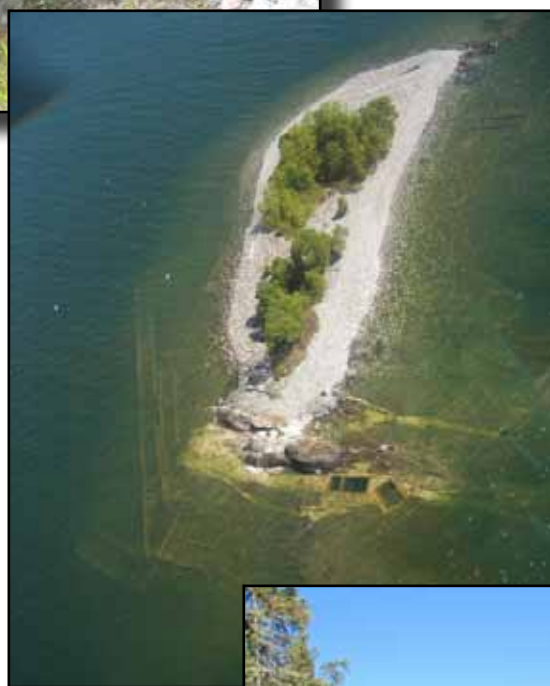


58th ANNUAL MEETING INSTITUTE ON LAKE SUPERIOR GEOLOGY

Thunder Bay, Ontario - May 16-20, 2012

Part 2 – Field Trip Guidebook



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58th ANNUAL MEETING

INSTITUTE ON LAKE SUPERIOR GEOLOGY



May 16-20, 2012

Thunder Bay, Ontario

HOSTED BY:

Pete Hollings

Chair

Lakehead University

Proceedings - Volume 58

Part 2 – Field Trip Guidebook

Edited by Pete Hollings, Al MacTavish and Bill Addison

Cover photos: Top - Neoarchean conglomerate in the Max Lake area, Hwy 527, Wabigoon Subprovince, Middle - Silver Islet Mine, Lake Superior, Right - Inspiration diabase sills, Chimney Lake near Armstrong (all photos courtesy of Mark Smyk).

58th INSTITUTE ON LAKE SUPERIOR GEOLOGY

VOLUME 58 CONSISTS OF:

PART 1: PROGRAM AND ABSTRACTS

PART 2: FIELD TRIP GUIDEBOOK

TRIP 1 & 13: SUDBURY IMPACTOCLASTIC DEBRISITES AT THUNDER BAY

TRIP 2: GEOLOGY OF THE SIBLEY PENINSULA

TRIP 3: LAC DES ILES MINE

TRIP 4: SHEBANDOWAN MINE AREA

TRIP 5: GEOLOGY OF THE THUNDER BAY AREA

TRIP 6: THUNDER BAY AMETHYST MINE

TRIP 7: BUILDING STONE TOUR OF DOWNTOWN PORT ARTHUR, THUNDER BAY

TRIP 8: HIGHWAY 527 TRANSECT

TRIP 9: REHABILITATION OF THE PAST-PRODUCING SHEBANDOWAN AND NORTH
COLDSTREAM MINE SITES

TRIP 10: GEOARCHAEOLOGY OF THUNDER BAY

TRIP 11: MIDCONTINENT RIFT INTRUSIONS

TRIP 12: MUSSELWHITE MINE

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

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Introduction, safety considerations and acknowledgements

Pete Hollings

Department of Geology, Lakehead University, Thunder Bay, Ontario, P7B 5E1, Canada

This volume is intended to serve not only as a guide for 58th ILSG field trip participants but also as a reference for those planning to revisit these areas at a later date. Consequently we have included UTM coordinates in the NAD 83 datum for the majority of stops, as well as instructions on how to reach them. For some of the stops on private land we have withheld the UTM coordinates to respect the privacy of the property owner. As some of the stops are on private and staked land, please be sure to obtain the land owners' permission before entering their land. For up-to-date information on land ownership please contact the Thunder Bay Resident Geologists' Office (807 475 1331). Sample collection is prohibited at some stops on private land or in Provincial Parks.

Many of the fieldtrips will be visiting stops along either major highways or busy logging roads. Please

take care when crossing or parking along these roads. For those field trips that are visiting active mine sites personal protective equipment will be required. Please notify the field trip leaders if you have any medical conditions that may be of concern during the trip. Each trip leader is equipped with a first aid kit and satellite/cell phone, so please notify them of any incident.

We would like to thank all authors who contributed to this field guide and also all those who provided comments and assisted with the running of the field trips themselves. We appreciate the assistance and cooperation of the exploration and mining companies in providing us access and information concerning their properties. We are particularly grateful to the Musselwhite and Lac des Iles mines for running field trips on their properties.

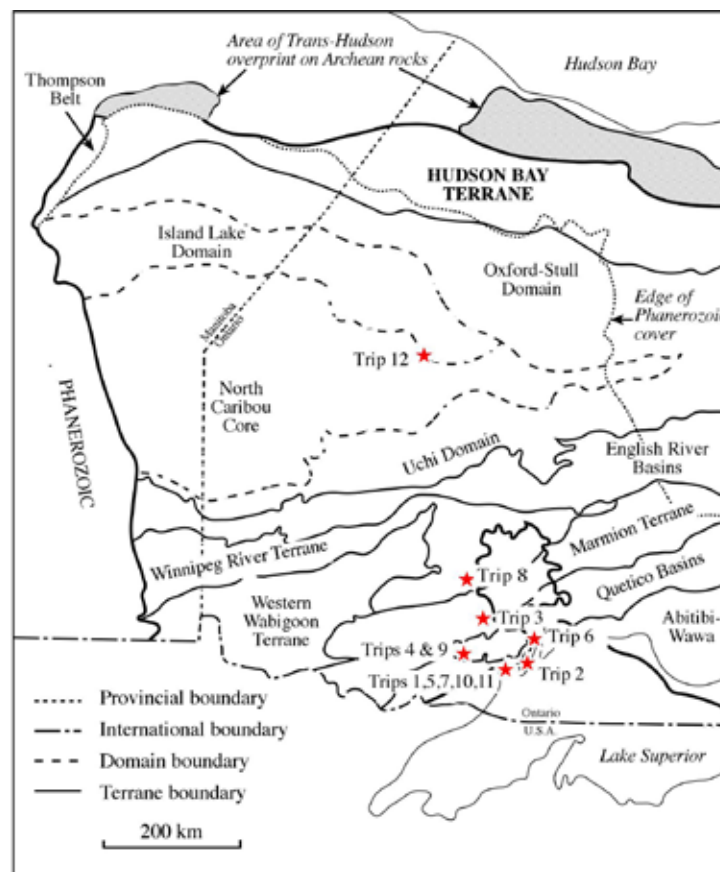


Figure 1. Map showing the general locations of field trips for the 2012 meeting.

Field trips 1 & 13 - Sudbury Impactoclastic Debrisites at Thunder Bay

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Abstract

Eight outcrops of chaotic debrisite containing ejecta from the 1850 Ma Sudbury impact event have been identified in and near the city of Thunder Bay, Ontario, 650 km west of the center of the Sudbury crater. Ejecta features include devitrified vesicular impact glass, spherules, accretionary lapilli, microtektites and tektites, and shocked quartz grains containing relict planar features including planar deformation features. The original volume of ejecta has been significantly reduced by carbonate replacement and recrystallization, so that today ejecta only make up ~ 20 % of the debrisite volume. Major debrisite components include ripped up clasts of carbonate grainstones, stromatolites and chert of the 1878 Ma Gunflint Formation. These Gunflint boulder to coarse sand-sized clasts commonly fine upward, in marked contrast to the chaotic nature of the remainder of the debrisite. Seven of the eight sites have had the upper portion of the impact layer removed by glaciation. The eighth site shows a complete stratigraphic section from the Gunflint Formation, up through the ejecta bearing layer, and into the overlying 1832 Ma Rove Formation.

The sequence of events deduced from these outcrops is as follows.

1. Mafic volcanic ash was deposited and reworked in a carbonate dominated, near-shore environment that supported microbial mat growth and stromatolites.
2. These areas were then subaerially exposed.
3. Upon impact, earthquakes fractured some stromatolites as well as the underlying Gunflint Formation chert and carbonate.
4. Impact-generated density currents (base surges) stripped the area of loose sediment and incorporated ripped up Gunflint chert-carbonate breccia clasts, before being deposited as a chaotic variable layer.
5. In an ensuing period of subaerial exposure lasting < 18 m.y., blocky, meteoric calcite cements formed in this material while weathering and erosive reworking modified the deposits.
6. The Rove Sea then transgressed the area depositing

the overlying Rove Formation carbonaceous shale.

Introduction

In 2005, Addison et al. documented an ejecta layer formed by an 1850 Ma (Krogh et al., 1984) impact event in cores drilled north of Lake Superior in Ontario and Minnesota. Features in this Sudbury impact layer (SIL) included planar shock features, notably planar deformation features (PDF) in quartz grains; accretionary lapilli; devitrified microtektites and tektites; and devitrified vesicular impact glass (DVIG). The ejecta were linked to the 1850 Ma Sudbury impact by their presence between an 1878 Ma Gunflint Formation tuff (Fralick et al., 2002) approximately 105 metres below the ejecta and tuffs variously dated at 1827 ± 8 Ma, 1832 ± 3 Ma and 1836 ± 5 Ma from the Rove and Virginia Formations 5 to 6 metres above the ejecta (Addison et al., 2005). Sudbury is the only known terrestrial impact from this time interval (Earth Impact Database, 2012) and an oceanic impact would not have produced the quartz- and feldspar-rich, craton-sourced ejecta.

The identification of the SIL has led to the discovery of about 30 additional ejecta-bearing drill core and outcrop sites (Fig. 1) in the Lake Superior region (Cannon and Addison, 2007; Pufahl et al., 2007; Jirsa et al., 2008; Cannon et al., 2010). Eight of them are in the northern Gunflint Formation outcrop area in and near Thunder Bay, Ontario (Fig. 1).

Locations are supplied for all sites except one in a private yard which is omitted to protect the owner's privacy. These new sites are 660-680 km from Sudbury. Using the crater radius of 130 km as determined by Spray et al. (2004), these new outcrops are 5.1-5.2r (crater radii) from the Sudbury crater center. The approximate boundary between proximal and distal ejecta is usually given as 5 crater radii (French, 1998) but this boundary is a transition zone, not a sharp line. The ejecta features seen on this field trip are consistently proximal, not distal.

This field trip will examine outcrops for macroscopic ejecta and non-ejecta features (Table 1) and relate

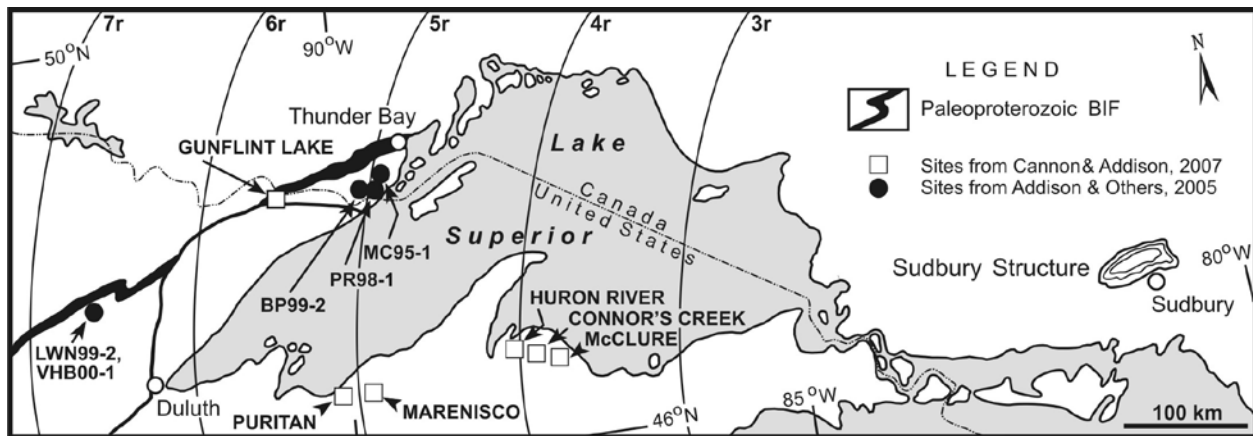


Figure 1. Approximate locations of some Sudbury impact layer (SIL) localities in the Lake Superior region. Concentric lines represent multiples of the final Sudbury impact crater radius of ~130 km as determined by Spray et al. (2004).

them to the dynamics of the Sudbury impact, the second largest and fourth oldest impact known on Earth (Earth Impact Database, Jan. 10, 2012). A large impact results in a sequence of events at the impact site which subsequently played out in the Thunder Bay area. The Earth Impacts Effects Program (impact.ese.ic.ac.uk/ImpactEffects/) allows an estimate of the time of delivery and magnitude of events from the impact by inputting: 1) distance from impact (660 km); 2) projectile diameter; 3) projectile density; 4) projectile velocity; 5) impact angle and; 6) target rock type (crystalline – granitic at Sudbury). A velocity of 25

km/s with an impactor diameter of 23 km, along with the other variables noted above, produces a final radius of 134 km, very close to the actual value of Spray et al. (2004).

It is interesting that the SIL thickness predicted by the model is not matched by reality at Thunder Bay. For instance, the maximum ejecta thickness seen is ~4 m at Hillcrest Park, where it was once thicker because the exposure top is erosively truncated. The only complete stratigraphic exposure of the SIL is 3.2 m. Likewise, complete SIL in drill cores BP99-2 and PR98-1 from ~35 km south of Thunder Bay do not

Table 1. Major effects of a Sudbury-sized impact at Thunder Bay, 660 km from its epicenter, as predicted by the Earth Impacts Effects Program (impact.ese.ic.ac.uk/ImpactEffects/). The field trip outcrops will show the effects of earthquakes and some of the types of ejecta features generated at the impact site. Ambiguous evidence of air blast will be seen at one site. Effects from the fireball have not been identified so far.

Feature	Feature Magnitude	Consequences
Fireball	Reaches max. of 1220 x brighter than Sun ~14.5 s after start of impact.	All subaerial organic matter ignites but this is of little consequence at 1.85 Ga since there was almost no terrestrial life to ignite and little, if any atmospheric oxygen.
Earthquakes	Magnitude 10.7 on Richter scale. They begin 2.2 minutes after impact starts.	Stronger than any ever recorded by humans. Surface bedrock fractured. Soils liquefy and shift, if present.
Ejecta	Ejecta begins arriving 6.5 minutes after start of impact & accumulates to 12 m thickness.	Ejecta mostly delivered by base surges but some arrives in ballistic trajectories. Predicted mean fragment diameter is 1.6 cm. Dust will continue settling months to years afterwards.
Air Blast	Supersonic wind (max velocity of 1200 m/s) arrives 33 minutes after start of impact.	Supersonic wind blast would, today, knock down whatever was left standing by the earthquakes and fires.

exceed 0.8 m in thickness. All of these values are well short of the model's predicted 12 m thickness which raises questions either about the model or about the SIL's post-depositional history or both.

Terminology and Features

Like other branches of geology, the geology of large extraterrestrial impacts has its own rapidly evolving vocabulary which is not widely known in the larger geological community. Therefore terms and specialized impact features are best defined or described and illustrated before seeing and discussing the outcrops on this field trip.

Ejecta

Ejecta is a collective name for anything thrown out of the crater during the impact. It includes target rock breccia clasts of all sizes from μm - to km-scale. It includes melt which cooled to form glass clasts of various shapes and sizes, most of them now devitrified. It also includes dust and glassy spherules which condensed from rock vapour ejected high into Earth's atmosphere and even above it.

Ballistic Ejecta Curtain

Many ejecta components initially travel outward as a curtain on a ballistic trajectory, most of it landing at about $2r$ from the crater center (French, 1998). There, this massive amount of material lands, severely abrading the landscape and incorporating the abraded material into an outward-rolling debris flow.

Debrisite

Shanmugam (2006) argues that tsunamite should not be used to describe tsunami deposits because it describes a process and does not deal with clast sizes like conventional sedimentary terminology e.g., sandstone, claystone, etc. He proposes the term "debrite" for tsunami deposits because of their wide variety of clast sizes. By inference, impact related deposits should not be called impactites. We choose debrisite as the best descriptor for these deposits which result from four sets of related energetic events: 1) the impact; 2) impact-induced earthquakes; 3) ejecta traveling in ballistic trajectories and; 4) ground-hugging density currents (called base surges in earlier literature). Thus, the SIL is composed of debris with clasts in the μm to metre size range and of variable origins. Even though all debrisite sites reported here contain ejecta, debrisite

is not synonymous with ejecta because ejecta features only comprise about 20 % of debrisite while localized areas of some outcrops seemingly lack ejecta.

Ground-hugging Density Currents (Base surges)

Much of the early impact literature applied volcanic terminology to impact generated deposits. In fact, a number of the SIL deposits south of Lake Superior were identified by searching the literature for pyroclastic deposits, then checking to see if ejecta features were present (W. F. Cannon, personal communication, 2008). Base surge is one such borrowed term but we use both base surges and the newer impact literature term ground-hugging density currents interchangeably.

Matrix (not ejecta)

The largest debrisite component by volume is carbonate matrix (Figs. 2A-D), with calcite > dolomite > ankerite. Carbonate has partially replaced most ejecta features (Figs. 3A-D, F) and likely obliterated many more of them. It has also infilled vesicles (Figs. 3A-D). Pervasive carbonate recrystallization, with crystals up to 5 mm maximum dimension, has further destroyed ejecta features (Figs. 3A & F, 2C). There are areas in the matrix comprising angular submillimetre to millimetre carbonate clasts composed of crystals $\leq 20 \mu\text{m}$ in size (Fig. 2). These clasts may represent Gunflint carbonate ground up in the turbulent events leading to deposition of the SIL but, if so, most show at least minor recrystallization planes.

Silica also replaces ejecta features (Fig. 3E). Silica is most visible as anastomosing chert at submillimetre-scales in most thin sections as well as centimetre-scale bands in outcrops. Such silica is usually microcrystalline and is clearly a post-depositional feature.

Today, the matrix comprises ~80 % of the debrisite volume leaving ejecta features at ~20 %. At the time of deposition the ratio of ejecta to matrix must have been a significant but unknown amount higher, judging from the volume of partially carbonate-replaced ejecta remnants.

Stromatolites (not ejecta)

Seven of the eight identified outcrop sites show: 1) debrisite lying directly on stromatolites (Fig. 4A) or microbialite mats and/or; 2) broken subcentimetre to decimetre-size stromatolite clasts within the debrisite. The stromatolite or microbialite clasts attest to violent events breaking them up and mixing them into the

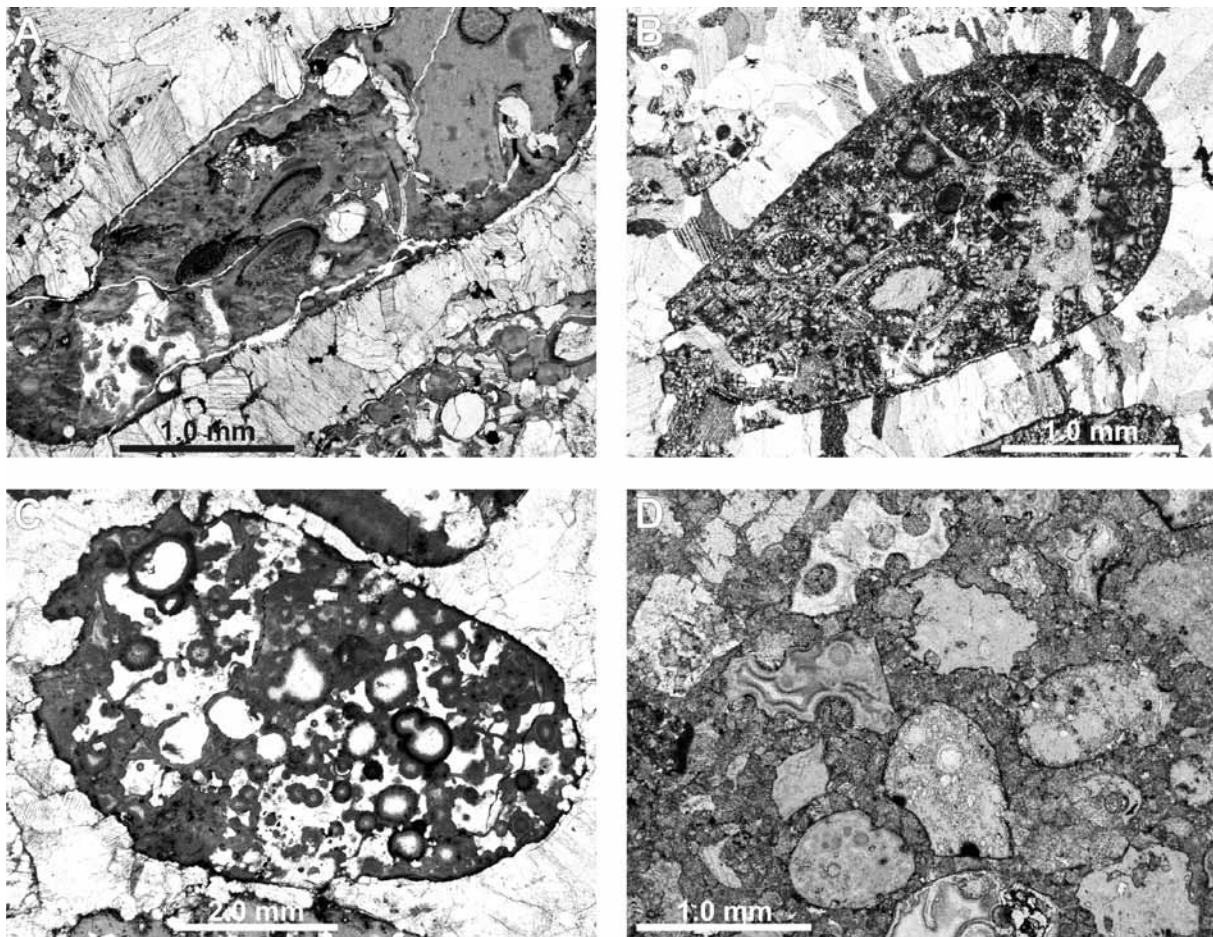


Figure 2. Splashform devitrified vesicular glass (DVIG) clasts in carbonate matrix. A – GTP site; plane polarized light (pp). B – GTP site; crossed polarizers (xp). C – Private yard site (pp). D – Atypical abundance of DVIG fragments, splashform or otherwise. Hwy 588 (pp).

debrisite.

Subrectangular Blocks of Upper Gunflint Chert-Carbonate (not ejecta)

Prior to highway reconstruction in 2011 the original Terry Fox exposure showed the upper 0.5 m of Gunflint chert-carbonate bedrock heavily fractured with the subrectangular blocks slightly separated from each other but still basically in situ (Fig. 5B).

Subrectangular blocks of Upper Gunflint chert-carbonate, commonly exceeding 0.5 m maximum dimension (Fig. 4B), are found at or near the base of most debrisites (Table 1). They, along with fractured chert clasts of all sizes, show upward fining within the chaotic debrisite. The chert-carbonate blocks usually have sharply angled corners, except at the Private Yard (Fig. 4B), GTP and BB sites where some subrounded blocks exist among angular blocks. None of the blocks show weathering rinds.

Anastomosing Silica and Agate with Mini-stalactites (not ejecta)

Hillcrest Park, with its 3.5-4 m thick debrisite, shows extensive post-depositional anastomosing chert deposits, some of which include banded agate in localized zones. These chert deposits flow around debrisite clasts but never cut through them. In places, the chert and agate has been deposited in debrisite voids. Near the top of the Hillcrest Park deposit, two vugs contain silica and agate stalactites 1-3 cm long (Fig. 4E). The Banning Bluff and Baseball Central sites and the DVIG-rich, recessively weathering layer at the Terry Fox site also show anastomosing chert but on a smaller scale than Hillcrest Park. Micro-scale anastomosing chert is seen in thin sections from all sites.

The agate layer at the top of the TF site shows digitate projections from both the base and top of the layer which may have been miniature stalactites and

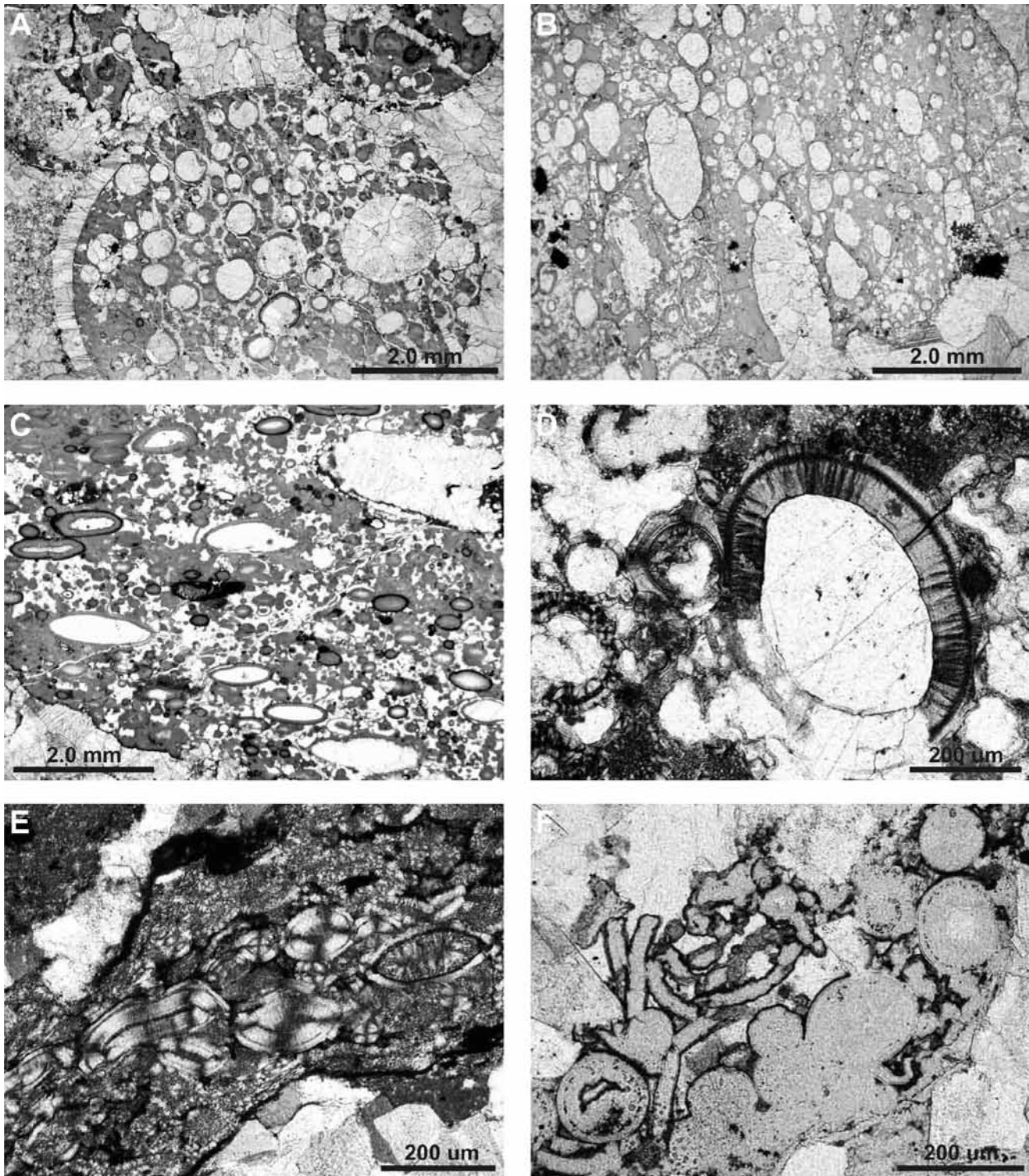


Figure 3. Varying degrees of vesicle deformation in devitrified vesicular impact glass (DVIG) clasts and spherules. A – undeformed vesicles infilled by calcite. Note recrystallized carbonate at top center and right; GTP Site; plane polarized light (pp). B – ovoid vesicles aligned subvertically and infilled with calcite. There is no evidence of lateral compression of the clast, so presumably the vesicles were deformed prior to deposition; GA site (pp). C – calcite-infilled, deformed spherules in cluster; Private Yard (pp). D – fibrous spherule rim, partially destroyed by carbonate replacement; GA site; crossed polarizers (xp). E – collapsed and partially collapsed spherules. Presumably the spherules were hollow prior to collapse. Hwy 588 (xp). F – fractured spherule rim bits and devitrified whole spherules. Note recrystallized carbonate at upper left and lower right; Hwy 588 site (pp).

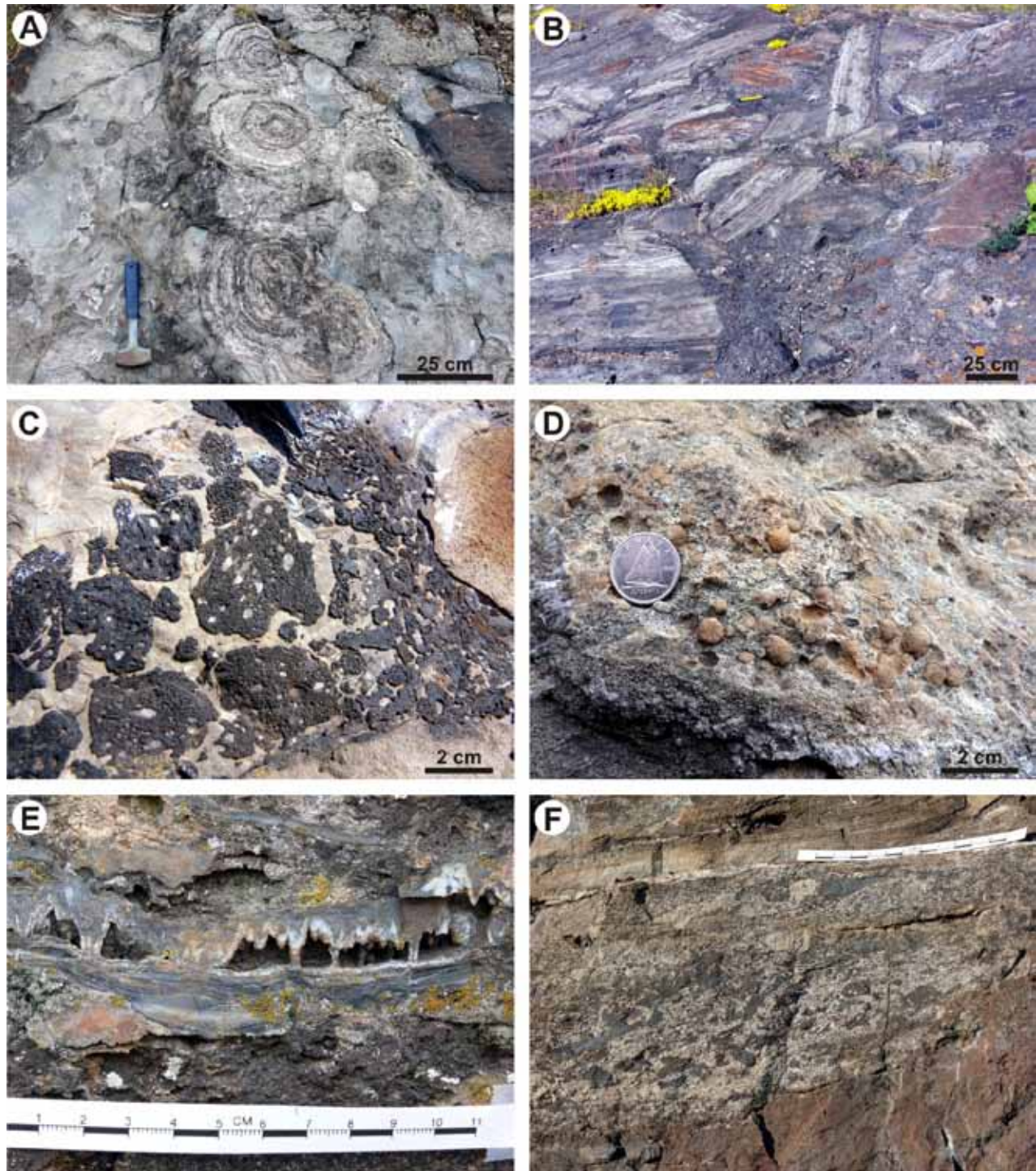


Figure 4. Various features of SIL sites. A – Gunflint Formation stromatolites exposed on a glacially truncated surface. While it is not recognizable in the photo, debrisite lies over stromatolites at upper right of the photo. Private Yard site. B – Angular to slightly subangular clast-supported Gunflint Formation breccia with a finer DVIG-rich and calcite-rich matrix, all of which lies directly on Gunflint stromatolites. The angular clasts suggest a short travel distance from their point of origin. C – DVIG clasts within a recrystallized calcite matrix. The silicate devitrification product supports growth of a black lichen, whereas calcite prevents lichen growth. The vesicles are calcite infilled. Private Yard site. D – Orange weathered accretionary lapilli in a recrystallized carbonate matrix. Hillcrest Park. E – Stalactites hanging from top of a vug with agate flowstone deposited on bottom of vug, an indicator of postdepositional subaerial exposure. Hillcrest Park. F – Ocean transgression sequence beginning with an iron-rich alteration profile at the bottom of the photo which marks the top of the debrisite. Above it are rip-ups composed of mudstones or clasts of the iron-rich alteration profile embedded in a carbonate matrix. The boundary between these two units marks a disconformity. The rip-up zone grades into siltstones of the lower Rove Formation at the top of the photo. Original Terry Fox site. Scale is graduated in centimetres.

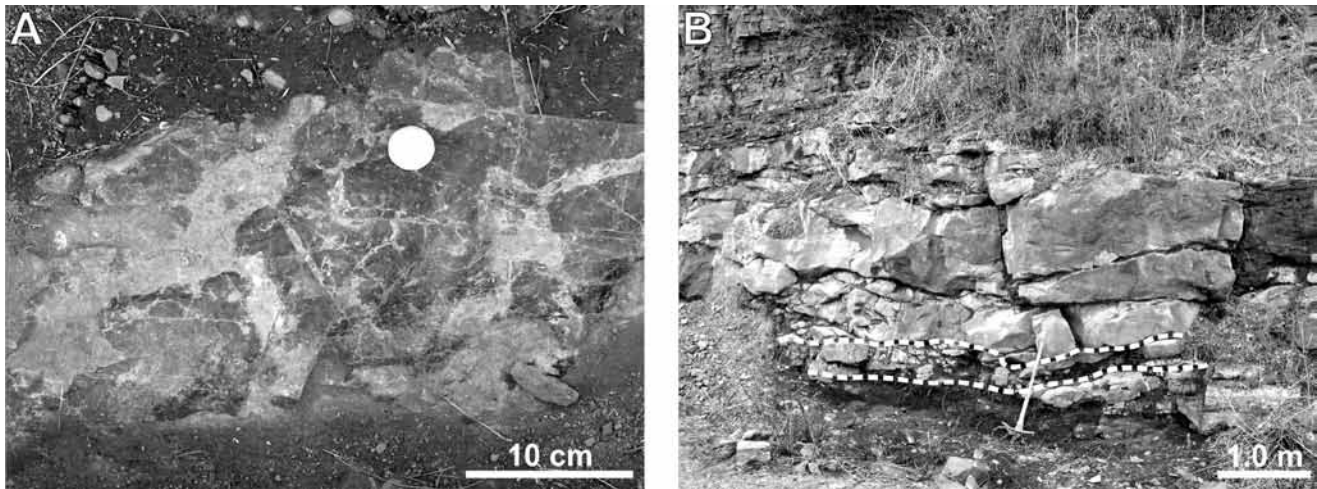


Figure 5. A – Earthquake fractured black Gunflint chert. The fractures are thought to have opened during passage of the dilational phase of the earthquake wave, allowing very fine-grained ankeritic sand (light gray) to fill the cracks preventing them from closing back up. Highway 588 site. B – Rectangular to subrectangular earthquake fractured uppermost Gunflint chert-carbonate clasts which delaminated along bedding planes (between dotted lines). These blocks are still more or less in situ, the base surge having failed to rip them up. The more blocky material on top of it is the SIL debrisite. This old Terry Fox site, was removed by highway reconstruction in 2011.

stalagmites at one point but, if so, silica deposition continued until they were all encased in a solid agate mass.

Unshocked Quartz and Feldspar Grains (some may be ejecta, some are not)

Both angular and subrounded detrital quartz and feldspar grains are found in the debrisite at all eight sites, with some thin sections showing as many as 10-20 grains per slide. The grains range in size from 40 μm to 800 μm , too small to be seen on this field trip. The angular grains tend to be at the small end of this size range, are shard-like and are likely ejecta. Subrounded grains are probably detrital sand picked up by the base surges flowing across the landscape. Neither grain type is seen in Gunflint Formation rocks.

Planar Features in Quartz Grains (ejecta)

A few quartz grains show planar features, some of them PDFs as defined by French (1998 and references therein). Nearly all of the planar features have been found within accretionary lapilli at the Hwy 588 site and Hillcrest Park (Figs. 6A-4D). Up to three intersecting sets of PDF are seen in quartz grains, which are typically 50-100 μm in size.

A single quartz grain from the Hwy 588 site has planar fractures (Fig. 6D) with their characteristic wide spacing and thick lines (French, 1998). A quartz grain from the Terry Fox upper iron-rich alteration zone

shows similar features.

The dark lines of PDFs are isotropic quartz glass formed when the high pressure shock waves instantaneously destroy the quartz crystal structure without melting it. They are diagnostic of extremely high pressure shock waves, only obtained in nature by impacts, but they are microscopic and, thus, not seen on this field trip.

Spherules (ejecta)

Most spherules seen near Thunder Bay are frozen melt droplets ejected during the impact. Spherule sizes range from 50 μm to 1 mm.

Except for very rare single spherules, all spherules are clustered, with cluster sizes ranging from 1 mm to 5 cm maximum dimension (Figs. 3E, 5F). Extrapolating from the two dimensions seen in thin section clusters to three dimensions, spherule numbers probably ranged from a few tens of spherules to perhaps 1,000 per cluster. Some spherule clusters contain quartz grains within them. Spherule clusters are the dominant ejecta feature by volume in these debrisites. Despite that, identifiable spherule clusters will not be seen on the field trip.

Spherules show prominent rims (Figs. 3D-F). Many rims are round and complete; however, other rims are variously deformed, ranging from slightly ovoid, through ovoid, to totally collapsed where all that remains are two flat rim layers squeezed together

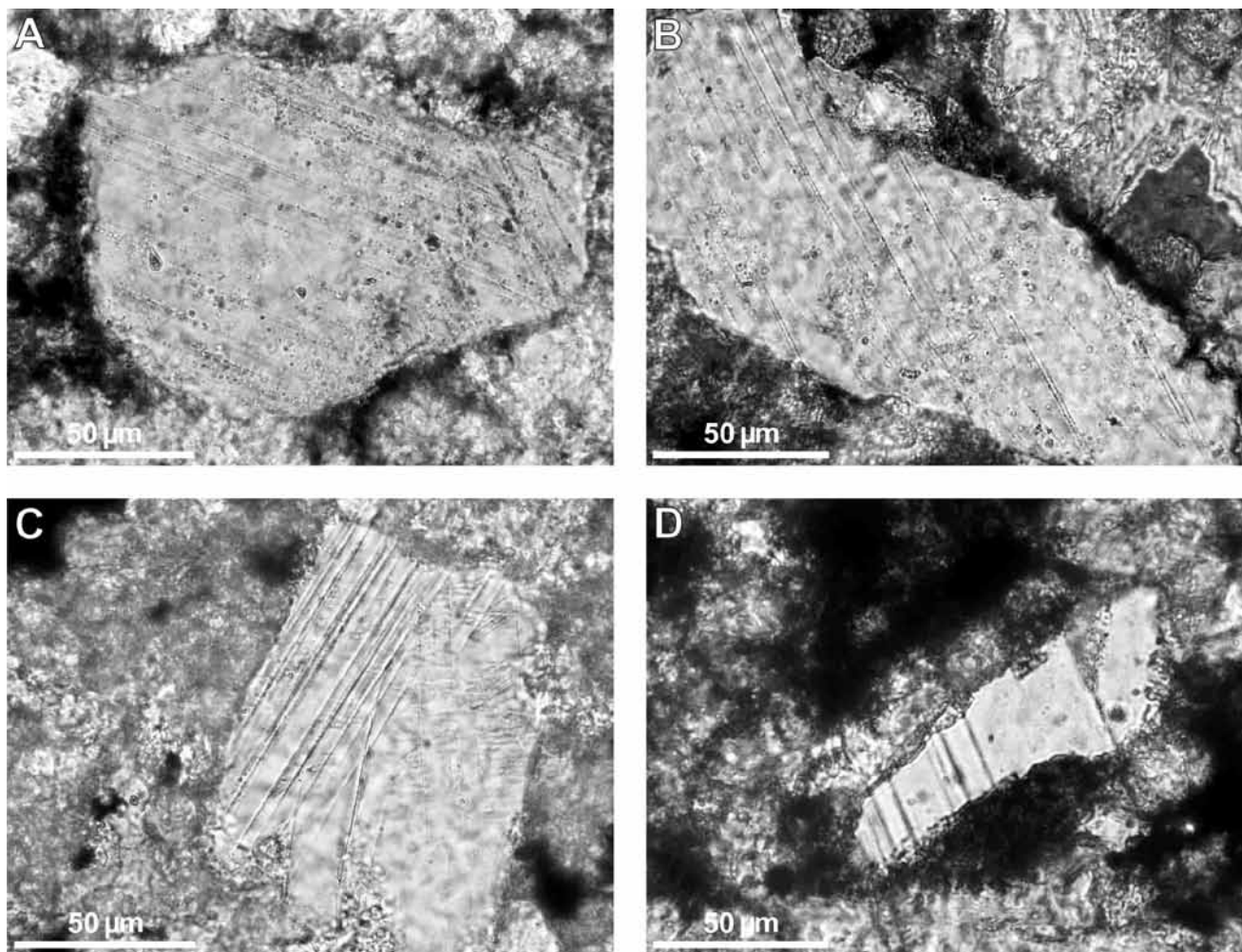


Figure 6. Planar features in quartz grains within accretionary lapilli. A – two PDF sets; Hillcrest Park; plane polarized light (pp). B – single PDF set; Hwy 588 site; crossed polarizers (xp). C – two PDF sets, with the less distinct set along the right side of the grain; Hillcrest Park (xp). D – planar fractures; Hwy 588 (pp).

(Fig. 3E), suggesting that these spherules were hollow. In some cases, a portion of the rim has fractured but remains attached to an otherwise almost intact spherule. Fractured rim pieces are also seen amongst intact spherules, or they are randomly oriented in a cluster, presumably at the site of a former spherule (Fig. 3F). Sometimes the spherules in an entire cluster are collapsed, but in other cases, only a few spherules within a cluster are collapsed. Clusters showing spherules with little or no deformation generally show spherules with 1-3 contact points with adjacent spherules indicating that they have experienced little post depositional compaction (Simonson, 2009).

Spherule rims range from amorphous features composed of unidentified clay minerals (Fig. 3F) to crystalline silica (Fig. 3E) and calcite rims, and crystalline rims of as yet unidentified minerals. Rims also vary in thickness. Some of the smallest clay-

replaced spherules are rimless or else the rims were destroyed during devitrification.

Spherules generally show one of two general core types. The first is featureless and composed of as yet unidentified clays. The second core type is crystalline, most commonly calcite, and less commonly silica. The cores are usually centered within the spherule but in some cases they are off-center. The outer boundary of the cores is typically smooth but some cores show botryoidal ingrowths from the core edge towards the center. In other cases carbonate replacement has produced uneven core boundaries.

Carbonate replacement of spherules is pervasive. Judging from the remnants, we estimate that >50 % of the spherules have been replaced by recrystallized carbonate, most commonly calcite and less so, dolomite and rarely ankerite. The volume of material

replaced by carbonate is probably significantly higher than this because portions of thin sections are pure recrystallized carbonate, offering no clue as to what the original material was in those areas.

Perhaps the most notable feature of the spherules and spherule clusters is their extensive morphological and compositional variability.

Devitrified Glass (ejecta)

There are three categories of devitrified glass: 1) devitrified vesicular impact glass (DVIG) clasts and, 2) rare microtektites (<1 mm in size) and tektites (>1 mm in size) and 3) spherules. It will be seen best at the Private Yard site (Table 1.1).

DVIG clasts are usually irregularly shaped (Fig. 4C) but splashform (streamlined) shapes are also present (Figs. 2A-C). They range in size from 1-2 mm up to 5 cm. Vesicles in DVIG are usually infilled with carbonate, most commonly calcite. Vesicle shapes range from round (Fig. 3A) to ovoid (Figs. 3B-C). If most vesicles in a clast are ovoid, they show a preferred orientation along their long axes (Figs. 3B, C). Vesicle size, whether within a single clast or between clasts, is also variable (Figs. 3A-C).

There are few positively identifiable microtektites and tektites in these deposits but carbonate-replaced microtektite and tektite shapes are more numerous. However, the Gunflint Formation has iron-rich chloritic granules that have many shapes in common with splashform microtektites and are the same size (average 0.8 mm). The two can only be distinguished if some remnant of their internal structure has not been replaced by carbonate. Microtektites and tektites show a blue-gray platy or granular fabric under crossed polarizers, whereas chloritic granules have a blotchy black appearance in plane polarized light. Thus, if a microtektite shape is totally carbonate replaced, there is no way of visually determining whether it was a microtektite or a Gunflint chlorite granule.

Accretionary Lapilli (have ejecta and non-ejecta components)

Accretionary lapilli consist of fine clasts of accreted target rock, usually quartz, some of which show shock induced planar deformation features (PDFs) and feldspar. The accretionary lapilli form in the base surge from the impact and as such incorporate non-target dust-sized particles picked up by the ground-hugging base surges. Impact-generated base surge dynamics

are poorly understood and super-computers are not yet powerful enough to model these complex flows (N. Artemevia, personal communication, 2008).

Accretionary lapilli and armored lapilli are found in outcrop only at Hillcrest Park and Hwy 588, and then only within localized areas of the larger exposure at each site. They range in size from 2-13 mm maximum dimension at Hillcrest Park and 5-25 mm maximum dimension at the Hwy 588 site. Lapilli show rounded to subrounded shapes (Figs. 7A-D). Lapilli fragments are present but rare, so they have undergone little breakage. The lapilli range from fairly uniformly gray accreted grains to ones with alternating dark gray, thick bands with thinner bands of very fine black amorphous material (Fig. 7B). The alternating dark gray and black laminations may be repeated up to two times in larger lapilli. The black laminations appear in ~ 35 % of lapilli.

By volume, the most common feature in lapilli is 10-50 µm carbonate crystals. Larger carbonate clasts, up to 2.5 mm maximum dimension, form the cores of the few lapilli in which cores are visible. Lapilli show carbonate recrystallization, but where this has occurred, it has enlarged the original crystals only marginally. In contrast, large recrystallized carbonate crystals up to 5 mm in size are abundant in the debrisite outside the lapilli abutting their outer margins, so some unknown factor has inhibited large scale carbonate recrystallization within lapilli.

All lapilli also contain 30-250 µm grains of quartz and feldspar, which are found scattered in the gray, coarser-grained areas of the lapilli. Quartz and feldspar grains comprise about 5% of the total lapilli grain population (carbonate grain clusters being the majority). Approximately 1% of the quartz grains exhibit PDFs, making them very difficult to find.

The features in the Hillcrest Park lapilli are more poorly defined than those at Hwy 588. The Hillcrest Park lapilli thin sections were prepared from weathered rock, whereas the Hwy 588 lapilli thin sections were prepared from unweathered rock.

Geologic Setting

The Sudbury impact occurred during a period of tectonic activity along the southern margin of the Superior craton. Prior to the Sudbury event, two interpretations for the area's geologic setting exist. Kissin and Fralick (1994), Hemming et al. (1995), Van Wyck and Johnson (1997) and Pufahl et al. (2004) see it

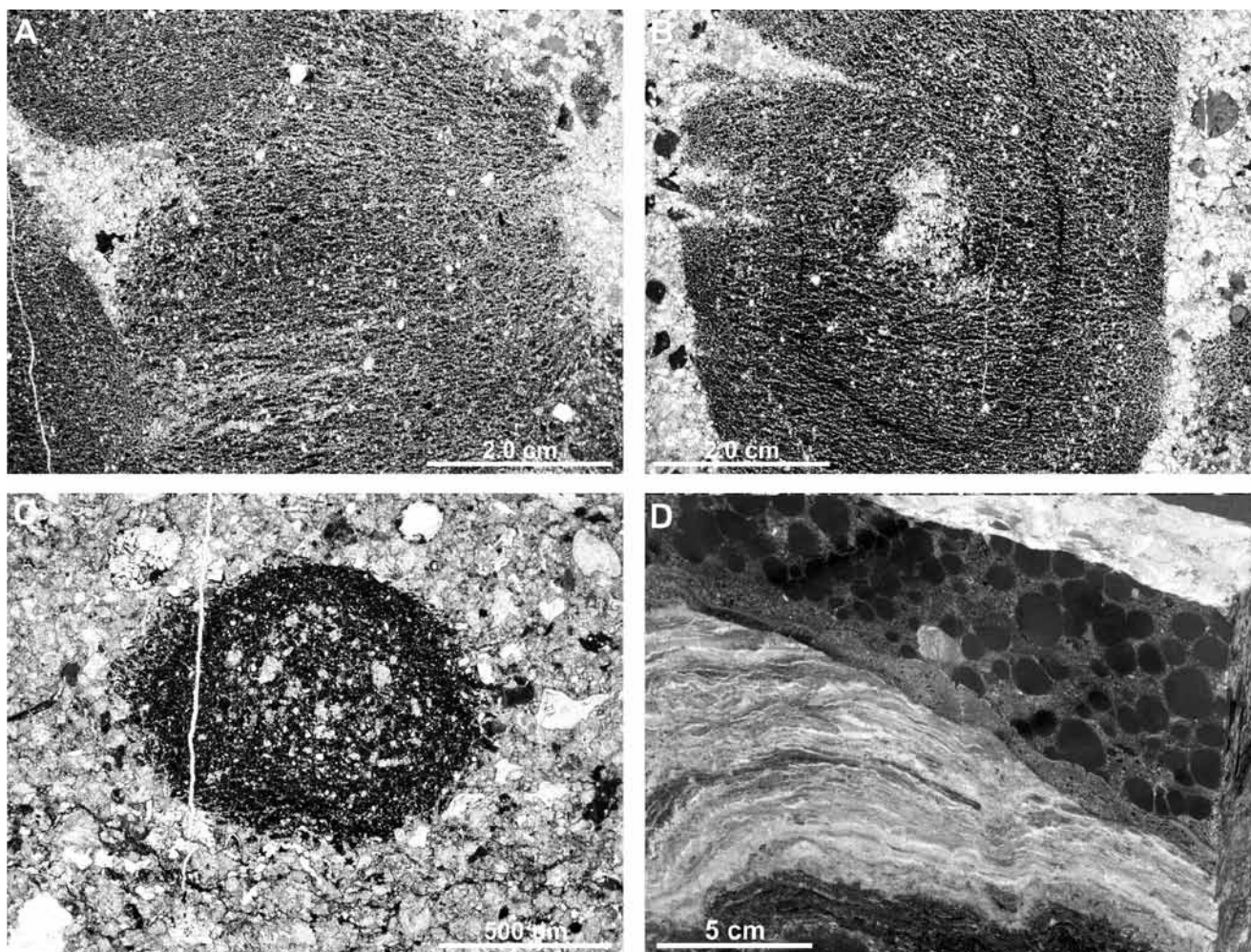


Figure 7. Accretionary lapilli. A – ‘stack of cards’ type accretionary lapilli; Hwy 588 site. Terminology after Schumacher and Schmincke (1991, 1995). B – armored, banded accretionary lapillus. Black band is an extremely fine-grained unidentified black substance. The nucleus is a clast of fine-grained, angular, fractured carbonate; Hwy 588 site. C – accreted material resembling an accretionary lapillus but < 2mm in diameter. Yancey and Guillemette (2008) have called such structures sublapilli, a term which we adopt. Hwy 588 site. D – 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; Hwy 588 site, polished surface. The gray component in all lapilli photos is primarily fine-grained, angular, fractured carbonate clasts whose individual crystals are usually < 10 µm maximum dimension. These clasts are typically < 50 µm in size but they may be as large as 500 µm. Quartz and feldspar grains are a minor component among the carbonate clasts within the lapilli. The black lapilli on the polished surface in D resemble those in A, B and C when seen in thin section. Lapilli from Hillcrest Park are not shown because they are heavily weathered and their features are less distinct.

as a backarc basin formed on this margin as extension, possibly subduction roll-back had caused an area of the continental crust to subside and be flooded. An alternative explanation is summarized by Schneider et al. (2002) and Schulz and Cannon (2007) involving successive island arc collisions and the development of a foreland basin which subsided to receive the Gunflint Formation sediments on its northern margin. Pufahl et al. (2010) described the backarc basin evolving into a foreland basin.

Chemical sediments (chert, iron oxides and iron carbonates) and volcanoclastics of the Gunflint Formation were deposited onto Archean basement rocks (Gill, 1926; Tanton, 1931; Moorehouse and Goodwin, 1960) in a nearshore marine setting and organized into fining- and coarsening-upwards successions (Fralick and Barrett, 1995) on this open, wave and tide dominated environment (Ojakangas, 1983; Fralick, 1988; Pufahl et al., 2000; Pufahl and Fralick, 2004). Shegelski (1982) interpreted

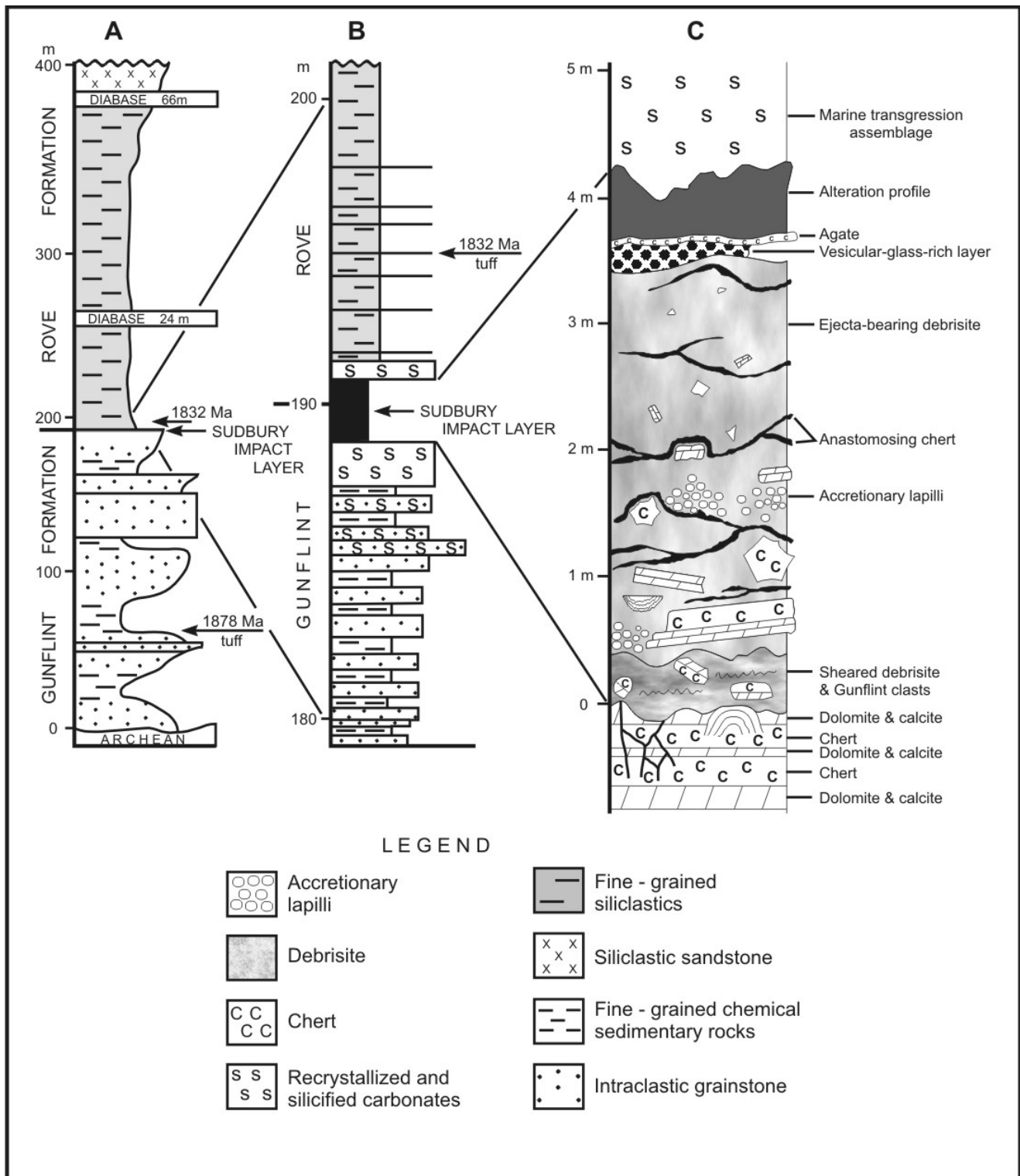


Figure 8. A – general stratigraphy of the Gunflint and Rove Formations showing location of the Sudbury Impact Layer (SIL) and the locations of dated zircon. B – more detailed stratigraphy for 10 m above and below SIL. C – composite cartoon of debrisite features from all eight SIL sites. No site shows all features.

stromatolites and carbonate at the top of the Gunflint Formation in the Thunder Bay area as a carbonate-rich lagoon environment marking the end of the Gunflint Formation. A depositional hiatus exists between the

top of the 1878 Ma (Fralick et al., 2002) Gunflint Formation and the overlying 1832 Ma (Addison et al., 2005) basal Rove Formation, probably caused by the 1860-1835 Ma (Sims et al., 1989) Penokean Orogeny

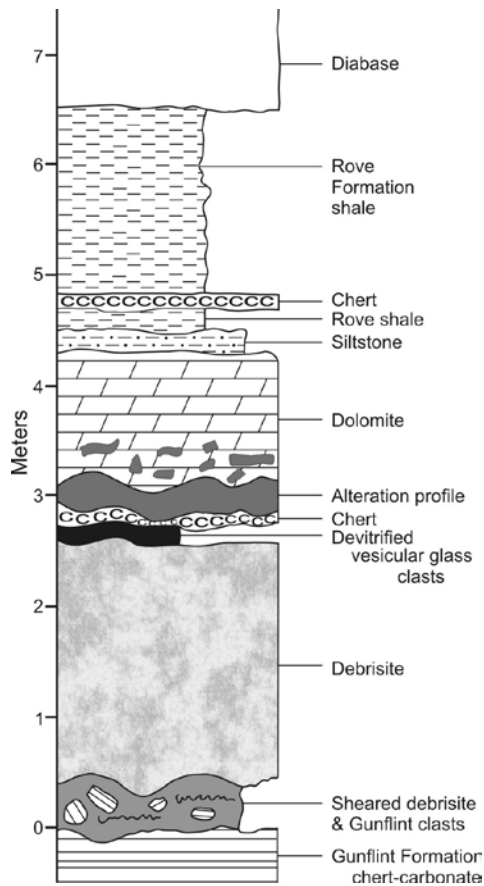


Figure 9. Cartoon stratigraphic column as seen at the pre 2010 Terry Fox Lookout rock cut, Hwy 11-17. This is the only complete outcrop exposure extending from the Gunflint Formation, through the Sudbury impact layer and up into the Rove Formation in the Thunder Bay area. Disconformities exist at the Gunflint Formation-sheared debrisite contact and the alteration profile-dolomite contact.

to the south, which resulted in crustal up-warping and withdrawal of the sea (Johnston et al., 2006; Cannon and Schulz, 2009). Alteration of this subaerial Gunflint surface, including development of meteoric calcite cement, silicification, and agate/pyrite veins and vugs, occurred during this time interval (Tanton, 1931; Fralick and Burton, 2008).

The SIL lies on this stromatolitic, silicified carbonate surface at the top of the Gunflint Formation (Figs. 8, 9). The SIL is overlain by carbonaceous black shale and grainstone of the Rove Formation which records the end of the Penokean Orogeny and the beginning of crustal relaxation and flooding of the area. The Rove sediment was likely eroded from the Trans-Hudson Orogen to the northwest (Maric and Fralick, 2005; Johnston et al., 2006). Today, the Gunflint and Rove Formations lie on a homocline dipping southeast towards Lake Superior at an average of 5° (Gill, 1926). These rocks remain unmetamorphosed except for localized zones adjacent to diabase sills and dikes (Tanton, 1931).

The Debrisite Outcrops

The outcrops located since 2005 (Fig. 10) contain ejecta features similar to those seen in the drill cores described by Addison et al. (2005; Fig. 1). However, there are significant variations in the particular ejecta features present from outcrop to outcrop (Table 2) and on a decimetre- to metre-scale within a single outcrop. Only the Terry Fox site (TF) displays a complete

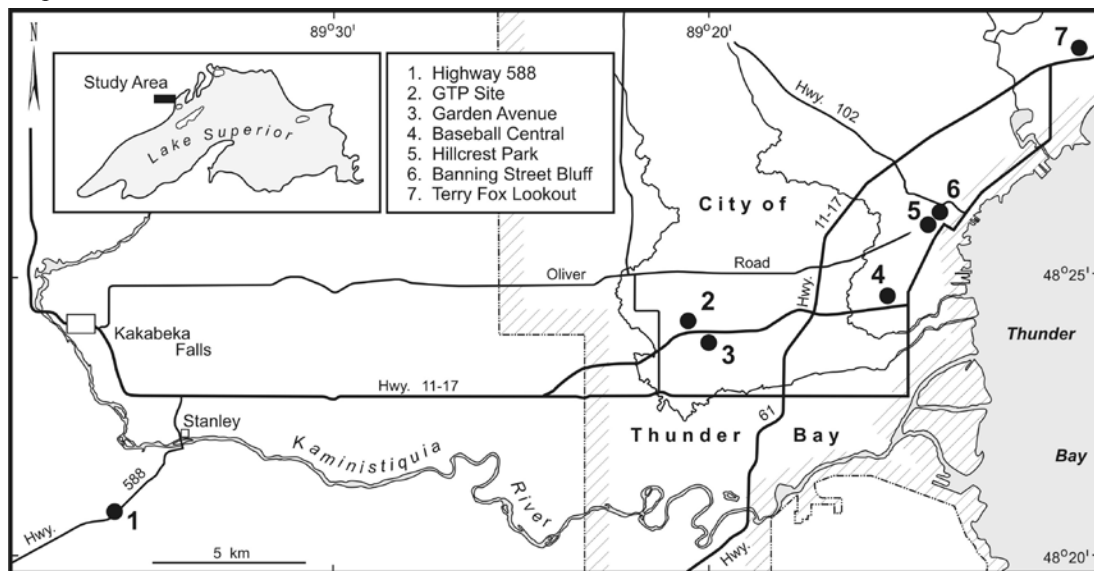


Figure 10. Debrisite containing Sudbury impact event ejecta in and near The City of Thunder Bay. Note: one site in a private citizen's yard is not shown to protect their privacy. Sites 2-6 are either on private property or in city parks and, as such, are "No Hammer" and "No Collecting" zones.

Table 2. Each debrisite site has a unique combination of ejecta and non-ejecta features that are summarized in Table1. Please note that “no” has two meanings: 1. the feature may not be present at all at that site and; 2. we have not found the feature but it may be present. For instance, more thin sections might show PDFs where none are currently found. “??” means that the evidence for this feature is weak and ambiguous. Again, more thin sections might resolve the ambiguity.

Site	Highway 588 ditches	GTP abandoned railway rock cut	Garden Avenue Quarry area	Private yard, Thunder Bay	Banning St. bluff below	Baseball Central, Central Ave.	Highway 11-17 at Terry Fox Lookout	Hillcrest Park, Thunder Bay
Stromatolites or microbialite mats in situ below debrisite base	yes	no	yes	yes	no	no	no	yes
Ripped up stromatolite or microbialite clasts in debrisite	yes	yes	??	yes	yes	no	no	no
Subrectangular Gunflint blocks, >0.5 m maximum dimension in debrisite	no	??	no	yes	yes	yes	no	yes
Fractured black chert or chert- carbonate more or less in situ below ejecta base	yes	yes	no	no	no	no	no	no
Angular chert clasts and shards, sub-cm to 3 dm max. dimension in debrisite	yes	yes	yes	yes	yes	yes	yes	yes
Post-depositional anastomosing chert in debrisite	no	no	no	no	yes	no	no	yes
Alteration profile (possible paleosol) below base of debrisite	no	no	no	yes	no	yes	yes	yes
Devitrified vesicular glass with carbonate in-filled vesicles	yes	yes	yes	yes	yes	yes	??	yes
Accretionary lapilli	yes	no	no	no	no	no	no	yes
Microtektites in debrisite thin sections	??	??	no	no	no	??	yes	no
Gunflint Formation iron granules	yes	yes	no	yes	yes	no	??	no
PDF in quartz or feldspar grains/shards	yes	no	no	no	??	no	??	yes
Isotropic quartz containing crystallites	no	yes	yes	no	??	yes	no	no
Subrounded to angular quartz & feldspar grains in debrisite matrix	yes	yes	yes	yes	yes	yes	yes	yes
Top of debrisite deposit visible	no	no	no	no	no	no	yes	no
Bottom of debrisite deposit visible	yes	??	no	yes	no	yes	yes	yes
Approximate debrisite thickness (m)	0.4	0.4	0.5	0.6	~ 2	~ 2	2.7	~ 4

stratigraphic column extending from the Upper Gunflint Formation, through the debrisite and up into the Rove Formation (Fig. 9). The other seven sites (Fig. 10) are all erosively truncated. Most of these sites are also briefly described in Jirsa et al. (2011).

All of the ejecta-bearing debrisites, except the TF site, are seen primarily in plan view and range in area from as little as 10 m² at the Highway 588 (Hwy 588) site to over 1000 m² at Baseball Central and Garden Avenue. Most sites also show some portion of themselves in cross-section. Preserved debrisite thickness ranges from 0.4 m at the Hwy 588 and Grand Trunk Pacific Railway (GTP) sites to 3.5-4 m at Hillcrest Park (Table 2).

The SIL shows four major components: 1) a matrix of carbonates, commonly dolomite and calcite and least commonly ankerite; 2) Gunflint Formation clasts in the submillimetre to metre size range; 3) ejecta and; 4) minor components of uncertain origin such as subrounded quartz and feldspar grains. The debrisites are chaotic, showing large variations in the percentage of the various components both in surface and cross-sectional exposures. Gunflint Formation clasts show upward fining when seen in cross-section thicknesses >1 m, whereas there is little evidence of upward fining in the other components.

N.B. The only sites on public land are Highway 588 and Terry Fox on Highway 11-17. Please respect “no hammering” and “no collecting” at all other sites.

Stop 1. Hillcrest Park

UTM coordinates: NAD83; 16U 0334728E / 5366952N

Hillcrest Park has debrisite exposures on a dip slope with a true thickness of 3.5-4 m, the thickest of all exposures. However, it was once thicker because it lacks a carbonate cap topped by shale, which marks the transition from the debrisite to the Rove Formation seen at the Terry Fox site and in drill cores (Addison et al., 2005). An intermittent, erosively truncated, 5-15 cm thick microbialite layer lying on Upper Gunflint chert-carbonate lies beneath the debrisite.

The Hillcrest Park lane cliff face shows four chaotic, undulating, largely ungraded lenses with one lens displaying a prominent U-shaped channel. The lenses become thinner upwards. Each shows a heterogeneous mix of features and a chaotic patchiness at decimetre-to metre-scales.

Weathered accretionary lapilli are present and are confined to a localized area comprising <5 % of the total exposure face. The patch of 3-13 mm diameter lapilli (Fig. 7D) is located 1.5-2.5 m above the base of the exposure. Planar features and PDFs are present in quartz grains within accretionary lapilli (Figs. 4A, 4C). Planar features have not been found in quartz grains outside accretionary lapilli within the debrisite matrix.

Scattered angular Gunflint chert and chert-carbonate rip-ups range in size from 0.5 m maximum dimension near the base of the deposit to 1-2 mm near the deposit top. One disintegrating heavily weathered round granitic boulder 33 cm across lies at the base of the debrisite.

Carbonate-replaced microtektites may be present based upon size and shape of some features. However, carbonate-replaced Gunflint Formation chlorite granules have sizes and shapes similar to microtektites making it impossible to distinguish between the two if they are totally carbonate replaced as described previously. So far, only one confirmed microtektite has been observed at Hillcrest Park compared to tens of carbonate-replaced microtektite or granule shapes.

Irregularly shaped DVIG clasts up to 2 cm maximum dimension are a common debrisite feature. Some DVIG vesicles are ovoid or totally flattened.

Spherules appear in clusters in which the spherules are frequently deformed or crushed. Many apparent spherule clusters are heavily altered by carbonate replacement making it difficult to determine whether the feature is carbonate-replaced DVIG or whether they are really spherules. Most ejecta features at Hillcrest Park are poorly preserved because of a combination of carbonate or silica replacement and weathering.

This is the best site to view post depositional anastomosing black chert and light gray and black banded agate in the debrisite. The agate usually appears to have been deposited in vugs. Centimetre-sized silica stalactites occur in two vugs (Fig. 4E) near the top of the debrisite exposure.

Stop 2. Private Yard

(no UTM coordinates to protect owner's privacy)

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 (Fig. 4B) and ejecta, primarily DVIG, which is surrounded and partially replaced by blocky calcite

cement (Figs. 4C, 3C, 2C). The debrisite remnant preserved here is 0-0.5 m thick and unconformably overlies stromatolites and chloritic grainstone of the uppermost Gunflint Formation (Fig. 4A). 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.

Stop 3. Banning Street Bluff (BB)

UTM coordinates: NAD83; 16U 0335129E / 5367236N

A bluff at the north end of Banning Street shows both SIL and Upper Gunflint Formation clast-supported breccia composed of cobble to boulder-size clasts as large as 3-4 m maximum dimension, separated in places by pyritic and carbonaceous black shale similar to the Rove Formation. There are both subrectangular Gunflint Formation chert-carbonate blocks and many-sided, nearly equidimensional chert blocks. As with Gunflint breccia and clasts at other sites, none show weathering rinds. Stromatolite clasts rest upside down and on their sides in the debrisite. The only ejecta-bearing debrisite lies at the base of the breccia pile, the inverse of the sequence at the GTP site.

Other non-ejecta features include millimetre-scale, sharply angular chert fragments. One chert fragment contains chloritic granules similar in shape and size to microtektites. Three ejecta-bearing 26 mm by 45 mm thin sections showed only one subrounded quartz grain. Postdepositional anastomosing chert is present but it is not nearly as common as at Hillcrest Park.

There is meager evidence of ejecta in the BB debrisite with DVIG clasts and both crushed and uncrushed spherule clusters being the most obvious ejecta features. A 250 µm clast showed one set of enigmatic planar features in a quartz crystal within it, plus a second crystal with two sets of possible relict PDFs. Microtektites were not observed in the BB thin sections.

We interpret this site to be a slide deposit which occurred after the SIL layer was lithified and after transgression by the Rove Sea.

Stop 4. Highway 11-17, Terry Fox Lookout (TF)

UTM coordinates: NAD83; 16U 340112E / 5372511N

The Terry Fox site today was created by Hwy 11-17 reconstruction in 2011 (Fig. 11). The old rock face, now removed for fill, was about 40 years old and its weathered surface showed a number of faint features brought out by the weathering (Fig. 12). Had we not had the benefit of the weathered surface we probably would not have been able to interpret the new rock face (Fig. 11). The following description is based on the now-removed outcrop. Given another 3-5 decades, the new face will probably resemble the old one.

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 in turn by a diabase sill (Figs. 9, 10). 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 (Fig. 13) or in thin section. Almost all detail, including any vesicles in possible DVIG 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-



Figure 11. The current Terry Fox outcrop is a nearly featureless gray carbonate face. Given several decades of weathering, faint features within the debrisite should begin to appear as in Figure 12. White vertical scale is 1 m. Rusty area is weathered rock.

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 (Fig. 4E), 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-size rip-up mudstone clasts and probable Gunflint Formations clasts (Fig. 4F). This is followed by millimetre- to centimetre-scale layered carbonate strata topped by a zone with a few poorly defined,

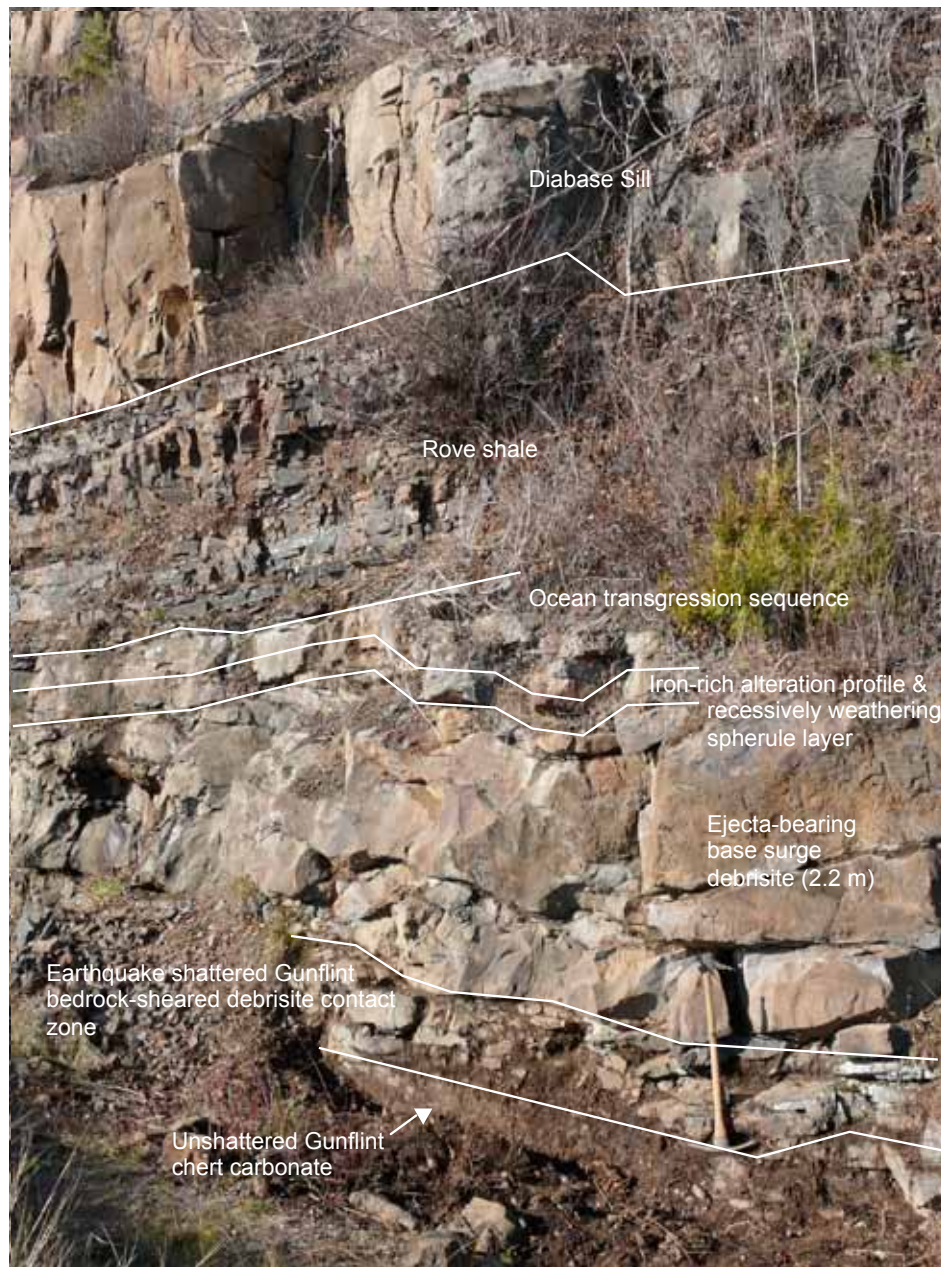


Figure 12. The weathered Terry Fox outcrop on Highway 11-17, Dec. 26, 2010. The weathered surface brought out faint features that had been carbonate replaced, notably devitrified vesicular impact glass (DVIG) shapes and tektites. Pick is 0.9 m tall.

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.

Stop 5. Grand Trunk Pacific Railway Rock Cut (GTP)

UTM coordinates: NAD83; 16U 0326399E / 5363836N

A cut through an outcrop knob on the abandoned GTP right-of-way, approximately 0.5 km east of Mapleward Road and north of Highway 11-17, shows about 4 m composed of Upper Gunflint Formation clast-



Figure 13. Weathered carbonate-rich debrisite surface of pre-2011 Terry Fox rock cut. The rounder light features are likely microtektites or DVIG, however, they are sufficiently indistinct that this is not certain. Many of the more angular features are probably large recrystallized carbonate (likely dolomite) crystals. The reddish tinge is due to a light iron staining.

supported, boulder-sized breccia lacking visible ejecta features topped by up to 0.4 m of DVIG-rich debrisite similar to that seen at the private yard. The Gunflint breccia blocks range from well rounded to angular. The largest clast is a rectangular calcite-cemented slab of grainstone 0.4 m thick by 5 m long containing an upside down stromatolite indicating that the slab is overturned. The small amount of matrix between blocks appears similar to the blocks but finer grained. The dip of the Gunflint beds increases westward, suggesting a nearby fault beneath overburden. The east side of the knob had exposures of fractured but in situ, sharp-cornered Gunflint chert-carbonate. The fractures remain closed and did not show any infilling of the ankeritic grainstone seen at Hwy 588. This exposure has now been destroyed by ATV traffic.

The overlying ejecta-bearing debrisite is dominated by irregularly shaped and splashform DVIG clasts (Figs. 2A & B). Vesicle shape ranges from round (Fig. 3A) to ovoid (Fig. 3B) to crushed (Figs. 3E & F) depending on the clast examined. All vesicles but the crushed ones are infilled with calcite. A small number of deformed spherule clusters are present. A single 1.0 mm accreted sublapillus (pellet) composed of quartz and feldspar grains, similar to accretionary lapilli, is present but accretionary lapilli (>2 mm diameter) are absent. Rounded to angular sub millimetre quartz grains are present.

Gunflint Formation clasts containing blotchy black

chloritic granules are abundant at this site but single granules are rare within the secondary carbonate cement.

Stop 6. Highway 588 (Hwy 588)

UTM coordinates: NAD83; 16U 0307539E / 5357977N

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 (Fig. 7D). 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 (Fig. 5A).

Thin sections prepared from the blasted material show a variety of ejecta features, the most obvious being accretionary lapilli (Figs. 7A-D) which have yielded quartz and feldspar grains showing planar deformation features (PDFs) and planar fractures (Figs. 6B & D). 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 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 (Figs. 3E & F).

Non-ejecta features include subrounded to round chert grains in carbonate cement, subcentimetre 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.

Optional stops - Garden Avenue Quarry (GA) and Baseball Central (BC).

These two outcrops will not be visited on the trip because their features are similar to the outcrops we will visit. However, should you wish to visit them their locations follow.

Garden Avenue Quarry, accessed off Hwy 11-17: UTM Coordinates: NAD83; 16U 032695E / 5363209N.

Baseball Central, accessed off Central Avenue: UTM Coordinates: NAD83; 16U 332333E / 5364300N.

Interpretation and Discussion

Do All Eight Sites Contain Sudbury Ejecta?

Only the Hillcrest Park and Hwy 588 outcrops contain shocked quartz with PDF sets indicating that their debrisites are of impact origin, which leaves the origin of the debrisites at the other six sites open to question. However, a variety of features (Table 1) indicate that all sites share a common origin, namely their common stratigraphic position, their chaotic nature, upward fining Gunflint Formation clasts, DVIG clasts, spherules, and iron-rich alteration profiles, to name the more obvious ones. Therefore, all eight outcrops are attributed to the Sudbury impact event.

So Extensive a Deposit, So Few Outcrops

These eight small outcrop areas appear anomalous given the large area over which debrisite must have been originally deposited. Initially, we anticipated finding the SIL along the entire exposed length of the contact between the Gunflint and Rove Formations. Not so.

The basal 9.8 m of the Rove Formation consists of siltstone and friable shale interspersed with forty-six, 2-15 cm thick, poorly lithified, greenish-gray tuff beds (Maric, 2006). Just a single year of weathering has reduced the tuff beds in exposed drill cores to flaky mush. Two further multi-layered tuffaceous zones occur in the 60 m above this basal zone. Thus, wherever the lower Rove Formation and ejecta-bearing debrisite were exposed, pre-glacial weathering, glacial gouging and post-glacial weathering have removed the easily

eroded lower Rove Formation plus the underlying SIL, leaving topographic depressions, most of which were subsequently post-glacially infilled by lakes, swamps or sediments. Where erosional surfaces have reached the Upper Gunflint Formation chert and chert-carbonate, these more resistant layers have provided a floor to further erosion which persists today. Thus, the few remaining ejecta-bearing debrisite outcrops are small erosional remnants of a once extensive SIL debrisite sheet exposed along the Rove-Gunflint contact.

Fractured Gunflint Formation Bedrock and Gunflint Rip-ups

In Ontario the Upper Member of the Gunflint Formation is a widespread “very complex unit” with “beds of ferruginous carbonate and chert” (Moorehouse and Goodwin, 1960). The chert ranges from chalcedony to microcrystalline quartz. We hypothesize that this surficial or near-surficial Gunflint chert-carbonate was fractured by the powerful earthquakes which arrived in the study area approximately two minutes after impact (Marcus et al., 2000, Earth Impact Effects Program: <http://impact.ese.ic.ac.uk/ImpactEffects/>), providing much of the sharply angular chert breccia subsequently embedded in the overlying debrisite (Fig. 8B). However, other sub centimetre chert clasts embedded in the debrisite are quite rounded, yet their appearance is indistinguishable from Gunflint chert. This suggests that such clasts are Gunflint chert and that they underwent extended travel and abrasion within a ground-hugging density current. These detrital clasts may have been produced by conventional erosional processes but this type of detrital material has never been noted elsewhere in the Upper Gunflint Formation. The Gunflint Formation outcrops sporadically from Thunder Bay to the Slate Islands in Lake Superior 165 km east of Thunder Bay (Sage, 1991). Thus, it is possible for small chert clasts to have traveled as much as 150 km in the violent base surge environment, perhaps producing the rounded clasts from ripped up earthquake-fractured chert.

The Hwy 588 (Fig. 5A) and Terry Fox (Fig. 5B) sites are alone in showing fractured but still in situ Gunflint Formation chert bedrock with ankeritic grainstone deposited in the cracks between the blocks. The depth of this widespread fracturing at the Hwy 588 site is unknown because no visible vertical section is present but, judging from blasted bedrock lining the ditch banks, fractures are estimated to have penetrated as much as 0.5 m based on what is seen in cross-section at the Terry Fox site. The GTP site also shows in situ

fractures outlining rectangular blocks of Gunflint chert-carbonate, but the fractures remain closed and are not infilled. The fractures at these Gunflint sites are smaller in scale than those in Archean granite attributed to Sudbury event earthquakes at Silver Lake, Michigan (3.8r) by Cannon and Schulz (2008). There, Paleoproterozoic sediments were injected into the Archean granite fractures. The shattering of the Hwy 588 chert may well be similar to events at Silver Lake where it is suggested that the dilational phase of seismic waves opened fractures, allowing emplacement of overlying soft sediments into the openings (Cannon and Schulz, 2008). Bedrock fractures in the Barton Creek dolomite at Albion Island, Belize are similarly ascribed to seismic fracturing during the Chicxulub event by Ocampo et al. (1996).

There is no evidence of in situ bedrock fracturing at other sites. However, there is ample evidence of fractured Gunflint carbonate and chert-carbonate and stromatolites in the form of rectangular blocks and clasts within the debrisite, at other sites. These blocks suggest that near-surface Gunflint Formation chert-carbonate and carbonate was seismically fractured and delaminated along bedding planes and that some of this fractured material was ripped up by base surges and incorporated into the debrisite. Similarly deposited angular to sub angular clasts are reported from the Chicxulub event in Belize (Kenkmann and Schönián, 2006).

The lack of alteration rinds on either rounded or angular Gunflint breccia fragments at any site suggests that they were not derived from pre-impact features such as weathered talus. However, alteration rinds typical of weathered surfaces are present on a rounded granite boulder and on an unidentified crystalline cobble at the base of the debrisite at Hillcrest Park and on two crystalline cobbles in the basal shear zone at the Terry Fox site. The presence of weathering rinds on these non-Gunflint clasts supports the view that Gunflint clasts, all of which lack weathering rinds, were derived from freshly earthquake shattered Gunflint bedrock and subsequently ripped up and incorporated into the debrisite by the base surge.

Deposits of Gunflint breccia and lapillistone are found between the Gunflint and Rove Formations at Gunflint Lake, Minnesota, 760 km (5.8r) from Sudbury (Jirsa et al., 2008, 2011). The areal extent of the 7 m thick Gunflint Lake debrisite is much greater than comparable Gunflint breccia zones at the BB and GTP sites at Thunder Bay. It is also nearly twice as

thick as any Thunder Bay site. This contradicts the general westward thinning of the SIL (Addison et al., 2005). If the Gunflint Lake deposits were emplaced by base surges, and the base surges had lost sufficient energy by the time they reached the Gunflint Lake area, the entrained debris could have piled up into thick 'ramparts' as described for end-of-flow Martian base surge deposits (Kenkmann and Schönián, 2006; Osinski, 2006; Mouginiis-Mark and Garbeil, 2007).

Basal Shear Zone

The slickensides seen at the base of the Terry Fox debrisite are similar to "highly chaotic shear planes often connected with polished and striated surfaces..." described for Chicxulub debris by Kenkmann and Schönián (2006) and Wigforss-Lange et al. (2007), and to "a thin basal shear zone" seen in the Stac Fada Member debrisite in Scotland (Amor et al., 2008). The slickenside striae are aligned at an azimuth of 140°, which supports the idea that the slickensides are related to drag shearing during deposition of the overlying fast moving base surge arriving from Sudbury, which lies at an azimuth of 108°. However, a fault just north of the new entrance to the Terry Fox Welcome Centre may also have produced these locally present slickensides.

Peritidal or Subaerial Depositional Environment?

It seems intuitive, with the chaotic SIL resting directly on microbialite mats and stromatolites, that it was deposited in peritidal lagoon environments that were subsequently reworked by tsunamis. As observations accumulated we were forced to re-examine this idea.

Was the Area Subaerial Prior to Debrisite Deposition?

The uppermost 30 m of the Gunflint Formation shows an upward-shoaling succession from thin fine grainstone layers in chemical and clastic mudstone to dominantly grainstone layers to thicker and coarser ripple-laminated and cross-stratified grainstones. These are overlain by the chert-carbonate, stromatolitic limestone, and grainstones, which are erosively terminated by the overlying debrisite. This falling stage sequence suggests that the area was nearly emergent prior to emplacement of the SIL, but it does not show that the area was subaerial at the time of SIL deposition.

Moorehouse and Goodwin (1960) noted that the uppermost Gunflint Formation is composed of a thin, calcite-rich unit which they designated the Limestone

Member. This is the material that directly underlies the debrisite in many locations. It is composed of iron-chlorite grainstone and iron-chlorite-rich layers in stromatolites that are cemented by blocky calcite. This calcitic cement is probably meteoric in origin (Fralick and Burton, 2008). This cement shows 100x vanadium and 10x uranium enrichment relative to lower Gunflint Formation background levels, indicative of a redox front in a subaerial environment (Fralick and Burton, 2008).

Was the Area Subaerial During SIL Deposition?

This is really a question of whether the SIL was deposited by base surges, by tsunamis, or by some combination of the two.

Despite the top of the Gunflint Formation being a gently sloping, recently subaerial shallow lagoon environment with an ocean towards the south, the evidence for these being tsunami-emplaced deposits is weak. We see no evidence of tsunami lithologic couplets created by wave runup and backwash (Nishimura and Miyaji, 1995; Scheffers and Kelletat, 2004; Fujino et al., 2006). Nor is there any evidence of sand or other particle injection into the Gunflint substrate, a feature created by very high dynamic pressures of large tsunamis (Le Roux and Vargas, 2005). On the other hand, like the K/P boundary, these deposits are “sedimentologically complex, differing in architecture and composition from place to place” (Smit et al., 1996). However, in the SIL the complexity is very localized compared to the K/P deposits and it lacks the multi-unit stratigraphy described by Smit et al (1996). Thus, these SIL debrisites are different from the K/P deposits which are ascribed to large tsunamis even though both share some common ejecta.

The heterogeneous distribution of devitrified glass clasts also argues against tsunami deposition of these deposits. These clasts range from tektites (once solid, non vesicular glass) with a density of about 2200 kg/m³ to highly vesicular clasts, some of which may have been able to float on water. With this range in density, some clast sorting based on density would be expected during tsunami deposition, with the highly vesicular clasts being preferentially laid down towards the top of the deposit and with a higher proportion of the least vesicular, denser clasts deposited towards the bottom of the debrisite. This should be especially notable in the waning stages of tsunami wave recession. We observe no evidence of this.

Tsunamis cannot be totally ruled out for the deposits

described here. If such tsunamis were weak and reworked only the topmost freshly deposited debrisite, the record of such could have been erased during postdepositional subaerial exposure (discussed below). But, had a tsunami reworked the top of base surge deposited debrisite, water would presumably have settled into the hot, non-reworked debrisite and we would expect to see fluid or vapour escape pipes. We have seen none, suggesting that even weak tsunamis never reached these deposits. The absence of key tsunami features casts doubt on the SIL debrisite being deposited or reworked by tsunamis.

Base surge deposit features are present. The TF site and Hillcrest Park are the only sites with sufficient thickness to see the structures typical of base surge deposits and the TF site is so heavily overprinted by dolomite recrystallization that structures within it are barely visible. U-shaped channels and massive bedding are among the typical volcanic base surge features (Hattson and Alvarez, 1973; Fisher and Schmincke, 1984; Gencalioglu-Kuşcu et al., 2007; Branney and Brown, 2011). Volcanic base surge beds show distinct upward fining (Fisher and Schmincke, 1984; Dellino et al., 2004). Upward fining in these SIL beds is either absent or, at best, ambiguous, with the notable exception of upward fining of Gunflint clasts in these otherwise chaotic features. Like tsunami deposition, key features of base surge deposits are obscured or missing because of the small scale of the outcrops relative to the scale of the SIL, or the glacial truncation of most outcrops, or, in the case of the TF site, the massive carbonate recrystallization which obscures so much detail.

The question of whether these debrisites were deposited by tsunamis or base surges cannot, at this stage, be unequivocally answered, but the evidence obtained to date supports base surge deposition of the SIL at Thunder Bay. Future work comparing features of the Thunder Bay sites to sites elsewhere in the Lake Superior region may help to answer the question more definitively.

Did the Area Remain Subaerial After Debrisite Deposition?

There is 5-6 m of sediment between the top of the 1850 Ma (Krogh et al., 1984) Sudbury impact event debrisite and dated zircons from three sites in the Rove Formation and one site in the correlative Virginia Formation with an age of 1832 Ma (Addison et al., 2005). This low sedimentation rate for an approximately 18 m.y. period suggests a depositional hiatus. The disconformity at

the debrisite-Rove Formation contact seen at the TF site (Fig. 8F) also supports this view. In addition, the silica stalactites at Hillcrest Park (Fig. 8E) could only have been produced in a subaerial environment. There is a sequence of three lithofacies at the top of the TF debrisite that offers important support for a prolonged period of subaerial exposure: 1) The lowermost of the three lithofacies is the recessively weathering, spherule and DVIG-rich layer. It suggests that a significant but unknown thickness of the carbonate component of the debrisite was leached away, leaving behind a winnowed spherule-DVIG residuum. 2) The 2-8 cm thick layer of agate and chert with upward and downward pointing projections within it, which may have been stalactites and stalagmites before the void was totally infilled by agate and chert, seems to be the product of leaching of overlying material and redeposition at this lower level. 3) The 5-30 cm thick iron-rich alteration profile at the top of the TF site may be a paleosol. It also contains spherules, DVIG, and microtektites. This paleosol hypothesis will remain so until further work tests the idea, but it is in the zone where a paleosol would be expected and the concept is consistent with the other interpretations.

There was a period of subaerial exposure after deposition of the SIL but its duration is unknown. The great mystery is how any unconsolidated debrisite survived such a long period of subaerial exposure.

How Did Any Debrisite Survive?

The reasons for debrisite survival are unknown but the Rove Formation suggests a possible mechanism for debrisite preservation. Its lower 9.8 m contains 46 tuff layers, which decrease in frequency from seven layers per metre at the base of the deposit to zero within that thickness (Maric, 2006). The combination of tuffs found in the Gunflint Formation below the SIL, combined with the 46 tightly spaced tuffs immediately above the SIL in the Rove Formation, suggests tuff deposition may have been ongoing during the depositional hiatus. If so, the tuffs may have borne the brunt of weathering during the period of subaerial exposure rather than the debrisite. The tuffs may also have provided silica leachate which was subsequently deposited as the anastomosing chert and agate throughout the middle to upper SIL seen at Hillcrest Park and at BB and BC sites, thus helping to preserve it.

Summary and Conclusions

Eight SIL outcrops containing ejecta from the 1850

Ma Sudbury impact event have been identified in and near the city of Thunder Bay, Ontario, north of Lake Superior. The SIL was likely deposited by base surges on a subaerially exposed carbonate succession forming the top of the Gunflint Formation. The primary debrisite component by volume is recrystallized carbonate in which Gunflint chert and chert-carbonate breccia and ejecta are embedded. Today, ejecta are a minor component of the total debrisite volume, however, at the time of deposition, it was undoubtedly greater because carbonate replacement has destroyed many features, while recrystallization of carbonate further obscured features. Ejecta features include shocked quartz grains with relict planar features including PDFs and planar fractures, unshocked quartz and feldspar grains, spherules, DVIG clasts, rare microtektites and tektites and accretionary lapilli. Seven of the eight sites have had some portion of the Sudbury impact layer (SIL) removed by glaciation and subsequent weathering. The eighth site near Terry Fox Lookout shows a complete stratigraphic section from the Gunflint Formation, through the SIL and up into the overlying Rove Formation. Disconformities appear at both the base and top of the SIL. The study area has had a complex history, summarized as follows.

1. The upper Gunflint Formation shows an upward fining sequence ending with mafic volcanic ash being deposited and reworked into a carbonate-dominated, nearshore environment supporting microbialite mat growth and stromatolites.
2. Regression of the Gunflint Sea was completed at some unknown time prior to the 1850 Ma Sudbury impact event. Prior to deposition of the SIL, blocky, meteoric calcite cements formed beneath the subaerial surface.
3. Approximately two minutes after the impact, violent earthquakes fractured and delaminated lithified portions of the Upper Gunflint Formation, as evidenced by still in situ fractured rock at the Terry Fox, Hwy 588 and GTP sites.
4. The earthquakes were followed by ground-hugging density currents (base surges) which stripped all unlithified material down to bedrock and ripped up, ground up and entrained some portion of the earthquake-fractured upper Gunflint Formation rock. The base surges then contained the following mixture of features: 1) clasts of fractured carbonate in the fine sand to fine gravel size range; 2) ripped up clasts of Gunflint fractured chert, chert-carbonate and stromatolites; 3) ejecta consisting of

DVIG, spherules, accretionary lapilli, tektites and microtektites, and quartz and feldspar grains and shards, some of which show planar features and PDFs; and 4) small clasts of uncertain origin.

5. The sharply angular nature of most Gunflint chert and chert-carbonate clasts indicates a relatively short travel distance. Slightly rounded chert-carbonate clasts are less common and probably traveled only slightly further from their source than the angular ones. None of these clasts show weathering rinds. Some submillimetre- and millimetre-scale chert clasts are well rounded and could have traveled as much as 165 km from the furthest east Gunflint Formation known today at Slate Islands.
6. Accretionary lapilli formed within the base surges. Some accretionary lapilli passed through zones with varying water vapour concentrations, allowing them to accumulate alternating coarser-grained layers and finer-grained layers. Armored lapilli are also present.
7. The debrisites deposited by these base surges show chaotic patchiness with significant changes in clast sizes and composition over metre and even centimetre distances within the deposits. The one exception to this chaos is an upward fining of Gunflint clasts within the otherwise chaotic debrisite.
8. The SIL was subaerially exposed after deposition as evidenced by anastomosing chert and agate within the debrisite, centimetre-scale agate stalactites in debrisite vugs and the tri-level lithofacies of relict winnowed spherule clusters, agate, and iron-rich alteration profile at the top of the TF site. Silica and carbonate replacement and recrystallization probably began during this period of subaerial exposure. An unknown quantity of debrisite was removed during this period of subaerial exposure.
9. Tuffs deposited on top of the debrisite may have provided sufficient protection to allow survival of some of the debrisite. They could also have provided leachate which led to the extensive deposition of anastomosing chert in the SIL seen at both megascopic and microscopic levels.
10. The Rove Sea then transgressed over the area, first depositing about one metre of carbonate and siltstone before the lower Rove Formation organic-rich mud began accumulating, intercalated with numerous volcanic ash layers.
11. Compaction and carbonate replacement during diagenesis probably continued destroying features in the debrisite. Silica replacement did the same to

a lesser extent. Carbonate recrystallization further destroyed or obscured features.

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Field trip 2 - Geology of the Sibley Peninsula

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Introduction

The Sibley Peninsula extends into Lake Superior, approximately 25 km east of the City of Thunder Bay (Fig. 1). The peninsula, approximately 52 km long and 10 km wide, separates Thunder Bay (the bay of Lake Superior, not the city) on the west from Black Bay on the east. It can be divided into two physiographic units, based largely on bedrock geology. Highlands or tablelands underlain by Mesoproterozoic Midcontinent Rift-related mafic sills intruding the Rove Formation and/or Sibley Group sandstones dominate the area west of Highway 587, rising as much as 380 m above Lake Superior at the Sleeping Giant. East of the highway, flat-lying Sibley Group siltstones and calcareous

sedimentary rocks result in a relatively subdued topography.

This field trip will examine a number of Paleo- and Mesoproterozoic sedimentary and igneous units from the base of the Sibley Peninsula to its tip (c.f., Franklin et al., 1982; Fralick et al., 2000). Most of the stops are road-cuts so caution must be exercised. Do not stand on the paved portion of the road and be aware of vehicular traffic at all times. Sample collecting is not allowed without a collecting permit in Sleeping Giant Provincial Park.

Archean basement below the Sibley Peninsula consists of metavolcanic and intrusive rocks of the

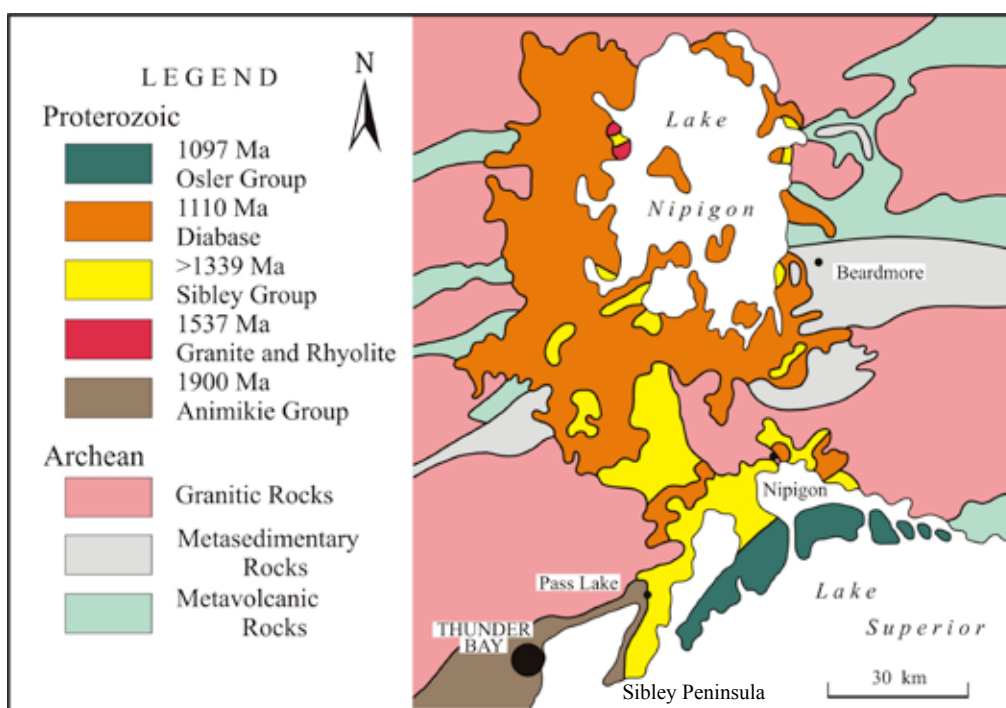


Figure 1. Regional geology east of Thunder Bay including the Sibley Peninsula south of Pass Lake.

Wawa-Abitibi Subprovince. These are nonconformably overlain by the chemical-clastic sedimentary units of the 1878+/-1 Ma Gunflint Formation (Fralick et al., 2002). The Animikie Basin, in which the Gunflint Formation was deposited, developed due to back-arc spreading (Fralick et al., 2002) on the southern margin of Superior Province and forms a southward-thickening wedge sedimented on the shelf during transgressive-regressive-transgressive cycles (Fralick and Barrett, 1995; Pufahl, 1996; Pufahl and Fralick, 2000, 2004). The 1850 Ma (Krogh et al., 1984) Sudbury ejecta layer occurs near, or in places at, the top of the Gunflint Formation. These units will not be seen on this trip, as they form the subcrop. The 1835 Ma Rove Formation (Addison et al., 2005) disconformably overlies the Gunflint Formation. It consists of a lower siltstone dominated unit meters to 10 meters thick. This is overlain by approximately 100 meters of black shale representing a starved succession. The upper eight hundred meters of the formation is dominated by turbiditic, progradational parasequences out-building from distal deltaic bars to the north-northwest represented by lenticular to flaser bedded sandstones at the top of the preserved succession (Maric and Fralick, 2005; Maric, 2006). Sediment was probably derived from the Trans-Hudson Orogeny that was underway to the northwest.

The Sibley Group (Fig. 2) lies disconformably on the Rove Formation, with the directly underlying shales showing the effects of Mesoproterozoic weathering. The age of the Sibley Group is poorly constrained. The best estimate is obtained from its polar wander position, which is the same as the 1450 to 1400 Ma Belt Supergroup (Elston et al., 1993, 2002; Evans et al., 2000). The Belt, which was deposited on the western side of North America, also records the same climatic fluctuations as the Sibley (Rogala et al., 2007). The Sibley basin originally developed as a down-sag accumulating conglomerates and sandstones filling topographic lows and then expanding to cover the area with sheet sandstones. The main sediment source was from the northwest, with Paleoproterozoic zircons dominating this population (Rogala et al., 2007). This indicates a probable source from the eroding Trans-Hudson highlands. Lacustrine conditions developed throughout the area, with the lake becoming more saline with time. As the lake shrank, strand-line stromatolitic dolostone was deposited with a sub-aerial weathered upper surface and terra rosa (soil) development in places. Next a period of basin instability occurs as it down-tilts to the north, the basement collapses into

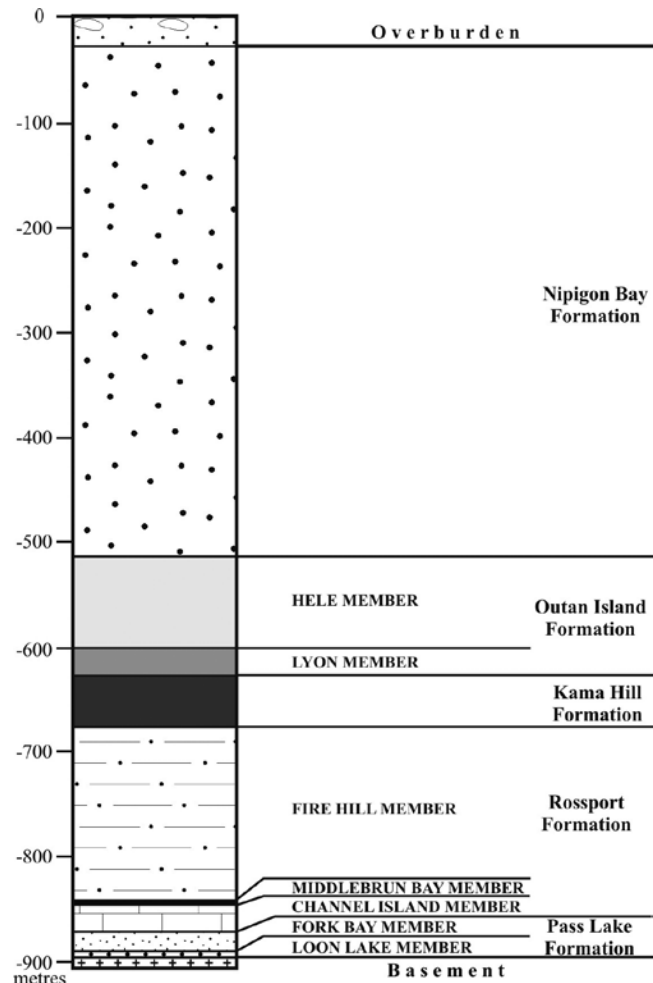


Figure 2. Stratigraphy of the Mesoproterozoic Sibley Group (From Rogala et al., 2007).

a half-graben and within the basin, sub-aerial mass-flows are generated (Rogala, 2003; Rogala et al., 2005, 2007). Sediment feed is now from the southwest, comprising considerable Mesoproterozoic zircons. This indicates that the Penokean area was not a barrier to sediment transport from the south. The above is the highest stratigraphic unit we will see on this trip. The descriptions below include overlying units.

The general geology of the Sibley Peninsula was most recently summarized by Carl (2011):

Gunflint Formation

Gunflint Formation sedimentary rocks make up a chemical-clastic assemblage whose upper portion was deposited 1878Ma (Fralick et al., 2002). This formation crops out close to the northern limits of the Sibley peninsula near Pass Lake, Ontario. At this site, rare folding is present in Gunflint sedimentary layers. This

folding is thought to be related to fold-and-thrust belt deformation caused by Penokean compression (Hill and Smyk, 2005). Despite the presence of these compression-related folds, most Animikie Group sedimentary rocks in Ontario are undeformed (Sutcliffe, 1991). Animikie Group sedimentary rocks in Ontario have been classified as having a sub-greenschist metamorphic grade (Easton, 2000) and are frequently considered to be unmetamorphosed for convenience of interpreting depositional environments.

Rove Formation

The Gunflint Formation was once thought to transition conformably into the overlying Rove Formation; however, a disconformity has been recognized between these two assemblages (Schulz and Cannon, 2007). The idea that the Gunflint and Rove Formations are discontinuous first became apparent when it was proposed that the Sudbury impact which occurred 1850Ma was a subaerial occurrence (Addison et al., 2005) in the Thunder Bay area (P. Fralick, pers. comm., 2011). This suggests there was a period during which no deposition was occurring that was coeval with the Sudbury impact (Schulz and Cannon, 2007). Using a volcanic ash layer near the base of the Rove Formation, Addison et al. (2005) determined an age of 1836Ma for basal Rove shales. This indicates deposition in the Animikie basin had resumed, and water was again present in this basin at 1836Ma. In the vicinity of the Sibley Peninsula, the Rove Formation has a shallow southeasterly dip (Horton, 1989) and a thickness greater than ~610m based on a diamond drill hole at Sibley Bay (Geul, 1973). Due to erosion, only the lower part of the Rove Formation can be found on the southern portion of the Sibley Peninsula. This basal portion of the Rove Formation consists mostly of black shales with interbeds of siltstone (Maric and Fralick, 2005).

Rove Formation shales sporadically outcrop on the western shoreline of the peninsula from Sawyer Bay northwards and are never more than a few metres above the 183m elevation of Lake Superior. Also on the peninsula's western shoreline, well-preserved Rove shale concretions can be seen next to the Kabeyun Trail between Clavet Bay and Hoorigan Bay. Similar Rove Formation sedimentary rocks are lacking on the

northeastern shores of the Sibley Peninsula due to the southeasterly dip direction of these rocks. On the southern tip of the Sibley peninsula, south of Perry Bay and Sawyer Bay, the Rove Formation is well represented. Here, sedimentary layers, consisting mostly of black shales, are present beneath the Sleeping Giant landform at elevations up to 369m. The occurrence of Rove Formation sedimentary rocks at such high elevations, when the south-easterly dip of this unit should cause it to be beneath the surface on the peninsula's southern tip, can be explained by the Silver Islet fault described by William Logan as a transverse dislocation that lets down the succeeding formation by several hundred feet (Logan, 1847). This fault displaced Animikie Group sedimentary rocks as well as other assemblages and played a key role in the genesis of ore at the historic Silver Islet mine site (Horton, 1989). Eventually, the long-lived Animikie Basin closed and deposition of Rove Formation sedimentary rocks ceased. This resulted in a gap in the rock record on the Sibley Peninsula and allowed for erosion of Animikie Group sedimentary rocks.

Sibley Group Sedimentary Rocks

Large-scale subsidence following doming related to the intrusion of the approximately 1540Ma Mesoproterozoic English Bay Complex (Hollings et al., 2004) created an ovoid depression known as the Sibley Basin (Rogala, 2003). Located to the north of the Sibley Peninsula, the English Bay Complex is a granite-rhyolite assemblage (Hollings et al., 2004) with an age of 1537Ma (Davis and Sutcliffe, 1985). The formation of the Sibley Basin post dates igneous activity of the English Bay Complex, with deposition in this basin probably beginning slightly before 1500 Ma (Rogala, 2003). Sediments deposited in the Sibley Basin comprise the Sibley Group, a relatively flat, unmetamorphosed assemblage divided into five distinct formations (Franklin et al., 1980). The lowest three formations of the Sibley Group are found in abundance throughout Sleeping Giant Provincial Park and make up the majority of the surficial geology of the Sibley Peninsula.

Pass Lake Formation

The oldest Sibley Group formation is the Pass Lake Formation which itself consists of two members: the Loon Lake Member and the Fork Bay Member (Cheadle, 1986). The Loon

Lake member is the lowermost assemblage of the Sibley Group and consists of conglomerate lenses that were deposited in depressions caused by erosion of the underlying rock (Franklin et al., 1980). This unit can be seen on the Sibley Peninsula directly adjacent to Pass Lake where it is in contact with the overlying sandstones of the Fork Bay Member. Cliff faces adjacent to Pass Lake are dominated by Fork Bay Member plane bedded sandstones which are also commonly seen in cliff faces close to the western shores of the Sibley Peninsula. In general, Fork Bay Member sedimentary rocks comprise sandstones which can be massive and well sorted, massive and poorly sorted, silty, laminated or rippled (Rogala et al., 2005). These sandstones are thought to have been deposited in a shallow, quiet, lacustrine environment (Franklin et al., 1980).

Rosspport Formation

Fork Bay Member sandstones are overlain by the Rosspport Formation, which consists of three members and ten facies associations. The Channel Island Member is the lowermost assemblage of the Rosspport Formation and consists of dolomitic mudstones, which gradually increase in abundance as the Pass Lake Formation transitions into the Rosspport Formation (Rogala, 2003). Many of the ten facies associations of the Rosspport Formation contain siltstones and carbonates which provide clues to the types of environments that once existed when many rocks of the Sibley Peninsula formed. A unique cyclic siltstone-dolomite lithofacies association is present in the Rosspport Formation, and has been interpreted to represent a shallow offshore environment where muds were deposited during wet periods and dolomite precipitated during periods of drought (Fralick et al., 2000). This suggests that the lake(s) in which sands of the Pass Lake Formation were deposited became progressively more saline and ultimately formed playa lakes (Rogala et al., 2005). This interpretation appears to be consistent with unpublished paleomagnetic results of G. Borradaile, which suggest at the time of sediment deposition, the Sibley Basin was near the Earth's equator where arid conditions would have dominated (Fralick et al., 2000). Directly above the cyclic siltstone-dolomite layers, dolomitic mudstones containing mud cracks and gypsum nodules are present. The occurrence of these

evaporites has been interpreted to represent a clastic sabkha environment (Fralick et al., 2000). The dolomitic mudstones are at times overlain by sporadic stromatolitic chert-carbonate lithofacies which are indicative of shallow water, near-shore environments. These laterally discontinuous stromatolitic facies are part of the Middlebrun Bay Member and may represent migrating shorelines of partially restricted bays (Rogala, 2003). Above the Middlebrun Bay Member is the Fire Hill Member which is the uppermost member of the Rosspport Formation. The Fire Hill Member can be extremely difficult to distinguish from the Channel Island Member (Rogala et al., 2005). For this reason, the previously characterized member assemblages (Cheadle, 1986) are now described as lithofacies associations (Rogala et al., 2005). The uppermost of these associations are primarily composed of a variety of conglomerates, sandstones, siltstones and mudstones. These uppermost sedimentary rocks were deposited onto a mudflat, with coarser, unsorted sediments and mud-chip conglomerates representing debris flows and slumping events (Rogala, 2003). Siltstones can be found at the top of the Rosspport Formation which grade into the overlying Kama Hill Formation.

Kama Hill Formation

The Kama Hill Formation is divided into four lithofacies associations. The first three consist of fine-grained sandstones and siltstones that are horizontally laminated, mud-cracked and rippled (Rogala et al., 2005). Horizontally laminated mudstones often cap these fine-grained sandstone and siltstone units. These four lithofacies associations are thought to represent a floodplain system which periodically contained ponds. Thicker units of fine-grained sedimentary rocks which sometimes contain wave ripples, likely indicate the presence of long-lasting ponds (Rogala, 2003). Flooding events which covered the floodplains probably increased in occurrence until the formation of the subaqueous Outan Island Formation (Rogala et al., 2005).

Outan Island and Nipigon Bay Formations

Although not present on the Sibley Peninsula, the Outan Island and Nipigon Bay Formations represent important final stages in the depositional history of the Sibley Basin. The Outan Island Formation consists of mudstone, laminated

sandstone/mudstone, siltstone, sandstone and conglomerate lithofacies associations (Rogala, 2003). The lower part of the Outan Island Formation has been interpreted as representing a deltaic environment, whereas its upper part represents a fluvial environment (Rogala, 2005).

As described by Rogala (2003), the Nipigon Bay Formation is the uppermost formation in the Sibley Group that consists of a cross-stratified sandstone lithofacies association and a horizontally laminated sandstone lithofacies association. These associations are thought to represent an ancient aeolian environment. This interpretation is supported by the high degree of sediment sorting seen in sandstones, as well as the presence of large-scale dune topography. The Nipigon Bay Formation was likely subjected to a semiarid to arid climatic regime and probably resembled a modern day desert.

The Sibley Group has a minimum depositional age of 1339 Ma based on an Rb-Sr isochron constructed by Franklin (1978). This age is presently the youngest depositional age determined for the Sibley Group, with final deposition in the Sibley Basin probably occurring shortly after this date. The final formation of Sibley Group sedimentary rocks signaled the beginning of a roughly 200 million year hiatus in rock formation on the Sibley Peninsula.

An expanded description of the lower Sibley Group was presented by Metsaranta (2006) in his M.Sc. thesis on the sedimentology, geochemistry and paleohydrology of these rock units:

Based on the analysis of lithofacies associations and stratigraphy, the following conclusions can be made:

1. The portions of the Sibley Group studied (lithostratigraphic Pass Lake and Rossport Formations) contain a variety of distinct lithofacies associations. These lithofacies associations can be divided into 4 informally defined allostratigraphic units which roughly correspond to existing lithostratigraphic subdivisions.

2. The lower clastic unit forms the base of the Sibley Group and contains the following lithofacies associations representing distinct depositional settings: boulder conglomerate-sandstone-calcrete (proximal ephemeral braided

streams), pebble to cobble conglomerate (ephemeral braided streams), trough cross-stratified sandstone (ephemeral braided streams), massive cobble conglomerate (transgressive lag, reworking of braided stream deposits during lacustrine transgression), green sandstone-siltstone (wave and storm influenced fluvial dominated deltas), planar cross-stratified sandstone (nearshore migration of large sandwaves), and thinning upward sandstone (beach and storm remobilized nearshore sandstone sheets).

3. The mixed siliciclastic-carbonate unit disconformably to conformably overlies the lower clastic unit and consists of the following lithofacies associations: red siltstone (non-saline lake), red siltstone-dolostone (perennial saline lake, distal from clastic sources) and red siltstone-dolomitic sandstone (perennial saline lake, proximal to clastic sources).

4. The upper clastic unit sharply overlies the mixed siliciclastic carbonate unit and consists of the sheet sandstone lithofacies association (ephemeral playa lake (?) or perennial lake with increased clastic supply with respect to underlying units), and the black chert-carbonate lithofacies association (shoreline). Subaerial exposure features are present at the top of the black-chert-carbonate lithofacies association and include the intraformational conglomerate lithofacies association (subaerial debris flows, intrusive and/or extrusive sedimentary breccias, terra rossa style soils, dissolution collapse breccias).

5. The mixed siliciclastic-carbonate-evaporite unit overlies the subaerial exposure surface at the top of the upper clastic unit. It consists of the massive dolostone (saline lake), the red siltstone-sulfate (wet evaporate-rich mudflats around lake margins) and the fine-grained sandstone (dry, evaporate poor mud and sand flats around lake margins) lithofacies associations.

Based on the stable isotope, Sr isotope and trace element data, the following conclusions can be made:

1. Overall, the geochemical data supports a non-marine origin for the Pass Lake and Rossport Formations.

2. Low Sr isotope ratios from calcrete in the

lower clastic unit suggest atmospheric deposition and weathering of Gunflint Formation carbonate bedrock was the primary source of cations for pedogenic carbonate rather than weathering of local silicate sources. Relatively ^{13}C -rich calcrete carbon isotopic composition suggests little organic contributions to soil CO_2 . REE geochemistry suggests calcretes precipitated from oxidizing non-marine water.

3. S, Sr, REE and Y data for the mixed siliciclastic carbonate unit support a lacustrine origin for these rocks. Variations in S isotopic composition may be related to changes in the composition of sulfides weathering to supply sulfate to the system. MREE enriched PAAS normalized REE patterns for dolostone samples differ from those found in other carbonate lithofacies and this probably relates to more reducing conditions in lake waters relative to surface waters supplying the lake. Stratigraphic variations in C and O for this unit were created by evaporation and/or residence time effects.

4. Slightly enriched $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values in stromatolitic units in both the upper siliciclastic unit and mixed siliciclastic-carbon-evaporite unit reflect a generally more arid evaporitic environment as compared to the mixed-siliciclastic unit. Shifts toward lighter $\delta^{13}\text{C}$ in pedogenic carbonates from these units probably reflect a contribution of dissolved organic carbon. REE data for these units is consistent with a non-marine, oxidizing depositional setting.

Carl (2011) continued to describe the igneous units on Sibley Peninsula thus:

Midcontinent Rift

At approximately 1.15 Ga, early stage mafic magmatism of the Midcontinent Rift (MCR) began (Heaman et al., 2007). Centered around Lake Superior, the MCR contains over one million cubic kilometres of mafic volcanic and plutonic rock (Klewin and Shirey, 1992). These rocks formed as the result of a major continental rifting episode (Shirey et al., 1994) that spanned at least 60 million years (Heaman et al., 2007). During this time, rifting of the Superior craton nearly resulted in the splitting of the North American continent and the formation of an ocean. This rifting event is mostly represented by flood basalts, which dominated the igneous

activity of the MCR and reached thicknesses in excess of 20km (Cannon et al., 1989). These basalts are generally tholeiitic in composition and are thought to be plume-sourced (Klewin and Shirey, 1992). Large mafic igneous bodies associated with the MCR are predominantly attributed to the upwelling of a mantle plume beneath the North American Continent (Burke and Dewey, 1973; Hollings et al., 2010a), with the most primitive magmas associated with the MCR being emplaced early in the rift's history (Hart and MacDonald, 2007). This upwelling plume was responsible for the formation of numerous volcanic and intrusive units. The Logan and Nipigon Sills, Osler volcanic rocks, and what some have speculated to be Pigeon River dikes (Sutcliffe, 1991) dominate the MCR exposures proximal to the Sibley Peninsula.

Logan Igneous Suite

The Logan Sills and Nipigon Sills are part of what is now known as the Logan Igneous Suite (LIS; Hollings et al., 2007a). The LIS contains various diabase sill formations located north of Lake Superior (Hollings et al., 2007a). Using geochemical data, sills to the north of Thunder Bay have been classified as Nipigon Sills, with sills south of Thunder Bay referred to as Logan Sills (Hollings et al., 2007a). The diabase sill exposures on the Sibley Peninsula occur on the peninsula's southern tip and make up the Sleeping Giant and Thunder Mountain landforms. For convenience, this sill will henceforth be referred to as the Sleeping Giant Sill (SGS). The SGS is located due east of Thunder Bay and has not been the subject of geochemical analysis in the past.

Logan Sills

Logan sills, such as those seen on Mt. McKay, commonly appear as mesas of the Nor'Wester Mountains found immediately south of Thunder Bay. These sills concordantly intrude sedimentary layers of the Rove Formation and have a gentle southwest dip (Sutcliffe, 1991). The Logan Sills have been classified as quartz-tholeiitic diabase that often contains labradorite, augite, pigeonite and iron-titanium oxides (Sutcliffe, 1991). Logan Sills are characterized by high TiO_2 and Gd/Yb_n when compared to Pigeon River dikes and Nipigon Sills (Hollings et al., 2010a). In hand sample, Logan Sills contain medium to coarse grains that are dark grey and ophitic-textured. Heaman et

al. (2007) determined an age of approximately 1115 Ma for the sill that caps Mt. McKay, which presently serves as the only reliable dating of a Logan Sill (Heaman et al., 2007). Hollings et al. (2010a) reported small Nb anomalies and fractionated REE patterns in samples from Logan Sills. These characteristics along with elevated TiO₂ values and low Mg# values are the best traits for distinguishing Logan Sills from other nearby sill suites (Smyk and Hollings, 2009).

Nipigon Sills

Located in the Nipigon Embayment (Sutcliffe, 1991), the Nipigon Sills represent a northern component of the MCR that may have formed in a failed arm of the rift (Richardson and Hollings, 2005). These sills may be up to 200m thick and intrude all other rocks in the area (Heaman et al., 2007). The Nipigon Sills are diabase comprised of medium-to-coarse-grained, lath-shaped, euhedral crystals of ophitic-textured plagioclase with abundant pyroxene as well as trace olivine and magnetite (Hart et al., 2005). The sills of the Nipigon Embayment were sometimes broadly referred to as Nipigon Sills and have recently been subdivided into five distinct sill suites based on geochemical analysis (Hollings et al., 2007a). These suites are the mafic Nipigon, Inspiration and McIntyre Sills and the ultramafic to mafic Jackfish and Shillabear Sills (Hollings et al., 2007a). The SGS is a mafic sill, and therefore only the mafic Nipigon, Inspiration, and McIntyre sills will be considered here. The Nipigon and Inspiration sills both have low La/Sm_n ratios with the Inspiration sills having elevated Gd/Yb_n ratios compared to the Nipigon sills (Hollings et al., 2007a). The McIntyre sills are distinguished by their low La/Sm_n ratios and intermediate Gd/Yb_n ratios compared to other sill suites (Hollings et al., 2007a). These suites can also be differentiated by plotting Mg# versus TiO₂ with the McIntyre sills having TiO₂ values elevated similarly to the TiO₂ values recorded for Logan Sills (Fig. 2.3) near Thunder Bay (Hollings et al., 2007a). Sills of the Nipigon Embayment range in age from roughly 1106 Ma to perhaps as much as 1159 Ma and are considered to be amongst the oldest magmatic expressions of the MCR (Heaman et al., 2007).

Pigeon River Dikes

The Pigeon River Dikes occur as a northeast trending swarm that formed as a result of MCR

magmatic activity. This dike swarm is located on the northern shore of Lake Superior from Grand Portage, Minnesota, to the Black Bay Peninsula in Ontario (Osmani, 1991). Pigeon River Dikes can either cross-cut Logan Sills or terminate against these sills. Geochemical data suggest the Pigeon River Dikes could not have acted as feeders to the Logan Sills (Hollings et al., 2010a). Ages of Pigeon River Dikes range from 1078Ma for a dike near Arrow River in Devon Township to 1141Ma for a dike in Crooks Township (Heaman et al., 2007). This span of over 60 million years is appreciably longer than the duration of formation for most plume-derived large igneous provinces (Hollings et al., 2010a). Pigeon River Dikes have an olivine-tholeiitic composition (Sutcliffe, 1991) and can be distinguished from Logan Sills based on their low TiO₂ and Gd/Yb_n values (Hollings et al., 2010a). These values are broadly similar to the Nipigon Sills on Gd/Yb_n versus La/Sm_n and Mg# versus TiO₂ diagrams (Hollings et al., 2010a). Pigeon River Dikes generally have northeasterly strikes and very steep dips to the south (Osmani, 1991). Pigeon River Dikes have been cross-cut by the geochemically distinct Cloud River Dikes located south of Thunder Bay (Smyk and Hollings, 2007). On the Sibley Peninsula, numerous dikes having northeasterly strikes are present (Tanton, 1924). Dikes of the Sibley Peninsula have been shown to have both normal and reversed polarities suggesting a range of ages for these dikes (Pesonen and Halls, 1983).

Osler Group

The Osler Group consists primarily of mafic rocks found along the north shore of Lake Superior, representative of basaltic flows that occurred approximately 1108-1105 Ma (Hollings et al., 2007b). These basaltic flows reached a thickness of approximately 3 km and can be frequently found in outcrop to the northeast of the Sibley Peninsula. Osler Group basalts have chondrite-normalized La/Sm_n ratios that range from 1.5 to 3.9 and chondrite-normalized Gd/Yb_n ratios ranging from 1.5 to 3.7 (Hollings et al., 2007b). Osler Group volcanic rocks are abundant on the Black Bay Peninsula, located east of the Sibley Peninsula. No volcanic rocks of any kind have been observed on the Sibley Peninsula.

Road log for field trip (NAD83, UTM Zone 16)

STOP NAME	STOP NO.	LANDMARK (0 Km)	DISTANCE (km)	NORTHING	EASTING
		Junction Highways 11-17 and 587	0.0		
Blende Creek					
Area: Gunflint Fm chert-carbonate	1A		0.6	5384392	369112
Gunflint Fm chert-carbonate	1B		1.4	5383837	369581
		CNR Trestle	2.1		
Rove Fm concretions	2		3.5	5382426	370841
		Watson Site (turn-off)	3.8		
Watson Site	3	n.a.		5382460	371737
Kettle	optional		4.8	5381279	371241
Brohm Site	optional		5.2	5380897	371236
Pass Lake (Animikie / Sibley disconformity)	4A		6.0	5380509	371853
Pass Lake (basal conglomerate)	4B		6.5	5380537	372294
Rosspport Fm. Siltstones	5		7.7	5380075	373001
Rosspport Fm. Lacustrine Sheet Sandstones	6		8.8	5378948	372877
Rosspport Fm., Lacustrine Channel Sandstones	7A		9.9	5377943	372842
Rosspport Fm., Lacustrine Channel Sandstones	7B		10.1	5377728	372868
		Pass Lake Crossroad	10.6		
Rosspport Fm., Subaerial Siltstone-Caliche	8		10.9	5376859	373095
		Jakobsen Road	11.2		

**Northwest-striking
dykes in RosSPORT
Fm.**

9A		14.2	5374104	372312
9B		14.3	5374052	372320
	Sleeping Giant Provincial Park boundary	14.4		
	Kay Lake	16.9		
	Portage Drive	17.2		
	Joe Creek trail	17.8		
	Joeboy Lake	19.5		
	Thunder Bay Lookout turn- off	21.1		

**East-northeast-
striking dyke in
RosSPORT Fm**

10		21.8	5368080	371382
	east- northeast- striking dyke	22.6	5367443	370954
	east- northeast- striking dyke	23.8	5366398	370595
	east-striking dyke	23.9	5366287	370537
	Sifting Lake trailhead	24.2		
	northeast- striking dyke	24.6	5365748	370280
	Lake Marie Louise (north end)	29.8		
	northeast- striking dyke	34.1	5357977	366961
	Lake Marie Louise campground	34.2		
	Plantain Lake trailhead	34.8		
	Silver Islet loop junction	35.7		
	Sawyer Bay trailhead	37.3		
	east- northeast- striking dyke	37.5	5355569	364889
	Silver Islet General Store	38.5		

Rove Fm and east-northeast- striking- diabase dyke	11	39.0	5354955	365488
east-northeast- trending dyke in Rove Fm. / Silver Islet view	12	39.9	5355427	366232
Sibley Creek (bridge)		40.0		
Loop T- junction		40.3		
Middlebrun Bay trailhead		40.7		
Loop T- junction		42.1		



Figure 3. Stop locations, northern Sibley Peninsula; north to left of image from Google Earth

Stop Descriptions

Stops 1A,B: Blende Creek area (Gunflint Fm.)

**UTM coordinates: NAD83; 16U A - 0369112E / 5384392N,
B - 0369581E / 5383837N**

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 grainstones weathering orangey-brown alternating with white chert layers (Fig. 4). In places the chert can be seen



Figure 4. Folded and faulted Gunflint strata on Highway 588, Stop 1B.

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 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 (Fig. 4). 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.

Recent examination of the Gunflint Formation near Pass Lake has led to the recognition of structures typical of Penokean (circa 1875 to 1835 Ma) fold-and-thrust belt deformation (Hill and Smyk 2005). Discrete bedding-plane faults with locally developed gouge and breccia can be traced laterally into horizontal, hanging-wall ramps with associated fault-bend folding. Fold-and-thrust belt deformation is caused by regional

compression. Previous workers had ascribed the folds to syn-sedimentary slumping and Keweenawan diabase sill emplacement and thought that they were attributable to local, rather than regional-scale, deformation. Displacement in fold-and-thrust belts tends to be localized along discrete bedding planes and not easily recognized. This may account for the perceived lack or absence of structures elsewhere in the Gunflint Formation (Hill and Smyk, 2005).

Penokean structures on the northern side of Lake Superior represent the northward migration of thrust faults into the foreland (passive margin Archean basement + Gunflint Formation) caused by hinterland collision to the south.

Stop 2: “Devil’s Flower Pots” (Rove Formation concretions)

UTM coordinates: NAD83; 16U 0370841E / 5382426N

Just north of Highway 587, a quarry face exposure of black, fissile Rove Formation shale displays lenticular and elliptical concretions, flattened along bedding planes (Fig. 5). 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



Figure 5. Lenticular concretion in Rove Formation shale, old quarry face, north of Highway 587, Stop 2. Concretion is almost 1 m in diameter.

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 from 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 $((\text{Mg},\text{Al})(\text{Fe},\text{Al})_4(\text{SO}_4)_6(\text{OH})_2 \cdot 20\text{H}_2\text{O})$; Resident Geologist's Files, Thunder Bay).



Figure 6. 1835 Ma Rove shale, note bleached zone overlying oxidized zone, overlain by the approximately 1450 Ma basal sandstones and conglomerates of the Loon Lake Member, Pass Lake Formation, Sibley Group, Stop 3.

Stop 3: Watson Road section (Pass Lake and Rove formations) - private property; permission is required to access

UTM coordinates: NAD83; 16U 0371737E / 5382460N

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.

The Rove shales immediately below the contact were subject to Mesoproterozoic weathering (Fig. 6).



Figure 7. Disrupted zone in the Rove shales underlying the Sibley Group, Stop 3. The origin of this structure is enigmatic, but may have been caused by fluid escape



Figure 8. Close-up of conglomerate shale contact, Stop 3. Note the strong oxidation of both units.

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 (Fig. 7). Very immature, iron oxide-rich conglomerates and sandstones of the Loon Lake Member, Pass Lake



Figure 9. Typical Loon Lake Member coarse-grained sediments at Stop 3. From the bottom to the top of the photograph: 1) matrix-supported conglomerate, probably a high-density mass-flow; 2) a one clast-thick, pebble-cobble lag developed at the top of this conglomerate through erosion. This may represent either an Aeolian deflation lag or one developed by water erosion of the fine-grained fraction. 3) a boulder-cobble, matrix-supported, mass-flow conglomerate. This was probably a very high-viscosity flow as the larger clasts were suspended near the top of the flow. 5) an upper flow regime parallel-laminated sandstone probably deposited by sheet-flood on the alluvial fans surface; 6) a clast-supported fluvial conglomerate.

Formation, overlies the Rove (Fig. 8). The conglomerate and sandstone layers are laterally discontinuous (Fig. 9), 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 the erosion of underlying units. This is sharply overlain by mature, well-sorted, medium-grained sandstones of the Fork Bay Member, Pass Lake Formation (Fig. 10). 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 (Fig. 11). Rare, odd features are present both in cross-sectional and bedding plane views in this outcrop (Figs. 12, 13). These may be dewatering pipes.



Figure 10. Sharp contact between the hematite-rich conglomerates of the Loon Lake Member and the well-sorted, buff sandstones of the Fork Bay Member, Pass Lake Formation, Sibley Group, Stop 3.



Figure 11. Medium- to coarse-grained, well-sorted sandstone bed of the Fork Bay Member, Stop 3. The majority of the bed is upper flow regime parallel laminated, with a reworked, cross-stratified top.

Stops (Optional): Kettle and Brohm Archaeological Site

UTM coordinates: NAD83; 16U 0371241E / 5381279N and 371236E / 5380897N respectively

Local archaeological sites are closely tied to paleo-shorelines, especially those associated with Lake Minong. When flooded to Minong levels (Fig. 14),



Figure 12. Bedding-plane view of odd concentric layering, Stop 3. The pattern is caused by erosion of the top of domed layers. These may be water escape structures with the basal portions of sand volcanoes preserved, or not.

Sibley Peninsula becomes a virtual island, connected to the mainland near Pass Lake with a series of baymouth bars, forming a spit between the sandstone cliffs (Geddes et al., 1987). The location of these sites may relate to the importance of this paleogeography in constricting the movement of caribou and other animals from the peninsula to the mainland. Although there is no organic preservation to confirm that these were caribou ambush and processing sites, it remains a compelling theory.



Figure 13. Cross-section through a domal structure in a Fork Bay sandstone, Stop 3.

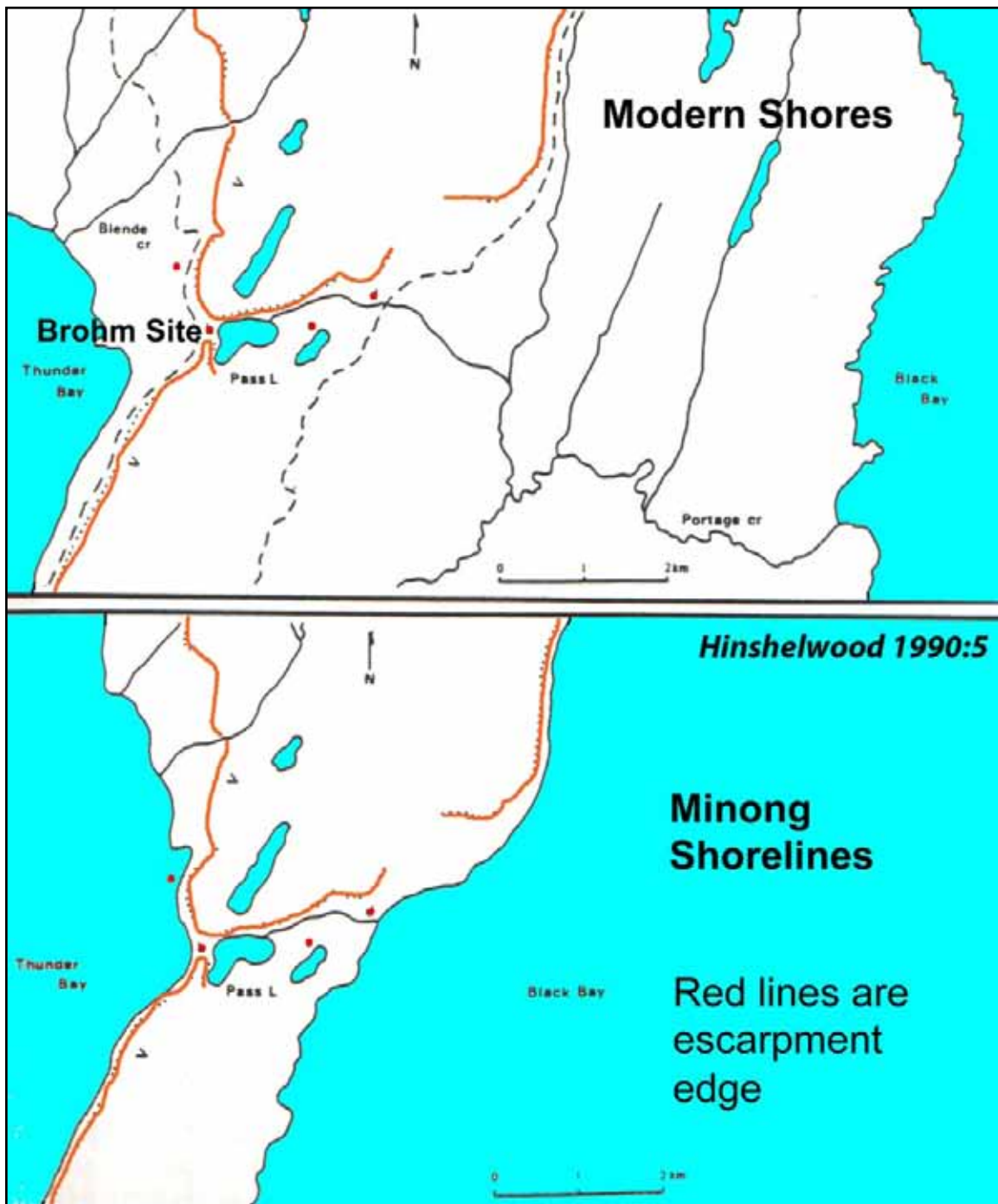


Figure 14. Reconstruction of shoreline near Pass Lake during Minong time (Hinshelwood 1990)

A historic plaque on a roadside pull-off describes the archaeological discovery:

In 1950, archeological investigations in this area uncovered a site which had been used as a workshop camp by a group of the earliest known people in this part of the Upper Great Lakes basin. Called Aqua-Plano Indians because they migrated from the western plains to fossil beaches of glacial and post-glacial lakes in

this region, they appeared about 9,000 years ago following the retreat of glaciers and the northward movement of plants and animals. They developed a distinctive tradition based primarily on large game hunting using weapons and specialized tools made of taconite, a stone that was obtained locally. Their way of life, which was closely related to the environment, disappeared as the climate grew warmer.

(http://www.ontarioplaques.com/Plaques_STU/Plaque_ThunderBay17.html)

This Quaternary geology of this area was also described by Geddes et al. (1987; Fig. 15 and 16):

[Figure 15] shows the arrangement of bars which separated Pass Lake from open water about 9000 years B.P. The strait between Black Bay and Thunder Bay appears to have followed a subglacial channel scoured into the rock floor, and which forms a narrow trough along the length of Pass Lake. Geomorphological evidence of the environmental conditions at the time of habitation will be seen in gravel pit sections of the baymouth bars. Cryoturbated layers show a marked vertical alignment of platy shale fragments, suggesting that the bar surfaces were exposed to intensely cold conditions as they accumulated. The grounding of small icebergs, as evidenced by the reorientation of bar features around depressions is also indicated ["Kettle" Stop].

Pass Lake section

Stop 4A: Disconformity between Pass Lake and Rove Fm.'s (UTM 371853E / 5380509N)

UTM coordinates: NAD83; 16U 0371853E / 5380509N

The rock cut adjacent to the railway at Pass Lake provides an excellent exposure of several types of flat-lying sedimentary rocks and their contact relationships. This cliff is the type section for the Pass Lake Formation of the Sibley Group. Exposure is almost continuous for about 3.2 km along the tracks and gives a stratigraphic thickness of 50 m.

The oldest rocks in the section, fine-grained, fissile shales of the Rove Formation (Animikie Group), are exposed at the northwestern end of the cliff exposure. Originally black, they were oxidized in pre-Sibley times, resulting in their purple and green colour. These rocks originated as 1835 Ma muds.

The Animikie Group rocks are disconformably overlain by coarser-grained sedimentary rocks of the Sibley Group, similar to what we observed at the

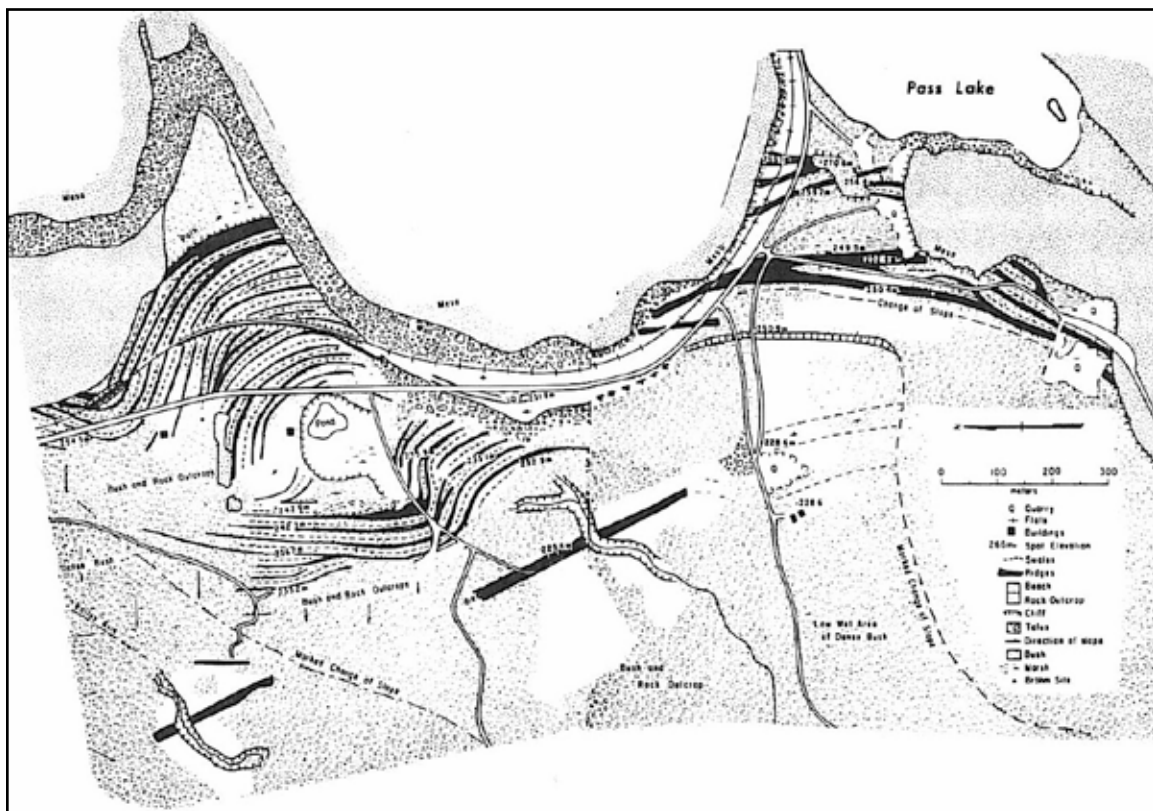
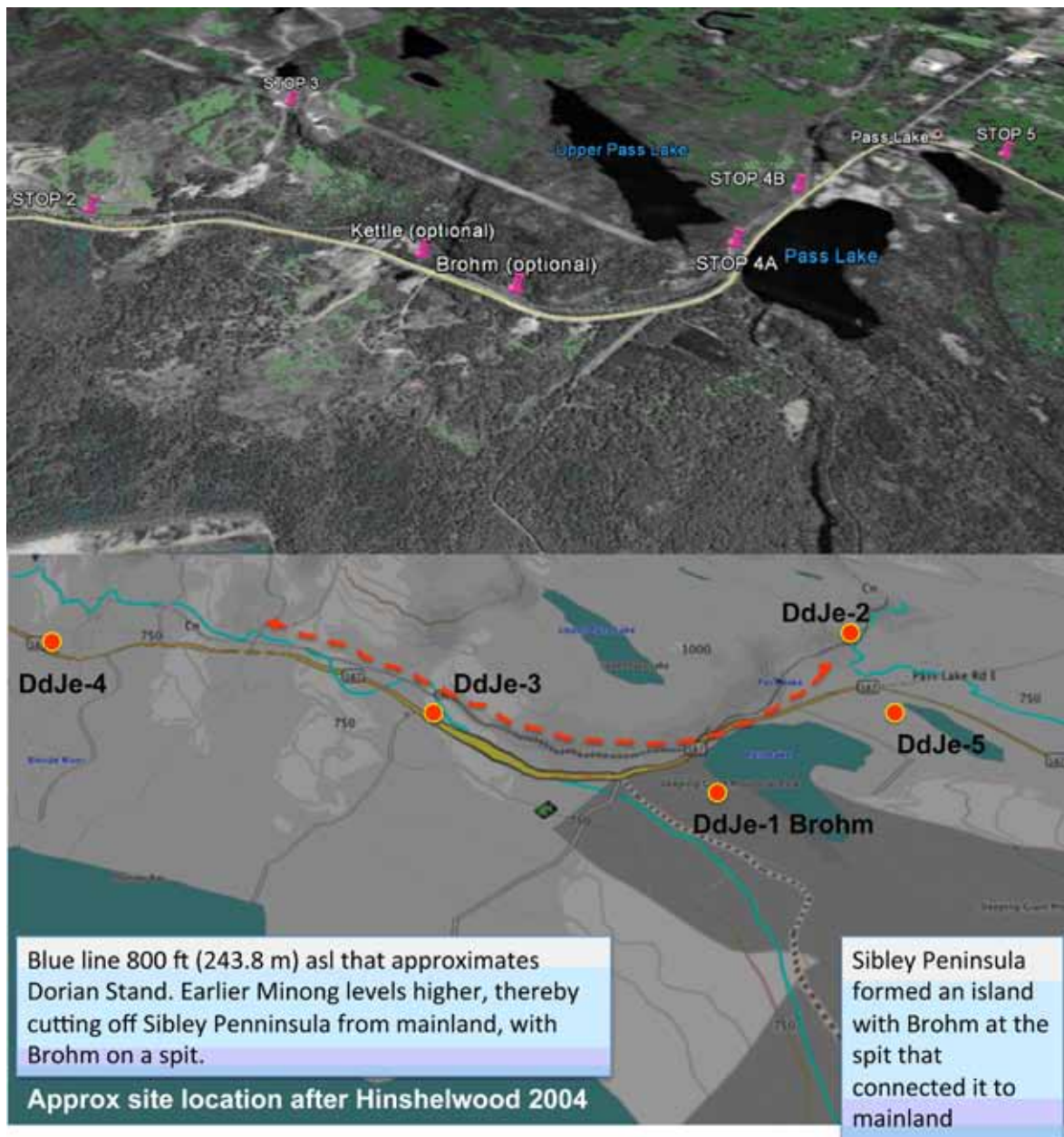


Figure 15. Baymouth bar complex in the vicinity of Pass Lake and Brohm archaeological site (Geddes et al. 1987). “Kettle” field trip stop corresponds to pond shown near the highway

previous outcrop. Immediately above the Animikie-Sibley disconformity, lenses of conglomerate comprise the base of the Pass Lake Formation (Fig. 17). This basal conglomerate contains angular to rounded pebbles, cobbles and boulders in a sandy matrix. These clasts are derived predominantly from erosion of Gunflint Formation chert and taconite; there are also clasts of quartz veins and Archean granite. (Gunflint rocks are exposed a few kilometres to the northwest on

Highway 587). This conglomerate attains a thickness of approximately 3 m at the southeast end of the exposure, where it is sharply overlain by massive to parallel laminated, buff-coloured Pass Lake sandstone (quartz arenite). The constituent sand grains consist mainly of quartz, with minor chert and feldspar, with calcite cement lower in the cliff and silica cement higher up. The sandstones were deposited in the near-shore of the lacustrine system.



Paleo sites near Pass Lake (Brohm).

Along glacial lake strands at base of escarpment (red dashed line) (cuesta).
 Speculation about a 'travel constriction' for caribou movement along the lake shore.
 Perhaps shoreline sites were ambush/intercept sites along animal travel routes.

Figure 16. Paleo-Indian sites, paleogeography and field trip stop locations (Hinshelwood, 2004)



Figure 17. Disconformity between weathered Rove shales and Pass Lake basal conglomerate, Stop 4A

Stop 4B: Pass Lake Formation

A cliff on the far side of the railroad tracks contains the type section of the Pass Lake Formation. 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 (Fig. 18) capped by siltstones on the top of the cliff. Individual beds are reasonably laterally continuous though sometimes lense 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. While examining the lithofacies in the field we will discuss the merits of each interpretation. 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.

Stop 5: Rossport Formation Siltstones

UTM coordinates: NAD83; 16U 0373001E / 5380075N

The dip of the strata to the south, probably developed due to block rotation during Mid-Continental rifting, allows us to observe higher levels of the Sibley Group as we drive down this stretch of the highway.



Figure 18. Thinning-upward sequence in Pass Lake sandstones, Stop 4B

The fining- and thinning-upward trend of the sandstones in the last section culminates in the massive siltstones we see at this stop. Not much can be said about these structureless red siltstones. They appear to have formed from rainout sediment in the offshore portion of the lacustrine system.

STOP 6: Rossport Formation Lacustrine Sheet Sandstones

UTM coordinates: NAD83; 16U 0372877E / 5378948N

Here we have another example of the offshore red siltstones, but with two sandstone layers in them (Fig. 19). The lower layer is actually composed of two amalgamated layers. These sheet sandstones probably represent large flood (storm) events during which the flow conditions were intense enough to transport the sand into the further offshore areas of the lake. Rare sedimentary structures, consisting of hummocky cross-stratification (Fig. 20), in the otherwise massive sandstones indicate storm generated currents deposited the sands. The presence of washed-out dunes at other locations indicates flow velocities were transitional from lower to upper flow regime, as opposed to the near shore sand sheets we looked at two stops back, which were deposited during upper flow regime conditions. The tops of the sand sheets were reworked by waves forming ripples.



Figure 19. Sandstone sheets deposited in the off-shore during storm events, Stop 6. Fairweather deposition produced the siltstone between the sandstone sheets.



Figure 20. Hummocky cross-stratification in a sheet sandstone, Stop 6. This type of layering is produced by the interaction of storm waves and an offshore flowing current carrying sand. These offshore currents, geostrophic flows, are produced by the storm surge draining away from land during the waning of the storm.

Stop 7A (west side of road): Rossport Formation Lacustrine Channel Sandstones

UTM coordinates: NAD83; 16U 0372842E / 5377943N

This outcrop consists of somewhat chaotic layering (Fig. 21). It appears to have several steeply dipping faults running through it. On the southern (down-road) side of the outcrop the sandstone layers abruptly abut against red siltstone. Sandstone layers near the top of the outcrop appear to lean downwards. There are also two lithologies present here that we have not seen so far. The most evident of these is purple shale. It has an interesting mineralogy compared to the red siltstone. The red siltstone has abundant potassium-rich micas and clays. The purple shale does not contain these but instead has potassium feldspar (SEM-EDX and XRD analyses). This implies an alteration where potassium enrichment drove the standard hydrolysis weathering reaction backwards. The other new rock type is dolostone. Beds of it resemble the sandstone, but it is easily distinguished on fresh surfaces. Try to figure out what is causing the deformation of the layering.

Stop 7B (east side of road): Rossport Fm., Lacustrine Channel Sandstones

UTM coordinates: NAD83; 16U 0372868E / 5377728N

The layers strike across the road, meaning this section should be similar to the one we just looked at.



Figure 21. Chaotic layering in an outcrop stratigraphically overlying and sheet sandstones and massive siltstones, Stop 7b.

It is not. Before reading further look at the outcrop and see if you can figure out what is going on.

The outcrop consists of three units. There are beds of sandstone at the top of the outcrop. Underlying these is a chaotic, poorly-sorted, jumble of siltstone, shale and dolomite clasts with a pebbly silt matrix. Underlying this is a thick layer of dolostone that has a very irregular upper surface (Fig. 22). In places the intraformational conglomerate extends down to ground level. Now that you possibly have a better idea what the lithologies are can you figure out what happened here?

The answer: You are looking at a karst surface on the dolostone. It was buried by a sub-aerial mass-flow deposit, possibly triggered by a change in the basin slope at this time, from down to the southeast to down to the northwest. The sand-sheets that also come in at this stratigraphic interval are flowing from the southeast (Cheadle, 1986; Rogala, 2007). There is also the possibility that the chaotic unit represents a collapse breccia.



Figure 22. Sub-aerial mass-flow deposits overlying a very irregular karst surface eroded into the dolostone at the bottom of the photo, Stop 7b. There is also a possibility that the intraformational conglomerate represents a collapse breccia.

Stop 8: Rossport Formation., Subaerial Siltstone-Caliche

UTM coordinates: NAD83; 16U 0373095E / 5376859N

This outcrop represents the highest level of the Sibley Group that we will see on the peninsula. Again we have an outcrop of massive siltstone. However, the internal structuring here is very different than those previously examined (Fig. 23). This would have been more evident a few years ago when the outcrops were buried by small, centimeter and less, chunks weathered out from them. In the interim people, including PWF, discovered that this material can be used to make decorative garden paths and it has disappeared. If you have worked in the southwestern badlands you may have seen modern examples of this type of material coating the ground. The small chunks are soil peds that are baked in the sun. They commonly have clay coatings called cutans on their sides. Further evidence that what we are looking at represents arid to semi-arid soil is the presence of the light green layers. These are rich in dolomite and represent dolocrete layers which form in soils in semi-arid environments. They are created where evaporation is greater than precipitation and there is a net upward movement of water due to capillary action. Evaporation causes the soil fluid to be super-saturated and precipitate carbonate. We can see the peds and dolocrete layers in these exposures (Fig. 23), denoting the large, probably shallow, lake had gone from this area for good.



Figure 23. Soils developed in the semi-arid environment of the Rossport Formation. The upper light greenish grey unit is a carbonate-rich horizon with some dolocrete. The red unit consists of soil peds, the small fragments it is disintegrating into.

STOPS 9A,B: Northwest-striking dykes in Rossport Fm.

UTM coordinates: NAD83; 16U 0372312E / 5374104N and 0372320E / 5374052N

Two narrow, parallel, northwest-striking diabase dykes intrude Rossport Formation siltstones at these two locations. Sampling by Hollings et al. (2009) identified profound geochemical differences between northeast- and northwest-striking dykes on Sibley Peninsula.

These dykes plot within, or very near the fields defined for mafic/ultramafic sills and intrusions on both Gd/Yb_{cn} versus La/Sm_{cn} and Mg# versus TiO₂ diagrams (Fig. 24). These fields are derived from mafic/

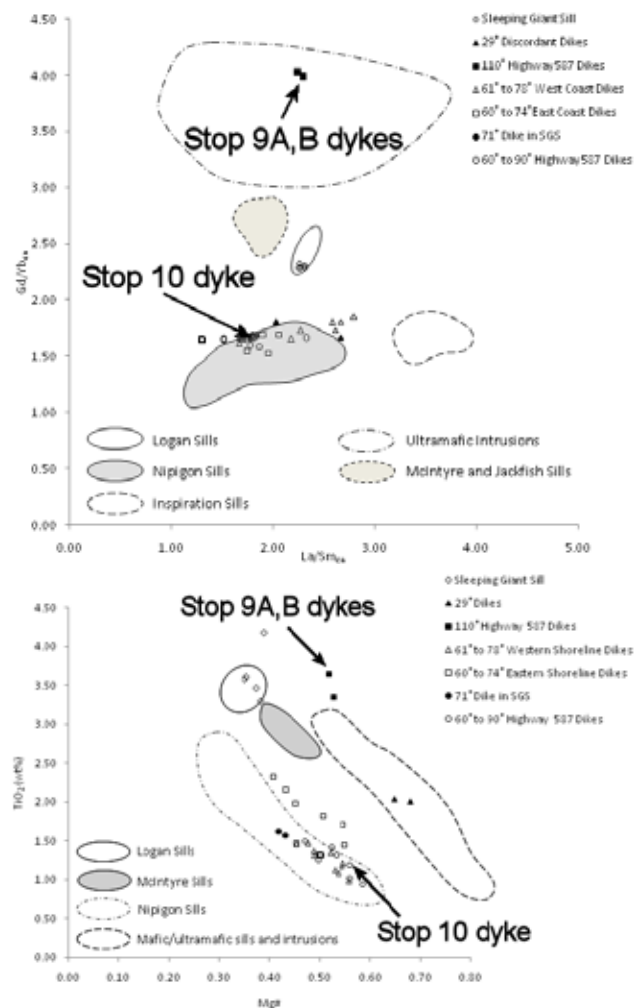


Figure 24. Gd/Yb_{cn} versus La/Sm_{cn} plot and TiO₂ versus Mg# plot for diabase dykes at Stops 9A,B and 10, as well as other mafic and ultramafic intrusions on the Sibley Peninsula and the northern Midcontinent Rift (Carl, 2011; Hollings et al., 2007a,b)

ultramafic intrusions located in the Nipigon embayment northeast of the Sibley Peninsula (cf. Hollings et al., 2007a,b). The northwest strike directions make them spatially unsuitable candidates for feeders of any of the mafic/ultramafic intrusions located near Lake Nipigon, northeast of the Sibley Peninsula (Carl, 2011). REE patterns for these dikes do not display large negative niobium anomalies compared to the REE patterns of east-northeast-striking dikes. The small negative niobium anomalies suggest a more primitive source for these northwest-striking dikes compared to the east-northeast striking dikes of the Sibley Peninsula. The SiO_2 weight percentages of these dykes is 48.7 and 49.4, respectively. TiO_2 and MgO weight percentages for these two dikes are roughly 3.35 and 3.65 and 7.74 and 7.19, respectively. No cross-cutting relationships were noted at the Highway 587 sample sites, and therefore a relative age relationship of these dikes to the east-northeast striking dikes (e.g., Stop 10) is unknown (Carl, 2011).

Stop 10: East-northeast-striking dyke in Rosspoint Formation

UTM coordinates: NAD83; 16U 0373382E / 5368080N

At this location, a steeply dipping, east-northeast-striking diabase dyke crosscuts Rosspoint Formation

siltstones (Fig. 25). There appear to be some localized contact metamorphic effects, including some hornfels and reduction of the iron in these hematite-rich sedimentary rocks. The geochemistry (Gd/Yb_{cn} versus La/Sm_{cn} , TiO_2 versus $\text{Mg\#}$; Fig. 24) of east-northeast-striking dykes along Highway 587 falls into the field previously defined by Nipigon sills (Carl, 2011; Hollings et al., 2007a,b). They may be correlative with Pigeon River dykes of similar orientation south of Thunder Bay.

STOP 11: Rove Formation and east-northeast-striking diabase dyke (UTM 5354955N, 365488E)

UTM coordinates: NAD83; 16U 0365488E / 5354955N

The unweathered Rove Formation can be seen here (Fig. 26). The dominance of siltstone raises the possibility that this is the lower portion of the Rove Formation. In fact the numerous siltstone layers imply that we may be looking at the lowest ten metres of the Rove Formation, as during initial transgression the more shore proximal location led to a more silt-rich interval. The fine parallel layering and lack of wave formed structures is interesting as it appears that no shallow water, coarser-grained lithofacies were deposited in the Rove. This may denote very rapid transgression or an arid climate limiting sediment delivery to the basin. Or



Figure 25 . East-northeast-striking diabase dyke crosscutting Rosspoint Formation siltstones, Stop 10, south of Joeboy Lake.



Figure 26. A grey siltstone dominated succession of the Rove Formation, Stop 11. This outcrop also contains thin dark shales and thicker, clay-rich sandstones. The Rove in this area either represents the basal siltstone-rich area or a portion of the turbiditic fan/ramp higher in the Formation. The diabase can be seen in the upper right of the photo.

possibly, as paleocurrents indicate the Rove Formation represents foreland deposits associated with the Trans-Hudson Orogeny, an early sediment starved phase was produced by development of a tectonic moat. This is common in other orogeny related black shale-turbidite

sequences such as the Martinsburg Formation, which was deposited in the Taconic foreland. The other possibility is that this section is considerably higher in the Rove Formation. The presence of sandstones indicates that this may be a portion of the submarine

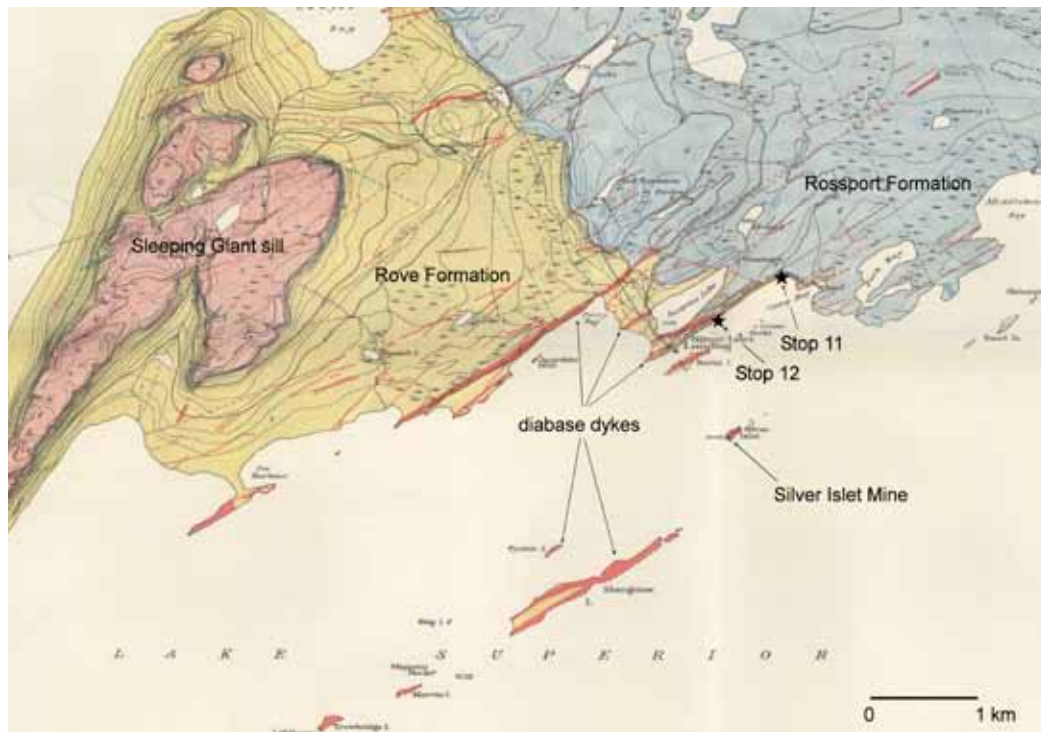


Figure 27. Geology of the southern Sibley Peninsula (Tanton, 1924, 1931) showing stop locations.

fan/ramp system that overlies the 100 meters of basal shales and siltstones.

STOP 12: East-northeast-trending dyke in Rove Formation / Silver Islet view

UTM coordinates: NAD83; 16U 0366232E / 5355427N

This stop affords us the opportunity to look across the waters of Lake Superior to the former site of the Silver Islet Mine, 1.8 km to the south. This tiny speck, barely rising above the waves, was once home to the most famous silver mine in the world (Figs. 28, 29). The epic story of its discovery, development and legacy was summarized by Mohide (1985):

The early regulations of the government with regard to the area of mining locations were, compared to modern ideas, generous to a fault. A mining location was required to be two miles in front by five miles in width, comprising an area of 6,400 acres. In 1856 a mining company had obtained from the Crown sixteen of these locations fronting at intervals on the north and northeast shores of Lake Superior, comprising in all 99,498 acres. The conditions required the grantee "to commence and bona fide carry on mining operations within a period of eighteen months", under penalty of forfeiture of the lands. But the company did not comply with the conditions, neither did the government exact the penalty. The grant reserved all mines of gold and silver and imposed a royalty varying from 2 to 10 per cent of the value of the ore extracted. In, as we may suppose, its anxiety to start things moving, the government abandoned the reservations of gold and silver, repealed the royalties, and forgave one-half of the purchase money of 80 cents per acre. The only offset on the part of the government was to levy a tax of two cents per acre on all lands granted previous to 1868. All restrictions having been removed and facing an annual tax of approximately \$3,140 the company apparently deemed it advisable to examine their lands for possible deposits of mineral. Their interest was probably quickened by the discovery of silver made by the McKellar brothers and others on the north shore of Lake Superior. The task of examination the company committed to Thomas MacFarlane, a well-known geologist and civil engineer. MacFarlane and his party set out in the spring of 1868 and cruised the

locations one by one. On the Jarvis location they found a vein of silver, upon which considerable work was afterwards done and a quantity of silver recovered.

Eventually the party arrived at the Woods location. He determined to make a complete study of the location and set his assistant, Gerald C. Brown, to survey the shore-line. While engaged in planting pickets on the islands in Lake Superior fronting the location, Brown landed on the tiny rock about the size of a ballroom to which Macfarlane afterwards gave the name of Silver Islet and here he noticed a vein carrying galena. Macfarlane thereupon visited the spot himself and put three of his men to work. On the north shore of the islet there was a vein having a width of 20 feet, which on the south divided into two branches, each seven to eight feet wide. On the 10th of July the first metallic silver was noticed by John Morgan, one of the exploring party, at the water's edge on the east or hanging side of the west branch of the vein, in the form of small nuggets. A single blast was sufficient to detach all the vein rock carrying ore above the surface of the water, but the ore was traced some distance out into the lake, where instead of scattered nuggets of native silver, large patches of veinstone rich in galena were visible, intermixed with small particles and large nuggets of silver. The thickness of the rich part of the vein varied from a few inches to two feet and by working in the icy water with crowbars some rich pieces of ore were broken off.

On the 15th of July three packages of the best specimens were shipped from Fort William, Thunder Bay, altogether 1,336 lbs. of ore having been obtained. This shipment was carefully weighed and sampled in the following December. Assays by Professor Chapman of Toronto, Dr. Hayes of Boston and Macfarlane himself, gave an average of 2,087 ounces Troy per long ton. Next year explorations were resumed on the rock, but winds and waves, together with the extreme coldness of the water, proved great hindrances. Nevertheless, by working with tongs and long-handled shovels in two to four feet of water, the party was able to raise and ship 46 half-barrels of good ore, weighing 9,455 lbs. valued by Macfarlane on the basis of his assays at \$6,751. Such results indicated a mine and amply justified further development. A shaft house and sleeping

and dining rooms for the men were erected and strong barriers of two-inch plank built to protect them from the furious gales and sweeping winds of Lake Superior. Inflowing water slowed up work in the shaft and the contracted area of the mine severely hampered operations. Weather conditions, however, had their compensations for when the winter set in, the frozen surface of the lake provided solid footing for the men who managed to raise some nine tons more of the ore. The total quantity of ore recovered up to this time was 28,073 lbs. which realized after being smelted, the sum of \$23,115. The mining company in March, 1870 sold not only the Woods tract, but all of its other locations, now 18 in number, to a new group, the price realized being \$225,000.

The new company formed the Ontario Mineral Lands Company and selected Captain William B. Frue as manager, who at once set to work in September of that year. Frue was a man of remarkable quality and is worthy of some words of mention. Not only was he a skilled and experienced mine superintendent, but he seems to have been of heroic calibre. During the six years beginning with 1870 production of silver from the mine amounted to 1,561,882 fine ounces, but output was lessening year by year, due to unfavourable changes in the vein as the lower workings were reached. The company got into financial difficulties and facing a heavy deficit, sold all its holdings to a new concern known as the Silver Islet Consolidated Mining and Lands Company, with a capitalization of 51,000,000.

The new company met with wonderful success, far exceeding their expectations. The first three levels were explored and much silver recovered from them. The mine, which had been allowed to fill up as far as the third level, was dewatered and work was carried on from the fourth to the tenth level. The results are thus described in the company's report for 1878: "Silver of unparalleled riches was found in the winzes, in the drifts and in the stopes and rich stamp-mill rock abounded in all workings, the vein north of the shaft being particularly productive". The year 1878 closed with an output of silver estimated at 724,632 ounces, of which 551,111 ounces was obtained from "packing ore" (i.e. one rich enough to be put up and shipped to the smelter in small packages) and 170,521 ounces from stamp mill concentrates. This estimate, however,

proved to be too large, for the amount realized was 5137,022 less than the value originally placed on the shipment. Frue had been succeeded as superintendent by Richard Trethewey, who decided to divert the vertical shaft to an incline following the line of the diorite dike, associated with the very rich ore of the earlier workings had been found. He carried it down to a further depth of 414 feet and drifts started at the bottom showed the vein to present highly promising conditions. Some very rich ore was encountered in a south drift. The hope was that the chimney of ore would again be tapped at the junction of the diorite and the slates, but although fugitive bunches of ore were met with, the mine failed to respond to expectations. Evidently the end was approaching. Minerals which frequently accompanied the silver, such as [zinc] blende, galena, pyrite, cobalt and nickel were found, but the silver was absent. Notwithstanding financial and mining difficulties, the work was continued until February, 1884. It was still intended to carry on, but a cargo of coal in charge of a drunken ship's captain had failed to arrive before the close of navigation and no course was open but to allow the mine to fill with water and cease operations.

The total production of the mine in ounces cannot be precisely stated, the records being incomplete, but the entire value is given at \$3,500,000. At the average price of silver during the 16-year period of operations say \$1.15 per ounce, this would represent a total output of about 3,044,000 fine ounces. To this may be added 16,769 ounces obtained during 1921 and 1922 by the Islet Exploration Company which removed a quantity of rich ore from the roof of the mine. The main production was from two very rich bonanzas, one of which was completely worked out in 1874, yielding over two million dollars. In shape this mass of ore resembled an irregular pear and consisted of arborescent silver. The second bonanza was found on the third level in 1878. It was remarkable for its width (5 feet solid across the breast) and for the occurrence of two previously unknown compounds [mixtures] of silver, huntelite and animikite. This deposit was phenomenal in its structure, the middle of the fourth level being sunk literally through solid silver, the metal projecting boldly from the four walls of the winze. In the breast it stood out in great arborescent masses in the shape of hooks



Figure 28. Silver Islet ca. 1900 (top; www.canadiangeographic.ca) and today.

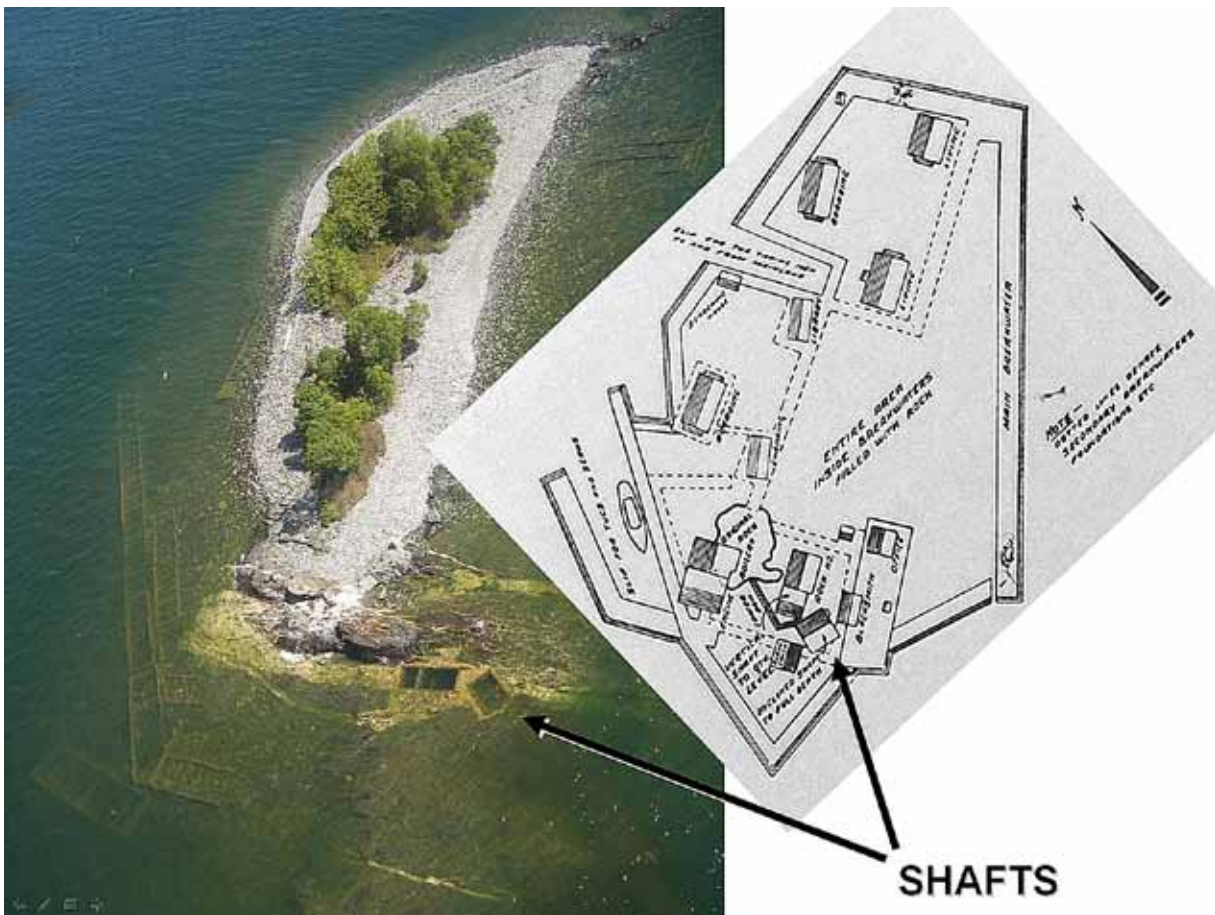


Figure 29. Aerial view of present-day Silver Islet. Site plan of old mine from Barr (1988). Note that the original island consisted only of the small outcrop at the lower end of the present island.



Figure 30. Stop locations, southern part of Sibley Peninsula. Image from Google Earth.

and spikes and in gnarled, drawn out and twisted bunches. The width of the deposit was over 10 feet and including the accompanying stamp rock, it yielded about 800,000 ounces of silver.

Probably nowhere, or at any rate nowhere in Canada, had mining been carried on under conditions so difficult as at Silver Islet, the area of which before enlargement by protective works, was no larger than a good-sized ballroom. Wind and water conspired to prevent an invasion of the tiny spot.

Silver Islet not only rose to fame as a prolific silver producer, having contributed half of all the silver in Canada in 1870's. The Frue vanner, still in use in different forms today, was developed at Silver Islet in 1872. The use of steam-powered diamond drills and the Burleigh piston-type, compressed air-powered rock drill, was pioneered there. The discovery of Silver Islet led to the development and prosperity of Port Arthur (now Thunder Bay) and spurred the exploration and settlement of northwestern Ontario (cf. Barr 1988).

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Field trip 3 - Lac des Iles Mine

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Regional Setting of the Lac des Iles Mine

The Lac des Iles Mine area is underlain by mafic to ultramafic rocks of the Neoarchean Lac des Iles Intrusive Complex (LDI-IC). The LDI-IC is part of the Lac des Iles Suite, whose mafic intrusive rocks generally range in age between 2686 and 2699 Ma (c.f. Stone, 2010). These rocks have intruded a variety of metamorphosed granitoid and supracrustal greenstone belt rocks (ca. 2.9 to 2.7 Ga in age) of the Wabigoon Subprovince of the Superior Province (Fig. 1). The LDI-IC lies immediately north of the boundary between the volcano-plutonic Wabigoon and metasedimentary Quetico subprovinces. The LDI-IC is the largest of a series of mafic and ultramafic intrusions that occur

along the Wabigoon-Quetico boundary and which collectively define a 30 km diameter circular pattern (Fig. 1).

There are three broad, temporally diverse settings for Archean platinum group element (PGE) mineralization west of Lake Nipigon (c.f. Smyk et al., 2002):

- (1) Pre-tectonic mafic to ultramafic subvolcanic(?) intrusions intimately associated and coeval with greenstone belts of various ages in the Wabigoon Subprovince;
- (2) Mafic intrusive rocks occurring within syntectonic to post-tectonic, diorite-monzodiorite-monzonite suites with sanukitoid affinity within the Wabigoon and Quetico subprovinces (e.g., Shelby Lake

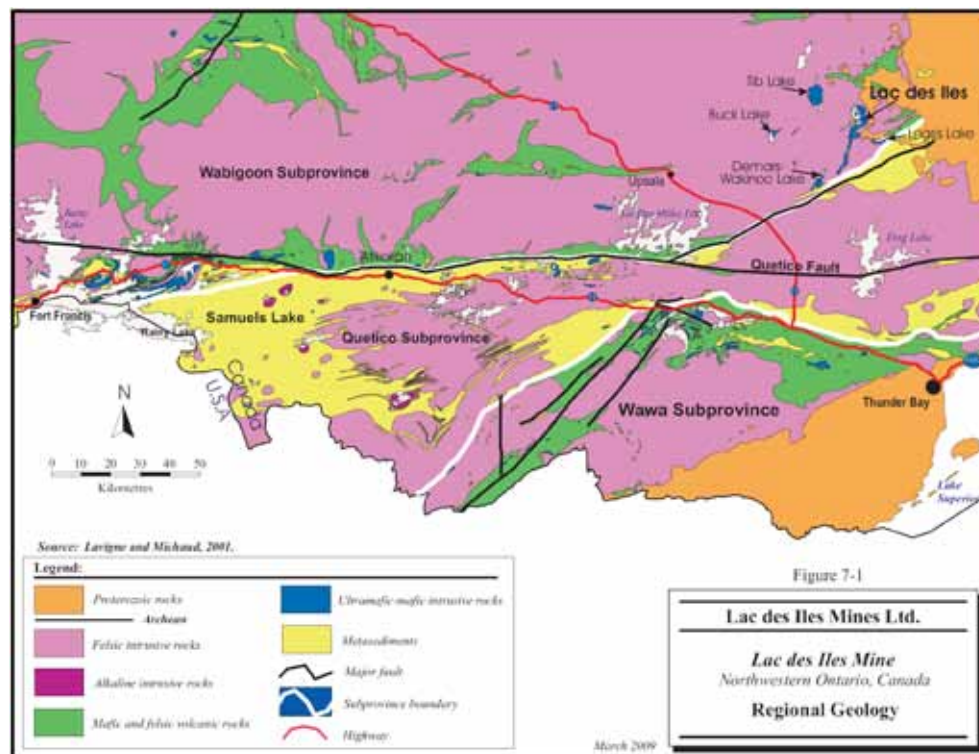


Figure 1. Regional setting of the Lac des Iles complex and related ultramafic and mafic intrusions within the Wabigoon Subprovince (from Lavigne and Michaud, 2002).

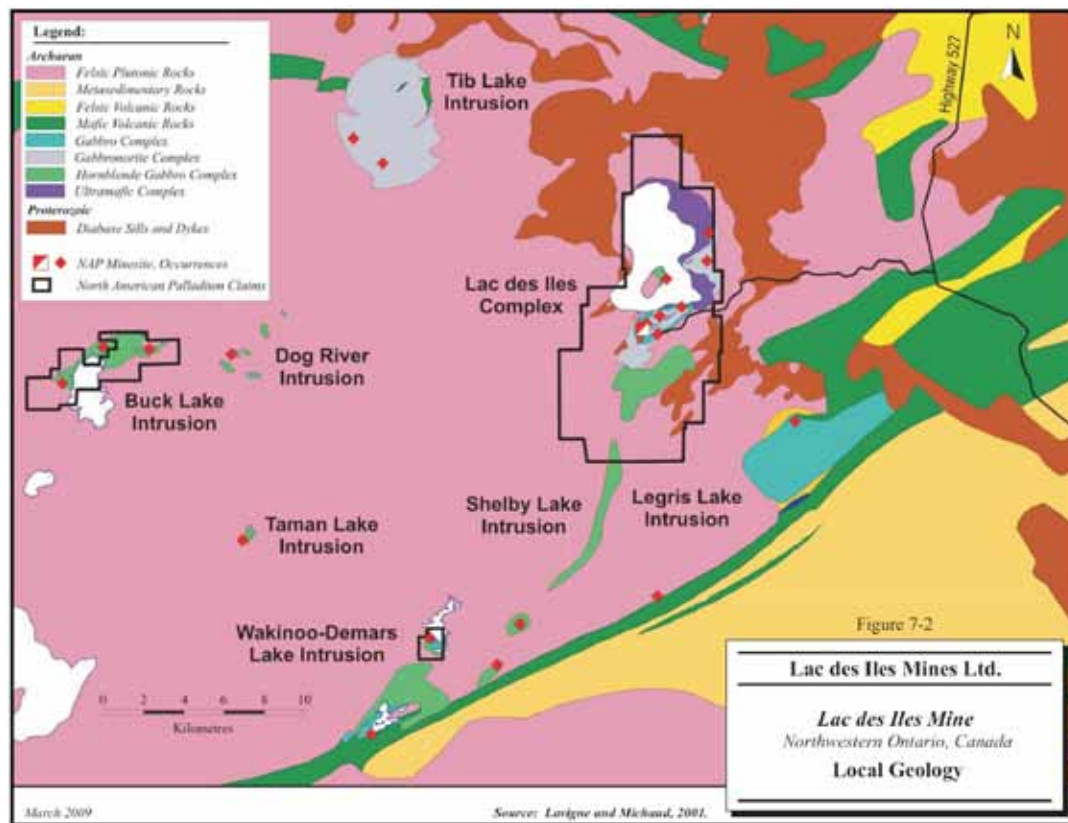


Figure 2. Geologic setting of the Lac des Iles complex and related ultramafic and mafic intrusions (from Lavigne and Michaud)

batholith, Stone et al., 2003; Roaring River Complex, Schnieders et al., 2002).

- (3) Posttectonic, mafic to ultramafic intrusions related to late plutonism in the Wabigoon Subprovince, hosted by gneissic tonalite-granodiorite (e.g., Lac des Iles suite).

Pre-tectonic, deformed mafic to ultramafic intrusions and/or coeval komatiitic metavolcanic rocks within greenstone belt assemblages may host copper-nickel-PGE mineralization. This broad classification is equivalent to the “komatiitic-associated” and “intrusions comagmatic with volcanic rocks” settings for copper-nickel-PGE ± chromium mineralization described by Fyon et al. (1992). Such deposits are characterized by remobilized, deformed and annealed, net-textured to massive sulfides. Examples include the past-producing Shebandowan Mine in the Shebandowan greenstone belt, west of Thunder Bay (8.64 million tonnes mined, grading 1.92% Ni, 0.98% Cu, 2.62 g/t Pt+Pd; Resident Geologist’s Files, Thunder Bay South District, Thunder Bay), and in the Obonga Lake belt, the Core Zone gabbro (2733±7 Ma; Tomlinson et al., 1999) and the Puddy Lake serpentinite (both may contain indications of such mineralization). Lavigne et al. (1991) reported

metal contents from the Puddy Lake serpentinite up to 5.02% Cu, 2.1% Ni, 415 ppb Au, 1500 ppb Pt and 3750 ppb Pd; cobalt values of 0.07% were reported from drilling in the 1960s (Lavigne et al., 1992).

PGE mineralization is also associated with rocks of the sanukitoid suite (ca. 2688 to 2690 Ma, Davis et al., 1990; Kamo, 2004; cf. Stern et al., 1989). The Shelby Lake batholith, which consists of hornblende leucogabbro (diorite) to monzodiorite and hornblende granodiorite to granite, contains disseminated sulfide zones in thin units of hornblende gabbro distributed along its northwestern margin (e.g., Turtle Hill and Stocker occurrences). Similar, somewhat larger sulfide occurrences have been described (Stone et al., 2003) at Wakinoo Lake, Towle Lake (e.g., Powder Hill, Stinger and Vande occurrences) and Legris Lake (e.g., Main and Poplar occurrences). Diamond drill holes in hornblende gabbro at Legris Lake contained 2.04 g/t Pd, 0.41 g/t Pt, 0.71 g/t Au, 0.42% Cu and 0.13% Ni over 9.95 m (News Release, Avalon Ventures Ltd. and Starcore Resources Ltd., November 10, 2000). The Roaring River complex (Stern and Hanson, 1991) consists of a variety of plutonic rocks including diorite, monzodiorite, monzonite, quartz monzodiorite and granodiorite, all of sanukitoid affinity; gabbroic and

pyroxenitic mega-inclusions occur in these phases. Grab samples of the Mere showing contain up to 1249 ppm Ni, 3159 ppm Cu and 1.1 g/t Pt+Pd+Au and the Leigh (boulder) occurrence returned up to 2067 ppm Ni, 1920 ppm Cu and 1.23 g/t Pt+Pd+Au (Schnieders et al., 2002 and references therein). Disseminated to locally net-textured chalcopyrite, iron sulfides, pentlandite and magnetite typically characterize PGE-

mineralized zones, which are commonly associated with intrusive contacts, polyphase intrusive breccia, and sheared and hydrothermally altered zones.

Mafic to ultramafic intrusions of the Lac des Iles suite (ca. 2686 to 2699 Ma; Stone, 2010 and references therein; Davis, 2003; Kamo, 2004) and their associated copper-nickel-PGE mineralization at North American

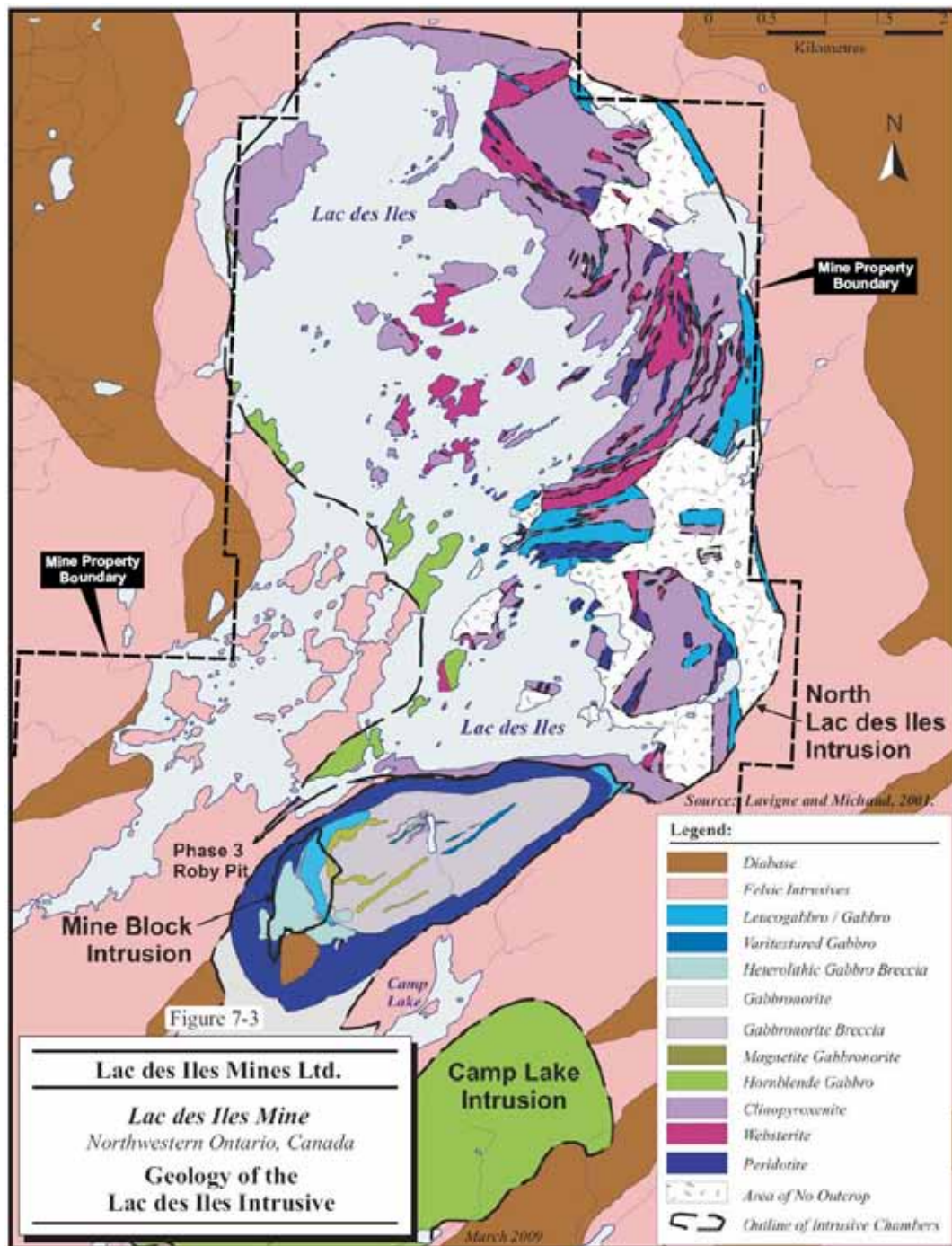


Figure 3. Geology of the Northern Ultramafic and Mine Block intrusions of the Lac des Iles Complex (from Lavigne and Michaud, 2002).

Palladium Ltd.'s Lac des Iles mine were most recently described by Stone et al. (2003). This suite includes the Buck Lake, Dog River, Taman Lake, Demars Lake, Bullseye and Tib Lake intrusions (see Fig. 2), as well as the Northern Ultramafic intrusion and Mine Block intrusion at Lac des Iles (Fig. 3). These leucogabbro and gabbro-norite intrusions ($\pm$ anorthosite, peridotite) range from 1 to 10 km in diameter and are considered to represent a continuum of the Quetico suite of mafic to ultra mafic intrusions (cf. MacTavish, 1999; Pettigrew et al., 2000).

Michaud (1998), Lavigne and Michaud (2002), and Lavigne et al. (2005) provided recent descriptions of the deposits in the Mine Block intrusion (Fig. 4). Platinum group elements are associated with disseminated Cu-Ni-sulfide minerals in the matrix of magmatic breccia, in varitextured to pegmatitic gabbroic rocks (which together represent the Breccia Zone), and also in pyroxenite that is part of the High-Grade Zone. Platinum group elements also occur with

sulfide-poor, varitextured to pegmatitic gabbro in the Roby and North Roby zones and are locally associated with strong silicate alteration (e.g., Roby Zone and portions of the High-Grade Zone; Lavigne et al., 2005). The Roby Zone is the product of multiple stages of intrusion, alteration and mineralization (Lavigne et al., 2005).

Hinchey et al. (2005) put forward a schematic model illustrating a deposit model for the history of mineralization at the southern Roby Zone (Fig. 5). The textures of the Lac des Iles deposit are similar to those of contact-type PGE deposits, but there are fundamental differences between the two. The Lac des Iles deposit is not localized near the contact between the host intrusion and the country rocks and evidence of the assimilation of the host rocks is lacking. Instead, the mineralization at Lac des Iles has many features in common with layered intrusion-hosted deposits, in which pulses of primitive magma introduced the PGE. Unlike the quiescent magma chambers of most

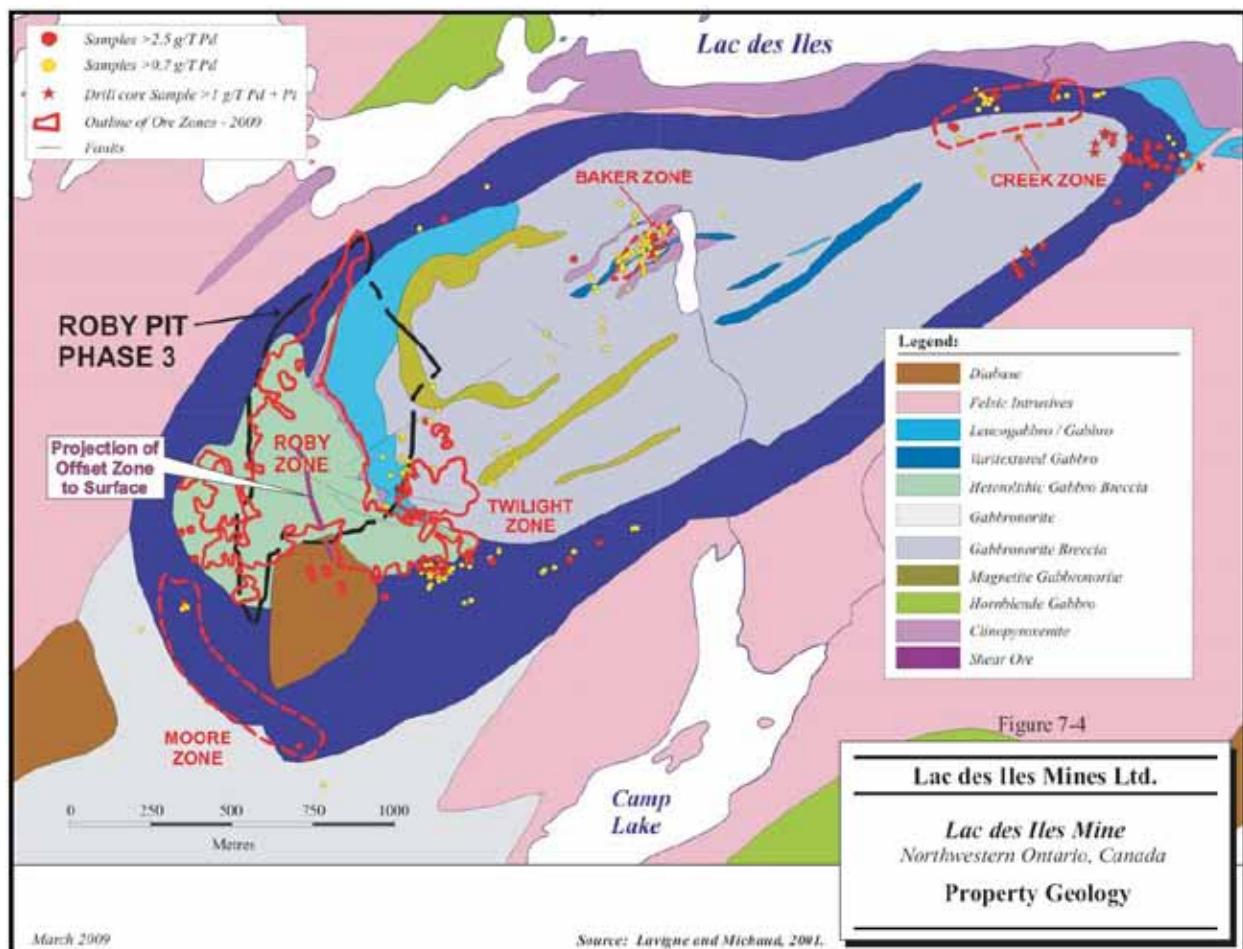


Figure 4. Geology of the Mine Block intrusion showing the main ore zones (from Lavigne and Michaud, 2002).

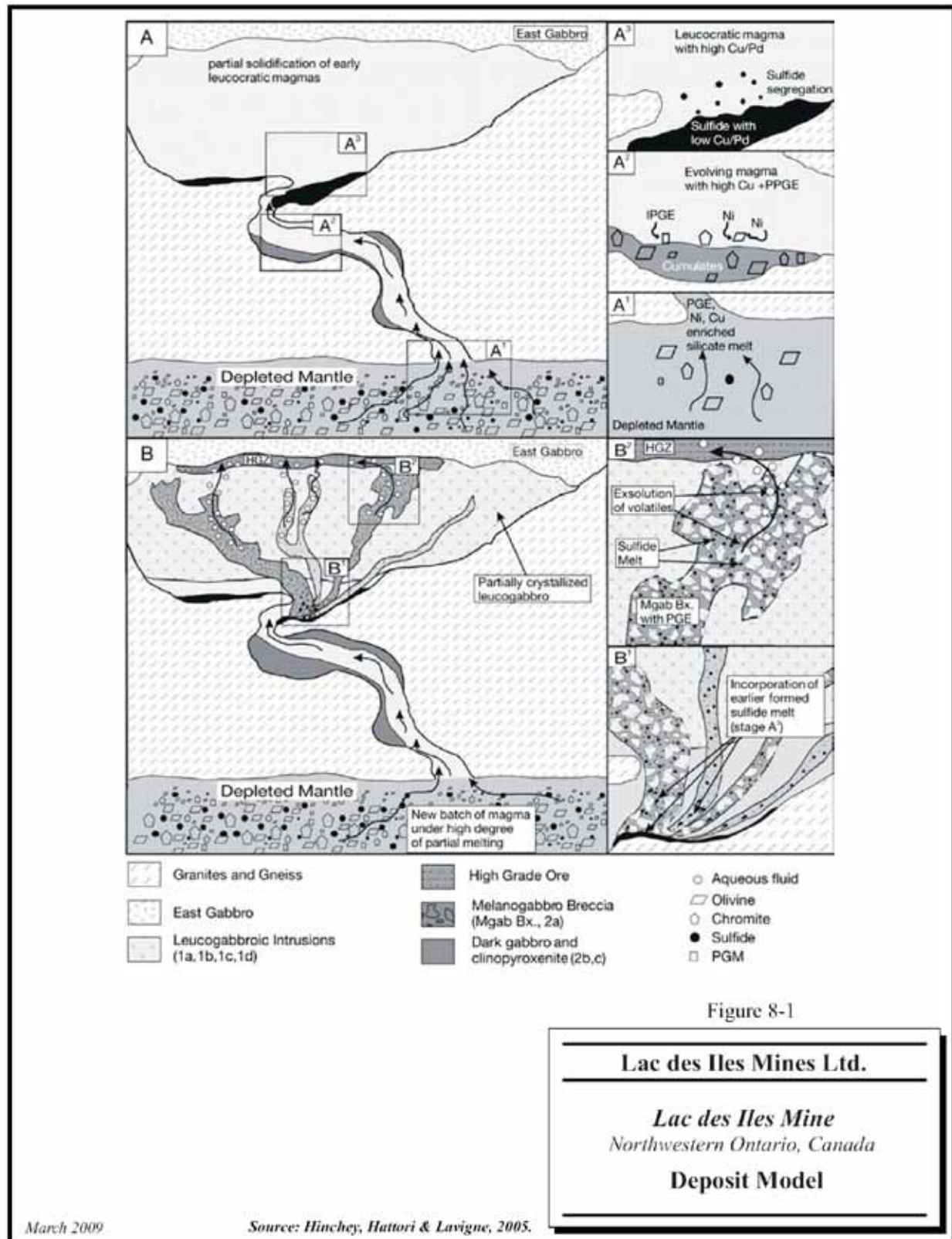


Figure 5. Schematic mineralization model for the Lac des Iles Mine (Hinchey et al., 2005)

layered deposits, the magmas at Lac des Iles were intruded energetically, forming breccias and magma-mingling textures. Magmas formed by a high degree

of partial melting in a depleted mantle source (Fig. 5, A1) became enriched in Cu, Pt, and Pd through fractional crystallization of olivine, chromite, and high-

temperature PGM (Fig. 5, A2), segregated sulfide melt that had low Cu/Pd ratios along the conduit and the base of the magma chamber (Fig. 5, A3), and solidified as the early leucocratic gabbros. A second episode of partial melting in the mantle source produced another batch of fertile magma. As with the early magma, this magma was enriched in Cu, Pt, and Pd through fractional crystallization (Fig. 5, A2). This magma incorporated the earlier sulfide melt and intruded forcefully into the partially crystallized leucocratic rocks (Fig. 5, B1), causing brecciation and magma mingling, and solidified as fertile melanocratic gabbro. Aqueous fluids that separated from the melanocratic magma percolated through the cumulates, partially dissolving Pd and concentrating it in the High Grade ore zone adjacent to barren East Gabbro (Fig. 5, B2).

History of Exploration and Mining

The Lac des Iles area was initially mapped by Jolliffe (1934) and later by Pye (1968), Watkinson and Dunning (1979), Sutcliffe and Sweeny (1985, 1986), Sweeny and Edgar (1987), and Stone et al. (2003). The regional geology has been summarized by Stone (2010). Economic interest in the area was sparked by the ground-truthed aeromagnetic anomalies. Significant palladium mineralization was first discovered in the Roby Zone in 1963 by a prospecting syndicate. Various exploration programs were undertaken over the next 25 years by a number of companies, including Gunnex Ltd., Anaconda Ltd., Texas Gulf Sulphur Co. Inc., and Boston Bay Mines Ltd. In 1990, Madeleine Mines Ltd., a precursor to North American Palladium Limited (NAPL), developed the property. After intermittent

production and continuing capital expenditures, commercial open pit production of the Roby Zone was achieved in December 1993. NAPL was formed through corporate reorganization.

In 2000, an expansion program began and a new mill was commissioned in the second quarter of 2001 to achieve its rated 15,000 t per day throughput in August 2002. From 1999 to 2001, an extensive drilling campaign identified mineralization at depth, below the ultimate pit bottom. The drilling identified 2 zones with potential for underground mining: the Roby Underground Zone and the Offset Zone.

On July 31, 2003, a positive pre-feasibility study for underground mining of the Roby Underground Zone (down-dip extension of the open pit Main Zone) was completed, and was followed by a feasibility study for underground mining in 2004. Development on the Roby Underground Zone started in 2004, with the ramp developed and the zone accessed in late 2005. Development muck was delivered to the concentrator in December 2005 and underground commercial production began in March 2006. The Offset Zone, discovered in 2000, was historically subdivided into the Offset High Grade Zone and the adjacent Roby Footwall Zone.

The Offset Zone is the fault-offset, down-dip extension of the Roby Underground Zone that was mined below the Roby open pit until October 2008. A number of surface and underground drilling programs have targeted the Offset Zone since 2001.

In 2008, a surface drilling program focused on exploring targets on the Mine Block Intrusion and on

Table 1. Production figures for Lac des Illes (MD&A, 2011)

	Unit	2011	2010	2009	2008
				Mine closed	
Ore Mined	Tonnes	1,830,234	615,926		*3,676,418
Waste Mined – Open Pit	Tonnes				6,964,501
Mill Throughput	Tonnes	1,689,781	649,649		3,722,732
Pd Head Grade	g/t	3.70	6.06		2.49
Pd Recovery	%	78.34	80.80		75.30
Pd Produced	Oz	146,624	95,057		212,046
Pt Produced	Oz	9,143	4,894		16,311
Au Produced	Oz	7,267	4,023		15,921
Ni Produced	Lbs	816,037	395,622		2,503,902
Cu Produced	Lbs	1,596,185	658,013		4,623,278

* Added; Ore Mined - Underground + Ore Mined - Open Pit

Table 2. Resource figures for Lac des Illes (MD&A, 2011)

Location	Tonnes (000's)	Pd (g/t)	Pt (g/t)	Au (g/t)	Cu (%)	Ni (%)	PdEq (g/t)
Measured Resources							
Roby Open Pit	3,722	1.99	0.23	0.17	0.075	0.065	3.77
Roby Stockpiles	763	2.09	0.19	0.18	0.060	0.079	3.79
Indicated Resources							
Roby Open Pit	2,565	2.20	0.24	0.18	0.078	0.068	4.07
Roby RGO Stockpile	13,365	0.97	0.12	0.08	0.034	0.056	2.00
Roby Underground	3,292	7.61	0.44	0.33	0.062	0.077	10.46
Offset Underground	12,331	5.02	0.38	0.37	0.114	0.133	8.28
Total M&I Resources	36,038	3.18	0.26	0.22	0.072	0.086	5.29
Inferred Resources							
Offset Underground	4,637	4.9	0.4	0.3	0.12	0.13	8.0

the Southeast Breccia Zone, situated adjacent to the southeastern corner of the open pit.

The Cowboy Zone was discovered in June 2009 during infill drilling of the Offset Zone to support a pre-feasibility study (news release, NAPL, June 25, 2009). It is located 30 to 50 m down-section to the west of the Offset Zone, extends for up to 250 m along strike and 350 m down-dip, and it remains open in all directions. Similar to the Offset Zone, the Cowboy Zone appears to consist of several mineralized subzones. Intersections include 5.10 g/t Pd over 4 m, 3.88 g/t Pd over 4 m, and 4.46 g/t Pd over 5 m.

Open pit mining of the Roby Zone began in 1993. The open pit was operated by conventional truck-and-shovel mining, with low- and high-grade material stockpiled near the on-site concentrator. In May 2004, LDI collared a portal in the northwest wall of the pit and ramped down to access the Roby Underground Zone that continues down-dip from the Roby Zone hanging wall below the pit. LDI began processing development muck from the Roby Underground Zone in December 2005. The ramp was extended around the pit to the north and the new portal was opened in the east wall in 2006. The Roby Underground Zone reached commercial production at 2000 t per day in April 2006. Operations were suspended in October, 2008 due to the global economic downturn and depressed metal prices. Palladium production at Lac des Iles Mine resumed in April 2010 (news release, NAPL, April 14, 2010). NAPL expects to produce 140 000 ounces of palladium

per year. Ore production from the Roby Underground zone is expected to increase to a target rate of 2600 t per day. A summary of ore mined is presented in Table 1.

Since production began in 1993 at Lac des Iles Mine, almost 42 Mt of ore have been processed, and approximately 2.3 million ounces of palladium produced (see Table 1). A Mineral Resource Summary (December 31, 2008) is given in Table 2 (McCombe et al., 2009).

Local and Property Geology

Many of the following excerpts have been modified after McCombe et al. (2009).

The mine lies in the southern portion of the Lac des Iles Intrusive Complex (LDI-IC) (see Figs. 2 & 3), in a roughly elliptical intrusive package measuring 3 km long by 1.5 km wide, termed the Mine Block Intrusive (MBI) (see Figs. 3 & 4). It hosts a number of PGE deposits and the most important of these is the Roby Zone with its three subzones: the North Roby Zone, the High Grade Zone, and the Breccia Zone.

The MBI comprises rocks with a very wide range of textures and mafic and ultramafic compositions, ranging from anorthosite to clinopyroxenite, leucogabbro to melanonorite, and includes magnetite-rich gabbro. Textures include equigranular, fine- to coarse-grained, porphyritic and pegmatitic, varitextured

units, and heterolithic gabbro breccias. These last three textural types are the most common host to PGE mineralization, including the Roby Zone. The MBI consists of two lithologically distinct domains. The oval-shaped domain immediately south of Lac des Iles is lithologically complex and contains widespread PGE mineralization, while the domain further to the south is dominated by massive, medium-grained, PGE-barren gabbro (see Figs. 2 to 4). Extensive stripping has disclosed that the interior of the oval-shaped domain has an abundance of monolithic and heterolithic breccia with an average composition of gabbro. Within this area, individual lithological units are not laterally extensive and are chaotically distributed. The most laterally continuous unit is a massive, medium-grained gabbro, referred to as East Gabbro (EGAB) (see Fig. 5). EGAB is adjacent to a varitextured gabbro “rim” to the west and more equigranular gabbro (GN) to the east. The varitextured rim is host to the Roby palladium deposit, where heterolithic gabbro breccia (HGABBX) commonly occurs as pipes and pods, and large blocks (~60 m) of varying composition. A pyroxenite unit (PYXT), at the contact between the EGAB and the HGABBX, is host to much of the High Grade Zone.

The principal rock types in the Offset Zone area include the following:

East Gabbro (EGAB) – is a well-known gabbro “marker unit” that is characteristically uniform and compositionally homogeneous. EGAB has very minor alteration, with local trace pyrite and epidote. It has no significant associated mineralization, and bounds the Roby Zone to the east. (i.e., hanging wall contact of the Roby Zone).

Heterolithic Gabbro Breccia (HGABBX) – the principal host for the Roby Zone, consisting of a melanogabbro to gabbro matrix with variable clast composition, ranging from leucogabbro to pyroxenite. Clast percentage varies commonly from 15 to 60%. This unit comprises most of the economic ore grade material in the open pit and underground reserves.

Varitextured Gabbro (VGAB) – the majority of rock types, excluding EGAB, have a varitextured counterpart. The VGAB varies from leucocratic to pyroxenitic, with grain sizes from fine to very coarse, to pegmatitic. The coarser-grained units form patches and “veinlets” within finer-grained counterparts.

Gabbro (GAB) – the most common gabbros in the MBI are medium grained and equigranular, but range from fine to coarse grained and may locally be

leucocratic to melanocratic.

Magnetic Gabbro (MTGAB) – medium-grained, equigranular gabbro occurs within the MBI and contains black, fine-grained, interstitial magnetite (typically < 20%); magnetite content ranges from trace amounts to local, narrow layers of 60 to 95% magnetite.

Pyroxenite (PYXT) – a steeply dipping, thin layer situated along the contact between the Heterolithic Gabbro and EGAB; it hosts the highest proportion of the High Grade Zone. This unit is responsible for much of the high PGE grades. Not all pyroxenites locally carry economic PGE grades.

Gabbro (GN) – a 20 to 50 m thick, steeply dipping slab located along the northwestern contact of the EGAB; it is also a host unit of the High Grade Zone, although to a lesser degree than the PYXT. The gabbro appears to be a gradational extension of the pyroxenite to the northeast of the mine site.

Gabbro Breccia (GNBX) – a palladium-mineralized (Twilight Zone) heterolithic breccia, similar to the HGABBX but without pegmatitic phases or varitextured gabbro; it occurs as a roughly cylindrical pod, approximately 150 m in diameter, completely enclosed by the EGAB.

Dikes – late, post-mineralization, mafic dikes vary from small, discrete bodies that occupy space within the modeled mineralized wire frames to large bodies that control the northern termination of the Offset Zone. A dike swarm approximately 30 m wide and trending approximately easterly was mapped at the southern extent of the Roby Zone.

Two major faults have been interpreted to influence the Offset Zone. The Offset Fault structure displaces the High Grade Zone down and approximately 300 m to the west. This fault, easily picked out in diamond-drill core, is often marked by extensive fault gouge, fracturing, alteration of adjacent country rock, and infilling by mafic dikes. The B2 Fault has recently been recognized and interpreted from the underground Offset Zone diamond drilling. It lies approximately 20 to 40 m below and parallel to the westerly dipping Baker Fault and is marked by narrow intersections of fault gouge, fracturing and late mafic dikes.

Mineralized Zones

The **Roby Zone** is a bulk-mineable, PGE-enriched disseminated sulfide deposit with a minimum north to south length of 950 m, and a width of 815 m, including the Twilight Zone in the southwestern portion of the

deposit. The Roby Zone consists of 3 distinct ore types: High Grade Ore (7.6% of volume), North Roby Ore (5.3% of volume), and Breccia Ore (87.1% of volume). The High Grade Ore is the primary ore type mined underground.

High Grade Zone ore is hosted mainly within a 15 to 25 m thick unit of locally sheared pyroxenite/melanogabbro. A host to high-grade PGE mineralization, it is located in the east-central portion of the Roby Zone, bounded by the barren EGAB hanging wall and HGABBX-hosted Breccia Ore to the west. The High Grade Zone is primarily confined to a 400 m long segment of the pyroxenite, although it does extend northward into the gabbro. The High Grade Zone, striking north-northwest to north-northeast, dips almost vertically near surface and flattens to nearly 45° at depth. Below the open pit, this zone is referred to as the Roby Underground Zone. The zone appears to be terminated down dip by a relatively shallow dipping fault, the Offset Fault.

The **Offset Zone**, a higher grade zone similar to the High Grade Zone, is located below the Offset Fault structure, where it is displaced down and approximately 300 m to the west. The Offset Zone can be split into 3 horizons and has been divided into 3 subzones: the High Grade (HG) Subzone; the Mid (MID) Subzone; and the Footwall (FW) Subzone.

High Grade Subzone mineralization is stratabound, along the contact between the EGAB and the mineralized HGABBX. Within the HGABBX, there is a high-grade core typically constrained to an easily recognized ultramafic unit, the pyroxenite. Width varies from 4 to 30 m, with an average of 15 m. Approximately 2% of the zone is occupied by late dikes (dilution). Less than 1% is occupied by shears and faults.

The **MID Subzone** is proximal to the HG Zone, generally sharing a common boundary in the centre sections and then splitting away near the top and bottom areas. Palladium grades within the MID Subzone can approximate the high grades found within the HG Zone. Apparent widths can vary from 4 to 90 m, with an average of 15 m. Approximately 4% of the zone is occupied by late dikes (dilution). Less than 1% is occupied by shears and faults.

The **Footwall Subzone** is a stand-alone band of higher grade mineralization that can be defined based on higher grade intersections within the Footwall varitextured gabbro mineralization. This subzone, located approximately 2 to 40 m from the MID Zone, was interpreted based on vertical continuity seen in the

drill hole intersections. It is discontinuous and sinuous in plan and has less of a defined areal extent than the other zones. Apparent widths can vary from 4 to 20 m, with an average of 7 m. Approximately 1% of the zone is occupied by late dikes (dilution). Other mineralized zones present within the MBI, as shown in Figure 2-4, are described below:

The **Twilight Zone** was removed with the mining of the open pit.

The **Baker Zone** is located approximately 1 km northeast from the Roby and Twilight zones and contains similar rock types and textures. Gabbro/norites have been intruded by east-northeast-trending, heterolithic melanogabbro breccia and lesser melanogabbro, leucogabbro breccia, varitextured gabbro and late pyroxenite dikes. Surface exploration has exposed the Baker Zone breccias and associated lithologies over a 150 by 55 m area. The heterolithic melanogabbro breccia hosts blebby to disseminated to narrow veinlets of sulphide with sporadic mineralization in the adjacent lithologies. The north-trending, shallowly westerly dipping Baker Fault appears to truncate the Baker Zone mineralization at depth. Extensive surface exploration by NAP occurred mainly from 1998 to 2001 and consisted of prospecting, stripping/trenching (including the main stripped area of approximately 200 by 120 m), channel sampling, geological mapping and ground induced polarization (IP) / resistivity surveys. Sixteen diamond-drill holes in 1998–99 tested the main portion of the Baker Zone over a 250 m strike length and to a maximum depth of 200 m. Subsequent exploration (trenching and diamond drilling) has tested possible strike extensions of the zone and the area below the Baker Fault.

The **Moore Zone** is a low-grade, presently uneconomic, mineralized zone approximately 500 m south of the current Roby pit with similar lithologies and textures to other MBI breccias. The central area of interest is a small breccia pod measuring approximately 200 m long, varying from approximately 15 to 115 m wide, which occurs within the massive, medium-grained gabbro/norite typical of the more southerly domain of the MBI. The main Moore Zone mineralization is located in the eastern portion of the breccia pod and appears to be structurally controlled (trending ~030°, dipping 70° east), ranging from 5 to 25 m thick. Prospecting, mapping, trenching, sampling and limited diamond drilling of the Moore Zone have indicated limited economic potential.

The **Creek Zone** is located approximately 2 km

northeast of the Roby pit in the northeastern nose of the MBI, near the contact with the north LDI-IC. Surface trenching has exposed the main portion of the Creek Zone in an area 90 m long by 10 to 40 m wide. It is dominated by low-sulfide breccias that have intruded the varitextured gabbro rim of the MBI. The breccias consist of approximately 90% GBNR clasts and only approximately 10% MGAB matrix. Unlike the Roby Zone, mineralization is not dominantly hosted by the breccia matrix but seems to occur within the pegmatitic gabbrobronite. Prospecting, mapping, trenching, sampling and limited diamond drilling of the Creek Zone have indicated limited mineralized potential.

Mineralization

Platinum group element and base metal mineralization at the Lac des Iles Mine appears to be dominantly stratabound along the contact between the EGAB and the mineralized HGABBX. Within the HGABBX, there is a high-grade core typically constrained to an easily recognized pyroxenite unit. Visible PGE mineralization is rare and its occurrence is difficult to predict. In general, economic PGE grades are anticipated within gabbroic to pyroxenitic rocks (in close proximity to the marker unit EGAB) that exhibit strong sausseritization of plagioclase feldspars, strong talcose alteration and association with either disseminated or blebby secondary sulfides. Higher PGE grades (mean – 7.89 g/t Pd, maximum – 55.95 g/t Pd) occur in those portions of the pyroxenite that are altered to an assemblage of amphibole (anthophyllite-actinolite-hornblende)-talc-chlorite. The PGE tenor is not proportional to the sulfide content, and samples free of visible sulfide often contain more than 10 g/t Pd. The high-grade mineralization is located primarily within the western, highly altered portion of the pyroxenite, since much of the pyroxenite between the barren EGAB and the High Grade Zone is low grade. The higher grade “High Grade Ore” is not restricted to the pyroxenite as it commonly straddles the pyroxenite/gabbro breccia contact to widths exceeding 250 m.

The majority of platinum-group minerals occur either interstitially to sulfides as cumulus grains or are associated with sulfides at sulfide-silicate boundaries, occurring as discrete mineral inclusions within secondary silicates of altered rocks (Sweeny 1989; Lavigne and Michaud 2001). Palladium and platinum mineralization within the High Grade Zone consists primarily of fine-grained PGE sulfide, braggite and the telluride minerals merenskyite and kotulskite (Sweeny, 1989; Lavigne and Michaud, 2001).

The platinum-group minerals at Lac des Iles Mine include the following (Lavigne and Michaud, 2001):

Braggite (Pt,Pd)S
 Kotulskite Pd(Te,Bi)₂
 Isometrieite Pd₁₁(Sb,Te)₂As₂
 Merenskyite PdTe₂
 Moncheite PtTe₂
 Palladoarsenide Pd₂As
 Sperrylite PtAs₂
 Stibiopalladinite Pd₅Sb₂
 Stillwaterite Pd₈As₃
 Vysotskite PdS
 Unnamed Ag₄Pd₃Te₄
 Unnamed Pd₅As₂
 Melonite, gold, pentlandite Pd in solid solution

Field Trip Stops

Roby Zone open pit
 Baker Zone
 North VT Rim Trenches
 Exploration office and diamond drill core
 Directional drilling sites and Devico Unit Mill

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Field trip 4 - Shebandowan Mine Area

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Introduction

This field trip will focus on two main aspects of the geology of the Lake Shebandowan area: the presence of a suite of metasediment and metavolcanic rocks usually described as being “Timiskaming-type” that unconformably overlies older Archean, “Keewatin-type” rocks; and the presence of numerous komatiitic ultramafic bodies within the underlying Keewatin metavolcanic rocks that in part are spatially associated with a Timiskaming-age pull-apart basin (Figs. 1 and 2).

Many of the Archean terrains within the Canadian Shield are host to an eo-Archean sequence of shoshonitic/

alkali metavolcanic and fluvial metasedimentary rocks that occupy pull-apart basins typically spatially related to major sub-province transcurrent boundary faults such as the Kirkland Lake-Cadillac Fault and the Porcupine-Destor Fault. This suite of rocks is commonly referred to as “Timiskaming-type” after the Timiskaming Group found within the Abitibi Terrain. Examples of “Timiskaming-type” include the Oxford Lake Group in Manitoba, the Haüy Formation in the southern and northern parts of the Abitibi, the Opemisca Group near Chibougamau, the Seine Group of the Wabigoon Subprovince, and the Shebandowan Group (Card, 1990).

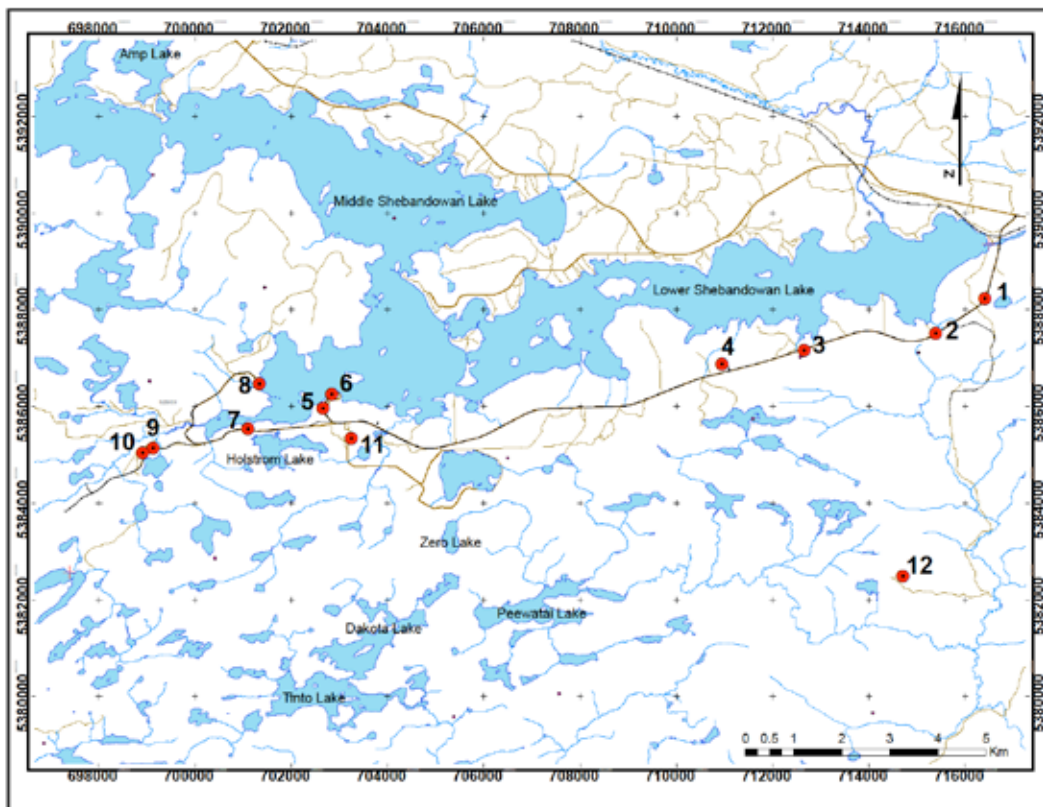


Figure 1. Shebandowan field trip stops

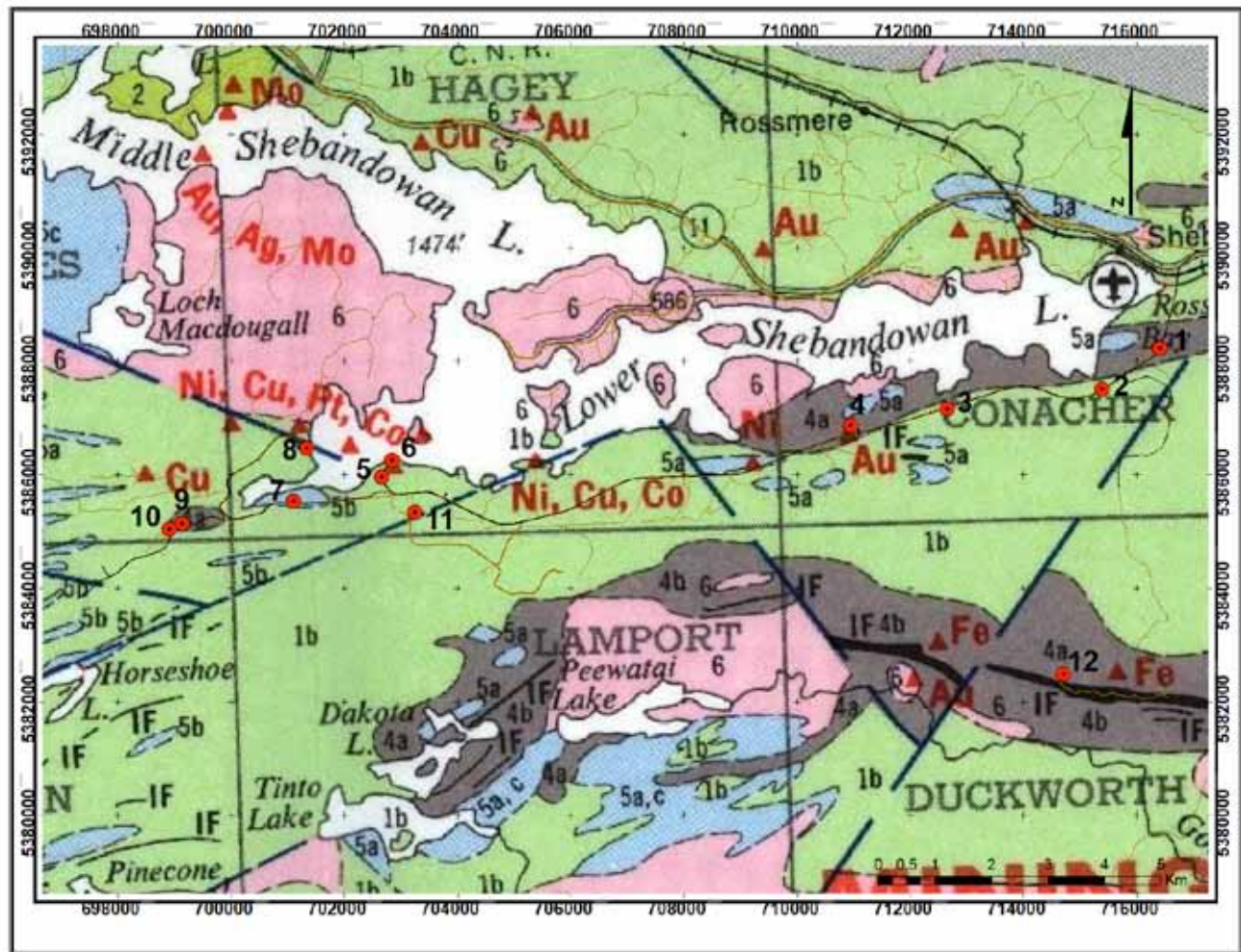


Figure 2 – General Geology – Shebandowan area. From Pye and Fenwick (1965)

Timiskaming-type rocks range in age from approximately 2700 Ma to about 2680 Ma (Card, 1990; Corkery et al., 2000). They unconformably overlie older metavolcanic sequences hereafter referred to as “Keewatin”. Timiskaming-type metavolcanic rocks usually are shoshonitic (high Al_2O_3 and K_2O with TiO_2 content < 1.3 wt.%) and are associated with fluvial meta-conglomerates and meta-sandstones.

The Timiskaming-type rocks in the Lower Shebandowan Lake area consist of calc-alkaline metavolcanic rocks and alluvial-fluvial meta-conglomerates and meta-sandstones. The metasedimentary rocks are immature, typically with cross-bedded arenites and conglomerates with some shale and ironstone. The metavolcanic rocks are calc-alkaline to alkali/shoshonitic subaerial volcanic rocks. They all display rapid facies changes and internal unconformities and typically only display late deformational and metamorphic events.

The komatiitic bodies of the Shebandowan area

have traditionally been considered intrusive bodies. While no spinifex textures, considered diagnostic of flow emplacement, have been found there are other features present that indicate they were deposited as flows. These include the fact that they are stratabound and commonly have ironstone or chert beds along one contact. In addition, generally accepted notions that ultramafics are intrusive does not take into consideration that the density of molten ultramafic rocks is too high to allow emplacement by density contrast and therefore must have relied more on processes such as over-pressure. This, combined with the fact these rocks are typically associated with extensional environments makes emplacement by intrusion, especially when they are conformable to local stratigraphy, extremely difficult to explain except by extrusive processes.

Road Log

Drive west from Thunder Bay on Highway 11-17. At Shabaqua turn left onto Highway 11 and then drive 12.4 km to the Junction of Highway 11 and the Shebandowan Mine Road. Turn Left.

0 km - Junction of Hwy 11 and Shebandowan Mine Road

0.7 km – bridge over Shebandowan River

1.9 km – Stop 1

3.1 km – Stop 2

5.7 km – Stop 3a

5.8 km – Stop 3b

5.9 km – Stop 3c

7.6 km – trail into Stop 4. Follow trail for 250 m, keeping to the left

15.9 km – gate barring entrance to Shebandowan mine site

16.4 km – junction with road to No. 1 shaft.

16.7 km – Stop 5.

17.1 km – Stop 6.

Return to turn off to No. 1 Shaft. Reset trip meter. Turn Right (west).

1.7 km – Stop 7.

2.9 km – junction with Otto Lake road – keep to the right.

4.9 km – junction with road – keep to the right.

5.5 km – Stop 8.

Return to Otto Lake road junction, reset trip meter and then turn right.

0.9 km – Stop 9.

1.2 km – Stop 10.

Return to Gate house.

Take the road to the right. Drive 325 metres and turn left and drive 120m to Stop 11.

Return to Gate House and turn right and head east.

13.2 km – Junction with Duckworth Road – Turn Right.

17.6 – Junction – keep to the right.

18.5 – Junction – keep to the left.

19.0 – Junction with I Zone Road – turn right.

20.5 – Stop 12.

Field Trip Stop Descriptions

Stop 1. Timiskaming-type metavolcanic debris flow with intercalated sediments.

UTM coordinates NAD83; 15U 0716408E / 5388223N

The relatively undeformed metavolcanic debris flows at this stop consist of poorly sorted sub-angular to well-rounded fragments in a fine- to medium-grained tuffaceous matrix. A characteristic feature is the presence of hornblende phenocrysts in both the fragments and the matrix. Locally what appears to be graded bedding within the debris flows is present. One such locality is on the east side of the road where the debris flows are in contact with an intercalated greywacke unit. Here fragments within the debris flow fine to the south. On the west side of the road the same greywacke unit shows evidence of folding (fold axis plunging vertically) with graded bedding in the north limb indicating a synclinal structure.

Compositionally the debris flows vary from basalt (<53% SiO₂) to rhyolite (Brown, 1985). Shegelski (1980) has shown that they represent a typical calc-alkaline volcanic suite with shoshonitic affinities.

The presence of red pigmentation in these debris flows has led to much speculation as to their depositional environment. Pigmentation has resulted in clasts and matrix ranging from grey-green to red in colour. Locally grey-green clasts exhibit hematized rims while others are totally hematized. There are several possible processes that may have produced this red colouration: hydrothermal alteration after lithification; magmatic differentiation; deposition in an Archean oxygenated atmosphere resulting in red bed formation; or secondary oxidation during the Proterozoic or Phanerozoic. By plotting the ratios of ferric iron oxide (Fe₂O₃) to ferrous oxide (FeO) against SiO₂, Shegelski (1980) ruled out magmatic differentiation. He then postulated that the red pigmentation was the product of red bed development.

Due to the variability in red pigmentation doubt still remains as to whether it is a product of red bed development. In particular is the spectrum of pigmentation, including red fragments in a grey-green matrix, green fragments in a reddish matrix and fragments that are partially red and partially green. It's due to this variability that has led others, such as Brown (1985), to believe the pigmentation is the product of varying degrees of hematization subsequent to deposition.

Stop 2. Timiskaming-type conglomerate.

UTM coordinates NAD 83; 15U 0715383E / 5387509N

At this location two facies of the epiclastic suite of Timiskaming-type rocks are exposed. The dominant rock-type is poorly sorted, highly foliated meta-conglomerate (Fig. 3). 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 meta-conglomerate to the west are meta-mudstone and meta-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 3. Timiskaming-Keewatin contact.

UTM coordinates NAD 83; 15U 712653E / 5387161N

Here we will examine what may possibly be the



Figure 3. Timiskaming Conglomerate at Stop 2

sheared unconformity between the Timiskaming-type metasedimentary rocks and the underlying Keewatin metavolcanic rocks (15U 0712637E / 5387160N). Relatively pristine Keewatin felsic metavolcanic rocks are exposed to the west (15U 0712548E / 5387139N). Note the intense deformation of the meta-conglomerate on the north side of the road with deformation intensity increasing to the south. On the south side we have exposed highly foliated and carbonatized rocks that may or may not be Timiskaming-type metasedimentary rocks in contact with relatively undeformed, possibly Keewatin fragmental rocks.

On the north side of the road (15U 0712742E / 5387227N) there are bands of conglomerate with a significant proportion of barren sulphide pebbles and cobbles. Also present are several felsic intrusive rocks that cut through at a low angle to the stratigraphy and which are also strongly carbonatized.

Stop 4. Timiskaming-type Monzonite.

UTM coordinates NAD 83; 15U 0710946E / 5386868N

On the north side of the mine road is an overgrown logging road. Follow it, keeping to the left, for 230 metres.

At this location, and several other outcrops to the east, we have exposed a small intrusion within the Timiskaming debris flow pile. This intrusion is interpreted to be a high level magma chamber that was parent to the volcanic system that produced the debris flows. Macroscopically the rock resembles the fragments found in the debris flows; both possess hornblende phenocrysts and have the same mottled green to red colourization. This pigmentation within an intrusive environment indicates that the reddish colourization was not the product of exposure to an oxygenated atmosphere but is more likely related to oxidizing deuteritic fluids. The restriction of this alteration to the Timiskaming-type igneous rocks, both intrusive and extrusive, limits other possible interpretations.

Stop 5. Deformed debris flows.

UTM coordinates NAD 83; 15U 0702659E / 5385962N

This outcrop consists of strongly deformed hornblende-phyric Timiskaming-type debris flows. Note the highly foliated nature reflecting its proximity to the Crayfish Creek Fault, approximately 100 metres to the north. Though strongly foliated note the many

similarities with the undeformed debris flows seen at Stop 1.

Stop 6. Shebandowan Mine Ultramafic Rocks.

UTM coordinates NAD 83; 15U 0702841E / 5386253N

This stop is beside the now capped production shaft used to extract the nickel-copper sulphide ore from the Shebandowan deposit. During its operation it produced 9.4 million tonnes grading 1.7% Ni, 0.9% Cu and 1.56 g/t total precious metals (Pt + Pd + Au).

The outcrop consists of serpentized peridotite (see photo below) with numerous narrow zones of talc-carbonate schist that form an anastomosing network. It is this unit that is host to the nickel-copper sulphides. On the north side of the outcrop is a feldspar porphyry dike, an apophysis of the Shebandowan Lake Stock to the north, cutting across the peridotite.

Stop 7. Ultramafic with iron formation.

UTM coordinates NAD 83; 15U 0701100E / 5385538N

This rock cut is through the contact between an ultramafic massive flow and an oxide-facies

iron formation. The south outcrop is serpentized peridotite and the iron formation is on the north side. There is another iron formation on the south side of this ultramafic unit (Fig. 4). Geological mapping by Morton (1982) found that tops in this area are to the south. Just to the west there are a series of outcrops of massive mafic flows with flow top breccias that confirm tops are to the south.

Stop 8. Discovery Point.

UTM coordinates NAD 83; 15U 0701334E / 5386459N

The first sign of nickel mineralisation of what eventually became the Shebandowan Mine was made in 1913 by Jules Cross. At the time he was mapping for the Ontario Department of Mines and while doing mapping along the shoreline noted signs of nickel and copper mineralisation. The original showing can still be seen in the form of several pits right at the water's edge at Discovery Point.

Stop 8a.

UTM coordinates NAD 83; 15U 0701350E / 5386512N

Walk east on the road 75 metres. On the left is the

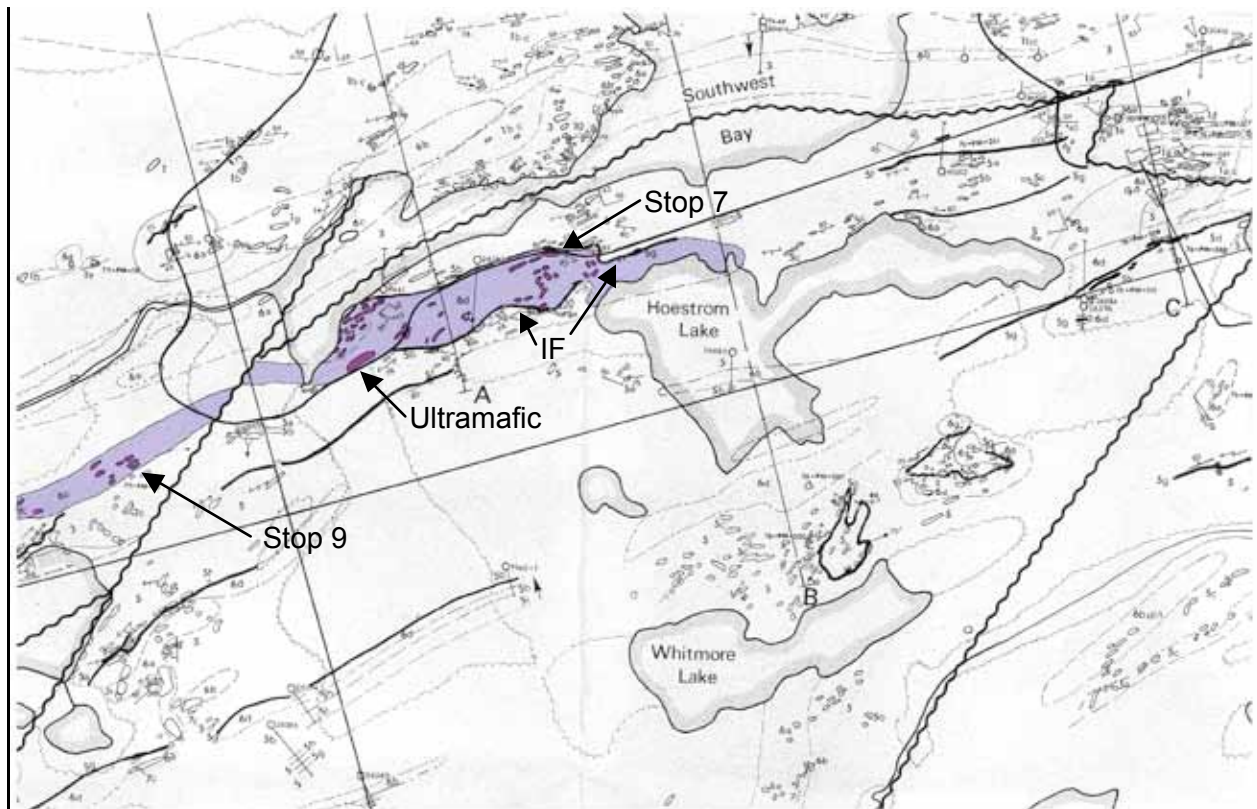


Figure 4. Detailed geology of the Southwest Bay area (Morton, 1982).

concrete cap over the No. 1 Exploration shaft that was used for initial access to the Shebandowan ore body.

Stop 8b

UTM coordinates NAD 83; 15U 0701323E / 5386451N

Return to vehicles. Walk 90 metres to the west until you see an old road on the left that works its way east down and along the hillside. Continue 120 metres to the bottom where there was once a boat landing at the waters edge. Work your way along the edge of the bay east for 50 metres to several old pits very close to the water. Here the mineralisation is at the north edge of the mine peridotite body where it is in contact with sheared and banded mafic metavolcanic rocks. Look for sheared peridotite with signs of nickel bloom (Fig. 5).



Figure 5. Cu-Ni Mineralisation at Discovery Point (Stop8b).

Stop 9. Contact between ultramafic flow and overlying felsic fragmental unit.

UTM coordinates NAD 83; 15U 0699131E / 5385139N

This ultramafic flow unit is a fine-grained peridotite with well-developed polyhedral jointing. It is cut by a felsic dike. Locally the contact with the overlying felsic debris flow is exposed and is characterised by a thin, strongly foliated black chert horizon capping the ultramafic, evidence that this unit was deposited as a flow.

Stop 10. Fragments of Black Chert in the base of the

debris flow.

UTM coordinates NAD 83; 15U 0698906E / 5385040N

Here we have several examples of angular pieces of black chert ripped up off the chert horizon capping the ultramafic (exposed just to the north on the other side of the old logging road) and incorporated into the base of the felsic debris flow unit.

Stop 11. Pillowed Mafic Feldspar Phyric Flows.

UTM coordinates NAD 83; 15U 0703244E / 5385343N

Here we have a unit that probably can be used as a stratigraphic marker horizon as it has been found at several locations over a 5 kilometre strike-length. It consists of obvious pillows with well developed selvages but with feldspar phenocrysts, some up to several centimetres across. Note how the crystals are smaller as you approach the pillow margins. Tops are consistently to the south.

Stop 12. Iron formation hosting felsic dike with gold-bearing quartz ladder veins.

UTM coordinates NAD 83; 15U 0714705E / 5382490N

At this stop we have Timiskaming oxide-facies iron formation intercalated with argillite intruded by a later felsic dike (Fig. 6). This dike is host to gold-bearing quartz ladder veins (Fig. 7). Fractures opened up in the dike due to the ductility contrast of the enclosing iron-rich argillites and the felsic dike. Later auriferous



Figure 6. Felsic dike cutting iron formation at Stop 12.



Figure 7. Auriferous quartz ladder veins within the dike at Stop 12.

fluids, likely carrying gold as thio complexes reacted with the iron oxides resulting in the formation of pyrite and precipitation of native gold. If one looks carefully at the exposed contact of the dike, where the iron formation has been eroded away and looking for pyrite concentrations you commonly will also find fine- to coarse-grained native gold. Figure 8 (from Aubut et al., 1990) shows the simplified geology of this location.

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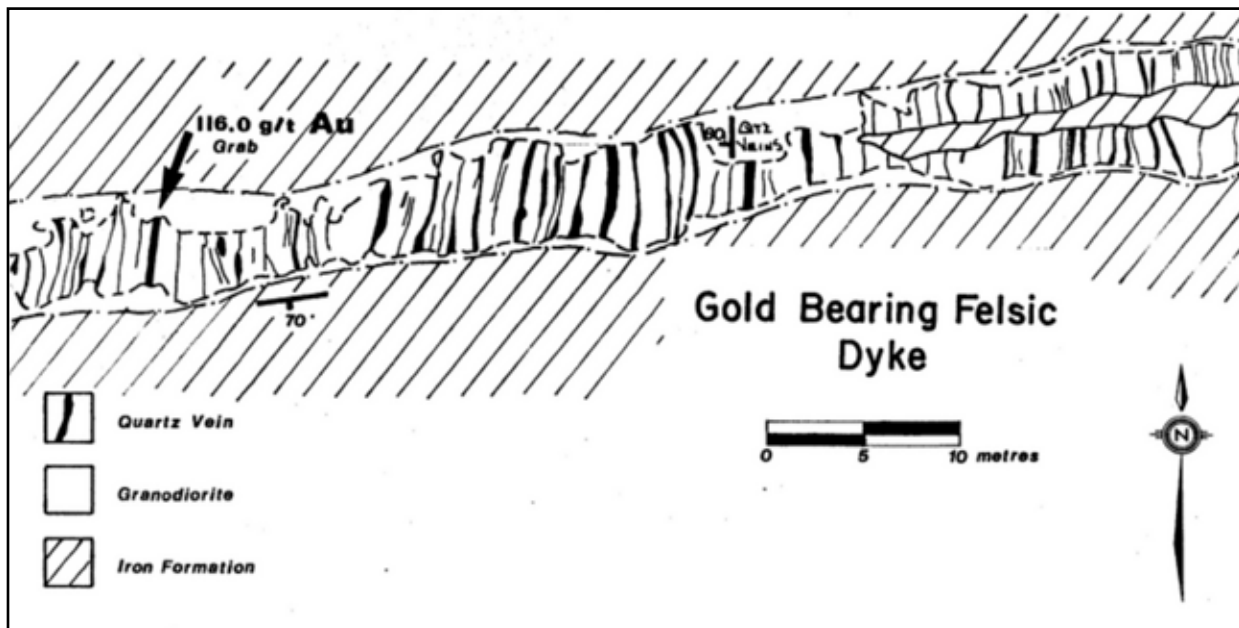


Figure 8. Simplified geology of Stop 12. From Aubut et al. (1990)

Field trip 5 - Guide to the Thunder Bay area

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Regional Geology

Thunder Bay is situated on the northern margin of the Southern Province of the Canadian Shield. The Southern Province consists of Proterozoic rocks which unconformably overlie Archean basement rocks of the southern Superior Province (cf. Tanton, 1931; Pye, 1969). Archean basement rocks of the Wawa Subprovince are predominantly granitoid plutons and slivers of greenschist- to amphibolite-facies supracrustal (i.e. greenstone belt) rocks (Williams et al., 1991; Fig. 1).

The Paleoproterozoic Animikie Group 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 towards the center of the Mesoproterozoic Midcontinent Rift (MCR) 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). These rocks grade 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).

The Sibley Group, exposed on the nearby Sibley Peninsula, has been subdivided into five formations;

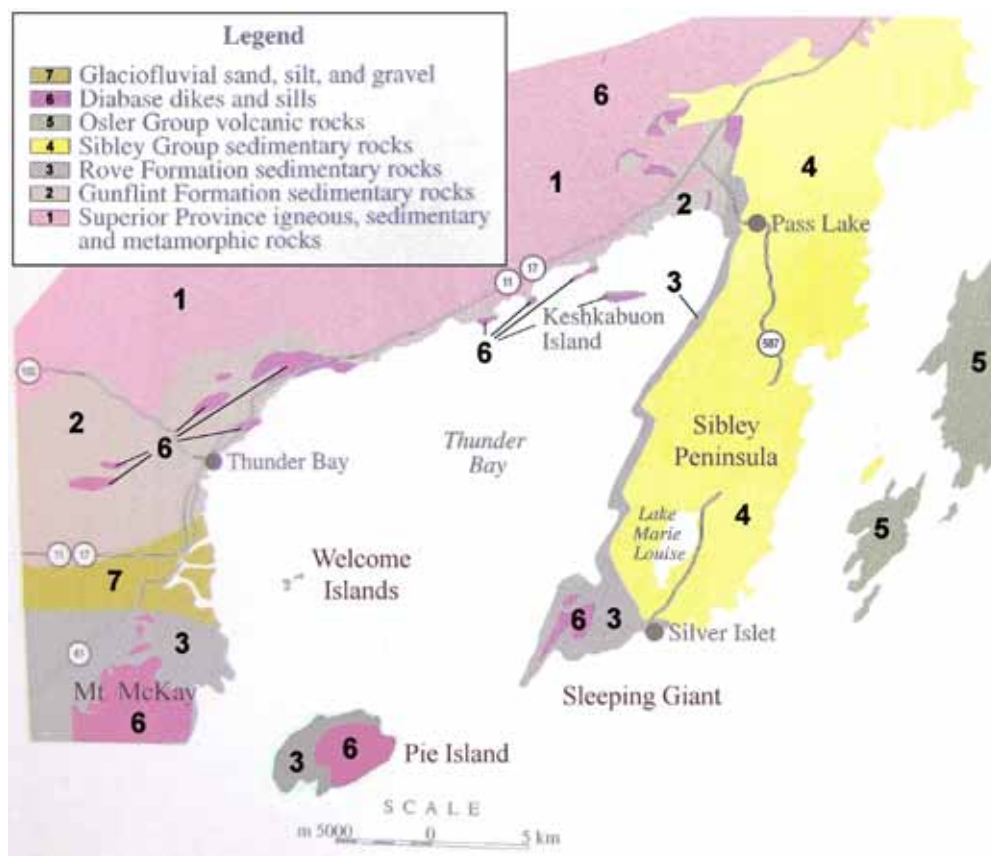


Figure 1. General geology of the Thunder Bay area, modified after Pye (1969)

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 (Kama Hill Formation). The Outan Island Formation represents the transition from subaqueous to subaerial conditions, and the Nipigon Bay Formation represents an aeolian environment (Rogala, 2003; Rogala et al., 2007). The depositional age for much of the Sibley Group is constrained between ~1340 and 1450 Ma.

The northern margin of the Midcontinent Rift (MCR) is dominated by hypabyssal rocks of the Mesoproterozoic Midcontinent Rift Intrusive Supersuite (Miller et al. 2002), which intrude Paleoproterozoic Animikie Group and Mesoproterozoic Sibley Group sedimentary rocks and Archean basement (Fig. 2). Older hypabyssal rocks (1124 Ma Seagull intrusion, 1109-1113 Ma sills; Heaman et al., 2007) predominate in the Nipigon Embayment (cf. Hart and MacDonald,

2007). 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. Osler Group rocks are also intruded by mafic dykes and intrusive complexes (1095 Ma Moss Lake gabbro, Heaman et al., 2007; 1089 Ma St. Ignace Complex, Smyk et al., 2006) which represent the youngest local MCR magmatism. Ages in this part of the MCR range from ca. 1140 Ma (Heaman and Easton, 2007) to ages younger than the magnetic polarity reversal that occurred between 1105 and 1102 Ma (Davis and Green, 1997). Hollings et al. (2007a) proposed that the term Logan Igneous Suite, which would fall within the Midcontinent Rift Intrusive Supersuite of Miller et al. (2002), should be applied to all the diabase sills in the area north of Lake Superior, with subdivision into the informal terms, Nipigon sills for the sills north of Thunder Bay, and Logan sills to the south.

Starting about 11,000 years ago (Ka), Wisconsinan ice melted back from its position in central Minnesota and Wisconsin, and quickly exhumed the Thunder Bay region, forming recessional moraines during brief still-stand periods (Phillips, 2004; Phillips et al., 1994).

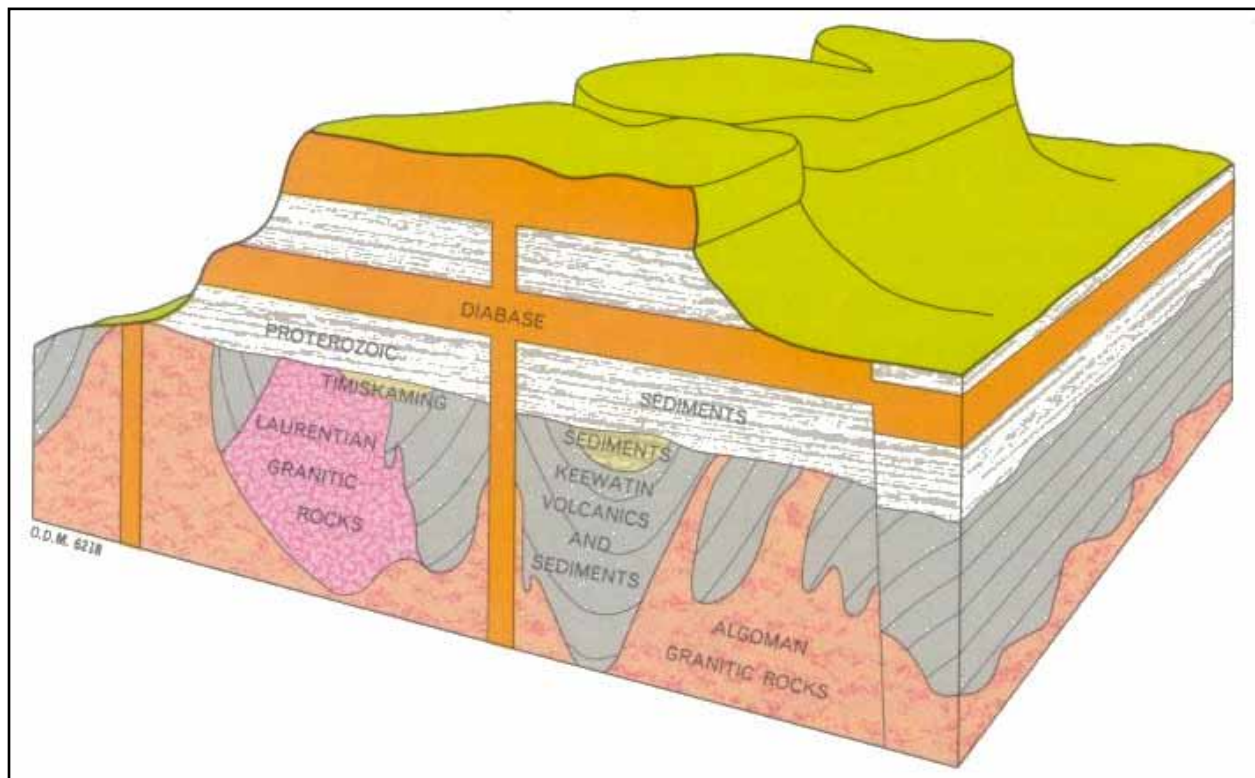


Figure 2. Schematic block diagram illustrating local stratigraphy (Pye 1969).

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 Re-advance). As that ice began to melt, glacial lakes were formed between the moraines and the retreating ice margins. As Superior ice melted, water levels 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 tool-kit 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).

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 most recently during the 46th Institute on Lake Superior Geology (e.g. Pufahl et al., 2000; Phillips et al., 2000).

Stop Descriptions

Stop 1A: Hillcrest Park area

UTM coordinates: NAD83; 16U 0334689E / 5366961N

On a clear day, Sibley Peninsula and the Sleeping Giant

Giant can be seen across the waters of Thunder Bay (Fig. 3). The 240 m high cliffs facing Thunder Bay are the highest in Ontario. The Sleeping Giant is capped by a Logan diabase sill which has intruded Rove Formation shale and sandstone. Other prominent mesas and cuestas include Pie Island and Mount McKay and the other hills of the Nor'Wester range to the south. All of these hills consist of Rove Formation sedimentary rocks capped by Logan sills. Isle Royale (Michigan), visible on the distant horizon, consists of Keweenaw basalt flows (Portage Lake Volcanics) associated with the Midcontinent Rift.

Raised beaches, which represent former lake levels of glacial Lake Minong, are visible within the city and extend westward up the Kaministiquia River valley. The most prominent of these in downtown Thunder Bay North, extending along Algoma Street, is associated with the Nipissing Great Lakes stage (ca. 5500 years ago), approximately 20 m above present-day Lake Superior.

The large bell at this lookout rests upon the "Upper Limestone" member of the Gunflint Formation. This

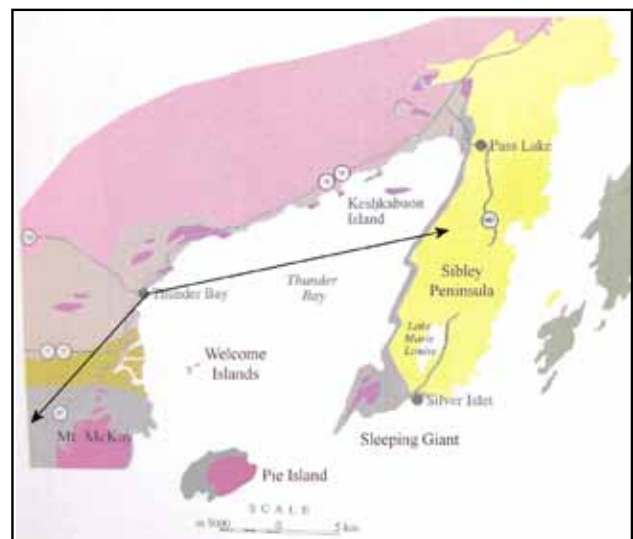


Figure 3. Panoramic view (ca. 130 °) of Thunder Bay from Hillcrest Park (image from <http://www.360cities.net/image/hillcrest-park-thunder-bay#106.03,-0.78,54.2>) Map above depicts field of view from this vantage point.

unit is now interpreted as carbonatized ejecta related to the Sudbury impact event (ca. 1850 Ma) [see Field Trip guide 1, this volume]. Proceed down the concrete stairs, turn right and walk along the laneway. The ejecta here is characterized by debrisite breccia and accretionary lapilli (Fig. 4). Calcareous beach rock contains algal bioherms and basal cryptalgal laminites with fenestral fabric which are overlain by coarse-grained, poorly sorted breccia.

Stop 1B: Debrisite Breccia

UTM coordinates: NAD83; 16U 334163E / 5366301N

(n.b. Private Property, ask for permission to access)

Another spectacular debrisite breccia (Fig. 5a) is exposed at the corner of Markland and Hill streets. This outcrop was mapped in detail by Shegelski (1982; Fig. 5b) and later described in the context of impact-related brecciation by Addison et al. (2010):

A bedrock exposure, ~5 m by 15 m, in a private yard...contains a spectacular exposure of Gunflint chert-carbonate breccia and ejecta, primarily devitrified vesicular impact glass, which are 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 ~30 cm below the erosive contact between the debrisite and the Gunflint bedrock.

The most common ejecta feature is devitrified vesicular impact glass clasts 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.

Our route then takes us west along Highway 11-17 toward the community of Kakabeka Falls. The route follows the Kaministiquia River valley, which is dominated by fluvio-lacustrine deposits related to proglacial lakes (Burwasser, 1977). Tills associated with Superior lobe ice predominate north of the valley, extending southward from the rolling hills of exposed Archean rocks. In contrast, the landscape south of the highway consists of flat plains underlain by flat-lying Gunflint and Rove Formation sedimentary rocks and Quaternary sediments, punctuated by diabase-topped cuestas. Just west of the junction of Highway 11-17 and the Highway 588 (Stanley) turn-off, glacio-fluvial gravels and sands of the Stanley delta formed where the Kaministiquia River entered into Lake Beaver Bay ca. 9.7 Ka (Phillips, 2004; Phillips et al., 2004). A well-formed bluff, representing a lower Beaver Bay phase (260 m / 853 feet) extends along the north side of the highway. The present-day river has deeply incised the delta.

The town of Kakabeka Falls is built on the floor (at 277 m ASL) of an old distributary of the Kaministiquia River which cut through a higher terrace level. This terrace (~300 m / 984 feet) represents the highest level of the Stanley delta. The Crane archaeological site is found on the west side of the river, where the old river entered Lake Beaver Bay (Phillips et al., 1994, 2000; Phillips, 2004).



Figure 4. Accretionary lapilli in debrisite, Hillcrest Park (Stop 1A).



Figure 5a. Debrisite breccia, corner of Markland and Hill streets, Stop 1B.

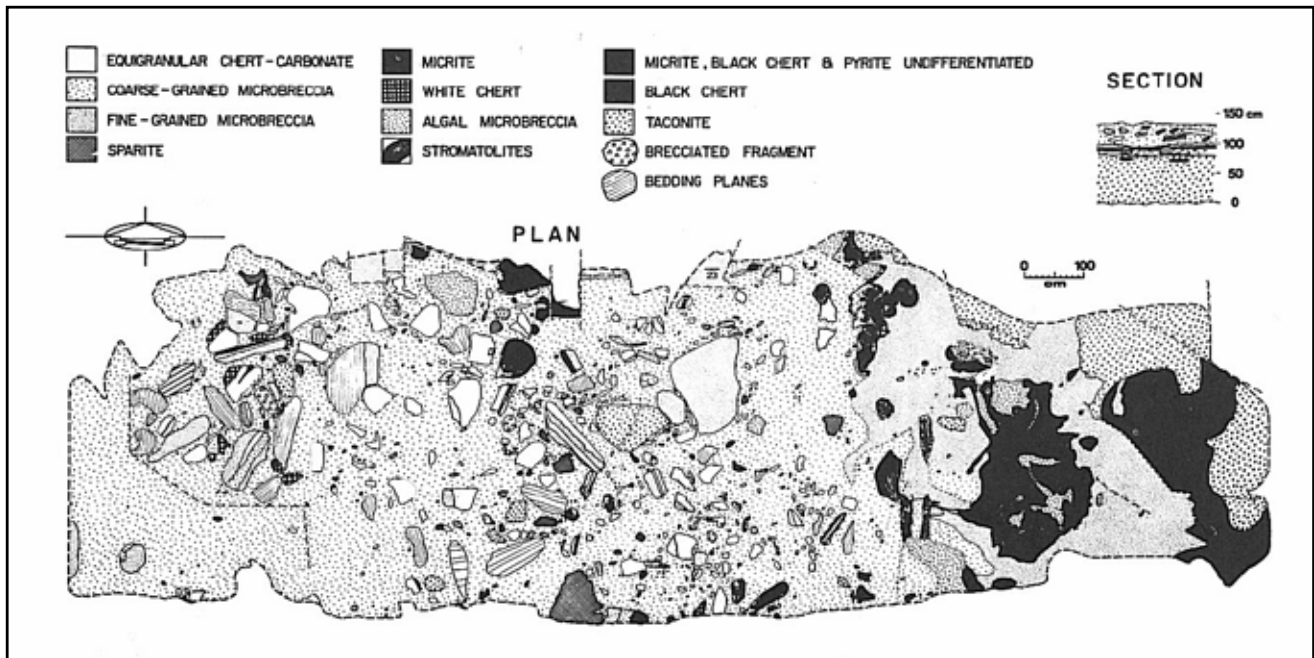


Figure 5b. Detailed map of the debrisite breccia outcrop at Stop 1B by Shegelski (1982).

Stop 2: Kakabeka Falls Provincial Park

Stop 2A: Junction of Highways 11-17 and 590

UTM coordinates: NAD83; 16U 0305176E / 5364668N

Walk uphill on the west side of Highway-590. Highway excavation has revealed the sharp unconformity between Archean, gneissic, felsic plutonic rocks and the overlying Paleoproterozoic Gunflint Formation. Large, domical stromatolites occur on “highs” on the eroded Archean basement (Fig. 6). The associated sedimentary rocks, representing intertidal, foreshore sedimentation, are laminated cryptalgal cherts overlain by a wavy bedded grainstone-micrite facies. The latter is capped by chaotic, slumped, laminated chemical sedimentary rocks which are probably cryptalgal. A few metres further uphill, overlying the previous sequence is a brecciated and slumped pyritic black chert. Gunflint conglomerate (Kakabeka Member), exposed in the river gorge a few hundred metres to the north, forms the basal member of the Gunflint Formation in this area. A major, northeast-trending fault has resulted in a down-dropping of the block to the southeast. The Kakabeka gorge, ~600 m to the east, exposes rocks much higher up in the Gunflint stratigraphy.

As noted by Pufahl et al. (2000) the first set of rapids above the highway bridge are formed by Archean

granitoids. The slow-water area to the south is underlain by the Gunflint Formation. Kakabeka conglomerate patchily overlies the Archean basement. Silicified stromatolites are developed on the conglomerate or directly on the granitic basement. This is the location from which samples collected in the 1950’s yielded the first Gunflint cyanobacteria described in the literature. The samples, from the silicified stromatolites, were described by Tyler and Barghoorn (1954).



Figure 6. Unconformity between Archean granitoids and Gunflint Formation, Stop 2A.

Stop 2B: Kakabeka Falls

UTM coordinates: NAD83; 16U 0305738E / 5364400N

(n.b. Entry/parking fee is required in Kakabeka Falls Provincial Park. No sample collecting permitted)

The park is dominated by a single, spectacular feature, Kakabeka Falls, which drops 39 m over sheer cliffs in Gunflint Formation sedimentary rocks. 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 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, lowermost chert-carbonate bed of the Gunflint Formation which forms a resistant cap rock to the softer underlying shales. Looking down the gorge, one can observe a lapilli-tuff member as a lighter grey unit near the base overlain by a thick sequence of black shales (Fig. 7). 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. As noted by Pufahl et al. (2000), this sequence

represents the major volcanoclastic horizon present in the upper Gunflint Formation and is traceable to the south as the Biwabik Formation through the Mesabi Range. Basalts outcropping approximately 30 km to the southwest are probably correlative with this unit.

The outcrop on the northern edge of the parking lot contains layers of banded 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 (cf. Pufahl et al., 2000).



Figure 7. The gorge below Kakabeka Falls, developed in Gunflint Formation shales and tuffs (www.Audreyhansen.com)

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Field trip 6 - Thunder Bay Amethyst Mine

Stephen Kissin

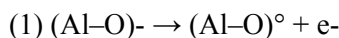
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Introduction

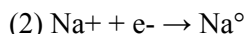
Properties of Amethyst

Amethyst, occurring in abundance in the Thunder Bay region, is a purple gemstone variety of 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 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.

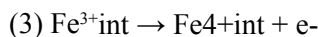
In a series of papers by Cohen and co-workers, culminating in a summary in Cohen (1989), a simultaneous sequence of reactions was proposed for the formation of amethyst.



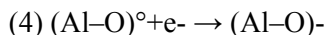
Ionizing radiation forms a hole center from oxidizing the substitutional Al-O bond.



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



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



Trapped hole center is satiated as $[\text{AlO}^\circ]$ is reduced via gaining the electron from step 3.

The presence of iron in 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^+ into interstitial sites. The color of amethyst is produced by absorption of light in the visible region of the spectrum

owing to the presence of Fe^{4+} , as originally shown by Cox (1977).

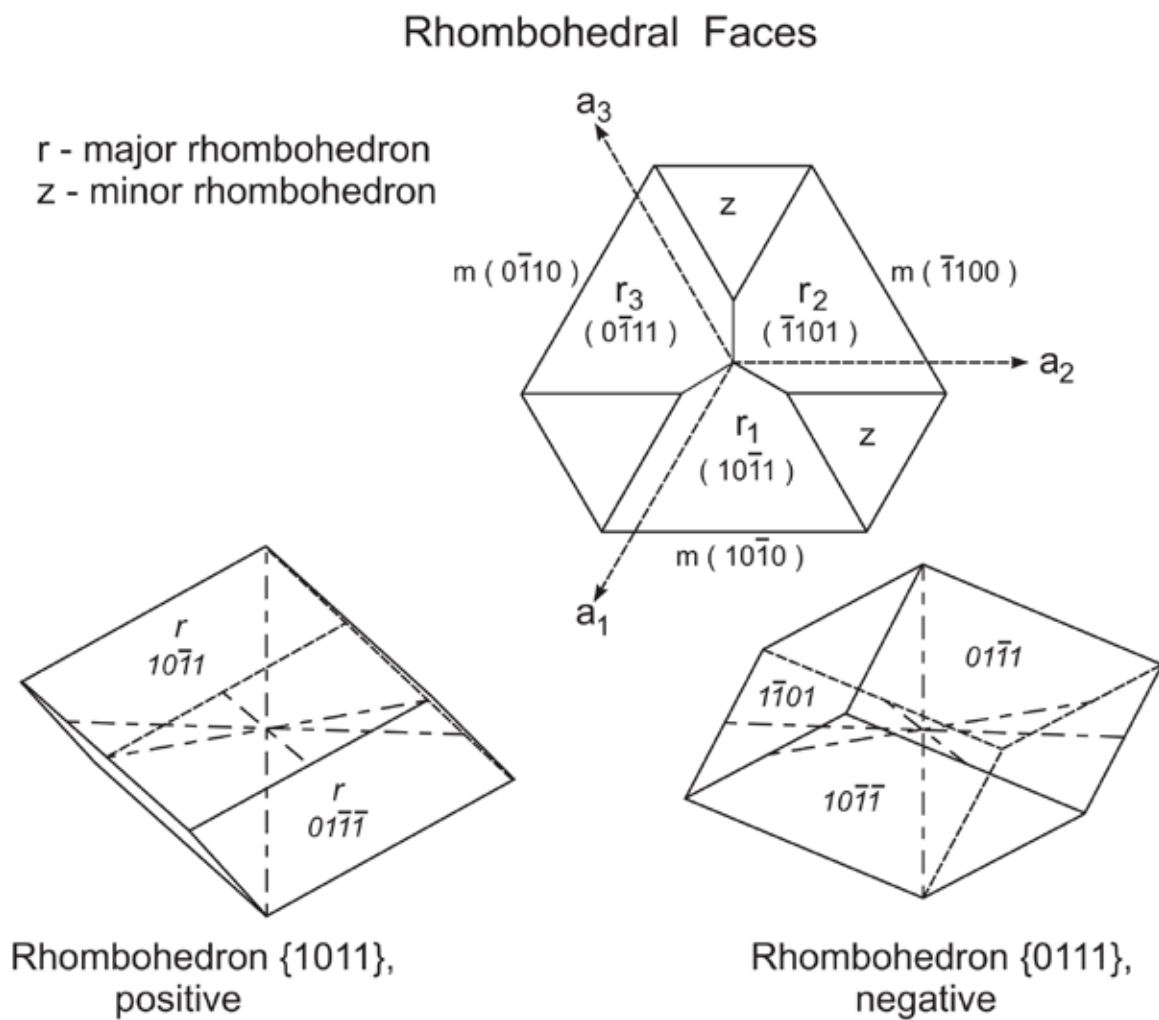
The proposed mechanism requires the coincidence of four geological conditions for the formation of amethyst:

- (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 $\{101\bar{1}\}$ and negative $\{011\bar{1}\}$ 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 (Fron del, 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; Fron del, 1962). The orientation of Brazil law twins in Figure 2 is typical 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 $\{101\bar{1}\}$, 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



Distribution of rhombohedral forms.

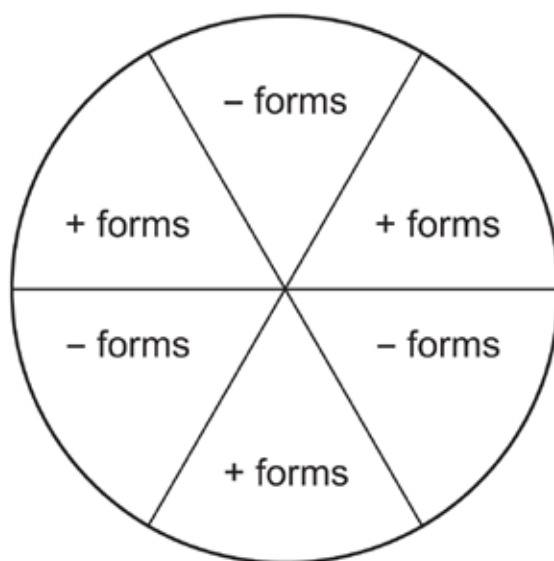


Figure 1. A typical amethyst crystal viewed perpendicular to the c-axis, illustrating the combination of positive $\{101\bar{1}\}$ and negative $\{011\bar{1}\}$ rhombohedra.

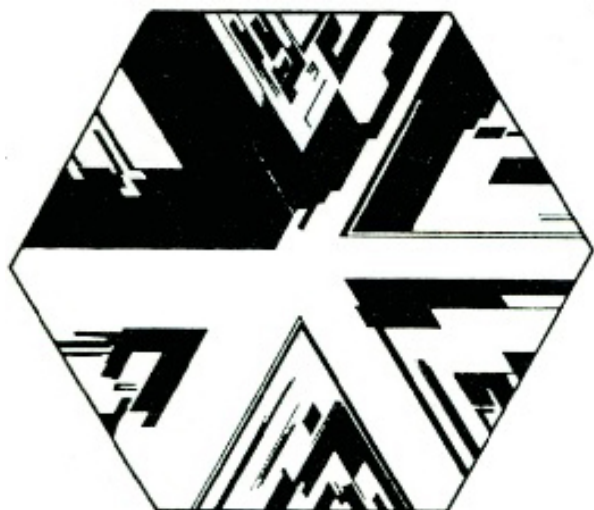


Figure 2. Etched basal α -quartz illustrating the typical occurrence of Brazil law twinning in which alternate bands contain left- and right-handed α -quartz (after Frondel, 1962).

incorporation of Fe^{3+} and that iron is preferentially concentrated along this composition plane in amethyst.

Amethyst Deposits in the Thunder Bay Area

In this 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. However, it was not until 1862 that amethyst was commercially exploited in various mines associated with lead-zinc and silver deposits. 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, inexpensive amethyst from Brazil.

In 1961, the construction of a road leading to a now abandoned fire tower revealed a very large amethyst deposit, which is now known as the Thunder Bay Amethyst Mine Panorama. In large vugs near the surface of the vein deposit, amethyst crystals 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). 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).

Geology of the 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 Franklin et al. (1986) and will not be repeated in detail here. The 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 Shebandowan 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 (1978a) 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 Thunder Bay Amethyst Mine, although its presence as abundant fragments in mineralized breccias within the vein system indicates that these sediments 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 is deposited 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 (Keweenaw) 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 $1537 \pm 10/-2$ 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 intracratonic 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

The Thunder Bay Amethyst Mine is located within a first-order strike-slip fault, which strikes at $90-100^\circ$ and dips steeply to the south. This fault is roughly parallel to one 2.1 km to the south, which strikes east-northeasterly 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 $162 - 150^\circ$ 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

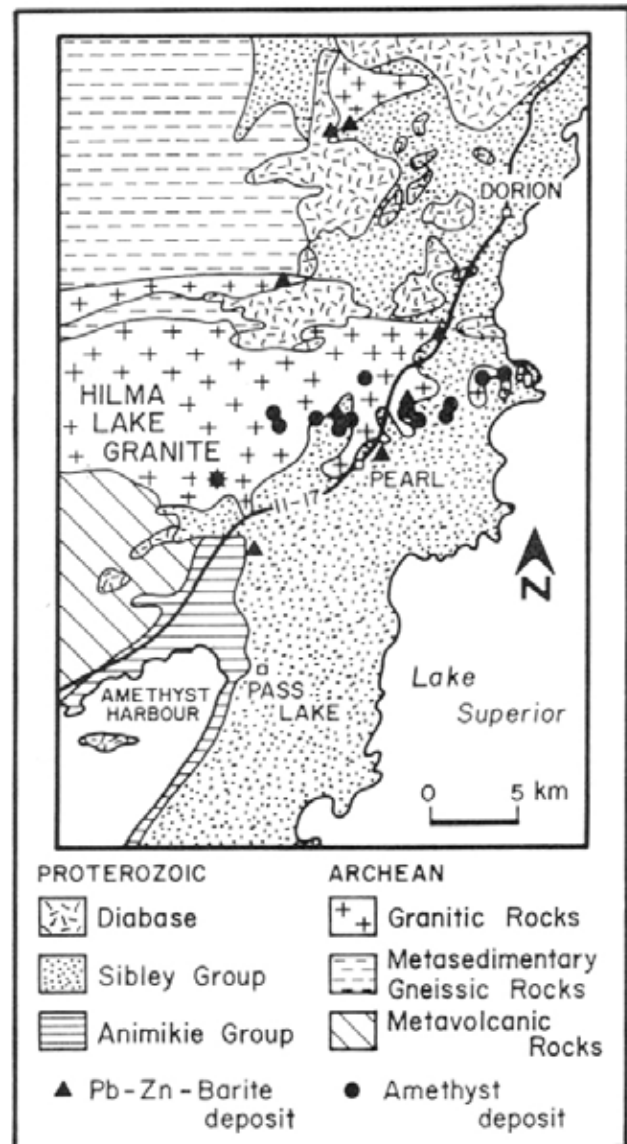


Figure 3. Local geology and location map of amethyst deposits and lead-zinc-barite deposits, showing the relationship of the former to the margin of the Sibley Group outcrop and the Hilma Lake granite. The location of the Thunder Bay Amethyst Mine is indicated by the star. Modified after Patterson (1985).

extension of the vein system to the east was developed, offset to the north by a few metres by a 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.

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

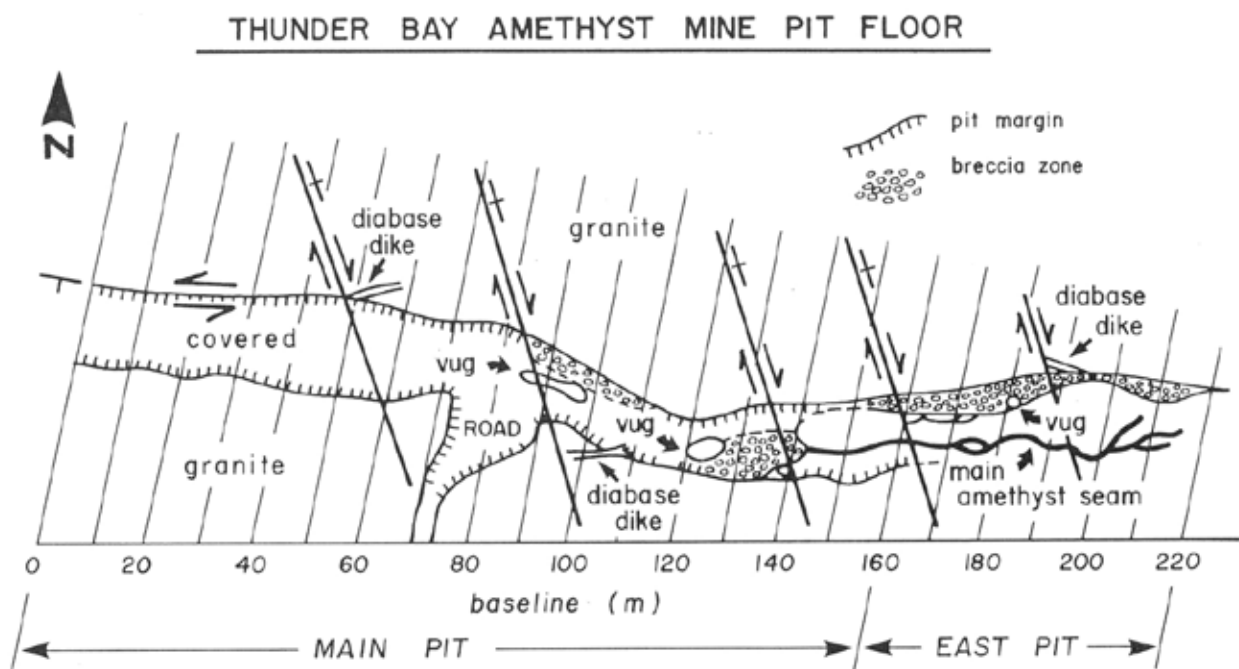


Figure 4. Diagram of the main pit of the Thunder Bay Amethyst Mine.

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. 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 several varieties of quartz occur in the Thunder Bay Amethyst Mine, including colorless quartz; chalcedony; amethyst; the yellowish variety, citrine; and the greenish variety, prasiolite or "greened

Table 1. Paragenetic sequences observed in the veins of the Thunder Bay Amethyst Mine

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.

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 1 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 baroque."

The compositions of specimens of amethyst by neutron activation analysis for selected trace elements (Table 2) 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.

Other non-sulfide minerals. Barite is rare in the

Table 2. Analyses of r-zones of amethyst for selected trace elements (in ppm; Kissin, 1997)

<u>Sample No.</u>	<u>Fe</u>	<u>Al</u>	<u>Ge</u>	<u>Ti</u>
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.

veins at the Thunder Bay Amethyst Mine, 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-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) 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 Thunder Bay Amethyst Mine were discussed in detail by McArthur et al. (1993). The conclusions of their study are given below; however, for details of the evidence, their paper should be consulted. Genetic speculations on the Thunder Bay Amethyst Mine 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 H₂S 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 $NaCl-CaCl_2-H_2O$ 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 Thunder Bay Amethyst Mine 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 Thunder Bay Amethyst Mine 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 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 Keweenaw 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 Keweenaw 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, and some additional work is underway in an effort to resolve this remaining question.

Summary

Field and laboratory studies of the Thunder Bay Amethyst Mine 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 an 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-CaCl₂-H₂O 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 $\delta^{18}\text{O}$ 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 $\delta^{34}\text{S}$ of -0.4-0.6 ‰ and -1.4 ‰ in chalcopyrite. These volumes are consistent with derivation from H₂S 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 H₂S 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 Thunder Bay Amethyst Mine 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 Hebert and Rossman (2008), who attributed the development of greenish-grey to greenish quartz to the presence of H₂O in the crystal. Our work (Klärner 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”. Work on this problem is continuing.

Road Log Airline Travelodge to Thunder Bay Amethyst Mine

Leaving the Airline Travelodge, 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.

At 5.7 km is the intersection with Oliver Road, which leads to Lakehead University about 2 km to the east. 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 on at 6.7 km, outcrops of a Logan diabase sill 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. Passing the junction of Red River Road (Highway 102) at 10.0 km, the expressway is on a level stretch marking the top of a Logan sill.

The expressway then passes downhill to the Current River at 15.9 km. 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 expressway ends and the highway is again cutting into a diabase sill. A fault trends along the highway offsetting the sill on opposite sides of the highway. A few hundred metres further 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 at 20.1 km, 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 metavolcanic rocks was exposed on the left of the highway. Hopefully, new road construction will expose this unconformity once again. The hill is formed by the outcrop of the Mackenzie granite, an unmetamorphosed and undeformed, late Archean pluton. The highway continues on top of the granite, which contains occasional roof pendants of Archean metavolcanics.

After crossing the Mackenzie River at 35.7 km sparse outcrops of granite are replaced by poorly exposed Gunflint Formation until just past the junction with Highway 587 at 49.6 km. Here, well-bedded red-stained carbonates of the Gunflint Formation crop out beside the highway. Passing onward to the East Loon Road at 55.2 km, turn left onto the road, then right on Bass Lake Road after 0.6 km. 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.

Amethyst Mine tour

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

The tour will pass through the operating mining area, which is not available to ordinary tourists. No collecting is allowed in this area. 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 the collecting area. Specimens are also for sale in the shop.

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Field trip 7 - Building stone tour of downtown Port Arthur, Thunder Bay, Ontario

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Foreward

This walking tour will examine a number of buildings in the downtown core of the Port Arthur portion of Thunder Bay. The buildings of interest are constructed from a range of stones, the majority of which were produced in northwestern Ontario. The walking tour will consider the geological and architectural features of each building.

Introduction (taken from Hinz et al., 1994)

The dimension and monument stone industry in northwestern Ontario has a long history and is linked to the development and prosperity of the region. One of the earliest commercial operations was located on Vert Island in Nipigon Bay of Lake Superior. The Mesoproterozoic Sibley Group yielded an attractive red sandstone which was extracted by the Chicago-Verte Island Sandstone Company. The stone was shipped to Chicago, Winnipeg, southern Ontario and other U.S. cities for construction uses. Development of some of the earliest quarries in the Marathon and Nipigon areas was directly related to the construction of the Canadian Pacific Railway in the late 1880's. Syenites surrounding Marathon and sandstones south of Nipigon were used in the construction of railway trestles to span the Black, Pic, Little Pic, Steel and Nipigon rivers. Today these trestles show very little wear and are a testament to the long-standing durability of the stones.

Although markets for dimension stone decreased in the early 1900's, production continued at the Simpson Island sandstone quarry (1900-1910) and at the Bannerman and Horne quarry (1912-15) near Ignace.

The next period of quarry development took place during the late 1920's to early 1930's. Five small scale quarries operated northwest of Marathon along the Canadian Pacific Railway. Black and brown granites were extracted and shipped to customers in Toronto, Buffalo, Chicago and Detroit. In 1932, the last of these quarries closed due to the loss of a market.

In 1948, the Vermilion Pink Granite Company opened a quarry approximately 12 km southwest of the town of Vermilion Bay. This highly popular pink granite began production in 1954 and continued sporadically under various names until 1991 when the quarry, now named Granite Quarriers (GQI) Inc., closed. In 1981, Nelson Granite Limited of Sussex, New Brunswick began production of an identical granite from a quarry immediately south of the highway from the Granite Quarriers Inc. site. This quarry has operated year round since that time and is still in production.

Currently, 2012, Nelson Granite Limited is the only stone producer operating in northwestern Ontario. Nelson Granite produces a range of colours including pink, yellow, green, brown and white granite from four quarries located north of Kenora and west of Vermilion Bay. Northwestern Ontario stone is shipped around the world for a range of uses including: building stone for interior and exterior uses; monumental stone; and landscape uses including pavers, benches and accent pieces.

Detailed descriptions of the historic quarries, their operations, geology and geotechnical test results are provided in Hinz et al. (1994). Descriptions of current producers are available in the Kenora portion of the Report of Activities 2010, (Lichtblau et al., 2011). A list of building stone quarries in northwestern Ontario is given in Table 1.

Geologic setting

Sandstones of the Sibley Group

The red and buff coloured sandstones utilized as building stone throughout Thunder Bay were sourced from quarries in the Nipigon to Rossport area. The sandstones represent lithologies hosted within the Pass Lake Formation of the Sibley Group, the following description is taken from Fralick et al. (2000).

“Sedimentary rocks of the Sibley Group discontinuously outcrop on the north shore of

Table 1. Building stone quarries of northwestern Ontario. The asterisk (*) denotes stones which will be viewed at the various field trip stops.

Sibley Group Sediments

Quarry	Rock Type	Years of Operation
Cooke Point	Marble	1931-1940 (?)
George Point	Grey sandstone	Late 1800's (?)
La Grange Island	Red sandstone	1882-1883 (?)
Nipigon River *	Variegated marble	1883-1910 (?)
Quarry Island	White sandstone	Late 1880's (?)
Ruby Lake	Variegated marble	1996-1998
Simpson Island *	Buff sandstone	1904-1912
Vert Island *	Red sandstone	1881-1912
Wolf River	Buff sandstone	1913-15; 1921-31

Granitic Rocks

Quarry	Rock Type	Years of Operation
Bulter Grey *	Grey granite	1892-1943, 1946-52, 1989
Cold Spring	Black granite	1931-38(?)
C.P.R.	Black granite	1880's (?)
Docker Township *	Pink granite	1981-present
Forgotten Lake	Yellow granite	1997-present
GQI Quarriers Inc.	Pink granite	1948-1991
Peninsula Granite	Brown granite	1880's-1927(?)
Pine Green	Green granite	1992-present
Red Deer Lake	Brown granite	1996-present
Redditt	White granite	2010-present

Note: The usage of the term “granite” in the building stone industry is used regardless of lithology (eg. granite, syenite, gabbro)

Lake Superior and around Lake Nipigon (Fig. 1). The flat-lying to gently dipping, clastic-carbonate succession occupies a broad oval area disconformably to unconformably overlying Mesoproterozoic, Paleoproterozoic and Neoarchean rocks. Sibley Group rocks are located in the Southern Province of northwestern Ontario.

The fluvial and lacustrine strata of the Sibley Group are divided into three subhorizontal formations: 1) the lowermost is the Pass Lake Formation; 2) the Rossport Formation overlies it, followed by 3) the Kama Hill Formation.

The Pass Lake Formation varies from 0 to 50 m

in thickness and consists of basal conglomerate and upward-thinning beds of quartz arenite. It was deposited in a shallow lacustrine environment (Franklin et al., 1980).”

Marble of the Sibley Group (taken from Fralick et al., 2000)

“This marble consists of contact metamorphosed, Mesoproterozoic, Rossport Formation (Sibley Group) dolostone and other, calcareous sedimentary rocks in the contact metamorphic aureole of Keweenaw diabase sills. It has previously been termed Nipigon River marble and was quarried from 1883 to ca. 1910

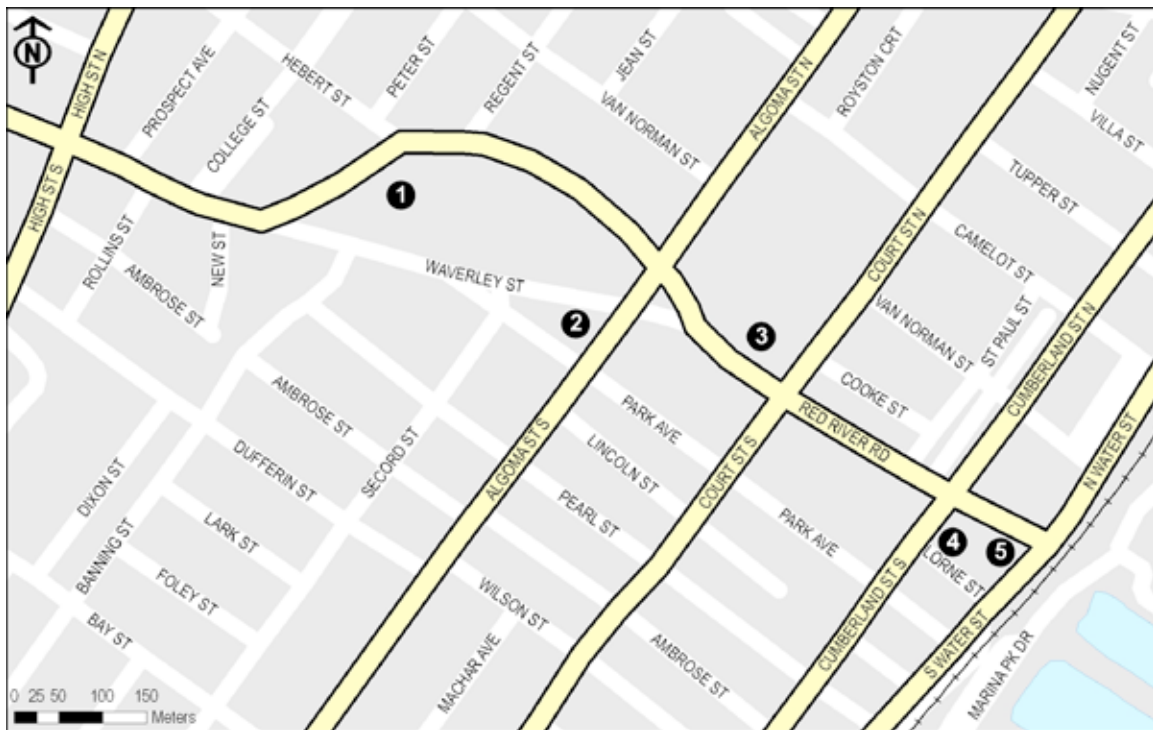


Figure 1. Field trip stop locations: 1. Port Arthur Collegiate Institute; 2. Trinity United Church; 3. Masonic Temple; 4. Former Hymer's Mens Wear; 5. Ontario Government Building.

at a site on the eastern side of the Nipigon River, approximately 6 west of the Ruby Lake quarry (Hinz et al., 1994). A similar hornfelsed unit is described in more detail at Stop 2-6.

Calcite, dolomite, epidote and opaque minerals were noted in thin section by Hinz et al. (1994) from the Nipigon River quarry."

Granites of the Wabigoon Subprovince (taken from Beakhouse et al., 1995)

"The Wabigoon subprovince is a 150 kilometre wide volcanoplutonic domain that has an exposed strike extend of 700 kilometres, extending an unknown distance beneath Paleozoic strata at either end.

The western Wabigoon region is characterized by interconnected, arcuate, metavolcanic 'greenstone belts' surrounding large elliptical batholiths. Granitoid rocks within the western Wabigoon region include large elliptical to multi-lobate batholiths that define the architecture of the greenstone belts as well as smaller stocks. Most of the large batholiths (Aulneau, Atikwa, Sabaskong) range compositionally from ultramafic to granitic but are predominantly tonalitic to granodioritic."

Both the Butler Grey and Vermilion Pink granites are sourced from intrusions within the Wabigoon subprovince.

Storey (1986) described the geology of the Butler Grey quarry: "The rock is massive, light grey to white, biotite granite (approximately 5% biotite). There are local variations in grain size and resultant colour variations. There are a few minor patch pegmatites. A very weak foliation trends north-northwest."

Storey (1986) also described the Vermilion Pink granite: "The rock was classified as quartz monzonite by Mattison (1952) and granite by Pryslak (1976). A modal analysis from Mattison (1952) plots as granite in the Streckeisen (1976) classification."

Field trip stops

Stops are located by UTM co-ordinates based on NAD 83, UTM Zone 16

The walking tour starts on the east side of the Port Arthur Collegiate Institute building. Park on Waverley Street on the south side of Waverley Park (0335224E 5367273N), walk to the east entrance of the Port Arthur Collegiate Institute.

Stop 1. Port Arthur Collegiate Institute (aka. P.A.C.I.; 401 Red River Road)

UTM coordinates: NAD83; 16U 0335208E / 5367358N

This building, constructed in 1909 of Simpson Island buff sandstone (Pass Lake Formation), is an example of the Queen Anne style that was in common use from the 1880s to the 1910s. When the school board made the initial planning for the building, it was decided that it should be “erected for posterity, and not be of the ‘shack’ order”, so they chose the stately Queen Anne style. The original P.A.C.I. can be seen in Figure 2. Alterations in 1925 resulted in the addition of four more classrooms, and more renovations to the north and south in 1953 and 1962 created other rooms. These alterations used Indiana Limestone in an attempt to blend into the original building. The younger, middle Mississippian (335-340 Ma) aged limestone is easily distinguished from the much older Mesoproterozoic Sibley sandstones by the prolific presence of marine fossils such as crinoids, bryozoa and gastropods (Fig. 3). A gymnasium planned in 1964 and completed in 1974, provoked controversy as its design was incompatible with the rest of the school.

Walk back to Waverley Street to view several

residential houses which incorporate Simpson and Vert island sandstones (eg. 369, 349 and 332 Waverley Street) walking east approximately 190 metres along Waverley Street to the Algoma Street corner.

Stop 2. Trinity United Church (30 Algoma Street South)

UTM coordinates: NAD83; 16U 0335432E / 5367226N

This building, completed in 1906, was formerly known as the Trinity Methodist Church, and became the Trinity United Church after the United Church of Canada was formed in 1925. Constructed of rough cut, Simpson Island buff sandstone (Pass Lake Formation), this structure is an example of the Late Gothic Revival style that was popular from the 1890s to the 1940s (Fig. 4). The unusual tower features very narrow windows (lancets), four buttresses, each capped with a pyramid shaped finial, and an extremely sharp hexagonal spire. The rest of the building also features very steeply pitched roofs, and arched windows in the Gothic style.

Walk north and cross Red River Road, approximately 90 metres, then cross Algoma Street and walk east approximately 125 metres.



Figure 2. The Port Arthur Collegiate Institute ca. 1909, from <http://images.ourontario.ca/>



Figure 3. Left - Fossiliferous Indiana limestone, arrow points to a fan-like bryozoan. Right - Enlarged view of the fan-like bryozoan.



Figure 4. Trinity United Church ca. 1930, from <http://www.hpd.mcl.gov.on.ca/>



Figure 5. Masonic Temple, Red River Road, Nipigon River marble displaying mat-like stromatolites.

Stop 3. Masonic Hall (262-270 Red River Road)

UTM coordinates: NAD83; 16U 0335596E / 5367166N

Built in 1910 and also known as the Shuniah Lodge, this stone, brick and concrete building replaced the old Masonic temple that was destroyed by fire in 1909. The first floor is made of cut Nipigon River marble (Rossport Formation) and the entrance features carved marble pilasters and decorative panels (Fig. 5). Originally there was a dome on the roof over the entrance, which has since been removed. The central portion of the building has a Mansard roof of French design. The building's windows are decorated with alternating round and triangular pediments above them. Commercial space occupies the ground floor, while the lodge is located above.

Continue east along Red River Road, crossing Court, St. Paul and Cumberland streets, approximately 300 metres. Cross Red River Road and proceed south for 60 metres to Lorne Street where you will see a red sandstone wall.

Stop 4. Former Hymer's Men's Wear (17 Cumberland Street South, Lorne Street wall)

UTM coordinates: NAD83; 16U 0335826E / 5366930N

The north wall of this building is constructed of Vert Island red sandstone, the building was constructed circa. 1900. The stone is a brick red sandstone which is part of the Mesoproterozoic Sibley Group, Pass Lake Formation (Fig. 6a). Syneresis cracks are evident on one block. These cracks are caused by subaqueous



Figure 6a (left). Red sandstone wall on the north side of the former Hymer's Men's Wear. 6b (right) Reductions band in Pass Lake formation sandstone.

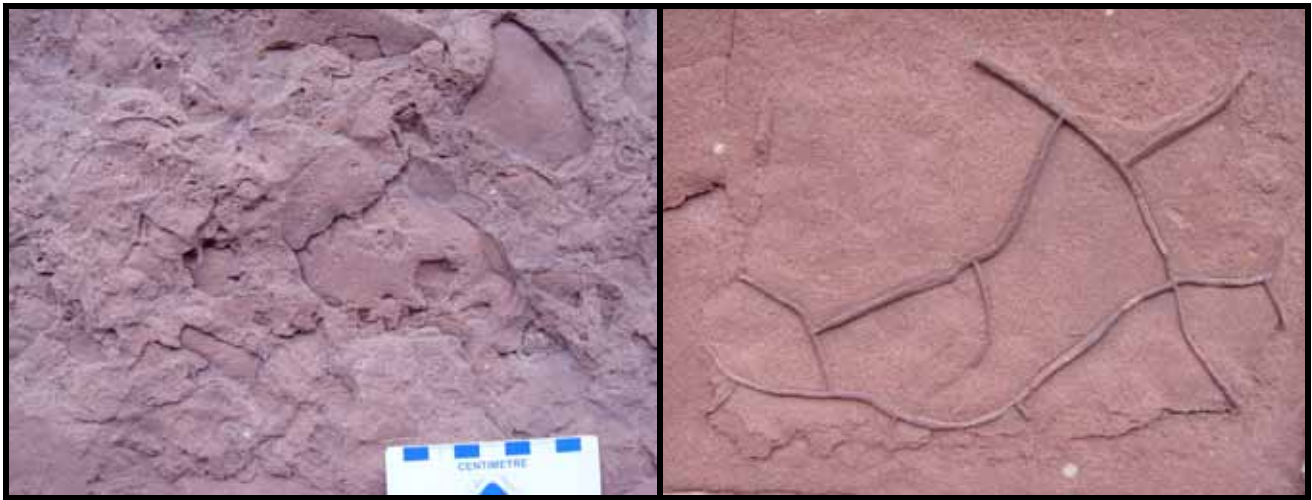


Figure 6c (left). Intraformational conglomerate, Pass Lake formation. **6d** (right). Syneresis cracks in Pass Lake formation sandstone.

shrinkage of sediments without dessication (Fig. 6d). Also visible in many blocks are intraformational conglomerate, buff coloured reduction spots and bands, ripples (Fig. 6b & c).

Walk 35 metres east along Lorne Street to the Ontario Government Building.

Stop 5. Ontario Government Building (189 Red River Road)

UTM coordinates: NAD83; 16U 0335847E / 5366925N

Opened in 1990, the Ontario Government Building was built to incorporate granite building stone from quarries in the Ignace and Vermilion Bay of northwestern Ontario, the stones included Butler Grey and Vermilion Pink (Fig. 7). The stones were used in the exterior and interior of the building and include polished, honed and flame finishes. The Ontario Government Building is described in the Ontario Architecture website (<http://www.ontarioarchitecture.com/postmodern.htm>):

“It is a classic Post Modern building in that it uses traditional architectural vocabulary in a new and impressive way. The front colonnade is a good example. The columns have neither bases nor capitals, but a decorative band level with the first floor lintels. There is an exaggerated cornice atop the architrave which has three horizontal bands. There is no ornament, not even fluting on the columns, and instead of marble, the columns are metal. Behind the colonnade, the building is a curtain wall of glass with an open concept foyer. Winding around the colonnade is a balustrade leading to other portions of the building and a landscaped front.”

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Figure 7. Entrance of the Ontario Government Building with the flame-finished Butler Grey granite on the exterior and features of the Post Modern architecture.

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Field trip 8 - A geologic transect across the Western Superior Province and Nipigon Embayment, Thunder Bay to Armstrong, Ontario

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Introduction

This field trip along Highway 527 between Thunder Bay and Armstrong describes a 250 km long transect through Archean rocks of the western Superior Province of the Canadian Shield and the Mesoproterozoic Nipigon Embayment. This transect extends from the Wawa Subprovince near Thunder Bay, across the entire width of the Quetico Subprovince and into the Wabigoon Subprovince, south of Armstrong (Fig. 1). The Wabigoon and Wawa (ca. 3.0 to 2.7 Ga) are volcano-plutonic subprovinces, containing a number of greenstone belts. The intervening Quetico Subprovince (ca. 2.7 Ga) consists of metamorphosed clastic sedimentary rocks, their high-grade metamorphic equivalents and derived granitic rocks.

These subprovinces are intruded or unconformably overlain in this area by a variety of Mesoproterozoic rocks of the Nipigon Embayment (Southern Province). These Mesoproterozoic rocks include Sibley Group (ca. 1.3 Ga) sedimentary rocks, the Badwater intrusive complex (ca. 1.6 Ga), and the English Bay felsic intrusive-volcanic complex (ca. 1.54 Ga). Voluminous mafic to ultramafic intrusive rocks related to the Mesoproterozoic Midcontinent Rift (ca. 1.1 Ga) predominate.

Day One of the trip will highlight representative Archean lithologies while Day Two will focus on

Mesoproterozoic rocks. 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. Please exercise caution when stopping and viewing roadside exposures.

Stott (2009) noted that the terminology of crustal subdivisions across the Archean Superior Province is slowly evolving such that regional lithologic subdivisions as subprovinces (Card and Ciesielski, 1986) are currently being reassessed in terms of terranes and adjacent, typically autochthonous domains (e.g., Percival and Helmstaedt, 2006; Stott et al., 2007). Subdivisions in the field trip area are shown in Table 1.

The most recent geological synopsis in the transect area was provided by Hart and MacDonald (2007):

The Nipigon Embayment is underlain, from north to south, by Archean rocks of the English River, Wabigoon, and Quetico subprovinces (Fig. 2). Much of the Embayment is underlain by a series of east-trending 2950 to 2700 Ma greenstone belts separated by 3000 to 2690 Ma intrusive rocks of the central and eastern Wabigoon subprovince (e.g., Blackburn et al., 1991). Tomlinson et al. (2004) proposed a north-south subdivision of the Wabigoon subprovince into the Winnipeg River and Marmion terranes based on isotopic data. The boundary between

Table 1. Geological subdivisions in the field trip area

Existing Nomenclature	Proposed Nomenclature
Wabigoon Subprovince	Winnipeg River Terrane (north)
	Marmion Terrane (south)
Quetico Subprovince	Quetico Basin
Wawa Subprovince	Wawa-Abitibi Terrane

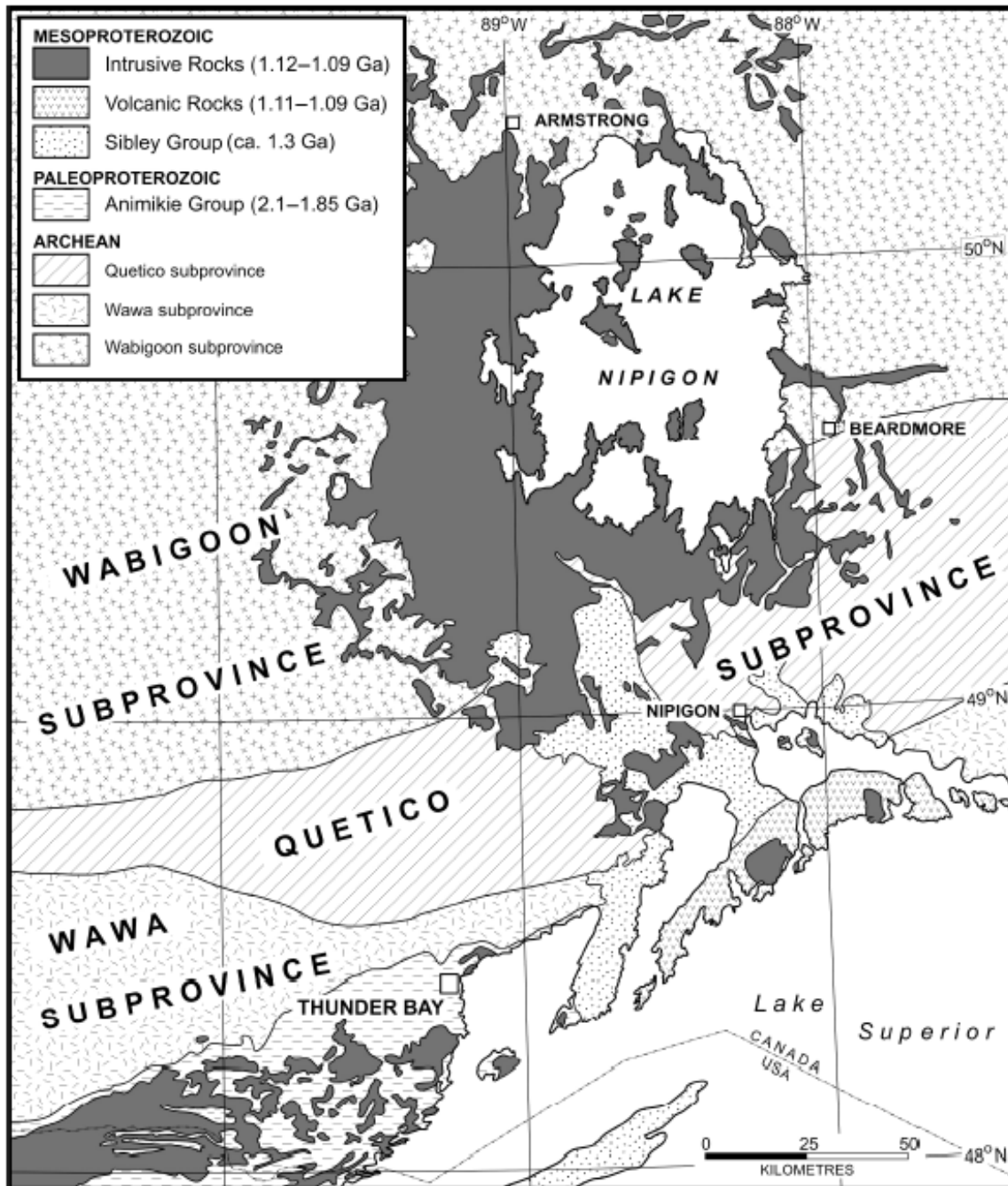


Figure 1. General geology of the field trip transect area, from Smyk and Franklin (2007)

the terranes is obscured by later granites west of Lake Nipigon, but is correlated with the Humboldt Bay high-strain zone to the east of the lake. The southern margin of the Wabigoon subprovince consists of a series of metasedimentary and metavolcanic belts separated from the granite–greenstone terrane by the Paint Lake Fault east of Lake Nipigon and the Max Creek Fault west of the lake (e.g., Williams and Stott, 1991; MacDonald et al., 2005). The boundary between

the Wabigoon and Quetico subprovinces is the Blackwater Fault to the east, south of Beardmore, and is located south of Peevy Lake to the southwest. The metasedimentary rocks of the Quetico subprovince are intruded by “I-type” tonalite to granodiorite and “S-type” muscovite and two-mica leucogranites (e.g., Williams, 1991; Breaks et al., 2003), with minor mafic to ultramafic and syenitic bodies (e.g., MacTavish, 1999; Pettigrew and Hattori, 2006). A minimum

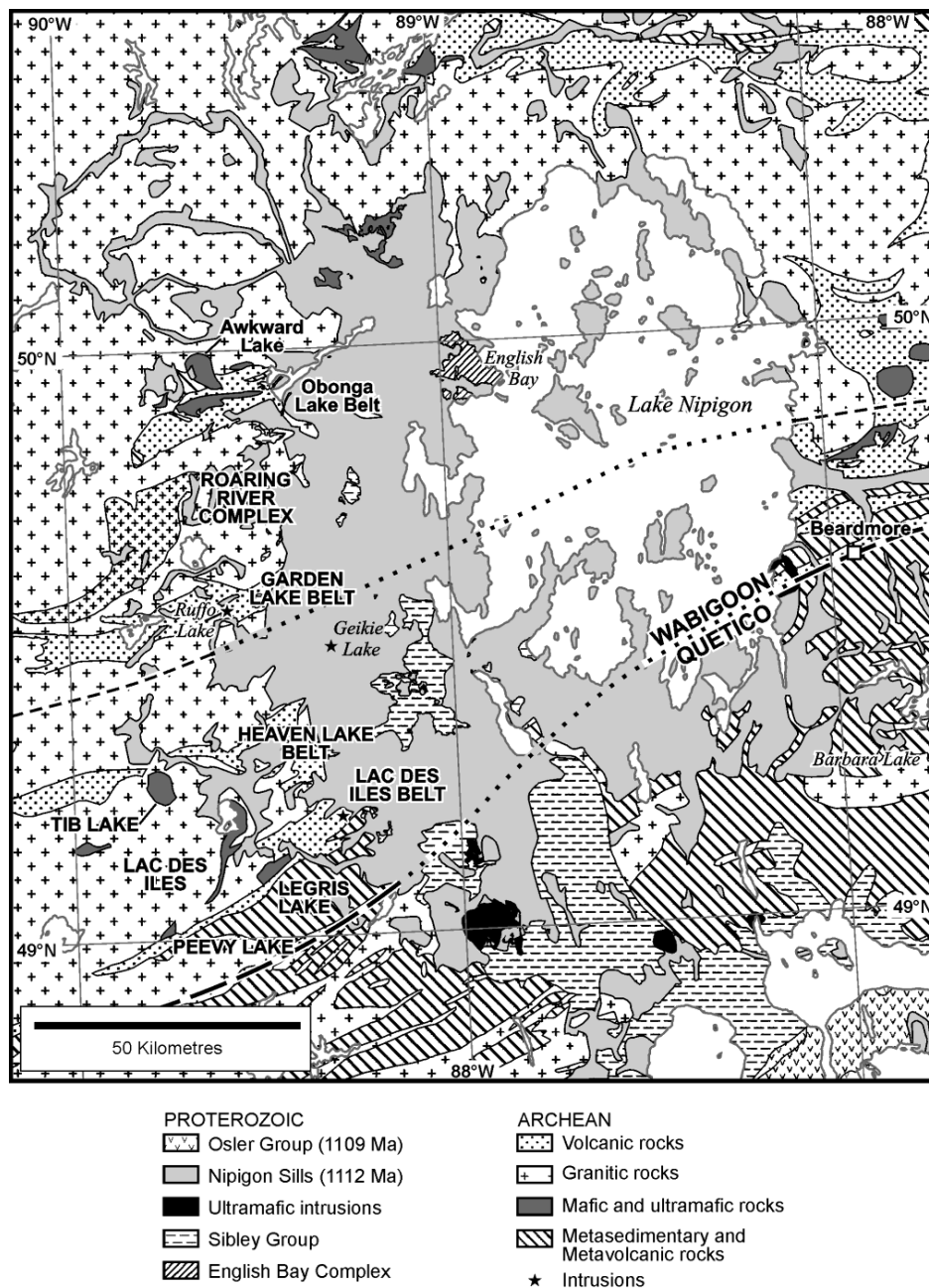


Figure 2. Hart and MacDonald's (2007) generalized geology of the Nipigon Embayment (modified from Ontario Geological Survey 1993; MacDonald 2004; Hart and Magyarosi 2004; MacDonald et al. 2005; Hart 2005) and the Archean rocks of the Wabigoon and Quetico subprovinces surrounding and underlying the Embayment.

age of deposition for the metasedimentary rocks is constrained by ages of 2665 ± 2 and $2653 \pm 3/-4$ Ma on leucogranites immediately southwest of the Nipigon Embayment (Percival and Sullivan, 1988), as well as the $2688 \pm 6/-5$ Ma Samuels Lake bodies (Pettigrew and Hattori, 2006), and the 2689 ± 2 Ma Black Pic monzodiorite (Zaleski et al., 1999). There are a number of late- to post-tectonic mafic to ultramafic intrusions in the central Wabigoon subprovince, including the

intrusions of the Lac des Iles area (e.g., Sutcliffe, 1987; Tomlinson et al., 2002; Stone et al., 2003). The various intrusions of the Lac des Iles area... have been collectively referred to as the Lac des Iles suite by Stone et al. (2003) and include the Northern Ultramafic and Mine Block intrusions of the Lac des Iles Complex (e.g., Hinchey et al., 2005). Previous workers have suggested that the intrusions of the Lac des Iles suite may be part of a contemporaneous magmatic event (e.g.,

Stone et al., 2003; Pettigrew and Hattori, 2006). A number of intrusions identified as a result of mapping (e.g., Hart, 2000) and diamond drilling (e.g., MacDonald et al., 2005) may be part of the same magmatic event.

A series of north-striking diabase dykes intrudes the Archean rocks of the Wabigoon and Quetico subprovinces. A paleomagnetic and geochemical study of the dykes located west of the Nipigon Embayment suggests that there are 2130–2120 Ma reverse-magnetized and 2110–2100 Ma normal-magnetized Marathon dykes (Ernst et al., 2006). A solitary northeast-trending

diabase dyke was found that may correlate with the 1140 Ma Abitibi swarm (Ernst et al., 2006).

Three Meso- to Paleoproterozoic [sic] lithologic units are located in the northwest portion of the Nipigon Embayment: the Badwater (Creek) intrusion, the Pillar Lake Volcanic rocks, and the English Bay Complex (Fig. 3). The mafic to felsic Badwater intrusion is unconformably overlain by flat-lying mafic pillowed volcanic rocks of the Pillar Lake volcanic unit (MacDonald, 2004). The English Bay volcanic-intrusive complex is located to the southeast, on the northwest shore of Lake Nipigon (e.g., Sutcliffe and Greenwood,

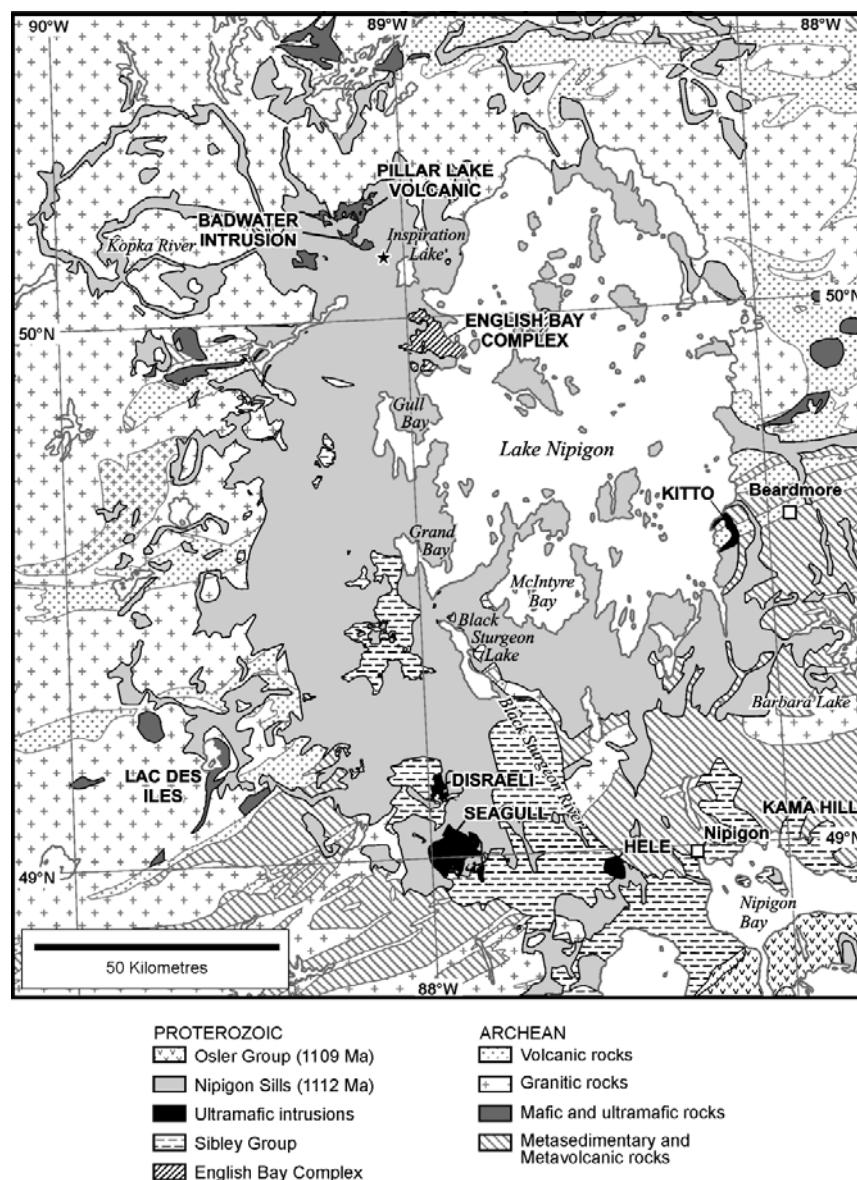


Figure 3. Hart and MacDonald's (2007) generalized geology of the Mesoproterozoic rocks of the Nipigon Embayment (modified from Ontario Geological Survey, 1993; MacDonald, 2004; Hart and Magyarosi, 2004; MacDonald et al., 2005; Hart 2005)

1985a). All three units appear to be localized along major regional structures.

Unconformably overlying the basement Archean and earlier Proterozoic rocks are the clastic and chemical sedimentary rocks of the Sibley Group. The Sibley Group is interpreted to have been deposited in a fluvial to shallow-lacustrine environment with transitions to playa lake and sabkha environments followed by reflooding of the basin, recorded in the middle to upper stratigraphic parts of the sequence (e.g., Franklin et al., 1980; Cheadle, 1986; Rogala, 2003). The thickest accumulation of Sibley Group rocks in the western portion of the Nipigon Embayment is within a half-graben, defined by the faults in the area of the Black Sturgeon River, thinning toward the west (e.g., Coates, 1972). Rogala et al. (2007) interpret the initial period of fault activity as representing a change from broad subsidence to active basin formation in the Lake Nipigon area prior to 1339 ± 33 Ma, and prior to formation of the Midcontinent Rift.

There are four sill-like mafic to ultramafic intrusions, with three (Disraeli, Seagull, and Hele) located to the south of Lake Nipigon and one (Kitto) located along the east side of the lake, emplaced into the Sibley Group and underlying Archean rocks (e.g., Sutcliffe, 1986, 1987; Hart and Magyarosi, 2004). Ultramafic-hosted PGE mineralization in the Seagull intrusion is located within discrete, laterally continuous zones, which a study by Heggie (2005), using whole-rock geochemistry, isotope geochemistry, and mineral chemistry, suggests have been caused by sulphur saturation of the magma during initial stages of emplacement, with zones higher in the intrusion probably reflecting influxes of less-evolved magma. Other examples of fine-grained, massive mafic to ultramafic sills occur scattered through the Nipigon Embayment. The thickest are the Jackfish sill in the northwest corner of Lake Nipigon and the Shillabeer sill south of the lake (e.g., MacDonald, 2004; Hollings et al., 2007a). The current outline of the Nipigon Embayment is defined by a series of diabase sills estimated to cover an area in excess of 20 000 km² (Sutcliffe, 1991). The shallow-dipping Nipigon diabase sills, ranging in thickness from <5 m to >180 m, intrude the mafic to ultramafic intrusions, Sibley Group, Pillar Volcanics, the English Bay Complex, and the underlying

Archean rocks (e.g., MacDonald, 2004; Hart, 2005). A lack of obvious textural, mineralogical, or geochemical variations within sill exposures hinders regional correlation of the sills and the development of a stratigraphic succession (e.g., Hollings et al., 2007). However, a recent airborne magnetic survey (Ontario Geological Survey, 2004), combined with new geochemistry and geochronology, has distinguished a number of distinctive sills with limited extents..., including the Inspiration sill (MacDonald 2004; Hollings et al., 2007a; Heaman et al., 2007) and the McIntyre sill (Richardson et al., 2005; Hollings et al., 2007; Heaman et al., 2007). Geological mapping suggests that the formation of the Nipigon Embayment was controlled by a series of north-, northwest- and northeast-trending faults that appear to correlate with prominent Archean basement structures (e.g., Hart, 2005; MacDonald et al., 2005). Interaction between these faults formed an asymmetric basin or half-graben in the southwest portion of the Nipigon Embayment as originally defined by Coates (1972). Franklin et al. (1980) suggested that the faults defining parts of the Nipigon Embayment represented a failed arm of the Midcontinent Rift, and Sutcliffe (1987) and Lightfoot et al. (1991) proposed that the ultramafic intrusions and Nipigon diabase sills intruded along these faults. Alternatively, the faults may be the result of subsidence following an anorogenic thermal event as proposed by Fralick and Kissin (1995) and Hollings et al. (2004). Further work by Rogala et al. (2007) indicates that most fault activity related to half-graben development was the result of broad subsidence that occurred more than 200 million years before the Midcontinent Rift, thus lending further credence to the interpretations of Fralick and Kissin (1995) and Hollings et al. (2004).

A synopsis of mineral deposits in the area was given by Smyk and Franklin (2007):

A variety of metallic and non-metallic mineral deposit types occur within Archean and Proterozoic rocks in the area encompassing the Lake Nipigon Region Geoscience Initiative. Archean deposit types include: Algoma-type banded iron formation-hosted iron (e.g., Lake Nipigon iron range) ; volcanogenic massive sulphide copper-zinc (e.g., Onaman-Tashota belt); ultramafic intrusion-hosted chromium (e.g., Puddy-Chrome lakes); mafic to ultramafic

Table 2. Mapping of the Precambrian bedrock geology of the transect area by the Ontario Geological Survey

Map Number	Map Area	Scale	Year	Authors
P2984	MacGregor Township, west half	1:15 840	1986	J.F. Scott, J.M. Seguin
ARM48C	Gorham Township and vicinity	1:63 360	1939	R.D. Macdonald
M2235	Disraeli Lake sheet	1:63 360	1971	M.E. Coates
M2172	Eayrs Lake-Starnes Lake area	1:63 360	1969	L. Kaye
M2136	Max Lake sheet	1:31 680	1967	E.G. Pye
P3580	Southwest Portion of the Nipigon Embayment	1:100 000	2006	Hart, T.R.
P3435	Lac des Iles Greenstone Belt	1:20 000	2001	Hart, T.R., MacDonald, C.A., Lepine, C.D.
P3434	Heaven Lake Greenstone Belt	1:20 000	2001	Hart, T.R., MacDonald, C.A., Lepine, C.D.
P3560	Cheeseman-Black Sturgeon Lakes Area	1:50 000	2005	MacDonald, C.A., Tremblay, E., ter Meer, M.
P3559	Kabitotikwia Lake Area	1:50 000	2005	MacDonald, C.A., Tremblay, E., ter Meer, M.
P3537	English Bay-Havoc Lake Area	1:50 000	2004	MacDonald, C.A., ter Meer, M., Lepage, L., Préfontaine, S., Tremblay, E.
P3536	Waweig-Wabinosh Lakes Area	1:50 000	2004	MacDonald, C.A., ter Meer, M., Lepage, L., Préfontaine, S., Tremblay, E.
P0962	Pashkokogan-Caribou lakes sheet	1:126 720	1974	Sage, R.P., Breaks, F.W., Stott, G., McWilliams, G., Bowen, R.P.

intrusion-hosted copper-nickel-platinum group element (PGE) (e.g. Lac des Iles); and pegmatite-hosted deposits of rare metals (Li, Ta, Be), uranium and molybdenum (e.g., Georgia Lake field; Black Sturgeon Lake; Anderson Lake, respectively). Mesothermal lode gold deposits are prominent in the Beardmore-Geraldton camp.

Superior-type iron formation occurs in Paleoproterozoic Gunflint Formation. "Red-bed" copper occurs in Mesoproterozoic Midcontinent

Rift-related Osler Group volcanic and interflow sedimentary rocks. Native copper and Cu-sulphides occur in Mesoproterozoic Sibley Group sedimentary rocks, adjacent to ultramafic intrusions. These mafic to ultramafic intrusions, associated with Midcontinent Rift magmatism, host copper-nickel-PGE deposits (e.g. Seagull, Great Lakes Nickel). Silver-bearing veins occur in Paleoproterozoic Animikie Group sedimentary rocks in proximity to Midcontinent Rift-related

mafic intrusions (e.g., Silver Islet; Silver Mountain). Lead-zinc-barite veins, uranium-bearing veins and amethyst vein- and replacement-type deposits may be co-genetic and formed at or near the unconformity between Sibley Group basal sandstone and underlying Archean granitic basement (e.g., Dorion; Black Sturgeon Lake; McTavish Township). The hydrothermal systems that produced all of these veins were probably driven by heat associated with Midcontinent rifting. Many occur in structures related to rift-bounding faults. Iron oxide-copper-gold deposits may occur near the English Bay intrusion.

Mapping of the Precambrian bedrock geology of the transect area by the Ontario Geological Survey (and its predecessors) ranges from detailed (e.g. 1:15 840) to reconnaissance-scale (1:250 000). Much of the detailed mapping focused on greenstone belts. Newer, comprehensive 1:50 000 mapping of the Nipigon

Embayment was undertaken as part of the Lake Nipigon Region Geoscience Initiative (Table 2).

Stop descriptions

Day One (Figs. 4 & 5)

Stop 1-1: Pillowed Metavolcanic Rocks, Wawa Subprovince

UTM coordinates: NAD83; 16U 0340852E / 5376037N

This is a typical exposure of greenschist-facies, massive to pillowed, locally vesicular mafic metavolcanic rocks, foliated at 260/80. The unit is cut by small shear zones, mafic xenolith-bearing feldspar porphyry dykes and north-northeast-trending calcite veins. Similar veins to the northeast contain lead and zinc sulphides and may be related to a Mesoproterozoic

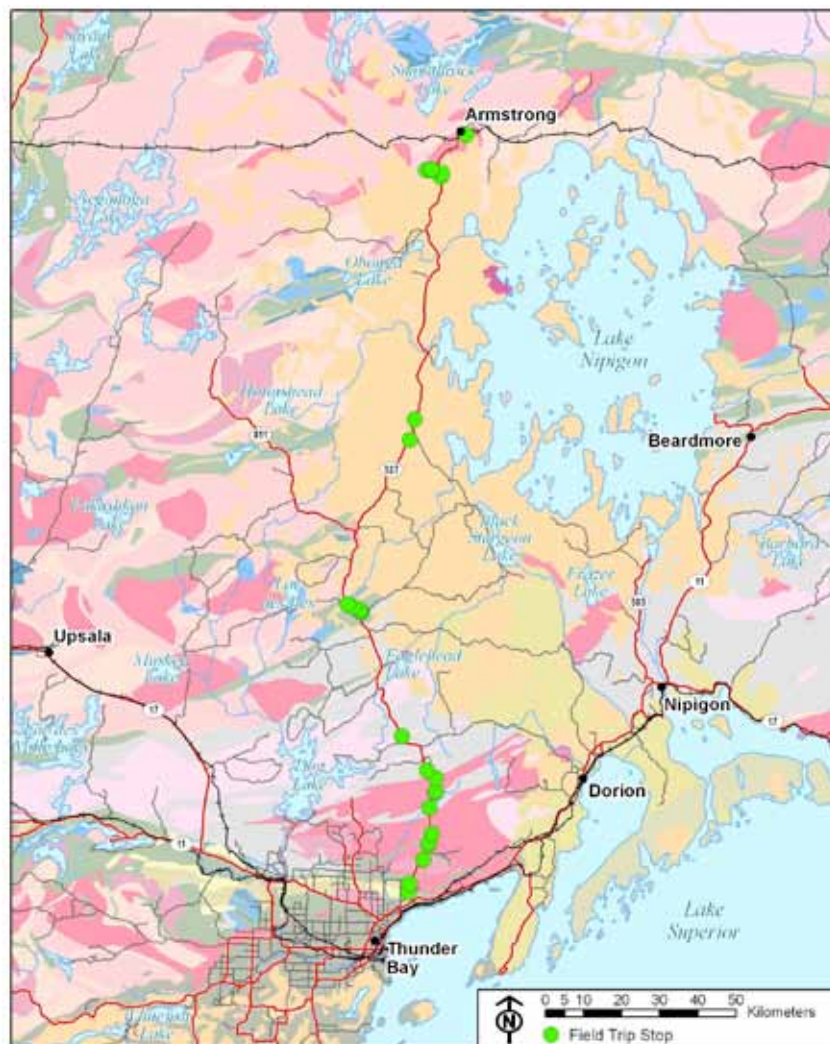


Figure 4. General geology of the transect area, showing the location of field trip stops along Highway 527

Road Log for Field Trip Stops

(N.B. Please exercise caution along highway and road right-of-ways.)

STOP NAME	STOP NUMBER	LANDMARK (0 Km)	DISTANCE (km)	NORTHING	EASTING
DAY ONE					
		Intersection of Hwy. 527 and Hwy. 11-17	0.0		
Wawa – Pillowed Metavolcanic Rocks	1-1		2.8	5376037	340852
		Mt. Baldy Road	2.9		
Wawa - Timiskaming Metasedimentary Rocks	1-2		5.1	5378261	341252
		Weigh Scales Compressor Station Road	5.2 6.5		
Wawa - Penassen Lakes Stock	1-3		12.4	5384413	344754
		Gibson's Road Magone Road	13.2 15.0		
Quetico - Metasedimentary Rocks	1-4A	Beaverlodge Road (exit Highway 527 to access stops 1-4A, 1-4B) Beaverlodge Road, north spur, 1.3 km west off Hwy. 527	16.1	5389006	346350
Wawa - Feldspar-phyric granitoid	1-4B	Beaverlodge Road, north spur, 1.0 km west off Hwy. 527		5388319	346239
		Kingfisher Lake Road	19.0		
Wawa – White Lily Lake Stock	1-5		19.7	5391178	347109
		Escape Lake Road	22.4		

		Barnum Road	25.3		
		Bush Road	26.8		
<i>Quetico</i> Pegmatite	1-6	100 m west off Hwy. 527		5398003	346508
		South Current River	27.9		
		Monday Lake Road	29.6		
<i>Quetico Fault</i> in Metasedimentary Rocks	1-7A		31.1	5401948	348057
<i>Quetico Fault</i> in Granitoid Rocks	1-7B		31.5	5402246	348117
		Shallowest Road	35.2		
<i>Quetico -</i> Pegmatitic Granite	1-8	100 m east of Hwy. 527		5405692	348152
<i>Quetico -</i> Migmatites	1-9		38.4	5407714	345756
		Orr's Place Store	40.0		
		Dorion Cut-Off	45.2		
		Pace Lake Road	49.9		
<i>Quetico -</i> Cordierite- bearing granitoids	1-10		51.1	5416579	339226
		Mott Lake Road	56.6		
		DeCoursey Lake	64.1		
		Eaglehead Lake Road	72.5		
		Pipeline	74.5		
		Fensom Lake Road	80.8		
		Mawn Lake Road	82.2		
		Max Lake Road	84.2		
		Camp 45 Road	87.5		
<i>Wabigoon -</i> Conglomerate	1-11		88.1	5448888	328624
		Max Creek	88.3		
<i>Wabigoon -</i> Tuff-breccia	1-12		88.6	5449317	328280
<i>Wabigoon -</i> Pillowed Mafic Metavolcanic Rocks	1-13		88.8	5449467	328159

Wabigoon - Brecciated Metavolcanic Rocks	1-14	90.7	5450297	326447
Wabigoon – Banded Iron Formation	1-15	92.5	5451015	324893
	LaChapelle Creek	93.7		
	Lac Des Iles Road	93.8		
	Poshkokogan Lake Road	103.6		
	Whistle Lake Road	104.9		

DAY TWO				
	Intersection, CNR Tracks and Hwy. 527, West end of Armstrong	0.0		
	Intersection, CNR Tracks and King St., East end of Armstrong	3.0		
	turn-off to quarry	3.3		
Pillar Lake Volcanics - Alarie Quarry	2-1	5.0	5573375	356068

	Intersection, CNR Tracks and Hwy. 527, West end of Armstrong	0.0		
	MacKenzie Lake Inn	2.5		
	Clearwater Lake Road	6.2		
	Frontier Road	7.8		
	Mattice Lake Road	10.0		

	Intersection of Mattice Lake Road with Highway 527	0.0		
	Bridge	0.8		
	Badwater Creek turnoff	3.8		
	Bridge	4.5		
Pillar Lake Volcanics – T-Junction, Mattice Lake Road	2-2	5.4	5564137	345619

Badwater Gabbro – Badwater Creek Road	2-3A	Intersection of Mattice Lake Road with Badwater Creek Road	0.0		
			0.6	5564306	347047
Badwater Gabbro - East of Badwater Creek Road	2-3B	(outcrops 30 m and 60 m east of road)	0.8	5564142	347043

Pillar Lake Volcanics - Chimney Lake	2-4	Intersection of Mattice Lake Road with Highway 527	0.0		
		outcrops 300 m east of highway)	2.4	5563345	349738
Pillar Lake Volcanics – Hwy. 527	2-5		3.3	5562320	349460
		Northern end of Waweig Lake	5.5		

Nipigon Diabase Sill, Highway 527	2-6	Bridge on Hwy. 527 at Gull River	0.0		
			20.2	5499194	342613
Sibley Group Sedimentary Rocks and Nipigon Diabase Sills	2-7	Kabi River Bridge	24.2		
		(outcrops 160 m southeast of highway)	26.8	5493814	341205

mineralizing event.

Stop 1-2: Timiskaming Metasedimentary Rocks, Wawa Subprovince

UTM coordinates: NAD83; 16U 0341252E / 5378261N

These steeply dipping, clastic metasedimentary rocks are massive, thickly to thinly bedded, quartz-rich arkosic sandstones with abundant mudstone fragments. Thinly bedded turbidites display graded bedding with tops to the south, indicating that this sequence (at 270/85°) has been locally overturned.

The northern Wawa Subprovince contains five major packages of sedimentary rock associations. There are: 1) cherts and iron formations deposited as inter-flow sediments in basaltic successions, 2) resedimented andesitic volcanic eruptive material forming areas of

graded beds interlayered in calc-alkaline successions, 3) thick turbidite sequences similar to the Quetico turbidites deposited when the Quetico trench was full and sediment gravity flows gained access to the Wawa ocean floor to the south (Fralick et al., 2006), 4) strandline to deep shelf deposits formed at 2692 Ma (unpublished U-Pb zircon age on volcanic ash) during a period of shoshonitic volcanism related to initial collision between a Wawa island arc and the Quetico accretionary complex, and 5) 2686 Ma fluvial conglomerates and sandstones shed into pull-apart basins during transpression during the main orogeny. The succession we are examining is reasonably non-descript, but most closely resembles units present in the shelf succession of the 2692 Ma assemblage. At other locations, hummocky cross-stratification present closer to shore in this system attests to the operation of geostrophic flows draining storm surges away

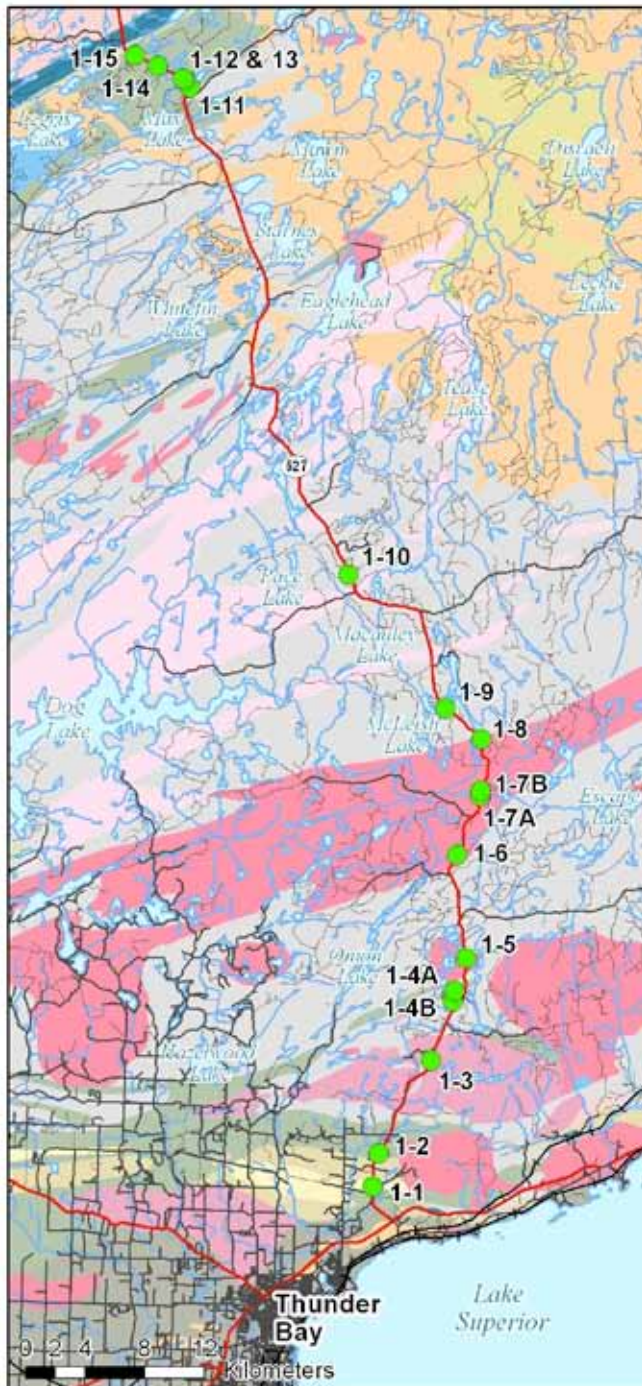


Figure 5. General geology and field trip stops, Day One (Archean)

from the coast. Below storm wave-base these flows deposited graded beds with the same appearance as slump induced turbidites. The lithologies present here are similar to these types of units, which are associated with hummocky cross-stratification and tidal flat deposits at other locations such as Finmark, west of Thunder Bay.

Stop 1-3: Penassen Lakes Stock, Wawa Subprovince

UTM coordinates: NAD83; 16U 0344754E / 5384413N

The Penassen Lakes stock is part of the Dog Lake granite chain, a linear series of at least six separate, possibly genetically related granitoid intrusions: Silver Falls, Trout Lake, Barnum Lake, Shabaqua, Penassen Lake and White Lily (Kuzmich et al., 2011; Fig. 7). They are emplaced in semi-pelitic to pelitic metasedimentary and gneissic rocks along the Quetico–Wawa subprovinces boundary north of Thunder Bay (Kehlenbeck, 1977). The intrusions form an approximately east-northeast-trending, evenly spaced chain spanning roughly 70 km, none of which have undergone geochronological analysis.

The magnetic signature of these intrusions suggests that they may be distinct from the typical S-type granites found within the Quetico Subprovince. Detailed geochemical and petrographic studies of the granites by Kuzmich (in progress) and Kuzmich et al. (2011) will provide additional insights into the origins and petrogenesis, as well as the development of the Quetico Subprovince as a whole.

The Penassen Lakes stock is a dark pink, massive, magnetic, medium-grained, quartz-monzodiorite to monzodiorite (Fig. 8). At this location, equigranular hornblende monzodiorite is cut by aplitic dykes and contains mafic xenoliths. Amethyst-bearing veins crosscut the stock at a locality approximately 600 m south along the highway.



Figure 6. Overturned turbidites, Stop 1-2

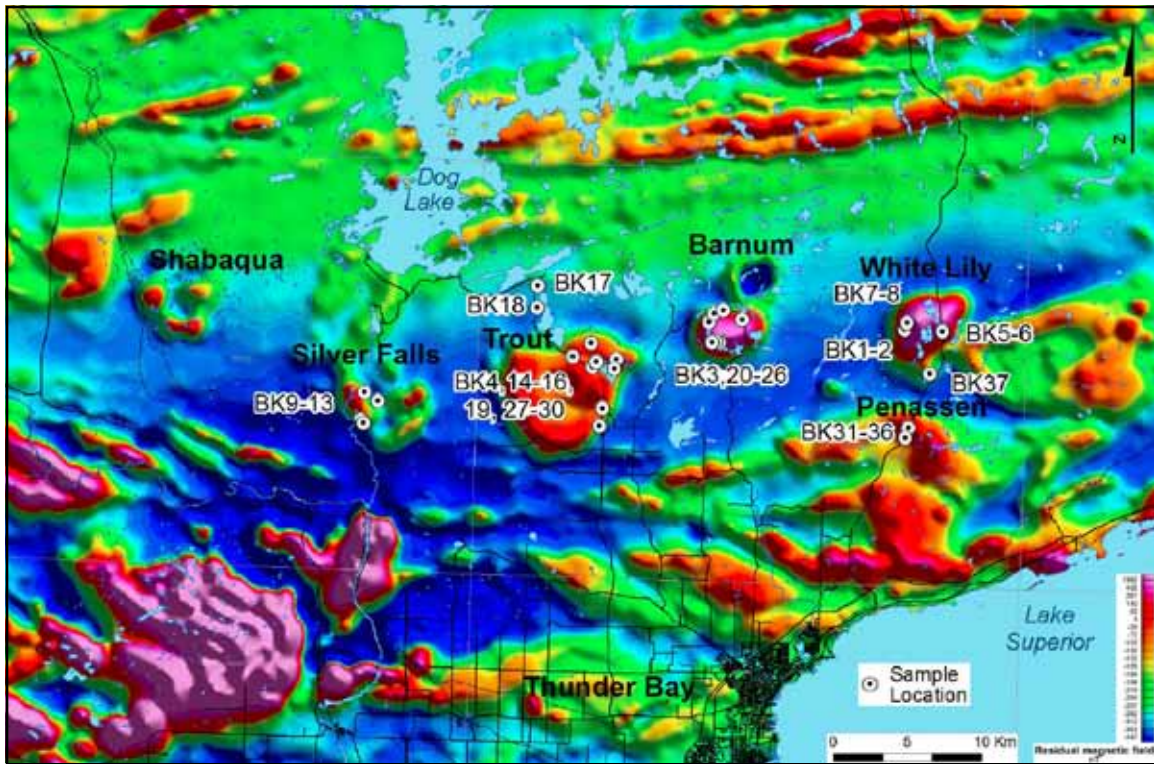


Figure 7. Aeromagnetic image of the area north of Thunder Bay, with Dog Lake intrusions labeled (Kuzmich et al., 2011)

Stop 1-4A: Metasedimentary Rocks, Quetico Subprovince

UTM coordinates: NAD83; 16U 0346350E / 5389006N

This site displays typical, thinly to thickly bedded Quetico metaturbidites (Fig. 9). Relict graded beds indicate southward younging and may have load casts at their bases. The fine-grained, pelitic tops appear to have porphyroblasts of what was tentatively identified as andalusite. Quartzo-feldspathic dykes host quartz veins in the necks of boudins. These dykes are locally cut by quartz-feldspar-muscovite pegmatite dykes, which may represent neosome generated by partial melting of the metasedimentary rocks. This corresponds to low- to medium-grade metamorphic assemblages outlined by Seemayer (1992) in this area along the southern margin of the Quetico Subprovince. Bear in mind that Wawa metavolcanic rocks have been noted north of this location, suggesting that the subprovincial boundary here may consist of intercalated panels of metavolcanic and metasedimentary rocks, intruded by late granitoids.

As described by Seemayer (1992), these low-grade metasedimentary rocks generally consist of quartz, plagioclase and biotite, with minor amounts of muscovite and chlorite. Porphyroblastic muscovite,

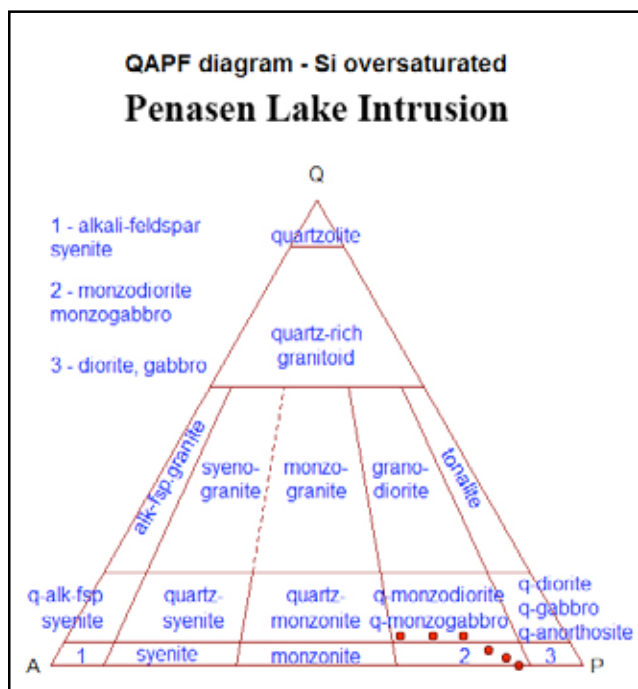


Figure 8. QAP diagram for samples of the Penassen Lakes stock (Kuzmich, in progress)



Figure 9. Graded metaturbidites and pegmatite dykes, Stop 1-4A

andalusite, garnet and cordierite are more common near contacts with granitoids and are attributed to contact metamorphism. Kehlenbeck (1977) noted hornfelsic textures in these contact zones.

Stop 1-4B: Feldspar-phyric granitoid, Wawa Subprovince

UTM coordinates: NAD83; 16U 0346239E / 5388319N

Just south of Stop 1-4A, there are exposures of an undeformed, medium-grained, feldspar-phyric granitoid with recessively weathered biotite and a moderate magnetic susceptibility. It may be a smaller, isolated intrusion related to one of the neighbouring (White Lily or Penassen Lakes) granitoid stocks.

STOP 1-5: White Lily Lake Stock, Wawa Subprovince

UTM coordinates: NAD83; 16U 0347109E / 5391178N

Kuzmich (in progress) has subdivided the White Lily intrusion into two separate phases: a fine- to

medium-grained, magnetic, dark pink, alkali feldspar-rich phase; and a medium- to coarse-grained, massive, magnetic, dark pink phase (Figs. 10 & 11). At this location, the intrusion consists of medium- to coarse-grained, K-feldspar-phyric, amphibole-bearing monzonite with a high magnetic susceptibility. Some slickensided surfaces were noted. The southern contact of the intrusion with metasedimentary country rocks is exposed approximately 700 m south along the highway in the vicinity of the Kingfisher Lake Road turn-off.

STOP 1-6: Pegmatite, Quetico Subprovince

UTM coordinates: NAD83; 16U 0346508E / 5398003N

Strongly peraluminous, muscovite-, cordierite- and garnet-bearing pegmatitic granite dikes, with local black tourmaline, were found to occur widely in the Quetico Subprovince along the Highway 527 from Walkinshaw Lake north to DeCoursey Lake (Breaks et al., 2003). Approximately 150 m west of Highway 527 and 40 m south of an old logging road, quartz-feldspar-muscovite pegmatite is exposed in a large whale-back outcrop. This locality, along with other pegmatites and related granitoid rocks, were described by Breaks et al. (2003):

Rare-element mineralization was discovered by the current survey within an extensive swarm of pegmatitic granite dikes at Onion Lake near Thunder Bay. The lens-shaped dikes of this swarm, as seen in the area near the junction of Highway 527 and the Barnett Lake road, occur as northeast-striking, whale-back glacial erosional

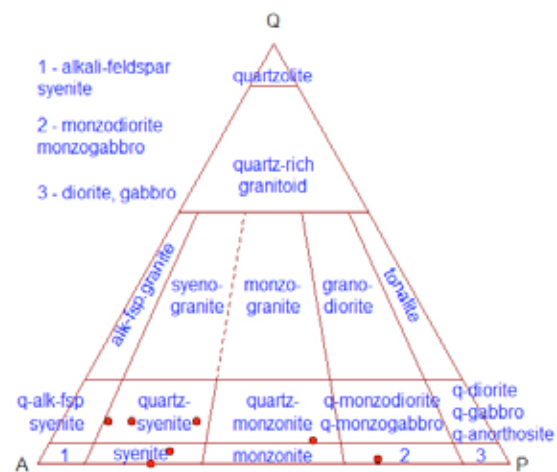


Figure 10. QAP diagram for samples of the White Lily Intrusion (Kuzmich, in progress)

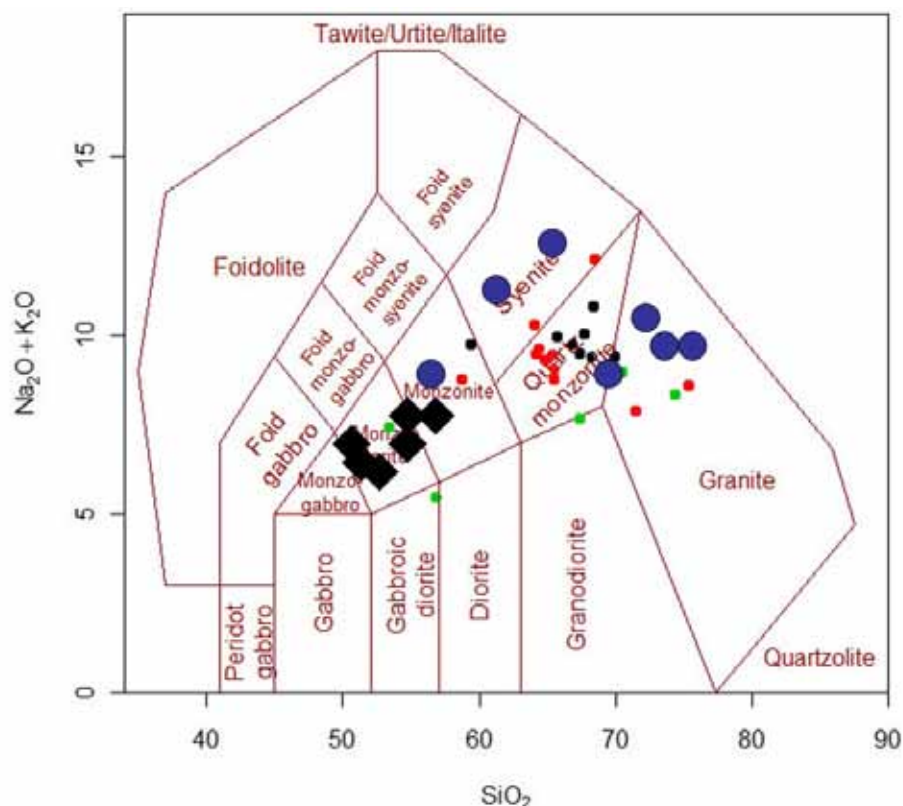


Figure 11. TAS diagram showing samples from Penassen Lakes (diamonds) and White Lily stocks (circles), from Kuzmich (in progress).

remnants that achieve a maximum size of 100 by 300 m. The internal units comprise:

- muscovite-rich potassic pegmatite
- quartz-rich patches with blocky K-feldspar, coarse muscovite books and sparse beryl
- fine-to medium-grained, garnet-biotite-muscovite granite
- garnet-biotite-muscovite pegmatitic leucogranite
- garnet and muscovite-garnet aplite

The quartz-rich patches locally contain pale green beryl up to 1 by 16 cm, as at locality 01-FWB-107 at Onion Lake (UTM 346199E, 5397916N, Zone 16). Black, tantalum-oxide minerals (ferrocolumbite: 27.31 weight % Ta_2O_5), up to 3 by 3 by 5 mm, were discovered at locality 01-JBS-52 (UTM 346512E, 5398007N, Zone 16) [STOP 1-6; Fig. 12] and apparently associated with local albitization of potassium feldspar megacrysts. Blocky potassium feldspar megacrysts up to 50 cm in diameter and muscovite books up to 10 cm in thickness were noted in the

potassic pegmatite and enclosed quartz-rich patches.

Dikes and foliation-concordant peraluminous granites and pegmatites were emplaced into Quetico Subprovince metasedimentary rocks during at least three intrusive episodes characterized by the following rock types:

- grey, garnet-biotite granite, fine- to medium-grained
- cordierite and garnet-cordierite granite
- sheets of pegmatitic leucogranite and associated quartz-rich patches

Stop 1-7A: Quetico Fault in Metasedimentary Rocks

UTM coordinates: NAD83; 16U 0348057E / 5401948N

As we near the Quetico Fault from the south, low-grade metasedimentary rocks give way to well-foliated schists which retain evidence of a layered sedimentary protolith but which may display incipient anatexis (Seemayer, 1992). They are characterized by a dominant, subvertical west-striking foliation. North of

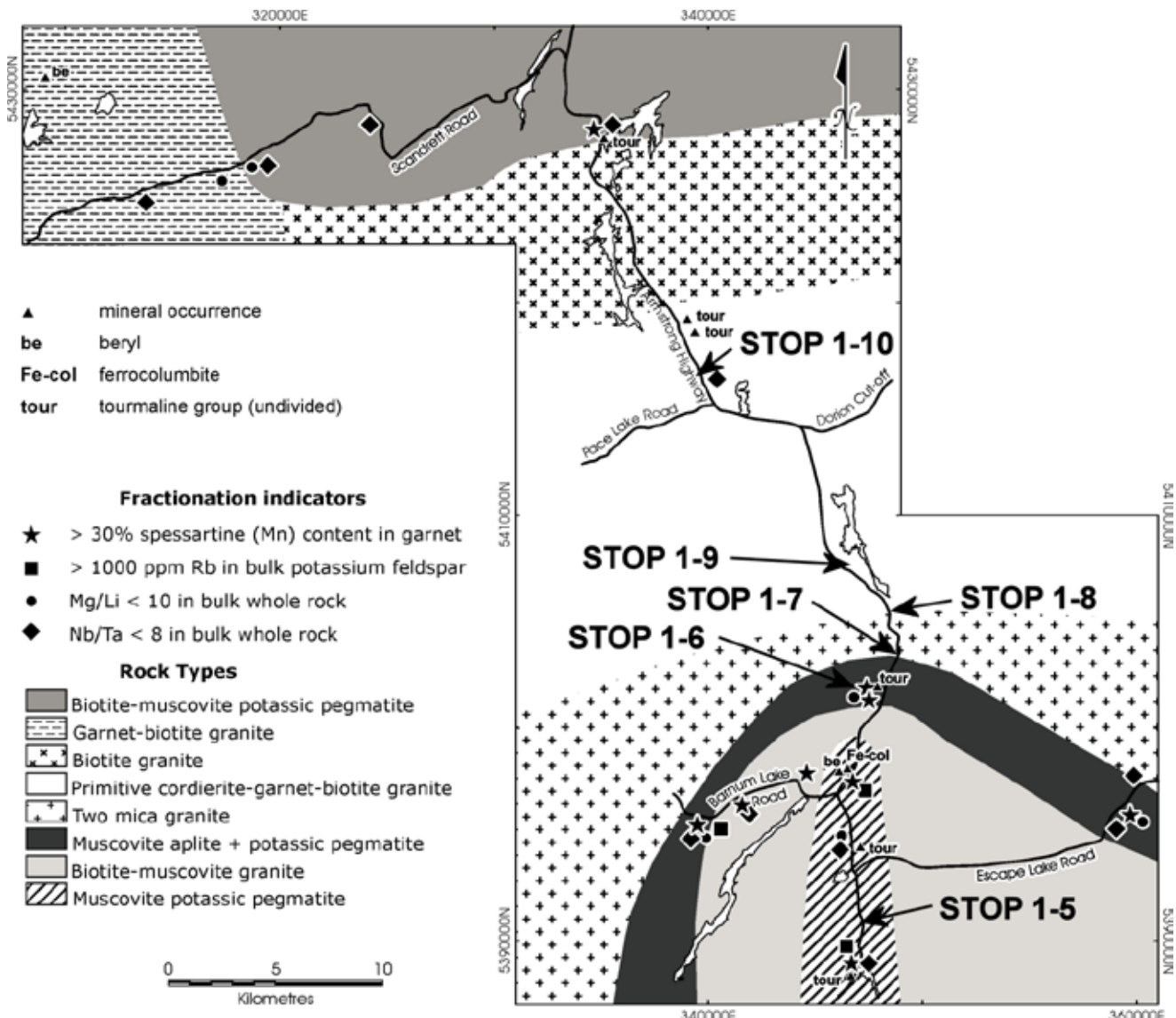


Figure 12. Geology and rare metal mineral occurrences along Highway 527 (after Breaks et al., 2003)

the Quetico Fault, migmatites predominate.

This section along Highway 527 was described by Seemayer (1992):

Mylonitic and cataclastic rocks of the subvertical Quetico Fault zone cut the migmatites of Block C [higher-grade subdivision]. The fault rocks outcrop for 2.5 km along Highway 527. Feldspars in mylonitized leucosome show a characteristic brick-red alteration colour which makes the fault rocks easy to identify. Recognizable stromatic migmatites in the fault zone show well-developed C-S fabric and abundant shear planes in the leucocratic layers. The fact that the migmatites have been sheared shows that final movement on the Quetico Fault post-dates

the migmatization and peak metamorphism, although the fault may have been initiated during an earlier period of transpression.

Mackasey et al. (1974) attributed a dextral displacement of 100 km to the Quetico Fault. Detailed examination of the fault from Rainy Lake to Highway 527 by Kennedy (1984) found evidence to support the dextral sense of strike-slip motion. Kennedy found that brittle deformation followed the predominantly ductile deformation within the fault zone. Purdon (1989) concluded that motion along the Quetico Fault northeast of Thunder Bay was of a complex nature. An early dip-slip component inferred from subvertical stretching lineations on foliation surfaces was

overprinted by slickenfibres resulting from dextral strike-slip motion. It is not surprising, then, that metasedimentary rocks showing incipient metamorphic differentiation are adjacent to well-segregated, stromatic migmatites separated by the fault, although the initial bulk composition of the rocks on either side of the fault was likely very similar.

A pervasive foliation (~250/70) is locally kink-banded. Granitic dykes and anastomosing, black patches of what may be pseudotachylite cross-cut the high-grade metamorphic rocks (Fig. 13).

Stop 1-7B: Quetico Fault in Granitoid Rocks

UTM coordinates: NAD83; 16U 0348117E / 5402246N

North of the ravine, a large rock cut displays red, coarse-grained to pegmatitic, K-feldspar granitoids. These granitoid rocks host numerous chlorite- and epidote-coated and slickensided joint surfaces and are locally well-foliated. Locally developed, shallowly west-plunging lineations and patchy pseudotachylite were also noted.

Stop 1-8: Pegmatitic Granite, Quetico Subprovince

UTM coordinates: NAD83; 16U 0348152E / 5405692N

Graphic-textured, pegmatitic granitoids are exposed

at a site where a bulk sample of feldspar was taken by local company, Thunderbrick Ltd. in 1981 (Assessment Files, Thunder Bay Resident Geologist's Office). Geochemical analyses and results of testing of its suitability for ceramics applications are not available.

Foliated pegmatitic rocks consist of quartz, feldspar, muscovite, garnet and biotite. Feldspar crystals may reach up to 60 cm in size. Garnet euhedra occur as disseminated crystals or trains of crystals. Biotite bands are also evident.

Stop 1-9: Migmatites, Quetico Subprovince

UTM coordinates: NAD83; 16U 0345756E / 5407714N

As noted by Seemayer (1992), stromatic (with lesser schlieric and agmatitic) migmatites predominate north of the Quetico Fault. Leucosome consists of quartz, plagioclase and perthite or microcline. Melanosome consists of biotite, quartz and plagioclase. Porphyroblasts of garnet, cordierite and sillimanite and mineral assemblages devoid of muscovite, but including alkali feldspar, reflect regional high-grade metamorphic conditions. Their occurrence is suggestive of diatexis of pelitic protoliths (Seemayer, 1992).

At this location, "lit-par-lit" migmatites, consisting of equal proportions of leucosome and melanosome display a gneissosity at 255/80° and schollen structure



Figure 13. Narrow zones of cataclasite and pseudotachylite associated with the Quetico Fault, Stop 1-7A



Figure 14. Schollen and schlieric migmatite, Stop 1-9

in large, rafted blocks. Quartzofeldspathic, pegmatite neosome dykes cut the migmatites (Fig. 14). Garnet porphyroblasts up to 1 cm in diameter are also noted.

Stop 1-10: Cordierite-bearing Granitoids, Quetico Subprovince

UTM coordinates: NAD83; 16U 039226E / 5416579N

Migmatitic rocks are intruded by cordierite-garnet-mica pegmatites along this stretch of highway, where relict bands and schollen of biotitic schists are still evident. These pegmatitic rocks were described in detail by Breaks et al. (2003):

Pegmatite sheets, at least 5 m thick, are evident on the Armstrong highway as at locality 01-FWB-105 near Keelor Lake (UTM 339218E, 5416563N, Zone 16). These sheets consist of coarse-grained, garnet-muscovite-cordierite granite that contain 15 to 20% cordierite crystals pervasively altered to soft, dark green-black pseudomorphs (Fig. 15). The coarse [-grained] granite is gradational into muscovite-rich, miarolitic cavity-bearing, pegmatite patches (blocky potassium feldspar;

muscovite, cleavelandite, quartz, brown black pyroxene and green fluorapatite).

STOP 1-11: Conglomerate, Wabigoon Subprovince

UTM coordinates: NAD83; 16U 0328624E / 5448888N

The 16 kilometer stretch of road that we have just



Figure 15. Cordierite crystals with incipient alteration along their margins, Stop 1-10



Figure 16. Location of field trip stops 1-11 to 1-15 (Google earth image). Scale is 1.2 km long.

traversed crosses a mainly metasedimentary terrain that is the western portion of the Beardmore-Geraldton area (Fig. 16). It is composed of three metasedimentary belts that are separated by metavolcanic belts. The northern metasedimentary belt is dominated by conglomerates and sandstones that were deposited by braided streams flowing from the Onaman-Tashota volcanic arc terrain to the north (Devaney, 1987; Fralick and Kronberg, 1997; Fralick et al., 1992; Fralick, 2003). The outcrop we will look at is part of this belt. The central metasedimentary belt to the east forms a northward younging, coarsening upward succession of ramp-fan turbidites (Barrett and Fralick, 1989) culminating in braid deltas interbedded with iron formation (Fralick and Pufahl, 2006). The same trend is present in the central metasedimentary belt here, though younging directions vary in the turbiditic portion due to isoclinal folding. The southern metasedimentary belt is composed of turbidites that are similar morphologically and geochemically to those in the central belt and the Quetico. This system served to deliver sediment to the Quetico trench as shown by these similarities and almost identical geochronology of their zircon populations (Fralick, 2003; Fralick et al., 2006). The main zircon population ranges from 2708 to 2698 Ma and represents erosion of synchronous calc-alkaline sub-areal volcanism occurring to the north. The clast composition reflects a source dominated by volcanic arc lithologies (mafic to felsic volcanics; diorite and other granitoids). Deep

seismic profiling reveals that this succession was overthrust onto the volcanic arc rocks and was itself overthrust by the Quetico metasediments. This agrees with previous work that concluded the Beardmore-Geraldton belt represents a forearc basin between the Onaman-Tashota subareal volcanic arc to the north and the Quetico accretionary complex to the south (Barrett and Fralick, 1989; Eriksson et al., 1994, 1997).

The outcrop we will examine forms a portion of the fluvial braided stream deposits that fed into the subaqueous portion of the basin to the south. Cobble-pebble conglomerates represent longitudinal gravel bars with interbedded coarse- to medium-grained sandstones formed as waning flow sand sheets on the bar tails. Channels are represented by the thicker sandstones, commonly containing trough cross-stratification, and some of the conglomeratic units. Small chute channels that cut into the bar tops during waning flow are represented by the sandstone lenses in the conglomerate. The rivers in this area were transitional to the south into sandy braided rivers, producing outcroppings of trough-cross-stratified, coarse- and medium-grained sandstone. This also formed in distributary mouth bars where the rivers flowed into the ocean to the south. The iron formations in the region are associated with this subaqueous, delta-top environment. In the marine basin to the south of the deltas turbidites accumulated in water that was shallow enough to allow storm reworking of the tops of these



Figure 17. Conglomerate, Stop 1-11

layers into dunes and ripples. Changes in river mouth position allowed muddy successions a few meters thick to develop during sediment starved intervals.

The sequence has been tectonised as evidenced by the flattening of the clasts in the conglomerate (Fig. 18). The mafic clasts have been affected the most and in places are reduced to ribbon-like shapes. In contrast, the granitoid clasts behaved as rigid bodies, tending to fracture rather than deform plastically.



Figure 18. Deformed conglomerate, showing rotation of competent granitoid cobbles and flattening of less-competent clasts, Stop 1-11

Stop 1-12: Tuff-breccia, Wabigoon Subprovince

UTM coordinates: NAD83; 16U 0328280E / 5449317N

A deformed intermediate tuff-breccia occurs within the dominantly mafic metavolcanic sequence near the intersection of Highway 527 and Kingdon Lake Road. Flattened, buff-coloured pyroclasts (lapilli to bombs) have aspect ratios ranging from 2:1 to >10:1 and define a strong foliation (050/50). Both pyroclasts and matrix are feldspar-phyric.

Stop 1-13: Pillowed Mafic Metavolcanic Rocks, Wabigoon Subprovince

UTM coordinates: NAD83; 16U0328159E / 5449467N

Pillowed mafic flows are best-exposed on the eastern side of the highway (Fig. 20). They are intensely foliated (055/52°) at the southern end of the outcrop. Light green-weathering bun and mattress pillows are locally vesicular/amygdaloidal and have darker green selvages; small re-entrants were noted. Pillows have been flattened into ovoid shapes with aspect ratios ranging from 5:1 to 10:1, precluding unequivocal “tops” determination.



Figure 19. Tuff-breccia, showing flattened pyroclasts, Stop 1-12

Stop 1-14: Brecciated Metavolcanic Rocks, Wabigoon Subprovince

UTM coordinates: NAD83; 16U 0326447E / 5450297N

perhaps both pyroclastic and autoclastic brecciation, is exposed on the eastern side of the highway. Flattened, buff (altered?) aphanitic fragments define a foliation at 050/55.

An enigmatic volcanic unit, showing aspects of



Figure 20 Pillowed basalt flow, Stop 1-13



Figure 21. General geology and field trip stops, Day Two (Mesoproterozoic)

Stop 1-15: Banded Iron Formation, Wabigoon Subprovince

UTM coordinates: NAD83; 16U 0324893E / 5451015N

A rusty-weathering sulphide-facies banded iron formation occurs within the volcanic succession. It is intercalated with tuffaceous and feldspathic, fragmental (pyroclastic?) rocks. The iron formation consists of pyrite, pyrrhotite + chalcopyrite and chert and has been

a focus of base metal exploration. Dome Exploration (Canada) Ltd. drilled this iron formation in 1972; no analytical results were reported (assessment files, Thunder Bay Resident Geologist's Office, Thunder Bay).

Day Two (Figs. 21 & 22)

Stop 2-1: Pillar Lake Volcanic Rocks, Alarie Quarry

UTM coordinates: NAD83; 16U 0356068E / 5573375N

The Pillar Lake volcanic suite was first recognized and mapped by Macdonald (2004) and is the subject of an ongoing study by Magee (in progress). Initial observations suggested that it largely consisted of a sequence of flat-lying pillowed and massive flows and autoclastic and hyaloclastic breccias. The volcanic rocks unconformably overlie the Badwater gabbro (~1599 Ma) and Badwater syenite (~1590 Ma) and are capped by Keweenawan Inspiration diabase sills (1159 + 33 Ma), yielding an apparent thickness of 20 to 40 m (Hart and Macdonald, 2007; Heaman et al., 2007). Titanite in an andesitic unit yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1129.0 + 4.6 Ma; no zircon nor baddeleyite was recovered in the sample (Heaman et al., 2007).

Reinvestigation of these volcanic rocks in 2010 (Smyk et al., 2011) was prompted by a new exposure south of Armstrong that had been created during ballast quarry development (Fig. 23). The quarry face exposes a ~15 m section of thin (0.5 to 2 m), flat-lying, variably altered, columnar-jointed, basaltic andesite flows, capped by a diabase sill. Individual flows may persist over the 130 m length of the exposure while others bifurcate and terminate as thin tendrils in flow breccia (Fig. 24). Autobrecciated zones are rubbly weathering and occupy the spaces between thin, pinching flows. Thin flow top breccias separate flows. Massive flows contain zones of pipe amygdulites at their bases and tops. The morphology and disposition of these flows is suggestive of an intercalated, subaerial pahoehoe and a'a flow succession.

These volcanic rocks are variably altered along fractures, flow contacts and within brecciated zones. This hydrothermal alteration is typically manifested as a beige to pink discolouration of the dark grey-black flows resulting from the destruction of primary ferromagnesian minerals and the introduction of alkali feldspar, sericite and quartz. Void spaces along joints and fractures and in vesicles has been occupied by large (< 3 cm), black, euhedral actinolite crystals.



Figure 22. Locations of stops 2-2 to 2-5 along Mattice Road and Highway 527 (Google earth image). Scale bar is 1452 m long.

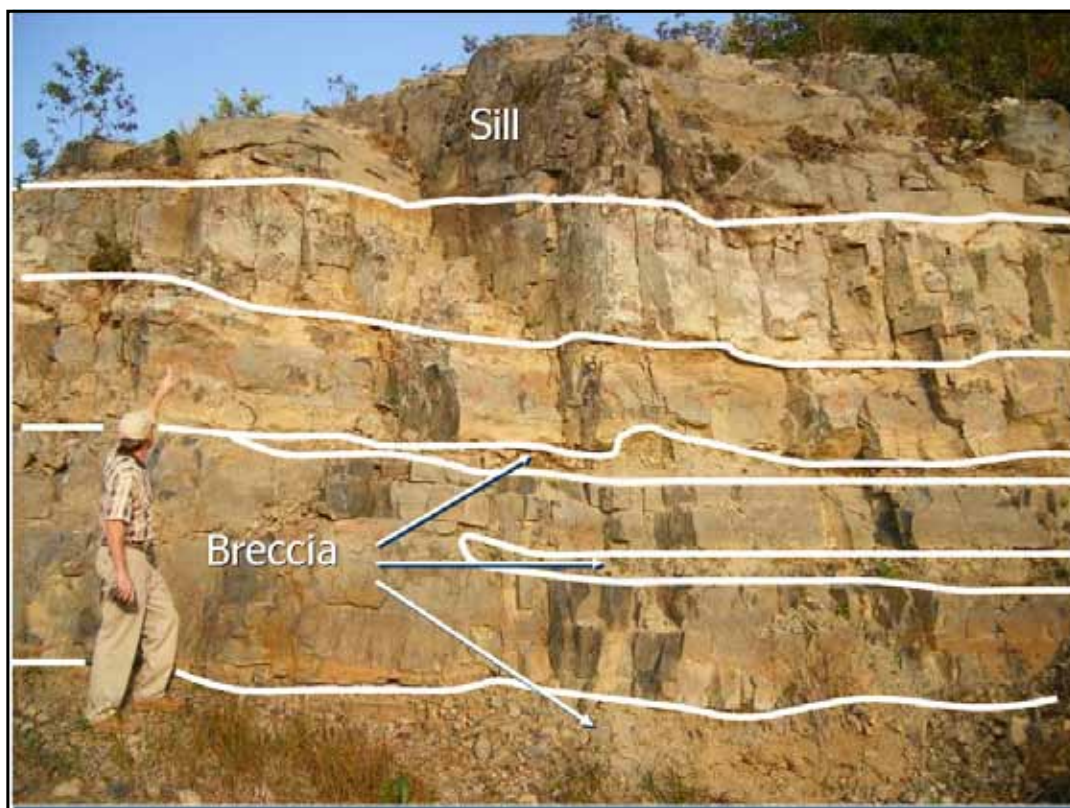


Figure 23. Thin, amygdaloidal (a'a?) flows separated by interflow breccia, quarry face, Stop 2-1. The flow succession is overlain by an Inspiration diabase sill.



Figure 24. Thin, bifurcating flows separated by rubbly breccia, quarry face, Stop 2-1. Scale bar (left) is 1 m long

Small acicular asbestiform crystals of edenite were also identified by x-ray diffraction analysis. Alteration is characterized by increases in Al_2O_3 , K_2O , Na_2O , and SiO_2 and decreases in CaO , Fe_2O_3 , MgO , P_2O_5 and TiO_2 (Fig. 25).

The trace element geochemistry of the sill that caps the quarry face is identical to that of the Inspiration sill identified by Hart and Macdonald (2007), suggesting that this sill is part of that intrusive suite (Figs. 26 & 27). Samples of the volcanic rocks display REE enrichment and negative Nb anomalies comparable to the range of samples of the Pillar Lake volcanic

suite reported by Magee (personal communication, 2011; Fig. 28). Samples of the Inspiration sills are geochemically indistinguishable from the Pillar Lake volcanic rocks and to least-altered pillowed samples with the lowest LOI values. The similarity between these two enigmatic suites suggests that they may be derived from the same source.

MacDonald and Tremblay (2005) distinguished the Inspiration sill, which exclusively overlies the Pillar Lake rocks, on the basis of its normal polarity and distinct geochemistry. Primary clinopyroxene in the Inspiration sill was commonly replaced by actinolite

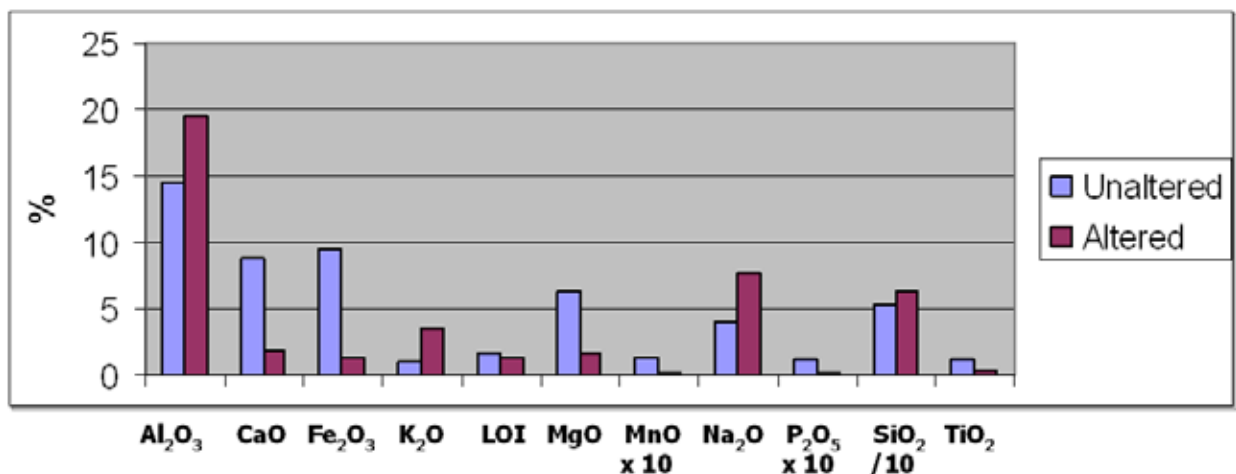


Figure 25. Comparison of major element chemistry, unaltered and altered basaltic andesite, Stop 2-1

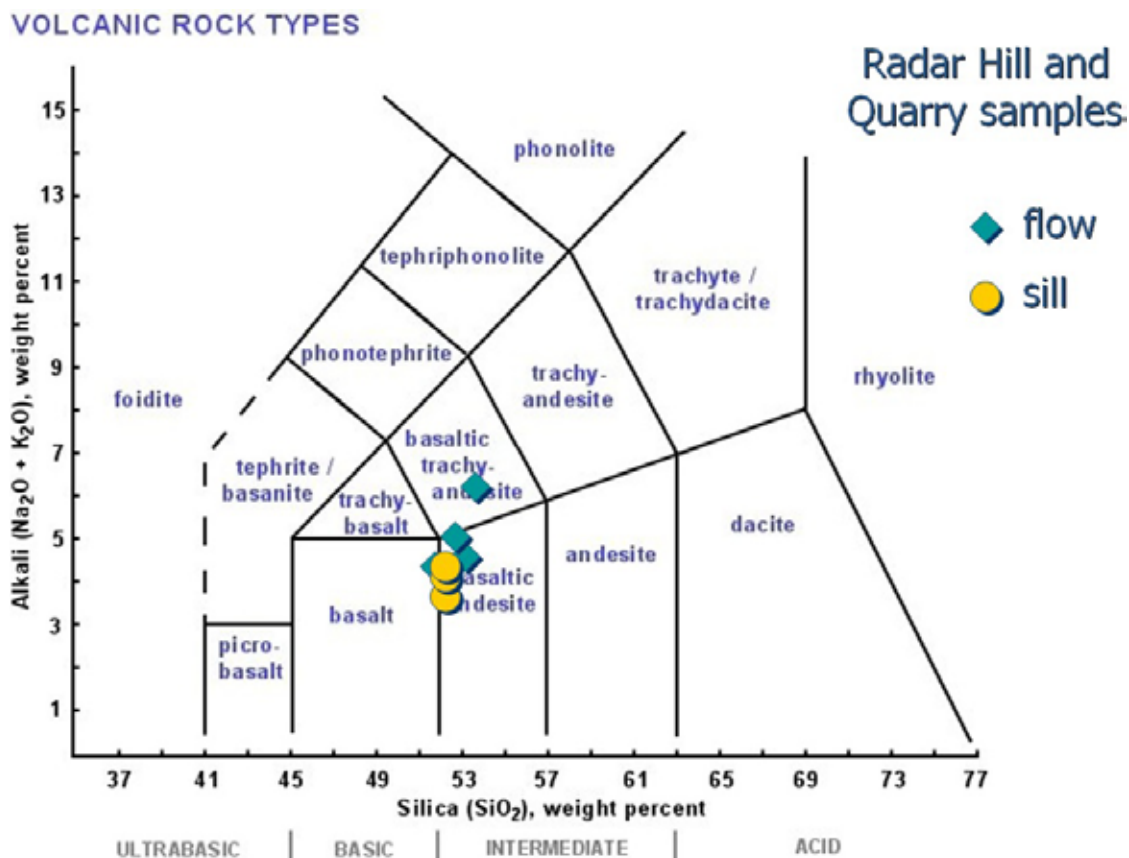


Figure 26. TAS plot for basaltic andesite (diamonds) and Inspiration diabase (circles), Stop 2-1

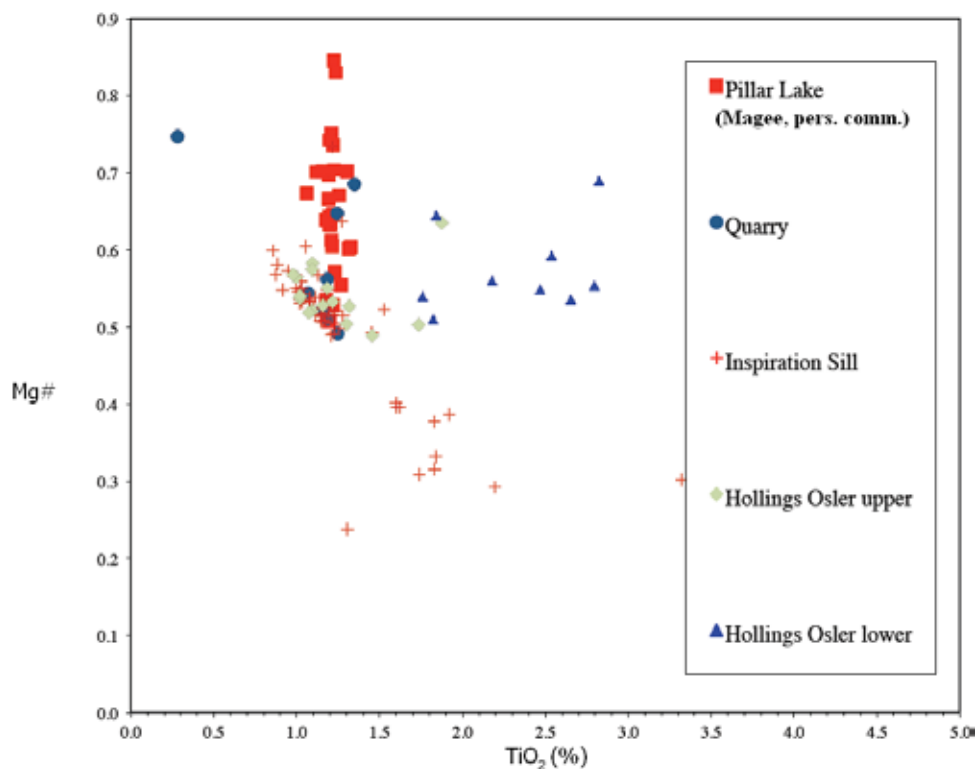


Figure 27. Plot of $\text{Mg\#}$ versus TiO_2 for Pillar Lake volcanic rocks collected by Magee (in progress; squares) and at Stop 2-1 by Smyk et al. (2011; circles), compared to Inspiration sill and Osler Group volcanic rocks

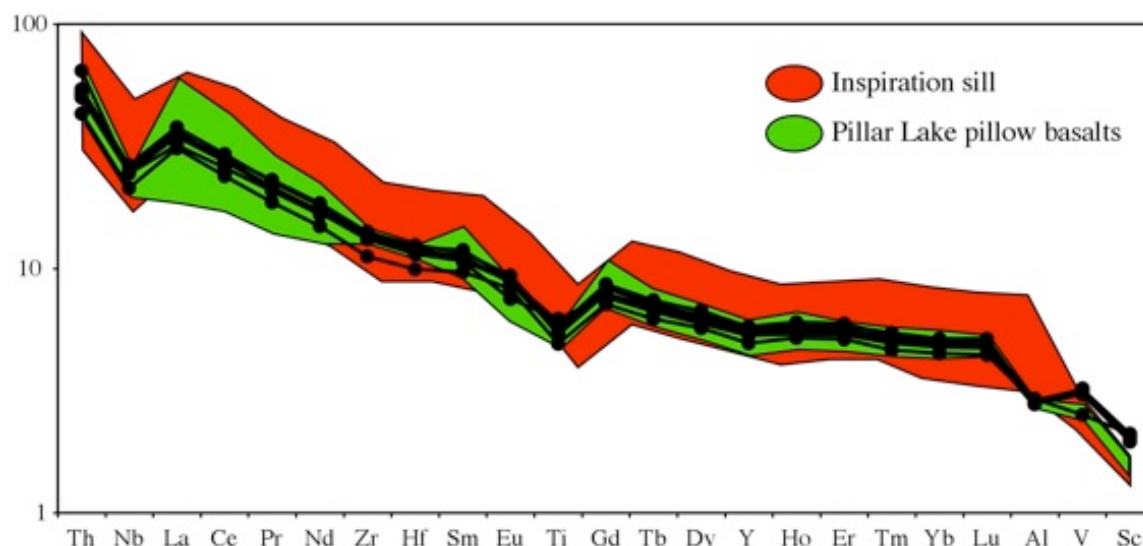


Figure 28. Chondrite-normalized REE plot for Pillar Lake volcanic rocks and Inspiration diabase (Smyk et al. 2011)

(Schandl, 2004; Fig. 29). MacDonald et al. (2005) also noted a remarkable lack of chilled margins on the Inspiration sill and suggested that it was older than the Nipigon sills. Later work by Heaman et al. (2007) determined an age of 1159 ± 33 Ma for the Inspiration sill, which is consistent with the early period of normal polarity. Smyk et al. (2011) proposed that the Pillar Lake basalts may, in fact, be coeval with the Inspiration sill which, in turn, may represent either a subvolcanic intrusion or possibly a massive, ponded flow / lava lake. This may account for the extensive alteration in the Pillar Lake basalts with the sill acting as an impermeable cap to hydrothermal fluids, which were concentrated in the underlying volcanic flows. The similarity of the Pillar Lake and Inspiration sill magmatism to other magmas of the Midcontinent Rift,

including the Nipigon sills and Osler volcanic rocks (Hollings et al., 2007), suggests that these rocks may represent a very early extrusive to subvolcanic phase of rift activity. The location of these rocks close to the ~ 1599 Ma Badwater gabbro, the ~ 1590 Ma Badwater syenite and the ~ 1540 Ma English Bay anorogenic granite, which have been interpreted as evidence for a long-lived crustal weakness (Hollings et al., 2004), may offer an explanation for how these magmas were erupted so early in the rift history.

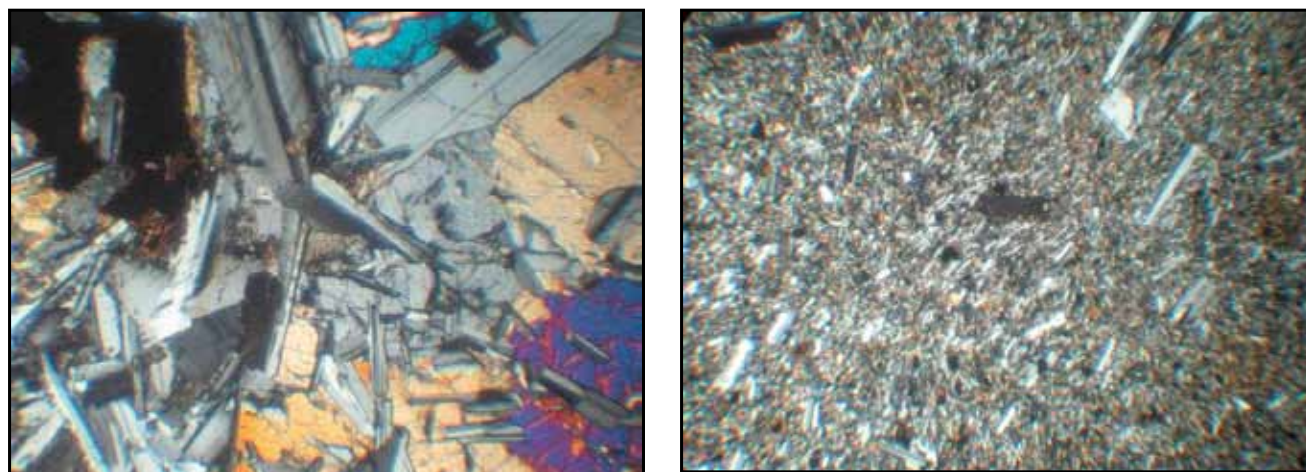


Figure 28. Photomicrographs of basaltic andesite (left) and Inspiration diabase (right), crossed nicols, FOV = 4 mm in both images, Stop 2-1.

Stop 2-2: Pillar Lake Volcanic Rocks, T-Junction, Mattice Lake Road

UTM coordinates: NAD83; 16U 0345619E / 5564137N

Several low-lying outcrops and large boulders cluster around a T-junction on the Mattice Lake Road. First discovered by MacDonald (2004), these Pillar Lake volcanic rocks display features indicative of submarine extrusion. Supporting evidence includes autoclastic breccias (flow/pillow breccia), hyaloclastite and pillow forms (Fig. 29). MacDonald (2004) described these rocks in this vicinity:

An apparently flat-lying succession of mafic metavolcanic rocks crops out near the north end of Pillar Lake. The unit extends over an approximately 20 to 25 km² area and consists of pillowed flow breccia, hyaloclastite breccia, pillowed and massive flows. Examination of outcrops from several locations would suggest that the apparently flat-lying unit has a thickness between 20 and 40 m. Lower portions consist of 3 or 4 alternating beds of pillowed flows with accompanying flow top breccia and or pillow breccia and hyaloclastite overlain by a 1 to 2 m thick layer of massive flow and/or sill. Locally, this unit displays slightly flattened 10 cm to 1 m diameter pillows with interflow hyaloclastite. Concentric jointing and tortoise shell-like cracks are common. Alteration of this unit typically consists of weakly to moderately pervasive to moderate patchy hematite alteration and local weak to moderate patchy sericite alteration. Alteration intensity increases with proximity to the north tip of Pillar Lake and along a creek leading between Mundell and Pillar lakes. The flat-lying and relatively undeformed nature of these flows, combined with the well-preserved nature of relatively delicate primary features such as hyaloclastite may suggest that these rocks are Proterozoic rather than Archean.

North of Pillar Lake, drilling in 2004 showed that the Pillar Lake volcanic rocks (unconformably?) overlie the 1599 Ma Badwater gabbro. The ongoing uncertainty regarding the absolute age of the Pillar Lake volcanic rocks was summarized by Heaman et al. (2007):

When first discovered in 2003, there was some question as to whether these volcanic rocks were Archean or Proterozoic in age, although their flat-lying character and alteration patterns

indicate that they are likely Proterozoic. The first sample ..., a mafic volcanic flow, was collected in 2003 to directly determine the age of these volcanic rocks. A total of about 30 colourless irregular zircon grains were recovered, but none of these have the physical characteristics readily attributed to volcanic zircon (generally too large). Two single zircon grains were analyzed but did not yield consistent results (Heaman and Easton, 2006) and were not considered to provide a good constraint on the time of volcanism. Subsequent mapping in the McLaurin Lake area revealed the presence of a flat-lying stratigraphic succession consisting of a lower, pillowed mafic volcanic flow, a thin sandstone horizon ...and an overlying sequence of mafic and andesitic flows ..., all overlain by an Inspiration diabase sill ... Because of the well-preserved stratigraphic relationships, this section was sampled in detail for geochronology, to at least bracket the age of volcanism, even if it were not possible to directly obtain an age from the volcanic rocks themselves.

A sample of fine-grained grey andesite from McLaurin Lake was collected to establish the age of the Pillar Lake volcanics in this region. There was no zircon or baddeleyite recovered from this sample; however, a modest amount of rutile and titanite was recovered. The multigrain fractions of rutile (1) and titanite (2) both have low uranium contents (8.3 and 7.6 ppm) and contrasting Th/U (0.222 and 27.818, respectively). The rutile fraction is concordant with a ²⁰⁶Pb/²³⁸U age of 1106.4 ± 2.8 Ma. The titanite fraction is also within error of concordia with a slightly

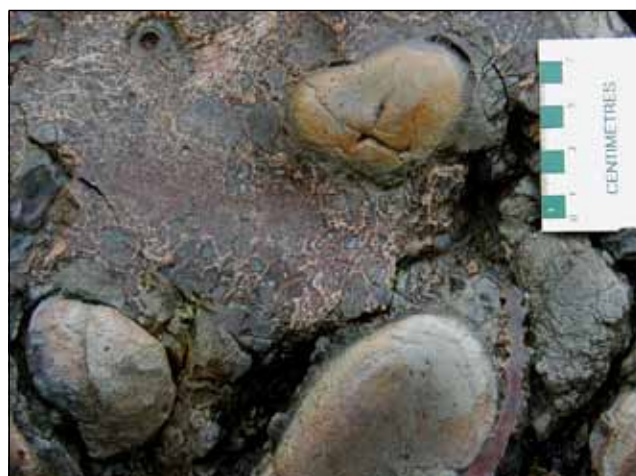


Figure 29. Autoclastic pillow breccia and hyaloclastite, Stop 2-2

older $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1129.0 ± 4.6 Ma. The titanite age of 1129 Ma is interpreted to be the best constraint on the age of the McLaurin Lake andesite. It could be a minimum age or it could closely mark the age of volcanism. The 1106 Ma rutile age is similar to the age of the McLaurin Lake diabase (reported previously) and could reflect thermal resetting during emplacement of the diabase (rutile has a much lower closure temperature to Pb diffusion of $\sim 400^\circ\text{C}$).

Stop 2-3A: Badwater Gabbro, Badwater Creek Road

UTM coordinates: NAD83; 16U 0347047E / 5564306N

Stop 2-3B: Badwater Gabbro, East of Badwater Creek Road

UTM coordinates: NAD83; 16U 0347043E / 5564142N

The Badwater gabbro is best-exposed north of Pillar Lake, where it is overlain by Pillar Lake volcanic rocks. It was dated by Heaman et al. (2007) at 1598.7 ± 1.1 Ma. It was intruded by the Badwater syenite at 1590.1 ± 0.8 Ma (Fig. 30). Drilling by East West Resource

Corporation in 2004 and 2008 indicated that gabbroic rocks underlie Pillar Lake volcanic rocks from at least west of Pillar Lake, east to McLaurin Lake (Middleton, 2004; Middleton and Bennett, 2008).

At this locality, the gabbro is typically coarse-grained and unaltered. Anorthositic bands, variations in cumulus and intercumulus minerals, and igneous foliation are suggestive of igneous layering (Fig. 31). Thin section petrography cited by Middleton and Bennett (2008) for a gabbro drilled 2.3 km east of this location listed a modal mineralogy as approximately.

Plagioclase (labradorite/bytownite) 55%

Clinopyroxene (augite?) 25%

Biotite 10%

Olivine (partly relict) 3%

Talc/sericite, minor iddingsite (after olivine) 2%

Amphibole (secondary, actinolitic) 2%

Opaque (magnetite?) 2%

(pyrrhotite?) 1 %

Clay? /sericite (after plagioclase) trace

Plagioclase forms mainly euhedral crystals up to about 4 mm long, with random orientations, partly



Figure 30. Badwater gabbro xenoliths in feldspar-phyric syenite, eastern shore of Pillar Lake



Figure 31. Badwater gabbro, displaying igneous foliation and anorthositic lens/layer, Stop 2-3B.

enclosing the mafic minerals. Pale brownish green clinopyroxene forms somewhat irregular, mainly subhedral crystals up to 4.5 mm long. Some crystals are partly altered, mainly around the rims, to minor secondary amphibole. Biotite forms ragged, irregular subhedral crystals mostly <2.5 mm in diameter, commonly wrapped around pyroxene, or interstitial to pyroxene and plagioclase. Olivine (or relict olivine) displays somewhat rounded or subhedral outlines up to almost 3 mm in diameter, commonly contained within pyroxene crystals or aggregates. The olivine is generally strongly fractured, with traces of minute secondary magnetite along the fractures. In places, the olivine is partly to locally completely replaced or pseudomorphed by very fine-grained talc/sericite or minor red-brown to greenish-brown iddingsite. Accessory opaque minerals appear to be mostly magnetite forming skeletal to irregular subhedral crystals up to 2 mm in diameter, interstitial to plagioclase, or associated with biotite and olivine or relict olivine. Sulfides (mostly pyrrhotite) form irregular subhedra up to 0.5 mm long and are also commonly interstitial to plagioclase and pyroxene, and are associated with biotite (Middleton and Bennett,

2008).

STOP 2-4: Pillar Lake Volcanics, Chimney Lake

UTM coordinates: NAD83; 16U 0349738E / 5563345N

This outcrop on the northern shore of Chimney Lake is enigmatic (Fig. 32). It appears to be a volcanic breccia, but with variable clast compositions and what appears to be clastic matrix material. It could represent a mass-flow resulting from topography created by volcanism or fault movements related to magma recharge, though the lack of reaction rims makes a lahar improbable. Could it possibly be a “diatrema”? We will discuss its genesis in the field.

A ~1m thick interflow sandstone unit described by Magee (in progress), approximately 1.5 km north of Chimney Lake consisted of basal sandstone, lithic arenite, and an upper quartz grain matrix breccia with locally derived basalt clasts. 50 detrital zircons from the interflow sandstone (Heaman et al., 2007) yielded a youngest concordant zircon of 1514 Ma. Dominant zircon populations fall between: 2700 to 2300



Figure 32. Fragmental unit, Pillar Lake Volcanics, Chimney Lake, Stop 2-4

Ma; 1950 to 1900 Ma; 1880 to 1780 Ma. The basal sandstone geochemical signature is andesitic (Magee, in progress).

STOP 2-5: Pillar Lake Volcanic Rocks, Highway 527

UTM coordinates: NAD83; 16U 0349460E / 5562320N

This road cut on the eastern side of Highway 527 provides a ~3 m section through flat-lying basalt flows with small (10 to 30 cm) ellipsoidal features, suggestive of bun pillows. However, Smyk et al. (2011) suggested that these features may be pahoehoe toes in cross-section. The suggestion that these were subaerial flows is supported by the discovery of ropy flow top in a dislodged boulder in the ditch at the base of the exposure (Fig. 33).

STOP 2-6: Nipigon Diabase Sill, Highway 527

UTM coordinates: NAD83; 16U 0342613E / 5499194N

An 85 m long, 12 m high road cut provides an exceptional exposure of a Nipigon diabase sill at this locality (Fig. 34). The diabase is massive, homogeneous, fine- to medium-grained and locally feldspar-phyric. Irregular joint faces have been infilled with coarse,

drusy quartz-chlorite-calcite-pectolite(?) + malachite veins. An orthogonal set of shallowly south-dipping and steeply north-dipping joints suggests that the sill may be shallowly south-dipping. The glacially polished surface on top of the road cut displays scattered pink, feldspathic (granophyric?) patches and pectolite(?) coated joint surfaces.

STOP 2-7: Sibley Group Sedimentary Rocks and Nipigon Diabase Sills

UTM coordinates: NAD83; 16U 0341205E / 5493814N

Sibley Group sedimentary rocks are typically preserved in this area of the Nipigon Embayment below diabase sills which protect them from erosion. At this locality, straddling an overgrown logging road, a 3 to 4 m section of white-weathering, thinly bedded Rosspport Formation dolomites are sandwiched between two thin Nipigon diabase sills (Fig. 35). As a result, the calcareous sedimentary rocks have been contact metamorphosed, resulting in the formation of metamorphic calc-silicates. Where Sibley Group rocks rich in carbonates, such as the Middlebrun Bay or Channel Island Members of the Rosspport Formation, are proximal to thick sills the mineralogy consists of sodium- and potassium-rich varieties of pargasite, tremolite, talc, magnesium-rich clinocllore and calcite.

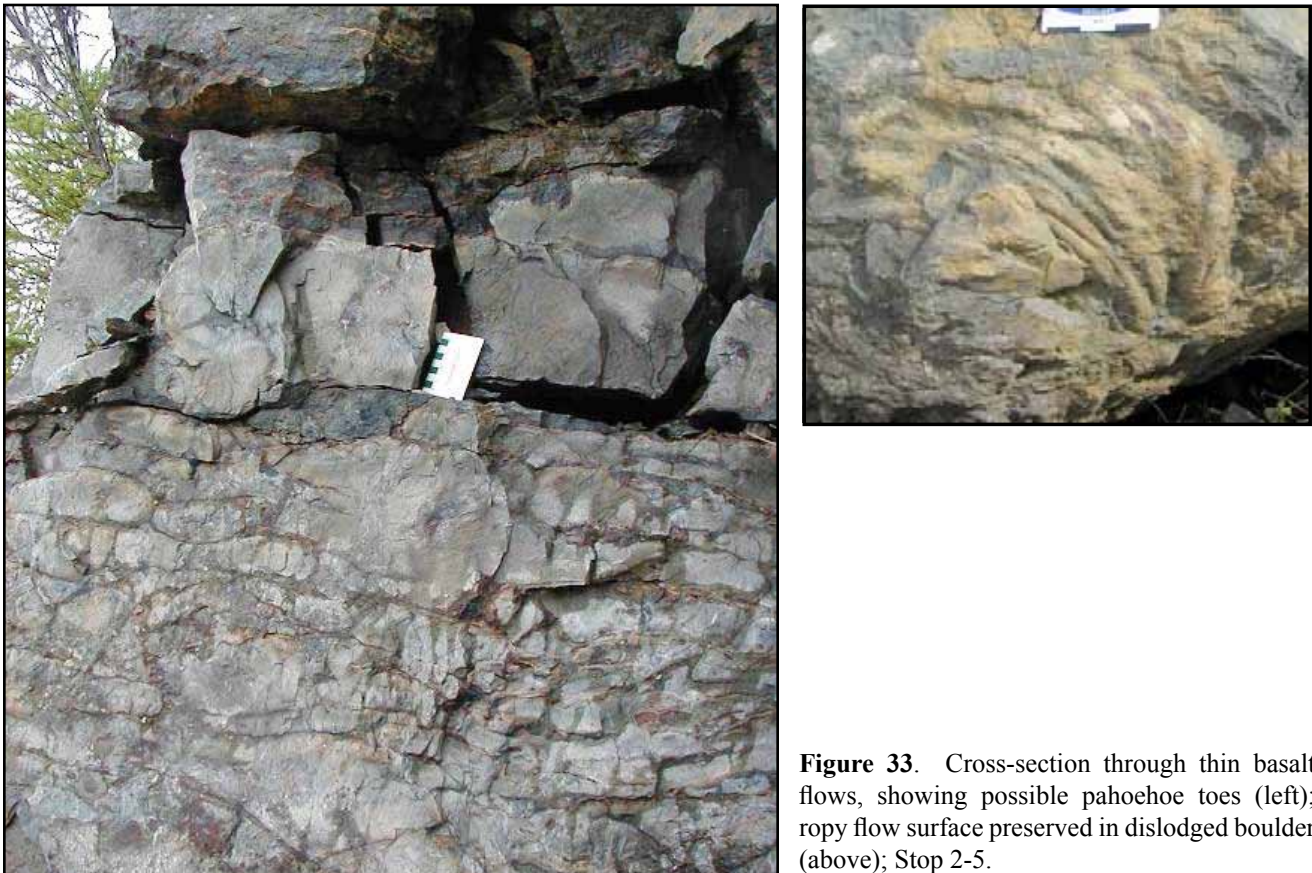


Figure 33. Cross-section through thin basalt flows, showing possible pahoehoe toes (left); ropy flow surface preserved in dislodged boulder (above); Stop 2-5.

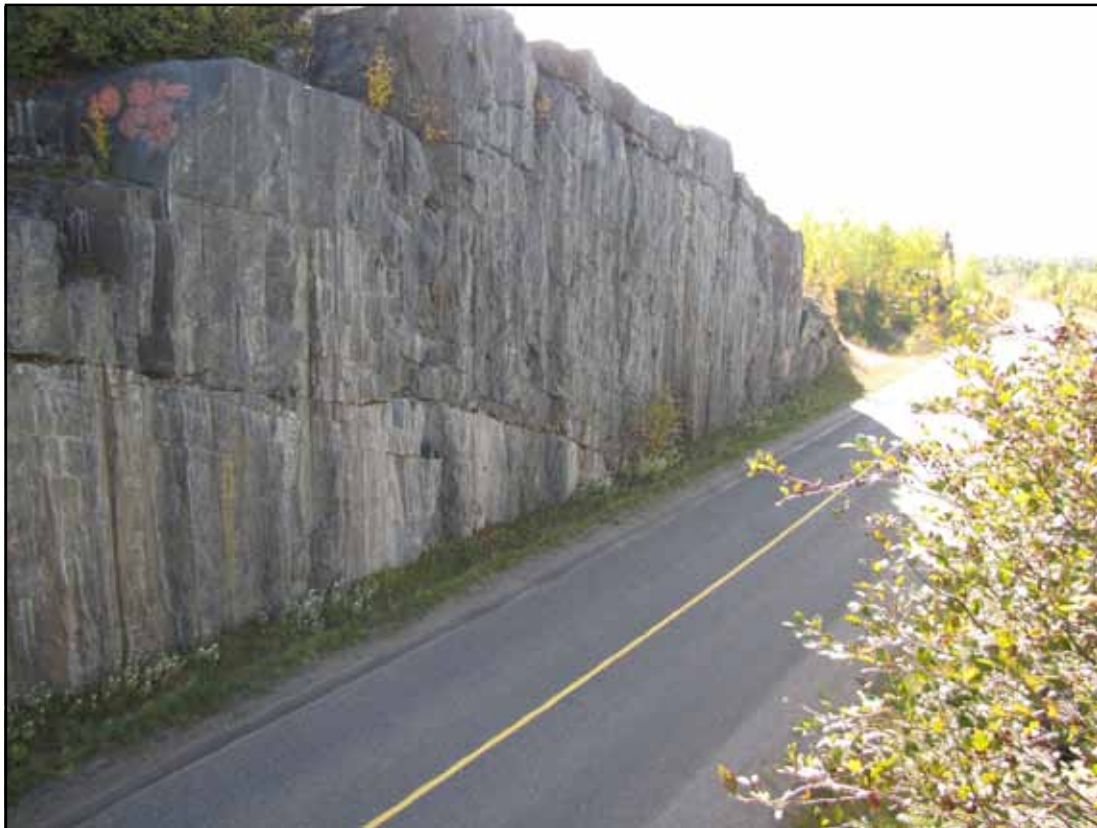


Figure 34. Inspiration diabase sill, showing shallowly south-dipping joints, Stop 2-6

Pargasite is common near the large sills, but tremolite, then talc, become important with distance away from the heat source (Rogala et al., 2005). This indicates temperatures in the 600°C dropping to 400°C range. The outcrop we are visiting appears to have initially been largely composed of dolomite. This combined with relict bedding and possible stromatolitic structures that are still evident indicates that this may represent the Middlebrun Bay Member.

On the other (western) side of the road, a flat-lying outcrop displays well-developed polygonal jointing (aka “tortoise-shell” texture) that are characteristic of the upper chilled contacts of these sills (Fig. 36). Feldspathic alteration along these joints results in their raised appearance. Numerous, recessively weathering ovoid pits may be the remnants of fluid- and volatile-rich pockets which migrated to the top of the cooling



Figure 35. Buff-weathering metadolostone, Stop 2-7



Figure 36. “Tortoise-shell” texture developed in polygonal joints at the chilled top of an Inspiration diabase sill, Stop 2-7

sill.

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Field trip 9 - Rehabilitation of the Past-Producing Shebandowan and North Coldstream Mine Sites

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Introduction

This trip will provide an overview of rehabilitation measures that have been implemented at two past-producing mines that are located in the Shebandowan greenstone belt west of the City of Thunder Bay (Fig. 1). The Shebandowan Mine operated between 1971 and 1998, producing 9.29 million tonnes of ore grading 1.75% nickel, 0.88% copper, 0.063% cobalt, 0.0533 oz/ton platinum group elements and 0.0575 oz/ton silver (Inco, 2001). The North Coldstream Mine operated between 1957 and 1967, producing approximately 2.5 million tonnes grading 1.97% copper, 0.012 ounces per ton gold and 0.22 ounces per ton silver (Golder Associates, 2002).

Both mines produced significant quantities of acid-generating tailings during their operational lives and provide a good illustration of historic and current

mining waste management practices for base metal mines, and the technologies that are employed to prevent and mitigate adverse water quality impacts.

Mine Rehabilitation Regulatory Framework

Mineral exploration, mine development and mine rehabilitation in the Province of Ontario are regulated under the Mining Act. Part VII of the Mining Act deals principally with the rehabilitation of mines and mining lands and was proclaimed in 1991 (with the most recent significant amendments occurring on June 30, 2000). Under Part VII, proponents of all advanced exploration projects and operating mines are required to file a certified Closure Plan including financial assurance to indicate the method, schedule and cost of all rehabilitation to be conducted on the site once



Figure 1. Location map illustrating Shebandowan and North Coldstream Mine sites relative to the City of Thunder Bay.

closure commences.

Closure Plans are not mandatory for historic mines that closed before 1991. However, all proponents of mining lands are responsible to ensure that any historic mine hazards on their property are progressively rehabilitated to prescribed standards. The minimum standards for mine rehabilitation are prescribed under Ontario Regulation 240/00 – Mine Development and Closure under Part VII of the Act. The Shebandowan Mine, which operated until 1998, is an example of a mine site that is being rehabilitated under a Mine Closure Plan, while the North Coldstream Mine, which closed before 1991, is currently undergoing progressive rehabilitation.

The Ministry of Northern Development and Mines (MNDM) is the government agency that is responsible for the administration and enforcement of the Mining Act.

Shebandowan Mine

Exploration and Development History

The following summary of the historical exploration and development of the Shebandowan Mine property is based on information compiled from the files of the MNDM Mines and Minerals Division office in Thunder Bay.

- 1913: Nickel-copper ore found at Discovery Point by prospector Julian Cross.
- 1936: International Nickel Company (Inco) purchased property.
- 1936-1965: Various surface exploration programs were carried out.
- 1966-1967: No.1 development shaft completed, underground diamond drilling.
- 1968: Inco announced decision to develop mine and supporting facilities at Shebandowan. No.2 (production) shaft and mill commissioned.
- 1969-1974: Forest debris created by road and on-site construction removed, topsoil stockpiled for later usage and contouring, and revegetation began (hydro-seeding of grass with oats/rye, straw mulch; 15 000 seedlings / trees).
- 1973: Mine and mill complex officially opened June 28.
- 1986-1988, 1992-1995: Operations temporarily suspended due to economic conditions.

1996: Mine Closure Plan accepted by MNDM.

1998: Shebandowan Mine permanently ceased operation and work began to implement the Mine Closure Plan. Work completed to date includes flooding of tailings basin, waste rock relocation to tailings pond, infrastructure demolition and removal, capping of mine openings, closure of two landfills, and revegetation of disturbed areas.

2001: Mine Closure Plan Amendment filed with MNDM.

2003: Option / joint-venture agreement signed between Inco (now Vale) and North American Palladium (NAP) to explore former mine property and environs.

2008: Underground exploration, ramp/decline advanced to collect bulk sample of ore from Shebandowan West deposit. This work was done under a separate advanced exploration Closure Plan filed by NAP. Operations were suspended due to the fall 2008 economic downturn.

2012: Vale continues to be responsible for the implementation of the Shebandowan Mine Closure Plan, while NAP is responsible for the Shebandowan West site. Denison Environmental has been contracted by Vale to carry out on-going site maintenance and rehabilitation activities.

Deposit Geology

The Shebandowan nickel-copper deposit is hosted in a serpentinitized peridotite sill that forms part of a mafic metavolcanic rock-dominated sequence (Morin, 1973; Osmani 1997). The ore body is also located near the southern margins of the quartz diorite Shebandowan Lake Stock. The northwest-trending Crayfish Creek Fault (a regional-scale dextral transcurrent structure) is located immediately south of the ore body. Rocks to the south of the fault form a separate domain consisting largely of intercalated felsic, intermediate and mafic metavolcanic rocks (Osmani, 1997).

The Shebandowan Mine ore body included three styles of mineralization: massive, breccia, and stringer ore (Osmani, 1997). Massive ore consisted of pyrrhotite, pentlandite, chalcopyrite, pyrite and magnetite. Breccia ore was comprised of pyrrhotite, chalcopyrite and pentlandite, and contains fragments of peridotite, mafic metavolcanics and granitic rock. Stringer ore occurred as stringers of chalcopyrite, pyrite and minor pyrrhotite and pentlandite in shear zones. The average width of the ore body was approximately 7.5 m (Inco, 2001) and was mined over a strike length

of approximately 3.5 km to a maximum depth of approximately 1000 m.

Field trip stops

Stop 1: Discovery Point

UTM coordinates NAD 83; 15U 0701365E / 5386525N

Nickel-copper ore was first discovered 99 years ago along the shoreline of Lower Shebandowan Lake at Discovery Point. The initial underground exploration shaft was sunk at this location by Inco during the 1960s. The shaft location is now marked by a vented reinforced concrete cap and is located near the west end of the ore body (workings extend approximately 500 m further to the west). The underground mine workings are mostly located below the lake, and extend a further 3 km to the east.

Rehabilitated vent raises are located on an island



Photo 1. Reinforced concrete cap, Shaft No. 1 - Discovery Point

that can be seen from the outcrop near the shoreline immediately east of the shaft, while the No. 2 production shaft was located on a point behind the island. The mining method used at Shebandowan was cut and fill



Figure 2. Satellite image of Shebandowan Mine site showing field trip stop locations.

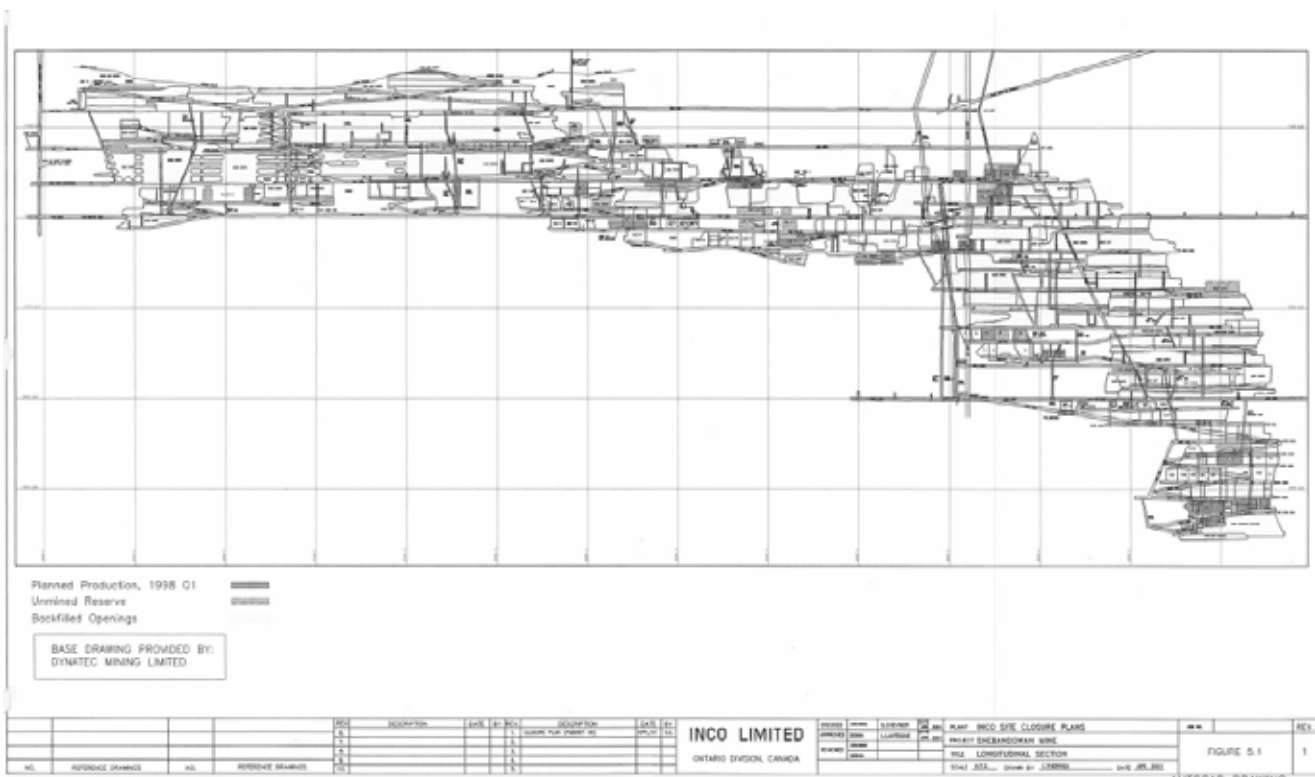


Figure 3. Shebandowan Mine longitudinal section from Closure Plan (Inco, 2001).

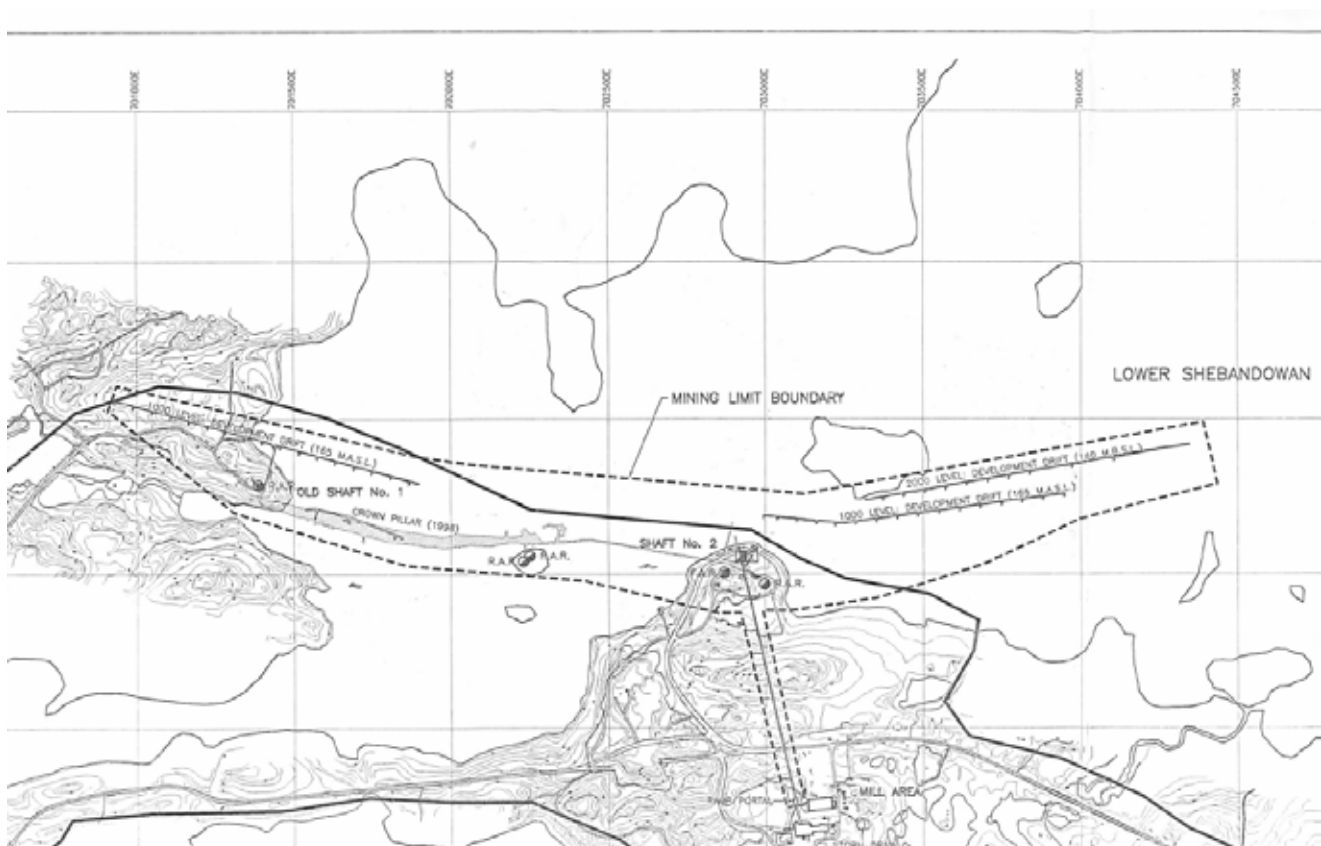


Figure 4. Plan view showing area of underground workings (Inco, 2001).

stoping, with the mine workings being backfilled with a cemented mixture of 60% tailings and 40% alluvial sand (30:1 ratio of backfill to Portland cement). The minimum crown pillar thickness beneath the lake is 33 metres and an engineering study has indicated that there are not likely to be any long-term rock stability issues.

Stop 2: Shebandowan West Prospect

UTM coordinates NAD 83; 15U 0700710E/ 5386900N

The Shebandowan West prospect consists of three shallow Ni-Cu mineralized zones (West, Road and D Zones) located immediately west and along strike with the Shebandowan Mine ore body. North American Palladium (NAP) <http://www.napalladium.com/English/projects/reserves-and-resources/default.aspx> reports measured and indicated resources of 1,292,000 tonnes grading 0.91% Ni, 0.62% Cu, 1.09 g/t Pd, 0.34 g/t Pt, and 0.23 g/t Au. During 2008, a ramp was advanced to collect a bulk sample for metallurgical testing. After the project was suspended, the portal and vent raise were backfilled to prevent inadvertent access to the underground workings. These are considered to be temporary measures that would need to be upgraded and certified if the proponent were to decide to permanently close-out the site.

To the west of the portal, NAP constructed an engineered containment pond to collect and retain water pumped from the underground workings. The pond is lined with high density polyethylene (HDPE) to prevent seepage. Because the ore is considered to



Photo 2. Ore pad and containment pond at Shebandowan West project site

have the potential to generate acid, a lined ore pad was also constructed adjacent to the pond in order to ensure that any contaminated runoff was not discharged to the natural environment. As per the requirements of the Closure Plan (North American Palladium, 2008), all ore was shipped off site for processing before activities at the site were suspended.

Stop 3: No. 2 Shaft Area

UTM coordinates NAD 83; 15U 0702850E / 5386235N

The clearing at this stop was previously the location of the Shebandowan Mine headframe and hoist structures. These were demolished and removed from the site in 2001. Similar to the No.1 shaft, the No. 2 shaft was subsequently capped with reinforced concrete.

The Mine Rehabilitation Code of Ontario requires that disturbed areas of mine sites be revegetated to stabilize surface soils, improve aesthetics, establish sustainable vegetation growth and support the



Photo 3. View of No. 2 Shaft area. Concrete caps and pump house are visible in centre of photo.

designated end use of the site. The No. 2 shaft area was contoured and seeded after the completion of demolition activities. Vegetation growth has been successful, with some native species (e.g., trees) already beginning to colonize the area.

The last remaining original building on the site is the former process water pumphouse located adjacent to the lake. This pumping equipment now serves as a source of water that is used during periods of low precipitation to ensure that the tailings pond remains saturated.

Stop 4: Mill Area

UTM coordinates NAD 83; 15U 0703230E / 5385400N

All buildings associated with the former mine/mill complex were demolished and removed from the site in 2003. Similar to the headframe area, the mine/mill complex area was seeded, and vegetation growth has been occurring. Revegetation efforts in this area have also included the planting of trees.

An acid-generating waste rock pile was previously located in the low-lying area at the southwest corner of the clearing. This waste rock was relocated to the tailings pond, and is now located under water. Residual groundwater quality impacts continue to be monitored in groundwater monitoring wells that have been installed to the south of the former waste rock area.

A storm water collection pond is located at the east end of the mill area clearing. Water collected in this pond contains elevated concentrations of metals



Photo 5. Storm water collection pond. Pump barge can be seen in centre of pond. Water is pumped to tailings pond at Dam No. 4.

(most notably nickel) leached from soils in the mill area. Groundwater in the vicinity of the pond is also impacted by residual petroleum hydrocarbons from a fuel spill that occurred in its vicinity during mine operations. Water from the storm pond continues to be pumped to the tailings pond for treatment. This will continue until the water quality meets regulatory requirements for direct discharge to the environment. Groundwater quality in the area downgradient of the storm pond also continues to be monitored for acid rock drainage (ARD) and petroleum hydrocarbon impacts. Groundwater monitoring must continue until geochemical stability has been demonstrated and there are no longer any significant risks of adverse impact to downgradient receiving water bodies.

Stop 5: Tailings Dam No. 4 Seepage Collection Pond

UTM coordinates NAD 83; 707780E, 5384550N

The Shebandowan Mine tailings were deposited in a 115 ha impoundment located approximately 1.5 km southeast of the mine/mill complex. The tailings contain a significant proportion of sulphide minerals and are considered to be acid-generating. Data presented in the 1996 Closure Plan (Inco, 1996) indicated that un-oxidized tailings contain up to 13.2% sulphur and have NP/AP (neutralization/acid generation potential) ratios of approximately 0.16. Values less than 1 indicate that a material is acid generating. As a result, the tailings impoundment closure design includes a permanent water cover to prevent sulphide oxidation and acid production. The tailings area is contained in a basin



Photo 4. View of rehabilitated mill area



Photo 6. Shebandowan Mine tailings pond. Dam No. 4 is located to right hand side of pond. Splitter dyke can be seen at centre of photo crossing pond. This dyke limits wave development in pond.

that is bounded by natural topography (bedrock outcrop areas) and six engineered dam structures.

Although the tailings dams are designed to retain water, some seepage occurs. Seepage that collects in a pond at the toe of Dam No. 4 contains elevated concentrations of iron and nickel. As a result, this seepage is pumped back into the tailings pond for treatment. Groundwater monitoring is carried out downgradient of all of the tailings dams to monitor the groundwater quality impacts resulting from seepage. Although there are elevated concentrations of iron, manganese and sulphate in these seepage plumes, there is no evidence of acidic drainage (Wesa, 2009).



Photo 7. Dam No. 4 seepage collection pond and pump house



Photo 8. Tailings pond spillway

Discharges from the tailings basin to the natural environment are regulated by the Ontario Ministry of the Environment under an Industrial Sewage Works approval. Since mine closure, the water quality in the tailings pond has improved to the point where no active treatment is required to meet the applicable effluent limits. However, monitoring will continue to be required until water quality meets the more stringent Provincial Water Quality Objectives.

Stop 6: Tailings Pond Spillway

UTM coordinates NAD 83; 0706315E / 5384850N

Discharges from the tailings pond occur through a spillway located at the east end of the impoundment. The spillway is excavated through solid bedrock near the north abutment of Dam No. 2. Discharge from the tailings pond is intermittent, and only occurs following times of significant precipitation or snow melt. When discharge is occurring, water quality monitoring must be carried out to demonstrate that the water quality meets effluent limits. The receiving water body is Gold Creek, which is part of the Matawin River watershed, which ultimately flows to Lake Superior.

North Coldstream Mine

Exploration and Development History

The following summary of the historical exploration and development of the North Coldstream Mine property is based on information compiled from the files of the MNDM Mines and Minerals Division office in Thunder Bay.

1870s: Copper mineralization discovered.

1951-1957: Coldstream Copper Mines Ltd. carried out exploration and development program.

1957: Production commenced.

1958: Operations suspended.

1959: Company re-organized and name changed to North Coldstream Mines Ltd.

1960-1967: Mine produced approximately 2.5 million tons of ore grading 1.97% copper, 0.012 ounces per ton gold and 0.22 ounces per ton silver.

1968: Mill and associated infrastructure and surface rights sold to Nelson Machinery.

1971: North Coldstream Mines changes name to Coldstream Mines.

1976: Coldstream goes into receivership.

1977: International Mogul acquired mineral rights.

1982: Conwest becomes mineral rights owner through

corporate amalgamation.

1991: Nelson Machinery placed into receivership.

1992-1995: MNMD ordered Nelson Machinery to submit a Closure Plan for mill site infrastructure, and Conwest to submit Closure Plan for tailings, mill yard. Subject to appeals, Mining Commissioner ruled that Nelson was responsible for rehabilitation of mill area, Conwest responsible for tailings and mine openings.

1996: Conwest acquired by Alberta Energy Company, re-named AEC West. Mineral rights later sold subject to an agreement that AEC West would retain responsibility for rehabilitation of tailings and mine workings. AEC West was subsequently re-named EnCana West and EWL Management (current corporate identity).

1998-2000: Tailings areas rehabilitated.

2000: MNMD rehabilitated mill site through Abandoned Mines Rehabilitation Fund.



Figure 5. North Coldstream Mine field trip stop locations.

2000-2002: Several mine openings to surface rehabilitated, fencing erected around open stopes.

2008-2012: Crown pillar stability investigations, additional tailings rehabilitation activities, work towards development of long-term monitoring program.

Deposit Geology

The North Coldstream deposit is located within an inferred s-folded metavolcanic and gabbroic rock sequence (Osmani, 1997). The gabbro intrudes along the contact between felsic and mafic metavolcanic rocks. The northeast-trending Burchell Lake Fault is located immediately west of the site.

The North Coldstream Mine ore body is a 120 x 300 m silicified zone located at the contact between the gabbro and mafic metavolcanic rocks. Osmani (1997) has interpreted the mineralized zone as silicified gabbro. The ore zone consists of a high density network of chalcopyrite and pyrite veinlets, and massive and disseminated mineralization within a siliceous host rock that resembles chert.

Field trip stops

Stop 7: TMA-1 Tailings Area

UTM coordinates NAD 83; 0678515E, 5386325N

Tailings Management Area 1 (TMA-1) was used for the deposition of North Coldstream Mine tailings until 1962 (Burns et al., 1999) and was the largest of two tailings deposition areas that were used during mine operations. The ore contained approximately 2 to 8%



Photo 10. TMA-1 tailings area in 1991.

chalcopyrite and roughly similar quantities of pyrite. Tailings sampling carried out by CANMET in 1994 indicated that the average sulphur content was 5.7%, and that they are acid generating (Burns et al., 1999). The majority of TMA-1 is located over permeable soil (sand and gravel) with a deep water table. This hydrogeologic setting has resulted in the development of a significant ARD plume in the groundwater immediately below and downgradient of the tailings. This contaminant plume has low pH and contains extremely high concentrations of sulphate and metals (e.g., Fe, Cu, Co, Mn, and Ni).

Groundwater from TMA-1 migrates in a westerly direction toward the Wawiag River, which is known to be a location of groundwater discharge from the overburden aquifer. The overburden aquifer is located in a west-trending bedrock depression that controls the ARD plume flow direction. The core of the ARD plume sinks toward the bottom of the aquifer between TMA-1 and the river. Prior to reaching the river, the



Photo 9. Aerial view of mine/mill complex and southwest end of TMA-1 tailings area in 1991



Photo 11. Embankment on west side of TMA-1 in 1998.



Figure 6. Approximate groundwater flow path from TMA-1.

deepest portions of the plume become confined below a silt confining layer. As a result, the most severely impacted groundwater does not discharge to the river. Nevertheless, measurable water quality impacts attributable to TMA-1 do occur in the Wawiag River (i.e., elevated levels of Fe and Co), especially during periods of low flow. Deep ARD-impacted groundwater beneath the confining layer changes flow direction and migrates in a southwest direction through another overburden-filled bedrock depression that parallels the Wawiag River toward Burchell Lake. The ultimate location where the deep ARD plume is believed to discharge to Burchell Lake is approximately 1.5 km offshore at a depth of 40 to 70 m (Golder Associates, 2011). To date, no significant impacts to the lake have

been documented as a result of the TMA-1 tailings plume.

In 1998-1999, a vegetated low permeability cover with a capillary break was placed over the TMA-1 tailings in an effort to reduce groundwater quality impacts. This work has resulted in reduced concentrations of sulphate and metals in the tailings impact plume. However, portions of the plume remain acidic and it is expected to take decades for the acid rock drainage plume to be fully rehabilitated. Between 2007 and 2010, EWL Management carried out several environmental and geochemical investigations to better characterize site conditions. Some of the key findings of this work are listed below.

- Shallow groundwater in the northern half of TMA-1 has neutral pH and lower sulphate and metal concentrations than in the southern half. As a result, the northern portion of the TMA-1 cover appears to be functioning as expected.
- September 2010 data for MW38B: pH = 7.4, sulphate = 1440 mg/L, Fe = 54 mg/L, Mn = 2.3 mg/L, Cu = <0.01 mg/L, Co = 0.0005 mg/L (Golder Associates 2010).
- Acidic groundwater continues to be present in the southern half of TMA-1, indicating that this portion of the cover was not performing as expected.
- September 2010 data for MW35B: pH = 3.3, sulphate = 3390 mg/L, Fe = 1000 mg/L, Mn = 11 mg/L, Cu = 3.6 mg/L, Co = 4.7 mg/L (Golder



Photo 12. Vegetated TMA-1 cover as it appeared in the summer of 2005.



Photo 13. Photograph taken in October 2011 during placement of cover over relocated tailings at TMA-1. Clay layer is visible to left, with overlying granular cover layer to right

Associates 2010)

- Significant aquifer recharge was occurring at the southeast end of TMA-1 immediately following major rainfall events. This is likely to have been responsible for the continued ARD generation in the south half of TMA-1.
- Storm drainage in the eastern perimeter spillway was continuing to show signs of acid generation.
- Two previously unidentified “orphan” tailings areas were found north and west of TMA-1. These were most likely related to mobilization from TMA-1 during historic storm events.
- Two small tailings areas located to the south of TMA-1 were identified as ongoing sources of ARD that required additional rehabilitation.



Photo 14. Reconstructed eastern perimeter spillway. East side of tailings relocation area can be seen to left.

During 2011, additional rehabilitation work was done on the site to address the on-going ARD issues. This work included the relocation of the orphan and southern tailings areas to TMA-1 and the reconstruction of the eastern perimeter spillway. The relocated tailings were placed over the eastern half of TMA-1 during the winter of 2011, and a low permeability clay cover was placed over them during the summer and fall. The eastern perimeter spillway was also reconstructed in 2011 to more effectively convey storm drainage around TMA-1. A key design element was the installation of a geosynthetic clay layer on the western side of the spillway to isolate drainage from the TMA-1 tailings. It is expected that this additional rehabilitation work will significantly improve storm drainage quality, reduce infiltration at the southeast end of TMA-1, and reduce contaminant loadings to the overburden aquifer.

Stop 8: TMA-2 Tailings Area

UTM coordinates NAD 83; 0679185E / 5386600N

From 1962 to the end of the mine life in 1967, approximately 500,000 tonnes of North Coldstream Mine tailings were deposited in Halet Lake, which is now known as TMA-2 (Burns et al., 1999). Prior to 1998, a 3 ha tailings beach was located at the former tailings discharge location at the southwest end of the lake. These tailings were generating acid and contributing metal loadings to the tailings pond and downstream receiving water bodies. The TMA-2 outlet drains north to Background Lake, which subsequently drains toward the Wawiag River and Burchell Lake.

During the summer of 1998, the majority of the TMA-2 tailings beach was relocated below water, with approximately 4,000 tonnes relocated to TMA-1 (Burns et al., 1999). The goal of the tailings relocation was to completely submerge the tailings and prevent further oxidation and acid generation. Suspended sediment, acidity and metals were released into the TMA-2 pond during tailings relocation. A temporary dam was constructed to prevent downstream discharge, and the pond was limed to reduce acidity and metal concentrations. Since completion of the tailings rehabilitation, water quality has stabilized, with neutral pH and low metal concentrations in the TMA-2 discharge (although copper and cobalt concentrations do slightly exceed Provincial Water Quality Objectives).

Following relocation, some of the tailings along the west margin of TMA-2 remain less than 2 m below the water surface. Tailings in the central basin are much



Photo 15. View of TMA-2 (Halet Lake) from former tailings beach area.

farther below surface, ranging in depth from 7 to 12 m (Golder Associates, 2011). Because some tailings are relatively close to surface, it is important that the water elevation is maintained at a level that maintains permanent saturation. Prior to 2008, water levels occasionally dropped during dry years to expose some of the tailings. However, water levels rose substantially

during 2008 due to the construction of a beaver dam on the outlet stream that regulated the water level and kept it well above the tailings. During the winter of 2012, EWL Management planned to construct an engineered water level control structure on the outlet stream in order to permanently maintain higher water levels in TMA-2.

A long-term water quality monitoring program will be carried out to monitor the success of the tailings rehabilitation efforts at the North Coldstream Mine. This monitoring program will include the sampling of surface water and groundwater monitoring wells.

Stop 9: Mine/Mill Area

UTM coordinates NAD 83; 0678120E / 5386040N

All surface structures on the mine/mill site were demolished and removed in 2000. By this time, the buildings had deteriorated to the point where they had become significant safety hazards. This work was performed by the Ministry of Northern Development and Mines utilizing the Abandoned Mines Fund.

Since the completion of the building demolition

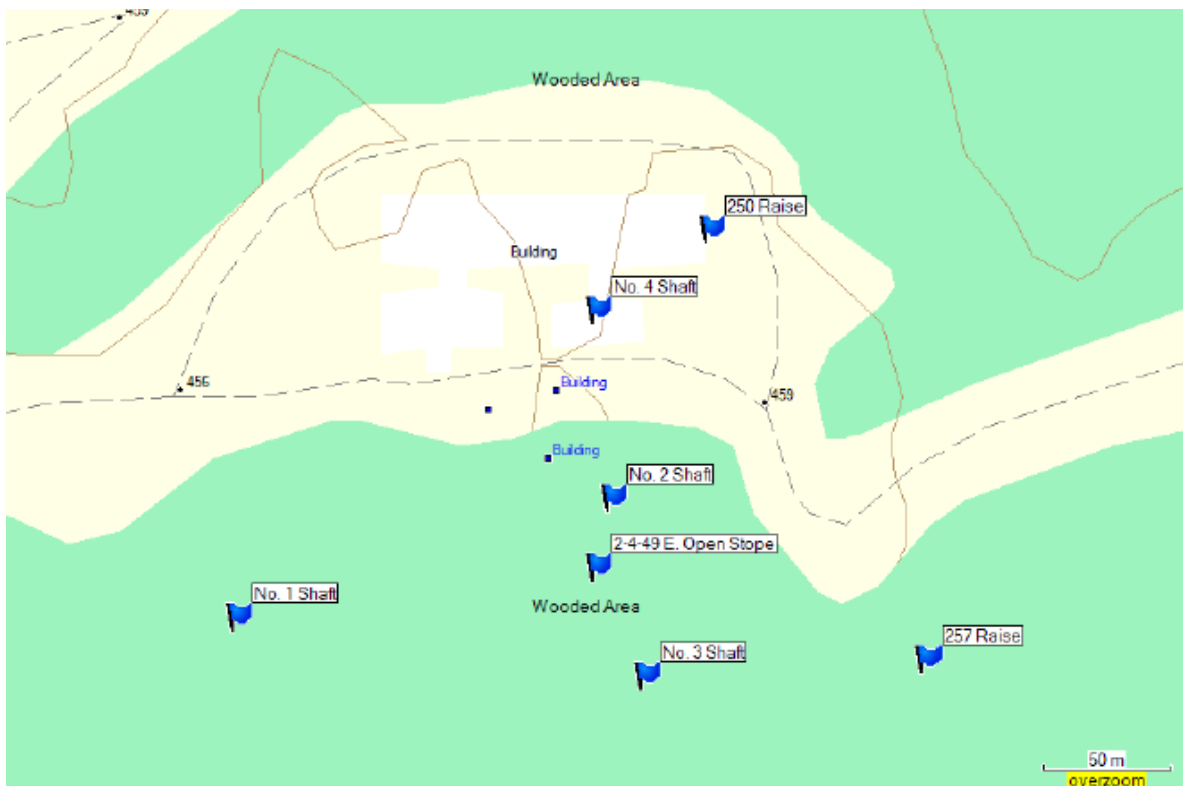


Figure 7. Map of North Coldstream mine/mill area showing locations of mine openings to surface. The locations of the former buildings are also shown.



Photo 16. Glory Hole

project, EWL Management and its predecessor companies have been working toward completing the rehabilitation of the mine openings to surface and the underground mine workings. By 2002, most of the shafts and raises had been permanently rehabilitated (Golder Associates, 2002). Certified reinforced concrete caps were constructed over the Nos. 3 and 4 shafts, and the 250 and 257 vent raises. The No. 2 shaft was backfilled to surface and its long-term stability has been certified. Areas with open stopes and/or unstable crown pillars have been fenced. These include areas near the No. 1 shaft, and above the 2-4-49 W and 2-4-49 E stopes. An open stope (known as the Glory Hole) is present in the fenced area around the 2-4-49 E stope.

EWL Management has carried out a crown pillar



Photo 17. No. 4 Shaft and 250 Vent Raise are located below fenced vent pipes. Fencing was installed to protect against vandalism.

investigation and is currently developing a final rehabilitation plan for all potentially unstable crown pillar areas. It is likely that this plan will include additional fencing and a rock stability monitoring program.

Stop 10: Burchell Lake “Ghost Town”

UTM coordinates NAD 83; 0677550E / 5386090N

This is a general interest stop to view the remains of the former town site of Burchell Lake. Although a number of larger buildings were removed following mine closure, many abandoned houses remain. Mine management homes were located in a separate development located further to the south. These buildings continue to be used as seasonal cottages.



Photo 18. Abandoned houses in former town site of Burchell Lake.

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Field trip 10 - Geoarchaeology of the Thunder Bay area

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Objectives

The field trip focuses on the deglaciation and lake level history of Thunder Bay and the immediately surrounding area (Fig. 1). In particular, we will examine evidence of Palaeo-Indian occupation along abandoned shorelines, river mouths and deltas of the Lake Minong stage of Lake Superior (circa 9.5 ka BP.). The trip includes a visit to the former mouth of the Current River at the north end of the city, where

the Simmonds and McDaid sites are located. We then travel east along Highway 11/17 to view several Palaeo-Indian sites currently undergoing salvage excavation along the path of highway development. The tour then returns to Thunder Bay to visit the Cummins site, and also the nearby Neebing R. sites. Next we visit the Rosslyn delta, on the Kaministiquia River, the western extent of Lake Minong. From there we will travel west to Kakabeka Falls, where earlier Lake Beaver Bay features will be examined. Finally, the trip will



Figure 1 Thunder Bay area orientation map.

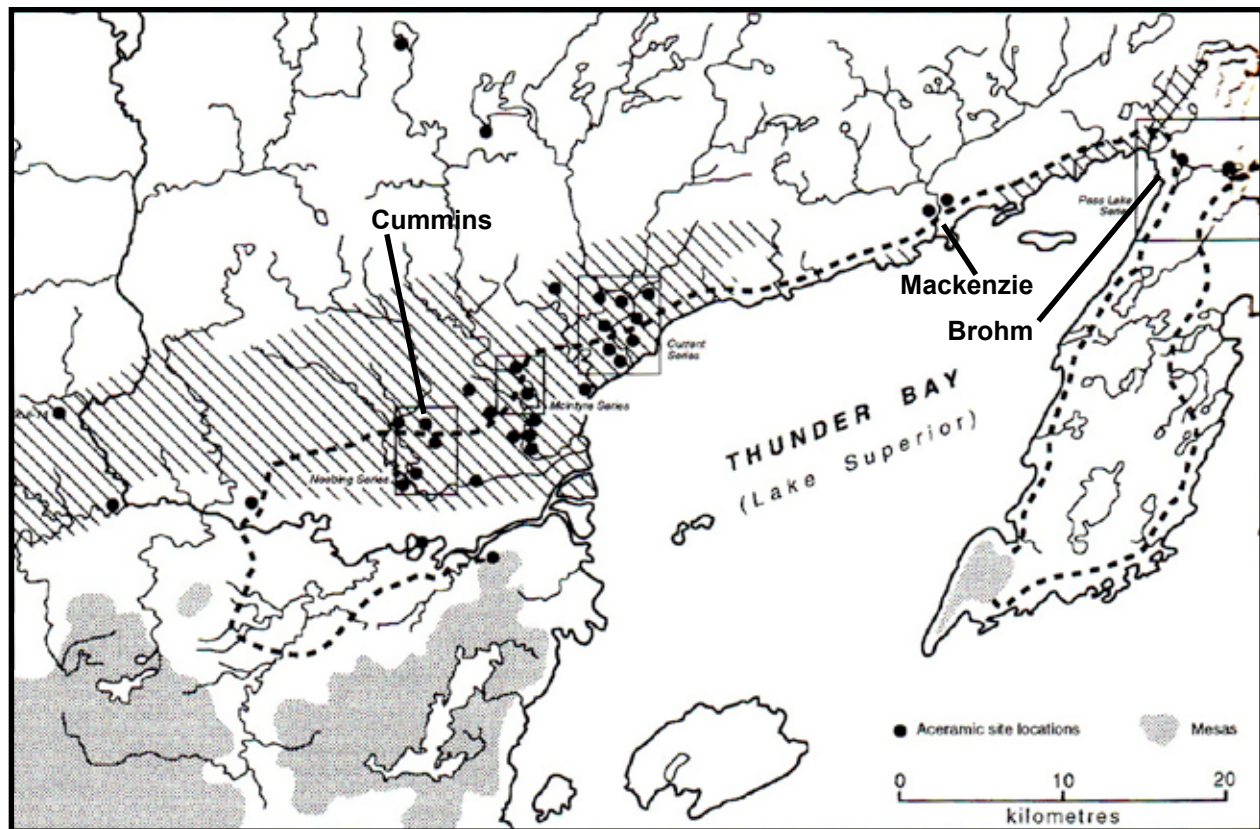


Figure 2 The spatial relationship of probable Plano and Archaic sites with Lake Minong shorelines (dashed line) and exposures of tool stone deriving from Gunflint Formation (hatched lines). After Hinshelwood (2004:234).

mount the Marks moraine, providing a view that will place the day's observations in context with pre and post Marquette ice marginal events and Palaeo-Indian presence.

Introduction

In the Thunder Bay area there is a strong, though not exclusive, relationship between Palaeo-Indian habitation, bedrock exposures of the favoured tool stone within the Gunflint Formation, and the abandoned shores of post-glacial lakes of the Superior basin (Fig. 2). Of particular importance are shores of Lake Minong, established about 9.5 ka B.P. This spatial relationship is perhaps illusory since it is clear that these people occupied a number of habitats, some far from ancient lakeshores, but in geomorphological mapping of these shorelines (often in the context of urban development), it is the lakeshore sites that have been most commonly found and reported.

Evidence accumulated since MacNeish's (1952) excavations at the Brohm Site (Pass Lake) on the Sibley Peninsula (Fig. 2) by archaeologists (Fox, 1975, 1980; Dawson, 1983b; Julig, 1984; Ross, 1997; Hinshelwood

2004) suggest that Palaeo-Indians migrated northeast from the Dakotas and Minnesota, and into the cul-de-sac formed between Lake Agassiz, the lakes of the Superior basin, and the retreating margin of the Laurentide ice sheet (Fig. 3). Lake Agassiz covered most of Manitoba and parts of Northwestern Ontario at its maximum extent, and contributed significantly to the complex hydrological sequence affecting the Lake Superior basin (Fig. 4). As such Lake Agassiz played an important role in the initial peopling of the greater part of northwestern Ontario since its spatial expanse shifted north over time with glacial retreat, and conditioning when land would have been available for northward human occupation (Fig. 5).

While some fluted projectile points (i.e. early Palaeo-Indian Clovis; Fig. 6) have been encountered as far north as central and perhaps northern Minnesota, none have been found yet in northwestern Ontario. Perhaps Clovis technology disappeared before the ice had retreated from the region, or that insufficient research has been done in the rugged uplands of the cul-de-sac that was first deglaciated (and therefore the most likely zone where such finds might be made).

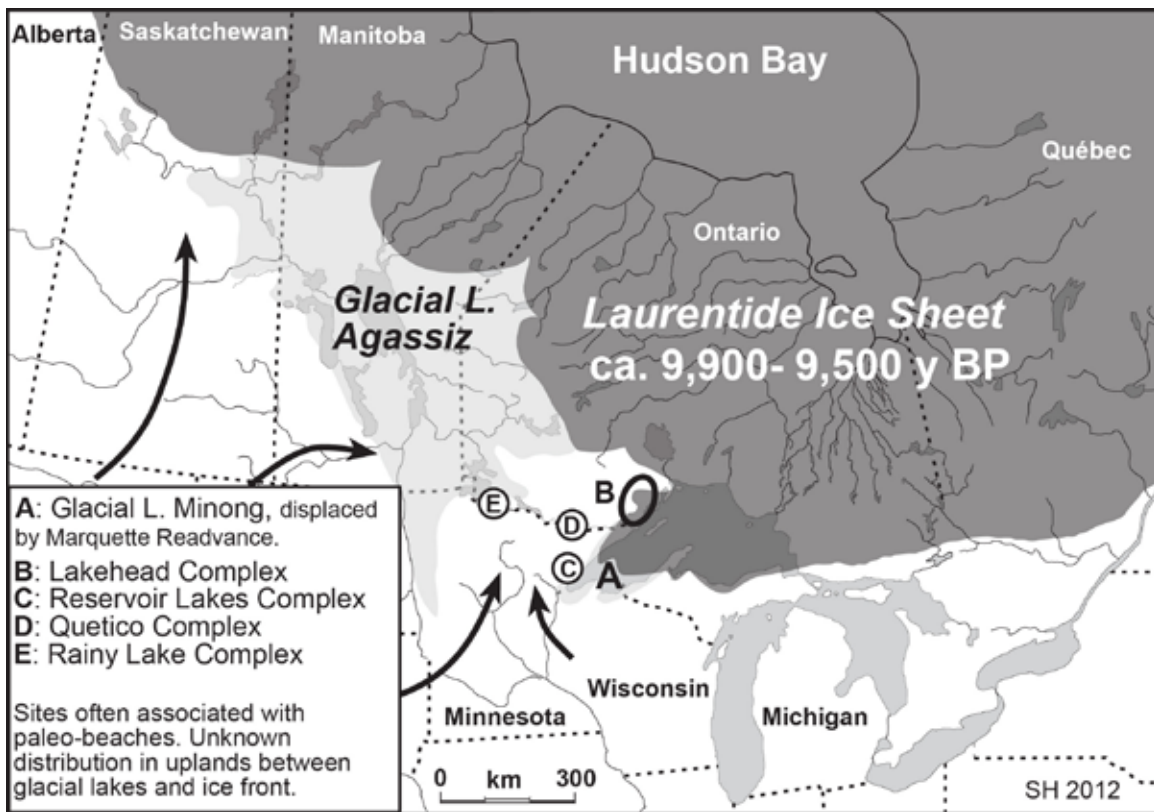


Figure 3 Proposed late Palaeo-Indian migration into the cul-de-sac formed between Lakes Agassiz and Minong and the Laurentide Ice Sheet (after Hamilton and Ross, 1997).

In any case, the Thunder Bay area has yielded unfluted lanceolate projectile points, probably deriving from several late Palaeo-Indian cultures (often collectively referred to as Plano) (Fig. 6). These represent hunting groups who pursued game throughout the cul-de-sac at some time after the Marquette readvance (ca. 9,900 to 9,500 y BP) (Fig. 3). While a range of projectile point types have been recovered, Ross (1997) has proposed that these sites contribute to the “Interlakes Composite”- consisting of a series of inter-related local populations who jointly utilized the deglaciated landscape at some point after ca. 9,500 years ago. This material culture is characterized in part by parallel-oblique flaked projectile points (Fig. 6a), and heavy use of siliceous stone deriving from the Gunflint Formation (i.e., Jasper Taconite, Gunflint Silica) that outcrops in the Thunder Bay area (Fig. 2). Other important raw materials include Knife Lake Siltstone, deriving from bedrock sources near the Minnesota/Ontario border, and a sparse array of non-local materials (including Hixton Silicified Sandstone from central Wisconsin and perhaps also Chalcedony from western North Dakota).

Local Palaeo-Indian sites exhibit a strong preference for bedrock lithic sources, with only a minor percentage

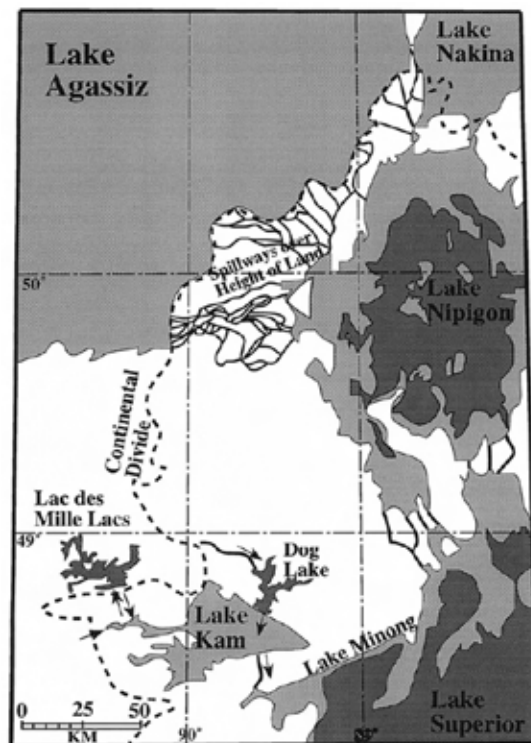


Figure 4 Repeated spills of water from Lake Agassiz into Lake Minong (Hamilton 1996:30 after Teller and Thorleifson 1983).

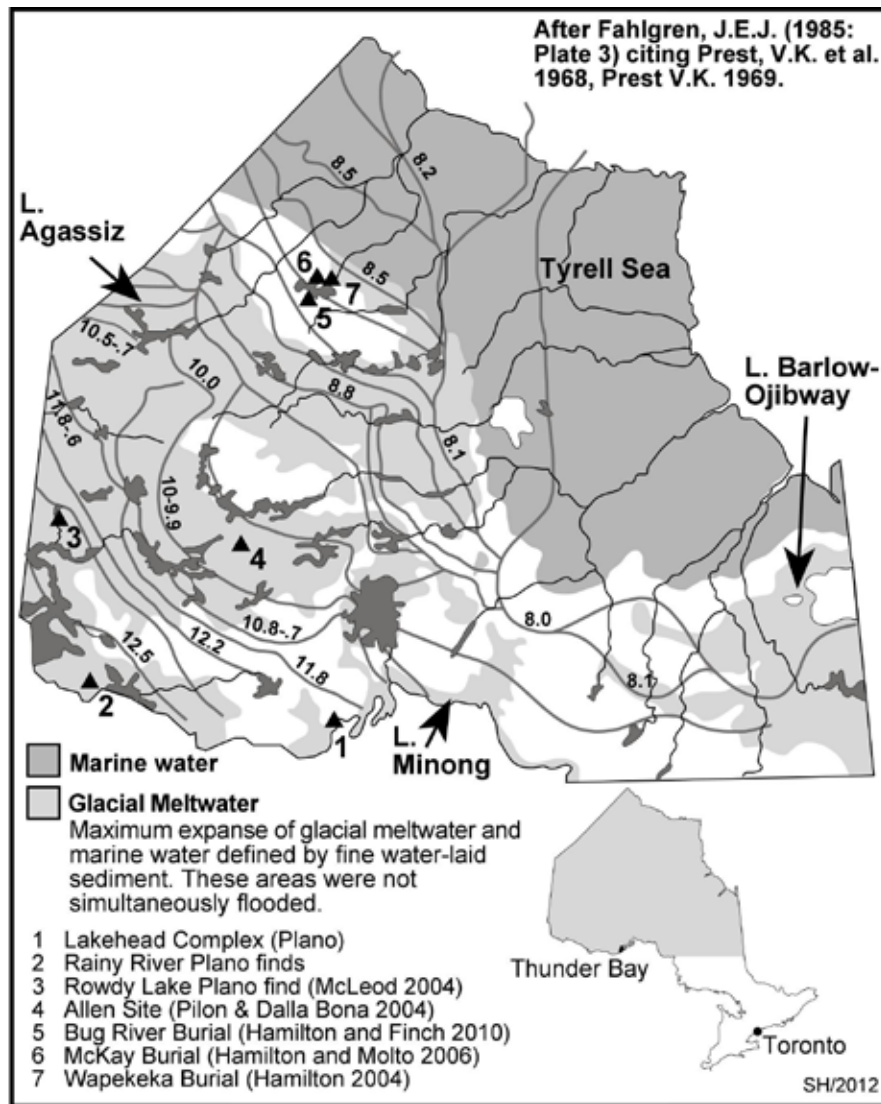


Figure 5 Northeastward retreat of the Laurentide glacier, with related northward shift of Lake Agassiz waters. A sparse array of early and middle Holocene archaeological sites suggest the time-transgressive northward migration of human populations (Hamilton nd).

of the raw materials suggesting use of cobble and pebble sources. This forms a sharp contrast to other cultures in the region dating to the Middle and Late Holocene. This resulted in some repeatedly used archaeological sites where exposures of suitable stone coincide with the ancient beaches of Lake Minong. The famous Cummins Site represents one such quarry/workshop that has yielded thousands of discarded flakes, blocks, preforms and other debris from tool/preform fabrication, but with a very low relative frequency of formal or informal tools. Other sites, that likely served as short-term camps, seasonal aggregation places, hunting/ambush sites, observation points, and other functions also dot the landscape. While they also yield much discarded stone, a somewhat higher relative frequency of lost, broken or discarded tools are

recovered, suggesting more generalized site functions. While the shorelines were not the only areas utilized on the early Holocene landscape, it is clear that they were important (likely used for seasonal ingathering), no doubt because of the spatial convergence of valued resources. Some of this site function variability is addressed at the various sites visited during the tour.

Deglaciation History

The broad details of the deglaciation of the Superior basin are shown in Figure 7. As ice of the Rainy River and Superior lobes withdrew from central Minnesota, a series of recessional moraines were left in place. The Vermilion moraine, trending northwest to southeast, across northeastern Minnesota, was followed by

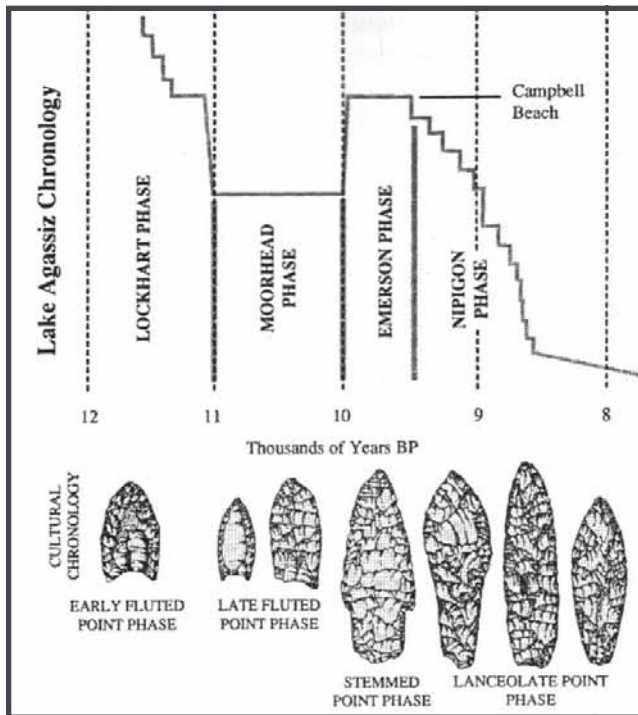


Figure 6 Pettipas' (2011) now-obsolete Lake Agassiz temporal sequence with proposed relationship to the Paleo-Indian cultural historical sequence. He revised the lake sequence in light of new data published by Leverington and Teller (2003) and Fisher (2005, 2008). We include it here because it offers a sense of archaeological conventional wisdom regarding the cultural sequence. This will soon be updated in light of ongoing research conducted at the Mackenzie I Site that has yielded a very large collection of projectile points. This will allow development of a more regionally relevant typology.

the Steep Rock moraine on the Canadian side of the border, and the Brule moraine (Fig. 7). Guided by these ice marginal positions, Lake Agassiz found an early eastern outlet through the Arrow/Whitefish lakes corridor (circa 11 ka B.P.) and, shortly after, through the Shebandowan lake corridor, ultimately using the Lake Nipigon spillways around 10.4 ka B.P. (Fig. 8a) to enter Early Lake Minong (Teller, 1985) which occupied the Superior basin.

There is no reason to believe that warming climate and biological regeneration upon the uplands unaffected by the Marquette readvance would not have enabled immigration of animal herds and people into the Thunder Bay region at this time. Only tenuous evidence of such early occupation has been reported (Phillips, 1993; Ross, 1994), though recent work in the Arrow/Whitefish corridor has confirmed a number of sites that may represent an earlier Palaeo-Indian presence (McLeod and Phillips, pers. communication,

1999). Unfortunately these surface finds have not yet been subjected to absolute dating.

A major event in the deglaciation history of the Lake Superior basin was the Marquette readvance (9.9 ka B.P.), during which the basin and its margins were briefly reoccupied by ice (Fig. 8b). Recessional moraines northeast of the Brule moraine were destroyed and more recent features (i.e., Marks Moraine, etc) formed in their place. This event also has considerable archaeological implications, since any evidence of occupation in the path of the Marquette readvance was likely buried or destroyed by the new ice cover.

Hudson Bay ice pushed southwestward into the Lake Superior basin, to halt on its southern shore (Drexler, Farrand and Hughes, 1983). Only in a small portion of Whitefish Bay near Sault Ste. Marie, in

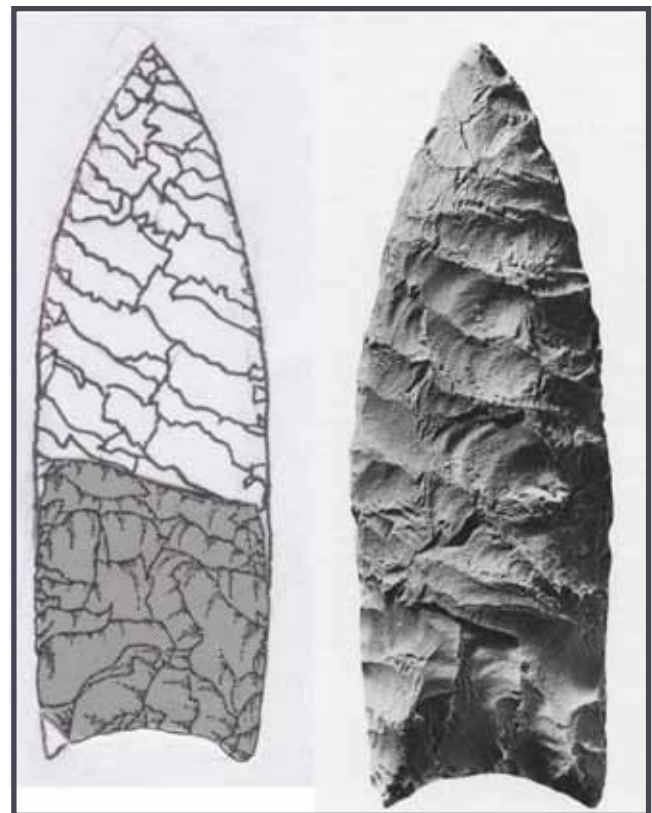


Figure 6a A sketch and photo of lanceolate projectile point style from the Brohm Site. Interlakes Composite sites generally yield a very small assemblage of projectile points reminiscent of a range of late Palaeo-Indian (unfluted) styles. However the Brohm and Mackenzie I Sites have yielded a high percentage of points with lanceolate blade form, basal indentation through repeated flaking, and edge grinding along the lower lateral portions of the blade. Particularly notable is the pattern of oblique parallel flaking that seems to be an important feature of the knapping strategy that appears unrelated to the functional utility of the tools.

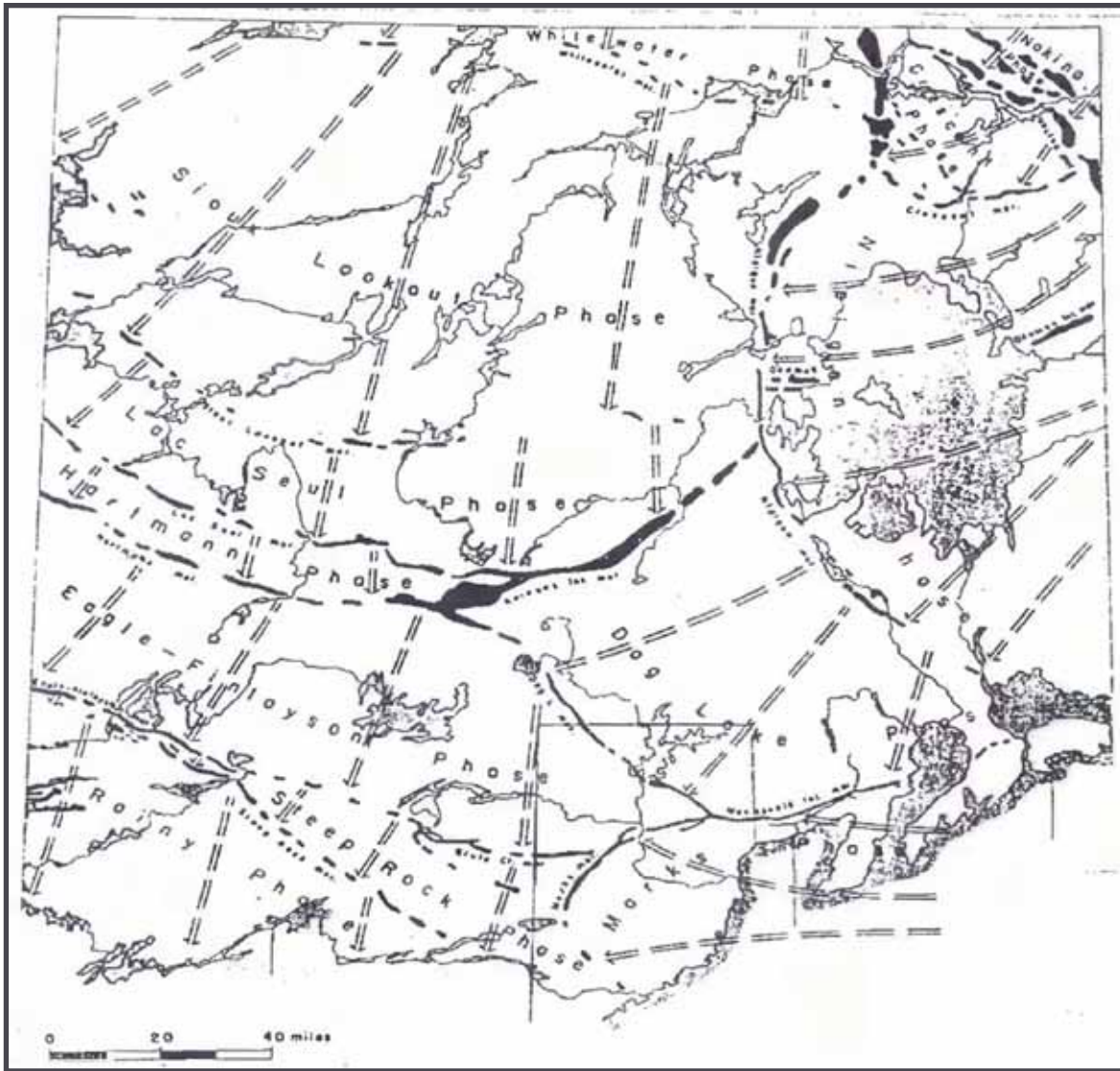


Figure 7 Map of Quetico-Nipigon area showing moraines and direction of ice movement during various phases (from Zoltai, 1965a).

the southeast corner of the basin, did Early Lake Minong remain an open lake (Farrand and Drexler, 1985). The Marquette lobe pushed westward against the steep Minnesota shore and, where less obstructed, flowed northwest up the Kaministiquia valley to the Marks moraine (Figs. 1, 7, 8). This prominent ridge, rising to over 470 m (1550') in places, curves from Lappe through Mokomon and around the northwest of Kakabeka towards the Pigeon River (Zoltai, 1963, 1965a, 1965b; Burwasser, 1977, 1980; Burwasser and Ferguson, 1980). Contemporaneously, to the west of the Lake Superior basin, the Patricia ice lobe pushed towards the Rainy River district and halted to deposit the Dog Lake moraine. This runs northwest from Lappe and holds up Hazlewood, One Island, Hawkeye and Dog lakes. East of Lappe, where the two moraines

meet, a line of glacial debris known as the Mackenzie interlobate moraine can be traced through Pearl and on to the Black Bay peninsula (Figs. 1, 8). A glacial lake, Lake Kaministiquia, was formed between the two ice margins (Teller, 1985) and, at Lappe, a huge sand and gravel delta was built by sediment pouring off the interlobate moraine. This scenario is still the subject of debate, however, and there is some field evidence that the Marks and Dog Lake moraines may be older features that were simply reoccupied by ice of the Marquette advance (Julig, McAndrews and Mahaney, 1990; Tickle, 1996; Noble, pers.comm.1999). Lake Agassiz, deprived of its eastern outlets by the advance of the ice, expanded in area and depth (the Emerson phase), and flooded catastrophically through the Clearwater and Athabasca valleys in Saskatchewan

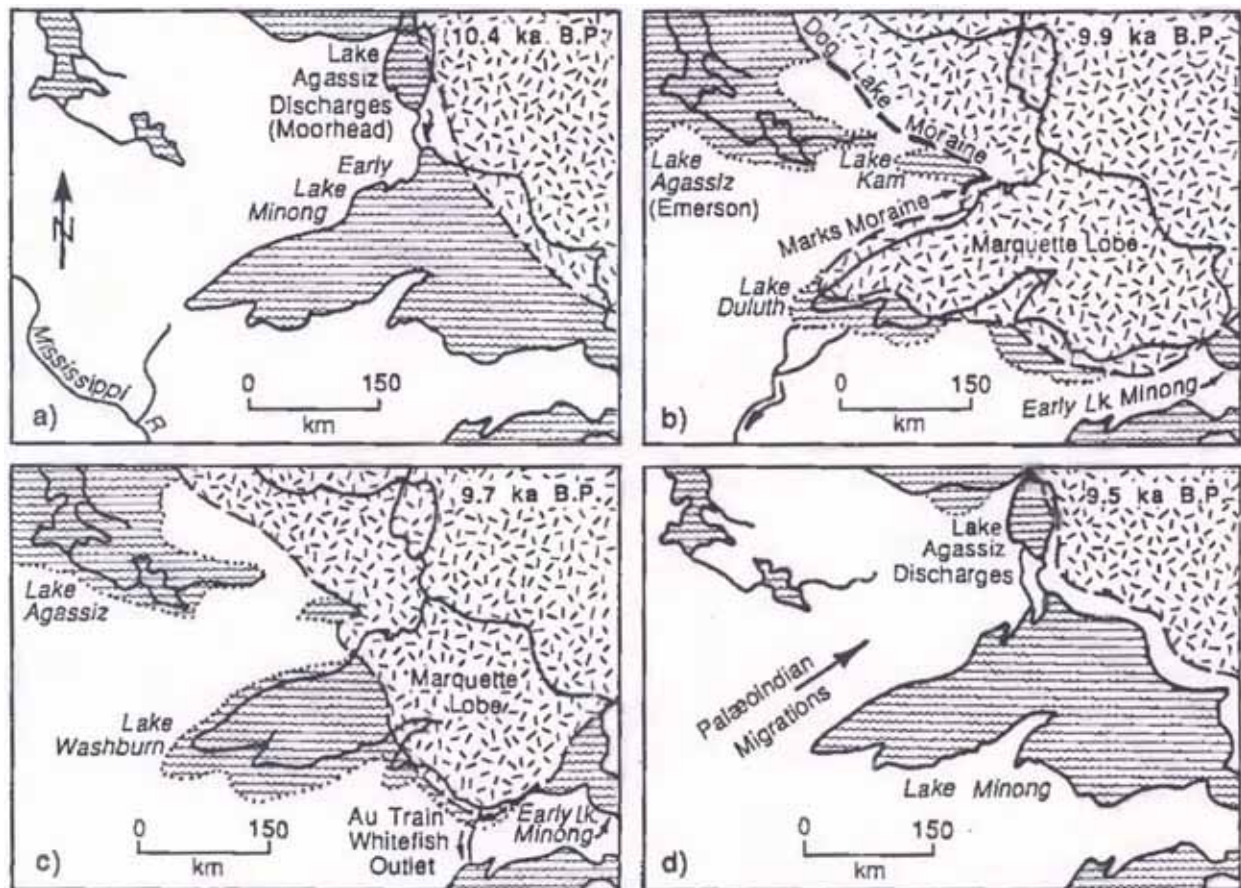


Figure 8 The retreat of the glaciers in the Superior Region between 10,400 and 9,500 years ago (from Phillips, 1993:95).

and Alberta into the Mackenzie River and the Arctic Ocean (Smith and Fisher, 1993).

The Marquette readvance obliterated evidence of the earlier phase of Lake Superior's shoreline history in most of the basin. Shoreline sites that prehistoric people might have occupied in the area of Thunder Bay before 9.9 ka B.P. may lie buried beneath Marquette deposits. In the four hundred years between 9.9 and 9.5 ka B.P., a period about which much more is known, ice withdrew from the Lake Superior basin (Figure 8c/8d). Then, again, Palaeo-Indian people followed game up the Interlakes corridor and into the Thunder Bay region, this time to settle, in part at least, on the shores of Lake Beaver Bay (Stuart, 1993) and Lake Minong (Phillips, 1988).

Shoreline History

At the peak of Wisconsin glacialiation, the northeastern margin of the Lake Superior basin was depressed by the weight of ice (isostatic depression) to a greater degree than the less heavily ice-loaded southwestern side. As a result, "rebound" (isostatic

recovery) since that time has been greater on the north shore of the Lake Superior basin than on the south shore. Lake shorelines which had been originally horizontal became progressively tilted along an approximate northeast axis, such that a shoreline of the same chronological age increases in altitude from southwest to northeast along the western shore of Lake Superior. A theoretical archaeological site at Grand Marais, Minnesota, found at 184 metres, just above the present storm beach, will be of similar age to another theoretical site in Terrace Bay, Ontario, at the 300 metre contour (117 metres above the present lake), with both sites lying on the same tilted shoreline (Fig. 9).

As the Marquette ice lobe wasted back, the eastern and western sides of the basin were exposed as separate entities (Fig. 8c). On the eastern side, a fairly stable Lake Minong, controlled by the height of the St. Mary's River outlet at Sault Ste. Marie, extended northwards up the Ontario shore and westwards along the Michigan shore as they were exposed by melting ice (Farrand and Drexler, 1985). In the enclosed western end of the basin, a distinct succession of ever larger but progressively lower level lakes formed (Lakes Duluth, Highbridge,

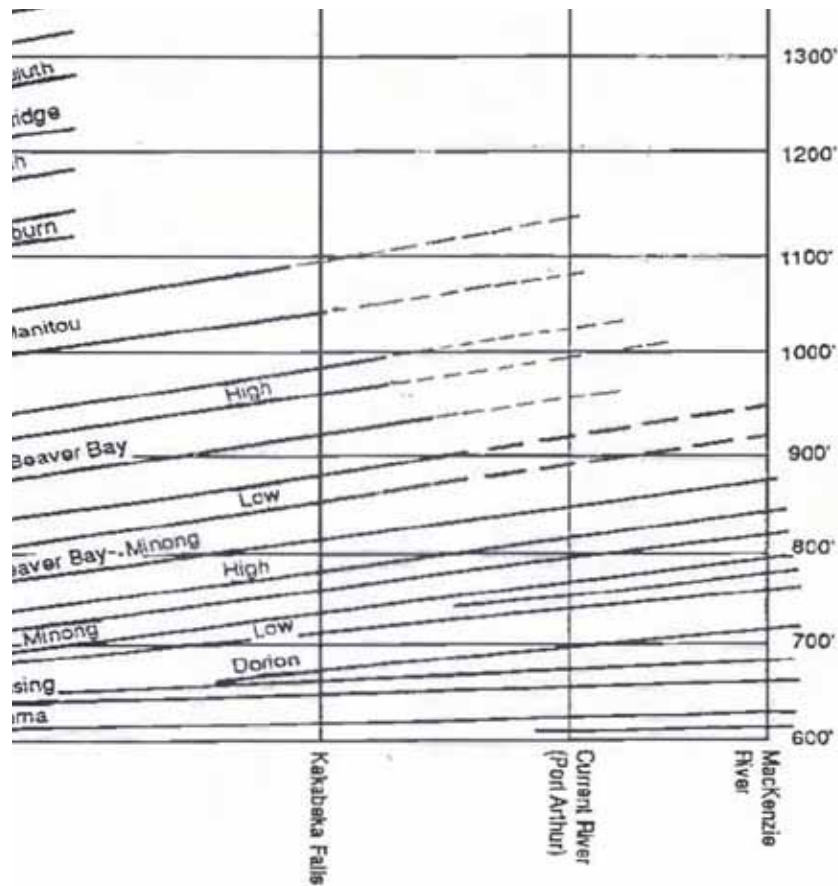


Figure 9 Shoreline diagram for the Pigeon River/Thunder Bay area (Farrand, 1960).

Moquah, Washburn, Manitou, and Beaver Bay), each extending further northeast up the Minnesota - Ontario shore and east along the Wisconsin shore (Farrand and Drexler, 1985). Along the land-based margins of these water bodies were formed various coastal features, such as bluffs and beaches, by which the shorelines can be traced today. As ice vacated the basin, the eastern and western lakes were united and the single shoreline of Lake Minong was formed around the basin about 9.5 ka B.P. (Figure 8d). For the next 1,500 years, water-levels in the Lake Superior basin declined as the St. Mary's sill was eroded down to bedrock. A staircase of Post-Minong shorelines were formed, the last and lowest of which was Lake Houghton, about 8.0 ka B.P.

By 8.0 ka B.P., due to isostatic uplift, the rising levels of Lake Huron had flooded into the St. Mary's River and reversed the flow (Larsen, 1987). This backflooding led to slowly rising water-levels in the Lake Superior basin and culminated in the Nipissing lake stage around 5,000 years B.P. Known as the pre-Nipissing transgression, this rise in water-level was imposed upon a still-tilting basin. On the north shore, east of Dorion, the rate of isostatic uplift remained more rapid than the rising waters of the pre-Nipissing period,

with the result being that the shoreline marking the Nipissing maximum level lies at a lower altitude than all older shorelines, including that of Lake Houghton. On the south shore the pre-Nipissing transgression was more rapid than isostatic recovery, and wave action "inherited" the features of older shorelines, modifying and destroying portions of them, including pre-existing shoreline archaeological sites (Phillips, 1977). This causes a chronological discontinuity which is at its greatest near Duluth and decreases towards Dorion where the Houghton shoreline appears above the present waterlevel (Fig. 10).

In Thunder Bay, evidence of Palaeo-Indian activities would normally have been traced as a continuum from the high Minong shoreline to the Houghton (which lies just below present water-level), but the pre-Nipissing transgression reoccupied the lower and later part of that record and cut a prominent bluff on which the General Hospital, the Court House, St. Joseph's Hospital, and Lakehead University are built. The Nipissing shoreline can be traced up the valley towards Mapleward road and across towards Mount McKay. Thus, the later part of the Palaeo-Indian record, the important transition into the Archaic period, and a good portion of the

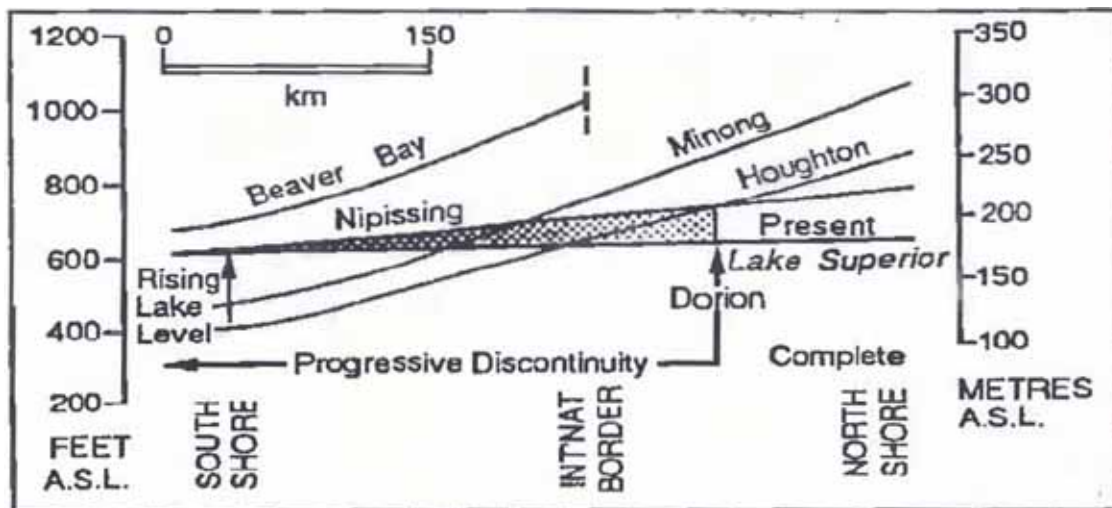


Figure 10 The pre-Nipissing transgression and the resulting loss of potential archaeological sites along the shores of Lake Superior (Phillips, 1993:100).

early Archaic is missing in Thunder Bay. Hinshelwood (2004) offers the observation that these shifting lake levels into the mid-Holocene suggests that some of the much older Plano lithic quarry and workshop sites might have been re-occupied during Archaic times. Several sites that are thought to be of Archaic age have also been encountered along the pre-Nipissing transgression strandline where it cuts through Lakehead University and adjacent properties.

Local History

Figure 11 represents an interpretation of the detailed history of the withdrawal of Superior (Marquette) ice from the Thunder Bay area, based on currently known information, though subject to revision. As ice wasted back from the Marks Moraine, shorelines on the south side of the moraine show that a body of water collected between the moraine and the ice front (Fig. 11a). This proglacial lake has been named Lake Cedar Creek (Jahnke, 1993), and there is tenuous evidence that it connected with high level lakes to the south, perhaps of Lake Duluth equivalence. As ice withdrew from the Kaministiquia embayment a small readvance to the Intola Moraine occurred (Fig. 11b). The higher levels of Lake Beaver Bay have been traced into this moraine (Stuart, 1993), and it is likely that Palaeo-Indian peoples entered the area at about this time, probably using the Marks moraine as a causeway. As ice withdrew further and water level declined to the lower levels of Lake Beaver Bay (Fig. 11c), the possibility of Palaeo-Indian occupation increases, and by the time Lake Minong was established in the Kaministiquia embayment, there is plenty of evidence to prove their presence (Fig.

11d). The shoreline diagram (Figure 9) shows that only as ice withdrew from the Thunder Bay region did the sequence of post-glacial lakes extend into the area. Beaver Bay shorelines can be found at Kakabeka, but only the lower levels extend eastwards through the city towards the Mackenzie River. A few notable sites are:

- 1) the Simmonds and McDaid sites at the mouth of the Current River,
- 2) the Hodder and Naomi Sites overlooking the Current River Mouth,
- 3) a site cluster around a possible Minong embayment near the present Mackenzie River,
- 4) the Biloski Site at the outlet of the McIntyre River into Minong,
- 5) Catherine, Neebing River and Cummins sites along the north side of the Kam embayment,
- 6) the Irene Site on a high bedrock-controlled upland overlooking the Minong shore, and
- 7) a collection of sites on the Rosslyn delta at the mouth of the Kaministiquia River.

While many of these sites are associated with Lake Minong shores, some landscape associations are less simple and deserve more analysis (Hamilton, 1996). Some Palaeo-Indian sites unrelated to shorelines occur at High Falls on the Pigeon river, on Dog Lake (McLeod, 1982), at Harstone Hill, near the junction of the Whitefish and Kaministiquia rivers and near Kakabeka Falls where the Kaministiquia River might have been most logically crossed. The Crane site near Kakabeka, found in a vegetable garden, provides a

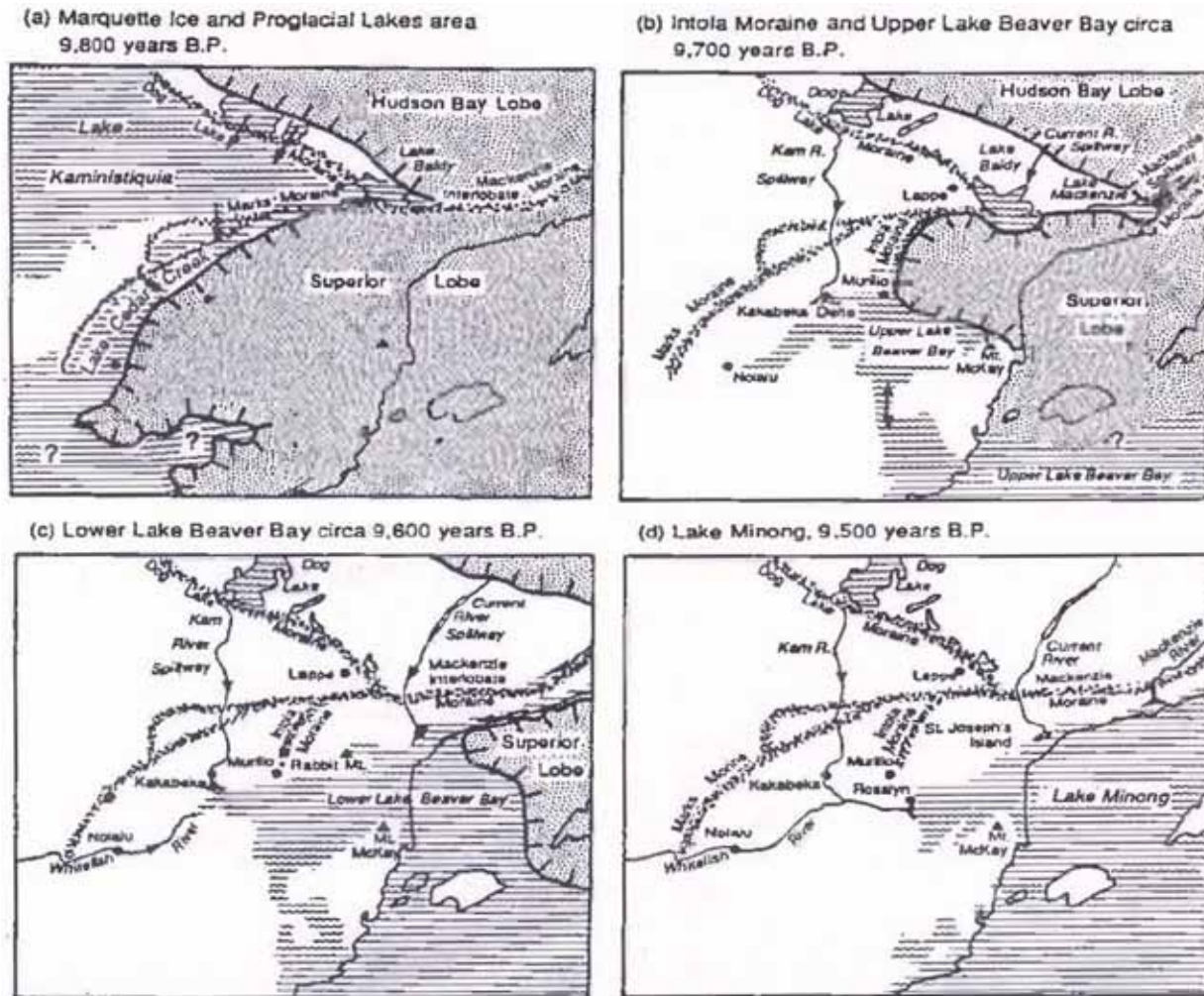


Figure 11 An interpretation of post-Marquette history in the Thunder Bay area (Phillips et al., 1994).

rich cache of beautifully crafted bifaces in a location apparently unrelated to any topographic feature, though on the surface of one of the higher Beaver Bay terraces.

Because of their antiquity and the acidic nature of the Boreal forest soils, only the lithic materials remain to be found at these sites, though it is very likely that many other natural materials would have been used.

Field Excursion Stops

After leaving the hotel, we will travel up Edward/ Golf Links Rd. This route takes us across the lower flats of the Kaministiquia River delta, with the bluff forming the Nipissing Transgression strand line occurring at Stop 1.

Stop 1 Golf Links Road-Thunder Bay Golf and Country Club.

UTM coordinates: NAD83; 16U 0331712E / 5365013N

While we will not leave the van, this location provides a view of the bluffs forming the strand associated with the Pre-Nipissing Transgression. To the left of Golf Links Road on the lower slopes of this bluff a taconite lithic scatter was encountered. Figure 23 provides a view of this beach feature, with several sites within the Lakehead University campus thought to be Archaic age. We will be returning along this route after descending down the Marks Moraine uplands, over the Upper Beaver Bay strand and the Lake Minong Strand to show the elevation contrast between these various beaches.

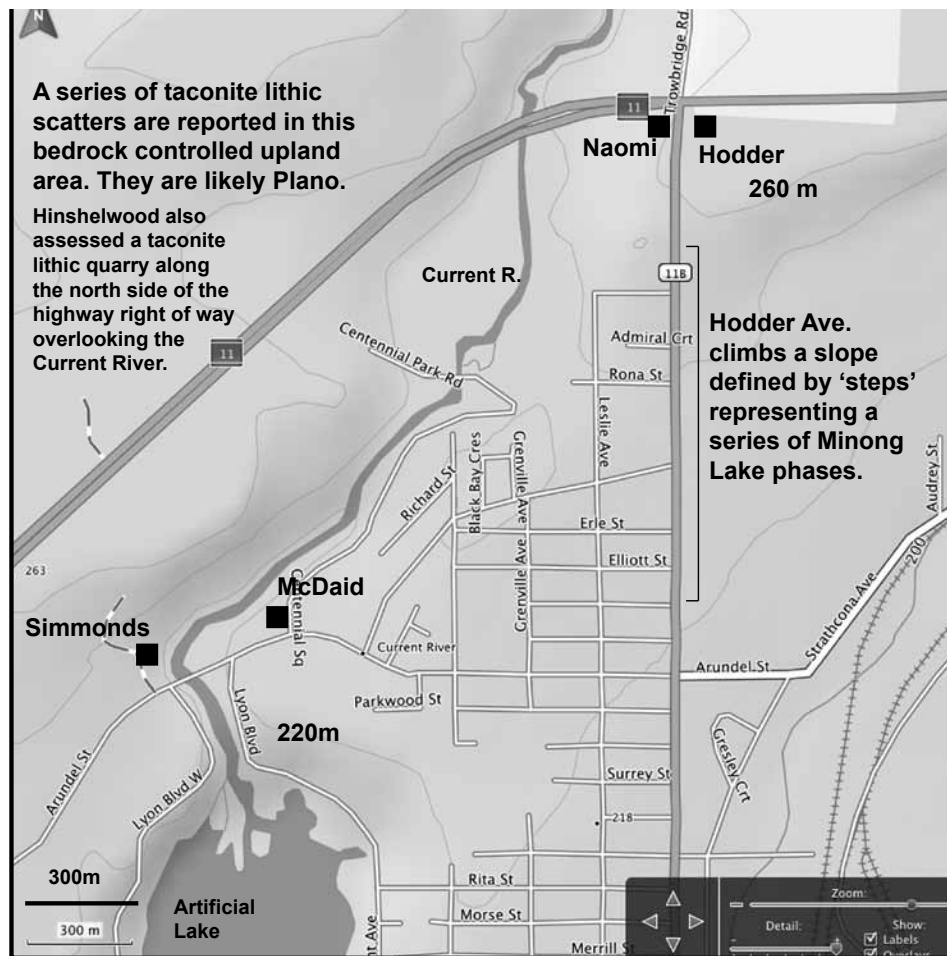


Figure 12 The Current River ‘mouth’ into Lake Minong, with the Naomi and Hodder Sites located on uplands well above glacial lakeshores. Note that Boulevard Lake is an artificial headpond for the dam and old hydro generating station on the Current River located downstream.

Stop 2 - Hillcrest Park Lookout.

UTM coordinates: NAD83; 16U 0334707E / 5366973N

We follow a route east along Oliver Road, and then north up High Street for a brief visit to Hillcrest Park from where the city can be seen in context with local topography and Lake Superior.

Stop 3 and 4 - Boulevard Park, the Bluffs and Centennial Park Rd. Figure 12

UTM coordinates: NAD83; 16U 0337219E / 5371002N

We travel northeast on High St., over the St. Joseph/Hillcrest island of Minong times, left on Balsam St, right on Hudson Ave and onto Arundel St. from which a left turn before Black Bay bridge will lead to the Bluffs Scenic Lookout (Stop 3). Walk down to Simmonds site. This site has been severely disturbed by park development and repeated cart track disturbance. We

then reboard to cross Black Bay bridge and turn left on Centennial Park Rd to visit the McDaid site (Stop 4), adjacent to the roadway.

The Simmonds (DcJh-4) and McDaid (DcJh-16) Sites

The present Current River runs in a channel incised into a gently lakeward dipping shelf of Gunflint formation that fronts a bounding rock wall, a structural feature, which now forms the ‘bluffs’ scenic lookout (Figs. 12, 13). In the Minong period, the river carried much water and sediment from inland proglacial lake flows and here evidence supports a major ‘bar building’ phase of coastal history, a series of river mouth spits or bars being formed on both sides of the river as water level generally declined.

The highest Minong shoreline lay against the bounding rock wall at approximately the 252 m (827’) level, but between 240 and 236 m (787-774’) a series

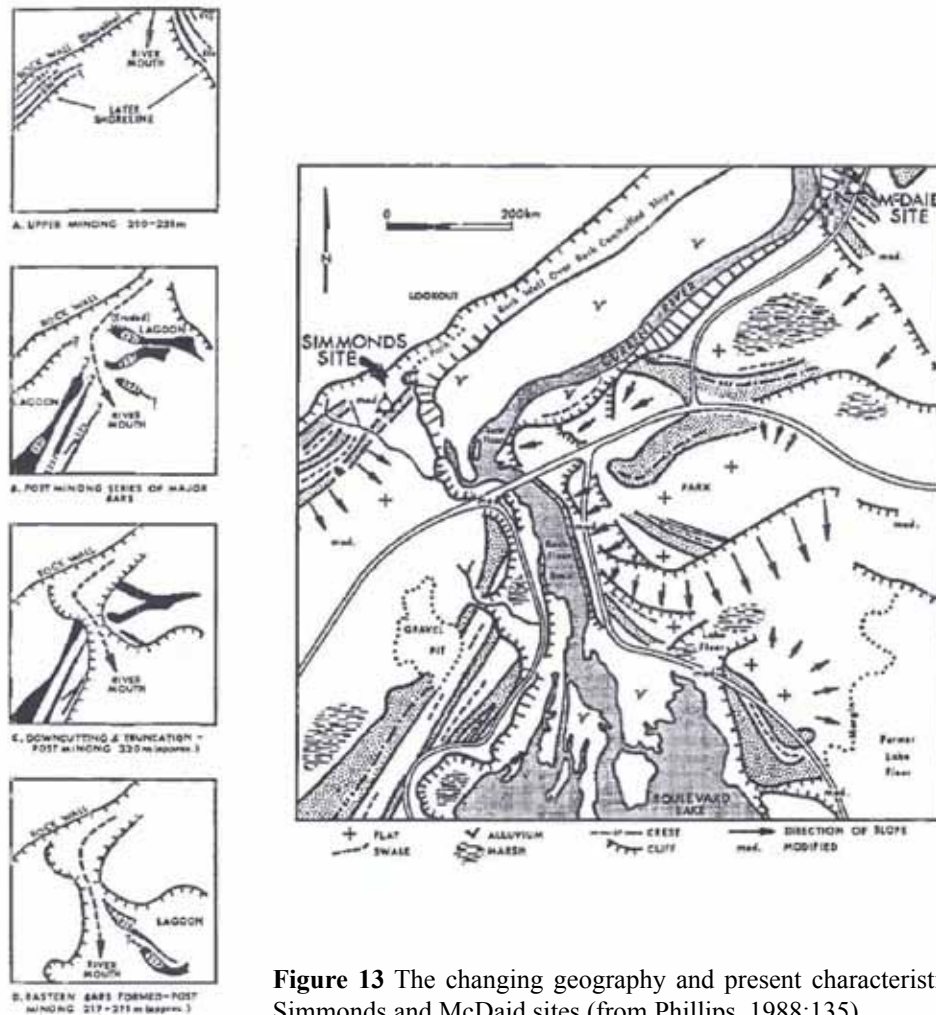


Figure 13 The changing geography and present characteristics of the Simmonds and McDaid sites (from Phillips, 1988:135).

of sand bars formed parallel to the wall on the west bank of the river and a matching series of bars on the east side curve sharply into the then river mouth from a source on the same rock wall east of the point where it is cut by the river. It appears that the river entered Lake Minong along the rock wall, forming offshore and beach bars which extend south-eastwards across the shallow McVickers embayment to the southwest. Simultaneously, longshore transport from the east built bars partially across the river mouth at times, only to be later truncated by fluvial action.

The Simmonds site on the west side, occurring on the parallel bars at about 236 m (774') is matched in elevation and position by the McDaid site on the eastern curving bars (Figs. 13, 14). Neither site is a long-term habitation site but show evidence of activity typical of a river mouth camping and fishing site. Interestingly, the major bar building episode appears to have been just subsequent to the occupation of these two sites, two large curving bars being formed at

231 and 227 m (758-745') on the east bank, the latter flat topped one largely a subaqueous feature that was probably contemporaneous with the supra-aqueous ridge form of the first. On the west bank a very long bar, now unfortunately truncated at its river mouth end, runs south west, in places broadening to over 100 m (328') in width. In almost text-like manner, the mean grain size and sorting characteristics along its length confirm that it prograded out from the mouth across the McVickers embayment, probably mostly in subaqueous form. Some evidence of Palaeo-Indian activity has been found on the crests of these newer bars but no sites equivalent to those named.

Stop 5 Naomi and Hodder sites. Figure 12.

UTM coordinates: NAD83; 16U 0338765E / 5372052N

Upon departure from the McDaid Site, the bus travels to Hodder Ave where it turns north to climb a long slope to its intersection with Highway 11/17

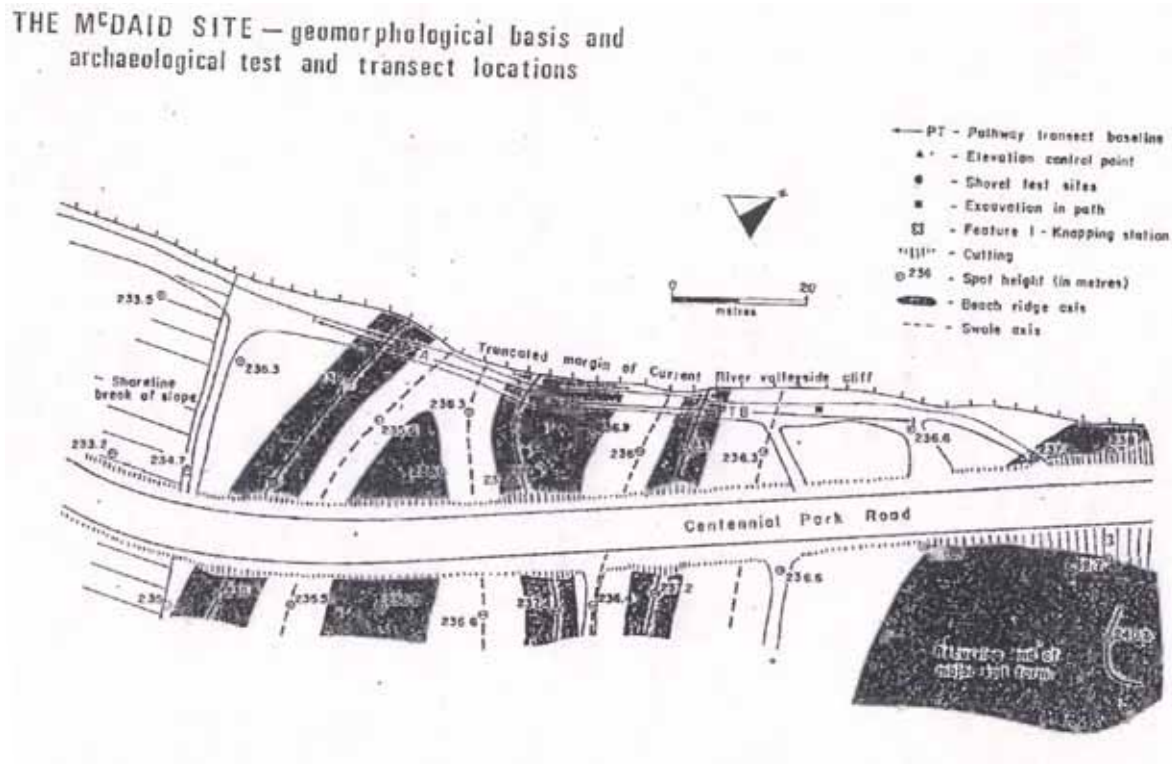


Figure 14 Geomorphic details of McDaid Site.

(Fig. 12). Hodder Ave crosses a series of ancient beach strands marking various phases of Lake Minong. At the top of this slope where Hodder intersects the highway, considerable construction is underway. We may not be able to park near the intersection nor leave the van, depending upon construction activity at the time of our visit.

The construction intercepted two archaeological sites located high on the bedrock-controlled uplands overlooking the Current River to the west and the waters of Lake Minong to the south. Both of these sites are far removed from modern or ancient water sources, and both are located on the north (leeward) side of the upland. Both the Naomi and Hodder Sites (Fig. 12) were salvage excavated as part of highway expansion by Western Heritage (an archaeological consulting firm based in western Canada). This position may have been calculated to offer protection from winds blowing off the glacial lake, and would have also provided a panoramic view of the extensive Current River valley to the north. If the landscape was comparatively open (taiga-steppe) then perhaps these sites might have offered a viewscape useful for observing game. These sites offer important 'cautionary tales' against undue emphasis on the apparent spatial association of Palaeo-Indian sites and former Lake Minong shorelines.

Clearly Plano settlement and land use was much more complex than first appearances would suggest.

The tour continues east along Highway 11/17 for about 20 minutes to the location of another cluster of Palaeo-Indian sites associated with the former outlet of the Mackenzie River into Lake Minong (Fig. 15). While two sites are found near the river bank (one on each high bank overlooking the Mackenzie River gorge), several other sites are associated with beaches and sand spits found along a shallow former embayment (Fig. 15).

Stop 6- Mackenzie Site Cluster (Figs. 1, 15)

UTM coordinates: NAD83; 16U 0355847E / 5377436N

Twinning of Highway 11/17 triggered extensive archaeological salvage excavations at several sites associated with high Minong Lake phases. Western Heritage is conducting ongoing salvage excavations. Thus, interpretation of these deposits remains preliminary and tentative. Again, because of active highway construction, it is not certain how close to this cluster of sites we will be permitted (safety and liability issues).

While the main site (Mackenzie I) is likely located

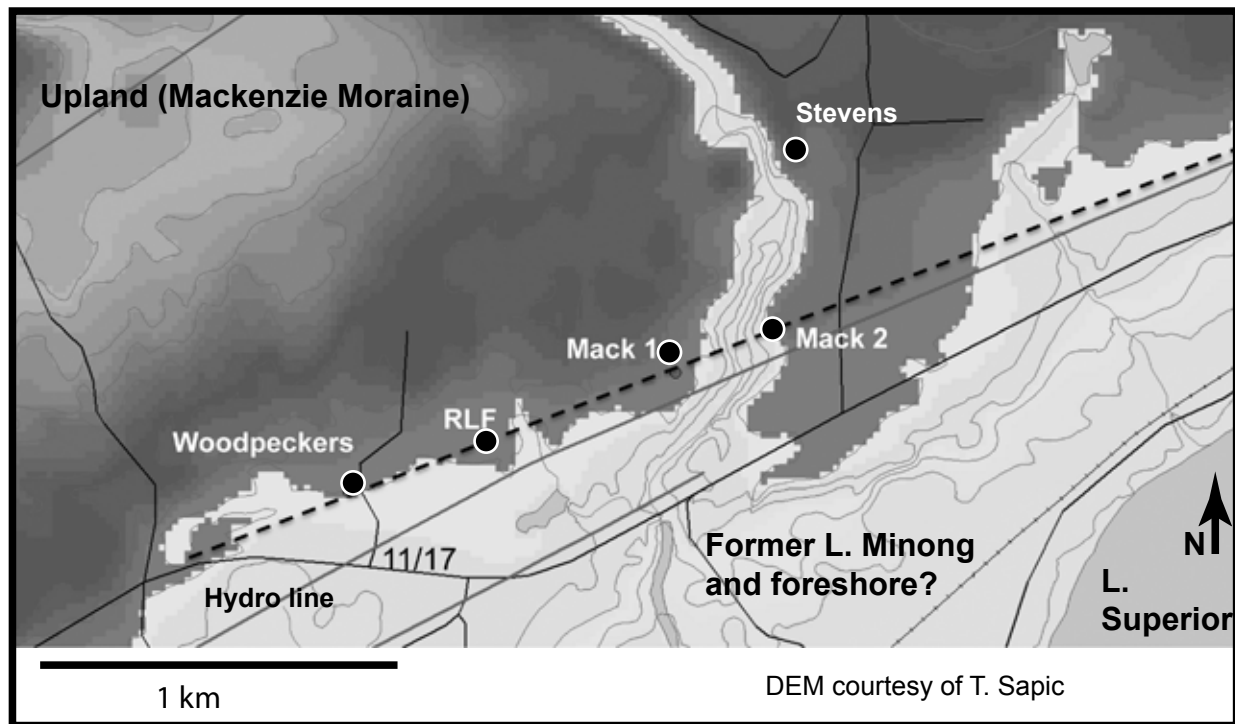


Figure 15 B/W version of a Digital Elevation Model of Mackenzie Site area, showing locations of Palaeo-Indian sites identified along highway construction corridor.

on a reworked deltaic deposit where the Mackenzie River channel entered Lake Minong, several others have been discovered upon beach strand features within a shallow embayment of Lake Minong (Fig. 15).

Archaeological inspection of where the hydro transmission lines cross the Mackenzie River in the 1970s led to the discovery of a Plano projectile point and two flakes in disturbed context. This site was named the Newton Site, and likely marks the southern extreme of the Mackenzie I Site (Fig. 16). The main part of the Mackenzie I Site remained undiscovered within the forest land to the north of the hydro transmission corridor until exploratory testing by Archaeological Services Inc. in anticipation of the construction of the new Mackenzie River bridge.

Several sites in this locality are found on sandy sediment, and are associated with bedrock knob exposures that might have once formed rocky coastal headlands. Longshore sediment accumulation from a Mackenzie River source seems to be the most likely explanation for these sandy beach features. Enhanced archaeological deposition (consistent with mixed function encampment) also seems to be associated with these protected zones. This is dramatically evident with the distribution of material culture at the Mackenzie I site where artifact processing and spatial analysis is the

furthest advanced. Also of note are several small and ephemeral springs that bisect the beach strandlines.

The most prominent encampment is the Mackenzie I Site, located on sandy sediments accumulated north and northwest of a bedrock exposure along the west side of the Mackenzie River gorge (Fig. 17). More than 2,500 square metres of this site has been excavated, making it one of the largest Palaeo-Indian excavations ever conducted in Canada (Fig. 18). Many thousands of pieces of debitage have been recovered. However, in sharp contrast to most other excavated local Palaeo-Indian sites that are dominated by debitage, the Mackenzie I site has also yielded a wide variety of tools, including upwards of 200 complete or fragmentary projectile points. This diversity of artifact types, coupled with the discontinuous clustering of material culture suggests a repeatedly used aggregation site where diverse activities were undertaken.

Because of the uncharacteristically rich recovery of diagnostic projectile points, this site will be an important type site for tool typologies useful for assessing Interlakes Composite in relation to other late Palaeo-Indian cultures throughout North America. The rich and diverse artifact recovery is enabling graduate student research addressing a wide range of topics (tool typology, lithic reduction strategies, sedimentology



Figure 16 Oblique air photo of new bridge over Mackenzie River, with Mackenzie I and II sites on each end. Early Lake Minong shorelines coincide with cleared area, with a main bank defined by white line.

and geoarchaeology, artifact patterning, activity areas and site function, etc.

While the Mackenzie River gorge is likely deeply downcut from its former configuration, the dense site distribution suggests repeated use of the embayment during Palaeo-Indian times. Notably, nothing except Palaeo-Indian diagnostic material has been encountered along the highway construction corridor, but reconnaissance by K.C.A. Dawson in the 1960s and 70s along the lower Mackenzie River valley revealed a succession of Archaic and Woodland era sites at successively lower outlets into Lake Superior.

Continued research in the Highway construction corridor has led to discovery, assessment and salvage excavation of several other sites, of particular note, the Woodpecker I and II and RLF Sites (Fig. 19). These sites are positioned upon the top of a sandy bank overlooking an extensive area of muskeg and beaver ponds. These wetlands are interpreted to be the former shoreline and shoals of Lake Minong, with the sandy/fine gravel bank forming the beach strand line.

The Woodpecker I and II sites were located by ASI during preliminary assessment of the highway corridor. These sites are initially interpreted to be

localized lithic scatters. Subsequent forest clearing and mapping greatly improves interpretative resolution in their paleo-hydrological context (Figs. 20 and 21).

The gently curving high bank illustrated in Figure 19 is interpreted to be the Lake Minong strandline, which forms a shallow embayment dotted with bedrock knob exposures that acted as rocky headlands along the former shore. These headlands broke the wave velocity along the shore, and allowed longshore accumulation of sediment on their leeward side. This resulted in a high raised bank with subtle berms and swales defining storm beach features. Upon one such berm is the small encampment/flaking station called the RLF site. The Woodpecker Sites are more complicated, with a bedrock dome enabling the development of a long spit or ridge of sandy sediment to accumulate to the west of the bedrock. This beach feature is bisected in at least three places by small streams that flow through underfit gullies across the beach strand. At issue is whether these streams are contemporaneous with Lake Minong water levels, or whether they are of mid to late Holocene derivation. Given the small drainage basin and rather small stream budget, coupled with the fact that no alluvial fan sediment accumulation is noted

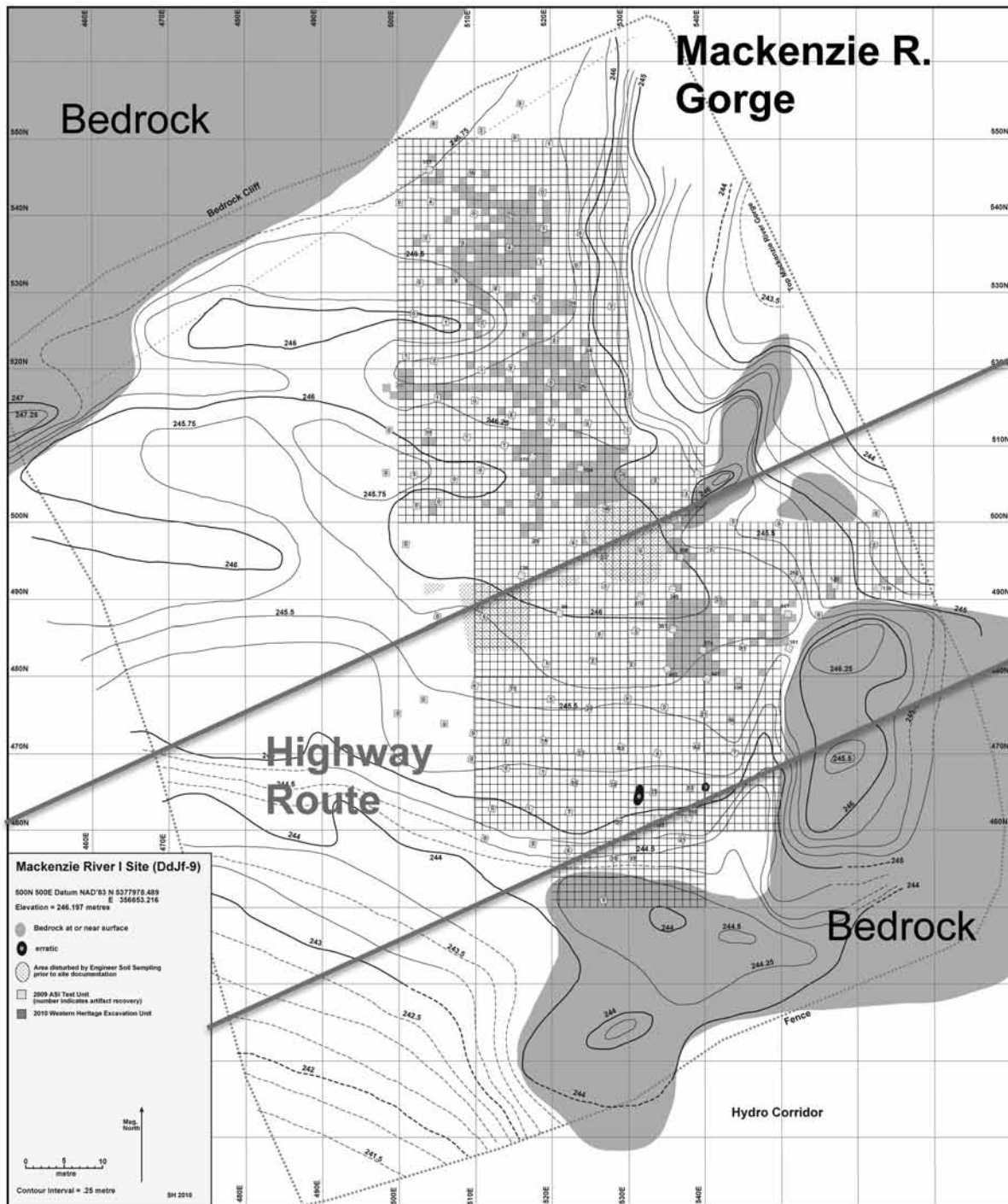


Figure 17 25 cm contour map of Mackenzie I Site showing bedrock with sandy/gravelly sediments between exposures. The extent of the block excavation is not fully documented here (see Fig. 18). Note the configuration of the contour lines suggests beach berms and back-beach swales. Well-sorted gravel deposits on the north side of main bedrock dome suggests former stream bed. Deep stratigraphic sections reveal deltaic sedimentation (well-sorted lens of fine pebbles, sands, and silts) overlaid by aeolian reworked fine sediment.

in the muskeg lowland below the beach strand, it is suspected that they are of early Holocene derivation. That is, larger stream-flows drained downslope off the Mackenzie Moraine uplands to the north, and either accumulated in the swales behind the storm beach, or

down and through the beach, with eroded sediment being carried away by wave action along the lakeward side of the beach strand line. When examining Figure 20 and 21, the most densely occupied portion of the Woodpecker I and II sites are associated with the banks

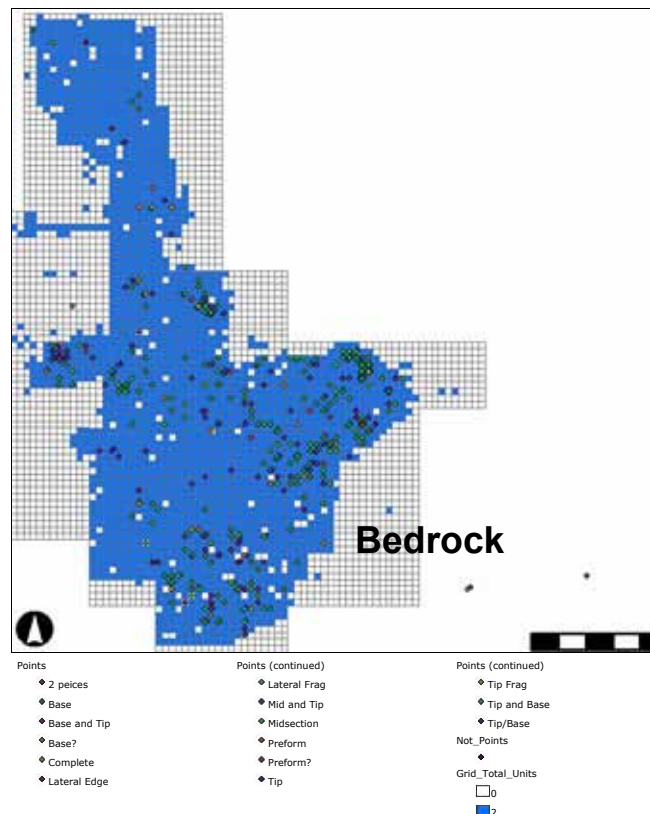
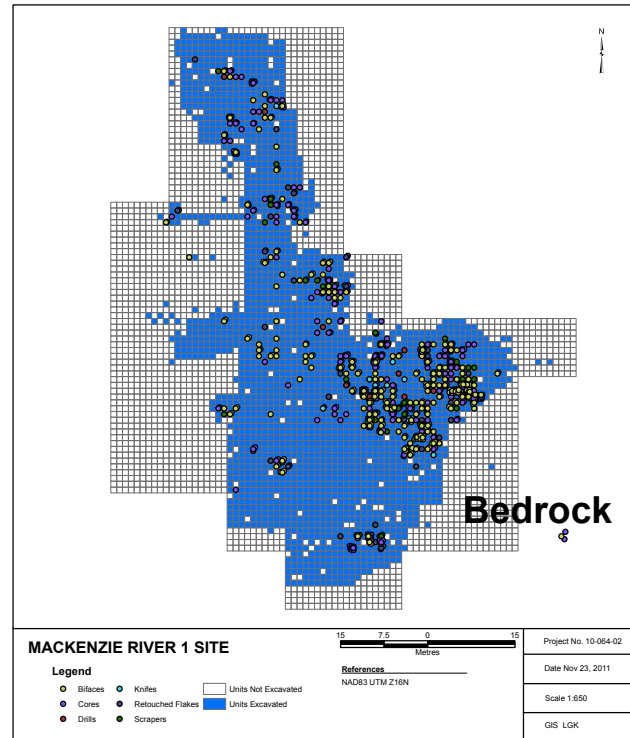
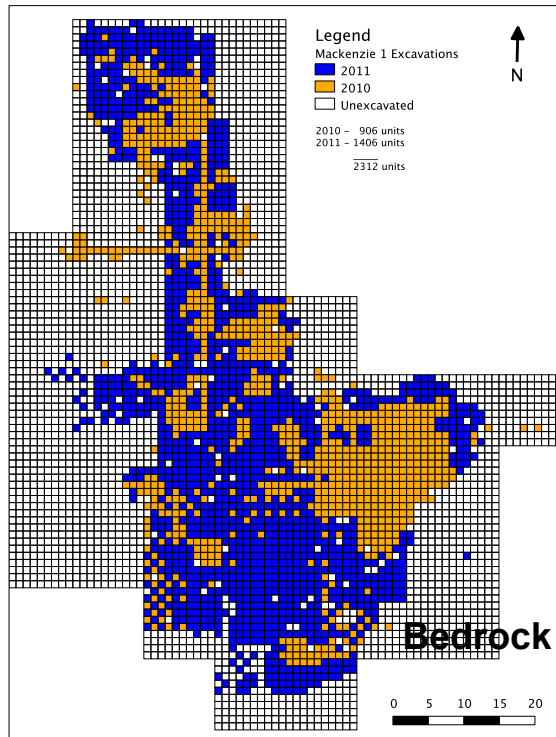


Figure 18 Preliminary plots of various tool types from the Mackenzie I excavations (Courtesy of Western Heritage). The plotted objects represent a very small fraction of the assemblage (formal and informal tools), while the many thousands of debitage, core fragments and other discarded debris is still being catalogued. The upper left image shows the extent of excavation in each of the two years of work at the site. When compared to Figure 17, it is evident that an important area for settlement was on the flat sandy surfaces to the immediate northwest of the main bedrock dome. This supports the speculation that these domes were important for breaking wave velocity and facilitating longshore sediment accumulation along their lee sides. We also speculate that these bedrock exposures were important considerations for settlement as they provided shelter from winds blowing off the lake. Such notions assume that the site was occupied while Lake Minong was at its high level.



Figure 19 View northeast from existing highway along the construction zone. Note the ‘bank’ in the clearing (left side of frame) that likely is the former Lake Minong strandline. In the foreground the current highway cuts through this former beach feature. Note the location of the Woodpecker I, II, and RLF sites along the beach strand line. The former is located on the northwest side of a bedrock knob, while the RLF site is a small flaking station located on a berm (perhaps a high storm beach) overlooking the main strand line.

of one such stream (Woodpecker I), or on the protected leeward side of the bedrock dome (Woodpecker II). Thus, occupation was densest along the protected leeward side of the bedrock dome and along the banks of the now dry stream bed.

The RLF Site was discovered after a bulldozer cleared the dense coniferous forest along the north highway lane, revealing taconite flakes. Subsequent test-pitting and block excavation revealed a series of small and localized lithic clusters that are interpreted to represent short-term camps or lithic reduction stations. Topographic mapping (Figure 22) facilitated interpretation of the site locality to represent a raised storm beach berm that overlooks a back-beach swale. This high storm beach also overlooks a second swale (to the south and lakeward) and a second slightly lower storm beach berm located on the top brink of the bank forming the most prominent Lake Minong strandline (Figure 22). The RLF site is at least 50 metres north (inland) from the main strandline and at least 100 metres removed from the nearest stream bed that bisects the former beach zone. While yielding much less artifact material than the others in the area, such sites remain scientifically important because their relatively simple and brief depositional history render them much more archaeologically interpretable.

Stop 7 McIntyre River outlet and the Biloski Site. Figure 23

UTM coordinates: NAD83; 16U 0332310E / 5367746N

Upon completion of our examination of the Mackenzie River area, we return to Thunder Bay, travelling down the 11/17 Expressway past a series of sites that have been located along the former shores of Lake Minong (and also shores representing the pre-Nipissing Transgression (Fig. 23). In the interest of time, we will just stop along the highway. On the north side the low Minong bluff is seen, with a housing development on top. The Biloski site occupies this surface and a small sand bar at the bluff foot, where a small embayment and secondary river channel once existed. This and several other nearby sites are likely Plano age, but the proximity of the pre-Nipissing Transgression strand line suggests the close spatial association of Plano and Archaic sites, consistent with Hinshelwood’s (2004) observations.

As the Expressway curves to the south, the area on the right holds a lengthy baymouth bar on the end of which lies the Catherine site, unseen from the road (Fig. 23). On the summit of Rabbit Mt. (275m, 902’), which rises behind the embayment, lies the Irene site (Fig. 23), a lookout site.

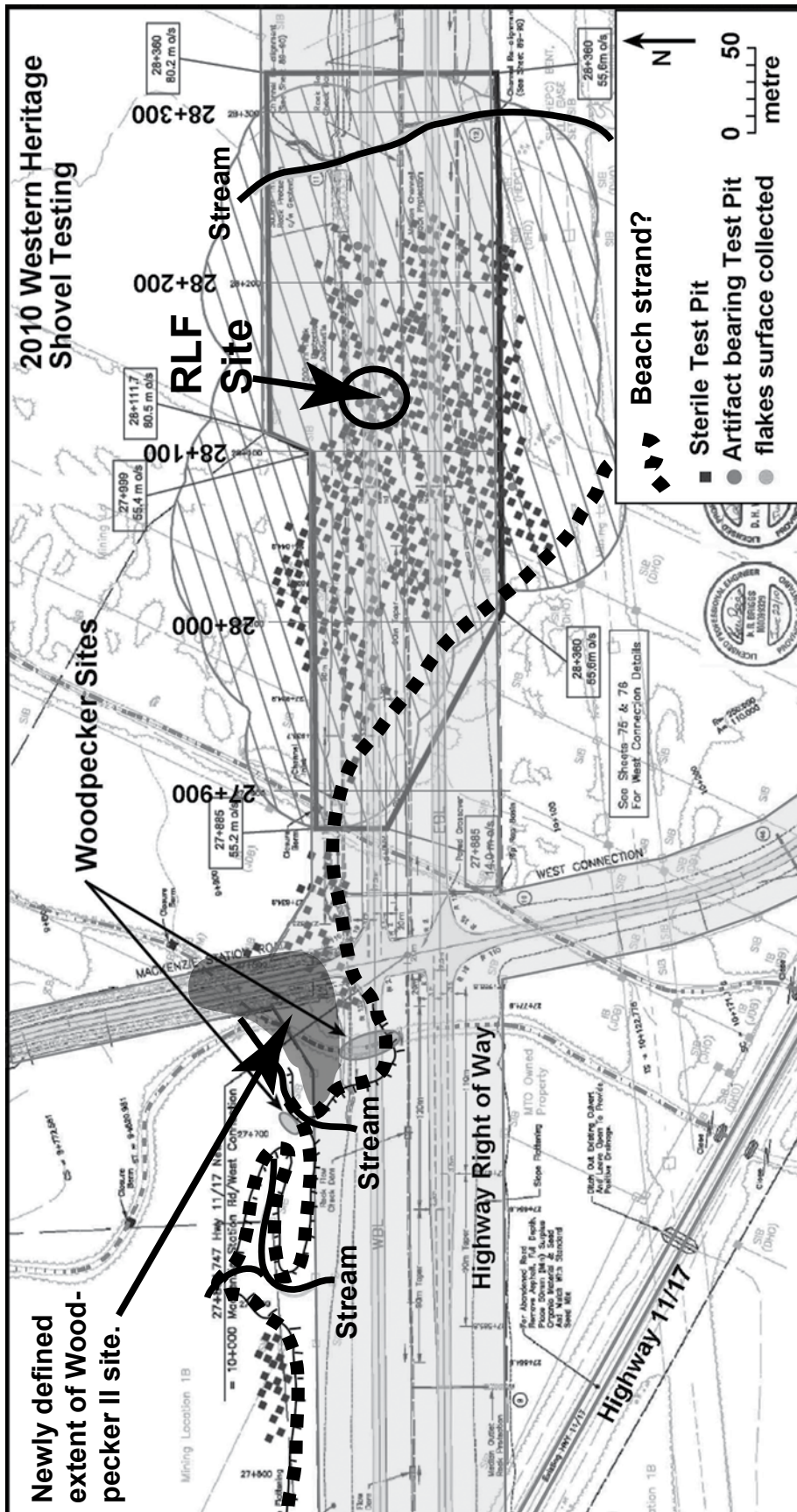


Figure 20 The Woodpecker and RLF sites are located upon sandy and fine gravelly sediments that form part of the Lake Minong strandline. Note the position of the sites in proximity to small relict stream beds that drain across the strand and into the projected Lake Minong. Noteably, no alluvial fan is associated with the lower end of either of the stream beds associated with the Woodpecker Sites, suggesting that it is not a recent stream that eroded through the pre-existing sandy ridge. Continued excavation at the Woodpecker site locality suggests a much more extensive deposit on the north and west side of the bedrock dome illustrated in Figure 21. This suggests encampment on flat sandy beach deposits that accumulated on the lee side of the proposed bedrock headland adjacent to a small spring flowing across the beach. The RLF site was discovered while archaeologists were walking the cleared north lane centre lane. Subsequent excavation revealed a small encampment/flaking station upon a raised berm overlooking the primary Lake Minong Beach strandline. While somewhat removed from the site, also note the small underfit stream bed located about 100 metres east of the site.

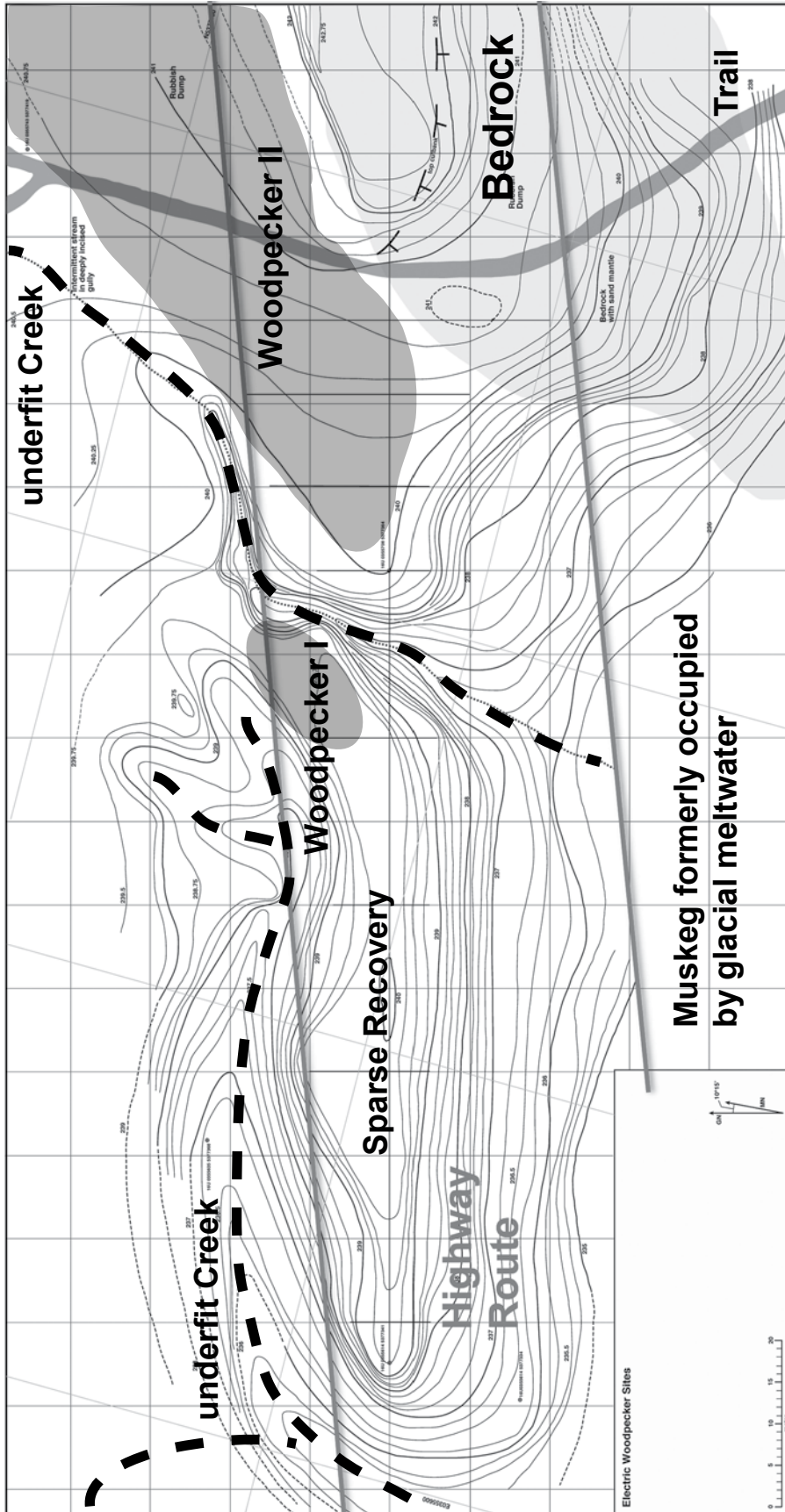


Figure 21 25 cm contour map of the Woodpecker Sites located on a raised sandy 'ridge' formed north and west of a bedrock exposure. Low muskeg ground to the south of sandy ridge likely contained Lake Minong waters. Several abandoned or underfit stream beds bisect the sandy ridge, and suggest streams flowing into Lake Minong across this raised beach formed by longshore sediment accumulation on the lee side of a bedrock headland protruding into Lake Minong. Ongoing geoarchaeological and OSL dating research is assessing the viability of this interpretation. Archaeological materials were initially discovered along the banks of one of these relict streams, and also in disturbed context along the cart trail that bisects the site. Continued archaeological investigation reveals an extensive deposit on the northwest or lee side of the bedrock and along the east bank of the small abandoned stream bed.

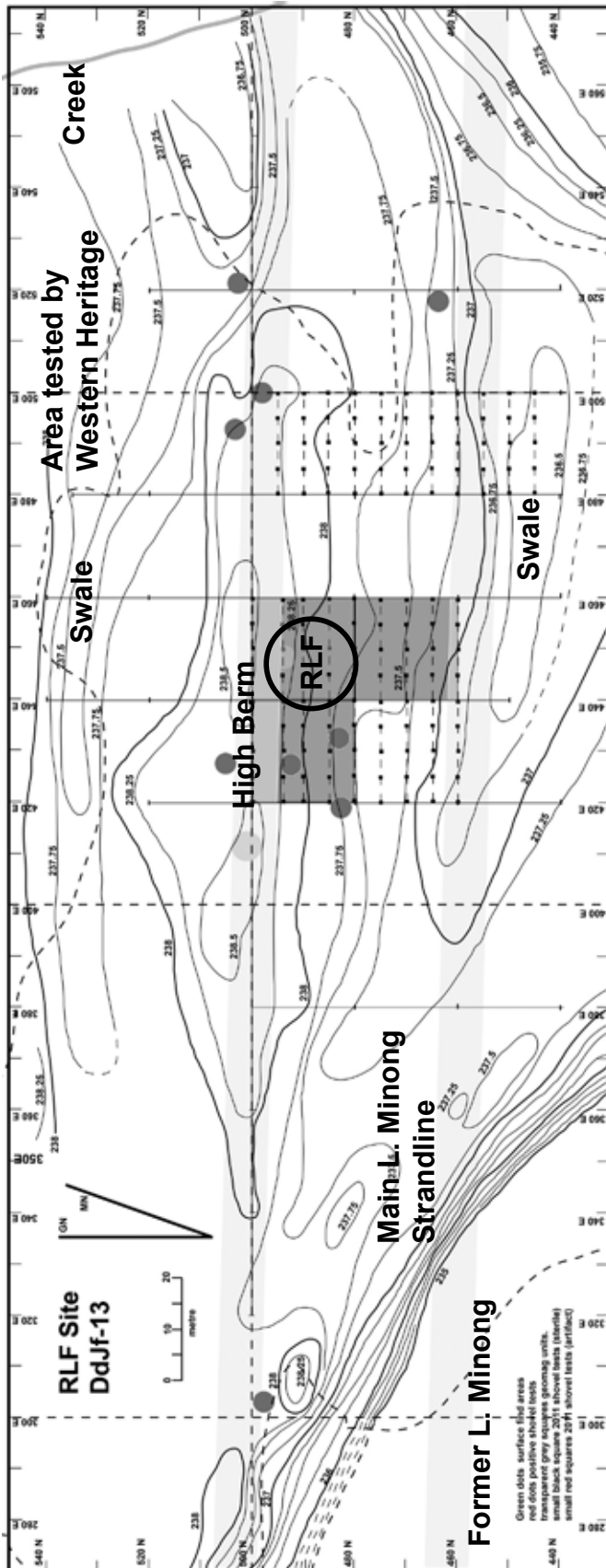


Figure 22 25 cm contour interval map of the RLF Site. This site was accidentally discovered by archaeologists while walking along the bulldozed north lane of the proposed highway. Lithic debitage exposed during bulldozing down trees led to shovel test reconnaissance, followed by Magnetic Gradiometer survey (Grey squares), and block excavation. Very localized dense clusters of debitage were encountered that suggests localized encampments or flaking stations. Much of the material appears to be located on the south flank of a raised berm (perhaps a sandy storm beach) about 50 metres north of the most prominent Lake Minong beach strand line. A small underfit stream bed is located ca. 100 metres east of the site. Locating such small and ephemeral (but highly interpretable) encampment zones is very difficult in forested conditions, particularly using conventional site prospecting techniques (5 metre interval shovel test pits). Western Heritage and Lakehead U. have been collaborating in experimentation of multi-proxy approaches to site discovery.

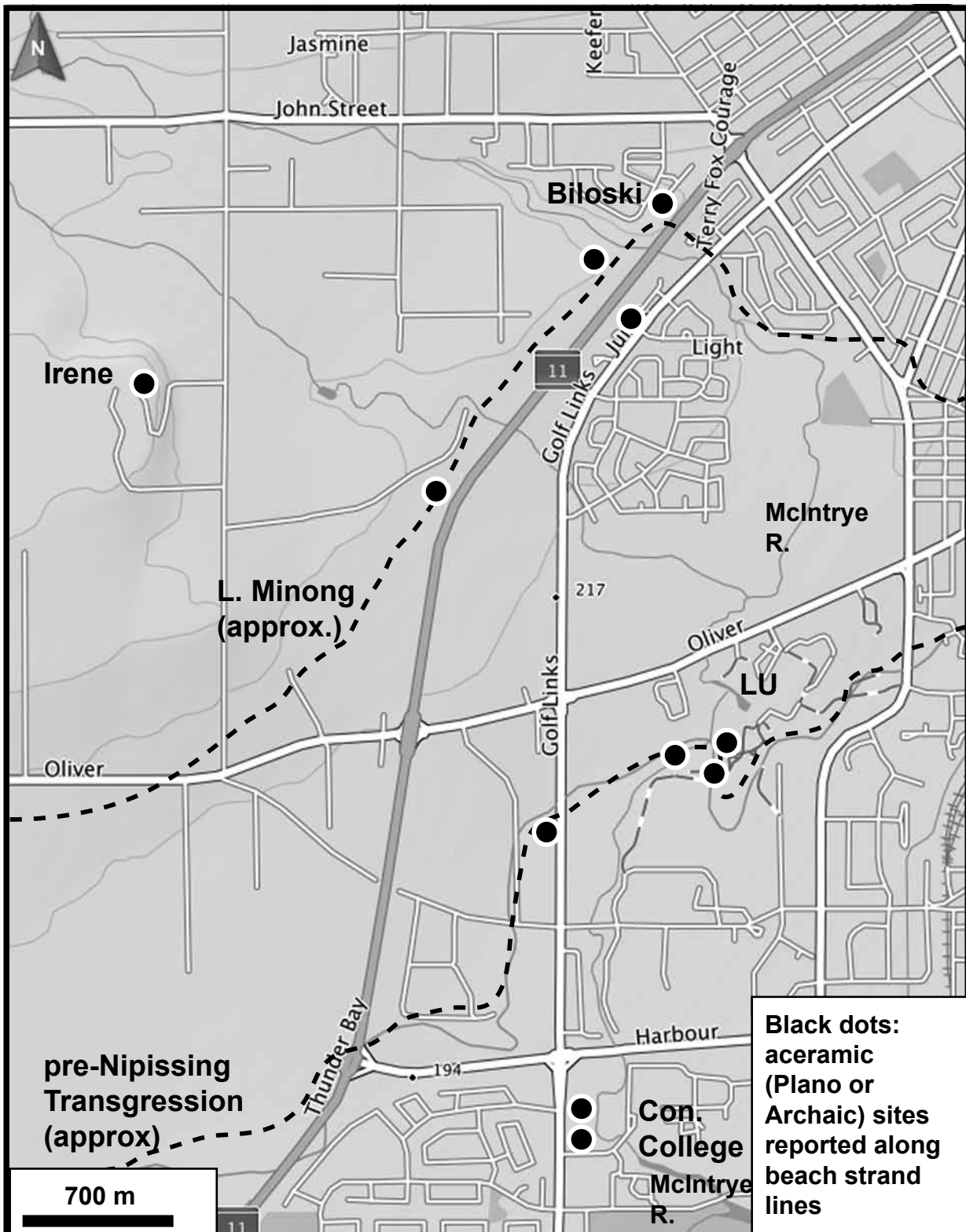


Figure 23 Archaeological sites near the McIntyre River. Some of these sites have confirmed or probable Plano affiliation, specifically those near the L. Minong Strandline. While not yielding diagnostic artifacts, it has been proposed that the sites near the pre-Nipissing Transgression are of Archaic affiliation. Also note the Irene Site, located on top of Rabbit Mountain, overlooking the Lake Minong shoreline.

Stop 8 - Mapleward Road, the Cummins Site. Figure 24

UTM coordinates: NAD83; 16U 0325670E / 5364042N

The route turns west on Oliver Road and within 2 km rises up the distinct Minong bluff. A left turn on Mapleward Road sees a gentle descent to the Minong shore again, and the Cummins site.

The Cummins Site (DCJi-1)

The best published Palaeo-Indian site in the region is the Cummins site (Fig. 24), that is one of a number of Plano sites in the area where the Neebing River flowed into Lake Minong. Reported by a prominent local collector (Hugh Cummins) in 1962, the site was most intensively excavated as part of Dr. Pat Julig's Ph.D. research. Initially considered a typical surface site, Julig et al. (1986) concluded that it is a rare stratified Palaeo-Indian site, under continued use over a long period of changing environmental conditions. The basic tool kit is reminiscent of Plains Plano culture (Dawson, 1983; Julig, 1984), and the recoveries imply diverse activities, albeit dominated by stone quarrying and knapping. This includes woodworking, fishing and beaver trapping in addition to the regular hunting of caribou and perhaps bison (McAndrews, 1982). Newman and Julig (1989) also attempted to extract and characterize blood residues from tools from this site, and propose a diverse diet. The Cummins Site was a major regional preform-making centre and the presence of exotic lithic components implies a broad geographical interaction with other groups in the region (Julig, 1984).

The Cummins Site occurs across the surface of several large sand and gravel bars which trend northeastwards from the then Neebing river mouth across a broad southerly gentle dip slope of local Gunflint shales, overlain by a thin water washed silty till (Figs. 24 and 25). In these shales occurs jasper taconite, the major source of tool making material. Figure 25 shows the fenced area of the site and the morphological details (Julig et al., 1990). Figure 26 shows the paleogeographic reconstruction of events believed to have formed the area (Phillips, 1982). Longshore transport from the Neebing river mouth built a series of progressive bars across the shallow water rock shelf, recurving into a minor river valley which formed a sheltered embayment to the west of a rock island. A further bar was built along the front of existing ones eventually crossed the embayment

in tombolo-like form to enclose a small lagoon, the Cummins Pond, after which lower and later shoreline features mimicked the established plan shape. A pollen core taken in the pond suggests this enclosure took place before 8.1 ka BP. (Julig et al., 1986).

The plan shape and structure of the bars is not compatible with a declining water level margin, indeed the accumulation and building up of these large features on a gentle shelf slope is unlikely without transgressive wave action. Even so, such action across a gentle shore slope would not ordinarily build up a large supra-aqueous bar without some initial encouragement to accumulate sediments along a line rather than disperse them in a sheet. The key to the existence of these bars in this location is underlying bedrock control. While Julig (1984) determined through resistivity and ground radar studies that variation in sediment character and bedrock irregularities could be traced, a more simple reconstruction of the bedrock profile is also possible. By surveying down the exposed bedrock dip-slope north of the site, the trend (see inset, Fig. 26) showed that beneath the site must lie a marked rock step, typical of many that occur in the present topography of these flat-lying shales, and often sharpened by wave action. Accumulation of sediments took place firstly against this rock step and subsequently over the top of it. It is possible that some till remained in the angle of the step. To the west of Mapleward Road an exposure of coarse sand and gravel with angular shale inclusions at the rear of the bar in contact with bedrock suggests overwash of the type that would be expected in this scenario. Once the linear feature was established, longshore sediment supply would extend its length and add to the lakeward face. However, overwash and building in elevation is characteristic of wave action either in an extreme storm event or a transgressive lake margin. Again, a 'bar building' period seems evident at Cummins. Julig (1984) found confirmatory evidence of habitation of these bars during the process of their accumulation, a buried layer of water worn taconite artefacts being associated with a period of overwash and construction. This suggests occupation of the site ... "when a lakeshore location offered advantages of longshore access, lagoon fishing and perhaps water transport" (Dawson, 1983b).

A cremation burial site was recovered from a sand quarry wall on the beaches marked by the triangle in Figure 25. This was erroneously cited as being located on the rock island by Dawson (1983) since Bill Ross (pers. comm.) reports information from J.V. Wright that suggests that a cartographic error resulted in its

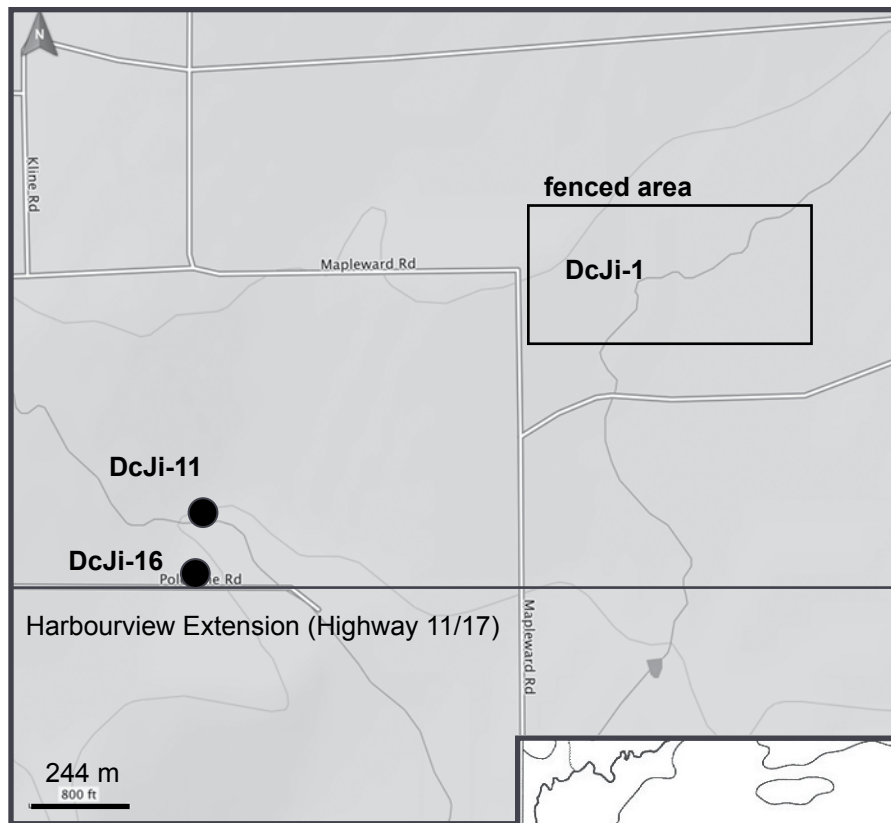
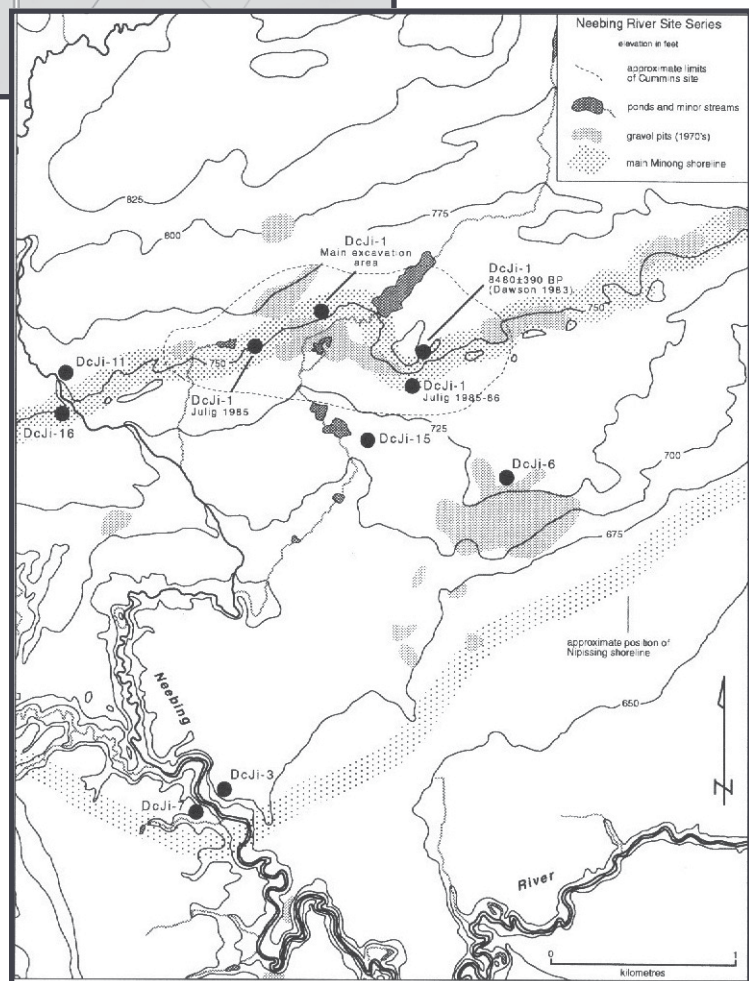


Figure 24 The Neebing River cluster contains a series of probable Plano and Archaic sites associated with shoreline features of either the Lake Minong Strandline, or the Nipissing Transgression strand. Perhaps the most important of this group is the Cummins Site (DcJi-1), a registered National Historic Site, a portion of which is provincially owned (within fenced area). Much of the balance of this site is slowly being destroyed by urban development. J.V. Wright, K.C.A. Dawson and P. Julig have conducted excavations at the Cummins Site. Julig's geoarchaeological examination, coupled with geomorphic mapping by B.A.M. Phillips and P. Fralick have enabled geomorphic interpretation. A Hinshelwood's salvage excavations at DcJi-16 demonstrate Archaic reoccupation of this Plano site (2004).



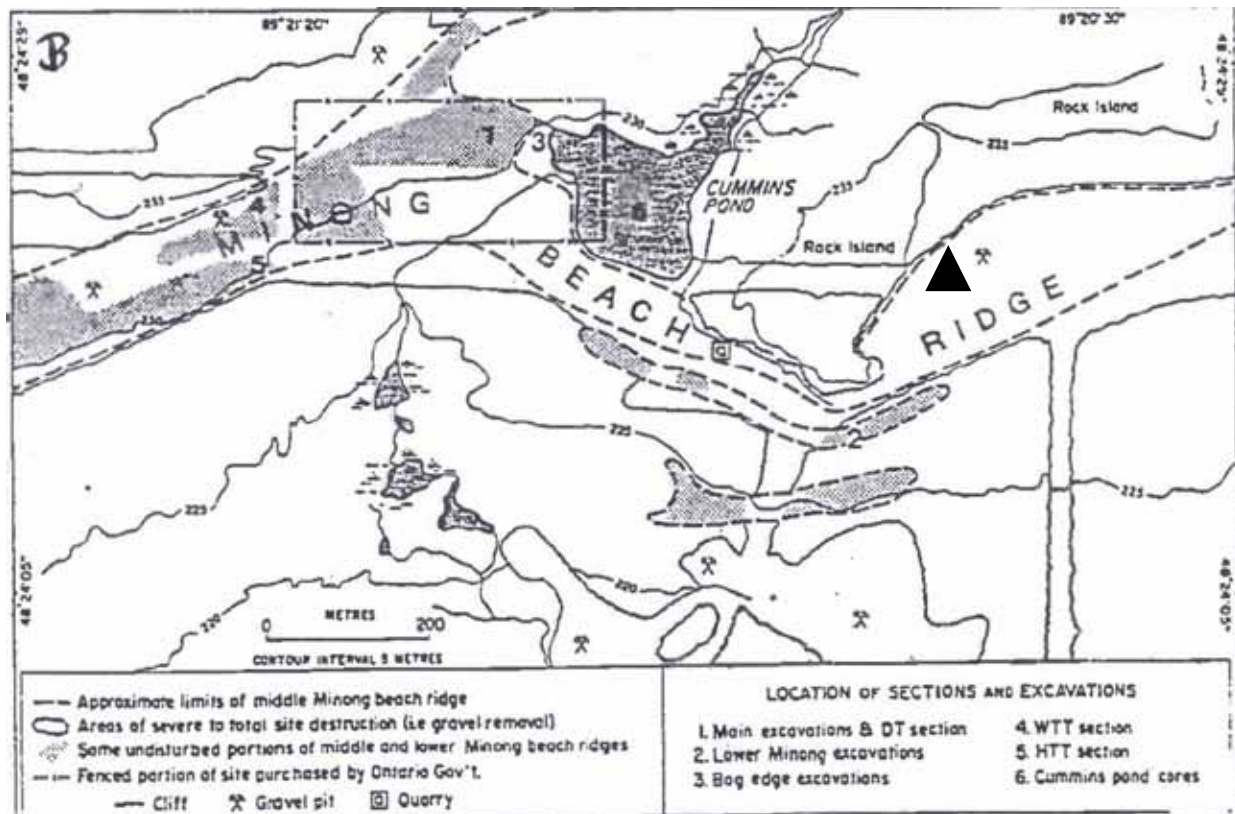


Figure 25 Cummins Site, with major paleo-lacustrine features (from Julig, McAndrews and Mahaney, 1990 based on Phillips, 1982). Black triangle marks the approximate location of the cremated human burial recovered by Wright in 1963 that was radiocarbon dated to 8480 ± 390 (NMC-1216).

misplacement, thereby miss-informing Dawson. This small sample was subjected to AMS radiocarbon dating, and the low collagen yield resulted in its complete consumption. The resultant date was $8,480 \pm 390$ years BP (NMC-1216). Note that the very large sigma associated with this date renders it a not particularly precise measure of the antiquity of the Cummins occupation, but it remains the oldest dated human remains recovered so far in Ontario.

Julig (1984) found artefacts both just below the aeolian sands that characteristically modified the topography of the shoreline features once the offshore shelf was exposed by lower lake levels (circa 8.0 ka BP) and in peat that accumulated in Cummins Pond. Indications are that long after the lake ceased to lap the beach face, the site remained used, at least until 7.5 ka BP (Julig et al., 1986).

A well-developed, beach deposit is exposed at the Cummins site. Figure 27 shows a schematic representation of the sandy cliff face where the strandline deposits are visible. The following quotation describes the sequence.

Low-angle, planar cross-stratified sands of

the foreshore dominate the exposure. Individual laminae dipping 3° to 12° lakeward (original swash-backwash surfaces) are arranged into packets which erosively truncate one another at very low angles. The planar cross-stratified sands are transitional both laterally and vertically into massive sands through a bioturbated zone. The bioturbated area represents a sparsely vegetated backshore environment while the massive sands were deposited as aeolian dunes.

The foreshore sands are erosively truncated in the eastern portion of the cliff by magnetite-rich sands. Internal structures indicate that they are also foreshore deposits. The magnetite-rich foreshore laminae were formed during regression when storm wave activity reworked the lower portion of the beach, erosively truncating the older deposits. During the storm events sand was removed and stored in offshore bars. In intervening periods of fair weather, sand moved from the offshore bars back onto the beach and was winnowing by small waves, producing the magnetite-rich lag deposits filling scour truncations. The beach assemblage is overlain

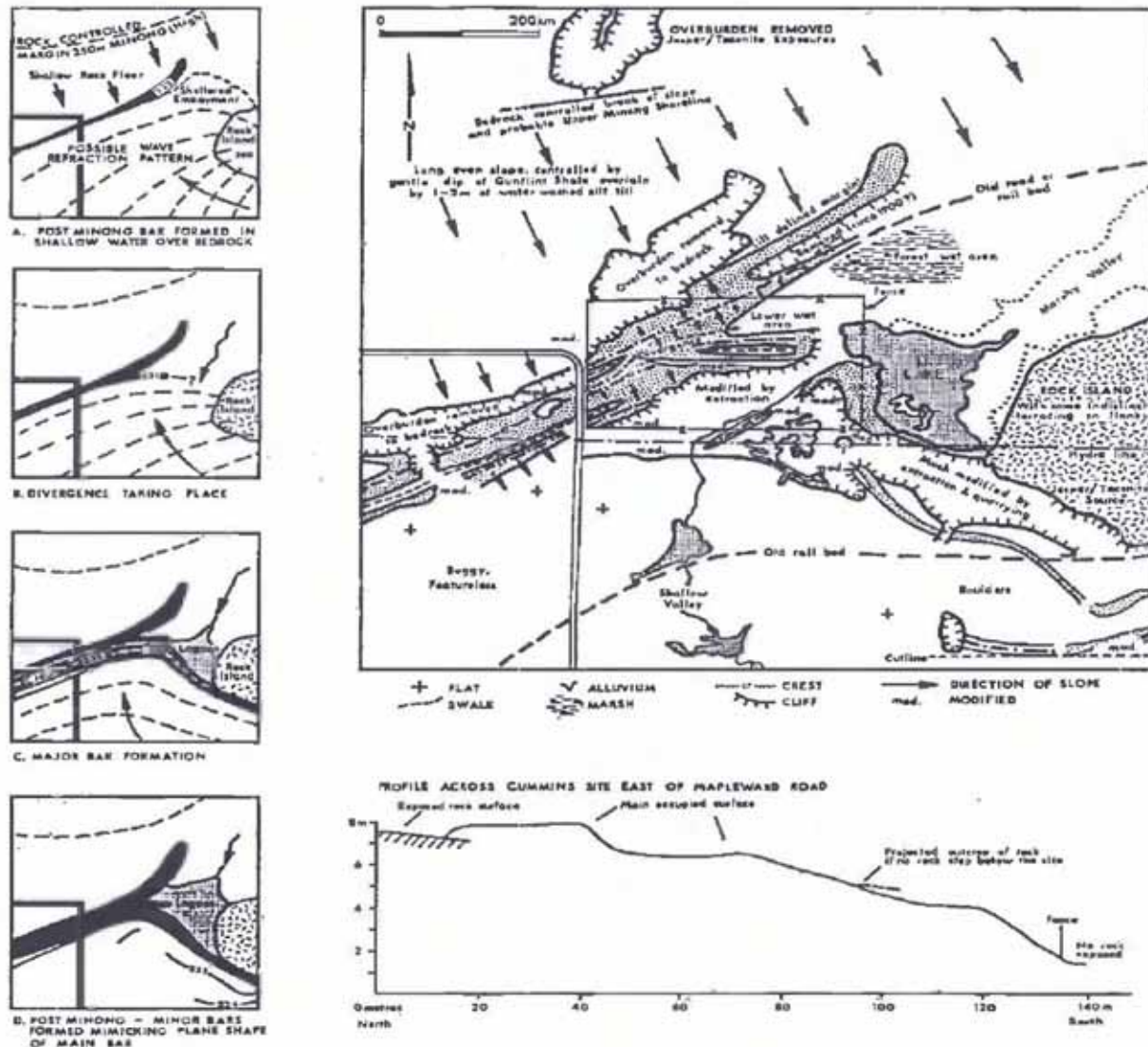


Figure 26 The changing geography and present characteristics of the Cummins Site (from Phillips, 1988:133).

by erosively based, trough cross stratified sands. These were formed after subaerial exposure of the area. Major rainfall events caused streams to flow off the adjacent rock knoll dissecting and reworking the upper layer of beach deposits.

Phillips, Fralick and Ross, 1987.
From INQUA Field Guide C-12,
Eds. Geddes, Kristjansson and Teller.

Stop 9. The Neebing River Site (DcJi-16). (Fig. 28)

UTM coordinates: NAD83; 16U 0324874E / 5363477N

In anticipation of highway development west from Thunder Bay, Andrew Hinshelwood conducted extensive excavations in the vicinity of the Neebing River south of the Cummins Site. One such site is

DcJi-16, located at the former outlet of the Neebing River into Lake Minong (Figs. 24 and 28). This stop on the side of the highway is approximately within the middle of this former site. In fact, the highway runs up the middle of the site, likely destroying most of the rivermouth feature upon which it was located. While the salvage excavation report has not been published, some of the relevant data is included in Hinshelwood's 2004 publication.

He identifies and excavates a series of habitation areas on the terraces that successively built up at the river mouth. This site strongly resembles the site context noted at the McDaid Site, but with an important new observation. Figure 28 reports the recovery of several Archaic affiliated objects in much the same context as the Plano materials. He proposes that this locality (and

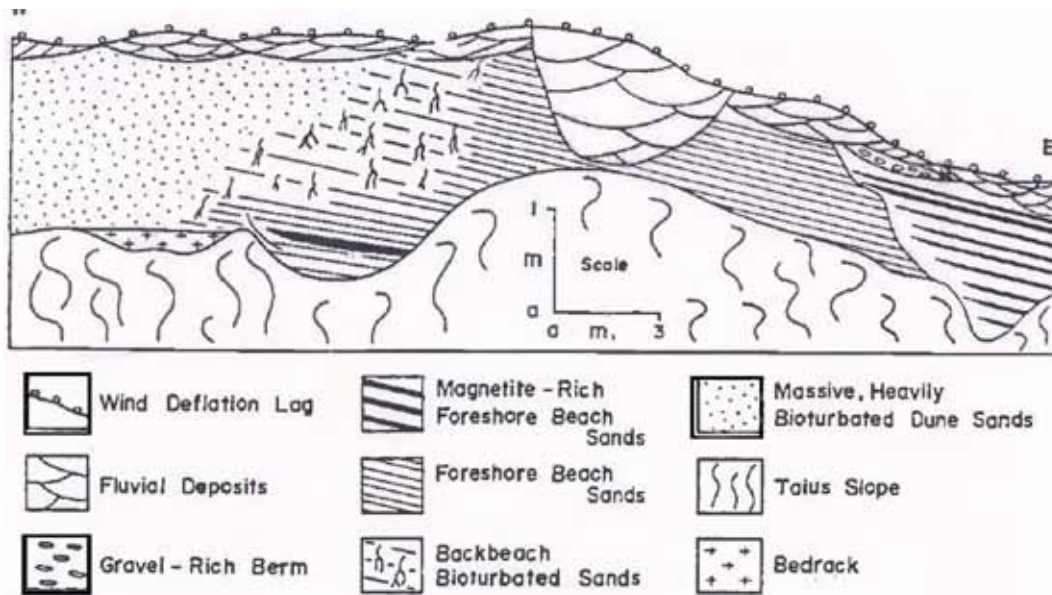


Figure 27 Schematic representation of the cliff face at the Cummins Site (Phillips, Fralick and Ross 1987).

others) were first occupied by Plano people when Lake Minong waters were nearby, but that it was reoccupied during Archaic times, perhaps because of its sandy riverbank sedimentary context coupled with nearby exposures of taconite. He also makes the point that Nipissing Transgression water levels had reflooded the Kaministiquia River delta to within about 2 km of DkJi-16, and that probable Archaic age sites are to be found to the south near the new outlet of the Neebing River into the Lake Superior Basin (Fig. 24).

Stop 10 (several) - The Rosslyn Delta. Figure 23:A.

UTM coordinates: NAD83; 16U 0317690E / 5360523N

The route continues west up the Harbour-view Expressway extension, past the Neebing River site (DkJi-16), and then southwest to once more intercept the bluff defining the Lake Minong shoreline at the Highway weigh scales station. Turning left off the highway, once on the terrace, the road leads south to several fields in which lithic material has been found.

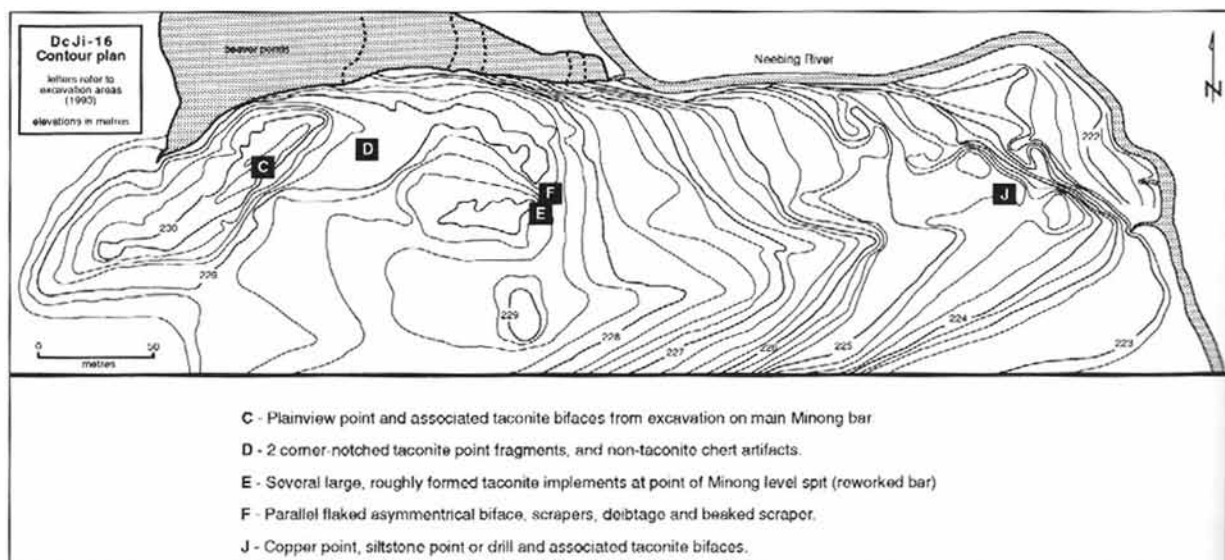


Figure 28 Contour map of DkJi-16, at the former outlet of the Neebing River into Lake Minong. Note its similarity to the placement of the McDaid Site (Fig. 14). Also note the mixed Plano and Archaic objects recovered on this site that Hinshelwood (2004) interpreted as evidence of re-occupation of this location.

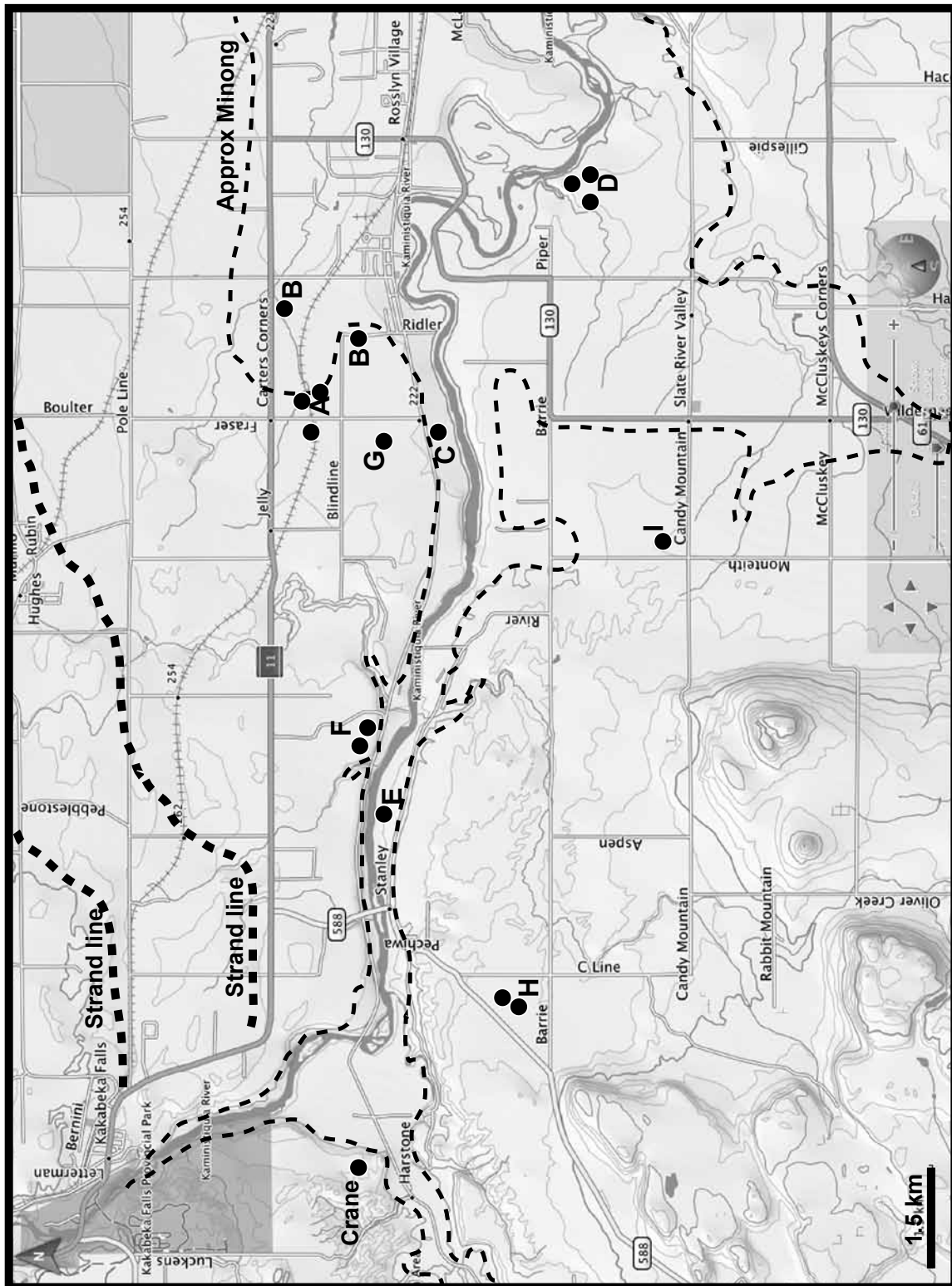


Figure 29 Overview map of selected sites within the upper Kaministiquia River Delta. Letters reference archaeological sites detailed in some of the following images. Dashed lines representing Strandlines are only approximate.

These sites, the Dairy Farm, Breukelman Evergreen and Halow A, B and C, will be viewed (Fig. 30) and then the route will follow the north side of the Kaministiquia River towards Stanley. Enroute the upper and lower Drezecky sites will be seen (Fig. 32), as well as the Pawlick site which can be seen on the south side of the river. From Stanley the route will climb the terraces of the Stanley delta and return to Hwy 11-17.

This stop provides a view north across the fields that contain a series of small lithic clusters associated with low sandy knolls around a shallow swale (Fig. 30). This swale coincides with the 750 foot contour line and suggests a shallow embayment of Lake Minong.

The following extract of text and figures derives from S. Hamilton (2000). The paper addresses the problem of developing archaeological predictive models in light of environmental transformation, and as yet incompletely understood past land use practises. Among the many challenges is the issue of development of temporally sensitive palaeo-environmental reconstructions, and also models of land use relevant to the cultures under consideration.

The apparent correlation between relict shorelines and Plano archaeological sites is well known in the Thunder Bay area where landscape evolution has been the subject of some study (Fig. 14). While many Plano shoreline archaeological sites are documented, the best known ones are lithic quarry and workshop sites such as the Cummins, Biloski, Simmonds, and McDaid Sites (Julig 1984, 1994; MacNeish 1952; Hinshelwood 1990; Phillips 1988, 1993). These large sites are well removed from modern shorelines, but are associated with the late Pleistocene shorelines of Glacial Lake Minong and Gunflint Formation bedrock exposures that are suitable for lithic tool production (Fig. 14; see Julig, McAndrews and Mahaney, 1990; Phillips 1988, 1993). However, these special-purpose sites do not characterize the full Palaeo-Indian settlement pattern. Rather, the dense recoveries from, and ready visibility of, these sites has resulted in their over-representation in the published archaeological literature. Archaeological reconnaissance upon agricultural fields within the Kaministiquia River delta has revealed a number of Plano or probable Plano sites in a wide range of landscape contexts (Hamilton, 1996) (Fig. 12). Many of these sites are found at or near Lake Minong shoreline elevations (i.e. 750 feet or 228.6 metres A.S.L.).

When placed in a hypothetical Lake Minong environmental context, they are found associated with:

- 1) springs flowing into sheltered coves of Lake Minong [Figs. 29:A, 30] (Breukelman Evergreen);*
- 2) on points of land that protruded out into Lake Minong to the northeast and southeast of the Breukelman Evergreen site cluster [Figs. 29:B, 30];*
- 3) upon low, well-drained sandy knolls surrounded by poorly drained floodplain/deltaic sediments [Figs. 29:C, 30] Halow C);*
- 4) along deltaic backwater channels [Figure 29:D, Figure 31] DbJi-8, DbJi-7, DcJi-28, DcJi-32); or*
- 5) on sandy storm beaches developed upon relict deltas [Figs. 29:D, Figure 31] DcJi-30, DcJi-31).*
- 6) raised Pleistocene terraces overlooking the present Kaministiquia River channel [Figs. 29:E, 30] Pawlick);*
- 7) on high valley rims that offer panoramic views [Figs. 29:F, 32] Drezecky E);*
- 8) along draws leading down to the Lake Minong shores [Figs. 29:G, 30] Halow A and B);*
- 9) along upland streams well removed from late Pleistocene shorelines [Figs. 29:H, 33] DcJj-12, DcJj-13); and*
- 10) upon well-drained upland knolls on what likely were formerly discontinuous permafrost uplands [Figs. 29:H, 33] Breukelman Field: lithic scatters A to E). Probable Plano lithic scatters have also been found upon isolated bedrock controlled knolls offering panoramic views of floodplains adjacent to the Lake Minong shoreline, [Figs. 29:I, 34] and at the top of the high bluff overlooking the gorge outlet of the Kaministiquia River into the shoreline floodplain of Lake Minong [Fig. 32].*

This range of landscape characteristics is certainly not exhaustive, but is sufficient to demonstrate the diverse microhabitats frequented by Plano people. These sites are consistent with other Lakehead Complex sites in terms of a very strong preference for Gunflint Formation lithic materials (Hamilton 1996). However, virtually

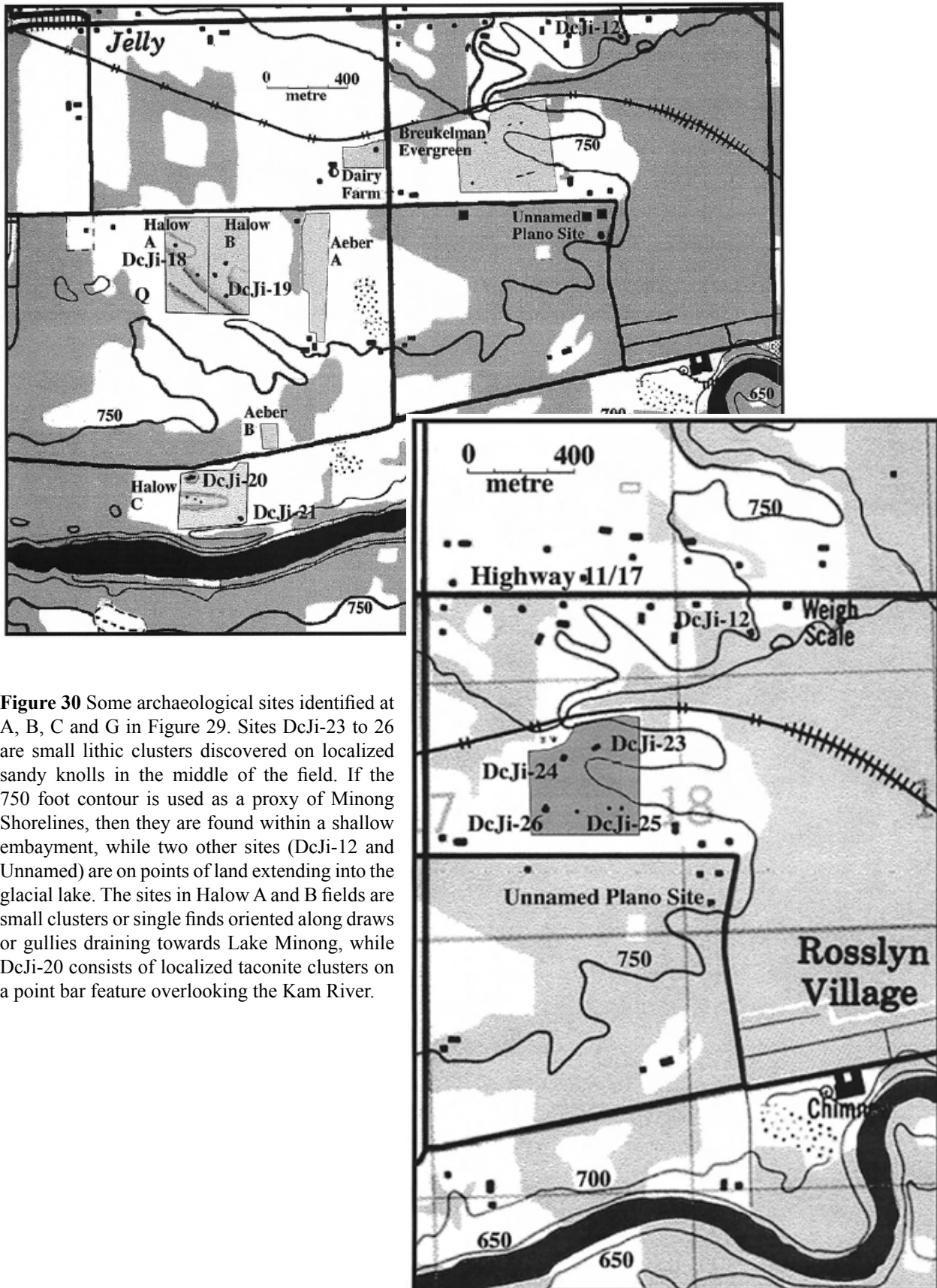


Figure 30 Some archaeological sites identified at A, B, C and G in Figure 29. Sites DcJi-23 to 26 are small lithic clusters discovered on localized sandy knolls in the middle of the field. If the 750 foot contour is used as a proxy of Minong Shorelines, then they are found within a shallow embayment, while two other sites (DcJi-12 and Unnamed) are on points of land extending into the glacial lake. The sites in Halow A and B fields are small clusters or single finds oriented along draws or gullies draining towards Lake Minong, while DcJi-20 consists of localized taconite clusters on a point bar feature overlooking the Kam River.

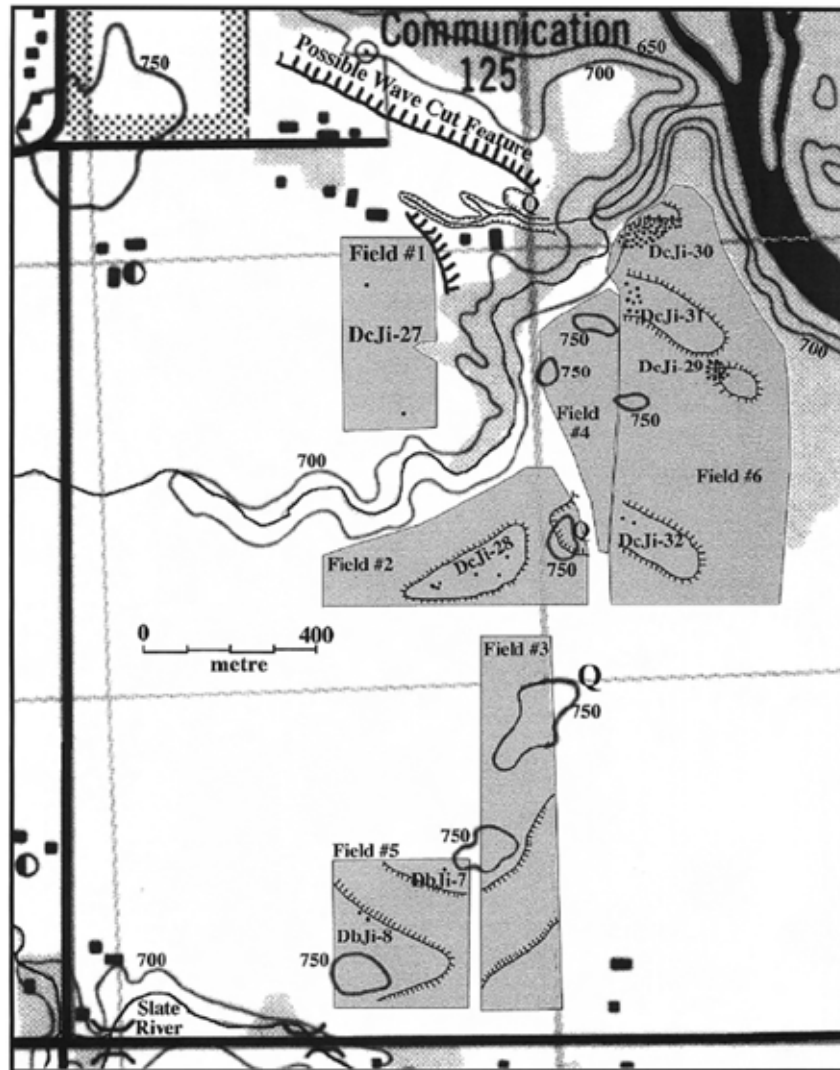


Figure 31 These sites referenced as D in Figure 29 are all located on localized sandy knolls within the cultivated field. Some are interpreted to be occupations upon a former storm beach (DcJi-29 to 30), while the others may reflect use of localized well-drained knolls within a former deltaic wetland at or immediately below the 750 foot contour.

all of these sites are well removed from bedrock exposures, are significantly smaller than the quarry-workshop sites (such as the Cummins Site), and yield many fewer artifacts. When compared to the quarry/workshop sites, these small sites also yield a much higher relative frequency of tools, preforms and utilized flakes compared to debitage (Hamilton, 1996). This indicates that the small sites represent encampments, hunting stands and food procurement sites, rather than lithic extraction and reduction stations. Such observations are hardly surprising, but they do serve as cautionary tales regarding the dangers of predicting site distribution on the basis of our current incomplete heritage inventory. These examples are also important in that they

demonstrate that modelling ancient human behaviour requires ongoing refinement, an understanding of human forager behaviour, and a well-developed sense of the nature and structure of the ancient landscape and its microhabitats."

Stop 11 Roadside stop to view Lower Beaver Bay strandlines Figure 37.

UTM coordinates: NAD83; 16U 0309392E / 5361819N

As we leave the Rosslyn delta (Minong, 9.5 ka B.P.)

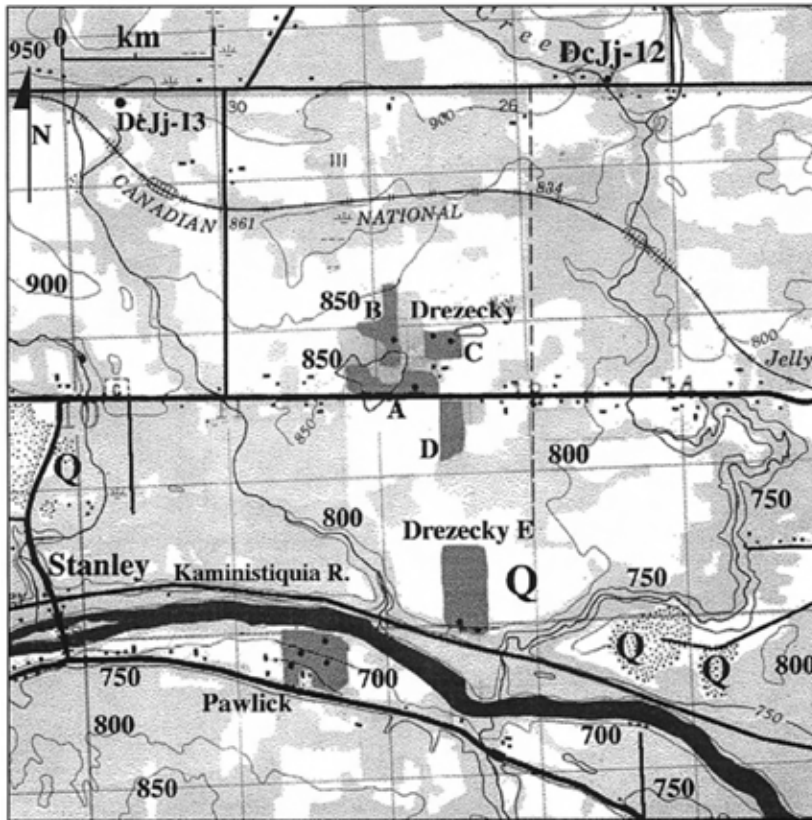
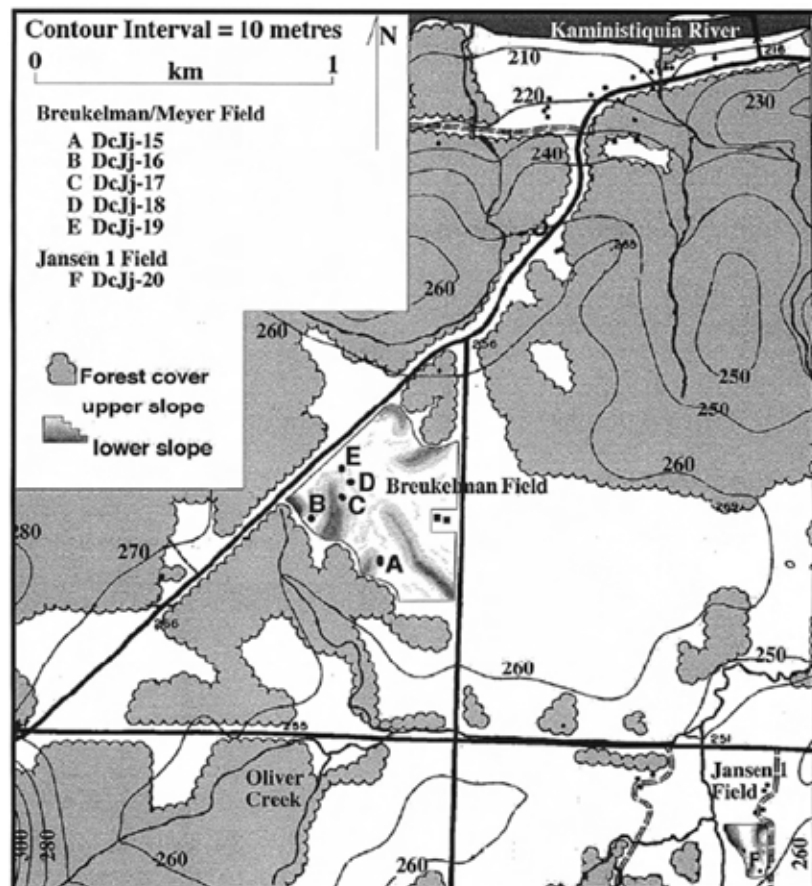


Figure 32 Location labelled F and G in Figure 29. Two small lithic scatters were encountered on the top brink of the gorge edge within Field Drezecky E. These are interpreted as game scouting sites given their commanding view across the lower gorge and Kam Delta to the south and east.

Figure 33 Location labelled H in Figure 29. The sites within the Breukelman and Meyer fields are found in localized clusters on top of sandy well-drained knolls surrounded by now-dry drainage channels. These sites are far removed from paleo-shoreline contexts and are interpreted as localized encampment zones within the interior region removed from glacial lake edges.



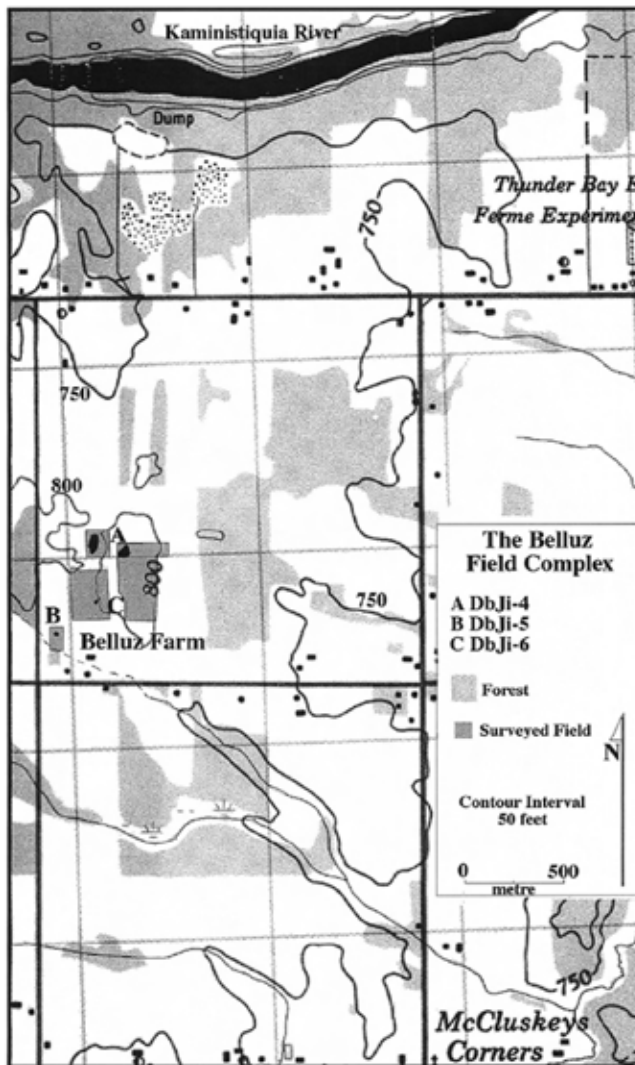


Figure 34 Location I in Figure 29. This set of small lithic clusters is located on the top brink of a localized upland facing north. It is interpreted as a hunting viewing location, whereby occupants could have watched the surrounding plains along the shores of Lake Minong (defined perhaps by the 750 foot contour line).

and move up the present Kaministiquia valley, one is in effect travelling back in time. The Stanley delta was built into Lake Beaver Bay (9.7 ka B.P.), the first Superior lake to occupy the area recently vacated by Marquette ice retreating from the Marks moraine.

Though perhaps less conspicuous, Palaeo-Indian sites of this period are also present. The larger step back in paleogeography is to consider the area that lies to the west of the maximum position of the Marks moraine, an area which remained unglaciated by the Marquette readvance and which was freed from ice around 11.0 ka B.P. Figure 35 shows the projected maximum position of the Marquette lobe and the large

area covered by the waters of a pre-Marquette lake (Early Lake Minong), now approximated by the 1400' contour. On the basis of some long known sites, both in the interior and on the Minnesota northshore, Phillips and Hill (1995) proposed that a Palaeo-Indian routeway along the Minnesota shore turned inland just east of Judge C. Magney State Park, towards North and South Fowl lakes and the Whitefish Lake region. A number of more recently discovered sites in the Whitefish - Arrow Lake corridor supports the contention that Palaeo-Indian peoples were present in the area perhaps before and during the Marquette advance, and that the corridor continued to be used into the post-Marquette periods of Lake Beaver Bay and Lake Minong. Though conjectural, there is the possibility that after the retreat of Marquette ice had begun, the Marks moraine itself provided a high ground routeway from the Whitefish Lake area along the north side of the Kaministiquia valley and into the area to the north and east of Thunder Bay.

Stop 12 - Kakabeka Falls

UTM coordinates: NAD83; 16U 0305771E / 5364428N

Just west of the junction of Hwy 11-17 and the Stanley turn-off, the huge structure of the Stanley delta is seen (Fig. 36). This delta was formed as the Kaministiquia River entered Lake Beaver Bay. A well formed bluff representing a lower Beaver Bay phase of 260m (853'), runs along the north side of the main highway, and the surface below it (250-240m; 820-787') is heavily exploited by the sand and gravel industry. The present Kaministiquia river has incised deeply into the delta, creating a terraced valley side. Kakabeka village is built on the floor (277m; 909') of an old distributary of the Kaministiquia river which cut through a higher terrace still (Fig. 36). Remnants of this terrace surface at around 300m (984') overlook the village. They represent the highest level of the Stanley delta, as it formed into Upper Lake Beaver Bay, and on the one on the west side of the Kaministiquia River, the Crane site was found. Kakabeka Falls is a major scenic attraction. It is formed as a result of a very resistant chert bed that caps the underlying, softer shales. It was just this sort of rock material that was desired for tool making. Today, a fairly spectacular gorge lies below the falls (Fig. 37). There has been some debate concerning the age of the gorge, since unless inherited and exhumed from a previous stage of river incision prior to the Marquette readvance, there is only post Beaver Bay time for it to develop its current grandeur.

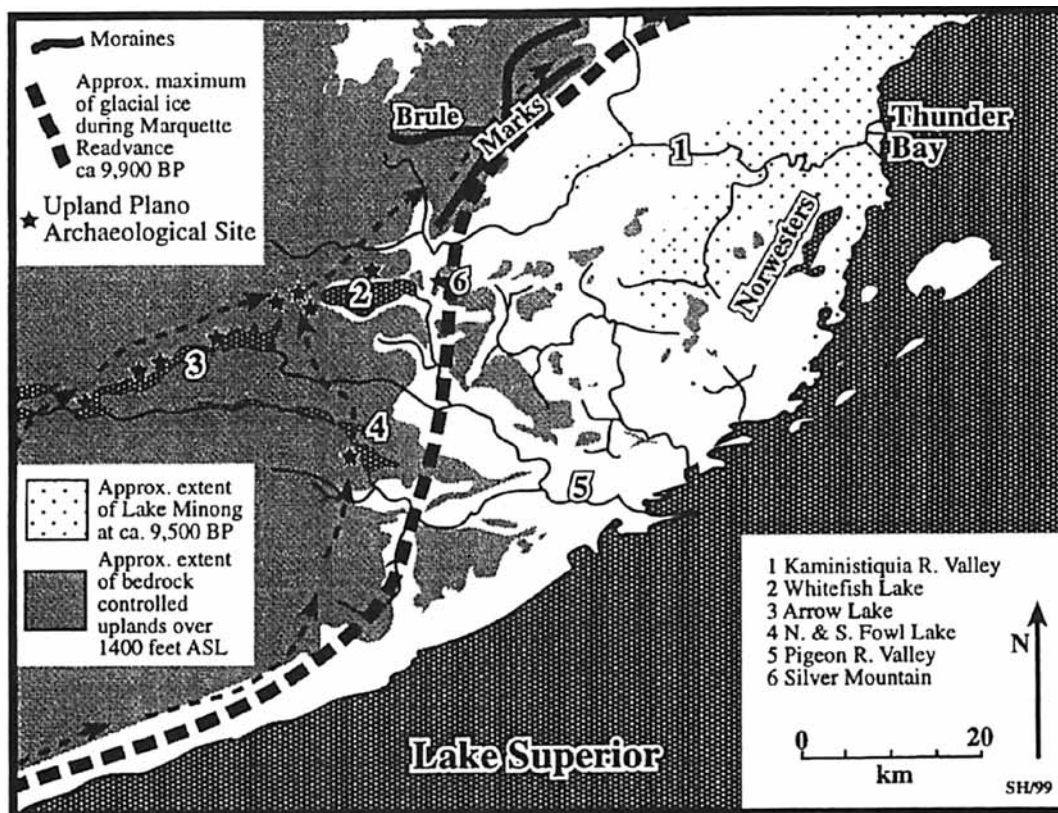


Figure 35 Plano and probable Plano sites found in the interior uplands around Arrow and Whitefish Lakes. Phillips and Hill (1995) proposed that these sites might predate or be contemporaneous with the Marquette Re-advance (Hamilton, 2000).

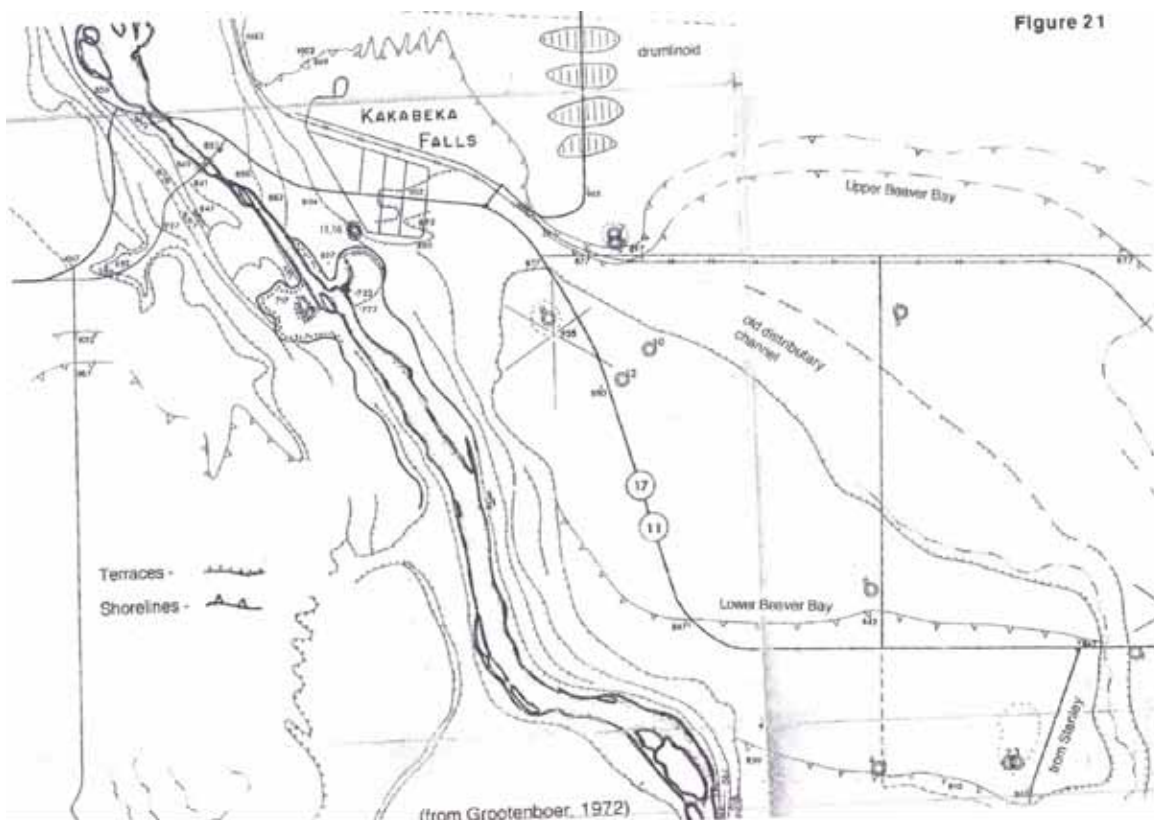


Figure 36 Strandline and other features between Stanley Corners and Kakabeka Falls.

An interesting abandoned falls and plunge pool on the west side of the gorge lies a short walk from the Park Information Centre, and represents a wide 'horseshoe' fall that would have been more spectacular than the present falls.

Stop 13 (several) - The Marks Moraine.

UTM coordinates: NAD83; 16U 0300280E / 5368976N

The route will turn on Hwy 590 just west of Kakabeka, run across the High Beaver Bay terrace remnant (look for the cemetery on the right) and due west until turning north towards the Marks Moraine. As the dirt road rises up the outer face of the Marks moraine it crosses two distinct benches at 426m (1400') and 442m (1450'). These shorelines of pro-glacial Lake Cedar Creek have been traced all along the outer slope of the Marks moraine (Jahnke, 1993) and represent still stands in a lake earlier than and isolated from the lakes of the Superior basin.

The Marks Moraine.

Stopping on the crest around 460m (1510'), the

view south over the Whitefish valley and the rugged borderland country of the mesa-like NorWesters is spectacular. It is not hard to envisage ice pushing its way up to the Marks moraine, nor is it hard to imagine pro-glacial Lake Cedar Creek occupying the narrow strip between the moraine and the retreating ice margin.

The moraine is not a simple structure. Although characterised by a till which contains diagnostic pieces of Sibley red sedimentary rocks, along much its length the Marks moraine joins and smothers isolated rock outcrops which probably greatly influenced the extent to which the ice pushed inland. The top of the moraine is remarkably flat in many places and is pockmarked with kettles. Spillway channels cross the feature in places suggesting that at one point Lake Kaministiquia to the north was a few metres higher than Lake Cedar Creek to the south. The moraine appears to have briefly existed as a narrow ridge between these two water bodies, and a possible post-Marquette routeway for animals and Palaeo-Indian people, from the unglaciated (Marquette) Arrow/Whitefish region and Interlakes corridor.

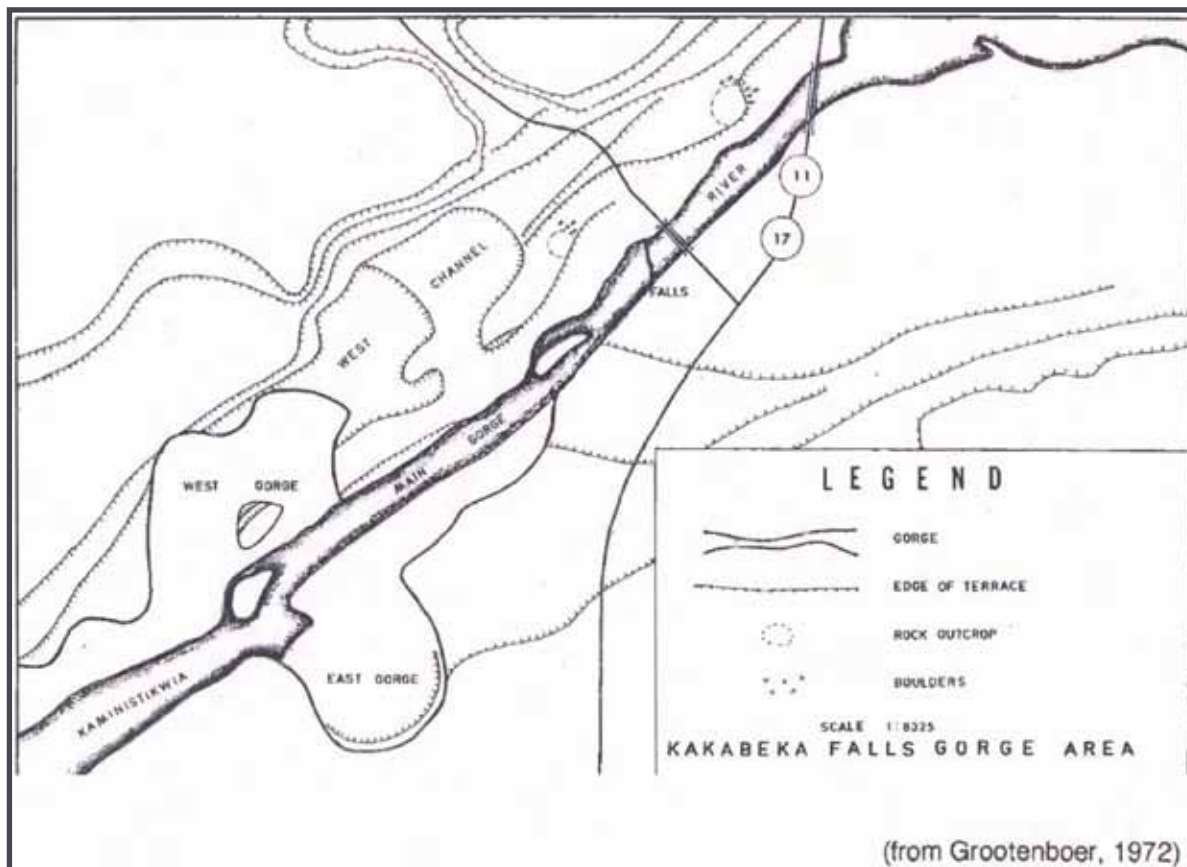


Figure 37 Kakabeka Falls and the Kaministiquia River gorge.

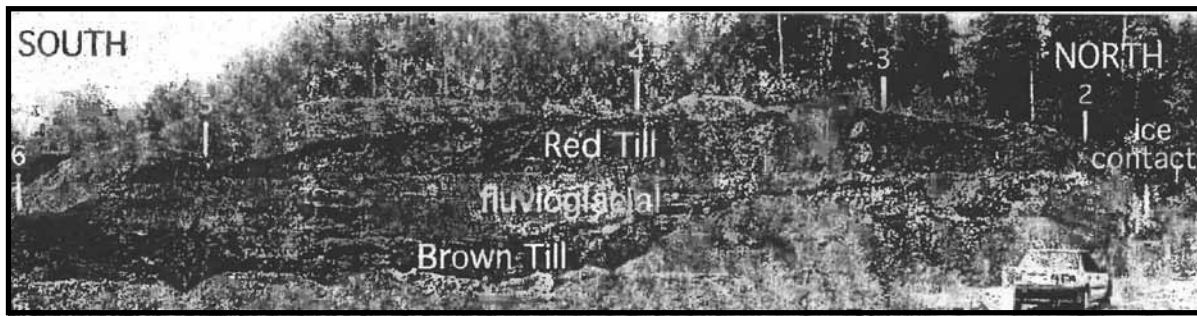


Figure 38 Cross section exposure of Conmee Pit (Tickle 1996).

The Conmee Pit and spillway.

Continuing north across the moraine and then turning east brings the excursion to a distinct channel cutting from north to south across the moraine surface. Here Conmee Township has several gravel pits. Apart from the huge boulder beds that suggest high discharges down this spillway at times, the pit reveals another key fact in understanding local deglaciation. Figure 38 shows a rare occasion when the western pit face was cleanly exposed. Tickle (1996) interpreted the sequence as consisting of two tills separated by fluvioglacial sediments. However, while one till was of diagnostic Sibley origin, the underlying one was of a northern provenance typical of the Patricia or Rainy River lobes. Elsewhere, at several places along the Marks moraine a thin Sibley till is plastered over fluvioglacial sediments. The observation suggests that the bulk of the Marks moraine may in fact be pre-Marquette in origin, with thin Marquette ice reoccupying the moraine. This has been the growing opinion of another local field worker (T. Noble, pers. communication, 1999) and looking at Figure 5, the possibility of the Marks moraine being the result of pushing the eastward part of the former Brule moraine is not unreasonable. Several exposures of contorted till structures also support this idea.

Marks Moraine scenic lookout.

Turning left out of the pit and again at the crossroad brings one to a point in the road from which a view east over the Kaministiquia valley, Thunder Bay city and the distant Sleeping Giant is obtained. The stop is useful only to impress viewers with scale of things. It is easy to imagine ice grinding past Mt. McKay and flowing up the valley, just as it is easy to imagine that a slight tilt of the present lake would bring the margin of the lake right up into the Kaministiquia and Whitefish valleys. Returning to the crossroad, the route will turn left down the face of the Marks moraine. The road is

straight, save for a jog around a kettle hole, and, lower down, rock is exposed in the fields and along the stream beds. Turning right at Hwy 11-17 brings the excursion back to Kakabeka Falls. From here the route will take Oliver Road back to the University. The road climbs out of Kakabeka over the Upper Beaver Bay terrace remnant and then runs due east to Lakehead University.

One landform of note on the way is a crag and tail feature, showing the flow of ice up the Kaministiquia valley. The 'Old Barn' restaurant lies on the tail and on the crag is the farmhouse, open for bed and breakfast (another time, eh!). Just east of Murillo but to the north of Oliver Road is the Murillo drumlin field, a series of long narrow forms parallel to the strike of the Gunflint shales, suggesting a good deal of bedrock control.

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Field trip 11 - Midcontinent Rift-Related Mafic Intrusions around Thunder Bay, Ontario

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Introduction

This field trip covers an area that has been the focus of much recent research. The guidebook has benefited from geochronologic and geochemical studies conducted as part of the Lake Nipigon Region Geoscience Initiative (Heaman et al., 2007; Hollings et al., 2007a,b, 2010) as well as recent undergraduate theses (Puchalski, 2010; Cundari, 2010; Carl, 2011) and on-going Masters Theses (Cundari, in progress). These studies have elucidated the nature of magmatism in the northern part of the Midcontinent Rift (MCR) and have been used to augment and refine our previous understanding of these magmatic events. This trip focuses on a variety of mafic intrusions associated with the MCR in and around Thunder Bay. These intrusions encompass changes in the nature of early to mid-stage MCR magmatism over a span of ~20 million years. They include Nipigon sills, Logan sills, Pigeon River dykes and the Riverdale sill. Contacts with Paleoproterozoic Rove and Gunflint formations sedimentary rocks are well-exposed in this area and illustrate some of the mechanisms of dyke/sill emplacement, as well as magma-wallrock interactions.

This guide builds upon those previously written and compiled by Franklin and Kustra (1972), Miller and Smyk (1995), Parker (2001), Miller et al. (2002) and Smyk and Hollings (2007).

Bear in mind that when visiting exploration or private properties, permission must be granted by the property owner. Current ownership information can be obtained from the Resident Geologist's Office, Ontario Geological Survey, in Thunder Bay. Please exercise caution along highway right-of-ways, near cliffs and along the lake shore.

Regional Geology

Situated within the Southern Province of the Canadian Shield, the field trip area is dominantly

underlain by Paleoproterozoic Rove Formation clastic sedimentary rocks to the south of Thunder Bay and Gunflint Formation in and to the north of Thunder Bay (Animikie Group), both of which have been intruded by MCR-related mafic intrusions. Previous mapping has been conducted by Tanton (1936a,b), Geul (1970, 1973) and Smith and Sutcliffe (1989). Detailed mapping, geophysical surveys and diamond drilling undertaken by exploration companies have provided additional detail and much-needed information about sub-surface geology. Figure 1 shows the generalized geology of the Thunder Bay area.

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

The area north of Thunder Bay displays less relief compared to the Logan Basin and has been described as peneplain by Tanton (1931). Mesoproterozoic diabase sills still provide the most dominant topographic features, as seen at the Silver Harbour Quarry (Stop 1), The Bluffs (Stop 2) and Mount McKay (Stop 3).

Animikie Group

Paleoproterozoic Gunflint and Rove sediments were deposited in the Animikie Basin, forming a southward-thickening wedge covering the southern margin of the Superior Province, which is truncated in east-central Minnesota and northern Wisconsin by Penokean magmatic terranes. Gunflint sedimentation

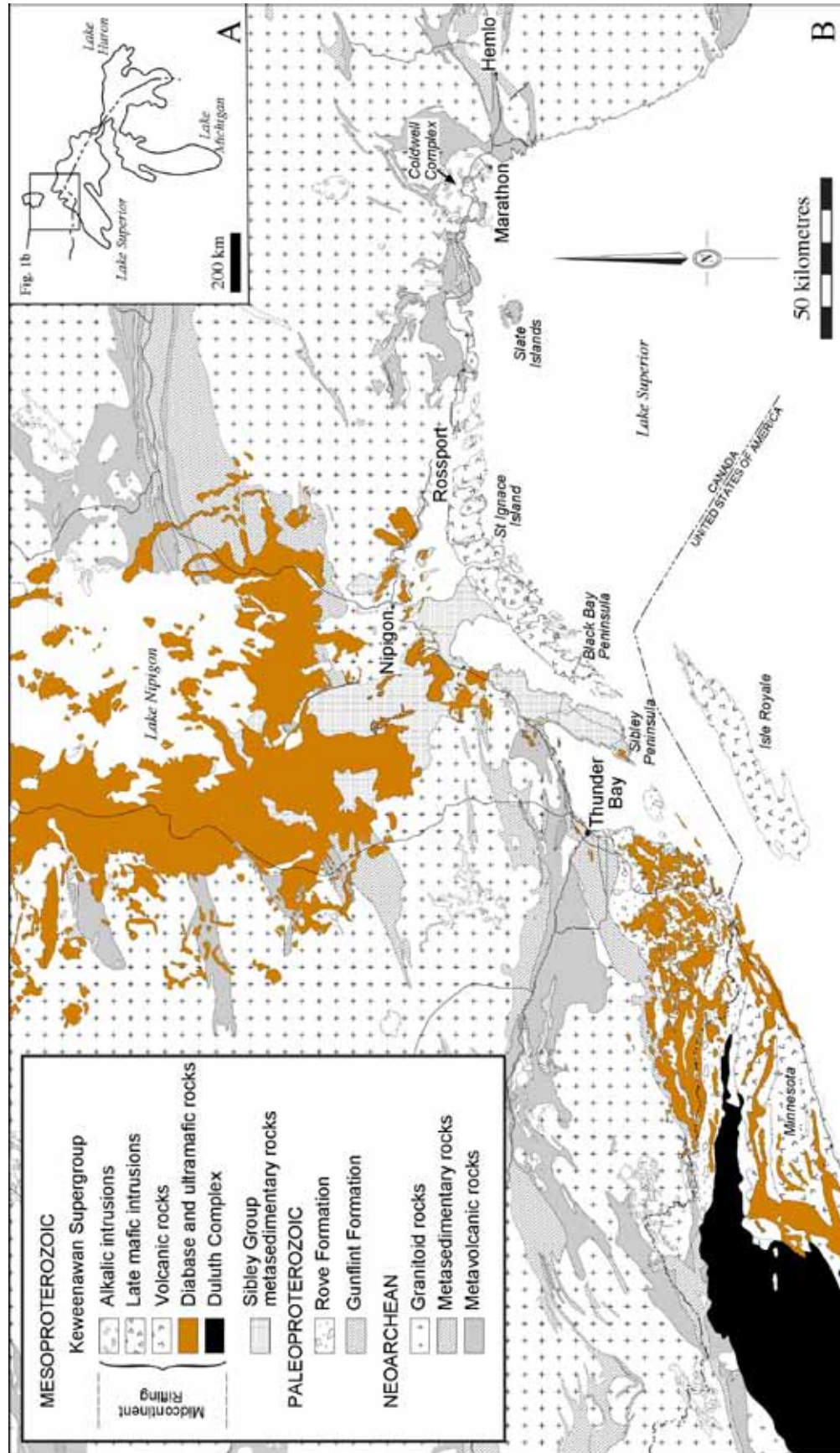


Figure 1: Simplified geological map of the Thunder Bay area. Modified after Pye and Fenwick (1965) and Carter et al. (1973).

began approximately 2.1 Ga and ceased approximately 1.85 Ga, prior to, or during the Penokean orogeny. The nature of the sediment varies considerably, ranging from volcanic through clastic to chemical precipitates which form thick successions of iron formation.

The Rove Formation is a turbidite-dominated shelf sequence, which overlies the Gunflint Formation (1878 $\pm$ 2 Ma; Fralick et al., 1998). It consists of a lower section of black, locally pyritic shales, which grades upwards into shales, interbedded with wackes. These clastic rocks were deposited by southeastward-moving turbidity currents, shed from the Archean craton to the north. The Rove has an approximate thickness of 500 to 600 m south and east of Thunder Bay, and thickens to the south. Rocks of the Rove Formation are flat-lying or dip gently to the southeast. The shales are thin-bedded, dark and fissile. Recent work by Amurawaiye (2001) and Maric and Fralick (2005) described a submarine ramp system in which the movement of coarse sediments into the deeper parts of the basin was mainly through the action of low- and high-density turbidity currents. Fair-weather and storm-generated currents dominated depositional activity at the edge of the basin. Amurawaiye (2001) concluded that approximately 70% of the Rove Formation locally consisted of organic shale.

The lower 100 to 150 m of the Rove Formation and the correlative Virginia Formation in Minnesota, consist of alternating shale-siltstone and black, pyritiferous shale successions, probably reflecting fluctuations in sea level (Maric and Fralick, 2005). These successions, and especially the upper black shale, likely represent a major condensed interval deposited in water $\sim$ 100 to 200 m deep. Lucente and Morey (1983) ascribed sedimentation of this interval to pelagic rainout of fine-grained sediment from dilute suspension or hemipelagic processes involving diffuse turbidity currents. The presence of abundant, sub-millimeter rip-up intraclasts also denotes the operation of sporadic bottom currents (Maric and Fralick, 2005). Tidal deposits present in correlative rocks to the south of Lake Superior confirms open connection to the ocean (Ojakangas et al., 2001). Above the upper, pure black shale interval, graded fine-grained sandstones are organized into a coarsening-upward succession approximately 100 m thick that is transitional into 400 m of medium-grained, sandstone-dominated, stacked parasequences (Maric and Fralick, 2005). This is overlain by lenticular to wavy bedded sandstones and shales with both wave and current ripples. The coarsening-upward to sandstone-dominated portion

of the Virginia and Rove Formations has been interpreted as a submarine fan (Lucente and Morey, 1983; Maric and Fralick, 2005) with the uppermost ripple laminated succession representing progradation of distal distributary mouth bars of a delta (Maric and Fralick, 2005). A sandstone sample from the submarine fan portion of the succession yielded a youngest U-Pb detrital zircon age of approximately 1780 Ma (Heaman and Easton, 2006). The predominantly Paleoproterozoic zircon population and paleocurrents indicating sediment derivation from the north, strongly suggest the Trans-Hudson Orogen was the source of the detritus (Morey, 1973).

Keweenawan Supergroup

Mesoproterozoic intrusive, volcanic and minor sedimentary rocks associated with the MCR collectively constitute the Keweenawan Supergroup. On the northern margin of the MCR, Keweenawan rocks include a variety of intrusive rocks and Osler Group volcanic rocks, which represent some of the earliest magmatism in the MCR. As shown in Table 1, ages range from ca. 1140 Ma (Heaman et al., 2007) to ages younger than the magnetic polarity reversal that occurred between 1105 and 1102 Ma (Davis and Green, 1997). A tabulated synopsis is provided below; bolded units occur within the field trip area.

The majority of mafic and ultramafic rocks in the Lake Nipigon and northern Lake Superior areas, including the Nipigon and Logan sills, appear to have been emplaced in a short, magnetically reversed, interval between ca. 1115 and 1100 Ma (Heaman et al., 2007). Emplacement of alkalic intrusions, such as the 1108 Ma Coldwell Complex (Heaman and Machado, 1992), and filling of much of the submerged part of the rift in Lake Superior, also occurred in this period. This was followed by a period of magnetically normal, waning mafic and felsic magmatism, between 1096 and 1085 Ma, that is preserved mainly along the Lake Superior shore by units such as the Crystal Lake (1099 $\pm$ 1 Ma), Moss Lake (1095 $\pm$ 2 Ma) and Blake (1095 $\pm$ 2 Ma) gabbros, and a Pigeon River dyke near Arrow River (1093 $\pm$ 3 Ma; Heaman et al., 2007).

Hypabyssal Mafic Rocks

Diabase sills, extending from the vicinity of Thunder Bay to east of Lake Nipigon, represent the northern remnants of the Midcontinent Rift, and have previously been referred to as the Logan sills (Stockwell et al., 1972), however, recent work suggests

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

Lithologic Unit	Locality / Age (Ma)	Reference(s)
St. Ignace Island Complex gabbro	St. Ignace Island / 1089.2 ± 3.2	Smyk et al.(2006)
Arrow River Dyke	Arrow River / 1078 ± 3	Heaman et al. (2007)
Pigeon River Dyke	Rita Bolduc / 1141 ± 20	Heaman et al. (2007)
Blake Gabbro	Blake Township / 1091.0 ± 4.5	Heaman et al. (2007)
Moss Lake Gabbro	Black Bay Peninsula / 1094.7 ± 3.1	Heaman et al. (2007)
Crystal Lake Gabbro	Great Lakes Nickel / 1099.6 ± 1.2	Heaman et al. (2007)
Mt. Mollie Dyke	1109.3 ± 6.3	Hollings et al. (2010)
Cloud river Dyke	1109.2 ± 4.2	Hollings et al. (2010)
Osler		
Osler Group rhyolite (central suite)	Agate Point / 1105 ± 2	Davis and Green(1997)
Osler Group rhyolite (lower suite)	Black Bay Peninsula / $1107.4 \pm 4/-2$	Davis and Sutcliffe (1985)
St. Ignace Island Complex Rhyolite	St. Ignace Island / 1107.2 ± 2.4	Smyk et al.(2006)
Coldwell Complex	Coldwell Complex / 1108 ± 1	Heaman and Machado (1987)
Logan Sills	Mt. McKay / 1114.7 ± 1.1	Heaman et al. (2007)
Nipigon Sills	Nipigon Embayment / 1114-1110	Heaman et al. (2007)
Ultramafic Intrusions	Nipigon Embayment / 1124-1113	Heaman et al. (2007)
Inspiration Sill	Lake Nipigon / 1141 ± 20	Heaman et al. (2007)
Marathon lamprophyre dykes	McKellar Harbour / $1145 \pm 15, -10$	Queen et al. (1996)

a geochemical difference between the sills to the north and south of the City of Thunder Bay (Hart, 2003; Hart et al., 2005). Hollings et al. (2007a) proposed that the term Logan Igneous Suite, which would fall within the Midcontinent Rift Intrusive Supersuite (Miller et al., 2002), should be applied to all the diabase sills in the area north of Lake Superior, with subdivision into the informal terms, Nipigon sills for the sills north of Thunder Bay, and Logan sills to the south.

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

Logan sills are recognized by their reversed magnetic polarity and generally take the form of columnar-jointed, thick sheets and sills whose geometry is

strongly controlled by the subhorizontal bedding of the country rock. They form conspicuous erosional remnants that create mesa and cuesta topography. From the international boundary area to Thunder Bay as many as six diabase sheets were emplaced nearly conformably into Animikie sedimentary rocks (Weiblen et al., 1972; Smith and Sutcliffe, 1987, 1989). Diamond drilling has also shown that stacked sills exist in the subsurface. For example, Dumont Nickel Inc. reported intersecting 14 gabbroic sills in a 705 m deep drill hole in central Pardee Township (Assessment Files, Thunder Bay South Resident Geologist’s District, Thunder Bay). North of the border, Smith and Sutcliffe (1989) reported sills up to 50 m thick, whereas in Minnesota, Jones (1984) studied four sills ranging from 50 to 160 m in thickness. Rare exposures of feeder dykes to sills and preserved sill terminations have been noted.

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

Nipigon sills are commonly massive, medium- to coarse-grained, olivine-tholeiitic diabase/gabbros (Sutcliffe, 1989; Hart and MacDonald, 2007). Nipigon sills are dominantly present throughout the Lake Nipigon area but have also been recognized in the Thunder Bay area (Hollings et al., 2007b). Nipigon sills are characterized by a massive, subophitic to ophitic, plagioclase and clinopyroxene texture with trace to 3% olivine and 1-2% modal magnetite (Hart et al., 2005).

Nipigon sills display a reverse magnetic polarity and generally form thick, columnar jointed sheets. Sills commonly intrude Sibley group sedimentary rocks but also can be found in contact with Archean rocks of the Quetico subprovince and the Marmion and Winnipeg River terranes. Sills often intrude earlier emplaced ultramafic units of the Nipigon Embayment as well as the 1129.0 ± 2.3 Ma Pillar lake Volcanic rocks and the 1129.0 ± 2.3 Ma English Bay Complex (Heaman et al., 2007) providing evidence for their emplacement during the second main phase of magmatism (Hart and MacDonald, 2007). The shallow dipping Nipigon diabase sills are estimated to cover an area in excess of 20 000 km² (Sutcliffe 1991) ranging in thickness from <5m to >180m (Hart and Macdonald, 2007).

Pigeon River dykes trend east-northeast to northeast and dip steeply to the southeast (Geul, 1970, 1973; Smith and Sutcliffe, 1989). Displacement and warping of the Rove Formation is evident along many of the dykes. Composite intrusions are noted in several dykes. Dyke widths average between 50 and 70 m, but may be as much as 150 m across in Ontario (Smith and Sutcliffe, 1987) and 500 m in Minnesota (Green et al., 1987). Forming northeast-trending, linear ridges, dykes can be traced semi-continuously for 15 km along strike. As noted by many workers, some dykes clearly crosscut Logan sills. However, Geul (1973) and Smith and Sutcliffe (1987) noted that others display somewhat ambiguous crosscutting relationships. In these latter cases, dykes may appear to merge with sills, suggesting that they were contemporaneous or that sills impeded the upward migration of the dykes. The presence of multiple sets of horizontal columnar jointing suggests the development of multiple or composite dykes.

The dykes typically consist of ophitic diabase that may be plagioclase-porphyritic. A typical, non-porphyritic olivine diabase consists of approximately 60% plagioclase (zoned labradorite; An_{55-70}), 20% augite + hypersthene, up to 15% olivine and up to 5% magnetite, ilmeno-magnetite and sulphides (Geul 1970, 1973). Average whole rock compositions of

Pigeon River dikes are moderately evolved ($Mg\# = 52$) olivine tholeiitic basalt.

The Riverdale sill was first characterized by Hollings et al. (2007b) as being geochemically and petrographically distinct from the surrounding Logan sills and the Nipigon sills to the north. Puchalski (2010) described the geochemical and petrographical characteristics. The Riverdale sill lies within the southern city limits of Thunder Bay, close to the northern boundary of the Logan basin. The unit displays a sill morphology exposed over an area approximately 6 km long and 2 km wide with true thickness unknown as the upper contact is not exposed (Puchalski, 2010). Exposures within a quarry on West Riverdale Road (Stop 4) display a thickness of 10 m where detailed sampling and subsequent geochemistry and petrography analyses were completed. Paleomagnetic work performed by Hollings et al. (2010) confirmed the unit to display a reverse polarity. The following is summarized from Puchalski (2010).

Rocks comprising the Riverdale sill are dominantly gabbronorites with lesser olivine gabbro present towards the centre of the intrusion. The gabbronorites are generally fine-grained and display no cumulate textures within any of the samples. Plagioclase typically occurs as subhedral laths with euhedral orthopyroxene and lesser clinopyroxene and olivine. Minor alteration is present in most samples as chlorite replacing pyroxene and sericite replacing plagioclase. The olivine gabbro samples lying toward the centre of the unit are petrographically similar to the gabbronorite samples except for a higher modal percentage of anhedral to euhedral olivine. Olivine grains are typically fine-grained but may range to medium grained. Variable amounts of serpentine is found replacing olivine.

A unit of mafic rock in Devon Township, south of Thunder Bay, was mapped by Tanton (1931) and was termed Rove Formation Basalts, but was subsequently mapped as a Logan diabase sill (Geul, 1970). Cundari (2010) described the detailed geological, petrographical and geochemical characteristics of the unit. The unit is exposed on a plateau 7 km long and 0.8 to 1.0 km wide. The unit is 4 to 6 m thick and is in apparent conformable contact with the underlying shales of the Paleoproterozoic Rove Formation, where a pronounced chilled margin consists of variolitic material up to 20 cm thick. The flow-top also exhibits a variolitic texture ~15 cm thick. The presence of ropy flow top and amygdules as well as quench textures, support a volcanic origin.

Major element chemistry reveals a tholeiitic, intermediate composition with samples plotting in the basaltic andesite to andesite fields as well as in the basaltic trachy-andesite to trachy-andesite fields on a TAS diagram. The unit typically has an intergranular texture consisting of randomly oriented plagioclase laths with interstitial chlorite, an alteration product of primary augite. Most samples contain minor serpentine (after olivine), opaque minerals, secondary quartz, oxides, pyrite and calcite. Amygdules are often present and are infilled with some combination of calcite, quartz, chlorite and pyrite. Lower flow contacts and flow-tops are typically glassy with abundant spherulites that sometimes coalesce into bands. This unit is definitively related to Midcontinent rift magmatism and is now referred to as the Devon Volcanics.

Contact Metamorphism

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

Rove Formation sedimentary rocks may be deformed along dyke contacts. As noted by Geul (1973) beds appear to dip toward the dykes or are "up-dragged" along dyke contacts. Deformed and fractured sedimentary rocks have been noted near sill terminations. Narrow, parallel tension gashes filled with quartz-feldspathic leucosome/neosome occur in metatectic, deformed siltstones.

Assimilation of country rock

Pink granophyric features have been well-documented in Logan sills to the west of Thunder

Bay which are attributed to in-situ assimilation of granitic material (Blackadar, 1956). This theory was reevaluated by Magnus (2010) who studied assimilation features in the Navilus and Terry Fox sills. Although these outcrops will not be visited on this field trip, they provide a sound explanation for some of the assimilation and sill-top features observed throughout the trip (i.e., Stops 2 and 5).

Two zones were reported to consistently appear around xenoliths present in the Navilus sill: a zone of quartz-feldspathic intergrowths, or "granophyre", adjacent to xenoliths, followed by a zone of pyroxene grains present on the interface between normal diabase magma and the granophyric zone (Magnus, 2010). Late-stage granophyric formations are also found interstitially between plagioclase and pyroxene grains with iron-oxides, likely representing immiscibility of a late-stage silica- and iron-rich liquid exsolved from the magma (Magnus, 2010). This premature exsolution of silica and iron-rich liquids from the magma are attributed to the introduction of silicate-rich xenoliths to the already silica-saturated, quartz-tholeiitic magma (Magnus, 2010).

Magnus (2010) further concluded that geochemical variation between diabase with inclusions and normal diabase was caused by late-stage fractionation during crystallization as noted by variable depletion in high field-strength elements (HFSEs) with LIL enrichment attributed to in-situ assimilation.

Discussion of Geochemistry

As part of the Lake Nipigon Region Geoscience Initiative, whole rock analyses were performed on a number of Midcontinent Rift-related intrusions south of Thunder Bay as well as the Lake Nipigon region. Subsequent research conducted at Lakehead University provided additional sampling of later-recognized units of interest (i.e., the Riverdale sill and the Devon Volcanics). Data from these studies, as well as data from previous mapping endeavors conducted by the Ontario Geological Survey, have been compiled in a database totaling 2400 spatially referenced points with whole-rock geochemical analyses. This database is currently being reevaluated by Cundari (in progress) to detect variability within units as well as put many of the obscure units in the context of the MCR.

Current discrimination of intrusive units associated with the MCR is through trace element patterns (e.g., primitive mantle-normalized spider plots, as well as measures of heavy and light rare-earth

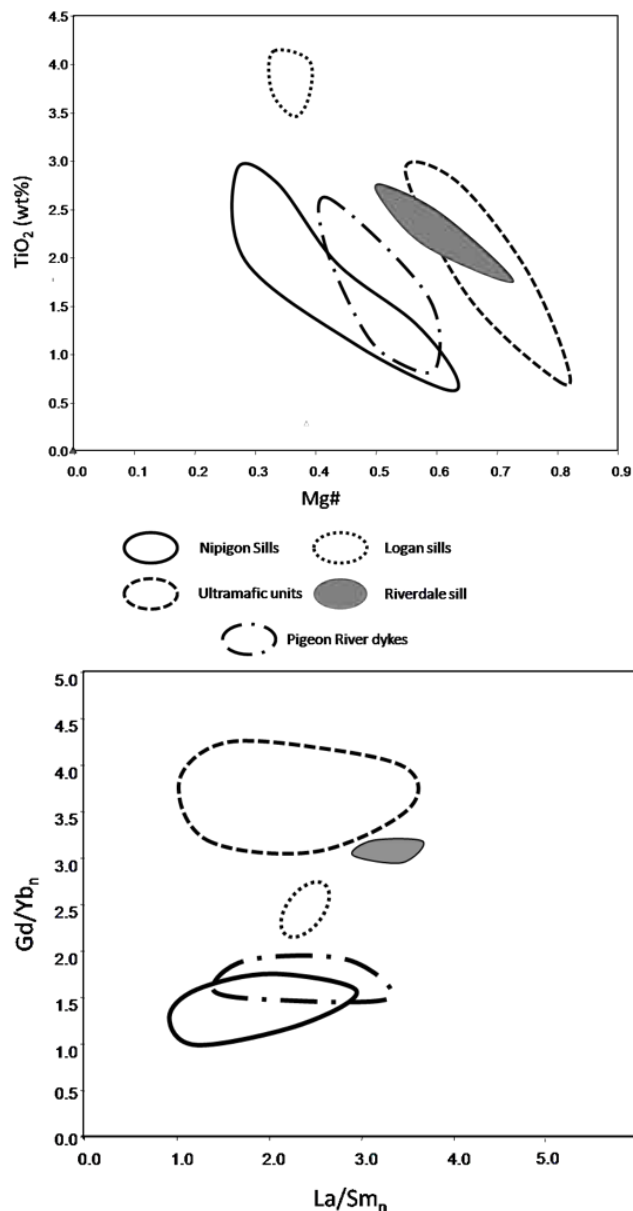


Figure 2: Discrimination diagrams for mafic and ultramafic intrusions near Thunder Bay. Data are from Hollings et al. (2007a) and Puchalski (2010). Normalizing values from Sun and McDonough (1989).

element abundances displayed by the plot of La/Sm_n (LREE) and Gd/Yb_n (HREE; Fig. 2). Major element abundances, i.e. $\text{Mg\#}$ vs. TiO_2 , also show distinct populations between units (Fig. 2; Table 2).

Nipigon and Logan sills show broadly similar morphological characteristics but can be distinguished from each other based on TiO_2 abundances. The Logan sills present to the south of Thunder Bay within the Logan Basin, (e.g., Mt. McKay, Stop 5) display higher TiO_2 content than the Nipigon sills (Stops 1 and 2)

present within the city of Thunder Bay and north around Lake Nipigon. This geographic distribution may be a result of either tapping different source regions at depth or the presence of a major compositional boundary (Hollings et al., 2007a).

Geochemical and petrographic data show no evidence for fractionation within the Riverdale sill, with only slight variation present towards the lower margin and the centre of the sill. Samples within 1 m of the contact display negative niobium anomalies interpreted to be a result of crustal contamination at depth, with samples towards the centre characterized as olivine gabbros. This suggests that the sill is composed of two pulses of magma, with the more-contaminated first pulse intruded by a less-contaminated second pulse (represented by the olivine gabbro pushing the contaminated magma towards the outer margins of the sill). The Riverdale sill is geochemically distinct based on its heavy rare-earth element abundances when compared to the surrounding Logan sills. Displaying a high Gd/Yb_n ratio denotes heavy rare earth element fractionation indicative of a deep-seated mantle melt sourced from below the garnet-spinel stability field (>100 km). As the Riverdale sill displays a Gd/Yb_n ratio of 3.0 – 3.5, it was likely sourced from this region suggesting it is genetically related to the ultramafic intrusions of the Nipigon Embayment (Puchalski, 2010).

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

Rare-earth element geochemistry of the Devon Volcanics show the unit to be relatively enriched in both HREEs and LREEs, similar to the ultramafic sills of the Nipigon Embayment as well as the Riverdale Sill (Hollings et al., 2007a, 2010). A primitive mantle-normalized REE plot shows that the volcanic unit is characteristic of an Ocean-Island Basalt, but with a negative niobium anomaly, most likely the result of lower crustal contamination. This evidence is further supported by an $\epsilon_{\text{Nd}(t=1100\text{Ma})}$ of -3.48, which also suggests contamination of the unit by a lower crustal source.

Table 2: Geochemical analysis of select intrusive rocks from the Thunder Bay area. Data from: A) Hart and Magyarosi (2004), B) Hollings et al. (2011).

	1	2	3	4	5	6	7	8
Intrusive unit	Logan sills		Nipigon sill	Hornfelsesd Rove	Riverdale sill	Riverdale sill	Pigeon River dykes	Pigeon River dykes
Description	Lower sill (upper contact)	Upper sill	Silver Harbor quarry	Squaw Bay road quarry	Olivine gabbro	Gabbro-norite	Whiskeyjack point N	Whiskeyjack point S
Sample	03TRH201	03TRH202	DB-167	DB-12	RP-8QB	RP-5Q	DB-10	DB-11
Source	A	A	B	B	B	B	B	B
SiO ₂	48.36	48.86	48.32	76.56	46.4	48.64	53.14	50.62
TiO ₂	3.65	3.59	0.91	0.45	2.04	2.94	1.23	1.22
Al ₂ O ₃	13.74	14.46	15.45	10.48	8.16	9.89	14.62	14.31
FeO ^t	14.62	15.00	11.73	4.88	12.96	13.94	10.90	11.32
MnO	0.21	0.18	0.19	0.02	0.16	0.2	0.18	0.19
MgO	4.4	4.05	7.88	1.48	12.53	7.14	5.49	6.58
CaO	7.12	7.45	10.85	0.2	9.24	8.51	9.75	8.65
Na ₂ O	2.81	3.43	2.11	2.11	1.51	3.09	2.64	2.85
K ₂ O	1.35	1.27	0.4	1.52	0.41	0.8	0.91	1.57
P ₂ O ₅	0.37	0.42	0.09	0.08	0.21	0.24	1.15	0.12
Volatiles	1.44	0.37	0.57	2.78	3.87	3.29	1.15	1.74
Total	99.7	100.76	99.81	101.09	98.93	100.24	101.38	100.45
mg#	23.13	21.26	40.18	23.28	49.16	33.87	33.50	36.76
(ppm)								
Cr	30	41	112	190	>1300	266	68	11
Co	34	36	62	12	93	67	45	47
Ni	81	88	131	35	406	92	95	95

The trace element characteristics of the volcanic unit suggest an origin in Keweenawan time as they are geochemically similar to units of the MCR (Hollings et al., 2007a) rather than Paleoproterozoic volcanic units of the Gunflint Formation.

Field trip stops

Stop 1: Silver Harbour Quarry

UTM coordinates: NAD83; 16U 0354388E / 5374970N

Location: Quarry adjacent to road cut at Silver Harbour boat launch. Silver Harbour Road off Lakeshore Drive.

Description: The first stop on the trip offers an excellent exposure of a Nipigon sill. Material was quarried from this locality to create many of the break-walls along this portion of the bay. Two localities are

of interest here: the first being the actual quarry which exposes a typical Nipigon diabase sill; the second is a road cut at the northern end of the quarry which displays some enigmatic, late-stage features. The road cut can be reached by a trail along the northwestern side of the clearing.

The features observed in Figures 4 and 5 appear to represent magma injected into the still-crystallizing sill. They are typically finer-grained than the surrounding medium- to coarse-grained, subophitic diabase. They appear as blobby masses or dykes with undulating contacts appearing to propagate up through the sill from depth. The amorphous form suggests that the surrounding sill material was not fully crystallized when these features were emplaced but must have been partially solid, as distinct contacts are preserved. These features may represent an episode of immiscible magma interaction.

Field trip stops - Coordinates and Descriptions

STOP NAME	STOP NUMBER	Description	NORTHING	EASTING
			(NAD 83)	
Silver Harbour Quarry sill	1	Quarry and adjacent road cut at Silver Harbour boat : launch	354388 E	5374970 N
The Bluffs	2	Flat-lying outcrop atop The Bluffs lookout; off Arundel St.	337266 E	5371048 N
Baked Rove outcrop	Optional	Sandstone quarry	333765 E	5354546 N
Pigeon River dykes (Chippewa park/Whiskeyjack point)	3	Two parallel, E-trending dykes cross-cutting Rove Formation sedimentary rocks; Whiskeyjack point, Lake Superior	336798 E	5355397 N
Riverdale sill (Robin's Donuts)	Optional	Easternmost exposure of Riverdale sill in Robin's donuts parking lot; Highway 61	326487 E	5355715 N
Riverdale sill Quarry	4	Riverdale sill exposed in contact with Rove sedimentary rocks; Candy Mountain Dr.	322410 E	5355212 N
Mount McKay	5	Mount McKay lookout on top of lower sill; hike to upper sill via trail is optional	331126 E	5357384 N



Figure 3: Satellite image of the Thunder Bay area showing field trip stops.



Figure 4: Photo of outcrop at Silver Harbour road cut showing immiscibility features. South side of road

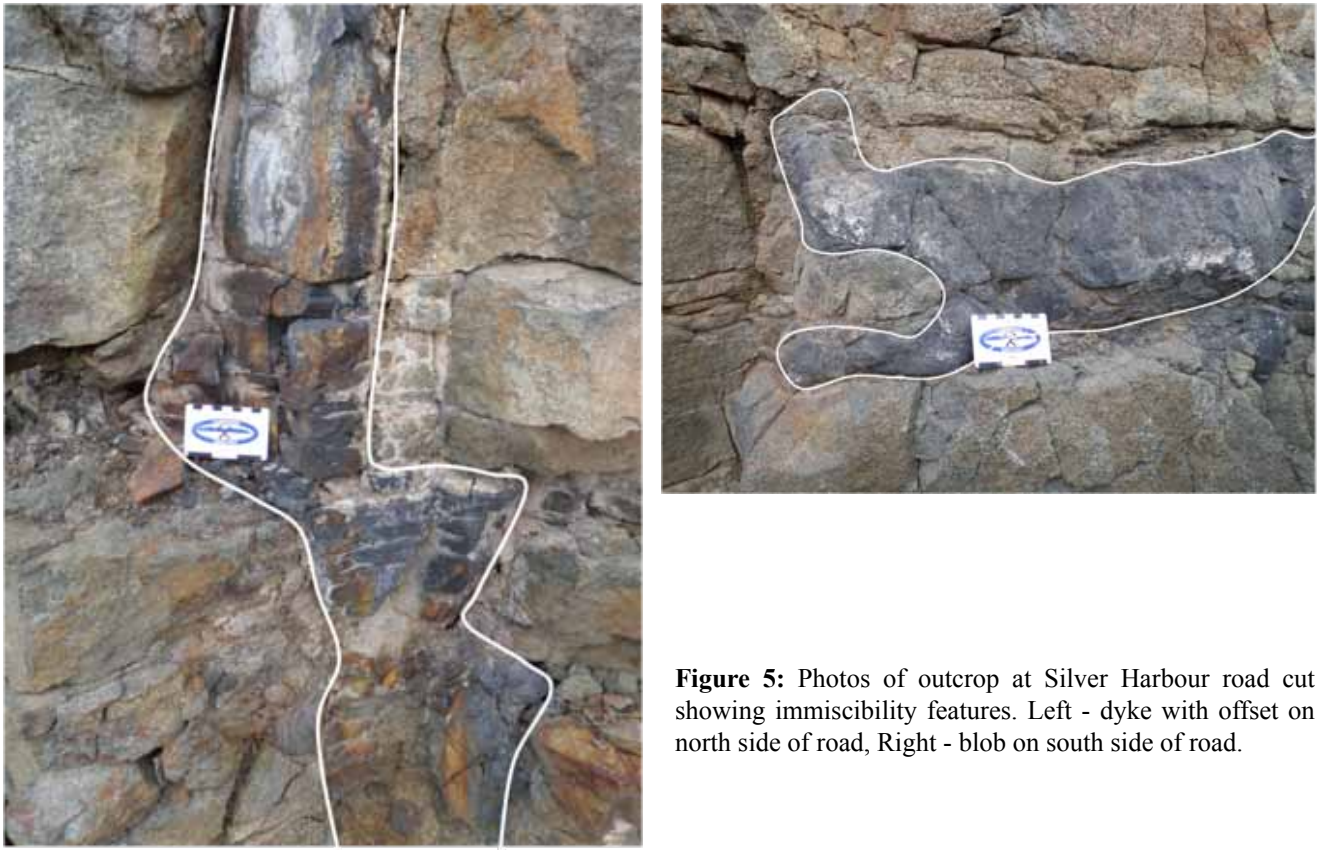


Figure 5: Photos of outcrop at Silver Harbour road cut showing immiscibility features. Left - dyke with offset on north side of road, Right - blob on south side of road.

Samples SP-RC-016 and SP-RC-018 from the younger intrusion have higher silica contents as well as larger loss on ignition values (Table 3) suggesting that the late-stage material was contaminated in silica and hydrous minerals possibly from the surrounding sedimentary rocks of the Gunflint Formation. This is further supported by the elevated light rare-earth elements abundances of these samples compared to the host sill (Fig. 6). Alternatively, the later-stage material may represent a slightly more fractionated product of a typical Nipigon sill melt. This is supported by the pronounced negative europium and titanium anomalies (Fig. 6) representing plagioclase and magnetite fractionation, respectively. The most plausible scenario is that the material injected into the sill underwent both processes whereby the material has residency time in

a shallow crustal magma chamber. Here it was able to fractionate plagioclase and magnetite as well as leach hydrous material including silica from the Gunflint Formation sedimentary rocks, resulting in the rare-earth element enrichment, elevated silica content and loss on ignition.

Stop 2: The Bluffs lookout

UTM coordinates: NAD83; 16U 0337266E / 5371048N

Location: The Bluffs lookout. Unmarked road off Arundel St. west of Lyon Blvd. W.

Description: This stop provides a good vantage point for the city of Thunder Bay and Lake Superior, looking southeast. The pronounced topographic high

Table 3: Major element chemistry for samples at Silver Harbour road cut. Data from Cundari (in progress)

Sample		SiO ₂	TiO ₂	Al ₂ O ₃	FeO	MnO	MgO	CaO	Na ₂ O	P ₂ O ₅	LOI	Total
SP-RC-016	Late-stage bleb	64.75	0.51	11.97	4.98	0.09	2.66	3.59	3.5	0.11	5.6	100.28
SP-RC-017	Nipigon sill	48.86	0.71	14.91	11.03	0.18	8.02	9.07	1.61	0.06	2.67	100.43
SP-RC-018	Late-stage dyke	58.03	0.5	10.9	10.92	0.15	4.17	1.54	0.81	0.08	8.83	99.52

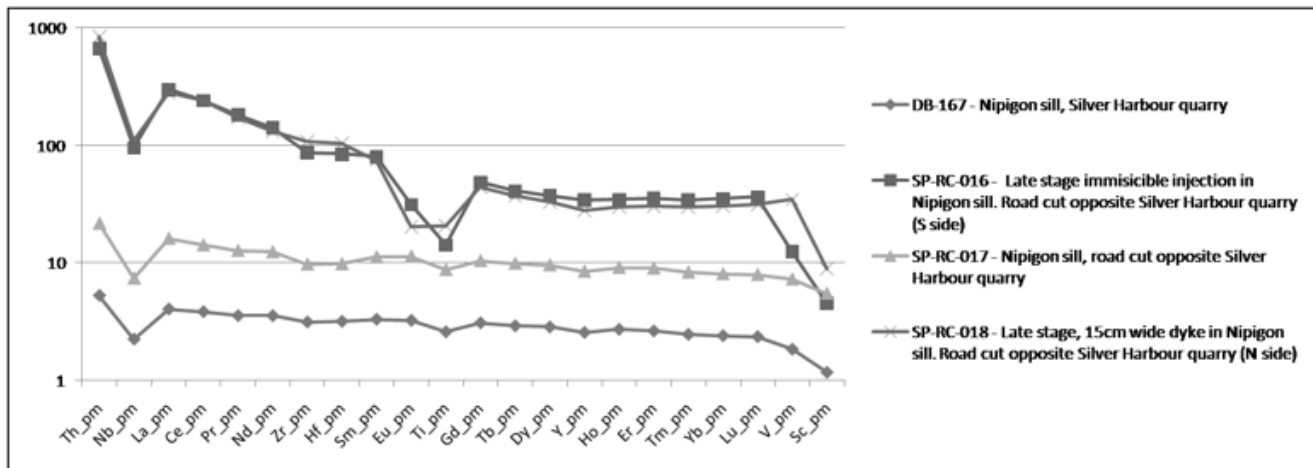


Figure 6: Primitive mantle-normalized trace element plots for samples from Silver Harbour quarry and road cut. Normalizing values from Sun and McDonough (1989). Data from Cundari (in progress).

that forms the lookout is situated on a diabase sill of Nipigon affinity. The flat-lying outcrop to the east of the lookout/parking lot exposes the top a sill, commonly characterized by feldspar-phyric patches (Fig. 7). The likely source of the feldspar-phyric patches is earlier-crystallized material or autoliths buoyantly rising to the top of the melt and crystallizing in situ.

Stop (OPTIONAL): Hornfelsed Rove Formation Sandstone

UTM coordinates: NAD83; 16U 0333765E / 5354546N

Location: Clearing off Squaw Bay road.

Description: This stop in Fort William First Nation provides excellent exposure of Rove Formation sandstone/wacke and one of the best examples

of hornfelsed sedimentary rocks resulting from Midcontinent Rift-related magmatism. In addition, a raised beach related to a higher stand (Minong or post-Minong stage?) of present-day Lake Superior is present towards the northeastern corner of the quarry (cf. Burwasser, 1977).

A massive, ~4 m thick bed of sandstone displays a characteristic hornfelsed texture. From afar, this looks like a diabase sill but upon closer inspection can be identified as a thick sandstone bed that may have been metamorphosed by an overlying (and possibly underlying) intrusion (Fig. 8). The SiO₂ content of this unit is 76.56 wt %.



Figure 7: Feldspar-phyric patches present towards the top of a Nipigon sill at The Bluffs; Stop 2.

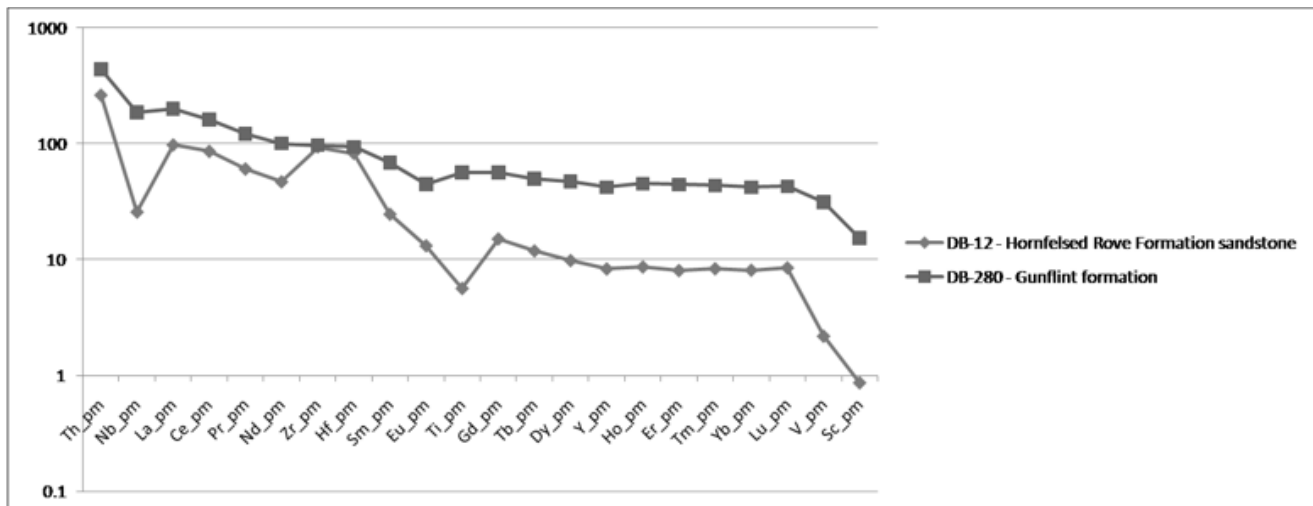


Figure 8: Primitive mantle normalized trace element plots for hornfelsed Rove Formation sandstone as well as Gunflint formation sedimentary rocks. Data from Hollings et al. (2011). Normalizing values from Sun and McDonough (1989)

Stop 3: Chippewa Park/Whiskeyjack Point

UTM coordinates: NAD83; 16U 0336798E / 5355397N

Location: Shoreline outcrops off Sandy Beach Rd. at Whiskeyjack Point

Description: Along the shoreline of Lake Superior at Whiskeyjack Point, two east-northeast trending Pigeon River-style dykes crosscut Rove Formation shale. This stop also provides a panoramic view of Thunder Bay including the Sleeping Giant, which is capped by a Logan diabase sill (Carl, 2011), as is Pie Island and the mesas to the south along the shoreline.

These dykes display a medium-grained, subophitic texture. Jointing, measured here perpendicular to dyke trend, can be used as a proxy to define approximate trends in other dykes where contacts are not observed.

Geochemically, the Pigeon River dykes display broadly similar trace element patterns to those of the Nipigon sills suggesting they are genetically related, possibly representing the feeders to the Nipigon sills. However, the wide considerable distance between the two suggests that this is unlikely. An alternative explanation has been presented by Hollings et al (2010) who have suggested that the Pigeon River dykes tapped the same long-lived mantle reservoir as the Nipigon sills that were present throughout the Midcontinent rifting.

Based on field relationships, Pigeon River dykes post-date Logan sills as several cross-cutting relationships have been documented throughout the

Logan Basin. This contradicts the geochronology data as Pigeon River dykes have been dated at 1141 ± 20 Ma for the Rita Bolduc dyke (UTM 310563E 53247021N; NAD83) and 1078 ± 3 Ma (UTM 296694E 5324134N; NAD83) for the Arrow River dyke whereas Logan sills are dated at 1114.7 ± 1.1 Ma (Heaman et al., 2007). Further geochronological and paleomagnetic work is ongoing to resolve these issues.

Stop (Optional): Riverdale sill at Robin's Donuts, Highway 61

UTM coordinates: NAD83; 16U 0326487E / 5355715N

Location: Outcrop in Robin's Donuts parking lot, Highway 61.



Figure 9: Photo of a Pigeon River dyke outcrop at Whiskeyjack point; Stop 11-3.

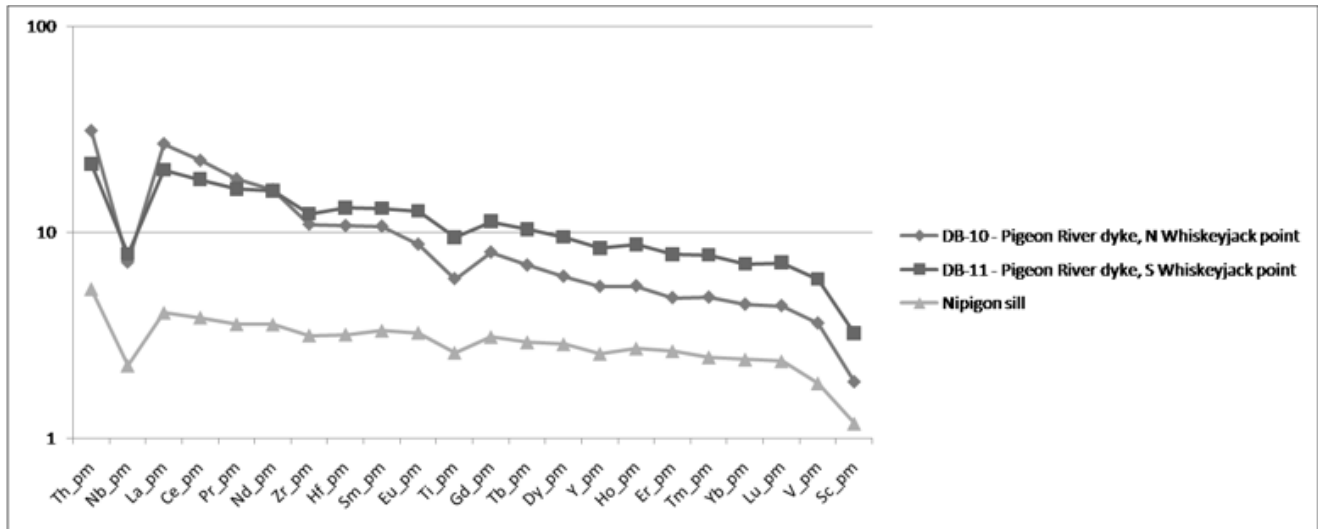


Figure 10: Primitive mantle-normalized trace element plots for Pigeon River dyes at Whiskeyjack Point, with Nipigon sill sample for comparison. Data from Hollings et al. (2011). Normalizing values from Sun and McDonough (1989)

Description: This exposure of the Riverdale sill represents the easternmost expression of the unit as described by Puchalski (2010). This low-lying, moderately weathered outcrop of the Riverdale sill gabbro-norite was the “discovery outcrop” for this unit (Smyk and Hollings, 2007).

Stop 4: Riverdale sill in Quarry

UTM coordinates: NAD83; 16U 0322410E / 5355212N

Location: Quarry at east end of Candy Mountain Dr.

Description: Sampling by Smyk and Hollings (2007) identified this as a Riverdale gabbro-norite sill in Rove Formation shale, wacke and minor tuffaceous units (Fig. 11). Subsequent detailed petrographic and



Figure 11: Photo of Riverdale sill quarry; Stop 4.

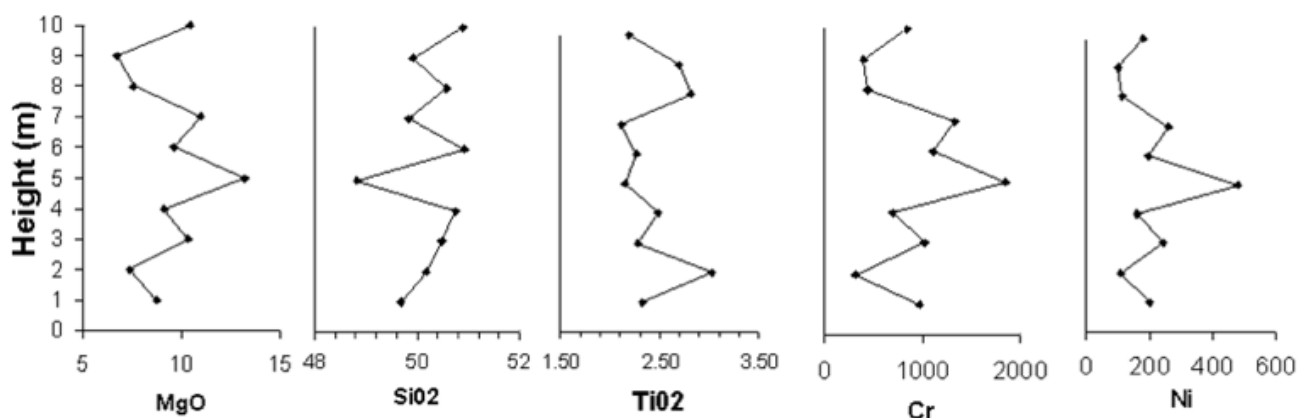


Figure 12: Height vs. elemental abundance for Riverdale sill quarry samples. TiO_2 , SiO_2 and MgO are in weight percent and Cr and Ni are ppm (Puchalski, 2010).

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 centre 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 1m 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 (Table 2). 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 (Fig. 12).

An olivine gabbro in the centre 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 centre 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 (Fig. 13). As the Rove shale displays significantly lower Nb/Nb^* and Gd/Yb_n values (Fig. 8) 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 (Fig. 14). 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. $\epsilon_{\text{Nd}}(T=1100\text{Ma})$ values of -1.6 to -1.9 for the Riverdale Sill are consistent with this model (Smyk et al., 2009).

Although the Riverdale sill is located near Logan

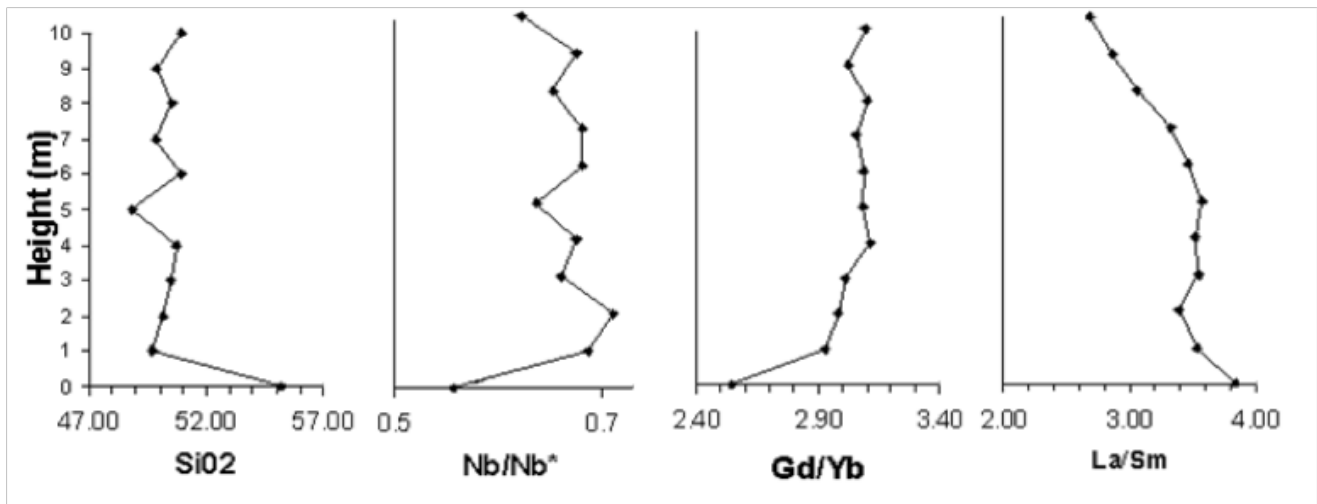


Figure 13: Height vs. elemental abundance (ppm) for Riverdale sill quarry samples. SiO₂ is in weight percent (Puchalski, 2010).

sills, it remains petrographically and geochemically distinct from them. 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 (Fig. 2). 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).

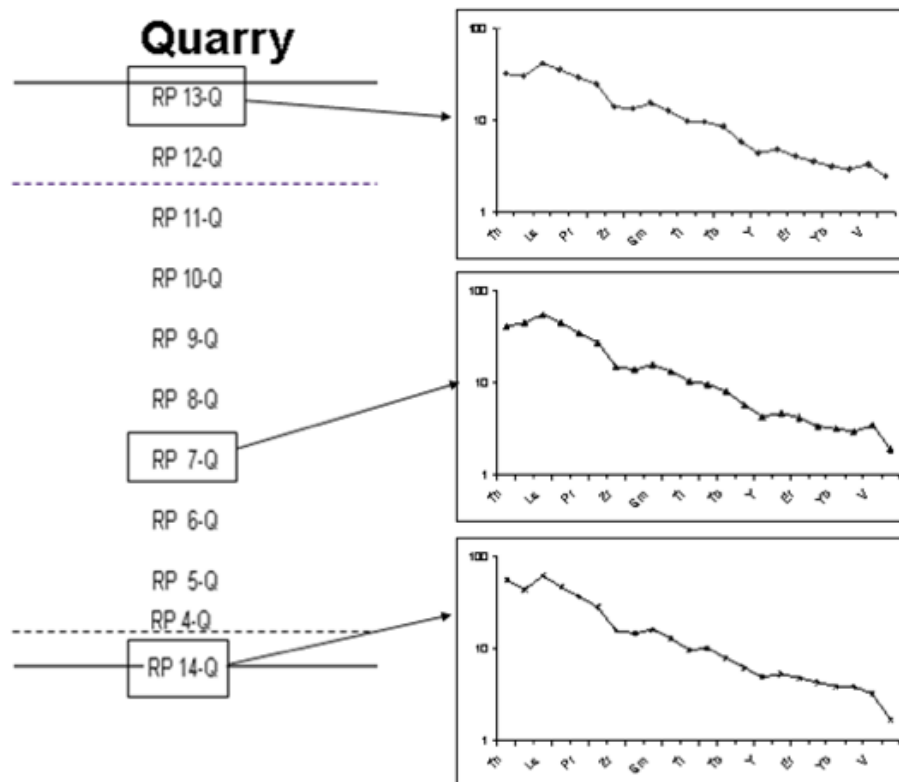


Figure 14: Primitive mantle-normalized trace element plots for successive samples through stratigraphy at the Riverdale sill quarry (Puchalski 2010). Normalizing values from Sun and McDonough (1989)

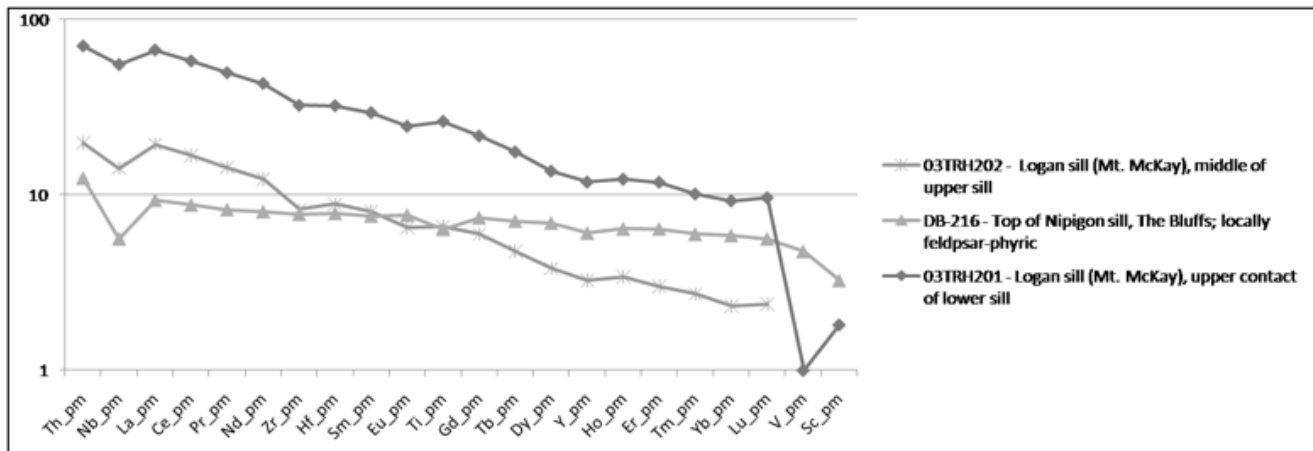


Figure 15: Primitive mantle normalized trace element plots for upper and lower sills at Mount McKay with Nipigon sill sample for comparison. Data from Hart and Magyarosi (2004) and Hollings et al. (2011). Normalizing values from Sun and McDonough (1989).

Stop 5: Mount McKay

UTM coordinates: NAD83; 16U 0331126E / 5357384N

Location: Mount McKay scenic lookout, off Mission Rd.

Description: This stop provides an exceptional view of the city of Thunder Bay as well as a great example of a stacked Logan sill sequence. The summit of Mount McKay at 482 m ASL is approximately 300 m higher than Lake Superior. Great exposures of Logan sills are abundant throughout the Logan basin south of Thunder Bay. From drill core, it has been reported that many sills are present at depth, as many as 14 as noted by Dumont Nickel Inc. who reported intersecting 14 gabbroic sills in a 705 m deep drill hole in central Pardee Township (Assessment Files, Thunder Bay South Resident Geologist's District, Thunder Bay). It is inferred that many of the Logan sills are underlain by additional sills but exposures of this are rare. Mount McKay provides the best example of a stacked sill sequence in outcrop. The geochemistry of samples from the two sills is presented in Figure 15.

The stop is centered on the lookout area, which represents the top of the lower sill at approximately 337 m ASL. Outcrop of the upper, ~60 m thick sill and adjacent, hornfelsed Rove wacke can be accessed by way of a hiking trail. If time permits, those interested in completing this hike to the upper sill may do so with extreme care. Feldspar-phyric patches similar to those observed at Stop 2 are present in an exposure of the fine-grained, upper, chilled contact of the lower sill found along a short trail to the west of the clearing next to a religious shrine. Polygonal jointing, characteristic of chilled contact zones, is well-developed in this

outcrop (Fig. 16).

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Figure 16: Photo showing polygonal jointing at the upper chill margin of the lower sill at Mount McKay; Stop 5.

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Field trip 12 - The Musselwhite Gold Deposit

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Summary

The Musselwhite gold mine in northwestern Ontario began production in the spring of 1997 with quoted reserves of 1.8 million ounces of gold. Exploration efforts since that time have successfully replaced mined reserves in most years and added substantially to them in several years. By the end of 2011 Musselwhite had produced 3.34 million ounces with remaining proven and probable reserves of 2.28 million ounces and a measured+indicated resource of 146,000 ounces and an inferred resource of 917,000 ounces.

Musselwhite is considered an orogenic gold deposit, hosted by tightly folded banded iron formation dated at ~ 2.98 Ga. It is located at the northern edge of the North Caribou Terrane (NCT), which forms the core of the North Caribou Superterrane (NCST) and the western Superior Province itself. The mine is adjacent to the ~2.86 Ga suture between the NCT and the Island Lake Domain to the north. Mineralization has been dated at 2.69 Ga, an age very close to that of gold occurrences elsewhere along the northern margin of the NCST. Unlike many orogenic gold deposits, but like a number of major BIF-hosted deposits, no major fault or shear zone that might have served as a pathway for mineralizing fluids has been found at Musselwhite. The current model for the formation of this deposit involves the development of mineralized high-strain zones along the steep limbs of the folded iron formation created during the flattening and folding event. This tectonic event was likely a result of the collision of the NCT with the Island Lake Domain 75km north of the mine, and/or the collision of the Northern Superior Superterrane ~200km to the north.

Field trip participants will have the opportunity to observe weakly deformed, shallow-dipping BIF at several outcrops west of the mine, followed by steeply dipping exposures of the BIF which hosts most of the ore at depth. Underground stops will be dependent on the faces available at the time of the tour but we hope to visit well mineralized, highly strained portions of the orebodies.

The following description of the geology is taken in part from Biczok (2007) and Biczok et al. (2012).

History of the Property

Recorded exploration in the mine area began in 1962 when brothers Harold and Alan Musselwhite of Kenpat Mines Ltd. discovered the small Kenpat gold showing in quartz veins on the north side of Opapimisk Lake as well as several showings in the iron formations on the south side. Over the next 5 years they conducted mapping, trenching and diamond drilling (12 holes totalling 773m). The Musselwhites re-staked the ground in 1973 and were subsequently financed by a syndicate of Dome Exploration, Canadian Nickel Co., Esso Minerals Canada Ltd. and Lacana Mining Corp. From 1976-1983 a major drilling program was undertaken in the West Anticline area followed by underground development and exploration in 1984. The West Anticline zone proved to be uneconomic and work soon shifted to the East Bay Synform area where the T-Antiform Zone was discovered by 1986. Exploration work carried on for another ten years and eventually the syndicate was reduced to two partners, Placer Dome (68%) and Kinross (32%). The T-Antiform ore zones had failed two early feasibility studies and failed to meet Placer Dome's economic thresholds in a third study. However, the project proponents used a risk analysis study to convince the company's board of directors that there was a very high probability of much more ore being found once the mine was in production and the go-ahead for construction was given in 1996 (Lewis, 1998). In 2002-3 the PQ Deeps ore zones were discovered followed by the Lynx Zone in 2010. After a series of corporate takeovers, Placer Dome's interest became the property of Goldcorp Canada Ltd. in 2006 who subsequently bought out Kinross' interest in 2007 and now hold 100%.

Regional Geology

The Musselwhite property covers a portion of the northwest-trending North Caribou Lake greenstone belt (NCGB) which is located in the northern margin of the Archean North Caribou Superterrane (NCST), adjacent to its internal boundary with the Island Lake Domain (Fig. 1). The NCST is a continental block consisting of ~3.0 Ga juvenile plutonic and minor volcanic rocks which underwent two periods of rifting

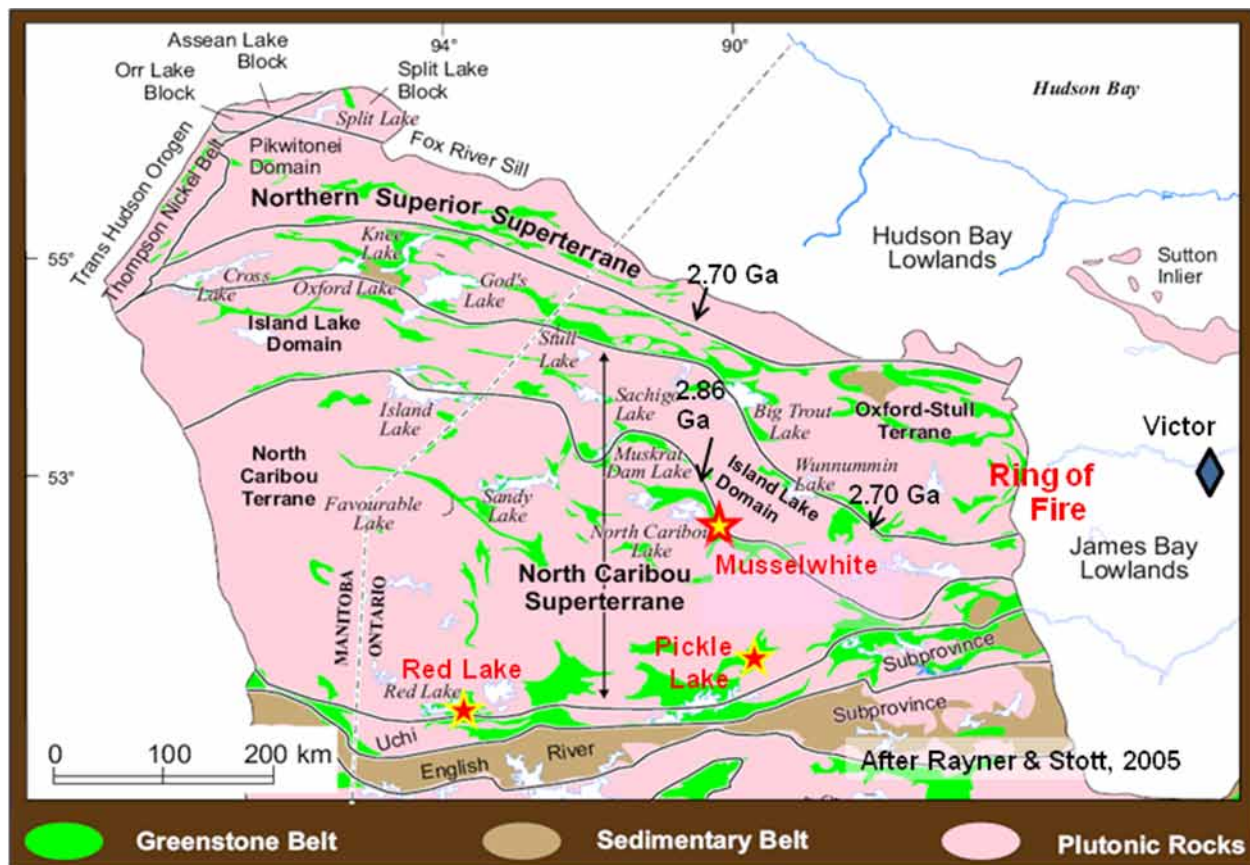


Figure 1. Tectonic setting of the Musselwhite gold deposit. After Rayner and Stott (2005)

and related deposition of arc sequences at 2.98-2.85 Ga and 2.85-2.71 Ga, followed by extensive reworking by continental arc magmatism at 2.75-2.70 Ga (Percival, 2007, and references therein). The North Caribou greenstone belt is one of the earlier arc sequences. Volcanic rocks of this belt have been dated at ~2982-2868 Ma and, more specifically, those in the mine area at 2.98-2.97 Ga (Breaks et al., 2001; Biczok et al., 2012; unpublished Musselwhite data).

The North Caribou greenstone belt (Fig. 2) has been mapped at various times by Satterly (1941), Emslie (1962), Andrews et al., (1981), and most recently a three year, multi-disciplinary effort in the mid-1980's by Breaks et al. (2001). These latter authors identified four dominantly volcanic rock suites in the Musselwhite mine area and these make up the **McGruer Assemblage** (Fig. 3):

North Rim Metavolcanic Suite (NRU): Occurs in the northeast corner of Opapimisk Lake and extends northwest from there over 60 km along the northern margin of the greenstone belt. It consists largely of mafic and lesser ultramafic volcanic rocks. A minor

felsic volcanic unit within this sequence was recently dated at 2868 Ma on Musselwhite's behalf, confirming a previous age of 2870 Ma (Davis and Stott, 2001).

South Rim Metavolcanic Suite (SRU): Occurs on the northwestern side of Opapimisk Lake and extends north and northwest from there more than 50 km along the southern margin of the NCGB. Regionally it is dominated by fine- to medium-grained, massive to pillowed basaltic flows with minor felsic and intermediate units and rare ultramafic units. Due to the paucity of felsic volcanic rocks in the SRU, only one age-date has been undertaken on rocks ascribed to the SRU by the OGS, that from an intermediate volcanic unit on the north shore of Opapimisk Lake and based on only two zircons. These rocks were dated at 2982 Ma, however, it is not certain that they actually belong to the SRU, they are more likely part of the OMU. A number of authors have interpreted the SRU as being the folded repetition of the NRU on opposite sides of a major synform forming the axial core of the NCGB. Recent lithogeochemical work at the University of Ottawa suggests that the NRU and SRU formed in different tectonic settings and are not

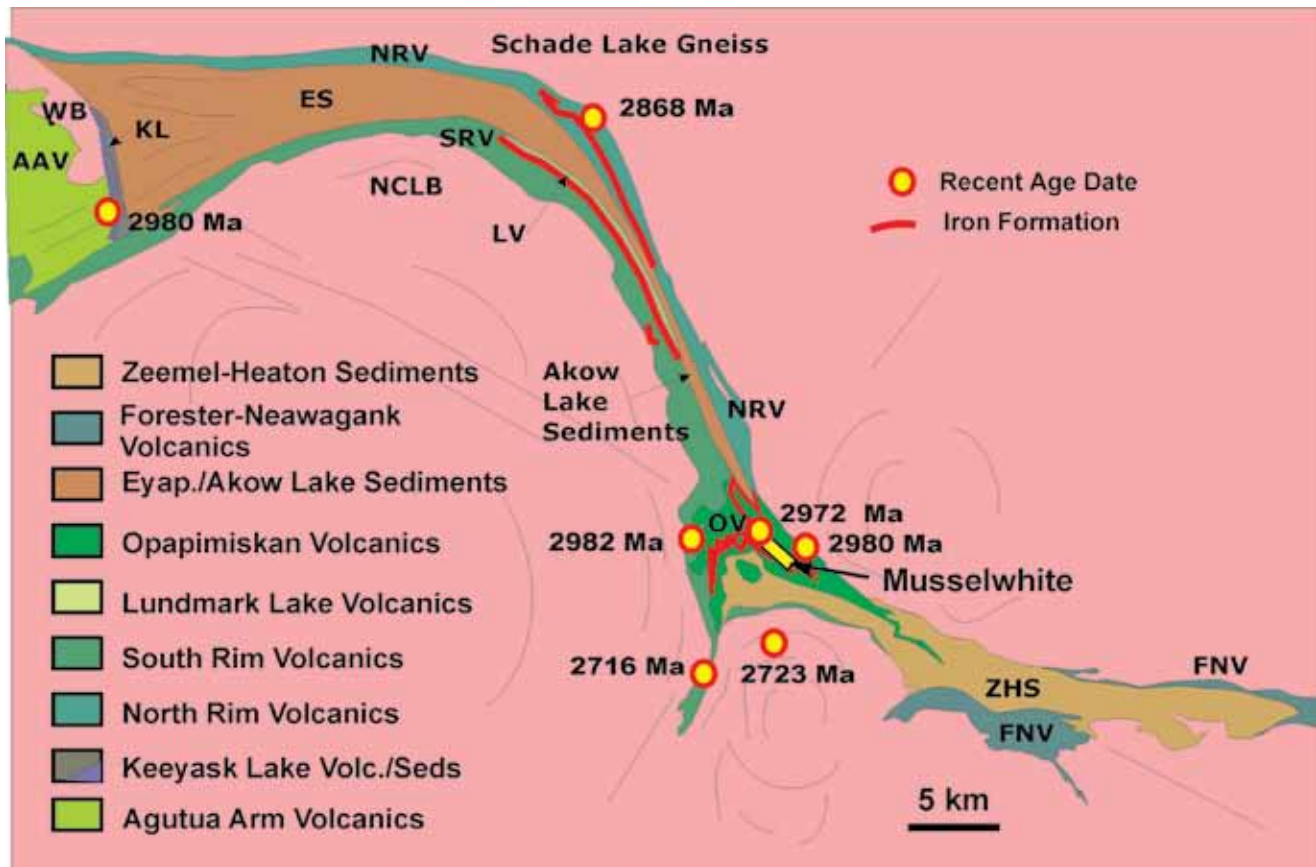


Figure 2. General geology of the North Caribou greenston belt with recent ages (after Breaks et al., 1987).

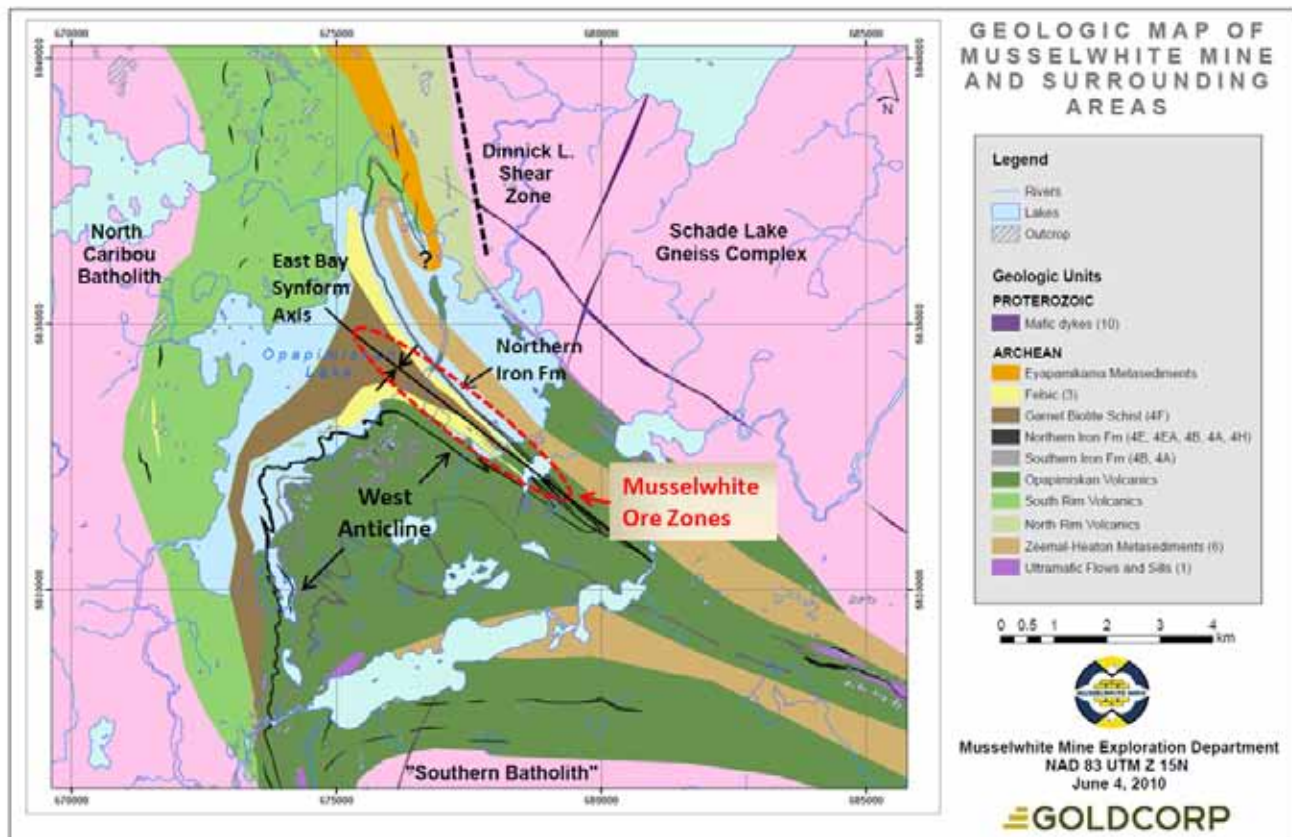


Figure 3. Geologic map of the Musselwhite mine area

equivalent (J. Duff, pers. comm.).

Drilling by the Musselwhite exploration department over the past 10 years has identified a thick sequence of felsic volcanic flows, tuffs and volcanoclastic units beneath Opapimiskan Lake and at depth below the mafic volcanic rocks exposed on the northwest shore of the lake. This felsic pile varies from coarse pyroclastic rocks to very fine-grained, massive units (flows or welded ash tuffs?) to biotite-rich volcanoclastic units. These features, and the sheer volume of the felsic units, indicate that there was a major felsic volcanic edifice at this location which overlies the Opapimiskan-Markop suite and lies in the area mapped as part of the South Rim unit. However, the continuation of intraformational iron formations common in the upper Opapimiskan-Markop suite into the lower section of the felsic volcanic rocks of the felsic suite implies, at the very least, that this is largely a conformable contact. Alternatively, the Opapimiskan-Markop volcanism may actually include the felsic and mafic rocks on the north side of Opapimiskan Lake. U-Pb age dating of

these felsic rocks is currently underway and may shed light on this issue.

Opapimiskan-Markop Metavolcanic Suite (OMU): Occupies the central portion of the NCGB in the Opapimiskan Lake area and is dominated by mafic to ultramafic flows with intercalated clastic and chemical sedimentary units, including the banded iron formations which host the Musselwhite gold deposit (described in more detail in the following section). The lower portion of the volcanic pile is dominated by ultramafic to high-Mg basalts, iron formation and lesser siliciclastic sediments, a “primitive” sequence common in many greenstone belts of the NCS and interpreted to be products of plume-related rifting by Hollings and Kerrich (1999). The upper portion of the OMU, above the main BIF horizons, consist of predominantly tholeiitic basalts. As noted above, it is currently unclear which unit the felsic volcanic strata located above these tholeiitic basalts (Fig. 4) should be assigned to, the OMU or SRU. Further geochronological and lithogeochemical work is ongoing to answer this

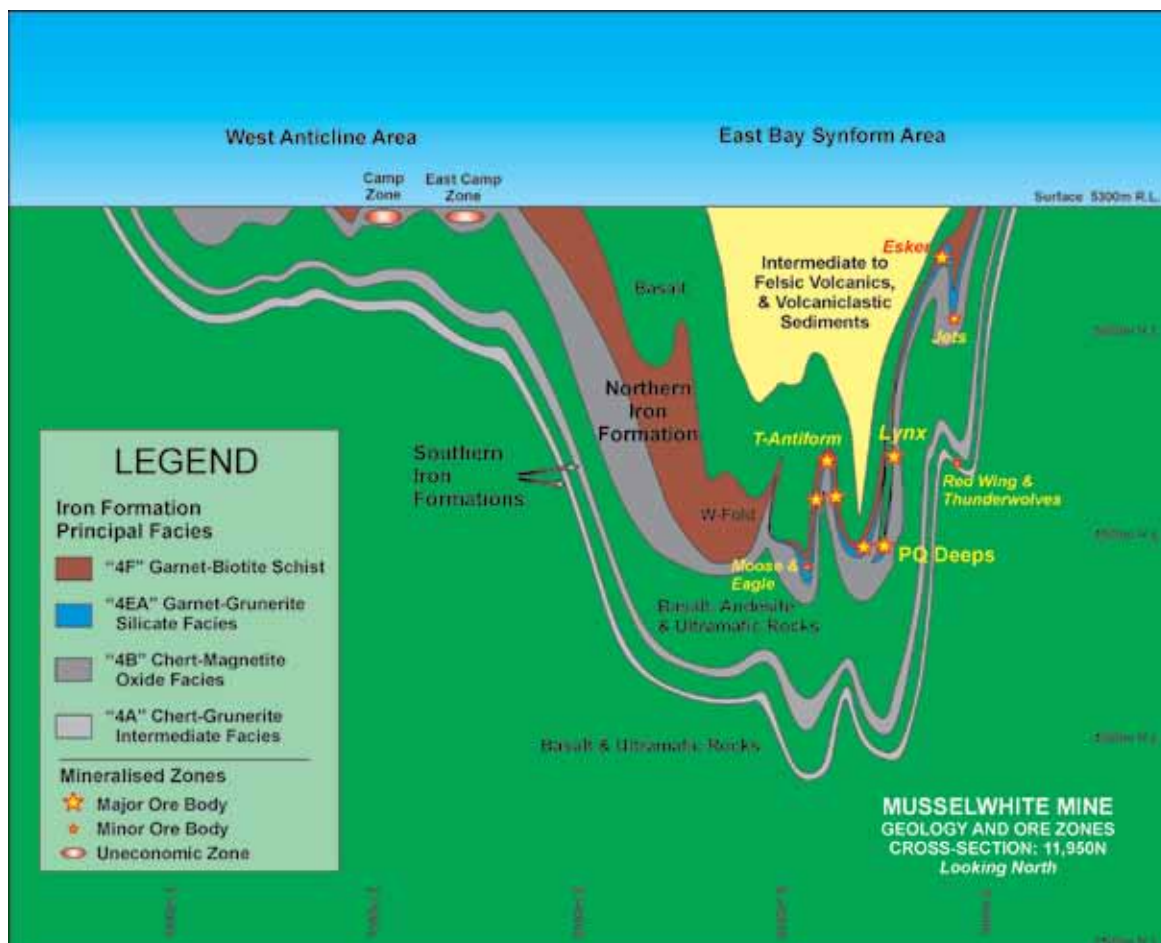


Figure 4. Generalized cross-section of the Musselwhite mine area and ore zones

question.

Forester Lake-Neawagank Metavolcanic Suite: Dominated by mafic and ultramafic volcanic units and occurs in the southeastern extremity of the belt.

Granitoid gneiss and intrusions

The greenstone belt is bounded to the north by the ~2.86 Ga Schade Lake granitic gneiss complex and various poorly-defined granitoid plutons within it. To the southwest is the ~2.85 Ga North Caribou Pluton and to the southeast is a poorly documented granitic batholith region assumed to also be ~2.85-2.86 Ga, but intruded by at least two 2.72 Ga plutons south of Musselwhite. These younger plutons are similar in age to those formed in a back arc position to the Confederation arc in the Uchi subprovince 200km to the south along the southern margin of the NCT. They are also similar to those found north of Musselwhite as far as the Hudson Bay Lowlands and potentially related to subduction of the Northern Superior Superterrane under the NCT. Further work is required to determine which of these suites the young plutons at Musselwhite belong to. Locally abundant S-type pegmatitic granite dykes occur throughout the area, particularly within areas underlain by metasedimentary rocks. These granites contain muscovite, garnet and tourmaline and

are assumed to have formed by the partial melting of the metasedimentary rocks at depth. They have been dated at 2716 to 2669 Ma and are the only intrusive rocks known in the area which overlap the age of the mineralization (Biczok et al., 2012).

Structural Geology

Three deformation events have been recognized in the NCGB by previous workers (Hall and Rigg, 1986; Breaks et al., 2001). While the time between each event may be uncertain, there is good field evidence for these three discrete episodes of variably oriented strain. The earliest event, D1, is typically manifested only by tight to isoclinal folds in the iron formations, which are typically refolded by D2/F2. An excellent exposure of a large refolded F1 fold occurs in the West Anticline area (Fig. 5) and a classic basin and dome interference fold pattern occurs within the BIF on “Grunerite Island” in Opapimiskan Lake (Fig. 6). D2 is by far the dominant deformation event in the area and has produced a near vertical, moderate to strong planar north-trending foliation throughout the area with variably developed lineations, boudinage, and mesoscopic folds (Breaks et al, 2001). Syn-D2 shearing is commonly developed parallel to F2 and locally produces well-developed rootless folds in the



Figure 5. F1 fold refolded by F2, West Anticline area.



Figure 6. Dome and basin interference fold pattern, Grunerite Island, Opapimiskan Lake.



Figure 7. Soft sediment slump features in BIF.

more highly strained margins of the iron formations. D3 is a relatively weak and localized event evidenced by minor warping and crenulation cleavages.

In the mine area, the strata have been folded into a broad antiform known as the West Anticline and the adjacent broad synform known as the East Bay Synform (Fig. 4). The West Anticline is an open fold with a relatively flat, undulating crest ~ 1 km across, featuring a series of smaller gentle folds that plunge to the north at $<5^\circ$ to $\sim 30^\circ$. Rocks in this area are commonly only weakly deformed and preserve a variety of soft-sediment features in the iron formations including slumping (Fig. 7). In contrast, the East Bay Synform is bounded by limbs that dip steeply between ~ 70 - 90° and the keel is host to two 2nd order antiforms, the “T-Antiform” and the “W-Fold”. The folds plunge fairly consistently at $\sim 12^\circ$ to the north. This is a considerably higher strain setting than the West Anticline and primary structures are rarely preserved here. Late shears and faults are common in the volcanic rocks coring this synform and these are typically pervasively biotitized and laced with 20-40% thin calcite veinlets. Mineralized high-strain zones are predominantly developed in the upper margin of the Northern Iron Formation within the steepest portions of the fold limbs. The T-Antiform, the adjacent synformal keel (known as the “PQ Deeps”), and the east (PQ) limb of the East Bay synform host the bulk of the gold mineralization at Musselwhite. A major sub-vertical fault at $\sim 3^\circ$ to the fold axes has sliced the eastern limb (known as the PQ limb) into two pieces over a 700m interval and displaced the western portion ~ 1.3 km to the south, forming a very large tubular sheath fold. This fault has fortuitously juxtaposed the ore zones over an interval of several hundred metres.

Metamorphism

Sedimentary rocks intercalated throughout the volcanic pile in the mine area commonly contain garnet +/- staurolite and the area is therefore considered to be of amphibolite grade. The greenschist-amphibolite isograd is thought to be located at least 5km from the mine to the north.

Mine Stratigraphy

Rocks hosting the Musselwhite gold deposit belong to the Opapimiskan-Markop Metavolcanic Suite (OMU) and have been subdivided into a detailed stratigraphy that is relatively consistent over the property, although major facies changes are locally observed along and

across strike. This “mine stratigraphy” is depicted in Figure 8.

The lower portion of the OMU consists mainly of komatiitic basalts and ultramafic flows/intrusions, with local high-Mg andesite flows. This presence of andesite, which is commonly bleached and highly biotitized in the mine area, is somewhat unusual in such a mafic-ultramafic sequence. Lithogeochemical analysis suggests that it formed by fractional crystallization of komatiite melt contaminated with either crustal TTG melts or felsic volcanic magma (Hollings and Kerrich, 1999). This sequence is overlain by two major banded iron formations (BIF) separated by 10-30m of mafic-ultramafic volcanics and local high-Mg andesite.

The Southern Iron Formation (SIF) is the lowermost BIF and is a relatively monotonous sequence of thinly laminated magnetite and chert. There is generally little or no silicate, sulphide, or other facies within this horizon in the mine area. The SIF commonly occurs in two principal horizons, 5 to 20m thick, separated by 5-10m of basalt. The SIF hosts a number of small mineralized zones in tightly

folded areas including the Ranger, Red Wing and Thunderwolves.

The Northern Iron Formation (NIF) sits ~20-30m above the SIF and is the main ore host at Musselwhite. This is a complexly layered horizon typically ~40m thick in total and consists of seven different facies thought to reflect varying proportions of clastic and chemical sedimentation combined with variations in the Redox conditions. Not every facies is always present across the drilled extent of the NIF, but where they are, the following stratigraphy is observed from the base to the top of the formation.

Unit 4H: This unit is a sulphidic iron formation, composed of 10-80% syngenetic pyrrhotite in a dark grey to black cherty argillite (Fig. 9). The continuity of the 4H is poor and where present its thickness varies from <10cm to as much as 10m.

Unit 4A: This unit is fairly ubiquitous throughout the mine area and is the most common basal unit of the NIF. It is composed of pale grey, weak-moderately magnetic chert, interlayered with 30-50%, diffuse bands of fine-grained, light yellow grunerite.

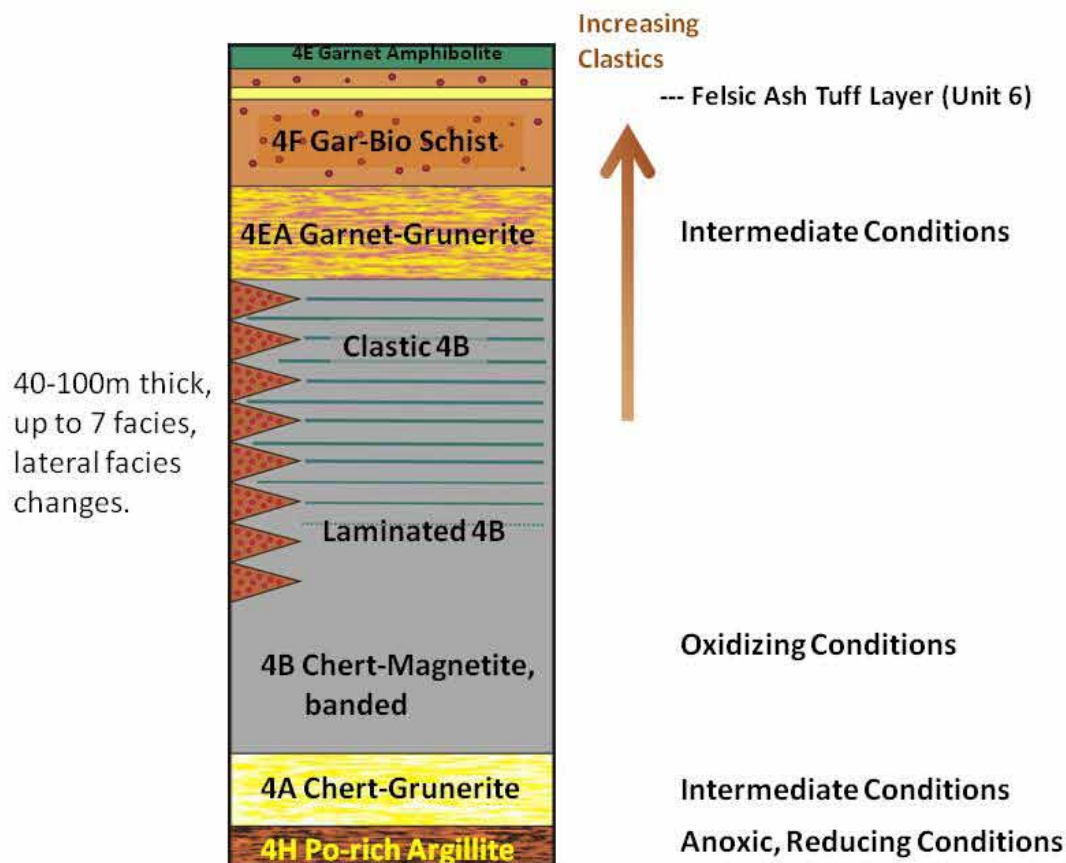


Figure 8. Detailed stratigraphy of the Northern Iron Formation including presumed Redox conditions during formation.



Figure 9. 4H, massive pyrrhotite exhalite with chert fragments. Drill core sample 5.1cm (2") wide.

Unit 4B: Composed of thinly laminated to thick-banded (~1-2cm) chert-magnetite oxide-facies BIF, the 4B is typically 20-30m thick and forms ~3/4 of the NIF (and usually all of the SIF). It can be subdivided into two varieties: a lower, thick-banded, relatively pure chert-magnetite BIF (Fig. 10), and an upper, thinly laminated, more clastic-rich variety consisting of alternating intervals, typically <1-2cm thick, of thin, diffuse laminae of magnetic chert, and more homogeneous, medium-dark green, fine-grained amphibole-rich layers (Fig. 11). With increasing stratigraphic height these amphibole dominant bands become garnetiferous and may contain up to 25% pale red garnets <2mm in diameter. The amphibole+/-garnet layers are readily affected by hydrothermal alteration. Adjacent to mineralized zones and/or major quartz veins they are commonly altered to massive, fine-grained black biotite, with an associated coarsening of the garnets.



Figure 10. 4B, thick banded chert-magnetite BIF. Drill core sample 5.1cm (2") wide.



Figure 11. Laminated 4B (chert magnetite) below, clastic 4B with thin layers of green amphibole-garnet above. Drill core sample 5.1cm (2") wide.

Unit 4EA: Pristine 4EA is a silicate iron formation composed almost entirely of massive bands of garnet-grunerite with ~20-30% bands of pale grey, moderately magnetic chert <1-2cm thick (Fig. 12). The garnet-grunerite layers contain ~30-55% almandine garnets, 1-4mm across, in a fine-grained matrix of pale yellow grunerite and minor fine-grained disseminated magnetite. The 4EA forms the main ore-host at Musselwhite and in mineralized zones it has undergone quartz flooding/veining, replacement of the original grunerite by green amphibole adjacent to the veins (hornblende or ferrotschermakite; Otto, 2002), significant coarsening of the garnets due to hydrothermal overgrowths, and pyrrhotite mineralization. There has been some debate over the years regarding the origin of the grunerite in the 4EA. Some have argued that it

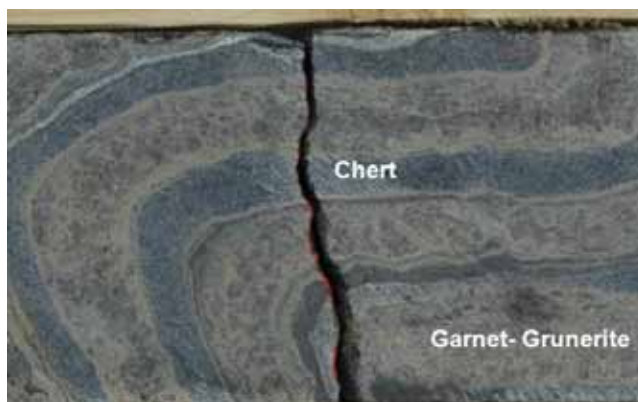


Figure 12. 4EA, garnet-grunerite silicate facies iron formation. Drill core sample 5.1cm (2") wide.

is the product of hydrothermal alteration of precursor magnetite + chert. While such alteration is locally observed on a small scale (mm to cm) in sheared, quartz veined portions of the 4B for example, it is difficult to envisage such a process producing the ~10m thick unit we see today that is continuous over many kilometers, always at the same stratigraphic heights, commonly has no evidence of hydrothermal alteration (such as quartz veins, calcic amphiboles, etc.), and displaying no gradation along strike into “fresh” magnetite+chert units. Perhaps most compelling is the nature of the BIF approximately 20km along strike to the north of Musselwhite in a greenschist grade region. Outcrops here are dominated by grunerite-chert layers with very little or no magnetite. What little magnetite is present occurs in very thin laminae, interbedded with grunerite, and delicately folded into complex patterns. It seems highly unlikely that the grunerite laminae here could have formed at the expense of the magnetite and still have preserved this delicate layering. The preferred explanation for the formation of the grunerite in the 4EA is that it is the product of metamorphism of the original iron-silica gels produced by seafloor hydrothermal vents.

Unit 4F: The 4F is a garnet-biotite +/- staurolite schist with an average Fe_2O_3 content of ~25-30% and thus qualifies as an iron formation itself (Fig. 13). Typically it contains 30-55% subhedral garnets, 1-5mm across, in a fine-grained matrix of 60-70% biotite with lesser quartz, feldspar and magnetite. Intraformational 4F horizons within the basaltic pile commonly contain up to 30-40% anhedral, light yellow staurolite grains 1-2mm across.

Unit 6: This is a thin, but semi-continuous unit, typically <1m thick, that occurs in the upper portion



Figure 13. 4F, garnet-biotite-(Qtz-Fd)±staurolite schist; ferruginous metapelite. Drill core sample 5.1cm (2”) wide.



Figure 14. 4E, garnetiferous amphibolite. Drill core sample 5.1cm (2”) wide.

of the 4F sequence in the NIF. It is a light beige-grey, siliceous, fine-grained, equigranular, very homogeneous rock composed of 20-30% finely dispersed biotite and 70-80% quartz-feldspar. Lithogeochemical analyses indicate that this unit is very similar to the local felsic volcanic rocks but has relatively elevated levels of V, Mn and Ba. It is interpreted as a meta-sediment derived from a waterlain felsic ash tuff.

Unit 4E: Where present, the 4E forms the uppermost unit of the NIF. It is generally a thin, <1m, fine-grained, massive, medium-dark green amphibolite containing 15-30% anhedral pale red garnets 2-4mm across (Fig. 14). It has little or no visible quartz or feldspar and averages 25% Fe_2O_3 .

The 4H, 4A, 4E and 4F all occur as discrete, “intraformational” horizons within the basaltic pile in addition to occurring within the NIF. These horizons are most commonly <2m thick but can swell to ten’s of metres within fold crests or keels. They are most abundant in the first 20-30m above the NIF and can be locally well mineralized.

Overlying the NIF is a variable thickness of basalts ranging from <2m in the Esker fold area to 30-50m in the T-Antiform area. There are little or no komatiitic basalts or ultramafic units above the NIF. In the past, the various basalts in the mine area were designated as “BVol” (for “basic” or “basement” volcanics) or “2Vol” for the units near the NIF. Given the uncertainty of defining a “basement” in the mine area, these terms have been abandoned in recent years in favour of a generic “Unit 2” and its variants for all basalts and andesites.

Mineralization and Alteration

Mineralization at Musselwhite is found largely within sub-vertical high-strain zones in the favourable iron formation units, primarily the silicate facies (4EA) and to a lesser extent the oxide facies (4B), where a number of the smaller ore zones are found (e.g., Jets, Ranger, Thunderwolves, lower portion of the PQ Deeps A-Block). Scattered small mineralized shears cut mafic volcanic rocks, however, to date these occurrences are of limited extent and uneconomic. Discrete mineralized shear zones up to 10m wide occur across a zone ~225m from the “Moose Zone” in the west through the T-Antiform and into the PQ Deeps in the east. Locally these shear zones coalesce into broad zones up to 40m wide (e.g. PQ Deeps C-Block). Individual mineralized shear zones tend to be quite persistent along strike, following the upper margin of the Northern Iron Formation and/or high strain zones within second order folds (e.g. the Jets Zone) for 2-3km. While the shear zones are primarily ductile, on both macroscopic and microscopic scales, there is evidence of associated brittle deformation that produced dilatant zones and allowed the infiltration of gold-bearing fluids.

Gold mineralization within the 4EA is generally accompanied by substantial quartz veining or flooding, pyrrhotite formation, green amphibole (hornblende-ferrotschermakite; Otto, 2002) replacement of the original grunerite-rich host (Fig. 15), a coarsening of the garnets, and local late-stage chlorite. The formation of pyrrhotite is thought to be a consequence of sulphidation reactions between the original gold-bearing bi-sulphide complexes and the iron-rich minerals of the BIF. The gold content is crudely proportional to the sulphur

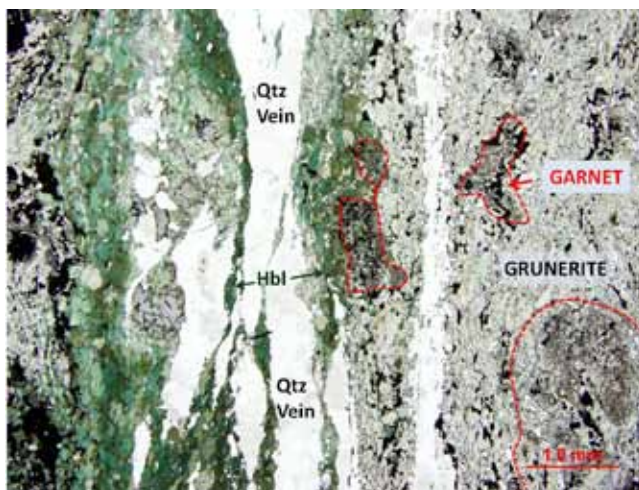


Figure 15. Photomicrograph of 4EA cut by quartz veins flanked by green amphibole replacement of grunerite (PPL).

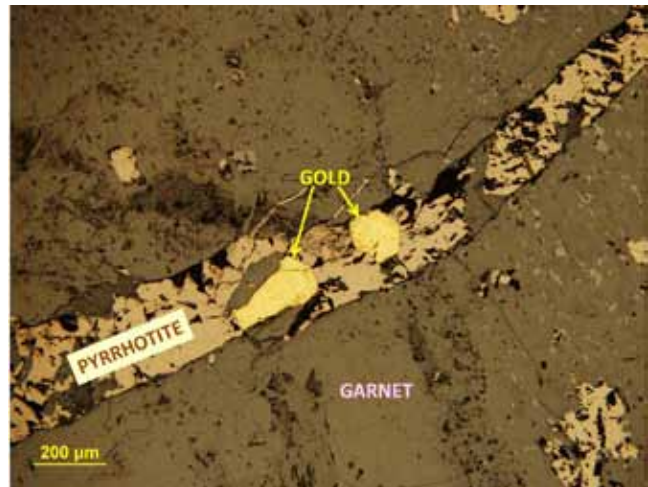


Figure 16. Reflected light photomicrograph of garnet cut by pyrrhotite-gold filled fracture.

content of the mineralization in the ratio of 5 g/t Au for each 1% of sulphur. Gold occurs as free-milling native gold, most commonly in pyrrhotite-filled fractures in garnets (Fig. 16), with lesser amounts in the pyrrhotite, green amphibole, and rarely in quartz veins. The clastic 4B underlying mineralized 4EA is commonly highly altered itself. The original very fine-grained, green amphibole-rich laminae are replaced by massive fine-grained biotite and 5-15% medium-grained secondary garnets; this alteration is especially common adjacent to quartz veins. In spite of the intensity of this alteration, it typically has nil to very low levels of gold or pyrrhotite.

Surface Field Trip Stops

Stop 1a: Trench #4, West anticline

Lithology: This stripped outcrop on the west side of the exploration road exposes the Southern Iron Formation and the adjacent ultramafic rocks. The BIF here is dominated by chert with lesser beds of a light grey magnetite-amphibole-rich unit not seen outside the West Anticline area. Only along the margins of the outcrop does one see the chert-magnetite BIF which is predominant on the rest of the property.

Comments: The change in the degree of strain is evident in this exposure as one crosses from the central, shallow-dipping area to the steeply-dipping, highly strained margins. The abundant small minor folds along the margins have an overall “Z” pattern with their axes plunging shallow to the (grid) north. On flat surfaces the tops of these folds appear as rootless or

intrafolial folds.

An unusual iron-rich chlorite schist, with a peculiar knobby or ovoid texture, appears to be locally interbedded with the cherty BIF as well as cross-cutting it. This unit averages ~40% Fe₂O₃, 15% Al₂O₃, 27% SiO₂ and 6% MgO. It commonly contains intergrown magnetite and tourmaline grains and has been interpreted as an iron-rich metasediment that was locally injected as “dykes” through cracks in the more lithified BIF above.

Stop 1b: Z-folded Chert-Magnetite BIF

On the opposite (east) side of the road from the large stripped outcrop discussed above is a small exposure of 4B (well banded chert-magnetite BIF) that displays metre-scale Z-folds. Minor folds like these throughout the West Anticline can be related to their position on the series of undulating folds that make up the crest of the major antiform.

Stop 2: Lakeshore Exposure of Gently Folded 4E/4EA

Lithology: This small exposure near the shore of Opapimiskan Lake was only rediscovered in the Fall of 2011. What little of it was exposed at that time consisted of gently folded 4E and/or 4EA belonging to the Northern Iron Formation, plunging to the north.

Comments: The outcrop was partially excavated in the Fall and will be power-washed in the spring prior to the field trip. It is expected to provide an excellent look at weakly deformed 4E / 4EA of the NIF, one of the few such exposures in the area.

Stop 3: PQ Limb Section through the Northern Iron Formation

Lithology: This is the only outcrop of the complete Northern Iron Formation found on the property and was created during overburden stripping operations related to the development of the PQ Shallows and the Ranger open pits. The exposures are part of the sub-vertical PQ limb, the eastern limb of the East Bay synform. The section begins with the eastern footwall which exposes deformed pillow basalts and passes through an almost complete section of the NIF, including a probable 4H, minor 4A, well-developed 4B, 4EA, 4F, Unit 6, and 4E. A small basaltic (“Bvol”) dyke is found within and roughly parallel to the trend of the 4EA.

Comments: there is a wide range of features to note

in these outcrops including:

- 1) The overall structural fabric is predominantly flattening rather than strike-slip movement or shearing.
- 2) There is a well developed zone of quartz veining and intense biotite alteration in the clastic 4B immediately below the 4EA. This type of alteration is typically barren and conversely biotite alteration such as this is relatively rare in well-mineralized 4EA.
- 3) The hanging wall basalts have a distinct pale pink-purple color due to the pervasive fine-grained biotite alteration. Biotite alteration, bleaching of the amphiboles, and a strong foliation are typical of the basalts adjacent to the contact of the Northern Iron Formation and indicates that strain was partitioned (focused) at this contact.
- 4) There is a northeast trending series of small-scale folds and cleavages scattered throughout the exposures. These may be part of the D3 deformation event.

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