



63rd Annual Meeting

Institute on Lake Superior Geology



**Wawa, Ontario
May 8 - 12, 2017**



Wawa

**Wild Goose
or
Land of the Big Goose
in Ojibway**



**Proceedings Volume 63
Part 2: Field Trip Guidebook**



Institute on Lake Superior Geology

63rd ANNUAL MEETING

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Proceedings Volume 63

Part 2: Field Trip Guidebook

Compiled by Theodore J. Bornhorst and Margaret J. Hanson

Cover Photos: Top left, Calcite crystals on pyrite from George W. MacLeod Mine, Wawa, collection of the A. E. Seaman Mineral Museum donate, photograph by George Robinson; iconic Wawa goose at junction of Trans-Canada Highway and Highway 101, image from <http://www.northernontario.travel/algoma-country/fun-facts-about-the-famous-wawa-goose>; Sir James Dunn open pit iron mine on the Eleanor Iron Range, photograph by Anthony Pace.

63rd INSTITUTE ON LAKE SUPERIOR GEOLOGY

VOLUME 63 CONSISTS OF:

PART 1: PROGRAM AND ABSTRACTS

PART 2: FIELD TRIP GUIDEBOOK

TRIP 1: ARCHEAN AND PROTEROZOIC GEOLOGY OF THE MARATHON-HEMLO AREA

Day 1: Geology of the Coldwell Alkaline Complex

Part 1: Transect through the Coldwell alkaline complex

Part 2: Marathon Cu-PGM deposit

Day 2: Geology of the eastern Schreiber-Hemlo Greenstone Belt in the vicinity of Heron Bay and Hemlo

TRIP 2: MORE UNUSUAL DIAMOND-BEARING ROCKS OF THE WAWA AREA

TRIP 3: GEOLOGY OF THE WAWA GOLD PROJECT

TRIP 4: GEOLOGY OF THE ISLAND GOLD MINE

TRIP 5: GEOLOGY OF THE RENABIE AREA

TRIP 6: KAPUSKASING STRUCTURAL ZONE AND BORDEN LAKE GOLD DEPOSIT

Reference to material in Part 1 should follow the example below:

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Part 1: Field Trip Guidebook

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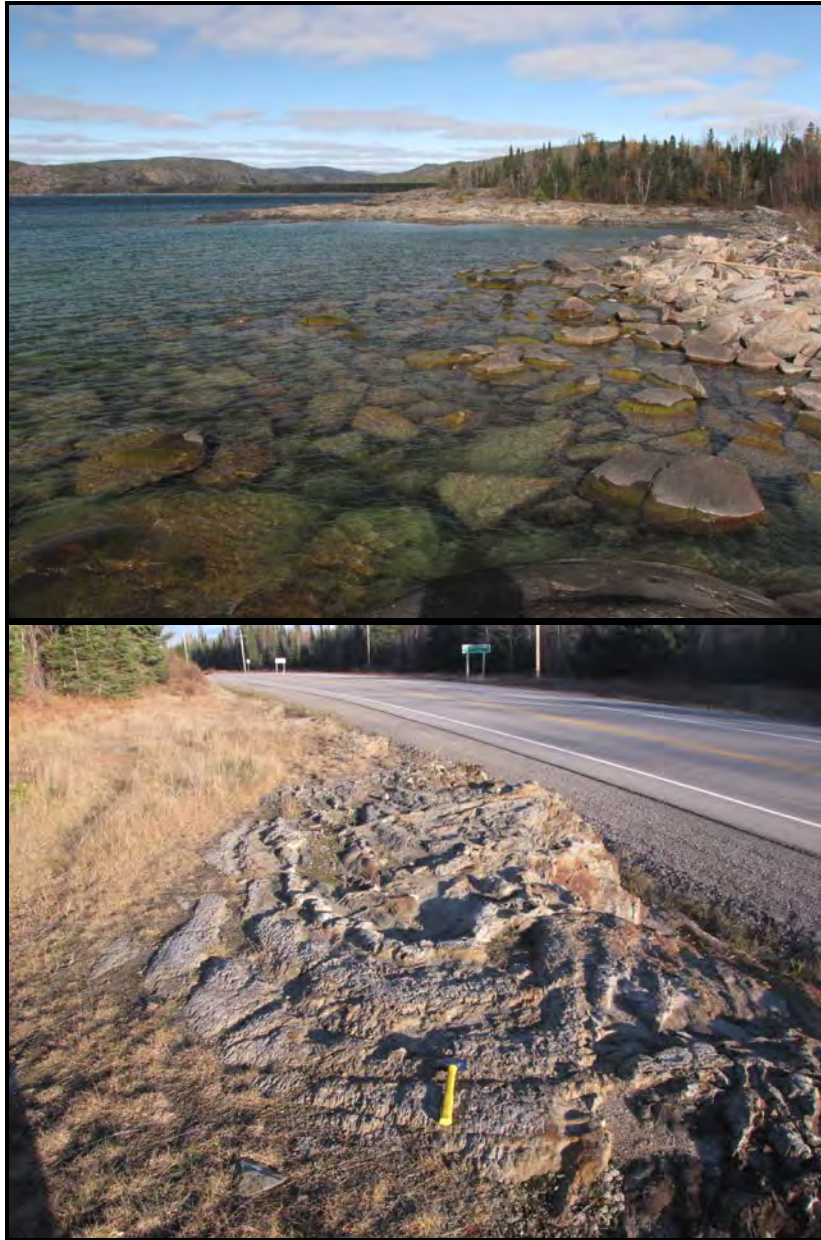
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Field Trip 1

Archean and Proterozoic geology of the Marathon-Hemlo area

Day 1

Geology of the Coldwell Alkaline Complex



The Coldwell Alkaline Complex portion (Day 1) of Field Trip 1 consists of two parts: 1) a west to east transect through southern part of the complex and 2) a visit to the Marathon Cu-PGE Deposit.

Part 1: Transect Through the Coldwell Alkaline Complex

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A variety of Mesoproterozoic, Midcontinent Rift-related alkalic and carbonatitic rocks occur within several intrusive complexes on or near the northern shore of Lake Superior (see Figures 1 and 2). They include the Coldwell and Killala Lake alkaline complexes, the Prairie Lake and Chipman Lake carbonatites, and numerous diatremes and related dikes in the vicinity of Dead Horse Creek (Sage 1982, 1985, 1987) (see Figure 2). These complexes are spatially localized and structurally controlled by the Trans-Superior Tectonic Zone (TSTZ), a north-northeast-trending structure that extends for over 600km and includes the Thiel Fault in Lake Superior (Klasner et al. 1982). Alkaline magmatism related to Midcontinent rifting occurred along the TSTZ from approximately 1.2 to 1.0 Ga, as is shown in Table 1 below:

Table 1

Lithologic Unit/Complex	Age(s) (Method)	Reference
Coldwell Alkaline Complex	1108 ± 1 Ma (U/Pb)	Heaman and Michado (1987)
Be-Zr Zone crosscutting Dead Horse Creek diatreme	1112.7 ± 4 Ma (U/Pb) ¹ 1128.7 ± 6 Ma (U/Pb) ²	Krogh and Wilkinson (M. Smyk pers. Comm. 1995)
Prairie Lake Carbonatite	1130 ± 10 Ma (Rb/Sr)	Pollock (1987)
Lamprophyre Dyke, McKellar Harbour	1145 ± 15/10 Ma (U/Pb)	Queen et al. (1996)
Gabbro (biotite), Killala Lake Complex	1185 ± 90 Ma (K/Ar)	Coats (1970)
Syenite, Killala Lake Complex	1050 ± 35 Ma (Rb/Sr)	Bell and Blenkinsop (1980)

(¹-2.49% discordant; ² 1.82% discordant)

It has been postulated that the TSTZ may represent part of a failed arm of a Keweenawan-age triple junction (Weiblen 1982; Mitchell and Platt 1982b) or the intersection of a late fracture system with the rift (Mitchell et al. 1983). Local alkalic and carbonatite complexes have been emplaced at inflections in the trends of major structural zones, or at sites of cross-faulting (Sage 1991). The Coldwell and Killala Lake alkaline complexes are both thought by some to have formed as the result of ring fracturing and caldera collapse. The abundance of observed xenolithic blocks and roof pendants suggests that these complexes are presently exposed at relatively high structural levels.

Similar ages for numerous mafic intrusions in the Nipigon Embayment (cf. Heaman et al. 2007) and the alkalic rocks of the Coldwell Complex (1108 ± 1 Ma; Heaman and Machado 1992) indicate the contemporaneous production of tholeiitic and alkalic magmas during Midcontinent rifting. The oldest magnetization, found in the gabbros and augite syenites on the eastern side of the complex, records a concordant pole position with reversed polarity at about 1109 ± 5 Ma on the Keweenawan segment of Precambrian apparent polar wander path (Lewchuk and Symons 1990). The localization of the alkalic magmatism off-axis, dominantly northeast of the central rift, prompted Heaman and Machado (1992) to suggest that this may have been a region of maximum lithospheric extension during rifting. U/Pb data (Heaman and Machado 1992) demonstrate that most rock units in the Coldwell Complex were emplaced within a relatively short time span (<3 million years) ca. 1108 Ma, and support the contention that the complex experienced relatively rapid cooling from initial emplacement temperatures to at least ~500° C.

Strontium-, neodymium- and lead-isotopic compositions of selected minerals from different phases of the complex (Heaman and Machado 1992) display considerable scatter, suggesting that their magmas had different isotopic compositions. The initial strontium- and neodymium-isotopic compositions of clinopyroxene and plagioclase from one of the earliest gabbroic phases are identical to data derived from primitive olivine tholeiites from the Midcontinent Rift and indicate that the majority of magmas, both

tholeiitic and alkaline, have a uniform, nearly chondritic isotopic composition (ibid). Samarium-neodymium data, supported by oxygen-isotopic and whole-rock geochemical data, indicate that crustal contamination played a small, varied role in the generation of the Coldwell magmas (Bohay 1997). In addition to small, variable amounts of assimilation of upper and lower crust, the parental plume magmas also interacted with the lithospheric upper mantle to a small degree (ibid).

Local alkalic and carbonatitic intrusive rocks host a variety of characteristic base, precious, titaniferous, phosphate, and rare metal occurrences (cf. Smyk and Sage 1995). They include the following:

1. Magmatic Cu-Ni-PGE ($\pm$ Au, Ag) in gabbros of the Killala Lake and Coldwell complexes;
2. Magmatic Ti-V-apatite deposits in the Eastern Border Gabbros of the Coldwell Complex;
3. Magmatic U, Nb (+ wollastonite, apatite) in the Prairie Lake carbonatite (Sage 1987);
4. Late-stage magmatic Nb-Y-F-family rare earth elements in syenite pegmatites (Alexander 2007);
5. A Be-Zr-U-Th-Y mineralized zone crosscutting the Dead Horse Creek diatreme (Smyk et al. 1993; Potter, 2004); and
6. Pb-Zn-Ag-mineralized quartz-carbonate veins (Kissin and McCuaig 1988).

The Coldwell Alkaline Complex (see Figure 3) covers an area of $\sim 580\text{km}^2$, making it one of the largest alkalic complexes in the world and the largest in North America. It was emplaced during the early stages of the Midcontinent rift system, which includes: early large and small mafic to ultramafic intrusions (i.e. Seagull Lake Complex, Thunder Bay North Complex); Keweenawan flood basalts, the Duluth Complex, the Nipigon and Logan sills, and a variety of non-diabase mafic to ultramafic dyke-rocks. The Coldwell Complex was mapped by Kerr (1910a, 1910b), Puskas (1967), and Walker *et al.* (1993b, 1993c), and comprises 3, superimposed ring sub-complexes or magmatic centers (Mitchell and Platt 1978) that young progressively (Centers 1 to 3) to the southwest (see Figure 4). Walker et al. (1993) and Barrie et al. (2002) dispute the series of ring dykes or sheeted cones interpretation and suggest that the complex is a composite lopolith or sill. The intrusive centres can be generally described as follows:

1. Center 1: Generally silica-saturated rocks with oversaturated residue; chiefly consisting of the Eastern and Western border gabbros (the oldest rocks within the complex) and later iron-rich augite syenite and syenite-syenodiorite (Mitchell and Platt, 1978, 1982; Mulja, 1989);
2. Center 2: Generally silica-undersaturated alkalic rocks with oversaturated residue; consisting of locally nepheline- and hastingsite-bearing miaskitic nepheline syenite, and numerous volumetrically minor alkaline lamprophyre and analcime tinguaites dykes (Mitchell and Platt, 1978, 1982; Laderoute, 1989; and Mulja, 1989); and
3. Center 3: Silica-oversaturated alkalic rocks with oversaturated residue; consisting of magnesio-hornblende syenites, quartz syenites, and minor granites (Mitchell and Platt, 1994; Lukosius-Sanders, 1988).

The mineralogy of the main lithologic units is listed in Table 2. The superimposition of the three intrusive centres and a complex, protracted magmatic history has produced a myriad of hybrid rocks, igneous breccias, and ambiguous crosscutting relationships.

The wide variety of lamprophyric and other dyke rocks occurring within the complex (as described by Mitchell and Platt, 1994) include (in order of emplacement):

1. Mafic ocellar lamprophyre (camptonitic variety)
2. Quartz-bearing, mafic lamprophyres (camptonitic variety)
3. Sannaite-type lamprophyres
4. Monchiquitic-type lamprophyres
5. Feldspar glomeroporphyritic and alkali basalt dikes
6. Analcime tinguaites (heronite)

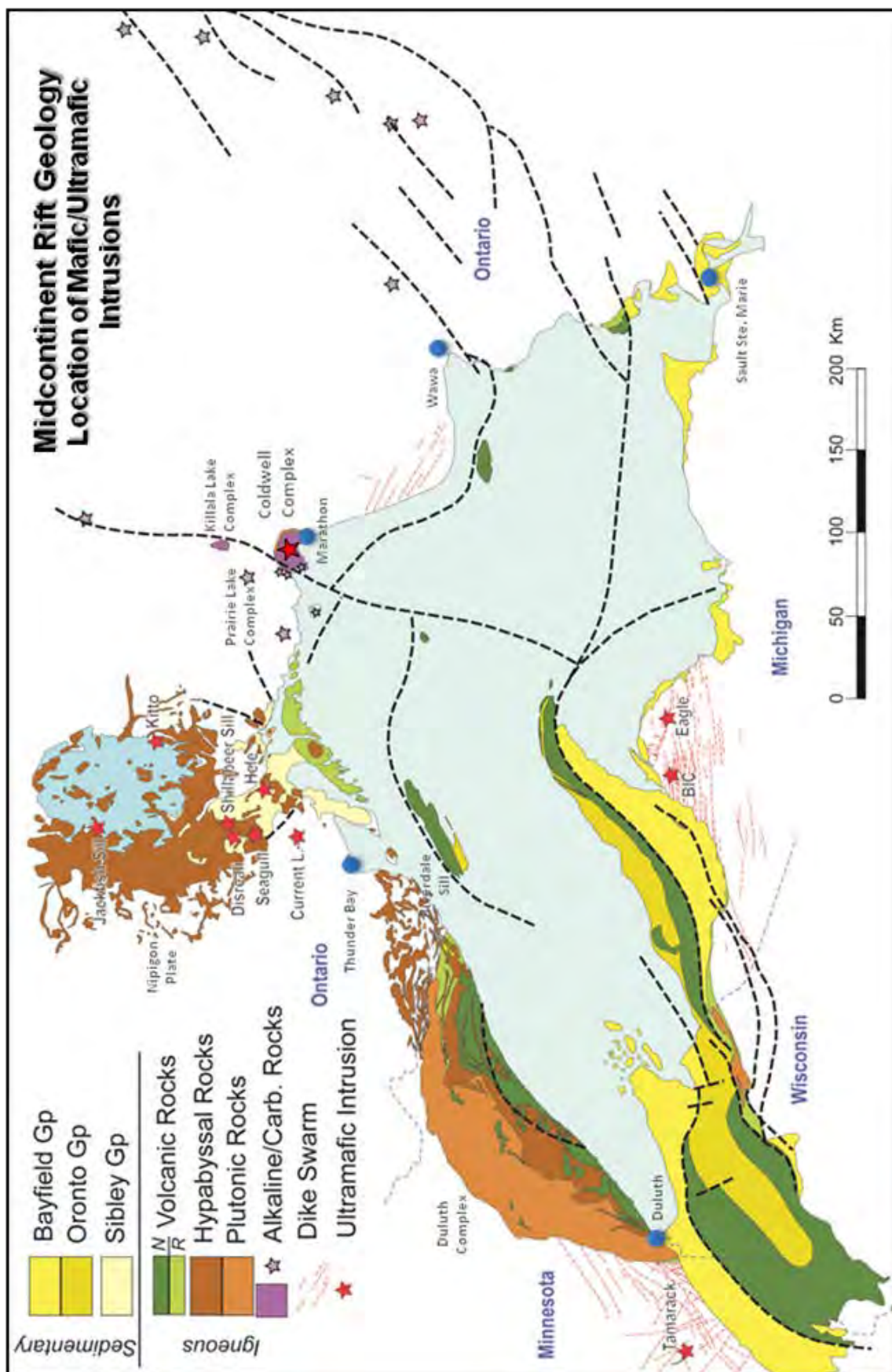


Figure 1. Mid-continent Rift geology and the locations of mafic/ultramafic intrusions (After Miller et al. 1995).

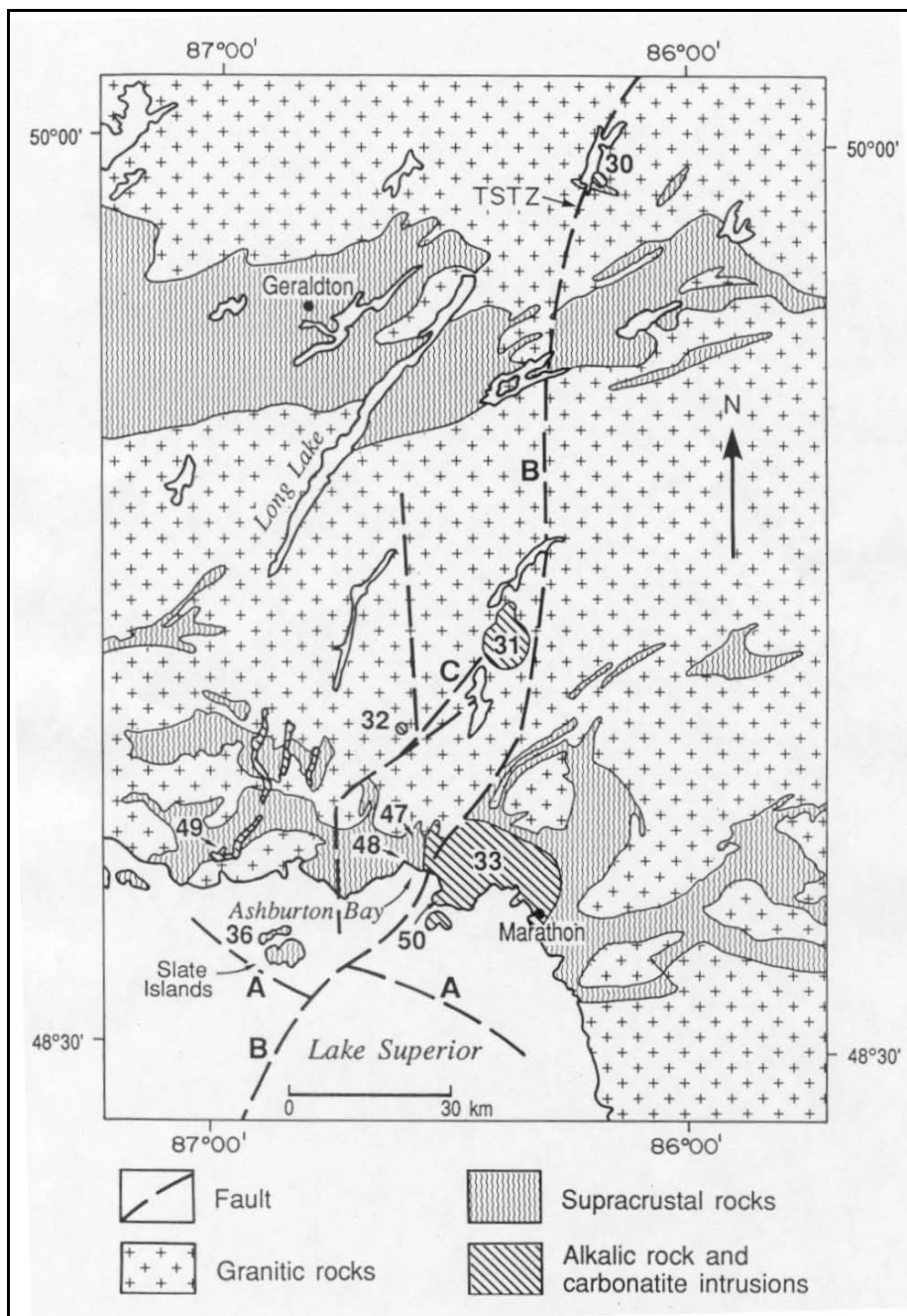


Figure 2. Regional geology (Sage 1991) in the vicinity of the Trans-Superior Tectonic Zone (TSTZ), extension of the Thiel Fault (B). Key to numbering: 30 – Chipman Lake fenites / carbonatite dikes; 31 – Killala Lake alkaline complex; 32 – Prairie Lake Carbonatite; 33 – Coldwell alkaline complex; 36 – Slate Islands; 47 – Dead Horse Creek diatremes; 48 – McKellar Creek diatreme; 49 – Gold Range Diatreme; 50 – Neys Diatreme; A – Michipicoten Fault; C – Killala Lake Deformation Zone.

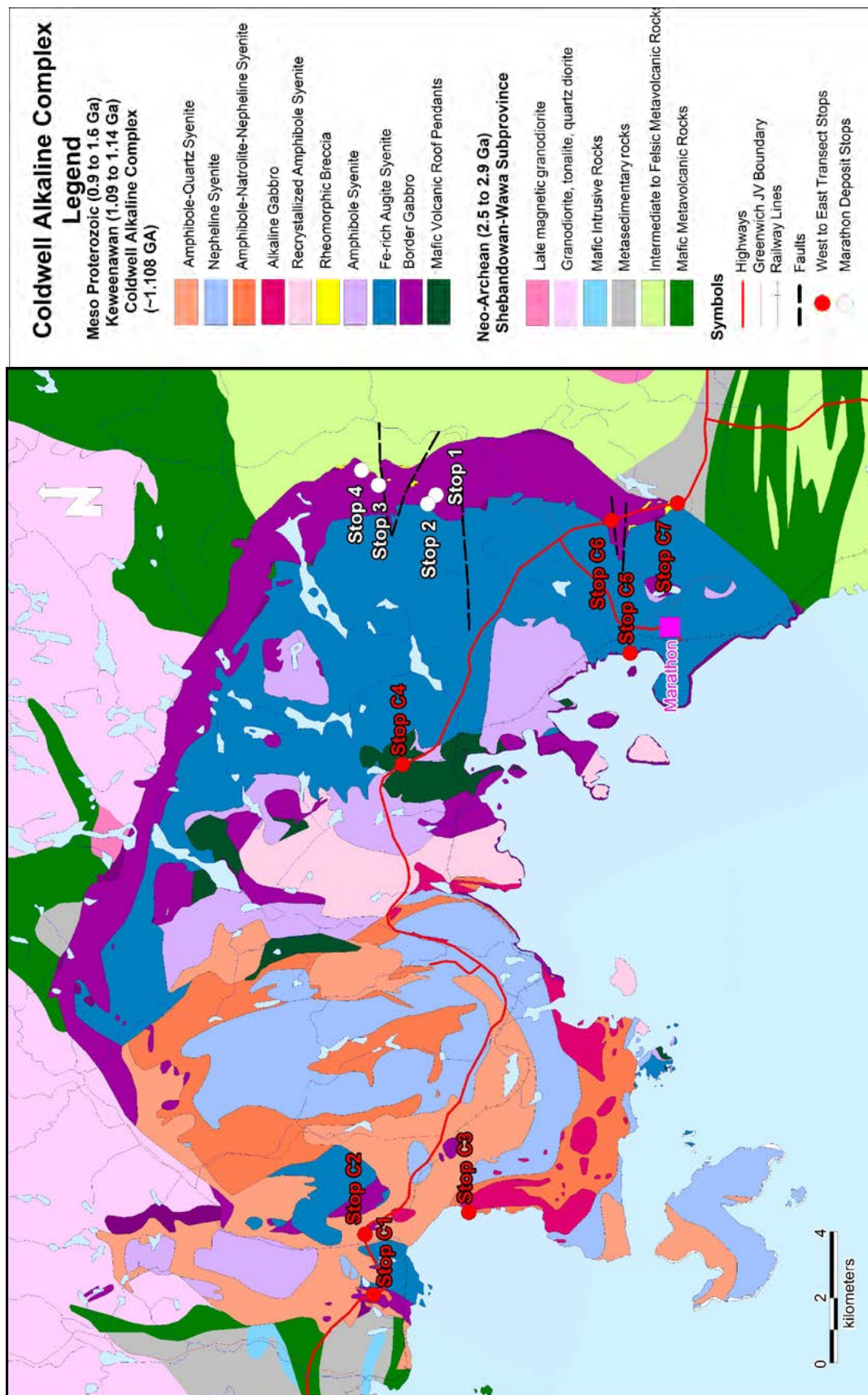


Figure 3. Generalized geology of the Coldwell Alkaline Complex (after Walker et al. 1993) with field trip stops.

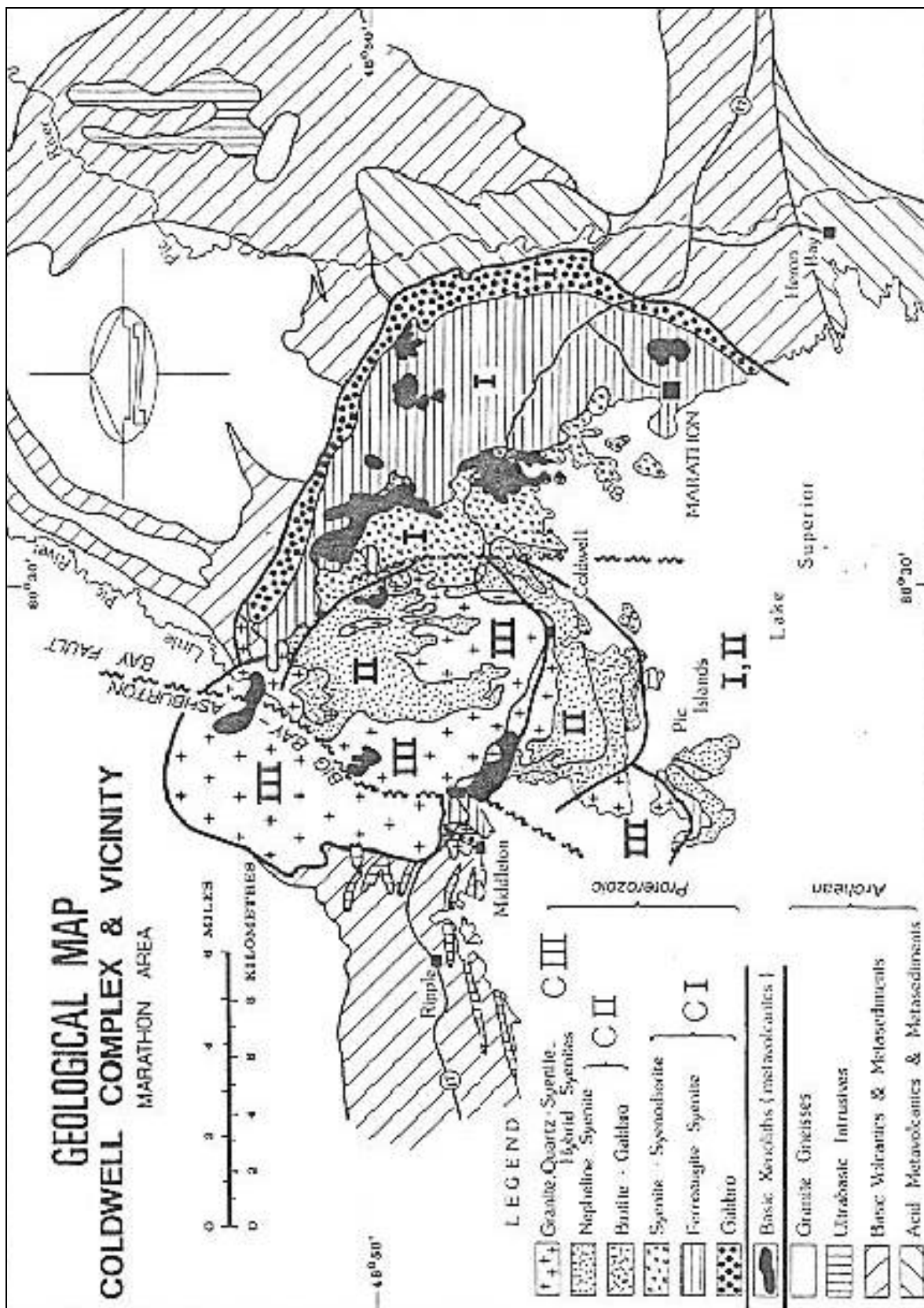


Figure 4. Coldwell Alkaline Complex magmatic centres: CI (Centre 1), CII (Centre 2), and CIII (Centre 3). Generalized geology after Mitchell and Platt, 1994.

Abundant large rafts and/or roof pendants of mafic volcanic rocks are mapped throughout the Coldwell Complex and in places exhibit horizontal extensional cooling cracks on a ten to hundreds of metres scale that are thought consistent by some workers with sub-horizontal bedding. For the most part the roof pendants may be the lowermost portions of the Keweenaw flood basalt sequences suggesting that the complex is barely unroofed and is exposed at a very shallow crustal level (Mitchell and Platt 1994; Sage and Watkinson 1995; Barrie et al. 2002). It is also highly probable that some, or most of the mafic rafts observed within the complex that could not be roof pendants are detached portions of chilled complex roof or wall rocks (fine-grained gabbros).

The mafic intrusive rocks occurring within Centers 1 and 2 are tabulated, with their associated mineralized zones, below:

Table 2

Intrusion (Centre)	Lithologic Units	Reference(s)
Eastern Gabbro (1)	Layered gabbro cumulates (olivine gabbro, gabbro, troctolite, anorthositic (leuco-) gabbro); Fe-Ti oxide $\pm$ apatite cumulates	Shaw (1994, 1997); Lum (1973); Barrie et al. (2002)
Western Gabbro (1)	Massive and layered series gabbro; olivine-bearing	Penczak (1992); Wilkinson (1983)
Two Duck Lake (1)	Gabbronorite, olivine gabbronorite, olivine-bearing gabbro, leucogabbro	Shaw (1994, 1997)
	Hornblende gabbro to monzodiorite; olivine ferrogabbro to ferrodiorite; olivine gabbro to diorite	Dahl et al. (1987)
Malpas Lake (1)	Amphibole-bearing olivine gabbro	Shaw (1994, 1997)
Geordie Lake (??)	Troctolite, olivine gabbro	Mulja (1989); MacTavish et al. (1987)
Alkalic gabbro (2)	Biotite gabbro	Mitchell and Platt (1982b)
	Biotite- and olivine-gabbro	Walker et al. (1993a)

Magmatic, gabbro-hosted Cu-Ni-PGE deposits in the Coldwell Complex have been the focus of much exploration and research for the past 60 years. Mineralized zones occur within the border gabbro at the eastern (Marathon deposit; Skipper Lake Zone) and western (Middleton occurrences) margins of the complex, and within its interior at Geordie Lake. The Geordie Lake mineralized zones, hosted by a younger (?) gabbro, are enriched in tellurium and silver and have higher Pd:Pt ratios (~19) (Mulja and Mitchell 1991) than the border gabbro-hosted deposits (~4) (Smyk 2001). Geochemical variations in mineralized zones in the Coldwell Complex are shown in Figure 5. A table of selected Coldwell Complex deposits and mineralized zones is shown below (see Table 3).

Table 3

Mineralized Zone	Grade / Significant Assays	Ore Mineralogy	Reference(s)
Marathon	Measured and Indicated In-Pit Resources : 114.8 Mt @ 0.775 g/t Pd, 0.228 g/t Pt, 0.083 g/t Au, 0.241% Cu, 1.567 g/t Ag; Proven and Probable In-Pit Reserves: 91.447 Mt @ 0.832 g/t Pd, 0.237 g/t Pt, 0.085 g/t Au, 0.247% Cu, 1.440 g/t Ag (January 2010)	Chalcopyrite > cubanite □ pyrite; hollingworthite, atokite-zvyaginstevite, sperrylite, Bi-kotulskite, michenerite, merenskyite, monceite, stibiopalladinite, paolovite, merteite II, palladoarsenide, unnamed (Pd ₅ As ₂), nickeline, majakite, argentine gold	Marathon PGM Corporation Ohnenstetter et al. (1991); Watkinson and Ohnenstetter (1992); Good and Crockett (1994a, 1994b)
Geordie Lake	Measured and Indicated Resources (above \$13.00/t cut-off): 32.42 Mt @ 0.61 g/t Pd, 0.04 g/t Pt, 0.05 g/t Au, 0.37% Cu, 2.93 g/t Ag	Chalcopyrite, bornite, pyrite, millerite, siegenite, pentlandite, galena, chalcocite, melonite, hessite, unnamed (Ag ₃ Te ₂), altaite, kotulskite, merenskyite, michenerite, sopcheite, Pd-bismuthotelluride, paolovite, Pd-arsenide, guanglinite, Pd-antimonide, sperrylite, electrum, Pd _{1.6} As _{1.5} Ni, AgSb ₄	news release, Marathon PGM Corporation, May 04, 2010 Mulja (1989); Mulja and Mitchell (1990, 1991)
Middleton		Chalcopyrite, pyrrhotite, pentlandite, sphalerite, pyrite	Penczak (1992)
Skipper Lake	average grade of 1.05 g/t Pd+Pt+Au over 12 m	Chalcopyrite, bornite, pentlandite, cobaltite, galena, chalcocite; telargpalite, polarite, kotulskite, taimyrite, merteite, zvyagintsevite, plumbopalladinite, majakite, tetraferroplatinum	MacTavish (2000)
Area 41	0.48 g/t Pt+Pd+Au over 202 m, incl. 1.23 g/t Pt+Pd+Au over 61 m	n/a	Benton Resources Corp.

Marked similarities exist between the mineralization style, geochemistry, and host rocks of Coldwell Complex-, Duluth Complex-, and the Crystal Lake gabbro-hosted deposits near Thunder Bay. Similarities include mineral textures, abundance and compositions, crystallization paths for the host gabbros, silicate-sulphide associations, trace-element trends and chalcophile element fractionation trends (Good and Crockett 1994a).

Research by Watkinson and Ohnenstetter (1992) and Good and Crockett (1994a, 1994b) produced debate between the relative importance of magmatic and hydrothermal processes in local copper-nickel-PGE mineralization processes. Watkinson and Ohnenstetter (1992) presented field, petrographic and mineral-chemical data that support the interaction of magmatic sulphide mineral assemblages with a chlorine-rich mixture of magmatic (deuteric) fluid and volatile species generated by the breakdown of assimilated xenoliths at low temperatures. However, Good and Crockett (1994a, 1994b) contended that element migration took place over only very short distances and that the original, bulk sulphides were not enriched in copper and PGE by later fluids.

The information within this field trip guide was taken from a variety of sources, including guidebooks from previous field trips to the Coldwell Complex: Puskas (1970); Loubat (1972); Mitchell and Platt (1977, 1982a, 1994); Smyk and Sage (1995), Smyk (2001), Smyk (2010), and unpublished field observations and mapping completed by A. MacTavish (1992). All UTM co-ordinates listed are NAD83 Zone 16.

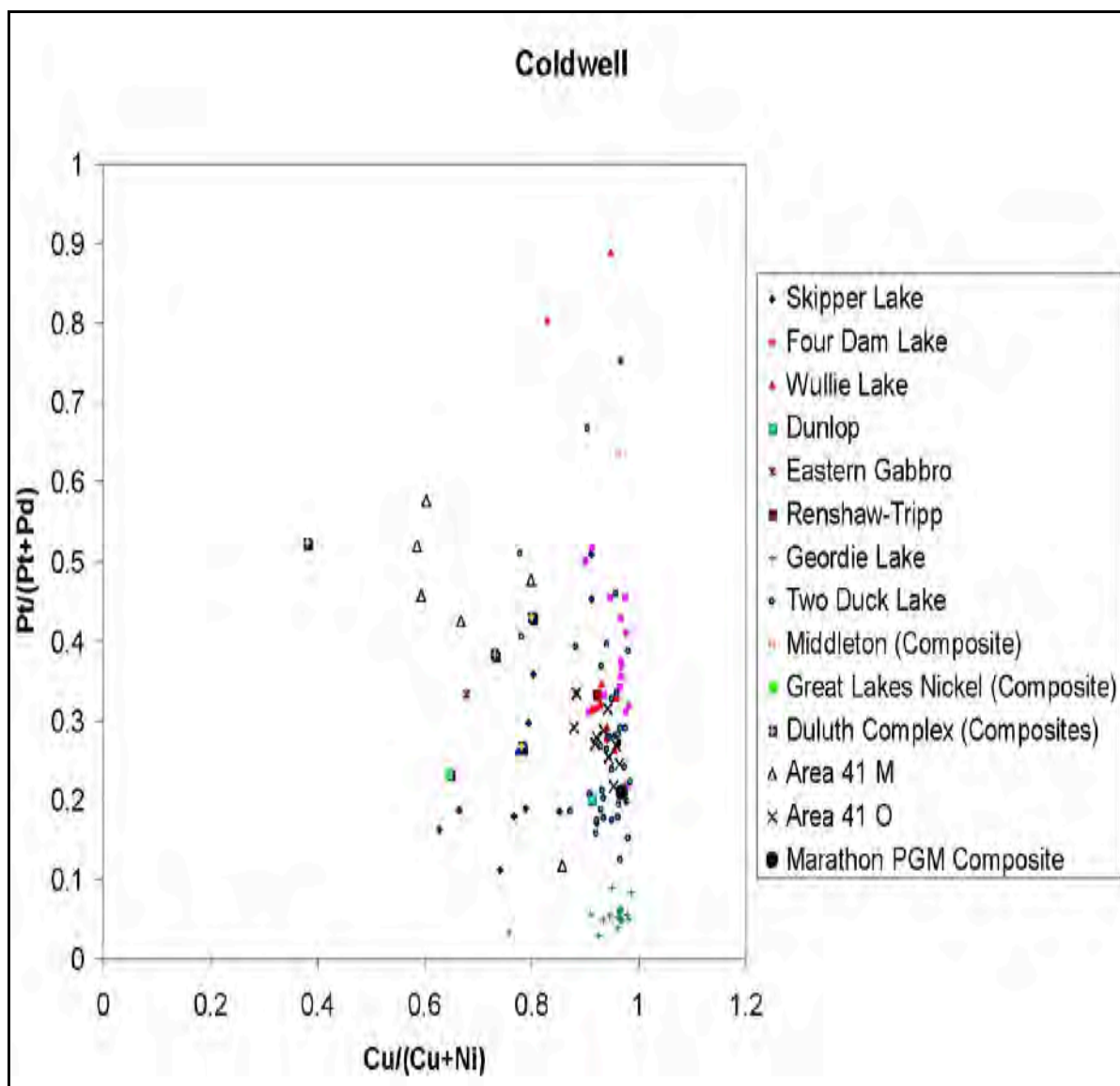


Figure 5. Discrimination plot for PGE-mineralized samples for Coldwell and other Midcontinent Rift-related intrusions. Data from Good (1992), MacTavish (unpublished data, Resident Geologist's Files, Thunder Bay), Watkinson et al. (1983), Wilkinson (1983), and unpublished data, Resident Geologist's Files, Thunder Bay. Duluth Complex composite data from Hauck et al. (1997).

Coldwell Alkaline Complex Transect - Field Trip Stops

STOP C1: Natrolite-Bearing Syenite and Massive Fe-Ti-oxides

Location: Lat: 48°47'45" N, Long: 86°39'15"W; UTM 525528E, 5405511N
29.4 to 29.9km west of the Highway 626 and Highway 17 junction

Description: This exposure displays natrolite-bearing, pegmatitic syenite (see Photo 1). Reddish orange natrolite (an acicular or prismatic zeolite mineral replacing nepheline) patches up to 15cm in diameter, crystals of perthitic feldspar up to 30cm in length, and crystals or black amphibole up to 25cm in length comprise the bulk of this syenite (see Photo 2). Mitchell and Platt (1994) reported accessory pleochroic clinopyroxene, zircon, titanite, and biotite. Natrolite has locally been ascribed to the hydrothermal alteration of primary nepheline and has also been referred to as “hydronepheline” by local workers. The syenite is intruded by a camptonite lamprophyre dike (Mitchell and Platt 1994) and also hosts large, medium-grained gabbro xenoliths (see Photo 3), up to 1m in thickness and sometimes up to 5m in length (west-side of the highway), that exhibit 1 to 2cm wide, dark reaction rims adjacent to the enclosing syenite. To the east, the pegmatitic syenite gives way to finer grained nepheline syenite in which chalky-weathering nepheline may be observed. Rare natrolite grains are also present. Farther east, a variety of equigranular and pegmatitic syenites are exposed.

Near the eastern end of the outcrop (UTM 525625E, 5404825N), a large xenolith of gabbro-hosted, massive titaniferous magnetite has been exposed. Minor clinopyroxene, plagioclase and apatite occur within the massive oxide unit. Analyses completed in 1951 and reported by Hinz and Landry (1994) indicated total iron and titanium values ranging between 33 and 45%, and 4.5 and 13.5%, respectively; phosphorus contents ranged up to 0.371%.



Photo 1. Pink, natrolite-bearing, pegmatitic augite syenite. Photo credit D. Campbell.



Photo 2. Pegmatitic syenite containing reddish orange patches of natrolite, light pinkish perthitic feldspar, and black amphibole. Photo credit A. MacTavish.



Photo 3. Large gabbro xenolith located on the west side of the highway. Please note that the xenolith has been crosscut by fine-grained syenite veins and that the syenite below the xenolith is varitextured to pegmatitic in texture, whereas the syenite above is medium- to coarse-grained. Photo credit A. MacTavish.

STOP C2: Little Pic River Breccia Zone

Location: Lat: 48°47'45"N, Long: 86°37'30"W; UTM 527478E, 5405531N
27.3 km west of the Highway 626 and Highway 17 junction

Description: These road cuts, particularly along the south side of the highway, expose spectacular intrusive breccias within the youngest rocks of the complex, along the east side of the fault zone that the Little Pic River occupies. The breccias often occur as semi-continuous, fragmented, elongate rafts (western end of southern rock cut, see Photo 4) that consist of angular to rounded blocks of fine- to medium-grained, equigranular, mafic (gabbroic?) rocks within a groundmass of pink, medium-grained, quartz syenite. In some cases blocks can exhibit both angular and lobate to cusped (amoeboid) margins (see Photo 5). The mafic rocks comprising the blocks were interpreted as oligoclase-bearing basalt by Mitchell and Platt (1982a). Subsequent discussion and study has led to the suggestion of perhaps 2 texturally discernable types of basic xenoliths, those with: (1) sharp, angular margins, and (2) those with lobate to cusped margins. In this model, the angular xenoliths represent synplutonic basalts which are now preserved elsewhere as megaxenoliths in younger intrusions. The cusped-margined xenoliths may represent the effects of mixing between 2 contemporaneous gabbroic/basaltic and syenite magmas (i.e., magma mixing or co-mingling). Cusped, possible chilled margins with quench-textured clinopyroxene, plagioclase and skeletal olivine have been noted in similar xenoliths to the south on the Coldwell Peninsula by G. Shore (personal communication with M. Smyk, 1995) and suggest the quenching of the basic magma against the cooler, syenitic magma. These are reasonable hypotheses and there are definitely at least 2 types and textures of xenoliths; however, they do not completely explain the presence of blocks exhibiting both margin types as observed by the senior author of this guide and shown in Photo 5. Texturally there also seems to be 3 different types of xenoliths: the most abundant are dark grey to black, very fine-grained xenoliths (Photo 4); medium-grained, greyish pink xenoliths with somewhat less distinct, but still relatively sharp margins (Photo 5), and several unusual zones where there is a subvertical zone of rounded, dark grey, amphibole-phyric xenoliths within a pinkish, mafic groundmass. Are these some sort of breccia dykes or just hybridized zones of xenoliths (see Photo 6; what do you think?)? Although isolated xenoliths are common, there are many areas within these outcrops where incipient or in-situ brecciation characterized by syenite dykes and “jig-saw puzzle/jig-saw fit” breccias are observed, where brecciated fragments can be fitted back together. Mirolitic cavities, up to several centimetres in width, contain euhedral quartz, feldspar, and calcite crystals.

The breccia zone persists to the east, towards the scenic lookout located 800m to the east. The south side of the highway is underlain by oligoclase gabbro and quartz syenite, while various, xenolithic-bearing syenitic rocks are exposed on the north side. These pyroxene- and amphibole-(ferro-edenite) bearing syenites contain xenoliths of alkali gabbro, alkali diorite and other, equigranular to porphyritic syenites. Near the lookout turnoff, gray, nepheline-bearing syenite intrudes the mafic rocks and contains orange natrolite. Sannaite and ocellar, camptonitic lamprophyre dikes have been reported near this site by Mitchell and Platt (1994) who proposed the following order of local emplacement:

Mg-hornblende syenite → contaminated Fe-edenite syenite → Fe-edenite syenite → quartz syenite (earliest → latest)

Lukosi-Sanders (1988) classified the local rocks as miaskitic, metaluminous syenites enriched in U, Th, REE and Zr. These syenites have affinities to A-type granites and have been interpreted to be the result of fractional crystallization of mantle-derived, basaltic magma (Lukosi-Sanders 1988; Mitchell et al. 1993).



Photo 4. Syenite outcrop on the south side of the highway containing blocks of mafic xenoliths often occurring in elongated, semi-continuous rafts. Photo credit A. MacTavish.



Photo 5. Elongated, jig-saw-fit xenolithic block exhibiting both angular and lobate/cusate (amoeboid) margins. Please note the lighter-coloured, pinkish-grey elongated xenolithic block with apparently sharp margins located below the dark grey block. This lower block appears to be coarser-grained and may possibly be different in composition. Photo credit A. MacTavish.



Photo 6. Rounded dark grey, amphibole-phyric xenoliths/inclusions within a fine- to medium-grained, pinkish mafic groundmass (breccia dyke?). Photo credit M. Puumala.

Stops C3: Prisoners Cove, Neys Provincial Park (Sample Collecting Prohibited!). The descriptions (unpublished mapping/field descriptions, MacTavish 1992) herein are for 14 sub-stops along the shoreline south of Prisoner's Cove; however, due to time constraints only the northernmost stops will be visited.

Location: Lat: 48°46'30"N, Long: 86°37'10"W; UTM 527984E, 5402537N;
23.5km west of the Hwy 626 and Hwy 17 junction; 2.8km south of Hwy 17 to the park headquarters and then south along the shoreline trail

General Description: The wave-washed, glacially polished outcrops along the shoreline of Lake Superior at Prisoner Cove and south for over a kilometre along the western side of the Coldwell Peninsula exhibit a variety of lithologic, textural, and crosscutting features that characterize much of the Center 2 magmatism in the Coldwell Complex. In its simplest sense, this composite stop displays the contact between alkalic biotite gabbro and amphibole-nepheline syenite, but the enigmatic effects of assimilation and hybridization have severely complicated and obscured many of the primary features. In all cases within the nepheline syenitic rocks exposed along the shoreline at this stop the nepheline has been completely altered to the zeolite mineral natrolite which weathers to orange-coloured pits.

Medium- to coarse-grained, olivine- and enclave-bearing, biotite gabbro comprises much of the eastern portion of the outcrops. Gabbro xenoliths occur within the syenite and within hybrid phases along their mutual contact, which trends roughly north-south, parallel to the shoreline. The outcrops often exhibit a pitted surface resulting from the preferential weathering of mafic enclaves consisting of biotite-olivine gabbro to biotite-clinopyroxene gabbro or leucogabbro (Walker et al. 1992) within a more syenitic groundmass. The syenitic groundmass consists of fine- to coarse-grained nepheline (altered to natrolite) syenite with minor acicular amphibole and poikilitic biotite. Mitchell and Platt (1994) have identified the amphibole as hastingsite.

Distinct to diffuse layering, a nebulous to locally distinct igneous foliation, and localized soft-sediment style magmatic deformation exists within the amphibole-nepheline syenite. Identifiable, undisturbed layering may be oriented parallel to the apparent syenite/gabbro contact and dips from very steeply to vertically in the north to shallow to moderate to the east (where measurable) in the south. Observed soft-sediment deformation features consist of flame structures, fluid-escape features, slump folds, and isolated well-layered syenite blocks surrounded by obvious fluid escape textures. Much past discussion has focused on whether the observed structures have resulted simply from igneous process, syn- or post-intrusion shearing, or a combination of these processes. The present author's strongly favour igneous processes since the observed fracturing is very localized, is late and brittle, and does not appear to have affected the foliation or layering within the surrounding rocks in any observable way. It is highly probable that the crystallizing magma chamber was often shocked by MCR tectonic activity. These earthquakes then caused the slumping of unstable crystallizing layers along chamber walls; allowed the isolation of broken, but relatively intact layered blocks; and allowed trapped deuteritic fluids formed during the fractionation process to escape upwards through the broken layers. Upon close examination the fracturing presently observed in outcrop obviously took place after the chamber was completely crystallized and was able to deform in a brittle manner.

Sub-Stop C3a (527980E, 5402571N): This area, located on the point to the north and west of the old flat-bottomed boats, mainly consists of foliated, wispy, hybridized amphibole-nepheline syenite with diffuse discontinuous "layers" (see Photo 7). The core of this outcrop is flanked to the northeast by a heterogeneous zone containing large numbers of rounded to angular, variably assimilated (metasomatized?) and disaggregated inclusions/xenoliths of biotite gabbro. Reaction rims around these inclusions are readily visible. Also the inclusion-rich zone, as a whole, seems to be enclosed within a diffuse reaction zone when compared to the hybrid syenites adjacent to the west. The western margin of the exposure is a medium- to coarse-grained hybridized syenite with numerous very coarse-grained to pegmatitic inclusions of amphibole-nepheline syenite. At the northwestern tip of the outcrop is an elongate, diffuse zone of apparently non-hybridized, non-foliated syenite (possibly the original parent syenite?).

Sub-Stop C3b (527950E, 5402535N): This stop, located 30m west-southwest of the old boats near the shoreline, consists of a 4 to 5m wide, west-northwest-striking, brittle fracture zone hosting a 70 to 100cm thick, dark greenish-grey, ocellar lamprophyre dyke at its northern margin near the water's edge. The ocellae present within the dyke are composed of reddish, recessive-weathering carbonate ($\pm$ zeolites?) which are elongated parallel to dyke margins (elongated by flow?). The lamprophyre dyke is also enveloped by a brick-red alteration halo that is not completely within the fracture zone and also extends into the unfractured hybrid syenites to the north for up to 5m. This red halo could be due to either hematization or K-alteration. Similar, subparallel fracture zones can also be observed 10m and 23m to the south.

Sub-Stop C3c (527966E, 5402475N): This stop is located ~50m east-southeast of Sub-stop C3b, and consists of a zone of large blocks (?) of coarse-grained, natrolite-bearing, biotite gabbro to biotite melagabbro that are surrounded by fine- to medium-grained amphibole-nepheline syenite containing diffuse gabbro xenolith ghosts. It is distinctly possible that this is not a zone of xenoliths/inclusions at all, but the exposed upper contact of an underlying biotite gabbro body that is part of the biotite gabbro body located about 40m to the southeast (see Sub-Stop C3e, below) where the syenite is observed to overly the gabbro. These blocks (?) are cross-cut by narrow horizontal and subvertical syenite veins and dykes (see Photo 8).



Photo 7. Wispy, relatively mafic in appearance, hybrid amphibole-nepheline syenite exhibiting diffuse discontinuous layers. Photo credit A. MacTavish.



Photo 8. Large biotite gabbro xenolith/inclusion (?) crosscut by veins and dykes of amphibole-nepheline syenite. Photo credit A. MacTavish.

Sub-Stop C3d (528004E, 5402440N): This location (45m southeast of Sub-stop C3c) consists of an irregular zone of gabbro xenoliths/inclusions, surrounded by fine- to medium-grained, weakly foliated syenite (see Photos 9 and 10). The xenoliths are in the process of being broken down in stages from originally angular, cohesive blocks to diffuse groupings of amoeboid to wispy mafic remnants within a mafic mineral-rich, hybrid syenite. This process is probably not assimilation in the strictest sense, but is more likely a process of chemical (rather than thermal) invasion through metasomatism that over time breaks down the xenoliths and then eventually disaggregates the mineral constituents of the blocks to the point where they are then assimilated into the syenite melt. The hybridized (?) syenite surrounding the xenoliths exhibit aligned amphibole grains that may indicate flow (?) around and between fragments. There are also places where there are noticeable (up to 15cm thick) halos surrounding zones of xenoliths that consist of aligned amphibole grains that are somewhat separated into diffuse bands.

Sub-Stop C3e (528014E, 5402433N): Located only 12m southeast of Sub-Stop 3d and consists of coarse-grained, knobby-weathering, biotite gabbro that has been cross-cut by numerous hair thin to 5cm thick, very fine- to fine-grained syenite stringers and veins and the occasional, larger, fine-grained to pegmatitic syenite vein (pegmatite is in centre of these veins) (see Photo 11). There are numerous leucocratic clots (oikocrysts?) of plagioclase (see Photo 12) throughout.



Photo 9. Biotite gabbro xenoliths/inclusions separating from and original larger block and beginning to assimilate (?) into the surrounding syenite melt through a process of metasomatism and disaggregation. Photo credit A. MacTavish.



Photo 10. Disaggregating biotite gabbro xenoliths exhibiting subparallel reaction haloes. Photo credit M. Puumala.



Photo 11. Biotite gabbro crosscut by fine-grained to pegmatitic syenite dyke (centre) and thinner syenite veins (centre left). Photo credit A. MacTavish.



Photo 12. Leucocratic clots of plagioclase (oikocrysts) within biotite gabbro. Photo credit A. MacTavish.

Sub-Stop C3f (527982E, 5402324N): This sub-stop (~115m west-southwest of Sub-stop C3e) consists of an irregular, variably assimilated zone of mafic (gabbroic?) xenoliths of highly variable size ranges. Many blocks are in the last stages of assimilation where the original xenoliths are now merely ghosts infilled with isolated mafic remnants and considerable numbers of hornblende grains.

Sub-Stop C3g (527978E, 5402292N): This location (~30m south of Sub-stop C3f) consists of locally well-developed modal layering within medium- to locally coarse-grained amphibole-nepheline syenite. The magmatic layering dips shallowly to the west and west-southwest at between 20 and 26° and there is a possible weak alignment of K-feldspar laths parallel to layering. The bases of the undulating modal layers are defined by a sharp increase in amphibole content. The best defined layering is near the lake with layering becoming increasingly more diffuse, disrupted, folded (slumping?), and contorted to the east until it becomes unrecognizable.

Sub-Stop C3h (527971E, 5402265N): At this location (~27m south of Sub-stop C3g) are 2, sub-parallel, aphanitic to fine-grained, ocellar lamprophyre dykes (see Photo 13) occupying a narrow southeast-striking fracture zone. The dykes dip to the northeast between 54° and subvertical. The ocellae (immiscible liquid droplets) are usually centralized within the dykes away from the strongly chilled dyke margins and are infilled with several minerals including apple-green and greyish minerals (zeolites?), and possibly white calcite.



Photo 13. Ocellar lamprophyre dyke in narrow fracture zone. Photo credit A. MacTavish

Sub-Stop C3i (527987E, 5402198N): At this location (~70m south of Sub-stop C3h) is a zone of leopard mottles in moderately mafic, often grain-size-layered (?) amphibole nepheline syenite.

Sub-Stop C3j (527971E, 5402265N): This location (~80m south of Sub-stop C3i) is, for lack of a better name, a “Layer Breccia Zone” where there has been strong disruption, localized rotation, and folding of original magmatic syenite layers. Finer-grained syenite containing acicular amphibole grains has flowed around the layer blocks and alignment of those amphibole grains mirrors flow directions. The zone is surrounded by a highly disturbed hybrid mixtite with few measurable features. Thinner blocks are composed of a series of thin modal layers of highly variable textures. The thicker layers are usually the coarsest, are sometimes size-graded, and contain glomerocrysts of K-feldspar (with include amphibole and natrolite after nepheline) up to 1.5cm in diameter surrounded by acicular amphibole grains and recessive-weathering altered nepheline.

Sub-Stop C3k (528000E, 5402039N): Located 77m south of Sub-stop C3j. This sub-stop consists of a well-layered block of amphibole-nepheline syenite (~6.5m by 3.5m in dimensions) that is surrounded by a highly distorted zone of fine- to very-coarse-grained (varitextured) syenitic material that appears to have flowed around the block. This isolated block is composed of a sequence of 3 thick layers where amphibole and K-feldspar are aligned subparallel to layer bases. The base of each layer is undulatory on the scale of a single very coarse feldspar crystal.

Sub-Stop C3l (528004E, 5402016N): A further 23m south of Sub-stop C3k is a zone characterized by well-developed syenite layering (see Photo 14), some possible magmatic channel scours, some localized soft-sediment-style deformation, and a few zones of intense, localized layer disruption. Most of the layers within the southern part of the zone are quite flat lying (18°W dip).



Photo 14. Well-developed, relatively flat-lying magmatic layering within amphibole-nepheline syenite. Photo credit A. MacTavish

Sub-Stop C3m (528011E, 5402006N): This sub-stop is located 12m southeast of Sub-stop C3l and is directly adjacent to it. It consists of a zone of disturbed and distorted syenite layering similar to that observed north of the isolated block observed at Sub-stop C3k. Contorted and convoluted layering is common and folding is observed locally with the disturbance increasing in intensity to the south. Most noticeable in this area is a deformed, crosscutting, texturally variable, vein-like body (see Photo 15) composed of mobile “flow-banded” material. The “vein” margins are often irregular, possibly due to volatile fluid seepage (?) and it is often cored by coarse-grained to pegmatitic veinlets and pods. It is possible that this structure has erupted from the nose of a slump fold.

Sub-Stop C3n (528020E, 5401977N): This final sub-stop is located 30m east-southeast of Sub-stop C3m and consists of a large slump-fold composed of medium- to very coarse-grained, modally- and normally grain-size graded syenite layers (see Photo 16). This was interpreted as slump folding due to the presence of at least 3 axial planar directions present within 3 separate folds all in close proximity to each other. Unfortunately since mapping was completed in 1992 this exposure has become partially obscured by the growth of lichen.



Photo 15. Crosscutting, vein-like body possibly resulting from the movement of volatile-rich magmatic fluids. Photo credit A. MacTavish.

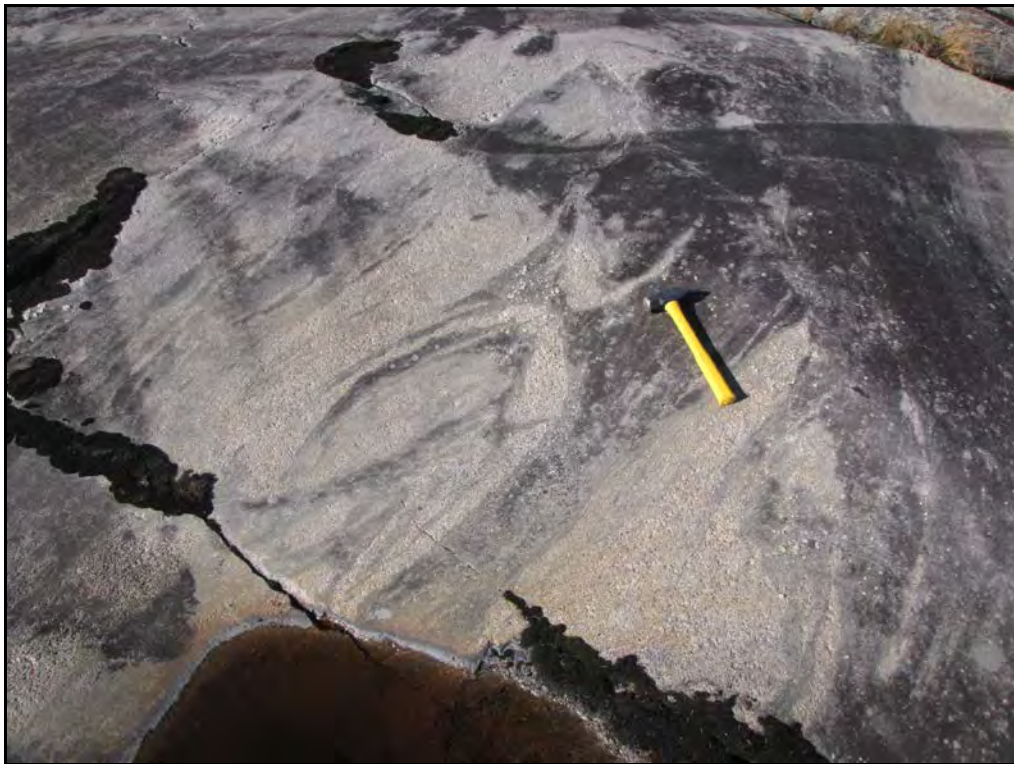


Photo 16. Lichen-obsured slump fold within layered amphibole-nepheline syenite. Photo credit A. MacTavish.

STOP C4: Hornfelsed Basaltic Roof Pendants, Wolf Camp Lake

Location: Lat: 48°47'40"N, Long: 86°26'05"W; UTM 541775E, 5404189N
8.6km west of the Highway 626 and Highway 17 junction

Description: Hornfelsed basaltic rocks overlying the complex were recognized early in its mapping by Tuominen (1967) and Puskas (1970) and likely represent a volcanic edifice that has been subsequently eroded (Sage 1986). Mitchell and Platt (1994) and Nicol (1990) have considered these basalts to have a tholeiitic lineage, contemporaneous with the Coldwell Complex. Fresh, metasomatized and hornfelsed, andesine-oligoclase basalt flows are estimated to attain a thickness of 5km (Mitchell and Platt 1994; Nicol 1990). Assimilation and brecciation of the flows by subsequent gabbroic to syenitic magmatism has resulted in the widespread development of basaltic xenoliths ranging from 1m to over 1km in size, comprising a roof pendant in the central part of the complex (Walker et al. 1992). Walker et al. (1992) subdivided these basaltic rocks into 3 main units:

1. Aphanitic to fine-grained, massive, locally amygdaloidal (?) / ocellar basalt;
2. Medium-grained, diabasic (ophitic) basalt; and
3. Aphanitic to medium-grained, feldspar-phyric, diabasic (ophitic) basalt.

At Wolf Camp Lake, aphanitic basalts contain round to amoeboid, epidote- and quartz-filled structures up to 2cm in diameter that have been interpreted as amygdulites (see Photo 17). Well-defined, amygdulite-bearing zones dip 8° to the southwest in this vicinity (Walker et al. 1992). The basaltic roof pendant is locally underlain and enveloped by feldspar-phyric amphibole syenite and Fe-rich augite syenite.



Photo 17. Amygdulites within the basaltic roof pendant located near Wolf Camp Lake. Photo credit D. Campbell.

STOP C5: Layered Fe-rich Augite Syenite (Alternate stop if time allows)

Location: Lat: 48°44'20"N, Long: 86°23'25"W; UTM 544782E, 5398443N

680m west along the shoreline of Lake Superior from the end of the James River industrial road along the waterfront in Marathon; OR 150m south of Carden Cove road, 0.3km past CPR tracks (park at 544864E, 5398750N)

Description: Broad expanses of glacially polished and wave-washed, massive Fe-rich augite syenite occur all along this part of the Lake Superior shoreline near Marathon. Fresh surfaces vary from dark green-brown to black, despite a buff to white weathered surface. Small dimension stone quarries were developed and produced in this area during the 1930's. Much of the stone was shipped to larger centres in the American mid-west and Toronto.

Fe-rich augite syenite (formerly referred to as ferroaugite syenite) comprises a large portion of the exposure in the eastern half of the Coldwell Complex. It appears to be a sheet-like intrusion that dips approximately 15° toward the center of the complex, sandwiched between the underlying Eastern Border Gabbro and an overlying, recrystallized amphibole-quartz syenite; it also intrudes the basaltic roof pendants (Walker et al. 1992; 1993a). Crystallization of the syenite inwards from its upper and lower contacts produced mineralogical and compositional variations across it (Walker et al. 1993a). Constituent minerals include iridescent, lathlike, cryptoperthitic feldspar (up to 30% interstitial), and variable amounts of fayalite, amphibole, aenigmatite, and rare quartz. Coarse-grained to pegmatitic portions of the syenite host a variety of REE-bearing fluoro-carbonates, quartz, chalcedony, and molybdenite. Iridescent feldspar, known locally as "spectrolite", was recently (2010) commercially extracted on a very small-scale from pegmatite at Shack Lake near Marathon.

Although this unit is typically massive, rhythmic to chaotic layering is locally developed and where observed commonly dips shallowly towards the centre of the complex. At this site, layering strikes at 070° and dips 60° north. The layering is unusual in that it is defined by an intercumulus mineral (augite) rather than by cumulus phases (feldspar).

STOP C6: Layered Eastern (Border) Gabbro

Location: Lat: 48°44'00"N, Long: 86°20'00"W; UTM 549199E, 5398010N

1.7 km east of the Highway 626 and Highway 17 junction

Description: Layering in the Eastern Border Gabbro shows distinct variations in style, is usually parallel to the eastern contact of the gabbro, and dips 20° to 60° toward the center of the complex (Shaw 1994; 1997). At this stop, layering strikes approximately north and dips west towards the rest of the complex at ~45°. This thickly layered sequence is underlain by massive gabbro near the contact with the Archean country rocks. The macrorhythmic layering is laterally discontinuous, pinching out over distances of 5 to 10m and contacts are sharp and conformable (Shaw 1994; 1997). Rhythmic layering is modal and has been related to variation in the respective proportions of plagioclase (An₆₀₋₃₅), augite (Fo₆₇₋₄₃), minor orthopyroxene (En₅₅₋₆₆), and Fe-Ti-oxides by Lum (1973). Modal plagioclase varies from approximately 60 to 80% in the leucocratic layers and 20 to 35% in the meso- to melanocratic layers (Shaw 1994). A second band of layered gabbro, separated from the first by massive gabbro, is exposed on top of the long rock cut (see Photo 18). Here, the macrorhythmic layering (see Photo 19) produces relatively thin (1 to 5cm) to medium thick (5 to 100cm) layers that can be traced for over 35m along strike. Layer contacts are sharp, locally scalloped and conformable. Trough cross-bedding has been noted on vertical faces by Shaw (1994). This stop is also close to the contact between the Eastern Border Gabbro and the Fe-rich augite syenite to the west. Pegmatitic syenite dykes intrude the gabbro at this locality and contain miarolitic cavities. McLaughlin (1990) has reported the presence of a variety of REE-bearing fluorocarbonates (bastnaesite, parisite, synchisite), Nb-bearing phases, and zircon in pegmatitic syenite with quartz, feldspar, and sodic amphibole.



Photo 18. Macrorhythmic layering within the Eastern Border Gabbro. Photo credit M. Smyk.



Photo 19. Macrorhythmic modal layering within the Eastern Border Gabbro. Photo credit A. MacTavish.

STOP C7 (Alternate Stop): Eastern Contact of the Coldwell Complex

Location: Lat: 48°43'10"N, Long: 86°19'30"W; UTM 549656E, 5396238N
3.3 to 3.6km east of the Highway 626 and Highway 17 junction

Description: A number of highway rock cuts and outcrops expose the eastern contact of the Coldwell Complex with the enclosing Archean greenstone belt country rocks. Center 1 gabbros, the oldest rocks of the complex, form a ring dyke that forms the eastern and northern margins of the complex where it is in contact with Archean supracrustal and granitoid rocks. The reverse magnetization of these gabbros (Lilley 1964) produces prominent magnetic “lows” on aeromagnetic maps. The most recent and comprehensive study of the Eastern Border Gabbro was conducted by Shaw (1994, 1997) who noted that more than 90% of the unit consists of layered gabbro.

At this location, varitextured, unlayered Eastern Border Gabbro is in contact with, and contains numerous xenoliths of, Archean metasedimentary rocks. This has produced hybrid and contaminated phases and rheomorphic breccia. Crosscutting Center 1 syenite dikes are commonly pegmatitic. Amethystine quartz, calcite, and molybdenite occur in vugs within this chaotic contact zone.

Disseminated iron- and copper-sulphides occur in biotite-rich, varitextured gabbro (Dunlop Occurrence), which has experienced sporadic exploration since the discovery of copper in the early 1950's. It was last drilled in 1992 by Noranda Inc. with the best assay intervals grading 0.35% Cu/6.0m and 0.42% Cu/4.0m, respectively (Resident Geologist's Files, Thunder Bay). A grab sample of rusty-weathering, moderately magnetic, fine- to medium-grained gabbro with coarse biotite and blebby chalcopyrite graded 5090ppm Cu, 494ppm Ni, 241ppm Zn, 8ppb Pd, 2ppb Pt and 22ppb Au (ibid). Overgrown pits are located just inside the tree line, west of the highway (UTM 549575E, 5396290N).

Shaw (1994; 1997), Walker et al. (1993a, 1993b, 1993c), Currie (1980), and Tucker (1995) have documented a number of occurrences of rheomorphic breccia associated with the Eastern Border Gabbro along its intrusive, basal contact with the Archean supracrustal country rocks. Breccia units are characterized by chaotic flow fabrics that surround flow-oriented clasts situated in a medium-grained, granitic matrix. This unit has been somewhat enigmatic, having been alternatively described by earlier workers as conglomerate and ignimbrite (Resident Geologist's Files, Thunder Bay). Similar exposures of this map unit also occur along the western contact of the complex, north of Middleton (cf. Wilkinson 1983).

Locally, pods of breccia vary from 20 to 75m in width and are up to 250m long. The breccia exposed along Highway 17 at this site contains mainly hornfelsed Archean clastic metasedimentary and metavolcanic rocks and massive vein quartz. In the vicinity of Two Duck Lake, the breccia contains fine-grained gabbro clasts (Tucker 1995). The breccia varies from clast- to matrix-supported; the matrix consists of equigranular quartz, feldspar, and minor biotite, clino- and orthopyroxene, and opaque minerals; and tourmaline and prehnite overgrowths have been noted (Tucker 1995). Rounded to angular clasts range in size from 0.5 to over 100cm and locally have developed 1 to 2cm wide, chlorite-rich reaction rims that are thickest where they are matrix-supported (Shaw 1994). Magnetite and quartz-feldspar-tourmaline veins cut both matrix and clasts. Quartz-feldspathic rinds and crosscutting veinlets have been interpreted to be the result of partial melting of the felsic material during assimilation. The close association between rheomorphic breccia and the Eastern Border Gabbro suggests that the intrusion of the gabbro led to the brecciation and partial melting of the country rocks (Shaw 1994, 1997; Tucker 1995).

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PART 2: MARATHON CU-PGM DEPOSIT

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Introduction to the Marathon Deposit:

The Cu-PGM sulphide mineralization of the Marathon deposit is hosted by the Two Duck Lake Gabbro, the latest mafic intrusive event and consequently the most continuous gabbroic body within the Eastern Gabbro Suite at the Marathon deposit.

The Eastern Gabbro Suite, located around the eastern and northern margin of the Coldwell, was composed initially of a thick sequence of tholeiitic basalt that was subsequently intruded by a much larger volume of leucocratic to ultramafic intrusions that caused contact metamorphism of the basalt to pyroxene-hornfels grade (Good et al., 2015). All of these units are represented at the Marathon deposit (Figure 1).

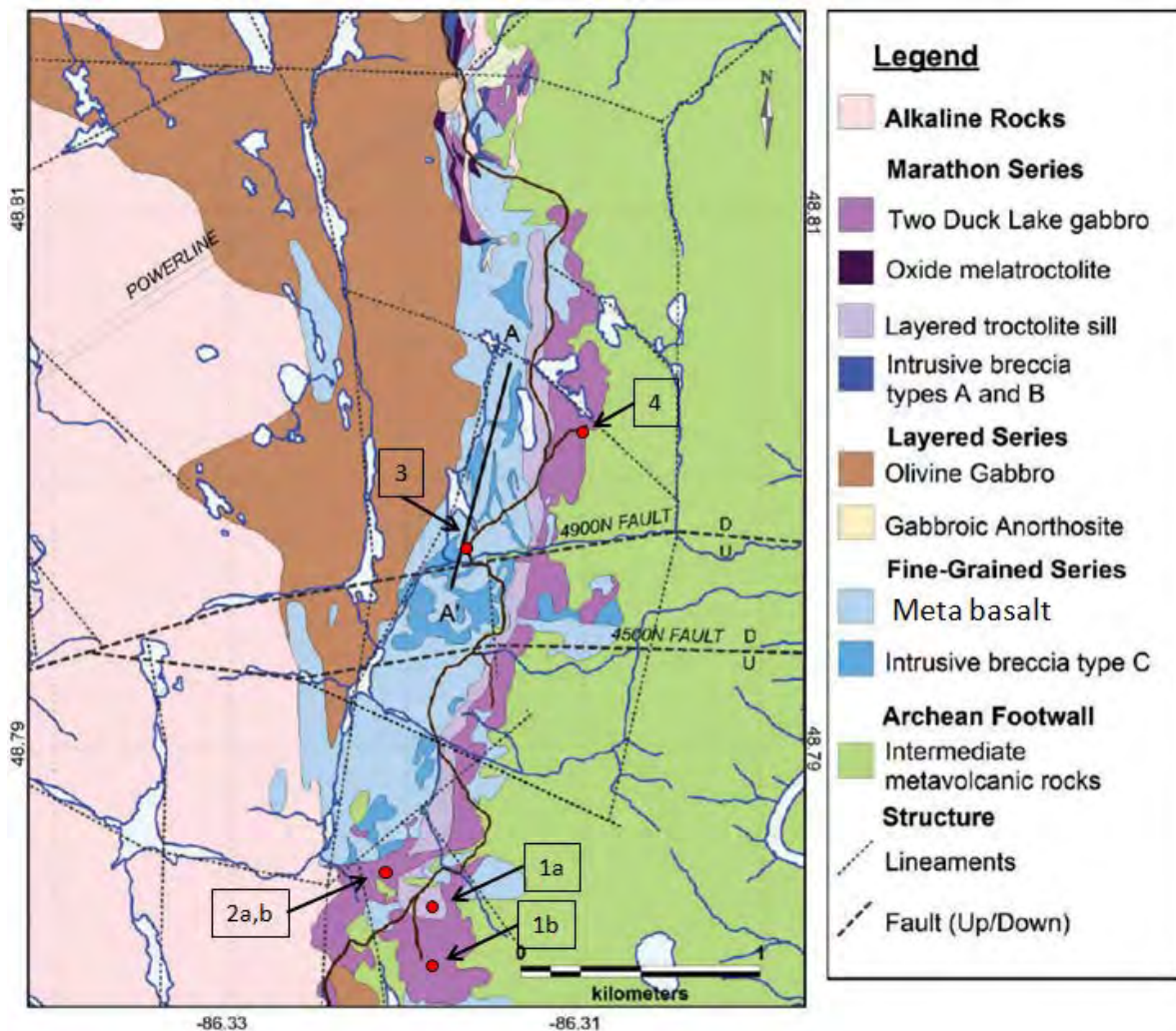


Figure 1: Geology of the Marathon deposit (after Good et al. 2015) highlighting location of field trip stops. Stops are marked with red dot and labelled as stop 1a, etc. Note two normal faults that correspond to strong surface lineaments (dashed lines)

The topography of the Coldwell is characterized by deep valleys and steep cliffs that form strong surface lineaments. Two lineaments at the Marathon deposit correspond to north dipping normal faults (north side down) with displacement of approximately 50 metres.

Two Duck Lake Intrusion:

The Two Duck Lake intrusion is irregular in shape and elongated north-south (Figure 2). The dip at the east contact is variable from nearly flat (at the south end) to vertical and locally over turned where the footwall overhangs the intrusion. The intrusion is composed of coarse-grained to pegmatitic olivine gabbro and troctolite. Modal layering is rare.

The TDL gabbro was interpreted to have formed by intrusion of a nearly homogeneous plagioclase crystal mush by Good and Crocket (1994). But recent work suggests the intrusion formed by accumulation of several pulses of magma in a conduit setting (Good 2010; Ruthart, 2012; Good et al., 2015; and Shahabi Far, 2017).

Multiple feeder channels were inferred by Good et al. (2015) to occur in the vicinity of several coincident features, including: deep V- or U-shaped channels in the footwall contact; topographic lineaments; very thick mineralized intervals; and irregular-shaped intrusions of olivine-magnetite-clinopyroxene-apatite.

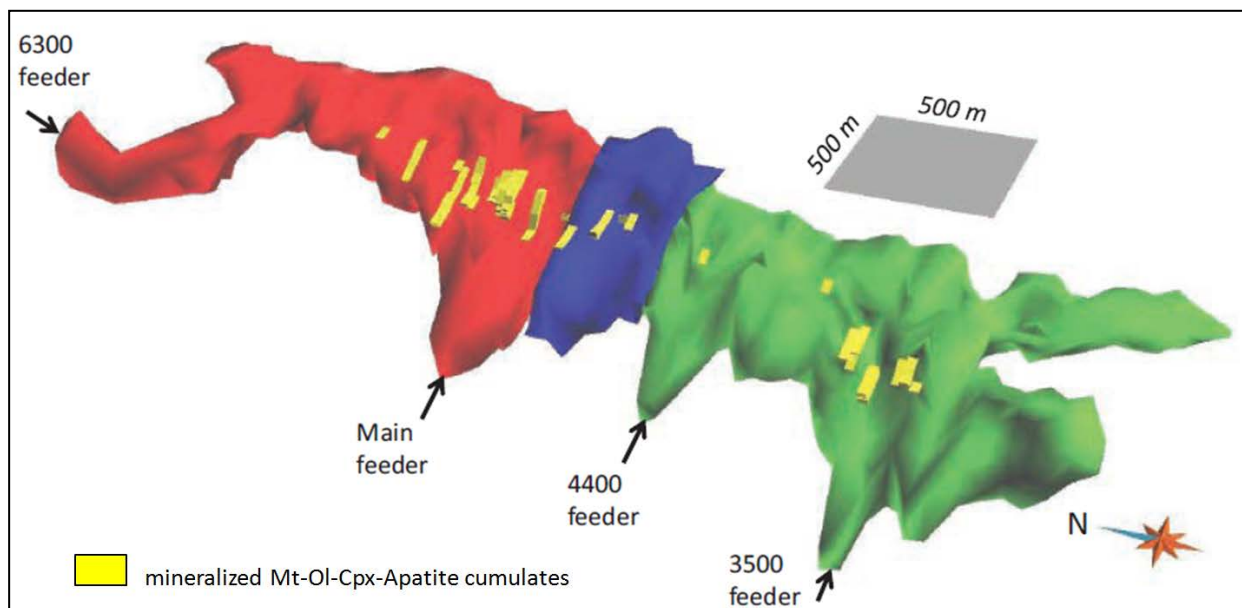


Figure 2: 3d isometric view of the Two Duck Lake intrusion (from Good et al. 2015). Three coloured portions indicate blocks that were offset by normal faults with north side down by up to 60 metres. Note that numerous intrusions of mineralized Mt-Ol-Cpx-Ap rock (yellow) occur in the vicinity of the 6300 feeder but are not shown.

Age relationships

Evidence suggests that all units in the Coldwell were emplaced within a short time interval between about 1108 Ma and 1105 Ma (Heaman and Machado 1992; Good et al. in preparation). Age relationships, based on cross-cutting contacts and U-Pb age dating for the various units are summarized in Figure 3.

The metabasalt is interpreted to correlate with Mamainse Point Volcanic Group 1 (Figure 4) which was emplaced at approximately 1108 Ma (Keays and Lightfoot, 2015).

The metabasalt was subsequently intruded by the following units, listed in order from oldest to youngest, layered troctolite sill of the Marathon Series, gabbroic anorthosite and olivine gabbro of the Layered Series, Two Duck Lake gabbro and various ultramafic units composed of magnetite +/- olivine +/- apatite +/- clinopyroxene of the Marathon Series, Malpa Lake intrusion, and syenite.

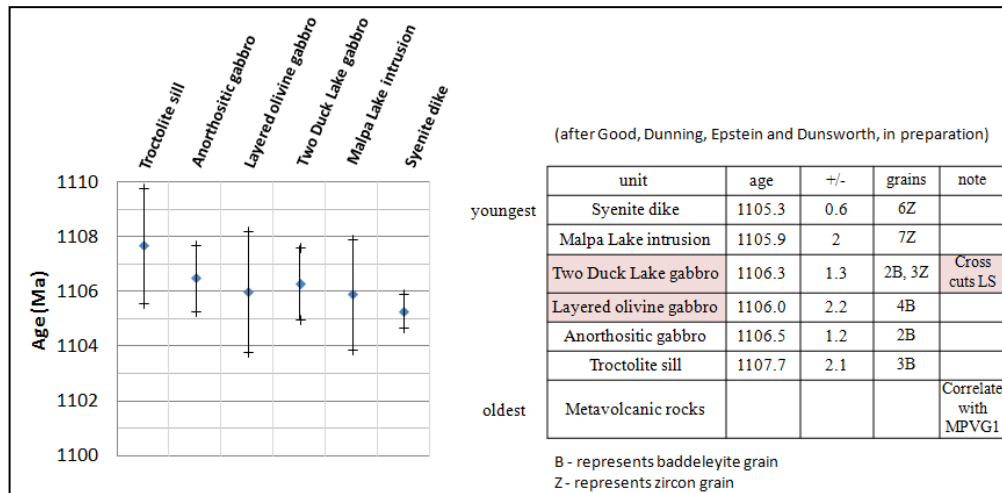


Figure 3: Relative timing of mafic metavolcanic and intrusive events (age dates after Good et al, in preparation) in the Eastern Gabbro Suite of the Coldwell Alkaline Complex.

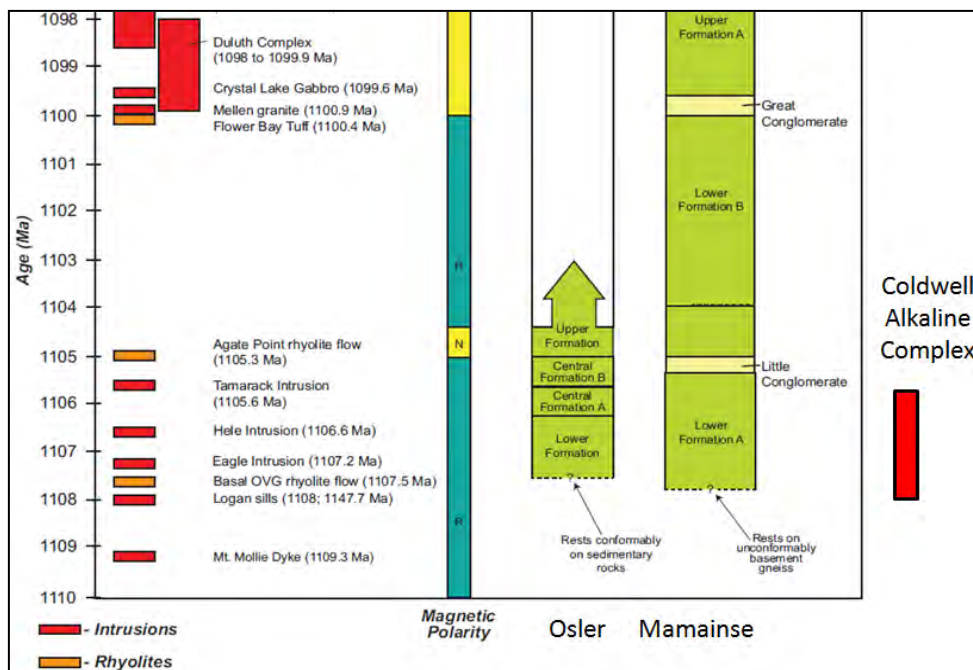


Figure 4: Correlation diagram showing range of ages for the Coldwell units compared to volcanic and intrusive units in the Midcontinent Rift (after Keays and Lightfoot, 2015).

Mineralization:

Disseminated sulfide mineralization is hosted by the Two Duck Lake gabbro and associated breccia (Figure 5) and occurs within several thick and continuous shallow-dipping lenses that parallel the footwall contact. The lenses are referred to as the Footwall, Main, and Hangingwall zones and the W Horizon. Sulfides in the Footwall, Main, and Hanging-wall zones consist predominantly of chalcopyrite and pyrrhotite with minor amounts of cubanite, bornite, pentlandite, cobaltite, and pyrite. Sulfides occur interstitial to primary silicates and also in association with hydrous silicates such as amphibole, chlorite, and minor serpentine (Watkinson and Ohnenstetter, 1992; Samson et al., 2008). Chalcopyrite occurs as separate grains or as rims on pyrrhotite grains. Some chalcopyrite is intergrown with highly calcic plagioclase (An70–An80) in replacement zones at the margins of plagioclase crystals (Good and Crocket, 1994; Shahabifar, 2016).

The W horizon is characterised by extreme PGE enrichment relative to Cu with several 2m thick drill hole intersections having 20 to 70 ppm Pd and Cu/Pd as low as 3. The best intersection contains 34 ppm Pd and 9.6 ppm Pt over 10 m. Mass balance considerations, assuming initial magma contained 10 ppb Pd, would require a magma column on the order of 34 km to generate the 34 ppm Pd in this interval.

The W Horizon is commonly difficult to identify in drill core because it typically contains only trace sulfides, but if sulfides are present, they consist of chalcopyrite and bornite with minor pyrrhotite and trace amounts of pentlandite, cobaltite, and pyrite (Ruthart, 2012).

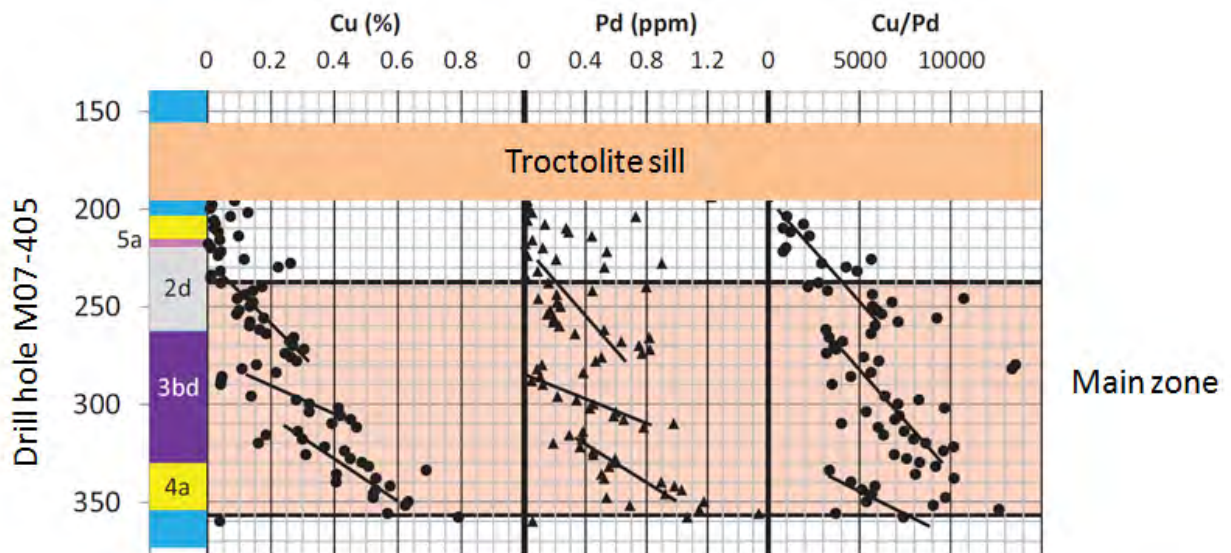


Figure 5: Stratigraphic section through the Main zone and overlying troctolite sill. Note the saw tooth pattern for Cu, Pd and Cu/Pd indicating individual pulses of sulphide-bearing crystal slurry. Unit 2d, breccia of metabasalt blocks and Two Duck Lake gabbro; unit 3bd, coarse grained ophitic and pegmatitic Two Duck Lake gabbro; unit 4a, breccia of footwall blocks and Two Duck Lake gabbro (from Good et al., 2015).

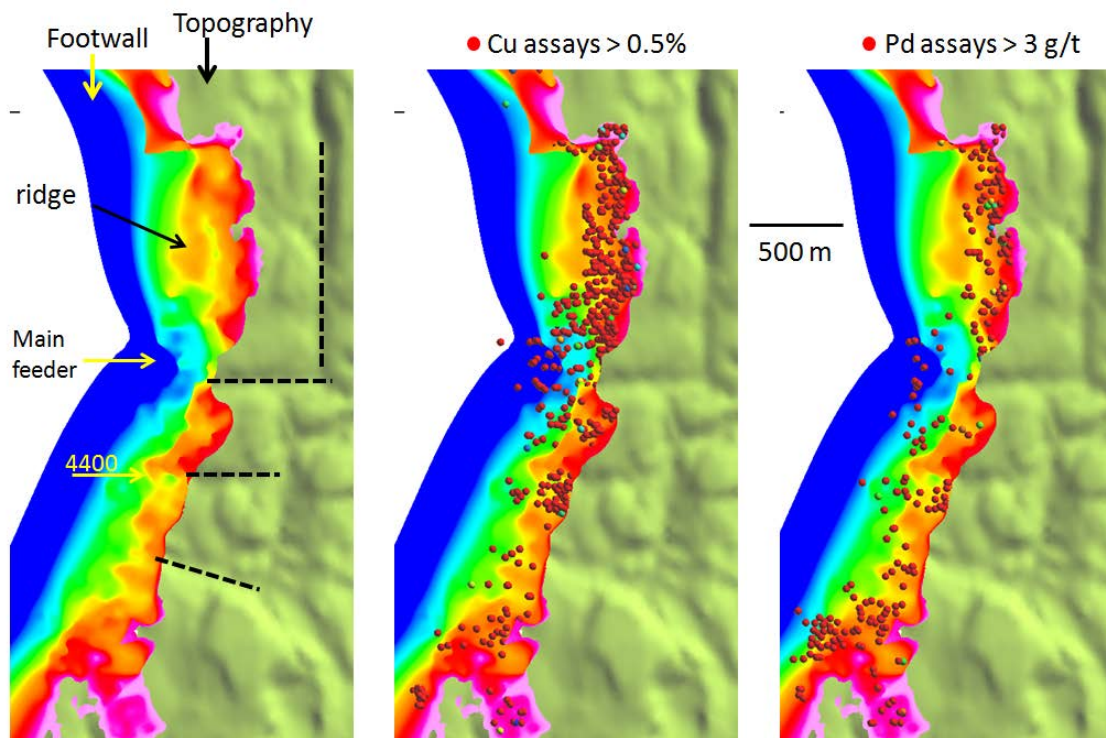


Figure 6: Three versions of top view for the Marathon deposit showing 3d topography (green surface) and contoured footwall surface models. Note the troughs and ridges (left hand image) correspond to surface lineaments. Note the higher grade assays for Cu (>0.5%) and Pd (>3 ppm) are aligned within zones that parallel troughs within the 3d footwall surface model.

Deposit Model:

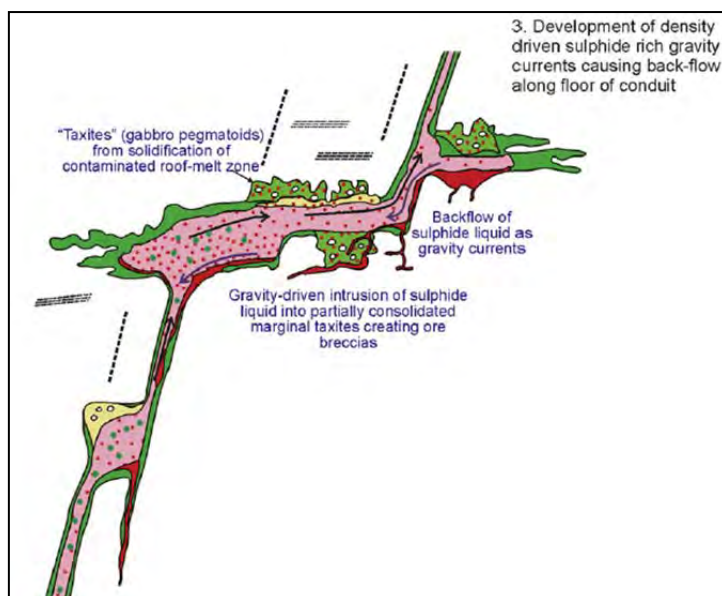


Figure 7: Step 3 of schematic illustration for magmatic plumbing system (after Barnes et al, 2016)

Evidence for magma conduit setting at Marathon include:

- association with volcanic rocks
- fault control
- brecciation and assimilation
- accumulation in trough setting
- flow through PGE upgrading
- tube shaped intrusion
- gravity driven back flow
- high heat flux

Marathon Cu-PGM Deposit - Field Trip Stops

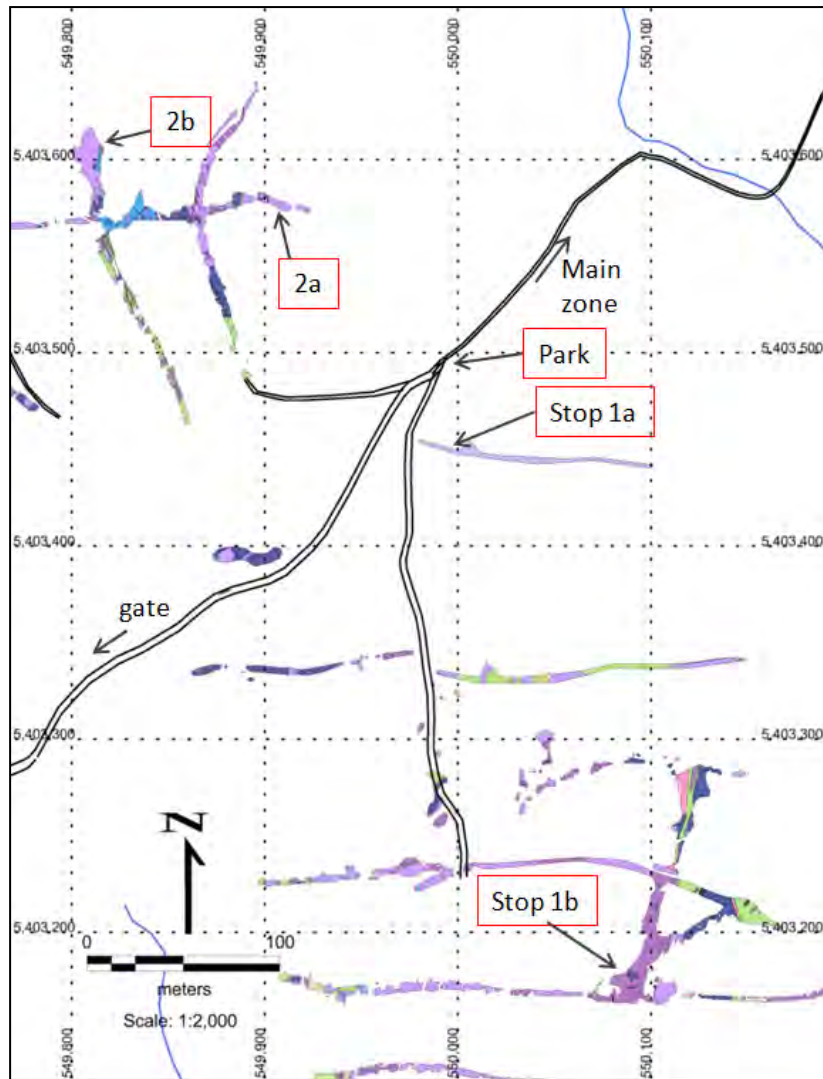


Figure 8: Map at south end of Marathon deposit showing location of stops 1 and 2.

Stop 1a: South end of Troctolite Sill (Figures 8 and 9b):

Trench exposure with coarse grained mottled augite troctolite shows large fresh oikocrysts of olivine (brown), clinopyroxene (black) and magnetite (black) and subhedral plagioclase (white).

The layered troctolite sill is an important marker horizon because it occurs just above the top of the Main Mineralized zone and is an indicator of the relative fault offset that occurred along E-W–trending normal faults at 5404500 and 5404900 North (Figure 1).

The sill dips moderately west at the north end, but flattens out in the south to sub-horizontal (Figure 10b). Note layering is approximately east west at stop 1a, consistent with rotation of sill from west dipping in the north to sub horizontal or north dipping in the south.

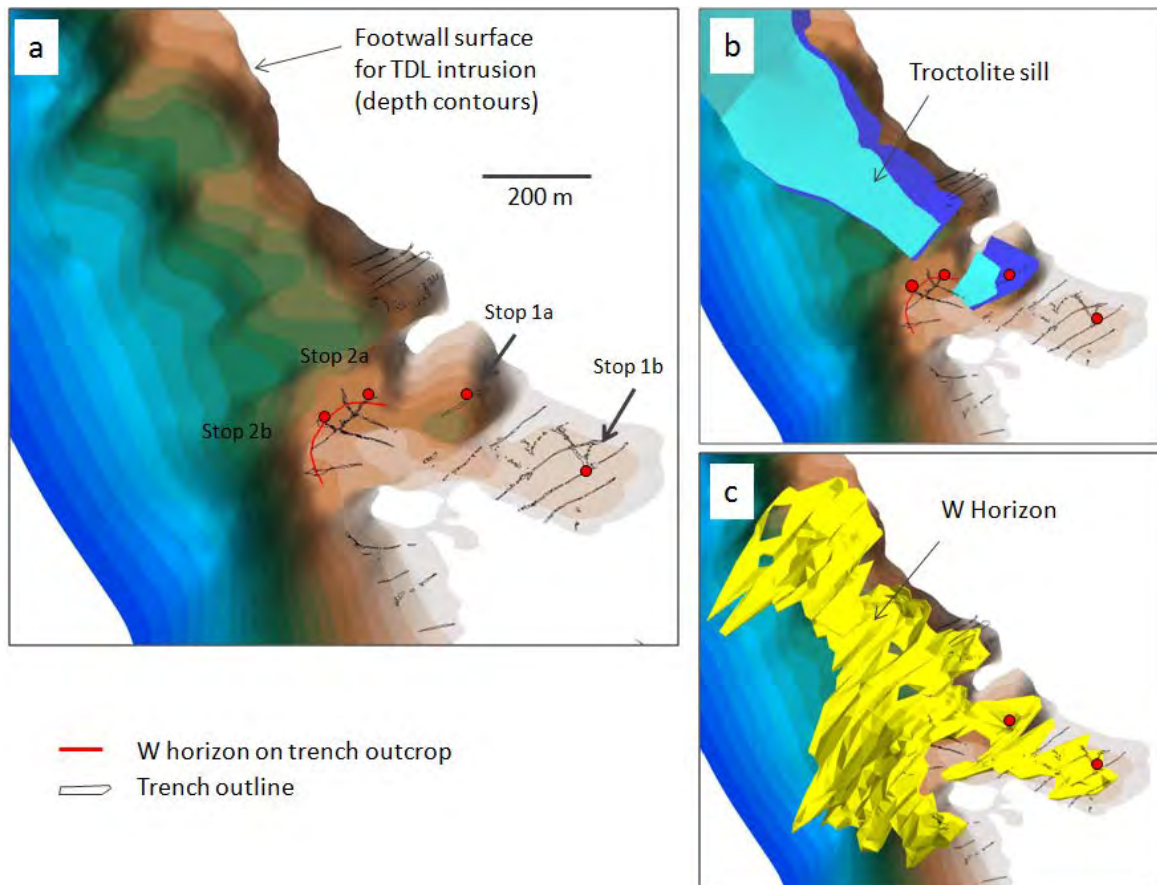


Figure 9: 3d image (iso view) of geology at south end of Marathon deposit showing location of stops 1 and 2 on trenched outcrops (black polygons): (a) shows orientation of footwall surface troughs approximately perpendicular to contact, and the red centre line of the W horizon at surface on the splat trench; (b) top (light blue) and bottom surfaces (dark blue) of the troctolite sill. Gap in the troctolite sill surfaces represents location where W Horizon and TDL gabbro cuts the through the sill; (c) surface model of W Horizon (yellow).

Stop 1b: South end of Marathon deposit (Figures 9, 11 and 12):

Trench outcrop exposure of Two Duck Lake gabbro shows shallow dipping W Horizon and Main zone type mineralization.

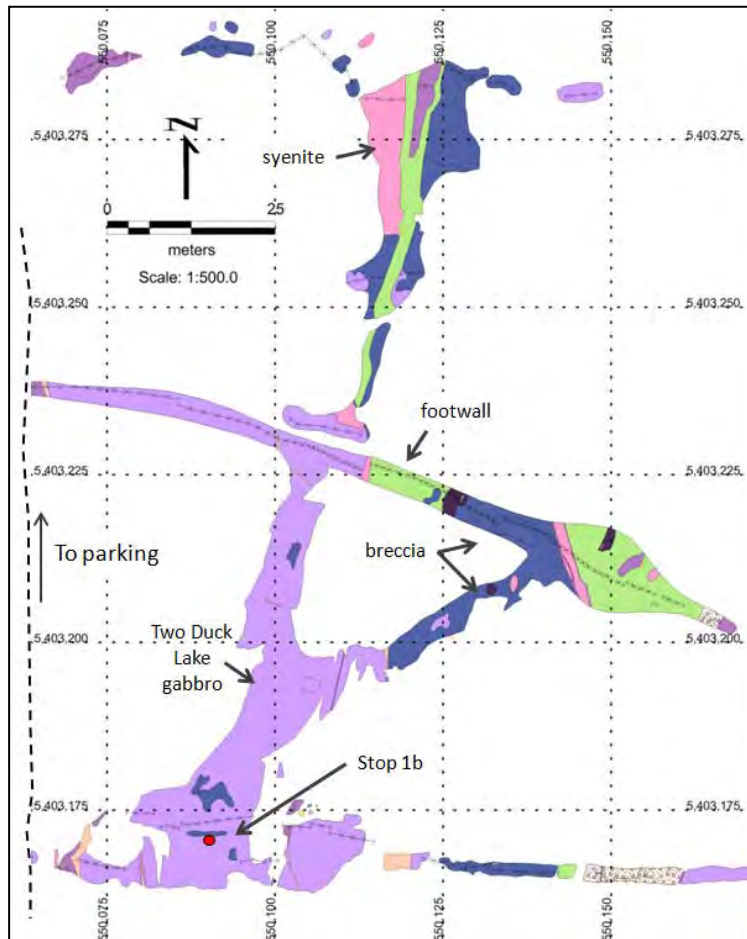


Figure 10: Plan map of trench at the south end of the Marathon deposit. Red circle marks location of historic trench (ca. mid 1960's) with high copper mineralization. The unit was not assayed for Pd until 2005.

The channel sample located just north of the red circle returned assays of 3.37 ppm Pd+Pt+Au, and 0.35% Cu over 18.6 m

East-west layering in TDL gabbro is visible just south of trench.

Note textural evidence for cross cutting intrusions.

Sample #	Interval (M)	Pd g/t	Pt g/t	Au g/t	Total PGM g/t	Cu %
44	1.53	5.20	1.44	0.70	7.34	0.86
45	1.45	1.92	0.68	0.17	2.77	0.13
46	1.63	2.15	0.71	0.22	3.08	0.23
47	1.42	3.26	0.69	0.35	4.29	0.50
48	1.15	3.90	1.03	0.56	5.49	0.36
49	1.29	4.48	1.13	0.33	5.95	0.47
50	1.03	2.51	0.56	0.17	3.23	0.50
51	1.1	2.60	0.81	0.26	3.67	0.47
52	1.53	1.11	0.39	0.09	1.58	0.26
53	1.48	2.91	0.74	0.21	3.86	0.50
54	1.51	0.10	0.06	0.02	0.19	0.04
55	1.75	0.25	0.11	0.04	0.40	0.04
56	1.68	1.38	0.42	0.12	1.92	0.24

East-west channel samples 44 to 56 (just north of red dot at stop 1b).

Interval average:
3.37 ppm Pd+Pt+Au, and
0.35% Cu over 18.6 m.

Stop 2a: Splat trench and the W Horizon

Figure 11:

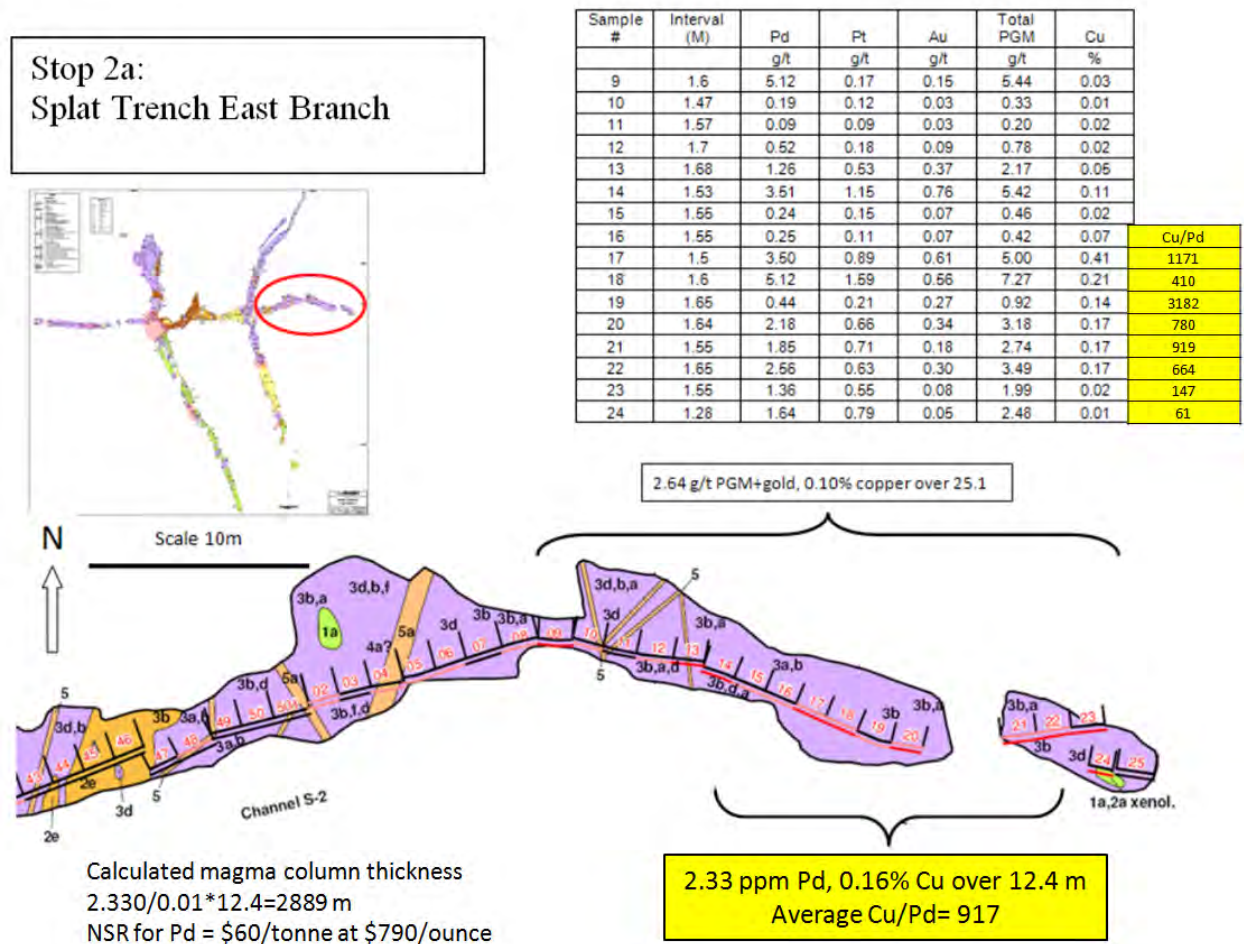


Figure 12: PGE rich samples from the W Horizon contain fresh clinopyroxene, olivine and plagioclase.

Top photo sample with 107 ppm Pd+Pt+Au and 203 ppm Cu.

Bottom photo sample with 70 ppm Pd+Pt+Au and 0.86 % Cu.

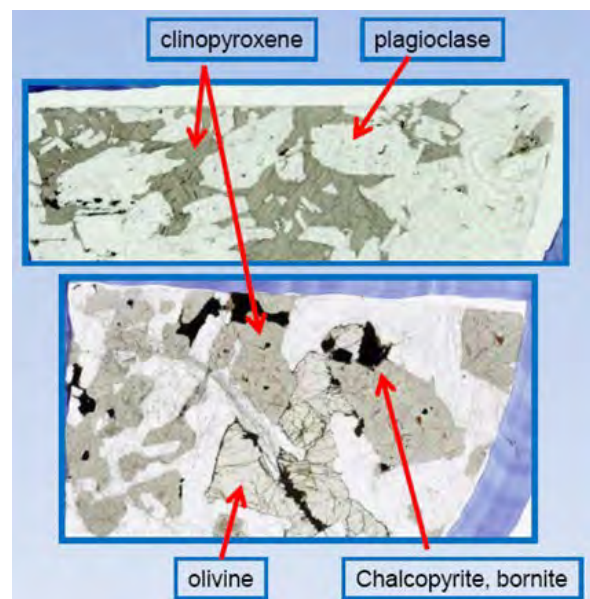


Figure 12:

Splat Trench - NW branch

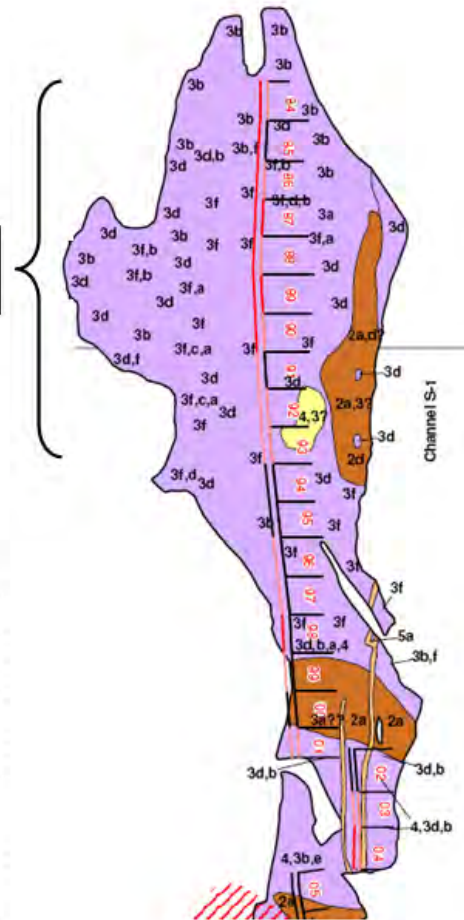


2.13 g/t PGM+gold,
0.36% copper over 17 m



Scale 10m

Sample #	Interval (M)	Pd (g/t)	Pt (g/t)	Au (g/t)	Total PGM+Au	Cu (%)
84	2	1.55	0.46	0.58	2.59	0.31
85	2.09	1.68	0.38	0.21	2.27	0.12
86	2	1.55	0.35	0.13	2.04	0.29
87	1.91	1.51	0.33	0.16	2.00	0.56
88	1.97	0.75	0.23	0.23	1.20	0.47
89	1.98	5.40	0.86	0.42	6.68	0.81
90	1.95	1.89	0.29	0.11	2.29	0.46
91	1.97	0.32	0.21	0.20	0.73	0.07
92	1.91	0.35	0.25	0.20	0.80	0.21
93	1.92	0.36	0.13	0.24	0.73	0.26



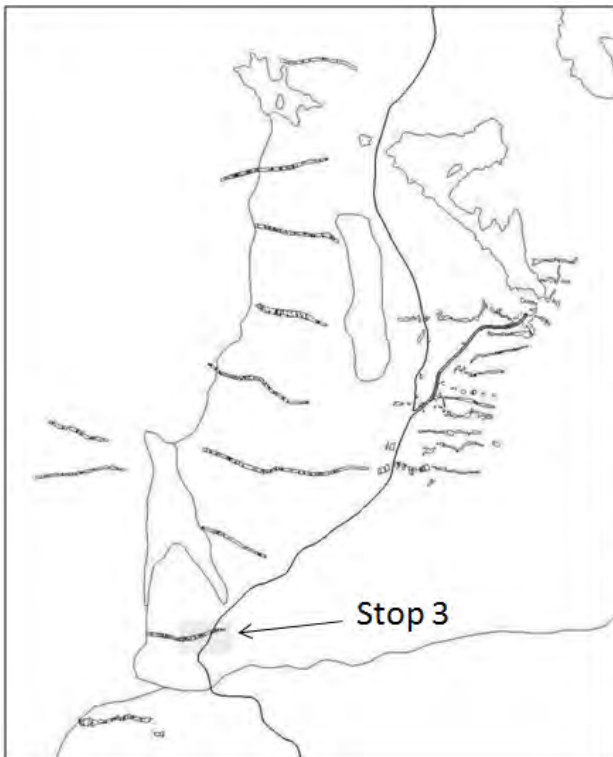
Stop 3: Ultramafic magnetite-olivine-clinopyroxene-apatite intrusions

Numerous pod shaped ultramafic apatite intrusions occur within the metavolcanic pile to the west of and stratigraphically above the Two Duck Lake intrusion. Similar intrusions occur within the TDL intrusion.

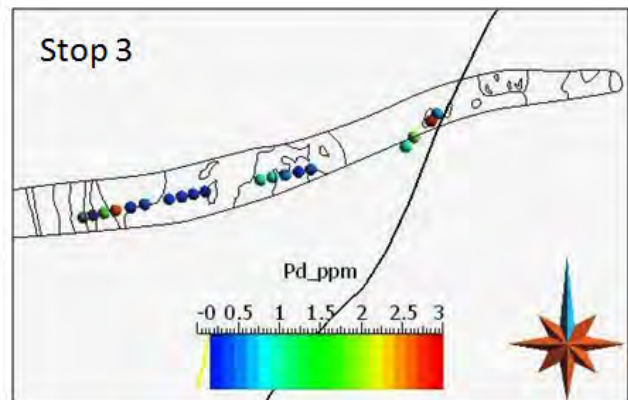
These units were previously referred to as reef type accumulations, but trenched exposures show they are discontinuous intrusive bodies that cut the metavolcanic and Layered Gabbro Series. The Four Dams and Willie Lake Cu-Pd occurrences are larger examples of this type of mineralization.

At the Marathon deposit, the ultramafic units occur as discontinuous pods that range in size from about 20 to 100 m along strike and less than about 10-20 m in thickness. The best intersections in surface outcrops include 9.3m at 1.6 g/t PGM+Au and 0.13% Cu and 4.94 m at 2.15 g/t PGM+Au and 0.29% Cu.

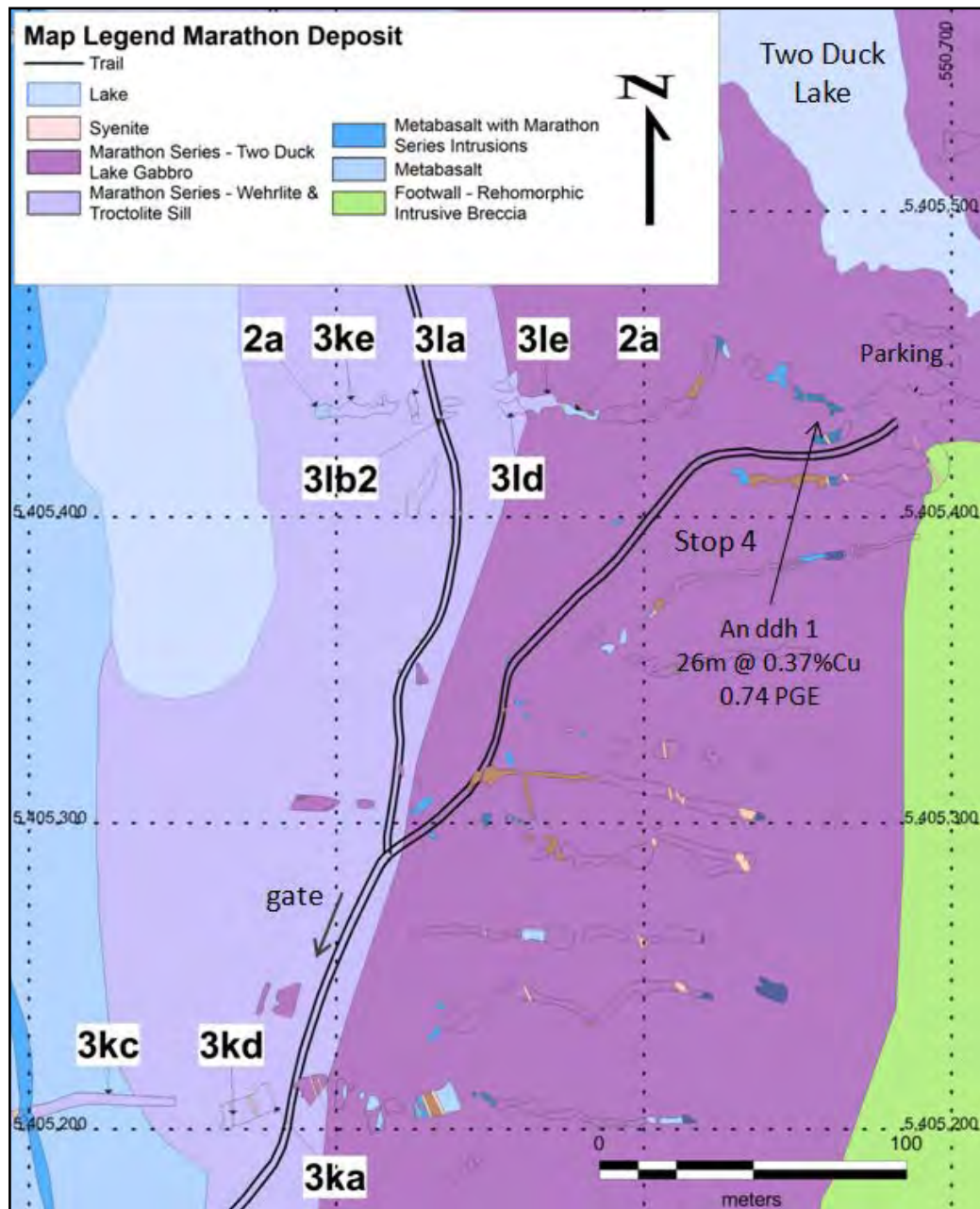
The ultramafic pods are concentrated in the vicinity above trough structures in the footwall and are potential indicators of underlying magma conduits. These intrusions are proposed to have formed by backflow of dense minerals in a conduit setting.



	Au ppb	Pt ppb	Pd ppb	Cu ppm	Ni ppm
701214	2.5	65	76	276	82
701215	42	109	518	1515	196
701216	353	614	3145	4447	465
701217	218	389	1849	3489	412
701218	125	212	1025	2219	374
701174	1097	431	2774	2211	417
701175	876	334	1495	1369	347
701176	295	269	213	389	236
701177	284	109	466	668	131



Stop 4: Main Zone at Two Duck Lake



MARATHON Cu-PGE DEPOSIT REFERENCES

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Field Trip 1

Archean and Proterozoic geology of the Marathon-Hemlo area

Day 2

Geology of the eastern Schreiber-Hemlo Greenstone Belt in the vicinity of Heron Bay and Hemlo

Mark Puumala and Mark Smyk
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and
Tom Muir
Ontario Geological Survey (retired)

INTRODUCTION

Day 2 of the Marathon-Hemlo field trip will begin in the Heron Bay area, just to the east of the contact between the Mesoproterozoic Coldwell alkaline complex and the eastern half of the Neoarchean Schreiber-Hemlo greenstone belt. The first three stops will highlight the lower-metamorphic grade (predominantly greenschist facies) supracrustal rocks and variable deformation that are characteristic of this portion of the belt. The trip will then proceed eastward to the world-renowned Hemlo gold camp (production to date of close to 22 Moz Au) where the bulk of the trip will be spent examining the strongly deformed and higher-metamorphic grade (amphibolite facies) supracrustal rocks that are exposed along Highway 17 adjacent to Barrick Gold Corporation's Williams Mine property. These roadside exposures provide an overview of the bedrock types, geological structures, mineralization and alteration that are present in the vicinity of the Hemlo gold deposit, and include outcrops of the up-dip projections of two gold-mineralized zones.

REGIONAL GEOLOGY

The Schreiber-Hemlo greenstone belt is located in the Neoarchean Wawa subprovince. The belt extends from Schreiber in the west, almost to White River in the east, and is bisected by the Mesoproterozoic Coldwell alkaline complex between the Little Pic River and Marathon. Consequently, the two halves of the greenstone belt are sometimes referred to separately as the Schreiber greenstone belt and the Hemlo greenstone belt (see Figure 1).

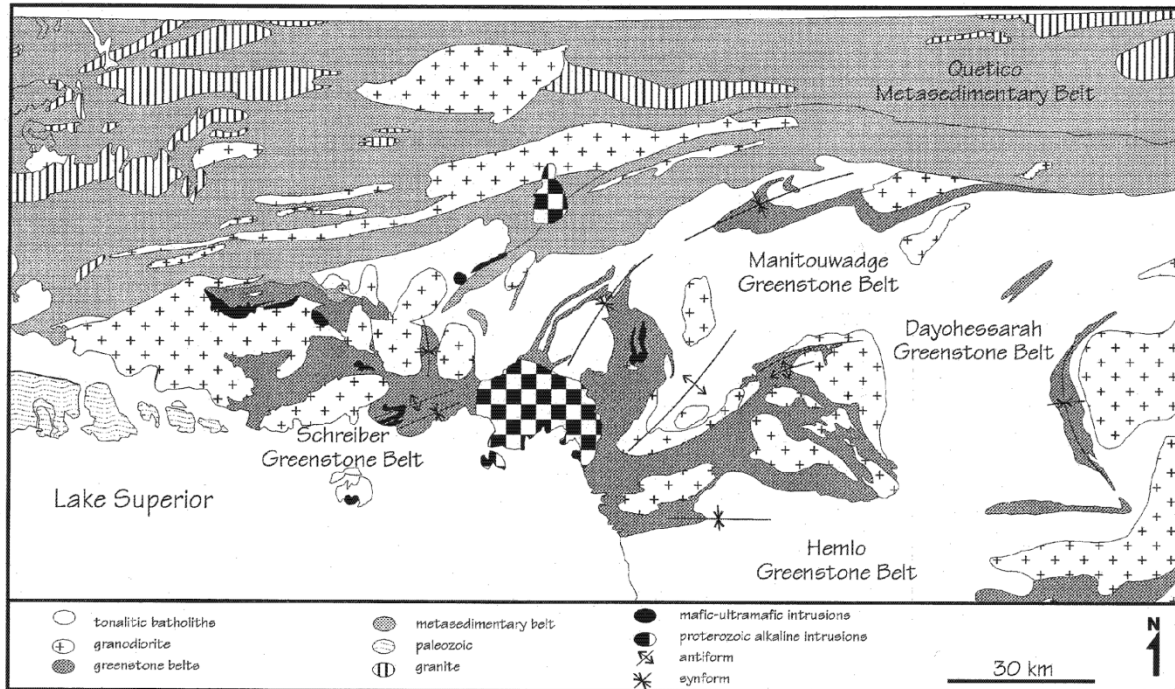


Figure 1. Regional geological setting of the Schreiber-Hemlo greenstone belt (from Jackson et al. 1998).

The western half of the Schreiber-Hemlo greenstone belt consists of a mixture of tholeiitic mafic metavolcanic rocks, calc-alkalic mafic to felsic volcanic rocks and sedimentary rocks whose stratigraphic and structural relationships have not previously been resolved (Williams et al. 1991). An Ontario Geological Survey (OGS) mapping project that is currently being undertaken in the western portion of the belt (Magnus and Walker 2015; Magnus and Arnold 2016) is designed to address this knowledge gap.

The eastern half of the Schreiber-Hemlo greenstone belt (i.e. east of the Coldwell complex) is much better understood (Muir 2000) and consists of three main units. The lowermost unit is dominated by tholeiitic metavolcanic rocks that are intercalated with lesser amounts of mafic to ultramafic intrusions and flows, and is older than 2697 Ma. This is overlain by calc-alkalic metavolcanic rocks (flows and pyroclastic rocks) and related intrusions that are between 2695 and 2688 Ma. This metavolcanic rock sequence is intercalated with metasedimentary rocks and is eventually superseded by an overlying metasedimentary rock dominated sequence (Muir et al. 1999).

The greenstone belt is bounded to the north and south by earlier (ca. 2720 Ma), gneissic granitoid rocks of the Black-Pic and Pukaskwa batholiths (see Figure 2), and has been intruded internally and along its margins by younger granitoid plutons ranging in age from 2697 to 2677 Ma. These include the ca. 2697 Ma Dotted Lake pluton, the ca. 2690 to 2684 Ma Cedar Lake pluton, Cedar Lake stock and Heron Bay pluton and the ca. 2679 to 2677 Ma Gowan Lake, Musher Lake and Bremner plutons (Beakhouse 2001).

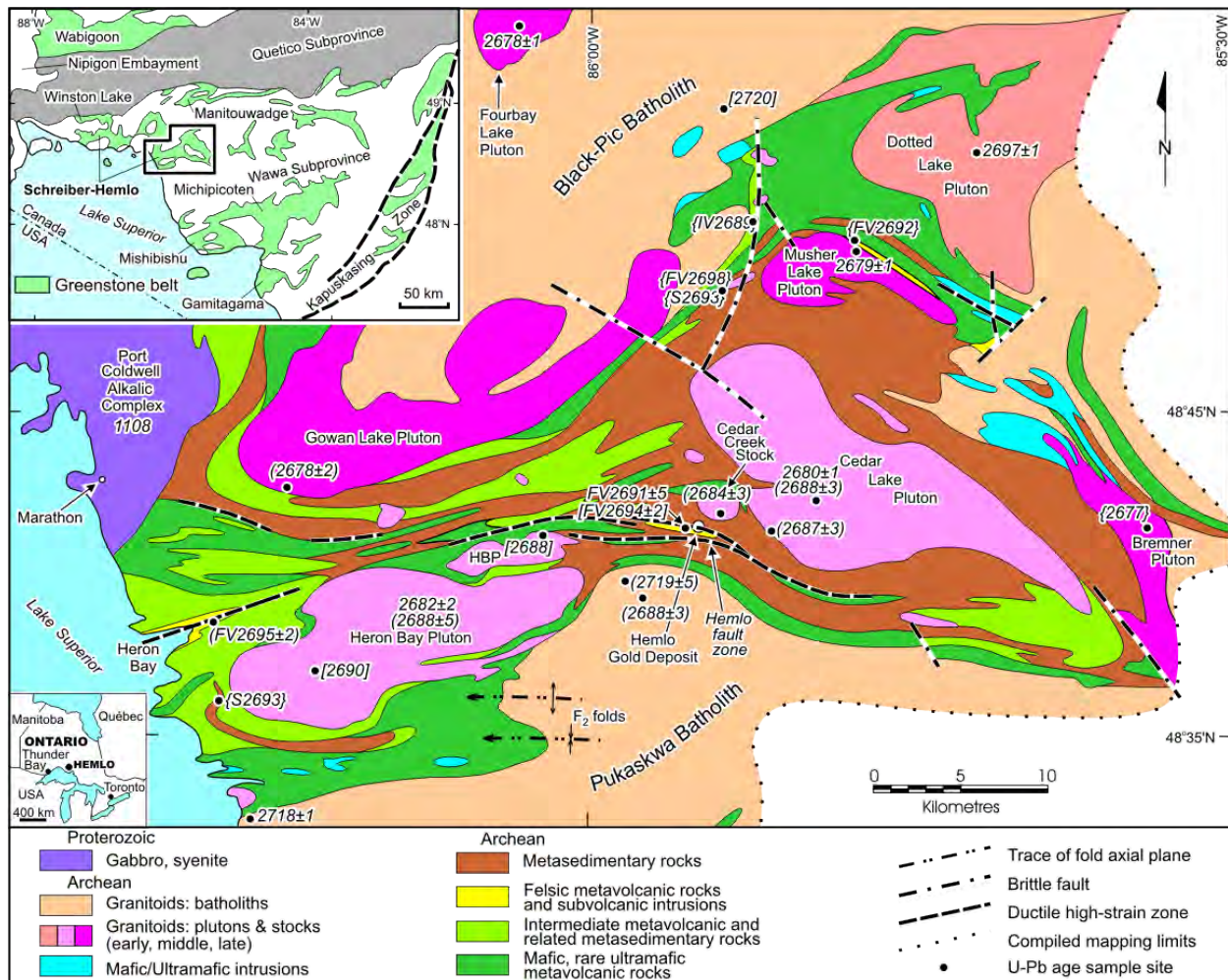


Figure 2. Generalized geology of the eastern half of the Schreiber-Hemlo greenstone belt (*from* Muir and Smyk 2006).

The eastern Schreiber-Hemlo greenstone belt has the morphology of an open synclinorium with complex internal structural patterns. It is interpreted to have undergone 2 main stages of structural deformation (Muir et al. 1999; Jackson 1998). Each deformation stage occurred over an extended period of time and may have consisted of multiple events (i.e., as documented by Muir (1997, 2003) in the immediate vicinity of the Hemlo gold deposit). The first phase of deformation ($D1_R$) affected rocks younger than 2693 Ma and resulted in the development of penetrative foliation that is generally subparallel to stratigraphy and intrusion boundaries. The second regional deformation event ($D2_R$) affected rocks as young as 2675 Ma and is considered to have been responsible for the development of map-scale folds with upright axial planar fabrics (Jackson 1998).

The eastern Schreiber-Hemlo greenstone belt is dominated by westward-plunging structural elements that suggest that the portion of the belt adjacent to the Coldwell complex exposes a shallower structural level of the belt than the eastern end (Muir et al. 1999; Jackson 1998). This

interpretation is consistent with the general eastward increase in metamorphic grade from greenschist facies in the Heron Bay area, to amphibolite facies in the Hemlo area (Thompson 2006).

Numerous high strain zones have been documented in the vicinity of the Hemlo gold deposit, with the most prominent of these being the Hemlo fault zone (Muir 1997). This high-strain zone may represent a portion of a regional-scale boundary fault that extends along the entire length of the greenstone belt between Heron Bay and Hemlo (Williams et al. 1991; Muir et al. 1999).

GEOLOGY OF THE HEMLO GOLD DEPOSIT

The Hemlo gold deposit is located 35 km east of Marathon in the south-central portion of the eastern Schreiber-Hemlo greenstone belt (see Figure 2). The following overview of the deposit geology is excerpted from Muir et al. (1995).

The Hemlo deposit is situated within supracrustal rocks in a southern bifurcated segment of the eastern part of the Schreiber-Hemlo greenstone belt. Page (1947a, 1947b; 1948; 1949) was the first geologist who recognized the Hemlo Fault as an important structural feature and identified the close co-planar relationship the fault had with the Lake Superior Shear Zone. He also recognized that all the gold discoveries at that time were hosted by the Lake Superior Shear Zone, and stated that the zone had been traced on surface for over 12 kilometres. Page also noted that the gold mineralization was associated with felsic porphyritic bodies and that the emplacement of these porphyries was related to a major structure which he termed the "Heron Bay-Hemlo Break".

The Hemlo deposit largely lies at or near the contact between felsic to intermediate quartz-feldspar-phyric rocks (pyroclastic and subvolcanic(?) varieties) and metasedimentary rocks. Here, the rocks generally strike at 290° to 295° and dip between 60° and 70° to the northeast.

The Hemlo deposit is presently interpreted by the authors as being hosted within 290°-striking, highly strained, transposed, and juxtaposed, lithotectonic supracrustal segments, which lie in a generally east-striking greenstone belt. The deposit cannot be demonstrated to be stratiform or stratabound. Sporadically distributed, anomalous gold mineralization has been noted, about several kilometres southeast and east-southeast of the Hemlo deposit on the Lac Minerals Limited, White River property, as being spatially associated with sericitic and pyritic rocks within what is interpreted as a brittle-ductile shear zone (Pan and Fleet 1988, 1989, 1990; Pan 1990).

Underground mapping and drilling have demonstrated the existence of parallel mineralized zones within both the metavolcanic and metasedimentary rocks, as well as mineralized zones which transect the metavolcanic-metasedimentary contact. The Hemlo deposit orebodies, collectively, extend for a strike length of about 3.7 km, a depth of 1.35

km, and an approximate down-plunge distance of 2.5 km. The main mineralized zone extends for a strike length of about 2.9 km, and a down-dip distance of 2.5 km (Harris 1989). The thickness of the main mineralized zone ranges from about 2 m in the David Bell Mine (Burk et al. 1986) to 50 m in the Williams Mine (Walford et al. 1986).

Several types of ore are delineated in each of the 3 mines, based largely on the predominant mineral(s) and/or textures present. Commonly, because of extensive metasomatism and deformation, the mineralized zones comprise rocks of equivocal protolith(s). Alteration, collectively, is in the form of widely various degrees of microclinization, sericitization, biotitization, silicification, carbonatization, albitization, pyritization, and tourmalinization. Significant amounts of barite of equivocal origin are locally present. Bright green vanadian muscovite (Harris 1989) is commonly present in the altered rocks, as is molybdenite. At least 2 ages of quartz veins can be found within the ore zones: some veins display considerable folding, attenuation, boudinage, and dismemberment, whereas others display minimal deformation. In some cases, outside the ore zone, there are numerous quartz veins which tend to display a lower degree of deformation.

Collectively, the ores are enriched in Au, Mo, Sb, Hg, As, Tl, V, and Ba. Gold is commonly disseminated along with molybdenite. Native gold grains are mercury rich and occur along quartz-feldspar and pyrite grain boundaries and fractures, as well as inclusions in, or rimmed with, several varieties of sulphide minerals including, rarely, pyrite and molybdenite (Harris 1989). Visible gold is not common overall, but does occur within quartz veins in feldspathized, molybdenite-bearing rocks, along molybdenite-green-mica-bearing fractures, in stibnite- and cinnabar-bearing quartz pods, and rarely in fractures in some of the plagioclase-porphyritic dikes. Molybdenite is the second most abundant sulphide, after pyrite, and occurs as fine- to very fine-grained, foliation-parallel blades, euhedral crystals, and platy masses mostly in association with silicate minerals, chiefly feldspar and quartz (Harris 1989).

Previous field guides for the Hemlo area have largely focussed on the Hemlo “Main Zone,” which continues to contribute to the production of the Williams Mine, and is known as the “B Zone” in current mine terminology. However, much of the current mine production is obtained from a distinct ore body known as the “C Zone.” The following description of the C Zone is excerpted from Langlais and Barber (2015).

The C Zone represents multiple sub-parallel lenses of irregular, generally narrow, gold mineralization. C Zone ore is stratigraphically different from the main zone and occurs in two broad geological domains, the porphyritic felsic metavolcanics and the intermediate to felsic volcanoclastic sediment unit. The open pit is located within the C Zone.

The general stratigraphy from south to north is Lower Metasedimentary rocks, Porphyritic Felsic Metavolcanics (Moose Lake Porphyry), Quartz Eye Muscovite Schist, Intermediate to Felsic Volcaniclastic Sediments (fragmental unit) and the Upper Metasedimentary rock sequence. Lower and upper denote the relative structural positions of the metasedimentary rock units as the younging directions are unclear. All of the major rock units are highly deformed with multiple events of deformation. Structural geology is complex. Rocks in the deposit area exhibit high strain. At the deposit scale, rocks in the area are tightly isoclinally folded. Most of the ore bodies occur on one or more limbs of these folds. Local drag folding can be seen in the ore. Occasional transverse faults offset ore and wall rock units up to a few meters, and there is some shearing along major contacts. Regional metamorphism is up to amphibolite grade. The deposit has also been cut by a number of north-south trending diabase and lamprophyre dikes which post-date mineralization.

Although much research has been carried out in the Hemlo area since gold production commenced in 1985, it is interesting to note that considerable debate continues around the application of a genetic model to the Hemlo gold deposit (especially for the Main Zone). Epithermal, intrusion-related (i.e., porphyry) and shear zone-hosted mesothermal models have all been suggested to explain the origins of this enigmatic deposit, with none having achieved widespread acceptance Muir (2002). Irrespective of the model used to explain its origins, it is clear that the Hemlo deposit's current morphology is largely related to deformation associated with the Lake Superior shear zone.

Hemlo Gold deposit Area Exploration and Development History

The chronology of events leading up to the discovery and development of the Hemlo mines is condensed from Muir et al. (1995). The subsequent development summary is largely based on Langlais and Barber (2015), with additional information obtained from the Thunder Bay South District Resident Geologist's files.

1869: Gold was discovered by Moses Pee-Kong-Gay near the present town of Heron Bay.

1920's: J. LeCours sank test pits on a mineralized shear zone near Hemlo station, 6 km southwest of the Hemlo deposit. Assays of up to 4.16 ounces gold per ton were reported. At about the same time, a group of claims was staked on a small quartz vein that returned low gold assay values just to the north of railway mile post 38 (measured from White River).

1930-31: J.E. Thomson mapped the area for the Ontario Department of Mines.

1937: Bowhill Mines shipped a 500 lb. (227 kg) test sample from the Heron Bay area that returned 0.30 ounce gold per ton and 1.53 ounces silver per ton.

1944-46: Prospector Peter Moses discovered a siliceous, mineralized shear zone approximately one-half mile (0.8 km) north of mile post 37 on the railway with samples returning assays up to 0.415 ounce gold per ton. Harry Ollmann and Dr. Jack K. Williams staked 11 claims in the discovery area (Ollmann-Williams property). Gold values were encountered in a large shear zone and stripping, trenching, and diamond drilling were completed. The claims were subsequently patented in 1947. After Ollmann died in December 1947, the claims were placed in Williams' name in trust by mutual consent.

1946-50: Adjoining claims were staked by consulting geologist Trevor Page. These claims, together with others staked by associates including Moses Fisher (J.E. Thomson's guide), subsequently became part of the Lake Superior Mining Corporation Limited property. Samples collected by Page assayed up to 0.13 ounce gold per ton. Lake Superior Mining Corporation carried out mapping, trenching, chip and channel sampling, and X-Ray diamond drilling on the Lake Superior shear zone. By 1950, a mineralized zone containing 31 543 tons to a depth of 300 feet (91 m) with a calculated width of 8.8 feet (2.7 m) and a cut grade of 0.22 ounce gold per ton was outlined.

1951: The Lake Superior property was optioned by Teck-Hughes Gold Mining Limited. 6 diamond drill holes totalling 2733 feet (833 m) were completed to add to the over 6000 feet of drilling completed to that time by Lake Superior Mining Corporation Limited. The size of zone No. 1 (formerly the 'A' zone) was increased to 76 653 tons at a grade of 0.27 ounce gold per ton.

1957: Teck Exploration Company Ltd. drilled 7 "packsack" drill holes totalling 289 feet (88 m).

1957: Bartley and Page wrote a geological report on the Hemlo area for the Canadian Pacific Railway.

1958: Cusco Mines Ltd. optioned the Lake Superior claims and carried out diamond drilling to test the main mineralized zone.

1960s: The Lake Superior property was staked intermittently by prospectors during the 1960's.

1973: The Lake Superior property was staked by J.E. Halonen for Ardel Explorations Ltd. Three holes totalling 789.9 feet (241 m) were drilled and the deposit tonnage was increased to 150 000 tons grading 0.21 ounce gold per ton above 60 feet (18 m) depth.

1976-77: Claims were staked by R.G. Newman to the west of the Williams property. The claims were investigated by Copper Lake Explorations Ltd. (soil and bedrock geochemical sampling).

1977-78: T.L. Muir of the Ontario Geological Survey mapped the Heron Bay and Hemlo areas, in 1977 and 1978, respectively (Muir 1982a, 1982b). Muir reported an occurrence (0.32 oz/ton Au), presently referred to as the “Highway Zone,” in altered felsic metavolcanic rocks several kilometres west of the main Hemlo deposit. This occurrence is probably in the vicinity of the mile post 38 discovery of the 1920's. Claim stakers and explorationists would later base much of their land acquisition during the staking rush on Muir's maps.

1979: Prospectors Donald McKinnon and John Larche staked the claims surrounding the 11 patented Williams claims.

1980: Corona Resources Ltd. (later International Corona Resources Limited) optioned the McKinnon and Larche claims and completed preliminary linecutting and geophysical surveys.

1981: Corona began a \$600 000 diamond drilling program in January. In March, R. Hughes and F. Lang optioned 156 claims which lie to the east and west of the Williams and Corona properties. These claims were put into the holding of their companies, Golden Sceptre Resources Ltd. and Goliath Gold Mines Ltd., who subsequently relinquished controlling interest to Noranda Exploration Company Ltd. In May, representatives of LAC Minerals Ltd. visited Corona's drill site and exchanged information pursuant to a possible joint-venture agreement. While negotiations with Williams' widow in Maryland for the Williams property were ongoing, diamond drilling was stepped back to the east of the outlined deposit. Drill hole 76 intersected a 10.5-foot (3.2 m) section grading 0.209 ounce gold per ton at a depth of 336.5 feet (102.5 m). This new, separate zone was the main Hemlo orebody. By August, 120 drill holes totalling 43 000 feet (13 106 m) had delineated 750 000 tons of rock grading 0.10 ounce gold per ton in the 'West' zone and had begun to indicate the much larger reserves of the 'East' or main zone. The Hemlo gold rush, ultimately involving 180 companies, ensued.

Both Corona and LAC had been actively negotiating for the Williams property. In July, Mrs. Williams accepted LAC's offer. Corona, citing a breach of a fiduciary agreement, sued LAC for ownership of the Williams claims. Teck Corporation subsequently entered into a joint venture agreement with Corona in December to develop what would become the David Bell Mine.

1982: LAC announced the discovery of the deposit on their property, which would become known as the Page-Williams Mine. Drilling by Goliath Gold Mines intersected the northward-dipping extension of the ore zone. The Goliath part of the deposit became, after a joint venture with Noranda Exploration Company Limited, the Golden Giant Mine.

1985: The David Bell, Golden Giant and Page-Williams mines commenced gold production.

1986-87: The Supreme Court of Ontario awarded the Page-Williams Mine to International Corona Resources Ltd. LAC appealed the decision to the Ontario Court of Appeal but continued to operate the mine under conditions imposed by the court. The Ontario Court of Appeal upheld the earlier decision in October, 1987. The Supreme Court of Canada later granted LAC the right to appeal the provincial court ruling.

1987: Golden Sceptre Resources Ltd., Goliath Gold Mines Ltd., and Noranda Minerals Inc. amalgamated their holdings and formed Hemlo Gold Mines Inc.

1989: The Supreme Court of Canada awarded the Page-Williams Mine, Canada's largest gold producer, to Corona who subsequently re-named it the Williams Mine (the mine operates under the name Williams Operating Corporation).

1991: Homestake Mining Corporation purchased the assets of International Corona Resources, including their interest in the Williams and David Bell Mines.

1992: Noranda Minerals Inc. transferred ownership of the Golden Giant Mine to Hemlo Gold.

1996: Battle Mountain Canada Ltd. acquired the Golden Giant Mine.

1998-99: Williams Operating Corporation acquired the surface and mineral rights to the Sceptre claims from Battle Mountain Canada to the 9450 elevation of the Williams Mine grid in 1998. In 1999, Williams also acquired the surface and mining rights on the Horizon claims from Battle Mountain Canada to the 10150 elevation of the Williams Mine grid. These acquisitions would permit pit expansion to the west, and allow evaluation of underground mining of the down dip extension of the C-Zone pit.

1999: Homestake's interest in the Williams and David Bell Mines was purchased by Barrick Gold Inc. in 1999. Milling operations ceased at the David Bell Mine when ore processing was transferred to the Williams mill.

2001: Ownership of Golden Giant changed to Newmont Canada Ltd.

2002: Williams acquired the surface and mineral rights from surface to the 10150 level on lease 273 and the remainder of lease 274 from Newmont Canada Ltd., providing an area for barren waste stockpiles from the expanded pit.

2006: Williams acquired the surface and mineral rights on lease 106623 from Newmont Canada Ltd. This acquisition allowed Williams to mine C Zone mineralization above the 9450 level as well as the down dip extension of the C Zone mineralization on the Interlake property.

2006: The Golden Giant Mine became the first Hemlo mining operation to close. The mine produced a total of 6,780,373 ounces of gold.

2008: Newmont and Williams entered into an agreement to allow Williams to extend its underground mining operations on the Williams property through a 60 m restricted area (Boundary Pillar).

2009: Barrick Gold acquires Teck's interests in the Williams and David Bell mines, giving Barrick sole ownership.

2010: Barrick acquires claims from Newmont that include the Golden Giant Mine workings.

2014: David Bell Mine closed.

2015: Barrick acquired additional claims from Newmont that are located immediately to the north and west of the Williams Mine. These acquisitions have opened up new opportunities for ore body expansion and exploration. Barrick's land holdings as of August 2016 are shown on Figure 3.

2017: The Williams Mine is currently owned by Barrick Gold Corporation and continues to produce from open pit and underground operations. 2016 production was 235 000 ounces Au from 3 408 000 tonnes milled (Barrick Gold Corporation, Q4 and Year-End Mine Statistics, February 15, 2017).

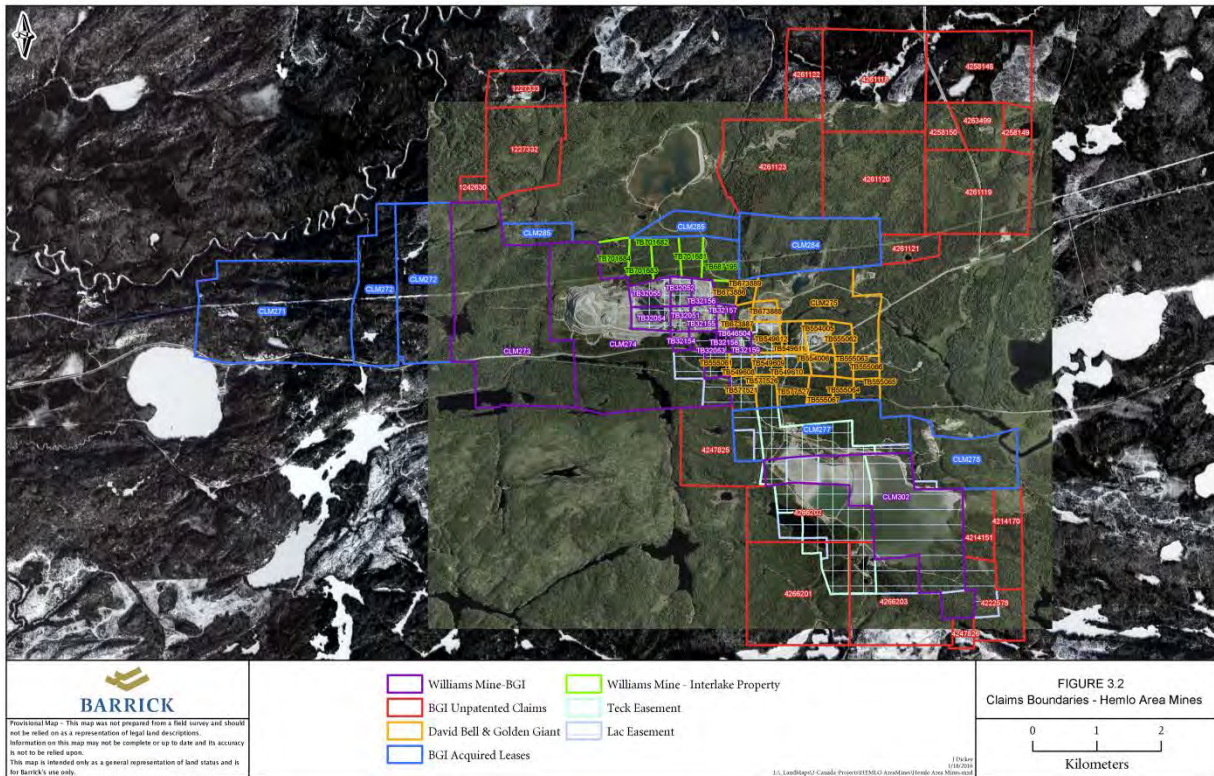


Figure 3. Barrick Gold Corporation's Hemlo Mine area land holdings (*from Williams Mine Closure Plan Amendment, August 2016*).

Hemlo Gold Deposit Production, Reserves and Resources

Cumulative gold production from the Hemlo gold camp to the end of 2016 has been 21 667 271 ounces Au (worth \$26 billion in \$US at today's gold price of \$1200/oz) from 108 983 846 tonnes of milled ore. The following graph (Figure 4) illustrates annual production from the three mines between 1985 and 2016. Production peaked at approximately 1.35 Moz in 1990.

Proven and Probable Reserve figures for the Williams Mine as of December 31, 2016, totalled 25 782 000 tonnes at a grade of 1.92 g/t Au for a total of 1 588 000 ounces Au. Measured and Indicated Resources currently stand at 58 897 000 tonnes at a grade of 0.908 g/ton Au for a total of 1 720 000 ounces Au (Barrick Gold Corporation, 2016 Mineral Reserves and Mineral Resources, February 15, 2017).

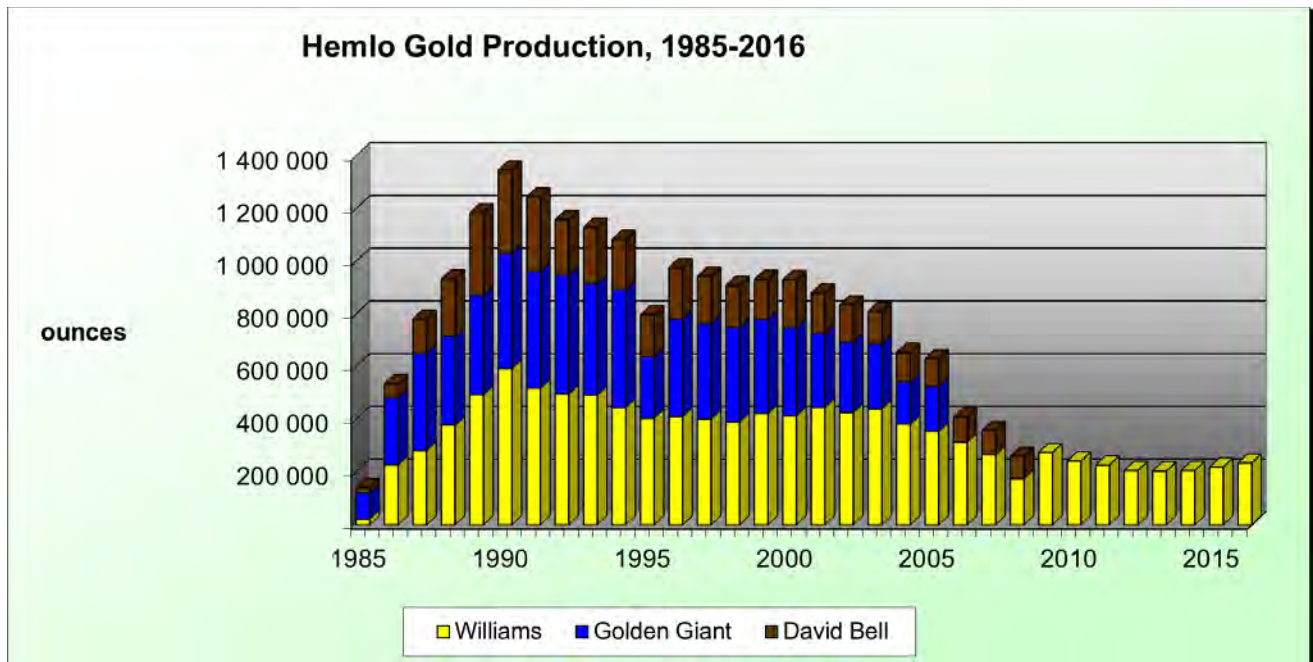


Figure 4. Graph illustrating annual gold production from the Hemlo camp to the end of 2016. Note that between 2009 and 2014 only the combined production statistics for the Williams and David Bell mines were published. These combined statistics were assigned to the Williams mine for the purposes of this graphic.

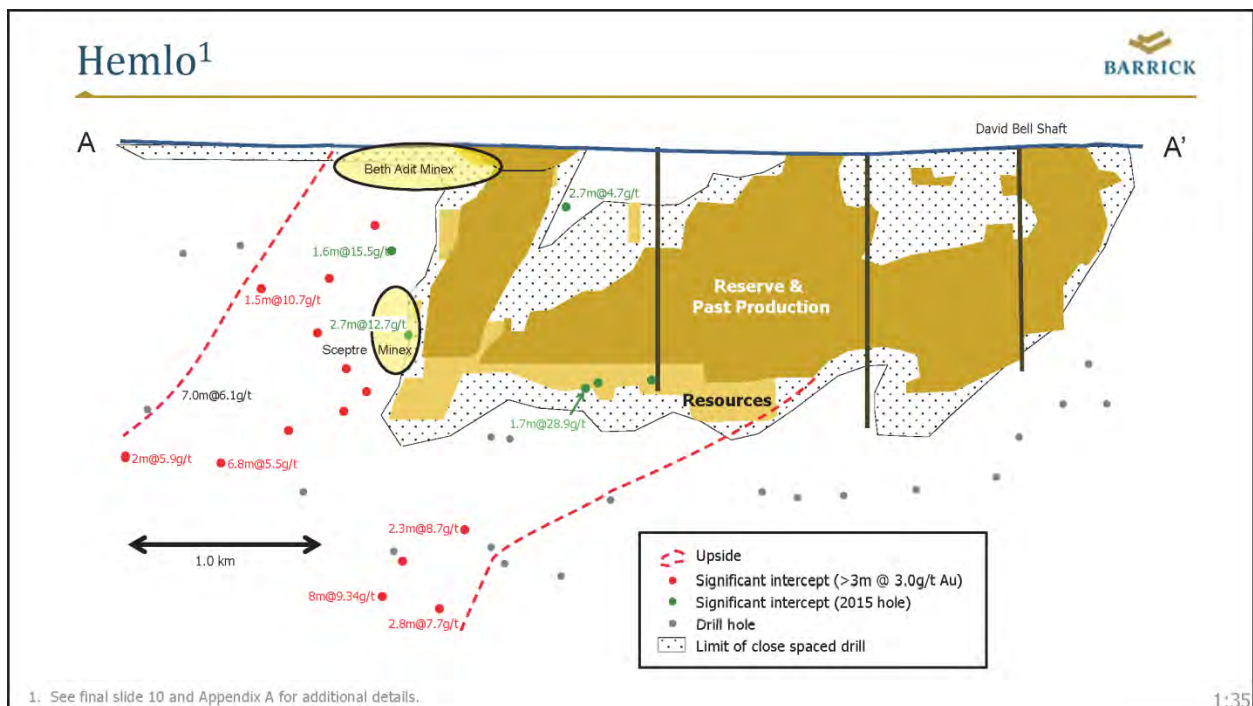


Figure 5. Long section view of the Hemlo deposit illustrating the location of the orebody (mined-out and reserves), mineral resources and areas of exploration potential along strike and down-plunge toward the west (from Barrick Gold Corporation, investor day presentation, February 22, 2016).

Field Trip Stops

Most of the field trip stop descriptions contained in this guide are mildly edited excerpts from an unpublished Ontario Geological Survey field trip guide that was prepared by Muir and Smyk (2006). Many of the coloured maps (unless otherwise noted) are also excerpted from that field guide. Field trip locations for the Heron Bay area are shown on Figure 6, while Hemlo area stops are shown on Figure 9.

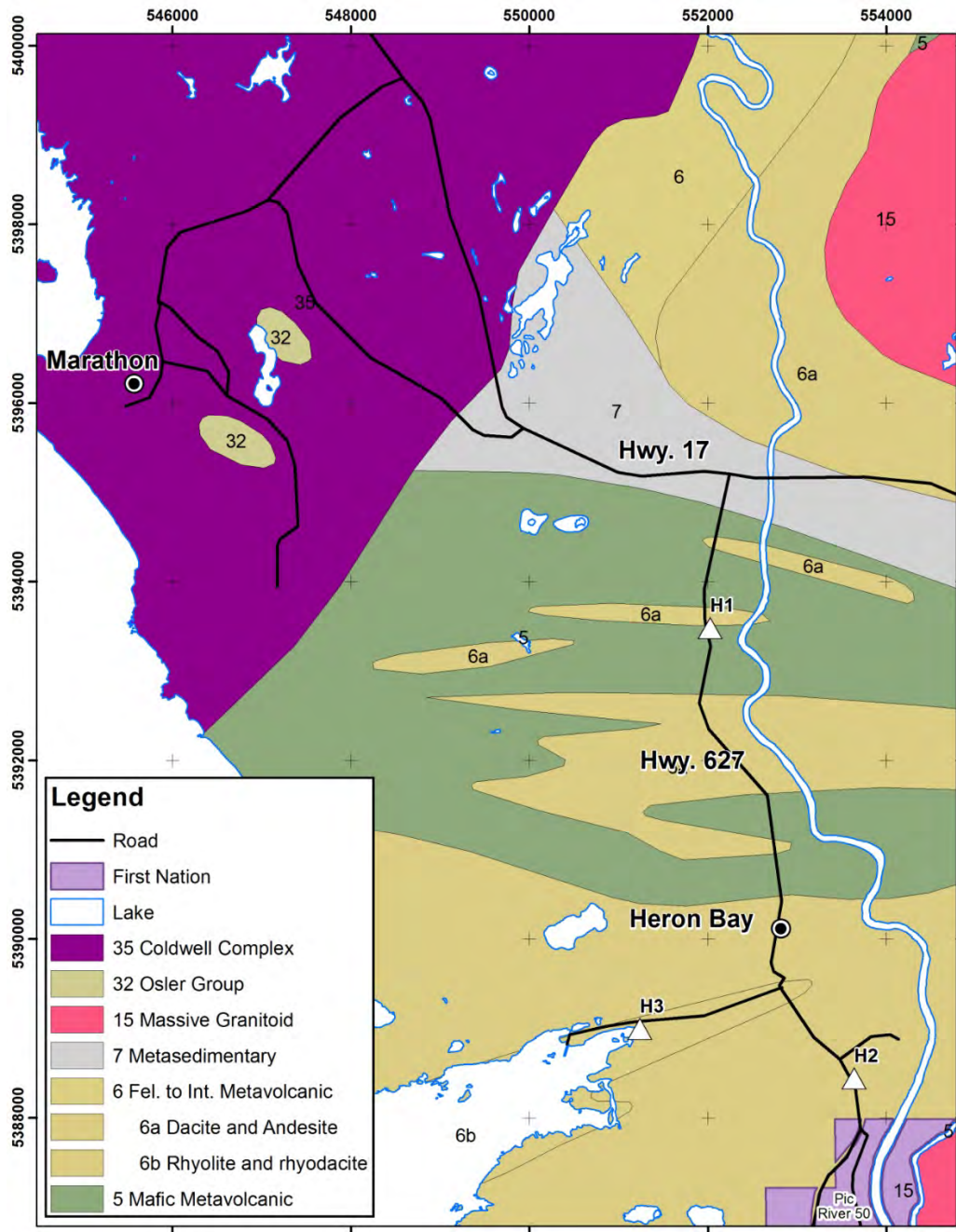


Figure 6. Geological map illustrating field trip stop locations (white triangles) in the Heron Bay area (geology from Ontario Geological Survey 2011)

Stop H1: Pillowed and massive mafic metavolcanic rocks

UTM Zone 16, 552024E, 5393482N

From Marathon, travel 6.8 km southeast on Highway 17 from the corner of Peninsula Road to the intersection with Highway 627. Then travel south for 1.7 km along highway 627 to an outcrop exposure on the west side of the highway, at a power line crossing.



Figure 7. Pillowed mafic metavolcanic rocks at stop H1.

At this stop are several small outcrops that display sections of a steeply dipping, thick mafic, tholeiitic flow. The more northerly ones consist of "massive," medium- to fine-grained basalt. One outcrop displays what appears to be a dike with a highly irregular orientation. Although large amphibole porphyroblasts give the impression that the rock is somewhat gabbroic, grain size decreases upwards through the flow to its aphanitic top, where small pillows and autoclastic breccias are developed. Although the pillows are somewhat flattened (see Figure 7), they have well-preserved selvages that indicate that the top direction is toward the south. The most southerly outcrop exposes a section that displays possible flow banding and what appears to be a flow top breccia or pillow breccia. Minor quartz + carbonate veins are present, particularly in the main outcrop.

Stop H2: Felsic pyroclastic rocks and “Heronite” dike

Continue 5.6 km south along Highway 627 (and through the community of Heron Bay) to an outcrop located on the west side of the highway. This stop is located approximately 300 m past the intersection with a dirt road that heads east from the highway.

UTM Zone 16, 553640E, 5388442N



Figure 8. Analcite tinguaita (heronite) dike cross-cutting felsic pyroclastic rocks at stop H2.

This stop provides an opportunity to view some of the felsic to intermediate pyroclastic rocks that occur in the immediate vicinity of Heron Bay. The metavolcanic rocks in this outcrop exposure have been cross-cut by an approximately 1 m wide Mesoproterozoic alkalic mafic dike that is likely to be related to the alkalic rocks of the Coldwell complex (see Figure 8).

The Heron Bay area pyroclastic rocks are quartz-plagioclase-phyric, and include pyroclastic breccia, tuff-breccia, lapilli-tuff, tuff and crystal tuff (Muir 1982). These rocks are mostly calc-alkalic dacite (intermediate), with some rhyolite breccias occurring near the lakeshore of Heron Bay). Overall, the fragments are subrounded to subangular, heterolithic in texture and composition, commonly more felsic than the matrix, and quartz-feldspar phyric. Some of the

fragments are more mafic than the matrix and include intermediate, feldspar $\pm$ quartz-phyric rocks, and mafic, aphyric rocks. Bluish quartz phenocrysts are locally common. The matrix consists of feldspar, quartz, sericite, $\pm$ chlorite. These rocks were dated at 2695 ± 2 Ma by Corfu and Muir (1989).

The alkalic mafic dike exposed in this outcrop has been classified as analcite tinguaitite (heronite). A number of dikes with this composition occur in the Heron Bay area and were first described by Coleman (1899; 1900). Interesting features include the presence of ocelli within the dike and fluorite mineralization in the metavolcanic rocks near the dike contact.

Stop H3: Deformed and altered felsic rocks along the shoreline of Heron Bay

Turn around and travel north on Highway 627 for 1.3 km back to the community of Heron Bay and turn west onto a gravel road just before the railway crossing. Then travel 1.7 km to a boat launching site on the shore of Heron Bay, Lake Superior.

UTM Zone 16, 551238E, 5388997N

The felsic metavolcanic rocks exposed along the shoreline at this stop provide an example of some of the deformation and alteration effects that are associated with the Heron Bay deformation zone. The following description of the Heron Bay deformation zone is excerpted from MacTavish and Osmani (1996).

The 'Heron Bay Deformation Zone' (HDZ) is a strong, locally intense, northeasterly trending, roughly 900 m wide zone, consisting of numerous, discrete, anastomosing shears, faults, and lineaments. Shearing has deformed the host rocks into a highly variable assemblage of quartz-sericite-carbonate $\pm$ chlorite schist. Hematization, silicification and iron carbonatization are locally common. Intermediate to felsic hypabyssal intrusive rocks (eg. porphyries) are often emplaced along the margins of the zone in a region of more brittle-ductile deformation. These dykes/sills are shallow dipping, and with the exception of the margins of a few bodies, are undeformed. This suggests that they were emplaced during the later stages of the regional deformation events.

A number of gold occurrences are found in association with the Heron Bay deformation zone. (e.g. Heron Bay Mine (Peekongay) and Bowhill Mines occurrences; Patterson 1984) One of those occurrences, known as the Screamer Zone is located approximately 200 m north-northwest of this stop. Gold mineralization at the Screamer Zone occurs in two variably brecciated quartz-carbonate veins. These veins occur within sheared and iron-carbonatized mafic metavolcanic host rocks at the contact with a feldspar porphyry dike. The dike itself intruded along the contact between mafic and felsic to intermediate metavolcanic rocks (MacTavish and Osmani 1996).

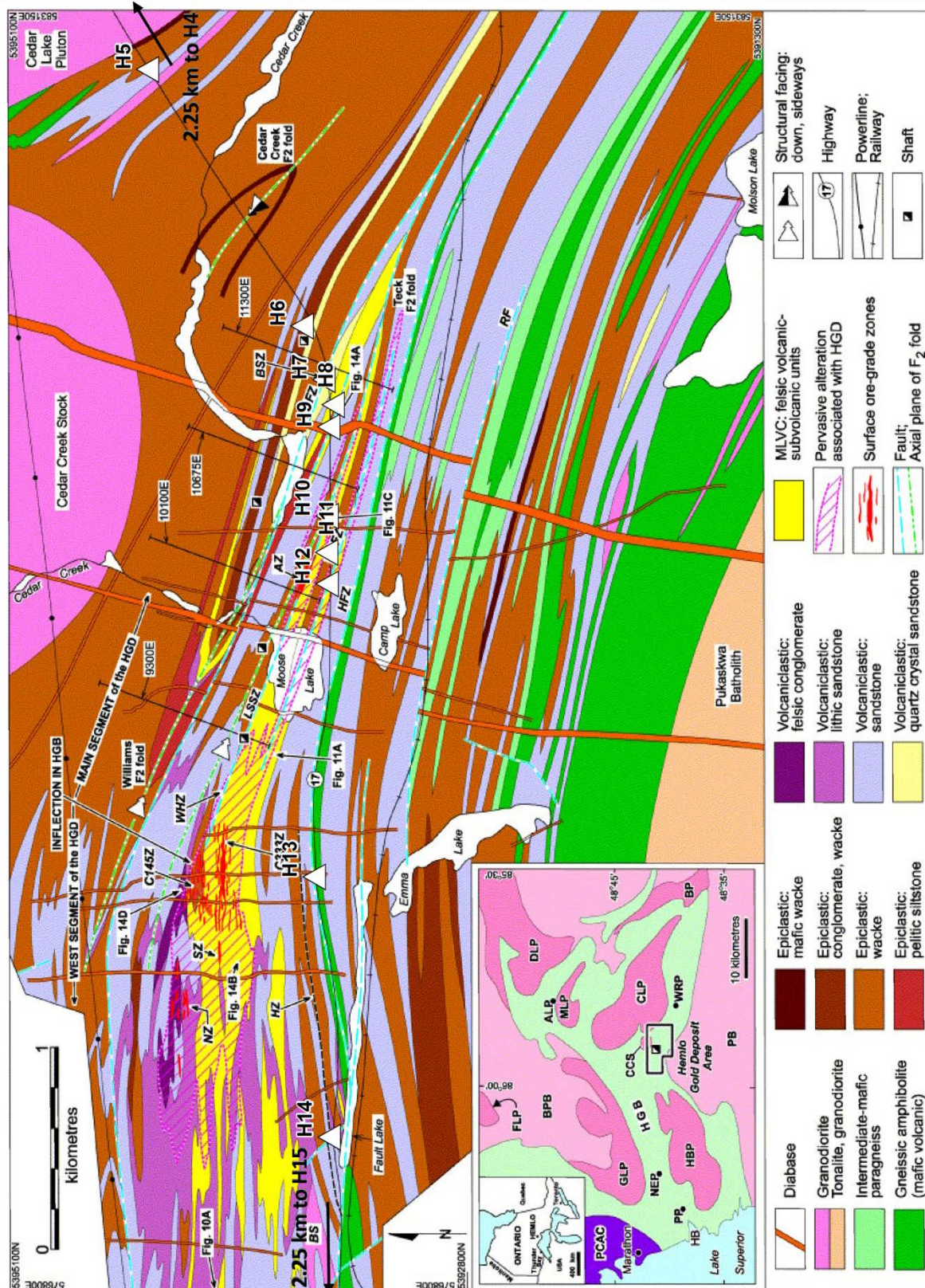


Figure 9. Geological map of the Hemlo gold deposit area showing field trip stop locations (white triangles). Stops H4 and H15 are approximately 2.25 km east and west of the map limits respectively. *Geology from Muir (2002).*

Stop H4: Cedar Lake pluton

Return to Highway 17 and then proceed east. After travelling for approximately 33 km along Highway 17, and just after passing the Williams Mine, you will see the intersection with Highway 614. Continue eastward along Highway 17 past this intersection for another 1.4 km to a location with road cuts on both sides of the highway.

UTM Zone 16, 585227E, 5395748N

The Cedar Lake pluton consists of massive to very weakly foliated medium-grained, microcline-megacrystic, hornblende-biotite granodiorite that contains mafic/ultramafic clots or xenoliths and diorite to monzodiorite inclusions and dikes (varies with location in pluton). The diorite and monzodiorite inclusions and dikes display a variety of relationships with the granodiorite including back-veining by the granodiorite. These relationships suggest a co-magmatic relationship (Beakhouse, 2001). The mafic-ultramafic and dioritic inclusions are commonly preferentially oriented, generally parallel or subparallel to the contact of the pluton (in a broad sense) and to the foliation in the granodiorite where it is discernible. The foliation locally deflects around the inclusions. The foliation may be related to original flow foliation; synchronous D2 deformation (i.e., syntectonic pluton); and/or a later D3 deformation. The pluton here has been intruded by a few intermediate-composition Archean dikes and a Proterozoic biotite lamprophyre dike. The Archean dikes are locally sheared with an apparent dextral sense. The granodiorite at Stop H4 was initially dated at 2688 ± 3 Ma (Corfu and Muir 1989). A more recent sample provided a somewhat younger age of 2680 ± 1 Ma (Beakhouse and Davis, 2005).

Stop H5: Cedar Lake Pluton western contact with metasedimentary rocks

Turn around vehicle (there is a gravel road on the south side of the highway that can be used as a turn-around site just east of stop H4) and travel west along Highway 17 for 2.6 to the Yellow Brick Road sign.

UTM Zone 16, 582891E, 5394626N

At this stop, the Cedar Lake pluton consists of a relatively magnetite-rich, fine- to medium-grained, biotite-hornblende granodiorite, which is massive except for a very weak foliation within several metres of the contact. A plagioclase-phyrlic, hornblende-biotite sheet exists to the southwest (Figure 10).

The granodiorite at Stop H5 is intruded by many aplite and pegmatite dikes, some of which are composite: generally, pegmatite tends to crosscut aplite. The dikes typically terminate against the country rock schist, although some display ductile deformation within the contact granodiorite. The strike of the dikes within the granodiorite tends to “rotate” clockwise toward the contact, consistent with a dextral component of displacement at the contact. To the north-northwest of here, dextral shear along the contact is related to D2, even though D2 is an overall sinistral event.

The country rocks within about 9 m of the contact consist of D3-crenulated mafic schist containing several locally dismembered, aplitic and granodioritic dikelets and stringers, which display refolded folds. This refolded nature of country rock layering and/or dikes in the immediate contact aureole is common with at least the Cedar Lake pluton and the Pukaskwa Batholith.

For approximately the next 150 m westward, the country rocks consist mostly of various sets of metawacke and metasiltstone “packages”, with locally isoclinally folded layering. There are numerous mafic to felsic dikes and several swarms of dikes, generally granitic. Boudinage in dikes is locally present. The granitoid dikes consist mostly of weakly foliated granodiorite, which is similar in grain size and composition to the marginal phase of the Cedar Lake Pluton, as well as some plagioclase-porphyritic dikes. Within this sequence of turbiditic sedimentary rocks is a thick, composite sheet (Figure 10) of foliated, medium-grained, plagioclase-porphyritic, hornblende-biotite granodiorite (2687 ± 3 Ma; Corfu and Muir 1989), and foliated, finer-grained, biotite-hornblende granodiorite, both of which were intruded by aplitic dikes and subsequently by quartz veins. At the structurally lower (western) contact of this sheet, apophyses of the porphyritic granodiorite are folded about a less steeply dipping, S_2 -like foliation, with attendant attenuation and boudinage which has taken place in 2 dimensions. The folds have been interpreted by others as F2 folds (i.e., dikes predated F2), but it is more likely that they represent a local contact strain phenomenon related to a progressive and complex D2 event.

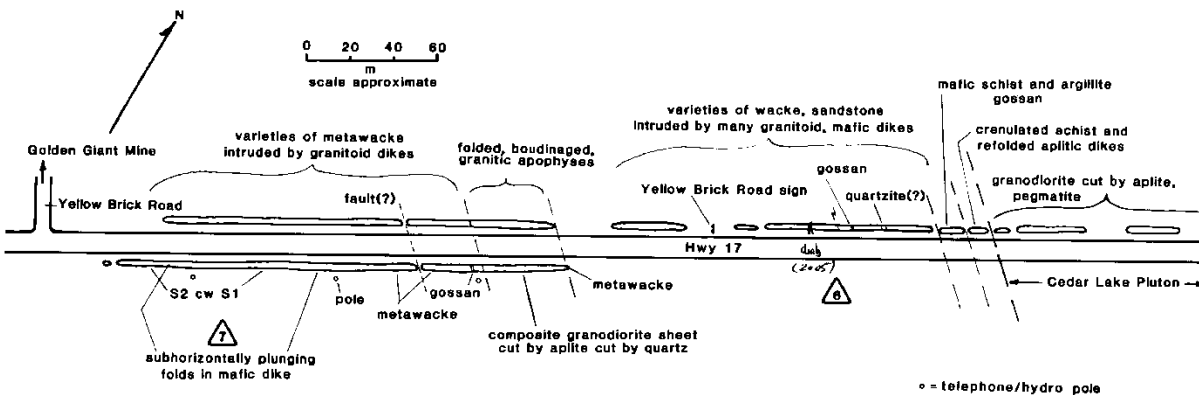


Figure 10. Sketch map of outcrops at the Cedar Lake pluton western contact (*from* Muir et al. 1995). Note that the number labels shown on this figure denote the 1995 field trip stop numbers.

Stop H6: Deformed metaconglomerate

Travel 1.5 km to an outcrop located on the north side of the highway approximately 150 m past the access road to the former David Bell Mine site.

UTM Zone 16, 581605E, 5393852N

The main outcrop in this set of 3 small outcrops consists mostly of 2 metaconglomerate “layers” (one with predominantly cobbles-boulders; the other with predominantly pebbles), separated by a medium- to coarse-grained wacke locally with entrained pebble- to cobble-sized clasts. This represents a weak, remnant bedding (S₀/S₁). Flattened clasts tend to be aligned parallel to S₂, and oriented lengthwise slightly clockwise with respect to the crude layering, consistent with this unit being on the northeast limb of the northwest-closing Williams fold (see Figure 9). This fold is part of the large-scale S-shaped Williams-Teck fold pair, formed during the D₂ sinistral shear event and best delineated by the Moose Lake volcanic complex (yellow unit; Figure 9). Note that some of the clasts have been deformed by F₃ folds (see Figure 11).



Figure 11. Deformed conglomerate (F₃ fold) at Stop H6.

Stop H7: Cedar Creek fault zone area

Travel 300 m west along Highway 17 to a series of outcrops on the north side of the highway.

UTM Zone 16, 581337E, 5393743N

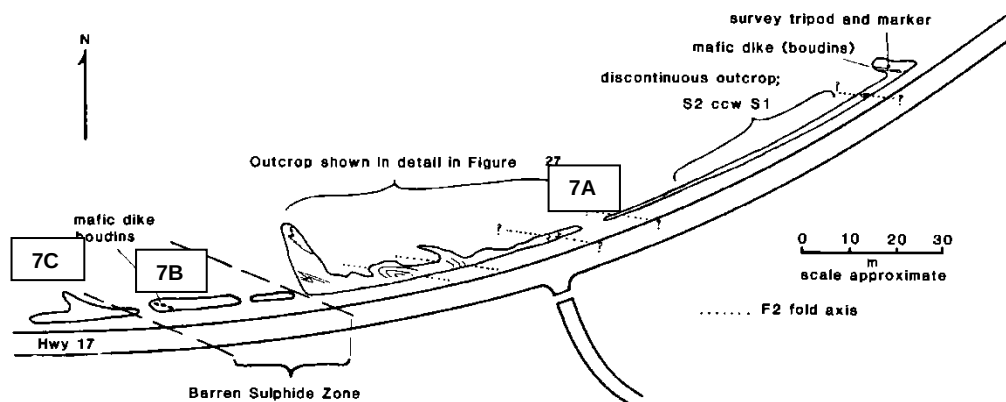


Figure 12. Sketch map of outcrops at stop H7 (from Muir et al. 1995).

Outcrop H7A: Folded, sheared, porphyroblastic metasedimentary units

Compositional layering in this outcrop is generally well defined and likely reflects modified original bedding. Folded layering defines F2 folds and retrograded porphyroblasts (cordierite?) are aligned parallel to S2. Other porphyroblast species found include garnet, and anthophyllite / cummingtonite. Evidence of superposed D3 dextral shear consists of: spaced shear bands and microfabrics with the apparent appropriate geometry and sense of fabric deflection. Some small Z-shaped folds are ambiguous F2 or F3 folds. Note that the orientation of the axial plane of the main F2 fold here is about 270°, whereas the overall orientation of the Williams fold axial plane is about 290°, possibly consistent with some back-rotation during D3. Some of the amphibole-rich (“calc-silicate”) layers reveal evidence of alteration, commonly on both sides of the layers, and this evidence increases toward the Barren Sulphide zone (Outcrop 7B). This suggests that all of the rocks may have been altered. Toward this zone, there is a general change to more thinly laminated pelitic(?) sedimentary rocks, which also show more evidence of alteration (pyrite, muscovite). An increase in the degree of strain toward the zone is evident from the presence of tighter F2 folds and boudinage in calc-silicate layers and a mafic dike (Figure 13). Here the tight F2 folds display S-shaped asymmetry.

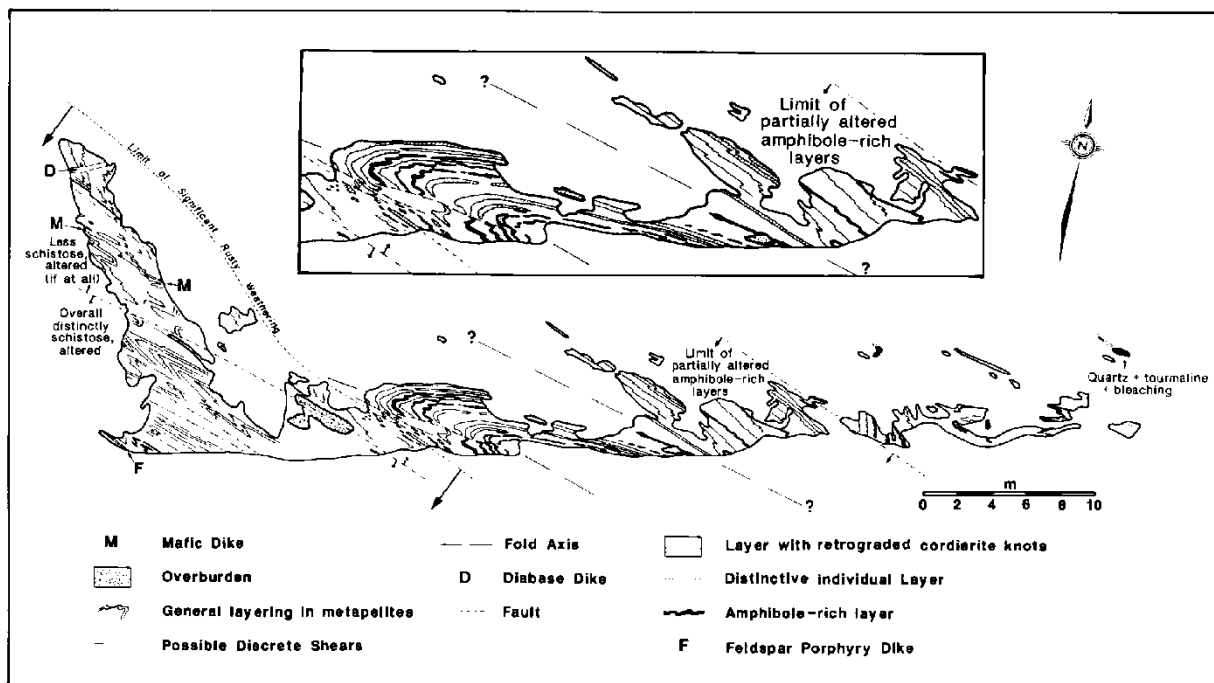


Figure 13. Geological map of outcrop area H7A (after Muir 1990).



Figure 14. F2 fold closure at Outcrop H7A.

Outcrop H7B: Barren Sulphide Zone (Cedar Creek fault zone)

This zone (at least 15 m thick) was also known in the earlier days as the Sucker zone. This is because it had all of the appearances of being a good candidate for a mineralized “horizon,” but in fact is essentially barren. The significant degree of oxidation precludes many unambiguous observations, but it appears to comprise sheared, pyritiferous, quartz-feldspar-sericite schist with small amounts of green mica and irregularly distributed, trace amounts of gold. The rock is possibly derived from the hanging wall pelitic/wacke metasediments (Outcrop H7A). Shearing is quite likely due to D3 based on fabric deflection, geometry of lozenges and the presence of inferred shear bands. Earlier involvement of D2 is possible. The Barren Sulphide zone is inferred to coincide with the Cedar Creek fault zone, one of several high-strain zones along this stretch of highway.

Outcrop H7C: Reworked felsic volcanoclastic rocks

These rocks consist of somewhat rusty weathering, layered, feldspathic, quartz-crystal-bearing rocks, interpreted to be derived from reworked felsic volcanic detritus of the Moose Lake volcanic complex (MLVC; Figure 9) which, in part, is spatially associated with the Hemlo gold deposit. The rocks do not appear to be the protolith to most if any of the muscovitic schists in the Barren Sulphide zone. Minor disseminated pyrite and rare green (vanadian?) muscovite are locally present in the structurally upper parts of this unit, indicating some alteration has taken place.

Stop H8: Bedded felsic fragmental rocks

Cross to the south side of the highway (while carefully watching for traffic) and walk a short distance to a low outcrop that is adjacent to the highway.

UTM Zone 16, 581210E, 5393696N

This is an outcrop of layered, quartz-plagioclase-phyrlic, heterolithic, felsic fragmental rocks that display a variety of clast sizes. The clasts are flattened parallel to S2 and are oriented counter-clockwise with respect to layering (consistent with being on the central limb of the Williams-Teck fold pair).

Stop H9: Felsic fragmental and pelitic sedimentary rocks

Carefully re-cross the highway and walk 100 m west to an outcrop on the north side of the highway.

UTM Zone 16, 581099E, 5393719N

The structurally lower part of Stop H9 consists of an enigmatic unit inferred to be altered, felsic fragmental rocks of volcanic origin. Alteration here is uncharacteristic, showing parts that are greenish, charcoal grey, and pink. Lenses are locally discernible, in part defined by the

distribution of the coloured minerals. Staining for K-feldspar shows microcline tends to be locally distributed around some of the mafic lenses and along some of the dominant cleavage. Note the high degree of strain in this outcrop (likely another, currently unnamed, high-strain zone).

The structurally upper part of this outcrop, which displays a sharp, locally slightly discordant contact with the felsic rocks, consists of hard, thinly laminated pelitic rocks that contain abundant staurolite, garnet and sillimanite (fibrolite). A very tight east-closing F2 fold in the layering is evident, similar in style to that in the pelitic rocks of Outcrop H7A. A possible west-closing fold (south part of unit) is enigmatic but, if present, would make this a Z-shaped fold pair. Some staurolite crystals have overgrown and incorporated the laminations. Clockwise rotation of many staurolite crystals, on both limbs of the F2 fold, as well as the orientation of fibrolite, locally about staurolite and counterclockwise to laminations, are inferred as D3-related features. Reddish alteration along undeformed fractures may be related to common alteration (hematitization and/or epidotization) found near diabase dikes, one of which lies at the west end of the outcrop.

Stop H10: Main Mineralized Zone

Travel 450 m west along Highway 17 to a low outcrop in the ditch on the north side of the highway.

UTM Zone 16, 580652E, 5393727N

This outcrop exposes the gold-mineralized rocks of what had been previously termed the West or Lake Superior zone. Although not directly connected with the main Hemlo orebody (aka “B” zone) it represents both its on-strike and up-dip projection that was the focus of the vast majority of exploration efforts prior to the discovery of the main orebody in May, 1981.

Figure 15 attempts to indicate various features that were visible several years ago. Close examination near the westernmost channel sample (outcrop “B”) reveals highly strained pyritiferous fragmental rocks with, among other things, sparse green (vanadian) muscovite (gMs) lenses and a general grey to bluish grey colour in parts of the matrix due to finely disseminated molybdenite, which is generally a “pathfinder” to gold in this deposit. Grab and channel sampling has returned up to 6 g/t Au. The outcrop contains appreciable barite crystals (Davis and Lin 2003).

The high degree of strain in this fragmental rock reflects that these rocks lie within the Lake Superior Shear zone, one of several high-strain zones in the Hemlo deposit area. This high-strain zone is up to 50 m thick. Complex fabric relationships with a possible chronologic interpretation are indicated in the inset at the upper part of Figure 15 pertaining to outcrop “B”. F3 folds (“flanking structures”) can be seen in the southeast part of outcrop “B”.

About 50 m west along the highway ditch is outcrop “A”, which displays the more “massive” quartz-plagioclase porphyry (QPP) that structurally underlies the Main Mineralized zone. Also evident is an example of one of many (in the area) swarms of plagioclase-phyrlic dikes. Note that the dikes are deformed: extension followed by shortening, reflecting boudinage related to D2 predated shortening related to D3.

Also note that in these outcrops and all other ones already seen and to be seen later today, that quartz veins are generally few and small; there typically being as many, if not more, small pods or lenses of quartz as there are veins. Full interpretation of kinetics is commonly enigmatic. This attests to the complex history of strain.

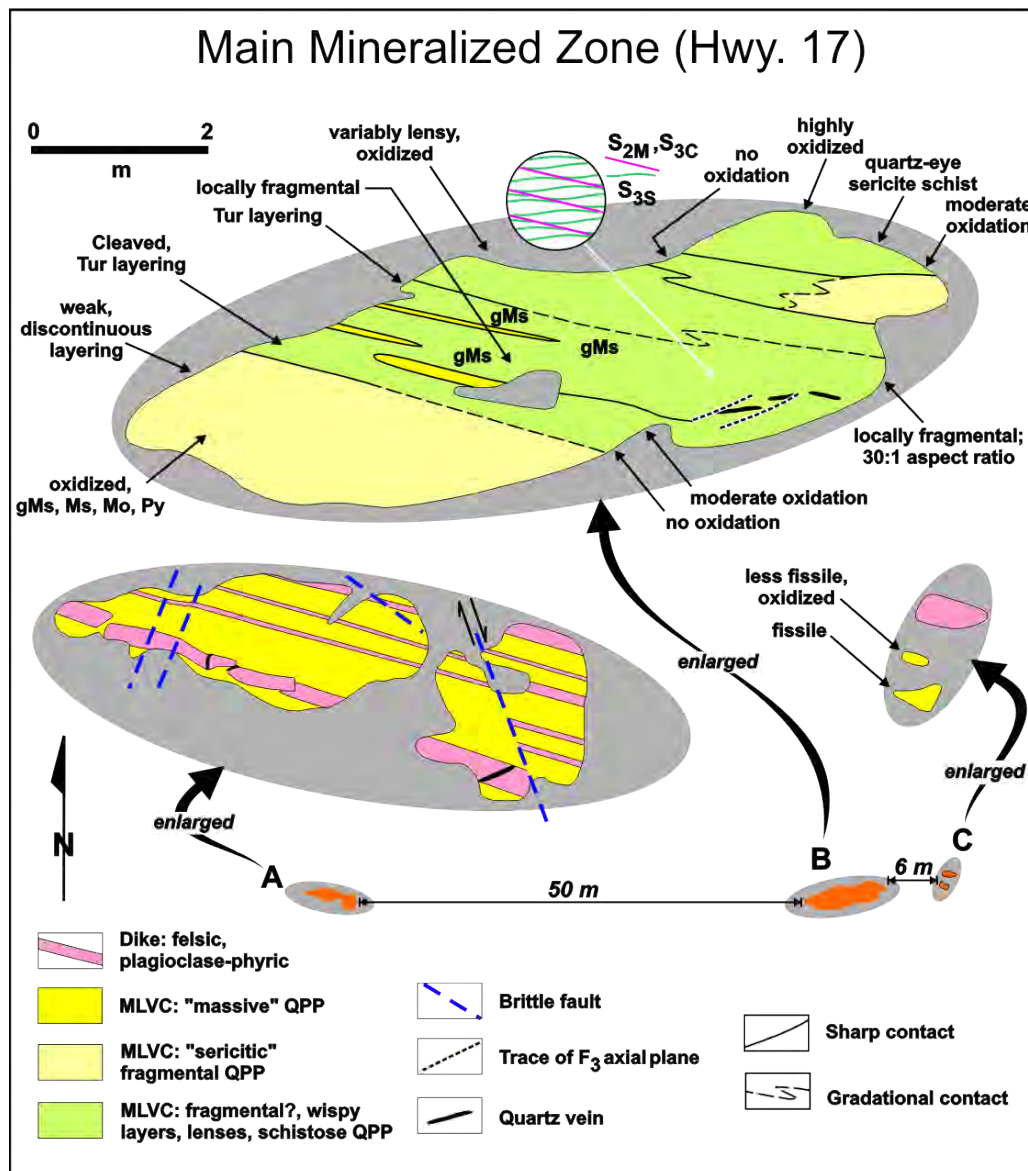


Figure 15. Main Mineralized Zone (also known as West Zone) geology (*after* Muir and Smyk 2006).

Stop H11: Inter-ore zones metasedimentary rocks.

Travel 180 m west along Highway 17 to a pair of outcrops on the north side of the highway that are bisected by a road that formerly accessed the Williams A Zone open pit.

UTM Zone 16, 580471E, 5393729N

The outcrop to the east of the A Zone pit road (see Figure 16) displays relatively feldspathic, turbiditic sedimentary layers, with well-preserved grain-size gradation indicating overturned, south-southwest facing units. The well-preserved nature of this outcrop, contrasts with the adjacent rocks (hanging wall and footwall) and suggests that it may represent a low-strain lithon.

To the west of the A Zone road, the sediments are darker grey and locally contain magnetite-bearing layers and discrete folds. At the western end of the outcrop, the layering is interpreted as transposed. The increase in degree of strain over an inferred 5 to 10 m across strike between outcrops is considerable and is inferred to reflect, in part, the Moose Lake fault zone. At least 2 types of feldspar porphyry dikes are exposed in this outcrop. Out of interest, the structurally lower one, locally called "popcorn" porphyry because of its coarser plagioclase phenocrysts, was used in the early stages of exploration as a cutoff marker unit for drilling collared to the north. Note that, ironically, the Lower Mineralized zone occurs structurally below this dike, and is exposed about 110 m further west along the highway.

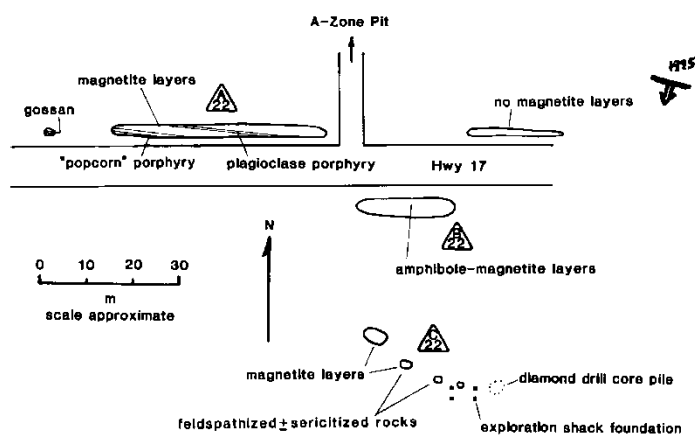


Figure 16. Sketch map of outcrops at Stop H11 (from Muir et al. 1995). The number labels shown on this figure denote the 1995 field trip stop numbers.

The A Zone pit can be seen through the fence located just to the north of the outcrops at this stop. Figure 17 illustrates the pre-mining geology of the surface expression of the A Zone. The ore zone occurs at the contact between metasedimentary and fragmental rocks.

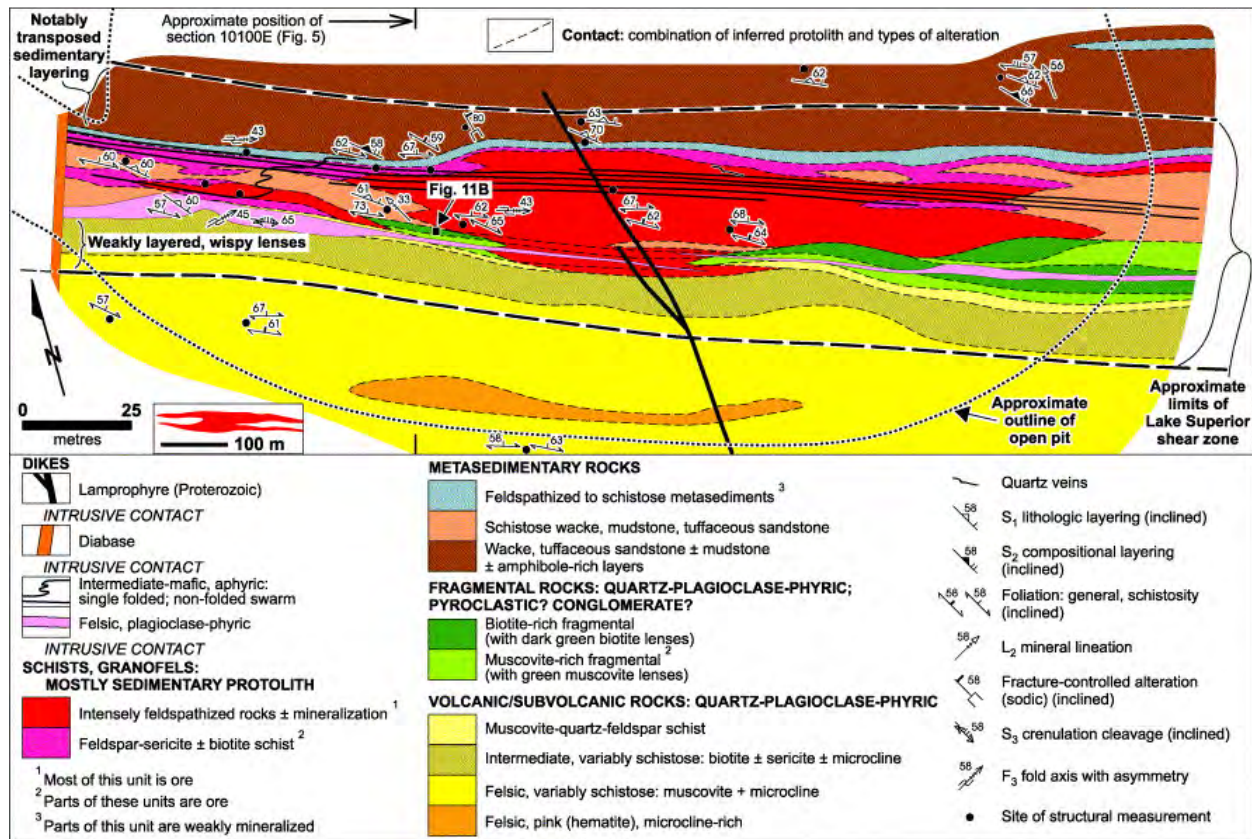


Figure 17. Pre-mining geological map of the A Zone. The approximate outline of the ore zone is shown in red (from Muir 2002; inset showing ore outline derived from Walford et al. 1986).

Stop H12: Up-dip projection of the Lower Mineralized Zone.

Travel 110 m west to another outcrop on the north side of the highway.

UTM Zone 16, 580311E, 5393727N

This outcrop is the up-dip projected zone of alteration associated with the actual Lower Mineralized Zone (which occurs at about 900 m depth here). However, the rocks at surface are altered, and locally contain low gold values, minor amounts of green (vanadian?) muscovite, pyrite, medium to coarse-grained tourmaline, muscovite, and rarely microcline porphyroblasts. Barite veinlets and layer-parallel seams have been reported (Schnieders, pers. com. 1994). Magnetite-bearing wacke, which is at the north part of the outcrop (see Figure 18), becomes muscovitic and possibly silicified adjacent to the quartz-plagioclase porphyry (QPP of the Moose Lake Volcanic Complex). The contact between the altered sedimentary rocks and the QPP is locally at an uncharacteristically high angle to the layering in part of the outcrop (folded? crosscutting?). Granodioritic, feldspar porphyritic, and intermediate to mafic dikes occur here.

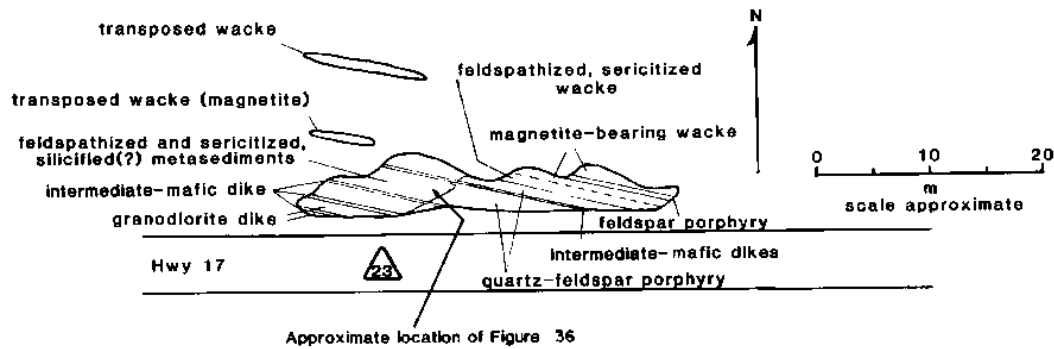


Figure 18. Sketch map of the up-dip projection of the Lower Mineralized Zone (from Muir et al. 1995). The number label shown on this figure denotes the 1995 field trip stop number.

Stop H13: Hemlo Fault zone.

Travel 1.5 km west along highway 17 (past the Williams Mine entrance road and tailings line overpass) to a steep road cut on the north side of the highway.

UTM Zone 16, 578843E, 5393791N

This exposure shows part of the fault zone (which is probably up to tens of metres thick), and features the structurally lower amphibolite/gneissic amphibolite (inferred to be derived from pillowed and massive mafic volcanic units, based on outcrops to the west) and the structurally overlying feldspathic sedimentary rocks. Mafic-ultramafic dikes occur largely as chlorite $\pm$ actinolite $\pm$ talc phyllonites, mostly within the amphibolitic rocks, and provide much of the evidence for relatively late D3 dextral shear, as can be seen by fabric relationships. Also occurring here are a few intermediate dikes (presently biotite schists) and a plagioclase-phyric dike swarm. Very-coarse-grained tourmaline crystals and crystal “clots” are associated with a feldspar $\pm$ quartz dike or vein. The outcrop, featured in Figure 19, was modified in 1989 (post-mapping by Smart 1988) by roadway blasting.

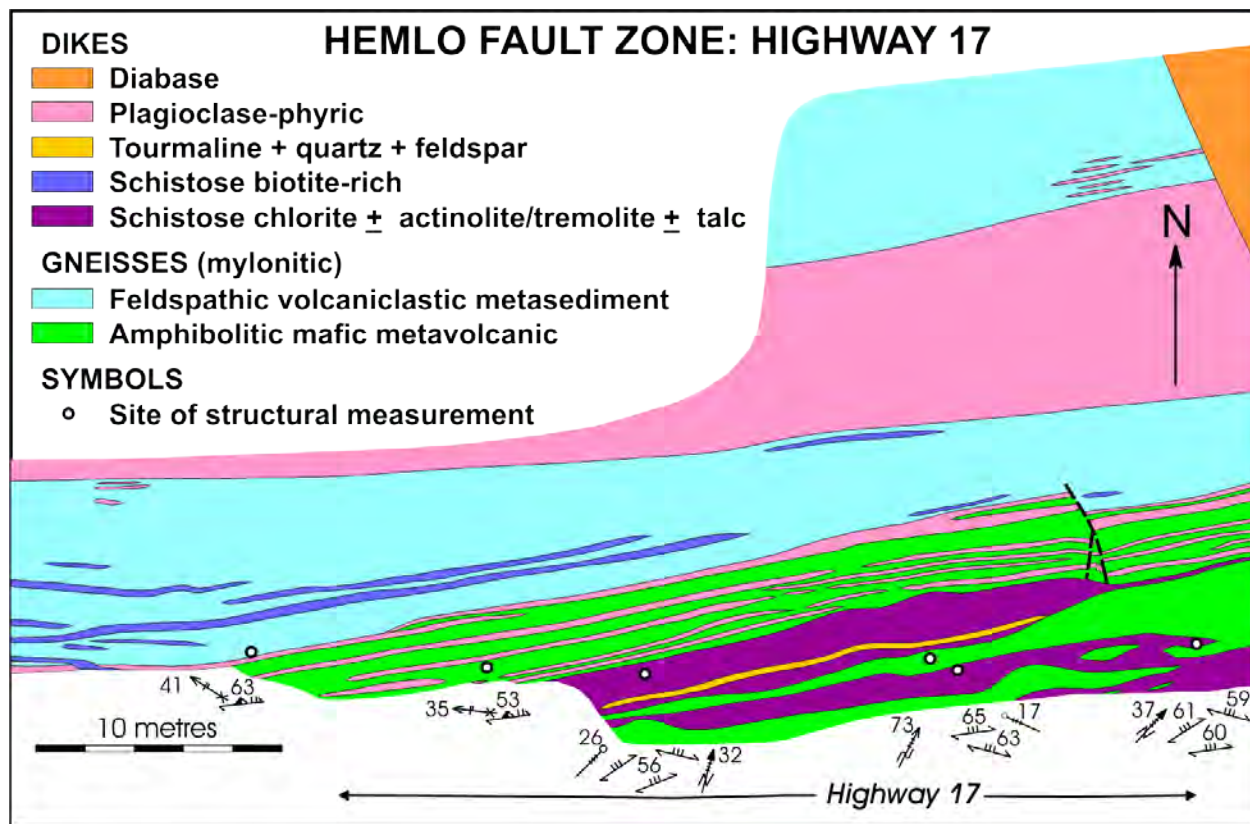


Figure 19. Geology of the Hemlo fault zone at Stop H13 (modified from Smart 1988).

Stop H14: Highway Zone

Travel 1.3 km west along Highway 17 to a large outcrop area on the north side of the highway.

UTM Zone 16, 577540E, 5393717N

This outcrop (see geological map, Figure 20) demonstrates the heterogeneous strain that occurs even away from the Hemlo gold deposit. It presumably represents the structurally upper part of the tens-of-metres thick Hemlo fault zone. The north part of the outcrop displays disrupted layering in feldspathic siltstone and arenite with S-shaped folds. As one approaches the highway (and hence the Hemlo fault zone), the units change to:

- 1) quartz-plagioclase-phyric, heterolithic and monolithic, felsic fragmental rocks, displaying contacts tightly interfolded(?) with wacke. The felsic fragmental rocks are pyritiferous, erratically auriferous and locally contain green (vanadian?) muscovite; and then
- 2) wacke-siltstone which displays well-developed laminations that are thinly spaced and inferred to be derived, in part, from a combination of primary layering and compositional differentiation along the main cleavage.

Very tight folds (south of the diabase dike) with S-shaped asymmetry are evident. One interpretation is that there is a composite S1/S2 fabric that is locally preserved and oriented

slightly clockwise to the dominant S_m (mylonitic) fabric. The S_3 flattening fabric (micaceous) can be locally seen within the S_m cleavage, counter-clockwise to it. The contact with the amphibolitic and mafic-ultramafic rocks seen in Stop H13 presumably lies to the south of the highway.

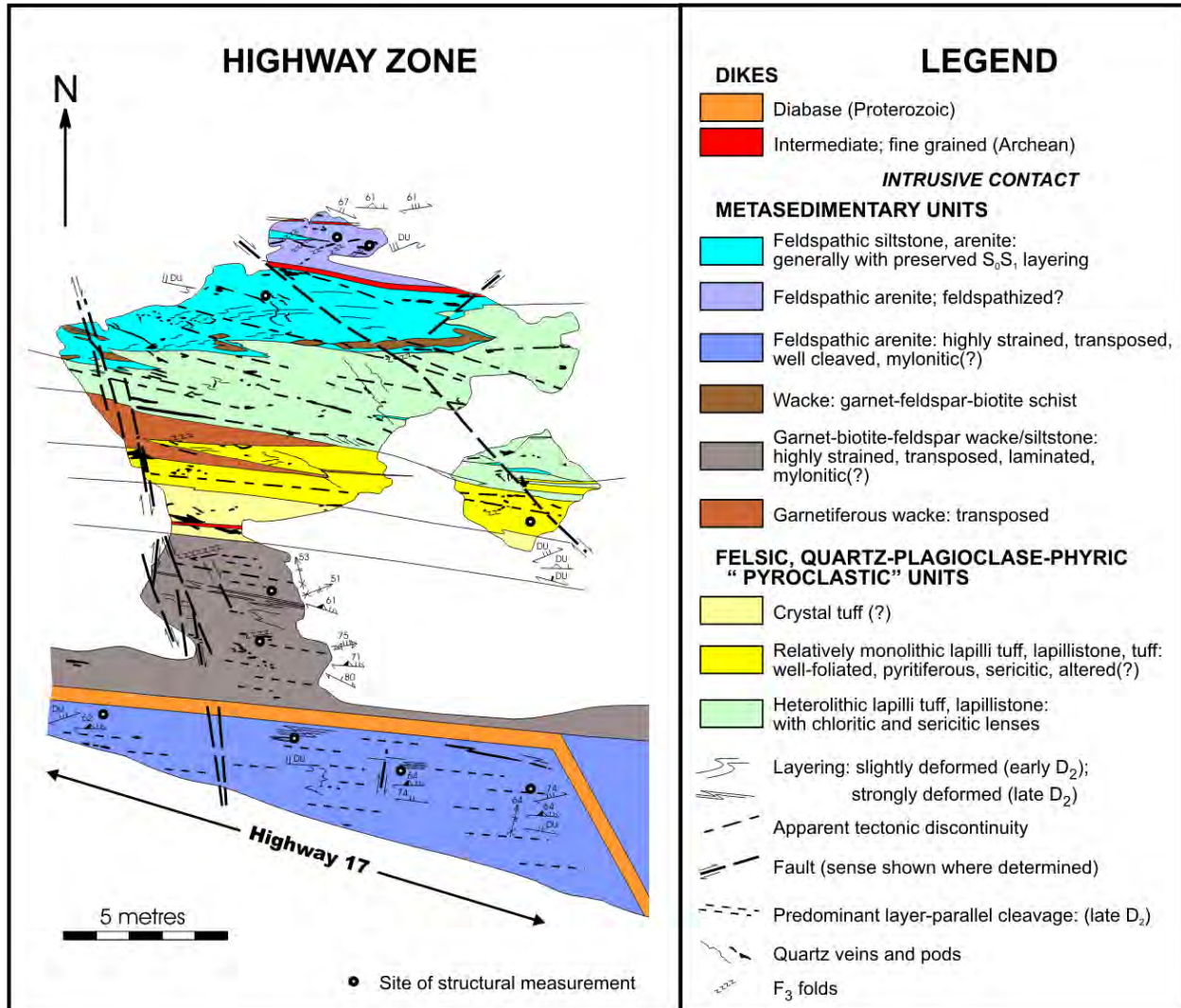


Figure 20. Geological map of the Highway Zone at Stop H14 (modified from Muir et al. 1995).

Stop H15: Homestake F3 fold.

Travel 3 km west along Highway 17 to a high road cut on the south side of the highway.

UTM Zone 16, 574529E, 5393489N

This outcrop (see geological map, Figure 21) is possibly north of the Hemlo fault zone, but ambiguity abounds. Several features in this otherwise over-lichened outcrop are of note:

- 1) Rare, calc-alkalic or “alkalic” pillows with feldspar phenocrysts and hornblende crystals, locally with amygdaloidal/vesicular textures;
- 2) Abrupt strain partitioning from slightly strained pillows (near the highway) to moderately strained pillows (D2) over a distance of less than 1 m southward;
- 3) A related volcanic breccia or volcanoclastic unit (fragments with similar composition and textures);
- 4) Adjacent feldspathic arenite and/or altered wacke;
- 5) A short, thin gossan, which returned 103 ppm Au and 100 ppm Mo (Resident Geologist’s files, Thunder Bay South District) – the highest molybdenite results outside of the Hemlo gold deposit proper and similar to some feldspathized tholeiitic pillows “along strike” to the west of the deposit;
- 6) Varieties of variably deformed metawacke and metasiltstone, locally with garnet ± magnetite ± staurolite porphyroblasts;
- 7) Examples of S2 and F2 structural elements overprinted by S3 and F3 structural elements (e.g., folded S2);
- 8) Numerous variably deformed (folded, transposed) dikes such as mafic schists; mafic biotite lamprophyre schists ± mafic-ultramafic xenoliths; a mafic net-veined or back-veined dike (mixture of previous 2 types?); and 2 types of intermediate dikes;
- 9) The xenolith-rich parts of the lamprophyre dikes are very similar to some of the diamondiferous intrusions found in the Wawa greenstone belt.

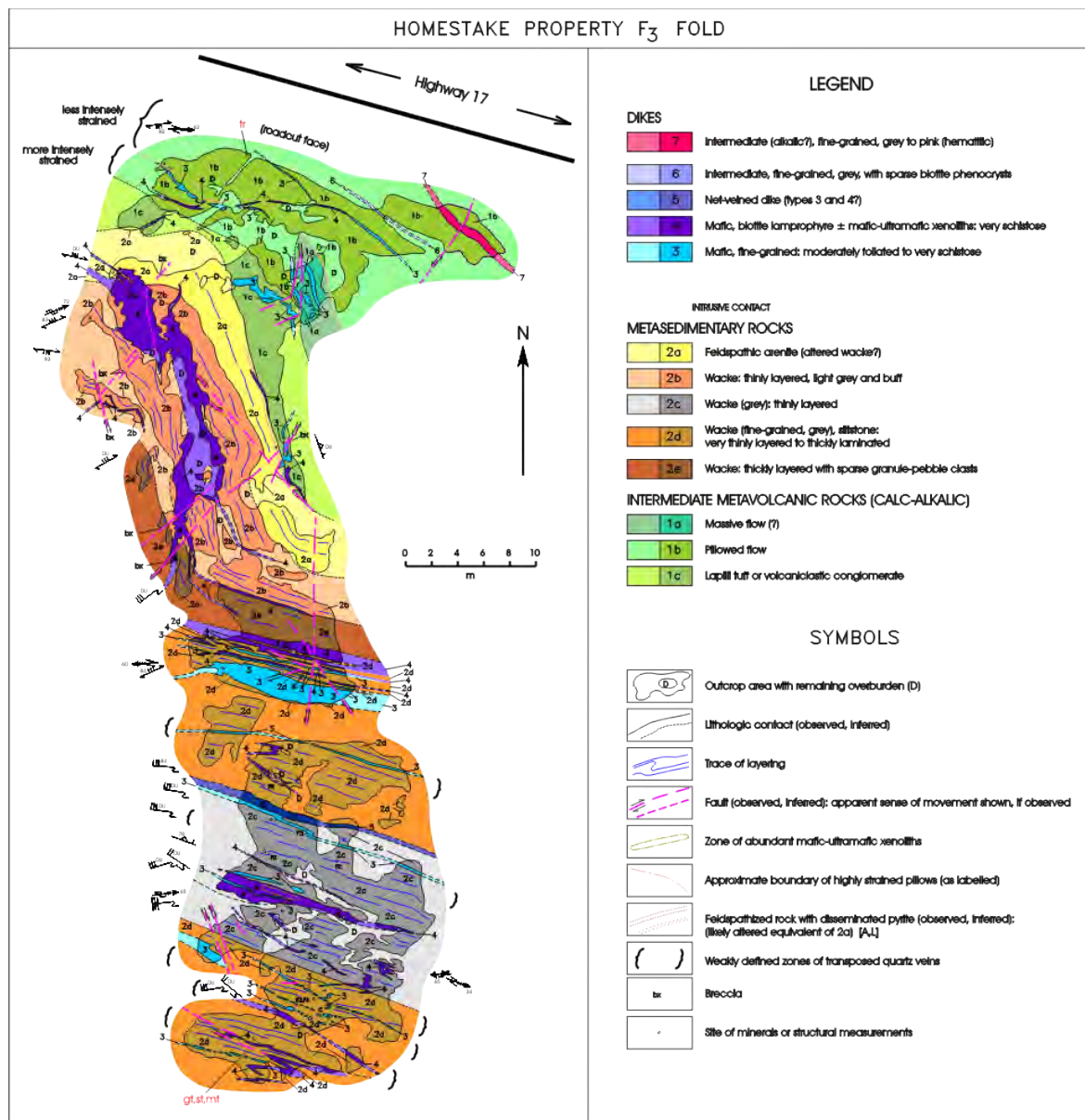


Figure 21. Detailed geology of the Homestake F3 fold exposure (modified from Muir et al. 1995).

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Field Trip 2

Unusual Archean Diamond-bearing rocks of the Wawa Area

by

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Resident Geologist's Program,
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Timmins, Ontario



On the cover (clockwise from top): Giant lower crustal to upper mantle xenoliths, Enigma Property, Menzies Township, Oasis Diamond Corporation Inc.; Diamonds from the Festival Property (photo courtesy of Pele Mountain Resources Inc.); Sandor Diamond Occurrence, Highway 17, Spider Resources Inc. & KWG Resources Inc.; Heterolithic diamond-bearing breccia, Engagement Zone, Northern Sierra Minerals Corporation

Unusual Diamond-bearing rocks of the Wawa Area

Introduction

In 1991, local prospector C. “Mickey” Clement recovered at least three alluvial diamonds from the Michipicoten River. Two of the stones were sent to the Department of Mineralogy, Royal Ontario Museum, where they were identified as industrial-grade diamonds with weights of 1.05 and 1.13 carats. Both stones were described as frosted and graphite-inclusion riddled.

In 1995, prospector Sandor Surmacz and geologist Marcelle Hauseux recovered diamonds from a bedrock occurrence on the Trans Canada Highway, approximately 20 km north of the town of Wawa. An 18.1-kg bulk sample of a xenolith-rich lamprophyre yielded 1 macrodiamond and 5 microdiamonds. All but one was gem quality.

Since then, more than 50 occurrences of diamondiferous bedrock have been reported in an area of approximately 30 km² in size, centred approximately 20 km north of the town of Wawa. The occurrences are hosted within a sequence of unusual, Archean-aged, heterolithic breccias. Historically, this sequence of rocks has received little exploration interest and was considered to have little economic significance.

This field trip will focus on exposures of a series of foliated lamprophyre dikes and associated heterolithic breccias outcropping in Lalibert, Leclair, Menzies and Musquash townships. This field trip guide represents a summary of information available at the time of writing and should not be considered the final analysis of these rock types. Much more research is required on these rocks. Active exploration and research is still underway on the properties included in this field guide. Given the limitations of time, the field trip will visit only some of the more accessible properties. Bear in mind that when visiting active exploration or mine properties, permission must be granted by the property owner. Current ownership information can be obtained from the Resident Geologist’s Office in Timmins, or the District Geologist’s Office in Sault Ste Marie, Ontario.

Geological Overview of the Michipicoten Greenstone Belt

The Wawa region lies within the Wawa subprovince of the Canadian Shield. The Michipicoten greenstone belt extends inland for approximately 150 km from the Lake Superior shore and has an average width of 38 km. The greenstone belt consists of supracrustal rocks of Archean age. Younger Archean granitic rocks surround the greenstone belt. Figure 1 shows a generalized geological map of the Michipicoten greenstone belt.

The oldest volcanic cycle is approximately 2900 Ma and is of limited distribution. This cycle is best developed in Esquega Township along the southern flank of the supracrustal assemblage. Portions of this metavolcanic cycle extend into eastern McMurray and western Lastheels townships. The base of this volcanic cycle consists of massive to pillowed komatiitic flows intruded by mafic sills. The ultramafic rocks are overlain by intermediate to felsic tuffs capped by thinly bedded chert-magnetite-sulphide iron formation (Judith-Kathleen Iron Formation). Intermediate to felsic metavolcanic tuffs below the Judith iron formation have been dated at 2889 ± 9 Ma (Turek et al. 1992).

Overlying the 2900 Ma cycle is a 2750 Ma volcanic cycle. This volcanic cycle is predominately composed of intermediate to felsic tuffs, breccias and flows. Porphyritic and spherulitic flows are not common and most of the intermediate to felsic metavolcanic rock is fragmental. The base of this cycle consists of high magnesium and high iron tholeiitic massive and pillowed flows. It lies conformably atop the Judith-Kathleen iron formation at the east end of Wawa Lake, but the basal unit is poorly exposed elsewhere. Overlying the mafic metavolcanic rocks is a sequence of heterolithic, intermediate to mafic breccia that has been traced for a distance of over 13 kilometers. This unit is in turn, overlain by a thick section of intermediate to felsic tuffs, breccias and massive flows that reaches a maximum thickness of approximately 2000 m below the Helen Iron Formation. The upper part of the intermediate to felsic metavolcanic rocks has been dated at 2749 ± 2 Ma (Turek et al. 1992). The Michipicoten (Helen) iron formation caps this

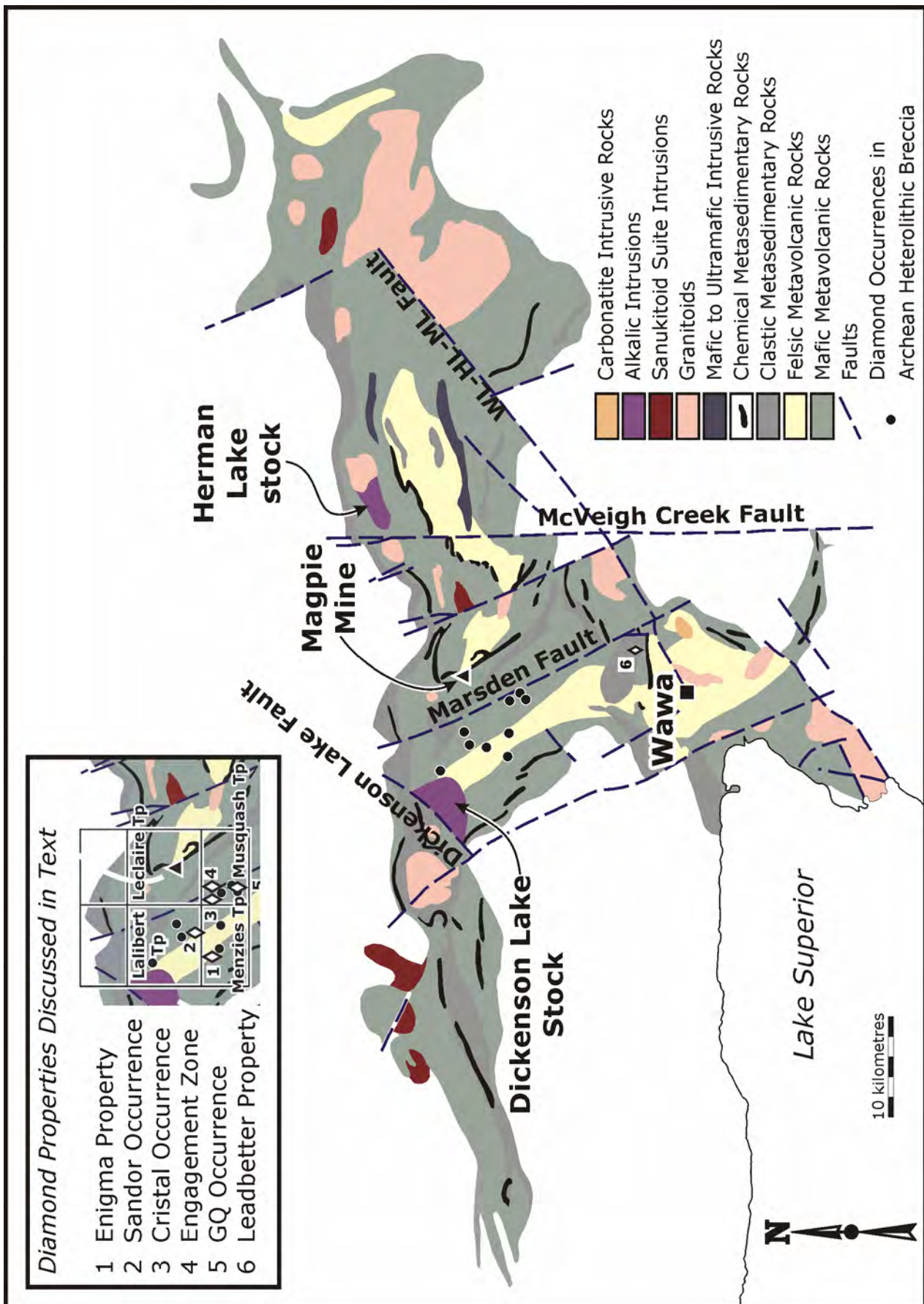


Figure 1. Generalized geological map of the Michipicoten greenstone belt showing some of the diamond occurrences (modified after Stott et al. 2002).

this volcanic cycle. It commonly exceeds 100 m in thickness and, in the vicinity of the past-producing Helen Mine; it has been tectonically thickened to over 300 m. This iron formation was the source of virtually all commercial iron ore production in the Wawa area from 1898 to 1998.

The youngest metavolcanic rocks in the area are those of the 2700 Ma volcanic cycle. These rocks underlie approximately fifty-percent of the Michipicoten greenstone belt and are found in the central and northern parts of the belt. The basal unit is composed of massive and pillowed mafic amygdaloidal flows that attain a maximum thickness of 5.5 km in the Goudreau area. This unit is overlain by intermediate to felsic metavolcanic rocks, or their stratigraphic equivalent, the Doré metasedimentary rocks. These metavolcanic rocks are typically composed of tuffs and monolithic and heterolithic breccias. Quartz + feldspar crystal tuff and an intermediate tuff from this volcanic cycle returned a U-Pb zircon age of 2701 ± 8 Ma (Turek et al. 1992) and 2701.4 ± 2.1 Ma (Ayer et al. 2003). The intermediate to felsic tuffs interdigitate with clastic metasedimentary sequences that include cross-bedded sandstone and a tonalite cobble conglomerate (Doré conglomerate). Corfu and Sage (1987, 1992) reported an age of 2698 ± 2 Ma for a tonalite clast in the Doré conglomerate and maximum ages of 2680 ± 3 and 2682 ± 3 Ma for sedimentary sequences in northern and central parts of the Michipicoten greenstone belt.

Geochronological and structural evidence indicates that sedimentation continued after cycle 3 volcanism and predated a major folding and faulting event. Arias (1996) noted that the rocks comprising cycle 3 in the central part of the Michipicoten greenstone belt are upside down and represent the overturned limb of a belt-scale recumbent nappe fold. This inverted limb has been refolded and imbricated by subsequent south-verging thrust faults, which caused local repetition of the stratigraphic sequence (Wilson 2004).

Felsic plutonism occurred synchronous with all of the major volcanic cycles and continued after volcanism ceased at Wawa. Plutonic rocks associated with cycle 1 volcanism include the Murray-Algoma porphyry (2881 ± 6 Ma) and the Regnery biotite granite of the Hawk Lake granitic complex (2888 ± 2 Ma). Both intrusions are situated in Esquega Township. The Jubilee granitic stock, located in McMurray Township, was dated at 2745 ± 3 Ma and is coeval with cycle 2 volcanism. Plutons associated with cycle 3 volcanism range in composition from tonalite through granodiorite and granite and have ages ranging from 2698 to 2693 Ma. These plutons are located south and west of the Michipicoten greenstone belt (Stone and Semenyina 2004).

The Kapuskasing Structural Zone extends east from the shore of Lake Superior, northeast through Kapuskasing and into the Hudson Bay Lowland. Local features interpreted to be associated with it include northeast-striking Proterozoic lamprophyre dikes (Sage 1994; Morris 1999).

Lamprophyre dikes of middle Proterozoic age are common in the region south of the Wawa-Hawk-Manitowik Lake Fault and rare to non-existent north of the fault. They are carbonate, biotite, and sometimes olivine-rich and usually less than 1.0 m in width. The dikes generally strike northeast. Dikes exposed in McMurray Township commonly have blue to blue-green sodic amphibole developed in their wall rocks. This mineral has been interpreted to be a product of fenitization. These dikes are likely spatially and temporally related to the emplacement of the Keewenawan-age Firesand River Carbonatite (Sage 1994).

North of the Wawa-Hawk-Manitowik Lake Fault, in the area extending west from the former Magpie Mine (Leclaire Township), to the east side of the Dickenson Lake Stock (Lalibert Township), there are a series of what appear to be narrow, biotite-amphibole-rich dikes that have been interpreted as Archean lamprophyres. These dikes commonly have large, rounded inclusions (lower crust to upper mantle-derived xenoliths) up to 3.0 m in size, the centres of which are often completely altered to radiating clusters of actinolite crystals.

Titanite from the matrix of one of these dikes returned an age of 2703 ± 42 Ma (Sage 2000). The date is interpreted to be a minimum age of intrusion. Subsequent dating of a zircon from a gneissic xenolith from the same dike returned an age of 2684.9 ± 1.4 Ma (Ketchum, Kamo and Davis 2003).

Several narrow (<10 m) kimberlite dikes also intrude the Michipicoten greenstone belt. These dikes are largely restricted to the eastern part of the greenstone belt and are spatially related to the northeast-trending Kapuskasing Structural Zone. Several of these dikes have been dated. Intrusions K115 and K121, located in Isaac Township, returned an average $^{87}\text{Rb}/^{86}\text{Sr}$ age of 1097 ± 7 Ma (Kaminsky et al. 2002). A kimberlite dike intersected in Pele Mountain Resources Inc.'s drill hole 97-34, drilled in Riggs Township, returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1197 ± 24 Ma (R. P. Sage, Ontario Geological Survey, unpublished, 2000; Wilson 2004).

Quaternary Geology

All Quaternary deposits within the Wawa area were deposited during the Late Wisconsin by the Labrador sector of the Laurentide Ice Sheet. Peat, recovered from a bog located within the surface of a terrace associated with the highest glacial lake in the Lake Superior basin, was radiocarbon dated at 9759 ± 170 BP. A caribou antler, recovered from near the surface of the delta of glacial lake Minong III, yielded a radiocarbon date of 8820 ± 80 BP (Morris 2001).

Bedrock striae indicate that there were two prominent ice flow directions. The oldest and most pervasive ice flow was south to southwest ($159^\circ - 240^\circ$). A later, weaker ice flow was to the southwest and west ($220^\circ - 290^\circ$). The younger set of striae was formed during the latter stages of glaciation as the ice sheet began to thin and bedrock topography began to influence the direction of ice flow.

Much of the overburden consists of a thin (< 1 m), discontinuous till veneer draped over the bedrock. Most of the area's thicker surficial deposits are located within bedrock controlled valleys where glaciolacustrine waters from the Lake Superior basin covered the area. In several of these valleys, the glaciofluvial material reaches thicknesses up to 32 m (Morris 2001).

Structural Geology of the Diamond-bearing Rocks

The greatest concentration of diamond-bearing rocks in the Michipicoten greenstone belt is constrained to a roughly 30 km² block of land coinciding with a D₁ recumbent nappe identified by Arias (1996). The nappe lies immediately north and adjacent to the Kapuskasing Structural Zone and has affected the two youngest rock sequences in the greenstone belt.

In cross section, a generalized structural section across the greenstone belt would be upside down and represents the overturned limb of a belt-scale recumbent fold. This inverted limb of the nappe fold has been refolded and imbricated by subsequent south-verging thrust faults. The result is that the geometry of the rocks hosting the diamondiferous bedrock is that of an inverted anticline (Arias and Helmstaedt 1990). The effect is to create a tectonic repetition of the diamond-bearing sequence of rocks. In simplified terms, the rocks hosting the diamonds in the Wawa area can be considered as a single, overturned fold limb that has been faulted along a series of thrust zones. As a result, the diamond-bearing breccias, the lamprophyres associated with them and the surface onto which these metavolcanic rocks were deposited has been repeated at least four times (Walker 2002).

The western limit of the nappe is likely the Dickenson Lake fault that passes along the west side of the Dickenson Lake stock. Extrusive lithologies similar to those hosting diamonds were recognized to the east of the stock by Sage (1993) during reconnaissance mapping. To date, no diamondiferous occurrences have been discovered to the west of the Dickenson lake fault. The eastern limit of the nappe is the Marsden Lake fault. Prior to 2002, rocks favourable to hosting diamonds had not been observed east of this structure (Wilson 2004). However, in 2002, Oasis Diamond Exploration Inc. made a discovery of diamonds on the east shore of the Magpie River and in 2004, diamondiferous occurrences of bedrock were found on the west shore of the Magpie River in Chabanel Township. These discoveries suggest that the potential for these host rocks extends farther eastward than previously believed. Figure 2 shows a graphic representation of the nappe structure.

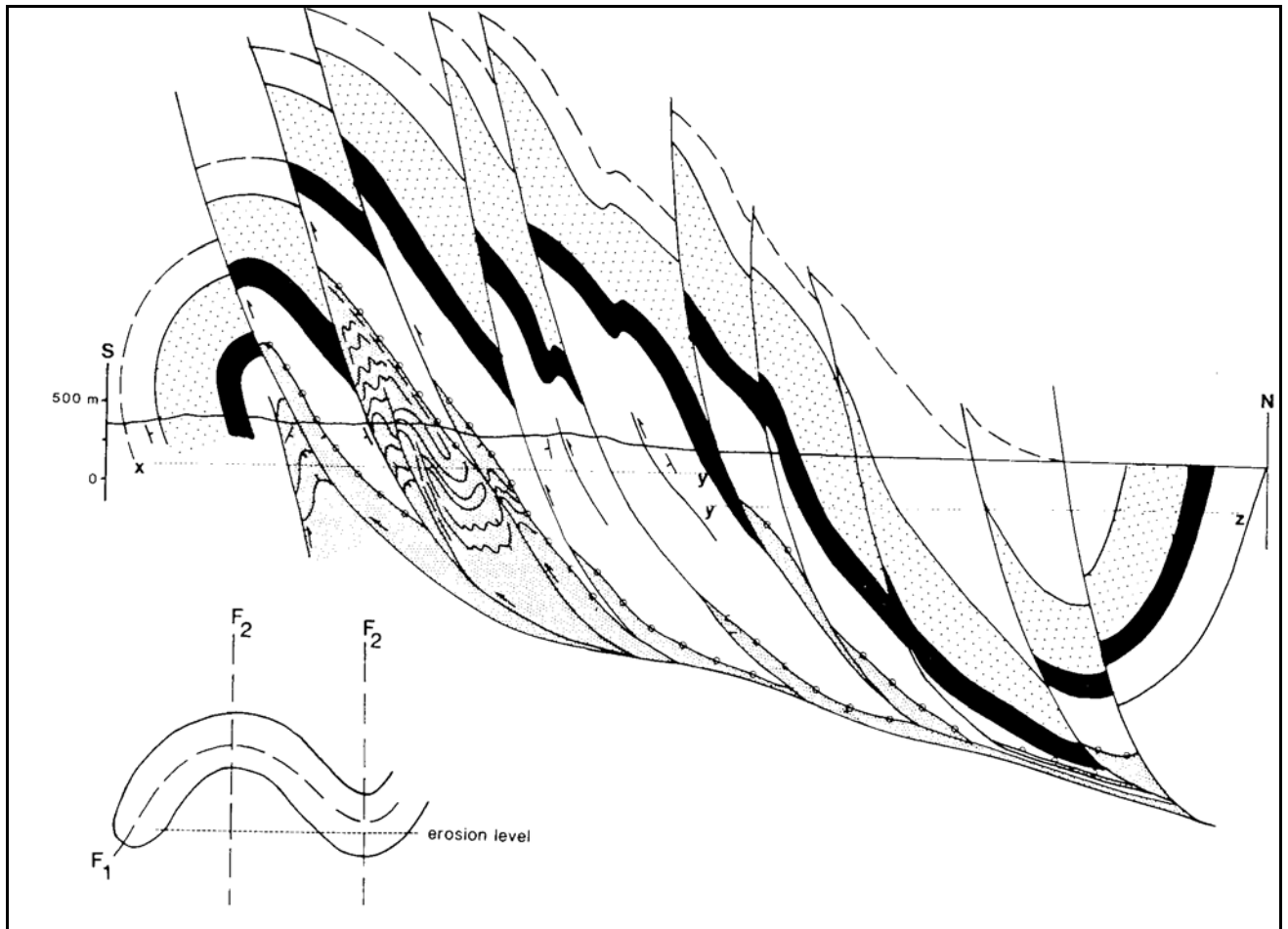


Figure 2. Composite structural section through the central part of the Michipicoten greenstone belt. Section X-Y from south-central Corbiere Tp. (Josephine Iron Range) to Andre Lake central Corbiere Tp. Section Y-Z is a schematic section from west central Lalibert Tp. The sketch in the lower left explains the present configuration of the belt as a regional nappe fold (F_1) refolded about F_2 . Imbricate thrusts are considered related to F_s (Arias 1996).

Description of the Diamond-bearing Rocks

The diamond-bearing breccia and associated lamprophyre is broadly distributed throughout Lalibert, Leclaire, Menzies and Musquash townships. On-going mapping by Pele Mountain Resources Inc. on the Festival property and by Nathalie Lefebvre on the GQ Property has helped to refine the classification of these rocks. Systematic exploration and sampling suggest that the individual diamond occurrences are part of a much larger suite of rocks and that diamonds occur primarily within discrete layers at the base of diamond-bearing zones.

North-northwest-trending diamond-bearing zones of breccia and lamprophyre are up to 1500 m in length and up to 800 m in width (Pele Mountain Resources Inc., press release, January 18, 2005). The breccia forms thick units (maximum true thickness is approximately 110 m) dipping to the northeast 30°. The lateral extent and thickness of the breccia unit is not well constrained, owing to the large-scale regional folding and thrusting (Lefebvre 2004). Figure 3 provides a detailed map of the southwest corner of the Festival Property showing a recent interpretation of these diamond-bearing zones.

The diamond-bearing rocks can be visually subdivided into two classes, lamprophyre (dikes and bodies of indeterminate morphology) and heterolithic or polymict breccias. It is often difficult to differentiate between the two classes since the lamprophyre dikes frequently contain an assortment of inclusions that



Figure 3. Detailed geology of the southwestern corner of the Festival Property (modified from Pele Mountain Resources, Inc. press release, January 18, 2005). The units shown in green consist of mafic to intermediate massive and pillowed flows, breccias and tuffs (Vaillancourt 2005b, Sage et al. 1982)

give them the appearance of breccia. The lamprophyre dikes cut the breccia units. Both lithologies have been metamorphosed to upper greenschist facies.

The breccia primarily consists of angular, pebble-sized, lithic fragments, mainly of volcanic composition, contained within a green to grey fine-grained matrix. The matrix grain size ranges from < 2 mm to 1 mm. At least eleven distinctive types of lithic fragments have been observed in the breccia and the fragments are irregularly distributed throughout the breccia. The clast population is primarily derived from rocks with which the breccia is intercalated. Most typically these clasts are mafic and felsic metavolcanic rocks and intermediate to mafic intrusive rocks. Other clast types include fragments of clast-supported breccia within matrix supported breccia, fragments of earlier matrix-supported breccia with fewer than 5% fragments and coated lithic fragments (Lefebvre 2004).

The breccia is characteristically massive, unstratified and poorly sorted with clast size ranging from sand to boulders up to 9 m. Primary sedimentary structures such as bedding and crude grading are rare (Lefebvre 2004).

Petrographic work by Lefebvre (2004) on the breccias identified a typical fragmental texture within a mineralogically variable matrix. Typically the groundmass is dominated by actinolite, but chlorite and biotite dominated groundmass are locally predominant. Juvenile magmatic material also was observed as discrete fragments and rims on other clasts in the breccia. Its petrography is distinct from the breccia groundmass. The juvenile magmatic material contains more abundant actinolite grains; fewer epidote and fine-grained plagioclase grains and more fine-grained oligoclase and muscovite; and more oscillatory-zoned hornblende.

Most exploration efforts over the past few years have concentrated on the breccias because they are considered to have the best potential for hosting commercial diamond deposits. Over the course of the last few years, explorationists have subdivided the diamond-bearing breccias into three separate facies. These facies are volcanic (pyroclastic), subvolcanic/intrusive breccia and hypabyssal facies. A variable, but distinctive proportion and composition of fragments and/or xenoliths characterizes each facies (Wilson 2004).

The volcanic facies contains breccia, lapilli- and ash sized fragments and consists of medium to thickly bedded pyroclastic air-fall deposits. They are characterized by angular to sub-angular Archean supracrustal fragments, some hypabyssal fragments and rare lower crustal to upper mantle xenoliths.

The subvolcanic/intrusive breccia facies are the most variable in texture and appearance. The rock is characterized by observed intrusive relationships, a high proportion of fragments and a close proximity to the volcanic facies. The fragment characterization is variable. The facies can contain all or some of the following fragment types: supracrustal fragments, crustal fragments and lower crustal to upper mantle xenoliths. This facies was once included with the lamprophyres but is now interpreted by various exploration companies as a debris flow.

The hypabyssal facies hosts variable proportions (<25%), of sub-rounded to rounded mantle xenoliths, as well as minor proportions of gneiss and/or trondhjemite fragments. This facies also was originally classified as lamprophyre but is now considered to be part of the debris flows (Lefebvre et al. 2003).

The brecciated unit(s) could also be a single or multiple diatreme(s) localized within Lalibert, Leclaire, Menzies and Musquash townships. A tectonic repetition of the diamond-bearing diatreme(s) has been achieved through the regional deformation events described previously. This regional deformation may help to explain the layered appearance observed at some of the diamond occurrences (J. Ayer, Ontario Geological Survey, personal communication 2006).

Lamprophyre occurs as narrow dike-like intrusions or in bodies of indeterminate morphology cross-cutting or intercalated with local country rocks. Dikes range in width from 50 cm to 2 m and may display 2-3 cm offshoots in some locations. Contacts with the host rocks vary from sharp and straight to highly irregular. No variation in grain size within the dikes has been observed and the no variation in colour or mineralogy

has been observed within country rock adjacent to the dikes. The lamprophyre is fine-grained, grey in colour and contains approximately 5-10% subrounded to subangular fragments. The fragment population is dominated by actinolite-rich monomineralic rocks, or by biotite-rich greenstone and hornblende-rich ultramafic rocks (Lefebvre 2004).

Lamprophyre is differentiated from the breccia by: a lower clast content; a predominance of highly altered, coarse-grained actinolite fragments; scarcity of wall rock fragments; the rounded shape of the xenoliths; and the presence of a weaker fabric (Lefebvre 2004).

The lamprophyre dikes post-date all other lithologies. The lamprophyre with indeterminate morphology predates most of the host lithologies since fragments of each are found within these lamprophyres. However, in some locations, lamprophyre fragments have been observed within intermediate and mafic intrusive rocks near the contact with lamprophyre (Lefebvre 2004).

Fragments within lamprophyre dikes commonly have a biotite-rich rim enclosing the xenoliths and the fragments found in the lamprophyres of indeterminate morphology do not. The lamprophyres of indeterminate morphology also show positive relief of less severely weathered xenoliths more so than the lamprophyre dikes. Lastly, the lamprophyre dikes have a less variable fragment lithology (Lefebvre 2004).

The lamprophyre is petrographically distinct from the breccia. It contains a lower abundance of clasts and fewer clast types. Unlike the breccia, the lamprophyre contains no juvenile magmatic material and oscillatory-zoned hornblende grains are rare. Detailed petrographic descriptions of the lamprophyre can be found in Lefebvre (2004).

Relationship between diamond content and lithology

Microdiamonds have been recovered from a wide array of breccias and lamprophyres in the Wawa area. Between 2003 and 2004, the Ontario Geological Survey (OGS) investigated a number of diamond occurrences during a mapping program conducted in Menzies and Musquash townships. Closer look at locations where bulk samples had been collected suggested that the bulk samples probably included more than one rock type. Diamonds were recovered from each of these bulk sample sites, but it would be difficult to establish from which of the rock types the diamonds were recovered. The OGS undertook a limited sampling program to further investigate the diamond content of specific lithologies (Vaillancourt et al. 2005a).

The OGS collected three small samples (95.21 kg total weight) from three different lithological units at the Cristal and Genesis diamond occurrences. These occurrences are on the Festival Diamond Property owned by Pele Mountain Resources Inc. Results from this limited sampling program reinforce the observation that microdiamonds are not restricted to a single unit. Microdiamonds were recovered from heterolithic breccia, both with and without ultramafic magma pockets, and from a fragment-free ultramafic dike. There is the possibility, however, that the diamonds recovered from the ultramafic dike are xenocrysts derived from diamond-bearing host breccia (Vaillancourt et al. 2005a).

The OGS concluded that the results from only three samples are not sufficient to draw irrefutable conclusions regarding the location of the microdiamonds. Collection and analysis of more, well constrained samples is necessary to further refine the diamond potential of specific host rocks.

In 2005, Spider Resources Inc. and KWG Resources Inc. conducted a similar bedrock sampling program to investigate the diamond-bearing potential of specific lithologies. Separate representative samples (approximate weight 16 kg each) of the matrix, xenoliths and run of mill (ROM) portions of the bedrock were collected from the Wawa Diamond Project and were sent for caustic dissolution. The results are as follows: matrix sample returned 67 diamonds (0.008 total ct), xenolith sample returned 244 diamonds (0.051 total ct) and the ROM sample returned 86 diamonds (0.006 total ct) (Spider Resources Inc., press release, February 20, 2006).

Geochronology

In conjunction with recent geological mapping in the area, the Ontario Geological Survey has been conducting geochronological work to help understand the nature and timing of the diamondiferous units and their host rocks within the Michipicoten greenstone belt. A felsic volcanic horizon hosting diamondiferous units returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2701.4 ± 2.1 Ma. Maximum $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 2685.1 ± 1.0 Ma and 2684.9 ± 1.4 Ma have been returned for diamondiferous lamprophyre dikes cutting the Catfish assemblage (2.7 Ga) intermediate to felsic metavolcanic rocks in Lalibert and Menzies townships (Ayer et al. 2003). A second sample of felsic lapilli tuff, part of the Catfish assemblage, adjacent to the Moet Occurrence contains zircons that returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2698.7 ± 1.1 Ma (Vaillancourt et al. 2004).

A sample was collected from the diamondiferous breccia at the Moet Occurrence in order to determine the age of brecciation. Five zircons were analyzed. The three oldest ages are 2687 ± 2 Ma, 2683 ± 2 Ma and 2681 ± 2 Ma. The two youngest zircons gave results that precisely overlapped one another at 2679.2 ± 2.1 Ma (Vaillancourt et al. 2004). Since this is the youngest zircon age obtained from the breccia, it represents either the time of crystallization or emplacement of the body if the zircons are magmatic, or a maximum time of emplacement if the zircons are xenocrystic (Vaillancourt et al. 2005a). If the zircons are xenocrystic, this age must still be close to the time of breccia emplacement since a lamprophyre dike from the GQ property returned a date of 2673 ± 8 Ma from titanite (R. P. Sage, Ontario Geological Survey, unpublished 2000).

The data indicate that the felsic metavolcanic rocks hosting the diamondiferous breccias are part of the Catfish assemblage. The maximum age for the diamondiferous breccias and the associated dikes is less than 2680 Ma. These absolute age constraints indicate that the breccias are not volcanoclastic units belonging to the Catfish assemblage (Vaillancourt et al. 2005a).

Zircons from a sample of felsic lapilli tuff from northwestern Menzies township returned an age of 2736.0 ± 0.8 Ma which is taken to represent the age of eruption and crystallization of the tuff. This age clearly indicates that the volcanic package underlying the iron formation in the western part of Menzies Township is part of the Wawa assemblage (2.75 Ga) and brackets the uppermost part of the assemblage at 2736 Ma (Vaillancourt et al. 2005a).

Geochemistry of the Diamond-bearing Rocks

Both Williams (2002) and Lefebvre (2004) conclude that the whole rock major element geochemistry is consistent with a calc-alkaline classification for both the lamprophyres and the associated breccia. Both authors also noted that the compositions of chromite in the Wawa metavolcanic rocks are in the range typical for lamprophyres and dissimilar to those in kimberlites and lamproites.

Whole rock geochemistry for the diamond-bearing rocks is tabulated in Sage (2000), Williams (2002), Lefebvre (2004), Stone and Semenyna (2004) and Vaillancourt et al. (2005c). Sage and Williams' work is specific to the diamond-bearing and non diamond-bearing lamprophyres. Work by the other authors relates to both the breccia and the lamprophyres. Whole rock geochemistry for kimberlites of the Wawa area can be found in Kaminsky et al. (2002).

Diamond characteristics

Lefebvre (2004) undertook a study examining a parcel of 80 macrodiamonds recovered from the volcanoclastic breccia on the GQ Property. Results from this work are summarized below and in De Stefano et al. (2006). Additional work on the morphology of the Wawa diamonds can be found in Stone and Semenyna (2004). Stachel et al. (2004) summarize results of analysis conducted on diamonds from the Genesis and Cristal diamond occurrences presently held by Pele Mountain Resources Inc.

Shape: The diamonds display a highly variable primary growth form. Most of the diamonds are either octahedral aggregates (44% of the population) or single octahedral crystals (26%). Single cubic and cubic-octahedral crystals and their aggregates, as well as macles form the remainder of the population. Forty-eight percent of the diamond population is single crystals and only 28 diamonds could be evaluated for crystal regularity. The majority of the diamonds also display some degree of distortion.

Colour and transparency: The diamonds included in this study are classified into colourless, brown, grey, black, yellow and white. No pink, green, violet or blue diamonds were observed. The colour distribution within the population is: colourless (48%), heterogeneous (24%), yellow (11%), black (3%), brown (10%) and grey (3%). The heterogeneity in colour is observed only in aggregates. The diamond population consists of 48% transparent crystals, 25% translucent crystals, 14% opaque crystals and 14% is a combination of opaque and translucent crystals. Transparent crystals are typically colourless and also comprise a few yellow octahedral single crystal and coarse aggregates as well as macles. Translucent crystals comprise all possible primary crystal forms and colours. Opaque crystals are mostly fine-grained aggregates which have black body colouring.

Resorption: Generally speaking, the diamonds have experienced low degrees of resorption. Only 21% of the diamond population displays extensive resorption. Some crystals (14%) exhibit non-uniform resorption where one part of the crystal is more strongly resorbed than another.

Inclusions: Mineral inclusions were identified in 58% of the diamonds. Both primary and secondary inclusions were observed. The mineralogy of the recovered primary inclusions is listed in descending order of abundance: olivine (Fo₉₂ and Fo₈₉), clinopyroxene (omphacite), plagioclase (albite and An-rich), orthopyroxene (En₉₃) and Fe-Ni sulphide (pentlandite).

Cathodoluminescence: The relative abundances of cathodoluminescence (CL) colours for the Wawa diamonds are: orange-red (46%), yellow (28%), orange-green (10%), green (6%), and other non-uniform colours (10%). None of the 69 diamonds examined displayed the more common blue CL.

Impurities: Fournier Transform Infrared (FTR) spectrometry was used to investigate the nitrogen and aggregation states for 41 diamonds. The majority of the diamonds have low nitrogen contents, < 300 ppm. The diamonds show two modes of nitrogen aggregation suggesting mantle storage at 1100 - 1170° C.

Diamonds from the Genesis occurrence are almost exclusively cubes including some fragmented, twinned and moderately resorbed cubes. Most of the crystals contain clouds. Fully transparent stones are dominantly brown although colourless stones also are common; one diamond was yellow in colour. Nitrogen concentrations range from below detection (<10 ppm) to 600 atomic ppm. Nitrogen aggregation is very low (Stachel et al. 2004).

Diamonds from the Cristal occurrence range from un-resorbed octahedra to highly resorbed dodecahedra. Octahedral and weakly resorbed octahedral stones dominate the population. About 25% of the population are irregular crystals, macles are common (15%) and about 5% of the diamonds show cubo-octahedral growth. The stones fall into two dominant colour classifications, colourless and a range of brown colouration. Nitrogen contents range from <10 – 560 ppm, but with only one exception nitrogen is ≤170 ppm. Nitrogen aggregation varies between 0 and 97% B-centre. Olivine is the most common mineral inclusion, followed by pyrope garnet and Mg-chromite (Stachel et al. 2004).

Origin of the Diamond Deposits

Based on published data on the diamond-bearing rocks at Wawa and Cobalt, Wyman et al. (in press) suggest that the tectonic setting of the deposits and nature of the host rocks indicate that the diamonds may be derived from the asthenospheric wedge and subducted slab at shallow depths (100 – 160 km) rather than the deep keels of Archean cratons associated with traditional diamond deposit types. Models of low-temperature Phanerozoic diamond formation in active subduction zones, or rapid uplift and emplacement of peridotite massif occurrences, can be adapted to the Archean deposits. The stability field of diamonds in

most Phanerozoic scenarios may be too deep to be accessed by the lamprophyric magmas. Shallow subduction, as proposed for these occurrences of adakitic-type rocks in the Wawa subprovince, could generate two different diamond stability windows at sufficiently shallow depths to account for their presence in lamprophyric magmas.

Wyman et al. (in press) states that any tectonic model for these Archean diamond occurrences must address several requirements. These requirements include

1. a deep source for oxidized metasomatic fluids that is activated prior to lamprophyre emplacement
2. a mechanism to isolate this isotopically aged and depleted source for tens or hundreds of millions of years until it is heated in the mantle during orogeny
3. a hybridized mantle source for primitive, hydrous, shoshonitic lamprophyres
4. sustained cold finger effect in the mantle to establish a shallow-mantle diamond stability window

Two theories of diamond origin are postulated by De Stefano et al. (2006). Both a cratonic and orogenic model of diamond formation are discussed in an effort to rationalize the observed diamond characteristics. The authors conclude that neither model fully explains all of the observed characteristics.

Field Trip Stops

Field Trip Road Log

Stop	Locality	km
	Intersection of Hwy 101 and Hwy 17	0
	Take Highway 17 north	
	Catfish Road forestry road – turn east	19.2
1	GQ Diamond Discovery	23.3
2	Northern Sierra Minerals Corporation Area B	23.9
3	Northern Sierra Minerals Corporation Engagement Zone	29.1
	Return to Highway 17 <i>reset odometer</i>	0
	Drive north to access road – turn east	6.2
	Park	6.6
	Walk eastward along trail	7.6
4	Moet Occurrence, Festival Property	7.6
	Return to Highway 17 <i>reset odometer</i>	0
	Drive north, park on shoulder of highway	4.3
5	Sandor Diamond Occurrence	4.3
	Continue north on Highway 17 to intersection of Highway 519, turn right	12.6
	Safely turn in parking area and return to Intersection of Hwy 17 and 519 <i>reset odometer</i>	0
	Drive south on Highway 17, park on shoulder of road	3.1
6	Dubreuilville Dike	3.1
	Continue south on Highway 17	44.9
	Turn left into access road to gravel pit, park	
	Walk south approximately 150 m	
7	Monchiquite Dike	45
	Drive north on Hwy 17 to Wawa	53.2
8	Wawa Motor Hotel	

STOP 1 - GQ Diamond Discovery Site

Northern Sierra Minerals Corporation Area A

UTM co-ordinates – 0665570E 5333291N NAD83

Several outcrops of diamondiferous breccia outcrop on the west side of a forestry road. This exposure is an example of the hypabyssal facies of the three identified diamond-bearing units. The rock cut displays the apparently conformable nature of these “lamprophyre” dikes. The most notable features of these outcrops are the actinolite-rich nature of the matrix and the presence of biotite-rich reaction rims around the xenoliths. It is frequently difficult to distinguish these dikes from the mafic to intermediate agglomeratic and tuffaceous host rocks. In the vicinity of the discovery area, located at the south end of Area A, the diamondiferous breccias are arrayed linearly along the logging road where the topography indicates a 5 – 10 m thick, northwest-trending dike (Cavey 2002).

A compilation of the geology of the GQ Property is shown in figure 4. Figure 5 provides a compilation of the diamond occurrences on the GQ Property.

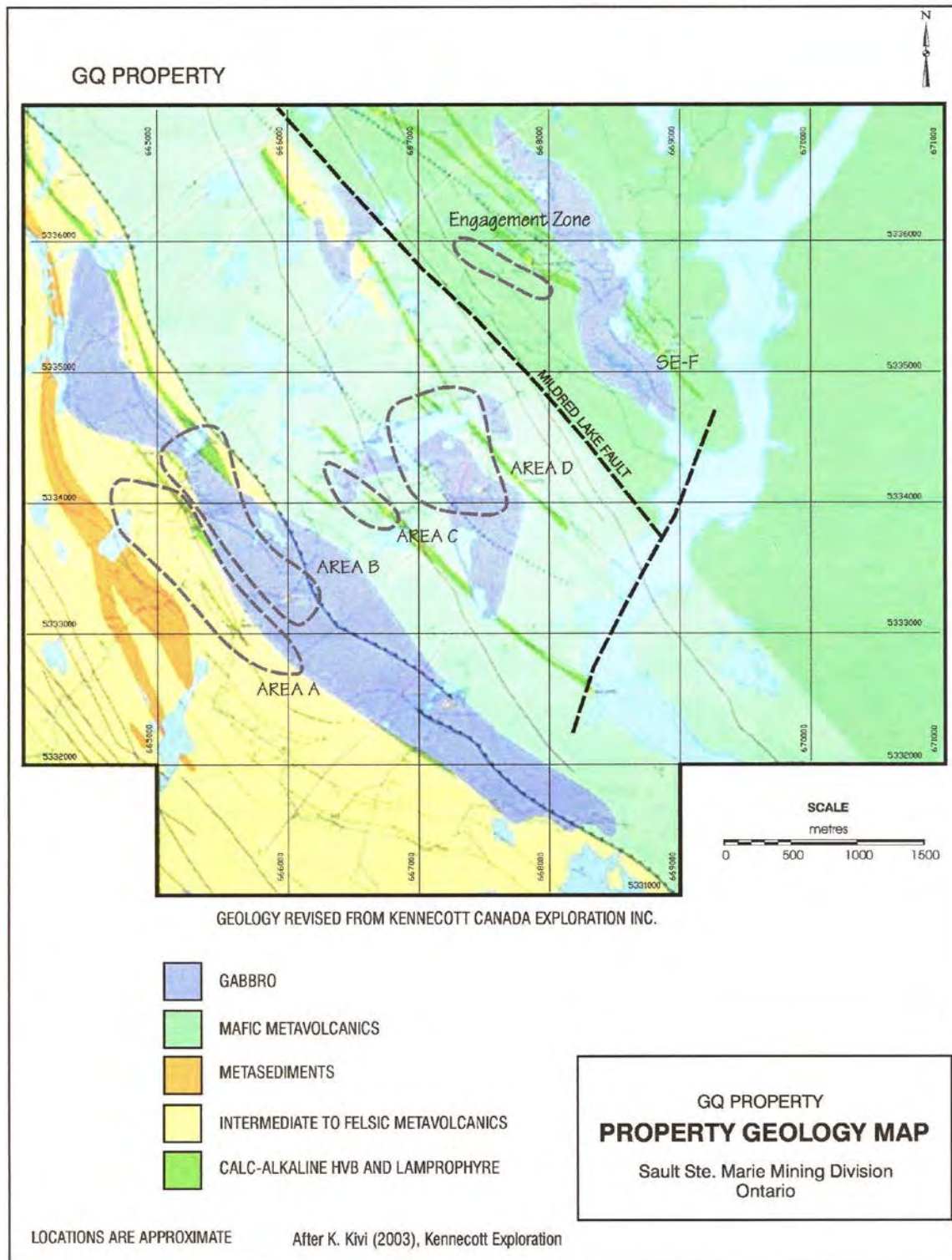


Figure 4. Geological compilation map of the GQ Property, Musquash Township, Northern Sierra Minerals Corporation (Cavey 2004).

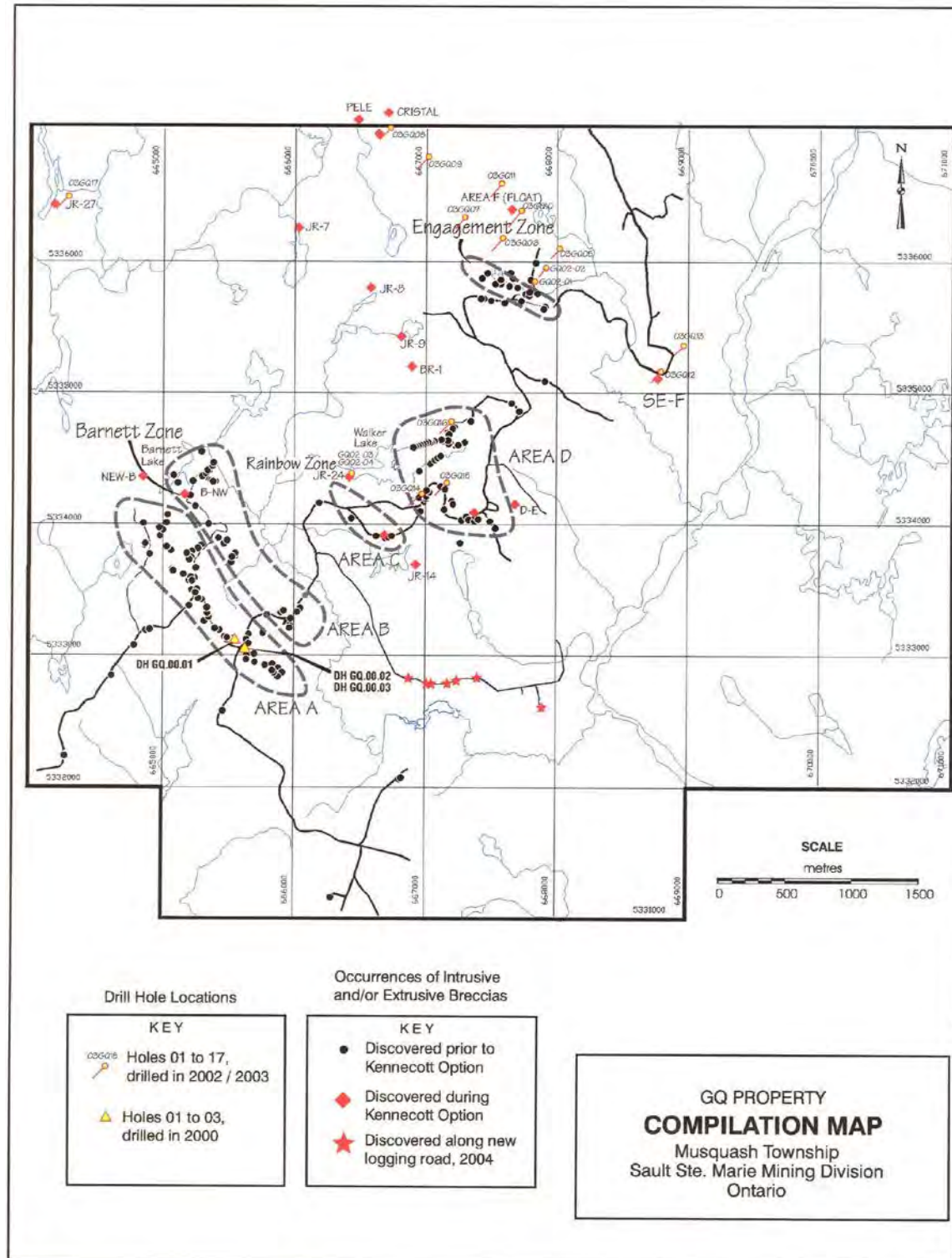


Figure 5. Occurrences of diamondiferous bedrock on the GQ Property, Musquash Township, Northern Sierra Minerals Corporation (Cavey 2004).

A relatively precise $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2674 ± 8 Ma was returned from a sample collected in 2000 by R. P. Sage (R. P. Sage, Ontario Geological Survey, unpublished 2000). The sample analysed was titanite. The titanite grains are considered to be a primary mineral and not xenocrystic.

In thin section, the rock is characterized by a green, medium-grained, granoblastic to decussate groundmass of actinolite, biotite, chlorite, plagioclase and accessory minerals. Amphibole or biotite macrocrysts up to 1 mm are common (Stone and Semenyina 2004).

Local prospectors T. Nicholson, J. Robert and M. Tremblay made the discovery in the fall of 1999. The first two bedrock samples (63.4 kg and 70.5 kg) collected were processed by Kennecott Canada Exploration Inc. in their Thunder Bay laboratory. According to Kennecott's report, the 63.4 kg sample yielded 45 diamonds, of which 10 were macro diamonds and 35 were microdiamonds. One of the macro diamonds measured 1.01 mm in one dimension. The 70.5 kg sample yielded 9 microdiamonds. All stones were white in colour and transparent in clarity.

Duplicate samples were collected by Band-Ore Resources Ltd. in early 2000 and were processed at SGS Lakefield Research Limited. A 54.6 kg sample yielded 98 microdiamonds. A confirmation sample from the same area yielded 98 microdiamonds from a 54.6 kg sample.

In 2000, Band-Ore Resources drilled 3 short holes (75 m total) at the discovery site. Table 1 details the diamond recovery results from the drill program. Only partial intervals from drill hole DDH GQ-00-3 were submitted for microdiamond recovery since portions of the core were used for thin sections, microprobe analysis and display purposes. In total 5 diamonds were recovered from DDH GQ-00-3, including one champagne coloured macro diamond and one white microdiamond from a sample weighing 7.5 kg. Drill sections for DD GQ-00-01 through 03 are shown on figure 6.

Table 1. Diamond recovery results from 2000 Band-Ore Resources Ltd. diamond-drilling program

Drill Hole	Sample No.	Sample Size (kg)	No. Macro Diamonds	No. Micro Diamonds
DDH GQ-00-1	Sample 1A	63.35		1
	Sample 1B	30.17		10
	Sample 1C	28.51		6
DDH GQ-00-2	Sample 2A	37.52	1	30
	Sample 2B	30.97		434
	Sample 2C	28.51		30

results compiled from Band-Ore Resources Ltd. press releases 2000

To date, the discovery site has yielded 746 diamonds, including 15 macro diamonds, from sample material weighing 785 kg. The largest diamond recovered exceeds 1.0 mm in size and the majority of the stones are gem quality, white, clear and transparent.

STOP 2 - Northern Sierra Minerals Corporation Area B

UTM co-ordinates – 0665425E 5334748N NAD83

The exposure on the east side of the forestry road provides an excellent exposure of the subvolcanic/intrusive breccia facies. Subrounded to rounded xenoliths dominate the vertical exposure (Figure 7). Field relationships between the intrusive breccia and other heterolithic breccias can be observed in several outcrops along the road.

Texturally, the subvolcanic/intrusive facies may resemble both the hypabyssal facies and the intrusive heterolithic breccias. The facies consists of mica and amphibole phenocrysts (<2mm) in a groundmass of mica, actinolitic amphibole and lesser albite, carbonate, sphene and oxides. Alteration includes variably chloritized mica while the other phenocrysts have been extensively altered to varying proportions of mica, albite and actinolitic amphibole.

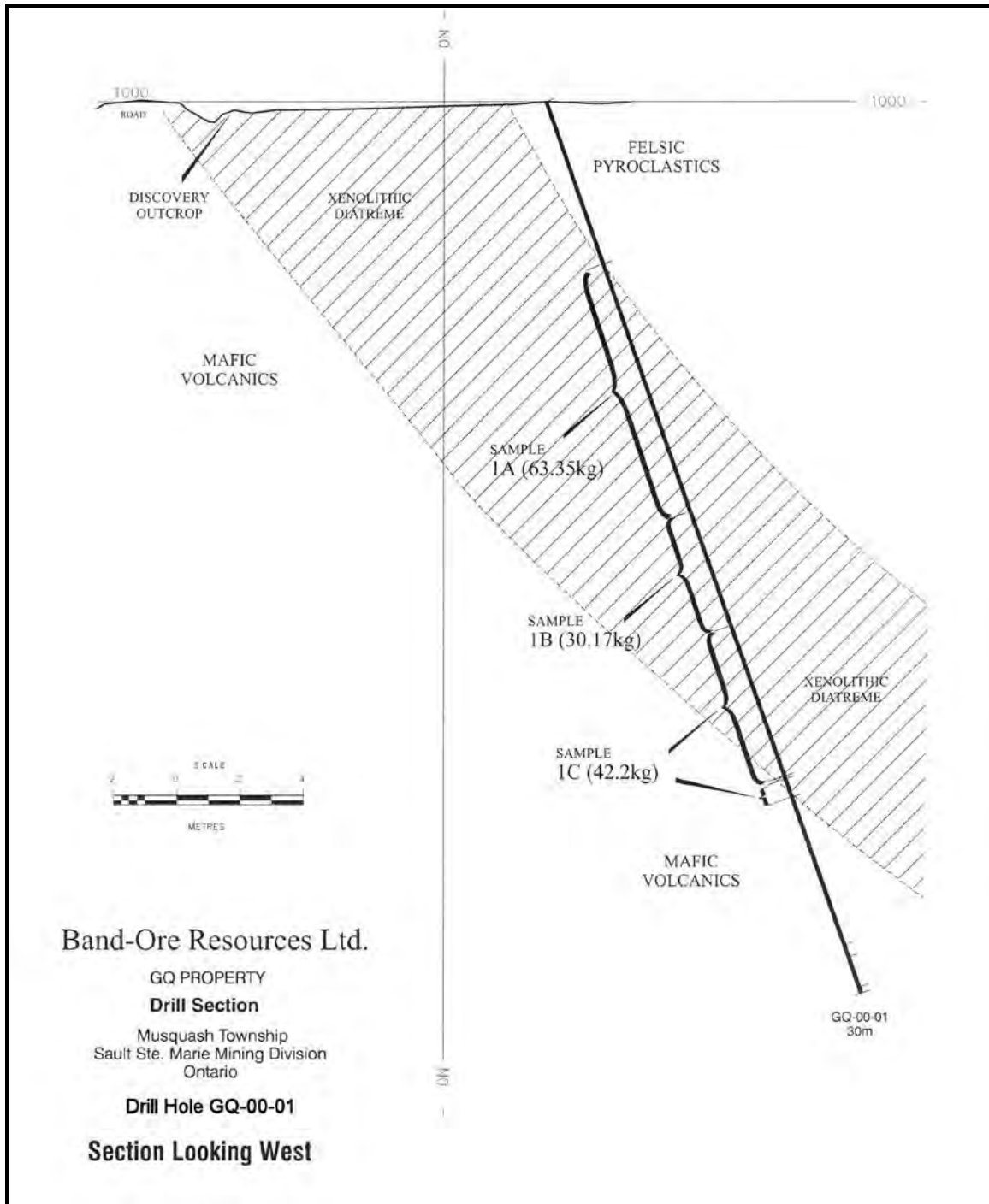


Figure 6. Drill hole sections GQ-00-01, GQ-00-02 and GQ-00-03, GQ Property, Musquash Township, Northern Sierra Minerals Corporation (Cavey 2002).

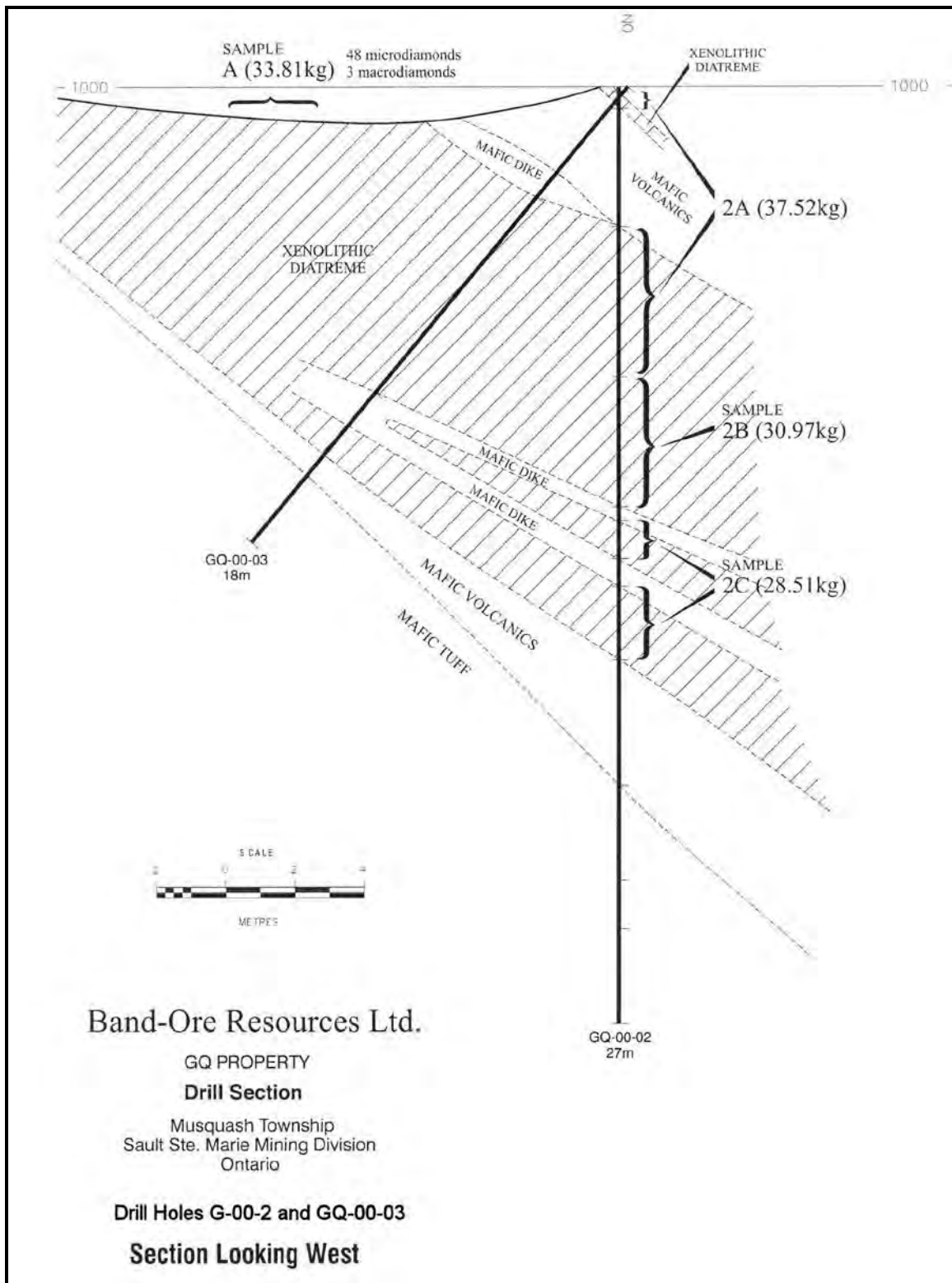


Figure 6 cont'd. Drill hole sections GQ-00-01, GQ-00-02 and GQ-00-03, GQ Property, Musquash Township, Northern Sierra Minerals Corporation (Cavey 2002).



Figure 7. Stop 2 - Northern Sierra Minerals Corporation Area B – rounded lower crustal to upper mantle xenoliths

Stone and Semenyna (2004) completed a petrographic examination of one of the ultramafic xenoliths from this site. The sample was dominated by a coarse, radiating to decussate, clear to pale green amphibole of tremolitic to magnesium-rich actinolite composition. Carbonate occurs locally and biotite is concentrated at the rims of the xenolith.

Band-Ore Resources Ltd. discovered area B in 2000. Thirty-three (33) reconnaissance samples were collected from this area and a total of 273 diamonds was recovered from 352 kg of material (Cavey 2002). One 24 kg sample returned 126 microdiamonds (1.37 mg total weight). No macro diamonds were recovered from Area B. Band-Ore Resources Ltd. completed only a reconnaissance sampling program, minor stripping and trenching in this area.

The Barnett Zone lies approximately 1.6 km to the northwest. It was discovered in the fall of 2001 by Kennecott Canada Exploration Inc. who completed a limited program of mechanical stripping, channel sampling and washing of outcrops over the area. A total of 27 outcrop channel samples (270 kg) were collected. The channel samples returned 330 microdiamonds and 3 macro diamonds. A single 24 kg sample of heterolithic breccia from the Barnett Zone returned 3 macrodiamonds and 123 microdiamonds.

A total of 273 diamonds (261 microdiamonds and 12 macrodiamonds) were recovered from 34 samples (352 kg) collected between September 2000 and July 2001 by Kennecott Canada Exploration Inc. in Area B (including the Barnett Zone).

STOP 3 - Northern Sierra Minerals Corporation Engagement Zone
Bulk Sample Site
UTM co-ordinates – 0667760E 5336073N NAD83

The main outcrop exposure consists of medium to thickly bedded pyroclastic air-fall deposits and debris flows described as heterolithic tuff-breccias that grade upwards to lapilli-tuff and tuff. The groundmass of the extrusive phase is fine-grained with variable proportions of relatively small (<2mm) altered phenocrysts, including chloritized mica. The groundmass consists primarily of actinolitic amphibole with rare to minor mica or granular albite. The pyroclastic rocks are mineralogically and compositionally similar to intrusive varieties, but have a significantly higher proportion of mica phenocrysts.

Both diamond-bearing intrusive and extrusive rocks host significant proportions of fragments derived from the local country rock. Rare to minor, deep crustal and upper mantle xenoliths, such as banded gneiss and extensively altered talcose ultramafic xenoliths are present. Fresh mantle rocks, such as lhertzolite, harzburgite and eclogite have not been identified.

The matrix material is typically fine-grained, green, weakly foliated actinolite schist. Albite is present, although it is less abundant (<10%) than in the lamprophyre dikes and implies a more ultramafic composition for the breccia matrix than for the lamprophyre dikes. Titanite is fairly abundant and calcite, epidote, apatite and sulphide minerals occur locally. Macrocrysts of actinolite are commonly observed. Rare macrocrysts of amphibole also are observed and are frequently altered to actinolite. The actinolite macrocrysts are probably metamorphic in origin, whereas the amphibole macrocrysts may represent accidental or cognate crystals (Stone and Semenyina 2004).

The Engagement Zone has a minimum strike length of 335 m and a horizontal width in excess of 75m. The zone strikes northwesterly and has a shallow dip to the northeast. This zone may represent the southeast extension of Pele Mountain Resources Inc.'s Cristal diamond occurrence located approximately 2 km to the northwest. In 2003, a 0.72 carat macrodiamond was recovered from a bulk sample collected from the Cristal. The geology of the Engagement Zone and sample locations are shown in Figure 8. A simplified cross section of the Engagement Zone is shown in Figure 9.

Band-Ore Resources Ltd. discovered this zone in January 2001. A 16 kg sample from a single angular boulder of diatreme breccia returned 128 microdiamonds. Four subsequent samples (96 kg total weight) returned 5045 microdiamonds and 65 macrodiamonds.

In 2001 a mini-bulk sample weighing 12.5 tonnes was collected under the supervision of Kennecott Canada Exploration Inc. and shipped to the Saskatchewan Research Council. The largest diamond recovered from this sample was a 0.254 carat, broken, white octahedral stone. Two additional bulk samples were collected in 2003. A 22 tonne sample tested an area where 6 channel samples (63 kg total) recovered 1752 stones. A 20 tonne sample tested an area where 5 channel samples weighing 41.8 kg returned 552 stones. The results from these bulk samples are shown in Table 2.

Table 2. Diamond results from Engagement Zone bulk samples (2001)

Occurrence	Sample Weight	Sieve +1mm	Sieve +2mm	Sieve +3mm	Sieve +5mm	Sieve +6mm	Total Diamonds	Total Carat Weight
Engagement Zone East	22.1 tonnes	1	4	4	2	1	12	0.375
Engagement Zone West	20.4 tonnes	2	3	3			8	0.155

results compiled from Band-Ore Resources Ltd. press releases 2004

To date, exploration on the Engagement Zone has included an orientation geochemical survey, geological mapping, channel sampling, trenching, bulk sampling and a 9-hole (1775 m) diamond drilling program designed to test the strike continuation of the zone. The drilling program demonstrated that thick diamondiferous breccia deposits can feather out and thin to a few centimeters thickness (Cavey 2003).

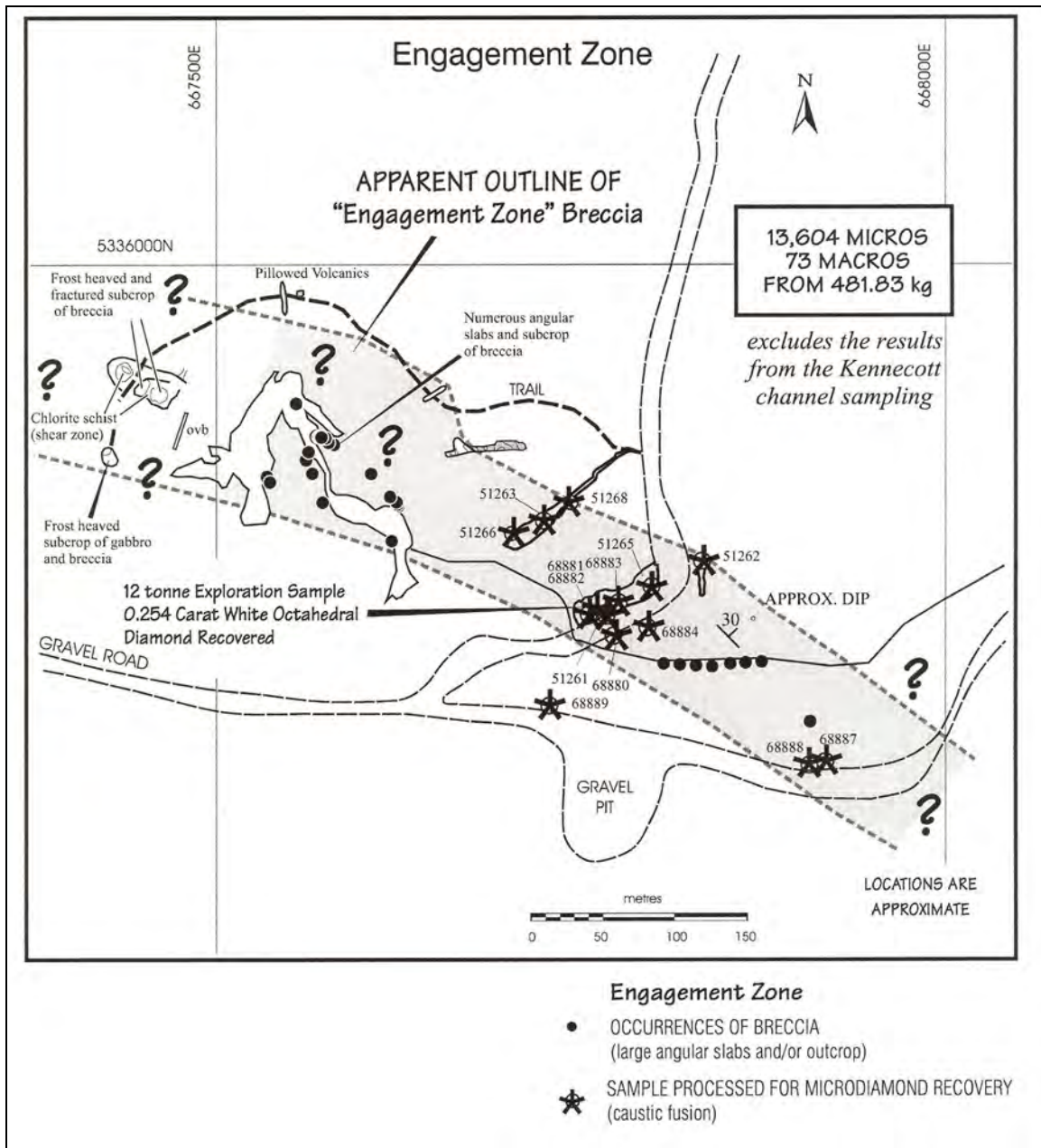


Figure 8. Stop 5 – Geology and sample locations at the Engagement Zone, GQ Property, Northern Sierra Minerals Corporation (Cavey 2002)

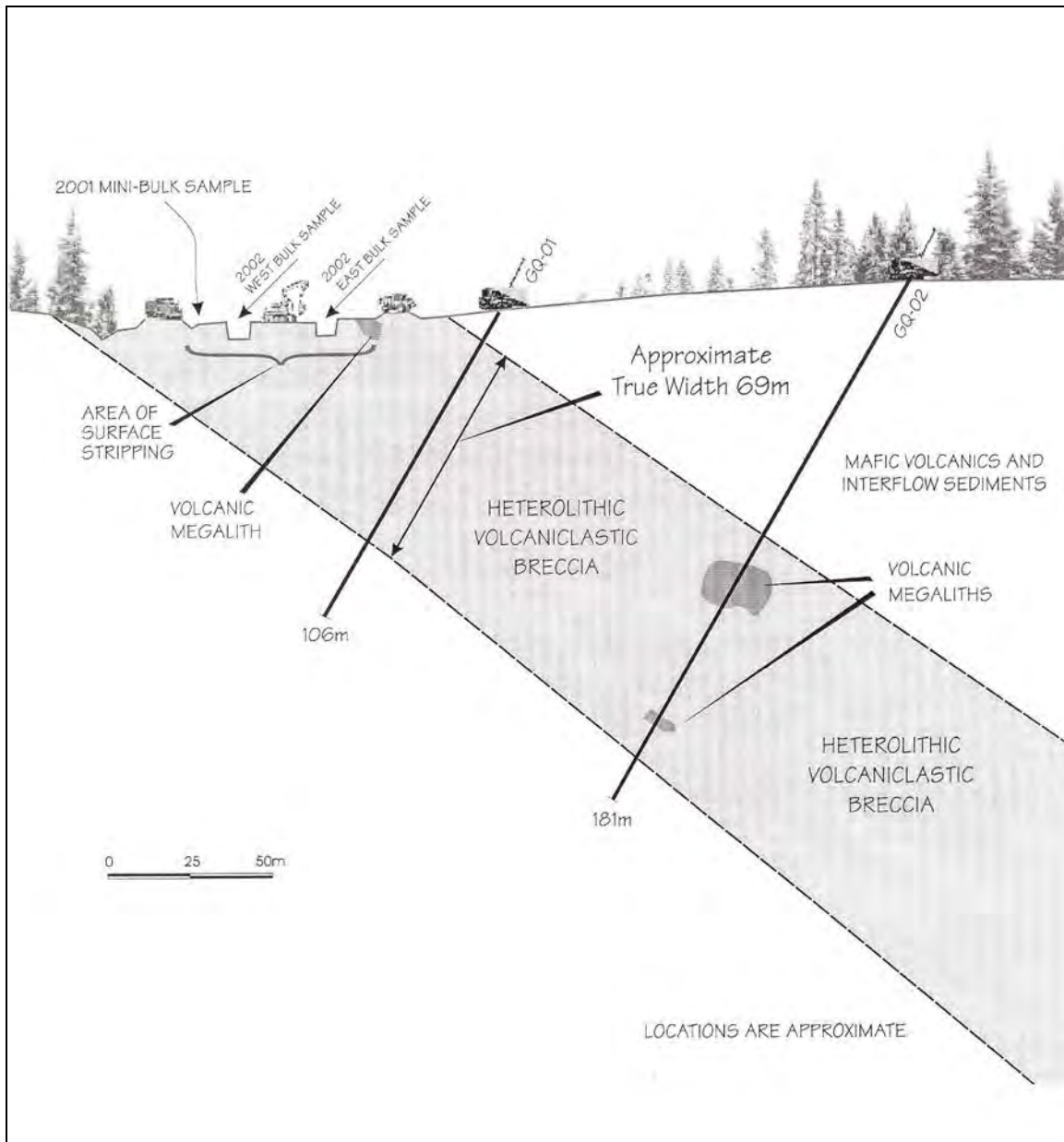


Figure 9. Simplified cross section through the Engagement Zone looking northwest (Cavey 2004).

STOP 4: - Moet Occurrence, Festival Property

Pele Mountain Resources Inc.

UTM co-ordinates – 0662709E 5338009N NAD83

The Moet occurrence is a large stripped outcrop that extends in a north-south orientation across the forestry road. The outcrop displays all three facies of diamond-bearing bedrock exposed over an area 500 m by 300 m. It is hosted within fine- to medium-grained mafic metavolcanic rocks with closely associated intermediate to felsic metavolcanic rocks and metasediments. The volcanic facies was found concentrated within a series of outcrop exposures along the west side of a 5-8 m north-trending ridge and the subvolcanic/intrusive breccias and hypabyssal rocks are present in several outcrops east of the volcanic facies. The subvolcanic/intrusive breccias are hosted in the metavolcanics and the volcanic facies appears to overlie these rocks, and are in turn overlain by intermediate to felsic metavolcanic rocks. The fragments

within the volcanic facies consist of country rock fragments with lesser crustal fragments and lower crust to upper mantle xenoliths. The hypabyssal facies displays primarily mantle xenoliths (Walker 2003). A detailed geological map of the Moet locality is shown in Figure 10.

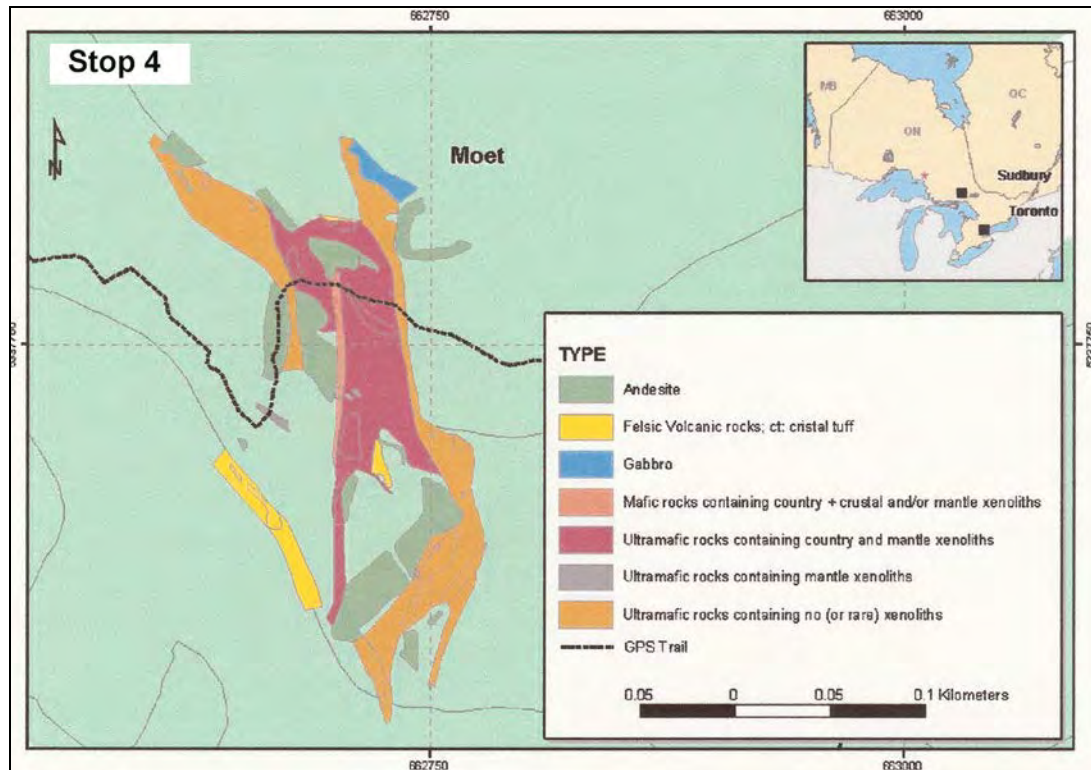


Figure 10. Detailed geological map of the Moet Occurrence, Festival Property, Pele Mountain Resources Inc. (Kjarsgaard et al. 2003).

The access trail to the outcrop passes through a sequence of intermediate to felsic tuffs and tuffaceous breccias. The composition of the fragments and that of the matrix are highly variable from felsic to intermediate and a combination of felsic matrix with intermediate fragments and vice versa is not uncommon. At a distance of approximately 1km, the trail passes through a sequence of mafic to intermediate tuffs and lapilli tuffs (Vaillancourt et al. 2005b).

An age date of 2698 ± 1 Ma was returned from sample of the felsic metavolcanic rocks at the occurrence. A sample of the breccia returned an age date of 2680 Ma (J. Ayer, Ontario Geological Survey, personal communication, 2004).

Discovered by Pele Mountain Resources Inc. in 2001, the occurrence initially gained interest because the breccia has a size distribution of diamonds that includes coarser sized diamonds from relatively small samples. For example, an 8 kg sample collected in 2001 recovered a total of 9 diamonds, 4 of which were in the +600 mesh fraction. Diamonds are consistently recovered from both the volcanic and subvolcanic facies at this showing. A summary of the results of the 2001 and 2002 sampling of the occurrence by Pele Mountain Resources Inc. is found in Table 3.

Table 3. Summary of the diamond results from the 2001-02 sampling of the Moet Occurrence

Facies	Weight (kg)	<425 mesh	>425 mesh	total diamonds
volcanic	117	48	0	48
volcanic	16	2	1	3
hypabyssal	24	0	0	0
hypabyssal	8.6	0	0	0
volcanic	9.3	7	0	7
volcanic	14.1	12	5	17
hypabyssal	9.3	0	0	0
subvolcanic breccia	32	30	1	31

results compiled from Pele Mountain Resources Inc. press releases 2002

Further exploration was conducted at the occurrence in 2003 when DeBeers Canada Exploration Inc. completed a detailed airborne geophysical survey, stripping, mapping and sampling program. A 47.8-tonne bulk sample from the site returned 5 diamonds with a total carat weight of 0.13. All diamonds were recovered from the +1 to +3 sieve class screens (Pele Mountain Resources Inc., press release, March 17, 2004).

STOP 5: - Sandor Diamond Occurrence

Spider Resources Inc. & KWG Resources Inc.

UTM co-ordinates – 0659805E 5342191N NAD83

The Sandor occurrence is the first confirmed occurrence of diamonds in bedrock in the Wawa area. The occurrence is located in a 4 m high road cut on the east side of the Trans Canada Highway (Highway 17). The dike is approximately 5 m wide, steeply dipping and strikes roughly parallel to the regional schistosity at 120°. The dark, greenish-grey rock weathers olive grey, is highly fractured, moderately carbonatized and is non-magnetic. It is composed of up to 40% actinolite replaced mantle xenoliths and supracrustal xenoliths. Towards the margins of the dike xenoliths are less common and the rock grades into an adjacent micaceous dike. Only remnants of the dike remain in situ. The dike is hosted by gabbros and intermediate to felsic crystal tuffs. A (308.6 kg) sample of the dike, collected by Spider Resources Inc. in 1997, returned a total of 97 diamonds comprising 1 commercial stone, 13 macrodiamonds and 83 microdiamonds.

A short walk into the forest from the top of the outcrop leads to a larger stripped area where field relationships between the host gabbro and the dike can be observed. A second, xenolith-bearing dike (occurrence LAL-3) is located at the north end of the outcrop. This dike is 2 m wide and closely resembles the Sandor occurrence. A 34.6 kg sample of this dike contained 1 microdiamond. A detailed geological map of the Sandor occurrence is shown in Figure 11. A compilation map of the geology of the Spider Resources Inc. and KWG Resources Inc. property is shown in Figure 12.

Using normative mineralogy, Sage (2000) concluded that this dike should be classified as a spessartite. A spessartite is defined as a lamprophyre composed of phenocrysts of green hornblende or clinopyroxene in a groundmass of sodic plagioclase with accessory olivine, biotite, apatite and opaque oxides.

Titanite and rutile from the matrix of the Sandor dike returned an age of 2703 ± 42 Ma (Sage 2000). The date is interpreted to be a minimum age of intrusion. Subsequent dating of a zircon from a gneissic xenolith from the Sandor dike returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2684.9 ± 1.4 Ma (Ketchum, Kamo and Davis 2003).

Spider Resources Inc. and joint venture partner KWG Resources Inc. have taken 3 mini-bulk samples along the 1 km strike extent of the dike since the fall of 2001. The three bulk samples had a combined weight of 7.61 tonnes and returned 11 commercial stones and 9 macro diamonds. These samples were tested only for macro diamond and commercial diamond content (Spider Resources Inc., *press releases*, February and March 2002).

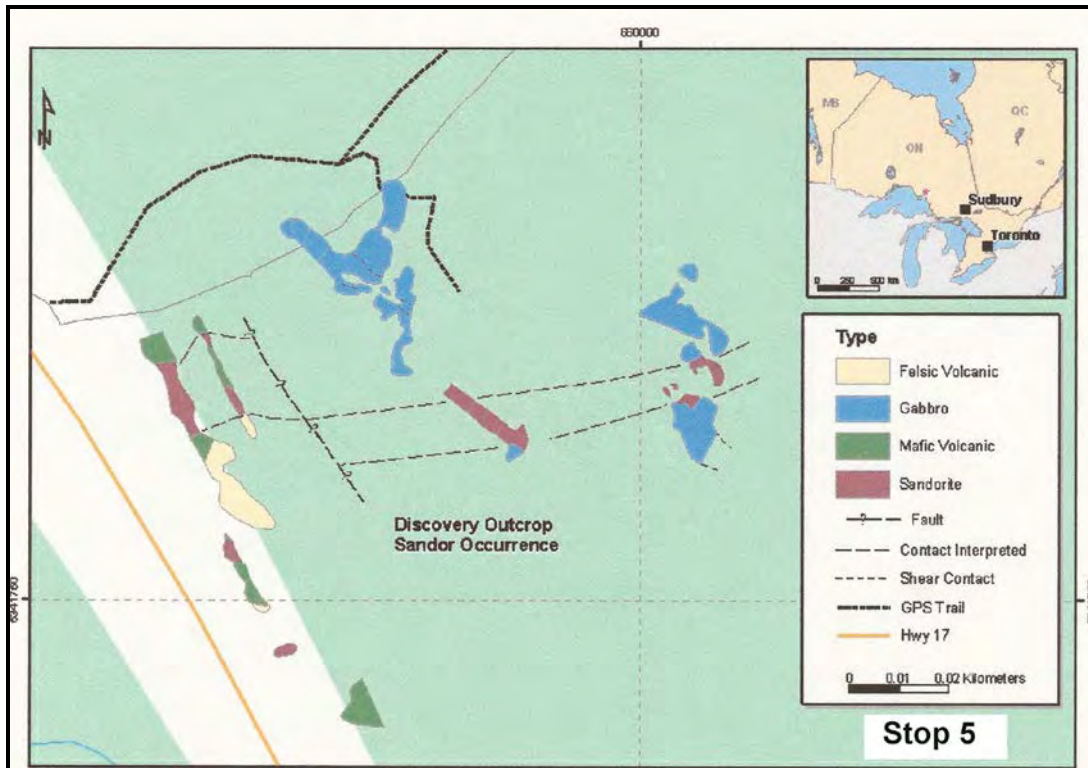


Figure 11. Detailed geology of the Sandor Occurrence, Spider Resources Inc. & KWG Resources Inc. (Kjarsgaard et al. 2003).

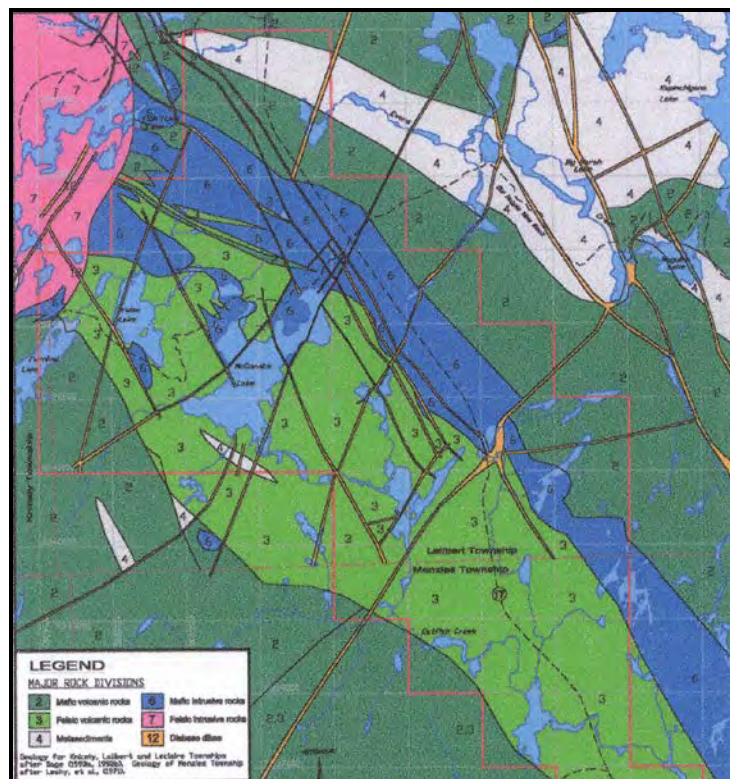


Figure 12. Generalized geology of Wawa Project, Spider Resources Inc. and KWG Resources Inc. (Spider Resources Inc. and KWG Resources Inc., CD-ROM presentation, update February 24, 2004)

STOP 6: - Dubreuilville Dike - Xenolith-rich lamprophyre
Spider Resources Inc. & KWG Resources Inc.
UTM co-ordinates – 0657251E 5347023N NAD83

The dike outcrops in a 1.5 m high exposure on the west side of the Trans-Canada Highway (Highway 17). In the area between the former Magpie Iron Mine in Leclaire Township to approximately the eastern contact of the Dickenson Lake Stock, there are a number of exposures of unusual dike rocks. These rocks are characterized by prominent round, to elliptical inclusions of actinolite or actinolite plus talc. Xenoliths are altered to fine-grained actinolite with or without talc, and some display zoning from talc core to an actinolite rim. The actinolite inclusions may consist of prismatic green crystals as large as 8 cm in length, which may be randomly oriented or radiating inward towards the core. The inclusions containing talc consist of a talc core with the prismatic to acicular actinolite projecting radially inward towards the core. The xenoliths are believed to represent at least two original mafic compositions which are likely to be originally of lower crust origin or deeper. A weakly developed regional schistosity crosses the dike implying an Archean age (Sage 1993, 2000). An example of one of the lower crustal to upper mantle xenoliths is shown in Figure 13.



Figure 13. Lower crustal to upper mantle xenolith, Dubreuilville Dike Stop 6, Highway 17 North.

This dike may be the one described by Higgins (1986). He reported that the dike consists of 60% euhedral amphibole, 20% biotite replacing amphibole and 15% plagioclase. Minor sphene and opaque minerals are present and chromite is reported from the core of the talc-bearing clasts. The bulk composition of the nodules is reportedly pyroxenite, but Higgins did not indicate whether their source was mantle or crust. The rounded outline of the clasts may reflect magmatic erosion during transport and the present mineralogy is the product of regional metamorphism (Higgins 1986).

STOP 7: - Monchiquite Dike

Michipicoten Post Provincial Park

UTM co-ordinates – 0663600E 5309223N NAD83

On the west side of Trans Canada Highway (Highway 17) two, 1-metre wide and one, 3-metre wide monchiquite dikes outcrop at the base of a 3 m high outcrop. The dikes are parallel, strike 060° and crosscut a portion of the Mission Stock. A 4 m wide, east-trending microsyenite dike also can be observed cutting the granodiorite and the monchiquite at this location. The north-northwest-trending Trembley Fault cuts across the north end of the outcrop (Massey 1985).

These lamprophyres are most commonly seen in road cuts along the highway and along the shore of Lake Superior. Typically, they are black when fresh and weather to an orange-brown colour. The dikes are usually narrow and are typically less than 1 m wide. Biotite-phyric and olivine-phyric varieties are most common, but pyroxene and feldspar phenocrysts also are observed. Often the dikes show evidence of multiple and composite intrusion, sometimes with tin screens of country rock trapped within the dikes. Lamprophyre dikes south of the Michipicoten River typically have a northeast trend (Massey 1985).

Lamprophyre dikes are commonly found cross cutting all lithologies south of the Wawa – Hawk – Manitowik Lakes Fault. Because of their ease of weathering they are infrequently seen in outcrop. However, the dikes are frequently found in underground mine workings. These dikes have a number of macrocrystic resemblances to the dikes seen here. As early as 1927, geologists working in the area had noted a similarity between these lamprophyre dikes and kimberlite (Gledhill 1927). Recent petrographic and mineralogical studies on several dikes in both McMurray and Lendrum townships have suggested that some of the geochemistry falls within the classification of type II kimberlites (orangeites) (Barnett 2001). To date, no diamonds have been recovered from these dikes.

These lamprophyre dikes are probably Proterozoic in age and are interpreted to represent Proterozoic alkalic magma emplacement into structures related to the Kapuskasing Structural Zone, perhaps consanguineous with the nearby Firesand Creek carbonatite complex. Rocks from the Firesand Creek carbonatite complex have a U-Pb date of 1078 ± 2.4 Ma (Sage 2000).

STOP 8: - Diamondiferous Conglomerates north of Wawa

This unit, referred to as the Leadbetter Conglomerate, is the main diamond-bearing unit on the property; it is comprised of clast supported polymictic conglomerate that is poorly sorted. Predominantly massive, the unit is rarely stratified and when it is, the bedding varies in thickness from 10-30 centimetres to over 1 meter with no obvious graded bedding. The sediments range in clast size from mud-sized particles to boulder-sized clasts which are sub-rounded to rounded, flattened and highly stretched. The matrix is very fine-grained and chlorite rich with a minor amount of quartzofeldspathic material. The clasts predominantly consists of volcanics, with a minor amount of non-volcanics (siltstone and chert).

The conglomerate appears to pinch out to the northwest in the project area, but thickens dramatically in the central and easter parts of the project area where surface exposures give an indicated thickness of up to 300 m. Deformation of this unit has resulted in stretched or elongated clasts. The degree of stretching is variable suggesting that there are local zones of intense deformation.

The diamondiferous conglomerate is interpreted to represent valley fill deposits in the proximal reaches of an alluvial fan complex (Wendland, 2009) with the matrix supported cobble and boulder conglomerates forming a debris flow sequence. An ultramafic component in the source for the matrix component of the conglomerate is indicated by immobile element geochemistry. The fact that a distinct suite of diamond indicator minerals consisting of chromite, olivine, and rare ilmenite, pyrope garnet and chrome diopside, have been recovered from the conglomerate makes it unique in comparison to the sparsely diamondiferous Wawa volcanoclastics located to the north which contain a paucity of conventional indicator minerals (Verley 2009).

STOP 9: - Contemplation of the rocks on the fireplace at the Wawa Motor Hotel

Acknowledgments

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Field Trip 3

Geology of the Wawa gold project

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1.0 Introduction

The fieldtrip on the Wawa Gold project combines an overview of the exploration and mining history in the Wawa Gold Camp and of the geology of the various mineralized structures of the property. The field trip is centered on the historic Surluga Road, along which most of the historic gold mines of the Wawa Gold Camp were exploited. The Wawa Gold Project is located 2 km east of the town of Wawa in Ontario and has a long exploration history that begun in the 1860s and has been discontinuously explored and developed since. This long period of activity resulted in the exploitation of 9 gold mines with preserved records of production (Fig. 1-1 and Table 1-1; Rupert, 1997); as well as many other small-scale mining operations without production records and the digging of numerous pits and the sinking of many shafts (Abandoned Mines Information System Database, 2017).

The main gold concentration defined so far on the Wawa Gold Project is the Surluga Deposit a NI 43-101 compliant inferred resource of 1,088,000 ounces @ 1.71 g/t gold (0.5 g/t gold cut-off; Ronacher et al., 2015; Table 1-2). The inferred Surluga Deposit resource is entirely hosted within a brittle-ductile shear zone named the Jubilee Shear Zone. Recently discovered gold zones not identified in the immediate vicinity of the inferred resource are not incorporated in the current resource model.

Table 1-1 Historic gold mine and gold production that were active on the Wawa Gold Project

Mine	Tonnes Milled	Gold Grade (g/t)	Gold Recovered (Ounces)	Production period(s)
Mariposa	8	72.99	19	1902-1903
Grace+Darwin	41,302	13.27	17634	1900-1903, 1907-1910, 1934-1937
Parkhill	114,096	14.81	54298	1929-1938
Van Sickle	8,372	6.34	1710	1933-1936
Cooper	4,435	11.42	1627	1897-1900, 1926-1939
Jubilee	107,930	4.29	36178	
Minto	57,335	12.56		
Surluga	86,082	3.12	8626	1964-1969
Totals	419,560	9.04	120,093	

Table 1-2 Mineral resource statement* of the Surluga Deposit effective May 26th, 2015 estimated by SRK consulting (Canada) Inc.

Resource Category	Cut-off Gold (g/t)	Quantity ('000 t)	Grade Gold (g/t)	Contained Metal Gold ('000 oz)
Inferred**				
Pit-Constrained	0.4	10,239	2.05	676
Outside Pit-constrained	0.4	8,630	1.07	298
Underground	2.5	955	3.73	114
Total	0.5	19,824	1.71	1,088

* Mineral resources are not mineral reserves and have not demonstrated economic viability. All figures are rounded to reflect the relative accuracy of the estimate. Composites have been capped where appropriate.

** Open pit mineral resources are reported at a cut-off grade of 0.40 g/t gold in relation with a conceptual pit shell constructed by SRK. Underground mineral resources include classified modelled blocks below the conceptual pit shell and above a cut-off grade of 2.50 g/t gold. Cut-off grades are based on a gold price of US\$1,250 per ounce, a gold recovery of 95 percent and a US\$:CAD exchange rate of 0.95.

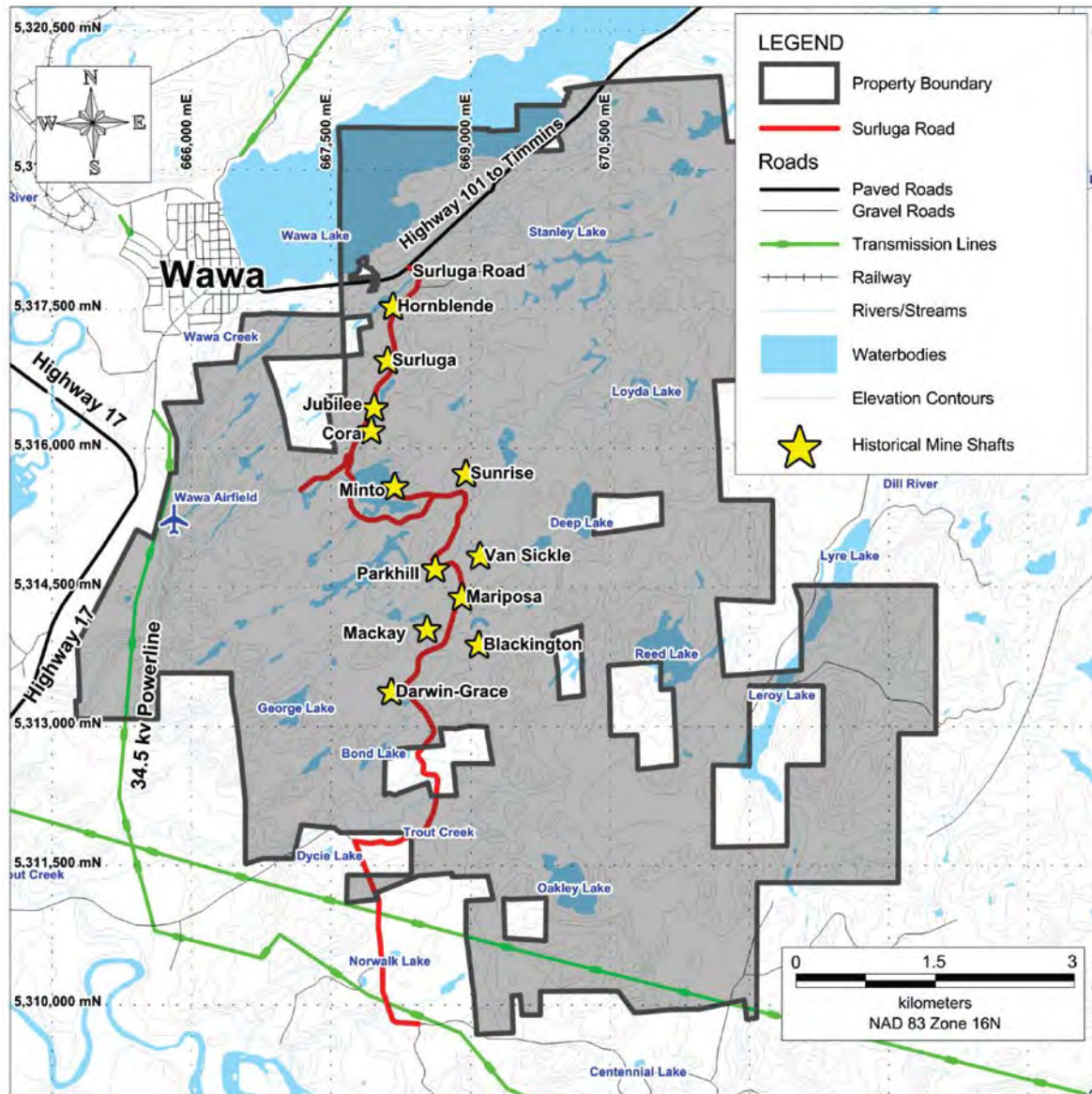


Figure 1-1 - Map showing the main historic mines as well as some of the main shafts and pits of the Wawa Gold Project

2.0 Overview of the property history

2.1 Discovery period - 1897 to 1910

The Wawa area has been explored for gold since the 1860s and gold was first discovered by William Teddy's wife in 1897 at Mackay Point along the south shore of Wawa Lake (Rupert, 1997; Frey, 1987). A staking rush followed the discovery and benefited from the change in claim staking adopted by the Ontario Government to encourage staking in 1895 (MacMillan and Rupert, 1990). This early rush period resulted in multiple discoveries.

Attempts to produce gold from bedrock started in 1897 with the sinking of many shafts and the digging of many test pits throughout the property. Between 1897 and 1898, gold was identified on the Jubilee Shear Zone and a 103-foot shaft was sunk by the Great Northern Mining Company Ltd. in the sericite schists

characteristic of the structure west of Jubilee Lake (Sage, 1994). Gold values encountered in that shaft were described as very low and the shaft was abandoned. In 1897, the Minto Mine was discovered by S. Berailldt who sold it to D. Tisdale. A 130-foot inclined shaft was sunk and the Minto mine operated until 1900. The Hornblende Shear Zone, located 300 metres west of the Jubilee Shear Zone, was discovered in 1899 by Mr. Peter Nissen. Two inclined shafts of 22 and 32 feet were sunk in the structure and in 1900, a test mill was constructed near Hornblende Lake by the Hornblende Mining Company. In 1902 and 1903, the Mariposa Gold Company sunk the 208-foot Mariposa shaft inclined at 80°NE in the footwall of the Mariposa Vein with two drifted levels at 100 and 200 feet (Sage, 1994).

Gold production in the discovery period started on a larger scale in 1900 following the discovery of the Grace vein. The Algoma Commercial Company sunk a 304-foot shaft and extracted 6,097 tons of ore between 1900 and 1903 after which, commercial gold production ceased (Sage, 1994). The Lepage Gold Mining Company resumed production in the Grace Mine between 1907 and 1910 and produced 4,260 tons of ore (Fig. 2-1).



Figure 2-1 - Grace Mine in 1908

Source: http://www.michiwawa.ca/albums/wawa/goldfields/goldfield_one.htm

2.2 Peak of mining activity - 1925 - 1938

During the period between 1910 and 1925, the Wawa Gold Project went into an exploration and production hiatus, with many land transactions between different parties and brief periods of field activity (Sage, 1994). The period extending from the mid-late 1920s to the late 1930s then saw the peak of mining activity on the property, with several mines in operation. By the late 1930s, 15 mines produced gold in the Wawa area (Frey, 1987). On the Wawa Gold Project, production records exist for eight of the mines that produced during that period (Cooper, Minto, Jubilee, Parkhill, Grace+Darwin, Mariposa and Van Sickle; Figure 1-1; Tables 1-1). The Cora vein, located in the Jubilee Shear Zone, was also briefly mined in the

Cora shaft in 1927. The other larger mine from that period that produced gold from the Jubilee Shear Zone was the Jubilee Mine that produced 107,930 tonnes @ 4.29 g/t gold. The largest producer of that period was the Parkhill Mine that produced 54,298 ounces of gold from 114,096 tonnes @ 14.81 g/t gold between 1929 and 1938 (Table 1-1; Figure 2-2).

PARKHILL (aka GOLDPARK)

This is a picture of the town of Parkhill. The headframe of the Parkhill mine can be seen in the background.

(Picture provided by Howard Whent)



Figure 2-2 - Mining town of Parkhill inhabited during the operation of the Parkhill Mine

Source: http://www.michiawawa.ca/albums/wawa/goldfields/goldfield_one.htm

2.3 Surluga Mine discovery and first mining operation - 1960 to 1980

Exploration and development activity substantially slowed down in the 1940s and the 1950s, but resumed in 1960 when Tom Surluga and W.D. Sutherland drilled 25 holes between 1960 and 1961 that resulted in the discovery of the Surluga Deposit (Table 5-7; Sage, 1994). In 1962, Sutherland and company formed Surluga Gold Mines Limited to continue the development of the Surluga Deposit (Table 5-7). Between 1964 and 1968, Surluga Gold Mines Limited sank a 950-foot shaft with 7 levels spaced by 150 feet that formed the Surluga mine. In 1967, Surluga Gold Mines Limited constructed a 750 ton per day mill on the property and started an extensive underground drilling program (Amalgamated reports, 41N15NE0036). The mill was in operation in 1968 and 1969 but, inadequate definition of reserves in the mine before its construction precluded its optimal operation. The mill was shut down in 1969 and underground development continued between 1969 and 1971 with the objective of defining enough reserves to operate the mill (Amalgamated reports, 41N15NE0036). In 1973, under strong debt pressure, the Surluga Gold Mine company was re-organized as Pursides Gold Mines that conducted an extensive exploration program between 1973 and 1975 (Amalgamated reports, 41N15NE0036). Pursides Gold Mines ultimately went bankrupt and was re-organized as Citadel Gold Mines Inc. ("Citadel") in 1980.

2.5 Second development of the Surluga Mine by Citadel Gold Mines - 1986 to 1991

In 1986, before additional reserves were defined and the exploration model updated, the Surluga mine was dewatered, the Surluga mine shaft refurbished, the mill reconstructed, and a 3 year program of surface and underground drilling and geological mapping started on the Surluga Deposit (Rupert, 1997). One exploration success is the discovery in 1989 of the Old Tom zone in the southernmost part of the

Surluga Deposit. During that period, Citadel also undertook an extensive exploration program of its property to find additional gold to feed the newly refurbished mill (Rupert, 1989a, b; Rupert and Leroy, 1989). This included diamond drilling of Root and Cooper-Ganley vein systems, stripping, trenching, channel sampling and geological mapping, as well as many airborne and ground geophysical surveys. In 1987, Citadel purchased from Duraine, the Parkhill and Grace-Darwin Mine properties (Rupert, 1997).

Development and exploration in the Surluga Deposit however stopped again in 1989 because of the mill inefficiency, lack of defined reserves to feed the mill, the un-optimized design of the mine, including the difficulties of mechanizing production and problems with dilution control because of the cryptic boundaries of the high-grade zones (E. Hoffman, pers. comm.).



Figure 2-3 Surluga mine in 2002 before site reclamation in the closure plan

Source: http://www.michiawawa.ca/albums/wawa/goldfields/goldfield_one.htm

2.6 Change of the exploration model for the Surluga Deposit - 1990 to 1996

In 1996, while reviewing the available data on the Surluga Deposit, Bowdidge (1996) postulated that the Jubilee Shear Zone is a large-scale structure up to 150 feet thick and contains widespread low grade mineralization; and evaluated that a substantial resource of low-grade mineralization may exist in structure. This evaluation changed the perception of the property and forms the basis of the exploration model now used by Red Pine Exploration.

3.0 Geology of the Wawa Gold Project

3.1 Regional Geology

The Wawa Gold Project is in the southern part of the Michipicoten greenstone belt, one of two greenstone belts that form the Wawa Subprovince of the Superior Province (Fig. 2-1). The Michipicoten greenstone belt was formed by three cycles of mafic to felsic volcanism associated with concomitant subvolcanic intrusions (Sage, 1994). Zircon U-Pb ages date volcanic Cycle 1 to 2.9 Ga, volcanic Cycle 2 to 2.75 Ga, and volcanic Cycle 3 to 2.7 Ga (Turek et al., 1992; Sage, 1994). In the southern part of the Michipicoten greenstone belt, two large intrusive complexes were emplaced during cycles 1 (Hawk Lake Granitic Complex) and 2 (Jubilee Stock). These syn-volcanic intrusions have been interpreted to delineate the centres of calderas and to be the intrusive equivalent of their host felsic to intermediate volcanic rocks (Sage, 1994). Like other greenstone belts within the Superior Province, the Michipicoten greenstone belt contains basaltic to komatiitic volcanic rocks. Throughout most of the Michipicoten Greenstone Belt, the hiatus between volcanic Cycles 2 and 3 was marked by the deposition of large banded iron formations. In the Wawa area, from their discovery in 1898 to the closure of the last iron mine of the region in 1998, these iron formations were explored and mined over a 100 year period and were at one point the largest source of iron in Canada. All the rocks of the Michipicoten greenstone belt are metamorphosed at greenschist facies and its volcano-plutonic sequences have been repeatedly deformed and folded.

Post-Archean magmatism includes diabase dikes and the emplacement of the Firesand River Carbonatite intruded along the Wawa-Hawk Lake-Manitowik Lake Fault System. The Wawa Gold Project is located within the southern part of the Michipicoten greenstone belt (Sherman, 2005).

A prominent structure in the southern Michipicoten greenstone belt is the Wawa-Hawk Lake-Manitowik Lake Fault System, which defines the boundary between a lamprophyre-rich domain to the south and lamprophyre-free domain to the north (Figure 2-1). The emplacement of the Firesand River Carbonatite along the Wawa-Hawk Lake-Manitowik Lake Fault System suggests that the fault is deep-seated, whereas the location of the Jubilee Stock and Hawk Granite Complex along the fault indicate that it may follow an older structure active during the formation of the greenstone belt.

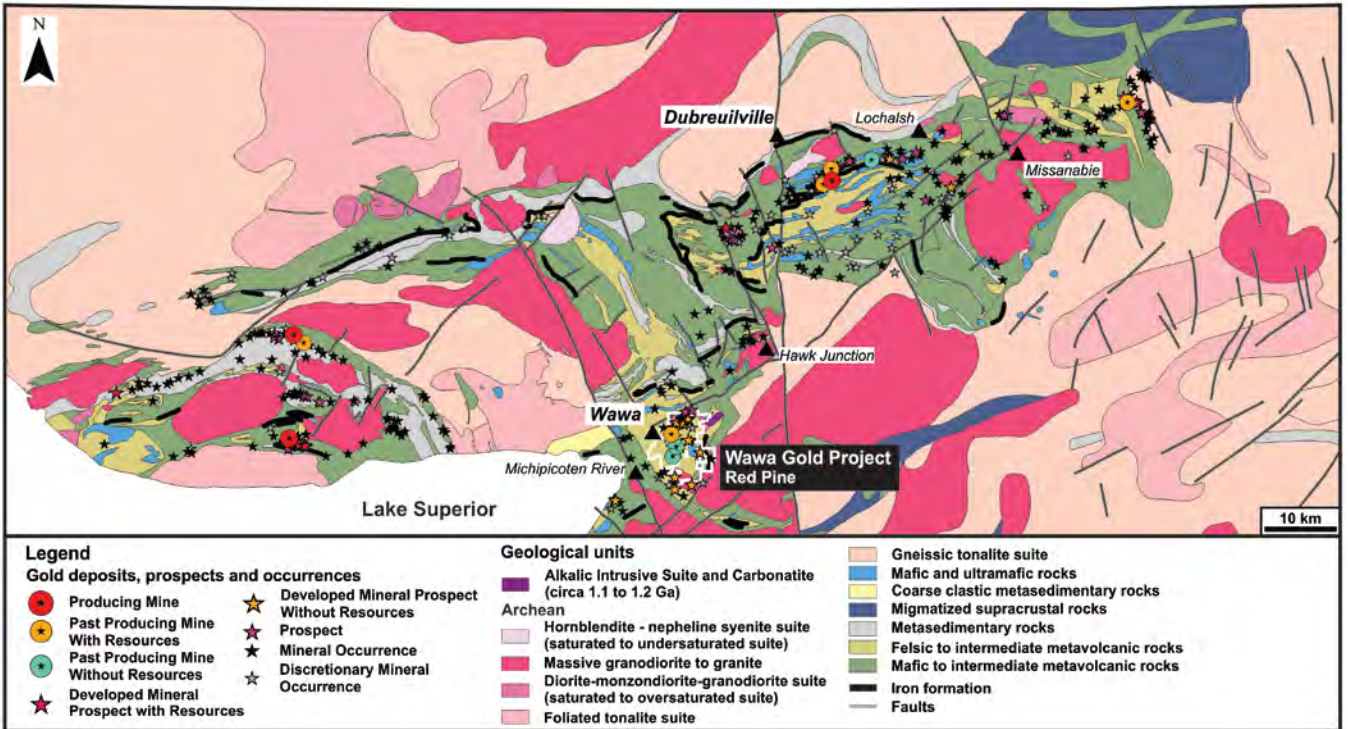


Figure 3-1 - Regional Geology of the Michipicoten Greenstone Belt and location of the Wawa Gold Project within the belt

Geology from Ontario Geological Survey (2011) open file MRD 126; Showing and deposit locations from Ontario Geological Survey (2014) Mineral Deposit Inventory

3.2 Property Geology

The Wawa Gold Project occurs in a mafic to felsic volcano-plutonic sequence formed during the second cycle of volcanism in the Michipicoten Greenstone Belt (Sage, 1994).

3.2.1 Jubilee Stock

The main intrusive complex of the property is named the Jubilee Stock that represents a key geological unit of the property for gold mineralization as all the largest historic mines (Darwin-Grace, Jubilee, Minto, Surluga, Parkhill) and most of the known gold showings are located within or at the margins of the Jubilee Stock (Fig. 3-2). The compositional and geometrical complexity of the Jubilee Stock, comprising many contacts zones between rocks of different rheology, are interpreted to be important controls on the geometry and distribution of the gold zones.

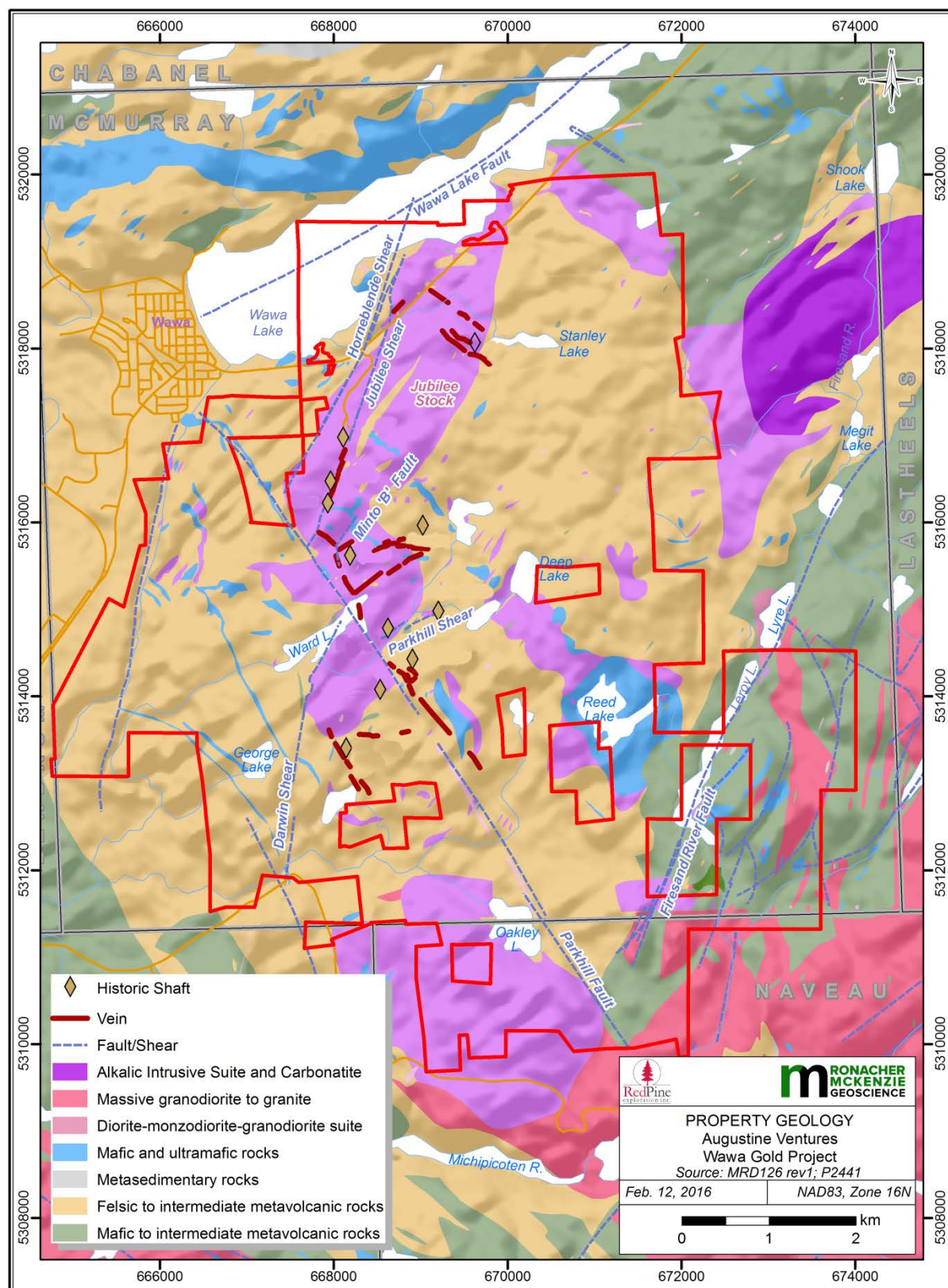


Figure 3-2 - Geology map of the Wawa Gold Property from Ronacher et al. (2016)

The Jubilee Stock is a composite calc-alkaline to transitional intrusive complex formed by many individual intrusions compositionally ranging from gabbroic-dioritic to granitic (Figs. 3-3a, b and 4a, b). Based on U-Pb ages on zircons (Sullivan et al., 1985), the coarser-grained intrusions of the Jubilee Stock was emplaced around $2,745 \pm 3$ Ma, whereas one porphyritic intrusions located at the margins of the stock was dated at $2,742 \pm 6$ Ma (Turek et al., 1992). Within error, these ages are contemporaneous to those of its host volcanic sequence dated at $2,746 \pm 11$ Ma and $2,744 \pm 10$ Ma (Turek et al., 1992). The main intrusive facies of the Jubilee Stock include: medium- to coarse-grained "diorite" (key unit historically used to define the Jubilee Stock), porphyritic intrusions with variable phenocryst assemblages (biotite, feldspar-biotite, feldspar, quartz-feldspar) generally emplaced at the margins of the intrusive complex, fine- to medium-grained mafic intrusions (gabbroic), zones of magma mixing and mingling where mafic to intermediate-felsic magmas intrude each other; and broad zones of intrusive breccia developed at the contact between the intrusive complex and its host volcanic rocks, but also along the contact zones between individual intrusions of the Jubilee Stock.

Mapping by Sage et al. (1982) shows that the core of the Jubilee Stock is curved-shaped into a sigmoid form, its long axis oriented at 20° , and has an approximate 6 x 1.3 kilometres surface expression. The actual extent of the Jubilee Stock remains however ill-defined; as the porphyritic intrusions typically forming the margins of the intrusive complex are texturally similar to some of the crystal-rich volcanic rocks of the volcanic sequence and their classification as intrusive or volcanic varied according to the generations of workers on the property. Recent re-mapping by Red Pine indicates that the surface extent of the Jubilee Stock is larger than the core zone currently illustrated on the geology maps of the property.

Medium-grained dioritic to granitic intrusions (Figures 3-5 and 3-6)

Medium-grained to coarse-grained dioritic to granitic intrusions form the diagnostic core zone of the Jubilee Stock. Historically to simplify the nomenclature of those units, all the operators of the property described the medium- to coarse-grained intrusions of the Jubilee Stock using diorite as a generic term. However, chemically based on the Pearce (1996) diagram, the coarser grained intrusions of the Jubilee Stock vary compositionally between diorite and granite/granodiorite (Fig. 3-3a). This is supported by the petrographic work from Sage (1994) that indicates some of the intrusions have modal 10-30% quartz, 40-55% plagioclase and 10-20% biotite, which indicates granitic composition.

Porphyritic intrusions (Figure 3-7)

Many porphyritic intrusions surround the core of Jubilee Stock and were hypothesized by Sage (1994) to occupy the ring fracture of a large caldera centered on the Jubilee Stock. In the contact zones between different intrusions, the porphyritic intrusions are often intermixed together and with intrusions of the medium- to coarse-grained diorite. The main primary phenocrysts assemblages observed in the porphyritic units are: feldspar, biotite-feldspar, quartz-feldspar and quartz. A compositional continuum and visual gradation between the medium- to coarse-grained diorite and intrusions of the feldspar-phyric, biotite-feldspar-phyric and biotite-phyric units are commonly observed, indicating the likely coeval emplacement of those units. Because of the variability in the mapping and logging of the porphyritic units, the porphyritic units of the Jubilee Stock remain undivided. The porphyritic intrusions are compositionally similar to the Jubilee Stock (Fig. 3-4).

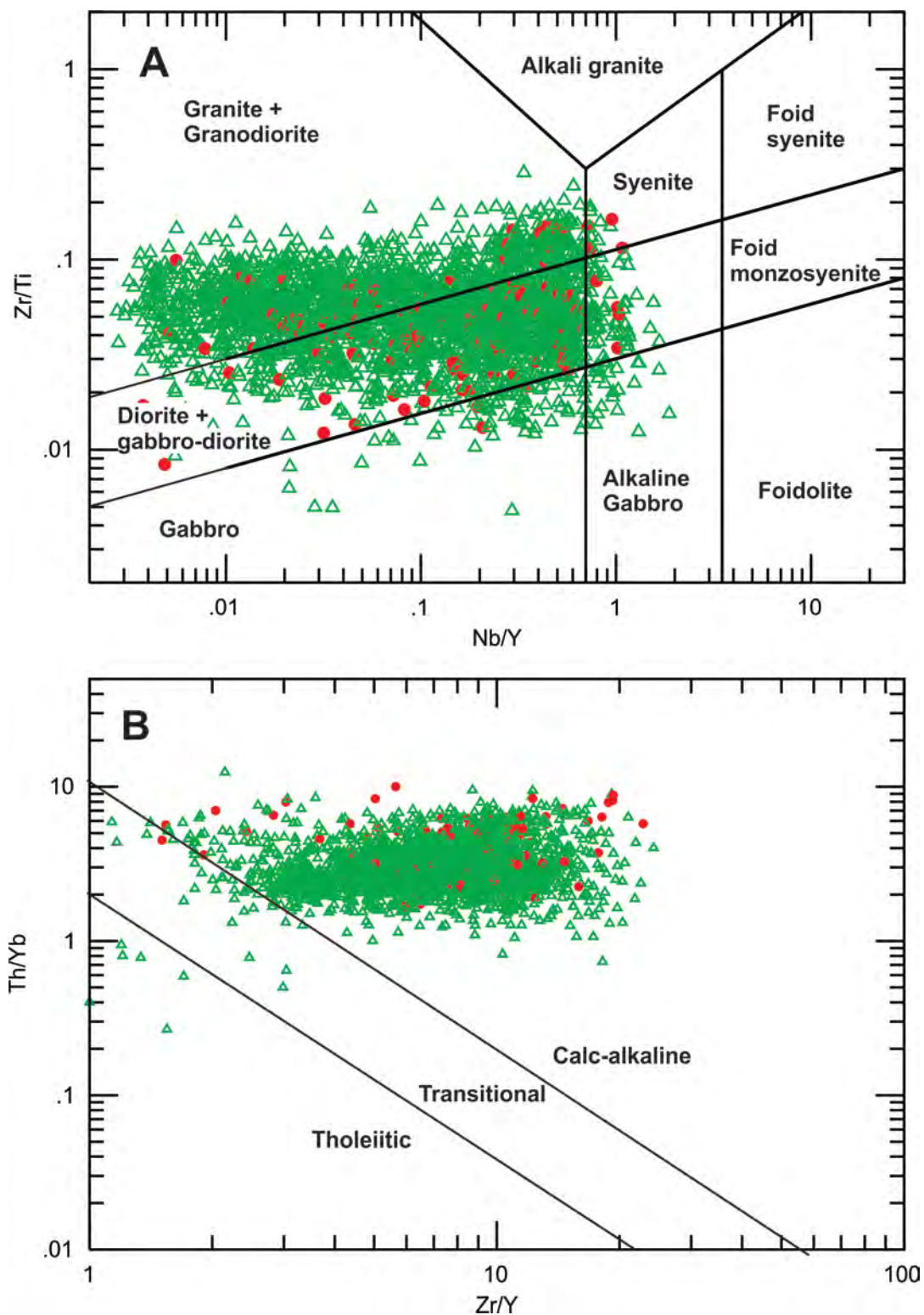


Figure 3-3 Geochemical classification diagram of the intrusions typical of the core of the Jubilee Stock. A) Zr/Ti versus Nb/Y discrimination diagram from Pearce (1996); B) Th/Yb discrimination diagram for discrimination of magmatic affinities from Ross and Bédard (2009)

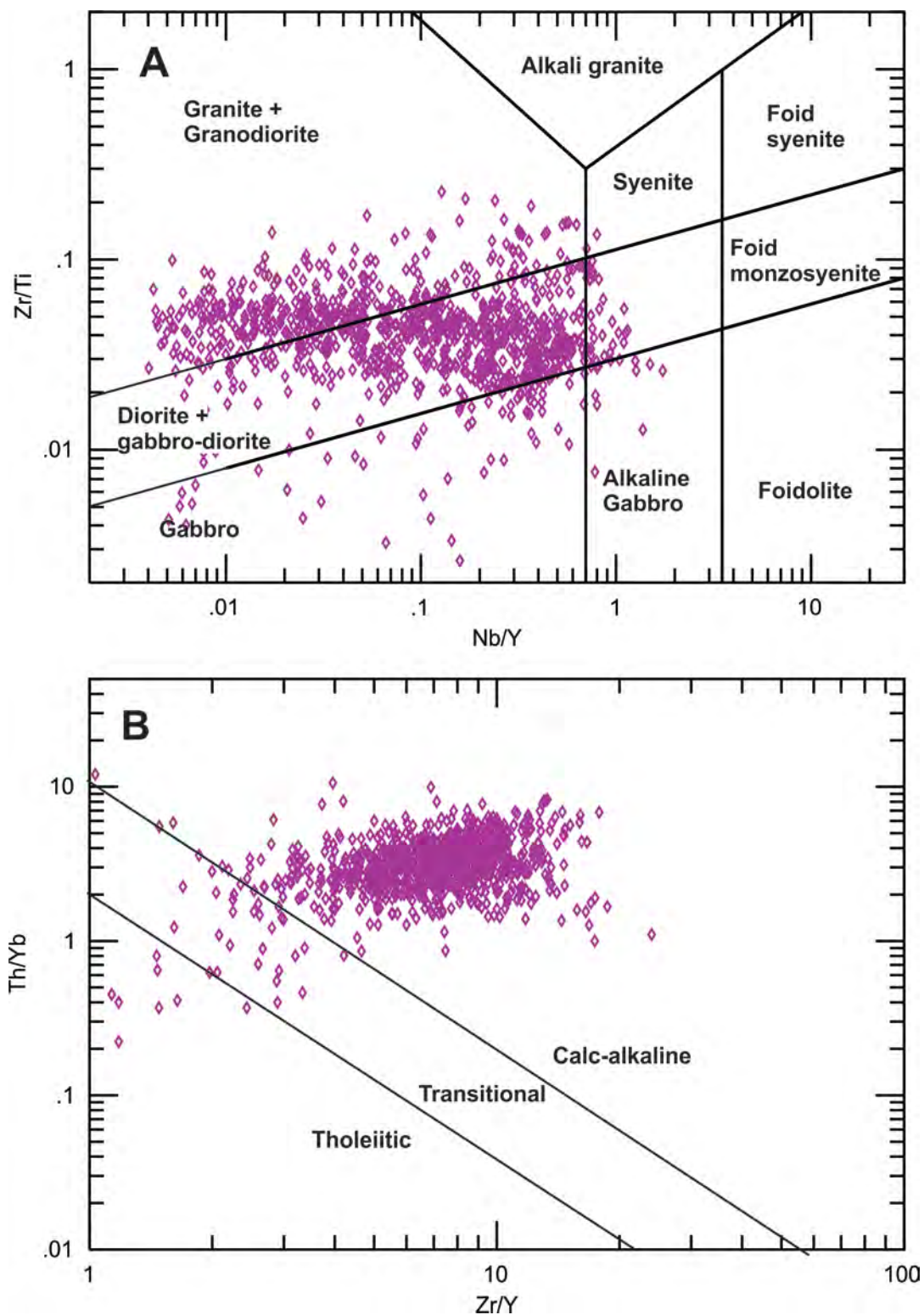


Figure 3-4 Geochemical classification diagram of the porphyritic intrusions observed in the Jubilee Stock. A) Zr/Ti versus Nb/Y discrimination diagram from Pearce (1996); B) Th/Yb discrimination diagram for discrimination of magmatic affinities from Ross and Bédard (2009)

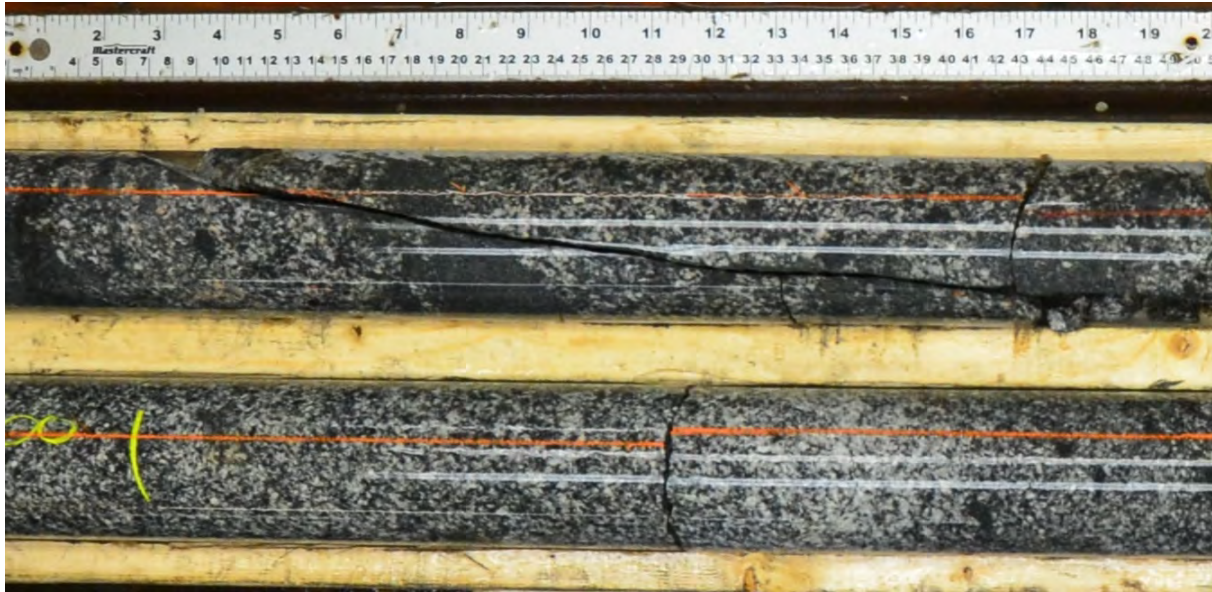


Figure 3-5 Medium- to coarse-grained facies of the Jubilee Stock "diorite" near the contact with the volcanic units containing enclaves of volcanic rocks



Figure 3-6 - Typical Jubilee Stock "diorite" in forming the core of the intrusive complex



Figure 3-7 Feldspar-quartz porphyritic intrusion exposed near the Surluga Deposit

Silicified/Albitized unit (Figure 3-8)

This unit corresponds to rock altered by a strong to intense sodic-silica alteration and encompasses altered diorite, volcanic units and porphyritic intrusions. This unit prevails in certain zones of the Wawa Gold Corridor and may correspond to the hornfelsed units described by Sage (1994) as occurring along some of the contacts between the Jubilee Stock and the volcanic rocks. In zones of intense alteration, the primary textures of the host rocks are generally destroyed, the unit becomes quite homogeneous and protolith identification is difficult. In the transitional zones, strong alteration fronts are seen to replace the host units. The predominant precursor unit is most likely fine-grained volcanic units intruded by the Jubilee stock in which albitization was preferentially partitioned.

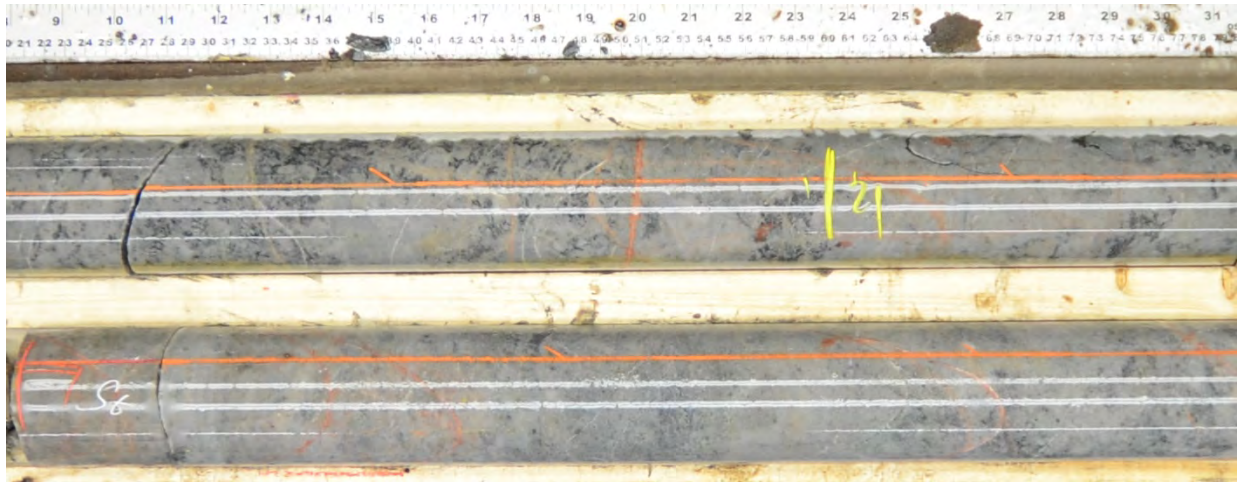


Figure 3-8 Albitized unit formed near the contacts between the Jubilee Stock and the volcanic units

Intrusive breccias (Figures 3-9 and 3-10)

Many of the contact zones between the intrusions of the Jubilee Stock, but also between different intrusive facies of the stock, are characterized by the formation of intrusive breccia zones. The breccia cement is typically composed of the coarser-grained facies granitic intrusions, whereas the fragments, predominantly of volcanic origin, are fine- to very fine-grained, vary considerably in size, ranging from a few millimetres to tens of metres, and some are partially assimilated by the dioritic magma. A report by Sage (1994) and noted by Red Pine geologists, this is making the mapping of this unit, especially in drill cores, particularly challenging.



Figure 3-9 Surface exposure of the intrusive breccia formed at the contact between the Jubilee Stock medium- to coarse-grained diorite and the volcanic units

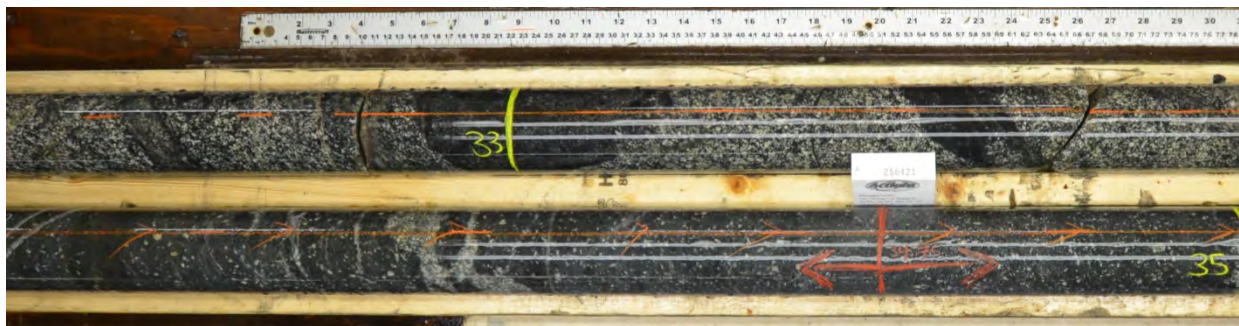


Figure 3-10 Intrusive breccia texture in drill hole and melanocratic feldspar-phyric unit in the contact zone between the Jubilee Stock coarse-grained diorite and the volcanic units

3.2.2 Gabbroic rocks

Tholeiitic to calc-alkaline mafic-intermediate to mafic intrusions were documented by Red Pine in the Jubilee Stock (Fig. 3-11). Visually, the mafic intrusions were described and discriminated based on the grain size of the core of the intrusion and the absence or presence of a porphyritic texture with feldspars being the main phenocrysts. The mafic intrusions observed in the Jubilee Stock include: coarse-grained gabbros pertaining to the tholeiitic suite (Fig. 3-12), fine-grained gabbros pertaining to the tholeiitic suite (Figure 3-13), and feldspar-phyric very fine-grained gabbro to gabbro-diorite generally pertaining to the calc-alkaline suite (Figure 3-14). Both the tholeiitic and the calc-alkaline mafic intrusions are deformed in the gold-bearing structures of the Wawa Gold Corridor. Walker (2011) also recognized, based on the observation of magma mixing textures between felsic and a mafic magmas in the Jubilee Stock, that some of the mafic intrusions are comagmatic with the stock. This observation is supported by the occurrence of calc-alkaline mafic intrusions in the Jubilee Stock. The intersection between some intrusions of the tholeiitic suite and the gold-bearing structure was also observed to form zones of preferential gold enrichments. In some of the mafic intrusions of the calc-alkaline suite in the Surluga Deposit, Ni-Cu mineralization can occur as disseminated cluster of pyrrhotite-chalcopyrite, the pyrrhotite likely intermingled with pentlandite (Fig. 3-14).

The largest mafic intrusion on the property is centered on Reed Lake and forms the Reed Lake mafic-ultramafic complex which is composed of diorite, quartz-gabbro, leuco- to mela-gabbro and pyroxenite (Fig. 6-2). Sage (1994) inferred that the combined trends of the long axis of 315° for the Reed Lake and 20° for the Jubilee Stock may suggest there were emplaced in a conjugate fracture system and that they are possibly petrogenetically related.

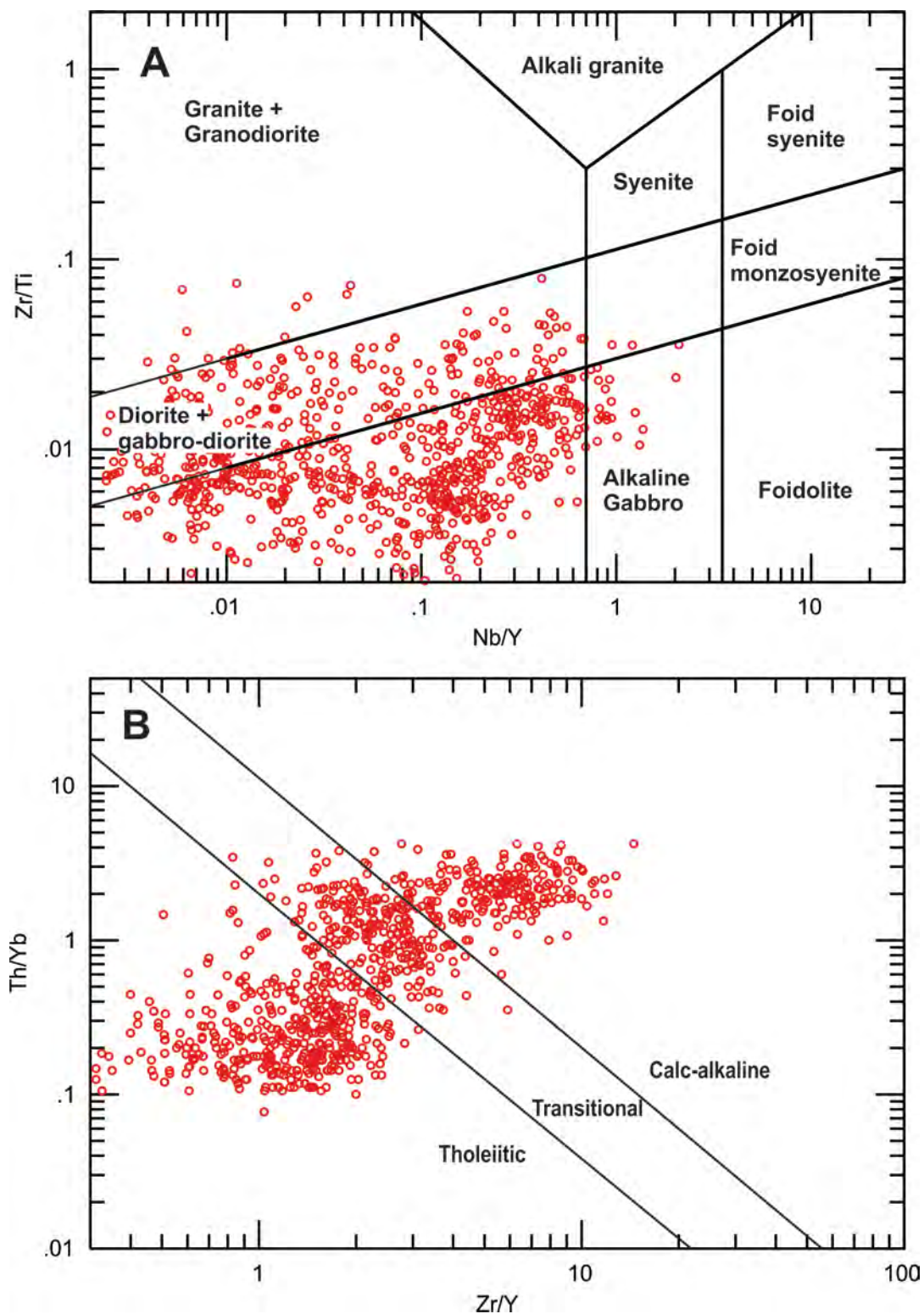


Fig. 3-11 Geochemical classification diagram of the mafic and mafic/intermediate intrusions observed in the Jubilee Stock. A) Zr/Ti versus Nb/Y discrimination diagram from Pearce (1996); B) Th/Yb discrimination diagram for discrimination of magmatic affinities from Ross and Bédard (2009)

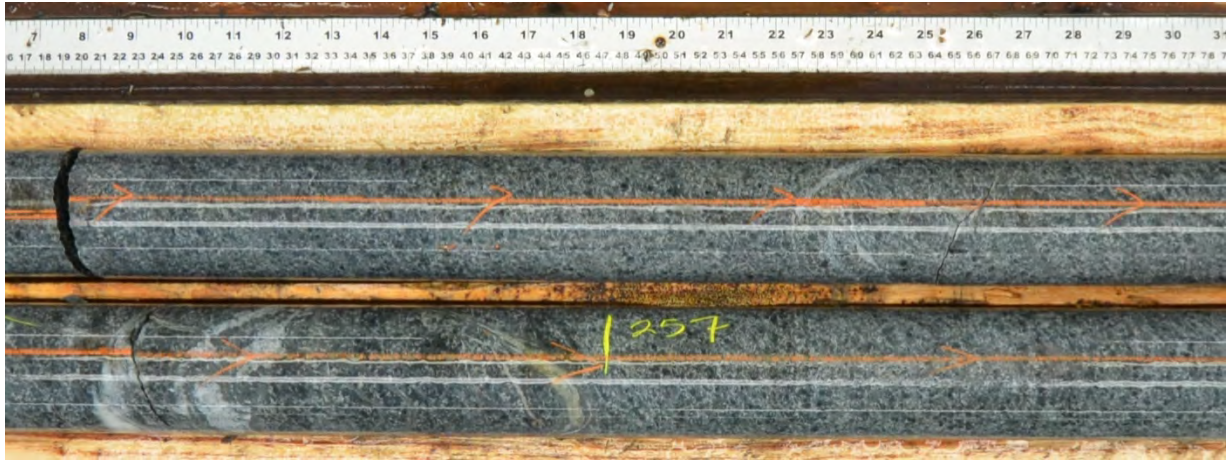


Figure 3-12 Coarse-grained tholeiitic gabbroic intrusion in the Jubilee Stock



Figure 3-13 Fine-grained tholeiitic gabbro dyke in the Jubilee Stock

3.2.3 Volcanic units

For most of the Wawa Gold Project where large surface mapping and drilling programs have been conducted, the descriptions of the volcanic units are constantly evolving depending of the geologist, exploration model, and time period. In many cases the sub-volcanic porphyritic intrusions, part of the Jubilee Stock, and the volcanic units, are confused and their classification inter-changed. To add to the difficulties of recognizing the volcanic units, in historic logs, in adjacent drill holes, some of the gold-bearing structures are variably described as sedimentary rocks, fragmental volcanic units, volcanic flows and porphyritic intrusions.

3.2.4 Lamprophyre dikes

Lamprophyre dikes are pervasive throughout the Wawa Gold Project and at least two generations of lamprophyre exists. One generation is late-stage and cut all the gold mineralized zones of the property. Dikes of that generation are black, porphyritic, medium-grained and strongly magnetic with a blue amphibole alteration halo. A possible set of lamprophyre is likely older. Dikes of that set are generally smaller, their primary mineralogy is partially to completely replaced, which gives them a dark- to pale-greenish color. One dike of this set is also possibly gold mineralized, indicating that some of the lamprophyre dykes could have been emplaced prior to the formation of the gold system. A few carbonatite dikes, likely related to the Firesand Carbonatite located a few hundred metres east of the northeastern corner of the property, were also observed in drill holes in the Surluga Deposit.

3.3 Late brittle faulting

The main brittle fault of the Wawa Gold Project is the NW-oriented and subvertical Parkhill Fault. Following Sage (1994), the Parkhill Fault is the southeastern extension of the northwest-striking Black Trout Lake Fault and is seen to truncate the Wawa Gold Corridor, which likely resurfaces to the south in the Darwin Shear Zone. The age of the Parkhill Fault remains uncertain and its intrusions by gabbroic rocks interpreted to be Archean indicate that it is possibly a long lived-structure in the area, even possibly formed during the evolution of the gold system. The interpreted late movement along the Parkhill Fault is left-lateral with an unknown vertical component.

4.0 Gold zones of the Wawa Gold Project

4.1 Attributes of the gold zones of the Wawa Gold Project

Gold mineralization is obvious throughout the Wawa Gold Project and is spatially related to the numerous shear zones, fractures and replacement zones of variable strike and dip. The Surluga Deposit forms the largest gold concentration currently defined on the Wawa Gold Project. The deposit is entirely hosted in the Jubilee Shear Zone that crosscut the various intrusive rocks of the Jubilee Stock. The Jubilee Shear Zone is a brittle-ductile structure consisting of a number of parallel, ~300–900 m long en-echelon segments that strikes northeast (018–034°) and dips (25–55°) to the southeast (Figs. 4 to 6). Its width ranges from 9 m to 75 m and extends from Wawa Lake to the northwest-trending Parkhill Fault (3.2 km). Helmstaedt (1988) interpreted that the gold-mineralized quartz veins predate some of the brittle-ductile shearing in the Jubilee Shear Zone and that the geometry of the high-grade zone of the deposit is controlled by a strong stretching lineation in the shear zone. Syn-deformation folding in the structure may also increase the thickness of the mineralized zone as parasitic folds, generally present in the thickest mineralized zones of the structure (Figure 6-19).

Red Pine Exploration demonstrated that the Jubilee Shear Zone is part of a larger deformation corridor that includes many other gold-bearing structures and zones of gold mineralization. This larger deformation corridor, in which tectonic foliations oriented NNE to NE (20°–45°) and dipping between 30° and 80° and stretching lineation trending 160°–190° and plunging 20°–35° are systematically observed (Fig. 7), was named the Wawa Gold Corridor. In addition to the Jubilee Shear Zone, the main gold-bearing structures of the Wawa Gold Corridor includes: the Darwin Shear Zone (interpreted southern extension Jubilee Shear Zone south of Parkhill Fault), the Grace Deformation Zone, the Hornblende shear zones (Hornblende Upper and Lower), the Parkhill Mine Shear Zone, the Minto shear zones (Minto A, B

and C), the Surluga Road Shear Zone, and the William gold zones. From north to south, the Wawa Gold Corridor is continuously defined between Wawa Lake and the Darwin-Grace Mine. Laterally, the Wawa Gold Corridor is defined between the Hornblende Shear Zone and the Minto B Shear Zone north of the Parkhill Fault, and between the Darwin Shear Zone and the Darwin-Grace Mine, south of the Parkhill Fault (Figs. 1, 3, 8-10). Throughout the Wawa Gold Corridor, gold mineralization predominantly occurred in brittle-ductile shear zones, but also occurs in micro-breccias formed into albitized zones of the Jubilee Stock intrusions, and into network of quartz tension veins.

Although the main orientation of the tectonic fabrics defining the Wawa Gold Corridor and some of its main gold zones is NNE (Hornblende, Jubilee, Minto B, Darwin), additional orientations exist for the corridor gold-bearing shear zones. This includes ENE-oriented shear zones (Minto C, Parkhill Mine Shear Zone, Mickelson Shear Zone, Root Vein structure) and WNW-oriented shear zones (Grace Deformation Zone, Minto A shear Zone). Domains of L-tectonite controlled by the stretching lineation trending 160°-190° and plunging 20°-35° without strong planar fabric development are also relatively common.



Figure 4 - Cora Shaft into the Jubilee Shear Zone



Figure 5 - Brittle-ductile shearing and tight parasitic folding of auriferous quartz veins in the Jubilee Shear Zone



Figure 6 - Zone of brittle deformation in the Jubilee Shear Zone



Figure 7 - Characteristic stretching lineation of the Wawa Gold Corridor preferentially partitioned in a mafic dike (William Gold Zone)



Figure 8 - Brittle-ductile deformation in the Minto B Shear Zone

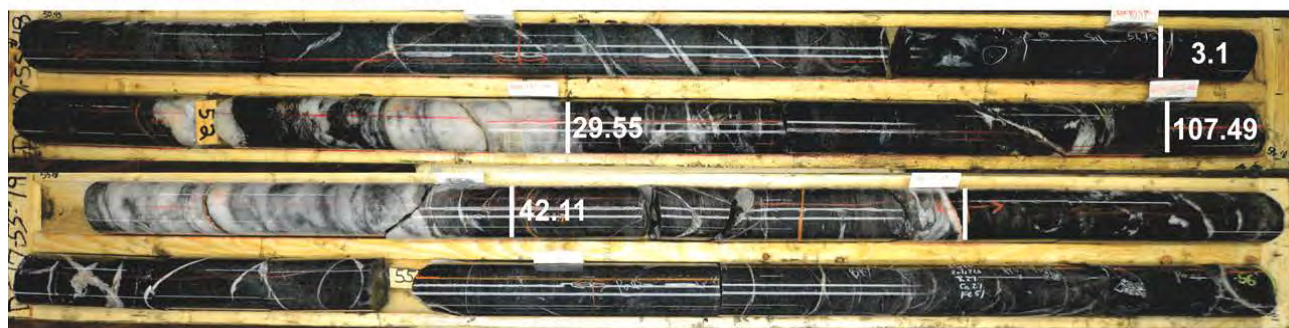


Figure 9 - Intersection of the Minto A Shear Zone, related to the Minto and Parkhill mines, and the Jubilee Shear Zone

DG-17-54 - 29.27 g/t Au over 3.4m



DG-17-55 - 42.22 g/t Au over 2.44m



DG-17-56 - 57.31 g/t Au over 3.14m & 3.56 g/t Au over 2.22m

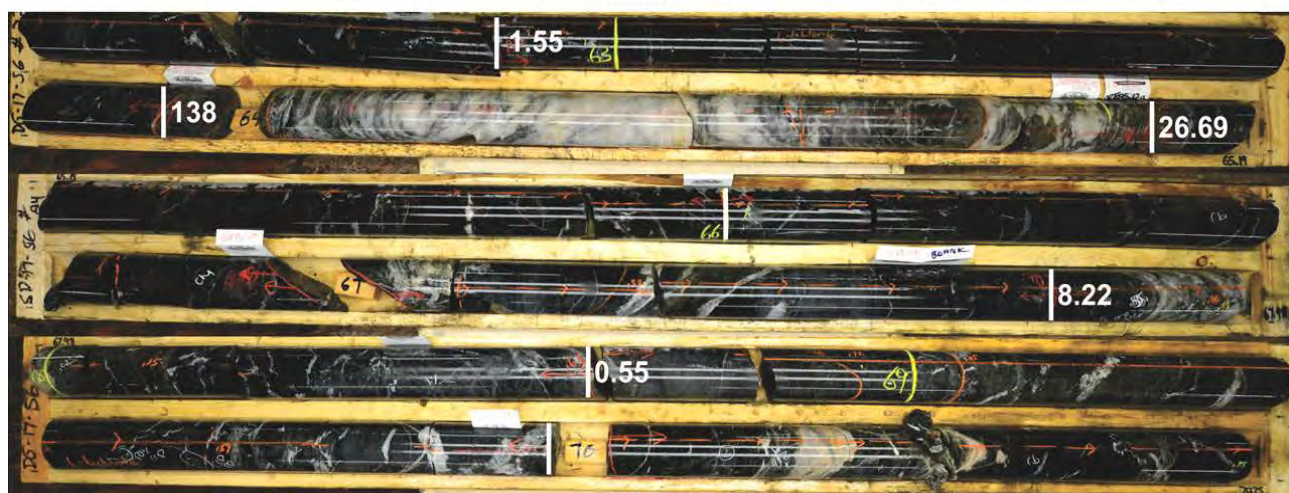


Figure 10 - Gold mineralization in the Grace Deformation Zone related to the historic Darwin-Grace Mine



Figure 6-19 - Folded high-grade quartz shear vein in the Mickelson Shear Zone

Channel sample contains 69.5 g/t gold over 0.7m

In most of the gold zones on the property, key visual indicators of gold mineralization include the presence of sulfides, potassic alteration (sericitic or biotitic) and development of quartz veins with accessory sulfides (Fig. 11).

The main style of gold mineralization in the Wawa Gold Corridor, in both the shear zones and the replacement zones, principally occurs as free gold associated with pyrite, accessory to minor pyrrhotite and minor to absent chalcopyrite deposited with silicification and sericitic alteration, and with or without quartz veining. Moderate to strong tourmalinization also occurred in some of those gold zones. The modal content of sulfides is generally around 0.2-3%, and locally up to 5-10% in the higher grade zones, which are almost systematically associated with quartz veins.

Another style of gold mineralization is characterized by free gold associated with abundant arsenopyrite with accessory to minor pyrite and/or pyrrhotite deposited with chlorite+biotite alteration, weak to moderate-strong sericite alteration and with or without quartz veining. Arsenopyrite in those gold zones can occur in modal content exceeding 3% (Grace Deformation Zone, Minto C shear zone). In the Jubilee Shear, Hornblende and Minto A+B shear zones, arsenopyrite-bearing or pyrite-bearing gold zones occur in close spatial association. Many zones with a transitional metal signature between the As-rich and other metal assemblages also occur in those structures (Fig. 12). In the transitional zones, pyrite is the main sulfide, but these zones exhibit a variable arsenic content between 100-1000 ppm. In the Grace and Minto C shear zones, arsenopyrite-rich gold mineralization predominates and is associated with abundant quartz veining in the higher grade zones of the structure (Fig. 10). In the Minto C shear zone, quartz veining is also relatively minor to absent; and does not seem to be the main control on gold concentration as high-grade gold in the structure is not spatially associated with quartz veining like in the Grace Deformation Zone (Fig. 13).

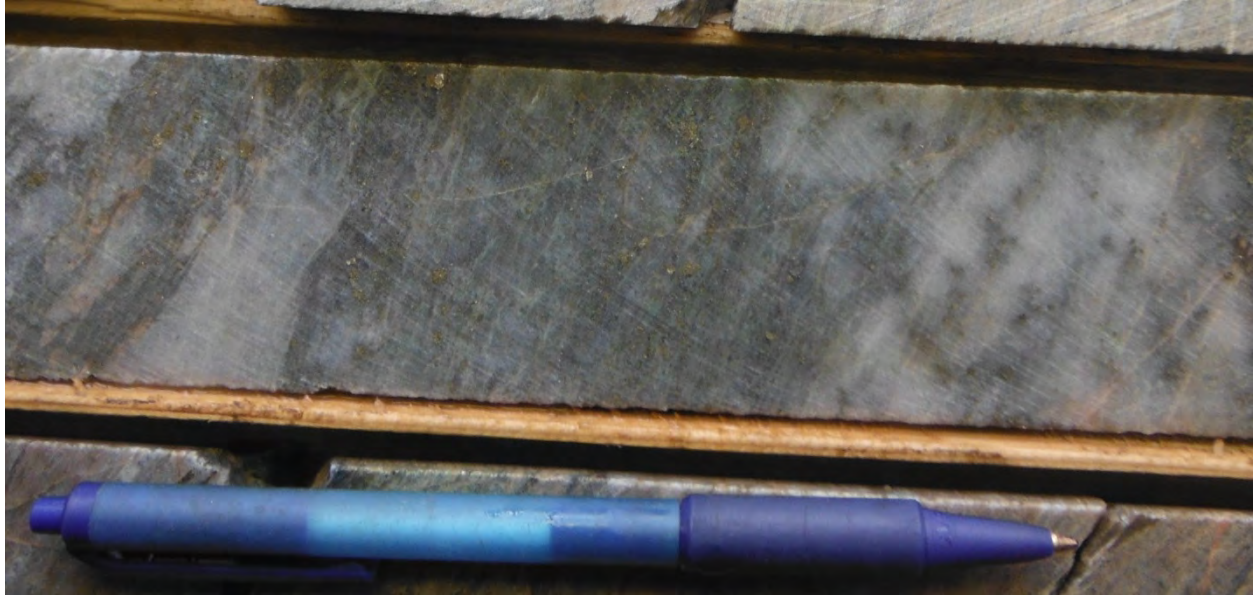


Figure 11 - Grey quartz vein with pyrite typical of the higher-grade zones of the pyritic gold zones of the Surluga Deposit

Drilling intersection contains 18.62 g/t gold over 0.48 m

A rarely observed sulfide paragenesis formed of pyrite with accessory sphalerite and galena deposited in quartz veins also occur in the Wawa Gold Corridor, and was so far only observed in the Surluga Deposit. However chemically Pb and Zn enrichments are characteristic of many of the property gold zones (Fig. 12).

The main network of gold-bearing tension veins recognized by Red Pine in the Wawa Gold Corridor is preferentially formed in the medium- to coarse-grained diorite and is not associated with obvious shearing (Fig. 13). Individual tension veins of the network are moderately to steeply dipping (50-80°) to the WNW to NNE and typically contain accessory tourmaline, accessory pyrite and pyrrhotite, minor chalcopyrite and locally visible gold. The variable development of tectonic fabrics in the tension veins of the property indicate a pre- to syn-tectonic formation of the gold mineralized tension veins.

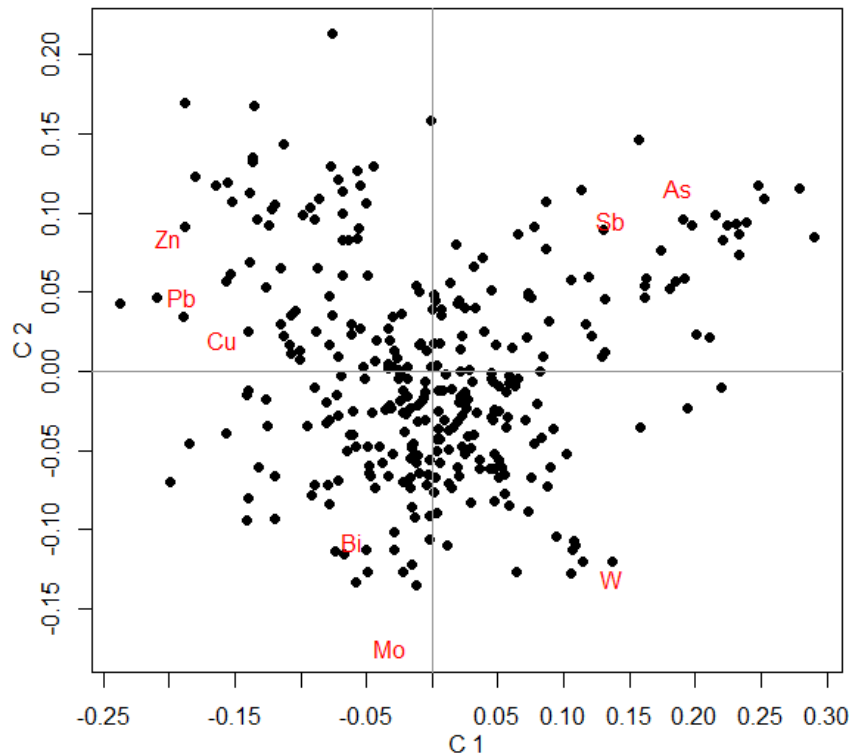


Figure 12 - PC1 vs PC2 biplot of log-centered transformed concentration of metals to illustrate the diversified metal assemblages observed in the gold zones of the Wawa Gold Corridor

Analyses from 326 drill core samples with Au > 1 g/t; because of the large number of analyses below detection, Ag, Se and Te were not used for the PCA.



Figure 12 - High-grade core of the Minto C Shear Zone formed after a strongly sericitized and silicified quartz-feldspar porphyry

Grab samples from this arsenopyrite-rich shear zone without ubiquitous quartz veins contains 17 g/t and 5.51 g/t gold



Figure 13 - Quartz tension vein with visible gold in the footwall of the Surluga Deposit
Drilling intersection contains 53.2 g/t gold over 1 m

The William gold zones form a style of gold mineralization that contrasts with the shear-hosted gold zones of the Wawa Gold Corridor and were first identified during the fall 2015 drill program. In the William gold zones, tectonic fabrics (planar and linear) are absent to locally strong, hydrothermal alteration spatially related to gold deposition is weak to strong and mineralization is associated with pervasive micro-brecciation of the albitized host rock (Fig. 13). The main geometrical controls hypothesized for the William gold zone mineralized zone are: weak ENE-striking and shallowly dipping (20° - 35°) shearing and the stretching lineation of the Wawa Gold Corridor that shallowly plunges to the ESE. An elevated arsenic background is diagnostic of these zones and likely relates to the finely disseminated pyrite interpreted to be related to gold mineralization. Because of the weak visual indicators associated with William-like mineralized zones, they were generally missed by all previous operators of the property.

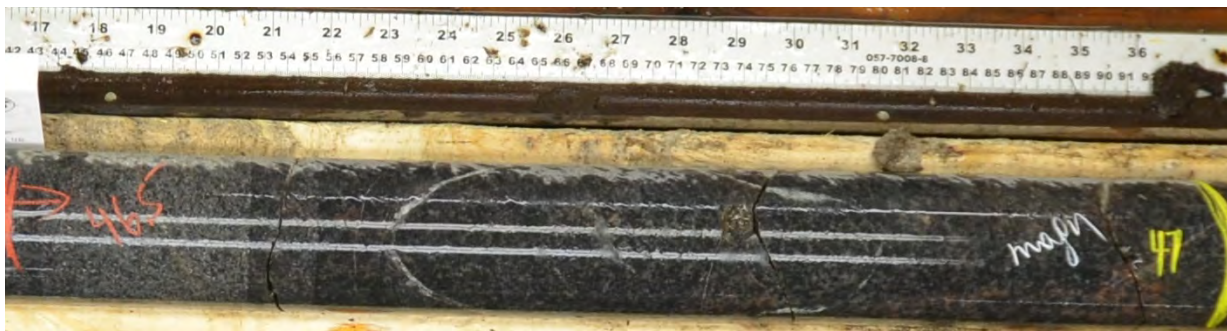


Figure 13 - William-like mineralization in the Jubilee Shear Zone hangingwall -
Drilling intersection contains 1.86 g/t gold over 1.15 metre

Three main controls on the spatial distribution and geometry of the gold zones were defined in the Wawa Gold Corridor and include:

- rheological contrasts between the various intrusive phases of the Jubilee Stock to partition and focus deformation and fluid circulation (Fig. 14);
- chemical contrast between units of the Jubilee Stock and more specifically the contact zones between gabbroic intrusions and the other intrusion types of the Jubilee Stock to partition and focus deformation and fluid circulation, and also to act as a chemical trap for gold; and
- the stretching lineation characteristic of the shear zones of the property.



Figure 14 - Strain partitioning in the Minto B shear zone where shearing is preferentially partitioned in the mafic rocks whereas the dioritic domain remains weakly deformed

4.2 Hydrothermal alteration associated with gold mineralization

Many mineralogical changes are associated with the formation of the gold-bearing shear zones. In mafic rocks, progressive chloritization and carbonatation (calcite and ankerite) with subsidiary tourmalinization and mariposite formation indicate increasing proximity to the gold-bearing shear zones (Fig. 2). In the intermediate to felsic precursors, the mineralogical changes are progressive sericitization, silicification and carbonatation. In intermediate to mafic rocks, magnetite is also deposited in the shoulders of the gold-bearing structures and can form haloes extending 25-30 metres away from the gold-bearing structure. In the Wawa Gold Corridor the hangingwall of the Jubilee Shear Zone the biotite alteration forms a distal halo extending tens of metres away from the structure that gets progressively chloritized with increasing proximity to the shear zone. Barren sericitization in sheared intermediate to felsic precursor rocks and barren chloritization and carbonatation in sheared gabbroic rocks also occurs in the gold-bearing structures of the property.

A pervasive hydrothermal imprint related to the emplacement of swarms of lamprophyre dykes is present throughout the property. Mineralogically it leads to the formation of riebeckite in the contact zones of the lamprophyre dykes as well as networks of iron carbonate (pos. siderite) veins with K-feldspar selvages.

5. Field Trip Stops

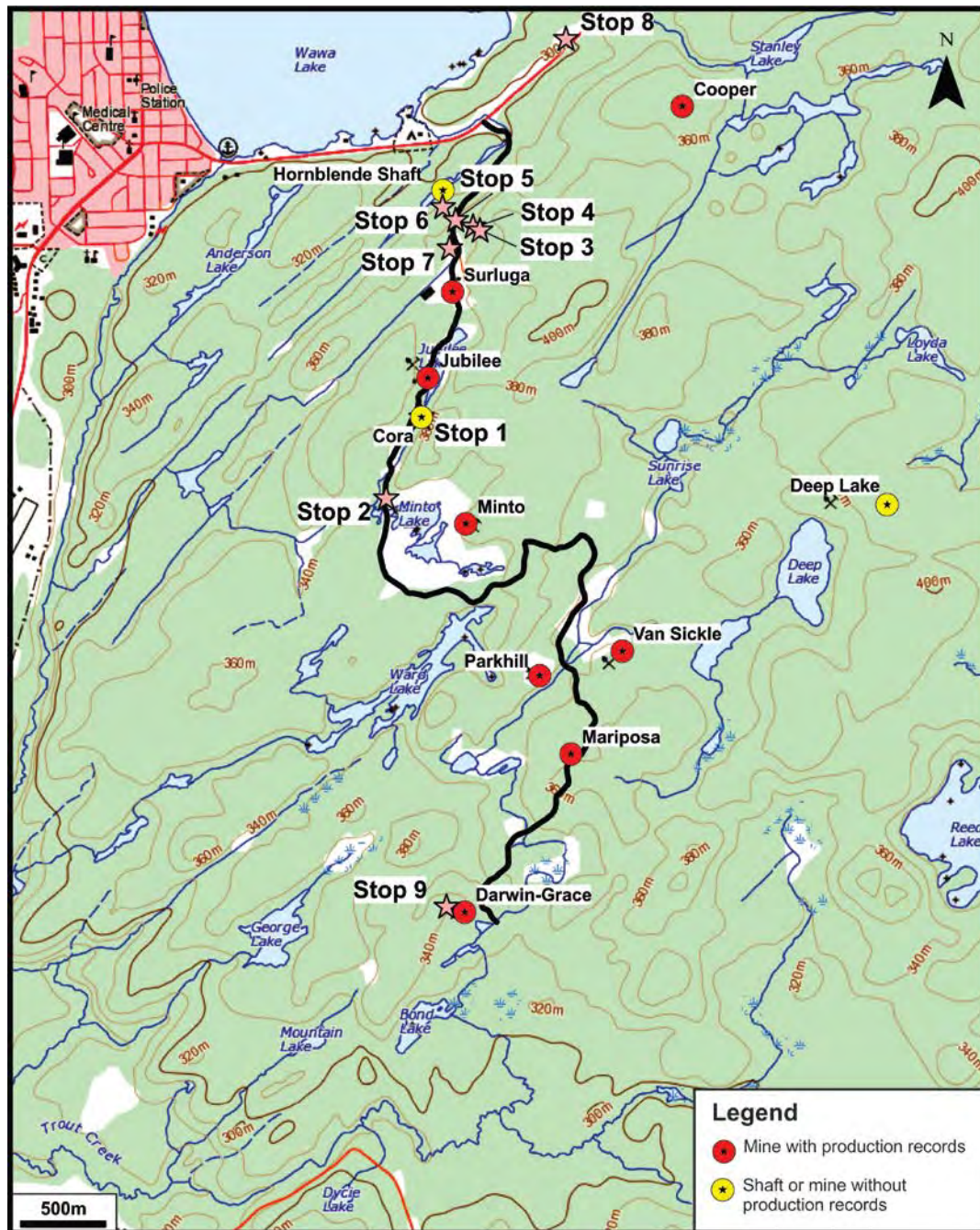


Figure 5-1 Stops of the fieldtrip on the Wawa Gold Project

Stop 1: Cora Shaft and Jubilee Shear Zone

Coordinates: UTM 16 NAD83 667930mE, 5316243mN

This stop provides an overview of the early mining history in the Jubilee Shear Zone, current host of the inferred resource estimated for the Wawa Gold Project, and what is now the Surluga Deposit. This stop is located just south of the small peninsula in Jubilee Lake where the historic Jubilee Mine was operated

between 1926 and 1938 (Fig. 5-1), and where the Cora Shaft was sunk on a quartz vein in the Jubilee Shear Zone in 1927. The Cora Shaft outcrop provides a good overview of the structural complexity of the Jubilee Shear Zone, of the geometry of some of the parasitic folds in the shear zone, and on the higher grade quartz veins formed in the shear zone. Individual grab samples from the Cora Shaft quartz vein contain up to 50.8 g/t gold.

JUBILEE LAKE MINE

Started in 1926, the Jubilee Mine was in production until 1938.

(Picture provided by Howard Whent)



Figure 5-1 Jubilee Mine during operation in the 1930s

Source: http://www.michiwawa.ca/albums/wawa/goldfields/goldfield_one.htm



Figure 4 Exposure of the Jubilee Shear Zone at the Cora Shaft

Stop 2: Jubilee Shear Zone Southern Extension

Coordinates: UTM 16 NAD83 667723 mE, 5315771 mN

This stop illustrates the control of the zones rheological contrasts between different intrusions of the Jubilee Stock has on strain partitioning, fluid circulation and gold concentration in the Jubilee Shear Zone. The Jubilee Shear Zone on the outcrop traverses a contact zone between a mafic dyke and the Jubilee Stock "diorite". Brittle-ductile deformation is strongly partitioned into the mafic rocks in the shear zone, whereas quartz veining and gold concentration preferentially occur along the contact zones between the deformed mafic rocks and the "diorite". Channel sample from the contact zone contains 6.74 g/t gold over 1 metre.



Figure 5 Exposure of the southern extension of the Jubilee Shear Zone

Stops 3 to 7 - Wawa Gold Corridor transect

This series of stops illustrates some of the main gold zones and gold mineralization types of the Wawa Gold Corridor in the northernmost extension of the Surluga Deposit. The transect begins in the Jubilee Shear Zone and also covers some of the various intrusive facies of the Wawa Gold Corridor.

Stop 3: Jubilee Shear Zone Northern Extension

Coordinates: UTM 16 NAD83 668273 mE, 5317339 mN

This stop provides an overview of the Jubilee Shear Zone in the northernmost extension of the Surluga Deposit. The trench also exposes the brittle to brittle-ductile deformation regime that were active in the Jubilee Shear Zone.

Stop 4: Tholeiitic Gabbro and Jubilee Stock "diorite"

Coordinates: UTM 16 NAD83 668231 mE, 5317362 mN

This stop shows at surface, one of the tholeiitic gabbro dyke on the property. The intersection between that gabbro dyke and the Jubilee Shear Zone spatially correspond to one of the main high-grade zones defined in the Surluga Deposit. Gabbro dykes of the tholeiitic series are considered as having a critical role on the spatial distribution of the higher-grade zones in the gold-bearing structures of the property.



Figure 6 Contact between the tholeiitic gabbro and an intermediate intrusion of the Jubilee Stock

Stop 5: Surluga Road Shear Zone

Coordinates: UTM 16 NAD83 668130 mE, 5317400 mN

This stop shows one of the shear zone, parallel to the Jubilee Shear Zone, discovered by Red Pine in the Wawa Gold Corridor. Similar to the other structures on the property, the intersection between the structure and mafic/diorite contacts forms a preferential zone for gold enrichment and quartz veining. The shear is adjacent to the Surluga Road, but was not tested for gold before the exploration program of Red Pine during the summer of 2016. Channel sampling on the exposed outcrop indicate that the shear zone contains 1.59 g/t gold over 7 metres.



Figure 7 Exposure of the Surluga Road Shear Zone next to the Surluga Road

Stop 6: Hornblende Shear Zone

Coordinates: UTM 16 NAD83 668054 mE, 5317474 mN

The Hornblende Shear Zone is one of the large gold-bearing structures in the footwall of the Surluga Deposit and two contrasting styles of mineralization are visible on the exposure. The northern part of the exposure is characterized by a high-strain domain with low quartz vein density and pervasive dissemination of pyrite. The southern domain is characterized by weak strain intensity, more abundant quartz veining. Drilling in the northern domain hole HS-15-28 contains 1.25 g/t gold over 15.35 metres, whereas hole HS-15-27, drilled in the southern domain, contains 3.11 g/t gold over 8.1 metres.



Figure 8 Hornblende Shear Zone exposure

Stop 7: William Gold Zone

Coordinates: UTM 16 NAD83 668093 mE, 5317233 mN

This stop illustrates the cryptic style of gold mineralization associated with the William gold zones. The controls on the spatial distribution of the gold on the outcrop, and within the William gold zones in general, remain to be clearly established. Tectonic fabrics of the Wawa Gold Corridor are moderately to weakly developed on the outcrop and well-evident in the mafic dykes in which a strong stretching lineation is developed, but the relation between the tectonic fabric and gold distribution could not be unequivocally established.

Stop 8: Root Vein

Coordinates: UTM 16 NAD83 668775 mE, 5318459 mN

This stop illustrates a gold-bearing quartz vein formed in a ENE-oriented shear zone. The overall style of gold mineralization characteristic of the Root Vein is similar to the structures mined in the Parkhill and Van Sickle mines. The outcrop also exposes at surface an example of a lamprophyre dyke that is part of the lamprophyre dyke swarm of the Wawa Gold Property; and provides good exposures of the porphyritic and coarser-grained intrusive facies of the Jubilee Stock, and of the tension veins formed in the Wawa Gold Corridor.



Figure 9 Core of the Root vein
Length of channel sample is 1.5 metres

Stop 9 Darwin-Grace Mine site and exposure of Grace Deformation Zone

UTM Coordinates: UTM 16 NAD83 668077 mE, 5313382 mN

This stop illustrates the site of the historic Darwin-Grace mine and an exposure of the Grace Deformation Zone. The exposure of the Grace Deformation Zone illustrates the strong brittle-ductile deformation and arsenopyrite flooding with quartz veining characteristic of that structure. Grab samples collected from the Grace Deformation on the outcrop contain between 1.27 g/t and 18.4 g/t gold.

Stop 10 Gold zones of the Wawa Gold Corridor (Red Pine Exploration core shack)

This stop, at the Red Pine core shack, will present a suite of drill core intersections to illustrate the main gold zones discovered so far in the Wawa Gold Corridor. This will include drill intersections from: the Jubilee Shear Zone, the Hornblende Shear Zone, the William gold zones, the Minto shear zones and the Grace Deformation Zone, and some of the gold-bearing tension veins.

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Field Trip 4

Geology of the Island Gold Mine

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Introduction

This field trip will provide participants with an overview of the local property geology of the Island Gold Mine which is situated 83 kilometers north-west of Wawa. The trip will start at the Island Gold Mine coreshack where core will be displayed from ore intercepts within the mine horizon across a 2 kilometer wide strike length and then the visitation of outcrop exposures along a 4 kilometer segment of the immediate mine stratigraphy. Elements of the Goudreau Lake Deformation Zone brittle-ductile deformation which include shearing, folding and veining will be seen through many of the exposures in the area including the gabbro sill hosting the Kremzar Deposit, highly Z folded Michipicoten Iron formation at the Morisson No. 1, the exposed south east extension of the Island Gold mine structure and within the Webb Lake Sill and Stock which hosts the Magino Mine and Argonaut Gold open pit resource. The Goudreau-Lochalsh area is a historic gold camp with the first gold discovery in 1918 by D. J. McCarthy and Mr. W. J. Webb within the Webb Lake Stock and later in 1926 by Patrice Kremzar on what is now the Richmond Island Gold Mine Property. In the 1930's the Cline Mine was in production 8 kilometers east of the Island Gold mine and eventually produced 63,000 ounces of gold.

Regional Geology

The Island Gold deposit located in the Wawa Sub-Province of the Superior Province of the Canadian Shield. The Wawa Subprovince is composed of numerous east-west trending sublinear supracrustal belts of greenstone enclosed by larger regions of granitoid rocks. Greenstone belts are dominated by metavolcanic and metasedimentary rocks ranging in age from 2.9 to 2.7 Ga (Percival, 2012).

The Island Gold Deposit occurs on the north central flank of the Michipicoten greenstone belt which has produced over 3 million ounces of gold since the first 'gold rush' in Wawa in the late 1890's. The greenstone belt has dimensions of 145 km by 45 km in dimensions and lies 150 km south east of the prolific Hemlo Mining Camp. It has been subdivided into three cycles of bimodal volcanism aged 2.9, 2.75 and 2.7 Ga respectively, referred to as the Hawk, Wawa and Catfish assemblages (Turek, 1994, Sage 1994). Volcanism in the lower and upper cycles are interpreted by Sage (1994) to have been Hawaiian and Plinian types respectively. Each of the three cycles is capped by pyrite bearing iron formation with an unconformity separating the volcanic rocks from the overlying iron formation in cycles 2 and 3. The Dore metasediments, composed of wackes and conglomerates, unconformably overlie the 2.7 Ga cycle and are the youngest rocks in the belt. The metamorphic grade varies from upper greenschist to lower amphibolite at the margins of the belt in proximity to the external granitoid rocks. The belt is interpreted to consist originally of a monoclonal sequence that was subsequently underwent complex thrusting, folding and shearing (Sage 1994). Arias and Helmstaedt (1990) interpret the present configuration of the belt as a regional nappe fold refolded about an F2 axis and then imbricately thrust into numerous slices.

Island Gold Mine Michipicoten Greenstone Belt

Legend:

- Proterozoic Plutonic and Volcanic Rocks
- Archean Granitoid Plutons
- Archean Metasedimentary Terranes
- Archean Greenstone Terranes
- Fault
- Past Gold Producer
- Gold Producer

Map Labels: Lake Superior, Marathon, White River, HEMLO, EAGLE RIVER MINE, ISLAND GOLD MINE, Canada, USA.

Scale: 25 km

Inset Map: Ontario, Quebec, Montreal, Toronto, 250km

Figure 1: Michipicoten Greenstone Belt (MGB)

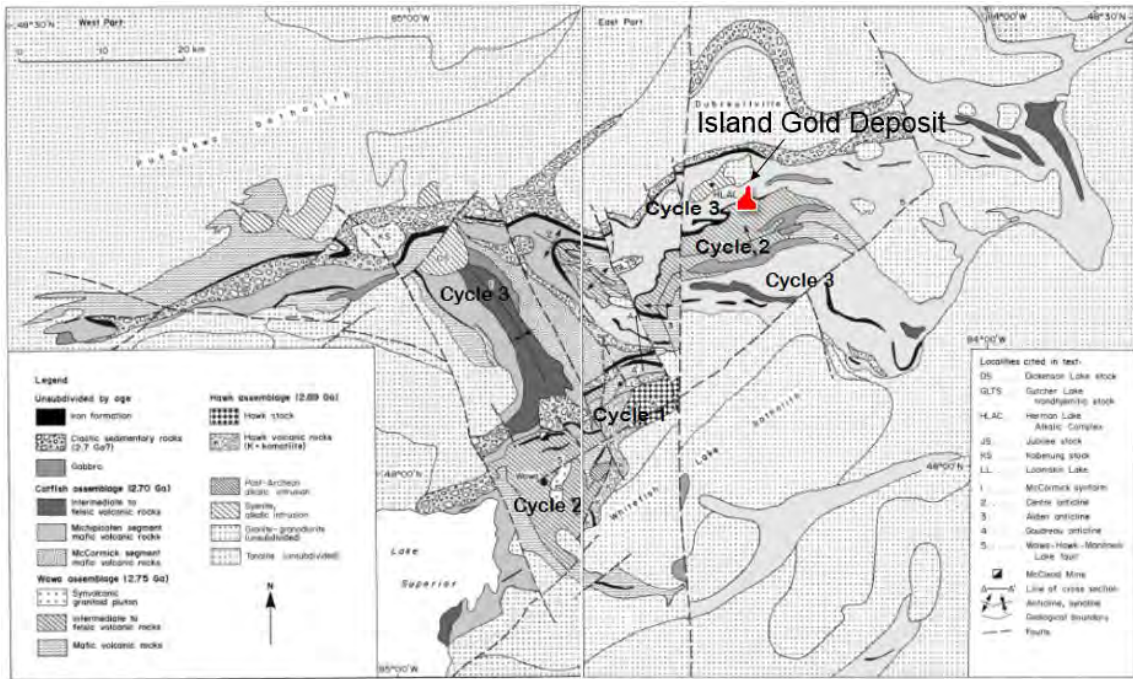


Figure 2: MGB Geology and Structure

The main structural control on most Au occurrences in the Goudreau-Lochalsh area in which the Island Gold deposit is situated is the Goudreau Lake Deformation Zone (GLDZ). This comprises a 30 kilometer long by 4.5 kilometer wide deformational structure described by Heather and Arias (1992) as a dextral oblique slip structure occurring near the interface between the Wawa and Catfish volcanic cycles. The deformation zone has a gentle sigmoidal shape and is not a focused single structure but a composite of smaller scale structures with development of deformation being more controlled by hosting geological units such as formational contacts where rock competency contrasts exist and less by external regional stress. (Dube B., and Gosselin, P., 2007). The Cradle Lake Deformation is situated south of the GLDZ on the south arm of the Goudreau Lake Anticline near the interface of cycle 2 and 3 volcanic assemblages.

The deposits of the Goudreau Lochalsh camp have a strong spatial correlation with a variety of felsic intrusive rocks including trondhjemite stocks and sills, felsic porphyries and dyke complexes which is common to many Superior Province gold deposits (Hodgson and MacGeehan, 1982). Exceptions occur including the Kremzar Deposit (47,000 oz. 1988-90) which is hosted by a regional gabbroic sill. The Island Gold deposit is situated within the cycle 2 volcanic assemblage however there is close spatial association with both mafic, felsic dykes and massive felsic volcanic rocks within the mineralized gold bearing corridor. Competency contrasts and rheologic variations between the more and less competent units provide planes of weakness along which shearing can develop, structural traps can open and gold bearing veins can subsequently be formed. The iron rich chemistry of proximal mafic dykes are also considered to assist in gold formation by providing a chemical trap in the Island Gold system. Other factors include the recognition of an antiform on the north limb of the Goudreau Lake Anticline in which the Island Gold mine sequence appears to be situated in adding to structural complexity of parasitic folding and flexural slippage and associated shearing and thrusting along fold limbs.

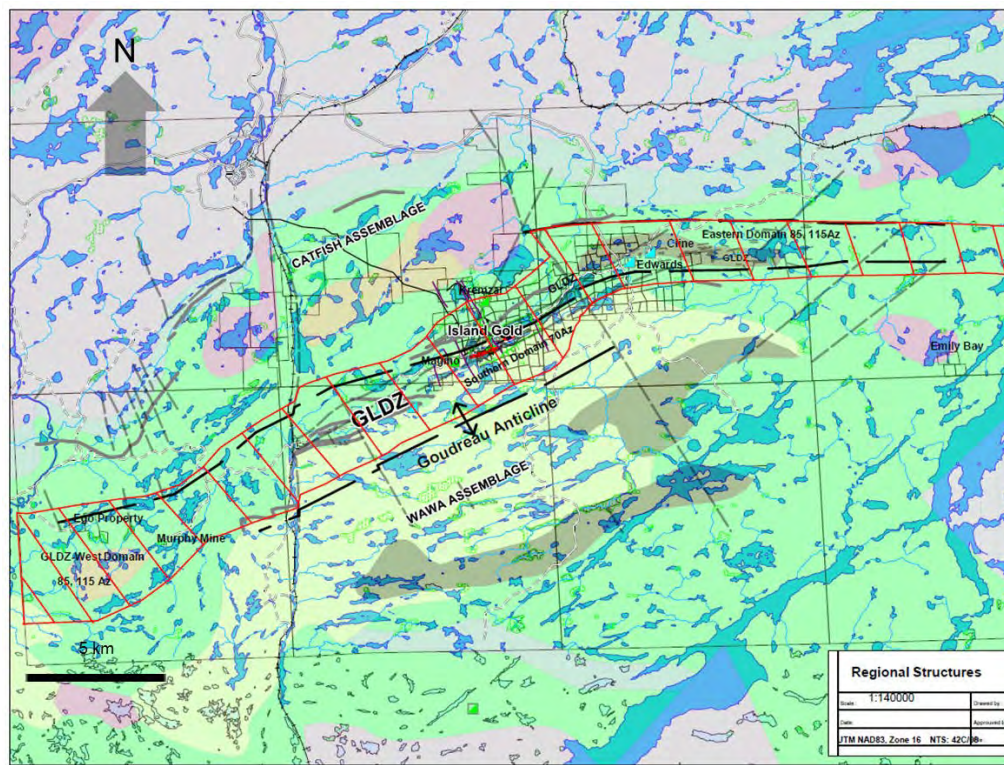


Figure 3: Goudreau Lake Deformation Zone (GLDZ)

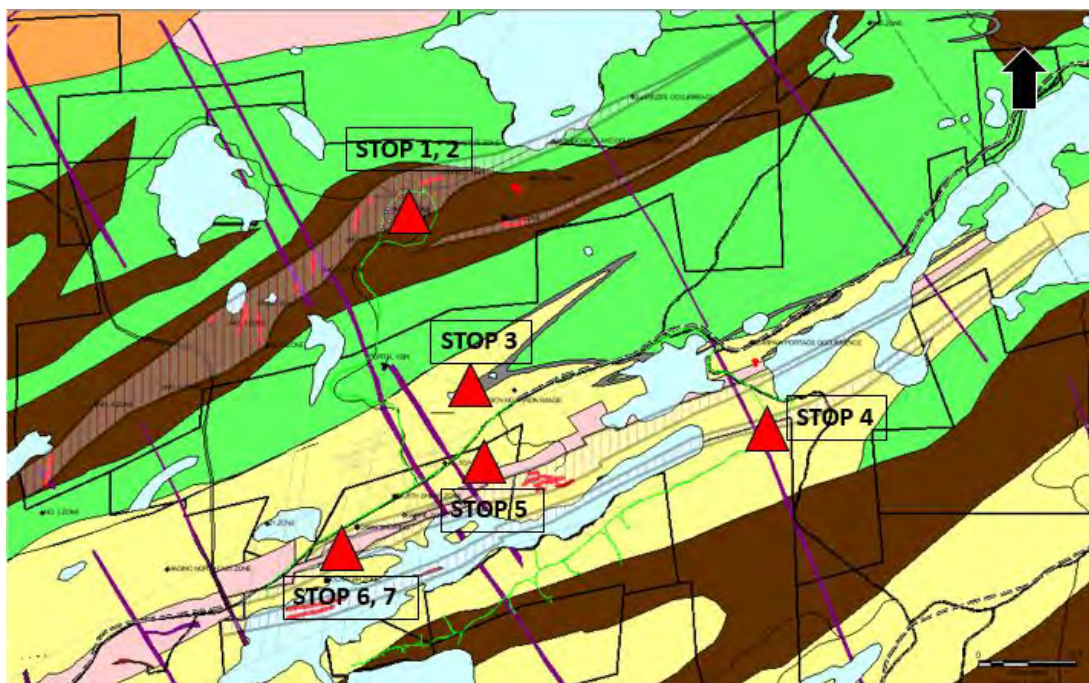


Figure 4: Island Gold Stop Locations

STOP 1: The Island Gold Mine Coreshack

Location: UTM NAD83 16U 690538mE, 5353660mN

Description: Since the beginning of production in 2007 at the Island Gold Mine Richmond Mines has produced 2.1 MT of ore at 6.46 g/t totaling 433,000 ounces of gold. The Island Gold Mine is an Archean Lode gold deposit and consists of seven main narrow sub-parallel east-northeast striking, steeply south dipping, stacked quartz vein systems carrying gold mineralization within envelopes of intense sericitization, silicification and carbonatization with 2% to 5% pyrite which can be up to 8 meters wide. (Adam et al., 2015). The mineralized zones are sub parallel and can anastomize thereby merging or splitting along strike. Main veins styles include laminated veins with a defined ribbon appearance (Fig.7), sheeted veinlet zones composed of millimeter to centimeter scale quartz carbonate stringers, massive smokey greyish veins (Fig.5) and several types of milky white extensional veins which can appear as blow outs, gashes or horizontal ladder veins. Quartz veins at depth can be over four meters wide (Fig.7). The mineralized corridor increases from 50 meters wide above the 400L to over 100 meters at depth. Gold mineralization discovered at depth now constitute the principal source of ore and represent the down dip continuation of the mineralized zones which were mined before 2016.

As of December 31, 2016, total Proven and Probable Reserves at the Island Gold Mine were 2.5 MT at a grade of 9.17 g/T for 752,200 gold ounces. The estimated Measured and Indicated Resources totaled 0.47 MT at a grade of 5.94 g/T for 91,450 ounces. Estimated Inferred Resources totaled 3.0 MT at a grade of 10.18 g/T for 995,700 ounces.

Diamond drill core will be displayed from intercepts across approximately 2 kilometers of the mine horizon which will include intercepts of mineralization, quartz veining and alteration together with accompanying longsection and cross sections for location and geology. Core representatives of footwall and hangingwall rock will be presented as well as representative rocks types within the sequence.

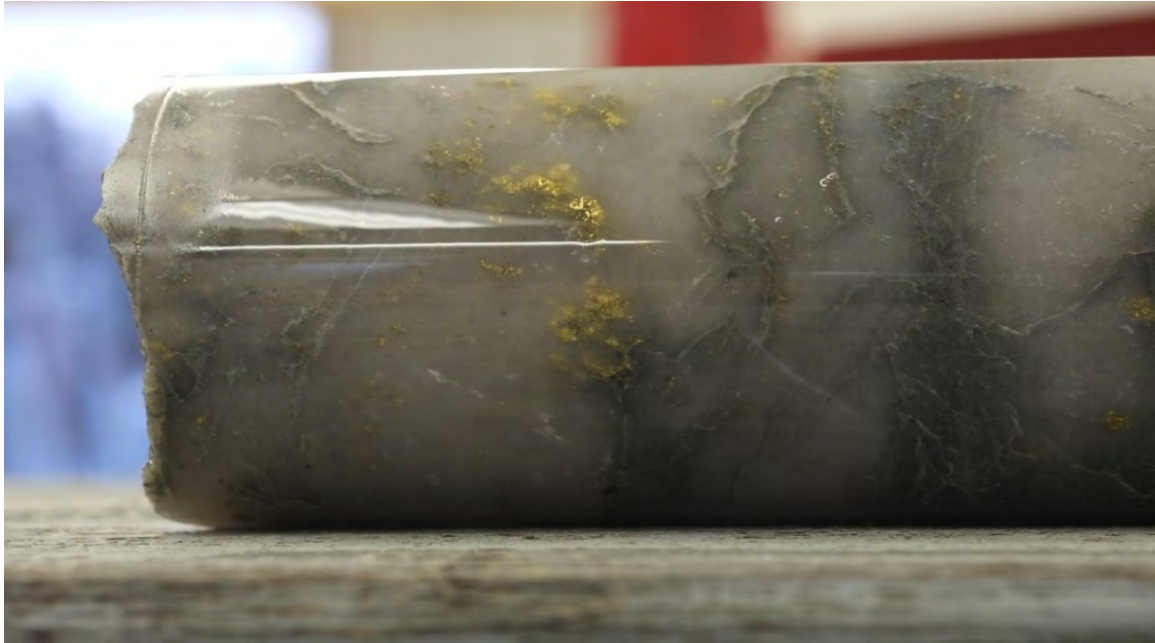


Figure 5: C zone quartz vein



Figure 6: GD-14-01C, 19.9 g/t over 3.9 meters true width, -1250 meters

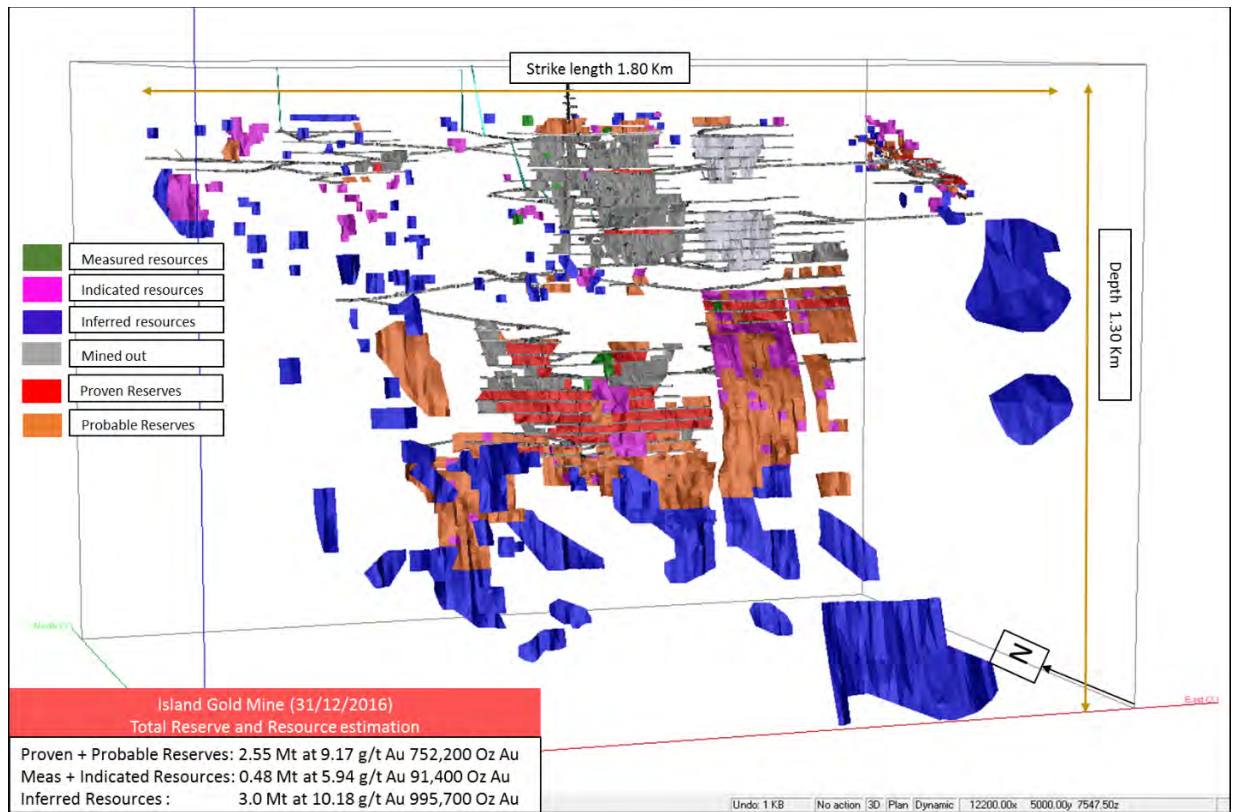


Figure 7: Isometric Diagram Island Gold Mine Looking North East



Figure 8: 800 Level quartz veins on 4 meter wide face (42.25g/t – 325.49 g/t uncut)

STOP 2: Heritage Outcrop

Location: NAD83 16U 690478mE, 5353601mN

Description: The Heritage outcrop is situated in the younger northerly Catfish Volcanic Assemblage (2.9 Ga) which is primarily mafic in character facing north, dipping steeply north and striking 060°-070°. This outcrop is the discovery outcrop of the Kremzar Mine which was in production by Canamax between 1988-1990 and produced a total of 47,000 ounces at a grade of ?. The Kremzar deposit is hosted by a regional gabbroic sill which extends across the property to the Maskinonge Lake fault to the east. Gold was first discovered on the property in 1926 by Patrice Kremzar in a piece of float on the old Pic road. The property contains over 17 gold showings of varying economic potential.

The Kremzar orebody occurs within sheared and altered gabbroic intrusive rocks. Kwok (1986) describes the alteration as an outer chlorite-biotite-carbonate zone and an inner zone dominated by sericite-biotite-potassium feldspar-Fe-carbonate. The mineralized sheared areas of rock weather a rusty brown. The main Kremzar mineralization is comprised of two sheared hosted quartz vein systems. The main mineralized zone called the R zone is situated at the west of the exposure and occurs within a 120-140° striking shear zone (dextral movement), dips steeply southwest and plunges 70° north east. A secondary B zone is situated to the north west of the exposure. Outside of the main mineralized shear zone containing the R and B zones the gabbroic rocks have undergone a complex system of brittle fracturing with fracture controlled Fe carbonate alteration and quartz filled fractures dominate the veining with the primary orientation at 30° (sinistral movement) and a secondary orientation of 120-140°, both orientations may contain anomalous gold values. Sharp boundaries exist between the altered sheared rock corridor and the surrounding unaltered gabbro which bound them.



Figure 9: Heritage Outcrop R zone

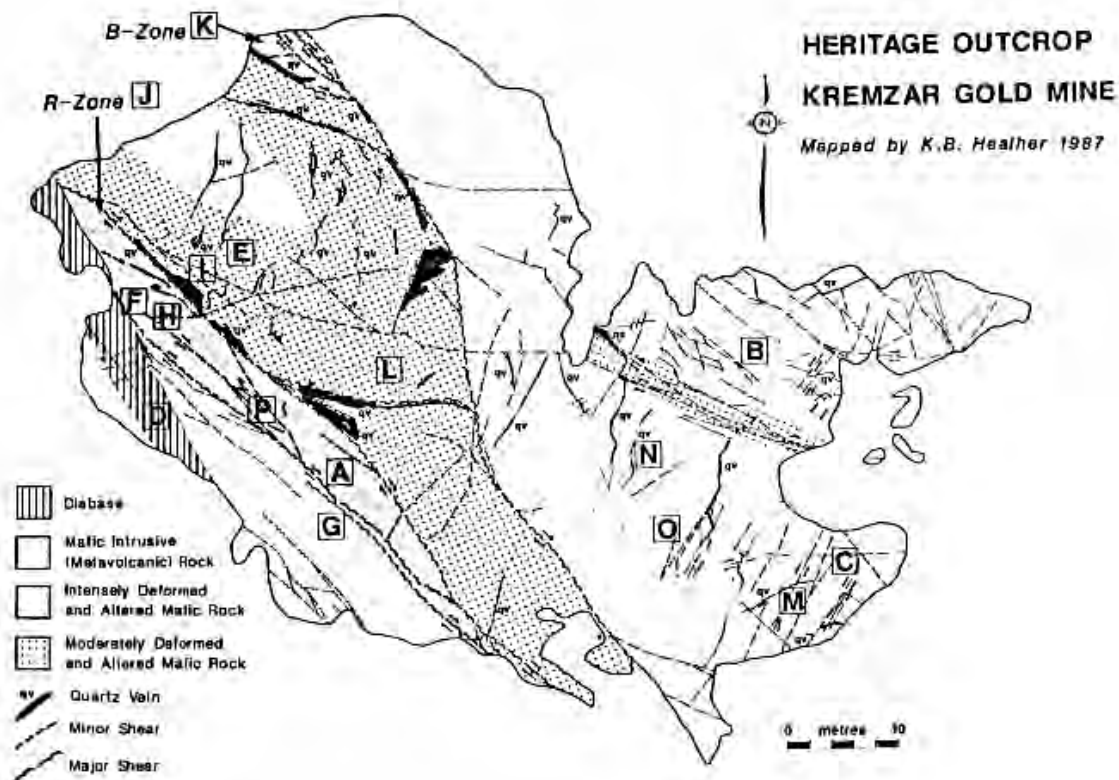


Figure 10: Heritage Outcrop Map

A – intensely sheared and altered gabbro

B – Undeformed gabbro

C – clot size coarse grained amphibole and plagioclase crystals

D – Diabase 160° azimuth

E- Outer alteration zone, cl-bt-cb

F – Inner alteration zone se-bt-Kspar-fe cb

G – sharp boundary between alteration and relatively unaltered rocks

H – 120°-140° striking shear zone

I – R zone vein

J – R zone

K – B zone

L – R and B zone appear to converge

M – brittle fracturing

N – myriad of narrow quartz filled fracture fillings

O – strong fe carbonate

P – brittle fractures strongly disrupted in ductile shear

STOP 3: Morrison No.1 Iron Formation

Location: NAD83 16U UTM 690775mE, 5352598mN

Description: This stop is part of the Goudreau Iron Range, a Michipicoten type iron formation, which occurs at the interface between Cycle 2 (Wawa Assemblage, 2.75 Ga) and Cycle 3 (Catfish Assemblage, 2.9 Ga) volcanics. The iron formation is in sharp conformable contact with the southerly crystal tuffs of the Wawa Volcanic assemblage which it overlies. The Goudreau iron formation is a thickly bedded exhalative chemical sediment which can contain a minimum of five facies (Sage 1994) including a lower siderite or calcite unit, pyrite unit, chert-magnetite unit, chert-wacke unit and an uppermost argillite-pyrite-graphite. At the eastern outcrop of the Morrison No.1 a stratigraphic sequence of siderite bearing ironstones (10m) overlain by massive pyrite (3m) in turn overlain by chert-pyrite-magnetite beds (20m) at the top (north) can be seen. Oxygen isotope data indicates iron formation deposition occurred under 200° C (LeSeeleur, 1980) suggesting a distal low temperature regime of formation and that depositing fluids were too low to transport base metals (Edmond et al, 1974).

Randomly oriented chloritoid porphyroblasts within the iron formation are prominent and range up to 1 cm in diameter occurring pyroclastic fragments as well as the footwall felsic volcanic rocks to the south. Chloritoid alteration is a common alteration mineral in the Wawa iron ranges. Chloritoid bearing pyroclastic fragments within the overlying the Morrison iron formation are present and common in outcrop. Lockwood (1983,1986) determined chloritoid-bearing rocks where a stratiform alteration pattern resulted from unfocused fluid discharge and overall low water/rock ratios under low pressure in subaqueous to subaerial conditions. Sage suggests that regionally the great strike extent of chloritoid bearing units indicates a hydrothermal system controlled by rock permeability along a broad aquifer.

Shear zones are present throughout the outcrop as narrow discrete zones displaying dextral movement. S and Z symmetry folds are also present in the Morrison No.1 exposure. Z folds may be interpreted as dextral shear folds and the S folds possibly suggesting parasitic folding related to a possible antiformal structure in the area. Minor folds in the iron formation are present and plunge shallowly to the west. Other exposure features of note include boudinaged mafic dykes which 'float' within less competent siderite ironstones but which appear to be one original mafic dyke unit which has been tightly Z folded. The Morrison No.1 has as well been highly Z folded in the immediate area as depicted by geophysical interpretation. Regionally along strike however the iron formation can be highly transposed or distended.

The gold mineralization is fracture controlled and quartz filled fractures can be seen developing at the west end of the outcrop area. A historic non-compliant 43-101 resource for the Morrison #1 is about 540,000 tons @ .05 g/t gold (T. Foster, 1982).



Figure 11: Chloritoid in Mafic Dykes



Figure 12: Sulphide Facies Iron Formation

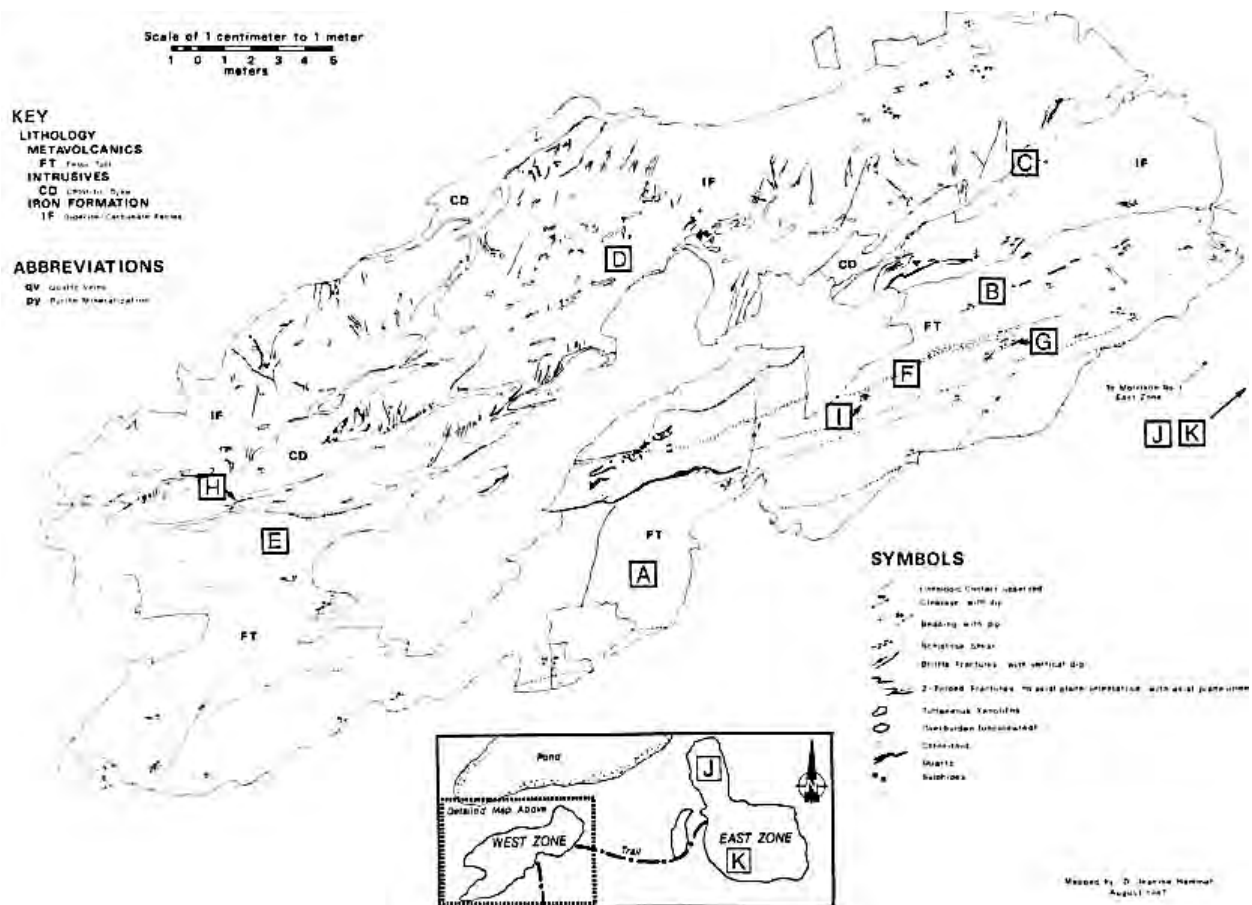


Figure 13: Morrison No.1 West Zone

A – footwall volcanic rocks

B – abundant chloritoid

C – boudinaged mafic dyke in siderite ironstone

D – py bearing ironstone + pervasive Fe-carbonate

E – strongly siderized tuffaceous rocks

F – discrete shears and dextral displacement

G – Z symmetry folds

H – S symmetry folds

I – crenulation cleavage suggesting oblique north side up dextral displacement

J – siderite overlain by pyrite overlain by chert-pyrite-magnetite at the top

K – narrow quartz veinlets crosscut iron formation

Stop 4: South East Mine Structure Stripped Area

Location: NAD83 16U 692318mE, 5352439mN

Description: This stripped area is situated at the south eastern end of Goudreau Lake. It represents the only exposed segment of the Island Gold mine structure to date as the producing mine is presently situated beneath Goudreau Lake. The stripped exposure consists of a sequence of fragmental volcanic rock units succeeded to the south by a highly sheared, friable zone of sericitized felsic rocks in turn succeeded by massive felsic volcanic rock units. Local mafic dykes can occur throughout the sequence. Quartz veining is present locally and can be up to a meters in width. The shear has been correlated with the eastern strike extension of the Island Gold mine structure. At this location the structure is dipping steeply north and is striking approximately 70°.

Felsic volcanic rocks in the southern part of the exposure area are a generally massive, fine grained, white weathered type and can contain clear colorless rounded quartz eyes up to 5%. The felsic volcanic rocks within the mine sequence to the west have a rock chemistry that straddles the border of the Rhyodacite-Dacite and Andesite field in the Winchester and Floyd (1977) diagram (T.Ciufo, 2017). The felsic rocks place in the dacite field of a TAS diagram (T.Ciufo, 2017). The fragmental volcanic rock unit in the north part of the stripped exposure is generally oligomictic, matrix supported and not well sorted. Variable amounts of whitish lensoid shaped lapilli fragments range in size from 4mm to several centimeters with compositions from felsic to intermediate with rare mafic clasts on occasion. Locally larger subangular to angular fragments up to 20 cm are present. The chloritoid alteration can commonly be seen within the fragments and matrix. Chloritoid bearing fragments can locally been seen suspended in chloritic units of either sediment or dyke rocks.

Rock competency contrast is suggested to have created planes of weakness and therefore promoted shear propagation along these boundaries as one part of the mechanism in the gold forming process at Island Gold. In the exposure this seems to be demonstrated as a highly sheared felsic rock zone occurs between a southerly massive felsic volcanic unit to the south and a 'less competent' volcanic breccia unit to the north.



Figure 14: Island Gold Mine Structure SE of Goudreau Lake

Stop 5: Webb Lake Sill (optional)

Location: NAD83 16U 689841mE, 5351605mN

Description: The Webb Lake Stock is a trondhjemite intrusion situated 250 meters north of the Island Gold Mine structure. The stock has been mapped by Sage (1994) as being two rock masses composed of a southerly and northerly lobe which are divided by an intervening unit of a fine grained mafic intrusive or possibly metavolcanic. The main body of the stock extends for a strike length of over 2 km but continues into the Richmond Property to the east with a more sill like geometry traceable for 5 kilometers. The Webb Lake Stock dips 60-70° north and strikes approximately 70°. The intrusive is generally fine grained, hypidiomorphic granular with 65% plagioclase and 35% quartz with minor carbonate, chlorite and tourmaline. Sericite is common and potassium feldspar is rare.

The outcrop exposure exhibits Z folds which display shallow 240° oriented fold hinges and north dipping axial planes. This structure is interpreted (K. Jellicoe, 2017) to have formed during D2 deformation during a NW-SE south-up dextral transpression. A late extensional tourmaline vein is present on the outcrop which cross cuts a previously developed D1 foliation as floods the planes of foliation.



Figure 15: Z fold on south flank of Webb Lake Sill – Vent Raise Outcrop

STOP 6: Webb Lake Trondhjemite Stock (Optional)

Location: NAD83 U16 689416mE, 5351240mN

Description: This stop is on the Webb Lake stock and hosting the gold mineralization at the Magino deposit. Gold was first found by Mr. D. J. McCarthy and Mr. W. J. Webb in 1917. Inferred reserves are estimated at 105.4Mt at 0.89g/t Au. The long axis of the north dipping stock is parallel to the regional supracrustal stratigraphy and has a strike length of approximately 2 kilometers and widths up to 300 m. The stock was emplaced in mafic volcanic rocks from the lower part of the Catfish assemblage (Cycle 3). It is called a granodiorite by Argonaut (43-101, 2016) or a trondhjemite by Sage (1994). The gold is hosted in numerous quartz veins and in quartz-flooded bands, parallel to the Goudreau Lake Deformation Zone (Argonaut 43-101, 2016). Silicification, sericitisation and an increase in pyrite content are typical around the mineralized veins and quartz flooded bands. This outcrop exhibits a mostly massive intrusive with numerous thin more ductile bands, and a few thin mineralized quartz veins that can be followed on a few meters.

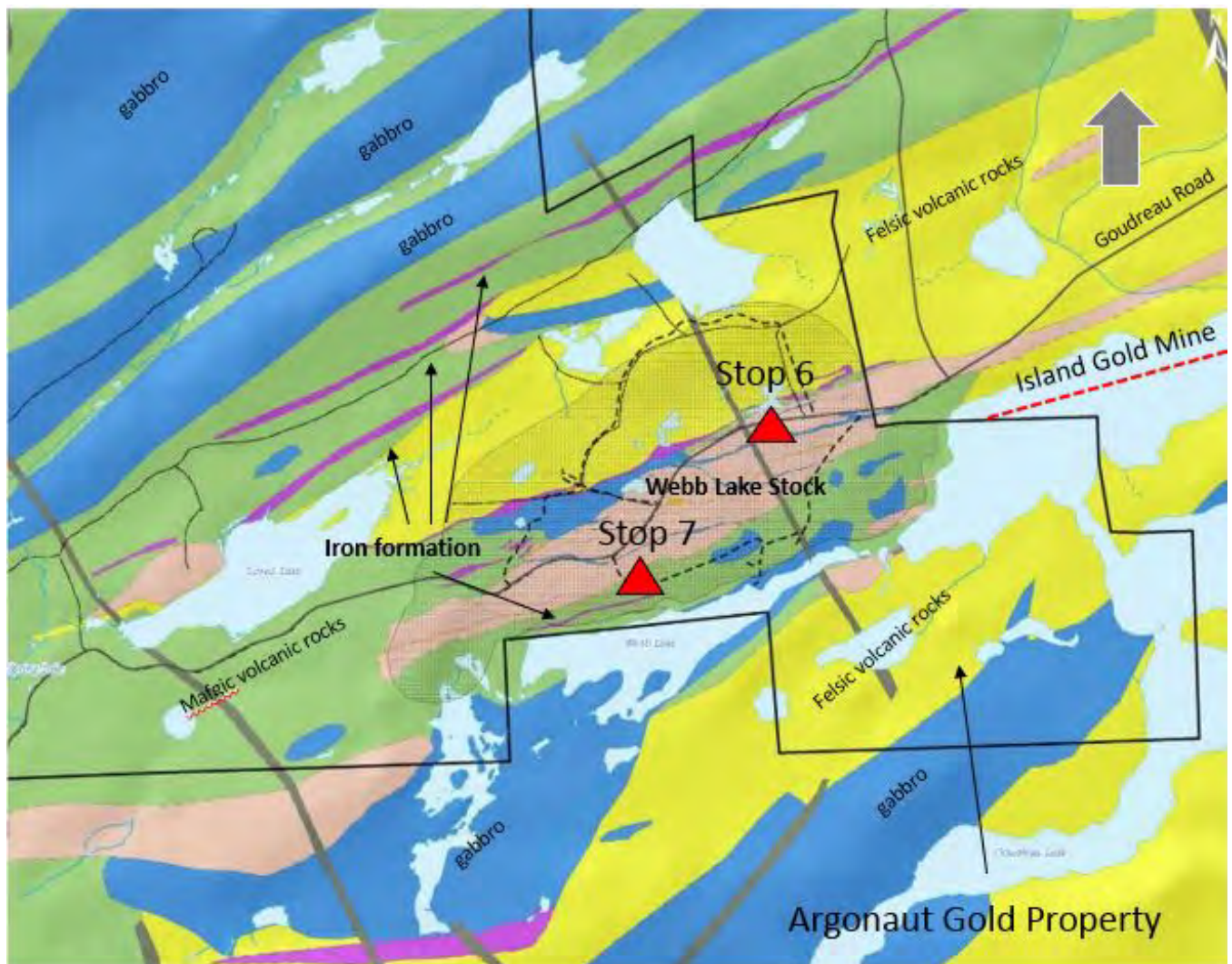


Figure 16: Webb Lake Stock Stop Locations



Figure 17: Webb Lake Trondhjemite, small scale auriferous veins



Figure 18: Webb Lake Trondhjemite C/S fabric

Stop 7: Iron Formation (Optional)

Location: NAD83 U16 688700mE, 5350900mN

Description: Upper level of the Michipicoten Iron Formation containing banded layers of chert, magnetite, pyrite with intercalated argillaceous cycle three mafic volcanic rocks. The Webb Lake stock is located less than 30 meters north.

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FIELD TRIP 5

Geology of the Renabie Mine area, Michipicoten greenstone belt

Lise Robichaud (Ontario Geological Survey)
Jordan McDivitt (Laurentian University)

Introduction

This field trip focuses on the northeast portion of the Michipicoten greenstone belt and includes the past-producing Renabie Mine, situated within the Missanabie–Renabie gold district (Figure 1). The bedrock geology of the field stops is based on recent mapping of the Renabie Mine area completed by the Ontario Geological Survey (Robichaud, McDivitt and Trevisan 2017; Robichaud et al. 2016; Robichaud et al. 2015; Robichaud, McDivitt and Trevisan 2015; Robichaud and McDivitt 2014). Additionally, detailed outcrop maps for Stops 3, 4, 5, and 15 are modified from a recently completed MSc thesis: *Gold mineralization in the Missanabie–Renabie district of the Wawa Subprovince (Missanabie, Ontario, Canada)* (McDivitt 2016a). An accompanying Miscellaneous Release of Data (McDivitt 2016b) also provides additional information. Readers should consult publications referenced in this guidebook for additional information and references. The order and number of stops that will be visited during this field trip will be determined by access, time, and weather concerns.

Regional Geology

The Michipicoten greenstone belt consists of successions of Archean metavolcanic and metasedimentary rocks intruded by Archean granitic rocks (Turek, Smith and Van Schmus 1982) and younger Proterozoic mafic dikes. The supracrustal rocks of the belt have been previously subdivided into 3 distinct volcanic cycles: 2900 Ma, 2750 Ma and 2700 Ma (Sage and Heather 1991; Heather and Arias 1992; Turek, Smith and Van Schmus 1982, 1984; Turek, Van Schmus and Sage 1988). New geochronological data from recent mapping suggest that there is a previously undocumented volcanic cycle ranging in age from 2731 to 2723 Ma (Kamo 2014, 2015, 2016), which indicates that the volcanic history is more continuous than was previously believed.

The Michipicoten greenstone belt has been previously interpreted as a continuation of the Abitibi greenstone belt west of the Kapuskasing Structural Zone (Ayer et al. 2010). In comparison, the Abitibi greenstone belt consists of stratigraphically continuous sequences of Archean metavolcanic rocks and metasedimentary rocks ranging from earlier than 2750 Ma to 2695 Ma followed, in Ontario, by 2 sedimentary basins: Porcupine-type at 2690 to 2682 Ma and Timiskaming-type at 2676 to 2670 Ma (Ayer et al. 2002; Ayer et al. 1999a; Ayer et al. 1999b; Ayer, Ketchum and Trowell 2002; Ayer et al. 2005). The second (2750 Ma) and third (2700 Ma) volcanic cycles of the Michipicoten greenstone belt, as well as the 2731 to 2723 Ma volcanism, may represent time-equivalent volcanism similar to the Abitibi greenstone belt; however, no volcanism of 2900 Ma (equivalent to the first cycle) has been described in the Abitibi greenstone belt.

Geological Setting of the Renabie Mine Area

This field trip targets the northeast portion of the Michipicoten greenstone belt including the past-producing Renabie Mine. This area is composed of metamorphosed mafic to felsic volcanic and volcanoclastic rocks, interspersed with metasedimentary rocks. Large-bodied, felsic to intermediate granitoids intrude the Archean supracrustal rocks. These include the Missinaibi Lake batholith, which

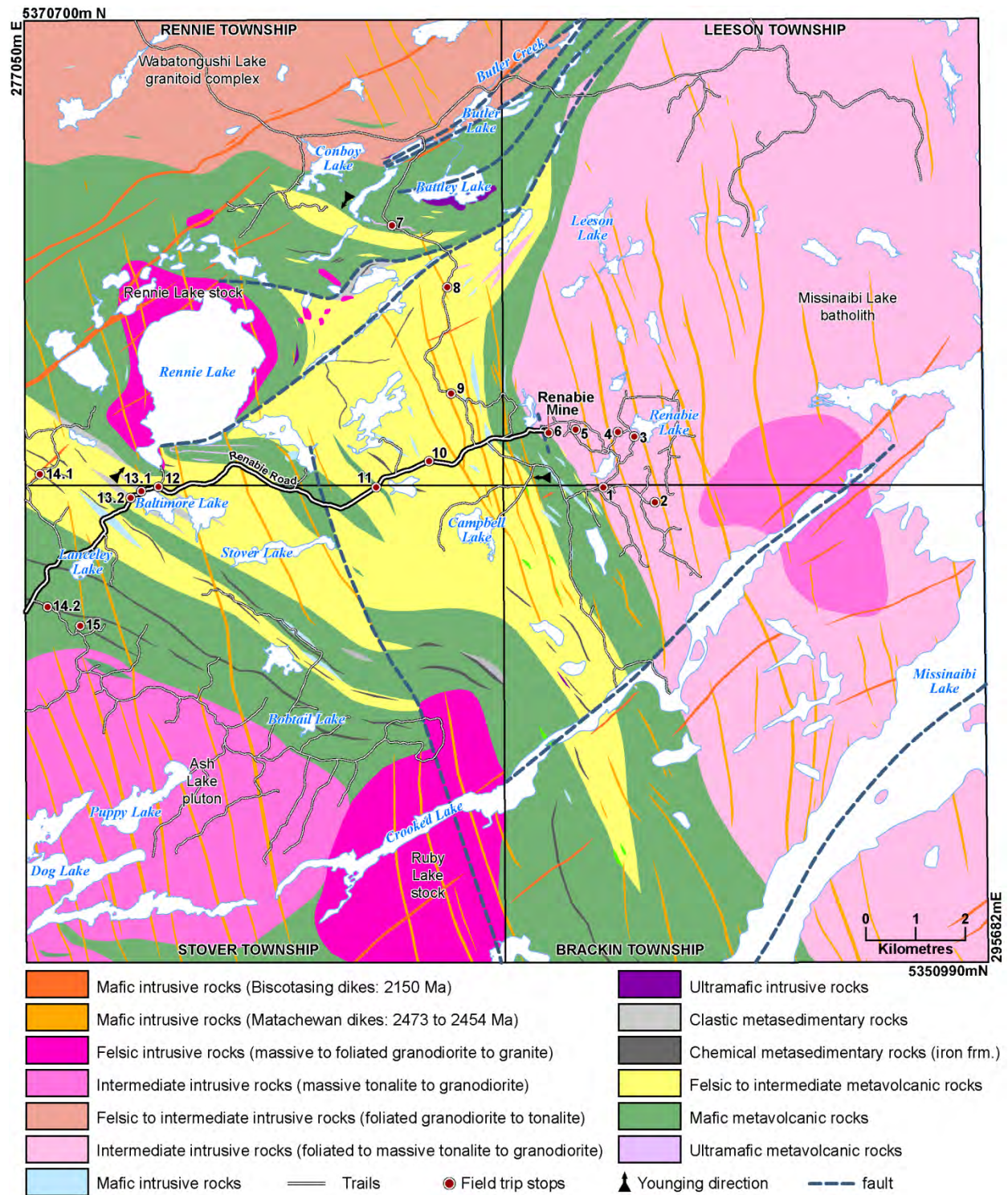


Figure 1. Simplified geological map of the northeast part of the Michipicoten greenstone belt based on recent mapping by the Ontario Geological Survey (Robichaud, McDivitt and Trevisan 2017; Robichaud et al. 2016; Robichaud et al. 2015). Rennie, Leeson, Stover and Brackin townships are included in the map area. Younging direction symbol indicates stratigraphic facing. Field trip stop locations are indicated on the map. The Renabie road and the Renabie Mine are also denoted for reference purposes. All UTM co-ordinates are provided in Zone 17 using NAD83.

dominates the eastern portion of the area, the Wabatonigushi Lake granitoid complex, which occupies the northern portion of Rennie Township, and the Ash Lake pluton, which occupies the southwest portion of Stover Township. Archean felsic stocks, sills, and dikes, as well as Proterozoic gabbroic dikes, are also intrusive to the Archean supracrustal rocks (see Figure 1).

The only past-producing gold zones in the Missanabie–Renabie gold district are located in the Renabie Mine area, within tonalite of the Missinaibi Lake batholith near its contact with metavolcanic rocks of the Michipicoten greenstone belt (see Figure 1). The gold zones were mined at the Renabie underground mine and at the C-zone, Nudulama, and Braminco open pits. Mining occurred over a 50 year period, from 1941–1991, and produced roughly 1.1 million ounces of gold at a grade of approximately 6.6 g/t (Turek et al. 1996; Callan and Spooner 1998). The mineralized zones from the Renabie Mine to the Nudulama East outcrop define an east-southeast trend, referred to as the “Renabie trend”, which strikes close to perpendicular to the regional foliation. Assay-defined ore shoots along the Renabie trend have eastern to east-southeastern trends and a moderate westerly plunge (McDivitt 2016a). The Braminco zone, which is located 1.1 km south of the Renabie trend, differs in orientation: the zone is parallel to the regional foliation, which strikes north-northwest and dips steeply to the west. Ore shoots within the Braminco zone plunge moderately to the west, similarly to those along the Renabie trend (Callan and Spooner 1998).

Metavolcanic Rocks

Mafic metavolcanic rocks are the dominant lithology in the central portion of the area and are interlayered with felsic to intermediate metavolcanic rocks and metasedimentary rocks (see Figure 1). Massive flows are predominant, but pillowed flows are observed locally. The pillows are well to poorly developed and are up to 1 m in diameter. The mafic metavolcanic flows are dark grey to black on fresh surfaces and are typically fine-grained and well foliated. The mafic metavolcanic rocks are typically characterized by greenschist-facies metamorphic mineralogy, although, near the margin of the felsic to intermediate intrusive rocks, the metamorphic mineralogy is representative of amphibolite facies. Where the Archean supracrustal rocks are wedged between the Missinaibi Lake batholith and the Wabatonigushi Lake granitoid complex, the rocks are locally migmatized.

Felsic to intermediate metavolcanic rocks occur largely in the southern half of Rennie Township and extend south into Stover and Brackin townships (see Figure 1). They are generally fine- to medium-grained volcaniclastic rocks dominated by tuffs and crystal tuffs. Tuff-breccias occur locally in the northern part of Stover Township. The fresh surface of all these felsic to intermediate rocks is light grey and weathers to a lighter grey to beige colour. Bedding is rarely observed and ranges in thickness from finely laminated (a few millimetres) to thickly bedded (up to 1 m). Turek et al. (1996) reported a U/Pb age of 2740 ± 8 Ma for felsic metavolcanic rocks in the southeastern corner of Rennie Township; a U/Pb age of 2730.9 ± 1.2 Ma for the same rocks was reported by Kamo (2016) and Robichaud, McDivitt and Trevisan (2017). A U/Pb age of 2728.6 ± 1.1 Ma was reported for the felsic metavolcanic rocks in Brackin Township (Kamo 2015; Robichaud et al. 2015). A younger U/Pb age of 2704.6 ± 2.1 Ma for the more northern package of felsic metavolcanic rocks, in the north-central portion of Rennie Township was reported by Kamo (2016) and Robichaud, McDivitt and Trevisan (2017).

Metasedimentary Rocks

Clastic metasedimentary rocks are generally restricted to the southwestern part of Rennie Township, extending to the central part of Stover Township although minor occurrences can also be observed in central Rennie Township. These rocks consist of buff grey, quartz-rich, thinly bedded siltstone, sandstone, and conglomerate. Siltstone is the dominant sedimentary unit, but sporadic beds of sandstone and conglomerate also occur. The conglomerate is matrix-supported, but contains large cobbles of predominantly tonalitic composition with a few mudstone and/or siltstone cobbles. Bedding thickness

varies between sediment types: the siltstone tends to be thinly to thickly laminated, the sandstone beds range from a few decimetres to a metre in thickness, and the conglomerate beds are typically several metres thick. Facing reversals are implied by cross-bedding and graded bedding observed locally in some of the sedimentary rocks, indicating that the area is folded. A U/Pb age of 2695 ± 3 Ma for conglomerates in the northwestern corner of Stover Township was reported by Davis (2016).

Iron formation is defined primarily by geophysical evidence (Ontario Geological Survey 1999, 2002a, 2002b, 2003, 2011). Iron formation extends predominantly from the southwestern part of Rennie Township to the central part of Stover Township. Minor iron formation also occurs in east-central Rennie Township and in the southern portion of Brackin Township. The iron formation typically consists of magnetite-rich layers interlayered with siltstone or sandstone, forming beds that range in thickness from thin laminations to thin beds (≤ 10 cm). Disseminated sulphides were also noted within some of the clastic layers.

Archean Intrusions

The most extensive Archean intrusion in the map area is the Missinaibi Lake batholith (see Figure 1). It occupies most of the eastern half of the map area and is primarily composed of medium-grained tonalite to granodiorite with minor occurrences of granite. The main mafic mineral is typically biotite, but the presence of hornblende is noted in varying amounts locally. The batholith is generally moderately to well foliated, although some areas are characterized by weak fabric development. Kamo (2015) reports a U/Pb age of 2720.8 ± 1.4 Ma for a tonalitic phase of the Missinaibi Lake batholith in Brackin Township. Turek et al. (1996) reported a U/Pb age of 2741 ± 21 Ma for a tonalitic phase of the Missinaibi Lake batholith in Leeson Township. Younger phases of the Missinaibi Lake batholith occur as plutons, dikes, and stock-like features. The largest occurrence of these younger phases is at the boundary between Leeson and Brackin townships near Crooked Lake (see Figure 1). Younger phases of tonalitic composition tend to be finer grained than the main phase of the batholith and, in some instances, display the regional foliation, although other examples are noted to be massive. A sample of one of these younger tonalitic phases returned an age of 2688 ± 2 Ma (Kamo 2015; Robichaud et al. 2015).

The northern portion of Rennie Township is occupied by the felsic intrusive rocks of the Wabatongushi Lake granitoid complex (see Figure 1). These rocks vary in composition from tonalite to granodiorite and typically contain less biotite than rocks from the Missinaibi Lake batholith. There is a 500 m thick boundary between the granitoid complex and the mafic metavolcanic rocks to the south. This boundary is characterized by *lit-par-lit* injection of the mafic metavolcanic by felsic intrusive rocks. A U/Pb age of 2700 ± 1 Ma for the Wabatongushi Lake granitoid complex was reported by Kamo (2016) and Robichaud, McDivitt and Trevisan (2017). Turek et al. (1996) reported a U/Pb age of $2688 \text{ Ma} \pm 14 \text{ Ma}$ for the same intrusion.

The Ash Lake pluton (see Figure 1) occupies the southwestern portion of Stover Township and is composed primarily of medium-grained tonalite to granodiorite with minor occurrences of granite. The main mafic minerals are typically biotite $\pm$ hornblende. The rocks are generally massive to moderately foliated, although the margins of the pluton are characterized by stronger foliation development. Turek et al. (1996) reported a U/Pb age of 2679 ± 5 Ma for the Ash Lake pluton in West Township, situated to the west of Stover Township. Frarey and Krogh (1986) reported a U/Pb age of 2684.5 ± 2.7 Ma for the Ash Lake pluton in Copenace Township, located southwest of Stover Township.

The Rennie Lake stock, which occurs in the southwest portion of Rennie Township (see Figure 1), is composed of massive granitic to granodioritic rocks. The main mafic mineral is amphibole with minor biotite. The rocks are light pink on fresh surfaces and weather to creamy pink. A U/Pb age of 2678 ± 4 Ma for the Rennie Lake stock was reported by Davis (2016) and Robichaud, McDivitt and Trevisan (2017). Turek et al. (1996) reported a U/Pb age of ca. 2668 Ma (from a single zircon) for the Rennie Lake stock.

The Ruby Lake stock is interpreted to be temporally related to the Rennie Lake stock. The Ruby Lake stock, which occurs in the southeast portion of Stover Township (see Figure 1), is composed of massive, nonfoliated granitic to granodioritic rocks. The core of the Ruby Lake stock commonly displays potassium feldspar megacrysts, but the texture has also been noted near the intrusion's contacts. The main mafic mineral is amphibole with minor biotite. The stock is light pink on fresh surfaces and weathers from creamy to dark pink. Turek et al. (1996) reported a U/Pb age of 2661 ± 11 Ma for the Ruby Lake stock.

Several smaller mafic and ultramafic intrusions occur in the supracrustal rocks in the area (see Figure 1). The mafic rocks are typically gabbroic in composition and are medium- to coarse-grained, foliated, and typically amphibolitized. The ultramafic rocks differ in composition (from peridotite to pyroxenite and hornblende) and are variably serpentinized. They are often massive, fine- to coarse-grained, and sometimes display large euhedral crystals of pyroxene or hornblende.

Proterozoic Intrusions

Proterozoic mafic dikes are late features in the area. Most dikes trend north-northwest; however, a few trend northeast. The Proterozoic dikes generally have sharp, linear aeromagnetic signatures and the dikes shown in Figure 1 are delineated mainly by using airborne magnetic data (Ontario Geological Survey 1999, 2002a, 2002b, 2003, 2011). They are predominantly fine- to medium-grained gabbros, which commonly display diabase texture, and occasionally contain plagioclase phenocrysts or glomerocrysts. The dikes sometimes contain trace amounts of pyrite. Based on their orientation, the north-northwest-trending dikes are interpreted to be part of the Matachewan dike swarm (2454 Ma: Osmani 1991; 2473 Ma: Heaman 1997) and the northeast-trending dikes have been interpreted to be part of the Biscotasing dike swarm (2150 Ma: Osmani 1991).

Structural Geology

Primary layering (S_0) is observed in the form of sedimentary bedding and pillowed flows. Asymmetrical layering, such as graded beds and cross-beds, are sometimes observed within the area. In the eastern portion of the map area, along the boundary with the Missinaibi Lake batholith, facing is to the west. In the northern mafic metavolcanic rocks, just south of the Wabatonigushi Lake granitoid complex, facing is to the southwest. The clastic metasedimentary rocks in the southwest corner of Rennie Township face northeastward.

The earliest foliation (S_1) is only locally defined and is consistently overprinted by the dominant regional foliation (S_2). Where it is observable, the S_1 foliation is axial planar to F_1 folds, which are defined by mineralized laminated quartz veins. Both the S_1 foliation and F_1 folds are varied in orientation resulting from effects of later overprinting deformation.

The S_1 foliation and F_1 folds are overprinted by tight, generally upright, F_2 folds. The regional S_2 foliation is axial planar to the F_2 folds and is northwest to north-northwest striking with a steep westerly dip in the Renabie Mine area. At the Pileggi No. 1 outcrop, in the northwest corner of Stover Township, the S_2 foliation is subvertical and strikes west-northwest (Figure 1). The S_2 foliation is well-developed in the Missinaibi Lake batholith, where it is best identified by differentiated domains of biotite, and in supracrustal rocks of the Michipicoten greenstone belt, where it parallels primary layering (S_0). The S_2 foliation also parallels the intrusive contacts of the supracrustal rocks with the Missinaibi Lake batholith and Ash Lake pluton. A regional L_2 mineral stretching lineation is developed in association with the S_2 foliation. In the Renabie Mine area the L_2 lineation parallels the orientation of ore shoots and has a moderate to steep westerly plunge.

Late, east- to east-southeast-striking, steeply south-dipping shear zones overprint the S_2 foliation along the Renabie trend. These shear zones are host to the gold-bearing laminated quartz veins that were the main focus of gold production in the Renabie Mine area. The shear zones formed during a D_3 deformation event and are characterized by a strong internal S_3 foliation developed within the quartz-sericite-pyrite alteration halos that surround the laminated veins. Shear sense indicators record sinistral-reverse, oblique slip movement along the shear zones during the D_3 event. The mineralized D_3 shear zones are exposed along the Renabie trend from west to east at the C-Zone, Nudulama, and Nudulama East outcrops (McDivitt 2016a) (see Figure 1).

Within the shear zones along the Renabie trend, S_3 defines east-northeast-trending, steeply plunging Z-shaped drag folds. These folds record a late dextral transcurrent reactivation of the shear zones during a D_4 deformation event. Similar Z-shaped F_4 folds overprint the S_2 foliation at the Pileggi No. 1 outcrop (see Figure 1), suggesting that a late dextral reactivation of mineralized zones with an easterly trend is a recurrent feature in the district (McDivitt 2016a).

The study area was also affected by late brittle faulting (see Figure 1). The presence of these late faults is primarily supported by offsets revealed through mapping, topography, and by geophysical interpretation. There are several major faults interpreted in the area, some of which show displacements of up to 1 km. A major sinistral fault trends north-northwest from the southeast corner of Stover Township to the south-central edge of Rennie Township, transecting the Ruby Lake stock and showing a displacement of approximately 1 km. Another major fault runs underneath Crooked Lake and shows dextral displacement of just under 1 km; however, it does not transect the Ruby Lake stock. There are several faults extending southwestward from the northwest corner of Leeson Township towards the Rennie Lake stock. Some show dextral displacement of 1 km, whereas others do not display obvious offsets.

Field Trip Stops

Notes:

- 1) *Some of these outcrops are adjacent to open pits; please refrain from crossing fenced areas to access outcrop.*
- 2) *This group is unlikely to visit all of the stops described below during this trip. A number of stop descriptions are included in this guide to provide context and for future visits to the region. Permission must be granted from mining companies for future access.*
- 3) *The order and number of stops that will be visited during this field trip will be determined by access, time, and weather concerns. These factors may preclude visiting some stops, and alternative locations may be substituted.*
- 4) *All locations are given in NAD 83, Zone 17N UTM coordinates.*

Stop 1: Braminco open pit

Location: UTM Zone 17, 288468E 5360854N

The Braminco open pit (see Figure 1) is hosted in the Missinaibi Lake batholith near the contact with the mafic metavolcanic rocks of the Michipicoten greenstone belt. It consists of a shear zone that is 10-15 m wide, striking 135-145° and dipping moderately to the southwest (Sage and Heather 1991).

Mineralization at Braminco is similar in character to mineralization along the Renabie trend and consists of shear zone-hosted laminated quartz veins associated with quartz-sericite-pyrite alteration zones and late, hematite-bearing alteration assemblages. However, the Braminco shear zone differs in orientation from the Renabie trend shear zones, which are more easterly in trend. The Braminco shear zone is interpreted to have formed as a sinistral D₂ shear zone that underwent a dextral reactivation during D₃ (McDivitt 2016a). It was mined primarily for its silica, but gold was also recovered (Wilson, A., OGS, personal communication, 2012).

Stop 2: Baltic D outcrop

Location: UTM Zone 17, 289990E 5360020N

The Baltic D outcrop (see Figure 1) contains gold-bearing laminated quartz veins (Photo 1A) surrounded by sericitized, sulphide-bearing wallrock (Photo 1B) that formed after biotite tonalite of the Missinaibi Lake batholith. The laminated quartz veins are shallow in their dip and define isoclinal, near-recumbent F₁ folds. A continuous S₁ cleavage is axial planar to the folds and is well-developed within the sericitized wallrock. The F₁-folded veins and S₁ cleavage are refolded by upright, open, northwest-striking F₂ folds (McDivitt 2016a). The regional S₂ foliation is axial planar to the F₂ folds. The S₂ foliation strikes northwest and dips steeply west; it is well-developed in the biotite tonalite surrounding the sericitized wallrock marginal to the veins. The regional L₂ stretching lineation occurs in association with S₂ foliation and the F₂ folds. The lineation plunges moderately northwest, subparallel to the hinges of the F₂ folds (McDivitt 2016a).

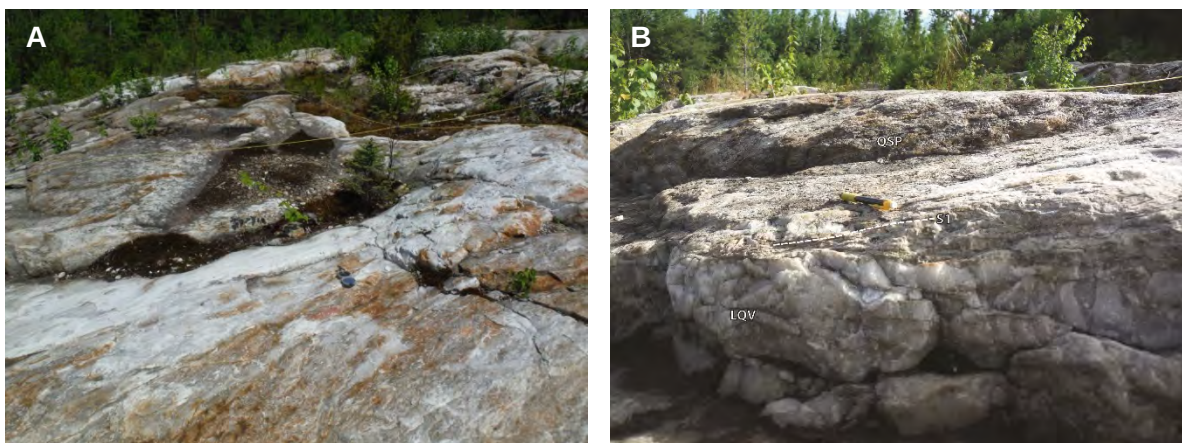


Photo 1. A) Overview photograph of the Baltic D outcrop displaying shallowly-dipping laminated quartz veins. Compass for scale is 6.9 cm wide, with sighting arm pointing north. **B)** Laminated quartz vein (LQV) adjacent to quartz-sericite-pyrite (QSP) schist, within which S_1 is developed. Pencil for scale is 0.8 cm wide and points north.

Stop 3: Nudulama East outcrop

Location: UTM Zone 17, 289028E 5361881N

The Nudulama East outcrop (Figure 2) is the easternmost outcrop defining the Renabie trend (see Figure 1). The outcrop is situated within the Missinaibi Lake batholith, approximately 2 km to the east of the contact with the Michipicoten greenstone belt. Mineralization is characterized by laminated quartz veins (Photo 2A) hosted within steeply dipping, east-southeast-trending D_3 shear zones. The shear zones are defined by strongly foliated (S_3) quartz-sericite-pyrite schist. Both the laminated veins and schist are mineralized. The veins typically contain 2-5% pyrite with minor chalcopyrite, molybdenite, galena, sphalerite, and rare visible gold. The surrounding schist contains similar amounts of pyrite, but other sulphide minerals are rare to absent. In addition to the laminated veins, late quartz $\pm$ chlorite $\pm$ epidote $\pm$ potassium-feldspar veins associated with hematite-bearing alteration assemblages occur at Nudulama East (Photo 2B, C). Where these veins are abundant, biotite within the host tonalite is pervasively chloritized, and the rock displays a distinct reddish hue because of the presence of hematite. A large quartz-chlorite breccia body is associated with the hematite-bearing alteration zone and cuts across the southern mineralized shear zone (Photo 2D, E). The breccia contains clasts of laminated quartz veins and sheared, hematite-altered tonalite enclosed within a matrix of vuggy quartz locally infilled by chlorite (Photo 2F) (McDivitt 2016a).

At Nudulama East, the S_2 foliation is oriented roughly perpendicular to the trend of the mineralized D_3 shear zones. The S_2 foliation and F_2 folds defined by laminated quartz veins and felsic dikes are overprinted by F_3 folds and transposed into parallelism with the D_3 shear zones. The F_3 folds have an axial planar S_3 foliation, which becomes the main foliation within the shear zones. Mutual overprinting relationships observed between hematite-associated veins and late D_3 structures suggest that the hematite-associated veins are syn- D_3 in timing (McDivitt 2016a). Neither the hematite-associated veins nor the quartz-chlorite breccia contain significant gold mineralization.

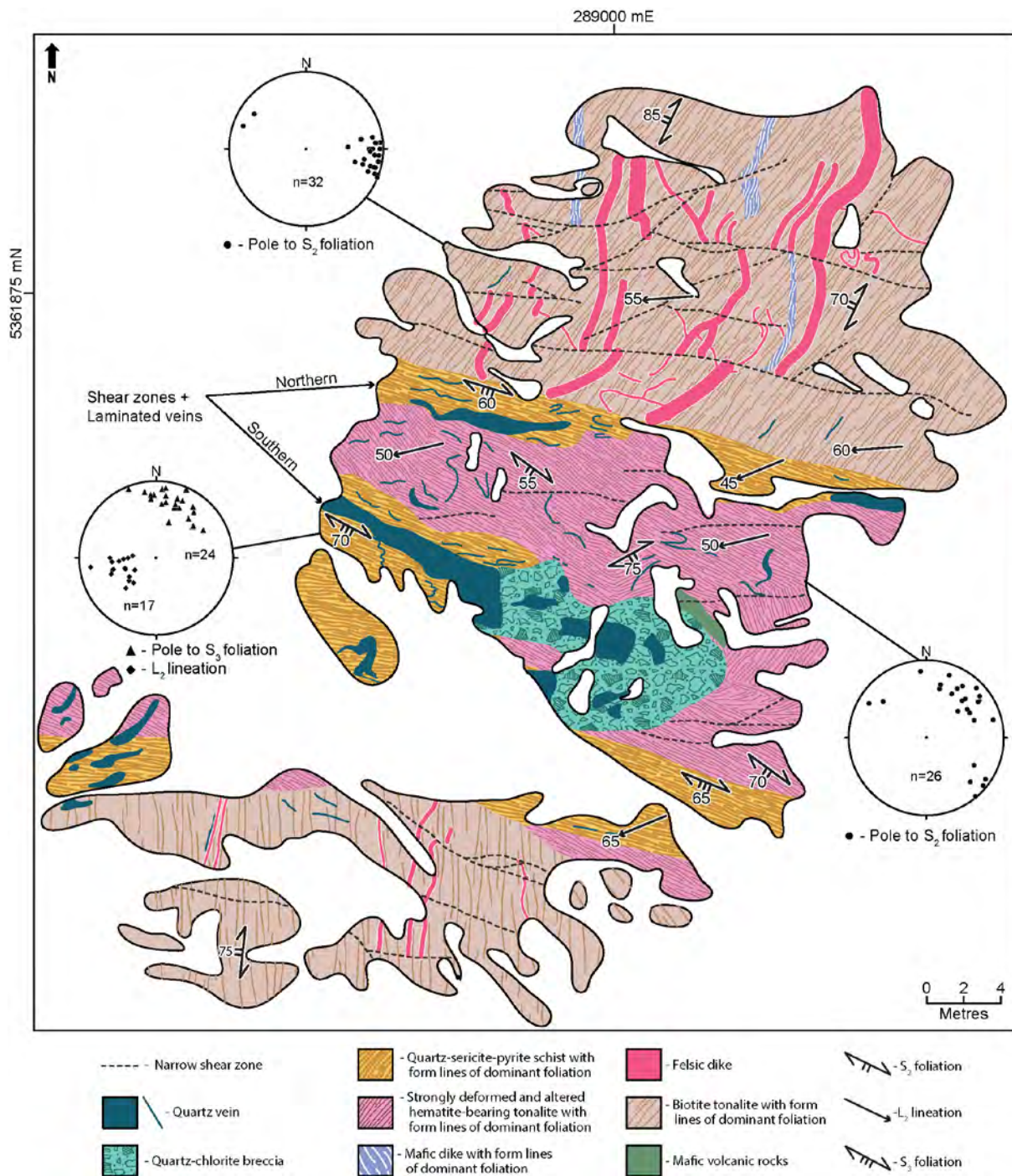


Figure 2. Detailed geological map of Nudulama East, modified from McDivitt (2016). Lower-hemisphere, equal-area stereographic plots showing the L_2 stretching lineation and poles to the S_2 and S_3 foliations. n = number of measurements.



Photo 2. Field photographs of the Nudulama East outcrop. **A)** Laminated quartz vein. Pencil for scale is 0.8 cm wide and points north. **B)** L₂-lineated, hematite-bearing altered tonalite. Pencil for scale is 0.8 cm wide and points north. **C)** Planar hematite-association alteration vein containing quartz-epidote infill mineralogy. Pencil for scale is 0.8 cm wide and points north. **D)** Hydrothermal breccia with fragments of foliated, hematite-altered tonalite surrounded by a quartz-chlorite matrix. **E)** A well-defined contact between a laminated quartz vein (LQV) vein within a quartz-sericite-pyrite shear zone, and a quartz-breccia vein containing clasts of hematite-bearing altered tonalite. Compass for scale is 7.1 cm wide, with sighting arm pointing north. **F)** Irregular-shaped hematite-associated alteration veins with quartz-chlorite infill mineralogy. Well-developed hematite-bearing alteration zones surrounds the veins. Pencil for scale is 0.8 cm wide and points north.

Stop 4: Nudulama – surrounding outcrop

Location: UTM Zone 17, 288623E 5361923N

The Nudulama outcrop (Figure 3) is located approximately 400 m west of the Nudulama East outcrop along the Renabie trend (see Figure 1). The outcrop is situated within the Missinaibi Lake batholith, approximately 1.5 km to the east of the contact with the Michipicoten greenstone belt. Mineralization at Nudulama is characterized by laminated quartz veins within an approximately 10 m wide D_3 shear zone that consist of strongly foliated quartz-sericite-pyrite schist (Photo 3A, B) (McDivitt 2016a). The laminated veins comprise very fine-grained saccharoidal quartz layers separated by thin sericitic laminations. The veins typically contain up to 5% pyrite with minor chalcopyrite, molybdenite, galena, sphalerite, and rare visible gold; the schist contains similar quantities of pyrite, but a relative absence of other sulphide minerals is noted. Both the laminated veins and schist consistently return significant gold values (i.e., >1 ppm), with higher grade samples containing in excess of 10 ppm gold. Late quartz $\pm$ chlorite $\pm$ epidote $\pm$ potassium-feldspar veins associated with hematite-bearing alteration assemblages occur at Nudulama, but they are less abundant than at Nudulama East. Steeply dipping, Z-shaped drag folds that trend east-northeast are defined by the S_3 foliation within the mineralized shear zone. These folds formed during a late dextral transcurrent reactivation of the shear zone during a D_4 deformation event (McDivitt 2016a).

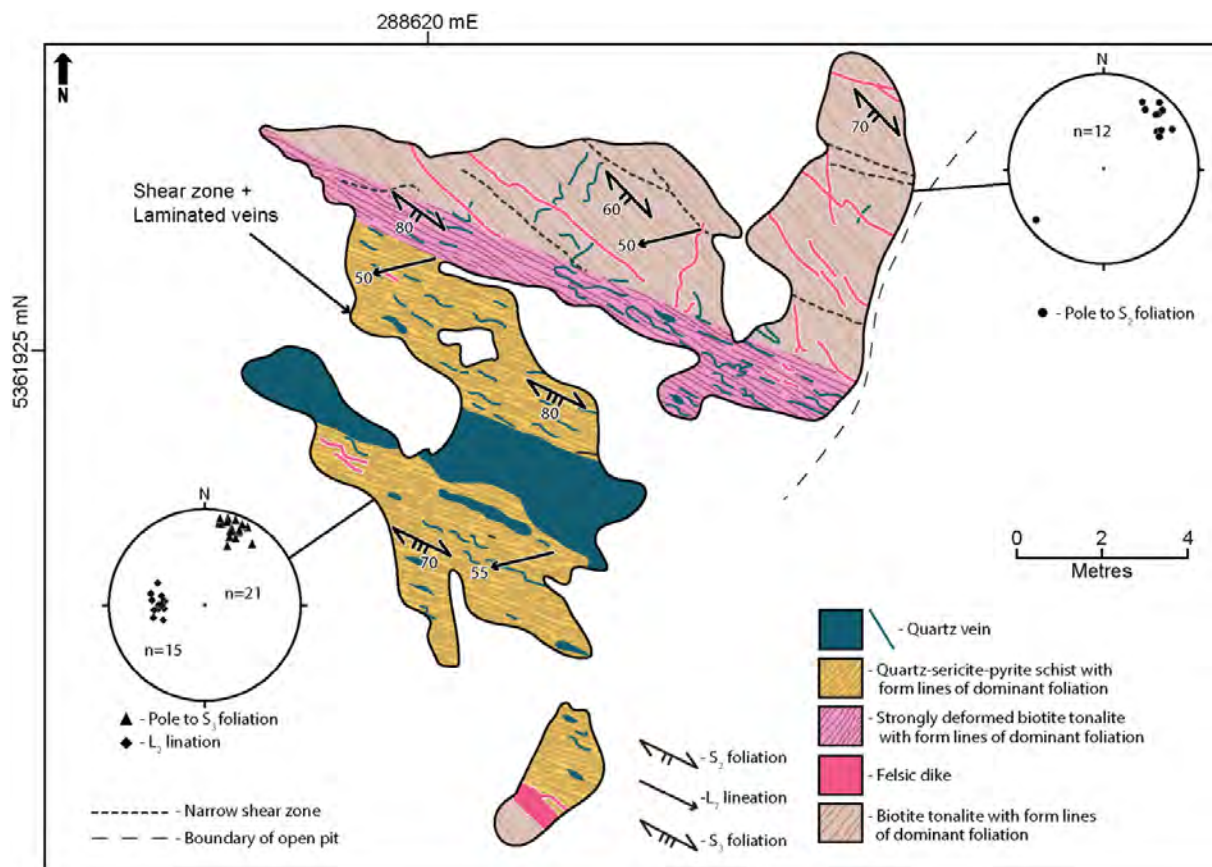


Figure 3. Detailed geological map of Nudulama, modified from McDivitt (2016). Lower-hemisphere, equal-area stereographic plots of the L_2 stretching lineation and poles to S_2 and S_3 foliations. n = number of measurements.

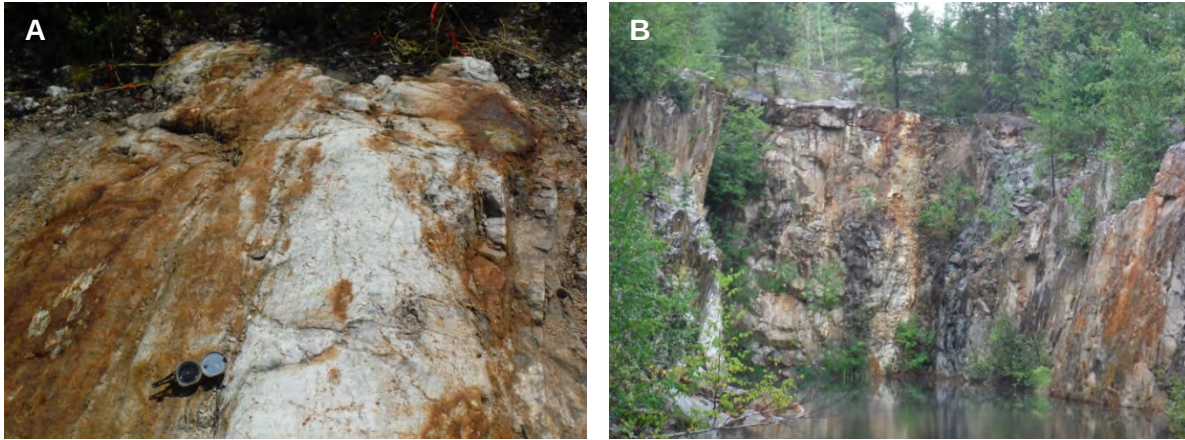


Photo 3. Field photographs of the Nudulama outcrop. **A)** Mineralized zone shows a laminated quartz vein surrounded by quartz-sericite-pyrite schist. Compass for scale is 6.9 cm wide, with sighting arm pointing north. **B)** East-facing, vertical wall at the Nudulama open pit (~10 m wide), displaying the same mineralized zone as in A. Photo taken looking north.

Stop 5: C-Zone outcrop

Location: UTM Zone 17, 287890E 5362050N

The C-Zone outcrop (Figure 4) is located approximately 750 m west of the Nudulama outcrop along the Renabie trend (see Figure 1). The outcrop occurs within the Missinaibi Lake batholith, approximately 750 m to the east of the contact with the Michipicoten greenstone belt. Mineralization at the C-Zone outcrop is similar in character to mineralization at Nudulama and Nudulama East outcrops: it consists of laminated quartz veins (Photo 4A, B) hosted within an east-trending, steeply dipping D_3 shear zone of variable width (5-10 m). The shear zone consists of strongly foliated (S_3) quartz-sericite-pyrite schist. Both the laminated veins and the schist are mineralized. The schist contains 2-5% pyrite; the veins contain similar amounts of pyrite but can also have small amounts of chalcopyrite, molybdenite, galena, sphalerite, and rare visible gold (McDivitt 2016a).

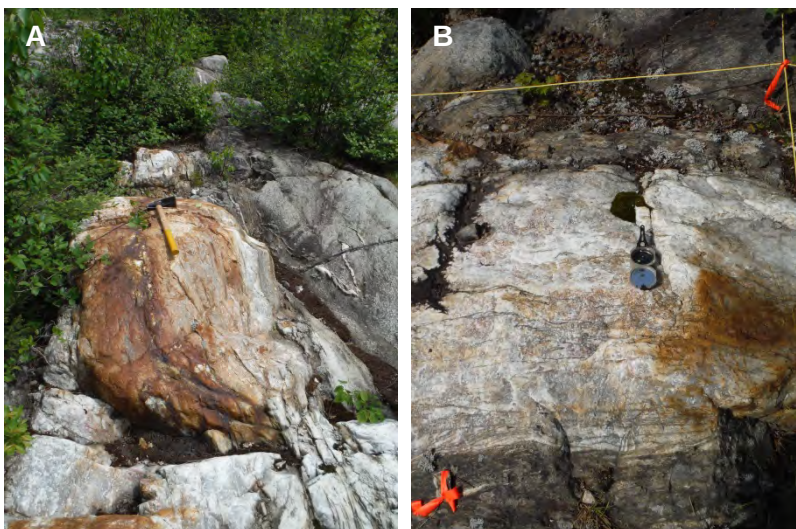


Photo 4. Field photographs of the C-Zone outcrop. **A)** The main laminated quartz vein. The hammer head points to the east (the hammer is 70 cm long). **B)** A close up photograph of a laminated quartz vein. Compass for scale is 6.9 cm wide, with sighting arm pointing north.

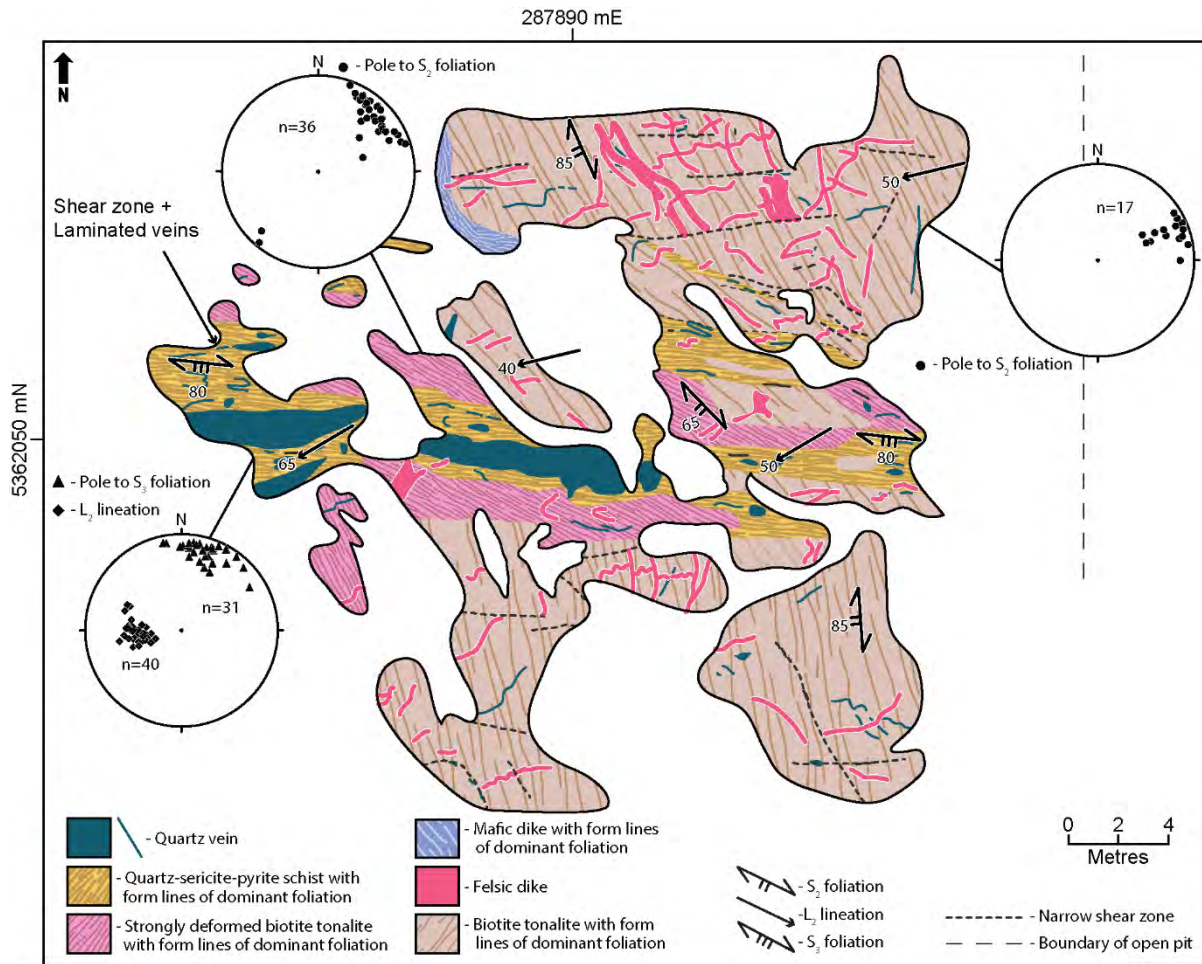


Figure 4. Detailed geological map of the C-Zone, modified from McDivitt (2016). Lower-hemisphere, equal-area stereographic plots of the L_2 stretching lineation and poles to S_2 and S_3 foliations. n = number of measurements.

Stop 6: No. 2 Shaft and contact outcrop

Location: UTM Zone 17, 287246E 5361990N

The site of the former Renabie Mine is the westernmost outcrop defining the Renabie trend. At the now defunct site of the No. 2 shaft (see Figure 1), the Shaft Fault offsets the contact between the Archean supracrustal rocks of the Michipicoten greenstone belt and the Missinaibi Lake batholith. The fault is not observed at surface and was primarily defined from underground workings. This fault, and others parallel to it, are defined as late, north-trending brittle faults that cut across the Renabie trend. These faults generally appear to have a sinistral offset. The Shaft Fault has been interpreted as an oblique-slip fault having a sinistral strike-slip component as well as a west-side-up (reverse), dip-slip component (Callan and Spooner 1998).

On the hill to the south of the No. 2 Shaft, the contact between the mafic metavolcanic rocks of the Michipicoten greenstone belt and tonalite of the Missinaibi Lake batholith can be observed. The contact trends roughly 140° , dips 70° to the southwest and parallels the regional S_2 foliation.

Stop 7: Ultramafic metavolcanic rocks

Location: UTM Zone 17, 284487E 5366237N

Ultramafic metavolcanic rocks (Photo 5) (see Figure 1) are seldom observed in the area. This outcrop displays talc and chlorite alteration and also contains trace amounts of pyrite. The foliation parallels the contact of the supracrustal rocks with the Wabatongushi Lake granitoid complex, which is situated roughly 1.4 km north of the outcrop.



Photo 5. Highly altered ultramafic metavolcanic rocks.

Stop 8: VMS-style alteration

Location: UTM Zone 17, 285343E 5364967N

This is a steep outcrop located on the west side of the road (see Figure 1). The main lithology consists of felsic tuffs (Photo 6A) with areas of crystal tuff. At the top of the exposure a thick (~10-20 cm) band of coarse-grained amphibole and garnet can be observed (Photo 6B). Disseminated pyrite can also be found within the alteration area.



Photo 6. A) Felsic metavolcanic rocks. **B)** Amphibole and garnet alteration with disseminated pyrite.

Although geochemical analysis did not return particularly good assay values for this outcrop, it does bear some similarities with the Conboy Lake occurrence (MDI42B05NW00021; Ontario Geological Survey 2017). The Conboy Lake occurrence (282745E 5366813N) is located in north-central Rennie Township, approximately 3 km to the northwest of this location. It is an historic zinc, silver and copper occurrence with minor gold mineralization, and is classified in the OGS Mineral Deposit Inventory (MDI) as a developed prospect with reserves. It has a long history of sporadic mineral exploration between 1939 and 2010, as summarized in Ontario Geological Survey (2017).

The Conboy Lake occurrence is found within younger felsic metavolcanic rocks (2704.6 ± 2.1 Ma; Kamo 2016; Robichaud, McDivitt and Trevisan 2017) that display pervasive sericite alteration. The zinc mineralization occurs as thick layers of massive sphalerite with disseminated pyrite and chalcopyrite (Robichaud, McDivitt and Trevisan 2015; Riley 1971, p.45-49; Ontario Geological Survey 2017).

Stop 9: Felsic crystal tuff

Location: UTM Zone 17, 285440E 5362893N

The felsic crystal tuffs (see Figure 1) at this outcrop are typical of the area. They are generally fine- to medium-grained with coarser crystals of either quartz or plagioclase. The fresh surface is light grey and weathers to a lighter grey to beige colour. Bedding is rarely observed in the felsic metavolcanic rocks and this outcrop is no different. This outcrop was sampled for geochronology in 2015, with a U/Pb age of 2730.9 ± 1.2 Ma reported by Kamo (2016).

Stop 10: Felsic tuff-breccia

Location: UTM Zone 17, 283838E 5361108N

Felsic tuff-breccias (Photo 7) (see Figure 1) are seldom observed in the area and are very similar to the conglomerates, but differ in their monolithic clast composition. The clasts are composed of felsic metavolcanic material and are angular. In some areas, the clast edges seem to be broken *in situ*, making it a mosaic breccia.

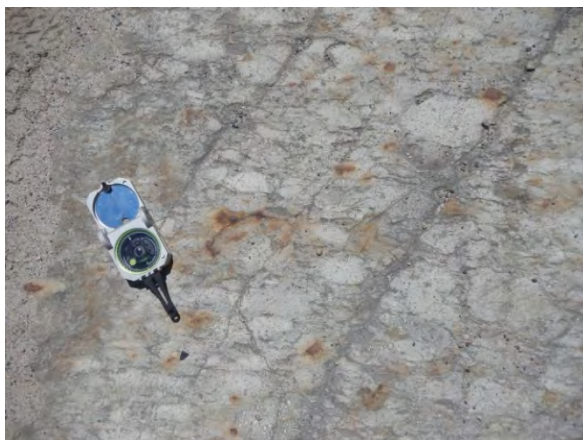


Photo 7. Felsic tuff breccia. Compass for scale is 7.1 cm wide, with sighting arm pointing north.

Stop 11: Pillowed mafic flows

Location: UTM Zone 17, 279434E 5361230N

This outcrop (see Figure 1) is located approximately 50 m to the north of the Renabie road; it may need to be cleared of the brush to see the features. The outcrop consists of mafic lava flows. Pillows can be observed, ranging in size from 30 to 60 cm.

Stop 12: Rennie Lake turnoff conglomerate

Location: UTM Zone 17, 279434E 5361230N

This polymictic conglomerate is situated on the Renabie road (Photo 8). The conglomerate is matrix supported, but contains large cobbles of predominantly tonalitic porphyry with a few mudstone and siltstone cobbles. The conglomerate is highly foliated and the clasts show an elongation fabric that parallels bedding. Cross-bedding is observed in the more sandy layers and indicates younging to the northeast. A U/Pb age of 2695 ± 3 Ma for the sandy portion of this outcrop was reported by Davis (2016).



Photo 8. Highly foliated conglomerate. Compass for scale is 7.1 cm wide, with sighting arm pointing north.

Stop 13.1: Baltimore Lake metasedimentary rocks and diabase dike

Location: UTM Zone 17, 279076E 5361116N

The Baltimore Lake metasedimentary rocks consist of buff-grey, quartz-rich, thinly bedded siltstone intercalated with lesser sandier layers. Small staurolite grains can be observed in some areas. Graded bedding (Photo 9A) can be observed on the outcrop and indicates facing is to the northeast.

The diabase dike is located at the southern end of the outcrop, near the water. The dike is interpreted to be of the Matachewan swarm because of its north-trending orientation. The contact between the metasedimentary rocks and the dike can be observed near the water's edge. The dike contains numerous plagioclase phenocrysts, and in one area, several large glomerocrysts (Photo 9B).

On the southern shore of Baltimore Lake, the metasedimentary rocks are more deformed. They are folded, making primary features difficult to discern (Photo 9C, D). The metasedimentary rocks are predominantly sandstones that are finely interbedded with silty layers, but some of the sandstone beds are up to 1 m in thickness. Conglomerates can also be observed on the southern shore of the lake.

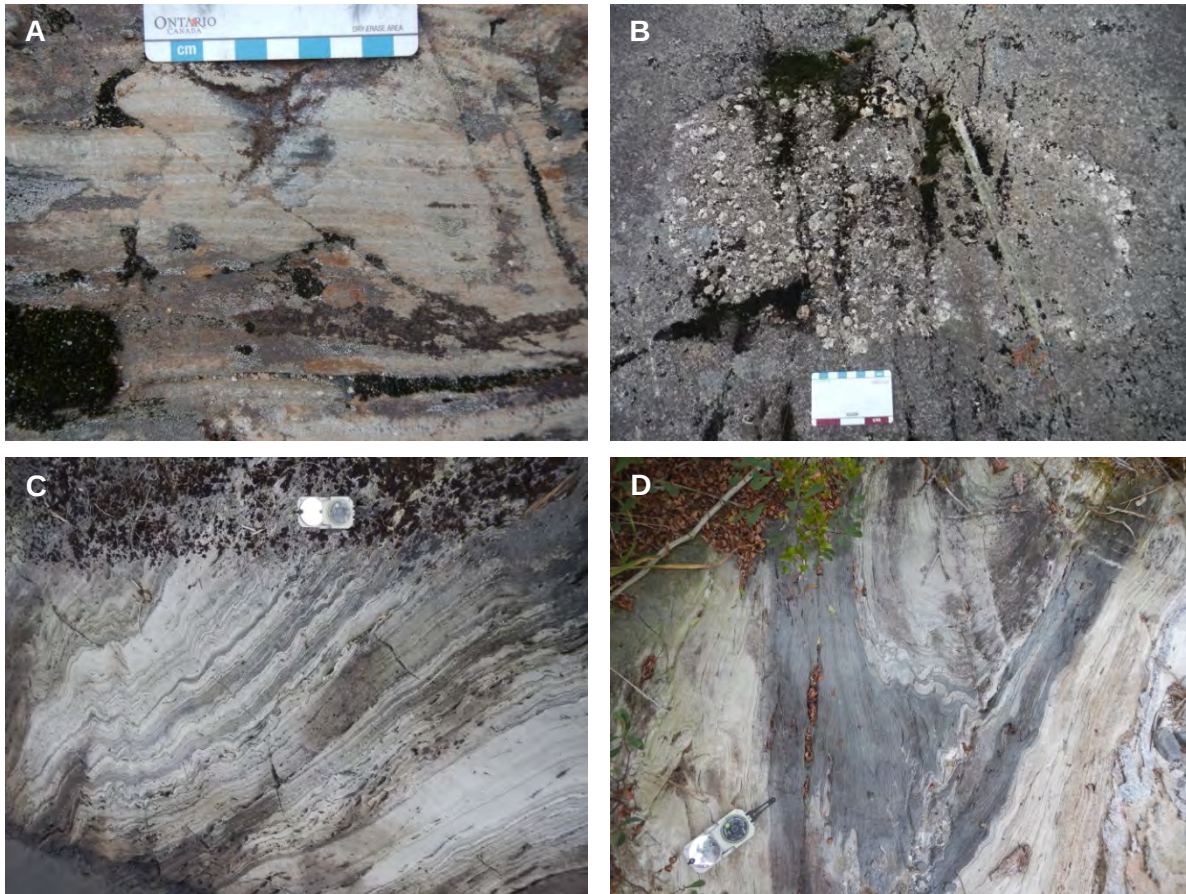


Photo 9. Field photographs of the Baltimore Lake area. **A)** Graded bedding indicates facing is to the top of the photo. Top of photo is to the northeast. **B)** Matachewan dike with plagioclase glomerocrysts. **C)** Interbedded sandy and silty layers showing parasitic axial planar folding. **D)** Large synclinal fold. Compass for scale is 7.1 cm wide, with sighting arm pointing north.

Stop 13.2: Baltimore Lake metasedimentary rocks

Location: UTM Zone 17, 278836E 5361004N

Approximately 250 m west of the last outcrop sits another large exposure of the same metasedimentary rocks. A very well-polished surface occurs in the middle of the Renabie road and offers the best exposure; however, there are many areas of interest across the outcrop. The metasedimentary rocks consist of buff-grey, quartz-rich, thinly bedded siltstone and sandstone. Siltstone is the dominant sedimentary unit, but sporadic beds of sandstone also occur. The siltstone tends to be thinly to thickly laminated, and the sandstone ranges from a few decimetres to 1 m in thickness.

Stop 14.1: Iron formation

Location: UTM Zone 17, 277276E 5361723N

From the Renabie road, go north past the quarry for approximately 1.3 km until a small east-trending trail is reached. Follow this trail for 1 km on foot (or by ATV); the outcrop will be on the south side of the trail.

The iron formation consists of magnetite-rich layers interbedded with sandstone (Photo 10A, B). Folds can be observed within the banded iron formation, and are especially visible in the central cavity where the third dimension can be observed. Disseminated pyrite and chalcopyrite are also noted.

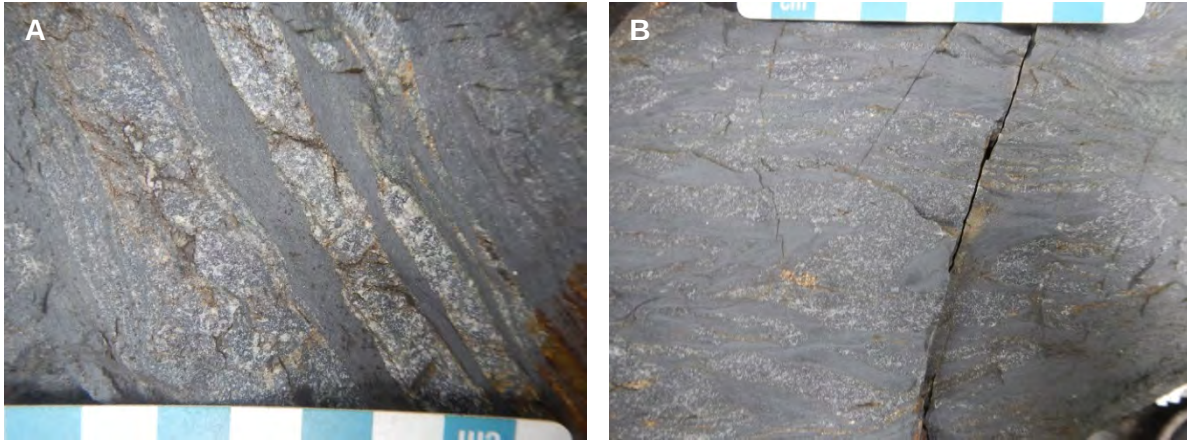


Photo 10. A) Banded iron formation displaying magnetite-rich layers interbedded with sandstone. **B)** Folded banded iron formation.

Stop 14.2: Iron formation

Location: UTM Zone 17, 276889E 5359035N

This iron formation is more readily accessible than that at the previous stop, but is not as well-exposed. This outcrop is located approximately 100 m down the road leading to the Pileggi No. 1 outcrop (Figure 1). There is also another small exposure 200 m further down the road.

The outcrop is dominated by the presence of gabbro; the iron formation occupies a small portion ($< 1 \text{ m}^2$) of this outcrop. The iron formation consists of magnetite-rich layers interbedded with sandstone.

Stop 15: Pileggi No. 1 outcrop

Location: UTM Zone 17, 277677E 5358487N

The Pileggi No. 1 outcrop (Figure 5) occurs approximately 120 m north of the main road (Figure 1). There is a small access trail that can be used to walk up to the outcrop. The outcrop consists of 2 separate exposures (A and B) that connect over a small ridge (see Figure 5 inset). The outcrop consists of metavolcanic rocks of the Michipicoten greenstone belt that have been intruded by feldspar-phyric and amphibole-phyric dikes. Matachewan dikes intrude the metavolcanic rocks. Gold mineralization occurs within deformed laminated quartz veins that returned values up to 22.8 ppm.

The laminated quartz veins are overprinted by isoclinal F_1 folds associated with an S_1 axial planar cleavage. The F_1 folds and S_1 cleavage formed during an early deformation event, which predated the development of the regional S_2 foliation (Photo 11A). The S_2 foliation (Photo 11B), which strikes

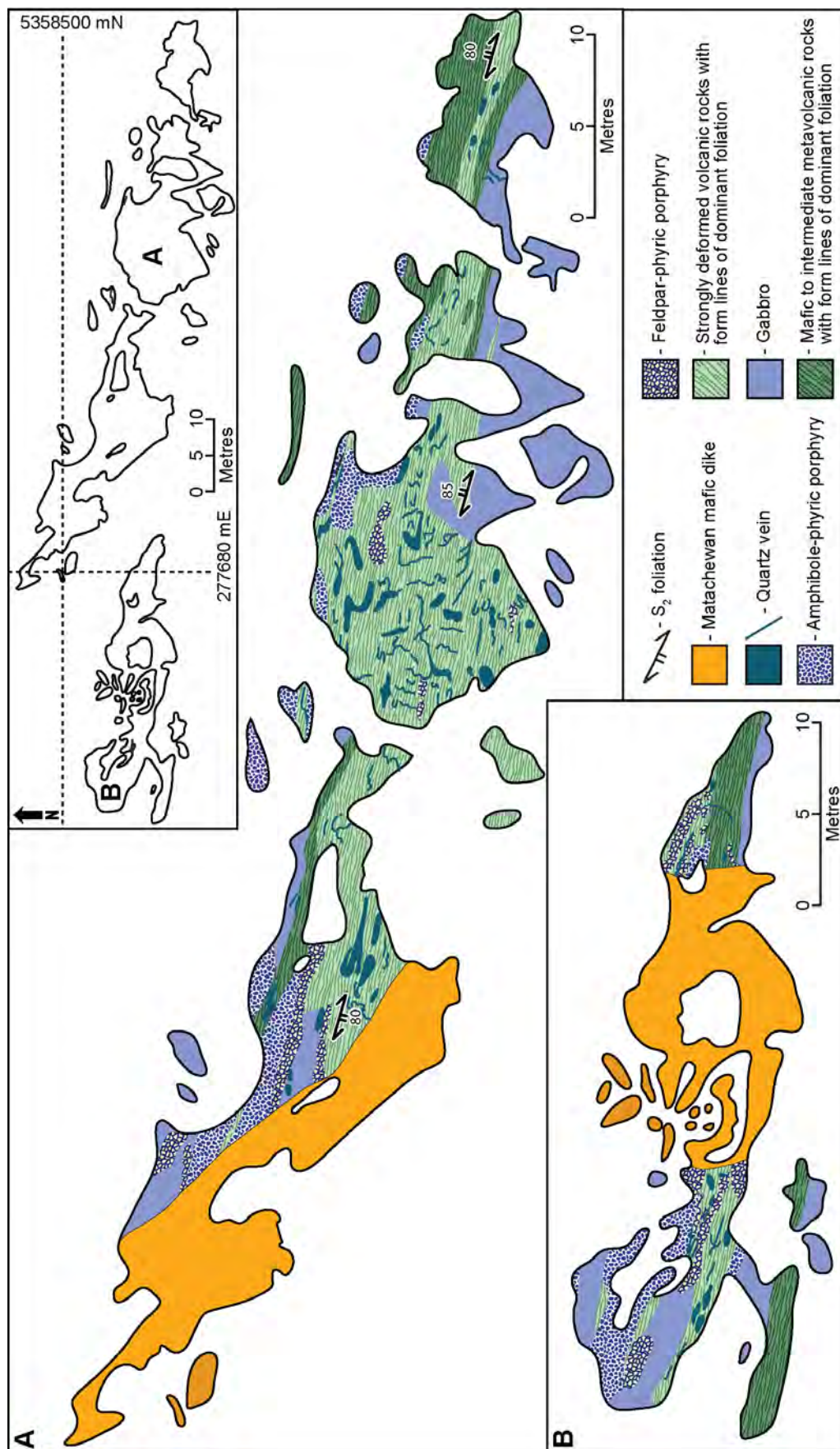


Figure 5. Detailed geological maps of Pileggi No. 1 north (A) and south (B) exposures, modified from McDivitt (2016). Inset map shows the relative location of the two exposures.

approximately 100° and dips steeply, is axial planar to an upright, shallowly plunging F₂ fold that occupies roughly half the width of the outcrop. The F₂ fold overprints the F₁ folded laminated veins and the S₁ cleavage. Parasitic folds in the hinge and limbs of the larger F₂ fold vary in plunge from subhorizontal to moderately plunging. The S₂ foliation is manifested as a disjunctive or crenulation cleavage in the hinges of F₂ folds, and as a slaty transposition cleavage along the limbs of the folds. Z-shaped drag folds and northwest-trending, subvertical quartz tension gashes overprint the S₂ foliation and the laminated veins (Photo 11C). The Z-shaped drag folds trend east-northeast and are steeply plunging. The quartz tension gashes are either straight or Z-shaped sigmoidal. The drag folds and the quartz tensions gashes are attributed to a late dextral reactivation of the mineralized zone during the D₄ deformation event (McDivitt 2016a). The variable shape of the tension gashes indicates that they were emplaced throughout the D₄ event, with the Z-shaped tension gashes being older than the straight ones. Unlike the laminated veins, the late tension gashes are not gold bearing.

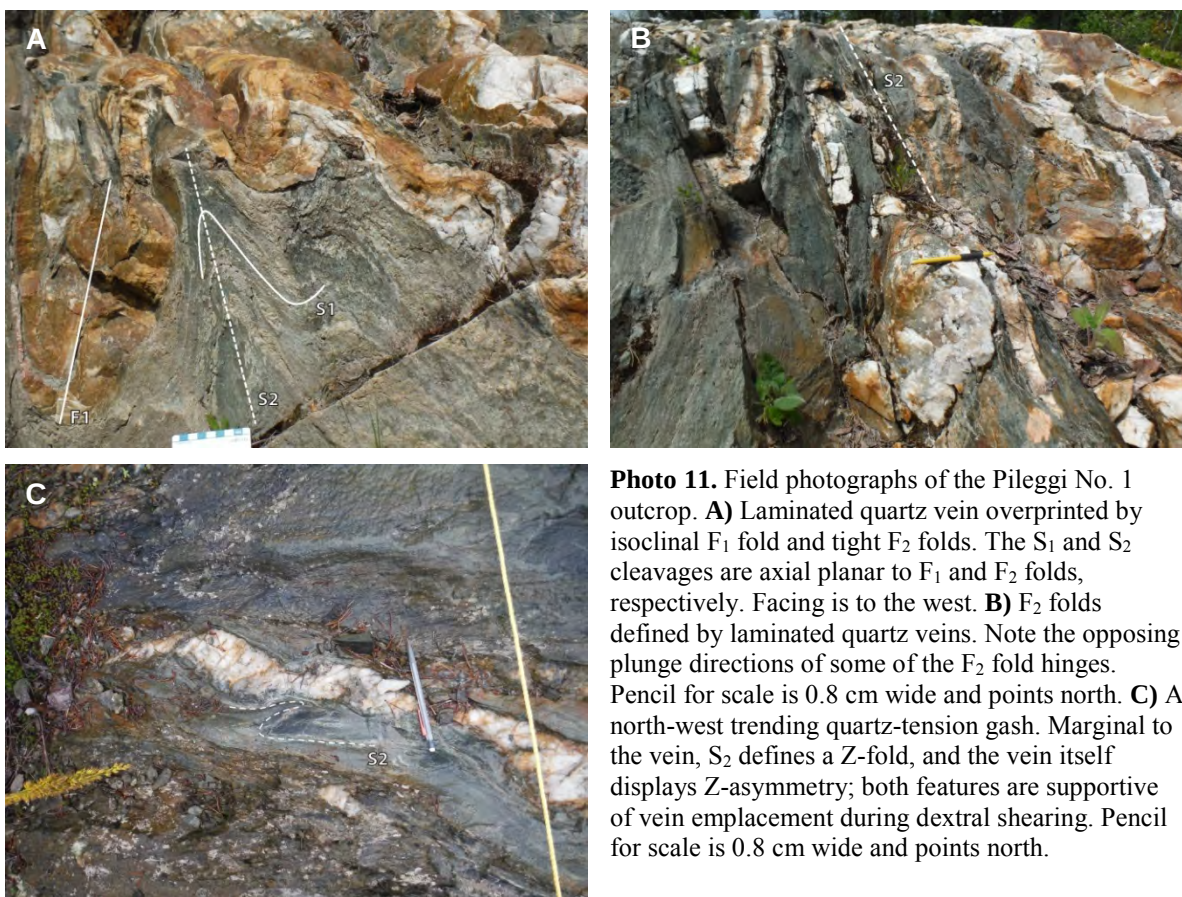


Photo 11. Field photographs of the Pileggi No. 1 outcrop. **A)** Laminated quartz vein overprinted by isoclinal F₁ fold and tight F₂ folds. The S₁ and S₂ cleavages are axial planar to F₁ and F₂ folds, respectively. Facing is to the west. **B)** F₂ folds defined by laminated quartz veins. Note the opposing plunge directions of some of the F₂ fold hinges. Pencil for scale is 0.8 cm wide and points north. **C)** A north-west trending quartz-tension gash. Marginal to the vein, S₂ defines a Z-fold, and the vein itself displays Z-asymmetry; both features are supportive of vein emplacement during dextral shearing. Pencil for scale is 0.8 cm wide and points north.

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Field Trip 6

Kapuskasing Structural Zone and the Borden Lake Gold Deposit

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Introduction

The Kapuskasing Structural Zone is a window into the continental crust, which was investigated thoroughly during the past 30 years. Due to its nature, it was suspected to be barren of any mineralization typical of greenstone belts. Recent findings raised more questions about it, and gave geoscience a black eye, especially after the discovery of the Borden Lake gold deposit.

This field trip will be an overview of the Kapuskasing Structural Zone (KSZ). The KSZ was the subject of an ILSG field trip back in 1987 (Percival 1987). This field trip will take place within reasonable distance of Highway 101. As a precautionary note, be aware of the traffic when exiting the vehicles and while crossing or walking along the highway or any roads. Field trip participants should also use caution on outcrops and ditches since they might prove slippery and steep.

Exploration and mining history

Early explorers were mostly looking at the greenstone belts on either side of the Kapuskasing Structural Zone. However, within the zone, exploration was mostly aimed at the carbonatite intrusions. In the late 1950's, the Nemegosenda Carbonatite Complex was explored by the Dominion Gulf Company. The company performed an aeromagnetic survey which showed a circular anomaly on the east side of Nemegosenda Lake. This anomaly was, in fact, the Nemegosenda Alkaline Complex. Through clever use of a ground magnetic survey in an area where outcrops were scarce, the Dominion Gulf Company singled out highs which required follow-up drilling. From 1954 to 1959, the company drilled over a hundred holes and even opened an adit of 190 metres into the main mineralized zone (D-Zone) which contained niobium (Archibald, 2008). The company also completed diamond drilling in the South-East Area, located 2 km southeast from the adit, which shows endowment in rare earth elements. The property was taken over by Musto Explorations Ltd. in 1987, who pursued the exploration on the South-East Area by retrieving core drilled in the fifties, split and re-assayed the core. The results were similar to those obtained in the fifties.

In the early nineties, Placer Explorations did a re-plotting of the D zone and South East Area. These plots were used to calculate a block model of the D zone. In 2008, Sarissa Resources Inc. acquired the Nemegosenda project, and re-evaluated the historic data. Sarissa Resources followed up with a series of diamond drill holes to confirm its location in the D zone. In the fall of 2016, the project was sold to Indo Global Exchange(s) Pte, Ltd.

The Lackner Lake carbonatite alkalic complex is another area which has undergone exploration in the past sixty years. Two patented claims, in McNaught Township, uncovered magnetite mineralization discovered over a century ago. In 1949, prospectors discovered strong radioactivity on these claims. Nemegos Uranium Corporation was formed to explore these and nearby claims. The exploration concentrated on developing apatite, magnetite, and uranium ore deposits. In 1951, an aeromagnetic survey was flown by Dominion Gulf Company, who followed up with staking anomalies located on the east flank of the structure. At the same time, niobium was discovered in samples assayed by Nemegos Uranium Corporation, but the results were not followed up. In 1954, Multi-Minerals Limited demonstrated the extensive niobium-bearing mineralization on the claims formerly owned by Nemegos Uranium Corporation. The company trenched and drilled in the southwest corner of the complex, which uncovered three zones of niobium mineralization. In 1970, Multi-Minerals Limited optioned one zone to Fetio Industrial Developments Limited who proceeded with an evaluation of possible production of titanium-iron and phosphate concentrates. The company shipped 1500 tons of concentrate to the United States for metallurgical testing. The property was then optioned to Mertec Resources Development in 1974, which was terminated in 1975. During 1978 to 1979, Multi-Minerals Limited drilled 6 holes totalling 2176 feet to test uranium potential of various mineralized zones outlined by previous work. 6378366 Canada Inc. and 6070205 Canada Inc. went back to compile data on the Lackner Lake carbonatite in 2008. A series of exploration projects were undertaken, including prospecting, sampling, ground radiometric survey and assaying. It was followed by an aerial magnetic and radiometric survey by Rare Earth Metals Inc. in 2009. The most recent work was completed in 2014 by Gold Crossing Limited and International Explorers & Prospectors Inc. who completed radiometric surveys, geological mapping, sampling, and assaying of rock and past cores.

The Shawmere Anorthosite complex was explored back in the 1990's. Purechem Ltd. did geological mapping, geochemistry, petrography and bulk sampling from 1993 to 1999 for aluminium silica. The property was then purchased by Southern Africa Corporation in 2001, which bulk sampled and performed more tests. In 2002-2007, Avalon Ventures Ltd. acquired the property and completed bulk sampling and sent more than 1000 tons of material for furnace trials and processing of anorthosite. The company, now under the name of Avalon Advanced Metals Inc., still holds the property.

The Borden Lake gold deposit was initially explored in the early 1990's by M. Tremblay. The prospector, with the help of Jacques Robert, completed a series of exploration projects including prospecting, sampling, electromagnetic survey and stripping in an area which featured Timiskaming age conglomerates. Getting positive results, the claims were acquired by Probe Mines Limited in 2010. From 2010 to 2013, Probe Mines Limited drilled over 250 holes on the property. Probe Mines Limited released an indicated resource estimate showing 9,262,000 tonnes at a grade of 5.39 g/t gold, and inferred resources estimate of 3,034,000 tons at a grade of 4.37 g/t gold for an underground operation (cut-off of 2.5 g/t gold), and an indicated 70,301,000 tons at a grade of 1.03 g/t gold and an inferred 247,000 tonnes at a grade of 0.8 g/t gold for an open pit operation (cut-off of 0.5 g/t gold; Dzick 2014). Goldcorp Inc. acquired the property by absorbing Probe Mines Limited in the summer of 2013. Goldcorp Inc. is currently drilling and planning for the possible opening of a mine in the near future. The latest resource estimate show 3.02 Mt at a grade of 5.77 g/t gold of indicated resources, 2.03 Mt at a grade of 5.49 g/t gold of inferred resources (Goldcorp Inc. 2016).

Geologic Setting

The Kapuskasing Structural Zone (KSZ) is located in the Superior Province, cutting it in two. It trends in a NNE direction, starting south of James Bay and dying off gradually south of Chapleau. While the rocks on each side are of greenschist to amphibolite facies, mostly composed of volcano-sedimentary belts with intrusives, the rocks of the KSZ are striking in contrast since they are of amphibolite to granulite facies. The KSZ is also characterized by a strong positive gravity and aeromagnetic anomalies that are trending NNE. The west contact of the KSZ is transitional while the east contact is outlined by the Ivanhoe Lake cataclastic zone. The structures within the zone are at an almost right angle with those from the surrounding belts, as pictured in aeromagnetic surveys (Figure 1).

The rock compositions within the structure consist of alternating bands of mafic gneiss, paragneiss, tonalitic gneiss and dioritic rocks which are elongated to the northeast (Percival 1981). Two discrete bodies of anorthosite form the Shawmere anorthositic complex (Thurston, Siragusa and Sage, 1977). All the rocks in the KSZ were deformed and metamorphosed to high-grade, and the contact relations among units are still blurry in understanding (Percival 1981).

Paragneiss is characterized by garnet, biotite, plagioclase and quartz. It is thought to be of sedimentary origin (Percival 1981). These paragneissic rocks contain up to 20% concordant tonalitic layers, and locally contain hypersthene. Paragneisses were observed to be richer in biotite southeast of the Shenango complex. A small unit of paragneiss west of Tom Smith Lake contains up to 10% muscovite as well as carbonate layers, while a small unit southeast of Carty Lake contains up to 5% graphite (Percival 1981). South of Nemegosenda Lake, along Highway 101, layers of varying composition on the 10-50 cm scale in fine-grained, biotite-poor, biotite-plagioclase-quartz gneiss hints that these rocks have an arkose protolith (Percival 1981). A recurrent component of many paragneiss outcrops are enclaves of mafic gneiss. These enclaves are comparable in composition and texture to the bigger mafic gneiss belts. The variation in composition of the paragneiss suggests an original sedimentary facies change (Percival 1981).

Mafic gneiss is distinguished by garnet-clinopyroxene-hornblende-plagioclase-quartz mineral assemblages and generally incorporates concordant tonalite layers (Percival 1981). Centimeter thick layers are made by different proportions of minerals. Discrete 2-4 km wide mafic gneiss belts are found in the Borden-Hellyer Lakes area, in a body adjacent south of the main anorthosite body.

Dioritic rocks occur dominantly within paragneiss belts. These homogeneous meta-igneous bodies consist of medium-to-coarse-grained hornblende-biotite-plagioclase rocks with up to 10% quartz as well as some clinopyroxene, orthopyroxene and rare garnet (Percival 1981). The texture of the dioritic rocks varies from foliated to flaser or augen, to gneissic. The largest body occurs in the Chapleau-Nemegosenda Rivers area, which also shows gabbro, hornblendite and pyroxenite layers from 10 cm to 2 m in width. Veinlets of quartz-monzonite constitute 15% of outcrop in the belt northeast of the Lackner complex, which possesses clinopyroxene-bearing leucosome (Percival 1981).

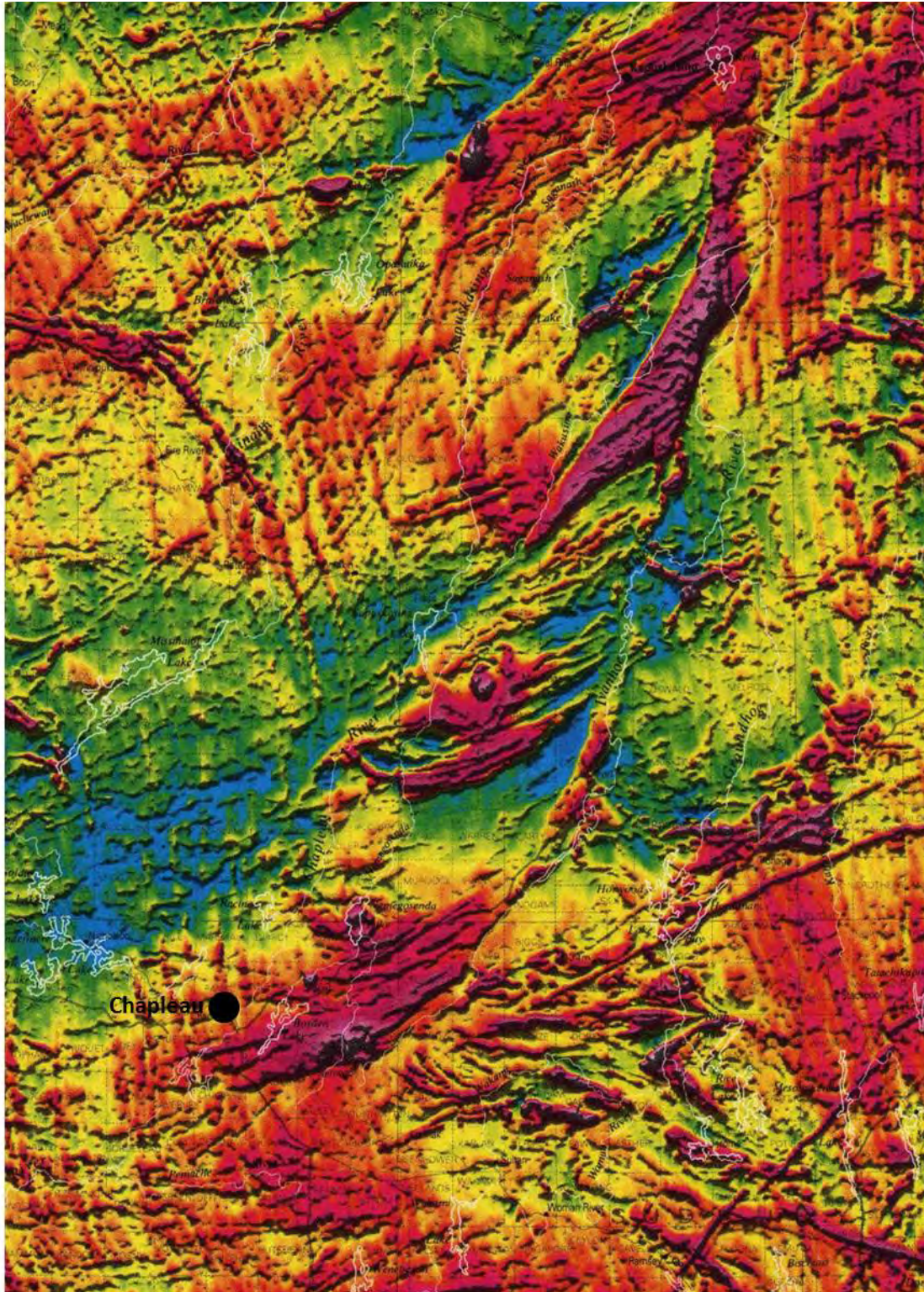


Figure 1: Aeromagnetic map of the Kapuskasing Structural Zone (modified from Gupta 1991).

Tonalitic rocks outcrop as discrete belts of gneissic and xenolithic bodies south of the Shawmere anorthosite complex (Percival 1981). Southeast and south of Carty Lake, there is a body of coarse garnet-hornblende-biotite-plagioclase-quartz gneissic tonalite hosting mafic and ultramafic granulite enclaves. Inclusions ranging from fist-sized to 1 m, constitute 15% of the rock and are of layered mafic gneiss, amphibolite, and garnet-orthopyroxene-hornblende-biotite rock. Schlieren and units of paragneiss up to 1 km wide are common. The proportion of garnet diminishes and the composition becomes granodioritic to the southwest along the belt. Enclaves encountered are mostly amphibolites with rare bright green cummingtonite-hornblende-biotite rocks (Percival 1981). Three eastward-thinning belts, between Nemegosenda and Lackner Lakes, are comprised of medium-to-coarse grained, thinly-banded gneissic tonalite (Figure 2) containing hornblende, biotite, plagioclase and quartz in variable proportions (Percival 1981).



Figure 2: Thinly-banded gneissic tonalite

The Shawmere anorthosite complex consists of a large, oval mass of anorthosite to anorthositic gabbro lying adjacent to and ESE of the KSZ underlying an area of 800 km², and a satellite Y-shape mass underlying an area of 90 km² (Thurston, Siragusa and Sage, 1977). A strongly foliated quartz diorite and monzonite underlying an area of 260 km² sits in between the two masses of anorthosite. The complex has an overall dimension of 84 km by 24 km, its main axis pointing in an ENE direction. The complex is divided into zones of contrasting structural and compositional characteristics: a megacrystic gabbroic anorthosite zone, an anorthosite zone, a banded zone and a border zone (Riccio 1981a , Riccio 1981b). The central portion of the complex is occupied by the megacrystic gabbroic anorthosite zone, which extends to the northeastern margin of the intrusion (Riccio 1981a, Riccio 1981b). It is dominated by

gabbroic anorthosite with subordinate anorthosite, anorthositic gabbro and gabbro, and minor melanogabbro (Riccio 1981a, Riccio 1981b). Rocks within this zone have large (1 to 50 cm) phenocrysts (Percival 1981). Grains of pyroxene with amphibole or garnet coronas are common within this zone (Percival 1981). Olivine appears rarely, at the center of orthopyroxene grains (Percival 1981). The northwestern part of the intrusion shows the anorthosite zone, which consists of granular aggregates of plagioclase, with minor amphibole and rare garnet (Riccio 1981a, Riccio 1981b). The southeastern margin of the complex is where the banded zone appears. It is characterised by bands of 1-30 cm thick of pure anorthosite, clinopyroxene-hornblende-plagioclase gabbro, sphene-garnet-amphibole-plagioclase-quartz rock, garnet-rich garnet-hornblende-quartz rock and minor ultramafic rocks (Riccio 1981 a, Riccio 1981b). The 50-100 m thick border zone separates anorthositic rocks from country rocks in the Carty-Harold Lake area and along the northwestern margin of the complex (Riccio 1981a, Riccio 1981b). This zone consists of foliated to gneissic garnetiferous amphibolite with 5-20% concordant to discordant tonalite bands (Percival 1981).

Mafic dikes have been observed in the KSZ, in two different swarms (Percival 1981). The Kapuskasing swarm is characterized by ENE-striking, SE-dipping dikes of 1-10 m wide, consisting of sparsely plagioclase-porphyrific, medium-to-fine-grained, ophitic, green-grey gabbro (Percival 1981). The second swarm is of NE-trending dikes with pitted weathering rusty surface which may be part of the Abitibi swarm (Percival 1981).

In the Kapuskasing Structural Zone, four alkalic rock-carbonatite complexes are identified. Three are located within 20 km of Borden Lake: Borden, Nemegosenda and Lackner Alkalic complexes. The fourth complex is the Shenango (Figure 2). All four are easily noticeable on aeromagnetic surveys, appearing like round highs on the maps. The Shenango Alkalic complex contains different rock suites than the Borden, Nemegosenda and Lackner complexes. The Shenango contains diorites, monzodiorites, quartz monzonites and granites (Sage 1987c), which differs from the other carbonatites found in the other intrusives of similar age; it is a silica-saturated series of rocks (Sage 1987c).

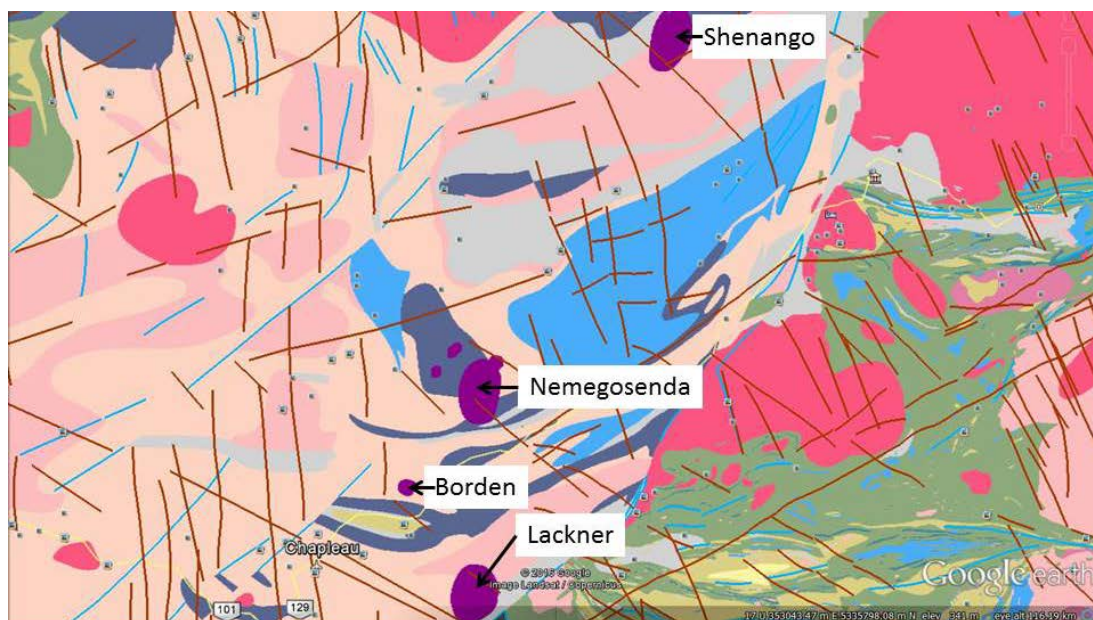


Figure 3: Alkalic complexes of the Kapuskasing Structural Zone

The Nemegosenda Lake Alkalic Rock Complex is made of fine-to-medium-grained syenitic to malignitic rocks which are limited along the southern border of the complex by an arcuate mass of coarse-grained nepheline syenite (Sage 1987b). The complex is contained within a fenitized envelope produced by metasomatism by alkali-iron-rich aqueous fluids from the crystallizing magma (Parsons 1961). Ijolites occur in the northwest corner of the complex, and a mass of gabbro rests along the northwest margin and an isolated band along the east flank. Parsons (1961) noted fresh unmetamorphosed nature of the gabbro units but classified them as country rocks. Sage (1987b) believed that they are an early phase of the alkalic magmatism, perhaps analogous to the gabbroic margins observed at the Port Coldwell and Killala Lake complexes located north of the northeast corner of Lake Superior.

The Borden Township Alkalic Rock Complex intruded Archean biotite-quartz-plagioclase gneisses which show fenitization in proximity to the carbonatite intrusion (Sage 1987a). According to Sage (1987a), sovite, silicocarbonatite and fenitized wall rock alternate throughout the length of the examined drill core. It suggests that the complex likely represents a succession of sovite and silicocarbonatite cone sheets emplaced into brecciated biotite-quartz-plagioclase gneisses (Sage 1987a).

The Lackner Lake Alkalic Rock Complex consists of core and peripheral nepheline syenites, separated by a medial, arcuate, partial ring of alkali mafic rocks like ijolite and malignite (Sage 1988). The leucocratic coarse to very coarse-grained, nepheline syenites of the core and periphery cannot be differentiated by mineralogy, textural or crosscutting field relations (Sage 1988). In proximity to the arcuate band of mafic rocks, the syenites are strongly to weakly trachytoid texture, are locally finer-grained, and contain abundant subangular to subrounded mafic xenoliths. The xenoliths, commonly biotite-rich, are derived from the more mafic phases of the complex which are cut by the nepheline syenites (Sage 1988). Dikes of magnetite-apatite cut both the mafic and syenitic phases of the complex, which is also cut by lamprophyre dikes (Parsons 1961).

Cataclastic rocks are present in the transition between the Western Abitibi subprovince and the Kapuskasing Structural Zone, especially in the southern Ivanhoe Lake-Ivanhoe River area. The area is characterized by mylonite and cataclasite (Percival 1981). Pseudotachylite and ultra-cataclasite veinlets are developed within the high-grade rocks of the KSZ, including anorthosite, northwest of Ivanhoe Lake (Riccio 1981a, b). The Ivanhoe Lake cataclastic zone, the zone affected by cataclasis, is 1-2 km in width and may have more than one discrete fault zone (Percival 1981).

Structure

The rocks of the Kapuskasing Structural Zone are characterized by gneissosity, defined by small to major differences in mineralogical composition of individual bands ranging from 1 to 90 cm in thickness (Thurston et al. 1977). The orientation of lithological contacts and gneissosity make up the prominent east-northeast structural grain of the Kapuskasing structural zone. Gneissosity in all rock types is folded or warped about gently-plunging (0-25°) northeast-trending axes (Percival 1987). The folds vary from northwest-facing monoclinal flexures to isoclinal with consistent "Z" sense asymmetry when viewed toward the east. Axial surfaces are rarely accompanied by a foliation defined by flattened quartz grains. The trend of fold axes and lineations is northeast-southwest throughout this part of the Kapuskasing zone, but plunge direction varies on a regional scale from dominantly southeasterly in the south to northeasterly in the north. Between these areas, lineations are within 10° of horizontal and abrupt changes in plunge

direction occur on the 100 m scale. Both regional and local plunge reversals can be related to a gently southeast-plunging warp axis (Percival 1987).

Plausible Explanations

To this day, the Kapuskasing Structural Zone has seen various explanations for its presence. The origin of the structure has been interpreted as “thinning of the granitic upper crustal layer” (Garland 1950), Mid- to Late Proterozoic rifting (MacLaren et al. 1968; Bennett et al. 1967; Burke and Dewey 1973; Thurston et al. 1977) a suture between the western and eastern Superior Province (Wilson 1968), dextral transcurrent faulting (Goodings and Brookfield 1992), and an east-verging thrust exposing an oblique crustal cross section (Percival and Card 1983; Percival 1986). Percival and West (1994) criticize these models since no simple model is capable of answering all the present features.

The suture model does not account for the correlativity in age and geology between the Abitibi and Wawa subprovinces. A horst model describes the geometry of the blocks within the KSZ, but the third dimension shows a better consistency with a compressional origin than extension. Also, the KSZ existed before the emplacement of the carbonatites and alkali rock complexes around 1100 Ma. The dextral transcurrent model is based on northeasterly aeromagnetic and map trends in the KSZ. Other explanations for these trends are possible, and the scale of any transcurrent displacement seemed to be minor (< 20km). The east-verging Early Proterozoic thrust fault seems to explain many features of the KSZ. However, the Matachewan dike swarm displacement noted along the southern end of the KSZ does not seem possible given its semicontinuous nature, which shows 55-70 km west-over-east apparent movement (Percival and West 1994).

Percival and West (1994) use a model which follows this explanation:

“Supracrustal rocks of the Kapuskasing Zone (2750-2700 Ma) were buried by younger supracrustal rocks, tectonic shortening, and intrusion of mid-crustal tonalities (2700-2660 Ma). Metamorphism began during this period and continued in response to magmatic heat and crustal collapse (2660-2625 Ma). The Borden Lake conglomerate was deposited in a pull-apart basin and transported into the hot deep crust as a sliver along a downward vector on a dominantly transcurrent fault. Kapuskasing rocks remained at depth, where they cooled slowly and underwent intermittent minor deformation (2625-2585 Ma). The Superior Province was eroded by 10 km on average, elevating Kapuskasing levels from approximately 30 to 20 km. Incipient breakup of the Superior craton (2500-2450 Ma) was recorded as new mineral growth at deep structural levels and by Matachewan dyke injection. A few additional kilometres of erosion preceded intrusion of Kapuskasing dykes (2040 Ma) into the intact crustal section. At approximately 1.9 Ga, when Kapuskasing levels had cooled to <300°C, stresses caused by plate collisions at the Superior margin were transmitted into the interior in the form of dextral transpression along northeast-trending and northwest-over-southeast thrusting, elevating Kapuskasing-level rocks [...]. Formation of a crustal root accommodated shortening in the lower crust [...]. Northwest-dipping normal faults and a conjugate set of strike-slip faults broke the Kapuskasing structure into separate blocks with variable geometry. During the relaxation phase, isostatic rebound reduced topography on the root and probably produced a few more kilometres of uplift along steep structures not coincident with the Ivanhoe Lake Fault.”

The recent discovery of the Borden Lake deposit shows that perhaps the story is wrong; maybe we are looking at mineralization put into place prior to the uplift.

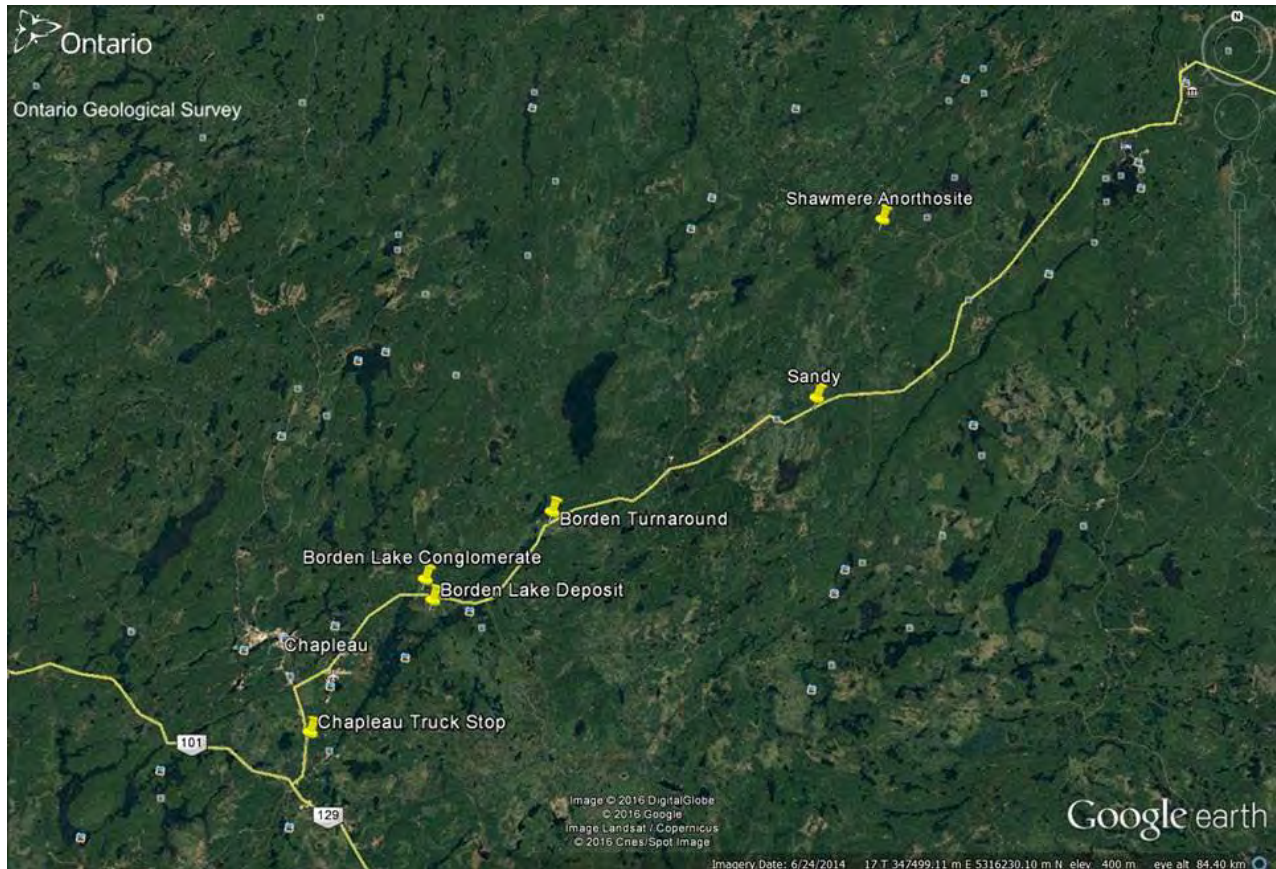


Figure 4: Field trip stops

Stop 1: Shawmere Anorthosite

Zone: 17, Easting: 366922 Northing: 5332558

Access: Take the Warren-Carty main haul road. Travel down the road for about 10 km. The site is accessible from the road through a wet overgrown section which leads to the bulk sample pit created by Avalon Ventures Ltd. back in 2007.

The rocks in the bulk sample pit (Figure 5) are a very coarse-grained anorthosite with 1-4 mm diameter grains, with sporadically some 3 cm diameter porphyroblasts although most are in the 1 cm diameter range. The porphyroblasts are xenomorphic, poikiloblastic and disseminated throughout the rocks. The grains in the matrix are equant and may be polygonized.



Figure 3: Shawmere Anorthosite pit

Stop 2: Sandy Outcrop

Zone: 17, Easting: 360841 Northing: 5318167

The outcrop shows granitic gneiss, with melanosome and leucosome in thin beds of 1 to 3 cm in thickness. The melanosome consists of amphiboles, biotite and calcic feldspars, while the leucosome consists of quartz, potassic and sodic feldspars. The folded and wavy beds underline migmatization of the rock. A diabase dike crosses the outcrop with seeming difficulty.

Stop 3: Borden Turnaround Outcrop

Zone: 17, Easting: 340126 Northing: 5309885

The outcrop (Figure 6) shows a seemingly partially undifferentiated granitic migmatite. Dark paleosomes are seen on the outcrop, composed of biotite and amphiboles (restites). Garnets are rather invisible, some trace sulphides are present. Behind the outcrop, on boulders broken off during the road construction, it is possible to see an “augen” (Figure 7) made out of a piece of possibly chert, which may suggest that the protolith may be a metasediment.



Figure 4: Turnaround Outcrop



Figure 5: "Augen" made of possibly chert, Turnaround Outcrop. Notice the zoning in the eye, which is perpendicular to the gneissosity of the rock.

Stop 4: Borden Lake Conglomerate

Zone: 17, Easting: 330441 Northing: 5305051

The Borden Lake Conglomerate is a deformed metaconglomerate (Figure 8) which was dated at 2664 ± 12 Ma (Percival et al. 1981), using tonalitic cobbles. The age date of the zircons is of a later deformation-metamorphic event. Some texts refer to it as a “pebble” conglomerate, but most clasts in the outcrop are more in the “cobble” range of sizes. The rock is vastly clast supported with a 10-15% matrix of quartz-biotite-hornblende-garnet. The cobbles are made of felsic metavolcanics, metasediments, granodiorite, tonalite, plagioclase-porphyritic meta-andesite and amphibolite, with some hornblende and quartz veins. The foliation is gently north dipping. They are elongated, forming rods lineated at 075°N , and a plunge of 026° (Figures 8 and 9). The metaconglomerate is accompanied spatially by an amphibolite and paragneiss to the south of the highway. Somehow, the amphibolite is the host of the gold mineralization of the Borden Lake deposit.

To the south of the outcrop, towards the highway, lies a felsic porphyry unit (Figure 10). The unit is grey in appearance with white “eyes” of quartz porphyroblasts. The porphyroblasts are up to 1 cm in diameter. Reddish garnets can be seen on the outcrop.



Figure 6: Borden Lake conglomerate, with elongated cobbles



Figure 7: Borden Lake conglomerate, down plunge view



Figure 8: Quartz porphyry, with quartz eyes up to 1 cm in diameter.

Stop 5: Chapleau Truck Stop Outcrop

Zone: 17, Easting: 321942 Northing: 5294079

The roadcut (Figure 11) shows sub-vertical orientation of mafic xenoliths in a medium grained tonalite, showing a “salt & pepper” texture. A pegmatitic aplite vein (~10 cm in thickness) crosscuts the outcrop. The abundance of mafic xenoliths seems to increase towards the south of the outcrop along the highway.



Figure 9: Truckstop outcrop with subvertical xenoliths and cut by a dark pink aplite vein

Stop 6: Borden Lake Deposit

The Borden Lake gold deposit lies at the intersection of the Wawa sub-province, the Kapuskasing Structural Zone (KSZ) and the Abitibi sub-province, specifically within the southernmost limits of the KSZ (Dzick 2014). It is located on a peninsula of Borden Lake about 20km from the town of Chapleau.

The deposit dips at 40-45° to the north-northeast and strikes to the west-northwest. This trend is bound to the north by an extensive package of metaconglomerates and intersected by a significant east-west striking fault. To the south it is bound by a similar package of metaconglomerates, which then grades into various garnet-boitite gneisses, amphibolites and granulite-grade gneisses.

The gold mineralization within the deposit is typically characterized by a higher-grade core surrounded by a lower-grade envelope, within a package of volcano-metasedimentary rocks of variable composition. The main sulphides are pyrite and pyrrhotite, with the latter typically dominating within the gold zone. The west-northwest portion of the deposit is generally lower grade with some higher grade pods. The east-southeast portion of the deposit hosts the High-Grade Zone (Dzick, 2014).

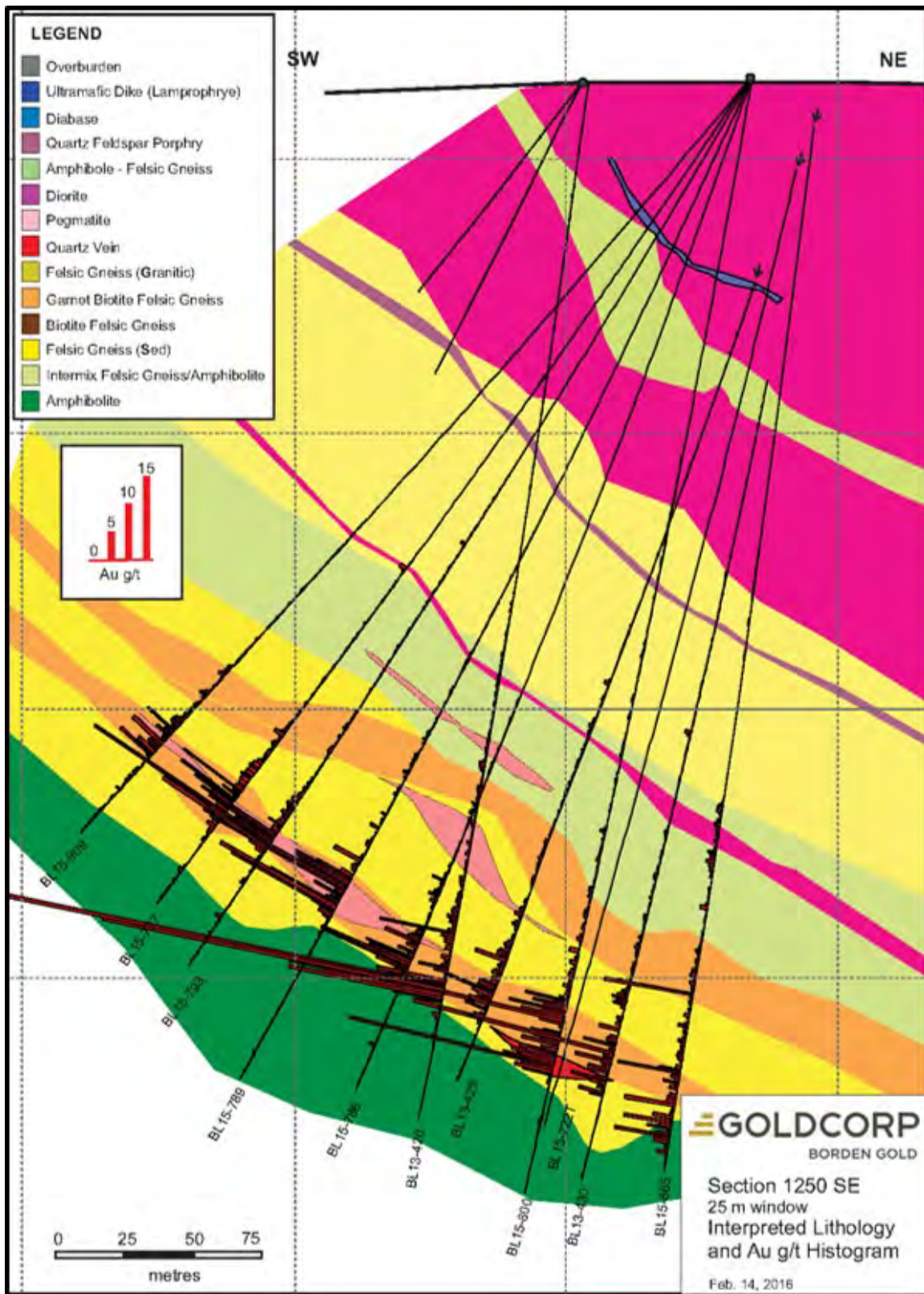


Figure 10

An interpreted geological drill section (Figure 12) intercepts the High-Grade Zone (HGZ) at the south eastern portion of the deposit. The general hanging wall sequence of rocks is an intermixed package of felsic intrusive (quartz-metadiorite to diorite) sills/dikes with felsic metavolcanic-metasedimentary gneisses. This upper sequence of felsic rocks is cross-cut or intermixed with a coarse grained porphyroblastic metagabbro (locally termed amphibole felsic gneiss) and amphibolite units. Sulphide mineralization in the hanging wall is dominantly pyrite in amounts ranging from trace to 1%. The appearance of pyrite and pyrrhotite mineralized amphibolites marks the transition from the hanging wall to the beginning of the gold-bearing zone and the start of the low-grade gold envelope.

The low-grade envelope (Figure 13) is typically characterized by intermixed felsic gneisses and amphibolites with an overall increase in sulphide mineralization to 1-2% pyrite/pyrrhotite. The increase in sulphide content generally coincides with a higher degree of strain and development of a foliation that becomes better defined and more intense down sequence. The transition from the low-grade envelope to the beginning of the high-grade core is marked by the presence of a unit that is locally termed the granitic gneiss. The highest grade portions of the gold zone correlate with deformed quartz veins and quartz pegmatites hosted within small sections of garnet-biotite gneiss and amphibolite. All main sequence units within the core can be silicified to varying degrees, with the units proximal to the quartz veins and pegmatites generally undergoing the strongest intensity of alteration. Visible gold is most often associated with quartz veining and areas of intense silicification.

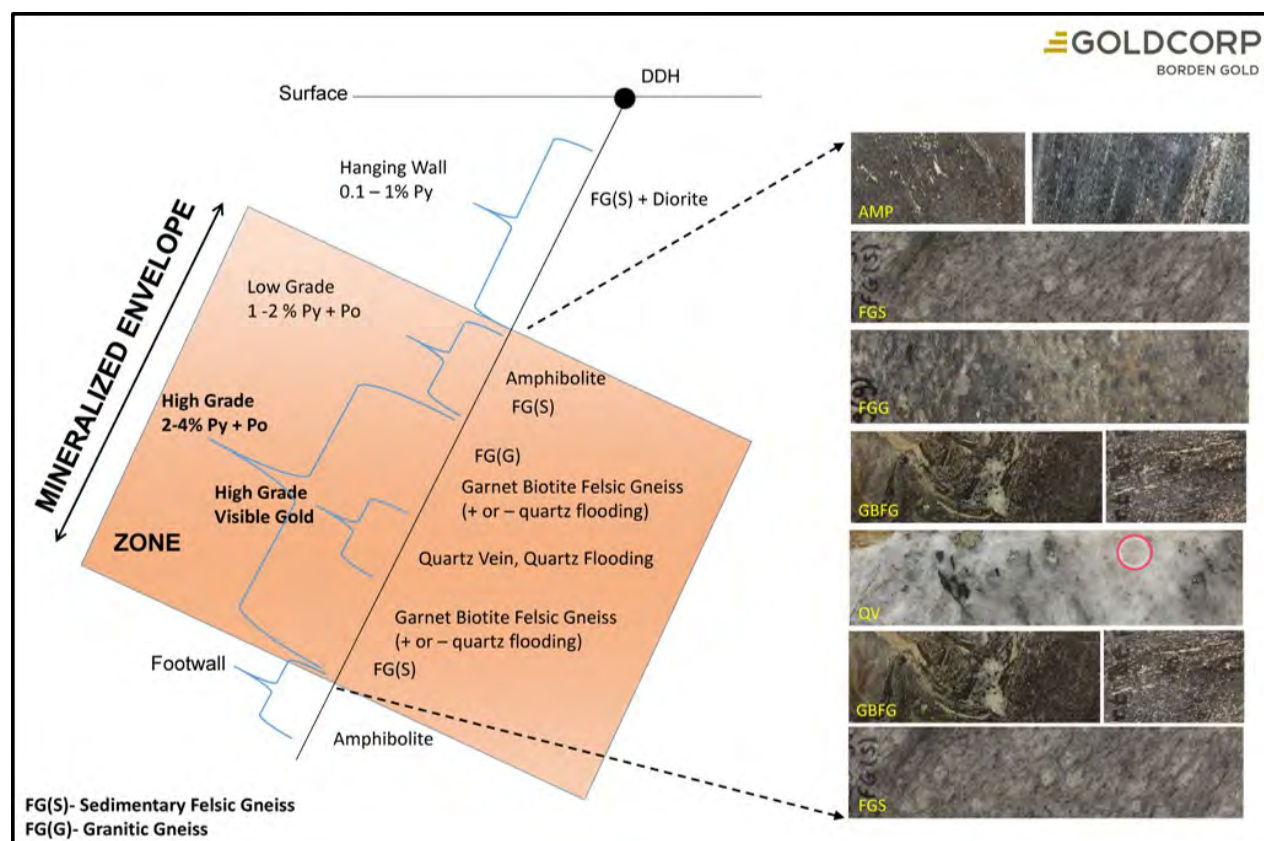


Figure 11

The Borden Lake deposit is located on a peninsula on Borden Lake. A volcano-sedimentary package of various composition hosts the gold mineralization, which occurs as a wide zone of disseminated and fracture-controlled sulphides. Pyrite and pyrrhotite are the main sulphides present, with pyrite being the most common. Gold occurs in a high-grade core surrounded by a low to moderate grade envelope, accompanied by minor silver. The grade of the core seems to be improving towards the southeast where it becomes the high-grade zone (HGZ) with average grades above 2.5 g/t gold (Dzick, 2014).

The northwest section of the deposit is characterized by sporadic silicification without lithological control or quartz veining. The southeast section, on the other hand, has a well-developed hydrothermal system with quartz flooding and potassic alteration which defines the HGZ. Various host rocks contain the mineralization, usually dominated by metasedimentary horizons and subordinate intrusives of felsic to intermediate composition, who display feldspathic, chloritic and biotitic alterations. Outcrops in the northwest rarely show visible gold, which is the opposite of the HGZ in the southeast, especially in the quartz-rich core. That core is present in the low-grade zone, sometimes attaining very high grades like the HGZ (Dzick, 2014).

The deposit showed continuity and has been consistently intersected along strike, reaching a length of 3.7 km (Dzick, 2014). The deposit remained open both to the northwest and southeast directions. The dip and plunge of the deposit are to the northeast and shallow southeast respectively. Gold mineralization seemed to be controlled by a ductile shear zone, which is more apparent in the HGZ. The mineralized zone has been confirmed to a vertical depth of approximately 650 m, and is up to 120 m wide (Dzick, 2014).

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