

51st ANNUAL MEETING

Nipigon, Ontario - May 24-28, 2005

INSTITUTE ON LAKE SUPERIOR GEOLOGY

Part 2 – Field Trip Guidebook





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HOSTED BY:

Mark Smyk and Pete Hollings

Co-Chairs

Ontario Geological Survey and Lakehead University

Proceedings - Volume 51

Part 2 – Field Trip Guidebook

Compiled and edited by Pete Hollings, Lakehead University

Cover Photos: Left - basaltic dyke cutting Osler volcanics on Wilson Island, Middle - Aerial view of the Black Sturgeon fault, Right - Ruby Lake near Nipigon, Northern Ontario.

51ST INSTITUTE ON LAKE SUPERIOR GEOLOGY

VOLUME 51 CONSISTS OF:

PART 1: PROGRAM AND ABSTRACTS

PART 2: FIELD TRIP GUIDEBOOK

TRIP 1: GEOLOGY AND FOLD MINERALISATION OF THE BEARDMORE-GERALDTON GREENSTONE BELT

TRIP 2: QUATERNARY GEOLOGY OF THE BEARDMORE – NIPIGON AREA

TRIPS 3 & 6: A STRATIGRAPHIC TRANSECT ACROSS THE NORTHERN FLANK OF THE MIDCONTINENT RIFT NEAR ROSSPORT

TRIP 4: GEOLOGY AND RARE ELEMENT PEGMATITES OF THE QUETICO SUBPROVINCE NEAR NIPIGON

TRIP 5: GEOLOGY OF THE BLACK STURGEON AREA

Reference to material in Part 2 should follow the example below:

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Table of Contents

Introduction, safety considerations and acknowledgements.....	1
Fieldtrip 1 - Geology and gold mineralisation of the Beardmore-Geraldton greenstone belt	3
Fieldtrip 2 – Quaternary geology of the Beardmore – Nipigon area	41
Fieldtrips 3 and 6 - A stratigraphic transect across the Northern flank of the Midcontinent Rift near Rossport	57
Fieldtrip 4 - Geology and rare element pegmatites of the Quetico Subprovince near Nipigon	71
Fieldtrip 5 - Geology of the Black Sturgeon Area	85

Introduction, safety considerations and acknowledgements

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This volume is intended to serve not only as a guide for 51st ILSG field trip participants but also as a reference for those planning to revisit these areas at a later date. Consequently we have included UTM coordinates in the NAD 83 datum for stops, as well as instructions on how to reach them. As some of the stops are on private and staked land, particularly on Trip 1, please be sure to obtain the land owners' permission before entering their land.

In what is perhaps a first for the Institute this year we are offering a fieldtrip onto Lake Superior. This creates a number of unique safety issues. Please exercise caution when getting in and out of the boats as the outcrops are often extremely slippery. Life jackets must be worn in the boats at all times. If you are planning to revisit these sites please be very careful as Lake Superior is a dangerous lake, waves can often be many metres high and even in mid summer fog can appear very quickly.

The other fieldtrips will be visiting stops along either

major highways or busy logging roads. Please take care when crossing or parking along these roads.

We would like to thank all the other authors who contributed to this field guide (Peter Barnett, Tom Hart, Phil Fralick and Steve Kissin) and also all those who provided comments and assisted with the running of the field trips themselves. We appreciate the assistance and cooperation of the exploration and mining companies in providing us access and information concerning their properties, particularly David Malouf, Roxmark Mines Limited.

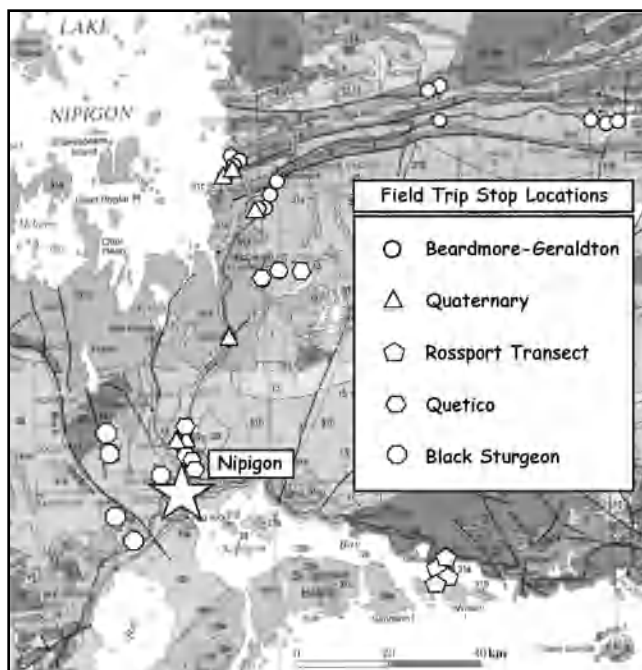


Figure 1. Map showing the location of the five field trips.

Fieldtrip 1 - Geology and gold mineralisation of the Beardmore-Geraldton greenstone belt

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Introduction

The Beardmore-Geraldton belt (BGB) is a Neoarchean metavolcanic-metasedimentary terrane at the boundary between the Quetico Subprovince and the eastern Wabigoon Subprovince of the Superior Province (Fig. 1a). The belt can be subdivided into six east-striking sub-belts, all of greenschist facies metamorphic grade (Fig. 1b; Devaney and Williams, 1989). They are: the northern metasedimentary sub-belt (NMB), northern volcanic sub-belt (NVB), central metasedimentary sub-belt (CMB), central volcanic sub-belt (CVB), southern metasedimentary sub-belt (SMB) and southern volcanic sub-belt (SVB) (Fig. 2). The northern units are not as laterally continuous as the southern units.

Previous studies have shown that although these belts are fault-bounded (Fig. 3), mainly along the

southern boundaries of the volcanic sub-belts, they probably reflect an original sedimentary assemblage deposited on a cratonic margin in environments ranging from alluvial fan-braid plain in the NMB, through fan delta-braid delta in the CMB to a submarine fan/ramp in the SMB (Devaney and Fralick, 1985; Barrett and Fralick, 1985; Devaney, 1987; Barrett and Fralick, 1989). Original continuity of this succession prior to tectonic disruption cannot be proven, but is supported by consistent stratigraphic trends and sedimentary structures that mostly young to the north. Although isoclinal folds are present on varying scales, particularly in the southeast part of the terrane near Geraldton, the overall structure of the belt appears to be initially one of stacked, imbricate, internally northward-younging sheets which have been interpreted as the product of accretionary wedge tectonics (Williams, 1986, 1987; Devaney and Williams, 1989; Williams and Stott,

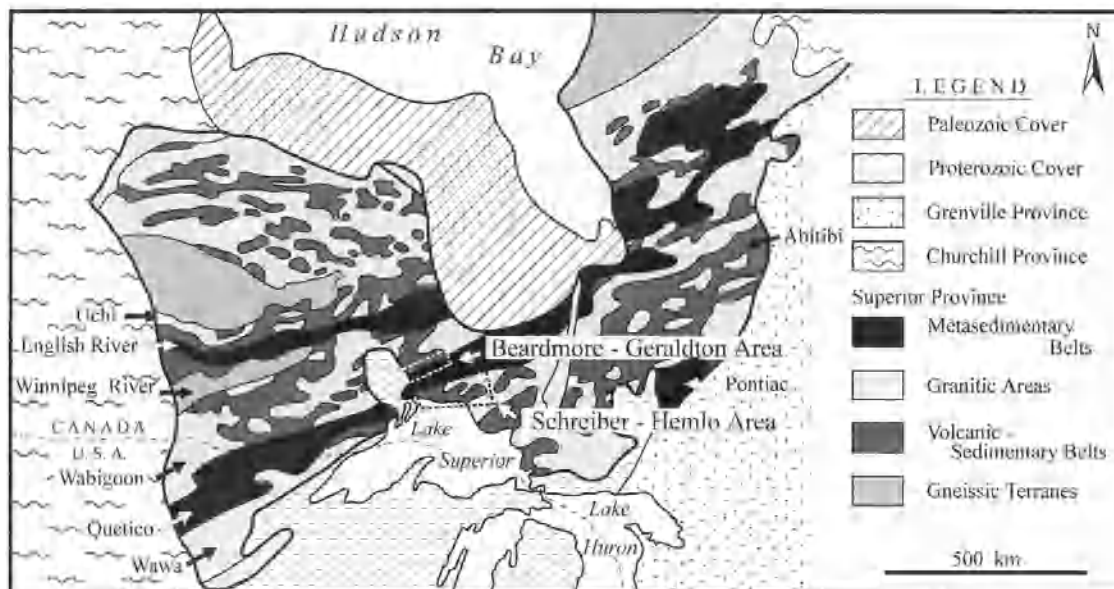
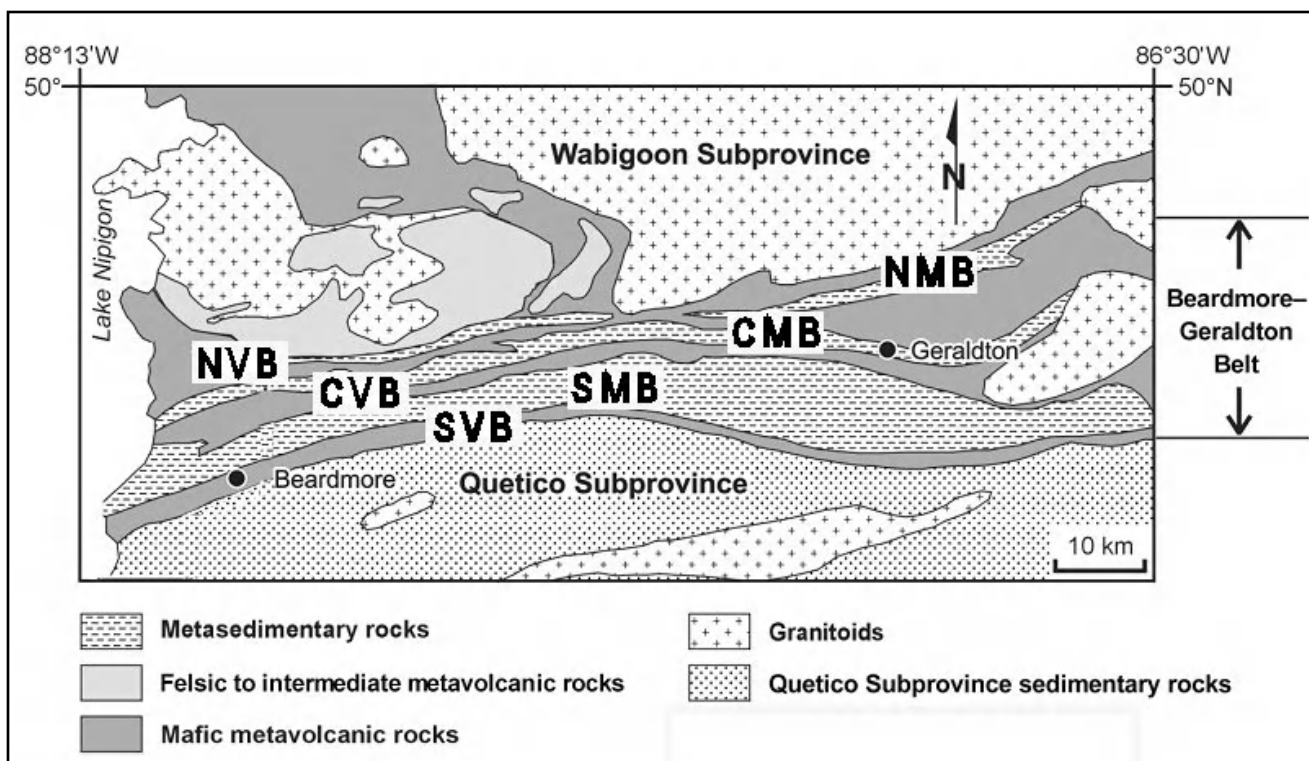
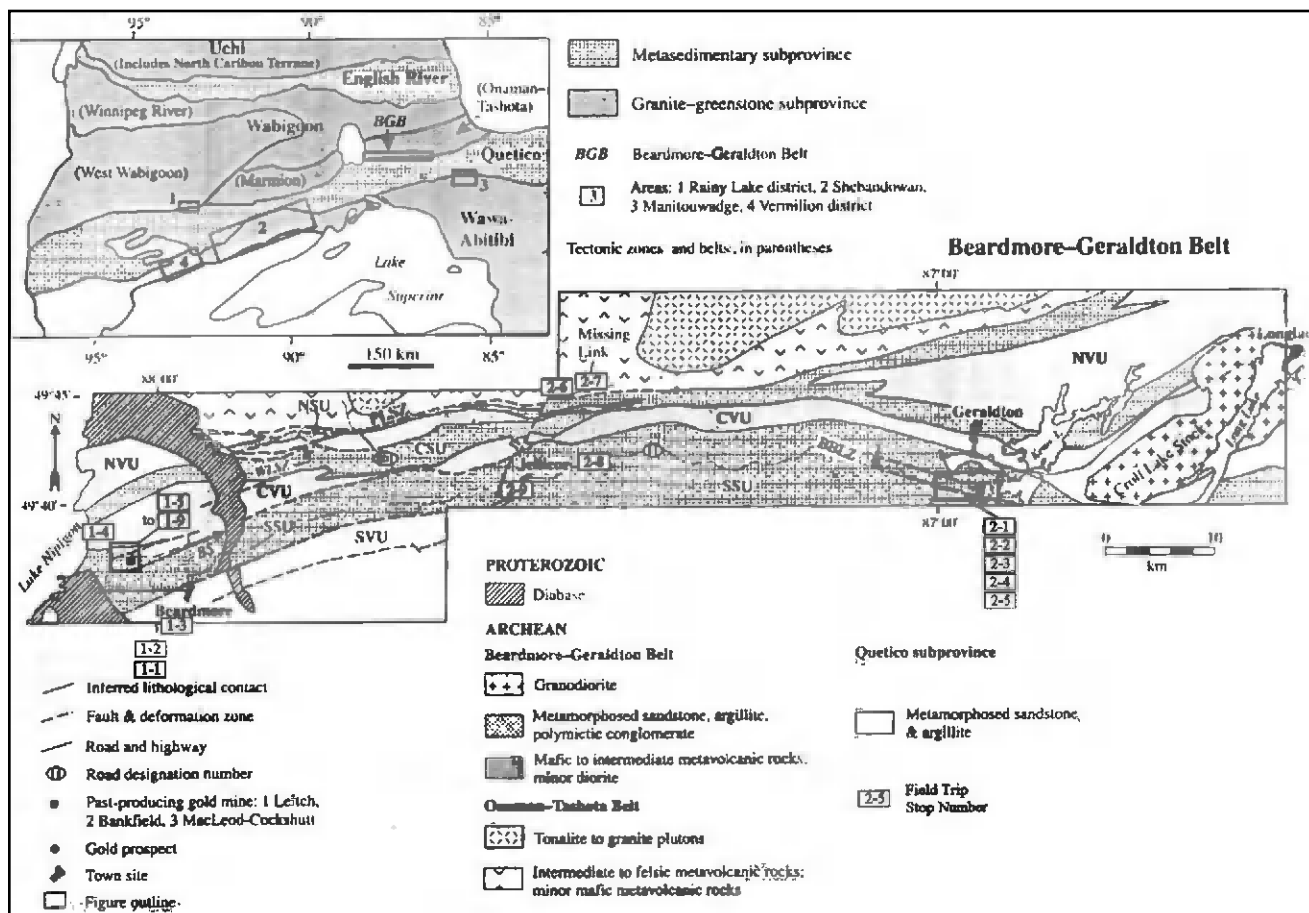


Figure 1a. Location of the Beardmore-Geraldton greenstone belt, Superior Province.



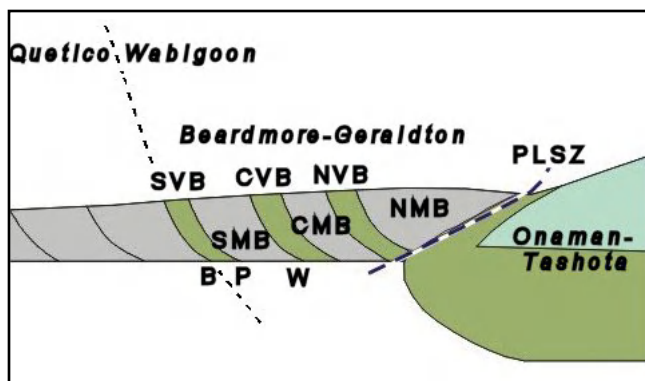


Figure 4. Schematic cross-section of the southern half of the eastern Wabigoon Subprovince showing imbrication of fault-bounded metavolcanic (VB's) and metasedimentary (MB's) sub-belts (modified after Stott and Davis, 1999); S = Southern; C = Central; N = Northern; PLSZ - Paint Lake Shear Zone; B - Blackwater Fault; P - Princess Lake Fault; W - Standingstone - Watson Lake Fault.

1991). Thus, the steeply dipping sheets probably represent the telescoped version of an original cratonic margin assemblage. South of the Beardmore-Geraldton belt, more highly metamorphosed, turbiditic sediments of the submarine margin and trench-fill are present in the Quetico Subprovince (Williams, 1986; Fralick et al., 1993).

The presence of the Onaman-Tashota volcanic arc terrane (ca. 2740 Ma; Stott and Davis, 1999) to the immediate north, the imbricate thrust structure of the Quetico metasedimentary rocks to the south (Williams, 1986, 1987), and the style of sedimentary depositional systems in the region strongly suggests that the Beardmore-Geraldton belt represents a fore-arc assemblage or a cross section of a complete island arc system (see "Depositional Environments") (Fig. 3). The mostly mafic volcanic rocks which form the basement to the sedimentary rocks are probably slivers of oceanic volcanic terranes which were accreted prior to development of the clastic wedge. Similar depositional and tectonic evolutionary trends in the Great Basin of California have been described by Dickinson and Seely (1979).

Igneous rocks

Mafic metavolcanic rocks of the southern metavolcanic sub-belt (SVB) consist of massive and pillowed flows with minor tuffs, lapilli tuffs, and tuff breccias and interflow chert-magnetite iron formations. The central metavolcanic sub-belt (CVB) consists of intermediate massive and pillowed flows with significant tuffs, lapilli tuffs and tuff breccias, and

minor interflow chert-magnetite iron formation. The northern metavolcanic sub-belt (NVB) is subdivided into the northern Bish Bay assemblage (BBA) and the southern Poplar Point assemblage (PPA) (Hart et al., 2002; Fig. 4). The BBA is composed of east-striking, mafic pillowed to massive flows and rare tuffs resembling the SVB. The PPA consists of northwest-striking, intermediate flows, tuff-breccias and tuffs resembling the CVB, with subordinate mafic massive and pillowed flows.

A number of rock types intrude the supracrustal rocks of the BGB, including a series of mafic to ultramafic, synvolcanic, intermediate to felsic synvolcanic, mafic post-tectonic intrusive rocks and diabase dikes (Hart et al., 2002). The synvolcanic gabbroic rocks form thin sills, subparallel to the strike of the mafic metavolcanic rocks of the SVB and the BBA. A large intrusive body located in the BBA is a composite intrusion with gabbroic and peridotitic phases in the southern and northern parts of the intrusion, respectively. A series of intermediate to felsic dikes and sills ranging from massive granodiorite to quartz-porphyritic, feldspar-

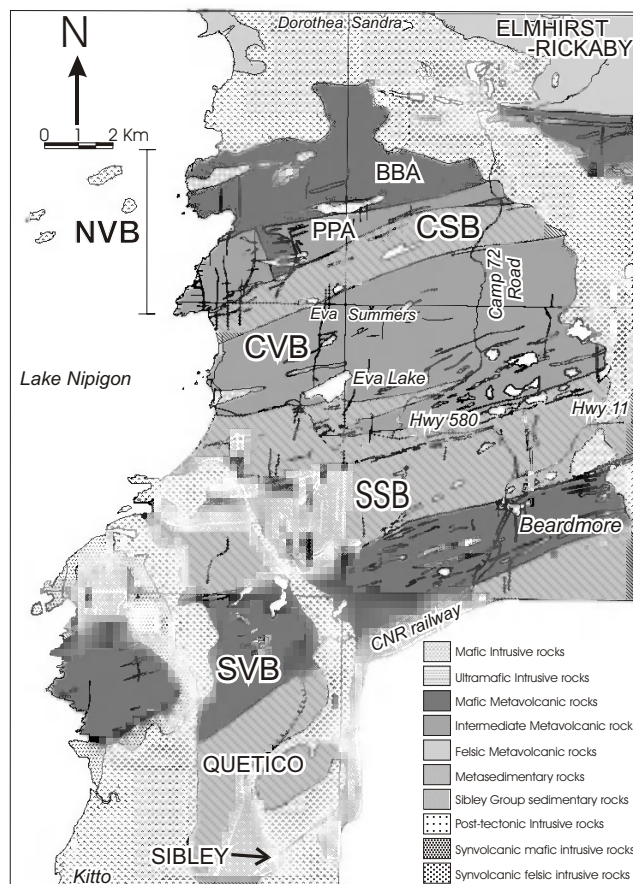


Figure 4. Geology of the Beardmore area (Hart et al., 2002); BBA - Bish Bay Assemblage; PPA- Poplar Point Assemblage.

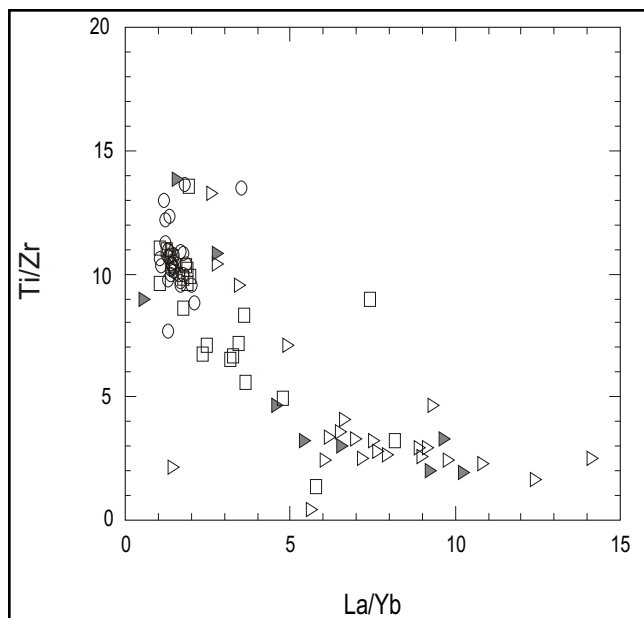


Figure 4. Zr/Ti versus La/Yb diagram for the metavolcanic rocks of the Beardmore area. Circles - SVB, open and filled triangles - CVB, squares - NVB. Data from Hart et al. (2002) and Tomlinson (1996).

porphyritic and feldspar-quartz-porphyritic phases occur within the metavolcanic rocks of the PPA. These units appear to have been emplaced along the regional foliation, although some bodies are subhorizontal in orientation. A feldspar-porphyritic granodiorite dike intrudes the mafic flows of the SVB and resembles the dikes of the PPA. Late, post-tectonic diorite sills predominantly occur within the metasedimentary and metavolcanic rocks along the contact between the SSB and CVB. Additional intrusions located along the northern and southern contacts of the PPA are generally undeformed diorite sills that have chilled contacts with the metasedimentary rocks. A swarm of narrow, generally north-striking diabase dikes intrude the supracrustal rocks and appear to be predominantly Paleoproterozoic in age. A series of Mesoproterozoic diabase sills of the Nipigon Sill Complex intrude all other supracrustal rocks of the BGB.

Geochemical groupings in the metavolcanic rocks span the three belts within the western portion of the BGB. The mafic metavolcanic rocks of the SVB and NVB are predominantly basalt with high Ti/Zr and low Zr/Y, and negative slopes on mantle-normalized, extended element diagrams characteristic of mid-ocean ridge basalt (Fig. 5; MORB; Tomlinson, 1996; Hart et al., 2002). A subset of basalt samples have a mixture of elemental abundances and ratios characteristic of subduction-related volcanic rocks

which led Tomlinson (1996) to suggest that these rocks formed in a back-arc basin, similar to the Lau Basin, where there are volcanic rocks with both arc and/or MORB geochemical characteristics. The intermediate metavolcanic rocks are generally confined to the CVB and PPA of the NVB and are predominantly andesite. The andesites have low Ti/Zr, high Zr/Y, and positive slopes with prominent negative niobium and tantalum anomalies on a mantle-normalized, extended element diagram, that are generally characteristic of calc-alkaline andesite formed in an island-arc environment (Fig. 5; Tomlinson, 1996; Hart et al., 2002). Two felsic metavolcanic rocks from the CVB and the PPA were analysed by Hart et al. (2002), and have low $(La/Yb)_n$ ratios with moderately high ytterbium values characteristic of FII felsic metavolcanic rocks (Leshner et al., 1986), indicative of a potential to host volcanogenic massive sulphide mineralization. Both samples have elemental abundances and ratios characteristic of active continental margin and/or the destructive plate margin felsic volcanic rocks (e.g., Wood, 1980). A sample of the flow from the PPA has a U/Pb age from zircons of 2724.9 ± 1.1 Ma and the flow from the CVB has a U/Pb age from zircons of 2724.9 ± 1.2 Ma (Hart et al., 2002).

The geochemistry of the intrusive rocks of the western portion of the BGB was characterised by Hart et al. (2002). The mafic intrusive rocks have elemental abundances and ratios that generally reflect the composition of the metavolcanic rocks that they intrude. Unfortunately these similarities also mean that the geochemistry does not aid in differentiating coarse-grained, gabbroic-textured flows from subvolcanic, gabbroic intrusions. Both the southern mafic and northern ultramafic phases of the composite intrusion in the BBA have similar tholeiitic geochemical characteristics. A few analyses that are available for the intermediate to felsic dikes and bodies indicate that these units have high Zr/Y and $(La/Yb)_n$ ratios characteristic of alkalic rocks, which may be interpreted to be the products of either an early Archean, post-tectonic intrusive event or part of a Proterozoic magmatic event related to the initial rifting during formation of the Nipigon Embayment. A number of post-tectonic, mafic intrusions located along the contact between the SSB and the CVB have Ti/Zr, Zr/Y and $(La/Yb)_n$ ratios similar to the andesite of the CVB, and generally characteristic of calc-alkaline rocks. The overlap in the geochemistry of these intrusions and the CVB andesite is interpreted to indicate that these two

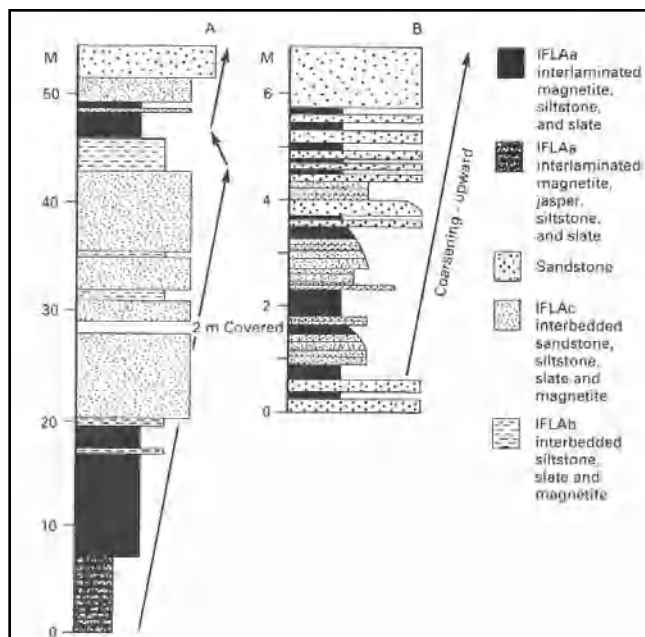


Figure 6. Stratigraphic sections of iron formation bearing units. A) Outcrop section measured at the Leitch Mine (for detailed location see Barrett and Fralick, 1985). B) Outcrop section measured at Solomon's Pillars (for detailed location see Fralick 1987). From Fralick and Barrett (1995).

rock types formed by similar petrogenetic processes, but the occurrence of intrusions in both metavolcanic rocks and metasedimentary rocks indicates that some of these intrusions are younger.

Sedimentary rocks

Lithofacies Associations

The NMB, the northern (uppermost) third of the CMB and the northernmost portion of the SMB are dominated by a conglomeratic assemblage with minor amounts of sandstone. The clast-supported conglomerates are poorly to moderately sorted, and almost always non-graded with a poorly to moderately sorted sand matrix. Bedding is defined by variation in average or maximum clast size between units, but it is commonly indistinct. Scouring is locally preserved, but most other primary features such as imbrication have been destroyed by deformation. Sandstones interbedded with the conglomerates commonly appear massive, but in some outcrops planar lamination and cross-stratification are present. They form lenses in conglomeratic beds; thin, irregular sheets blanketing conglomeratic beds; wedges abutting conglomeratic beds; and thick units separating conglomerates. Clast types in the conglomerates are almost exclusively igneous, representing a suite of rocks similar to those present in the Onaman-Tashota volcanic terrane to the

north.

The turbiditic association of the SMB can be divided into a clastic association and a chemical association, with a high proportion of oxide-facies, banded iron formation (BIF) layers. In the chemical association, clastic interbeds are generally less than several centimetres thick, and range in grain-size from silt to coarse sand. Upward-thickening and -coarsening trends over several metres are locally present, as at Solomon's Pillars and the Leitch Mine near Beardmore (Fig. 6; Barrett and Fralick, 1985). Within the overall upward trend, oscillations between silts, sands, and iron formation occur. Depending on the relations between these types of beds, four iron formation lithofacies associations (IFLA) can be defined:

- IFLA-a consists of dominantly magnetite-rich sediment with millimetre- to centimetre-scale, graded or ungraded silt interbeds (Fig. 7).
- IFLA-b comprises centimetre-scale graded to sharply bounded silt beds, either contiguous or separated by millimetre-thick laminae of magnetite-rich sediment.
- IFLA-c represents sand-rich composite units up to about 1m thick, generally consisting of thin, stacked, ungraded, laminated sand beds. These composite units consist mainly of medium- to coarse-grained sand. They are separated by intervals of IFLA-b up to 15cm thick.
- IFLA-d consists of framework-supported, polymictic conglomeratic beds up to a few metres thick, interbedded with sandstone and minor iron formation, or interbedded with fairly thick iron formation and thin-bedded sands.

Conglomerate contain mainly mafic to felsic volcanic and granitic clasts. Although flattened clasts indicate that IFLA-d outcrops are tectonically thinned, their associations are primary, with the conglomeratic units erosively cutting down into BIF-sandstone packages. Transitions between IFLA-a, -b and -c can be gradual or abrupt. Some silt-sand successions containing iron formation exhibit intervals of thicker and well-graded clastic beds. They form structured sections up to several metres thick within successions that are otherwise generally disorganized.

Clastic units in the lower two thirds of the CMB and the SMB are divisible into three lithofacies associations: a thin-bedded, turbidite-dominated

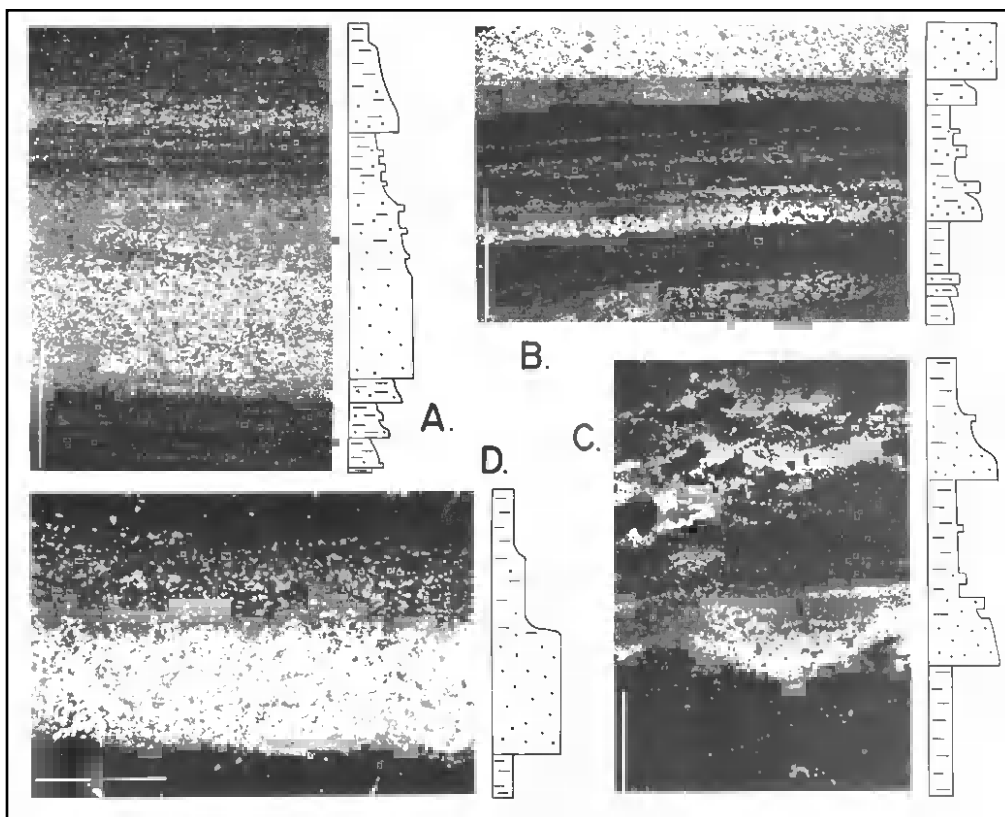


Figure 7. Photomicrographs of siltstone associated with laminated, magnetite-rich sediment (IFLA-a). Logs to right of photos outline layer contacts and schematically illustrate grain size variations. Dots and dashes represent siliciclastic and magnetite-rich components, respectively. Scale bars: A.) 5.0 mm; B.) 2.0 mm; C.) 3.0 mm; D.) 2.0 mm. All images are in crossed nicols. From Barrett and Fralick (1985).

association (LA2); a medium-bedded, turbidite-dominated association (LA3); and a thick-bedded association (LA4). LA2 consists mostly of graded, <10 cm thick siltstone and/or sandstone beds that are either unstructured or thin and fine upwards over 1 to 3 m. LA3 is divisible into two types. LA3a consists of medium- to coarse-grained sandstones with sharp bottom and top contacts. Parallel lamination is present near the tops of some of the otherwise massive beds. These successions are unstructured. LA3b is similar to LA3a except these beds are organized into either upward-thickening or upward-thinning trends. Thick, poorly graded sandstones dominate LA4. The beds usually have a coarse sand or pebbly base, grading into a thick, poorly sorted, massive central area. They are often abruptly capped by thin, fine-grained sandstone. Irregular, erosional bases and scattered rip-up clasts are common.

Structured, thinning- and fining-upward sequences, metres to tens of metres thick, are present in the area south of Beardmore. The successions are topped by CDE and/or DE turbidites which are abruptly overlain by massive grain flows/high-density turbidites with

internal inverse- to normal-graded, conglomeratic bands. Pebbles present in the conglomerates are mainly felsic igneous rocks (extrusive and intrusive), while rip-up clasts are not the expected mudstone or siltstone, but rather clay- and silt-rich, fine-grained sandstone.

Load structures are ubiquitous throughout the area. Commonly, the base of one unit sags into the underlying beds. Locally, multiple internal loads are developed, usually in the B division. These loads sag into the A division, in places extending into the underlying beds.

Depositional Environments

Deposition of the clastic and chemical sediments in the Beardmore-Geraldton belt occurred within a narrow time frame. Crosscutting felsic intrusive rocks provide a minimum age for the sediments of $2691 \pm 3/-2$ Ma (Anglin, 1987; Anglin et al., 1988). U-Pb detrital zircon geochronology on sandstone samples from the region consistently give youngest zircon ages in the range 2701 ± 1 Ma to 2696 ± 2 Ma (Hart et al., 2002; Don Davis, unpublished data, personal communication, 2004). These ages are very similar to

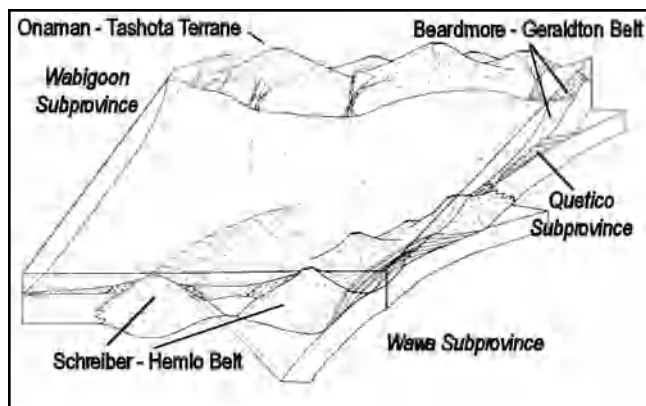


Figure 8a. 2700 Ma paleogeography of the region surrounding Beardmore-Geraldton. The island arc in the south represents the Schreiber-Hemlo area of Wawa Subprovince. The trench system to the north of this arc is now metaturbidites of the Quetico Subprovince. The sediments feeding from the southern flank of the Wabigoon cratonic arc in the north formed the Beardmore-Geraldton succession. From Fralick and Barrett (1991).

ages of detrital zircons from the Quetico Subprovince and a more distal turbiditic assemblage, the McKellar Harbour Formation, outcropping in the northern Wawa

Subprovince on Lake Superior (Purdon, 1995). In addition to geochronology, lithofacies associations and geochemistry indicate that the depositional system probably extended from the cratonic margin (i.e., Wabigoon Subprovince), across the arc-trench gap and trench (i.e., Quetico Subprovince) to the outboard ocean floor (i.e., Wawa Subprovince) immediately prior to the commencement of collision of an island arc system to the south.

The NMB is represented by sandstone and conglomerate deposited in a fluvial fan, and/or braid plain environment (Devaney and Fralick, 1985). The CMB displays this type of assemblage in its northern third, with successions in the southern portion similar to those described above for the SMB. Thus, the paleo-shoreline exposed by the present level of erosion lies within the CMB. All of the sediments in the SMB are marine, except for rare conglomerates at the top of the succession and throughout the section in the highly deformed Geraldton area. The turbidites were deposited in fore-arc environments ranging from delta-

front to submarine fan or ramp. Many unstructured sandstone sequences are interpreted as having accumulated on a submarine ramp that was locally transected by channels feeding deeper fan lobes (Barrett and Fralick, 1989). An interpretive view of the overall paleoenvironment is shown in Figures 8a and b.

Sediments deposited in the Quetico trench to the south are similar in many respects to the submarine ramp deposits of the adjacent fore-arc basin, but there are differences. AB and ABC

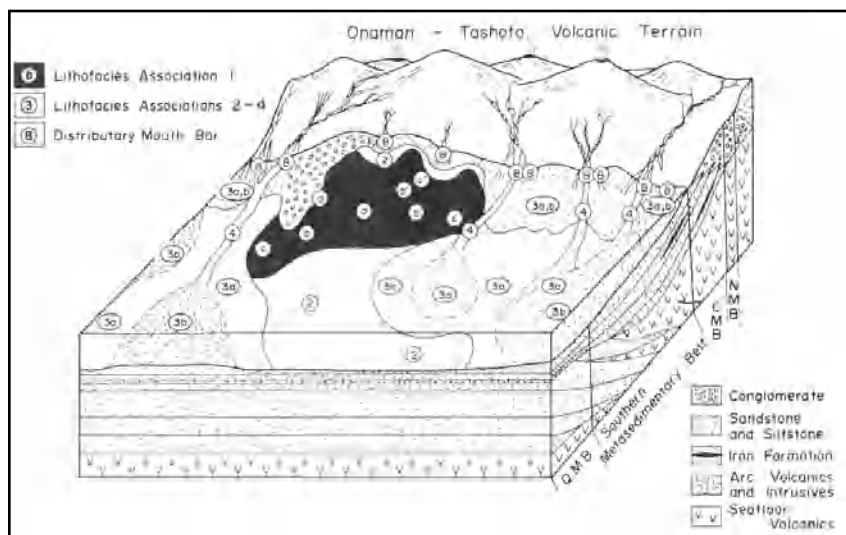


Figure 8b. Suggested depositional environments for sedimentary rocks in the Beardmore-Geraldton belt. The subaqueous portion of this diagram is depicting environments mostly present in the Southern Metasedimentary Belt. The Central Metasedimentary Belt represents mainly fan-delta environments, and the Northern Metasedimentary belt a braidplain environment. The entire sedimentary assemblage in all three belts forms a large-scale progradational system. In the hinterland is the Onaman-Tashota volcanic arc terrain which supplied the detritus. This sediment was transported across braidplains to distributary mouth bars. Because of rapid accumulation rates and the tectonically active environment, gravity flows and slumps were frequently triggered. The resultant grain-flows produced the vertically unstructured, poorly graded, mud-poor, medium- to coarse-grained deposits of the ramp succession. More structured sandstone sections formed on locally developed fan lobes and thick grain-flows filled the submarine feeder channels. Iron formation was deposited immediately offshore from the channel mouths on braid deltas during periods of low sediment influx. The meter-scale coarsening-upwards cycles present in the iron formation record the effects of delta lobe outbuilding. Progradational events may result in the stacking of coarsening-upwards iron formation-bearing, subaqueous parasequences or the delta top braid system may erode through the uppermost chemical sediments during maximum outbuilding. The iron formation is limited to the shore-proximal zone indicating a precipitation model where nutrients delivered to the system by rivers caused heightened cyanobacterial growth rates in the near-shore, similar to modern systems. The oxygen produced by photosynthesis caused precipitation of iron oxides in these oxygen oases. From Barrett and Fralick (1989)

turbidites dominate the trench fill, whereas grainflows and A turbidites characterize the more proximal, fore-arc environment. Iron formation is present in the fore-arc basin associated with thinner-bedded, finer-grained sediments deposited in the shallow near-shore portions of the deltas. Iron formation accumulated when transgressions caused intervals of clastic sediment starvation in the proximal pro-delta and distributary bar environments. Iron formation is absent from sediments deposited further offshore in the Beardmore-Geraldton belt and the Quetico trench deposits. The limited distribution of oxide-facies iron formation (occurring only in shore-proximal areas) indicates that biological activity thrived in the near-shore waters where the commonly limiting nutrients, nitrogen and phosphorous, are delivered at higher levels due to river input. It is difficult to prove that iron compound precipitation was driven by free oxygen liberated during photosynthesis by cyanobacteria, but this would be a logical assumption.

Rare successions that coarsen and thicken upward may represent limited establishment of discrete fan lobes in the distal fore-arc basin and Quetico trench

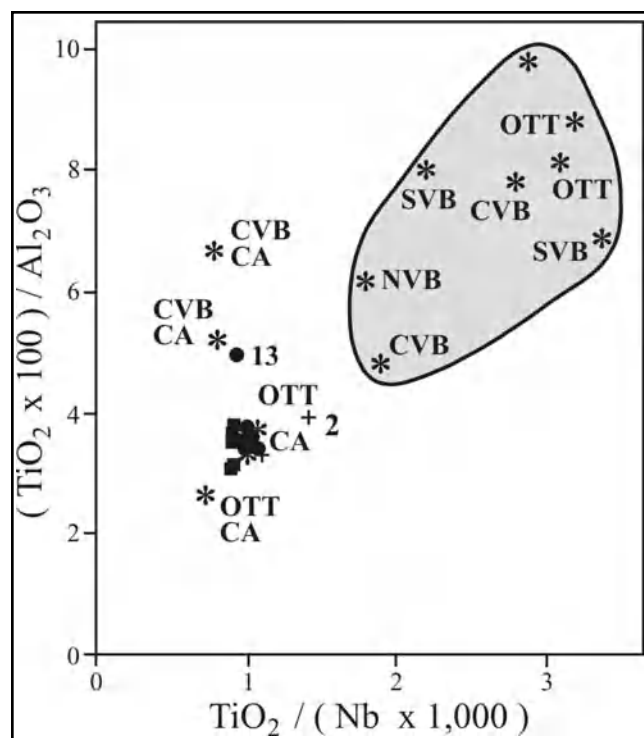


Figure 9. Ratio plot of chemically immobile elements in sandstones showing derivation of the sands deposited in the Beardmore-Geraldton terrain and the Quetico Subprovince from the Onaman-Tashota calc-alkaline volcanic arc. From Fralick and Kronberg (1997); Fralick (2003); igneous data from Tomlinson et al. (1993) and Kresz and Zayachivsky (1989).

floor, but for the most part the trench fill was dominated by overlapping, interfering fans fed by multiple channels extending from the shallow marine portion of the fore-arc basin. One of these channels outcrops south of Beardmore (Stop 1-2). The channel is filled with thick, upward-thinning and -fining successions that grade from conglomerate and coarse sandstone grainflows or turbidites deposited from high density currents near channel bases, to CDE and DE, low-density current turbidites that fill abandoned channels near their tops.

Sediment supplied to the fore-arc basin and the trench was derived from erosion of felsic and intermediate igneous material, with a large contribution probably coming from resedimentation of unconsolidated volcanoclastic material. Clast compositions indicate that the Onaman-Tashota arc terrane to the north was the source (Devaney, 1987; Devaney and Williams, 1989). Grain size and lithofacies trends also strongly support a source to the north. In addition, whole rock geochemistry conducted on sandstone samples from the Beardmore-Geraldton area and the Quetico Subprovince further confirm that the Onaman-Tashota terrane is the only igneous complex in the area which could have supplied the majority of the detritus (Figs. 9 and 10; Fralick and Kronberg, 1997; Fralick, 2003). The ease with which volcanoclastic material can be physically eroded from subaerially erupted debris could have led to short subaerial residency periods, and thus lowered the extent of chemical weathering and production of clays. This is probably the major factor contributing to the rarity of E division fines in the deposits.

Structure

The discussion of the structural development is taken largely from recent work and synthesis conducted by Lafrance et al. (2004). Their treatment of the topic is the most recent structural study, outlining three phases of structural overprinting of the volcano-sedimentary assemblage and ascertaining the timing of each.

After deposition of the clastic succession the area was tectonised by thrust faulting, regional folding and dextral shearing. Thrust faulting resulted in the regional volcanic and sedimentary packages being vertically juxtaposed in an imbricated thrust stack (Devaney and Williams, 1989). This represents D1 and may be related to uncommon, early, F1 folds. The youngest detrital

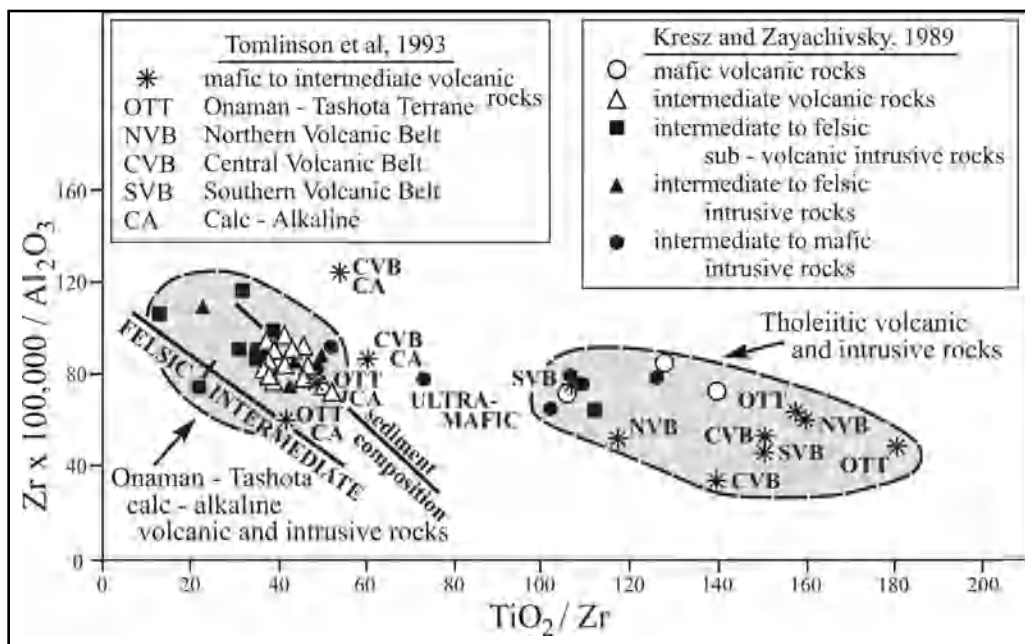


Figure 10. A ratio plot similar to Figure 9 using Zr. From Fralick and Kronberg (1997); Fralick (2003); igneous data from Tomlinson et al. (1993) and Kresz and Zayachivsky (1989).

zircon recovered from the sedimentary units is 2696 ± 2 Ma (Hart et al., 2002) and this places a maximum age on the thrusting event.

The D2 event is recorded by the development of tight to isoclinal folds and a flattening strain fabric responsible for transposed bedding and flattening of clasts and pillows. A homoclinal, north-younging panel of regional extent developed at this time and represents the sheared-off southern limb of a larger syncline. D2 deformation affects altered and gold-mineralized porphyry dykes in the syn-tectonic Croll Lake stock which U-Pb geochronology indicates are $2691 \pm 3/-2$ Ma (Anglin, 1987; Anglin et al., 1988). An age of 2699 ± 1 Ma for a Au-mineralized feldspar porphyry dyke at the Hard Rock Mine and identical ages of 2690 ± 1 Ma for two phases of the Croll Lake stock put constraints on the timing of major deformation and hydrothermal activity in the belt (Corfu, 2000).

The final event, D3, occurred as regional transpression developed in the compressive framework of the area. Vertical bed orientations developed during D2 did not refold but rather were overprinted by a steeply dipping, regional cleavage. Partitioning of the strain, during east-west dextral shear, between less competent argillites and more competent sandstones and porphyries resulted in cleavage refraction near lithologic contacts. The pervasive cleavage developed in the Paint Lake shear zone at this time shows a progressive rotation towards the orientation of the zone (Lafrance et al., 2004; DeWolfe et al., 2000).

This is in contrast with the Barton Bay Lithotectonic Zone (BBLZ) where the S2 fabric was reactivated to accommodate the D3 shear. Some folds were generated during this interval but they tend to be smaller Z-folds, overprinting limbs of regional F2 folds. Shear zones active at this time were dextral with nearly horizontal displacements.

To sum up the results of Lafrance et al. (2004):

- Thrusting along the southern margin of Wabigoon Subprovince in the Beardmore-Geraldton area between 2696 Ma and 2691 Ma resulted in the large-scale assemblage of what was to become the belt structure.
- The rocks were then modified by a compressive event (D2), probably between 2692 Ma and 2686 Ma, which steepened the beds to a near-vertical position, caused extensive flattening and formed large-scale fold structures.
- D3 is more poorly constrained in time but may be synchronous with 2670 to 2650 Ma peak metamorphic conditions in Quetico Subprovince. It is characterized by the development of dextral shear zones and small-scale Z-folding.

Gold mineralisation

Overview

Gold was first discovered in the Beardmore-Geraldton Greenstone Belt in 1925 (Mason and MacConnell,

Table 1 - Gold mineralisation in the Beardmore-Geraldton area. From Mason and McConnell (1983).

Mine	Years of production	Ore milled (tons)	Gold produced (Oz)	Average grade (Oz/ton)	Silver produced (Oz)
Bankfield	1937-42, 1944-47	231,009	66, 417	0.29	7,590
Brengold	1941, 1949	46	134	2.91	
Crooked Green Creek	1980-1984	1,455	471		
Hard Rock	1938-1951	1,458,375	269,081	0.18	9,009
Jellicoe	1939-1941	10,620	4,238	0.40	145
Leitch	1936-1968	920,745	847, 690	0.92	31,802
Little Long Lac	1934-54, 1956	1,780,516	605,499	0.34	52,750
MacLeod-Cockshutt	1938-1968	10,337,229	1,475,728	0.14	101,388
Magnet Consolidated	1938-43, 1946-52	359,912	152,089	0.42	16,879
Maloney Sturgeon	1937	1	73	73	16
Maylac	1946-1947	1,518	792	0.52	46
Mosher Long Lac	1962-1966	2,710,657	330,265	0.12	34,604
Northern Empire	1934-41, 1949	425,866	149, 493	0.35	19,803
Orphan (Dikdik)	1934-1935	3,525	2,460	0.70	1,558
Sand River	1937-1942	157,870	50,065	0.32	3,628
Sturgeon River	1936-1942	141,123	73,438	0.51	15. 922
Talmora-Long Lac	1942, 1948	6,634	1,417	0.21	36
Tashota-Nipigon	1935, 1938	51,200	12,356	0.24	14, 527
Theresa	1935-38, 1941-1943, 1945, 1950-53, 1955	26,120	4,785	0.18	202
Tombill	1938-42, 1955	190, 622	69, 120	0.36	8,595
Total			4,115,611		318,500

1983). The Beardmore-Geraldton greenstone belt has produced approximately 4.1 million ounces of gold and over 300,000 ounces of silver. Most of this production came from two distinct gold camps at Beardmore and Geraldton, respectively. The balance of gold production came from several small mines scattered 30 km northeast of Beardmore and one mine south of Longlac. Table 1 (Mason and McConnell, 1983) is a production summary of all past gold producers.

Gold in the field trip area occurs dominantly in quartz-carbonate veins and sulphide replacement zones cutting all the major Neoproterozoic rock types of the area. The vein systems are most typically hosted by: 1) clastic sedimentary rocks; 2) contacts between clastic sedimentary rocks and albite porphyry; 3) interbedded iron formation and clastic sedimentary rocks (Macdonald, 1984); 4) interflow iron formation and mafic metavolcanic rocks; and 5) post-tectonic mafic intrusive rocks (Hart et al., 2002).

Structures Related to Mineralization

The fine-grained sedimentary rocks in the southern portion of the area are isoclinally folded with a well-

developed, sub-vertical, east-striking cleavage. The fold axes vary in trend and plunge: in the west, the fold axes plunge west at moderate angles (Hart et al., 2002; Lafrance et al., 2004); near Jellicoe fold axes are vertical; and as we move progressively to the east the fold axes return to a westward plunge (Kehlenbeck, 1983). Shear discontinuities along axial planar surfaces also become more prevalent to the east (Kehlenbeck, 1983). Earlier work also outlined a later deformational event. It formed chevron folds with cleavage co-planar to the axial surfaces (Kehlenbeck, 1983; Anglin and Franklin, 1985). The second deformational event of Kehlenbeck (1983) is probably spatially and genetically related to regional, east-trending fault zones and is thus at least partially the same event as D3 of Lafrance et al. (2004).

Geraldton Camp

Gold mineralization in the belt has resulted from the introduction of hydrothermal fluids in zones of high crustal permeability. Permeability was generated by prolonged, multiple periods of deformation which focused not only fluids, but magmatic activity and intrusions. In the Geraldton camp (Fig. 11), a major zone

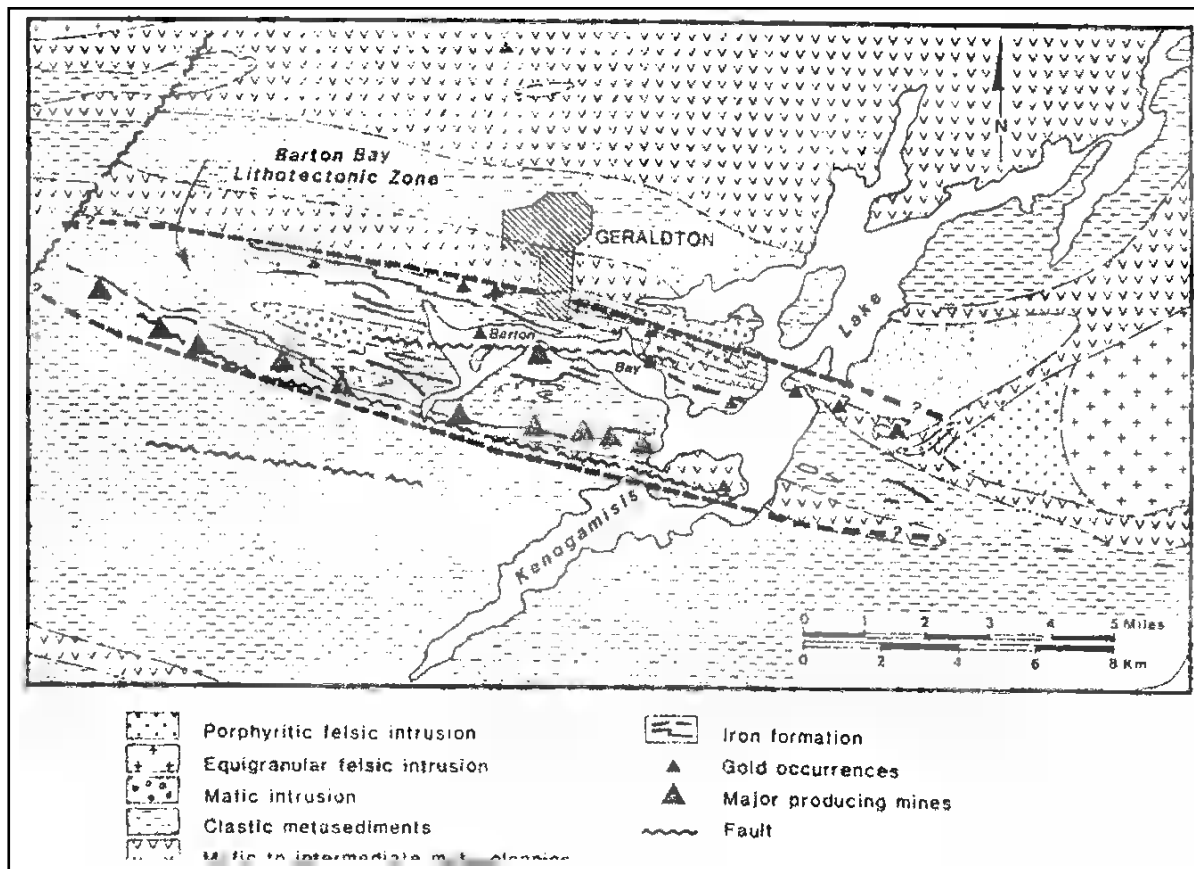
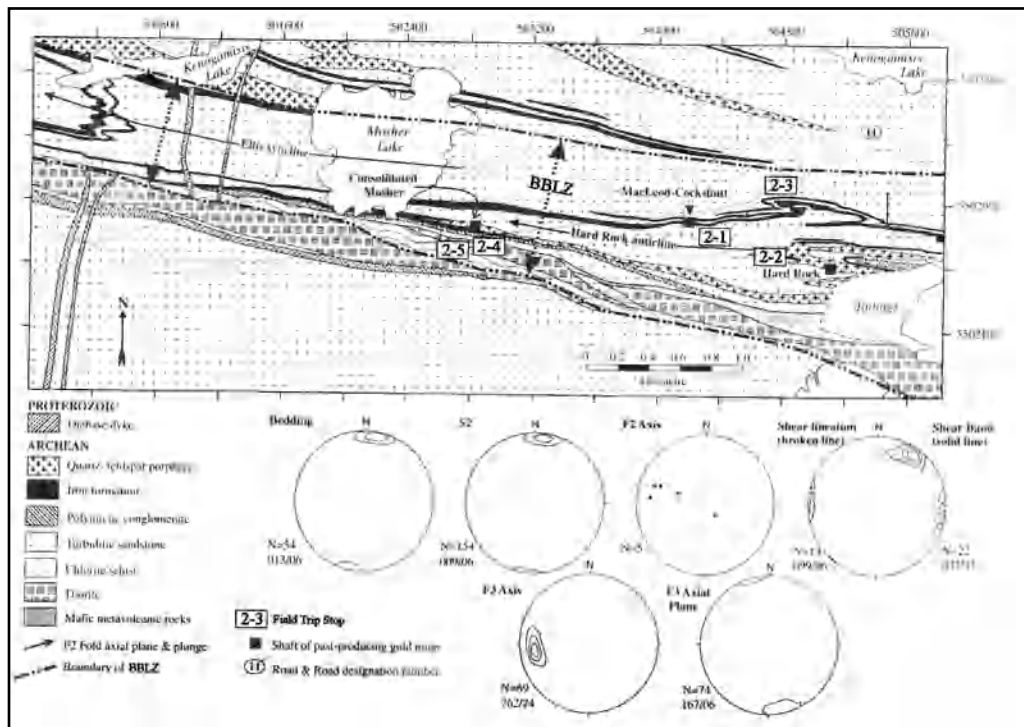


Figure 11. Spatial association of gold deposits to the Barton Bay lithotectonic zone north of the Bankfield-Tombill Shear Zone. Modified after Macdonald (1985).

of deformation in which the gold mines are located has been alternatively termed: the Bankfield-Tombill Fault Zone (Pye, 1952; Horwood and Pye, 1955); Tombill-Bankfield Deformation Zone (Lafrance et al., 2004);

and Barton Bay Deformation Zone (Williams and Stott, 1991). In acknowledging not only the deformation within this zone, but also its lithologic variety, we will use the term Barton Bay Lithotectonic Zone (BBLZ)

(Fig. 12).



Horwood and Pye (1955) noted that the area adjacent to and north of the BBLZ not only has S-folds related to the regional folding pattern but also has Z-folds. The Z-folds in this area only occur along the north side of and proximal to the BBLZ. This led Pye (1952), Horwood

Figure 12. Structure and Stop locations within the Barton Bay Lithotectonic Zone. After Lafrance et al. (2004).

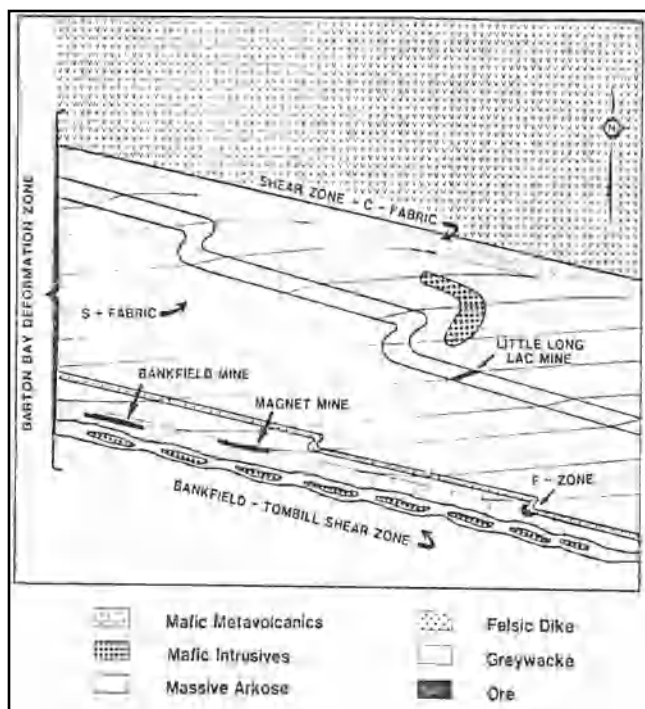


Figure 13. Fault-related fabric development in the Barton Bay Lithotectonic Zone near Geraldton. Modified after Colvine et al. (1984); from Scott (1985).

and Pye (1955) and Buck and Williams (1984) to hypothesize that the second deformational event was related to the development of east-trending faults. This was confirmed by Lafrance et al. (2004) (their D3) and the faulting was attributed to transpression causing dextral shear throughout the region.

Most mineralized occurrences in the Geraldton camp lie in a zone of deformation to the immediate

north of, and genetically linked to, the Barton Bay Deformation Zone (BBDZ) (Fig. 13). This zone of deformation varies from 1000 to 3000 m in total width (Lavigne, 1983; in press), while the crush zone of the Bankfield-Tombill Fault proper ranges from meters to hundreds of meters in width. The crush zone has been intensely silicified (Pye, 1952), carbonatized (Anglin and Franklin, 1985) and contains minor amounts of gold (Pye, 1952).

Numerous Z-folds on various scales were formed in the deformation zone. Auriferous vein systems in the MacLeod-Cockshutt and Hard Rock mines are hosted by one of the Z-folds. This structure plunges shallowly west (Fig. 14; Horwood and Pye, 1955) and is mimicked by minor parasitic folds in the BIF. The parasitic Z-folds are generally less than 5 m in amplitude (Macdonald, 1982, 1983a) and are commonly cut by quartz-carbonate veins subparallel to axial traces. Displacement zones, also subparallel to axial traces, developed at lithologic contacts on isoclinal fold limbs, also host auriferous veins. Veins parallel to axial planes of folds are relatively straight and non-sinuuous whereas those orientated at an angle to axial planes of Z-folds are highly pygmatically folded.

Beardmore Camp

Gold mineralization in the Beardmore area is hosted by quartz-carbonate and sulphide-bearing quartz veins located in foliation-parallel, carbonatized, brittle-ductile shear zones in either clastic metasedimentary rocks of the SMB or in mafic metavolcanic rocks, proximal to

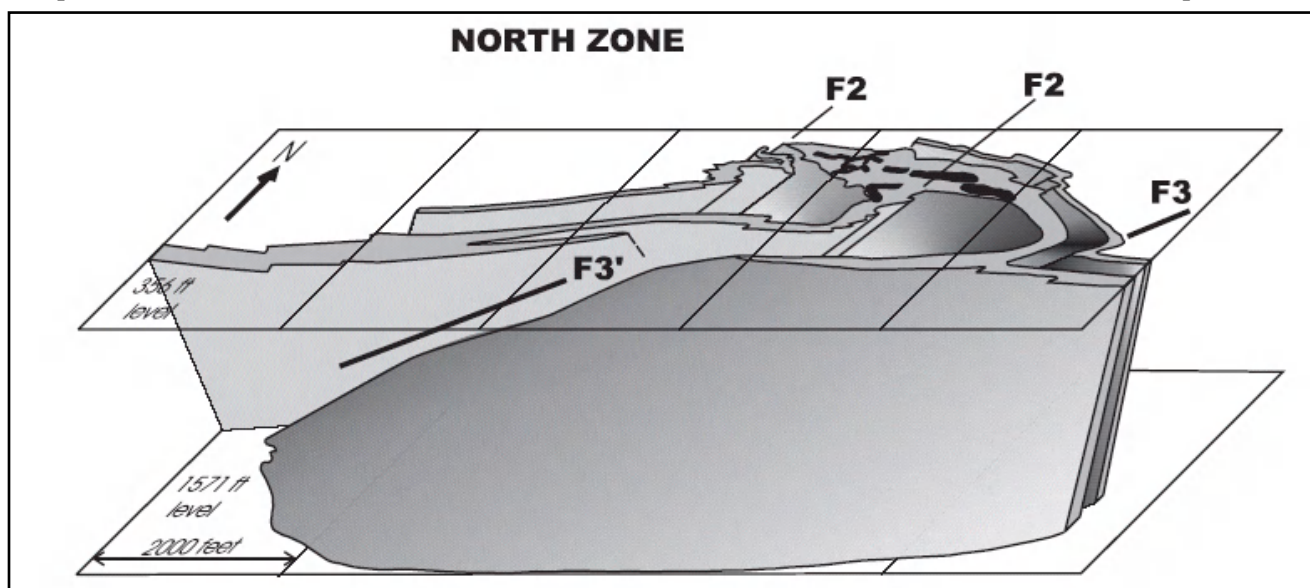


Figure 14. Block diagram of the North Zone at the MacLeod-Cockshutt and Hard Rock mines drawn by Lafrance et al. (2004) using level mine plans published in Horwood and Pye (1955). Diagram shows the overprinting of a S F2 fold by a Z F3 fold on the north limb of the Hard Rock anticline. Ore pods are shown in black.

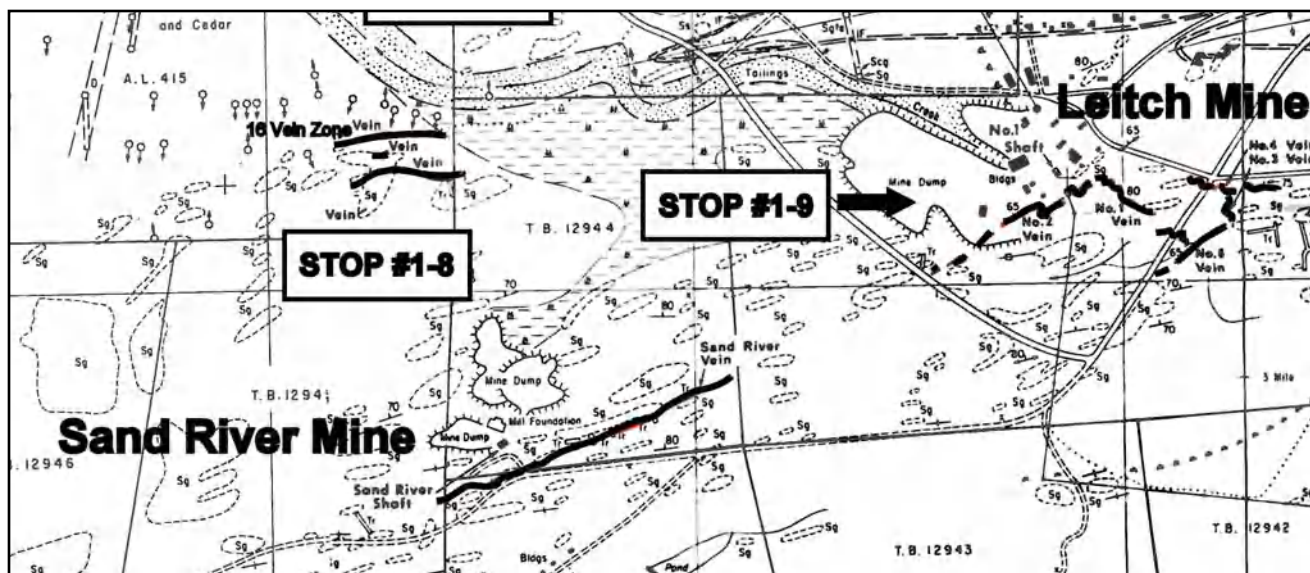


Figure 15. Map showing the surface trace of mineralized quartz veins on the Sand River - Leitch properties. After Ferguson (1967a).

BIF of the SVB (Hart et al., 2002). Gold occurrences are also hosted by quartz veins in post-tectonic intrusive rocks and disseminated sulphides in carbonatized shears in metavolcanic and metasedimentary rocks associated with intrusive rocks.

Regional mapping by Hart et al. (2002) and a composite longitudinal section of the Sand River Mine and Leitch Mine vein system based on that of Ferguson (1967a,b; Fig. 15) indicates that: 1) the Leitch Mine No. 2 vein and the Sand River Mine vein are the same vein; 2) the vein is foliation-parallel and is at a low angle to the regional fold axis at a strike of 240° and a dip of 65° ; 3) the vein contains multiple ore shoots that are a series of dilatant zones controlled by the intersection of two regional fabrics such that the ore shoots plunge moderately steeply at about 320° and have good down-plunge continuity. The veins are hosted by the ductile-brittle shear system which is part of the Standingstone Lake fault (Hart et al., 2002; a.k.a. Standingstone Lake – Watson Lake fault of Bruce and Laird, 1937; Mackasey, 1975; Shanks, 1993), the western extension of the Oxaline Lake – Watson Lake fault (e.g., Williams, 1986). This fault trends about 075° northeast, and is topographically expressed by a swampy valley that includes Standingstone Lake and its associated tributaries, north of Highway 580, east of the Leitch Mine. The fault is manifested by the development of numerous small, northeast-striking, carbonatized shear zones. Lafrance et al. (2004) defined the structure hosting this Sand River vein as being parallel to the axial plane of this fold and S3. The Standingstone Fault has been interpreted to be a

normal, south-side down fault (Carter, 1987) that has experienced multiphase deformation from dip-slip to dextral strike-slip (Williams, 1987).

The Northern Empire Mine is hosted by a series of *en échelon* and composite quartz veins containing differing amounts of pyrite, arsenopyrite, pyrrhotite and chalcopyrite, hosted by carbonatized, sheared metavolcanic rocks and interflow iron formation of the SVB near the Empire fault and the SVB–SMB contact. The quartz veining occurs in brittle features of the deformed iron formation. The Empire Fault trends approximately 060° northeast, is steeply south dipping (but slightly shallower than the average dip of the overturned metavolcanic flows) and is marked by a thin fault gouge of black graphite bordered by a few feet of sheared rock (Benedict and Titcomb, 1948). Previous interpretations (Benedict and Titcomb, 1948; Shanks, 1993) placed the Empire Fault adjacent to the SVB–SMB contact and Shanks (1993) interpreted kinematic indicators to suggest a south side-up sense of motion. Hart et al. (2002) suggested that there are 2 distinct, subparallel faults: the Empire Fault to the east, and the Princess Lake Fault to the west, splaying off the SVB–SMB boundary. Both of these resulted in the formation of zones of brittle deformation in interflow iron formations in mafic metavolcanic rocks of the SVB that host auriferous quartz stringers containing arsenopyrite, pyrite, pyrrhotite and chalcopyrite (Hart et al., 2002).

Co-incidence of structures related to the fault separating the BBA and the PPA northwest of

Beardmore, and the location of post-tectonic, mafic intrusion appear to have controlled the emplacement of quartz veins hosting gold, copper and/or molybdenum mineralization (Hart et al., 2002). The Tyson #1 gold occurrence (Mason and White, 1986) and the Tyson #2 molybdenum occurrence (Hart et al., 2002; Mackasey 1975) are hosted by quartz veins in the post-tectonic intrusion. Copper mineralization is hosted by sheared, carbonatized lapilli tuff within the fault (Hart et al., 2002). The fault trends approximately 075° to 080°, is marked by intense carbonatization and weak to moderate ankerite and calcite-flouding, and may have controlled the emplacement of the intrusion.

Vein Systems

Auriferous veins are typically 2 to 10 cm wide and primarily composed of quartz plus ankerite, dolomite and other carbonate species. Pyrite, arsenopyrite, pyrrhotite, sphalerite and chalcopyrite are locally present. The gold occurs as small blebs and fracture fills within individual quartz and sulphide crystals and in voids between crystals. It has also been observed as fine threads extending along cleavage planes in carbonate minerals (Armstrong, 1943).

The auriferous veins often show evidence of brecciation and multiple generations of quartz. The brecciated zones are richer in gold than non-brecciated areas (Bruce, 1935, 1937a,b; Horwood and Pye, 1955). Crack-seal-textured ore veins, which predominate at the Leitch and Sand River mines, are indicative of successive periods of dilation and fluid influx. These veins contained minor and varied amounts of pyrite, arsenopyrite, chalcopyrite and tetrahedrite, scheelite

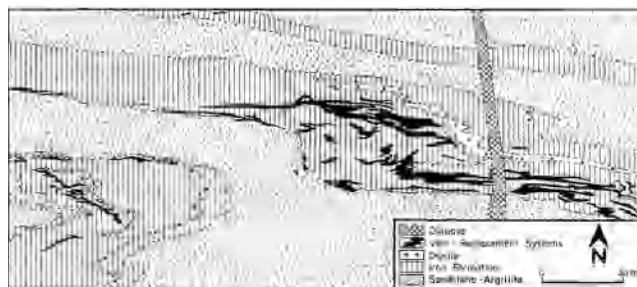


Figure 17. Plan of the North Zone (No. 30 Vein system), 250-foot level, Hard Rock Mine. After Horwood and Pye (1955).

with septa of sericitic and carbonatized wall rocks.

Alteration Halos

Substantial alteration halos surround auriferous veins (Fig. 16). Carbonatization (ankeritic dolomite) dominates in host lithologies with low concentrations of iron-bearing minerals (e.g., clastic sedimentary rocks) and fracture zones associated with both major and minor faults. Siltstones have been found to contain up to 20% CO₂ (Anglin and Macdonald, 1984). Carbonatization is ubiquitous in the clastic sedimentary rocks at the MacLeod-Cockshutt and Hard Rock mines. However, 300 m along strike away from this structure, carbonate alteration is rare (Macdonald, 1983a). Siderite supercedes ankeritic dolomite as the major carbonate phase within 5 cm of the vein margin.

In addition to carbonatization, pyritization occurs in iron-rich rocks (e.g., BIF) adjacent to veins. These pyritic halos not only replace the magnetite/hematite laminae in BIF but also sandstone and siltstone laminae in the Geraldton area. The pyritization generally extends only centimetres to tens of centimetres away

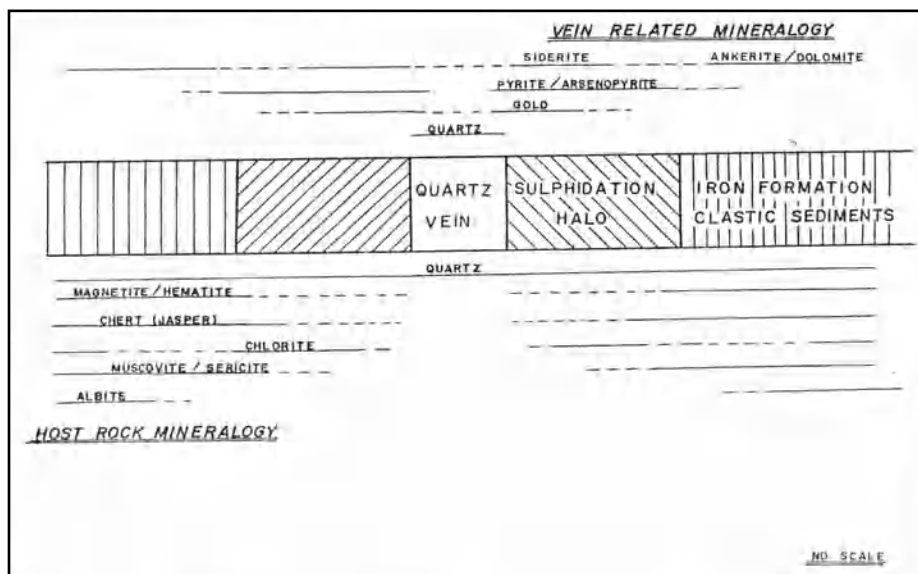


Figure 16. Host rock, alteration halo and vein mineralogy for veins in iron formation-siliciclastic assemblages. From Scott (1985).

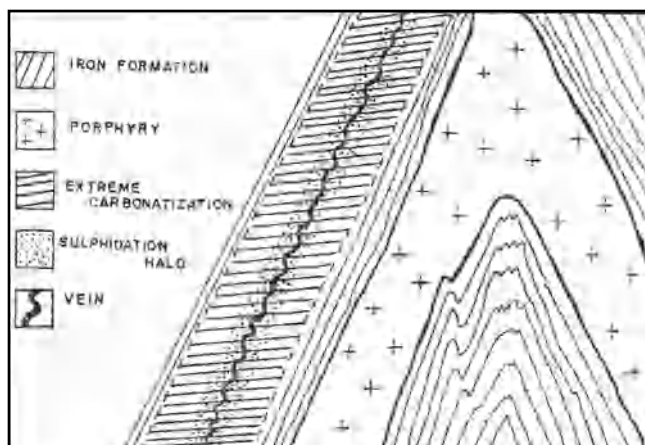


Figure 18. Schematic representation of vein-related alteration, Hard Rock Mine (Scott, 1985); carbonatized zone is approximately 1 m wide.

from individual veins. However, where densely packed systems of veins exist sulphide bodies meters by tens of meters in size have developed (e.g., MacLeod-Cockshutt-Hard Rock mines; Horwood and Pye, 1955; Figs. 17 and 18).

The vast majority of the gold associated with the vein systems in BIF is contained within sulphides in alteration halos. The gold generally occurs as sub-microscopic particles and minute blebs within the pyrite (Pye, 1952; Horwood and Pye, 1955; Mackasey,

1976; Mason and McConnell, 1983; Scott, 1985). Concentrations of gold may reach in excess of 15,000 ppb. In contrast to this, the surrounding, sulphide-free BIF consistently contains less than 20 ppb gold (Anglin and Macdonald, 1984; Fig. 19).

Generation of Hydrothermal Fluids and Timing of Mineralization

Past studies have suggested the gold-bearing fluids in the Geraldton camp were derived either from a granodioritic body (Croll Lake Stock) lying to the immediate east of the area (Horwood and Pye, 1955; Macdonald, 1983b) or by metamorphic redistribution of gold contained in the banded iron formation (Boyle, 1976; Mackasey, 1975). Arguments can be made against gold originating in the BIF. Sulphide-free BIF in the area generally contains less than 20 ppb gold (Anglin and Macdonald, 1984). This amount is much lower than the average values of 61 ppb Au for high-Fe tholeiites and 78 ppb Au for volcanic-associated BIF which were obtained from a volcanic pile lying in Superior Province northwest of the Beardmore-Geraldton greenstone belt (Cowan and Crockett, 1980). Obviously, if volcanic and related chemical sedimentary rocks in the map area have similar gold contents (and sparse data indicate they do) they would provide a more adequate source of gold. In addition, alteration halos around the veins indicate the fluids were in marked disequilibrium, both in composition and Eh, with the BIF signifying that they were derived from an external source.

An alternative model to those outlined above is the genesis of the fluids through metamorphic de-watering. Studies conducted by Horwood and Pye (1955) indicated the fluids which deposited the auriferous veins contained Ag, As, Au, C, Ca, Co, Hg, Mg, Pb, S, Si, Te, W, Zn and possibly Fe and Ti. This assemblage is similar to the elements enriched in and near auriferous veins studied by Phillips and Groves (1983), Kerrich and Hodder (1982) and Kerrich (1981). These authors interpreted the fluids which formed those deposits to have been generated during burial and metamorphic de-watering of a volcanic pile.

The spatial association between the east-striking fracture systems and the gold mineralization has led previous authors to believe that the faults and related minor structures served as passageways along which the fluids moved (Horwood and Pye, 1955; Macdonald, 1984; Anglin and Franklin, 1985). Structural control of

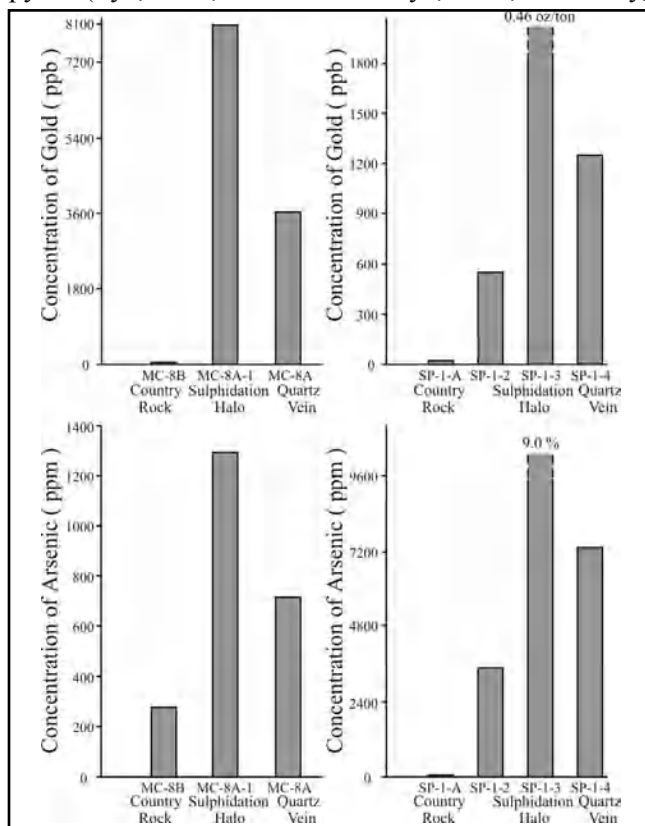


Figure 19. Gold and arsenic distribution associated with veins at MacLeod-Cockshutt (MC) and Solomon's Pillars (SP). From Scott (1985).

vein systems, and by inference, fluid migration paths, is common in the Superior Province (Rigg and Helmstaedt, 1981; Fryer et al., 1979; Stott and Schnieders, 1983; Poulsen, 1983; Durocher and Hugon, 1983; Hodgson, 1983). The fluids were probably generated at depth and utilized the permeability developed in deformed zones to gain access to higher crustal levels. Permeable fault zones provide conduits to the fluids allowing them to move upwards towards lower-pressure/temperature zones.

Episodic and protracted development of dilatant zones and shearing is indicated by the open space-filling nature of the veins and their brecciation. Fluids migrated towards dilation zones enveloping fault zones, driven by a pressure gradient and seismic pumping (Sibson et al., 1975). The same model has been put forward to explain structurally controlled gold veins in the western Wabigoon Subprovince (Poulsen, 1983).

The development of abundant carbonate and iron sulphide in the reaction halos around veins indicates the fluid could not have been very acid (Krauskopf, 1967; Boyle, 1969) and was probably near neutral (Phillips and Groves, 1983). The formation of massive amounts of carbonate minerals also denotes a very CO₂-rich fluid. This conforms to data from other studies which advance H₂O- CO₂-dominated fluids as of major importance in the formation of Archean lode gold deposits (Kerrick, 1981; Kerrich and Hodder, 1982). Fluid inclusion studies of the MacLeod-Cockshutt deposit also indicated a CO₂-rich fluid at a temperature in excess of 380°C was involved in forming at least some of the veins (e.g., Macdonald, pers. comm., 1985). Recent research has suggested that major gold deposits may be the result of prolonged mixing of two fluids: a hot, reducing, orogenic fluid that migrates along major regional fault structures; and a colder, oxidized, gold- and base metal-enriched, magmatic fluid associated with late, mantle-derived plutons (c.f. Neumayr et al., 2004; Rogers et al., 2004).

Gold mineralization in the Beardmore area has not been studied to the same degree as in the Geraldton area, but much of the preceding discussion is probably applicable there as well. However, there are a number of points specific to the Beardmore area. An intermediate to felsic intrusion comparable to the Croll Lake Stock has not been identified. There are a number of post-tectonic mafic intrusions, and one of these intrusions along the faulted BBA-PPA boundary hosts gold and molybdenite mineralization (Tyson occurrences) and

there is copper mineralization in the adjacent fault (Hart et al., 2002). Mafic intrusions of comparable composition are located to the west of the Sand River Mine, south of the interpreted trace of the Standingstone Lake Fault, but intrusions are not spatially associated with mineralization in either the Leitch or Sand River mines. The Standingstone Lake Fault may be part of the fault system separating the SMB and CVB as it, in part, parallels this contact, but it also is the approximate location of the reversal in top directions in the metasedimentary rocks of the SMB (Hart et al., 2002). Axinite, a borosilicate similar to tourmaline, is located in brittle structures in mafic metavolcanic rocks along strike with the Leitch – Sand River vein to the east, and along the BBA-PPA contact to the north (Hart et al., 2002). Although axinite is not associated with gold mineralization, as in the Chibougamau (Dubé and Guha, 1993) and Cadillac areas (Bardoux et al., 1990), it may be a late-stage mineral associated with an intrusion that may have supplied hydrothermal fluids during the gold-mineralizing event. However, the axinite may be unrelated to the gold mineralizing event and reflecting an unexposed pegmatite or the intrusion of the diabase sill (e.g., Ozaki, 1972).

Gold mineralization at the Leitch and MacLeod-Cockshutt mines, the two largest, past-producing gold mines in the Beardmore–Geraldton Belt, is associated with D3 brittle shear zones and folds, overprinting regional F2 folds (e.g., Lafrance et al., 2004). The plunge of the ore zones is parallel to F3 fold axes and to the intersection of D3 shear zones with F2 and F3 folds. At the deposit scale, the plunge of the ore zones is similar in orientation to the plunge of F3 axes (e.g., North Zone, MacLeod-Cockshutt Mine) and to the intersection of D3 shear zones with F3 folds (e.g., Leitch Mine; Hart et al., 2002; Lafrance et al., 2004) and F2 folds (e.g., F Zone, MacLeod-Cockshutt Mine; Horwood and Pye, 1955). Lafrance et al. (2004) provided an estimate for the age of syn-D3 gold mineralization ranging from 2686 Ma (i.e., the age of the pre-D3 Ottortail granite in the Rainy Lake district) to 2640 Ma (i.e. the minimum age for the formation of D3 structures and high-temperature metamorphism in the Quetico subprovince; Zaleski et al., 1999). On a subprovince scale, regional folds that are cut by D3 dextral shear zones are promising targets for discovering the next generation of large gold deposits (ibid).

Stops

Preface

The Beardmore-Geraldton area has been the subject of numerous field trips and published and unpublished field trip guide books, including those by Mason et al. (1985), Fralick and Barrett (1991) and Williams and Stott (1991). Bear in mind that when visiting active exploration or mine properties, permission must be granted by the property owner. Current ownership information can be obtained from the Resident Geologist's Office in Thunder Bay. Please exercise caution along highway, road right-of-ways and lake shores.

Field Trip Road Log

Stop	Locality	km
<i>Beardmore - South</i>		
	Intersection of Hwy. 11 & CNR (Beardmore)	0
1-3	Blackwater fault/pillowed volcanic rocks	1.3
1-2	Quetico metasedimentary rocks	5.7
1-1	Sibley-Quetico unconformity	13.6
<i>Beardmore - West</i>		
	Junction of Hwy. 11 & 580	0
	East Leitch turn-off	5.8
	Longlac Superior strips	3.7
	72 Road turn-off	5.3
1-9	Leitch mine turn-off	7.1
	Sand River Mine turn-off	7.4
1-8	Sand River 16 Vein Zone	8.3
1-7	Leitch BIF	8.4
1-6	Eva Creek: conglomerate/BIF	9.3
1-5	Eva Creek: pyroclastic rocks	9.6
	Million Dollar Road turn-off	10.4
	High Hill Harbour turn-off	12.0
1-4	Poplar Lodge pillowed flows, G & G Goodman driveway (access to lakeshore)	14.0

Geraldton gold camp

	Junction Hwys 11 & 584	0.0
2-1	MacLeod - Cockshutt headframe	0.0
2-2	Porphyry Hill	
2-3	Glory Hole BIF	
	Junction Hwys 11 & 584	0.0
	Ashmore/Errington boundary	0.5
	Mosher Mine turn-off	1.3
2-4	Conglomerate	1.3
2-5	Gabbro	1.6
	Gravel Pit turn-off	2.3
	Powerline crossing	3.8
	Wintering Road turn-off	4.7
	Powerline crossing	5.9
	McClellan Strip	6.4
	Magnet Mine turn-off	8.5
	Bankfield Mine turn-off	10.6
	Key Lake/Malouf Vein	11.8
	Wild Goose Lake Park	27.0

Jellicoe-Kinghorn Road

2-8	Southern Sedimentary belt sandstone	39.6
	Jellicoe Post Office	47
2-9	Pillowed mafic flows	47.2
	Junction Hwy 11 & Kinghorn Road	0
	Ara Lake sign	1.7
	CNR tracks	2.9
	Sturgeon River Bridge	5.3
	6 km Road turn-off	6.0
2-6	Missing Link extension	7.2
	Missing Link turn-off	7.7
	Missing Link parking spot	+0.4
2-7	Missing Link (NE corner)	+0.2

Stop 1-1 - Archean – Proterozoic Unconformity

UTM coordinates - 0423431E 5483881N

The angular unconformity between east-trending, steeply dipping Neoarchean Quetico metasedimentary rocks and flat-lying Mesoproterozoic Sibley Group



Figure 20. Angular unconformity between steeply dipping Quetico metaturbidites and overlying Sibley Group sandstone-carbonate units, Highway 11 (Stop 1-1).

sedimentary rocks is exposed on both sides of Highway 11. The phyllitic Quetico rocks display layer-parallel slickensides (bedding-parallel fault?).

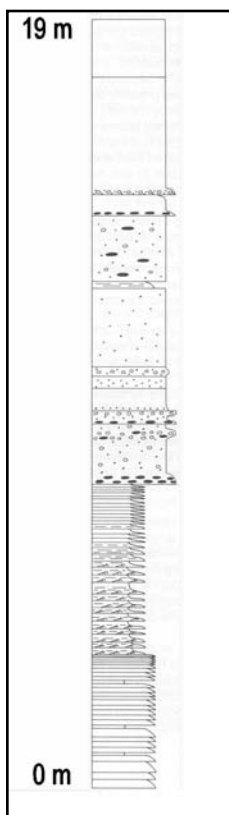
This outcrop consists of metasedimentary rocks unconformably overlain by a thin layer of Sibley Group sandstone (Fig. 20). The underlying rocks were folded to a near vertical position during the Kenoran Orogeny, prior to Mesoproterozoic deposition of the Sibley sedimentary units. The Sibley here consists of an upper, 30 cm thick, orthoquartzite separated from the Archean basement by 0 to 30 cm of silicified carbonate. The carbonate occurs in fractures in the substrate up to 60 cm below the paleosurface. This gives the distinct impression that the carbonate is a well-developed, caliche zone. Caliche soils form in semi-arid settings where the evaporation rate exceeds the precipitation rate. Irregular laminations in this layer are up-buckled in places, forming domical structures, probably the result of expansion during crystallization. It is difficult to say much about the overlying orthoquartzite. It is not even known if it is wind- or water-lain. Similar, thin sandstone layers occur both to the east and west of Lake Nipigon, separating basement from Mesoproterozoic (Keweenaw) diabase sills. They are correlated with the Sibley simply on the basis of their probable Mesoproterozoic age.

Stop 1-2 - Quetico Metasedimentary Rocks

UTM coordinates - 0428530E 5489570N

Most of the Quetico consists of an unstructured submarine ramp system (i.e., an environment where large, long-lasting channels do not exist and sediment

Figure 21. Stratigraphic column of Quetico metaturbidites, Stop 1-2 (Fralick et al., 1992)



delivery routes switch rapidly, producing a chaotic structuring to the deposits). This outcrop provides one of the few examples of structured channel outbuilding in this ramp setting. A thinning- and fining-upward succession of Bouma ADE, CDE and DE turbidites is sharply overlain by a thick-bedded unit of ABCD turbidites with abundant clay rip-ups and extra-basinal pebbles (Fig. 21). The lower, 8 m-thick, fine-grained assemblage is the upper portion of a subaqueous channel feeding sediment from the Beardmore-Geraldton forearc basin to the Quetico trench. It is erosively overlain by the basal section of the next channel complex.

The phyllitic rocks that occur at the northern margin of the Quetico Subprovince south of Beardmore are strongly sheared, exhibiting rust-spot elongation, quartz fibre lineation, and crinkles, some of which can be attributed to non-parallel, cleavage-bedding and cleavage-vein intersections (Williams and Stott, 1991). None of these intersections appear to have reliable or consistent tectonic significance.

Stop 1-3 - Blackwater Fault / Pillowed Metavolcanic Rocks

UTM coordinates - 0430174E 5493162N

Deformed, pillowed mafic metavolcanic rocks of the Southern Volcanic Sub-belt are exposed along the sides of Highway 11 just south of Beardmore. They form the southern margin of the Beardmore-Geraldton greenstone belt and the eastern Wabigoon Subprovince in this area, where it is in fault-bounded contact with Quetico metasedimentary rocks to the south. The Blackwater River Fault (BRF) is a regional-scale structure which separates the two subprovinces. It is likely a dextral, transcurrent fault, similar to other sub-belt-bounding structures in the Beardmore-Geraldton belt. It has been traced for over 100 km and underlies the valley occupied by Camroad Creek, a few hundred



Figure 22. Deformed pillow basalt near Blackwater Fault, Highway 11 (Stop #1-3).

metres south of this field trip stop. Immediately north of the BRF, Shanks (1993) noted a series of sub-vertical, non-penetrative planar surfaces, interpreted to be either slip surfaces or C-planes. Estimation of the finite strain ellipsoid shape by Williams and Stott (1991) using pillow shapes and selvage thickness indicated plane strain. Localized, small-scale c-s fabrics were noted by Williams and Stott (1991) in schistose selvages, indicating a north over south sense of displacement.

Light grey-green weathering basalt flows are foliated at approximately $240^{\circ}/80^{\circ}$ north. They are locally phyllitic, friable and gossanous. Despite flattening and aspect ratios of 3:1 to 10:1, pillows give northward top directions. Pillows range in size from 0.5 to 1.0 m and have recessively weathered selvages (Fig. 22). They contain flattened vesicles near their rims and are locally variolitic. Isolated patches of autoclastic(?) breccia contains 0.10 to 0.25 m-sized, lenticular fragments. The flows are more massive to the south and are crosscut by rusty, “bull-white” quartz veins. Joint surfaces are coated with quartz and calcite. A rusty quartz vein occupying the sheared contact between basalt and sandstone to the north was sampled but returned no gold nor silver. A 3m wide diabase dyke trends $\sim 155^{\circ}$ and cuts the supracrustal rocks.

Stop 1-4 - Poplar Lodge Pillowed Metavolcanic Rocks

Private property access to lake shore; please ask permission from cottage owners

UTM coordinates - 0421034E 5499814N (northern end of section), 0421089E 5499665N (southern end of section)

Weakly deformed, northward-younging, pillowed

intermediate to mafic metavolcanic rocks of the Central Volcanic Sub-belt are exposed along the shoreline of Lake Nipigon, just north of Poplar Lodge. They have been described by numerous workers, including Coleman (1907), Laird (1937), O’Brien (1985), Tomlinson (1996) and Hart et al. (2002).

Massive, vesicular/amygdaloidal and/or pillowed basalt/andesite, heterolithic breccias, as well as porphyritic and flow-banded units were described by O’Brien (1985). Individual flows in the CVB are commonly a few metres thick, with a few examples that are tens of metres thick with pillows that are generally weakly deformed to undeformed. In this outcrop, the pillows may be greater than 1 m in diameter (Fig. 23). Pillow selvages in these flows are generally thicker than in the mafic metavolcanic rocks. Internal features include amygdaloidal and vesicular flow laminae textures and feldspar phenocrysts, with some pillows in flows to the south containing concentric flow laminae defined by the alignment of vesicles approximately 1 mm in diameter.

Pillow breccias occur as a transition from the pillowed flows to the hyaloclastite-rich, pillow fragment breccias, and are exposed on the western end of the outcrop as it dips into the lake. The pillow breccia portion of the units is generally 1 m wide, with the pillows very abruptly diminishing in size away from the main body of the flow. A good example of this type of breccia is exposed in the bay to the south, ~ 150 m from this location, but is only accessible when the water levels are low. Pillow fragment breccia forms the major part of the unit, and is generally a few metres in width adjacent to the flow front. This transition is visible near the waterline. The breccia thickens to the



Figure 23. Pillow basalt, Poplar Lodge, Lake Nipigon (Stop 1-4).

west, which may suggest that the direction of flow was towards the west. Fragments generally consist of sub-angular parts of pillows and rounded fragments or amoeboid fragments and small pillows in hyaloclastite matrix. Although the pillow breccias have been classified as volcanoclastic rocks, a portion of the units could be classified as hyaloclastite or rocks that are the result of fragmentation of the magma due to interaction with water (Cas and Wright, 1987; Batiza and White, 2000). There is no depth restriction to the formation of the pillow breccias, as the breccias have been observed in both shallow water and on seafloor sites.

Heterolithic breccias consist of predominantly 0.5 to 10 cm sub-angular to angular clasts of intermediate metavolcanic rocks, but minor clasts of feldspar-porphyrific, felsic metavolcanic rocks, cherts and mafic metavolcanic rocks overlie pillow breccias in a few locations along the lake shore. Generally, the breccias lack any internal structure. Although some areas appear to have bedding structures, they generally cannot be traced for any distance. The matrix is generally highly chloritized and some of the smaller, more porous clasts may have been totally chloritized; reaction rims along clasts margins may be developed

Major element and trace element analysis indicated that the pillowed flows are tholeiitic basalt and andesite (O'Brien, 1985), characteristic of mid-ocean ridge basalt (Tomlinson, 1996; Hart et al., 2002).

Stop 1-5 - Eva Creek Pyroclastic and Sedimentary Rocks

UTM coordinates - 0424037E 5498780N

Dacitic pyroclastic rocks along the southern margin



Figure 24. Deformed felsic pyroclastic breccia, Highway 580 (Stop 1-5).

of the CVB are exposed on a small ridge 50 m east of Highway 580 (Fig. 28). A massive, aphyric, monomictic tuff-breccia has an iron-carbonatized matrix and is cut by epidote veins (Fig. 24). It is foliated at 065°/78° south and hosts narrow shear zones with crenulated fabrics and quartz veins. There is a suggestion of flattened pillows at the northern edge of the outcrop. A north-trending diabase dyke cuts the supracrustal rocks.

An outcrop on the west side of Highway 580, across from Stop #1-5, exposes typical, medium-bedded turbidites of the SMB. These consist predominantly of A, B, D and very minor E divisions. The minor E division reflects very little clay in the sediment transport system, a logical consequence of sediment feed being mostly derived from active subareal, pyroclastic volcanism to the north in the Onaman-Tashota terrain (Fralick, 2003). The interval of time between eruption, erosion and sedimentation was not sufficient to produce significant amounts of chemical weathering (i.e., clay). The relative lack of C (rippled) divisions is more difficult to account for. One possibility is that higher ocean temperatures resulted in a thinner, laminar sublayer which fine sand was capable of disrupting and, since ripples need a laminar sublayer to form, the C division was suppressed. The non-regular bed thickness variation up through this section reflects rapid changes of the position of sediment feed areas into the subaqueous portions of the basin, resulting in a somewhat chaotic interlayering of thin and thick turbidites.

Stop 1-6 - Eva Creek Conglomerate / Banded Iron Formation

UTM coordinates - 0423925E 5498505N

Exploration of the local iron ranges began in earnest around 1900 when several claims in what was to become the Sand River and Leitch mines area were staked and surveyed (Laird, 1937). In 1919, the Lake Superior Ore Company was formed to option and drill a portion of the Central range. At this time, the Kokoko camps on claim A.L. 413 were built. Drilling carried on until 1920. Coleman (1907) reported that drilling returned assays in the 40 to 50% Fe range, with negligible amounts of deleterious sulphur and phosphorus. Laird (1937) characterized the iron deposits of the Central range:

“The beds have been crumpled and intricately folded,

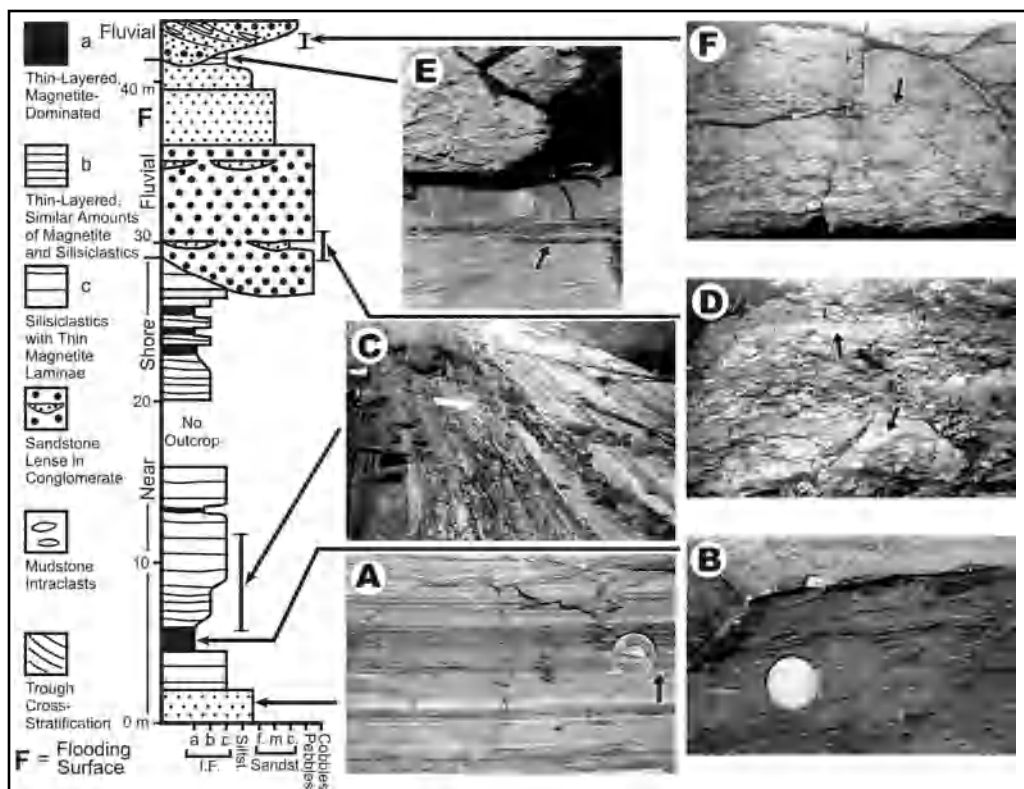


Figure 25. Stratigraphy, Eva Creek section (Stop 1-6).

with the result that they have the appearance of great thickness in places. The formation is well-banded and consists of thin layers of red jasper alternating with steel-grey hematite, the whole being irregularly interbedded with dark slaty bands. The jasper-hematite bands are not commonly more than a few feet in width and in places they occur as mere streaks in the slaty material. Magnetite is not a prominent constituent of this band, but in places it is present in quantities sufficient to cause considerable magnetic disturbance.”

This is one of the best examples of the stratigraphic position of BIF in the belt. The lower portion of the section consists of interlayered magnetite, siltstone



Figure 26. Conglomerate associated with BIF, Eva Creek (Stop 1-6).

and sandstone in various ratios from chemical sediment-dominated to clastic-dominated (Fig. 25). This is erosively truncated by a conglomerate. The clast-supported, polymictic conglomerate (Fig. 26) contains diffuse zones of larger and smaller clasts, lenses of medium-grained sandstone and crescent scours and sand shadows along internal surfaces where sandstone lenses are concentrated. This is overlain by a thick, medium-grained sandstone, which is, in turn, succeeded by a clastic-BIF succession. A coarse-grained sandstone-conglomerate assemblage erosively cuts down into this package, removing the upper metre in places. This outcrop highlights the juxtaposition of BIF and erosively based, fluvial channels. The fluvial system was periodically extending into a nearshore dominated by thin, graded siltstones and sandstones with iron oxide laminae. The depositional architecture is composed of a series of prograding parasequences in the littoral zone with BIF developed on flooding surfaces. Commonly iron-rich BIF directly overlies the flooding surface and is upwardly transitional into more clastic-dominated units. These may be erosively incised by the next channel assemblage. Off-delta, deeper-water turbidite units do not contain BIF. This strongly indicates that BIF deposition was the result of photosynthetic oxidation in the nearshore, producing an oxygenated oasis near nutrient-rich waters emanating from river mouths.

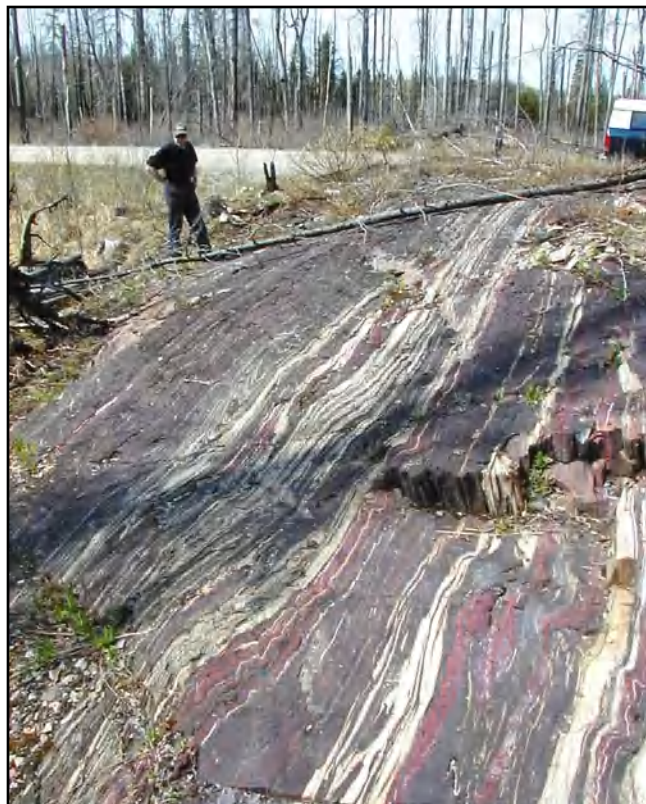


Figure 27. Deformed BIF, Highway 580 (Stop 1-7).

Stop 1-7 - Leitch Banded Iron Formation

UTM coordinates 0424600E 5497867N

This is an excellent exposure of oxide-facies BIF interlayered with siliciclastic layers. Unlike many BIF's in this region, hematite, not magnetite, dominates this exposure. At first glance the rocks give the impression that there is intense shearing along the cleavage direction disrupting the layering (Fig. 27). Closer examination may lead to a different conclusion. The sharp-sided, clay-poor, medium-grained sandstones form lenses (averaging 1 cm thick), cutting saucer-shaped scours into the 1 mm-thick, stacked iron oxide laminae. That is, the sandstones are lenticular to flaser-bedded. Some lenses appear to be ripple-laminated, although the cleavage cutting through at a similar angle makes identification problematic. This type of siliciclastic depositional system is typical of distal portions of distributary mouth bars in deltaic settings. The interlayering of iron formation indicates relatively rapid precipitation of the chemical sediments in a nearshore environment.

Stop 1-8 - Sand River Mines Property - #16 Vein system

UTM coordinates - 0424339E 5497563N

The following history has been modified from a summary reports prepared by Bevan (2004) for Roxmark Mines Limited (who are the current property owners) and Mason and White (1986). Laird (1937) provided the first geological description of the property and its exploration activity.

History of the Sand River and Leitch Mines

The first discovery of gold in the Standingstone River area west of Beardmore was made on what became the Sand River Mine property in June, 1934 by Russell Cryderman (Laird, 1937). The Discovery or Number 1 vein averaged only 13 inches in width, but was remarkably persistent along strike, being exposed in cross-trenches and surface pits over a length of 2100 feet. The vein was traced to the northeast onto the Leitch Mine in 1935 (Fig. 28). Stripping and trenching were carried out by the newly incorporated Sand River Gold Mining Company on the Number 1 vein. Servicing was by two three-compartment shafts, 2950 feet apart in a northeasterly direction: the westerly Sand River shaft to 2656 feet and the Leitch main shaft to 3006 feet, with a winze from the 19th (2875-foot) level to the 30th (4525-foot) level. A road connecting the two properties with Beardmore was constructed in 1936.

Production started with milling facilities at 75 tons per day at both properties in 1937. Between 1937 and 1942, Sand River processed 157,870 tons at a recovery grade of 0.32 ounce gold per ton, yielding 50,065 ounces gold, to their 9th (1150-foot) level. Mining was by conventional shrinkage stoping with an approximate width of 3.0 feet. The Sand River operation ceased in 1942. Leitch Mines processed 80 tons per day, milling 176,535 tons at a recovery grade of 0.71 ounce gold per ton in the same period with a high-grade resuing approach.

Northern Empire Mines Limited, a subsidiary of Newmont Mining Corp., began milling operations in Beardmore in 1934 with a maximum of 186 tons per day in 1939 and ceased in 1941. They then acquired the Sand River property on a ten-year lease. They drifted out on two of the three un-mined Sand River levels above the diabase sill. Sufficient encouragement led them to collar a winze on the Sand River 11th or 1450-foot level, designed to sink through the diabase.

The winze was stopped at 1506 feet when war broke out in Europe. Northern Empire resumed operations at Sand River in 1944 as the Undersill Gold Mining Company. Undersill de-watered and deepened the Sand River shaft through the flat-lying, 600-foot thick diabase sill to 2180 feet. The shaft was bottomed at 2656 feet. They explored the 17th or 2610-foot level with a shaft crosscut of approximately 450 feet and 2340 feet of drifting, of which 1300 feet was eastward to the Leitch boundary, and 1040 feet was westward. The 1947 Leitch Gold Mines Limited Annual Report stated that Northern Empire's Undersill Gold Mining Company had encountered "good ore" on the Sand River 17th (2610-foot) level, an extension of the high-grade Leitch Number 2 vein. Northern Empire, unable to capitalize on its 1944 to 1947 development with gold at US\$35 per ounce, closed down the Undersill Gold Mining operation in 1954 and concentrated development at its Magnet Mine, near Geraldton. Leitch acquired the Sand River property in the same year to protect the depth extension of their steeply west-plunging Number 2 vein. Leitch developed the Sand River property by an internal winze at the Leitch Mine from their 19th (2875-foot) level to the 30th (4525-foot) level. Leitch's production from the Sand River - Undersill Gold Mining Company's holdings totalled 376 283 tons at a recovery grade of 1.11 ounces gold per ton from 1955 to mine closure in 1965. This was an appreciable improvement of grade with depth from 530 112 tons at 0.79 ounce gold per ton from 1938 to 1954 above the 19th level. Environmental requirements were completed on the Leitch Mine and the Sand River property was returned to the original owners in 1968.

In 1940, exploration by Halport Gold Mines outlined a parallel vein system, named the #16 vein system. The following table covers the history of this discovery.

Historical Summary - #16 Vein System

1940 - Halport Gold Mines located a number of surface veins on Claim AL 415 at the eastern end to the boundary with Sand River Claim TB 12944.

1944 - Leitch acquired the Halport holdings.

1945 - 22 diamond drill holes totalling 16 510 feet were drilled.

1950 - A long crosscut on the 8th level was initiated because surface drilling had indicated ore at this horizon.

1951 - Further crosscutting and some drilling was

carried out.

1952 - Further crosscutting and drifting to open up the #16 vein. Drilling of 51 underground holes totalling 11,116 feet, mainly on the 8th level.

1953 - Further drilling and stoping on the #16 vein - production and milling of 3,598 tons grading 0.56 oz. Au/ton.

1969 - Teck Corporation (together with Highland-Bell Limited and F.E. Hall) purchased control of the Leitch Gold Mines property.

1971 - The entire Leitch assets were acquired by Teck Corporation.

1946 - Teck Corporation carried out exploration work until 1981 over the Leitch property, including line cutting, VLF-EM magnetometer surveys, 393 holes of overburden drill sampling and 29 short diamond drill holes totalling 3,454 feet and covering a strike length of 1,400 feet. One hole was abandoned.

1987 - Sand River-Cryderman diamond drilling program. 10 holes totalling 2,030 feet were drilled on Claim TB 12944 covering the #16 vein to the east.

1988 - Teck Corporation drilling program in joint venture with San Paulo Explorations Inc. 24 diamond drill holes totalling 22,098 feet, of which three were abandoned, one hit the open stope and one intersected the vertical diabase dyke and was terminated. Airborne and ground geophysical surveys and reverse circulation drilling were also undertaken until 1989.

2003 - Acquisition of Sand River property by Roxmark Mines Limited; Advanced Exploration project initiated.

In 2003, Roxmark Mines Limited acquired the Sand River, Leitch and East Leitch properties from Rio Fortuna, Teck Cominco and Kinross, respectively (News Release, Roxmark Mines Limited, November 28, 2003) and began an Advanced Exploration program. In 2004, two vein systems were exposed by trenching, stripping, channel sampling and stockpiling at the East Leitch and Sand River #16 zones, respectively. Stripping of these areas had been recommended in preparation for taking bulk samples (ibid). Geological mapping and sampling was planned to gain a better understanding of the mode of occurrence and identify other possible veins and extensions in the #16 vein system. Further exploration in the form of diamond drilling in five areas was recommended as follow-

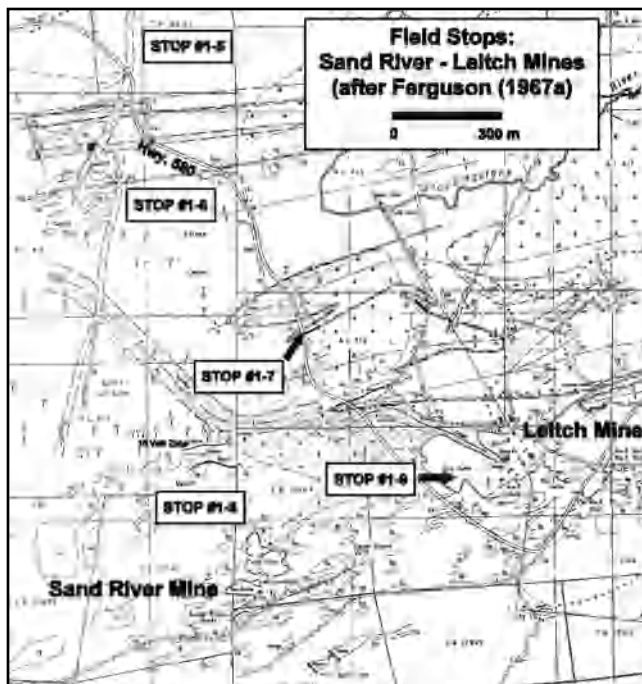


Figure 28. Geology of the Sand River-Leitch mines area after Ferguson (1967a), showing Stop locations.

up work in preparation for underground exploration and development. Bulk samples will be processed at Roxmark's 200 ton-per-day Northern Empire Mill in Beardmore.

Roxmark reported the results of samples taken from the #16-vein system on the Sand River Mine property (ibid, December 16, 2004). The Company is continuing with its plan for the surface bulk sample program and in particular, the in-fill sampling on the #16-1 and #16-5 veins. Channel samples were cut every three feet. All composite assays used a minimum width of 1.0 feet. The #16-1 West Vein returned an average assay of 1.737 ounces Au per ton over a strike length of 42 feet. The #16-1 East Vein returned 0.347 ounce Au per ton over 37 feet. The recent assay

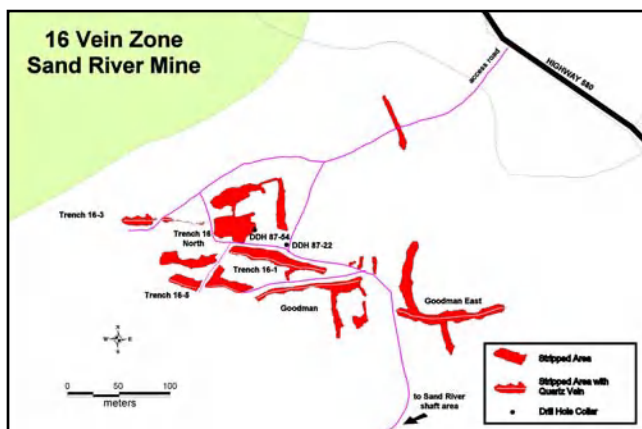


Figure 29. Map of stripping and veins, Sand River 16 Vein zone (Stop 1-8).

results confirm the historical sampling data completed by Teck Corporation. Furthermore, the current assay results confirm the historical drill holes that intersected the #16-1 Vein. Recent Roxmark assays of the #16-5 West vein were 1.058 ounces Au per ton over a 26.5 foot strike length. The #16-5 East Vein returned 1.094 ounces Au per ton over a strike length of 35.0 feet.

Geology of the #16 Vein Area

The Leitch-Sand River area (Figs. 15 and 28), most recently mapped by Hart et al. (2002), is predominantly underlain by clastic metasedimentary rocks of the SMB. Thickly bedded, southward-younging, overturned, feldspathic sandstone and siltstone have been well-exposed by recent stripping operations in the vicinity of the #16 North zone (Fig. 29). Bedding has east-northeasterly to east-southeasterly orientations with steep, northerly dips. Sub-parallel shear zones are distinguished by well-developed, slaty S3 cleavage, as well as local sericitization and carbonatization. Oblique cleavage – bedding relationships are locally exposed. Cleavage is refracted through interbedded siltstone and sandstone beds (Fig. 30).

A number of auriferous quartz veins have been exposed in the immediate area (Figs. 28 & 29). Visible gold occurs within fractures in quartz veins and most commonly is found along thin seams or septa of sericite and chlorite, along with arsenopyrite, pyrite and ankerite in composite, "crack-seal"-textured veins. Tetrahedrite and sphalerite are also reported to occur in these fractures. Pink-beige scheelite is relatively abundant.

The Leitch Mine No. 2 vein and the Sand River Mine vein are the same vein striking southwest parallel to a

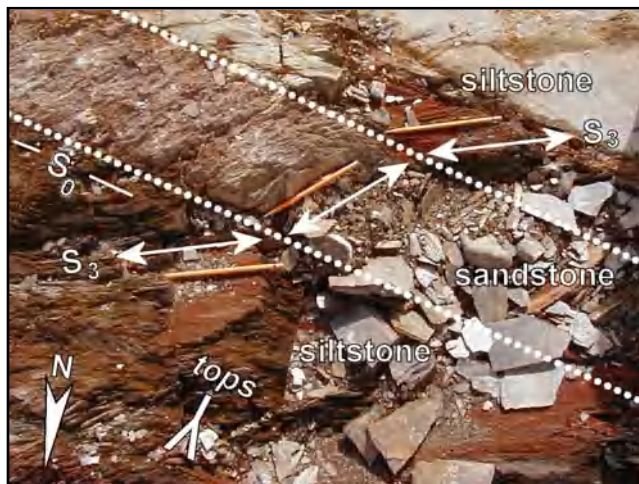


Figure 30. Refraction of cleavage in sandstone/siltstone, Sand River 16 Vein zone (Stop 1-8).

subvertical to vertical tectonic fabric which is subparallel to the original bedding in the clastic metasedimentary rocks (Hart et al., 2002). The vein contains multiple oreshoots that plunge at about 320° in the plane of the 65° northwest dipping vein, and these shoots are controlled by the intersection of the first fabric with a second regional tectonic fabric. Lafrance et al. (2004) defined the strike of the vein as 240° and dip 65°, with plunges to the oreshoots being consistent with the F3 fold axes, parallel to the Standingstone - Watson Lake Fault which was likely developed in two parallel, D3, dextral shear zones. Ore zone plunges are consistent with F3 fold axes. Hart et al. (2002) identified that the potential for additional gold mineralization would be very high in structures subparallel to parallel to the Sand River Mine – Leitch Mine No. 2 vein, such as the No.16 veins that are exposed north of the Sand River Mine shaft.

Stop 1-9 - Leitch Mine

UTM coordinates - 0425302E 5497484N

This stop allows trip participants to scour Leitch Mine dump piles for representative samples of gold-bearing quartz veins and clastic sedimentary rocks. Visible gold occurs within fractures in quartz veins and most commonly is found along thin seams or septa of sericite and chlorite, along with arsenopyrite, pyrite and ankerite in composite, “crack-seal”-textured veins. Tetrahedrite and sphalerite are also reported to occur in these fractures. Pink-beige scheelite is relatively abundant. Leitch Mine operated continuously from 1937 to 1965. Mill clean-up ensued between 1966 and 1968. One of the highest-grade gold mines in Canada during its lifetime, the mine yielded 861,981 ounces of



Figure 31. No. 1 vein, Leitch Mine, October 1935 (Laird 1937; Stop #1-9).

gold from 1,022,360 milled tons at a recovered grade of 0.915 ounce gold per ton. 31,802 ounces of silver were also produced at a grade of 0.035 ounce silver per ton. 64.3 tons of tungsten ore were produced at a average grade of 3.95% WO₃. Figure 31 gives us a glimpse of the property at the time of its discovery.

Day Two - Geraldton Area

Past-producing gold mines in the Geraldton camp (~ 3.1 million ounces) are confined to the northern part of the SMB and are localized by the Barton Bay Deformation Zone (BBDZ). The BBDZ (Fig. 12) is a 1 km wide, high-strain zone that extends from the Hard Rock and MacLeod-Cockshutt mines west along Highway 11 to the Bankfield Mine (Pye, 1952; Lavigne, in press; Lafrance et al., 2004). Shear is distributed heterogeneously across the BBDZ. In contrast to the mafic metavolcanic/mafic intrusive rocks to the north, and the monotonous sequence of sandstone to the south, the rocks within the BBDZ (mafic to intermediate intrusive rocks, felsic porphyry dykes and sills, sandstone/pelite, conglomerate and BIF) are complexly intercalated and have high competency contrasts. As noted by Lavigne (in press), only a very few areas in the belt have such lithologic heterogeneity. All of the aforementioned rocks in the BBDZ, except for the conglomerates, host gold. Although Geraldton is often considered to be a BIF-hosted gold camp, it is important to note that less than 30% of the gold came from BIF-hosted ore (Macdonald, 1982). Most of the production came from clastic sedimentary rocks, particularly at or near the contact with feldspar porphyry intrusions. Anglin and Franklin (1985) cited a more consistent spatial relationship between gold mineralization, deformation zones and felsic intrusive rocks.

Stop 2-1 - MacLeod-Cockshutt Mine

UTM coordinates - 0504200E 5502770N

The history of the MacLeod-Cockshutt Mine (Fig. 32) has been summarized by Horwood and Pye (1955) and Mason and White (1985). In 1931, following the discovery of gold by W.W. Smith on Discovery Point, Kenogamisis Lake, F. MacLeod and A. Cockshutt staked the ground adjoining the Hard Rock Gold Mines Limited property to the west. Surface exploration led to the discovery of gold-bearing quartz veins in 1931. The discovery of larger mineralized zones in 1933

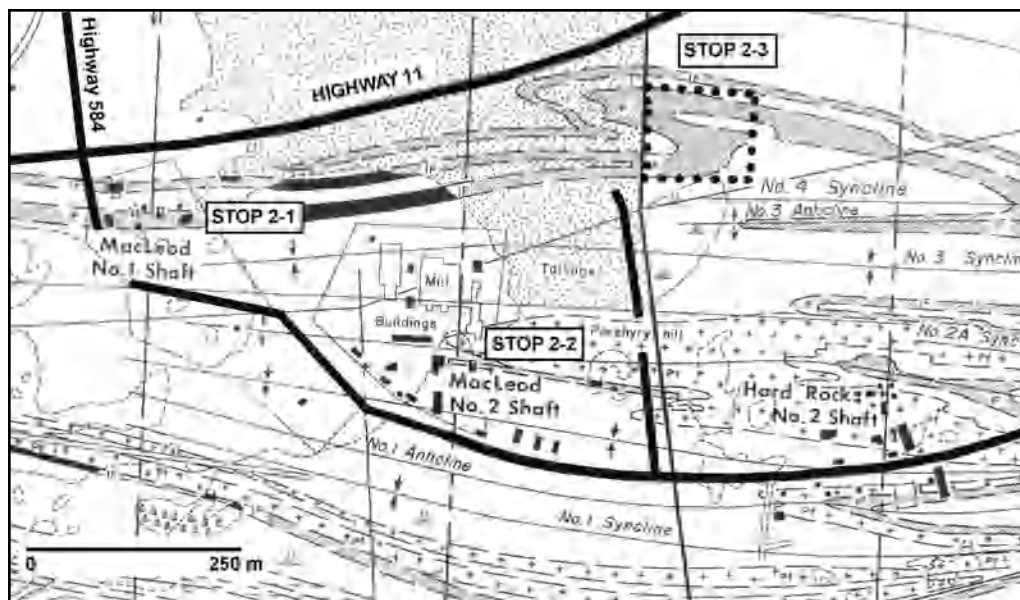


Figure 32. Map of the MacLeod-Cockshutt - Hardrock mines area (modified after Ferguson 1967). Pf = albite porphyry; IF = banded iron formation; sandstone is unlabelled and unhatched.

led to the organization of a new company, MacLeod-Cockshutt Gold Mines Limited. In 1934, shaft-sinking began here at the Number 1 shaft; the Number 2 shaft, 600 m to the southeast, was sunk in 1936. The MacLeod-Cockshutt Mine became the fifth producing gold mine in the Little Long Lac area on April 19, 1938 when a mill with a rated capacity of 600 tons per day was brought into operation. In 1967, MacLeod-Cockshutt Gold Mines Limited, Consolidated Mosher Mines Limited and Hard Rock Gold Mines Limited were amalgamated to form MacLeod Mosher Gold Mines Limited. Underground operations continued until July, 1970. The mine had produced 1,546,980 ounces of gold at an average grade of approximately 0.14 ounces of gold per ton. This total accounts for about half of all the gold produced by the 10 mines in the Geraldton gold camp between 1934 and 1970.

In the 1980's, Lac Minerals Ltd. (now Barrick Gold Corporation) undertook studies of existing underground reserves at the MacLeod-Cockshutt and neighbouring Hardrock mines and carried out lithogeochemical sampling (Gray, 1993). Starting in 1987, Lac conducted ground geophysical surveys, followed by 77 diamond drill holes, totaling approximately 50,000 feet. Targets, especially those with open pit potential, were investigated (e.g., Hardrock D and F; North and South Porphyry; Porphyry Hill zones). In 1992, Asarco Exploration Company of Canada Limited entered into a 5-year earn-in agreement with Lac and in 1993 carried out a program of reverse circulation overburden drilling and diamond drilling, the latter largely focusing on the near-surface portion of the F-zone and targets along the plunging nose of the albite porphyry (ibid). As a result

of this work, a geological resource was calculated by Horvath (1993) for the Porphyry Hill, West and East pits:

Pit Resource: 1,920,000 tons grading 0.079 ounce gold per ton (with strip ratio, including overburden, of 4.76 to 1)

Ramp Resource: 1,160,000 tons grading 0.127 ounce gold per ton

Asarco continued their exploration campaign into 1994, completing reverse circulation holes in overburden; sonic holes in tailings; and an additional 40,000 feet of diamond drilling, mostly on the aforementioned targets (Gray, 1994). Cyprus Canada Inc. assumed Asarco's role in the Lac Minerals agreement in 1996 and drilled 24 holes, leading to the discovery of the B-zone (Mason and White, 1997). The agreement ended in 1997 and the properties have been dormant since then. Lac Minerals Properties Inc. began a rehabilitation program in 1996 that is still active.

In the first description of the fledgling Little Long Lac (i.e. Geraldton) camp, Bruce (1935) proposed three types of gold deposits:

- (1) Shear zones in sedimentary rocks, along which narrow, but closely spaced quartz veins occur in parallel planes;
- (2) Irregular veins of quartz accompanied by pyrite, filling fractures in iron formation; and
- (3) Zones of pyritization and silicification in both in sedimentary and intrusive rocks.

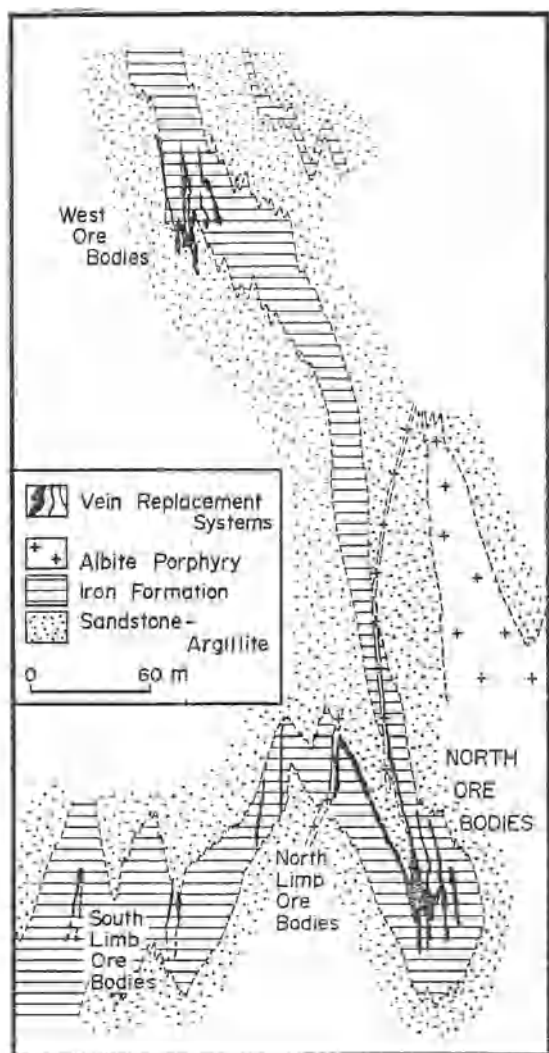


Figure 33. Vertical north-south section, MacLeod-Cockshutt Mine. After Horwood and Pye (1955).

Six main ore zones were exploited at the MacLeod-Cockshutt Mine. Horwood and Pye (1955) classified the MacLeod-Cockshutt orebodies and zones of mineralization into three distinct types:

- (1) Quartz veins and mineralized zones
- (2) Irregular, massive sulphide-quartz lenses in a folded series of sandstone and banded iron formation
- (3) Quartz stringer zones in sandstone and subordinate quartz diorite.

The first ore type, exemplified by the South, No. 210 Quartz Vein, No.'s 516, 517 and 519 Drift zones, is manifested by numerous but small lenses of semi-massive sulphides (coarse pyrite + arsenopyrite, sphalerite and chalcopyrite) scattered along the contacts of intrusions and metasedimentary rocks; by disseminated, auriferous sulphides in porphyry; and by minor quartz-carbonate veins and stringers cutting both

the porphyry and the adjacent metasedimentary rocks.

The second ore type is exemplified by the North, North Limb, South Limb and West zones (Fig. 33). Ore consists of masses, lenses and tongues of almost massive sulphides that replace folded wall rock, particularly laminae of incompetent, sheared sandstone and magnetite beds in banded iron formation. Sulphides, principally pyrite and arsenopyrite, also replace rocks adjacent to steeply dipping, east-trending shear fractures or shallowly west-dipping, north-trending tension fractures. Fractures are commonly healed by late(r)-stage quartz. It was reported that replacement-type ore averaged approximately 0.5 ounce gold per ton, while fracture-filling ore typically returned between 1 and 3 ounces gold per ton.

The third ore type, best exemplified by the F-zone, consists of numerous quartz veins and stringers localized within a wide zone of shearing and fracturing along the northern side of the albite porphyry. The veins and stringers range in width from < 1 to 10 cm. They appear to be largely confined to sandstone and lean BIF nearer to surface and occur in altered quartz diorite at lower levels to the west. Generally, quartz stringers are bordered by buff-coloured, sericitized and carbonatized sandstone with fine- to coarse-grained pyrite and subordinate arsenopyrite. In iron-bearing wall rock, sulphides are more predominant and may



Figure 34. No. 1 Shaft headframe, MacLeod-Cockshutt Mine (Stop 2-1).



Figure 35. Deformed magnetite BIF, MacLeod-Cockshutt Mine (Stop 2-1; coin is 2 cm in diameter).



Figure 36. West-plunging folds in BIF, MacLeod-Cockshutt Mine (Stop #2-1).

form small replacement lenses. Although the quartz veinlets contain visible gold, much gold is also associated with replacement pyrite and arsenopyrite.

The recently exposed outcrop at the base of the No.1 shaft headframe (Fig. 34) displays highly deformed BIF and sandstone, part of a band that extends 600 m east onto the Hard Rock Mine property and hosts the North ore zone there (Stop 2-3). The steeply dipping BIF contains primary magnetite, but displays some

secondary pyrite in association with iron carbonatization and quartz veining. Transpressive/oblique compression has produced open Z-folds and localized S-folds, with evidence of re-folding and late(r) transposition (Figs. 35 and 36). Consistently shallow (15° to 25°) plunges are observed on fold hinges. Lafrance et al. (2004) noted several, small isoclinal F1 folds that were folded by parasitic F2 S-folds, resulting in a strong, rod-like, coaxial lineation that plunges 20° to 40° west, subparallel to the strike of the S2 axial-planar cleavage ($100^{\circ}/85^{\circ}$ southwest) of regional folds.

The BIF consists of a finely parallel-laminated sequence of alternating, magnetite-rich layers and silt- to very fine sand-rich layers. The sand layers rarely show grading due to deformation, although some graded examples show increasing magnetite content in their clay-rich tops, which are gradational to the overlying, pure chemical sediment layer. On the microscopic scale, the magnetite layers have small amounts of chert and detrital minerals, but are otherwise mineralogically pure. Although deformation limits the depositional interpretations that can be drawn in this area, other exposures near Beardmore (e.g., Stop 1-6) indicate that the iron-rich sediments accumulated in the shore-proximal positions near the mouths of rivers in braid-delta complexes.

Stop 2-2 - Porphyry Hill, Hard Rock Gold Mines Limited

UTM coordinates - 0504810E 5502850N

In 1935, Hard Rock Gold Mines Limited began surface work and diamond drilling on the “Porphyry Hill” section of the property and discovered gold-mineralized zones in and along the northern tongue of porphyry (Horwood and Pye, 1955). The discovery of scattered, finely disseminated pyrite and numerous, small quartz stringers prompted underground follow-up, resulting in the definition of the No. 1 and No. 2 vein systems. In addition, a small, high-grade shoot outcropped near the western property boundary. The X-vein system was defined underground (3.5 feet wide, 110 feet long, with an average grade of 0.049 ounce gold per ton), within an inclusion of sandstone and lean BIF in albite porphyry (ibid).

Asarco’s proposed Porphyry Hill open pit was designed to recover shallow ore from several zones, including the F-zone (sandstone-BIF-hosted; northern porphyry contact-hosted); Porphyry zone (sheared



Figure 37. Contact between albite porphyry and sandstone/BIF, Porphyry Hill (Stop 2-2).

porphyry-hosted); South Porphyry zone (southern porphyry contact-hosted) and the Southern Iron Formation zone. Horvath's (1993) geological resource estimate for the Porphyry Hill open pit included an estimated high-grade resource of 1.02 million tons at a grade of 0.100 ounce gold per ton, and a low-grade resource of 0.48 million tons at a grade of 0.034 ounce gold per ton (for a total of 1.5 million tons grading 0.079 ounce gold per ton).

Deformed porphyry dykes and sills intruding sandstone and BIF are exposed at this site (Fig. 37). Crowded albite and minor quartz phenocrysts occur in a fine-grained matrix of quartz, feldspar, sericite, calcite and chlorite. An age of 2698.6 ± 1.3 Ma was obtained on a Au-mineralized "crowded albite" feldspar porphyry dyke at the Hard Rock Mine (Corfu, 2000).

Stop 2-3 - Glory Hole BIF, Hard Rock Gold Mines Limited

UTM coordinates - 0504862E 5503143N

Deformed, gold-bearing BIF is exposed along the edges of a fenced-off, water-filled glory hole, just south of Highway 11 that represents the mined, near-surface portion of the Hard Rock North ore zone. BIF occurs with siltstone/sandstone, porphyry sills and diorite sills and dykes. This orebody consisted of numerous, irregularly shaped replacement lenses of sulphides distributed in an east-trending zone along the northern limb and in the trough of a minor synform having a plunge of approximately 30° west (Horwood and Pye, 1955). The open stope mimics the shape of a F3 Z-fold defined by folded BIF horizons (Macdonald, 1988). Individual lenses were not of great lateral and vertical extent, but had lengths of up to 300 feet and widths up to 30 feet. Collectively, these sulphide lenses formed an ore zone which persisted and plunged onto the

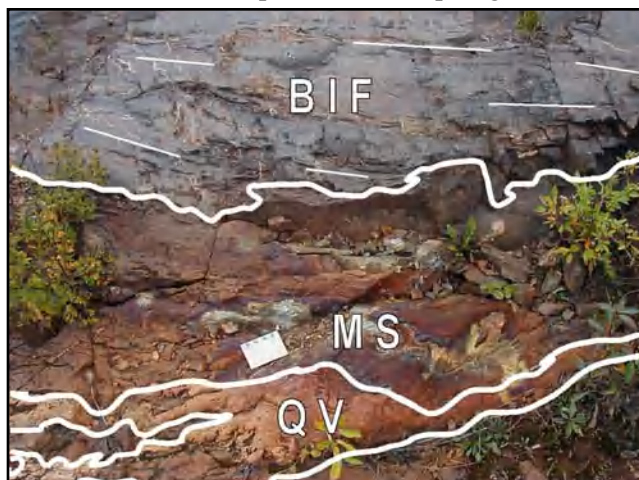


Figure 38. Oxide-facies banded iron formation (BIF) cut by massive sulphide (MS)-rich alteration envelope surrounding quartz-carbonate vein (QV), Hard Rock Mine (Stop 2-3).

MacLeod-Cockshutt property to the west. This is the same BIF that is exposed below the No. 1 headframe (Stop 2-1), approximately 600 m west of this location.

The so-called massive sulphide lenses contain up to 65% combined pyrite, arsenopyrite and pyrrhotite, the remainder being quartz. Minor sphalerite, chalcopyrite and galena, as well as accessory ankerite, calcite and tourmaline has also been noted. Replacement, controlled by quartz-filled fractures and bedding planes, is locally pervasive, destroying primary lithologic features (Fig. 38). Pervasive carbonatization accompanies replacement and gold mineralization. Replacement lenses are localized in zones of intense deformation



Figure 39. Pre- and post-rehabilitation views of the MacLeod-Cockshutt tailings at the junction of Highways 11 and 584, showing new Heritage Interpretative Centre.

along contacts between sandstone and BIF, and in the noses of small, asymmetric folds in an orientation consistent with the axial-planar cleavage in these folds (Anglin and Franklin, 1985). North zone ore mined to the mid-1940's averaged approximately 0.25 ounce gold per ton.

Stop (optional) - Geraldton Reclamation Project

In the late 1990's, Barrick Gold Corporation, in conjunction with the Town of Geraldton and government agencies embarked on a tailings reclamation project. This Tourism Development Project led to the development of the Heritage Interpretative Centre and a 9-hole expansion of the Kenogamisis Golf Course. The following description has been gleaned from the website of one of the designers (<http://www.gsd.harvard.edu/people/faculty/schwartz/projects.html>).

In the Geraldton reclamation project, mine tailings were reshaped for both aesthetic and economic reasons (Fig. 39). 14 million tons of tailings from the mines covered 170 acres to a depth of 27 feet. In order to spur economic redevelopment, the Town made the decision to make something of the tailings by improving their appearance and adding opportunities for visitors. Design alternatives involved sculpting the flat pile into compelling sculptural landforms which serve as a dynamic roadway edge and a gateway to the town.

Trails invite one to walk, bird watch, mountain bike, snow board, sled or snowmobile. Technical constraints were key to the final form of the earthwork. The different types and sizes of earth-moving equipment and their turning radii provided guidelines for the grading plan. A primary objective in the project was to balance cut and fill, and to maintain a maximum total earth moving of 150 000 cubic meters. Cut is kept to a minimum as arsenic levels are higher toward the bottom of the pile. There is a cap at the bottom of the pile of tailings, and there is a maximum of an additional 5 m that can occur on top. Standing water has been considered as a design element, but the water table has been respected by the re-grading. Storm drainage is maintained and the proposed earthwork does not impede sight lines for traffic safety. Six to twelve inches of peat topsoil were added to disturbed areas to aid in revegetation. A planting plan for the project focuses primarily on native grasses, especially those golden in color. The Geraldton project reveals the power of design to remake a wasteland into a new landscape. Even more than an earthwork, this landform is also a cultural artifact, highlighting the location and role of mining in the life of the town.

Stop 2-4 - Conglomerate, Highway 11

UTM coordinates - 0502906E 5502908N

Low outcrops on the south side of Highway 11 opposite the turnoff for the Consolidated Mosher Mine display deformed conglomerate, sandstone and BIF of the SMB within the BBLZ. As noted by Williams and Stott (1991), strain is exhibited by differentiated layering in pelitic rocks and elongated clasts in the conglomerates, especially in less-competent, mafic clasts (Fig. 40). Abundant, shallowly west-

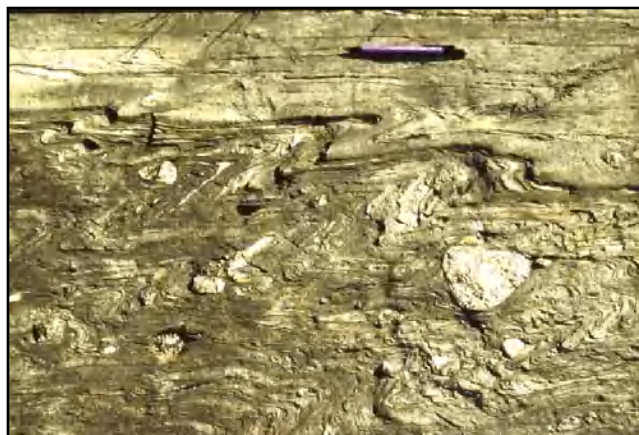


Figure 40. Sheared conglomerate in BBLZ, Highway 11 (Stop 2-4; Pen is 15 cm long.)

plunging, Z-folds with attendant axial-planar fabrics, back-rotated pebbles, pressure fringes and intraclast fractures support the contention that the BBLZ is a major dextral, ductile shear zone (Williams and Stott, 1991). Lafrance et al. (2004) noted that the long axes of deformed granitic clasts in polymictic conglomerate are parallel to the axial plane of F3 folds. S2 strikes 100° in the sandstone matrix away from the clasts, but as it wraps around the clasts, it changes orientation to that of the elongated granitic clasts, resulting in a geometry similar to dextral, asymmetrical strain shadows around rigid objects (ibid).

Polymictic, clast-supported, cobble-pebble conglomerate with deformed, lenticular lenses of deformed sandstone is interbedded with thicker, medium-grained sandstone sheets. They collectively represent a gravelly, braided fluvial system with sand-dominated channels, separated by gravelly longitudinal bar complexes. The bars were active during flood events with small sand channels within the bars developing as chutes during falling stage. In these types of braided systems, more massive, coarser-grained gravels developed in the bar-head region of the bars. Bar-tails are often composed of interlayered sand and thin gravel layers, dissected by secondary channels.

Stop 2-5 - Gabbro, Highway 11

UTM coordinates - 0502530E 5502860N

Approximately 300 m west of Stop 2-4, outcrops of deformed gabbro lie on the south side of Highway 11, just east of a small pond. The amphibolitized, medium-grained gabbro is cut by a network of dark green, chloritic zones < 1 cm wide in which hornblende megacrysts are elongated. These anastomosing shear bands separate lozenges of relatively undeformed gabbro, imparting an disconcerting impression of a pillowed basalt flow. With increasing degrees of deformation, the shear bands become sub-parallel, forming a C fabric (Macdonald et al., 1990). In the most deformed rock, a third (C') fabric develops at 45° to the C fabric. The northwestern orientation of C' is indicative of dextral shear. Elsewhere on the outcrop, sharply bounded, metre-wide zones of fine-grained greenschist represent mylonitized gabbro (Williams and Stott, 1991). Buck and Williams (1984) considered such variations in the preservation of primary textures, versus the development of shear fabrics, to be a function of the inhomogeneity of strain and not of the protolith.

Stop 2-6 - Missing Link Extension, Kinghorn Road

UTM coordinates - 0466744E 5511220N

The Missing Link Extension occurrence was discovered by Beardmore prospector Myron Nelson in 1986 and optioned to Freewest Resources Inc. in 1990 (Resident Geologist's Files, Thunder Bay North District, Thunder Bay). It is situated on the west side of the Kinghorn Road, just north of the Paint Lake Deformation Zone (PLDZ). The PLDZ is a 500 m-wide high-strain zone at or near the northern boundary of the Beardmore-Geraldton belt (Mackasey, 1976; Reilly, 1988; DeWolfe, 2002; Lafrance et al., 2004). The shear zone extends for over 40 km, from Lake Nipigon to the late, sinistral Jellicoe Fault, where it is offset by 500 m. The shear zone can be traced for ~ 9 km on the east side of the fault before it is lost under overburden. At the Missing Link property (1 km to the northeast; Stop 2-7), a northern splay/extension of the PLSZ has been exposed by mechanized stripping.

At Paint Lake (DeWolfe, 2002), the PLSZ is parallel to lithological contacts between polymictic conglomerate (NMB) and volcanic rocks (NVB), and forms the boundary with the Onaman-Tashota belt to the north. Shear sense indicators are observed in both conglomerate and volcanic rocks. Several granitic clasts in conglomerate are bounded by dextral, asymmetrical strain shadows filled with quartz fibers. Steeply dipping dextral shear bands cut across S3 (Lafrance et al., 2004).

The property straddles intermediate to mafic metavolcanic rocks on the north and clastic metasedimentary rocks to the south, separated by a major, east-striking structure interpreted to be the PLSZ (Resident Geologist's Files, Thunder Bay North District, Thunder Bay). Stripping and trenching exposed two sub-parallel zones of alteration and mineralization within sheared and isoclinally folded andesite. The orientation of the major fold axis is generally sub-parallel to the shear fabric, which strikes between 080° and 090° and dips steeply south to vertically. The fold is believed to plunge shallowly to the east. The Main Zone was exposed over a strike length of 40 m and a width of 12 m. Sub-zones were recognized on the basis of variations in the relative amounts of chloritization, carbonatization, silicification, quartz-carbonate veining and sulphide minerals. Finely disseminated arsenopyrite, pyrite and chalcopyrite were noted. Values of up to 0.20 ounce gold per ton were returned

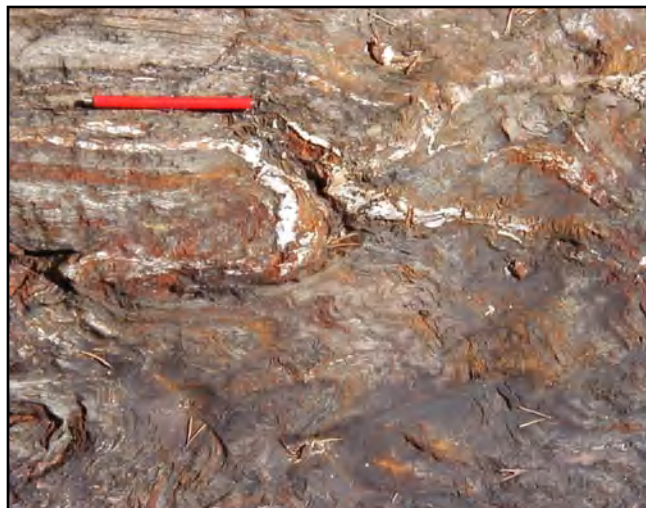


Figure 41a. Crenulated and altered metavolcanic phyllonite with quartz-carbonate veins, Paint Lake Shear Zone, Missing Link property (Stop 2-7).



Figure 41b. Crenulated and altered metavolcanic phyllonite, Paint Lake Shear Zone, Missing Link property (Stop 2-7).

in sulphide-rich grab samples.

Stop 2-7 - Missing Link Property / Paint Lake Shear Zone

UTM coordinates - 0467535E 5511479N

The Missing Link occurrence was described by Mason et al. (1989). Stripping, prospecting and a

variety of surveys were undertaken on the property by Beardmore prospector, Nolan Cox ca. 1986-1988. Older, undocumented trenches (ca. 1950's) were discovered during the course of this work. The property straddles intermediate to mafic metavolcanic rocks on the north and clastic metasedimentary rocks to the south, separated by a major structure (at 080°) interpreted to be an extension or splay of the Paint Lake Shear Zone (PLSZ) (Lafrance et al., 2004). At the Missing Link, phyllonites (mafic metavolcanic protolith?) display intense microfolding of S3 on millimetre-scale, producing a fine L3' crenulation lineation along S3 (ibid) (Figs. 41a,b).

DeWolfe et al. (2000) stated that gold mineralization and associated hydrothermal alteration were focused along second- and third-order fault splays off of the PLDZ. Second-order splays generally occur south of the PLDZ, often following lithologic contacts and refracting into an orientation sub-parallel to the PLDZ. Third-order splays are concentrated within the Northern Volcanic Sub-belt, where they are oriented both to the northeast and southeast.

Gold mineralization at the Missing Link property is associated with sheared and altered mafic metavolcanic rocks. The deformation zone has been exposed over a width of 60 to 90 m by stripping. Disseminated, foliation-parallel, euhedral pyrite and arsenopyrite within zones of both shearing and quartz-carbonate veining, host gold. Pervasive carbonatization and more localized, erratically distributed silicification and hematitization have also been noted (Mason et al., 1989). Three zones (North Zone; Shaft Pit and Baseline Trench) were sampled by N. Cox. The North Zone strikes 085° and dips 61° south and has returned assay values of up to 0.256 ounce gold per ton over 1.8 m. Grab samples collected by Mason et al. (1989) from all the zones returned up to 0.14 ounce gold per ton. The property was optioned by Homestake Mineral Development Company in 1989, who conducted a number of geochemical, geological and geophysical surveys. Only low gold values (e.g. 2.06 g/t Au over 1.0 m) were returned and in 1990 the property was returned to the vendor (Resident Geologist's Files, Thunder Bay North District, Thunder Bay). Placer Dome Inc. optioned the property in 1991 and completed just over 1000 m of drilling in four holes. Despite delineating a zone approximately 175 m wide within which intensely sheared and altered rocks occurred, no intersections of economic significance were encountered (ibid).



Figure 42. Load casts at base of sandstone (arrows), Highway 11 (Stop 2-8).

Stop 2-8 - Sandstone, Southern Sedimentary Sub-belt, Highway 11

UTM coordinates - 0468062E 55005080N

Thickly bedded, north-younging, overturned turbidites are exposed on the south side of Highway 11 opposite the Kinghorn Road turn-off. They display a variety of well-preserved sedimentary structures, including graded bedding, ball-and-pillow structures, flames and scours (Fig. 42). Some sandstone beds display a weakly developed cleavage.

This section of turbidites contains graded AE and AA beds up to 8m thick (Figs. 43 and 44). Most grading is in the basal few centimeters where the amount of very coarse sand grains decreases rapidly and in a zone approximately three-quarters of the way up through the bed where there is a rapid decrease from medium- to

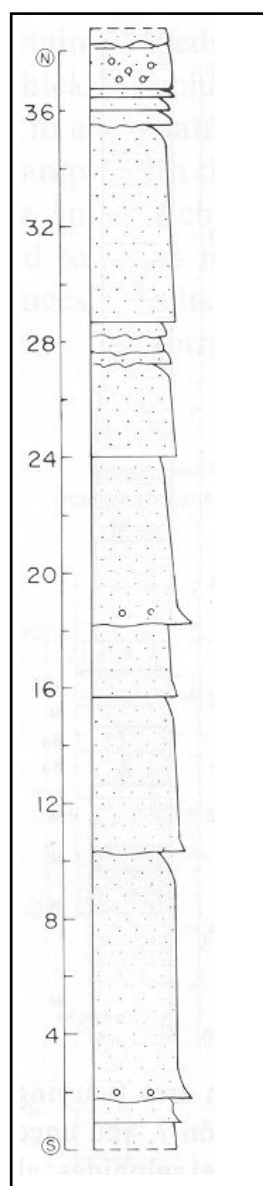


Figure 43. Stratigraphic column of sandstone outcrop (Stop 2-9; scale in metres).

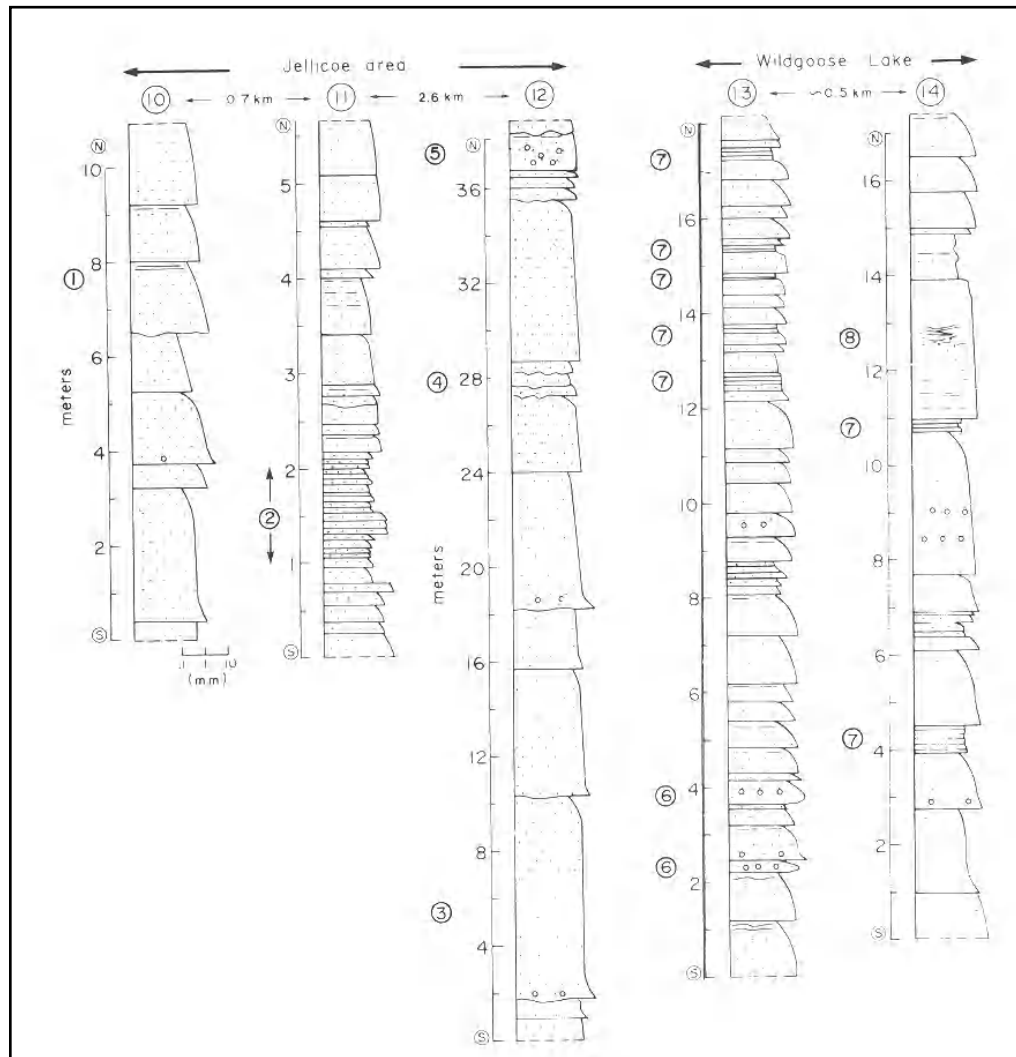


Figure 44. Measured sections of the turbidite assemblage in the Southern Metasedimentary Belt between Jellicoe and Geraldton. Note the extreme thickness of some of the beds. These units can be divided into a thin-bedded association, a medium-bedded association and a thick-bedded association. From Barrett and Fralick (1989).

fine-grained sand. This results in beds that are quite monotonous, with little change in their grain size through most of their thickness. These sediments were transported by typical, high-density turbidity currents and grainflows which deposit sediment in source-proximal locations. The fine-grained sand tops on the massive beds were possibly the result of resultant turbulence after the main turbidity current swept through the area. Currents moving water in behind the tail of the turbidity current would sweep sand from the surrounding bottom onto the coarser turbidity current deposits.

Stop 2-9 - Pillowed Mafic Metavolcanic Rocks, Highway 11

UTM coordinates - 0460740E 5503325N

Pillowed and massive, greenschist-facies, mafic metavolcanic rocks occur on the south side of Highway 11, 200 m west of the Jellicoe post office. They occur in a small, fault-bounded enclave of metavolcanic rocks with clastic metasedimentary rocks of the Southern Sedimentary Sub-belt (Bruce, 1937; Mackasey, 1976). They display a non-penetrative schistosity, on which there is a steeply plunging mineral and stretching lineation. Parallel to the schistosity are shear zones in which a non-penetrative fabric is forming at an angle to the zone, becoming parallel to it and more intense near the axis of the zone, characteristic of S-C fabric relationships (Williams and Stott, 1991). The shear sense based on this local relationship suggests south side-down. In graphitic metapelites and BIF exposed in road cuts 200 m west, the stretching lineation is well-developed.

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Fieldtrip 2 – Quaternary geology of the Beardmore – Nipigon area

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Introduction

In association with the International Quaternary Association in 1987 (INQUA'87) an excursion guidebook C-12: *Quaternary features and scenery along the North Shore of Lake Superior* was created (Geddes et al., 1987). Unfortunately the trip was never run. This current field trip and field trip guide utilizes much of the information presented in the INQUA'87 field trip guide particularly as it pertains to the Beardmore – Nipigon area.

Although this part of Ontario is pivotal to many large-scale glacial histories and theories of the Great Lakes Region, North America and global climate change, very little if any detailed mapping of the Quaternary geology has been done. One exception is the work of Thorleifson and Kristjansson (1993) in the very northern part of the field trip area (scale 1:100,000). In addition, Mollard and Mollard (1981) undertook a terrain analysis (scale 1:100,000) of a large part of the field trip area and produced a map of some of the major landforms and material types that occur there. The field trip area has been displayed on several regional maps including Zoltai (1965a), Sado and Carswell (1987), Barnett et al. (1991) and Sado et al. (1994). In addition, regional studies on selected aspects of interest to Quaternary researchers, particularly associated with ancestral lake features and deposits (e.g., Farrand, 1960; Teller and Thorleifson, 1983, 1987; Teller and Mahnic, 1987) have been done. Zoltai (1965b) described and discussed the formation of some of the major landforms and deposits of the Nipigon area.

During the trip we will visit several sites and landforms that display clues to the glacial history of the area and hopefully discussions on how they relate, support or conflict, with several of the regional histories proposed in the literature will ensue.

The fieldtrip route basically follows Highway 11 north from Nipigon to Beardmore and then northwest on Highway 580 to Poplar Lodge on Lake Nipigon and

then returns to Nipigon along the same route (Fig. 1).

The route follows the east branch of the Nipigon River, initially across the Nipigon lowlands then along the Pijitawabik Canyon cut through Nipigon Sills and onto more-typical Canadian Shield terrain north of Beardmore.

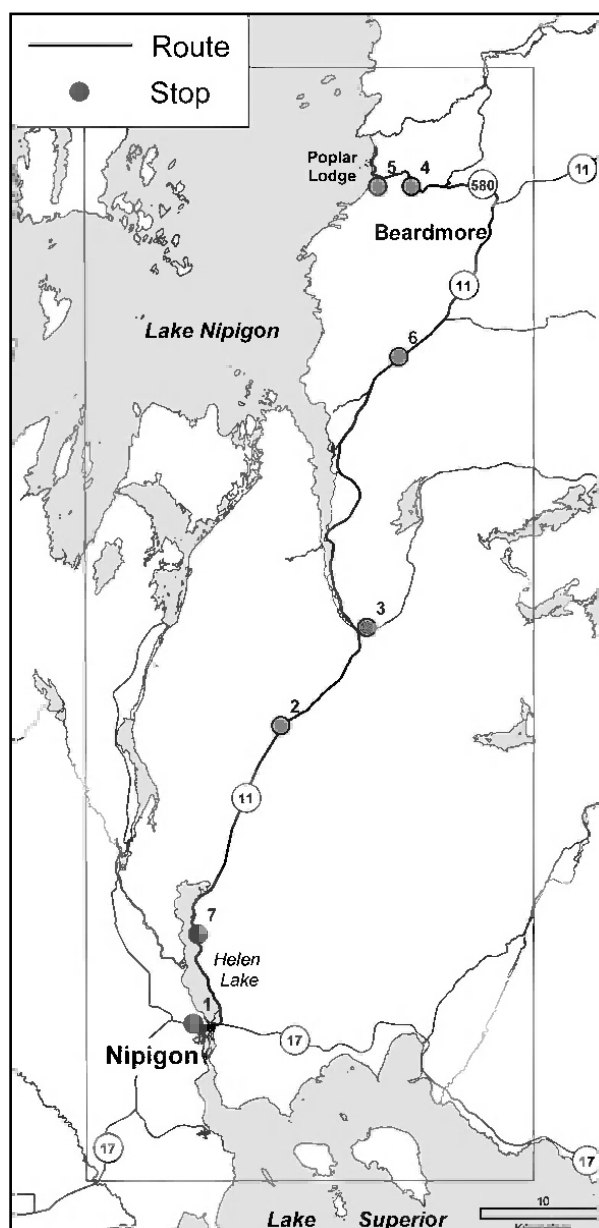


Figure 1. Location, route and stops. Rectangle is for reference between the various regional figures.

Regional geological setting

During the fieldtrip we will be travelling between Lake Superior and Lake Nipigon following a now abandoned drainage way used during deglaciation between these two lakes. The present drainage divide is located between McKirdy and Shamrock Lake within the Pijitawabik Canyon. Keemle Creek via a series of small lakes drains northward into Pijitawabik Bay of Lake Nipigon (approximately 260 m or 852 ft asl). An unnamed east branch of Cash Creek drains southward from McKirdy Lake into Cash Creek and onto Helen Lake and then into the Nipigon River to Lake Superior (approximately 183 m or 600 ft asl). The main branch of the Nipigon River is the present-day outlet of Lake Nipigon.

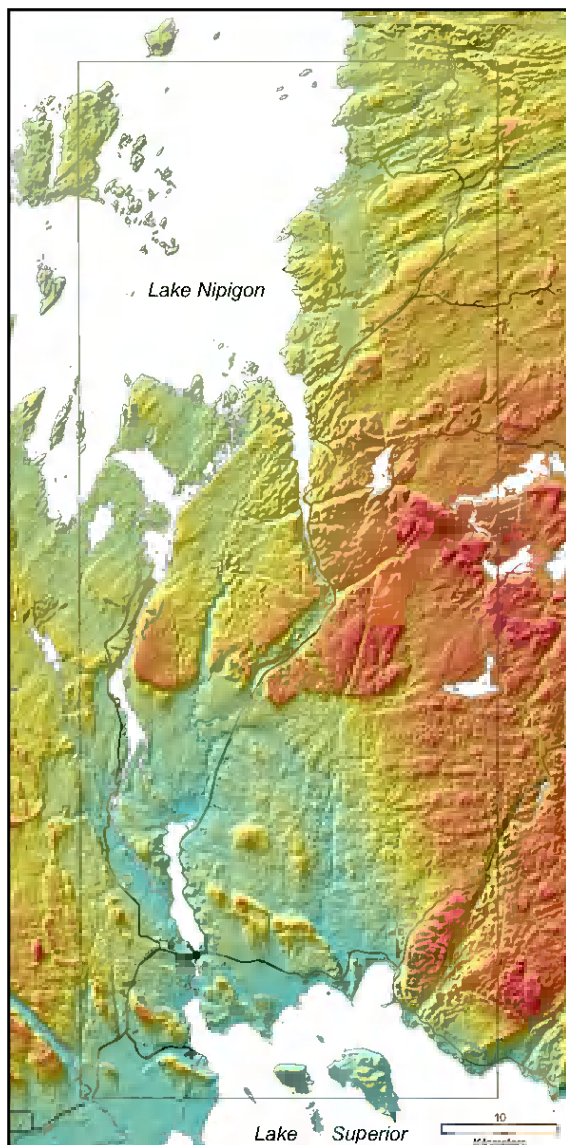


Figure 2. Fused digital elevation model (DEM) and hillshaded DEM. Rectangle is for reference between the various regional figures.

The fieldtrip area, located on the Canadian Shield, is an irregular, subdued, gently rolling, bedrock-dominated terrain underlain by Precambrian igneous and metamorphic rocks (Figs. 2, 3). The area immediately south of Lake Nipigon, however, is not typical shield terrain. It has been referred to as the Nipigon Plain by Bostock (1970) and “is formed on nearly flat-lying Proterozoic gabbro sills and sediments surrounding Lake Nipigon” (Bostock, 1970, p.16). Combined past tectonic activity, differential weathering and glacial/ glacial meltwater erosion of the Proterozoic sills and Sibley Group metasedimentary rock surfaces has produced a unique landscape of buttes and mesas and steep-walled valleys along the margins of the Nipigon Plain. To the east of Lake Nipigon, beyond the extent of the sills, in the area underlain by Archean igneous



Figure 3. Bedrock geology of the fieldtrip area (OGS 2003).

and metamorphic rocks, more typical Shield terrain is prevalent.

Farrand (1960) has referred to the southern part of the fieldtrip area as the Nipigon Lowlands. Here, the irregular surface of the Canadian Shield has been subdued and partially infilled by sand, silt and clay of glaciolacustrine origin producing a broad dissected plain. Farrand (1960) recognized at least 4 terraces or steps within this lowland at various elevations (692, 800, 907-912, 950-952 and 980 ft or about 211, 244, 276-278, 290, 299 m) that he related to ancestral lake levels in the Lake Superior basin.

Relief over the entire region exceeds 350 m (about 1150 ft). Local relief in areas of Archean rocks is commonly between 15 and 50 m (50 and 165 ft). Along the eroded edges of sills that surround the Nipigon Plain it can exceed 100 m (about 328 ft). Local relief of over 100 m (about 328 ft) also exists along the walls of the Pijitawabik Canyon.

Bedrock geology

The study area contains rocks of the Superior Craton and the Southern Province (Fig. 3). Major geological subdivisions include the Wabigoon and Quetico subprovinces (Blackburn et al., 1991; Williams, 1991), the Proterozoic Sibley Group and Nipigon Diabase Sills (Sutcliffe, 1991).

Metasedimentary and granitic rocks dominate the rocks of the Quetico Subprovince in the southern part of the fieldtrip area. Mafic metavolcanic rocks dominate the Wabigoon Subprovince with granodiorite to granite intrusions forming the inter-greenstone belt parts.

Diabase/gabbro in the form of sills and dikes is the dominant rock type in the Nipigon embayment or the north-central part of the fieldtrip area (Fig. 3). Nipigon sills are massive bodies that vary in thickness from 150 to 200 m. Internal textural zoning indicates that the sills are single cooling bodies (Blackburn et al., 1991). The Nipigon sills in the area are very similar texturally and monotonous in appearance. They overlie and interfinger with Sibley Group sedimentary rocks that outcrop or subcrop within the southern part of the Nipigon embayment. The Nipigon sills formed contemporaneous with the initiation of the Lake Superior Midcontinent rift age-dated at approximately 1108 Ma (Blackburn et al., 1991).

With the possible exceptions of some lateral and

vertical outliers the age of the Sibley is constrained between the youngest detrital zircon in a sample from the Kama Hill Formation (1420 Ma; L. Heaman pers. comm. 2004) and a Rb-Sr isochron from authogenic minerals in the Rossport and Kama Hill Formations (1339 Ma; Franklin, 1978). The Sibley consists of unmetamorphosed red-bed sequences consisting of conglomerates, sandstones and shales. They are believed to have accumulated in a subsiding graben prior to the formation of the midcontinent rift (Sutcliffe, 1991).

Greenstone belts east of Lake Nipigon are part of the Eastern Wabigoon Subprovince and have been traced beneath the diabase and sedimentary units (Blackburn, 1991; Smyk et al., 2005).

Quaternary geology

The surficial (Quaternary) geology of the fieldtrip area south and east of Lake Nipigon has been investigated only at a regional scale despite its importance to the overall understanding of the glacial history of Ontario and in particular to mineral exploration using unconsolidated sediments (drift exploration).

Regional maps at various scales that cover the area include Zoltai (1965a), Sado and Carswell (1987), Barnett et al. (1991) and Sado et al. (1994). Zoltai (1965b) described and discussed the formation of some of the major landforms and deposits of the Quetico-Nipigon area. Mollard and Mollard (1981) undertook a terrain analysis of the study area and produced maps of some of the major landforms and material types that occur there (Fig. 4). Several theses and papers have been written on significant aspects or features of Quaternary age from within the area of interest (Farrand, 1960; Teller and Thorleifson, 1983, 1987; Teller and Mahnic, 1987).

Zoltai (1965a, 1965b), Elson (1967), Prest (1970), Dredge and Cowan (1989) Barnett (1991) and Sado et al. (1994) have summarized regional concepts of the glacial history surrounding Lake Nipigon. There is very little agreement as to the details of the sequence and timing of events. The deposits in the area probably represent ice movement during the Michigan Subepisode (Late Wisconsinan), however; it may reflect ice cover during the entire Wisconsin Episode or approximately the last 100,000 years.

The Laurentide Ice Sheet flowed southwestward during its maximum extent about 20,000 years

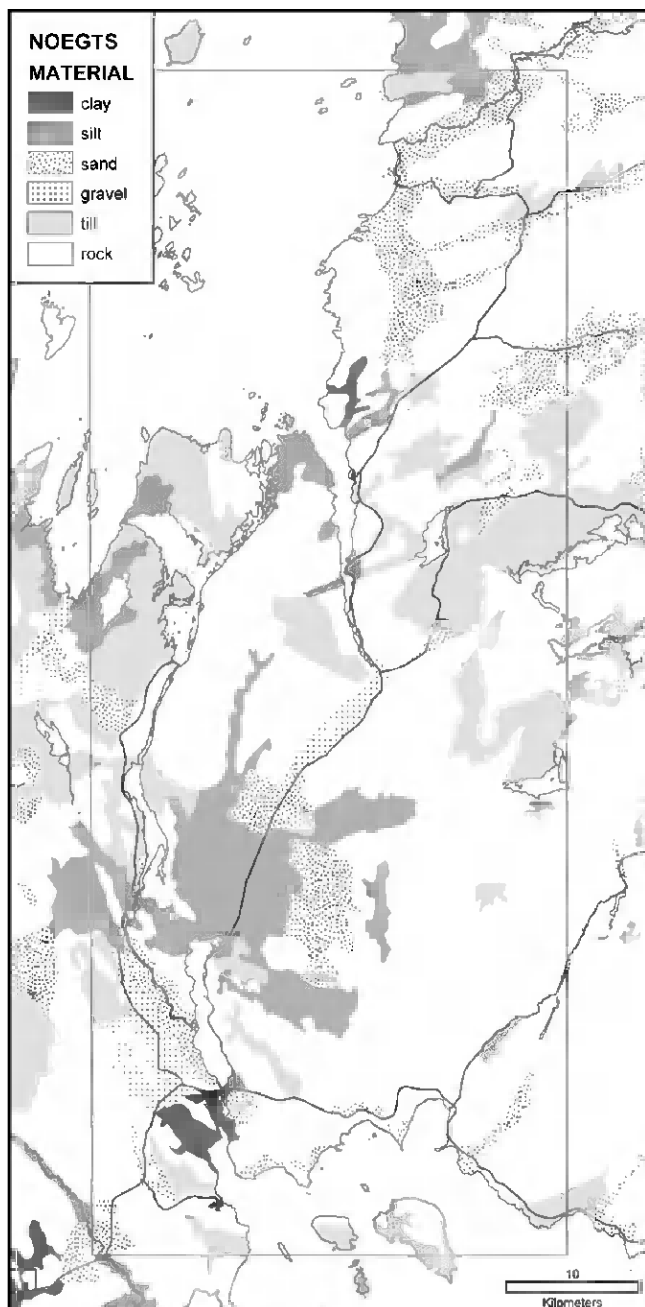


Figure 4. Surficial sediments in the fieldtrip area (modified from Mollard and Mollard, 1981).

ago. During deglaciation and possibly during initial ice movement into the area, ice flow direction was controlled on a regional scale by topography. During these times ice flow followed broad topographic lows, especially the Great Lakes basins and most likely Lake Nipigon, producing lobate ice margins that could have possibly behaved somewhat independently (Fig. 5).

Prest (1970), following the work of Zoltai (1965b), suggested that an active ice margin in northwestern Ontario fronted glacial Lake Agassiz during ice margin retreat. They (Zoltai, 1965b; Prest, 1970) postulated several alternating stages of ice margin retreat and

re-advance to produce most of the major moraines in northwestern Ontario. The Marks, Mackenzie and Dog Lake moraines have been suggested to have formed at about the same time, about 10,000 years ago (Burwasser, 1977; Drexler et al., 1983) and are probably correlative to the Grand Marais Moraines that formed during the Marquette advance (Drexler et al., 1983). With further ice margin recession and a suggested major re-advance, the Lac Seul moraine formed west of the fieldtrip area (Prest 1970). Zoltai (1965b) suggested that this ice margin formed the northern part of the Kaiashk Interlobate Moraine (Fig. 5).

Prest (1970) suggested that with further ice margin retreat, the Sioux Lookout and Whitewater moraines formed west of Lake Nipigon. These events were followed by a westward advance in the Lake Nipigon basin that produced the Nipigon Moraine (Stop 1) along the west and southern side of the lake (Zoltai, 1965a, b). Prest (1970) suggests that only the northern part of this moraine formed following the Whitewater Moraine and that the southern part is older. Zoltai (1965a, b) identified several other moraines east of the Nipigon Moraine in the Lake Nipigon basin that appear to be younger than the Nipigon Moraine. The Nakina Moraines that are located along the northeastern edge of the Lake Nipigon basin mark the last significant ice margin positions to directly affect the region.

To complicate the history even more is the suggestion that glacial Lake Agassiz waters drained catastrophically into the Nipigon/Superior basin several

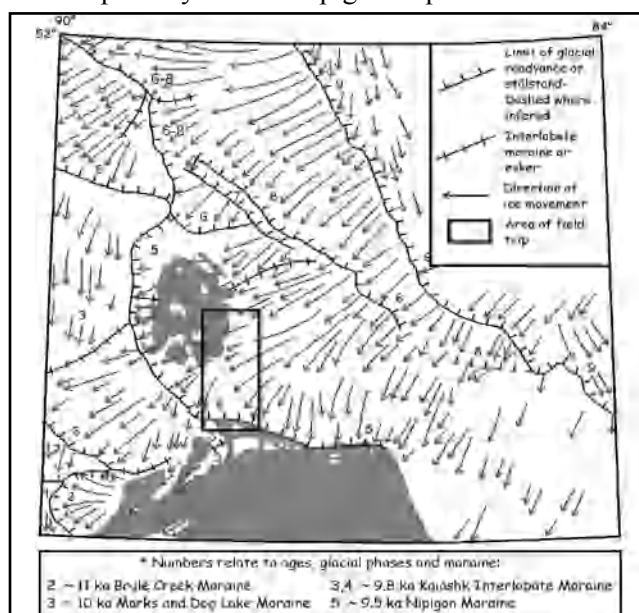


Figure 5. Ice flow direction and ice margins in the Lake Nipigon area (Sado et al., 1994).

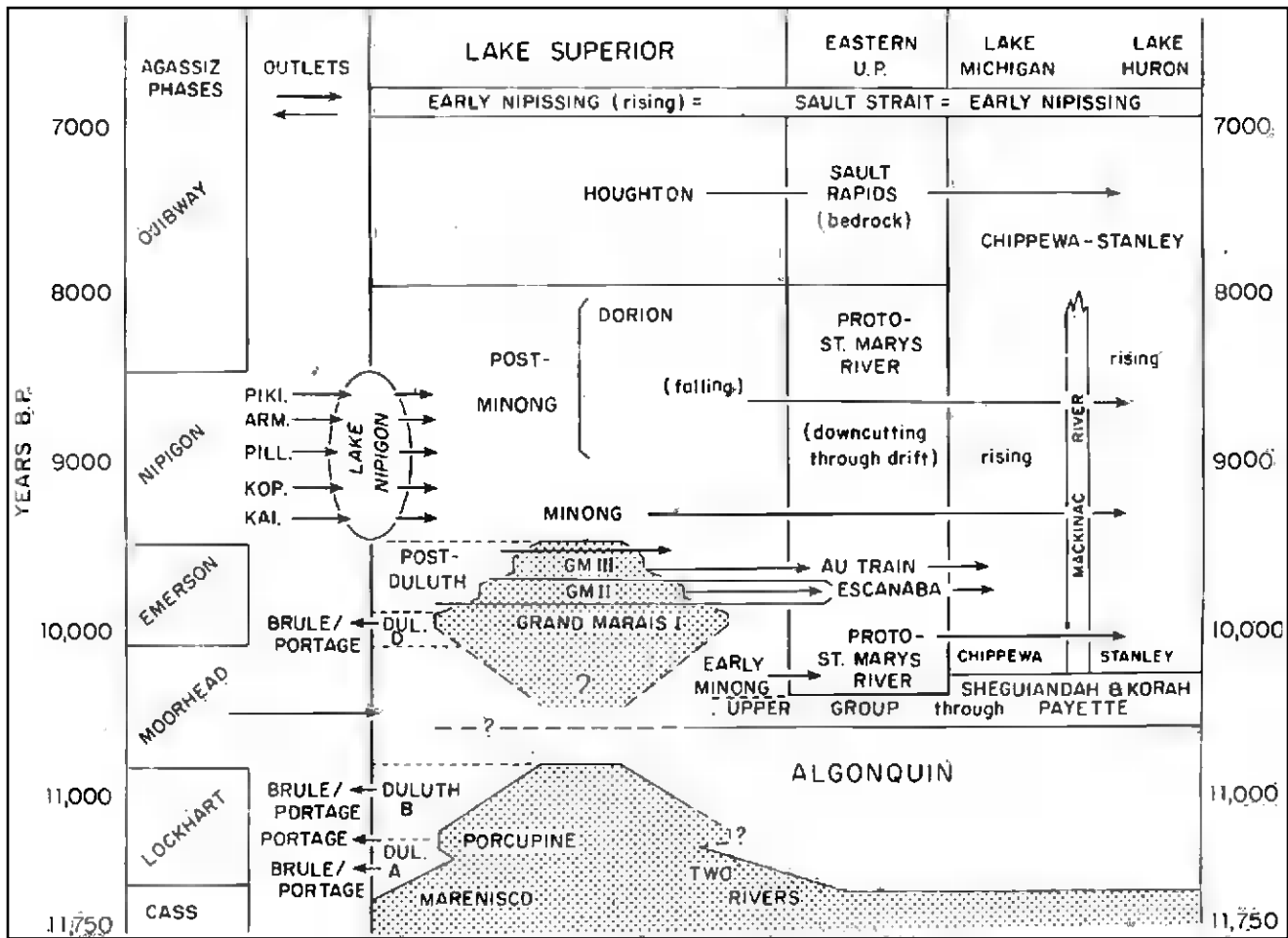


Figure 6. Chart of ice-sheet fluctuations and glacial lake phases (from Farrand and Drexler, 1985).

times during the deglaciation of the Lake Nipigon and Superior basins. Farrand and Drexler (1985) produced a chart summarizing some of the ice sheet fluctuations and their relation to glacial lake phases in northwestern Ontario and the upper Great Lakes basins (Fig. 6).

Agassiz drainage into the Nipigon and Superior basins occurred both before and after the Marquette maximum (Farrand and Drexler, 1985). A network of 5 groups of channels (Teller and Thorleifson, 1983, 1987) carried drainage from glacial Lake Agassiz following the Marquette advance (Fig. 7). And "as postulated by Teller and Thorleifson (1983), the outflow through the Nipigon channels must have been catastrophic" (Farrand and Drexler, 1985, p.23). These "catastrophic discharges from Lake Agassiz were responsible for cutting down the Lake Minong sill" (Farrand and Drexler, 1985, p.17).

The Nipigon Moraine marks a former ice margin located along the north shore of Lake Superior and has been traced along the southern and western side of Lake Nipigon (Fig. 5). It is commonly associated with ancestral glacial Lake Minong and as such believed



Figure 7. Eastern outlets of Lake Agassiz. Outlet number 26 is the Pijitawabik Canyon (from Teller and Thorleifson, 1983).

to have formed about 9500 years ago (Farrand and Drexler, 1985). Farrand's regional work on abandoned shoreline features along the north shore of Lake Superior (Farrand, 1960) is the most comprehensive field study on shorelines to date in the area (Figure 8). Farrand (1960) identified evidence of glacial Lake Minong about 25 km north of the Nipigon Moraine (Stop 3). And recently Slattery (2003) identified deltas that formed along the Nakina Moraines built to Minong III (?) or post-Minong levels at elevations of 360 m in the Longlac area.

If catastrophic Agassiz drainage carved the outlet at St Marys River, why do shore features of Minong or fairly high Post Minong levels exist well behind the Nipigon Moraine, a feature that in part post-dates the opening or use of many of the earlier eastern outlets?

The proposed eastern outlet events and features were based on several coarse-grained deposits associated with the large deep steep-walled valleys that occur throughout the region (Teller and Thorleifson, 1983, 1987). The Pijitawabik Canyon (Stop 3, Outlet number 26, Fig. 7) was proposed as one of the passageways for this meltwater (Teller and Thorleifson, 1983, 1987) and the Pijitawabik delta (Stop 2) was inferred to have formed during Agassiz drainage events (Geddes et al., 1987).

An alternative explanation for the development of most, if not all, the steep-walled valleys (e.g., Pijitawabik Canyon) is that they were initially formed in part by differential erosion along fractures and faults within the diabase/gabbro sills prior to glaciation. The valleys were then subsequently carved by subglacial meltwater floods based on the similarities of this channelled landscape to that of the Channelled Scablands of eastern Washington State (Baker et al., 1987). Many of these valleys contain glacial sediments indicating that the bedrock channels themselves were not carved by postglacial Lake Agassiz drainage events, but already existed. Agassiz outflow may have reoccupied some of these steep-walled channels that were carved by the subglacial meltwater flood(s).

The complex and still poorly understood deglacial history of the Nipigon basin does not make drift exploration easy. The events and features discussed above could have resulted in many areas having multiple directions of ice movement due to the shifting of ice centers through time that will require linking evidence of ice movement to specific tills (Fig. 4). Determining

the history of meltwater and ancestral lakes in the study area is also very important to interpret the results of drift exploration programs.

Studies of tills deposited in the Beardmore-Geraldton area (Thorleifson and Kristjansson, 1993) and in the Hemlo-Matheson-White River area (Geddes and Kristjansson, 1986; Karrow and Geddes, 1987; Hicock, 1988) have brought to light another factor that affects mineral exploration using till (drift exploration) in the fieldtrip area.

The till in the Beardmore area appears for the most part to have been deposited subglacially by an actively flowing glacier (ideal for mineral exploration using till). However, where the till is thick it is commonly rich in clasts of Paleozoic carbonate and Proterozoic metasedimentary rocks derived from Hudson Bay Lowlands some 100 km or greater away (Stop 6). The content of carbonate in the till matrix is also high (up to 50%; Thorleifson and Kristjansson, 1993). This "exotic" till tends to grade into locally derived till (Stop 4) at surface in the area surrounding the community of Beardmore.

Thorleifson and Kristjansson (1993) attribute the abundance of far-traveled debris to:

- 1) the high susceptibility of carbonate rocks to erosion
- 2) low erodibility of granitic rock which outcrop between the Beardmore area and the Hudson Bay Lowlands
- 3) the short distance of transport over the erodible greenstone belt rocks and
- 4) a zone of vigorous ice flow (ice streaming).

Thorleifson and Kristjansson (1993) suggested that an ice stream would be required to produce and distribute the carbonate-rich or exotic till to the Beardmore-Geraldton area based in part on the regional distribution in a radiating tongue parallel to ice flow. They also suggest that the exotic till was transported beneath the glacier by pervasive shear (deformable bed). They argued that "deposition by lodgement of basal debris seem contradicted by carbonate clast preservation" (Thorleifson and Kristjansson, 1993, p.57). Hicock (1988), however, records evidence within exotic till exposures of discrete not pervasive shear that would suggest lodgement as the process of deposition, although, he also suggested ice streaming to produce the exotic till.

Geddes and Kristjansson (1986) suggested that

englacial transport of the carbonate debris over long distances was required. Hicock (1986) suggested that englacial load under dynamic ice conditions can be transferred to the glacier bed in the leeside of upland settings.

Regardless of the mechanism of entrainment, transport and deposition of the carbonate-rich till, it has been suggested that the “standard rules of glacial comminution as applied in mineral exploration may not be applicable” (Karrow and Geddes, 1987, p.368). And that care be taken to recognize and avoid this exotic till when sampling for mineral exploration purposes (Thorleifson and Kristjansson, 1993).

The field trip stops were chosen to give the participants an overview of the Quaternary geology of the area between Lake Superior and Lake Nipigon and to introduce them to some remarkable landforms and scenery that exists in this part of Ontario.

For the INQUA’87 fieldtrip guide, Geddes et al. (1987) prepared a figure showing the profiles of water planes and shoreline features recognized by Farrand (1960) in the Lake Superior basin between Thunder Bay and Kama Bay. This figure appears as Figure 9 with a few modifications for reference purposes.

Stops

Stop 1 – Nipigon Moraine (J. Nichols Sand and Gravel, Nipigon)

UTM coordinates – 0406748E 5430513N

Note - permission from the land owner is required for access to this stop.

The Nipigon Moraine marks a former ice margin of the Laurentide Ice Sheet that fronted a large glacial lake. It is generally accepted that this lake was glacial Lake Minong and that its formation occurred about 9500 yr BP (Farrand and Drexler, 1985). In the Nipigon area it is a discontinuous ridge that rises more than 40 m above the surrounding lake plains. Elevation at the site is approximately 280 m asl. The ridge rises to the west where it becomes flat-topped at an elevation of about 295 - 296 m asl. The form here does not appear to be primary (i.e. delta) but the result of subsequent erosion and deposition (wave cut and wave built platform). A lower plain occurs to the north of the moraine here at about 290 m asl. These features raise the question of what was the actual water level in the Superior basin during the formation of the Nipigon Moraine.

An ice marginal feature built along the Nipigon

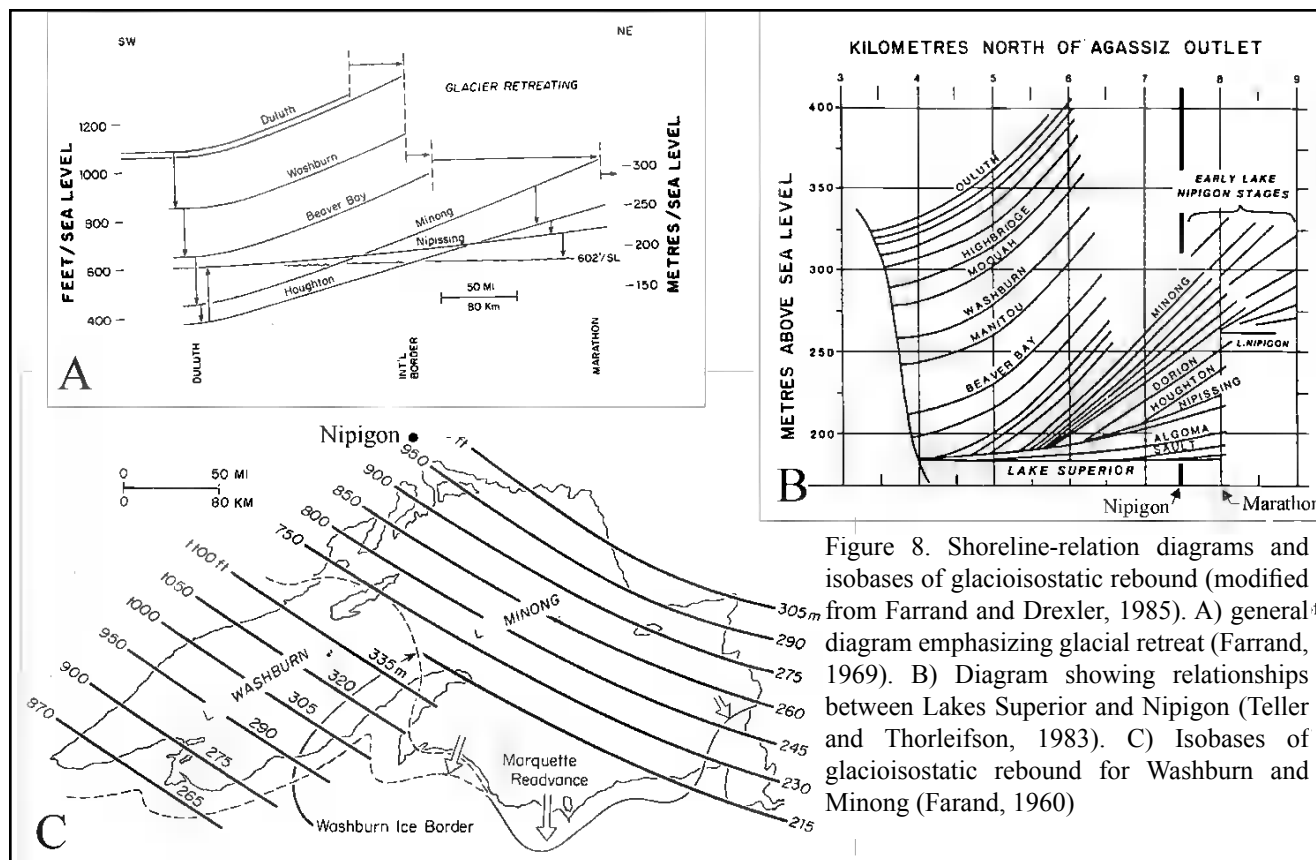


Figure 8. Shoreline-relation diagrams and isobases of glacioisostatic rebound (modified from Farrand and Drexler, 1985). A) general diagram emphasizing glacial retreat (Farrand, 1969). B) Diagram showing relationships between Lakes Superior and Nipigon (Teller and Thorleifson, 1983). C) Isobases of glacioisostatic rebound for Washburn and Minong (Farand, 1960)

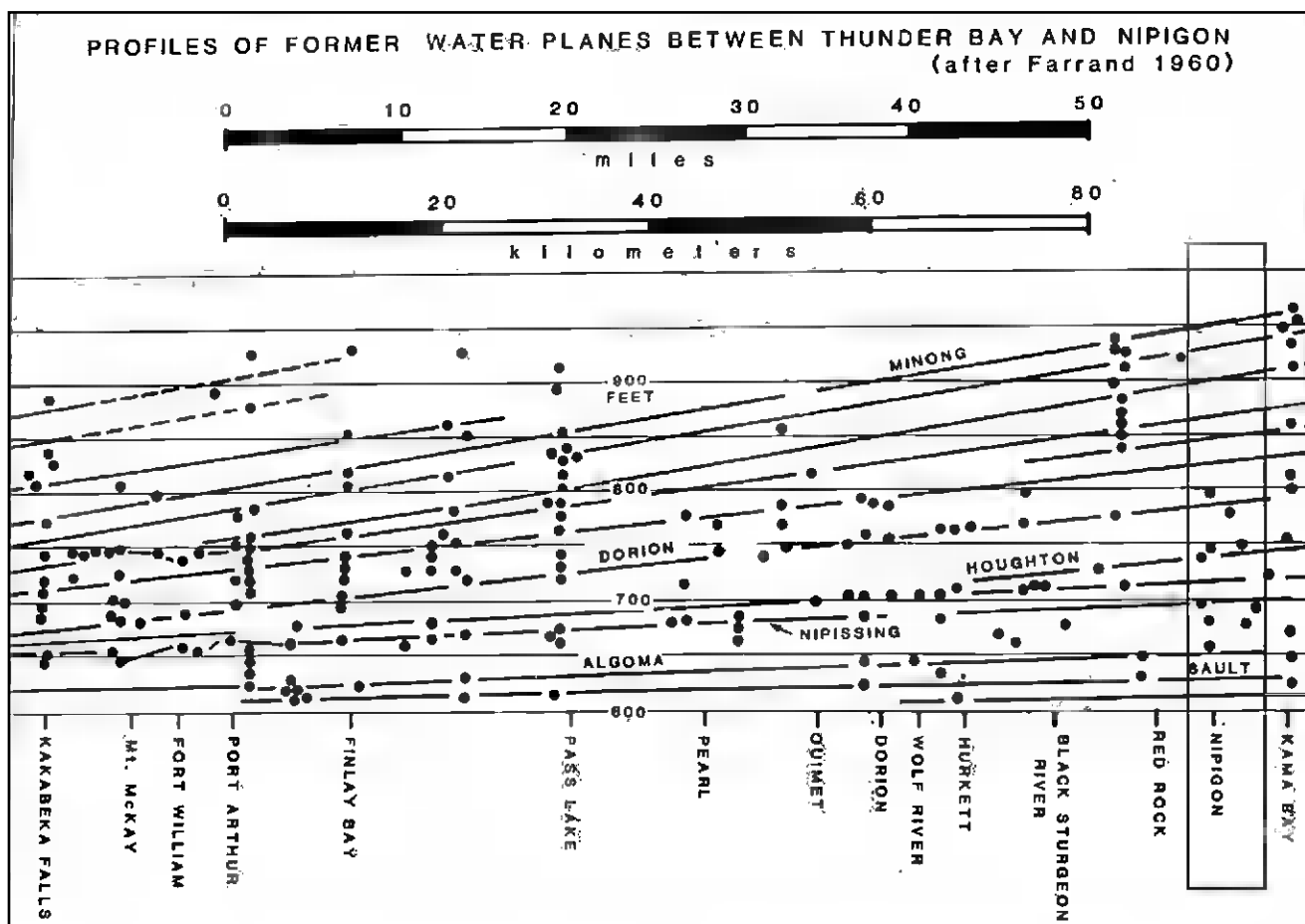


Figure 9. Profiles of former water planes in the Lake Superior basin from Thunder Bay and Kama Bay (after Farrand, 1960; Geddes et al., 1987).

Moraine south east of Black Sturgeon Lake that has a delta form occurs at an elevation of between 300 and 304 m. This feature would fall near to the isobase passing through Nipigon, essentially making the diagram one isobar off (Fig. 8C).

Shore features identified as ice-marginal deltas at Marathon and Wawa, that were reported to be built into glacial Lake Minong (Geddes et al., 1987), may not be deltas at all, but subaqueous fans later planed off by waves of a subsequent lake. This is the case at Wawa (S. Slattery, personal communications, 2003). The sedimentology of the feature at Marathon remains to be studied in detail before such a conclusion can also be made but preliminary observations are not consistent with a deltaic origin.

The Jim Nichols Trucking Limited sand and gravel pit provides a look at the internal structures of the sediments making up the Nipigon Moraine. In general the ridge here is made up of a fining upward and outward sequence of stratified sand and gravel. The lower part or core is composed of steeply dipping beds of gravel and sand (Fig. 10). These avalanche or foreset beds

probably formed during rapid growth by sediment gravity flow processes. The beds generally dip to the south or across the ridge form. Trough cross-bedded (dunes) sand, pebbly sand and pebble gravel overlie the steeply dipping beds. The aforementioned sediment sequence is common in subaqueous fans (Shaw, 1985). Near the top and along the ice-proximal side (north-side) minor flowtill lenses are present. Faults and post-depositional slumping are common along the proximal side of the ridge likely formed following the removal of the glacier support and sediment supply.

The sequence is interpreted as an ice marginal subaqueous fan. If so, it only places a minimum (280 m asl) for the lake that existed during its formation. As mentioned previously, the water here during formation could be up to 20 m deep. A shorebluff and bouldery gravel beach bar of a younger lake occurs along the roadway at the entrance of the pit at approximately 250 m asl (Figs. 10 and 11).

* *Alternative stop - look-out over community of Nipigon from the Nipigon Moraine (UTM coordinates - 0407343E 5430274N).*

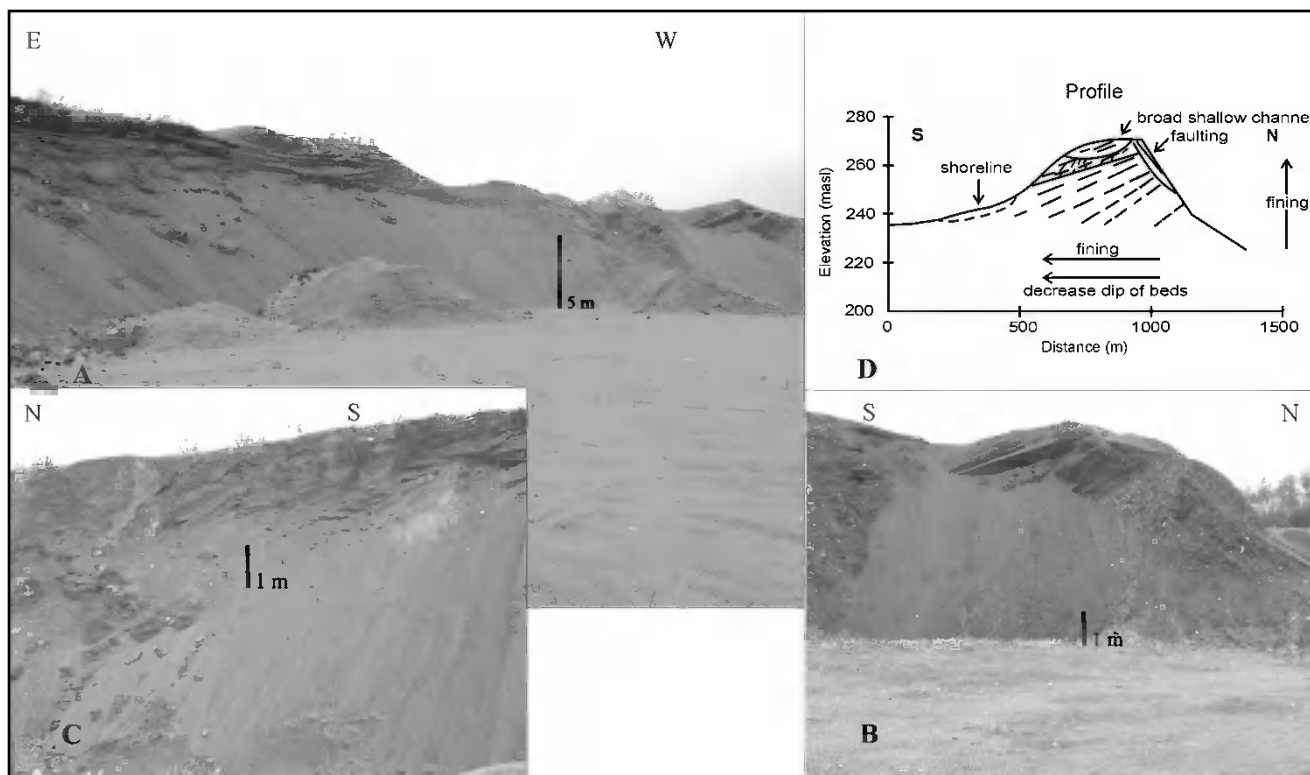


Figure 10. Internal structures of the Nipigon Moraine, Jim Nichols Trucking Ltd sand and gravel pit: A – longitudinal view, B – steeply dipping avalanche bedding, C – faulting and slumping along northern slope of the ridge and D – schematic cross-section of the Nipigon Moraine at this site.

Stop 2 – Pijitawabik delta

UTM coordinates – 041382E 545403N

The route from Stop 1 to Stop 2 crosses the Nipigon lowlands along the east side of Helen Lake. Glaciolacustrine fine-grained rhythmites occur beneath the surface of the plain until just north of Mignet Creek. At the break in slope about 400 m north of the creek,

Farrand identified a Dorion shore feature at 800 ft asl or about 244 m (Figs. 11 and 12). Another prominent bluff, a Post-Minong shoreline, occurs at an elevation of between 907 and 912 ft (276-278 m) asl (Farrand, 1960).

Farrand (1960, p.194) suggested that the “very extensive, near level surface of sand and some rather heavy gravel” about 25 km north of Nipigon is also a

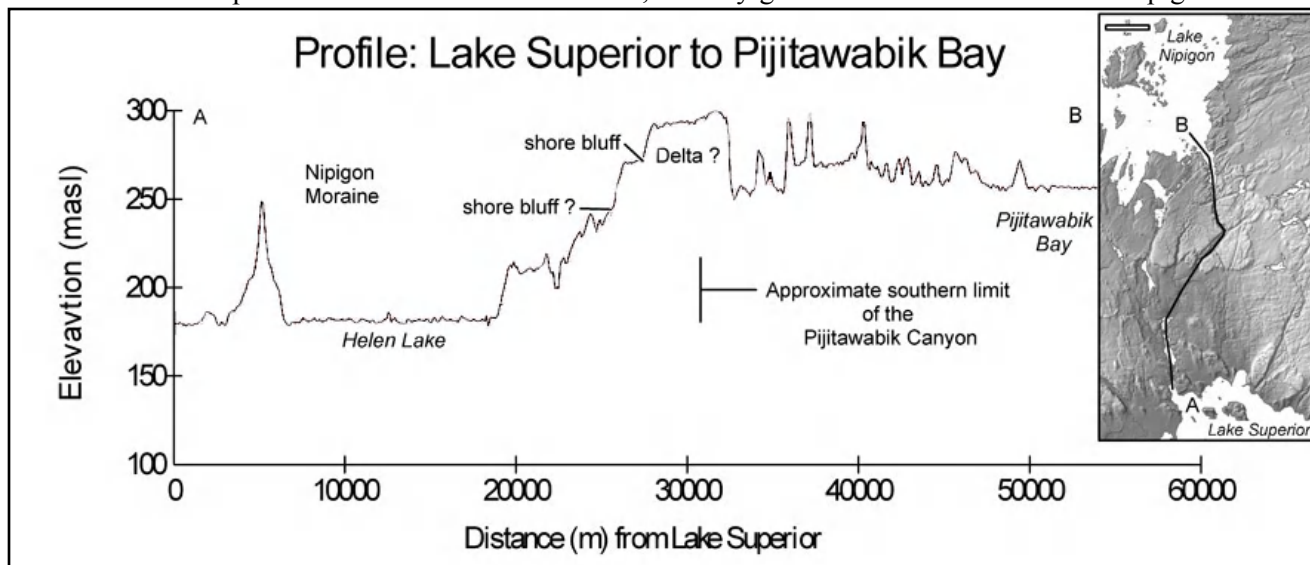


Figure 11. Profile from Lake Superior to Pijitawabik Bay of Lake Nipigon following the Pijitawabik Canyon. Stop 1 is on the crest of the Nipigon Moraine, Stop 2 at the Pijitawabik delta and stop 3 within the Pijitawabik Canyon. Several features controlled by ancestral water levels in the Superior basin can be seen.

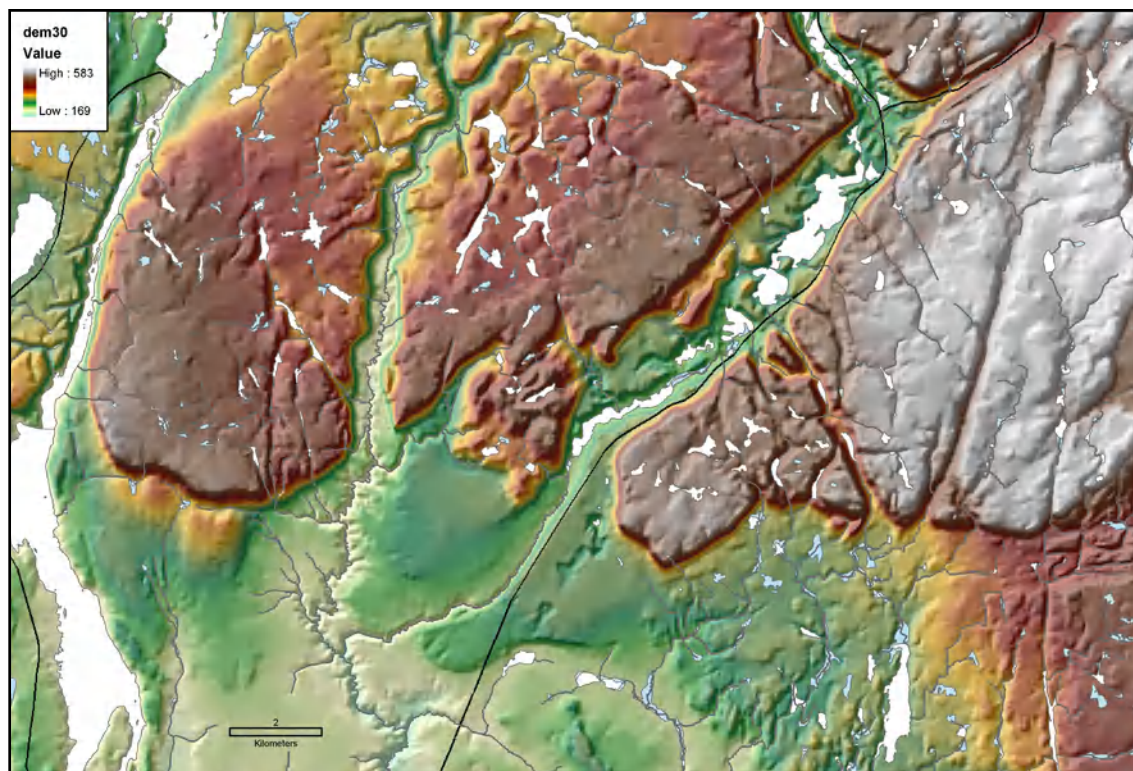


Figure 12. A fused DEM and hillshade of the Pijitawabik delta and the southern part of the Pijitawabik Canyon.

product of glacial Lake Minong (Figs. 11 and 12). This feature has an elevation of 950-952 ft or about 290 m asl. At 980 ft or 299 m asl, Farrand (1960, p.194) identified another possible Minong feature within the Pijitawabik Canyon. Both these features are well inside or north of the Nipigon Moraine.

The following quote from Farrand (1960) sums up some of his conclusions as to the origin and relationships that these deposits have with lakes in the Superior and Nipigon basins. “The 980-foot gravel accumulation lies nearly on the isobase of Terrace Bay where the Minong beaches lie at 950 to 1000 feet. Moreover, the abundance of sediment in the Pijitawabik Bay canyon implies the presence of the ice sheet either in Pijitawabik Bay itself or at least in the Lake Nipigon basin”. “These facts lead to the conclusion that the Nipigon basin was ice-filled (or nearly so) in Minong time. As lake levels fell in the Superior basin and the ice sheet retreated from the Lake Nipigon basin these basins apparently contained separate water bodies connected only by the ancestral courses of the Nipigon and Black Sturgeon rivers. In other words, there appears to be no validity in the assumption generally made (Stanley, 1932, p.135; Leverett, 1927, p. 1.7) that the Nipigon and Superior basins were occupied at any time by one continuous lake” (Farrand, 1960, p.82).

The initial conclusions made by Farrand appear to

be sound, however convincing evidence that Stanley’s or Leverett’s assumptions had “no validity” was not substantiated.

Down-cutting of the delta to the 290 m level may have been caused by a drop in base level as water in the Superior basin lowered in response to erosion at the outlet sill at Sault Ste Marie. The delta may have been fed from the glacier margin that remained within the Pijitawabik Canyon farther up the valley. Alternatively, during basin (Nipigon/Superior) separation, as an outlet stream from the lake Nipigon basin, or possibly built by an Agassiz drainage event through the valley.

Stop 3 – Pijitawabik Canyon

UTM coordinates – 0420816E 5462100N

This stop may be a series of stops along the route

The Pijitawabik Canyon is a deep, steep-walled valley cut into Nipigon diabase/gabbro sills. It is about 20 km long with a prominent near right-angle bend about 11 km from its south end. It can be considered to be made up of two segments; a southern segment that is about 2 km wide and a northern segment that is only about 1 km wide (Figs. 13 and 14). Local relief along the valley walls of both segments can exceed 150 m.

The valley is controlled by major structures in the

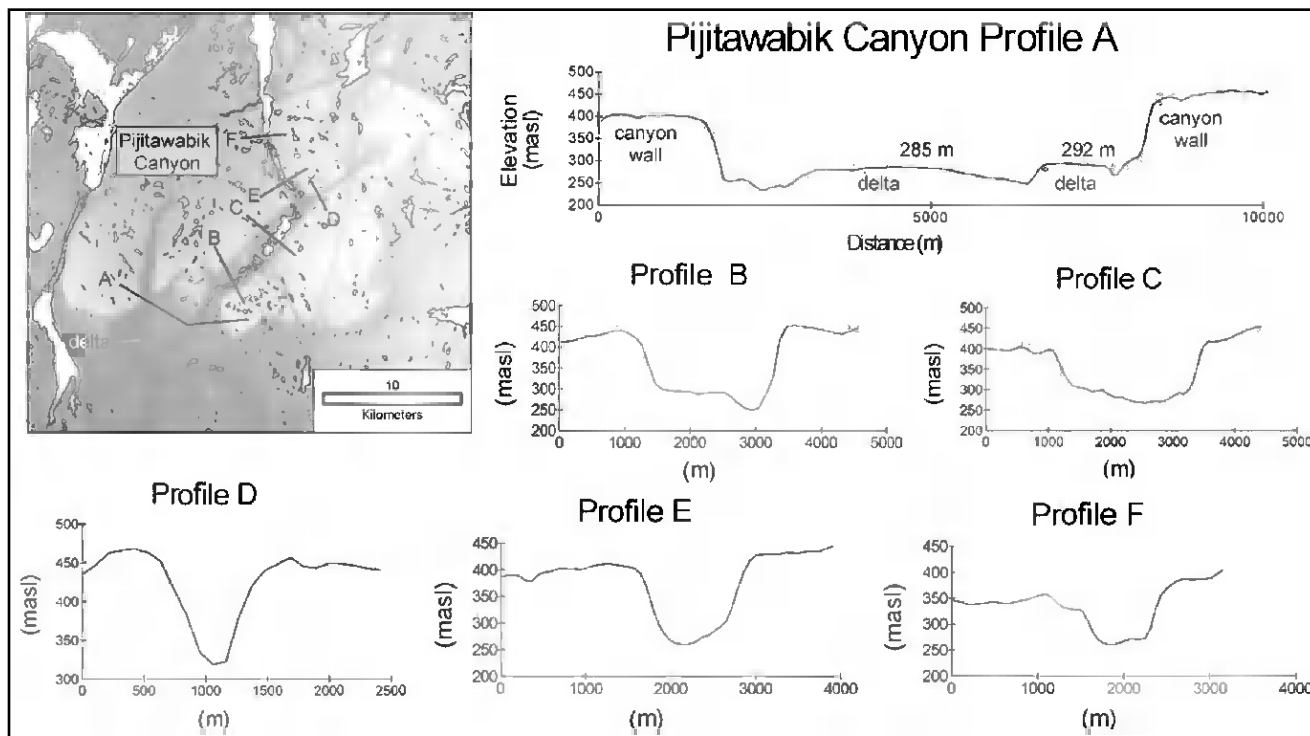


Figure 13. Selected profiles across the Pijitawabik Canyon.

bedrock. In addition, the rock characteristics of the Nipigon sills are such that weathering can be intense along fractures, producing grus-like residual soils that are easily removed by erosion.

The Pijitawabik delta emerges from the southern end of the canyon and deltaic sediments occur within the canyon walls (Figs. 11, 12 & 13). Remnant terraces of sand and gravel (Figs. 12 & 13) fall from an elevation of about 305 m at the canyon bend to the delta surface at about 299 m at the mouth of the canyon. The ice margin feeding this system must have been just north of the bend as suggested by Farrand (1960), in the vicinity of this stop. The side canyon, Profile D (Fig. 13) is cut down to the level of the terraces only, so major down-cutting here, must have occurred during

delta building.

Stop 4 – Sand River Property, Roxmark Mines Limited

UTM coordinates – 0424364E 5497579N

Note - permission from the land owner is required for access to this stop.

In 2003, Roxmark Mines Limited acquired the Sand River, Leitch and East Leitch properties from Rio Fortuna, Teck Cominco and Kinross, respectively (News Release, Roxmark Mines Limited, November 28, 2003) and began an advanced exploration program. In 2004, trenching and stripping at Sand River exposed vein systems filled with quartz.

Visible gold occurs within fractures in quartz veins and most commonly is found along thin seams or septa of sericite and chlorite, along with arsenopyrite, pyrite and ankerite. Channel samples were cut every three feet. All composite assays used a minimum width of 1.0 feet. The #16-1 West Vein returned an average assay of 1.737 ounces Au per ton over a strike length of 42 feet. The #16-1 East Vein returned 0.347 ounce Au per ton over 37 feet.

The Sand River area was most recently mapped by Hart et al. (2002) and is underlain predominantly by clastic metasedimentary rocks. Thickly bedded, southward-younging, feldspathic wacke, sandstone,



Figure 14. View of the Pijitawabik Canyon looking north toward Lake Nipigon (photo courtesy of P. Kor, OMNR).



Figure 15. Polished bedrock surface containing 2 sets of striations (photo courtesy of M. Smyk, OGS).

and argillaceous rocks have been well exposed by the recent stripping operations. Bedding has east-northeasterly to east-southeasterly orientations with steep, northerly dips. Well-developed, slaty cleavage, as well as local sericitization and carbonatization distinguish sub-parallel shear zones. Oblique cleavage - bedding relationships are locally exposed. Cleavage is refracted through interbedded siltstone and sandstone beds.



Figure 16. "Local till" overlain by silt and clay rhythmites at the Sand River Property (photo courtesy of M. Smyk, OGS).

The rock surface has been polished and striated with at least two ice flow directions indicated (Fig. 15). The oldest set is oriented between 265-270°N and the youngest set at 290°N. Minor meltwater erosion forms are also present. The edges of the stripped area and trenches expose up to 2 m of glacial and postglacial sediments.

In one section, an interesting sequence of sediments is exposed. Resting directly on the bedrock surface is a thin (30 to 50 cm) layer of stony, slightly silty, sand diamicton interpreted as subglacial till (Fig. 16). Clast content is high and is composed of many locally-derived lithologies, dominated by metasedimentary rock fragments. The till is typical of areas of thin till within the Canadian Shield allowing for variations in lithology and stoniness. It has less than 2% carbonate content in the silt/clay fraction of the matrix. A sample of till from this layer contained 69% sand, 28% silt and 3% clay-sized particles (2 microns). In comparison a sample of till from Stop 6 contained 45% sand, 49% silt and 6% clay and had a matrix carbonate content of about 27%.

Overlying the till layer is 30 to 50 cm of rhythmically bedded silt and clay. These sediments were deposited in an ancestral lake located in at least the Lake Nipigon basin during ice marginal recession. A channel sample of the rhythmites contained 7% sand, 77% silt and 16% clay and its carbonate content was about 30%.

The rhythmites have been overturned toward the northwest (fold axis ~30°N) and are overlain by a thin layer of diamicton containing sorted lenses and wisps



Figure 17. Overturned rhythmites and overlying diamicton (photo courtesy of M. Smyk, OGS).



Figure 18. Transverse dune at stop 5. Inset shows internal structure (photos courtesy of M. Smyk, OGS).
of sand and gravelly sand (Fig. 17).

The diamicton may represent a minor fluctuation of the ice margin; overriding the glaciolacustrine rhythmites followed by ice marginal debris flow sedimentation (flowtills) or may be debris flows related to slumping or flowing of debris downslope; a post glacial event. The direction of overturning of the rhythmites, however, is similar to the direction of the younger set of striations and perpendicular to the slope direction at the site.

Stop 5 – Nipigon dunes

UTM coordinates – 0421102E 5497420N

At this stop a road cut exposes the internal structures of a transverse dune that probably formed along the shore of an ancestral lake in the Nipigon basin (Fig. 18). Thorleifson and Kristjansson (1983) mapped out a shorebluff several metres above present-day Lake Nipigon.

The second part of the stop is to a small field of dunes developed on an abandoned lake bottom that presently has an elevation of about 296 m asl. The route follows along the shorebluff mapped by Thorleifson and Kristjansson (1983), then crossing a subtle break in slope at about 178 m asl on our way to the dune field.

Stop 6 – Carbonate till

UTM coordinates – 0423418E 5483886N

In the 7 m exposure on the west side of Highway 11, approximately 6 m of light brownish grey, massive to fissile, slightly silty to silty, very fine to coarse grained sand with pebbles, cobbles and boulders, interpreted as

till, overlies the bedrock surface. The bedrock surface is striated with the striations oriented at 245°N.

The bedrock directly below the till is composed of fine to very fine grained sandstone, 20 to 30 cm thick, resting on Quetico metasedimentary rocks. The Archean - Proterozoic unconformity, an angular unconformity between the flat-lying Sibley Group sedimentary rocks and east-trending steeply dipping Quetico metasedimentary rocks, is exposed on both sides of Highway 11 (*UTM 0423431E 5483881N*).

The till fizzes when a weak solution of hydrochloric acid (~10%) is applied. It contains about 27% carbonate (11% calcite and 16% dolomite) in the silt/clay fraction (Chittick analysis). The till at Sand River (Stop 4) contained less than 2% carbonate. A sample of till from this site contained 45% sand, 49% silt and 6% clay-sized particles (2 microns). In comparison a sample of till from the Sand River Property (Stop 4) contained 69% sand, 28% silt and 3% clay.

This exposure of thick, carbonate-rich “exotic till” is one of the furthest west exposures. This exotic till is best exposed farther to the east, between Geraldton and Jellico especially around Wildgoose Lake. It underlies a spectacular streamlined surface (Fig. 19) where “drumlins up to 1 km in length are scattered across the area, but the principal morphological feature on thick till deposits are flutes about 1 km wide and over 20 km long” (Thorleifson and Kristjansson, 1983, p.54).

Zoltai (1965a) noted that the Nipigon Moraine, west of Lake Nipigon, corresponded to the extent of carbonate-rich till and Karrow and Geddes (1986) used the extent of carbonate-rich till to suggest that the



Figure 19. Streamlined till surface near Wildgoose Lake (photo courtesy of OMNR 74-4929-18-156).

Nipigon Moraine was the correlative of the Chapleau Moraine north of Lake Huron.

Stop 7 – Nipigon Lowland Rhythmites

UTM coordinates – 0407200E 5437660N

This stop is to look at some of the sediments that underlie the Nipigon lowland. At this site, 0.6 to 1 m of apparently massive slightly clayey silt to silt containing freshwater molluscs overlies greater than 14 m of silt and clay rhythmites (Fig. 20). The upper 6 m of the section is well exposed in a drainage ditch just east of Highway 11. The rhythmites are slightly contorted at the top of the unit then become horizontally bedded with depth. Concretions have formed around rootlets below 4.5 m.

In the woods to the east, a small abandoned sand and gravel pit occurs at the base of the bedrock-controlled hill and may possibly mark the position of the shore and level of the lake that the molluscs inhabited. Farrand (1960) identified a couple of terraces east of the highway and Helen Lake that he attributed to Nipissing/Algoma.

Acknowledgements

The writer would like to thank Mr James Nichols, Jim Nichols Trucking Ltd., Nipigon and Mr. Dave Malouf, Roxmark Mines Ltd., for kindly giving permission to visit their properties.

Mr Phil Kor (OMNR) and Mark Smyk (OGS) provided most of the photographs used in the fieldguide



Figure 20. Example of fresh water molluscs collected from Stop 7.

and Mark kindly provided text on the bedrock geology for several of the field trip stops. Roxanne Corcoran (OGS) produced some of the line illustrations. All of their help is much appreciated.

In addition, the indirect contributions of the contributors to the 1987 INQUA fieldtrip guide (Geddes et al., 1987) and Dr. W.R. Farrand (1960) for his work on ancestral lake levels in the Superior basin improved the content of this field guide. This fieldtrip guide was produced with the permission of the Director, Ontario Geological Survey.

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Fieldtrips 3 and 6 - A stratigraphic transect across the Northern flank of the Midcontinent Rift near Rossport

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Safety

As this trip will be taking place on Lake Superior it will be weather dependant and could be cancelled or curtailed at very short notice. An alternative trip to the Winston Lake mine will be undertaken in case of bad weather. Please exercise caution when getting in and out of the boats as the outcrops are often extremely slippery. Life jackets must be worn in the boats at all times. It will probably be very cold out on the lake so please dress warmly.

Regional geology

Archean granites, outcropping along the shoreline near Rossport, are unconformably overlain by strata of the Gunflint Formation (Fig. 1). These sediments were deposited on a south facing shelf at approximately 1878 Ma (Fralick et al., 2002). The Formation consists of a Lower Member composed of basal stromatolitic bioherms overlain by ankeritic, interclastic grainstones. A regressive, karstified surface caps the northern portion of this assemblage (Fralick and Barrett, 1995) and is succeeded by the Upper Member. It begins with a repetition of the underlying lithologies to which, higher in the succession, are added carbonaceous shales, tuffs and rarely mafic volcanic rocks. These chemical and fine-grained siliciclastic sediments record parasequence development on a storm-dominated shelf (Pufahl and Fralick, 2004) forming the relatively stable portion of a back-arc basin (Kissin and Fralick, 1994; Hemming et al., 1995) prior to compressive northward thrusting of the arc at approximately 1860 to 1835 Ma. As the compression of the Penokean Orogeny waned the sea again transgressed over the area depositing black, carbonaceous shales of the Rove Formation. This depositional cycle lasted from 1835 Ma (Kissin et al., 2003; Addison et al., 2005) till less than 1780 Ma ago (new detrital zircon laser ablation ICP MS data).

The lower portion of the Gunflint Formation in the

Rossport area is poorly exposed. The limited outcrop of the Lower Gunflint is similar to the succession comprising the thin, basal Kakabeka Conglomerate and overlying interclastic grainstones present in exposures to the west near Thunder Bay. Good exposure of the Upper Gunflint exists on Quarry Island and consists of possible basaltic flow rocks with associated stromatolites, overlain by a succession of medium- to coarse-grained, graded, sandstone beds. The geochemistry of the sandstones is similar to Archean rocks to the north indicating probable derivation from this source. Black shales, lithically correlative with the Rove Formation, outcrop on an island approximately 5 km to the west. The shales do not overly the turbidite succession on Quarry Island where arenites of the Sibley Group disconformably rest on an erosion surface at the top of the Gunflint sandstones.

The basal unit of the Sibley Group is the Pass Lake Formation. It is composed of the conglomeratic Loon Lake Member and the overlying sandstones of the Fork Bay Member (Cheadle, 1986). Where the basal conglomerates are present they either represent: 1) large channel fills cutting down into sandstones to shales with abundant caliche zones, or; 2) more laterally continuous conglomerates interbedded and overlain by parallel laminated, medium-grained sandstones. The former represent channel gravels in braided fluvial systems and the latter coarse-grained strandline deposits. The overlying Fork Bay sandstones likewise record both braided fluvial deposition and subaqueous, strand proximal sand-sheet development. In addition to upward thinning and fining successions developed during transgressive systems tract formation, other sandstone assemblages thicken and coarsen upwards representing progradational, delta lobe outbuilding. It is not known whether the water body was lacustrine or marine, though preliminary isotopic analyses (C, O, S and Sr) are more in agreement with a hypersaline lacustrine setting for at least the later history of the water mass. This is consistent with Cheadle's (1986) findings

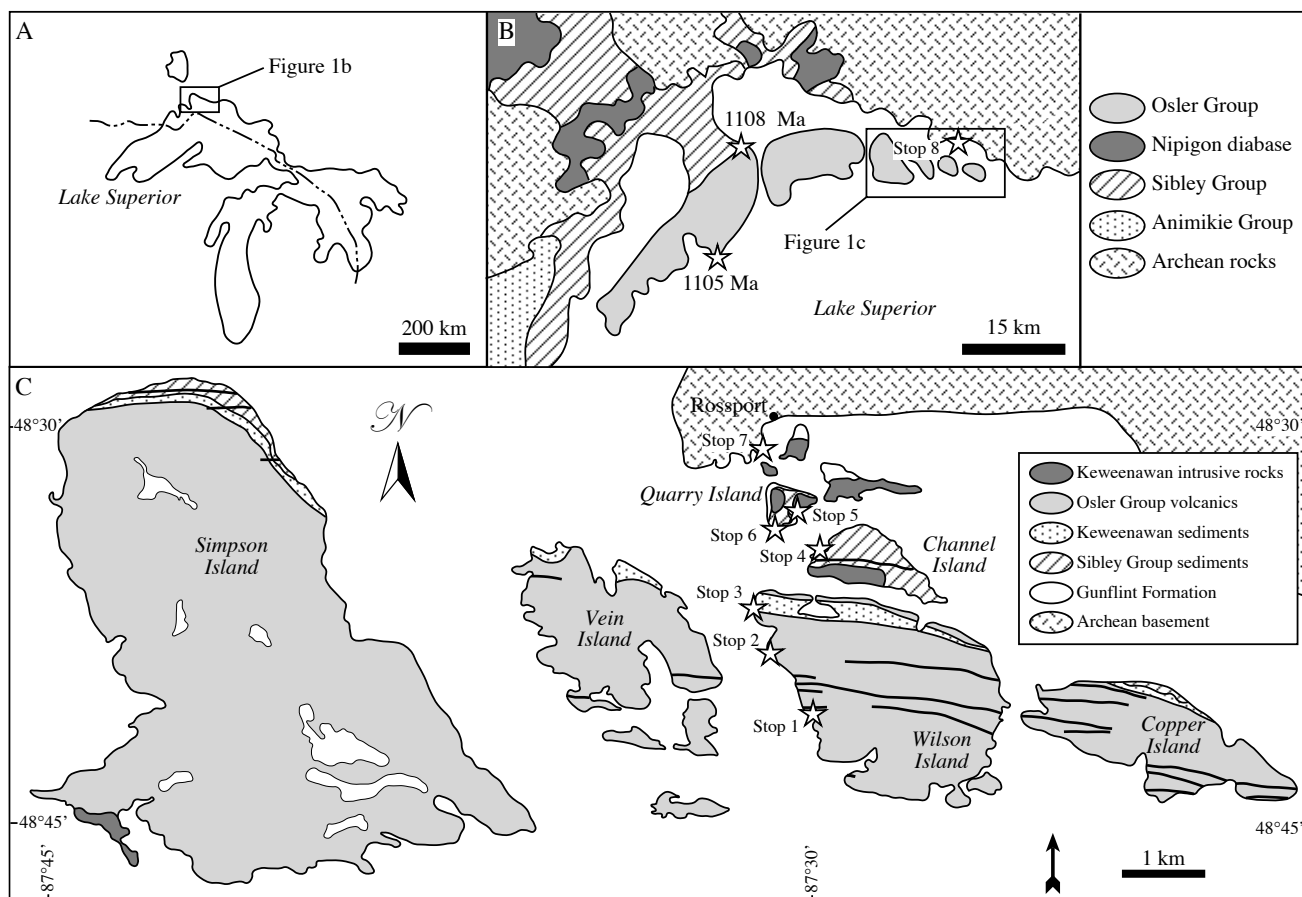


Figure 1. Map showing the location of the fieldtrip area. B) Regional geology map showing the extent of the Osler Volcanic Group. Age data from Davis and Sutcliffe (1985) and Davis and Green (1997). Modified after Sutcliffe (1986). C) Geological map of the Osler Volcanic Group showing sample locations. Modified after Giguere (1975).

based on regional paleogeography. The increasing salinity of the water resulted in dolomite, minor gypsum and rarer barite and celestite precipitation mixed with mud deposition. A red and green banding developed in this assemblage due to periodic anoxia of the bottom sediments. The final desiccation of the lacustrine basin is recorded by the development of strandline microbial bioherms (stromatolites) which are overlain by either a terra rosa (red, wind-delivered soil) or subareal, intraformational, mass-flow conglomerates. This is succeeded upwards by mudstones with abundant gypsum nodules representing mudflats formed in an arid climatic setting where hypersaline groundwaters precipitated gypsum. Together all these fine-grained sediments comprise the Rossport Formation. It is overlain by the Kama Hill Formation; a coarsening upwards deltaic succession recording flooding of the basin and development of a more humid climate. The age of the Pass Lake and Rossport Formations can be bracketed between laser ablation MS youngest ages on detrital zircons of 1600 Ma and a Rb-Sr isochron of

1339 Ma (Franklin, 1978). Youngest detrital zircons in the Kama Hill Formation are 1420 Ma.

The Sibley Group is very well exposed along the shorelines of the islands off Rossport. The basal disconformity can be seen about two thirds of the way up the cliff face on the western side of Quarry Island where it overlies graded sandstone beds of the Gunflint Formation. Blocks of medium-grained sandstone were extracted from the cliff face on the island for use in the construction of buildings in Thunder Bay. These sandstones are medium- and large-scale planar cross-stratified and may represent a sandflat composed of transverse bars in a braided stream or small barchan, eolian dunes. Rare pebbles indicate the former may be the case but this evidence is not conclusive. Channel and Copper Islands contain excellent exposures of the lacustrine rocks with outbuilding of channel systems along the paleolake margins. One of the best outcrops of the strandline stromatolitic carbonates occurs on Channel Island and will be visited during this fieldtrip. On Copper Island the Rossport Formation is overlain

disconformably by pebbly, fluvial conglomerates of the basal Osler. Thirty kilometers to the west the uppermost unit of the Sibley, the Nipigon Bay Formation, underlies the same disconformity. This highlights the fact that approximately 600 m of erosive downcutting occurred in the Rosspoint area before the basal Osler was deposited.

The oldest rift-related rocks on which U-Pb age determinations have been performed lie along the northwestern portion of the rift. These include, from NE to SW, the alkaline intrusives of the Coldwell Complex (1108±1 Ma, Heaman and Machado, 1992), the lower Osler Group volcanics (1108+4/-2 to 1105±2 Ma, Davis and Sutcliffe, 1985; Davis and Green, 1997), the

Logan Sills (1109+4/-2 Ma, Davis and Sutcliffe, 1985), the lower portion of the North Shore Volcanics (1108±2 Ma, Davis and Green, 1997), the Swamper Lake Gabbro and Nathan's Series intrusives (1107 Ma, Paces and Miller, 1993) and the lower portion of the Powder Mill Group (1107±2 Ma, Davis and Green, 1997). These older units outcrop on the rift flanks where erosion has removed the younger rift sequence or, in the case of the Coldwell and Logan Sills, are intruded into older rocks immediately north of the rift.

The Osler Volcanic Group comprise a ~3km thick sequence (Cannon et al., 1989) lying unconformably above Sibley Group metasediments (Fig. 2). The volcanic sequence is overlain and intruded by the

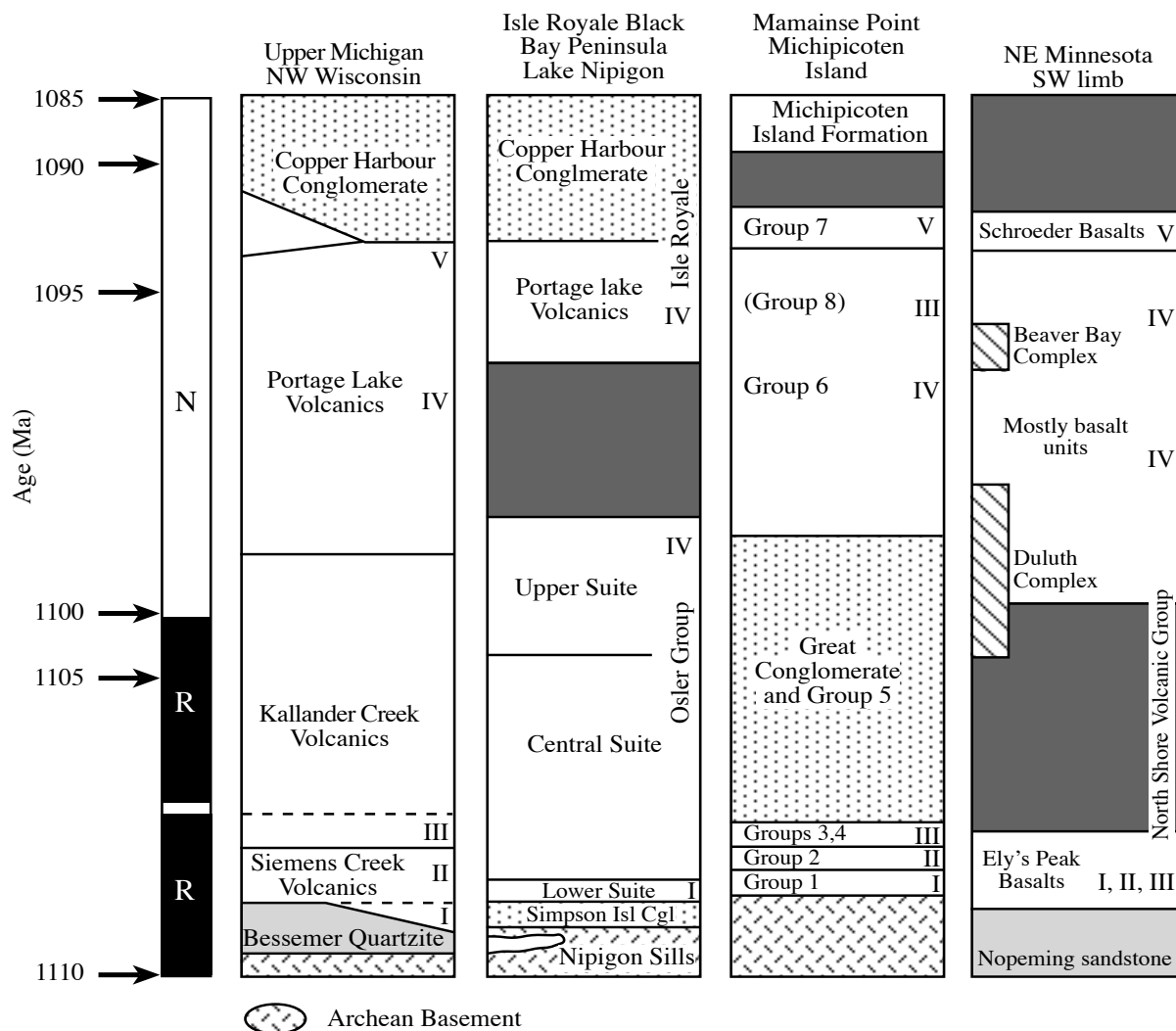


Figure 2. Schematic correlations of volcanic rocks of the Midcontinent Rift based on the stratigraphic position of distinctive basalt sequences, magnetic polarity and absolute age where possible. Modified after Nicholson et al. (1997). Dashed lines in Upper Michigan section separate lower and upper members of Kallander Creek and Siemens Creek volcanics. Left hand column shows magnetic polarity. Roman numerals I-IV refer to five distinctive laterally extensive basalt compositions identified on the south shore of western Lake Superior. Where equivalent basalt compositions occur in other stratigraphic successions, the appropriate Roman numeral is noted (see Nicholson et al., 1997 for data sources). Shaded regions represent intervals in which contacts are covered or obscured by plutonic rocks.

St. Ignace Island Volcanic-Plutonic Complex an intercalated sequence of basaltic rocks and rhyolitic flows (Sutcliffe and Smith, 1988). Detailed descriptions of the Osler Group have been provided by McIlwaine and Wallace (1976) and Lightfoot et al. (1991). Generally the mafic flows of the Osler Group consist of massive to amygdaloidal flows, with locally developed ropey tops and pahoehoe textures (Sutcliffe and Smith, 1988). The flows range in thickness from 5cm to 30m (Lightfoot et al., 1991) with a regional dip of ~6-15° S (Giguere, 1975; Lightfoot et al. 1991; Hollings et al., 2005). The majority of the exposed section is magnetically reversed with only the upper 100m displaying a normal polarity (Halls, 1974). The contact between the two units is marked by the presence of the Puff Island conglomerate and a discordance between the basalt flows above and below the contact. This has been interpreted as representing a significant break in the eruption history. A felsic porphyry near the base of the Osler Group has yielded an age of 1107.5±4/-2 Ma (Davis and Sutcliffe, 1985) whereas zircons from the Agate Point rhyolite towards the top of the reversely magnetized sequence have yielded an age of 1105±2 Ma (Fig. 1b; Davis and Green, 1997).

Within the Osler Group interflow sediments are typically thin and of limited extent. Field descriptions of the sedimentary successions appear in Giguere (1975) and McIlwaine and Wallace (1976). They show that there are two main zones of sedimentary rocks within the Osler Group. One occurs near the base of the volcanic pile. The other is present approximately 2700 meters higher in the succession marking the paleomagnetic reversal.

Lightfoot et al. (1991) in a study of the Osler Volcanic Group exposed along the shores of the Black Bay Peninsula to the west of the field trip location proposed that the major and trace element geochemical data could be used to subdivide the flows into an Upper, Central and Lower Suite although the boundaries between the suites were not clear cut. While the geochemical compositions of the Central (750-900m) and Upper suites (1900-3000m) overlap their Lower Suite (0-750m) is distinguished by elevated Mg numbers (0.55-0.7 versus 0.3-0.6), lower Al₂O₃ (8-12 wt% versus 13-17wt%), lower Th/Nb ratios (0.09-0.70 versus 0.3-0.6) but higher Gd/Yb_n ratios (3.5-4.5 versus 1.6-2.6). Nicholson et al. (1997) concluded that there were five geochemically distinct flood-basalt compositions within the Mid-continent rift that are

common to most sections and appear in approximately the same stratigraphic order (Fig. 2) They recognized a lower suite in the Siemens Creek Volcanics (Basalt Type 1; Fig. 2), which they suggest is analogous to the Lower Suite of Lightfoot et al. (1991). In the United States this unit is less than 100m thick whereas the Lower Suite of Lightfoot et al. (1991) is ~750m thick. They report a narrow range of $\epsilon_{\text{Nd}(1100)}$ values for the Siemens Creek Volcanics of -0.7 to +0.7. Nicholson et al. (1997) further suggest that there is a broadly recognizable suite of basalts above this (Basalt type II) which includes the upper Siemens Creek Volcanics and in the upper part of the Grand Portage lavas (Fig. 2). The suite is characterized by slight negative Nb anomalies and a range of $\epsilon_{\text{Nd}(1100)}$ values of -1.4 to -6.9. They suggest that this may be analogous to the most primitive members of the Central Suite of Lightfoot et al. (1991).

The volcanic flows of the Osler Group on Wilson Island are all basalts or basaltic andesites (SiO₂ = 47-56 wt%; MgO = 5-16 wt%; Hollings et al., 2005). The basalts are characterized by LREE enrichment (La/Sm_n = 1.5-3.9) in conjunction with moderately fractionated HREE (Gd/Yb_n = 1.5-3.7) and slight positive to moderately negative Nb anomalies (Nb/Nb* = 0.56-1.13; Hollings et al., 2005; Fig. 3). Major and trace element data show trends of increasing SiO₂ and decreasing MgO and display strong positive correlations between La/Sm_n, Th/La and Th/Nb with height (Fig. 4). This correlation is most pronounced above 400m with samples from the base of the stratigraphy displaying more or less constant values of these ratios (Fig. 4). Measured ¹⁴³Nd/¹⁴⁴Nd ratios for the seven Osler basalts analysed range from 0.511857-0.512286 with $\epsilon_{\text{Nd}(t=1106\text{Ma})}$ of +0.3 to -5.3 (Hollings et al., 2005). The high incompatible element abundances, in conjunction with LREE enrichment and strongly fractionated HREE are comparable to modern OIB, albeit at lower absolute abundances (Fig. 3). When compared to other flood basalt sequences the more primitive basalts from this study closely resemble basalts from the Parana-Etendeka flood basalt sequence (Fig. 4; Gibson et al., 2000). The ϵ_{Nd} data from the most primitive members of the Osler Group is consistent with an enriched mantle plume rather than a contaminated depleted mantle source, given the lack of trace element evidence for contamination in these samples. Depleted mantle at 1100 Ma would have had a positive ϵ_{Nd} perhaps as high as +6 whereas an

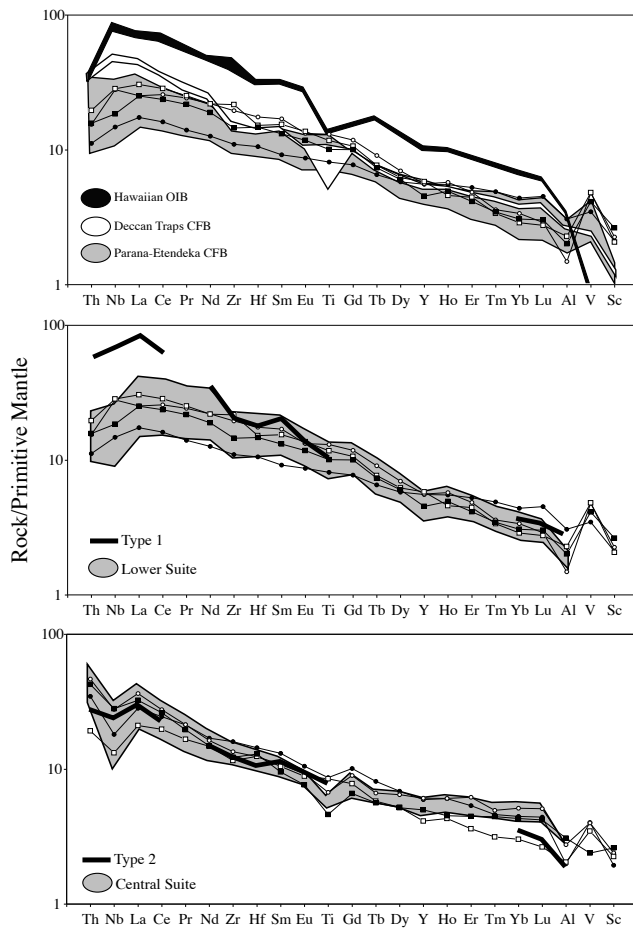


Figure 3. Comparison of primitive mantle normalized plots from the Osler Group with A) Phanerozoic OIB and Continental Flood Basalts (CFB) and B & C) the Lower and Central suites of Lightfoot et al. (1991). From Hollings et al. (2005).

enriched plume source would have $\epsilon_{Nd} \sim 0$ (Nicholson and Shirey, 1990; Shirey et al., 1994). Up sequence the basalts are characterized by higher SiO_2 , Th and La/ Sm_n abundances in conjunction with increasingly negative Nb and Ti anomalies and $\epsilon_{Nd(1106)}$ values of -4 to -5. This is consistent with contamination of these basalts

by an older lithospheric component characterized by pronounced LREE enrichment, high Th abundances but generally unfractionated HREE (Hollings et al., 2005).

The sedimentary successions near the base of the Osler Group constitute the Simpson Island Formation and have recently been described in detail by Hollings et al. (2005). They are composed of a Lower Member dominated by trough cross-stratified, medium-grained sandstones directly overlying basement and an Upper Member with a greater variety of siliclastic units. The Lower Member sits on an irregular, erosional surface cut into the underlying quartz arenites of the Nipigon Bay Formation, Sibley Group. A massive pebble-cobble conglomerate overlies the unconformable surface and is in turn overlain by decameter-scale layers of coarse-grained and pebbly sandstone (Fig. 5, Section 1). Sandstones are parallel laminated, commonly have cross-stratified tops, more rarely contain pebbly transverse ribs and chute and pool-like structures. The central portion of the succession is composed of trough cross-stratified, medium-grained sandstone organized into a stacked assemblage of lenses. Pebble stringers and pebbly sandstones commonly occur on the deeper portions of curving set boundaries (Fig. 5, Section 2). Massive pebble-cobble conglomerates sharply overlie the sandstone succession (Fig. 5, Sections 2 and 4). The conglomerates contain trough cross-stratified, medium-grained sandstone lenses; decameter- to meter-scale wedges of planar cross-stratified sandstone and are interbedded with assemblages of trough cross-stratified sandstones up to one meter thick. Another assemblage of trough cross-stratified sandstone, similar to the one in the central portion of the succession, caps the basal sedimentary assemblage (Fig. 5, Sections 3 and 4). Clast lithologies in the pebble-cobble population are dominated by quartz, chert, various types of volcanic

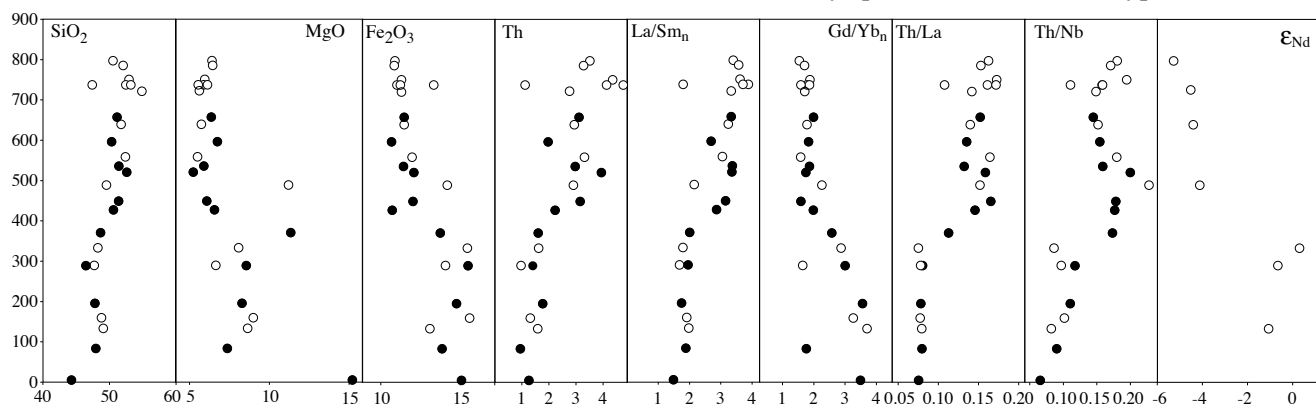
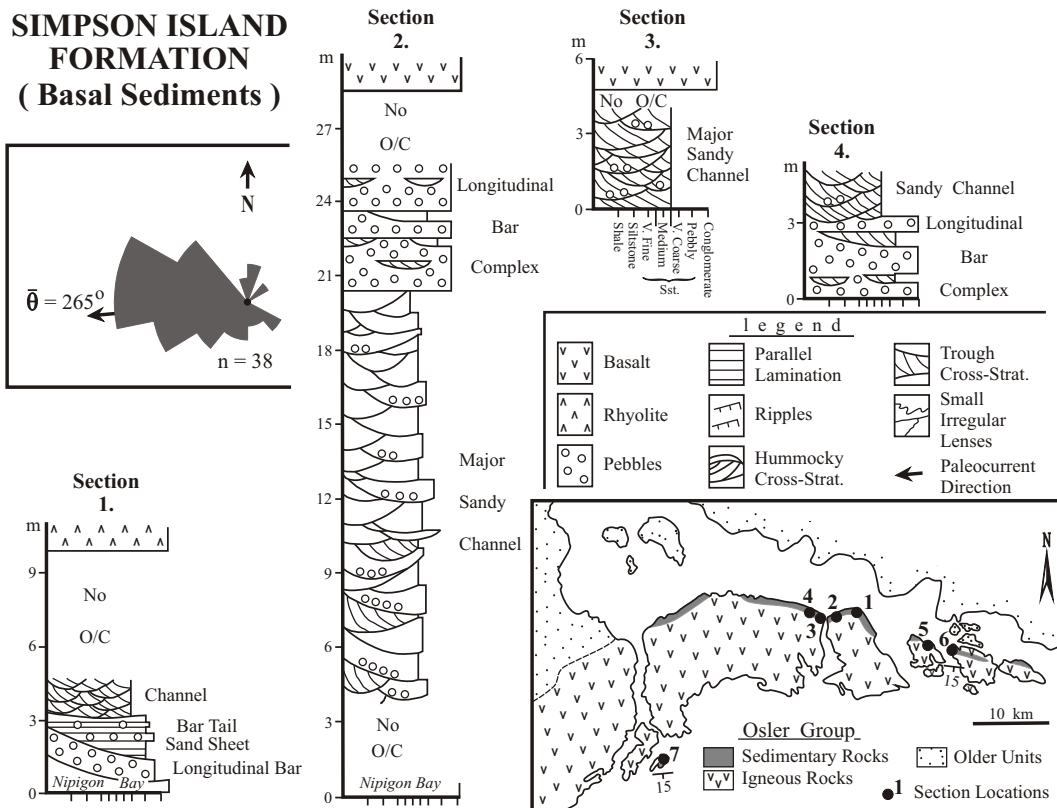


Figure 4. Geochemical stratigraphy of the Osler Group on Vein and Wilson Islands. The stratigraphic position of the samples has been calculated assuming a dip of 10° parallel to the section. From Hollings et al. (2005).



INTERFLOW SEDIMENTS

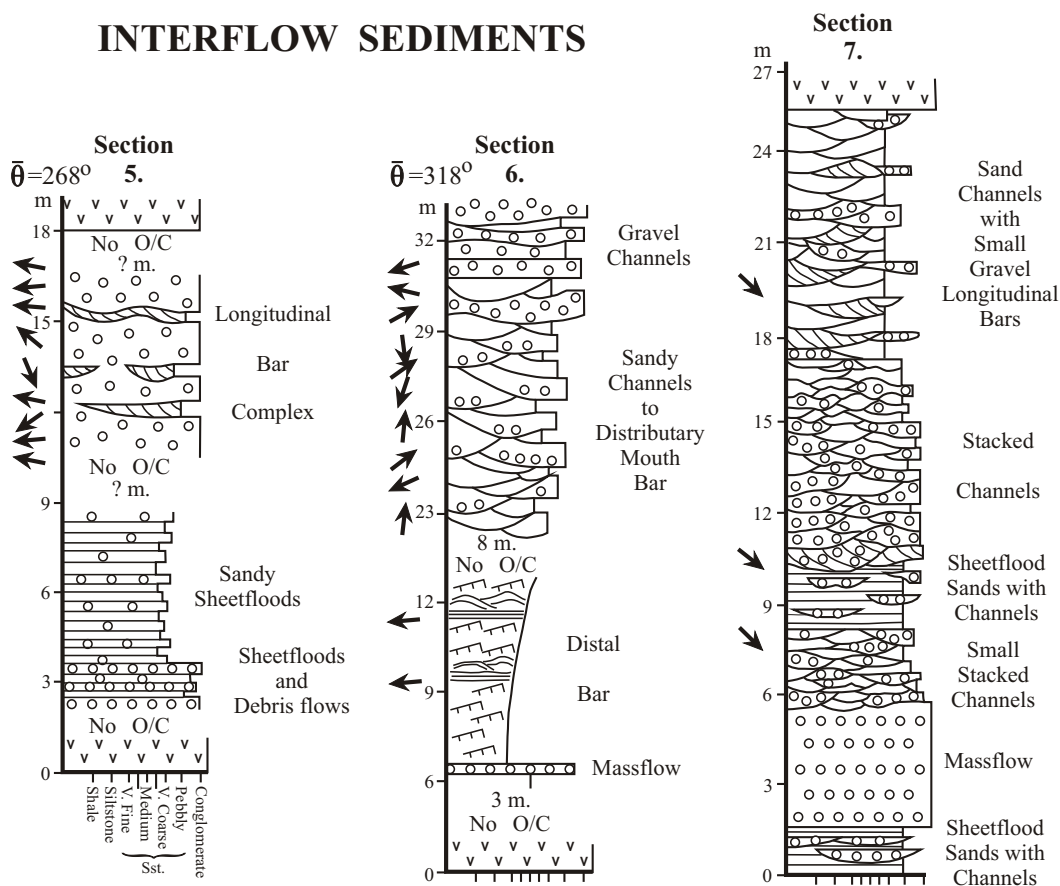


Figure 5. Sections of sedimentary rocks in the Osler Group. Sections 1 through 4 are the basal sedimentary succession of the Lower Member, Simpson Island Formation, at different locations (see inset map). Sections 5 and 6 are of the Upper Member, Simpson Island Formation, interlayered with basal basalt flows. Section 7 is the sedimentary assemblage near the top of the Osler Group on Puff Island. From Hollings et al. (2005).

rocks, red siltstone, metamorphosed granite and at the east end of the outcrop belt, on Copper Island, a higher proportion of unmetamorphosed red granite. Current indicators show flow was to the west, averaging 265°.

A sedimentary assemblage also occurs near the upper limit of outcrop of the Osler Formation, at the top of the magnetically reversed interval (Fig. 5, Section 7). These interflow sediments are located on Puff Island and overly a felsic porphyry with a U-Pb age of 1105 Ma (Davis and Green, 1997). They contain: sharp sided assemblages of laterally continuous, pebbly, coarse-grained sandstone beds with caliche horizons which are scoured into by small lenses of conglomerate; large scours filled with trough cross-sets over a meter thick; stacked assemblages of irregular lenses filled by trough cross-stratified, coarse-grained, pebbly sandstone; and, poorly sorted, disorganized, massive boulder-cobble conglomerate. Clasts are all volcanic, ranging from quartz-feldspar porphyries to mafic compositions. Paleocurrents on large-scale sedimentary structures consistently show flow to the southeast.

The Simpson Island Formation is composed of a laterally continuous sedimentary succession up to 25 meters thick and discontinuous sedimentary units up to 30 meters thick interlayered with the basal basalt flows. The lowest sedimentary beds fill channelways cut into the underlying sandstones of the Nipigon Bay Formation. The channel fills and overlying sedimentary assemblage represent a braided stream system, similar to the South Saskatchewan model (Miall, 1978), where dunes composed of coarse-grained sand migrated down the channels and gravelly longitudinal bars with chute channels and bar edge sand wedges form the higher relief areas (Fig. 5). The fluvial interpretation is consistent with Tanton (1931) and McIlwaine and Wallace (1976). Clast lithologies indicate debris was mainly derived from erosion of local lithologies.

Stops

The trip will depart from the public dock at Rosport and will undertake a traverse through the stratigraphy of the Midcontinent Rift, starting with the youngest rocks of the Osler Volcanic Group, proceeding through the Sibley and Gunflint Formations and finishing with a look at the granites of the Archean basement (Fig. 1). In order to make the most of the calmer weather typical of early mornings we will first travel for approximately 30 minutes to the most southerly outcrop on Wilson

Island.

Stop 1 – Osler volcanics, Wilson Island

UTM coordinates – 0462794E 5402095N

Flows of the Osler Group on Wilson Island are typically >1m thick, frequently amygdaloidal towards the top and bottom of the flow with a massive core and rarely displayed a pahoehoe texture on the flow surface. The basalts are characterized by clinopyroxene and plagioclase phenocrysts in a groundmass of plagioclase, augite and Fe-Ti oxides. Rarely, basalts from the base of the sequence contained pseudomorphed olivine phenocrysts. The basalts have all been subjected to low-grade metamorphism ranging from zeolite to prehnite-pumpellyite facies (McIlwaine and Wallace, 1976). At this stop, approximately 500m above the basal conglomerates, are exposed a sequence of thin rubbly basalt flows ~50cm thick with rare massive flows ~2-4m thick and thin interflow sediments (Fig. 6). Geochemically basalts at this outcrop are similar to the Central Suite of Lightfoot et al. (1991). Sedimentary units are predominantly quartz sandstones with thin shale partings. These are best interpreted as sands washing into small hollows on the surface of the flow units with a mud drape settling out towards the top of the layer. In places units that appear to have been deposited on the surface of basalt flows connect into sub-vertical cooling cracks in the flows. Sediment filled cooling cracks can be up to 2m deep (Fig. 7).

Stop 2 – Osler Volcanics, North end of Wilson Island

UTM coordinates – 0461810E 5403438N

Exposed at this outcrop, are the lower flows of the Osler Volcanic Group ~300m above the conglomerates



Figure 6. Thin interflow sediments between basaltic flows, Stop 1 on Wilson Island.



Figure 7. Approximately 2m thick mafic flow cut by sediment filled cooling crack. Stop 1, Wilson Island.

of the Upper Simpson Island Formation. The basaltic flows are generally massive, ranging in thickness from 1-3m. Flow tops vary from rubbly to well-developed pahoehoe textures (Fig. 8). The basalts are vesicular and amygdaloidal and in places the vesicles are elongated giving them almost a pipe-like appearance. Geochemically basalts at this outcrop are similar to the Central Suite of Lightfoot et al. (1991). In some areas the basalt flows incorporate xenoliths of a more vesicular material which have both sharp and diffuse contacts (Fig. 9) At the north end of the outcrop the flows are cut by a 2-3m wide mafic dyke. This dyke is geochemically distinct from the flows but comparable to the older diabase intrusions in the vicinity of Lake Nipigon.

Stop 3 – Upper Simpson Island Formation, Daylight Point, Wilson Island

UTM coordinates – 0461450E 5404650N

A sedimentary assemblage of the Upper Member occurs on Wilson Island, overlying approximately 50 meters of basal basalt. This coarsening upwards succession has oscillation rippled, very fine-grained sandstones at its base (Fig. 5, Section 6). These coarsen upwards by the addition of increasing amounts of medium-grained, parallel laminated to hummocky cross-stratified to oscillation rippled, decimeter-scale



Figure 8. Pahoehoe texture at Stop 2, Wilson Island.



Figure 9. Amygdaloidal xenoliths in basalt flows at Stop 2, Wilson Island.

sandstone beds (Figs. 10, 11). A covered interval separates this assemblage from overlying medium- to large-scale, trough cross-stratified, coarse-grained sandstones to conglomerates (Fig. 12). Paleocurrent indicators show flow to the west, though with a higher variance than other sections. Clast lithologies are probably locally derived from both Archean and Proterozoic sources. The fine-grained sandstone near the base of the section represents a wave modified delta-front (ie., a distributary mouth bar of a small delta). The presence of small, dish-shaped scours suggests a shallow water environment with no large channels. The upper part of the sequence represents badly organized river deposits with gravelly, longitudinal bar forms and channels filled with sand. The planar cross sets at the base of the cliff were formed by transverse bars, while the trough cross beds represent migrating dunes.



Figure 10. Fine-grained red sandstones with thin shale partings forming the base of the deltaic deposits at Stop 3.

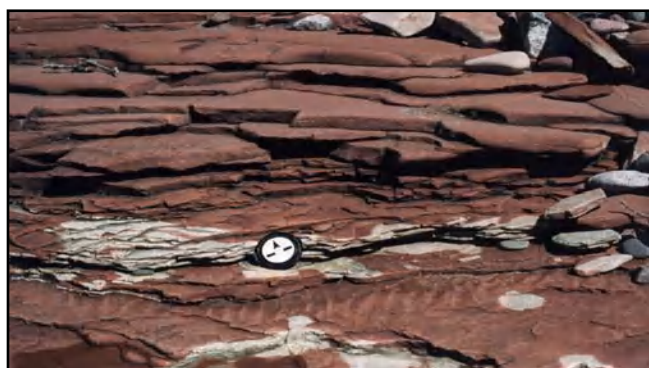


Figure 11. Oscillation ripples with overlying hummocky, medium-grained sandstone showing the effects of wave reworking on the sediments forming the delta front at Stop 3.



Figure 12. Cross-stratified sandstone and massive conglomerate forming the upper portion of the prograding deltaic succession present in the Upper Member of the Simpson Island Formation present at Stop 3. These sediments represent a longitudinal bar-channel complex of a braided stream.

Stop 4 – Sibley Group, Mary Ann Bay, Channel Island

UTM coordinates – 0462831E 5405801N

Note - this is a very small outcrop and access to the rocks may be limited if lake levels are high. Please take care moving from the landing point to the outcrop.

A succession of grey dolomites interbedded with

red siltstones outcrop on the shoreline here (Fig. 13). These are interpreted to be part of the cyclic facies (Channel Island Member) of the Rossport Formation as exposed at Kama Hill. The dolomite and minor gypsum indicate a hypersaline environment interpreted to be lacustrine because of the multiple deltas and sand sheets building in from a variety of directions. The cyclic facies is overlain by stromatolites and interbedded thin, sandy, carbonate storm layers (Fig. 14) of the Middlebrun Bay Member. This meter thick assemblage is similar to Recent sabkha deposits on the south shore of the Persian Gulf, and in particular the open water ponds on this sabkha where sandy storm layers are well preserved. The upper few centimeters of the stromatolitic unit is altered to a grey-green layer that represents a weathered horizon interpreted to



Figure 13. Interlayered red siltstones and dolostones (lower unit underlying the more massive strandline carbonate with overlying mass-flow deposits) were deposited in a saline lake away either temporally or spatially from areas of coarse sand influx. The colour banding reflects the position of the redox boundary as the sediments accumulated. The grey layers commonly have slightly higher dolomite contents possibly reflecting higher organic productivity leading to more photosynthetically mitigated carbonate precipitation (higher dolomite content) and heavier organic loading to the sediment (redox boundary moving upward to at or above sediment water interface). Stop 4, Mary Ann Bay.



Figure 14. Stromatolitic layering (smooth mat with small pinnacles) with interbedded coarse silt to very fine-grained sand storm layers (white). This is typical of strandline to sabkha environments and especially open water ponds on the sabkha. Some of the storm layers were remobilized in the form of clastic dykes and sills.

have formed as the sequence became subaerial and the stromatolites weathered in situ. This weathered zone is traceable throughout the basin with the strandline deposits below it commonly containing well developed tepee structures. The carbonates are overlain by a mass-flow unit with intraformational clasts of red siltstone, sandstone and dolostone up to boulder size. Although the contact between the carbonates and the mass flow units is locally obscured by the intrusion of a sill, the transition is interpreted to represent a minor time gap based on the weathered zone which expands to thick terra rosa (soil) deposits at other locations.

Stop 5 – Gunflint Formation, Quarry Island

UTM coordinates – 0462305E 5406730N

A succession of sandstones and mafic volcanic rocks outcrop on the south shore of Quarry Island. On the northeastern end of the outcrop area a gabbro, probably related to the Midcontinental Rift, is exposed. Next to this is a small outcrop of stromatolites with a box-like appearance (Fig. 15). The rectangular to square outline of the mounds contrasts with the round to oval appearance of classic stromatolites, though their organic origin is exemplified by the high angle layering, which, when projected into the area now eroded, can be seen to form mounded structures. Areas between the stromatolites are infilled with coarser siliciclastic sandstones and cherty clasts. The next outcrop of



Figure 15. Odd shaped structures of probable stromatolitic origin. within the Gunflint Formation. Stop 5, Quarry Island

Gunflint volcanic rocks is problematic. Mafic volcanic flow rocks occur interbedded with Upper Gunflint lithologies southwest of Thunder Bay. These are also associated with stromatolites that developed on the firm substrate of the flow tops. Thus, the igneous rocks in the Gunflint assemblage on Quarry Island could be correlative to the other flow rocks, but it is difficult to conclusively show that these rocks are extrusive. Possible flow banding is present, as are areas of finer and coarser material in individual units. The igneous rocks are overlain by medium- to coarse-grained sandstones with bed thicknesses averaging approximately 30 cm. The sandstones are dark in appearance giving the impression they were derived from mafic detritus, but their geochemistry indicates an intermediate source similar in composition to the Archean crust to the north. The layers are excellently graded (Fig. 16) with the only sedimentary structure being sporadically developed parallel lamination. Beds such as these are commonly thought of as typical turbidites and the deposits ascribed to reasonably deep water. However, it must be remembered that graded bedding simply means a decelerating flow deposited the bed, which can occur in any water depth. These beds may be tempestites, ie. beds formed by storm events, in this case in water deeper than storm wave base but certainly not anything approaching abyssal depths. Or they may have formed



Figure 16. Well graded sandstone layers. Though these are probably turbidites they were not deposited in excessively deep water and may, in fact, represent distal tempestites. Stratigraphic position is problematic as these beds may belong to either the Upper Gunflint or Rove Formations. Stop 5, Quarry Island.

from inter- or overflow sediment-water plumes off river mouths, though lack of current reworking of bed tops makes this unlikely. Alternatively they may represent prodelta deposits formed by slumping of the delta front. All of these environments are relatively shallow which agrees with the presence of stromatolites not far stratigraphically below the graded beds. Another



Figure 17. Unusual markings on the bedding planes of the graded sandstones. Stop 5, Quarry Island.



Figure 18. A second example of unusual marking observed at Stop 6. The origin of these markings is unclear. Stop 5, Quarry Island

interesting point concerning these clastic units is that although such sandstones are common in the upper Rove Formation they are not present in the Gunflint at any other location. Thus, their stratigraphic position is debatable. The third unusual attribute is the presence of difficult to interpret structures on some bedding planes. Series of enechelon small crack-fill like features cut across bedding planes (Fig. 17). In addition a jellyfish-like impression was found on a bedding plane (Fig. 18). This feature had five-fold symmetry, similar to echinoderms, but the age of the rocks and its presence in sandstone leads to the distinct possibility that it was manufactured.

Stop 6 – Pass Lake Formation, Quarry Island

50 meters inland from the last shore outcrop (Stop 5) where the iron train rail lies across the cobble-boulder beach - **this stop will be time dependant.**

This outcrop consists of a cliff-face in sandstones, which were quarried, and the blocks produced used in the construction of buildings in Thunder Bay (Fralick et al., 2000). Here we see the basal sediments of the Sibley Group, the Pass Lake Formation. The Pass Lake forms a diverse group of basal coarse clastic deposits representing environments ranging from braided fluvial through to subaqueous sand sheets. The medium-grained sandstones present in this cliff are organized into a series of large-scale planar cross-stratified sets with normal to low dip angles. Sorting is fairly good and only one pebble has been found in the succession. Assemblages such as this pose a dilemma in formulating an interpretation of their depositional environment. Both aeolian sand dunes and sandflats composed of



Figure 19. Basal conglomerate of the Gunflint exposed at Gut Point. Photo courtesy of Mark Smyk.

transverse bars in braided rivers are capable of creating such an organization of lithofacies. The presence of the pebble is significant as only freak wind-storms, such as tornadoes, can move material of this size, and these do not form sand dunes. So, it is more likely that these deposits are subaqueous but this rests only on a slim piece of evidence.

Stop 7 – Basal conglomerate of the Gunflint, Gut Point

UTM coordinates – 0461610E 5408587N

The unconformity and basal Gunflint are exposed at Gut Point as a thin, discontinuous veneer along the lakeshore on top of Archean basement (Fig. 19). The basement is a medium-grained, equigranular granite, which has been altered (saussuritized/chloritized) beneath the basal Gunflint. The basal conglomerate is up to 30 cm thick and occupies depressions in the paleo-erosion surface in the basement. The conglomerate



Figure 20. Matrix supported basal conglomerate of the Gunflint Formation containing rounded pebbles of quartz, chert and lithic fragments. Photo courtesy of Mark Smyk.



Figure 21. Black chert with gossan within the basal conglomerate of the Gunflint Formation at Gut Point. Photo courtesy of Mark Smyk.

is matrix-supported, with subangular to rounded pebbles of white, sugary quartz, lesser cherty and lithic fragments and minor jasper in a medium-grained, sandy matrix (Fig. 20). A black, pyritiferous chert breccia is marked by a conspicuous gossan (Fig. 21). Sulphide mineralization may be related to a persistent, parallel fracture set at 115°. Fractures may host quartz-calcite-barite veins ranging from < 1 to 20 cm wide as well as vein breccia. A conjugate fracture set at right angles to the first is locally developed. Silicification adjacent to the veins has preserved a thin (1 to 5 cm) veneer of Gunflint from being eroded (especially the carbonate units). A 2m wide diabase dyke strikes at 115° through the outcrop.

Stop 8 – Archean basement, Selim Point

UTM coordinates – 0469219E 5409146N

From the dock in Rosport return to Highway 17 and head east for ~5 km. Turn right on to Lakeshore Drive just west of Whitesand Provincial Park. Follow the dirt road to a parking spot opposite a small tombola (Fig. 22). The porphyritic granite exposed here is Archean in age and part of the Wawa Subprovince. The area was mapped by Carter (1988) who described the rocks as porphyritic pink, hornblende + biotite alkali feldspar granite, a phase of the Whitesand Lake Batholith. The porphyritic “facies” is surrounded by massive pink and grey phases of alkali feldspar granite that is not exposed at this locality. The batholith is about 8 x 16 km in size and intrudes the Schreiber greenstone belt; no radiometric date has been generated. Feldspar phenocrysts are typically 3-4 cm across (Fig. 23), subhedral to euhedral and in places appear to display localized alignment suggestive of flow banding.



Figure 22. Porphyritic Archean granite, Selim Point.



Figure 23. Feldspar phenocrysts in Archean porphyritic granite at Selim Point.

Acknowledgements

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Fieldtrip 4 - Geology and rare element pegmatites of the Quetico Subprovince near Nipigon

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Overview

The Quetico Subprovince of the Superior Province is situated between the Wabigoon and Wawa volcano-plutonic subprovinces that bound the Quetico on its northern and southern margins, respectively. This east-trending subprovince has a fairly consistent width of 70 km and consists predominantly of metasedimentary rocks and their migmatitic and anatectic derivatives (Williams, 1991). In general, the boundaries of the Quetico, whether or not they may be primary and/or tectonic, have been mapped as steeply dipping surfaces across which there is commonly a distinct contrast in lithology. In northwestern Ontario, the Quetico has been defined to exclude mafic metavolcanic rocks and derived migmatites; these rocks are instead assigned to the bounding Wawa and Wabigoon subprovinces.

An overview of the lithologic, metamorphic, structural and tectonic characteristics of the Quetico Subprovince has most recently been provided by Easton (2000):

“The intensity of metamorphism varies within the subprovince, such that rocks marginal to the subprovince tend to be at lower grade than in the interior. The lowest metamorphic grade is found along the northern boundary with the Wabigoon subprovince (Pirie and Mackasey, 1978). Locally, subgreenschist- to greenschist-facies rocks occur along the southern boundary (Borradaile, 1982), but typically, there is a rapid rise in metamorphic grade north of the Wawa subprovince, especially north of Manitouwadge, where a belt of metasedimentary granulites occurs within the Quetico subprovince close to, and parallel with, the northern margin of the Wawa subprovince (Coates, 1968; Williams and Breaks, 1989, 1990; Pan et al., 1994). As a result, grade distribution is asymmetrical, with the maximum in temperature and pressure occurring south of the central Quetico, locally coincident with the southern margin.”

In contrast, Seemayer (1992) also described an asymmetric metamorphic grade distribution that had metamorphic grade increasing from south to north across the Quetico, southwest of Lake Nipigon. The southern margin was characterized by greenschist-facies rocks, the central portions were at amphibolite facies and the northern margin displayed an abrupt decrease in grade adjacent to the Wabigoon Subprovince to the north. The Quetico Fault, which is normally situated at the northern margin of the Quetico, lies well within the Quetico southwest of Lake Nipigon (Seemayer, 1992). Seemayer (1992) determined temperatures based on garnet-biotite thermometry ranging from 526°C at the southern margin of the Quetico Subprovince, increasing asymmetrically to a maximum of 714°C, then falling sharply at the northern margin to 517°C.

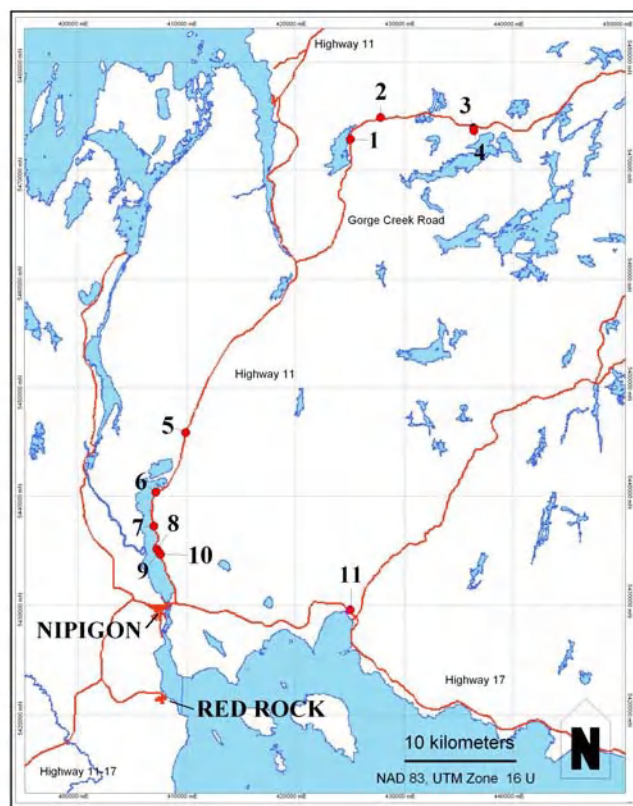


Figure 1. Field trip stop locations.

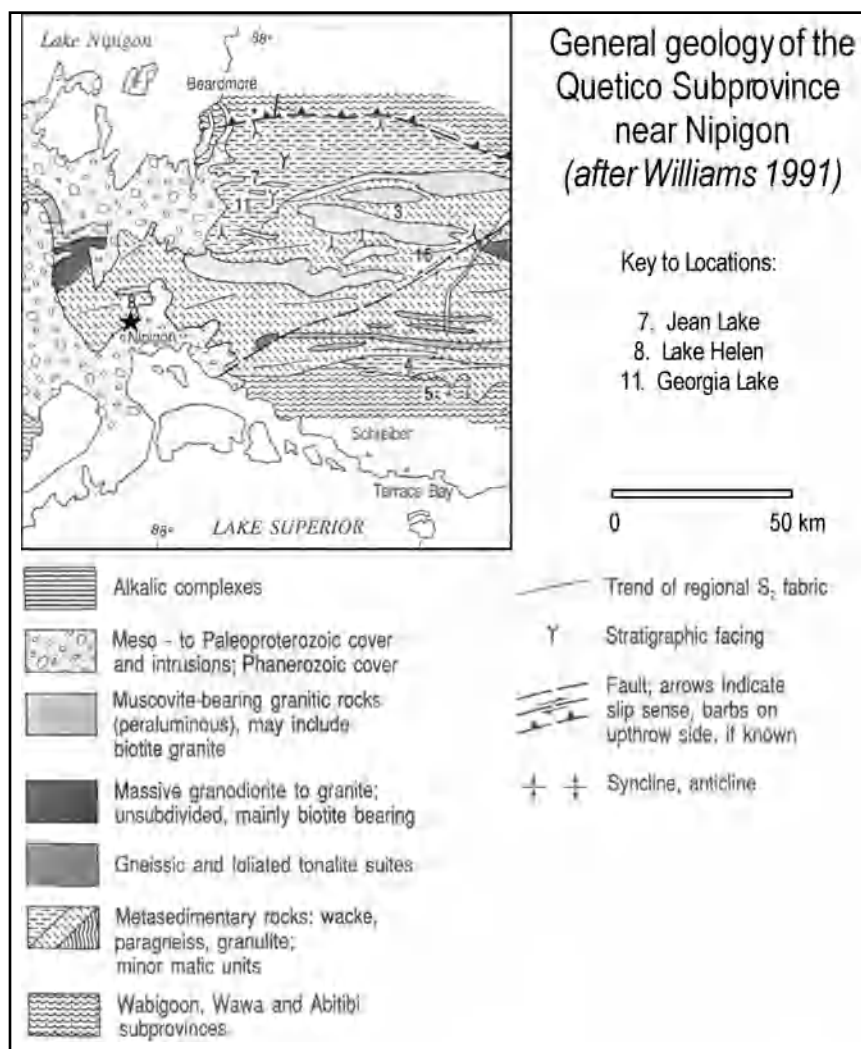


Figure 2. Generalized regional geology (after Williams, 1991).

Pressure calculated at near peak temperature was 5 ± 1.5 kbar.

Easton (2000) described the regional metamorphic conditions and metamorphic history:

“P–T conditions increase from west to east, for example, 500°C and 2.5 kbar at the Minnesota border west of Thunder Bay (Percival, 1989), to 700–780°C and 5.4–6.1 kbar adjacent to the Kapuskasing structural zone (Percival and McGrath, 1986; Percival, 1989). Typical conditions in the central region are on the order of 620°C and 3.3 kbar (Percival, 1989). Granulites north of Manitouwadge yield 680–770°C and 4.4–6.4 kbar (Pan et al., 1994; Percival, 1989). The regional variation in P–T can be ascribed to a relatively shallow level of erosion in the west (<10 km) and a deeper level in the east (>12 km) (Percival et al., 1985). Rocks located east of the Kapuskasing structural zone are believed to be generally at upper-amphibolite-facies conditions (Williams, 1991).”

“Evidence for an earlier, medium-pressure,

low-temperature, pre-tectonic or early syntectonic metamorphism comes from four areas within the subprovince. In the Atikokan region, and in northern Minnesota, both at the northern margin of the Quetico subprovince, an early *M1* metamorphic peak between *D1* and *D2* produced Ky–St–Bt assemblages (Ayres, 1978; Tabor et al., 1989). Kyanite inclusions in plagioclase within Grt–Sil–Bt–Pl–Qtz schist near Raith, north of Thunder Bay, have been reported by Percival et al. (1985). Kehlenbeck (1976) also presented textural evidence for a polymetamorphic history along the northern margin of the Quetico subprovince north of Thunder Bay. Again, along the northern margin of the subprovince, in the Beardmore–Geraldton area (Williams, 1989), amphibolite-facies conditions were attained prior to *D2* deformation, with the distribution of facies being structurally controlled within thrust-bounded panels (Williams, 1991).”

“In contrast, the main phase of regional metamorphism (*M2*), which produced the observed

map-pattern (Fig. 2), occurred late syntectonically (Sawyer, 1983; Williams, 1991). The general sequence of isograds, based on the appearance of diagnostic assemblages in pelites, is Chl–Ms–Bt, Grt+And+Sil, Grt+Crd+Sil, *in situ* granitic leucosome, and Opx (Pirie and Mackasey, 1978; Percival and Stern, 1984). The common occurrence of Grt–And in metapelites in the western Quetico subprovince is diagnostic of bathozone 2 (<3.4 kbar; Carmichael, 1978), whereas the presence of Sil–St in the eastern Quetico is diagnostic of bathozones 3 and 4 (3.4–5.5 kbar)."

"As noted by Williams (1991), tectonic thickening of the sedimentary pile and intrusion of minor I-type granitic rocks occurred prior to the thermal acme. Most of the large pre- to syntectonic granitic bodies are peraluminous and have sedimentary sources but display little evidence of thermal contact metamorphism; one exception is the South Beatty Lake pluton in the northern Quetico subprovince (Pirie & Mackasey, 1978). Steeply dipping thermal gradients, local increases in temperature around large plutons, and the general association of the highest-grade rocks with abundant generation of leucosome, indicate that the source of heat was a combination of burial, upward magmatic transport, and tectonism."

"In the northern Quetico, M1 metamorphism is estimated to have occurred between 2698 Ma, the maximum age of sedimentation (Davis et al., 1990) and 2688±4 Ma, the age of emplacement of the late syntectonic Blalock pluton (Davis et al., 1990). In the southern Quetico, M1 occurred after 2690 Ma, the maximum age of sedimentation (Zaleski et al., 1999). The timing of M2 metamorphism is less well constrained, and may have been protracted. In the Manitouwadge area, Zaleski et al. (1999) constrained regional D2 deformation to 2680–2677 Ma, and suggested that migmatization in both the northern Wawa and southern Quetico subprovinces occurred after 2679 Ma, broadly coincident with D3 deformation. This inference is consistent with observations elsewhere in the Quetico that contact aureoles around late plutons, dated at 2671±2 to 2665±2 Ma, as well as late granitic pegmatites dated at 2653±4 Ma, overprint regional metamorphic fabrics (Percival and Sullivan, 1988; Percival, 1989). North of Manitouwadge, Pan *et al.* (1998) reported a U–Pb zircon age of 2666±1 Ma from a granitic pegmatite concordant with respect to the D3 fabric, and suggested that the regional amphibolite-facies metamorphism occurred between 2671 and

2665 Ma, consistent with the ages cited above. The timing of peak granulite-facies metamorphism north of Manitouwadge appears to be some 15 Ma younger, on the basis of U–Pb zircon ages of 2650± and 2651±3 Ma from a mafic granulite and a tonalitic leucosome, respectively (Pan et al., 1998). Zaleski & van Breemen (1997) reported that titanite ages young with increasing metamorphic grade, ranging from ~2686 Ma in the southern Manitouwadge greenstone belt to ~2640 Ma in the southern Quetico, suggesting that the thermal effects of regional metamorphism may have lasted over ~30 million years, from 2677 to 2640 Ma, in the higher-grade parts of the Wawa and Quetico subprovinces. On the basis of their regional geological and geochronological studies, Zaleski et al. (1999) concluded that "M2 metamorphism occurred after the tectonic juxtaposition of the Quetico and Wawa subprovinces."

This field trip will cover the southern two-thirds of the Quetico Subprovince, from south of Beardmore to Nipigon (Fig. 1). As mentioned above, there is an asymmetric distribution in metamorphic grade, with a gradual progression from greenschist-facies, clastic metasedimentary rocks near the northern contact with the Wabigoon Subprovince; to lower amphibolite-facies schists and gneisses; through to upper amphibolite-facies migmatites and derived granitic rocks near the southern contact with the Wawa Subprovince near Nipigon. Thermal and pressure maximum occurs south of the center of the Quetico. The metamorphic character is of high-temperature/low-pressure (Abukuma-type) metamorphism, associated with the abundant intrusion of granitoid rocks and the regional distribution of migmatites derived from the metasedimentary rocks (Kamineni et al., 1988; Percival and McGrath, 1986; Percival, 1989; Williams, 1989). Peak metamorphic assemblages in the field trip area suggest conditions >650°C and 5 kbar, corresponding to bathozones 4 to 5 (14 to 15 km deep in the crust) (Easton, 2000; Percival, 1989).

The nomenclature of Mehnert (1968) has been used in describing migmatites in this field guide.

Overview of the Georgia Lake Pegmatite Field

The Georgia Lake pegmatite field, as described by Pye (1965) and Mulligan (1965) represents the largest concentration of rare element mineralization in the

Table 1. Estimated reserves for the Georgia Lake pegmatite field.

Deposit	Estimated reserves (tons)	Average grade (% Li ₂ O)
Aumacho No. 1 (Brink)	759,475	1.65
Aumacho No. 2		
Aumacho No. 3	96,000	1.50
Conway	1,830,000	0.96
Jean Lake No. 4 (Parole)	1,689,000	1.30
McVittie	261,000	1.03
Nama Creek North Zone	2,784,000	1.11
Nama Creek South Zone	1,508,332	0.96
Jackpot	2,000,000	1.09
Vegan No. 2 (Newkirk)	750,000	1.38
Total	11,677,807	1.14

Superior Province of Ontario (Breaks et al., 2003a). It comprises 38 rare element occurrences and 10 spodumene-bearing, pegmatite deposits (total resource of 11.7 million tons grading an average of 1.14% Li₂O; Pye, 1965; Table 1; Fig. 3) were discovered during an exploration rush in the 1950's (Pye, 1956).

As delineated by Breaks et al. (2003a), the field covers approximately 1200 km² (32 km by 37 km) and is hosted by upper greenschist- to lower amphibolite-facies, clastic metasedimentary country rocks. Four types of rare element pegmatites (using the classification of Cerný (1991a) have been identified locally; Breaks



Figure 3. Generalized regional geology (after Breaks et al., 2003a).

Table 2. Subdivisions of the Georgia Lake pegmatite field, after Breaks et al. (2003a)

Pegmatite group	Type	Examples
Postagoni River	Albite-spodumene	McVittie; Dunning; Nama Creek
Lake Jean	Albite-spodumene	Foster; Giles; Trans
Barbara Lake	Albite-spodumene	Vegan; Newkirk; Point; Jackpot
Forgan Lake	Albite-spodumene	Forgan; Lucky Lake (Lun-Echo)
Cosgrave Lake	Beryl±petalite sub-type	Swanson; MNW
Gathering Lake	Albite	<i>Newly discovered, unnamed</i>

et al., 2003a):

- 1) Beryl-type
- 2) Albite-spodumene-type
- 3) Complex-type/petalite-subtype
- 4) Albite-type

The Georgia Lake pegmatite field was initially subdivided by Milne (1962) and then Zayachivsky (1985) and Kissin and Zayachivsky (1985) into the Northern, Central and Southern groups. More recent data and mapping conducted by Breaks *et al.* (2003a) has led to a revised, six-fold subdivision (Table 2).

The two pegmatites that will be visited on the field trip, the Dunning pegmatite (Stop 2) and the Foster pegmatite (Stop 4), are examples of the albite-spodumene-type.

Zayachivsky (1985) divided the felsic plutonic rocks in the Georgia Lake area into three groups (Fig. 4):

- (1) Two-mica leucogranites, occurring as a large plutonic mass south of the pegmatite field (*cf.* Glacier Lake Batholith) and as smaller, satellite intrusions;
- (2) Kilgour Lake Group granitoids centered around a small gabbroic unit near Kilgour Lake; and
- (3) Tonalitic sills, distributed throughout the pegmatite field.

Zayachivsky (1985) suggested that the two-mica leucogranites and tonalitic sills were derived from the partial melting of metapelites and metawacke, respectively (*i.e.* S-type granitoids). The Kilgour Lake rocks were presumed to be the products of the fractional crystallization of a mafic magma generated in the upper mantle/lower crust (*i.e.* I-type granitoids).

Breaks et al. (2003a) described two distinct, fertile, peraluminous, S-type granite plutons (MNW and Barbara Lake stocks) that occur within metasedimentary country rocks. They consist of virtually identical, grey, medium-grained, generally homogeneous,

biotite-muscovite and muscovite-biotite granite and granodiorite.

The Glacier Lake Batholith (GLB) is a large mass of medium- to coarse-grained, S-type granite and pegmatitic granite situated immediately south of the northern metasedimentary part of the Quetico Subprovince, extending east of Highway 11 at least 50 km. It dominantly consists of grey, medium- to coarse-grained biotite and muscovite-biotite granite, which locally contains minor tonalitic and migmatitic metasedimentary enclaves. Generally massive, this rock may display a local, weak foliation developed in 1 to 2 cm-thick, shallowly lineated veins of fibrolite, muscovite and quartz that appear to have formed during aluminous hydrothermal alteration controlled by late shearing (Breaks et al., 2003a). Gradation into

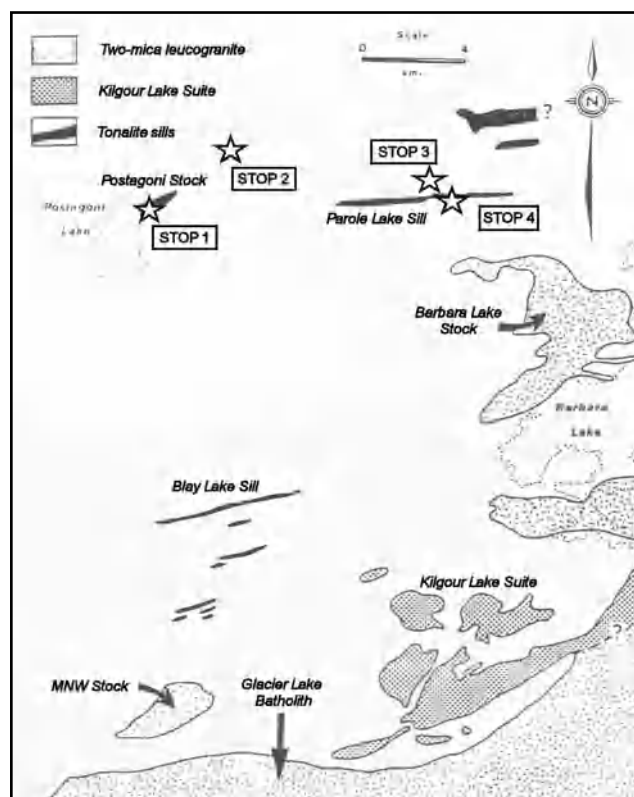


Figure 4. Plutonic rocks in the Georgia Lake area (after Pye, 1965; Zayachivsky, 1985; Breaks et al., 2003a).

consanguineous pods of tourmaline-muscovite potassic pegmatite up to 1 m in diameter occurs locally. There are several episodes of S-type granitic magmatism in the GLB. Undeformed, tourmaline-cordierite or tour-muscovite dikes are the youngest and most chemically evolved units (F. Breaks, Ontario Geological Survey, personal communication, 2004).

Pye (1965) undertook the first lithogeochemical survey for rare elements in the aforementioned felsic plutonic rocks in the Georgia Lake area. Many anomalies, ranging from 100 to 400 ppm Li, were noted.

Two modes of source fluid derivation may have operated during the emplacement of the Georgia Lake pegmatites (Zayachivsky, 1985). Pegmatites of the Central Group were derived by igneous fractionation of a granitic parent near Barbara Lake and emplaced as low-viscosity, volatile-rich melts in a direction away from the source area. This is supported by a fractionation trend defined most clearly by the concentration of Rb and Cs in perthitic microcline and muscovite across the Central Group. The pegmatites of the Northern Group may have been the result of direct anatexis of lithium-rich Quetico metasedimentary rocks. Breaks et al. (2003a) described the MNW pegmatite (petalite sub-type) as the most evolved of the local pegmatites, based on its geochemistry, mineralogy, including the occurrence of manganotantalite. A general rule for rare element-bearing Ontario pegmatites is that tantalum mineralization is associated with lithium mineralization in the form of spodumene and/or petalite (*ibid*).

Prediction of the source of lithium deposits in “fertile” parental granites is a somewhat uncertain art. The essential parental granite is a siliceous, peralkaline leucogranite, but the lithium content of a prospective parental granite is not highly indicative of the presence of lithium-rich pegmatites (Cerný, 1991a). Data summarized by Stewart (1978) and more recently by Cerný (1991a) indicate that Li-contents in parental granites are typically <100 ppm. Moreover, analyses of bulk compositions of spodumene- and petalite-bearing pegmatites yield a closely grouped mean of 1.53 weight % Li_2O (Stewart, 1978) and range between 1.5 and 2.0 weight % Li_2O (Cerný, 1991b). These data indicate that differences between spodumene- and petalite-bearing pegmatites are caused by pressure and temperature differences during crystallization rather than differences in bulk composition. Cerný (1991b) suggested that exploration for rare-element pegmatites

can be pursued by the identification of dispersion haloes of rare alkali metals (Li, Rb, Cs) in country rocks and some other elements such as Be, Sn and Cu. Hidden pegmatite targets in promising terrain may thus be located.

Uraniferous Dykes

Uranium-bearing pegmatite dykes are associated with felsic plutonic rocks in this part of the Quetico Subprovince. Several occurrences have been documented by Scott (1987), including some in the vicinity of Nipigon. White, albite-rich, biotite-quartz±muscovite±apatite pegmatite dykes up to 30 m wide intrude gneissic and migmatitic rocks. Yellow and yellow green stain (secondary uranophane ($\text{Ca}(\text{UO}_2)_2\text{Si}_2\text{O}_7 \cdot 6\text{H}_2\text{O}$)) characterize many of the uraniferous dykes. Uraninite appears to be the primary uranium-bearing mineral (Franklin, 1978). The Lake Helen occurrence is situated on the western shore of Lake Helen, opposite the roadside rest area on Highway 11 (Stop 7). Trenching and sampling of this occurrence in 1967 by Aggressive Mining Limited yielded values up to 0.135% U_3O_8 (*ibid*). Such dykes are considered to be the primary source of uranium that occurs in structures at or near the Archean / Proterozoic unconformity in the Nipigon Basin.

Stops

Field Trip Road Log

Stop	Locality	km
Lake Jean stops		
	Intersection of Hwy.'s 11 & 17 in the Town of Nipigon	0
	Take Hwy 11 north	37.4
	Gorge Creek turn-off - head east	
	<i>Reset</i>	
	Postagoni River Bridge	12.4
1	Postagoni Stock	14.4
2	Dunning pegmatite	18.4
	Powerline crossing	19.1
3	Quetico metasedimentary rocks	28
4	Foster pegmatite	28
Lake Helen stops		
	Intersection of Hwy.'s 11 & 17 in the Town of Nipigon	0
	Take Hwy 11 north	
5	Glacier Lake Batholith	17.1

6	Biotite leucogranite	10.8
7	Migmatite - roadside rest area	7.4
	Pull-off area	5.9
8	Pegmatitic granite	5.2
9	Pegmatitic granite	4.85
10	Pegmatites in migmatite	4.6
	Intersection of Hwy.'s 11 & 17 in the Town of Nipigon	0
	Take Hwy 17 east	
11	Migmatites (optional)	18.4

Stop 1 - Postagoni Stock

UTM coordinates - 0425041E 5472869N

Large, locally derived boulders of leucotonalite of the Postagoni Stock (*a.k.a.* Postagoni Lake sill) occur on the eastern side of the forest access road. Pye's (1965) mapping identified a porphyritic granitic intrusion approximately 1.6 km long and 700 m wide in metasedimentary country rocks. The McVittie pegmatite cuts the stock north of Dive Lake; it was visited during an ILSG field trip in 1990 (Kissin, 1990). Both the McVittie and Dunning (Stop 2) pegmatites are part of the Northern Group (Zayachivsky, 1985) or Postagoni Lake Group (Breaks et al., 2003a), comprising unzoned albite-spodumene-type pegmatites.

Zayachivsky (1985) described this sill as a coarse-grained tonalite with closely packed crystals of plagioclase up to 1.5 cm long, locally imparting a porphyritic texture. This texture occurs throughout the sill except near its contacts with the metasedimentary country rocks where a several metre-wide chilled margin has developed. The chilled margin texture resembles that of the porphyry dykes described by Pye (1965). A modal analysis of the Postagoni Stock was given by Zayachivsky (1985; Table 3).

Table 3. Modal analysis of the Postagoni Stock.

Mineral	Modal abundance(%)
Plagioclase	66.6
K-feldspar	5.6
Quartz	25.7
Muscovite	0.4
Biotite	1.6
Epidote	0.1
Zircon	Trace

The Postagoni Stock and the other local tonalitic sills are invariably in sharp contact with the country

rocks and display no foliation nor significant internal variation. An U-Pb zircon age of 2684.8 ± 2.1 Ma was determined for the Postagoni Stock (D. Davis, University of Toronto, unpublished data).

Stop 2 - Dunning Pegmatite

UTM coordinates - 0427823E 5474882N

This small roadside outcrop exposes the Dunning pegmatite dyke, which has an estimated width of 2 to 3 m at this location and intrudes Quetico metasedimentary rocks. Woolverton (1956) noted that pegmatite dykes on the Dunning property were exposed intermittently in a zone extending over 600 m in length.

This dyke is part of Zayachivsky's (1985) Northern Group, which are characterized by medium- to very coarse-grained, perthitic microcline and spodumene in a fine-grained matrix of quartz, muscovite and albite, with trace amounts of apatite and garnet. They are typically unzoned to poorly zoned and commonly contain aplitic stringers, bands and pods, within and parallel to dyke contacts.

This dyke is a typical albite-spodumene-type pegmatite (Fig. 5). Green spodumene crystals, up to 4.5 cm long, are aligned perpendicular to the dyke



Figure 5. Drusy spodumene, Dunning pegmatite (Stop 2).

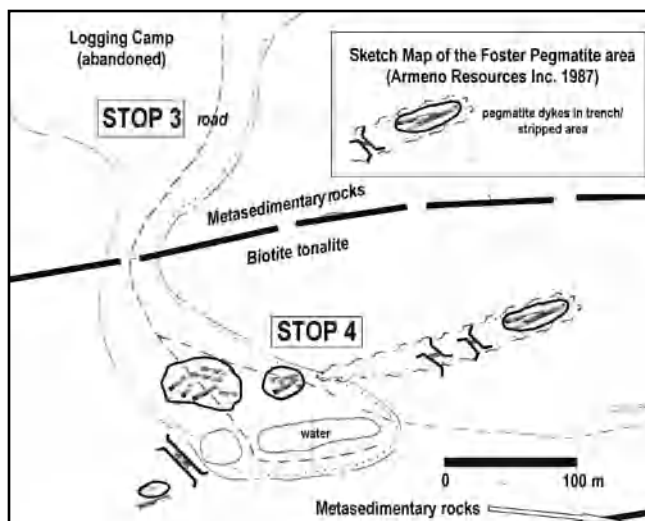


Figure 6. Sketch map of Foster pegmatite area (Stops 3 & 4).

contact. This phenocrystic spodumene typically contains several percent iron and is commonly altered to a dark green to black alteration product of chlorite and sericite (*aka* “rotten spodumene”). Some altered, brown spodumene is also present, along with quartz, albite ($\pm$ cleavelandite), beige alkali feldspar, green muscovite and minor oxides. Pye (1965) reported that the lithia contents of the two dykes on the Dunning property were low and of no economic importance.

Stop 3 - Quetico Metasedimentary Rocks

UTM coordinates - 0436345E 5473931N

Medium-grade clastic metasedimentary rocks typical of the north-central part of the Quetico Subprovince are exposed in a small outcrop north of the Foster pegmatite (Fig. 6). South of the Beardmore-Geraldton belt, more highly metamorphosed, turbiditic sediments



Figure 7. Graded bedding in metaturbidites, Foster area (Stop 3).

of the submarine margin and trench-fill are present in the Quetico (Fralick et al., 1992; Williams, 1989). 10 to 20 cm thick sandstone-siltstone beds strike 100° and likely dip steeply to the south, based on observations in the area by Pye (1965). Graded bedding suggest tops to the south (Fig. 7). Pye (1965) noted that biotite-quartz-feldspar schist such as this are most commonly granoblastic; garnet porphyroblasts have also been noted. Some lenticular pods of quartzo-feldspathic leucosome are present.

Stop 4 - Foster Pegmatite

UTM coordinates - 0436379E 5473665N

A series of stripped and washed outcrops reveals a swarm of narrow pegmatite dykes cutting an east-trending sill of massive, biotite tonalite, termed the Parole Lake sill by Zayachivsky (1985; Fig. 8). The Foster deposit was discovered in 1956 and was tested by diamond drilling later that year by Towagmac Exploration Co. Ltd. (Walter, 1957; Pye 1965; Table 4). This work indicated that the lithium deposit was too low-grade and the property was allowed to lapse.

Armeno Resources Inc. explored the property (including the neighbouring Lew deposit) from 1986 to 1989, carrying out stripping, geological, magnetometer and electromagnetic surveys, as well as diamond drilling (Assessment Files, Thunder Bay North District, Thunder Bay). The lone drill hole intersected biotite tonalite with numerous quartz-rich, pegmatitic dykes ≤ 1 m thick with up to 5% spodumene; no chemical analyses were reported (*ibid*).

The main Foster dyke strikes approximately 085° and dips 80° to 85° south. It is exposed continuously



Figure 8. Pegmatite dykes in tonalite, Foster (Stop 4).

Table 4. Results of drilling at the Foster deposit (Walter, 1957).

Drill Hole Number	Location (feet)	Core length (feet)	Lithia Content (%)
7	3+50 west	29.0	0.26
6	4+50 west	30.0	0.55
9	6+50 west	31.3	1.04
		20.0	0.58
10	7+50 west	32.7	0.58

over 75 m and averages 9 m in thickness. To the west, it splits into a number of thin, parallel dykes. In the most westerly drill hole (T-10), 7 individual dykes, 0.3 to 2.4 m wide were intersected, separated by sections of host tonalite a few centimetres to a metre wide (Pye, 1965). The deposit contains 10 to 15% spodumene, which is for the most part, unaltered. Similar to other deposits in the Lake Jean Group, crystals are prismatic and perpendicular to dyke contacts and may extend across the entire dyke (*ibid*). Milne (1962) listed the Foster pegmatite in the Jean Lake sub-group (with the Giles, Camp and Lew pegmatites) on the basis of well-developed spodumene orientation and the occurrence of narrow, branching and continuous, longitudinal aplite dykes.



Figure 9. Drusy spodumene, Foster pegmatite (Stop 4).

These are classic albite-spodumene-type pegmatites. Fresh green to dark green-black, altered spodumene crystals may reach lengths of 8 cm (Fig. 9). Other constituent minerals include K-feldspar, albite ($\pm$ cleavelandite) and quartz, minor apatite and garnet, and tiny, sparsely distributed, black tantalum-niobium-bearing oxide minerals (F. Breaks, OGS, personal communication, 2004). A locally developed border zone (≤ 5 cm) consists of quartz + feldspar.

Stop 5 - Glacier Lake Batholith Leucogranite

UTM coordinates - 0409911E 5445897N

Large road cuts on both sides of Highway 11 provide excellent exposures of the Glacier Lake Batholith (GLB). At this location, it consists of white, massive to locally foliated, muscovite < biotite, medium- to coarse-grained granite. Localized pods of tourmaline-biotite-muscovite-potassium feldspar pegmatite occur within the granite. These pods may reach 1 m in diameter and locally contain tourmaline-quartz intergrowths. Numerous, curvate, fibrolite - muscovite $\pm$ tourmaline veins up to 1 cm thick also occur in the host granite. Purple fluorite is exposed on a fractured outcrop face on the west side of the highway (Fig. 10).

Stop 6 - Glacier Lake Batholith Leucogranite/ Migmatite

UTM coordinates - 0407163E 5440408N

This is another example of the southern margin of the Glacier Lake Batholith where it is in contact with metasedimentary migmatite (Fig. 11). White, biotite-muscovite leucogranite contains rare, blue-green apatite. Foliation is developed in feldspathic segregations and biotitic seams. Leucogranite dykes and migmatite locally exhibit a *lit-par-lit* structure. Tight to isoclinal folds have developed in the quartz-biotite-feldspar schist (Fig. 12). All rocks display boudinage and folding. Folded leucosome suggests



Figure 10. Fluorite on fracture surface in sericite- and fibrolite-bearing Leucogranite (Stop 5).



Figure 11. Contact between folded metasedimentary migmatite and Leucogranite (Stop 6).

an early and protracted deformation history (Fig. 13). Metasedimentary migmatite enclaves are common in the white, S-type pegmatitic granites along the highway. Note that less evolved S-type granite may contain biotite as the only mica.

Narrow diabase dykes cut the country rocks on the western side of the highway.

Stop 7 - Migmatite

UTM coordinates - 0406986E 5437331N

Migmatites are exposed on the shore and islands of Lake Helen and on a large, glacially streamlined,

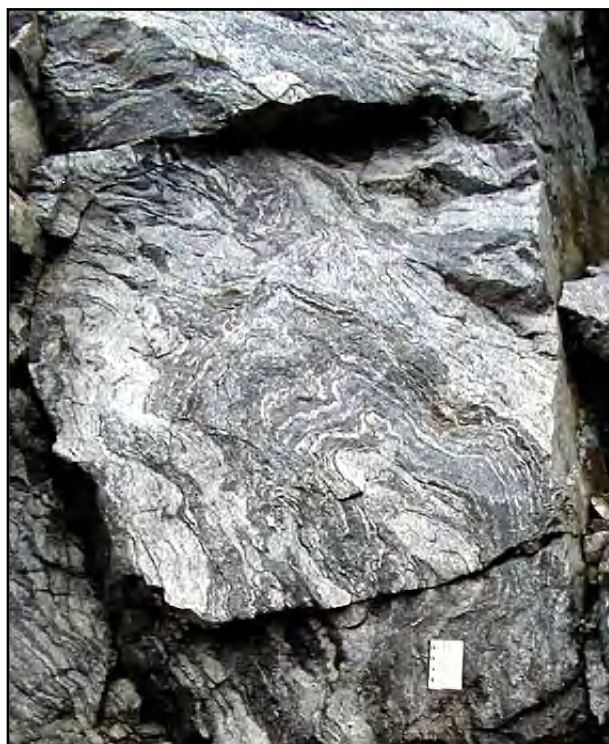


Figure 12. Folded metasedimentary migmatite (Stop 6).



Figure 13. Folded metasedimentary schist, incipient leucosome development (Stop 6).

whaleback outcrop behind the highway rest stop/pull-off area. There is quite a bit of variation in the relative amounts of leucosome and restite (Fig. 14). Boudinaged and pygmatically folded leucosome pods and veins occur in a quartzo-feldspathic with narrow, biotitic restite septa.

Stop 8 - Pegmatitic Granite

UTM coordinates - 0407272E 5435208N

This outcrop shows a transition in magmatism from S-type (*i.e.* Glacier Lake Batholith) to I-type granitoids. Pink, coarse-grained, biotite granite here consists of pink, coarse-grained, perthitic K-feldspar, quartz, coarse-grained biotite, and brown, altered plagioclase (up to 4 cm long). These weakly peraluminous, pegmatitic, biotite granites are relatively primitive and are younger than the white, two-mica leucogranites. Such rocks are probably of I-type origin and typically are metasedimentary enclave-free. Bulk rock levels of



Figure 14. Lit-par-lit migmatites with large metasedimentary inclusions (schollen) (Stop 7 area).



Figure 15. Pegmatitic K-spar - quartz – biotite granite (Stop 8).

rare-elements are very low: 128 ppm Rb, 2.63 ppm Cs, 1.8 ppm Nb, 0.39 ppm Ta (F. Breaks, OGS, personal communication, 2004).

A K-feldspar-megacrystic, pink biotite granite (Fig. 15) with a shallowly east-dipping foliation and metasedimentary xenoliths is intruded by a more massive granite at this locality. The intrusive contact is embayed and scalloped, suggestive of co-mingling magmatic textures (Fig. 16). The megacrystic, foliated granite may represent an earlier, xenolith-bearing phase that was intruded by the massive granite while still relatively warm and plastic.

Stop 9 - Pegmatitic Granite

UTM coordinates - 0407458E 5434906N

A large, bare, whaleback outcrop on the east side of the highway consists of coarse-grained to pegmatitic, massive, homogeneous pink granite (*i.e.* the younger granitic unit at Stop 8; Fig. 17). Crystals or interstitial patches of quartz, biotite and locally sericitized, K-feldspar average 2 to 3 cm in size (Fig. 18). Individual



Figure 16. Scalloped contact (arrow) between foliated biotite granite (top) and pegmatitic granite (bottom) (Stop 8).



Figure 17. “Whaleback” outcrop of coarse-grained to pegmatitic granite; note large (>60 cm) feldspar megacryst (circled) (Stop 9).

feldspar megacrysts may attain lengths of over 60 cm.

Stop 10 - Pegmatite Dykes in Migmatite

UTM coordinates - 0407587E 5434690N

Approximately 200 m south of the massive pink granite, white pegmatite dykes intrude fine-grained metasedimentary schist. This dark, fine-grained, feldspathic, biotite schist displays a well-developed foliation and minor folds, indicative of a long, protracted ductile deformation history, perhaps coeval with dyke emplacement. Note the sub-horizontal mineral lineation. Flattening of feldspars, producing small-scale, augen structures, is also indicative of high-temperature (>500° C) deformation.

The boudinaged and annealed, sericite- and biotite-bearing dykes range from a few centimetres (*lit-par-lit* structured) to several metres in thickness. They are mineralogically and texturally interesting, containing



Figure 18. Coarse-grained to pegmatitic granite (Stop 9).

both cordierite and quartz-tourmaline intergrowths with alkali (locally sericitized) feldspar. Garnet is conspicuous by its absence.

Breaks et al. (2003b) described potassic, biotite pegmatite that grades into a medium-grained, biotite granite near this stop (UTM: Easting 408339; Northing 5433016). The pegmatite contains coarse, euhedral K-spar; quartz; prismatic, medium- to coarse-grained, black tourmaline (schorl-dravite; Fig. 19); and fine-grained, green and blue fluor-apatite (0.3 to 0.9 weight % MnO). A biotite-rich, metasomatic contact occurs between the biotite granite and diabase. Graphic, coarse-grained cordierite (< 2 cm) and tourmaline occur in the granite near a diabase contact.



Figure 19. Quartz-tourmaline intergrowth in pegmatite dyke (Stop 10).

Stop 11 - Migmatites (Optional)

UTM coordinates - 5429614N 0425006E

This optional stop displays schollen (raft)-structured migmatites in which folded and schlieric, mafic, paleosome xenoliths have been highly deformed and disrupted (Fig. 20). There is a relatively high proportion of quartz-feldspathic neosome matrix to the xenoliths, suggesting high(er) degrees of partial melting. Patches and dykelets of coarse-grained, biotite-quartz-feldspar neosome contain garnet, cordierite and large (>10 cm), euhedral green apatite crystals.

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Dr. Fred Breaks (Ontario Geological Survey) is thanked for the provision of unpublished data on pegmatites and host rocks. Dr. Mary Louise Hill (Lakehead University) provided useful insight into the metamorphic and structural history of the high-grade

metamorphic rocks. John Scott (Ontario Geological Survey) generated maps for this report and provided information on local uranium mineralization.



Figure 20. Schollen structure in migmatite (Stop 11).

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Fieldtrip 5 - Geology of the Black Sturgeon Area

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Introduction

The Nipigon Embayment is an approximately 19,000 km² area of Mesoproterozoic igneous and sedimentary rocks centred on Lake Nipigon, north of Lake Superior, approximately 100 km northeast of Thunder Bay, Ontario (Fig. 1). The Embayment is underlain by Archean rocks of the Wabigoon Subprovince to the north, and the Quetico Subprovince to the south. This field trip will be concentrating on the Proterozoic and underlying Archean rocks of the southern portion of the Embayment.

The first geological work in the Lake Nipigon area was conducted by Bell (1870) for the Geological Survey of Canada. Portions of the area were examined by McInnes (1896), Parks (1901), Coleman (1909), and Wilson (1910). Geological mapping by Coates (1972) covered the area south of Lake Nipigon at a scale of 1:63,360. The adjacent area to the south was mapped

by McIlwaine and Tihor (1975a,b) at 1:15,840 scale. The area surrounding Lake Nipigon, and extending to the southern edge of Black Sturgeon Lake, was mapped at a scale of 1:50 000 by Sutcliffe (1982a,b; 1985a,b,c). Sutcliffe (1986) completed a doctoral thesis on the intrusive rocks of the Lake Nipigon area, including the Nipigon diabase sills and ultramafic intrusions of the Disraeli and Seagull-Leckie Lakes areas. Franklin et al. (1980), Cheadle (1986), and Rogala (2003) have carried out detailed investigations of the stratigraphy of the Sibley Group.

Bedrock mapping and research studies were undertaken between 2003 and 2005 as part of the Lake Nipigon Region Geoscience Initiative (LNRGI). The LNRGI is a geoscience-based geological data acquisition and compilation program managed by the Ontario Prospectors Association (OPA) and financially supported by the Northern Ontario Heritage Fund Corporation (NOHFC). The major goals of the initiative are to conduct research and collect new geoscience data that will help to explain the geological history of the area, and identify areas of high mineral potential that will attract exploration to the region. The initiative included bedrock mapping projects (MacDonald, 2004; MacDonald and Tremblay, 2005), airborne magnetic and radiometric geophysical surveys (Ontario Geological Survey, 2004a), ground gravity surveys (Ontario Geological Survey, 2004b), and targeted surficial geochemical and lineament studies (e.g., Barnett, 2004; Dyer, 2004), a paleomagnetic study (e.g., Ernst et al., 2005) and geochronological studies (e.g., Heaman et al., 2005). Mapping of the southern portion of the Nipigon Embayment was completed by the Ontario Geological Survey as part of its commitment of in-kind support to the LNRGI (Hart and Magyarosi, 2004; Hart, 2005a). Lakehead University, as a partner in the Initiative, contributed a detailed sedimentary basin reconstruction (e.g., Metsaranta and Fralick, 2004) and petrochemical characterization of the mafic and ultramafic intrusions (e.g., Heggie and Hollings, 2004; Richardson and Hollings, 2005).

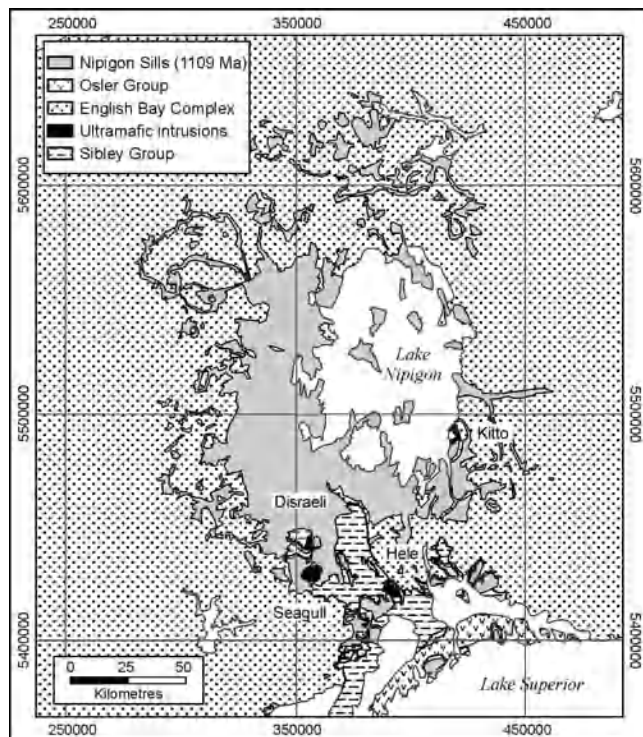


Figure 1. Mafic and ultramafic intrusions of the Nipigon Embayment.

Mineral exploration and prospecting in the area has been sporadic with the first recorded activity being the exploration for iron ore in the early 1900s (Coates, 1972). Exploration in the early 1900s also resulted in discovery of lead and zinc mineralization to the southwest, and limited production (Shklanka, 1969). The discovery of sporadic copper mineralization near Disraeli Lake in 1965 resulted in a surge in copper and the other base metals in the late 1960s (Coates, 1972). Exploration for uranium has been cyclical, with the greatest activity occurring in the middle to late 1970s coinciding with the discovery of large unconformity related deposits in other locations including the Wollaston Basin of Saskatchewan (Scott, 1987). A uranium occurrence was discovered by R.H. Sutcliffe during the 1981 field season and subsequently acquired and explored by Uranerz Exploration and Mining Corporation Limited until 1985 (Scott, 1987). A number of companies are actively exploring for uranium mineralization at the time this guide was being prepared. Several companies investigated the potential of sedimentary-hosted base metal mineralization in the Sibley Group in the early 1990s, but other than an airborne geophysical survey completed by Cominco in 1993, little of the work was

filed for assessment. Exploration for PGE mineralization intensified after the 1998 discovery of mineralization in the Seagull/Leckie Lake intrusion (Osmani and Rees, 1998). Initial results of diamond drilling indicated the presence of a basal pyroxenite hosting up to 1.71 ppm Pt and 1.87 ppm Pd over 2.1 m (Durham, 2000). Re-sampling completed in early 2004 defined two zones higher in the intrusion with up to 1.52 ppm Pt and 1.78 ppm Pd over 3.0 m and 1.56 ppm Pt and 1.87 ppm Pd over 3 m in the upper and lower zones respectively (East West Resource Corporation, 2004a). Drilling in late 2004 intersected a mineralized zone consisting of 2.38 g/t Pt, 2.65 g/t Pd, 0.19 g/t Au, 0.18 g/t Rh, 0.39 g/t Os, 0.34 g/t Ir and 0.08 g/t Ru (Pd:Pt ratio of 1.11) associated 0.19% Cu and 0.20% Ni over 3.21 metre (Platinum Group Metals Ltd., 2005).

General Geology

The southern portion of the Nipigon Embayment is underlain by Archean rocks of the Quetico subprovince that are unconformably overlain by Proterozoic sedimentary rocks of the Sibley Group (Fig. 2). Rocks of the Quetico Subprovince and the Sibley Group are

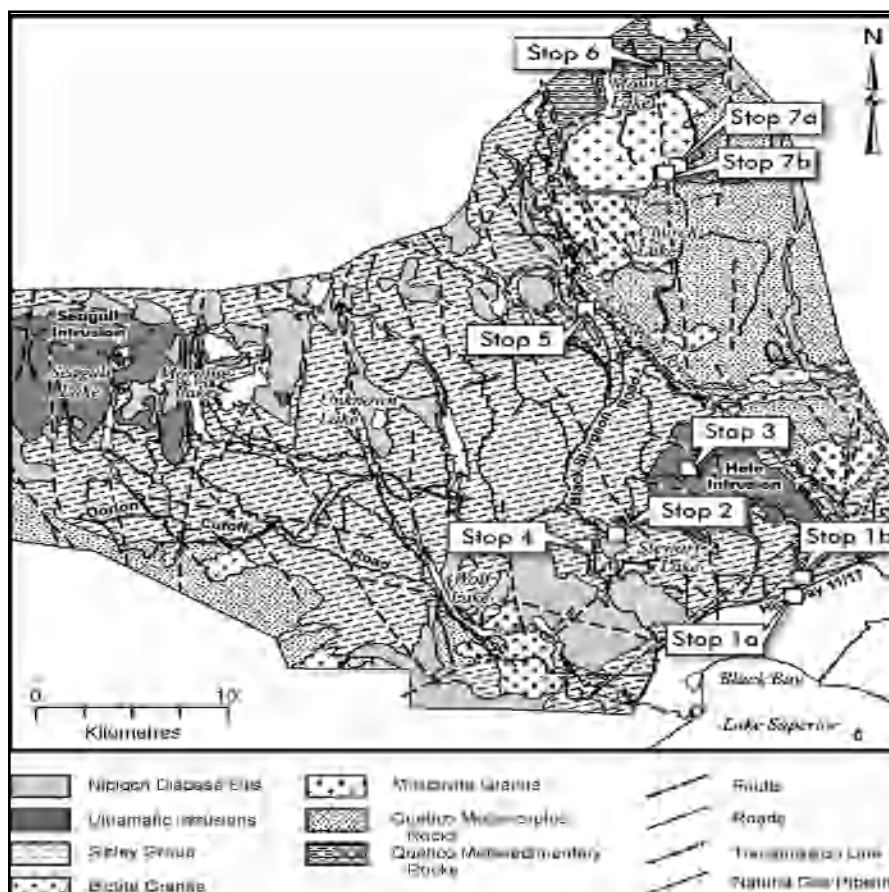


Figure 2. Geology of the southern portion of the Nipigon Embayment.

intruded by Proterozoic mafic to ultramafic intrusive rocks in the Disraeli and Seagull Lakes, and Hele and Kitto township areas (Fig. 2). Mafic sills of the Nipigon Diabase Sill Complex intrude all other rock types in the map area.

Rocks of the Quetico Subprovince consist of biotite and/or andalusite schists that are gradually replaced towards the south by amphibolites (Hart, 2005a). The rocks to the north have a schistosity oriented at approximately 050° to 060°, with near vertical dips. In the south, leucogranites intrude the amphibolite or biotite gneisses and form complex mixtures that have been classified as migmatite. The airborne magnetic survey indicates a regional fabric with about an 080° to 090° orientation in the migmatitic rocks (Ontario Geological Survey, 2004). A number of areas east of the Black Sturgeon River contain a high percentage of coarse grained to pegmatitic irregular bodies and dykes of the muscovite granite suite. Irregular bodies of medium to coarse-grained rocks of the biotite granite suite intrude the amphibolites and leucocratic phases form the leucosome of the migmatitic rocks.

Widely spaced, weakly deformed and metamorphosed, north trending diabase dikes intrude the Archean rocks of the Quetico Subprovince. A paleomagnetic study of similar dykes located to the west, in the area of Highway 527, by Ernst et al. (2005) has interpreted these dykes to be part of the early Proterozoic 2121 – 2101 Ma Marathon dyke swarm.

The Proterozoic Sibley Group sedimentary rocks unconformably overlie the Archean rocks of the Quetico Subprovince, and are relatively flat-lying, with dips of less than 5° and strikes that are usually difficult to determine. Rocks within this area have been subdivided into three relatively flat-lying formations, the lower Pass Lake, the middle Rossport, and the upper Kama Hill (e.g., Cheadle, 1986; Fig. 3). The Pass Lake Formation in this area is best observed in drill core, and consists of conglomerates, quartz arenites, and minor dolomites, with rare outcrops of the upper sandstone members. A sandstone of the Pass Lake Formation has a youngest detrital zircon grain with a U-Pb age of 1670 Ma (Heaman et al., 2005). The Rossport Formation consists of mudstone, dolomite and siltstones of the lower Channel Island Member, limestone and stromatolitic limestone of the Middlebrun Bay Member, and predominantly muddy variably carbonatized siltstones and mudstones of the upper Fire Hill Member (see Fig. 3). The Kama Hill

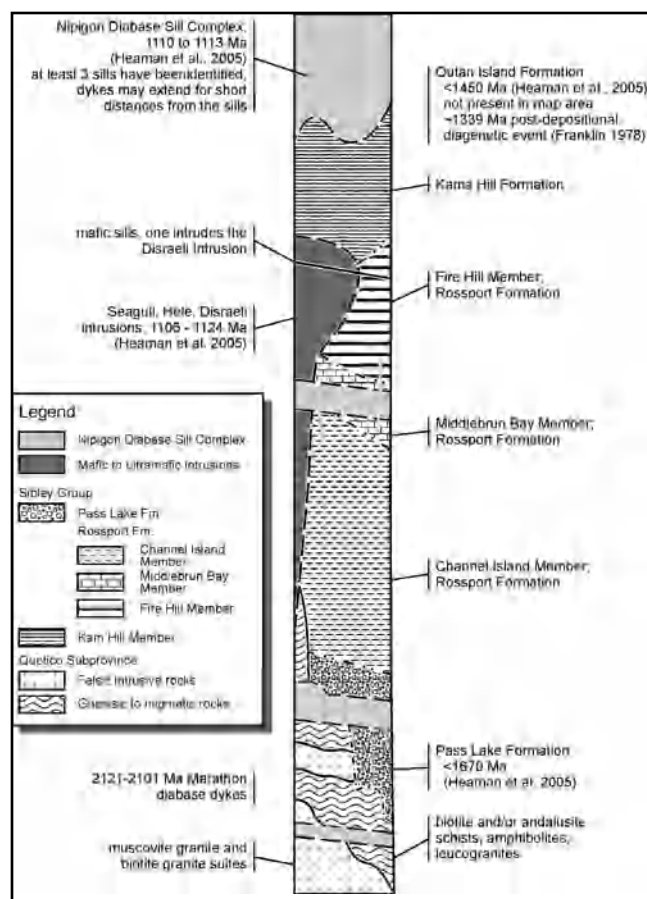


Figure 3. Generalized stratigraphic column for the southern portion of the Nipigon Embayment.

Formation is the uppermost formation of the Sibley Group exposed in the map area and consists of laminated mudstones and siltstones with sandstone prominent in the upper section. A sandstone of the Outan Island Formation, which overlies the Kama Hill Formation has a youngest detrital zircon with a U-Pb age of 1450 Ma (Heaman et al., 2005). A sandstone of the Nipigon Bay Formation, which overlies the Outan Island Formation, has a youngest detrital zircon grain with a U-Pb age of 1657 Ma (Heaman et al., 2005). A Rb/Sr age of 1339 +/- 33 Ma based on analyses of a combination of Rossport and Kama Hill formation samples by Franklin (1978), probably denotes a post-depositional diagenetic event. Cheadle (1986) estimated up to 200 m of topographic relief on the Archean paleosurface at the time of deposition of the Sibley Group. Although the bedding is generally flat-lying on an outcrop scale, these extremes in topographic relief probably controlled sedimentation processes on a local scale complicating stratigraphic correlations (P. Fralick and B. Rogala, Lakehead University, personal communication, 2003). Regionally, the Sibley Group thickens to the south under Lake Superior with two additional formations

overlying the Kama Hill formation (Rogala, 2003), and thins to the north and west (e.g., Cheadle, 1986).

Mafic to ultramafic intrusions occur in the Disraeli and Seagull-Leckie Lakes, Hele and Kitto township areas (Fig. 1), and all four are composed of a pyroxene peridotite core with an irregular olivine gabbro zone along the margin (e.g., Hart et al., 2002b; Hart and Magyarosi, 2004). The Seagull Intrusion has been studied in the greatest detail, and petrography indicates that it consists predominately of cumulate textured dunites, lherzolites, and olivine websterite, with minor olivine-hornblende pyroxenite (Heggie and Hollings, 2004). An olivine gabbro forms an irregular border zone to the ultramafic rocks, and is the unit most commonly exposed in outcrop. All of the intrusions have a monzogabbro, or granophyre, unit occurring as irregular pods or bands in the olivine gabbro, but are distinguished from it by the presence of abundant pink feldspar and minor amphibole. An olivine gabbro from the Hele Intrusion has a U-Pb baddeleyite age of 1106 ± 1.5 Ma with one fraction yielding a 1119.4 Ma age (L. Heaman, University of Alberta, personal communication, 2004). Samples from the Seagull Intrusion have U-Pb ages from zircon and baddeleyite of between 1105 and 1124 Ma (L. Heaman, personal communication, 2004). Samples from the Disraeli Intrusion have U-Pb ages from baddeleyite and zircon of 1111 Ma (L. Heaman, personal communication, 2004). Samples of an olivine gabbro from the Kitto Intrusion has a U-Pb baddeleyite age of 1117 Ma, but has baddeleyite fractions with a range of ages comparable to the other intrusions (L. Heaman, personal communication, 2004). Diabase sills have chill contacts with olivine gabbro of the Seagull, Hele and Kitto intrusions.

A group of fine grained, massive mafic sills with a higher olivine content may intrude the Sibley Group sedimentary rocks. Some of these sills are spatially associated with the ultramafic intrusions, as in the Moraine Lake and Hele Township areas, and probably represent extensions of the intrusions (Hart, 2005). In the Shillabeer Lake and Kama Hill areas, similar sills are not associated with known ultramafic intrusions (e.g., Richardson and Hollings, 2005; Hart, 2004) and may be comparable to the Jackfish Sills described by MacDonald (2004) in the English Bay area. One of these sills intrudes the Disraeli Intrusion suggesting that some examples of this group post-date the larger ultramafic intrusions (Hart, 2005). This group of sills

may be difficult to distinguish from the diabase sills without petrography and lithogeochemistry.

A series of shallow-dipping Proterozoic diabase sills intrude all other rock units in the map area. The sills display a range of U-Pb zircon ages from 1110 to 1113 Ma from sills located west of Lake Nipigon (Heaman et al., 2005). The diabase sills have variable internal subdivisions, a generalized section from top to bottom would contain: 1) chill, 2) fine-grained variably amygduloidal, 3) magnetite-rich, 4) medium to coarse-grained, 5) coarse to very coarse-grained, 6) fine-grained, variable amygduloidal, and 7) chill (Hart and Magyarosi, 2004). These subdivisions resemble those proposed by Sutcliffe (1986). Massive, medium- to coarse-grained feldspar and pyroxene diabase is the most common rock type and generally lacks a sub-ophitic or diabasic texture. Although these rocks should properly be classified as gabbro the diabase classification has been applied to all rocks associated with the sills to avoid confusion with other intrusions in the area. The sills range in thickness from less than 5 m to greater than 180 m and there appears to be at least two major sills greater than 100 m in thickness. However, erosion and block faulting hinders the correlation of the sills and thus the determination of the original number of sills.

Quaternary ice-marginal deltaic deposits of the Nipigon Moraine form an extensive cover along the Black Sturgeon River and the area immediately to the west of the river (Barnett, 2004). The rest of the map area further from the river is covered by differing thickness of till that is reported by Barnett (2004) to be mainly locally derived.

Regional metamorphism of the Archean rocks of the Quetico Subprovince varies from upper greenschist to lower amphibolite facies, with limited areas of migmatite. Hornfels metamorphism, in zones up to 10 m wide, is variably developed in the Sibley Group in contact with the peridotites, and occasionally the diabase sills. Hematite and specular hematite is associated with some late north-trending fracturing in both Archean and Proterozoic rocks, and appears to be a useful indicator of the north-trending fault systems in this area, where most structures are hidden by talus and glacial deposits (Hart and Magyarosi, 2004).

Metasedimentary rocks of the Quetico Subprovince have experienced multiple phases of deformation. Schistosity is well developed along the bedding planes

obscuring primary structures but the rocks appear to be the metamorphosed equivalents of the greywackes and siltstones present in the Beardmore-Geraldton greenstone belt (Hart et al., 2002a). North- and northwest-trending regional fault trends have displaced the Proterozoic Sibley Group, and a northeast-trend may only represent minor late movement along previous Archean structures (Hart and Magyarosi, 2004; MacDonald, 2004). The north and northwest trending faults, define the Black Sturgeon Fault Zone, which has been interpreted to be *en échelon* resulting in the formation of a block faulted asymmetric basin or graben (e.g., Coates, 1972). There appears to have been little lateral displacement based on the correlation of the iron formations across Black Sturgeon Lake in the area north of the current map area (Hart and Magyarosi, 2004). The north trending faults are interpreted to be regionally continuous structures that may correlate with north-trending structures in the Gull Bay area, approximately 150 km to the north. The northwest-trending faults display vertical movement with no apparent horizontal displacement, and may be traced to the west into the adjacent Wabigoon Subprovince. Accurate determination of displacements on these structures is not possible as they have been obscured by talus slopes and glacial deposits.

The timing of the different faults is not known, and it is probable that they have been reactivated. There have been several proposals made to account for the structural history of the area, with most focussing on the Sibley Group sedimentary rocks. Coates (1972) proposed a half graben with the Black Sturgeon River as the eastern boundary that also served as the main control on the deposition of the Sibley Group. Franklin et al. (1980) proposed that a failed arm of the Mesoproterozoic Keweenawan Midcontinent Rift created a basin, which included the half graben. Sutcliffe (1986) further developed the failed arm model as a major control on the emplacement of the peridotite intrusions and the diabase sills, and highlighted the existence of a subparallel fault system along the Nipigon River to the east. Alternatively, Fralick and Kissin (1995) and Hollings et al. (2004) proposed that the thermal upwelling that formed the English Bay Complex was followed by relaxation and formation of an intracratonic basin in which the Sibley Group was deposited.

The failed arm model predicts the presence of a dike swarm (e.g., Ernst and Buchan, 2003), in this case,

north-trending and centred on Lake Nipigon. The paleomagnetic and geochemical study of the north-trending dykes located to the west of the Nipigon Embayment by Ernst et al. (2005) suggests that they are part of the 2101 to 2121 Ma Marathon dyke swarm, not Keweenawan. If the identification of these dykes as part of the Marathon swarm is indeed correct, then combined with the lack of a north trending dyke swarm north or east of the Nipigon Embayment, a key element of the failed arm model appears to be absent.

The presence of Pass Lake Formation rocks at the base of a number of diamond-drill holes in the map area suggests that the half graben was present before, or was formed during, the deposition of the Sibley Group, sometime between 1670 and 1339 Ma (Heaman et al., 2005; Franklin, 1978). The occurrence of Pass Lake Formation at the base of Moseau Mountain, on the east side of the Black Sturgeon River, suggests that fault activity continued throughout the entire period of sedimentation, or that there has been significant post-sedimentation fault re-activation. At this time, there is no definitive means of distinguishing between these models although, the location of Pass Lake Formation rocks at higher elevations to the east could mean that sedimentation originally extended across much of the Nipigon Embayment, consistent with the first model.

Timing and magnitude of displacement along faults related to the emplacement of either the ultramafic intrusions or diabase sills is also difficult to determine. The high degree of fracturing and hematite alteration in some of the diamond drill core from the 1106-1124 Ma Seagull Intrusion (Heaman et al., 2005) suggests post emplacement fault activity. Extensive hematite and the occurrence of uranium mineralization in north trending structures has been dated at 1090 ± 20 Ma (Ruzicka and LeCheminant, 1984). Current interpretations suggest that faulting has resulted in displacement of igneous horizons within the Seagull intrusion, but it is just as possible that the intrusion was emplaced into a pre-existing structure to produce its current form. Pre-existing fault structures probably, in part, controlled the emplacement of the 1111-1113 Ma Nipigon sills (Heaman et al., 2005), as has occurred in other mafic sill complexes (e.g., Leaman, 1975). However, due to the lack of a distinctive marker in the area, it is difficult to determine if the apparently disjointed nature of many of the sills is the result of the sills ramping upward along pre-existing structures and subsequently being eroded away, or if it is a product of late post-

emplacement fault displacement.

Geochemistry

Mafic to Ultramafic Intrusive Rocks

Hele Intrusion

The Hele Intrusion may be subdivided into olivine gabbro, olivine melagabbro, and peridotite based on their MgO contents, with the peridotite having between 19 - 29 wt.% MgO and the olivine gabbro having MgO contents of less than 11 wt.% MgO. Elemental variations in diamond drill core samples appear to indicate that the Hele Intrusion is composed of more than one pulse of magma, resulting in an interlayering of the peridotite and melagabbro. Pink feldspar may occur as a minor mineral phase in the olivine gabbro, and is a major constituent of the monzogabbro, or granophyre. The presence of pink feldspar usually correlates with higher K₂O content that may be up to 7.42 wt.%, with no appreciable change in the Fe₂O₃ contents compared to the other olivine gabbro. Although the high K₂O is possibly related to late stage igneous fractionation, it is considered more likely that these values are a result of assimilation of Sibley Group rocks and that the pink colour is a result of potassic alteration rather than hematization.

Total rare earth element (REE) and high field strength (HFSE) concentrations in the peridotite

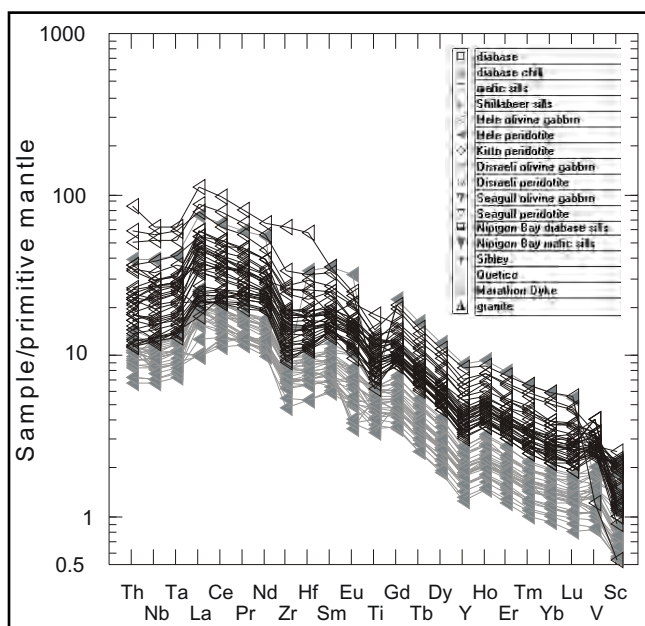


Figure 4. Mantle-normalized extended element diagrams of the olivine gabbro, melagabbro and peridotite of the Hele Intrusion (normalizing factors from Sun and McDonough, 1989).

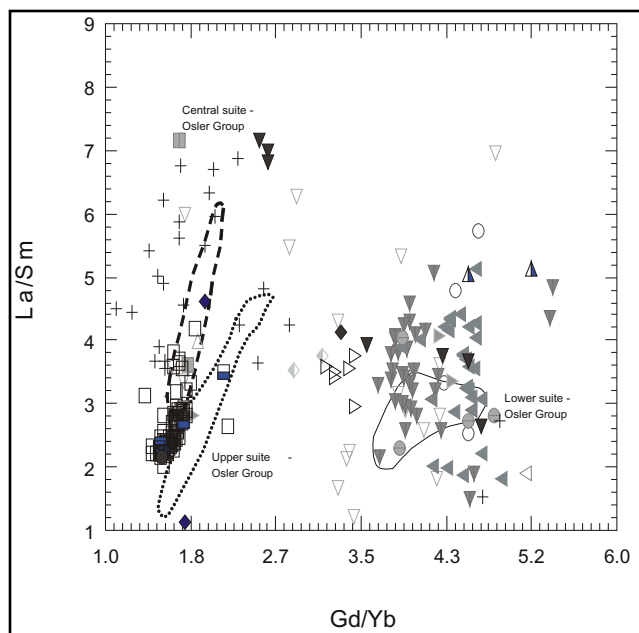


Figure 5. Gd/Yb versus La/Sm diagram of the mafic to ultramafic intrusive rocks and the Nipigon diabase sill complex with fields for the volcanic rocks of the Osler Group (after Lightfoot et al., 1991). Legend for symbols, see Figure 4.

samples are generally lower than in the olivine gabbro or olivine melagabbro (Fig. 4). All of the rocks have weak negative TiO₂ and Zr anomalies on the mantle normalized extended element diagram (*see* Fig. 4), and these anomalies are more pronounced than negative Nb-Ta anomalies for the Seagull and Disraeli intrusions (Hart and Magyarosi, 2004) although all three intrusion have flattening of the trends between Th and La. All three units have [La/Yb]_{mn} (mantle normalized) and Gd/Yb ratios which are distinctly higher than the diabase sills, and plot toward the high end of the field defined by the Lower Suite of the Osler Group volcanic rocks (Hart, 2005; Fig. 5). Rocks of the Hele Intrusion have a wider range of Gd/Yb ratios than the rocks from the Disraeli and Seagull intrusions, but overlap with the values for the other intrusions. The peridotite and olivine gabbro have REE and HFSE ratios that are comparable to ocean island basalts and most of the rocks from the Disraeli and Seagull intrusions in contrast to the diabase sills (Fig. 6).

The peridotite and gabbro are not enriched in nickel, and the peridotites follow an olivine-controlled igneous fractionation trend, which would be consistent with a parental MgO concentration of approximately 13 wt.% (Fig. 7). Both units have a flat positive slope on a primitive mantle normalized PGE diagram with enrichment in Ir and Pt (Fig. 8), compared to the trends for the diabase sills that follow a trend similar to that of

most continental flood basalts (Hart et al., 2002a).

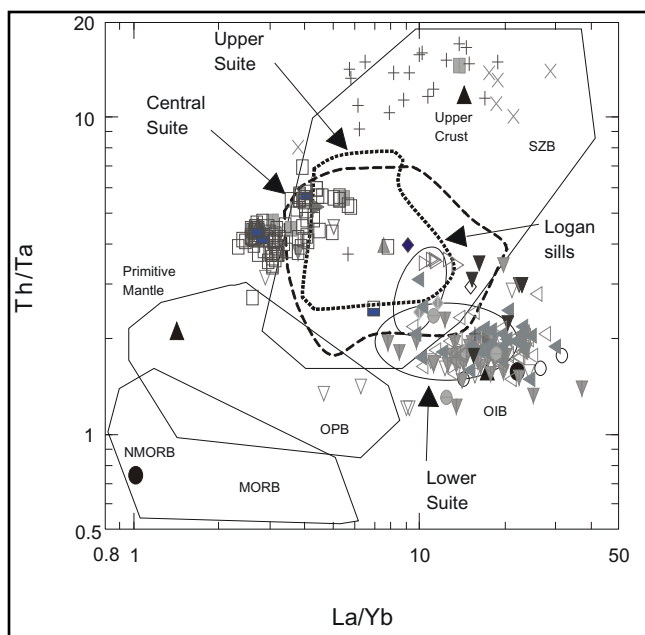


Figure 6. La/Yb versus Th/Ta diagram of the mafic to ultramafic intrusive rocks and the Nipigon diabase sill complex with fields for the volcanic rocks of the Osler Group and the Logan diabase sills south of Thunder Bay (after Condie et al., 2002). Legend for symbols, see Figure 4.

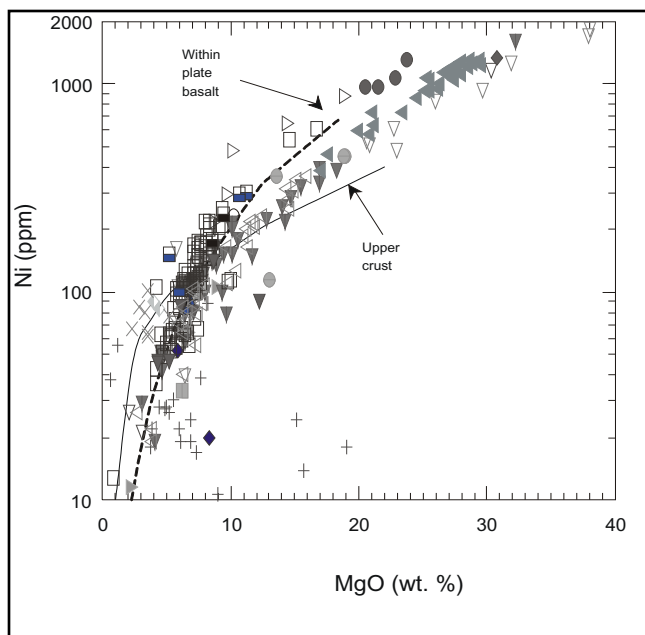


Figure 7. MgO versus nickel diagram of the mafic to ultramafic intrusive rocks and the Nipigon sill complex (after Lightfoot et al., 2001). Legend for symbols, see Figure 4.

Nipigon Diabase Sill Complex

The diabase sills plot along a line indicative of fractionation controlled mainly by plagioclase and clinopyroxene, as suggested by Sutcliffe (1987). The diabase sills display a range in REE and HFSE concentrations that will require further investigation

(Fig. 9), but the $[La/Yb]_{mn}$ are distinct from those of the rocks of the mafic to ultramafic intrusions. Most of the diabase samples have weak negative TiO_2 anomalies, but no apparent Zr anomalies, on the mantle normalized extended element diagram (see Fig. 9), and pronounced negative Nb-Ta anomalies. Some HFSE and REE ratios for the diabase show some similarities to the Central and Upper Suite volcanic rocks of the Osler Group as has been suggested by Sutcliffe (1987) (see Fig. 5).

The Nipigon diabase sills are not enriched in nickel, and follow an igneous fractionation trend (see Fig. 7). PGE contents of the diabase are similar to the trend followed by most continental flood basalts, and comparable to the PGE contents and trends noted for the diabase sills from the Beardmore area by Hart et al. (2002a; Fig. 10).

Mapping and geochemical sampling by Sutcliffe (1986) and Hart et al. (2002a) suggested that the sills do not represent single cooling units. However, the results of detailed sampling of a number of continuous sill sections from diamond drill core suggest that some sills may be single cooling units. Whereas others have chemical variations suggesting a more complex history of emplacement that may include multiple pulses of magma that did not re-equilibrate before cooling. However, the variations are subtle and generally not distinctive enough to be evident in a group of outcrop samples. The diabase sills have HFSE and REE ratios with some similarities to subduction zone basalts and the Central and Upper Suites of the Osler Group (see Fig. 6).

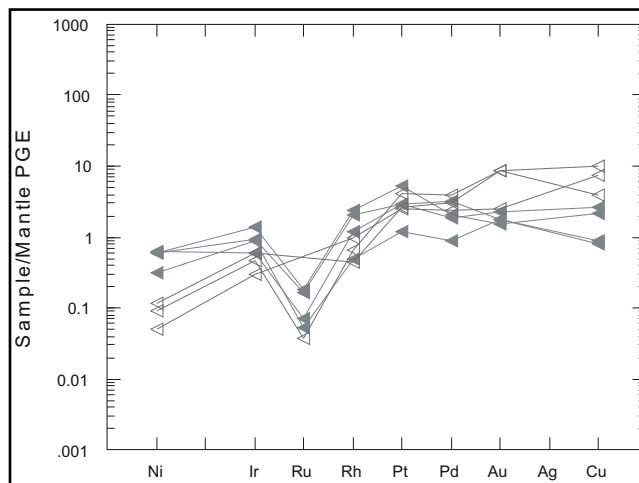


Figure 8. Primitive mantle-normalized platinum group element diagrams of the olivine gabbro, melagabbro and peridotite of the Hele Intrusion (normalizing factors from Barnes et al., 1987). Legend for symbols, see Figure 4.

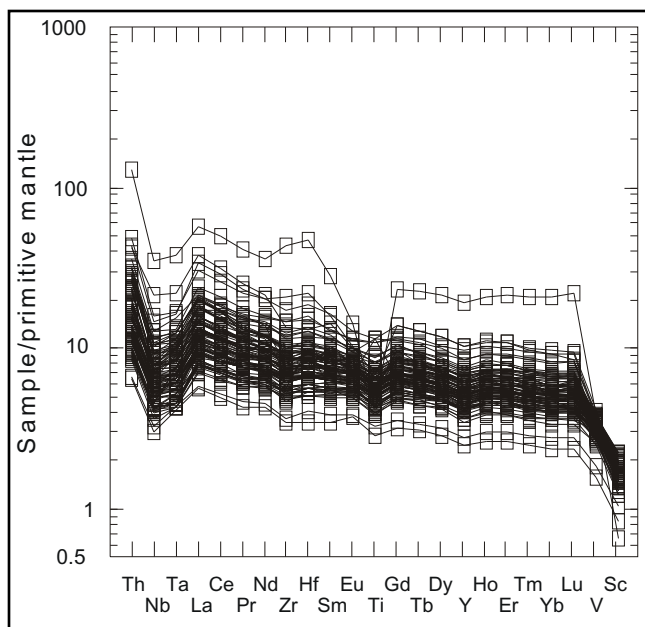


Figure 9. Mantle-normalized extended element diagrams of the Nipigon Diabase Sill Complex (normalizing factors from Sun and McDonough, 1989). Legend for symbols, see Figure 4.

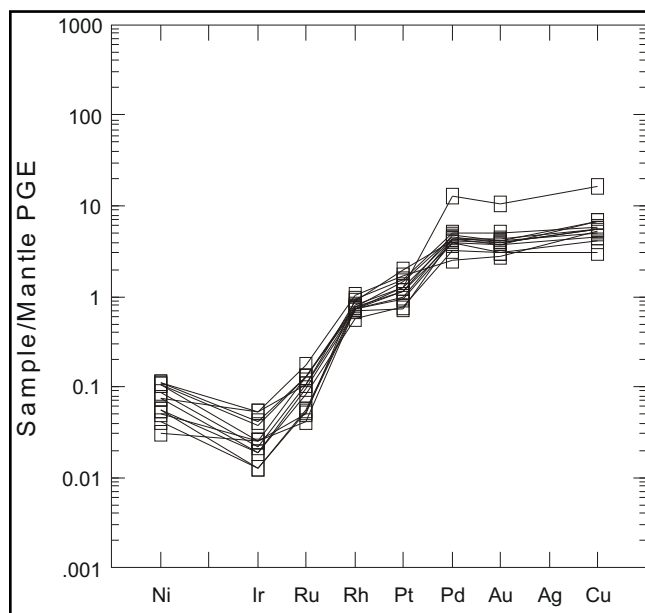


Figure 10. Primitive mantle-normalized platinum group element diagrams of Nipigon Diabase Sill Complex (normalizing factors from Barnes et al., 1987). Legend for symbols, see Figure 4.

The Nipigon diabase sills have lower TiO_2 , Zr/Y, La/Yb values than the Logan diabase sills located to the south of Thunder Bay (see Fig. 6) and have a regional distribution that extends south to at least the north side of the City of Thunder Bay (Hart, 2004). Several samples collected from the area south of Thunder Bay have geochemical characteristics comparable to the Nipigon diabase sills (see Fig. 6), but further investigation is required. This regional geographic distribution of high

and low TiO_2 rocks is similar to regional distributions observed in some flood basalt provinces suggesting that the Logan and Nipigon sills may be part of a single intrusive event.

The presence of variants of the sills that have specific geochemical and/or geographic distribution has resulted in the introduction of a variety of new, informal, names for the diabase sills in the past few years, such as the distinction between the Logan sills in the Thunder Bay area and the Nipigon sills around Lake Nipigon. In addition, new sill types, such as the Inspiration sills have also been identified (e.g., MacDonald, 2004; Richardson and Hollings, 2005). This does cause a problem, however, in that the historical term Logan Sills has been used exclusively in the paleomagnetic literature, and in much of the international literature, for all of the diabase sills north of Lake Superior. Until these terminology issues are resolved, it is recommended that the historical nomenclature of Logan Sills should be applied to all the diabase sills in the area north of Lake Superior, with subdivision into the informal terms, Logan sills in the south and Nipigon sills to the north. The Nipigon sills may be further subdivided into several chemical variants, such as the normal Nipigon, Inspiration, and Jackfish-like sills. Clarification of this nomenclature is contingent on further geochemical, petrological and geochronological work.

Mineralization

Platinum Group Elements

Hele Intrusion

Mineralization has not been located in either of the two diamond drill holes or in outcrop of the Hele Intrusion. The peridotite and olivine gabbro have similar PGE contents with Pt/Pd ratios of 1.03 to 4.63 with the highest values being 75.87 ppb Pt, 50.24 ppb Pd and 24.07 ppb Au in an olivine gabbro (Hart, 2005). The peridotite and olivine gabbro also have similar base metal contents with 18 to 1320 ppm Ni, 17 to 459 ppm Cu, and 72 to 280 ppm Zn.

Seagull Intrusion

Mineralization in the southern portion of the Nipigon Embayment occurs as three reef style zones hosted by the Seagull Intrusion. The upper zone contains up to 1.52 ppm Pt, 1.78 ppm Pd, 0.34 ppm Os, 0.11 ppm Rh, 0.23 ppm Ir, 0.15 ppm Au, 0.58% Cu, and

0.34% Ni over 3 m (East West Resource Corporation 2004a,b). The lower zone contains up to 1.56 ppm Pt, 1.87 ppm Pd, 0.20 ppm Os, 0.10 ppm Rh, 0.18 ppm Ir, 0.15 ppm Au, 0.55% Cu, and 0.35% Ni over 3 m (East West Resource Corporation 2004a,b). The best intersection in the basal zone, close to the intrusion Quetico contact, is 1.71 ppm Pt and 1.87 ppm Pd over 2.1 m associated with 10% disseminated to net textured sulphide minerals including pyrite, pyrrhotite and minor chalcopyrite (Durham, 2000). Diamond-drill holes completed in 2004 intersected a zone near the base of the intrusion consisting of 2.38 g/t Pt, 2.65 g/t Pd, 0.19 g/t Au, 0.18 g/t Rh, 0.39 g/t Os, 0.34 g/t Ir, 0.08 g/t Ru, 0.19% Cu and 0.20% Ni over 3.21 metre (Platinum Group Metals Ltd., 2005).

Mineralization in the basal zone is present near the lower contact of the intrusion, and is interpreted to be caused by sulphur saturation of the magma during initial stages of emplacement (Heggie and Hollings, 2004). The upper and lower zones are associated with more primitive whole rock geochemistry and mineralogy possibly reflecting an influx of less evolved magma. The mineralogy has been characterized by Heggie and Hollings (2004) and the following details are from that source. Nickel and copper sulphide minerals occur as disseminated bravoite and chalcopyrite respectively, found interstitial to olivine. Copper is also present as veinlets of native copper cross cutting sulphide minerals. Platinum group minerals are located at the grain boundaries of NiFe sulphide minerals, and are dominated by sperrylite, keithconnite, stibiopalladinite, and a copper palladium alloy.

Stops

Field Trip Road Log

Stop	Locality	km
	Junction of Highway 585 and Highway 11/17 in the Town of Nipigon	0
	Red Rock	5.5
	Black Sturgeon River	16
1a	Rock quarry north of Big Squaw Creek Road cut	19
1b	Big Squaw Creek Road Cut	19.5
	Junction of Highway 11/17 and the Stewart Lake Road	26.5
	<i>Reset</i>	
	Junction of the Stewart Lake Road and Fowlkes Lakes roads	5.9

Stop	Locality	km
	<i>Reset</i>	
2	Hele Intrusion peridotite	2.9
	Junction of the Fowlkes Lakes and Driftstone Lake roads	6.3
	<i>Reset</i>	
3	Hele Intrusion olivine gabbro	3.9
	Return to Stewart Lake Road	
	<i>Reset</i>	
	Junction of Stewart Lake and Black Sturgeon roads	3.8
	<i>Reset</i>	
4	Black Sturgeon Road	3.3
5	Eagle Mountain	20.3
	Junction of Black Sturgeon and Camp 42 roads	22.5
	<i>Reset</i>	
	Junction of Camp 42 and Church Road	14.1
	<i>Reset</i>	
6	Andalusite schist	7
	<i>Reset</i>	
	Return to south to junction of Camp 42 and Church Road	7
	<i>Reset</i>	
	Road to left	0.75
7a	Granite-dabase contact. South on trail	0.2
7b	Pegmatite	12

Trip starts at the junction of Highway 585 and 11/17 in Nipigon. Proceed southwest on Highway 11/17.

5.5 km - Rossport Formation

Oxidized units of the Rossport Formation of the Sibley Group are exposed in the hillside on the east side of the highway. These rocks unconformably overlie Archean rocks of the Quetico Subprovince, and are intruded at the top of the hill by a diabase sill of the Nipigon Sill Complex

16 km - Black Sturgeon River (Photo 1)

There are two major regional fault trends, north and northwest, that define the structural zone referred to as the Black Sturgeon Fault Zone. The combination of these two trends has been interpreted to be *en échelon* resulting in the formation of an asymmetric basin or graben, as originally proposed by Coates (1972). A series of subparallel north-trending faults are interpreted to be regionally continuous structures that can be traced

by geophysics and topography for over 150 km north into the area of the English Bay Complex. A series of northwest-trending faults can be traced for similar distances to the west into the Wabigoon Subprovince. Apparent flexures in the north-trending faults probably represent intersections with northwest-trending faults. Vertical displacements on these faults are varied (may be up to 350 m) and are visible along the Black Sturgeon River, interpreted by Coates (1972) to be the half-graben hinge. There appears to have been little lateral displacement on these faults based on the correlation of the Archean iron formations that extend across Black Sturgeon Lake (Hart and Magyarosi, 2004). Corridors of north-trending faults cut, and may have controlled the emplacement of, the Seagull and Disraeli peridotite intrusions, and possibly the Hele peridotite intrusion.

19 km - turn right onto natural gas pipeline access road on the north side of the highway and immediately turn left and follow trail along the north side of the highway. Proceed 200 m to the edge of the gravel pit



Photo 1. The erosional valley of the Black Sturgeon River following a northwest fault, with the Archean to the right (east) and Hele Intrusion to the left (west). The valley is approximately 800 m wide. Photo courtesy of J. Scott.

and park vehicles. Walk for 115 m along trail heading north and skirting the edge of the gravel pit.

Stop 1a - Rock Quarry

UTM coordinates – 0396320E 5416664N

This stop is located on the south side of a rock quarry in a diabase sill which produced fill for highway construction.

This outcrop is the upper contact zone of the diabase sill that extends to the south into the Big Squaw Creek road cut. This contact zone is a relatively rare variant of the typical chilled contact consisting of pebble sized, rounded xenoliths of quartz and feldspar in a biotite-rich fine-grained diabase. Contact relationships observed to the south, in the next stop, indicate that the diabase is intruding sedimentary rocks of the Sibley Group. The composition of the country rock probably influences the formation of this relatively rare contact type, as the diabase sills commonly have glassy chilled contacts. This type of contact zone has been interpreted to be a result of the partial melting and assimilation of country rock xenoliths resulting in a quartz-feldspar restite and a biotite-rich diabase. Similar quartz-feldspar xenoliths in a hornfels carbonate-rich sediment are observed along some contacts in the road cut to the south suggesting that the presence of a carbonate-rich unit may be important to the formation of this contact style. However, a similar contact zone is present in a diabase sill intruding rocks of the Quetico Subprovince without any obvious evidence of Sibley sediments, about 10 km southwest of Beardmore, near the Warnesford Quarry



Stop 1a. Fine-grained, massive, biotitic diabase matrix containing angular to subrounded pebble-sized xenoliths of coarse-grained quartz and white to pinkish feldspar. Markings on scale card are at 1 cm intervals.



Stop 1b. The east block of hornfels units of the Rossport Formation surrounded by diabase sill along the north side of Highway 11/17.

on the CNR railway.

In other areas, intrusion of diabase and ultramafic sills into carbonate-rich Sibley sediment has resulted in the formation of variolitic sills with a chilled margin composed in part by densely packed fine varioles and a core of widely spaced large varioles. Variolitic textures have been observed southwest of Moraine Lake, in a sill cut by diamond drill hole NI92-7 located west of Disraeli Lake, and in and drill hole HE02-01 below the Hele Intrusion.

Return to the highway and turn right.

Proceed west for 600 m, driving past the outcrop, located on the north side of the highway, and turn left into the entrance to a private dump to park.

Stop 1b - Big Squaw Creek road cut

UTM coordinates – 0396114E 5416374N

A diabase sill which has intruded and metamorphosed blocks of Rossport Formation mudstone and dolostone is exposed in a rockcut along the north side of Highway 11/17, in east Stirling Township. The eastern block appears to have been lifted, resulting in a westerly dip, with diabase intruding underneath and from the east (see photo above). However, the block is abruptly truncated by a vertical contact to the west, and a less than 25 cm thick sill intrudes along the bedding eastward from this contact. The ambiguity of the injection direction of the diabase may be a result of the rock cut being oblique to the injection direction, and the block may not be totally separated from the surrounding Sibley Group. The western block is composed of sedimentary rocks comparable in composition to those in the east

block but has a chaotic structure with no preservation of bedding, and is probably a detached, deformed and metamorphosed xenolith

Metamorphism in the sedimentary rocks decreases away from the contact, and the sedimentary rocks close to the contact contain an assemblage of diopside, phlogopite, clinocllore, pargasite, and calcite (Hart, 2005). In some areas along the contact there are irregular zones to isolated pebble sized clasts of quartz and microcline within the calc-silicate hornfels. The quartz and microcline may represent a partial melt of the sedimentary rocks by the diabase, and may be equivalent to the xenoliths observed at the previous stop.

A calc-silicate hornfels or skarn (Meinert, 1992) may be developed along the contact between the diabase sills and the Sibley Group sedimentary rocks, but is ubiquitous along the contacts between the peridotites and Sibley Group contacts (Hart, 2005). Where present, the hornfels is varied in width, extending up to 10 m into the sedimentary rocks, although an approximately 60 m thick layer of hornfels occurs in the Rossport Formation between the two diabase sills in diamond-drill hole ST02-01 (Hart and Magyarosi, 2004). The effects of metamorphism are most pronounced in the Rossport Formation, with the mudstone and siltstones beds commonly developing a light green, mauve, reddish gray to light gray banding. Dolomite-rich beds form light green, white to buff marbles dependent on the original mud content. Radial texture in 3 to 8 cm wide zones occur along some sill contacts, with radially oriented amphibole of the 0.2 to 0.5 cm in diameter.

During his regional study of the Nipigon diabase sills,

Sutcliffe (1986) identified two common assemblages of fine-grained, calc-silicate minerals, calcite-tremolite-forsterite (clinocllore) and calcite-diopside-forsterite at lower and higher metamorphic grades, respectively. The forsterite is altered to serpentine. No dolomite and only minor quartz were observed. Rogala (2001) completed a study of interbedded mudstones and dolomites of the cyclic facies of the Channel Island Member and lower Middlebrun Bay Member of the Rossport Formation, intersected by diamond drill hole NI92-5 located west of Disraeli Lake. These rocks contained a mineral assemblage of pargasite, tremolite, talc, clinocllore, calcite and barite with varying amounts of hematite and apatite, halite, sylvite, gypsum and dolomite. The halite and sylvite were interpreted to be a result of evaporite minerals present in the Sibley Group and the presence of these minerals probably controlled the formation of pargasite (Rogala, 2001). The barite was interpreted to have been introduced into the Sibley Group by later fluid associated with the lead-zinc veining.

Rogala (2001) noted that one sill contact had a thin band of pyrite with chalcopyrite, magnetite, ilmenite, galena, and rare gold and silver. High gold has been noted in a sample from the McVicars Resources Muskrat Lake diamond drill hole ML01-3 at 239.8 m and also in samples from along the shore of Lake Nipigon (A. Richardson, MSc student, Lakehead University, personal communication, 2004). Past workers have suggested that the copper mineralization may be related to the intrusion of the diabase or ultramafic intrusions (e.g., Coates, 1972). However, it is not clear if the metals are igneous or sedimentary in origin.

The diabase is commonly medium brown to brownish grey, massive, medium- to coarse-grained feldspar and pyroxene with trace to 3% olivine and 1 to 2% magnetite. Medium- to coarse-grained diabase forms the majority of the sills and generally lacks a sub-ophitic or diabasic texture and should properly be classified as gabbro, but to avoid confusion with other intrusions in the area the diabase classification has been applied to all rocks associated with the sills. Portions of this outcrop contain ophitic textures consisting of pyroxene oikocrysts up to 2 cm in diameter, and in diamond drill hole the coarse-grained ophitic textured diabase is often interlayered with sub-ophitic diabase. Calculations by Sutcliffe (1986) suggested that the diabase sills were emplaced at shallow crustal levels, at a lithostatic load of between 0.06 and 0.43 kilobars. The sills were estimated to have crystallized over a temperature range

of 1100° to 800°C, adjacent wall rocks were estimated to have reached 665°C, and the metamorphic mineral assemblages suggest temperatures of above 540°C.

26.5 km - west along Highway 11/17, and turn right on to the Stewart Lake Road - reset.

5.9 km follow the Stewart Lake Road for 4.2 km north and then west after a right-angled turn for an additional 1.7 km. Turn right onto the Fowlkes Lakes Road - reset.

2.9 km – north on the Fowlkes Lakes Road to where a skidder trail departs to the west.

Stop 2 - Hele Peridotite

UTM coordinates – 0386762E 5420119N

Park the trucks and proceed by foot along the skidder trail for 500 m to the outcrop.

Located in the southeast corner of Hele Township, the Hele Intrusion covers approximately 40 km² and has a maximum thickness of 130 m, as defined by diamond drilling. Peridotite is very poorly exposed in the central portion of the body, east of Driftstone Lake, and is interlayered with olivine gabbro and feldspathic peridotite in diamond drill core. The peridotite is best exposed in an area of clear cut logging located on the northern edge of Stirling Township as a highly weathered and serpentized rock with a high magnetic susceptibility overlain by a diabase sill. The peridotite is dark green, medium to coarse-grained, massive, variably ophitic, and composed of olivine and pyroxene with variable plagioclase and trace to 1% fine-grained reddish brown mica (Hart, 2005). Numerous, subparallel serpentine and chlorite-rich fractures result in a banded or layered appearance to the otherwise massive peridotite. Contacts between the peridotite, feldspathic peridotite and olivine gabbro observed in diamond drill core are generally gradual transitions in grain size and olivine and plagioclase content. There appears to be more than one band of peridotite interlayered with the melagabbro and feldspathic peridotite suggesting internal subdivisions similar to the Seagull Intrusion.

The Hele Intrusion appears to be a relatively flat-lying sill intruding the Sibley Group with no known feeder zone, although there are fewer diamond drill holes with which to constrain the geometry of the intrusion. Geophysical inversion modelling of the airborne magnetic data by Desmond Rainsford



Stop 2. Example of parallel fractures filled with serpentine and minor biotite in the peridotite of the Disraeli Intrusion. Pencil is approximately 15 cm long.

(geophysicist, Ontario Geological Survey) supports this interpretation (Hart, 2005). Inversion modelling of the ground gravity data indicated the presence of an anomaly located along the western edge of the surface expression of the intrusion and extending to the southwest. This anomaly may represent an extension of the ultramafic intrusion, but recent diamond drill results in the area of the Seagull Intrusion intersected a magnetite rich skarn associated with a similar gravity anomaly.

The Seagull Intrusion hosts PGE mineralization and differs from the Hele Intrusion in having a lopolithic shape, and intruding along or close to the contact between the Quetico Subprovince and the Sibley Group. The central portion of the intrusion appears to be saucer-shaped, but this may be a product of post-intrusion faulting, and a flat-lying sill along the periphery of the body. An underlying, or peripheral, feeder structure for the Seagull Intrusion has not yet been identified.

6.3 km – junction with the Driftstone Road and turn right - reset.

3.9 km – east along the Driftstone Road.

Stop 3 - Hele Intrusion - olivine gabbro

UTM coordinates – 0390358E 5424134N

The olivine gabbro to olivine melagabbro forms much of the exposed area of the Hele Intrusion. The gabbro is fine to medium-grained, medium to dark green, massive and composed of pyroxene, up to 30% plagioclase, and up to 40% olivine. Trace to 1% fine

grained, reddish brown mica is common in the olivine melagabbro in the central areas of the intrusion. An absence or very low abundance of mica and a lower concentration of olivine in the olivine gabbro which forms the northern portion and margins of the intrusion makes it difficult to distinguish these rocks from the olivine-bearing diabase sills. The gabbro contains pods of monzogabbro in outcrop (see photo below) and is interbanded with monzogabbro in the upper portion of the diamond drill holes.

The monzogabbro, or granophyre, is generally medium to coarse-grained, and massive with a variable amphibole content and distinctive pink feldspar or groundmass that occur in irregular shaped pods up to a metre in diameter. These pods are best exposed along the logging trails in the northeast portion of the intrusion, and are hosted by medium to fine-grained olivine gabbro containing minor pink feldspar. Less commonly, the monzogabbro is fine-grained, and massive reddish pink with minor quartz. This type of monzogabbro is present in irregular pods and dykelets with sharp contacts, with and without the coarser grained monzogabbro, in the western portion of the intrusion. Some of the best examples are in the central part of the intrusion, close to a small exposure of Sibley Group sedimentary rock that is in contact with the upper portion of the intrusion. In diamond drill core, the two types of monzogabbro are generally closely associated and in some examples the finer grained variety is rimmed by the coarser grained rock. The abundance of pink feldspar correlates with elevated K_2O values, and has been interpreted to be the result of the assimilation of



Stop 3. Contact between olivine gabbro and an irregular pod of monzogabbro / granophyre, in the western portion of the Hele Intrusion. The pen magnet is approximately 16 cm long.

Sibley Group sedimentary rocks (Hart and Magyarosi, 2004). The finer grained monzogabbro may be a poorly assimilated sediment xenolith and the coarser grained monzogabbro may be the result of assimilation of the volatile rich sedimentary rocks.

Turn vehicles around and return by the same route to the corner of the Driftstone Road and the Stewart Lake Road 10.2 km to the south.

Turn right onto the Stewart Lake Road.

3.8 km - Stewart Lake Road ends at the Black Sturgeon Road.

Turn right onto the Black Sturgeon Road and drive 3.3 km north.

Stop 4 - Black Sturgeon Road

UTM coordinates – 0385313E 5418737N

Stop at kilometer 9 of the Black Sturgeon Road and look to the east across the clear cut, a diabase sill intrudes the Rosspport Formation of the Sibley Group.

Jointing in the diabase sills may be either arcuate or linear. The arcuate joints formed by cooling of the magma vary in strike direction by up to 70° and are commonly observed on horizontal surfaces. Jointing is best exposed in an area to the north, of the clear cut within the 1999 burn. Jointing located close to the sill contacts forms crude polygon joint patterns with “T” or flattened “Y” junctions, and the interiors of the sills lack the “Y” junctions commonly present in polygonal jointed unit (e.g., Aydin and DeGraff, 1988). This lack of “Y” junctions suggests a slow cooling rate which would be in agreement with the 315 to 560 year cooling times for the sills estimated by Sutcliffe (1986).

Looking to the northeast, an offset in the basal

contact of the sill is evident, which appears to be a result of later movement along a northwest-trending structure. Joints are rare in the outcrops forming the interior of the sills, and when present are generally linear in nature. The column-like jointing is probably a result of late stage fracturing of the cooling magma combined with later fault activity rather than a true columnar jointing.

20.3 km – *this portion of the Black Sturgeon Road is underlain by units of the Sibley Group and outcrops are rare.*

Stop 5 - Eagle Mountain

UTM coordinates – 0385387E 5433471N

Stop at the 20.3 km in the open space on the east side of the road for a view of Eagle Mountain.

Eagle Mountain is an approximately 10 km long ridge located west of the Black Sturgeon River and is the first of a series of north-trending fault bounded ridges located further west. The ridges are commonly capped by a diabase sill that has intruded units of the Kama Hill Formation. The lower portions of the ridges are generally poorly exposed but appear to be composed of units of the Rosspport Formation. In some areas, as in Eagle Mountain, a second diabase sill intrudes the lower portions of the ridge. There were three diamond drill holes that were completed on the lower slope of Eagle Mountain that provide a stratigraphic section though the lower portion of the Sibley Group. Two of these holes intersected a mafic dyke, which is also exposed in a ravine on the central portion of the ridge. The 5 to 10 m wide fine-grained massive dyke dips approximately 50° south. This dyke appears to be related to a series of fine grained mafic sills with a higher olivine content. Although, these sills



Stop 4. Nipigon diabase sill intruding units of the Rosspport Formation located along the Black Sturgeon Road. Photo is looking to the southeast.



Stop 5. Looking north from the top of Eagle Mountain, with units of the Kama Hill Formation intruded, overlain, by a diabase sill. The sill – sediment contact is approximately along the upper tree line in the cliff.

are geochemically similar to the ultramafic intrusions, they do not appear to be associated with any known ultramafic intrusion (Hart, 2005; Richardson and Hollings, 2005).

22.5 km – (km 26 on the Black Sturgeon Road) turn right at the load levelers, on to the Camp 42 Road - reset.

14.1 km - continue east and then north on the Camp 42 Road until the junction with the Church Road - reset.

7 km – north along the Church Road.

Stop 6 - Andalusite schist

UTM coordinates – 0388613E 5449432N

These rocks consist of metamorphosed feldspathic wacke, lithic wacke, and siltstone with metamorphic grade increasing progressively toward the south (Hart 2005). The metawackes and metasiltsstones are light grayish brown to medium gray, with 3 to 30 cm thick beds, and a schistosity parallel to bedding resulting in the variable preservation of original textures. These rocks commonly contain 5 to 10% biotite, trace to 1% very fine-grained garnet and, a trace subhedral mineral tentatively identified as cordierite. Specific bands are composed of individual to multiple beds containing 10 to 15% coarse- to very coarse-grained, subhedral to euhedral andalusite. The andalusite occurs disseminated through some beds and occasionally concentrated along parallel bands that appear to represent the original bedding plane. This distribution probably reflects the distribution of clay minerals, with higher Al_2O_3 content,

within the wackes (Hart and Magyarosi, 2004). The presence of andalusite, with minor cordierite, suggests a low pressure, high temperature metamorphic event possibly related to the emplacement of the felsic intrusive rocks. The andalusite pseudomorphs are partially replaced by a mica indicating some degree of retrograde metamorphism.

These metasedimentary rocks closely resemble the clastic metasedimentary rocks of the Quetico and Wabigoon subprovinces in the Beardmore area, east of Lake Nipigon (Hart et al., 2002a). For this reason, these rocks are interpreted to have accumulated in a trench environment of an accretionary complex comparable to the depositional environment of the Beardmore-Geraldton greenstone belt (e.g., Williams, 1989). The transition from the southern Wabigoon to Quetico subprovince is located approximately 25 km to the north, where a series of iron formations appear to correlate with the iron formations of the Beardmore-Geraldton belt (e.g., Sutcliffe, 1986). However, the iron formation exposed on the east side of Black Sturgeon Lake (Hart and Magyarosi, 2004) is comparable to the chert-magnetite iron formation in the Mawn Lake area (Coates, 1972) rather than the chert poor units of the Beardmore area.

Turn vehicles around

7 km - proceed back along the Church Road to the junction with the Camp 42 Road

0.750 km – trail to left, proceed south up hill along the trail

0.200 km – park and walk right, to the west, for ~10 m



Stop 6. Coarse-grained andalusite pseudomorphs in metamorphosed feldspathic wackes and siltstones of the Quetico Subprovince.

Stop 7a - Granite – Diabase Contact

UTM coordinates – 0390536 E 5442913N

The contact between the leucocratic muscovite granite of the Quetico Subprovince and the Mesoproterozoic diabase sill of the Nipigon Sill Complex is exposed in this flat-lying outcrop. The medium to coarse-grained leucocratic muscovite granite occurs as rounded xenoliths up to 20 cm in length in the chill zone of the diabase sill (see photo below). The sill is composed of a fine to very fine-grained, glomeroporphyritic diabase with polygonal jointing and some areas have the glassy chill zone preserved. This outcrop is interpreted to be the upper contact of the diabase sill which is dipping shallowly to the north, and the hill to the southwest is capped to the same diabase sill. This same sill caps the hill above the outcrop at stop 7b.

About 20 m to the right or east of the trail, an outcrop of the Quetico Subprovince granite is cut by a northeast-trending, approximately 8 cm wide, aphanitic diabase dyke. Dykes may occur close to the diabase sill contacts generally extending for less than a few metres, but may in some cases extend for a few tens of metres. This dyke is thought to be injected from the diabase sill which is located beneath this outcrop.

1.60 km – return to the Camp 42 Road and continue back to the gravel pits.

Stop 7b - Pegmatite and Nipigon Sill

UTM coordinates – 0389764E 5442111N

The outcrop is located approximately 300 m east of the road, and is accessible by a trail leading from the back of the small gravel pit.



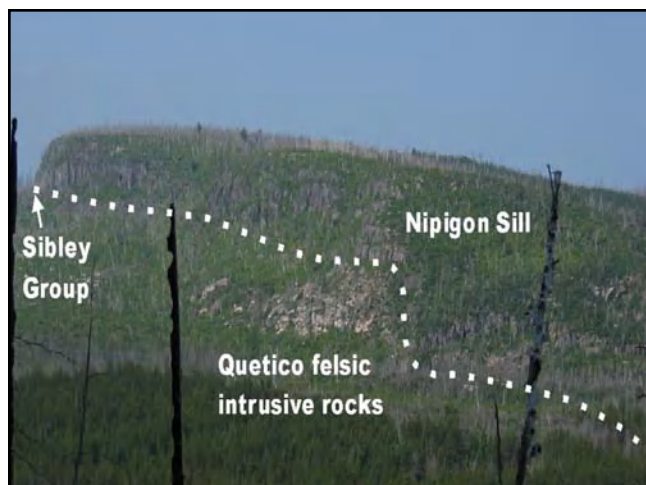
Stop 7a. Xenoliths of Quetico granite in the glassy chill zone of a Nipigon diabase sill.

The west dipping outcrop consists of a muscovite granite dyke intruding biotite schists, and the hill above is capped by a diabase sill. Muscovite granite dykes in this area are generally a few metres in width but may be up to a few tens of metres wide, and are usually irregular in form and orientation although generally following the fabric of the surrounding country rock. This outcrop is composed of pegmatitic quartz, potassium feldspar, and plagioclase with graphic intergrowths of quartz and feldspar. Black, coarse-grained to pegmatitic tourmaline occurs as tabular grains and as graphic intergrowths with quartz. Fine grained subhedral, blue-green apatite occurs as an accessory mineral located along the contacts with the biotite schist country rock. This dyke is one of a number of dykes intruding the metamorphic rocks adjacent to the leucocratic muscovite granite.

A body of leucocratic muscovite granite occurs to the north and northwest and is light gray, pinkish gray, to white, massive, and medium to very coarse-grained with occasional pegmatitic sections, composed of quartz and potassium feldspar, and plagioclase with typically less than 5% muscovite (Hart, 2005). Irregular to tabular shaped xenoliths of metasedimentary and schistose rocks up to a few metres in diameter are present throughout, but appear to be more common towards the margins of the body. Numerous muscovite granite dykes intrude the massive body and may constitute up to 30% of the outcrop. These dykes are commonly a few metres in width, but may be up to a



Stop 7b. Black tourmaline in the leucocratic muscovite granite dyke.



Stop 7b. Northeast dipping diabase sill observed to the west of the pegmatitic dyke outcrop.

few tens of metres. The granite dykes are composed of pegmatitic quartz, potassium feldspar, and plagioclase and frequently contain graphic intergrowths of quartz and feldspar, black, coarse-grained to pegmatitic, tabular tourmaline usually graphically intergrown with quartz and plumose intergrowths of muscovite-quartz. Trace to 1% fine-grained, disseminated, subhedral, reddish brown garnet, fine grained light green mica or rosy quartz was identified in a few of the dykes.

Contact relationships suggest that the leucocratic muscovite granite and pegmatite dykes intruded the country rocks rather than being products of partial melting of the host country rocks. However, the initial mode of formation is not known and these rocks may have formed by partial melting of sedimentary rocks at depth.

The pegmatitic, muscovite leucogranite body and

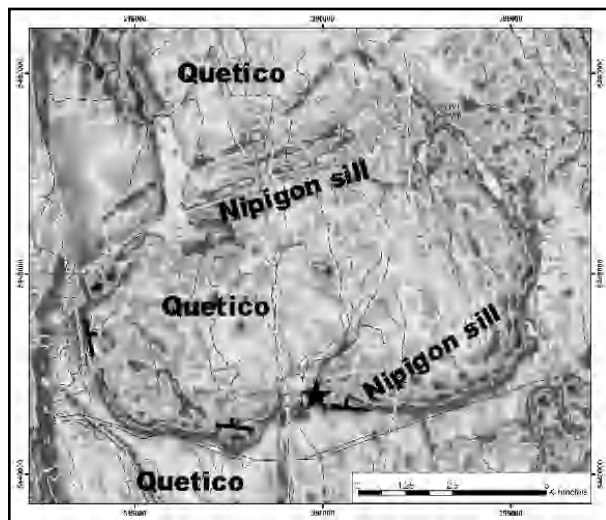


Figure 11. First vertical derivative of the total field magnetic data the inward dipping saucer shaped diabase sill. Star marks the location of field trip Stop 7.

dykes contain intergrowths of tourmaline and quartz, and muscovite and quartz commonly observed in rocks that have been classified as fertile granites, or granites that have the potential to host rare-element mineralization (Breaks et al., 2003). Additional work is on-going to characterise these bodies. The muscovite leucogranite occurs along the regional trend extending from Georgia-Barbara lakes area, about 60 km to the northeast, to the DeCoursey Lake and Onion Lake areas, approximately 45 km to the southwest (Breaks et al., 2003), and this area may represent a similar concentration of fertile granites.

Looking west across the valley, a diabase sill can be observed to dip shallowly to the northeast, whereas along the Black Sturgeon River the same sill can be observed to dip shallowly east (see Photo below). The shallow inward dipping orientation of this sill, forming a broad saucer with a width of about 5 km, is also visible in the airborne magnetic pattern (Fig. 11) (Ontario Geological Survey 2004). A number of these saucer-shaped sills, with widths of up to 6 km and interiors occupied by older rock types, are evident from airborne magnetic surveys both in this area and in sills intruding Archean rocks of the Wabigoon Subprovince to the northeast and northwest. Variation in sill orientation over short distances is common in other mafic sill complexes (e.g., Tasmania: Leaman, 1975; Karoo: Chevailler and Woodford, 1999). This change in orientation may be due to a number of factors including a change in orientation of the principal stress direction (Gretener, 1969) during sill emplacement, and the sill may follow pre-existing structures such as faults. There appear to be at least three major sills (100 to ~250 m), and possibly a number of thinner (5 to 25 m) sills in the area, based on a combination of outcrop and diamond drill hole information. However, correlation between outcrops and diamond-drill holes is complicated by the saucer shape of some sills, the presence of block faulting that may post-date sill emplacement, and the lack of distinctive geochemical differences between the sills.

12 km – back to the junction of the Camp 42 and Black Sturgeon roads

26 km – south to Highway 11/17 along the Black Sturgeon Road – bypassing the Stewart Lake Road

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