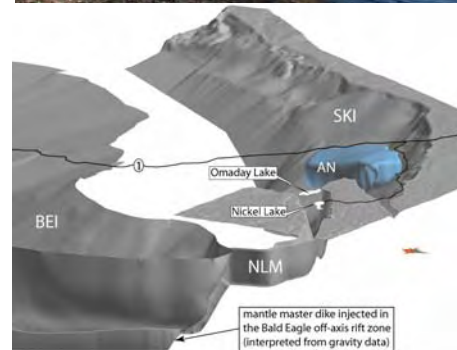


53RD ANNUAL MEETING



INSTITUTE ON LAKE SUPERIOR GEOLOGY

LUTSEN, MINNESOTA MAY 8 – 13, 2007



**PROCEEDINGS VOLUME 53
PART 2 – FIELD TRIP GUIDEBOOK**



INSTITUTE ON LAKE SUPERIOR GEOLOGY

LUTSEN, MINNESOTA MAY 8 – 13, 2007

HOSTED BY:

LAUREL G. WOODRUFF, U.S. GEOLOGICAL SURVEY
MEETING CHAIRPERSON

JAMES D. MILLER, JR., MINNESOTA GEOLOGICAL SURVEY
FIELD TRIP COORDINATOR

PROCEEDINGS VOLUME 53

Part 2 Field Guidebook

Compiled by James D. Miller, Jr. (MGS) and Dean M. Peterson (NRRI)

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53rd Annual Institute on Lake Superior Geology

FIELD TRIP 1

IGNEOUS STRATIGRAPHY OF THE POPLAR LAKE INTRUSION (formerly Nathan's Layered Series)

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Introduction

The Duluth Complex is the largest exposed intrusive component of the Mesoproterozoic Midcontinent Rift, covering an area of over 5,000 km² in northeastern Minnesota (Fig. 1-1). The Complex comprises a more or less continuous mass of mafic to felsic plutonic rocks that extends in an arcuate fashion from Duluth to nearly Grand Portage. These rocks were emplaced as sheet-like intrusions into the lower section of a comagmatic volcanic edifice (the North Shore Volcanic Group), which itself was erupted onto a peneplained surface composed of Paleoproterozoic sedimentary rocks and Neoarchean granite-greenstone assemblages. The Duluth Complex is generally subdivided into four major series on the basis of intrusive age, dominant lithology, internal structure, and structural position within the complex (Miller and others, 2002). In order of younging, these are:

Felsic Series – massive granophyric granite and smaller amounts of intermediate rock that occurs as a semicontinuous mass of intrusions strung along the eastern and central roof zone of the complex and was emplaced during early stage magmatism (~1108 Ma).

Early Gabbro Series – layered sequences of dominantly gabbroic cumulates that occurs along the northeastern contact of the Duluth Complex and also was emplaced during early stage magmatism (~1108 Ma)

Anorthositic Series – a structurally complex suite of foliated, but rarely layered, plagioclase-rich gabbroic cumulates that was emplaced throughout the complex during main stage magmatism (~1099 Ma).

Layered Series – a suite of stratiform troctolitic to ferrogabbroic cumulates that comprise at least 11 variably-differentiated mafic layered intrusions and occurs mostly along the base of the Duluth Complex. These intrusions were emplaced during main stage magmatism, but generally after the anorthositic series.

The Poplar Lake Intrusion, which is the focus of this field trip, and the Crocodile Lake intrusion to the east comprise the only two known intrusions in the Early Gabbro Series. The early emplacement of these intrusions was first implied by cross-cutting field relations mapped by Nathan (1969), who recognized that the Poplar Lake intrusion units are cut on their west end by rocks now recognized as belonging to the Layered and Anorthositic Series. An early intrusive age was also implied by their reversed magnetic polarity (Beck, 1970), which is similar to the lowermost lavas of the North Shore Volcanic Group (Green, 1972). U-Pb ages of zircons from the base of the Poplar Lake intrusion confirm its early emplacement by yielding an age of 1106.9±0.6 Ma (Paces and Miller, 1993). This age is similar to lower reversed-polarity lavas of the Midcontinent Rift (1109-1107 Ma, Davis and Green, 1997) and to the reversed polarity Logan Sills (1108(+4/-2) Ma; Davis and Sutcliffe, 1985), which occur in the footwall of the Poplar Lake Intrusion (Fig. 1-1).

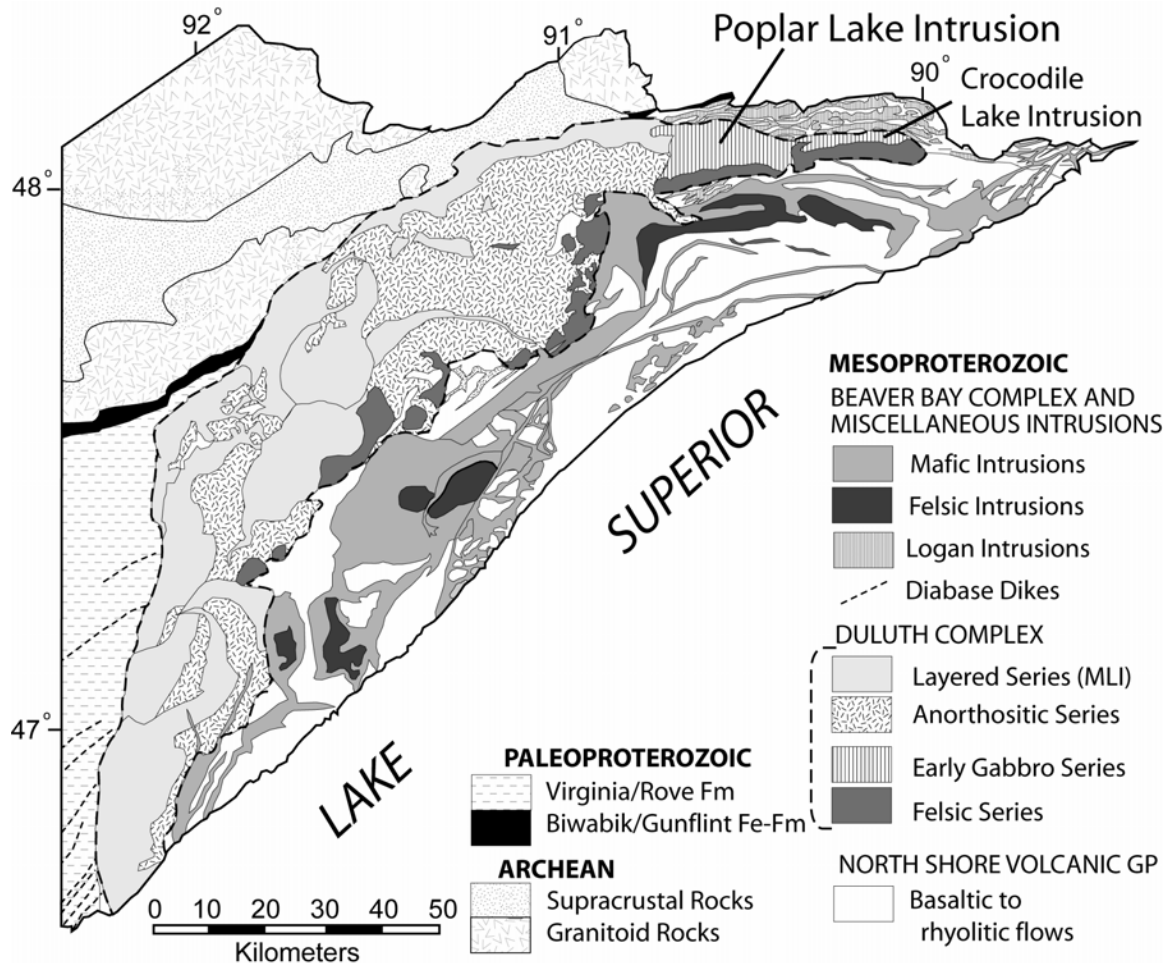


Figure 1-1. Generalized geology of northeastern Minnesota showing the locations of the Poplar Lake and Crocodile Lake intrusions, which comprise the Early Gabbro Series of the Duluth Complex.

Previous Studies

The geological survey of Minnesota (1872-1901), directed by N.H. Winchell and assisted by U.S. Grant, H.V. Winchell, and A. H. Elftman, visited the Gunflint Trail area many times over the latter years of the survey. A geologic map of Cook County (Fig. 1-2), as well as more detailed (1"= 2mi) plates of the Gunflint Lake and Rove Lake areas were published in volume 4 of the Final Report (Winchell, 1899). These maps do not distinguish the various types of gabbro, but they do accurately portray the basal contact of the gabbro with the Rove formation and gabbro-granophyre contact just south of Winchell Lake.

Frank Grout and coworkers conducted reconnaissance field mapping in Cook County in a period between 1913 and 1953. This work was compiled soon after Grout's death in 1958 in MGS Bulletin 39 (Grout, Sharp, and Schwartz, 1959). In township maps of the Poplar Lake area (Fig. 1-3), Grout distinguished gabbro, oxide-rich gabbro, anorthositic gabbro, intermediate

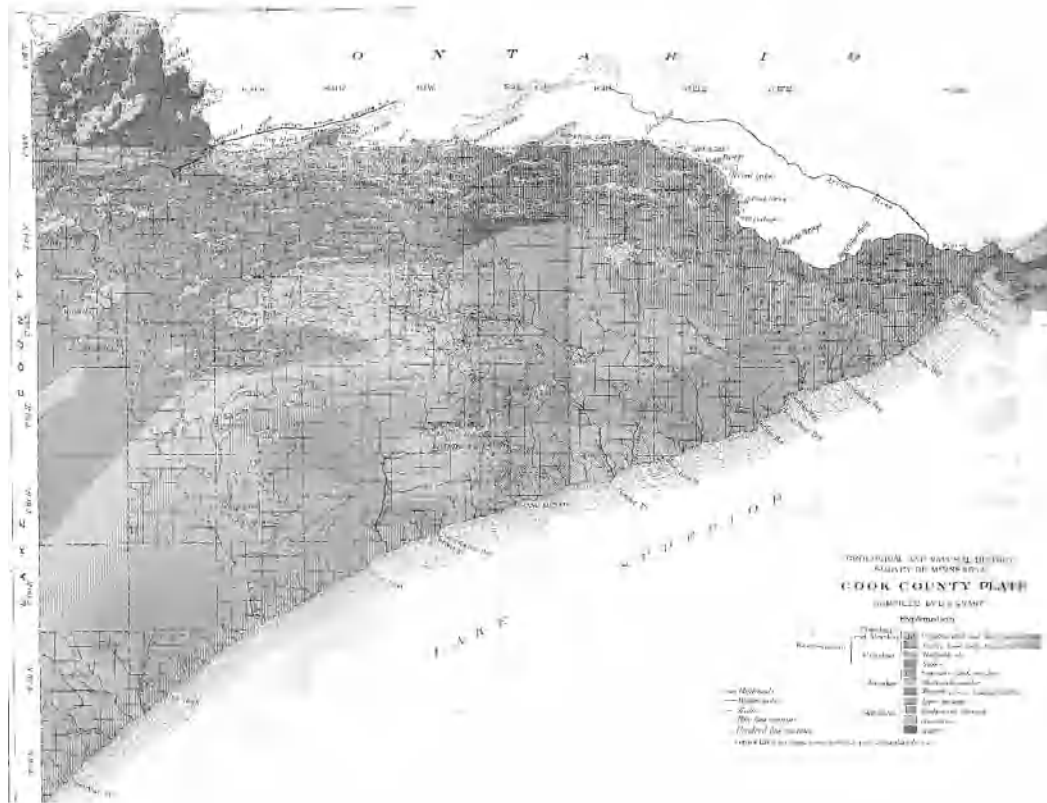


Figure 1-2. Geologic map of Cook County by U.S. Grant from the Minnesota Geological Survey Final Report vol. 4 (Winchell, 1899).

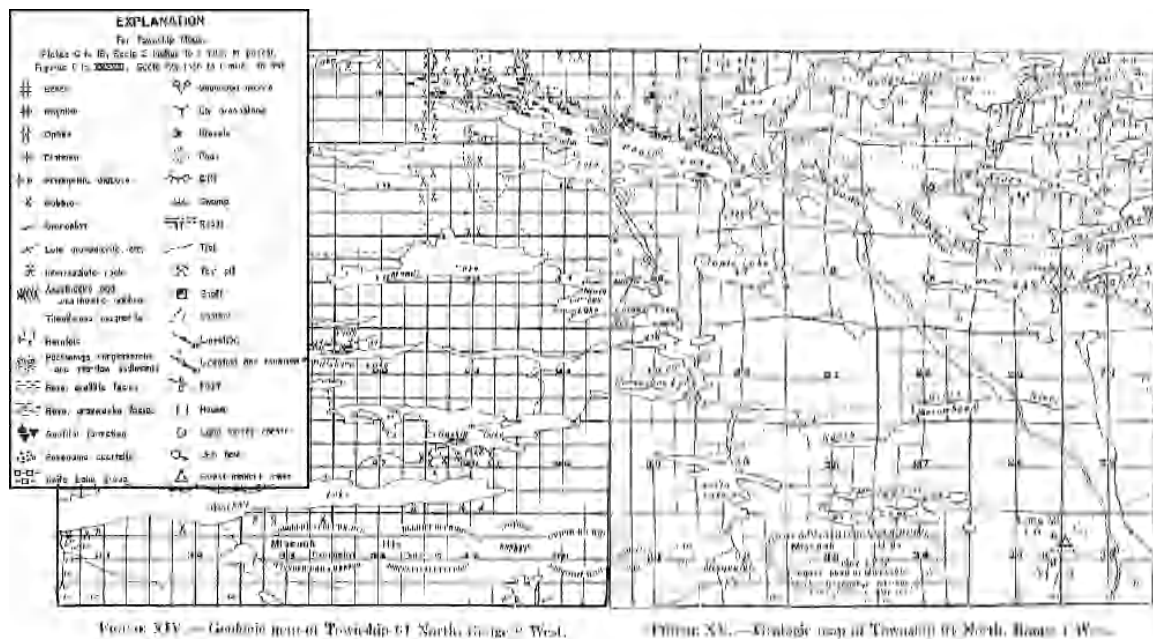


Figure 1-3. Reconnaissance township geologic maps of the Poplar Lake area from Grout, Sharp, and Schwartz (1959)

intrusive rocks and granophyre and noted the moderate southerly dip of layering and foliation. The oxide-rich intervals of the Poplar Lake intrusion were previously cited by Broderick (1917) and by Grout (1950) as potential titanium and iron ore resource.

By far the most complete study of the Poplar Lake intrusion comes from the PhD dissertation work of Harold Nathan (1969). Nathan mapped the bedrock geology of the Duluth Complex in the Gunflint Lake, South Lake, and Hungry Jack Lake and followed this up with a very thorough petrographic study. Reconnaissance geologic maps of these quadrangles were published nearly a decade later (Morey and Nathan, 1977; Mathez, Nathan, and Morey, 1977; and Morey and Nathan, 1978). Phinney (1972) briefly summarized Nathan's work for the MGS Centennial Volume in a section describing the northern prong of the Duluth Complex (Figure 1-4). In that same volume, Davidson (1972) first adopts the name "Nathan's layered series". Later, Weiblen and Morey (1980) referred to the intrusion as the "Layered Series of Nathan" and Weiblen (1982) used the term "Nathan's Layered Series". In the new geologic map compilation of the Duluth Complex (Miller and others, 2001) and the companion report (Miller and others, 2002), the informal name of Nathan's Layered Series was replaced by the Poplar Lake intrusion to acknowledge the location where most of the lithologies recognized by Nathan (1969) are exposed and accessible.

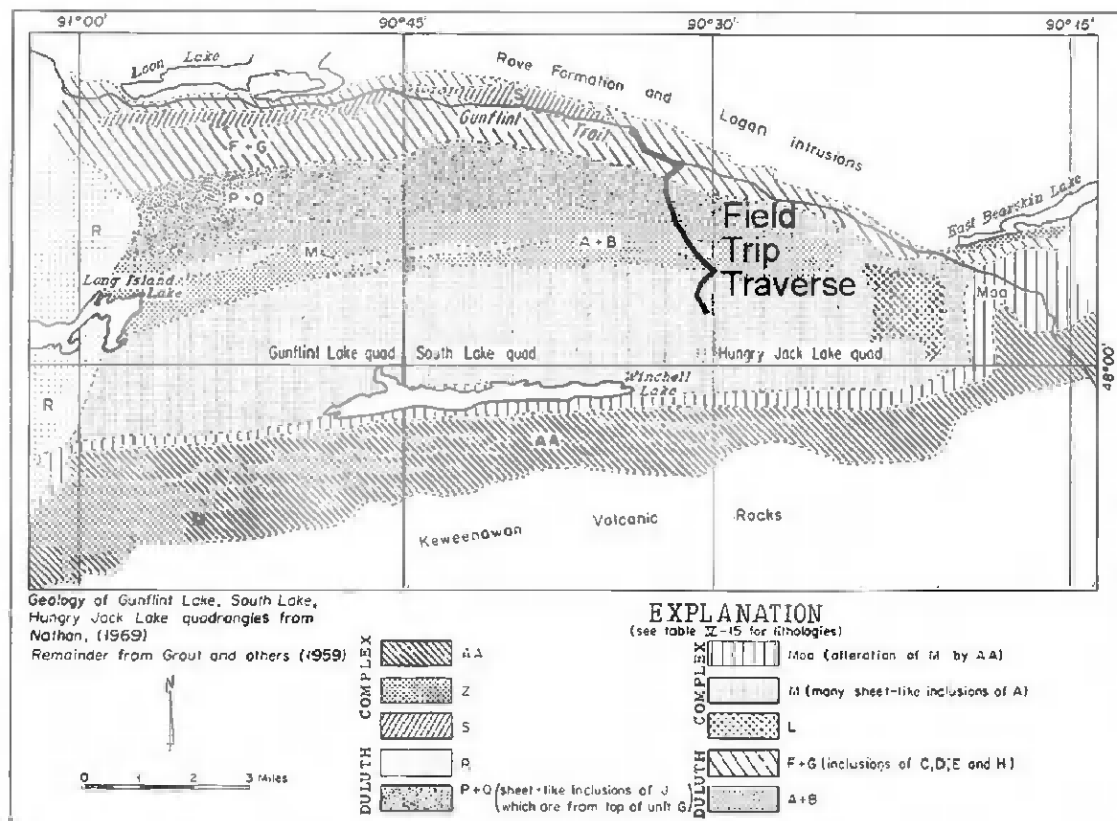


Figure 1-4. Generalized geology of the Poplar Lake intrusion as portrayed by Phinney (1972) based on mapping by Nathan (1969) and Davidson (1972).

Except for limited reconnaissance mapping and geochemical sampling by Jerde (2001), little new mapping or petrologic research has been devoted to this very interesting early gabbroic intrusion of the Duluth Complex. Therefore, the igneous stratigraphy and unit descriptions given for this field trip through the lower half of the intrusion will largely follow Nathan's (1969) conventions. One notable exception is that whereas Nathan (1969) recommended naming rocks by listing the minerals present

in the rock in decreasing order of abundance (e.g., a plagioclase-augite-olivine rock), we have adopted more conventional modal rock name nomenclature (e.g., an olivine gabbro; see Fig. 1-5 for the modal classification scheme used here). Also, we will point out where our petrographic analysis of samples from certain field stop locations differ from Nathan's descriptions.

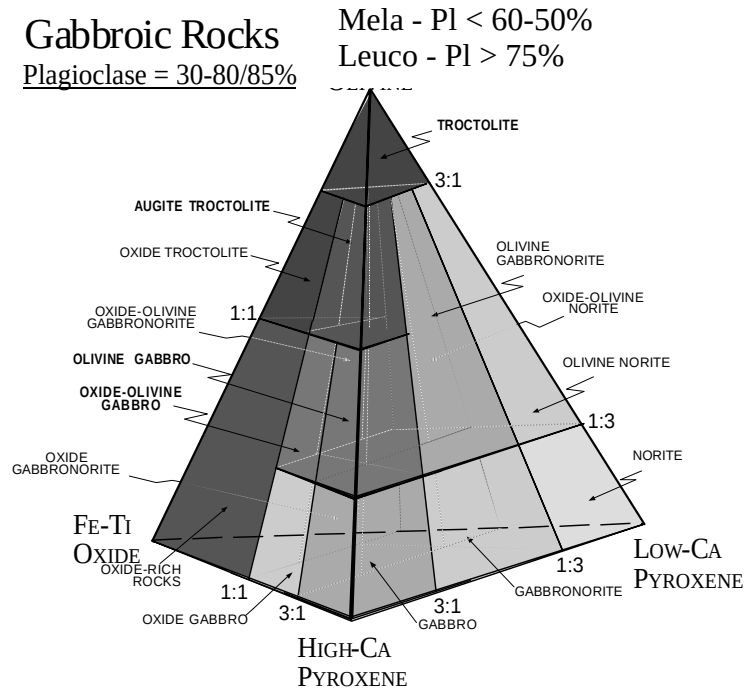


Figure 1-5. Modal classification for gabbroic rocks (Pl 30-80/85%) used in this guide. Prefixes mela- or leuco- (e.g. melatroctolite, olivine leucogabbro) are used when plagioclase abundance is <60 to 65% or >75%, respectively. Taken from Miller and others, 2002.

Igneous Stratigraphy and Petrologic Interpretations of the Poplar Lake Intrusion

Nathan (1969) distinguished 27 different units in his mapping of the three quadrangles that include the Poplar Lake intrusion (Fig. 1-4). He labeled these alphabetically from A through AA in order of decreasing age. Nathan interpreted the relative ages of the units on the basis of fine-grained margins, cross-cutting intrusive relationships, inclusions, and/or thermal effects.

The various map units Nathan (1969) identified within the Poplar Lake intrusion can be generally grouped into 6 major lithologies:

- 1) oxide-rich, gabbro to olivine gabbro: F, G, H, T, U, V
- 2) olivine gabbro to troctolite: P, Q
- 3) gabbro to olivine gabbro: A, B, M
- 4) anorthositic rock types: J, some M, some Q, S
- 5) intermediate to felsic rock types: K, O, AA, Maa
- 6) fine-grained mafic rocks (inclusions?): C, D, E, N, X, Y

Davidson conducted reconnaissance mapping in quadrangles to the south of the three that Nathan (1969) mapped and had some different interpretations of the upper units of the Poplar Lake Intrusion (Davidson, 1972, 1977a, 1977b; Davidson and Burnell, 1977). Davidson (1972) interpreted the extensive mass of granophyre at the top of the Poplar Lake intrusion, which Nathan (1969) designated as unit AA (Fig. 1-4), to be a separate intrusion that belongs to the felsic series of the Duluth Complex. Like Nathan (1969), he interpreted the massive granophyre to be one of the youngest intrusion of the Duluth Complex based on observations that granophyre dikes and irregular masses commonly cut most other rock types. Nathan's unit Maa, which underlies the granophyre mass and is composed of a highly altered assemblage of intermediate rock types, is interpreted to be an altered and contaminated phase of unit M caused by the late intrusion of granophyre (unit AA) (Fig. 1-4). Davidson (1972, 1977a, 1977b; Davidson and Burnell, 1977) interpreted these intermediate rocks (unit grd) to be marginal phases of the granophyre. Another reinterpretation of Nathan's (1969) mapping by Davidson (1972) and others (Morey and Nathan, 1978; Morey and others, 1981) is that Nathan's unit R is actually part of the Tuscarora Intrusion and thus belongs to the layered series (or the troctolite-olivine gabbro series as Davidson (1972) called it).

In the most recent map compilation of the Duluth Complex (M-119; Miller and others, 2001), several reinterpretations were made of both Nathan's (1969) and Davidson's (1972) map interpretations. One change is the suggestion that Nathan's unit S, a gabbroic anorthosite that cross cuts lower units of the intrusion, is actually part of the younger anorthositic series. More significantly, whereas both workers considered the granophyre mass, now termed the Misquah Hills granophyre, to be one of the youngest intrusions in the Duluth Complex, it is now considered to be one of the oldest. Major evidence for this comes from U-Pb age dating, which indicates a crystallization age of 1106 ± 6 Ma (Vervoort and Wirth, 2004). Consequently, the intermediate rocks of Nathan's unit Maa and Davidson's unit grd, which underlie the granophyre, are now seen to represent hybridization between partially melted granophyre and underlying mafic magmas of the Poplar Lake intrusion. Miller and others (2002) suggested that the granophyre acted as a low-density trap, which caused the Poplar Lake intrusion to underplate and partially melt the felsic mass.

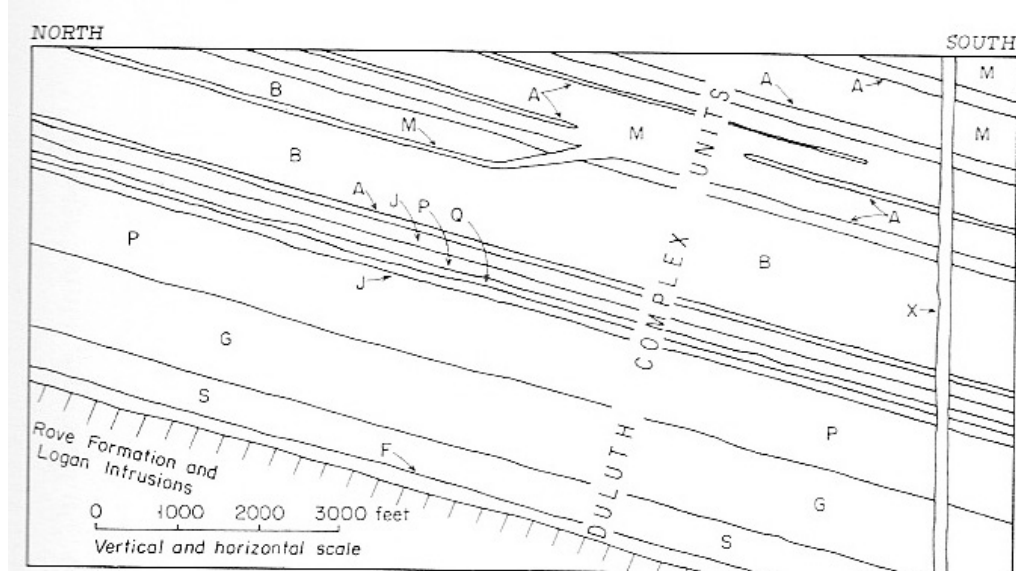


Figure 1-6. Diagrammatic cross-section through the central part of the Poplar Lake intrusion. Figure from Phinney (1972, Fig. V-33) based on data from Nathan (1969).

Examining the stacking of Nathan's (1969) units, it is immediately clear that he interpreted the emplacement of the Poplar Lake intrusion to be a complex multistage process (Figs. 1-6 and 1.7). He envisioned various sill-like intrusions to have been multiply emplaced at seemingly random levels in the intrusion. Nathan (1969) interpreted most intrusive units to have crystallized in place with little evidence of fractional crystallization. This rather unconventional interpretation of the igneous stratigraphy of a mafic layered intrusion compels some skepticism as to Nathan's model. For example, the sulfide-mineralized heterogeneous basal unit (F), commonly the earliest formed rock type in other Duluth Complex layered intrusions, is thought to have been emplaced after considerable volumes of A and B were emplaced. Many of the fine-grained gabbro units have limited areal extent and appear similar to hornfels basalt inclusions commonly found throughout other Duluth Complex intrusions. Moreover, the anorthositic units are mostly seen as late, however, anorthositic rock are typically some of the earliest intrusive phases throughout most of the Duluth Complex. Nevertheless, until a rigorous follow-up field and petrologic of this intrusion is conducted, Nathan's multiple emplacement hypothesis stands as the current model.

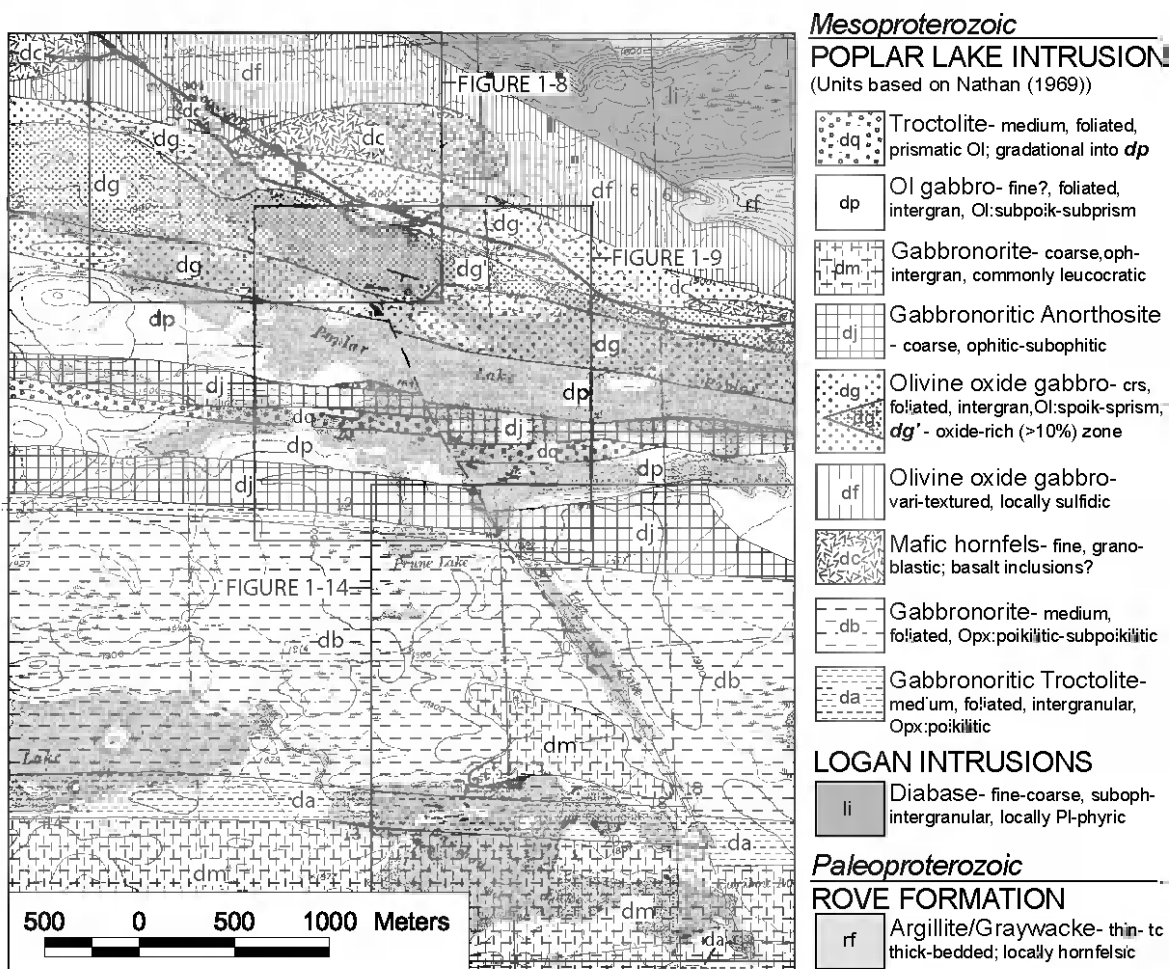


Figure 1-7. Geologic map of the field trip area based on Morey and Nathan (1977). Unit labels after Nathan (1969), but standard modal rock names (see Fig. 1-5) interpreted from Morey and Nathan's (1977) descriptions. Locations of more detailed maps of stop locations (Figs. 1-8, 1-9 and 1-14) are shown by inset boxes.

FIELD TRIP 1- DAY 1

Field Stop Descriptions

Upon arriving at Rockwood Lodge and unpacking gear, we will set out to investigate roadcuts along a one-mile stretch of the Gunflint Trail for the rest of the morning.

STOP 1-1. BASAL MARGIN UNITS – F, C, G

Location: Roadcuts along Gunflint Trail between entrance and exit of Cook County Road 82. South Lake 7.5' quadrangle (T64N, R2W, Sec. 1).

UTMs: start – 683540E 5326180N; end – 684950E 5325380N

Description: Roadcuts along this section of the Gunflint Trail expose some of the rock types composing the marginal zone of Nathan's layered series (Fig. 1-8). Beginning from the junction of the Gunflint Trail with County Road 92, we will proceed east along the north side of the Trail back toward the Rockwood Lodge entrance.

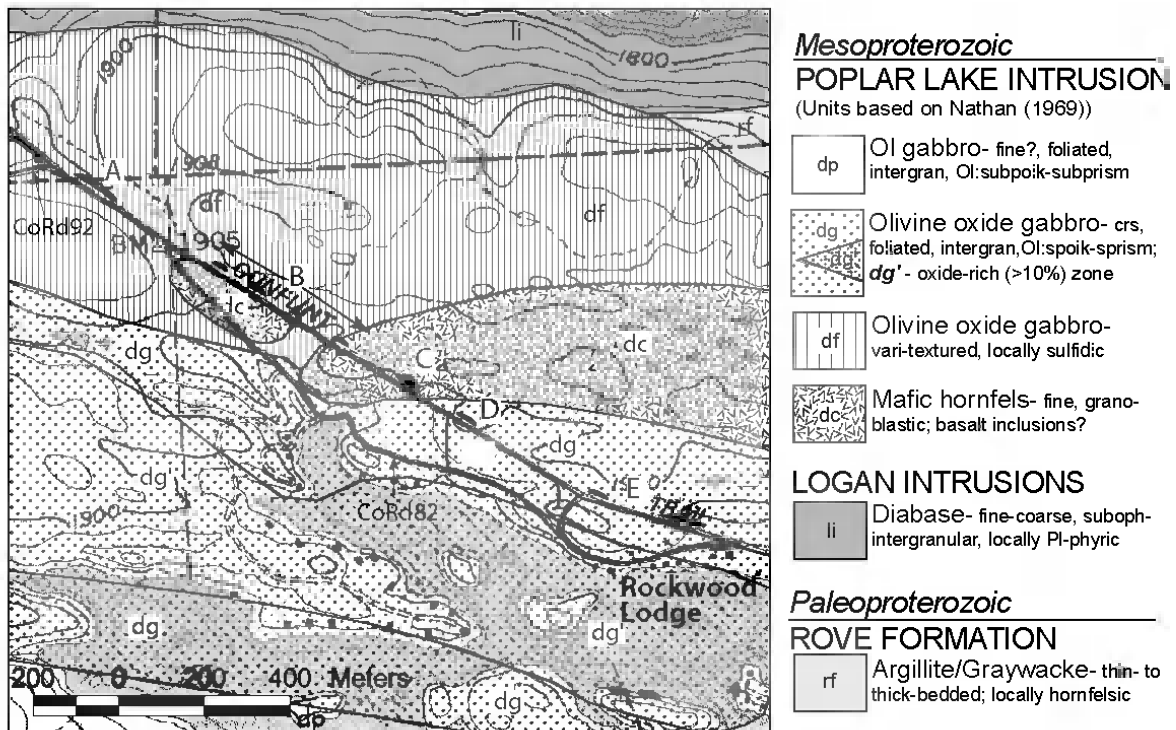


Figure 1-8. Geology of the Gunflint Trail area northwest of the Rockwood Lodge (based on Morey and Nathan, 1977). Roadcuts A – E are described in the text for Stop 1-1.

Roadcuts A – These exposures show a more homogeneous and sulfide poor phase of Unit F. The rock is a coarse-grained ophitic oxide olivine gabbro. Similar exposures occur in the roadcuts on the south side of the Trail, west Co Rd 92 junction. This is the location of geochronology sample NLS-5, which yielded a U-Pb zircon age of 1106.9 ± 0.8 Ma (Paces and Miller, 1993).

Roadcuts B – This series of a low, deeply weathered roadcuts and low pavements are more typical of unit F. About 150 m from the junction is of coarse-grained, decussate, ophitic, biotitic oxide-olivine gabbro. This noncumulate, vari-textured (taxitic) gabbro forms the lowermost unit of

Nathan's (1969) layered series (his unit F). Locally, Cu-Fe sulfide is present and gives rise to the very rusty appearance of the outcrop. Sulfide mineralization in the lower contact zone is characteristic of many parts of the Duluth Complex (Bonnichsen, 1972; Miller and others, 2002). As is interpreted from the Cu-Ni-PGE sulfide deposits of the NW Duluth Complex, the probable source of the sulfur are the shale and graywacke of the Paleoproterozoic Rove Formation, which forms the footwall of the gabbro. The lack of significant sulfide mineralization here compared to the NW area may indicate a lower sulfur content to the sediments or depletion of sulfur by earlier intrusion of the Logan Sills.

Roadcut C – At this prominent roadcut, a very fine grained gabbro is exposed. It has irregular areas of coarser gabbro scattered throughout and some pockets of sulfide. This is Nathan's unit C, which he suggested could be recrystallized inclusions of Keweenawan lavas, Logan sills, or more likely, the chilled contact of a precursor intrusion (his units A and B). In thin section, the rock is a strongly recrystallized, poikiloblastic gabbro with small (<5 mm) high-density oikocrysts of augite and hypersthene enclosing granular plagioclase. Because the texture of Logan sills remains unaffected as the contact with the layered series approached (Nathan, 1969), it is unlikely that this body is such an inclusion. Rather, the zones rich in ovoid clots of coarse feldspar are similar in appearance to what have been interpreted elsewhere in the complex as recrystallized amygdulites (Bonnichsen, 1972), and this inclusion is most likely a basalt flow. The abundance of these types of inclusions and the relative lack of quartz- or cordierite-bearing inclusions of Rove Formation suggest that the focus of emplacement of Nathan's layered series was at the horizontal to shallow-dipping discontinuity between Early Proterozoic sedimentary rocks and Keweenawan lava flows.

Roadcuts D and E About 700 m farther east past a bend in the road, another deeply weathered roadcut exposes a medium- to coarse-grained, moderately laminated oxide troctolite with ophitic augite. This belongs to Nathan's unit G which he characterized as "coarse-grained, olivine-plagioclase and augite-plagioclase rocks with strongly foliated plagioclase and abundant iron oxides (Fe-Ti oxides) . . . biotite is conspicuous . . . density graded layering is prominent" (Nathan, 1969, p. 68). Although they form structurally lowest cumulates in the layered series, Nathan interpreted this unit to be younger than structurally higher troctolitic to gabbroic units on the basis of a complex sequence of discordant relationships (Units A and B, Fig. 3.21). The validity of Nathan's intrusive stratigraphy clearly needs to be tested.

--Return to Rockwood Lodge for lunch and prepare for afternoon canoe trip.--

WARNING: The ice just went out (we hope) on the lake and so the water is just above freezing. We will take some time over the lunch hour to instruct everyone on basic canoe safety and what to do in the event that someone dumps in the water. We have BWCA-experienced assistants on the trip and we will bring special equipment to guard against hypothermia. Despite these precautions, the best preventative is using extreme care and common sense.

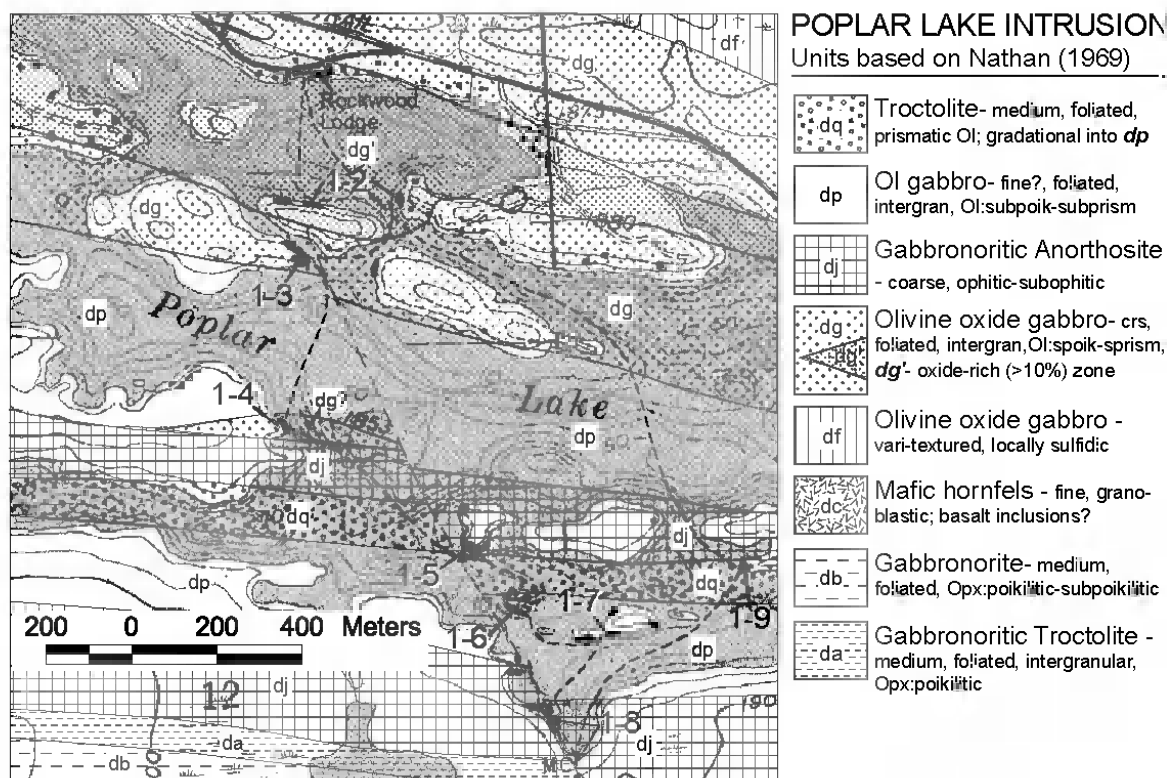


Figure 1-9. Geologic map of the central Poplar Lake area showing field stop locations for the afternoon of Day 1. Dashed line shows canoe route. Geology after Morey and Nathan (1977).

STOP 1-2. UNIT G' – OLIVINE OXIDE GABBRONORITE

Location: Float by shoreline ledges on island about 300m due south of Rockwood Lodge.

UTMs: 684800E 5325000N

Description: These lichen-covered, grussy outcrop ledges are composed of medium-grained, well-foliated, oxide gabbro typical of Unit G', an oxide-rich (locally >20%) interval within Unit G. Paddling to the east, a prominent outcrop can be seen at the western point of a peninsula. The well-developed foliation gives rise to a sheet jointing that dips about 20° to the south.

--Canoe around the east end of this island then south into a narrow channel. Head WSW to eastern point of another island which has a sizeable dock for disembarking. --

STOP 1-3. UNIT G – OXIDE OLIVINE GABBRONORITE

Location: Private residence and dock at east point of island; please request permission from owners to view exposure.

UTMs: 684905E 5325355N

Description: The main rock type exposed over this outcrop point is a medium coarse-grained, foliated, intergranular, oxide olivine gabbronorite. This exposure is typical of Unit G. One curious feature is the occurrence of a 20- to 30-cm-thick, fine grained sill of a similar mineralogy

and texture exposed in the cliff face back from the boat dock. The lower contact is rather sharp but the upper contact is gradational (Fig. 1-10A). Throughout the sill and especially near the upper contact, coarse crystals and crystal clusters of pyroxene and plagioclase occur (Fig. 1-10B).

Petrographic observations across this sill show that it differs from the coarser host in several subtle ways. Olivine in the fine-grained gabbro is commonly subpoikilitic compared to granular olivine in the host gabbro. Low-Ca pyroxene is less abundant in the fine-grained gabbro (5% compared to 20%) and is dominantly hypersthene rather than inverted pigeonite. Perhaps this represents a new pulse of magma into the chamber or it marks venting event from the magma chamber causing decompression quenching of a water-saturated evolved magma. The presence of up to 2% biotite and the generally similar mineralogy and texture of the fine-grained and medium-grained gabbros suggest the latter explanation is more likely.

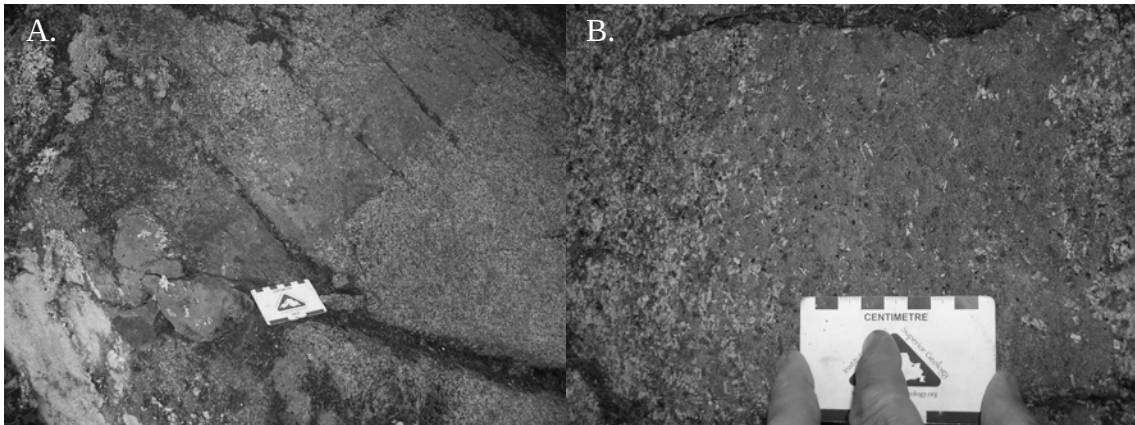


Figure 1-10. A) Fine-grained gabbro sill in medium-grained gabbro host; B) Close-up of piece of float from the sill show dispersed coarse plagioclase and pyroxenes in the fine matrix.

--Canoe about 400m due south to northernmost point of eastern end of peninsula. Limited landing spots, so this stop is optional,--

STOP 1-4. UNIT G OR P?- LAYERED OLIVINE GABBRO (Optional)

Location: Point of land in southwestern Poplar Lake near the far north central part of Sec. 12.
UTMs: 684785E 5324469N

Description: Nathan implies by his map and unit designation that this as a thin inclusion? of unit G enclosed between younger units P and J (Fig. 1-9). The exposure here is a medium- to medium fine-grained, moderately layered, well-foliated, intergranular oxide olivine gabbro. A thin section sample from here shows a medium fine-grained oxide olivine gabbro that has a very granular (granoblastic?) texture and is lacking in low-Ca pyroxene. This rock has similarities to Nathan's Unit P by being fine, olivine-rich, and oxide poor. The lack of Opx makes it distinct from either Unit P (commonly contains some poikilitic hypersthene) and especially Unit G (typically contains significant amounts of inverted pigeonite).

--Canoe about 550m to the east-southeast to island. Land canoes against outcrops on the southeast and southwest sides of the island.,--

STOP 1-5. UNIT J – OLIVINE OXIDE LEUCOGABBRO

Location: Island in southern part of Poplar Lake.

UTMs: 685210E 5324200N

Description: Exposures on the south side of the island are composed of coarse-grained, foliated, ophitic to intergranular olivine oxide leucogabbro, which Nathan classifies as Unit J. Plagioclase ranges from 70 to 85% of the rock and is the only consistently cumulus phase. Augite and Fe-Ti oxide are the dominant mafic minerals and vary from anhedral granular to poikilitic. Minor olivine (3-5%) is also variable in texture from subprismatic (Fig. 1-11A) to subpoikilitic. Layers and pods rich in granular Fe-Ti oxides and pyroxene are evident on the southwestern corner of the island (Fig. 1-11B). In poor exposures on the north side of the island (downsection), granular (cumulus) augite, olivine, and oxide are present and the rock seems less leucocratic.

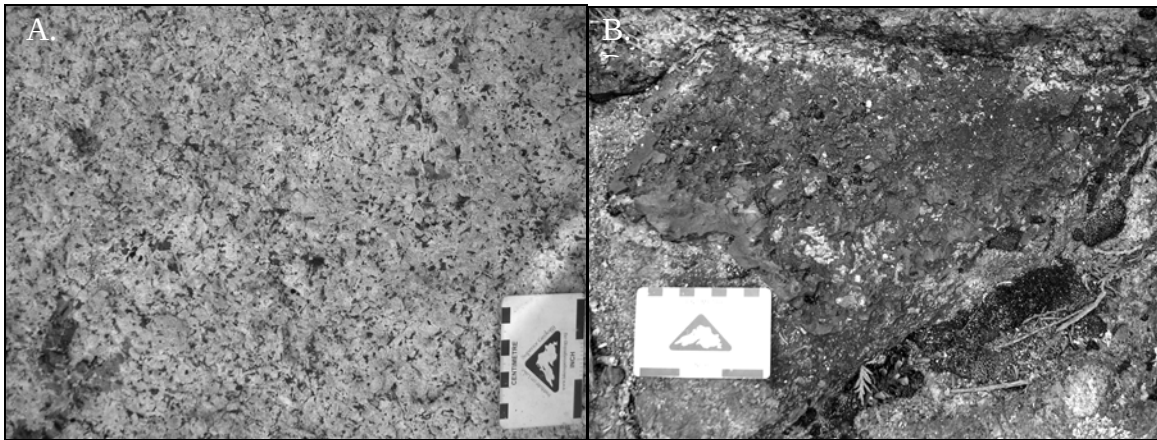


Figure 1-11. A) Common texture of leucogabbro with granular mafic phases (Ol, Aug, FeOx). B) Massive Fe-Ti oxide layer in leucogabbro.

-- Paddle ~175m to the southeast to small island with campsite. Land canoes against outcrops on the south side or in cove on east side of the island. Welcome to Happy Island!--

STOP 1-6. UNIT P – OLIVINE GABBRO

Location: Island in southern part of Poplar Lake.

UTMs: 685310E 5324050N

Description: “Happy Island” is composed of two east-west oriented outcrop areas. Exposures on the south side of the island are composed of medium-grained, well-foliated, intergranular olivine gabbro with variable abundances of Fe-Ti oxide (1-7%). Olivine occurs as small (<1cm), subpoikilitic clots or as prisms 0.5 to 3 cm long (~1:10 aspect ratios; Fig. 1-12) that tend to be concentrated in thin layers. This rock type is consistent with Nathan's description of unit P.

Morey and Nathan's (1977) map shows a contact between unit P to the south and unit Q, a troctolite, to the north running through the middle of the island. However, outcrop exposures and petrographic observations of two samples from the northern end show the rock to be mostly an oxide olivine gabbro like Unit P. The only significant differences are a slightly coarser grain size, subtle textural and modal layering, and a slightly higher concentration in inverted pigeonite. Nothing on the north half looks like an augite-poor troctolite, which characterizes Unit Q (See Stop 1-9).

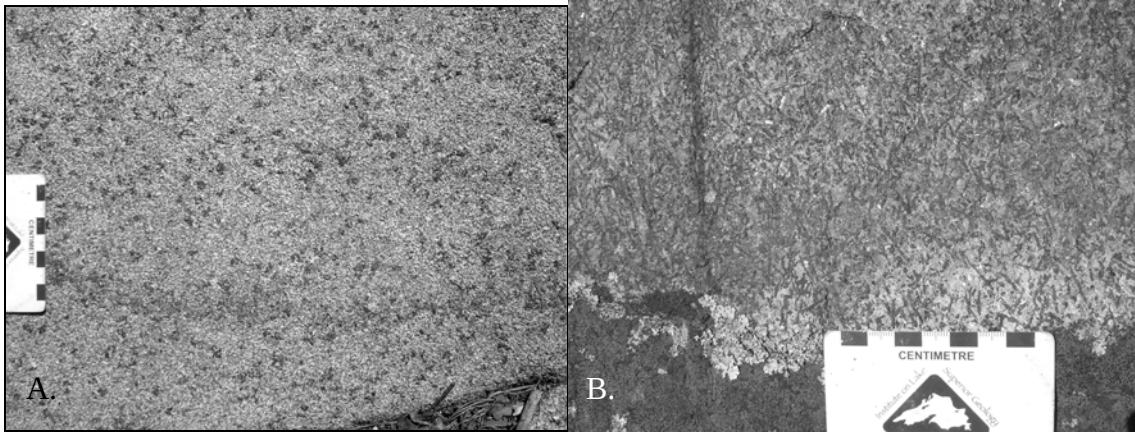


Figure 1-12. Textures of olivine found in Unit P: A) 5mm subpoikilitic clots (Stop 1-7); B) 1cm long prismatic olivine (Stop 1-6)

-- Paddle ~200m to the east-southeast to the south side of an E-W island. Land canoes against outcrops on the south-central of the island. --

STOP 1-7. UNIT P – OLIVINE GABBRO

Location: Island in southern part of Poplar Lake.

UTMs: 685532E 5323990N

Description: Island is composed of two outcrop areas on the south side of the island, between which access is difficult. Both exposure areas are composed of medium-grained, well-laminated, intergranular olivine gabbro (PAOf cumulate) as observed on the south side of the island at Stop 3-3. Locally, 2-5-cm clots of coarse pyroxene and oxide are present and elsewhere are pods (inclusions?) of leucogabbro (Unit J?). Again olivine occurs as small subpoikilitic clots or as prisms up to 3 cm long (Fig. 1-12). Because prismatic olivine tends to be concentrated in thin layers along which joints tend to develop, this texture is preferentially displayed on the outcrop face. Note that on the plane of layering, prismatic olivine is commonly oriented parallel to the strike of layering. This observation and the delicate prismatic habit of the olivine argues against crystal settling by density currents.

-- Canoe ~250m to the southwest to steep cliffs lining the bay leading the portage to Lizz Lake. Float along the cliffs and then land canoes at the portage. --

STOP 1-8. UNIT J –LEUCOGABBRONORITE-GABBROIC ANORTHOSITE

Location: Bay leading to portage (to Lizz Lake) at south shore of Poplar Lake.

UTMs: 685460E 5323700N

Description: Exposed in sheer cliffs along the west side of the bay leading to the portage to Lizz Lake is leucogabbroic rock typical of Unit J. Lichen and moss cover make it difficult to see much of the rock. A thin section of a sample collected here shows the rock to be a medium coarse-grained ophitic leucogabbronorite. In some places, pockmarked surfaces are evident and indicate the presence of large oikocrysts of poikilitic olivine. This texture is very well

displayed in the exposure on the west side of the portage trail (Fig. 1-13). Here the olivine oikocrysts are up to 5 cm across. Ophitic augite is also evident here in oikocrysts up to 10 cm across.

Interestingly, this rock type is very similar to the dominant lithology of the Anorthositic Series of the Duluth Complex. It would worthwhile determining the magnetic polarity of this rock type to see if it has a typical Poplar Lake reversed polarity, or is if is normal like all known Anorthositic Series rocks.



Figure 1-13. Mottled surface of ophitic olivine gabbroic anorthosite at Stop 1-8. Clots are formed from both olivine and augite oikocrysts.

-- Canoe ~600m to the northeast past the eastern point of island where Stop 1-7 is located. Continue northeast to middle point of large island. Not sure how many canoes can land here so this is an optional stop.. --

STOP 1-9. UNIT Q –TROCTOLITE (OPTIONAL)

Location: Middle point of west side of large island in southern Portage Lake. .

UTMs: 685850E 5324150N

Description: According to a thin section sample of this outcrop that shows a medium-grained, well foliated troctolite with small olivine grains, this site is more typical of Nathan's Unit Q.

-- Canoe northwest back to Rockwood Lodge (~1.6 kilometers). End of Day 1.—

FIELD TRIP 1- DAY 2

Depart from Rockwood Lodge and return to portage to Lizz Lake (Stop 1-8). Portage canoes to Lizz Lake landing then double back on portage trail to view first stop of the day.

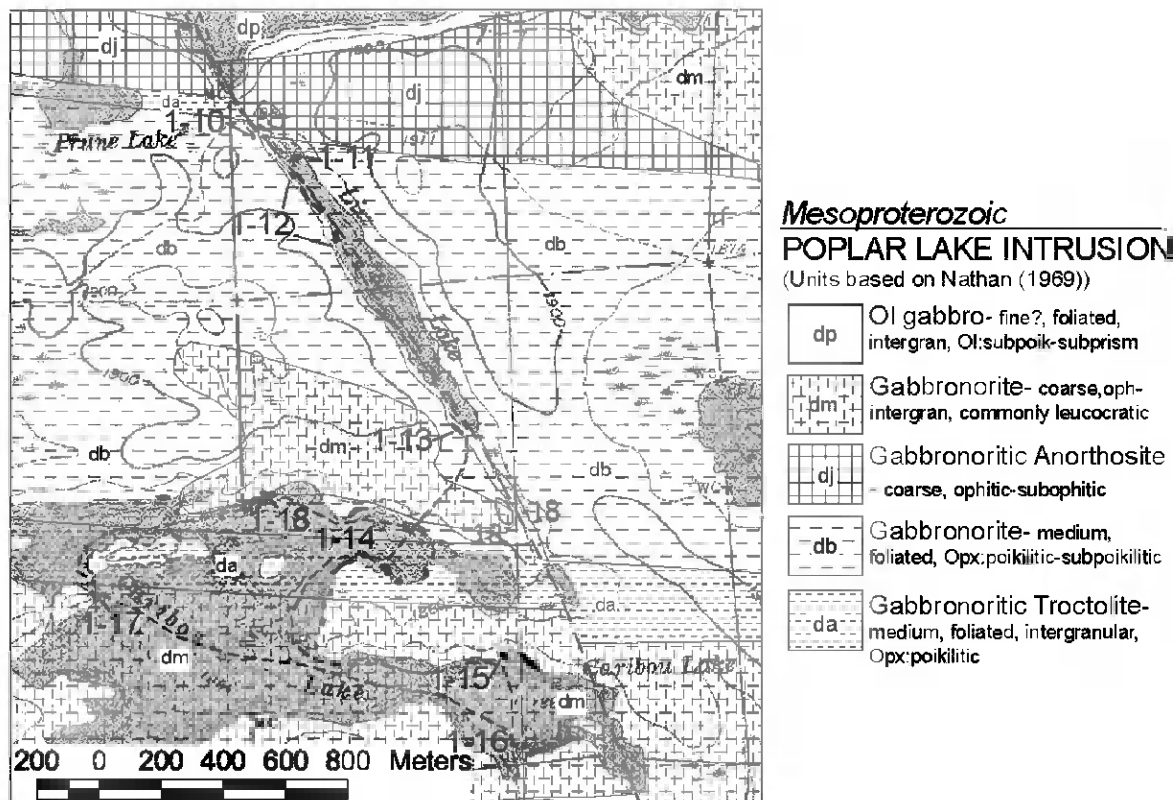


Figure 1-14. Geologic map of the Lizz and Caribou Lakes area showing field stop locations for Day 2. Dashed line shows canoe route. Geology after Morey and Nathan (1977).

STOP 1-10. UNITS J AND B AND THE LIZZ LAKE FAULT

Location: Lizz Lake portage trail just 50m south of junction with Banadad trail.

UTMs: 685500E 5323600N

Description: Exposed on the east side of the trail is a medium coarse-grained ophitic oxide gabbronoritic anorthosite typical of leucocratic Unit J. About 10m to the west of the trail is a poorly exposed ledge of medium-grained, foliated, gabbronorite with poikilitic hypersthene. The rock contains no olivine or Fe-Ti oxide. This rock type is typical of Nathan's Unit B. The juxtaposition of these two distinct rock types across this gap is the best evidence of right lateral offset across a fault that Morey and Nathan (1977) inferred through here and down the elongate trend of Lizz Lake (Fig. 1-14). The Lizz Lake fault is the only fault that Nathan (1969) identified in the interior of the Poplar Lake Intrusion.

-- Canoe a short distance to the outcrop island about 100 m to the southeast of the portage--

STOP 1-11. UNIT J? – GABBRONORITIC ANORTHOSITE

Location: Island with abundant outcrop on north end of Lizz Lake.

UTMs: 685685E 5323490N

Description: Good exposures over most of this island show the rock to be a medium coarse-grained, poorly foliated, ophitic, gabbronoritic anorthosite similar to the Unit J outcrop at Stop 1-10. However, Nathan has this exposure mapped within Unit B. This may indicate that the Unit J-Unit A/B contact is farther south on the east side of the Lizz Lake fault. Alternatively, perhaps this is an inclusion in Unit B though this would violate Nathan's interpretation of Unit B being an early phase of the Poplar Lake intrusion. More mapping in this area is needed to resolve this.

-- Float along a series of cliff faces on the western shore of Lizz Lake to view Stop 12 exposures--

STOP 1-12. UNIT B – GABBRONORITE

Location: Series of outcrop ledges along the northwestern shoreline of Lizz Lake.

UTMs: 685715E 5323345N

Description: Although largely obscured by lichen, the ledges along this shoreline display a homogeneous, medium-grained, well-foliated gabbronorite with $\text{Opx} \geq \text{Cpx}$. Orthopyroxene (hypersthene and inverted pigeonite) is typically subpoikilitic to poikilitic, whereas clinopyroxene (augite) is typically anhedral granular (Fig. 1-15A). In one well exposed location beneath a tree tip-up, a subtle modal layering was recognized (Fig. 1-15B).

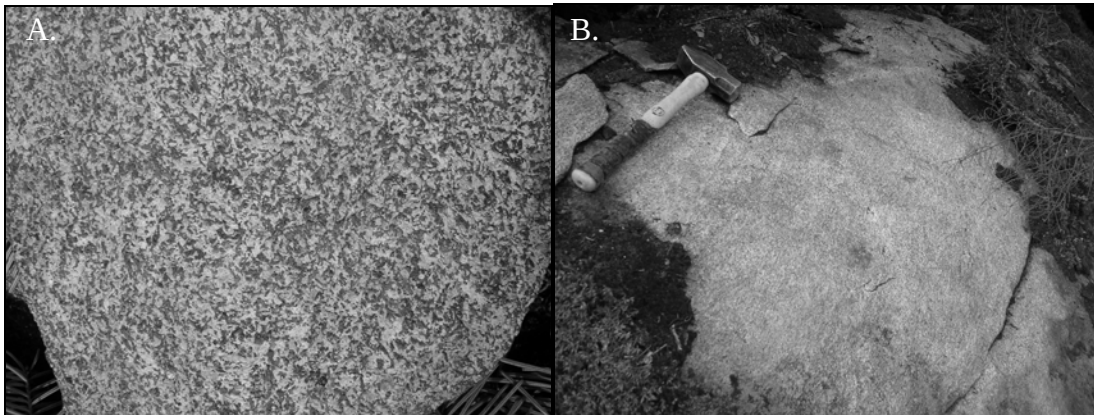


Figure 1-15. A) Texture of the gabbronorite typical of Unit B. B) Subtle modal layering common observed under a tip-up at Stop 1-12.

-- Canoe to the south end of Lizz Lake to portage to Caribou Lake. From the portage landing, head west about 8m to a rock cliff --

STOP 1-13. UNITS B AND X? – GABBRONORITE AND FINE GABBRO

Location: Outcrop ledge southwestern corner of Lizz Lake near portage to Caribou Lake.

UTMs: 686195E 5322605N

Description: Most of this exposure is a very fine-grained mafic rock in sharp contact with a medium-grained gabbro. Thin sections of the fine-grained rock shows it to be a felty-textured, oxide gabbro and the medium-grained gabbro is a well foliated, intergranular gabbro with poikilitic olivine. Nathan mapped this area as largely unit B with an occurrence of Unit X, which he defines as a fine-grained Pl-porphyrific oxide gabbro. By its unit letter designation, Nathan interpreted this as young intrusive unit.

The medium gabbro is similar to typical unit B, except for the poikilitic olivine. Perhaps it is a hybrid between Units B and A. The fine-grained rock being an intrusion into B would be consistent with its felty texture. One would expect the rock to display a granoblastic texture if it was a hornfels inclusion. We will investigate the exposure to ascertain whether the fine-grained rock is an intrusion or an inclusion.

-- Take the portage to Caribou Lake. Canoe 180m south to a campsite fronted by a large outcrop pavement--

STOP 1-14. UNIT A – OLIVINE GABBRONORITE

Location: Outcrop pavement at campsite on south shore of northeast arm of Caribou Lake.

UTMs: 685960E 5322155N

Description: Very homogeneous exposure of medium-grained, foliated, olivine gabbro with poikilitic inverted pigeonite. Some of the inverted pigeonite oikocrysts can be up to 30 cm across.

This exposure is identified as Unit A by Morey and Nathan (1977), which they describe as being a fine-grained troctolite with up to 9% inverted pigeonite and 3% augite. The thin section observed from this outcrop contains 8% augite and 12% inverted pigeonite as well as 20% olivine. Given that unit A occurs at many stratigraphic levels in the upper part of the Poplar Lake intrusion, it is perhaps not surprising that it is variable in mode,

-- Canoe ~1200 m to the southeast arm Caribou Lake to outcrop-fronted campsite on northern shore--

STOP 1-15. UNIT M – AUGITE NORITE

Location: Outcrop pavement at campsite on north shore of southeastern arm of Caribou Lake.

UTMs: 686347E 5321875N

Description: Homogeneous outcrop of medium-grained, well-foliated, intergranular augite norite. Thin section show that 25% of rock is anhedral to subprismatic granular inverted pigeonite with anhedral granular augite comprising about 10% of the rock. The prismatic inverted pigeonite weathers brown which makes it appear similar to olivine in outcrop (Fig. 1-16). The rock is devoid of olivine and contains only trace Fe-Ti oxide.

Nathan and Morey (1977) describe unit M as a coarse-grained, poorly foliated, gabbronorite with average mafic mineral concentrations of up to 2% oxide, 13% inverted pigeonite, and 24% augite.

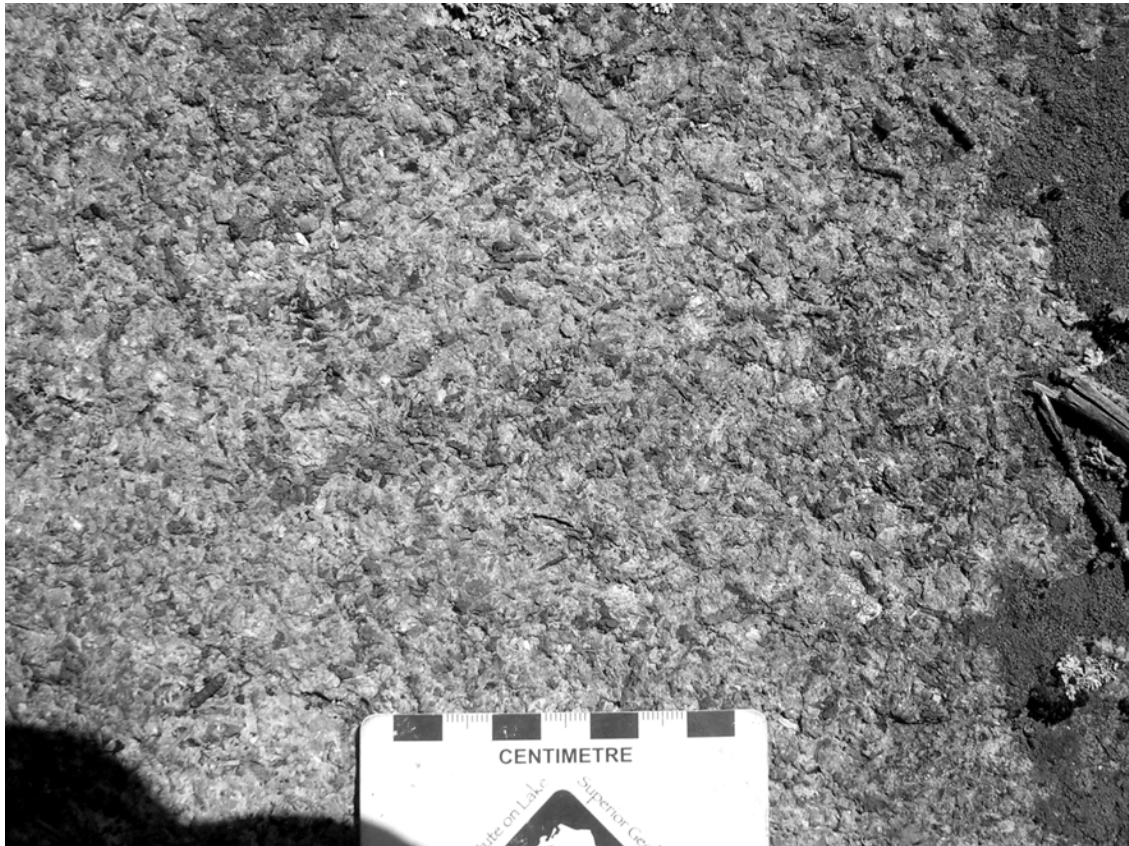


Figure 1-16. Texture of the prismatic augite norite at Stop 1-15.

-- Canoe ~400 m south to the portage to Horseshoe Lake --

STOP 1-16. UNIT M – GABBRONORITIC ANORTHOSITE

Location: Small outcrop at portage landing to Horseshoe Lake in southeastern Caribou Lake.

UTMs: 686375E 5321490N

Description: A small outcrop at portage landing appears to be an example of what Nathan cites as the “plagioclase-rich facies” of Unit M, which typically contains coarsely poikilitic pyroxene (Morey and Nathan, 1977). A thin section of this rock shows it to be medium coarse-grained, poorly foliated, ophitic gabbronoritic anorthosite.

-- Canoe ~1500 m to a campsite on a peninsular point in western Caribou Lake --

STOP 1-17. UNIT M OR A? – OLIVINE GABBRONORITE

Location: Outcrop-fronted campground on peninsula in western Caribou Lake.

UTMs: 685030E 5322070N

Description: Nathan includes this exposure within Unit M, but the rock is clearly olivine-bearing and is more similar to Unit A lithologies. A thin section from this location reveals a medium-grained, foliated, ophitic olivine gabbro with about 17% granular olivine, 13% poikilitic inverted pigeonite, and 4% subophitic augite. Basically, it is similar to the olivine gabbro observed at Stop 1-14, which is mapped as Unit A. Morey and Nathan (1977) place the Unit M – Unit A contact just north of this peninsula (Fig. 1-14), suggesting that it should be farther south.

-- Canoe north around island and then east; continue past small peninsula on north shore to campsite; total distance approximately 800m. --

STOP 1-18. UNIT M OR B – LAYERED GABBRONORITE

Location: Outcrop pavement at BWCA campsite on the north shore of Caribou Lake.

UTMs: 685575E 5322400N

Description: Generally a medium coarse-grained, subophitic to intergranular leucogabbro (Fig. 1-17A) locally with cm-scale mesocratic to melanocratic layers (Fig. 1-17B).

This exposure is situated right at a contact between Unit B and an irregular body of Unit M. Given its coarser grain size and generally leucocratic composition, it seems closer to Unit M.

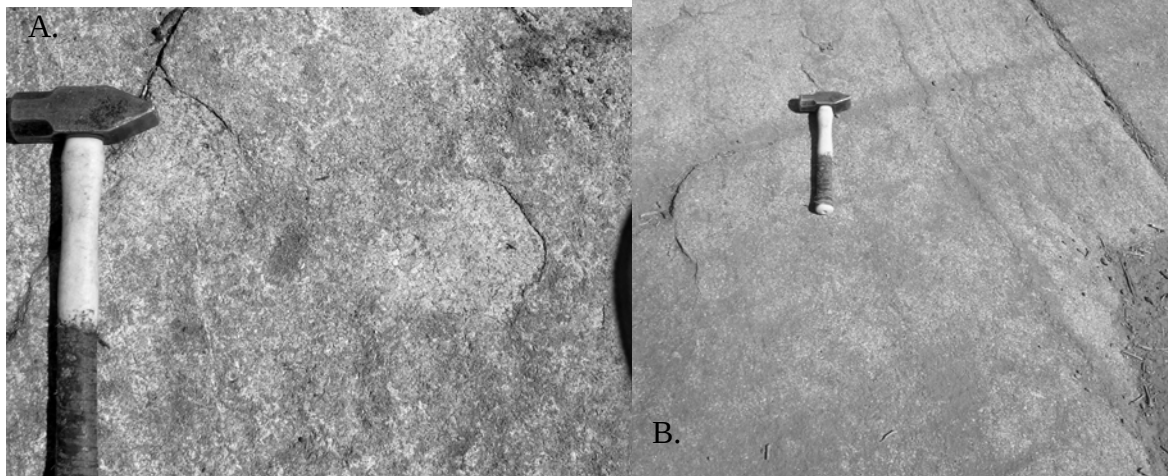


Figure 1-17. A) Texture of the leucogabbro at Stop 1-18. B) Cm-scale wavy modal layering displayed at Stop 1-18.

-- Continue along north shore to Lizz Lake portage. Return to Rockwood Lodge. --

END OF FIELD TRIP 1

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53rd Annual Institute on Lake Superior Geology

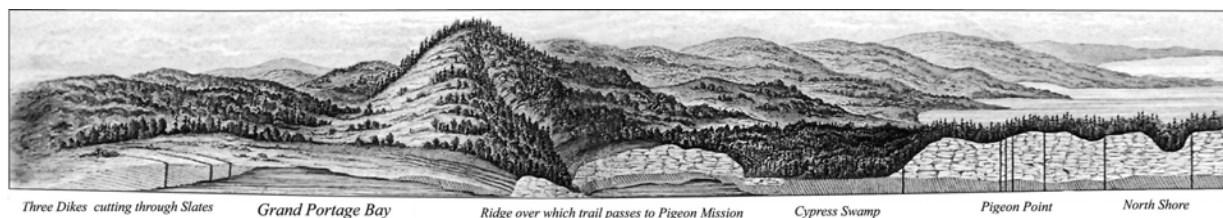
FIELD TRIP 2

GEOLOGIC AND CULTURAL HISTORY OF THE GRAND PORTAGE NATIONAL MONUMENT

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U.S. Geological Survey

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Brian A.M. Phillips
Lakehead University, emeritus



Engraving of Grand Portage area made from field sketch done in 1849 by survey party from U.S. General Land Office. Illustration is included in report by David Dale Owen in 1852. Engraving shows physiography of the area as well as geologic cross sections.

GEOLOGIC AND CULTURAL HISTORY OF THE GRAND PORTAGE NATIONAL MONUMENT

William F. Cannon, U.S. Geological Survey

Brian A.M. Phillips, Lakehead University, emeritus

Including a section on cultural history excerpted from a report by Douglas Birk, Senior Archeologist emeritus, Institute for Minnesota Archeology

Introduction

The Grand Portage National Monument, maintained by the U.S. National Park Service, consists of a restoration of an 18th century fort and trading post on the shore of Lake Superior and the 8.5 mile portage trail extending northwestward from the lake shore to the Pigeon River. The unrestored Fort Charlotte lies at the northwest end of the trail, on the south bank of the Pigeon River, at the head of a long series of rapids and waterfalls that made this long portage necessary in order to continue westward canoe travel from Lake Superior. The portage trail was the principal highway to the west for 18th century explorers, settlers, and fur traders who crossed the portage from Lake Superior to the Pigeon River, then into the westward draining Rainy River system, and eventually deep into the northern Great Plains and Rocky Mountains. It was over this trail that the wealth of furs was brought on their way to Montreal and eventually to Europe. Those who crossed the trail were largely responsible for the exploration and development of the Canadian west and northern U. S. Great Plains. The early fur traders, the Voyageurs, made their annual trek over the portage in late summer and then began their long canoe journey to the west to trade with native peoples for furs over the ensuing winter. They returned over the portage the following summer to the trading post to deliver their bounty of furs to representatives of the North West Company (NWC) who annually traveled from Montreal to the Grand Portage trading post.

The geology of the Monument and surrounding countryside consists of: 1) the Rove Formation, a Paleoproterozoic clastic sequence of shale and sandstone, 2) mafic dikes and flows formed in the Mesoproterozoic Midcontinent Rift, and 3) glacial and post-glacial deposits produced by the last tongue of Pleistocene glaciers that occupied the western Lake Superior basin and a variety of lake shoreline features formed by a series of post-glacial lakes that inundated the area shortly after deglaciation.

During this trip we will tour the restored fort and trading post, discuss the European and Native American history of the site, and examine bedrock and glacial to post-glacial geologic features of the area, particularly as they affected the human history through the unique set of physiographic features that they produced. Figure 2.1 is a map of the Monument showing the location of restored buildings on Grand Portage Bay, and three geologic field trip stops.

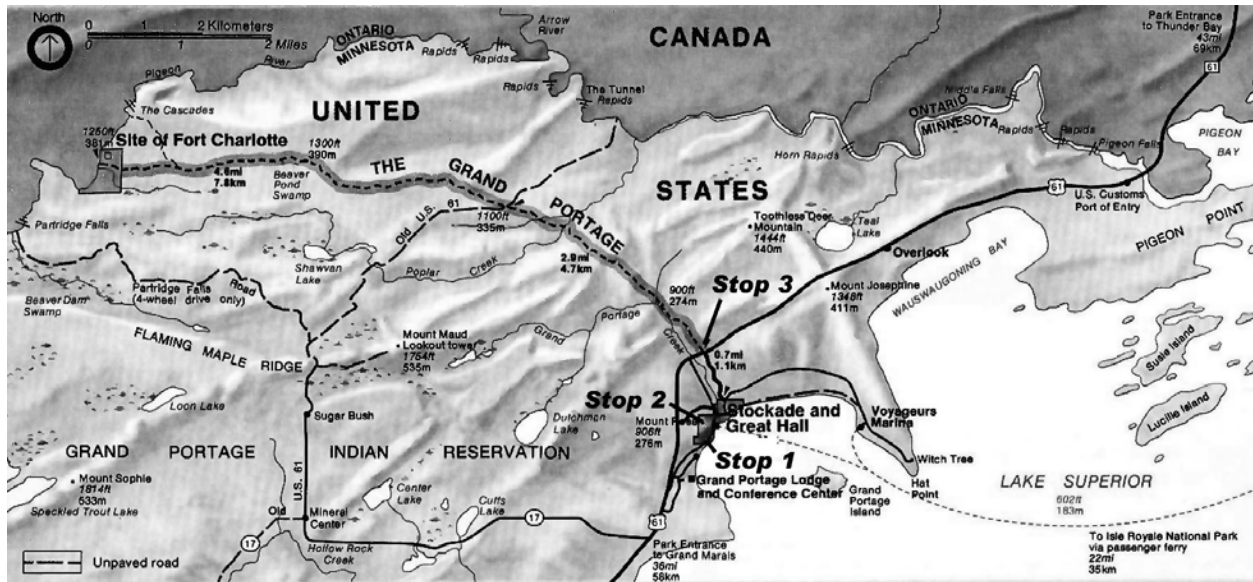


Figure 2.1. Map of the Grand Portage region showing the location of the National Monument and portage trail. Geologic field trip stops are also shown.

Cultural history

(This account of cultural history is excerpted and slightly modified from Douglas Birk, 2005, National Register of Historic Places, Grand Portage)

Grand Portage derives its name and much of its historical significance from a formidable portage, more than eight miles long, which links the Great Lakes with an interconnected series of inland waterways to the northwest. Key to the development of the North American fur trade, the Grand Portage gave Canadian explorers access to the northern Plains and to travel routes extending to the Arctic and Pacific Oceans. In the late eighteenth century the portage became a hub of seasonal coalescence and transshipment activities that linked suppliers in far-off places like England, the West Indies, and Brazil with a network of trading posts set in mosaic Native hunting territories throughout western Canada. The setting of Grand Portage, on the west side of Lake Superior and near the mid-point of a 3,000-mile commercial canoe transportation system, enhanced its role in channeling, shaping, and controlling the northwestern Montreal fur trade. Inspired by these qualities, one historian has declared that Grand Portage was once “probably the single most important fur trade location in the world.”

North American Indian peoples knew the use of canoes and portaging thousands of years before the arrival of Europeans, and they initially opened the Grand Portage route. French colonials traveling west from Montreal first arrived at Lake Superior early in the seventeenth century. Having learned of the Grand Portage from Indian informants by at least 1722, the French began crossing it in 1731 to advance their affairs and alliances in more remote areas to the Northwest. From then through the mid-1800s, the portage attracted a colorful succession of French, British, and American fur traders. Between 1780 and 1802, at the height of the commercial traffic, Grand Portage was a western headquarters and regional trade center of the North West Company. It was also the scene of an annual summer gathering that made it one of the busiest nodes of human activity in the midcontinent. By promoting the mercantile interests of France and Great Britain, the Grand Portage facilitated the spread of European colonialism and the development of the nations of North America. In the nineteenth century Grand Portage also played an important part in settling the disputed boundary between Minnesota and Ontario, which is to say between Canada and the contiguous United States in the region west of Lake Superior.

The most enduring human presence at Grand Portage is that of aboriginal North Americans, including, most recently, the Ojibwe, Algonkian-speaking peoples with Woodland cultural traditions. Details surrounding the emergence of the Ojibwe people or the Ojibwe identity are often debated. Some traditional accounts indicate that, in the seventeenth and eighteenth centuries, the Saulteur (Ojibwe) of the eastern Lake Superior region extended their territorial prerogatives westward with the advancing fur trade. A group since known as the Grand Portage Ojibwe established ties to Grand Portage by the 1730s. They were active in that region throughout the fur trade era, and they remain a vital community today. All landscapes, resources, sites, properties, themes, and contexts that are now part of the Grand Portage National Monument story are in some way linked to the Grand Portage Ojibwe. The National Monument was formed through their invitation and cooperation, and efforts to maintain and interpret historic properties and traditions at Grand Portage have long been assisted by that local band or by individual members of the band.

French and later British traders commonly entered the Great Lakes-Northwest trade by traveling west from Montreal. The use of Indian guides, birchbark canoes, and a vast network of established Indian canoe routes and portages assisted their movements (Morse, 1969; Birk, 1994). From the beginning, the fur trade built on entrenched Indian exchange practices while catering to Indian preferences and needs (e.g., White, 1982, 1987). Over time, as the business of accruing and transporting goods grew more complex and exchange frontiers expanded westward, the systems of trade were institutionalized. Certain individuals, like voyageurs, became specialists in the trade hierarchy just as certain places gained wider distinction as nodes or corridors of commerce. The most influential settlements were those that played strategic roles in the flow of workers, provisions, merchandise, and information (e.g., Hirth, 1976).

Grand Portage became such a place in the eighteenth century. Through regular use, the Grand Portage emerged as a hub for local trade and as a transshipment center--a gateway community that linked markets and linear transportation lanes in the east with branching trade routes and mosaic trade districts to the northwest (Birk, 1984). Between 1731 and 1804 tons of supplies and furs were shuttled over the portage; some, in and out of the warehouses eventually built at either end of the trail. After 1760 the portage became a general rendezvous. It was a veritable beehive of activity during summers, but in winters the outposts there were comparatively quiet and staffed only by small crews engaged in local trading and facilities management (Gilman, 1992). At the height of the trade, around 1800, Grand Portage was the western headquarters of the North West Company (NWC) and the rival XY Company (XYC), two of the largest commercial establishments in North America. When the NWC and XYC moved their operations north to Kaministiquia (later Fort William, Ontario) at the start of the nineteenth century, Grand Portage lost its identity and its status as trade center, abruptly becoming remote to the main channels of trade and communication and less important to the outside world.

Grand Portage is a place of majestic scale and natural beauty with many prominent landmarks. The lakefront area faces Grand Portage Bay, a horseshoe-shaped inlet that forms the deepest natural indentation on the Minnesota coast of Lake Superior (Schwartz, 1928). Flanking the northeast side of the bay is the rolling ridge of Hat Point (*Pointe au Chapeaux*), backed farther inland by the promontory of Mount Josephine, the top of which towers nearly 750-feet above the surface of the lake. The opposite, southwest edge of the bay is framed by the sloping headland of Raspberry Point (*Pointe à la Framboise*). About a mile offshore, in the center of the bay and partly shielding the bay from the big lake, is Grand Portage Island (also known as Pete's Island, Sheep Island, *Isle aux Mouton*, etc.) (Gates, 1965; Grout and others, 1959). Near the lakefront within the Grand Portage National Monument is an imposing hill known as Mount Rose. The rounded and rocky summit of the hill rises several hundred feet above the bay and overlooks all of the historic properties clustered along the shoreline there (Winchell, 1899; Thompson, 1969; Gilman, 1992), as well as the mouth of Grand Portage Creek, a stream channel that drains from the interior highlands and skirts the base of Mount Rose before entering Lake Superior. Mount Rose once

served as a vantage to watch for approaching watercraft, the earliest such documented use perhaps occurring in 1767 (Carver, 1956).

The Grand Portage is the first leg of a remarkable inland canoe route that passes from Lake Superior through a chain of lakes and rivers along the present international boundary between Ontario and Minnesota. That mainline route, sometimes called the “Voyageur’s Highway,” links with a vast network of other branching canoe trails in the hinterlands beyond (Morse, 1969). Few viable inland water routes emanate from the west side of Lake Superior, and the Grand Portage-Pigeon River route early proved to be the most direct and efficient gateway to the “border lakes” region and the far Northwest (Gilman, 1992). The Pigeon River is a relatively short stream whose headwaters at the rim of the Superior basin are only about 50 miles west of Grand Portage Bay and enters Lake Superior about five miles to the northeast of Grand Portage. The river is navigable by canoe for much of its length above Fort Charlotte. For 20 miles below Fort Charlotte, however, the Pigeon River is impassable. That stretch follows an eastward course through a sinuated series of rapids, canyons, and falls before reaching the big lake. The terrain on the Canadian side of the lower Pigeon River is too rugged and the distance between the lake and navigable parts of the river too great for portaging to be practical there (Schwartz, 1928; Buck, 1931; Burpee, 1931; Morse, 1969, Birk, 1998).

The ascent from Lake Superior through the Grand Portage and up the Pigeon River to the height-of-land separating the Laurentian and Hudsonian basins is the steepest part of the entire historic canoe route that once ran between Montreal and Athabasca (Morse, 1969; Birk, 1998). Prior to the time the NWC made improvements to the trail, heavily outfitted canoe brigades could take up to ten days to move from Lake Superior to the Pigeon River (Wallace, 1934). During the heyday of the fur trade, with the NWC installed at Grand Portage, the same carry might take from five to seven days (Gilman, 1992). Some men loaded with packs or kegs weighing 180 pounds or more were able to make a round trip on the portage within six hours (Buck, 1931). Because porters often moved at a trot and took short rests en route, their actual progress was about as expedient as that of an unencumbered walker, who, under optimal conditions and without rests, might cross the portage in 2.5 hours (Thompson, 1969). Early fur traders tried using horses and oxen on the portage but under normal conditions in warmer seasons of the year, the trail was best suited for human porters (Burpee, 1931; Thompson, 1969; Gilman, 1992). Animal-drawn carts or sleds were more commonly used during the “off season” (other than summers) or during the nineteenth century, when traffic on the portage was greatly reduced. Diminished forest cover along some parts of the portage in the 1800s opened extensive views of the surrounding terrain and perhaps changed the surface environment of the trail (e.g., Winchell, 1899). In 1858, for example, the trail was said to be dry and in good condition, making it passable for oxen teams (Hind in Dawson, 1968).

The Fort Charlotte site complex involves the remains of fur trade facilities that once stood along the Pigeon River at the head of the portage on either side of the mouth of Snow Creek. Evidence suggests that the NWC conducted operations along the river north of the creek (between the creek and the whitewater rapids), while the XYC and perhaps other earlier firms occupied an area along the river south of the creek. To reach these various facilities the portage trail probably forked somewhere east of the Pigeon River so that one branch of the trail led to the south side of the creek and the other continued along the north side. The main trail, the one most used and over the greatest length of time, was that on the north side. Underwater archeological investigations conducted at the main Pigeon River landing suggest that, through about 1780 or so, the main portage intersected the river very close to the mouth of Snow Creek. With the formation and growth of the NWC, the portage landing migrated downstream, first to the place where the NWC built a canoe landing and quay or “dock” (Winchell, 1899), and later to where the trail meets the river today, at a point about midway between Snow Creek and the head of the whitewater rapids (Birk, 1975; Birk and Wheeler, 1976; Wheeler and others, 1975; Birk, 1979). Modern recreational use of the landing, along with floating ice, fluctuating water levels, river currents, and log drives on the

Pigeon River, may have damaged early landing structures or associated offshore archeological deposits at Fort Charlotte (Birk, 1975).

French traders frequented the portage between 1731 and 1760 (Nute; 1944; Woolworth and Wooworth, 1982) and perhaps earlier (e.g., Burpee, 1931), yet suspected French-era deposits or structures at Grand Portage Bay have eluded archeological discovery. Surprisingly little information about French activities at Grand Portage is found in available written records either. Nevertheless, some scholars suggest that French traders erected buildings at Grand Portage before 1760 (e.g., Warren, 1957; Buck, 1931; Babcock, 1940; Nute, 1944; Woolworth, 1967; Woolworth and Woolworth, 1982), that they had facilities at both ends of the portage as early as 1732 (Woolworth and Woolworth, 1982), or that they maintained structures at each end of the portage from ca. 1732 through 1760 (Woolworth and Woolworth, 1982). An unsubstantiated “French post” at the Pigeon River landing even has an assigned cultural resource inventory number (CR-105) (Woolworth and Woolworth, 1982). Other researchers are skeptical of such varied interpretations (e.g., Burpee, 1931; Thompson, 1969; Gilman, 1992). The apparent lack of French-colonial materials on the lakefront at Grand Portage Bay west of Grand Portage Creek hints that the greatest potential for such materials there may lie in areas east of the creek (Brown, 1937). To date, the only documented French colonial archeological loci are those identified during field investigations at the Pigeon River landing (Birk and Wheeler, 1976) and at the suspected Parting Trees pose on the Grand Portage trail (Douglas A. Birk, personal observation). Barring future archival discoveries, most questions regarding French presence at Grand Portage may only be answered through further archeological inquiry.



Figure 2.2. Artist's conception of North West Company Depot at Grand Portage

The earliest known trading houses at Grand Portage Bay were those established by British traders in about 1768. The houses were built west of Grand Portage Creek at the place later developed for the NWC depot (Nute, 1940; 1944). Other trade facilities were later constructed along the lakefront and, by 1793, the NWC alone was said to have sixteen buildings within its fort. The NWC also claimed dominion over adjoining parcels at the bay and may have opened facilities at the Pigeon River at the time of its formation (1783). Rival traders soon followed the NWC's example by building a “hangard or store” on the Pigeon River in 1785 (Gates, 1965; Thompson, 1969). Being more abundant and pervasive, material

evidence of British presence at Grand Portage has proved easier to find, identify, and interpret than that relating to earlier French colonial operations. Some features noted or implied in early historical records have yet to be identified through archeology. For example, the NWC had horses, cattle, hogs, and sheep at Grand Portage (Thompson, 1969; Gilman, 1992), which likely necessitated the use of barns, stables, pens, or corrals. The central clearing at Grand Portage Bay could have been used for pasturage, and grazing livestock there may actually have given the NWC a reason and excuse to fence the perimeters of that opening. To overwinter livestock required labor-intensive preparations. Hay was said to be abundant at Grand Portage in the late 1700s, but it had to be cut and stacked by hand. At the same time, cold damp fog, along with high levels of ground moisture and lingering winters, often caused harvested hay to rot (e.g., Thompson, 1969). Gardening was similarly impaired by weather and soil conditions. Potatoes were the only food crop typically grown with any success (e.g., Thompson, 1969; Woolworth and Woolworth, 1982). The size, location, and arrangement of the NWC's barns, corrals, pastures, and gardens are presently unknown.

Grand Portage National Monument

Grand Portage was designated a National Historic Site in 1951. In 1958 Grand Portage National Monument was established to commemorate and preserve a premier site and route of the 18th century fur trade that led to pioneering international commerce and exploration in North America as well as cultural contact between Ojibwe and other Native societies and the North West Company partners, clerks, and canoe-men. The Monument was also established to work with the Grand Portage Band in preserving and interpreting the heritage and lifeways of the Ojibwe people. Grand Portage National Monument is a homeplace of tribal and family history and cultural persistence. Grand Portage National Monument contains reconstructed buildings and well-preserved archeological remains of several fur trading posts instrumental in the exploration of the West and in the economic history of the United States and Canada. The national monument contains the entire length of the portage that marked the entrance into the interior of western Canada. The national monument is significant because of the fundamental interrelationship of Ojibwe heritage and fur trade history. Grand Portage National Monument protects, commemorates, and interprets a reconstructed fur depot of the North West Company, a rendezvous site for international commerce and canoe route for transcontinental exploration, Native heritage, natural scene, and history of cross cultural contact and accommodation between traders, Ojibwe, and other participants in the fur trade.

Today nearly 90,000 visitors pass through the reconstructed stockade and related buildings each year. Three reconstructed historical buildings currently serve these visitors. Based on archeological excavations and research from 1938 through the present, the reconstructed buildings have been furnished in the period of the late 1700's. Technology from the time can be viewed in the Ojibwe Village and Voyageur Encampment staffed by National Park Rangers. The Great Hall and kitchen buildings contain reproduction furnishings similar to those used by late 18th century North West Company business partners, clerks, voyageur guides and Ojibwe families. In a room of the Great Hall from June until September, Ojibwe artisans create and sell beaded designs and birchbark basketry popular to Indian people of the Lake Superior region. From late May until early September interpretive programs are presented around the historical site. Subjects include historic cooking and baking, Ojibwe craft demonstrations, historic black powder musket firings, lever fur press operation, historic gardens and conducted walking tours. An annual highlight is Rendezvous Days, held the second week of August, during which the annual meeting of company representatives from Montreal, fur traders and local native people is recreated.

Geologic history

The geology of Grand Portage is discussed in regard to Precambrian bedrock features, a relatively uncomplicated story based on both Paleoproterozoic and Mesoproterozoic rock units, and Pleistocene and

Holocene history representing a dynamic story of geologically rapid landscape change caused by post-glacial changes in the levels of ancestral Lake Superior. These latter changes overlapped with early human occupation of the area and undoubtedly had a major influence on human activity. The most recent geologic investigations of the monument were done in 2001 and funded by the National Park Service. Results are in administrative reports to NPS: (1) the bedrock geology was studied by W.F. Cannon and Laurel G. Woodruff of the USGS; (2) glacial and post-glacial geology was surveyed by Brian A.M. Phillips of Lakehead University. Much of the following descriptions are taken from those reports.

Bedrock geology

The bedrock geology of extreme northeastern Minnesota, including the Grand Portage National Monument, consists of Precambrian rocks of two separate ages. The bedrock geology of the Grand Portage region is shown in Figure 2.3. Most of the area is underlain by sedimentary rocks of the Rove Formation, formed during Paleoproterozoic time, approximately 1,850 million years ago. These rocks have been slightly tilted toward the southeast but otherwise are remarkably little altered for rocks of such antiquity. A second period of rock formation occurred during Mesoproterozoic time, approximately 1,100 million years ago. At that time diabase dikes were emplaced into the Rove Formation. These dikes are of two distinct generations separated by a short time interval and exhibit contrasting magnetic polarity. The oldest are known as the Grand Portage dike swarm and are generally east-trending features no more than a few tens of feet thick. Slightly younger dikes of the Pigeon River swarm are much more voluminous and support the higher ridges of the region such as Mt. Rose and Mt. Josephine. The dikes form a roughly orthogonal array of NE-trending and NW-trending bodies.

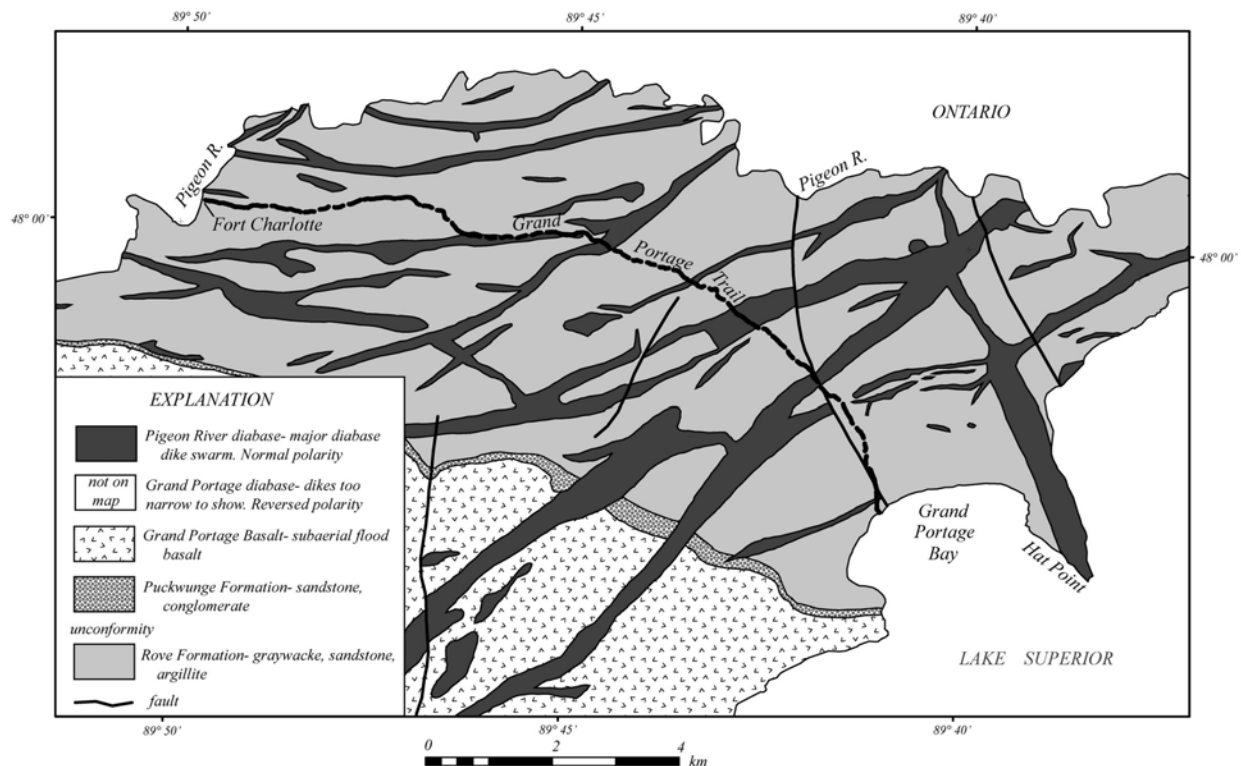


Figure 2.3. Bedrock geologic map of the Grand Portage area in northeastern Minnesota. Geology extracted from digital files by Miller and others, 2001

The essentials of the geologic relationships were established by the earliest geological reconnaissance of the region conducted by J.G Norwood in 1849 (Owen, 1852). An etching from Owen's report (see title

page) illustrates in a schematic manner the geology of the Grand Portage area. It shows large diabase bodies that form both sills, as on Pigeon Point, and dikes, as on Hat Point. "Slates" of the Rove Formation in Grand Portage Bay are cut by thin diabase dikes.

There are three principal references on the geology of the region. First, Grout and Schwartz (1933) studied the Rove Formation and diabase intrusions and presented the first detailed maps of the Grand Portage area. In addition to a compilation of the regional geology, they presented maps of individual townships showing the geology and the location of bedrock exposures observed during their studies. Later, Grout, Sharp, and Schwartz (1959) published a report on the geology of Cook County, which provided some updated information on the Grand Portage area. Finally, in 1969, Morey published a detailed examination of the Rove Formation in northeastern Minnesota and adjacent parts of Ontario. This publication provides the best description and interpretation of the geologic history of the Rove and includes some specific observations within the Monument, particularly at Mt. Rose.

Rove Formation

The report by Morey (1969) provides an excellent description of the Rove Formation. The abstract of that report is presented below as a summary of the regional character of the Rove and some of our specific observations within the Monument follow.

According to Morey (1969, p. 1-2):

"The Middle Precambrian Rove Formation, the upper part of the Animikie Group, is estimated to be at least 3,200 feet thick and is exposed between northwestern Cook County, Minnesota and the Thunder Bay district, Ontario. It is a sequence of graywacke, argillite, locally abundant intraformational conglomerate, quartzite, and carbonate rocks. The Formation was deposited sometime between 2.0 and 1.7 b.y. ago in a northeast-trending basin, the configuration of which was controlled by a pre-existing structural grain.

Detailed mapping in the 7 ½ minute South Lake quadrangle combined with a field and laboratory study of approximately 150 other scattered stratigraphic sections provide a basis for the recognition of three informal lithic units. From oldest to youngest these are: (1) lower argillite, 400 feet thick; (2) transitional beds of interbedded argillite and greywacke, 70 to 100 feet thick; and (3) thin-bedded greywacke, as much as 2,700 feet thick.

It is concluded that the argillite and associated greywacke-sandstone and graywacke-siltstone units were deposited in moderately deep, quiet water. Repeated greywacke sedimentation units indicate sediment transport and deposition by turbidity currents. A sedimentation unit reconstructed from composite sections consists of (1) a basal conglomeratic graywacke, (2) a structureless unit which grades indistinctly into (3) a graded graywacke that is overlain by (4) a laminated graywacke, which may be modified by (5) small-scale cross-bedding, or (6) contorted bedding. Any one or several of these may be absent, but the units are always overlain by (7) an argillite.

Post-depositional soft-sediment structures such as load casts, flame structures, clastic dikes, bed pull-aparts, overfolds, and micro-faults indicate rapid deposition of Rove sediments, active bottom currents, and post-depositional deformation, implying a significant paleoslope.

A detailed analysis of paleocurrent directional indicators such as groove casts, flute casts, dendritic ridges, and cross-bedding shows that the turbidity currents had a southerly trend about perpendicular to the axis of the Rove basin. However, ripple marks, winnowed lag deposits at the tops of many greywacke beds, and possibly some festoon-type cross-bedding show that the turbidites were later modified by bottom currents that trended southwesterly or parallel to the axis of the basin.

The heavy minerals of the Rove are characterized by epidote-group minerals, apatite, sphene, and tourmaline, and are typical of older Precambrian igneous rocks now exposed north of the present Rove outcrop area.

Thin-section and x-ray analyses of 200 samples show that the graywackes consist of angular, poorly sorted grains of clastic quartz and plagioclase (An₁₀-An₂₀) embedded in an argillaceous matrix that now consists of quartz, chlorite, and muscovite. The fine-grained fissile argillite and mudstone have the same mineralogy and microtextures as the greywacke.

Erosion subsequent to pre-Keweenaw tilting removed an unknown amount of the formation prior to deposition of Lower Keweenaw sedimentary rocks. The intrusion of Middle Keweenaw mafic igneous rocks caused local metamorphism of the Rove Formation to a variety of mineral assemblages now assigned to the pyroxene- and hornblende-hornfels facies, but the remainder of the formation is essentially unmetamorphosed.”

An additional study of the Rove has been reported recently (Maric and Fralick, 2005) and presents a somewhat different picture from Morey's 1969 description. Based on examination of a series of drill holes near Thunder Bay, they defined an internal stratigraphy consisting of: 1) a 150 m-thick basal unit of carbonaceous shale with lesser siltstone and fine-grained sandstone beds, which grades upward into, 2) a unit about 350 m thick containing more than 100 individually coarsening upward beds of sandstone with interlayered shale. This sequence is overlain by the uppermost preserved unit of 3) black shale with thin rippled sandstone beds. They interpreted the lower carbon-rich sediments to have been deposited in a sediment-starved anoxic basin in which very slow sedimentation rates resulted in accumulation of only 100 m of black shale during tens of millions of years of sedimentation. The upper coarser-grained units were deposited by a southward prograding turbiditic to shelf system.

Most, if not all, of the Rove Formation exposed in the Grand Portage Monument is part of the upper thin-bedded greywacke unit described by Morey (1969). Most exposures consist of flaggy, fine- to medium-grained greywacke, with minor interbeds of argillite. The unit tends to split into slabs 1-2 inches thick parallel to indistinct bedding. The extensive outcrops and roadcuts on the north side of Highway 61 where the portage trail crosses the highway (Fig. 2.1, Stop 2.3) are typical of many of the other exposures along the trail to the north. The exposures on Mt. Rose (Stop 2) differ in that bedding of the greywacke is somewhat thinner and more distinct. Features such as cross-bedding and contorted bedding are present, although rare. An additional significant difference is the abundance of units of fine-grained fissile argillite as much as several tens of feet thick. The Rove Formation immediately adjacent to the Mt. Rose diabase dike is strongly metamorphosed. Original sedimentary textures are largely obliterated and metamorphic minerals, including biotite and clots of a mineral preliminarily identified as cordierite, have formed in the original greywacke matrix.

Over the past few years, the age of the Rove Formation has been clarified by radiometric dating and the results provide a more precise age estimate than the 300 m.y. span indicated by Morey's study in the 1960's. Thin volcanic layers near the base of the Rove Formation near Thunder Bay, Ontario have been dated by the U-Pb zircon technique and indicate that deposition began somewhat before 1835 Ma. The dated volcanic layer is about 6m above the base of the formation and has an age of 1836+/- 8 Ma (Addison and others, 2005). The base of the Rove in that same area lies on a layer of ejecta believed to have originated from the major 1850 Ma meteorite impact at Sudbury, Ontario (Addison and others, 2005), further helping to bracket the age of the basal beds of the formation between 1850 and 1836 Ma. An additional age constraint comes from radiometric dates on detrital zircons from a bed about 400 m above the base and near the top of the preserved part of the Rove about 15 km north of Grand Portage. Concordant ages of individual grains are as young as 1777 Ma, providing a maximum age for deposition of the middle part of the Rove Formation (figure 2.4) (Heamon and Easton, 2006). A minimum age is not well constrained at this time.

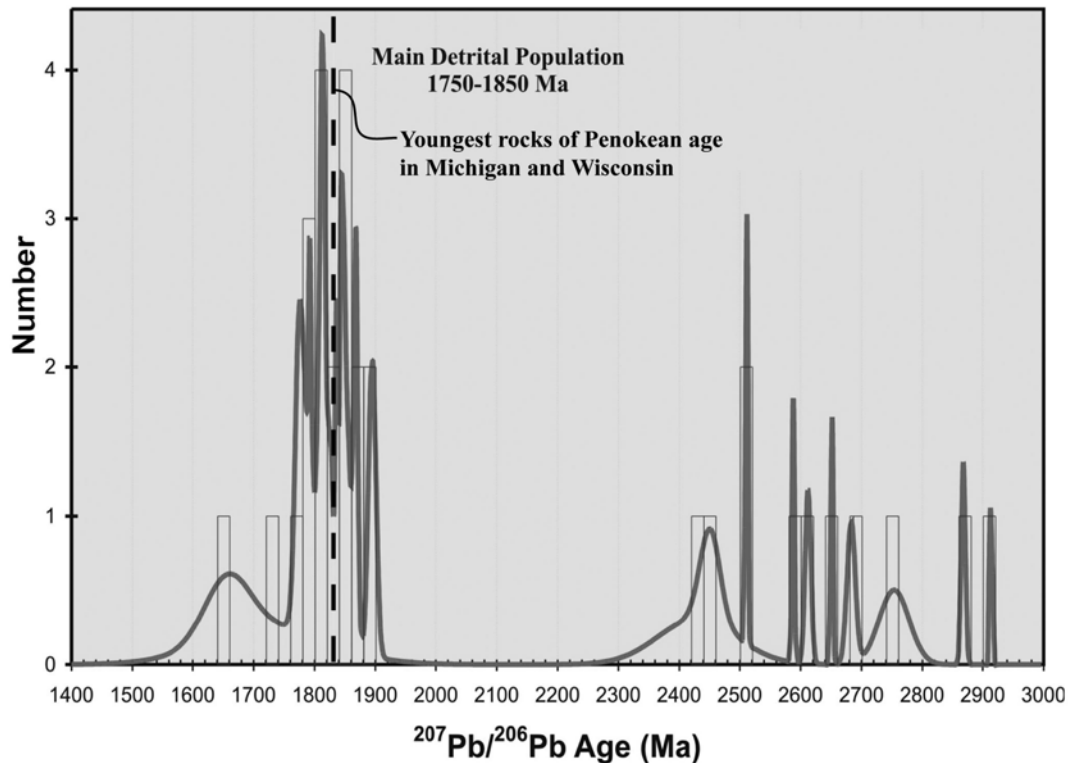


Figure 2.4. Age spectrum for detrital zircons in the Rove Formation presented by Heamon and Easton (2006, figure 33). Spectrum indicates that a majority of detrital zircons were derived from a source younger than rocks of the Penokean orogeny to the south, and have come from a terrane with an abundance of zircons with ages of 1850 to 1750 Ma.

These new age data require a rethinking of traditional interpretations of the Rove, both in terms of stratigraphic correlations and tectonic setting of deposition. The Rove Formation has been long considered to be the stratigraphic equivalent of major turbidite sequences to the south, such as the Thomson Formation in east-central Minnesota and the Michigamme Formation in northern Michigan. Recent discovery of the Sudbury ejecta layer in Michigan near the base of the Michigamme Formation confirms that the onset of sedimentation was nearly synchronous across the region at 1850 Ma. However, the duration of sedimentation between Michigan and northern Minnesota/Ontario appears to be drastically different. The 1777 Ma concordant detrital zircons from the Rove are from a bed about 400 m above the base of the formation. Although this is near the top of the preserved formation in the sample area, a total thickness nearly 1000 m was estimated for the Rove (Morey, 1969). Therefore, sedimentation may have lasted well beyond the 1777 Ma maximum age. In Michigan, it is well established that the Michigamme Formation was deposited, deeply buried by tectonic overthrusting, metamorphosed, and intruded by post-tectonic granite by 1835 Ma (Schneider and others, 2002; Schneider and others, 2004), which is the approximate age of the volcanic layers near the base of the Rove. Thus, much of the thick clastic sequence of the Rove appears to be entirely younger than the Michigamme Formation. The Rove has commonly been considered to have been deposited in the foreland sedimentary basin north of the 1890-1835 Ma Penokean orogen, but it now appears that sedimentation was almost entirely after the Penokean deformation ceased in Michigan and Wisconsin. The detrital zircon suite in the Rove, as shown in figure 4, consists dominantly of grains in the 1850-1750 Ma age range rather than Archean zircons, indicating that a source area with rocks of that age was well established and contributing the majority of detritus to the basin during at least the younger part of Rove deposition. Detrital zircon analyses of the Thomson and Rove Formations in Minnesota yielded somewhat similar results (Kerber, 2006). Although no zircons younger than about 1800 Ma were found in either formation based on analyses of a single sample from

each, the great preponderance of zircons are in the range 2000-1800 Ma and Archean zircons are scarce). Thus, based on presently available data, it appears that the preponderance of detritus contained in the vast turbidite deposits of the Rove, Virginia, Thomson, Tyler, and Michigamme Formations across the western Lake Superior region was derived from a source terrane(s) with ages in the approximate range 2000-1750 Ma.

A widely accepted model for this turbidite deposition has been a tectonic foreland in which subsidence was being driven by northward overthrusting of arc rocks of the Wisconsin magmatic terranes, now exposed in northern Wisconsin, and that these rocks provided much of the detritus of which the turbidites are composed. This model is still acceptable for the southern parts of the basin, but complications arise in applying it to our current understanding of the Rove Formation. Relationships in Michigan and Wisconsin indicate that tectonic loading by overthrusting onto the craton margin was completed by 1835 Ma (Schneider and others, 2002). However, deposition of the basal units of the Rove was just beginning at that time. Deposition of the Rove continued for at least 60 million years after that time, long after the end of the Penokean orogeny, and basin subsidence of about 1 km occurred during that period. What was the driving mechanism for that subsidence? An additional enigma is provided by a wealth of paleocurrent data from the Rove which indicates a consistent southerly sediment transport (Morey, 1969). This direction is inconsistent with a sediment source from the Wisconsin magmatic terranes. The preponderance of Paleoproterozoic detrital zircons suggests that much of the Rove detritus was derived from a Paleoproterozoic orogen north of the Lake Superior region, perhaps the Trans-Hudson orogen in part, which contains rock units as young as 1775 Ma. Thus the Rove Formation no longer fits well into the model of Penokean foreland sedimentation because the time of sedimentation and sediment source directions are inconsistent with that model.

The nature and cause of the Rove basin is a newly emerging topic for further study. Because so much of the Rove sedimentation is significantly younger than the Penokean orogeny and paleocurrent data indicate that the Penokean orogen is not the source for the Paleoproterozoic-aged detritus that seems to make up most of the Rove, it is necessary to divorce the Rove, and perhaps at least part of the correlative Virginia Formation, from the suite of features ascribed to the Penokean orogeny, and seek a cause in younger orogenic events. It appears, based on the most recent data, that the basal parts of the Rove Formation composed of black shale deposited in a sediment-starved basin may be a product of Penokean foreland basin development. However, the remaining coarser clastic deposits of the upper part of the Rove reflect a reactivation of the basin during a time spanning the 1775 Ma maximum age of the middle part of the formation. A recent reinterpretation of the tectonic assembly of the north-central part of the US (NICE Working Group, in press) indicates that the mid-geon 17 Yavapai orogeny extends eastward from its originally defined extent in the southwestern US into the southern part of the Lake Superior region. A belt of gneiss domes, granite plutons, and metamorphic nodes also developed across central Minnesota and northern Michigan at that time (Schneider and others, 2004; Holm and others, 2005). Thus, it has been established recently that major mid-geon 17 tectonic activity occurred only about 200 km south of the Rove basin. Is this tectonic episode responsible for reactivation of the older Penokean foreland basin and deposition of the Rove Formation? Although these Yavapai-age events might in some poorly understood manner have caused basin subsidence and also provide a source of detritus of the same age as that established for the Rove, the enigma of southerly-directed sediment transport in the Rove of material of that age remains unresolved.

Grand Portage diabase dikes

A set of thin diabase dikes was described by Green and others (1987) that they named the Grand Portage dike swarm. These dikes are present throughout the Grand Portage region and intrude the Rove Formation as well as the Middle Proterozoic volcanic rocks exposed south of Grand Portage. More than 50 of these dikes have been identified, some of which are shown on maps in Green and others (1987) and

in the digital compilation by Miller and others (2001). The dikes have a relatively uniform trend from $N65^{\circ} - 90^{\circ}$ E. The Grand Portage dikes tend to be quite strongly magnetized, but acquired their magnetic properties as they cooled during a period when the earth's magnetic field was inverted. This reversed magnetic polarity interferes with the present earth's magnetic field resulting in distinctive linear negative anomalies on magnetic maps. This characteristic makes them quite easy to identify, even in areas where they are not exposed. A combination of surface exposures and magnetic expression has allowed us to identify 11 new dikes of the Grand Portage swarm that cross the monument. The most prominent of these are a set of five dikes that are exposed along the east side of Mt Rose (one is just south of the Monument boundary, the other four are on U.S. Park Service property). Each dike stands as a low to prominent ridge as much as 10 m wide as a result of differential erosion between the relatively hard diabase and much softer Rove Formation. One of these dikes shown in Figure 2.5a is probably the dike that inspired the somewhat fanciful etching that appeared in Owen's 1852 report (Figure 2.5b).

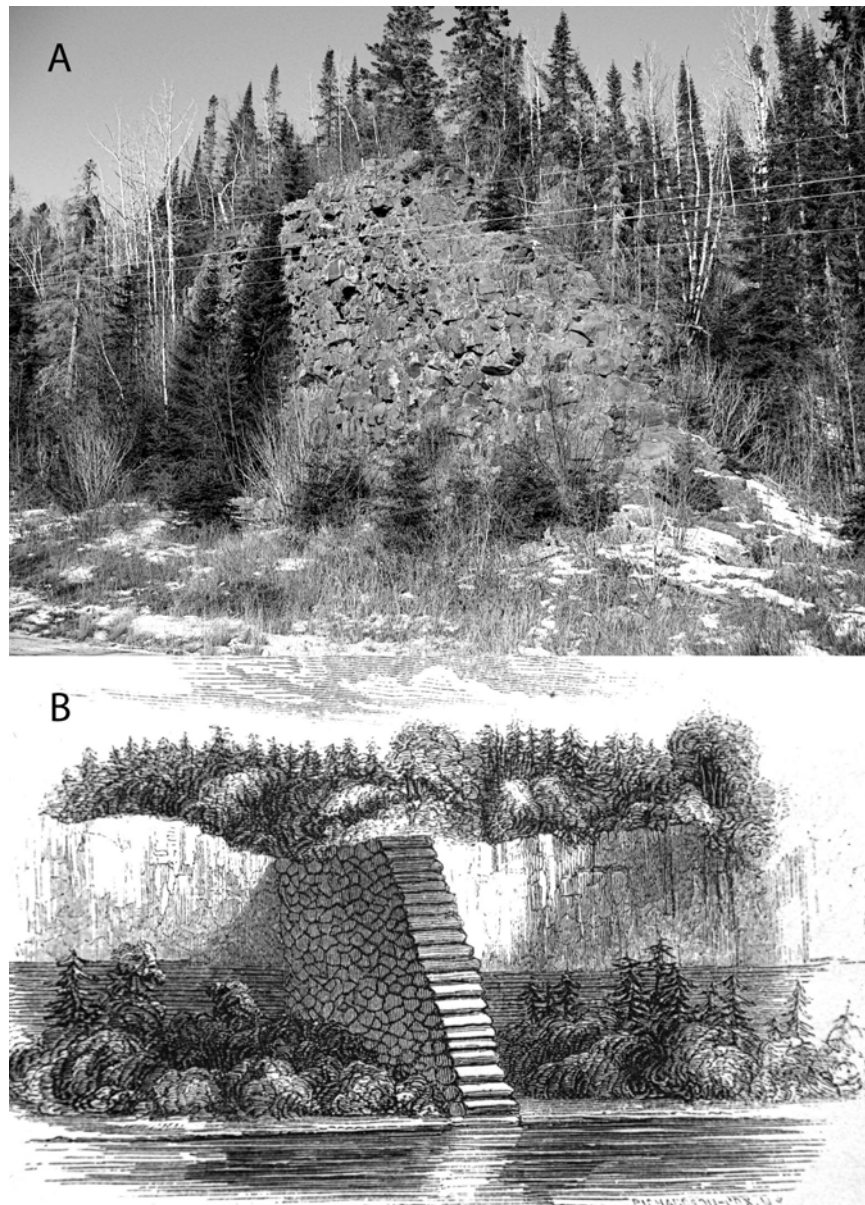


Figure 2.5. a- Diabase dike of Grand Portage swarm at south edge of Monument property. b- Etching from Owen (1852) based on field sketch, probably of the dike in a.

The magnetic profile shown in Figure 2.6 illustrates the sharp negative magnetic anomalies caused by these dikes. Three other similar anomalies are shown on a magnetic profile along the Grand Portage trail. Although dikes are not exposed at the surface in these areas they are inferred with confidence to occur beneath the surficial cover material. Two other dikes were identified in outcrop just north of Highway 61 but their magnetic expression is less distinct.

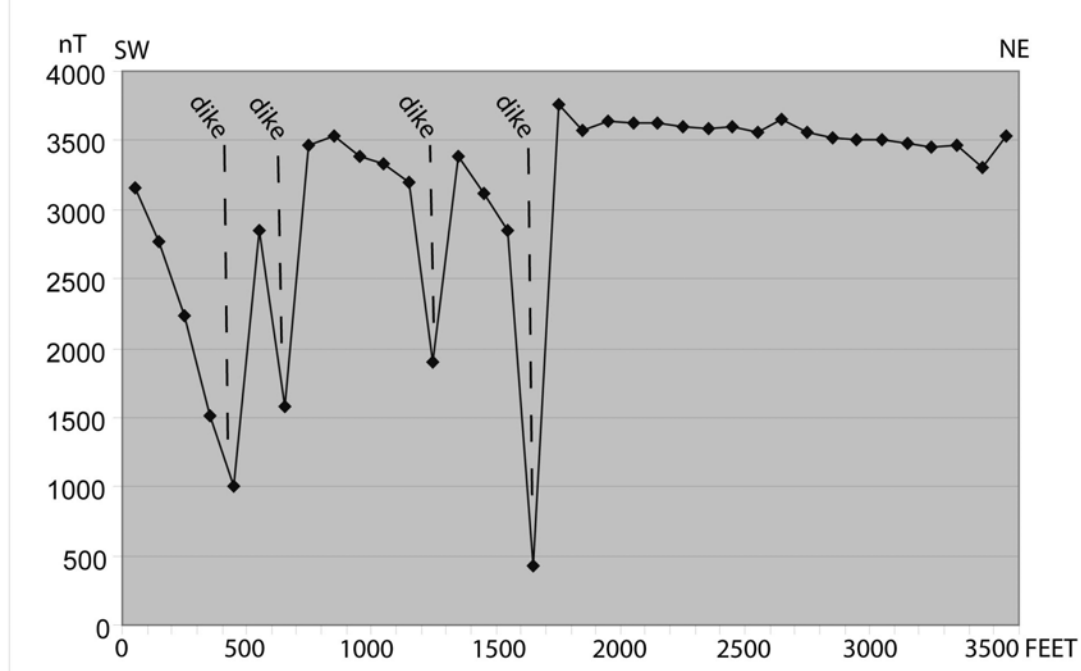


Figure 2.6. Magnetic profile measured at 100 foot spacing along roadway northward from south boundary of Monument. The four sharp negative anomalies are caused by the four Grand Portage dikes exposed along the roadway.

Pigeon River diabase dikes

The Pigeon River diabase dike swarm (Green and others, 1987) is compositionally similar to the Grand Portage dikes but is slightly younger and volumetrically much more prominent. The Pigeon River dikes trend mostly about N65°E and less commonly about N40°W. Because of their hard, massive nature they form the prominent, steep-sided ridges of the region, including Mt. Rose. Because they have normal magnetic polarity, these dikes tend to produce positive anomalies on magnetic maps, but they are apparently somewhat variably magnetized so cannot be recognized in magnetic data as readily as the Grand Portage dikes.

Absolute ages of Grand Portage and Pigeon River dikes

The Grand Portage and Pigeon River diabase dikes have not been dated directly, but their age can be closely approximated because of their relationships with other nearby units that have been dated. A regional correlation of rock units related to the Midcontinent Rift and a summary of radiometric ages was published by Nicholson and others (1997). The Grand Portage dikes are known to cut the Hovland and Grand Portage basalts south of the Monument. Those basalts were erupted about 1,110 to 1,105 million years ago. The shift of the earth's magnetic field from reversed to normal polarity occurred about 1,100 years ago. The intrusion of the Grand Portage dikes, therefore, must have occurred after eruption of the Grand Portage basalts, but before the magnetic field reversal in the interval 1,105 to 1,100 million years

ago. The Pigeon River dikes, which have normal magnetic polarity, are younger than 1,100 million years. They may be related to the Lutsen basalts that were erupted about 1095 to 1090 million years ago.

Glacial and post-glacial geology

The effect of ice

Because ice advances reached as far south as Illinois, Iowa, Nebraska, and Kansas numerous times during the Pleistocene it is safe to assume that the Grand Portage area has been affected by the passage of several ice lobes over the last 500,000 years. The general direction of ice movement was from the north and northeast, and present-day landscapes in the borderlands first were sculpted of sub-glacial processes of plucking, abrasion and the high pressure flow of sub-glacial meltwater. Next, the processes of deglaciation resulted in deposition of ablation tills and a variety of fluvioglacial landforms (eskers, kames, ice-contact deltas) and sediments (sorted and unsorted silts, sands and gravels). Meltwater, temporarily ponded between the waning ice margin and enclosing topography, formed short-lived pro-glacial lakes that changed in area and depth as the configuration of the ice margin changed. Many of these lakes existed long enough to deposit substantial layers of lacustrine clays, and many ended their existence by draining catastrophically down spillways exhumed by the melting ice. Inundation of the land recently revealed by ice melt was common as various lake bodies occupied the Lake Superior basin at different times and water levels, depending upon the rate of water supply and the loss through such outlets as were available.

In the Grand Portage area, the present topography is the product of the effects of glacial ice. Rock types more resistant to glacial processes remain upstanding and less resistant rock materials have been more deeply eroded and removed. However, ice must be considered as a very viscous liquid, not only exerting its influence on the underlying topography, but also being influenced by that topography. Though ice crossed the area from generally north to south, the already upstanding diabase ridges that characterize the area will have deflected more recent ice events. As shown in R.P. Sharp's map (Fig. 2.7), glacial striae, the grooves ground into solid rock by the scraping action of sub-glacial debris, are found sub-parallel to the ridges rather than orthogonal to them (as in R5E) (Sharp, 1953).

The most effective ice erosion is performed early in ice advance. Later, under increasing thickness of ice, erosion of the underlying ground surface all but ceases. Shearing takes place between upper layers of ice sliding over lower layers of ice that are trapped within the topographic depressions. This trapped ice becomes largely stagnant and therefore ineffective at erosion. The topography of the Grand Portage area suggests that this scenario is quite likely to have taken place.

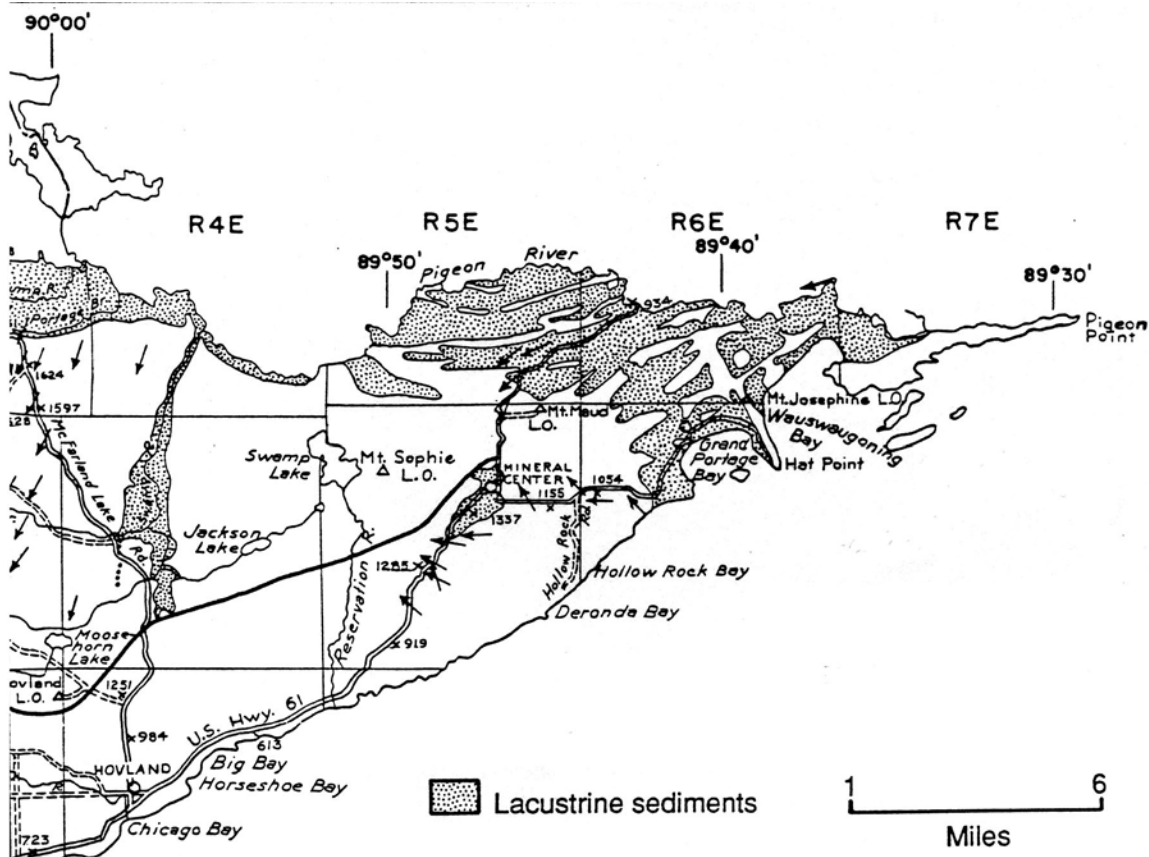


Figure 2.7. Part of R.P. Sharp's map of Cook County, MN (Sharp, 1953) showing glacial striae directions (arrows).

Sub-glacial water flow

It is likely that the topographic differentiation between the hard intrusive rocks and the less resistant sedimentary rocks began in pre-glacial time by the time-honored agents of runoff, wind and chemical weathering. At this time too, the conspicuous breaks in the ridge features are likely to have first been formed. A number of major faults, along which detectable vertical movement has taken place, cut across the area (Fig. 2.3), but many unmapped smaller faults and shatter zones also are present. These linear weaknesses in the hard rock materials are the first to be exploited by the agents of denudation, and once initiated, processes such as running water become focused in these locations, exacerbating the erosional effects.

The role of sub-glacial water flow was particularly important. Unrecognized by earlier workers studying glacial geomorphology, water flowing beneath and within ice is often under considerable hydraulic pressure. Unaffected by gravity, such flow can run counter to the slope or topographic grain of the underlying surface. The force of water, carrying subglacial debris and running at the interface of the ice and ground surface, is a highly effective erosive tool. Scouring of rock surfaces and the erosion of steep sided channels into the rock surface (called tunnel valleys) can take place. In post-glacial landscape glacial meltwater flow in these valleys might be seen to run uphill, cross divides or run contrary to the drainage pattern. Some are used by present drainage, but many remain dry or hold lakes and bogs. Associated with these sub-glacial features are eskers, which represent a final stage of deposition within the tunnels. These eskers might run along the valleys in places or climb over the margins onto ground alongside the main channel.

Isostatic uplift

Although the mechanism for post-glacial isostatic uplift or rebound remains inadequately explained by geophysics, the effect of this mechanism can readily be observed and measured. The process began as ice margins began to wane and withdraw, and continues to the present day. Greater uplift occurs where greater depression was made by a greater weight of ice. As a result, models for post-glacial rebound see greater and longer continuing uplift towards the center of the former ice mass.

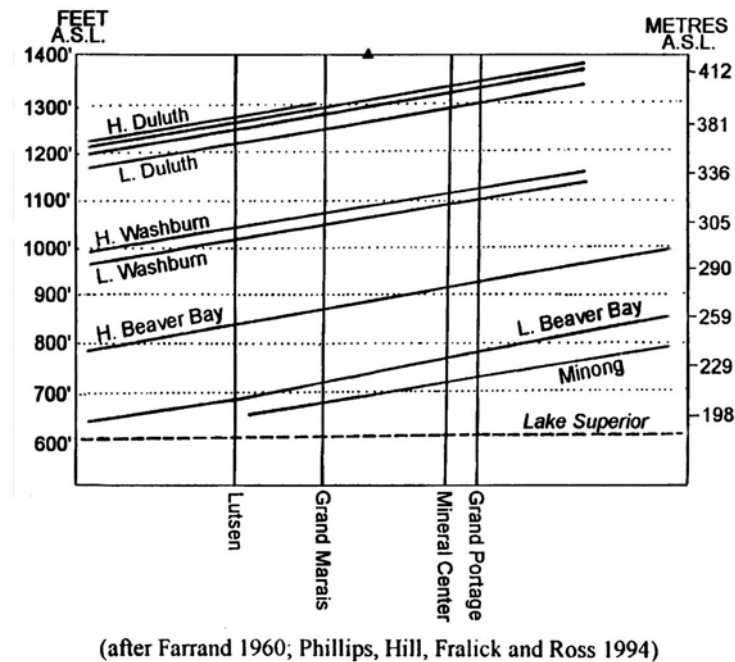


Figure 2.8. Shoreline diagram for the Grand Portage area showing present elevations of former lake shores.

Along the Minnesota north shore the axis of uplift is sub-parallel to the shore, and currently, uplift at the Pigeon River is about 250 feet greater than at Duluth. Consequently, an initially horizontal line, such as a lake shoreline, is now tilted such that its elevation rises over this range between Duluth and the Pigeon River. The sequence of lakes that occupied the Superior Basin as ice withdrew and isostatic uplift occurred resulted in a staircase of shorelines found along the north shore, diverging to the northeast (Fig. 2.8). Thus in Cook County, the vertical distance between these former shoreline features is greatest at Grand Portage. A more detailed account of pre-historic water levels can be found in Phillips (2001, pp.8-28).

Evidence of former lake margins

Typically, the presence of a lake margin, sustained for a period of time sufficient for waves to do work, is represented either by an erosional form such as a bluff, or by a depositional form such as a beach. In addition, rivers entering the lake body tend to deposit fluvial sediments in the form of deltas, or at least a series of river mouth bars that characteristically curve out of the river mouth and along the shore. Both features mark the position at which rivers formerly met with the shore of the lake.

Streams along the Minnesota north shore have had to respond to steady isostatic uplift and falling water levels. Typically, this has resulted in the lengthening of stream courses and the deep incision of their beds as they struggle to meet the lakeshore at grade. Not surprisingly, the north shore abounds in gorges, waterfalls and rapids as outcrops of more resistant rock inhibit equal downcutting along the stream course. Another characteristic feature that results from isostatic rebound is the building of deltas at successively lower elevations as the stream course extends, each delta in part built from the material eroded from an earlier one as a result of incision into the former delta surface.

The Grand Portage Trail crosses a number of these features, each denoting a period of lake level stability lengthy enough for a defined landform to be formed (Figs. 2.9, 2.10). The features have significance for finding Paleo-Indian artifacts, since it is established that early people tended to favor shoreline and river mouth locations for their temporary habitation and activities.

The succession of shorelines on the Grand Portage Trail

As Marquette ice, the youngest glacial lobe in the western Superior basin, melted back and exhumed the basin successively from southwest to northeast, the highest post-Marquette shorelines are those of Lake Duluth, the earliest phase of which was restricted to the Duluth area. When ice still covered the Thunder Bay region, Lake Duluth first extended across the international border into the southern part of the Nor'wester mountains. At Grand Portage, the high Duluth level was at 1340 feet and although Mt. Maud was now part of the mainland, no part of the Grand Portage Trail was above water.

Lake Name	Elev. Fasl.
Epi Duluth	1440
Duluth H	1340
Duluth M	1325
Duluth L	1290
Sub Duluth	1245
Highbridge	1200
Moquah	1150
Washburn H	1110
Washburn L	1080
Manitou H	1020
Manitou L	970
Beaver Bay H	920
Beaver Bay 2	890
Beaver Bay 3	845
Beaver Bay 4	810
Beaver Bay L	775
B Bay-Minong	750
Minong	715
Post-Minong 1	700
Post-Minong 2	675
Post-Minong 3	665
Post-Minong 4	650
Houghton	582
Nipissing	636
Algoma	621
Sault	610
Superior	602

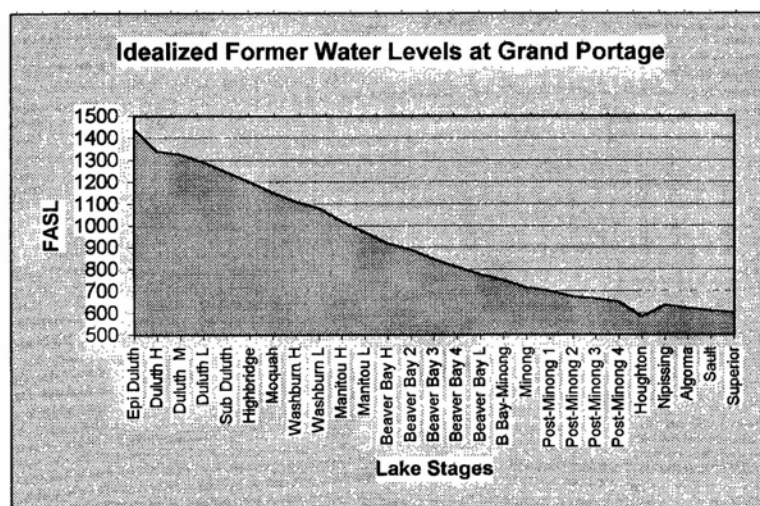


Figure 2.9. Idealized former water levels at Grand Portage.

The gradual withdrawal of ice uncovers recently glaciated ground. If, as in this case, the ground is lower than the ice marginal (pro-glacial) lake, it is immediately inundated and protected from exposure to weathering and erosion, even obtaining a modest cover of pro-glacial lake (glaciolacustrine) clays or silts if under water long enough for offshore sediment to accumulate.

Portions of the Grand Portage Trail first appeared above water during the later Low Lake Duluth phase (1290 feet). A good portion of the Trail north of Old Hwy 61 lay above water, including the uppermost part of the narrow ridge along which the Trail runs for a mile or more. This part of the Trail, having been inundated only briefly, is not overlain by lacustrine sediments. The thin till is wave washed, and bare rock is exposed in many places. A pattern can be seen emerging here. High areas, subject to inundation for a lesser period, are less likely to be covered by lacustrine sediments. Lower areas, subject to a long period under water may accumulate lacustrine sediments, though only if a source of such sediment is available. To a large extent, what happens is related to the degree of shelter or exposure to wave action. This will vary locally as the water level lowers. Sheltered areas may show little modification of the existing sediments, but exposed areas are likely to show winnowing of the glacial sediments, removal of finer particles and a lag of larger clasts left on the surface. Sufficient wave action may have time to sort and transport sediment, forming recognizable beach or bar features where longshore transport is consistent. Extreme wave action may have the energy to erode and form soft-sediment bluffs, or even undercut and fashion solid rock into wave-abraded forms. In summary, as the already complicated topography of the Grand Portage Trail area slowly emerged, an ever-changing pattern of wave refraction around the shoals, islands and peninsulas led to a localized sculpting of the topography. At the Lake Highbridge level (1200 feet), the present Grand Portage Trail was divided into two clearly different portions. The Trail above Old Hwy 61 was above water, while that below the highway was still inundated. This dual character is still very visible today. The character also occurs into the High Lake Washburn phase (1110 feet), with the steep section of the Trail immediately above Old Hwy 61 forming a fairly stable promontory, down which the water level fell in stages (Figure 2.10). Meanwhile, Mt. Josephine emerged, as did higher parts of the narrow diabase ridges, the links gradually joining as water level fell.

During the High Lake Washburn phase (Fig. 2.10A), two 'gaps' were noticeably still inundated. The eastern one marks the line of the Grand Portage fault, and the Grand Portage Trail passes through the western one. At this time, the Poplar Creek basin had become significantly protected from wave action, though, as in present Lake Superior, water flow through the two gaps was likely intensified by seiche and meteorological events, leading to possible scouring and some transfer of sediment to one or both sides of the 'channel'. The lakeward side of the emerging ridges probably experienced the full force of Lake Superior wave action, so the transfer was likely to have been mostly towards the Poplar Creek basin.

Within the relatively calm waters of the basin, lacustrine sediment would have accumulated liberally. It is during the High Lake Manitou level (1020 feet) (Fig. 2.10B) that the 'gaps' became dry. On both sides of the gaps, some crossing feature such as a beach or a bar was likely, though if earlier scoured to bedrock, little sediment might have been available to form these. If soft sediments formed a shoreline feature in such an emerged gap, it would be named a 'tombolo', in that it joined mainland to a former island. This is a feature only occasionally seen on the shore of present Lake Superior.

During the High Lake Manitou level, the basin of the Grand Portage Creek also became a more sheltered reentrant, and the upper part of the creek course probably flowed. At the subsequent High Lake Beaver Bay level (920 feet), a distinct Grand Portage Bay emerged and Grand Portage Creek was a sheltered river mouth extending far up valley. This character was less pronounced by the time of Mid Lake Beaver Bay (845 feet) (Fig. 2.10C), when the summit of Mt. Rose appeared as a small island, but the potential for river mouth habitation by Paleo-Indian people is good during the five Beaver Bay lake stages. As Mt. Rose became a larger island during the Lower Lake Beaver Bay stage (775 feet), it probably protected the river mouth and the potential for habitation was sustained. At this time, wave refraction around Mt. Rose would have been quite complex and the area immediately behind the steep island would have been very protected.

Lake Minong, dated as about 9,500 BP, is notable as the first of the succession of lakes to occupy the whole of the ice-free Superior basin, and its shoreline can be traced around the present lake. Lake

Minong time is also notable in the Minnesota-Ontario borderlands as a period in which evidence of Paleo-Indian activity became more pronounced than before. In the Grand Portage area, the Minong shoreline occurs at 715 feet. A defined bluff forms a deep 'V' shaped reentrant into the valley of the Grand Portage Creek, which the Grand Portage Trail now surmounts. At this time, Mt. Rose ceased to be an island and further contributed to the sheltered nature of Grand Portage Bay (Fig. 2.10D). The potential for Paleo-Indian occupation around this sheltered river mouth is high.

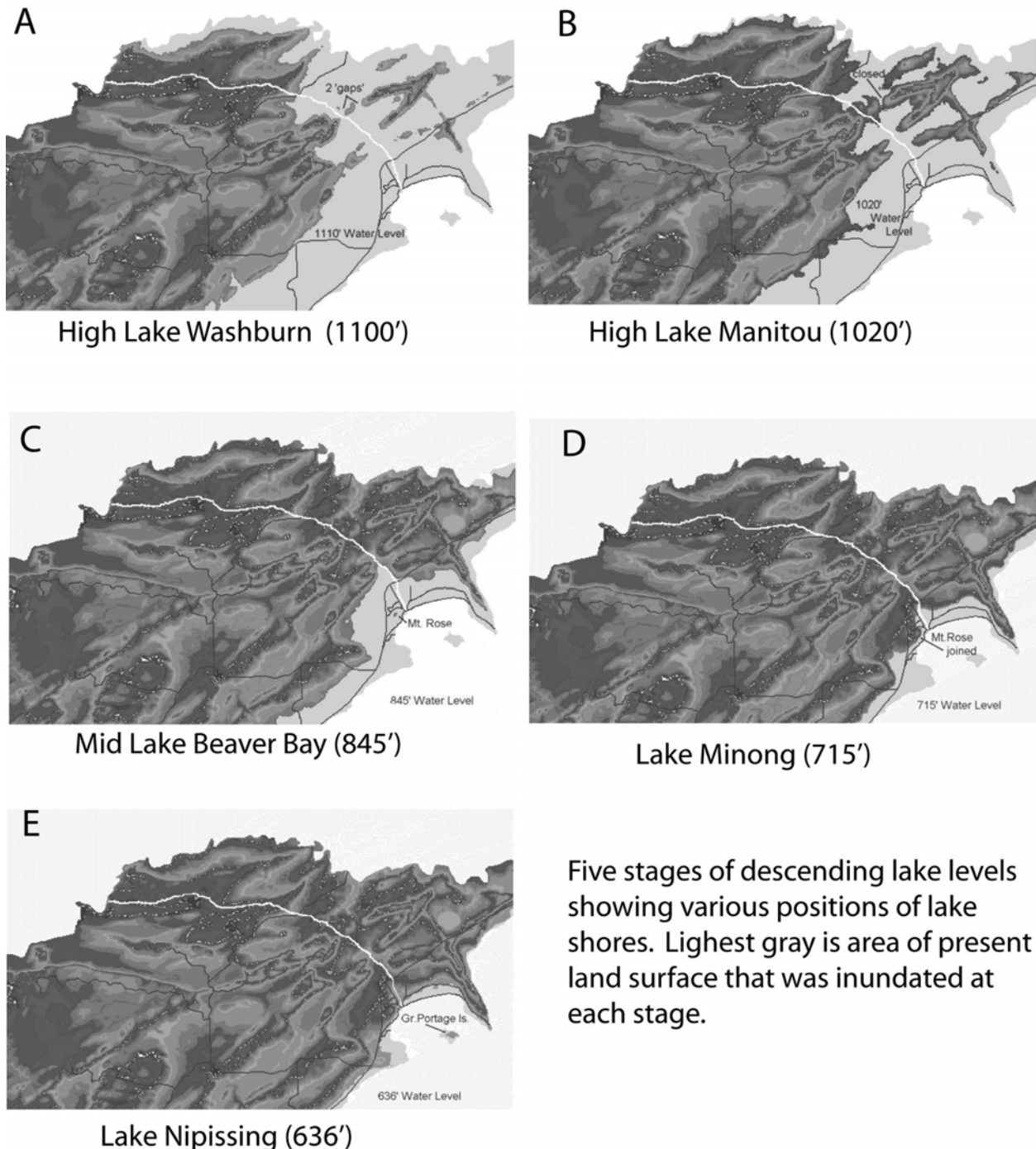


Figure 2.10. Reconstruction of former shorelines at five levels of post-glacial lakes in the Grand Portage region.

The Post-Minong period was a time when pro-glacial Lake Agassiz flowed through a series of eastern outlets into the Nipigon basin, and thence into Lake Superior and the lower Great Lakes. At least five episodes of catastrophic discharge have been recognized (Teller and Thorliefson, 1983). Each had the potential to briefly raise the level of the Post-Minong lakes, and erode the glacial sediments that swathed the St. Mary's river outlet to Lake Huron. Ultimately, the outlet was eroded to bedrock and the Lake Houghton phase was established as the lowest of the succession of post-glacial lakes in the Superior basin.

Water level declined to a low position of 582 feet in Grand Portage Bay, placing the shoreline adjacent to Grand Portage Island, which separated two shallow bays. Little if anything remains of this portion of history, because there followed a steady rise of water level in the Superior basin as the St. Mary's outlet was isostatically raised. This transgression drowned the recently exposed Post-Minong surface and culminated in the Nipissing shoreline, which in places truncates and reoccupies older shoreline features. In the Grand Portage area the Nipissing rose to 636 feet, preserving the Minong and the upper Post-Minong shorelines (Fig. 2.10E). Near Grand Marais, the Nipissing inherited the older Minong bluff, refreshing it, and further south the Minong was destroyed and drowned by the Nipissing transgression. Subsequently, water level fell to the Lake Algoma level (621 feet) and the Sault level (610 feet) before declining to present lake level (602 feet). Features formed by these more recent water levels run through the area of the Stockade, though more recent shore erosion has removed the lowest.

In summary, the present Grand Portage Trail traverses a landscape that has been fashioned first by ancient tectonic activity that emplaced the pattern of dikes, which are responsible for the basic topographic character of the area. Secondly, multiple glaciations have ground across the ridges and basins, steepening and smoothing rock surfaces, and depositing varying depths of sandy, coarse tills, particularly in lower areas. Thirdly, a succession of post-glacial lakes has fully submerged the course of the Trail. As the water level gradually declined, the Trail topography progressively emerged. These lakes have left some evidence of their brief periods of stillstand in the form of wave-washed sediments, erosional bluffs and constructional beaches and bars. They have also deposited a variable thickness of lacustrine clays and silts, particularly within the basins and valleys of the bedrock topography.

The geological underpinnings of human history of the Grand Portage region

Early human activity in any region is strongly influenced by physical barriers to transit and by the easiest solutions to the impediments that those barriers represent. Geological events that occurred from as little as ten thousand years ago to as much as 1.1 billion years ago largely predetermined the course of European history in the Grand Portage region because they controlled the physiography and the unique physiographic characteristics of Grand Portage that made it the obvious location for the "highway to the west". These events not only dictated that the portage was necessary, but determined the best route for the portage, and, on a broader scale, were responsible for the enormous advantage to westward travel inherent in making the portage.

European exploration and settlement of the continental interior was strongly guided by the relatively easy water access provided by the Great Lakes as far west as western Lake Superior. Access farther west into the deep interior of the northern Great Plains, Canadian prairies, and beyond was facilitated by additional relatively easy water routes up the Pigeon River, over the continental divide, and through the Rainy River system westward to major north-flowing drainages and their tributaries. The shortest distance from Lake Superior to waters navigable westward by canoe is over the Grand Portage. This short route to the west was known and used by Native Americans well before the first Europeans arrived in the early 1600's.

But the stretch of high ground from the lakeshore to the upper reaches of the Pigeon River was the most prominent physical barrier along the entire westward route from Montreal to the Rocky Mountains. In fact, the length and difficulty of the portage lead to a mutiny among the first westward-bound French exploration party that arrived at the site in 1731.

The portage was necessary because of a long series of rapids and falls along the lower reaches of the Pigeon River which falls about 660 feet from the upper end of the portage down to the level of Lake Superior. But a remarkable feature of this area is that once past the 8.5 mile portage, the way west was through the relatively flat water of the upper Pigeon River and a series of interconnected lakes. Within only about 40 miles from the west end of the portage one leaves the Lake Superior drainage, crosses the continental divide, and enters the Rainy Lake watershed which drains ultimately to Hudson Bay. Thus, three regional features: 1) Lake Superior and the easy route that it provides for water-borne commerce from the east, particularly Montreal, 2) the steep gradient of the Pigeon River in its lower reaches that made canoe transit impossible, and 3) the very small size of the Lake Superior watershed which, once exited, provided essentially unlimited canoe access to the interior of North America, all played fundamental roles in making the Grand Portage the most favored and easiest route for westward travel.

The present basin of Lake Superior is the exhumed sediment-filled central valley of an ancient rift that formed 1.1 billion years ago. At that time a great rift valley, known as the Midcontinent rift, began to split the North American continent. The rift was filled first with tens of kilometers of volcanic rocks, mostly basalt such as the Grand Portage basalts exposed just south of the Monument, and finally by many kilometers of sediments, now mostly red sandstones. Figure 11 shows the trace of the axis of the Midcontinent rift which faithfully follows the southward concave shape of the present lake. The sandstones in the central part of the rift presented an easily erodible substrate for glacial scouring in contrast to the much more durable igneous rocks along the rift flanks, which today stand as the prominent ridges and uplands of the region. As a result, repeated glaciations during the Pleistocene gouged the deep valley now occupied by Lake Superior. Thus, Lake Superior and its favorable transportation route owe their existence to these two ancient events of rifting and glaciation. But both of these events combined to create the unfavorable situation of the extensive series of rapids and waterfalls along the lower reaches of the Pigeon River, which required the arduous “grand portage” to circumvent them. The Pigeon, along with all other major rivers along the northwest shore of Lake Superior, reach the lake over a series of rapids and falls that extend nearly to the lake shore. This is in contrast to the more typical morphology of rivers which commonly have extended reaches of slack water upstream from their terminus and thus provide waterborne access well inland. It is again the repeated glacial scouring along the ancient rift axis that created this situation. Numerous lobes of ice flowed southwestward through the western Lake Superior basin and oversteepened the topographic gradient near the present lake shore. This steep gradient extends well below the present lake level into a bedrock trough whose bottom is more than 1500 feet below the lake and about 1000 feet below sea level. In the relatively short period since the last glaciation, about 10,000 years, the rivers have not had adequate time to modify this glacially induced gradient and thus plunge down this steep topography until they intersect the modern level of Lake Superior making upstream canoe travel from the river mouths impossible.

A still more fundamental geologic phenomenon is responsible for the highly unusual situation of a remarkably small watershed surrounding Lake Superior thus allowing a short distance of transit from the lake to outward flowing drainages. Modern interpretations of the geodynamic causes of the Midcontinent rift suggest that the region was underlain by a newly developing plume of upwelling mantle which caused both intense basaltic volcanism along the rift and the active splitting of the continent (Hutchinson and others, 1990; Cannon and Hinze, 1992). As the new plume waned in intensity, welding of cooling mantle material to the base of the crust, a so called rift pillow, resulted in an unusually thick crust in the Lake Superior region. Interpretation of seismic refraction surveys shows that present day crust beneath the lake is as much as 55 km thick in contrast to a typical thickness of about 40 km in surrounding regions (Halls,

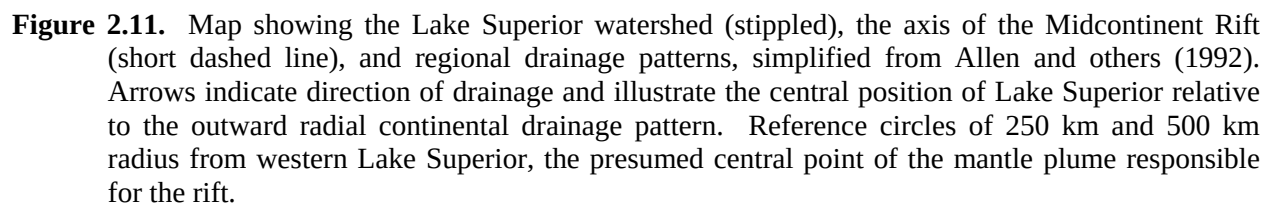
1982). The isostatic effect of this thick crust is to maintain the Lake Superior region as a subtle topographic high (Allen and others, 1992). Examination of the continental scale drainage patterns (Fig. 11) shows that the Lake Superior watershed is anomalously small in relation to the size of the lake, and that major river systems drain radially away from the lake as a result of this isostatic bulge. Thus Lake Superior is, in a sense, a hub from which numerous possible outward draining water routes are available once the Superior divide is crossed. The Grand Portage is the shortest of these available routes.

On a smaller scale, geologic features also played a significant role in determining the course of the portage and both its upper and lower terminus. Grand Portage Bay is the most sheltered harbor anywhere along the northwest shore of Lake Superior and thus was a naturally favorable location for developing the trading post at the western terminus of the Great Lakes transportation route. The bay is underlain by relatively non-resistant sedimentary rocks of the Rove Formation which make up most of the low country in the region. The sheltering headlands are formed from resistant igneous rocks. To the northeast, Hat Point and the high ridge extending inland is a reflection of the Mt. Josephine dike of the Pigeon River swarm. To the southwest, basalt flows of the Grand Portage Volcanics support the topographic high that forms the southern shore of the harbor. These same volcanic rocks also underlie Grand Portage Island that lies across the mouth of the bay and provides additional shelter.

In order to traverse westward from Grand Portage Bay a way needed to be found through the imposing steep-sided northeast-trending ridges that are supported by the resistant diabase of the Pigeon River dikes. A gap through these ridges is now followed by Grand Portage Creek, which flows along a fault trace (see Fig. 2.3). Although this fault has only minor offset, it apparently created weakened rocks that were either eroded by pre-glacial streams or were plucked by glacial ice or subglacial streams. In any case, this topographic gap is readily seen from Lake Superior and likely attracted the earliest humans in the area as a relatively easy course to the west.

At the western terminus of the portage, local geology also played a role in determining the most favorable location at which canoes could be easily embarked and disembarked. Fort Charlotte was established at a point where a diabase dike, probably of the Pigeon River swarm, intrudes the Rove Formation and trends across the Pigeon River. This resistant dike forms a small topographic rib which creates a small rapid on the river that could not be crossed by canoes and, perhaps more significantly, impounded an upstream reach of still and unusually deep water creating a small natural harbor.

These features illustrate the importance of the geologic history of the region and its influence on present physiography. It was both local and regional physiography that dictated the course of human occupation and exploitation of the Grand Portage region and its enormous significance in the 18th century history of the northern part of the continent. This “geological predetermination” of human events serves as a prime example of the interaction between the physical challenges encountered by early human exploration and settlement of a region and the ingenuity and industry with which they adapted to those challenges and opportunities.



GEOLOGICAL STOPS

The following are brief descriptions of specific geological features that can be observed in several different areas of the Grand Portage National Monument. Locations of stops are shown in Figure 2.1.

Stop 2.1. Stockade to Visitor Center: Grand Portage diabase dikes, Rove Formation, and glacial lake features.

Shoreline features of some of the lower post-glacial lakes are visible near the Stockade. A distinct feature that runs along the shore zone and through the Stockade is the gentle bluff of the Sault stage (610 feet). To the east of the Grand Portage Creek, the bluff curves lakeward on Premiers Point, and runs northeast through the picnic area, parallel to the shore until rising erosional bluffs of Lake Superior truncate and eliminate it. Offshore, a lake floor of large boulders that fringes the east side of the sandy creek channel indicates the former extension of Premiers Point. Just above the Sault shoreline lies another curving bluff, the Algoma (621 feet). To the west of Grand Portage Creek, the Sault bluff runs through the Stockade enclosure. The Great Hall building is constructed on the bluff, such that the lakeward side lies at the foot and the inland side lies on the top. The bluff forms a marked step through the Ojibwe Village exhibit to the west of the Stockade and is then truncated again by rising erosional bluffs near the Visitor Center. Behind the Canoe Warehouse in the Ojibwe Village exhibit, a small section of Algoma bluff has survived the disturbance of road construction.

As you walk the roadway between the stockade and visitor center (under construction in 2007) you will cross four diabase dikes of the Grand Portage swarm (Green and others, 1987). A fifth dike crops out just south of the Monument boundary and is especially prominent as a free-standing wall of diabase about 20 feet tall, a classic geomorphic dike form. Figure 5 compares a recent photograph of this dike with an etching from Owen's 1852 General Land Office report. All of the dikes form linear ridges up to a few tens of feet wide as a result of being much more resistant to erosion than the Rove shale into which they are intruded. The topography created by the dikes was probably emphasized by wave erosion along the glacial Lake Nipissing shore when water level was a few tens of feet higher than present Lake Superior. Pocket beaches were probably formed on the Rove Formation between the resistant dikes although talus from the steep south flank of Mt. Rose now covers these areas.

Individual dikes here are 10-20 feet thick, fine grained, massive, and commonly show chilled contacts against the Rove Formation. The dikes were formed during early stages of development of the Midcontinent rift. In addition to the Rove, they also intrude volcanic flows of the Grand Portage Volcanics to the southwest of the Monument. The dikes are rather strongly magnetized and have a reversed remanent polarity. This makes them easy to find and identify by magnetic surveying because they appear as linear negative anomalies. A magnetic profile measured along the road with a proton precession magnetometer in 2001 shows four distinct magnetic lows that correspond to the four dikes that cross the road (Fig. 2.6). A profile along the portage trail, also done in 2001, found several additional dikes that had not been known previously.

The dikes can be dated rather precisely by a combination of geologic relationships and magnetic polarity. They cut the Grand Portage Volcanics that were extruded early in the history of the Midcontinent rift and acquired their remanent magnetization before the magnetic reversal, which is recognized widely in igneous rocks of the Midcontinent rift, at about 1,100 Ma.

Stop 2.2. Rove Formation and Pigeon River diabase on Mt. Rose

The Mt. Rose Trail provides a well exposed section of the Rove Formation. The summit is underlain by a diabase dike of the Pigeon River swarm and a contact metamorphic halo in the Rove near the diabase contact. The following descriptions are keyed to numbered placards along the trail. Much of the east-facing flank of Mt. Rose, over which the trail climbs, consists of steep bluffs of bedrock exposures of the Rove Formation. The steepness is probably due to wave erosion during higher level stands of post-glacial lakes. The Nipissing transgression reoccupied the base of Mt. Rose around 636 feet, with wave action vigorously cutting a bluff seen from Stop 2.1 on the Mt. Rose Nature Trail. The base of the bluff is now mostly covered by more recent talus. Cut in shale, no abrasion notch is expected, but an angular undercut can be seen in places to the south of the Trail.

The Trail follows a bench at approximately 678 feet between lookout Stops 5 and 6 that faces Lake Superior and which would have experienced strong wave action at the post-Minong water levels. Large accumulations of blocky talus below a sharp bluff between Mt. Rose Nature Trail Stops 7 and 10 represent the work of waves during the period of Lake Minong (715 feet). In several places, a convincing undercut is seen, though the base of the rock bluff is almost everywhere covered by more recent talus accumulation.

At 775 feet, at the time of Lake Low Beaver Bay, Mt. Rose was an island separated from the mainland by a shallow strait. On the Mt. Rose Nature Trail, there are one or two rock benches that could have been wave modified by this level of Beaver Bay, but the strata outcrops in nearly horizontal form and it is not possible to distinguish between wave formed features and structural control.

The Rove Formation along the trail is mostly flaggy sandstone in which prominent bedding partings emphasize bedding at inch-scale. Cross bedding is seen rarely and a few beds show soft-sediment slump structures. The sandy units are separated by thinner units of finer-grained fissile argillite. Beds are nearly flat lying with dips of only a few degrees toward the southwest. Within 10 or 20 feet of the contact with the diabase, the Rove is distinctly baked in the contact aureole. Closest to the contact, bedding is largely obliterated and the sediments are recrystallized to a hornfels containing biotite and clots of a secondary mineral in the matrix that is tentatively identified in thin section as cordierite.

The diabase is mostly medium-grained and massive. The nearly vertical southeastern contact can be seen near the top of steep bluff as well as in pavement outcrops near the summit. The rock seen here is typical of the Pigeon River diabase dikes in the region in that it forms dikes hundreds of feet thick, has normal magnetic polarity and forms the prominent high, steep-sided ridges that are the most distinctive elements of the local topography.

Stop 2.3. Rove Formation and Grand Portage dikes along the Beaver Bay shoreline.

At this stop, both bedrock and post-glacial lake features are displayed. On the north side of Highway 61 at the trail crossing there are extensive exposures of the Rove Formation and two diabase dikes of the Grand Portage swarm. The unusual amount of bedrock exposure is probably a result of wave washing along the coastline of Beaver Bay stage lakes. This shoreline erosion stripped the cover of glacial till or lake sediment that probably originally mantled the bedrock, as it does over most of the surrounding area. The Rove in this area is fine-grained gray sandstone that breaks into approximately inch-thick flaggy slabs. Bedding dips about 5 degrees to the southeast, typical of the Rove over most of the region. Vertical diabase dikes of the Grand Portage swarm can be traced across the outcrop area and into the roadcut exposures along Highway 61.

Just north of the bedrock outcrop, the Grand Portage Trail follows a terrace along the 816-foot contour, the inner edge of which is bounded by an indistinct bluff at 828 feet. The surface of the Trail is

noticeably roughened with large boulders, suggesting scour and winnowing of the finer sediments. It seems likely that this remnant terrace represents the floor immediately within the river mouth as it was in Mid Beaver Bay time (Fig. 2.12). As it is traced up valley, the bluff rises to 831 feet and 834 feet as would be expected within the river mouth. The Trail drops down a bluff from the 816 foot level into a small boggy embayment. This represents the shore of the 4th level of Beaver Bay (810 feet, between Mid and Low) and is a strong feature. The Trail rises briefly again to the former terrace (now 822 feet) and then descends into a broad embayment, backed by the continuation of the same 4th level of Beaver Bay before rising again to the terrace, (now 834 feet). A tributary coming in from the east breaks the bluff.

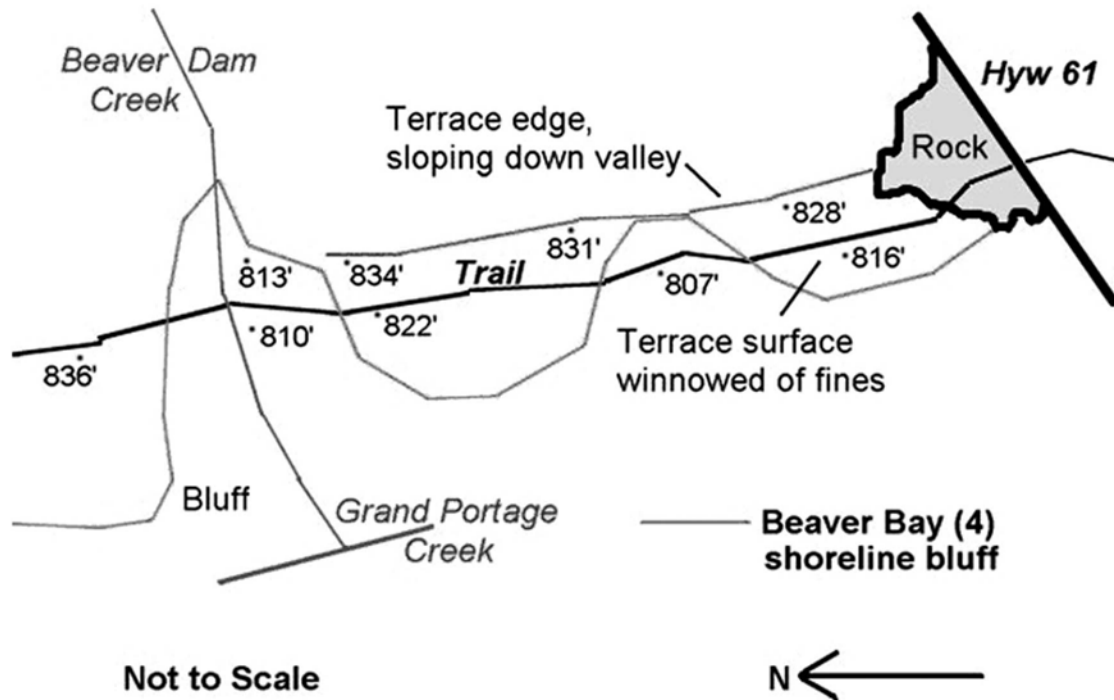


Figure 2.12. Sketch map of the area along Grand Portage Trail north of Highway 61 showing outcrop of the Rove Formation and the trace of the bluff of the Beaver Bay stage shoreline.

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FIELD TRIP 3

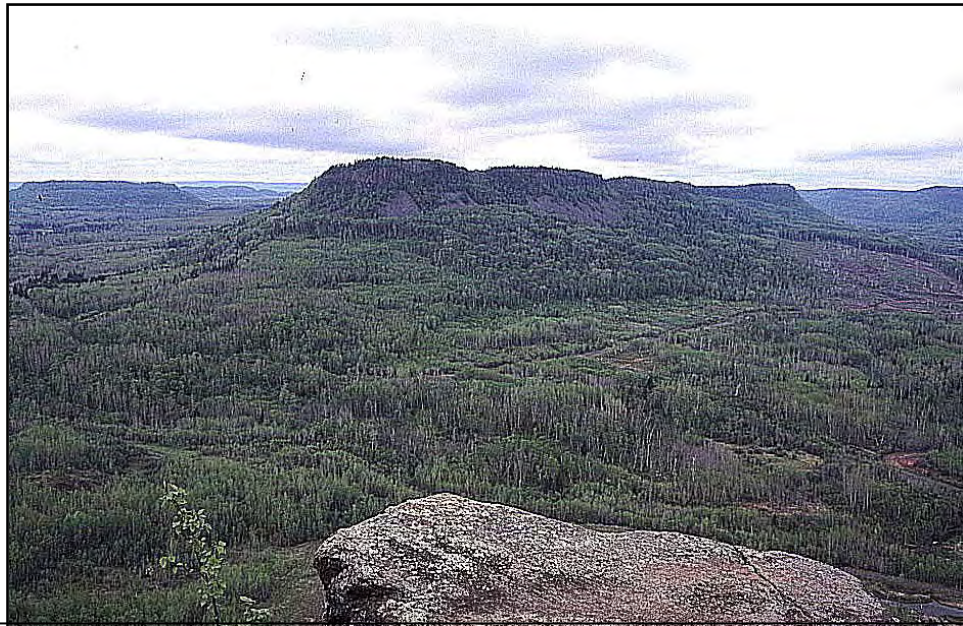
MIDCONTINENT RIFT-RELATED MAFIC INTRUSIONS NORTH OF THE INTERNATIONAL BORDER

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Frontispiece: View looking southwest from the top of ridge (Crystal Lake Gabbro) at the Great Lakes Nickel property. Mesas are underlain by Logan diabase sills which have intruded Rove Formation sedimentary rocks.

INTRODUCTION

This field trip covers an area that has been the focus of much research and economic interest. The guidebook has benefited from recent geochronologic and geochemical studies conducted as part of the Lake Nipigon Region Geoscience Initiative. These studies have elucidated the nature of magmatism in the northern part of the Midcontinent Rift (MCR) and have been used to augment and refine our previous understanding of these magmatic events. This trip focuses on a variety of mafic intrusions associated with the MCR north of the International Border near Pigeon River. These intrusions represent changes in the nature of early to mid-stage MCR magmatism over a span of ~20 million years. They include Logan sills, Pigeon River and other dykes and the layered Crystal Lake Gabbro, host to the Great Lakes Nickel copper-nickel deposit. Contacts with Paleoproterozoic Rove Formation sedimentary rocks are well-exposed in this area and illustrate some of the mechanisms of dyke/sill emplacement, as well as magma-wallrock interactions. These interactions may play an important role in local metallogenesis.

This guide book builds upon those previously written and compiled by Franklin and Kustra (1972), Miller and Smyk (1995), Parker et al. (2001) and Miller et al. (2002). Bear in mind that when visiting exploration or private properties, permission must be granted by the property owner. Current ownership information can be obtained from the Resident Geologist's Office, Ontario Geological Survey, in Thunder Bay. Please exercise caution along highway right-of-ways, cliffs and mine workings.

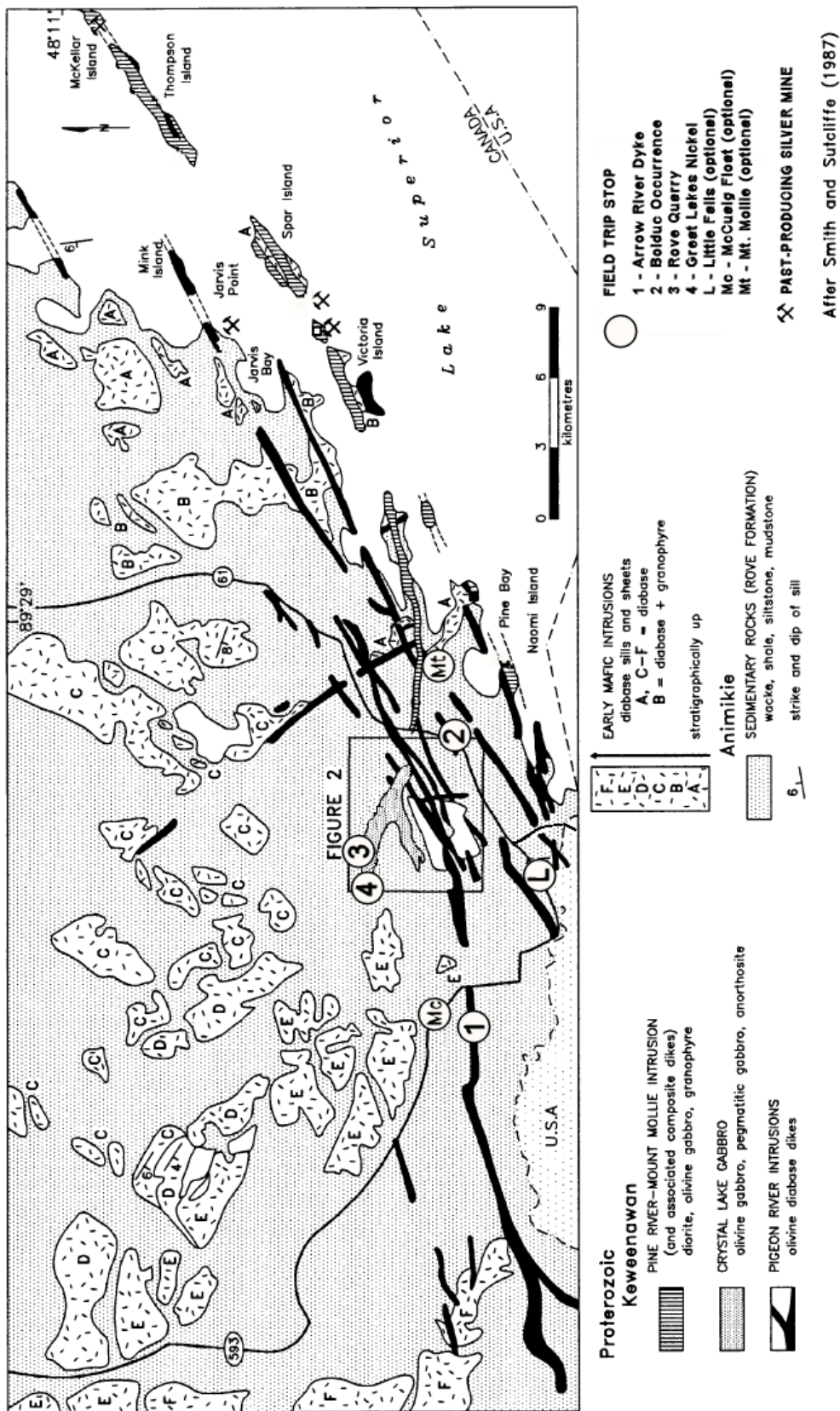
REGIONAL GEOLOGY

Situated within the Southern Province of the Canadian Shield, the field trip area is dominantly underlain by Paleoproterozoic Rove Formation clastic sedimentary rocks (Animikie Group) which have been intruded by MCR mafic intrusions (Figure 3-1). Previous mapping has been conducted by Tanton (1936a,b), Geul (1970, 1973) and Smith and Sutcliffe (1989). Detailed mapping, geophysical surveys and diamond drilling undertaken by exploration companies have provided additional detail and much-needed information about sub-surface geology.

This area is a rugged, upland area of diabase-capped mesas and ridges that occupies a 70 km by 30 km, northeast-trending topographic feature between Thunder Bay and the Minnesota border, termed the "Logan Basin" by North (2000). Logan Sills underlie mesas that commonly rise 150 m above valleys underlain by deeply eroded, flat-lying, Rove Formation sedimentary rocks (see Frontispiece). Northwest of the Logan Basin, Archean granitoid rocks of the Superior Province form low, rolling hills. Southeast of the Logan Basin the topography is dominated by northeast-trending, linear ridges underlain by Pigeon River dykes.

Animikie Group

At approximately 2.45 Ga a rift to passive margin developed along the southern edge of the Superior Province when a land-mass to the south separated (Fralick and Miall 1989). Later closure of the resultant ocean led to the deposition of the Animikie and North Range Groups as a backarc basin developed (Pufahl and Fralick 1995, 2004; Hemming et al. 1995; Van Wyck and Johnson 1997; Pufahl et al. 2000), which collision later transformed into a foreland setting (Hoffman 1987; Southwick and Morey 1991; Hemming et al. 1995; Ojakangas et al. 2001; Maric and Fralick 2005).



The Animikie Group in the study area is represented by the Rove Formation, which overlies the Gunflint formation and has an approximate thickness of 500 to 600 m south and east of Thunder Bay, thickening to the south. Rocks of the Rove Formation are flat-lying or dip gently to the southeast. Recent work by Amurawaiye (2001) and Maric and Fralick (2005) described a submarine ramp system in which the movement of coarse sediments into the deeper parts of the basin was mainly through the action of low- and high-density turbidity currents. Fair-weather and storm-generated currents dominated depositional activity at the edge of the basin. Mud and ash particles accumulated during tranquil periods (*ibid*). Amurawaiye (2001) has stated that approximately 70% of the Rove Formation locally consists of organic shale whose hydrocarbon contents have now been degraded.

The lower 100 to 150 m of the Rove Formation and the correlative Virginia Formation in Minnesota consist of alternating shale-siltstone and black, pyritiferous shale successions, probably reflecting fluctuations in sea level (Maric and Fralick 2005). These successions, and especially the upper black shale, likely represent a major condensed interval deposited in water ~100 to 200 m deep. Lucente and Morey (1983) ascribed sedimentation of this interval to pelagic rainout of fine-grained sediment from dilute suspension or hemipelagic processes involving diffuse turbidity currents. The presence of abundant, sub-millimeter rip-up intraclasts also denotes the operation of sporadic bottom currents (Maric and Fralick 2005). Tidal deposits present in correlative rocks to the south of Lake Superior (Ojakangas et al. 2001) confirm open connection to the ocean. Above the upper, pure black shale interval, graded fine-grained sandstones are organized into a coarsening-upward succession approximately 100 m thick that is transitional into 400 m of medium-grained, sandstone-dominated, stacked parasequences (Maric and Fralick 2005). This is overlain by lenticular to wavy bedded sandstones and shales with both wave and current ripples. The coarsening-upward to sandstone-dominated portion of the Virginia and Rove Formations has been interpreted as a submarine fan (Lucente and Morey 1983, Maric and Fralick 2005) with the uppermost ripple laminated succession representing progradation of distal distributary mouth bars of a delta (Maric and Fralick 2005). A sandstone sample from the submarine fan portion of the succession yielded a youngest U-Pb detrital zircon age of approximately 1780 Ma (Heaman and Easton 2006). The predominantly Paleoproterozoic zircon population and paleocurrents indicating sediment derivation from the north (Morey 1973), strongly suggest the Trans-Hudson Orogen was the source of the detritus.

Keweenawan Supergroup

Mesoproterozoic intrusive, volcanic and minor sedimentary rocks associated with the MCR collectively constitute the Keweenawan Supergroup. On the northern margin of the MCR, Keweenawan rocks include a variety of intrusive rocks and Osler Group volcanic rocks which represent some of the earliest magmatism in the MCR. As shown in Table 3-1, ages range from *ca.* 1140 Ma (Heaman and Easton 2006) to ages younger than the magnetic polarity reversal that occurred between 1105 and 1102 Ma (Davis and Green 1997). A tabulated synopsis is provided below; bolded units occur within the field trip area.

Most mafic and ultramafic rocks in the Lake Nipigon and northern Lake Superior areas, including the Nipigon and Logan sills, appear to have been emplaced in a short, magnetically reversed, interval between *ca.* 1115 and 1100 Ma (Heaman and Easton 2006). Emplacement of alkalic intrusions, such as the 1108 Ma Coldwell Complex (Heaman and Machado 1992), and filling of much of the submerged part of the rift in Lake Superior, also occurred in this period. This was followed by a period of magnetically normal, waning mafic and felsic magmatism, between 1096 and 1085 Ma, that is preserved mainly along the Lake Superior shore by units such as the Crystal Lake (1099±1 Ma), Moss Lake (1095±2 Ma) and Blake (1095±2 Ma) gabbros, and a Pigeon River dyke near Arrow River (1093±3 Ma; Heaman and Easton 2006).

Table 3-1: Geochronology data of MCR-related rocks in Northwestern Ontario.

Lithologic Unit	Locality / Age (Ma)	Reference(s)
St. Ignace Island Complex gabbro	St. Ignace Island / 1089.2 $\pm$ 3.2	Smyk et al. (2006)
Arrow River Dyke	Arrow River / 1078 $\pm$ 3	Heaman and Easton (2006)
Pigeon River Dyke	Rita Bolduc / 1141 $\pm$ 20	Heaman and Easton (2007)
Blake Gabbro	Blake Township / 1091.0 $\pm$ 4.5	Heaman and Easton (2006)
Moss Lake Gabbro	Black Bay Peninsula / 1094.7 $\pm$ 3.1	Heaman and Easton (2006)
Crystal Lake Gabbro	Great Lakes Nickel / 1099.6 $\pm$ 1.2	Heaman and Easton (2006)
Pine Point – Mt. Mollie dyke	n/a	
Osler Group rhyolite (central suite)	Agate Point / 1105 $\pm$ 2	Davis and Green (1997)
Osler Group rhyolite (lower suite)	Black Bay Peninsula / 1107.4 $\pm$ 4/-2	Davis and Sutcliffe (1985)
St. Ignace Island Complex Rhyolite	St. Ignace Island / 1107.2 $\pm$ 2.4	Smyk et al. (2006)
Coldwell Complex	Coldwell Complex / 1108 $\pm$ 1	Heaman and Machado (1987)
Logan Sills	Mt. McKay / 1115 $\pm$ 1	Heaman and Easton (2006)
Nipigon Sills	Nipigon Embayment / 1114-1110	Heaman and Easton (2006)
Ultramafic Intrusions	Nipigon Embayment / 1124-1113	Heaman and Easton (2006)
Pigeon River	Crooks Twp. / 1141 $\pm$ 20	Heaman and Easton (2006)
Inspiration Sill	Lake Nipigon / 1141 $\pm$ 20	Heaman and Easton (2006)
Marathon lamprophyre dykes	McKellar Harbour / 1145 $\pm$ 15/-10	Queen et al. (1996)

Hypabyssal Mafic Rocks

Diabase sills, extending from the vicinity of Thunder Bay to east of Lake Nipigon, represent the northern remnants of the Midcontinent Rift, and have previously been referred to as the Logan sills (Stockwell et al. 1972), however recent work suggests a geochemical difference between the sills to the north and south of the City of Thunder Bay (Hart 2003; Hart et al. 2005b). Hollings et al. (2007) proposed that the term Logan Igneous Suite, which would fall within the Midcontinent Rift Intrusive Supersuite of Miller et al. (2002), should be applied to all the diabase sills in the area north of Lake Superior, with subdivision into the informal terms, Nipigon sills for the sills north of Thunder Bay, and Logan sills to the south.

Logan sills generally consist of fine- to coarse-grained, ophitic to intergranular, quartz tholeiitic diabase/gabbro (Smith and Sutcliffe 1987; Geul 1970, 1973). Coarse-grained, intergranular gabbro, locally rich in granophyric mesostasis, is common in the interior of the thicker sills. The upper sections of the diabase sills are commonly plagioclase-porphyritic, containing as much as 60% phenocrysts. Chilled margin and bulk compositions are iron-rich, quartz-tholeiitic basalt. Compositional and textural variation in sills has been noted by North (2000) and Beskar (2001) in Blake Township, where vari-textured, "taxitic" gabbro has been described.

Logan sills are recognized by their reversed magnetic polarity and generally take the form of columnar-jointed, thick sheets and sills whose geometry is strongly controlled by the subhorizontal bedding of the country rock. They form conspicuous erosional remnants that create mesa and cuesta topography. From the international boundary area to Thunder Bay as many as six diabase sheets were emplaced nearly conformably into Animikie sedimentary rocks (Weiblen et al., 1972; Smith and Sutcliffe 1987, 1989). Diamond drilling has also shown that stacked sills exist in the subsurface. For example, Dumont Nickel Inc. reported intersecting 14 gabbroic sills in a 705 m deep drill hole in central Pardee Township (Assessment Files, Thunder Bay South Resident Geologist's District, Thunder Bay). North of the border,

Smith and Sutcliffe (1989) reported sills up to 50 m thick, whereas in Minnesota, Jones (1984) studied four sills ranging from 50 to 160 m in maximum thickness. Rare exposures of feeder dykes to sills and preserved sill terminations have been noted (e.g., Stop 3-3).

The textural stratigraphy, which varies from a lower, ophitic zone to an upper pegmatitic zone, indicates that in most cases, the sills cooled as single units, probably over a period of 200 to 500 years (Smith and Sutcliffe 1989). Chilled contact zones are developed against sedimentary country rocks; sedimentary xenoliths are rare.

Pigeon River dykes trend east-northeast to northeast and dip steeply to the southeast (Geul 1970, 1973; Smith and Sutcliffe 1989). Displacement and warping of the Rove Formation is evident along many of the dikes. Composite intrusions are noted in several dykes (add reference). Dyke widths average between 50 and 70 m, but may be as much as 150 m across in Ontario (Smith and Sutcliffe 1987) and 500 m in Minnesota (Green et al. 1987). Forming northeast-trending, linear ridges, dykes can be traced semi-continuously for 15 km along strike. As noted by many workers, some dykes clearly crosscut Logan sills. However, Geul (1973) and Smith and Sutcliffe (1987) noted that others display somewhat ambiguous crosscutting relationships. In these latter cases, dykes may appear to merge with sills, suggesting that they were contemporaneous or that sills impeded the upward migration of the dykes. The presence of multiple sets of horizontal columnar jointing suggests the development of multiple or composite dykes. The dykes typically consist of ophitic diabase that may be plagioclase-porphyritic. A typical, non-porphyritic olivine diabase consists of approximately 60% plagioclase (zoned labradorite; An₅₅₋₇₀), 20% augite ± hypersthene, up to 15% olivine and up to 5% magnetite, ilmeno-magnetite and sulphides (Geul 1970, 1973). Average whole rock compositions of Pigeon River dikes are moderately evolved (mg# = 52) olivine tholeiitic basalt.

A 15-km long, northwest-trending diabase dike, termed the **Arrow River dyke** by Smith and Sutcliffe (1989) crosscuts Pigeon River dikes in Ontario. This dike and two shorter, similarly oriented dikes are composed of intergranular, quartz diabase that is commonly plagioclase-phyric. They are collectively known as the Arrow River dykes, despite the fact that they occur well north of the Arrow River. ***The authors recommend that they be named Cloud River dykes.*** Confusion has arisen because the term “Arrow River dyke” has also been used colloquially to denote a prominent Pigeon River dyke that occurs between the Arrow River and Pigeon River in southern Devon Township (i.e. Stop 3-1).

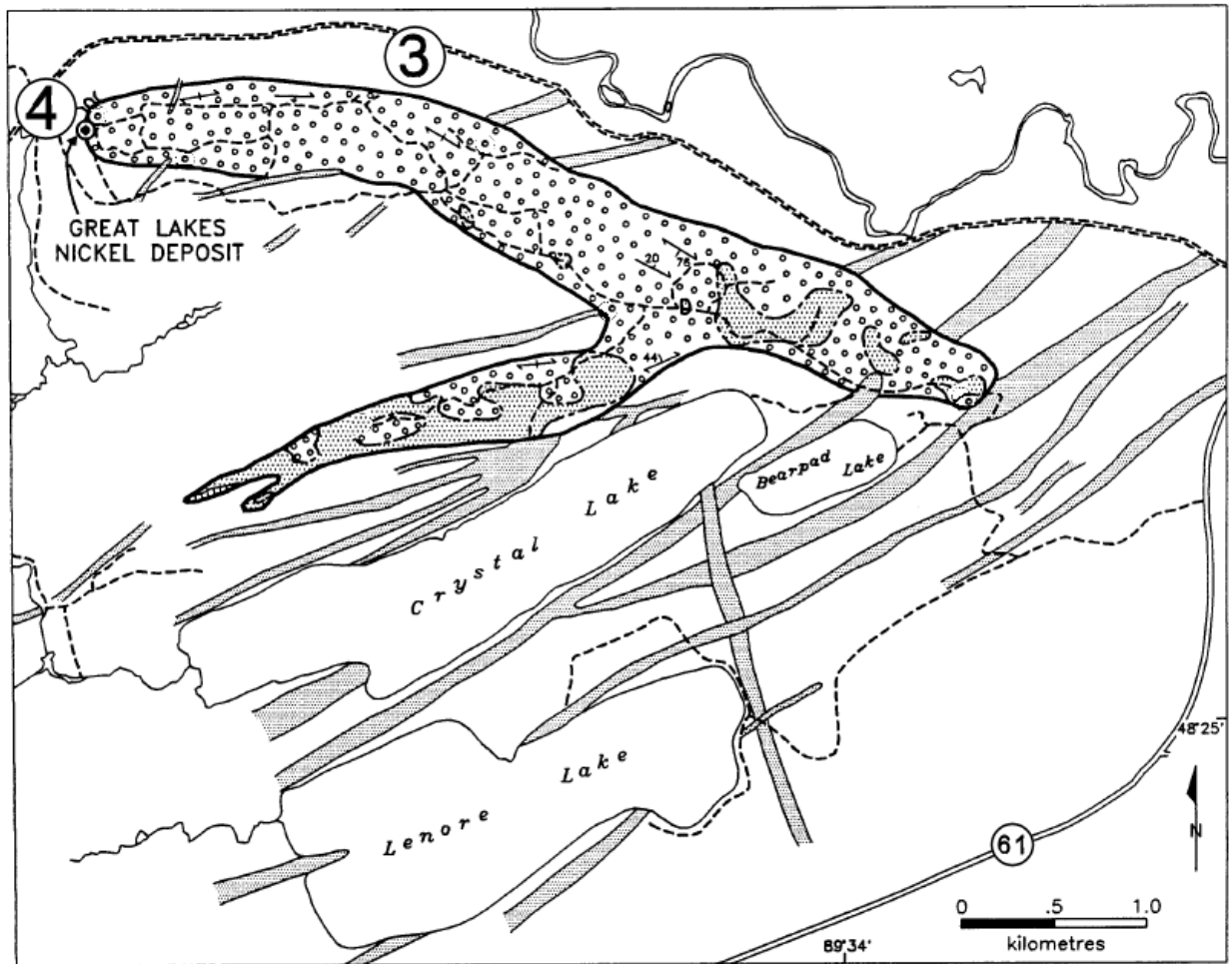
The youngest intrusions in the area tend to be more irregularly shaped and internally zoned. One of these is the **Crystal Lake gabbro**, which is Y-shaped in plan view, with a west-northwest-striking limb 5km long and an east-northeast-striking, southern limb 2.75 km long (Figure 3-2). Internal layering and foliation suggest that the surface geometry of the northern limb may result from the tilting of a canoe-shaped body, open on its western end (Smith and Sutcliffe 1987, 1989). The intrusion was subdivided into three major zones by Reeve (1969) and Geul (1970) and has been further subdivided by Smith and Sutcliffe (1987, 1989) and Cogulu (1990) into four major, roughly equivalent, lithologic zones:

- (1) an **upper zone** (60 to 80 m) of sulfide-barren troctolite, olivine gabbro and anorthositic gabbro;
- (2) a **middle zone** (30 to 42 m) of cyclic, layered anorthositic and olivine gabbro, Cr-spinel-bearing anorthosite, and olivine gabbro (Figures 3-3 and 3-4);
- (3) a **lower, unlayered zone** (50 m) of vari-textured gabbro and leucotroctolite, which hosts the bulk of the Cu-Ni-sulfide deposit; and
- (4) a **basal zone** (1 to 7 m) of fine-grained, chilled melagabbro and hornfelsed country-rock xenoliths.

Table 3-2: Geochemical analyses of select intrusive rocks from the Border Area

Intrusive Unit:	<u>Logan Sills</u>		<u>Grand Portage</u> <u>dikes</u>	<u>Mt Josephine</u>	<u>Pigeon River</u> <u>dikes</u>
Source:	A	B	C	D	C
Description:	Chilled Margin	Bulk Comp	Avg. of 17 dikes	Composite dike	Average of 14 dikes
SiO ₂	49	50.1	52.5	47.9	48.9
TiO ₂	3.4	3.6	2.48	3.66	1.65
Al ₂ O ₃	13.1	13.1	13.4	11.8	16.3
FeO ^t	15.6	14.3	13.3	16.2	11.5
MnO	0.22	0.16	0.19	0.24	0.18
MgO	5.6	3.9	3.68	4.04	6.46
CaO	7.45	7.2	6.72	8.9	10.21
Na ₂ O	2.52	3.4	3.17	2.63	2.42
K ₂ O	1.16	1.5	1.79	1.29	0.52
P ₂ O ₅	0.38	0.43	0.48	0.53	0.18
Volatiles	0.92	2.5	-----	-----	-----
Total	99.35	100.19	99.11	99.6	99.8
mg#	38.4	32.2	34.1	33	50.1
<u>Trace Elements (ppm)</u>					
Cr	50	---	44	50	100
Ni	80	---	61	22	136
Rb	---	---	55	28	15.7
Sr	700	---	420	359	280
Zr	250	---	281	280	---
Hf	---	---	9	---	2.94

Table from Miller and Smyk (1995); Sources: (A) Geul (1970), Table 3, Sample 2; (B) Jones (1984), Table A-I, Sill A; (C) Green *et al.* (1987), Table II; (D) Green (1986), Appendix E, Sample GP-60.



Middle Proterozoic

Crystal Lake Gabbro

- olivine gabbro, troctolite
- gabbro, leucogabbro, pegmatitic gabbro, anorthosite
- diorite
- chilled gabbro

Pigeon River Dikes

- olivine diabase

Early Proterozoic

Rove Formation

- argillite, wacke

- layering (strike dip)
- foliation (strike dip)
- lithological contact
- adit
- roads



FIELD TRIP STOP

- 3 - Rove Quarry
- 4 - Great Lakes Nickel

After Smith and Sutcliffe (1987)



Figure 3-3: Polished slab of massive Cr-spinel layers in contact with medium-grained gabbro, Great Lakes Nickel property (Stop 3-4). White bar is 4 cm long.

Figure 3-4: Igneous layering in the Crystal Lake Gabbro, exposed in the upper cliff section on the Great Lakes Nickel property (Stop 3-4). This is part of the middle zone of the intrusion, characterized by cyclic, layered anorthositic and olivine gabbro, Cr-spinel-bearing anorthosite, and olivine gabbro.



Projecting east of the Crystal Lake gabbro and arcing to the northeast to form a string of islands in Lake Superior is a major composite dike called the **Pine River-Mount Mollie** dyke (Geul 1973). Smith and Sutcliffe (1989) described this 35 km-long body as consisting of gabbroic, dioritic and granophyric rocks. Gabbroic rocks in the margins of the dike commonly have modal layering and sulphide mineralization similar to that in the Crystal Lake gabbro. Inward from the gabbro, quartz-bearing diorite cores much of the dike. In the central and western portions, granophyre occupies the core of the dike and shows contact relationships with diorite that indicates mixing of felsic and mafic liquids (Smith and Sutcliffe 1987, 1989).

As described by Geul (1973), the Pine River-Mount Mollie Gabbro comprises two segments of an east-striking intrusion which appears to postdate the Pigeon River dykes. The poorly exposed Pine River segment apparently connects to the eastern end of the Crystal Lake Gabbro. Magnetometer survey data and drilling indicate that the Pine River and Mount Mollie segments are connected beneath deep overburden. Field relationships and diamond drilling show that the dip of the gabbro contacts varies from near vertical to 35° north, indicating that the body has the shape of a tilted, truncated cone in cross-section. The exposed width ranges from approximately 60 m (true thickness) to 300 m (apparent thickness), depending upon the attitudes of the contacts. The gabbro has a distinct mottled appearance and is further distinguished from most other diabases in the area by larger grain size, leucogabbroic composition, and degree of mineralization. Diamond drilling logs summarized by Geul (1973) defined a crude zonation in the gabbro, based on variations in grain size, feldspar (i.e. anorthosite) and sulphide content.

Thin section petrography by Geul (1973) indicated that the gabbro ranges in composition from anorthositic gabbro to olivine gabbro and quartz gabbro. Deuteric alteration is slight to intense and consists of secondary amphibole, biotite, saussurite, sericite, and hematite and appears to be closely related to the presence of an intergranular granophyric phase. Bodies of granophyre cut the gabbro in several locations and appear to represent the differentiated end product of the gabbroic magma. The gabbro is composed essentially of pyroxene and zoned labradorite.

DISCUSSION OF GEOCHEMISTRY

As part of the Lake Nipigon Region Geoscience Initiative whole rock analyses were performed on a number of the intrusions south of Thunder Bay; these data can be combined with published (Cogulu 1990), unpublished data for some of the dykes in the region (Larry Hulbert, Geological Survey of Canada, personal communication, 2006) and data from exploration projects (e.g., Rosatelli 2002), in order to assess variations in geochemistry.

On a primitive mantle-normalized diagram, the samples of Crystal Lake gabbro and Pigeon River dykes show similar patterns (Figure 3-5). Pigeon River dykes from the Arrow River (Stop 3-1) and Rita Bolduc occurrence (Stop 3-2) are similar to the range of data from other dykes of the Pigeon River swarm sampled in Lake Superior (Victoria Island, Cloud Bay and Jarvis Point (Figure 3-6; Larry Hulbert, Geological Survey of Canada, personal communication, 2006). However, samples of the Pigeon River dyke swarm from Lake Superior are characterized by higher Th abundances than those from the study area, resulting in more pronounced negative Nb anomalies. The only published analysis for the Arrow Dyke swarm (Rosatelli 2002) is indistinguishable from the Pigeon River swarm. By contrast, samples from the Mount Mollie dyke are characterized by elevated La/Sm_n ratios, higher Th contents and more pronounced Nb anomalies (Figure 3-6). This may be a reflection of greater crustal contamination during emplacement or possibly a distinct mantle source. Regardless, this geochemistry provides another means for distinguishing these dykes.

When the data from the dyke swarms are compared to the regional data set generated for the sills and intrusions of the Lake Nipigon embayment (Hollings et al. 2007) it can be seen that Pigeon River dyke swarm geochemistry more closely resembles that of the sills of the Nipigon suite than that of the ultramafic intrusions or the Logan sills (Figure 3-6). By contrast, the Mt Mollie dyke geochemistry appears to be transitional between that of the Nipigon sills and Inspiration sills. Additional isotopic and geochronological studies will be required in order to further investigate the relationships between these MCR-related intrusions.

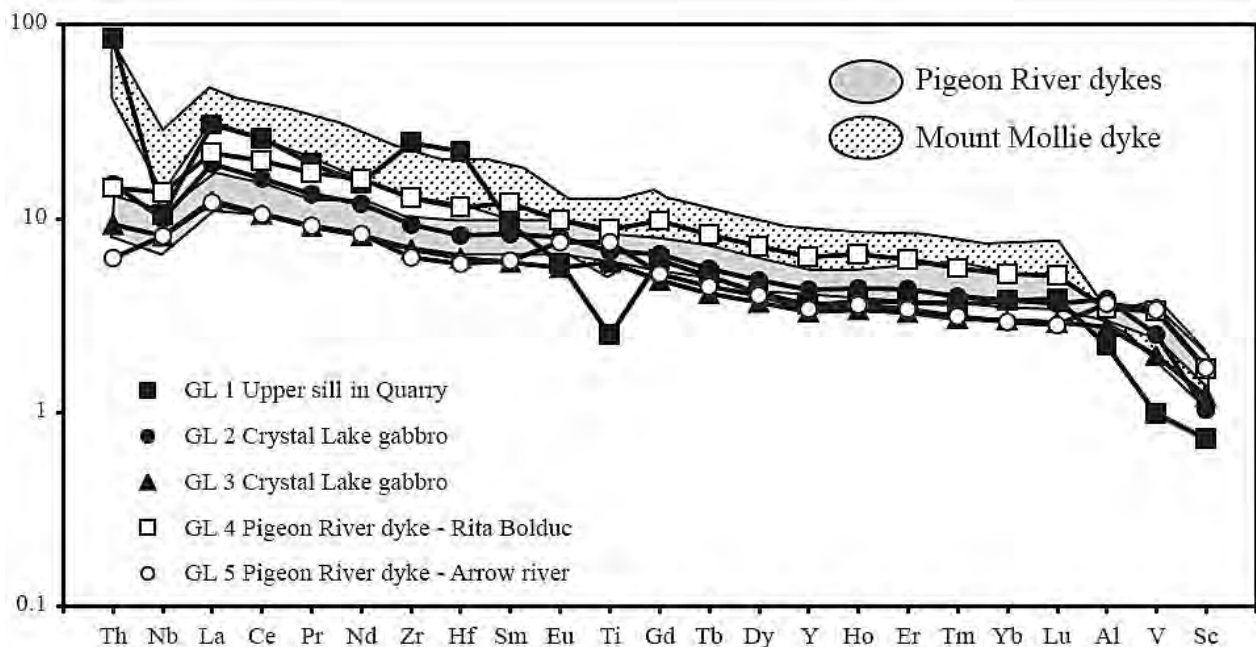


Figure 3-5: Primitive mantle-normalized plots for mafic intrusive rocks in the field trip area. Fields from data contributed by L. Hulbert, Geological Survey of Canada, personal communication (2006). Normalizing values from Sun and McDonough (1989).

CONTACT METAMORPHISM

There is a remarkable range in the reported intensity and nature of contact metamorphic effects in Rove sedimentary rocks at diabase dyke and sill contacts, owing mainly to the subjectivity of the mapper and the exposures in question. Geul (1973) noted that sedimentary hornfelsic rocks are restricted to a narrow zone of baking between 2 to 10 cm wide at diabase dike contacts. Metamorphosed siltstone displays two stages: first: slight recrystallization of biotite aggregates in an incipient hornfelsic texture; and second, a more complete recrystallization of biotite, surrounded by pale sericitic aggregates, set in a quartzo-feldspathic matrix. Franklin (1970) suggested that contact effects existed up to 8 m from sill contacts and possibly up to 23 m. They were manifested as microporphyroblasts of mica and chlorite (“spotted alteration”), graphite destruction and the conversion of pyrite to pyrrhotite. Geul (1973) noted that minute particles (< 0.01 mm) of oxide and sulphide minerals are locally abundant in the contact zone.

Rove Formation sedimentary rocks may be deformed along dyke contacts. As noted by Geul (1973) beds appear to dip toward the dykes or are “up-dragged” along dyke contacts (e.g. Stop 3-2). Up-dragging of Rove Formation rocks was also noted along the southern contact of an east-striking Pigeon River dyke at Arrow Rapids by Geul (1970). Deformed and fractured sedimentary rocks have been noted near sill terminations (e.g. Stop 3-3). Narrow, parallel, tension gashes filled with quartzo-feldspathic leucosome/neosome occur in metatectic, deformed siltstones at Stop 3-2 (Figure 3-8).

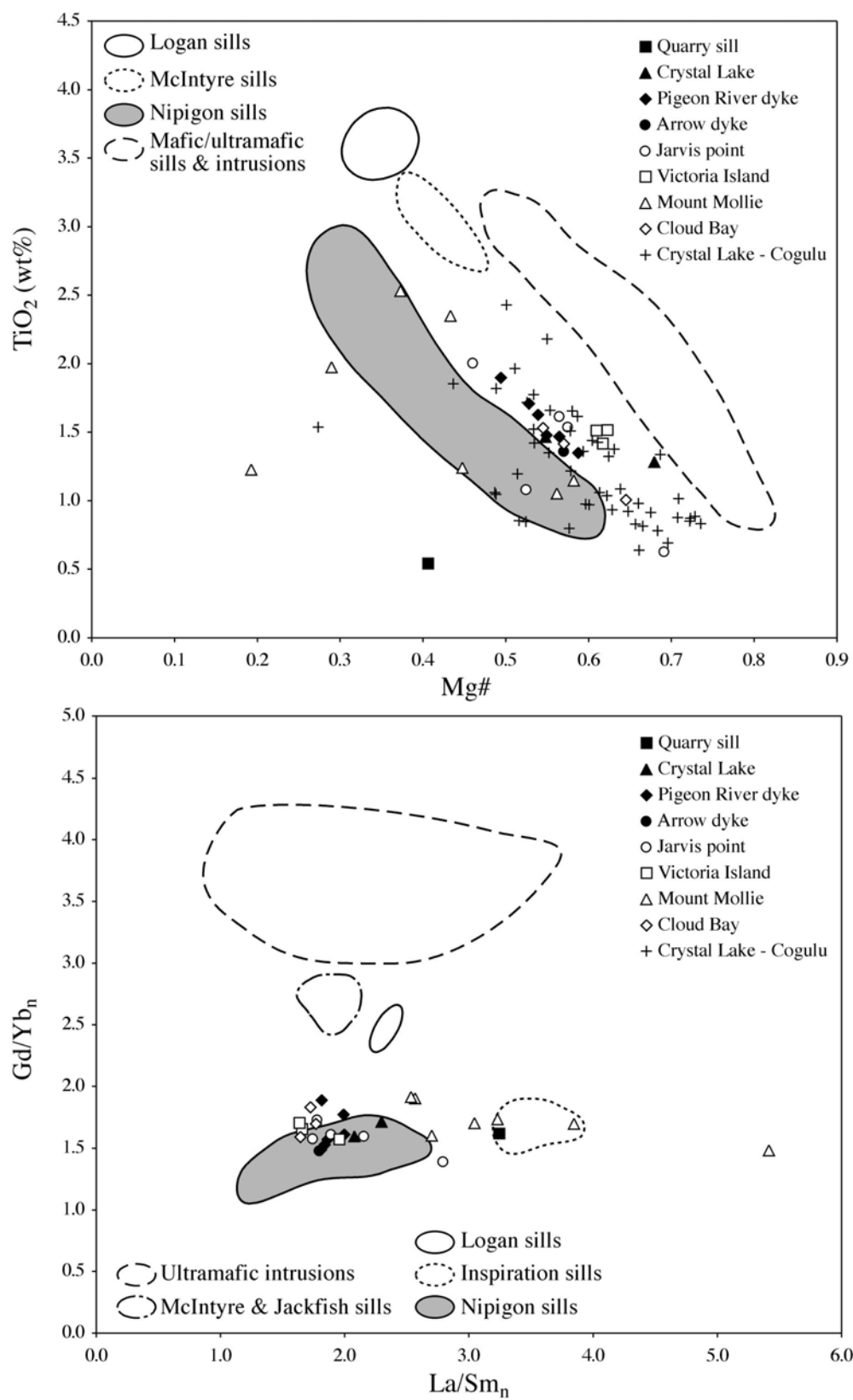


Figure 3-6: Discrimination diagrams for mafic and ultramafic intrusions near Thunder Bay. Data are from Hollings et al. (2007) and Hart (2002).

MINERAL EXPLORATION

Active prospecting was first carried out during the latter part of the nineteenth century, following the discovery of silver vein deposits along the shoreline and islands of Lake Superior (Island Silver Belt) and to the northwest (Mainland Silver Belt; Oja 1967). Small silver mining operations took place on Spar and Jarvis Islands and at the Prince Mine. Barite was produced from veins on McKellar Island and Jarvis Island. Silver-bearing veins are in or near crustal-scale, extensional listric faults that formed during MCR extension and are spatially associated with MCR-related mafic intrusions (Franklin et al. 1986).

Copper-nickel-mineralized boulders were found in 1903 by J. A. McCuaig in Pardee Township (Geul 1970). This discovery led to a concerted exploration effort to find their source (Geul 1970, 1973). Local copper-nickel occurrences were soon described by Tanton (1935, 1937). United States Smelting, Refining and Mining Company conducted exploration in 1936, and were followed by Frobisher Exploration Company in 1942. In the summer of 1952, J.S. Brodie and T.W. Page examined a large outcrop of gabbroic rocks, 6 km northeast of the original float discovery on the property now held by the Great Lakes Nickel Corporation Limited. Following this discovery, several blocks of ground were staked in Crooks Township by Whitegate Mining Company Limited, J. S. Strickland, and others, in an effort to cover the interpreted eastward extension of the favourable host rock. In 1967, Canadian Exploration Limited investigated the Strickland-Whitegate claims in southwestern Crooks Township and Anaconda American Brass Limited prospected the Pine River-Mount Mollie area as part of a regional reconnaissance program.

In 1968, Phelps Dodge Corporation of Canada Limited explored and drilled claims in the Pine River-Mount Mollie area of Crooks Township. In 1989, International Platinum Corp. drilled a 3592 m hole in central Crooks Township. A renewed interest in platinum group elements in the late 1990s led to a reevaluation of the area's potential and a flurry of exploration and drilling by companies like BHP Billiton World Expl. Inc., Falconbridge Ltd., McVicar Minerals Ltd., North Atlantic Nickel Corp. and Dumont Nickel Corp. A Noril'sk-style model was employed, as had been earlier suggested by Lightfoot and Lavigne (1995) and others in promoting the area's mineral potential.



Figure 3-7:
Columnar joints in
the Crystal Lake
Gabbro, exposed
above the upper adit
on the Great Lakes
Nickel property
(Stop 3-4). The adit
opening (lower left)
is approximately 3m
wide.



Figure 3-8: Narrow,
parallel tension
gashed filled with
quartz-feldspathic
leucosome/neosome
in metatectic,
deformed siltstone
near contact with
Pigeon River dyke,
Stop 3-2

FIELD TRIP STOPS: ROAD LOG AND DESCRIPTIONS

STOP NAME	STOP NUMBER	LANDMARK (0.0 km)	DISTANCE (km)	NORTHING	EASTING
				(NAD 83)	
	<i>Reset Trip Odometer</i>	Visitor Centre @ Border	0.0 / travel north		
		Junction with Hwy. 593	2.3 / turn left, travel west		
Middle Falls Dyke	Optional	Middle Falls	4.5	5321108	304872
		Junction with Old Border Road	10.9 / turn left, travel west 500 m		
		Junction with Logging Road	11.4 / turn right, travel northwest 3.8 km		
		Arrow River bridge	14.4		
Arrow River Dyke	3-1		15.2	5324134	296694
	<i>Reset Trip Odometer</i>	Junction of Hwy. 593 and Old Border Road	0.0/ left turn, travel north		
		Junction with Hwy. 597 (Pardee Road)	1.7 (on right)		
McCuaig Float Occurrence	Optional		3.8 / 150 m north of hwy. (flagged trail)	5325546	299009
	<i>Reset Trip Odometer</i>	Junction of Hwy's. 593 and 61	0.0 / turn left, travel north		
Rita Bolduc Occurrence / Pigeon River Dyke	3-2		4.4	5324701	310563

		Junction with Great Lakes Nickel Road	6.6 (gated, on left)		
		Pine River Bridge	6.75		
		Junction with Memory Road	6.9 / turn right, travel east 3.1 km		
		Trail to Mount Mollie lookout	10.0 / trail on east side of road		
Mount Mollie Lookout	Optional		Follow trail ~300 east to top of ridge	5325444	313843
		Return to Highway 61; turn left, travel south	Travel south 300 m to Great Lakes Nickel Road (gated, on right)		
	Reset Odometer	Junction of Hwy. 61 with Great Lakes Nickel Road	0.0 / travel west		
Rove Formation Quarry	3-3		4.7 (south side of road)	5328687	306401
Great Lakes Nickel Property	3-4		7.5 (western end of ridge; cliff face)	5328392	303612

STOP (OPTIONAL): Middle Falls

Location: South side of Highway 593, 4.5 km west of Visitor Centre at Border on Highway 61 (UTM 304872E / 5321108 N; NAD83) (N.B. On older maps, this location is referred to as Little Falls.)

Description: This roadside pull-off offers a picturesque view of Middle Falls, an 18 m high waterfall underlain by a resistant, west-northwest-trending, diabase dyke cutting Rove Formation clastic sedimentary rocks. The dyke is exposed in a roadcut on the highway opposite the pull-off area. It is cut by a number of parallel, rusty carbonate veins.

STOP 3-1: Pigeon River Dyke (a.k.a. the “Arrow River Dyke”)

Location: Road cut, 15.2 km west of Visitor Centre at Border on Highway 61, via Old Border Road (@ 10.9 km) and logging road (@ 11.4 km) (UTM 296694E / 5324134N; NAD83)

Description: This stop is situated on top of a prominent northeast-trending ridge which extends for over 20 km and rises up to 150 m above the surrounding countryside. The ridge is underlain by a Pigeon River dyke which, owing to its location, has been colloquially referred to as “the Arrow River dyke”, although it is not part of the northwest-trending Arrow River dyke swarm of Smith and Sutcliffe (1989).

The exposed dyke consists generally of massive fine- to medium-grained diabase. Conspicuous, pyroxene-phyric patches have diffuse to indistinct contacts with the finer-grained diabase, suggesting that they may represent fluid-rich pockets or cognate xenoliths. (Similar textures are noted at Stop 3-2 as well.)

A very small amount (~50 grains) of tiny baddeleyite fragments and blades were recovered from a sample of diabase at this location (Heaman and Easton 2006). A larger number of 40–100 µm colourless resorbed zircons were also recovered, but these were interpreted to be of xenocrystic origin. The U/Pb results are from one small multi-grain baddeleyite fraction. This baddeleyite fraction has a low uranium content (188 ppm U) and a low Th/U ratio (0.022); the latter is typical for magmatic baddeleyite. This baddeleyite analysis displays slight reverse discordance (–1.4%) and has a $^{206}\text{Pb} / ^{238}\text{U}$ age of 1092.7 ± 2.6 Ma, which is interpreted as a good estimate for the minimum age of the Arrow River diabase dike (*ibid*).

STOP (OPTIONAL): McCuaig Float Occurrence

Location: North side of Highway 593, 3.8 km west of junction of Highway 593 and Old Border Road / 1.7 km west of junction of Highways 593 and 597 (Pardee Road) (UTM 299009E / 5325546 N; NAD83)

Description: This stop comprises a number of pits and trenches developed in a hummocky, boulder-strewn hillside. Copper-nickel-mineralized boulders were found in 1903 by J. A. McCuaig (Geul 1970). In 1936, a magnetometer survey was conducted by United States Refining and Mining Company. In 1942, Frobisher Exploration Company Limited completed a second magnetometer survey. In 1949-1950, J. A. McCuaig (original discoverer’s son) carried out some minor trenching and sampling and drilled a 200 foot subvertical hole which reportedly showed the mineralized rock was in the form of float boulders. A small bulk sample was tested by Falconbridge Nickel Mines Limited. In 1967, Hollinger Consolidated Gold Mines Limited carried out geophysical surveys. Hanna Gold Mines Limited completed a vertical hole nearby. In 1990, Fleck Resources Ltd. carried out magnetometer and geochemical surveys along with limited prospecting, trenching and sampling just north of this occurrence (Resident Geologist’s Files, Thunder Bay South District, Thunder Bay).

Tanton (1935, 1937), McCuaig (1950) and Geul (1970) have described mineralized, vari-textured diabase, gabbroic anorthosite and norite. Interstitial to blebby pyrite, pyrrhotite, chalcopyrite and pentlandite characterize the sulphide mineralization. Assay results are shown below:

<u>Reference</u>	<u>Cu (%)</u>	<u>Ni (%)</u>	<u>Co (ppm)</u>	<u>Pd (ppb)</u>	<u>Pt (ppb)</u>
Geul (1970)	0.23	.018	n/a	Trace	n/a
Resident Geologist Files, OGS	1.76	0.72	190	n/a	610
Tanton (1935)	0.62	0.34	n/a	n/a	5140

STOP 3-2: Rita Bolduc Occurrence (Pigeon River Dyke)

Location: Highway 61, 4.4 km north of the junction between Highways 593 and 61; (UTM 310563E / 5324701N; NAD83)

Description: This highway roadcut (Figure 3-9) exposes the southern contact and an almost complete cross-section of a sulphide-mineralized Pigeon River dyke. The property was explored with a self potential survey and trenching in 1957 (Geul 1973). It was part of a property that was acquired and explored by Falconbridge Limited and McVicar Minerals. The main sulphide showing is situated along the south edge of a Pigeon River olivine diabase dike, carrying disseminated to massive copper-nickel-sulphide mineralization in a fracture zone, up to 35 feet in width and extending for a distance of at least 100 feet along the dike (Geul 1973). A grab sample of massive to disseminated pentlandite, pyrrhotite with minor chalcopyrite returned 0.70% Cu and 0.35% Ni and trace amounts of Pd (*ibid*). Net-textured pyrrhotite and chalcopyrite blebs occur locally. Grab samples collected from sulphide-mineralized sections of the dyke on the west side of the highway yielded the following results:

Sample	Au (ppb)	Cu (ppm)	Ni (ppm)	Pd (ppb)	Pt (ppb)
01-BRB-01		1442	707	28.03	21.84
01-BRB-02	14.29	880	505	17.55	13.90
01-BRB-03	25.02	6517	2664	23.16	n.d.

(Resident Geologist's Files, Thunder Bay South District, Thunder Bay)

Hornfelsed, rusty weathering, Rove Formation sedimentary rocks are exposed along the southern dyke margin. They are fissile and display an orthogonal joint pattern. Beds have been up-folded along the contact, suggesting either dip-slip fault motion and/or high magma flow rates. Similar “up-drag” features were noted on the southern contact of the Arrow River dyke at Arrow Rapids by Geul (1970), also in association with a copper-nickel sulphide-mineralized fault/fracture zone. Narrow, parallel tension gashed filled with quartzo-feldspathic leucosome/neosome occur in metatectic, deformed siltstone; infilled fractures may produce leucosome dykes (Figure 3-8). They are oriented parallel to the dyke contact and occur several metres away from the contact. Rare, small (< 1 cm diameter), round chloritic patches may represent incipient “spotted” alteration, similar to that described by Franklin (1970).

Geochronologic sampling by Heaman and Easton (2006) at this location showed that the least magnetic fraction contained a large amount of colourless euhedral apatite, pyrite and a small amount of baddeleyite and zircon. A very small amount of tiny tan baddeleyite blades and fragments was recovered and a number of composite baddeleyite/zircon grains were identified. Although not that common, primary igneous baddeleyite overgrown by magmatic zircon has been reported previously and is consistent with the phase relationship for crystallization of these two minerals in a mafic magma where the silica activity is progressively increasing throughout crystallization. The four zircon analyses have varied uranium and thorium contents (163–1104 ppm and 34–237 ppm, respectively), but have high Th/U (1.17–1.81), consistent with primary magmatic zircon crystallizing from a mafic magma. The four zircon analyses are discordant (8.7–11.7%) and have $^{207}\text{Pb}/^{206}\text{Pb}$ ages that vary between 1050 and 1090 Ma. The baddeleyite fraction (#5) is least discordant (3.8%) and contains relatively low uranium (213 ppm) for baddeleyite. The 1133.8 Ma $^{207}\text{Pb}/^{206}\text{Pb}$ age obtained for this baddeleyite fraction is interpreted as a minimum estimate for the age of this dike. A Model 1 regression treatment of three zircon (#1, #2, #3) and the baddeleyite (#5) yields an upper intercept age of 1141 ± 20 Ma, which is interpreted as the best estimate for the emplacement of the Pigeon River dike (Heaman and Easton 2006).

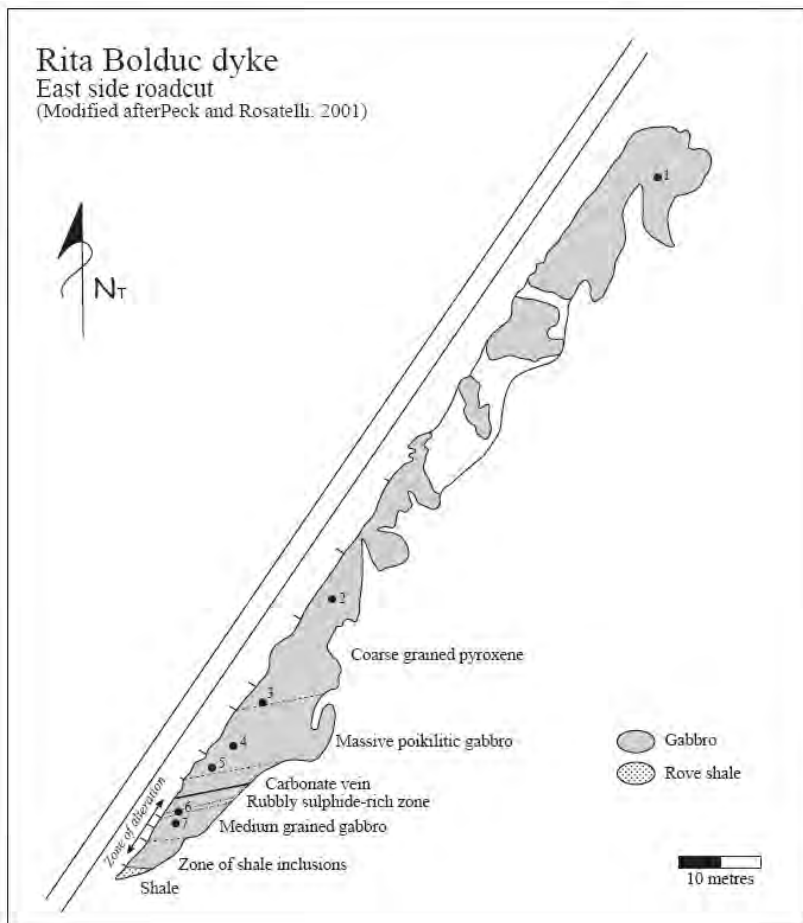


Figure 3-9: Sketch map of the road cut exposing a Pigeon River dyke on the eastern side of Highway 61 at Stop 3-2. Geology modified after Peck and Rosatelli (personal communication, 2001) in Parker et al.(2001). Geochemical sample locations are shown as numbered black dots.

STOP (OPTIONAL): Mount Mollie Lookout

Location: Approximately 300 m north along old road bed, 3.1 km east of Highway 61 on Memory Road. The junction of Memory Road and Highway 61 is 6.9 km north of the junction of Highways 593 and 61, just north of the Pine River bridge (UTM 313843E / 5325444N; NAD83)

Please be advised that this stop is located on private land. Permission must be granted by the land owner. Contact the Resident Geologist's Office in Thunder Bay for further information.

Description: This stop, atop a prominent northeast-trending ridge, affords a panoramic view of Pine Bay, Lake Superior and the surrounding, ridge-dominated topography. This and other nearby ridges are underlain by ophitic Pigeon River diabase dykes. Pigeon Point, Minnesota (type locality for pigeonite) can be seen on the horizon to the south. Northwest-trending ridges, underlain by Arrow River dykes, extend inland from south of Big Trout Bay, just north of this location. A long (>1 km) and challenging trail extends north and east from this vantage point to copper-nickel-mineralized occurrences in the Pine River-Mount Mollie dyke (ca. UTM 314748E / 5325640N; NAD27), which is approximately 60 m wide at that location. Heslop (1968) noted blebs of pyrrhotite-pentlandite-chalcopyrite and magnetite-ilmenite intergrowths at this occurrence. Supergene(?) native copper has also been noted there.

STOP 3-3: Rove Formation Quarry, Great Lakes Nickel Limited Road

Location: 4.7 km west of Highway 61 on the unmarked Great Lakes Nickel Road (private access). The GLNR extends west from Highway 61 at a point just south of the Pine River bridge, 6.6 km north of the junction of Highways 61 and 593 (UTM 306401E, 5328687N; NAD83).

Description: A dormant shale quarry on the south side of the Great Lakes Nickel Limited access road provides a representative section of nearly flat-lying Rove Formation turbidites, diabase dykes and a sill. The quarry face is capped by a 1.1 m thick wacke bed that is underlain by a 1.1 m thick, conformable diabase sill in the main part of the quarry. Approximately 12 m of thinly to thickly bedded wacke and shale are exposed below the sill. The wackes are typically massive, and display sharp, planar contacts with the recessively weathered shale units. Scours of underlying shale are locally developed. Some of the shale interbeds display sediment loading features and distinctive sole features and are locally flaser-bedded. Another, 22 cm wide diabase dyke intrudes the sedimentary rocks east of the sill/dyke exposure, strikes ~060° and dips ~80° east.

The columnar-jointed diabase sill is fine- to medium-grained and has conspicuous, grey chilled margins. It pinches out on the eastern end of the quarry face where it apparently “steps down” through the sedimentary rocks and merges with a near-vertically dipping diabase (feeder?) dyke. This dyke is weakly magnetic, strikes 030° and has an exposed thickness of approximately 4m. A sample collected near the chilled margin returned the following analysis:

Sample	Al ₂ O ₃	CaO	Fe ₂ O ₃	K ₂ O	LOI	MgO	MnO	Na ₂ O	P ₂ O ₅	SiO ₂	TiO ₂	TOTAL
01-BGL-01	14.51	9.15	13.92	0.78	1.08	5.31	0.2	2.69	0.15	51.78	1.52	101.1

(Resident Geologist's Files, Thunder Bay South District, Thunder Bay)

In contrast a sample of the upper sill was considerably more silica rich (GL-1; SiO₂ = 72 wt%) and is characterized by elevated La/Sm_n ratios (3.2) and Th contents (7 ppm vs. <1.3 in other dykes and sills). This likely reflects localized assimilation of Rove wacke by the sill during emplacement. A thin veneer of the hornfelsed sedimentary host rocks is locally preserved against the chilled dyke margin. The diabase is plagioclase-phyric and contains small sub-rounded shale xenoliths. A tendril of diabase extends west from the dyke (below the sill) and forms a bulbous, pillow-like termination with radial cooling fractures. The adjacent shale is chaotically folded and crenulated, suggesting that shearing accompanied intrusion. Two metres further west, an isolated ovoid (45 cm by 75 cm) “blob” of diabase intrudes the shale. Similar fingering termination features were described at the periphery of sills by Pollard et al. (1975). At finger terminations, the host rock is wedged aside and compacted and along the edges of the fingers, buckling and shearing of strata are common. Pollard et al. (1975) ascribed finger formation to the instability of the advancing interface between a viscous magma and a more viscous host rock. Antonellini and Cambray (1992) also studied a possible structural control on local sill emplacement.

A block of wacke was selected at this location for a detrital zircon study (Heaman and Easton 2006). The detrital zircon ages form two distinct age clusters, with the majority of grains having ²⁰⁷Pb/²⁰⁶Pb ages in the 1780–1880 Ma range. A second population consists of Archean grains, most of which are in the 2600–2700 Ma range. The oldest grain has a ²⁰⁷Pb/²⁰⁶Pb age of 2914 Ma. The youngest grain, with a ²⁰⁶Pb/²³⁸U age of 1731 Ma places a maximum age on the depositional age of this wacke. This grain is 5% discordant, consequently, a better estimate for a maximum depositional age is provided by a 1777 Ma concordant detrital zircon grain. This interpretation is supported by the abundance of grains (n=10) between 1796 and 1777 Ma, and is consistent with U/Pb ash bed ages of 1836±5 and 1832±3 Ma from the basal Rove Formation (Addison et al. 2005).

The section exposed in the quarry walls represents a portion of the middle of the almost kilometer-thick Rove Formation (P. Fralick, Lakehead University, personal communication, 2007). The lower, underlying Rove is dominated by approximately 120 m of carbonaceous shale. This gradationally coarsens and thickens upwards through 100 m by the addition of thin sandstone

turbidites, culminating in beds similar to the ones exposed in the quarry. Proceeding upwards, above the rocks exposed in this section, ripples become more abundant in the sandstones until the assemblage is dominated by lenticular to flaser-bedded, wave- and current-rippled units. The Rove Formation represents a period of sediment starvation followed by progradation of a prodeltaic turbidite apron and distal delta top distributary mouth bar complex. Sections in cored drillholes, equivalent to the exposed strata here, are composed of parasequences that coarsen and thicken upwards over meters to a few tens of meters by the addition of thicker sandstone beds. These represent prograding turbidite lobes that are rapidly abandoned as the subaqueous feeder channel switches position. A period of sediment starvation follows with mud accumulation. This is ended by the gradational outbuilding of another lobe as the feeder channel moves closer to this area again. The thicker shales overlying the sand-rich tops of the parasequences are carbonaceous and pyritic, compared to the thinner shales forming the tops of individual turbidites. This denotes that the clays forming the thicker shales separating lobe outbuilding periods were deposited slowly allowing the remains of microbes to become common in the sediment. During diagenesis the microbes degraded, forming strongly reducing conditions that led to formation of diagenetic pyrite in the sediment. In contrast, the more rapid deposition of clay between turbidity current events during lobe outbuilding produced lower concentrations of carbon and much less pyrite. The difference between the two types of shale can easily be seen due to staining of the pyrite-rich rock during oxidation. Complete Bouma Sequences are rarely exhibited by the turbidites as the rippled C-division is commonly not present. However, flute marks produced by scouring of the viscous turbidity current into the muddy bottom and subsequent filling with sand, are ubiquitous (Figure 3-12). Commonly, in other turbiditic formations, flute marks are randomly developed on the bedding surface, but here they are in places arranged geometrically in diagonally offset rows. They also take on unusual elongate shapes, more akin to groove marks produced by something dragging across the bottom. The cause of the abnormal patterns and shapes is not known.

Figure 3-10: View, looking south, of the eastern end of the dormant quarry in Rove Formation sedimentary rocks on the Great Lakes Nickel road, Stop 3-3. A diabase feeder dyke (left-center db) branches out into two sills: a main one, near the top of the quarry face (top-center db) and midway up the face, where it fingers and terminates (white box corresponds to Figure 3-11). The white dashed line separates oxidized, pyritic shale below and relatively pyrite-free shale above.





Figure 3-11: Diabase (db) “finger and blob” in Rove Formation wacke, Stop 3-3. The diabase was injected from a feeder dyke to the left of the photograph. A propagation fracture, with attendant folding and shearing, is denoted by the dashed line.



Figure 3-12: Casts of flute marks on the bottom of a sandstone layer, Stop 3-3. Some have the classic “V” shape with their deepest area at the apex of the V. The V opens in the down-current direction. Others are drawn out and appear to have migrated down-current as they cut, producing a groove rather than a V.

STOP 3-4: Crystal Lake Gabbro / Great Lakes Nickel Property

Location: 7.5 km west of Highway 61 on the unmarked Great Lakes Nickel Road (private access). The GLNR extends west from Highway 61 at a point just south of the Pine River bridge, 6.6 km north of the junction of Highways 61 and 593 (UTM 303612E, 5328392N; NAD83).

Description: A switch-back road extends from the base of a 140 m high ridge to the lower adit, excavated at the contact between Rove Formation country rocks and the base of the Crystal Lake gabbro. An overgrown trail continues upward to the upper adit in “taxitic”, varied-texture gabbro and talus boulders. The upper portions of the intrusion, including layered rocks, are visible from this vantage point but are inaccessible for safety reasons. A stockpile of mineralized gabbro is available for sampling on the road at the base of the ridge. A second dump, further up the road, consists mainly of Rove Formation shales and argillites with abundant zeolite and carbonate veins.

The discovery of copper- and nickel-mineralized float boulders in this area early in this century led to a concerted exploration effort to find their source. United States Smelting, Refining and Mining Company conducted exploration in 1936, and were followed by Frobisher Exploration Company in 1942. In the summer of 1952, J.S. Brodie and T.W. Page examined a large outcrop of gabbroic rocks, 6 km northeast of the original float discovery. Prospecting soon indicated a Cu-Ni-Pt-mineralized zone and the ground was staked for Mattawin Gold Mines. Due to lack of operating capital, the Mattawin property was optioned to Falconbridge Nickel Mines in late 1952. A trench was subsequently excavated to test the lower gabbro contact. Although trench samples returned anomalous copper and nickel values, they did not warrant further investigation and the option was allowed to lapse. Additional work by J.S. Brodie drew attention to the area north of the trench. The property was optioned to R. Barker and W. Dawidowich in 1954. Six diamond drill holes, totalling 1058 m, were completed. One 55-foot section returned 0.54% Cu and 0.18% Ni with some precious metal values. Mogul Mining Corporation Limited held the property from 1954 to 1957 and drilled seven holes, totalling 1693 m. Intersections of the mineralized zone averaged 9 to 12m and assayed 0.9% combined Cu and Ni.

Late in 1964, Great Lakes Nickel Corporation Limited acquired the property and initiated their exploration program in June, 1965. From 1965 to 1970, 47 803 m of surface drilling was completed; 19 underground holes, totalling 392 m, were also drilled. Underground drilling was conducted from a newly constructed, 37 m long adit, driven into the base of the hillside. In addition, Thunder Bay Nickel Mining Corporation Limited drilled 16 holes, totalling 13 579 m, on the down-plunge extension of the deposit, 2.5 km east of the main workings. In 1972, access and development work was undertaken by Great Lakes Nickel to further test the deposit. This work included the excavation and driving of a 522 m development portal and drift and over 12 000 m of surface and underground drilling. Plant site surveys, bulk sampling, metallurgical and feasibility tests were also conducted, financed largely by the Swedish company, Boliden Aktiebolag. By 1974, plans were made to mine the deposit at an initial rate of 1.8 million tons per year (subsequently increased to 2.5 million tons per year). Up to that point, about \$10 M (Canadian) had been spent on the property on 58 689 m of surface drilling, 26 182 m of underground drilling and the driving of the adit, which eventually reached a length of 1041 m. This work had outlined a deposit of 32.8 million tons grading 0.36% Cu and 0.20% Ni, with a further potential reserve of 40 million tons of about the same grade.

However, cost escalations, high interest rates and uncertain metal prices forced suspension of mine development in October, 1974. Rising interest in platinum-group elements in the mid-1980's prompted Fleck Resources Ltd. to re-evaluate the deposit whose reserves then stood at 45.6 million tons at a grade of 0.334% Cu and 0.183% Ni. Between September, 1986 and February, 1987, Fleck completed geological mapping and sampling, as well as the re-logging and assaying of more than 9144 m of drill core. Six holes were drilled to test the deposit for its PGE potential. Sampling by Fleck returned the following assays on a 3.7 million ton portion of the deposit: 0.006 oz./ton Pt,

0.57% Cu, 0.027 oz./ton Pd, 0.264% Ni, 0.003oz./ton Au, 0.016% Co, 0.04 oz./ton Ag. Based on February, 1987 metal prices, this material was valued at \$34.29 (Can.) per ton (Fleck Resources Ltd., Annual Report, 1987). Great Lakes Nickel has undertaken a comprehensive, due-diligence review of the property (Hubacheck 2001) in order to generate a valuation opinion and summary for potential investors or joint-venture partners.

The geology of the Crystal Lake gabbro has been described by Geul (1970), Reeve (1969), MacRae and Reeve (1968), Whittaker (1986) and Smith and Sutcliffe (1987, 1989). It intrudes Rove Formation argillaceous and arenaceous rocks. The intrusion was subdivided into three major zones by Reeve (1969) and Geul (1970) and has been further subdivided by Smith and Sutcliffe (1987, 1989) and Cogulu (1990) into four major, roughly equivalent, lithologic zones:

- (1) an **upper zone** (60 to 80 m) of sulphide-barren, troctolite, olivine gabbro and anorthositic gabbro;
- (2) a **middle zone** (30 to 42 m) of cyclic, layered anorthositic and olivine gabbro, Cr-spinel-bearing anorthosite and olivine gabbro (Figure 3-4);
- (3) a **lower, unlayered zone** (50 m) of vari-textured gabbro and leucotroctolite, which hosts the bulk of the Cu-Ni-sulphide deposit; and
- (4) a **basal zone** (1 to 7 m) of fine-grained, chilled melagabbro and hornfelsed country rock xenoliths.

Orthocumulate rocks predominate in the Crystal Lake intrusion; adcumulates occur only in the cyclic, layered rocks, in association with Cr-spinel mineralization (Cogulu 1990). Plagioclase, with lesser olivine and Cr-spinel (Figure 3-3), is the main cumulus mineral, while clinopyroxene, magnetite and Cu-Ni-sulphides are intercumulus (Cogulu 1990; Smith and Sutcliffe 1987, 1989). Total REE, LREE/HREE ratios, and sulphide content decrease from the basal zone to the cyclic sequence, while the modal proportion of olivine increases (*ibid*). Smith and Sutcliffe (1987, 1989) have suggested the following crystallization sequence:

Cr-spinel → olivine → plagioclase → clinopyroxene → magnetite → apatite.

Biotite is epitaxial to olivine and augite, and it may also rim intercumulus sulfides and oxides (Cogulu 1990).

Sulphides are disseminated, interstitial and included grains and droplets. Pyrrhotite, chalcopyrite, cubanite and pentlandite are the main sulphides. Accessory minerals include violarite, troilite, niccolite, maucherite, native bismuth, mackinawite, bornite, millerite, nickeloan pyrite, sphalerite and marcasite (Cogulu 1993a). Adcumulate and orthocumulate Cr-spinels have been subdivided by Cogulu (1993b) into compositionally and texturally distinct groups with complex re-equilibration histories. Heterogeneity between Cr-spinels was also noted by Whittaker (1986).

Local sulphur sources exist in local sedimentary rocks, as well as in the Archean basement. Data from the Duluth Complex indicated similar sulphur-isotopic signatures exist between basement sedimentary rocks and Ni-Cu-sulphides (Mainwaring and Naldrett 1974). Similar S-isotopic signatures also exist between Crystal Lake gabbro and Rove Formation shales (J. Franklin, personal communication, 1999.) Franklin (1970; personal communication, 2001) has also noted contact effects in Rove shales below diabase sills which include the conversion of pyrite to pyrrhotite, the loss of carbon in the spotted contact aureole and the rearrangement of clay mineral structures. Eckstrand and Cogulu (1986) presented Se/S data which supported the contention that sulphur in the Ni-Cu-sulphide minerals at Crystal Lake was derived from the barren, Se-poor, sedimentary sulphides of the Rove Formation. Sulphide concentrates from the Rove Formation and the lower parts of the intrusion (which host the sulphide deposit) yielded lower Se/S ratios than those gleaned from the Cr-spinel-bearing, cyclic layered zone whose Se/S ratios correspond closely to chondritic values (~ 200 to 600×10^{-6}), implying that the sulphur was likely mantle-derived.

Stratigraphic Unit	(Se/S) x 10 ⁻⁶
Cyclic Layered Zone	213 - 687
Lower Unlayered Zone	70 - 151
Basal Contact Zone	28 - 114
Rove Formation (Footwall)	2 - 118

(Eckstrand and Cogulu 1986)

A grab sample of non-magnetic, massive subhedral pyrite with minor, interstitial, fine-grained silicates taken from the lower adit area, near the basal contact of the Crystal Lake Gabbro, returned 116 ppb Au, 688 ppb Pd, 14 ppb Pt, 13494 ppm Cu, 4756 ppm Ni and 255 ppm Zn (Resident Geologist's Files, Thunder Bay).

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53rd Annual Institute on Lake Superior Geology

FIELD TRIP 4

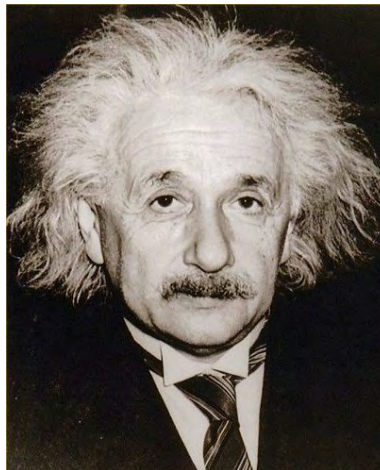
GEOLOGY OF THE NICKEL LAKE MACRODIKE AND ITS ASSOCIATION WITH CU-NI-PGE MINERALIZATION IN THE NORTHERN SOUTH KAWISHIWI INTRUSION, DULUTH COMPLEX, NORTHEASTERN MINNESOTA

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“Imagination is more important than knowledge”

Albert Einstein

INTRODUCTION

Mineral deposits are concentrations of specific ore-minerals that society utilizes in immeasurable ways. They have formed from Archean times up to the present, and vary greatly in commodity, mineralogy, alteration, trace element signature, geophysical properties, and grade-tonnage. Individual ore deposits are always unique, and this uniqueness arises from two main sources: 1) fundamental differences in geologic processes and environments; and 2) local, site-specific, geologic variations and associated bounding geometries. The seemingly limitless number of permutations of these (and other) features of mineral deposits defies imagination. Geologists employed in the search for ore deposits have, over the last century, developed the intellectual concept of **Ore Deposit Models**, which seek to organize many of

these variables, lump individual mineral deposits into classes, and establish criteria to aid in mineral exploration (Peterson, 2001a). Such models may be strictly *empirical* – a collection of observable facts associated with the occurrence of certain metals in economic proportions – or *genetic* – which attempts to describe the physical and chemical processes responsible for the development of an ore deposit and its related empirical features.

Advances in our collective understanding of ore deposit types is nonlinear, rather it occurs in fits and starts. Take, for example, the great advance in knowledge – and thus mineral exploration models and exploration success – for porphyry copper deposits in the 1960s and 1970s (Lowell and Guilbert, 1970), volcanogenic massive sulfide deposits in the 1970s and 1980s (Franklin et al., 1981), and mesothermal lode gold deposits in the 1980s and 1990s (Hodgson, 1993). These advances were the product of many factors, which conceivably the most important was collaborative ore deposit research between the mineral industry, academia, government agencies, and research organizations.

We are currently in the midst of a similar great advance in knowledge of magmatic Ni-Cu-PGE deposits (Arndt et al., 2006; Barnes and Lightfoot, 2006; Eckstrand and Hulbert, in prep.). Perhaps the most important impetus for this current revolution was the fall of the Iron Curtain, which opened the door in the early 1990s to some of the world's greatest magmatic Ni-Cu-PGE deposits, i.e., the Noril'sk-Talnakh deposits in Russia and Jinchuan deposit in China, to economic geologists trained in open western societies (especially Canada). In addition, the 1993 discovery of the magmatic Ni-Cu deposit at Voisey's Bay, Labrador, Canada, and its inferred origin as a magmatic feeder dike, has brought about a revolution in our collective understanding of the origin of magmatic Ni-Cu-PGE deposits. This revolution can simply be stated as ***“find the magmatic feeder dike and/or channelized magma flow zones”*** to the ore-bearing mafic/ultramafic intrusion(s). A magmatic conduit that experienced repeated influxes of magma appears to be the key to the formation of high-grade, world-class, Ni-Cu-PGE deposits (Naldrett, 1997).

This field trip is truly the result of over two decades of dedicated Duluth Complex research, almost entirely funded by the State of Minnesota and the University of Minnesota, by geologists of the University of Minnesota Duluth's (UMD), Natural Resources Research Institute (NRRI). Steve Hauck, director of the NRRI's Economic Geology Group (EGG), is here thanked for his decades of dedicated research and seeing through that the EGG remains financially solvent through a never-ending series of budgetary crises. As well, the NRRI's Mark Severson – who has logged well over 900,000 feet of Duluth Complex drill core – has without a doubt the more knowledge than anyone on the geology of the mineralized zones within the deposits of the Duluth Complex. The authors of this guidebook have simply built upon this intellectual capital in new ways, including geochemical and 3-D modeling of drill core data, and especially the most basic geologic endeavor, detailed field mapping.

FIELD TRIP TENETS

Many basic attributes of the Duluth Complex Cu-Ni-PGE sulfide deposits resemble those of deposits at Noril'sk, Russia, Jinchuan, China, and Voisey's Bay, Canada that are associated with sulfide mineralization in intrusive feeder zones. Such attributes include shallow tholeiitic intrusions associated with plateau basalt volcanism, external sedimentary sources of sulfur, and openness to repeated magma influx and expulsion. However, the biggest difference between these world-class magmatic ore systems and the deposits of the Duluth Complex is the lack of significant nickel-rich (Ni>Cu) massive sulfide orebodies at Duluth. A critical attribute of the high-grade Noril'sk-Talnakh and Voisey's Bay deposits, not **previously positively identified** in the Duluth Complex, is the location of a magmatic conduit, i.e., the feeder zone.

Several geologists have previously identified possible conduits that may have fed the Partridge River (PRI) and South Kawishiwi (SKI) intrusions of the Duluth Complex. Severson and Zanko (unpub. data) suggest that the Grano fault might mark a possible feeder zone for the Local Boy ore zone at the northeast

end of the PRI. Thériault et al. (2000) postulated that a PRI conduit was present somewhere between the Wetlegs and Dunka Road deposits. Another possible PRI feeder zone may have been along the prolongation of the Siphon fault, which is a Paleoproterozoic growth fault (Graber, 1993) that may have been reactivated during emplacement of the Duluth Complex (Severson and Hauck, 1997). Several authors have suggested the presence of a sub-vertical magmatic feeder beneath the Bald Eagle intrusion (BEI) based on field relations (Weiblen and Morey, 1980) and geophysical attributes (Chandler, 1990).

Peterson (2001b) interpreted the systematic variation in Cu-Ni-PGE mineralization in the Maturi deposit and its extension east to Maturi Extension, as indicative of magma input from the east-northeast via an arcing macrodiike (herein first termed the Nickel Lake Macrodiike (NLM)) that connects the deep-seated source of the BEI and the SKI. Subsequent mapping and research (Peterson, 2002a-f, 2006a, Peterson et al., 2004, Peterson and Hauck, 2005) has built on this model, which recently culminated with the publication of a new detailed bedrock geology map of the area (Peterson et al., 2006). If correct, this model predicts that a Voisey's Bay-type Ni-Cu-PGE massive sulfide body may exist at depth in the area where the NLM meets the SKI, south-southeast of the Spruce Road deposit (see Figs. 4.2 and 4.3). This field trip will investigate evidence that the NLM is a feeder to the SKI, and thus is one of the principal conduits (only?) that brought Cu-Ni-PGE from the Earth's mantle and/or lower crustal magmatic staging chambers into the Earth's upper crust via the NLM into the SKI. This will be accomplished through scientific discussions (hopefully heated) on numerous outcrops (Day 1) as well as visualization of the geology in 3-D and displays of recently drilled cores by Duluth Metals Limited (Day 2).

The fundamental tenet of this field trip is to convey to the participants the notion that science gives one the ability to **imagine reality**. Herein, *science* is geologic research of the Duluth Complex (geologic mapping, drill core logging, geochemical studies, and exploration drilling), and *reality* is new understanding how this magmatic system concentrated and enriched known and potential concentrations of Ni-Cu-PGE at the base of the NLM in the field trip locale as well as adjacent areas of the SKI. The field trip leaders ask the participants (and others who may use the guidebook in the future) to use their imaginations throughout the field trip (or subsequent field excursions) to think about a few basic known and possible realities of the Duluth Complex:

- 1) The Duluth Complex is perhaps the world's largest untapped resource of Ni-Cu-PGE, with multi-billion tons of geologic resources estimated to be worth >1 trillion dollars, (Peterson, 2006c);
- 2) The general geologic setting of the deposits in the Duluth Complex is similar to other world class Ni-Cu-PGE mining camps hosted by rocks in rift settings (Noril'sk-Talnakh, Jinchuan);
- 3) Overall, the ratio of Ni to Cu in Duluth Complex deposits average 1:3;
- 4) Worldwide, the ratio of Ni to Cu in similar deposits averages between 1:1 to 2:1;
- 5) By analogy, there seems to be an enormous mass of missing Ni-rich mineralization;
- 6) At depth, the Nickel Lake Macrodiike may host a large percentage of this missing Ni in the SKI.

MAGMATIC NI-CU-PGE ORE DEPOSIT MODEL

The basic starting point to begin our quest to understand the significance (both geologic and economic) of the Nickel Lake Macrodiike is a quick review of the magmatic Ni-Cu-PGE ore deposit model. There are literally hundreds of recent publications in the geological literature that deal with this important class of ore deposit, and readers interested in this topic may find that Barnes and Lightfoot (2006), Arndt et al. (2006), Eckstrand and Hulbert (in prep.), Naldrett, (1989, 1997, 1999), Naldrett et al. (2000), Li and Naldrett (1999), Lightfoot et al. (1994) and references therein are excellent reviews that describe magmatic Ni-Cu-PGE deposits in general, and major deposits in particular.

Magmatic Ni-Cu-PGE sulfide deposits occur as sulfide concentrations associated with a variety of mafic to ultramafic rocks in four major geological settings: 1) rifts and continental flood basalt settings (Noril'sk-Talnakh, Russia; Duluth Complex, Minnesota; Jinchuan, China); 2) meteorite impacts (Sudbury, Ontario, the only mining camp in this class); 3) komatiite lava flows and related intrusions (Thompson, Manitoba; Raglan, Québec; Kambalda, Australia; Pechenga, Russia); and 4) a variety of miscellaneous tholeiitic intrusions (Voisey's Bay, Labrador; Lynn Lake, Manitoba). The ores are enriched in sulfur, iron, nickel, copper, cobalt, and the platinum group elements (Pt, Pd, Rh, Ru, Ir, and Os) and may contain minor Ag, As, Au, Bi, Hg, Pb, Sb, Se, Te, and Zn. Grade-tonnage diagrams for magmatic Ni-Cu-PGE sulfide deposits/camps are presented in Figure 4.1 which highlights the major camps/deposits listed above.

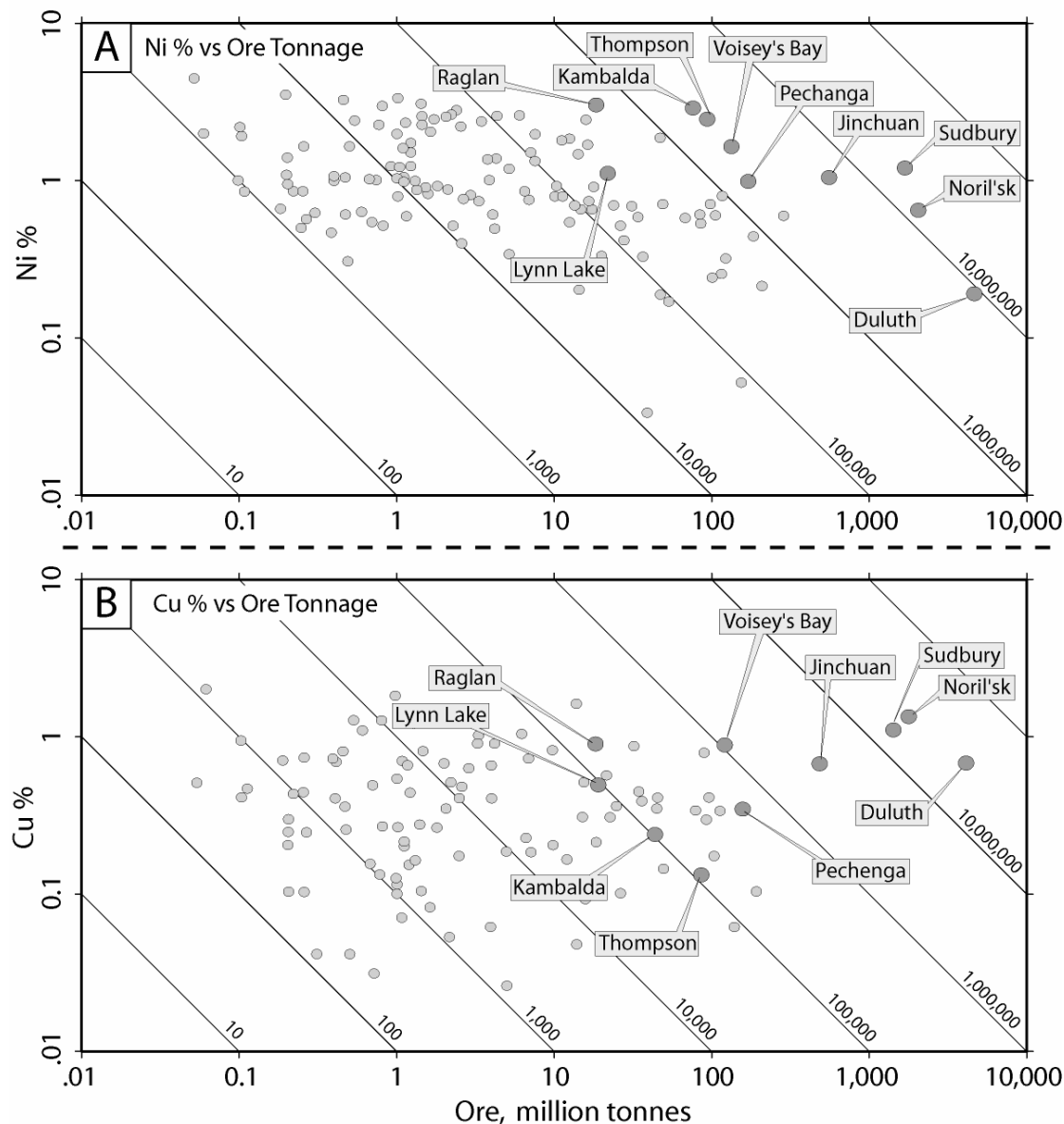


Figure 4.1. Tonnage and Ni grades of magmatic Ni-Cu sulfide deposits; B. Tonnages and Cu grades of magmatic Ni-Cu sulfide deposits. Inclined contours show quantities of contained metals (tonnes) in each figure. Figure modified from Eckstrand and Hulbert, in prep.

The basic intrinsic geological features characteristic of a vast majority of magmatic Ni-Cu-PGE sulfide deposits include: (1) olivine-rich magmas; (2) proximity to a major crustal fault; (3) sulfide-bearing country rocks; (4) chalcophile element depletion in related intrusive or extrusive rocks; (5) field and/or geochemical evidence of interaction between the magma and the country rocks; and (6) presence of, or proximity to, a magma conduit (Naldrett, 1999). Fundamental geologic processes and constraints that together leads to the formation of these deposits include:

- 1) Deposits form as the result of segregation and concentration of droplets of liquid sulfide from mafic or ultramafic magmas;
- 2) Chalcophile elements from the silicate melt partition into the droplets as a result of turbulent magma flow;
- 3) An appropriate physical environment is required so that the sulfide liquid mixes with enough magma to become adequately enriched in chalcophile metals;
- 4) Sulfides must cluster in a restricted locality, generally due to the influence of gravity, so that the resulting metal concentration is of ore grade;
- 5) Massive sulfide concentrations form a high-temperature monosulfide solid solution (MSS);
- 6) As the MSS cools, it exsolves minerals and fractionates;
 - a. Forms a solid cumulate mass of pyrrhotite-rich massive sulfide (enriched in Fe, Ni, Co, Ir, Ru, Rh).
 - b. Forms a liquid residuum that is enriched in Cu, Pd, Pt, Au, and other minor elements, including As, Bi, Te, Sb (which will crystallize later into minerals as the system cools).
- 7) For some time, the Cu-Pd-Pt residual liquid can move and form high grade ore shoots/deposits
 - a. i.e., footwall veins in Sudbury can be 30 wt. % Cu and multi-ounce/ton Pt + Pd.

Possibly the greatest recent advance in understanding Ni-Cu-PGE sulfide deposits has been the appreciation of coherent and compelling scientific arguments that have shown that magma dynamics play a key role in the concentration and metal enrichment of sulfide minerals in these deposits (Naldrett, 1997). These arguments build on the long held notion that sulfide-rich orebodies achieve their concentrations mainly through the settling of sulfide droplets in magmas due to the effects of gravity. The new fundamental tenet has been the realization that **decreases in the flow rate of magmas**, principally due to the geometry and obstructions in magmatic conduits, is the major factor in settling entrained sulfide droplets and forming sulfide-rich orebodies (Eckstrand and Hulbert, in prep.). These geometries include, for example, widened parts of magmatic feeder dikes, the lowest zones of undulating basal zones of conduits, and the location where such dikes enter larger magma chambers. Geologists that are engaged in Ni-Cu-PGE exploration projects should use their imaginations to envision magma dynamics like a person fly-fishing along a stream, and search for deposits behind the large boulders and in the slow-moving pools below rapids where the big rainbow and brown trout are most likely to be found.

REGIONAL GEOLOGIC SETTING, DULUTH COMPLEX

The Duluth Complex and associated intrusions of Keweenawan age (~1.1 billion) in northeastern Minnesota constitute one of the largest mafic intrusive complexes in the world, second only to the Bushveld Complex of South Africa (Miller et al., 2002). These rocks cover a 5,700 square kilometer arcuate area associated with the two strongest gravity anomalies (+50 and +70 milligals) in North America, that imply intrusive roots more than 13 kilometers deep (Allen and others, 1997). The comagmatic flood basalts and intrusive rocks underlying most of northeastern Minnesota were emplaced during the development of the Mesoproterozoic Midcontinent rift, which can be traced geophysically from exposures in the Lake Superior region along a 2,000 kilometer-long, segmented, arcuate path to

Kansas and Lower Michigan. The Duluth Complex is defined as the more or less continuous mass of mafic to felsic plutonic rocks that extends for >275 kilometers in an arcuate fashion from Duluth nearly to Grand Portage (Fig. 4.2). It is bounded by a footwall of Paleoproterozoic sedimentary rocks and Archean granite-greenstone terranes (Peterson and Severson, 2002), and a hanging wall largely of comagmatic, rift related flood basalts and hypabyssal intrusions of the Beaver Bay Complex (Fig. 4.2). In genetic terms, the Duluth Complex is composed of multiple discrete intrusions of mafic to felsic tholeiitic magmas that were episodically emplaced into the base of a comagmatic volcanic edifice between 1108 and 1098 Ma.

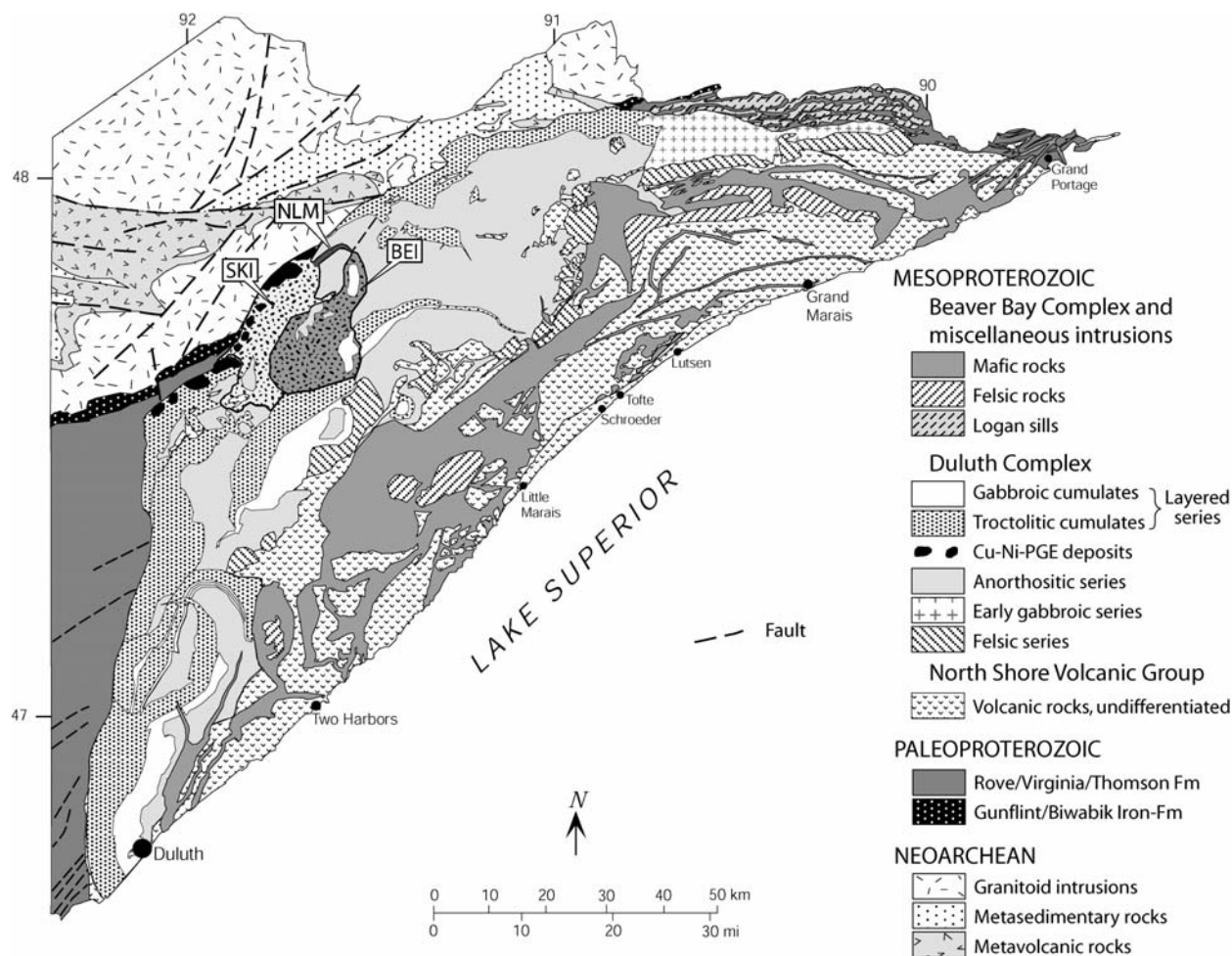


Figure 4.2. Generalized geologic map of northeastern Minnesota. Highlighted intrusions include the Bald Eagle (BEI) and South Kawishiwi (SKI) intrusions, as well as the linking Nickel Lake Macrodiike (NLM) (modified from Miller et al., 2002).

The geology of the Duluth Complex and adjacent areas has recently been described in two major publications by the Minnesota Geological Survey (MGS). These include a 1:200,000 scale regional bedrock geological map of northeastern Minnesota (Miller et al., 2001), and a comprehensive written description of the geology depicted on this map (Miller et al, 2002), commonly referred to as the “bible” by geologists working on Duluth Complex geology. Readers’ interested in more detailed descriptions of the geologic setting of the Duluth Complex should begin their quest for knowledge by downloading these publications from the MGS website (<http://mgssun6.mngs.umn.edu/pub2/>). Within the nearly continuous mass of intrusive igneous rock forming the Duluth Complex, four general rock series are distinguished on the basis of age, dominant lithology, internal structure, and structural position within the complex.

Felsic series—Massive granophyric granite and smaller amounts of intermediate rock that occur as a semicontinuous mass of intrusions strung along the eastern and central roof zone of the complex, emplaced during an early stage magmatism (~1108 Ma).

Early gabbro series—Layered sequences of dominantly gabbroic cumulates that occur along the northeastern contact of the Duluth Complex, emplaced during early stage magmatism (~1108 Ma).

Anorthositic series—A structurally complex suite of foliated, but rarely layered, plagioclase-rich gabbroic cumulates emplaced throughout the complex during main stage magmatism (~1099 Ma).

Layered series—A suite of stratiform troctolitic intrusions that comprises at least 11 variably differentiated mafic layered intrusions that occur mostly along the base of the Duluth Complex. These intrusions were emplaced shortly after the Anorthositic series (~1099 Ma).

This field trip will investigate rocks of the Layered Series – the SKI, NLM, and by implication the BEI – and Anorthositic series rocks in outcrops within and along the margins of the NLM. It is hoped that discussions on the outcrop (Day 1), coupled with examinations of selected sections of mineralized Duluth Metals Limited drill core from the SKI, and visualization of the geology in a 3-D presentation (Day 2) will bring new insight to the geology and mineral potential of the field trip area to the participants. Prior to these investigations, the quick descriptions of the local geology, basal contact-associated styles of Cu-Ni-PGE mineralization, and calculated Cu-Ni grade-tonnage geologic resources within this area that follow will give the field trip participants a better appreciation of the significance of the NLM and its inferred potential for hosting great quantities of Ni-rich sulfide mineralization at depth.

LOCAL GEOLOGIC SETTING

Robust field and geophysical data suggest that the emplacements of the BEI and SKI may be closely linked (Weiblen and Morey, 1980). At the northern margin of the BEI, a macrodiike of well-foliated troctolite (herein termed NLM) arcs northwest to southwest and merges with the middle of the northern SKI (Fig. 4.3). Green et al., (1966) mapped the macrodiike as part of the SKI, but its composition is very similar to the troctolitic phase of the BEI. Peterson (2001b) proposed a model for the mineralization in the SKI whereby the initial emplacement of the intrusion was formed by sulfide-contaminated magmas that emerged from the NLM and flowed southwest between a footwall of Archean granite and a hanging wall of Anorthositic series rocks. Miller et al. (2002) present an important interpretive model for the emplacement of the BEI and SKI via a common feeder system, as well as depict the origin of sulfur saturation of basal SKI magmas (Fig. 4.4) due to contamination from Paleoproterozoic sedimentary rocks (Ripley, 1986). Peterson et al. (2006) completed detailed geological mapping along the western end of the NLM and adjacent SKI, and have defined distinct mappable units in both of these troctolitic bodies.

BALD EAGLE INTRUSION

The BEI is a large (4.5 to 16.5 km x 31 km) troctolitic to gabbroic body that was emplaced partially within Anorthositic series rocks, the SKI, and the Greenwood Lake Intrusion (Fig. 4.2). Weiblen (1965) mapped the well-exposed northern portion of the intrusion and showed that it consists of an outer zone of troctolite and an inner zone of olivine gabbro. In the poorly exposed southwestern portions of the intrusion, field mapping by Green et al., (1966) and Foose and Cooper (1978) showed the BEI and SKI in direct conformable contact. Steep foliation and modal layering (Weiblen, 1965; Green et al., 1966) integrated with a distinct gravity anomaly over the northern BEI imply that the northern part of this intrusion is funnel shaped and necks down to a steep feeder dike. Weiblen and Morey (1980) interpreted the limited cryptic variation (Weiblen, 1965), the steep dip of lamination and layering, and adcumulate nature of the BEI as indicative of its being an open conduit to higher intrusions and perhaps volcanic flows.

A Plan View

Intrusion Base in 3-D

SKI - South Kawishiwi Intrusion

NLM - Nickel Lake Macrodiike

BEI - Bald Eagle Intrusion

Important Xenolith

AN - Anorthosite

Cu-Ni-(PGE) Deposits

SR - Spruce Road

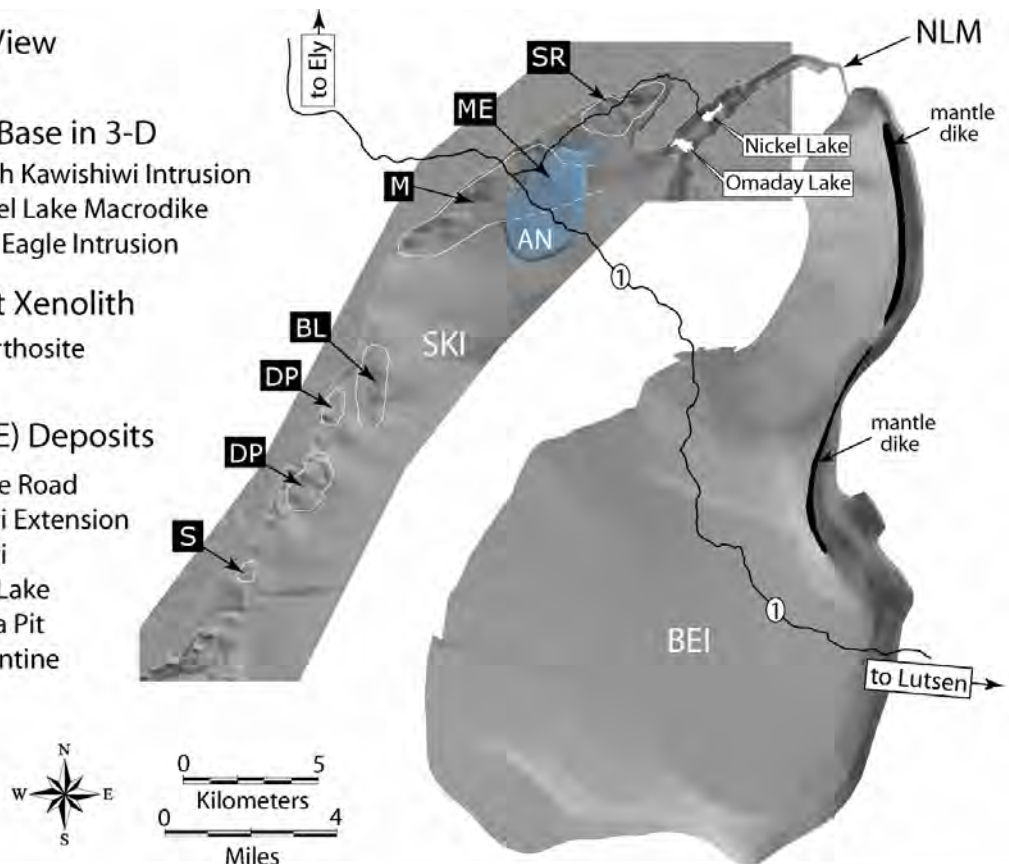
ME - Maturi Extension

M - Maturi

BL - Birch Lake

DP - Dunka Pit

S - Serpentine



B 3-D View

looking down to the southwest

Note: scale varies in view

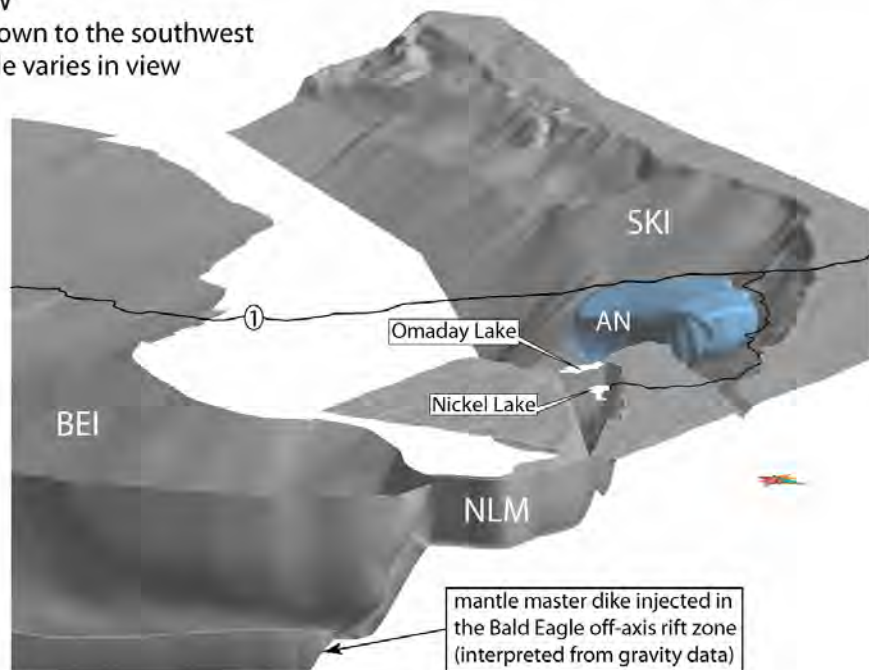


Figure 4.3. Simplified views of an integrated 3-D model of the BEI, NLM, and SKI of the Duluth Complex. A. plan view and B. view to the southwest. Model surfaces built from drill hole piercing points, detailed geological mapping, and interpretation of gravity and aeromagnetic data.

Petrologic observations and geophysical interpretations (Chandler, 1990; Chandler and Ferderer, 1989) suggest that the BEI and SKI were emplaced by successive overplating of magmas from a common feeder centered on the northern BEI and extending along the trace of the NLM that links the BEI and SKI. A model that depicts the origin of the BEI and SKI via a common dynamic feeder by Miller et al. (2002) is presented in Figure 4.4. In a related analogy, Cartwright and Møller-Hansen (2006) have shown that interconnected sill complexes transect the middle to upper crust over a vertical distance of 8-12 km offshore of Norway. The geometry of the gravity and magnetic anomalies of the BEI, as well as the overall Midcontinent Rift is very similar to the pattern of the seismic reflections profiles of active ridge systems (Vislova, 2003). In detail, the geophysical expressions of the BEI have the same shape and dimensions as the “bulls’ eye” pattern of low velocity seismic reflection anomalies along the East Pacific Rise. These anomalies are interpreted to define regions of melt concentrations, i.e., active magma chambers. These data suggest that the BEI could be a “frozen” dynamic magma chamber (Weiblen et al., 2005; Peterson and Hauck, 2005).

The BEI has posed unresolved questions concerning its origin and magmatic significance since its discovery in 1961 (Weiblen, 1965; Weiblen and Morey, 1980; Miller et al., 2002). A number of its characteristics contrast markedly with those of the other mapped intrusions in the Midcontinent Rift: 1) it has a well-defined intrusive contact in Anorthositic series rocks around its northern perimeter; 2) there is a subtle, but recognizable metamorphic contact effect on these anorthositic gabbros; 3) a primary magmatic foliation is well defined by mineral orientation and discoid segregation of plagioclase from mafic phases; 4) foliation measurements define a steeply-dipping asymmetric funnel with the foliation paralleling the contact and grading from steep to horizontal inward; 5) the intrusion consists of two cumulus units, an outer troctolite and inner olivine gabbro; and 6) there is only minor (< a few %) intercumulus material in the cumulates, i.e., clinopyroxene and iron oxides (Weiblen et al., 2005).

SOUTH KAWISHIWI INTRUSION

The South Kawishiwi intrusion (SKI), together with the similar sized Partridge River intrusion (PRI) immediately to the south, are most renown for hosting the largest tonnage of Cu-Ni sulfide mineralization in the world (Naldrett, 1997). The realization that the SKI hosts vast quantities of Cu-Ni mineralization over 50 years ago has lead to the publication of numerous geologic maps, (Green et al., 1966; Bonnicksen, 1974; Foose and Cooper, 1974; Miller et al., 2001; Peterson, 2002e, f; Peterson et al., 2004; Peterson, 2006b; Peterson et al., 2006), articles (Bonnicksen et al., 1980; Weiblen and Morey, 1980; Ripley, 1986; Chandler and Ferderer, 1989; Lee and Ripley, 1996; Hauck et al., 1997; Peterson, 2001b) theses (Weiblen, 1965; Vislova, 2003; Marma, 2003), and reports (Phinney, 1969; Phinney, 1972; Listerude and Meineke, 1977; Morey and Cooper, 1977; Foose, 1984; Dahlberg, 1987; Dahlberg et al., 1989; Kuhns et al., 1990; Severson, 1994; Zanko et al., 1994; Hauck et al., 1997; Peterson, 1997; Peterson, 2001c; Miller et al., 2002; Peterson, 2002d; Patelke, 2003; Severson and Hauck, 2003).

The SKI is shallow dipping (~20° to the east-southeast) sill-like intrusion dominantly composed of troctolitic cumulates that are exposed in an 8- x 32-kilometer arcuate band along the northwestern margin of the Duluth Complex (Fig. 4.2). Footwall rocks include the Paleoproterozoic Virginia Formation in the Serpentine and Dunka Pit deposits, the Paleoproterozoic Biwabik Iron Formation in the Dunka Pit and Birch Lake deposits, and the Archean Giants Range batholith from the northern Birch Lake deposit north to the Spruce Road deposit (see Fig. 4.3 for deposit locations). The presence of shallow-dipping Biwabik Iron Formation inclusions as far north as the Spruce Road deposit indicates that the majority of Paleoproterozoic units were assimilated and removed from the footwall during emplacement of the SKI, leaving the Giants Range batholith as the dominant footwall rock type. Alternately, the Virginia and Biwabik Iron Formations may simply have been largely eroded prior to the development of the Mid Continent Rift. Also present as inclusions in the SKI are mafic volcanic hornfels (North Shore Volcanic Group), quartz sandstone hornfels (either the Puckwunge or Nopeming sandstones), and anorthosite (of

the Anorthosite series). Anorthositic series rocks abut the SKI on the northeast – and enclose an interpreted SKI feeder dike (the NLM) that extends farther northeast – the PRI forms the southern sidewall of the SKI, and the BEI and Anorthositic series rocks overlie the SKI to the east (Fig. 4.2).

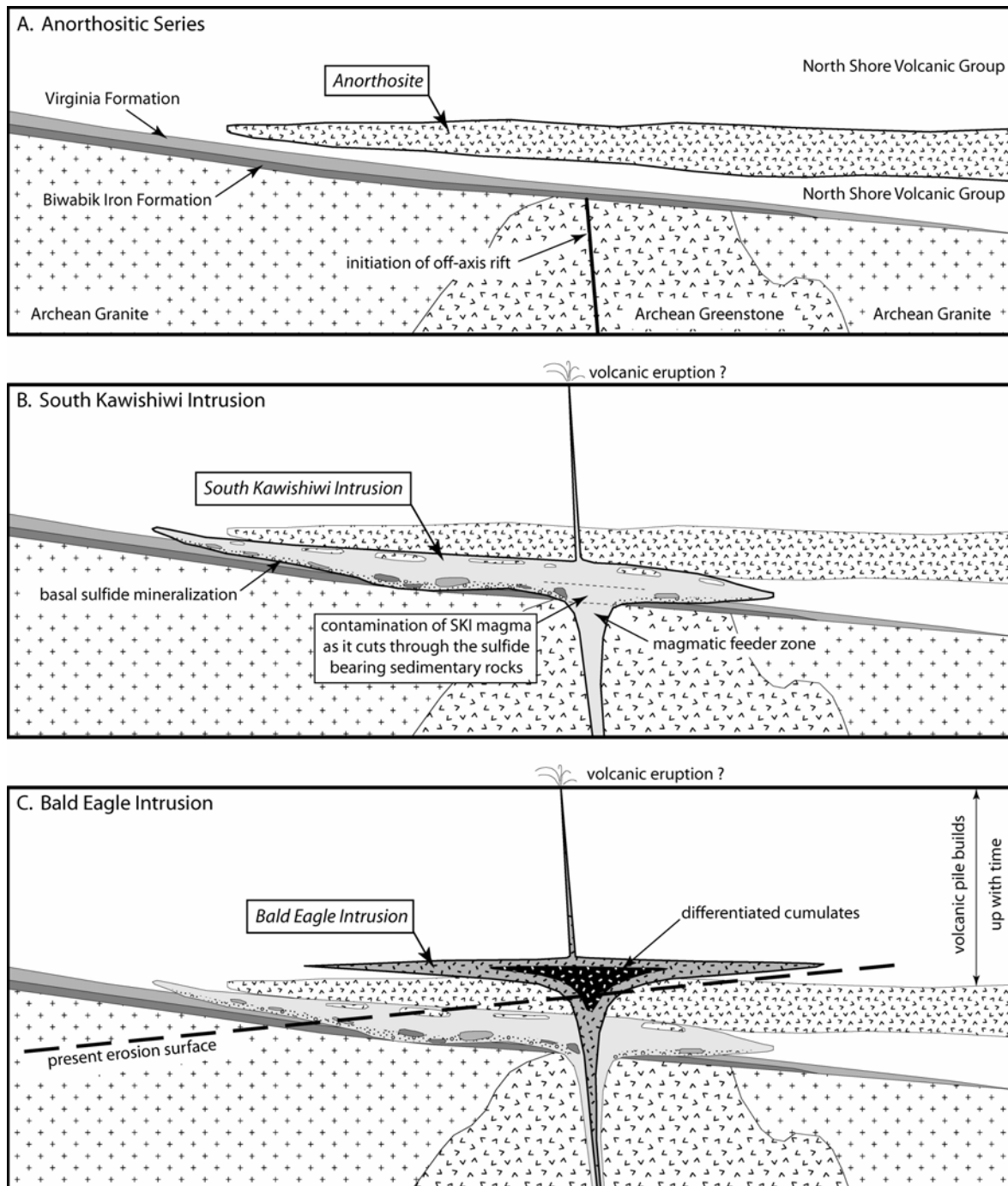


Figure 4.4. Interpretive model which depicts the emplacement and mineralization mechanisms of the South Kawishiwi and Bald Eagle intrusions via a common feeder system. A) Intrusion of plagioclase crystal mushes into volcanic rocks to create Anorthositic series rocks; B) Intrusion of the SKI below Anorthositic series rocks, with the early basal units reaching sulfur saturation via contamination from Paleoproterozoic sedimentary rocks; and C) Intrusion of the BEI above Anorthositic series rocks. Modified from figure 6.15 of Miller et al. (2002).

On the regional Duluth Complex map of Miller et al. (2001), the SKI is subdivided into five major map units. These are, from the base upward,

1. Heterogeneous **sulfide-bearing** troctolite, gabbro, and norite with localized hornfels inclusions;
2. A thick unit of subophitic to ophitic augite troctolite;
3. Discontinuous and localized layers of poikilitic leucotroctolite;
4. A thick homogeneous sequence of ophitic troctolite; and
5. A thick uppermost sequence of homogeneous troctolite that contains numerous anorthositic layers.

Severson (1994) and Zanko et al. (1994) further subdivided the SKI into 17 different lithostratigraphic units that are present in over 180 drill holes over a strike length of 31 kilometers. Sulfide mineralization is confined to the BH, BAN, UW, and U3 units near the base of the intrusion, and to a lesser extent the U1, U2, and PEG units. Major marker horizons that are correlated in drill holes include three horizons with abundant cyclic ultramafic layers (U1, U2, and U3 units) and a pegmatite-bearing unit (PEG unit) that was initially recognized by Foose (1984). The understanding of the significance of a large anorthositic inclusion (Fig. 4.3), originally intersected in six deep drill holes east of the Maturi deposit, and its role in magma dynamics of the SKI has been a key feature in the development of an exploration model for Duluth Metals Limited's Maturi Extension deposit (Peterson, 2001c).

Basal SKI Mineralization

Basal mineralization within the South Kawishiwi intrusion has traditionally been divided into five distinct deposits: 1) Serpentine, 2) Dunka Pit, 3) Birch Lake, 4) Maturi, and 5) Spruce Road. Recent drilling by Duluth Metals Limited has confirmed the existence on an additional economically significant deposit east of Maturi first envisioned by Peterson (2001c), named the Maturi Extension deposit (Fig. 4.3). Although the style of mineralization in all of the deposits is dominated by disseminated Cu-Ni sulfides, differences occur between the deposits in igneous stratigraphy, sulfide mineralogy, Cu-Ni and PGE grade, mineralization thickness, and contained tonnes. In addition to mineralization spatially associated with the base of the intrusion, a distinct zone of Cu and PGE-enriched mineralization occurs thousands of feet above the base of the intrusion within linear zones in the South Filson Creek deposit (Kuhns et al., 1990). Compilation and analysis of drill hole assay data by Peterson (2001, 2002a-d) has led to new understanding of two distinct styles of mineralization associated with the base of the SKI. These distinctive styles of mineralization are spatially coherent, i.e., the boundaries between them are linear (Fig. 4.5), and Peterson (2001b, c) informally termed them *open* and *confined*, which are described in more detail below.

"Open" - vertically extensive (can be > 450 meters) mineralization with low - high Cu-Ni grade and low Au+PGE grades. Cu-Ni grades commonly increase towards the basal contact although mineralized zones are typically erratic in their spatial extent and grade. Restricted zones of massive sulfide locally occur at, and/or immediately below, the basal contact. The erratic pattern of mineralization in part mirrors the lithologic heterogeneity of the basal units and may reflect repeated input of small pulses of barren and sulfur-contaminated magma. Examples of this "Open" style include the Spruce Road, Serpentine, and Dunka Pit deposits. The Serpentine deposit is unique within this group as it contains significant tonnage of pyrrhotite-rich massive sulfide at the basal contact that is associated with an immediate footwall sulfide source (Zanko et al., 1994).

"Confined" - vertically restricted (< 150 meters) mineralization with moderate - high Cu-Ni grades and moderate to very high (locally) Au+PGE grades. Cu-Ni grades typically are the highest near the top of the mineralized zone (upper BH into U3) and gradually decrease with depth toward the basal contact. Only limited zones of massive sulfide occurring at, and/or immediately below, the basal contact have been identified. For example, the upper portion of the mineralized zone within the Maturi deposit (which averages ~150 feet thick) commonly exhibits copper values nearing 1.0%

that decrease to ~0.25% at the basal contact. The spatial continuity of both the igneous stratigraphy and Cu-Ni-PGE grades of this style of mineralization point toward larger sustained inputs of magma (and/or more turbulent input of magma, thus higher fractionation of base- and precious-metals into the sulfide fraction) than the "Open" style. Examples of the "Confined" style include the Maturi, Maturi Extension, and the Birch Lake deposits.

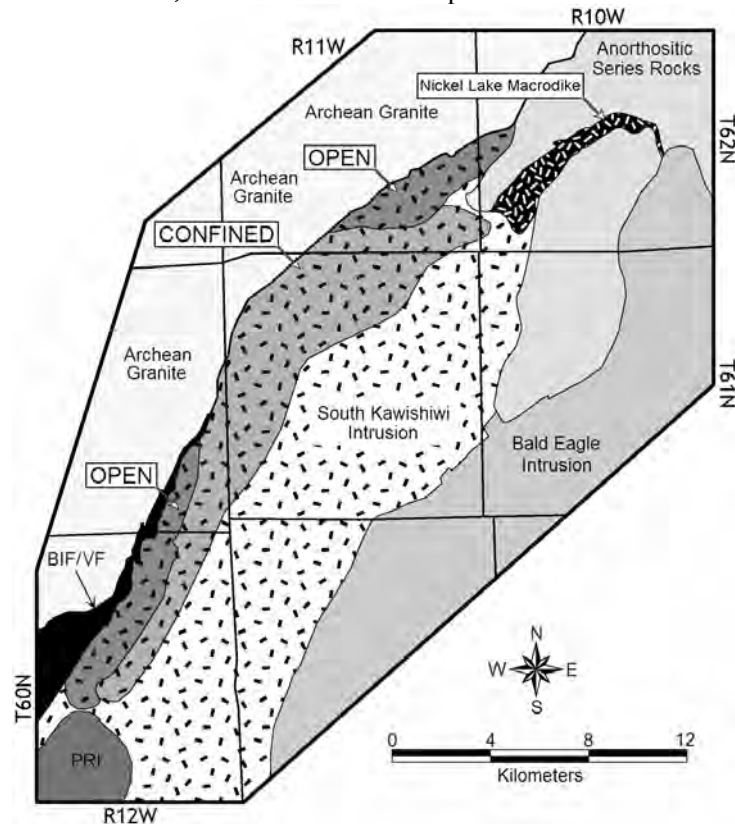


Figure 4.5. Location map of Open and Confined styles of mineralization in the South Kawishiwi Intrusion.

Recently completed (Peterson, 2002a, 2006a, b, c) and ongoing research on the distribution of Cu-Ni-PGE in the SKI integrates detailed geological mapping, drill hole logging, assay compilation, grade-tonnage calculation, recalculation of assay data to 100% sulfide, and 3-D visualization in order to imagine the coupled magmatic dynamics and mineralization history of the area. The most profound difference in the Open and Confined styles of Cu-Ni-(PGE) mineralization is perhaps best revealed in plots of assay data that have been recalculated to 100% sulfide compositions (Kerr, 2001, Naldrett et al., 2000). Such plots give one clues to the complicated history of immiscible sulfide droplets within SKI magmas, and could be used to calculate R-factors (herein the mass transfer of chalcophile elements between immiscible sulfide and silicate liquids in magmas) of the mineralization systems. As well, discrimination of different styles of mineralization and their internal metal budgets lets one come to grips with item 3 of the previously described Ni-Cu-PGE ore deposit model, that states, *“An appropriate physical environment is required so that the sulfide liquid mixes with enough magma to become adequately enriched in chalcophile metals”*. Geochemical plots of drill core assay data recalculated to 100% sulfide for the Open and Confined styles of mineralization are presented in Figure 4.6, and reveal profound differences in the metal budgets of the sulfide mineralization. Any genetic ore deposit model and/or mineral exploration model used to search for additional mineralization in the SKI must attempt to explain these differences in context with the overall NLM-SKI magmatic system, i.e. its timing, geometry, and magmatic plumbing.

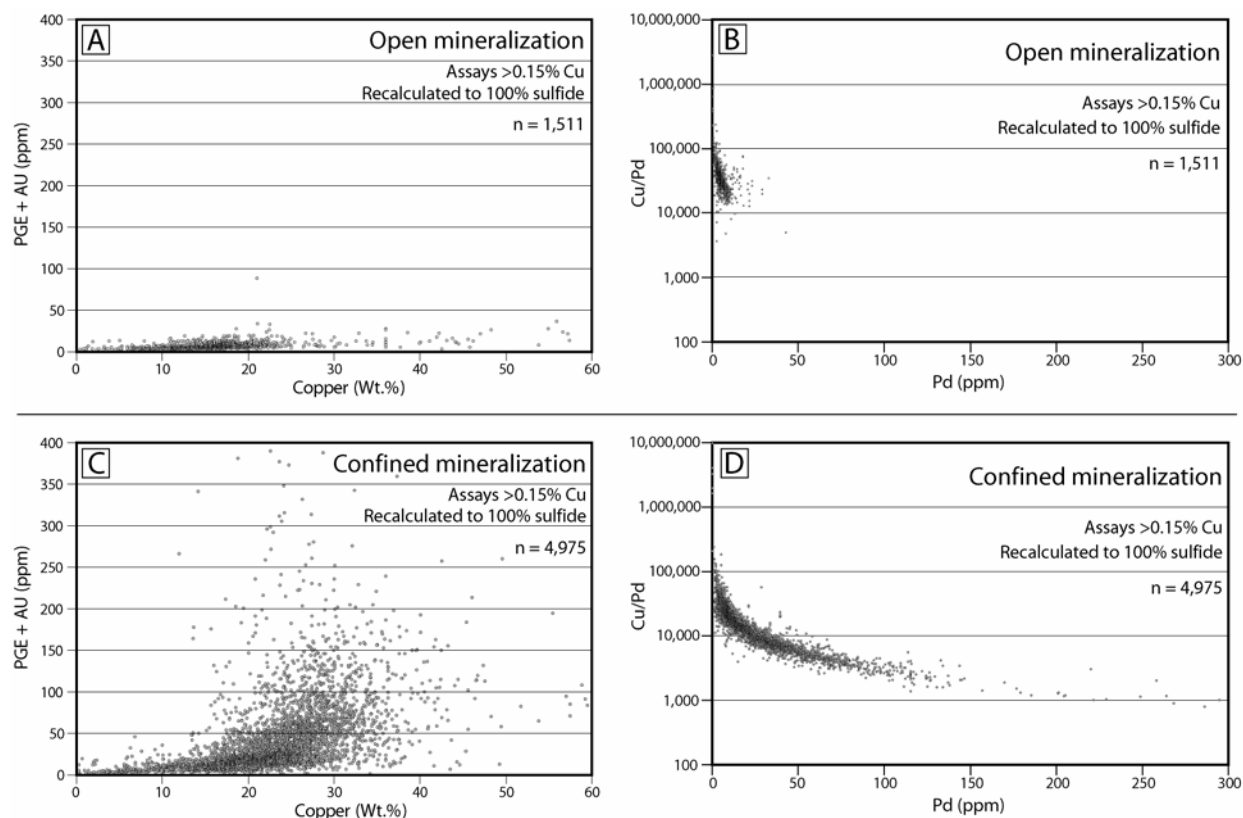


Figure 4.6. Geochemical plots utilizing recalculation of assay data to 100% sulfide (Kerr, 2001) to discriminate variations of metals in sulfide minerals in the two major styles of basal contact associated mineralization in the SKI. A) Open-style PGE+Au vs. Cu plot, B) Open Style Cu/Pd ratio vs. Pd plot, C) Confined-style PGE+Au vs. Cu plot, D) Confined-style Cu/Pd ratio vs. Pd plot.

Recent research in developing Cu and Ni grade maps (Peterson, 2002d) for deposits in the SKI, coupled with the realization that there are two distinct styles of basal-contact associated mineralization, has led to the publication of inferred grade-tonnage estimates of Open and Confined styles of mineralization (Peterson, 2002c). Grade and tonnage data categorized into the Open and Confined styles of Cu-Ni mineralization within the South Kawishiwi intrusion are given in Table 4.1, which used all publicly available assay data in the year 2001. These mineral resource estimates have been calculated by the senior author (Peterson) following as much as possible the definitions and guidelines adopted by the Canadian Institute of Mining, Metallurgy, Petroleum (CIM "Standards on Mineral Resources and Reserves") in August 2000 (Postle et al.). Due to the fact that they rely on historic data from a variety of different sources, the level of data verification expected by the guidelines was not possible in all cases. Inherent uncertainties in the estimation and accuracy of these mineral resource estimates are a function of the quantity and quality of the available drill hole assay data and the quality of the methods used to determine them, which for the Spruce Road deposit are outlined in Peterson (2002d).

It is hoped that the data presented so far in this field guide, once integrated together in one's mind, leads to the conclusion that understanding the geologic history of the NLM may lead to profound advances in our understanding of Cu-Ni-PGE mineralization in the SKI. Such understanding may ultimately lead to the discovery of the "Missing Nickel" in the Duluth Complex Cu-Ni-PGE deposits, thus redefining the district as the Duluth Complex Ni-Cu-PGE deposits.

Table 4.1. Cumulative inferred mineral resource estimates for the Open and Confined styles of mineralization in the South Kawishiwi intrusion. Open style modeled to 1500 Ft. (457m.) and Confined style modeled to 500 Ft. (152m.) above base.

Copper Cutoff	Open Style Mineralization			Confined Style Mineralization		
	Cu %	Ni %	Tonnes	Cu %	Ni %	Tonnes
1.05	1.043	0.559	114,000	1.035	0.233	3,200,000
1.00	1.029	0.415	229,000	1.010	0.222	24,232,000
0.95	0.969	0.388	800,000	0.989	0.227	37,034,000
0.90	0.933	0.324	1,715,000	0.962	0.233	52,579,000
0.85	0.924	0.327	1,943,000	0.936	0.238	69,038,000
0.80	0.844	0.313	4,572,000	0.893	0.244	100,585,000
0.75	0.818	0.304	6,058,000	0.852	0.243	139,905,000
0.70	0.758	0.290	11,773,000	0.802	0.239	206,200,000
0.65	0.685	0.246	33,262,000	0.748	0.232	316,386,000
0.60	0.644	0.227	61,608,000	0.693	0.222	505,669,000
0.55	0.597	0.210	122,874,000	0.641	0.211	785,936,000
0.50	0.547	0.194	245,862,000	0.590	0.197	1,230,797,000
0.45	0.502	0.184	448,519,000	0.553	0.187	1,667,886,000
0.40	0.461	0.171	747,645,000	0.517	0.176	2,181,785,000
0.35	0.421	0.158	1,154,558,000	0.482	0.165	2,754,206,000
0.30	0.374	0.143	1,862,655,000	0.446	0.153	3,428,584,000
0.25	0.329	0.129	2,894,340,000	0.408	0.141	4,254,298,000
0.20	0.286	0.114	4,343,339,000	0.369	0.128	5,230,432,000
0.15	0.242	0.099	6,371,617,000	0.329	0.115	6,406,364,000
0.10	0.201	0.083	8,950,828,000	0.283	0.099	7,998,811,000
0.05	0.156	0.065	12,606,986,000	0.219	0.078	10,943,215,000
0.01	0.107	0.045	19,087,646,000	0.131	0.047	18,987,747,000

Note:

- 1) Cutoff grade intervals calculated from summation of modeled data (see Peterson, 2002d). For example, the 0.55 Cu cutoff includes all gridded data that falls between 0.575 and 0.525 wt. % Cu.
- 2) Cu%, Ni%, and tonnes represent cumulative addition of data from the Cu cutoff value and all data of higher grade.
- 3) Low Cu cutoff data dominantly represents barren rock within the mineralized zone of the Open style and barren rock above the mineralized zone within the Confined style.
- 4) Tonnes rounded to the nearest 1,000.
- 5) All intervals not assayed were assigned Cu and Ni values of 0.00 wt% and integrated into the model.

NICKEL LAKE MACRODIKE

Detailed geological mapping at a scale of 1:5,000 by the authors and Chris White, a Masters Candidate in the Department of Geological Sciences at the University of Minnesota Duluth, was completed in the late summer and fall of 2006, and published at a scale of 1:10,000 by the NRRI (Peterson et al., 2006). This map, available online at <http://www.nrri.umn.edu/egg/REPORTS/MAP200604/MAP200604.html> is the foundation upon which the field component of this trip will be based. During the course of this mapping, approximately 1,000 outcrops along nearly 100 kilometers of field traverses were examined to identify and confirm the internal lithologic variability, contact relationships, and structure of the western extent of the NLM, the adjacent SKI, and bounding rocks of the Anorthositic series. The authors wish to acknowledge Dr. Paul Weiblen (emeritus professor of geology at the University of Minnesota) for his keen insight on the geology of the area and Dr. George Hudak and undergraduate student Jeremiah

Gowey of the University of Wisconsin Oshkosh for assistance in mapping outcrops around and south of Omaday Lake. As well, a one day field excursion to Nickel Lake with Dave Peck (Anglo American), Harry Noyes (Encampment Resources), and Theodore DeMatties (consulting geologist) prior to the mapping campaign developed new insight on identifying dynamic magmatic systems in the field to the senior author (Peterson).

Additional reconnaissance mapping in early November by Dean Peterson was completed to field check compiled outcrop locations depicted on the 1957 INCO map of the Spruce Road Deposit and the 1968 Hanna Mining map of the South Filson Creek deposit (both of which are publicly available in the DNR archive at Hibbing, Minnesota). The reconnaissance mapping confirmed the location of gossanous Cu-Ni bearing INCO outcrops and reconfirmed the outstanding field mapping of all types of Duluth Complex rocks by Hanna Mining Company geologists of the late 1960s.

The NLM is a northwest to southwest-trending (Fig. 4.3), steeply dipping, asymmetric troctolitic and gabbroic intrusion interpreted to be a feeder dike for the northern portions of the SKI. The macrodike is interpreted to be located within a major rift-parallel normal fault (down to the southeast) now obscured by intrusion of NLM igneous rocks. Regional southward tilting (based on the deep level of erosion of the northern Bald Eagle Intrusion directly east of this area) leads to the interpretation that the southwest end of the NLM (near Omaday Lake) is structurally higher than the northeastern portion of the dike, and represents the location where magma flow changed from dike-like to sill-like, as it exited the dike – thus the magma velocity slowed – and entered the growing SKI magma chamber. Excellent potential exists for Ni-Cu rich massive sulfide at the basal contact where the dike enters the SKI (Section 31, T62N, R10W). The dike is composed of three main units: 1) inclusion-rich, locally sulfide-bearing, heterogeneous troctolite (unit N-Th); 2) layered troctolite, melatroctolite, and dunite (unit N-Tl); and 3) a late, cross-cutting, coarse-grained to pegmatitic oxide-rich, olivine-gabbro to melagabbro (unit N-xG).

Description of NLM Map Units

The basis for all of the field aspects of this field trip is the aforementioned recently published bedrock geologic map of the NLM (Peterson et al., 2006). As well, the description of “Field Trip Stops” to follow differ from most geology field trip guidebooks in that we are simply going to take some walks in the bush, mostly along logging roads and snowmobile trails, look at numerous outcrops of the NML and adjacent rocks, and discuss the geology. We are NOT GOING TO SPECIFIC OUTCROPS to try to make a case for our interpretations; instead we urge you to use your imagination while we look at outcrops and add to the conversation. Instead of writing detailed descriptions of specific outcrops we’ll visit during the “stops” in the field trip, the authors have instead decided to simply copy verbatim the *Description of Map Units* from the map NRRI/MAP-2006-04 (Peterson et al., 2006) below, and use this as a reference during the field trip. The rocks of the NLM include, generally from youngest to oldest:

Oxide Gabbro (N-xG) - Dark-grey, coarse-grained to pegmatitic, recessive weathered, oxide-rich (magnetite and ilmenite), olivine-gabbro to melagabbro. Contains small inclusions of anorthosite (unit N-Ai), basalt (unit N-Bi), and troctolite (unit N-Th). Interpreted to be the youngest phase of the dike based on inclusion types and cross-cutting relationships.

Layered Troctolite to Dunite (N-Tl) - Grey to black, medium-grained, well-layered troctolite, melatroctolite, and dunite. Lamination of plagioclase and olivine parallel to modal layering is commonly observed as well as igneous scours and crossbedding. Small inclusions of anorthosite (unit N-Ai) and basalt (unit N-Bi) rare. Layering possibly developed as the up-welling magma streams through the dynamic (expanding) feeder dike. A constant temperature appropriate to plagioclase-olivine crystallization is maintained by a balance between the heat content of the incoming magma plus the heat of crystallization and the heat loss through the chamber walls.

Plagioclase and olivine are left behind and oriented/segregated on the walls of the expanding chamber (Weiblen, pers. comm.). Sulfide noted locally in outcrop in the SE corner of Section 31.

Heterogeneous Troctolite (N-Th) - Light to dark grey, medium- to coarse-grained, inclusion-rich, heterogeneous troctolitic rocks with local igneous scour structures. Unit composed of intermixed troctolite, anorthositic-troctolite, melatroctolite, and gabbroic phases surrounding numerous local country rock (unit N-Ai) and exotic (units N-IF and N-Bi) inclusions that are elongate parallel to the macrodiike, as well as hosts the sulfide-bearing unit N-Ts. Interpreted to be the initial highly dynamic magmatic phase of the dike that carried exotic inclusions from deep in the crust to their present level.

Sulfide-Bearing Troctolite (N-Ts) - Rusty weathered, medium- to coarse-grained, sulfide-bearing, heterogeneous troctolitic and gabbroic rocks. Generally forms recessive weathering, Fe-stained, gossanous outcrops near, but not at, the northern margin of the macrodiike. The current extent of this unit on the map is confined to those areas with outcrop, which in reality may be much more extensive as these outcrops generally end along linear swampy areas.

Anorthosite Inclusion (N-Ai) - Light-grey, medium- to coarse-grained troctolitic-anorthosite, commonly with 1-2 cm poikilitic olivine pits. The large anorthosite inclusions at the southwest end of the dike (around Omaday Lake) are interpreted to represent a "logjam" of blocks that quit moving due to the decreased speed of the macrodiike magmas as they entered the South Kawishiwi Intrusion magma chamber. Includes blocks within the SKI adjacent to the Nickel Lake macrodiike.

Basaltic Hornfels Inclusion (N-Bi) - Grey, fine-grained, steeply-dipping to vertical, granoblastic, locally magnetic, massive to amygdaloidal basaltic hornfels. Includes a highly magnetic block within the SKI adjacent to the Nickel Lake macrodiike.

Biwabik Iron Formation Inclusion (N-IF) - Well-bedded, steeply-dipping, recrystallized (layered magnetite and pyroxenite) iron-formation commonly with disseminated Cu-Ni sulfides. Forms an intense localized positive magnetic anomaly.

Anorthositic Series - Subsuite of the Duluth Complex composed predominantly of plagioclase cumulates displaying complex internal structure and lacking obvious signs of in situ differentiation. Occurs throughout the Duluth Complex as anorthosite, troctolitic-anorthosite, and gabbroic-anorthosite, commonly poikilitic.

Anorthositic Rocks Undivided (A-tA) - Mixed group of anorthositic cumulates occurring as large sill-like masses and as inclusions within troctolitic cumulates. Common rock types include troctolitic-anorthosite, leucotroctolite, anorthosite, and olivine-bearing gabbroic-anorthosite. Olivine ranges from 2 to 15 percent in mode and from granular to poikilitic in texture, with oikocrysts ranging from 1 to 3 centimeters in diameter. Plagioclase mode ranges from 75 to 95 percent and varies from being non-foliated to well-foliated. Inclusions range in size from a few centimeters to elongate bodies hundreds of meters long that are parallel to foliation in the enclosing troctolite.

DAY 1, FIELD TRIP STOPS

Traverse #1

The location of the first field trip traverse (#1) is given in Figure 4.7. This walk in the bush begins within numerous outcrops of the bounding Anorthositic series rocks (map unit A-tA) on the northern margin of the NLM. The trail will take us southwest into inclusion-rich heterogeneous troctolitic rocks (map unit N-Th) and into southwestern most zone of known Cu-Ni-(PGE) mineralization in the NLM (map unit N-Ts), which in this location is associated with a large, sub-vertical inclusion of Biwabik Iron Formation (map unit N-IF). One must try to imagine from where such an inclusion came from (see Figure 4.4), how

it relates to the model that the NML is a feeder dike to the SKI, why it and other large mapped inclusions in the NML (Fig. 4.7) are concentrated at the southwestern end of the NML (magma velocity). Think back to the previously described fundamental tenet of Ni-Cu-PGE magmatic sulfide deposits, “... realization that decreases in the flow rate of magmas, principally due to the geometry of the conduit, is a major factor in settling entrained sulfide droplets and forming sulfide-rich ore bodies...”.

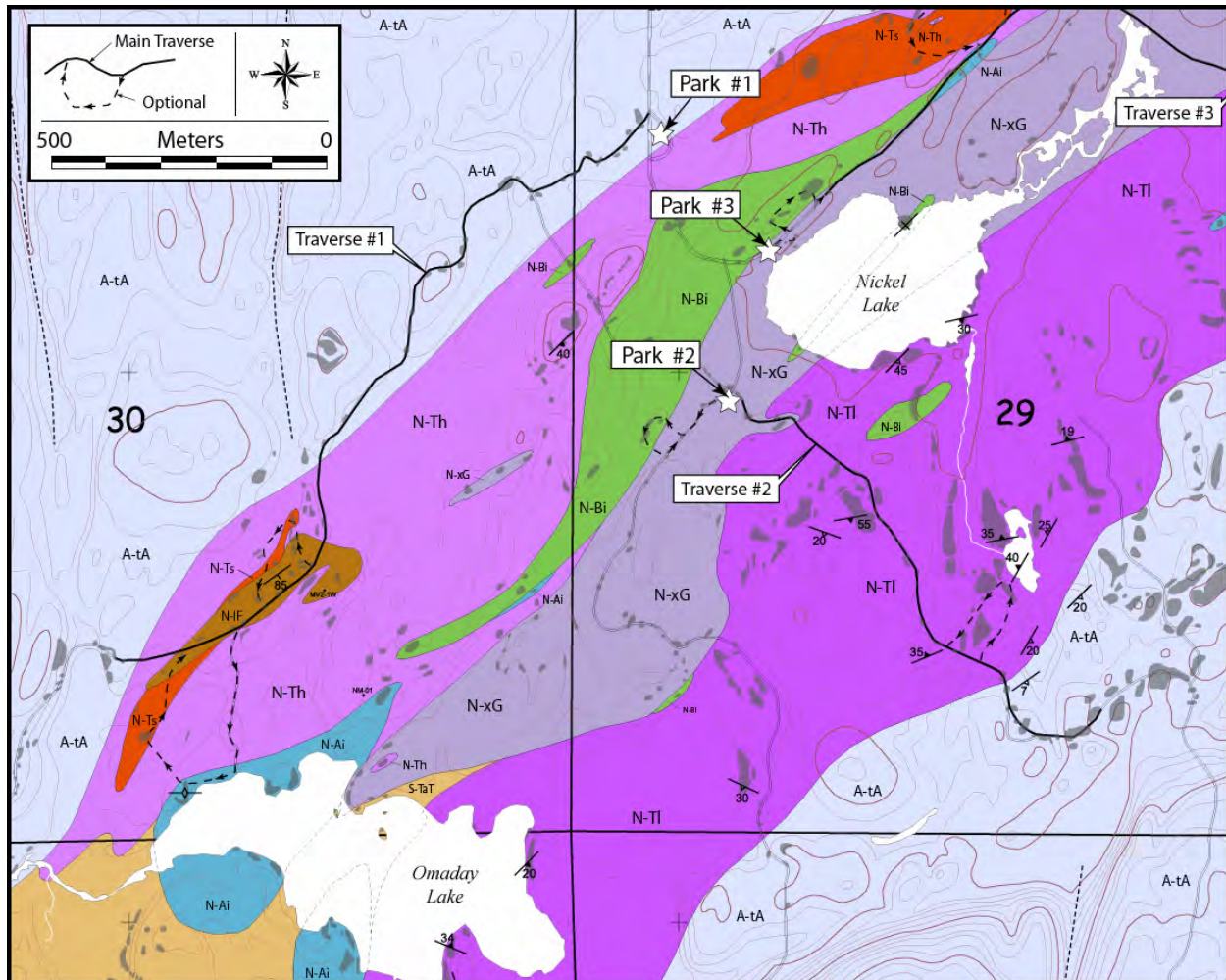


Figure 4.7. Bedrock geology map of portions of the southwestern end of the Nickel Lake Macrodike in Sections 29 and 30, T62N, R10W. Dark lines are superimposed locations of field trip traverse #1, and traverse #2 along logging roads and snowmobile trails. Dashed lines represent short traverses through the bush to additional known outcrops.

Traverse #2

The location of the second field trip traverse (#2) is presented in Figure 4.7. This walk to the southeast along a logging road/snowmobile trail begins in coarse-grained, oxide-rich olivine gabbro to melagabbro of map unit N-xG. Contact relationships between the N-xG and layered troctolitic to dunitic rocks (map unit N-TI) can be complex, and we'll investigate these relationships early on the traverse. The bulk of this traverse will be through the widest section (>650 m) of unit N-TI that has been mapped in the NLM. Careful attention should be directed towards igneous textures in the N-TI, and what they imply to magmatic processes (expanding dike with time, modal layering, pasting plagioclase and olivine phenocrysts on dike walls, solidification fronts, etc...). Optional traverses shown include a walk to

outcrops of the very large (~ 1.5 km long) basalt inclusion (map unit N-Bi), and to exceptional exposures of layered troctolite and dunite around a small beaver pond.

Traverse #3

The location of the third field trip traverse (#3) is presented in Figure 4.8. This walk to the north-northeast of Nickel Lake will traverse through most of the main units identified in the NLM, including map units N-Th, N-xG, N-Tl, N-Ts, N-Bi, and N-Ai. An important walk through the bush (optional traverse on Figure 4.8) will visit several Cu-Ni-(PGE) mineralized outcrops near the northwestern margin of the NLM, and allow us to view some spectacular exposures, around the margin of a drained beaver pond, of the Anorthositic series rocks immediately northwest of the NLM. The authors cannot speak too strongly on the importance of finding these types of exposures (totally free of lichen, moss, trees, etc...), early in a mapping program, as they provide proxies for subsequent mapping of outcrops deep in the bush, where trees, shrubs, shade, forest litter, dirt, and black flies partially obscure exposures and/or ones willing to observe them.

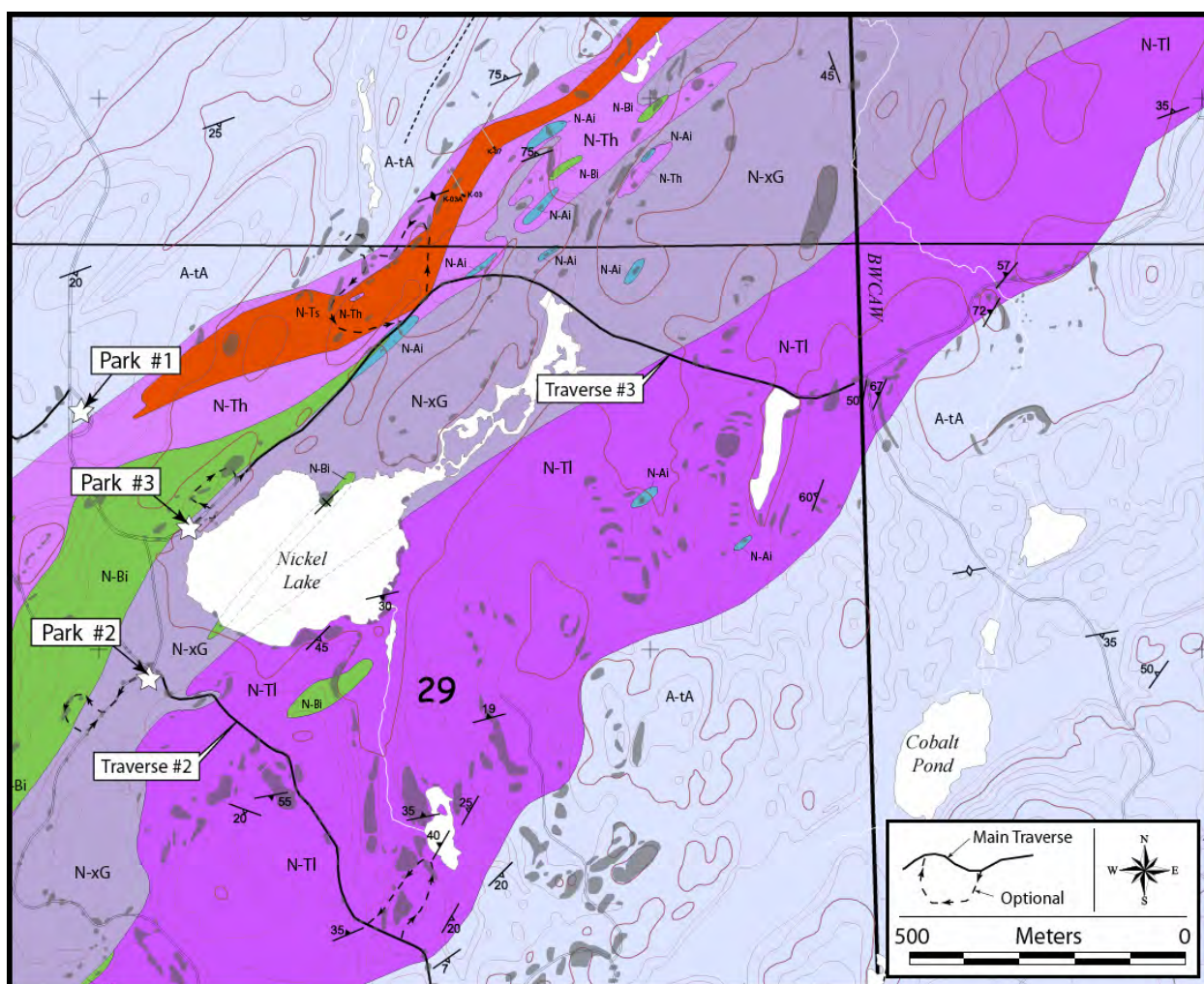


Figure 4.8. Bedrock geology map of portions of the Nickel Lake Macrodike in the vicinity of Nickel Lake, Section 29, T62N, R10W. Dark line in the center of the image is the superimposed location of field trip traverse #3 along a logging road. Dashed line in the northwest quadrant represents a short traverse through the bush into the Cu-Ni-PGE mineralized northern zone of the NML. Note that beaver dams along the main traverse may cause very wet and/or impassable conditions.

Field relationships that offer evidence that a dike-like mafic to ultramafic intrusion is a conduit through which magma ascended upwards in the Earth's crust all lead back to the fundamental tenets of the Ni-Cu-PGE ore deposit model. Such relationships provide evidence that the rocks formed in a dynamic, sulfide-bearing magmatic system (once again, think like a person fly fishing a trout stream) that include: 1) early phases should be inclusion-rich (some of which should be from a deeper crustal level) and form as the igneous conduit breaches upwards into the Earth's crust; 2) imbrication and/or elongation of entrained country rock inclusions parallel to igneous foliation; 3) igneous scour structures; 4) prominent steeply-dipping igneous foliation and localized disruption due to magmatic injection; 5) cross-bedding of modal layering; 6) evidence of sulfide mineralization; and 7) evidence that that magma velocity varies. One of the authors' goals of the field trip is to expand this list based on conversations on the outcrop with the participants. Outcrop photographs given in Figure 4.9 show a few of these lines of evidence for the dynamic nature of the NLM.

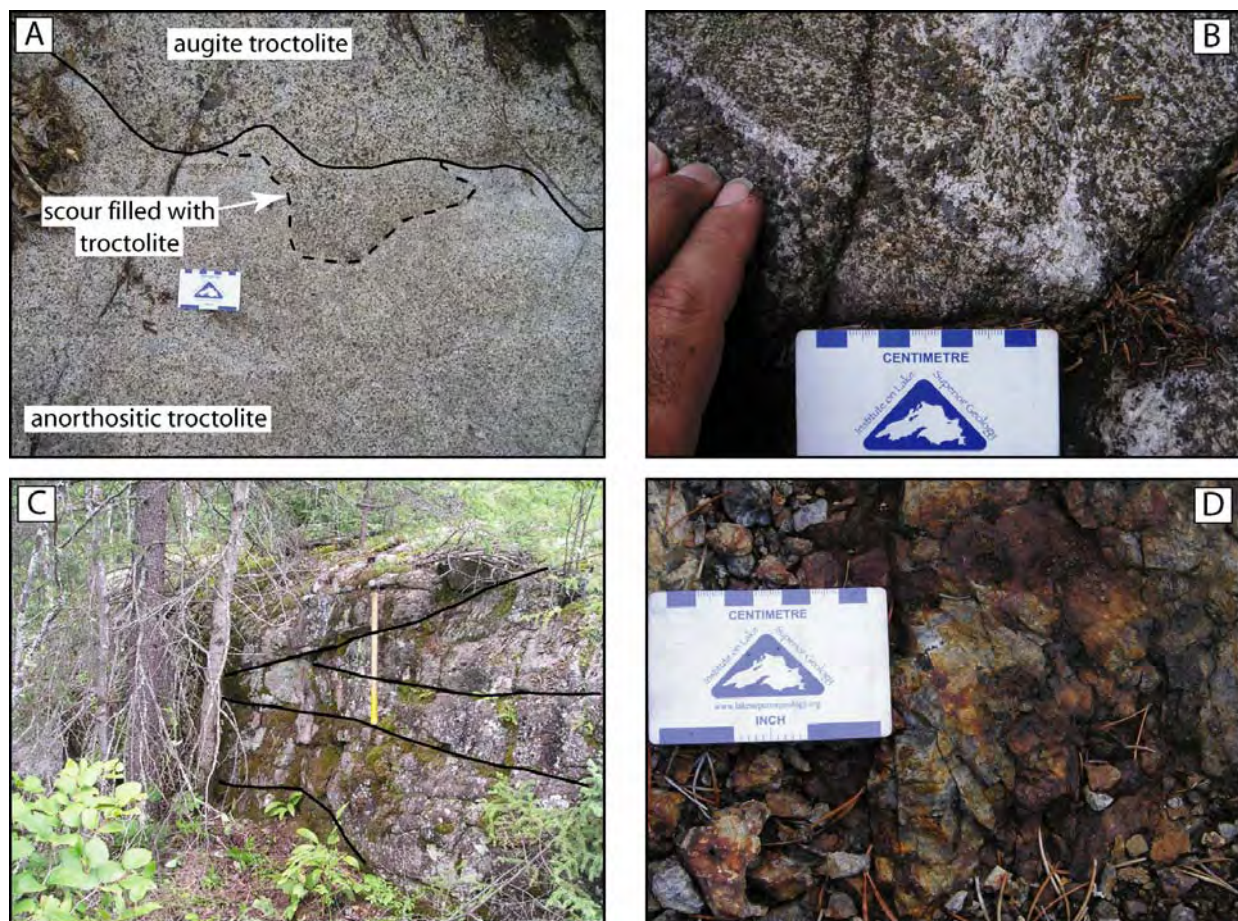


Figure 4.9. Photographs of selected outcrops that give us clues to the dynamic nature of the magmatic processes which ultimately led to the formation of the NLM, and by interpretation, the northern portion of the SKI and its associated Cu-Ni-PGE mineralization. A) Scour structure in map unit N-Th. B) Disrupted igneous foliation along the southern margin of map unit N-Tl. C) Cross-bedding of troctolitic rocks in map unit N-Tl. D) Cu-Ni sulfide mineralization in map unit N-Ts.

DAY 2, 3-D VISUALIZATION AND DRILL CORE DISPLAYS

Day two of this field trip will be spent at Duluth Metals Limited's field office and drill core logging facility in Ely, Minnesota. The day will be split up into two principal activities: 1) 3-D visualization of subsurface geological features of the Nickel Lake macrodiike and South Kawishiwi intrusion utilizing the computer program gOcad (geologic object computer aided design) and a Geowall (bring your camera because we'll all be wearing those funny looking 3D glasses), and 2) examination of selected core intervals from a number of holes drilled by Duluth Metals Limited over the last year.

3-D GEOLOGICAL MODELING AND VISUALIZATION

The science of geology uses a variety of tools to study the earth. However, the basis for every type of geologic study is fundamentally rooted in observations made of rocks in their natural habitat – “**in the field**”. Geologists that do not have an intimate appreciation of the power and fundamental nature of field geology cannot, in turn, appreciate coherent and compelling field-based scientific arguments from which all other geologic interpretations grow. This basic tenet may never be truer than for geologists engaged in mineral exploration.

The advance in computer technology over the last twenty years has revolutionized all aspects of our lives. One such advance in geology has been the development of sophisticated 3-D geological software that, **if used correctly, i.e., created and maintained by geologists who understand the rocks**, can be an outstanding tool for letting geologists interpret data (field observations, drill hole data, etc...) into the subsurface where direct observation is impossible. The first author of this guidebook (Peterson) integrates many types of geological data into the computer program gOcad®, which is perhaps the most sophisticated 3-D geological modeling software available (see websites <http://www.gocad.org/www/> and <http://www.earthdecision.com/>).

3-D models of active mines and/or advanced exploration projects benefit from continuous validation and upgrading of the underlying database, as well as the production of regional geological syntheses, integrating new geological, geophysical, geochemical, geotechnical, and geohydrological models into a single platform. The natural outgrowth of 3-D geological models is their extrapolation away from areas with large amounts of data to more remote areas with less and/or no data, and can assist in the definition of new exploration targets. For geologic teams working in a mineral exploration setting, the main advantages of using integrated 3-D geological models are to (from Fallara et al., 2006):

- 1) *Share the information*
 - a. Conveys the data and their interpretation in an immersive format
 - b. Avoids the loss of knowledge and interpretation which may be,
 - i. Distributed within various locations within the company
 - ii. Filed in disorganized ways
 - iii. Never filed and existing only in a geologists memory
- 2) *Be a catalyst in the development of geologic knowledge*
 - a. Easily integrates data in a common format
 - b. Preserves data that is easily shared, seen, and analyzed
- 3) *Hastens problem solving throughout the company*
 - a. Adaptable to team-driven resolution
 - b. Easy access to geologists and management
 - c. Allows for the shared comprehension of the data

- 4) *Accelerates the process of data integration and interpretation*
 - a. Define potential exploration targets with reduced uncertainty
 - b. Focuses work on interpretation
 - c. Optimizes data subsets
- 5) *Direct access to data and manipulation within the gOcad[®] software*
 - a. Integrated 3-D querying within the geological model
 - b. Saves time and money, rapid validation of data, and uncertainty resolution

A 3-D display of the first author's gOcad modeling of geologic data from the Duluth Complex in general, and the SKI, NLM, and BEI in particular, will be presented in Duluth Metals Limited's drill core logging facility at the start of Day 2 of this field trip. A simple screen dump out of gOcad of some of this data is presented in Figure 4.10.

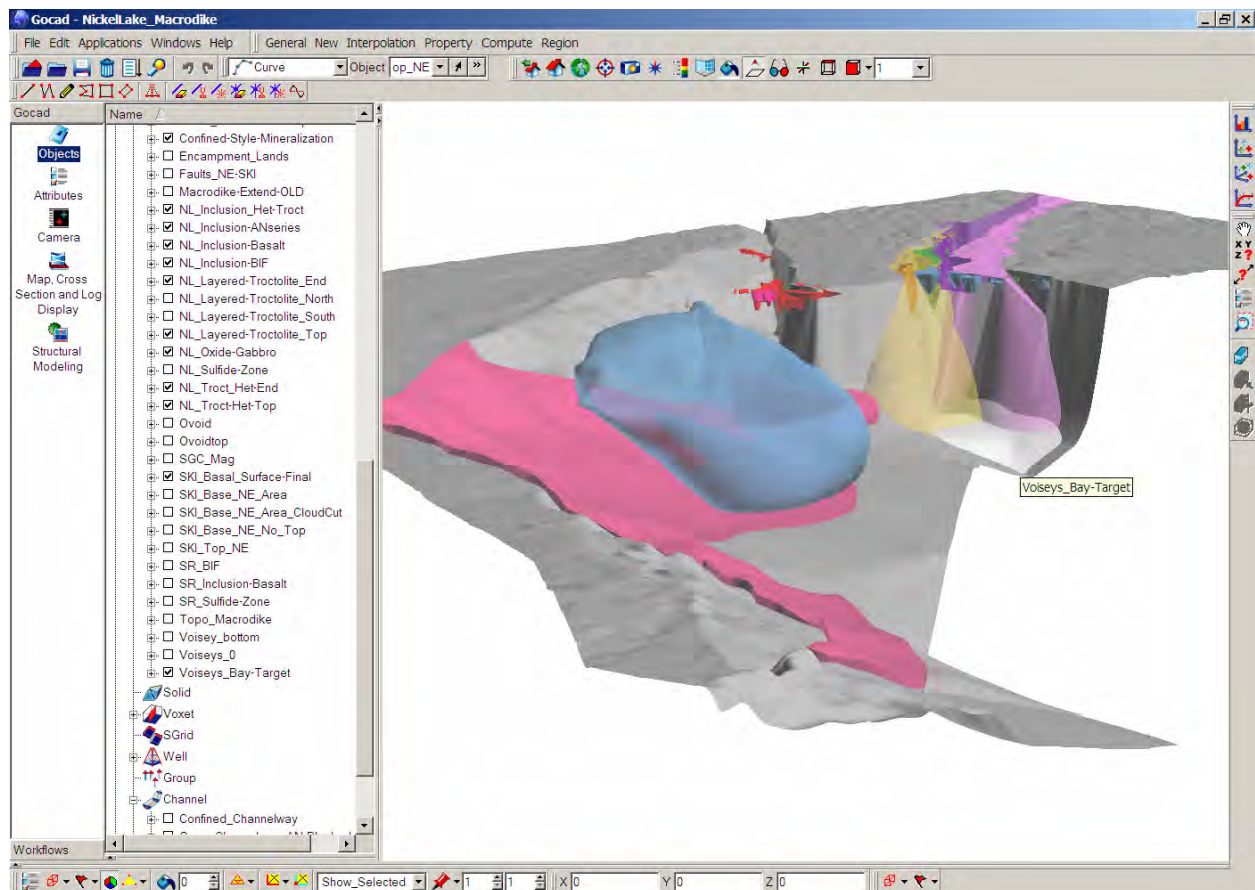


Figure 4.10. Print screen image of the SKI and NLM gOcad model (Peterson, unpublished data).

DULUTH METALS LIMITED DRILL CORE DISPLAYS

Recent drilling (Fig. 4.11) and core logging by Duluth Metals Limited east of the Maturi deposit in 2006 and 2007 has confirmed numerous stratigraphic units described by Severson (1994), especially the sulfide-bearing PEG, U3, BH, BAN, and upper GRB units.

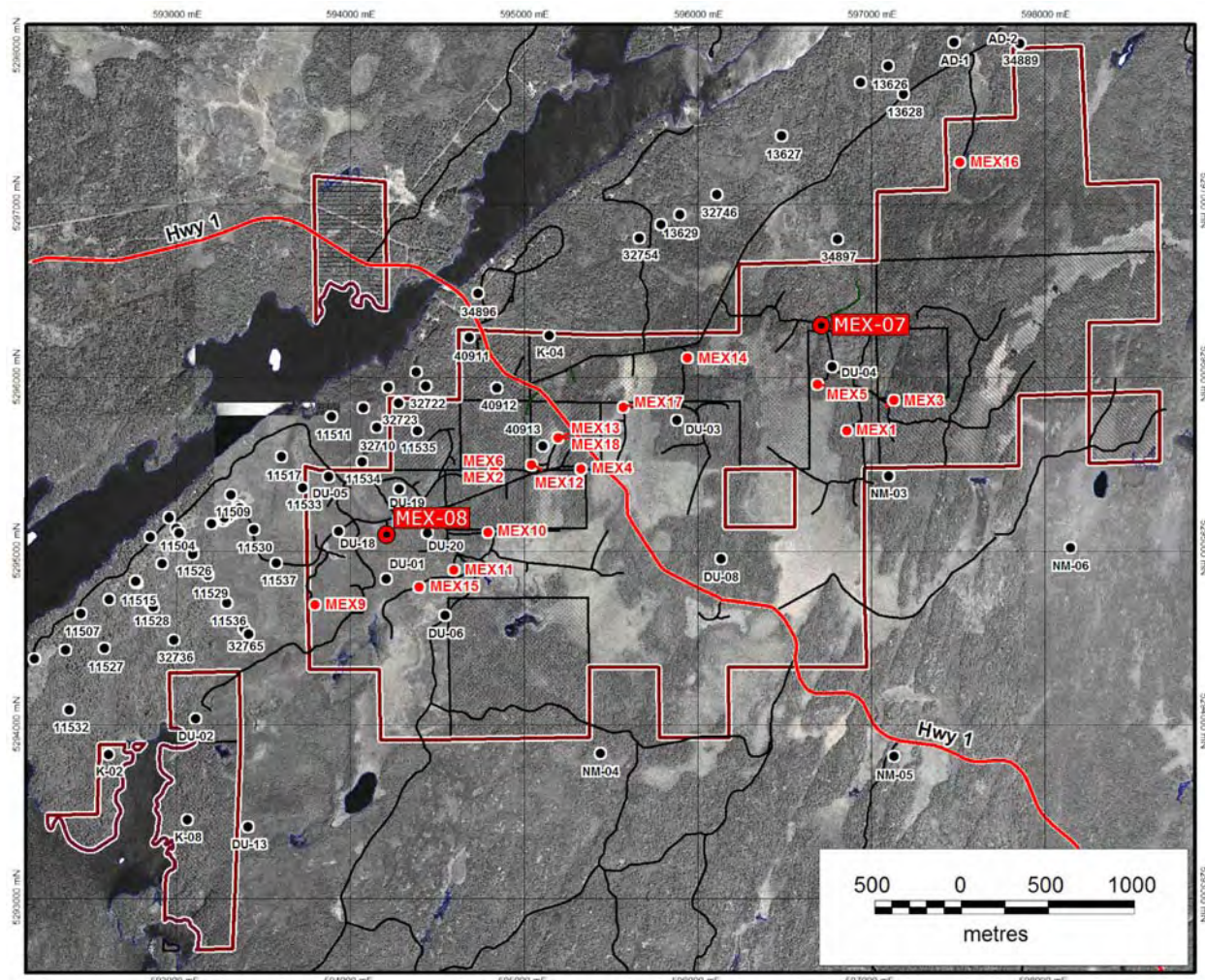


Figure 11. Property map of Duluth Metals Limited. Included are historic (black) and recent (red) drill holes.

The numerous correlative units of the SKI from historic and recent drill holes include the following:

Main AGT – Thick zone of homogenous augite troctolite to olivine gabbro.

AT&T – Thick zone of homogeneous medium-grained anorthositic troctolite with troctolites.

AT(T) – Homogeneous anorthositic troctolite grading to troctolite and lesser amounts augite troctolite.

T-AGT – Homogeneous medium-grained troctolite and augite troctolite.

Pic – Thin horizon of medium-grained picrite that is rarely developed, and thus usually uncorrelative.

Upper Gabbro – Heterogeneous, medium- to coarse-grained, oxide-rich gabbro and olivine gabbro.

AN Group – Combination of coarse-grained homogenous anorthosite, troctolitic anorthosite, anorthositic troctolite, and medium-grained homogeneous gabbro, and olivine gabbro of the Anorthositic series of the Duluth Complex.

Basalt Inclusion – Fine-grained, typically magnetic, basalt hornfels with sharp external contacts and locally preserved stretched amygdules.

U2 – Ultramafic horizon of locally serpentinized, medium-grained picrite and troctolite with sharp contacts.

PEG – Heterogeneous pegmatoidal and pegmatitic troctolite and anorthositic troctolite with minor amounts of augite troctolite, olivine gabbro, and troctolitic anorthosite. The PEG unit typically is weakly sulfide-bearing with sulfide-barren rocks above.

U3 – Ultramafic zone of locally serpentized, sulfide-bearing, medium-grained dunite, picrite, and troctolite with sharp contacts. This unit is the PGE-dominate horizon at the Birch Lake deposit.

BH – Basal heterogeneous zone that is the main sulfide-bearing unit of the SKI. Consists of fine- to medium- to coarse-grained and pegmatitic troctolite and to a lesser extent anorthositic troctolite and augite troctolite.

BAN – Basal augite troctolite and norite which is the contaminated zone between the BH and underlying GRB. Consists of fine- to medium-grained, sulfide-bearing augite troctolite and norite.

GRB – A variety of rocks from the footwall Archean Giants Range Batholith. Rock types include medium- to coarse-grained hornblendite, diorite, and porphyritic or nonporphyritic monzonite to quartz monzodiorite. Sulfide mineralization is usually restricted to the upper 20 feet of the GRB, but can exist in small concentrations over a hundred feet into the granitic footwall.

Simplified drill hole logs portrayed as stratigraphic columns from two holes recently drilled in the eastern (MEX-07) and western (MEX-08) portions of the Maturi Extension deposit are presented in Figure 4.12. In addition, reported Cu-Ni-PGE grades from Duluth Metals Limited for these holes are given in Table 4.2. The stratigraphy observed in MEX-07 is typical of the eastern exploration area and the deep holes in this region (>3,000 feet) that are situated below a very large pillar/xenolith of Anorthosite series rocks (see Fig. 4.3). Channelized magma flow out of the NLM and under this Anorthosite block (think of eddies due to overlying friction as a fisherman would) has been interpreted by Peterson (2001b, c) as one of the driving mechanisms for enhanced turbulent flow of sulfide-bearing magmas, thus forming one of the elements of the Ni-Cu-PGE ore deposit model, “*an appropriate physical environment is required so that the sulfide liquid mixes with enough magma to become adequately enriched in chalcophile metals*”. The stratigraphy observed in MEX-08 correlates with many of the drill holes in the western exploration area (as well as Franconia Minerals’, Maturi deposit) immediately to the west of the Anorthosite series pillar/xenolith. Mineralization in both of these holes consist of chalcopyrite-dominated disseminated sulfides (typically 2-5% total sulfide) restricted to a confined zone directly above the footwall and contain moderate to high Cu, Ni, and PGE concentrations.

If one would **very simply** (unlike the data depicted in Figure 4.6, which used the strict rules of Kerr, 2001) recalculate these reported grades (Table 4.2) to 100% sulfide, i.e. as a specific gravity concentrator would do as a first step at an active mine, then one would have to multiply the assay data by a factor between 20 to 50 (5% to 2% total sulfide minerals in the assayed drill hole intervals) to imagine the true reality that is the enrichment of Cu-Ni-PGE in the sulfide minerals of the Maturi Extension deposit. Such recalculations lead to grades (at 100% sulfide) of ~33% Cu, ~9% Ni, and ~11 – >82 g/t PGE+Au for the sulfide fraction of the Maturi Extension deposit, which point to these preliminary conclusions:

- 1) The extensive sulfide mineralization at the Maturi Extension deposit (as well as the Maturi, and Birch Lake deposits, i.e., the *Confined* mineralization of Peterson 2001, 2002a-d) is similar at a recalculated 100% sulfide grade to the metal budget of the fractionated residuum of a MSS;
- 2) One can imagine such mineralization (the *Confined*-style of Peterson 2002a) initially forming from incorporation of the Cu-PGE residuum of a fractionating Ni-rich MSS at the junction of the NLM and SKI into magmas streaming out of the NLM and into the SKI.
- 3) A seemingly robust geologic model of a dynamic magmatic system that incorporates the BEI, NLM, and the SKI exists herein that vector towards an exploration target for a Ni-rich massive sulfide body (the “*missing Ni-rich MSS*”) at the junction of the NLM and the SKI.

Table 2. Reported mineralized intervals for holes MEX-07 and MEX-08 drilled by Duluth Metals Limited on the Maturi Extension property in 2006.

MEX-08	From (Ft)	To (Ft)	Total (Ft)	Cu %	Ni %	Pd (g/t)	Pt (g/t)	Au (g/t)
Overall	2032	2236.5	204.5	0.69	0.22	0.313	0.121	0.108
including	2052	2142	90	0.81	0.24	0.365	0.147	0.056
including	2202	2236.5	34.5	0.89	0.28	0.352	0.12	0.096
MEX-07	From (Ft)	To (Ft)	Total (Ft)	Cu %	Ni %	Pd (g/t)	Pt (g/t)	Au (g/t)
Overall	2543	2798	255	0.429	0.129	0.370	0.183	0.077
including	2608	2673	65	0.824	0.275	0.906	0.444	0.173
including	2628	2663	35	0.898	0.310	1.004	0.472	0.177

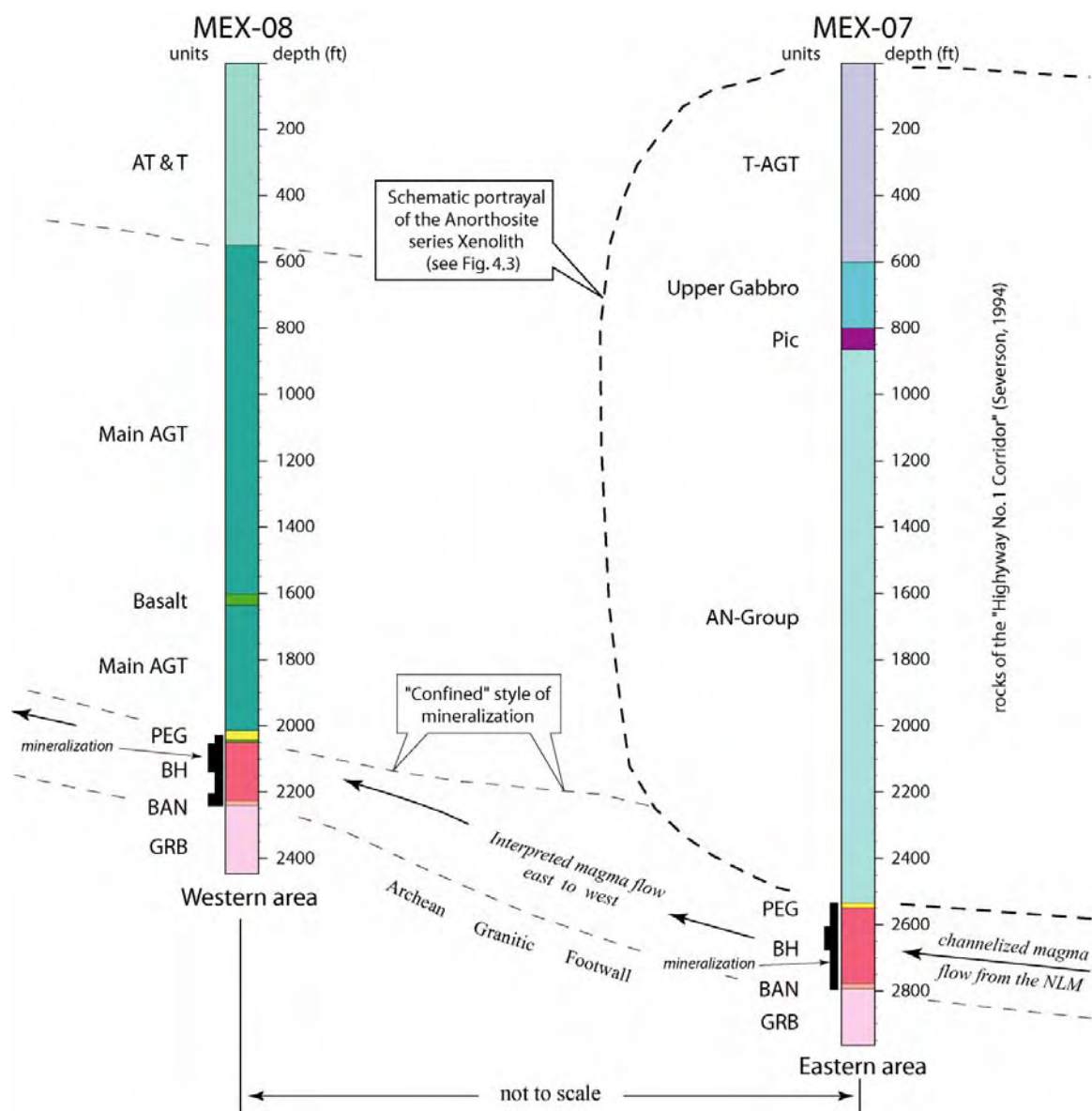


Figure 4.12. Stratigraphic sections of holes MEX-07 and MEX-08 drilled by Duluth Metals Limited on the Maturi Extension property in 2006.

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53rd Annual Institute on Lake Superior Geology

FIELD TRIP 5

Geologic Highlights of New Mapping in the Upper Southwestern to Northeastern Sequence of the North Shore Volcanic Group and Beaver Bay Complex

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View northeast across Pork Bay on the North Shore

INTRODUCTION

The Minnesota Geological Survey is continuing to map the Keweenawan bedrock geology of 1:24,000-scale quadrangles near the North Shore of Lake Superior, in northeastern Minnesota. The first phase of mapping from 1985-1993, partially supported by the USGS COGEOMAP program, focused on the Beaver Bay Complex in the central part of the region (Fig. 5-1; Miller, 1988; Miller and others, 1989, 1993, 1994; Boerboom and Miller, 1994). More recent mapping, supported by the USGS STATEMAP program, has focused on quadrangles that intersect the shoreline, where the bedrock is predominantly volcanic rocks. This mapping phase started at the outskirts of Duluth in 2001, and to date has proceeded to the Lutsen quadrangle, having skipped past the COGEOMAP quadrangles. Ten quadrangle-scale geologic maps have been published to date from this work (Boerboom and others, 2002a, 2002b, 2003a, 2003b; 2006; 2007; Boerboom and Green, 2004; 2005, 2006; Miller and others, 2006), and at least two more are planned (Fig. 5-1).

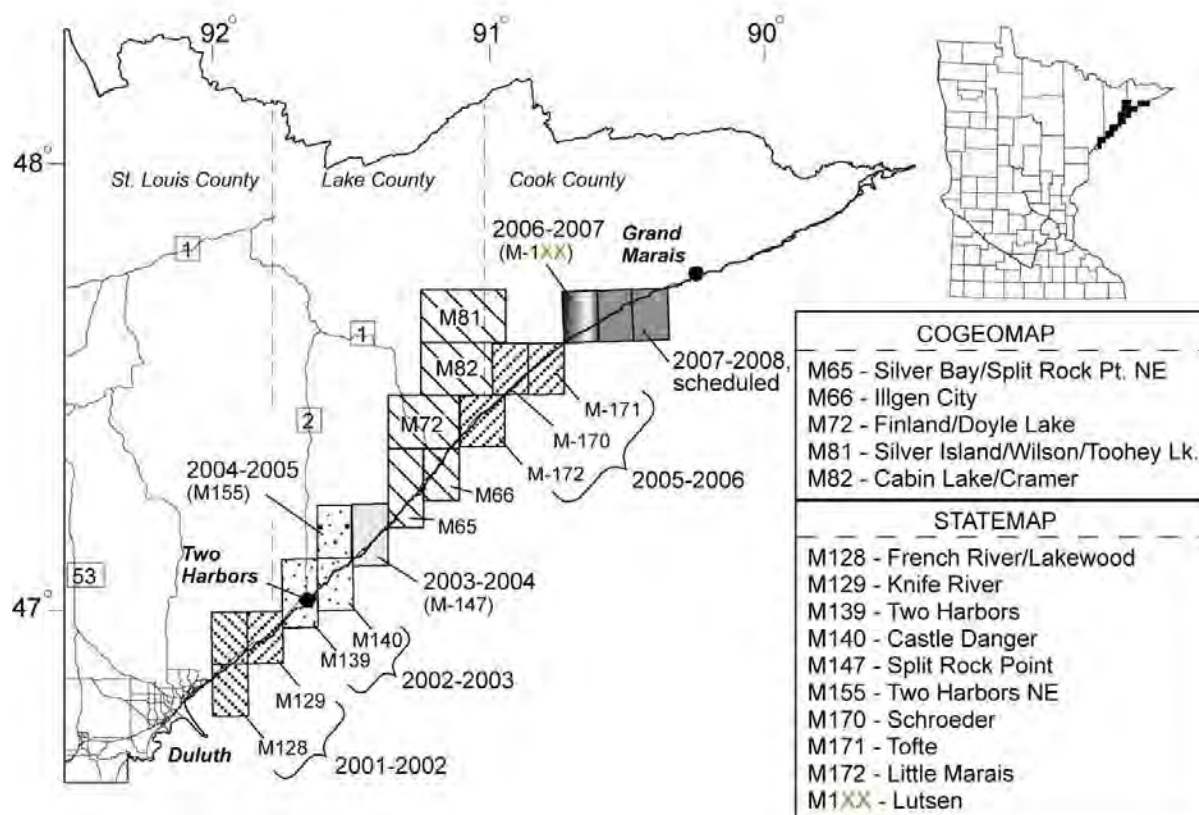


Figure 5-1. North Shore 7.5' quadrangles mapped by the MGS through the USGS-sponsored COGEOMAP and STATEMAP programs. M-numbers refer to MGS Miscellaneous Map Series.

A trip similar to this, which covered the Upper Southwest sequence of the North Shore Volcanic Group (NSVG), was conducted for the 2004 Institute on Lake Superior Geology meeting held in Duluth. The trip for this year's meeting will cross the transition from the uppermost Southwest limb to the uppermost Northeast limb, as well as the basal Schroeder-Lutsen sequence of the NSVG (Fig. 5-2).

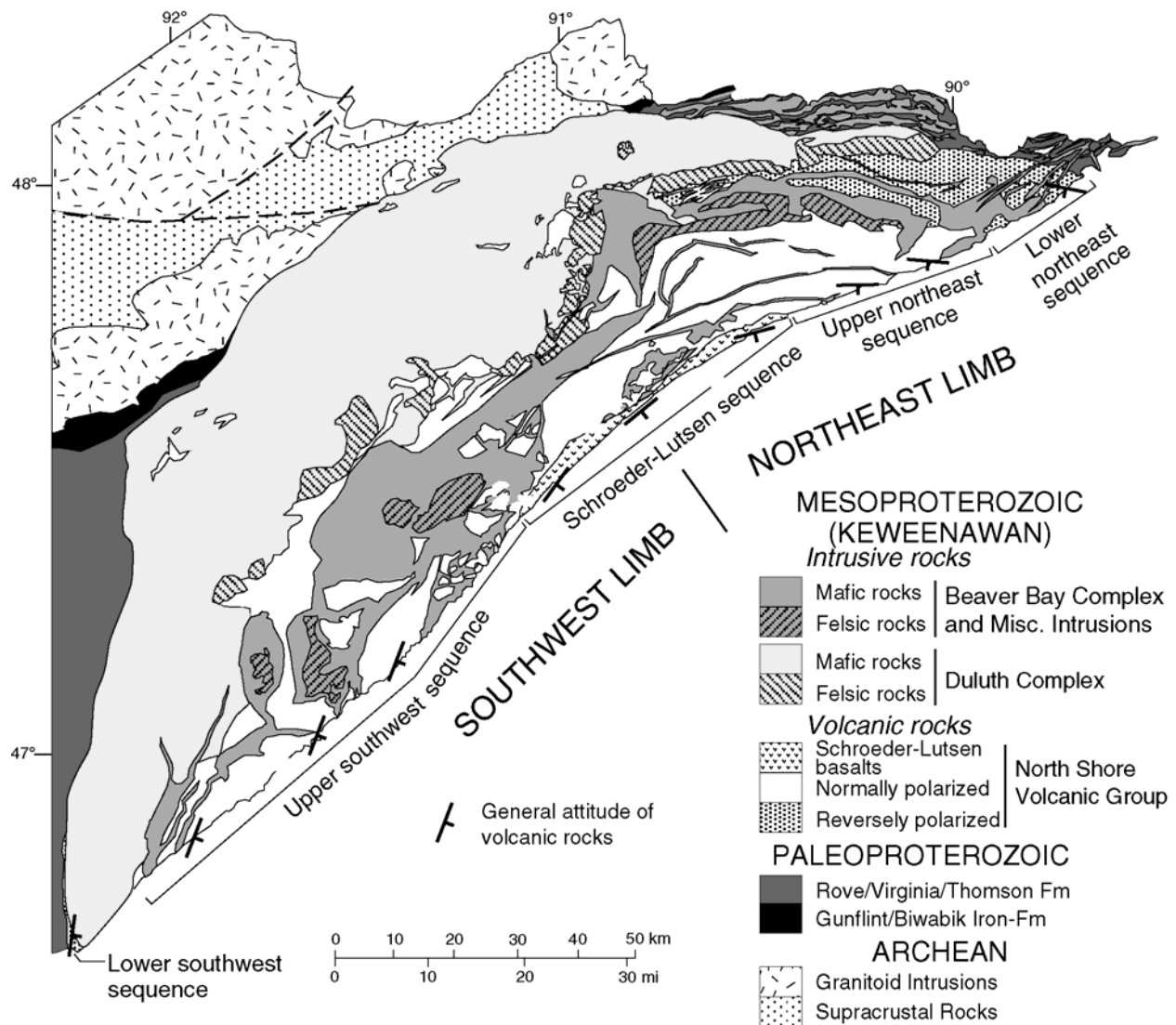


Figure 5-2. Simplified geology of northeastern Minnesota showing the major sequences of the North Shore Volcanic Group (taken from Fig. 5.2 of Miller and others, 2002).

VOLCANIC AND SEDIMENTARY ROCKS OF THE NORTH SHORE VOLCANIC GROUP

This overview is a slightly modified excerpt from the MGS Report of Investigations 58, Chapter 5: "Volcanic and sedimentary rocks of the Keweenaw Supergroup in Northeastern Minnesota" by John Green (Miller and others, 2002).

Magmatic activity related to the 1.1 Ga Midcontinent Rift produced a more than 10 kilometer thick edifice of lava flows and subvolcanic intrusions that are exposed along Minnesota's north shore of Lake Superior. The lava flows and minor sedimentary rocks are referred to as the North Shore Volcanic

Group (Goldich and others, 1961) and the intrusive rocks are variably assigned to the Duluth Complex, the Beaver Bay Complex, and miscellaneous intrusions of the Midcontinent Rift Intrusive Supersuite. Although previous publications have subdivided the North Shore Volcanic Group into informal volcanic suites and distinctive flows (Green, 1972, 1982), the 1:200,000-scale map (M-119) produced by Miller and others (2001) and accompanying report (Miller et al., 2002) represents the first time that the NSVG has been subdivided into coherent lithostratigraphic units on a geologic map. A brief description of the North Shore Volcanic Group and associated interflow sedimentary rocks of the Keweenaw Supergroup is given here.

Rock classification, recognition, and textures

As a coherent tholeiitic compositional suite, the volcanic rocks of the North Shore Volcanic Group can be described using only a few rock names (Fig. 5-3). The most primitive rocks are olivine tholeiites, which form an iron-enrichment trend with further evolution. They display ophitic textures and pahoehoe surfaces nearly everywhere. Most olivine tholeiites are aphyric, but those that are porphyritic contain dominantly plagioclase phenocrysts, less commonly olivine. Transitional basalts contain somewhat higher alkalis and other incompatible elements than the olivine tholeiites, but generally not enough to classify them as alkalic. Their texture is typically intergranular and fine- to medium-grained. Porphyritic varieties generally contain small phenocrysts of plagioclase, olivine, clinopyroxene, and magnetite. The reversed-polarity Hovland lavas, however, are characterized by transitional basalts (grading to basaltic andesites) that contain abundant, large, tabular, plagioclase phenocrysts. Transitional basalt flow surfaces are generally smooth (pahoehoe), although a few show breccia tops.

The basaltic andesites and andesites (greater than 52 percent SiO_2) are tholeiitic, rather than calc-alkaline; they show iron enrichment and contain only anhydrous ferromagnesian minerals. These rocks are typically fine-grained and intergranular to felty or pilotaxitic, and many contain small phenocrysts of plagioclase, olivine, clinopyroxene, and magnetite. The andesites generally weather to a red-brown color, and have flow-brecciated (aa) tops but not bases. Many flows that contain 50 to 55 percent silica show millimeter-scale oxidation lamination (Green, 1989) parallel to the base. A few highly iron-enriched flows, separable only by chemical analysis, can be called ferroandesites.

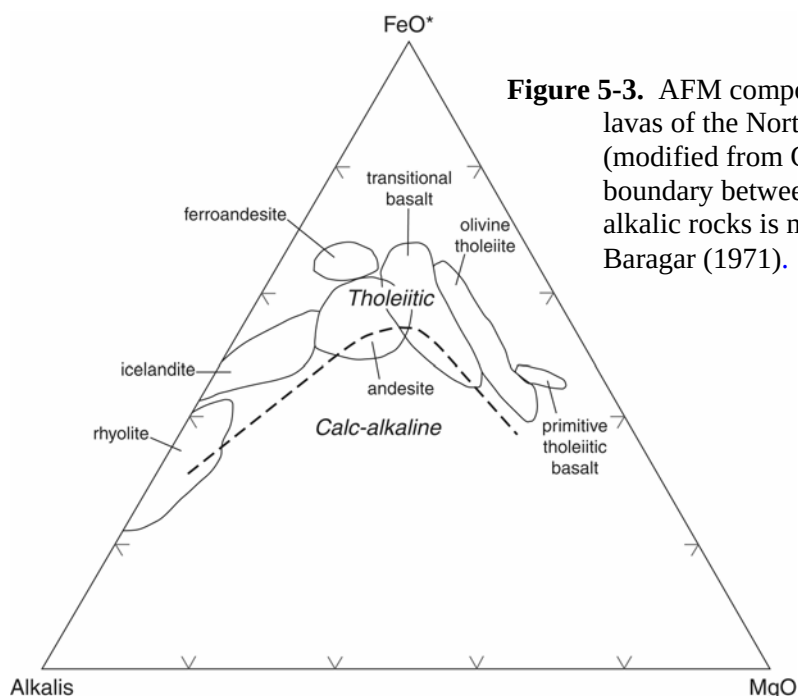


Figure 5-3. AFM compositional diagram for the lavas of the North Shore Volcanic Group (modified from Green, 1982). The boundary between tholeiitic and calc-alkalic rocks is modified from Irvine and Baragar (1971). $\text{FeO}^* = \text{FeO} + 0.9\text{Fe}_2\text{O}_3$.

Carmichael (1964) first used the name "icelandite" for rocks intermediate in character between andesites and rhyolites in the Tertiary lavas of eastern Iceland. They might be considered the tholeiitic equivalent of calc-alkaline dacite in orogenic suites. Other examples of these rocks have been described from the Galapagos (McBirney and Williams, 1969) and the Miocene of Nevada–Oregon (Wallace and others, 1980). Very large flows of similar composition in the Etendeka volcanics of Namibia have been referred to as quartz latites by Milner and others (1992). Icelandites in the North Shore Volcanic Group (Green and Fitz, 1993) are characterized chemically by SiO₂ contents ranging from 60 to 68 percent, high FeO* (averaging 7 percent), K₂O + Na₂O values between 6.5 and 9 percent, a potassium/sodium atomic ratio of about 0.9, and an Mg number [Mg/(Mg + Fe) atomic] averaging 0.14. Petrographically, North Shore Volcanic Group icelandites grade continuously from the andesites to somewhat paler colors (brown or tan), but have a similar phenocryst assemblage. Quartz and alkali feldspar are common in the groundmass but never occur as phenocrysts. Flowtop features (crusty to coarsely brecciated) indicate that the icelandites were erupted as lavas.

The rhyolites have higher silica and total alkali contents, and lower FeO than the icelandites. They are generally light gray, pink, or red. Several are very thick, extensive, and voluminous (up to several hundred cubic kilometers; Green and Fitz, 1993). Although most rhyolites are porphyritic (phenocrysts of quartz, alkali feldspar ± plagioclase, and altered fayalite ± ferroaugite), some lack quartz and alkali feldspar phenocrysts, and rare flows are aphyric. Groundmass textures are fine-grained holocrystalline, typically with a meshwork of platy quartz paramorphs after primary tridymite, which may show a flow structure. A "snowflake" texture is common, in which poikilitic quartz patches (coalesced ex-tridymite grains) enclose small, dusty alkali-feldspar grains. In quartz-phyric flows, these poikilitic quartz patches are in optical continuity with adjacent quartz phenocrysts (Green, 1990). Outcrop-scale flow structure, including folding, is common near flow tops and bases. In some flows, distinct fiamme, deformed to varying degrees, are recognizable. These imply an explosive eruption that produced a pyroclastic flow, which welded and remobilized to produce a rheoignimbrite.

Structure and Lithostratigraphy

As stated above, the Midcontinent Rift volcanic rocks and interbedded redbeds in northeastern Minnesota comprise the North Shore Volcanic Group (Goldich and others, 1961; Green, 1972, 1977, 1982; Basaltic Volcanism Study Project, 1981). In general, these rocks form an arcuate stack that is slightly tilted toward the southeast and forms the roof rocks into and under which the Duluth Complex and associated hypabyssal intrusions were emplaced. At the southwest end of the North Shore Volcanic Group near Duluth, the volcanic rocks strike north with a 10° to 20° easterly dip; at the northeast end at Grand Portage, the flows strike east–west with a 10° southerly dip (Fig. 5-2). Thus, traveling northeast along the Lake Superior shore northeast from Duluth, and southwest from Grand Portage, one encounters successively higher flows in the volcanic stratigraphy until the Tofte–Lutsen area in southern Cook County, where the highest exposed flows crop out. Exposure is generally excellent along the eroding lakeshore, and along the lower, high-gradient stretches of tributary streams, providing good control on the stratigraphy. A stack of volcanic rocks approximately 9.7-kilometers-thick has been measured in the "southwest limb" (Table 5-1), and another stack of volcanic rocks about 7.2-kilometers-thick has been measured in the "northeast limb" (Table 5-2, Miller and others, 2002). This implies nearly continuous subsidence during the rifting process. The difference in stratigraphic thickness between the two limbs reflects major complications in the central area, which appears not to have subsided at the same rate as in the limbs, and into which many of the subvolcanic intrusions were emplaced. Except for the capping Schroeder–Lutsen basalt sequence, no stratigraphic unit can be traced from one limb to the other; each limb has its own stratigraphic column.

To aid in the correlation of intrusive and extrusive rock units throughout the Midcontinent Rift system, their paleomagnetic polarity has been used. Nearly all of the igneous and sedimentary rocks associated with the Midcontinent Rift were formed either during an earlier, reversed-polarity interval or a succeeding normal-polarity interval. Thus, in each limb of the North Shore Volcanic Group, the lower stratigraphic

units show reversed polarity, and the upper sequences show normal polarity. This polarity reversal forms the basis for distinguishing upper and lower sequences in the northeast and southwest limbs. U-Pb zircon dates demonstrate that the reversed-polarity magmatism occurred mainly in the time interval from 1108 to 1107 Ma; whereas, around the Lake Superior basin, normal-polarity magmatism occurred mainly in the interval from 1099 to 1094 Ma (for example Davis and Paces, 1990; Paces and Miller, 1993; Davis and Green, 1997). These two pulses were separated by a magmatically inactive time (at least in the upper crust), which appears to be expressed as a slight unconformity in the volcanic sequence on the north shore. However, because intrusions subsequently penetrated along this horizon in the North Shore Volcanic Group, this unconformity has not been recognized in outcrop. The Duluth Complex separates the upper and lower sequences of the southwestern limb (Fig. 5-2, Table 5-1).

One other significant gap in the stratigraphic continuity of the North Shore Volcanic Group occurs near the stratigraphic top, where the Schroeder–Lutsen sequence overlies the upper units of the northeast and southwest sequences (Fig. 5-2, Tables 5-1 and 5-2). In the northeast limb, southwest of Grand Marais (Fig. 5-3), the Schroeder–Lutsen sequence unconformably overlies the Good Harbor Bay lavas, which are now recognized to include the Terrace Point basalt, as well as the Cut Face Creek Sandstone and the Good Harbor Bay andesites. The Terrace Point basalt contains a granite xenolith dated at 1096.7 ± 0.9 Ma and thus must be younger than that. In the southwest limb, the Schroeder–Lutsen sequence overlies andesites and icelandites of the Onion River lavas (Boerboom and Green, 2006; Boerboom and others, 2006) as well as a sandstone and conglomerate unit (the Little Marais conglomerate) that in turn rests in sharp angular unconformity atop structurally disturbed flows of the Belmore Bay lavas near Little Marais. The Onion River lavas form a coherent package of dominantly intermediate-composition lavas that have only recently been mapped in detail, and their stratigraphic thickness is not taken into account in Table 5.1. They are located in approximately the same stratigraphic position as the Belmore Bay lavas, and may be continuous with them, but their relationship has been obscured by emplacement of the Beaver Bay Complex. Furthermore, the gently dipping Schroeder basalts have not been penetrated by the abundant hypabyssal intrusions of the Beaver Bay Complex that complicate the underlying volcanic sequence in this mid-shore area (Green, 1992; Miller and others, 1993). Attempts to date the Schroeder–Lutsen sequence have been unsuccessful to this point.

The five major lithostratigraphic sequences comprising the two limbs of the North Shore Volcanic Group are further subdivided into informal lithostratigraphic units (Table 5-1 and 5-2). Some of these units are individual flows of distinctive lithology and/or substantial thickness and lateral extent. Most are suites of lava flows that have distinct lithologic characteristics or that are separated by intrusions. Some lava formations contain distinct flows or sedimentary rock units within an otherwise homogeneous package of lavas. Such units are given informal names (such as the Silver Beaver rhyolite within the Baptism River lavas, Manitou transitional basalt within the Schroeder basalts, and Indian Camp sandstone within the Lutsen basalts). See Chapter 1 of RI58 (Miller and others, 2002) for more details on stratigraphic nomenclature of Keweenaw rock in northeastern Minnesota.

Table 5-1. Generalized stratigraphy of the southwest limb of the North Shore Volcanic Group showing U/Pb ages (Davis and Green, 1997; Green and others, 2001). Positions of intrusions denote approximate stratigraphic level affected and not age of emplacement.

Thickness(m)	Lithostratigraphic units		Lithologic character	U/Pb age
9735	Total section			
	150	Carlton Quarry lavas (<i>fault bounded</i>)	basalt, andesite, and rhyolite flows	1094.3±2.0
945	<u>Schroeder–Lutsen sequence (normal polarity)</u>			
	900	Schroeder basalts	ophitic olivine tholeiite basalt flows; includes Manitou transitional basalt and Pork Bay breccia	
	<45	Little Marais conglomerate	polymict volcanic conglomerate and sandstone	
	<i>angular unconformity</i>			
8275	<u>Upper southwest sequence (normal polarity)</u>			
	565	Bell Harbor lavas	mostly quartz tholeiite basalt and basaltic andesite flows	
	100	Palisade Head rhyolite	gray-pink, porphyritic rhyolite flow	1096.6±1.7
	<i>Beaver Bay Complex</i>			~1096
	700	Baptism River lavas	mixed lavas, mostly basalt; includes 165-meter-thick Silver Beaver rhyolite	
	20	Silver Bay porphyritic basalt	ophitic basalt flow with abundant large plagioclase phenocrysts	
	730	Gooseberry River basalts	mixed basalt flows, mostly ophitic	
	<i>Lafayette Bluff, Silver Creek diabase intrusions</i>			
	315	Two Harbors basalts	mixed aphyric basalt flows; quartz tholeiite flows at base	
	550	Larsmont basalts	ophitic olivine tholeiite flows	
	<i>Stony Point–Knife Island diabase sheet</i>			
	1500	Sucker River basalts	mixed basalt flows, mostly ophitic	
	1350	Lakewood lavas	mostly basalt flows; rhyolite, icelandite, and ferroandesite at base	
	<i>Lester River diabase sill</i>			
	1285	Lakeside lavas	mixed basalt, andesite, icelandite, and rhyolite flows	1098.4±1.9
	<i>Endion diabase sill</i>			
	1160	Leif Erickson Park lavas	mixed basalts, andesites	
	<i>Duluth Complex</i>			~1099
370	<u>Lower southwest sequence (reversed polarity)</u>			
	370	Ely's Peak basalts	porphyritic, diabasic, and ophitic basalts; pillowed and pyx-phyric basal flow	
	>8	Nopeming Sandstone	white to tan quartzite and conglomerate	
	<i>angular unconformity</i>			
		Thomson Formation (Paleoproterozoic)		

Table 5-2. Generalized stratigraphy of the northeast limb of the North Shore Volcanic Group showing U/Pb ages (Davis and Green, 1997; Green and others, 2001). Positions of intrusions denote approximate stratigraphic level affected and not age of emplacement.

Thickness(m) ages	Lithostratigraphic units		Lithologic character	U/Pb
7359	Total section			
325	<u>Schroeder–Lutsen sequence (normal polarity)</u>			
		Lutsen basalts	olivine tholeiite basalts, mostly ophitic; includes Indian Camp sandstone	
	<i>angular unconformity</i>			
3998	<u>Upper northeast sequence (normal polarity)</u>			
	130	Terrace Point basalt (within upper Good Harbor Bay andesites)	ophitic, olivine tholeiite basalt	younger than 1096.7±0.8
	100	Cut Face Creek sandstone	red, laminated, ripple-marked sandstone	
	131	Good Harbor Bay andesites	brown, porphyritic basaltic andesites	
	<i>Beaver Bay Complex – Beaver River diabase and Leveaux ferrodiorite</i>			
	122	Breakwater basalt	brown, columnar-jointed basalt flow	
	348	Grand Marais felsites	pink to gray porphyritic rhyolite and felsite	
	335	Croftville basalts	intergranular basalt and andesite flows, thick interflow sandstone	
	250	Devil’s Track rhyolite	aphyric, intergranular rhyolite flow	
	70	Maple Hill rhyolite	porphyritic rhyolite flow	
	274	Red Cliff basalts	ophitic olivine tholeiite flows, some plagioclase-phyric	
	366	Kimball Creek rhyolite	porphyritic rheognimbrite	
	539	Marr Island lavas	mixed basalt, tholeiitic andesite, and icelandite flows	
	198	Naniboujou basalts	intergranular basalt flows	
	235	Devil’s Kettle rhyolite	porphyritic lava flow	1097.7±1.7
	900	Brule River lavas	interbedded basalt and rhyolite flows	1100.2±1.9
	<i>Brule Lake-Hovland gabbro</i>			
3036	<u>Lower northeast sequence (reversed polarity)</u>			
	1932	Hovland lavas	mostly plagioclase-phyric basalt flows, some rhyolite and andesite	1107.7±1.9
	67	Red Rock rhyolite	red/tan, porphyritic rhyolite	1107.9±1.8
	92	Deronda Bay andesite	tan/brown, porphyritic andesite	
	945	Grand Portage lavas	basalt lava flows, pillowed at base	
BASE				
60	60	Puckwunge Sandstone	tan/white, cross-bedded quartz sandstone	
	<i>slight angular unconformity</i>			
		Rove Formation (Paleoproterozoic)		

Physical Volcanology

The volcanic rocks of the Midcontinent Rift, including the North Shore Volcanic Group, represent one of the world's oldest and best-preserved examples of plateau lavas. However, they contain a greater thickness of flows and, in the North Shore Volcanic Group, a higher proportion of evolved compositions **than typical plateau lavas**. They are similar physically and chemically to the Tertiary lavas that make up eastern Iceland (Sigvaldason, 1974; Walker, 1974; Green, 1977; Wood, 1978), and they formed similarly over a plume at another major rift. The rocks also resemble the late Tertiary and Quaternary volcanic rocks of the southern Snake River Plain, Idaho and southeastern Oregon, because of their interbedded basalts and large rhyolites (for example Bonnicksen and Kauffman, 1987; Link and Hackett, 1988; Reidel and Hooper, 1989; Manley, 1996).

The basalts range in character from typical flood flows as voluminous as tens of cubic kilometers to more modest, "plains-type" flows (Greeley, 1982) and thin flow units less than a meter thick. At Duluth in the southwest limb, and Grand Portage in the northeast limb, the lowest flows in the volcanic sequences are pillowed, and thus inferred to have erupted subaqueously; however, nearly all of the other flows were erupted subaerially. The flows show different physical characteristics, closely tied to their chemical compositions and viscosities (Green, 1977, 1989; Green and Fitz, 1993). Olivine tholeiites, which dominate the North Shore Volcanic Group, all have pahoehoe surfaces, with or without ropy structures. Other physical characteristics of the various rock types were previously discussed in the "Rock classification, recognition, and textures" section of this chapter.

All of the flows ranging in composition from basalts to icelandites were erupted as lavas. The rhyolites are notable in their abundance relative to other plateau-lava sequences, their size (up to several hundred cubic kilometers), and extent (Green and Fitz, 1993). Several rhyolites show textural evidence of rheomorphic flow after eruption as ash-flow tuffs, though some were lavas. One of the largest, the Devil Track rhyolite in Cook County, which is as thick as 250 meters and can be traced for 40 kilometers along strike, has ambiguous features that make its mode of eruption difficult to discern; it may be a lava flow. Nearly all the icelandites and rhyolites show evidence of an unusually high temperature of eruption, such as magmatically crystallized groundmass tridymite. The evidently low viscosity of these large rhyolites is attributed to their high temperature, high iron and fluorine contents, and low oxidation state (Green and Fitz, 1993).

Geochemistry and Chemostratigraphy

The North Shore Volcanic Group constitutes a subalkalic, tholeiitic suite that ranges continuously from rather primitive olivine tholeiite to rhyolite, and shows a strong iron-enrichment trend (Fig. 5-3; also Basaltic Volcanism Study Project, 1981; Green, 1982; Brannon, 1984). However, relative abundances are strongly bimodal; basalts are greatly predominant, but rhyolites make up 10 to 25 percent of the section. The basalts show trace element and isotopic evidence of derivation mostly from a mantle plume (Nicholson and others, 1997), whereas most of the rhyolites include major contributions from partial melting of the Archean basement (Vervoort and Green, 1997). The most common basalt type, ophitic olivine tholeiite, is generally aluminum-rich (16 to 18 percent Al_2O_3). The most primitive flows have Mg numbers of about 0.65 to 0.68.

The basal few flows in both limbs of the North Shore Volcanic Group have a unique geochemical and petrographic character. Typically they contain augite phenocrysts, are aluminum-poor, and are rich in both compatible (chromium and nickel) and incompatible elements (titanium, phosphorus, and lanthanum) with steep chondrite-normalized La/Yb ratios. This suggests derivation by relatively small-fraction melting of the initial plume head (Nicholson and others, 1991, 1997; Green, 1995).

In general, there is little stratigraphic regularity of compositional change within the North Shore Volcanic Group, with the following exceptions. In the middle of the upper southwest sequence, there is a marked upward progression toward more primitive compositions through a 3.4-kilometer section from rhyolite east of the Lester River into a thick group of primitive olivine tholeiites in the Knife River–Two Harbors area (Brannon, 1984). This includes the Lakewood lavas, the Sucker River basalts, and the Laramont basalts. In contrast, in the lower northeast sequence, the approximately 1-kilometer-thick basal

Grand Portage lavas progress upsection from basalt to increasingly evolved compositions, ending with Red Rock rhyolite (Green, 1995). As mentioned above, the Schroeder–Lutsen sequence, the youngest in the North Shore Volcanic Group, is composed almost entirely of olivine tholeiites.

All of the North Shore Volcanic Group has been affected to some degree by hydrothermal/burial metamorphism. The more permeable (fractured, vesicular) tops and bases of the flows have undergone considerable mineralogical change (deposition of amygdale minerals, alteration of primary minerals), but in many cases the massive flow interiors are remarkably little-altered. Where alteration has approached equilibrium, mineral assemblages range from lower greenschist facies at the base of the North Shore Volcanic Group to zeolite facies at the top (Schmidt, 1993; Schmidt and Robinson, 1997).

Interflow Sedimentary Rocks

Clastic redbed strata occur at many horizons within the North Shore Volcanic Group (Jirsa, 1984). They are lenticular and range in thickness from a few centimeters to about 100 meters. As these rocks are relatively soft and erodible compared to the adjacent volcanic flows, they are mostly covered and are exposed only along actively eroding sites such as streambeds and the lakeshore. However several interflow sandstones have recently been recognized from water well cutting samples (Boerboom and others, 2007). Overall, the sedimentary rocks are predominantly red to brown, well sorted sandstone, with minor conglomerate, siltstone, and shale. Conglomerate beds are most abundant in the midshore area from Little Marais to Lutsen, either on or in close proximity to a prominent gravity low that is at a right angle to the North Shore. This gravity low is thought to reflect an uplifted crustal block (White, 1966), and intrusive rocks along this trend are known to contain large xenoliths of Archean crust (Boerboom, 1994). Although the relative timing and effects of uplift on the adjacent volcanic rocks are not fully understood, the association of interflow conglomeratic rocks with this structure implies that uplift may have been active during volcanism.

Compositionally, these redbeds are mainly immature lithic arkose and feldspathic lithic arenite (see Fig. 5.3 in Miller and others, 2002). The angular to subrounded clasts are mainly plagioclase, mafic to felsic volcanic rock fragments, clinopyroxene, and Fe-Ti oxides; devitrified or replaced volcanic ash particles and shards are present in a few beds. Quartz is uncommon to absent. The framework grains have been variably cemented with hematite, calcite, prehnite, and a variety of zeolites, depending on the local hydrothermal/burial/contact metamorphic conditions. In some places hydrothermal minerals have replaced many or most of the clasts.

A few of these redbed units have thicknesses in excess of 25 meters. These include a cross-bedded sandstone in Leif Erickson Park in Duluth (35 meters), which disconformably overlies an eroded basalt flow; the Little Marais conglomerate (and sandstone) exposed in the Manitou River area near Little Marais (as thick as 45 meters); the Indian Camp sandstone (68 meters) northeast of Lutsen; the Cut Face Creek sandstone southwest of Grand Marais (100 meters); and other newly recognized units to the north of and lower in the stratigraphy than the Cut Face Creek sandstone (See Boerboom – abstract volume for this meeting). The Cut Face Creek sandstone can be traced in outcrop for at least 4 kilometers along strike, and topographic and aeromagnetic data indicate the newly-recognized sandstones to the north extend for at least 10-15 kilometers along strike. Of these, the Little Marais conglomerate occurs at the base of the Schroeder–Lutsen basalt sequence. The Cut Face Creek sandstone underlies the Terrace Point basalt, which is now recognized to occur in the upper part of the Good Harbor Bay sequence (Boerboom and Green, 2007), and the sandstones intersected by well drilling to the north occur between basaltic lava flows tentatively grouped with the Croftville basalts. The sandstone in these units is typically planar- or cross-bedded, and some beds are ripple-marked or mud-cracked. The rocks are inferred to be dominantly fluvial, deposited by moderate-gradient, east- to southwest-flowing streams from sources nearly entirely within the subsiding Midcontinent rift basin (Jirsa, 1984).

Many flow-top breccias of andesite and basalt with aa structure contain laminated red sandstone as a matrix because sand filtered down from the flow surface. Similarly, red,

laminated sandstone and siltstone form clastic dikes or crevice-fillings a few centimeters wide in the upper parts of some lava flows.

GEOLOGY OF THE BEAVER BAY COMPLEX

The Beaver Bay Complex (BBC) is a hypabyssal, multiple-intrusive igneous complex that was emplaced into the upper part of the NSVG over a 600-km² area in northeastern Minnesota (Fig. 5-4). Much of this area was the focus of detailed bedrock mapping by the Minnesota Geological Survey between 1985 and 1994 (Miller, 1988; Miller and others, 1989, 1993, 1994; Boerboom and Miller, 1994). Recent and ongoing mapping in the eastern BBC (Albers, 2006; Boerboom and others, 2006) will better established the geology of this part of the complex. Three general areas of the BBC, southern, northern, and eastern, are distinguished on the basis of distinctive rock types and intrusion form (Miller and Chandler, 1997). The relationship of BBC intrusions to other subvolcanic intrusions within the NSVG (Fig. 5-4) is unclear, because of poor exposure to the southwest and insufficient mapping to the northeast. Within the mapped area of the BBC, thirteen intrusive units have been identified that represent at least six major intrusive events (Miller and Chandler, 1997). Most intrusive activity forming the BBC occurred around 1096 Ma based on U-Pb dates of 1095.8±1.2 Ma for a Silver Bay intrusion, the youngest unit of the BBC, and

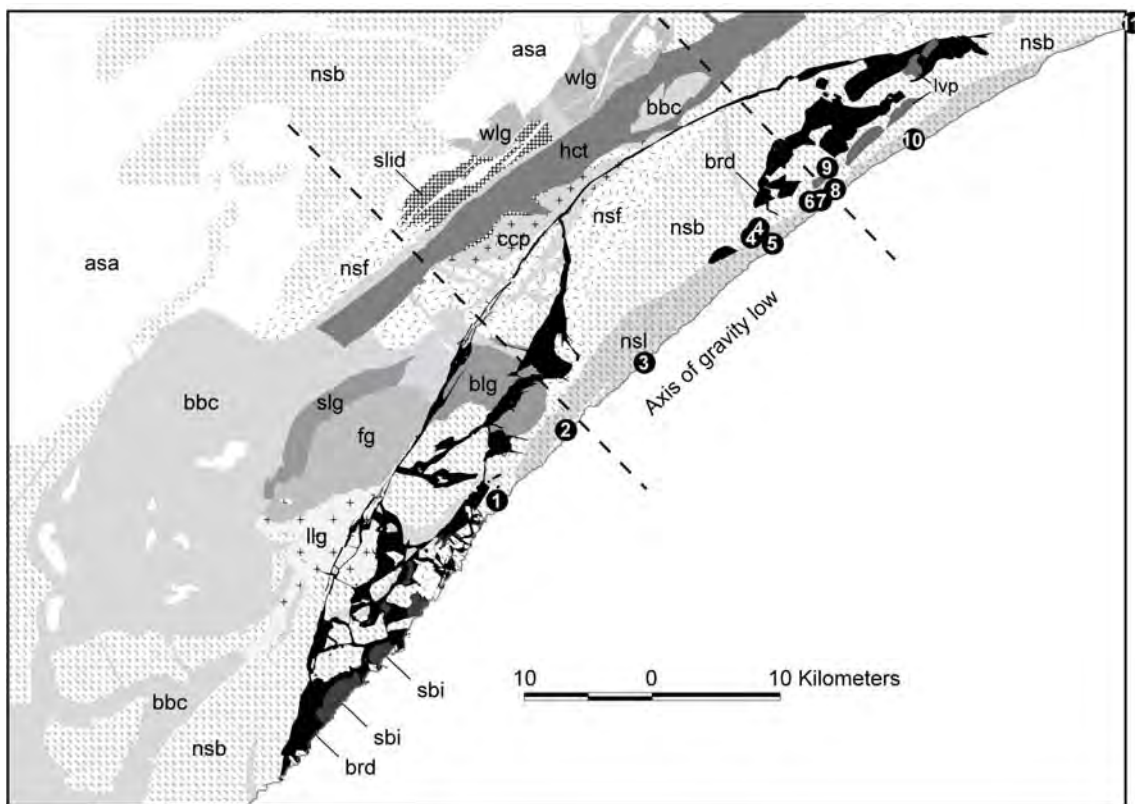


Figure 5-4: Geology of the southern and northern Beaver Bay Complex (after Miller and others, 2001). Units labels are: nsl - NSVG Schroeder basalts; nsb - NSVG basaltic volcanics; nsf - NSVG felsic volcanics; asa - anorthositic series of the Duluth Complex; wlg - Whitefish Lake granophyre; slid - Shoepack Lake inclusion-rich diorite; ccp - Cabin Creek porphyritic diorite; hct - Houghtaling Creek trocolite; blg - Blesner Lake gabbro; llg - Lax Lake gabbro; fg - Finland granophyre and qtz ferromonzonite; slg - Sonju Lake troctolite and gabbro; lvp - Leveaux porphyritic ferrodiorite; brd - Beaver River diabase; sbi - Silver Bay intrusions. See map M-119 for more details. Locations of field trip stops are shown by black dots with corresponding stop number in white.

1096.1±0.8 Ma for the Sonju Lake intrusion (Paces and Miller, 1993). Whether activity overlapped the main stage of Duluth Complex magmatism at 1099 Ma is unknown, because attempts to date the oldest component of the BBC were not successful (Paces and Miller, 1993). The boundary between the BBC and Duluth Complex is generally marked by a northeast-trending keel-shaped intrusion in the northern BBC (Houghtaling Creek troctolite – hct on Fig. 5-4) that separates largely dike and sill intrusions of the BBC to the southeast from massive granophyric granite and extensive areas of structurally complex gabbroic anorthosite to the northwest that are typical of the roof zone of the Duluth Complex.

The range of BBC parent magma compositions is similar to the olivine tholeiite and transitional basalt compositions that dominate the NSVG (Fig. 5-5A). Moreover, like the NSVG, the sequence of intrusion of BBC magmas generally involved progressively more primitive compositions. Compositional variations within the various intrusive units developed as a result of *in situ* magmatic differentiation (Fig. 5-5B), assimilation of footwall rocks, and/or composite intrusions of evolved magma from deeper staging chambers (Fig. 5-5C). The tightly clustered trend of BBC parent magma compositions evident on an AFM diagram (Fig. 5-5A) and the systematic variation of other elemental abundances suggest that all mafic BBC magmas evolved from a common olivine tholeiitic primary magma type. Such a primary composition, which is approximated by the most primitive, high-Al olivine tholeiites of the NSVG, is thought to have given rise to most MCR magmas, especially in later stages of magmatism (Green, 1983;

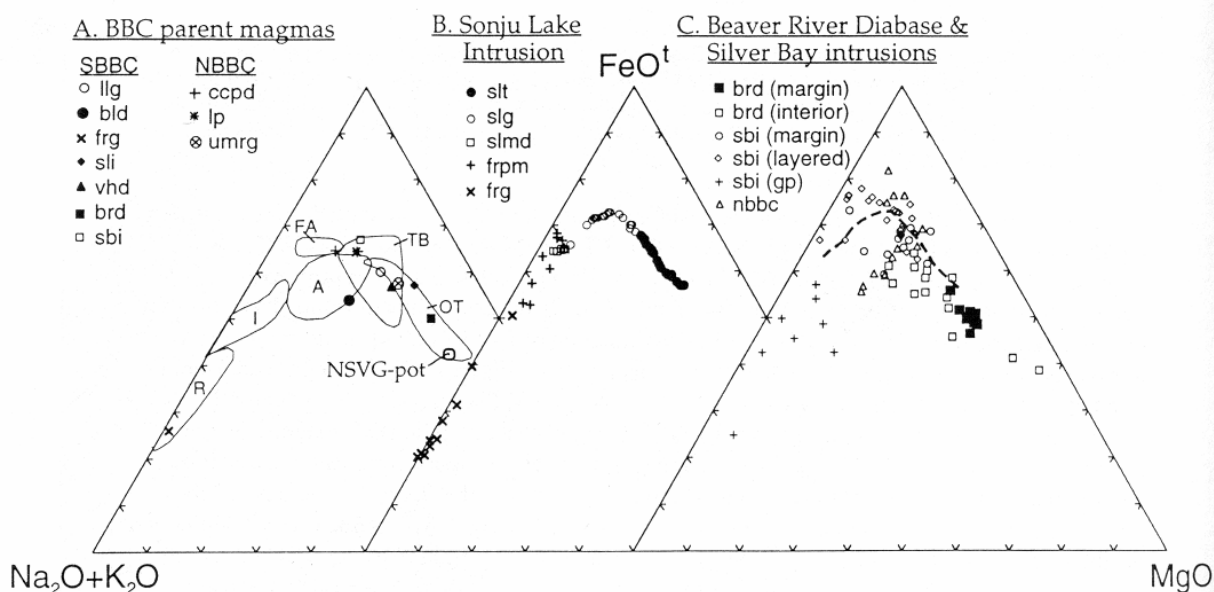


Figure 5-5: AFM diagrams of Beaver Bay Complex (BBC) intrusions. A) Plot of estimated parental magma compositions to various BBC intrusions (see Miller and Chandler, 1997 for unit descriptions and details) compared to major NSVG lava compositions: OT - olivine tholeiite, TB - transitional basalt, A - andesite, FA - ferroandesite, I - icelandite, R - rhyolite (after Green, 1983). NSVG-pot is a primitive NSVG olivine tholeiite composition. The composition of the Leveaux porphyry is indicated by the label – lp. B) Calculated liquid line of descent of the Sonju Lake intrusion through troctolitic (slt), gabbroic (slg-slmd) and monzodiorite (slmd) intervals of the layered sequence. Also plotted are whole rock compositions of Finland granite (frg) and quartz ferromonzodiorite (frpm). C) Whole rock composition plots of Beaver River diabase (brd; ophitic margins and coarse subophitic interiors distinguished), Silver Bay intrusions (sbi; coarse marginal facies, layered ferrogabbroic cumulate interiors, and granophyric compositions distinguished), and composite intrusions from the northern BBC (nbbc). Sonju Lake intrusion differentiation trend (dashed line) is also shown.

Miller and Weiblen, 1990; Klewin and Shirey, 1992). That even the most primitive of the BBC intrusions, the Beaver River diabase, is significantly evolved from a primitive olivine tholeiite composition (Fig. 5.5C) indicates that all BBC parent magmas were generated in turn by magmatic differentiation of such a primary composition in deeper staging chambers. Petrologic modeling of some BBC intrusions and other hypabyssal bodies that intruded the NSVG (Jerde, 1991) suggests that most magmas experienced multistage, polybaric fractionation between their extraction from the mantle and their subvolcanic emplacement. Although the available radiometric ages do not indicate an overlap of magmatic activity between the BBC and the Duluth Complex (Paces and Miller, 1993), additional dates and more detailed petrologic studies may show that some Duluth Complex intrusions acted as the final levels of staging and differentiation of some BBC-bound magmas.

The focus of emplacement of BBC intrusions appears to have migrated toward the rift axis and toward higher stratigraphic levels with time, perhaps reflecting plate drift and thickening of the volcanic pile. Over the exposed extent of the BBC, intrusion shapes appear to have been controlled by a shallow crustal ridge which trends northwest across the BBC. The presence of this buried crustal ridge is indicated by a pronounced saddle in the gravity high over northeastern Minnesota (Chandler, 1990) and the presence of Archean granitic gneiss, biotite schist, metagraywacke and granodiorite inclusions in early intrusions over the gravity low (Boerboom, 1994). The broad network of dikes and sheets characterizing the southern BBC becomes tightly focused into a narrow zone of subparallel dikes in the northern BBC which is situated over the gravity minimum (Fig. 5-4). The eastern BBC opens up again into thick sheet intrusions. Based on geologic, geochronologic, geophysical, and geochemical evidence, Miller and Chandler (1997) suggested that the BBC, particularly the youngest Beaver River diabase dike and sheet network, acted as a magma conduit and structural boundary to the formation and infilling of the western end of the Portage Lake Volcanic basin during the main to late stages of rift volcanism and graben formation. On the North Shore, these volcanic rocks are represented by the Schroeder-Lutsen basalts (Fig. 5-4).

For more detailed information on these and other units of the Beaver Bay Complex see reports by Miller and Chandler (1997) and Miller and others (2002, Chapter 7) and the 1:24,000 bedrock geologic maps of the area (Miller, 1988; Miller and others, 1989, 1993a, 1994; Boerboom and Miller, 1994; Boerboom and others, 2006)

This trip will have stops in the Beaver River diabase and the Leveaux porphyritic ferrodiorite. The Beaver River diabase typically forms prominent, lumpy topographic highs due in part to the large, resistant inclusions of anorthosite contained in it (e.g. Carlton Peak, stop 4B). It occurs as a series of interconnected set of dikes and sills composed predominantly of ophitic olivine gabbro. The northern and western limit is a dike that is sharply defined by outcrop and aeromagnetic data, and the area below this arc is a series of large, bifurcating dikes and thick, gently southeast-dipping sheets that are likely erosional remnants of a former semi-continuous sill (Figure 5-4). The upper and lower margins of the larger diabase sheets commonly contain inclusions of nearly pure anorthosite such as at Carlton Peak, and less commonly, large inclusions of granophyric granite.

The Leveaux porphyritic diorite (LPD) is a semi-continuous, hypabyssal sill, at least 90 meters thick, that intruded into upper NSVG sequence rocks and was later intruded by the Beaver River diabase (Albers, 2006). The LPD makes up five prominent, cuesta-like ridges that trend parallel to Lake Superior over a 20 km distance and numerous smaller knob-like outcrops that occur as large inclusions within the younger Beaver River diabase. Examples of the Leveaux cuestas are Moose and Eagle Mountains, which can be seen uphill from the Lutsen Resort where this meeting is being held.

Recent detailed bedrock geologic mapping by Albers (2006) and Boerboom and others (2006) has subdivided the LPD into four units: the Upper Contact Zone, the Porphyritic Zone, the Aphyric Zone, and the Sparsely Porphyritic Unit. The Upper Contact Zone is a thin (~10m), sparsely porphyritic (~5%), moderately vesicular, and locally amygdaloidal fine-grained ferromonzodiorite that forms the upper part of the LPD sill and is often not exposed due to erosion. The Porphyritic Zone is a thick (40-60m), fine- to

medium fine-grained ferromonzodiorite with 30-40% plagioclase megacrysts that composes the midsection of the LPD sill. The Aphyric Zone is a thin (4-19m), fine-grained ferromonzodiorite with rare plagioclase megacrysts (<3%) that occurs at the base of the sill. The Porphyritic-Aphyric Transition (PAT) is a gradational interval that separates the porphyritic and aphyric zones over a thickness of 10-40 cm. The Sparsely Porphyritic Ferrodiorite unit, which contains 5-20 % plagioclase megacrysts, occurs as outliers to the main LPD ridge, and may represent feeder dikes the LPD sill. The orientation of the PAT and sheet joints from the aphyric section indicate a gentle (10-20 degrees) southeast dip. Stops 6 and 9 will examine the porphyritic and upper contact zones of the Leveaux ferrodiorite.

FIELD TRIP 5 STOP DESCRIPTIONS

The general location of all stops are shown on Figure 5-6. A more detailed location map is included with each stop description, showing a portion of the appropriate 1:24,000 quadrangle map. Also listed are UTM coordinates (nad '83), township-range-section, and the 7.5' quadrangle name.

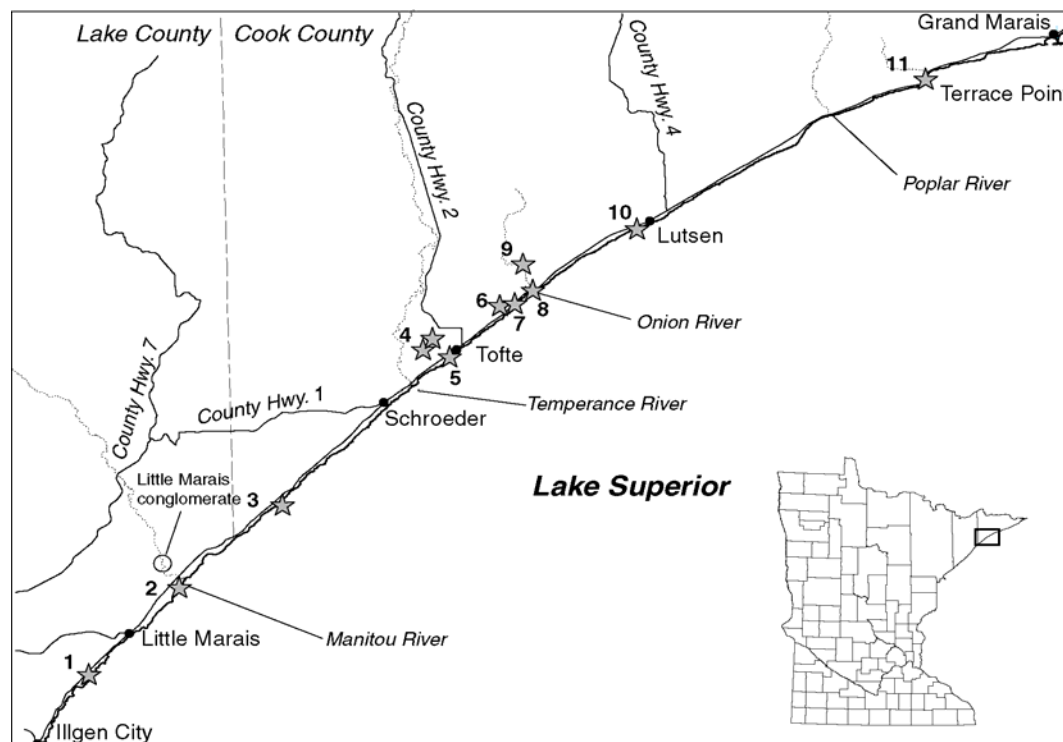


Figure 5-6. Simple map showing relative locations of field trip stops, listed below.

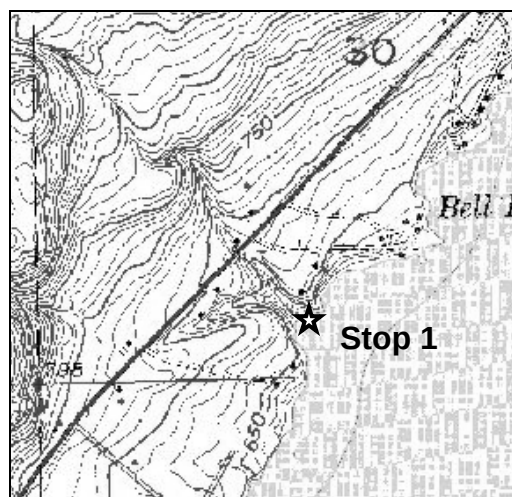
- Stop 1. Kennedy Landing/Bell Harbor – conglomerate below SLB
- Stop 2. Caribou River - Pork Bay breccia and overlying basalt
- Stop 3. Sugar Loaf Cove – Schroeder-Lutsen sequence basalt
- Stop 4. Carlton Peak – Rhyolite and anorthosite
- Stop 5. Tofte Town Park – Schroeder-Lutsen sequence
- Stop 6. Springdale Hill – Porphyritic phases of Leveaux diorite
- Stop 7. Base of Schroeder-Lutsen sequence
- Stop 8. Onion River – Good Harbor Bay andesites
- Stop 9. Oberg Mountain -Leveaux porphyritic diorite and shoreline view
- Stop 10. Conglomerate
- Stop 11. Cut Face Creek – Terrace Point basalt, sandstone, and Good Harbor Bay andesite

---Leaving Lutsen Resort, head southwest on Hwy 61. Proceed approximately 26.5 miles to overpass over prominent creek. We will take the creek bed down to the shore and return to the road by private driveway. This is private property, so permission to access the shore is required. .---

Stop 5-1. Unconformity Conglomerate at Kennedy Landing/Bell Harbor.

UTM coordinates (nad '83) 640454E, 5274226N

T.57N., R.6W., Sec.30, Finland quadrangle



Highlights: ophitic basalt (Schroeder-type), polymict conglomerate, angular unconformity, felsic lithic tuff, agate-bearing quartz tholeiite flow, normal faults, interflow basalt clast conglomerates.

Description: From the cliff face at the west end of this bay, several volcanic and structural features are visible as noted in Figure 5-7. Exposed in the cliff face is an ophitic basalt which is the basal flow of the Schroeder-Lutsen sequence (OB). Underlying this basalt is a polymict volcanogenic conglomerate (PCg). Various types of basalt dominate this conglomerate, but it also contains clasts of felsite, granophyre, lithic tuff, and autolithic

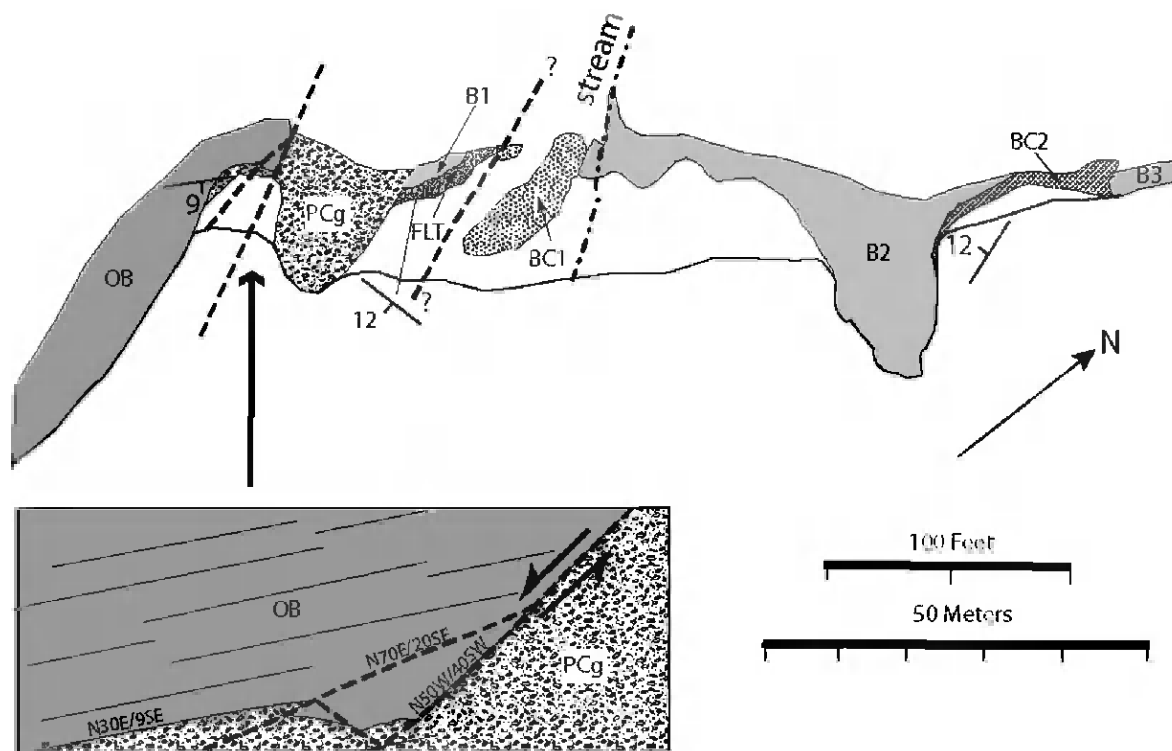


Figure 5-7. Plan view of the geology exposed at Stop 5-1. Unit abbreviations are: OB – ophitic basalt, PCg – polymict conglomerate, B1-3 – basalt flows, FLT – felsic lithic tuff, BC1-2 – basalt clast conglomerate. Inset show fault relationships between OB and PCg units in western cove.

shale fragments. In the back wall of the cove between the ophitic basalt ledge and the point of conglomerate, two splayed faults are exposed (Fig. 5-7, inset). On the northeast side of the point of polymict conglomerate is in sharp angular contact with a thin sheet of basaltic andesite (B1) and underlying a felsic lithic tuff (FLT) with imbricated clasts of porphyritic rhyolite. The contact between the polymict conglomerate and the basalt/tuff appears to be an unconformable erosional surface, but a fault contact cannot be discounted. .

Continuing northeast, this basalt/lithic tuff sequence also abruptly give way across a small gap (fault?) to a basalt clast conglomerate (BC1). Near the outlet of a stream, the BC1 conglomerate grades downsection to a basalt breccia flow top of quartz tholeiite basalt flow (B2). This flow is colored a deep army green and contains abundant vugs and amygdulites of agate and amethyst. Northeast of the point formed by flow B2, another basalt clast conglomerate (BC2) caps the upper part of the next underlying lava flow (B3).

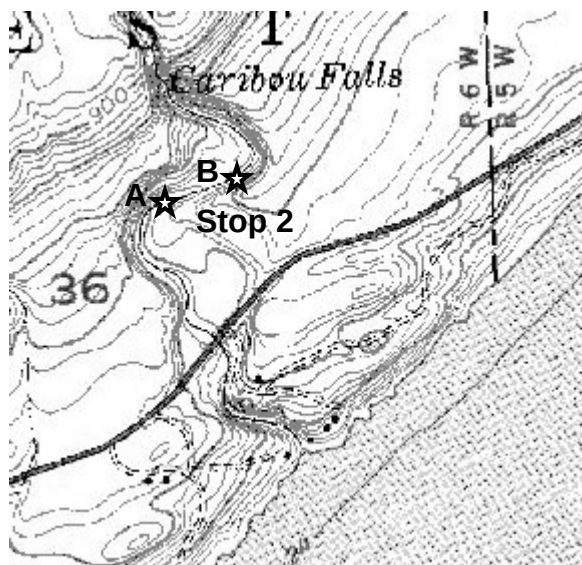
Discussion: The polymict conglomerate, which is termed the Little Marais conglomerate (Miller et al., 1993; Miller et al., 2006), is interpreted to occupy an angular unconformity that separates the Schroeder-Lutsen sequence from the Upper Southwestern sequence of the NSVG. It can be traced at least 10 km to the northeast into the Little Marais quadrangle. Farther to the east, this unconformity develops a conformable relationship with the underlying volcanics and is marked by a thick sandstone unit (Cutface Creek Sandstone) (see Stop 5-11).

---Return northeast on Highway 61. Approximately 7.5 miles after crossing the Caribou River, turn left into parking area. Take the Superior Hiking trail path upstream about 300m to low-lying outcrops on the rivers edge .---

Stop 5-2. Pork Bay Breccia and Manitou River basalt.

UTM coordinates (nad '83) Area A-648250E, 5258850N; Area B-648480E, 5258950N

T.58N., R.6W., Sec. 36, Little Marais quadrangle



Highlights: Volcaniclastic breccia (Pork Bay breccia), thin olivine tholeiite flows, transitional basalt (Manitou River basalt)

Description: Location A- Exposed in low-lying outcrops on the south bank of the river is a brick red, polymict, matrix-supported, volcaniclastic breccia – the Pork Bay breccia. It is composed of subangular to subrounded volcanogenic clasts up to 40 cm across in an unbedded, sandy matrix (Fig. 5-8A). Clasts are composed dominantly of massive to scoriaceous and oxidized to fresh basalt and minor amounts (<5%) of other rift-related rock types (e.g., gabbro, felsite, volcanic sandstone). The matrix consists of granule to medium sand-sized volcanic fragments cemented by calcite and zeolite.

Location B – Progressing upstream about 200 m to where the river make a sharp turn to the left (north), units overlying the Pork Bay Breccia are exposed (Fig. 5.8B). In the streambed are several thin (<1m) amygdaloidal olivine tholeiite flows oriented N50E/~20°SE. Exposures of volcaniclastic breccia occur

adjacent to the thin basalts, but contacts are not exposed. At the base of the cliff face on the southern shore are poor exposures of the breccia which presumably overlie the thin basalts. The overlying this breccia and exposed over most of the 8m-high cliff face is a massive, slightly Pl-porphyrific transitional basalt – the Manitou River basalt. At the basal contact, this flow is strongly chilled possibly irregular pockets of glass.

We will not have time to continue upstream to Caribou Falls, however, a later visit is encouraged. The falls is interpreted to represent a major fault scarp that juxtaposes the Pork Bay breccia (downstream) with olivine tholeiitic basalts of the Schroeder-Lutsen sequence. Despite ample exposure in the falls area, a well-identified fault zone could not be identified leaving open the possibility that the contact may be steep lithologic contact rather than a structural one.

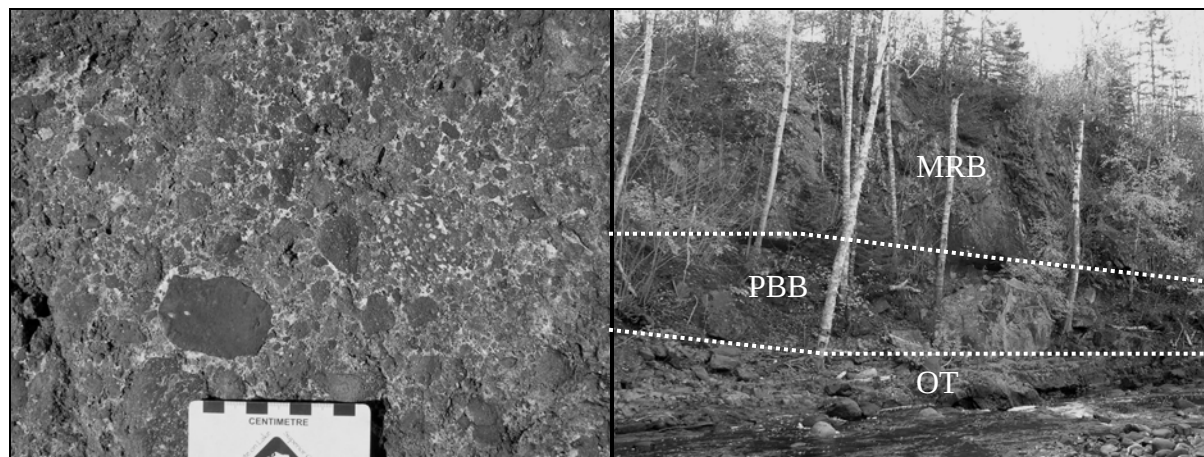


Figure 5-8. A) Close-up photo of the polymict clasts in the Pork Bay breccia; whitish interstitial mineral is mostly calcite. B) Exposures in southern stream bank at Location B – OT – thin olivine tholeiite flows in streambed, PPB – breccia exposed in lower cliff face, MRB – massive, transitional basalt exposed in upper cliff face.

Discussion: The origin of the Pork Bay Breccia is enigmatic. Various ideas put forth to explain its formation include that it is a lahar or some type of mass wasting deposit, that it represents a collapse breccia within a caldera, and that it represents deposits of an explosive cinder cone. None of these completely explain the salient attributes of the unit or its relationship with adjacent units. These attributes include: the wide variety of volcanic, intrusive and sedimentary clast types, the subrounded nature of the clasts, the porous sandy matrix, the lack of bedding, the thickness of the breccia (up to 60 m), the occurrence of variable thicknesses of tholeiitic basalts (2- 100m) between the breccia and the Manitou River basalt, and the occurrence of basalt flows within the upper parts of the breccia (as seen at location B). Given these attributes, we will discuss possible modes of origin on the outcrop.

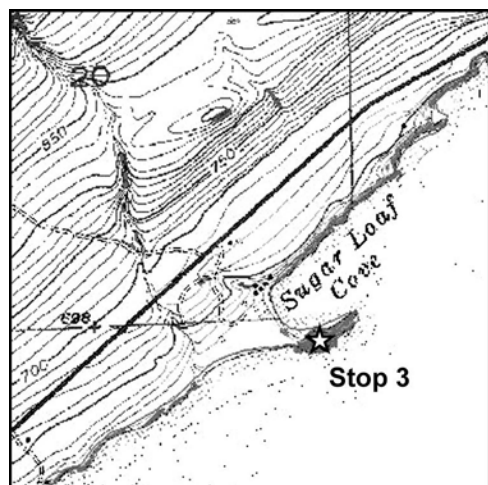
---Continue northeast on Highway 61. Approximately 2.7 miles from the Caribou River, turn right into Sugar Loaf Cove Interpretive Center (look for a brown highway sign). Park in the lot and follow trails down to Lake Superior shore.---

Stop 5-3. Olivine tholeiitic basalt flows at Sugarloaf Cove.

UTM coordinates (nad '83) 652044E, 5261120N

T.58N., R.5W., Sec. 29, Little Marais quadrangle

No hammers please!



Note: This stop is on the property of the Sugarloaf Cove Interpretive Center Association and a State of Minnesota Scientific and Natural Area. It has been established to preserve the natural features of the beach and point for all to enjoy. Please leave your hammers behind and take only pictures, not samples.

Highlights: Morphological features of olivine tholeiitic basalt flows.

Description: The lava flows exposed at and near this stop are typical of the Schroeder-Lutsen basalts. Flows exhibit features characteristic of fluid, low-viscosity pahoehoe olivine tholeiites (Fig. 5-9). Columnar joints and sandstone-filled clastic dikes (sand-filled cracks) are common in the upper part of the massive flow interiors.

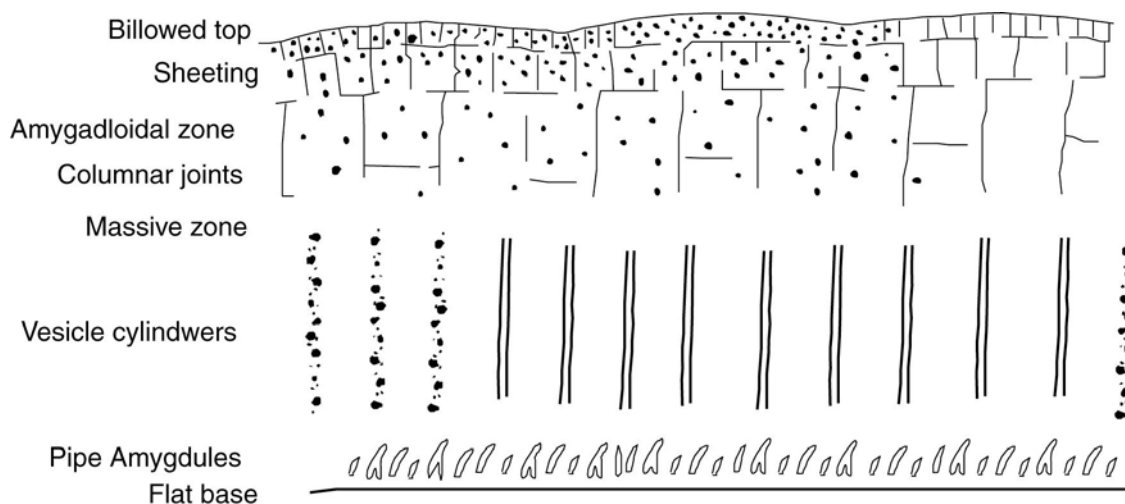


Figure 5-9. Idealized internal structure of olivine tholeiite-composition lava flows. From Green, 1989.

Six lava flows are exposed on the point (Fig 5-10A). The about 10 m of the upper part of flow 1, an ophitic basalt, is exposed on the southwest end of the point. Flows 2 and 3 are 1- to 2-meters thick and display bent pipe amygdules and oxidized ropey flow tops (Fig. 5-10B). Flow 4 is 2-3 meters thick and displays well developed columnar joints, ropey flow tops, and abundant clastic dikes at its eastern extent. Flows 5 and 6 define a trough structure atop flow 4 in the midsection of the point. At the northeast end, flow 5 occurs as two thin (~1m) flows and over 5 meters of ophitic basalt of represent the lower part of flow 6.

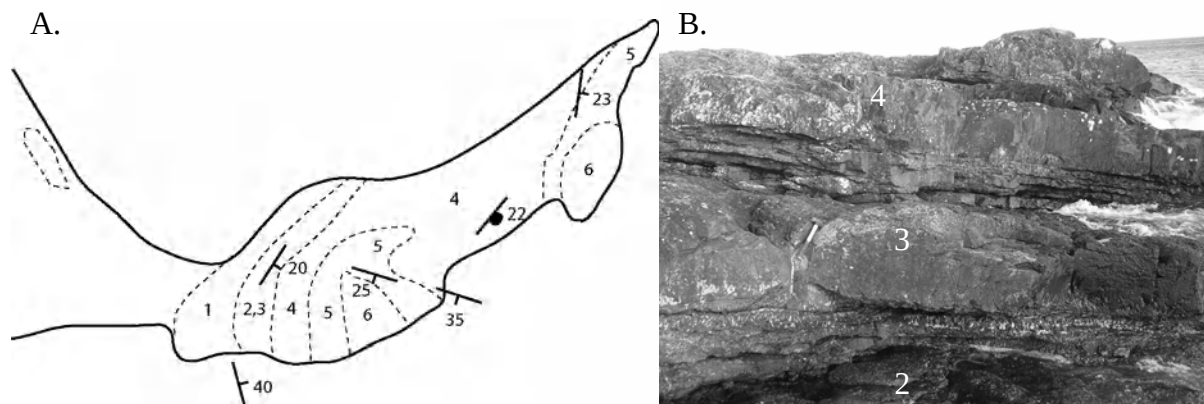


Figure 5-10. A) Sketch of distribution and orientation of six flows on Sugarloaf Cove Point. B) Photo of thin lava flow 2-4 at the west end of the point.

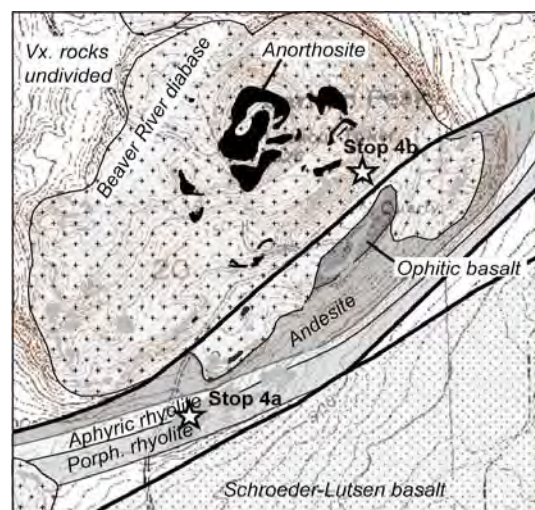
Discussion: These olivine tholeiite basalt flows are typical of the Schroeder-Lutsen sequence, the uppermost sequence of the North Shore Volcanic Group. In this area the Schroeder-Lutsen basalts unconformably overlie a thick sequence of mafic to felsic volcanic rocks, informally termed the Cross River lavas (Boerboom and Green, 2006), that strike approximately N55E and dip 10-20° SE, in contrast with the nearly shore-parallel strike of the Schroeder-Lutsen basalts (see Fig. 5-15, stop 7). The occurrence of these olivine tholeiite flows at the top of the NSVG highlights the fact that the NSVG generally becomes more primitive (less evolved) up section.

---Return to Highway 61 and drive NE approximately 7 miles to Temperance River. Turn left on small gravel road 0.8 miles NE of Temperance River across from National Forest Service Sign. Drive 0.95 miles up hill, over outcrops in road bed (Schroeder basalt) and park; Stop 4A outcrops just east of road.--

Stop 5-4A. Carlton Peak rhyolites.

UTM coordinates (nad '83) 660737E, 5271317N

T.59N., R.4W., Sec. 20, Tofte quadrangle



Highlights: Aphyric and porphyritic rhyolite flows.

Description: There are several low outcrops in this area of rhyolite (at the south edge) and porphyritic rhyolite (at the north edge). There are also outcrops of andesite to the northeast of the rhyolite, and of ophitic basalt to the north of that. The porphyritic rhyolite is grayish-pink, with 5-8% phenocrysts of K-feldspar, quartz, oxidized iron silicate minerals, and magnetite. The aphyric rhyolite is light gray to purplish-red, with rare small phenocrysts of quartz, K-feldspar, altered mafic silicates, and magnetite. Small vuggy quartz-lined amygdules and tension gashes are common throughout.

Discussion: The rhyolites here are part of the Carlton quarry sequence, which includes ophitic basalt at the top,

andesite, porphyritic rhyolite, and aphyric rhyolite at the base. This sequence has been intruded by the Beaver River diabase (stop 4B). This entire set of flows dips moderately northwest, and is presumably fault-bounded. The porphyritic rhyolite has an age of $1,094 \pm 2.0$ Ma (Green and others, 2001), the youngest age obtained to date from Keweenaw rocks in Minnesota, and is comparable in age and stratigraphic position to rhyolite in the Porcupine Volcanics in Michigan and to the upper part of the Portage Lake Volcanics (Green, 2002 and references therein). The position of the Carlton Quarry lavas within the NSVG is somewhat enigmatic, given the age, and fortuitous gaps in outcrop preclude full understanding of the relationships of these lavas to the rest of the NSVG.

---Continue on up the gravel road to a large flat area below the main quarry face, or turn left on the way and wind your way up to the quarry wall. If you go up there, BE CAREFUL!---

Stop 5-4B. Carlton Peak anorthosite and Beaver River diabase.

UTM coordinates (nad '83) 661309E, 5272060N

T.59N., R.4W., Sec. 20, Tofte quadrangle

Highlights: Carlton Peak anorthosite, Beaver River diabase (BRD).

Description: In the quarry staging area are multiple large blocks of pale green to locally purple, coarse-grained anorthosite, and blocks of weathered, dark green, Beaver River diabase. The anorthosite contains scattered 2-3 cm diameter dark clots of altered poikilitic olivine, but otherwise is composed of nearly 100% fresh, glassy plagioclase.

The anorthosite is well exposed in natural outcrops at the top of Carlton Peak, which requires a moderately strenuous hike of approximately one mile on the Superior Hiking Trail, accessed by driving 2.8 miles north of Tofte on County Highway 2 (Caribou Trail) to the well-marked Superior Hiking Trail parking lot. An alternative short hike to the top of Britton Peak, located just east of and above the parking lot, leads to an inclusion of the same type of anorthosite, and a clear view of Carlton Peak. Other places on the shore where these anorthosite inclusions can be easily seen are at Split Rock Lighthouse State Park (see guidebooks by Miller and others, 1987, Stop 1 and Boerboom and others, 2004, Stop 2-8), and Tettegouche State Park. The easiest place of all is a newly-constructed road cut on Highway 61 in Silver Bay, which has clean exposures of anorthosite inclusions within the BRD.

Discussion: Carlton Peak is a prominent topographic knob formed by the Carlton Peak anorthosite, which occurs as large inclusions in the BRD. Here and at many other places along the North Shore, anorthosite forms rounded, bare knobs because it is very massive, contains few fractures, and is more resistant to chemical weathering than the surrounding diabase (Fig. 5-11). Individual anorthosite inclusions at Carlton Peak range up to 1100 feet in diameter, and are among the largest of the known anorthosite inclusions in the BRD.

The anorthosite inclusions throughout the BRD are round to angular in shape and vary in size from individual plagioclase crystals to large inclusions such as at Carlton Peak, and some are brecciated and recrystallized (Morrison and others, 1983). They are all composed of 90-99% coarse-grained bytownite to labradorite, and minor Mg-rich olivine, orthopyroxene, and less commonly clinopyroxene. The inclusions are most common near the bottoms of BRD sills, and in sharp contact with the BRD, but the BRD is not chilled against them.

Anorthosite inclusions of this type are different from anorthositic-series rocks of the Duluth Complex, but there are similar inclusions of anorthosite within the anorthositic series. The highly disordered structural state of plagioclase and the absence of chilled margins in the BRD against the anorthosite indicate these inclusions were derived from a lower to mid-crustal source (Miller and Weiblen (1990).

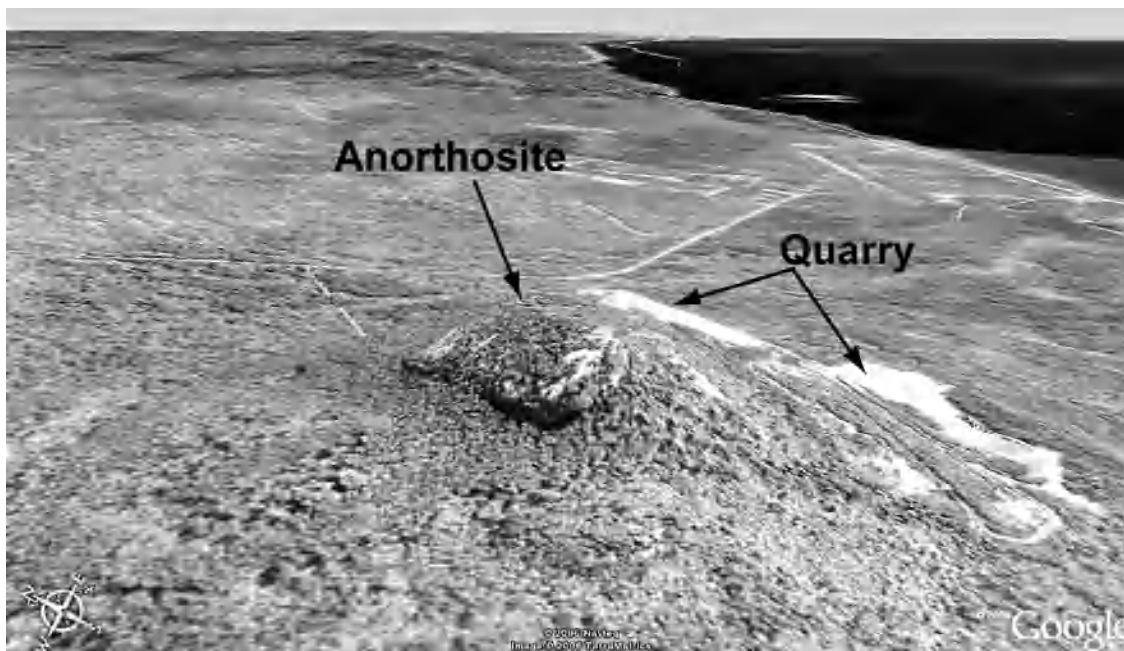


Figure 5-11. Google earth oblique image of Carlton Peak anorthosite, looking northeast up the shore. From this perspective it appears to be a large slab-shaped inclusion.

Isotopic and trace-element compositions of the anorthosite xenoliths imply that they may be pre-Keweenaw in age (Morrison and others, 1983), but the data are ambiguous. As an alternative, Miller and Weiblen (1990) suggest a crystal-mush theory in which significant proportions of Keweenaw anorthosite may have been generated by flotation of plagioclase under high pressure in the deep crust prior to Beaver Bay Complex magmatism at 1,096 Ma. Plagioclase cumulates formed under deep crustal conditions would probably be distinctive in character and composition, compared to anorthosite cumulates formed in the shallow crust. The ambiguous isotopic compositions of the inclusions may indicate that the anorthosites formed from Keweenaw magmas that were contaminated by older crust, rather than older anorthosites being contaminated by interaction with Keweenaw magmas.

The Beaver River diabase (BRD) forms an extensive network of dikes and sills, and is one of the youngest intrusive phases of the Beaver Bay Complex. The BRD extends from Split Rock Point to just west of Grand Marais, in a large arc that is approximately 57 miles (90 km) long (Figure 5-4). Most of the diabase is fine-grained ophitic olivine gabbro, but the centers of some of the thicker sills (up to 150 meters) grade into coarse-grained subophitic to intergranular gabbro.

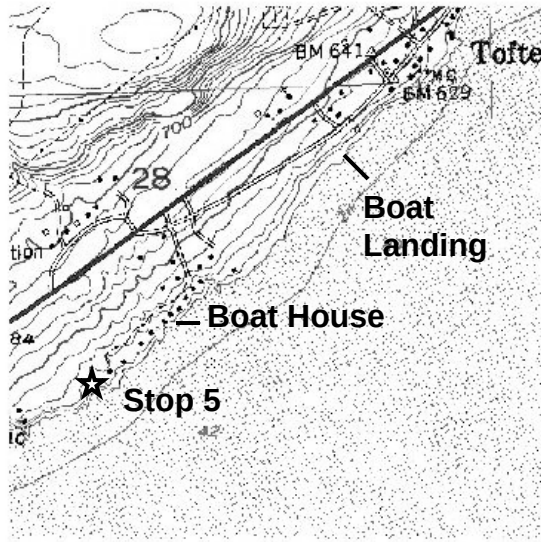
Historic sidelight: In 1902, anorthosite from an inclusion similar to this, in present-day Split Rock Lighthouse State Park, was quarried by a group who mistakenly thought the anorthosite was corundum. In fact there is actually a high point of land on Lake Superior in the park, held up by anorthosite, known as Corundum Point. Another quarry site was near the Baptism River at Illgen City. Their plan was to sell the “corundum” on the east coast, but they had to abandon that plan and turn to other ventures. This company went on to become the Minnesota Mining and Manufacturing Company, now simply 3M!

---Drive back to Highway 61 and head northeast towards Tofte for 0.8 miles, then veer right onto Tofte Park Road (24). Take first private drive to the right and head to shore OR continue to Tofte town park and boat access on right after about ¼ mile. From boat landing, head southwest about 1 km along semi-continuous outcrop of tholeiitic lava flows (~7 flows). Stop is 250m southwest of old boathouse.---

Stop 5-5. Flow features and feeder dikes in Schroeder basalts

UTM coordinates (nad '83) 661800E, 5270200N

T.59N., R.4W., Sec. 28, Tofte quadrangle



Highlights: Ophitic olivine tholeiite basalt flows, columnar jointing, amygdaloidal feeder dikes.

Description: Accessing the shore at the old boathouse (662000E, 52700350N), a smooth billowing flow contact dipping gently to the northeast is exposed at the boat ramp. Heading southwest about 30 meters, a north-trending 0.5m-wide feeder dike is exposed in the amygdaloidal flow. The dike is zoned in the degree of oxidation and in the concentration of vesicles and amygdules (Fig. 5-12A).

Continuing about 130 meters to the southwest over the polygonally-jointed surface of an ophitic basalt flow, a domal flow contact with a thin flow is encountered. Passing across this dome back onto

the ophitic basalt, one can see a feeder dike breaking through the amygdaloidal top of the underlying flow and feeding the ophitic basalt. The feeder dike is lined with pipe amygdules that are curved upward (Fig. 5-12B).

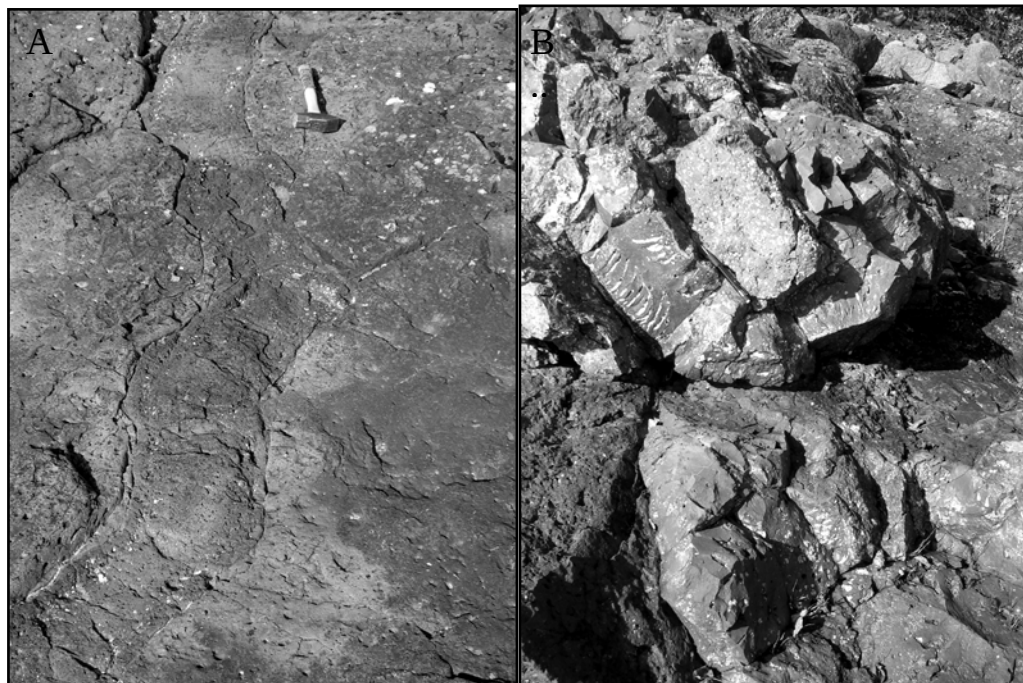


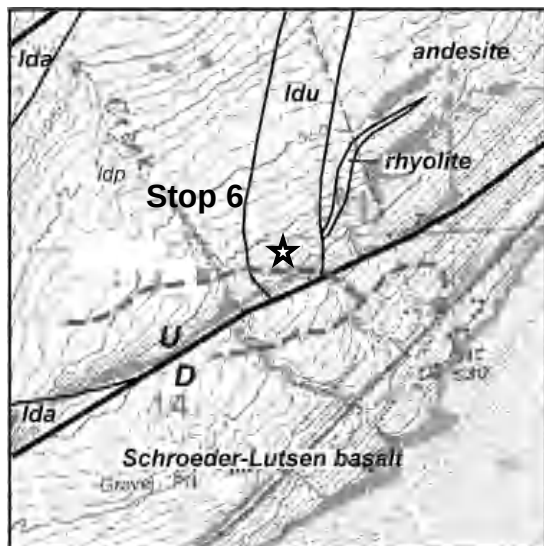
Figure 5-12. A) Feeder dike with zonation in oxidation and amygdule concentration. B) Feeder dike with upturned pipe amygdules.

---Continue northeast on Tofte Park Road to Highway 61. Turn right (northeast) and go three miles to Leveaux Ridge road on the left, across from the Chateau Leveaux. Go up Leveaux Ridge Road a short distance (~0.25 mi)., then turn right on Overlook Trail. Head uphill(~0.3 mi), park where road curves to west near driveway on right (21 Overlook Tr.). ---

Stop 5-6. Porphyritic Phases of Leveaux Porphyritic Diorite (LPD)

UTM coordinates (nad '83) 665850E, 5274260N

T.59N., R.4W., Sec. 14, Tofte quadrangle



Highlights: Amygdaloidal olivine tholeiite basalt of Schroeder Lutsen sequence, inferred fault zone, amygdaloidal, sparsely porphyritic upper zone and porphyritic zones of the Leveaux porphyritic sill.

Description: Exposures in a low ledge and the drainage ditch south of the driveway are moderately amygdaloidal intergranular basaltic andesite. Several mineralized northeast-trending shear zones cut the rock.

On the opposite side of the driveway is a 5' roadcut of sparsely plagioclase porphyritic and amygdaloidal, fine-grained ferrodiorite. Centimeter-sized plagioclase phenocrysts compose about 5% of the rock as does a similar abundance of quartz amygdules (Fig. 5-13A). This rock type continues along this roadcut to the next driveway (#33).

Opposite the bridge across a substantial creek are pavement outcrops of strongly porphyritic (30-40%) ferrodiorite which are typical of the main Porphyritic Zone of the LPD sill (Fig. 5-13B)



Figure 5-13. A) Outcrop appearance of sparsely porphyritic and amygdaloidal ferrodiorite typical of the Upper Zone of the Leveaux ferrodiorite. B) Outcrop appearance of strongly porphyritic ferrodiorite typical of the main Porphyritic Zone of the Leveaux ferrodiorite.

Discussion: This exposure of the Leveaux porphyritic diorite sill is the only location where the upper contact zone is preserved. Along most of the cuesta mountain exposures, the upper part of the sill is

eroded into the porphyritic zone (see Stop 8). The vesicular/amygdaloidal character of the upper contact zone speaks to the shallow level of emplacement of the LPD into the volcanic edifice. The sparsely porphyritic nature of this zone is interpreted by Albers (2006) to have been caused by flow differentiation of phenocrysts out of the upper contact zone during emplacement and subsequent rapid crystallization.

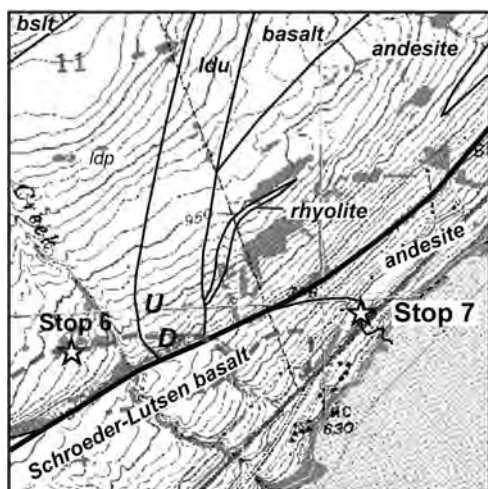
Also, this inferred fault is a major fault zone which can be traced for 10's of kilometers subparallel to the shore.

---Go back to Highway 61 and continue NE 0.16 miles to roadcuts on uphill side of Highway. Park at edge of highway. WATCH OUT FOR HEAVY TRAFFIC!. ---

Stop 5-7. Base of Schroeder-Lutsen Sequence and top of underlying andesites

UTM coordinates (nad '83) 666381E, 5274345N

T.59N., R.4W., Sec. 13 Tofte quadrangle



Highlights: Contact between base of Schroeder-Lutsen basalt sequence and underlying Good Harbor Bay andesites. This stop will be at a roadcut, but if permission can be obtained a better outcrop may be visited down on the lake shore. The description below is of the lakeshore outcrop, but most of the features in it can be seen on the roadcut.

Description: Upper part of andesites is composed of amygdaloidal andesite fragments up to 50 cm in size, surrounded by reddish-brown sandstone. The andesite clasts are angular, and vary in shape from equant to slab-like; the latter commonly oriented near vertical. Sand makes up 30-40% of the volume of the upper 3 meters of the breccia, and the breccia is capped by a cross-bedded sandstone wedge that is cut off by erosion on a cliff corner (Fig. 5-14).

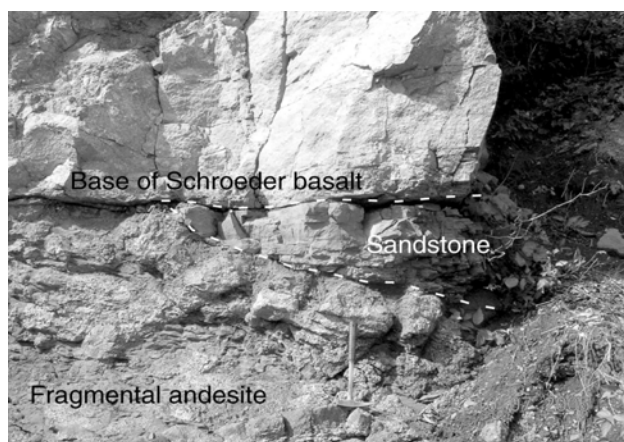
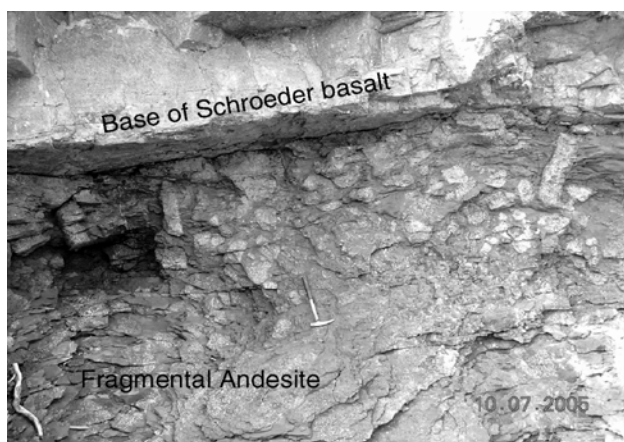
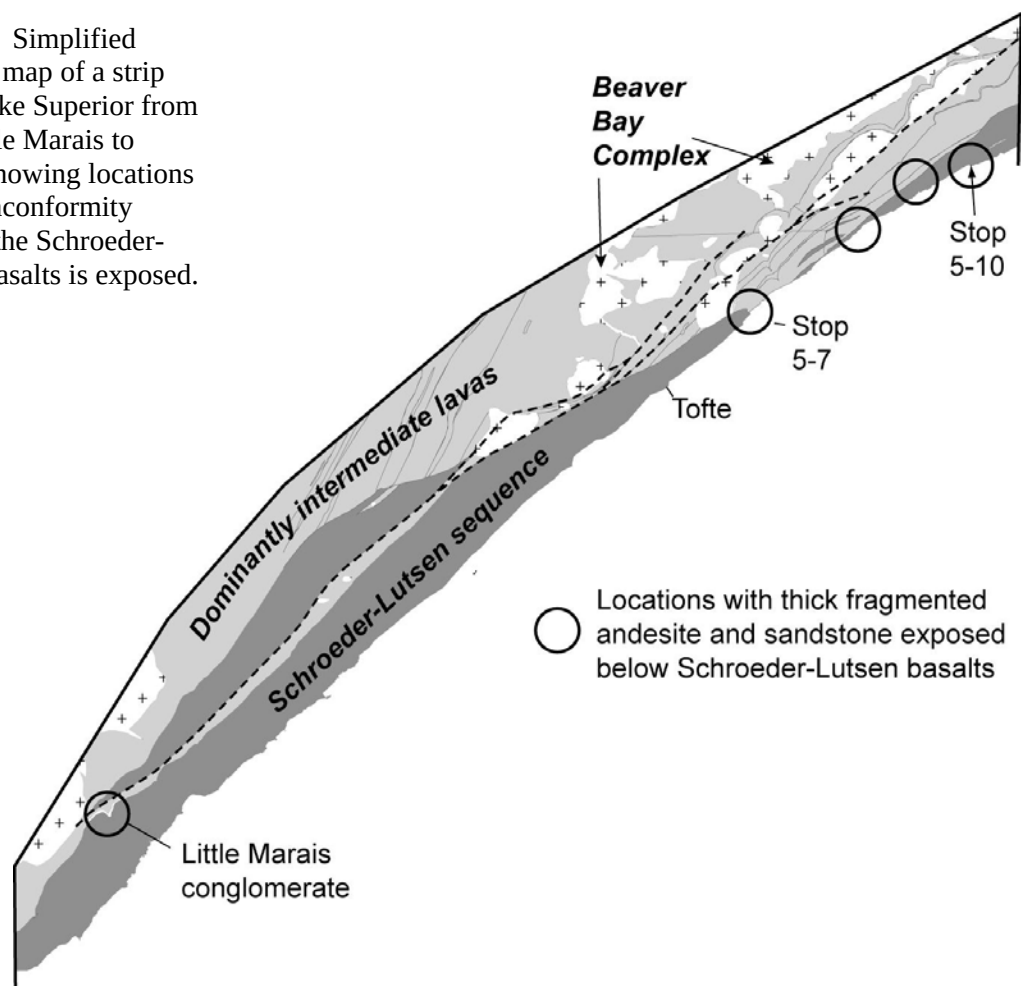


Figure 5-14. Photographs of base of Schroeder basalt and underlying fragmental andesite, from outcrop on shoreline below stop 7. Hammer is approximately 45 cm long.

The base of the Schroeder basalt flow undulates slightly and contains pipe amygdules. In places, if the light is right, one can make out vague, bulbous lobe-like forms approximately 30 cm in diameter at the base of the flow.

Discussion: There are two more exposures between here and Lutsen similar to this one, all characterized by large, boulder-sized and blocky amygdaloidal andesite clasts surrounded by sandstone, with a thin, discontinuous sandstone cap. These are also similar to andesite exposed beneath the Little Marais conglomerate, approximately 15 miles to the southwest (Fig. 5-15; Miller and others, 2006). The Little Marais Conglomerate is at least 60 feet thick, and composed mostly of sandstone, but contains some silt/shale beds, as well as large rounded boulders of basalt near the base and large angular blocks of basalt higher in the section that are ‘floating’ in sandstone. In all cases the underlying fragmental andesites are unusually thick, with substantially larger clasts than typical andesitic flow-top breccias. Despite the fact that the Schroeder-Lutsen sequence is clearly unconformable to the underlying rocks, as made evident by the discordant strike directions between the two (Fig. 5-15), the andesite flow tops below are all fragmental, at least in the available exposures. The upper surface of the Good Harbor Bay andesites may have been a slightly eroded surface made up of clinkery lava flows with an irregular topography, covered by sedimentary rocks of variable thickness. Local fault escarpments that formed prior to eruption of the Schroeder-Lutsen basalts may account for the unusually thick Little Marais conglomerate, with blocks of basalt fallen off the fault escarpment into the sedimentary pile.

Figure 5-15. Simplified geologic map of a strip along Lake Superior from near Little Marais to Lutsen showing locations where unconformity beneath the Schroeder-Lutsen basalts is exposed.



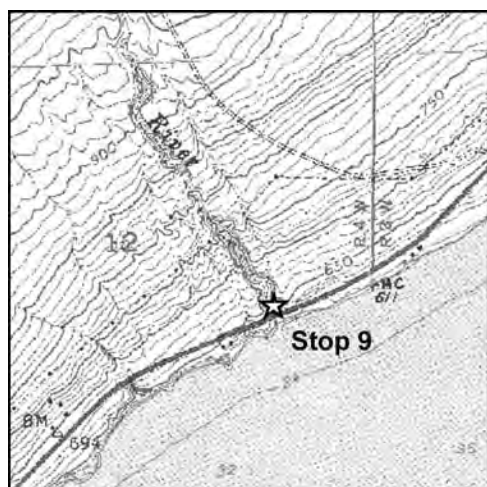
---**DIRECTIONS:** Continue northeast on Highway 61 about 0.7 miles. Just past the Onion River, pull into the Ray Berglund State Scenic Memorial Wayside. Park there and walk SW along highway to Onion River. An excerpt from MN Dept. of Transportation website: “The Onion River derived its name from a Paul Bunyan legend. The legend says that wild onions grew in such abundance in the area, that Paul Bunyan’s loggers shed tears while cutting lumber”.---

Stop 5-8. Morphology of andesite lava flows in the Onion River valley.

UTM coordinates (nad '83) 667525E, 5275197N

T.59N., R.4W., Sec. 12 Tofte quadrangle

**The extent to which these flows can be seen is dependent upon the water level in the Onion River. If the river is high, some outcrops can be accessed on the highway and on the shoreline near the river mouth.*



Highlights: Andesite flow features, Terrace Point basalt, minor faults.

Description: This stop will go through a multiple series of andesitic lava flows which exhibit classic flow features (Fig. 5-16). The andesite is porphyritic (plagioclase, clinopyroxene, some magnetite), with a matrix of fine felty plagioclase, interstitial K-feldspar, granular clinopyroxene, and blocky Fe-Ti oxides.

A sand-filled flow-top breccia that caps the first flow in the river series is exposed on the roadcut just SW of the Onion River and on the shore at the river mouth. The base of the same flow, which exhibits classic andesite oxidation-lamination texture, from which the attitude of the flow can be reliably measured, is exposed in the river bed below the highway.

A 3 meter wide brittle fault (attitude N45E, 85°N), with reverse sense of movement (north side down approximately 2 meters) is exposed about 200 feet north of the highway on the east bank. The fault zone contains brecciated andesite fragments surrounded by coarse white calcite and pinkish-orange laumontite. This fault projects back southwest across the river to a till-covered portion of the slope. Next to the fault, upstream, is another thick sand-filled flow-top breccia at the top of the next underlying flow.

Proceed upstream past a total of five andesite flows. The sixth flow is the western end of the Terrace Point basalt flow (see also Stop 11 and Fig. 5-18). Although the top of the flow is fragmental similar to the andesite flows downstream, it is internally ophitic (very rare for an NSVG andesite), and contains rare fresh and glassy, blocky plagioclase phenocrysts. The next flow below this ophitic basalt is probably icelandite in composition, but lacking chemistry has been lumped with the andesites. After a series of andesite and icelandite flows, the river makes a sharp bend to the east where it intersects a prominent fault marked with slickenside surfaces that forms a high escarpment on the south side of the river. Above this is a narrow canyon that can be traversed in late summer when the water level is lowest, through a series of thin, irregular olivine tholeiite basalt flows. Traverse upstream until a snowmobile trail bridge is encountered; there is a small trail along the east side of the river that goes south from the snowmobile trail back to the parking area.

Discussion: These flows were original termed the Onion River lavas by Boerboom and others (2006), with all the lava flows from Highway 61 to beyond the snowmobile bridge lumped together. Subsequent mapping to the east (Boerboom and others, 2007) has shown that the Onion River lavas are actually the western end of the Good Harbor Bay andesites. To the east the Terrace Point basalt is underlain by the Cut Face Creek sandstone, but that apparently pinches out and is no longer present here.

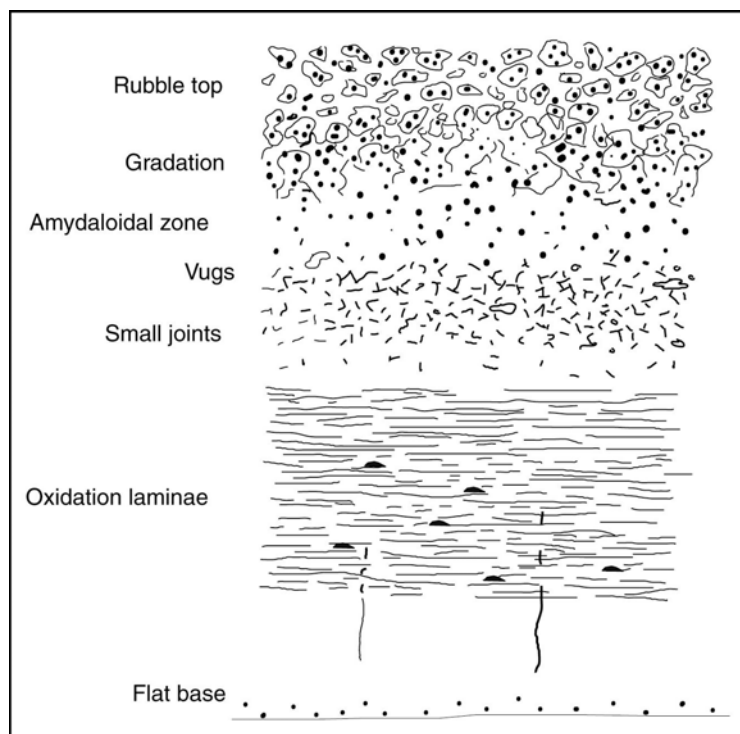


Figure 5-16. Idealized flow structure of quartz tholeiite to andesite-composition lava flows. From Green, 1989.

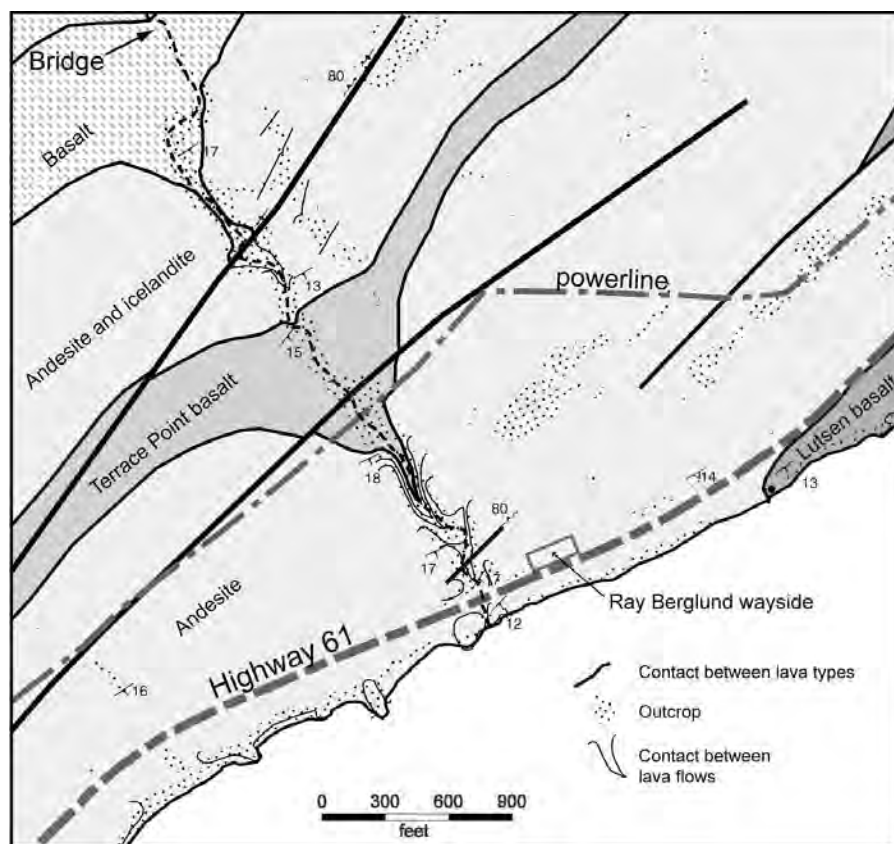


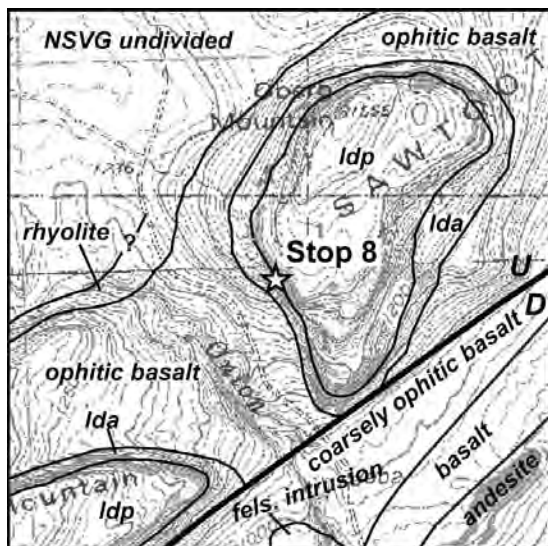
Figure 5-17. Geologic map of the Onion River lavas.

---Continue 0.4 miles northeast on Highway 61 to the Onion River Road. Take the Onion R. Rd 1.7 miles to parking area for the Superior Hiking Trail. Take Oberg Mtn trail about 0.6mi (~1000m) to the southwest overlook . ---

Stop 5-9. Porphyritic Zone of Leveaux Porphyritic Diorite (optional stop)

UTM coordinates (nad '83) 666960E, 5276900N

T.59N., R.4W., Sec. 1 Tofte and Honeymoon quadrangles



Highlights: Porphyritic Zone of the LPD, vista of Leveaux Mountain and shoreline.

Description: At the southwestern crest of Oberg Mtn, exposures of typical porphyritic ferrodiorite of the upper Porphyritic Zone of the LPD are exposed. The weathering of the outcrop accentuates the porphyritic texture of the white plagioclase against a brownish matrix.

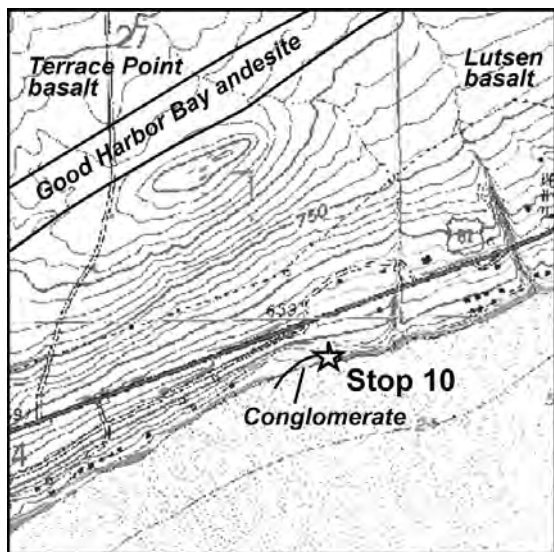
From this vantage point, the southeast-dipping sill can be seen in the topographic expression of Leveaux Mountain visible across the Onion River valley to the southwest. Farther southwest just off shore, Gull and Bear Islands can be seen (on a clear day) near the Taconite Harbor power plant, about 10 miles away. These islands are formed by outliers of LPD and represent the southernmost exposures of the Leveaux porphyritic diorite sill.

---DIRECTION: Continue east on Highway 61 for about 2.5 miles to the Poplar River. 600 feet past the Poplar River, veer right onto County Road 35 (gravel). Drive east on this road 1 mile and park at pull off where road joins back onto Highway 61. Walk straight down to lake through the woods to small cove.---

Stop 5-10. Interflow conglomerate.

UTM coordinates (nad '83) 673942E, 5279298N

T.63N., R.3W., Sec. 24 Lutsen quadrangle



Highlights: Interflow conglomerate, flow contact.

Description: Small, 2 meter thick exposure of reddish-brown, poorly bedded, moderately well sorted and rounded, cobble conglomerate exposed below a thick flow of Lutsen basalt. The conglomerate is exposed as a small erosional window through the Lutsen basalts.

Cobbles in the conglomerate are mostly less than 5 cm in diameter, in a matrix of coarse gritty sand and white calcite cement. In order of abundance, the clasts are composed of ambiguous brown, fine-grained mafic to intermediate volcanic rocks, flow-banded rhyolite, amygdaloidal basalt possibly similar to the adjacent Lutsen basalt, reworked sandstone, and rare agate. In thin section many of the clasts are composed entirely of a felty-matted, gray-birefringent mineral, presumably some type of secondary zeolite; others are replaced by

calcite in irregular alteration zones and clots. Less altered clasts exhibit relict flow banding typical of NSVG rhyolites, and relict textures typical of NSVG mafic to intermediate volcanic rocks. The contact with the overlying flow of dark green, ophitic Lutsen basalt is very sharp and straight, measured at N70E, 15° SE, typical for the NSVG.

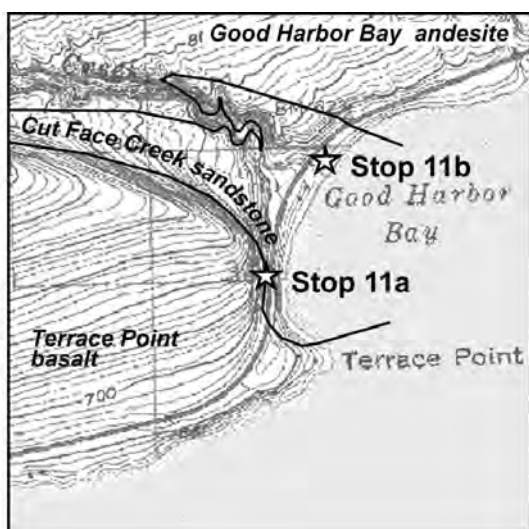
Discussion: Further northeast up the shore there are some reddish-brown, fine-grained, interflow sandstones contained entirely within the Lutsen basalts, as scattered 2-3 meter thick and one 75 meter thick bed. Although the conglomerate at this stop is likely not more than a few meters thick and may similarly be contained entirely within the Lutsen basalts, it does not outcrop in nearby stream exposures where anticipated if it were a conformable layer in the Lutsen basalts. More likely, this exposure is a window eroded through the base of the Schroeder-Lutsen sequence, and the conglomerate marks the unconformity between the Good Harbor Bay andesites and the Lutsen basalts (see Fig. 5-15, stop 7). The conglomerate contains few if any clasts derived from the texturally distinct Lutsen basalts, but rather is composed predominantly of felsic-intermediate composition volcanic rocks, basalt unlike Lutsen basalt, and some reworked sandstone. Thus the source for the conglomerate is apparently dominated by older volcanic and sedimentary rocks that are similar in composition to those observed inland from the Schroeder-Lutsen sequence. The conglomerate may represent local deposition along a minor fault escarpment, or a proximal-facies high-energy fluvial deposit filling topographic depressions on the andesites.

---Drive northeast on Highway 61 approximately 13 miles to Historical Marker/Geologic Marker at top of steep hill (Terrace Point) at curve in road. Park at wayside, stop is across Highway at large roadcut. BE CAREFUL CROSSING HIGHWAY.---

Stop 5-11A. Cut Face Creek Roadcut

UTM coordinates (nad '83) 691780E, 5289170N

T.61N., R.1W., Sec. 34 Good Harbor Bay quadrangle



Highlights: Thick interflow sandstone, deformation features in sandstone at base of flow, shale rip-up chips, dessication cracks, Terrace Point basalt flow, fragmental basalt intruded and overrun by basalt flow.

Description: In high roadcut on north side of highway is an obvious contact between the Cut Face Creek Sandstone and the overlying Terrace Point basalt flow. The sandstone overlies the Good Harbor Bay andesites (stop 11b).

Sandstone –Jirsa (1984) measured approximately 73 meters of sandstone and 3 meters of shale in this section of the Cut Face Creek sandstone, but reported overall that nearly 30 percent of it is composed of thinly bedded, graded layers of fine-grained sand, silt, and clay. He

reports both symmetrical and asymmetrical ripple marks, and bimodal paleocurrent distribution, and concluded that the Cut Face Creek sandstone was deposited in a fluvial-lacustrine environment. Unlike nearly all the other exposed interflow sandstones within the NSVG, planar cross-bedding is predominant, but some trough cross beds are present near the top of the section. Other features that may be visible in

the sandstone are dessication cracks filled with sandstone or coarse pink zeolite minerals, and rip-up textures.

The base of the sandstone is exposed to the north in the Cut Face Creek valley, where it overlies the Good Harbor Bay andesite (see stop 11B).

Basalt – The Terrace Point basalt, which overlies the sandstone, is a major ridge-forming unit from here to the southwest, forming ‘sawtooth mountains’. It is a distinctive flow characterized by a dark green color, white thomsonite amygdules, uniform 3-4mm ophitic texture, and scattered small glassy plagioclase phenocrysts. In general the contact with the sandstone is sharp and straight, but in places the sandstone has been slightly deformed by the basalt flow.

At the south end of the outcrop is a unit of scoriaceous, fragmental basalt that is intruded and by the Terrace Point flow. Similar rock types and relationships have been observed to the southwest, also near the flow base. The breccia contains 1-100 cm angular basalt fragments that are both massive and amygdaloidal, and scattered large blocks of massive basalt. At this locality and others, sub-volcanic dikes of Terrace Point basalt that intrude the fragmental basalt are slightly chilled, and contain small amygdules stretched parallel to, and columnar joints perpendicular to, curvilinear dike margins. The fragmental unit is interpreted as a cinder cone or lahar-type deposit that was subsequently intruded and overrun by the thick Terrace Point basalt flow.

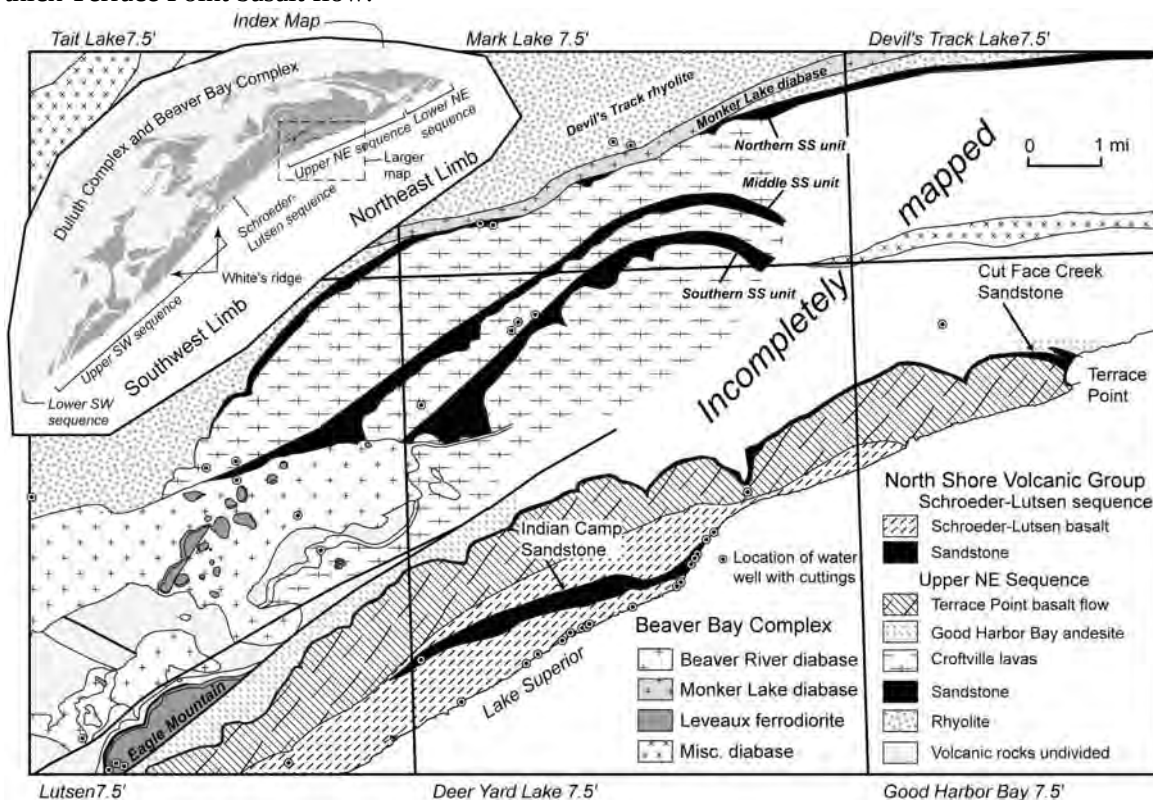


Figure 5-18. Simplified geologic map showing distribution of interflow sandstones and major volcanic units in the Lutsen to Terrace Point area.

Discussion: The Terrace Point basalt flow has been considered (e.g., Green, 1972) to be the basal flow of the Schroeder-Lutsen sequence. However, recent detailed mapping to the southwest (e.g. Boerboom and Green, 2006, Boerboom and others, 2006 and 2007) has shown that the Terrace Point basalt is actually contained within the Good Harbor Bay lavas. On the Onion River continuous exposures show that the top of the Terrace Point basalt is fragmental and overlain by more andesite (Stop 8), and is clearly an

integral part of the flow sequence. Also, the base of the Schroeder-Lutsen sequence overlies both the Terrace Point basalt and andesites in the Good Harbor Bay lavas (Figure 5-15, stop 7). Thus, based on new mapping, the Terrace Point flow and the Good Harbor Bay andesites are now considered to be part of a continuous series termed the Good Harbor Bay lavas.

The Cut Face Creek Sandstone was considered to represent clastic deposition during a hiatus in volcanism prior to eruption of the Terrace Point/Schroeder-Lutsen basalts. It was also considered to be somewhat unique in that it was one of the few thick sandstone units in the northeast limb of the North Shore Volcanic Group, along with the 68 meter thick Indian Camp sandstone (which is within the Schroeder-Lutsen basalts). However, recent remapping has shown that there are likely at least three more substantially thick sandstone units that occur within lower series of lava flows to the north of the Good Harbor Bay lavas (Fig. 5-18; Boerboom and others, 2007). Thus, it is now recognized that the Cutface Creek and Indian Camp sandstones, exposed because of more active erosion near the Lake Superior coast, are only one part of a larger set of thick sandstone units, some of which are located over 5 miles inland.

The recognition of thick interflow sandstones throughout several different series of lava flows at different stratigraphic heights implies active sedimentation during a prolonged period of volcanism. Fragmental flow tops in the Good Harbor Bay Lavas contain abundant sandstone infillings, and thin, discontinuous, layers and crack fillings of sandstone are common in the Schroeder-Lutsen basalts. Elevated levels of clastic deposition during active volcanism may account for the relative abundance of sand at the tops of the lava flows, and the thick interflow sandstones may have formed during periods of volcanic quiescence, but continued basin subsidence. Available data seem to indicate that the thicker sandstones may vary in thickness along strike, consistent with deposition onto an irregular lava surface.

---Drive northeast on Highway 61 approximately ¼ of a mile to the Cut Face Creek wayside located to the right on the lake shore, just past Cut Face Creek. Park at wayside. Either cross highway to small roadcuts across from wayside (WATCH TRAFFIC), or if the creek is low walk back on beach to Cut Face Creek and proceed upstream in creek valley.---

Stop 5-11B. Good Harbor Bay andesite

UTM coordinates (nad '83) 691953E, 5289526N

T.61N., R.1W., Sec. 34 Good Harbor Bay quadrangle

Highlights: Good Harbor Bay andesites, Cut Face Creek Sandstone



Description: Small outcrop directly across highway from parking area and to northeast, and outcrops at the mouth of Cut Face Creek, are fine-grained, brownish-gray, sparsely porphyritic andesite. These are part of the Good Harbor Bay andesites, a series of andesitic flows that stretches from near Grand Marais to Tofte. The andesite here is overlain by the Cut Face Creek Sandstone, and the contact is exposed in numerous locations in the meandering Cut Face Creek valley. The basal contact appears to be sub-conformable with the andesite on a local scale, but on a regional scale the upper andesite surface probably undulates. Small pebbles of massive to amygdaloidal andesite are common in the lower 3 meters of the sandstone, typically in 3-25 cm thick planar cross-bedded pebbly beds. The upper part of the andesite typically contains stretched amygdules, and has cracks filled with sandstone (Fig. 5-19). Like the exposures in the Onion River (Stop 9), the andesites here have rubbly, sand-filled flow top breccias with fragments of stretched amygdaloidal andesite.

Figure 5-19. Photograph of base of Cut Face Creek sandstone (white arrow), showing clastic dike (black arrow) filling a crack in underlying andesite. Hammer in circle is 45 cm long.

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53rd Annual Institute on Lake Superior Geology

FIELD TRIP 6

GEOLOGY ALONG THE GUNFLINT TRAIL

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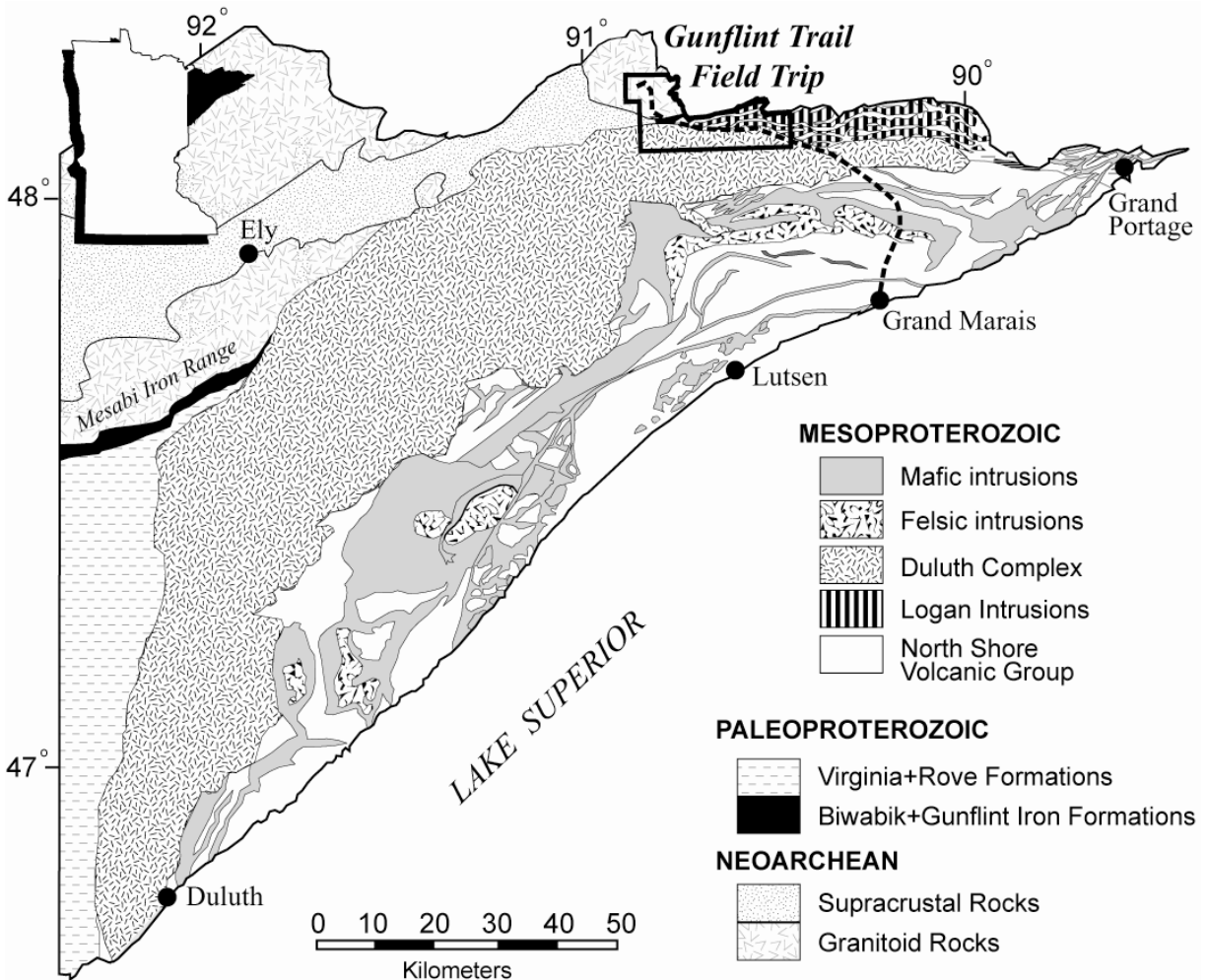


Figure 6.1. Regional geologic map of northeastern Minnesota showing the location of Gunflint Trail (modified from Jirsa and Miller, 2005). Inset box outlines Figure 6.4. Field trip area lies at the junction of Neoproterozoic, Paleoproterozoic, and Mesoproterozoic terranes.

INTRODUCTION

The Gunflint Trail provides access to the Boundary Waters Canoe Area Wilderness (BWCAW), which lies a short distance off either side of the road. The western end of the trail also transects a nearly complete geologic section of Archean, Paleoproterozoic, and Mesoproterozoic rocks (Fig. 6.1), and a diversity of well displayed contact relationships. In this area, the Archean greenstone-granite terrane of the Wawa subprovince of Superior Province is represented by a succession of metavolcanic rocks (ca 2700 Ma) intruded by the Saganaga Tonalite (ca 2689 Ma). The Archean rocks are unconformably overlain by Paleoproterozoic sedimentary strata of the Animikie Group (ca 1870-1830 Ma), which includes the Gunflint Iron Formation. Mesoproterozoic rifting is manifest in hypabyssal dikes and sills of the Logan intrusions (ca 1115 Ma), and several phases of the Duluth Complex (ca 1100 Ma), emplaced into both Archean and Paleoproterozoic rocks. Despite this exceptionally well exposed record of Precambrian history, little geologic mapping has been conducted in this region since the 1970's. As a result, this field guide relies heavily on prior mapping (Grout, 1929; Mathez and others, 1977; Morey and Nathan 1977; 1978; Morey and others 1981) and field trip guides (e.g., Weiblen and others, 1971; Miller

and others, 1987). Although the individual stops are different, much of this guide is modified only slightly from earlier publications, with permission from the authors.

Mapping in this region by the Minnesota Geological Survey has resumed, after a 30+ year hiatus. The new mapping will benefit from geological techniques and insights acquired during the intervening decades, particularly those resulting from the work of Canadian geologists. Furthermore, exposures in some areas are more accessible due to road building and recent forest fires. Regarding the latter, many parts of the BWCAW and adjacent lands experienced a megastorm in 1999 that resulted in large tracts of downed and tangled trees. Controlled burns and wild fires in some of the devastated areas—including the 2006 Cavity Lake fire—laid bare many outcrops that once were deeply concealed in forest, lichen, and moss.

GEOLOGIC SETTING

Neoproterozoic

The oldest rocks exposed in the region are Neoproterozoic metavolcanic and metasedimentary rocks—probably equivalent to, but not demonstrably continuous with, the Ely Greenstone. Although the supracrustal successions are dissected by faults and intrusions, some correlation can also be made with adjacent terrane in Ontario (Fig. 6.2). Most recent mapping in this terrane was conducted by Jirsa and Miller (2004) in the 1:100,000-scale, Ely-Basswood Lake map sheet that lies just west of the Gunflint Trail. A mapping thesis by Vervoort (1987) delineated geology in the JAP Lake area, which lies between that map sheet and the Gunflint Trail exposures. Rocks of the terrane are divided into a number of fault-bounded segments, each having distinct geologic characteristics that cannot be easily correlated from place to place. The Gunflint Trail area exposes the eastern end of what Gruner (1941) referred to as the Gabimichigami segment. Supracrustal rocks in this segment include variably pillowed, hypabyssal mafic metavolcanic flows (**Stop 6-1**), and andesitic pyroclastic rocks. The supracrustal rocks are cut by hornblende-phyric intrusions that are similar in composition to the associated andesite, and on this basis, inferred to be more or less contemporaneous with them (Fein, 1977). Based on stratigraphic facing directions established from pillowed metabasalt flows, the supracrustal sequence forms an east-trending and steeply south dipping and younging homocline. The Saganaga Tonalite was emplaced along the northern edge of the supracrustal succession. The metamorphic grade of greenstone along this contact is locally increased from greenschist facies that is typical of much of the belt, to amphibolite grade, and foliation is well-developed. The western edge of the Gabimichigami segment is terminated by a north-northeast-trending fault that juxtaposed greenstone against sedimentary and volcanic rocks of the Knife Lake Group. The Knife Lake Group includes the Ogishkemuncie conglomerate that contains detrital clasts of the Saganaga Tonalite. This distinctive sequence of conglomerate, sandstone, and alkalic rocks is interpreted to have been deposited in a complex array of successor basins developed along early-formed faults (Jirsa and Miller, 2005) at some time after emplacement of the Saganaga Tonalite.

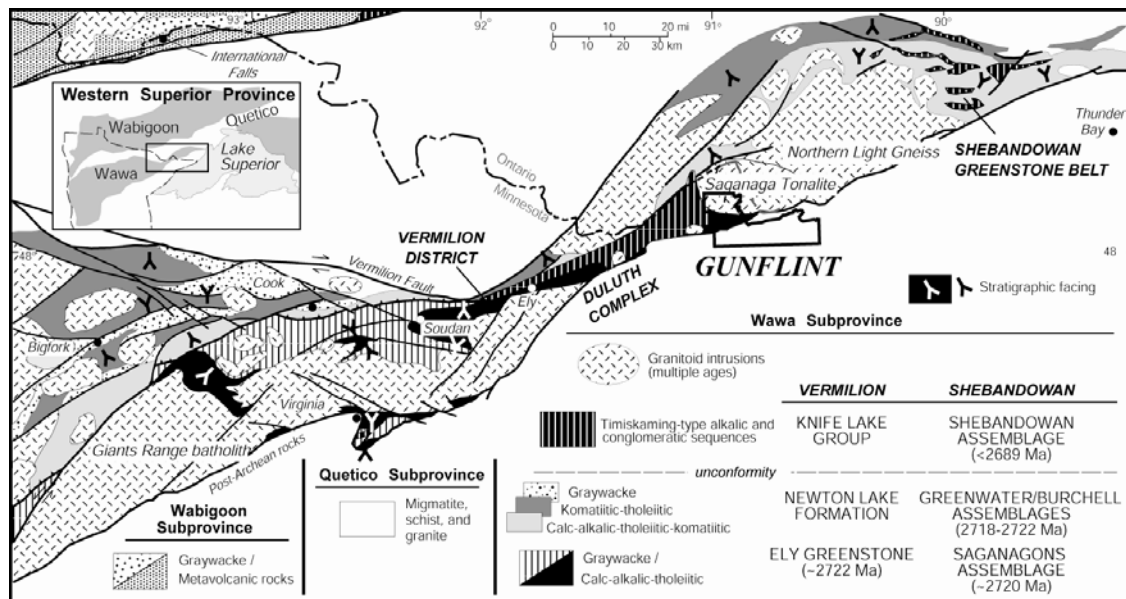


Figure 6.2. Geologic setting of Archean bedrock in northeastern Minnesota and adjacent Ontario (modified from Jirsa and Miller, 2005). Inset map shows location relative to subprovinces of the Superior Province.

Lacking detailed geochronologic data for this immediate area, much of the temporal distinction between various geological elements of the Neoproterozoic bedrock is based on the correlation of fabrics that resulted from three major phases of deformation, denoted D_1 , D_2 , and D_3 , with U-Pb zircon dates acquired elsewhere. All three deformation events are the result of N-S- to NW-SE-directed compression. The timing of D_1 deformation is bracketed between deposition of the volcanic and clastic rocks of the Wawa subprovince at about 2722 Ma (Peterson and others, 2001), and emplacement of the Saganaga Tonalite at about 2689 Ma (Corfu and Stott, 1998). Folds in the Ely Greenstone and related rocks attributed to D_1 deformation are truncated by faults associated with Knife Lake strata, indicating that the latter is synchronous with or post-dates deposition and early deformation of the Ely. As such, the Knife Lake Group is inferred to be a Timiskaming-type sequence temporally equivalent to the Shebandowan assemblage exposed in adjacent parts of Ontario (Corfu and Stott, 1998). D_2 deformation effected all of the Archean supracrustal units and can be crudely bracketed by U-Pb dates of intrusions in the Giants Range batholith to the southwest that place the regional deformation and metamorphic event between about 2674 Ma and 2685 Ma (Boerboom and Zartman, 1993). D_3 deformation produced faults in the low grade supracrustal and intrusive rocks of the Wawa subprovince, and folds and metamorphic fabric in granitic and migmatitic rocks of the Quetico subprovince to the north.

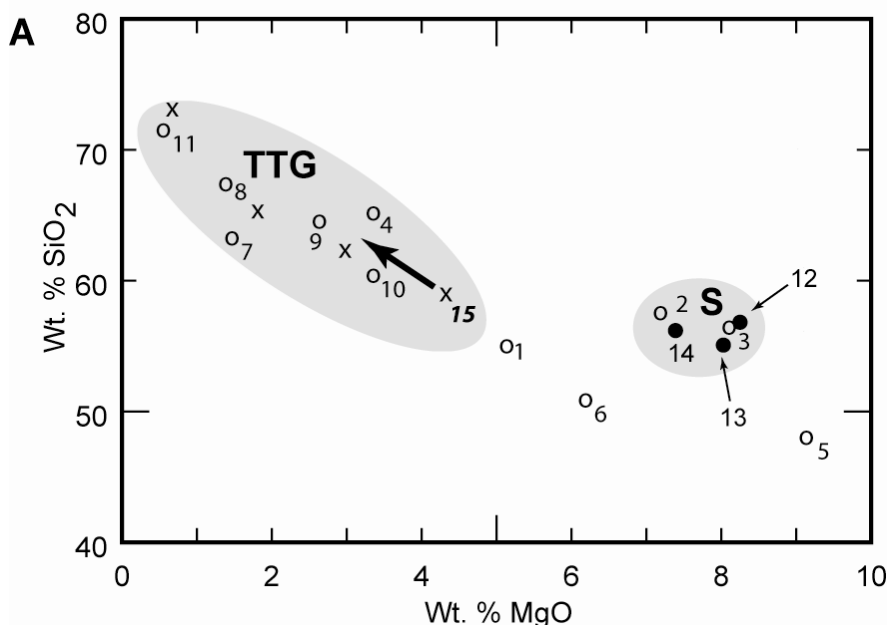
The batholith generally referred to as the Saganaga Tonalite was emplaced into metavolcanic and gneissic granitoid country rocks. Igneous fabric is best developed in the border zones; but overall, the intrusion is moderately foliated and has a shallow eastward lineation. U-Pb age dates include 2689 ± 1 Ma for the tonalite; whereas, 2 phases of the Northern Light Gneiss, which lies in contact with the Saganaga along its northeastern boundary in Ontario, are dated at 2707 ± 2 Ma and 2750 ± 2 Ma (Corfu and Stott, 1998). On the basis of geochemistry, Vervoort (1987) infers that the Saganaga batholith represents the magma source for dacitic lavas that overlie mafic and ultramafic rocks in the JAP Lake area.

The Saganaga batholith was divided by Grout (1929) into a medium- to coarse-grained tonalitic phase that makes up the majority of the batholith (**Stops 6-2 and 6-3**), and a more mafic border phase (**Stops 6-4 and 6-5**). Grout (1929, p.568) defined this border zone approximately one mile in width along the south, east, and south sides of the batholith. In the Gunflint Trail area, the border phase is more irregular in

distribution and consists of strongly foliated, medium to coarse-grained, quartz-bearing hornblende diorite to granodiorite that generally lacks quartz eyes.

The border phase is also marked by a wide variety of inclusions. Grout (1929) suggested that some of the inclusions in the Saganaga batholith are the products of igneous processes that ranged from assimilation to the mixing of mafic and felsic magmas. Alternatively, Hanson (1972) suggested that some of the included mafic material could be a product of assimilation of greenstone country rocks during pluton emplacement. Field evidence showing the presence of nearly all gradations of inclusions between mafic and felsic magmatic-looking components, and the presence of quartz phenocrysts in some mafic inclusions, supports the former interpretation for many of the inclusions. The results of analyses of two mafic inclusions acquired recently by the authors are summarized in Figures 6.3A, B, and C. These data indicate that the mafic inclusions (Fig. 6.3D) have the common chemical characteristics of the sanukitoid suite of Archean granitoid rocks. Thus, we infer that the inclusions studied are autoliths derived from a separate, primitive sanukitoid-magma. These data support Grout's earlier interpretations. This is not to say that country rock inclusions are not present—some fine-grained, dark-green, and variably cleaved inclusions do occur very near the contact with enclosing metavolcanic rocks (**Stop 6-6a**). To date there has been no systematic study of the mafic components of the Saganaga batholith. As pointed out by Didier and Barbarin (1991) in the conclusion of their multi-authored volume, further study of enclaves will contribute to the advancement of the overall petrogenesis of granites. The Saganaga batholith provides ample opportunity for such studies.

Figure 6.3. Geochemical comparison of Saganaga batholith, mafic inclusions, and related rocks.

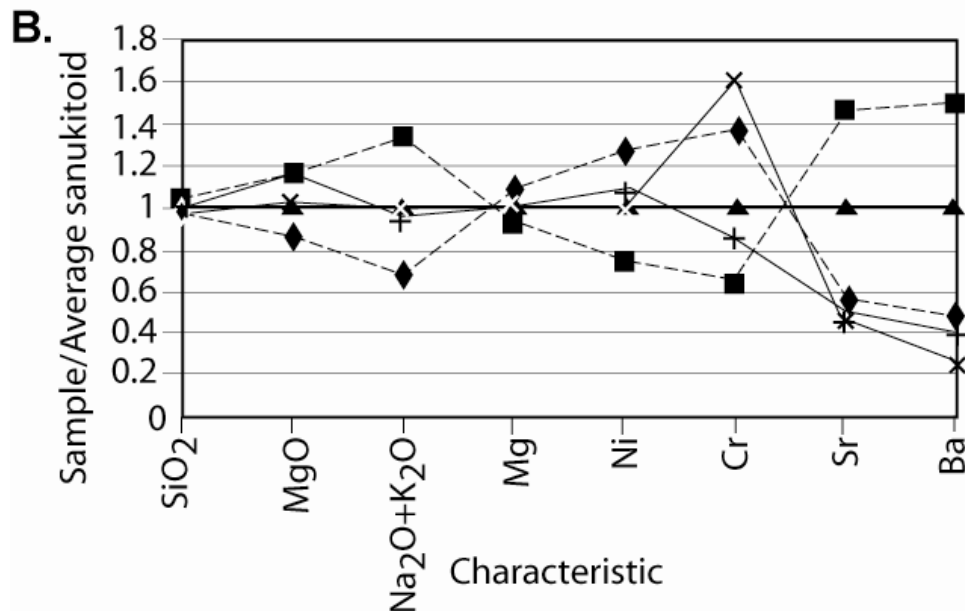


A. SiO₂-MgO plot. Abbreviations and data sources: **TTG**=Compositional field of the Archean Tonalite-Trondhjemite-Granodiorite suite; **S**=Compositional field of primitive Archean sanukitoid samples from the southwestern Superior Province (Stern and others, 1989).

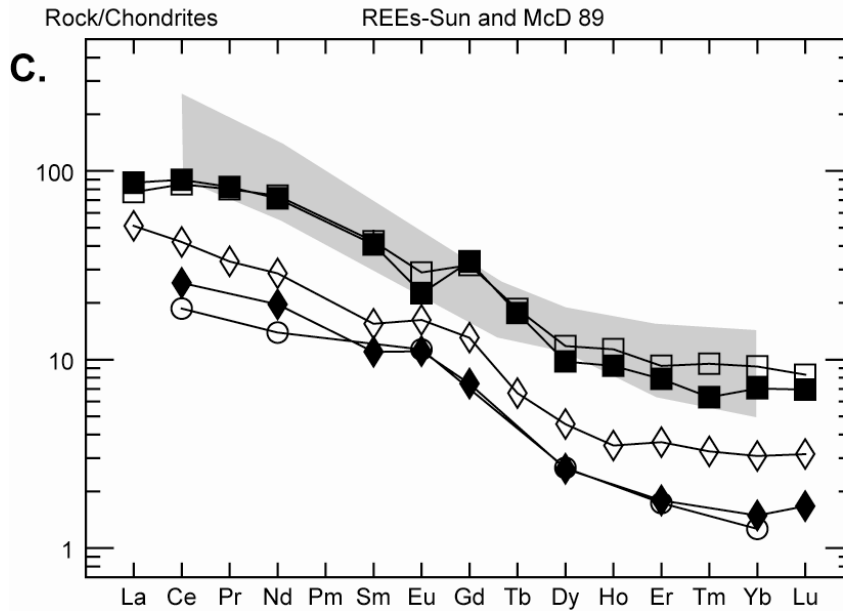
Open circles: 1 Ely greenstone; 2 and 3 mafic inclusions at Stop 6-5; 4 border phase of the Saganaga Tonalite at Stop 6-5; 5, 6 Ely greenstone; 7-10 tonalite; 11 granodiorite. Analyses 1-4 were acquired for this study (on-file with Minnesota Geological Survey); analyses 5-11 are from Arth and Hanson, 1975, Table 2, p. 331, analyses 1-2, 19-22, and 24 respectively. Analyses 12-14 are from Stern and others, 1989, Table 1 p. 1695, samples 1-3.

Crosses: analyses of the Roaring River Complex, Superior Province, Canada from Stern and Hanson (1991, Table 2, p. 220). They have demonstrated that incremental equilibrium crystallization of a melt

with the composition of analysis 15 (sample #7, op cit) can produce the three members of the TTG suite indicated. The two analyses of mafic inclusions at Stop 6-5 (#s 2 and 3) plot in the primitive sanukitoid field, and their chemistry is similar to other chemical characteristics of the sanukitoid suite (Figs. 6.3B and C). These data imply that the mafic inclusions must either be deformed xenoliths of a sanukitoid body or segregations from a consanguineous melt.

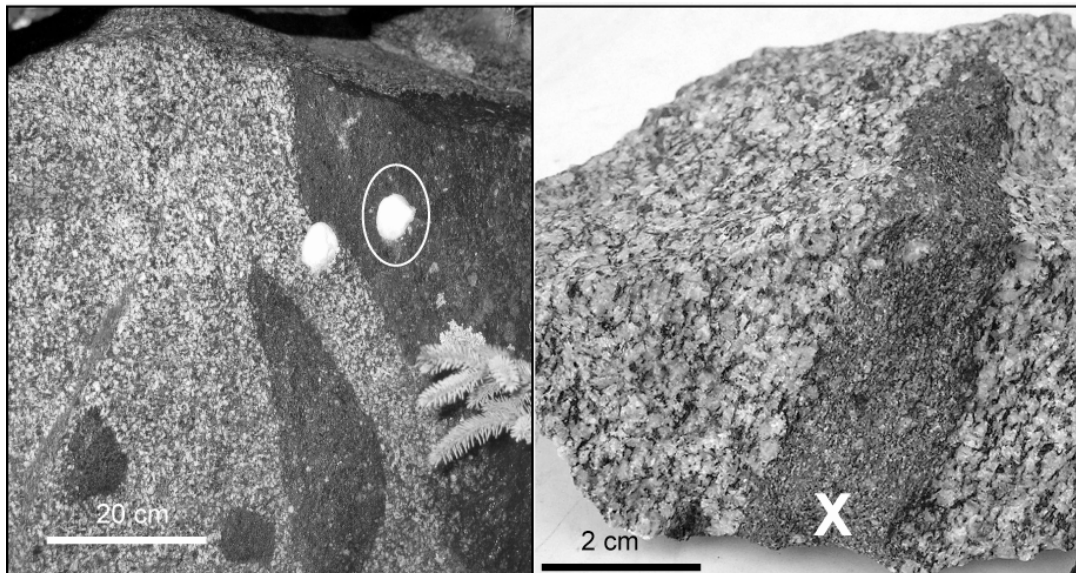


B. Chemical characteristics of selected sanukitoid suite rocks (data from Stern and others, 1989) and mafic enclaves in the Saganaga batholith (data from this study). **Solid squares, dashed line** = range of high-silica sanukitoid; **Solid diamonds, dashed line** = range of low-silica sanukitoid; **Solid triangles** = average of high- and low-silica sanukitoid; **Pluses** = high-silica mafic inclusion (#2 from Fig. 6.3A), in the Saganaga batholith; **Crosses** = low-silica mafic inclusion (#3) in the Saganaga batholith. The analyses are normalized to an average of high and low values for *each chemical characteristic* of several sanukitoid analyses. Data for the mafic inclusions lie well within the range of the sanukitoid values, except for Cr in mafic inclusion #3, which plots just above the high range of the sanukitoid rocks, and Ba and Sr for both mafic inclusions that plot slightly lower than the low range of the sanukitoid rocks.



C. REE analyses.

Open squares = Saganaga mafic enclave from this study (#2 from Fig. 6.3A), **Solid squares** = Saganaga mafic enclave from this study (#3), **Open diamonds** = Saganaga tonalite this study (#4), **Solid diamonds** = Saganaga tonalite of Arth and Hanson, 1975, sample 19, **Open circles** = Saganaga tonalite of Arth and Hanson, 1975, sample 20. The REE data for the mafic enclaves in the Saganaga batholith are distinct from that of the tonalite samples, and lie roughly within the sanukitoid field (shaded) as defined by Stern and others (1989, Fig. 10, p.1702). These data support the conclusion from the chemical characteristics (Fig.6.3B), and the $\text{SiO}_2\text{-MgO}$ (Fig. 6.3A).



D. Photographs of mafic inclusions analyzed for this study. Both inclusions occur in the strongly foliated, granodioritic border phase of Saganaga batholith at Stop 6-5. LEFT=inclusion #2, material for analyses shown on Fig. 6.3A-C was taken from circled (snow-filled) drill hole. RIGHT=Sample 3: material for analyses taken from bottom of sample below X.

Paleoproterozoic

Animikie Group

Sedimentary rocks of the Paleoproterozoic Animikie Group are exposed in an east-trending belt that extends from Thunder Bay on Lake Superior to a point 12 miles (19km) west of Gunflint Trail, where the belt is truncated by the Mesoproterozoic Duluth Complex (Fig. 6.4). The Animikie Group in this area consists of basal conglomerate and iron-bearing strata of the Gunflint Iron Formation, overlain by slate and siltstone of the Rove Formation. The sequence is broadly correlative with Animikie strata exposed to the southwest along the Mesabi Iron Range. The rocks form a homocline that dips gently southward, except where deformed by folding, faulting, and emplacement of Mesoproterozoic intrusions. For example, local folding, inferred to be associated with emplacement of the Logan intrusions, may be the product of magmatic shouldering into the relatively ductile sedimentary country rocks. Sequence "inflation" by the Logan sills may explain the observation that the dip of Paleoproterozoic rocks increases from 10 degrees away from the contact with Duluth Complex, to 60 degrees locally near it. Faults are present locally, but few have displacements greater than 50 feet. A notable exception is the block-shaped Lookout fault (Fig. 6.5) that lies adjacent to the Gunflint trail. As much as 200 feet of uplift on the west and south is speculated (Morey and others, 1981). In addition, the dip of Animikie strata on the west side of the fault is much steeper than that on the east, which explains in part the difference in the widths of map units apparent from the geologic map. Much of the complex-looking fold pattern east of the fault is likely an artifact of complicated surface topography and shallowly dipping units. The Lookout fault displaces both Animikie rocks and the Logan intrusions, but the timing of offset relative to emplacement of the Duluth Complex is unclear.

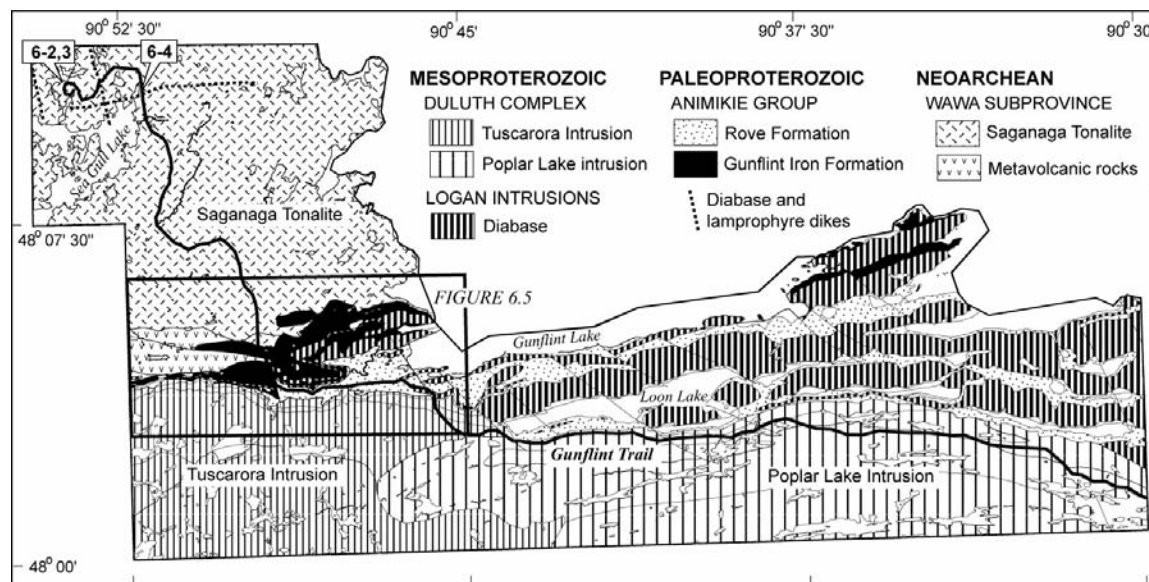


Figure 6.4. Geologic map along the northwestern end of the Gunflint Trail, modified from published quadrangle maps (Morey and Nathan, 1977; 1978; and Morey and others, 1981), and mapping by Grout (1929). The locations of stops 6-2, 6-3, and 6-4 are shown—other field trip stops lie within the outline corresponding to Figure 6.5.

Across the border in Canada, the Gunflint Iron Formation is essentially unmetamorphosed and consists of quartz, hematite, iron carbonates, greenalite, and magnetite. A thorough review and sedimentological interpretation of these rocks can be found in Pufahl and Fralick (2000), Fralick and others (2002), and references therein. In Minnesota, the original mineralogic character of the iron-formation was modified by contact metamorphism adjacent to Mesoproterozoic intrusions, and many of the fine-scale textural and mineralogical attributes used to subdivide the iron-formation in Canada have

been destroyed. Consequently, a four-fold nomenclatural scheme that emphasizes various bedding aspects that are readily visible has been used for the Gunflint since the early work of Broderick (1920). The scheme was originally devised for the Biwabik Iron Formation on the Mesabi Iron Range the Mesabi Iron Range. It defines four informal members—Lower cherty, Lower slaty, Upper cherty, and Upper slaty. These terms emphasize the relative proportions of strata dominated by thickly bedded and granular (cherty) beds consisting of rounded sand-size intraclasts, vs. thinly bedded and very fine-grained (slaty) beds. The Lower and Upper members are grouped on the published geologic map that shows iron-formation (Morey and others, 1981) and on Figure 6.5.

The Lower cherty member ranges in thickness from 15 to 45 ft (4.5-13.5m). A conglomerate containing clasts of Archean granite and greenstone set in a matrix of feldspathic quartzite or stromatolite-bearing chert occurs locally (**Stop 6-6b**). The basal conglomerate is absent in most of the area, and a persistent thick-bedded to massive, magnetite-rich unit 5-15 ft (1.5-4.5m) thick (**Stop 6-6c**) composes the base of the Lower cherty member and unconformably overlies Archean rocks. The magnetite-rich layer is in turn overlain by a thick-bedded, chert-rich, magnetite-poor unit about 15 ft (4.5m) thick.

The overlying Lower slaty member is 80-95 ft (24-29m) thick. The lowermost 10 ft (3m) is a black, thinly bedded, nearly magnetite-free argillite composed predominantly of volcanically derived material. This may equate with the Intermediate slate unit known on the Mesabi range. Massive and cherty beds that resemble the upper part of the Lower cherty member occur above the argillite, but they pass abruptly upward into a sequence of thick, chert- and silicate-rich beds with sparse magnetite intercalated with intervals of thinly laminated silicate-rich beds. The uppermost 50 ft (15m) of the Lower slaty member consists of thinly bedded to laminated slate containing various silicates and 20-35 percent magnetite (**Stop 6-7a**).

The Upper cherty member is approximately 50 ft (15m) thick. The gradational contact between the Lower slaty and the Upper cherty members is marked by the appearance of irregularly bedded to lenticular chert-rich layers and by thin irregular layers consisting almost entirely of magnetite. The upper part of the Upper cherty member is characterized by several thick units of granular chert that contain microbial structures, intraformational conglomerate, and abundant magnetite. (**Stop 6-7c**).

The Upper slaty member is approximately 150 ft (45m) thick. Thick lenticular beds of chert containing disseminated magnetite occur in the lower few meters, but most of the member consists of thin-bedded to laminated, silicate-rich units that are intercalated with intervals of thinly laminated graphitic argillite and centimeter-thick beds of relatively pure chert. The upper 10 ft (3m) are nearly magnetite-free and consist of limestone and chert interbedded with argillite (**Stops 6-8a, 6-8b**).

The Rove Formation gradationally overlies the Gunflint Iron Formation and is at least 3,200 ft (970m) thick, although estimates of thickness are complicated by multiple Mesoproterozoic intrusions and likely fault repetition. It can be divided into a lower argillaceous unit, a transitional sequence, and an upper thin-bedded graywacke-rich unit (Morey, 1969). The lower argillite consists of intercalated argillaceous siltstone, silty argillite, and carbonaceous argillite (**Stop 6-9**). The transitional sequence separates dominantly argillaceous rocks below from an interval containing intercalated beds of coarse- and fine-grained graywacke. The sandy beds contain many primary sedimentary structures indicative of deposition by turbidity currents that flowed dominantly southward.

Contact metamorphic effects on the Gunflint Iron Formation were studied in detail by Floran and Papike (1975, 1978), and metamorphism of the Rove Formation was described by Labotka and others (1981). The sizes of the metamorphic aureoles adjacent to the Logan intrusions (as much as 33 ft (10m) wide) and their mineral assemblages are directly related to the thicknesses of the sills. Adjacent to the Duluth Complex, three metamorphic zones have been distinguished:

- 1) An outer zone of slightly metamorphosed iron-formation consisting of quartz, iron carbonate, minnesotaite, stilpnomelane, and hematite partially replaced by magnetite.
- 2) A 1.2 mile (2km) wide intermediate zone of moderately metamorphosed iron-formation containing grunerite-cummingtonite, hornblende, and actinolite, as well as quartz and fine-grained magnetite.

- 3) A proximal zone of highly metamorphosed iron-formation occurring within 0.3 miles (0.5km) of the contact with the Duluth Complex and composed of quartz, magnetite, iron-rich pyroxenes, and fayalite.

By contrast with the iron-formation, contact-related recrystallization of the Rove Formation is much less pronounced, even very near the Duluth Complex. The aureole is marked by a complex mixture of rock types suggestive of partial melting within a meter or so of the contact. In this contact zone, argillaceous rocks retain large-scale layering, but have a vague granoblastic texture. Individual layers contain cordierite and hypersthene locally, with minor biotite, ilmenite, plagioclase, K-feldspar, and olivine. At varying distances away from the contact, but generally no more than 100 ft (30m) away, biotite is the only well-developed metamorphic mineral in the pelitic rocks. Calcareous beds near the contact contain minor grossular garnet.

A U-Pb age of 1878 ± 1 for the Gunflint Iron Formation was determined by Fralick and others (2002) from reworked volcanoclastic zircons in Ontario exposures. Subsequent geochronologic study of the Rove and equivalent Virginia Formations has considerably complicated and protracted the timeline of Animikie Group deposition. For example, dates by Addison and others (2005) of detrital zircons near the base of the Rove Formation yielded ages of 1827 ± 8 and 1836 ± 5 Ma, and 1832 ± 3 from a similar stratigraphic position in the Virginia Formation. All of these samples were acquired from several meters above what Addison and others infer is an ejecta layer derived from the Sudbury impact event (ca 1850). Recently acquired detrital zircon dates by Heaman and Easton (2005) indicate that upper parts of the Rove Formation are as young as 1777 Ma. Wirth and others (2006) report ca 1790 Ma detrital zircons in the Rove and equivalent Thomson Formations. Collectively, these data imply that deposition of strata assigned to the Animikie Group spanned some 100 million years, and hints at the likelihood of several major unconformities.

The tectonic and depositional setting of the Gunflint and other semi-contemporaneous Lake Superior iron-formations remains somewhat contentious (summarized for all iron-formations in Clout and Simonson, 2005). Modern interpretations include deposition in one or more foredeep basins developed cratonward of crustal loading during Penokean orogenesis (Morey and Southwick, 1995). More recently, Pufahl and Fralick (2004) have inferred from sedimentological evidence that deposition occurred along a south-sloping continental margin undergoing subsidence in a back-arc extensional setting. Geochemical evidence for within-plate signatures of volcanic rocks associated with iron-formation in Michigan suggested to Schulz and Cannon (in press) a similar interpretation of back-arc spreading. From this and geochronologic work, they infer that deposition occurred in a broad foreland having local extensional basins hosting iron-formation. Following iron-formation deposition, turbidites of the Rove Formation and equivalent strata were deposited in a rapidly subsiding basin, with subsidence presumably driven by crustal loading by the northward migrating fold and thrust belt. A hiatus boundary between iron-formation and overlying turbidites identified in Ontario and Minnesota (Addison and others, 2005), and Michigan (Schulz and Cannon, in press) roughly coincides with the ca 1850 Sudbury impact event.

Mafic dikes

Mafic dikes emplaced into the Saganaga Tonalite are prominent on aeromagnetic maps as two sets of positive linear anomalies trending northward and eastward. Exposures mapped along the north-trending magnetic trajectories (Fig. 6.4) show that the western of the two is diabasic, and the eastern consists of lamprophyre. The latter was the subject of work by O'Brien (1982) who mapped the 70 foot thick dike, reprocessed aeromagnetic data, and conducted petrologic work. He classifies the lamprophyre dike as camptonite, defined by its content of kaersutite (calcic amphibole), titanbiotite, and titanaugite phenocrysts in a plagioclase-rich groundmass. The precise age of these dikes is unclear, but a Paleoproterozoic age is likely. K-Ar dates of biotite by Goldich and others (1961) yielded an age of about 1750 Ma, which they attributed to late thermal resetting during alteration of the dikes. Petrographic and geochemical analyses by O'Brien indicate that much of the alteration was deuteric—having formed during cooling of the dike. On this basis, O'Brien speculated that the K-Ar date may be an accurate estimate for emplacement. A second set of aeromagnetic anomalies—trending east and slightly less pronounced than

the north-trending ones—have not been investigated in detail. The roadcut at **Stop 6-4** exposes diabase inferred to lie along one of the anomalies.

Mesoproterozoic

Mesoproterozoic mafic intrusions comprise the remaining exposures in the Gunflint Trail area (Figures 6.4 and 6.5). The rocks represent the early magmatic stages of the Midcontinent Rift system. The apparently earliest of these are diabasic sills and dikes emplaced into the Animikie strata and collectively referred to as the Logan intrusions. A baddeleyite age of 1115 ± 1 Ma is reported from a sample of a Logan sill near Thunder Bay (Heaman and Easton, 2005). The sills are intruded, with slight angular discordance, by medium to coarse-grained gabbro and troctolite of the Poplar Lake and Tuscarora intrusions of the Duluth Complex. The Poplar Lake is part of the early gabbroic series of the complex. A basal gabbroic unit of the intrusion yielded a date of 1106.9 ± 0.8 (Miller and Severson, 2002). Field relationships indicate that the Poplar Lake is intruded by the Tuscarora intrusion, which is considered to be part of the layered series of Duluth Complex.

Logan Intrusions

The Logan intrusions are exposed along a series of prominent, east-trending ridges formed by the differential erosion of diabase sills and sedimentary rocks, particularly the Rove Formation. Individual sills are as much as 1100 ft (33m) thick, and can be traced along strike for several kilometers. Branching and merging of individual sills is common, and many sills thicken and thin down-dip. Some sills terminate against joints and inferred faults. Locally, fractures in the Rove Formation are occupied by thin dikes, which give a box-work configuration to the hypabyssal intrusions. Rock types include aphyric basalt, fine- to medium-grained diabase with ophitic clinopyroxene enclosing plagioclase, plagioclase cumulates, and granophyre (Jones, 1984). Chilled margins form sharp contacts with, and locally contain inclusions of, the country rocks (**Stop 6-10**). Diabase coarsens to medium-grained near the center of individual sills, and clinopyroxene is ophitic throughout. Minor differentiation is manifest in accumulations of plagioclase or granophyric intergrowths (quartz, sodic plagioclase, and orthoclase) in upper parts of sills. The configuration of Logan intrusions closely follows that of the gently folded Animikie Group strata, implying that minor ductile deformation occurred during the Mesoproterozoic.

Duluth Complex

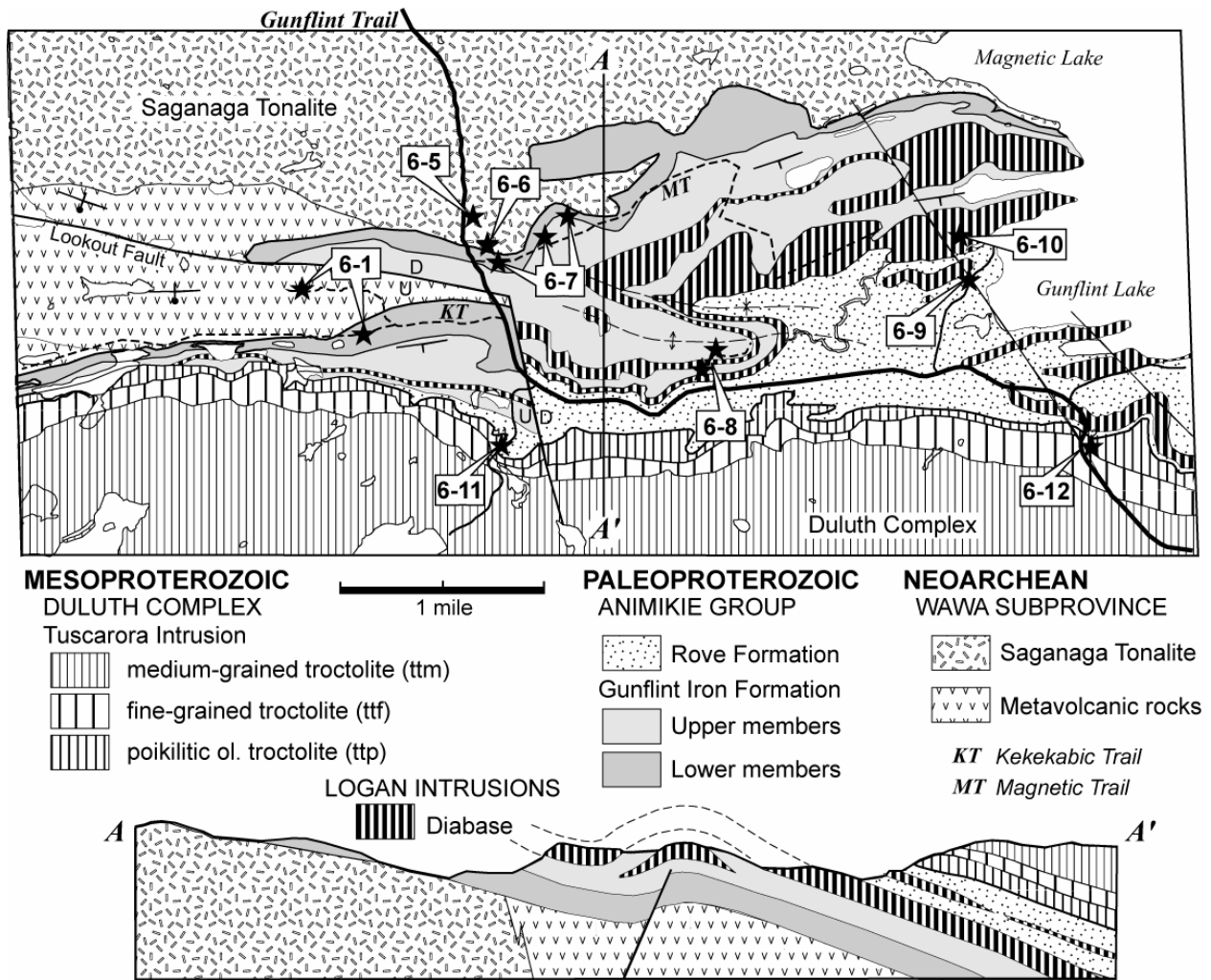
The Duluth Complex is a sequence of generally discordant plutonic rocks consisting of many separate intrusions. Two of these intrusions, the Poplar Lake and Tuscarora, are exposed along the Gunflint Trail. The Poplar Lake intrusion, formerly referred to as Nathan's layered series, consists of interlayered gabbroic cumulates, with minor amounts of troctolitic and anorthositic cumulates. Rocks of the Poplar Lake have reversed magnetic polarity, and thus are broadly correlative with lower lavas of the North Shore Volcanic Group and with the Logan intrusions. As originally defined by Nathan (1969), the Poplar Lake intrusion is composed of at least 27 sheet-like units of mafic cumulates and intermediate to felsic rocks. More detail about the Poplar Lake intrusion can be found in this volume (Miller and Jerde, 2007; Field Trip #1). The Tuscarora intrusion irregularly cuts across the layered gabbro of the Poplar Lake intrusion (Morey and Nathan, 1978). The basal part of the Tuscarora intrusion consists of a fine-grained, augite-poikilitic, olivine gabbro (unit *ttf* on Fig. 6.5) (**Stops 6-11 and 6-12**). Within 0.3 mi (0.5km) of the basal contact, fine-grained troctolite coarsens to medium-grained (unit *ttm*). The troctolite units consist of 65-70 percent plagioclase and 10-15 percent olivine. Relative amounts of poikilitic augite and iron-titanium oxides vary locally. Orthopyroxene mantles olivine and occurs in symplectic intergrowth with plagioclase. Biotite is locally present in association with iron-titanium oxides. Modal layering is well developed and generally concordant with unit boundaries that dip gently to the south—typically more shallowly dipping than the subjacent Animikie Group strata. The *ttp* and *ttf* units commonly contain chalcopyrite, pyrrhotite, and minor pentlandite interstitial to plagioclase and olivine. The sulfide concentrations are subeconomic, but locally form mappable zones (**Stop 6-12**).

FIELD TRIP 6

Stop Descriptions

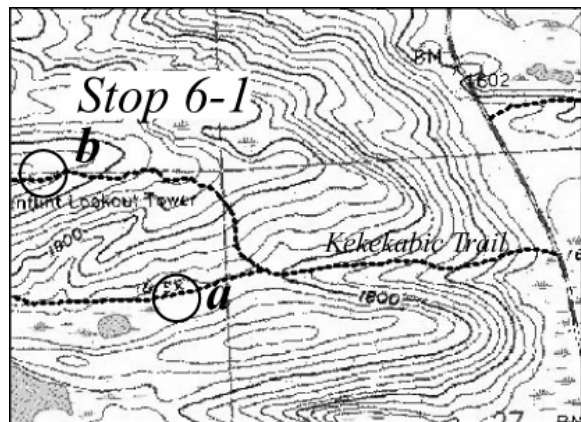
NOTES:

- 1) This group will not visit all of the stops described below. A number of stop descriptions are included in this guide to provide context and for future visits to the region.
- 2) Small maps showing location of individual stops are modified from USGS 7.5-minute (1:24,000-scale) topographic maps, which are named below.
- 3) All UTM coordinates are given in NAD 83.



STOP 6-1

Neoarchean pillowed basalt and basal Gunflint Iron Formation



Location: T. 65 N., R. 4 W., sec. 28, NW, NE;
Kekekabic Trail to Lookout (Fig. 6.5)

Long Island Lake quadrangle

6-1a = So-called "Paulson Mine" at UTM
660,504E/5,328,673N (NAD83)

6-1b = Lookout at UTM 660,857E/5,328,364N

[Requires 2 mile hike]

Description:

Follow the Kekekabic hiking trail from the parking lot, westward approximately 0.6 mi. to the junction with a north-bound trail that leads to the lookout. The main Kekekabic trail parallels the base of the Gunflint Iron Formation, and the north-facing slope immediately south of the trail contains exposures of the Lower slaty member. The iron-formation has been strongly metamorphosed in this area and now consists of various assemblages of quartz-grunerite-fayalite-magnetite and quartz-cummingtonite-grunerite-pyroxene-magnetite.

6-1a. Continuing westward about 800 ft from the trail junction, several test pits can be seen, including one that is fenced and labeled "Paulson Mine 1893," but probably isn't (the mine apparently lies some distance farther west). This and other test pits are developed in the Lower cherty member of iron-formation, and waste piles contain abundant pyrrhotite, other sulfide minerals, and magnetite.

6-1b. Returning to the junction with the north-bound trail and following it up-slope, there are numerous exposures of Neoarchean metabasalt. Several small outliers of iron-formation containing iron-silicates and magnetite can be found along the route, implying that this south-facing surface is very near the unconformable contact between Archean and Paleoproterozoic rocks. About **120** feet north of the junction, a ledge of fine-grained diabase crosses the trail. Preliminary analyses (this study) indicate that it has a komatiitic composition. Intrusions of this composition are also found in the Newton Lake Formation, some 30 miles to the southwest, and high-Mg tholeiitic basalt flows and pyroxenitic to peridotitic sills were described by Vervoort (1987) in the JAP Lake area two miles to the west. At and near the lookout, metabasalt flows with pillow structures and autobreccia fabrics are exposed. Judged from pillows—which have been moderately flattened by regional D₂ deformation—bedding trends east-northeastward, is steeply south dipping and stratigraphic facing is southward.

STOP 6-2

Felsic phase of the Neoproterozoic Saganaga Tonalite cut by a small mafic dike.



Location: T. 66 N., R. 4 W., sec. 31, NW, SW;
Campsite #18 (Fig. 6.4)
Munker Island quadrangle
UTM: 656,333E/5,335,730N

Description: Most of the outcrops in this area consist of gray, massive to trachytoid-foliated, medium to coarse-grained tonalite, having plagioclase in much greater abundance than microcline (Fig. 6.6). Large quartz phenocrysts, or eyes, as much as 1 cm in diameter are characteristic of this phase, which is typical of 90% of the batholith. Quartz eyes are polycrystalline aggregates, in which each crystal has a different optical orientation. Quartz also occurs as an interstitial mineral to subhedral plagioclase (An₂₀₋₂₈). Small amounts of microcline occur as antipathetic exsolution in plagioclase, as rims on plagioclase, and as small interstitial grains. Hornblende is the dominant ferromagnesian mineral, together with minor amounts of augite, biotite, epidote, and chlorite.



Figure 6.6. Texture typical of the Saganaga Tonalite, including lineated quartz "eyes" (medium gray). The small dike of aphanitic mafic rock in this exposure has not been sampled, but is inferred to be related to larger north-trending diabasic and lamprophyric dikes that form prominent north-trending anomalies on aeromagnetic maps.

The pronounced foliation in the Saganaga Tonalite was inferred by Grout (1933) as a primary flow fabric (trachytoid). The tonalite is inferred to have been emplaced into Archean metavolcanic rocks shortly after early (D₁) deformation, based on a U-Pb date of 2689±1 in Canadian exposures (Corfu and Stott, 1998). As such, it experienced major regional metamorphism and transposition associated with D₂ deformation. As with many large plutons in such terranes, the debate remains unresolved about whether the fabrics are wholly magmatic, wholly tectonic, or some hybrid of the two.

STOP 6-3

Neoarchean Saganaga Tonalite with rounded dioritic to granodioritic inclusions



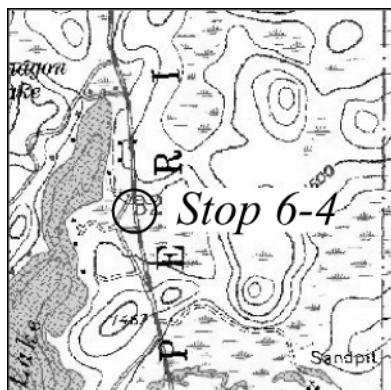
Location: T. 66 N., R. 4 W., sec. 31, SW, NW;
Campsite #13 (Fig. 6.4)
Munker Island quadrangle
UTM: 656,219E/5,336,079N

Description:

Equigranular tonalite with characteristic quartz eyes, containing a wide variety of inclusions. The term "inclusion" has a tortured usage—we prefer to use the term to apply to material that has a contrasting composition or appearance from its host, regardless of origin, size, shape, degree of assimilation, or extent of equilibration with the enclosing host magma. Inclusions may represent xenolithic blocks of country rock incorporated into the Saganaga Tonalite, or cognate phases of the intrusion (autoliths). The inclusions found here have not been studied in detail. They present a unique opportunity for further petrologic and mineral chemistry research to explore their complex origin. The shape of inclusions at this stop is related to either magmatic or tectonic processes. In either case, one must keep in mind that 2-dimensional shapes are dependent on the orientation of exposure-surface relative to that of the inclusions. This exposure, for example, provides only one surface—it is impossible to say from this whether the inclusions are discoid, sub-spherical, or egg-shaped.

STOP 6-4

Granodioritic phase of Neoarchean Saganaga Tonalite with inclusions; cut by diabase dike



Location: T. 66 N., R. 4 W., sec. 32, near center;
Roadcut on both sides of Gunflint Trail (Fig. 6.4)
Conners quadrangle
UTM: 658,575E/5,335,837N

Description:

Exposures on the west side of the road consist of pinkish to gray hornblende granodiorite to granite inferred to be a border phase of the Saganaga Tonalite and containing inclusions that are both more felsic and more mafic than the enclosing rock (Fig. 6.7). For example, the large angular block shown in Fig. B

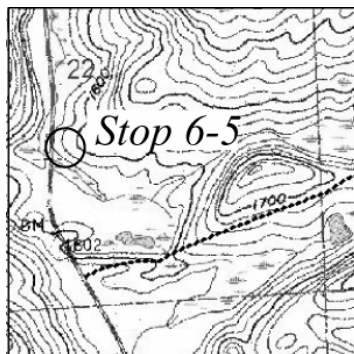
consists of quartz-eye-bearing tonalite, much like rock that is typical of the main phase of the intrusion. Foliation is poorly developed and likely magmatic in origin. This exposure demonstrates that the Saganaga is a composite batholith that, despite its apparent homogeneity, consists of quite varied magmatic phases, particularly near its border. Tonalite on the east side of the road is cut by a fine-grained diabasic dike several meters in width. Although this location lies some distance east of the prominent north-trending aeromagnetic trends associated with dikes that are described in the text above, the dikes presumably are related.



Figure 6.7. Granodioritic phase of Saganaga batholith, containing many and varied inclusions. Large, lighter colored block in center of exposure is inferred to be an autolith of quartz-eye-bearing tonalite similar to the major phase of the batholith.

STOP 6-5

Border phase Neoproterozoic Saganaga Tonalite with flattened inclusions and well-developed foliation in contact zone with Neoproterozoic metabasalt



Location: T. 65 N., R. 4 W., sec. 22, SE, NE, SW; Roadcut on east side of Gunflint Trail (Fig. 6.5)
Long Island Lake quadrangle
UTM: 661,834E/5,329,257N

Description:

This location exposes the border phase of the Saganaga batholith, characterized by a granodioritic composition, general lack of quartz eyes, and an abundance of dioritic inclusions consisting of varied proportions of hornblende, pyroxene, biotite, plagioclase, and minor quartz (Fig. 6.8). The irregular

ovoid and discoid shape of inclusions is oriented subparallel to well developed, steeply dipping and east-trending foliation. Hornblende crystals and aggregates define a prominent lineation plunging shallowly to the east. Petrology indicates that much of this fabric appears magmatic, yet foliation may be a hybrid of approximately coaxial magmatic flow and regional tectonic deformation (D_2). This is typical of the border zone of the intrusion against Archean metabasaltic country rocks, which presumably lie in the low ground just to the south.

A preliminary comparative geochemical study, summarized in Fig. 6.3 above, indicates that the mafic inclusions are not partially assimilated, recrystallized, and tectonically deformed country rock volcanic xenoliths as implied by some earlier workers. They have the common chemical characteristics of the sanukitoid suite of Archean granitoid rocks. Thus, we infer that the inclusions studied are autoliths derived from a separate, primitive sanukitoid-magma.



Figure 6.8. Granodioritic border phase of Saganaga Tonalite containing mafic inclusions (see discussion associated with Fig. 6.3).

STOP 6-6

Contact zone of Neoproterozoic Saganaga Tonalite and metabasalt; unconformably overlain by Paleoproterozoic conglomerate of the Lower cherty member of Gunflint Iron Formation



Location: 3 exposures, all in T. 65 N., R. 4 W., sec. 22, NW, SW, SE; Bush path off Gunflint Trail [individual UTM coordinates given below] (Fig. 6.5)
Long Island Lake quadrangle

Description:

6-6a [UTM: 661,925E/5,329,065N] Archean metavolcanic rocks containing abundant granitic sheets and dikes, presumably related to border phases of Saganaga Tonalite (Fig. 6.9). The boundary between tonalite and metabasalt has been mapped in many places as a fault (Weiblen and others, 1971). These exposures do not preclude that possibility, but they imply that passive emplacement of the intrusion has also occurred, at least locally.



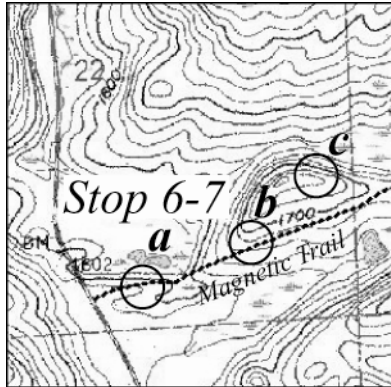
Figure 6.9. Outcrop of border zone of Saganaga Tonalite. Dark gray material probably represents various phases of the intrusion; lighter gray, wedge-shaped area in the foreground is inferred to be metavolcanic country rock.

6-6b [UTM: 661,965E/5,329,062N] Conglomerate developed at the gently southward dipping unconformity between Neoproterozoic intrusive and metavolcanic rocks and the overlying basal part of the Paleoproterozoic Animikie Group. The unit, regionally known as the Kakabeka Conglomerate, is present only locally on the western end of the Gunflint. In most places, the Lower cherty member of iron-formation lies directly on eroded Archean surfaces. This small outcrop is one of the few places where the conglomerate is exposed and accessible along the contact. The conglomerate is greenish gray, poorly bedded, and contains subangular to subrounded fragments of Saganaga Tonalite and related granitoid rocks, metabasalt, and quartz, in a granular siliceous matrix.

6-6c [UTM: 661,965E/5,329,037N] Walking southward from the basal conglomerate is a low "step-up" onto southward dipping strata of the Lower cherty member of Gunflint Iron Formation.

STOP 6-7

Stromatolitic cherty Gunflint Iron Formation



Location: 3 exposures, all in T. 65 N., R. 4 W., sec. 22 [specific locations given below], Magnetic Trail (Fig. 6.5)
Long Island Lake quadrangle

CAUTION and ADVICE: This is a fairly long hike, approximately 1 mile round-trip; please be prepared with water and other field needs. Although this is not in the BWCAW, it does lie within Superior National Forest and is frequented by hikers. For this reason, and to preserve scientific value, please be respectful in matters of hammering and sampling.

Description:

6-7a [Sec. 22 SW, SE; UTM: 662,034E/5,328,8895N] Thin-bedded, fine-grained, chert-amphibole-magnetite-bearing strata assigned to the upper part of the Lower slaty member of Gunflint Iron Formation. Beds strike ENE and dip generally less than 8 degrees southward.

6-7b [Sec. 22 SE, SE; UTM: 662,401E/5,329,093N] Near the edge of this large controlled burn area are several exposures of stromatolitic iron-formation—inferred to be stratigraphically near the base of the Upper cherty member of the Gunflint Iron Formation. Irregular domal and laminar forms are present. Some float blocks of basal conglomerate can be found here, presumably glacially transported southward from the basal unconformity (Fig. 6.10).



Figure 6.10. Boulder of basal conglomeratic unit of Gunflint Iron Formation.

6-7c [Sec. 22 NE, SE; UTM: 662,583E/5,329,257N] Inferred to be in the Upper cherty member of Gunflint Iron Formation. Crest of ridge exposes abundant 3-dimensional views of stromatolites, intraformational conglomerate, and stromatolite "hash," all in a peloidal to ooidal, cherty grainstone matrix. Given the apparent mineralogic replacement and moderate metamorphic grade, little of the original carbon-based material is likely present. Despite this, examples of nearly all morphological forms of stromatolites can be found, including columnar, domal, and laminar (Fig. 6.11).

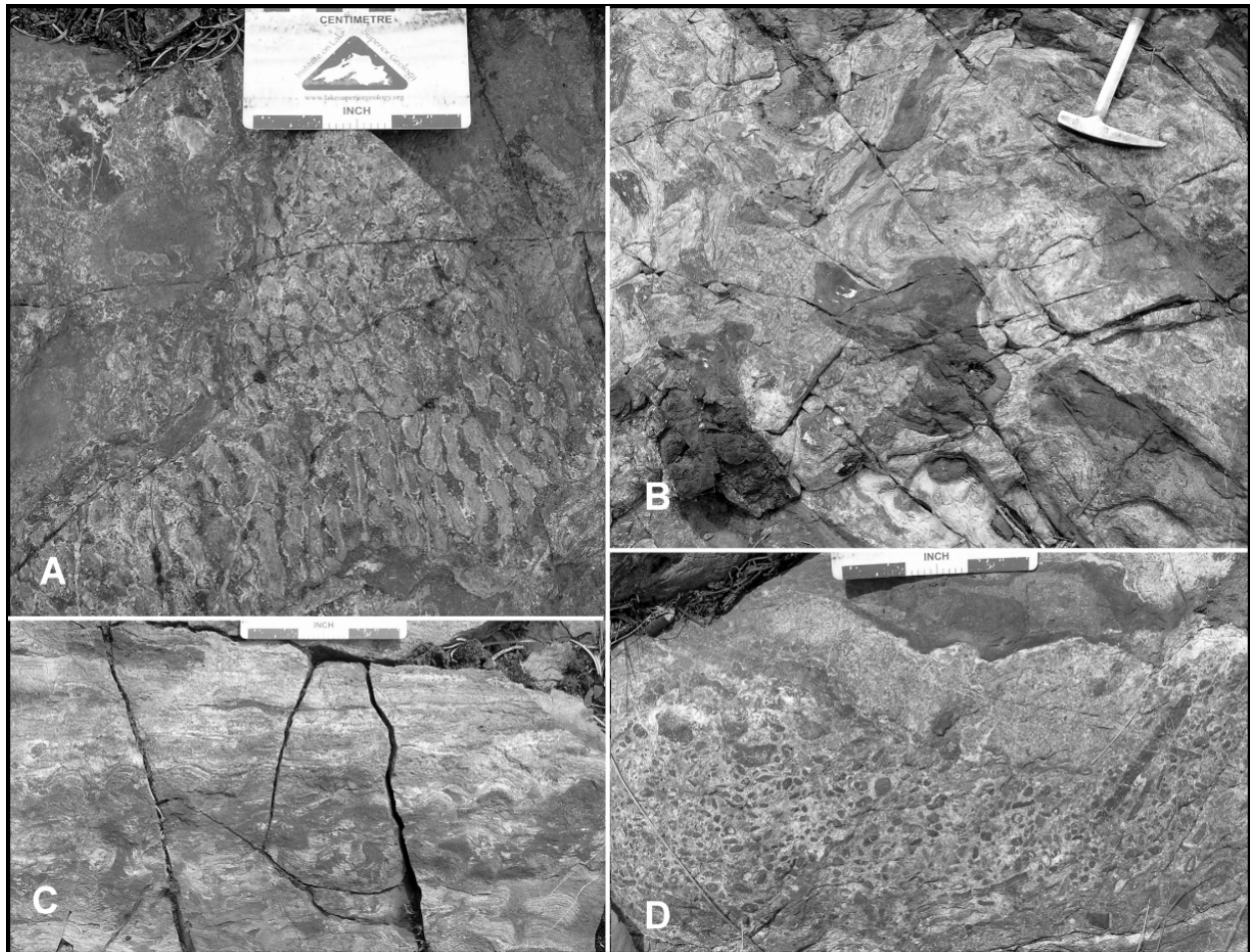
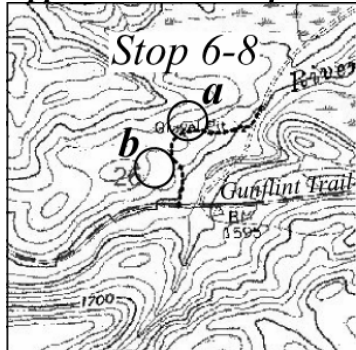


Figure 6.11. Cherty, stromatolitic Gunflint Iron Formation. A=Oblique surface of small columnar stromatolites; B=Horizontal surface of irregular domal stromatolites; C=Vertical section of laminar stromatolites; D="Stromatolite hash."

STOP 6-8

Upper member Gunflint Iron Formation



Location: 2 exposures in T. 65 N., R. 4 W., sec. 26 SW, NE;
Gravel pit north of Gunflint Trail (Fig. 6.5)
Long Island Lake quadrangle

Description:

6-8a [UTM: 663,754E/5,328,212N] Exposed here is a dip-slope composed primarily of granular (cherty) and laminated (slaty) strata of the uppermost Gunflint Iron Formation. The slope defines the southern limb of a large, east-plunging anticline, outlined by interdigitated iron-formation and nearly concordant Logan sills (Fig. 6.5). The gentle dip of this limb illustrates the observation that low angle folding and moderate-relief topography are responsible for the complex map pattern here on the east side of the Lookout fault.

The bedding surface is marked by what we tentatively infer to be syneresis cracks. The cracks, now filled with quartz, occur both concentrically and radially around a central, apparently raised core within a single granular layer of siliceous iron-formation (Fig. 6.12). Syneresis cracks are defined generally as shrinkage cracks formed by dewatering in a gel or colloidal suspension. They differ from septarian cracks that may develop in a similar way, in that the latter typically occur in concretions. Surprisingly diverse interpretations can be found in the literature about syneresis cracks (summarized in Pratt, 2001). There is, however, general agreement that they represent localized tensional failure during sediment dewatering. The explanation for localized semi-brittle response to what likely were formation-wide stresses—caused by compaction or vibration due to syn-sedimentary earthquakes—is more contentious. It has been ascribed variously to the localization of cements, locally increased pore pressure, or zones of granular sediment made coherent by "microbial glue." It is interesting to note that syneresis structures are more prevalent in Precambrian and Cambrian rocks than younger ones. This may be due in part to more uniform organic bonding of clays in younger strata, which reduced the occurrence of stress-localization.



Figure 6.12. Syneresis cracks on eroded bedding surface of thinly bedded Gunflint Iron Formation.

6-8b [UTM: 663,637E/5,328,055N]

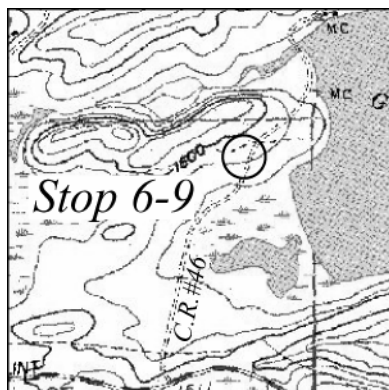
Metamorphosed carbonate-rich iron-formation. This small exposure is likely an outcrop, based on the observation that carbonate strata are seen elsewhere in the uppermost part of the Gunflint, and it is properly positioned relative to stratigraphy as depicted on the geologic map (Morey and others, 1981). The rock appears to be a metamorphosed breccia containing fragments of thinly layered carbonate. Mineralogy is not well known, as the exposure was only recently discovered. Based on published studies (Floran and Papike, 1975) it may contain some combination of ferrohypersthene, fayalite ($\pm$ quartz), grunerite-cummingtonite and perhaps garnet (Fig. 6.13).



Figure 6.13. Uppermost carbonate facies of Gunflint Iron Formation. Large, light-colored crystals may be grunerite-cummingtonite; dark spots appear to be garnet.

STOP 6-9

Paleoproterozoic slate of the Rove Formation



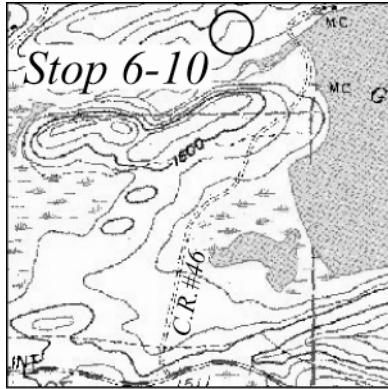
Location: T. 65 N., R. 4 W., sec. 25 NE, NE;
Road cut on County Road #46 about 0.5 mi. N of
Gunflint Trail (Fig. 6.5)
Long Island Lake quadrangle
UTM: 666,750E/5,328,753N

Description:

The small roadside outcrop exposes slate that is typical of the argillaceous parts of the Rove Formation. Bedding strikes eastward and dips gently to the south.

STOP 6-10

Mesoproterozoic Logan Intrusion



Location: T. 65 N., R. 4 W., sec. 24 SE, SE;
Bush trail off County Road #46 about 0.8 miles
north of Gunflint Trail (Fig. 6.5)
Long Island Lake quadrangle
UTM: 665,674E/5,329,111N

Description:

Most of the Logan intrusions consist of medium-grained diabase, as seen along the trail leading to this outcrop. By contrast, this exposure—inferred to lie near the roof of the sill—consists of deformed and partially assimilated xenolithic fragments of Rove Formation slate and siltstone in white-weathering, feldspathic, locally porphyritic and granophyric rock (Fig. 6.14).

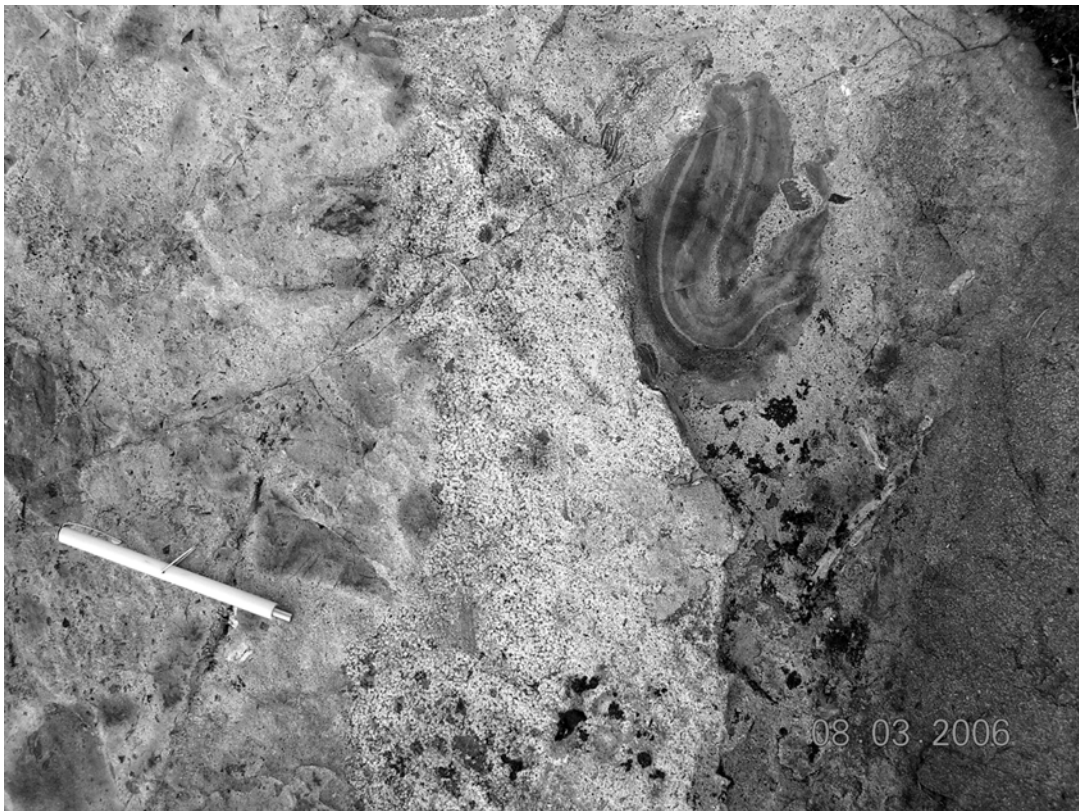
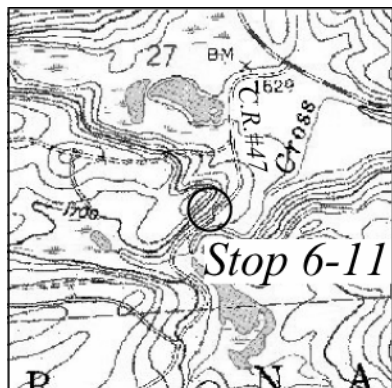


Figure 6.14. Irregular border phase at upper part of one of the Logan Intrusions, containing irregular inclusions of Rove Formation slate country rock.

STOP 6-11

Mesoproterozoic Tuscarora intrusion of Duluth Complex—atypical border phase



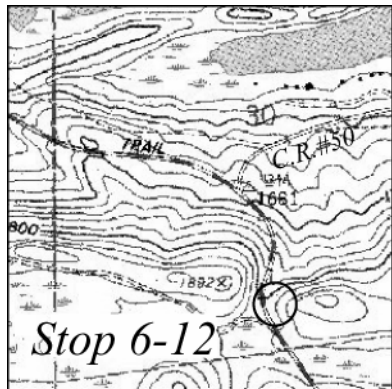
Location: T. 65 N., R. 4 W., sec. 27 SW, SE;
Roadcut on Cross Lake road (CR#47) (Fig. 6.5)
Long Island Lake quadrangle
UTM: 662,074E/5,327,457N

Description:

A confusing exposure of the lower units (*ttp*, *ttf*) of the Tuscarora intrusion. The outcrop consists of intergranular to ophitic gabbro and augite troctolite, with pods and veinlets of coarse mafic pegmatite and shear bands containing sulfide mineralization. Spheroidal weathering produced "core stones" locally.

STOP 6-12

Mesoproterozoic Tuscarora intrusion of Duluth Complex



Location: T. 65 N., R. 3 W., sec. 30 SW, SE;
roadcut on Gunflint Trail east of CR#50. (Fig. 6.5)
Long Island Lake quadrangle
UTM: 666,638E/5,327,433N

Description:

Just south of the parking pull-off is the rather poorly exposed intrusive contact between Rove Formation and units *ttp* and *ttf* of the Tuscarora Intrusion (Fig. 6.5 explanation; Morey and others, 1981). Unit *ttf* is a typical example of Cu-sulfide mineralized augite troctolite that is found at the base of the Duluth Complex here and in the Hoyt Lakes-Kawishiwi area to the southwest. It contains disseminated pyrite, pyrrhotite, and chalcopyrite. In the 1970's, International Nickel Company (INCO) drilled 7 holes in the basal Duluth Complex (Tuscarora and western Poplar Lake intrusions) to evaluate potential for Cu-Ni mining. All of these holes lie along the basal part of the intrusion within a few miles east and west of this stop. The archived drill cores were studied by Mogessie (1976) and Mogessie and others (1976).

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