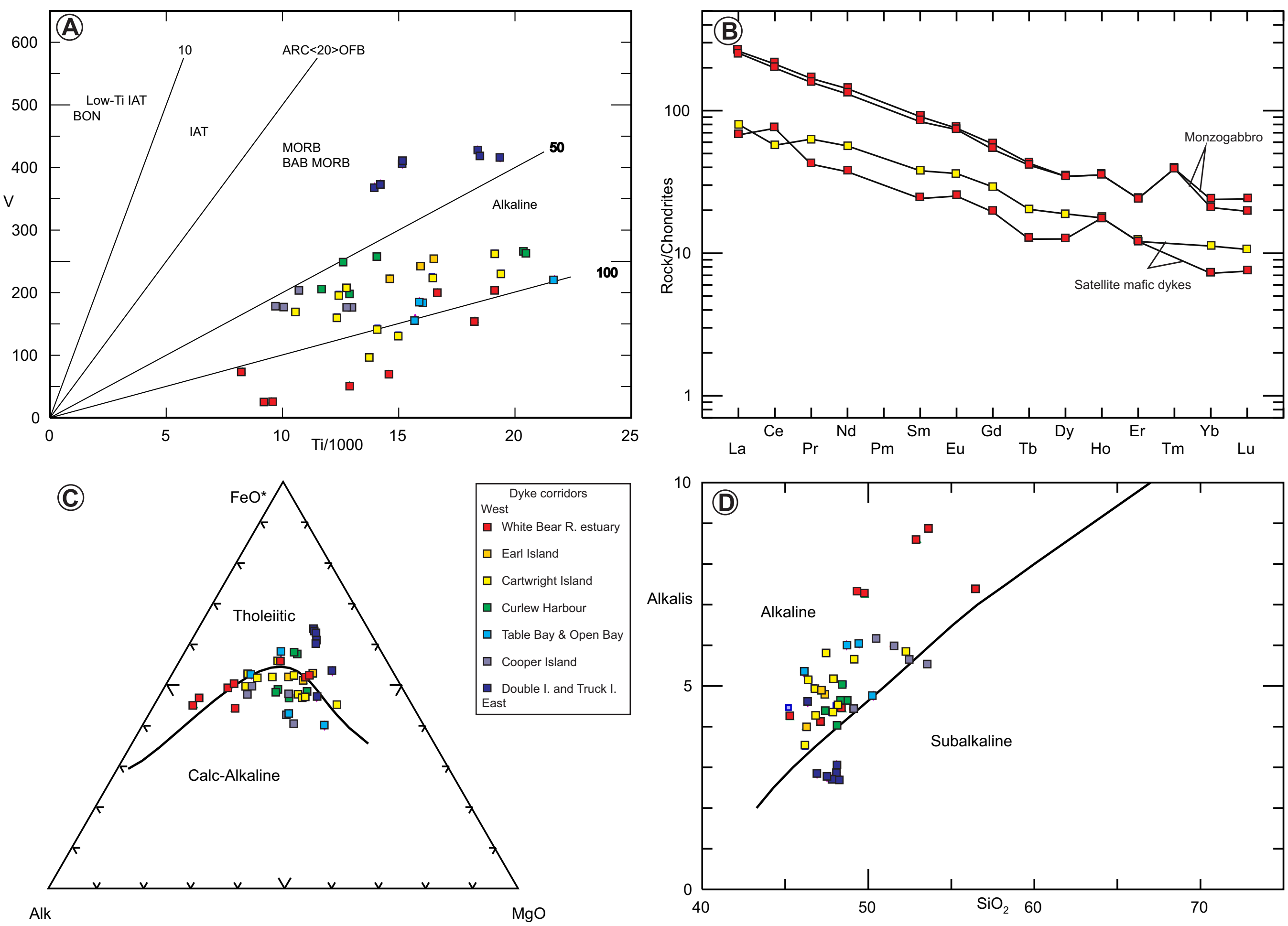


Figure 18.4. *Lithogeochemical characteristics of the Long Range dykes.*

detail by others. They were first recorded by Piloski (1955), subsequently mapped by Eade (1962), investigated by Brinco in 1982 and re-examined by Meyer and Dean (1986). Many of the quartz veins are related to brittle faulting and, because of a common trend, can be linked to the same fracture system as that utilized by the Long Range dykes. There is little doubt that these quartz veins are related to Iapetan rifting.

Most localities in the St. Lewis River area designated as Unit Nq are based on Eade's (1962) mapping (EA61-034, EA61-036, M61-090, M61-099b, R61-096), with the addition of some sites recorded during ground traversing during the 1987 mapping (CC87-100, JS87-107). Eade described the rock at locality EA61-034 as a quartz-feldspar pegmatite, typically heavily stained by hematite and showing some rusty patches. It is characterized by abundant fractures, shears and slickenside surfaces and is clearly related to a zone of major brittle deformation. At locality R61-096, pure, white quartz veins are described as merging northward into one huge quartz vein, stained reddish by hematite. Inclusions of the country-rock gneiss are contained within the quartz veins.

The quartz vein at M61-099b is reported to be 20' wide and trending 055°. This occurrence is not indicated on Eade's map, the information having been gleaned from notes on the aerial photograph that was used by Eade's assistant (GSC unpublished archives). Recalling that quartzite occurs at M61-099a (*cf.* Section 13.3.3.4), which is only 850 m to the north-northwest one might be suspicious that either the quartz vein or the quartzite has been misidentified. Both observations are valid, however. The identity of the quartzite is substantiated by observations during the 1987 mapping and examination of two thin sections (VN87-275, EA61-717; *cf.* Section 13.3.3.4). The trend of the quartz vein is somewhat atypical and there are no obvious associated topographic lineaments. Interestingly, 5.3 km from the 055°-trending quartz vein, on a bearing of 060°, a fault-brecciated quartz vein was described at data-station CC87-100 (note not CG87-100).

Four thin sections from this locality (CC87-100A, -100B, -100C, -100D) comprise angular fragments of quartz having a wide range in size and criss-crossed by veins of polygonal quartz in all directions (Plate 18.7A). Other felsic minerals obvious in the thin sections are plagioclase and poorly twinned microcline, both more apparent in the stained slab than in thin section. Minor biotite and an opaque mineral are also present. Locality CC87-100 is also on a strong north-northeast-trending lineament that coincides elsewhere with fault breccia observed during ground traverses, so it is possible that the alignment of the trend of the quartz vein at M61-099b with CC87-100 is fortuitous (or the alignment with the north-northeast-trending lineament; or that both are significant and related).

Meyer and Dean (1986) described the rocks as quartz and quartz-feldspar veins. They note that the veins display evidence of multiple intrusion and brecciation, having breccia fragments that range from large and angular to small and subrounded. Other minerals noted are minor hematite, magnetite, epidote and bright orange K-feldspar speculated to be adularia. Meyer and Dean's interest in the quartz veins was for their potential as a silica resource. Further details of their economic potential and evaluation by industry are given by Gower (2010c).

Separate from the occurrences described above, vein quartz was recorded on the middle part of the St. Lewis River (JS87-107). The orientation of the quartz vein was not recorded, but topography suggests either a north-northeast-trending feature, or, on a more regional scale, on an east-northeast-trending lineament defined by the southern branch of the St. Lewis River. Farther to the southwest, this was interpreted as a brittle fault by Gower *et al.* (1993). The rock exhibits an unusual breccia texture in thin section that is suggestive of a diatreme. Although the mineral assemblage and stained-slab appearance point toward a psammitic metasediment (*cf.* Section 13.1.3.3), quartz and feldspar grains are fractured, and very small, angular fragments of the same material infill the fractures.

18.2.2.2 Upper Alexis River

An interesting example of vein quartz is located in the Upper Alexis River area (Plate 18.7B) and is well documented by field description, field photographs and stained slabs. The locality is within about 200 m of a major Long Range dyke (unequivocally confirmed from its stained slab). The vein quartz consists of angular, brecciated blocks of randomly oriented quartz veins and stringers. Alteration includes sericitization, chloritization, silicification and hematization. Three stained slabs show considerable variation in K-feldspar content (0–50%), and common euhedral pyrite.

Neither the contact of the quartz vein nor the Long Range dyke is mentioned in field notes as being exposed, so, given that the Long Range dyke here is a K-feldspar-bearing, coarse-grained gabbro that is typical of the dykes that are 10's or even 100's of metres wide, it is possible that the quartz veins and dyke are in contact. This, in turn, may suggest genetic linkage between the two, thus providing another example of the close relationship between brittle-faulting, vein quartz formation and Long Range dyke emplacement, as indicated by: i) north-northeast-trending veins in the St. Lewis River area, and ii) extremely xenolithic dykes at data-station R61-146 in (what may be) the White Bear River estuary dyke, and at VO81-131 in the Curlew Harbour dyke.

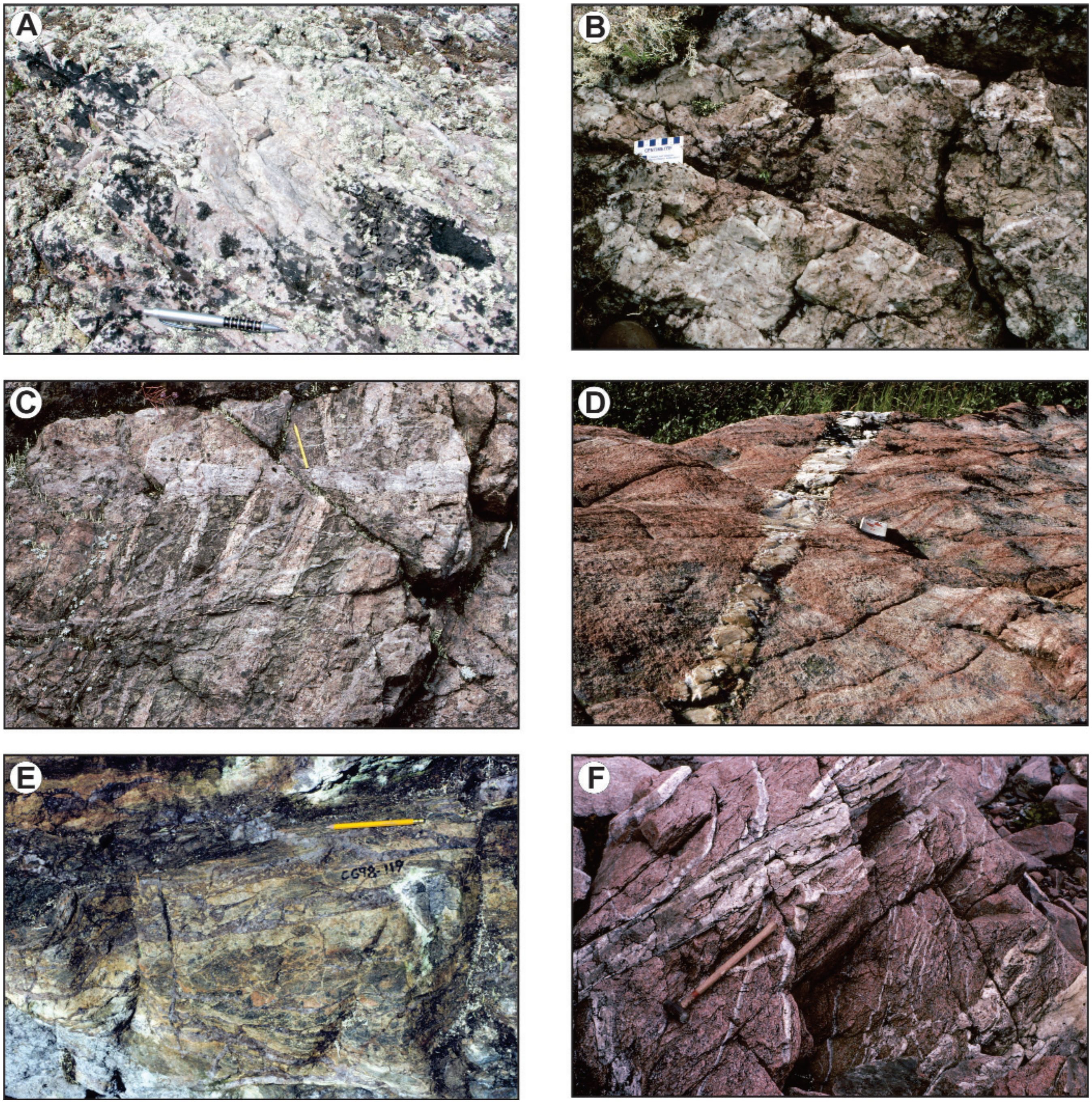


Plate 18.7. *Examples of quartz veins and associated fault breccciation. A. Fault-brecciated quartz vein (CC87-100), B. Brecciated quartz vein (DD91-080). Long Range dyke nearby – outside of image frame (*see *text), C. Quartz vein showing brecciation and multiple injection (CG87-213), D. Simple quartz vein associated with altered granite (CG84-309), E. Fault breccia showing silicification, epidotization, chloritization and hematization (CG98-119), F. Multiple quartz veining associated with altered granite (CG85-652).*

18.2.2.3 Upper Paradise River Area

In the Upper Paradise River area, about 11 km southeast of the intersection of the White Bear River estuary Long Range dyke with the highway (Figure 18.2), two major quartz veins were reported by van Nostrand (1992). Some uncertainty exists regarding their orientation; van Nostrand (1992) reports a trend of 060° for both, but his field notes record 026° and 050°. They are indistinct on aerial photographs, but a trend of about 045° is indicated. The quartz veins are described as 4- to 5-m-thick, white-weathering, fine to very fine grained, associated with fault breccia, and with alteration zones containing chlorite, epidote and muscovite (VN91-404, VN91-406, DD91-132). The veins are interpreted by van Nostrand (1992) to be related to faulting along the southeast side of the Sandwich Bay graben.

Another sizable outcrop of vein quartz is situated about 14 km west-southwest of the southernmost quartz vein mentioned above (at DE91-155, DE91-156). Stained slabs show it to be strongly deformed, which, despite being included in this section, may imply that it is older and unrelated.

Examples of other quartz veins and quartz vein breccia in the area are shown in Plate 18.7C, D.

18.2.2.4 No-Name Lake Area

East of No-Name Lake, a silicified fault breccia at one outcrop (CG98-119) plus two outcrops of vein quartz (CG98-120, CG98-121), form hills having a common northeast-trending (050°) alignment. The fault breccia comprises green-, white- and red-weathering, fine grained, sheared and brecciated rock that has been extensively silicified, hematized, chloritized and epidotized (Plate 18.7E). Deformation and alteration have rendered the original protolith unrecognizable. The fault breccia is at least 50 m wide and has been extensively injected (in an irregular, anastomosing manner) by quartz veins, ranging from 1 mm to 3 cm wide.

In a thin section (CG98-119), recognized minerals are: i) sericitized and saussuritized plagioclase, ii) K-feldspar associated with quartz and hematite in cavities, iii) granular and euhedral epidote, iv) titanite, some grains of which are abnormally large and associated with euhedral epidote, v), fibrous, colourless to green tremolite–actinolite, vi) opaque minerals that include both oxide and sulphide, vii) quartz, in veins and cavities, and viii) chlorite.

The two other outcrops, to the southwest, differ in that they consist almost entirely of vein quartz, rather than fault breccia. At data station CG98-120, the vein quartz occupies a northeast-trending linear zone at least 100 m wide and 0.5 km long that is easily seen on satellite and aerial photographic imagery. The vein quartz is mostly massive, although some shear zones are present.

The outcrops are interpreted to be related to a single brittle fault that is possibly a southwest extension of the northwest side of the Sandwich Bay graben.

18.2.2.5 Island of Ponds

Extremely fine-grained, strongly hematized and recrystallized fault breccia with later quartz veins was found on the northwest coast on the Island of Ponds (VN85-619), and fault breccia with abundant quartz veins recorded on Eagle Island, 4 km farther west (Plate 18.7F). The latter locality is one of the few in southeast Labrador where transparent euhedral quartz crystals were found during mapping.

A line linking the two outcrops is parallel to the prevailing brittle faulting trend in that area, but, apart from that, there is no compelling reason to link them conceptually. Also, despite being designated Unit Nq, there is no evidence that they are Neoproterozoic, other than weak deformation.

18.2.3 CARBONATE VEINS (Nc)

Apart from one isolated example west of St. Lewis Inlet (a quartz–calcite vein at VN87-093, situated on an interpreted north-northeast-trending brittle fault; Figure 18.2), carbonate veins are spatially associated with either the Lake Melville rift system or the Sandwich Bay graben. There seems little doubt that they developed during the rifting events. No photographs are available in the author’s collection.

18.2.3.1 Lake Melville Rift System

Three carbonate vein occurrences were recorded during Stevenson’s 1:250 000-scale mapping in the Rigolet–Groswater Bay region. Two are located on the north side of The Backway and associated with an east–west brittle fault. One is noted in field notes as ‘blobs’ and suggested to be remnants of a metamorphosed, segmented vein (SGJ69-068). Field notes for the other site note that the carbonate vein cuts earlier fabrics (SGJ68-069). The third one is associated with a brittle fault on the north side of the Lake Melville rift system and recorded to be 0.5” (1.25 cm) thick (SGJ68-002).

On the east side of Etagaulet Bay (south side of Lake Melville), Emslie (1975 field notes) recorded several calcite veins ‘up to 1.2” wide’ at EC75-195, and ‘some over 1’ thick, strike 150°/90°’ at EC75-260. Emslie also reported clastic dykes from the same area (EC75-235).

18.2.3.2 Sandwich Bay Graben

All the carbonate veins known in the Sandwich Bay graben are situated at the southern end of Sandwich Bay, on its western side. The first observation of carbonate veins in the Sandwich Bay graben was by Cherry (field notes, MC77-091), who merely commented that they postdate pegmatite injection at the site. During 1:100 000-scale mapping, four other occurrences were discovered (MW84-091, 093, 094, 095). The veins trend between 175° and 190° and are vertical or dip steeply to the west. Thickness of veins was not recorded. The five occurrences are aligned roughly north–south, more-or-less the same as the average of their individual trends, and parallel to the Earl Island Long Range dyke, situated 2 km to the east. Carbonate veins in the same area that do not line up with the above occurrences were also found at VN84-002, close to the Earl Island Long Range dyke. A fault breccia 11 km west of the community of Paradise River veined by carbonate (thin section VN84-264) is probably part of the same brittle fault system.