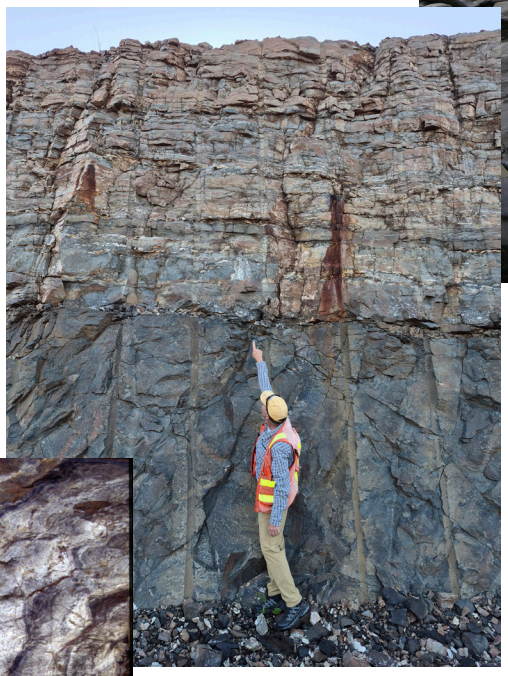
72nd Annual Meeting

Thunder Bay, Ontario - May 21-22, 2026

Institute on Lake Superior Geology

Part 2 – Field Trip Guidebook







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72tnd Annual Meeting

Institute on Lake Superior Geology

May 21-22, 2026

Thunder Bay, Ontario

HOSTED BY:

Mark Puumala and Peter Hinz

Co-Chairs

Ontario Geological Survey (Retired)

Proceedings - Volume 72

Part 2 – Field Trip Guidebook

Compiled and edited by Pete Hollings

Cover Photos: Top - Keweenawan diabase dyke on Lake Superior shoreline near Thunder Bay, Middle - Archean-Paleoproterozoic unconformity, Highway 11-17, near Pass Lake turnoff, Bottom - Colloform stromatolite, Gunflint Formation, Kakabeka Falls

72nd Institute on Lake Superior Geology

Volume 72 consists of:

Part 1: Program and Abstracts

Part 2: Field Trip Guidebook

Trips 1 & 4: "Classic" Geological Sites in the Thunder Bay Area

Trip 2: Geology of the Quetico Supprovince North of Thunder Bay

Trip 3: Gold Deposits of the Shebandowan Greenstone Belt

Trip 5: Structural Geology and Gold Mineralisation of the Mine Centre Area

Trip 6: Amethyst Deposits of Thunder Bay

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

Poulsen, K.H., 2026. Archean Geology and Metallogeny of the Rainy Lake Wrench Zone. In; Hollings, P. (Ed.), Institute on Lake Superior Geology Proceedings, 72nd Annual Meeting, Thunder Bay, Ontario, Part 2 - Field trip guidebook, v.72, part 2, 3-31.

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Introduction - considerations and acknowledgements

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and

Mark Puumala

Resident Geologist Program, Ontario Geological Survey, Ministry of Energy & Mines,

Thunder Bay, Ontario (Retired)

This volume is intended to serve not only as a guide for the 72nd ILSG field trip participants but also as a reference for those interested in reprising the trips at a future date. In order to facilitate this, trip leaders have provided UTM coordinates in the NAD 83 datum for stops, as well as plain word descriptions for locating each trip stop. It should be noted that some stops are located on private land or registered mining claims. As such, individuals visiting these stops are advised to obtain the land holders' permission prior to entering their property. If in doubt, we recommend contacting the Resident Geologist Program office in Thunder Bay for further information about current property ownership.

This year's slate of field trips include stops located either on provincial highways or busy logging roads which can create safety issues. For those participating in facilitated trips at this year's meeting, make sure to pay attention to the trip leaders' safety orientation at the start of the trip, and follow any stop-specific safety instructions. For individuals using the field trip guide for future private tours it is advisable to be wary of road traffic and exercise extreme caution. Please take care when crossing or parking at the sides of these roads.

The organizing committee would like to thank all the field trip leaders who authored and contributed to this field guide along with those who provided comments and/or assisted with the running of the trips themselves. Field trip leaders and authors include Howard Poulsen, Riku Metsaranta, Gaetan Launay, Dorothy Campbell, Justin Jonsson, Vittoria D'Angelo, Mark Smyk, Mark Puumala, Steve Kissin and Greg Paju.

The Committee thanks participating exploration companies and mine operators for their cooperation and assistance in providing access and information in regards to their properties, as well as their staff time for leading the tour participants on their respective properties. Participating companies include Delta Resources, Gold X2 Mining, Amethyst Mine Panorama and Diamond Willow Amethyst Mine.



Figure 1. Map illustrating general locations of ILSG 2026 field trips. Symbols are labelled with numbers that correspond to the trip numbers (1 to 6) used in the meeting program and field trip guidebook.

Trips 1 & 4 - "Classic" Geological Sites in the Thunder Bay Area

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Introduction

The geology of the Thunder Bay area features a variety of Archean and Proterozoic rocks of the Superior and Southern Provinces of the Canadian Shield, respectively, as well as unconsolidated deposits and landforms associated with Quaternary glacial and post-glacial processes. This field trip features examples of many of these rocks and features, providing an overview of the varied geology the area has to offer. A number of field guides (e.g., Pye, 1969; Kustra et al., 1977; Franklin et al., 1982) have covered the Thunder Bay area, including those written for the 46th (e.g. Pufahl et al., 2000; Phillips et al., 2000) and 58th Institute on Lake Superior Geology annual meetings (e.g. Fralick et al., 2012; Smyk, 2012; Phillips et al., 2012; Cundari et al., 2012). These guides contain descriptions of some of the field trip stops covered in this guide and they will be referenced appropriately. This guide also benefits from ongoing local research and mapping conducted by the Ontario Geological Survey, Geological Survey of Canada and Lakehead University.

Day One of this trip features exposures north and east of Thunder Bay, while those of Day Two are located south and west of the City. Bear in mind that this trip marks the first time that many of these stops have been visited and described as part of a formal field trip. This is especially true of stops along Highway 11-17, whose expansion ca. 2010-2012 produced many remarkable new exposures. Please exercise caution when stopping and viewing roadside outcrops. Permission or admittance may need to be obtained to visit some stops; this will be outlined in the guide when necessary.

Regional Geology Overview

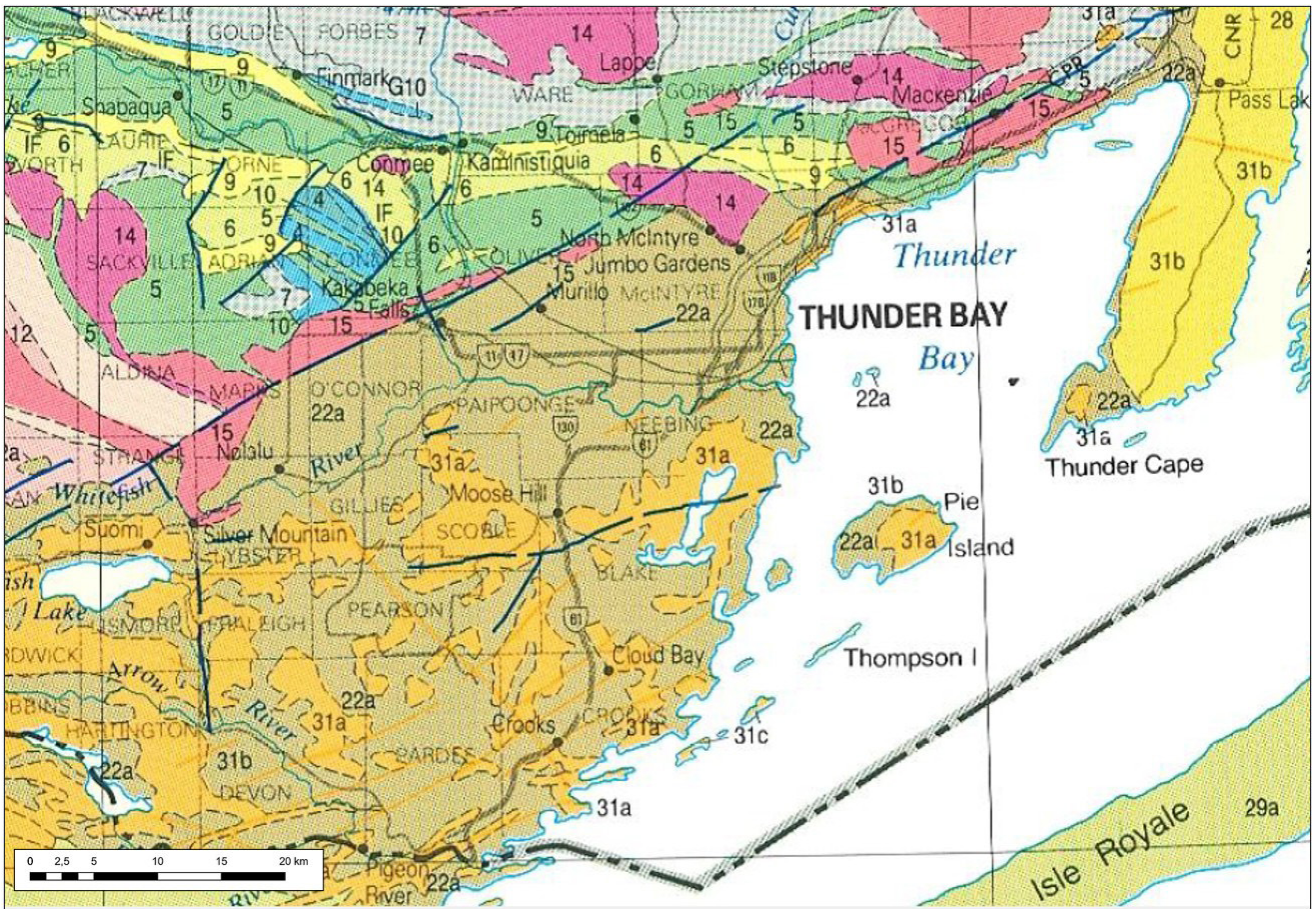
Precambrian Geology

The Thunder Bay area straddles the boundary between Archean rocks of the Superior Province and Proterozoic rocks of the Southern Province (Figure 1). In the vicinity of Thunder Bay, Superior Province rocks comprise volcano-plutonic rocks of the Neoarchean Wawa Subprovince and metasedimentary and granitoid rocks of the Neoarchean Quetico Subprovince, bounding the Wawa to the north.

Locally, the supracrustal rocks of the Wawa Subprovince have been subdivided into the Greenwater and Shebandowan assemblages (Williams et al., 1991). The ca. 2.72 Ga Greenwater assemblage consists of a north-younging sequence of mafic to felsic metavolcanic rocks with subordinate interbedded clastic and chemical metasedimentary rocks. Mafic metavolcanic rocks within this assemblage consist predominantly of tholeiitic to calc-alkalic pillowed flows. The intermediate and felsic metavolcanic sequences are calc-alkalic and consist predominantly of coarse-grained pyroclastic deposits and massive to feldspar-phyric flows. The ca. 2.69 Ga Shebandowan assemblage is a younger, possibly fault-bounded (Williams et al., 1991) sequence of sub-alkalic to alkalic, predominantly coarse-grained pyroclastic metavolcanic rocks with interbedded coarse- to fine-grained, commonly well-preserved, proximal metasedimentary rocks. Intrusions within the Wawa Subprovince supracrustal assemblages consist of narrow felsic dikes, syenitic to tonalitic pre- to syn-tectonic plutons, minor gabbro bodies and scattered narrow mafic dikes.

Rocks of the Neoarchean Quetico Subprovince abut the Wawa Subprovince to the north. They consist mainly of clastic metsedimentary rocks (turbiditic wacke, arkose, quartz arenite, slate and argillite) as well as post- to syn-deformational, syenitoid to granitoid plutons (cf. Metsaranta, 2022; Metsaranta and Walker, 2019). Migmatization becomes common in the rocks towards the northern portion of the area as metamorphic grade increases (Williams, 1991).

The Southern Province consists of Proterozoic



MESOPROTEROZOIC

31 **Mafic and related intrusive rocks (Keweenawan age)**[l]
31a *Logan and Nipigon sills (1109 Ma):* diabase sills
31b *Pigeon River and Pukaskwa swarms:* diabase dikes

29 **Osler Gp.; Mamainse Point Fm., Michipicoten Island Fm.**
29a Basalt and associated conglomerate and arkose
29b Rhyolite, quartz feldspar porphyry; associated conglomerate and arkose

28 **Sibley Gp.:** conglomerate, sandstone, shale

PALEOPROTEROZOIC

22 **Sedimentary rocks**
22a *Animikie Gp.*[m]: wacke, shale, iron formation, limestone, minor volcanic rocks

NEOARCHEAN

INTRUSIVE ROCKS

15 **Massive granodiorite to granite:** massive to foliated granodiorite to granite
15a Potassium feldspar megacrystic units

14 **Diorite–monzonite–granodiorite suite:** diorite, tonalite, monzonite, granodiorite, syenite and hypabyssal equivalents (saturated to oversaturated suite)

13 **Muscovite-bearing granitic rocks:** muscovite-biotite and cordierite-biotite granite, granodiorite-tonalite

12 **Foliated tonalite suite:** tonalite to granodiorite—foliated to massive

11 **Gneissic tonalite suite:** tonalite to granodiorite—foliated to gneissic—with minor supracrustal inclusions

10 **Mafic and ultramafic rocks**[q]: gabbro, anorthosite, ultramafic rocks

SUPRACRUSTAL ROCKS

9 **Coarse clastic metasedimentary rocks**[r]: mainly coarse clastic metasedimentary rocks, with minor, mainly alkalic, mafic to felsic metavolcanic flows, tuffs and breccias

8 **Migmatized supracrustal rocks**[eg]: metavolcanic rocks, minor metasedimentary rocks, mafic gneisses of uncertain protolith, granitic gneisses

7 **Metasedimentary rocks**[eg]: wacke, arkose, argillite, slate, marble, chert, iron formation, minor metavolcanic rocks
7a Paragneisses and migmatites[s]
7b Conglomerate and arenite

6 **Felsic to intermediate metavolcanic rocks**[gt]: rhyolitic, rhyodacitic, dacitic and andesitic flows, tuffs and breccias, chert, iron formation, minor metasedimentary and intrusive rocks; related migmatites

5 **Mafic to intermediate metavolcanic rocks**[gt]: basaltic and andesitic flows, tuffs and breccias, chert, iron formation, minor metasedimentary and intrusive rocks, related migmatites
5a Andesitic flows, tuffs and breccias with minor rhyolites[u]

4 **Mafic to ultramafic metavolcanic rocks**[gt]: mafic metavolcanic rocks with minor komatiite, minor metasedimentary and pyroclastic rocks

Figure 1. Generalized geology of the Thunder Bay area; geology from Ontario Geological Survey Map 2542, Bedrock Geology of Ontario, West-Central Sheet, scale 1:1 000 000 (1991).

rocks which unconformably overlie or intrude Archean basement rocks of the southern Superior Province (cf. Tanton, 1931; Pye, 1969). North and west of Lake Superior, the Southern Province comprises:

1) Paleoproterozoic (ca. 1.8 Ga) Animikie Group sedimentary and minor volcanic rocks;

2) Mesoproterozoic (ca. 1.4 Ga) Sibley Group sedimentary rocks; and

3) Mesoproterozoic (ca. 1.1 Ga) Midcontinent Rift volcanic and intrusive rocks.

The Animikie Group, exposed in Ontario, Minnesota, Wisconsin and Michigan, is represented locally by the Gunflint Formation and overlying Rove Formation. These dominantly sedimentary formations constitute a largely unmetamorphosed, undeformed, homoclinal succession which dips shallowly to the southeast. The Gunflint Formation is a chemical-clastic assemblage which yielded a U-Pb age from reworked volcanic ash of 1878.3 ± 1.3 Ma (Fralick et al., 2002). Rocks containing intraformational breccias, accretionary lapilli, spherules and shocked quartz that occur near the top of the Gunflint Formation are interpreted to represent ejecta from the Sudbury impact event that occurred at *circa* 1850 Ma (Addison et al., 2005; Krogh, Davis and Corfu 1984). The Gunflint Formation grades upward into turbiditic sandstone and shales of the Rove Formation south of Thunder Bay. U-Pb zircon ages from ash beds in the basal Rove Formation yielded 1836±5 and 1832±3 Ma (Addison et al., 2005). A sandstone sample from the submarine fan portion of this succession yielded a youngest detrital zircon U-Pb age of approximately 1780 Ma (Heaman and Easton, 2006), but this relatively young age is widely considered to be problematic and not reflective of the true age of these rocks. Sedimentation in this part of the Animikie basin, widely thought to represent the distal foreland of the Penokean Orogen, likely ended ca. 1800 Ma or earlier. However, Maric (2006) suggested that the Rove (and correlative Virginia) Formation represents the transition from a sediment-starved basin, with exceedingly slow deposition rates, to active deltaic progradation with sediment probably derived from the Trans-Hudson orogenic zone to the north.

The Sibley Group, exposed on the Sibley Peninsula and farther north, has been subdivided into five formations; detailed descriptions of each formation have been reported previously (Franklin et al., 1980; Cheadle, 1986; Rogala, 2003; Rogala et al., 2005, 2007). The overall sedimentary environment indicates a fluctuating climatic scenario, in which the Sibley Group was deposited in a lacustrine system (Pass Lake Formation) that gradually evolved into a saline playa lake environment (Rossport Formation). As the climate progressively became drier, a sabkha-type environment developed (Fire Hill Member of the Rossport Formation). After a break in time, the Kama Hill and Outan Island formations represent outbuilding of a large deltaic complex to the north (Jones et al., 2022), and the Nipigon Bay Formation represents an aeolian environment (Rogala, 2003; Rogala et al., 2007). The depositional age for much of the Sibley Group had been constrained between ~1340 and 1450 Ma.

The northern margin of the Midcontinent Rift is dominated by mafic hypabyssal rocks of the Midcontinent Rift Intrusive Supersuite (Miller et al. 2002), which intrude all Proterozoic rocks and Archean basement. South of Thunder Bay, Logan (1106.3+2.0 Ma; Smith et al., 2025) diabase sills predominate. Nipigon diabase sills (1108.2+0.9 Ma; Bleeker et al., 2020) occur in and north of the City, and form the bulk of the Nipigon Embayment. Volcanic and minor sedimentary rocks of the ca. 1108 to 1105 Ma Osler Group (Davis and Sutcliffe, 1985; Davis and Green, 1997) are exposed to the east on Black Bay Peninsula and on offshore islands in Lake Superior. While all aforementioned rocks are related to the Early Magmatic Stage (*ca.* 1110–1103 Ma; Miller and Nicholson, 2013) of Midcontinent Rift development, younger intrusions (ca. 1097-1092 Ma; Smith et al., 2025) are associated with a magmatic episode that followed the emplacement of the 1099 Ma Duluth Complex. Three main domains were suggested by Smith et al. (2025) in describing the northern flank of the Midcontinent Rift west of Thunder Bay and Lake Nipigon, namely, from south to north: (1) a gently south-(southeast-)tilted Midcontinent Rift margin; (2) a pronounced basement arch just north of Thunder Bay, likely representing a flexural bulge; and (3) the erosional remnant of the Lake Nipigon rift-and-sag basin, preserving the Sibley Group intruded by extensive Nipigon diabase sills (Figure 2).

Quaternary Geology

The first Pleistocene ice sheet in the Thunder Bay region, ca. 1 Ma, moved over and stripped a deeply weathered, relatively flat bedrock landscape (Zaniewski et al., 2020). During the Pleistocene, possibly ten or more major advances and retreats of ice took place, each with its own history of advancing and retreating lobes of ice. The region's present landscape is the product of interplay between three major ice lobes (i.e. Patricia or Rainy River Lobe, from the north; the Hudson Bay Lobe, from the northeast; and the Superior Lobe, from the east) originating from three accumulation centers

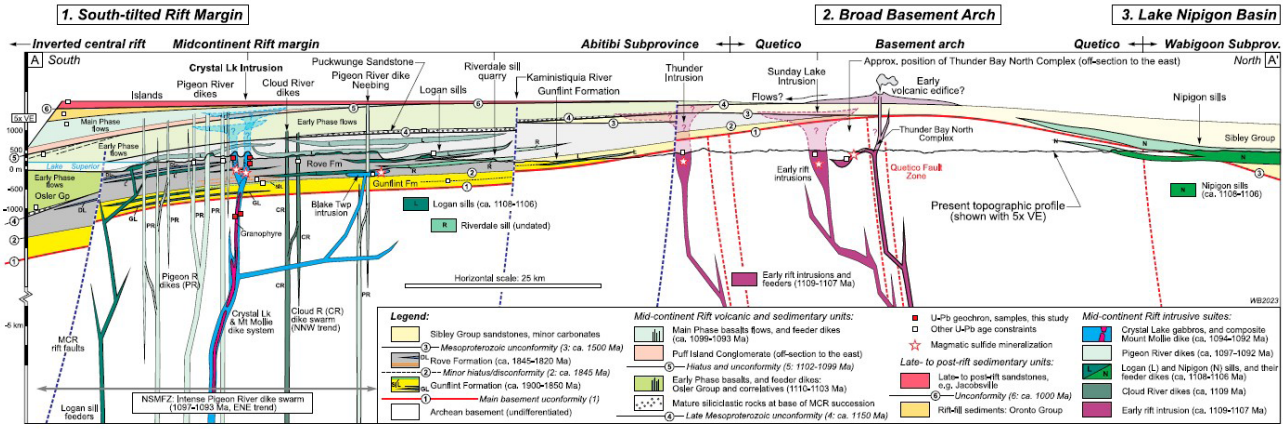


Figure 2. West-looking cross section through the northern flank of the Midcontinent Rift, just west of Thunder Bay and Lake Nipigon, from Smith et al. (2025)

during Wisconsin glaciation. Only the final event, comprising the Marquette Readvance (ca. 11 500 Ka) and its subsequent retreat, is understood in local detail (ibid).

Starting about 11,000 years ago (Ka), Wisconsin ice melted back from its position in central Minnesota and Wisconsin, and quickly exhumed the Thunder Bay region, forming recessional moraines during brief stillstand periods (Phillips, 2004; Phillips et al., 1994). The Lake Superior basin was occupied by Early Lake Minong, the shoreline of which is found close to the 1400-foot (427m) contour in the borderland area. About 10 Ka, ice re-advanced from north of Lake Nipigon, sweeping across the Superior Basin (Marquette Readvance). As that ice began to melt, glacial lakes were formed between the moraines and the retreating ice margins. As Superior ice melted, water levels progressively lowered, forming a series of shoreline features down-slope and depositing thick lacustrine clays. Superior ice withdrew to the north of Lake Nipigon around 9.5 Ka, and for the first time since the Marquette Re-advance, the Superior basin was occupied by a single lake, Lake Minong. This lake level extended up the Kaministiquia embayment to Rosslyn, where a large delta structure was built. The Minong shoreline runs through the upper part of the city, being particularly evident in Boulevard Park where river mouth bars and terraces of the Current River are seen. The Minong shoreline in the city is strongly associated with Palaeo-Indian sites, the Cummins Site being the best-known. It is likely that as water levels fell, these early people moved down from the Arrow-Whitefish Lakes area into the Kaministiquia embayment. Little remains of the toolkit of these people other than a variety of knapped lithic tools made from taconitic chert that occurs in the local Gunflint Formation (cf. Hamilton, 1996).

Field Trip Stop Descriptions - Day One

Day One begins with visits to a number of locations northeast of the City, clustered around the northern end of Thunder Bay of Lake Superior (Figure 3) and ends near and within the City (Figure 4). This small area is underlain by a variety of rocks that record almost three billion years of local geologic history, spanning from the Neoarchean (ca. 2.7 Ga) to the Paleoproterozoic (ca. 1.8 Ga) and Mesoproterozoic (ca. 1.4 and 1.1 Ga) and perhaps to the Mesozoic (ca. 100 Ma). Unconsolidated glacial and post-glacial deposits and features attest to a long-lived, Pleistocene glaciation record. All GPS coordinates are NAD83, UTM Zone 16.

STOP 1-1: Blende Lake Unconformity (0367703 E / 5383357N)

This exceptional highway rock cut, like many others on this stretch of Highway 11-17, was exposed by new highway excavations ca. 2012. This ~700 m-long exposure features the unconformity between Neoarchean Wawa metavolcanic and gabbroic rocks and Paleoproterozoic sedimentary rocks of the lower Gunflint Formation (cf. Scott, 1990; Figure 5). Basement rocks here have also been described by Landman (2021) as coarse-grained amphibolite, interpreted as a mafic intrusion which has undergone amphibolite-facies metamorphism.

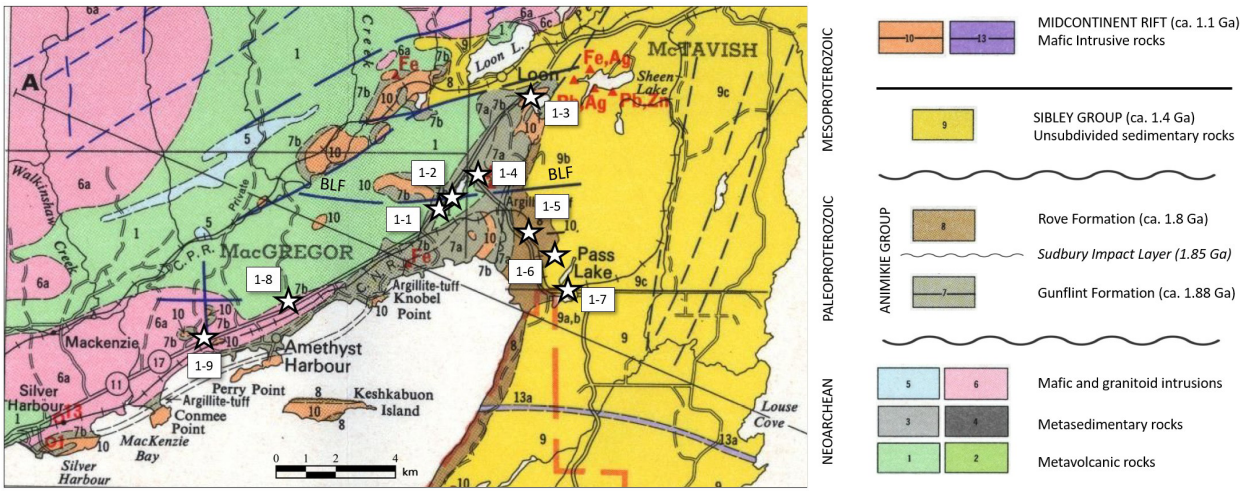


Figure 3. Generalized geology of the northern end of Thunder Bay of Lake Superior, showing the first 9 field trip stop locations of Day 1. Geology from Map M2232 (Carter et al., 1973). BLF – Blende Lake Fault

This exposure was described by Metsaranta and Kurcinka (2022), as part of an ongoing Ontario Geological Survey (OGS) bedrock mapping project of the Animikie Basin near Thunder Bay:

> *"...chloritized Archean felsic to intermediate intrusive rocks are locally overlain by at least 4 stromatolite mounds comprising thinly laminated black to red chert [Figure 6]. The stromatolite mounds have a height of up to 30 to 50 cm and similar widths. The top of one mound is marked by a thin stylolitic band [Figure 7]. The areas between stromatolite mounds comprise silicified grainstones that locally contain sulphide nodules up to 5 cm in diameter. The grainstones enclosing the stromatolite form medium to thick beds characterized by medium- to large-scale trough cross-stratification. At this locality, an east-dipping and roughly north-striking small displacement thrust fault puts Archean basement rocks above Gunflint Formation rocks. The fault appears to displace Midcontinent Rift–related quartz-carbonate-sulphide veins indicating*

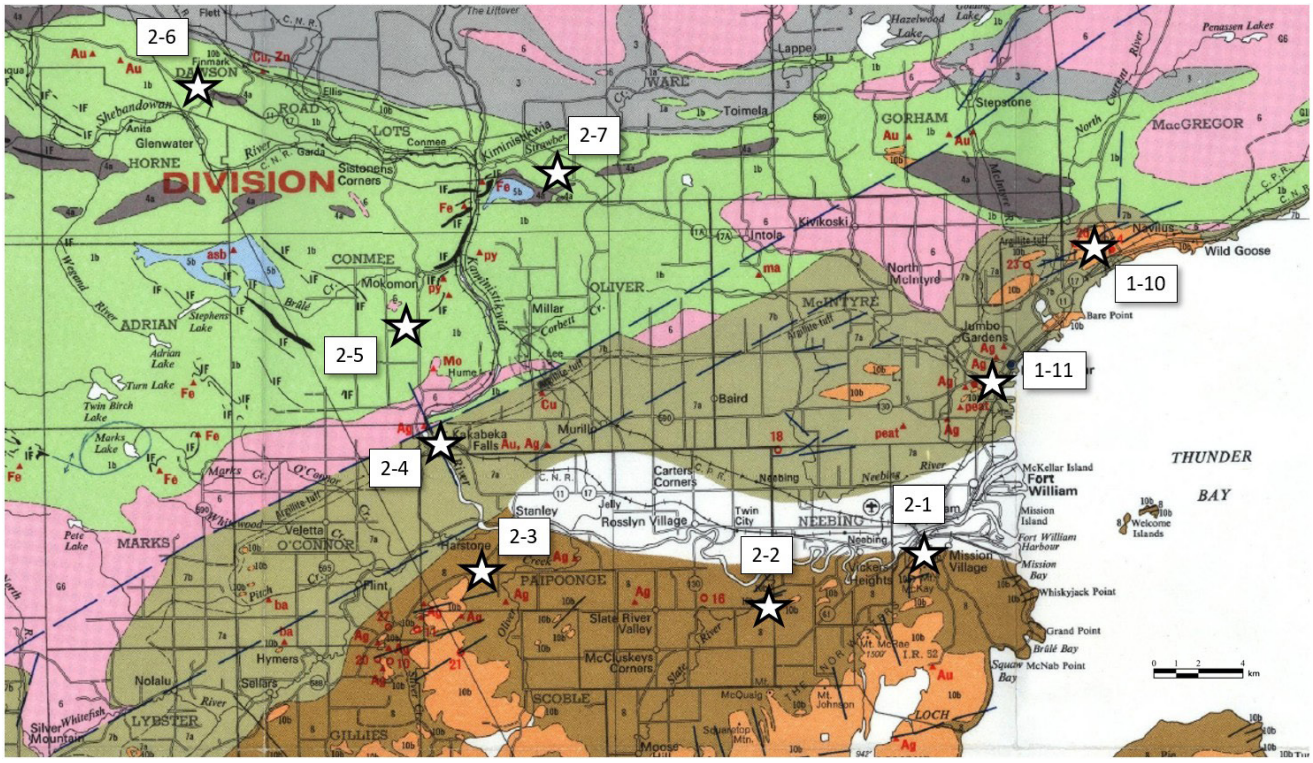


Figure 4. Field trip stop location map, showing Day 1 stops 1-10 and 1-11 and Day Two stops. (See Figure 3 for map legend).



Figure 5. Paleoproterozoic Gunflint Formation sedimentary rocks unconformably overlying Archean basement, east side of Highway 11-17, STOP 1-1.

that the thrusts may be related to rift inversion; however, this field relationship is equivocal."

These ferroan dolomite and siderite grainstones (medium-grained, sand-sized iron carbonates), referred to as granular iron formation, are common in the Thunder Bay region, dominating the near-shore of the Animikie Basin (see Fralick et al., 2012; STOP 1-4).

The thin basal conglomerate (*aka* Kakabeka Member, Figure 8) of the Gunflint Formation is discontinuously distributed along the paleosurface. As described by Metsaranta and Kurcinka (2022), the conglomerate has a clast-supported texture, consisting of coarse-grained sand, granules, pebbles and rare cobbles in a sandy matrix. Quartz, pink granitoid, clastic metasedimentary and mafic metavolcanic clasts were noted.

Kup et al. (2025) also noted that the conglomerate



Figure 7. Stylolites in Gunflint Formation grainstone, STOP 1-1, visible as a black serrated band to the right of the scale card.



Figure 6. Mound-shaped stromatolites, with onlapping grainstones, resting on chloritized Archean basement, STOP 1-1. The stromatolite has a black chert core, and red, jaspilitic outer layers. Photo from Metsaranta and Kurcinka (2022).

varies from 0 to 30 cm in thickness here, and is usually absent from the local topographic "highs" (i.e. knobs or ridges of the Archean basement), but may thicken in depressions in the paleosurface. The basal conglomerate lag may contain large, well-rounded boulders, up to 0.5 m in diameter; these have been observed only on the northwest side of the highway, opposite STOP 1-1. Black, cherty bands (0.1–1.5 cm) occur locally in parts of the Kakabeka conglomerate (ibid). Recent geochronologic study of the conglomerate shows that the main population of detrital zircons is consistent with derivation from local Neoarchean intrusions (R.



Figure 8: Quartz pebble-rich Kakabeka conglomerate, STOP 1-1. Photo from Metsaranta and Kurcinka (2022).

Metsaranta, personal communication, 2026) like the Mackenzie granite (ca. 2672 Ma; Puumala et al., 2015).

The stromatolites which occur in the basal unit of this Gunflint section (Unit 1 of Kup et al., 2025) were described by Kup et al. (2025):

> *"Unit 1 of the Gunflint Formation is characterized by the common occurrence of cherty, domical to columnar stromatolites within the basal 2 m of the formation. Relatively large stromatolite heads tend to occur preferentially on knobs or ridges of the Archean basement, usually metres apart from one another. Smaller stromatolite heads (5–15 cm in diameter or height) are present locally, even on top of the Kakabeka conglomerate. Tabular, finely undulatory, microbialite-like structures (commonly <10 cm thick) may occur in places, or laterally connected to stromatolite heads. The microbialite-like structures may weather to a reddish colour, similar to some of the weathered stromatolite heads. Overall, the colour of the domical to columnar cherty stromatolites ranges from red, yellow, white, grey and black, depending on the iron content and the nature of weathering. Relatively large (usually fresh) stromatolite heads tend to be jet black near their centre and change to lighter grey and white toward their edges; however, edges themselves are commonly marked by red and/or yellow banding. The stromatolites are most easily seen and accessible toward the southern ends of the outcrop, directly above the road ditch level."*

Optical and SEM imaging data collected by Kup et al. (2025) suggest that well-preserved Gunflint-type microfossils (both filamentous and coccoid types) tend to occur in sporadic pockets in samples collected from this locality.

STOP 1-2: Blende Lake Fault (0368005 E / 5383802N)

The northern end of the same rock cut, approximately 500 m north-northeast of STOP 1-1, exposes the east-northeast-striking Blende Lake Fault (cf. Scott, 1990). Rusty fault gouge occurs between Archean gabbroic rocks to the south and folded, flaser-bedded Gunflint wacke and siltstone, cored by a 3 m-wide calcite ± quartz vein / vein breccia which contains Gunflint fragments (Figure 9). Base metal sulphides, including galena, have been noted. The fault is also exposed on the other side of the highway; Gunflint rocks are folded next to the fault there as well.

The east-northeast orientation of the Blende Lake Fault is similar to other structures on and north of the Sibley Peninsula, some of which host gabbroic dykes



Figure 9. Rock cut exposure of the Blende Lake Fault, east side of Highway 11-17 (STOP 1-2), separating folded Gunflint Formation rocks (left) from Neoarchean basement (right). The fault zone is cored by calcite vein / vein breccia; fault gouge flanks the vein. Field notebook for scale.

attributed to the waning stages of Mesoproterozoic Midcontinent Rift magmatism. Scott (1990) noted that Gunflint rocks, normally flat-lying or gently southeast-dipping, are folded and brecciated to a large extent in the area between Blende Lake and O'Connor Point on Lake Superior. Folding described by Moorhouse (1960) east of Blende Lake, was attributed to far-field Penokean fold-and-thrust deformation by Hill and Smyk (2005), prior to the recognition of the Sudbury Impact Layer and associated deformation in the Animikie Basin. Koroscil (2013) noted that thrust faults, once ascribed to Penokean deformation, cut the SIL at the Terry Fox Monument (STOP 1-10) and thus may post-date the Penokean. Landman (2021) noted that:

> *"Later Proterozoic features, including the Blende Lake fault, have a common strike of east-northeast, which aligns with the orientation of the 1.1 Ga Mid-Continent Rift in Thunder Bay. This similarity is further reflected by the Blende Lake fault being oriented subparallel to silver veins related to the Mid-Continent Rift. Similarities between orientations of brittle structures in the [Neoarchean] amphibolite and Gunflint Formation suggest that the Mid-Continent Rift*

in Thunder Bay may have reactivated some Archean-aged, orogenic-related faults and shear fractures. Minor folding in the Gunflint Formation truncated by the Blende Lake fault, as well as reverse reactivation along the plane, may be evidence of compression during the later stage of the Mid-Continent Rift."

STOP 1-3: Mirror Lake Turn-off (0370744 E / 5387262N)

This stop displays a number of quite enigmatic features that are still being evaluated in the context of evolving local geologic ideas.

The hillside exposes iron-rich, folded and brecciated Gunflint Formation sedimentary rocks that have been intruded by a Nipigon diabase sill. Tight to isoclinal, plunging to recumbent folds have developed in certain parts of the otherwise ~flat-lying, thinly to thickly bedded, taconitic, martite(?)-bearing grainstones (Figure 10).

As is the case at STOP 1-2, the cause of the deformation in the Gunflint rocks is a matter of debate. There is growing support that local folding and brecciation may be related to far-field effects generated by the Sudbury meteor impact ca. 1850 Ma., post-dating the end of Gunflint deposition by perhaps ca. 20 My and preceding the onset of Rove sedimentation. Rocks interpreted as being part of the Sudbury Impact Layer were noted in geotechnical drilling ~ 1 km north of this location (P. Fralick, personal communication, 2025). Despite the fact that the Sudbury impact structure is ~660 km away and that the most dramatic deformation is usually crater-proximal (i.e. within ~5 crater radii), Addison and Brumpton (2012) noted that the Thunder Bay area would have still experienced dramatic impact-induced effects, including magnitude 10.7 earthquakes and likely tsunamis. Alternatively, it has also been suggested that some of this deformation may be related to the dominantly extensional stress regime associated with Midcontinent rifting. Local compressional (contractional) structures may form within relay zones between overlapping normal fault tips, particularly as the fault segments grow, interact, and prepare to connect. While normal fault systems are dominated by horizontal extension (pulling apart), the 3D interaction and rotation of blocks in the relay zone (or "relay ramp") can create local stress perturbations that lead to shortening, folding, and antithetic faulting (cf. Camanni et al., 2023). Further work is required to better map the extent and character of folding and brecciation in order to suggest deformation mechanisms and causative factors.



Figure 10. Recumbent fold in Gunflint Formation chert-carbonate rocks, STOP 1-3. Folded bedding planes are traced by dashed lines. A thin veneer of Phanerozoic(?) conglomerate (cgl) occurs on outcrop surfaces and in crevices.

Rove shales and wackes (e.g. STOP 1-5), mapped ~2.5 km south of here by McIlwaine (1975), are not deformed. They overlie Gunflint rocks and are disconformably overlain by sandstones of the Pass Lake Formation of the Sibley Group (STOP 1-6).

Two other enigmatic rocks are exposed at this location; both are conglomerates. One conglomerate occurs as thin coatings plastered on exposed outcrop surfaces and in fractures in the folded Gunflint rocks (Figures 10 and 11). It is a brown, poorly sorted, sandy, matrix-supported unit. Sibley Group (ca. 1.4 Ga) sedimentary rock clasts, ranging from sub-angular to rounded pebbles and cobbles, predominate. Most recognizable are rust-red calcareous siltstones of the Rossport Formation (with their characteristic pale reduction spots) the base of which occurs approximately 75 m stratigraphically above the Gunflint exposed here. It must also be noted that medium-grained mafic igneous clasts appear to be ca. 1.1 Ga Nipigon diabase

Given the presence of Sibley and Nipigon diabase



Figure 11. Close-up of thin veneer of conglomerate on Gunflint substrate, showing reddish-orange Rossport Formation siltstone and Nipigon diabase fragments.

clasts, the conglomerate must postdate at least Midcontinent rifting, the hitherto youngest geologic event in the local lithologic record. Although no Phanerozoic rocks have been documented in the Thunder Bay region, Cretaceous rocks have long been known to overlie the Biwabik Formation (time-correlative with the Gunflint) on the Mesabi iron range of northern Minnesota (e.g. Bergquist, 1944); "soft" iron ores there formed there during the Cretaceous. Paleomagnetic studies by Purucker (1983) in the Eldorado Beach – Nelson roads area, ~6.5 km southwest of this location, suggested that secondary enrichment of Gunflint and Mesabi iron ores took place at approximately the same time between Aptian and Cenomanian time (ca. 125-94 Ma). In their study of anthraxolite in the Gunflint Formation in the Kakabeka Falls area, Hayatsu et al. (1983) identified two very distinct macromolecular materials. These two hydrocarbon fractions were thought to represent derivation from sediments of two vastly different ages: an older one, characterized by heavier aromatic ring compounds, derived from Gunflint-aged organic remains; and another, aliphatic fraction derived from Cretaceous (or possibly Jurassic) sediments. Cretaceous microfossils were described in lateritic "buckshot" ore in the Archean Steep Rock Lake iron deposit near Atikokan (Machado, 1987) that underwent Mesozoic karstification, weathering and/or enrichment (cf. Fralick and Riding, 2015). Interestingly, the exposure of stromatolitic Gunflint rocks on the west side of Highway 11-17, just opposite the Mirror Lake turnoff, only 200 m away from STOP 1-3, is notably rusty and apparently more oxidized than the vast majority of Gunflint rocks. This may represent Mesozoic paleoweathering, raising the possibility that Cretaceous deposits and paleoweathering effects may have once extended as far east as the Thunder Bay area. Any rocks that survived Pleistocene glaciation may survive in isolated patches or have been as yet unrecognized.

The second conglomerate (Figures 12A and 12B) is similarly exposed as a plastered veneer draped on the exposed outcrop face. Unlike the other unit, it appears to be clast-supported and monomictic; angular, dark grey, fine-grained, shaly Gunflint fragments are cemented by calcite. This monomictic clast population suggests local derivation, perhaps a talus deposit created and cemented during the Pleistocene.

Remnants of a Nipigon diabase sill form prominent, cuesta-like hills in the vicinity of this stop and around Deception and Mirror lakes (McIlwaine, 1975). Smooth, glacially polished surfaces with striae are visible at the road level in this cliffside exposure.

STOP 1-4: Gunflint Formation, Blende Creek area (0369581E / 5383837N)

This stop description, featuring deformed chert-carbonate units in the Gunflint Formation (Figure 13), is taken from Fralick et al. (2012):

> *"Along Highway 587, rock cuts display thinly bedded, generally flat-lying sedimentary rocks of the Gunflint Formation. The outcrops we have driven past are composed of ankerite and siderite grainstones (medium-grained, sand-sized iron carbonates) referred to as granular iron formation (GIF). These are common in the Thunder Bay region, dominating the near-shore of the Animikie Basin. The iron carbonate grains were produced by wave erosion of carbonate precipitates and represent storm deposits in the near-shore. The iron may have precipitated as a carbonate in this shore-proximal zone due to photosynthesizing bacteria removing CO2 from the water and thus increasing the pH and driving the carbonate phase into supersaturation. The outcrop we are looking at has these carbonate*

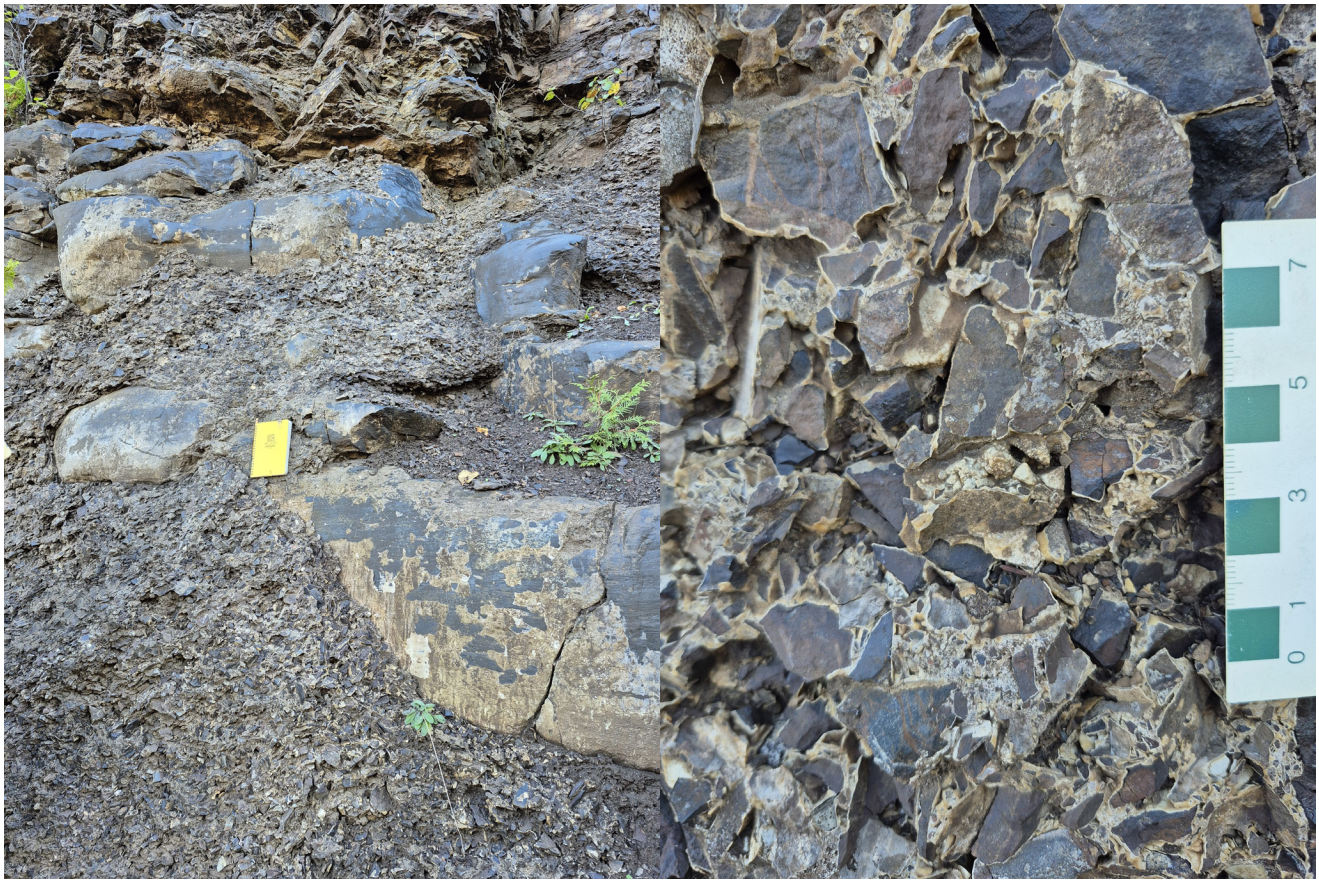


Figure 12. A (left). Clast-supported, calcite- cemented conglomerate veneer over Gunflint, STOP 1-3. B (right). Close-up view of conglomerate / breccia in Figure 12A.

grainstones weathering orangey-brown alternating with white chert layers. In places the chert can be seen replacing the carbonate but other layers appear to be primary chert. In the older literature an outcrop such as this would be ascribed to deeper water due to less evidence of current activity. However, because of its shore proximal location it probably formed in a



Figure 13. Folded Gunflint Formation grainstones, north side of Highway 587, STOP 1-4, with locally axial-planar quartz-carbonate veins.

quieter water location near the strand-line, i.e., a sheltered lagoonal area behind on offshore bar.

These exposures are somewhat unique in that the rocks are folded; elsewhere, they are undeformed. The hinge zones, where the majority of stress is focused, are commonly fractured. These fractures may be occupied by quartz-calcite veins following the vertical axial plane. The outcrop to the west hosts numerous veins and vein breccias that strike between 40° and 45° and dip almost vertically to the southeast. These breccias contain sparry calcite, drusy quartz and also altered shale fragments, suggesting that these Rove Formation rocks likely occurred above this section during vein emplacement. A thin, northwest-dipping diabase dyke intrudes the Gunflint rocks at this location and is, in turn, cut by these veins."

Deformational features in Gunflint Formation rocks near Pass Lake have been previously ascribed to Penokean fold-and-thrust activity in the foreland (i.e. passive margin Archean basement + Gunflint Formation; Hill and Smyk, 2005). These include discrete bedding-plane faults with locally developed gouge and breccia that can be traced laterally into horizontal, hanging wall ramps with associated fault-bend folding. Previous workers had also ascribed folding to syn-sedimentary slumping and Keweenawan diabase sill emplacement and thought that they were attributable to local, rather than regional-scale, deformation. As introduced at STOP 1-3, there is growing support for the contention that such deformation may be related to the Sudbury impact event ca. 1850 Ma.

STOP 1-5: "Devil's Flower Pots" (Rove Formation concretions) 0370841E / 5382426N

This stop description is taken from Fralick et al. (2012):

"Just north of Highway 587, a quarry face exposure of black, fissile Rove Formation shale displays lenticular and elliptical concretions, flattened along bedding planes [Figure 14]. These structures form during diagenesis, following initial compaction and dewatering of the sediments. They represent a concentration of a cementing agent (e.g., silica, calcite) focused during the migration of fluid through the sediments. They often are nucleated around a piece of organic material or other foreign object, which creates a perturbation in fluid flow with a distinct chemistry. Because the cementing agent in this case is more resistant to weathering, these concretions stand out of the soft shale and may commonly completely detach form their host rock. Groundwater and surficial water flow through the shale has led to the dissolution and subsequent precipitation of a variety of low-temperature minerals (e.g. carbonates, sulphates, hydroxides) that occur as white and yellow encrustations on the bedrock surface. One of the more unusual of these secondary minerals is yellow magnesium aluminocopiaptite ($(Mg,Al)(Fe,Al)_4(SO_4)_6(OH)_2.20H_2O$; Resident Geologist's Files, Thunder Bay)."



Figure 14: Flowerpot-shaped Rove Formation concretion on wall of inactive shale quarry at STOP 1-5.

STOP 1-6: Edwards Road section (Pass Lake and Rove formations) 0371737E / 5382460N *(n.b. private property; permission is required to access)*

This stop provides us with an excellent stratigraphic section that extends upward from the top of the Animikie Rove Formation into the Pass Lake Formation, the lowermost formation of the Sibley

Group. The youngest detrital zircons in the Sibley Group are ca. 1.4 Ga (Rogala et al., 2007). An Rb–Sr isochron age of 1339 ± 33 Ma was determined on dolomitic mudstones from the Rossport and Kama Hill formations (Franklin 1978; Franklin et al. 1980). Recent studies of some of the concretions (quartz-carbonate ± various very fine-grained impurities and inclusions) in the Pass Lake Formation, associated with late advanced diagenesis, had enough uranium to generate an age of 1483±4 Ma (W. Bleeker and H. Rochin-Banaga et al., unpublished data / personal communication, 2025; Figure 15). Together with the ca. 1500 Ma youngest detrital zircons (SHRIMP data on 3 samples; ibid), this provides a greatly improved age constraint, just marginally younger than 1500 Ma, on the deposition of the lower part of the Sibley Group.

This stop description is taken from Fralick et al. (2012):

> *"A private access road extending up the mesa provides an excellent 150 m long section exposing the disconformity between the Rove Formation and the overlying basal conglomerate and sandstones of the Pass Lake Formation [Figures 16 and 17]. The Rove shales immediately below the contact were subject to Mesoproterozoic weathering. Geochemical investigations have outlined an oxidized zone below the contact grading to a more reduced zone with abundant chlorite a few tens of centimeters lower in the section. In one area what may be a dewatering or degassing structure strongly deforms the shale. Very immature, iron oxide-rich conglomerates and sandstones of the Loon Lake Member, Pass Lake Formation, overlie the Rove. The conglomerate and sandstone layers are laterally discontinuous, with some conglomerates in clast-support (fluvial deposits) and some in matrix-support (sub-aerial debris-flow deposits). Successions such as this in the Sibley are typical of arid to semi-arid alluvial fans (Cheadle 1986), though this would have been a very small one. The abundant hematite probably denotes a deep water table. Clasts are locally derived from*

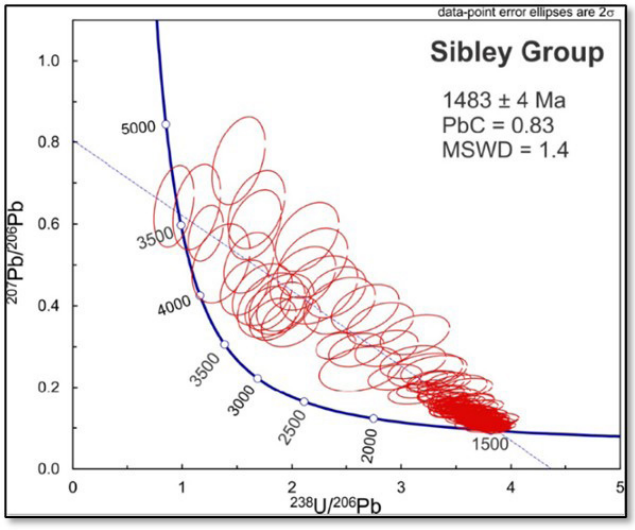


Figure 15. U-Pb concordia diagram presenting geochronological data for the Sibley Group (W. Bleeker and H. Rochin-Banaga et al., unpublished data, 2025)



Figure 16. Cobble-sized clast of Gunflint Formation stromatolitic jasper/chert visible in the Loon Lake Member conglomerate exposed along the Edwards Road section, STOP 1-6.



Figure 17. Disconformable contact (just above hammer) between weathered green Rove Formation shales and hematite-rich, basal conglomerate and sandstone of the 6-7 m thick, Loon Lake Member (Pass Lake Formation), STOP 1-6. Overlying, well-sorted, buff sandstones of the Fork Bay Member form the top of the exposure.

the erosion of underlying units. This is sharply overlain by mature, well-sorted, medium-grained sandstones of the Fork Bay Member, Pass Lake Formation. Detrital zircon geochronology and paleocurrents (Cheadle 1986; Rogala et al. 2007) indicate that the major source of this sediment was the Trans-Hudson highlands. The travel distance accounts for its maturity compared to the locally derived underlying conglomerates. The sandstone was deposited as sheet flows into the shallow nearshore of a lacustrine system that had flooded the area (Cheadle 1986; Rogala 2003; Metsaranta 2006; Rogala et al. 2007). These sandstone layers are laterally continuous, massive to parallel-laminated, in places with trough cross-stratified or rippled tops [Figure 18]. Rare, odd features are present both in cross-sectional and bedding plane views in this outcrop. These may be dewatering pipes."

Fralick et al. (2012) noted that the matrix-supported conglomerate was probably deposited as a high-density mass-flow while the boulder-cobble, matrix-supported conglomerate probably represents a very high-viscosity mass flow as the larger clasts were suspended near the top of the flow. Upper flow regime, parallel-laminated sandstones were probably deposited by sheet-floods on an alluvial fan's surface and are interbedded with clast-supported fluvial conglomerate.

The top of the hill affords a tremendous view of Thunder Bay, Sibley Peninsula and offshore islands. A south-dipping diabase sill forms the prominent cuesta of Caribou Island. Glacial erosional features are exposed on this outcrop surface. These include two sets of glacial striae at 040˚ and ~060˚ and subparallel arrays of crescentic gouges and chatter marks/lunate fractures (Figure 19).



Figure 18. Medium- to coarse-grained, well-sorted sandstone bed of the Fork Bay Member, top of hill, STOP 1-6. The majority of the bed is upper flow regime parallel laminated, with a reworked, cross-stratified top.

En route to STOP 1-7, the highway traverses a series of baymouth bars that formed as lake levels fell from the Lake Beaver Bay stage (ca. 11 to 10.5 Ka) to the Lake Minong stage (ca. 10.5 to 8.5 Ka), connecting what had been the "island of Sibley" to the mainland near Pass Lake (Zaniewski et al., 2020; Geddes et al., 1987; *see* Fralick et al., 2012). This new connection formed an ideal natural trap for Palaeo-Indian hunters to use. The materials excavated at the Brohm archaeological site, on the top of the main baymouth bar, were all hunting-related projectiles and scrapers, many made onsite from chunks of jasper taconite that they carried with them from quarry sources (Zaniewski et al., 2020; MacNeish 1952).



Figure 19. Glacial erosional features exposed in the sandstone outcrop surface at the top of the hill, STOP 1-6. These include glacial striae (dashed arrows), concave up-ice crescentic gouges (CG) and concave down-ice chatter marks (CM).

STOP 1-7: Pass Lake section (Loon Member conglomerate) 372282E / 5380560N

The cliffs adjacent to the abandoned railway at Pass Lake is the type section for the Pass Lake Formation. Exposure is almost continuous for 3.2 km along the tracks and provides a ~50 m-thick stratigraphic section. Rove Formation shales, exposed at the northwestern end of the cliff exposure, disconformably underlie the Pass Lake Formation but are not exposed here. This cliff face, a popular destination for local rock climbers, exposes the basal Loon Member conglomerate and overlying, buff sandstones of the Fork Bay Member (Pass Lake Formation; Figure 20). A description was provided by Fralick et al. (2012):

> *"The basal conglomerate thins and thickens laterally, pinching down to pebbly sandstone in places. Clasts are generally surrounded and dominated by local Gunflint Formation lithologies. The matrix is poorly sorted. The conglomerates are overlain by a thinning-upward sequence of sandstone beds capped by siltstones on the top of the cliff. Individual beds are reasonably laterally continuous though sometimes lens out. They are dominated by upper flow regime parallel lamination with occasional ripples and small-scale dunes on their tops.*
>
> *Both alluvial fan-braided fluvial and shallow lacustrine (Cheadle, 1986; Franklin et al., 1980, respectively) depositional environments have been proposed. The bedding organization of the conglomerates exposed here is somewhat different than those observed earlier. This opens the possibility that the conglomerates at this location were reworked by wave activity during initial lacustrine flooding.*
>
> *The sandstone beds again represent sheet-floods, forming sand-flats in the shallow lake. The thinning- and fining-upward sequence of sandstone beds is a classic example of a transgressive succession showing decreased sand supply through time as the shoreline moves further away from the area."*

Overlying red-orange Rossport Formation siltstones begin to outcrop approximately 1.2 km east of STOP 1-7 (see Fralick et al., 2012 for stop descriptions).



Figure 20. Pass Lake section exposure of Loon Lake Member conglomerate overlain by Fork Bay Member sandstone at STOP 1-7.

STOP 1-8: Neoarchean Pyroclastic and Clastic Sedimentary Rocks 360568E / 5379943

This 100 m-long rock cut on the northwest side of Highway 11-17 was exposed by highway construction ca. 2012. It features a remarkable exposure of Neoarchean pyroclastic and clastic sedimentary rocks of the Shebandowan greenstone belt that strike ~140˚ and dip steeply northeast (Figure 21). These supracrustal rocks are intruded by granitoid rocks of the McKenzie granite and may represent a large pendant within the intrusion. The exposure was the subject of an undergraduate thesis by Bjorkman (2014), from which most of the descriptions will be gleaned.

Bjorkman (2014) identified 13 lithofacies/lithologic units in this complex section:

This exposure exemplifies the close connection between Neoarchean pyroclastic activity and sedimentation (Figure 22). Bjorkman (2014) suggested that these rocks were deposited in a vent-proximal environment, a contention supported by the presence of graded ash beds, high-velocity base surge deposits and impact structures from pyroclastic bombs (Figure 23).

Lithofacies / Interpretation	Description
1 Lahar – debris flow	Unsorted, clast-supported, conglomerate; clasts are pebble-sized to small boulder-sized within a medium-grained, sand-sized matrix
2 Ash fall	Fine-grained to medium-grained, sandy, continuous beds; parallel laminated with no cross-bedding, with an average thickness of 1 cm or less
3 Fluvial reworking and base surge deposits	Cross-bedded and graded, medium-grained sandstone beds, which occur in alternating sequences; this unit has a sharp basal contact with the lower Lithofacies 2 and a poorly defined contact with upper units. The transitions from the graded beds to the cross-stratified beds are distinct.
4 Ash fall	Parallel-laminated, continuous layers of medium-grained, graded beds, more rarely observed to be cross-bedded at very low angles.
5 Lag deposit	Clast-supported conglomerate in which clasts are very uniform and commonly cobble-sized. The clasts are flattened and oval-ellipsoidal, with rounded edges. The long axes of the clasts occasionally have tail-like tips.
6 Slump	Disturbed beds of material very similar to that found within Lithofacies 4. The parallel-laminated strata are disrupted by failure of the slope and are truncated by an angular disconformity of the overlying unit. There are rubble blocks adjacent to the truncated strata. These blocks of failed beds lie along the base of this facies, with no evidence of sorting after the failure. There is no grading in the matrix, which is massive, medium-grained sandstone.
7 Lahar	Repetitive sequences of normally graded, medium-grained sandstone beds, gravelly matrix supported beds, and non-graded massive beds composed of medium-grained sand. Discontinuous, lens-shaped beds are very common. The normal graded beds are composed of coarse-grained sandstone, which grades into fine-grained tops of beds. These often have eroded, scoured tops, with very distinctly defined bases. Cross-bedding is common.
8 Lahar / channelized debris flow	Massive graded conglomerate with angular to very rounded and moderately flattened clasts. The clasts appear monomictic and range in size between 5-15 cm, the majority being 12 cm by 7 cm. The clasts make up 70% of the total composition, while the matrix is mostly a uniform medium-grained sandy composition, with 10% coarser sand-sized fragments. The unit is on average 3-5 m wide.
9 Ash tuff	Fine-grained, sand-sized matrix with medium-grained and subhedral porphyritic feldspar crystals. The weathering surface is very irregular as the feldspars stand-out from the matrix. The unit occurs sporadically, locally intruded by dark green material. The average thickness varies between 50 cm to 100 cm. There is no grading throughout this unit, and the unit conforms to the same stratigraphy as the surrounding units, which is most often Lithofacies Association 1.
10 Hornblendite	Medium-grained dyke which crosscuts stratigraphy. The largest of the intrusive dikes, it can be traced through the entire outcrop. It is distinguished by the very irregular shape of its contacts with the host rock. The matrix consists of green equigranular, subhedral crystals in the middle of the intrusion and lighter altered plagioclase crystals along finer-grained contacts. The body intrudes (brecciates) itself where the dike dilates. Other smaller dikes crosscut this one. Cobble- to boulder-sized xenoliths were noted.
11 Hornblendite	Medium-grained, green-grey mafic dyke, 5-7 cm wide, with equal amounts of mafic and felsic minerals. The dike cuts through the green intrusive veinlets.
12 Hornblendite	A set of dark green-grey dykes, up to 0.5 m in width, striking approximately the same direction as Lithofacies 10; may be sill-like intrusions, wispy and infiltrating intrusions which engulf clastic material. This unit commonly contains wall rock xenoliths, which are very sharp and angular.
13 Syenite	Medium- to coarse-grained, subvertical and east-southeast-striking dykes. The dykes are the youngest rock type in the outcrop and are noted regionally. They are composed of red feldspar, amphibole, biotite, and quartz. The red feldspar gives the rock a brick-red colour.

A combination of subaerial and shallow subaqueous conditions likely existed at the time of deposition, with fluvial reworking and deposition occurring during periods of volcanic dormancy. Phreatomagmatic processes, similar to those that produce maar craters, likely predominated.

The calc-alkalic geochemistry (Figure 24), pyroclastic volcanism and subaerial/shallow water to fluvial clastic sedimentary rocks suggest that these rocks are part of the younger Shebandowan assemblage

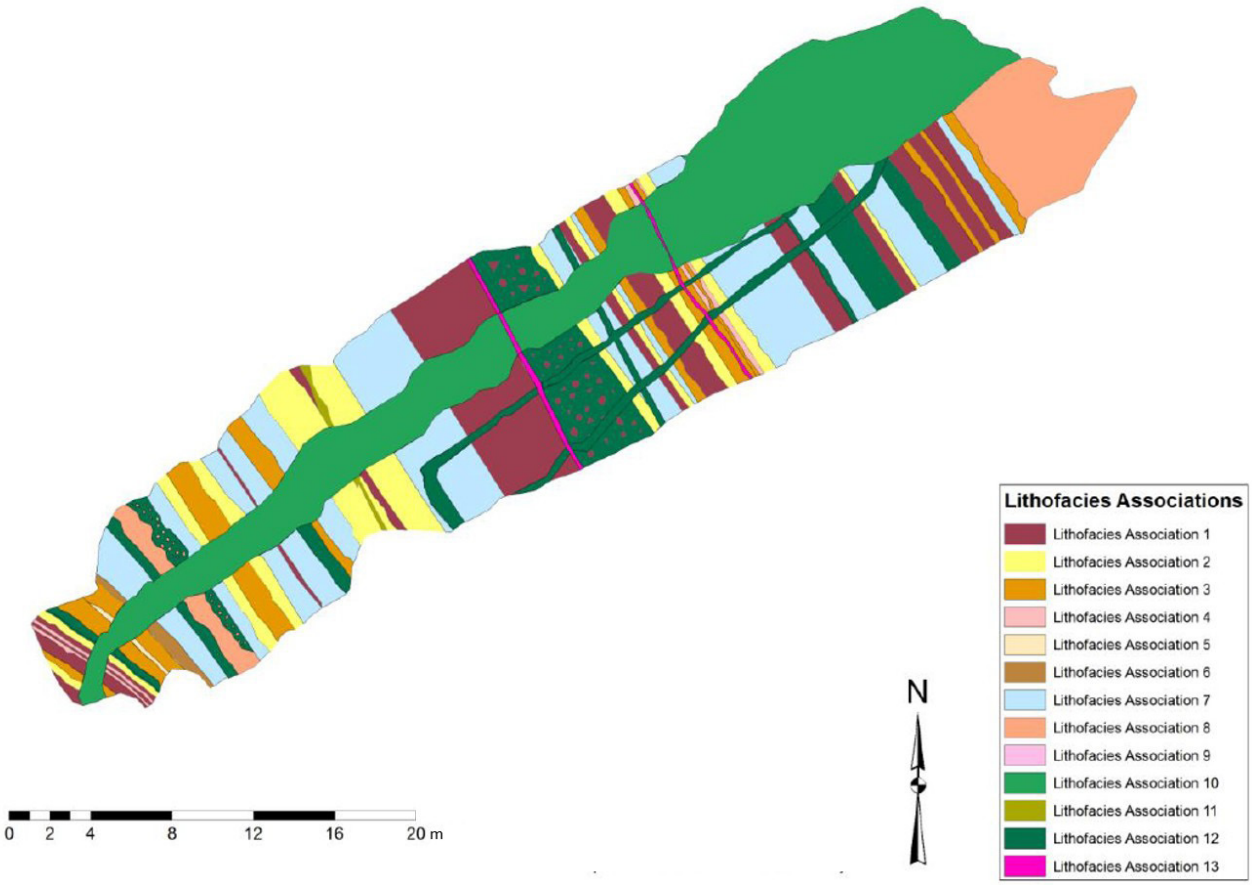


Figure 21. Map of the main outcrop, STOP 1-8 (Bjorkman, 2014). Lithofacies Association 1 through 9 are (resedimented) pyroclastic and clastic sedimentary units; lithofacies association 10 through 14 are crosscutting dykes.

of the Wawa subprovince. Shebandowan rocks (ca. 2690-2680 Ma), unconformably overlying the older (ca. 2720 Ma) Greenwater assemblage rocks, were deposited in fault-bounded, pull-apart basins during regional transpressive (D2) deformation.

North and south of the main, supracrustal-dominated outcrop, pink granitoid rocks associated with the McKenzie granite occur (Figure 25). The McKenzie granite is approximately 22 km long (east-west) by 3.2 km wide (north-south) and has been divided into two segments that are separated by a fault (Scott, 1990). Based on the mapping of Scott (1990), and the interpretation of aeromagnetic data, Metsaranta (2015) suggested that the McKenzie granite comprises multiple distinct intrusive bodies, and referred to the western segment as the Mount Baldy intrusion.

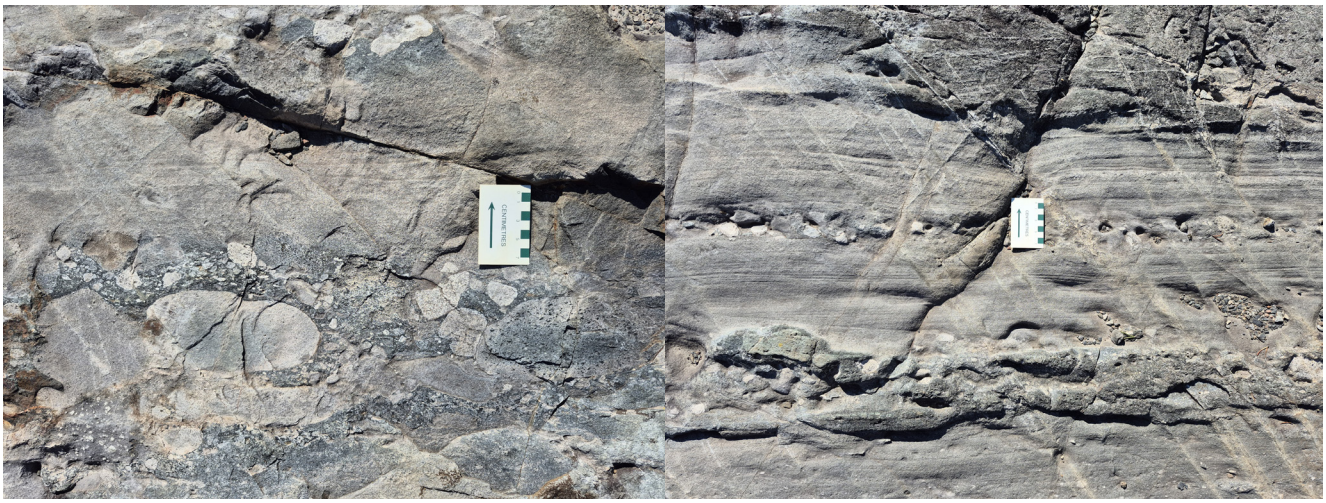


Figure 22. Resedimented pyroclastic material as conglomeratic beds and lenses, STOP 1-8.

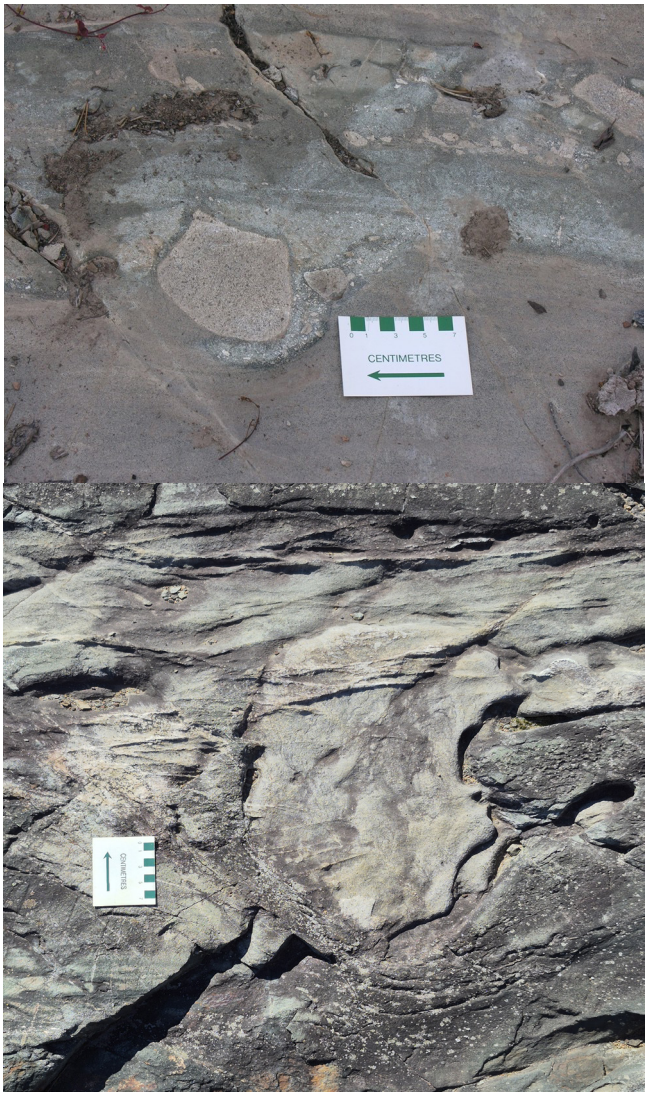


Figure 23: Large pyroclasts, commonly with attendant bomb sags, STOP 1-8.

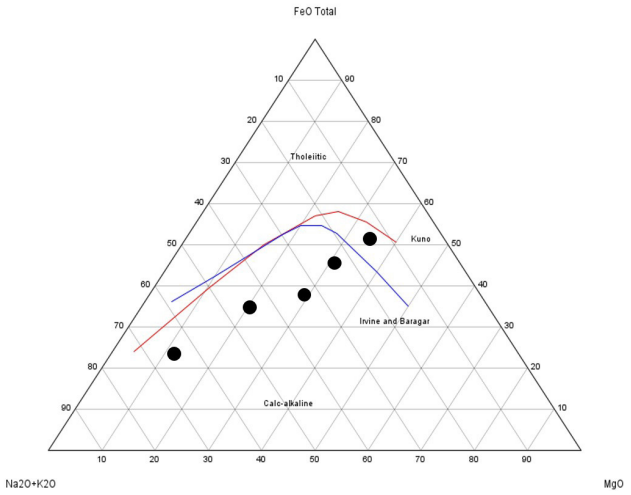


Figure 24: AFM plot of samples collected and analyzed by Bjorkman (2014), showing the calc-alkaline nature of the resedimented pyroclastic rocks at STOP 1-8.



Figure 25. Photo illustrating cross-cutting relationships at STOP 1-8. The granitoid dyke in the bottom half of the photo that cross-cuts all lithologies, including a hornblendite dyke (top center), is associated with the nearby McKenzie granite.

The Neoarchean, S-type McKenzie granite (Hughes, 2016) is primarily a peraluminous quartz monzonite, with mineral assemblages characterized by microcline-plagioclase-quartz-muscovite-biotite with minor amounts of inequigranular hornblende, chlorite, titanite and rarely calcite. The McKenzie granite exhibits a peraluminous geochemistry, with SiO_2 contents ranging from 63.8 to 68.2 weight % along with enrichment in light rare earth elements and fractionated heavy rare earth elements, decreasing trends of major oxides, transition metals and high field strength elements. Scattering of the large ion lithophile elements on discrimination diagrams is likely due to remobilization during chlorite, sericite and carbonate alteration (Hughes et al., 2017). It is proposed to have formed in a similar way to the model proposed for the later stages of the genesis of the nearby Dog Lake Granite Chain, which involved partial melting of a mantle wedge beneath the Wawa-Abitibi island arc. The proposed late-stage emplacement model is consistent with recent U-Pb geochronology (Puumala et al., 2015) that indicated that the McKenzie granite was emplaced at 2672.6 ± 1.5 Ma (zircon, U/Pb thermal ionization mass spectrometry). These S-type melts, formed from the partial melting of metasedimentary rocks, may have interacted with I-type melts, allowing for the variations in geochemical and petrological data that are observed in the McKenzie granite, such as the presence of hornblende, that are not common for standard S-type granites.

STOP 1-9: Gunflint / Archean Unconformity, Crystal Beach 358661E / 5378895N

This road cut on the northwest side of Highway 11-17 near Crystal Beach provides another outstanding exposure of the contact between Archean basement and unconformably overlying Paleoproterozoic Gunflint Formation, similar to that exposed at STOP 1-1. However, there are a number of unique features here that warrant description and examination.

The basement at this location is the Neoarchean McKenzie granite (2672 Ma) which has been conspicuously altered 1 to 3 m below the unconformity. The original pink granite has been altered to dark green chlorite (+ clays?) up to 2 m; alteration intensity increases upward towards the unconformity. Unaltered pink, K-spar-phyric granite gives way to altered versions in which the matrix is incipiently to completely chloritized, leaving relict, unaltered K-spar phenocrysts. The phenocrysts have also been replaced (saussurite + clays + chlorite) in the most intensely altered granite, leaving only relict quartz (Figure 26).

The correlation between alteration intensity and proximity to the unconformity suggests that the altered rocks may represent a regolith/saprolite. Such alteration is often interpreted as a combination of ancient subaerial weathering (true paleosols) and later fluid migration from the overlying iron formation. Geochemical studies by Yip (2016) and Fralick (personal communication, 2024) at this location suggest that iron-rich Gunflint fluids replaced and masked the geochemical signature of the original paleosol. Similar alteration characteristics were described by Kronberg and Fralick (1992), who noted that alteration of ferromagnesian minerals in felsic Archean rocks southwest of Thunder Bay was apparently due to diffusion of iron-rich, Gunflint-derived fluids across the Proterozoic -Archean unconformity, consistent with slow mineral-fluid exchanges under diagenetic or low-grade metamorphic conditions. Chemical changes in mafic minerals include additions of iron, manganese, and water and losses of silica, calcium, and magnesium. They concluded that these chemical changes occurred as Gunflint fluids diffused into underlying rock over a time frame of 10^5-10^7 years.

Spalling of overlying Gunflint rocks has exposed a section of smooth, bare basement paleosurface (Figure 27). The contention of Pre-Gunflint weathering is supported by the occurrence of boulder-sized, spheroidally weathered, altered granitic corestones on the paleosurface, where they are enveloped by Kakabeka Member (basal) conglomerate and saprolite/regolith, and are draped by Gunflint grainstones (Figure 28). Kakabeka conglomerate infills depressions in the paleosurface and fractures that extend down into weathered basement. The conglomerate here consists largely of resistate quartz pebbles in a chloritic,



Figure 26. Selected hand samples of McKenzie granite from STOP 1-9, showing progressive alteration (chloritization) from unaltered (left) through incipient and pervasive matrix replacement (second and third from left, respectively) to complete replacement of matrix and K-spar phenocrysts (far right).



Figure 27: Smooth, curved, bare Archean basement paleosurface (accentuated in half-shadow above yellow field notebook), exposed below overlying, draped Gunflint grainstones, STOP 1-9.

saprolitic/regolith matrix. Although this is perhaps the first documented example of corestones in the Gunflint or Biwabik formations, Paleoproterozoic (ca. 1.85 Ga) weathering-produced corestones in the Flin Flon-Creighton area of Manitoba and Saskatchewan were documented by Sindol et al. (2020).

Many of the joint surfaces and fractures in this exposure are covered and infilled by vein minerals, mainly quartz/amethyst, barite, fluorite, calcite with rare base metal sulphides (pyrite, chalcopyrite, galena, acanthite). These veins constitute the 7Z amethyst occurrence (Figure 29), first explored ca. 1890 (Ontario Mineral Inventory, https://www.geologyontario.mines.gov.on.ca/mineral-inventory/MDI52A10SW00007).

The following description of the occurrence is excerpted from Puumala et al. (2015).

> *The 7Z amethyst occurrence is hosted in a vein system and/or breccia zone that strikes 050° and is located approximately at the unconformity between sedimentary rocks of the Paleoproterozoic Gunflint Formation and Neoarchean intrusive rocks of the McKenzie granite stock. The amethyst-bearing vein system has been exposed in a series of 3 historic trenches over a strike length of 180 m.*
>
> *The portion of the vein system exposed in the southwestern and central trenches is hosted by rocks of the Gunflint Formation, while the veins exposed in the northeastern trench occur at the unconformity and are hosted by both Gunflint and granitic rocks. Gunflint rocks are strongly silicified adjacent to the veins. The exposed width of the vein system is approximately 10 m.*
>
> *The majority of the amethyst-bearing veins strike 050° (i.e., parallel to the breccia zone) with near-vertical dips. The vein widths are variable, ranging from centimetre- to metre-scale. In the Gunflint Formation rocks, a near-horizontal set of narrow veins also occurs along bedding plane fractures. A third set of narrow, approximately north-striking veins, was also observed immediately to the south of the main breccia zone in road cuts along the north side of Highway 11-17.*
>
> *The amethystine quartz in this vein system shows a wide variation in colour, ranging from light pink (i.e., rose quartz) through to deep purple. Colourless to white quartz and smoky quartz are also abundant. Veins hosted by granite tend to contain lighter coloured amethyst, while deep purple amethyst and smoky quartz are most likely to be found in the southwestern trench, which is hosted by Gunflint Formation rocks. Most amethyst crystal points are on the order of 1 cm wide. However, much larger crystals were observed in some vugs. Crystals hosted in the Gunflint Formation rocks commonly have a surface coating of hematite.*

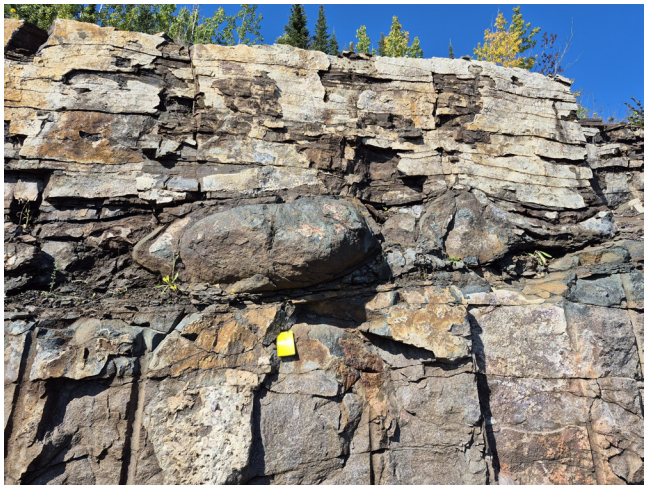


Figure 28. Spheroidally weathered, chloritized Neoarchean granitic corestone boulders resting on the paleosurface at the Paleoproterozoic-Archean unconformity, STOP 1-9. The corestones are enveloped by saprolitic sediments and conglomerate/regolith, and overlain by draping Gunflint grainstones.

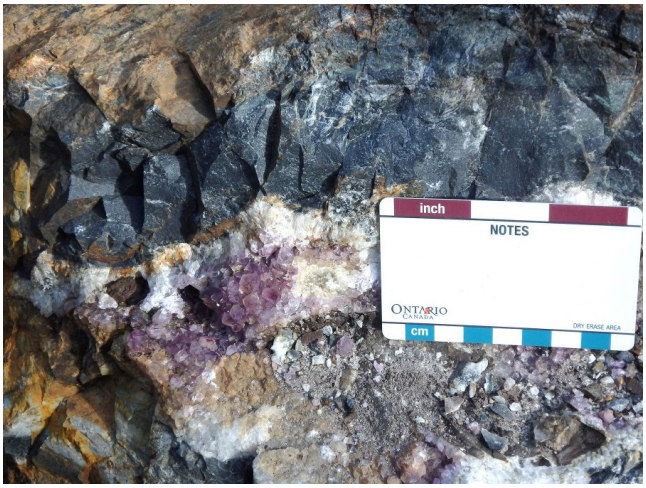


Figure 29. Amethyst-bearing vein hosted in the Gunflint Formation at the 7Z occurrence.

Although recent sampling has reported no significant silver values, a local newspaper reported in 1890 that 7Z was “a veritable mountain of amethyst with rich surface signs of silver” (ibid). This vein system is an

example of a broad group of silver-bearing, carbonate-quartz veins that typically occur in Animikie Group sedimentary rocks, often in close association with the Archean-Proterozoic unconformity and Midcontinent Rift-related diabase sills (Oja, 1967; Franklin et al., 1986; Kissin, 1992). They likely formed from metamorphically generated fluid from in the Midcontinent Rift and expelled along rift-bounding faults (Smyk and Frankin, 2007).

STOP 1-10: Terry Fox National Historic Monument 339836E / 5372406N

This stop includes opportunities to view outcrop exposures near the Terry Fox National Historic Monument and lookout that **c**ommemorates Terry Fox's 143-day, 5373-km Marathon of Hope run to raise money for cancer research in 1980, which continues to inspire global fundraising efforts.

A number of rock types and features are exposed in the road cuts that flank the highway and access road near the monument (Figure 30). A prominent, columnar-jointed Nipigon diabase sill (Terry Fox sill; Magnus, 2012; Magnus and Kissin, 2010) intrudes and caps these Rove and Gunflint formation sedimentary rocks. As a result, this site displays a complete stratigraphic section from the Gunflint Formation, through the Sudbury Impact Layer (SIL) and up into the overlying Rove Formation. Disconformities appear at both the base and top of the SIL (Addison and Brumpton, 2012).

The description of the SIL (Figures 31 and 32) at this location by Addison and Brumpton (2012) was largely based on a former outcrop exposure that was removed during highway reconstruction in 2011.

"This is the only outcrop showing a complete ~ 3 m cross-section of the ejecta-bearing debrisite layer extending from Gunflint chert-carbonate up into the basal Rove Formation, which is overlain



Figure 30. This quarried rock face adjacent to the Terry Fox Lookout road displays a cross-section that includes (bottom left to top right) the Gunflint Formation, Sudbury Impact Layer (SIL), Rove Formation and Nipigon diabase. A Midcontinent Rift-related normal fault exhibiting approximately 4 to 5 metres of vertical displacement is visible near the left margin of the quarry face and is highlighted with a dashed line.



Figure 31. Rocks of the Sudbury Impact Layer (grey) and Rove formation (black) are visible in this photo from STOP 1-10. Geologist's hand is located at the top of the Sudbury Impact Layer.

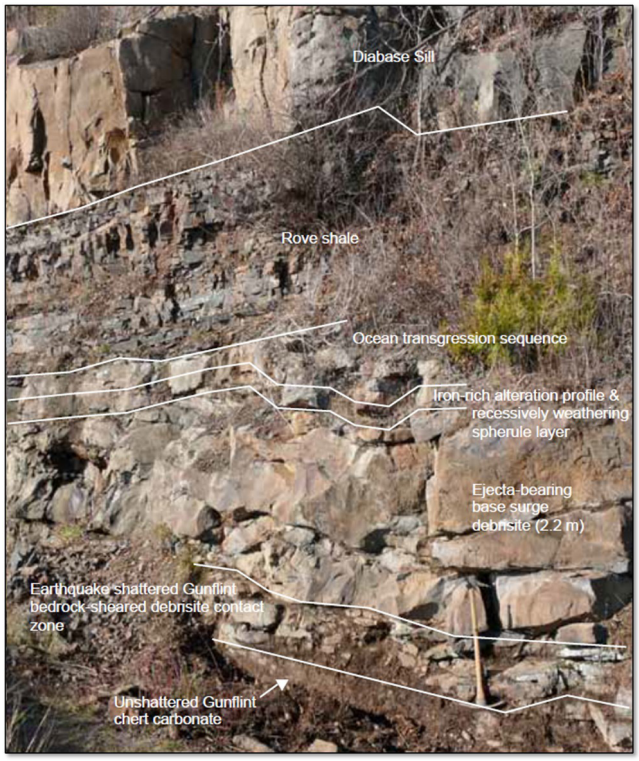


Figure 32. The weathered outcrop (now gone) at STOP 1-10 in 2010 (Addison and Brumpton, 2012). Carbonate-replaced devitrified vesicular impact glass shapes and tektites were then visible on the weathered surface. The Ocean Transgression Sequence is composed of ankerite grainstones identical to those of the Gunflint and probably represents a limited transgression millions of years prior to the deposition of the Rove Formation (P. Fralick, personal communication, 2026).

in turn by a diabase sill. An iron-rich alteration profile, heavily replaced by secondary pyrite, lies ~ 1 m below the base of the debrisite and a few metres northeast of the main outcrop.

The basal SIL is a recessively weathering, locally sheared, clastic layer about 0.5 m thick containing crushed spherule clusters, some of which are aligned subvertically instead of in the usual subhorizontal position. Several sets of subhorizontal slickensides, whose striae are aligned at a 140° azimuth, are found at various levels within this layer. Postdepositional anastomosing chert has replaced much of this basal sheared layer, obliterating considerable structural detail. Non-ejecta features include centimetre to millimetre-sized angular chert clasts and angular, subrounded to round Gunflint Formation iron carbonate clasts plus two rounded crystalline rocks with prominent alteration rinds. The presence of clasts with weathering rinds reinforces the idea that Gunflint clasts lacking such rinds were freshly fractured by impact-generated earthquakes before being incorporated into the debrisite.

The main body of the 2.2 m thick debrisite lies in sharp contact over the basal sheared clastic unit. It is so heavily replaced by recrystallized dolomite that any possible ejecta features are only seen as vaguely outlined shapes on weathered surfaces or in thin section. Almost all detail, including any vesicles in possible DVIG- [devitrified vesicular impact glass] shaped clasts, has been destroyed. Tektites and microtektites may be present, based upon shape and rare faint devitrification textures. A single, polycrystalline, rounded quartz grain shows faint planar features. Both angular and rounded millimetre-scale chert clasts are also present, but not common.

A 5-20 cm thick undulating, dark brown, recessively weathering, spherule-cluster-rich layer appears as a groove across the cliff face at the top of the dolomite-replaced debrisite. This mass of spherule clusters is much more concentrated than seen at any other location or than is suggested by faint shapes in the main dolomite-replaced layer immediately beneath it. These concentrated clusters seem to be the residuum of a thicker layer. Plentiful, thin anastomosing post-depositional chert strands weave through this spherule-rich material but on a much finer scale than at Hillcrest Park.

Red-brown agate 3-8 cm thick lies on top of the spherule-rich layer. Laterally discontinuous vertical digitate projections extend down from the top and project up from the base of this agate layer. They are similar in shape and size to the agate stalactites in vugs at Hillcrest Park, except that in this case the spaces between the projections were subsequently infilled by more agate. The red-brown colour is similar to that of the iron-rich alteration profile overlying it but it is a less saturated hue.

An iron-rich alteration profile on top of the spherule-rich layer, consisting of hematite has been largely replaced by secondary pyrite. Prominent deformed spherule clusters are locally present. The total thickness of all these ejecta-bearing layers is 3 m.

The top of this iron-rich layer marks a return to carbonate deposition. The basal 10-15 cm of this 80-100 cm thick carbonate zone is unstratified and shows dark, angular, commonly rectangular, millimetre-centimetre-sized rip-up mudstone clasts and probable Gunflint Formations clasts. This is followed by millimetre- to centimetre-scale layered carbonate strata topped by a zone with a few poorly defined, laterally discontinuous beds containing centimetre-scale, angular carbonate clasts.

The carbonate then makes an abrupt transition to 10-15 cm of gray siltstone and is overtopped by 10-15 cm of black, rusty weathering shale characteristic of the Rove Formation. The black shale is interrupted by 5 cm of chert before returning to 0.9-1.2 m of black, rusty weathering shale which is overlain in turn by a diabase sill more than 8 m thick. The shale is less friable than typical lower Rove shale, probably the result of low-grade metamorphism induced by the overlying sill."

A 050˚-060˚-striking normal fault, perhaps related to Midcontinent Rift-related extension, has displaced Animikie rocks and diabase 4 to 5 m. Koroscil (2013) identified thrust faults, mainly expressed as small discrete bedding plane faults with few kinematic indicators or piercing points to quantify the displacement. Thrust faults within the SIL were

identified by slickenlines or slickenfibres on fault plane surfaces. The faults can be traced along strike until they are either covered by overburden or cut by a prominent normal fault which displaces all units in the hanging wall down to the south several metres (ibid). Reid-Sharp (2016) described faults, related damage zones and calcite vein breccias along the highway ~2 km northeast of STOP 1-10. Normal faults that transect Gunflint Formation ± Archean basement rocks, strike east-northeast and dip to the southeast, were also ascribed to extension during Midcontinent rifting.

The past-producing Thunder Bay Silver Mine is situated between the highway and the Terry Fox access road. Discovered in 1866 by P. McKellar, it was mined underground until 1874 via four shallow (8 to 21 m deep) shafts (Ontario Mineral Inventory, https://www.geologyontario.mines.gov.on.ca/mineral-inventory/MDI52A06NE00005).

Mineralization occurs in calcite-quartz veins that are hosted in chert-carbonates (Gunflint Formation) and shales (Rove Formation). The host rocks strike 034/22 southeast in the vicinity of the vein but subhorizontally 30.5 m to the northwest beneath a diabase sill 12.2 m thick. A 3 m wide composite vein or stockwork system consisting of 2.5 cm wide quartz-carbonate veinlets lies within and parallel to a fault that also strikes 034/65 northwest. Ore was mined locally over the total length of 182.9 m. Native silver and acanthite occurred in pockets 7.6 - 45.7 cm thick by 1.8 -12.2 m in length, the silver being in leaves and grains irregularly distributed in a gangue of quartz, with some calcite, galena, sphalerite, and pyrite. A second vein of calcite occurs in a parallel fault 6.1 m southeast of the composite vein (ibid; Sergiades, 1968). When first opened, two orebodies were found, one next to the north or hanging-wall and one in the middle (Tanton, 1931). The ore was brought to a stamp mill at the mouth of the Current River, 4 km south of the mine (Figure 33). Production totaled an estimated $20 000 (Bowen, 1911), or approximately 15 000 ounces of silver.



Figure 33. Stamp mill of the Thunder Bay Silver Mine at the mouth of the Current River, ca. 1880.

STOP 1-11: Sudbury Impact Layer, Markland and Hill Streets 334163E / 5366301N (*n.b. Private Property, ask for permission to access. Be very careful not to step on any plants. No hammers are allowed.)*

Another spectacular debrisite breccia of the Sudbury Impact Layer is exposed at the corner of Markland and Hill streets. This outcrop was mapped in detail by Shegelski (1982; Figure 34) and later described in the context of impact-related brecciation by Addison et al. (2010) and Addison and Brumpton (2012, Figure 35):

> *"A bedrock exposure, about 5 m by 15 m, in a private yard in Thunder Bay contains a spectacular debrisite exposure composed mainly of Gunflint chert-carbonate breccia and ejecta, primarily DVIG [devitrified vesicular impact glass], which is surrounded and partially replaced by blocky calcite cement. The debrisite remnant preserved here is 0-0.5 m thick and unconformably overlies stromatolites and chloritic grainstone of the uppermost Gunflint Formation. An iron-rich alteration zone exists approximately 30 cm below the erosive contact between the debrisite and the Gunflint bedrock.*
>
> *DVIG clasts are up to 2 cm across. Vesicles range from round to ovoid to nearly flat. Angular quartz and feldspar grains, chert shards, and chloritic granules are also present. Quartz grains with PDFs have not been found here."*

The SIL is also exposed nearby at Hillcrest Park and along Banning Street.

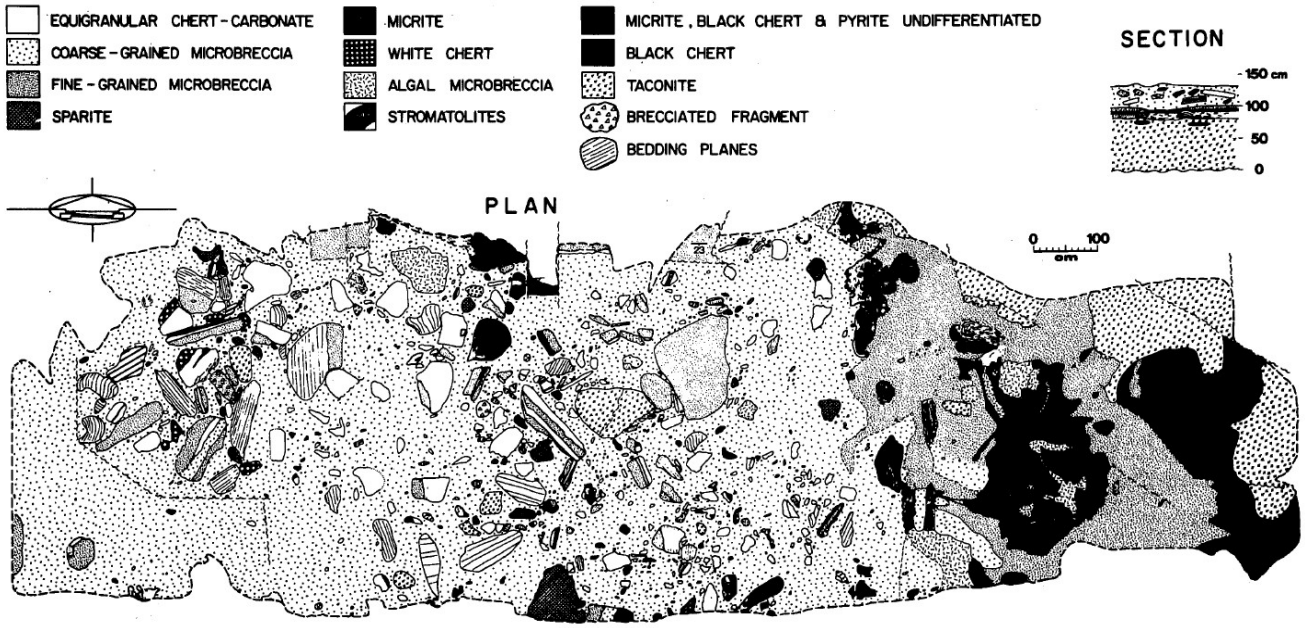


Figure 34. Detailed map of the debrisite breccia outcrop at STOP 1-11 by Shegelski (1982)

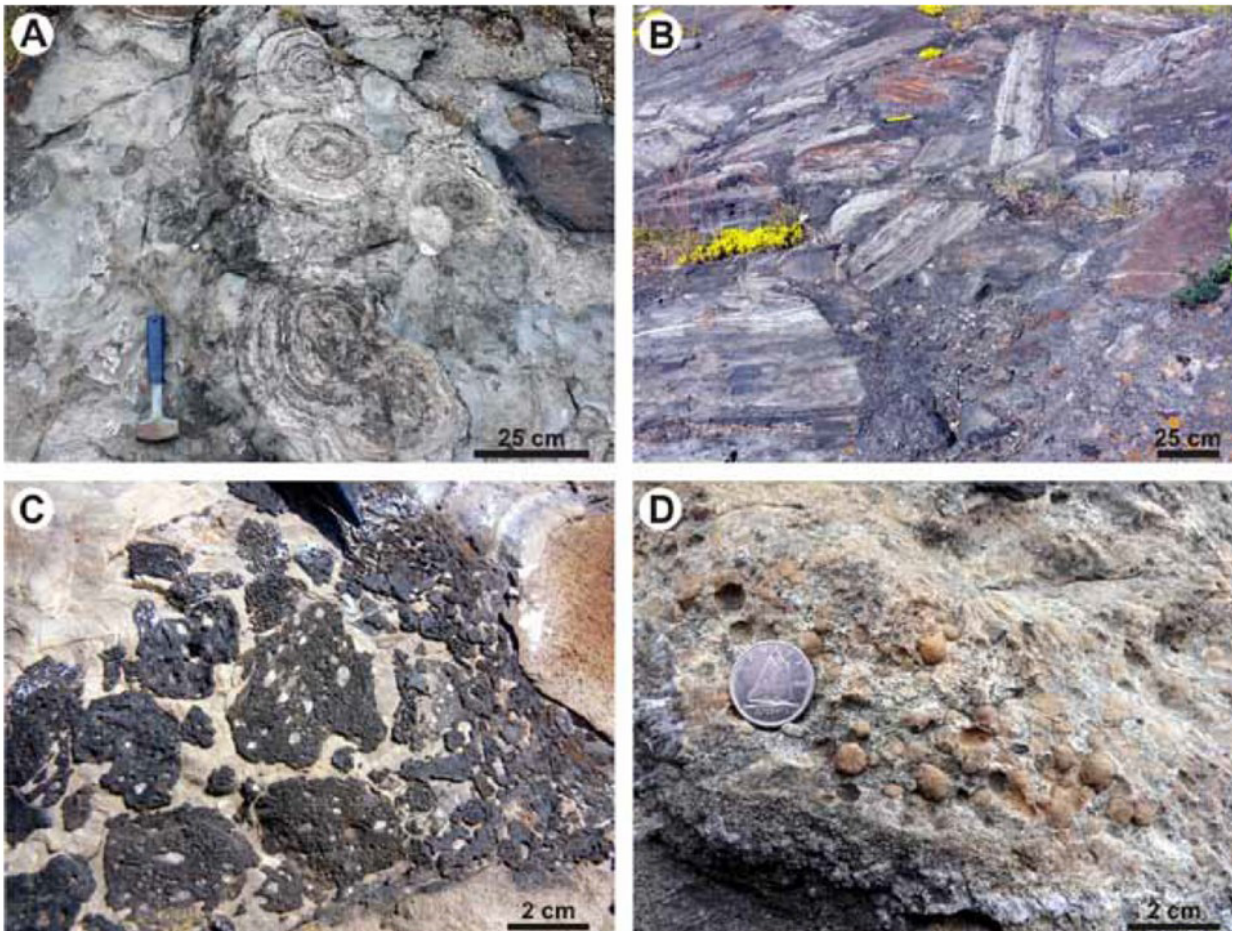


Figure 35. (from Addison and Brumpton, 2012) A – Gunflint Formation stromatolites exposed on a glacially truncated surface, STOP 1-11. While it is not recognizable in the photo, debrisite lies over stromatolites at upper right of the photo. B – Angular to slightly subangular clast-supported Gunflint Formation breccia with a finer DVIG-rich and calcite-rich matrix, all of which lies directly on Gunflint stromatolites, STOP 1-11. The angular clasts suggest a short travel distance from their point of origin. C – DVIG clasts within a recrystallized calcite matrix, STOP 1-11. The silicate devitrification product supports growth of a black lichen, whereas calcite prevents lichen growth. The vesicles are calcite infilled. D – Orange, weathered accretionary lapilli in a recrystallized carbonate matrix, Hillcrest Park.

Field Trip Stop Descriptions - Day Two

Day Two comprises stops to a variety of locations south and west of Thunder Bay (Figure 36). This area is underlain by a variety of rocks that record almost three billion years of local geologic history, spanning from the Neoarchean (ca. 2.7 Ga) to the Paleoproterozoic (ca. 1.8 Ga) and Mesoproterozoic (ca. 1.4 and 1.1 Ga). Unconsolidated glacial and post-glacial deposits and features attest to a long-lived, Pleistocene glaciation record. All GPS coordinates are NAD83, UTM Zone 16.

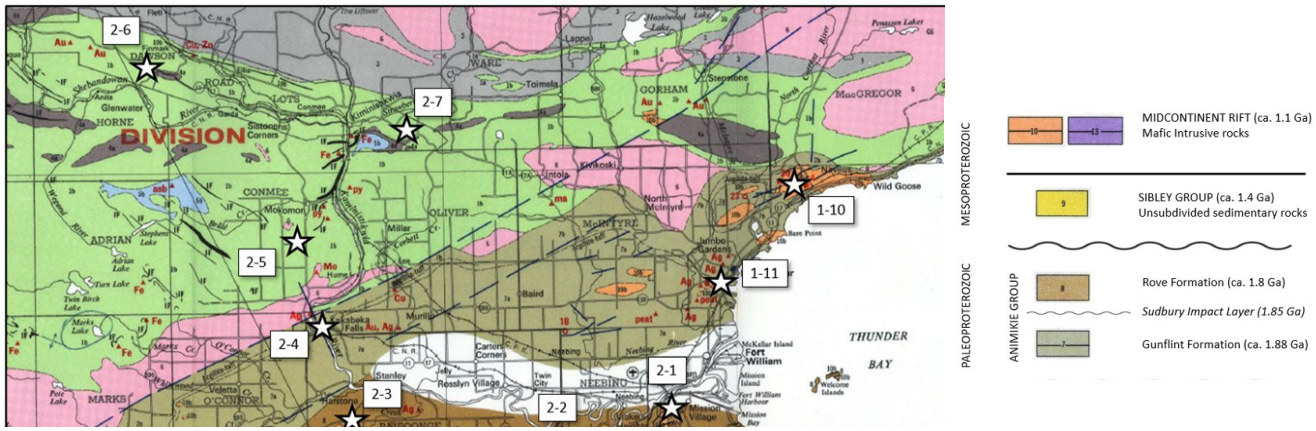


Figure 36. Generalized geology of the Thunder Bay area, showing Day Two field trip stop locations. Geology from Map M2232 (Carter et al., 1973).

STOP 2-1: Mount McKay Lookout (Anemki Wajiw) 0331126E / 5357384N *(n.b. admission via a gate operated by Fort William First Nation)*

Our first stop provides not only a panoramic view of Thunder Bay and surrounding area, but also stacked Logan sills which have produced the iconic mesa topography of Mount McKay and other similar mesas to the south and west in the Animikie-underlain Logan basin, collectively known as the Nor'westers. This location had previously been described by Cundari et al. (2012).

Mount McKay is also known as Anemki Wajiw ("Thunder Mountain") in Ojibwe. The summit, at 482 m ASL, is approximately 300 m higher than Lake Superior. The stop is centered on the lookout area (Figure 37), which represents the top of the lower sill at approximately 337 m ASL. The upper, ~60 m thick, columnar-jointed sill and adjacent, hornfelsed

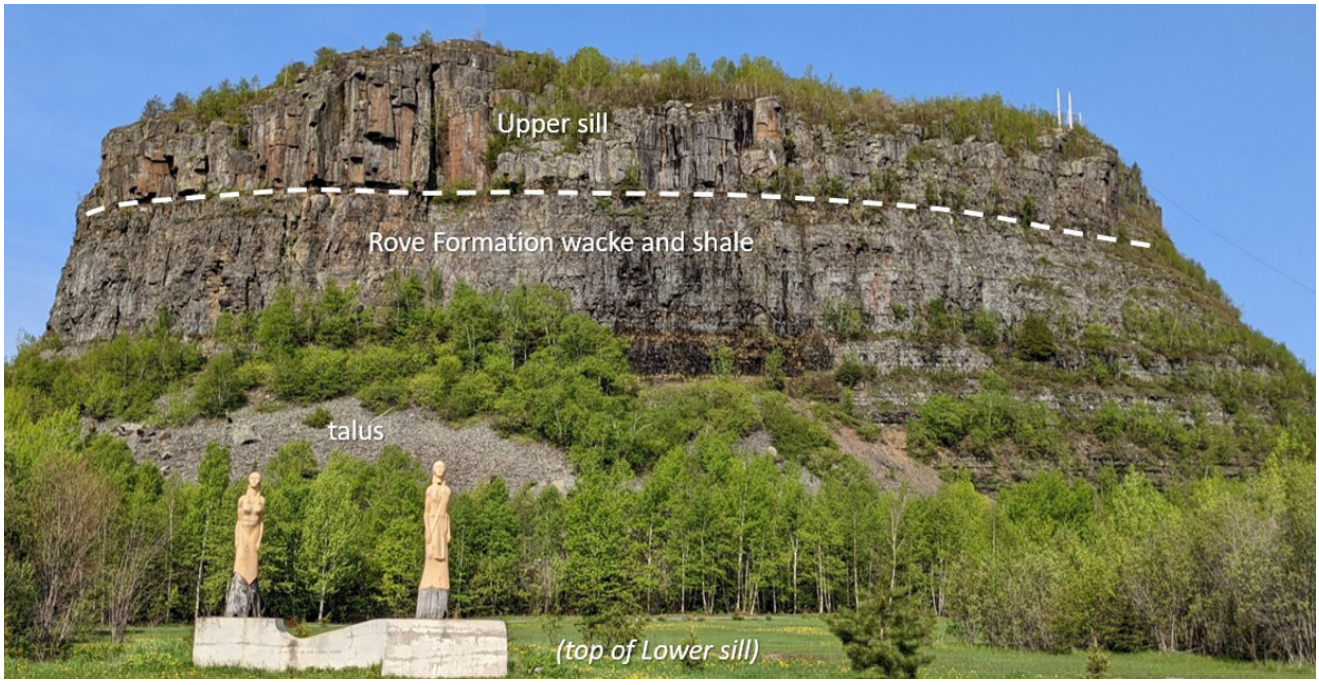


Figure 37. View of the top of Mount McKay, capped by the ~60 m-thick, upper Logan diabase sill. The Lookout level is underlain by the top of the lower sill.

Rove Formation wacke can be accessed by way of a hiking trail which leads to the summit. Although stacked sills have been encountered in drilling, few examples exist in surface exposures. As many at 14 sills were reported, for example, in a 705 m-deep drill hole in central Pardee Township by Dumont Nickel Inc. (Assessment Files, Thunder Bay South Resident Geologist's District, Thunder Bay).

The rugged topography (Figure 38) has produced extensive colluvial deposits and talus slopes. Unconsolidated, sandy lacustrine and fluvio-lacustrine deposits occur below the bedrock- and colluvium-predominated slopes. Abandoned shoreline escarpments and beach bars, visible between the lookout and Lake Superior, reflect higher post-glacial lake levels (Burwasser, 1977).

A tentative age of 1114.7 ± 1.1 Ma was determined from a Logan sill on Mount McKay, using a limited selection of very small baddeleyite grains (Heaman et al., 2007). Resampling of Zr-enriched, pegmatoidal, upper portion of the Logan Sill capping Mount McKay and re-analysis of baddeleyite and magmatic zircon yielded an age of 1106.3 ± 2.0 Ma (Bleeker et al., 2020). This, and similar ages elsewhere, led Bleeker et al. (2020) to favour a relatively sharp onset of high-volume mafic-ultramafic magmatism in the Midcontinent Rift at ca. 1110 to 1106 Ma.

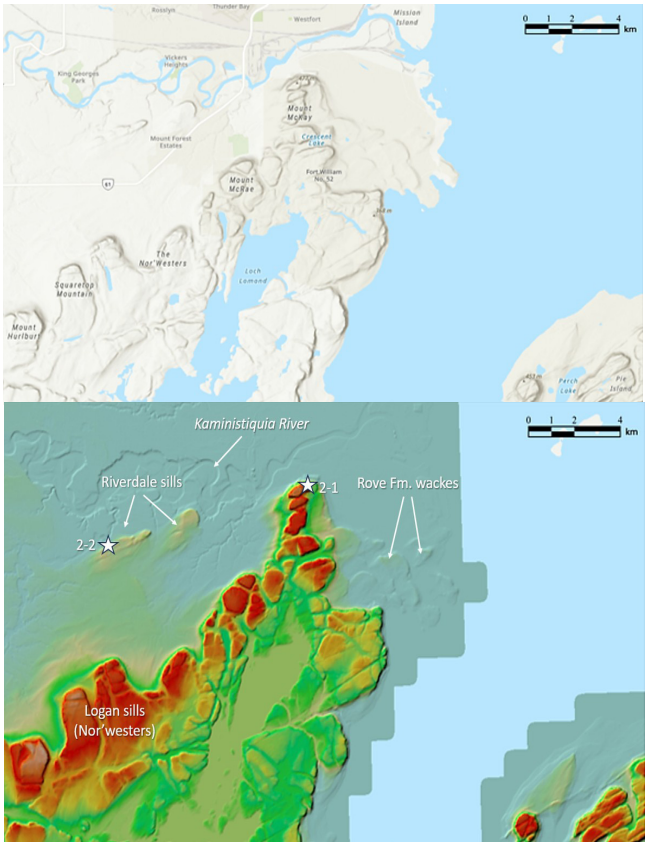


Figure 38. Shaded relief LiDAR image of area south of Thunder Bay, showing topographic relief (i.e. gently south-dipping mesas/cuestas) resulting from erosion-resistant mafic sills and, to a lesser extent, siliceous wackes in the Rove Formation (data from https://www.arcgis.com/apps/mapviewer/index.html?url=https://ws.geoservices.lrc.gov.on.ca/arcgis5/rest/services/Elevation/FRI_DTM_SPL/ImageServer). STOPS 2-1 and 2-2 are also shown.

Feldspar-phyric patches, common near upper chilled sill contacts, are present in an exposure of the upper, chilled contact of the lower sill along a path to the west of the clearing (Figure 39).



Figure 39. Polygonal jointing in upper chilled surface of lower diabase sill, STOP 2-1.

Logan sills generally consist of fine- to coarse-grained, ophitic to intergranular, quartz tholeiitic diabase/gabbro (Smith and Sutcliffe, 1987; Geul, 1970, 1973). Coarse-grained, intergranular gabbro, locally rich in granophyric mesostasis, is common in the interior of the thicker sills. Geochemical data from sampling of the upper and lower sills by Hart and Magyarosi (2004) are provided in Figure 40. These sills represent the northernmost known extent of Logan diabase sills near Thunder Bay. Nipigon diabase sills occur within the city and extend northward to the Nipigon Embayment.

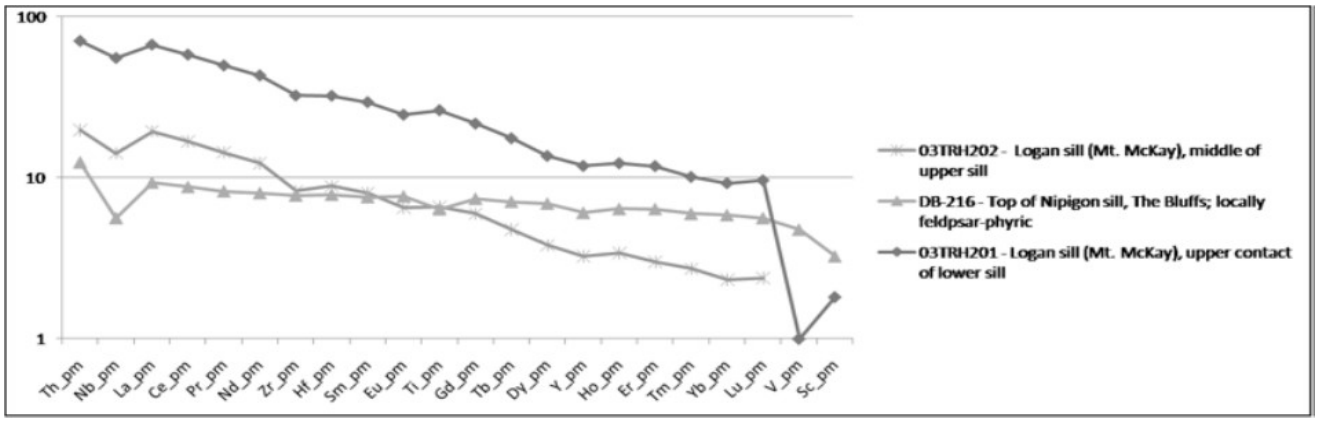


Figure 40. Primitive mantle-normalized trace element plots for upper and lower sills at Mount McKay with Nipigon sill sample for comparison. Data from Hart and Magyarosi (2004) and Hollings et al. (2011). Normalizing values from Sun and McDonough (1989).

Nipigon sills are characterized by generally lower incompatible trace element abundances, lower TiO_2 content, and a distinct negative Nb–Ta anomaly. They typically have lower Gd/Yb_{cn} ratios compared to Logan sills. Logan Sills are characterized by higher TiO_2 and higher Gd/Yb_{cn} ratios, indicating a greater degree of heavy rare earth element fractionation (cf. Hollings et al., 2010). Riverdale sills (STOP 2-2) are geochemically distinguishable from both Nipigon and Logan diabase (Figure 41).

STOP 2-2: Riverdale Quarry 322418E / 5355233N

(n.b. Private property; permission is required to access. Caution advised on site due to slip and fall risks associated with steep slopes and vertical rock faces)

This former shale quarry exposes a ~20 m-thick section of the lower Rove Formation, overlain by a ~12 m-thick Riverdale, columnar-jointed, gabbronorite sill related to Midcontinent Rift magmatism (Figure 41).

This location was previously described by Cundari et al. (2012):

"Sampling by Smyk and Hollings (2007) identified this as a Riverdale gabbronorite

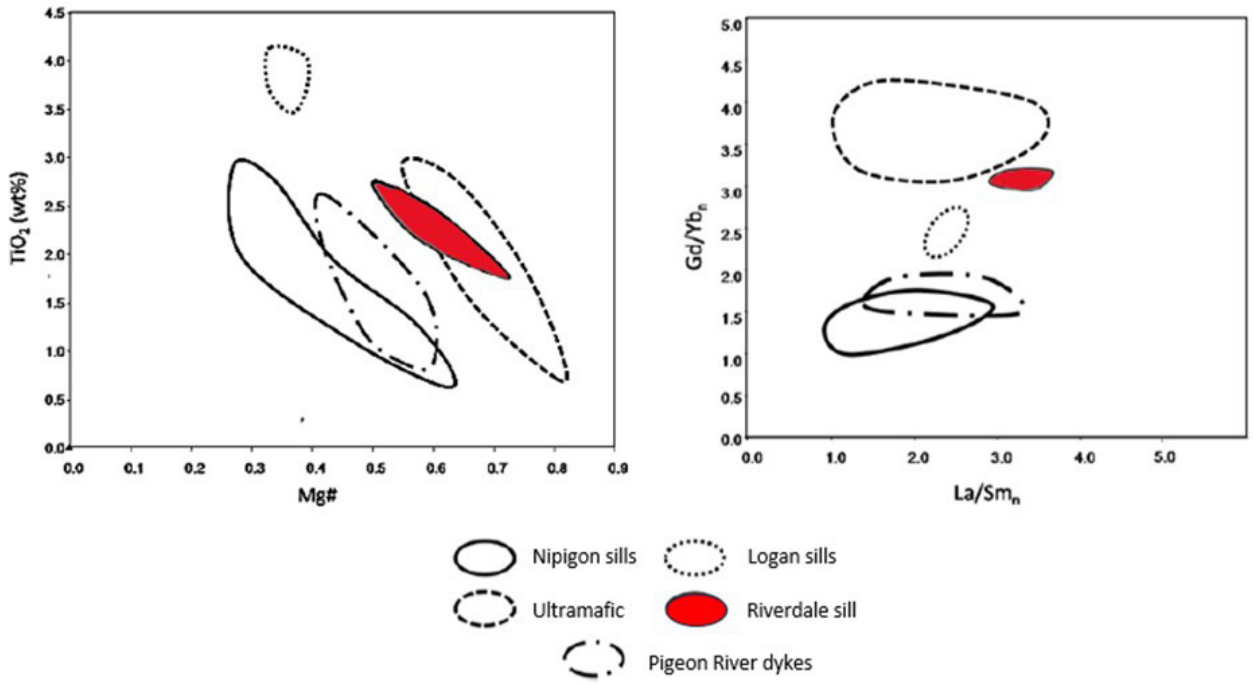


Figure 41. Discrimination diagrams for mafic and ultramafic intrusions near Thunder Bay (from Cundari et al. 2012). Data are from Hollings et al. (2007a) and Puchalski (2010). Normalizing values from Sun and McDonough (1989).



Figure 42. Riverdale gabbronorite sill capping section of Rove Formation clastic sedimentary rocks, Riverdale Quarry. (Photo taken in 2008. Arrow points to geologist for scale.)

sill in Rove Formation shale, wacke and minor tuffaceous units. Subsequent detailed petrographic and geochemical analyses were carried out by Puchalski (2010). Samples were taken through stratigraphy at the quarry to investigate composition and contamination, as well as to test whether the sill had undergone differentiation. The following section provides a concise summary of those findings.

The mafic intrusive rocks within the quarry are dominantly classified as gabbronorites with olivine gabbro present towards the center of the sill. The gabbronorites are generally fine-grained with plagioclase occurring as subhedral laths. Orthopyroxene is present in greater abundance than clinopyroxene, occurring as anhedral to subhedral crystals. Varying degrees of alteration are manifested as sericitization of plagioclase and chloritization of pyroxene. The olivine gabbro is texturally similar to the gabbronorite, albeit with a higher modal percentage of fine-grained, anhedral to euhedral olivine. In most samples, olivine is replaced by serpentine, producing secondary quartz and calcite, as well as minor magnetite. Alteration is significantly greater in the narrow, chilled margin at the contact. Pyrite occurs throughout the unit; minor chalcopyrite has also been noted.

Sampling for whole rock major and trace element geochemistry was undertaken by Puchalski (2010) throughout the 10 m exposure at 1-m intervals. Olivine gabbro samples display broadly similar trace element characteristics to those of the gabbronorite samples. Differences lie within the major element abundances; olivine gabbros are lower in SiO_2 and elevated in MgO, Cr, Co, and Ni compared to the gabbronorite samples. The sill does not display any evidence for

differentiation as shown by the erratic trends of MgO, SiO_2, TiO_2, Cr and Ni through stratigraphy. An olivine gabbro in the center of the sill displays elevated MgO, Cr, and Ni values as well as a lower abundance of silica when compared to the surrounding samples. This is likely the result of a slightly more primitive magma intruding the center of the sill. The lack of chilled margins between the olivine gabbro and the gabbronorite suggest that the sill had not fully crystallized when the second pulse intruded. A sample of a 60 cm wide north-trending diabase dyke which intrudes the sill near the western end of the quarry is geochemically comparable to the surrounding Logan sills.

Contamination by the Rove shale is evident in samples taken from close to the contact (<1 m above the contact). These samples display higher SiO_2 values as well as lower Nb/Nb and Gd/Yb_n values than the rest of the unit. As the Rove shale displays significantly lower Nb/Nb* and Gd/Yb_n values than that of the surrounding gabbronorite. The Rove shale is the likely source of this contamination signature. Two different pulses of magma are recognized within the Riverdale sill, based on contamination signatures denoted by negative niobium anomalies. The less-contaminated samples are typically found towards the core of the intrusion with rocks above and below displaying a greater degree of contamination. Samples taken within 60 cm of a shale xenolith do not display a distinct negative niobium anomaly. This shows that the source of contamination responsible for the negative niobium anomaly is not the Rove shale but is likely a crustal component from depth. εNd (T=1100 Ma) values of -1.6 to -1.9 for the Riverdale Sill are consistent with this model (Smyk and Hollings, 2009).*

Although the Riverdale sill is located near Logan sills, it remains petrographically and geochemically distinct from them [Figure 42]. Geochemical discrimination based on La/Sm_n (LREE) vs. Gd/Yb_n (HREE) shows characteristics

similar to those for the ultramafic units of the Nipigon Embayment (e.g., Disraeli, Kitto, Hele and Seagull), closely resembling the mafic to ultramafic Jackfish sill. The Jackfish sill is finer-grained and displays a higher modal abundance of olivine than the Nipigon sills surrounding it (Hollings et al., 2007a). This suggests that the Riverdale sill may be genetically related to the ultramafic and mafic to ultramafic units of the Nipigon Embayment. This is consistent with the reversed polarity of the Riverdale sill (Hollings et al., 2010)."

A number of features are visible at or near the exposed upper and lower sill contacts (Figure 43). Calcite-filled vesicles define a crudely developed

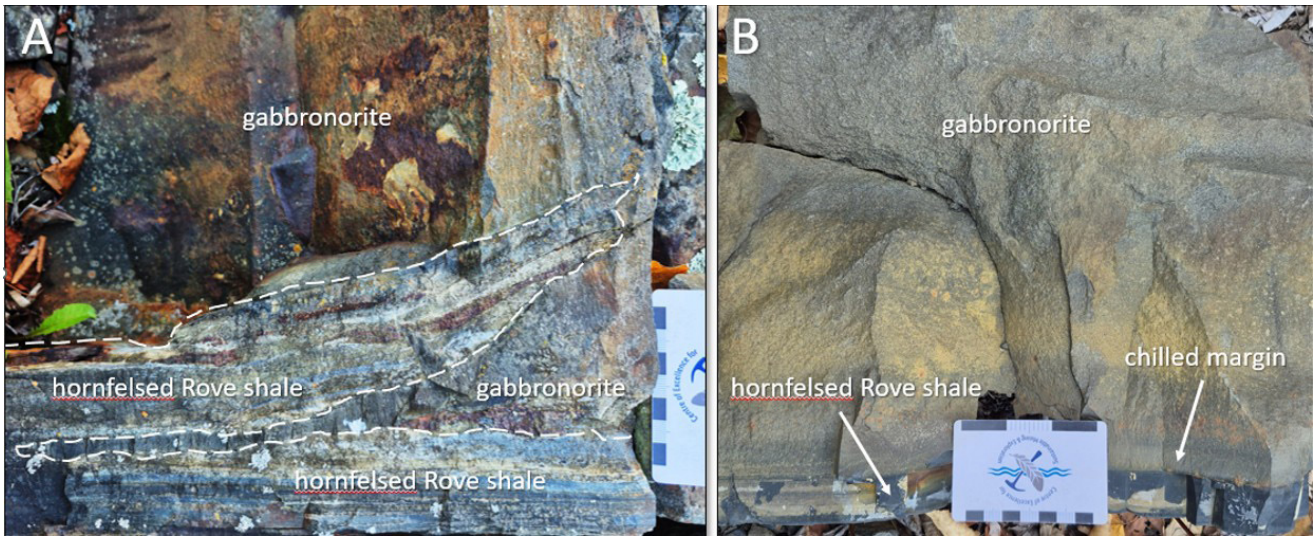


Figure 43. Lower gabbronorite sill contact. (A) Delamination of Rove shales by injection of gabbronorite sill magma; (B) Chilled margin of gabbronorite sill against hornfelsed Rove shale.

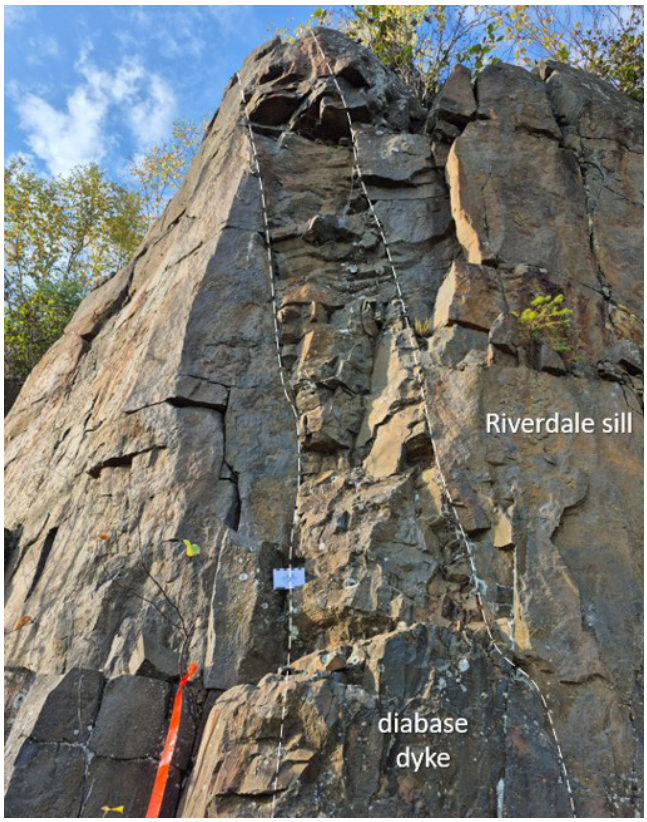


Figure 44. Narrow diabase dyke with Logan sill-like geochemistry intruding Riverdale gabbronorite sill, Riverdale Quarry. Scale card straddles the eastern dyke contact.

layer/joint filling(?) in medium-grained gabbronorite, ~1 m above the lower sill contact. Stoping and delamination of Rove shales is also evident here. Thin, parallel chilled margins, perhaps representing multiple influxes of magma, occur above the lower sill contact. A narrow (75 cm) diabase dyke with Logan sill-like geochemistry intrudes the Riverdale gabbronorite sill near the western end of the quarry exposure (Figure 44). Glacial striae are visible on exposed outcrop surfaces at 060˚ and 075˚.

STOP 2-3: Sudbury Impact Layer, Highway 588 0307539E / 5357977N

This stop, while having lost much of the best exposure of the Sudbury Impact Layer (SIL) due to ongoing highway construction, still provides an opportunity to view some of the features associated with the SIL in the affected ankeritic, Gunflint Formation chert-carbonate rocks. The SIL was also intersected a few metres below surface in a shallow drill hole, collared in Rove Formation shales in an abandoned quarry approximately 300 m south-southwest of STOP 2-3. This location was previously described by Addison and Brumpton (2012):

"When first observed in 2000, the Hwy 588 outcrop was a bedrock exposure in the ditch on the northwest side of the highway, 2.4 km southwest of the hamlet of Stanley. It was a glacially polished and striated surface showing erosively truncated stromatolites up to 0.5 m diameter, some of which were surrounded by accretionary lapilli 3-25 mm in diameter [Figure 45]. Ankeritic grainstone and chloritic grainstone surrounded other stromatolites. This exposure was subsequently blasted to deepen the ditch and the blasted rock now lines the ditch slopes, giving a highly fragmented cross-section and plan view of the exposure. Since then, we have exposed bedrock in the ditch about 50 m southwest of the first exposure. It shows a glacially striated surface of exposed stromatolites and shattered, but in-situ black chert with an ankeritic grainstone filling in the cracks. The chert is assumed to have fractured during the compressional stage of impact-triggered earthquake waves with the fractures then opening during the dilational wave phase. Fine granular material then fell into the openings, preventing them from closing and subsequently the material was lithified.

Thin sections prepared from the blasted material show a variety of ejecta features, the most obvious being accretionary lapilli which have yielded quartz and feldspar grains showing planar deformation features (PDFs) and planar fractures. Planar features have not been found in larger subrounded and angular quartz and feldspar grains contained within the debrisite generally as opposed to within accretionary lapilli. This is the only site in which DVIG [devitrified vesicular impact glass], is not the most obvious ejecta feature within the debrisite. In fact, no DVIG has been observed, however carbonate and silica replaced clusters of spherules are present.

Non-ejecta features include subrounded to round chert grains in carbonate cement, sub-centimetre stromatolite fragments and mudstone and shale rip-ups. Chloritic, blotchy, black Gunflint Formation granules, similar in shape and size to microtektites, are present within the carbonate cement. Carbonate-replaced microtektite shapes are present but since they lack residual internal structure, it is impossible to determine if they were microtektites or carbonate-replaced Gunflint chlorite granules."

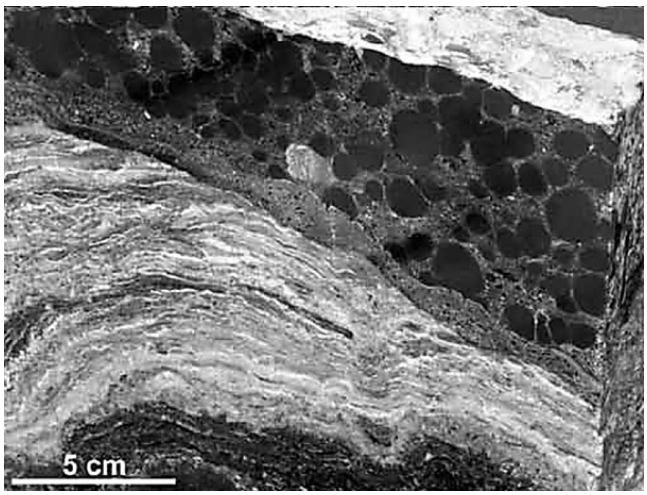


Figure 45. Accretionary and armored lapilli draped unconformably over a stromatolite, composed of silicified carbonate, which was abraded to its present configuration likely by a base surge immediately preceding the deposition of the lapilli; STOP 2-3, polished surface. The gray component is primarily fine-grained, angular, fractured carbonate clasts whose individual crystals are usually <10 μm. These clasts are typically <50 μm but they may be as large as 500 μm. Quartz and feldspar grains are a minor component among the carbonate clasts within the lapilli. (caption modified from Addison and Brumpton, 2012).

STOP 2-3: Kakabeka Falls Provincial Park 0305738E / 5364400N; 0305178E / 5364663 ***(n.b. Entry/parking fee is required in Kakabeka Falls Provincial Park. Sample collecting and hammers are NOT permitted.)***

Two stops at Kakabeka Falls provide an opportunity to see both a thick section of Gunflint Formation rocks exposed in the gorge of the Kaministiquia River, and the basal, stromatolite-bearing units of the Gunflint unconformably overlying Neoarchean granitoid basement. This location was previously described by Pufahl et al. (2000) and Smyk (2012).

The park is dominated by a single, spectacular feature, Kakabeka Falls, which drops 39 m over sheer cliffs in Gunflint Formation sedimentary rocks (Figure 46). Kakabeka is an aboriginal word meaning "steep cliffs". The age of the river gorge below the falls is still debated. If none of it existed prior to the glacial Lake Beaver Bay stage, then it is less than ca. 9700 years old. The portage around the falls contains artifacts ranging from the Paleoindian to the historic (fur trade) periods.

The falls owes its existence to the thin chert-



Figure 46. Kakabeka Falls and gorge, cut into flat-lying Gunflint Formation shales. Photo from https://hikebiketravel.com/a-trip-to-kakabeka-falls-near-thunder-bay/.

carbonate bed which forms a resistant cap rock to the softer underlying shales. The river gorge is composed of a sequence of volcaniclastic shales (less-resistant, darker units) and tuffs (more-resistant, lighter coloured units). This sequence represents the major volcaniclastic horizon in the upper Gunflint Formation that is traceable to the south through the Mesabi Range. Note that shale is the predominant lithology in the Kaministiquia sections and this is, in fact, typical for the Gunflint Formation in general throughout the Thunder Bay region.

Samples of lapilli tuff and reworked tuffs from the middle of the Gunflint Formation, collected by Fralick et al. (2002) at Kakabeka Falls yielded a euhedral zircon population with a U-Pb age of 1878.3 ± 1.3 Ma, believed to be nearly synchronous with the depositional age.

The outcrop on the northern edge of the parking lot contains layers of banded/ribbon chert-carbonate within black, fissile shale. The alternating, dark grey chert and brown siderite-ankerite layers display slump and soft-sediment deformation features. Microscopic examination of banded chert-carbonates reveals delicate lamination in the chert which resembles the "ribbon texture" of algal mats. The interlayered carbonate bands contain complex, microspherical structures which likely resulted by nucleation from a gel state. Local thick beds of carbonaceous siderite (2-3 wt% carbon) form carbonate iron formation; contemporaneous deposition of carbon and carbonate suggests biological activity during iron deposition. Studies of the Gunflint Formation have described this type of sediment as forming in a deep, quiet water environment. However, similar carbonate sequences are often attributed to deposition in intertidal or lagoonal subtidal environments (Pufahl et al., 2000).

The rapids visible north of the highway bridge are formed by Archean granitoids. The slow-water area to the south is underlain by the Gunflint Formation. The basal conglomerate (Kakabeka member) is patchily preserved on Archean basement here. Silicified stromatolites are developed on the conglomerate or directly on the basement. This is the location from which samples collected from silicified stromatolites in the 1950's yielded the first documented Gunflint cyanobacteria (Tyler and Barghoorn, 1954).

The rock cut on Highway 590 immediately south of the intersection with Highway 11-17, west of the Kaministiquia River bridge, expose cherty carbonates at the base of the Gunflint Formation where it rests unconformably over Neorchean granitoid basement. Large-form stromatolites are developed at the unconformity (Figures 47 and 48). The stromatolitic, ribbon carbonates are abruptly overlain by a grainstone succession. Black anthraxolite veinlets and void fillings occur with vein quartz in the chert-carbonate rocks.

Anthraxolite and pyrobitumen in the Gunflint Formation (Figure 49) has been noted and studied numerous researchers, including Tanton (1931), Ellesworth (1934), Goodwin (1956), Kwiatkowski (1975), Barghoorn et al. (1977), Hayatsu et al (1983), Mancuso et al. (1989), Rutter (2014), Rasmussen and Muhling (2019), and Rasmussen et al. (2021).



Figure 47. Colloform stromatolite (left of hammer) in Fe-rich carbonate grainstones, Highway 590 exposure, STOP 2-4. The stromatolite was situated approximately 40 cm above the unconformity with Neoarchean basement granitoid rocks. Unfortunately, the stromatolite spalled from the outcrop face ca. 2016. Coin is 2.5 cm in diameter.



Figure 48. Detailed view of the large, silicified, colloform stromatolite in Figure 47.

Rasmussen et al. (2021) suggested that stromatolitic, black Gunflint cherts were saturated in syn-sedimentary oil. Thermally altered oil (pyrobitumen) occurs in the stromatolites and intercolumn sediments, fills pores and fractures, and coats detrital and diagenetic grain surfaces. Hayatsu et al. (1983) described two very distinct macromolecular materials in the Gunflint anthraxolite that suggested that the Thunder Bay area was once covered by Cretaceous or Jurassic marine sediments, similar to those documented in the Mesabi range of Minnesota.

STOP 2-5: Briggs Drive Gravel Pit 0304790E / 5369390N (n.b. Private Property, contact Township of Conmee for access permission)

This stop not only highlights some interesting glacial sediments, but also provides an opportunity to examine large boulders of a variety of local rock types that have been transported by glacial ice and meltwater. This location was described by Bajc (2000):

"At this stop, we will be looking at a section of Superior lobe drift. The exposed sequence consists of approximately 6 to 7 m of well sorted, steeply-dipping sand and gravel of Superior provenance overlain by 2 to 3 m of silty Superior lobe till [Figure 50]. Clasts in the lower glaciofluvial unit consist primarily of Proterozoic metasedimentary rocks. Numerous cobbles and boulders belonging to the Gunflint Formation and Sibley Group are recognizable. Foreset beds dip steeply to the north and are likely of deltaic origin. The delta was probably built proglacially into an early phase of glacial Lake Kaministikwia. The feature therefore represents a location at which the advancing Superior Lobe stalled prior to reaching its maximum position at the Marks Moraine.

The delta is actually located within only 3 km of the Superior lobe limit and occurs at an elevation of about 375 m asl, 85 m below the maximum elevation of Lake Kaministikwia. The delta was



Figure 49. Void-filling, conchoidal anthraxolite and quartz in sideritic Gunflint grainstone, Highway 590 exposure, STOP 2-4.



Figure 50. View, looking west, of steeply dipping, gravelly foreset beds of a delta constructed along the margin of the advancing Superior lobe, Briggs Drive gravel pit, STOP 2-5. Silty Superior lobe till caps the sequence.

overridden by the Superior lobe resulting in the truncation of the foreset unit and removal of the topset beds. Several metres of silty, Superior Lobe subglacial till was deposited on top of the sands and gravels.

Of particular significance is the occurrence of large angular to rounded boulders on the pit floor. The boulders were extracted from the lower glaciofluvial unit and, in some cases, do not appear to have been transported very far. Some of the larger boulders measure several metres across and still display striated surfaces. The boulders are derived from both Archean and Proterozoic source rocks. Several boulders of sulphidized iron formation and massive pyrite of Archean age were discovered in the boulder piles. One of the boulders measured over 1 m in diameter and consisted of massive pyrite and magnetite with 10 to 15% sphalerite disseminated in pyrite-rich sections. Sphalerite was also concentrated along fractures and adjacent to quartz veinlets throughout the rock. Two samples from the pyrite-rich zones returned values of: 1) 5.13% Zn, 18 ppm Cu, 19 ppm Pb, 260 ppb Au and 0.5 ppm Ag; and 2) 2.85% Zn, 16 ppm Cu, 20 ppm Pb, 245 ppb Au and 0.5 ppm Ag. A sample from the magnetite-rich zone returned values of 850 ppm Zn, 25 ppm Cu, 5 ppm Pb, 25 ppb Au and <0.2 ppm Ag. A second sulphidized iron formation boulder measuring approximately 0.5 m in diameter and consisting almost exclusively of pyrite, returned values of 140 ppm Zn, 8 ppm Cu, 11 ppm Pb, 710 ppb Au and <0.2 ppm Ag.

There are two possible source areas for the boulders. Superior lobe striae in the immediate vicinity of the pit are oriented at 320 to 330° Az. If the boulders were eroded and transported by Superior ice, then there is a 5 km window towards the southeast from which they could have been derived. Proterozoic metasedimentary rocks outcrop beyond the 5 km limit. Alternatively, the boulders could have initially been eroded by northern ice from a source to the north-northeast of the pit then remobilized by the Superior lobe.

Exploration work during the early 1900s along the lower reaches of Brule Creek, 4 to 5 km north-northeast of the gravel pit, by B.L. Morrison, the Davis Sulphur Company and General Chemical Company resulted in the discovery of "seven lenticular masses of brecciated, banded iron formation, in which pyrite has replaced a considerable part of the rock" (Carter 1990). The largest of these masses has a maximum width of 23 m and is 244 m long. Other discoveries include a 21-m wide body of pyrite containing magnetite and pyrrhotite and a 9 m wide by 15 m long zone of magnetite-pyrite-jasper ironstone. It is possible that the boulders found within the gravel pit [Figure 51] were derived from this area and that sphalerite was not recognized in the rock. It is not yet clear whether the sulphides indicate proximity to a VMS style zone of mineralization. Further work is required to assess the mineral potential of this area."

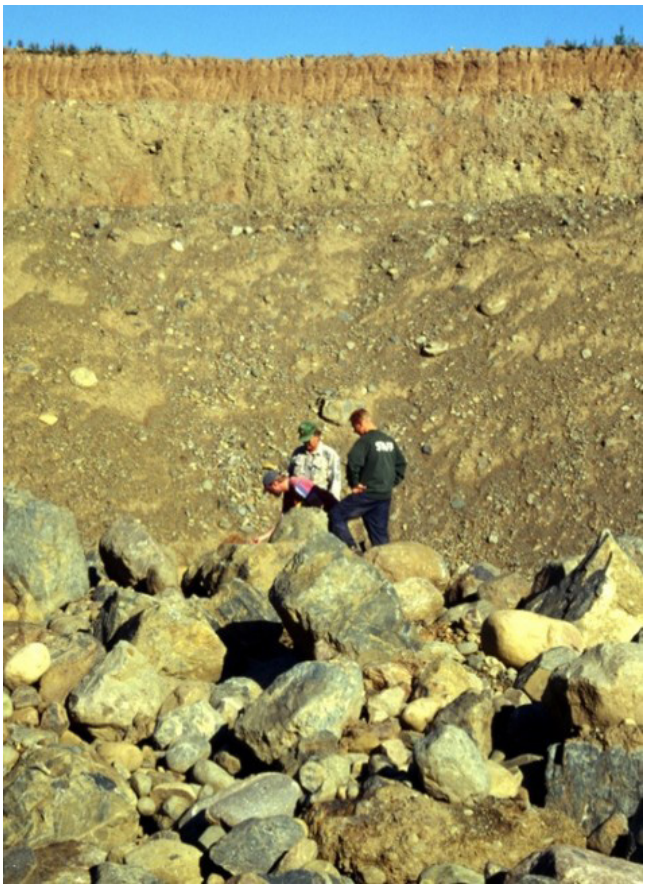


Figure 51. Pile of oversized boulders of a variety of local Archean and Proterozoic rock types, Briggs Drive gravel pit, STOP 2-5. Reddish silty Superior lobe till is visible at the top of the pit wall. Photo taken ca. 1999.

STOP 2-6: Temiskaming sedimentary rocks, Finmark 293709E / 5383903N

This stop, having long been a "must-see" for local geology students, has benefited from new exposures

on the south side of Highway 11-17 that were created during highway expansion and ballast quarry development ca. 2019. The outcrops on both sides of the highway expose excellent examples of clastic Neoarchean "Temiskaming-type" metasedimentary rocks (ca. 2690-2695 Ma) of the Shebandowan assemblage that still display many primary sedimentary features that provide clues as to the depositional environment. The Timiskaming-type successions of the SGB are interpreted to have been deposited in subaerial to shallow marine environments (Shegelski, 1980).

The ballast quarry immediately to the south has been developed in mafic, Neoarchean metavolcanic rocks of the Greenwater assemblage (ca. 2720 Ma) which presumably underlie the clastic rocks unconformably or are in fault contact with them. Parker (1980) noted that reversals of top directions and the presence of both easterly and westerly plunging minor folds, suggest that one or more episodes of folding have occurred. Detrital zircon geochronology by Corfu and Stott (1998) confirmed that the metasedimentary rocks in the Finmark area ($\leq$2691$\pm$3 Ma) and in the southern part of Adrian Township ($\leq$2700$\pm$4 Ma) are younger than the Greenwater assemblage.

Because of the remarkable preservation of primary sedimentary features, this stop has been the focus of study for many years, including theses by Parker (1980) and Koebernick (1996). The metasedimentary sequence here comprises interbedded sandstone-siltstone-mudstone sequences which alternate with thick deposits of cross-stratified sandstones (Parker, 1980). The interlayered sequences contain many of the primary sedimentary structures characteristic of tidal flat deposits, such as flaser bedding, lenticular bedding, herringbone cross-bedding, mud cracks, mud drapes, and bipolar paleocurrent indicators. Parker (1980) noted that the clastic sedimentary rocks are composed of feldspar, rock fragments, quartz, and some mafic minerals. Modal analysis revealed that most of the sandstones in the area are arkosic arenites. Lithic fragments are felsic to intermediate and predominantly calc-alkalic volcanics, with lesser amounts of other igneous grains and sedimentary rock fragments. This led Parker (1980) to suggest that the clastic rocks probably represent immature detritus from proximal volcanic centers.

A detailed study of the "Temiskaming-type" clastic rocks at this location and along this section of Highway 11-17 by Koebernick and Fralick (1995) and Koebernick (1996) documented sedimentary structures and bed sequences consistent with shallow water, coastal sedimentation in three major depositional environments: tidal strandline, the shoreface, and the offshore (e.g. Figure 52). Koebernick (1996) noted:

> *"The three environments and associated sub-environments record processes reflective of differing current activity which controlled and*



Figure 52. Stacked, trough cross-bedded sandstone beds with tangential foreset laminae, south side of Highway 11-17, STOP 2-6. Ripples are preserved on bedding surfaces (lower photo).

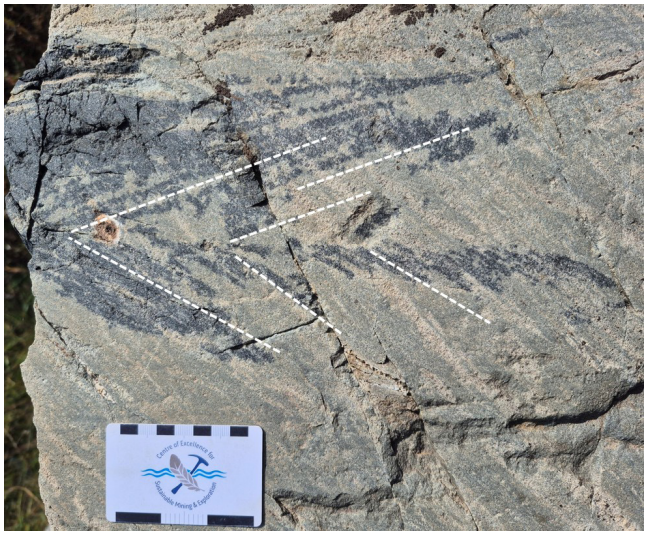


Figure 53. Dark alteration holes defining herringbone cross bedding in sandstone, north side of Highway 11-17, STOP 2-6.

influenced deposition. The tidal environment was dominated by bidirectional tidal currents [Figure 53]. Deposition In the shoreface was predominated by unidirectional wave-produced currents which overprinted prevailing tidal current activity, in the distal portions of the shoreface environment though, deposition was once again controlled by tidal currents. In the offshore, deposition was controlled by storm currents which generated distinctive beds of hummocky cross-stratification. The tidal environment is composed of many sedimentary structures similar to those present in Phanerozoic and present-day tidal sequences. In the tidal flat sub-environment, vertical sequences of flaser, lenticular, wavy and coarsely interlayered bedding reflect current velocity fluctuations Intimately tied to spring - neap tidal cycles. The tidal channel sub-environment lacks many of the features characteristic of tidal channels described in the literature; such as extensive point bar development. Instead, the tidal channels of the study area appear to represent sequences deposited in relatively straight channels. Migration of sand waves and dune fields deposited the cross-stratified lithofacies of the shoreface environment. Similar to a high-energy, non-barred coastline, the proximal portion of the shoreface lacks any evidence of beach development. Instead, the shoreface records a rapid and discontinuous transition from the tidal strandline environment. Hummocky cross-stratification (HCS) [Figure 54], parallel-

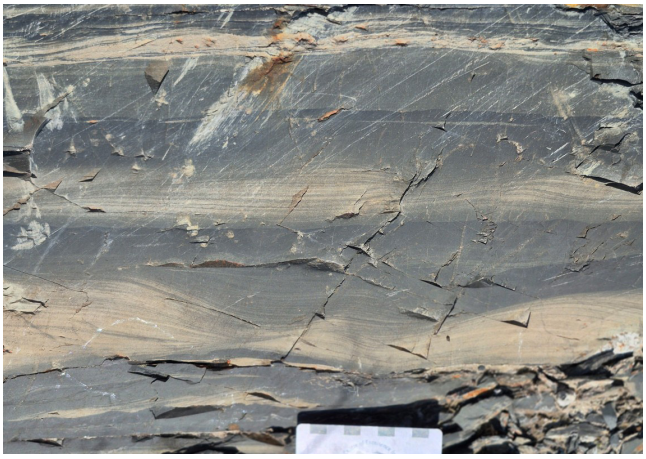


Figure 54. Hummocky cross-stratification, which is only formed and preserved by storm waves in depths between fair weather wave base and storm wave base.

laminated and massive sandstone beds as well as siltstone and mudstone beds typify the offshore environment [Figure 55]. The HCS differs greatly in thickness and internal structure from HCS described in the literature. The HCS in the study area reflects restricted and/or variable sediment supply and flow conditions. A paleotidal range was determined from the sediments of the tidal environment. The range indicated a mesotidal



Figure 55. Thinning- and fining-upward sequence, showing transition from medium-bedded sandstones to a mudstone/siltstone-dominated package with thin sandstone interbeds, south side of highway, STOP 2-6. This likely represents deepening of the water, going from nearshore coarse-grained sands moved around on the bottom as dunes by fair-weather waves and currents, to deeper water deposits below fair-weather wave base, representing tempestites (i.e. hummocky cross-stratified storm deposits and graded beds formed below storm wave base by the same geostrophic flows that formed hummocky cross-stratification in shallower water; P. Fralick, personal communication, 2026).

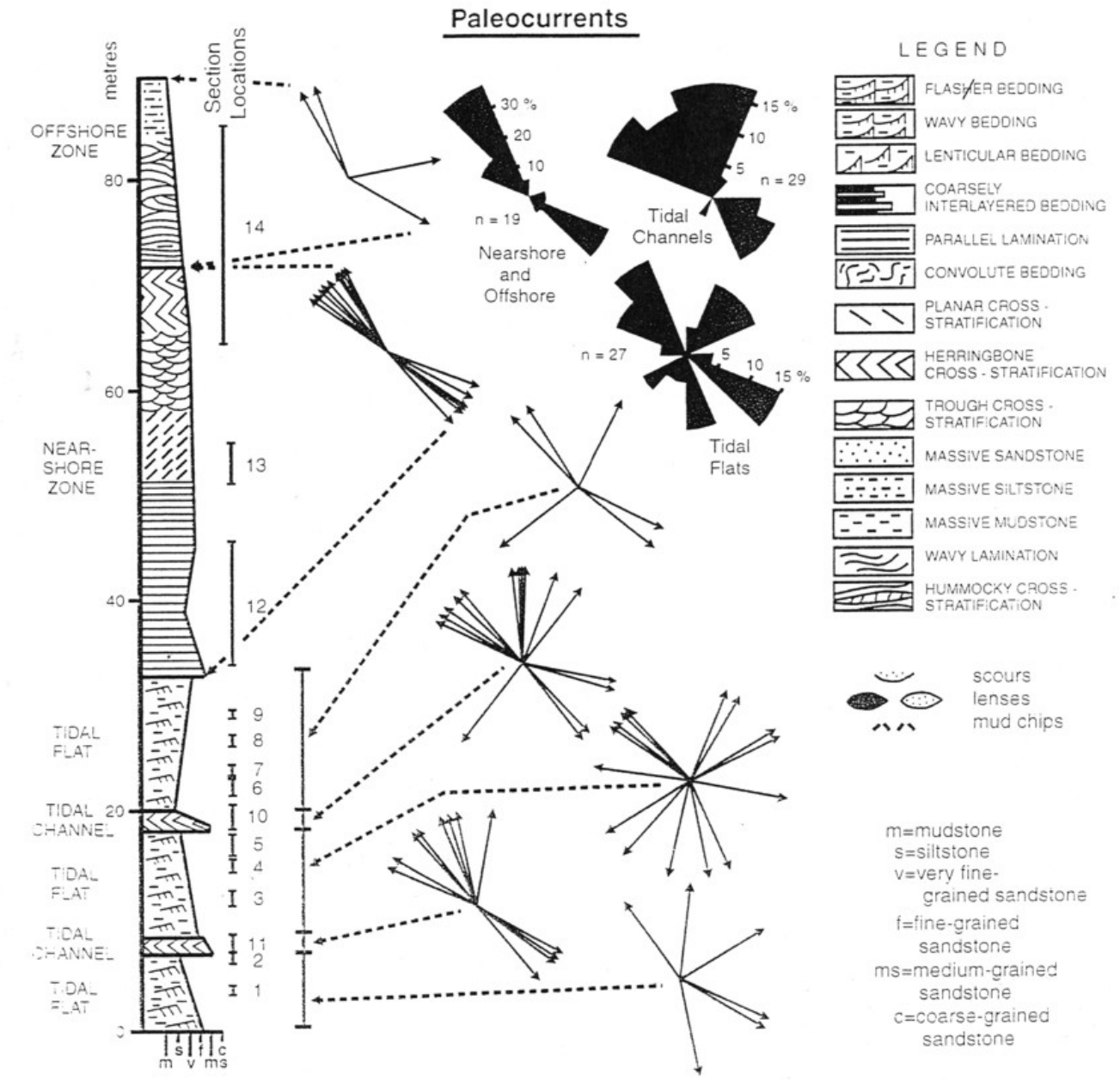


Figure 56. Stratigraphic column of outcrops on north side of Highway 11-17, STOP 2-6, showing primary sedimentary features and paleocurrent measurements (P. Fralick, personal communication, 2026).

environment and is comparable to Precambrian tidal ranges reported in the literature. Tidal rhythmites, present on the tidal flats, suggest a length of 26 days for the Neoarchean lunar month. Currents which deposited the tidal rhythmites produced both semi-diurnal and diurnal sediment sequences [Figure 56]."

As noted above, in spite of the remarkable preservation of sedimentary structures, the Shebandowan assemblage sedimentary rocks in this area have experienced tectonic deformation and display features that include minor folds and younging reversals. Evidence of shearing and brittle deformation, including quartz-carbonate veining, can also be observed, especially in the easternmost portions of the outcrop exposure on the south side of the highway (Figures 57 and 58).

Bedding-cleavage relationships indicative of folding are visible in the outcrops at this location. Bedding orientations vary from approximately 325/70 northeast on the north side of the highway to 110/70 south on the south side. The cleavage-bedding relationship is most easily observed in the thin-bedded mudstones and siltstones south of the highway, where the cleavage



Figure 57. Quartz-carbonate veining in altered and deformed rocks at the east end of the outcrop area on the south side of Highway 11-17 at STOP 2-6.



Figure 58. Photo illustrating small-scale folds, shears and sigmoidal tension fractures in thin-bedded siltstone and mudstone at STOP 2-6.

orientation is approximately 085/85 south. The change in bedding orientation relative to cleavage, when combined with northward-younging indicators (e.g., graded bedding), indicate the probable presence of an anticline axis a short distance to the north. This interpretation is consistent with previous geological mapping completed by Carter (1985).

The structures observed here may have developed during the same tectonic events that gave rise to orogenic gold mineralization at the nearby Eureka Gold Deposit, which is currently being explored by Delta Resources Limited. Eureka is located approximately 4 km to the west-northwest of here, and the deposit occurs within a structural corridor known as the "Shebandowan structural zone." Gold mineralization at Eureka also has a close spatial association with the unconformity between the Greenwater and Shebandowan assemblages. Delta Resources has outlined the Eureka Gold Deposit over a 2.5-kilometre strike length, and to a vertical depth of 300 metres. Mineralization occurs over true widths ranging from 10 to 100 metres, and the deposit remains open in all directions.

Gold is hosted by multiple generations of quartz-ankerite-pyrite veinlets that generally range from 1 mm to 10 cm wide and cross-cut multiple lithologies. Wider quartz veins up to 4.5 metres wide, and gold-bearing silica-flooding zones are also found within the deposit. Host rock alteration is characterized by intense, texture-destructive ankeritization, silicification, albitization and sericitization combined with trace to 2% disseminated pyrite and trace arsenopyrite. The altered rocks typically contain anomalous gold. Feldspar-phyric monzonite to diorite dikes also have a close spatial association with the mineralization and are locally altered (https://www.deltaresources.ca/delta-1-gold-project/).

STOP 2-7: Pillowed Basalt, Mud Lake 315029E / 5376770N

No trip would be complete without pillowed basalt! These roadside exposures along Highway 102 near Mud Lake display tholeiitic, mafic to intermediate volcanic



Figure 59. Pillowed basalt flow, STOP 2-7, showing well-preserved, close-packed pillows and hyaloclastite-filled inter-pillow spaces.

rocks of the Neoarchean Greenwater assemblage, situated near the Quetico–Wawa subprovince boundary (Brown, 1995; Brown and Fogal, 1995). In this area, the degree of pillow preservation varies considerably.

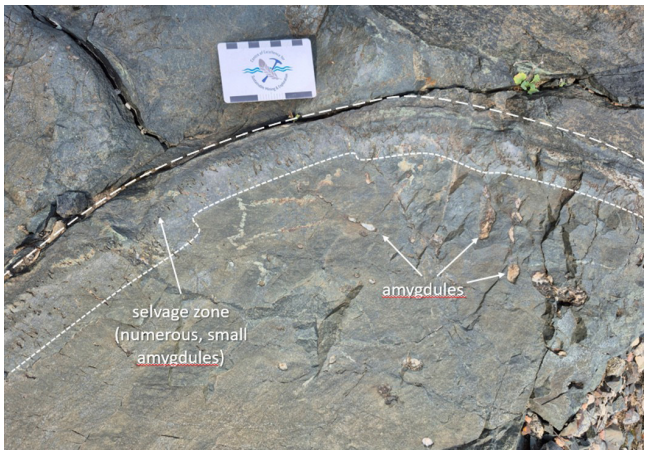


Figure 60. Close-up of pillow, STOP 2-7, showing contact of chilled upper selvage (large dashed line).

However, well-formed, close-packed pillows, ranging from 10 by 15 cm to 30 by 60 cm in size, are locally preserved (Figure 59). Where discernible, younging directions within this unit are consistently to the north. Carbonate-filled amygdules, generally less than or equal to 1 mm in size and constituting up to 10% of the pillows by volume, are commonly present, radiating outward from the core of the pillows (ibid; Figure 60).



Figure 61. Isolated-pillow breccia, STOP 2-7, showing amoeboid to angular pillow fragments in hyaloclastite-rich matrix.

North-younging, pillowed flows exposed at STOP 2-7 display well-preserved primary features, including autoclastic breccias (e.g. pillow breccia, inter-pillow hyaloclastite; Figure 61), close packing and pillow cusps, and calcite-filled amygdules. Larger, ovoid amygdules occur sparingly in the cores of pillows, while smaller, more numerous, pipe-like amygdules tend to be concentrated near pillow selvages. Pillows typically range between 25 cm and ~1m in size.

The Mud Lake Cu-Zn occurrence (Ontario Mineral Inventory, https://www.geologyontario.mines.gov.on.ca/mineral-inventory/MDI000000002310) can be observed in a roadside outcrop located a few hundred metres northwest of STOP 2-7 along the highway. Pyrite, minor chalcopyrite and rare sphalerite are finely disseminated throughout, and adjacent to, a sericitized northeast-trending zone of shearing hosted within chemical metasedimentary rocks interbedded with felsic and intermediate pyroclastic metavolcanic rocks (Brown, 1995; Brown and Fogal, 1995). The mineralization was first uncovered during construction along Highway 102 in the mid-1970s. A grab sample collected in 1975 by staff of the Resident Geologist's office, Thunder Bay, yielded values of 0.24% Cu, 0.87% Zn, 0.12 ounces Ag per ton and 0.005 ounces Au per ton (Fenwick and Scott, 1976).

The felsic metavolcanic rock unit adjacent to the copper- and zinc-mineralized horizon yielded a U-Pb age of 2718±3 Ma (Corfu and Stott, 1998).

A magnetic lamprophyre dyke, ≤ 2m wide, crosscuts the pillowed flows at 065˚-080˚ and dips steeply north. Some of these late Neoarchean intrusions in this area were classified as kersantites (i.e. calc-alkaline, biotite-plagioclase-bearing lamprophyre) by Brown (1995).

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Trip 2 - Geology of the Quetico Subprovince and Shebandowan greenstone belt north of Thunder Bay

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Introduction

This field trip examines the geology of the southern Quetico Subprovince (QS) and its tectonically intercalated contact with the Shebandowan greenstone belt (SGB) north and west of the City of Thunder Bay. Much of the content of this guidebook is informed by a multiyear, 1:50 000 scale bedrock mapping project that is being carried out in the area by the Ontario Geological Survey. The area encompassed by this guidebook represents the southern half of the multiyear bedrock mapping project area (see Figure 1). At the time of this field trip, a new bedrock geology map of the southern half of the project area and associated data are in preparation. Fieldwork on the northern half of the larger project area should be completed during the summer of 2026. In total, the new mapping will cover an area of approximately 4200 km^2 of which approximately 70% had never been mapped at the 1:50 000 scale prior to this work. Some of the results of this bedrock mapping are summarized in interim publications (Metsaranta 2015; Metsaranta and Walker 2019; Metsaranta and Hamilton 2020, Metsaranta and Kamo 2021, Metsaranta 2022, Launay and Metsaranta 2023, Launay and Metsaranta 2024).

The field trip will focus on the Archean geology of the area depicted on Figure 1. Although this is a one-day field trip, we have included 15 stops dispersed over a large area. We will not be able to visit all stops in one day. The order of the stops is organized from west to east in the SGB, followed by a south to north traverse along Highway 527 across the QS. Stops are labelled as "Optional" or "Planned". "Planned" outcrops are the stops we will endeavour to visit during the field trip. Optional stops are included to put into perspective many of the "Planned" stops. As they are easily accessible, participants can visit these "Optional" outcrops on their own. As we are attempting to visit a high number of outcrops over a large area in one day, time spent on each outcrop may be limited. UTM coordinates used throughout the guidebook are NAD 83 Zone 16.

Background Regional Geological Context

The Quetico Subprovince is a vast geological entity that extends, at minimum, from central Minnesota to western Quebec. In "subprovince-style" subdivisions of the Superior Province (e.g. Card and Cielieski 1986, Williams 1991) the QS is bounded to the north by the Wabigoon Subprovince and to the south by the Wawa Subprovince. In more recent subdivisions (e.g. Percival et al. 2006, 2012; Stott et al. 2010) of the Superior Province into "terranes" and "domains" the Quetico Subprovince is commonly referred to as the Quetico basin or Quetico terrane and it is bounded to the south by the Wawa-Abitibi terrane and to the north by the Western and Eastern Wabigoon terranes and the Marmion terrane. In this guidebook, we will refer to the "Quetico" as the Quetico Subprovince (QS) to avoid any interpretive tectonic implications. Similarly, rather than discussing subprovinces or terranes bounding the QS to the south, we will simply refer to the Shebandowan greenstone belt (SGB).

The detailed geology of the QS (as a whole) is somewhat poorly understood. Systematic OGS mapping of large portions of the QS has not been carried out previously at 1:50 000 or 1:20 000 scale. Consequently, accurate bedrock geology maps of much of the QS do not exist, nor do large scale regional geochronology or geochemistry datasets that are tied to geological mapping. General regional geological syntheses of the QS are provided by Percival (1989) and Williams (1991). Additional synoptic descriptions of the QS are included in Percival et al. (2006, 2012) and these include a summary of existing geochronological constraints. Additional influential studies on the metamorphic history of the QS include Pan, Fleet and Heaman (1996); Valli et al. (2004) and a recent PhD study by Rehm (2025) among others.

The QS has historically been interpreted to have been deposited in a fore-arc setting (e.g., Percival 1989, Williams 1991). In reality, the tectonic setting is likely more complex. Geochronology indicates most of the QS was deposited after circa 2700 Ma. However,

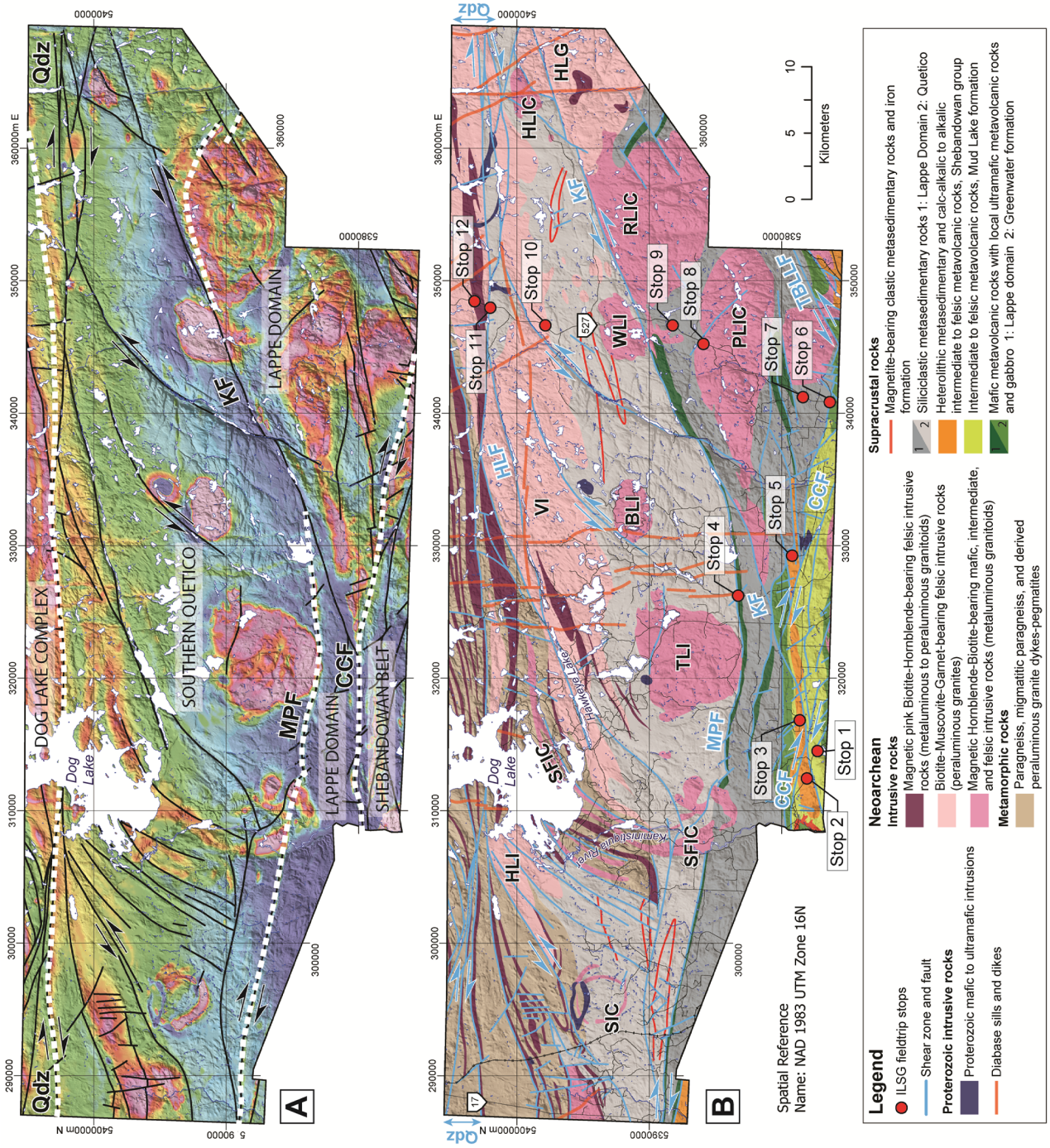


Figure 1 (A) Total magnetic field image of the map area (Ontario Geological Survey 2017), underlain with lidar imagery (Ministry of Natural Resources and Forestry 2023). (B) Geological map (modified from Launay and Metsaranta, 2023) of the field trip area showing the location of stops presented in this guidebook. Note that Stops 13-15 are just to the north of the area portrayed by this map. Geological abbreviations: BLI, Barnum Lake intrusion; CCF, Crayfish Creek fault; CLI; HLG, Hilma Lake granite; HLI, Hadwen Lake intrusion; HLIC, Hades Lake intrusive complex; KF, Kingfisher fault; MFP, Moving Post fault PLIC, Penassen Lakes intrusive complex; QDZ, Quetico deformation zone; RLIC, Roll Lake intrusive complex; SFIC, Silver Falls intrusive complex; SIC, Shabaqua intrusive complex; TBLLF, Thunder Bay–Loon Lake fault.

constraints vary by location (see discussion in Percival et al. 2006; and references therein) with some authors indicating deposition between approximately 2698 Ma and 2696 Ma and others indicating deposition after approximately 2692 Ma. Maximum depositional ages are poorly constrained because of the limited availability of representative and consistent detrital zircon datasets across the Quetico Subprovince, making stratigraphic and tectonic interpretations challenging. A framework for deformation and metamorphism summarized in

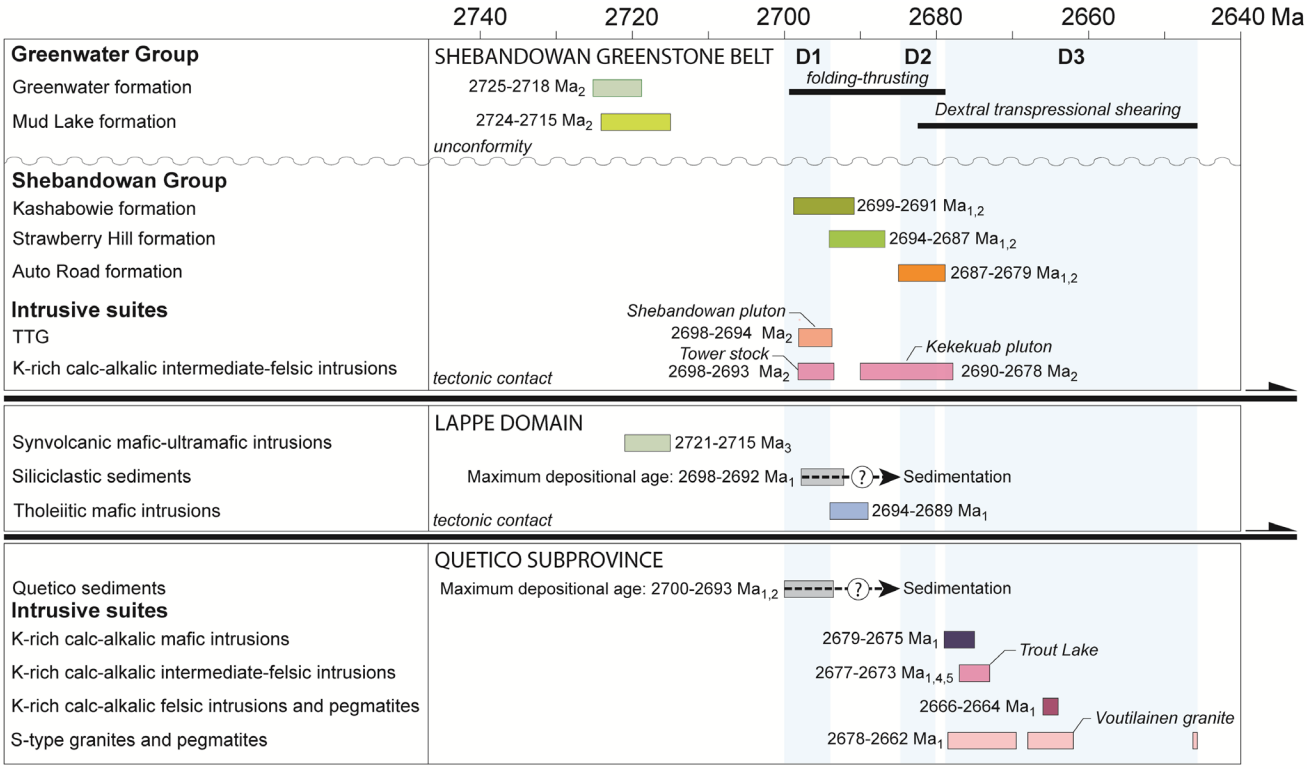
Percival et al. (2006 and references therein) suggests polyphase deformation and metamorphism spanned from close to the time of deposition to approximately 2650 Ma. Williams (1991) described four discrete deformation episodes affecting the QS. Although work on the metamorphic history of the Quetico differs regarding details, most work converges on the prevalence of subprovince-wide high temperature, low pressure metamorphism, which was likely preceded in some places by early medium-pressure and temperature metamorphism. Metamorphic grade increases in the eastern part of the QS where it reaches granulite-facies conditions (see Pan et al. 1998). Intrusive rocks are abundant in the QS, their characteristics are explored in the field guidebook and general characteristics are summarized by Williams (1991). Much attention has been paid to the boundary between the QS and the Wabigoon subprovince (e.g. the Beardmore-Geraldton greenstone belt), however, much less has been paid to the geology of the southern boundary.

The SGB (and correlative greenstone belts in Minnesota) is a relatively narrow, arcuate greenstone belt, that extends from the Pass Lake area (east of Thunder Bay) to Northern Minnesota. The SGB is described in detail in Williams et al. (1991). A more recent geochronology based tectonostratigraphic framework for the SGB was proposed by Corfu and Stott (1998) and this remains in common usage as a stratigraphic and structural framework. Minor modifications to the Corfu and Stott (1998) framework have been added by Lodge (2016). In contrast to the QS, much of the SGB has been mapped by the OGS at 1:20 000 scale. However, much of this mapping was carried out prior to technological advancements like the widespread use of U-Pb geochronology, routine high precision trace element geochemistry, access to lidar imagery and high-resolution airborne magnetic data. That said, the OGS also has an on-going multiyear bedrock mapping project in progress to map much of the eastern part of the SGB at 1:20 000 scale (e.g., Lodge 2014; Ratcliffe 2016, 2017, 2019).

The general greenstone belt-wide tectono-stratigraphic framework for the SGB described by Corfu and Stott (1998) includes *circa* 2720 Ma aged rocks of the Greenwater assemblage (mainly tholeiitic mafic metavolcanic rocks and lesser ultramafic metavolcanic rocks, mafic-ultramafic intrusive rocks and metasedimentary rocks), *circa* 2718 Ma age rocks of the Burchell assemblage (mainly calc-alkalic felsic to intermediate metavolcanic rocks), *circa* 2695 Ma aged rocks of the Kashabowie assemblage (mainly calc-alkalic intermediate metavolcanic and metasedimentary rocks), *circa* 2690Ma aged rocks of the Shebandowan assemblage (calc-alkalic intermediate metavolcanic rocks, shallow marine and fluvial metasedimentary rocks) and younger than *circa* 2682 Ma aged rocks of the Auto Road assemblage (conglomerate). Corfu and Stott (1998) envisaged a structural history of D_1 thrusting that interleaved the Greenwater assemblage along with the Burchell assemblage with the Kashabowie assemblage followed by a regional unconformity overlain by younger rocks of the Shebandowan and Auto Road assemblages deposited during D_2 transpression. D_2 transpression ceased by about 2680 Ma. According to the framework of Corfu and Stott (1998) intrusive rocks in the SGB include older gneissic tonalitic rocks to the south of the belt with ages as old as approximately 2750Ma, syn-Greenwater assemblage mafic-ultramafic intrusions, syn-Kashabowie assemblage tonalite, syn-Shebandowan assemblage monzodiorite-granite (Tower stock) and post-tectonic *circa* 2680 Ma aged intrusions of biotite-hornblende bearing diorite to granodiorite. Locally, evidence for magmatic rocks with ages around 2710 Ma are present in some parts of the SGB, e.g. Kabaigon porphyry (Corfu and Stott 1998).

Geology of field trip area

This fieldtrip will examine the geology of four distinct geological domains (*see* Figure 1): 1) the northeastern part of the Shebandowan greenstone belt, 2) the Lappe domain, 3) the southern Quetico domain and 4) the Dog Lake injection complex (Figure 1). Figure 2 is a geological timeline summarizing important events affecting the different geological units in the QS-SGB boundary zone and southern QS compiled from unpublished OGS data and various other sources. Synoptic reviews of these domains are given below. Additional details are provided in the field trip stop descriptions and will be augmented by discussions in the field.



1. Launay and Metsaranta in prep; 2. Corfu and Stott, 1998; 3. Corfu., 2000; 4. Wang et al., 2020; 5. Kamo, 2013

Figure 2 Geological timeline summarizing the main volcanic, sedimentary, intrusive, and structural events affecting the Shebandowan greenstone belt and the Quetico Subprovince. Ages are compiled from Corfu and Stott, 1998; Corfu, 2000; Kamo, 2013; Wang et al., 2020 and preliminary OGS data. Age bars include analytical uncertainties.

Shebandowan greenstone belt (Stops 1, 2, 3, 5 and 6)

This field trip guidebook only examines a narrow portion of the northern margin of the SGB that was covered by our mapping (Figure 1). In this area, the northern boundary of the SGB is generally east-striking dextral shear zone interpreted to be the eastward extension of the Crayfish Creek fault. (Figure 1). Within this area of the SGB, we recognize four mappable units at the 1:50 000 scale which are subdivided into two informal groups, each comprising two informal formations. The older Greenwater group (a less repetitive Group-level name should be devised) consists of the Greenwater and Mud Lake formations and the younger Shebandowan group consists of the Strawberry Hill and Auto Road formations. Although not specifically identified in our area of mapping, we would include the "Kashabowie assemblage" (e.g. Corfu and Stott 1998) with the Shebandowan group. These units correspond with the "older" and "younger" portions of the SGB as described by previous workers (e.g., Corfu and Stott 1998, Lodge 2016) in most respects. However, we feel that moving towards a "sub-assemblage level" nomenclature is warranted to begin a framework for more detailed characterization of supracrustal rock variability at the regional scale. This is particularly true for rocks of the Shebandowan assemblage (in the sense defined by previous workers) which is lithologically heterogeneous.

Greenwater group

The Greenwater formation (Stops 1 and 6) consists mainly of tholeiitic mafic metavolcanic rocks, minor ultramafic metavolcanic rocks, synvolcanic gabbroic intrusions and minor clastic and chemical metasedimentary rocks. We do not have specific age constraints on the Greenwater formation; however, we infer that it is part of the "older" SGB based on lithological similarities with rocks of the Greenwater assemblage *sensu stricto*. At more detailed mapping scales, the Greenwater formation could likely be further subdivided (e.g. mafic dominated vs ultramafic dominated portions).

The Mud Lake formation (Stop 1) consists mainly of calc-alkalic fragmental volcaniclastic rocks and locally coherent flows of felsic to intermediate composition. In the area, Corfu and Stott (1998) determined the age of this unit to be 2718 +/- 3 Ma. These are likely equivalent to Burchell assemblage of Lodge (2016).

Shebandowan Group

The Strawberry Hill formation (SHF, Stop 2) typically consists of coarse-grained polymictic breccias characterized by hornblende phenocrysts in a dark matrix. The SHF is typically mafic to intermediate and calc-alkalic. It typically has elevated magnetic susceptibility and displays pink to red hematite alteration and/or pale green epidote alteration. Locally, it also occurs as more massive, hornblende-phyric mafic flows or shallow intrusions. At some localities, breccias or flows are associated with thinly layered tuffs, and reworked tuffs with clear pyroclastic textures such as bombs that deform layering. Geochronology by Corfu and Stott (1998) and our own work indicate deposition/eruption of the SHF at *circa* 2690 Ma. Although the bulk of SHF rocks are undeformed, they are locally cut by narrow ductile shear zones, and it is in sheared contact with older rocks. The SHF forms a thin, but important marker unit that can be correlated across much of the central SGB in the area depicted by Figure 1. Breccias of the SHF are compositionally similar to, and of the same age as the Tower stock (just west of the area depicted in Figure 1) which hosts low-grade disseminated, intrusion related gold mineralization and includes marginal breccias similar to the SHF (e.g. Carter 1992).

The Auto Road formation (ARF, Stops 3 and 5) comprises mainly polymictic, matrix supported conglomerate. Local occurrences of trough cross-stratified, likely fluvial sandstones, are also considered part of the ARF. Calc-alkalic mafic metavolcanic rocks, including apparently pillowed flows are locally intercalated with ARF sedimentary rocks. The ARF was deposited after 2682 ± 3 Ma based on geochronology in Corfu and Stott (1998) and as such, it clearly postdates the SHF.

Dextral shear zones are a common feature of the SGB. Although it is not clear in Figure 1, geophysical patterns in adjacent areas like the LD and the SGB outside of our mapping area suggest that D_2 shear zones post date D_1 thrust faults. Sinistral northeast-trending shear zones in the SGB such as the Kingfisher and Thunder Bay-Loon Lake faults are relatively younger than the dextral shear zones based on mapping inferred off-sets.

Potassium-rich calc-alkalic suite intrusions (PRCAS) form a minor component of the SGB as shown on Figure 1. These include massive to weakly foliated hornblende-biotite-magnetite quartz monzonite to monzogranite dominated intrusions of unknown age. These may be related to similar intrusions in the SGB like the Kekekaub pluton (*circa* 2680 Ma) or perhaps correlate with the Tower stock (*circa* 2690Ma).

Lappe Domain (Stops 4, 7 and 9)

The Lappe domain comprises mainly metasedimentary rocks (wacke and siltstone) similar to those of the southern Quetico subprovince to the north. Its southern boundary is the Crayfish creek fault whereas it is bounded to the north by the Moving Post fault. The LD is characterized by thin fault bounded panels of mafic metavolcanic and mafic intrusive rocks comparable to the Greenwater formation intercalated with the metasedimentary rocks. Where best preserved, these mafic panels contain pillowed mafic flows, local banded iron formations, local thin ultramafic schists (sheared flows or thin sills) and local sulfidic mudstones. The margins of the mafic panels are commonly sheared and often preserve well developed steeply plunging stretching lineations indicating dip-slip, probable thrust motion. Locally LD rocks are folded, however we do not have a sufficient coverage of detailed younging data or outcrop scale fold observations to determine the nature of folding in the LD.

Although we do not have direct age constraints on metavolcanic rocks in the mafic panels, preliminary data suggests some thin gabbro bodies in LD metasedimentary rocks are intrusive (i.e. not structurally interleaved). These provide minimum age constraints for LD sedimentation; combined with preliminary detrital zircon data, and considering analytical uncertainties, LD sedimentation is bracketed between about 2698 and 2689 Ma. At another locality, Corfu (2000) determined and age of *circa* 2718 Ma for a gabbro within one of the mafic panels in Ware township. If this age is reliable, then some of the mafic rocks in the LD correlate with the Greenwater formation. Another hypothesis to consider is that this age could reflect zircon inheritance. Regardless of the age of the mafic panels and whether they represent tectonic slivers of older rocks, or if they are part of the "stratigraphy", observed thrusts faults indicate that the boundary between the SGB and the QS likely represents a zone of fold-thrust deformation. The timing of LD sedimentation corresponds with the timing of deposition of the Kashabowie assemblage in the SGB (*see* Corfu and Stott 1998). Observed thrust

faults in the LD may therefore correspond to "D_1" of Corfu and Stott (1998) and this interpretation links the timing of early QS deformation with early deformation in the SGB.

East of the Kingfisher fault, large, multiphase, PRCAS intrusive complexes, (the Penassen Lakes intrusive complex and the Roll Lake intrusive complex) were emplaced into the LD. These intrusions consist of an early mafic phase (hornblendite to monzogabbro) roughly coeval with intermediate a hornblende-biotite-magnetite monzodiorite to quartz monzonite phase and a late phase that is volumetrically dominant and composed of biotite monzogranite. Compositionally similar rocks are found to the north in the QS and the Dog Lake injection complex and ages determined for the different phases are consistent regionally as summarized in Figure 2. Locally, peraluminous granitic pegmatites occur in the northern LD (e.g. Walkinshaw pegmatites) but are not associated with any obvious parental granite.

LD metasedimentary rocks are typically low metamorphic grade, dominated by biotite or chlorite, quartz, plagioclase assemblages. However, near intrusions they locally contain contact metamorphic porphyroblasts of andalusite and/or cordierite and in some areas experienced partial melting.

Southern Quetico domain (Stops 10, 11 and 12)

The southern Quetico domain comprises mainly metasedimentary rocks (wacke and minor siltstone) with rare intermediate tuffaceous horizons, rare mafic tuff (possibly boninitic). Bedding and foliations in the southern QS are typically east to northeast trending and "D_2" folds typically have east to northeast trending axial surfaces. Southern QS metasedimentary rocks gradually become more recrystallized towards the north. Metamorphic assemblages are typically dominated by biotite and locally biotite-garnet. Andalusite- and cordierite-bearing metamorphic assemblages are also relatively common and may related to proximity to intrusions. Staurolite-bearing assemblages are present locally but rare.

The southern QS was intruded by numerous potassium-rich calc-alkalic suite intrusions (PRCAS). These are depicted on Figure 1 and include from west to east, the Shabaqua intrusive complex, the Silver Falls intrusive complex, the Trout Lake intrusion, the Barnum Lake intrusion, the Whitelily Lake intrusive complex and the Hades Lake intrusive complex. These intrusions and intrusive complexes are variably complex mixtures of mafic (hornblendite-monzogabbro), intermediate (monzodiorite-quartz monzonite) and felsic (monzogranite) phases. S-type granites are also common in the southern QS, these include the Hilma Lake granite, the Voutilainen intrusion and the Hadwen Lake intrusion. Peraluminous granitic pegmatites are also abundant and spatially related to S-type granite bodies.

Dog Lake injection complex and Quetico deformation zone (Stops 12, 13, 14 and 15)

Strain and degree of metasedimentary rock recrystallization increase abruptly in the vicinity of the Quetico deformation zone (QDZ) in the northern part of the area. In this domain, QS metasedimentary rocks are recrystallized and comprised mainly of biotite-quartz-feldspar+/- magnetite paragneiss and locally also sillimanite-cordierite-garnet bearing paragneiss. The Dog Lake injection complex (DLIC) is characterized by paragneiss intruded by a high volume of both peraluminous and HPCAS intrusions (injections) that were emplaced synchronously with intense dextral transpression along the QDZ. Intrusions of both suites are commonly schlieric with strong fabrics defined by schlieren and magmatic minerals. At the contact with HPCAS intrusions, paragneisses are commonly strongly magnetic which likely resulted from their oxidation by fluids exsolved from these HPCAS intrusion triggering the crystallisation of magnetite. These zones are also particularly rich in biotite which locally give the paragneisses the appearance of melanosome. Migmatitic rocks are a volumetrically minor component of the DLIC and comprise mainly patchy metatexite likely related to heating by the high volume of intrusive rocks in the area.

Syn-tectonic fabrics are ubiquitous in intrusive rocks of the DLIC. Strong fabrics are generally steep and east striking to east-northeast striking. Dextral shear bands with strikes of approximately N290-300 are common as are approximately N40 striking and N15-20 striking subvertical sinistral shear bands. C-S fabrics are common and suggest syn-emplacement dextral strike slip and locally north side up thrust components of motion. Shearing related fabrics in the paragneisses have the same orientations and kinematics and syn-tectonic emplacement fabrics in the granitoid rocks. Folding of gneissosity is common and folding

tends to be gently plunging. Intense, steeply dipping, east striking mylonitic zones are present locally. Late brittle-ductile strike slip and thrust motion occurred along the QDZ locally overprinting ductile fabrics. This pattern is repeated at map scale where the sinistral north to northeast-trending shear zones branches onto dextral east-trending QDZ in the Dog Lake area (*see* Figure 1). These observations suggest a transition from an earlier ductile transpressive deformation to later more brittle-ductile deformation.

Field trip stop descriptions

Stop 1 (Optional) - Mafic metavolcanic rocks of the Greenwater formation and felsic to intermediate metavolcanic rocks of the Mud Lake formation

315073E 5376695N (Greenwater formation)

314610E 5377220N (Mud Lake formation)

Park on the shoulder of Mud Lake Road. Wear reflective vests, stay on shoulder or in the ditch. Traffic is heavy on Hwy 102 so be very cautious crossing the highway.

A series of outcrops in this area shows mafic metavolcanic rocks considered to be part of the Greenwater formation and felsic metavolcanic rocks of the Mud Lake formation. These two formations are typical of the older part of the Shebandowan greenstone belt. From the junction of Hwy 102 and Mud Lake Road, outcrops immediately to the east and west are mainly mafic metavolcanic rocks of the Greenwater formation. These rocks display well preserved volcanic features such as prominent pillows (Figure 3A) and local pillow breccias. The pillowed flows are subvertical and striking to N75. Based on pillow cusps, they appear to young northward. Quartz-epidote veining, local red-pinkish alteration and east striking brittle-ductile shear zones can also be seen at this outcrop. Younger, mica-phyric mafic lamprophyre dikes are also present.

Farther west, closer to the north end of Mokomon Lake (314610E, 5377220N) felsic metavolcanic rocks of the Mud Lake formation comprising tuff breccias, lapilli tuffs and locally flows are present in outcrops located on the north side of the highway. These felsic metavolcanic rocks host a thin, semi-massive sulfide (sphalerite, pyrite, pyrrhotite) horizon known as the Mud Lake VMS occurrence (Figure 3B). The rocks

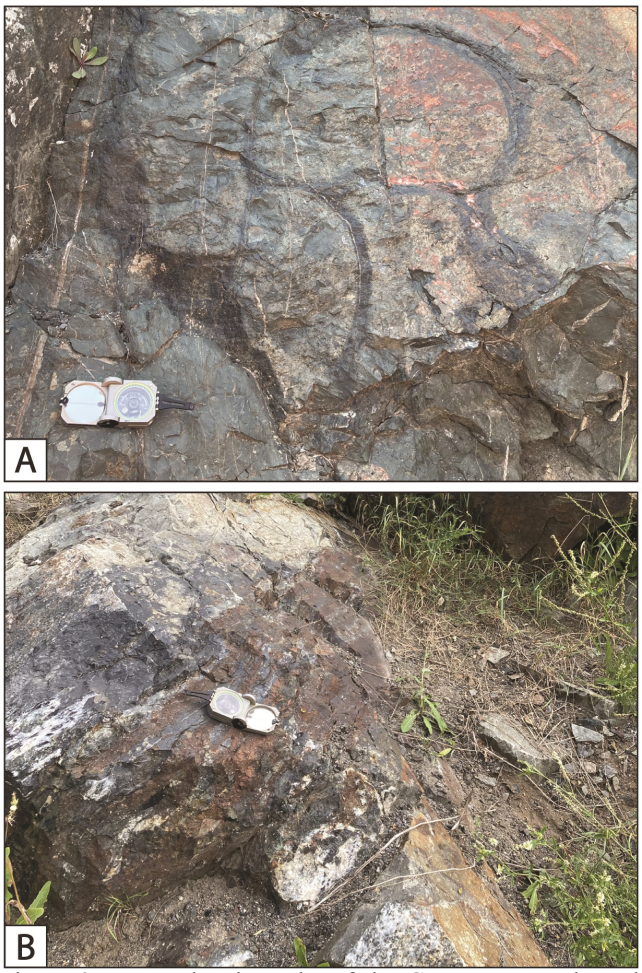


Figure 3. Metavolcanic rocks of the Greenwater and Mud Lake formations. A) Large pillows in mafic flow, Greenwater formation. B) Sulfide mineralization in felsic metavolcanic rocks, Mud Lake formation.

here are strongly foliated (260/75) with local shearing (234/80) and locally cut by rusty shallow dipping faults. Felsic rocks from this outcrop have an age of 2718 +/- 3 Ma based on Corfu and Stott (1998). The age of Greenwater formation (Greenwater assemblage) rocks throughout the greenstone belt is mainly inferred from adjacent felsic to intermediate units and mafic-ultramafic intrusions. Although the contact between these two units appears sharp at this location, mafic rocks of the Greenwater formation are intercalated with felsic to intermediate rocks of the Mud Lake formation in other locations, suggesting that the contact may have originally been gradational.

Stop 2 (planned) - Strawberry Hill formation, Shebandowan group

312462E 5378109N

Park along the access road to the quarry on the southern side of Dawson Road. Participants must always wear reflective vests and remain on the road shoulder, as traffic along this section is relatively dense and driver visibility is poor.

This stop exposes a representative outcrop of the Strawberry Hill formation of the Shebandowan group. It consists of a massive, undeformed, mafic to intermediate calcalkaline, hornblende-phyric, matrix-supported breccia (Figure 4A). The breccia is polymictic, containing angular to subrounded clasts composed predominantly of plagioclase-phyric, medium- to coarse-grained pink monzonite, along with subordinate clasts of mafic and intermediate volcanic rocks, all set within a fine-grained, dark-green matrix.

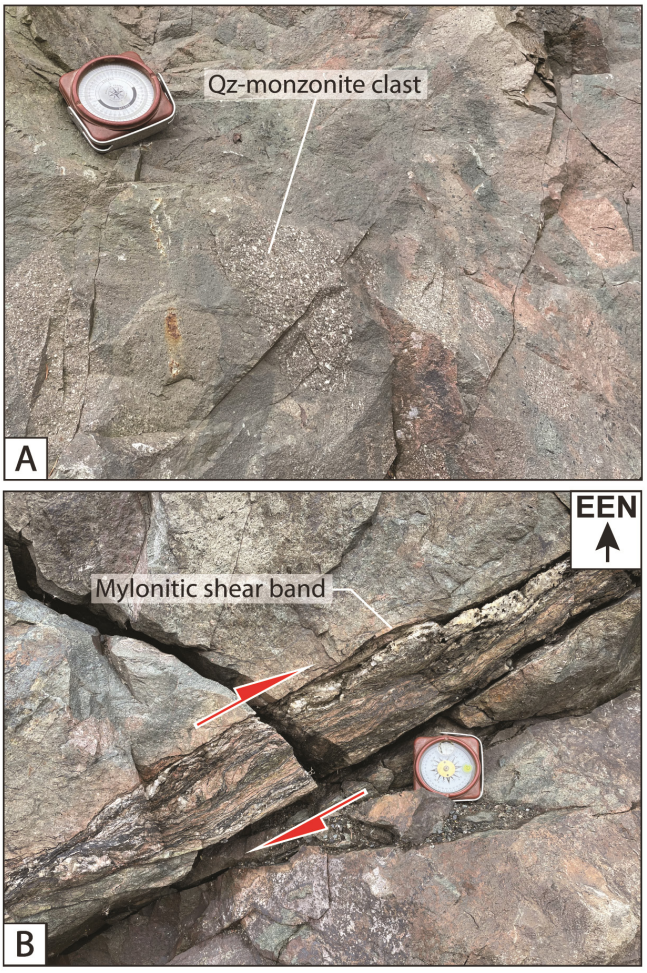


Figure 4 Representative photographs of the Strawberry Hill formation outcrop (Stop 2). (A) Massive poorly sorted matrix supported polymictic intermediate breccia. (B) Discrete mylonitic shear band with asymmetric kinematic indicators (C-S fabric) indicating a north-side-up sense of shearing.

Clasts of monzonitic composition locally preserve an internal magmatic foliation, highlighted by the alignment of plagioclase phenocrysts. This magmatic fabric suggests that the intrusion from which the clasts were derived was likely emplaced syn-tectonically. The absence of visible bedding, combined with the generally angular to sub-rounded nature of the clasts, suggests formation in a high-energy volcanic environment, possibly associated with a cryptodome, and likely proximal to the volcanic source. Alternatively, this unit may represent a subvolcanic magmatic breccia, comparable to the breccias around the Tower stock described by Carter (1992).

Based on TIMS U–Pb ages reported by Corfu and Stott (1998), magmatism related to this breccia likely occurred at approximately 2692±6 Ma. Thus, the SHF is likely contemporaneous with the 2690 ± 3 Ma Tower stock (Corfu and Stott 1998)

Although the unit is largely massive and lacks visible foliation, discrete mylonitic shear bands are locally present (Figure 4B). These shear bands are commonly associated with carbonate alteration and quartz–carbonate veining. The shear zones generally strike northeast and locally preserve kinematic indicators consistent with a thrust motion, indicating a northsideup sense of shearing (Figure 4B).

The breccia is also cut by multiple generations of quartz–carbonate veins, indicating postdepositional brittle deformation coeval with hydrothermal activity. These veins locally offset both clasts and matrix but do not significantly disrupt the overall massive character of the breccia.

Stop 3 (Planned) - Auto Road formation, Shebandowan group

316851E 5378641N

Park at the entrance of the private dirt road south of Korpela Road. As Korpela Road is narrow, please ensure that your vehicle does not obstruct traffic or restrict access along the road.

This stop exposes a representative outcrop of the Auto Road formation of the Shebandowan group. It consists of a strongly foliated, poorly sorted, matrixsupported polymictic conglomerate. The conglomerate contains predominantly pebble to boulder-sized clasts of plagioclasephyric, medium to coarse-grained pink monzonite and monzogranite, many of which are

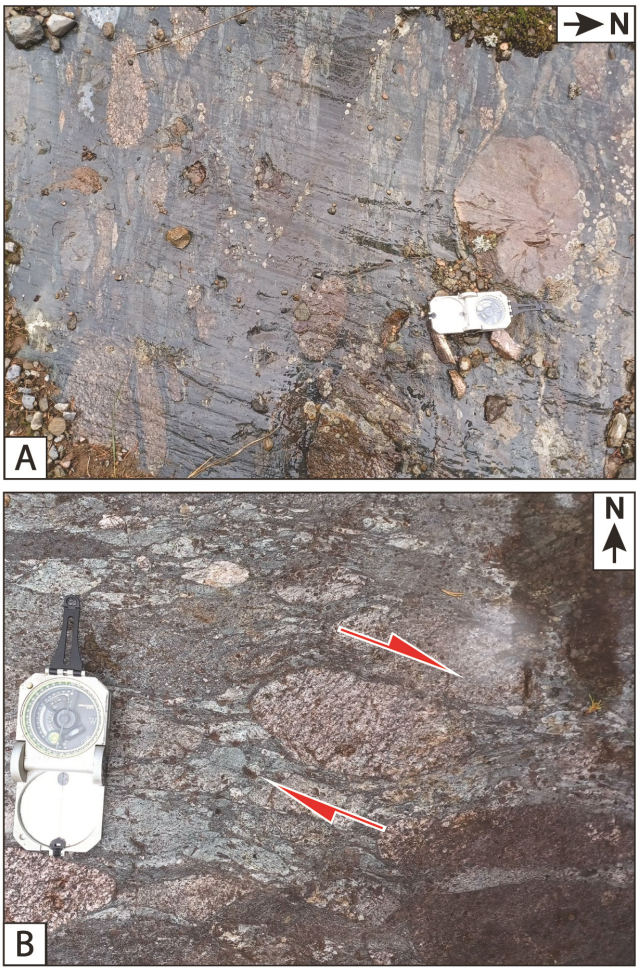


Figure 5 Representative photographs of the Auto Road formation outcrop (Stop 3). (A) Strongly foliated polymictic conglomerate. (B) Dextral asymmetrical C-S fabric wrapping around a clast of pink monzogranite. Note the discordant internal foliation within the clast.

strongly flattened parallel to the foliation (Figure 5A). The matrix is sandy, medium to coarse-grained, and characterized by a darkgreen color.

Several monzonitic and monzogranitic clasts preserve an internal foliation that is discordant with the matrix foliation (Figure 5B), indicating that these intrusive rocks were deformed prior to erosion and deposition. This relationship suggests a preAuto Road formation deformation event affecting the source intrusions.

The conglomerate exhibits a strong east–west trending foliation and is overprinted by a northwest-trending (approximately N300°) dextral shearing, highlighted by the development of asymmetric C–S fabrics wrapping around the clasts (Figure 5B).

Preliminary ages obtained from a foliated intrusive clast constrain the maximum age of the preAuto Road formation deformation event to approximately 2688 ± 0.8 Ma, whereas TIMS U–Pb ages reported by Corfu and Stott (1998) indicate a maximum depositional age of 2682 ± 2 Ma for the conglomerate. Together, these ages suggest that the intrusions from which the clasts were derived were deformed after 2688 Ma, prior to erosion and deposition of the conglomerate sometime after 2682 Ma. The conglomerate was subsequently deformed and affected by dextral shearing likely related to the Crayfish Creek fault.

Stop 4a (Planned) - Sheared mafic rocks, medium-bedded wackes and thrust deformation at the Northern boundary of Lappe domain.

326238E/ 5383267N

Park along Moving Post Road or in the parking lot of the old Lappe Store if permission is obtained. Wear reflective vests, be mindful of traffic, shoulders of the road are narrow and visibility is poor.

This outcrop illustrates strongly deformed, tholeiitic mafic metavolcanic rocks (Figure 6A) of the northernmost mafic metavolcanic unit of the Lappe domain. Mafic metavolcanic rocks at this exposure are characterized by N250 striking north-dipping strong foliation and well-developed northward plunging lineation. The apparent dip-slip motion along the shear zone is north-side-down. In its present geometry, the true kinematics of the dip-slip motion along this structure is equivocal (Figure 6B). However, an interesting and commonly repeating pattern along strike is that metamorphosed north dipping wackes located north and south of this mafic metavolcanic unit consistently young southward indicating that the whole stratigraphic succession is overturned. This pattern was also observed in the past OGS mapping campaign of MacDonald (1939, observe printed version of old map). If our interpretation of the kinematics of this structure and the facing direction in the bounding metagreywacke units are correct, a possible interpretation of this structure is that it represents an originally northward verging thrust that was subsequently steepened and overturned.

The stratigraphic relationship between the mafic metavolcanic rocks at this locality and surrounding wackes is not well constrained. Commonly both contacts of the mafic unit are sheared and there is a paucity of rocks suitable for geochronology in the exposures that we have mapped. Reliable younging

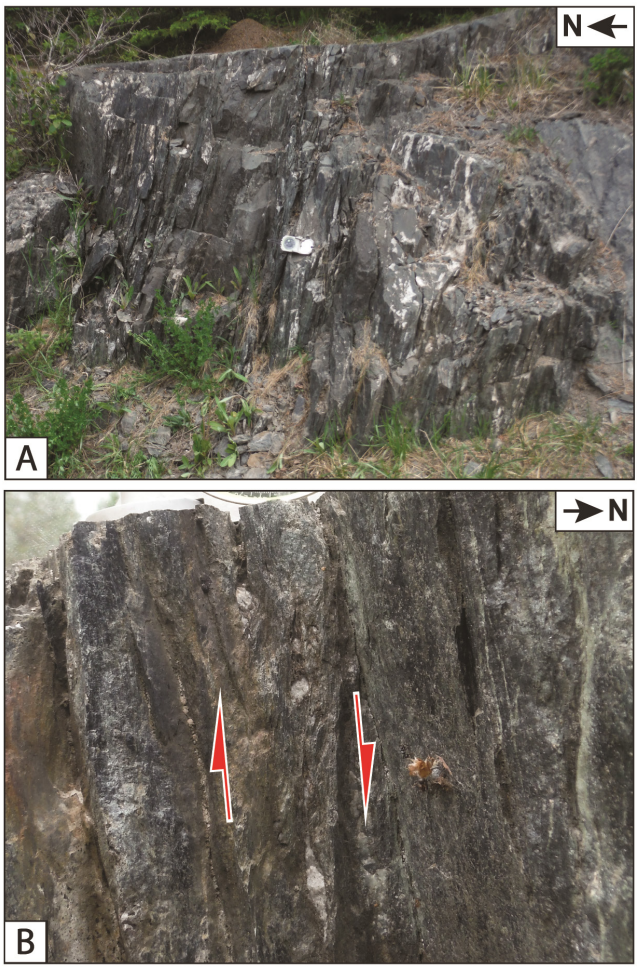


Figure 6 Representative photographs of sheared mafic volcanic flow from the Lappe Domain (Stop 4). (A) Strongly foliated and sheared mafic volcanic flow. (B) C-S fabric wrapping around quartz eyes indicating north-side down sense of shearing

indicators within the mafic metavolcanic unit are almost nonexistent. At one locality, farher to the east, southward younging pillows were observed along a similar structure in a similar setting. We have not observed strong evidence of stratigraphic continuity between the surrounding wackes and the mafic unit. In this case, observed younging directions do not argue against stratigraphic continuity, however the sheared nature of contacts makes interpretation difficult.

A thin, mica-phyric lamprophyre dike is also present at this locality. Mafic lamprophyre dikes are common throughout the area in the SGB, LD and QS. They have not been successfully dated and their contact relationships and relationships to structures and other intrusive suites is difficult to interpret as in many places relative age relationships are contradictory. This may indicate multiple generations of lamprophyric mafic intrusion are present regionally.

Stop 4b (Optional) - Southward younging Lappe domain metasedimentary rocks

326207E 5382509N

This optional outcrop is located approximately 750 m south of Stop 4a along Dog Lake Road. It exposes mediumbedded wackes displaying a southward younging direction, as indicated by graded bedding and load casts (Figure 7A and 7B). The southward younging of these clastic metasedimentary rocks occurs near the mafic metavolcanic rocks and associated thrust fault observed at Stop 4a, providing important constraints on local stratigraphic facing and structural relationships.

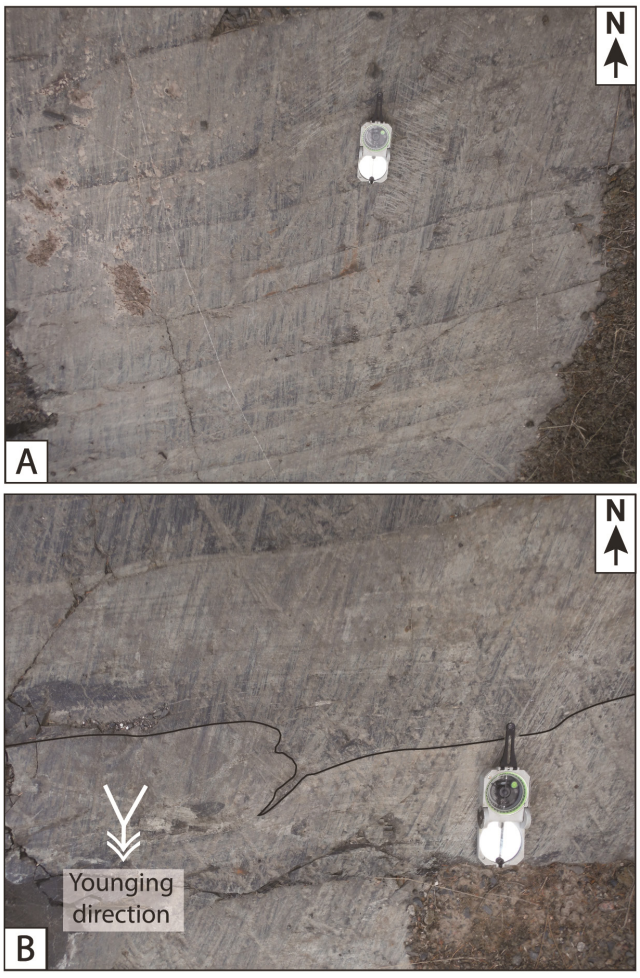


Figure 7 Clastic sedimentary rocks of the Lappe Domain (Stop 4b). (A) Medium bedded wackes with graded beds. (B) Flame structure showing a southward younging direction.

Stop 5 (Optional) - Cross bedded sandstone, Auto Road formation, Shebandowan group

329269E 5379186N

Park along the south side of Peterson Road, near the entrance to the private residential access road located on the curve. This outcrop is situated on private property. Authorization from the landowner is required prior to accessing the outcrop, and participants must ensure that permission has been obtained before entering on the property.

This large outcrop exposes a well-preserved section of sandstone in sheared contact with a calc-alkaline mafic volcanic flow (Figure 8A). The sandstone displays a variety of well-developed sedimentary structures, including crossbedding and channelized geometries, indicative of a fluvial depositional environment (Figure 8B). The sandstone locally contains plagioclase phenocrysts, suggesting a potential juvenile volcaniclastic component.

The mafic lava flow is locally pillowed and occurs in sheared contact with the sandstone (Figure 8C). The shear zone is characterized by a penetrative east–west-striking foliation that dips steeply to the south. A well-developed stretching lineation, plunging steeply (~55°) to the east, is observed on foliation planes. Locally, kinematic indicators are preserved and indicate a dextral sense of shearing. Together with the steeply plunging lineation, these observations suggest a dextral transpressional deformation with a top-to-the northwest thrust component.

Younging directions determined from cross bedding indicate a consistent southward younging across the outcrop. The sandstone is also crosscut by late, narrow mafic dikes, indicating post-depositional magmatic activity (Figure 8D).

Although samples collected for U–Pb geochronology

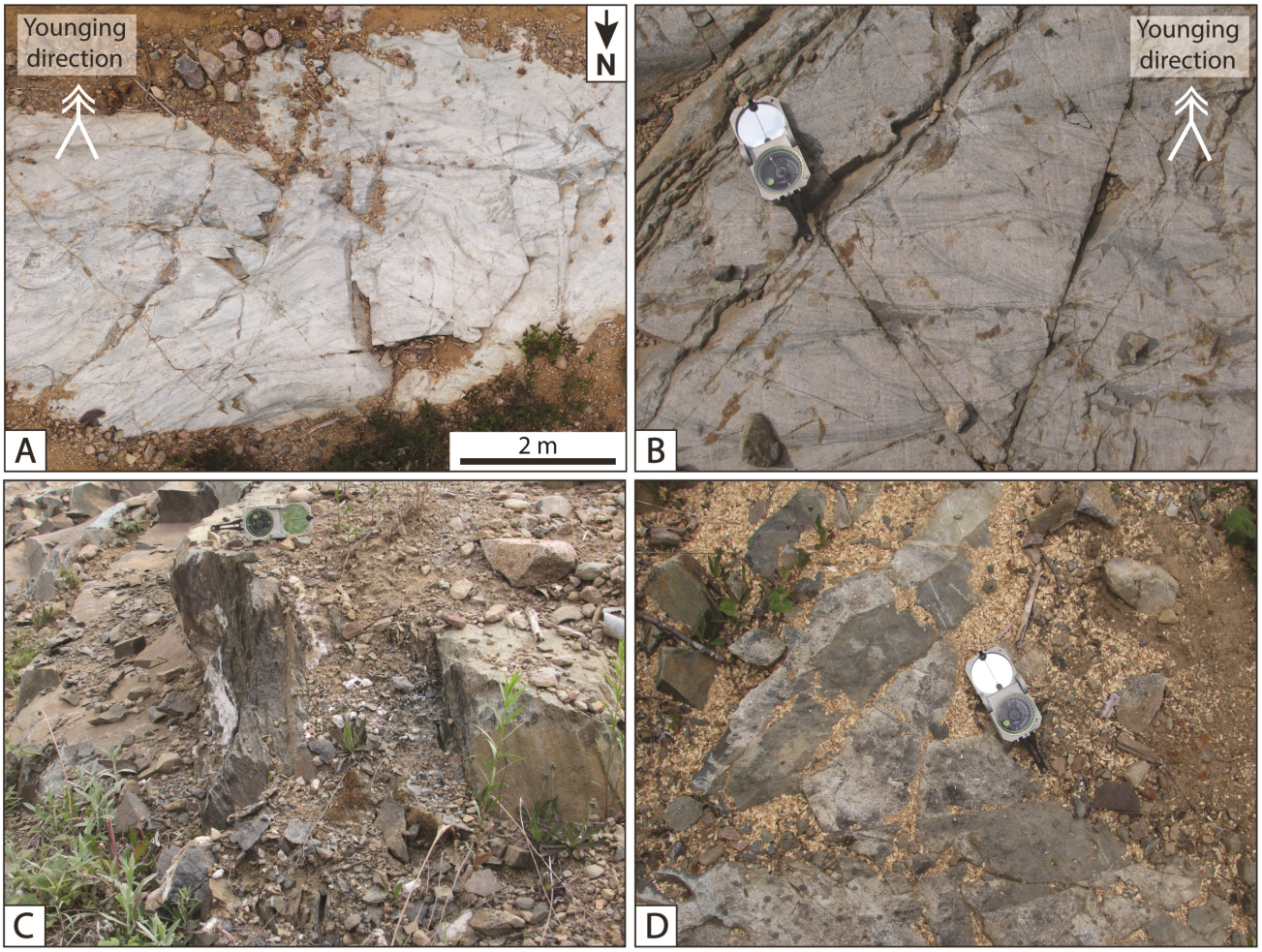


Figure 8 Outcrop of cross-bedded sandstone of the Auto Road formation (Stop 5). (A) Aerial drone photograph showing crossbedding and channel structures. (B) Close up on cross-stratified sandstone with southward younging direction. (C) Strongly sheared and foliated mafic volcanic flow occurring within the sandstone. (D) Late mafic dikes crosscutting sandstone.

did not yield datable mineral phases, the sandstone is interpreted to be part of the Shebandowan group, most likely the Auto Road formation, based on its characteristic fluvial depositional environment.

Stop 6 (Planned) - Greenwater assemblage mafic metavolcanic rocks, eastern extension of Crayfish Creek deformation zone

340868E 5376210N

Park on the shoulder of Mount Baldy Road at its junction with Hwy 527. Wear high visibility vests. The shoulder here is narrow and logging truck traffic can be heavy. Be careful if crossing the highway. Footing is uneven and there are commonly garbage and glass in the ditches.

This outcrop represents the sheared contact between the Shebandowan greenstone belt (to the south) and the Lappe domain (to the north) and may represent the eastward extension of the Crayfish Creek deformation zone.

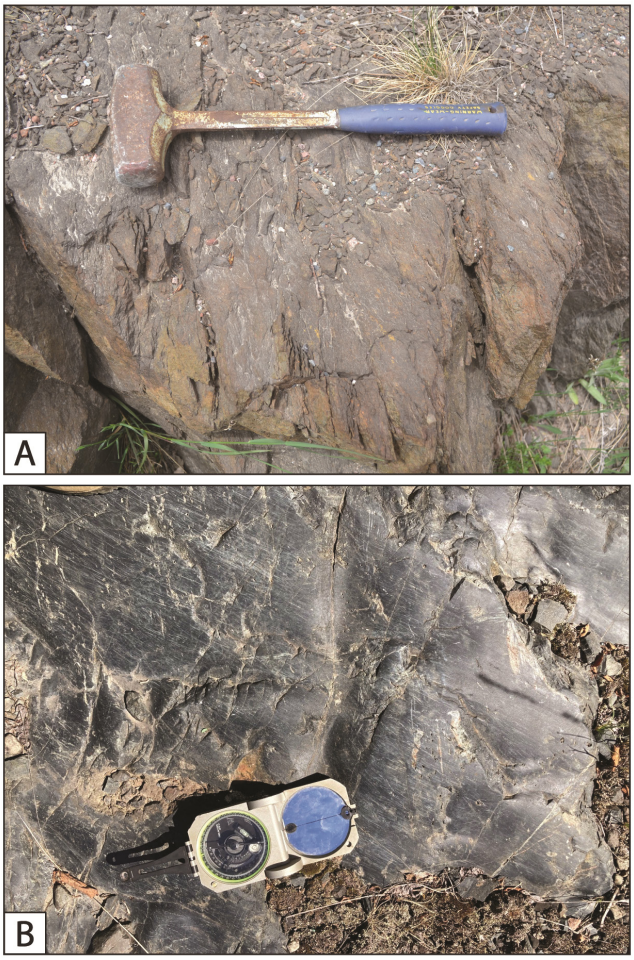


Figure 9. Greenwater formation metavolcanic rocks (Stop 6). (A) ankerite altered mafic-ultrmafic schist. (B) Massive, plagioclase-phyric mafic flow.

The northern part of the outcrop consists mainly of east-striking, steeply south-dipping, sheared, ankerite altered, magnesium-rich mafic rocks (Figure 9A). During mapping, kinematics of the shearing at this locality were not determined confidently. Precise identification of protoliths in the northern part of the outcrop is problematic as most primary features were obliterated. However, towards the south, rock types are well preserved and include massive, fine- to medium-grained mafic volcanic flows (Figure 9B), mafic pillowed flows, a thin pyrite-bearing nodule black mudstone, and quartz-feldspar porphyritic intermediate dikes. Enigmatic weakly boudinaged dikes of mafic to ultramafic composition locally cut the main shearing fabric in the northern part of the outcrop. Minor Proterozoic calcite veins are also present.

Lappe domain metasedimentary rocks to the north were deposited after *circa* 2700 Ma and perhaps after *circa* 2690 Ma, depending on the interpretation of preliminary detrital zircon data. Based on geochronology performed elsewhere in the SGB, the Greenwater formation is inferred here to have an age of *circa* 2720 Ma. Therefore, this shear zone juxtaposes rocks that differ in age by at least 20 million years. At this locality, shearing could represent a transpressive dextral reactivation of an earlier shear zone that interleaved units of disparate age. Note that there does not appear to be a rhyolitic unit equivalent to the Mud Lake formation north of the Greenwater formation as seen in Stop 1. This feature could be attributable to either a fault-related subtraction or a lateral stratigraphic discontinuity.

The "QFP" dikes cutting the Greenwater formation here have the same appearance as dikes commonly observed in the Lappe domain and in the southern Quetico. Unfortunately, these dikes have proven difficult to date due to the lack of mineral phases amenable to U-Pb geochronology. Pyritic black shales like those at this locality occur locally in the Greenwater and Mud Lake formations.

Stop 7 (Optional) - Lappe domain metasedimentary rocks

341256E 5378262N

Park at the "weigh-scale" on the west side of Highway 527. Wear high visibility vests and use a high level of caution crossing the highway to the outcrop. Ditches also have uneven footing and garbage.

This outcrop illustrates low metamorphic grade rocks of the southern Lappe domain. Here ~ 250° striking, north dipping but southward younging metasedimentary rocks include thin- to medium-bedded intercalated siltstone and sandstone (Figure 10A) overlain by very thickly bedded poorly sorted volcaniclastic sandstone (Figure 10B). Near the base of the thick sandstone bed, there is a folded horizon of thinly interbedded sandstone and siltstone. This may represent syn-depositional soft deformation. Locally, sparse carbonate nodules are present in the outcrop; these have been interpreted as early diagenetic features. Farther north, calc-silicate nodules (amphibole, epidote, locally garnet) are common in the QS and probably represent more metamorphosed equivalents of these carbonate nodules.

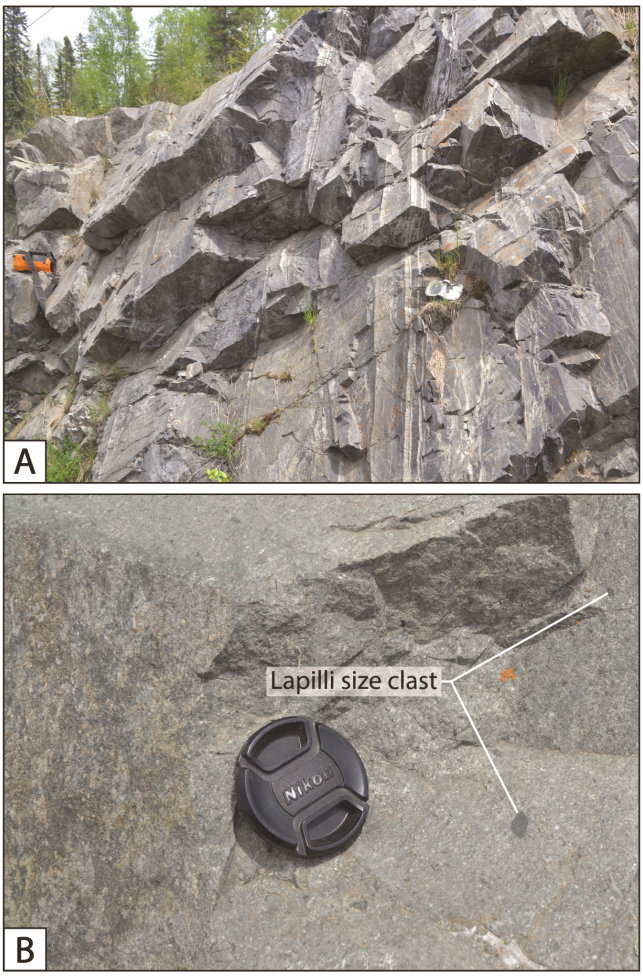


Figure 10. Lappe domain metasedimentary rocks at Stop 7. (A) steeply dipping, southward younging, thin- to medium-bedded, low metamorphic grade siltstone and sandstone. (B) Poorly sorted, lapilli and intraformational-sedimentary-clast bearing thick-bedded volcaniclastic sandstone.

As alluded to in the description of Stop 6, Laser-ablation-ICP-MS zircon geochronology was carried out on samples from this outcrop. This data suggest deposition of these rocks about the same time as much of the Shebandowan group. However, analytical precision does not permit precise chronostratigraphic correlation with the Strawberry Hill or Auto Road formations. Maximum depositional ages for tidally influenced shallow marine sediments in the Finmark area have maximum depositional ages of about 2691 Ma based on limited population, single crystal ID-TIMS geochronology reported by Corfu and Stott (1998). The Finmark metasedimentary rocks occur in a similar structural setting based on geophysical interpretation.

Stop 8 (Optional) - Northern margin of the Penassen Lakes intrusive complex

345318E 5385843N

This stop requires parking on the shoulder of Hwy 527. Use hazard lights and traffic cones to increase visibility. Wear reflective vests and be mindful again of the traffic on Hwy 527. Be also mindful of soft shoulders when parking and walking.

This outcrop represents part of the northern contact of the Penassen Lakes intrusive complex. At this outcrop early hornblende monzodiorite is crosscut by intermediate aged hornblende quartz monzonite, which is cut by late pink leucocratic biotite monzogranite dikes (Figure 11A and 11B). Also visible are narrow mica-phyric mafic lamprophyre dikes and xenoliths of metawacke and possibly mafic metavolcanic rocks.

Multiphase, oxidized (magnetite bearing) intrusive complexes are a common feature of the SGB, LD and QS (*see* Figures 1 and 2). Early mafic phases of these intrusive complexes were emplaced between about 2677 and 2670 Ma, whereas later pink monzogranite phases appear to be about 5 million years younger (Figure 2). Although relative and absolute age differences between different phases of these intrusions are observed, we refer to them collectively as the potassium-rich calc-alkalic suite.

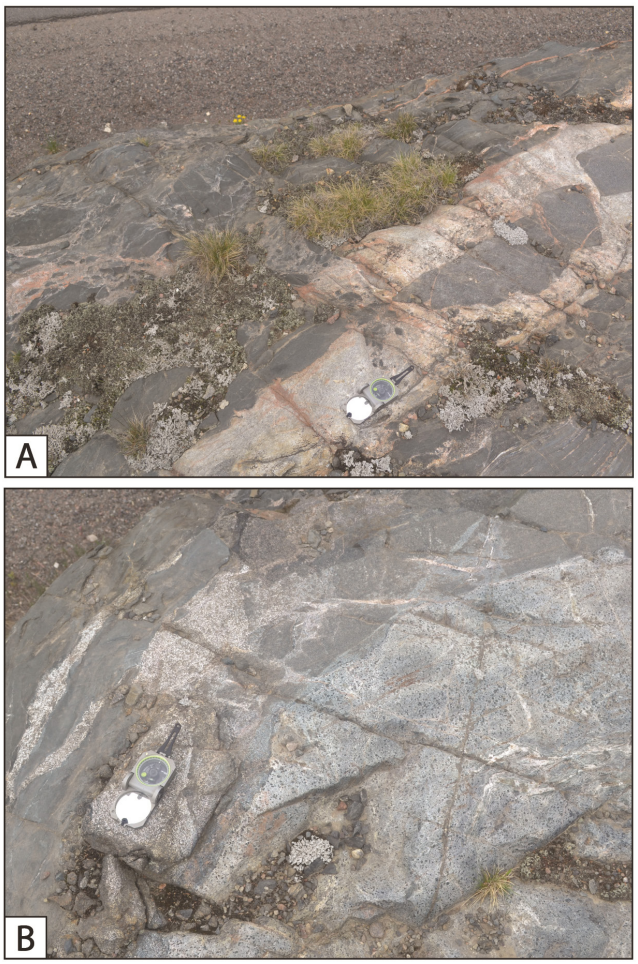


Figure 11. Northern margin of the Penassen Lakes intrusive complex at Stop 8. (A) Metasedimentary country rock fragments intruded by grey monzodiorite, crosscut by pink quartz monzonite to monzogranite dikes. (B) Amphibole-phyric quartz monzonite with xenoliths of metamorphosed wacke.

We consider this area part of the Lappe domain as we are south of the northernmost mapped mafic metavolcanic panel and the inferred eastward extension of the Moving Post fault. The Penassen Lakes intrusive complex and the similar Roll Lake intrusive complex to the north appear to post date the Moving Post fault based on geophysical interpretation and the presence of amphibolitic mafic pillowed flows occurring as large, strongly deformed inliers within the RLC and the emergence of a thin shear zone bounded mafic panel on the east side of the RLIC (Figure 1). This provides a clear relative age constraint on the Moving Post fault. These rocks are not, however, post-tectonic as we will see in later stops.

Stop 9 (Planned) - Lappe Domain metasedimentary rocks and Walkinshaw peraluminous granitic pegmatites

346697E 5388265N

Turn to the east on side road located just south of Stop 9 coordinate (Seagris Rd; no sign). If this side road is too rough for vehicles, use the road leading to summer camps on the northern end of Walkinshaw Lake, farther north. Wear high visibility vests, stay on shoulder or in ditch. Use extreme caution when crossing the road. Outcrops are relatively tall at this locality, be mindful of the potential for falling rocks in places.

This long outcrop contains relatively undeformed metasedimentary rocks (Figure 12A and 12B) in the northern part of the Lappe domain along with peraluminous granitic pegmatite dikes locally containing green mica, black tourmaline (Figure 12C and 12D) and disseminated molybdenite. These pegmatites we refer to as the Walkinshaw pegmatites. This area of metasedimentary rocks is surrounded by several large intrusive complexes (Potasssium-rich calc-alkalic suite), the Whitelilly Lake, Roll Lake and Penassen intrusive complexes, as well as several minor, sub-concordant, pink monzogranite bodies that are too small to map at 1:50 000 scale.

At this locality sedimentary features are well preserved (*see* Figure 12A) and the overall strain appears low. Foliation-bedding orientation relationships and measured intersection lineations suggest that folds in this area likely plunge steeply. Locally, thin shear zones have steep lineation plunges. Finer-grained beds locally have well developed porphyroblasts of andalusite (Figure 12B) and some cordierite, both are commonly replaced by muscovite. These assemblages are consistent with high temperature-low pressure metamorphism. At this locality we interpret that the observed metamorphic assemblage results from proximity to the many large intrusions in the area.

The Walkinshaw pegmatites are somewhat enigmatic. They occur in relatively low metamorphic grade rocks and there is no clear peraluminous "parent" granite nearby. A speculative explanation could be that small volume melts were locally produced by partially melting adjacent to contacts of nearby intrusions (PRCAS). There is local evidence for such melts, but only in very small volumes.

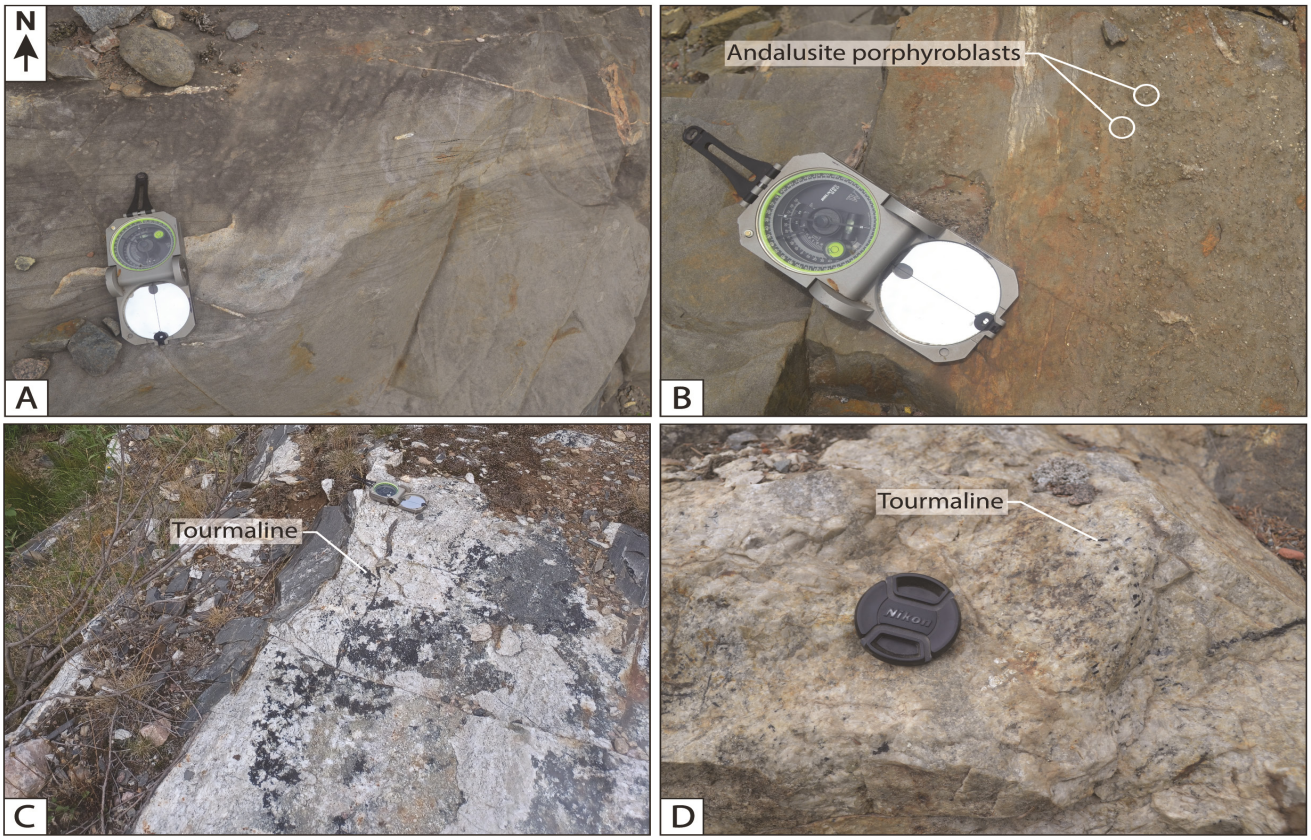


Figure 12. Lappe domain metasedimentary rocks and the "Walkinshaw" peraluminous granitic pegmatites (Stop 9). (A) northward younging wacke bed based on scouring and normal grading. (B) Andalusite porphyroblast-rich bed. (C) and (D) Black-tourmaline-rich subconcordant muscovite pegmatite dike.

Stop 10 (Planned) - Onion Lake pegmatites and folded Quetico metawacke

346817E 5398027N

Turn east on Cliff Rd (no sign) and drive about 250m down the dirt logging road and park to one side. Be aware of potential for logging traffic or other road users. Wood ticks are common at this site in spring and early summer. Use caution while walking around on the uneven ground.

This outcrop shows a clean exposure of a peraluminous granitic pegmatite typical of pegmatites we refer to as the Onion Lake pegmatites. At this stop, a biotite-muscovite-garnet bearing pegmatite-aplite dike (Figure 13A and 13B) strikes roughly northeast and appears to post-date a strong foliation affecting typical Quetico Subprovince metasedimentary rocks. The pegmatite displays prominent interlayering of pegmatitic and aplitic rock and unidirectional solidification textures (Figure 13A). This pegmatite is not far south of the contact of a peraluminous granite we refer to as the Voutilainen intrusion. Several large "whalebacks" of pegmatite are present in this area. Although not entirely clear at this locality, the Onion Lake pegmatites are boudinaged and commonly have internal fabrics defined by magmatic phases implying a syn-tectonic emplacement. Regionally, the northeastward strike of pegmatite contacts is parallel to sinistral shear zones which are interpreted to be antithetic structures related to the overall dextral shearing related to the Quetico deformation zone. Granites in this area occur near the southern margin of prominent deformation related to the QDZ. At least two distinct generations of peraluminous granitic pegmatites are present in the southern Quetico. A second generation of pegmatites, younger than the one at this locality crosscut at a high angle the foliation related to the Quetico deformation zone and therefore are post-tectonic. These later pegmatites are much less abundant, and we will not be able to examine the younger generation of pegmatites on this trip.

Metasedimentary rocks at this locality are strongly deformed. Relatively steeply plunging, east-northeast striking z- to m-folding is revealed by prominent quartz-feldspar veins (Figure 13C). These quartz-feldspar veins are very common in the southern QS.



Figure 13. Onion Lake peraluminous granitic pegmatite and folded Quetico metasedimentary rocks (Stop 10). (A) Pegmatite-aplite layering, unidirectional solidification textures, biotite-muscovite pegmatite and garnet aplite (B) Close-up of abundant garnet in aplitic phase, (C) boudinaged S-type granite dike sub-parallel to axial plane of "D_2" folds in Quetico metawacke.

Based on cross-cutting relationships they appear to predate most of the mapped intrusive suites. Note the narrow, boudinaged peraluminous granitic dike emplaced sub parallel to the axial plane of the folds appearing to crosscut the veining (Figure 13C). Outcrop scale folds of sedimentary layering are relatively uncommon in the southern Quetico making overall understanding of fold-geometries somewhat difficult. These folds are likely "D_2" in the nomenclature of Williams (1991) and limbs of similarly oriented folds are elsewhere postdated by dextral shearing (D_3). A prominent linear magnetic anomaly, caused by a relatively magnetic wacke unit, shows a clear regional z-folding pattern and this is likely the general geometry of "D_2" folding in the southern QS. In this area, many traverses across-strike documented well preserved younging indicators that show multiple reversals over relatively short distances. These reversals likely represent parasitic folds in hinge zones of larger z-fold enveloping surfaces. Towards the north, in the Quetico deformation zone, fold orientations change and tend to be east-striking, upright or slightly inclined with gentle plunges. These folds have been interpreted as being part of the D_3 event.

Stop 11 (Optional) - Mylonite and late brittle-ductile deformation Quetico deformation zone

348059E 5401951N

Pull vehicles over near the north end of long outcrop and park on the shoulder of Hwy 527. Keep the duration of stop relatively short. If longer stop required park on logging road just to the south. Wear reflective vests, use hazard lights. Traffic can be heavy.

At this locality, highly deformed metasedimentary rocks and sheared and boudinaged muscovite pegmatites are present. Locally, north dipping brittle structures offset pegmatite dike contacts with a north over south sense of displacement (Figure 14A). Pale

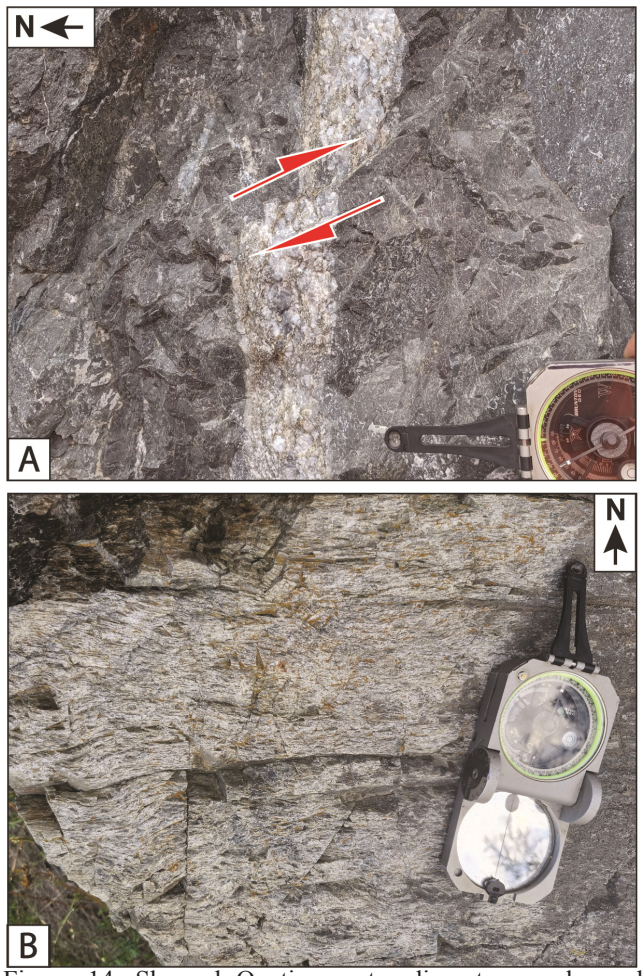


Figure 14. Sheared Quetico metasedimentary rocks and peraluminous granitic pegmatites, Quetico deformation zone (Stop 11). (A) brittle minor off-set thrust fault post-dates dextral shearing related to the ductile phase of the QDZ. (B) Pale green, siliceous mylonite related to the Quetico deformation zone.

green siliceous rocks exposed at the north end of the outcrop are likely a mylonitic band (Figure 14B) within the larger Quetico deformation zone. To the south, dark grey rocks are strongly deformed wacke.

Stop 12 (Planned) - Intermediate and felsic phases of potassium-rich calc-alkalic intrusive suite and syn-tectonic schlieric S-type granite

348541E 5403237N

Pull over on the shoulder of Hwy 527. Use hazard lights, wear reflective vests, use extreme caution when crossing the highway.

This large outcrop illustrates contact relationships between intermediate and felsic phases of the potassium-rich calc-alkalic intrusive suite (PRCAS) and a schlieric biotite-rich peraluminous granite. This outcrop represents the transition from the southern QS domain to the south and the Dog Lake injection complex to the north.

Grey amphibole-biotite-magnetite quartz monzonite at this outcrop is crosscut by later pink, moderately magnetic biotite monzogranite (Figure 15A and 15B). The quartz monzonite is moderately foliated and syn-tectonic. The quartz monzonite and the biotite monzogranite are compositionally similar to bigger, typically less strained intrusions located farther south e. Conversely, these types of intrusions are highly strained to the north in the Dog Lake injection complex. Preliminary geochronology suggests that despite highly variable degrees of strain, PRCAS intrusions have comparable ages in the Lappe domain, southern Quetico Subprovince and in the Dog Lake injection complex. Weakly deformed intrusions like the Trout Lake and Barnum Lake intrusions are essentially contemporaneous with highly sheared syn-tectonic equivalents in the DLIC.

At the north end of the outcrop, syn-tectonic, schlieric, low magnetic susceptibility peraluminous leucogranites are in contact with rocks of the PRCAS intrusions. C-S fabrics in the granite, and narrow sheared bands (Figure 15C and 15D) indicate north side up thrusting with a dextral strike-slip horizontal component during emplacement. East-northeast striking foliations locally bear shallowly plunging stretching and mineral lineations indicating strike slip motion. These lineations may have formed under brittle-ductile conditions after the main phase of ductile syn-granite shearing. The relative age of the peraluminous intrusion and the intermediate PRCAS phase is not immediately clear at this exposure. Regionally, the intermediate phase of the PRCAS slightly predates the bulk of S-type granite intrusions, and pink biotite monzogranites are typically younger. In some areas, hybridization of magmas have been documented.

Stop 13 (Planned) - Shear-hosted leucogranitic injections, Dog Lake injection complex

346928E/ 5407112N (this stop is not on Figure 1)

From Highway 527, turn right onto Hiccup Road (a logging road). Park on the curve near the stack of logs. This road is not active at the time of writing, and traffic is expected to be minimal.

This outcrop exposes paragneiss of the Dog Lake complex intruded and variably digested by large volumes of syntectonic leucogranitic injections (Figure

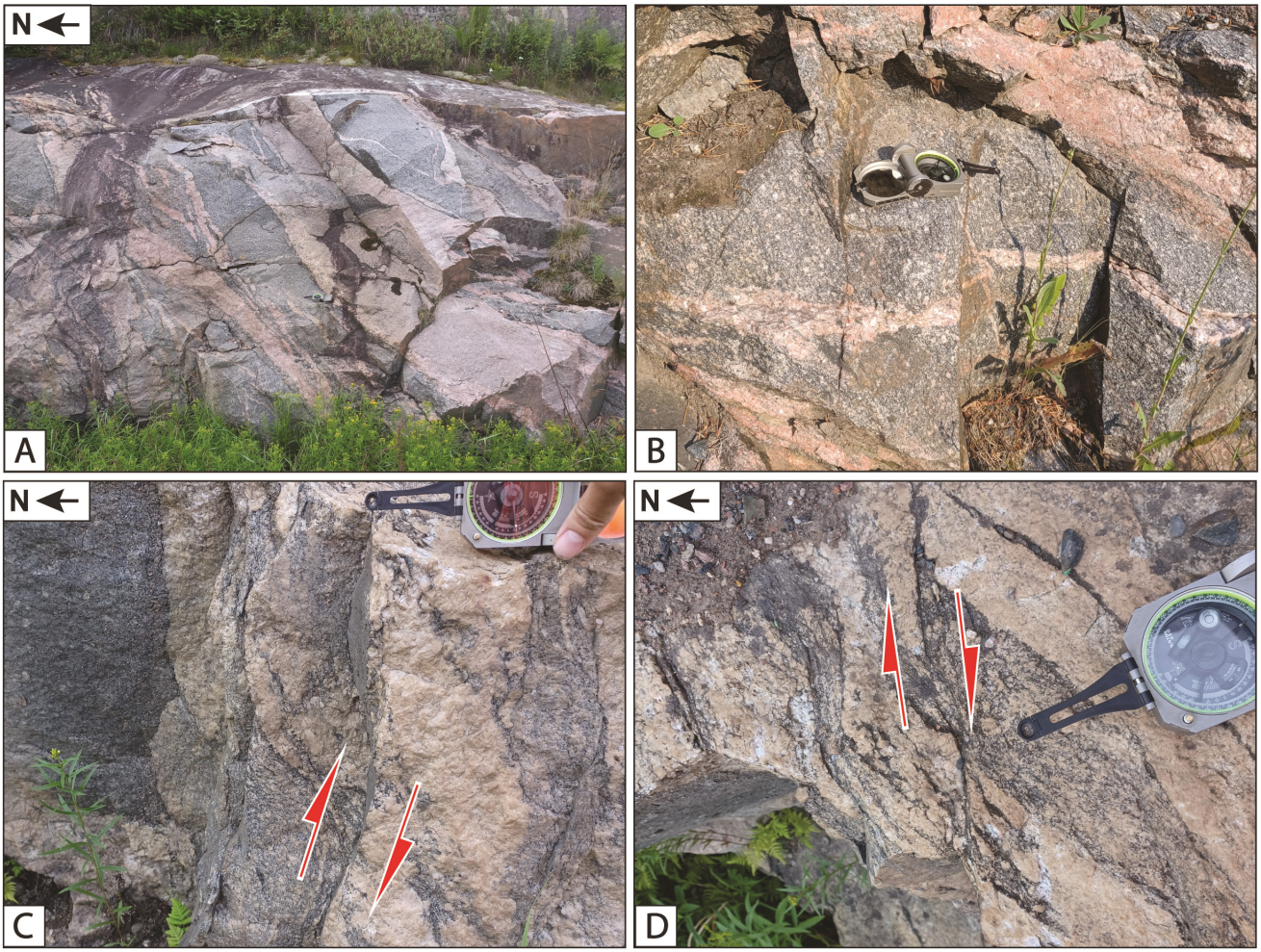


Figure 15. Potassium rich calc-alkalic suite intrusions and schlieric peraluminous granite at Stop 12. (A) Grey, foliated, feldspar porphyritic, hornblende-biotite quartz monzonite cross but by dikes of pink biotite monzogranite; (B) Porphyritic texture in quartz monzonite. (C) and D) syn-emplacement C-S fabrics defined by biotitic schlieren in biotite-rich peraluminous granite indicate a north-side up thrust component of motion (C) and dextral strike slip component of motion (D) during peraluminous granite emplacement.

16A), producing raft and schollen textures that can be easily confused with diatexite, and may therefore lead to misleading interpretations. The outcrop displays variable degrees of assimilation and digestion of the sedimentary host rocks by granitic melts, locally producing biotite schlieren within the leucogranite.

Leucogranite injections are heterogeneous, ranging from coarse-grained to porphyritic textures (Figure 16B). They occur as sheeted to irregular intrusions that clearly exploit east-west trending foliation planes and northwest-trending dextral shear bands within the paragneiss host rock. This strong structural control indicates syntectonic emplacement facilitated by regional dextral zones related to the Quetico deformation zone. Locally, antithetic sinistral shear bands are also observed.

Wider and more homogeneous granite injections locally contain garnet and pegmatitic segregations (Figure 16C and 16D). The presence of these pegmatitic segregations indicate a high degree of melt fractionation, which is incompatible with in situ partial melting. This interpretation is supported by chondritenormalized REE patterns, which display a strongly fractionated geochemical signature and variably developed negative Eu anomalies (Figure 17), comparable to those observed in the S-type granite intrusions of the Quetico subprovince. These features indicate that granitic melts had already undergone significant plagioclase fractionation at depth forming cumulates before extraction of the residual melt along shear zones.

Although some textures locally resemble those seen in migmatite, their interpretation as true anatectic melts is not supported by the mineralogy.

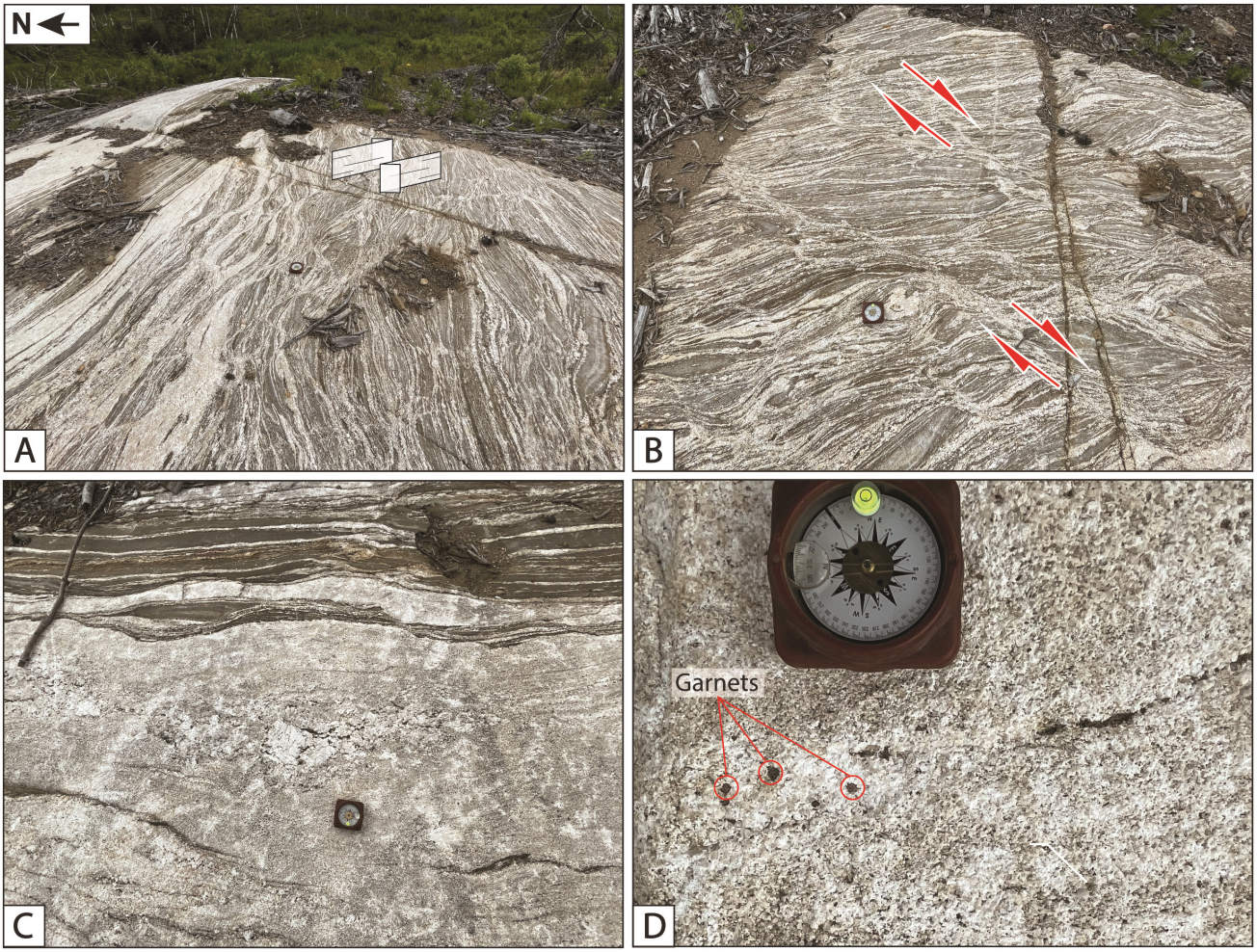


Figure 16 Photographs of shear-hosted leucogranitic injections of the Dog Lake injection complex (Stop 13). (A) Dense network of leucogranitic injections emplaced along foliation planes and dextral shear bands. Note the presence of paragneiss rafts within wider injections (B) Close-up view of dextral shear bands. Note the presence of biotite-rich schlieren in some leucogranitic injections. (C) Schlieric garnet-bearing dyke of leucogranite. Note the presence of pegmatitic pods indicating fluids segregation. (D) Close up on garnets within the dyke of leucogranitic.

Melanosome-looking parts are mostly only composed of biotite and lack typical peritectic mineral phases (e.g., garnet, cordierite, sillimanite) expected from insitu partial melting of metasedimentary protoliths.

Instead, this outcrop is interpreted as a migration zone for granitic melts generated deeper in the crust, which were channeled upward along regional dextral shear zones. During ascent, these melts assimilated sedimentary host rocks, resulting in the development of characteristic schollen and schlieric textures observed at this outcrop.

The large volume of ascending granitic melt likely induced high-temperature, low-pressure metamorphic conditions in the surrounding paragneiss, as evidenced by the development of sillimanite–cordierite assemblages. Locally, the paragneiss also experienced limited partial melting, interpreted to have been induced by thermal input associated with the emplacement of this large volume of granitic melts.

Stop 14 (Planned) - Patchy metatexites, Dog Lake injection complex

346711E 5406946N (this stop is not on Figure 1)

From the previous stop, return to Highway 527 and cross the highway onto Doodie Road. Park at the entrance of the road. The outcrop is located on the north side of the road.

This outcrop exposes an interlayered biotite–quartz–feldspar and garnet–biotite–quartz–feldspar migmatitic paragneiss, interpreted as a patchy metatexites (Figure 18A). Peritectic garnet is present within leucosome patches, providing clear evidence for insitu partial melting of fertile pelitic layers. The estimated melt proportion is relatively low (approximately 5–10%).

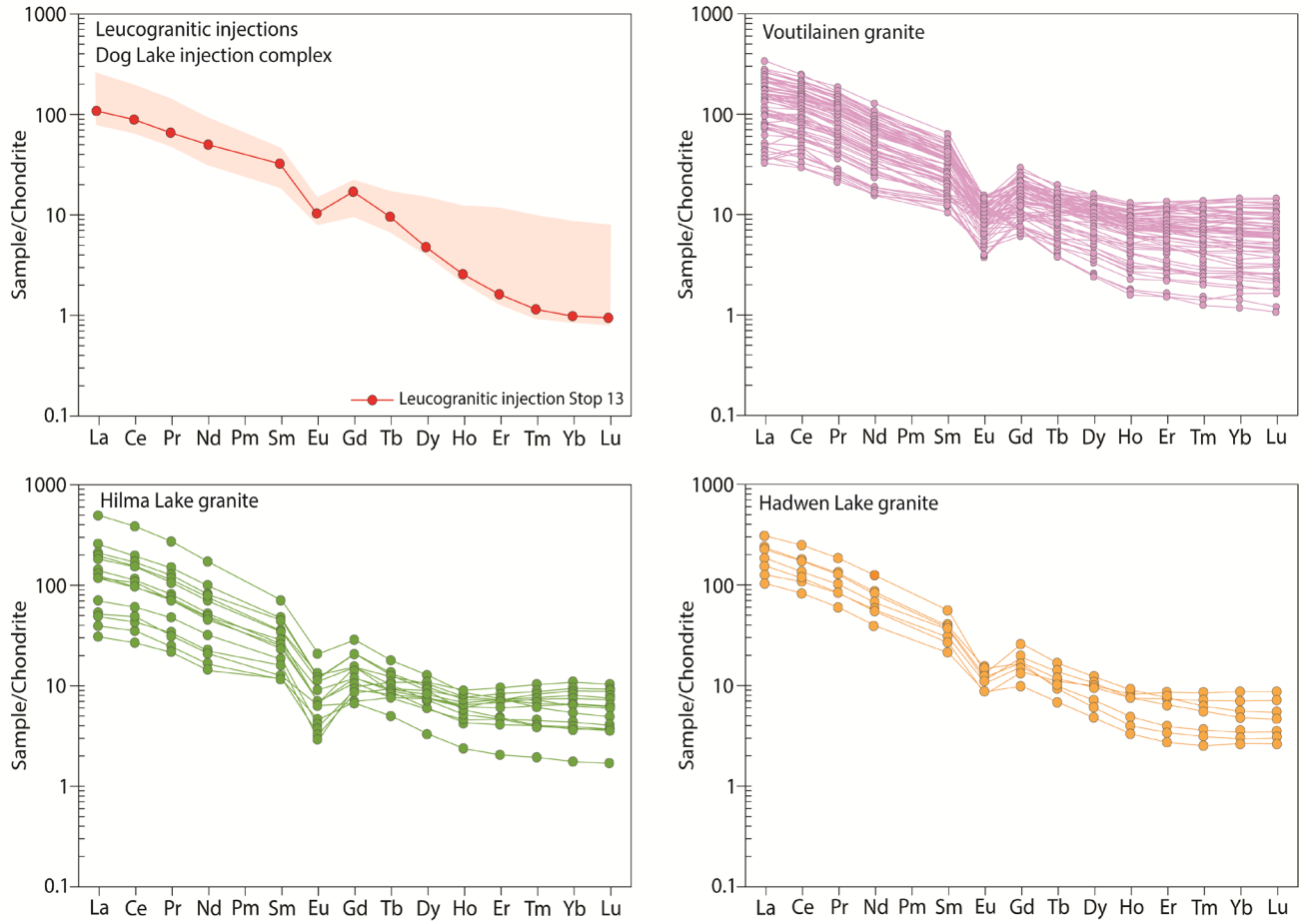


Figure 17 Chondrite normalized rare earth element (REE) patterns of leucogranitic injections from the Dog Lake injection Complex compared with Stype granite intrusions occurring in the Quetico Subprovince.

Bedding is largely preserved, indicating low melt connectivity and limited melt extraction (Figure 18B).

This migmatitic paragneiss is gently folded, with fold axes plunging eastward. The east–westtrending foliation dips steeply to the north. The paragneiss is cut by boudinaged dikes of white, coarse-grained leucogranite, similar to the leucogranitic injections observed at the previous stop.

This outcrop highlights a significant volumetric contrast between the limited amount of melt generated in situ within the paragneiss and the much larger volume of granitic injections observed at the previous stop (Stop 13), despite that the two localities are separated by only ~300 m. This contrast is a strong indication that the granitic melts observed elsewhere in the Dog Lake injection complex were not produced in situ but instead represent migrated and fractionated melts generated at deeper crustal levels, which were subsequently channeled upward along regional shear zones. The spatially restricted partial melting and migmatitization observed at this outcrop therefore do not represent the source of the Stype granites but rather reflect a thermal response to advected heat associated with the emplacement of large volumes of granitic melt in nearby shear corridors, rather than a widespread regional anatexis.

Stop 15 (Optional) - Syn-tectonic schlieric pink monzogranite, Dog Lake injection complex

347031E 5405626N (this stop is not on Figure 1)

Continue driving south along Doodie Road for approximately 1.5 km. This road is not active at the time of writing; however, road shoulders may be narrow or unstable, so vehicles should be parked directly on the road where it is safe to do so.

This final stop exposes a representative outcrop of syntectonic, magnetitebearing pink monzogranite (Figure 19). The intrusion displays a heterogeneous texture, ranging from porphyritic to locally pegmatitic, highlighting the high fluid content of the granitic melt during its emplacement. The granite contains biotite schlieren, resulting of the complete digestion of the paragneiss host rock. Locally, relict structures of the

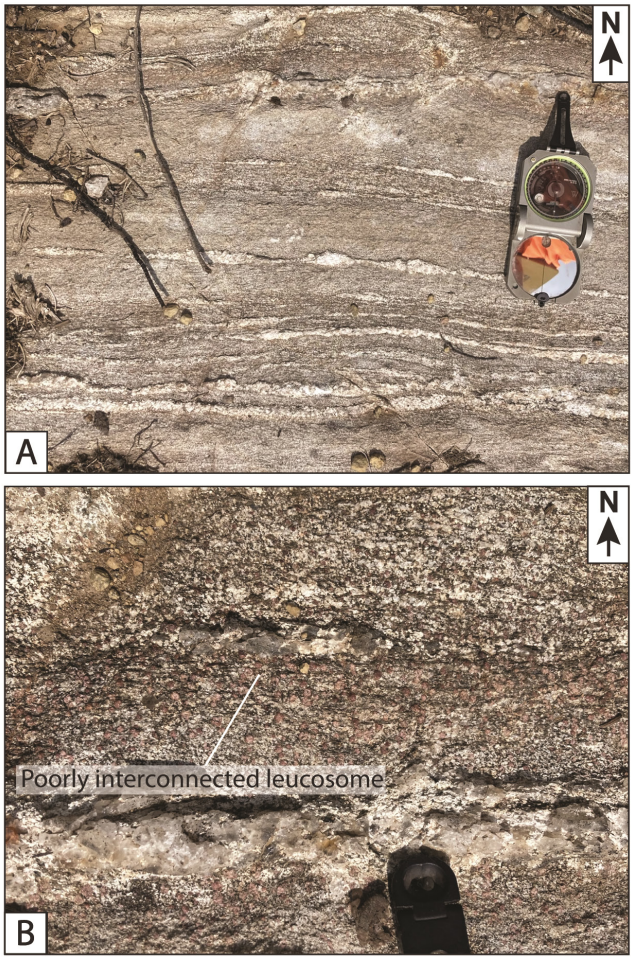


Figure 18 Representative photographs of garnet bearing, patchy metatexites from the Dog Lake Injection Complex (Stop 14). (A) In situ partial melting localized within fertile pelitic layers of the paragneiss. Note that primary bedding structures are largely preserved. (B) Close up view of photo A. The upper layer shows a clear lack of segregation between leucosome and melanosome indicating low melt connectivity.

original sedimentary bedding can still be inferred within these schlierenrich domains. The intrusion is strongly foliated and affected by welldeveloped dextral shear bands and antithetic sinistral shear bands, along which pegmatitic pods are locally emplaced.

This outcrop illustrates how regional dextral shear zones within the Dog Lake complex have acted as efficient pathways for multiple types of granitic melts, which derived from different sources in the lower crust. The pink monzogranite suite is interpreted as the final, most fractionated product of the potassiumrich calcalkaline intrusive suite.

These relationships reinforce the interpretation that the Dog Lake complex represents a major migration zone, where granitic magmas produced at depth were focused, transported, and emplaced along

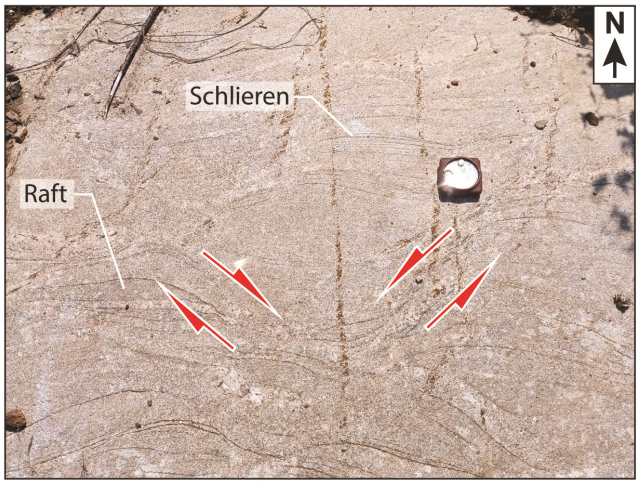


Figure 19 Photograph of syntectonic, schlieric biotite pink monzogranite (Stop 15). Note the presence of dextral and antithetic sinistral shear bands.

shear corridors. Melt migration was facilitated and channelized by regional transpressional deformation acting in the Quetico Subprovince and its adjacent greenstone belts.

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Trip 3 - Geological assemblages, regional structural framework and tectonic evolution of the Neoarchean Shebandowan greenstone belt

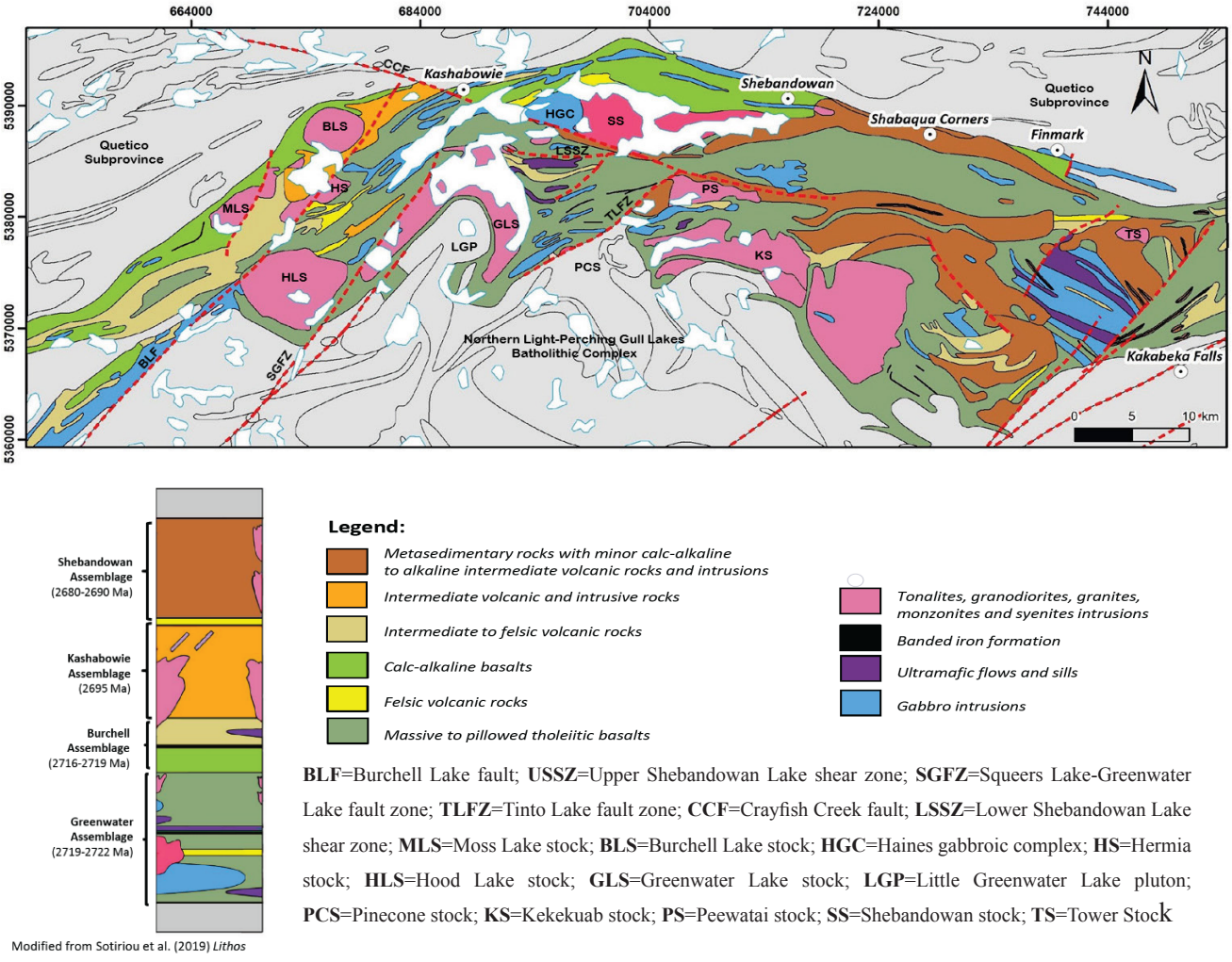
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Introduction

This trip provides an overview of the geological assemblages, regional structural framework, and tectonic evolution of the Neoarchean Shebandowan Greenstone Belt (SGB) and their relationship to gold and base metal mineralization. The SGB is situated in the western Wawa Subprovince (Superior Province) and extends 150 km from the Ontario–Minnesota border in the west to northeast of Thunder Bay in the east (Figure 1). The SGB is locally in fault contact with the Quetico Subprovince to the north and bounded by the older (2750 Ma) Northern Light–Perching Gull Lakes batholith (tonalitic gneiss) and younger granitic intrusions to the south (Lodge 2016).

The SGB is characterized by a complex history of early rifting, subduction-driven volcanism, tectonic accretion, and later transpressional deformation. The SGB comprises three main assemblages and two primary deformation events (Williams et al. 1991; Stott and Corfu 1991; Corfu and Stott 1998; Percival 2006; Lodge 2016; Reynolds et al. 2023; Dorval et al. 2026):



BLF=Burchell Lake fault; **USSZ**=Upper Shebandowan Lake shear zone; **SGFZ**=Squeers Lake-Greenwater Lake fault zone; **TLFZ**=Tinto Lake fault zone; **CCF**=Crayfish Creek fault; **LSSZ**=Lower Shebandowan Lake shear zone; **MLS**=Moss Lake stock; **BLS**=Burchell Lake stock; **HGC**=Haines gabbroic complex; **HS**=Hermia stock; **HLS**=Hood Lake stock; **GLS**=Greenwater Lake stock; **LGP**=Little Greenwater Lake pluton; **PCS**=Pinecone stock; **KS**=Kekekuab stock; **PS**=Peewatai stock; **SS**=Shebandowan stock; **TS**=Tower Stock

Figure 1. Regional Geology of the Shebandowan greenstone belt (*modified from* Kuster, Lesher and Houlé, 2022; *modified from* Sotiriou et al 2019; Lodge 2016; Osmani 1997a; Corfu and Stott, 1998).

Greenwater assemblage (2722-2719 Ma): volcanic suites characterized by thick sequences of tholeiitic mafic volcanic rocks, ultramafic flows (komatiites) and sills, iron formations, mafic intrusions, and minor FII- and FIII-type felsic volcanic rocks.

Burchell assemblage (2719-2716 Ma): calc-alkalic, dominantly intermediate volcanic rocks and lesser FI-type felsic volcanic rocks and with no known ultramafic sills or intrusions. This subdivision was first defined by Williams (1991) on the basis of younging directions but rejected by Corfu and Stott (1998) due to lack of chronological distinction and re-interpretation of structural architecture. Lodge (2016) interpreted more recent higher-precision geochronology as supporting a similar subdivision to that of Williams (1991).

Kashabowie assemblage (2695 Ma): syn-D_1, represents renewed activity on the SGB after a long hiatus. It is less voluminous and more evolved than Greenwater assemblage. Calc-alkaline to alkalic intermediate/felsic volcanic rocks, associated diorites, tonalites (e.g., Shebandowan Lake Pluton), tectonically interleaved with older 2720 Ma volcanic suites. This subdivision was first introduced by Corfu and Stott (1998) as part of their re-interpretation of older assemblage classifications.

D_1 Compressional Deformation event (2695 and 2690 Ma): associated with calc-alkaline magmatism and intra-arc deformation (thrust-stacking and interleaving).

Shebandowan assemblage (2690-2680 Ma): syn-D_2 Timiskaming-type assemblage, unconformably overlies older Greenwater assemblage, composed of calc-alkalic to alkalic volcanic rocks and associated coarse clastic Timiskaming-type sedimentary rocks, iron formation and late sanukitoid plutons.

D_2 Transpressional Deformation event (2685–2680 Ma): marked the final accretionary phase of the Wawa subprovince evolution of the Superior Craton, termed the Shebandowanian phase of the Kenoran Orogeny (Stott and Corfu 1991). This late-stage dextral transpression and oblique-slip deformation represents the development of Timiskaming-type pull-apart basins and regional Timiskaming-aged structures.

Auto Road assemblage (<2682 Ma): distinctly younger sedimentary assemblage in the SGB, dominated by conglomerate-sandstone units (with clasts of volcanic and granitoid origin). Corfu and Stott (1998) describe the assemblage as a small sedimentary basin, informally termed the "Auto Road assemblage".

The western limb of the SGB is often divided from the central and eastern portions of the belt by an informal north-south boundary roughly, at the town of Kashabowie (Figure 1). There are differences in the distribution of assemblages between the west and east sides: the Kashabowie assemblage is situated mostly along the western limb, the Shebandowan assemblage is situated on the central-eastern side, and the Auto Road assemblage is restricted to a small area on the eastern side.

The Greenwater/Burchell assemblage(s) make up the large majority of the preserved supracrustal rocks, despite comprising just ~6 million years of the >40 million-year evolution of the SGB (Figure 1). The volcanism recorded by these assemblages appears to have been two-stage: an extensional plume-rift setting recorded by the Greenwater assemblage followed by a compressional subduction-arc setting recorded by the Burchell assemblage (Figures 2, 3; Lodge 2016).

The deposition of chemically distinct Kashabowie assemblage volcanic rocks occurred after a ~21 million-year hiatus, recording a later compressional subduction-arc setting (Figure 3). These rocks are contemporaneous with the D_1 structural event, which involved the interleaving and thrust-stacking of the Kashabowie and Greenwater units (Reynolds et al. 2023).

Subsequently, the Shebandowan assemblage, represents the final stages of the Shebandowan accretionary event. These "Timiskaming-type," deposits unconformably overlie the Greenwater assemblage. They are characterized by a mix of clastic sediments (conglomerate, sandstone), iron formation, and calc-alkalic to alkalic volcanic rocks (Figure 3), interpreted to have formed in transtensional, pull-apart basins along the flanks of transpressional uplifts (Reynolds et al. 2023). Due to their tectonic setting, these rocks are strongly associated with structurally controlled, late-orogenic gold mineralization (Figure 4).

Based on the spatial distribution of southwest-

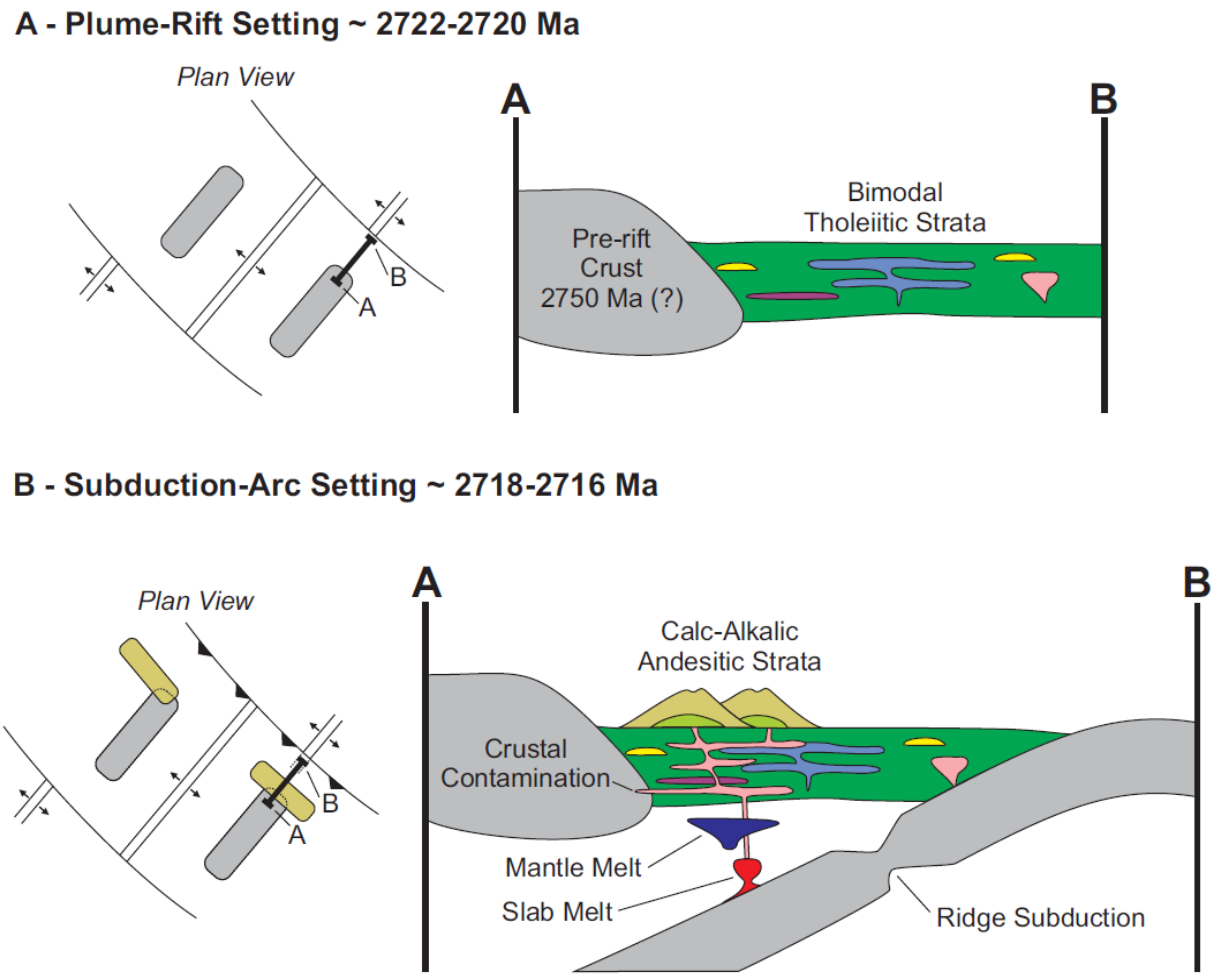


Figure 2. Schematic illustration of possible tectonic evolution of the Shebandowan greenstone belt in both plan view and crustal cross section (*from* Lodge 2016). Colors of units correspond to legends in Figure 1. Note sketch is not to scale. (A) Initial plume-dominated tectonic setting forming Greenwater assemblage. (B) Subduction-dominated tectonic setting forming Burchell Assemblage. A change in plate motion results in the initiation of subduction and formation of a calc-alkalic arc dominated by andesitic strata. Subduction of ridge results in high geothermal gradient and melting of slab to produce adakitic melts. Hybridization of mantle and slab derived melts results in magnesian andesites (Mg# > 50) *from* Lodge 2016.

trending metavolcanic rocks on the western limb, Osmani (1997a) defined three distinct geological units that remain in use by current explorers (Figure 6):

- Central Felsic Belt (CFB): a <5 km-wide core of the Burchell/Kashabowie assemblage.
- Northern Mafic Belt (NMB) and Southern Mafic Belt (SMB): mafic metavolcanic rocks of the Greenwater assemblage, flanking the CFB to the north and south respectively.

There are three past-producing mines in the SGB: the North Coldstream copper mine (1957-1967) and the Ardeen gold mine (1932-1936, 1942) in the western part of the belt, and the Shebandowan nickel-copper-PGE-cobalt mine (1971-1998) in the eastern part of the belt (Figure 5).

Tectonic associations provide spatial context for mineral prospectivity in the SGB (e.g. Lodge et al. 2015, Lodge 2016, Reynolds et al. 2023). Magmatic Ni-Cu-PGE mineralization occurs in the Greenwater assemblage mafic-ultramafic intrusive rocks, notably the sill-hosted deposit comprising the past-producing Shebandowan mine. The mine operated for most of 1971-1998, producing 9.29 Mt at 1.75% Ni, 0.88% Cu, 0.06% Co and 1.83 g/t PGEs. Clusters of magmatic sulfide occurrences also occur in the Haines gabbro (~7 km northwest of the mine) and in the Bateman Lake area (~40 km east of the mine).

Although no economic deposits that are definitively of volcanogenic massive sulfide (VMS) affinity exist in the SGB, several prospects exist in spatial association with Greenwater/Burchell assemblage felsic metavolcanic rocks. The North Coldstream copper-gold-silver deposit, located 10 km southwest of Kashabowie on the western arm of the SGB, is a

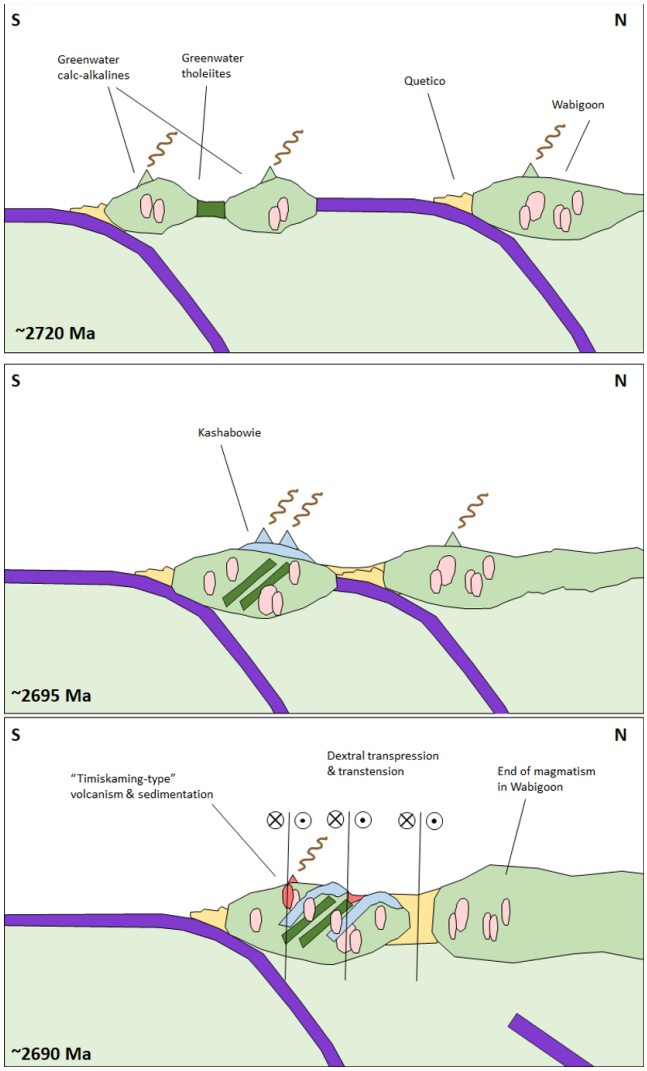


Figure 3. Evolution of Greenwater/Burchell, Kashabowie, and Shebandowan assemblages (*from* Reynolds et al. 2023).

past-producing (2.48 Mt at 1.87% Cu, 0.28 g/t Au, and 5.53 g/t Ag from 1957-1967) atypical deposit variably interpreted to be intrusion-related (e.g. Farrow 1994) or volcanogenic (Reynolds et al. 2023); it is currently being explored by Gold X2 Mining Inc., who tentatively interpret the deposit as a sheared, remobilized VMS system.

In more recent years, orogenic gold has become the main focus of mineral exploration in the SGB. Gold is primarily controlled by late tectonic D_2 structural zones, in contrast to lithologically controlled magmatic and VMS mineralization associated with Greenwater/Burchell assemblages. Gold mineralization is generally hosted within ductile-brittle shear zones, particularly near regional fault zones (Figure 4) or adjacent to "Timiskaming-type" unconformities (Figure 14).

Gold is typically hosted by quartz-carbonate-pyrite veins and veinlet networks cross-cutting all lithologies. On this field trip, we will look at some specific examples these structural zones:

- Moss Gold Deposit (Gold X2 Mining Inc.) and the 111 Zone (Bold Ventures Inc.): gold mineralization occurs near regional fault zones, within sheared diorites, felsic dykes/sills and mafic to intermediate metavolcanic rocks.
- I-Zone (Delta Resources Limited): gold-bearing quartz ladder veins within a felsic dyke (brittle, extensional), intruding Timiskaming iron formation.
- Eureka Zone (Delta Resources Limited): a key target for gold exploration at the unconformity

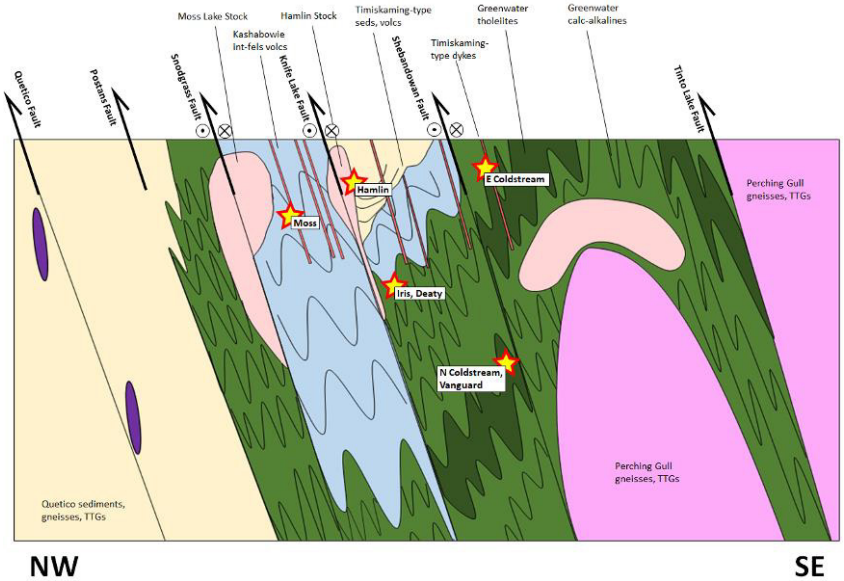


Figure 4. Schematic section of western Shebandowan greenstone belt (from Reynolds et al. 2023).

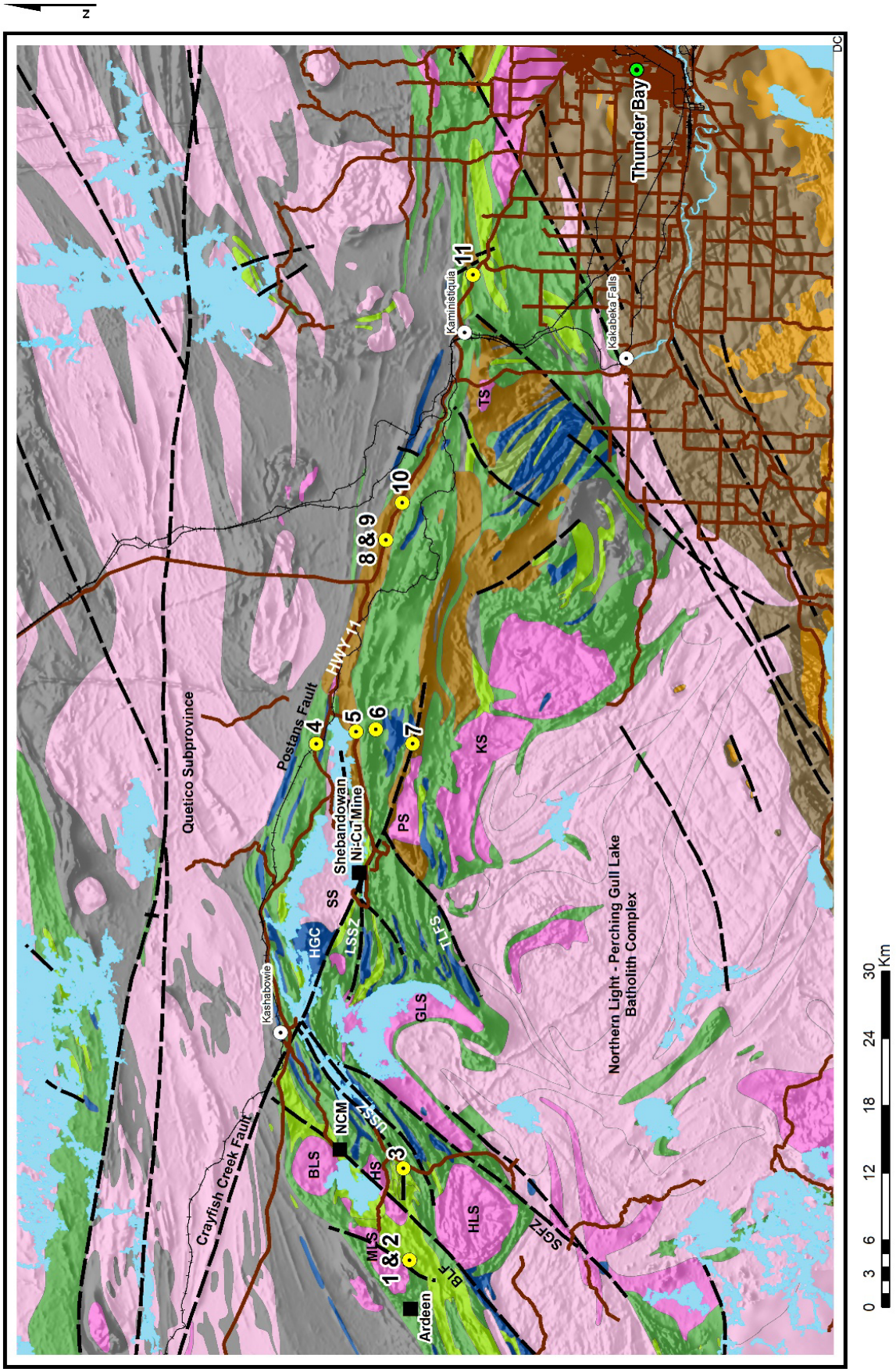


Figure 5. Geology map of the Shebandowan greenstone belt showing location of field trip stops. NCM: North Coldstream Mine, *See* Figure 1 for all other abbreviations.

between the Greenwater and Shebandowan assemblages, marking a "Timiskaming-type" unconformity.

Stops 1 to 2 - Moss Gold Deposit (Gold X2 Mining Inc.)
Permission is required from company to access sites

Gold X2 Mining Inc. (Gold X2) is exploring the Moss gold deposit, a high-tonnage low-grade deposit (Figure 6), located 100 km west of Thunder Bay on the western limb of the SGB. Gold X2 recently completed a Preliminary Economic Assessment, releasing an updated resource estimate as of January 16, 2026, (Dorval et al. 2026) for the deposit as follows:

- Indicated: 2.125 Moz Au at 1.03 g/t with 3.160 Moz Ag at 1.53 g/t
- Inferred: 3.910 Moz Au at 0.97 g/t Au with 6.273 Moz Ag at 1.55 g/t

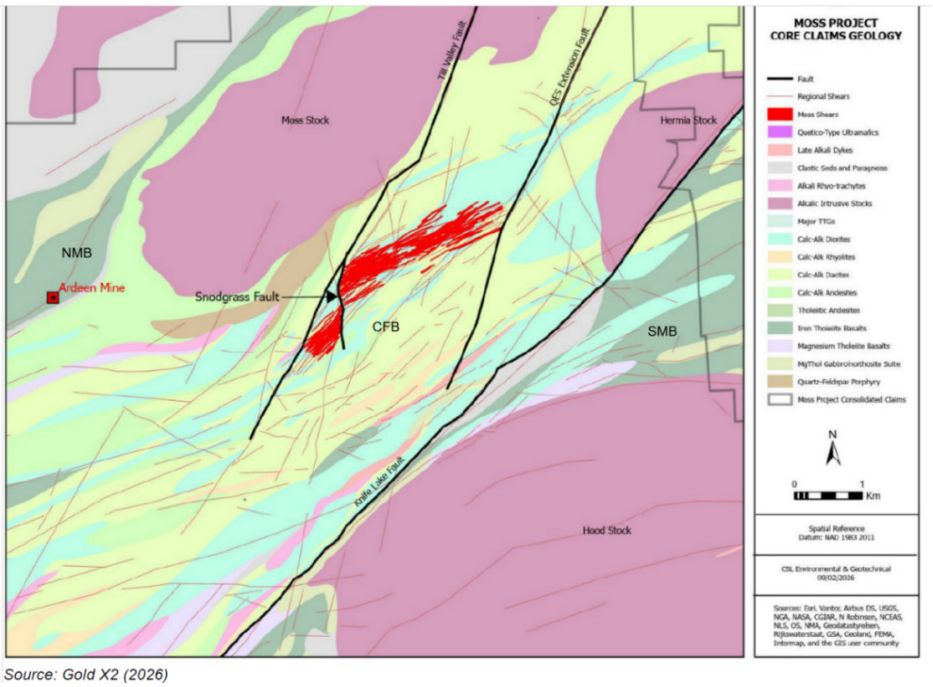
Engineering trade-off studies & design work is underway and a feasibility study is anticipated for Q3 2027 (Gold X2 Mining Inc., *Corporate Presentation*, April 12, 2026).

The Moss deposit is structurally controlled and situated within intermediate to felsic metavolcanic rocks of the Central Felsic Belt (CFB; Figure 6). Primarily hosted by sheared diorite (Figure 7), the deposit developed during and after intense ductile deformation, with 2 distinct tectonic-hydrothermal events identified (Reynolds et al 2023; Dorval et al. 2026). Alteration occurs in different styles and intensities but is generally composed of albite, biotite, sericite, chlorite, carbonate, epidote and pyrite (typically 2-10% of the rock; locally up to 15%). Gold mineralization occurs in complex arrays of small-scale quartz-carbonate-pyrite veinlets, breccias, and stockworks with higher grades within more intense, narrow shear zones (Nwakanma 2024; Dorval et al. 2026). The sulfide assemblage is dominated by pyrite, with minor chalcopyrite, sphalerite, and molybdenite. Rare, high-grade tellurides are associated with the high-grade gold mineralization (Reynolds et al. 2023; Dorval et al. 2026).

Stop 1. Moss Gold Deposit (Portal)

N83 Z15 U 668730E 5379177N

At this stop, highly sheared diorite and feldspar porphyry has been variably silicified, chloritized, hematized, sericitized and sulphidized (Figure 7). The outcrop is highly fractured and exhibits a network of narrow quartz-carbonate-pyrite veinlets. The now closed-off portal, developed in the mid-1980s by Tandem Resources Limited and Storimin Exploration Limited, lead to historical underground workings and gold zones at the 230-foot (70 m) level (Figure 8).



Source: Gold X2 (2026)

Figure 6. Geology map shear hosted Moss Lake Deposit (in red) *modified from* Dorval et al. (2026).

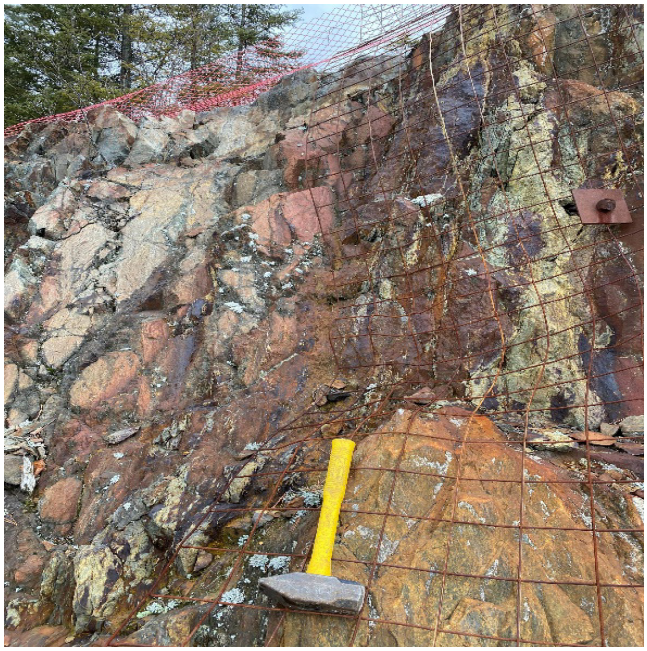


Figure 7. Gold-mineralized diorite at the Moss deposit that has been variably silicified, chloritized, hematized, sericitized and sulphidized.

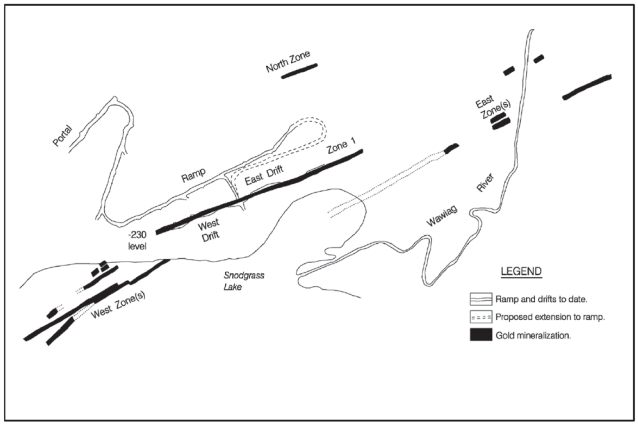
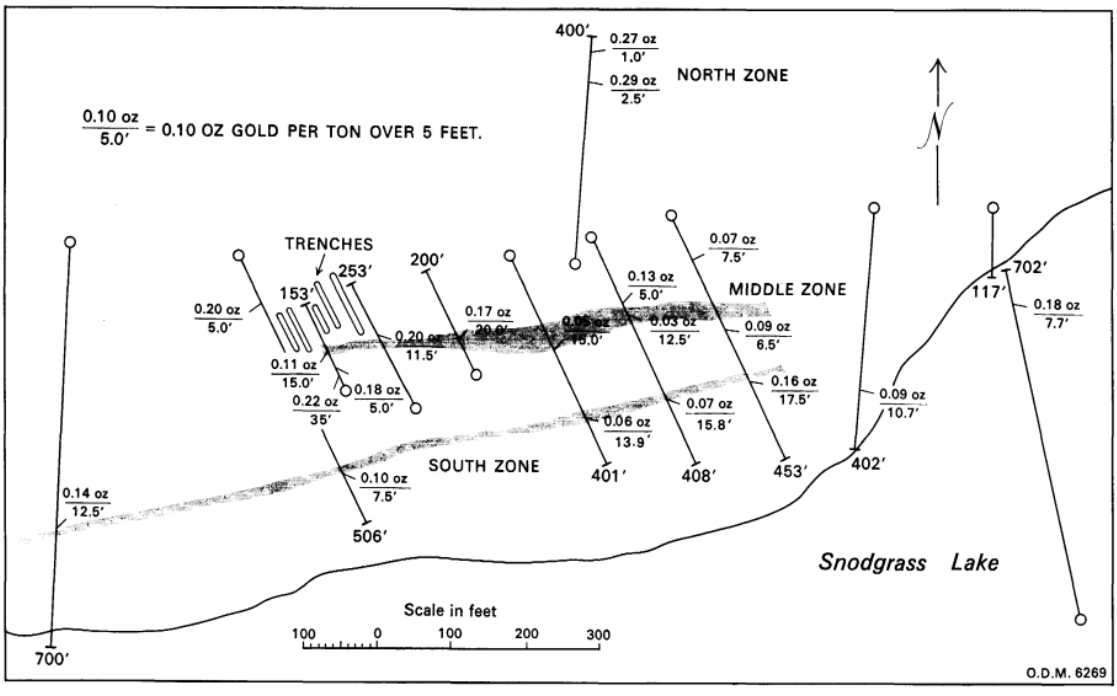


Figure 8. Underground plan of the 230-foot (70 m) level, showing gold-bearing zones of the Moss deposit (*from* Osmani 1997a; *modified from* an underground plan of Tandem Resources Limited - Storimin Exploration Limited, 1989).

Stop 2. Discovery Outcrop

The Moss property has a long history of exploration dating back to 1936, when Mining Corporation of Canada completed 5 trenches that exposed a zone of mineralization later known as the Main Zone (often referred to in historical records as the Snodgrass showing). Gold was initially identified in a mineralized zone hosted by sheared dacite and felspar porphyry near the northern contact with diorite. The zone measured approximately 25 feet (7.6 m) in width and 600 feet (180 m) in length. In 1945, Lobanor Gold Mines Limited followed up with 12 diamond drill holes which ultimately led to the development of the Moss deposit (Figure 9).

Subsequently, more than 30 companies explored various smaller sections of the property that were consolidated in 2014-2016 by Wesdome Gold Mines Ltd. In May 2021, Gold X2's predecessor (Goldshore Resources Inc.) acquired the Moss Gold property from



Plan of drilling on the Minoletti Snodgrass property; modified from a plan by Lobanor Gold Mines Limited, 1945.

Figure 9. Map showing location of initial trenches, drill holes and gold assay results by Lobanor Gold Mines (1945) (*from* Harris 1970).

Wesdome. In 2024-2025, the property was further expanded by: i) purchasing the "Coldstream claims" and acquiring Kesselrun Resources Ltd., whose Huronian project claims include the past-producing Ardeen mine, ii) staking the Hillcrest property (Crayfish Creek Fault extension) and claims covering the Squeers-Greenwater Fault Zone extension, to the north and south of the Moss deposit, respectively, and iii) optioning Sky Gold's Star Lake property, based on OGS gold-in-till anomalies (Figure 10). These expanded land holdings are strategic and a testament to the importance of regional structures for gold exploration.

Stop 3. 111 Au Zone Trench - Burchell Lake Au-Cu Property (Bold Ventures Inc.)

N83 Z15 U 676840E 5380320N

Permission is required from company to access site

The Burchell Lake Au-Cu property, located 95 km west of Thunder Bay, is adjacent to Gold X2's Moss property to the west. While the property hosts multiple Au and Au-Cu showings, this stop focuses on Bold's newly discovered 111 Au Zone (Figure 11 and 12). Initial grab samples returned 59.9 g/t and 68 g/t Au (Figure 11). Sampling at the 111 Au Zone by the Regional Resident Geologist (2025) returned up to 61.2 g/t Au and >1.2% Cu. Early assay results from 2026 drilling at the 111 Au Zone, BL-26-001 returned 0.42 g/t Au over 19 m, including 1.1 g/t Au over 5.0 m, and 2.7 g/t Au over 1 m.

At this stop, silicified mafic to intermediate metavolcanic rocks are crosscut by a northeast-trending anastomosing shear zone (Figures 11, 12). A 14 m-wide halo of anomalous gold (*see* red dotted outline on Figure 12) has been outlined, flanked with zinc and copper mineralization. Gold mineralization is associated with disseminated pyrite and stringers of chalcopyrite, hosted in strongly silica- and sericite-altered metavolcanic rocks. Locally, the rock is characterized by intense shearing and alteration obscuring the protolith, potentially a sheared and highly silicified metavolcanic rock or diorite. A narrow, relatively undeformed felspar porphyry occurs adjacent to the shear zone (Figure 12). Osmani (1997b) mapped this location as a felsic metavolcanic-dominated portion of the Southern Mafic Belt, though there was no outcrop exposure at the 111 Au Zone at the time of his mapping. Corfu and Stott (1998) reported a U–Pb zircon age of 2721 ± 1 Ma from a felsic metavolcanic flow less than 2 km to the northeast, interpreted to represent its eruption age. The mafic metavolcanic

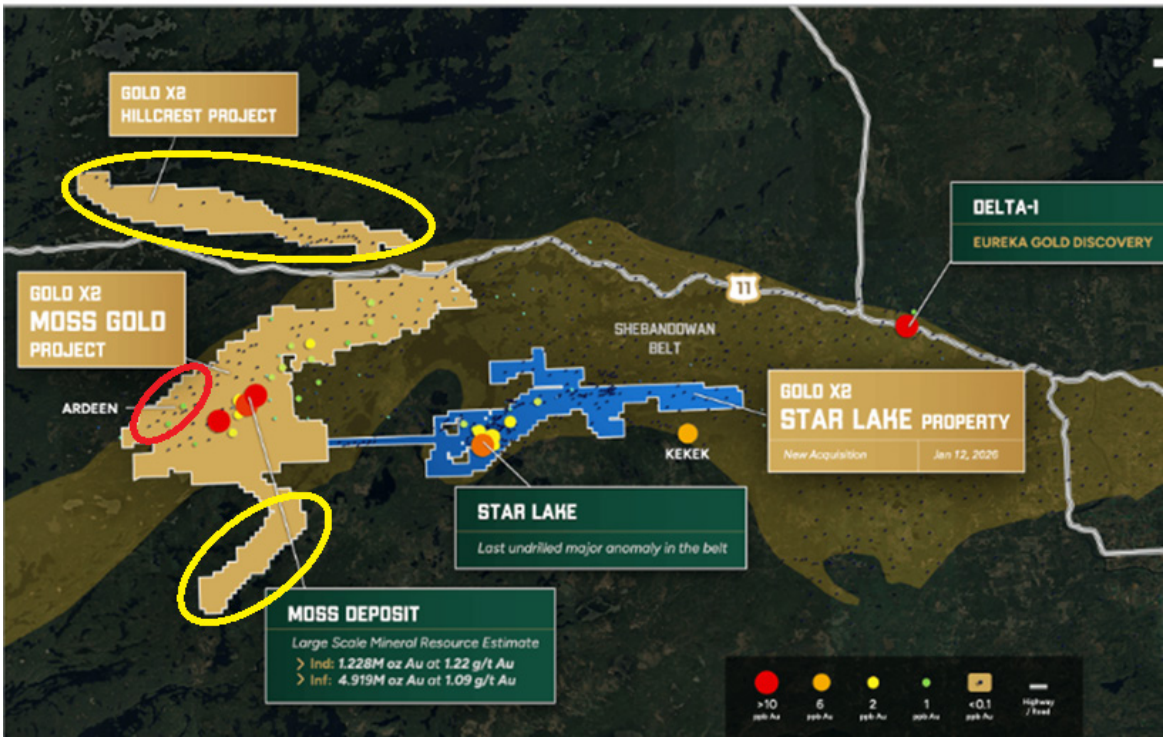


Figure 10. Map showing Gold X2's 2025-2026 land acquisitions: Kesselrun's Huronian project with the past-producing Ardeen Mine (red oval), Hillcrest and Squeers-Greenwater projects (yellow ovals), and Sky Gold's Star Lake property (blue) with a cluster of gold in-till anomalies (orange and yellow dots).

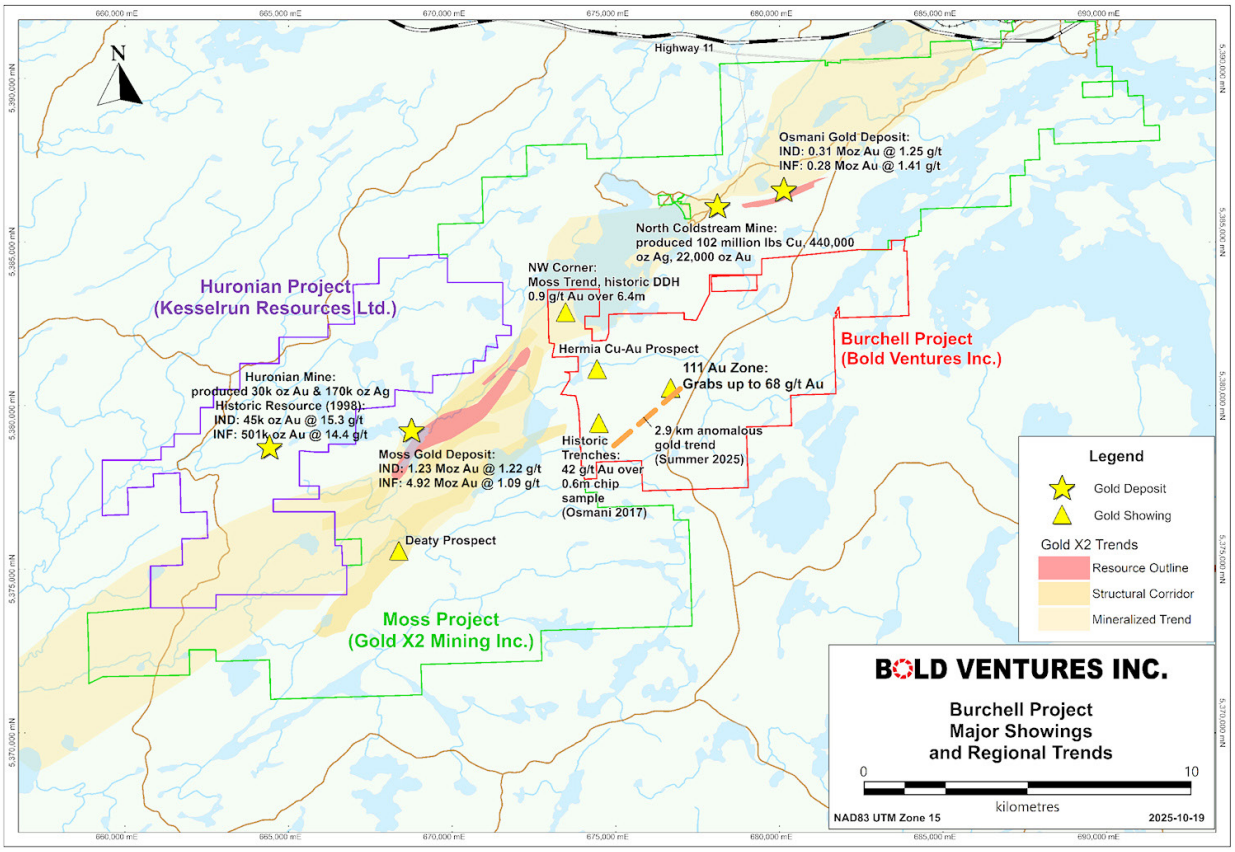


Figure 11. Map of land position, major showings, and 111 Au Zone trench highlighted with red oval (*from* Bold Ventures Inc., *news release,* October 20, 2025).

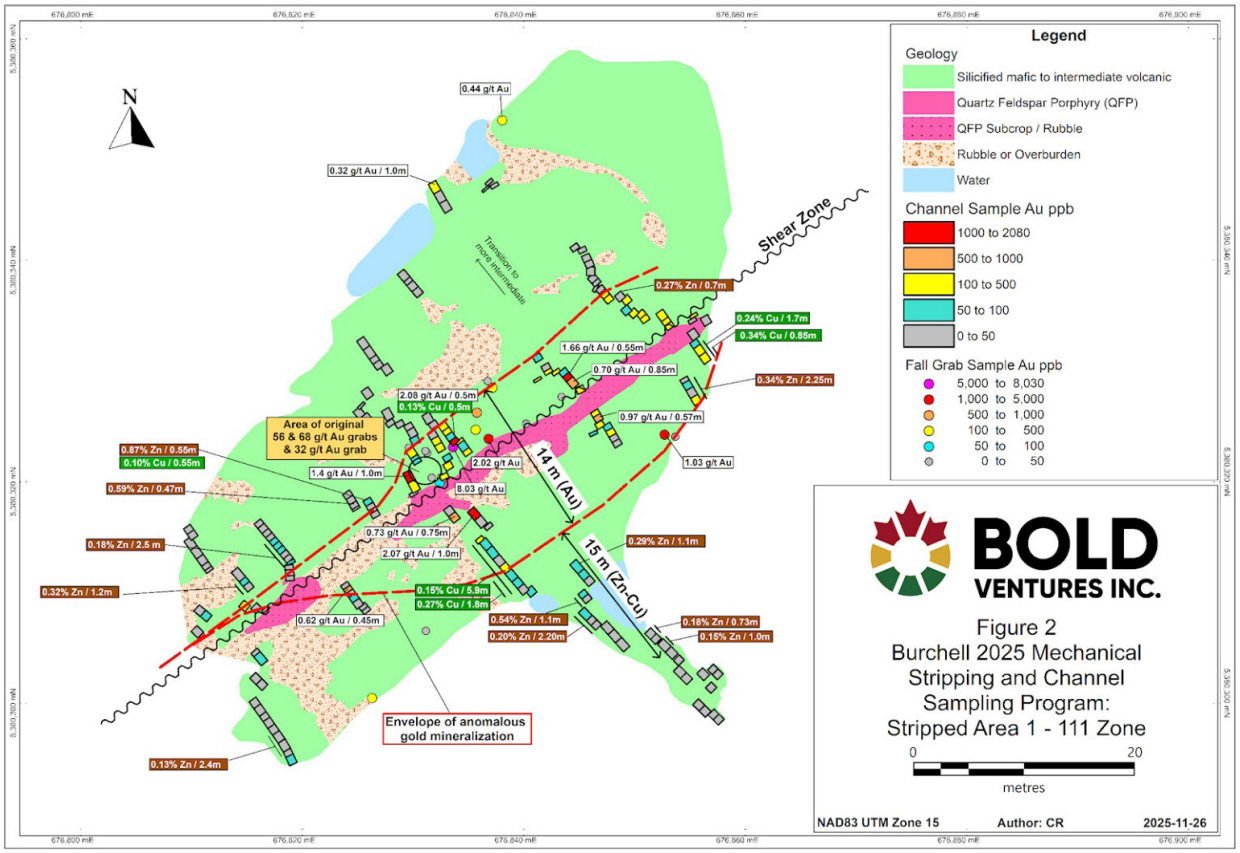


Figure 12. Geology map showing the 111 Au Zone with channel sample results for gold, copper and zinc (*from* Bold Ventures Inc, *news release,* October 20, 2025).

rocks here may either be tectonically interleaved with or conformable with the felsic metavolcanic rocks mapped by Osmani (1997b).

Stop 4. Pillowed vesicular basalt at Swamp River

N83 Z15 U 714018E 5390896N

This outcrop is an example of typical Greenwater assemblage tholeiitic mafic metavolcanic rocks. The outcrop is glacially polished with well-defined striations that trend 25°. Glassy pillow selvages and abundant vesicles are well preserved at this location. Pillows are deformed (~10:1 aspect ratio) in the same orientation as foliation, striking 100° and dipping steeply south. Original mineralogy is replaced by a typical greenschist facies assemblage of chlorite, hornblende, sericite, saussurite, and carbonate (Morin, 1973). Pillow selvages appear to have been loci for fluid movement, as evidence by localization of pyrite and carbonate to the selvages. Morin (1973) mapped these pillows as younging to the north-northeast – can you see this?

Stop 5. Timiskaming-type conglomerate

N83 Z15 U 715392E 5387505N

From Aubet and Campbell (2012).

At this location two facies of the epiclastic suite of Timiskaming-type rocks are exposed. The dominant rock type is poorly sorted, highly foliated conglomerate (Figure 13). Note the heterolithic nature of the fragments, including minor Keewatin-type red jasper fragments. This particular outcrop is highly deformed with the clasts being stretched, forming a well-developed lineation plunging steeply to the southeast. Note the abundant iron carbonate alteration within the sandy matrix. In fault contact with the conglomerate to the west are mudstone and siltstone. Here we have near vertical mineral lineations normal to rolls on the bedding planes. This unit is finely bedded with grading, although present, obscured by the deformation.

Stop 6. Autoclastite

N83 Z15 U 715698E 5385810N

This location is an example of ultramafic metavolcanic rocks of the Greenwater assemblage, featuring a flow-top breccia with a mixture of transported sub-angular, blocky clasts exhibiting some nice examples of random spinifex-textures and



Figure 13. Timiskaming Conglomerate

variolitic textures. The clasts range in size from 0.5 cm to 20 cm in diameter. Although this particular outcrop was not mapped by Rogers (1995), Rogers and Berger (1995) reported other nearby ultramafic metavolcanic units to be generally narrow (<50 m thick) and discontinuous (<1 km along-strike). Both olivine and pyroxene spinifex have been reported in the eastern SGB (e.g. Hinz 2018).

Stops 7 to 10. Delta-1 Au Property (Delta Resources Limited)

Permission required from company to access sites

The Delta-1 Gold property (formerly Shabaqua Gold Project) is located near Shabaqua, 50 km west of Thunder Bay. The area has a long history of exploration dating back to 1930s, where numerous companies and prospectors carried out prospecting, geological, geochemical, and geophysical surveys, trenching, sampling and diamond drilling programs.

While the Eureka deposit is the company's flagship, Delta Resources significantly expanded the Delta-1 property in 2024 by acquiring more than a dozen properties from numerous companies and prospectors. The Delta-1 property now has multiple gold prospects and occurrences covering a 35-km strike extent of several regional-scale structural zones, near or at the unconformity between Shebandowan (Temiskaming-

type) metasedimentary rocks and Greenwater metavolcanic rocks (see black dotted lines in Figure 14).

Stop 7. I-Zone gold-bearing quartz ladder veins

N83 Z15 U 714705E 5382490N

Modified from Aubet and Campbell (2012)

The I-Zone (and associated gold showings) is an exploration target situated proximal to the Crayfish Creek Fault (Figure 14), a major regional structure currently presenting as brittle but likely with a protracted brittle-ductile history. The I-Zone gold occurrence consists of felsic dikes intruding Timiskaming oxide facies iron formation intercalated with argillite. The felsic dikes are host to gold-bearing quartz-tension/ ladder veins with 3%-5% pyrite and localized visible gold (Figure 15). Fractures opened up in the dike due to the ductility contrast of the enclosing iron-rich argillites and the felsic dike. Later hydrothermal fluids, likely carrying gold reacted with the iron oxides resulting in the formation of pyrite and precipitation of native gold.

Historical findings at the site include Landore Resources' 1995 drill program, which intersected 4.32 g/t Au over 41 m, 4.53 g/t Au over 14.4 m, and 4.36 g/t Au over 20.4 m. Additionally, a 2008 mini-bulk sample conducted by Mengold Resources yielded an average grade of 9.9 g/t Au. Portofino Resources Inc. reported 2020 sampling at the I-Zone returned up to 45.9 g/t Au with 6 of 14 samples returning more that 5 g/t Au (Portofino Resources Inc., news release, November 17, 2020).

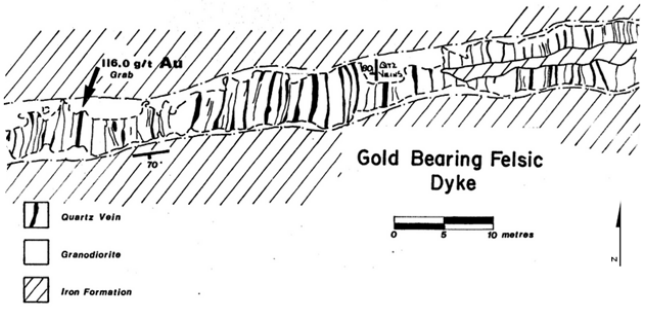


Figure 15. Simplified geology at the I-Zone (modified from Aubut et al., 1990).

Stop 8 Eureka Zone (2024 Delta-1 Eureka Trench above drill hole D1-23-60)

N83 Z16 U 290200E 5385348N

In 2017, Doug Parker and Barbara D'Silva generated renewed interest in gold exploration in the Shabaqua area, on the eastern limb of the SGB. The Parker-D'Silva team followed up on historical data and OGS gold-in-till anomalies with prospecting, mechanical stripping, and rock sampling, which successfully led to the discovery of the Eureka Gold Zone (Figure 16).

In 2019, Mr. Parker optioned the property to Delta Resources Inc. (Delta). Since then, Delta has advanced the project with 140 diamond drill holes (totaling more 40 000 m), confirming a mineralized zone for more than 2.5 km along strike and to a depth of 400 m (Figure 17). The Eureka Zone is situated adjacent to the unconformity between Shebandowan (<2690 Ma) and Greenwater (2720 Ma) assemblages. The

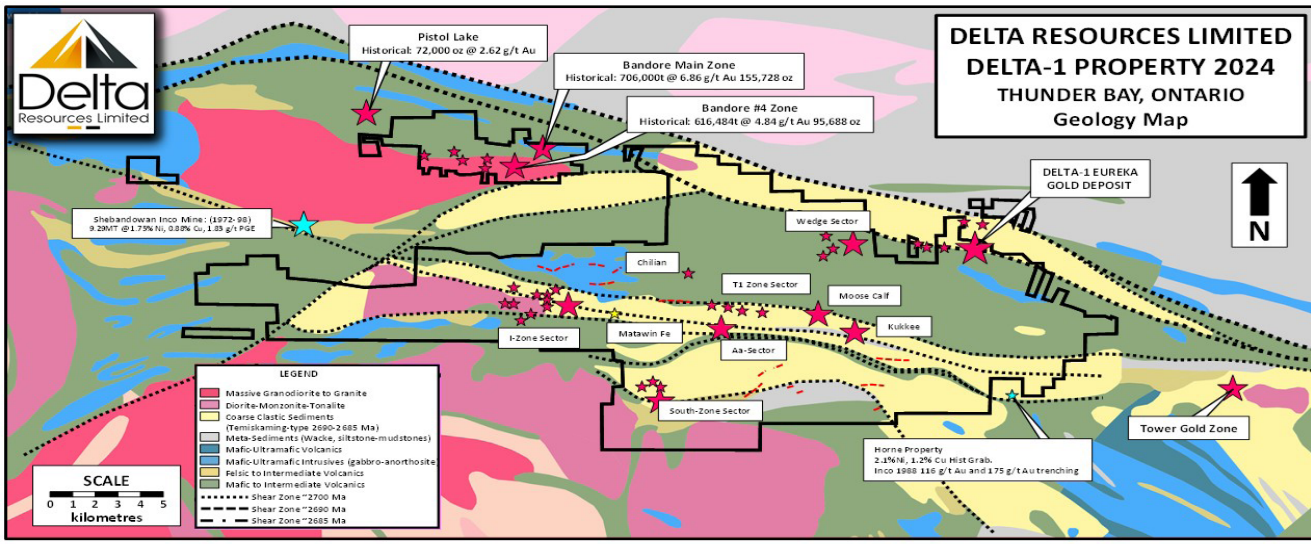


Figure 14. Geology map showing Delta 1 Gold property (black outline), regional structural zones (black dashed lines) and gold showings (red stars) situated at or near the uniformity between Greenwater-Shebandowan assemblages (*from* Delta Resources Inc. website, *Projects; Delta-1; Regional and Property Geology*, April 2026).



Figure 16. Doug Parker's 2017-18 prospecting and mechanical trenching programs generated renewed interest in gold exploration in the Shabaqua area.

unconformity between the Greenwater (ultramafic and mafic to intermediate metavolcanic rocks) and Shebandowan (Temiskaming-type metavolcanic and metasedimentary rocks) assemblages has a close spatial association with gold occurrences, widely known as prospective for gold exploration (Figure 14).

At this stop, Delta's 2024 trenching program exposed an 11 m surface section of the Eureka Gold Zone, directly above drill hole D1-23-60. This drill hole returned an intersection of 1.79 g/t Au over 128.5 m (including 2.16 g/t Au over 97.5 m), while channel sampling from the surface trench returned an average grade of 1.23 g/t Au over 11 m (Delta Resources Inc., *news releases*, September 12, 2023, and September 25, 2024). Gold mineralization at the Eureka Zone is hosted by a stockwork of 1 mm to 10 cm wide quartz-ankerite-pyrite veinlets. The quartz-ankerite-pyrite gold veinlets crosscut all lithologies and are hosted within a 300-400 m wide corridor of ankerite-silica-sericite altered rocks. The Greenwater assemblage host rocks at this stop are comprised of mafic metavolcanic and ultramafic flows, weathered to a dark rusty brown with rock textures nearly or completely obliterated (Figure 18).

Stop 9. Bylund Trench

N83 Z15 290490E 5385211N

At this stop, stockworks of gold-bearing quartz-ankerite-pyrite veinlets are situated within a broader 300-400 m carbonate-sericite-silica-altered halo that hosts anomalous/low-grade gold mineralization. The mineralized trend strikes ~110° and dips approximately 50-55° north. Mineralization is hosted within Greenwater assemblage rocks – most commonly, a feldspar-phyric tholeiitic basalt. That unit is not seen at this trench; what we see here a silica-rich rock that has historically been interpreted as chert but is being re-evaluated by the company as at least partially comprising highly silicified metavolcanic rocks. On the northeastern end of the trench, silicified komatiite or komatiitic basalt is present, displaying beautiful spinifex texture. Intense pyrite-ankerite alteration is widespread, but gold grades from this trench are relatively low.

Three generations of quartz-carbonate veins are visible at this trench: i) NE-trending, steeply NW-

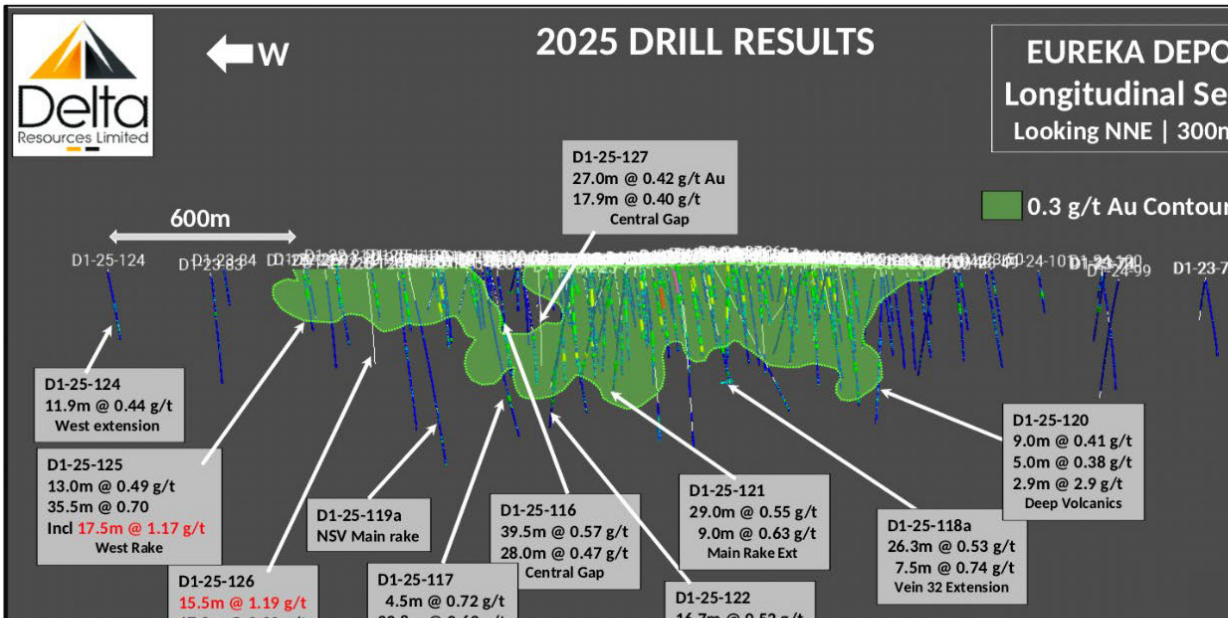


Figure 17. Longitudinal section of the Eureka Zone



Figure 18. Eureka Zone stockwork of 1 mm to 10 cm wide gold-bearing quartz-ankerite-pyrite veinlets hosted by Greenwater assemblage rocks are weathered to dark rusty brown with rock textures nearly or completely obliterated.

dipping, ii) NW-trending, steeply dipping, and iii) NE-trending, shallowly dipping. Vein sets 1 and 2 are conjugate and are post-dated by vein set 3. Vein set 1 is the main gold-bearing set.

Stop 10. Finmark Metasedimentary Rocks

N83 Z16 U 293525E 5383950N

From Puumala and Cundari (2023)

At this stop we will have an opportunity to view a remarkably well-preserved roadside exposure of Shebandowan assemblage clastic metasedimentary rocks. The following description of these rocks is provided by Carter (1990).

> *The rocks are mainly thinly bedded, the beds ranging in thickness from 5 cm to 12 cm. Primary sedimentary structures comprising load casts and flame structures, small-scale ripple structures, and cross bedding, are well developed in these rocks in the road exposures along Highway 11-17 about 2.5 km west of the eastern boundary of Horne Township, and in the outcrops immediately southeast of these.*

Parker (1980) indicates that the "Finmark metasedimentary belt" consists of sandstone-siltstone-mudstone sequences that alternate with thick units of cross-stratified sandstone. These sequences display many of the primary sedimentary structures that are characteristic of tidal flat (e.g., rhythmic layering, lenticular, wavy and flaser bedding) and tidal channel (e.g., herringbone cross stratification, large scale cross-stratification) depositional environments respectively. Petrology of these rocks indicates that the primary sediment source was a felsic to intermediate volcanic terrain (Parker 1980). Characteristics of the clasts and rock fragments are consistent with a proximal Shebandowan assemblage sediment source.

Stop 11. Auto Road Assemblage (Optional)

N83 Z16 U 313838E 5377726N

The Auto Road assemblage comprises a small sedimentary basin in south-central Ware township. It was first provisionally subdivided by Corfu and Stott (1998) on the basis of a U-Pb in youngest detrital zircon age of 2682±3 Ma – this is 9 m.y. younger (and outside of error provisions) than the youngest detrital minerals (zircon and titanite) in the Shebandowan assemblage. The assemblage is affected by D_2 deformation and therefore provides a lower constraint on both sedimentation and regional transpression in the SGB. Corfu and Stott (1998) comment:

> *The map pattern suggests that this conglomerate-sandstone unit is interbedded with Greenwater assemblage basaltic units (Brown, 1995), yet the polymictic conglomerate includes feldspar-hornblende-phyric volcanic clasts typically found within the Shebandowan assemblage. Also common are coarse granitoid cobbles as well as clasts of various volcanic lithologies. The results for sandstone sample Au presented below demonstrate that this is indeed one of the youngest supracrustal units of the Shebandowan greenstone belt as well as of the neighboring Quetico Subprovince, justifying its separate designation.*

At this location, felsic intrusive, chert, and felsic to mafic intrusive clasts of up to 40 cm in size are deformed (up to ~5:1 aspect ratio) by D_2 transpression. Mafic clasts are deformed to a roughly uniform degree, while felsic clasts vary from similarly deformed as mafic clasts to nearly undeformed.



Figure 19. Polymictic conglomerate of the Auto Road assemblage.

Acknowledgements

We would like to thank Gold X2 Mining Inc., Bold Ventures Inc., and Delta Resources Limited for permission to access parts of their properties and for their generous time, knowledge and support in preparing for this field trip.

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Trip 5 - Archean Geology and Metallogeny of the Rainy Lake Wrench Zone

K. Howard Poulsen

Geological Consultant

A little learning is a dangerous thing;
Drink deep, or taste not the Pierian spring:
There shallow draughts intoxicate the brain,
And drinking largely sobers us again.
Fired at first sight with what the Muse imparts,
In fearless youth we tempt the heights of Arts,
While from the bounded level of our mind
Short views we take, nor see the lengths behind;
But more advanced, behold with strange surprise
New distant scenes of endless science rise!
So pleased at first the towering Alps we try,
Mount o'er the vales, and seem to tread the sky,
The eternal snows appear already past,
And the first clouds and mountains seem the last;
But, those attained, we tremble to survey
The growing labors of the lengthened way,
The increasing prospects tire our wandering eyes,
Hills peep o'er hills, and Alps on Alps arise!
Alexander Pope, 1711

Foreword

Rainy Lake is a body of fresh water which straddles the border between Ontario, Canada and Minnesota, USA. It includes approximately 2600 km of rocky shoreline plus more than 1600 islands and covers an area of approximately 930 square kilometers. Rainy Lake is fed from the east by the Seine River waterway and is drained westward by the Rainy River which leads to the even larger Lake of the Woods and the Winnipeg River system (Fig. 1). For centuries it has been part of the historic water link between the Atlantic and Arctic watersheds: it was known as Tekamaniwen to the indigenous inhabitants of the region and as Lac a la Pluie to the French voyageurs and fur traders. Rainy Lake and Lake of the Woods are remnants of the vast glacial Lake Agassiz which formed by melting of the Wisconsin continental ice sheet approximately 13,000 years ago. The predominately Archean bedrock in the Rainy Lake region (Figs. 1, 2) is now exposed in arched, glacially-sculpted outcrops within areas of generally thin and discontinuous surficial cover overgrown by boreal forest.

My first visit to Rainy Lake was in summer 1966 when I helped a geophysical operator evaluate the long-wire electromagnetic survey method for our employer Dr. Ray Oja, a geological consultant working out of Thunder Bay. The equipment test was focused on the Grassy Portage Bay property of Noranda Mines Ltd. which included copper mineralization on their Halkirk

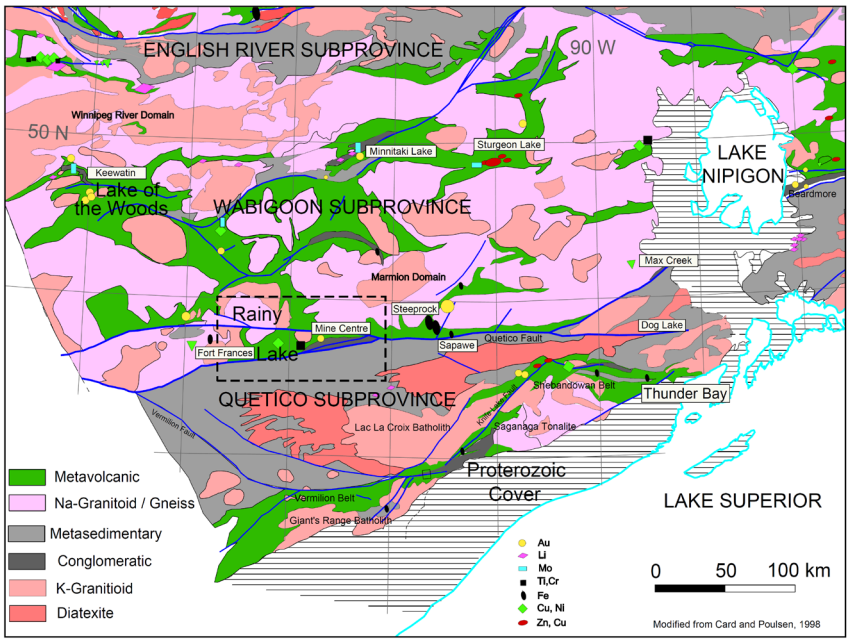


Figure 1: Southwestern Superior Province with locations of selected mineral deposits. The area of Figure 2 is outlined by the dashed rectangle.

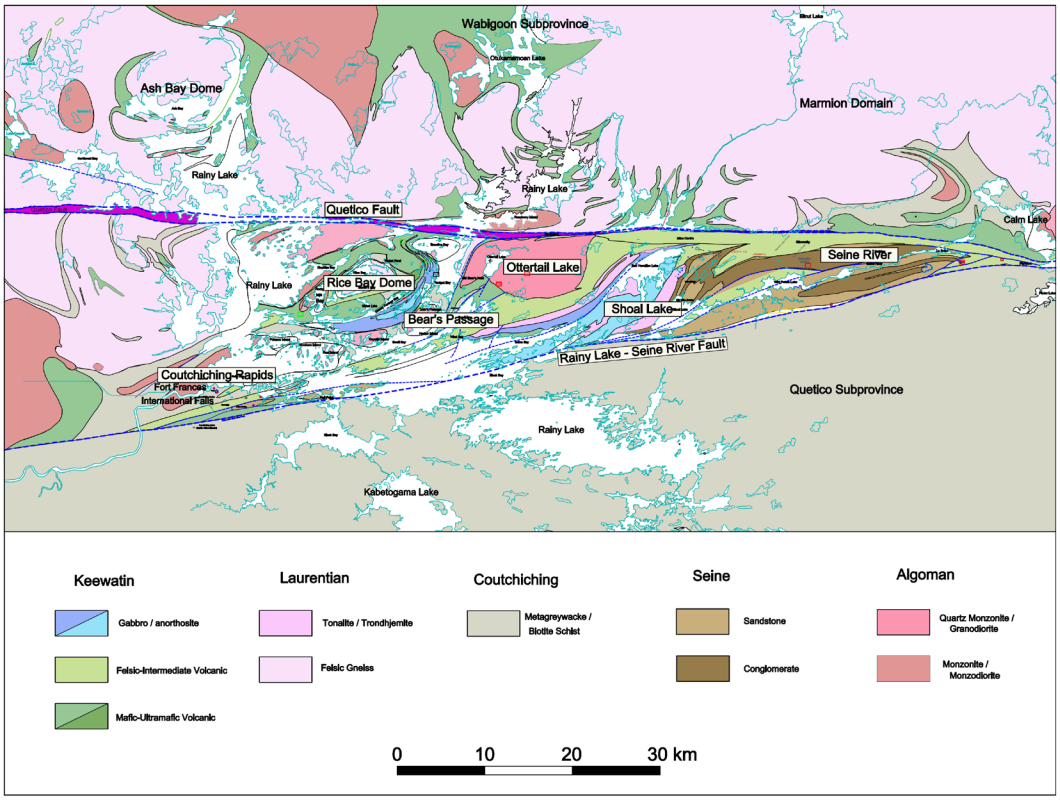


Figure 2: Rainy Lake Geology

– Watten (Northrock) prospect: C.J. Hodgson who later became one of my thesis supervisors at Queen's had completed his MSc thesis on this deposit in 1959. I returned to the property in the early 1970's with Dr. Mel Bartley who was then consulting for Northrock Mines and I helped him log a section of diamond drill core from the deposit which is located on the south flank of a feature known as the Rice Bay Dome (Fig. 2). Mel, who was a well-regarded geologist and one of the founders of Lakehead University, also consulted around that time for George Armstrong of Fort Frances. George was a successful highway construction contractor who, along with Mike Hupchuk, was also an avid part-time prospector. They had discovered Zn mineralization in 1971 near Pocket Pond east of Rice Bay. Mel and I made a site visit to Pocket Pond in fall 1972 and I prepared a report on the geophysical data for the property (Poulsen, 1973). At that time, I noted the existence of abundant outcrops along the drill roads near Armstrong's trenches which were extremely large for the time - they had been excavated by his road construction crew!

Jim Franklin, for whom I had been a research assistant at Lakehead University, left in July 1975 to join the GSC in Ottawa while I became a full-time technician in the geology department. I also began an independent look at roadside outcrops around Thunder Bay with a view toward identifying a possible thesis topic with Dick Ojakangas, the well-regarded sedimentologist at Duluth who was also famous for his Finn jokes. Around the same time, however, I told Jim about the interesting geology and the massive sulfide mineralization at Rainy Lake and we decided to investigate further. Armed with Andrew C. Lawson's classic GSC Memoir 40 as well as Fred Harris' more recent Ontario geology maps as guides, we visited several outcrops on Rainy Lake in summer 1976. In particular we re-visited the Pocket Pond property where we confirmed my original observation that, based on pillow shapes, the strata appeared to be overturned. We also visited Lawson's classic outcrops at Bear's Passage using a Zodiak boat only to realize when we got there that they now are located at a boat launch site which is easily accessible by road! We nonetheless had also found the key outcrops and agreed that the staurolite-bearing metasedimentary rocks are also overturned. We later met John Wood of the Ontario Geological Survey who was mapping to the east at Mine Centre and he showed us outcrops of Lawson's Seine conglomerate and of gold-bearing quartz veins near Bad Vermilion Lake.

Once Lakehead University gained approval for its own M.Sc. program in geology, I applied to become the first (part-time) graduate geology student at Lakehead to study the problem of the apparently

overturned stratigraphy around the Rice Bay Dome and Bear's Passage. The thesis work began in 1978 under the supervision of Manfred Kehlenbeck and the newly hired structural geology professor, Graham Borradaile. The resulting structural analysis showed that, contrary to the long-standing interpretation of the Rice Bay Dome as originally proposed by Lawson, the evidence was clearly in favour of downward facing folds and extensive stratigraphic inversion but it was not easy to convince others of this: I had, somewhat unwittingly, stumbled into a larger problem that had already dominated the discussion of the Archean rocks northwest of Lake Superior for many decades.

The Institute on Lake Superior Geology was initiated in 1955 as an annual meeting, most times with companion field trips, to discuss developments in geological understanding of both the U.S. and Canadian side of Lake Superior. Early meetings emphasized iron ore deposits which were of the greatest economic importance at that time but evolved into an exploration of a much more eclectic range of topics. Mel Bartley and Ed Pye of Port Arthur were among the early participants and Samuel S. Goldich, a pioneer of geochronology, was a founder and frequent contributor. I attended my first ILSG meeting at Madison, Wisconsin in 1973. Among the speakers were Paul Sims and Klaus Schultz of the U.S. Geological Survey and Don Davidson and John Green of the University of Minnesota at Duluth. A memorable and perhaps prophetic moment came during a presentation on Proterozoic stratigraphy of the Lake Superior area by the mild-mannered John Green who suggested that the Puckwunge Formation should be excluded from the Keweenawan Group. The proposal brought loud and angry condemnation by the short, red-faced Sam Goldich even before the talk was completed. As it turns out, although Goldich was a painfully shy and quiet individual in social situations, he was equally fierce and combative in professional settings. This proved to be the case again in 1976 at the ILSG meeting at St. Paul Minnesota. I attended Goldich's excellent field trip to the Archean gneisses of the Minnesota River Valley, including a visit to a small outcrop in a swamp where he believed he had sampled and analysed the oldest rock on Earth as reported by that time. When one of his former graduate students questioned the statistical validity of Goldich's data regression, the offender was told in no uncertain terms that, if he didn't like the method, he could just leave the field trip immediately! At that time Goldich was a member in high standing of an international group of geologists with a strong interest in Precambrian geochemistry and geochronology. The Canadian leader within this group was Alan M. Goodwin of the University of Toronto who had conceived of and organized the multidisciplinary Superior Geotraverse Project which ran from 1970 to 1978. Near the project's end Goodwin organized the Archean Geochemistry Conference in summer 1978 to highlight the significant results. This meeting involved the "who's who" of Precambrian geochemistry at the time and, after a series of conference presentations at the Quetico Centre, the group headed west to Fort Frances-International Falls. Sam Goldich and Zell Peterman led a one-day field trip on route to illustrate aspects of the geology at Rainy Lake. Peterman had completed his MS thesis with Goldich in 1959 on the metasedimentary rocks of the Rice Bay Dome and now was a geochemist with the US Geological Survey in Denver. With prior arrangement by Jim Franklin who was a formal participant, I was able to tag along unofficially and silently as a beginning graduate student. The emphasis at each stop was placed on the chemical composition of rock units as recorded on hand-written file cards which Goldich drew from a deck at each outcrop. At one exposure there was considerable debate about whether a xenolith-rich lamprophyre dike might actually be a new locality of the Seine conglomerate but the important localities at Bear Passage and Pocket Pond were not part of the field trip. The classical stratigraphic interpretation of the eminent geologist Andrew C. Lawson was adhered to and I was not in any position to offer an objection.

I made my first formal presentation on "Polyphase Deformation of Archean Rocks at Rainy Lake, Ontario" on May 10th of the following year at the Institute of Lake Superior Geology meeting at Duluth where I made the case for overturned strata around the Rice Bay Dome, contrary to Lawson's original interpretations. Sam Goldich, who often referred reverently to "Professor Lawson", was upset by this, so much so that he was unable to speak to me about it in person: he sent a delegation of Zell Peterman and Paul Sims instead to voice his displeasure. Both were apologetic and conciliatory and asked if it would be possible for me to arrange a field trip to visit the outcrops in question and this was set for later in the summer. In addition to Goldich, Sims and Peterman, the field trip participants included Dick Ojakangas

and David Southwick from Minnesota as well as the Ontario geologists John Wood, Charlie Blackburn and Dick Beard. The event did not start well with Goldich clearly muttering something to the effect of "young punks don't know anything" but the situation improved somewhat with successive stops. The second to last was at Lawson's famous Bear's Passage exposure: the group hadn't fully had a chance to look closely at the outcrop before Dick Ojakangas, with a pointing of his thumb, indicated the southwestward direction of younging of the northwestward-dipping graded beds. This prompted Goldich to become agitated and declare that he didn't believe graded bedding was a reliable criterion: Ojakangas, who had completed his PhD on the Cretaceous turbidites of the Great Valley Sequence in California, responded calmly that he had measured at least a thousand similar beds there without conflict. The final stop was at the exposure of pillow basalt near Pocket Pond where the field relationships proved to be even more convincing. Paul Sims, a no-nonsense individual whose standard dinner included a martini, a rare steak and a salad, followed by a foot-long cigar, had remained quiet throughout the day but, when confronted with the outcrop, he turned to Goldich and said "there's no question about this Sam, the section is overturned". This prompted Goldich to smile, walk over and shake my hand, saying "well young man, you showed me something important today that I didn't know before – let's go back to Fort Frances and drink some of that "Canadian" beer". He was always cordial to me from that point onward and made a point of connecting again in the field the following season. That one-day field trip was essential in demonstrating the credibility of the field observations and a second presentation on overturned Archean successions at the 1980 ILSG meeting at Eau Claire met with little resistance. A companion journal paper which previously had been rejected by the editor was ultimately accepted for publication with minor revisions by the Canadian Journal of Earth Sciences on June 17, 1980. The involvement of representatives from the Ontario and Minnesota geological surveys also proved to be important. Charlie Blackburn and John Wood later asked me to incorporate many of the field stops into one leg of a multi-day OGS-led excursion on Western Wabigoon Geology for the May 1982 GAC Meeting at Winnipeg. Dick Beard also lobbied hard for the funding of my subsequent work for Sandy Colvine of the Mineral Deposits Section of the OGS on the mineral deposits of the Mine Centre-Fort Frances area.

Much of the field work for the OGS had been completed by 1981 and formed the basis of a third ILSG presentation at International Falls in 1982. Dave Southwick of the Minnesota survey was the organizer of the meeting and late in 1981 asked me if I would lead a related field trip focussed on the mineral deposits of the area. I agreed and we set a limit of 25 participants but this was a period of renewed in interest in gold exploration so registration quickly filled up. Dave contacted me again in the New Year and asked if we could double the limit to 50 participants and I reluctantly agreed but that limit was also reached in a short time so Dave developed a waiting list which grew to more than 20 requests. He contacted me once more to ask whether I would accommodate additional participants if he could find a Greyhound bus and provide shuttle vans and drivers to speed up the logistics at some field stops. I once again agreed and an exhausting, but gratifying, one-day, 10-stop field trip was delivered to 77 participants on May 5, 1982. Two weeks later, John Wood and I also led a field trip with a structural-stratigraphic focus as part of the larger excursion organized by Charlie Blackburn for the Winnipeg meeting.

What follows is an attempt to illustrate the historical development of ideas about Rainy Lake geology using outcrops in the Mine Centre – Fort Frances corridor (Fig. 2). It includes a concise historical overview and a summary account of the highlights of the regional geology followed by updated descriptions of representative field localities. The motivation for doing this involves three main considerations. The first is purely practical and involves the precision and accuracy of outcrop locations. After the passage of more than forty years, some of the sites described in the guidebooks of 1982 are difficult to re-locate, especially for someone without prior knowledge of the area. Most mapping at that time was largely carried out on non-rectified aerial photographs with considerable local distortion and the modern digital tools of geographic positioning were not available. Furthermore, destruction of vegetation in some areas and new growth in others has rendered past bush trails and landmarks to be obscure, if not impossible, to recognize, even for the author of the guidebooks. Abandonment of old bush roads has also given way to new gravel access roads and the widening of highways has compromised some outcrops while exposing new ones. A second consideration is the currency of information and ideas.

The work in the late 1970's and early 1980's was done within a limited context without much consideration of comparable situations globally: this is somewhat ironic because one of Lawson's goals had been to use this region as a global type example of an Archean granite-greenstone belt. It is also important to recognize that new contributions have been made to the understanding of the geology of this area since the original guidebooks were written. The final consideration is historical. In its time the "Seine-Coutchiching problem", rooted in the simple but laborious task of geological field mapping, influenced discussions among geologists world-wide but many of the details about the background to the debate have not been adequately recorded, especially in the context of the outcrops themselves. Every effort has been made to avoid duplication of points which are adequately covered in the past guidebooks for this region and the material presented below is meant to serve mainly as a source of information for field trip leaders and new researchers to draw on to supplement the existing documents.

The Seine-Coutchiching Problem

The discovery of gold at Lake of the Woods in 1878 followed by the building of the C.P.R. line prompted Arthur Selwyn, director of the Geological Survey of Canada, to instruct one of its senior mappers to begin a survey of the geology of this area. Part of the task given to Robert Bell and his young assistants, A.C. Lawson and J.W. Tyrell, was to have the geology carefully worked out as a type locality for the "so-called Huronian system" (Zaslow, 1975, p. 184). Bell left Lawson and Tyrell at Bigstone Bay in spring 1883 to map the shoreline geology and topography while he surveyed a line northward toward Red Lake. When he returned, he checked their results and directed them to work separately, Lawson on geology and Tyrell on topography, before they all reconvened at Rat Portage (now Kenora) at the end of the season. By the beginning of the 1884 field season, Andrew Cowper Lawson had graduated from the University of Toronto with the gold medal in natural science and was put in charge of the project. By then, at age 23, he had been taken on staff at the Geological Survey of Canada and, during that field season, he continued the geological work at Lake of the Woods. A.E. Barlow and W.H.C. Smith independently mapped the topography southward toward Rainy Lake. Lawson's geological report and maps resulting from the work conducted at Lake of the Woods were published in 1885, questioning the approach the Survey had taken in the mapping Precambrian rocks up to that time (Zaslow, 1975). Lawson (1885) argued that the greenstone which he termed "Keewatin" is clearly intruded by foliated granitoid rocks. He termed these "Laurentian" in keeping with the original GSC terminology introduced by its first director W.E. Logan to denote quartzo-feldspathic basement gneiss upon which all supracrustal rocks had been deposited. Although Lawson's productivity was admired, his geological interpretations were doubted by more senior geologists.

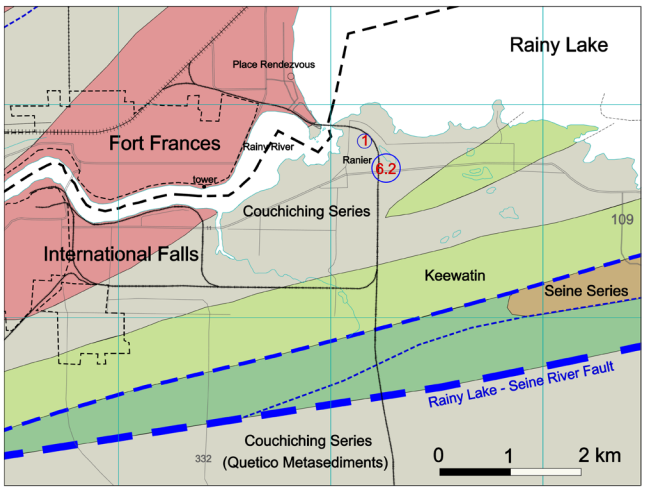


Figure 3: Simplified geology of the Coutchiching Rapids area using the colour scheme of Figure 2.

Perhaps because of the attention he gained and the fact that he had by now received an M.A. from the University of Toronto, Lawson was able to prevail on Selwyn to support his further academic advancement. He was allowed to attend courses during the winter in the U.S.A.: he was the first of many GSC geologists to follow this course of action for decades to come. Lawson, with Smith as topographer, began by mapping the canoe routes between Lake of the Woods and Rainy Lake in 1885 and in 1886 focussed on systematic mapping of the Rainy Lake area (Fig. 2). The following season, Lawson filled in the details in the Rainy Lake district while Smith moved eastward along the Seine River Route and southward to the Canada-U.S. border. Lawson recognized a series of metasedimentary rocks exposed along the Coutchiching rapids at the outlet of Rainy Lake into Rainy River at modern-day Fort Frances and International Falls (Fig. 3). He traced these rocks farther northeastward from the type locality into the Rice Bay and Bear's Passage areas (Figs. 4, 5) where the evidence suggested that rocks of the Coutchiching series are even older than the Keewatin (Fig. 6a). This

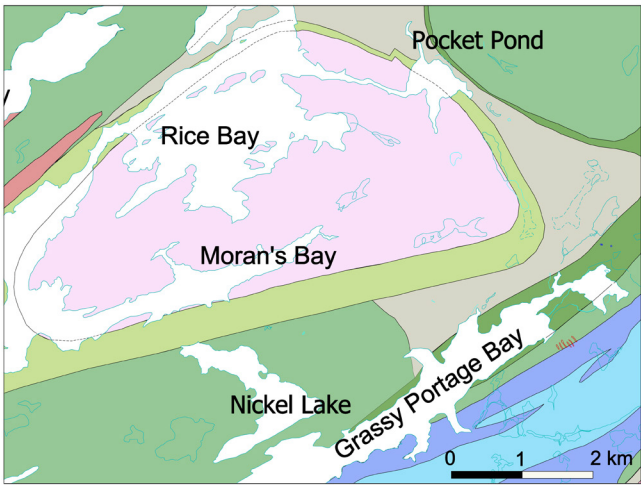


Figure 4: Simplified geology of the Rice Bay area

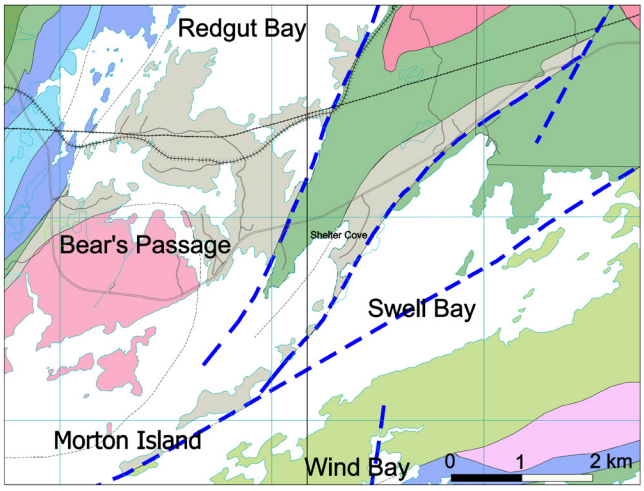


Figure 5: Simplified geology of the Bear's Passage area. Note that "Bear's Passage" refers to the strait linking Swell Bay to Redgut Bay but the terms "Bear Pass" or "Bear Passage" have also been used over time to describe the nearby area.

only added further to his dispute with GSC management leading to the heavy editing of his map and first report on Rainy Lake geology (Lawson, 1888). He also provided additional evidence that, rather than being a fundamental basement gneiss, the Laurentian rocks are actually deformed and metamorphosed intrusions that show evidence of cross-cutting the supracrustal rocks. A hand-written version of the report was also submitted for his PhD thesis at John's Hopkins University (Lawson (1888) where he applied the relatively new technique of optical petrographic description to thin sections from his field samples.

Although the accomplishments of Lawson and his colleague W.H.C. Smith were significant, it was the resulting geological interpretation that met continued resistance from management and the 1887 report was heavily edited (Saslow, 1975). Lawson, however, largely prevailed and showcased his results at the Fourth International Geological Congress at London in 1888 and the American Association for the Advancement of Science meeting at Toronto in 1889. During the 1889 field season Lawson completed the mapping of the Hunter Island Sheet southeastward of Rainy Lake with Smith but resigned from the Geological Survey of Canada in spring 1890. He worked for a while as a geological consultant in Vancouver but soon accepted a faculty position at the University of California at Berkley, where he pursued an illustrious career for the

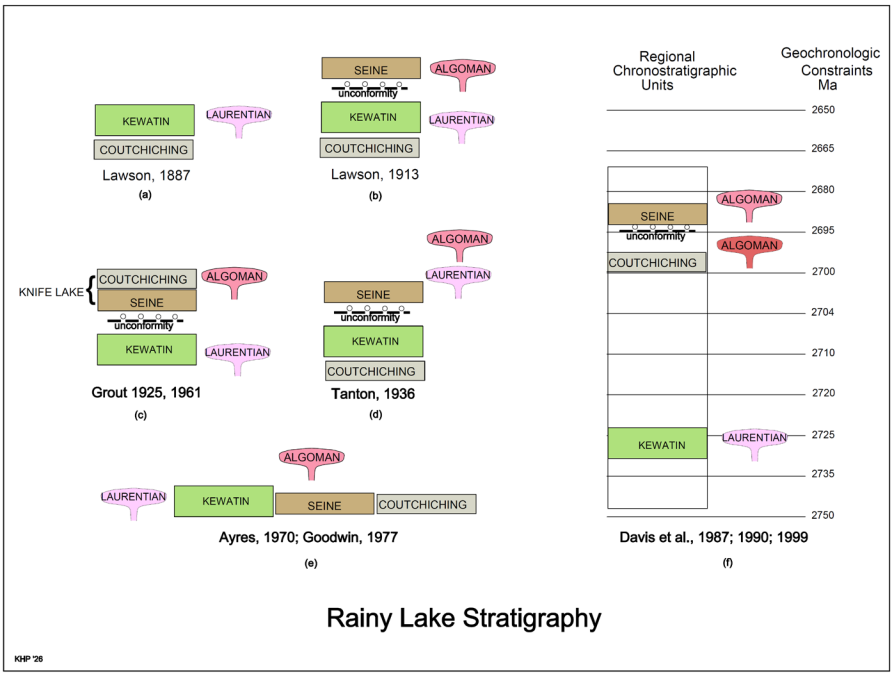


Figure 6: Portrayals of stratigraphic order at Rainy Lake (1887-1999).

next sixty years.

Although increasingly accepted overall, some aspects of Lawson's interpretation of Rainy Lake geology continued to be questioned by his peers. In particular, Coleman (1898) noted that the sedimentary sequence at Shoal Lake (Fig. 7), portrayed by Lawson as belonging to the Coutchiching, includes conglomerate with abundant rounded clasts of both Keewatin greenstone and Laurentian granitoid rocks which are also exposed nearby. The U.S. Survey, which was responsible for mapping the southward extensions of the area covered by Lawson and Smit, took particular exception to Lawson's interpretations. A special committee on stratigraphic nomenclature for the Lake Superior region was therefore convened by the U.S. Geological Survey and the Geological Survey of Canada and the resulting report was published in the Journal of Geology (Adams et al., 1905). It was critical of Lawson's interpretation and suggested that there is evidence for the Coutchiching rocks to be younger than the Keewatin, a point that was later re-affirmed Minnesota by Van Hise and Leith (1909).

Lawson was irate over the findings of the special committee and R.W. Brock, who was by then the director of the Geological Survey of Canada, invited Lawson to re-study the key parts of his original Rainy Lake map sheet in 1911. Lawson was also given the mandate to examine the rocks farther east along the Seine River toward Steeprock Lake and Sapawe. Several practical developments had ensued since the first mapping, including a gold rush to Mine Centre in the 1890's, construction of the CNR south line through the area circa 1906 and major forest fires in the region in 1910, generating much new bedrock exposure. By then Lawson was in mid-career and had gained pre-eminence in many aspects of geological science in the western U.S. so that, when he produced his famous Geological Survey of Canada Memoir 40 in 1913 and an accompanying map in 1914, they were accepted without revision. In the memoir he reaffirmed his interpretation of the field relationships in the Rice Bay and Bear Passage areas where he observed the Coutchiching metasedimentary rocks to dip at moderate angles below Keewatin strata. He also issued a bitter challenge to the members of the special committee (Memoir 40, p.13-14): *"The facts here recited in regard to this line of contact, particularly near the railway on the shores of Bear Passage and the south end of Redgut Bay, taken in connexion with the relations of the Coutchiching to the granite, appear to me to prove conclusively the superposition of the Keewatin upon the rocks mapped by me as Coutchiching in the report of 1887. I invite the attention of the International Committee and of the U.S. Geological Survey to this section and challenge them in view of the facts there*

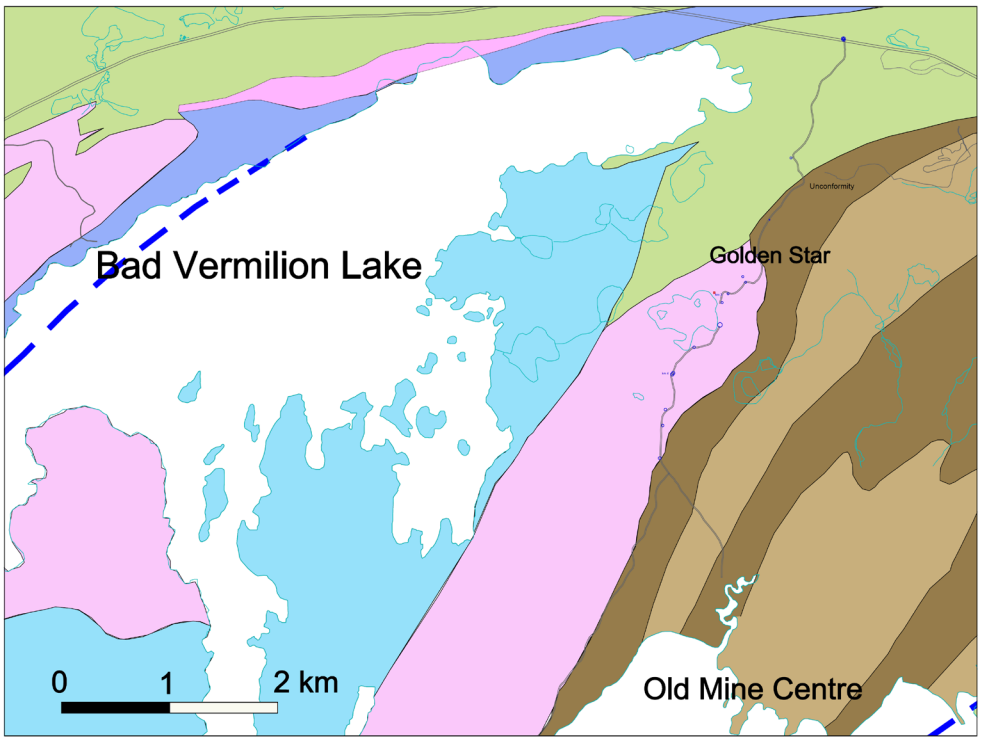


Figure 7: Bad Vermilion Lake area

apparent and easily accessible, to deny the relations of the Keewatin and Coutchiching as I mapped and described them a quarter of a century ago. The fact that these eminent authorities have denied in toto the existence of the Coutchiching series as a constituent member of the Archean below the Keewatin, without any attempt to verify the very explicit statement of the evidence in regard to this section contained in the report of 1887 places them in a curious light from the point of view of scientific method."

While forcefully retaining his contention that the Coutchiching rocks at Rice Bay (Fig. 4) and Bear's Passage (Fig. 5) are positioned below the Keewatin, Lawson also admitted to an error in the Bad Vermilion Lake area (Fig. 7). There he proposed the new term "Seine Series" for the conglomeratic metasedimentary rocks that he had previously included as a basal part of the Keewatin. He now confidently placed the basal Seine conglomerate unconformably above both the Laurentian granitoid rocks and the Keewatin metavolcanic rocks in that area (Fig. 6b). He also noted the presence of trough crossbeds in the sandy portions of the Seine series near Old Mine Centre (Fig. 7) and used the newly recognized criterion of determining the directions of stratigraphic younging using their shapes. This allowed him to define a synclinal fold within the Seine sedimentary sequence in a narrow belt extending eastward along the Seine River (Fig. 2). He also recognized that the Seine Series locally extended farther eastward beyond the Rainy Lake area. In so doing, he mistakenly classified sedimentary rocks at Sapawe (Fig. 1, then known as Iron Spur) as part of the Seine Series. The granitoid Blalock stock cuts metasedimentary rocks discordantly at that locality so Lawson introduced the new term "Algoman" for such intrusions which he believed to be generally younger than the Seine (Fig. 6b). Although subsequent studies at Sapawe support Lawson's contention of a late-tectonic intrusion, they also have consistently portrayed the intruded sedimentary rocks there as part of the Coutchiching rather than the Seine. Nonetheless, Lawson offered other acceptable field and petrographic distinctions that argue for the existence of a younger set of Algoman granitoid intrusions in the Rainy Lake area proper. They tend to contain higher proportions of K-feldspar than the dominantly sodic tonalitic rocks which comprise the Laurentian. He also mapped a narrow band of conglomerate near Hopkins Bay, west of Rice Bay (Fig. 2), and tentatively correlated it with the Seine sequence: at that locality he also presented strong evidence that the conglomerate is cut by younger Algoman granitoid rocks.

As he had in the 1880's Lawson used the International Geological Congress, this time at Ottawa in 1913, to promote his revised view of Rainy Lake geology and Precambrian stratigraphy in general. A debate was staged between Lawson and C.K. Leith to present arguments for and against the findings of the special committee: as later recalled by Leith, Lawson had a "slashing style" and *"while I came out feeling I had presented the facts, I also felt Lawson had chewed me up and thrown me to the wolves"* (Dott, 2001, p.1007). Lawson's views were further solidified during the 1913 International Congress field trip that he led on Sunday, August 17th for approximately 90 participants who had traveled by C.N.R. to the Mine Centre and Bear's Passage train stations after similar visits at Iron Spur and Steep Rock Lake the day before. At Mine Centre, participants were given the option of riding in horse-drawn wagons from the station to the Golden Star Mine along the Shoal Lake Road or taking a short boat ride across Bad Vermilion Lake to a walking trail leading to the mine (Fig. 7). At Bear's Passage, one group was assigned to a boat trip which visited lakeshore outcrops along Redgut Bay and Bear's Passage (Fig. 5) while others made a traverse though a similar geological section exposed relatively new rock cuts along the C.N.R. railway line. A carefully prepared itinerary and field guide for both of the historic localities (Uglow, 1913) allowed Lawson to illustrate the nature of each of his five stratigraphic units and his observations on the relationships among them.

Lawson's revised interpretation of the geology of the Lake of the Woods and Rainy Lake regions prevailed for another decade (Bruce, 1925) before the idea that there was still a "problem" was revived by F. F. Grout of the Minnesota Geological Survey. Grout (1925) reviewed the field relationships at the type locality of the Coutchiching near International Falls (Fig. 3) and at outcrops which Lawson had assigned to the Seine in the area of Neil Point farther to the east. Grout affirmed Lawson's use of cross-bedding as an indicator of stratigraphic younging at Neil Point but offered an alternative overall interpretation which placed both the Seine and Coutchiching above the Keewatin. He then moved northeastward across the international boundary to a locality south of Bear's Passage known as Morton Island (Fig. 5). There he used the newly recognized

field criterion of graded bedding to deduce that the direction of stratigraphic younging in the Coutchiching is northward and away from the Keewatin volcanic rocks. He also visited the key localities at Rice Bay, Bear Passage, Shoal Lake and also at Jackfish Lake southwest of Steeprock Lake. In all cases he raised objections to Lawson's positioning of the Coutchiching and placed it above both the Keewatin and the Seine (Fig. 6c). The problem expanded in scope when, during a subsequent field examination with T.L. Tanton of the Geological Survey of Canada, it became apparent that Grout had made a significant observational error on the Minnesota side of the boundary by misinterpreting an intrusion breccia to be a conglomerate of sedimentary origin. Tanton (1927) took great pains to publicly point this out at a Geological Society of America Precambrian Symposium, noting that the error was made by "a Minnesota geologist". Grout, in a discussion of Tanton's paper, duly acknowledged his own mistake but also stated combatively that a "structure section sketched in the field by Tanton shows more errors than any before". Rather than resolving the problem, these exchanges only served to accentuate it. Some years later, Tanton (1936) made a comparable error by misinterpreting the Seine conglomerate at Shoal Lake to have been intruded by the Laurentian granitoid rocks rather than being deposited above it (Fig. 6d).

Apart from his local mistake, Grout's overall arguments found some traction and provided the impetus for additional field work. J.E. Hawley, a graduate of the University of Wisconsin and a professor at Queen's University, was well versed in stratigraphy and structural geology. Along with a review of the Shoal Lake area, he conducted a study of the Seine and Coutchiching eastward though Jackfish Lake and past Sapawe. He concluded (Hawley, 1930) that part of the problem was, in some localities at least, that the contacts between the Coutchiching and Keewatin are arguably occupied by faults so that attitudes of strata alone provide inconclusive evidence of stratigraphic order. F.F. Grout also remained influential at that time and recommended the Seine-Coutchiching Problem to P.L. Merritt who conducted a study of the entire corridor from Rainy Lake eastward along the Seine River watershed for his Ph.D. thesis at Columbia University (Merritt, 1934). His conclusions concerning the two metasedimentary sequences supported Grout's interpretation (Fig. 6c) and he suggested that, with the notable exception the clastic rocks at Rice Bay and Bear's Passage, the term Coutchiching should be abandoned altogether and that all other sedimentary units should be included in the post-Keewatin, Seine Series above the basal conglomerate. Like Hawley, Merritt also provided detailed documentation of a fault contact between the Keewatin and sedimentary units at various localities and traced a continuous fault from Calm Lake eastward through Sapawe as far as Dog Lake, 60 km north of Thunder Bay: this is known today as the Quetico Fault. He also proposed (Merritt, 1934, p. 371) that "the fault movement along the contact is believed to combine a horizontal shear with an associated overthrust to the south". Grout had a further influence on the expanding Seine-Coutchiching problem in that he inspired Francis Pettijohn, his field assistant during the work at Rainy Lake, to take on pioneering work in the study of Archean sedimentary rocks in general. Pettijohn did his undergraduate and graduate work at the University of Minnesota and, in accepting the Penrose Medal for 1975 from the Geological Society of America, he acknowledged the importance of Grout's tutelage as well as the short time that he spent studying with A.C. Lawson at Berkley in 1927-28 to learn more about the alternative view. Pettijohn's Ph. D. thesis documented the Abram Lake conglomerate in the Minnitaki Lake area in part because it resembled both the Seine conglomerate at Rainy Lake and the Ogishke conglomerate at Knife Lake Minnesota (Fig. 1). He later summarized his findings at all three of these localities as well as at several others in the northern Lake Superior region (Pettijohn, 1937) to also conclude that the majority of sedimentary units are arguably younger than the Keewatin. He also raised the possibility, however, that not all of the units included in the Keewatin need be of the same age and that local intercalation between volcanic and sedimentary rocks may locally be possible. As important as these insights ultimately proved to be, Pettijohn's resolution of the Seine-Coutchiching problem also called for the abandonment of both of Lawson's local units (Seine and Coutchiching) in favour of an overarching "Knife Lake Series" composed of similar rocks which had precedence of definition in the geological literature in Minnesota. This also reinforced Grout's views and a summary paper on the topic (Grout et al., 1951) notably included a section entitled "No Coutchiching Recognized in Minnesota". Nonetheless, the "Seine-Coutchiching problem" was kept alive intermittently well into the 1970's even to the extreme point of academic speculations that no unconformities existed

at all and that there was simply a proximal to distal lateral facies equivalence (Fig. 6e) among broadly age-equivalent Keewatin, Seine and Coutchiching rocks (Bass, 1961; Ayres, 1971; Mackasey et al., 1974; Goodwin, 1977).

Sam Goldich who was a graduate of the University of Minnesota rejoined that institution in 1948 as a professor and director of the Rock Analysis Laboratory where he and Alfred Nier gained international reputations as pioneers in isotope geochemistry and geochronology. An outcome of that work was the application of geochronological methods to the resolution of stratigraphic problems in the Lake Superior region (Goldich et al., 1961; Goldich, 1968). Goldich compromised on the question of the Seine-Coutchiching problem by favouring the term "Knife Lake Group" over "Seine Group" above the Keewatin but also allowed for the possibility (using a question mark for emphasis) of the existence of Coutchiching metasedimentary rocks below it. Goldich tackled the geology of the Rainy Lake area head on by supporting three field-based M.S. theses at the University of Minnesota (Alt, 1959; Frye, 1959; Peterman, 1959) and the resulting maps and samples became the basis for on-going geochronology and geochemical studies (Peterman et al, 1972; Goldich and Peterman, 1980). With time, Lawson's original terminology and interpretation of stratigraphic order was largely supported by the data but with the added implication that all of the constituent rock-forming events took place in less than 100 million years with only local evidence for younger post-metamorphic retrogression.

At the time of the studies by Goldich and his colleagues an important fact remained: apart from Grout's work in Minnesota, no geologist other than A.C. Lawson had mapped systematically in the Rainy Lake area. This had been undertaken by him at a scale of one inch to four miles in 1885-87 and, with the assistance of H.C. Cooke and R.C. Wallace, at one inch to one mile in 1911. New mapping in greater detail was therefore ultimately undertaken by the Ontario Division of Mines beginning in the 1970's (Davies, 1973; Blackburn, 1973; Harris, 1974; Wood et al., 1980 a, b; Fumerton, 1985). The outcrop mapping of Fred Harris is perhaps the most notable because it provided an advanced level of lithostratigraphic detail, at a scale of one inch to ½ mile, while covering the historically controversial Rice Bay and Bear's Passage areas. He also provided new local evidence for stratigraphic younging in the Keewatin strata including the first recorded use of the shapes to pillows in basaltic flows in this area. Harris (1974) avoided the use of the historical stratigraphic terms but his table of formations tends to support Lawson's original interpretation of metasedimentary biotite schists at the base of the sequence. John Wood provided a comparable level of mapping at Mine Centre (Wood et al., 1980 a, b) with a focus on the Seine and Coutchiching metasedimentary rocks (Wood, 1980). Companion studies of the geology in the Minnesota portion of the Rainy Lake area were conducted under the auspices of the Minnesota Geological Survey and the US Geological Survey (Ojakangas, 1972; Southwick, 1972; Southwick and Ojakangas, 1979; Southwick and Sims, 1980).

Poulsen (1980) made extensive use of the report and maps of Harris (1974) as a foundation for structural and metamorphic studies in the Rice Bay and Bear's Passage areas. It eventually became clear, however, that one of the flaws in Lawson's original interpretation was that it relied on the assumption that structural superposition of the Keewatin above the Coutchiching equates to stratigraphic superposition as well in rock packages that are arguably overturned (Poulsen et al., 1981). This led to a revised interpretation of stratigraphic order (Fig. 6f) but one without geochronological constraint.

The full essence of the Seine-Coutchiching problem was ultimately clarified by application progressively improved methods of U-Pb analysis of zircons (Davis, 2023). Strategic sampling of each of Lawson's five lithostratigraphic units across several sites where field relationships had been established (Davis et al., 1989; Davis et al., 1990; Fralick and Davis,1999) provided the results that form the basis of the current chronostratigraphic chart for the area (Fig. 6f). Keewatin metavolcanic rocks and Laurentian metaplutonic rocks were shown to be of similar age (circa 2727 Ma) whereas detrital zircons from the Coutchiching suggested a younger age (circa 2700 Ma) and a granitoid clast and detrital zircons from the Seine conglomerate even younger (<2693 Ma). The age of crystallization of the Algoman intrusions was estimated to be in the range of 2693 to 2684 Ma. In total, the geochronological constraints have shown that most of the historical interpretations, including those of Lawson, had both merits as well as flaws whereas the lateral facies concepts which were so widely and uncritically accepted in the 1970's have proven to be entirely untenable. The net result, however, is that

Lawson's placement of the Coutchiching beneath the Keewatin on the grounds of the dip of strata alone was the main source of geological error. It was not that he did not understand the difference because, in Memoir 40, he appears to have been the first geologist to have compare directions of dip to the directions of younging (way-up or bedding top) in cross-bedded arenite of the Seine sedimentary unit. Lawson (1913) did acknowledge that he had learned about the utility of truncated cross-bedding from his field assistant J.D. Trueman, then a graduate student who in turn had been taught this by W.O. Hotchkiss at the University of Wisconsin (Dott, 2001). Hotchkiss was also familiar with upward-fining grain size variation in sandstone mudstone sequences but this was not yet in common use and therefore the significance of graded bedding, as preserved at Bear Passage area, was not yet appreciated by Lawson in 1913. It took the work of F.F. Grout (1925) to demonstrate that the graded beds at Morton Island indicate that the Coutchiching beds there stratigraphically overlie the Keewatin. Lawson also certainly would have been aware of the stratigraphic use of pillowed volcanic flows, as advocated by Morley E. Wilson (1913) in Memoir 39 of the Geological Survey of Canada, but was of the opinion that this method was unsound because he believed that pillows, then referred to as ellipsoidal structures, were of intrusive origin (Lawson, 1912).

As one looks back, the Timiskaming-Keewatin problem evolved along similar paths as another great geological debate that played out during much the same time frame, the Highlands controversy of Scotland. That problem also involved many observers who were focussed on small, geographically separated parts of a bigger problem and it has been said that, in many cases, they did not know what they did not know (Oldroyd, 1990). A case in point is the famous anecdote concerning T.L. Tanton of the Geological Survey of Canada and E.B. Bailey of the British Geological Survey (Dott, 2001). Tanton, who had graduated from the University of Wisconsin in 1915 under the supervision of C.K. Leith, led a group of Princeton geologists on a tour of Rainy Lake as part of their geological trip across Canada by rail in 1927. The group included two eminent guests from overseas, L.W. Collett from Switzerland and E.B. Bailey of Britain (Bailey, 1927). Tanton demonstrated the utility of cross-bedding and graded bedding to determine way-up in metasedimentary rocks, using Lawson's examples from the Seine Group at Shoal Lake and Grout's Coutchiching outcrops at Morton Island respectively. This resulted in Tanton's inclusion as a participant on a reciprocal visit to Scotland where he convinced Bailey that the Dalradian strata at Ballachulish are overturned (Tanton, 1930; Bailley, 1930, Dott, 2001). Another point of communality between the two controversies is that the reputations of the observers, especially Sir R.I. Murchison in the Highlands and A.C. Lawson at Rainy Lake, tended to get in the way of the geological facts. This should not overshadow the reality, however, that in his first, youthful burst of mapping from 1882 to 1889 and in his mid-career re-study from 1911 to1913, Lawson identified the five lithological building blocks which are representative of the architecture of virtually every Archean greenstone belt in the world. As Oldroyd (1990) has pointed out for the Scottish Highlands controversy, it is not only about who was right and who was wrong but it is also about the process of narrowing in on a consensus view based on the facts at hand. Similar sentiments were expressed by Lawson himself in the introduction to his 1913 report which fuelled the Seine-Coutchiching problem in the first place. The overall lesson of the problem seems to be that: "*Science is never 'settled' but evolves by the accumulation of facts, new ideas and vigorous open discussion and debate. Consensus is irrelevant in science; only truth matters."* (Dewey and Ryan, 2022, p.1834).

Rainy Lake Wrench Zone

Poulsen (1986b) introduced the term "Rainy Lake wrench zone" to distinguish rocks between the E-W Quetico Fault and the ENE Rainy Lake – Seine River Fault from the Quetico metasedimentary belt to the south and the main mass of the Wabigoon granite-greenstone belt to the north (Fig. 8). The rationale for highlighting the wrench zone involved many different geological aspects (Poulsen, 1986b) but the most prominent are the distinctive lenticular. s-shaped lithostratigraphic domains which merge with the discordant boundary faults. Broadly similar patterns are also evident in the steep metamorphic foliation which affects the Seine as well all of the older lithostratigraphic units. This is also the area in which the Seine – Coutchiching problem mainly played out and where generations of geologists contributed to the understanding of diverse aspects of its geology. It is also the focus of this field guide which can be used to illustrate the major lithostratigraphic

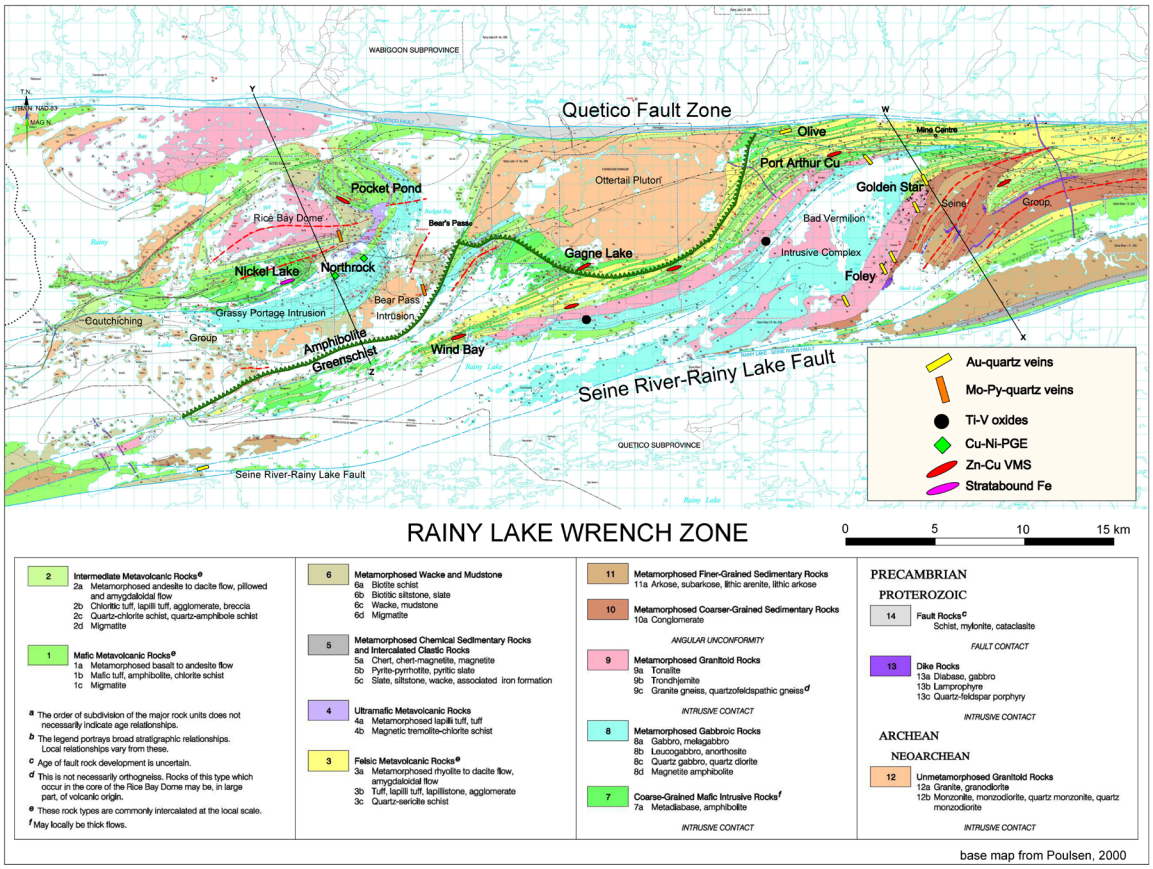


Figure 8: Simplified geological map of the Rainy Lake Wrench Zone.

units that comprise the wrench zone as well as the related topics of deformation, metamorphism and metallogeny.

Keewatin

Lawson applied the term "Keewatin Series" to all of the Archean metavolcanic rocks in the Rainy Lake area mainly to distinguish them from foliated quartzo-feldspathic rocks of probable plutonic origin. He initially did this in a descriptive way (Lawson, 1885) but his Rainy Lake reports (Lawson, 1887; Lawson, 1913) also provided petrographic detail and genetic interpretation. The Keewatin rocks within the wrench zone include lithofacies which are common to Archean greenstone belts in general. Mafic volcanic rocks predominate in the northwestern part of the zone, particularly at Windy Point, Nickel Lake and Pocket Pond. The rocks at these locations were metamorphosed to amphibolite facies assemblages so that primary features are difficult to document in the resulting foliated mafic tectonites. In places where strain is moderate it is relatively easy to identify pillows and varioles but the level of distortion in many places (Fig. 10a) makes it difficult to confidently use the shapes of pillow to confidently define directions of younging (Borradaile and Poulsen, 1981). A notable exception is at Pocket Pond (Fig. 9a) where adequate evidence of stratigraphic polarity is preserved (Fig. 10b). Felsic metavolcanic rocks predominate in the southeastern part of the wrench zone where they are commonly intercalated with rocks of andesitic composition (Fig.10c), The rhyolitic rocks are commonly quartz-phyric and included both coherent (Fig. 10d) and volcaniclastic (Fig. 10e) facies.

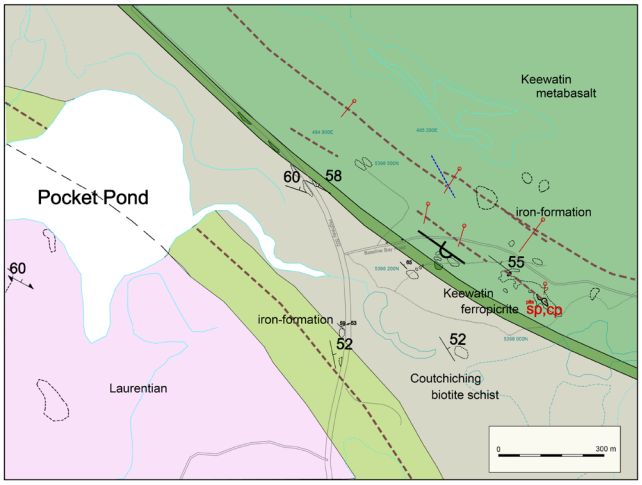


Figure 9: Pocket Pond locality

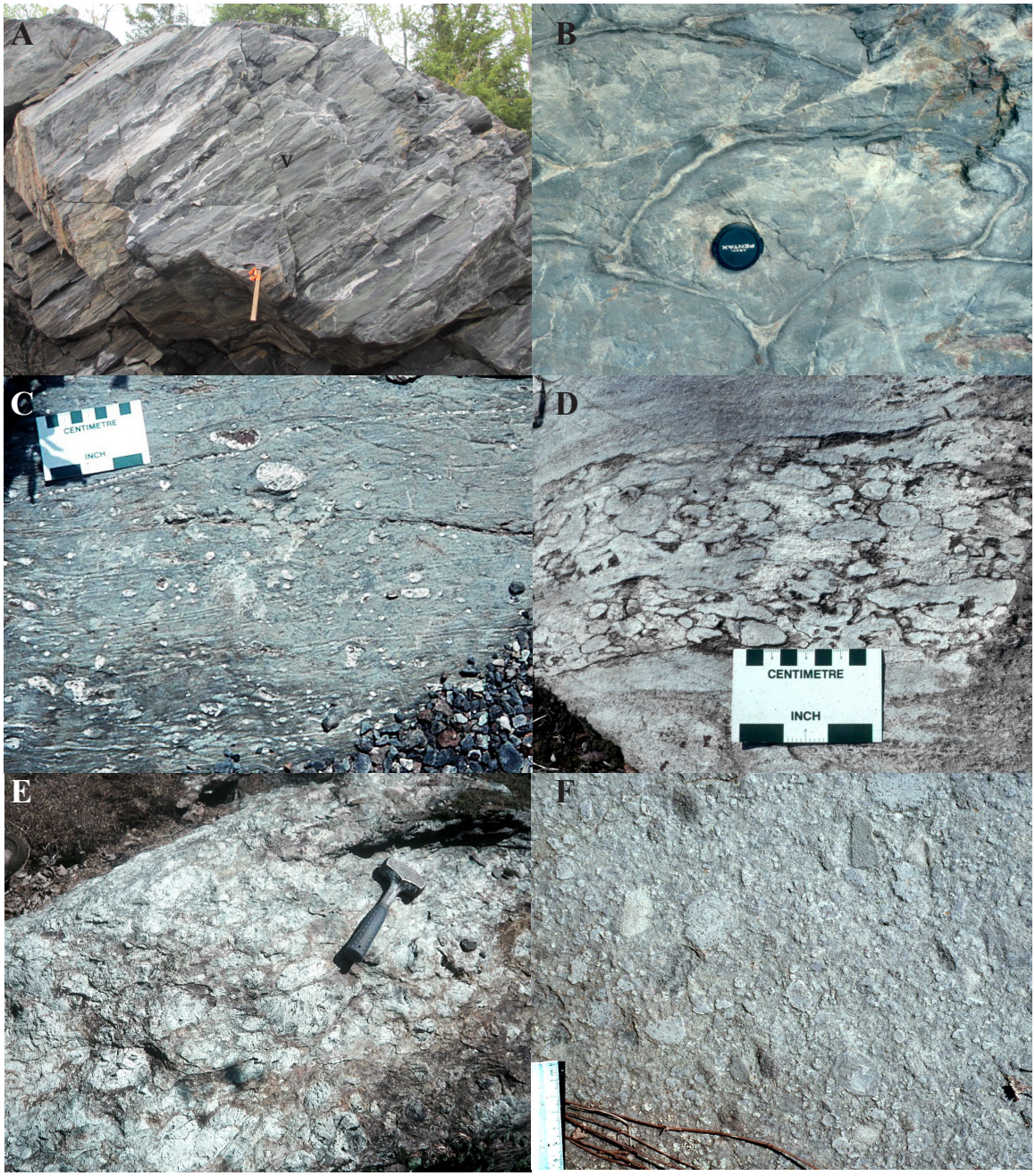


Figure 10: a) westward plunging shape lineation defined by deformed pillows in metabasalt, Nickel Lake area; a) Pillow basalt, Pocket Pond; c) Amygdaloidal basalt, Port Arthur Copper; d) spherulitic rhyolite, Ottertail east; e) Rhyolite Breccia, Wind Bay; f) volcaniclastic ferropicrite, Belacoma area.

One outstanding unit within the Keewatin sequence is composed of a distinctive ultramafic volcaniclastic rock (Fig. 10f) which is exposed in the Grassy Portage Bay area (Fig. 4). The unit was first recognized by Harris (1974) who classified it as an intermediate volcanic rock, mainly because of its common bright green, chloritic appearance along with volcaniclastic textures. Poulsen (1980) prosaically termed it "magnetic green rock" which is composed mainly of Mg-chlorite plus actinolite and magnetite. He provided lithogeochemical analyses to show that the rock has an ultramafic bulk composition but incorrectly classified it as a komatiite, a rock type with which it shares only some chemical similarities. He also compared the unit, both chemically and texturally to the better-known picritic Steep Rock Ashrock approximately 100 km to the east and suggested that their separation might be due to dextral displacement on the Quetico Fault (Fig. 1). Steve Schaefer conducted a study of the ultramafic units at both localities and confirmed their volcanic origins (Schaefer and Morton, 1991). He also provided the acronym GUP (Grassy Portage

Ultramafic Pyroclastic) for the rocks at the Rainy Lake locality. Goldstein and Francis (2008) pointed out the differences in the chemical composition of this unit compared to komatiites: the GUP shows higher FeO, TiO_2 and incompatible elements (e.g., Nb) as well as displaying fractionated rather than flat rare earth element patterns. Goldstein and Francis (2008) reclassified the rocks as pyroclastic ferropicrites, noting that they are examples of relatively rare Fe-rich volcanic varieties that were likely derived from partial melting a mantle source that was enriched in Ti and rare earth elements. A further characteristic of the GUP is that it contains microdiamonds which were discovered in 2008 by MetalCORP Limited at the Beaver Pond Occurrence (Hinz et al., 2010). All of these observations have significantly improved the understanding of this unusual volcanic unit but questions remain regarding its stratigraphic position and regional significance. Despite the remarkable similarity to the Dismal Ashrock at Steeprock, the notion of a strike-slip separation of the same unit is still feasible but not fully demonstrated. Tomlinson et al., (2003) reported a maximum age of 2780.4 +/-1.4 Ma for the Dismal Ashrock based on analyses of inherited zircons and argued that it is feasible for it and overlying basalts (Witch Bay formation) to be as young as other sequences in the Western Wabigoon Subprovince: by extension, this would include the mafic-ultramafic volcanic units in the Grassy Portage Bay area. If the ages of the ultramafic volcanic rocks at the two distant localities prove to be different, however, it would mean that an alternative explanation for their similarity would have to involve operation of similar processes at different times. In that case the communality might be sought in the mantle composition and depth that led to the formation and deposition of these unusual rocks.

Interflow metasedimentary rocks comprise a common but volumetrically small component of the Keewatin sequence. Although in places these rocks could be mistaken as providing evidence for interdigitation with Coutchiching biotite schists or with volcaniclastic rocks of intermediate composition, in most cases, they are arguably metalliferous, synvolcanic sedimentary units which range from pyritic mudstone and minor sandstone, to chert-magnetite banded iron-formation (Fig. 11a) and pyritic massive sulfide deposits (also termed sulfide facies iron-formation). The sulfide-bearing varieties were targets for possible sulfur production in the period around World War I when, particularly at Nickel Lake, they were noted to contain anomalous concentrations of Co-Ni-Zn-Cu. In places zinc is also a locally anomalous component and, at Pocket Pond, the small sphalerite lenses discovered in the 1970's are associated with iron-formation intercalated with metabasalt (Fig. 9). From a strictly geological perspective the presence of laterally extensive interflow units proves valuable for establishing a sense of stratification within the Keewatin because they not only can be mapped discontinuously in outcrop and drill core but cane be easily traced accurately by magnetic and electromagnetic surveys.

A.C. Lawson's 1914 geological map of Rainy Lake also includes two mafic plutonic rock types which he regarded to be part of the Keewatin sequence: extensive units of what he termed hornblende gabbro in the Grassy Portage Bay area (Fig. 4) and anorthosite in the Bad Vermilion Lake area (Fig. 5). Subsequent mapping has shown that he underestimated the total volumes of mafic plutonic rock in both cases and this was with good reason. It is now well-understood that thick, mafic submarine lava flows are capable of slow cooling rates to produce what can easily be accepted as a "gabbro-textured" facies that grades vertically and laterally over short distances into finer grained basaltic rocks, making their visual distinction from plutonic equivalents difficult. Furthermore, where amphibolite facies metamorphism has affected mafic volcanic rocks, recrystallization tends to coarsen the texture and obscure primary features: this is certainly the case in the northwestern western part of the Rainy Lake Wrench Zone. Finally, considerable local variations in textural detail are common in layered mafic intrusions (Fig. 11b, c, d) that that are difficult to map at a reconnaissance scale. What Lawson did map, however, were two of most extensive, distinctive and homogeneous plutonic phases, leucogabbro in the Grassy Portage Intrusion (Fig. 11e) and coarse anorthosite in the Bad Vermilion intrusion (Fig. 11 f). Detailed mapping by Hodgson (1959) at Grassy Portage Bay and by Harris (1974) at both localities provided much better definition of the full extents of these intrusions. Ashwal et al. (1983) undertook a more advanced petrological and geochemical study of the Bad Vermilion anorthosite and concluded that it represents the remnants of a subvolcanic magma chamber from which aliquots of magma had been extracted as extrusive lava flows. Poulsen and Hodgson (1985) reviewed the disposition of the different phases of both intrusions and the sulfide

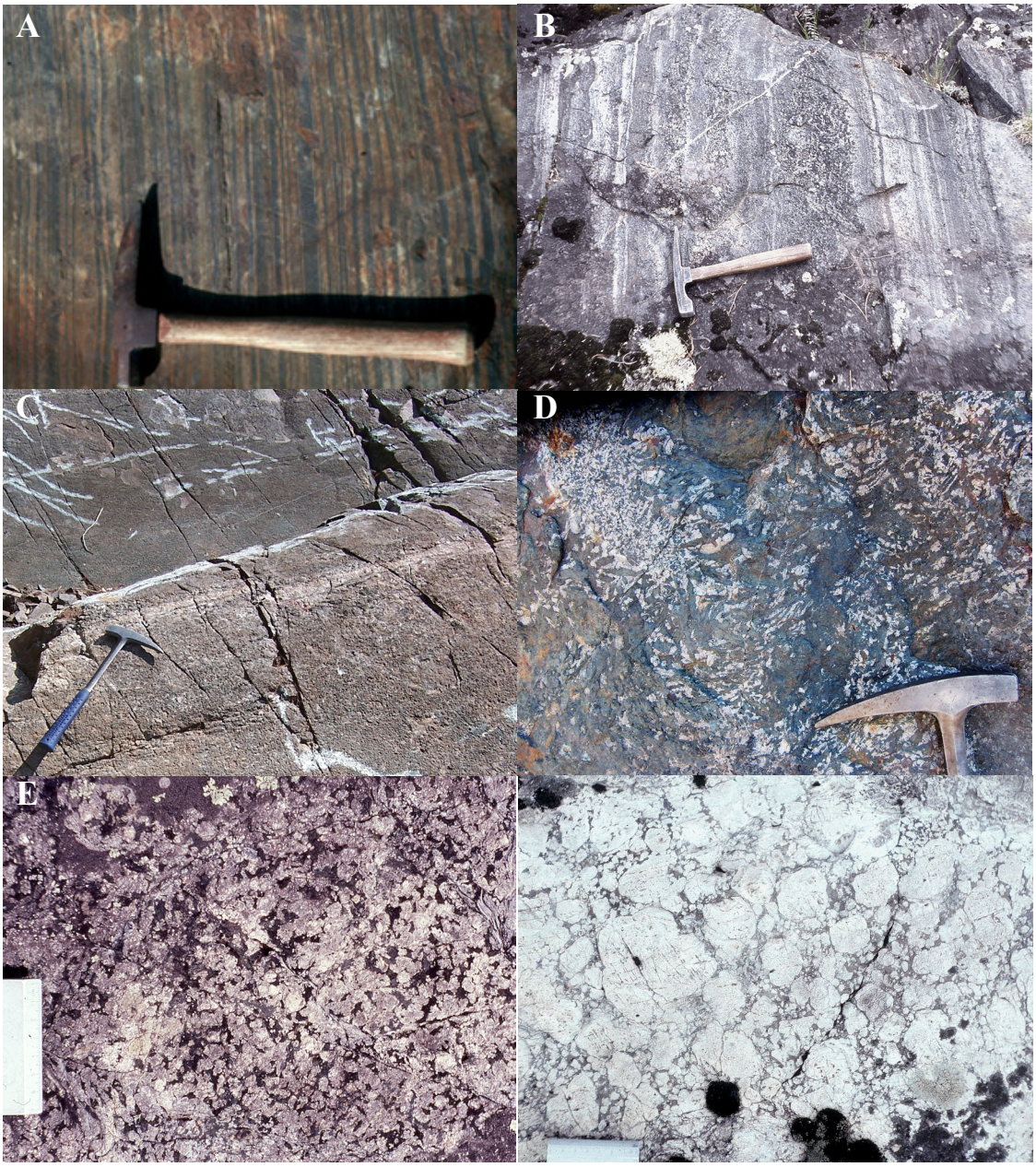


Figure 11: a) chert-magnetite iron-formation, Pocket Pond; b) thin-layered gabbro, Grassy Portage intrusion, west side of Redgut Bay; c) thick-layered gabbro-melagabbro, Northrock East trenches, Grassy Portage intrusion; d), Pegmatitic Gabbro, Northrock E trenches, Grassy Portage intrusion; e) leucogabbro, Grassy Portage intrusion; f) coarse anorthosite, Bad Vermilion intrusion, Scott Islands (the edge of the compass in the bottom left measures 10 cm)

and oxide mineralization within them, providing support to the idea that they are examples of syn-volcanic layered intrusions resulting from cumulus growth and magma fractionation. Both intrusions also received attention for their economic potential during exploration programs for Cu-Ni-PGE sulfide and Ti-V oxide mineralization (Hinz et al., 2010). The Bad Vermilion Intrusive complex and the surrounding metavolcanic rocks have recently been described as an arc-related "ophiolite" sequence (Wu et al., 2016) but this is highly unlikely given the dominance of rhyolite in the volcanic section and the absence of both peridotite and sheeted dikes.

The ages of the Keewatin units were largely unknown until the mid-1970's despite many attempts to apply modern geochronological methods (Goldich, 1968; Tilton and Grunenfelder, 1968; Hart and Davis, 1969; Peterman et al., 1972). At that point improvement in analytical precision and accuracy allowed U-Pb geochronology on carefully constrained samples to impact stratigraphic interpretations (Davis, 2023). Davis et al. (1988) applied these methods in the Rainy Lake Wrench zone to show that the units which were historically classified as Keewatin formed around 2727

Ma in a remarkably short interval of five to six million years (Fig. 5f). This included direct analysis of rhyolitic rocks from both the northwestern and southeastern part of the zone as well as the indirect constraint of mafic (Grassy Portage Gabbro) and felsic (Mud Lake trondhjemite) that cut the volcanic rocks. More recent attempts to provide additional ages of volcanic rocks near the Bad Vermilion Intrusion have proven to be unsuccessful in light of the lower analytical precision and accuracy of the methods employed (Wu et al. 2016).

Coutchiching

Lawson's 1914 map of Rainy Lake outlines three areas of Coutchiching rocks labeled as "mica schist, paragneiss and phyllite". The most extensive area occurs south of the Seine River Fault in Quetico Subprovince where the term Quetico metasedimentary rocks also applies (Fig. 12a). The other two major localities are located within the Rainy Lake Wrench Zone: a southern belt extending from Fort Frances-International Falls northeastward along Swell Bay and a northern one as a partially annular zone within the Rice Bay Dome (Figs. 2, 3, 4). Grout (1925)

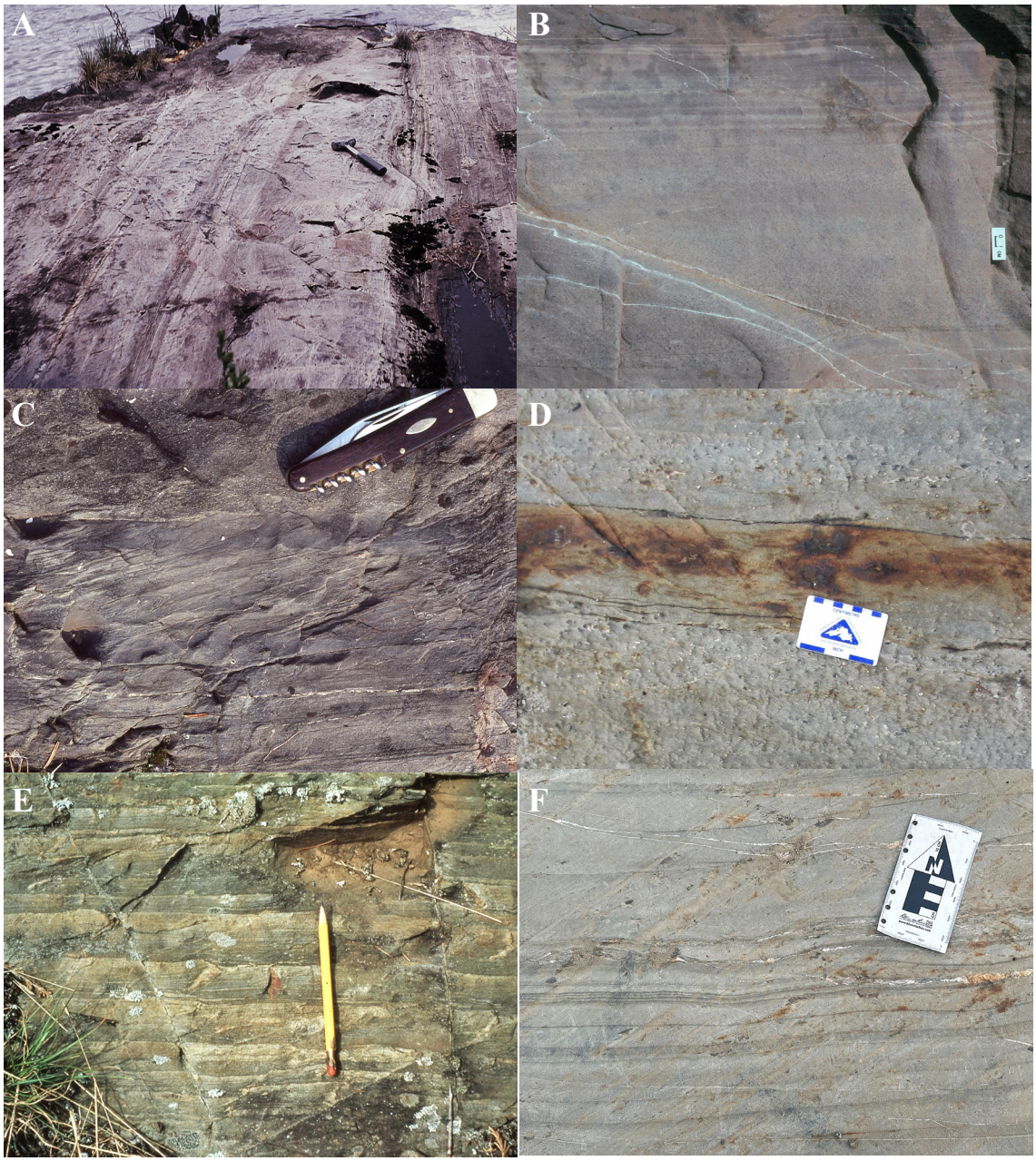


Figure 12: a) Quetico metasedimentary rocks, Bleak Bay area; b) thick-bedded wacke, Sandpoint Island c) graded beds cut by ENE cleavage, Morton Island (N to top of photo); d) knotty biotite schist containing retrograded staurolite and garnet, Great River Road; e) graded beds, Bear's Passage boat launch; f) graded beds in greenschist facies turbidites, Old Station Road (diagonal lines are glacial striae).

who was the first to recognize graded bedding in the metasedimentary rocks in the Swell Bay belt (Fig. 12b, c) and to apply it to establish local stratigraphic polarity at Morton Island (Fig. 13). This was confirmed by Merritt (1934) who also notably interpreted the colour banding of graded rocks in the Swell Bay as "varves". R.W. Ojakangas who often lamented that the Canadian glaciers had been cruel to Minnesota re-examined the sparse exposures of the Coutchiching rocks at Ranier, Minnesota near Lawson's type locality (Fig. 3). He described the rocks there as metagreywacke noting that they originally consisted of alternating beds of greywacke sandstone (or simply wacke) and mudstone deposited, not by glacial processes but by turbidity currents on submarine fans (Ojakangas, 1972; 1982). Most authors who have studied the belt of Coutchiching rocks along Swell Bay have also noted that they been clearly intruded by younger granitoid rocks (Algoman) and that the rocks on the northern shore of Swell Bay display amphibolite facies, pelitic metamorphic assemblages (Fig. 12d) involving biotite, muscovite, garnet, cordierite and staurolite (Ojakangas, 1982; Poulsen, 1980). The higher metamorphic grade has also been implicated by many authors for obscuring primary features such as graded bedding in the metasedimentary rocks.

The most contentious interpretations of the stratigraphic significance the Coutchiching rocks in the Swell Bay corridor result from observations in the Bear's Passage area (Figs. 4, 14). The Keewatin at this locality consists of a northwestward-dipping section composed of the upper part of the southeastward-younging Grassy Portage layered intrusion overlain by a thin unit of what are arguably mafic metavolcanic rocks. The staurolite-bearing metasedimentary rocks, although locally folded, also dip to the northwest and are cut discordantly by granodiorite of the Bear's Passage intrusion (Fig. 4). Although minor reversals in polarity of grading suggest local folding within the Coutchiching rocks in the Bear Passage area, a good quality exposure at their contact with Keewatin strata (Fig. 12e) demonstrates that the metasedimentary rocks are overturned (Poulsen, 1980). This plus the observations at Pocket Pond (Fig. 9) and Morton Island (Fig. 13), provides the geological evidence in favour of the Coutchiching being younger than the Keewatin. The most conclusive evidence, however, ultimately came from U-Pb analyses of detrital zircon in biotite schist near Tunnel Bay and in well preserved greenschist grade metagreywacke (Fig. 12f) northeast of Shelter Cove (Fig. 6) which represents the northeastward extension of the exposures at Morton Island (Davis et al. 1989). The age of the Coutchiching is constrained by the youngest detrital zircon grains at approximately 2704 Ma and by the circa 2692 Ma age of across-cutting felsic dike (Fig. 5 f). The presence of much older detrital grains (2930, 2940 and 3060 Ma) also suggested a potential contribution of detritus to the Coutchiching from a source area comparable to the Marmion domain north of the Quetico Fault extending in the Steeprock Lake area (Fig. 1). Similar conclusions were reached for the Quetico metasedimentary rocks by Davis et al. (1990).

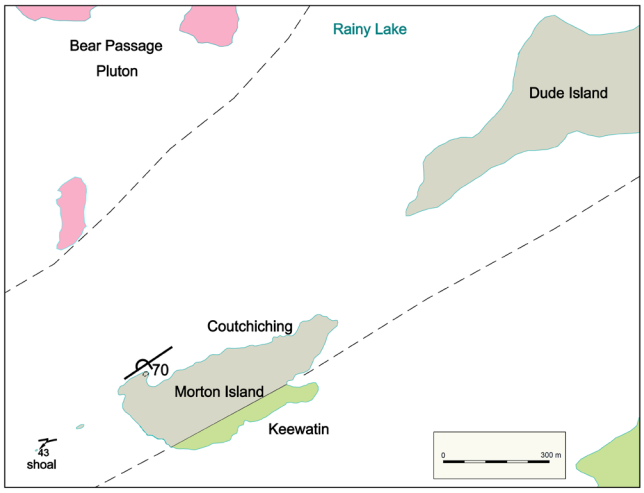


Figure 13: Morton Island locality (adapted from Poulsen and Wood, 1982)

Laurentian

Lawson (1887) used the term Laurentian for variably foliated granitoid rocks in general but by 1913 he only applied it at only three localities, Bad Vermilion lake,

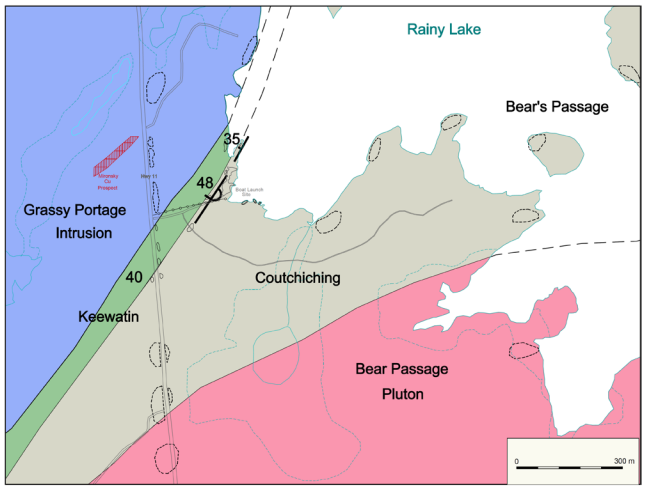


Figure 14: Bear's Passage locality

Rice Bay and Grassy Island near Neil Point (Fig. 2). The Laurentian granitoid rocks at Bad Vermilion Lake occupy three sinuous bodies that are broadly co-spatial with the Bad Vermillion layered intrusion (Fig. 7). The sodic granitoid rocks range from tonalite (Fig. 12a, b) to trondhjemite (Fig. 12c) in composition (Goldich and Peterman, 1980) and likely occupy the remnants of sills that are broadly concordant with stratigraphic layering in the surrounding northward-younging Keewatin volcanic rocks (Fig. 7). It is noteworthy that Lawson (1887) was the first to suggest that they might be syn-volcanic, subvolcanic intrusions. This point was verified by Davis et al. (1989) who established nearly identical ages of 2728 Ma for the intrusive Mud Lake trondhjemite near the Stellar gold deposit and an overlying rhyolite west of the Port Arthur copper deposit. The Laurentian rocks at Grassy Island likely represent an isolated remnant of the same stratigraphic section to the southwest (Fig. 2). A noteworthy characteristic of some outcrops of tonalite, particularly near gold-bearing quartz veins, is a quartz-rich sericitic rock (Fig. 12b) that was termed "protogene" by the early gold explorers in the region and results from plagioclase-destructive metamomatism related to carbonatization associated with brittle-ductile shear zones in the tonalite (Diamond and Marshall, 1990).

The Laurentian rocks exposed in the core of the Rice Bay Dome (Fig. 4) are much more difficult to interpret, in part due to overprinting deformation and amphibolite facies metamorphism. Lawson's 1914 map classified them to include as an inner body of granite and granite gneiss with an intrusive relationship with an outer annulus of Coutchiching biotite schist. Subsequent petrographic studies documented the distinctions among the lithologies (Frye, 1959; Peterman, 1959) and the term "paragneiss" was ultimately given to the innermost rocks. The existence of a large pluton was questioned and both the paragneiss and biotite schist were considered to be different components of the Coutchiching (Peterman, 1959; Peterman et al. 1972). By the same token, however, a small volume of deformed quartz-feldspar dikes and sills were shown to cut the paragneiss within the dome (Peterman et al. 1972). Harris (1974) took much the same approach and, apart from areas where the minor granitoid dikes and sills were particularly abundant, he mapped most of the interior of the Rice Bay Dome as being composed mainly of "biotite-feldspar-quartz schist" which he also assigned to the lower metasedimentary unit (i.e. the Coutchiching). Goldich and Peterman (1980) continued to view the rocks in the interior of the Rice Bay Dome as being composed of paragneiss derived from epiclastic sedimentary rocks but they also presented chemical data to show that they are different from the Coutchiching biotite schists and metagreywackes. Poulsen (1980) used the non-genetic term "grey gneiss" for rocks in the interior of the Rice Bay Dome (Fig. 15d, e) and also showed that they are fundamentally different in chemical composition from the annulus of biotite schist that envelopes them (Fig. 4). The minor deformed quartz-feldspar porphyry dikes (Fig. 15 e, f) are in, turn, different in chemical composition from both the grey gneiss and biotite schists (Poulsen. 1980; Goldich and Peterman, 1980).

Dick Ojakangas (personal communication, circa 1980) provided the novel suggestion that some of the grey gneisses actually may have been felsic volcanic rocks rather than felsic intrusions. This prompted Poulsen (1984) to opt for the uninspiring descriptive term quartzo-feldspathic gneiss to distinguish the Laurentian rocks from the Coutchiching biotite schists. Davis et al., (1989) reported a U-Pb zircon age of 2725+/-2 Ma from a sample of the quartzo-feldspathic gneiss near Moran's Bay (Fig. 4) to demonstrate its probable chronological equivalence with both the Keewatin rhyolite and the Laurentian Mud Lake trondhjemite in the Bad Vermilion Lake area. One of the notable lithogeochemical attributes of the biotite-rich Laurentian gneisses within the Rice Bay dome is their local deficiency in Na and Ca and their excess in Mg and Fe relative to their high silica and low Ti contents (Goldich and Peterman, 1980; Poulsen, 1980). One explanation for this is that they were locally subjected to plagioclase-destructive metasomatism which would also explain the presence of staurolite, andalusite and/or cordierite within them at specific sites. Such alteration in well-known in the environments of volcanic-associated massive sulfide deposits. Beakhouse (1984) evaluated this possibility in the western part of the Rice Bay dome where he identified the metamorphic assemblage quartz-chlorite-garnet-anthophyllite-staurolite with possible large relict grains of cordierite at one locality and common garnet over a larger area. Teck Corporation subsequently verified these mineralogical anomalies with further mapping and lithogeochemical surveys to conclude that the alteration is likely related to pyritic massive sulfide mineralization within an iron-

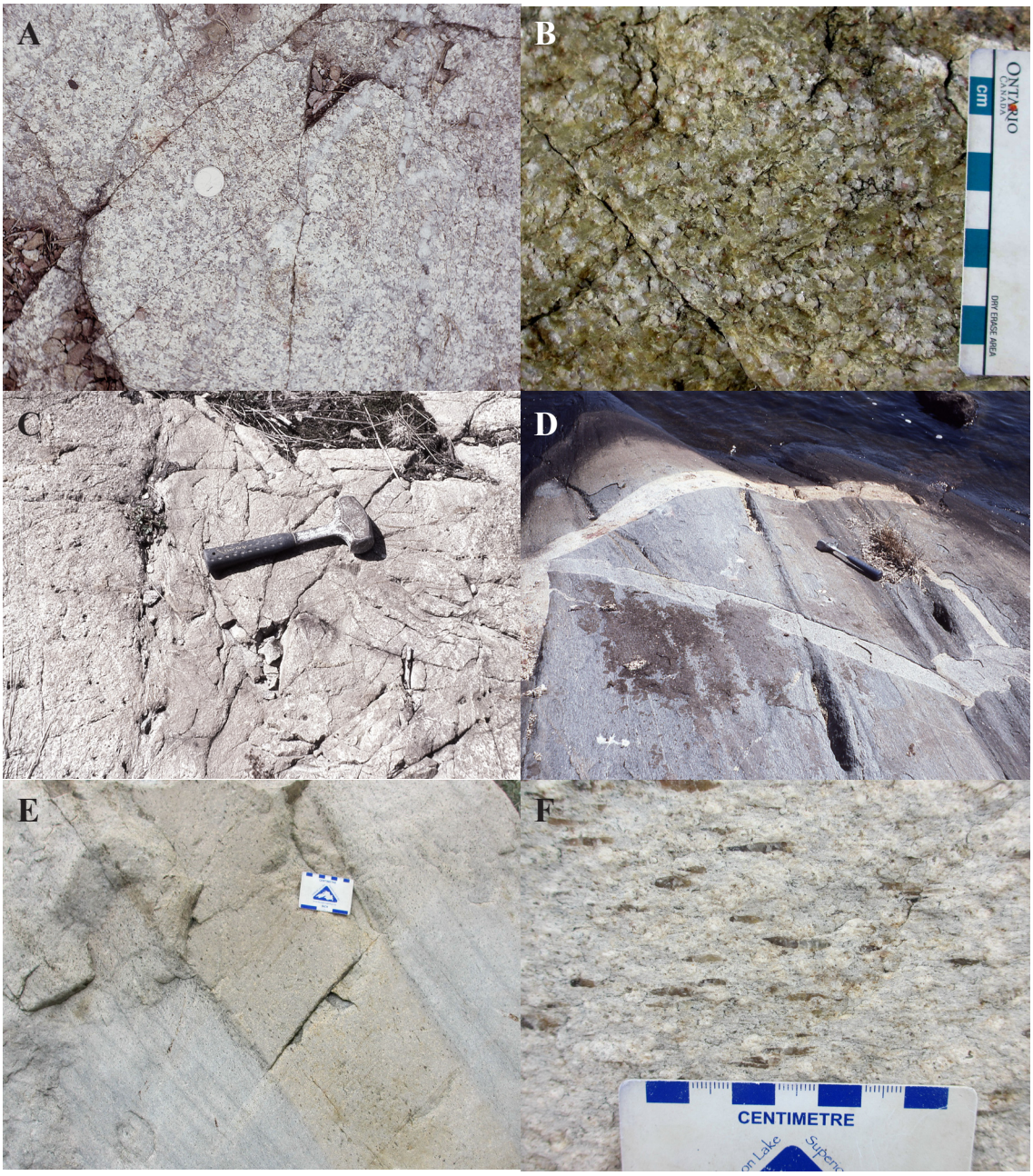


Figure 15: a) Bad Vermilion tonalite, Mine Centre; b) Sericitized tonalite ("protogene"), Mine Centre c) Mud Lake trondhjemite, Stellar gold property; d) grey quartzo-feldspathic gneiss cut by leucocratic dikes, Rice Bay, e) quartz-phyric grey gneiss cut by quartz-feldspar-phyric dike, Laurentian gneiss unit, Moran's Bay; f) deformed quartz (dark) and feldspar phenocrysts in qfp dike, Moran's Bay

formation unit near the outer part of the Rice Bay dome (Alderman, 1988).

In summary, despite incremental advances in establishing the geological facts concerning the Laurentian gneiss of the Rice Bay dome, considerable uncertainty remains about its origin. It has been established to be age equivalent and compositionally similar to both the Keewatin and Laurentian rocks in the Bad Vermilion Lake area but much study is required to establish its stratigraphic significance with respect to the Coutchiching and Keewatin rocks which structurally overlie them. The weights of evidence suggest, however, that the definitively intrusive aspects of the dome are attributable to the minor volume dikes and sills for no absolute ages have been established. On lithogeochemical grounds they may represent a phase on the younger Algoman intrusive suite (Goldich and Peterman, 1980).

Seine

Of all of Lawson's many achievements at Rainy Lake, it was arguably the recognition of the sedimentary rocks of the Seine Series, the interpretation of their probable depositional paleoenvironment and the demonstration of a high-angle unconformity beneath them that have best withstood the test of time. Although the overall level of exposure is uneven, the critical localities where this is best illustrated are located between Shoal Lake and Bad Vermilion Lake in the Mine Centre area (Fig. 7). In particular, exposures of the basal conglomerate (which Lawson termed "fanglomerate") near the Golden Star Mine (Fig. 16) and the overlying arenite facies exposed on islands in Shoal Lake to the south provided the diagnostic evidence for Lawson's arguments.

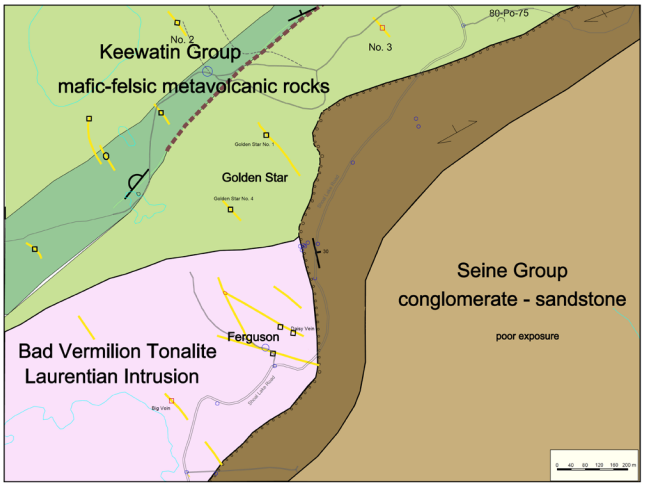


Figure 16: Simplified geology of the Golden Star locality.

Rocks of the Seine series occupy the area to the southeast of the trace of the unconformity and the underlying rocks of the Laurentian and Keewatin are located to the northwest. The S-shaped configuration of the unconformity trace is likely meaningful, not only because it mimics larger patterns in the wrench zone as a whole (Fig. 8) but also because the north-south segment reflects lower than average intensity of superimposed strain. Lawson was the first to note that this is in part responsible for the convincing preservation of contact relationships. The basal Seine conglomerate dips shallowly southeastward whereas Lawson showed that an interflow chert-carbonate unit within the Keewatin dips moderately northward. A relatively minor refinement (Pouslen and Wood, 1982) is that pillowed metabasalt overlies the chert-carbonate marker and indicates a northwestward-younging for the Keewatin rocks. In other words, there is evidence for back-to-back younging across a high-angle unconformity. Although weakly aligned due to overprinting strain, a critical point of observation is that clasts in the basal conglomerate show no evidence of a prior metamorphic foliation (Fig. 17a). The derivation of the coarse gritty matrix of the basal conglomerate from the underlying Laurentian tonalite is also clearly evident when compared to the intrusive rocks below the nonconformity. The shallow dipping basal conglomerate (Fig. 17b) to which Lawson ascribed an alluvial origin has been mapped along a persistent ridge of fair outcrop (Fig. 16) but topographically recessive arenite which overlies it to the east is poorly exposed.

The Seine arenite unit is well-exposed at Shoal Lake where cross-bedded sandstone (Fig. 17c) provides stratigraphic polarity as well as supporting the common interpretation of a fluviatile origin. Cross-bedded sandstone (Fig. 17d) also can be traced farther eastward along the Seine River (Fig. 2) where it can be demonstrated to be overlain by an upper unit of coarse, polymictic conglomerate (Fig. 17e) and, in some cases, intercalated with it (Fig. 17f). The uppermost conglomerate unit is notable for an abundance of granitoid clasts and Davis et al. (1989) reported an age of 2696.1+5/-3 demonstrating that it was sourced in a granitoid body that was much younger than the Laurentian which provided detritus for the basal Seine Conglomerate. Davis (1990) further constrained the depositional age of the sand-sized fraction from arenite at Horsecollar Junction (Fig. 2) by noting the presence of abundant detrital zircons with a U-Pb age of approximately 2693 Ma., effectively the same age as the Bear Pass pluton. This fact contradicted Lawson's original contention that all of the Algoman intrusions could be defined on the basis of the fact that they are younger than the Seine (see below). Nonetheless, Lawson's original interpretation of the Seine Series mainly as a product of Archean alluvial and fluvial sedimentary processes has been reinforced and elaborated upon by several authors (Ojakangas, 1972; Wood, 1980; Fralick and Davis, 1999; Czech and Fralick, 2002). Although his language was somewhat dense, the overall message of Lawson's paleoenvironmental interpretation is paraphrased as follows: *"it seems a fair inference that the conglomerate represents a gravelly flood plain… The distribution of the conglomerate … indicates the course of a river (following) the dominant structural lines … at a time which antedates the intense complication which folded*



Figure 17: a) basal Seine conglomerate inter-clast quartz grit derived from underlying tonalite, Golden Star arear; b) stratified basal conglomerate, Golden Star area; c) trough cross-bedded Seine arenite in plan view, Shoal Lake showing younging toward the top of the photo; d) deformed cross-bedded Seine arenite and pebble conglomerate in cross-section view, Seine River Bridge; e) deformed polymictic conglomerate in cross section view, east of Mine Centre; f) sandstone interbed in coarse upper Seine conglomerate in plan view west of Wild Potato Lake. Note the angle between bedding (arrows) and foliation.

and deformed the conglomerate" (Lawson, 1913, p.62). In other words, he envisioned the location of Seine conglomerate and arenite to have been controlled by syn-sedimentary faults to account in part for it's elongate map pattern (Fig. 2).

Algoman

Lawson's 1914 map portrays five different varieties of intrusive rocks at Rainy Lake in decreasing order of perceived age which he classified with the term Algoman: basic facies of syenite, syenite gneiss, granite and granite gneiss, banded and streaked gneiss and porphyroid gneiss. Harris (1974) made similar distinctions which allowed the least deformed Algoman rocks to be discussed in terms of three distinct spatial and compositional suites: the Rocky Islet Bay complex, the Swell Bay intrusions and the large and conspicuous Ottertail Lake Intrusion (Fig. 3). The first of these are dominated by quartz monzonite syenite and mafic syenite and are commonly feldspar-phyric (Fig. 18a,

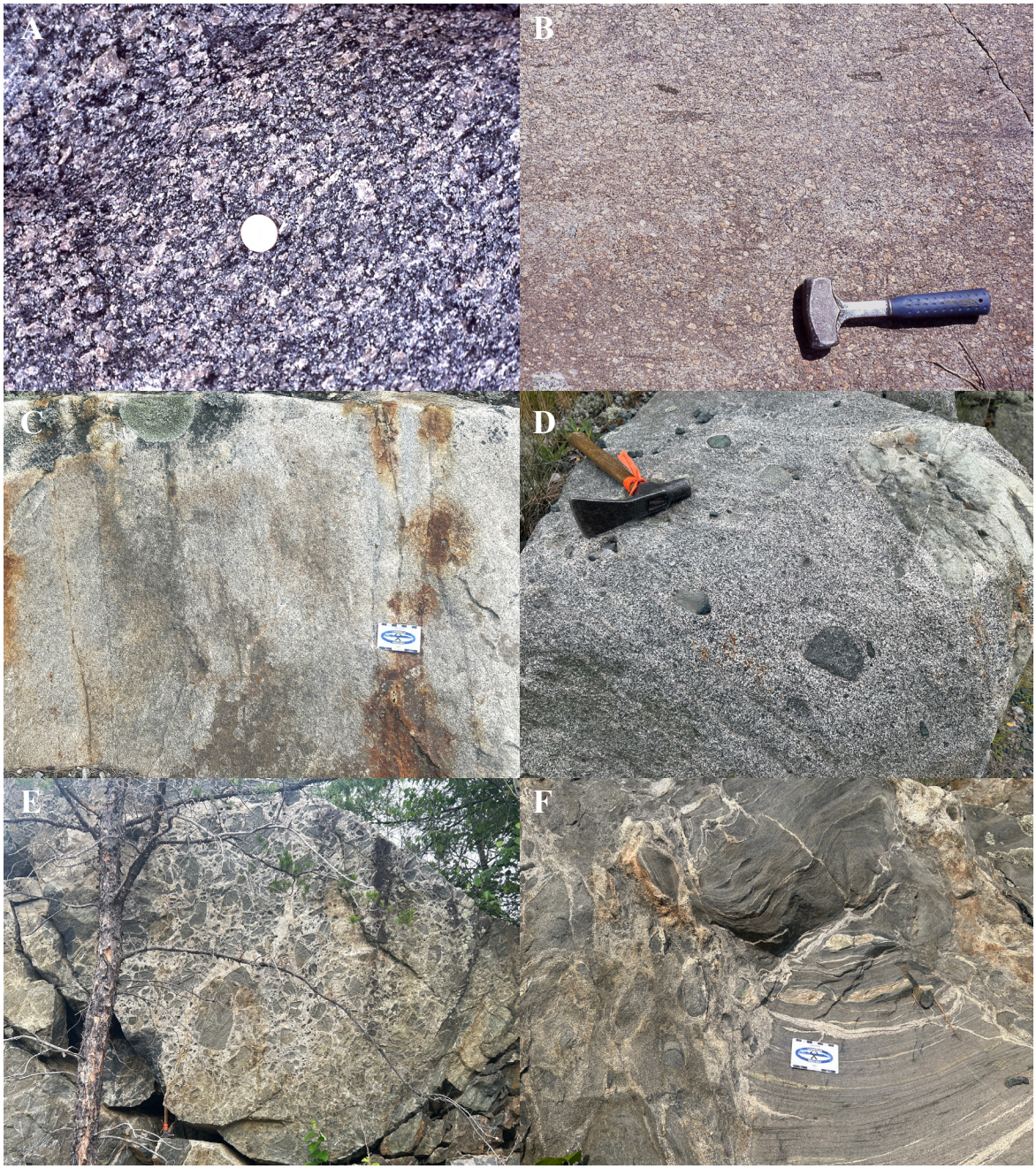


Figure 18: a) feldspar-phyric quartz monzonite, Raspberry Island; b) porphyritic quartz monzonite, Rocky Islet Bay; c) granodiorite cut by vertical sheeted quatz-pyrite veins, Bear's Passage; d) xenolithic monzodiorite, Ottertail Lake intrusion; e) Intrusion breccia with granitoid matrix, western Ottertail Lake intrusion; f) incipient brecciation and granitoid infilling of metamorphic tectonite, Ottertail Lake intrusion.

b), The Swell Bay intrusions, exemplified by the Bear Pass Pluton (Fig. 18c) are composed mainly of quartz monzonite and granodiorite (Goldich and Peterman; 1980). Some of these intrusions are compositionally zoned with mafic to intermediate margins and felsic interiors (Cram, 1923; Harris, 1974). The Ottertail Lake intrusion is also compositionally zoned from marginal hornblende-biotite quartz monzonite to interior leucocratic quartz monzonite in the interior (Goldich and Peterman, 1980): wallrock xenoliths are common in the marginal phase (Fig. 18d) and internal magmatic breccias (Fig. 18e, f) are well developed in what Lawson interpreted to be roof pendants of deformed and metamorphosed Keewatin rocks.

Goldich and Peterman (1980) demonstrated that the Algoman intrusive rocks commonly contain abundant K-feldspar and have much higher Sr contents than Laurentian tonalite and trondhjemite. The Ottertail Lake intrusion, also with high overall Sr content, displays a fractionation trend of increasing Rb:Sr

ratio toward its interior. Shirey and Hanson (1984, 1986) and Stern et al. (1989) further defined specific lithogeochemical characteristics of the Algoman rocks at Rainy Lake to show that they are also distinctive from other granitoid rocks at a global scale. Relative to their intermediate silica content (55-60%), they contain anomalous Mg, Sr, Ba, Ni, Cr and are strongly enriched in light rare earth elements. Stern et al. (1989) proposed their formation from hydrous melting of mantle that had been enriched large ion lithophile elements though prior metasomatism. Davis (1990) provided an estimate of 2693 +/- 2 Ma age for the Bear Pass pluton and coupled with the 2686+2/-1 Ma age of the Ottertail Lake intrusion (Davis et al. 1989), demonstrated that the sanukitoid magmatism spanned the time bracket for inferred for deposition of the Seine conglomerate and arenite above a profound angular unconformity. A recent study by Bjorkman et al. (2024) has demonstrated the widespread distribution of the sanukitoid suite of rocks across the Wabigoon Subprovince, including the Ottertail Lake Intrusion. This has been interpreted to represent a significant shift in magmatism at approximately 2690 Ma that can be explained by metasomatism and magmatism in a supra-subduction setting leading to collisional deformation and metamorphism that is commonly attributable to the Kenoran Orogeny.

Deformation and Metamorphism

The emphasis on protoliths and stratigraphic relationships that has historically dominated the discussion of the geology does not outweigh the fact that most of the rocks are clearly metamorphic tectonites as well. Lawson (1913) recognized this and attributed commonly observed foliation and lineation ("pencilling" in his terminology) to compressive deformation related temporally to the Algoman granitoid suite. Rocks in the southeastern part of the wrench zone have been metamorphosed to greenschist facies mineral assemblages and rocks of the amphibolite facies are dominant in the northwest (Fig. 8). Significant areas of retrograde metamorphism have also been noted (Peterman et al. 1972; Poulsen, 1984) and this has been taken to be the explanation why most geochronological approaches have yielded unreliable protolith ages. It is likely that the overall distribution of preserved prograde assemblages is the result a combination of both local contact and regional dynamothermal metamorphism. The common existence of minor structures of dynamothermal metamorphic origin such as foliation (Fig. 12c, 17f, minor folds (Fig. 19a, b) and lineation (Fig. 10a) are reflections of local strain. Rheological contrasts within and among lithological units have also been well established to be important in controlling the local strain intensity in the Rainy Lake area, particularly in the Seine conglomerate (Hsu, 1971; Jackson, 1982; Czeck et al., 2009). The highest strains are also common in features which are arguably shear zones in which strong foliation is accompanied by asymmetric distribution of foliation (Fig. 19c, d, e, f) that mimics the overall structural pattern in the wrench zone as a whole (Fig. 8). Following the lead of Peter Hudleston (1986) in the Vermilion district of Minnesota, dynamic interpretations invoking dextral transpression have been invoked by several authors to explain the overall structural style of the Rainy Lake wrench zone (Poulsen, 1986b; Borradaile et al., 1988; Poulsen et al., 1992; Czeck and Hudleston, 2003; Fernandez et al., 2013).

Beyond the local importance of dynamothermal metamorphic fabrics, however, the larger structural features in the wrench zone also of considerable interest. Foremost among these is the angular unconformity at the base of the Seine sedimentary sequence in the southeastern part of the zone (Fig. 20a) and it also provides an ideal temporal reference point for understanding the deformational and metamorphic history of the area. As illustrated above, the fact that lithic clasts in the basal conglomerate above the unconformity show no evidence of pre-depositional metamorphic fabrics yet clasts throughout the Seine have been variably strained during post-depositional dynamothermal metamorphism is an important one. It illustrates the insufficiency of using the development of foliation alone as a means of tracking a protracted structural history. A second notable structural aspect at Rainy Lake is the stratigraphic evidence for significant overturning of beds in the northwestern part of the zone (Fig. 20b). Poulsen (1980) suggested that this might have resulted from the overprinting of early recumbent folds by younger upright ones but, given the observation that the first-formed foliation in these rocks is also folded in the Rice Bay dome, the possibility of late-overturning of what may have been at one time steep strata can't be entirely ruled out. A third topic of importance is the fact that, since their recognition in the 1930's, there also has been a great deal of attention paid to the major faults that define the boundaries of

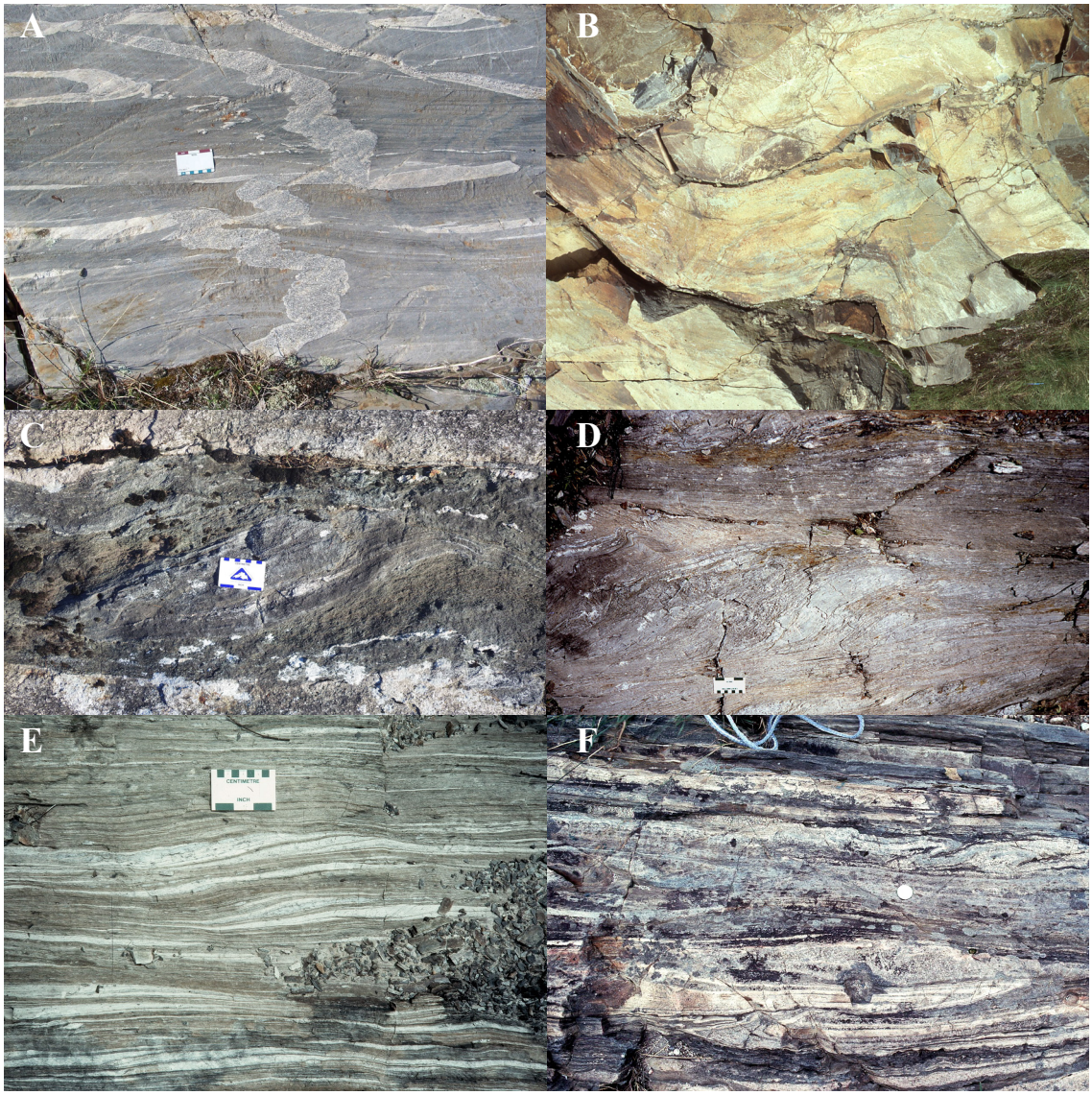


Figure 19: a) shortening and transposition of felsic dikes cutting Coutchiching biotite schist, north of Noden Causeway; b) folded felsic sills, Great River Road; c) asymmetric boudinage of the interior of a mafic dike relative to its foliated margins, Noden Causeway; d)) asymmetric boudinage in felsic metavolcanic rocks south of the Olive gold mine, e) asymmetric shapes of clasts in Seine meta-conglomerate adjacent to the Rainy Lake – Seine River Fault south of Seine River Bridge; f) tight asymmetric folds in mylonite, Little Turtle Lake landing.

the wrench zone. The rocks that now help to define the Quetico Fault at Rainy Lake were originally mapped by Lawson (1913) as part of a narrow belt of "porphyroid gneiss" extending westward from Little Turtle Lake at Mine Centre to Cheery Island. He recognized that the red porphyroid gneiss *"has a pronounced cataclastic structure and that the schistosity of the rock is referable to deformation involving shearing of the mass" (Lawson, 1913, p.94*). He stopped short of relating the rocks to a fault, however, interpreting them instead to represent the deformed southern margin of a large granitoid batholith: this is somewhat ironic because he is the geologist who, by this time, had named the San Andreas Fault and had compiled the definitive technical report on the Great San Franciso Earthquake of 1906. By the time F.R. Harris remapped the area, however, it had been recognized that the rocks here belong to the greater than 350 km long Quetico Fault based on the interpretation linears on air photo mosaics (Parkinson, 1962). Harris (1974) went on to describe the rocks in the fault as crushed granite, augen gneiss and mylonite and, like Lawson before him, locally showed gradational contacts with adjacent banded gneissic rocks which he termed migmatite. Kennedy (1984) studied 14 sites along the entire Quetico Fault, including 3 in the Rainy Lake wrench zone, and concluded that the mylonitic foliation on average resulted, not strictly from cataclastic processes, but from ductile flattening

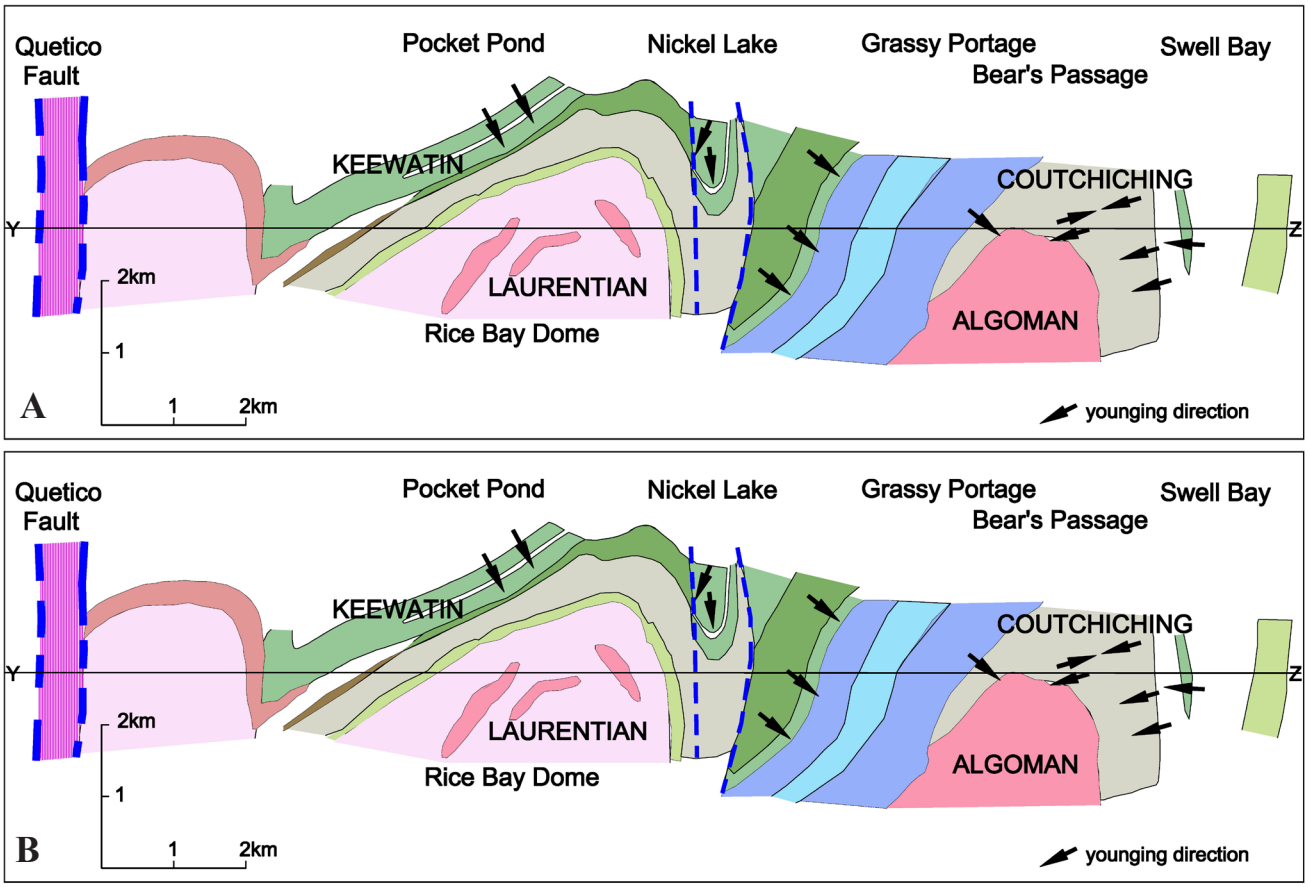


Figure 20: Schematic cross-sections through a) Rice Bay – Bear's Passage and b) the Bad Vermilion – Shoal Lake areas. See Figure 8 for the locations of the sections (adapted from Davis et al., 1989). The Quetico fault is located at the northern end of both sections.

based on measured axial ratios of deformed mineral aggregates and object-object strain estimates. She also used quartz c-axis fabric measurements and analysis of brittle micro-faults and ductile shear zones to argue for overall dextral displacement on the fault. Kennedy (1984) showed that the microfaults and minor shear zones dominantly strike NW and have dextral shear sense. She further argued that transition from ductile behaviour (mylonite) to brittle is consistent with the current level of exposure representing deformation at a crustal depth of 10-15 km. Borrradaile and Kennedy (1982) also showed evidence of flow-banding in veins of pseudotachylite at Crowrock Inlet as evidence of frictional melting in the fault zone. Peterman and Day (1989) reported a Rb-Sr isochron age of 1947+/-23 Ma to suggest that the pseudotachylite from both the Quetico and Seine River faults resulted from Proterozoic reactivation of the Archean faults.

Metallogeny

A commonly understated geological feature of the Rainy Lake wrench zone is the simple abundance of mineral occurrences within it in comparison to the adjacent areas on either side. Poulsen (2000b) enumerated 88 of them in total and demonstrated that they include examples that are representative of multiple deposit types (Figs. 8, 21) which, in turn, are thought to relate to multiple geological processes. Syngenetic deposits include stratabound metalliferous sediments in the mafic sections of the Keewatin including banded iron formation, pyritic massive sulfide deposits with locally anomalous zinc sulfides (Nickel Lake and Pocket Pond). Numeous Zn-Cu occurrences (Port Arthur Copper, Lochart Lake, Wind Bay, Gagne Lake, Pidgeon) demonstrably possess the descriptive of volcanic-associated massive sulfide deposits in general. Basal Cu+/-Ni sulfide mineralization (North Rock) and magnetite+/ilmenite mineralization (Seine Bay, Mironsky) is clearly associated with the Grassy Portage and Bad Vermilion Lake layered gabbroic intrusions (Poulsen and Hodgson, 1984). Quartz-pyrite-molybdenite veins show a spatial association with Algoman granitoid rocks and sheeted veins of this type within the Bear Pass Pluton are similar in style to those in the deeper parts of granitoid-related

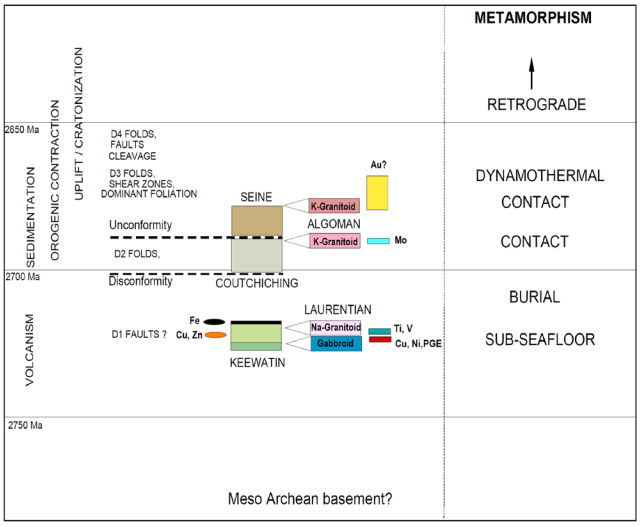


Figure 21: Metallogenic Summary of the Rainy Lake Wrench Zone

Phanerozoic vein and stockwork deposits. The gold-bearing quartz veins in the Mine Centre area which were the focus of a gold rush in the 1890's (Coleman, 1894; Winchell and Grant, 1895), are readily classified in modern terms as "orogenic" deposits characterized by ribbon quartz, carbonate-sericite alteration and spatial control by minor shear zones (Poulsen, 1986a).

A recurring question about the metallogeny of the Rainy Lake wrench zone concerns the apparent absence of economically viable mineral deposits compared to the numerous occurrences. While it is true the there is strong similarity between the make-up of the rocks in the Rainy Lake wrench zone and the central volcanic complexes at Chibougamau, Val d'Or and Noranda in the Abitibi subrprovince, the discrepancy in metal endowment may simply be explained in the context of the geological deposit types. For example, the metal endowment of syngenetic massive sulfide systems is thought to be negatively influenced by shallow water environments, the lack of a well-defined lithocap or by cooler upwelling fluids and this might apply to Rainy Lake. A notable characteristic of the orogenic Au-quartz veins at Mine Centre the kinematic evidence for strike-slip stress conditions for vein formation at Rainy Lake in contrast to conditions for reverse faulting allowing for higher fluid pressure at Val d'Or in the Abitibi Subprovince (Poulsen et al., 1992).

Road Log and Field Stops

A traverse which follows Highway 11 along the Rainy Lake wrench zone provides an opportunity to examine representative outcrops which illustrate its main geological features. The western starting point of the road log **Km 0.0** (92.4) is at the lookout tower on the waterfront in Fort Frances and the eastern ending point **Km 92.4** (0.0) is at the highway bridge across the Seine River near Crilly. The highway distances are approximate and, although the stops are described from west to east, they can be visited in any order depending on topical interests.

The lookout tower at Fort Frances is located on the north shore of the Rainy River near its outlet from Rainy Lake (Fig. 22). The rock exposures which Lawson (1887) originally chose as a type locality of the Archean metasedimentary biotite schist at the Coutchiching Rapids were flooded upon construction of the power dam to the west of here circa 1906. Since then, representative outcrops that illustrate the Coutchiching Group have been described nearby at Ranier, Minnesota by Ojakangas et al. (1982, Stop 1) and Jirsa and Hemstad (2010, Stop 6-2).

Drive east along Front Street and join Highway 11 and continuel eastbound from Fort Frances. Lake Road intersects the highway at **Km 1.9** (90.5). Continue through the land of the Couchiching First Nation past Couchiching Drive at **Km 3.3** (89.1). Note the discrepancy between the modern spelling compared that of the geological unit which was based on the version used topographically circa 1887. Continue past the C.N.R. Railway Crossing (**Km 5.5** (86.9)) and over the crest of the Noden Causeway bridge and continue past the intersection with a side road to the north marked "Scenic Lookout". This sideroad (**Km 8.0**, 84.4) leads to stop 13 of Czeck and Poulsen (2010). Continue eastward on Highway 11 and turn in to the next (unmarked) sideroad (**Km 8.8**, 83.6) which leads northward to a parking area beneath the hydro tower. This is **STOP NC (Noden Causeway)**.

This is an instructive stop (Fig. 23) in that this is one of the many islands in Rainy that would have been mapped both topographically by triangulation by W.H.C. Smith and geologically by Andrew C. Lawson in the 1880's. The rocks here consist mainly of foliated quartz monzonite of which Lawson first assigned to the Laurentian but later revised to the Algoman intrusive suite which he described as "mica syenite" belonging to a larger Pukamo Island intrusion (Lawson, 1913). Harris (1974) correlated these rocks with the Rocky Islet Bay Complex west of Rice Bay which are comprised mainly of felsic to intermediate granitoid rocks of variable composition. The main unit

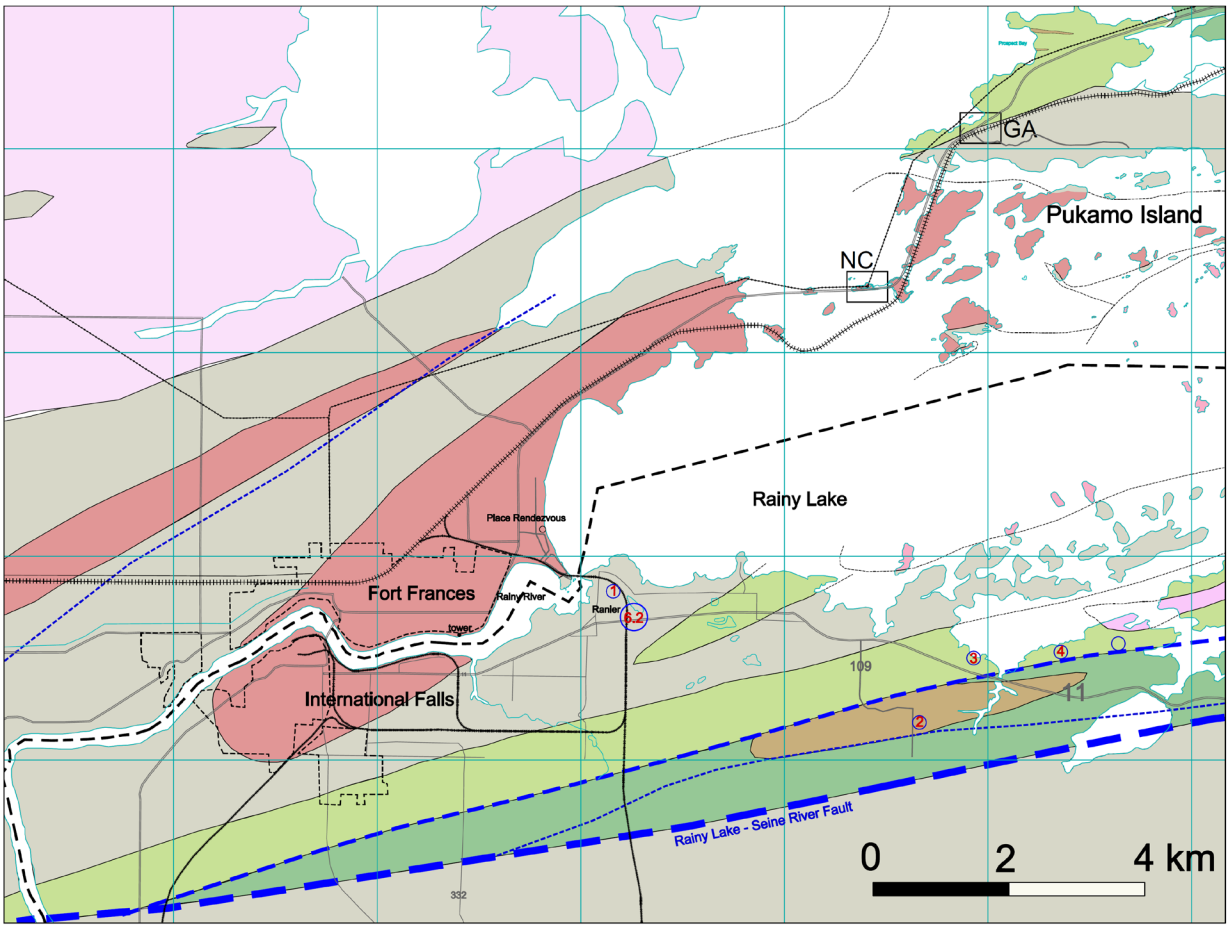


Figure 22: Simplified geology of the Fort Frances segment. Field stops NC – Noden Causeway; GA – George Armstrong Drive

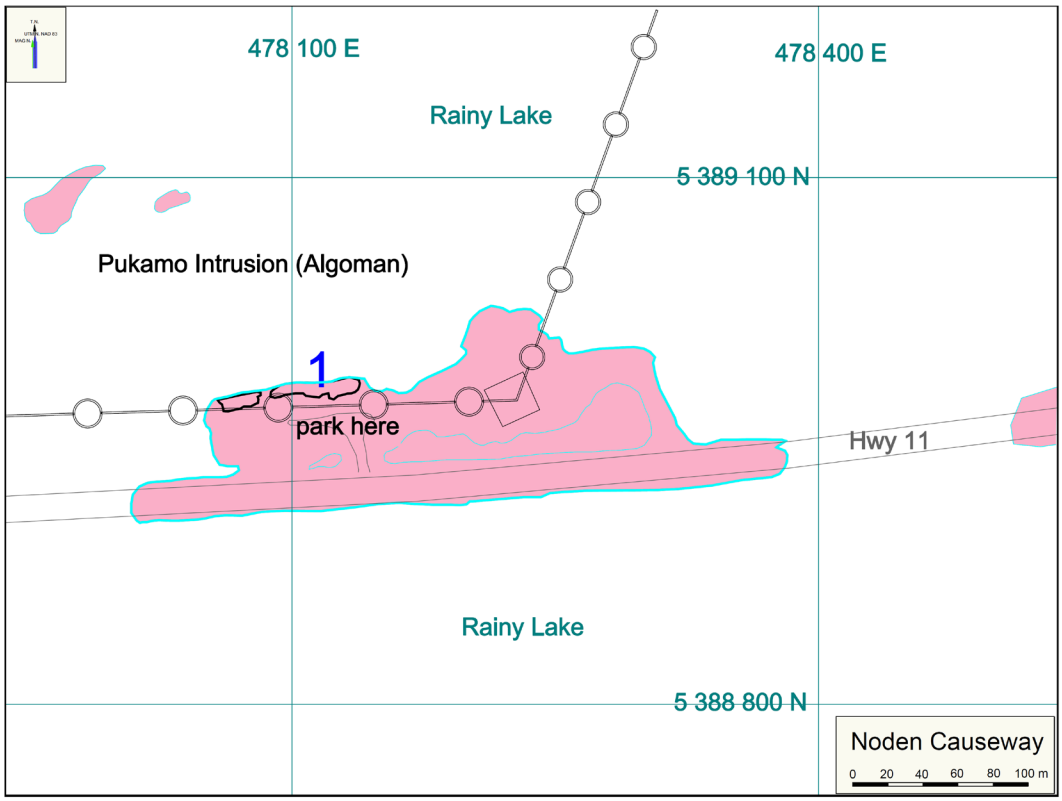


Figure 23: Noden Causeway stop (NC)

here is cut by a variety of dikes which offer contrasts in structural competence compared to the surrounding granitoid rock. Note the s-shaped asymmetric foliation pattern in one of the mafic dikes (site 1) that mimics the regional structural pattern of the Rainy Lake Wrench Zone as a whole (19c). Return to Highway 11 to resume the road log.

Km 11.8 (80.6) – George Armstrong Drive intersects highway 11 from the east; turn in and park near the mailboxes to examine **STOP GA (George Armstrong)**. This is also stop 11 of Czeck and Poulsen (2010) and Point of Interest 27 described in Pye (1968).

This is the first area of significant exposure of the Coutchiching rocks northeast of their type locality at Fort Frances. Although the nature of the contact with the Keewatin rocks is obscure (Fig. 24), it is still a good place to examine the differences between the metasedimentary biotite schists which are cut by felsic intrusive rocks (site 1) and the metavolcanic

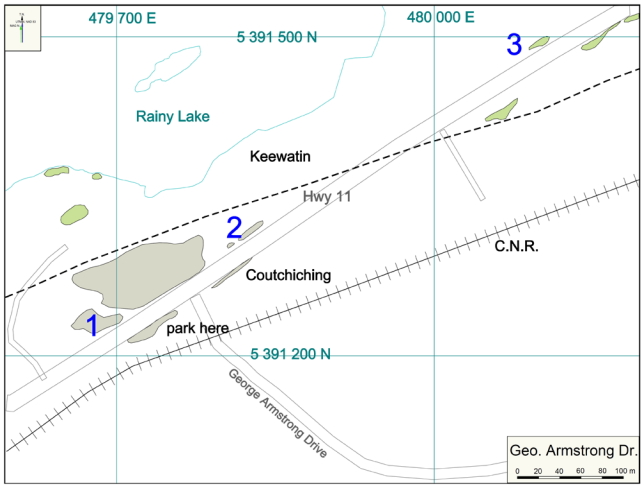


Figure 24: George Amstrong Drive stop (GA)

amphibole-biotite schists (site 3). Both units are now metamorphic tectonites which exhibit moderate to high strain but the variability of layer thickness in the metasedimentary units is consistent with their inferred origin as submarine turbidites (Ojakangas et al., 1982). Further evidence for the superimposed strain is evident at (site 2) where at least four generations of dikes cut the metasedimentary rocks and display the variable effects of folding and boudinage depending on their structural competence and pre-strain orientation with respect to bedding (see also Czeck and Poulsen (2010) and Druguet et al. (2008).

Continue eastward along Highway 11 past Commissioners Bay which is the location of a zircon sample from a Keewatin felsic which yielded a U-Pb age of approximately 2727 Ma (Davis et al., 1989).

Km 18.4 (74.0) – Windy Point Bridge

Km 21.1 (71.3) – outcrops on both sides of Highway 11. This **is STOP SM (Sims)** and corresponds in part to the Windy Point locality described by Pye (1968).

The outcrops here) display deformed pillowed and variolitic metabasalt which is a dominant lithology within the Keewatin volcanic sequence on the flanks of the Rice Bay Dome (Fig. 25). It is important to examine the exposure (site 1, Fig. 26)) carefully in three dimensions because primary pillow shapes which are inherently variable are further distorted by superposition of a moderate amount of tectonic strain. This result is log-shaped pillows with long axes that plunge moderately westward (Fig, 10a). The effects of the strain can be further appreciated by examining the cm-scale light-coloured patches that stand out against the darker amphibolitic background of the metabasalt (especially at site 2). They are varioles which predictably would have formed originally as spherical patches due to devitrification of glassy volcanic rock but here their shapes reflect their tectonic distortion with a flat aspect corresponding to a foliation and a long axis which plunges westward in the foliation. Note also that the dark pillow selvedges offer rheological contrasts with the rest of the basaltic material so that the down-plunge elongation is also expressed in places in the outcrops by boudinage of individual pillows. Pye (1968) described these outcrops without reference to their volcanic origins at all while still emphasizing the lineation and the sets of joints perpendicular to it. Even where the pillows are clearly defined the considerable strain makes it difficult to draw satisfactory conclusions about primary stratification and directions of younging. Harris (1974) and Poulsen (1980) suggested, albeit with some doubt, that the stratigraphic section in this area faces downward and eastward.

Km 24.9 (67.5) – The Nickle [sic] Lake Shores Road which intersects Highway 11 from the south leads to **STOP NL (Nickel Lake)**. This was stop 1 of Poulsen (1982).

This area illustrates the fact that, although the term Keewatin is synonymous with metavolcanic protoliths, it also contains clastic and chemical interflow sedimentary units which include oxide, sulfide, carbonate and silicate facies of iron-formation. These rocks are important from at structural point of view in that they typically have sharp magnetic and

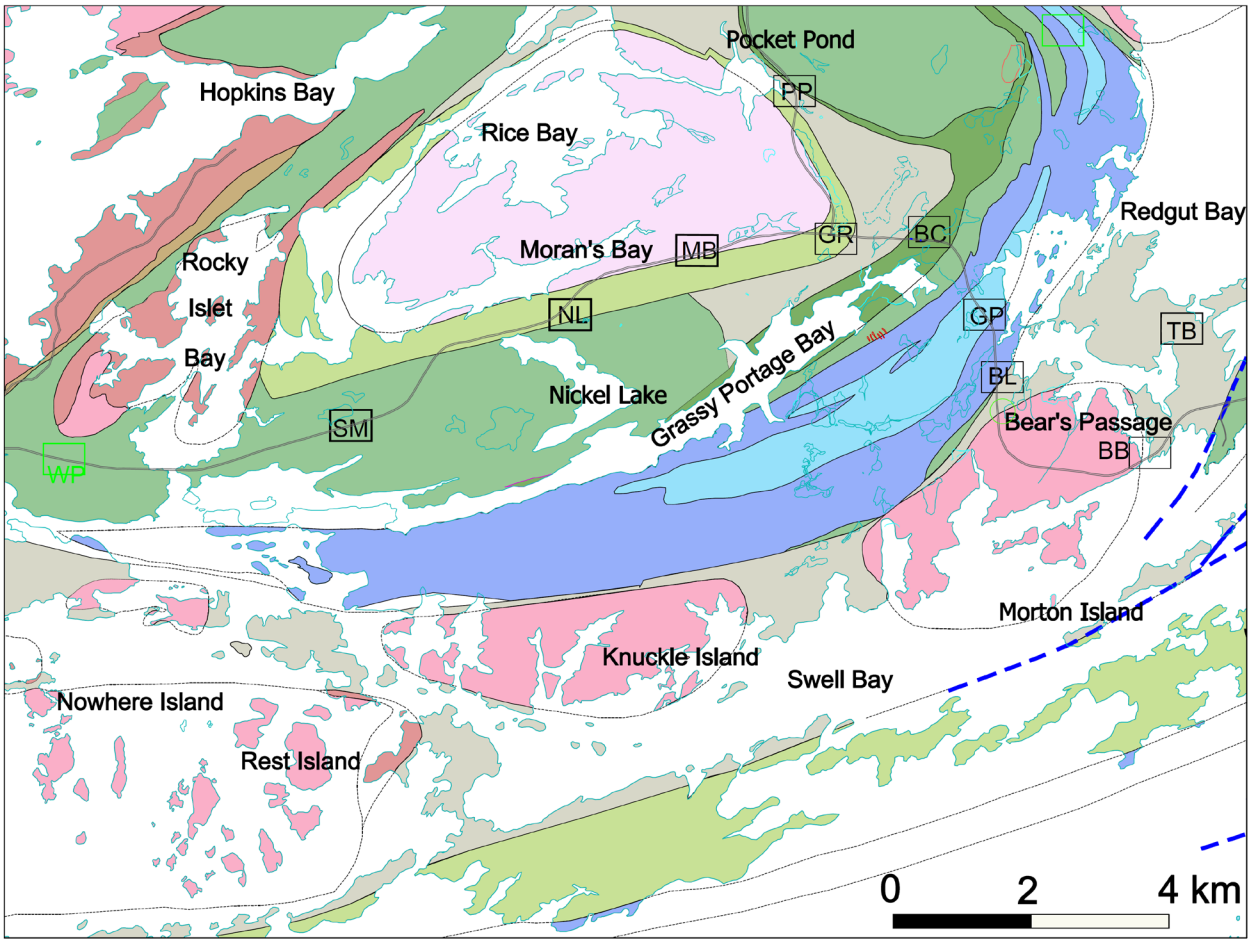


Figure 25: Simplified geology of the Swell Bay segment. Field Stops: SM-Sims; NL-Nickel Lake; MB – Moran's Bay; GR-Great River Rd.; PP-Pocket Pond; BC- Belacoma; GP-Grassy Portage; BL- Bear's Passage boat launch; BB- Bear's Passage bridge; TB-Tunnel Bay

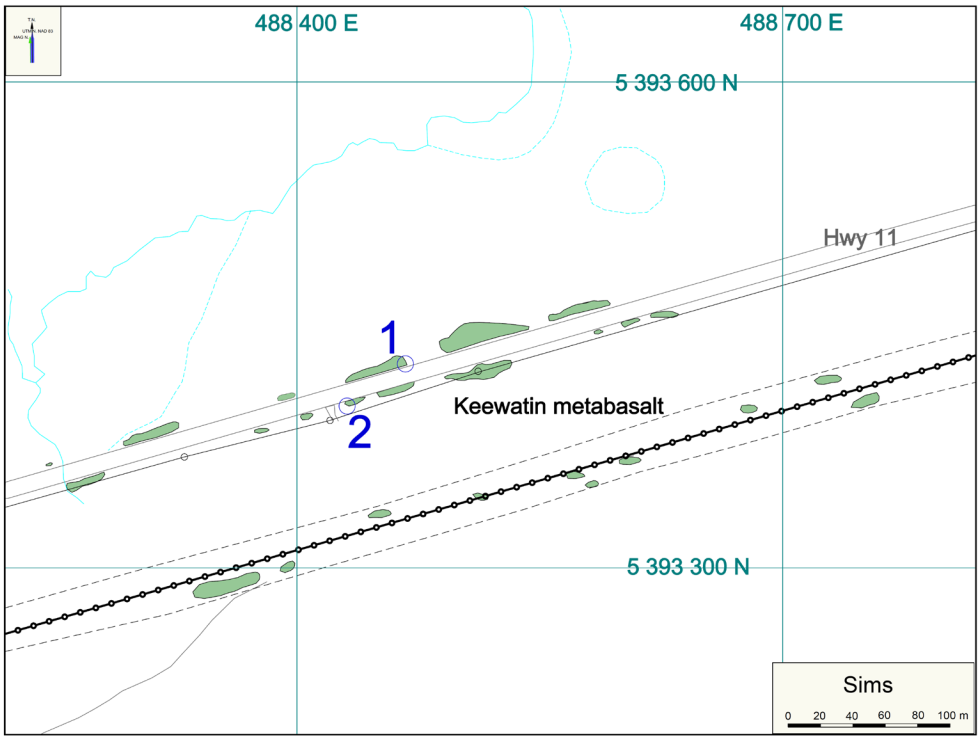


Figure 26: Sims stop (SM).

electromagnetic geophysical responses which aids in the definition of their position in areas of poor exposure. The rock here (Fig. 27, site 1) is typically referred to as chert-magnetite, banded iron formation (BIF) and is a lithology that is commonly folded at all scales. At Nickel Lake the iron-formation defines a structural synform (historically the Nickel Lake Syncline) which plunges shallowly westward along axes coincident with those of the minor folds and with the axes of maximum elongation in the adjacent volcanic rocks. The curved traces of folds observed here in a down-plunge view was originally interpreted by Poulsen (1980) to present a type 3 (coaxial) fold interference pattern. It is equally possible, however, that they result from a single deformation with a strong westward plunging linear component of strain (i.e. L-s tectonite).

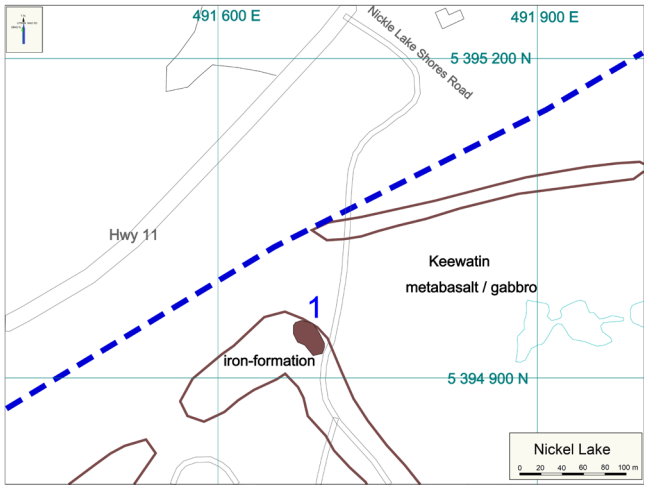


Figure 27: Nickel Lake stop (NL).

Km 26.9 (65.5) – outcrops on both sides of highway 11 but a particularly large one on the south side. This is **STOP MB (Moran's Bay)** and is described as Stop D.1 in Poulsen and Wood (1982).

The outcrop is located on the south limb of the prominent antiformal Rice Bay Dome (25). It provides ample illustration of the rocks Lawson (1914) mapped as Laurentian granite and granite gneiss in the interior of the dome (Fig. 28). Both Lawson (1913) and Harris (1974) interpreted the unit to be at least in part intrusive into the mantling Coutchiching metasedimentary rocks but the details remain in considerable doubt. R.W. Ojakangas was the first to suggest that the wispy banded, quartz-phyric, grey, foliated quartzofeldspathic can also be interpreted as a deformed rhyolite. This unit yielded a U-PB zircon age of 2725+/-2 Ma (Davis et al., 1989). It is cut by more competent sheets of coarser quartz-feldspar porphyry (Fig. 15e) which could potentially represent coeval subvolcanic intrusions or younger sills Algoman which are responsible for the cross-cutting relationships. Both lithofacies display prominent polycrystalline quartz aggregates which are likely deformed phenocrysts which help define both the tectonic foliation and a prominent lineation which plunges shallowly eastward at this locality (Fig. 15f).

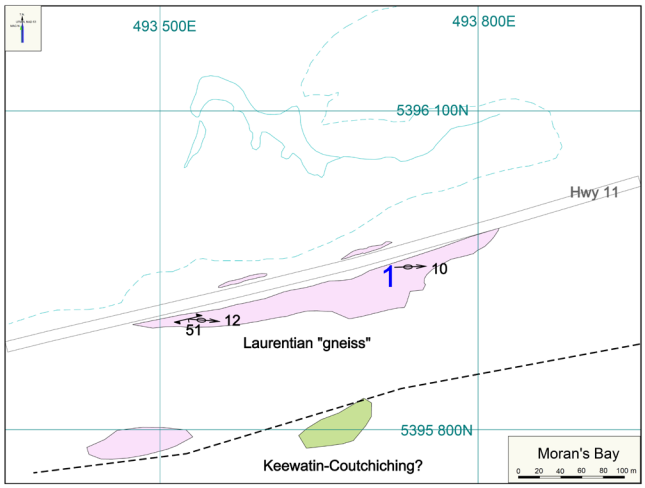


Figure 28: Moran's Bay stop (MB)

Km 29.0 (63.4**)** intersection between Highway 11 and Highway 502 (Fig. 25). This is **STOP GR (Great River Rd.)** which corresponds to Stop 10 of Czeck and Poulsen (2010).

Folded quartz-phyric intrusions on the north side of Highway 11 west of the intersection (site 1, Fig. 29) cut amphibole-biotite schists containing local iron-formation which were included with the Coutchiching biotite schist on the maps of Lawson (1914) and Harris (1974) but which show greater similarity to Keewatin units elsewhere. The porphyritic felsic intrusions have been generally included in the suite of Laurentian intrusions but the molybdenite-bearing quartz veins exposed here are also a characteristic of Algoman intrusions elsewhere. Despite these uncertainties of interpretation and the somewhat transitional nature of the contacts, it is clear these rocks serve to separate the inner core of the Rice Bay dome from a structurally higher annular band of moderately southeastward-dipping Couchiching biotite schists which are well exposed approximately east of the intersection (site 2).

It is also possible to make a short side-trip form this intersection northward along highway 502 for 2.2 km to its intersection with the Baseline Bay side road which enters from the east. This is **STOP PP (Pocket Pond)** and corresponds to Stop D.2 of Poulsen and Wood (1982).

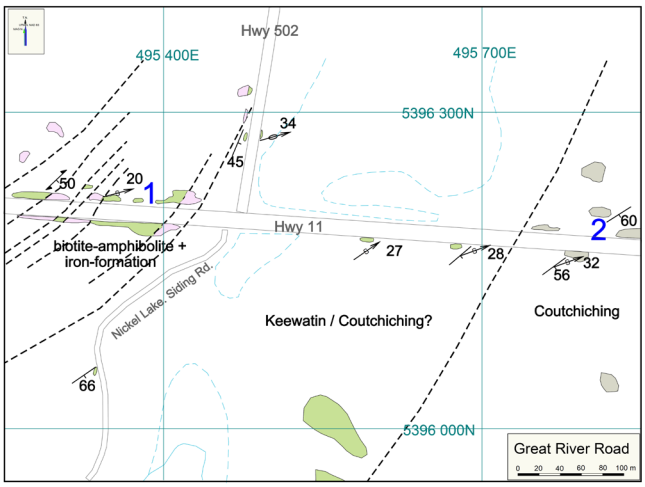


Figure 29: Great River Road stop (GR)

The critical outcrops (Fig. 30, site 1) that demonstrate the overturned stratigraphic section on the northern limb of the Rice Bay dome are now heavily overgrown and no longer instructive. A good sense of the nature of the northeastward-dipping contact between the Coutchiching metapelites and the distinctive green, magnetic ultramafic unit which here represents the Keewatin volcanic rocks can still be observed along Highway 502 (site 2). Continuity of the lithostratigraphic units and their moderate northeasterly dips in this area were established with the assistance of ground magnetic and electromagnetic surveys and by diamond drilling which targeted Cu-Zn mineralization associated with the interflow iron-formation units in the section. Although the contacts among the units are sharp and well defined there is no conclusive evidence for them to be erosional-depositional in origin but the evidence for an overturned volcanic sequence is sound (Fig. 30).

Return to Highway 11 and resume the road log.

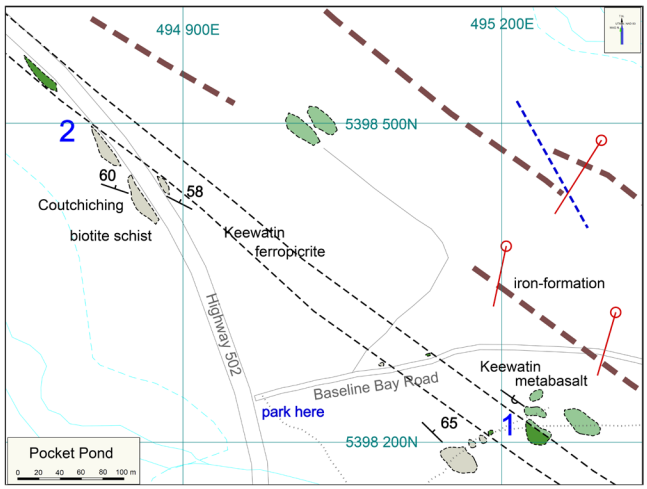


Figure 30: Pocket Pond stop (PP)

Km 30.2 (62.2) – bush road and outcrops on north side of Highway 11. This is **STOP BC (Belacoma)** corresponding to stop 3 of Poulsen (1982, stop D.3 of Poulsen and Wood (1982) and stop 5 of Hinz (2010).

This is a continuation of the Coutchiching-Keewatin contact which extends southward from Pocket Pond and westward to Nickel Lake and sharply defines the eastern closure of the Rice Bay Dome. The volcaniclastic ferropicrite unit here is exposed over a wider area than at Pocket Pond and the full nature of the contact is uncertain. The ultramafic rocks near the contact with the structurally underlying Coutchiching biotite schists (Fig. 31, site 1) are foliated as but appear to be progressively less deformed eastward (sites 2 and 3). Nonetheless, graded bedding of reasonable quality suggests the Coutchichiing strata are overturned in support of the observations at Pocket Pond. The cluster of outcrops near the beaver pond (site 3) have been documented by Schaefer and Morton (1991), Goldstein and Franceis (2008) and Hinz (2010) and the inference is that this unit is composed of relatively rare mantle-derived ultramafic coherent and pyroclastic rocks that locally contain well-preserved accretionary lapilli (Fig. 10f).

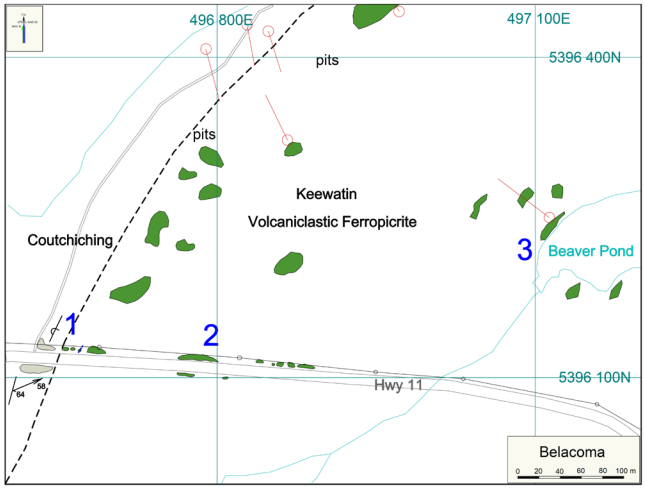


Figure 31: Belacoma stop (BC)

Km 31.3 (61.1) C.N.R. overpass

Km 31.9 (60.5) – numerous outcrops on both sides of Highway 11; safe parking is available beneath the powerline on the west side of the highway (Fig. 32). This is **Stop GP (Grassy Portage)** and corresponds to Stop D.4 of Poulsen and Wood (1982).

The gabbroic rocks exposed here are part of the metamorphosed Grassy Portage layered mafic intrusion and include plagioclase-rich leucogabbro (site 1) that gives way southward to garnet-bearing quartz-

ferrodiorite (site 2). The garnets are metamorphic porphyroblasts that likely crystallized owing to the favourable bulk composition of the diorite which has a higher Fe/Mg ratio and silica content than the leucogabbro. Lawson's 1914 map of the area portrayed the leucogabbro as "hornblende gabbro" alone as an intrusion within the Keewatin while including the gabbro and melagabbro to the north and the garnet-bearing quartz diorite to the south as Keewatin metavolcanic rocks. This inferred symmetry led to his interpretation of a synclinal axis centred on the leucogabbro but Harris (I974), Poulsen (1980) and Poulsen and Hodgson (1986) recognized all three lithofacies as distinctively different phases of a single layered mafic intrusion that shows progressive southward, upward in a stratigraphic sense, chemical and mineralogical fractionation.

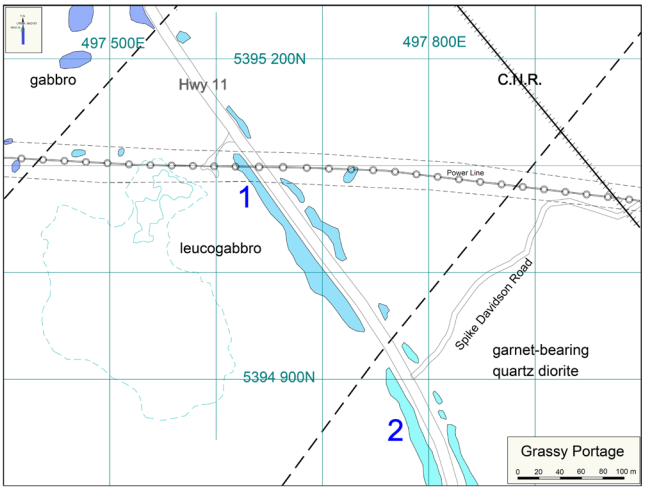


Figure 32: Grassy Portage stop (GP)

Km 33.6 (58.8) - the side road on the east side of Highway 11 leads to the boat launch at Bear's Passage where parking is available at the lakeside (Fig. 33). This is **STOP BL (Bear's Passage Boat Launch)** corresponding to Stop D.5 of Poulsen and Wood (1982) and locality 20 of Uglow (1913).

This critical area of outcrop illustrates one of the most contentious points of the Seine-Coutchiching problem. The Keewatin rocks which are cut locally by a foliated lamprophyre dike structurally overlie gabbroic rocks of the Grassy Portage layered intrusion (site 1.) The Coutchiching rocks are staurolite-bearing biotite schists (site 2) and locally display evidence of primary graded bedding with is enhanced by the distribution of porphyroclasts in upper parts of individual beds. Graded bedding which can be observed directly adjacent to the relatively sharp Keewatin-Coutchiching contact (site 3) is consistent with southward younging in the meta-turbidites and contradicts the structural order of the rocks based on dip alone. It is, however, consistent with the southward younging implied by the fractionation within the Grassy Portage layered mafic intrusion.

Return to Highway 11 to resume the road log

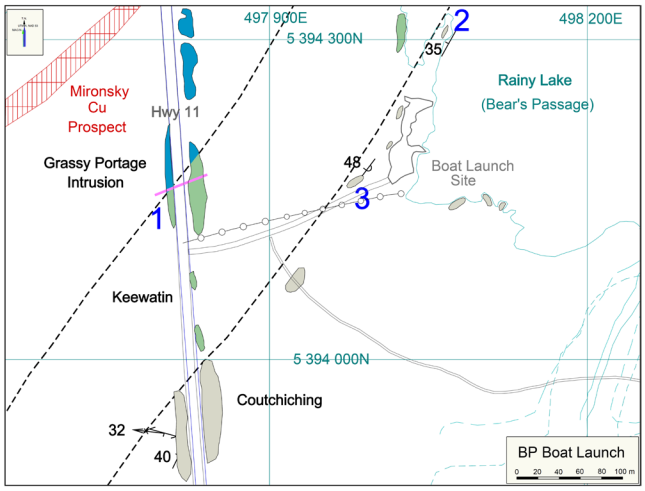


Figure 33: Bear's Passage Boat Launch stop (BL)

Km 36.4 (56.0) – Taylor's Road intersects Highway 11 from the north

Km 37.0 (55.4) – parking area and scenic view on South side of the highway (Fig. 34). This is **STOP BB (Bear's Passage Bridge)** corresponding to Stop 7 of Poulsen (1982) and Point of Interest 2 of Pye (1968).

The eastward dipping Coutchiching biotite schist, as exposed on the north side of the highway (site 2). is cut by granodiorite of the Bear Pass Pluton which contains sheeted quartz-pyrite-molybdenite veins which are exposed on both sides of the bridge (sites 1 and 3) The view southward from the lookout features Swell Bay and the belt of Keewatin volcanic rocks to the south of it. The Keewatin-Coutchiching contact is located on Morton Island to the southwest.

Km 37.7 (54.7) – Bear Pass Road intersects Highway 11 from the north. From this location it is possible to make a side trip to **STOP TB (Tunnel Bay)** by driving northward for 1.3 km to the C.N.R. tracks and taking the first dirt road uphill to an exposure of Coutchiching metasedimentary rocks (Fig. 35). This area is near Tunnel Bay and localities 5 and 6 of Uglow (1913).

These outcrops are located on the eastern limb of the antiformal culmination in the Bear's Passage area. The demonstration of the existence of the antiform

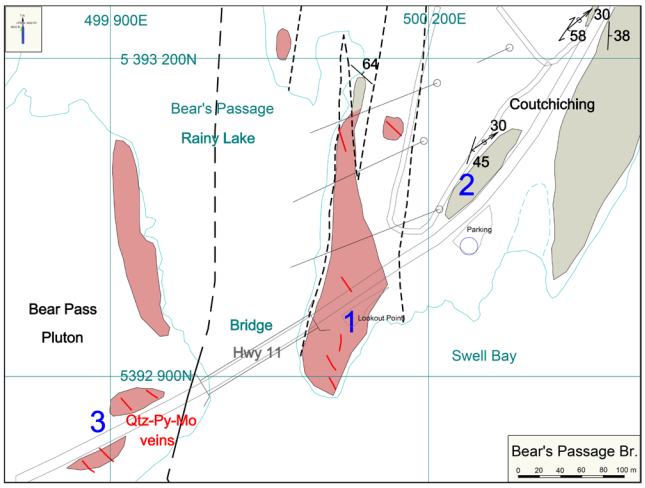


Figure 34: Bear's Passage bridge stop (BB)

was essential to Lawson's (1913) interpretation of the Coutchiching strata in the interior of this structure. It is also the location where D.W. Davis first demonstrated the effects of zircon inheritance from the Coutchiching metasedimentary rocks by felsic dikes related to the Bear Pass Pluton. One of the dikes near the stop of a steep outcrop can be viewed to the east at (site 2). An additional point of interest in these exposures (sites 1 and3) is that the main foliation is locally crenulated by a steep, northwest striking, transecting cleavage (S3 of Poulsen, 1980), which is particularly prominent in a 2 km-wide northwesterly trending corridor through this area. Although locally dominant at the mesoscopic and mircroscopic scales, where crenulation of the main biotite-rich foliation and rotation of metamorphic porphyroblasts are both evident, the effects of this deformation at the macroscopic scale are negligible.

Return to Highway 11 to resume the road log

Km 40.8 (51.6) – Old Station Road intersects Highway 11 from the north (Fig. 36). This is **STOP**

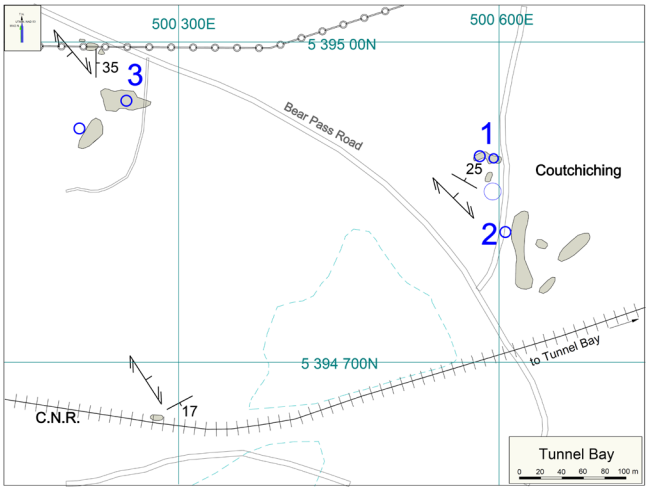


Figure 35: Tunnel Bay stop (TB)

OS (Old Station Road) where the field relationships are comparable to those at Morton Island (Stop D.6 of Poulsen and Wood (1982)).

Highway 11 at this locality (Fig. 37) is approximately parallel to the strike of stratification in the Coutchiching meta-sdedimentary rocks as well as to their mapped contact with Keewatin meta-volcanic rocks (Harris, 1974). The overall dip of bedding is steep to the southeast and in places a steep cleavage with a more northerly strike transects bedding to form a moderately eastward plunging intersection lineations. Exposure is plentiful but the clearest features of the Coutchiching beds are illustrated in flat outcrops on the south side of Highway 11 (site 1). The rocks here are metamorphosed to greenschist facies assemblages and primary features are reasonably well preserved: polarity in graded beds consistently indicate a northward direction of younging which is away from Keewatin volcanic rocks which are exposed at the shore of Rainy Lake south of here.

Km 43.3 (47.9)) – Ottertail Landing Road intersects Highway 11 from the north.

Km 46.2 (46.2) – a side road to a communications tower intersects Highway 11 from the north: turn in and park (Fig. 38). This is **STOP OW (Ottertail West).** The field relationships exposed in the outcrops east of the intersection on the north side of Highway 11 are comparable to those at stop 1 of Czeck and Poulsen (2010) which is located approximately 1 km to the west.

This is an area in which a roof pendant composed of foliated metavolcanic and metasedimentary schists has been variably incorporated into granitoid rocks of the Ottertail Lake intrusion. The outcrops here provide a rare case where highway improvement has also resulted in outcrop improvement. A marginal phase of the Ottertail Lake intrusion (site 1) is composed of diorite containing abundant mafic xenoliths (Fig. 18d). Magmatic breccias (Fig. 18e) are well exposed along the highway to the east (site 2) and, at one location nearby, a narrow NNE-striking mylonitic zone cuts the intrusive rocks. The most critical point made by Lawson (1913) and most observers since is that there is abundant visual evidence for intrusion of felsic magma into previously foliated metamorphic tectonites.

Km 47.5 (44.9) – the outcrop on the north side of the road was sampled by D.W. Davis to yield a U-Pb zircon age of 2686+/-3 Ma for this part of the Ottertail Lake intrusion.

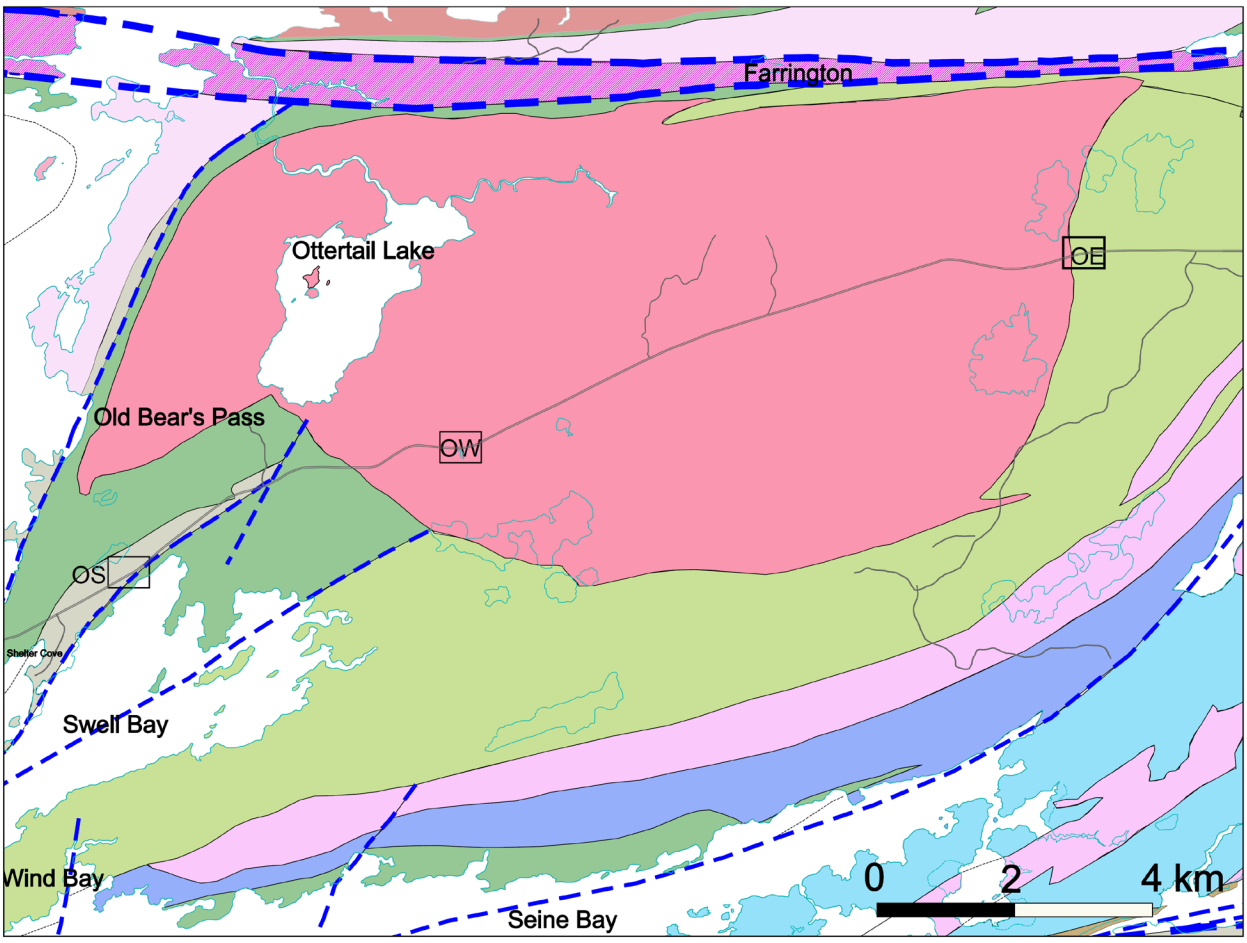


Figure 36: Simplified geology of the Ottertail Lake segment. Field Stops: OS- Old Station Rd; OW: Ottertail Lake West; OE-Ottertail East

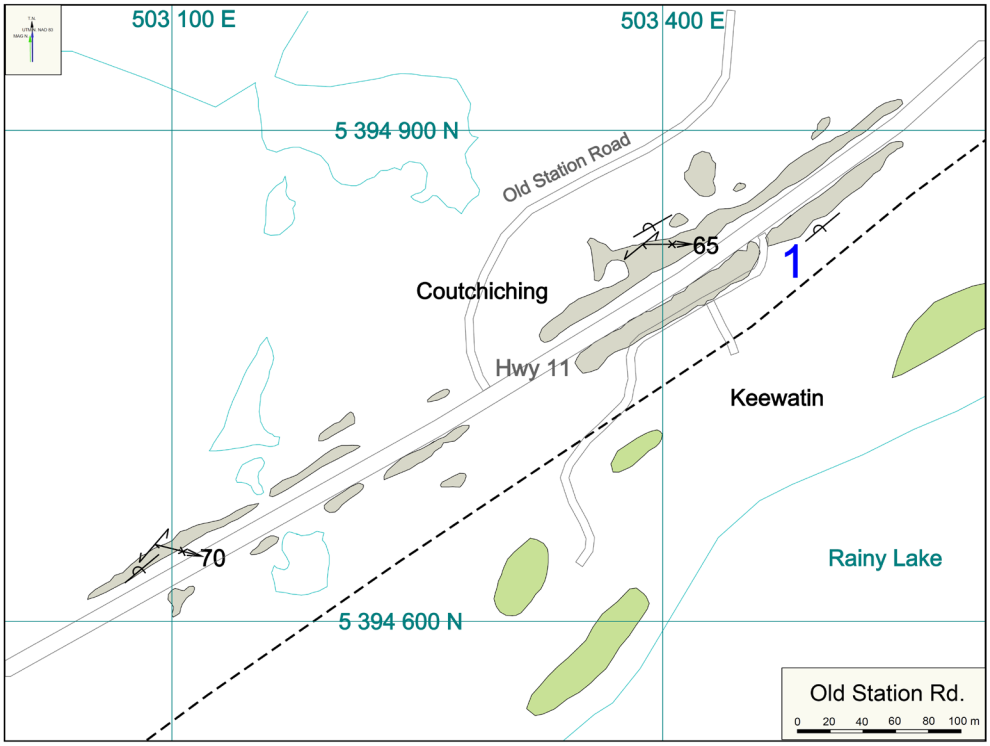


Figure 37: Old Station Road stop (OS)

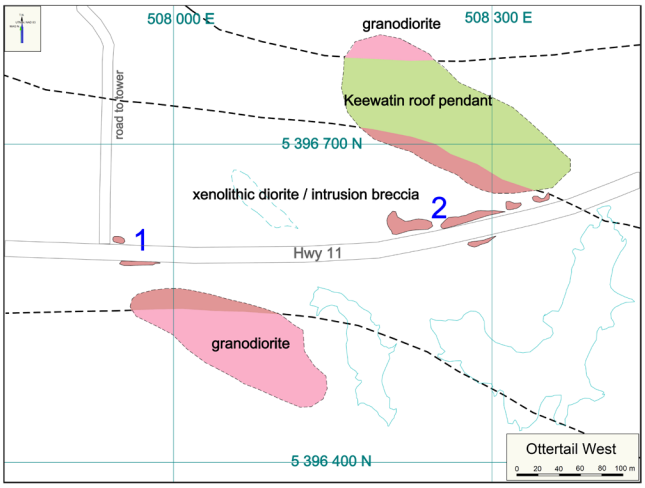


Figure 38: Ottertail West stop (OW)

Km 48.2 (46.2) – Pearson's Road intersects Highway 11 from the north

Km 53.9 (38.5) – the outcrops of Ottertail Lake intrusive rocks exposed here are described as Stop 2 in Czeck and Poulsen (2010).

Km 56.1 (36.3) – outcrops on both sides of the road expose the eastern margin of the Ottertail Lake Intrusion. This is **STOP OE (Ottertail East)** and corresponds in part to Stop D.7 of Poulsen and Wood (1982).

The easternmost outcrop on the north side of Highway 11 (site 1, Fig. 39) exposes deformed spherulitic and flow-banded rhyolite that is common in the Keewatin volcanic section in this part of the belt. It is cut by granitoid phases or the Ottertail Lake Intrusion, including a distinctive feldspar-phryic variety containing xenoliths (site 2). The abundance of xenoliths decreases westward in these outcrops (site 3).

Km 56.4 (36.0) intersection of Highway 11 and

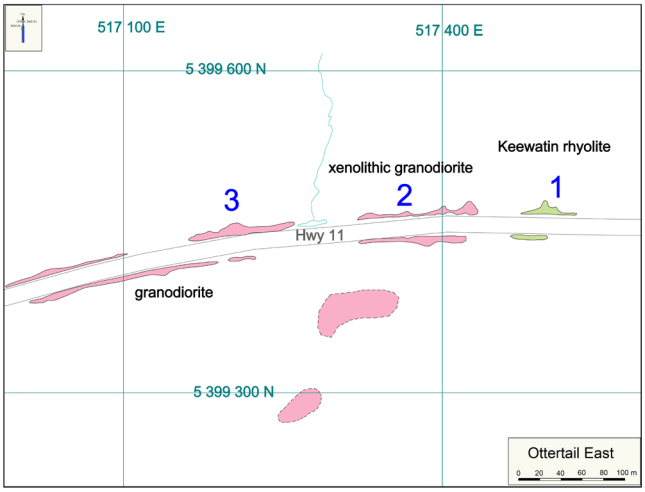


Figure 39: Ottertail East stop (OE)

Turtle River Road.

Km 58.8 (33.6) Patten Park picnic area

Km 62.0 (30.4) – Low outcrops are present on both sides of highway (Fig. 40) and a rusty waste dump is visible across a marshy area on the north side of Highway 11. This is **STOP PA (Port Arthur Copper)** which was stop 9 of Poulsen (1982) and Point of Interest 22 in Pye (1968).

Access the rusty area from the west side of the water and cross a small Beaver Dam to reach the large area of exposure (Fig. 41). The main mineralized lithology is composed of foliated amygdaloidal andesite (Fig. 10c) containing disseminated and semi-massive lenses of pyrite, chalcopyrite and sphalerite (site 1). Stratified, rusty felsic volcanic rocks are exposed on the north side of the outcrop area (site 2). This is but one of several occurrences of syngenetic sulfide deposits hosted by the felsic portions of the Keewatin volcanic section extending more than 25 km southwestward beyond Wind Bay. It is also noteworthy that the base metal deposits are located up-section northward from the syn-volcanic Bad Vermilion Lake intrusive complex (Fig. 40).

Km 63.4 (29.0 side road intersects Highway 11 from the north

Km 67.0 (25.5) -the Mine Centre Road intersects Highway 11. This road can be followed north to Little Turtle Lake by travelling for 1.0 km to Government Road and continuing .5 km to the C.N.R. tracks. Bear right at the intersection with Queen St. and follow the dirt road to the public boat launch site. The outcrops near the shoreline constitute **STOP LT (Little Turtle Landing)** which corresponds to Stop D.9 of Poulsen and Wood, 1982).

Lawson (1913) mapped the rocks that are exposed here as a distinctive lithological unit which he described as "porphyroid gneiss". In doing so, he effectively defined a 60 km E-W segment of what is now known as the Quetico Fault without explicit reference to faults but certainly recognized the overall significance of the rock type in "that it has a pronounced cataclastic structure and that the schistosity of the rock is referable to deformation involving shearing of the mass" (Lawson, 1913. P.94). Today the lithologies which he described are regarded as variably deformed fault rocks which include protomylonite (site 1) which is exposed in the outcrop east of the parking area and mylonite (site 2) along the shore of Little Turtle Lake.

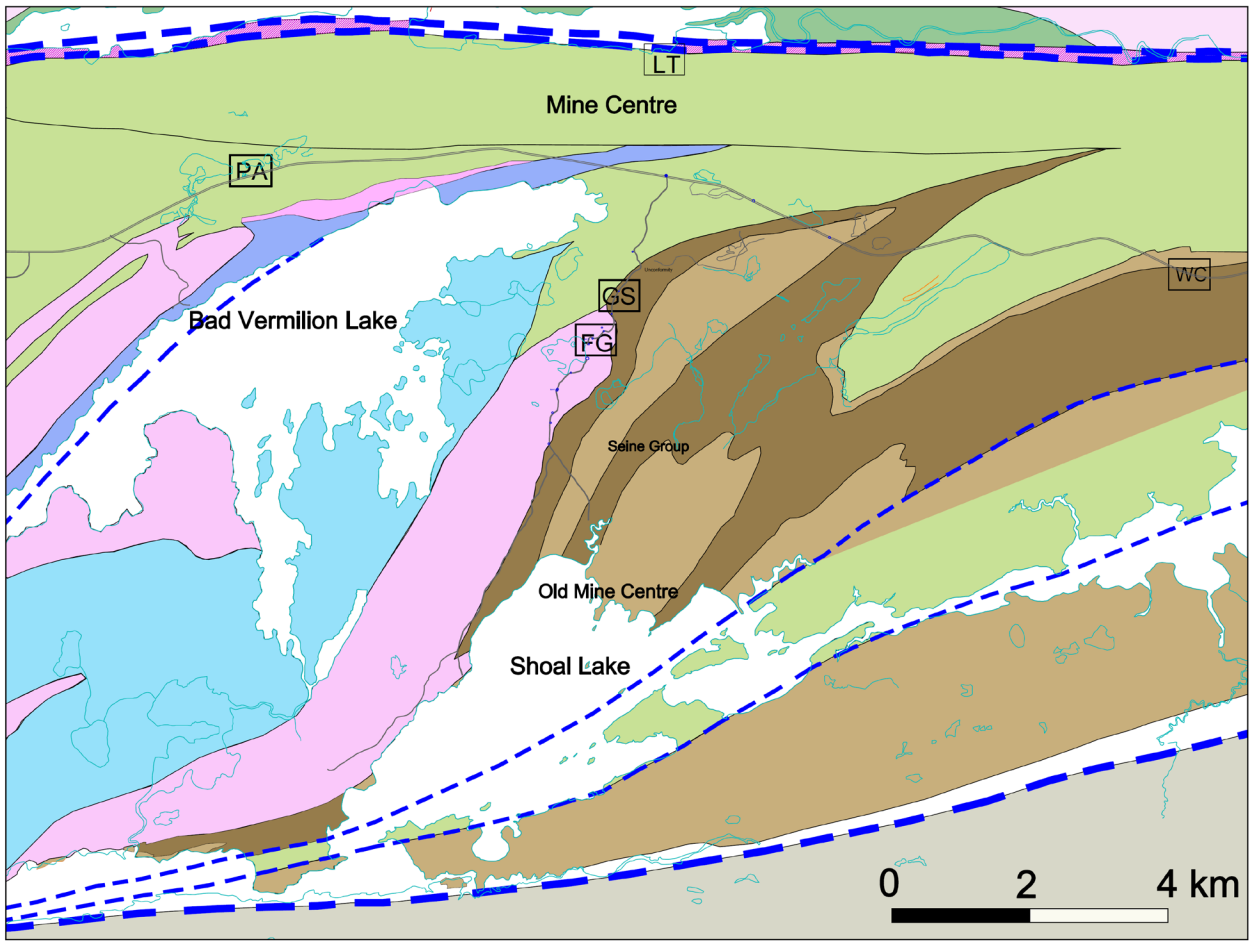


Figure 40: Simpified geology of the Mine Centre segment. Field Trip Stops: PA-Port Arthur Copper; LT-Little Turtle landing; FG: Ferguson; GS-Golden Star; WC-Windy City Rd.

Regrettably, a recently constructed dock partially obscures the best exposure of the folded mylonite (Fig. 19f) as described in Poulsen and Wood (1982).

Return to Highway 11 and resume the road log

Km 68.2 (24.2) The Shoal Lake Road meets Highway 11 from the south (Fig. 40). This road leads to what is arguably the most significant geological feature in the entire belt – the angular unconformity at the base of the Seine Group metasedimentary rocks. Follow the (in places rough) Shoal Lake public road southward for 3.3 km to a point where it is met from the east by a recently constructed but as yet uncompleted sideroad. The is **STOP FG (Ferguson)** and outcrops in this area

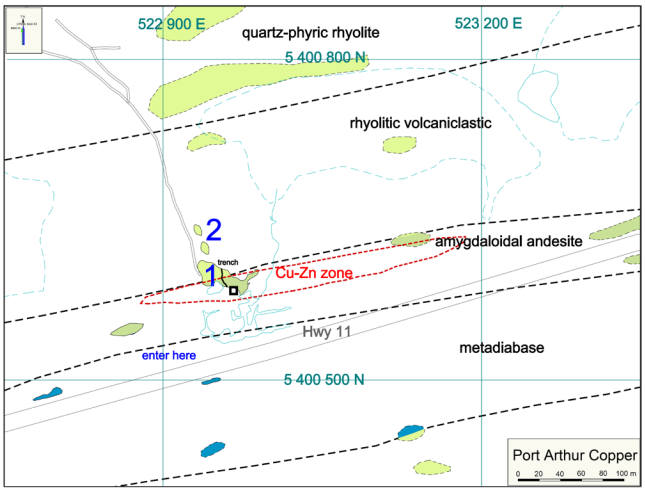


Figure 41: Port Arthur Copper stop (PA)

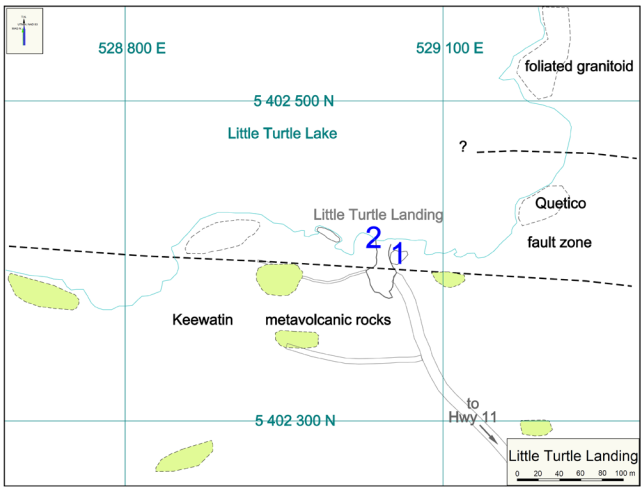


Figure 42: Little Turtle Landing stop (LT)

offer a good view of the Bad Vermilion tonalite which Lawson assigned to his Laurentian suite of granitoid rocks.

The recent excavation (site 1) has exposed the tonalite and small quartz veins with adjacent sericite-ankerite alteration of the style exposed at the Ferguson gold prospect to the north (site 2). Lawson (1913) demonstrated conclusively that the tonalite cuts both the Bad Vermillion gabbro-anorthosite to the west and Keewatin volcanic rocks to the north which include moderately northward dipping interflow chert-carbonate units and a northward younging unit of pillow basalt.

Turn around and proceed back northward along the Shoal Lake Road for 2.2 km to a small rise with a low outcrop on the east side; pull to the right side of the road and park as safely as possible (Fig. 44). This is

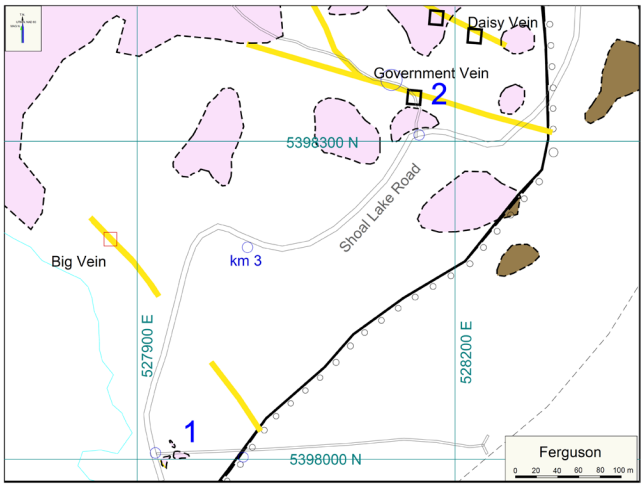


Figure 43: Ferguson stop (FG)

STOP GS (Golden Star) and the site of Stop D.10 of Poulsen and Wood (1982) and the contact described by Uglow (1913) as being marked by "brown flags" for the International Geological Congress Field Trip led by Lawson. The field relationships here have also been described and discussed more recently as Stop 1 of Czeck and Fralick (2020).

The base of the Seine Group here dips gently eastward at high angle to stratification in the Keewatin rocks. Much of the outcrop (site 1) is now grown over but five small patches have been recently cleaned to clearly show the west to east transition from quartz-bearing tonalite a), tonalite sand with rare clasts (b) to angular conglomerate (c) with interstitial sand (fanglomerate of Lawson) to polymictic pebble and cobble conglomerate (d, e). Although there is evidence of a weak tectonic foliation superimposed on the conglomerate (Fig. 17a, b) it is also clear that even the least competent lithic clasts possessed no tectonic fabric at the time of deposition across strata with a pre-existing steep dip.

Return northward to Highway 11 and continue eastward along it.

Km 76.5 (15.9) an unmarked bush road meets

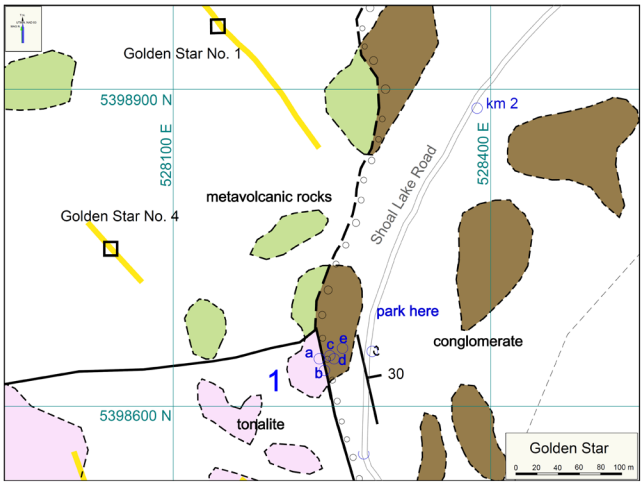


Figure 44: Golden Star stop (GS)

Highway 11 on the south side; pull in and park. This is **STOP WC (Windy City road)**. The increasingly overgrown leads southward from here for approximately 500 metres to a sign which explains how a windstorm in 1988 flattened trees over a seven km^2 area resulting in its nickname of "Windy City". Reclamation of the area resulted in local removal of shallow glacial overburden to produce two-dimensional pavement exposures of cobble to boulder conglomerate which show the rheological effects of superimposed strain. These outcrops comprise the "Forest Tour" Stop 5 of Czeck and Poulsen (2010)) can be reached by continuing another 250 m southward beyond the sign and following the second sideroad to the southwest (approximate UTM NAD 83 Zone15 N: 536 800E, 5 398 500N). The outcrops exposed at the intersection along highway at its intersection with the Forest Tour Road, however, make for a good and easily accessible substitute stop.

The outcrops occur along both sides of the highway and serve to illustrate three important aspects of Seine Group as a whole. The first is the distinction between the two main lithofacies: polymictic clast-supported conglomerate (site 1) versus thick-bedded, locally cross-bedded, arenaceous sandstone (site 2) which occupies the middle part of the Seine stratigraphic

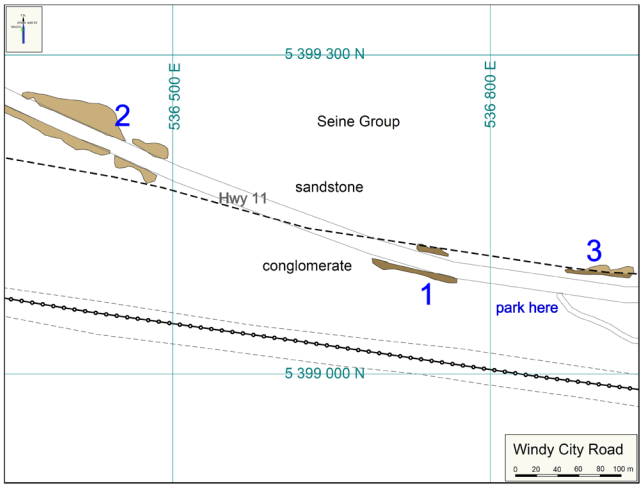


Figure 45: Windy City Road stop.

section – most evidence suggests that the strata young southward toward the polymictic conglomerate units. Second, a good three-dimensional view of the shape fabrics shows both elements of both foliation and eastward plunging lineation as well as the rheological differences in response to the bulk strain by clasts of different original composition and grain size. Third, a population of granitoid clasts is particularly noticeable in this part of the Seine stratigraphic section and these were commonly assumed to have been sourced in the Laurentian granitoid suite. A sample from this area (site 3) was collected and analysed by D.W. Davis to demonstrate that the age of a granitoid clast was actually consistent with that of rocks which are regarded to be part of the Algoman suite (Davis et al., 1989).

Km 78.4 (14.0) the Manion Lake Road meets Highway 11 from the north (Fig. 46).

Km 82.2 (10.8) Horsecollar Junction – the road to the south leads to the Seine River village and the outcrops the deformed conglomeratic facies of the Seine Group on the north side of the highway constitute Stop 5 of Czeck and Fralick (2002).

Km 92.1 (0.3) the Crilly Road meets Highway 11 from the north.

Km 92.4 (0.0) Highway bridge across the Seine River (Fig. 47). This is **STOP SR (Seine River Bridge)** and is also described as Stop D12 of Poulsen and Wood (1982) and Stop 5 of Czeck and Fralick (2002).

The outcrop southeast of the bridge provides an excellent visual representation in cross-section of the mixed arenite-conglomerate facies of the Seine Series. The overprinting steep foliation corresponds to pronounced shape fabrics in clasts at high angle to bedding in pebble conglomerate and the comparable shortening across the foliation is manifested by steepening and distortion of the foresets in the cross-bedded sandstone units. The beds dip shallowly northward and this also corresponds to the inferred

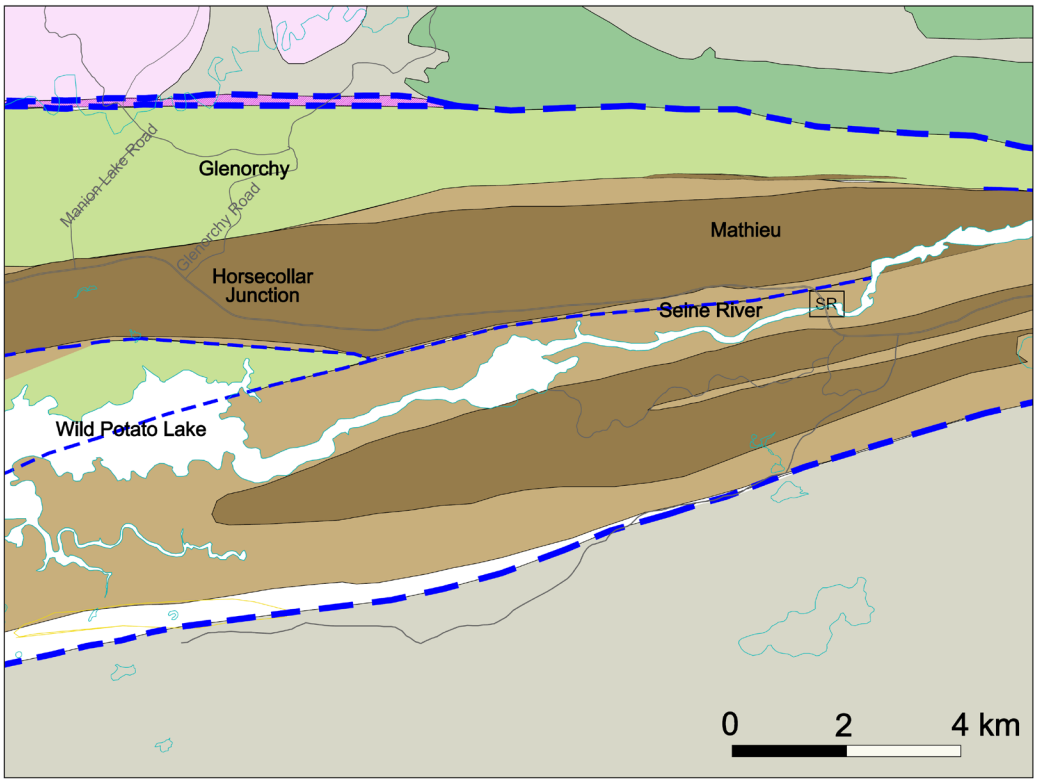


Figure 46: Seine River segment

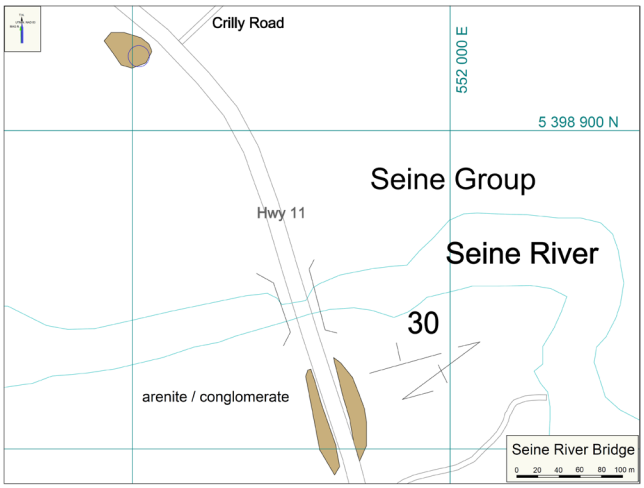


Figure 47: Seine River Bridge stop.

direction of stratigraphic younging. Although not formally defined as a type locality for the Seine Series, the outcrops here are arguably a good reference locality.

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Trip 6 - Amethyst Deposits of Thunder Bay

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Introduction

Properties of Amethyst

Amethyst, occurring in abundance in the Thunder Bay region, is purple gemstone variety of α-quartz. It has been known for some time that an iron impurity in quartz is the underlying source of amethyst coloration (Holden, 1925). However, incorporation of iron of alone cannot account for the formation of amethyst, as many varieties of quartz contain trace amounts of iron, yet amethyst is relatively rare, and large deposits of amethyst are very rare.

In a series of papers by Cohen and co-workers, culminating in a summary in Cohen (1989), a simultaneous sequence of reactions was proposed for the formation of amethyst.

(1) $(Al–O)^- \rightarrow (Al–O)^\circ + e^-$

Ionizing radiation forms a hole center from oxidizing the substitutional Al-O bond.

(2) $Na^+ + e^- \rightarrow Na^\circ$

Electron from step 1 is trapped by an interstitial alkali metal ion.

(3) $Fe^{3+}int \rightarrow Fe^{4+}int + e^-$

Induced ionizing radiation forms a trapped hole center via oxidizing the interstitial Fe^{3+}.

(4) $(Al–O)^\circ + e^- \rightarrow (Al–O)^-$

Trapped hole center is satiated as [AlO°] is reduced via gaining the electron from step 3.

The presence of iron is positions interstitial with respect to the SiO_4 framework was established by Adekeye and Cohen (1986), in noting its correlation with pervasive Brazil law twinning in colored sectors of amethyst crystals. Data on incorporation of the alkalis Na, K and Li and trivalent Al and Fe in quartz were reported by Deer et al. (1963), who further noted that the incorporation of Al^{3+} (and presumably Fe^{3+}), is compensated by the incorporation of Na^+ or Li^+ interstitial sites. The color of amethyst is produced by absorptions of light in the visible region of the spectrum owing to the presence of Fe^{4+}, as originally shown by Cox (1977).

The proposed mechanism requires the coincidence of four geological conditions for the formation of amethyst:

(1) The incorporation of Fe and Al, as well as Na or Li. This is not a limiting condition, as the small concentrations of these trace elements are readily available in hydrothermal solutions.
(2) A source of ionizing radiation, either from U and Th or ^{40}K in order to produce the defects in Fe and Al.
(3) Deposition at generally rather shallow depth such that oxidizing conditions prevail and iron is in the form of Fe^{3+}.
(4) Deposition with a temperature range for the stability of Fe^{4+}, the source of amethyst coloration.

The mechanism proposed above is consistent with observed data and provides a logical mechanism for the formation of amethyst. However, Rossman (1994) noted that there are unestablished factors in the model such that its acceptance is tentative.

Crystal forms expressed in amethyst are invariably simple, consisting only of combined positive {101} and negative {011} rhombohedra. The faces of one of the forms are generally largely and are designated as the major rhombohedron *r*, and the other form is designated as the minor rhombohedron *z* (Fig. 1). The only other form occasionally observed is the ditrigonal prism *m* (Frondel, 1962).

Amethystine coloration is unevenly distributed in the crystal, generally with concentration in the major rhombohedral forms, in which Brazil law twins are also concentrated (Fig. 2; Frondel, 1962). The orientation of Brazil law twins in Figure 2, is typical

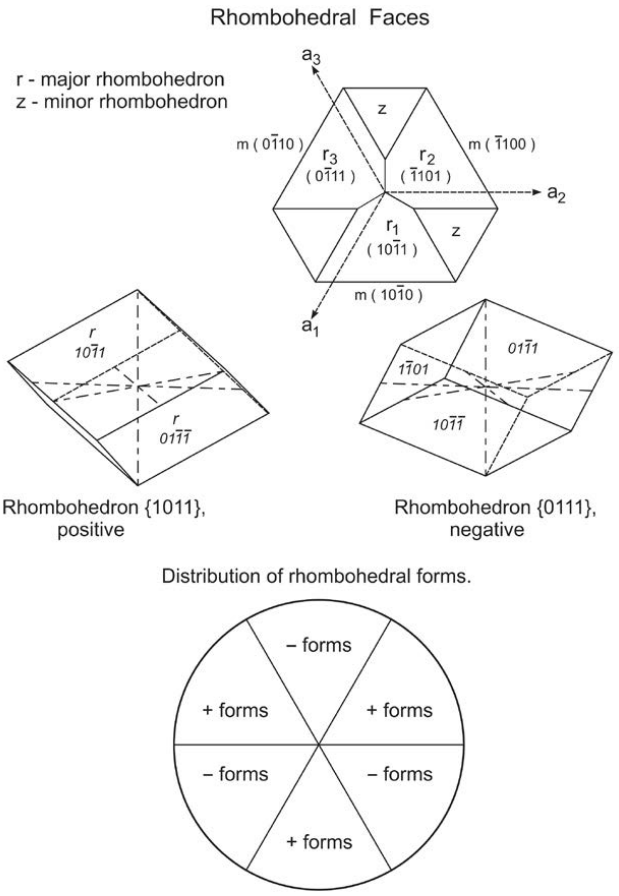


Figure 1. A typical amethyst crystal viewed perpendicular to the c-axis, illustrating the combination of positive $\{10\bar{1}1\}$ and negative $\{01\bar{1}1\}$ rhombohedra.

of their occurrence in α-quartz; however, in amethyst the twins are polysynthetic with a typical width of 0.1 mm (McLaren and Pitkethly 1982). The twin plane of the Brazil law is {101}, which separates right-handed and left-handed orientations of quartz. McLaren and Pitkethly (1982) demonstrated that the composition plane of the Brazil law twin provides space for incorporation of Fe^{3+} and that iron is preferentially concentrated along this composition plane in amethyst.

Amethyst's Name and Colour Origins

The word amethyst has its origins from the ancient Greek word *amethysto*s which may be translated as "not drunken", from the Greek *a-*, "not" + *methustos*, "intoxicated", as the gemstone was believed to prevent or lessen the effects of drinking alcohol.

There is a common theme regarding the mythological origin of amethyst's purple colouration. Bacchus (Dionysus to the Romans); the Greek god of wine-making, orchards, fruit, vegetation, fertility, festivity, insanity, ritual madness, religious ecstasy, and theatre,

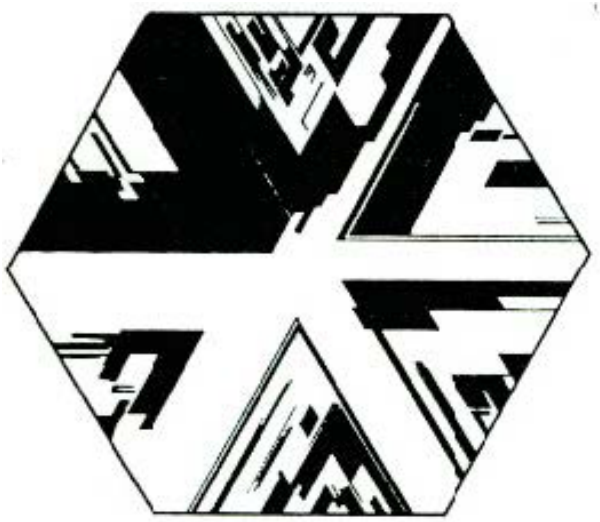


Figure 2. Etched basal α-quartz illustrating the typical occurrence of Brazil law twinning in which alternate bands contain left- and right-handed α-quartz (after Frondel, 1962).

pursuing a maiden named Amethyste, who was refusing his affections. Amethyste prayed to the gods to remain chaste, a prayer answered by the goddess Artemis (Diana to the Romans), who transformed her into a white stone. Bacchus humbled by Amethyste's desire to remain chaste, poured wine over the stone as an offering, dyeing the crystals purple.

In another variation the god was insulted by a mortal, and vowing to slay the next mortal who crossed his path in retaliation created fierce tigers to carry out his wrath. The hapless mortal a young woman, Amethystos, was on her way to the shrine of the goddess Diana, when the tigers fell upon her. Her life was spared by the goddess, but the price was being transformed into a statue of pure quartz. Seeing what his anger had done, a remorseful Dionysus was so moved that tears of wine poured from his eyes onto Amethystos, staining her stature purple.

Despite the belief in this origin story, there are no ancient texts supporting the myth, as compared to the ancient period that supposed birthed this story, it's quite recent as it was written in 1569 by the French Renaissance poet Rémi Belleau (1528–1577), in the poem "L'Amethyste, ou les Amours de Bacchus et d'Amethyste" (Amethyst or the loves of Bacchus and Amethyste; Belleau, 1576).

Amethyst Deposits in the Thunder Bay Area

In his summary of the history of amethyst in the Thunder Bay area, Patterson (1985) reported that as early as 1642, Radisson described the use of

"torquoise" as a gemstone by local indigenous peoples. Amethyst was an associated mineral in most of the lead-zinc and silver mines, and attracted the interest of a few prospectors. In the early 1860s, the McEachern brothers prospected for amethyst in the Thunder and Black Bay areas. In 1862, they mined two tons of amethyst crystals, which they barged to Toronto to sell in that city. About the same time, a shipment of amethyst from the Thunder Bay area was sold in Montreal. The success of the mineral as a valued item for sale even in an unprocessed state and the ease of mining encouraged other prospectors and developers to try searching for and producing amethyst (Garland 1994).In the 1880s, amethyst was mined northeast of Thunder Bay in a place now called Amethyst Harbour. Interest in Thunder Bay amethyst declined around the turn of the century with the development of deposits of high quality and inexpensive amethyst from Brazil.

The deposit that became known as the Amethyst Mine Panorama was originally discovered in 1935. When the fire tower was built in the 1950s, near Elbow Lake in McTavish Township, the large amethyst veins were uncovered by the roadbuilders. In the early 1960s, the area was staked, and trenches exposed the veins in what is now the open pit for the mine. In large vugs near the surface of the vein deposit, amethyst crystal of spectacular size were obtained. The development of the deposit with wide-spread sales and distribution of specimens revitalized interest in amethyst in the region (Sinkankas, 1976; Garland, 1994). The interest and activity in amethyst deposits in the Thunder Bay area led to the proclamation in 1975 designating amethyst as Ontario's provincial gemstone (Patterson 1985). A comprehensive report on amethyst deposits and mining activity in the Thunder Bay area was completed by Garland (1994).

The interest and activity in amethyst deposits in the Thunder Bay area led to the Mineral Emblem Act in 1975 designating amethyst as Ontario's provincial Mineral Emblem (Ontario, 1990; Patterson, 1985), with the 50th anniversary taking place in 2025.

There are currently 15 amethyst quarries authorized to produce under the Ontario Ministry of Nature Resources *Aggregate Resources Act* within two areas northeast of Thunder Bay (Campbell et al. 2024). Twelve of these authorized amethyst extraction sites are in McTavish Township and are accessible from Highway 11-17. The other three authorized quarries are located in the Tartan Lake Area (north of MacGregor Township) in an area that is accessed via the Magone Lake Road from Highway 527. Four quarries operate as tourist attractions that are open to the public on a seasonal basis. A listing of these amethyst quarries, including information about their products and services (where available), is provided in Table 1 (Campbell et al., 2024).

Geology Of Amethyst Mine Panorama (Thunder Bay Amethyst Mine)

Geologic Setting

The geological setting of the mine is complex, as an Archean and a Proterozoic record are preserved in the area. This record has been recently reviewed by Sutcliffe (1991) with an update by Addison et al. (2010) and will not be repeated in detail here. The Amethyst Mine Panorama (Thunder Bay Amethyst Mine) is hosted in the Archean Hilma Lake granite of McCrank et al. (1981). This pluton lies on the boundary of the Quetico Subprovince and the Wawa Subprovince, with typical greenstone lithologies on its southern margin and gneissic metasedimentary rocks on the northern margin. The Hilma Lake granite in the vicinity of the mine consists predominantly of monzonite, with compositional variation along the trend monzonite-quartz monzonite-granite-granodiorite and pegmatite and pegmatitic textural variants (Jennings, 1985). Jennings' study indicates that monzonite had been cut first by granodiorite, then by pegmatite, with some metasomatic alteration of early monzonite toward granodioritic composition.

At the Greenwich Lake uranium occurrence, a vein-type occurrence located 10 km to the northwest, Franklin (1978) noted the presence of quartz monzonitic pegmatites containing 60-100 ppm U in the form of uraninite. As these pegmatites are apparently comagmatic with the Hilma Lake granite, its uranium-rich character is likely a general feature. The Proterozoic rocks were deposited on the eroded Archean surface; however, the Animike Group (Gunflint and Rove Formations) is missing in the vicinity of the amethyst mine. As indicated by Franklin et al. (1980), the Mesoproterozoic Sibley Group progressively onlaps Archean terrain in a northerly direction. The Sibley Group is presently absent in the vicinity of the Amethyst Mine Panorama, although its presence as abundant fragments in mineralized breccias within the vein system indicates that these sediments

Table 1. Amethyst quarries in the Thunder Bay Area authorized to produce under the *Aggregate Resources Act* (from Campbell et al. 2024)

Deposit Name and Ownership	**Location (Licence)**	**Products and Services**
Amethyst Mine Panorama *Precious Purple Gemstones Ltd.*	McTavish Township (622921)	Tourist attraction (pick-your-own and mine tours), specimens, decorative and landscaping stone, and tumbling stone, jewellery, giftware, carvings, faceted gemstones www.amethystmine.com/
Blue Points Amethyst Mine *Jordan Vivian*	McTavish Township (624926)	Tourist attraction (pick-your-own), specimens, decorative stone, aquarium stone www.tripadvisor.ca/Attraction_Review-g155017-d3334892- Reviews-The_Blue_Point_Amethyst_Mine-Thunder_Bay_Thunder_Bay_District_Ontario.html (lynswan@lakenet.com – email)
Diamond Willow Amethyst Mine Big Pearl, Sward Lake *B. Leroux and C. Fayle*	McTavish Township *3 permitted quarries,* (626151, 625922, 626134)	Tourist attraction (pick-your-own and mine tours), specimens, decorative and landscaping stone, slabs, tumbling stone, jewellery and giftware www.diamondwillowamethyst.com/
Keetch Quarry / Boulder Creek Amethyst Quarry *L. Harasym*	McTavish Township (77956)	Tourist attraction (pick-your-own), specimens https://mininglifeonline.net/company_page_487.html
Assiniboia Amethyst Mine *P. and T. Smitham*	McTavish Township (626091)	Not open to the public, but may be visited by invitation only. Contact: https://assiniboiaamethystmine.weebly.com/
Bill's Old Amethyst Mine *K. Zytaruk*	McTavish Township (607322)	Not advertised
Canadian Shield Amethyst Mine *K. Zytaruk*	McTavish Township (616261)	Not advertised
Danbill Mine *Auralite 23 Mine and Company Inc.*	Tartan Lake Area (20227)	Specimens, polished and tumbled stone, jewellery, tiles and countertop stone www.auralite23canada.com/home.html
Gunnard Project *M. Noyes and J.A. Gavin*	McTavish Township (625989)	Not advertised
Loon Lake Technical Services Quarry *Loon Lake Technical Services*	McTavish Township (625067)	Not advertised
Purple Haze Mine *Auralite 23 Mine and Company Inc.*	Tartan Lake Area (624879)	Specimens, giftware, jewellery, decorative and landscaping stone from former owners at www.purplehazeamethyst.com/. Transferred to new ownership in late 2022.
Roll Lake Amethyst *Auralite 23 Mine and Company Inc.*	Tartan Lake Area (624838)	Specimens, polished and tumbled stone, jewellery, tiles and countertop stone www.auralite23canada.com/home.html
Windy Ridge Amethyst *L. Kowtuski*	McTavish Township (625831)	Email: windyridge@live.ca

were present as basement cover during the forming of the deposit.

The significance of the Sibley Group is unclear in the face of contradictory evidence concerning its age and depositional setting. Franklin et al. (1980) noted that the Sibley Group isdeposited at the location of a failed arm of an r-r-r triple junction, although they admitted to uncertainty as to the contemporaneity of sedimentation and rifting. Although some features of the Sibley Group are suggestive of a rift-filling deposit, the whole-rock Rb/Sr age of 1339±33 Ma (Franklin, 1978b) is approximately 200 Ma prior to the main stage of rifting of the Midcontinent (Keweenawan) Rift (Van Schmus et al., 1982). Cheadle (1986), however, concluded on the basis of sedimentological studies that the Sibley Group was not deposited in a classical aulocogen, but represents a deposit on a sagging crust preceding rifting. The Sibley Group was more recently dated by U/Pb geochronology in zircons in a basal rhyolite unit at 1537+10/-2 Ma (Davis and Sutcliffe, 1985). This timing makes a relationship with the Midcontinent Rift event unlikely, and Hollings et al. (2004) proposed that the Sibley Basin formed due to effects of a plume track that created an infracratonic basin.

Other deposits located at or near the Sibley -Archean unconformity include the Dorion lead -zinc -barite veins (Fig. 3). The ore-depositing solution was considered to be a basinal, connate brine by Franklin and Mitchell (1977), an interpretation supported by the fluid-inclusion studies of Haynes (1988). As illustrated in Figure 3, there is a close spatial relationship between the lead-zinc-barite veins and the amethyst, and both are spatially related to the Sibley-Archean unconformity.

Geological features of the mine

Amethyst Mine Panorama is located within a first-order strike-slip fault, which strikes at 90- 100° and dips steeply to the south. This fault is roughly parallel to one 2.1 krn to the south, which strikes east-northeasterly (McIlwaine, 1971) and has a vertical displacement of at least 125 m (Jennings, 1985), forming a major boundary to the Sibley Group's depositional basin. The strike-slip fault hosting the amethyst deposit is offset by seven first-order strike-slip faults, five of which are illustrated in Figure 4, which strike 162 - 150° and dip vertically producing en echelon, pull-apart structures in the main fault. These structures are filled by breccias of granitic country rock and Sibley Group sedimentary rocks with large proportions of void space. The brecciated fault was subsequently mineralized by hydrothermal solutions. At least two periods of mineralization occurred, as an early generation of amethyst was clearly brecciated and subsequently coated by a second generation of amethyst.

Figure 4 is an illustration of the state of the mine in 1987. At present, the main pit configuration is basically the same but has been deepened. In that year, an extension of the vein system to the east was developed, offset to the north by a few metres strike-slip fault. Jennings (1985) subdivided the mineralization patterns into three basic types: (i) open fracture fillings, (ii) breccias with tectonic and collapse subtypes, and (iii) "honeycomb" veins.

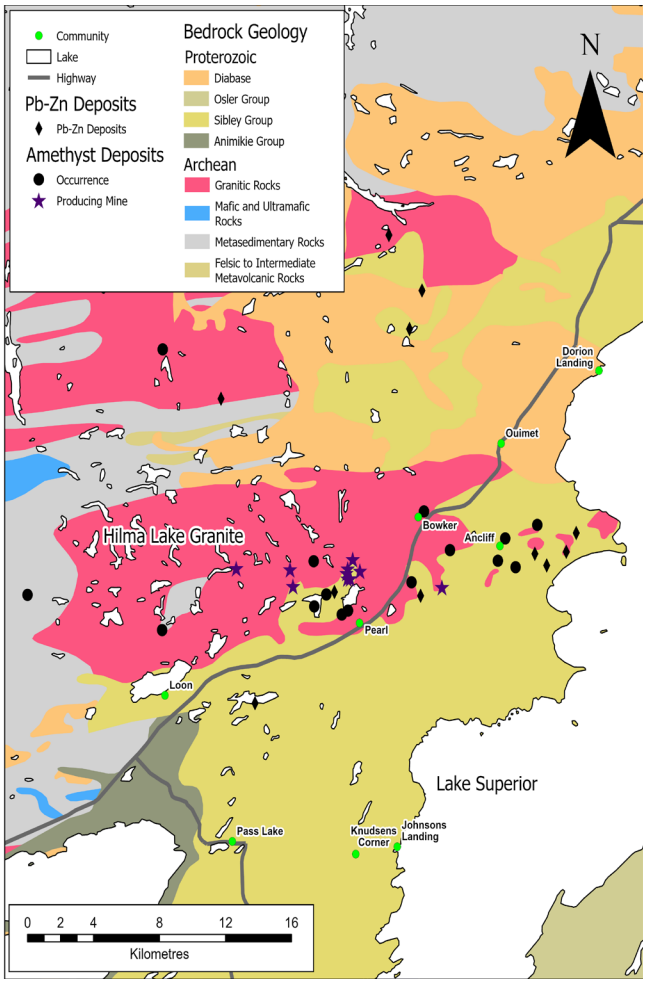


Figure 3. Local geology and location map of amethyst deposits and lead-zinc-barite deposits, showing the relationship of the former to the margin of the Sibley Group outcrop and the Hilma Lake granite. The location of the producing Thunder Bay Amethyst Mine's are indicated by stars. Bedrock geology and mineral occurrence locations modified from Ontario Geological Survey (2011; 2026).

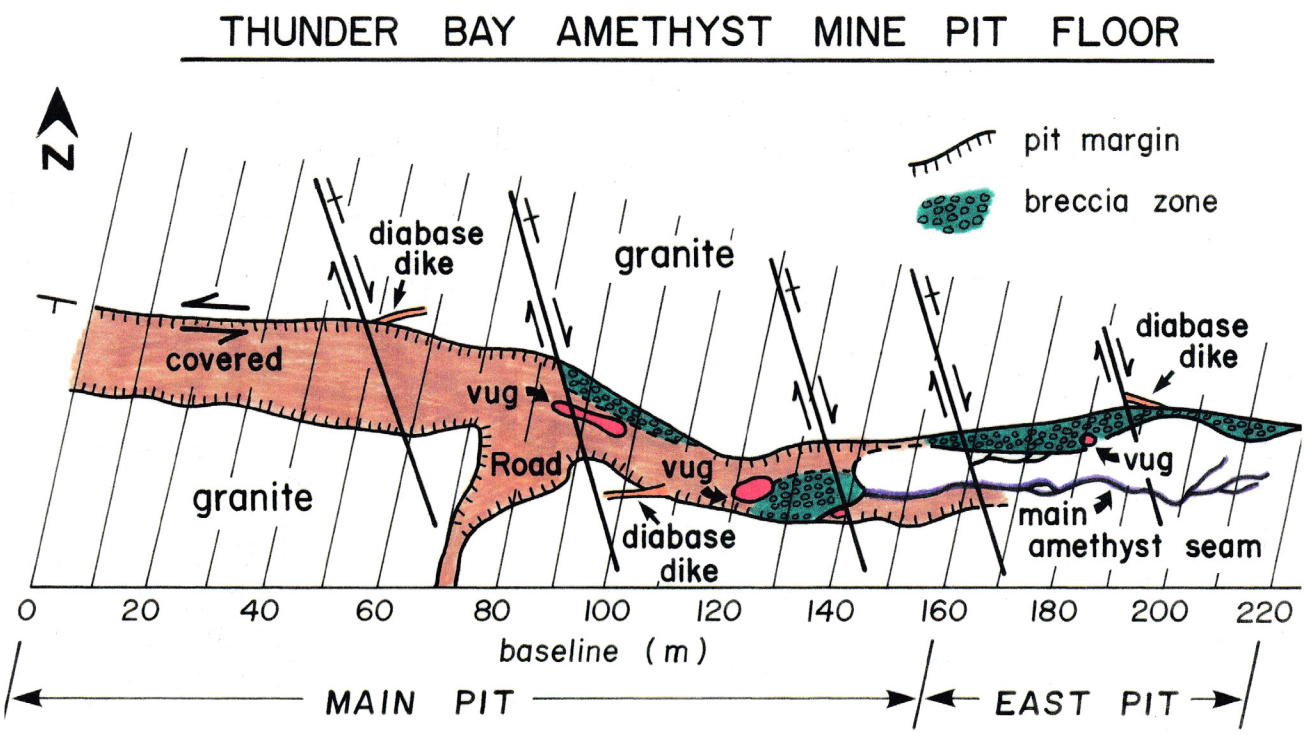


Figure 4. Diagram of the main pit of Amethyst Mine Panorama (Thunder Bay Amethyst Mine).

The strike directions of these veins are strongly clustered in two groups, one slightly west of north and parallel to the second stage of strike-slip faulting, and one easterly, parallel to the principal directions of faulting.

Open fracture fillings are common in the shallower zones of the deposit where low lithostatic pressure permitted the maintenance of open fissures following faulting. The veins are widest near the edges of collapsed breccias and at the intersections of oblique shears with the main fault zone. Fracture-fill mineralization occurred at the crystal-fluid interface as quartz crystals grew outward from the fracture walls. The crystals formed as parallel to radial growths with long crystal axes oriented perpendicular or subperpendicular to the growth surface. The crystal size invariably increases outward, and outward growth from opposite fractures resulted in an interlocking comb structure of euhedrally terminated crystals. This vein type may also contain vugs up to 2-3 m in diameter with large quartz crystals up to 10-15 cm in prism diameter.

Tectonic breccias are here attributed to fault movement, as opposed to brecciation caused by collapse with variable degrees of fluid action. Some breccia fragments are surrounded only by a later portion of the paragenetic sequence, suggesting that multiple fault motion during the mineralizing event has occurred. Breccia fragments of this type are invariably angular and may consist of fragments of earlier deposited vein material, which may have been thermally bleached.

Collapse brecciation is not always differentiated from tectonic brecciation, and some collapse breccias have undergone subsequent tectonic brecciation and vice versa. Evidence of collapse brecciation is seen in the occurrence of Sibley Group lithologies not present in the mine area now, together with granite and diabase as breccia fragments. Sibley Group fragments are particularly abundant within channel- or pipe-like structures in which fluid transport and abrasion have produced subangular to subrounded fragments, which have undergone an appreciable degree of sorting. Collapse-breccia fragments are typically coated with successive layers of chalcedony, colorless quartz, and amethyst, producing a cockade structure. Vugs have developed in open space produced in the breccia in which crystals with prism diameters of up to 10 cm have grown. Honeycomb veins are the result of quartz crystallization that has occurred in all directions from small nuclei, usually chalcedony, hematite, or silicified granite fragments, rather than from a fracture wall. The amethyst and quartz are more massive than in the other types of veins, but the growth is chaotic.

Mineralogy

Amethyst and other varieties of quartz.

Several varieties of quartz occur in Amethyst Mine Panorama, including colorless quartz; chalcedony; amethyst; the yellowish variety, citrine; and the greenish variety, prasiolite or "greened amethyst". Smoky quartz has very limited development . The only variety of gemstone interest is amethyst, although the occurrence of the other varieties has aided in establishing the sequence of deposition. A grading system based on estimated intensity of coloration and clarity of specimens is in use at the mine, and this system has also aided in establishing the paragenetic sequence. Thus, the intensity of coloration may be from I (lightest) to IV (darkest) and clarity from ***a*** (clear) to *f* (opaque). Table 2 lists typical paragenetic sequences in an older sequence, which is present as breccia fragments in a younger sequence presently occupying the veins. The prasiolite in stages 4 and 5 of the older sequence appears to be thermally bleached amethyst on the basis of both its appearance and experimental evidence that heat-treated amethyst can be transformed to prasiolite (Lehmann and Bambauer, 1973).

In the younger sequence, late-stage variations are noted, particularly as cappings to stage 5. A distinctive variety called "*black gem*", a dark, brownish-black amethyst, is apparently characteristic of larger crystals grown in vugs in which iron-enriched, late-stage fluids were trapped. These frequently have final growth zone that contains abundant hematite inclusions, such that recent sales of such material has been called "Thunder Bay red". It was this material, recovered in the early development of the deposit that led to the notorious statement by Sinkankas (1976, p. 204): "By far most of the amethyst is unsuited for lapidary purposes, with very little being free from flaws and hence useless for faceted gems or even baroques." Figure 5 illustrating cut and faceted gemstone demonstrates the error in Sinkankas' statement.

The compositions of specimens of amethyst by neutron activation analysis for selected trace elements (Table 3) revealed the presence of subequal concentrations of Fe and Al. As well, low concentrations of Ge were sought based on absorption spectra that indicated its presence. The low Ti concentrations are perhaps related to the spotty occurrence of rutile needles in the amethyst, needles occurring when concentrations are relatively greater.

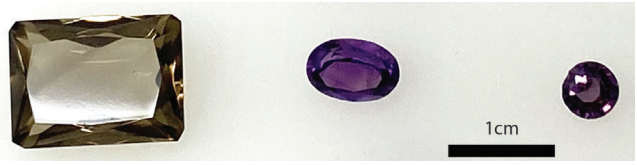


Figure 5. Cut and faceted smoky quartz (top left) and amethyst from Amethyst Mine Panorama (Thunder Bay Amethyst Mine). Photo by S. Kissin.

Table 2. Paragenetic sequences observed in the veins of Amethyst Mine Panorama

Stage	Type	Thickness (cm)
Older sequence (observed as breccia fragments)		
1	Chalcedony	0.1 - 0.3
2	Colorless quartz	0.8 - 1.0
3	Chalcedony	0.1 - 0.2
4	Prasiolite	2.0 - 3.0
5	Prasiolite	2.0 - 7.0
Younger sequence (main vein deposits)		
1	Chalcedony and hematite	0.1 - 0.3
2	Colorless to smoky quartz	0.5 - 1.0
2a	Amethyst I (top of stage 2)	0.5 - 0.8
3	Amethyst III - IV	2.0 - 3.0
4	Amethyst II - III	2.0 - 12.0
5	Amethyst II - IV	2.0 - 14.0
6	Black gem	Up to 4.0

Notes: Variations observed include (i) late-stage greenish and yellowish-amethyst; (ii) late-stage smoky quartz; (iii) discontinuous hematitic and milky quartz capping to crystal terminations; and (iv) development of black gem in crystals, deposited in vugs.

Table 3. Analyses of r-zones of amethyst for selected trace elements (in ppm; Kissin, 1997)

Sample No.	Fe	Al	Ge	Ti
DZS1*	217	447	0.5	n.d.
LZS2	102	393	0.5	0.01
BSS3	273	369	1.0	n.d.
TPS4	368	249	0.5	n.d.

**DZS1 evidently contains solid inclusions, as high concentrations (in ppm) were noted; e.g. Ta 0.329, W 0.38, Eu 0.349, Sr 89.43, Zr 1.02, Nb 0.13, Ba 2985.26, La 3.85, Ce 0.35, U 0.387. All samples contain small, but detectable quantities of Co, Ni, Ga, Rb, Nb, Zr, Mo, Sn, Sb, Cs, La, Pr, Nd and U.*

Other non-sulfide minerals.

Barite is rare in the veins at the Amethyst Mine Panorama, although it is abundant in other amethyst mines of the district, where it follows the final stage of quartz deposition. It was not observed in the course of the present study, but has been noted in the mine.

Calcite is fairly common in thin, monomineralic veins, but was not observed within the amethyst-bearing veins. The genetic link between the calcite veins and amethyst veins, if any, is unclear. Hematite is abundant as minute- (< 0.1 mm diam.) solid inclusions in stage 5 of amethyst deposition and occurs sporadically at other stages of deposition as well. Hematite occasionally occurs as a daughter mineral in fluid inclusions, particularly in stage 5 of crystallization. Rutile occurs as needles that transect the growth zones of the quartz in scattered locations within the mine. The orientations of the needles are apparently random; however, the possibility of crystallographically controlled growth directions has not been considered in detail. Native copper occurs in association with copper and copper-iron sulfides.

Sulfides

The common base-metal sulfides pyrite, chalcopyrite, galena, and sphalerite occur in small amounts throughout the vein succession and as veinlets and replacement bodies in altered granitic wall rock. Copper -iron sulfides, however, are predominant, and a sequence of the minerals cuprite-native copper-chalcocite-covellite associated with hematite and pyrite was documented by McArthur et al. (1993). The copper -iron sulfides exhibit typical replacement textures (atoll structures, core-and-rim relationships) in occurrences both in amethyst growth stages and in wall rock. The assemblages bornite+pyrite and chalcopyrite+pyrite and chalcopyrite+pyrite occur in wall rock only; however, spatial relations of wallrock sulfides to the veins do not reveal any pattern, owing in part to their scarcity. Malachite is present as a supergene product derived from these hypogene copper minerals.

Wall-rock alteration mineralogy.

Hematitization, chloritization, and kaolinitization are prominent in envelopes surrounding the veins within zones of brecciated granite; however, the alteration extends only a few centimetres into the granites outside of the zone of brecciation. Intense hematitization occurs fairly generally in altered rock nearest the amethyst veins. The strongly hematitized zone is generally only a few centimetres thick, but weaker hematitization is notable throughout the altered zone. Outward from the hematized zone is an irregular zone of highly chloritized rock ranging from a few to a few tens of centimetres thick. Sometimes associated with the chloritization is diffuse epidotization, which produced a pistachio green tint over zones up to a metre wide.

Kaolinitization is widespread and pervasive through the breccia zone, imparting a chalky, white appearance to relict feldspars. Other clay minerals, e.g., montmorillonite and illite, may also be present; however, they have not been sought in a detailed examination. The pervasive kaolinitization has allowed weathering to penetrate into the brecciated zone, resulting in a soft and loosely aggregated matrix in which the near-surface exposures of the amethyst are contained. The nature of this matrix has enabled a good deal of the amethyst to be mined with a minimum of blasting. The hematite-chlorite-epidote alteration assemblages in the presence of ubiquitous quartz are characteristic of the propylitic alteration typical in many hydrothermal ore deposits. The kaolinite and other clay minerals are characteristic of the argillic

alteration of hydrothermal ore deposits. The two alteration types are analogous at least in their relative timing to early peripheral propylitic alteration, which is overprinted by argillic alteration stemming from downward-infiltrating meteoric water.

Genesis of the deposit

The genesis of the deposits of the Amethyst Mine Panorama were discussed in detail by McArthur et al. (1993) in the light of evidence obtained in their study. The conclusions of their study are given below; however, for details of the evidence, their paper should be consulted. Genetic speculations on the Amethyst Mine Panorama are hampered at the outset by questions as to the timing of amethyst deposition, as discussed in the Introduction. The spatial and geochemical affinities of the amethyst deposits with the Dorion lead-zinc-barite veins and the relationships of both to the depositional margin of the Sibley Group sediments suggest that all three are interrelated. Franklin and Mitchell (1977) proposed that the lead -zinc -barite veins formed when, during diagenesis and settling of the Sibley Group sediments, metal-bearing brines were formed when expelled connate waters mobilized metals from the Sibley Group sediments and(or) weathered granitic basement rocks below the Archean-Proterozoic unconformity. The solutions thus formed would have hypothetically migrated through the basal Pass Lake Formation aquifer to escape at basin-marginal faults. Precipitation of sulfide, carried in chloride- and sulfate-bearing solution, occurred because of mixing of the relatively oxidized solution with H_2S gas trapped at the Pass Lake Formation pinch-out.

The amethyst deposits seem to be a variant of the lead-zinc-barite type of deposit in which the temperature was lower than that of the sulfide-rich lead-zinc-barite veins. The initially oxidizing to later reducing character of the solution is similar to that proposed for the lead-zinc-barite veins, but the relation to Pass Lake Formation pinch-outs is not present in most amethyst deposits. Rather, the amethyst deposits are generally hosted in granitic basement often with no Sibley Group sediments present. The amethyst deposits are richer in dissolved silica, having gained this component through the kaolinitization of feldspar during hydrothermal alteration of granitic country rock. As the amethyst deposits formed near the present or former unconformity with the Sibley Group, local reduction of the solution would have tended to occur as H_2S was released during thermal breakdown of organic matter in the sediments. The quantity of sulfides precipitated would have been limited not only by the relatively small amount of H_2S produced but also by the lower metal content of the solutions as compared with those depositing the Dorion lead-zinc-barite veins.

The latter characteristic is inferred by a comparison of the results of this study with those of Haynes (1988) on the Dorion lead-zinc-barite veins. He found that fluid inclusions from these deposits are $NaC1$-$CaC1_2$-H_2O type on the basis of microthermometry and direct analysis of decrepitates. However, the fluid inclusions depositing sulfides are significantly more saline than those at the Amethyst Mine Panorama in that they contain daughter salts. The more saline and higher temperature (105-203°C) fluid inclusions indicate that solutions that they represent would have had a better metal carrying capacity as chloride complexes. The similarity of the solution components to those at Amethyst Mine Panorama lends support to the idea that the same event formed both types of deposits. The solutions depositing amethyst would have been cooler and less saline variants of those that formed the lead-zinc-barite veins. If the two types of deposit are genetically linked, both suffer from the problem of lack of knowledge of the timing of ore deposition. The maximum age of both is 1339 Ma, the whole rock Rb/Sr age of the Sibley Group (Franklin, 1978b), as both types of veins cut Sibley Group rocks and contain breccia fragments of them. Franklin and Mitchell (1977) did not suggest a specific timing for formation of the Dorion lead-zinc-barite veins; however, their suggested mechanisms for creation of the deposit favor a timing soon after the deposition of the Sibley Group sediments. The expulsion of pore water called upon would presumably occur during late diagenesis. However, as there is no evidence to suggest that the Sibley Group sediments have ever been deeply buried, the source of heat is a problem. If the timing of deposition were close to the formation of the Sibley depositional basin, it is possible that a thermal anomaly, perhaps augmented by seismic pumping, in the lower crust was responsible for both phenomena.

Haynes (1988) suggested that the Dorion lead-zinc-barite veins formed either in the environment of Keweenawan rifting or later, possibly in the Paleozoic. There is no geological evidence for activity in the Paleozoic in the western Lake Superior region, and the style of mineralization associated with Keweenawan

events is different (silver deposits associated in part with Ni-Co arsenides; Franklin et al. 1986). Our preferred hypothesis is that the lead-zinc-barite veins and amethyst veins are associated with the timing of formation of and deposition in the Sibley basin. We, therefore, believe that these deposits are distinct from silver deposits in the Thunder Bay area and formed at a somewhat earlier time. The timing is, however, not at all certain. An attempt was made to directly date amethyst deposition by U/Pb age determinations on rutile needle inclusions in amethyst (Heaman and Easton, 2006). An age of 887±40 Ma with 68.6% discordance was determined on a very small sample with low uranium content. The authors indicated that these results should be viewed with caution as a large lead correction was needed. This age does not coincide with any known geological events in the area. As well, an attempt was made to date the cross-cutting diabase dikes, but was unable to yield any results.

Summary

Field and laboratory studies of the Amethyst Mine Panorama reveal the following:

(1) The vein system hosting amethyst deposits was formed by mineralization of an east-west-striking, steeply dipping strike-slip fault, opened into en echelon pull-apart structures by a series of later strike-slip faults, also dipping steeply and intersecting the first-formed fault at high angles. Much open space with brecciated and vuggy textures resulted. Breccia fragments include granitic host rock and Sibley Group sedimentary rocks, implying that the latter were present as a thin cover at the time of mineralization, although they are erosionally removed from the mine area at present. At least one early generation of amethyst is included as breccia fragments, indicating that fault movement continued during mineralization.

(2) At least two phases of amethyst crystallization separated by a period of brecciation are present. The older sequence contains five stages of quartz growth, the latter two of which were originally amethyst, but were thermally bleached to prasiolite by the influx of hot solutions that deposited the younger sequence of quartz. The younger sequence contains five and occasionally six stages of deposition, beginning with a stage of chalcedony and a stage of colorless quartz, followed by amethyst. Both sequences of deposition are traceable throughout the mine.

(3) Sulfide minerals including pyrite, chalcopyrite, galena, and sphalerite accompany amethyst deposition as small mineral inclusions and occur, as well, as veinlets and replacement bodies in altered granitic wall rock. Copper and copper-iron sulfides are most abundant and, together with native copper and cuprite, Eh-pH relationships indicate that the solutions forming the deposit were initially rather oxidizing and weakly acidic. In the course of crystallization, the solution became more reducing and slightly more acidic.

(4) Fluid-inclusion studies indicate that in the younger sequence of quartz deposition, homogenization temperatures range from 146.5 to 114.7°C (mean 132.1°C) as contrasted with 91.2-40.9°C (mean 68.4°C) for amethyst. Eutectic temperatures of frozen inclusions indicate that the solution was of the NaCl-CaCl-H_2O system, with possible concentration of an additional halide salt component in late-stage fluids. Few inclusions contain daughter minerals, and those found are hematite and sphalerite in late-stage fluids. Final melting temperatures indicate a trend of decreasing salinity in later growth stages.

(5) Oxygen isotopic determinations on quartz indicate a range of $\delta^{18}0$ outside that of juvenile waters and end-member basinal brines. Progressive mixing of basinal brine with local meteoric water is suggested.

(6) Sulfur isotopic analyses of pyrite yield $\delta^{34}S$ of -0.4 to 0.6 ‰ and -1.4 ‰ in chalcopyrite. These volumes are consistent with derivation from H_2S gas liberated by thermal action protection on organic material involving iron. The values are similar to those of the sulfur contained in sulfides in the Dorion lead-zinc-barite veins.

(7) The presence Sibley breccia fragments cemented by quartz indicates that the veins cannot be older than 1339 Ma, the Rb/Sr age of the unit. However, a younger limit cannot be established at present.

(8) On grounds of similarity in geological setting, proximity, composition of the ore-depositing solution, and sulfur isotopic composition, the amethyst veins are believed to be genetically related to the Dorion lead-zinc-barite veins. Both are believed to have been formed by solutions expelled and mobilized during diagenesis and compaction of the Sibley Group. The lead-zinc-barite veins formed in fractures at or near the margin of the Sibley depositional basin from solutions that were both hotter and more saline than those

depositing amethyst. Amethyst-depositing solutions travelled longer distances in granitic basement, dissolving silica by alteration of feldspar. Although the amethyst-depositing solutions probably carried less metal as chloride complexes than did the solutions forming the lead-zinc-barite veins, less H_2S at the site of deposition was probably the most significant factor causing a low sulfide content in the amethyst veins.

(9) The temperature conditions under which amethyst forms appear to have a high temperature limit; at the Amethyst Mine Panorama this limit is no higher than approximately 115°C and may be as low as approximately 90°C. Temperatures as high as approximately 145°C but possibly as low as 115°C may be sufficient to thermally bleach earlier generations of amethyst in the influx of hot solutions. However, this theory of thermal bleaching has been recently criticized by Herbert and Rossman (2008), who attributed the development of greenish-grey to greenish quartz to the presence of H_2O in the crystal. Our work (Klarner and Kissin, 2011) confirms the presence of water in IR absorption spectra; however, the water is largely contained in fluid inclusions, which are abundant and of secondary origin. Use of the highly focus beam of an FTIR microscope has shown that molecular water is of low and nearly identical concentration in both amethyst and "greened amethyst". Experiments by Goetz (2014) demonstrated that heating at 250ºC for extended periods did not result in bleaching of amethyst, disproving that the 145ºC temperature caused bleaching of amethyst. This problem is unresolved at present.

Geology of the Diamond Willow Amethyst Mine

Unlike the years of extensive research undertaken at Amethyst Mine Panorama, the other known amethyst deposits in the Thunder Bay region are not well studied and with most information available being from Garland (1994)

The original Diamond Willow Amethyst Mine was staked by Gunnard Noyes, in the 1960s with the mine initially operating in the 1970s. Following Gunnard's passing in 1988, the mining leases were split into two parcels per inheritance and turning the original mine into the current Diamond Willow and Blue Points Amethyst Mines, owned by a son and daughter, respectively. The Diamond Willow Amethyst Mine operated until 2007 and was subsequently closed until 2015 (Garland, 1994).

The currently producing Blue Points Amethyst Mine is the eastern extension of the original Diamond Willow mine and operates two and three-quarters of the four pits situated along the breccia zone (Fig. 6). The centre pit is the original and the largest, almost 60 m long and 4 m deep. A fence divides the pit between the two mines. The current Diamond Willow Amethyst Mine; the western extension of the original namesake mine site, operates one and one quarter of the four pits situated along this breccia zone (Fig. 7; Garland, 1994).

This mineralized and well developed breccia zone occupies a vertically dipping fault zone, trending approximately 090° and extends for almost a kilometre. The fault separates Sibley Group conglomerates of the Pass Lake Formation from Sibley Group mudstone of the Rossport Formation (Garland, 1994).

The current Diamond Willow Amethyst Mine is located at the western end of this fault/breccia zone, separating the Rossport Formation mudstones on the south from the Pass Lake Formation conglomerates on the north side. Both the mudstone and the conglomerate are well-layered, giving them a blocky appearance (Garland, 1994).

The breccia zone varies from 1 to 5 m wide, and is characterized by a quartz rich core and fragments of wall-rock material. In general, the fragment density increases away from the core, but is always matrix supported, the fragments are angular and representative of the wall rocks.

Within the breccia, amethyst filled vugs can attain sizes of over 1 m and are lined with large, dark purple crystal points up to approximately 7.5 cm in diameter. The vugs also tend to be filled with a dense red clay; fault gouge, consisting of finely ground quartz, feldspar, chlorite, and biotite (Vos, 1982; Patterson, 1985; Garland, 1994). which must be removed in order to mine the amethyst.

Light violet to a very dark, nearly black purple amethyst forms an extensive druse covering along the south wall, crystallizing between the mudstone and the breccia. The crystal points in this druse tend to be small, but are very well-formed, yielding excellent mineral specimens. Like Amethyst Mine Panorama, the amethyst crystals are sometimes coated with a layer of reddish brown hematite. Galena occurs as seams of crystals 1 cm in size, within the quartz at the west end of the exposed breccia zone, chalcopyrite-rich zones

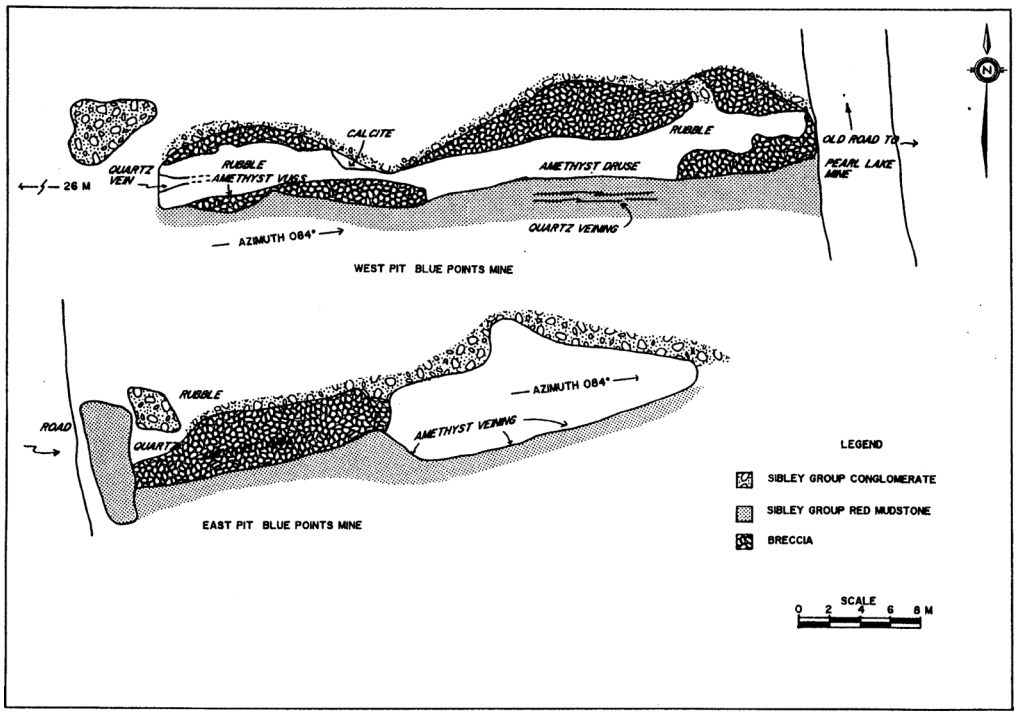


Figure 6. Plan view of the Blue Points amethyst mine.

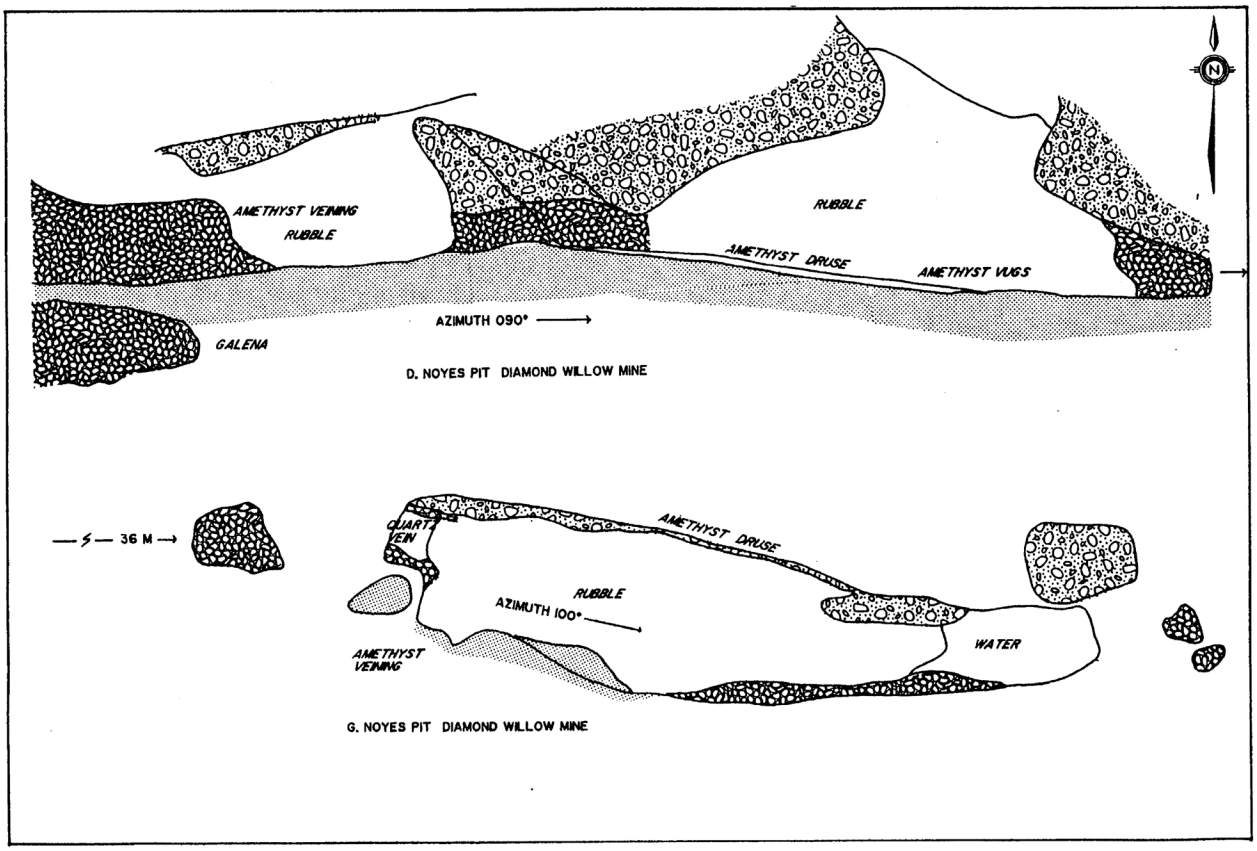


Figure 7. Plan view of the Diamond Willow amethyst mine, refer to Figure 6 for legend.

are associated with rusty stained or clear quartz crystals (Vos, 182; Garland, 1994).

Road Log Lakehead University to Amethyst Mine Panorama and Diamond Willow Amethyst Mine

Leaving Lakehead University, we will follow the portion of the Trans-Canada Highway 11-17 which is the Thunder Bay Expressway. The flat terrain is the remnant of the bottom of the Nipissing stage of ancestral Lake Superior, and proceeding northeasterly, we pass upward through strandlines of the receding Pleistocene lake.

Lakehead University itself is underlain by the Gunflint Formation at or near the top of the unit. Shaly rocks near the top of the formation are exposed in the bed of the McIntyre River that flows through the campus; however, in recent years blocks of rock containing the Sudbury ejecta debrisite were excavated during construction of new student residences. These placed in various places around the campus as ornamentation or barriers to vehicular traffic.

Continuing, outcrops of a Logan sill diabase are exposed on the left side of the expressway. These sills form the caps of the prominent mesas south of town and underlie the high ground in the northern section of Thunder Bay, formerly the city of Port Arthur. Passing the junction of Red River Road (Highway 102), the expressway is on a level stretch marking the top of a Logan sill.

The expressway then passes downhill to the Current River. In proceeding downhill outcrops of Logan sill diabase, Gunflint shale and Gunflint carbonate are successively exposed. The carbonate is ankeritic and is oxidized to yellowish orange. Climbing uphill from the Current River bridge, the highway is again cutting into diabase sill. A fault trends along the highway offsetting the sill on opposite sides of the highway. A few hundred metres farther along the highway, the sill is dropped downward by a fault trending perpendicularly to the highway.

Recent work has shown that this sill, known locally as the Terry Fox sill, is a Nipigon sill (Magnus and Kissin 2010). Nipigon sills, which occur from here northeasterly to the Lake Nipigon area, are somewhat younger than Logan sills and can be distinguished on the basis of their trace element composition. Proceeding downhill and rounding a curve to the left, there is a high bluff on the left capped by a prominent diabase sill. The sill has intruded the top of the Gunflint Formation and the overlying Sudbury debrisite layer, which is capped by a thin remnant of Rove Formation shale. This is the only outcrop known in the Thunder Bay area that contains the complete debrisite layer. The east end of this outcrop is bounded by a fault that dropped down the section.

Continuing onward, high ground on both sides of the highway are capped by sills; the sill on the right was extensively quarried for railway bed ballast and large stone for construction of the breakwall in the harbor. After the junction with Highway 527, the highway climbs the hill locally known as KOA hill. Prior to a widening of the highway about a decade ago, the angular unconformity between the Gunflint Formation and steeply dipping Archean metavolcanics was exposed on the left of the highway. The hill is formed by the outcrop of the Mackenzie granite, an unmetamorphosed and undeformed, late Archean pluton. The highway continues on top the of granite, which contains occasional roof pendants of Archean metavolcanics.

After crossing the Mackenzie River sparse outcrops of granite are replaced by poorly exposed Gunflint Formation until just past the junction with Highway 587. Here, well-bedded red-stained carbonates of the Gunflint Formation crop out beside the highway. Passing onward to the East Loon Road, turn left onto the road, then right on Bass Lake Road. Continue to the turn off on the right to the private road to the mine. Proceeding along the mine road, it climbs steeply up from the Sibley basin onto the Archean Hilma Lake granite, ascending along a border fault surface.

At the top of the grade, there is a chance to view Lake Superior with Black Bay, the Black Bay Peninsula and the Sibley Peninsula, clear weather permitting. A few more kilometres brings the road to the mine.

To get to the Diamond Willow Mine, head back to Highway 11-17, and turn east towards Nipigon. Travel for approximately 13.4 km and turn left onto 5 Rd S, then make a right and drive to 5 Rd N,, crossing the railbed and make a left onto a private dirt road. Continue on this road staying right for 2.56 km, until a "Y" junction is reached and stay left until you reach the Diamond Willow Amethyst Mine parking area.

Amethyst Mine Tours

Note: Safety boots or shoes recommended. No sandals or open-toed shoes.

The tours will pass through the operating mining areas, which is not available to ordinary tourists. No collecting is allowed in these areas After visiting the mining area, there will be an opportunity to look for specimens in a designated collecting area. The charge for specimens is by weight. Hammering or chiseling is not permitted in Amethyst Mine Panorama's collecting area, hammering and chiseling are only permitted at Diamond Willow Amethyst Mine; however, only hammers up to 2 lb. max, are permitted and absolutely no sledge hammers and safety glasses must be worn when using hammers or tools while collecting. Specimens are also for sale in the shops.

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