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FIELD TRIP GUIDEBOOK

TECTONIC ASSEMBLY OF THE PALEOPROTEROZOIC FLIN FLON BELT AND SETTING OF VMS DEPOSITS

(FIELD TRIP B1)

by

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INTRODUCTION

Deciphering the interrelations between mafic and felsic magmatism, deformation and sedimentation in Precambrian greenstone belts is critical to unraveling how continental crust and its endowment of mineral deposits were generated and preserved. Detailed mapping and research on greenstone belts has shown them to be complex amalgamations of volcanic, sedimentary and plutonic rocks of diverse ages and origins that are variably deformed and metamorphosed. There is no such thing as a 'type' greenstone belt or greenstones, nor are they all formed in the same tectonic environments (*cf.* de Wit and Ashwal, 1996). Perhaps the most important step towards resolving the origin and tectonic significance of greenstone belts is to stop considering them as coherent single entities, but instead to focus on their diversity and complexity in terms of tectonostratigraphy, geochronology, deformation and metamorphism.

Through an integrated program of geological mapping and laboratory research, the NATMAP Shield Margin Project has shown that the Flin Flon 'greenstone' belt (Fig. 1) comprises 1.92-1.88 Ga tectonostratigraphic assemblages derived from a variety of tectonic environments and amalgamated to form an accretionary collage prior to the emplacement of 1.87-1.83 Ga granitoid plutons (Lucas *et al.*, 1996). The plutons and coeval volcanic rocks are associated with younger arc(s) imposed on the collage, resulting in the development of a microcontinent by 1.85-1.84 Ga that probably included the adjacent Hanson Lake Block and Glennie Domain (Fig. 1; *cf.* Ansdell *et al.*, 1995). Seismic reflection profiling and research associated with the LITHOPROBE Trans-Hudson Orogen Transect have shown that subsequent intracontinental deformation (1.83-1.69 Ga) related to collision with Archean cratons and post-collisional tectonics only served to dismember this microcontinent (Lucas *et al.*, 1993; Lewry *et al.*, 1994; White *et al.*, 1994).

This field trip will focus on the geology, tectonic history and mineral deposits of the Flin Flon belt (Fig. 2). It has been designed to build the history of the Amisk collage and Snow Lake arc assemblage incrementally, so that the complex interrelations between magmatism, sedimentation, deformation and metamorphism can be understood in the framework of the new predictive tectonic model. We hope to (1) demonstrate that the Amisk collage was assembled relatively early in the evolution of the Flin Flon belt, (2) document the relations between the Amisk collage and the Snow Lake arc assemblage, and (3) illustrate both the local geologic setting and tectonic context of the VMS deposits.

Overview of the Flin Flon belt

Historically, the stratigraphy of the Flin Flon belt has been described in terms of two stratigraphic groups, Amisk Group volcanic rocks and Missi Group continental sedimentary rocks (Bruce, 1918; Harrison, 1951). The Flin Flon belt in the area between the Sturgeon-weir River (Saskatchewan) and Reed Lake (Manitoba) is now recognized as a collage (*Amisk collage*) of distinct tectonostratigraphic assemblages that was assembled prior to the emplacement of voluminous granitoid plutons and regional deformation related to the *ca.* 1.8 Ga Hudsonian orogeny (Lucas *et al.*, 1996). Each tectonostratigraphic assemblage is a distinct package of rocks in terms of its

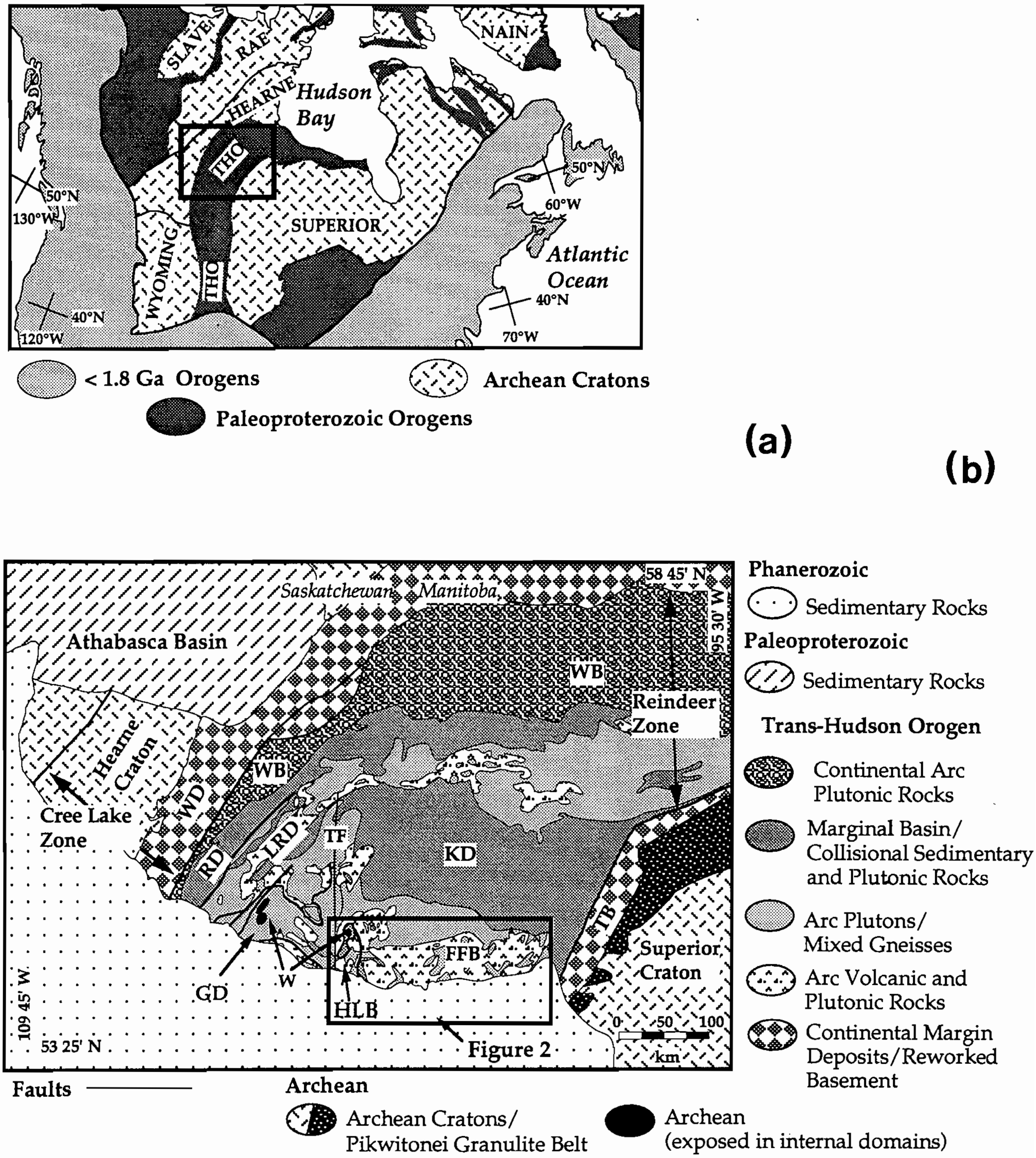


Figure 1: (a) Location of the Trans-Hudson Orogen (THO) and other >1.8 Ga Proterozoic orogens relative to the Archean provinces of continental North America, Greenland and western Scandinavia (pre-Atlantic reconstruction), after Hoffman (1989). Autochthonous platform cover is not shown. Location of Figure 1b is outlined in heavy box. (b) Map of the Trans-Hudson Orogen, after Hoffman (1988). "W" indicates location of Archean basement windows in the Reindeer Zone. FFB: Flin Flon belt; GD: Glennie Domain; HLB: Hanson Lake Block; LRD: La Ronge belt; KD: Kisseynew Domain; RD: Rottenstone Domain; TB: Thompson belt; TF: Tabbernor Fault Zone; WB: Wathaman-Chipewyan Batholith; WD: Wollaston Domain. From Lucas et al. (1996).

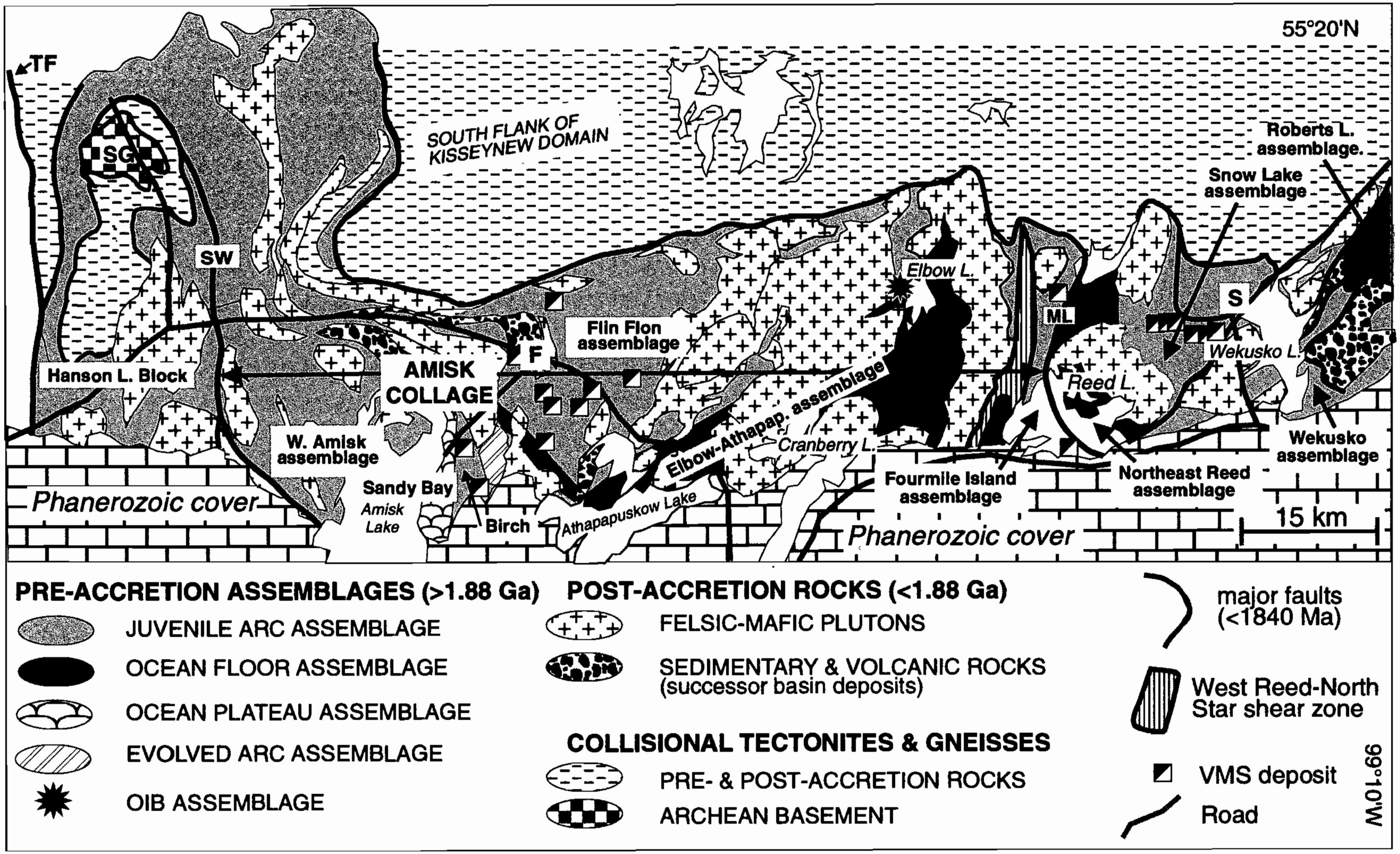


Figure 2: Map of the Flin Flon Belt showing the extent of the Amisk collage, major tectonostratigraphic assemblages and plutons, and location of mined VMS deposits. ML: Morton Lake fault; SW: Sturgeon-weir shear zone. F: Flin Flon; S: Snow Lake. Modified from Lucas et al. (1996).

stratigraphy, geochemistry, isotopic signature, age and inferred plate tectonic setting (see below; Lucas *et al.*, 1996). 'Tectonostratigraphic assemblage' is not necessarily equated with 'terrane' nor is it implied that each assemblage is a fragment of a unique plate. The basis for rejecting Amisk Group as the means of describing the 1.92-1.88 Ga volcano-plutonic rocks is that they do not form a stratigraphic 'group' in any sense of the word. The Amisk collage is bound by collisional thrusts (*ca.* 1.83-1.80 Ga) on its west (Sturgeon-weir shear zone; Ashton and Lewry, 1994), north (south flank of the Kisseynew Domain; Zwanzig, 1990; Lucas *et al.*, 1994; Norman *et al.*, 1995) and east sides (Morton Lake fault; Syme *et al.*, 1995), but extends beneath Phanerozoic platform cover to the south (Leclair *et al.*, 1993). The Flin Flon belt comprises the Amisk collage as well as the Snow Lake assemblage and Hanson Lake Block. The relations between the Flin Flon and Snow Lake VMS camps are discussed below.

Four main tectonostratigraphic assemblage types are recognized (Syme, 1990; Syme and Bailes, 1993; Watters *et al.*, 1994; Stern *et al.*, 1995a, 1995b): isotopically-juvenile oceanic arc (1.90-1.88 Ga), ocean floor (*ca.* 1.90 Ga), oceanic plateau/ocean island, and isotopically-evolved arc (1.92-1.90 Ga). Early (*ca.* 1.90 Ga) tholeiitic arc sequences were rifted, creating intra-arc basins in which oceanic basalts, turbidites and calc-alkaline and shoshonitic volcaniclastics accumulated at 1.89-1.88 Ga (Syme and Bailes, 1993; Reilly *et al.*, 1995; Stern *et al.*, 1995a). The tectonostratigraphic assemblages were juxtaposed in an accretionary complex at *ca.* 1.88-1.87 Ga, probably as a result of arc-arc collision(s) (D_1, Table 1; Fig. 3; Lucas *et al.*, 1996). Accretionary collage structures are largely obliterated by subsequent deformation and metamorphic events (D_2-D_5, Table 1), but can be inferred where cut by calc-alkaline plutons related to a 1.866-1.838 Ga arc. Coeval subaerial volcanism is recorded in *ca.* 1.87-1.85 Ga calc-alkaline to shoshonitic volcaniclastic sequences (Syme, 1988; Lucas *et al.*, 1996; Stern *et al.*, in prep.). Unroofing of the accretionary collage, development of a paleosol and deposition of alluvial-fluvial sedimentary rocks (Missi suite; Bailes and Syme, 1989; Holland *et al.*, 1989) occurred *ca.* 1.85-1.84 Ga (Ansdell *et al.*, 1992; Ansdell, 1993), coeval with the waning stages of post-accretion arc magmatism (Fig. 3; Stern and Lucas, 1994; Whalen and Hunt, 1994; Lucas *et al.*, 1996). Development of the Kisseynew turbidite basin (Kisseynew Domain, Fig. 1) was synchronous with continental sedimentation in the Flin Flon belt (David *et al.*, 1993, 1996; Machado and Zwanzig, 1995).

The transition from Kisseynew basin extension to collisional collapse occurred rapidly at about 1.840 Ga, although sedimentation and magmatism continued through to ~1.830 Ga (Fig. 3; Ansdell and Norman, 1995; Ansdell *et al.*, 1995; David *et al.*, 1996). The 1.84-1.83 Ga rocks, the Kisseynew basin turbidites (Burntwood suite; Zwanzig *et al.*, 1995) and their basement (Amisk collage, Snow Lake assemblage) now structurally overlie the Amisk collage along the south flank of the Kisseynew Domain (Fig. 1; Harrison, 1951; Zwanzig, 1990; Lucas *et al.*, 1994; Norman *et al.*, 1995; Connors, 1996). Following collisional thickening and peak metamorphism at 1.83-1.80 Ga, the Flin Flon belt experienced protracted intracontinental deformation to *ca.* 1.69 Ga (see below).

Table 1. *Deformation episodes in the Amisk collage.*

Episode	Structures	Magmatism	Metamorphism	Age (Ma)	Tectonic context
D1	S1, L1 (tectonites, mylonites)	None (?)	?	1880-1870	intraoceanic accretion
D2	S2, F2 (Vick Lake synform) S1/S2 (tectonites, mylonites)	Mafic to felsic dykes, sheets, plutons	Subgreenschist to amphibolite ('contact' to regional)	1870-1840	Intra-arc shortening, uplift/erosion of Amisk collage; development of Kisseynew back-arc basin
D3	S3, F3; shear bands (sinistral and dextral); high-angle shear zones; SW-vergent thrusts (e.g., Morton Lake thrust zone)	None (1840-1830 magmatic belt in Snow Lake assemblage, S flank of Kisseynew Domain	Regional peak metamorphism (subgreenschist to amphibolite facies) at 1820-1805 Ma	1840-1805	Regional collisional shortening and thickening via SW-vergent thrusts and folds; high-angle shear zones in Amisk collage
D4	SW-vergent thrusts (e.g., Sturgeon-weir shear zone); F4 folds and kinks, high-angle shear zones	Pegmatites, leucogranites	Retrograde	1805-1770 (?)	Post-collisional thrusting, folding; transpression of Amisk collage
D5	Brittle to ductile shear zones/faults	None	Retrograde	1770-1690	Post-collisional NW-SE shortening and longitudinal extension

References: Ansdell and Connors (1994), Ansdell *et al.* (1995), Ansdell and Norman (1995), Ashton (1992), Ashton *et al.* (1992), Bailes and Syme (1989), David *et al.* (1993, 1995), Digel and Gordon (1993, 1995), Fedorowich *et al.* (1995), Gordon *et al.* (1990), Heaman *et al.* (1992, 1993), Lucas *et al.* (1994), Reilly *et al.* (1993, 1994), Ryan and Williams (1993, 1994, 1995), Stauffer (1990), Stauffer and Mukherjee (1971), Stern and Lucas (1994), Syme (1994, 1995), Thomas (1992).

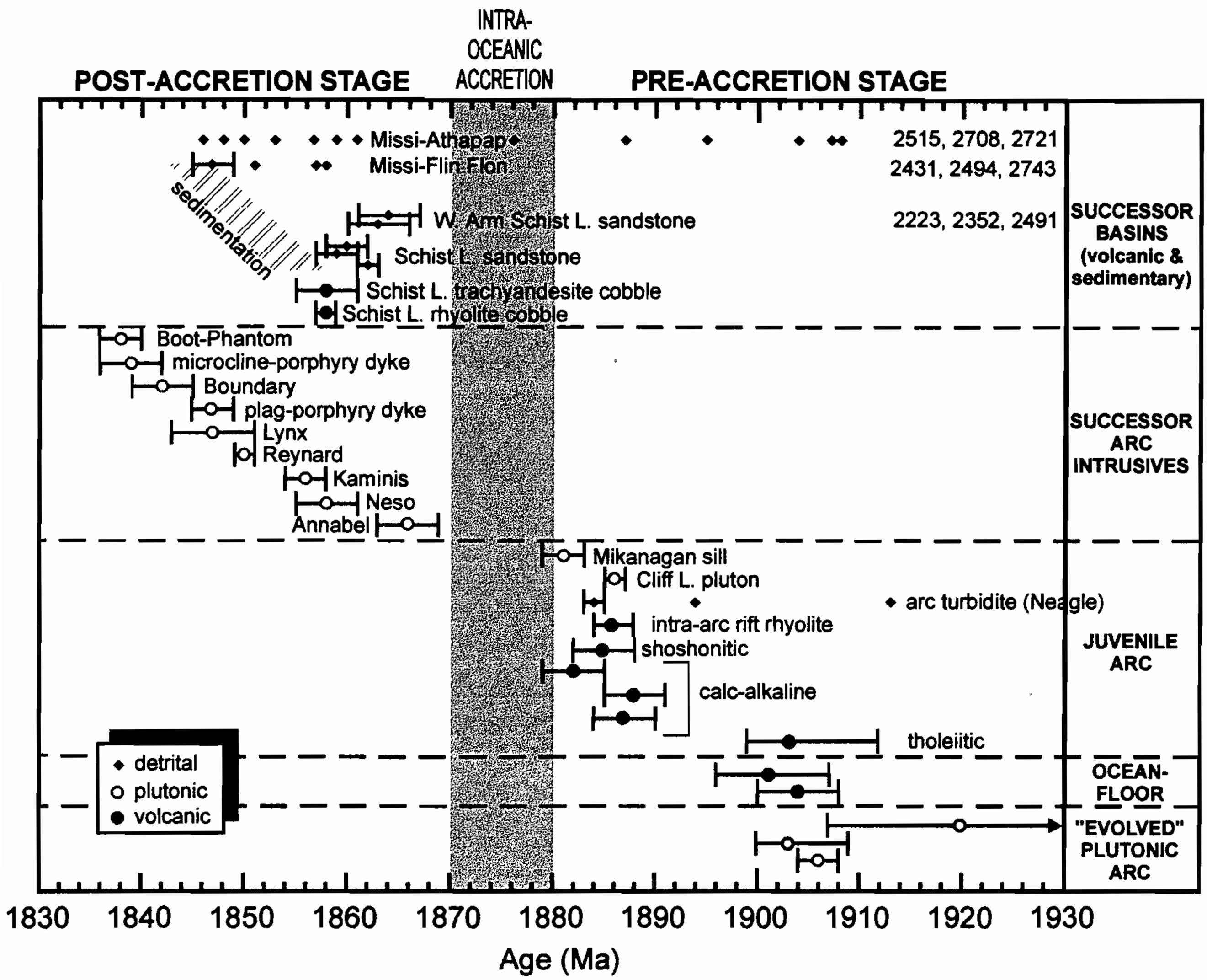


Figure 3: Summary of precise U-Pb zircon age constraints for both pre- and post-accretion rocks. Sources of data: Ansdell, 1993; Ansdell et al., 1992; David et al., 1993; David and Syme, 1995; Gordon et al., 1990; Heaman et al., 1992, 1993; Stern and Lucas, 1994; Stern et al., 1992, 1993, 1995b, unpublished data; Syme et al., 1991. Shaded area indicates age range for D1 accretion event.

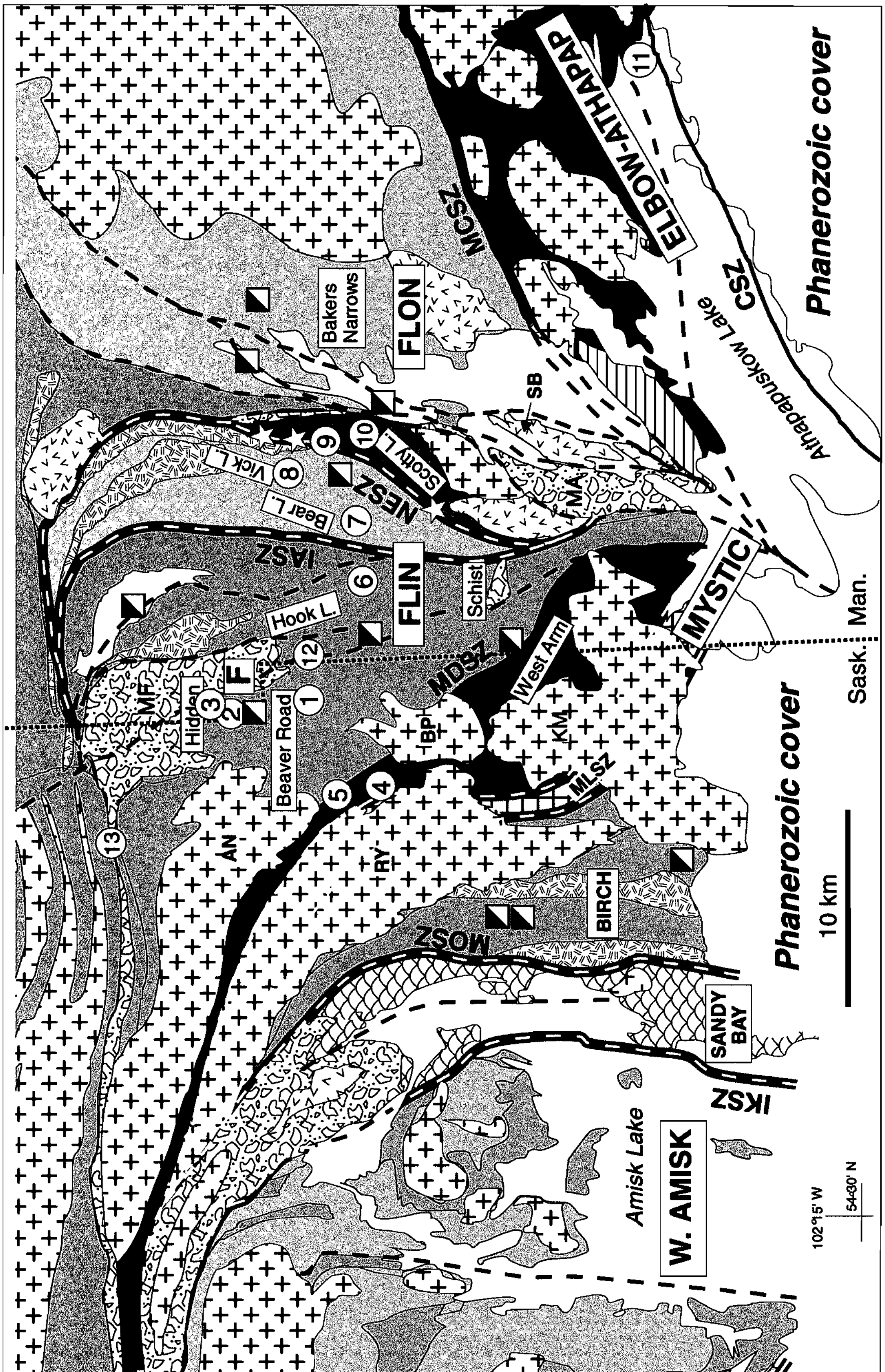
The Flin Flon belt in Trans-Hudson Orogen

The Flin Flon belt forms the southeastern part of the Trans-Hudson Orogen's internal zone (Reindeer Zone, Fig. 1), which is composed predominantly of juvenile (mantle-derived) 1.9-1.8 Ga crust (Lewry, 1981; Stauffer, 1984; Syme, 1990; Thom *et al.*, 1990; Stern *et al.*, 1995a, 1995b). The tectonomagmatic history of Flin Flon belt can be viewed in terms of the two major stages that characterize the tectonic history of the Reindeer Zone as a whole: (1) convergent margin (oceanic) tectonics and magmatism and occurred between 1.92 Ga and 1.83 Ga; and (2) intracontinental tectonics associated with collision and post-collisional convergence between bounding Archean cratons between 1.84 and 1.69 Ga (Bickford *et al.*, 1990; Gordon *et al.*, 1990; Machado, 1990; Stern and Lucas, 1994; Ansdell and Norman, 1995; Fedorowich *et al.*, 1995; Heaman *et al.*, 1994; David *et al.*, 1996). Initial collision of Reindeer Zone juvenile terranes (Fig. 1) with Archean blocks/cratons occurred at *ca.* 1.84 Ga (*cf.* Ansdell *et al.*, 1995; Connors, 1996; and references therein). Collisional deformation, marked by peak metamorphism, penetrative deformation and SW-thrusting throughout the Reindeer Zone (Lewry *et al.*, 1990; Zwanzig, 1990; Norman *et al.*, 1995), culminated at *ca.* 1.820-1.805 Ga.

LITHOPROBE seismic reflection images of Reindeer Zone crustal structure (Lewry *et al.*, 1994; Lucas *et al.*, 1994; White *et al.*, 1994) provide a means of determining the fate of the Flin Flon belt and juvenile terranes following terminal collision with Archean blocks/cratons. During collision, Reindeer Zone juvenile terranes (including the Flin Flon microcontinent) were sliced into 10-15 km thick imbricates (*e.g.*, Amisk collage) and juxtaposed between Kisseynew Domain allochthons and underthrust Archean basement (Lewry *et al.*, 1990; Lucas *et al.*, 1994). The extent of Archean basement underlying the Reindeer Zone has been inferred from isotopic study of post-collisional (<1.80 Ga) intrusive rocks (Bickford *et al.*, 1990, 1992) and seismic reflection images (Lewry *et al.*, 1994).

The Amisk collage is bound by major collisional thrust structures such as the Sturgeon-weir shear zone and the Morton Lake fault zone (Fig. 2). The Sturgeon-weir shear zone represents a 'floor thrust' and roots into the Pelican décollement zone overlying Archean basement (Lewry *et al.*, 1990; Ashton and Lewry, 1994; Heaman *et al.*, 1995). In contrast, the Morton Lake thrust zone forms part of the 'roof thrust' system that carries Kisseynew Domain turbidites and slices of its 'Flin Flon' basement on top of the Amisk collage. Although the Amisk collage is a coherent thrust imbricate, it was segmented into strike-slip fault blocks that experienced heterogeneous internal deformation (Stauffer and Mukherjee, 1971; Bailes and Syme, 1989; Lucas, 1993; Ryan and Williams, 1994, 1995, 1996; Syme, 1995), in part obscuring structures related to pre-1.83 Ga tectonics.

Post-collisional structures (D_5, Table 1) include a conjugate set of NNW to NNE-trending high-angle faults (in part reactivating older structures) that offset peak metamorphic mineral isograds (Bailes and Syme, 1989; Digel and Gordon, 1993, 1995; Ryan and Williams, 1994, 1996) and down-dropped the remnants of the successor basins (Fig. 4). More easterly-trending late faults were responsible for juxtaposing the low grade Flin Flon belt against higher grade rocks to the south of the shield (Lucas *et*



ELBOW-ATHAPAP
Phanerozoic cover
CSZ
Athapapuskow Lake
MCSZ
Bakers Narrows
FLON
SB
NESZ
Scotty L.
Vick L.
Bear L.
IASZ
FLIN
Schist
MYSTIC
MA
Hook L.
West Arm
MDBZ
F
Hidden
Beaver Road
MF
BP
KM
MLSZ
AN
RY
Sask.
Man.
Phanerozoic cover
10 km
BIRCH
MOSZ
SANDY BAY
IKSZ
Amisk Lake
W. AMISK
102°15' W
54°30' N
1
2
3
4
5
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7
8
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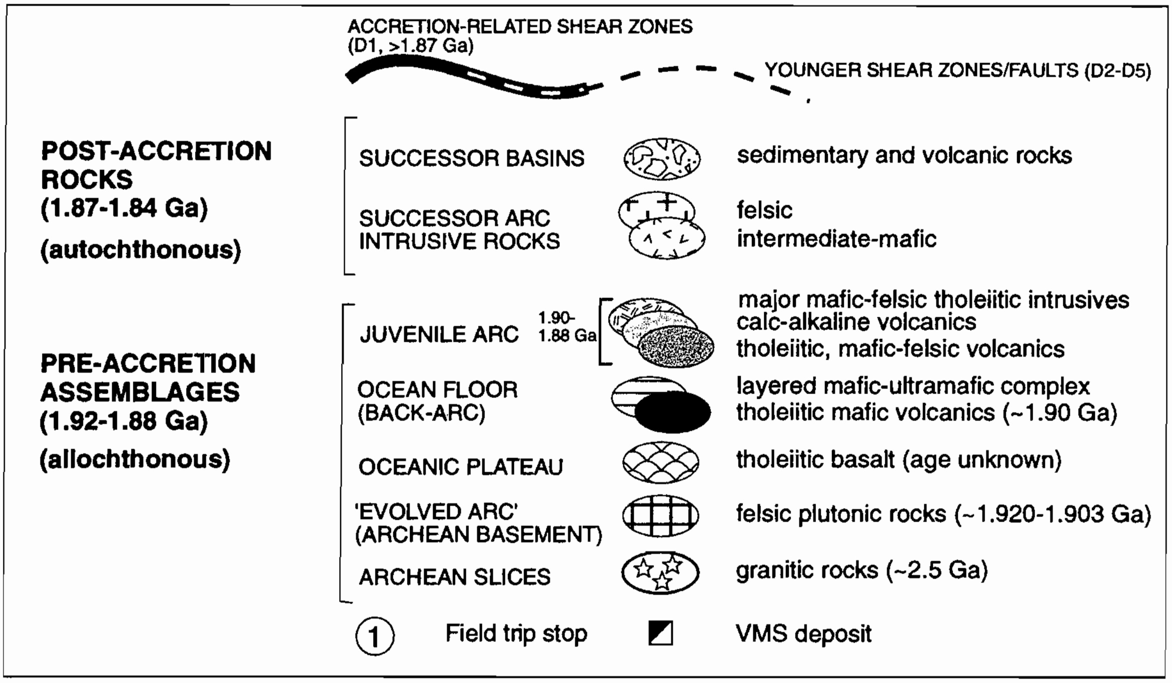


Figure 4: Tectonic assemblage map of the central portion of the Flin Flon belt, highlighting pre-accretion tectonostratigraphic assemblages and post-accretion (successor arc) plutons, volcano-sedimentary basins and faults (Bailes and Syme, 1989; Lucas, 1993; Syme and Bailes, 1993; Syme et al., 1993; Reilly, 1994; Reilly et al., 1994). Pre-accretion assemblages: W. AMISK (arc), FLIN FLON (arc), ELBOW-ATHAPAP (ocean-floor), MYSTIC (evolved arc plutons). Named suites (e.g., Hidden) are components of assemblages and are referred to in the text. F: Flin Flon Mine; town of Flin Flon. Shear zones: IKSZ: Iskwasoo shear zone; MOSZ: Mosher Lake shear zone; MLSZ: Mystic Lake shear zone; MDSZ: Meridian-West Arm shear zone; IASZ: Inlet Arm shear zone; NESZ: Northeast Arm shear zone; CSZ: Cranberry shear zone. Plutons: RY: Reynard Lake pluton; KM: Kaminis Lake pluton; BP: Boot-Phantom pluton; AN: Annabel Lake pluton; Successor basins: SB: Scheiders Bay basin; Schist: Schist Lake suite; MF: Flin Flon basin (Missi suite); MA: Athapapuskow basin (Missi suite). Modified from Lucas et al. (1996).

al., 1994; Leclair *et al.*, 1994; Syme, 1995). The youngest structure in the Flin Flon belt is a NNW-trending brittle fault (Ross Lake fault, Fig. 4) that has been dated at 1691 Ma ($^{40}Ar/^{39}Ar$ age on K-feldspar; Fedorowich *et al.*, 1995).

Tectonostratigraphic Assemblages

Arc Assemblages

Juvenile arc rocks in the Flin Flon belt can be divided into geographically separate segments, each of which is 15-50 km across. From west to east these are the Hanson Lake block, West Amisk assemblage, Birch Lake assemblage, Flin Flon assemblage, Fourmile Island assemblage, Snow Lake assemblage and Wekusko assemblage (Fig. 2). These are separated by major faults or intervening ocean-floor, turbidite, or older basement assemblages. The arc segments are internally complex, comprising numerous fault-bounded and folded volcanic suites (*e.g.*, Bailes and Syme, 1989; Table 2). It is likely that the rocks within each assemblage were closely related in their original setting, but correlation of volcanic stratigraphy between the segments has proven impossible. It is unclear whether the segments represent the fragmented parts of a formerly single arc, or were generated in completely different arcs (*e.g.*, Syme *et al.*, 1995).

The Flin Flon arc assemblage (Figs. 2, 4) contains mostly mafic volcanic rocks that were deposited in a subaqueous environment, but some pyroclastic rocks may have been erupted in a very shallow marine or subaerial setting (Bailes and Syme, 1989; Dolozi and Ayres, 1991). Basalt and basaltic andesite flows dominate the assemblage (Fig. 5). Pillow fragment breccias and mafic pyroclastic rocks are locally abundant, forming units a few metres to hundreds of metres thick. Rhyolite flows and associated felsic volcaniclastic rocks occur sporadically throughout predominantly mafic successions. The volcanic sequences represent a proximal facies with respect to source vents, as shown by the thicknesses of the volcanic components, rarity of sedimentary interbeds, and the abundance of synvolcanic intrusive rocks petrographically and geochemically similar to the volcanic rocks. Stratigraphic sequences are complex, typically displaying a wide variety of rock types, interfingering relationships, lenticular units, and abrupt facies variations. Regional metamorphic grade at Flin Flon increases northwards, towards the Kisseynew metasedimentary gneiss belt (Fig. 2), from prehnite-pumpellyite through greenschist to amphibolite facies (Bailes and Syme, 1989; Digel and Gordon, 1995), but most rocks have greenschist facies mineral assemblages (actinolite-chlorite-quartz-epidote-albite).

The Snow Lake arc segment (Fig. 2) is similar to the Flin Flon segment except for more extensive hydrothermal alteration, a higher proportion of volcaniclastic rocks and higher metamorphic grade (Bailes and Galley, 1991). The arc volcanic rocks were deposited under subaqueous conditions, and like the Flin Flon segment, the sequence also includes some material derived from shallow marine to subaerial pyroclastic deposits. Much of the Snow Lake segment is well preserved despite polyphase deformation and regional metamorphism from middle greenschist to middle amphibolite facies.

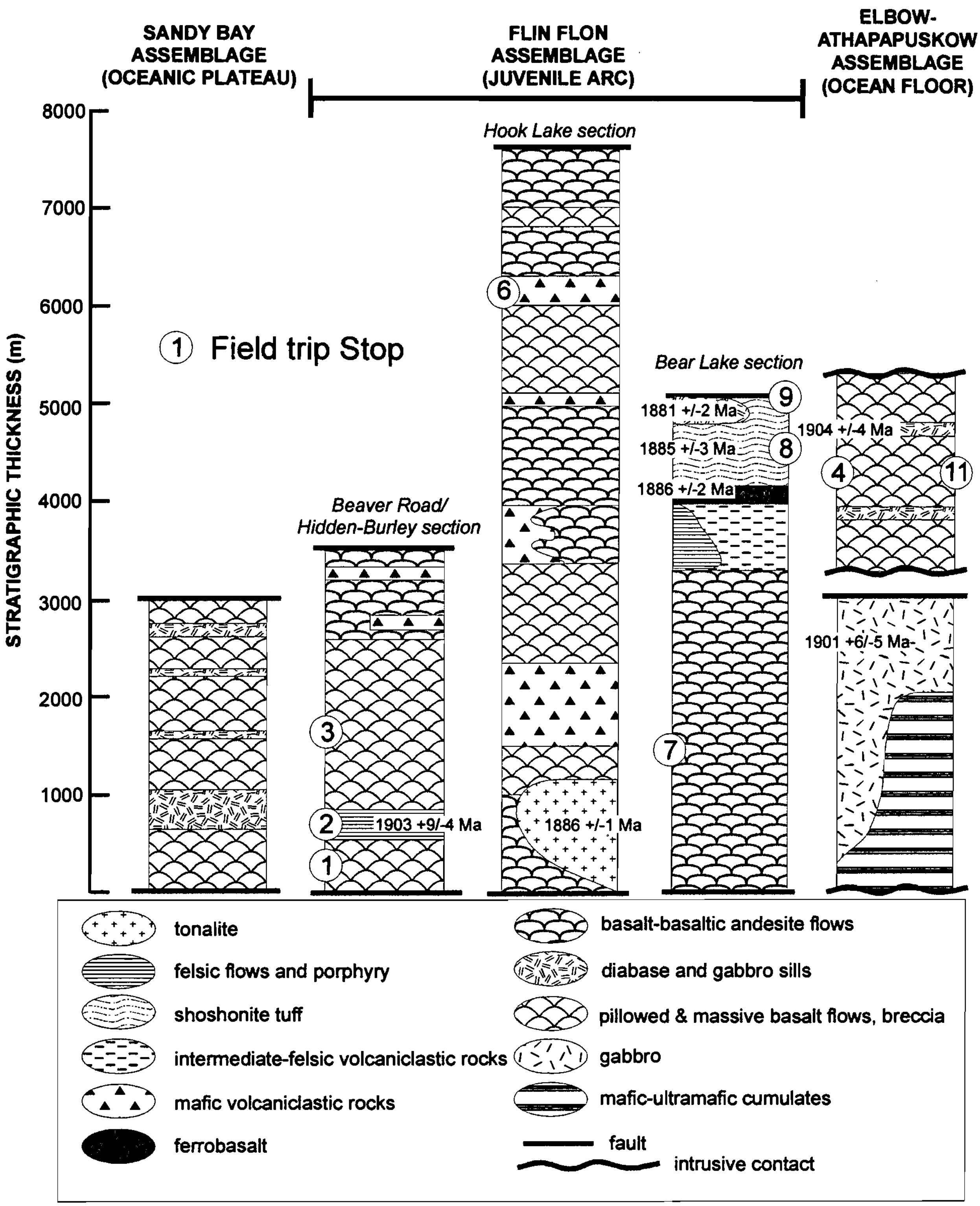


Figure 5: Representaive stratigraphic columns for juvenile arc, ocean floor and ocean plateau assembages. Sandy Bay: Reilly et al. (1994); Beaver Road/Hidden Burley: Bailes and Syme (1989), Thomas (1989); Hook Lake and Bear Lake: Bailes and Syme (1989); Elbow-Athapapuskow: Syme (1988, 1991). Modified from Lucas et al. (1996).

Table 2 *Summary of arc assemblage suites discussed in the text: Flin Flon, Snow Lake and West Amisk segments of the Flin Flon belt.*

Suite (reference)	Thickness	Components major; *minor*	Eruptive setting
Flin Flon assemblage			
Beaver Road (Thomas 1989) ***STOPS 1, 2***	2.5 km	• pillowed and massive mafic flows and breccias • volcaniclastic rocks • *rhyolite flows* • *intrusive porphyry*	subaqueous
Hidden Lake (Bailes and Syme, 1989) ***STOP 3***	3.3 km	• pillowed basalt flows (lower part) • pillowed basalt, basaltic andesite • pillow fragment breccia, mafic tuff • *andesite, rhyolite*	subaqueous
Bear Lake (Bailes and Syme, 1989) ***STOP 7***	5.5 km	• pillowed and massive basaltic andesite, pillow fragment breccia (3.3 km), overlain by: • caldera-fill rhyolite, felsic-intermediate volcaniclastics rocks	shoaling subaqueous mafic shield; caldera basin
Hook Lake (Bailes and Syme, 1989) ***STOP 6***	7.5 km	• pillowed and massive basalt, basaltic andesite • pillow fragment breccia • scoriaceous tuff and breccia	subaqueous, shallow subaqueous to subaerial
Bakers Narrows (Bailes and Syme, 1989)	3.8 km	• pillowed and massive basalt, basaltic andesite • rhyolite flows	subaqueous
Vick Lake (Bailes and Syme, 1989) ***STOP 8***	900 m	• shoshonite tuff with turbidite bedforms	subaqueous deposition; erupted in a shallow subaqueous to subaerial setting
Snow Lake assemblage			
Welch Lake (Bailes and Galley, 1991) ***STOPS 16, 17, 18***	>3 km	• pillowed basalt and basaltic andesite • rhyolite	subaqueous
Snell Lake (Bailes and Galley, 1991)	500 m	• porphyritic pillowed and massive basalt to andesite • *rhyolite at base*	subaqueous
Moore Lake (Bailes and Galley, 1991)	1 km	• pillowed aphyric basalt • amoeboid pillow breccia, pillow fragment breccia	subaqueous
West Amisk assemblage			
West Amisk (Reilly, 1993)		• andesitic tuff and flows • dacite and rhyolite tuff • *basalt*	subaqueous to subaerial

Volcanic rocks in the West Amisk arc segment (Figs. 2, 4) consist predominantly of andesitic tuffs and flows, dacitic and rhyolitic tuffs, and subordinate basaltic flows (Reilly, 1993 and references therein). A high-level, mafic to felsic volcanic complex interpreted as an emergent volcano is also present in this segment (Ayres *et al.*, 1981). Not discussed in this guidebook are arc rocks in the Hanson Lake Block (Maxeiner *et al.*, 1993), Fourmile Island assemblage (Syme *et al.*, 1995), and Wekusko Lake assemblage (Bailes, unpublished).

Geochemistry

Arc volcanic rocks within the Flin Flon, Snow Lake and West Amisk arc assemblages have been grouped into 'suites', comprising rocks that are part of locally coherent stratigraphic sequences (Table 2; Stern *et al.*, 1995a). The mafic-intermediate (<63% SiO_2) suites are subdivided into tholeiitic, calc-alkaline, shoshonitic and rare boninitic magma series (Fig. 6). Most suites are tholeiitic (TH), having high FeO^*/MgO at a given silica value. Calc-alkaline (CA) differentiation trends are displayed by the Bakers Narrows and West Amisk suites; the Bear Lake suite shows transitional tholeiitic/calc-alkaline characteristics. Some samples from the Welch Lake suite at Snow Lake show calc-alkaline tendencies, but these rocks may actually be members of a boninite (BO) series. Many of the TH series rocks have K_2O<0.5 wt.%, and are classified as low-K tholeiites; the CA rocks have 0.5-2.0 wt.% K_2O, and overlap the low- and medium-K fields (Fig. 6). Rocks from both series are characterized by relatively high Na_2O contents.

Tholeiitic series rocks (*e.g.*, Stops 1-3, 6) have the low to moderate Ni and Cr contents typical of island arc tholeiites (Stern *et al.*, 1995a; Syme, 1990). Low abundances of Ti, Zr, Hf, Nb, Y, and middle and HREEs exhibited by TH series rocks (except Moore Lake suite) are the most significant trace element characteristics of the arc assemblage, similar to modern oceanic island arc tholeiites (*e.g.*, Gill, 1981; Fig. 7, 8). For the most primitive rocks these values are only 1-2.5x their abundances in the estimated mantle source of modern N-MORBs (*e.g.*, Sun and McDonough, 1989; Pearce and Parkinson, 1993). These characteristics are likely due to their derivation from a highly refractory mantle source, more refractory than that assumed for modern N-MORBs. The extreme extent of the HFSE depletion exhibited by the Flin Flon TH series is observed in the island arc tholeiites of Tonga-Kermadec (*e.g.*, Ewart and Hawkesworth, 1987) and Fiji (Gill, 1987).

Calc-alkaline suites within the Flin Flon arc segment (Bear Lake suite (Stop 7) and Bakers Narrows suite) have low HFSE abundances and ratios, similar to the TH suites (Stern *et al.*, 1995a). The CA rocks, however, do not show marked enrichment in TiO_2 and FeO^* with fractionation (Fig. 6). They are exclusively LREE enriched, with similar La abundances compared to TH rocks but a more restricted range for HREEs (Fig. 6). MORB-normalized profiles are generally similar to the TH series rocks, with negative Nb, Zr-Hf, ±Ti anomalies, and positive Th spikes (Fig. 8). The West Amisk CA suite is markedly different from the Flin Flon segment CA suites in its higher Ni contents, and higher Zr and Nb contents at equivalent levels of fractionation (Fig. 7).

Shoshonite series rocks (SH: Vick Lake suite, Stop 8) plot within the basaltic trachyandesite and trachyandesite fields on a silica *versus* Na_2O+K_2O plot, and the

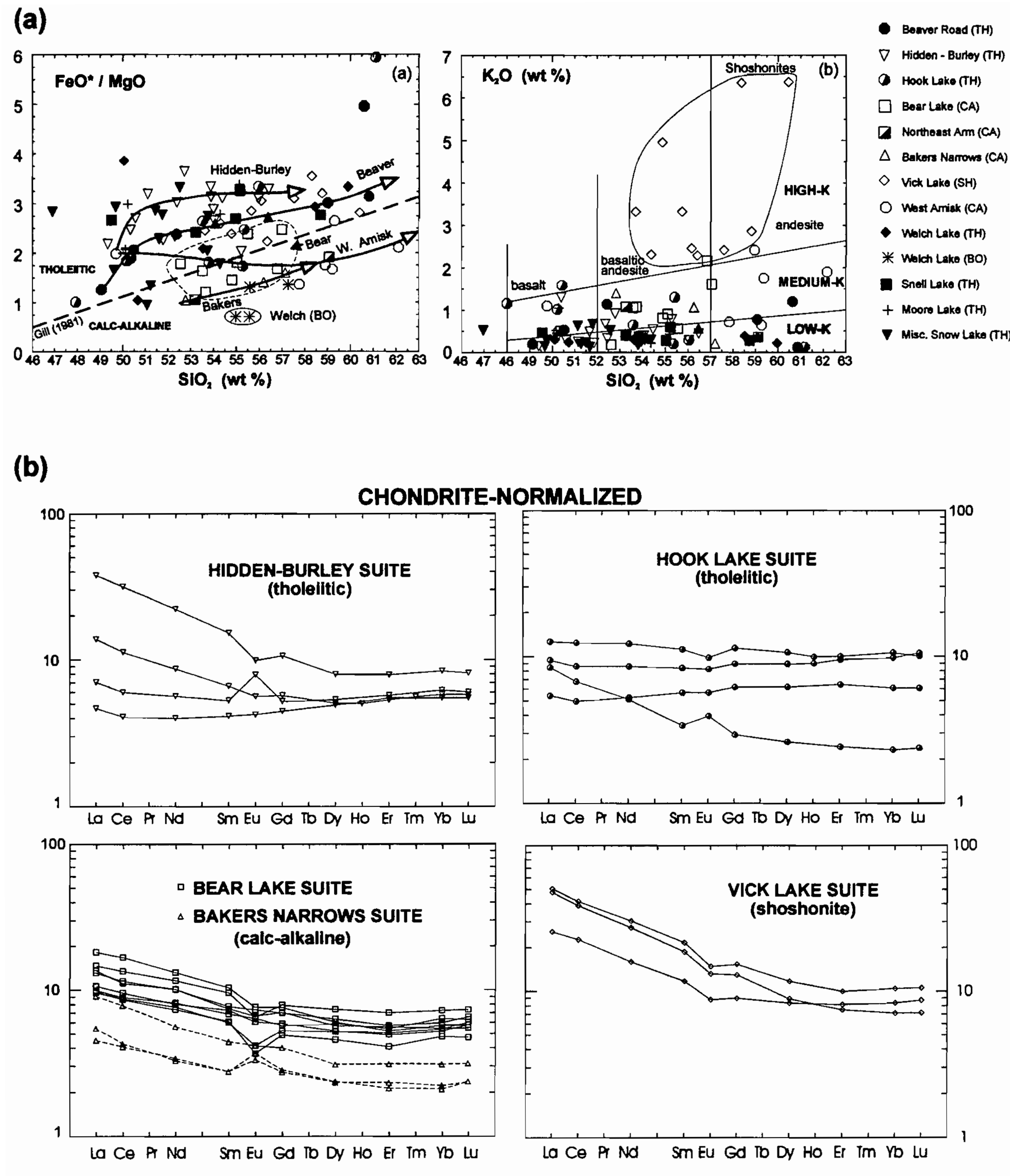


Figure 6: (a) Classification diagrams for the arc assemblage. Fractionation trends (arrows) are for the named suites in Table 2. From Stern et al. (1995a). (b) Chondrite-normalized REE patterns for arc assemblage whole-rocks at Flin Flon. From Stern et al. (1995a).

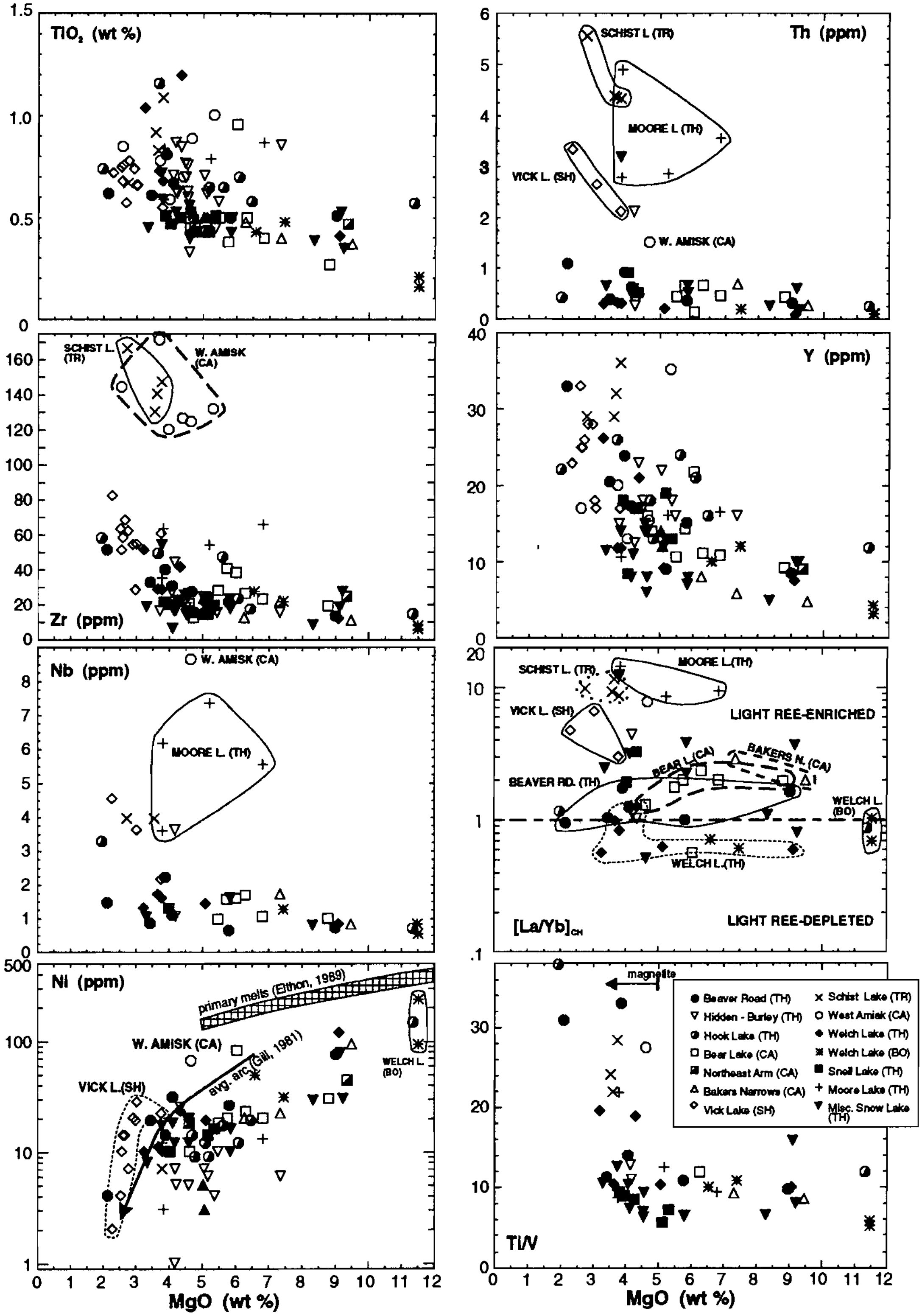


Figure 7: MgO vs. immobile minor and trace elements. In MgO vs. Ni, the curve for average orogenic rocks is from Gill (1981), and field of primary mantle melts is from Elthon (1989). From Stern et al. (1995a).

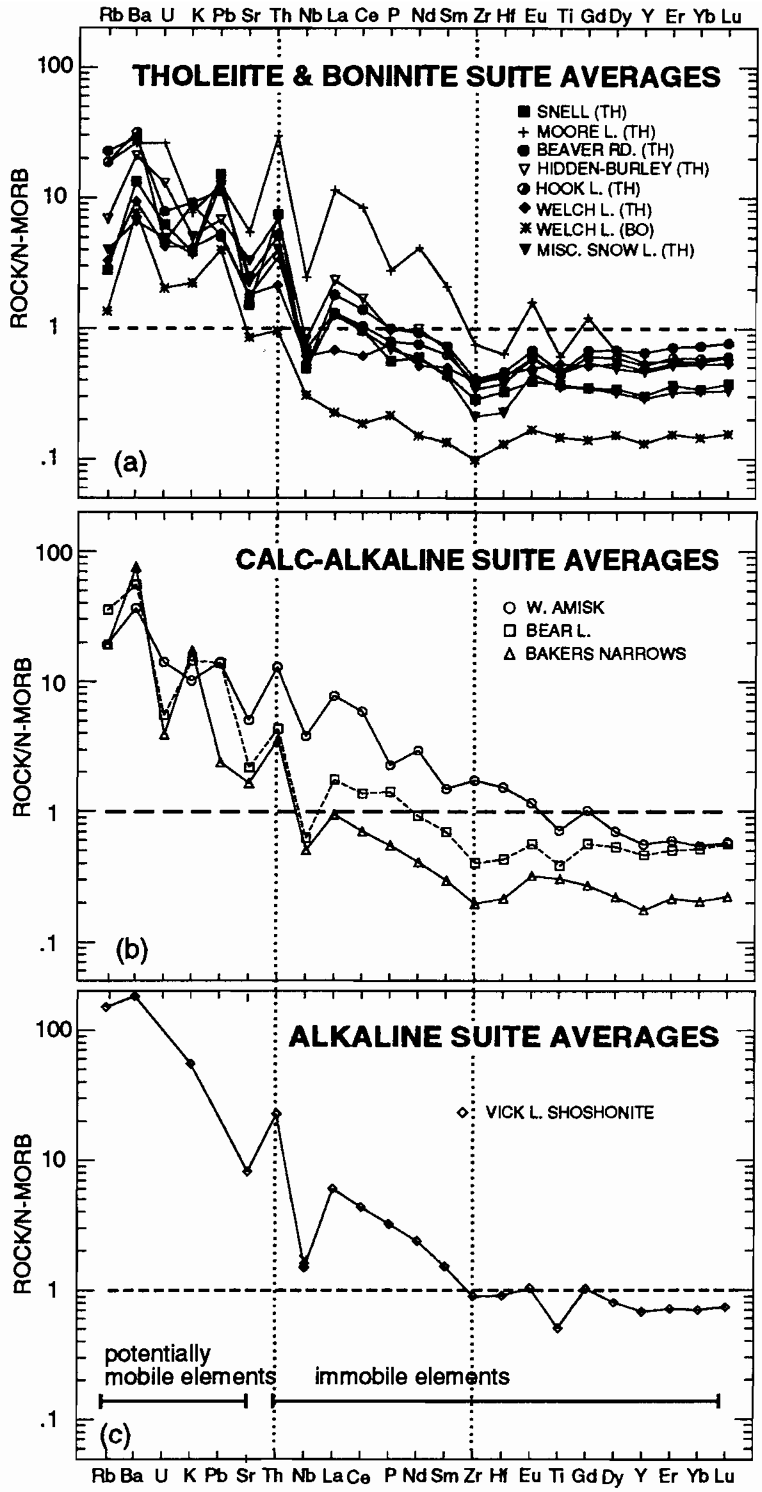


Figure 8: MORB-normalized (Sun and McDonough, 1989) incompatible element diagrams for arc assemblage suite averages. All data are by ICP mass spectrometry, except for Ti which is by XRF. Elements are in two groups, immobile elements on the right (Th-Lu) and more mobile elements on the left (Rb-Sr); within each group, elements are arranged in order of decreasing incompatibility in MORB-source mantle to the right. From Stern et al. (1995a).

high-K field on the silica *versus* K_2O plot (Fig. 6), quite distinct from all other rocks within the Flin Flon belt (Stern *et al.*, 1995a). The SH series rocks are characterized by greater enrichment in LREEs and flat to slightly fractionated HREE profiles (Fig. 6). The geological setting, age, and geochemistry of the Vick Lake SH suite suggest modern analogues erupting late in the evolution of oceanic arcs (*e.g.,* Morrison, 1980; Gill and Whelan, 1989).

Boninites are characterized by a combination of high MgO and high SiO_2 contents (*e.g.*, Hickey and Frey, 1982). Boninites are currently recognized only in the Snow Lake arc segment, within a 200 m-thick unit near the stratigraphic top of the Welch Lake basalt (see Snow Lake area Stops). Two samples with 54.5-56% SiO_2 and ~11 wt.% MgO plot within the boninite field on a SiO_2 *versus* MgO diagram (Stern *et al.*, 1995a). The Welch Lake boninites have the high CaO contents and high CaO/Al_2O_3 ratios characteristic of the lesser-known group of high-Ca boninites (Crawford *et al.*, 1989). Boninites have the lowest REE abundances of all rocks at 1-2x chondrites, with slightly LREE depleted to dish-shaped pattern. A forearc eruptive setting is suggested for the Welch Lake boninite, similar to more recent counterparts (Tongan forearc; Upper Pillow Lavas of the Troodos ophiolite) which, like Welch Lake, are intimately associated with low-Ti tholeiitic basalts.

Initial epsilon Nd (ε_{Nd}) values for TH series rocks range from -0.4 to +4.3, but intra-suite variability is somewhat smaller (Stern *et al.*, 1995a). Simple mixing calculations illustrate that most ε_{Nd} values for the TH series rocks in the Flin Flon arc assemblage can be explained by the addition of ~1-8% of a recycled late Archean crustal component to juvenile arc magmas. There is no systematic variation in the apparent proportion of older crustal component in the volcanic rocks with crystallization age. For the Flin Flon segment TH rocks, sediment subduction rather than intracrustal recycling was the major process (Stern *et al.*, 1995a). However, intracrustal recycling of older crust may locally have been an important process. The rocks from the Snow Lake assemblage tend to have consistently low initial ε_{Nd} values (<+3.1) compared to rocks from the other segments (Stern *et al.*, 1995a). Archean xenocrysts in a Snow Lake rhyolite (David *et al.*, 1996) suggest that arc magmas at Snow Lake were exposed to more older crust than at Flin Flon, supporting the idea that Snow Lake represents a separate arc.

Ocean-Floor Assemblage

Juvenile ocean-floor assemblages include basalt sequences and related mafic-ultramafic complexes (Figs. 2; 5). The main occurrence of ocean-floor basalts is a semi-continuous belt between the Elbow Lake and Athapapuskow Lake areas, Manitoba, termed the 'Elbow-Athapapuskow assemblage' (Fig. 2; Stern *et al.*, 1995b). The Elbow-Athapapuskow ocean-floor assemblage is up to 25 km wide and has a strike length of 100 km (Syme and Bailes, 1993; Syme, 1994; Stern *et al.*, 1995b); it extends beneath the Phanerozoic cover for at least another 75 km (Leclair *et al.*, 1993). This assemblage is everywhere either in fault contact with arc volcanic rocks, or the contact with arc rocks is stitched by younger plutons. Systematic mapping (Syme, 1988, 1991, 1992, 1993, 1994) has shown the Elbow-Athapapuskow assemblage to consist entirely of subaqueous basalt and related intrusives, in which there are no known occurrences of felsic volcanic rocks, terrestrial sediments, or older crystalline basement. Gabbro and diabase dykes and sills are common, although no sheeted dykes are known.

The ocean-floor basalts occur as laterally coherent units, 4 km to >60 km in strike length, that have stratigraphic thicknesses of 0.3-3.0 km, each having characteristic weathering colour, flow morphology, alteration assemblage and geochemistry (Syme, 1991, 1992; Stern *et al.*, 1995b). These units are informally termed 'formations' (listed in Fig. 9; see Stern *et al.* (1995b) and Syme (1995) for details). Elbow-Athapapuskow ocean-floor volcanism was concurrent with arc magmatism, demonstrated by the identical crystallization ages of a synvolcanic diabase sill on Athapapuskow Lake (1904 ±4 Ma: Stern *et al.*, 1995b) and Flin Flon assemblage arc tholeiites (1904 +6/-3 Ma: Mine Rhyolite, David *et al.*, 1993).

Kilometre-scale layered mafic-ultramafic bodies within the Elbow-Athapapuskow assemblage (Syme, 1988, 1991, 1992, 1994; Syme *et al.*, 1995; Williamson and Eckstrand, 1995) may represent dismembered syn-volcanic plutonic complexes (Stern *et al.*, 1995b). Igneous zircons recovered from the mafic-ultramafic complex at Claw Lake (southeast of Elbow Lake, Fig. 2) have a U-Pb age of 1901 +6/-5 Ma (Stern *et al.*, 1995b), establishing that these layered mafic-ultramafic plutonic complexes are coeval with ocean-floor volcanism in the Elbow-Athapapuskow assemblage.

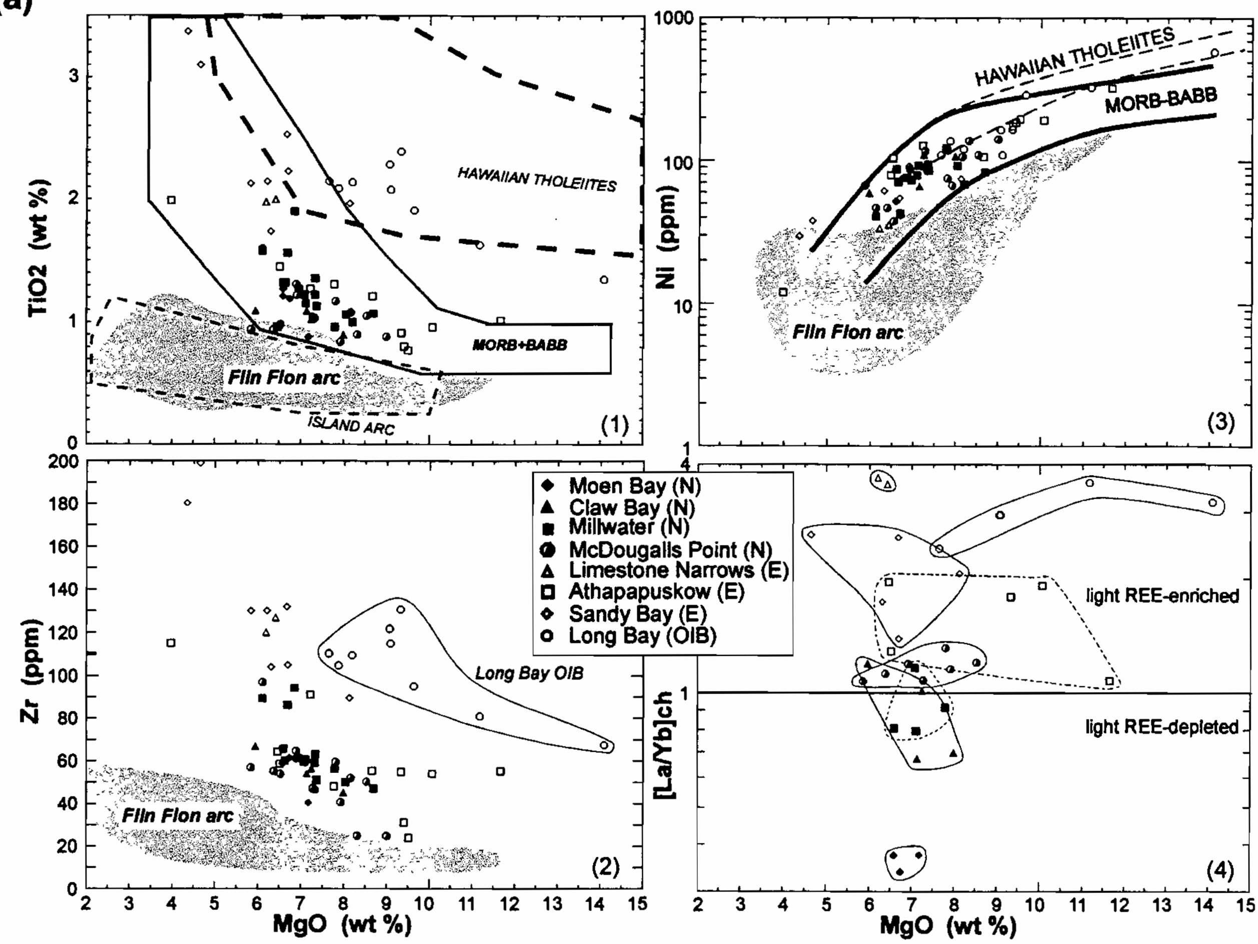
Geochemistry

The ocean-floor basalts are dominantly sub-alkaline with MgO contents typical of modern MORBs, falling mostly in the range 6-10 wt.% (Stern *et al.*, 1995b). The ocean-floor basalts have higher TiO_2, Zr, and Ni compared to the coeval arc volcanic rocks of equivalent MgO (Fig. 9). The ocean-floor basalts can be subdivided into two main types, using the ratios of the immobile trace elements (*e.g.*, Zr, Y, Nb, Th, REEs; LeRoex, 1987): 1) N-type (similar to 'normal' MORB); 2) E-type (similar to 'enriched' MORB). The Flin Flon E-type ocean-floor basalts have geochemical features in common with 'enriched', 'transitional', or 'plume' MORBs (*e.g.*, Sun *et al.*, 1979; Basaltic Volcanism Study Project, 1981; Schilling *et al.*, 1983; LeRoex *et al.*, 1983, 1985; LeRoex, 1987). Each of the basalt formations has a restricted range in composition, allowing them to be classified into N- or E-types.

N-type basalts have major element and incompatible trace element characteristics within the range exhibited by modern N-MORBs (Stern *et al.*, 1995b). They differ from modern N-MORBs in their tendency to lower TiO_2 and higher FeO^* contents; some have higher Th/Nb ratios. The lower TiO_2 and higher Th/Nb ratios suggest affinities of some of the rocks to back-arc basin basalts from the Mariana Trough, North Fiji Basin or Lau Basin (*e.g.*, Sinton and Fryer, 1987; Hawkins and Melchior, 1985; Price *et al.*, 1990; Falloon *et al.*, 1992). Initial ε_{Nd} values for N-type ocean-floor basalts as a whole range from +3.4 to +5.4, but are homogenous within individual formations (Stern *et al.*, 1995b).

E-type basalts (including Athapapuskow basalt, Stop 11) have higher TiO_2 than the N-types and are comparatively enriched in Th, Nb and LREEs (Fig. 9; Stern *et al.*, 1995b). The E-type Athapapuskow basalts are the least fractionated of all the ocean-

(a)



(b)

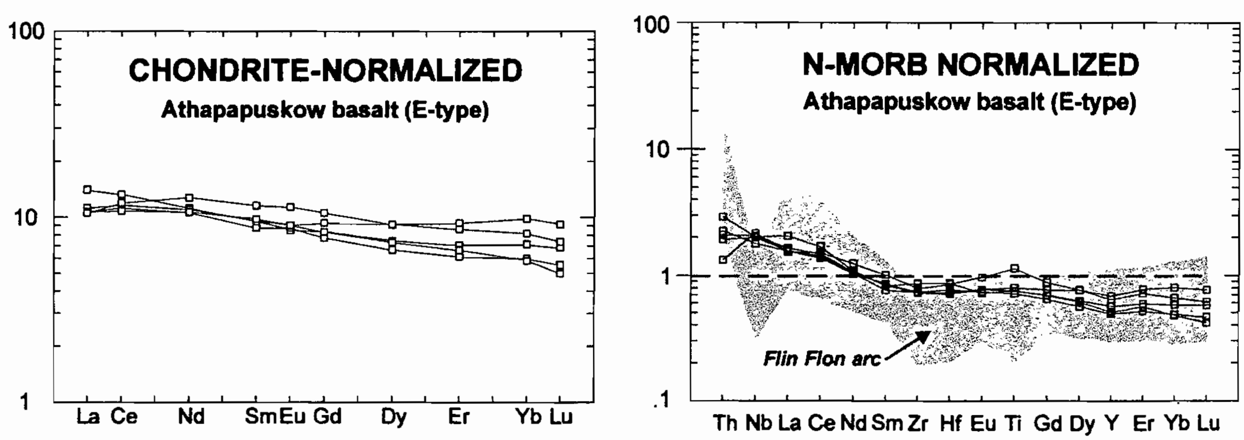


Figure 9: (a) Selected plots for the Flin Flon ocean-floor and ocean island basalts. Shaded field is for Flin Flon tholeiitic arc volcanic rocks (Stern et al., 1995a). BABB: back-arc basin basalt; MORB: mid-ocean ridge basalt. (1) MgO vs TiO_2, with dashed fields for modern intra-oceanic rocks from numerous published data sources; (2) MgO vs. Zr; (3) MgO vs. Ni; (4) MgO vs. chondrite-normalized La/Yb. From Stern et al. (1995b). (b) Chondrite-normalized REE and N-MORB-normalized trace element patterns for Athapapuskow basalt (Stop 11). From Stern et al. (1995b).

floor basalts, with several samples having between 9 and 12 wt.% MgO and 150-300 ppm Ni.

Similar to modern MORBs, the ocean-floor basalts of the Elbow-Athapapuskow assemblage are mostly non-primitive compositions, and there is relatively limited chemical variability within individual formations (Stern *et al.*, 1995b). Some show no evidence of either a continental crust or arc signature, confirmed by their strongly positive initial ε_{Nd} values. Other basalt formations, however, show dual features of MORB-like major element geochemistry and arc trace element signature (Stern *et al.*, 1995b), characteristics of some basalts in modern intra-oceanic back-arc basins (*e.g.*, Saunders and Tarney, 1991).

Ocean plateau and Ocean island basalt (OIB) assemblages

Sandy Bay E-type basalt (ocean plateau)

The Sandy Bay assemblage (Figs. 2, 4) is a *ca.* 3 km thick, monotonous sequence of basalt flows and syn-volcanic sills of unknown age (Fig. 5; Reilly *et al.*, 1994). These basalts are tectonically separated from the basalts of the Elbow-Athapapuskow assemblage by the Mosher Lake shear zone (Fig. 4).

Sandy Bay basalts are uniformly tholeiitic and plot in the E-MORB/tholeiitic OIB field on a Zr-Th-Nb diagram (Stern *et al.*, 1995b). The basalts are geochemically distinct from those of the arc and ocean-floor assemblages (Fig. 9). Trace element ratios range from N-MORB-like to OIB-like. REE patterns range from slightly to moderately LREE enriched. Two samples of Sandy Bay basalt that bracket the trace element compositional range of the sequence yield identical initial ε_{Nd} values of +4.5. The subaqueous basalts show no arc signature, and the trace element characteristics hint at plume-related magmatism (Stern *et al.*, 1995b). The Sandy Bay basalts may represent a fragment of an oceanic plateau, seamount, or other off-axis volcanic construct. Whatever their original tectonic setting, the basalts in this structurally-isolated block can not be related in any first-order way the basalts of the Elbow-Athapapuskow assemblage.

Long Bay conglomerate (ocean island basalt)

An isolated occurrence of conglomerates consisting principally of basaltic detritus is in possible unconformable contact with arc assemblage rocks at Elbow Lake (Fig. 2; Syme, 1991). The basaltic clasts are scoriaceous to strongly amygdaloidal, commonly display vesicle banding and preserve ropy or crenulated internal contacts in some clasts. These features are consistent with subaerial eruption of the basalts, although subaqueous debris flow bedforms suggest that the conglomerates were deposited in a marine setting as density current (proximal turbidite) flows (Syme, 1991).

The geochemical characteristics of these basalt clasts are remarkably similar to tholeiitic ocean island basalts (OIBs: *e.g.*, Hawaii; Stern *et al.*, 1995b). In addition to their high MgO and low Al_2O_3 contents, the trace element ratios and REE patterns are also similar to Hawaiian tholeiites (Fig. 9). The basalts have modest overall LREE enrichment, with concave-downward LREE profiles and HREE that are strongly fractionated. The Long Bay OIBs have smooth N-MORB normalized trace element patterns and primitive Th/Nb ratios (<0.1), suggesting these magmas had no significant interaction with older continental crust or metasomatized arc mantle. The Long Bay OIBs may have been related to hot-spot magmatism taking place on a remnant arc ridge, coeval with oceanic spreading in the Elbow-Athapapuskow back-arc basin.

Isotopically-evolved Proterozoic rocks/Archean rocks

The evidence for isotopically-evolved arc plutonic rocks in the central part of the Flin Flon belt is limited to the highly deformed Mystic Lake assemblage (Fig. 4; Reilly, 1992; Syme *et al.*, 1993). The Mystic Lake assemblage comprises massive to layered, amphibolite-grade tonalite, granodiorite and diorite. The geochemistry of Mystic Lake assemblage plutonic rocks is typical of calc-alkaline arc plutons (Stern, unpublished data). The distinguishing features of these rocks are their geochronology and isotopic signatures. U-Pb zircon dating has indicated that the tonalites range in age from ~1920 to 1903 Ma, and also that they contain both zircon cores and probable xenocrystic zircons that are Neoarchean in age (Stern and Lucas, 1994). Initial ε_{Nd} (-3.1 to -6.1) and $^{87}Sr/^{86}Sr$ (0.704-0.709 [mantle = ~0.702]) compositions indicate that the tonalites are highly evolved and non-juvenile (Stern *et al.*, 1992, 1993). Coupled with the evidence for xenocrystic zircons, this suggests that Archean basement was involved in the generation of these plutonic rocks (Stern, unpublished data). The Mystic Lake assemblage is interpreted as a tectonic slice of the middle crust of an arc built on Archean crust, possibly a microcontinental fragment (Lucas *et al.*, 1996).

Archean crustal fragments represent a minor (<<1%) but important component in the Amisk collage. Granitoid rocks, dated at 2497 and 2518 Ma (David and Syme, 1994) and with an average initial ε_{Nd} composition of -6.9 (Stern *et al.*, 1995a), occur as fault-bound lozenges (10-100 m wide x 100's of metres long) within the Northeast Arm shear zone (Stops 9, 10: Fig. 4; Lucas, 1993). The shear zone juxtaposes imbricates of the Flin Flon arc assemblage (and other assemblages?) characterized by contrasting stratigraphy, geochemistry, Nd isotopic signature and structural history.

Accretion

Juxtaposition of the arc and oceanic (back-arc) assemblages (D_1) in the central part of the Flin Flon belt (Fig. 4) is relatively well constrained to have occurred between 1882 and 1866 Ma, the age of the youngest volcanic rock and oldest 'stitching' pluton, respectively (Stern and Lucas, 1994). As discussed above, the pre-1866 Ma D_1 deformation in the Flin Flon belt resulted in the development of the Amisk collage (Table 1). Field study of shear zones bounding tectonostratigraphic assemblages in the central Flin Flon belt (Figs. 2, 4), first recognized through geological mapping by the Manitoba and Saskatchewan Geological Surveys (Bailes and Syme, 1989; Syme, 1988, 1995; Syme *et al.*, 1993; Thomas, 1989, 1990; Reilly, 1992, 1994; Reilly *et al.*, 1993), has revealed a complex and protracted tectonic history involving multiple episodes of magmatism, sedimentation, deformation and metamorphism (Table 1; Lucas *et al.*, 1996). These shear zones include (from west to east) the Iskwasoo, Mosher Lake, Mystic Lake, Meridian-West Arm, Inlet Arm, Northeast Arm (Fig. 4), the eastern four of which will be examined in the field trip. D_1 geometries are not easily resolved because virtually all stratigraphic contacts and early tectonic foliations in the 1.92-1.88 Ga assemblages dip steeply, having been deformed during D_2-D_5 deformation events.

Juxtaposition of arc and oceanic (back-arc) assemblages at 1.88-1.87 Ga may have occurred by either (1) thrust faulting in an accretionary complex, with steepening of structures during D_1 and D_2; or (2) juxtaposition of assemblages along strike-slip faults. It is plausible that both the juvenile and evolved arc assemblages in the central Flin Flon belt area were derived from the same arc system (*e.g.*, Flin Flon, Birch, Mystic Lake, West Amisk), and were interleaved with back-arc slices during D_1 accretion. Possible plate fragments ('terranes') are represented by the central Flin Flon belt arc assemblages (Flin Flon, Birch, West Amisk) and the Elbow-Athapapuskow ocean-floor assemblage, as well as the Snow Lake assemblage and Hanson Lake Block arc rocks. The regional map pattern consists of major, NNE-striking 'terranes' that extend south below the Phanerozoic cover and have strike lengths well in excess of 100 km (*e.g.*, Snow Lake arc assemblage; Elbow-Athapapuskow assemblage; Hanson Lake Block).

An intraoceanic setting for assembly of the Amisk collage is suggested by (1) the very limited scale of older (Archean) crust in the collage (<<1% of exposed area), (2) the interleaving of assemblages derived from at least two or three (micro)plates, (3) isotopically-juvenile post-accretion arc magmatism, and (4) lack of evidence for a major uplift and exhumation event at *ca.* 1.88-1.86 Ga (Lucas *et al.*, 1996). The apparent absence of melange associated with the proposed early structures is consistent with the inferred intraoceanic tectonic setting for the Amisk collage, well away from heavily-sedimented regions adjacent to thick continental plates and/or collision zones.

Post-accretion (successor) arcs and basins

The 1.92-1.88 Ga tectonostratigraphic assemblages were intruded by felsic to mafic intrusive rocks and overlain by sedimentary and volcanic rocks following initial development of the Amisk collage (Lucas *et al.*, 1996). In the central part of the Flin Flon belt (Fig. 4), the plutons range in age from 1866 to 1838 Ma (Fig. 3), in composition from gabbro to granite (Syme *et al.*, 1993), and in deformation state (exclusive of D_3-D_5 strain) from mylonitic (oldest) to isotropic (youngest). The sedimentary and volcanic rocks are preserved in grabens and synformal basins that formed during later intracontinental deformation (<1.83 Ga, D_3-D_5), but primary stratigraphic relations with both the older assemblages and the plutons have been well documented (Stauffer and Mukherjee, 1971; Bailes and Syme, 1989; Stauffer, 1990). These rocks accumulated in a variety of relatively small basins, and span the age range of the plutons (1.86-1.84) as well as that of voluminous turbidites in the adjacent Kisseynew Domain (Fig. 2; David *et al.*, 1993, 1995; Ansdell *et al.*, 1995).

1.87-1.84 Ga Plutons

The 1.87-1.84 Ga intrusive rocks are predominantly hornblende-bearing, medium-K calc-alkaline diorite-tonalite-granodiorite plutons, with lesser high-K gabbro-monzodiorite-granite plutons (Syme, 1988; Ansdell and Kyser, 1992; Whalen, 1992; O'Hanley and Kyser, 1994). The plutons can show significant internal compositional variation and zoning (Whalen, 1993). Trace element geochemistry indicates that the intrusive rocks have a typical arc signature, with high field strength element depletion and light REE enrichment (*cf.* Syme, 1988; Ansdell and Kyser, 1992; Whalen, 1993). Mafic magmas are represented throughout the 1.87-1.84 Ga interval and suggest that the calc-alkaline magmatism was directly linked to the addition of mantle melts to base of the collage crust. The 1.87-1.84 Ga plutons have uniformly juvenile initial ε_{Nd} compositions despite the fact that some cut isotopically-evolved assemblages (Mystic Lake assemblage, Fig. 4). This suggests that there was relatively little involvement of older crustal components in the generation of the plutons, perhaps not surprising in that older crust (incl. Mystic Lake Assemblage) comprises a very minor portion of the Amisk collage.

Calc-alkaline plutons largely mask the Amisk collage assemblages to the east of the central Flin Flon belt map area (Fig. 2; Whalen, 1992; Morrison and Whalen, 1995) and range in age from 1876 Ma to 1845 Ma; one late pluton has a U-Pb titanite age of 1826 Ma (Whalen and Hunt, 1994). These plutons are similar to those in the central part of the belt in terms of their geochemistry (Whalen, 1992), juvenile Nd-isotopic signature (Stern, unpublished data) and deformation state (Ryan and Williams, 1994, 1995, 1996; Morrison and Whalen, 1995). Integrating the geological, geochemical and isotopic data from across the ~100 km width of the 1.87-1.84 Ga plutonic belt (Fig. 2), Lucas *et al.* (1996) proposed that the calc-alkaline magmatism was related to an arc built on top of the Amisk collage (Fig. 10).

1.87-1.84 volcanic and sedimentary rocks

Volcanic, volcaniclastic and sedimentary rocks that are younger than the 1.92-1.88 Ga tectonostratigraphic assemblages have been documented across the central part of the Flin Flon belt ('successor' basin deposits, Figs. 2, 4). These rocks may represent the remnants of depositional basins that formed alluvial aprons and fluvial systems. Successor basin deposits fall into two contrasting types: older (>1.85 Ga) subaerial to submarine volcaniclastic and epiclastic deposits, and younger (<1.85 Ga) subaerial deposits derived from erosion of successor arc volcanic and plutonic rocks as well as the older tectonostratigraphic assemblages. Basement to the older basins has yet to be found, whereas younger basin sandstones and conglomerates unconformably overlie Amisk collage assemblages and post-accretion arc plutons (Bailes and Syme, 1989; Holland *et al.*, 1989; Stauffer, 1990).

An example of the older deposit type is the Schist Lake conglomerate-sandstone unit ("Schist", Fig. 4), composed entirely of volcanic detritus, including hornblende-plagioclase phyric trachyandesite boulders and rhyolite and felsic porphyry cobbles (Syme, 1988; Syme *et al.*, 1993). U-Pb zircon geochronology has yielded an age of 1858 Ma for a trachyandesite boulder (David *et al.*, 1993), 1858 Ma for a felsic porphyry cobble, and 1862-1859 Ma for detrital zircons in a sandstone sample (Fig. 3; Stern *et al.*, in prep.). The nearby 1858 Ma Neso Lake pluton (Fig. 4; Syme *et al.*, 1991) is a multi-phase intrusion that could be the plutonic root to the volcanic source of the trachyandesite boulders (Syme, 1988, 1991). The Schist Lake basin is interpreted as part of an alluvial-fluvial system built on the post-accretion arc and receiving detritus from one or more subaerial volcanic edifices.

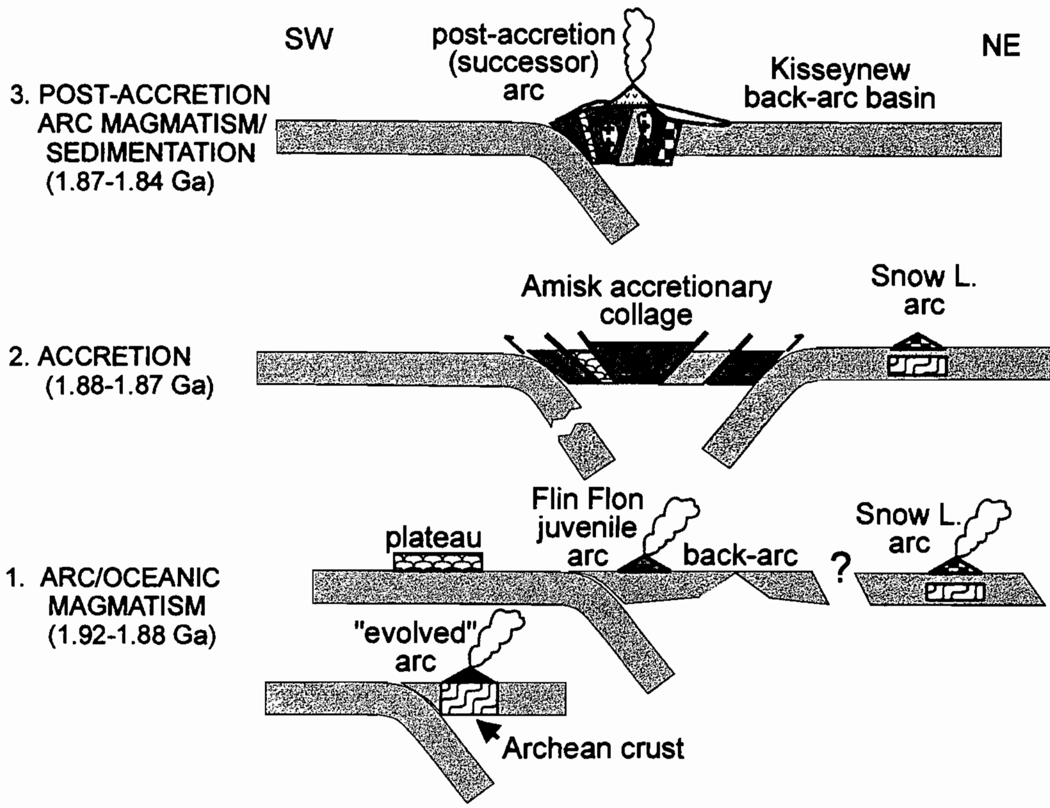


Figure 10: Cartoon illustrating the tectonic evolution of the Flin Flon belt (ca. 1.92-1.84 Ga). From Lucas et al. (1996).

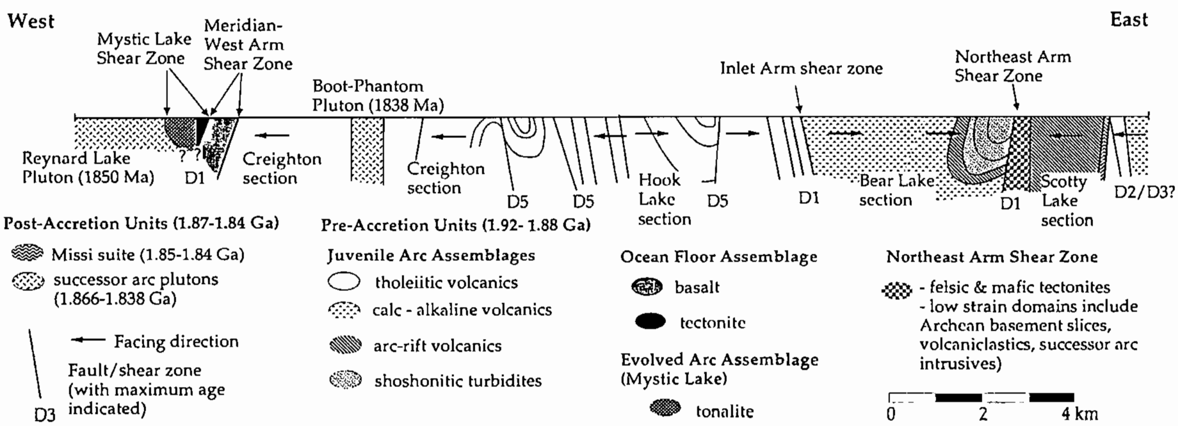


Figure 11: Schematic vertical cross-section across the central Flin Flon belt map area, constrained by surface dips only. Section illustrates the steeply-dipping structure of the Amisk collage and the predominance of strike-slip faulting through much of the deformation history (D2-D5). Folds are F2/D2 (Bear Lake and Creighton sections) and F3/D3 (Hook Lake section) in age. From Lucas et al. (1996).

The younger successor basin deposits are characterized by >2 km thick packages of sandstone and conglomerate deposited in alluvial and fluvial environments (*Missi suite*, Fig. 4; Bailes and Syme, 1989; Syme, 1988; Stauffer, 1990; and references therein). U-Pb analysis of detrital zircon populations in the sandstones (youngest zircon is 1846-1847 Ma; Ansdell *et al.*, 1992; Ansdell, 1993) and cross-cutting intrusions (1842 Ma; Heaman *et al.*, 1992) has bracketed sedimentation to ~1845 Ma in the central Flin Flon belt (Fig. 3). The *ca.* 1845 Ma volcanic and siliciclastic rocks were deposited in nonmarine basins associated with braided river systems (Bailes and Syme, 1989; Stauffer, 1990) that initially dissected the volcanic edifice of the post-accretion arc and later its basement (Amisk collage and plutons).

Deformation

Several field relations indicate that the D_1 shear zones were active (reactivated?) during 1.87-1.84 Ga arc magmatism, suggesting that D_2 represents an intra-arc deformation event (Table 1). First, the Mystic Lake and Meridian West-Arm shear zones contain abundant layer-parallel felsic and mafic sheets with a variably developed shear zone foliation. The felsic sheets range in age from *ca.* 1.87 Ga to 1.84 Ga (Stern *et al.*, 1993) and show a decrease in penetrative strain and an increase in obliquity relative to the shear fabric with decreasing age (see below). Second, the older plutons (*e.g.*, 1866 Ma Annabel Lake, 1856 Ma Kaminis Lake and 1850 Ma Reynard Lake plutons; Fig. 4) contain tectonic fabrics that parallel those in the shear zones (Reilly, 1992). Development of an amphibole lineation and growth of oriented hornblende in the necks of boudinaged pyroxenite veins indicates that metamorphism is synchronous with D_1-D_2 deformation (Table 1). Lucas *et al.* (1996) suggest that metamorphism resulted primarily from heat advected into the accretionary collage during plutonism (*i.e.*, 'regional' contact metamorphism).

Syn-magmatic D_2 shear zones effectively span the entire east-west width of the Amisk collage (Fig. 2), suggesting a regional deformation episode that was accommodated along a series of fault/shear zone strands (Lucas *et al.*, 1996; Morrison and Whalen, 1995; Ryan and Williams, 1994, 1995, 1996; Syme, 1993, 1994). It is likely that D_2 structures (*e.g.*, folds, steep belts, shear zones) and associated topography controlled the pattern of fluvial drainage systems and associated Missi suite sedimentation at 1.85-1.84 Ga.

An unresolved question is what caused the major uplift and erosion episode that resulted in the unroofing of the Amisk collage/plutonic basement by *ca.* 1845 Ma. The timing of this episode is bracketed by the change in character of the successor basin deposits at *ca.* 1.86-1.85 Ga, with older ones only containing volcanic detritus and Missi suite basins containing compositionally-diverse volcanic and plutonic detritus. This major uplift event may reflect thickening of the crust/lithosphere due to the combined effects of D_2 deformation and post-accretion arc magmatism. D_2 shortening may be associated with a collision between the Amisk collage and another terrane (Lucas *et al.*, 1996). A number of collisions occurred in the interval 1.86-1.85 Ga across the Trans-Hudson Orogen, thus providing a number of possible sources for the voluminous 1.85-1.84 Ga turbidites of the Kisseynew Domain (*cf.* Ansdell *et al.*, 1995).

STOPS IN THE FLIN FLON AREA

Introduction

The central part of the Flin Flon belt, stretching from Amisk Lake to Athapapuskow Lake (Figs. 2, 4), contains the full range of 1.92-1.88 Ga tectonostratigraphic assemblage types, 1.87-1.84 Ga post-accretion arc rocks and successor basin deposits, and structures related to all major deformation events (D_1-D_5) that are found across the belt as a whole. It is marked by relatively little deformed, low metamorphic grade (sub-greenschist to greenschist facies: Bailes and Syme, 1989; Digel and Gordon, 1993, 1995) rocks that have been mapped in significant detail over the past 100 years (see Harrison, 1951; Bailes and Syme, 1989; Syme *et al.*, 1993; and references therein). In addition, it contains the largest Cu-Zn VMS deposit in the belt (see below) as well as numerous smaller but economic deposits that are uniquely associated with juvenile arc assemblages (Syme and Bailes, 1993; Reilly, 1995). The field trip stops have been arranged to (1) present a geologic cross-section of the area from the Mystic Lake assemblage in the west to the Elbow-Athapapuskow assemblage in the southeast (Figs. 4; 11), (2) unravel the tectonic history of the area, and by analogy the Flin Flon belt as a whole, by examining critical stratigraphic, magmatic and structural relationships for time intervals descending from 1.90 to 1.80 Ga in successive stops (Fig. 3); and (3) document the geologic and tectonic context for the VMS deposits at Flin Flon proper and throughout the Flin Flon arc assemblage. Key issues that will be continuously referred over the course of the field trip are the age and tectonostratigraphic affinity of units, the local structural context, and the relations between local, regional and crustal-scale structures as understood through the LITHOPROBE seismic reflection profiles.

Stops in the Flin Flon area (Fig. 4) that will primarily focus on aspects of the 1.92-1.88 Ga tectonostratigraphic assemblages and VMS deposits include:

- juvenile arc rocks of the Flin Flon assemblage (Stops 1-3, 6, 7, 8)
- stratigraphic setting of the Flin Flon mine (Stops 1-2)
- MORB-like (arc-rift?) rocks of the Scotty Lake section, Flin Flon arc assemblage (Stop 10)
- juvenile ocean floor (back-arc) rocks of the West Arm and Elbow-Athapapuskow assemblages (Stops 4, 11)

Stops in the Flin Flon area that will address the D_1 accretion event (1.88-1.87 Ga) and overprinting D_2 deformation and magmatism (1.87-1.84 Ga) are:

- Meridian-West Arm and Mystic Lake shear zones (Stop 5)
- Northeast Arm shear zone (Stop 9)

Post-accretion arc plutons (1.87-1.84 Ga) and successor basin sedimentary rocks (*ca.* 1.845 Ga) will be highlighted in several stops:

- 1850 Ma Reynard Lake pluton (Stop 5)
- Missi suite sedimentary rocks and the sub-Missi unconformity and paleosol (Stop 12)

Collisional structures (1.84-1.80 Ga) within the Amisk collage will complete the Flin Flon portion of the field trip and set the stage for the first two Snow Lake area stops:

- D_3 structures along the Northeast Arm shear zone (Stop 9)
- Fold-foliation relations in the Missi suite (Stop 12)
- D_4 Annabel Lake shear zone (Stop 13)

VMS deposits in the Flin Flon assemblage

The Flin Flon greenstone belt is one of the largest Proterozoic volcanic-hosted massive sulphide (VMS) districts in the world, in which more than 160 million tonnes of sulphide has already been mined from 24 deposits (Fig. 2). All of the mined sulphide deposits in the Flin Flon belt are associated with the juvenile arc volcanic rocks (Syme and Bailes, 1993), whereas the ocean-floor assemblage contains no economic deposits (although several important showings are known). Knowledge of the physical and geochemical characteristics of the assemblages in the Amisk collage is thus crucial for effective base metal mineral exploration. The physical volcanology of the juvenile arc assemblage is dominated by subaqueous effusive and volcaniclastic components (documented in Bailes and Syme, 1989), and will be demonstrated at Stops 1-3, 6-8, and 10. For a detailed discussion of arc geochemistry, see Stern *et al.* (1995a), Syme and Bailes (1993), and Syme (1990).

VMS deposits in the arc assemblage at Flin Flon (Fig. 4) occur in tholeiitic and calc-alkaline suites dominated by basalt and basaltic andesite. The VMS deposits are stratigraphically associated with rhyolite, at major stratigraphic and compositional "breaks". The breaks can be recognized by contrasting major element, trace element and isotopic characteristics of the underlying and overlying mafic rocks. Most VMS deposits are underlain by mafic, intermediate or felsic volcaniclastic rocks, and have discordant footwall chloritic alteration zones. Together, these characteristics suggest that VMS deposits occur in depositional basins produced by intra-arc extension or rifting; examples include:

1. In the early primitive arc Beaver Road/Hidden-Burley suites, evidence for intra-arc extension is manifest by 'breaks' in the mafic stratigraphy that contain local unconformities and volcaniclastic/epiclastic rocks. Rhyolites and associated VMS deposits (*e.g.*, Flin Flon and Callinan deposits (discussed below); Syme and Bailes, 1993) also tend to occur at such breaks, suggesting that arc rifting played a role in the generation of these Cu-Zn deposits.
2. In the Bear Lake section, a calc-alkaline stratovolcano/caldera sequence hosting the Cuprus and White Lake VMS deposits is overlain by a unit of submarine ferrobasalt (1886 Ma, Gordon *et al.*, 1990). The ferrobasalts have an N-MORB geochemical signature (Stern *et al.*, 1995b, unpublished data) and are interpreted to be related to arc rifting. The Cuprus and White Lake deposits are hosted by graphitic sedimentary rocks, and are overlain directly by the basaltic arc-rift sequence. Deposition of the VMS deposits thus is linked in time to the initiation of arc rifting.

Stop descriptions

Stops 1, 2: Stratigraphic setting of the Flin Flon and Callinan VMS deposits

The host stratigraphy of the past-producing Flin Flon mine (Fig. 12) and newly-producing Callinan deposit will be examined in two localities. Stop 1 ('Millrock Hill', Figs. 4, 12, 13) is a section through mine stratigraphy located south of the orebody. A key felsic volcaniclastic unit in this section has features which suggest a relatively proximal depositional setting. Stop 2 (northeast of the HBMS smelter) lies between the Flin Flon and Callinan deposits, and displays a more distal facies in the same volcaniclastic unit that occurs on Millrock Hill. These relationships, and important map-scale facies relationships in the footwall sequence, demonstrate that the Flin Flon massive sulphides accumulated in a volcaniclastic basin.

The Flin Flon deposit was the largest VMS deposit in the Flin Flon belt (62.4 million tonnes), an order of magnitude larger than other deposits. The Flin Flon and Callinan deposits occur at the stratigraphic contact between the Beaver Road and Hidden-Burley suites (Table 2), both of which are early tholeiitic members of the primitive oceanic Flin Flon arc assemblage. The sulphides are contained within a 200 m thick package of rhyolitic flows and breccias (Mine rhyolite) between two basalt formations: South Main basalt in the stratigraphic footwall and Hidden Lake basalt in the stratigraphic hanging wall (Fig. 14, 13). A discordant zone of chlorite-rich altered rocks occurs stratigraphically beneath the massive sulphide orebody.

South Main basalt in the stratigraphic footwall of the Flin Flon deposit is at least 700 m thick (Figs. 13, 14). It represents a relatively shallow water volcanic construct, composed of distinct proximal (flow) and basin (breccia) facies, wherein pillowed flows display an abrupt lateral transition northwards into pillow fragment breccia. The basaltic breccias underlie the massive sulphide deposit and are overprinted by the footwall hydrothermal alteration zone. A plausible explanation for this facies relationship, and an angular unconformity exposed in the footwall succession, is that an intra-arc rift associated with arc extension is recorded in the footwall sequence. In this model, the Flin Flon and Callinan deposits are located in a rift basin filled with volcaniclastic rocks (Fig. 14).

Rhyolitic rocks in the mine stratigraphy include two distinct associations. The older of the two are complex rhyolitic bodies which cut across flow contacts in South Main basalt and are interpreted as high-level domes (South Main rhyolite domes: Syme and Bailes, 1993; Fig. 14). The domes have bulbous upper portions up to 150 m across connected to narrower feeder dykes. Features indicating near-surface emplacement include the presence of pumice breccias in the tops of the domes and flow banding parallel to dome margins. The younger, and more important, felsic body is a composite rhyolite flow termed Mine rhyolite (1904 +6/-3 Ma; David *et al.*, 1993). Mine rhyolite is the principal component of the heterogeneous package of rocks between the footwall (South Main) and hanging wall (Hidden Lake) basalt successions (Fig. 14), and hosts the Flin Flon and Callinan VMS deposits (Price, 1978; Koo and Mossman, 1975).

The Flin Flon orebody (now mined out) contained 62.4 million tonnes of sulphide grading 2.2% Cu, 4.1% Zn and 2.6 g/tonne Au. The sulphide orebody was

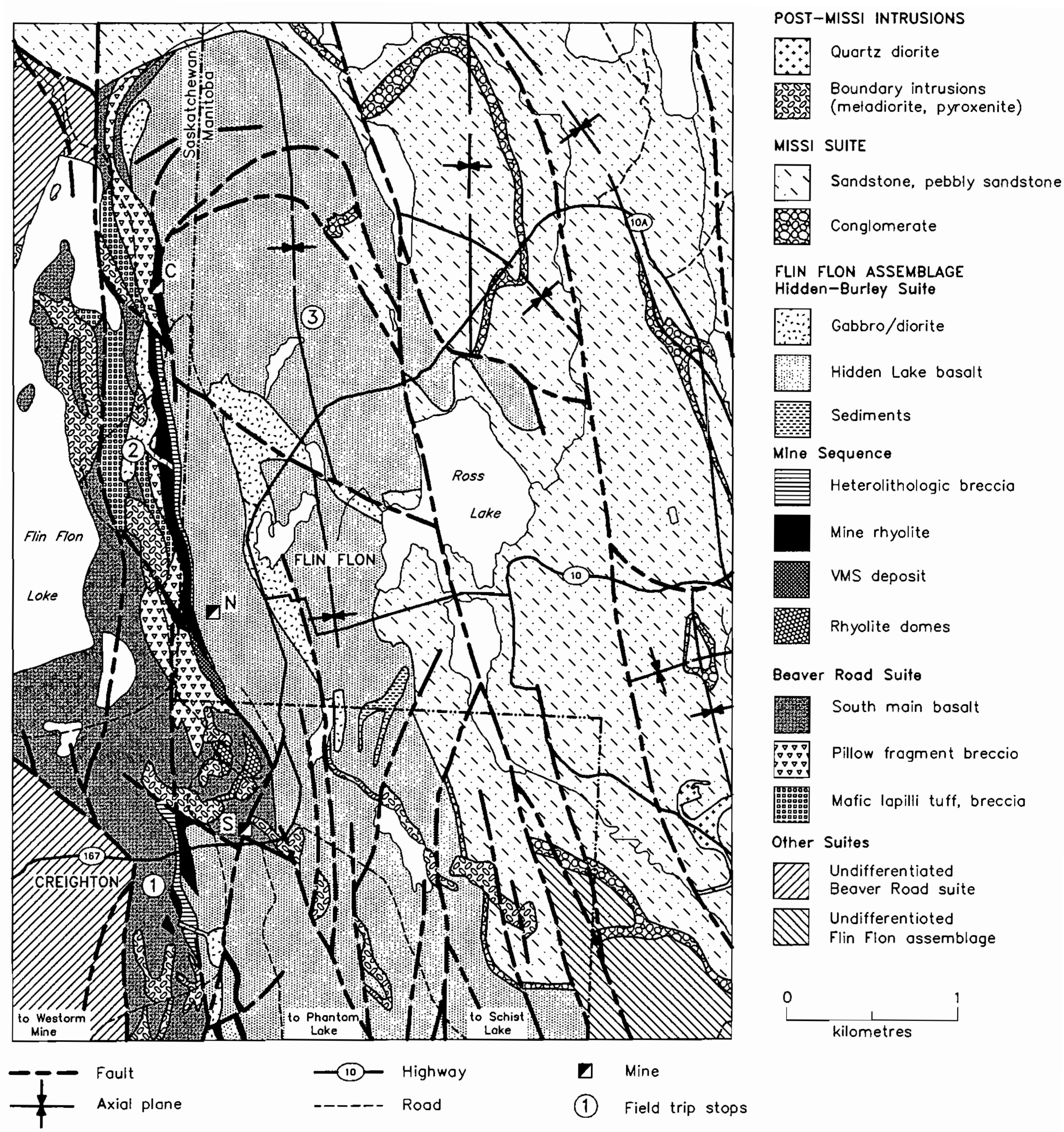


Figure 12: Simplified geology of the Flin Flon - Creighton area, showing the stratigraphic sequence hosting the Flin Flon and Callinan VMS deposits. In the stratigraphic footwall are South Main basalt and equivalents (dark), pillow fragment breccia and mafic volcaniclastic rocks (patterned). Rhyolite (black) dominates the heterogeneous package of felsic rocks hosting the massive sulphide deposits. Hidden Lake basalt (grey screen) forms the stratigraphic hanging wall of the deposits. The mine stratigraphy is folded about the NNW-trending Hidden Lake synform. S: South Main shaft, Flin Flon mine; N: North Main shaft, Flin Flon mine; C: Callinan deposit. Stop 1: Millrock Hill; Stop 2: NW of HBMS smelter; Stop 3: Hidden Lake. The NW-trending fault shown offsetting stratigraphy south of South Main shaft is conjectural. It is only one of several solutions that have been proposed to explain the stratigraphic relations between South Main shaft and Millrock Hill, in an area where bedrock is covered by infrastructure (roads, mine buildings, aggregate plant; see also Thomas, 1994). Modified from Syme et al. (1993).

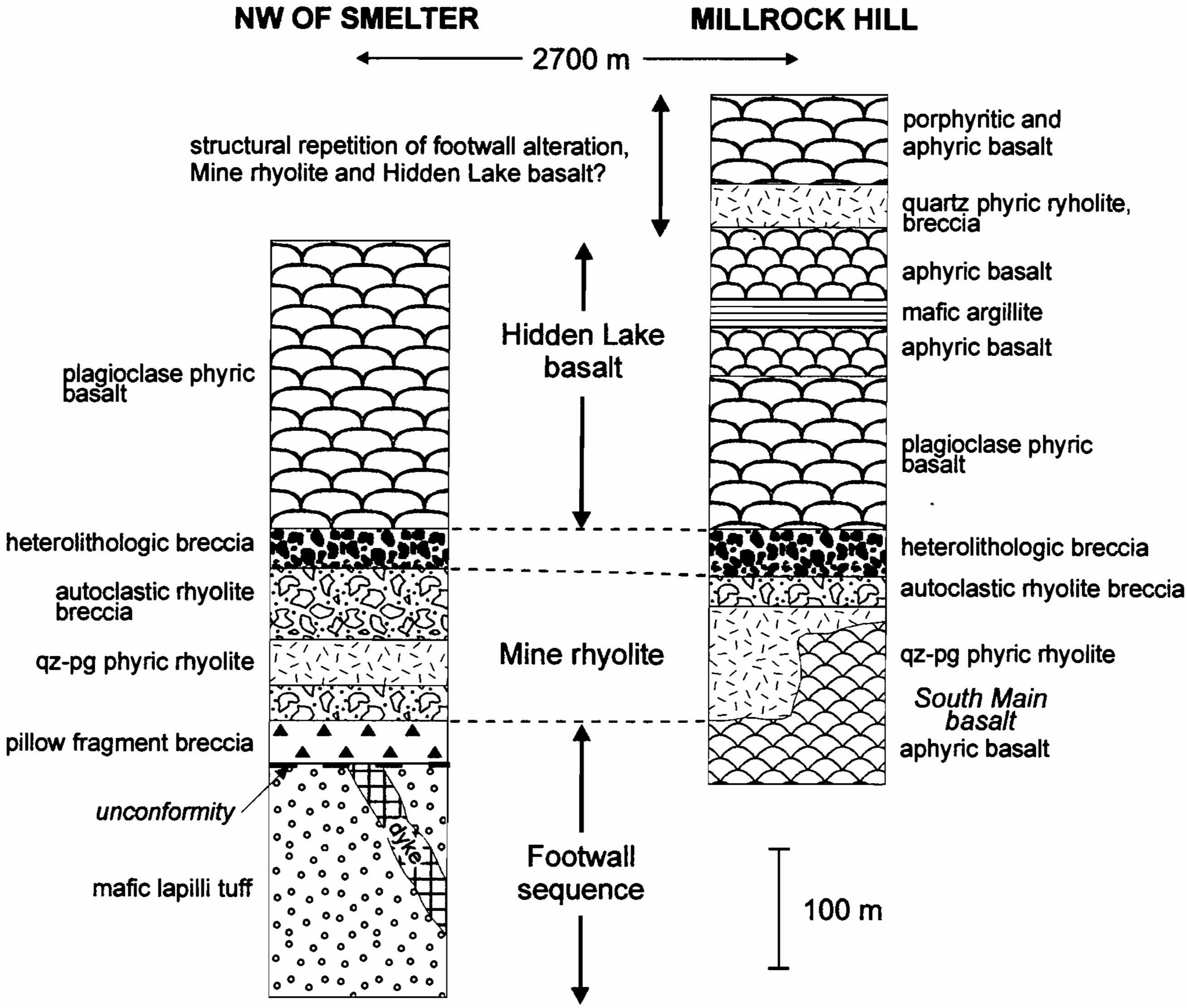


Figure 13: Comparative stratigraphic sections for 'Millrock Hill' (Stop 1; modified from Thomas, 1994) and northwest of the HBMS smelter complex (Stop 2; after Bailes and Syme, 1989). The Flin Flon Cu-Zn massive sulphide VMS deposit is hosted by Mine rhyolite. The footwall sequence on 'Millrock Hill' is proximal (effusive), whereas that NW of the smelter is basinal (volcaniclastic). Heterolithologic breccias overlying Mine rhyolite (the so-called "millrock") show a proximal to distal facies relationship parallel to that displayed by the mafic rocks in the stratigraphic footwall of the deposit.

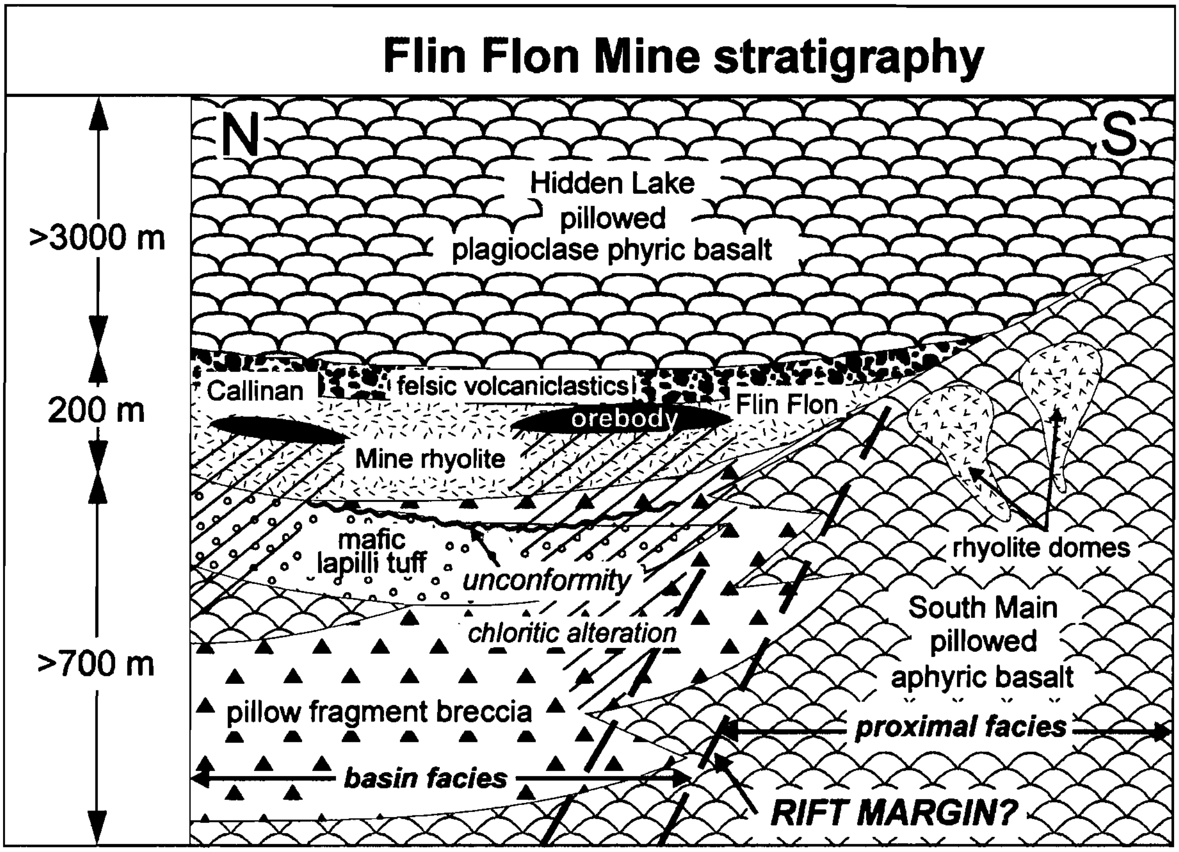


Figure 14: Schematic representation of stratigraphic relationships for the juvenile arc package that hosts the Flin Flon and Callinan VMS deposits. South Main basalt and Hidden Lake basalt are members of the tholeiitic Beaver Road and Hidden-Burley suites, respectively. A clear facies transition between proximal (effusive) and basinal (volcaniclastic) facies exists in South Main basalt, between South Main shaft (equivalent to the right side in this diagram) and northwest of the smelter (left). An intra-arc rift is one explanation for the facies distribution and angular intra-volcanic unconformity in the footwall rocks. Rift-bounding fault(s) may have provided pathways for the hydrothermal fluids responsible for the massive sulphides. Modified from Syme and Bailes (1993).

predominantly massive with chalcopyrite concentrated near the base and sphalerite, sometimes banded with pyrite, more common toward the hanging wall (Byers *et al.*, 1965; Howkins and Martin, 1970). The six ore zones strike 330° and dip 70° east for 1650 m down plunge. The zones average 270 m long, 21 m thick and 450 m in vertical extent. An extensive alteration zone occurs in the footwall, composed of chlorite with lesser amounts of sericite, talc and carbonate (Koo and Mossman, 1975). These altered rocks contain relict amygdales and quartz phenocrysts showing they were originally mafic volcanic rocks and rhyolite. Disseminated chalcopyrite and pyrite occur in the altered footwall rocks and formed approximately 30% of the ore mined (Gale and Eccles, 1988). This mineralization occurred below the massive sulfides, in a stringer zone hosted by Mine rhyolite and South Main breccia. Metal ratios in massive and disseminated ores are similar (Koo and Mossman, 1975).

A thin unit of heterolithologic rhyolite breccias overlies Mine rhyolite. At Stop 1, this breccia is coarsely fragmental and is interpreted as a series of thick debris flow beds. Rhyolite fragments are highly angular and may have been produced by phreatic explosions. At Stop 2 the volcaniclastic unit is thinly bedded and composed of much finer grained volcaniclastic material. The north to south, proximal to distal facies relationship exhibited by this unit parallels the pillowed flow - volcaniclastic facies relationship in South Main basalt (Fig. 13).

The 3.3 km thick Hidden Lake-Burley Lake sequence represents resumption of mafic volcanism following the development of the proposed rift basin, emplacement of the Mine rhyolite and VMS orebody (Figs. 14, 13). The sequence is subdivided into lower and upper members (Bailes and Syme, 1989). The lower member of the sequence (Hidden Lake basalt) is dominated by basalt flows, while the uppermost member of the sequence (Burley Lake basaltic andesite) includes basalt, basaltic andesite and rare andesite. Major element variation in this sequence forms a continuous magmatic trend (Fig. 6), with the LREE-enriched Burley Lake flows being the more differentiated, upper part of the sequence. Samples from the upper part of the footwall Beaver Road suite have higher FeO^{*}/MgO ratios and lower MgO than basal flows in the Hidden Lake-Burley Lake suite. The fact that the two basalt formations are chemically distinct is consistent with stratigraphic evidence of a hiatus in basaltic volcanism at the interval associated with the Flin Flon massive sulphides.

Structural studies in the Flin Flon Mine area show no evidence for major faults along the Flin Flon-Callinan ore horizon (Gale *et al.*, 1996). Although deformed and disrupted by mesoscopic and map-scale structures related to all deformation episodes, the stratigraphy of the ore horizon sequence appears to be essentially 'intact' from footwall to hanging wall (*cf.* Thomas, 1994). The significance of this is that Mine horizon can be traced from Callinan along strike past the Flin Flon mine and to the south, where it may be folded by a large southeast-plunging Z fold and is cut by a number of discrete faults (Thomas, 1994; Gale *et al.*, 1996).

Stop 1: Flin Flon Mine horizon, Millrock Hill

Thomas (1994) mapped Millrock Hill in detail (1:5 000 scale) to examine the stratigraphic relationship between several rhyolitic bodies and the productive Mine rhyolite to the north. The composite stratigraphic section of Millrock Hill (Fig. 13), the description of the geological units, and stratigraphic/structural relationships are summarized below by Brian Reilly (formerly with the Saskatchewan Geological Survey), from Thomas (1994) and additional observations by Reilly (1995).

Stratigraphy

The composite stratigraphic section (Fig. 13) consists of, in ascending stratigraphic order, South Main basalt, rhyolitic rocks that are interpreted to represent the stratigraphic equivalent of the Mine rhyolite, heterolithologic volcaniclastic rocks, and Hidden Lake basalt.

South Main basalts weather dark green to brownish green, but are rusty brown to dull green where altered (Bailes and Syme, 1989). The lower part of the section contains up to 140 m of aphyric pillow flows, flow breccias, and amoeboid pillow breccias, and subordinate plagioclase phyric flows. Pillows are highly amygdaloidal, locally exhibiting a bimodal amygdale size distribution and pipe amygdales along pillow margins, as at South Main shaft (Bailes and Syme, 1989, Syme and Bailes, 1993).

Overlying the South Main basalts are a number of separate rhyolitic bodies. Quartz (± plagioclase) porphyries up to 75 m thick form cross-cutting to conformable units that display a general upward increase in the abundance and size (up to 10 mm) of phenocrysts. The rhyolites contain flow banding and local pumice-rich upper zones. Coarse autoclastic rhyolite breccias, up to 20 m thick, locally overlie massive rhyolite. The presence of intrusive and conformable contacts, flow-like features and pumice suggest these rhyolites may represent high-level cryptodomes, locally transitional into eruptive flows. This unit is correlated with Mine rhyolite (Fig. 13; Stop 2).

'Millrock' volcanic breccia (a generic term for coarse felsic fragmental rocks: if one is standing on such breccias, a mine mill is commonly in view), up to 33 m thick overlies the porphyritic rhyolite. It contains a heterolithologic clast population including abundant block-sized, quartz phyric rhyolite and subordinate mafic fragments, supported by a fine-grained mafic matrix. Variations in the abundance and size of rhyolite fragments defines crude layering and grading. The rhyolite fragments are typically slab-like and highly angular. Re-entrant fractures displayed by some fragments suggest the unit may be the product of episodic phreatic explosions (by analogy with features in the Joliette breccia, Noranda; Dimroth and Rocheleau, 1979).

The heterolithologic 'Millrock' breccia is overlain by up to 128 m of plagioclase phyric pillowed basalt flows, succeeded by 20 to 34 m of aphyric massive to pillowed basalt flows and flow breccias (Fig. 13). These basalts are distinguished from the South Main basalts by their distinctive buff weathering color, absence of pipe vesicles, absence of hydrothermal alteration, abundance of plagioclase phenocrysts, and lower specific gravity. These light-coloured basalts can be correlated on the basis of lithology and stratigraphic position with Hidden Lake basalt in the stratigraphic hanging wall of the Flin Flon massive sulphide deposit (Fig. 13; see also Stop 3).

A unit of finely laminated to massive, fine-grained mafic to intermediate volcanic argillites with local normal grading and cross-lamination occurs within the Hidden Lake section on Millrock Hill. These are well exposed in the parking lot at the South Main shaft. In the area around the aggregate crushing plan southeast of the South Main shaft, the argillites are overlain by approximately 60 m of chloritic, altered aphyric basalt flows and flow breccias. Alteration increases stratigraphically upward towards the base of a quartz phyric massive rhyolite between Millrock Hill and the West Arm Mine road. This rhyolite, which is up to 24 m thick, is interpreted to be stratigraphically equivalent to the Mine Rhyolite (D. Price, Hudson Bay Exploration and Development, pers. comm.). The rhyolite is overlain by intercalated aphyric and plagioclase phyric pillow basalt flows; however, a thin unit of heterolithologic mafic breccia containing small rhyolite clasts is present above the rhyolite unit at one location. This bedded fragmental rock is similar to the Railway volcaniclastic unit (Stop 2; Bailes and Syme, 1989; Thomas, 1992). The presence of a package of chloritic altered basalt, quartz phyric rhyolite, and Railway volcaniclastics within the interpreted hanging wall Hidden Lake basalt suggests structural repetition (Fig. 13). This will be discussed further in the following section.

Structure

At a macroscopic scale, the earliest penetrative fabric (S_1) is variably preserved at Millrock Hill as a moderately- to well-developed schistosity intersecting primary layering. The S_1 foliation strikes northeast and dips shallowly to the south, and appears to be axial planar to an early set of tight to isoclinal recumbent minor folds (D_1) which are best defined by sets of folded quartz veins observed elsewhere in the region. Early shear zones, subparallel to the S_1 fabric and ranging between a few centimetres and several metres wide, suggest an overall reverse or thrust movement to the north. Bulk displacement appears to be small on these structures, which are probably multiple high strain zones developed on strands along D_1 fold limbs.

A second penetrative foliation (S_2), which typically dips moderately to steeply east-northeast and locally moderately to the west, is variably developed. The S_2 fabric is axial planar to the southeast-plunging, northerly-trending, upright D_2 Hidden Lake Syncline. Along the limbs of D_2 folds, the earlier S_1 foliation is progressively transposed parallel to the S_2 fabric, resulting in a composite S_1/S_2 foliation. S_1/S_2 relationships are shown in South Main basalt amoeboid pillow breccias and in fragments of the heterolithologic breccia. Similarly, early S1 shears are folded around D_2 structures and transposed on D_2 fold limbs. Northwest-trending (S_2) and north- to northeast-trending (S_3) shears show predominantly sinistral reverse and dextral normal movement, respectively. These shears clearly disrupt stratigraphy on a mesoscopic scale.

A moderately southeast-plunging S_2 extensional lineation pervasively developed throughout the area is broadly collinear to S_1/S_2 intersection and D_2 fold axes. This composite linear fabric is manifested by mineral and mineral aggregate domains, rodding of clasts, amygdales, and pillows and boudin axes.

The poor strike continuity and local repetition and of lithologic units in the Millrock Hill area is interpreted to be a result of polyphase folding and extensive faulting. Quartz phyric rhyolite bodies occurring between Millrock Hill and the Flin Flon Mine open pit follow stratigraphic layering at a macroscopic scale on the west limb of D_2 Hidden Lake syncline and appear to be stratigraphically equivalent. The east limb of this D_2 fold may appear east of Ross Lake where a west-facing sequence of altered rhyolite is overlain by a heterolithologic mafic volcanic breccia containing rhyolite fragments, similar to the Railway volcaniclastic unit and Millrock breccia outcrops. The possibility that the orebodies were structurally repeated during D_1 folding remains unclear.

Stop 2: Flin Flon Mine horizon, northwest of the HBMS smelter complex.

This section (Fig. 13) contrasts in a number of ways with the section at Millrock Hill (Stop 1):

- the footwall basaltic rocks are fragmental rather than effusive in character,
- Mine rhyolite is clearly effusive, and composed of 2-3 flow/flow breccia units,
- the overlying heterolithologic volcaniclastic unit is thinner bedded and finer grained than at Stop 1.

Evidence for an angular unconformity in the footwall volcaniclastic section will be examined, relating to the theme of arc extension, the production of intra-arc basins, and deposition of massive sulphides.

Plagioclase phyric mafic volcaniclastic rocks

The local stratigraphic sequence at Stop 2 begins with a thick bedded unit of yellow-buff weathering, plagioclase phyric mafic volcaniclastic rocks comprising interbedded plagioclase-rich lapilli tuff and heterolithologic breccia (Figs. 13, 14). The breccia beds are poorly sorted, fragment-supported, and vary with respect to abundance of plagioclase phenocrysts and size of fragments. Primary features in this unit are considerably obscured by recrystallization and alteration. Similar coarsely plagioclase phyric mafic volcaniclastic rocks occur elsewhere in the stratigraphic footwall of the Flin Flon deposit (Bailes and Syme, 1989; Thomas, 1994; Syme *et al.*, 1993). These rocks may be unrelated to South Main basalt, produced from a separate volcanic centre, and remobilized as subaqueous density flows into an intra-arc basin.

An angular unconformity between the yellowish-weathering mafic lapilli tuff and the overlying, green-weathering South Main basalt pillow fragment breccia (Fig. 13) is exposed on two outcrops. The unconformity surface is sharp, unfaulted and truncates bedding in the tuff at a high angle. Thin (1 m) mafic sills in the bedded tuff and breccia are also clearly truncated at the unconformity; the dykes have chilled margins and amygdaloidal cores. The unconformity may represent a period of synvolcanic faulting contemporaneous with the period of massive sulphide deposition, and may have served to focus the discharge of metal-bearing hydrothermal solutions.

Pillow fragment breccia

Pillow fragment breccia forms a unit up to 400 m thick directly underlying the Flin Flon orebody (Fig. 14). A distinct lateral facies change in South Main basalt has been mapped between South Main shaft and this locality (Bailes and Syme, 1989). South Main pillowed basalt flows grade, through a transition zone of intercalated amoeboid pillow breccia, pillow fragment breccia, and subordinate thin pillowed flows north of the orebody, to pillow fragment breccia beneath Mine rhyolite south of the orebody. Petrographically, fragments in the pillow fragment breccia are identical to South Main basalt, clearly indicating that these units are facies equivalents (Bailes and Syme, 1989).

The breccias weather dark green, are monolithologic, poorly sorted, clast supported and massive, without recognizable bedding contacts. Fragments (up to 40 cm) are subangular to rounded, equant, and variably amygdaloidal. Range in amygdale size and abundance is identical to those in the South Main basalt.

Stratigraphically beneath the Flin Flon orebody the breccias are variably altered (Fig. 14). In "least altered" varieties the matrix is chlorite-rich and fragments vary from relatively unaltered to partially silicified or epidotized. With increasing intensity of alteration the fragments also are strongly chloritized. Other forms of alteration include epidotization and local strong sericitization accompanied by 5-20% disseminated pyrite. Surface exposure within the alteration zone is very poor; the most detailed information comes from studies conducted in mine workings (*e.g.,* Koo and Mossman, 1975).

Mine rhyolite

A porphyritic, light yellow-weathering rhyolite flow composed of two or three distinct phases overlies the pillow fragment breccia (Figs. 13, 14) and hosts the Flin Flon orebody (Price, 1977; Koo and Mossman, 1975). The rhyolite has been traced on surface, in underground workings, and in exploration drilling for approximately 3 km. It is a maximum of 150 m thick northwest of the smelter complex and thins to both north and south.

The basal portion of the rhyolite is a breccia composed of angular to subrounded rhyolite fragments. The rhyolite contains 1-2% euhedral quartz phenocrysts (0.4-1.2 mm) and 4% euhedral tabular plagioclase phenocrysts (0.4-2 mm). The main phase of the Mine rhyolite flow contains 3-4% euhedral quartz phenocrysts (0.2-2.5 mm) and 7% euhedral plagioclase phenocrysts (0.2-1.2 mm). It is readily distinguished on outcrop from the basal phase by the presence of quartz phenocrysts 2 mm or more in diameter. The main phase has two components, a lower zone of flow banded massive rhyolite (43 m thick) and an upper unit of rhyolite flow breccia. The flow breccia contains tabular slabs of flow banded rhyolite ranging up to 1.9 m long by 20 cm wide with the same phenocryst population as the massive rhyolite. These flow banded slabs comprise 5-20% of the breccia and have a more or less consistent attitude, striking 190° and dipping 55-60° west. The slabs are in a rhyolite breccia matrix composed of angular to subangular rhyolite fragments in a dark grey rhyolite matrix.

Koo and Mossman (1975) have observed that the orebodies of the Flin Flon massive Cu-Zn sulphide deposit occur within, as well as directly above or below, the Mine rhyolite. Within the mine they outlined a chloritic alteration zone 2000x1000x200 m which transects the underlying pillow fragment breccia and Mine rhyolite, and terminates against Hidden Lake basalt in the stratigraphic hanging wall (Fig. 14).

Mine rhyolite and South Main domes are separate units and not part of a single rhyolite complex. Mine rhyolite has a relatively flat REE pattern [$(La/Yb)_n = 1.3$), with a moderate negative Eu anomaly ($Eu/Eu^* = 0.66$), whereas the South Main dome rhyolite has a flat profile with significantly higher absolute REE contents, and a large negative Eu anomaly ($Eu/Eu^* = 0.24$; Syme and Bailes, 1993). ε_{Nd} values of basalts and rhyolites throughout the Flin Flon section show limited variation, ranging from +3.2 to +4.3; the ε_{Nd} value for Mine rhyolite, +4.3, is the highest measured value (Stern *et al.*, 1992). At Snow Lake, 150 km east of Flin Flon, VMS deposits are similarly associated with periods of isotopically primitive rhyolitic magmatism (Stern *et al.*, 1992).

Heterolithologic volcaniclastic rocks

Bedded volcaniclastic rocks with abundant rhyolite fragments comprise a 35 m thick unit which overlies the Mine rhyolite ('Railway volcaniclastics', Bailes and Syme, 1989; Figs. 13, 14). Beds are 10 cm - 1 m thick, parallel sided to lenticular, and are defined by grain size differences in sandy beds and by variable fragment abundance in coarser beds. Some lenticular beds are clearly scours that cut underlying beds. Size grading, either normal or reverse to normal type, occurs in some beds. These bedforms suggest that the volcaniclastic rocks were deposited from subaqueous density currents.

In coarse beds the predominant fragment type is a yellow weathering porphyritic rhyolite corresponding to the upper phase of the Mine rhyolite. The rhyolite fragments are up to 1-18 cm long and are crudely lenticular in shape. A subsidiary fragment type is sparsely plagioclase phyric, amygdaloidal basalt containing 1% euhedral plagioclase phenocrysts. The rhyolite and basalt fragments are contained in, and supported by, a matrix composed of poorly defined rock and crystal fragments.

Railway volcaniclastic rocks at Stop 2 are stratigraphically equivalent to the 'Millrock' breccias at Stop 1 (Fig. 13). Both are in a broad sense subaqueous debris flow deposits, but they differ in their sedimentological characteristics, suggesting that they are different facies:

- the heterolithologic breccias at Stop 1 are considerably coarser grained and thicker bedded than the poorly sorted sandy-textured rocks at Stop 2, and are clearly a more proximal facies;
- rhyolite fragments at Stop 1 are highly angular. They could not have been transported for any great distance in a debris flow and retain delicate re-entrant angles. Fragments at Stop 2 are considerably less angular, likely due to abrasion during longer transport.

The proximal (Stop 1) to distal (Stop 2) relationship exhibited by these felsic volcaniclastic rocks parallels the effusive to fragmental facies relationship documented in the underlying South Main basalt (Fig. 13). The intra-arc basin containing these volcaniclastic rocks also contains the Flin Flon and Callinan deposits.

Stop 3: Hidden Lake basalt in the stratigraphic hanging wall of Flin Flon Mine

Introduction

Hidden Lake basalt is typical of juvenile arc assemblage subaqueous mafic constructs at Flin Flon. It conformably overlies the felsic volcaniclastic rocks at Stops 1 and 2. However, unlike units in the stratigraphic footwall of the Flin Flon deposit, Hidden Lake basalt displays no south to north facies variation (Fig. 14). It represents a new episode of mafic volcanism post-dating the arc extension and rifting responsible for the footwall volcaniclastic basin.

Hidden Lake basalt (Bailes and Syme, 1989) is the lowermost part of the Hidden-Burley stratigraphic sequence (>3 km thick; Fig. 5). Like other tholeiitic series rocks in the Flin Flon arc assemblage, it displays low abundances of Ti, Zr, Hf, Nb, Y, and middle and HREEs (Stern *et al.*, 1995a). These characteristics are likely due to their derivation from a mantle source that was more refractory than that assumed for modern N-MORBs. Hidden Lake basalt has Nd isotopic characteristics that are close to 'juvenile' in nature, suggesting that it could have incorporated, through subduction of sediments or intracrustal contamination, less than 1% of a recycled late Archean crustal component (Stern *et al.*, 1995a).

Hidden Lake basalt is highly proximal in nature (*cf.* Williams and McBirney, 1979), comprising thick pillowed flows, abundant synvolcanic mafic dykes and sills, and minor pillow fragment breccia, interflow tuff and sediments. The inferred depositional setting is marine: the general absence of intercalated phreatomagmatic breccias suggests that the flows were emplaced at depths >300 m (phreatomagmatic style dominates eruptions when water depth is less than about 300 m: Jones, 1979, 1970; Sigvaldason, 1968; Allen, 1979).

The pillowed flows commonly show irregular transitions from pillowed lava to amoeboid pillow breccia. Such breccias are composed of hyaloclastite containing abundant involuted ('amoeboid') pillows up to 1 m in length, elongate parallel to flow contacts. The pillows have complete selvages and commonly contain quartz-filled amygdales. The sandy-textured hyaloclastite is derived from *in situ* quenching and thermal fragmentation of the glassy rims on amoeboid pillows (Bailes and Syme, 1989). Amoeboid pillows are hyaloclastite-matrix supported, but there is evidence elsewhere that they be interconnected in the third dimension. Some amoeboid pillows are rooted in the pillowed (or massive) portion of the flow.

Concentrically banded lava tubes and synvolcanic basalt dykes with banded margins are common in Hidden Lake basalt. The banding in both lava tubes and dykes is interpreted to be the result of passage of multiple pulses of magma through the structures.

Structural setting

The Hidden Lake basalt sequence forms the core to a controversial regional fold structure termed the Hidden Lake synform (Fig. 15; Stockwell, 1960; Bailes and Syme, 1989). Recent work by Gale *et al.* (1996) has shown that the geometry of the Hidden Lake synform is not consistent with that of F_1 folds involving the Missi suite in the Flin Flon area (Flin Flon Creek, Pipeline; *cf.* Stauffer and Mukherjee, 1971), implying that the folding event responsible for the synform either pre-dates or post-dates F_1. There does not appear to be an expression of the Hidden Lake fold in the Missi suite sedimentary strata to the north of the Club Lake 'fault', although the F_1 Flin Flon Creek syncline has a concave-to-the-south axial trace (Stauffer and Mukherjee, 1971). The concave-to-the-south shape of the Club Lake contact, coupled with the southeast plunge of all basement and cover structures in the area (Fig. 15), requires that development of the Hidden Lake synform post-dated F_1. However, the regional S_2 foliation cuts across the hinge of the Hidden Lake synform. Gale *et al.* (1996) interpreted the Hidden Lake synform as an early F_2 structure, possibly coeval with

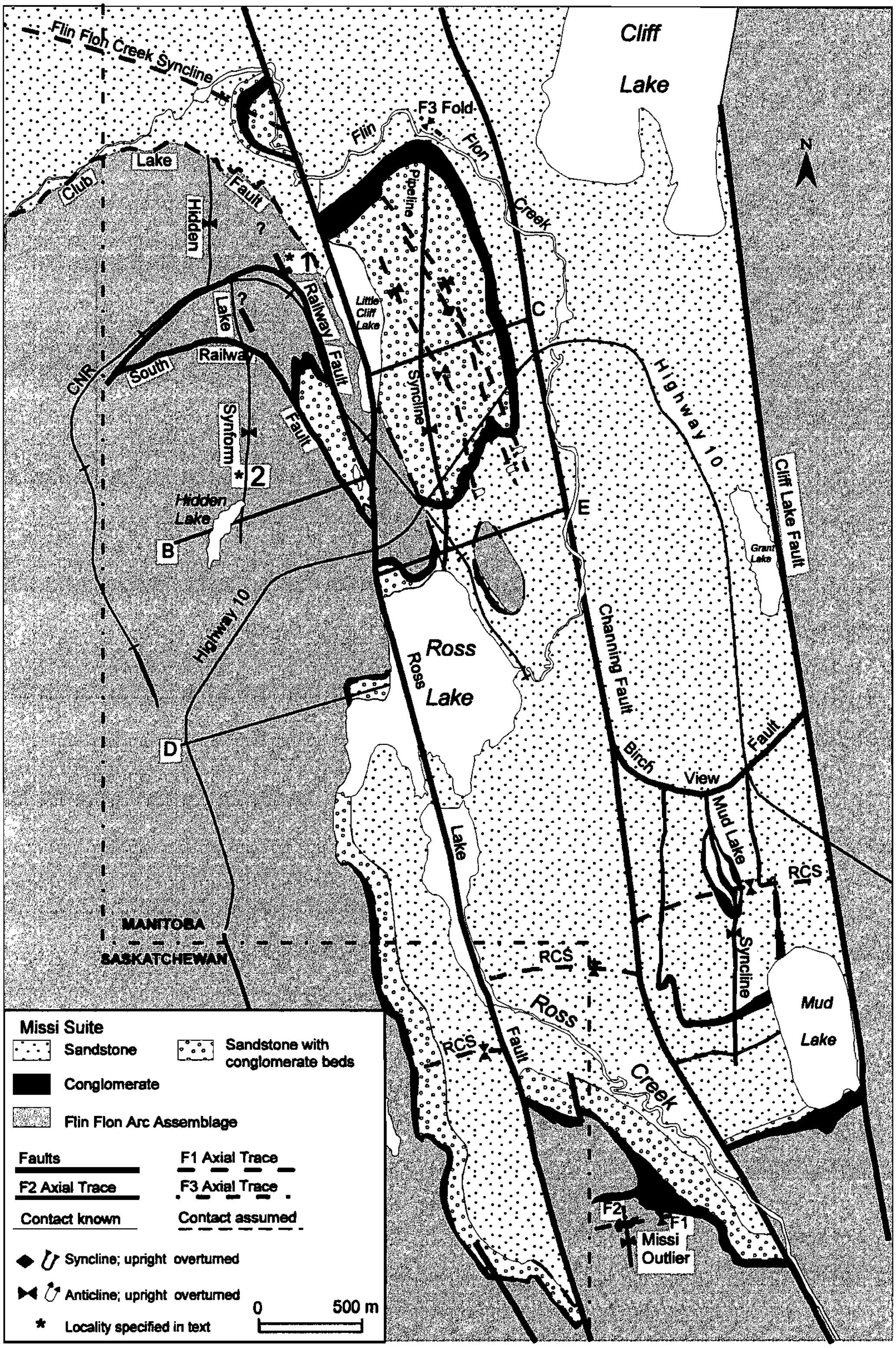


Figure 15: Structural geology of the Flin Flon area (Gale et al., 1996). Location of fold axes from Gale et al. (1996), Stockwell (1960), Bailes and Syme (1989), Stauffer and Mukherjee (1971) and Thomas (1992). RCS: Ross Creek syncline.

initiation of the F_2 folds in the Missi suite. The map pattern for the preserved Railway fault segments (Fig. 15) suggests that they initiated as F_1 thrust faults and were folded during early F_2 development of the Hidden Lake synform. The apparent presence of two adjacent F_2 synforms (Hidden Lake and Pipeline) without an intervening antiform probably results from later faulting (*e.g.*, Ross Lake-Little Cliff Lake faults, Fig. 15).

The Club Lake fault, separating the Hidden Lake basalts from structurally underlying Missi suite rocks, is best interpreted as a deformed unconformity (Fig. 15; Gale *et al.*, 1996). Its geometry requires F_1 recumbent folding of the basement-cover contact. Two packages of Missi suite rocks outcrop on the eastern limb of the Hidden Lake synform, isolated from the principal Missi basin to the east and south (Fig. 15). Documentation of the unconformity at the base of these two Missi outliers, as well as along the Club Lake fault, suggests that the basement-cover contact may not project significantly above the current erosion surface on the eastern limb of the Hidden Lake synform. The inferred geometry of the unconformity between the small Missi outlier (locality 1, Fig. 15) and the Club Lake fault trace suggests north or northeast vergence on the F_1 basement-cored nappe (Gale *et al.*, 1996), consistent with the asymmetry of the F_1 Flin Flon Creek syncline (Ambrose, 1936). Thrust-sense (*i.e.*, northward) shear may have occurred along the overturned basement-cover contact, but it is difficult to distinguish between basement on cover overlap due to folding or faulting.

Stauffer and Mukherjee (1971) estimated ~1067-1830 m of northwestward thrust displacement on the Club Lake fault. Recently, Hudson Bay Exploration and Development Co. Ltd. have drilled into Missi sandstones structurally beneath the Hidden Lake basalts in several holes (T. Baumgartner and D. Price, pers. comm., 1995). At locality 2 (Fig. 15), the basement-cover contact was encountered at a depth of 2100 m. Further drill hole information suggests that the contact dips moderately to the south or southeast, consistent with its dip along the east-west segment of the Club Lake fault. The amount of fold/fault overlap of basement on cover depends on the kinematics of deformation, which have not been fully resolved to date. However, Gale *et al.* (1996) suggest that if basement on cover overlap is primarily the result of F_1 folding (N/NE-verging), then overlap is in the range of 500-2000 m.

Stop description

Hidden Lake basalt underlies the large, elevated area of clean bedrock exposure north of the town of Flin Flon. This area was logged during the early days of mining in the area, and subsequent run-off removed virtually all of the thin soil cover. As a result the basalt is superbly exposed, allowing examination of complete lava flows and flow stratigraphy. From the parking area we will traverse 500 m north across pillow basalt outcrops to a locality directly northeast of Hidden Lake (Fig. 12). Here the volcanic stratigraphy trends northwest youngs to the southwest. The axial trace of the north-trending Hidden Lake synform lies within 100 m to the west. Three flows will be examined (Fig. 16):

1) The stratigraphically lowermost pillowed flow is >20 m thick, and displays both vertical and lateral (northwest to southeast) facies transitions from pillows to amoeboid pillow breccia. The lateral facies transition implies that the primary flow direction or paleoslope was to the southeast (in present day co-ordinates). In its original orientation

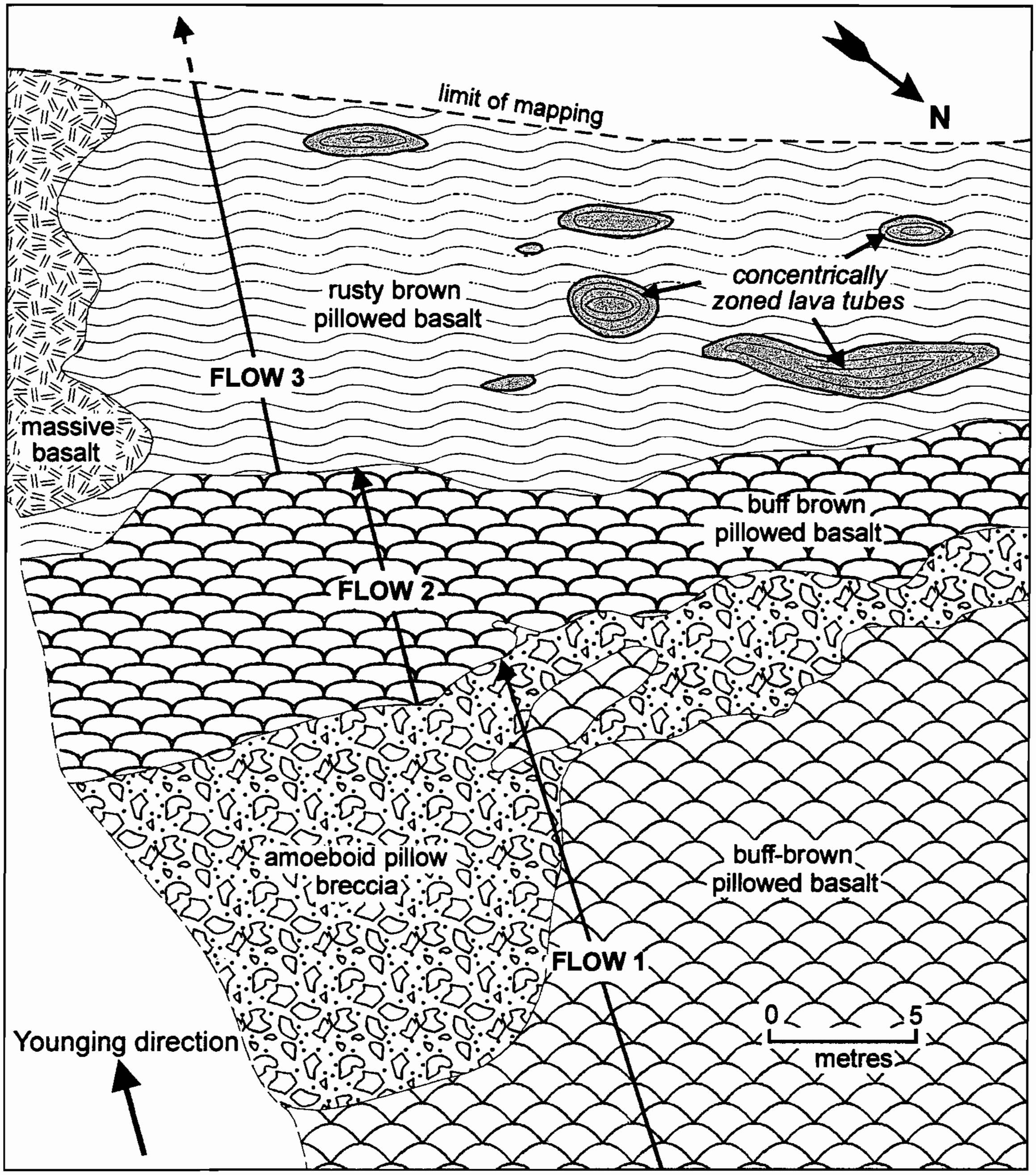


Figure 16: Outcrop map for Stop 3, Hidden Lake basalt. The local volcanic stratigraphy youngs to the southwest. Three pillowed basalt flows are exposed: 1) Flow 1 is >20 m thick, and displays a vertical and lateral (NW-SE) facies transition from pillows to amoeboid pillow breccia; 2) Flow 2 is 5-10 m thick and has no amoeboid pillow division; 3) Flow 3 is >15 m thick and has a distinctly darker weathering colour than underlying flows. Flow 3 has a SE-NW lateral facies transition from massive to pillowed material, and contains numerous concentrically banded lava tubes within its pillowed facies. See text for discussion.

this flow had a near-vertical 10 m scarp of pillowed material that abruptly passed laterally into amoeboid pillow breccia. Pillow fragments and complete pillows derived from the edge of the coherent, pillowed part of the flow occur in the amoeboid pillow breccia.

2) The overlying pillowed flow is relatively thin (5-10 m), and shows the internal zonation in pillow size that is typical for many of the arc assemblage basalt and basaltic andesite flows at Flin Flon. Pillows at the base of the flow are large and elongate ("mattress" pillows, Dimroth *et al.*, 1978), and grade upward in the flow to small pillows at the flow top.

3) The third flow in the sequence is a thick (>15 m) composite massive to pillowed variety, distinctly different from the underlying flows. It weathers dark rusty brown, in contrast to the buff-weathering underlying flows; the basal pillows are very large (megapillows); and the flow is characterized by the presence of several concentrically zoned lava tubes. Several of the lava tubes are exposed in oblique section (on horizontal outcrop surfaces) and in cross section (on steeply-dipping outcrop surfaces). The lava tubes plunge gently (15-25°) to the southeast (azimuth 150-160°), a geometry which suggests that they originate in the massive part of the flow. This flow displays a southeast to northwest lateral transition from massive to pillowed facies, suggesting a flow direction or paleoslope to the northwest. The composition, morphology and flow direction contrasts between this flow and the underlying flows suggests that they may have issued from separate vents on the sea floor.

A number of important aspects regarding the detailed structure of mafic flows are demonstrated by these three flows:

- amoeboid pillow breccias are not “mafic agglomerates” or debris flows. They are a common flow top facies observed in many subaqueous mafic flows. Amoeboid pillow breccia/hyaloclastite facies of flows can at least locally attain significant thickness (15 m in the example of Flow 1).
- flow orientation (strike and dip) in proximal arc mafic piles must be determined with caution, due to the large scale and irregular nature of the component flows. Contacts between pillowed facies and amoeboid pillow breccia facies of a flow are complex and can vary abruptly in attitude. Measuring the orientation of lithologic contacts (*i.e.*, pillows/breccia) <u>within</u> a flow is highly unreliable with regard to the orientation of volcanic stratigraphy (*e.g.*, Fig. 16, Flow 1). However, the contact between amoeboid pillow breccia and the pillowed portion of the <u>overlying</u> flow is generally parallel to volcanic stratigraphy (*e.g.*, contact between Flows 1 and 2).
- information on flow direction (*i.e.*, location of source vents) can be derived from detailed examination of flow contacts and facies. In proximal settings such as for Hidden Lake basalt, apparently contradictory information may result from the presence of multiple source vents, and/or seafloor topography (paleoslope) that is significantly modified with the emplacement of each new flow.

Stop 4: West Arm basalt

Tholeiitic basalt flows and synvolcanic gabbro-diorite sills are the chief components of West Arm basalt, a member of the juvenile ocean-floor assemblage south and west of Flin Flon (Fig. 4). Flows are massive to pillowed and are locally plagioclase phyric. Pillow breccias are locally common in this unit in Saskatchewan, but are rare in the Westarm Mine area of Manitoba. The basalts are medium to dark weathering, and contain upper greenschist to lower amphibolite facies mineral assemblages. West Arm basalt is in tectonic contact with the Flin Flon arc assemblage along the Meridian-West Arm shear zone (Fig. 4). The West Arm Cu-Zn massive sulphide deposit subcrops beneath Schist Lake and is hosted by argillites which are interpreted to lie within the Flin Flon arc assemblage. To the west, the West Arm assemblage exhibits increased deformation within the Mystic Lake shear zone and is in tectonic contact with the Mystic Lake assemblage. Stratigraphic facing in the basalts is uniformly southwest towards the Kaminis Lake pluton (Fig. 4).

At Stop 4, relatively undeformed pillowed basalts are intercalated with minor amygdaloidal sheet flows 1-1.5 m thick. The main features to observe include: amygdales of quartz ± carbonate, concentric thermal contraction fractures, radial pipe amygdales, and epidote ± quartz ± feldspar domains. These epidote-rich domains are typical of West Arm assemblage basalts and are generally concentrated in the cores of pillows or locally within the pillow interstices. The epidote-rich domains are interpreted as a product of hydrothermal seafloor alteration and are useful in identifying protoliths when the basalts are more highly strained.

Stop 5: Meridian-West Arm shear zone

Introduction

The structural relations between the Flin Flon, West Arm and Mystic Lake assemblages and 'stitching' intrusive rocks are well exposed in the Meridian Creek area, Saskatchewan (Fig. 17). Systematic bedrock mapping of this area was carried out by the Saskatchewan Geological Survey (Reilly, 1990, 1991, 1992; Thomas, 1989, 1990, 1991). Basalts and gabbros of the West Arm ocean floor assemblage are separated from diverse volcanic, volcaniclastic and intrusive units of the Flin Flon arc assemblage by the Meridian-West Arm shear zone. The West Arm assemblage rocks are in tectonic contact with evolved arc plutonic rocks of the Mystic Lake assemblage along the Mystic Lake shear zone, which engulfs the entire Mystic Lake assemblage and rocks of the Birch arc assemblage to the west. West Arm assemblage units appear to be progressively eliminated from south to north between the Mystic Lake and Flin Flon assemblages (Fig. 4). The timing of initial juxtaposition of these assemblages along the two shear corridors (D_1), as well as subsequent syn-magmatic deformation (D_2), is indicated by cross-cutting relations with various U-Pb dated intrusive rocks (Fig. 17; Stern and Lucas, 1994; Lucas *et al.*, 1996). Three outcrops will be examined at this stop: (1) Meridian-West Arm shear zone tectonites derived from West Arm assemblage units and intrusive sheets, (2) Mystic Lake shear zone tectonites derived from felsic intrusive sheets, and (3) weakly foliated Reynard Lake pluton (1850 ±1 Ma; Stern *et al.*, 1993) with screens of Mystic Lake shear zone tectonite.

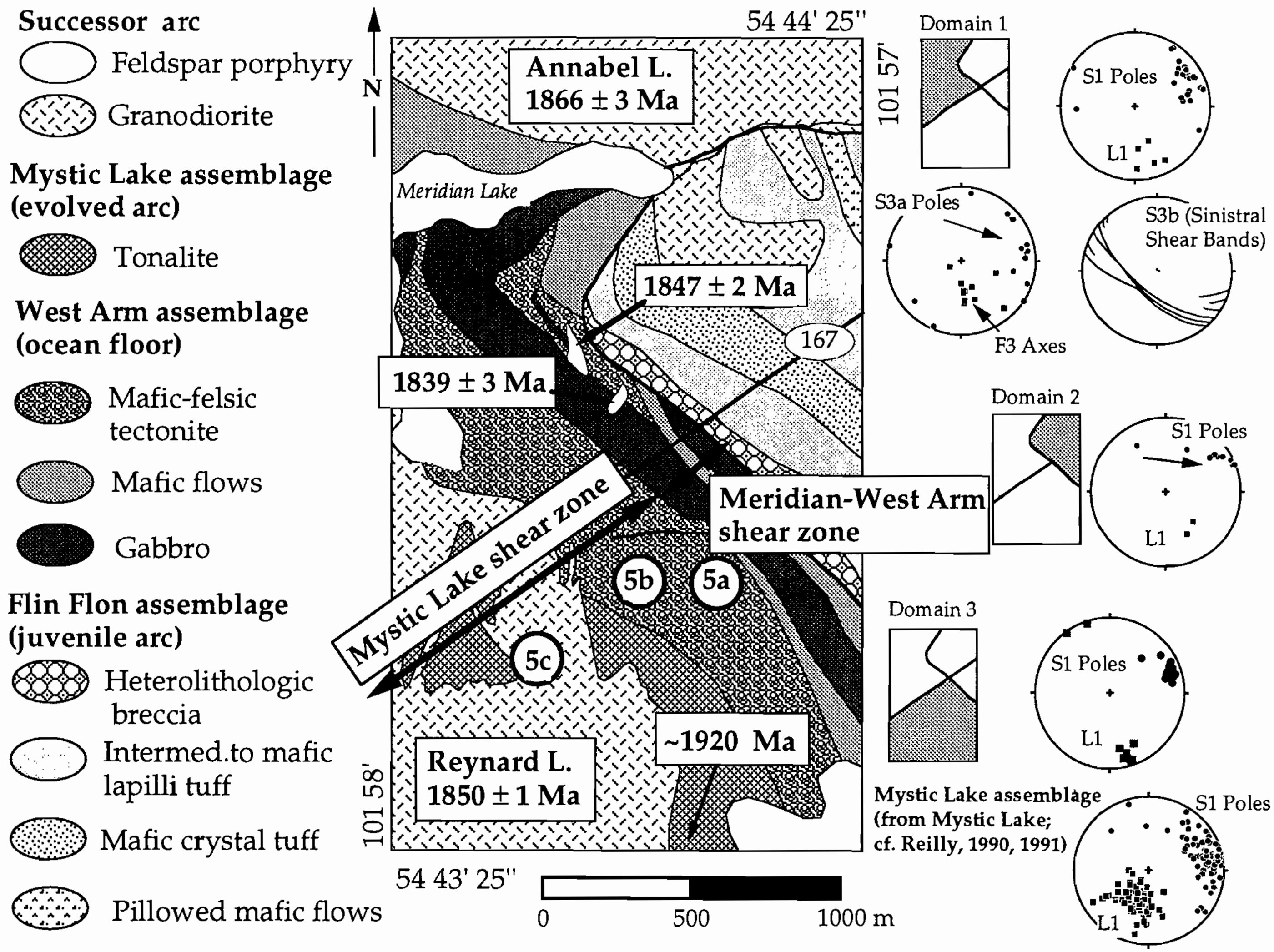


Figure 17: Detailed map of the Meridian Creek-Hutton Bay (Meridian Lake) area (Stop 5), after Thomas (1990, 1992). Meridian-West Arm shear zone includes westernmost 50 m of the Flin Flon arc assemblage and all of the West Arm assemblage units in the map area. The Meridian-West Arm shear zone merges with the Mystic Lake shear zone in the Mystic Lake assemblage units. U-Pb age determinations from Stern et al. (1993) and Stern and Lucas (1994). Equal area stereonets display structural data from Reilly (1990, 1991) and Lucas et al. (in prep.). Fabrics correlate with deformation events as follows: $S_1/L_1 = D_1$-D_2. Location maps for stereonet data show structural domains relative to Hwy. 167 and the boundary between the Flin Flon and West Arm assemblages south of Meridian Lake.

Both the Meridian-West Arm and Mystic Lake shear zones contain a SSE-striking, steeply west-dipping penetrative foliation (S_1) associated with D_1-D_2 deformation. It is defined by the crystallographic alignment of inequidimensional minerals (biotite, amphibole), "flattened" quartz grains and/or quartz-feldspar ribbons (Fig. 17). Within intensely deformed units, the S_1-parallel compositional layering is defined by sheared and/or isoclinally folded (*i.e.*, transposed) quartz veins and both mafic and felsic veins and sheets (Reilly, 1991). Boudinage of veins and compositional layers occurs in both horizontal and vertical planes, suggesting that the shear zones accommodated a component of flattening strain. Many of the felsic veins (tonalite, aplite, pegmatite) appear to have been emplaced during the ductile deformation event responsible for S_1 and the transposed layering (Lucas *et al.*, 1996). True mylonites occur in narrow (cm-scale) ductile shear zones in both felsic and mafic intrusive rocks (Reilly, 1991). S_1 is associated with a shallow to moderately SSE-plunging extension lineation (L_1, Fig. 17), generally defined by quartz-feldspar rods (polycrystalline aggregates) although locally by amphibole. Development of an L_1 amphibole lineation and growth of oriented hornblende in the necks of boudinaged pyroxenite veins indicates metamorphism synchronous with D_1-D_2 deformation (Table 1), probably related to plutonism (*i.e.*, 'regional' contact metamorphism; *cf.* Lucas *et al.*, 1996).

The penetrative early ductile fabrics in the Meridian-West Arm and Mystic Lake shear zones (S_1, L_1) are overprinted by heterogeneously developed structures attributed to D_3 (Fig. 17; Table 1). The two principal types of ductile overprinting structures are folds and related crenulation cleavages (F_3, S_{3a}) and shear bands (S_{3b}). Mesoscopic (mm to m-scale) folds of S_1, compositional layering and veins occur within the shear zone, and have both Z and S asymmetries. Spaced, steeply dipping axial planar crenulation cleavages (S_{3a}; Fig. 17) are generally developed in the hinge regions of the folds. Higher strain zones where S_1 is transposed into this fabric are locally present. F_3 fold axes trend SSE and have moderate plunges in general, albeit with significant variability (Fig. 17). The folds in general are upright structures with a moderate southward dip to their enveloping surface. The shear bands (S_{3b}) are narrow (mm- to cm-width) zones of strong ductile deformation which reorient the S_1 foliation and produce further grain size reduction and grain alignment at greenschist-grade conditions. The shear bands generally form an angle of ±10-30° to S_1, with the strike of dextral shear bands clockwise (east) of S_1 and that of sinistral shear bands counterclockwise of S_1 (west; Fig. 17). Locally, shear bands form an almost penetrative cleavage, and are associated with crenulation-like folds; shear band development may have followed folding and crenulation cleavage formation.

Outcrop 1

The Meridian-West Arm shear zone forms a relatively narrow (<250 m) high strain corridor marked by greenschist-grade, foliated and lineated tectonites derived principally from the West Arm assemblage (Fig. 17). Detailed study of the Meridian-West Arm shear zone has shown a consistent contrast in deformation history between 'hanging wall' (West Arm assemblage) and 'footwall' (Flin Flon assemblage) units. In the footwall, deformation is localized in a narrow zone (10-50 m wide) in which the shear zone foliation (S_1) is generally oblique to bedding but shows a progressive clockwise reorientation with increasing proximity to the contact (Fig. 17). This relation suggests a dextral component of shear along the contact between the tectonostratigraphic assemblages. In contrast, the West Arm assemblage transposed into a 100-250 m wide band of laminated mafic-felsic tectonite and mylonite adjacent to the Flin Flon assemblage. The felsic layers in the tectonite are derived from intrusive sheets, a feature it shares in common with in the Mystic Lake shear zone. Kinematic indicators associated with the Meridian-West Arm shear zone tectonites are infrequently observed but include dextrally-extended veins, asymmetrically-extended and back-rotated boudins, and dextral C/S fabrics developed in protomylonites (Thomas, 1990; Reilly, 1992; Lucas *et al.*, 1996).

A minimum age for the Meridian-West Arm shear zone is given by the age of the cross-cutting Annabel Lake pluton (Fig. 4; see Thomas, 1993 and references therein). It is a large, variably foliated and lineated granodiorite intrusion that tapers to the northwest as it is caught up in the *ca.* 1800 Ma Annabel Lake shear zone (Ashton, 1992; Reilly *et al.*, 1993; Fedorowich *et al.*, 1995). The pluton cross-cuts at map scale the Flin Flon arc assemblage, the Meridian-West Arm shear zone and tectonized West Arm assemblage of the easternmost part of the Mystic Lake shear zone (Fig. 17; Lucas *et al.*, 1996). The Annabel pluton is in general foliated throughout, with a well developed to mylonitic foliation in its northwest portion related to the Annabel Lake shear zone. Along Meridian Lake (Fig. 17), it is well foliated to mylonitic within about 25-50 m of its margin; the foliation is parallel to the Meridian-West Arm shear zone foliation and probably formed during continued, post-emplacement deformation along the shear zone. Stern and Lucas (1994) reported a U-Pb zircon age of 1866 Ma for the granodiorite. Cessation of deformation along the Meridian-West Arm shear zone is bracketed by the ages of two feldspar porphyry dykes: an 1847 Ma plagioclase-phyric dyke cuts across the shear zone at a low angle but contains the S_1 foliation whereas an 1839 Ma K-feldspar-phyric dyke cuts across the shear zone at a high angle (Fig. 17). This result is consistent with the 1838 Ma age of the Boot-Phantom pluton (Heaman *et al.*, 1992), which also cross-cuts the shear zone at a high-angle and does not contain the S_1 foliation (Thomas, 1989).

Features to note at this outcrop include:

- Laminated, felsic-mafic tectonites derived from the West Arm ocean floor assemblage and syn-tectonic intrusive sheets: contrast with deformation state in West Arm basalts viewed at Stop 4.
- Evidence for dextral kinematics (west-side-up and to the north relative to the Flin Flon assemblage): asymmetrically-extended pyroxenite veinlets with back-rotated boudins; note also that amphibole is replacing clinopyroxene.

Outcrop 2

The Mystic Lake shear zone is a 2-3 kilometre wide band of foliated to mylonitic rocks that is cored by Mystic Lake assemblage plutonic rocks but includes significant widths of adjacent Birch arc assemblage and West Arm ocean floor assemblage volcanic rocks (Figs. 4, 18; Lucas *et al.*, 1996). It is cut by the 1856 Ma Kaminis pluton and the 1850 Ma Reynard Lake pluton (Stern and Lucas, 1994), and thus must predate emplacement of successor arc plutons and deposition of the Missi suite. However, both

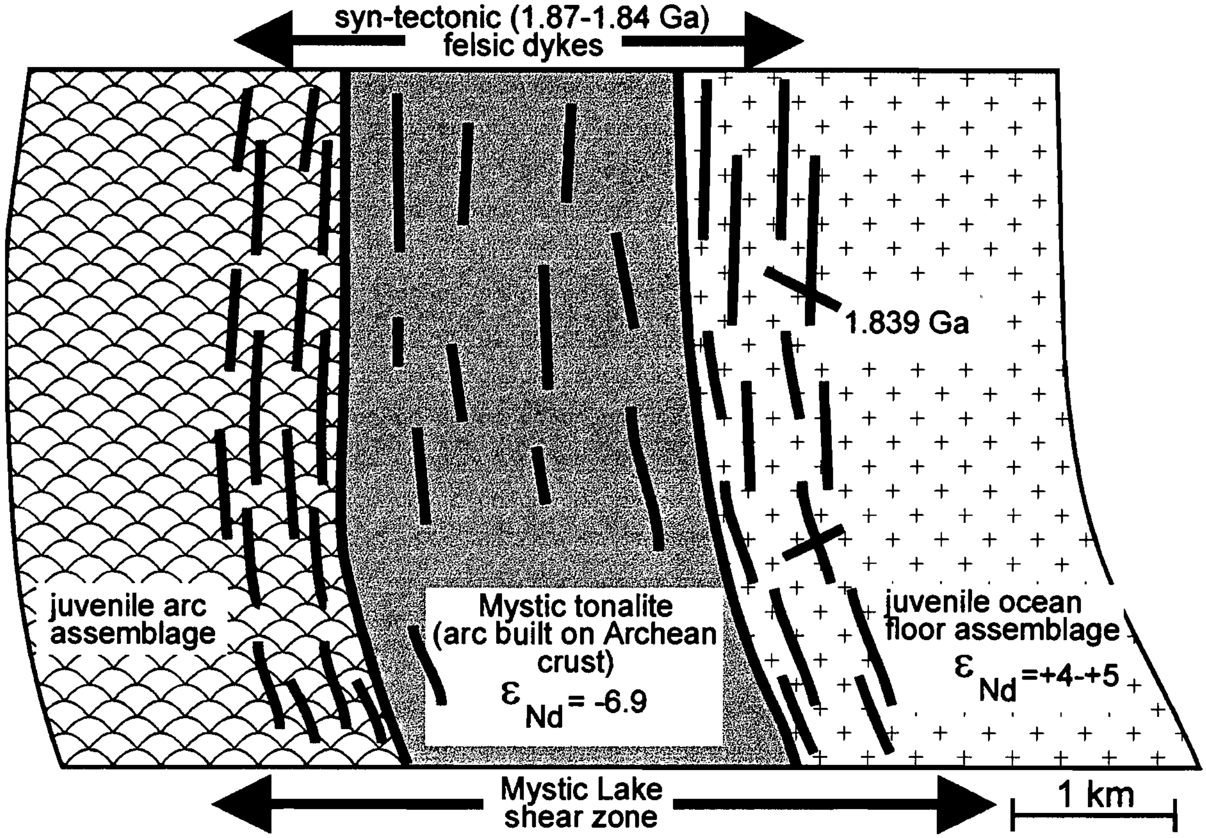


Figure 18: Cartoon illustrating the relations between the Birch Lake arc assemblage and West Arm ocean-floor assemblage. These ca. 1.9 Ga assemblages are structurally juxtaposed along the Mystic Lake shear zone, a 2-3 km wide band of foliated to mylonitic rocks that is cored by Mystic Lake assemblage plutonic rocks but includes significant widths of adjacent arc and ocean-floor rocks. Syn-tectonic felsic sheets are spatially associated with the deformation corridor. U-Pb zircon ages (Lucas and Stern, 1994; Lucas et al., 1996) of both syn-tectonic and cross-cutting intrusions constrains the age and duration of deformation in the Mystic Lake shear zone (minimum 1856 Ma; cessation of deformation by 1839 Ma).

of these plutons contain weakly to moderately developed tectonic fabrics that parallel those in the shear zones that they cross-cut (Reilly, 1992). This observation suggests that the shear zone was active during successor arc magmatism, consistent with the observation of abundant deformed layer-parallel intrusive sheets within the shear zone itself (Fig. 18). Note that D_1 represents accretion-related deformation *sensu stricto* and that D_2 represents subsequent syn-magmatic deformation (Table 1), although D_1 fabrics cannot be systematically distinguished from those related to D_2 in the Mystic Lake shear zone (Lucas *et al.*, 1996).

The transitions into the Mystic Lake shear zone from both the eastern and western wallrocks are characterized by: (1) increase in strain, generating L-S tectonites and resulting in the transposition of all lithological contacts into parallelism with the S_1 foliation; (2) appearance of centimetre to metre-scale, felsic to mafic veins and intrusive sheets (predominantly felsic), generally deformed and oriented sub-parallel to the wallrock tectonite foliation; (3) increase in the proportion of intrusive sheets to over 75% of outcrop, with only rare screens of wallrock remaining; and (4) appearance of Mystic Lake assemblage meta-plutonic rocks (tonalite-diorite), themselves interlayered with deformed intrusive sheets as in the wallrocks. The Mystic Lake-West Arm and Mystic Lake-Birch contacts are tectonic where observed, although mylonitized intrusive sheets generally separate Mystic Lake assemblage rocks from those of the two bounding assemblages (Lucas *et al.*, 1996).

Features to note at this outcrop include:

- Laminated, predominantly felsic tectonites derived from syn-tectonic intrusive sheets.
- Range in nature of felsic sheets, from aphyric, fine grained 'rhyolitic' material to medium grained granodiorite-tonalite.
- Deformed quartz veins showing complex kinematic histories.
- Overprinting folds, shear zones and related fabrics associated with D_3-D_5 deformation.

Outcrop 3

The Reynard Lake intrusion is a zoned gabbro-diorite-granodiorite body that intrudes the Birch, Mystic Lake and West Arm assemblages, as well as tectonites derived from each of these assemblages in the Mystic Lake shear zone (Fig. 4; Reilly, 1990, 1991; Thomas, 1989, 1990, 1991). Stern *et al.* (1993) obtained ages of 1849 +3/-2 Ma on a marginal gabbro phase and 1850 ±3 Ma on a K-feldspar porphyritic granodiorite phase. Heaman *et al.* (1992) also obtained an age of 1850 ±2 Ma for a gabbroic body called the Wekach gabbro at the margin of the Reynard Lake pluton. Together, these data suggest intrusion of the Reynard Lake pluton and related gabbro bodies at 1850 ±1 Ma (Stern *et al.*, 1993).

Features to note at this outcrop include:

- Reynard Lake granodiorite cutting layered felsic-mafic tectonites related to the Mystic Lake shear zone and including misoriented, angular tectonite blocks. The granodiorite also cuts across folds of the tectonite layering.

- Development of a weak foliation in the Reynard granodiorite, probably associated with the waning stages of D_2 deformation.

Stop 6: Stratigraphic section in the Hook Lake block

Introduction

Flin Flon arc assemblage stratigraphic sequences are marked by large-scale intercalation of effusive, fragmental and intrusive rocks. Such units are on the scale of 10s to 100s of metres thick. The sequences are internally complex, in which successive units vary with respect to flow or bed morphology, phenocryst content, and composition. It is this variability that distinguishes arc assemblage rocks from the monotonous basalt (± diabase) successions in the Elbow-Athapapuskow ocean-floor assemblage (*e.g.*, Stop 11).

The outcrop-scale stratigraphy at this Stop mirrors the map-scale characteristics of the arc assemblage. Here we will examine the components and bedforms of a mafic pyroclastic unit typical of the upper part of the Hook Lake suite stratigraphy (Figs. 5, 4). Mafic pyroclastic rocks are common in the Hook Lake and Flin Flon suites, but are less common elsewhere within the Flin Flon arc assemblage.

Flows in the sequence are part of the juvenile arc tholeiitic series, characterized by HFSE depletion and generally flat REE patterns (Stern *et al.*, 1995a). Primary augite phenocrysts from this sequence have very low contents of Ni, Ti and Cr typical of arc tholeiites (Turnock and Syme, unpub. data), consistent with the whole-rock geochemistry (Stern *et al.*, 1995a). Hook Lake suite rocks have 'juvenile' Nd-isotopic characteristics, plotting within the Flin Flon MORB-OIB field (*i.e.,* overlapping with the likely mantle source), implying that the influence of significantly older (*e.g.,* Archean) crust was minimal (Stern *et al.*, 1995a).

The rocks to be examined are part of a 300 m thick unit comprising intercalated pillowed basalt flows, scoria-rich tuff and pillow fragment breccia (Fig. 19). The environment of deposition for these rocks is constrained to be subaqueous by the presence of pillowed flows throughout the sequence. However, the physical characteristics of juvenile components in the tuffaceous members of the sequence (discussed below) suggest that the tuff was erupted in a subaerial or very shallow water environment, and deposited in shallow water close to the source vent(s).

Stop description

The outcrop area is located in the Hook Lake block (Bailes and Syme, 1989) defined as the fault-bounded stratigraphic package lying southeast of Flin Flon (Fig. 4). Parking is on an abandoned road on the south side of Highway 10, just west of the south end of Manistikwan (Big Island) Lake (Fig. 19). We will walk through an abandoned quarry to outcrops on the south rim.

This area lies within the subgreenschist (prehnite-pumpellyite) zone of metamorphism: the prehnite-out, pumpellyite-out, actinolite-in isograd lies 1.5 km to the north (Digel and Gordon, 1995; Bailes and Syme, 1989). On this particular outcrop both prehnite and pumpellyite are rare to absent. The rocks are virtually unfoliated, and primary textures are exceptionally well preserved. The Inlet Arm fault (defining the

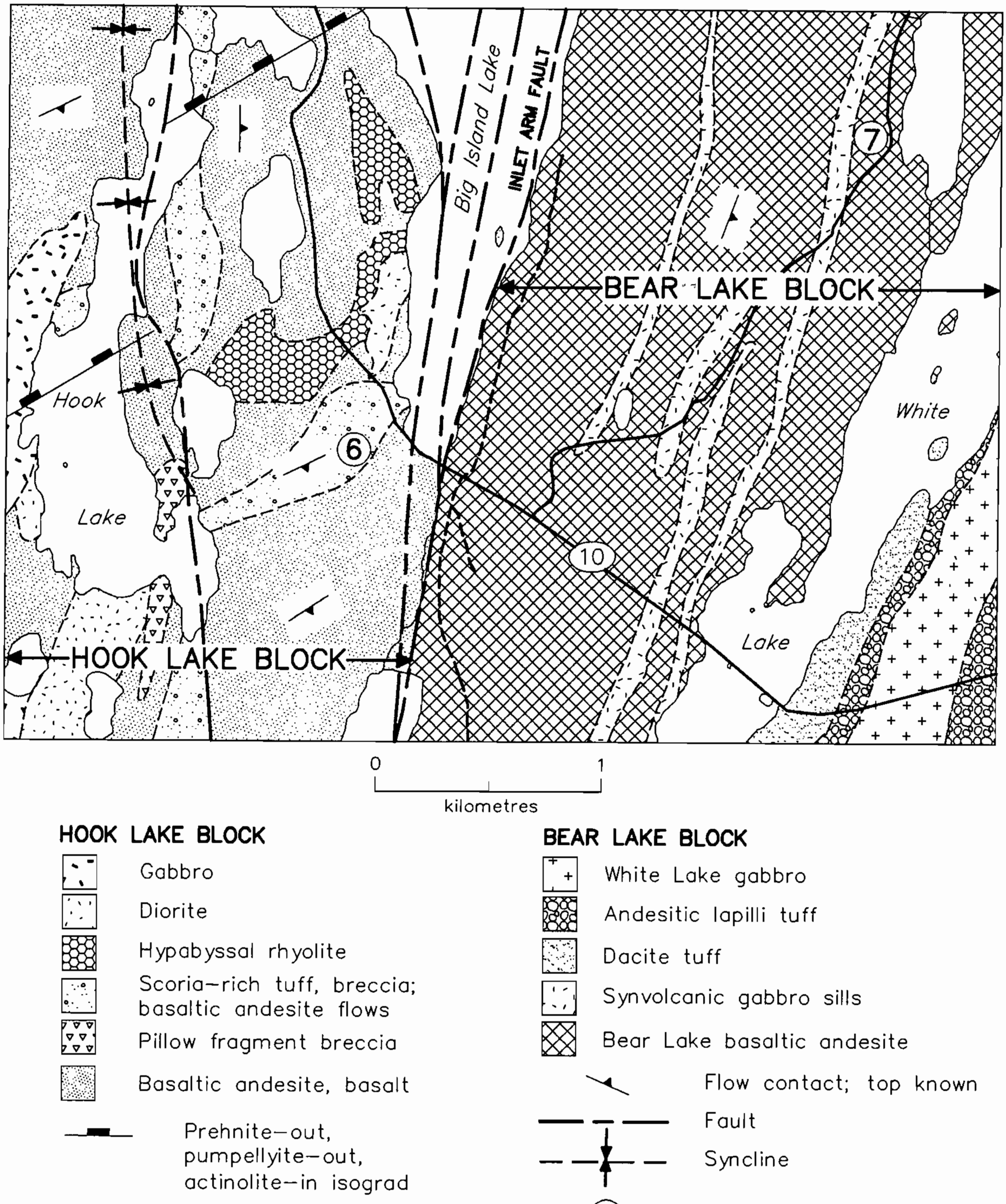


Figure 19: Simplified geological map of the Hook Lake area, southwest of Flin Flon. The Hook Lake block (a tholeiitic juvenile arc suite) is juxtaposed against the Bear Lake block (a transitional tholeiitic-calc-alkaline juvenile arc suite) by the Inlet Arm shear zone. Fabrics associated with the Inlet Arm shear zone are most strongly developed along the east shore of Big Island Lake, but extend as the principal foliation across the entire width of the Bear Lake block (Lucas et al., 1996). The structure labeled 'Inlet Arm fault' forms the western edge of the Inlet Arm shear zone, but is younger than the S1 shear fabric. Stop 6: Scoria tuff, tuff-breccia, pillowed basalt in the subgreenschist portion of the Hook Lake block; Stop 7: deformed Bear Lake basaltic andesite. Modified from Bailes and Syme (1989).

western edge of the Inlet Arm shear zone in the Bear Lake block: see Stop 7) is approximately 500 m to the east.

A line has been flagged (azimuth 133°) within and adjacent to the area shown on Figure 20. The flagged line is 210 m in length, begins at pillowed flow 1 and terminates in pillowed flow 9. Approximately 170 m from the start of the main line, a secondary flagged line (azimuth 240°) trends off to the right for 45 m.

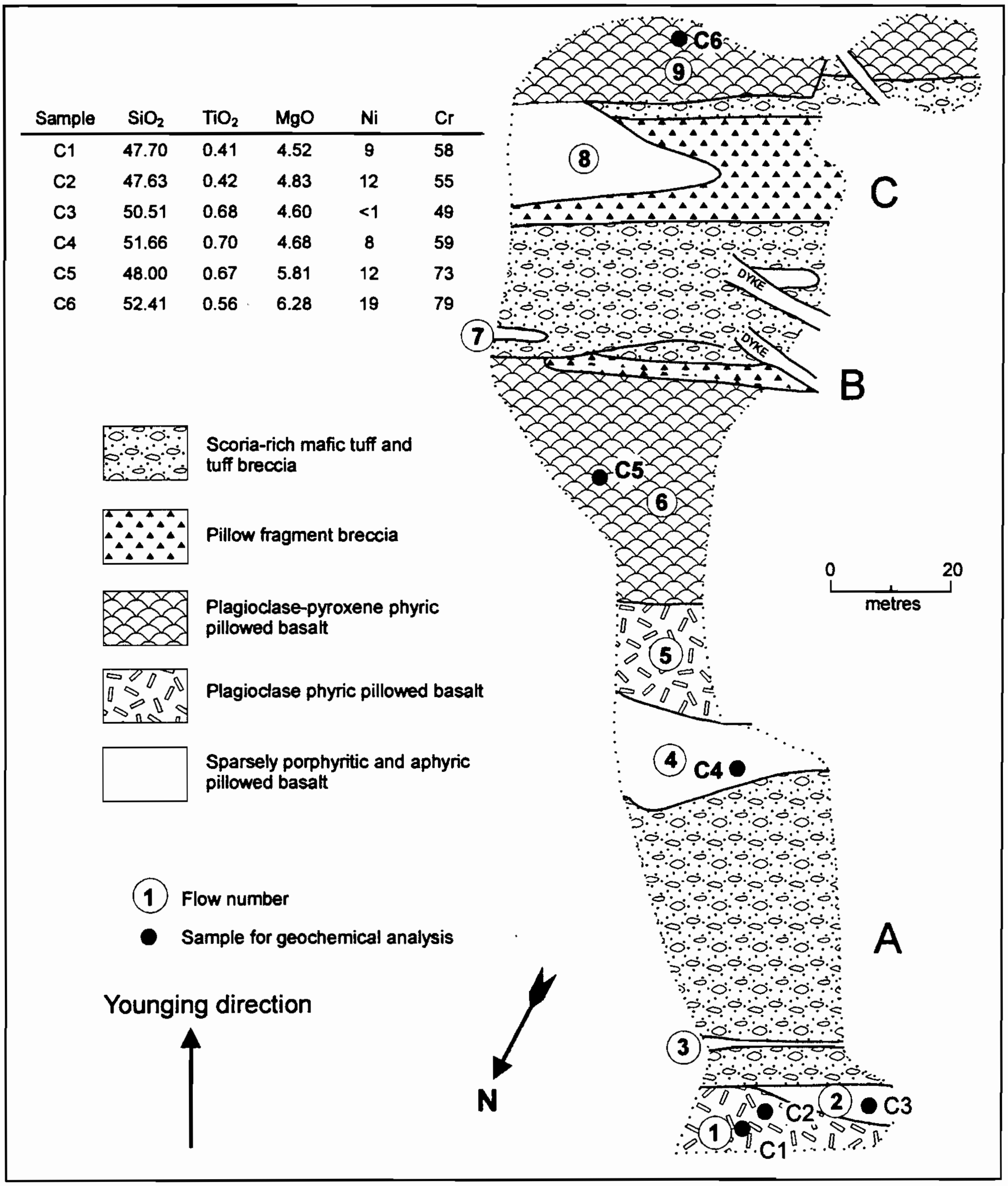
Pillowed flows are 2-40 m thick, and contain 5-30% plagioclase phenocrysts and 1-5% amphibole pseudomorphs after pyroxene phenocrysts. Flows on this outcrop are basalts and basaltic andesites (Fig. 20) with 4.52-6.28 wt.% MgO (note that both SiO_2 and MgO increase up-section). Some flows abruptly terminate in the exposed section. Amygdales range in size up to 6 mm but vary in size and abundance between flows and between pillows in a single flow. Concentric bands of carbonate-filled amygdales occur in the margins of pillows in Flow 6. Pillow fragment breccia is directly associated with flows 6 and 8; amoeboid pillow flow-top breccias are absent.

Pillow fragment breccia beds are 3-5 m thick, contain fragments to 20 cm, and include both monolithologic and heterolithologic types. Monolithologic breccias can in some instances be shown to contain fragments identical to the directly underlying basalt flow. Breccia beds have fragment-supported bases and matrix-rich tops but do not display normal size grading. Fragments are angular, and in many cases are clearly fragments of pillows that have disintegrated along radial and concentric fractures, producing characteristic pie-shaped clasts. Proximal breccias may contain complete, unfragmented pillows.

Scoria tuff and breccia beds (<1-13 m thick) are typically less than 5 m thick. Beds less than 1 m thick are normally graded with respect to plagioclase crystal size. Beds thicker than 1 m are commonly reverse graded at the base (Fig. 21). Tuff and breccia beds include A and AB types, and in some instances contain "trains" of accessory blocks (Fig. 21). The occurrence of scoria in Bouma-type beds intercalated with pillowed basalt indicates the pyroclastic material was transported and deposited by subaqueous density currents.

Scoria particles range in size from about 1 mm to 6 cm. The primary vesicularity and grain shapes of scoria are well preserved (Fig. 21). Scoria grains are framework supported with vesicles and inter-particle voids filled predominantly with carbonate. The shapes of scoria particles are controlled by a combination of vesicle walls, fracture surfaces, and chilled droplet margins. The particles are interpreted to have been produced during explosive subaerial magmatic eruptions on a volcanic island. During eruptions, rapidly vesiculating magma droplets in the eruption column were fragmented by internal gas pressure. Many of these hot vesiculated fragments may have been subsequently broken by thermal shock on contact with seawater.

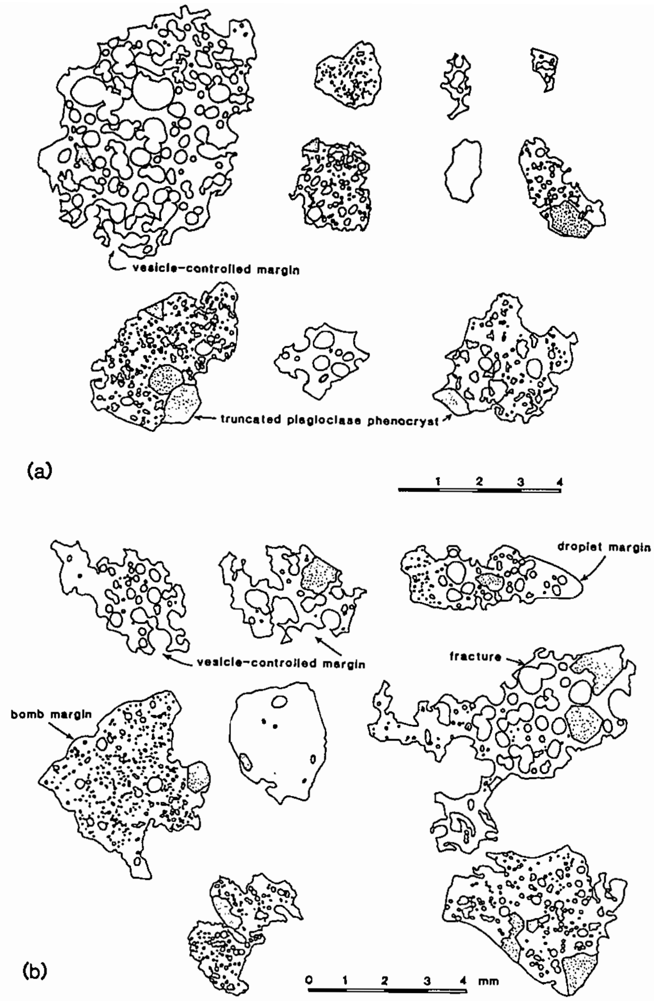
Accessory blocks (Fig. 21) are interpreted to be ballistic fragments. Individual blocks are angular, up to 40 cm across, and are fragments of amygdaloidal pillowed basalt that can be unambiguously matched to specific flows lower in the sequence (*e.g.*, blocks in unit 'A' (Fig. 20) match flows 1 and 3). The blocks commonly occur in discontinuous "trains" (1 block thick) and as isolated fragments; both "trains" and isolated blocks can occur at any level within a tuff bed (Fig. 21). The blocks are much

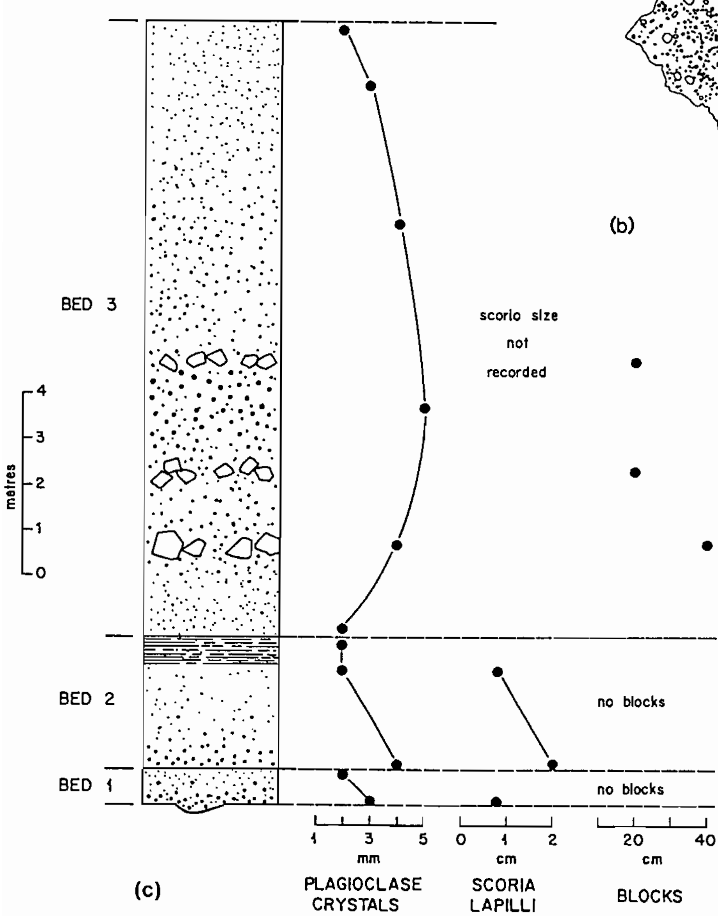


Sample	SiO_2	TiO_2	MgO	Ni	Cr
C1	47.70	0.41	4.52	9	58
C2	47.63	0.42	4.83	12	55
C3	50.51	0.68	4.60	<1	49
C4	51.66	0.70	4.68	8	59
C5	48.00	0.67	5.81	12	73
C6	52.41	0.56	6.28	19	79

Figure 20: Stratigraphic sequence at Stop 6, Hook Lake block. Subaqueous pillowed basalt flows are intercalated with bedded scoria-rich mafic tuff and tuff breccia. Pillowed flows vary from aphyric to coarsely porphyritic, but are all geochemically similar arc tholeiites, characterized by low HFSE (e.g., Ti), Ni and Cr (oxides in wt.%, Ni and Cr in ppm). A: scoria-rich tuff, beds 0.8-13.4 m thick; thicker beds contain trains of accessory blocks that can be matched to flows 1 and 3. B: tuff breccia and pillow fragment breccia both contain abundant pillow fragment blocks derived from underlying flow 6. The tuff-breccia has a scoria-rich matrix, whereas the pillow fragment breccia has no scoria in the matrix. C: pillow fragment breccias are thick bedded (5-6 m) and heterolithologic. Geology modified from Corkery (pers. comm., 1983). From Bailes and Syme (1989).

(a) Tracings of scoria particles in scoria-rich lapilli tuff. These 1-5 mm particles have predominantly vesicle-controlled margins, and were produced during magmatic eruptions in a very shallow water or subaerial environment. Some particles are bounded by fractures or chilled droplet margins. (a) from 6 m above base of bed 3 in (c); (b) from 1.5 m above base of bed 2.





(c) Bedforms with turbidite-like sequence of internal divisions indicate that the pyroclastic material was deposited from subaqueous density currents. Grading displayed by these three scoria tuff beds near the base of the section is demonstrated by variation in the maximum size of plagioclase crystals and scoria lapilli. Bed 1 is normally graded (Bouma A type), bed 2 is normally graded - parallel laminated (AB type), and bed 3 is reverse to normally graded with three crude layers of accessory blocks (up to 40 cm). The latter are fragments derived from pillowed flows lower in the sequence.

Figure 21: Bedforms and nature of ash particles, scoria-rich mafic tuff, Stop 6. From Bailes and Syme (1989).

larger and denser than the matrix scoria and were clearly not in hydraulic equilibrium with the density current that deposited the scoria. The accessory blocks may have been torn from vent walls by periodic phreatic/phreatomagmatic explosions which punctuated the dominantly magmatic, scoria-producing, eruptions.

Stop 7: Inlet Arm shear zone / Bear Lake basaltic andesite

Introduction

The Inlet Arm shear zone separates several major packages of arc volcanic and volcaniclastic rocks within the juvenile Flin Flon assemblage (Figs. 4, 22, 23; Bailes and Syme, 1989; Lucas *et al.*, 1995). As with most major structures in the area, the Inlet Arm shear zone is demonstrably long-lived, including early ductile deformation overprinted by both ductile and brittle structures including fracture arrays and pseudotachylite. We distinguish the Inlet Arm *fault* from the Inlet Arm *shear zone*: the fault represents a metre-scale zone of brittle-ductile deformation effectively corresponding to the actual contact between the Bear Lake block and blocks to the west (Bailes and Syme, 1989), whereas the shear zone is a kilometre-scale band of tectonite and foliated stratigraphy that is primarily developed in the Bear Lake section but is systematically eliminated along strike to the south by the fault (Fig. 4).

A penetrative foliation (S_1; Fig. 22) is mapped throughout the Bear Lake block but clearly intensifies towards the western boundary, resulting in a 50-100 m wide band of well laminated mafic tectonite (Lucas *et al.*, 1995). It is defined by flattened amygdales and pillows and at high strains, flattened epidosite domains which generally correlate with pillow cores in less deformed rocks (Bailes and Syme, 1989). A moderately- to steeply-plunging extension lineation (L_1) is associated with the S_1 foliation (Fig. 22), although chocolate tablet boudinage of competent domains indicates an overall flattening strain. The lineation is defined by quartz and feldspar rodding in quartz and felsic veins associated with the tectonites. The S_1 foliation is locally associated with cm-scale isoclinal folds of quartz, epidote and felsic veins, which consistently show an S-asymmetry where preserved. F_1 fold axes appear to be parallel to the L_1 extension lineation. A 10 cm-scale F_1 structure appears to fold an earlier foliation, suggesting that S_1 itself may be a crenulation cleavage. The S_1 foliation contains sinistral kinematic indicators within the horizontal plane, including rotated porphyroclasts derived from quartz veins and epidosite domains in the Bear Lake basalts.

The S_1 tectonites contain mylonites derived from felsic dykes along the highest strain portion of the Inlet arm shear zone, adjacent to the Hook Lake section to the west (Fig. 4). Unfortunately, these dykes have not yet yielded any zircon for U-Pb geochronological study. However, the tectonites are also cut by a plagioclase phyric felsic dyke close to the trace of the younger Inlet Arm fault. U-Pb analysis of zircons recovered from this dyke has yielded an imprecise age of 1843 +31/-16 Ma (Stern, unpublished data), which represents a minimum age for D_1-D_2 deformation along this corridor. This result is consistent with the relations between dyke emplacement and D_1/D_2 deformation along the Meridian-West Arm shear to the west, where cessation of D_2 deformation is bracketed between 1847 and 1839 Ma. The 1843 Ma dyke in the Inlet

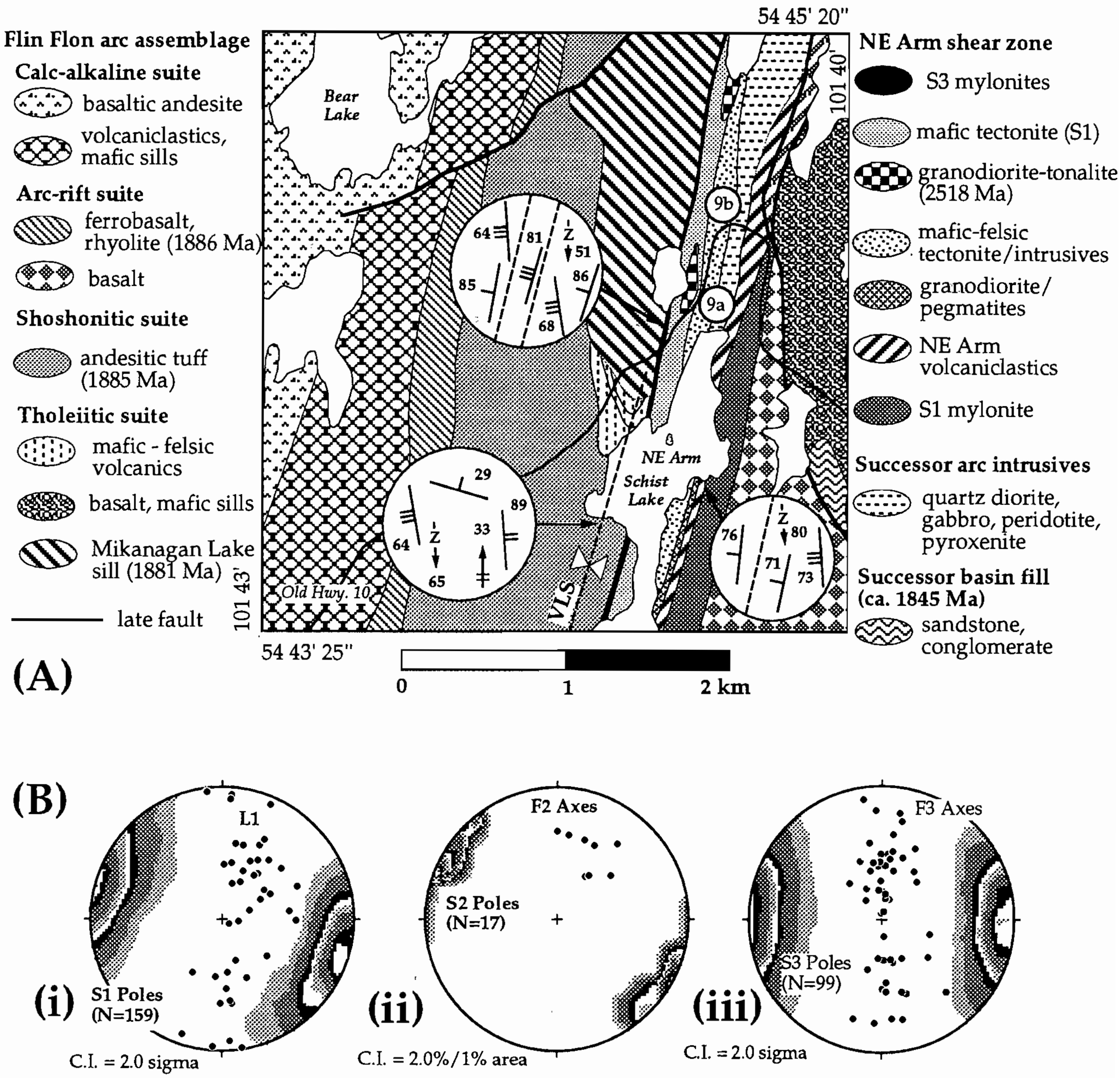


Figure 22:(a) Detailed map of a portion of the Northeast Arm shear zone. After Bailes and Syme (1989), Syme et al. (1993) and Lucas et al. (in prep.). Foliations illustrated in inset circles (S_1 - one 'tick'; S_2 - two 'ticks'; S_3 - three 'ticks') are associated with D_1-D_2, D_2 and D_3, respectively. Bedding/compositional layering is generally parallel to S_1. Arrow with two 'ticks' is a F_2 (D_2) fold axis and arrows with a 'Z' are F_3 (D_3) dextral fold axes. VLS: Vick lake synform. U-Pb geochronology: 2518 $^{+7}/_{-4}$ Ma (zircon, David and Syme, 1994); 1886 ±2 Ma (zircon, Gordon et al., 1990); 1885 ±3 Ma (zircon, Stern et al., 1993); 1881 ±2 Ma (titanite, David et al., 1993). Successor basin sandstone/conglomerate age inferred to be ca. 1845 Ma (Ansdell et al., 1992; Ansdell, 1993).
(b) Equal area stereonet plots for structural data from the Inlet Arm and Northeast Arm shear zones.


Bear Lake
White Lake
Old Hwy. 10
8
9
10
10
C
W
SYNCLINE
NORTHEAST ARM SHEAR ZONE
Lake
Schist
Athapapuskow Lake
0 500
metres

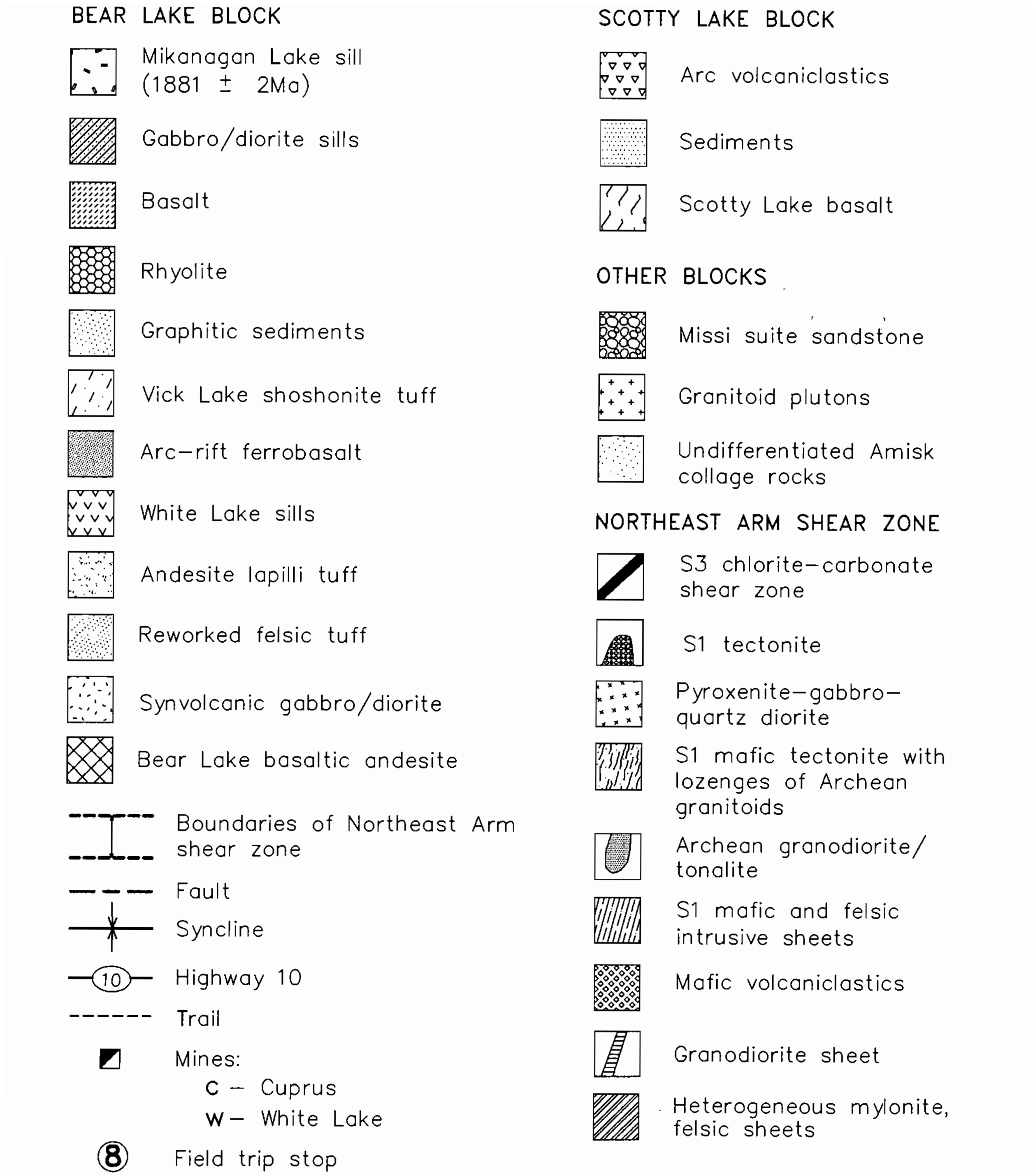


***Figure 23:** Generalized geology of the Northeast Arm of Schist Lake. The Northeast Arm shear zone is a 0.5-1 km wide deformation corridor within the Flin Flon assemblage that juxtaposes substantially different stratigraphic packages (Bailes and Syme, 1989; Lucas, 1993; Lucas et al., 1996). It forms the eastern boundary to the juvenile calc-alkaline Bear Lake block and the western boundary to a section of tholeiitic (arc-rift?) basalts with relatively evolved initial ε_{Nd} compositions (Scotty Lake block; Bailes and Syme, 1989; Syme et al., 1993; Lucas et al., 1996). VMS deposits in the Bear Lake block are: C - Cuprus mine; W - White Lake mine. Stop 8: Vick Lake shoshonite tuff; Stop 9: Northeast Arm shear zone, Archean granodiorite; Stop 10: Scotty Lake basalt. Modified from Syme et al. (1993) and unpublished map by Syme, Lucas and Stern (1994).*

Arm shear zone appears to have been emplaced at a low angle to S_1 and S_3, but does not contain the S_1 foliation. It folded by F_3 Z-folds and has a weakly developed S_3 foliation (Fig. 22). Given that it appears to cut across S_3 locally but is itself folded by F3, we speculate that it may have been emplaced during F_3/S_3. Fragments of both the dyke and shear zone tectonite are found in foliation-parallel breccia seams that post-date F_3/S_3 but are themselves locally deformed by late Z-folds.

Stop description

Bear Lake basaltic andesite consists of 3.3 km of buff-weathering flows, related breccias and dykes, at stratigraphic base of the east-facing Bear Lake block (Bailes and Syme, 1989). Pervasive epidotization has affected a large part of the southern and western portions of the unit, coinciding with the zone of most intense deformation associated with the Inlet Arm shear zone. This epidotization has largely destroyed primary structures and rendered flow contacts unrecognizable. Outside the area of strong epidotization primary structures and flow contacts are well preserved and readily identified.

The outcrop shows well foliated Bear Lake basaltic andesite, with the S_1 fabric attributed to deformation associated with the Inlet Arm shear zone. The outcrop is approximately 1400 m east of the trace of the Inlet Arm fault. The epidosite cores to the pillows are more resistant to deformation than non-epidotized pillow margins and inter-pillow hyaloclastite. Although virtually all of the Bear Lake basaltic andesite section is marked by the S_1 foliation related to Inlet Arm shear zone, facing indicators consistently show tops to the east.

Stop 8: Vick Lake shoshonitic tuff

Introduction

Intra-arc tectonic processes are recorded in the stratigraphic components of some of the fault-bounded successions in the Flin Flon arc assemblage. The Bear Lake block (Bailes and Syme, 1989), defined as the stratigraphic package contained between the Inlet Arm fault and Northeast Arm shear zone (Figs. 23, 24), contains a particularly clear example of intra-arc rifting. The lithologic components of the Bear Lake block are particularly well exposed, allowing more stratigraphic interpretation and reconstruction than is generally possible in less well-exposed areas.

The Bear Lake block contains three contrasting volcanic associations: 1) a 4 km thick, mildly calc-alkaline arc sequence; 2) 200 m of ferrobasalt; and 3) 1 km of basin-fill shoshonitic tuff with turbidite bedforms (Bailes and Syme, 1989; Syme and Bailes, 1993; Stern *et al.*, 1995a). These components are interpreted to record the construction and subsequent rifting of part of an oceanic arc. At this stop (Fig. 23) we will examine the topmost member of this succession, the shoshonitic Vick Lake tuff, to establish the deep basinal nature of these rocks.

Bear Lake basaltic andesite forming the faulted base of the arc succession (Figs. 23, 24) represents a shoaling subaqueous shield volcano >3.3 km thick. This volcanism apparently ended with caldera collapse of the shallow-water, upper portions of the subaqueous shield. Caldera formation was abruptly succeeded by effusion of

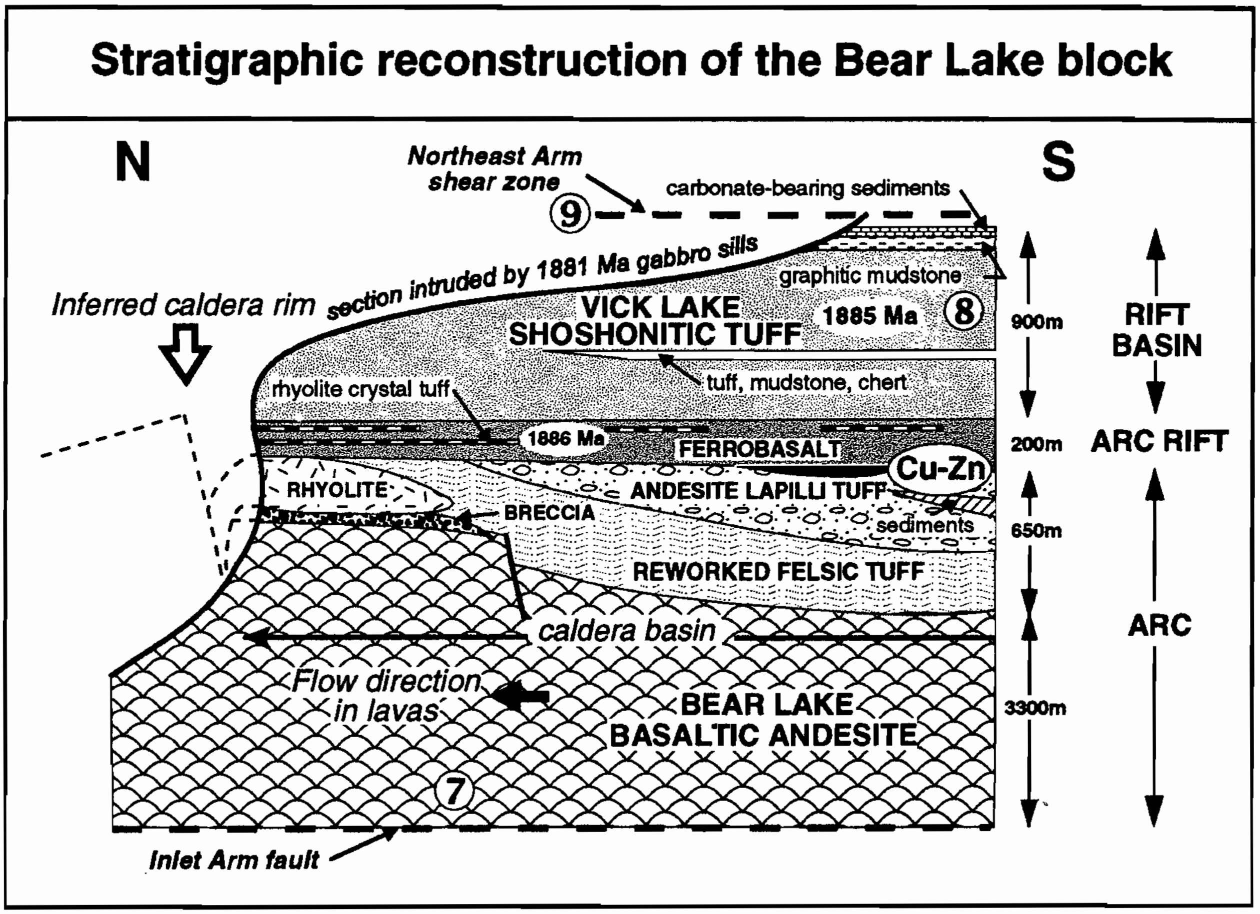


Figure 24: Stratigraphic relationships in the Bear Lake block, Flin Flon arc assemblage. The section contains three major groups of rocks: (1) transitional tholeiitic-calc-alkaline Bear Lake suite (comprising Bear Lake basaltic andesite, caldera-fill rhyolite, felsic volcaniclastic and intermediate volcaniclastic rocks), (2) arc-rift ferrobasalt, and (3) rift basin Vick Lake suite shoshonite tuff. The Cuprus and White Lake VMS deposits occur in a unit of graphitic mudstone at the top of the calc-alkaline sequence, overlain directly by arc-rift ferrobasalt. Stratigraphic locations of field trip stops 7, 8 and 9 are shown. Modified from Syme and Bailes (1993).

intracaldera subaqueous rhyolite flows and contemporaneous infilling of the southward-deepening basin by felsic and intermediate volcaniclastic rocks. Graphitic mudstones, cherts and stratabound massive sulphides (Cuprus and White Lake mines; Bailes and Syme, 1989; Syme and Bailes, 1993) were deposited at the top of the calc-alkaline sequence, in sub-basins which may have heralded an intra-arc rifting event. This rift event is represented in the Bear Lake block stratigraphy by a 150-200 m thick ferrobasalt formation (Bailes and Syme, 1989) with N-MORB characteristics (Stern *et al.*, 1995a; Lucas *et al.*, 1996; Stern, unpublished data). Deposition of the sulphides may have occurred in the earliest stages of arc extension, and are thus plausibly associated with an episode of high heat flow subsequently manifest by the extrusion of basalts derived from a MORB-like mantle. Inter-flow rhyolite crystal tuff beds in the ferrobasalt have a U-Pb zircon age of 1886 ±2 Ma (Gordon *et al.*, 1990), establishing the age of the rifting. The resulting arc-rift basin was subsequently filled with 900 m of fine-grained shoshonitic pyroclastic material deposited from subaqueous density currents (Stop 8 - Vick Lake tuff). This shoshonitic material has a U-Pb age of 1885 ±3 Ma (Stern *et al.*, 1993), clearly indicating that the rifting and shoshonitic volcanism were virtually coeval.

Samples from the Vick Lake tuff suite plot within the basaltic trachyandesite and trachyandesite fields on a silica *versus* Na_2O+K_2O plot, and the high-K field on the silica *versus* K_2O plot (Fig. 6), distinct from all other rocks within the Flin Flon belt (Stern *et al.*, 1995a). K_2O contents range from 2.3-6.5 wt.%, Na_2O from 1.0-5.0 wt.%, and K_2O/Na_2O ratios from 0.4-7.0. The high K_2O contents are interpreted to reflect magmatic values, supported by the strong positive correlation of K_2O with less-mobile, incompatible elements such as Th and La. The shoshonitic rocks are distinct from the tholeiitic and calc-alkaline rocks in most other respects on Harker variation diagrams, having higher Al_2O_3 (16-21 wt.%), lower MgO (2.3-3.0 wt.%), and higher P_2O_5 contents (0.3-0.45 wt.%) at any given silica value. The HFSE characteristics of the Vick Lake shoshonite suite partly overlap those of the tholeiitic and calc-alkaline series at Flin Flon, but extend to more extreme values for some elements and ratios. They are characterized by greater enrichment in LREEs [$(La/Yb)_n$ = 3-7; La_n = 25-40], and flat to slightly fractionated HREE profiles (Fig. 6). The rocks have particularly high Ba (640-2000 ppm), Sr (450-1000 ppm), and Th (2.1-3.4 ppm), features consistent with a shoshonitic affinity (Morrison, 1980).

The tuff is composed almost entirely of juvenile pyroclastic material: shards, pumice, plagioclase crystals and crystal fragments, vitric and microlithic microclasts, amphibole crystals and crystal fragments, and a fine grained recrystallized matrix (Fig. 25; Bailes and Syme, 1989):

- Shards (0.05-0.4 mm) are aphyric and have equant to tabular shapes with grain boundaries defined by straight to conchoidal fracture surfaces and vesicle walls. Vesicles (<0.05 mm) are round, oval or tube-shaped, and are filled with very fine grained epidote, chlorite, feldspar or quartz. Shards are recrystallized and replaced by a mixture of albite, subordinate quartz, epidote, chlorite and sericite.
- Pumice granules are typically less than 3 cm long, 0.4-6 mm in the ash-sized matrix of most beds and up to 30 cm in some pumice-rich beds. Pumice fragments

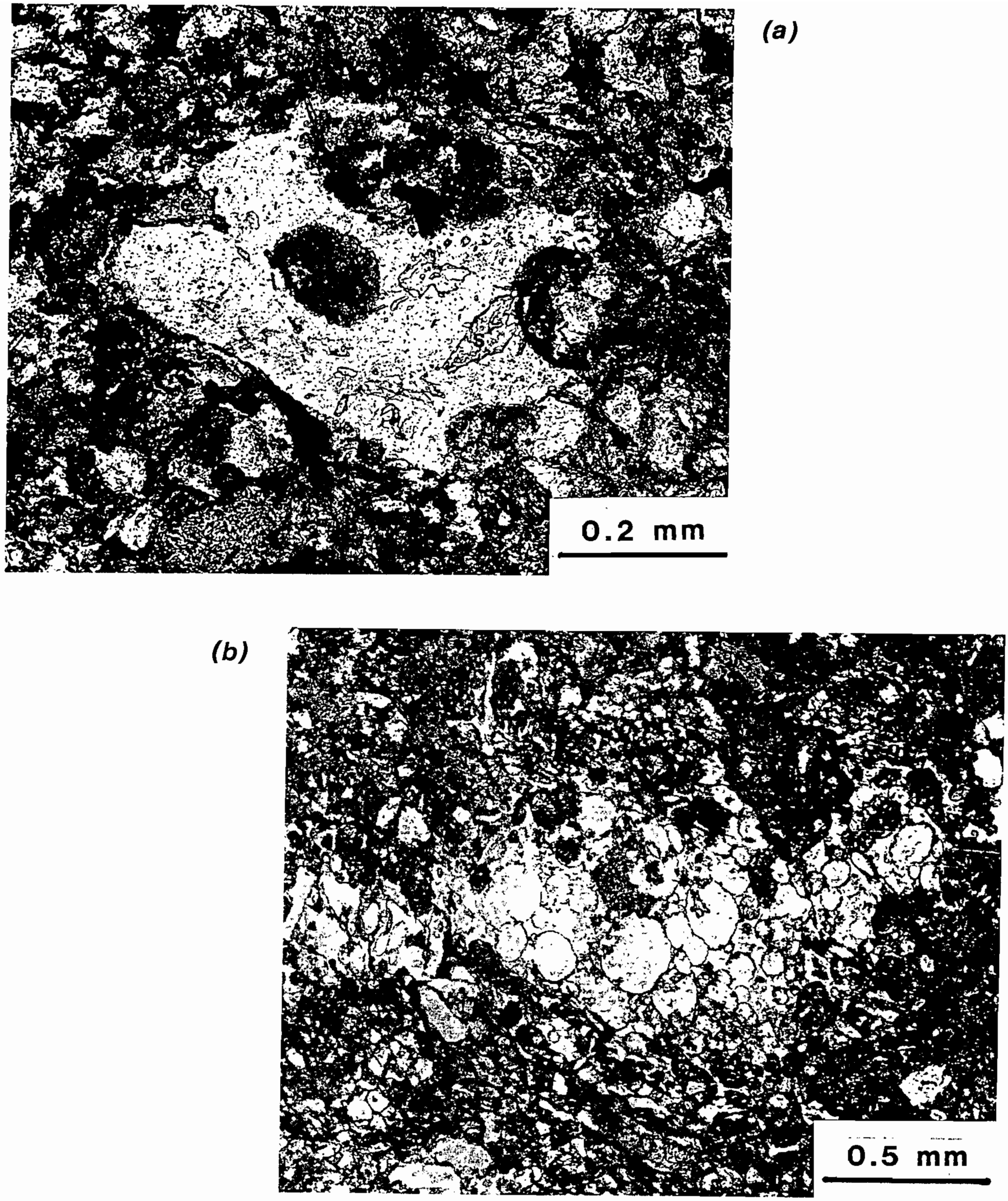


Figure **25**: *Photomicrographs of principal components of Vick Lake shoshonite tuff. (a) Large shard displaying conchoidal boundaries that cut through vesicles. Matrix consists of smaller shards and plagioclase crystal fragments. The original glass has been replaced by polygranular albite, sericite and epidote. (b) Pumice granule with high vesicularity. Vesicles are filled with quartz and albite. Both in plane polarized light. From Bailes and Syme (1989).*

contain plagioclase phenocrysts similar in size and shape to unbroken crystals in the matrix. The shape of pumice granules and blocks is variable and ranges from oval, equant to irregular. Vesicles (50-80%; <0.4 mm) are round, oval or tube-shaped, filled by a polygranular mosaic of feldspar and quartz.

- Euhedral plagioclase crystals and angular crystal fragments (0.1-2 mm) comprise approximately 15-40% of tuff beds. Normal size and abundance grading of plagioclase is typical in most beds. Broken euhedral prisms of magmatic hornblende comprise 0-3% of tuff and are typically concentrated in the bases of beds.
- Aphyric or plagioclase phyric microclasts (<2 mm) are common in the lower parts of graded beds where they comprise up to 15% of the tuff. These clasts are interpreted as juvenile glass fragments, bounded largely by conchoidal fracture surfaces, with shapes subsequently smoothed by abrasion. Subrounded microlitic lithic clasts (<0.5 mm) are interpreted as accessory components, in that they have a crystalline pilotaxitic texture of plagioclase microlites.

The Vick Lake tuff sequence is well bedded and broadly upward fining (Bailes and Syme, 1989). The basal 250 m is characterized by abundant thick beds containing large pumice fragments, but such fragments are rare in the upper two-thirds of the unit. Bed thickness also decreases upward within the unit. Within the uppermost 100 m Vick Lake tuff is interlayered with pyritic, graphitic mudstones. For the unit as a whole, beds are 0.2-18 m thick and average 1 m. The beds have a Bouma-type internal zonation similar to greywacke turbidites, comprising one or more of a graded "A" division, parallel laminated "B" division, ripple laminated "C" division, parallel laminated "D" division, and very fine grained, structureless "E" division top (Fig. 26). Most of the beds are AB(E) types. In the mid- to upper part of the "A" division, and rarely in the "B" division, many beds contain discrete pumice-rich layers (1 mm to 5 cm in width) composed of small (<10 mm) pumice granules (Fig. 26). Thin pumice-rich layers are discontinuous along strike, whereas thicker layers (≥5 cm) are continuous along strike for at least 5 m and show only minor variations in thickness. Grading, defined by an upward decrease in the size of plagioclase crystal fragments, is continuous from base to top of beds. Note that this normal size grading is continuous across the pumice granule layers, indicating that the layers are integral parts of the beds and do not represent separate depositional events.

Vick Lake tuff is interpreted to have erupted in a subaerial environment, following a period of extensive calc-alkaline island arc and arc-rift volcanism. Ash particle morphology and the abundance of pumice suggests that the tephra was produced in shallow water or subaerial phreatomagmatic and plinian eruptions (Fig. 27). The pyroclastic material likely entered the water column from base surges, pyroclastic flows and fallout. It was transported by turbulent subaqueous density currents from the flanks of the subaqueous portion of the cone into an adjacent back-arc basin, forming Bouma-type “turbidite” bedforms. The presence of beds up to 19 m thick suggests that the sequence is proximal to source vents. A number of individual beds may have been deposited during a single eruption event, but the 900 m thick sequence represents a large number of eruption events relatively closely spaced in

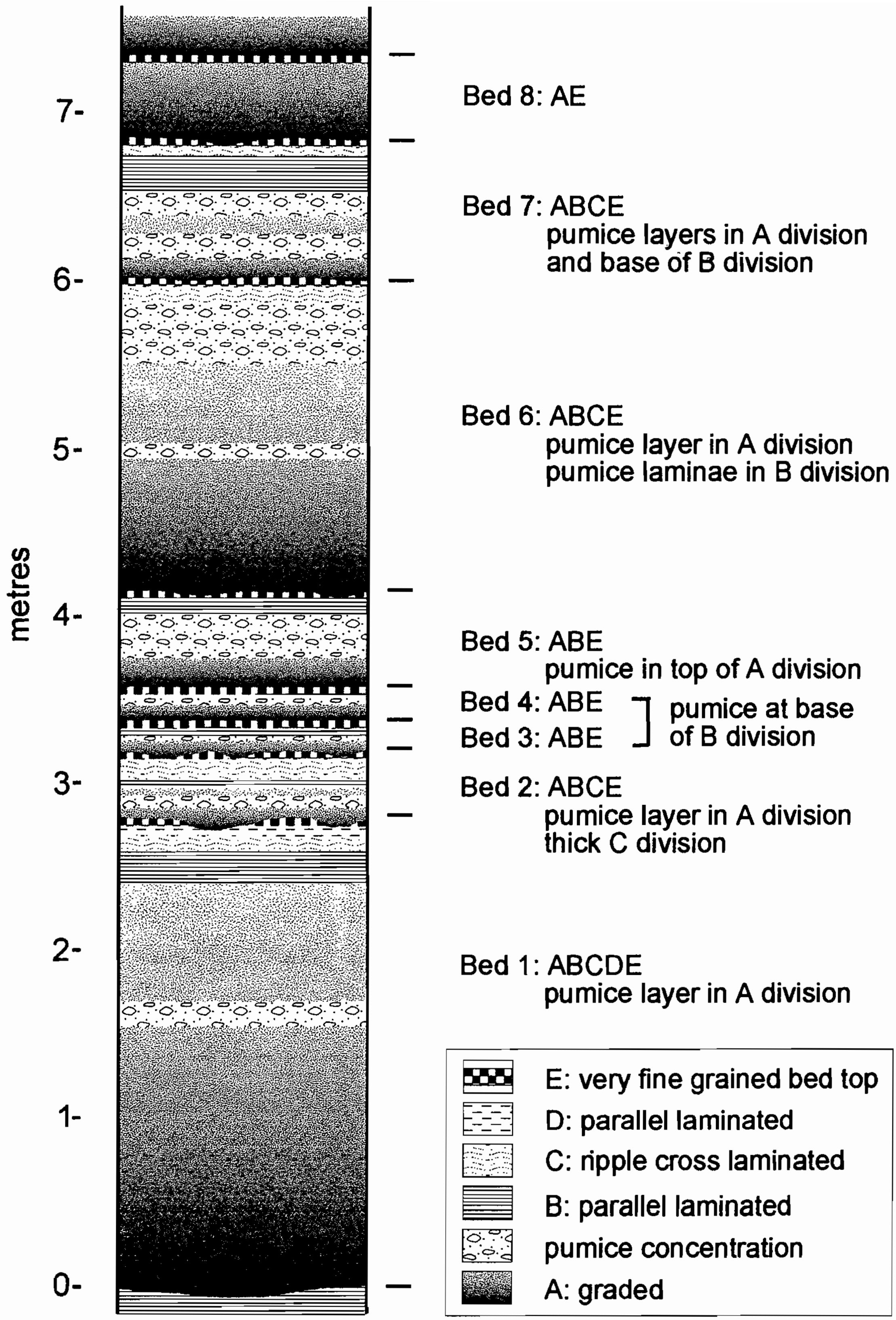


Figure 26: Sequence of 8 measured beds, Stop 8, Vick Lake shoshonite tuff. The tuff is composed of particles (glass shards, crystal fragments, pumice) produced by explosive volcanic processes in a shallow subaqueous to subaerial environment. Turbidite bedforms indicate that this pyroclastic material was deposited from turbulent subaqueous density currents, probably in a marine basinal setting.

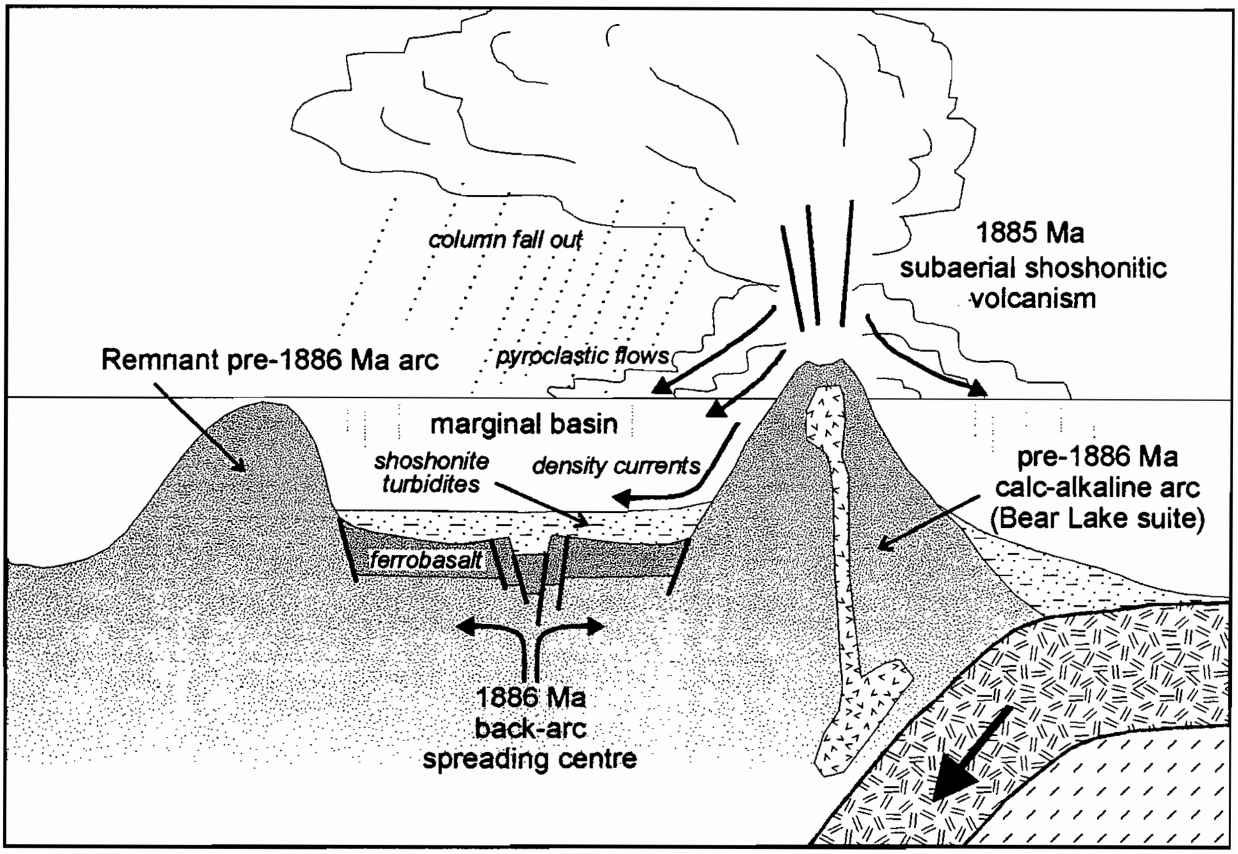


Figure 27: Model for the tectonic setting, eruptive mechanism and depositional environment of the Vick Lake shoshonite tuff (Stop 8). The morphology of constituent ash shards and pumice granules in Vick Lake tuff suggest the tephra was erupted in a shallow subaqueous to subaerial environment, following a period of extensive submarine calc-alkaline island arc and arc-rift volcanism. The shoshonitic material was deposited in an adjacent back-arc basin from turbulent density currents, giving rise to turbidite bedforms.

time. The general upward fining of the sequence, and gradual increase and ultimate dominance of pelagic sedimentary components in the upper 180 m of the unit indicates a gradual cessation of pyroclastic volcanism, with increasing proportions of graphitic shales deposited between eruptions. Modern analogues of the Vick Lake shoshonite suite (Stern *et al.*, 1995a) erupt late in the evolution of oceanic arcs (*e.g.,* Morrison, 1980; Gill and Whelan, 1989).

Stop description

Stop 8 is located approximately 700 m above the base of the unit. A sequence of eight beds, ranging from 0.20 to 2.80 m thick, are exposed at the top of the cliff-forming outcrop on old Highway 10 (Fig. 26). These beds include ABCDE, ABCE, ABE and AE types, most of which have pumice layers in the A or B division. Bases of beds are commonly scoured into the E divisions of underlying beds, in some instances removing all of the very fine grained bed tops (*e.g.*, base of bed 1, Fig. 26). Flame structures ornament the bases of some beds. Plagioclase crystals and crystal fragments display mesoscopic normal size grading (*e.g.*, bed 1: plagioclase microclasts are up to 1 mm at base, 0.75 mm at 1 m, and 0.5 mm at 2 m). The cream-weathering, very fine grained E division bed tops likely represent settling of the finest ash through the water column between eruptions, and are particularly prominent by virtue of their colour and grain size contrasts. Delicate ripple laminae are preserved in the C divisions of beds 1, 2 and 6.

Stop 9: Northeast Arm shear zone with lozenges of Archean tonalite

Introduction

The Northeast Arm shear zone is a 0.5-1 km wide deformation corridor within the Flin Flon assemblage that juxtaposes substantially different stratigraphic packages and contains tectonic units unique to the Amisk collage (Figs. 4, 23; Bailes and Syme, 1989; Lucas, 1993; David and Syme, 1994). It forms the eastern boundary to the juvenile calc-alkaline Bear Lake section and the western boundary to a section of tholeiitic (arc-rift?) basalts with relatively evolved initial ε_{Nd} compositions (Scotty Lake basalt; Bailes and Syme, 1989; Syme *et al.*, 1993; Lucas *et al.*, 1996). Three distinct tectonostratigraphic packages which display tremendous strike-length continuity have been mapped in relatively low strain domains within the shear zone, although they are transformed into mylonite at their margins. These packages are characterized by contrasting stratigraphy, geochemistry, Nd isotopic signature and structural history. The westernmost one includes the lozenges of tonalite-granodiorite dated at *ca.* 2.5 Ga (David and Syme, 1994) which are intruded by mafic dykes that may be related to the Scotty Lake basalts (Fig. 23). The central low strain package is dominated by mafic and felsic dykes and granodioritic intrusive sheets, with dyke compositions and grain sizes vary from rhyolitic/granitic to variably porphyritic basalt/diorite. The easternmost unit consists of a sequence of volcaniclastic rocks dominated by well-bedded lapilli tuff and breccia and plagioclase crystal tuff, with a calc-alkaline arc geochemical signature (*Northeast Arm volcaniclastics*: Bailes and Syme, 1989; Stern *et al.*, 1995a).

The tectonostratigraphic complexity of the Northeast Arm shear zone and the presence of low strain domains and late brittle faults makes the detailed correlation of structures difficult. However, the overall structural history of the Northeast Arm shear zone can be described by three principal deformation events (Fig. 22, Table 1): **D_1**: S_1/L_1 foliated wallrocks, tectonites and mylonites (equivalent to Inlet Arm S_1 foliation); **D_2**: F_2/S_2, developed principally in the easternmost Bear Lake block and manifest at map-scale by the Vick Lake synform; and **D_3**: consistent Z-asymmetry metre-scale folds (F_3) and associated crenulation cleavage (S_3), and a major mylonite band forming the western boundary to the shear zone. S_1 represents a regional, penetrative metamorphic foliation that is associated with a moderately north- or south-plunging extension lineation (Fig. 22), and is axial planar to rare folds of primary layering. Within the Northeast Arm corridor, S_1 mylonite bands separate the wall rocks and internal low strain packages, varying in width from metres to hundreds of metres.

D_2 structures are well developed within and adjacent to the Northeast Arm shear zone. The F_2 Vick Lake synform is a relatively tight structure, plunging NNE at 36° (*Northeast Arm syncline* of Bailes and Syme, 1989), that folds primary layering and S_1 (Fig. 22). The synform occurs solely in the Bear Lake section, as its eastern limb is cut-off by the S_3 mylonite band that marks the western boundary to the Northeast Arm shear zone. Minor F_2 folds vary from chevron style to isoclinal and are associated with a NNE-striking, subvertical S_2 cleavage (Fig. 22) which locally transposes S_1 and primary layering. Intense S_1/S_2 strain at the eastern edge of the Bear Lake section suggests that the actual Northeast Arm shear zone involved Bear Lake units and that shear zone deformation spanned the D_1 and D_2 episodes. Felsic to mafic intrusive sheets occur throughout the Northeast Arm corridor, and both predate and postdate D_1 and D_2 deformation. A composite pyroxenite-quartz diorite pluton, interpreted as a post-accretion arc intrusion (*i.e.*, 1.87-1.84 Ga), cuts mylonites (S_1/S_2) derived from the western Northeast Arm package containing the dated Archean rocks.

The geometry of Northeast Arm and Inlet Arm shear zones during D_2 is an important question, albeit not easily resolved due to overprinting deformation. The D_1 episode may have generated moderate to steep dips, although the enveloping surface to the F_2 folds in the Inlet Arm-Northeast Arm area (Fig. 22) suggests a shallow (~30°) dip for S_0/S_1 prior to F_2. Steep belts were generated during F_2 as a result of rotation of S_0/S_1 and S_2 foliation development. The fact that S_2 high strain zones are asymmetrically developed with respect to F_2 structures (*e.g.*, Vick Lake synform/Northeast Arm shear zone) suggests that the F_2/S_2 steep belts may have accommodated further D_2 deformation.

D_3 structures can be traced across the Bear Lake section from the Inlet Arm shear zone to the Northeast Arm shear zone and eastward into the Scotty Lake block (Lucas *et al.*, 1995, 1996). The most conspicuous structures are metre-scale, chevron-style folds (F_3) with a consistent Z-asymmetry and an associated crenulation cleavage (S_3). F_3 folds are upright and plunge both north and south (Fig. 22), with microfolds occasionally showing both north and south plunges on the same outcrop. The variable plunge of the F_3 folds suggests either (1) deformation of complex pre-D_3 structure, (2) rotation of F_3 axes during progressive D_3 deformation, and/or (3) post-D_3 folding about gently west-plunging axes. The crenulation cleavage varies from spaced to penetrative and generally strikes north and dips steeply to the east or west (Fig. 22).

Outcrop- and map-scale S_3 high strain zones occur throughout the Bear Lake section and Northeast Arm corridor, and are marked by development of an intense S_3 crenulation cleavage and, ultimately, a differentiated S_3 layering with a chlorite-carbonate mineral assemblage. The high strain zones have a NNE-orientation, thus suggesting a dextral C-S relation with the S_3 foliation (see top inset circle, Fig. 22), consistent with the Z (dextral) asymmetry of the F_3 folds. An 5-20 m wide S_3 mylonite strand forms the western margin of the Northeast Arm shear zone and truncates the F_2 Vick Lake synform (Fig. 23, 22). The D_3 structures are thought to be associated with dextral transcurrent shear localized along the Northeast Arm corridor and transpression of both the Northeast Arm and its wallrocks (Lucas *et al.*, 1995, 1996). Post-D_3 structures include conjugate kink bands, box folds, ductile to brittle shear zones and conjugate brittle fracture sets (NW-sinistral, NE-dextral; Lucas, 1993).

Stop description

Two stops will be made to examine lithologies and structures associated with the Northeast Arm shear zone (Fig. 22, 9a, 9b). The first stop (9a) will feature a rather extensive section of outcrop that includes one of the dated Archean tectonic slices (David and Syme, 1994), as well as mafic tectonites (S_1, S_2) most likely derived from basalt-gabbro sequence. The second stop (9b) highlights D_3 structures developed in laminated felsic-mafic tectonites (D_1-D_2) of the Northeast Arm shear zone.

Stop 9a:

The first outcrop is a variably-deformed granodiorite-tonalite that is cut by mafic dykes, themselves strongly deformed. U-Pb zircon geochronology of this body has yielded an age of 2518 Ma (David and Syme, 1994) and with an average initial ε_{Nd} composition of -6.9 (Stern *et al.*, 1995a). There are a number of fault-bound lozenges (10-100 m wide x 100's of metres long) within the Northeast Arm shear zone (Fig. 22; Lucas, 1993), all intruded by mafic dykes that may be related to a sequence of tholeiitic (arc-rift?) basalts with relatively evolved initial ε_{Nd} compositions (-1 to +2 at 1900 Ma; *cf.* Stern and others, 1995b) found immediately east of the shear zone (Scotty Lake basalt, Stop 10; Fig. 4).

The Archean tonalite-granodiorite is in tectonic contact with mafic rocks to the west along the lake shore. Note the highly deformed nature of the mafic rocks, even in contrast to the tonalite-granodiorite, and the evidence for multiple foliations in the mafic tectonites. Felsic dykes cut the mafic rocks and are themselves deformed, displaying evidence for asymmetric boudinage and back-rotation consistent with dextral shear (D_3?). The mafic rocks have a composite S_1-S_2 foliation, and as the D_3 fault contact with the Bear Lake section (Mikanagan sill, Fig. 22) on the west side of Beaverhouse Lake is approached, it is overprinted by an S_3 crenulation cleavage that eventually completely transposes the older fabric. The S_3 crenulation cleavage is highlighted in Stop 9b.

Stop 9b:

The unit bounding the Archean slices/mafic tectonite package to the east is a laminated mafic-felsic tectonite that appears to be largely derived from syn-tectonically emplaced intrusive sheets. These vary in composition from aphyric and quartz phyric felsic ('rhyolite') to tonalite-granodiorite to mafic-intermediate (basalt-andesite), and are inferred to be 'post-accretion arc' (1.87-1.84 Ga) or syn-D2 (Table 1) in age. At this outcrop, the intense strain precludes identification of primary textures and intrusive relations in general; most intrusive sheets/veins are at a low angle to foliation and compositional layering. However, observations in lower strain domains of the same unit along the Northeast Arm shear zone show subparallel mafic and felsic veins of varying deformation state, suggesting that at least some of the veins are emplaced at a low angle to compositional layering and that D_2 deformation and magmatism were broadly coeval.

The most striking feature of this outcrop is the well-developed S_3 cleavage, which is oblique to the composite S_1-S_2 foliation and compositional layering, and axial planar to F_3 folds. The steeply-dipping S_3 crenulation cleavage is oriented approximately 20-30° counterclockwise (northward) with respect to layering (*cf.* inset circle, Fig. NE Arm) and almost forms a differentiated cleavage. The moderately to steeply north-plunging F_3 folds show a consistent Z-asymmetry at this outcrop and along the length of the Northeast Arm corridor, suggesting dextral shear. As discussed above, D_3 deformation appears to be localized in narrow high strain zone that truncates stratigraphy and structures in the Bear Lake section to the west (Figs. 22, 23). The penetrative S_3 cleavage in this high strain zone is clockwise (NNE) from S_3 at this outcrop, suggesting a dextral 'C/S fabric' relationship at map scale. The S_3 cleavage is deformed by kink-bands that form a conjugate set.

Stop 10: Scotty Lake basalt

Introduction

Stratigraphic packages that differ substantially in tectonic setting and isotopic character are juxtaposed by corridors of deformation within the Flin Flon assemblage. One such corridor is the Northeast Arm shear zone (Figs. 4, 23: Stop 9). It forms the eastern boundary to the juvenile calc-alkaline to shoshonitic Bear Lake section (Stops 7, 8) and the western boundary to a section of tholeiitic (arc-rift?) basalts with relatively evolved initial ε_{Nd} compositions (Scotty Lake basalt; Bailes and Syme, 1989; Syme *et al.*, 1993). The purpose of Stop 10 is to contrast these relatively evolved arc-rift MORB-like basalts with the juvenile oceanic rocks west of the Northeast Arm shear zone.

Lithologic components in the Scotty Lake section record arc rifting and rift sedimentation. The dominant unit, Scotty Lake basalt, forms a homoclinal, west-facing sequence 1350 m thick. To the east and southeast the basalt is truncated by a relatively young fault ("old Highway 10 fault") and to the northeast, at Whitefish Lake, it is terminated by the Scotty Lake fault (Bailes and Syme, 1989). Scotty Lake basalt is overlain by a thin unit of sedimentary rocks (Fig. 23; Bailes and Syme, 1989), marking the end of arc-rift volcanism in the section. The sediments are in turn overlain by intermediate to mafic, porphyritic volcaniclastic breccias plausibly related to arc-type volcanism (Syme, Lucas and Stern, unpublished map). The stratigraphic position setting of the sedimentary and volcaniclastic rocks suggests that they are rift-fill deposits.

Scotty Lake basalt is composed of aphyric, rusty brown weathering, Fe-rich basalts and related synvolcanic dykes and sills. Four main flow types are present: 1) simple massive flows, 2) massive flows with hyaloclastite/amoeboid pillow breccia tops, 3) compound flows with massive bases, narrow pillowed divisions and hyaloclastite tops, and 4) pillowed flows. Pillowed flows are restricted to particular stratigraphic subunits, and are best exposed at Scotty Lake and along Highway 10 (Fig. 23; see also Bailes and Syme, 1989).

Scotty Lake basalt contains a suite of aphyric and porphyritic, mafic to intermediate, non-vesicular dykes which locally compose up to 100% of exposures. The dykes are entirely confined to Scotty Lake basalt and do not occur within the overlying sedimentary or volcaniclastic formations. These dykes range in width from a few centimetres to 150 m, and, although many are subconcordant to stratigraphy, cross-cutting relationships are common. Equigranular basalt dykes that are similar to host basalts in weathering colour, grain size and composition dominate the hypabyssal assemblage. These dykes have a hypidiomorphic granular texture comprising a mosaic of tabular, randomly oriented plagioclase, stubby pyroxene, interstitial quartz, and disseminated subhedral Fe-Ti oxides.

Stop description

At this location we will examine a dominantly pillowed part of the Scotty Lake succession. Pillows are 0.3-2 m in maximum dimension and are ovoid, bun-shaped or irregular in shape. Selvages are narrow (0.5-1 cm) and thermally fractured. Some pillows have spherulitic margins. Vesicularity is more pronounced in pillowed flows than in massive flows, but nevertheless amygdales compose <10% of pillows. Amygdales (1-5 mm) occur in the spherulitic marginal zones of some pillows and in the cores of others.

Hyaloclastite, best developed at the tops of massive Scotty Lake basalt flows, is locally present in the interpillow voids of pillowed flows. Hyaloclastite granules are bounded by curviplanar thermal fracture surfaces and are replaced by quartz, epidote, chlorite, actinolite and leucoxene. These secondary minerals occur in zones that pseudomorph the pre-metamorphic distribution of sideromelane glass and secondary palagonite (Bailes and Syme, 1989).

Stop 11: Athapapuskow basalt

Introduction

The Amisk collage (Fig. 2) is composed of two areally extensive 1.9 Ga assemblages: the arc assemblage (*e.g.*, Stops 1, 2, 6, 7, 8) and ocean-floor assemblage (Stops 4, 11). At this Stop we will examine a member of the ocean-floor assemblage at Cranberry Portage (Fig. 28).

Juvenile ocean-floor assemblages include a number of distinct basalt sequences and related mafic-ultramafic complexes (Stern *et al.*, 1995b). The basalts occur in a semi-continuous belt between the Elbow Lake and Athapapuskow Lake, and are collectively termed the 'Elbow-Athapapuskow assemblage' (Fig. 2; Stern *et al.*, 1995b). The Elbow-Athapapuskow ocean-floor assemblage is up to 25 km wide and has a strike

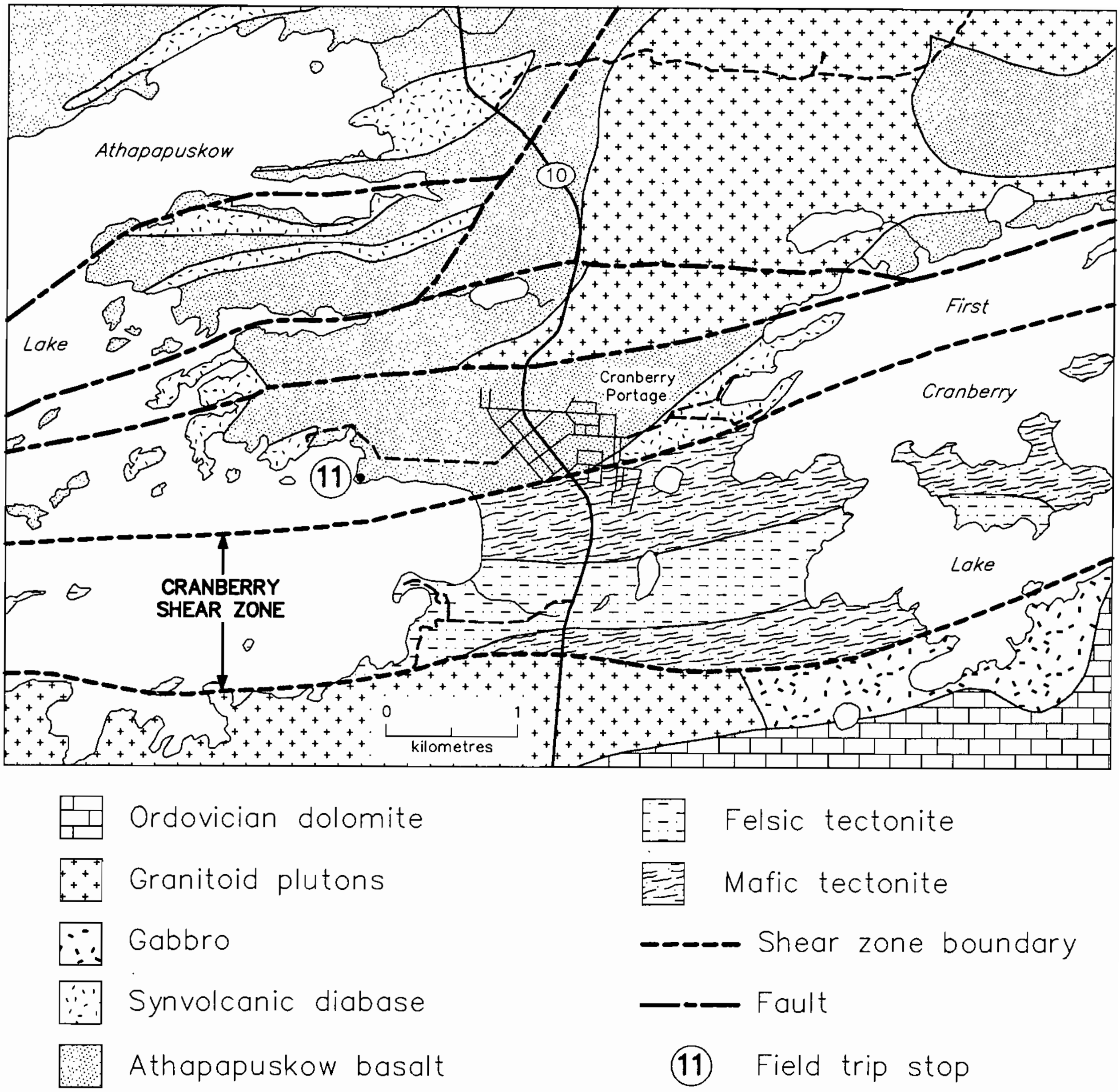


Figure 28: Simplified geological map of the Cranberry Portage area, modified from Podolsky (1951, 1958) and Syme (1993). The area is underlain by Athapapuskow basalt, a member of the Elbow-Athapapuskow ocean-floor assemblage, which was intruded by granitoid plutons and subsequently transected by the Cranberry shear zone (Syme, 1993, 1995; Ryan and Williams, 1996). Stop 11: outcrop of Athapapuskow basalt sheet flows, located on the property of Athapap Lodge, 145 Brydges Ave. NW, Cranberry Portage.

length of 100 km (Syme and Bailes, 1993; Syme, 1994; Stern *et al.*, 1995b); it extends beneath the Phanerozoic cover for at least another 75 km (Leclair *et al.*, 1993). This assemblage is everywhere either in fault contact with arc volcanic rocks, or the contact with arc rocks is stitched by younger plutons. Systematic mapping of the assemblage (Syme, 1988, 1991, 1992, 1993, 1994) has shown it to consist entirely of subaqueous basalt and related intrusives, in which there are no known occurrences of felsic volcanic rocks, terrestrial sediments, or older crystalline basement. Gabbro and diabase dykes and sills are common, although no sheeted dykes are known.

Ocean-floor basalts in the Elbow-Athapapuskow assemblage occur as laterally coherent units, 4 km to >60 km in strike length, that have stratigraphic thicknesses of 0.3-3.0 km, each having characteristic weathering colour, flow morphology, alteration assemblage and geochemistry (Syme, 1991, 1992, 1995; Stern *et al.*, 1995b). Such basalt units are considered informally as 'formations' (*e.g.,* Athapapuskow basalt, Millwater basalt, Fig. 9). Most display abundant evidence for seafloor hydrothermal alteration (*e.g.*, epidosite domains and veins, interpillow chert; Syme, 1991, 1992), consistent with eruption in a ridge setting. All rocks have regional greenschist grade metamorphic mineral assemblages (actinolite-chlorite-epidote-albite), locally overprinting amphibolite facies assemblages associated with the contact aureoles of 1.87-1.84 Ga plutons.

At Stop 11 flows assigned to the Athapapuskow basalt 'formation' are exposed (Fig. 28) on the property of Athapap Lodge. Athapapuskow basalt is named for its exposure in the large south basin of Athapapuskow Lake, but the unit trends northeast for 50 km through the Cranberry lakes (Syme, 1988, 1993), and southwest under Paleozoic cover (Leclair *et al.*, 1993). Athapapuskow basalt is a minimum of 1-2 km thick, and contains abundant diabase, gabbro, and rare ultramafic sills. The age of Athapapuskow basalt is constrained by the U-Pb zircon age of a contained synvolcanic diabase sill (1904 ±4 Ma, Stern *et al.*, 1995b). Athapapuskow basalt is thus coeval with the arc assemblage tholeiitic package that hosts the Flin Flon and Callinan massive sulphide deposits (1904 +6/-3 Ma, Mine rhyolite; David *et al.*, 1993).

Athapapuskow basalt has the geochemical characteristics of a "primitive" E-type MORB (Stern *et al.*, 1995b). These basalts have higher TiO_2 than the N-types and are comparatively enriched in Th, Nb and LREEs (Fig. 9). The E-type Athapapuskow basalts are the least fractionated of all the ocean-floor basalts, with several samples having 9-12 wt.% MgO and 150-300 ppm Ni; their 'enriched' trace element signatures compared with N-type basalts thus cannot be explained by fractional crystallization. N-MORB-normalized trace element patterns for Athapapuskow basalts have smooth negative slopes (Fig. 9), and plot in the E-MORB field on a Zr-Nb-Th basalt discrimination diagram (Stern *et al.*, 1995b). Th/Nb ratios in the basalts (<0.1) show no evidence of either a continental crustal or arc signature. These basalts have neither passed through sialic crust nor have been derived from metasomatised sub-arc mantle; this conclusion is confirmed by their strongly positive initial ε_{Nd} values (+3.1 to +3.8; Stern *et al.*, 1995b).

Stop description

Athapapuskow basalt flows occur in three different flow morphologies (Syme, 1988). Thin (1.5-5 m) massive sheet flows are the dominant flow type. They have chilled bases, non-amygdaloidal basal zones, highly vesicular central zones and flow tops containing large (to 2 cm) amygdales. Thick massive flows (>30 m) contain only sporadic 2-8 mm amygdales but have strongly amygdaloidal flow tops. Pillowed flows are rare, and occur intercalated with the other flow types. The eruptive environment is subaqueous, demonstrated by the occurrence of the intercalated pillowed flows. Epidote-dominated alteration features such as ovoid epidosite domains, epidote-filled amygdales, epidotized amygdaloidal flow tops, and epidote veins are common. These features are consistent with synvolcanic hydrothermal alteration in the porous portions of subaqueous flows.

A sequence of six thin (1-1.5 m) massive basalt flows exposed at Stop 12 (Fig. 29) are characteristic of the dominant, thin-flow facies of Athapapuskow basalt. Flow contacts are sharp, planar and in some instances recessive weathering. Bases of flows are chilled against underlying flows, and pipe vesicles to 10 cm long occur at the bases of some of the thicker flows. Upper parts of flows are strongly amygdaloidal (to 1 cm) and are characterized by intense epidote alteration. Flow tops are planar to slightly ropy.

Stop 12: Missi suite stratigraphic and structural relations with Amisk collage

Introduction

The Missi suite continental sedimentary rocks of the central Flin Flon belt (Fig. 4) form a cover sequence to the 1.92-1.88 Ga tectonostratigraphic assemblages (Bruce, 1918; Ambrose, 1936; Stockwell, 1960) similar to that of the Temiskaming sequences in the Superior Province. Key features of the Missi suite siliciclastic rocks are: (1) unconformable deposition on deformed rocks of the Amisk collage as well as on successor arc plutons; (2) development of an oxidized paleosol (regolith) at the unconformity (Holland *et al.*, 1989); (3) removal of significant stratigraphic section along the (angular) unconformity (ca. 2 km; Bailes and Syme, 1989); (4) presence of clasts derived from the 1.92-1.88 Ga assemblages (*e.g.*, pillowed basalt, iron formation), successor arc plutons (medium- to coarse-grained granitoid rocks) and jasper (Bailes and Syme, 1989; Syme, 1987; Stauffer, 1990); and (5) rare trachyandesite sills (Syme, 1988); and (6) laminated argillites (Syme, 1988). Together, these features suggest that the Missi suite sedimentation occurred during post-accretion arc magmatism on an uplifted and deeply incised terrain (Bailes and Syme, 1989; Stauffer, 1990). Depositional environments included alluvial fans and braided river systems. It is likely that D_2 structures (*e.g.*, folds, steep belts, shear zones) and associated topography controlled the pattern of fluvial drainage systems and associated Missi suite sedimentation at 1.85-1.84 Ga (Lucas *et al.*, 1996).

The Missi sandstones at Flin Flon were deposited at *ca.* 1845 Ma, bracketed by the age of the youngest detrital zircon (1847 Ma; Ansdell *et al.*, 1992; Ansdell, 1993) and the oldest cross-cutting intrusion (1842 ±3 Ma Boundary Intrusion; Heaman *et al.*, 1992). The unconformity cuts through a significant amount of basement section and is

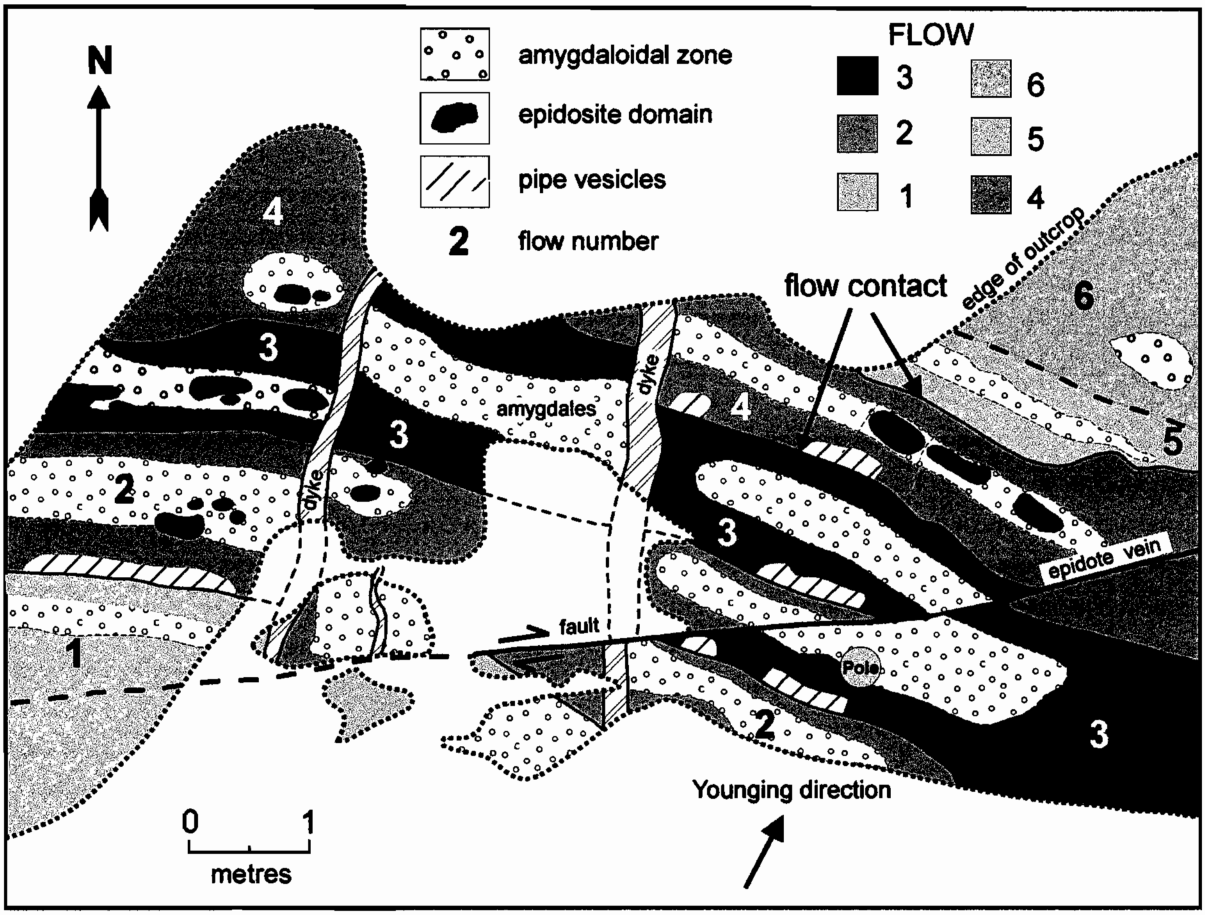


Figure 29: Outcrop map for Stop 11, Athapapuskow basalt. The local volcanic stratigraphy youngs to the north. Six massive basalt flows (numbered 1 to 6) are completely or partially exposed, ranging in thickness from 1 to 1.5 m. Flows 3 and 5 are plagioclase phyric and the remainder are aphyric. Thin sheetflows such as these are typical of Athapapuskow basalt, but thick (e.g., 30 m) massive flows and pillowed flows also occur in this formation. These thin flows commonly have chilled bases with pipe vesicles, non-amygdaloidal lower zones, highly amygdaloidal centres, and weakly amygdaloidal tops. The flow top surface is in some instances crenulated or ropy (e.g., top of flow 4). Epidosite domains are concentrated in the amygdaloidal (porous) parts of the flows and are interpreted as synvolcanic seafloor alteration features. Aphyric basalt dykes cut across the flows at a high angle; their relationship to the flows is unknown. This outcrop lies 500 m north of the margin of the Cranberry shear zone, and contains a dextral fault occupied by an epidote vein. Some flow contacts (e.g., between flows 5 and 6) are sheared.

markedly angular (Stockwell, 1960; Bailes and Syme, 1989), suggesting that the basement was deformed prior to or during erosion and sedimentation, consistent with regional constraints (Lucas *et al.*, 1995, 1996). The Flin Flon arc assemblage basement and Missi sedimentary cover rocks at Flin Flon were deformed and metamorphosed at greenschist to lower amphibolite grade conditions (Ambrose, 1936; Bailes and Syme, 1989; Digel and Gordon, 1995). Fedorowich *et al.* (1995) present biotite and hornblende Ar-Ar data that indicate peak metamorphism at 1820-1790 Ma in the Flin Flon area, coeval with regional peak metamorphism and deformation across the Trans-Hudson Orogen (*e.g.*, Gordon *et al.*, 1990; Ansdell and Norman, 1995; David *et al.*, 1996).

The Missi suite rocks at Flin Flon are preserved in a large fold interference basin that is cut by NNW-trending faults (Fig. 15). Stauffer and Mukherjee (1971) documented two generations of folds and one regional foliation in the Missi cover sequence within the study area as well as a number of dominantly NNW-trending, steeply-dipping oblique slip faults. The oblique slip faults are associated with both ductile (*e.g.*, Channing Fault) and brittle-ductile (*e.g.*, Ross Lake fault) fault zone structures and range in age from 1.82 to 1.69 Ga (Fedorowich *et al.*, 1995). Recently, Gale *et al.* (1996) documented three generations of folds (F_1, F_2, F_3) and associated foliations (S_1, S_2, S_3) in the southern part of the Missi basin. The absolute age of these structures is not resolved (all post-date the 1842 Ma Boundary Intrusions) but it is thought that the F_1 and F_2 structures are associated with regional D_3 deformation, the F_3 folds with D_4 deformation, and the ductile-to-brittle NNW-trending faults with D_3 to D_5 deformation (*e.g.*, Cliff Lake fault, Ross Lake fault: Fig. 15; Bailes and Syme, 1989; Gale *et al.*, 1996).

The S_1 foliation is not extensively developed throughout the study area and tends to be preserved in isolated lenses of fine-grained shale or thinly laminated tuff beds (Gale *et al.*, 1996). A number of F_1 folds have been identified on the basis of facing reversals, map pattern and overprinting by the S_2 regional foliation; only one macroscopic F_1 fold has a well developed S_1 axial planar foliation (Missi outlier, Fig. 30). Examples of F_1 folds include the Ross Creek syncline and the Flin Flon Creek syncline (Fig. 15; Stauffer and Mukherjee, 1971).

The S_2 foliation is a penetrative cleavage that generally strikes north to northeast and dips shallowly to moderately to the east. In the southern portion of the Missi basin, S_2 has a northerly attitude with a steeper eastward dip. Throughout the Flin Flon area, S_2 represents the principal flattening fabric in the Missi cover sequence (Stauffer and Mukherjee, 1971; Gale *et al.*, 1996), and developed at peak metamorphic conditions (Ambrose, 1936; Bailes and Syme, 1989; Fedorowich *et al.*, 1995). S_2 is axial planar to the F_2 Pipeline syncline (described below) and Mud Lake syncline (Fig. 15; Stauffer and Mukherjee, 1971). The S_3 foliation is a spaced, domainal cleavage that strikes northwest, dips steeply to moderately northeast and is heterogeneously developed throughout the area (Gale *et al.*, 1996). It is axial planar to rare F_3 folds.

A regional lineation characterizes the northern half of the Missi basin in the Flin Flon area, occurring in both the sedimentary cover and volcanic basement (Stauffer and Mukherjee, 1971; Bailes and Syme, 1989; Gale *et al.*, 1996). It is defined by the elongation of pillows, clasts within the conglomerate and volcaniclastic beds, and quartz and feldspar grains in the matrix of the sandstones and granites, as well as by the intersection of S_2 and S_3. In areas where the lineation is well developed, conglomerate clasts display aspect ratios of up to 7:1 (Stauffer, 1990). Stauffer and Mukherjee (1971) and Fedorowich *et al.* (1995), who associate the lineation with S_2 and growth of peak metamorphic assemblages (*i.e.*, L_2). An important feature of the Flin Flon area structures is the parallelism of virtually all linear features, including mesoscopic and macroscopic fold axes, elongation lineations, mineral lineations, and intersection lineations. Gale *et al.* (1996) suggest that the parallelism of these structures is independent of strain intensity, suggesting that the fold hinges have not been rotated into parallelism with the extension direction.

Stop description

A small outlier of Missi suite sedimentary rocks occurs just to the south of the principal Missi basin (Fig. 15) and contains evidence for all three deformation events that affect the Missi suite rocks (Gale *et al.*, 1996). Macroscopic folds are easily delineated by tracing two distinctive conglomerate units in the dominant pebbly sandstone facies (Fig. 30), while mesoscopic folds are well developed in beds with cm-scale hematite laminations. S_1 is a bedding-parallel cleavage that is axial planar to an east-west trending F_1 fold but is only developed in rare fine-grained shale lenses between sandstone beds and within the regolith developed at the unconformity. S_2 crenulates S_1 and is axial planar to open to tight folds which trend north-south and have axial planes that dip moderately to the east. S_3 cleavage strikes northwest and dips steeply to the northeast but is not obviously related to any macrosopic folds in the outlier. Clasts within the sandstone and conglomerate beds are flattened in S_2 where S_3 is absent, but are aligned in S_3 where it is developed. The Missi units in the outlier do not contain an extension lineation.

Gale *et al.* (1996) inferred that the outlier's basinal shape is a result of a type 1 interference pattern (Ramsey, 1967) caused by F_1 and F_2 folding. The macroscopic F_1 syncline trends east-west and has an overturned northern limb and a steeply north-dipping axial plane (Fig. 30). The southern limb generally dips moderately to the north but is locally overturned to the south due to F_2 folding (Fig. 30). These observations suggest that the F_1 fold has a southward vergence, but it is most likely a second-order fold on the flanks of a larger F_1 anticline that closes to the south (Gale *et al.*, 1996).

Key points to observe at this Stop include:

- Unconformity between Flin Flon arc assemblage basalts and overlying Missi suite sandstone. The unconformity is characterized at this location by a 1-2 m thick section of purple, hematiferous regolith derived from the basalts. The base of the Missi section contains a lag of basalt cobbles, and is overlain by cross-bedded sandstones.
- Overprinting cleavage relations (S_1, S_2 and S_3) in fine-grained shale layer.
- Macroscopic F_2 folds of bedding with S_2 as the axial planar cleavage.

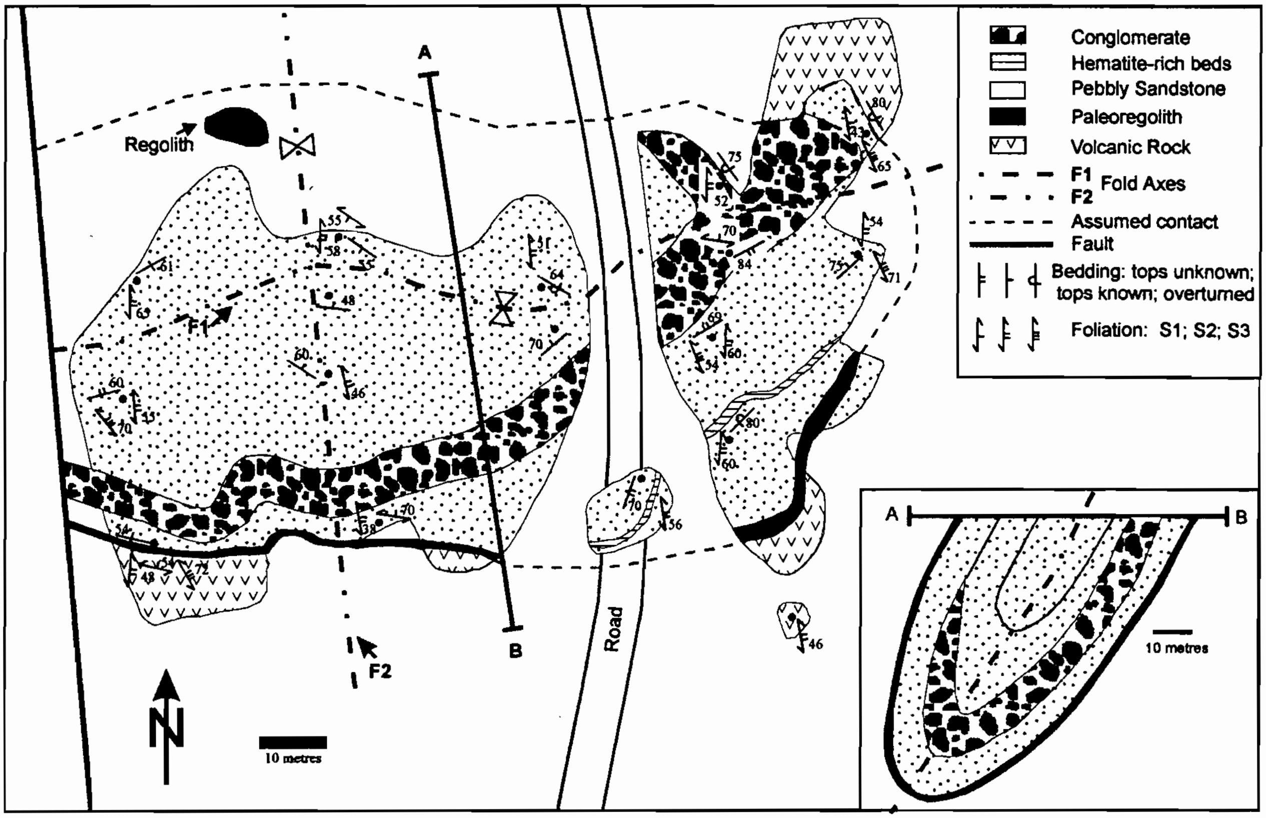


Figure 30:Map of the Missi suite outlier south of the principal Flin Flon Missi fold interference basin. Inset shows a schematic cross-section of the outlier. D. Gale (unpublished).

- Mesoscopic F_2 folds developed in hematite laminations. Note the Z-asymmetry and the first-order relationship between fold wavelength and the layer thickness.

Stop 13: Annabel Creek shear zone

The following description of the setting and characteristics of the Annabel Creek shear zone was contributed by Ken Ashton (Saskatchewan Geological Survey).

Introduction

The Annabel Creek section transects the F_3 Annabel Lake Synform (ALS; Byers and Dahlstrom, 1954), which exposes most of the rock types that make up the high-grade northwestern part of the Flin Flon volcanic belt (*i.e.,* Attitti Block). Annabel Lake and the creek extending to the east mark a lineament produced by the D_4 Annabel Lake Shear Zone (ALSZ), which is about 1 km wide at this point (Figure 31). The ALS is continuous with the F_2/F_3 Flin Flon Basin to the east, but has been highly attenuated and rotated into this north-dipping orientation by displacement along the ALSZ. The D_{4-5} Embury Lake Antiform was at least passively initiated by this displacement, although it may have been tightened by later northwest-southeast shortening.

Byers and Dahlstrom (1954) placed the boundary between the Flin Flon and Kisseynew domains several kilometres to the north where primary features became obliterated due to the effects of metamorphism and deformation. Parslow and Gaskarth (1985) suggested that the ALSZ (they used the name Pargas Shear Zone) be used to mark the domain boundary but it becomes more difficult to trace in the increasingly recrystallized rocks between Maskunow and Granite Lake and certainly does not truncate the volcano-plutonic units of the Flin Flon Domain to the west where they are folded around the D_{4-5} Wildnest Lake Synform. In fact, by tracing these units, the Flin Flon volcanic belt can be shown to be continuous with higher-grade equivalents to the northwest (Ashton *et al.*, 1987), which were historically placed in the Attitti Block (Macdonald, 1981). Based on this relationship, the boundary with the Kisseynew gneiss belt has been pushed about 10 km northward where it represents the transition from a dominantly volcano-plutonic terrain in the south to the Burntwood wacke/Missi arenite-dominated Kisseynew terrain to the north.

The northwestward increasing metamorphic gradient, which developed during D_3 time, is steepened in this area by displacement along the D_4 ALSZ. The isograd marking the first appearance of the assemblage sillimanite-garnet-biotite in wackes approximately coincides with this steep gradient. The metamorphic grade throughout this section is lower-middle amphibolite facies. Missi rocks on the north shore of Annabel Lake about 1 km to the west contain staurolite. Many of the mafic rocks in the section contain hornblende-garnet assemblages and the wackes contain rare sillimanite-garnet-biotite. Therefore, the ALSZ is not a sharp discrete discontinuity separating low-grade Flin Flon volcanic belt rocks from high-grade Kisseynew gneiss belt rocks.

From a regional tectonic perspective, the ALSZ is thought to have formed *ca.* 1805 Ma in response to collision between the largely Archean Sask Craton, seismically profiled by LITHOPROBE (Lewry *et al.*, 1994) and exposed in the Pelican Window

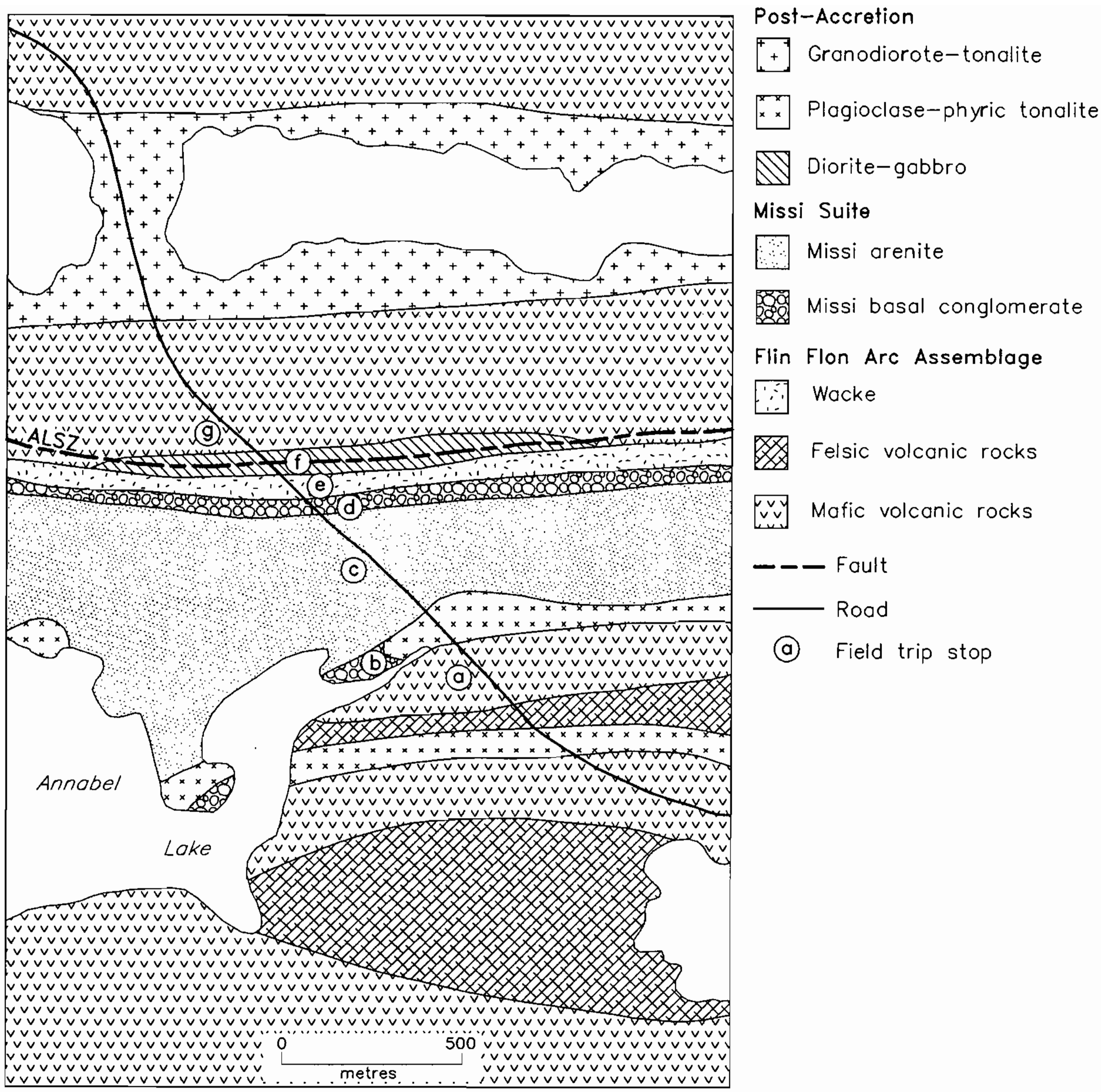


*Figure **31**: Generalized geology in the area of Stop 13, Annabel Creek shear zone (ALSZ). Modified from Thomas (1993).*

(Lewry *et al.*, in prep.), and a volcano-plutonic protocontinent spanning the once-continuous Flin Flon-Hanson-Glennie belt (Ansdell *et al.*, 1995). The collision caused rocks of the protocontinent to be thrust southwestward over the Archean craton along the Pelican Décollement Zone (PDZ), accounting for the westward increase in metamorphic grade and depth of crustal exposure approaching the Pelican Window from the east (Ashton and Digel, 1992). The ALSZ developed late during the collisional event as an accommodation zone allowing the hot, ductile rocks to the north to slide southwestwards past the cold, brittle low-grade rocks to the south. The apparent volume problem in the north is accommodated by differential displacement along the Sturgeon-weir Shear Zone, which is rooted in the PDZ, and by extensive D_{3-4} folding.

Stop description

A number of short stops are planned along both sides of the Hanson Lake road (Fig. 31).

(a) Start at the large outcrop of mafic rocks on the west side of the road south of the Annabel Creek bridge. Thomas (1993) mapped these as part of a plagioclase-phyric basalt unit containing quartz-feldspar amygdales. They were assigned the stratigraphic term Triangle Lake basalts and are considered to be part of the Flin Flon Assemblage.

(b) Go back to the highway and cross the bridge over Annabel Creek to the parking area. Several small outcrops of basal polymictic conglomerate and interlayered arenite belonging to the 1848 (Andsell, 1993) to 1842 ±3 Ma (Heaman *et al.*, 1992) Missi suite are intruded by transposed dykes of tonalitic plagioclase porphyry. The conglomerate is essentially the same as that seen at the Missi unconformity (Stop 12) but the effects of deformation and metamorphism make recognition a little more difficult. Note the doubly folded conglomerates at the northernmost small outcrop. The interference is thought to result from D_4 overprinting of earlier D_3 minor folds. An attempt to date a felsic porphyry dyke cutting Missi suite rocks about 1 km west of this site on Annabel Lake, using conventional U/Pb techniques, yielded scattered zircon data which were attributed to an inheritance problem. Sphene gave an upper intercept age of 1830 +10/-8 Ma (Ashton *et al.*, 1992) which is now interpreted as reflecting an early thermal (regional metamorphic?) event in the Flin Flon volcanic belt.

(c) Proceed northward along the east side of the road to a large outcrop of garnet-biotite-muscovite quartzofeldspathic schist. This is Missi arenite in the core of the F_3 syncline. A close look will reveal abundant magnetite which is characteristic of amphibolite facies Missi rocks (the hematite seen in the low-grade rocks is metamorphosed to magnetite). Byers and Dahlstrom (1954) defined the ALSZ on the basis of these rocks thinking that their muscovite-rich nature indicated alteration associated with shearing. Back-rotated asymmetric pull-aparts related to the shearing event are partially developed in quartz veins cutting the outcrop.

(d) Crossing the road and proceeding north, another outcrop of Missi arenite is crossed before encountering the intensely folded basal polymictic conglomerate on the north limb of the F_3 syncline at the edge of the trees.

(e) Continue along the west side of the road to a cleared area across the highway from the shooting range access road. The exposed garnet-biotite gneiss is somewhat different than the Missi garnet-biotite arenites seen earlier. These rocks contain no magnetite and little or no muscovite. Rare sillimanite has been found in thin argillaceous layers and biotite and garnet are abundant. They are compositionally similar to wackes of the Burntwood Group in the Kisseynew Domain but can be traced continuously eastward to Embury Lake where they have been treated as part of the "Amisk" stratigraphic sequence. Similar wackes are found interlayered with Amisk tuffaceous rocks west of Amisk Lake where they are part of the West Amisk assemblage (termed the Welsh Lake Assemblage by Reilly, 1993). Preliminary detrital zircon studies of the turbidites have failed to uncover zircons younger than 1884 Ma (Fig. 3; Heaman *et al.*, 1993). Together with their apparent stratigraphic position, underlying the Missi suite at this locality, this suggests that the wackes are broadly coeval with the volcanic rocks.

(f) A short walk to the north towards the knobby-weathering outcrop crosses a contact with mafic rocks, at least some of which are dykes. Sinistral layer-parallel offsets can be observed locally. The knobby-weathering outcrop consists of a mafic mylonite containing alternating layers of hornblende amphibolite, carbonate and silicified(?) material representing the main zone of ALSZ displacement.

(g) [optional] Crossing the road near the "shooting range" sign provides access to several more interesting outcrops. Most of the slip in these rocks has occurred along the carbonate layers. The amphibolite has behaved in a more competent fashion by first becoming boudinaged into pinch-and-swell structures and then rotating during sinistral shear into back-rotated asymmetric pull-aparts. The pull-aparts defined by quartz veining noted in the Missi arenites, and the northeast-dipping replacement veins in the plagioclase porphyry are aligned perpendicular to this northeast-southwest shortening, further indicating a sinistral sense of shear. Continued deformation has resulted in northwest-plunging folds with northeast-dipping axial planes. This phase of folding is observed throughout this east-west zone between the Flin Flon and Kisseynew belts and marks the southwest-verging D_4 event. A short walk eastward along strike to the edge of the cleared area reveals the gabbroic precursor to the sheared rocks. The ALSZ likely developed at or slightly after the peak of metamorphism since diopside and sphene would likely have formed had this composition been subjected to the middle amphibolite facies peak metamorphic conditions inferred from the metasedimentary assemblages. It appears to have developed at amphibolite facies conditions since individual metamorphic hornblende grains rather than primary pyroxene are being deformed at the edges of discrete shears in the gabbro. Garnet porphyroblasts in the necks of amphibolite boudins further indicate that the early extension took place at amphibolite grade.

BRIDGING THE 'OCEAN' BETWEEN FLIN FLON AND SNOW LAKE

Between Cranberry Portage and Snow Lake, Highway 39 lies immediately south of the northern edge of Phanerozoic cover (Fig. 2) and consequently there are no exposures of Precambrian bedrock (however, flat-lying Ordovician Red River Formation dolomite is exposed in numerous road-cuts). Although the field trip has no stops in this covered portion of the Flin Flon belt, this area contains key tectonostratigraphic components and structural relations bearing on the assembly of the Flin Flon belt. Inasmuch as much of the exposed Shield just north of the highway is dominated by Elbow-Athapapuskow ocean-floor assemblage basalts, we are in effect crossing the oceanic realm between separate and possibly unrelated arc segments at Flin Flon and Snow Lake.

The area between Cranberry Portage and Snow Lake has been the focus of recent 1:20 000 mapping (Syme, 1990, 1991, 1992; Gilbert, 1992, 1993), 1:50 000 mapping (Morrison and Whalen, 1995; Syme *et al.*, 1995), reconnaissance-scale investigations (Syme, 1993, 1994), thematic studies (Connors, 1996; Williamson and Eckstrand, 1995) and theses (*e.g.*, Ryan and Williams, 1996; Morrison, submitted). This recent work has built upon a solid foundation of existing mapping (Bailes, 1980). The following sections summarize some of the relationships that have major implications for the structural evolution of the Flin Flon belt.

Elbow-Athapapuskow assemblage

The supracrustal component of the Flin Flon belt between Cranberry Portage (Athapapuskow Lake) and Reed Lake (Fig. 2) is dominated by ocean-floor (back-arc) basalts of the Elbow-Athapapuskow assemblage. It is important to note that the assemblage as a whole is neither lithologically nor geochemically homogeneous (Syme, 1988, 1992, 1995; Stern *et al.*, 1995b). It is composed of at least six distinct 'formations', 300 - >1100 m thick by up to >60 km in strike length. Each 'formation' is internally more or less homogeneous, displaying a consistent set of lithologic and geochemical characteristics throughout its extent. These consistent field and compositional characteristics allow confident correlation of far-flung basaltic sequences (*e.g.*, Syme, 1993, 1994). For mapping purposes, different basalt 'formations' can be readily distinguished by characteristic combinations of:

- flow morphology (*e.g.*, massive (Stop 11) *vs.* pillowed)
- weathering colour (*e.g.*, dark green *vs.* light buff)
- pillow characteristics (pillow size and shape, vesicularity, selvage thickness, presence of inter-pillow chert)
- alteration assemblage (*e.g.*, epidote-dominated *vs.* carbonate-dominated)
- phenocrysts

Contacts between 'formations' in the Elbow-Athapapuskow assemblage are invariably structural (Syme, 1995). At Elbow Lake (Fig. 2), where three ocean-floor 'formations' and the sole ocean-island basalt occurrence were first documented (Syme, 1990, 1991, 1992), each suite is structurally isolated from other suites.

The southwest-trending Cranberry shear zone (Syme, 1993, 1995) transects the Elbow-Athapapuskow assemblage between Elbow and Athapapuskow lakes (Fig. 2). For much of its length the shear zone juxtaposes two distinct 'formations', McDougalls Point basalt on the southeast side of the zone and Athapapuskow basalt on the northwest side of the zone. McDougalls Point basalt (Syme, 1991, 1992) is a buff-brown weathering pillowed N-type MOR basalt with slightly elevated "arc-like" Th/Nb ratios, similar to back-arc basin basalts from the Mariana Trough, North Fiji Basin or Lau Basin (Stern *et al.*, 1995b). Athapapuskow basalt (Stop 11) is a dark green weathering, massive, magnesian E-type MOR basalt, the least fractionated of all ocean-floor basalts, showing no evidence of either a continental crustal or arc signature (Stern *et al.*, 1995b). These thick (>1 km) extensive basalt formations likely tapped a variety of mantle sources, perhaps in a back-arc setting (Stern *et al.*, 1995b), and may have been juxtaposed early in the accretion of the Amisk collage by a "proto"-Cranberry shear zone.

Reed Lake

A fundamental problem in the eastern part of the Flin Flon belt has been the relation between the Amisk collage, which includes the Flin Flon arc assemblage and contained VMS deposits, and the arc-type metavolcanic rocks at Snow Lake (Fig. 2). Significant stratigraphic, geochemical and isotopic differences between the arc rocks at Flin Flon and those at Snow Lake (Stern *et al.*, 1995a) suggest that these segments may represent the remnants of unrelated arcs (Lucas *et al.*, 1996). A reconnaissance 1:50 000 study in the Reed Lake area (Syme *et al.*, 1995) provided structural evidence for a tectonic contact between the Amisk collage and Snow Lake assemblage.

A major (kilometres wide), regionally extensive tectonite belt exposed on western Reed Lake (West Reed-North Star shear zone; Fig 2) juxtaposes rocks of oceanic affinity on the west (Reed Lake mafic-ultramafic complex) with rocks of arc affinity on west-central Reed Lake (Fourmile Island assemblage; Fig. 2). It thus marks the eastern termination of the Elbow-Athapapuskow ocean-floor assemblage. Taken in this light, the West Reed-North Star shear zone may well be a bounding structure (at map scale) to the collapsed Elbow-Athapapuskow back-arc basin. The West Reed-North Star shear zone is interpreted as a relatively early structure that developed within the Amisk collage (Syme *et al.*, 1995), possibly coeval with early movement on the Elbow Lake shear zone-Centre Lake tectonite (*ca.* 1.87-1.86 Ga; Ryan and Williams, 1996; K. Ansdell, pers. comm., 1995).

The Fourmile Island assemblage includes at least 5.5 km of subaqueous mafic, intermediate and felsic flows and heterolithologic volcaniclastic rocks (Syme *et al.*, 1995) on western Reed Lake (Fig. 2). Preliminary geochemistry from this assemblage (Fig. 32) clearly indicates that these rocks are similar to Flin Flon and Snow Lake arc assemblage rocks, having low contents of HFSE (*e.g.*, Ti, Zr) compared to ocean-floor rocks in the Elbow-Athapapuskow assemblage. They range in composition from basaltic andesite to dacite (some of the 'dacites' may be partially silicified basaltic andesites), and are weakly tholeiitic in character.

The Northeast Reed assemblage is an areally extensive sequence of pillowed basalt on northern to eastern Reed Lake (Fig. 2). This mafic volcanic sequence, distinguished by its lithological homogeneity, is similar in field character to some of the basalt 'formations' in the Elbow-Athapapuskow ocean floor assemblage. The sequence is >4.3 km thick, comprising >90% pillowed flows, rare diabase dykes or sills, and no volcaniclastic rocks. Similar aphyric pillowed to massive basalt exposed between Morton, File and Woosey lakes is also provisionally assigned to the Northeast Reed assemblage. Collectively these basalts are tholeiites geochemically similar to basalts in the Elbow-Athapapuskow ocean-floor assemblage, marked by higher HFSE contents than rocks in the Fourmile Island assemblage (Fig. 32).

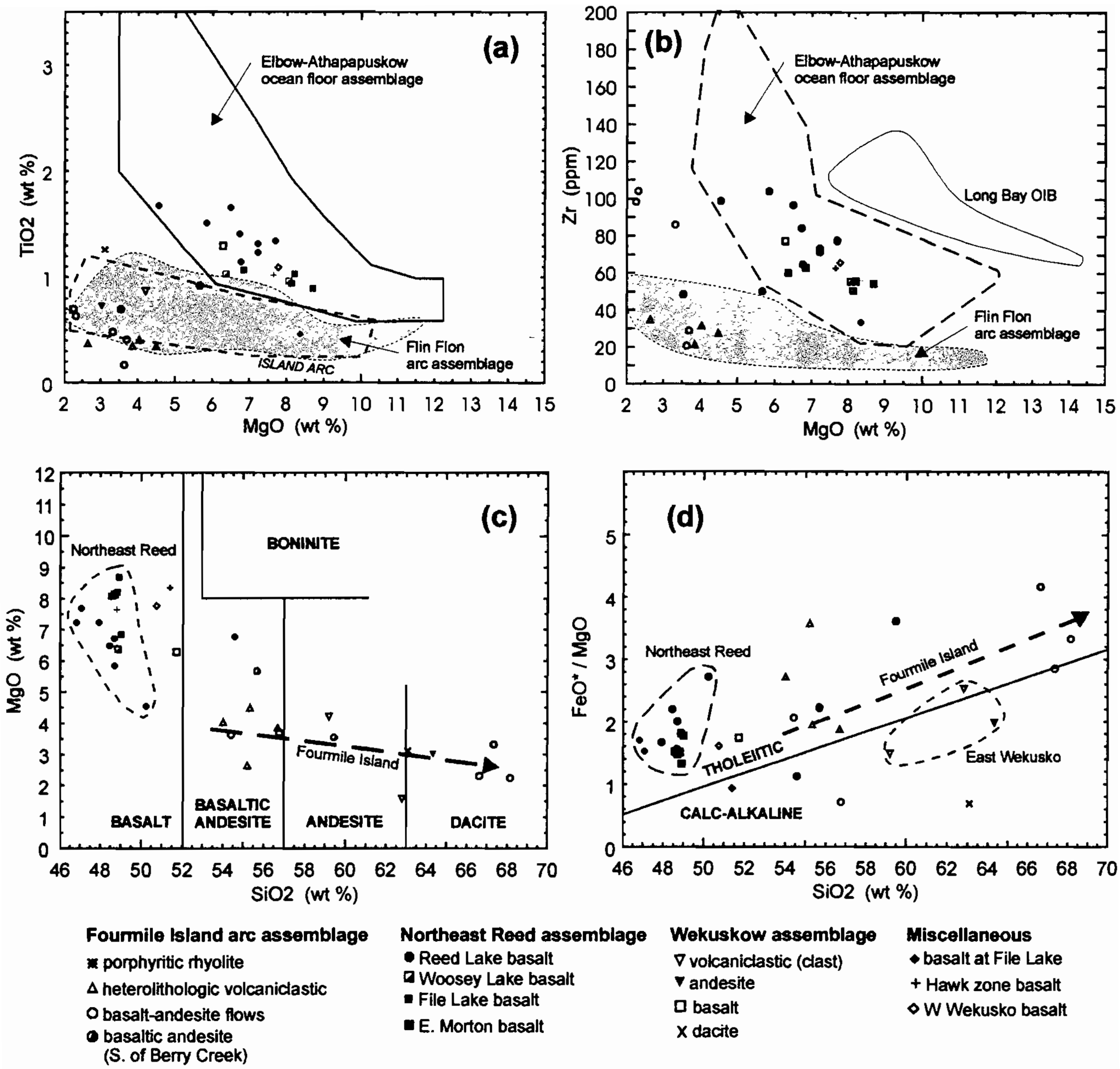


*Figure **32**: Selected geochemical plots for volcanic assemblages in the Reed Lake area. (a) MgO vs. TiO2: this discrimination diagram after Stern et al. (1995b) clearly separates Flin Flon belt arc rocks (shaded field) from Elbow-Athapapuskow ocean-floor basalts (solid line). Fourmile Island and Wekusko assemblages have the low TiO2 contents typical of arc rocks. The Northeast Reed basalts plot in the ocean-floor field, with higher TiO2 at any given MgO content than arc rocks; (b) MgO vs. Zr: ocean-floor basalts have higher Zr at any given MgO content than arc rocks (after Stern et al., 1995b); (c) SiO2 vs. MgO: the Fourmile Island and Wekusko arc assemblages display considerably more compositional diversity than the Northeast Reed ocean-floor assemblage; (d) SiO2 vs. FeO*/MgO: Northeast Reed ocean-floor basalts are tholeiitic; the trend for Fourmile Island assemblage has higher FeO*/MgO than calc-alkaline assemblages in the Amisk collage.*

The Reed Lake area is split into two domains by the Morton Lake thrust zone ('ML' on Fig. 2). Discovery of the Morton Lake thrust zone requires a revision of the location of the leading edge of the south flank of the Kisseynew Domain, as defined by the fault contact between allochthonous File Lake Formation turbidites and autochthonous Amisk collage rocks (Harrison, 1951; Connors, in prep.). The Morton Lake thrust juxtaposes a footwall (autochthonous) domain, comprising the Fourmile Island assemblage, West Reed-North Star shear zone and Reed Lake mafic-ultramafic complex, and a hanging wall (allochthonous) domain, consisting of the oceanic-affinity Northeast Reed assemblage, the composite Reed Lake pluton, and the Snow Lake arc assemblage. The fault zone itself includes the File Lake formation imbricate on Reed and Morton lakes. These relations are significant in that they suggest that the Snow Lake arc assemblage is contained in a southwest-verging allochthon emplaced after deposition (*ca.* 1.84 Ga) of the File Lake formation. This structural interpretation is consistent with earlier suggestions (based on lithologic, geochemical and isotopic criteria) that the Snow Lake and Flin Flon VMS camps are unrelated and represent remnants of different and distinct volcanic arcs that have been structurally juxtaposed.

STOPS IN THE SNOW LAKE AREA

Introduction

The relationship of volcanic rocks in the Snow Lake area to those already examined in the Flin Flon area (Fig. 2) is not only of academic interest but, because of the contained VMS deposits, is also of economic importance. In the past Snow Lake area volcanic rocks have simply been interpreted as the more highly metamorphosed and deformed equivalents of those at Flin Flon (Harrison, 1949; Bailes, 1971; Galley *et al.*, 1990). However, recent geochemical (Stern *et al.*, 1995a), isotopic (Stern *et al.*, 1992), U-Pb zircon age dating (Ansdell and Connors, 1994; David *et al.*, 1996) structural (Krause and Williams, 1994a, 1994b, 1995; Connors and Ansdell, 1994, 1996; Connors, 1996) and tectonostratigraphic (Lucas *et al.*, 1996; Syme *et al.*, 1995) studies suggest a more complicated relationship between these two assemblages. These studies indicate that although arc volcanic rocks at Snow Lake are comparable in age (*ca.* 1.89 Ga; David *et al.*, 1996) to those at Flin Flon (1.904-1.885 Ga), they are not simply their more recrystallized equivalents because they:

- consistently exhibit lower ε_{Nd} (<+3.1) than arc volcanic rocks at Flin Flon (Fig. 33; Stern *et al.*, 1992, 1995a),
- locally contain inherited Archean zircon xenocrysts whereas those at Flin Flon do not (David *et al.*, 1996),
- are structurally imbricated with *ca.* 1.85-1.83 Ga sedimentary rocks that are not present in Flin Flon equivalents (David *et al.*, 1996; Krause and Williams, 1994a; Syme *et al.*, 1995),
- are intruded by a prominent *ca.* 1.84-1.83 Ga suite of calc alkaline granitic magmas and are not intruded by the *ca.* 1.87-1.845 Ga suite of "successor arc" granitic plutons that charcterise the Flin Flon arc assemblage and the Amisk collage (Lucas *et al.*, 1996), and
- exhibit metamorphic and structural features that are more akin to rocks of the Kisseynew domain than those of the Amisk collage of the Flin Flon belt (Syme *et al.*, 1995).

The current working hypothesis is that the Snow Lake arc assemblage developed separately from that at Flin Flon, possibly upon an Archean microcontinental fragment (Stern *et al.*, 1995a; Lucas *et al.*, 1996), and was tectonically juxtaposed against accreted rocks of the Amisk collage during *ca.* 1.84 -1.81 Ga southwest-directed collision of rocks of the Kisseynew Domain with those of the Flin Flon belt (Syme *et al.*, 1995). During this collision event, tectonic slices of *ca.* 1.85 - 1.83 Ga, post-accretion sedimentary rocks (mainly Burntwood suite) were structurally imbricated with allocthons of pre-1.88 Ga volcanic rocks. The most westerly slice of the post-accretion sedimentary rocks, occupying the Morton Lake fault zone (Figs. 2, 34), is currently interpreted to be the structural contact between the Amisk collage accretionary complex and the various Kisseynew-type allochtons that characterize the Snow Lake area to the east (Syme *et al.*, 1995; Lucas *et al.*, 1996).

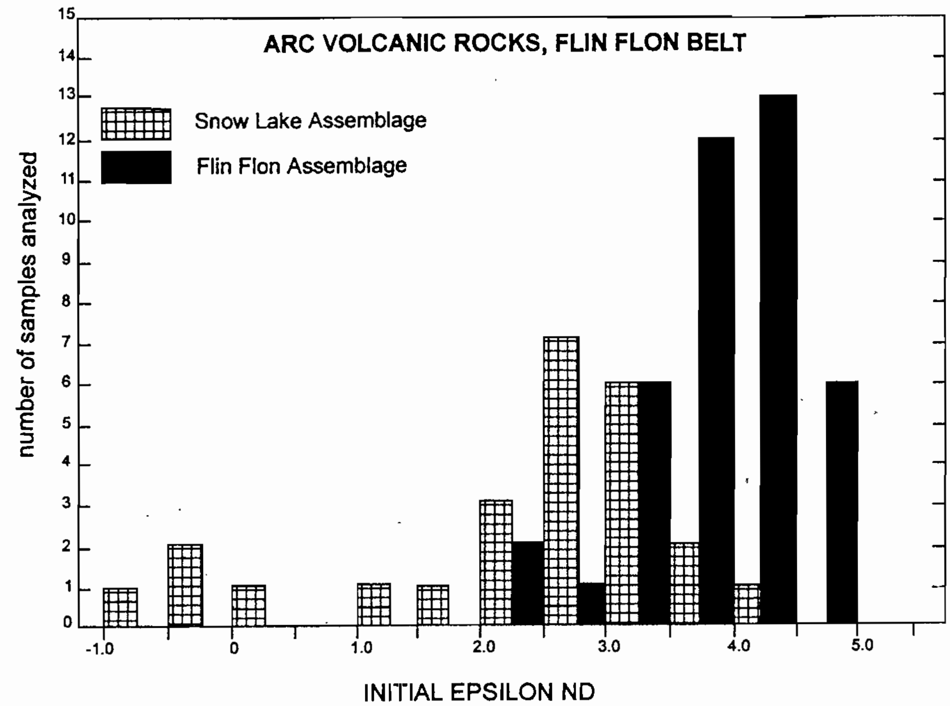


Figure33: Histogram comparing the initial ε_{Nd} values of analyzed rocks from the Flin Flon and Snow Lake arc assemblages (from Stern et al., 1992). The lower overall ε_{Nd} values for Snow Lake arc assemblage rocks is interpreted by Stern et al. (1992, 1995a) to be due to intracrustal contamination by older (Archean?) material. This is one indication that the Snow Lake arc assemblage may have developed separately from that at Flin Flon.

Whereas deformational features examined during the Flin Flon portion of the field trip were largely related to 1.88-1.87 Ga intraoceanic accretion of the Amisk collage, and subsequent deformation coeval with post-accretion arc magmatism (1.87-1.84 Ga), those to be observed at Snow Lake formed during 1.84-1.78 Ga collisions. Four post-1.84 Ga deformational events are recognized in the Snow Lake area (Kraus and Williams, 1995). The first event produced tight to isoclinal, *ca.* 1.84 Ga, recumbant F_1 folds (Bailes, 1980a; David *et al.*, 1996) and related thrust faults (Connors and Ansdell, 1994; Connors, 1996). The second event produced isoclinal, *ca.* 1.82-1.81 Ga, syn- to post-metamorphic F_2 folds and thrust faults (Krause and Williams, 1994a, 1995; Kraus and Menard, 1995). F_3 produced upright, open northeast trending post-metamorphic folds (*e.g.* Threehouse synform; Fig. **36**) and F4 includes east-trending open folds (Kraus and Williams, 1994a). A series of brittle-ductile shear zones (*e.g.* Berry Creek fault) deform and offset earlier structures. Syme *et al.* (1995) and Kraus (pers. com., 1996) suggest that the Berry Creek fault may have been a long-lived structure that was successively reactivated during post-collisional deformation, regional cooling and exhumation (1.8-1.7 Ga; Fedorowich *et al.*, 1995). U-Pb zircon and titanite ages that constrain major episodes of magmatism, sedimentation, plutonism, deformation and metamorphism in the Snow Lake area are summarized in Figure 35.

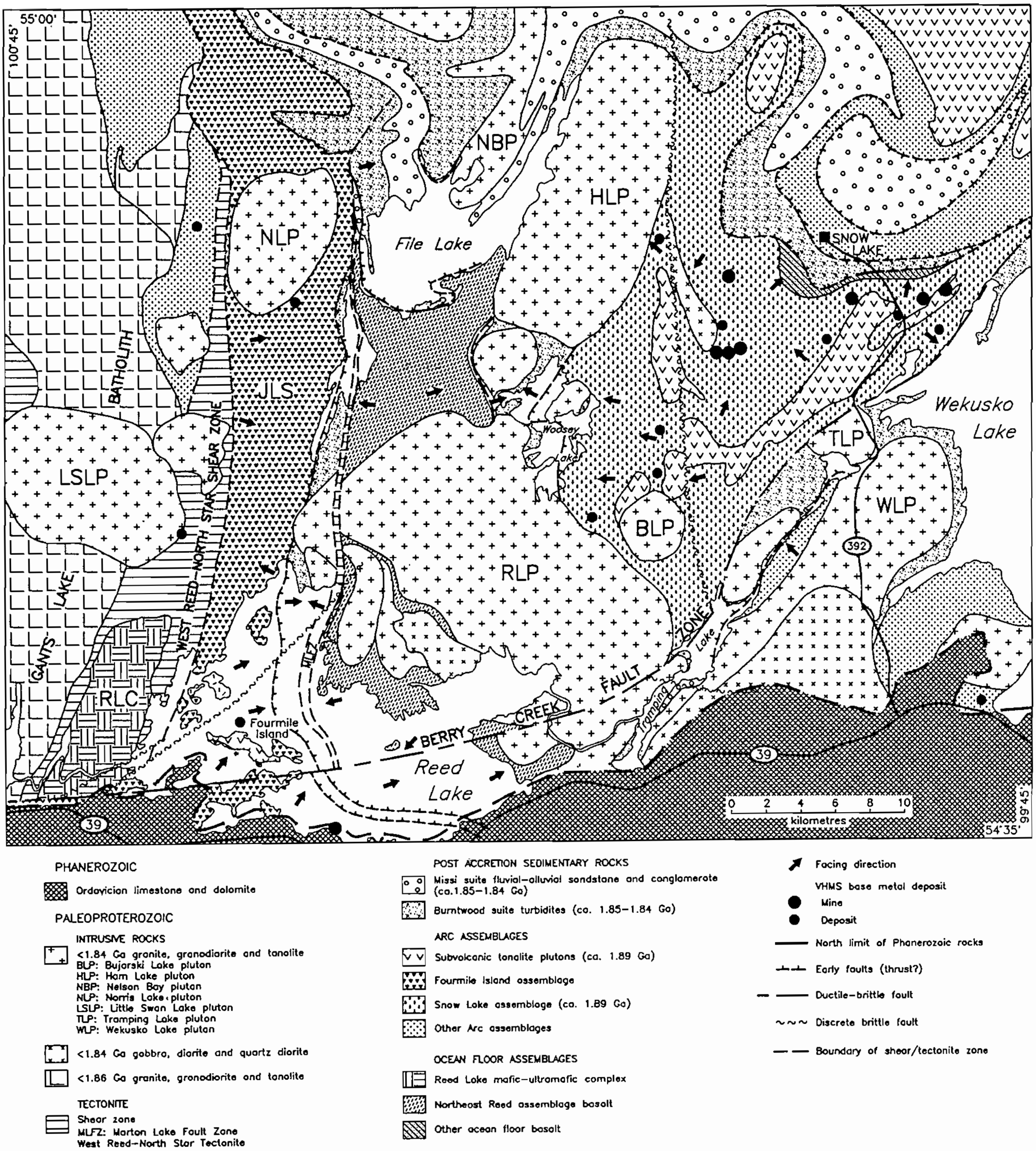


Figure 34: Generalized geology of the Reed Lake-Wekusko Lake area modified from Syme et al. (1995), Morrison and Whalen (1995), Bailes et al. (1994), Froese and Moore (1980), Rousell (1970) and Stanton (1945). The Morton Lake fault zone (MLFZ) is interpreted to be the structural contact between the Amisk collage, examined during the Flin Flon portion of the field trip, and the Snow Lake area. The Snow Lake area is characterized by a structural style and by lithologies that are more comparable to the Kisseynew Domain than those observed in the Amisk collage. The Snow Lake area consists of a series of Kisseynew-type allocthons of volcanic and sedimentary rocks.

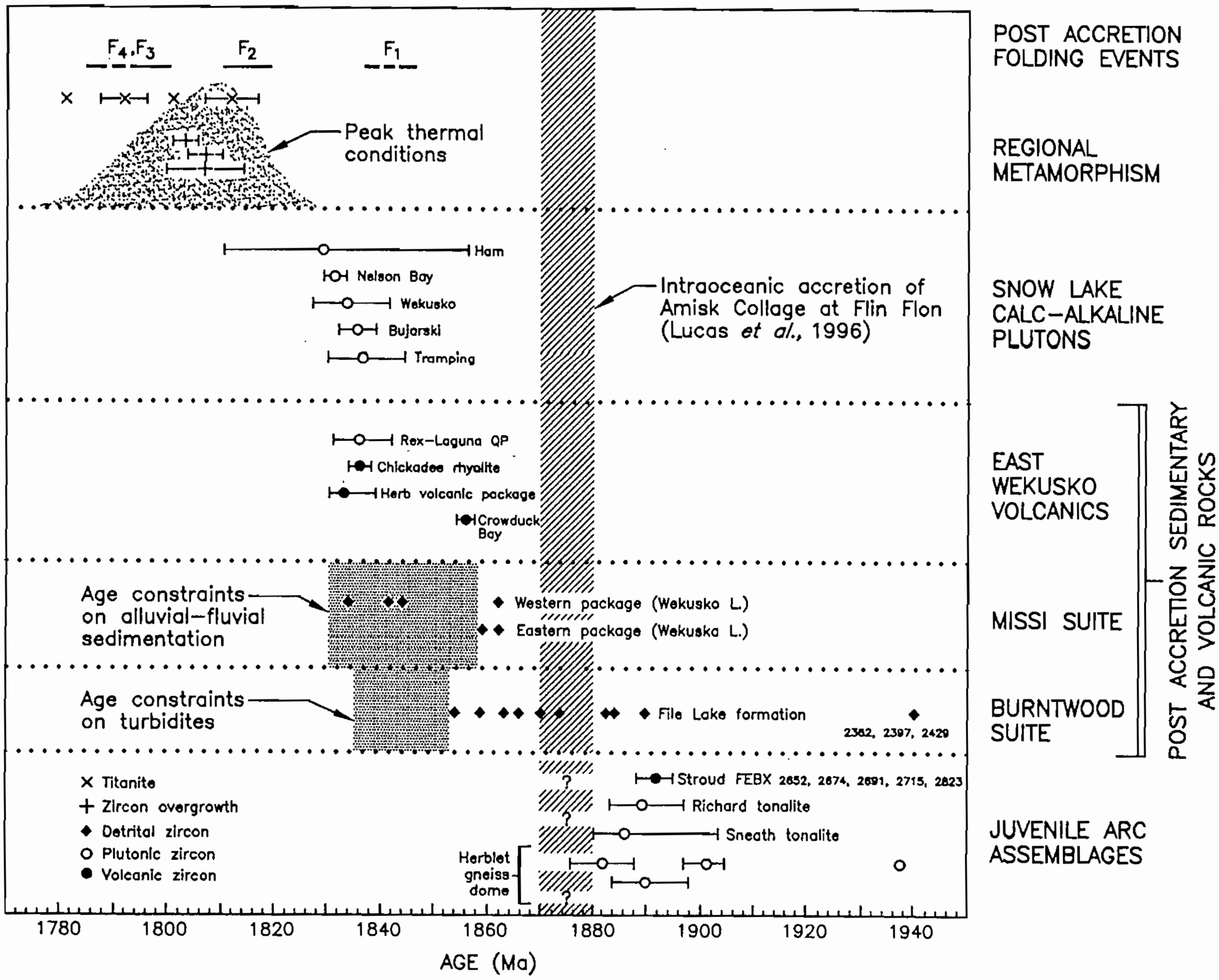


Figure 35: Summary of U-Pb zircon and titanite ages that constrain major episodes of magmatism, sedimentation, plutonism, deformation and metamorphism that effected Snow Lake rocks (from David et al., 1996).

Post-Accretion sedimentary and volcanic rocks

Introduction

Ca. 1.85-1.83 Ga sedimentary and volcanic rocks that post date the 1.88-1.87 Ga intra-oceanic accretion events of the Amisk collage outcrop widely in the Snow Lake area, and can be traced directly to the north into highly metamorphosed and recrystallized paragneisses and orthogneisses typical of the Kisseynew Domain (Bailes and McRitchie, 1978). The origin of and deformation of these post accretionary rocks (Fig. 35) is of particular interest as it bears on the fundamental tectonic relationship between the largely volcanic rocks of the Flin Flon Belt and the mainly sedimentary rocks of the Kisseynew Domain.

Stops 14 and 15 in greywacke turbidites of the Burntwood suite are included to show the primary sedimentary characteristics of this post-accretion sedimentary unit, to show structural and metamorphic features recorded in these well-layered rocks, and to demonstrate features that are critical to unravelling tectonstratigraphic events at Snow Lake.

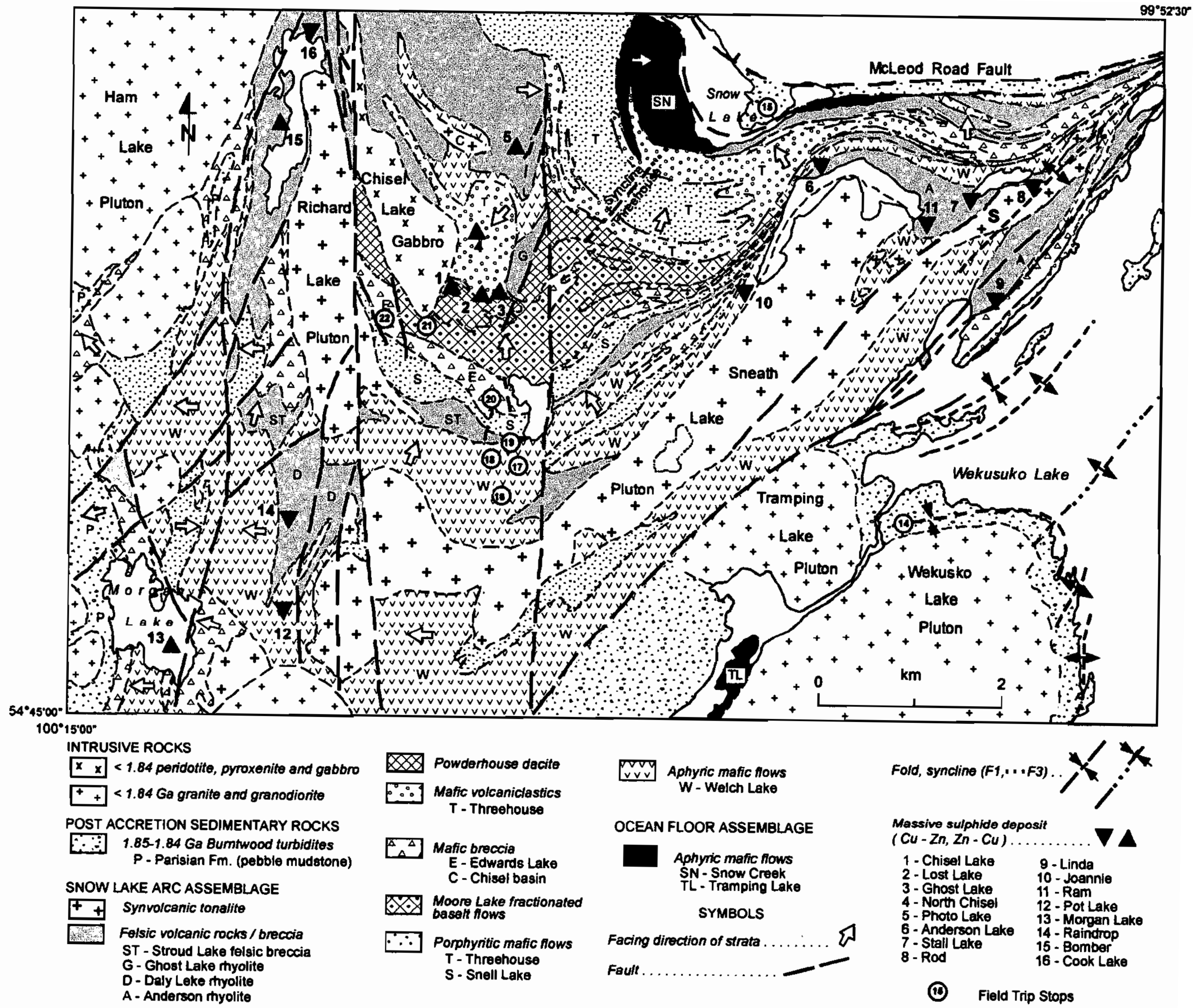


Figure 36: Simplified geology of the Snow Lake area. Circled numbers identify location of field trip stops. Bold triangles mark location of massive sulphide deposits. A schematic geological cross section of the Snow Lake arc assemblage, from south southwest to north northeast across this figure, is shown in Figure along with the location of Stops 15 - 22. A more detailed geolgoical map, showing the location of Stops 16 - 22, is given in Figure .

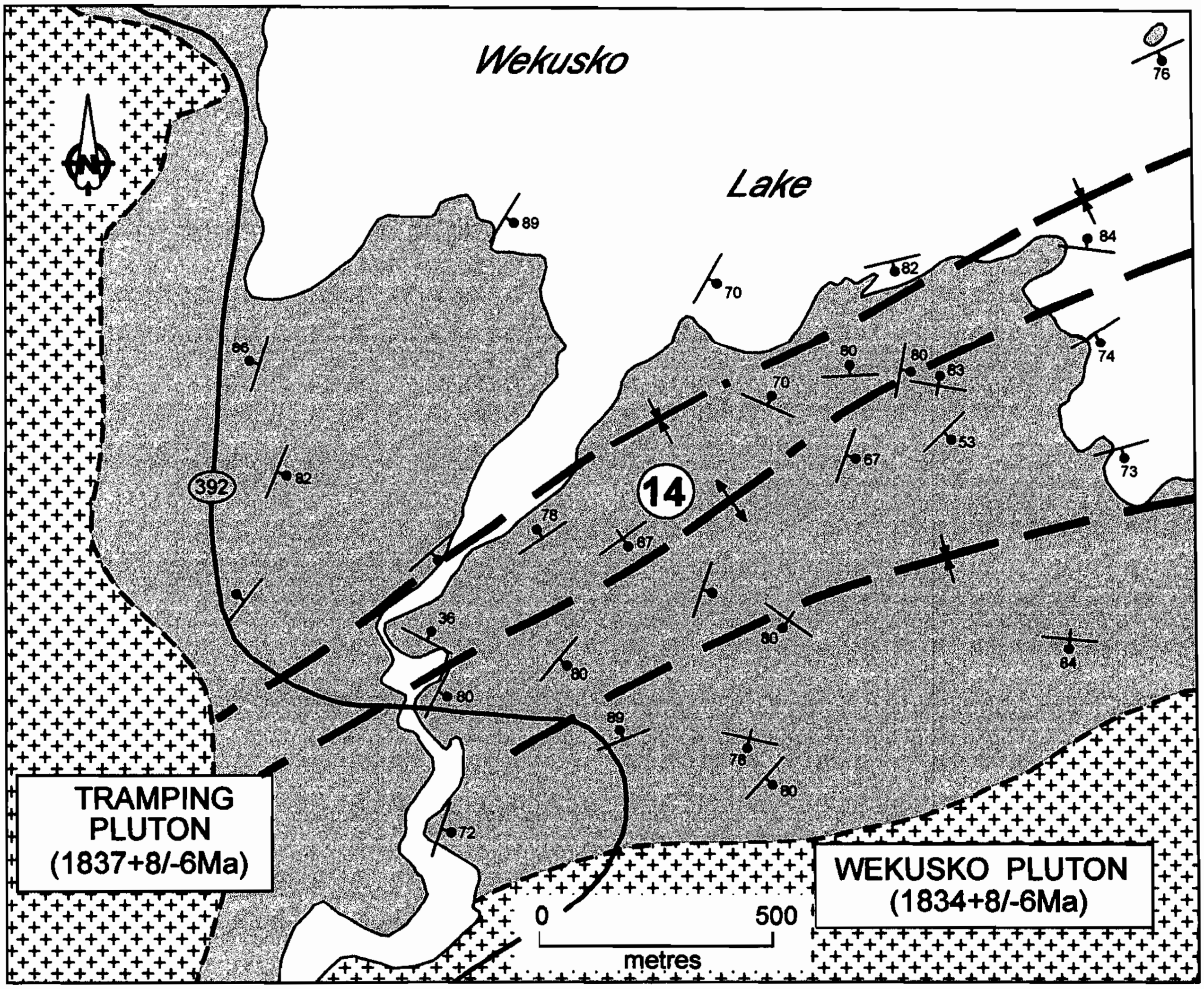


Figure 37: Location of Stop 14 on the north limb of an ENE trending isoclinal F_1 antiform in Burntwood suite turbidites. The F_1 folds are truncated to the west by the Tramping Lake granodiorite pluton and cut by dykes of Wekusko Lake granodiorite. Stop 14 is located on a firebreak trail 300m east of the Wekusko Falls campground storage compound.

STOP 14: Post-Accretion greywacke turbidites, Wekusko Lake

Burntwood suite pelitic gneisses with upper almandine-amphibolite facies mineral assemblages are the major component of the Kisseynew Domain (Bailes and McRitchie, 1978; Bailes, 1980a). At this stop (Figs. 36, 37), adjacent to Wekusko Lake, a relatively unrecrystallized sequence (greenschist facies mineral assemblages) of Burntwood suite greywacke, siltstone, and mudstone is exposed. They display well defined, 5 to 100 cm thick beds, composed of fine to medium sand-sized detritus, and exhibit bed forms and sedimentary structures consistent with deposition by turbidity currents. Sedimentary features to be observed at this stop include normal size grading, scour channels, load structures, rip ups, calcareous concretions, and zoned Bouma A, AB and ABE beds. The outcrop is located on a fire break 500m east of the Wekusko Falls campground park storage compound.

Beds at this Stop strike easterly and dip vertically. They display two cleavages. One cleavage is bedding-parallel, and is interpreted to be related to east-trending, steep-plunging, map-scale F_1 isoclinal folds (Fig. 37). The other cleavage is at a high angle to bedding (strike of 020° with a steep dip) and is interpreted to have developed during open F_3 folding. The F_1 folds are trucated (see Fig. 37) by the Tramping Lake pluton (1837 +8/-6 Ma; David *et al.*, 1996) and are cut by dykes of the Wekusko Lake pluton (1834 +8/-6 Ma; Gordon *et al.*, 1990, Bailes, 1992).

The Burntwood suite was deposited between 1835 Ma (cross cutting plutons) and 1842 Ma (age of the youngest detrital zircon, Machado and Zwanzig, 1995), approximately the same age as the ca. 1845 Ma Missi suite fluvial-alluvial arkoses and conglomerates (Ansdell, 1993) examined earlier at Stop 12. Southward coarsening of Burntwood suite turbidites (Bailes, 1980b; Syme *et al.*, 1995) and northward fining of conglomeratic Missi suite rocks (Harrison, 1951; Bailes, 1971; Bailes and Syme, 1989; Ansdell *et al.*, 1995) suggests that the two suites were, at least locally, sedimentary facies equivalents. This interpretation is supported by the presence of local turbidite bedforms in ca 1.84 Ga Missi suite arkosic sediments on Tramping Lake and well rounded, 'Missi-type' clast populations in pebble- and cobble-rich Burntwood suite turbidites on Reed Lake (Fig. 34; Syme *et al.*, 1995).

U-Pb dating of Burntwood suite detrital zircons from Wekusko and File lakes by David *et al.* (1996) indicate three main age populations: 1874-1859 Ma, 1890-1885 Ma and 2429-2362 Ma. The 1874-1859 Ma and the 1890-1885 Ma detrital zircon age populations are consistent with derivation of detritus by unroofing of accreted intraoceanic volcanic and intrusive rocks of the Amisk collage, as well as subsequent successor arc calc-alkaline plutons and extrusive equivalents. The 2362-2429 Ma detrital zircons in the Wekusko Lake Burntwood suite suggest that a cratonic terrane, younger than the bounding *ca.* 2.7-2.8 Ga Archean Superior craton, was present in the source area of the turbidites. There is no evidence to date of input of detritus from the bounding *ca.* 2.7-2.8 Ga Superior craton, suggesting that juxtaposition of the Flin Flon and Kisseynew domains with the Superior craton occurred later than 1.845-1.835 Ga.

Ansdell *et al.* (1995) have suggested a back arc tectonic setting for the Burntwood suite and the Kisseynew Domain. They suggest that the post-accretionary sediments likely reflect significant uplift across the Trans-Hudson orogen at ca. 1.86-1.85 Ga, possibly from the arc-arc and arc-continent collisions (e.g. Lewry, 1981; Bickford *et al.*, 1990; Lucas *et al.*, 1996).

STOP 15: Post-accretion meta-turbidites, Snow Lake

Structural and tectonic analysis (Kraus and Williams, 1994a, 1994b, 1995; Connors and Ansdell, 1994; Connors, 1996) and recent mapping (Bailes *et al.*, 1994; Syme *et al.*, 1995) indicate that the present distribution of Burntwood suite rocks in the Snow Lake area reflects thrusting that occurred during southwest-directed convergence between the Kisseynew Domain and Flin Fon Belt. During this deformation, the Burntwood suite (and other Kisseynew Domain lithologies) were thrust onto and imbricated with Flin Flon Belt volcanic rocks. Thus, slices of Burntwood suite turbidites commonly mark the location of tectonic boundaries between various allocthons of Flin Flon Belt stratigraphy. Stop 15 is located in one of these tectonic slices of Burntwood suite turbidites (Fig. 34, 38). This slice of Burntwood suite rocks is bounded to the south by the Snow Lake fault and to the north by the McLeod Road fault; both these faults are semiconformable to stratigraphy and are interpreted to be possible thrust structures (Kraus and Williams, 1994a, 1995).

Stop 15 consists of isoclinally folded Burntwood suite greywacke, siltstone and mudstone turbidites. The turbidites contain almandine-amphibolite facies metamorphic mineral assemblages, including prominent porphyroblasts of staurolite (up to 8 cm), garnet (up to 5 mm), and biotite (up to 5 mm). Despite recrystallization these meta-turbidites preserve primary sedimentary features such as normal size grading and local mudstone rip-ups. Metamorphic porphyroblasts at this outcrop are larger in finer grained, clay-rich portions of beds and, thus, display inverse metamorphic size gradation relative to primary sedimentary size grading.

At this stop the effects of successive episodes of deformation in the Snow Lake area are recorded in the well-bedded meta-turbidites by minor folds, cleavage relationships, and fabrics formed during porphyroblast growth and deformation. As part of a PhD study at the University of New Brunswick, Jurgen Kraus studied this and other outcrops and has established the presence of four phases of folding (F_1-F_4); three of them (F_1-F_3) are reflected in structures and fabrics at this stop. The description of this stop and the relationships between fabrics is based on material supplied by Jurgen.

Stop 15 is located near the trace of a mappable isoclinal F_1 syncline (Fig. 38). Minor F_1 folds are only rarely exposed, but are present here as moderately open to isoclinal, metre-scale structures. Although these particular folds do not display an axial planar S_1 cleavage, a prominent bedding-parallel cleavage (interpreted to be an S_1 fabric) is common elsewhere. Minor F_1 fold axes and a related mineral lineations plunge moderately to steeply in northeasterly to easterly directions. F_1 folds on this outcrop are overprinted at a moderate to high angle by an S_2 differentiated crenulation cleavage, which dips steeply in northeasterly directions. Differentiation of the S_2 crenulation cleavage into muscovite septae and quartz-rich microlithons has formed a penetrative schistosity or a domainal cleavage that presently anastomises around porphyroblasts. The presence of S_2-crenulated S_1 inclusion trails in porphyroblasts, suggests that porphyroblast growth was coeval with early stages of isoclinal F_2 folding (Kraus and Williams, 1994b). Continued F_2-deformation is suggested to be responsible for curvature of S_2 muscovite septae around porphyroblasts, as well as prominent pressure shadows and quartz-filled pull-aparts displayed by large staurolite porphyroblasts. The S_2 fabric is refracted across the relatively incompetent pelitic layers, where it typically displays a smaller S_0-S_2 dihedral angle than it does in more competent greywacke beds (Fig. 39). Open crenulations of the S_2 fabric, occuring selectively in pelitic layers, are interpreted by Kraus and Williams (1994b) to be related to large-scale F_3 folds (*e.g.* Threehouse syncline). At the north end of the outcrop, a small, Z-assymetrical, F_3 minor fold overprints a minor F_1 fold and the S_2 cleavage. Bedding-parallel, quartz-filled faults locally offset F_1 fold limbs and place beds with opposite facing directions together.

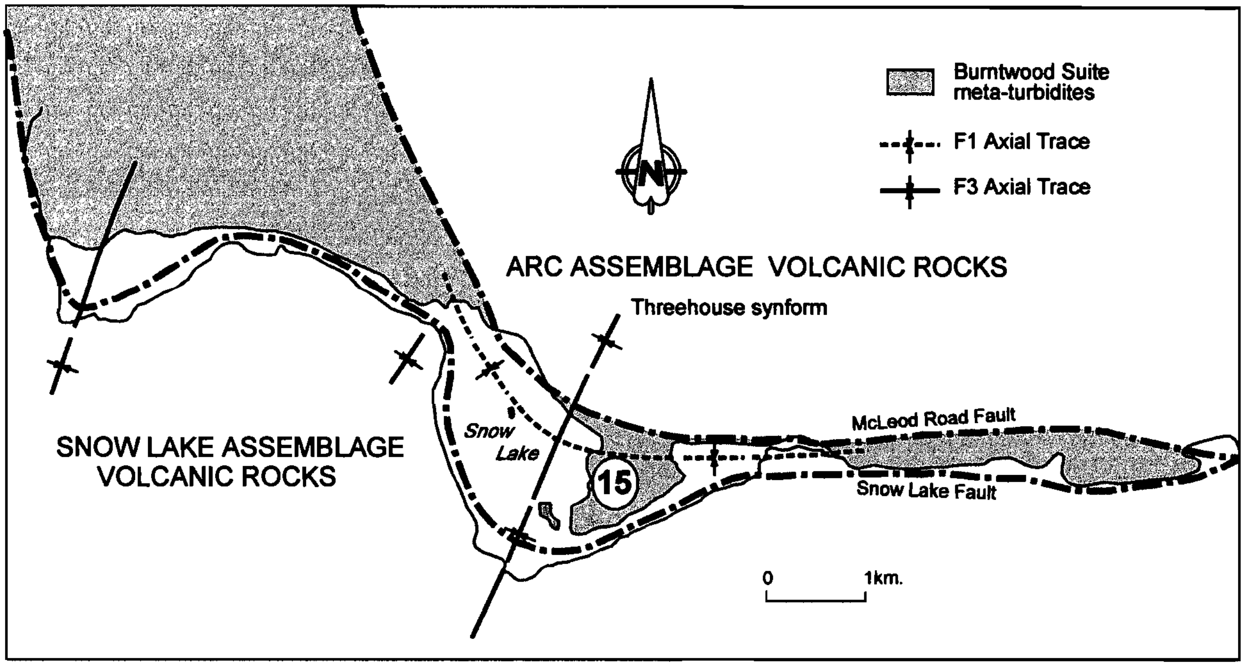


Figure 38: Location of the Stop 15 in a tectonic slice of Burntwood suite metaturbidites bounded to the south by the Snow Lake fault and to the north by the McLeod Roac fault. Stop 15 is located near the trace of an F_1 synform and on the east limb of the F_3 Threehouse synform.

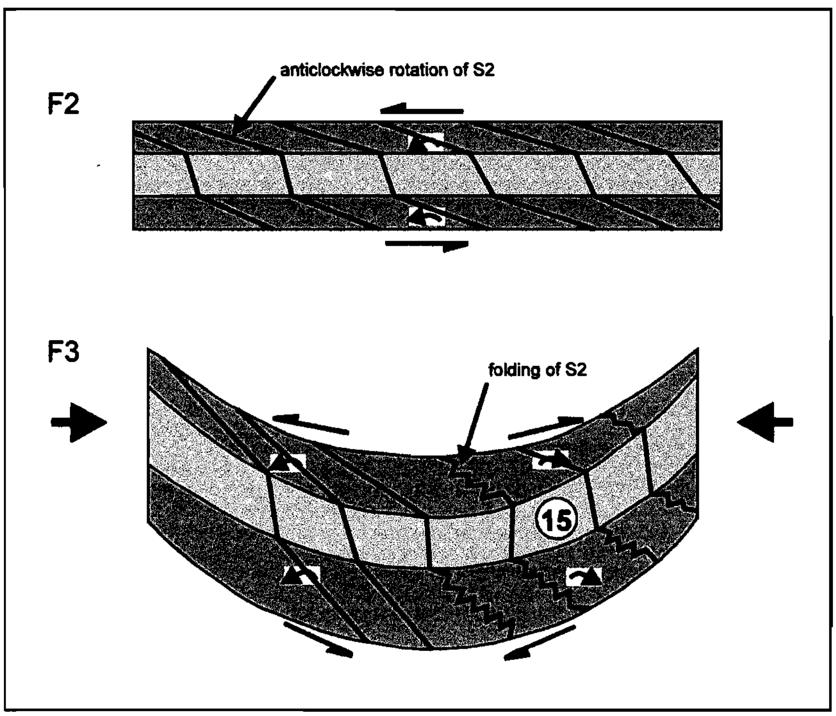


Figure 39: Two layer model illustrating the modification of S_2 during F_2 and F_3 from Kraus and Wlliams (1994b). Note the partitioning of deformation in competent (light grey) and incompetent (dark grey) layers.

These faults are early as they are crenulated by S_2. Refraction of the S_2 differentiated cleavage in incompetent layers (pelitic beds), is suggested to reflect bedding-parallel shear (Fig. 39; Kraus and William, 1994b).

Garnet-biotite thermometry using the calibration of Kleemann and Reinhardt (1994) indicates a temperature of 536 ± 11^0 C for a population of 24 garnet-biotite pairs from a sample of the Burntwood suite at Snow Lake (Kraus and Menard, 1995). A pressure of approximately 4 kb for the same sample was determined by Kraus and Menard (1995) using the garnet-biotite-muscovite-plagioclase barometer of Ghent and Stout (1981). Sparse geothermal and geobarometric data from metamorphic minerals in altered volcanic rocks south of the Snow Lake fault (Zaleski *et al.*, 1991, Menard and Gordon, 1995) suggest fairly constant peak temperatures of approximately 550^o C in conjunction with an increase in pressure for 10 km south across strike of tectonostratigraphic units. Krause and Menard (1995) suggest that this is consistent with a uniform metamorphic regime for the volcanic rocks to the south of the fault and for the Burntwood suite slice at Snow Lake. This apparent absence of a metamorphic break across the Snow Lake fault, suggests that it is an F_1 structure, in contrast to the post-peak metamorphic F_2 McLeod Road fault (Krause and Menard, 1995). Peak metamorphic conditions were reached in the Snow Lake area at *ca.* 1.81 Ga (Fig. 35) according to a U-Pb age dating of metamorphic zircon and titanite by David *et al.* (1996).

Snow Lake arc assemblage

Introduction

Isotopically juvenile *ca.* 1.89 Ga arc volcanic rocks at Snow Lake (Fig 2) comprise one of several allocthons in a thrust stack (Syme *et al.*, 1995) formed during *ca.* 1.84-1.80 Ga southwest-directed collision between the Kisseynew Domain and the Flin Flon Belt (Fig. 102). The > 6 km thick juvenile oceanic arc sequence at Snow Lake records in its stratigraphy and geochemistry (Bailes and Galley, 1996) a temporal evolution from a relatively more primitive to an evolved arc (Fig. 40). Similar to the Flin Flon arc assemblage, the Snow Lake arc assemblage is an important host for VMS base metal mineral deposits (Figs. 34, 36). VMS deposits at Snow Lake can be subdivided into Cu-rich, Zn-rich and Cu-Zn-Au types. Cu-rich deposits, mainly at Anderson and Stall lakes (Figs. 36, 40), occur in a flow-dominated, bimodal (basalt-rhyolite) sequence characterised by primitive arc tholeiite. Zn-rich types (*e.g.* Chisel Lake) occur in a volcaniclastic-dominated, relatively more evolved sequence (Fig. 36, 40). A recently discovered Cu-Zn-Au rich deposit at Photo Lake (Bailes and Simm, 1994) also occurs in the more evolved arc sequence but within a rhyolite-dominated section (Fig. 36). One of the universal characteristics of arc volcanic sequences containing associated VMS deposits is the widespread occurrence of hydrothermally altered rocks. In the Snow Lake area these altered rocks (Fig. 41) are readily recognized by their unique mineral assemblages that formed during subsequent (1.81 Ga) regional metamorphic recrystallization.

The compositionally bimodal primitive arc sequence includes Welch basalt/basaltic andesite, various rhyolite complexes, and the voluminous subvolcanic

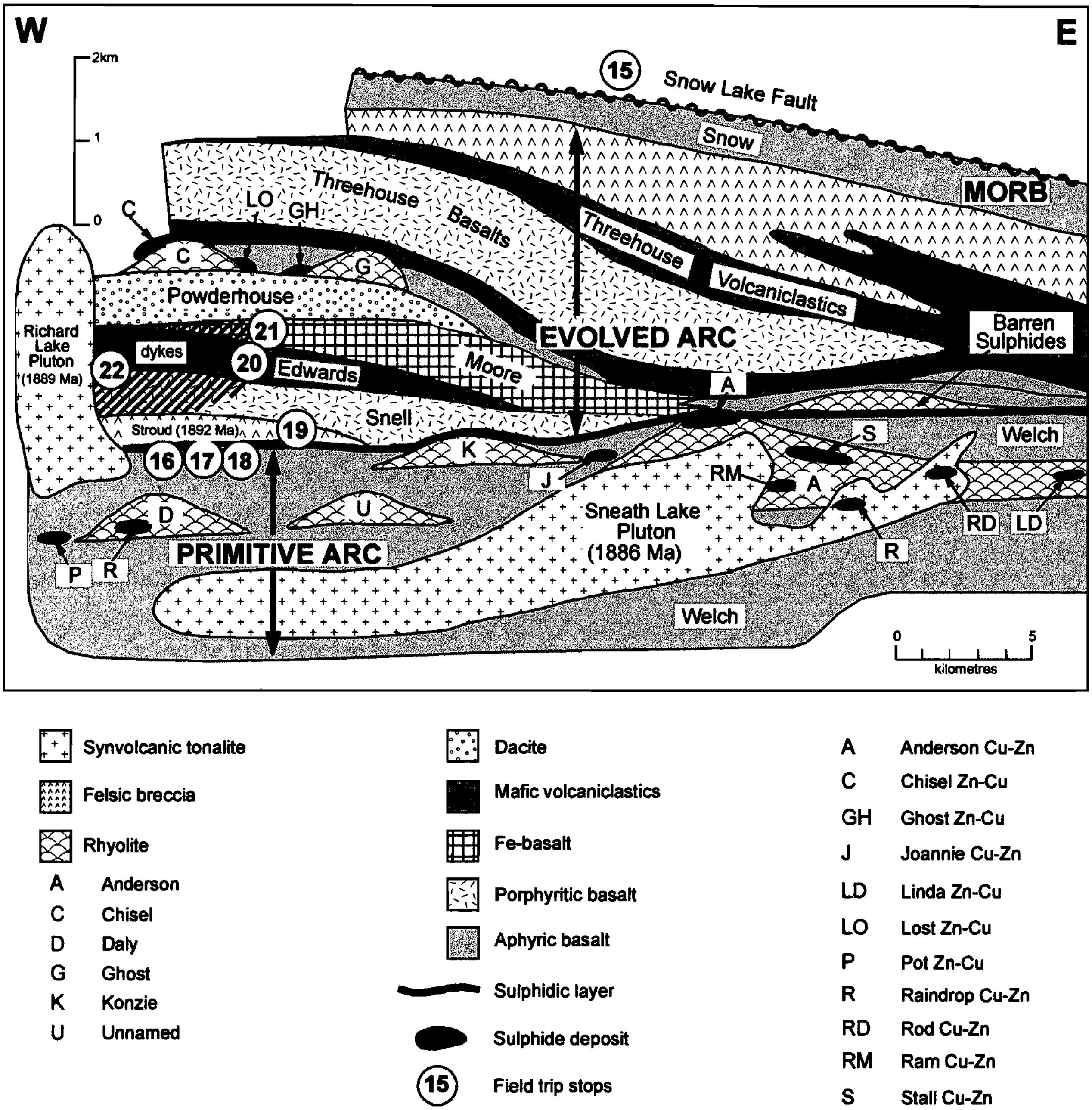


Figure 40: Schematic geological cross section of Snow Lake arc assemblage with location of field trip stops. The Snow Lake arc assemblage is subdivided into a Primitive and Evolved arc. The Primitive arc is a bimodal basalt-rhyolite sequence that includes the subvolcanic Sneath Lake tonalite intrusive complex. The Evolved arc is a more heterogeneous sequence, with up to 50% volcaniclastic detritus, that includes a prominent subvolcanic dyke complex and the Richard Lake tonalite pluton. The Evolved arc section is capped by basalts with MORB chemistry. Cu-rich VMS deposits occur associated with rhyolite flows in the Primitive arc and Zn-rich deposits with rhyolite flows in the Evolved arc.

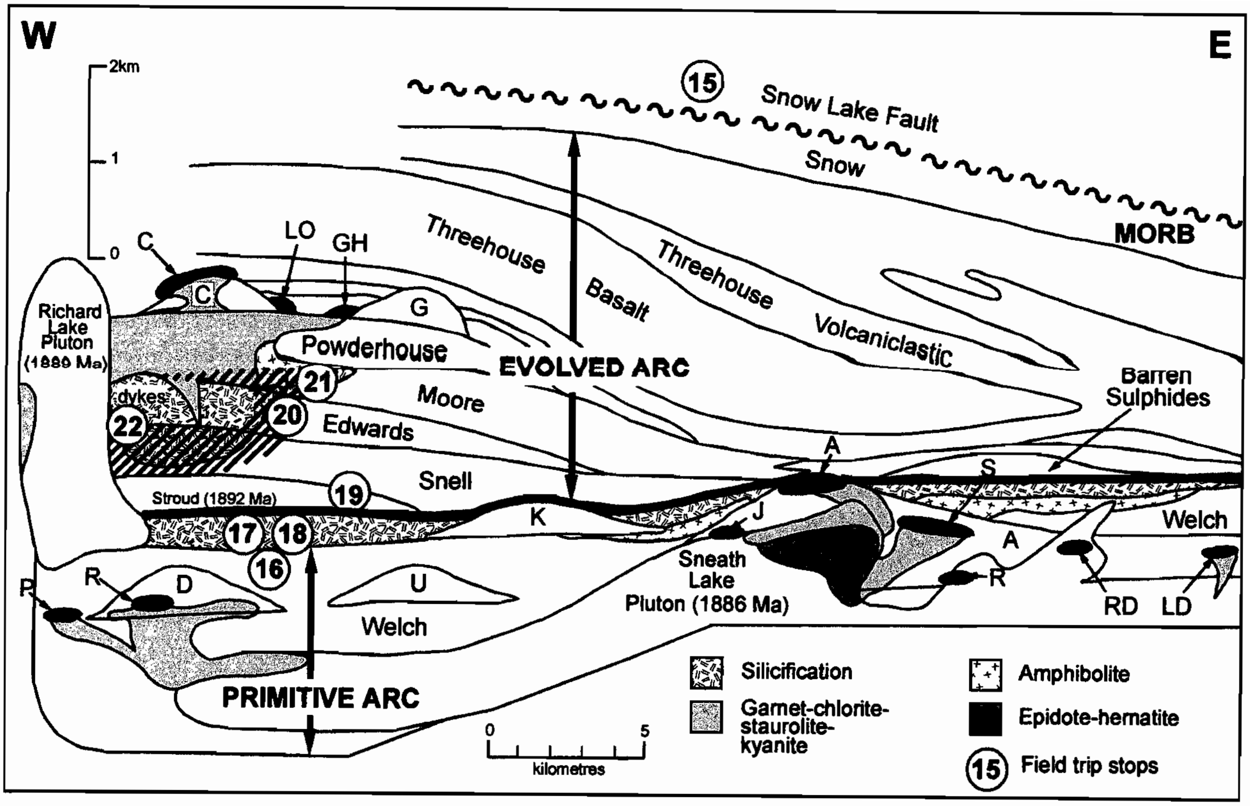


Figure 41: Schematic cross section of the Snow Lake arc assemblage showing the distribution of hydrothermally altered rocks. Field trip stops are shown by circled numbers.

Sneath Lake tonalite pluton (Fig. 40). The primitive arc rocks host three past-producing mines (Anderson Lake, Stall Lake and Rod) that contained 10.1 million tonnes of ore with an average grade of 4.15% Cu and 0.53% Zn (Bailes and Galley, 1996). The primitive arc also hosts the uneconomic 13 million tonne Linda deposit grading 0.3% Cu and 0.79% Zn (Minova, pers comm, 1992), the Raindrop zone grading 3.52% Cu and 0.82 % Zn over 8.36 m (Hodges and Manojlovic, 1993), the Pot deposit grading 1.43% Cu and 4.5% Zn over 2.1 m (HBED, pers com., 1972) and the small uneconomic, stockwork-type Joannie and Ram zones (Figs. 36, 40).

The > 3 km of Welch formation consists of subaqueous low-Ti tholeiitic basalt to andesite flows, and a subunit of high-Ca boninite (Stern *et al.,* 1995a; see Stop 16). High-Ca boninites are relatively rare in modern settings, having only been identified within the northern end of the Tongan forearc (Falloon *et al.*, 1989) and in the upper Pillow Lavas of the Troodos ophiolite (Duncan and Green, 1987). On the basis of the high-Ca boninite and low-Ti basalt association, Stern *et al.* (1995a) suggest that the primitive arc sequence and contained Cu-rich VMS deposits formed in a forearc tectonic setting.

Although volumetrically minor, subaqueous rhyolite flow complexes are extremely important because Cu-rich VMS deposits in the primitive arc section are rhyolite-hosted (Fig. 40). The deposits are spatially associated with regionally extensive zones of hydrothermal alteration that stratigraphically overlie the subvolcanic Sneath Lake tonalite pluton (Fig. 41), further suggesting that there is a direct link between VMS mineralization and felsic magmatism (Walford and Franklin, 1982; Bailes *et al.,* 1991; Bailes and Galley, 1996). This hypothesis is supported by geochemical similarity of the deposit-hosting rhyolites and the Sneath Lake tonalite (Fig. 42; Bailes and Galley, 1996). Although both the rhyolite and tonalite display the same flat REE profiles as the Welch low-Ti tholeiites (Fig. 43), it is significant that ε_{Nd} values of the rhyolite/tonalite magmas (+3.7) are more primitive than those of the Welch basalt (+3) (Fig. 44). This precludes a direct relationship through fractionation between the Welch basalt and the felsic magmas. The implication is that geochemically primitive felsic magmas may be important for the generation of these base metal-rich VMS deposits. Note also that Snow Lake primitive arc rhyolites (Fig. 42) do not display the Eu depletion anomalies typical of many ore-hosting rhyolites in the Archean Abitibi belt (Lesher *et al.*, 1986).

The integral role played by the Sneath Lake tonalite in VMS generation is most clearly evident in the Anderson Lake area where there is a steady increase in both intensity of alteration of supracrustal rocks and of the Cu-Zn ratio (Cu/Cu+Zn) of contained VMS deposits, from 27 (Linda) to 97 (Anderson Lake), as distance to the pluton decreases. The latter is interpreted to result from higher temperature for pluton-proximal portions of the regional synvolcanic hydrothermal system (Bailes and Galley, 1996).

The evolved arc sequence is compositionally bimodal, containing a series of mafic flows (e.g. Snell, Moore, Vent, Threehouse), felsic flows (Powderhouse dacite, Chisel rhyolite, Ghost rhyolite, Photo rhyolite) and subvolcanic felsic intrusions (*e.g.* Richard Lake tonalite) (Figs. 40, 46). Unlike the basalt-dominated, bimodal, primitive arc section, the evolved arc is characterized by a series of thin, discontinuous units, including a large number of volcaniclastic deposits. The high proportion of volcaniclastic rocks (between 40 to 50 %) and local unconformities (Bailes and Simms, 1994) in the evolved arc section suggests that considerable topographic relief was developed, possibly as a result of intra-arc extension and rifting.

Basalt and basaltic andesite units in the evolved arc section range in thickness from 100 m to >1000 m, and all share the arc-signature "decoupling" of LIL and HFS elements shown by the primitive arc Welch basalts and basaltic andesites. They also display geochemical characteristics that are consistent with formation in a more mature or evolved arc setting (Fig. 45a). For example, the geochemistry of the Snell suite, which displays anomalously high Th contents and the highest apparent proportion of "old" Nd (-0.4 to +2.4 ε_{Nd}), of all the arc rocks in the Flin Flon belt (Fig. 45b), is best interpreted as having been acquired through intracrustal contamination (Stern *et al.*, 1992, 1995a). The presence of inherited late Archean zircons (*ca.* 2.7 Ga) in the directly underlying Stroud rhyolite unit (David *et al.*, 1996) suggests that this contamination may have been in part from Archean crust in the basement to the Snow Lake arc segment. The stratigraphically higher Moore Lake basalt, although displaying a return to isotopically primitive ε_{Nd} values (+2.5), is anomalous in having distinctly higher abundances of Ti, Th, U, Zr, Y, Nb, P_2O_5, Sr and light REE at equivalent MgO than the primitive arc Welch suite. Plank and Langmuir (1988) and Woodhouse and Johnson (1993) point out that incompatible trace element enrichments in arc volcanic rocks, such as those shown by the Moore basalt, may reflect eruption through thicker

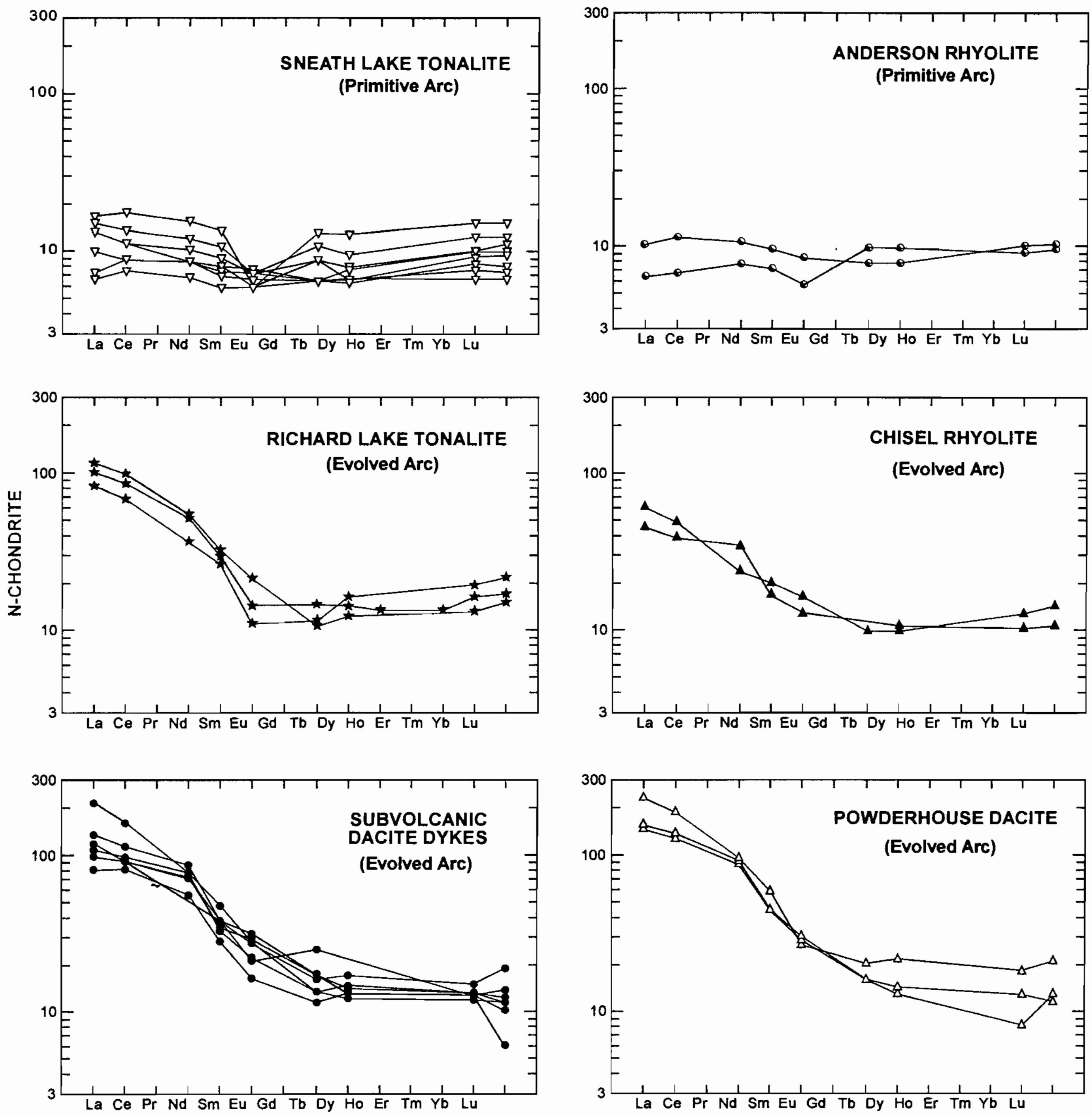


Figure 42: Chondrite-normalized REE patterns of ore-hosting Anderson and Chisel rhyolites compared to synvolcanic Sneath Lake and Richard Lake tonalites. Subvolcanic dacite dykes have REE values that are comparable to those displayed by the Powderhouse dacite. Note the absence of negative Eu anomalies for the VMS-hosting Anderson and Chisel ryholite complexes and for the subvolcanic Sneath and Richard Lake tonalite plutons.

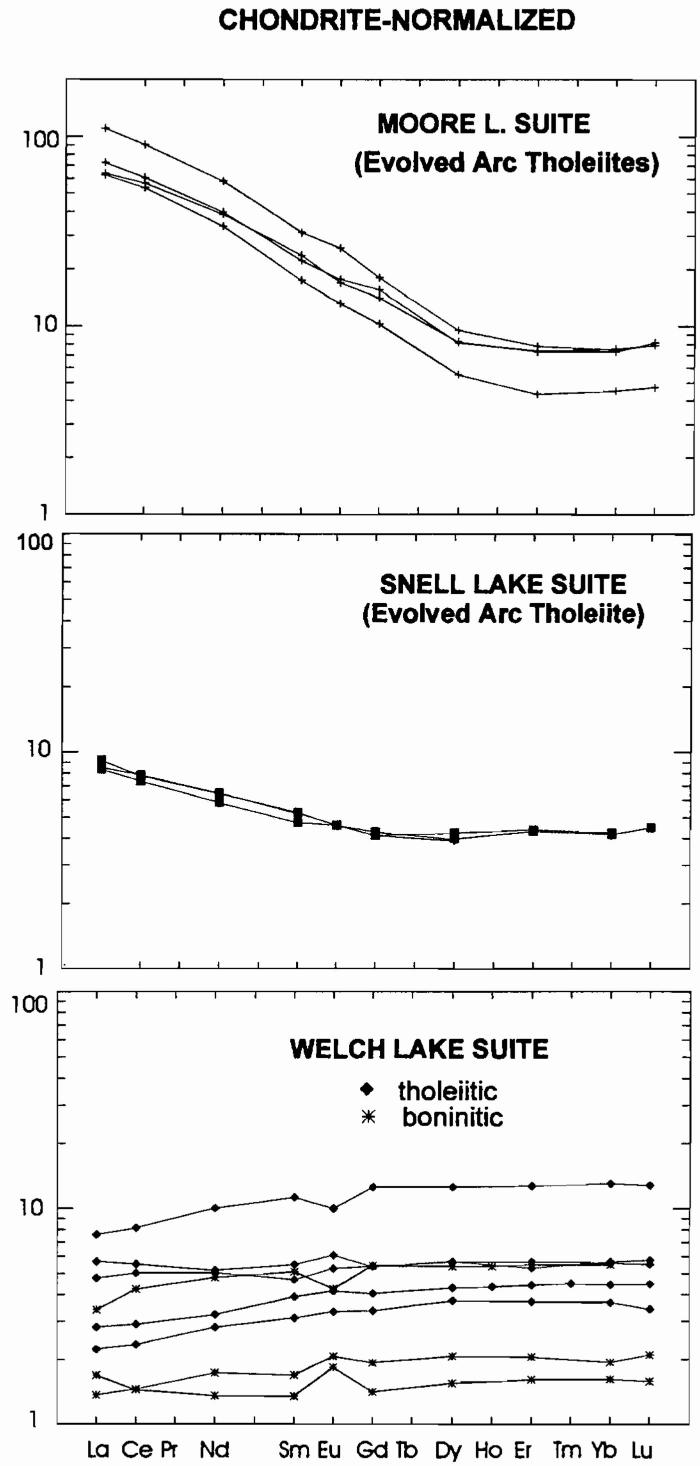


*Figure **43**: Chondrite-normalized REE patterns of major basalt and basaltic andesite units at Snow Lake from Stern et al. (1995a).*

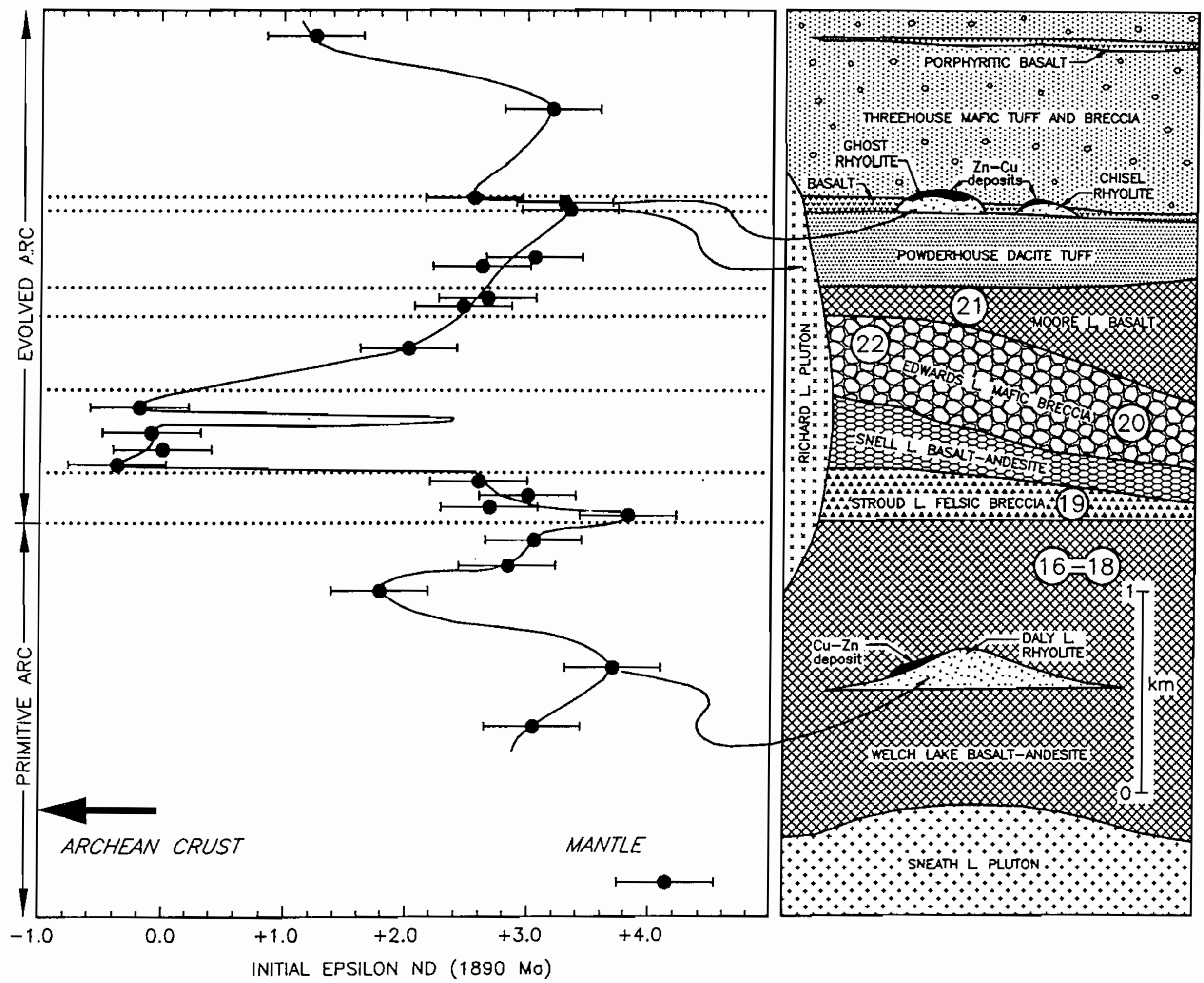


*Figure **44**: Isotopic and geological stratigraphy of the Chisel Lake section of the Snow Lake arc assemblage (modified from Stern et al., 1995a). Field trip stops are shown by circled numbers. Note the isotopically more primitive character (more positive values) of felsic magmas than the associated mafic magmas that precludes the felsic magmas from being derived through fractionation of the parent magma that produced the basalt flows.*

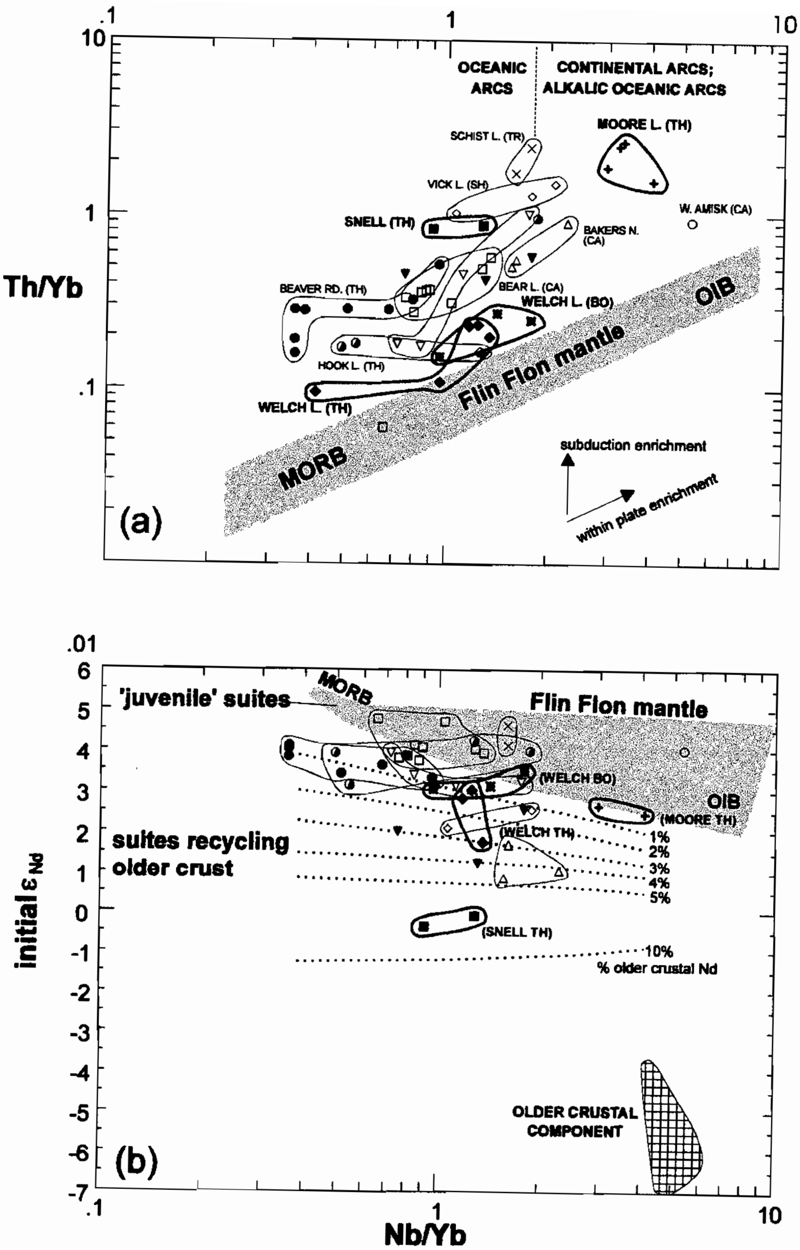


Figure 45: Basalts of the Flin Flon Belt on plots of: a) Th/Y versus Nb/Yb and b) initial ε_{Nd} versus Nb/Yb (from Stern et al., 1995a). Note the low ratios of Th/Y for Primitive arc basalts (Welch tholeiite, Welch boninite) and the much higher ratios for Evolved arc basalts (Snell tholeiite, Moore tholeiite). The low initial Nd values for the Snell Lake mafic flows is interpreted by Stern et al. (1995a) as a consequence of intracrustal contamination by older crustal material.

lithosphere (*e.g.* mature arc) as a consequence of lower average extents of melting due to the shorter length of the adiabatic melting column.

Four Zn-rich VMS deposits are located in the evolved arc sequence, approximately 2 km stratigraphically above its base (Fig. 40). Three deposits (Chisel, Lost and Ghost) have been mined and contained a total of 7.8 Mt (production plus reserves) grading 0.42% Cu and 10.06% Zn. The fourth deposit (Chisel North) contains 2.6 Mt grading 0.23% Cu and 8.9% Zn and has been drilled off, but is yet unmined. A recently discovered Cu-Zn-Au rich VMS deposit at Photo Lake, 4 km north of Chisel Lake (Fig. 36), has uncertain affinity to the Chisel area Zn-rich deposits (Bailes and Simms, 1994). The Chisel Lake Zn-rich VMS deposits are located along a linear alteration zone that is interpreted to be the expression of a synvolcanic fault (Galley *et al.*, 1993).

Zn-rich VMS deposits in the evolved arc, like those in the primitive arc section, are stratigraphically related to felsic magmatic activity: they are directly underlain by the Powderhouse dacite and are either hosted by or spatially associated with subaqueous rhyolite flows (Chisel, Ghost; Fig. 40). Also in the stratigraphic footwall is an extensive system of subvolcanic dacite dykes that are geochemically indistinguishable from the Powderhouse dacite, as well as the subvolcanic Richard Lake tonalite pluton that is geochemically identical to the rhyolite flows. The felsic magmas are characterized by enrichment in incompatible trace elements (*e.g.*, Th, U, Zr, Y, Nb, P_2O_5, and light REE). This presumably reflects derivation from the same fertile magma source as postulated for the Moore basalt, but ε_{Nd} values for the Chisel-Ghost rhyolite/Richard Lake tonalite (+3.3) are more primitive than for the associated Moore basalt (+2.5; Fig. 44), again precluding direct fractionation from a common parent magma. Thus the VMS-hosting rhyolite flows and subvolcanic tonalite plutons in both the primitive and evolved arc sequences display an isotopic character that is distinct from that of associated basalts. This observation, when considered in conjunction with the association of mineralization episodes with arc rifting events, suggests that the felsic rocks may be generated by deep crustal melting triggered by high heat flow (*cf.* Barrie, 1995). VMS-hosting evolved arc rhyolites do not display an Eu depletion anomaly, suggesting that this widely used criteria for identifying ore-hosting rhyolites in the Archean Superior Province (Lesher *et al.*, 1986) must be used with caution in the Snow Lake segment of the Paleoproterozoic Flin Flon belt.

The preponderance of Cu-rich VMS deposits in the primitive arc section compared to the abundance of Zn-rich deposits in the evolved arc sequence is clearly associated with the change from primitive (tholeiitic and boninite suites, thinner crust) to more evolved (thicker crust) magmas. In this light the change in metal ratios may reflect an overall reduction in high level heat flow due to crustal thickening. An alternative explanation is that the voluminous volcaniclastic deposits in evolved arc section resulted in more freely circulating fluids and a cooler hydrothermal regime, one that was incapable of maintaining the higher temperatures needed to form Cu-rich VMS desposits.

Stops 16 to 18 will examine both the low-Ti tholeiite basalts and boninite flows (Fig. 40, 46) that characterize the primitive arc sequence, as well as a prominent zone of sea floor, feldspathization/silicification that effects the upper 500 m of this succession Fig. 41, 47). Stops 19 to 22 emphasize the volcaniclastic units that dominate the evolved arc section (Fig. 40, 46), as well as the relationship between subvolcanic intrusions, subseafloor hydrothermal activity and alteration in these permeable rock formations.

STOP 16: Welch Lake boninite, Primitive Arc

Introduction

Boninites are rare in modern arcs and almost unknown in the Paleoproterozoic. The lowermost of three north-facing mafic flows at this stop (Figs. 46, 48) displays the chemical characteristics of a high-Ca boninite (Stern *et al.*, 1995), and the overlying two flows display boninite-like characteristics (*e.g.* higher MgO, Ni and Cr values than most tholeiitic rocks at Snow Lake). High-Ca boninites in the primitive arc sequence at Snow Lake suggest that this sequence may have been deposited in a forearc eruptive setting (Stern *et al.*, 1995a). Similar to modern counterparts, boninites in the Welch Lake suite are intimately associated with low-Ti tholeiitic basalts.

Location A: "High-Ca boninite"

The "high-Ca boninite" (flow 1) is over 35 m thick, aphyric, pillowed, and a distinctive pale pistachio green weathering colour. Pillows are blocky, display narrow (<5 cm) selvages, are 30-80 cm in diameter, contain 10-15% quartz and carbonate amygdales (0.5-3 mm), and are surrounded by light tan coloured, recrystallized interpillow hyaloclastite. A 1 m wide synvolcanic porphyritic mafic dyke, that intrudes between but never across pillows, is texturally identical to the overlying flow 2.

Location B: Contact between flows 1 and 2

The contact between flow 1 and 2 is covered by a narrow (<0.5 m wide) gully. The upper 4 m of flow 1 displays higher vesicularity and thicker domains of interpillow hyaloclastite than at location A. Flow 2 is easily distinguished from flow 1 by its pyroxene (amphibole-replaced)-plagioclase phyric character, medium to dark grey green weathering colour, and smoothly curving pillow morphologies. Pillows in flow 2 are locally imbricated consistent with flow transport from SW to NE (present geographic coordinates). Pillows in flow 2 display narrow selvages and up to 1 cm of rusty weathering interpillow hyaloclastite. Pillow size is relatively constant from bottom to top of this 5m thick flow. Southwest of location B, flows 1 and 2 are separated by up to 6 m of laminated mafic scoria tuff and lapilli tuff which includes a narrow porphyritic pillowed flow that is locally only one pillow thick.

Location C: Contact between flows 2 and 3

Flows 2 and 3 are separated by a 0.3-3 m unit of mafic scoria tuff and lapilli tuff that displays both reverse and normal size grading, and, in one locality, current lamination; the latter indicates the same SW to NE component of transport as the pillow imbrication in flow 2. The tuff/lapilli tuff contains local accessory clasts that are identical to the underlying flow. Flow 3 is similar in textural and composition to flow 2, but has a slightly lower phenocryst population than flow 2 and contains 1-3% quartz-filled radial pipe vesicles (up to 5 cm in length) that are not observed in flow 2. The radial

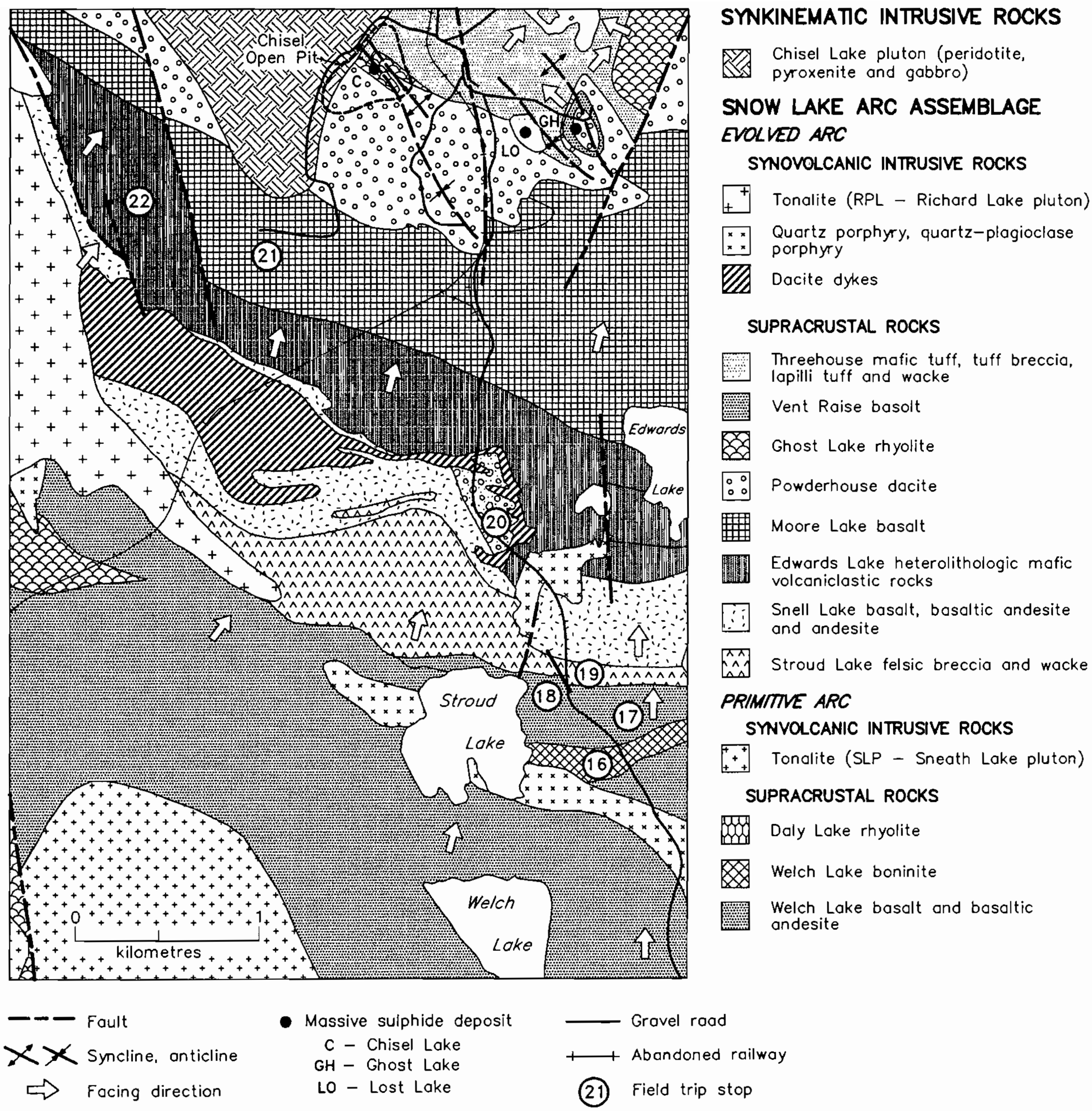


Figure46: Geology of the Chisel Lake area showing location of Stops 16 - 22. Stops 16 - 18 are located in basalt -basaltic andesite flows of the Primitive Arc. Stops 19 - 22 are located in volcaniclastic rocks that typify the Evolved Arc.

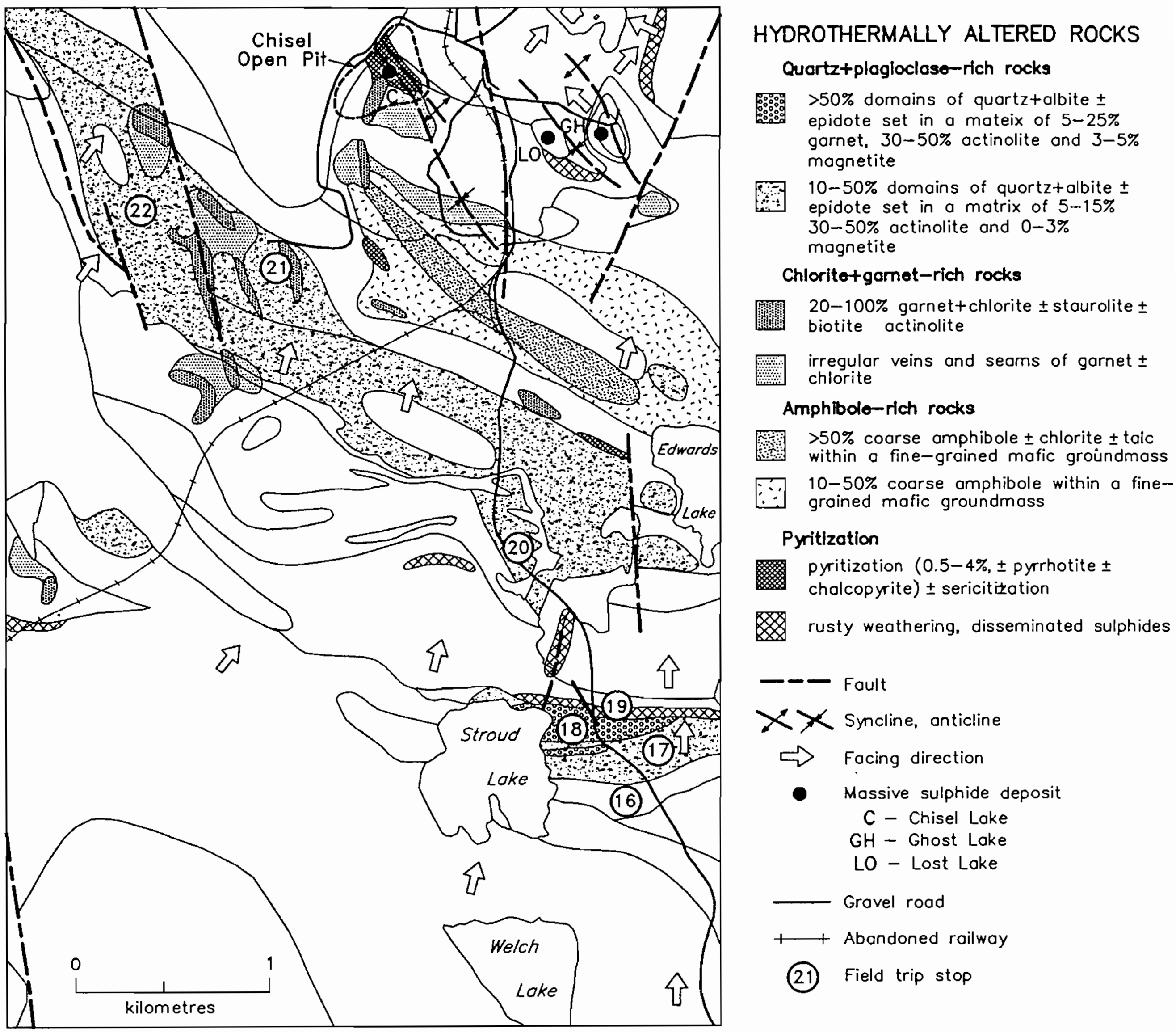


*Figure **47**: Large portions of the Chisel Lake section were hydrothermally altered contemporaneous with active volcanism. This figure shows the distribution of hydrothermally altered rocks and the location of Stops 16 - 22.*

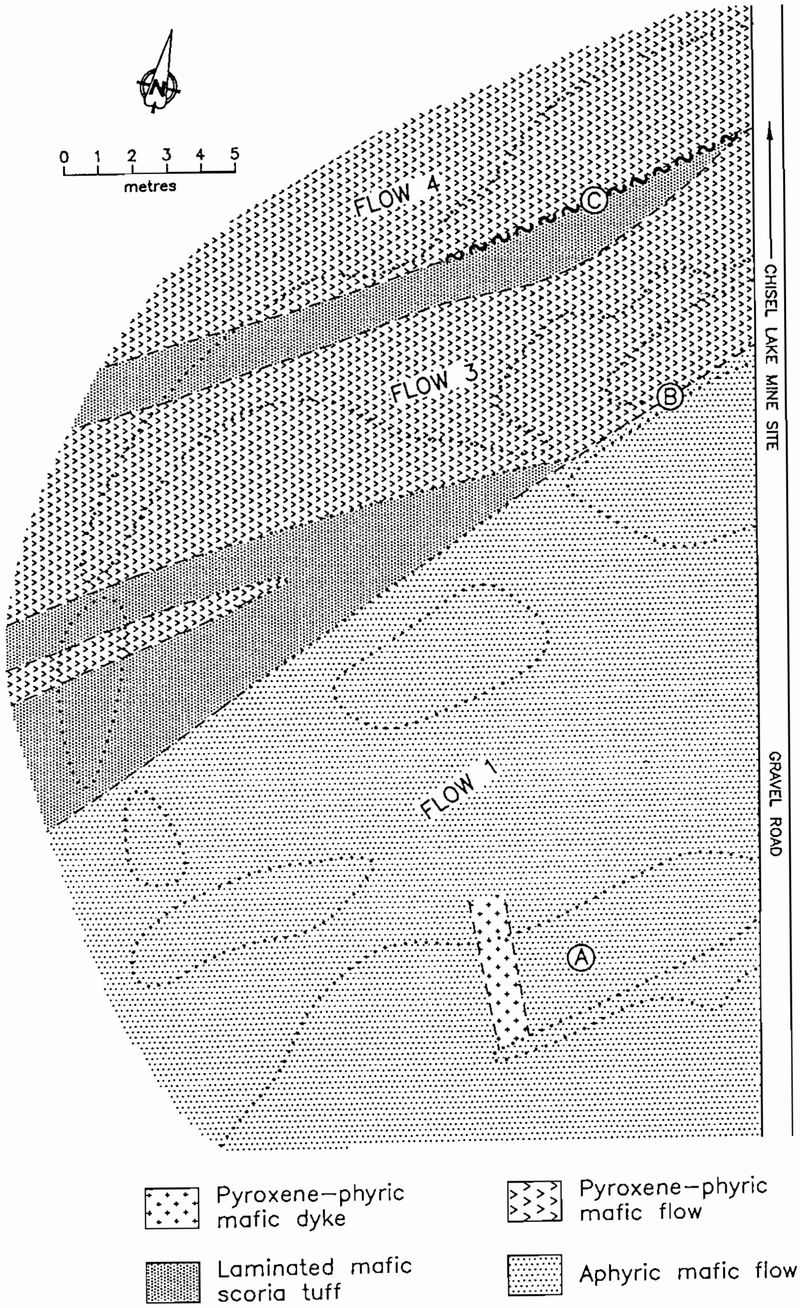


*Figure **48**: Series of boninite and boninite-like flows at Stop 16 in the Welch suite of the Primitive arc sequence, Snow Lake. The three locations to be examined are shown by circled letters.*

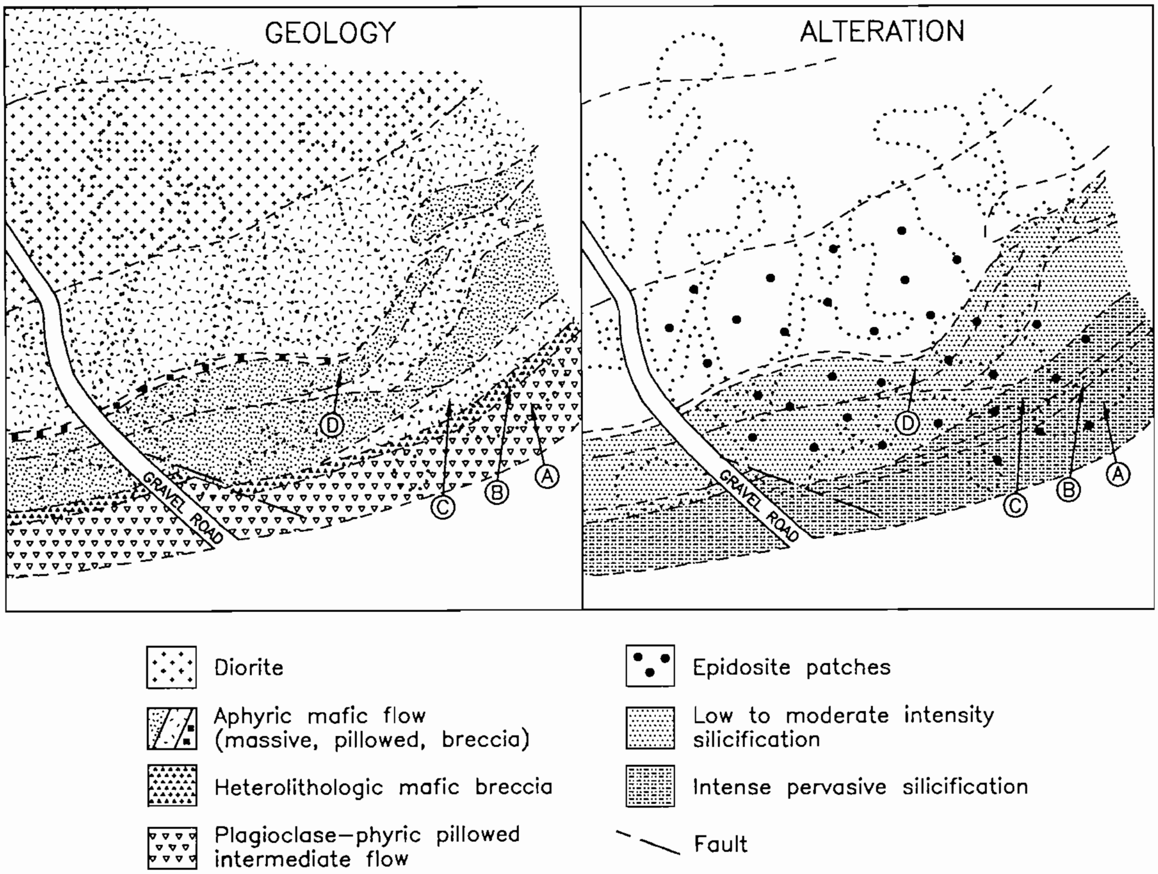


*Figure **49**: Series of variably feldspathized/silicified Welch suite low-Ti tholeiite flows at Stop 17, Snow Lake. The distribution of geological units and alteration types is from mapping by Roger Skirrow (1987).*

pipe vesicles in flow 3 and the intercallated scoria-rich tuffs suggest a moderately shallow water depositional environment (Moore and Schilling, 1973; Jones, 1969, 1970).

STOP 17: Welch Lake low-Ti basaltic andesite and andesite.

Introduction

Low-Ti arc tholeiites of the Welch Lake suite form a >3 km thick sequence that dominates the Primitive arc section (Figs. 36, 40, 46). They display high contents of large ion lithophile elements (LIL; eg. Rb, Ba, Th, Sr), low contents of high field strength (HFS; eg. Hf, Ti, Zr, Y) elements, and very low Ni and Cr relative to N-MORB (Figs. 7, 8); these features characterize subduction-related magmas (Gill, 1981; Tarney et al., 1981) formed within modern oceanic arc tectonic settings. The Welch Lake arc tholeiites display flat to slightly LREE depleted patterns (Fig. 43), similar to those displayed by the boninitic flows but with higher overall REE values (3-10X chondrite).

The feldspathized/silicified mafic flows at this stop are part of a 0.5 x >20 km semiconformable alteration zone at the top of the Primitive arc section in the Welch Lake formation (Fig. 41). The alteration zone underlies a thin zone of <5m sulphidic sediments known locally as the Foot-Mud horizon. The close proximity of the alteration zone and the directly overlying Foot-Mud unit suggests that the geothermal activity responsible for the alteration zone may also have produced the sulphidic sediments. The Foot-Mud unit postdates the prominent Cu-rich VMS deposits in the Primitive arc sequence, which are associated with rhyolite flow complexes lower in the stratigraphic sequence (Fig. 40).

At this stop, located less than 300 m below the top of the Primitive arc section (Fig. 40), we will examine a series of moderately feldspathized and silicified low-Ti arc tholeiite flows (locations A, B, C and D). The stop descriptions are based on work by Skirrow (1987), who studied alteration of these flows as part of his M.Sc. thesis at Carleton University.

Alteration

Feldspathized/silicified pillows at this stop are strongly zoned, generally with an outer 1-2 cm wide selvage of hornblende-oligoclase-quartz, a 10-20 cm thick silicified margin, and an interior zone with 10-30 cm diametre domains of yellow-green clinozoisite and quartz. Elemental gains and losses during alteration of pillows, calculated using the method of Grant (1986), are summarized below from Skirrow (1987):

	Gains	Losses	Immobile	Inconsistent
Selvage	Fe_2O_3T, MgO, Zn	SiO_2, Na_2O	TiO_2, Zr	CaO, K_2O, Y, Cu,V
Margin (silicified)	SiO_2, Na_2O, Y, Zn, Cu?	Fe_2O_3T, MgO, V, CaO	TiO_2, Zr, Al_2O_3	K_2O
Core (epidotized)	SiO_2(small), Y?, CaO, Fe_2O_3,	Na_2O, FeO, Zn	Zr, Al_2O_3,	MgO, TiO_2, Fe_2O_3T

Pillow margins show gains in SiO_2 that cannot be accounted for by redistribution of SiO_2 within a particular pillow; therefore, a source of SiO_2 external to the pillow is indicated. An overall small loss of Fe_2O_3(T) and MgO from altered pillows is likely.

Epidotization of pillow cores involves large gains of CaO, oxidation of ferrous to ferric iron and large losses of Na_2O. Chemical changes in zones of patchy epidotization are similar to that displayed in epidotized pillow cores.

Interpretation

Skirrow (1987) interpreted alteration of mafic flows at this stop to be a product of interaction of diffuse, near-seafloor silica-rich hydrothermal fluids with still hot Welch Lake mafic lava flows. The heat source producing the silica-rich hydrothermal discharge is suggested by Skirrow (1987) to have been the subvolcanic Sneath Lake tonalite intrusive complex. The silica could have been derived locally from alteration of glassy pillow selvages and interpillow hyaloclastite, or from lower parts of the volcanic pile (Skirrow and Franklin, 1994).

The observed zonation of alteration in pillows at this stop is interpreted by Skirrow (1987) to have been produced by rapid cooling of the outermost parts of the lava pillows against cold seawater, producing a glassy selvage, and establishing a large temperature gradient between the selvage and the hot pillow margin and core. Silica was precipitated in the pillow margins as the temperature of the invading silica-rich hydrothermal fluid was raised to >350° C (Kennedy, 1950). Fe, Ca and possibly Mg were leached from the pillow margins almost simultaneously with silica precipitation, promoted by high temperature and initially low pH (Seyfried & Bischoff, 1977). The greater intensity of alteration at the tops and bottoms of pillows, observed at locations A and C, suggests that there was greater heat retention near these relatively flat surfaces than at their pillow sides where the radius of curvature of the pillow surface is smaller. Skirrow suggests that silicification of the pillow margin probably sealed the hotter interior from further interaction with the externally-derived SiO_2-rich fluids, but redistribution of elements internal to the margin may have led to epidotization of the pillow core.

Feldspathization/silicification of pillows and flows is variable. A broad correlation between intensity of alteration with portions of pillows and flows that display abundant thermal contraction cracks suggests that altering fluids may have gained access to the interior of pillows and flows via these cracks.

Location A: Intensely feldspathized/silicified pillows, plagioclase-phyric mafic flow

At this location the upper 15 m of a strongly feldspathized/silicified plagioclase-phyric mafic flow, overlain to the north by heterolithologic breccia (location B), is exposed. Pillows in this flow are relatively large, contain 5-10% plagioclase phenocrysts, and exhibit quartz-filled concentric thermal contraction cracks. Vesicles increase slightly in abundance towards the top of the flow (northwards) and are filled with quartz or quartz-epidote. Some of the features accompanying feldspathization/silicification of a pillowed mafic flow are well displayed at this outcrop.

Altered pillows at this outcrop show a characteristic zonation from margin to core, as follows: a) a 1-2 cm wide medium grey selvage composed of hornblende + oligoclase + quartz and, in places, garnet + chlorite + biotite; b) a 10-20 cm silicified light grey margin containing quartz + oligoclase + hornblende + magnetite + garnet; c) and a core with 10-30 cm diameter yellowish-green, epidotized domains composed mainly of clinozoisite and quartz. The silicified margin typically includes white weathering intensely silicified domains (5-20 cm) composed of quartz + oligoclase (+

magnetite + hornblende), generally localized near the top and bottom of pillows. In strongly altered pillows these white weathering zones of intense silicification continue completely around the pillow forming "doughnut" structures (this feature is well developed at Stop 18).

Patches of yellow-green epidozite on this outcrop are surrounded by light grey to white domains of silicification similar to those at the margin of altered pillows. Thermal contraction cracks pass through both the silicified and epidotized domains without deviation indicating alteration was an approximately constant volume exchange of elements.

Location B: Fragments of feldspathized/silicified pillows in a heterolithologic breccia bed

A normally graded, heterolithologic breccia bed overlies and fills depressions in the plagioclase-phyric pillowed mafic flow of location A. The breccia bed has a massive base and indistinctly stratified top, contains angular to subangular clasts up to 40 cm in diameter, grades upwards into granule-sized detritus, and is framework-supported.

The important feature at this location is the presence of previously feldspathized/silicified fragments of plagioclase-phyric pillows (identical to the flow at location A) in this directly overlying breccia bed. This indicates that feldspathization/silicification of the underlying flow (location A) preceded deposition of the heterolithologic breccia bed. For this reason Skirrow (1987) interpreted the alteration to have occurred very close to, or at, the seafloor, probably while flows were still hot.

Location C: Feldspathized/ silicified pillows in aphyric mafic flow

A less intensely feldspathized/silicified pillowed mafic flow than that at location 1 is exposed here. The zonation of alteration within the pillows is almost identical to that at location 1. However, here pillows are smaller, selvages are more hornblendic, and intensely silicified patches (and associated thermal contraction cracks) occur closer to the selvages.

Location D: Patches of feldspathization/ silicification and epidotization in a massive aphyric flow

A massive 20 m thick mafic flow with distinctive patches of feldspathization/silicification and epidotization overlies, and is separated from, the silicified flow at location C by several massive and pillowed flows, including a pillow fragment breccia. Alteration patches at location 4 are sinuous, typically 0.3 to 1.5 m long, and elongate parallel to thermal contraction cracks (i.e. they are subparallel to lower contact of the flow). Most are concentrically zoned from an outer dark green hornblende-rich periphery, to a middle light grey quartz, plagioclase, hornblende and magnetite-bearing silicic zone, to an interior light yellowish green core composed of clinozoisite, quartz and minor sphene and carbonate. Some alteration patches are connected and appear to form an anatomizing network. Thermal contraction cracks in this flow are typically filled by hornblende, but where they pass through the silicic peripheries of alteration patches they are filled with quartz and plagioclase.

STOP 18: Feldspathized/silicified Welch Lake basaltic andesite

An excellent example of a feldspathized/silicified aphyric pillowed Welch Lake basaltic andesite flow is exposed at this Stop. The alteration is part of the same zone of feldspathization/silicification observed at Stop 17. Overlying Stroud Lake felsic breccias of the Evolved arc section are unaffected by the prominent feldspathization/silicification, supporting the interpretation that the alteration formed very close to, or at, the paleo seafloor.

Proceeding west (along strike) from the road, weakly to moderately feldspathized/silicified pillows are crossed on route to the intensely altered pillows. At the stop, white weathering zones of intense feldspathization/silicification form a complete ring or "doughnut" around the inner margin of pillows, and enclose less altered or epidotized pillow cores. The form and chemical composition of the alteration is very similar to that at Locations A and C of Stop 17. The origin of the alteration is considered to be the same as that previously described for the rocks at Stop 17.

STOP 19: Stroud Lake felsic breccia

The transition from the Primitive arc to Evolved arc is characterized by an abrupt change from a sequence dominated by subaqueous mafic flows (Stops 16-18; Figs. 40, 46) to one dominated by units composed of heterolithologic volcaniclastic detritus (Stops 19-22). It is also marked by distinctive changes in geochemical character of associated mafic flows (discussed below).

At this stop, located 50 m above the base of the Evolved arc sequence, a 29 m section of north-facing, well-bedded, monolithologic to heterolithologic Stroud Lake felsic breccia and wacke is exposed. The southern 11 m consists of a series of <5 cm and up to 2.6 m thick intermediate to felsic wacke beds that display bed forms and sedimentary structures characteristic of turbidity and fluidized sediment flows. Normal size grading, scour channels, rip-ups, synsedimentary faults, load structures and beds with A, AB and ABD Bouma zonation are present. The northern 18 m of the outcrop consists of a single (?) felsic bed characterized by distinctive, large quartz and plagioclase phenoclasts. The lower 8.5 m of this bed is massive and texturally uniform with 10% quartz phenoclasts (2-15 mm) and 15-20% plagioclase phenoclasts (2-5 mm) in a fine grained matrix. The upper 9.5 m is clearly fragmental and is possibly a separate bed. It includes a wide variety of over 30 cm diameter subangular to subrounded felsic blocks, including some that are texturally identical to the lower 8.5 m of the bed. The felsic blocks, as well as phenoclasts of quartz and plagioclase in the intervening detrital matrix, all display normal size grading. The top of the bed is a fine grained felsic sandstone.

The Stroud Lake felsic breccia is both overlain by and locally intercalated with Snell Lake subaqueous mafic flows. These mafic flows display anomalously high Th contents (Fig. 45) and the highest apparent proportion of "old" Nd (-0.4 to +2.4 ε_{Nd}; Fig. 44), of all the arc rocks in the Flin Flon belt (Fig. 45b). Stern *et al.* (1992, 1995a) attribute these features to intracrustal contamination.

David *et al.* (1996) analyzed zircons from a sample from a 1 m thick felsic wacke bed located at the southern end of this outcrop. The sample yielded two zircon sample populations. One suite defined a discordia line with an upper intercept of 1892 ± 3 Ma, interpreted to be the age of crystallization of felsic material forming the felsic wacke. Within error this age is identical to that determined for the Sneath Lake (1886+17/-9 Ma) and Richard Lake (1889+8/-6 Ma) tonalite plutons, consistent with their interpretation as subvolcanic intrusions (Bailes *et al.*, 1991). The second population yielded inversely discordant Pb-Pb ages of 2652 Ma and 2674, and Pb-Pb discordant ages of 2715 Ma, 2823 Ma and 2691 Ma that David *et al.* (1996) interpret as inherited. The *ca.* 2.7 Ga inherited zircons support the interpretation by Stern *et al.* (1992, 1995a) that the high proportion of "old" Nd in the intercalated Snell mafic flows is due to intracrustal contamination, possibly through contamination by Archean crust in the basement to the Snow Lake arc segment.

The coarsely porphyritic felsic blocks in the 18 m thick bed at this stop are texturally and geochemically indistinguishable from one of the phases of the Primitive arc Sneath Lake tonalite. This suggests that the Stroud Lake felsic breccia may have been derived by uplift and erosion of the underlying Primitive arc section coincident with outpouring of the isotopically anomalous Snell Lake mafic flows.

STOP 20: Feldspathization/silicification associated with dacite dykes in the Edwards Lake formation

The Edwards Lake formation consists of up to 1200 m of fine grained mafic volcaniclastic sediments and heterolithologic mafic volcanic breccia. The lower 550 m consists of fine grained mafic volcaniclastic sediments (observed at this Stop). The upper 650 m consists of heterolithologic mafic breccias that will be examined at Stop 22. Because much of this volcaniclastic formation is altered (Fig. 47) it has been interpreted to have been a hydrothermal aquifer (Skirrow, 1987; Bailes and Galley, 1989; Skirrow and Franklin, 1994). The alteration is spatially associated with an array of subvolcanic mafic to felsic sills, dykes and plugs that intrude both the Edwards Lake heterolithologic mafic wacke/breccia and the underlying Snell Lake mafic flows. Most prominent among the synvolcanic intrusions is a large swarm of aphyric and plagioclase phyric dacite dykes (Figure 40). The dacite dykes are identical in texture and chemistry to the Powderhouse dacite tuff (Figs. 40, 42), the unit that directly underlies the Chisel-Lost-Ghost-North Chisel Zn-Cu sulphide deposits (Bailes and Galley, 1996).

At this stop we will examine feldspathized/silicified fine grained mafic wackes of the Edwards Lake formation and demonstrate the relationship between this alteration and dyke emplacement. The mafic wackes are typically massive and featureless, such that altered varieties are difficult to distinguish from the dykes. However, at this stop the dacite intrusions are sheet-like bodies, cross-cut bedding, and have sharp, commonly flow-banded margins. They are creamy white in colour and are plagioclase-phyric with a fine grained groundmass of quartz, plagioclase and minor biotite. Plagioclase phenocrysts are commonly enveloped or partially replaced by biotite giving dykes a spotted appearance. Adjacent to dacite dykes the host mafic volcanic wacke is weakly to intensely feldspathized/silicified (Fig. 50). Silicified rocks are mottled shades of grey in colour, are composed of plagioclase, quartz and hornblende, and are slightly darker

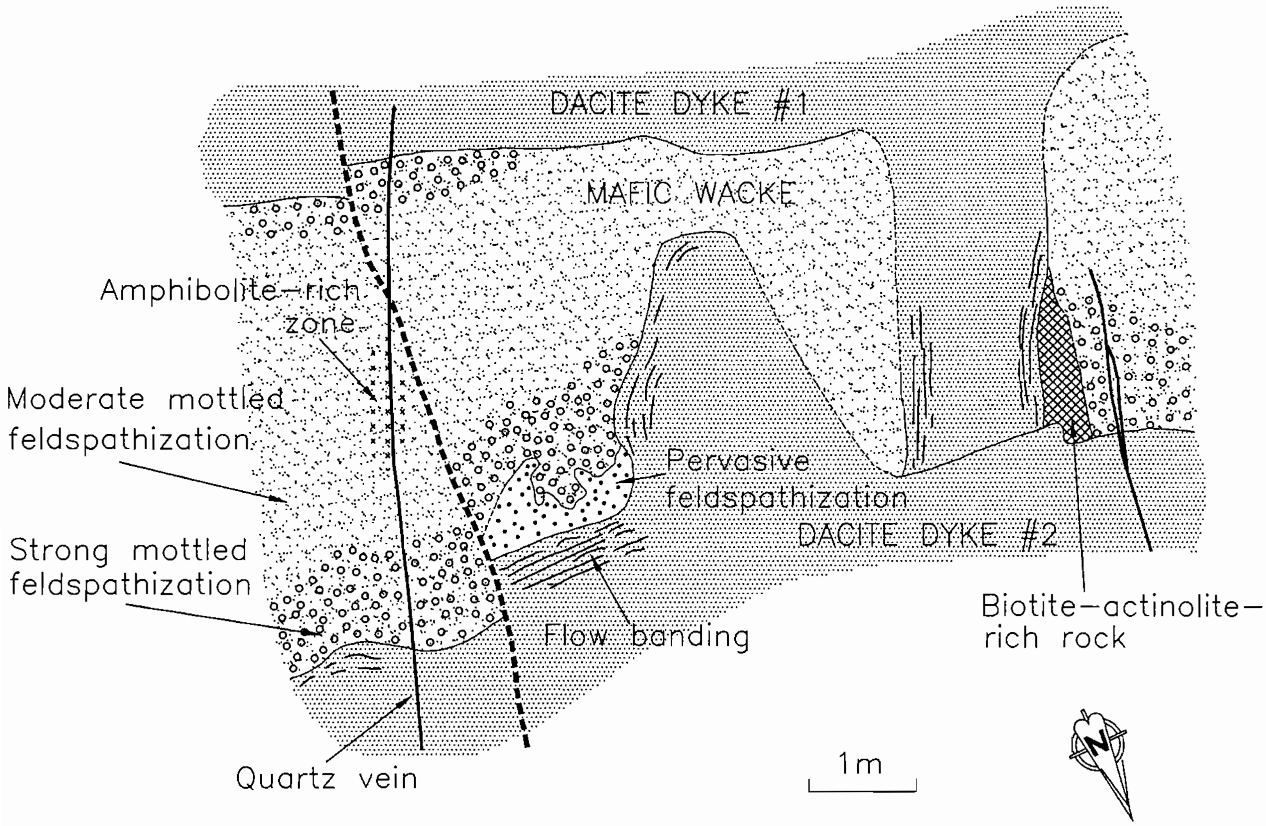


***Figure 50**: Sketch of the distribution feldspathized/silicified Edwards Lake volcaniclastic rocks adjacent to synvolcanic dacite dykes at Stop 20, Snow Lake. The dacite dykes are feeders for the Powderhouse dacite, a unit that forms the footwall to Zn-rich VMS deposits at Chisel Lake.*

in colour than the adjacent dacite dykes. More than one age of dacite dykes have intruded the mafic wackes at this stop (Fig. 50)

Mass balance calculations by Skirrow (1987) of elemental gains and losses in feldspathized/silicified rocks indicate that during feldspathization/silicification leaching of Fe and Mg accompanied the addition of SiO_2. Kennedy (1950) has shown that SiO_2 solubility decreases with rising temperature between 350°C and 420°C at pressures of 800 bars, and Mottl *et al.* (1979) have demonstrated that leaching of Fe and other transition metals from basalt by seawater is most significant at temperatures greater than 380° C in experimental systems. Based on these experimental results Skirrow (1987) suggested that felsic dykes at this outcrop were emplaced into a SiO_2-rich hydrothermal system, locally heating the fluids to greater than 350° C, resulting in feldspathization/silicification adjacent to the dyke margins.

These dacite dykes are the intrusive equivalent of the Powderhouse dacite, and this implies that the alteration at this outcrop, although broadly similar to the seafloor feldspathization/silicification at Stops 17 and 18, took place at depths of up to 2 km below the paleo seafloor. Although the dykes may not have been an adequate heat source to raise the overall geothermal gradient, (Bailes and Galley, 1996) speculate that they played an important role in focusing hydrothermal activity that was capable of leaching metals from host strata.

STOP 21: Highly altered Moore Lake mafic breccia

At this outcrop an east-trending unit of variably feldspathized/silicified monolithologic Moore Lake formation mafic volcanic breccia (Figs. 46, 47) is cross-cut, at a high angle, by a prominent zone of chlorite-rich altered rocks characterized by abundant porphyroblasts of staurolite and garnet in a chlorite-biotite-rich groundmass. The crosscutting chlorite-rich alteration zone at this stop is possibly related to a synvolcanic fault. Least altered breccias at this stop include complete pillows as fragments and broken pillow pieces, and are interpreted to be pillow fragment breccias derived from cold fragmentation and slumping of Moore Lake basalt flows. Chemical analyses of pillow fragments from this breccia display the prominent light REE enrichment that characterizes Moore Lake basalt flows.

We will first examine the least altered breccias in the centre of the outcrop (location A) and then proceed to the most altered portion at the east end of the outcrop (location B).

Location A: Feldspathised/silicified and chlorite-rich altered mafic monolithologic breccia

The breccia at this location is coarse with many fragments up to 40 cm in diameter and rare blocks up to 1 m in size. It is composed of medium grey, light grey and white weathering blocks which appear to be a heterolithologic mixture of mafic to felsic fragments but which are actually variably feldspathized/silicified aphyric mafic fragments. The primary mafic composition of fragments is manifest by their morphology; most are broken pieces of pillows, some are complete pillows, and one large block comprises a piece composed of several intact pillows. Many fragments contain quartz amygdales. Rare slabby fragments of altered laminated mafic tuff (interflow sediments?) are present; they must have been lithified prior to incorporation into the breccia.

Pillow fragment pieces vary from relatively unaltered medium grey blocks to white weathering completely altered clasts composed almost entirely of quartz and feldspar. Many fragments are zoned with a less altered core (composed of amphibole, garnet, quartz, and fine grained plagioclase) and a 1 to 3 cm wide white weathering rim (composed of quartz, feldspar, minor amphibole and magnetite). Inter-fragment matrix is a completely recrystallized mixture of black amphibole, anhedral garnet, plagioclase, quartz and minor euhedral magnetite.

The alteration of breccias at location 1 occurred in two stages. The early stage involved feldspathization/silicification during which Fe and Mg were removed and SiO_2 was added. This stage of alteration is represented by replacement of mafic pillow fragments by quartz and feldspar. Zones of feldspathization/silicification completely rim margins of broken pillows indicating that this alteration postdated deposition of the breccia. The second stage of alteration consisted of addition of Fe and Mg and is represented by garnet, amphibole and chlorite-rich rocks. This stage of alteration began with the overgrowth of permeable interfragment domains, and proceeded to alter the previously silicified pillow fragments. The zone of Fe-Mg metasomatism becomes more intense towards location B. Proceeding east towards the cross-cutting zone of intensely altered rocks at location B, the breccia is gradually overprinted by Fe-Mg-rich minerals such as garnet, chlorite and amphibole. Initially, the garnet-chlorite-amphibole alteration-assemblage is confined to the interfragment domains (i.e. zones of high permeability) but as location B is approached the fragments themselves are also affected. The altered breccia just west of the staurolite-bearing zone is a garnet-amphibole-biotite-rich rock with numerous white quartzofeldspathic fragments. The interfragment domains are composed amphibole, biotite, garnet and plagioclase. The fragments contain up to 5% garnet but virtually no other ferromagnesian minerals.

Location B: Staurolite-garnet-chlorite-biotite-rich altered rocks

The strongly altered staurolite-rich rocks at location 2 form a cross-cutting north-trending zone (Fig. 47). The contact of this zone of alteration is rapidly gradational to the west into the less altered equivalents exposed at location 1. The most strongly altered rocks are characterized by subhedral to euhedral honey brown to dark brown staurolite porphyroblasts (up to 5 cm long). In this zone, fragments are much less conspicuous than to the west, and where preserved are composed of quartz and feldspar with up to 50% anhedral garnet. The remainder of this zone is composed of 20% subhedral to euhedral staurolite, anhedral to euhedral garnet, biotite, chlorite, and 50% fine grained quartz and plagioclase. It is not suprising that this rock was misidentified as a pelitic schist by early geologists (Harrison, 1949; Williams, 1966).

STOP 22: Feldspathization/silicification in Edwards Lake mafic heterolithologic breccia

The upper 550 m of the Edwards Lake formation consists of heterolithologic mafic boulder, cobble and pebble breccias intercalated with mafic wacke. The breccias are composed mainly of plagioclase- and plagioclase-pyroxene phyric mafic fine grained fragments with lesser amounts of other mafic and felsic volcanic clasts. Most breccia beds in the Edwards Lake formation are 1-20 m thick; Skirrow (1987) reports one 60 m bed. Organization of breccia beds is consistent with deposition by subaqueous debris flows (Bailes, 1986; Skirrow, 1987).

An estimated 30 to 40% of the mafic volcaniclastic rocks in the Edwards Lake formation have been feldspathized/silicified (Figs. 41, 47), with the alteration involving replacement of the mafic detritus by varying amounts of fine grained quartz and feldspar to produce a white or grey weathering rock. The feldspathization/silicification is more or less restricted to the Edwards Lake formation and has not significantly effected overlying and underlying formations, with the exception of some of the Moore Lake mafic breccias (*e.g.* Stop 21), suggesting that alteration was probably controlled by high primary permeability of this volcaniclastic unit, and that hydrothermal fluids causing the feldspathization/silicification probably moved laterally.

Feldspathization/silicification in the Edwards Lake formation has a number of different forms. One is the selective alteration of clast margins and, in some instances, the entire fragment. This type of alteration of mafic fragments is easily recognized because of the contrast between less altered dark cores and quartz-plagioclase rich light coloured margins; some of these feldspathized/silicified fragments also display an inner core of epidosite. A second type of alteration (mottled feldspathization/silicification) consists of millimetre-to centimetre-size oval to irregular domains composed of fine grained quartz-plagioclase-(epidote). Typically these oval alteration domains are distributed randomly throughout both matrix and fragments, indicating that the alteration postdated deposition of the volcaniclastic rocks. Other less prominent types of feldspathization/silicification include selective alteration of interfragment matrix, of wallrock adjacent to fractures , and of rocks adjacent to dykes.

Over 20% of the outcrop at this stop (Fig. 51a) consists of narrow (~1-4 m) dykes that range from mafic to felsic in composition (Fig. 51b). Many of the dykes have irregular shapes with some displaying amoeboid protrusions into the host volcaniclastics of the Edwards Lake formation. The dykes are interpreted to be synvolcanic, an interpretation that is consistent with their morphologies and with their intimate spatial relationship to alteration phenomenon in the Edwards Lake formation both regionally and at an outcrop scale.

The large outcrop at this stop was mapped in detail (Fig. 51) during the summer of 1990 by James Scoates. The complex interplay between various dyke lithologies and between the dykes and the hydrothermal alteration phenomena is described by Galley and Scoates (1990). The least altered volcaniclastic rocks (location A) are examined first, followed with some examples of the different forms of feldspathization/silicification (locations C-D) that are well exposed on this large outcrop. The examples have been chosen to illustrate the character of the alteration, but there is no reason to restrict your attention to these examples.

Location A: Mafic heterolithologic boulder breccia and synvolcanic dykes

In general the level of alteration is less pronounced at the northwest end of the outcrop area, such that the original character of the volcaniclastic rocks is apparent. At location A relatively unaltered, 17 m thick, heterolithologic, mafic boulder breccia bed is

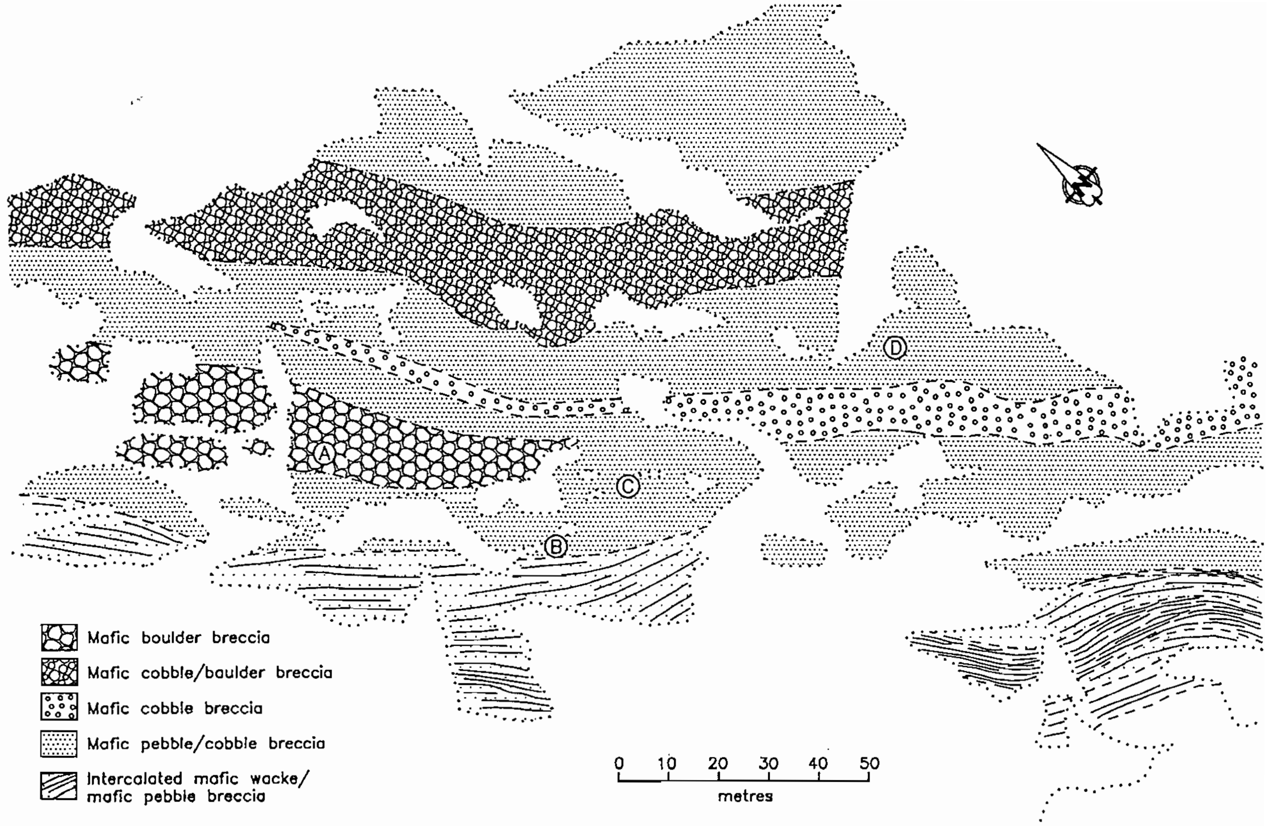


Figure 51a: Distribution of Edwards Lake volcaniclastic units at Stop 22 after removal of intrusive rocks. Modified from Galley and Scoates (1990).

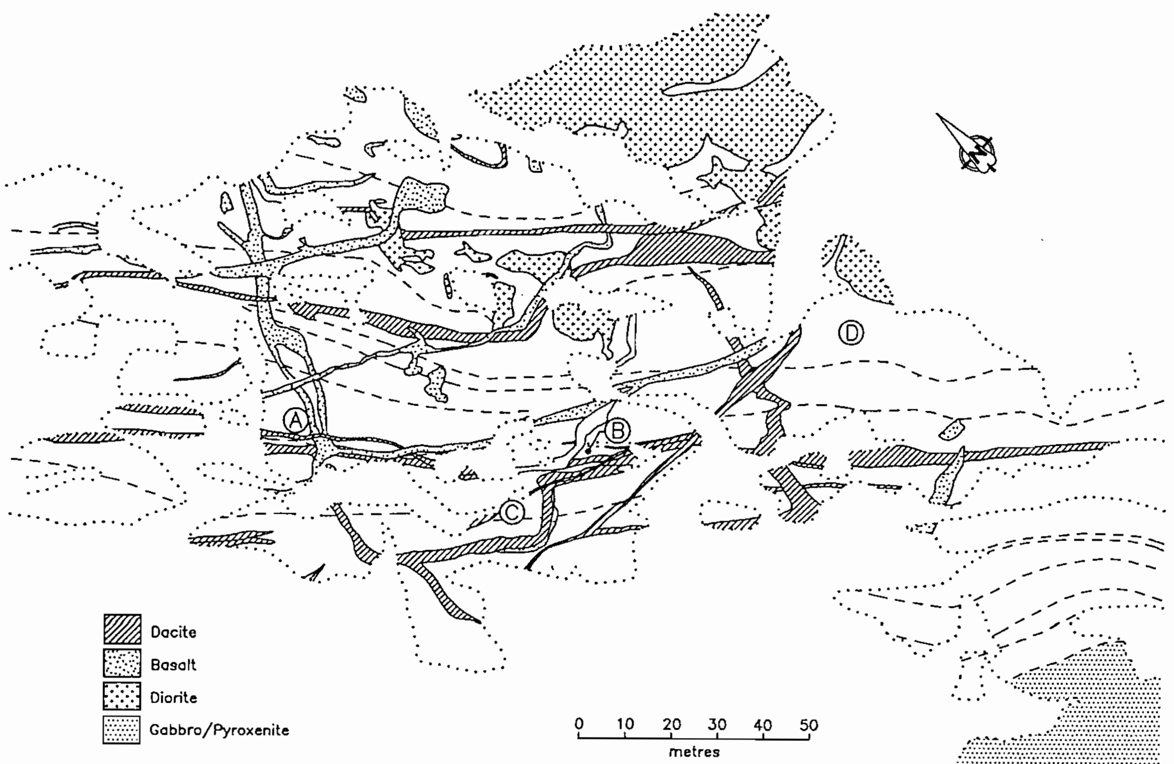


Figure 51b: Distribution of intrusive rocks at Stop 22. The dacite, basalt and diorite dykes are part of a subvolcanic dyke swarm intruded during the Evolved arc episode. Modified from Galley and Scoates (1990).

exposed. The breccia is composed of framework supported subangular to subround boulders up to 1.5 m in diameter. Most of the large boulders are porphyritic mafic blocks.

The boulder breccia is intruded by a dark grey green weathering mafic (basalt?) dyke that displays amoeboid margins and thermal contraction cracks, features that are consistent with intrusion of this dyke into the breccia while it was still unconsolidated and water-saturated. The mafic dyke both cuts and is cut by dacite dykes, indicating a complex chronology of dyke emplacement.

Location B: Mottled feldspathization/silicification spatially associated with a mafic dyke

At this location >60% of the outcrop is composed of various mafic to felsic synvolcanic dykes. Both the dykes and the host mafic pebble wackes display considerable alteration. The relationship between the alteration and the dykes is most evident where a narrow mafic dyke intruding well bedded mafic wacke displays an adjacent 1 to 3 cm zone of pervasive felspathization/silicification. In addition the adjacent mafic wackes are characterized by numerous small oval domains of fine grained quartz and plagioclase up to 50 cm from the dyke margin.

Feldspathization/silicification adjacent to the mafic dyke is interpreted to be due to decreased silica solubility caused by rising temperature generated by dyke intrusion, identical to the mechanism described for this type of alteration adjacent to the dacite dykes at Stop 19.

Location C: Orbicular quartz-plagioclase rich domains overprinting both matrix and fragments

Orbicular quartz-plagioclase alteration structures overgrow both fragments and interfragment areas at this location indicating that alteration took place after deposition of the breccia. The orbicular alteration structures occur in two size ranges, 0.2 to 0.4 cm and 1 to 4 cm, but are otherwise identical. The small and large alteration structures, which occur together, are developed in irregular domains that cut across fragment boundaries.

Location D: Zoned alteration in breccia fragments

Pervasively altered breccia fragments at this location display a wide variety of compositional zoning. Fragments typically have an epidosite-rich core surrounded by a quartz-feldspar-rich rim, but some fragments show a more complex zoning that can include actinolite rich cores, actinolite-garnet rich margins or, in some instances, both. In one fragment an amphibole-rich zone overprints earlier formed silica-rich and epidote-rich zones suggesting a complex interplay between the altering hydrothermal fluids and individual fragments. The breccia is cut by altered fractures filled and replaced by garnet-actinolite-chlorite.

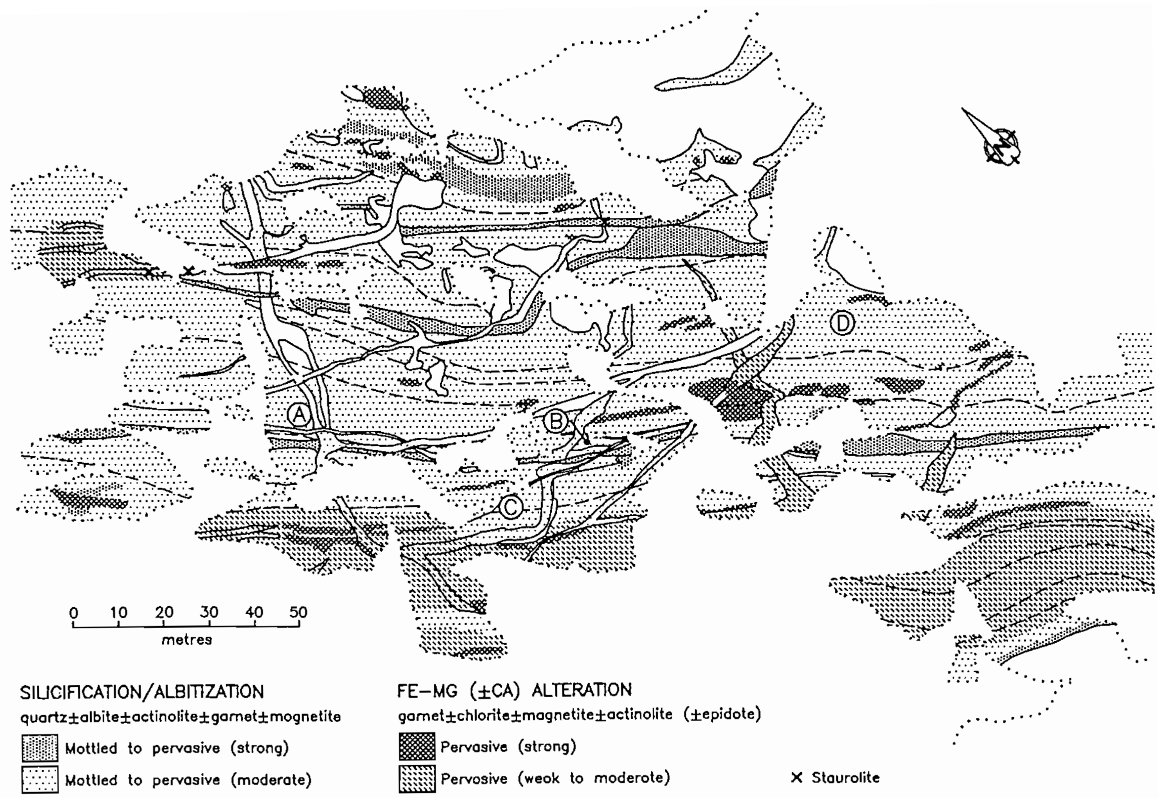


Figure 5Ic: Distribution of hydrothermally altered rocks in Edwards Lake heterolithologic mafic breccia at Stop 22. Modified from Galley and Scoates (1990).

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