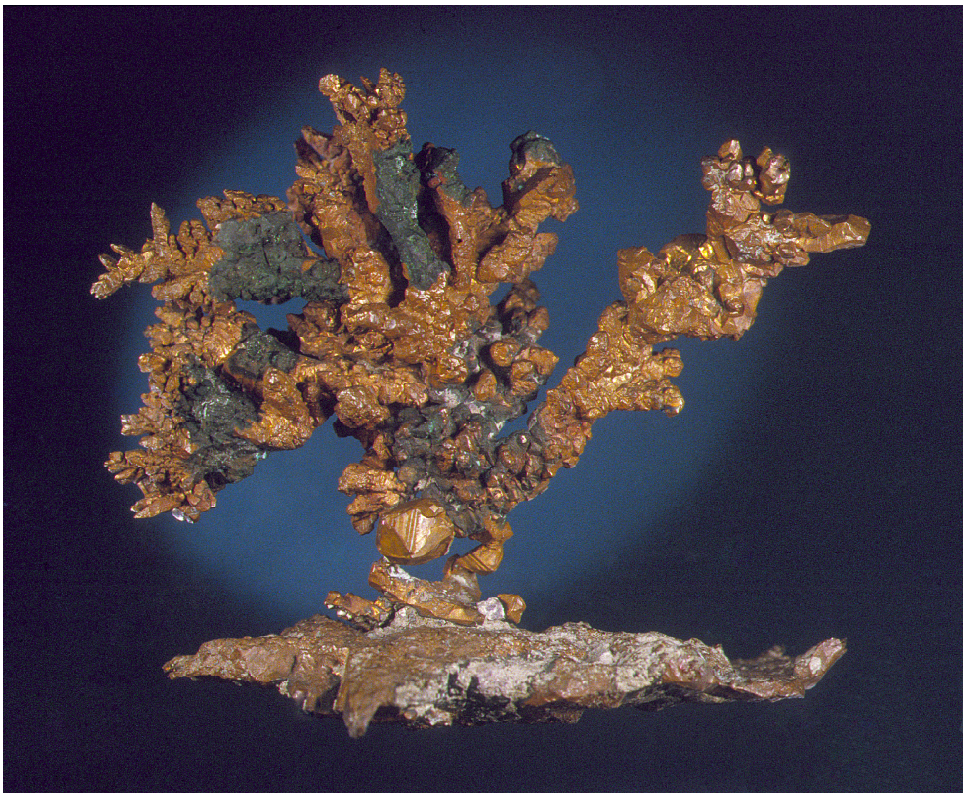
# Institute on Lake Superior Geology

## 59th Annual Meeting

**Houghton, Michigan May 8 - 11, 2013**



Proceedings Volume 59
Part 2 - Field Trip Guidebook

**Editors: Theodore J. Bornhorst and Robert J. Barron**

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**Institute on Lake Superior Geology**

# 59TH ANNUAL MEETING
# MAY 8-11, 2013
# HOUGHTON, MICHIGAN

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# Proceedings Volume 59

# Part 2 – Field Trip Guidebook

***EDITED BY THEODORE J. BORNHORST AND ROBERT J. BARRON***

**Cover Photo: Native copper from the Central Mine, Keweenaw Peninsula, Michigan. Collection of the A.E. Seaman Mineral Museum. Photograph by George Robinson.**

# 59TH INSTITUTE ON LAKE SUPERIOR GEOLOGY

## PROCEEDINGS VOLUME 59 CONSISTS OF:

**PART 1: PROGRAM AND ABSTRACTS**

**PART 2: FIELD TRIP GUIDEBOOK**

- **TRIP 1:** GEOLOGIC OVERVIEW OF THE KEWEENAW PENINSULA, MICHIGAN
- **TRIP 2:** CALEDONIA MINE, KEWEENAW PENINSULA NATIVE COPPER DISTRICT, ONTONAGON COUNTY, MICHIGAN
- **TRIP 3:** GEOLOGY OF SILVER MOUNTAIN, HOUGHTON COUNTY, MICHIGAN
- **TRIP 5:** GEOLOGY OF THE KEWEENAWAN SUPERGROUP, PORCUPINE MOUNTAINS, ONTONAGON AND GOGEBIC COUNTIES, MICHIGAN
- **TRIP 6:** GEOLOGY AND ENVIRONMENTAL SITE CONDITIONS OF THE COPPERWOOD DEPOSIT, GOGEBIC COUNTY, MICHIGAN

**Reference to material should follow the example below:**

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Peter Hollings
Department of Geology
Lakehead University
Thunder Bay, ON
P7B 5E1
CANADA
peter.hollings@lakeheadu.ca

**ILSG website: http://www.lakesuperiorgeology.org**

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# Field Trip 1

# Geologic Overview of the Keweenaw Peninsula, Michigan

Theodore J. Bornhorst
*A.E. Seaman Mineral Museum, Michigan Technological University, 1404 E. Sharon Avenue, Houghton, MI 49931*

Robert J. Barron
*Department of Geological and Mining Engineering and Sciences, Michigan Technological University, 1400 Townsend Drive, Houghton, MI 49931*

## Introduction

The geology of the far western Upper Peninsula of Michigan consists of three temporally distinct episodes. During the Mesoproterozoic, between about 1.15 and 1.03 Ga, up to 30 km of Keweenaw Supergroup volcanics and clastic sediments filled an intracratonic rift, the Midcontinent Rift (MCR) (Figs. 1 and 2) (Heaman et al., 2007; Davis and Paces, 1990; Cannon et al., 1989). After a 500 million year period of erosion, the MCR rocks were buried by Phanerozoic sedimentary rocks from about 500 Ma to 175 Ma (Catacosinos et al., 2001). Pleistocene continental glaciations, beginning about 2 million years ago, removed the Phanerozoic rocks from the Keweenaw Peninsula leaving only a few outliers. About 10,000 years ago, as the last remaining glaciers retreated, and they left behind a variety of unconsolidated clastic sediments. The geologic evolution of the far western Upper Peninsula is illustrated in cartoon form in Figure 3.

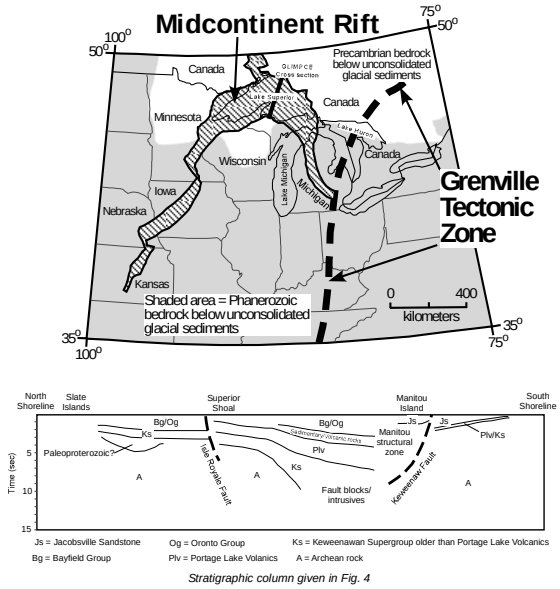



Figure 1: Generalized bedrock geologic map of the Midcontinent Rift. Grenville tectonic zone after Cannon (1994) and interpretative cross-section across the Lake Superior segment of the Midcontinent rift after Cannon et al. (1989).

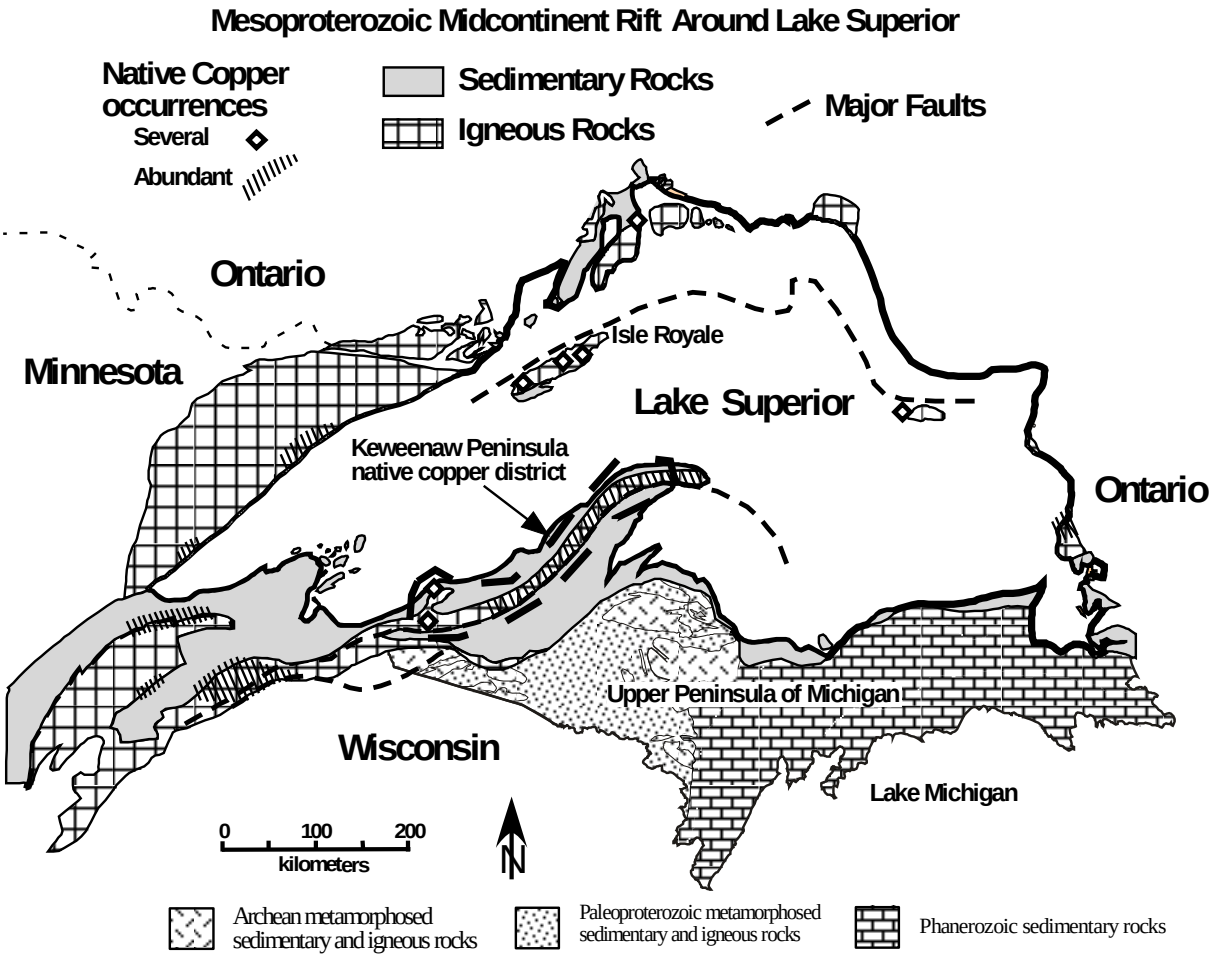



Figure 2: Generalized bedrock map showing the exposed rocks of the Midcontinent Rift around Lake Superior and the bedrock of the Upper Peninsula of Michigan. Locations of concentrations of native copper are shown around Lake Superior. Modified from Bornhorst and Barron (2011).

In the strictest sense, the geographic area of the Keweenaw Peninsula proper extends northeast of a NW-SE line drawn through L'Anse (Fig. 4), however, the term Keweenaw Peninsula has also been applied to the area containing MCR rocks farther to the south. Bornhorst and Barron (2011) used the name Keweenaw Peninsula native copper district to describe native copper deposits hosted by MCR rocks as far south as the White Pine Mine. The geologic description in this field trip guide is restricted to the Keweenaw Peninsula proper and does not include the southern bedrock such as at Silver Mountain described in Field Trip 3 (this volume) or Cambrian to Devonian rocks at Limestone Mountain described by Milstein (1987).

The descriptions of the geology of the Keweenaw Peninsula provided here were modified from a combination of Bornhorst and Barron (2011), Bornhorst and Lankton (2009), and Bornhorst and Rose (1994), and Bornhorst et al. (1983). Specific citation or quotation is not given in all instances.

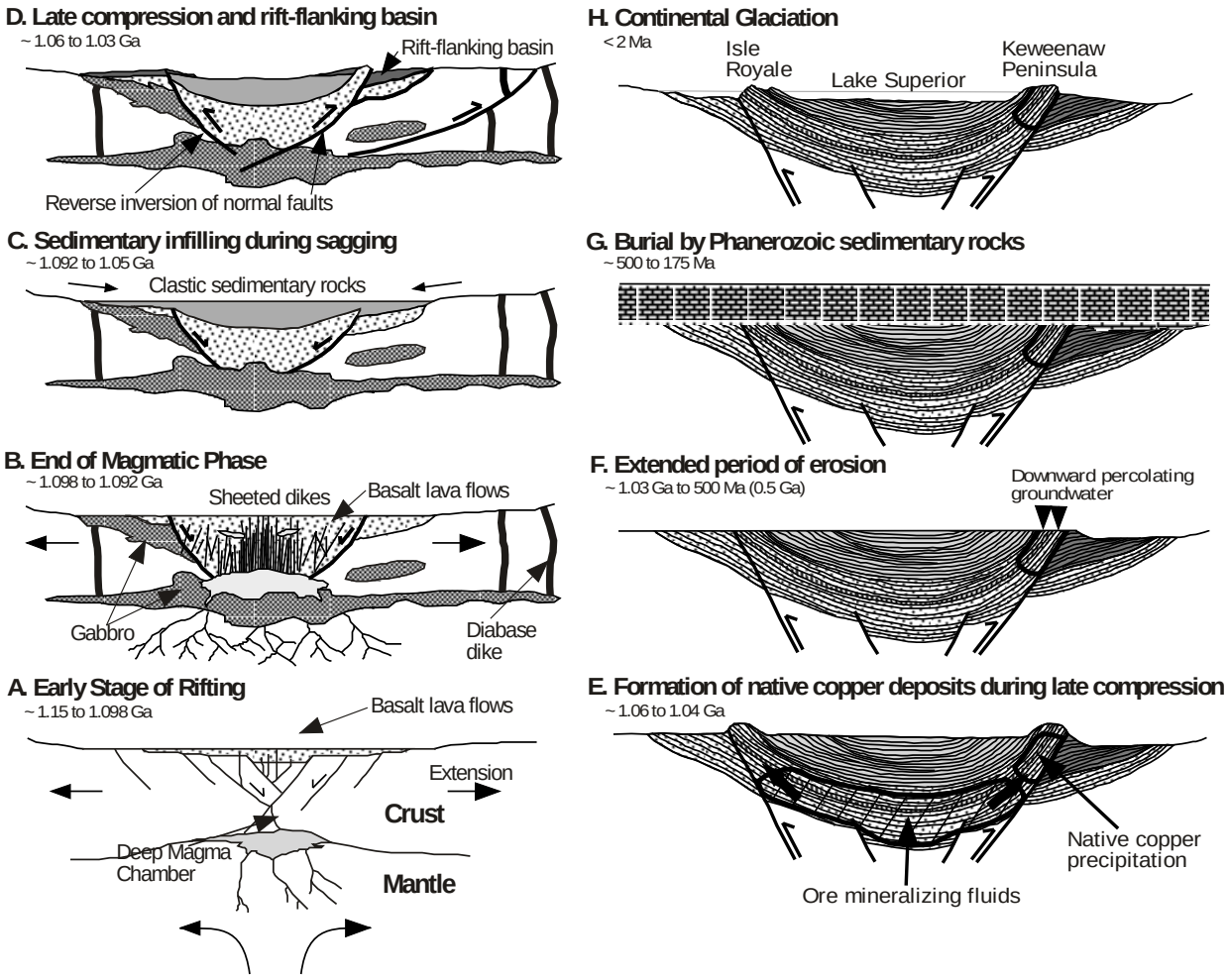



Figure 3: Cartoon NW to SE cross sections from Minnesota (left) to the Upper Peninsula (right) illustrating the progressive geologic evolution. Modified from Bornhorst and Lankton (2009).

## Midcontinent Rift Lithologic Units

The Keweenaw Peninsula is located on the southern margin of the Lake Superior segment of the MCR (Figs. 1and 2). The rock units that are associated with the MCR have been termed the Keweenawan Supergroup (Fig. 5). These rocks were deposited from about 1.15 and 1.03 Ga (Heaman et al., 2007; Davis and Paces, 1990; Cannon et al., 1989). The MCR beneath Lake Superior is filled with up to about 30 km of volcanic rocks (Figs. 1 and 3) (Hinze et al., 1990; Cannon et al., 1989; Cannon, 1992).

The MCR geology of the Keweenaw Peninsula can be divided into northwest-dipping, rift-filling volcanic and clastic sedimentary rocks located on the northwest side of the Keweenaw Peninsula and flat to low-dipping, rift-flanking clastic sedimentary rocks located on the southeast side (Fig. 4). These two contrasting lithologic settings are separated by the Keweenaw Fault which was originally a graben-bounding fault, but today is a high-angle reverse fault (Fig. 3).

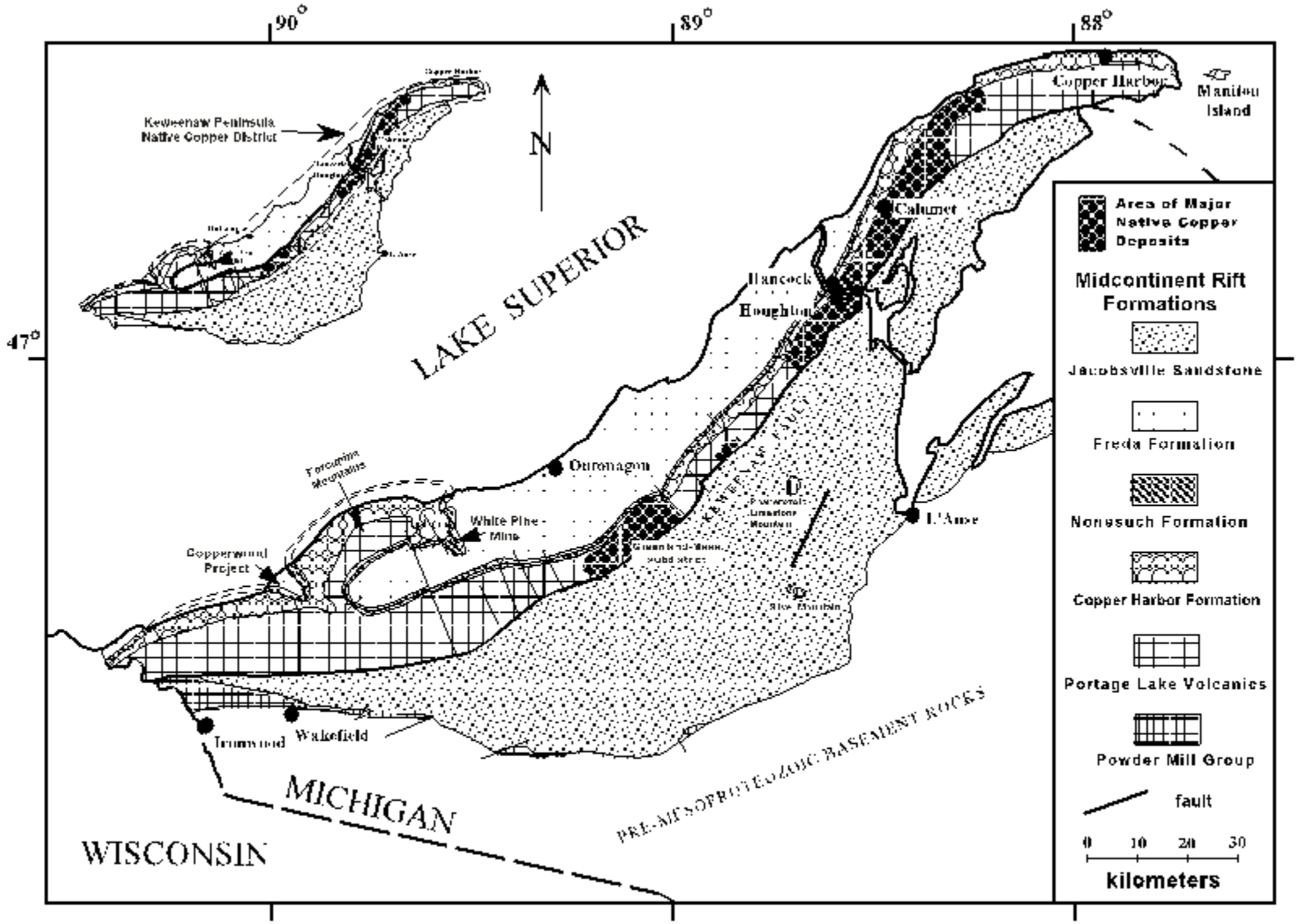



Figure 4: Geologic map of the far western part of the Upper Peninsula of Michigan showing the location of the Keweenaw Peninsula native copper district.

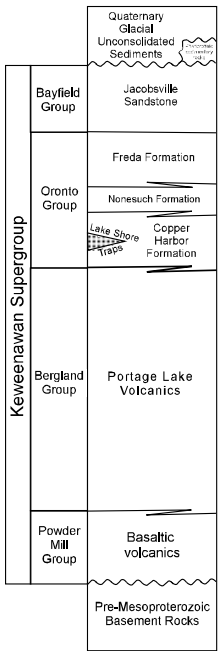



Figure 5: Lithostratigraphic bedrock units of the Mesoproterozoic Midcontinent rift system of Michigan.

*Portage Lake Volcanics*

The Portage Lake Volcanics (Figs. 4 and 5) is a 2,500 to 5,200 m thick formation dominantly composed of subaerial basalt lava flows with less than 1 % by volume intermediate to felsic volcanic and subvolcanic rocks located stratigraphically near the base of the exposed formation. Less than 5 % by volume is stratigraphically scattered interflow reddish-colored conglomerate and sandstone units that are greater in abundance towards the top of the formation (Butler and Burbank, 1929; White, 1968). The base of the formation is truncated by the Keweenaw Fault. The lavas flowed from fissure vents that tended to be located nearer the axis of the rift zone which produced a layered succession of flood basalts comparable to the rift zones of East Africa and Iceland (e.g., Nicholson, 1992 and reference therein). The Portage Lake Volcanics erupted over 2 to 3 million years from 1,096.2+/-1.8 (Copper City flow, Fig. 6) to 1,094.0+/-1.5 (Greenstone flow, Fig. 6) (Paces and Miller, 1993; Davis and Paces, 1990).

There are more than 200 individual basaltic lava flows in the exposed Portage Lake Volcanics which are typically aphyric, Mg-rich, high-Al olivine tholeiites (Paces, 1988). The most abundant type of basalt flows are olivine tholeiites, followed by primitive olivine tholeiites and quartz tholeiites. Iron-rich olivine tholeiites are generally lesser in abundance (Table 1). The thicker lava flows are compositionally stratified due to magmatic differentiation after eruption, especially the Greenstone flow, which is the thickest individual flow in the formation (Cornwall, 1951a and b; Broderick, 1935; Broderick and Hohl, 1935). The composition of the basalts is cyclical with minor and major cycles superimposed on an overall trend toward more primitive compositions towards the top of the formation. The basalt magmas were derived by partial melting of sub-continental upper

mantle with an overall stratigraphically upwards trend towards younger, to more primitive basalt compositions as a result of less contamination by crustal rocks (Paces, 1988; Paces and Bell, 1989). The repeated magmatism at the rift axis and progressive crustal thinning provided pathways for transport of magma to the surface creating less extended contact with crustal rocks and hence, less contamination. The youngest rocks of the Portage Lake Volcanics in the Keweenaw Peninsula have compositions similar to MORB suggesting the MCR nearly formed an ocean basin. The major geochemical cycles are due to fractional crystallization and replenishment in large magma chambers near the crust/mantle interface whereas the minor cycles are due to closed system fractional crystallization in small magma chambers within the crust (Paces, 1988). The Portage Lake Volcanics were likely derived by partial melting of trace element enriched plume-related mantle (Nicholson et al., 1997; Nicholson and Shirey, 1990; Paces and Bell, 1989).

Table1: Average and representative geochemical data for least altered lavas of the Portage Lake Volcanics (from Paces, 1988). Tholeiites were grouped by Ni content.

| | Primitive olivine tholeiite | Olivine tholeiite | Olivine tholeiite | Intermediate olivine tholeiite | Iron-rich olivine and quartz tholeiites | Andesite | Dacite | Rhyolite |
|---|---|---|---|---|---|---|---|---|
| Ni (ppm) | 400-300 | 300-250 | 250-200 | 200-100 | 100-15 | | | |
| | n=5 | n=9 | n=14 | n=8 | n=6 | n=1 | n=1 | n=1 |
| **Wt.%** | | | | | | | | |
| $SiO_2$ | 47.82 | 47.34 | 48.03 | 48.55 | 49.94 | 56.39 | 68.44 | 77.89 |
| $Al_2O_3$ | 15.89 | 15.27 | 15.32 | 15.12 | 13.28 | 13.78 | 15.17 | 12.77 |
| FeOt | 9.77 | 11.82 | 12.32 | 12.86 | 14.91 | 9.87 | 4.46 | 1.11 |
| MgO | 12.44 | 11.69 | 9.85 | 9.06 | 7.78 | 5.52 | 1.14 | 0.17 |
| CaO | 10.58 | 10.24 | 10.16 | 9.65 | 6.64 | 5.10 | 1.40 | 0.04 |
| $Na_2O$ | 2.04 | 2.10 | 2.25 | 2.31 | 2.91 | 3.94 | 4.74 | 3.67 |
| $K_2O$ | 0.19 | 0.22 | 0.33 | 0.42 | 1.43 | 2.27 | 3.86 | 4.28 |
| $TiO_2$ | 0.98 | 1.13 | 1.35 | 1.60 | 2.34 | 1.83 | 0.51 | 0.08 |
| $P_2O_5$ | 0.16 | 0.19 | 0.22 | 0.25 | 0.36 | 1.00 | 0.19 | 0.01 |
| MnO | 0.14 | 0.16 | 0.16 | 0.18 | 0.24 | 0.30 | 0.08 | 0.01 |
| **PPM** | | | | | | | | |
| Ni | 326 | 279 | 231 | 172 | 54 | 10 | 7 | 5 |
| Cu | 37 | 51 | 73 | 86 | 126 | 5 | 13 | 61 |
| Zr | 78 | 85 | 101 | 126 | 212 | 430 | 573 | 145 |

FeOt=total Fe as FeO

All observed basalt lava flows in the Portage Lake Volcanics were erupted subaerially and consist of a massive (vesicle-free) interior capped by a vesicular and/or brecciated flow top. The only evidence for involvement of water during eruption is a single thin hyaloclastic unit in the upper part of the formation (locally termed the ashbed). Subaerial eruption resulted in degassing of volatiles, notably $SO_2$ (Cornwall, 1951c). The lava flows range in thickness from 1 to 450 m with most of them between 10 to 20 m thick (Paces, 1988; White, 1960). Most of the lava flows cannot be traced along strike with confidence although a few such as the Scales Creek, Kearsarge, and Greenstone flows have well documented lateral continuity flows (Fig. 6). The Greenstone flow has been correlated down dip across the Lake Superior syncline to Isle Royale (Longo, 1982; Huber, 1975). The uppermost 5 to 20% of the tops of most (89 %) individual lava flows are vesicular with between 5 and 50% vesicles (White, 1986). The tops of 21 % of the flows are brecciated with clasts of vesicular basalt. The vesicles in most lava flows within the Portage Lake Volcanics, except for the stratigraphically uppermost, are filled with secondary minerals and are amygdules. Thus, amygdaloids are lava flows with vesicle-only tops and fragmental amygdaloids are those with vesicular and brecciated tops.

There are the minor amounts of andesite, dacite, and rhyolite lava flows and subvolcanic plutons that interfinger with and cross cut the basalts of the Portage Lake Volcanics (Table 1). Most of these occur in the stratigraphically lowermost portion of the Portage Lake Volcanics. A few dikes of intermediate composition and a diorite stock at Mt. Bohemia intrude the exposed Portage Lake Volcanics. The rhyolitic volcanic setting is analogous to the shield-type central volcanoes of Iceland (Nicholson, 1991).

Interflow sedimentary units are important stratigraphic markers in an otherwise monotonous succession of basalt lava flows that can be traced up to 90 km along strike and so many of them are given informal names (see Fig. 6). They consists of red-colored, well-lithified, pebble-to-boulder conglomerates with lesser amounts of interbedded sandstone and occasional significant amounts of siltstone and shale ranging in thickness from a few cm up to about 40 m (Merk and Jirsa, 1982; White, 1968; Butler and Burbank, 1927). The typical conglomerate has an exposed interflow lithology characterized by sub-rounded to angular pebbles in a sandy matrix. Clast size varies from pebbles to boulders and clast lithologies are predominantly felsic, although there is considerable variation within and between specific beds reflecting diversity in source terrane. Within the interflow Calumet and Hecla Conglomerate, Kalliokoski and Welch (1985) interpreted a subunit as a caliche soil profile. The interflow clastic sedimentary beds were deposited during intervals of volcanic quiescence most likely in terrestrial alluvial fans in an arid to sub-arid climate. Deposition was on top of the shallow-dipping to flat-lying lava flows by streams flowing from the topographic high on the margin of the MCR toward the center of the rift basin (now under Lake Superior) (White, 1968).

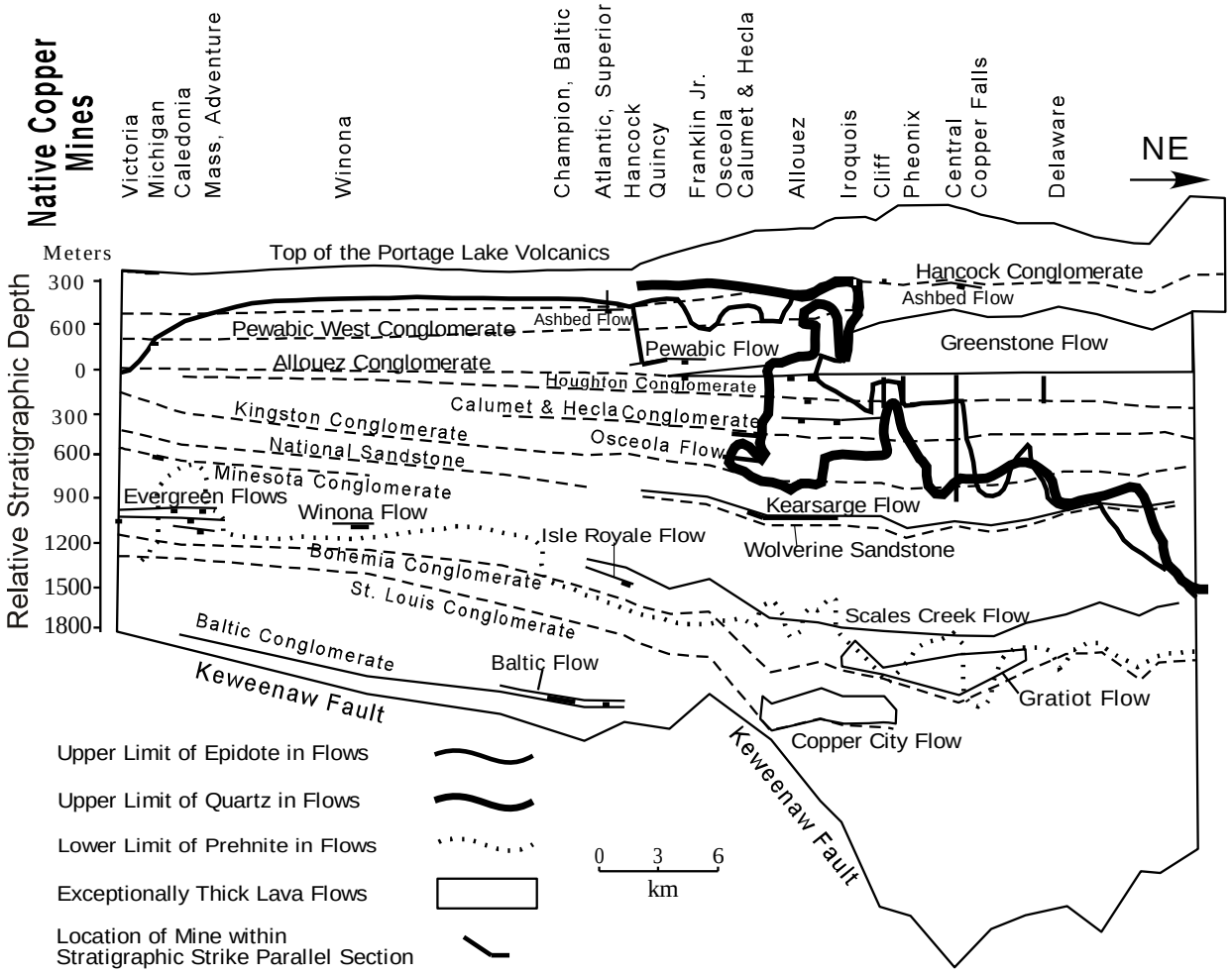



Figure 6: Generalized stratigraphic position parallel to strike of informal units within the Portage Lake Volcanics. Modified from Stoiber and Davidson (1959). Figure 5 shows location of Greenland-Mass subdistrict (Michigan, Caledonia, Mass, Adventure Mines) and Copper Harbor.

*Copper Harbor Formation*

The Copper Harbor Formation is the oldest formation in the Oronto Group and conformably overlies and interfingers with the Portage Lake Volcanics (Figs. 4 and 5). It consists of red-brown clastic sedimentary rocks with a maximum exposed thickness 2,000 m. The Copper Harbor Formation in the Keweenaw Peninsula includes a succession of subearially deposited lava flows informally named the Lake Shore Traps. The depositional environment of the Copper Harbor Formation was deposited in a prograding coalescing alluvial fan complex with proximal-to-distal braided stream and sheet flood facies on the alluvial fans to distal sand flats and flood plain facies (Elmore, 1984). The climate was probably arid with flashy seasonal streams. The highlands from which the Copper Harbor Formation were derived to the southeast, now buried under the Jacobsville Sandstone.

Conglomerates and sandstones are the dominant lithologies in the Copper Harbor Formation. The formation fines distally and up section, reflecting a waning sediment supply due to progressive erosion of the source area (Elmore, 1984). The poorly-sorted clasts in the conglomerates range in size from granules to boulders that are subrounded to rounded and are mostly volcanic in origin and have a ratio of mafic-to-intermediate + silicic composition of about 2:1 (Daniels, 1982). The conglomerates include clast-supported and matrix- supported varieties; some of the latter are diamictites. The conglomerates are interpreted as high-energy channel deposits on a coalescing alluvial fan (Elmore, 1984). The diamictites are debris flow in origin. Sandstones are predominantly red-brown, subangular-to-angular lithic graywackes with volcanic lithic fragments. These exhibit current-ripples, trough-cross beds, current and parting lineations, and reduction spots. Sandstone interbeds are more common in the upper 2/3 of the formation. The abundant calcite cement in the conglomerate and coarse sandstone was probably deposited as vadose carbonate or caliche (Kalliokoski, 1986). Thin red-colored siltstone and shale interbeds have desiccation cracks and are interpreted as filling abandoned channels on the alluvial fan surface. In the Copper Harbor area, there are also laminated cryptoalgal carbonate beds and ooid lenses occurring within the same general stratigraphic position. These are laterally-linked contorted layers in shale-siltstone that are draped over cobbles and are found as poorly developed mats in coarse sandstone (Elmore, 1983). The laminated carbonate beds are algal stromatolite (genus *Colleria*). The stromatolites formed in shallow, medial fan lakes and possibly abandoned channels on the alluvial fan surface (Elmore, 1983).

The Lake Shore Traps (Lane, 1911), an informal member of the Copper Harbor Formation (Fig. 5), are well exposed near the tip of the Keweenaw Peninsula where the unit is composed of 31 lava flows and one interflow conglomerate about 600 m thick (Paces and Bornhorst, 1985). The composition of the Lake Shore Traps is different than the underlying Portage Lake Volcanics reflecting the change from active rift-filling magmatism to active rift-filling clastic sedimentation with little to no magmatism. The rocks range from Fe-rich olivine tholeiitic basalt at the base to Fe-rich olivine-bearing tholeiitic basaltic andesites to tholeiitic andesites. Strato-geochemical relationships can be explained by a combination of fractional crystallization, parental magma replenishment, and wall rock assimilation (Paces and Bornhorst, 1985). Davis and Paces (1990) report a U-Pb age on zircon of 1087.2 +/- 1.6 Ma for the Lake Shore Traps.

*Nonesuch Formation*

The Nonesuch Formation conformably overlies and locally interfingers with the Copper Harbor Formation (Figs. 4 and 5). It consists of dominantly black-to-gray-to-green fine clastic sedimentary rocks with a maximum exposed thickness 240 m. Exposures in the Keweenaw Peninsula are limited with the best exposure at the Hancock Campground on M-203. This formation will not be visited for this field trip. The Nonesuch Formation was deposited in a generally anoxic lacustrine environment ranging from marginal lacustrine (sandflat-mudflat) to lacustrine to lacustrine-to-fluvial subenvironments (Elmore et al., 1989).

Siltstone and shale are the dominant lithologies with lesser very-fine sandstone and minor carbonate laminates. While gray (reduced) color characterizes most of this formation, the stratigraphic upper beds have more red-brown colors (Bornhorst and Williams, in press). Well-laminated to massive

black to dark-gray siltstone and shale were deposited in the lacustrine subenvironment. The lacustrine lithologies at the base of the Nonesuch Formation host economic quantities of chalcocite and native copper at the now closed White Pine Mine (Mauk et al., 1992) and chalcocite at the Copperwood project (Bornhorst and Williams, in press; and Field Trip 5 this guidebook). A thin carbonate laminate yielded a Pb-Pb isochron age of 1,081 ± 9 Ma (Ohr,1993)

*Freda Formation*

The Freda Formation is the youngest formation of the Oronto Groupand overlies the Nonesuch Formation with no explosed top (Figs. 4 and 5). The contact between the Freda and Nonesuch Formations is gradational. The exposed thickness is greater than 3,700 m, however, it is poorly exposed except along the Lake Superior shoreline. This formation will not be visited for this field trip. The Freda Formation is presumably overlain by the Jacobsville Formation. The Freda Formation was deposited in an environment characterized by shallow meandering streams (Daniels, 1982).

Red-brown fine to very-fine sandstone, siltstone, and mudstone are the dominant lithologies in the Freda Formation. Fining-upward sequences occur on the scale of a few meters. Based on regional correlations the age of the Freda Formation is likely 1,060 to 1,040 Ma (Cannon, 1992).

*Jacobsville Sandstone*

The Jacobsville Sandstone is the youngest Mesoproterozoic bedrock formation in the Keweenaw Peninsula (Figs. 4 and 5). Its stratigraphic relationship with other units is uncertain. It occurs in a contiguous geographic region bound on the northwest by the Keweenaw Fault and to the southeast, it angularly unconformably overlies Paleoproterozoic and Archean rocks (Fig 3). The Jacobsville Sandstone is estimated to be more than 2,900 m thick and the top is not exposed (Kalliokoski, 1982). The Jacobsville Sandstone was deposited in an environment characterized by shallow meandering streams (Kalliokoski, 1988). The formation occurs in a rift-flanking basin and at least part of it was deposited during active reverse movement along the Keweenaw Fault.

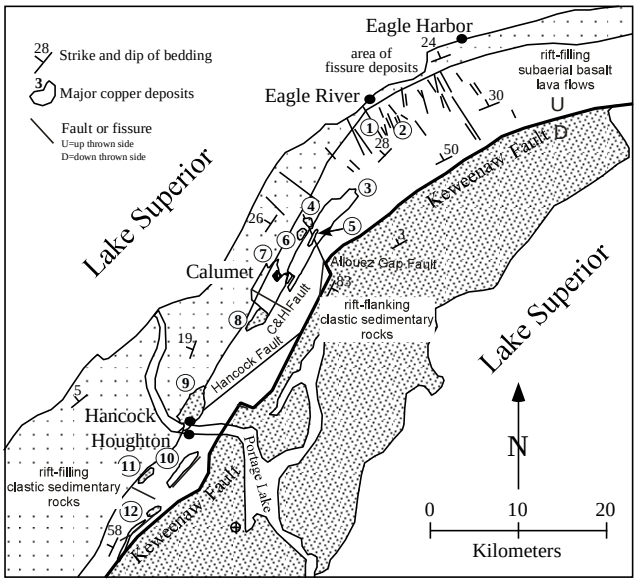
Red to red-brown sandstone is the dominant lithology with lesser amounts of red-brown conglomerate, siltstone, and shale. The sandstone varies from subarkose to quartz sublithic arenite although there are some beds of arkose and quartz arenite (Kalliokoski, 1982). Rounded-to-subrounded, very-fine to coarse sand grains of quartz, feldspar, and lithic fragments occur in massive to cross-bedded, fining-upward sequences. Quartz grains show evidence of volcanic and metamorphic origin. Ripple marked bedding surfaces and cross-bedding are common in some localities. The sandstone varies in color from red to a cream-white or purplish-red color; cream-white color occurs as spherical reduction spots and layers that tend to follow bedding or fractures. Conglomerate is more common in localities near the Keweenaw Fault or near the unconformable contact. Near the Keweenaw Fault, pebble to boulder sized clasts in the conglomerates are composed of felsic and mafic volcanic rocks, similar to Keweenaw Supergroup lithologies. Near the unconformable contact, clast lithologies are of locally derived chemically resistant debris such as quartz and iron formation. There are no interbedded volcanic rocks or cross-cutting igneous dikes within the Jacobsville Formation and while the older age is constrained the upper age is not. The

Jacobsville Formation is approximately 1.06-1.04 Ga to 1.03? Ga (Cannon, 1992).

**Midcontinent Rift Structure**

The last episode of the MCR was characterized by a compression of the continent. This compression transformed original graben-bounding normal faults into reverse faults, reactivated other extensional rift-related faults/fractures, and produced new compression-only faults/fractures and folds. Cannon et al. (1993) have determined that compression occurred at about 1,060+/-20 Ma. The probable cause of this event was continental collision along the Grenville front (Fig. 1) beginning as early as 1.08 Ga and ending by 1.04 Ga (Cannon, 1994; Cannon and Hinze, 1992; Hoffman, 1989).

The rift-filling Keweenawan Supergroup strata dips moderately northwesterly toward the center of the rift (Lake Superior) (Figs. 3 and 7). Their dip angles increase toward the exposed stratigraphic base which is truncated by the Keweenaw Fault. The present day dip of the strata within the MCR is a combination of syn-depositional downwarpage and tilting in response to reverse movement along




① Name of Deposit Given in Table 2

Figure 7: Simplified geologic map showing the location of the major deposits within the Keweenaw Peninsula native copper district, Michigan. Table 2 provides the names and production for the numbered deposits. The areas shown on the map are the mined out down-dip portion projected to the surface. All of the native copper mines are hosted by the Portage Lake Volcanics. Modified from Bornhorst and Barron (2011).

Table 2: Production from 1845 to 1968 of refined copper from native copper deposits (after Weege and Pollock, 1971).

| Name of Deposit | Million lbs Produced Refined Copper | Location Number Shown on Figure 7 |
|---|---|---|
| Calumet & Hecla Conglomerate | 4,229 | 7 |
| Kearsarge Flow Top | 2,263 | 3 |
| Baltic Flow Top | 1,845 | 12 |
| Pewabic Flow Top | 1,077 | 9 |
| Osceola Flow Top | 578 | 8 |
| Isle Royale Flow Top | 341 | 10 |
| Atlantic Ashbed | 143 | 11 |
| Allouez Conglomerate | 73 | 6 |
| Houghton Conglomerate | 38 | 4 |
| Kingston Conglomerate | 20 | 5 |
| Greenland-Mass Subdistrict | 72 | See Figure 3 |
| Other Flow Top and Conglomerate Deposits | 137 | |
| Cliff Fissure | 38 | 1 |
| Central Fissure | 53 | 2 |
| Other Fissure Deposits | 123 | |
| **District Total** | **11,030** | |

the Keweenaw Fault produced by continental compression. Bedding in the rift-flanking Jacobsville Sandstone dips less than $5^{O}$ in most areas, except near the Keweenaw Fault, where dips steepen in response to drag along the fault. Compression-related deposition produced the Jacobsville Sandstone.

There are many faults/fractures in the Keweenaw Peninsula. Some of these were exclusively formed during extension of the MCR when graben-bounding normal faulting was prominent along the margin (Fig. 3). However, most faults/fractures were likely reactivated by or related to the compressional event that inverted the major graben-bounding fault, the Keweenaw Fault, into an overall high-angle reverse fault (Cannon et al., 1989; White, 1968). The Keweenaw Fault strikes and dips more or less parallel to the bedding of the truncated Portage Lake Volcanics (Fig. 7) and is not necessarily one fault, as it is a zone with branches up to 0.8 km from the main fault (Butler and Burbank, 1929). Although the Keweenaw Fault would make an ideal conduit for movement of hydrothermal fluids, there are no native copper deposits along it similar to other ore bearing districts where the main faults are not well mineralized. The rocks within and adjacent are altered especially by paragenetically late hydrothermal fluids. Several reverse faults occur oblique to the strike of bedding. In the Eagle River area, high-angle faults with displacement from 0 to 200 m, fault-controlled native copper veins are common (Butler and Burbank, 1929). The Allouez Gap fault bisects the largest lava flow top hosted native copper deposit in the district and was likely a significant conduit for native copper mineralizing hydrothermal fluids (Bornhorst, 1997). The Allouez Gap fault may have been a reactivated original rift fault. Faults were the principal pathway for the upward movement and focusing of ore fluids into the stratabound lava flow tops in the Baltic and Isle Royale deposits (Broderick, 1931) as well as those in the Greenland-Mass subdistrict (Field Trip 2, this guidebook). Faulting occurred before, during and after deposition of native copper along

with associated alteration minerals based on fault brecciation and re-cementation of alteration minerals. There is a close relationship between faulting/fracturing produced by or reactivated by compression. These compressional structures acted as pathways for native copper mineralizing hydrothermal fluids.

Broad open synclines and anticlines, with wavelengths of around 10 km and various orientations, are superimposed on the regional dip. Faults with displacement and mineralized tension breaks are common near the crests of anticlines (Butler and Burbank, 1929). These post-depositional folds are likely related to the Keweenaw Fault (White, 1968).

**Keweenaw Peninsula Native Copper District**

Active copper mining occurred from 1845 to 1968 in the Keweenaw Peninsula native copper district. The estimated pre-mining geologic resource for the district is 19.7 billion lbs of copper and small quantities of temporally and spatially associated native silver (Bornhorst and Barron, 2011). The major ore producing horizons are located in a 45 km-long belt in the Keweenaw Peninsula (Figs. 5 and 7) and in a subdistrict to the southwest (Field Trip 2, this guidebook). Accompanying native copper and silver, the only economic metallic minerals, were a suite of nonmetallic alteration minerals (Fig. 8). Sulfide minerals are uncommon in the native copper deposits; chalcocite only occurs in trace amounts. Pyrite, an acid-producer when exposed to oxygenated waters, is absent. Several chalcocite deposits of unknown connection to the native copper deposits are hosted by the stratigraphically older Portage Lake Volcanics; the largest of these contains roughly 230 million lbs of copper (Maki and Bornhorst, 1999). These will not be discussed here.

*Native Copper Ore Bodies*

Ore bodies in the Keweenaw Peninsula are tabular, stratabound concentrations of native copper in Portage Lake Volcanics host rocks with sufficient original porosity including brecciated and amygdaloidal flow tops (58.5% of production) and interflow conglomerate beds (39.5% of production). Secondary porosity occurs along fractures/faults host veins (about 2% of production). Since the deposits represent important stratigraphic horizons, the host rocks were given informal member names (Butler and Burbank, 1929). Several mines with different names often worked the same deposit or same lithostratigraphic unit. About 85% of the total district production came from four deposits: Calumet and Hecla Conglomerate, top of the Kearsarge lava flow, top of the Baltic lava flow, and the top of the Pewabic lava flow.

The most common host rocks for native copper deposits are brecciated flow tops (fragmental amygdaloid) as their original porosity was typically much greater than vesicular (amygdaloidal) flow tops (White, 1968). The stratabound flow top deposits are “sandwiched” between a footwall consisting of the same flow as the mineralized flow top and hanging wall of barren massive basalt interior in the succeeding lava flow. Native copper is often more abundant near the top and bottom of the brecciated/fragmental amygdaloid interval of the flow top, however, in rich ore shoots, the entire brecciated/fragmental amygdaloid flow top contains significant amount of copper. As brecciated/fragmental amygdaloidal grades downward into massive basalt, it becomes deficient in native copper. In some cases, ore shoots are located in tongues of brecciated flow tops within massive basalt (Weege and Schillinger, 1962). The lateral and vertical distribution of

brecciated/fragmental amygdaloid within the top of the lava flow is irregular and hence, so is the grade of copper. In general, mined stope heights are from 3 to 5 m. Ore shoots are elongate, but also occur in a wide variety of shapes, with widths of 30 to 150 m and down dip lengths from 50 m to 600 m (White, 1968). The strike length for major ore bodies ranges from 1.5 to 11 km with down dip mineralization extending from 1.5 to 2.6 km below the surface on the inclined deposit. (Butler and Burbank, 1929; White, 1968).

Although interflow conglomerate beds make up only a small volume of the Portage Lake Volcanics, about 40 % of the district productions were hosted by them. These deposits were also tabular and stratabound, just like the flow top deposits. They are "sandwiched" between a footwall consisting of the top of the underlying lava flow and hanging wall of barren massive basalt interior in the underlying lava flow. The porosity of underlying brecciated/fragmental amygdaloid lava flow top is greatly decreased by silt and sand filling the origin open space between fragments. Native copper tends to be concentrated along specific stratigraphic bands that are 0.5 to 5 m thick (Weege et al., 1972). Interflow conglomerates overall host a significant fraction of the districts production. Specifically, the Calumet and Hecla Conglomerate was by far the largest single native copper deposit producing 4.2 billion lbs, as compared to the next largest deposit hosted by the Kearsarge flow top which produced 2.3 billion lbs (Fig. 7 and Table 2). The Calumet and Hecla Conglomerate was mined along a strike length of 4.9 km, and down-dip 2.8 km. The productive area corresponds to a thickening of the conglomerate from less than 1 m up to 6 m (Butler and Burbank, 1929; Weege et al., 1972). Ore grades decrease with depth where the width of the conglomerate is greater where essentially the same amount of copper is distributed throughout a greater volume. (Butler and Burbank, 1929). The highest grades correspond to beds where there is relatively little fine interstitial material or where interstitial spaces are filled with coarse sand or small pebbles (Weege et al., 1972). Thus, localization of native copper ore is dependent on sedimentary environmental factors.

The first mines in the district were developed on tabular steeply dipping deposits that cross cut bedding at high angles. The veins have widths of up to 3 m or more (Butler and Burbank, 1929). Veins are not single tabular bodies, but rather a series of parallel of anastomosing filled open spaces. While brecciation within the deposit is common, gouge is not present (Butler and Burbank, 1929). These adjacent lava flow tops and conglomerates are mineralized. The distribution of native copper in veins is more erratic than in either lava flow top or conglomerate deposits. The richest ore veins tend to be spatially associated with the intersections of the vein and well-oxidized lava flow tops (Butler and Burbank, 1929). Native copper occur as both finely disseminated with associated alteration minerals and as masses weighing many tons. These vein deposits are of slight economic importance in the district. Several small vein deposits are localized just beneath the thickest basalt flow in the district. A good example is the Greenstone flow in which hydrothermal fluids moved up along the cross fractures until blocked by the very thick impermeable massive interior of the Greenstone Flow.

There are veins spatially and genetically associated with the stratabound lava flow top or conglomerate deposits; these veins occur along faults that intersect major deposits such as the Baltic and Isle Royale (Broderick, 1931). This suggests that ore fluids moved upward along faults and outward into the permeable flow tops. The intersection of subsidiary faults with locally thick permeable horizons is a key factor in concentrating ore such as the Kearsage deposit (see Stop 4 and Fig. 12). White (1968) suggested that for the movement of ore fluids to occur, permeability due to fracturing was more important than primary permeability. Faults and small fractures cutting massive interior of lava flows were likely important for upward transport of ore fluids also. Overlapping of successive lava flows and minor unconformities suggests that simple up-dip movement of ore fluids was not likely without a network of fractures.

*Secondary Hydrothermal Minerals*

The rocks within the Keweenaw Peninsula native copper district were pervasively altered by low-temperature, low-pressure hydrothermal/burial metamorphic fluids. Alteration was most intensely associated with the native copper deposits, although to some degree, secondary hydrothermal minerals occur in all rocks of the Portage Lake Volcanics. Areas in the Keweenaw Peninsula more distal to the native copper deposits were less altered. The intensity and degree of alteration also varies as a function of position within lava flows; the massive interiors of lava flows being much less altered whereas the lava flow tops are relatively more altered. Lava flows in close proximity to cross cutting features tend to be more altered. The minerals occur as amygdule and vein fillings, and as whole rock replacements. Within the Portage Lake Volcanics, some original igneous minerals are present in the massive interiors of flows, but secondary minerals exist in the massive interiors of all flows regardless of their thickness. While the massive interiors of lava flows contain secondary minerals, their original igneous geochemical composition is often only slightly modified by secondary hydrothermal processes.

There are over 100 different secondary alteration minerals in the Keweenaw Peninsula; most of them are related to hydrothermal process and some are related to supergene processes. Only about 24 alteration minerals are common. Native copper with small quantities of native silver represents over 99% of the metallic minerals in the mined ore bodies of the district. Most of the native copper carries a small amount of arsenic in solid solution (typically less than 0.2 % arsenic in total copper + silver + arsenic; Broderick, 1929). Copper-nickel arsenides occur in veins that are paragenetically late (Moore, 1971; Stoiber and Davidson, 1959; Butler and Burbank, 1929). Within the native copper deposits, paragenetically late chalcocite occurs as small veins cutting lava flow top deposits, and as coatings on joints containing calcite in conglomerate deposits (White, 1968).

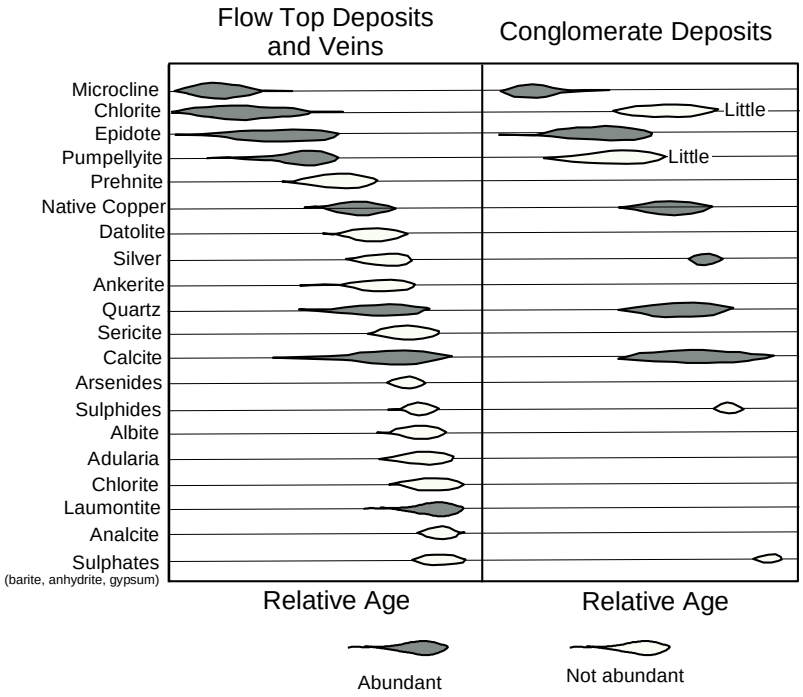



Figure 8: Paragenesis and relative abundance of secondary hydrothermal alteration minerals in the Keweenaw Peninsula native copper district. Modified from Butler and Burbank (1929).

District-wide there is a well-defined mineral paragenesis (Fig. 8), although individual deposits may not exactly follow the district-wide timing of precipitation. There is a general spatial variation of hydrothermal minerals in the Calumet area of the district (Fig. 9). Epidote and the appearance of quartz are spatially associated with major native copper deposits (Stoiber and Davidson, 1959). A detailed study by Stoiber and Davidson (1959) of the Kearsarge deposit shows that native copper is much more irregularly distributed than secondary mineral zones, but there is a general correlation with the abundance of native copper associated with the variation of quartz and microcline (see Stop 4 and Fig. 14). On the tip of the Keweenaw Peninsula, the suite of hydrothermal alteration minerals consists of low temperature zeolite minerals except within about 750 m of the Keweenaw Fault where there is epidote (Cornwall, 1955; Cornwall and White, 1955). Amygdule-filling minerals are equivalent to zeolite and prehnite-pumpellyite metamorphic facies. The hydrothermal /metamorphic mineral zones dip more gently towards Lake Superior than the strata, implying that the strata were tilted prior to hydrothermal alteration (Livnat, 1983; Broderick, 1929). The paragenetic succession of alteration minerals at the Kearsarge deposit begins with low-temperature (80 to 100$^{o}$C) agates followed by higher temperature native copper and temporally associated minerals (about 225$^{o}$C) and the final stage is superimposition of lower temperature late-stage barren laumontite and calcite. This progression represents a waxing and waning of the hydrothermal system. Much later, the native copper has been altered by oxidized groundwaters generating copper oxide minerals.

Native copper mineralization is younger than the Copper Harbor Conglomerate, which hosts rare veins of calcite and native copper (see Stop 6). White (1968) interpreted the age of native copper mineralization as after the deposition of parts or all of the Freda Sandstone. Minor amounts of

native copper occur within the lower beds of the Jacobsville Sandstone near Rice Lake. Based on field relations, hydrothermal alteration is younger than deposition of rift-filling strata and at least some of the rift-flanking Jacobsville Sandstone. The absolute age of hydrothermal alteration is between 1060 and 1047 Ma (+/- ~ 20 Ma) (Bornhorst et al., 1988). This age is consistent with the approximate age of 1060 Ma for regional continental compression that caused reverse faulting along the Keweenaw Fault (Cannon et al., 1993). Thus, the age of hydrothermal alteration is about 1070 to 1040 Ma contemporaneous with regional continental compression and some 10 to 30 million years after eruption of the Portage Lake Volcanics.

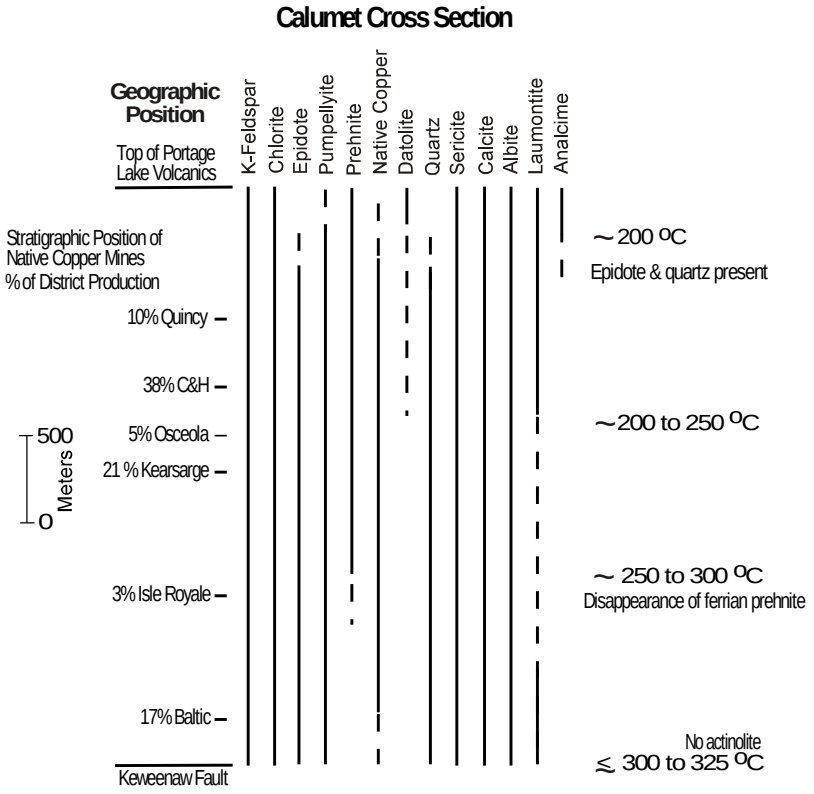



Figure 9: Distribution of prominent secondary hydrothermal alteration minerals in the Portage Lake Volcanics in a cross-section in vicinity of Calumet at the center of the major deposits of the Keweenaw Peninsula native copper district. Data compiled from Livant (1983) and Stoiber and Davidson (1959) and modified from Bornhorst and Rose (1994).

*Genesis of the Native Copper Deposits*

Native copper occurs throughout the MCR in Wisconsin, Minnesota, and Ontario (Fig. 2) which suggests a regional distribution of mineralizing hydrothermal waters. The regional Cu-bearing hydrothermal fluids can be best explained by their generation during burial metamorphism of rift-filling basalts with temperatures reaching a thermal maximum 10 to 30 million years after the end of widespread rift magmatism (Fig. 3). The coincidence of regional continental compression with the thermal maximum provided an integrated paleohydrologic system through reactivated and new faults and fractures. This allowed the upward movement of hydrothermal fluids to

focus in sites of future copper deposits at the very time of greatest fluid availability (Bornhorst 1997). During generation of the regional hydrothermal ore fluids, a few ppm of copper was leached from the rift-filling basalt strata (Jolly, 1974; White, 1968; Stoiber and Davidson, 1959). Simple calculations demonstrate that only a 3 km down dip volume of the 10 km thick rift-filling basalts along the 45 km strike length of the major copper deposits need be leached to generate sufficient copper for deposition, or in other words, the rift-filling basalts are a viable source rock for the copper. The hydrothermal fluids were low in sulfur since most of the sulfur in the buried rift-filling source rocks was degassed during eruption so subsequent deposition up dip in the same sulfur-poor rocks preferentially lead to deposition of native copper rather than copper sulfides. Precipitation of native copper was caused by mixing of ore fluids with cooler, more dilute shallower fluids, ore fluid-rock reactions, and cooling of ore fluids. The localization of large native copper deposits within the Keweenaw Peninsula rather than elsewhere in the MCR may be controlled by favorable geometric orientation within the regional continental compression stress field. Since much of the MCR strata is buried, perhaps another area of native copper deposits remains hidden.

**Phanerozoic**

The last events in the geologic development of the MCR in the Keweenaw Peninsula were the formation of the native copper deposits and deposition of the Jacobsville Sandstone during regional continental compression at 1.06 to 1.04 Ga; the Jacobsville deposition may have continued after compression until 1.03? Ga (Fig. 3). The Keweenaw Peninsula was subsequently subjected to a 500 million year period of erosion, from about 1.03 Ga to 0.5 Ga, 500 Ma and multiple kilometers of rock were eroded exposing the native copper deposits at the surface (Fig. 3) (Bornhorst and Robinson, 2004). Downward percolating groundwaters supergene altered native copper and produced a suite of including cuprite, tenorite, malachite, and chrysocolla. The rocks of the Keweenaw Peninsula were subsequently buried by Paleozoic sedimentary rocks associated with the Michigan basin beginning about 500 Ma (Fig. 3).

Over the past two million years, the Keweenaw Peninsula was subjected to several continental glacial periods which removed all of the overlying Paleozoic sedimentary rocks with the exception of a Paleozoic outlier slightly south (Fig. 3). After the last glacial episode, the native copper deposits were exposed at roughly the same erosional level as at 500 Ma or the end of the Precambrian. The continental glaciers sculpted the bedrock of the Keweenaw Peninsula and when the last glacier retreated about 10,000 years ago, it left behind a variety of unconsolidated glacial-related sediments that included entrained boulders of native copper. The glaciers carved out the topographic low the Lake Superior basin corresponding to the less competent clastic sedimentary rocks under the center of the MCR. After the glaciers retreated, very large volumes of water filled this topographic low and initially all but the highest land elevations were underwater of a large glacial lake. The glacial lake levels successively dropped over time to the current level of Lake Superior (Farrand 1960). As the lake levels receded humans populated the area.

**Objectives of Field Trip**

This field trip is designed to provide a geologic overview of the Keweenaw Peninsula and the Keweenaw Peninsula native copper district. The Mesoproterozoic MCR bedrock and hosted native copper deposits are unconformably overlain by unconsolidated Pleistocene glacial sediments.

The descriptions and stops of this field trip provide glimpses into both of these distinct geologic events. The MCR rocks and native copper deposits are the focus of this field guide. Figure 10 provides the regional geologic setting of the Stops.

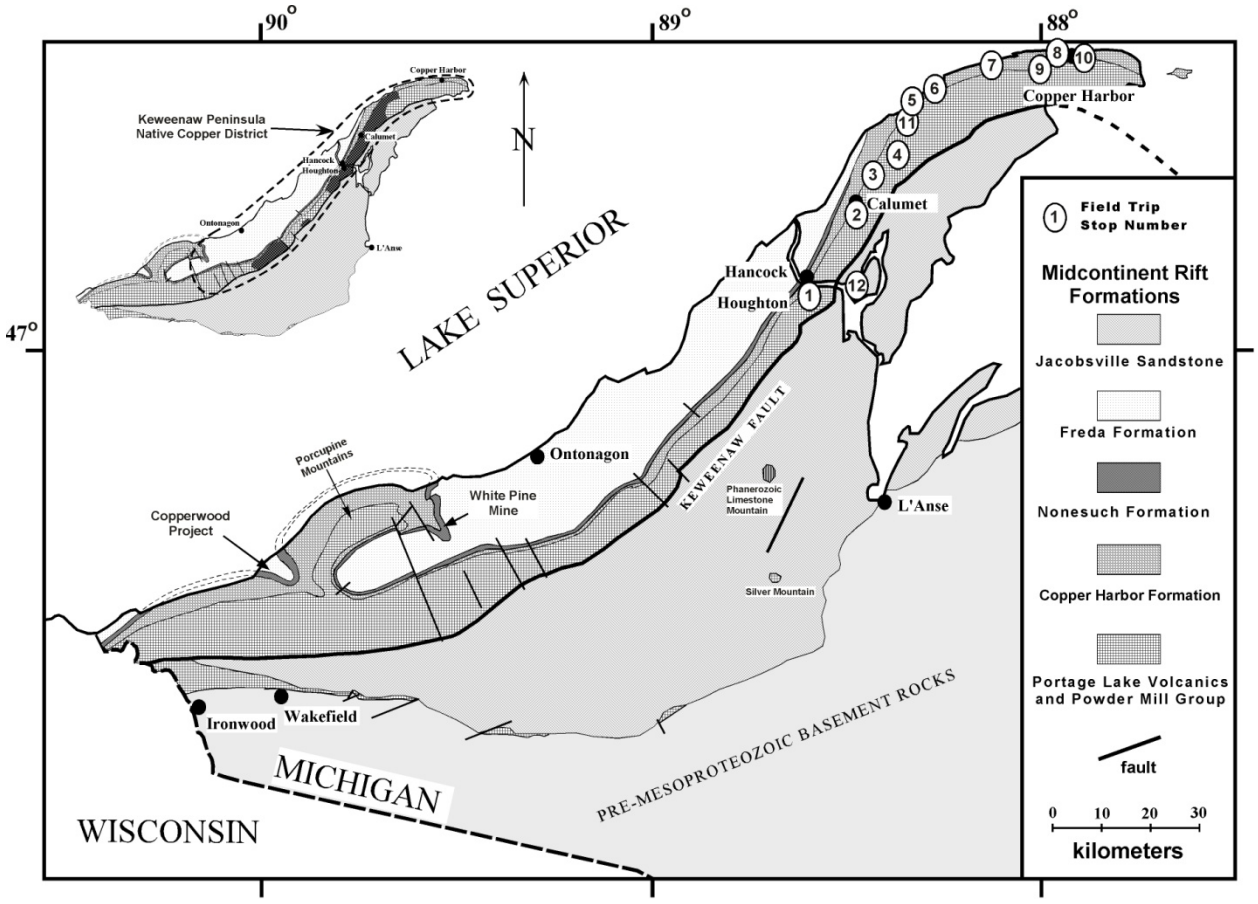



Figure 10: Geologic map of the far western part of the Upper Peninsula of Michigan showing field trip stops.

## Stop 1: Razorback Center

*<u>Directions</u>: Drive west through downtown Houghton on US-41to south M26. Drive 0.9 miles (1.4km) to Sharon Ave. at first stoplight. Turn left on Sharon Ave. for 0.1 miles(0.2km) to Razorback Dr. Turn right on Razorback Dr. for 0.1 miles (0.2km) to strip mall on left built on top of a small hill. Turn left and outcrop exposure is behind the strip mall. [UTM 5218745N 160379747E (NAD27 CONUS)]*




Figure 11: Photograph of the rock cut at Razorback Center looking northwest.

The rock cut at the edge of the parking lot in the rear (south side) of Razorback Center, Houghton provides an excellent example of the characteristics of subaerial basalt lava flows that comprise the Portage Lake Volcanics, the host rock unit for native copper deposits of the Keweenaw Peninsula native copper district and the rock unit that holds up the spine of the Keweenaw Peninsula (Fig. 10). This of exposure of subaerial lava flows is located stratigraphically between the Calumet and Hecla and Kingston Conglomerates (Fig. 11). It is also located between the stratigraphic level of the Isle Royale and the Quincy Mines. The lava flows strike about N30$^{o}$E and dip about 55$^{o}$ to the northwest (towards Lake Superior). The rock cut is at an oblique angle to the strike of the lava flows. When facing the exposure, the stratigraphic top is towards the northwest or toward the right. At the far eastern end, or far left, only the amygdaloidal top of the oldest basalt lava flow in this

rock cut is exposed. Stratigraphically upwards, towards the west/right, this flow top is overlain by a thick section of dark-gray to black massive basalt representing the interior of a prominent lava flow. Progressively, the abundance of amygdules increases upwards and the color of the basalt changes to greenish tones reflecting an increased degree of alteration. This zone represents the amygdaloidal top of the prominent lava flow in this rock cut. The amygdules tend to be concentrated along layers and near the upper contact; they coalesce into a continuous now filled open space. The contact between the amygdaloidal top of this prominent lava flow and the massive basalt of the overlying flow is well exposed (Fig. 11). Along most of the exposed contact, amygdaloidal basalt lies directly below the massive basalt indicating the lava flow had a smooth top (pahoehoe lava flow), however, at the level of the parking lot, the planar contact bends and there is a small zone of brecciated flow top. The entire cross section of the prominent lava flow is exposed in the Razorback Center rock cut. A typical lava flow in the Portage Lake Volcanics is between 10 to 20 m thick, the prominent lava flow at Razorback Center. Stratigraphically further upwards, towards the west, the prominent lava flow is overlain by a thick section of dark-gray to black massive basalt representing the interior of the overlying flow (Fig. 11). On the far western end, there is amygdaloidal basalt representing the top of this overlying flow; an almost complete cross section of this flow is exposed here.

Volcanic textures and structures at Razorback Center are typical of subaerial lava flows within the Portage Lake Volcanics. The basalts are mainly olivine tholeiites erupted as thick, ponded subaerial lava sheets. The very top and bottom of such lava flows typically consist of aphanitic chilled basalt. The contact between the underlying and overlying lava flows occurs where amygdules disappear abruptly and the overlying flow consists of massive basalt. The upper surface of the main flow was brecciated slightly by movement of lava after the formation of an upper crust, but rapidly grades downward to a non brecciated, highly vesicular flow top. The layered nature of amygdules in the prominent flow here at Razorback Center is likely a result of preferential accumulation of vesicles along laminar flow planes. The flow top breccia is laterally discontinuous for this flow. Slow cooling of the lava flow caused solidification toward the flow interior at a rate which allowed development of subophitic to ophitic textures (large oikocrysts of clinopyroxene enclosing a felted framework of An-rich plagioclase and intergranular olivine). The resulting massive, non-vesicular flow interior constitutes about two-thirds of the flow.

The effects of regional hydrothermal alteration can be observed within the amygdaloidal flow tops. The massive interiors are much less altered except along fractures. The original plagioclase in the massive basalt has been replaced by albite and the mafic minerals by chlorite, pumpellyite, and iron oxides. The massive interior of the flow is much less altered than the flow top which represents a relatively impermeable horizon in the paleohydrologic system except in the vicinity of selected fractures. The pseudomorphic alteration minerals in the massive interior of the basalt are similar to those which fill the amygdules. The amygdules here are filled with a variety of secondary minerals including: calcite, chlorite, epidote, prehnite, pumpellyite, quartz (not in order of abundance), and traces of native copper. Late stage laumontite abundantly fills some amygdules.

## Stop 2: Float Copper US-41 Calumet

*<u>Directions</u>: Get back onto M26 at the traffic light and turn right. Stay on north M26/US41 and cross the bridge over to Hancock. Drive through downtown Hancock and continue 10 miles (16km) north to Calumet. Float copper is located on the left side of US41/M26 past the first traffic light just before Red Jacket Rd. [UTM 5233057N 160390423E (NAD27 CONUS)]*

A float copper boulder weighting 4,263 kg (9,392 lbs) is on display at Stop 2 (Fig. 10). This mass of glacially transported native copper was found in 1970 about 4.5 miles SW of Calumet in less than three feet of surficial sediments. Native copper deposits of the Keweenaw Peninsula were exposed at the bedrock surface at the time of the last period of Pleistocene glaciations. The glacial ice plucked masses of malleable native copper from the tabular lodes and fissures which were subsequently smoothed and flattened by abrasion from other rocks carried by the glacial ice. When the glaciers retreated about 10,000 years ago, unconsolidated rock debris (rounded boulders to clay sized material) were left behind by the melting ice including masses of native copper such as this one "floating" among the unconsolidated rock debris. While some of the rocks in the glacial deposits are from far north of the Keweenaw Peninsula, most of them are recognizable as from local MCR strata exposed in the Keweenaw Peninsula. The large float copper masses could not have moved far from their source, but smaller masses have been transported quite far and have been found in Lower Michigan and Wisconsin. The largest known float copper was discovered in the early 2000s and weighed about 25 tons (50,000 lbs) near the Houghton County airport; it was cut into smaller masses and sold to be smelted and refined. Most pieces of float copper are small, ranging from a few to 50 cm across. The famous example of float copper was the Ontonagon boulder, a 3,700 pound specimen visited by numerous explorers and finally removed from the Keweenaw to the nation's capital in 1843. The Ontonagon boulder is part of the Smithsonian's collection.

This and other float copper masses have been surface altered by oxygenated groundwater and shallow precipitation since the glaciers retreated. This surface alteration consists of forms of copper including cuprite (copper oxide; $Cu_2O$), tenorite (copper oxide; CuO), malachite (hydrated copper carbonate; $(Cu_2(CO_3)(OH)_2)$ and rarely azurite (hydrated copper carbonate, $(Cu_3(CO_3)_2(OH)_2)$. Even when small cm sized masses of float copper are cut, the typical surface alteration is less than one mm thick; native copper is highly resistant to surface weathering. Float copper makes an attractive decorator specimen when a part of the surface is polished and buffed exposing shiny copper color.

The basalt mine rock buildings are part of the Keweenaw National Historical Park. The park was established on October 27, 1992, by U. S. Congress Public Law 102-543. The enabling legislation ascertained that the Keweenaw was nationally significant because of: its unique geology; the prehistoric use of its copper by Native Americans; the importance of the region as a leading copper producer and developer of new technologies; its long history of corporate paternalism; and because it became home to so many European ethnic groups that migrated to the United States. Older mining districts typically had only single-industry economies and when the mines shut down, the communities suffered major contraction. In 1910, nearly 40,000 people resided within a few miles of Stop 2 whereas now, fewer people live in all of Houghton County.

The idea that maybe the future of Calumet resided in its past was generated in the late 1980s; history could be "sold" to revitalize the community by increasing tourism. The national park itself only owns a few structures in Calumet, including these, and instead relies on public and private partners termed Keweenaw Heritage Sites. The heritage sites contain and interpret significant cultural and/or natural resources that together with park assets help tell the story of copper mining in the Keweenaw Peninsula. The Quincy Mine property on the edge of Hancock and the A.E. Seaman Mineral Museum on the campus of Michigan Tech are two among multiple Keweenaw Heritage Sites.

## Stop 3: Bumbletown Hill

**THE ROCK PILES DESCRIBED FOR THIS STOP ARE ON PRIVATE PROPERTY AND PERMISSION IS REQUIRED ACCESS THEM.**

*Directions: Continue on US-41 3.6 miles (5.8 km) past headquarters of the Keweenaw National Historical Park denoted by park sign and large specimen of glacial float native copper to Bumbletown Rd. and turn left (west). Drive about 0.4 miles (0.6 km) to rock pile. [UTM 5237960N 160393345E (NAD27 CONUS)]. To access the overlook leave the rock pile and continue on Bumbletown Road west about 0.5 miles (0.8 km) to overlook at the top of the hill near towers. [UTM 5238215N 160392860E (NAD27 CONUS)]*

The description of this stop is reproduced from Bornhorst and Barron (2011).

The Allouez conglomerate (informal member) is one of a small number of interflow clastic sedimentary horizons within the Portage Lake Volcanics visible in the rock pile at this stop (Fig. 10). This particular conglomerate bed can be traced along strike from the tip of the Keweenaw Peninsula, to at least the Mass area, a strike length of more than 120 km (Fig. 6). The Allouez conglomerate is stratigraphically just below the Greenstone flow, arguably the largest basalt flow in the world, within the Portage Lake Volcanics. The rock piles at the base of Bumbletown Hill are from the Allouez Mine. The Allouez conglomerate consists of mostly red-colored conglomerate with lesser amounts of sandstone and siltstone. The largest contained boulders at this locality are about 65 cm in diameter and the median size is about 8 cm. A pebble count of boulders more than 20 cm across gave the following results: mafic rock, mostly amygdaloidal basalt, 16%; quartz porphyritic rhyolite, 36%; feldspar porphyritic rhyolite, 11%; and granophyre, 37% (White, 1971).

The mines on the Allouez conglomerate yielded only about 75 million pounds of refined copper (Table 2). Some evidence of native copper mineralization can be seen in rocks at this stop. Occasionally, one can find a specimen with native copper filling the void space between clasts and grains. Calcite and chlorite are the dominant pore-filling secondary minerals visible on this rock pile. Thin black veinlets cutting the Allouez conglomerate consist of calcite with chalcocite "dust." While supergene alteration resulting from the downward percolation of groundwater is not common in most the native copper deposits, at this stop, supergene alteration minerals are common including chrysocolla, malachite, and cuprite.

From the overlook on a clear day, Isle Royale may be seen 80 km to the northwest and the Huron

Mountains may be seen beyond Keweenaw Bay, 60 km to the southeast. The land slopes very gradually to the northwest toward Lake Superior, as it does throughout most of the length of the Keweenaw Peninsula. The area is underlain mainly by conglomerates and sandstones of the Copper Harbor Formation dipping at about $20^{o}$. The southeastern flank of the Keweenaw Peninsula has a steeper slope at the skyline, following approximately the line of the Keweenaw fault. The low-lying plain between the fault and Keweenaw Bay is underlain by flat-lying Jacobsville Sandstone.

Bumbletown Hill is located on the southwest side of the Allouez Gap, a NW- to SE-trending valley. The valley follows the Allouez Gap fault, a zone of faults and fractures, along which the Portage Lake Volcanics and Keweenaw fault, are offset. At this gap, the strike of the Portage Lake Volcanics swings from about $N35^{o}E$ to $N50^{o}E$ (Figs. 5 and 7). Almost every permeable horizon near the Allouez Gap fault contains above average amounts of native copper; nowhere else in the district are there so many mineralized beds (Fig. 7). About 60% of the district production can be linked to the fault as a primary pathway for ore fluids. The fault bisects the Kearsarge deposit (see Fig. 12), which was the second largest copper producer in the native copper district. The line of rock piles demarking the many mines along the Kearsarge deposit is a little more than 1,500 m southeast of Bumbletown Hill. The Kingston Mine, a small deposit that produced 20 million pounds of copper (1963 to 1968; one of the most recent native copper mines to open and last to close), is bisected by the Allouez Gap fault. About 1,200 m $N65^{o}E$ of the hilltop, the Houghton conglomerate and the Iroquois flow produced 33 million pounds of copper.

Looking northeast along the strike of the Portage Lake Volcanics, one can see the cuesta form of the ridge upheld by the Greenstone flow. To the right of the ridge, the more distant hills are formed by lava flows lower in the Portage Lake Volcanics sequence. At Bumbletown Hill, the Greenstone flow is only 85 m thick, but the flow thickens abruptly to more than 400 m near end of the cuesta ridge. It dips northward at about $25^{o}$ toward the center of the Lake Superior. The Greenstone flow can be traced along much of the Keweenaw Peninsula and has been stratigraphically and geochemically correlated with a similar unit on Isle Royale, 90 km away, on the opposite side of the rift. Thus, the areal extent of this great flow exceeds 5,000 $km^2$, and its volume is on the order of 800 to1,500 $km^3$ (Longo, 1983). The geochemical composition of the Greenstone flow magma is more evolved than typical olivine tholeiites of the Portage Lake Volcanics.

## Stop 4: Seneca Mine Rock Pile

**THE ROCK PILES DESCRIBED FOR THIS STOP ARE ON PRIVATE PROPERTY AND PERMISSION IS REQUIRED ACCESS THEM.**

*Directions: Drive back to US41/M26 on Bumbletown Rd. and turn left. Continue northeast on US-41/M26 0.6 miles(0.9km) to B St. Turn Left on B St. Drive about 0.3 miles(0.5km) to rock pile. [UTM 5238775N 160394119E (NAD27 CONUS)]*

The Kearsarge lode was worked by the Seneca Mine, one of multiple mines which produced native copper from the top of the Kearsarge basalt lava flow over a strike length of more than 12 km and down-dip as much as 2,500 m (Figs. 10 and 12). About 1,026 million kg of refined copper were produced at an average grade of 1.05% Cu, making the Kearsarge deposit the largest flow top hosted deposit and the second largest producer in the district behind the C&H Conglomerate mines (Table 2). Production of copper from the Kearsarge lode began in 1887 and stopped in 1967.

The Kearsarge lava flow has been recognized for a distance of about 55 km along strike and dips between 35 and 40$^{o}$ NW (Fig. 12). It lies directly above the Wolverine Sandstone (Fig. 6). The amygdaloidal and/or brecciated top of the Kearsarge flow ranges from near zero up to 10 m in thickness. The productive top has an average thickness of around 2 m and consists of brecciated basalt (individual fragments of amygdaloidal basalt are generally less than 15 cm in greatest dimension). The brecciated basalt grades downward into amygdaloidal basalt with amygdules concentrated in layers. Further downward, the top grades into a zone of fewer and larger amygdules, and then into aphyric massive basalt in the interior of the flow. Just below the brecciated and/or amygdaloidal top of the flow, there is distinct plagioclase porphyritic basalt. The abundance and size of the plagioclase phenocrysts in this zone is variable, but they can make up a large percentage of the rock, with phenocrysts up to 2.5 cm in length. This zone is probably the result of plagioclase *in situ* floating during surface crystallization of the flow. Specimens with abundant plagioclase phenocrysts can be found on this rock pile.

The basalt itself in the Kearsarge flow is well oxidized. Albitized and pumpellyitized basalt consists of pseudomorphically replaced plagioclase set in a fine-grained to cryptocrystalline groundmass. Original igneous minerals were replaced in areas where alteration was intense. Olivine is almost invariably completely replaced while other igneous mineral are replaced by alteration minerals to varying degree.

The amygdule and interfragmental space-filling gangue minerals in the Kearsarge lode are generally (in order of most to least abundant): calcite, epidote, K-feldspar, quartz, and lesser amounts of chlorite, prehnite, pumpellyite, laumontite, and sericite. Native copper is closely associated in time and space with the secondary amygdule minerals (Stoiber and Davidson, 1959). Paragenetically, chlorite; epidote; microcline; and prehnite are early-formed minerals, and the latest-formed minerals are quartz; native copper; calcite; and chlorite (Fig. 14). A zonal stratabound arrangement of amygdule minerals in the Kearsarge deposit is seen in the Ahmeek Shaft No. 3 (Fig. 15). The zoning may be explained by deposition of secondary minerals from a hydrothermal solution moving along a permeable channel.

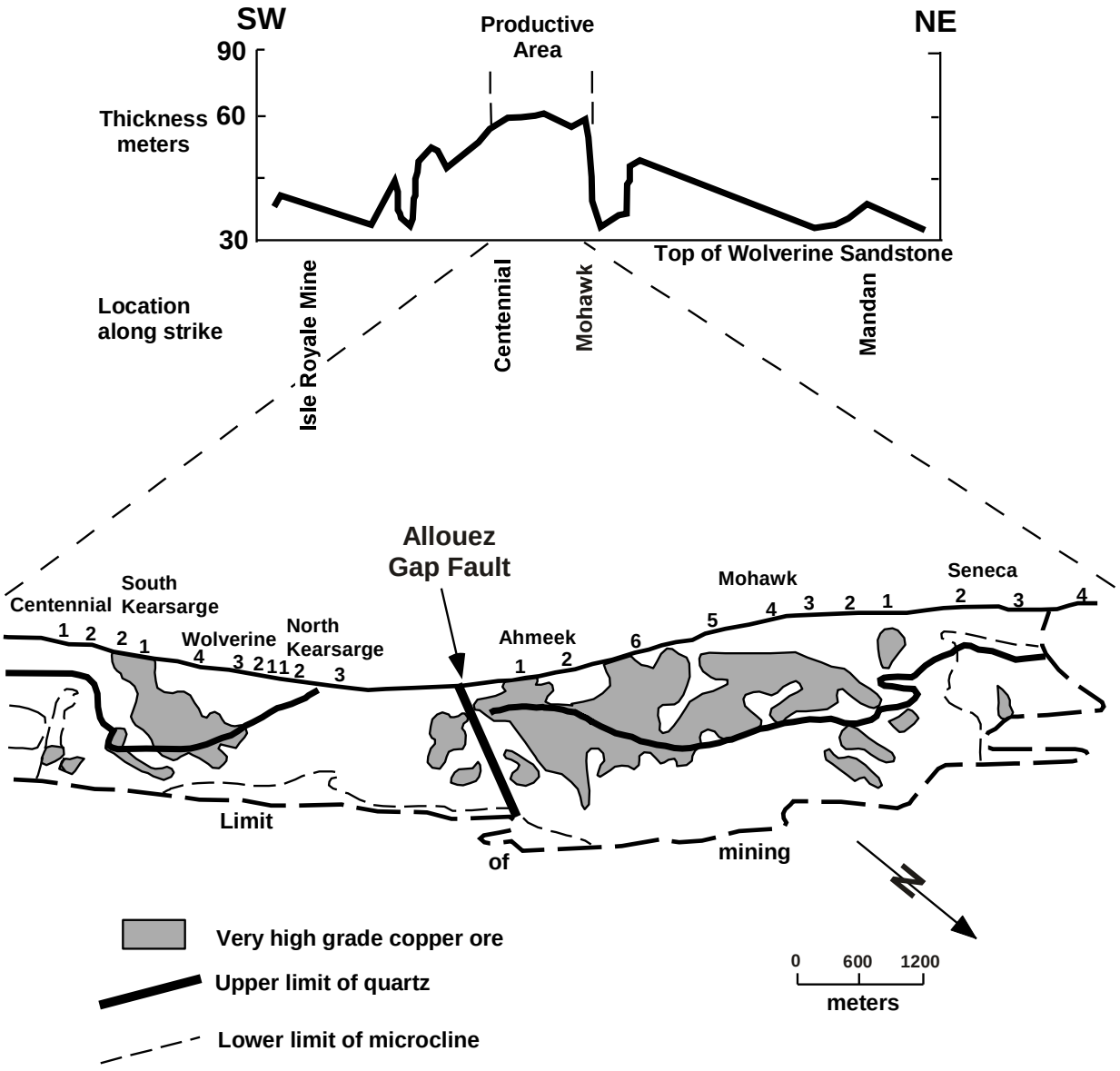



Figure 12: Thickness of the Kearsarge lava flow from showing the location of the productive area where the top of the flow is thickest. Modified from Butler and Burbank (1929). The most productive area corresponds to the thickest part of the flow which is bisected by the Allouez Gap fault. Bottom diagram is a down-dip strike parallel section project to vertical showing distribution of higher grade native copper ore and occurrence of important alteration minerals. Modified from Stoiber and Davidson (1959). Abundance of quartz in amygdules is greater than 10 % on the down-dip side (lower) of the line shown and K-feldspar is absent on the down-dip side (lower) of the line shown. The Kearsage flow dips about 35 to 40$^{o}$ NW and all data are projected.

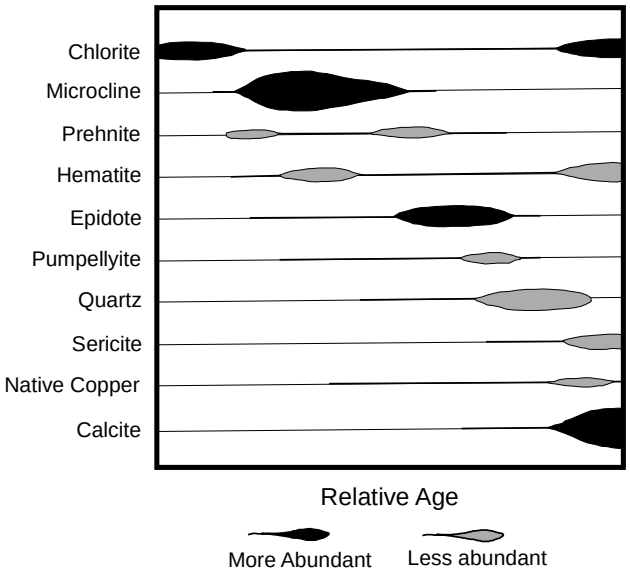
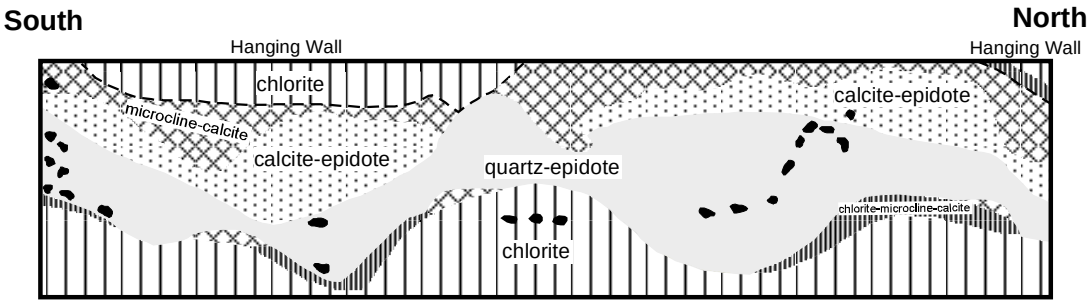



Figure 13: Paragenesis of secondary hydrothermal alteration minerals in the Kearsarge deposit at the Wolverine No. 2 Mine.




**Mineral Assemblage Band**

| Volume Percent Amygdule Filling | chlorite | chlorite-microcline-calcite | microcline-calcite | quartz-epidote | calcite-epidote |
|---|---|---|---|---|---|
| Chlorite | 100 | 69-74 | 0-3 | 0 | 0 |
| Microcline | 0 | 15-25 | 45-82 | 0 | 0 |
| Epidote | 0 | 0-1 | 5-10 | 90-96 | 12 |
| Calcite | trace | 0-5 | 0-47 | 0-1 | 87 |
| Quartz | 0 | 0-5 | 0-8 | 4-9 | 1 |
| Pumpellyite | 0 | 0-6 | 0-trace | 0 | trace |

Figure 14: Cross section of the top of the Kearsarge lava flow (amygdaloid) deposit showing the distribution of secondary hydrothermal amygdule-filling alteration minerals at the Ahmeek Mine, 35th level, 400 to 500 ft south of the shaft. Modified from Stoiber and Davidson (1959). Data from the back and walls are projected to a horizontal plane. There is a barren laumontite-quartz-calcite zone not shown here.

Chlorite and microcline would have been deposited first, along the outer limits of the solution channel; followed by quartz and epidote in the center of the channel; and finally, deposition of calcite in the remaining openings. This observation is consistent with the paragenetic relationships seen in individual samples. No strict correlation exists between the stratabound zoning and the grade of native-copper mineralization (Stoiber and Davidson, 1959). The amygdule minerals and grade of copper mineralization vary with depth. Within the upper limit of quartz (Fig. 12), the quartz content is typically about 15 % of open space fillings although it is considerably less than 10% at shallower depths. The lower limit of microcline may also mark the limit of significant copper mineralization. The amount of native copper present is much more irregular than variation of the mineralized zones.

The Allouez Gap Fault bisects the thickest segment of the Kearsarge Flow along its 55 km strike length (Fig. 12). Higher grades and production occur northeast of the fault where fractures with orientations that parallel the fault are more abundant. Within the Allouez Gap Fault zone, early epidote and quartz were brecciated and recemented by calcite, quartz, and native copper. After another episode of brecciation, the fault zone was recemented again with calcite; quartz; and lesser laumontite (Butler and Burbank, 1929). Movement along the fault occurred before, during, and after deposition of native copper. The fault apparently was a conduit for transport of ore fluids to the permeable flow top. The coincidence of this fault with the relatively thick flow top resulted in the second largest deposit in the district.

The Seneca Mine is an excellent locality to study the character of a representative basaltic flow top hosted native copper deposit. Specimens of massive basalt, massive basalt abundant plagioclase phenocrysts,and amygdaloidal basalt can be found on this rock pile. Masses of native copper are readily collectable especially when using a metal detector. Open-space filling minerals (amydgules and between breccia fragments) that occur in the lode can be found on the rock pile. Stoiber and Davidson (unpublished data) made a quantitative analysis of open-space filling minerals for the Seneca Mine rock pile and found open-space filling minerals consisted of: calcite, 57%; red feldspar 8%; pink feldspar 15%; epidote, 17%; prehnite, trace; pumpellyite, trace and quartz, trace. Many specimens contain multiple minerals and illustrate paragenetic relationships.

## Stop 5: Eagle River Falls

*<u>Directions</u>: Continue northeast on US-41/M26 9.5 miles (15km) to Phoenix and turn left on M26. Continue 2.3 miles (3.7km) to Eagle River and park by the bridge. [UTM 5251824N 160402193E (NAD27 CONUS)]*

The water falls of Eagle River is near the contact between the top of the Portage Lake Volcanics and the base of the Copper Harbor Formation (Fig. 10). The contact dips about $30^{o}$ NNW. The beds strike roughly parallel to the shoreline of Lake Superior; the orientation of the Keweenaw Peninsula changes from NE in vicinity of Houghton to ENE at Eagle River to E-W near the tip. The tholeiitic basalt subaerial lava flows just below the contact are pahoehoe with ropy upper surface. The orientation of the ropes indicates that the flow was erupted from a vent to the north geographically under Lake Superior. That the ropy flow top is preserved suggests that little erosion occurred between deposition of the last of the lava flows of the Portage Lake Volcanics and the Copper

Harbor Formation. The Copper Harbor Formation consists of red-brown rhyolite-pebble conglomerate, but includes many sandstone and even some shale beds. Under the bridge, one can get a good view of the lithology of the lower part of the Copper Harbor Formation. The Copper Harbor Formation was deposited in an alluvial fan shed off of a highland area to the SE (opposite of Lake Superior) likely buried under the rift-flanking Jacobsville Sandstone (Elmore, 1984).

This contact marks an abrupt change in the geologic evolution of the Midcontinent rift. Below this contact there is a thick succession of basalt subaerial lava flows with more than 200 individual flows and a cumulative thickness of about 5,000 m, thus magmatic activity dominated the Midcontinent rift at that time. Abruptly above the contact lava flows are strikingly absent and clastic sedimentation dominated the Midcontinent rift. While generally absent, a last gasp of magmatic activity will be seen at Stop 10 where a thin package of mafic to intermediate volcanic rocks, the Lake Shore Traps, are interfingered within the Copper Harbor Formation.

## Stop 6: Great Sand Bay

*<u>Directions</u>: Continue driving northeast on M26 through Eagle River for 4.8 miles(7.7km) until the Great Sand Bay overlook. [UTM 5253974N 160406985E (NAD27 CONUS)]*

The description of this stop is reproduced with minor modifications from Bornhorst and Barron (2011).

The Great Sand Bay overlook provides a beautiful view of Lake Superior (Fig. 10). Very large volumes of water filled the Lake Superior basin as a result of melting of the glaciers, turning it into a glacial lake. The levels of the glacial lakes depended on the position of the ice front, outlets, and crustal rebound (a result of removing the weight of the ice). There are 15 lake stages recognized in the Lake Superior basin (Farrand 1960). As the lake levels receded to the current level of Lake Superior, more and more of the Keweenaw Peninsula emerged. At the road level, the sand dunes are remains of the Lake Nipissing Stage (4,000 to 5,000 years ago) when the lake level was about 9 m (30 feet) higher than today. After lake stages at about 3,200, 2,000, and 1,000 years ago, the waters receded toward the present level termed Lake Superior.

The underlying bedrock is the Copper Harbor Formation. In the Keweenaw Peninsula there is a succession of basalt lava flows interbedded near the middle of the formation (see Stop 8). The massive interiors of these lava flows are more resistant to erosion than the underlying and overlying conglomerates and sandstones of the Copper Harbor Formation. As a result, harbors such those at Eagle Harbor and Copper Harbor are maintained by lava flows visible at their mouths. While not visible, lava flows occur at the mouth of Great Sand Bay too.

There are many extensive underwater fissure vein deposits which cross cut the Eagle River shoals located about 0.5 to 1 km offshore. Many of them are often quite rich in native copper and can contain long continuous stringers protruding up to 1.5 m in height and extending almost 6 meters in length. Most of veins are less than 50 cm in width and are primarily composed of quartz or calcite with minor amounts of laumontite , datolite, prehnite, and traces of silver. Veins will locally contain

clay pockets which can produce well defined copper crystal specimens. The largest copper specimen ever recovered underwater was a massive 17 ton unattached copper boulder in July of 2001. It was recovered from one of these vein deposits north of Jacobs Creek in about 9 m of water. To date, there have been 36 underwater copper veins discovered from the eastern tip of Great Sand Bay to Eagle River, about 3.2 km west.

## Stop 7: Hebard Park

*<u>Directions</u>: Continue driving 10.5 miles (16.9km) east on M-26 until arriving at Hebard Park conglomerate exposure on left. [UTM 5258659N 160428890E (NAD27 CONUS)]*

The description of this stop is reproduced from Bornhorst and Barron (2011).

The Copper Harbor Formation is exposed along the Lake Superior shoreline at Hebard Park (Fig. 10) and is stratigraphically above the Lake Shore Traps (Fig. 5). The lithologies at at this stop consist of interbedded conglomerates and sandstones that characterize the Copper Harbor Formation. Clast-supported conglomerate beds consist of rounded, cobble- to boulder-sized clasts with a matrix of coarse sand-sized subangular grains cemented with carbonate and iron oxide. Clasts are predominantly of silicic volcanic rocks, with subordinate basalt, pyroclastic, plutonic, and metamorphic rocks. Several finer grained interbeds higher in the exposed section exhibit crossbeds, current lineations, current ripples, parting lineation, and reduction spots. In particular, one should note the calcite-rich cemented zones that may represent vadose carbonate or paleocaliche (Kalliokoski, 1986). There is a thin continuous zone of laminated cryptoalgal carbonate, laterally-linked stromatolite, that is draped over cobbles and contorted layers in mudstone-siltstone.

## Stop 8: Hunter's Point Park

*<u>Directions:</u> Continue driving 2.4 miles (3.8km) east on M-26 to North Coast Rd. and then turn left. Drive 0.3 miles (0.4km) to Harbor Coast Lane and turn right. Drive 0.3miles (0.4km), park at the end of the road and walk down to shoreline. [UTM 5258263N 160432253E (NAD27 CONUS)]*

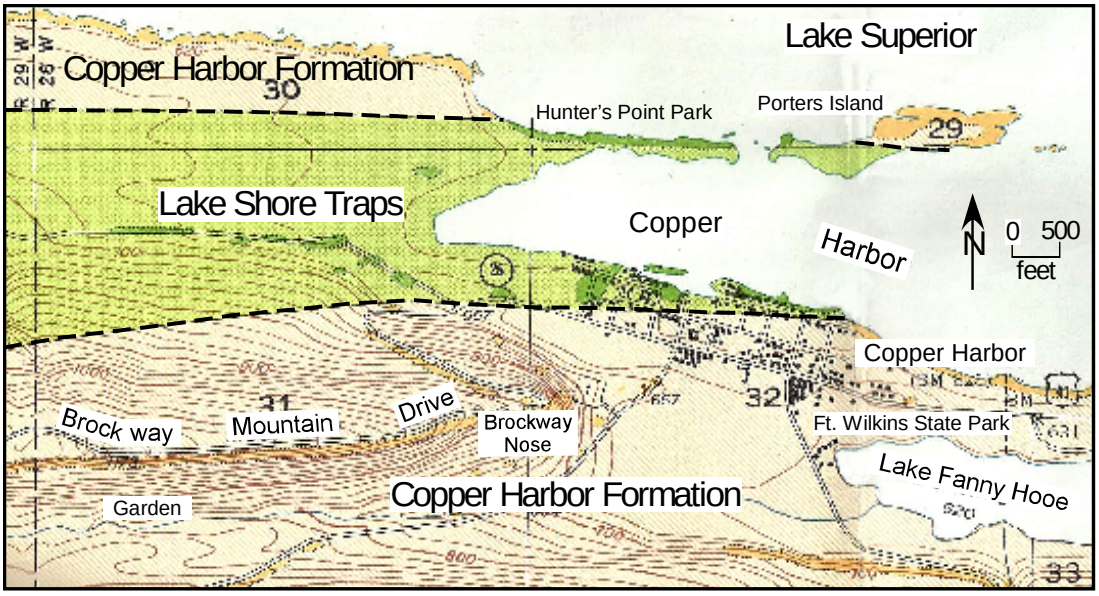



Figure 15: Geologic map of the Copper Harbor area taken directly from Cornwall (1955) showing the location of Hunter's Point (Stop 8), Brockway Nose (part of Stop 9), and Fort Wilkins Historic State Park (Stop 10).

Hunter's Point Park was established in 2005 when funding provided by the Michigan Natural Resources Trust Fund and many generous private donors (www.hunters-point.org) allowed the land to be purchased. Prior to becoming an official park the point was a popular hiking destination for visitors (Fig. 10 and 15). The land owners subdivided the area for residential housing which would have restricted public access without its conversion into a park. The name of Hunter's Point is uncertain but it could have been named after A.W. Hunter, an early resident in the town of Copper Harbor who purchased the point from the U.S. Government.

The Copper Harbor Formation is overall composed of volcanogenic clastic sedimentary rocks, dominantly conglomerates with lesser sandstone, siltstone, and shale such as observed at Stop 7. These rocks were deposited in a fining upward prograding alluvial fan complex (Elmore, 1984). Typically conglomerates are composed of clasts with a ratio of mafic-to-intermediate+felsic composition of about 2:1 (Daniels, 1982). Towards the tip of the Keweenaw Peninsula, the Copper Harbor Formation is informally subdivided into an inner (land side) "member" and an outer (lake side) "member." Between these two "members" there is a thin succession of interbedded lava flows collectively known as the Lake Shore Traps. The Lake Shore Traps consist of Fe-rich olivine tholeiite, basaltic andesite, and andesite lava that were erupted during the waning stage of volcanism within the MCR; the youngest flows tend to be more intermediate in

composition. At 1087.2 +/- 1.6 Ma (Davis and Paces, 1990), the Lake Shore Traps are among the youngest magmatism within the MCR. The thickest section of the Lake Shore Traps is about 15 km to the east at the tip of the peninsula. Volcanologically, the lower lava flows are interpreted as erupted as ponded sheets while the upper lava flows erupted on a low positive slope such as a shield volcano. The Lake Shore Traps were subaerially erupted pahoehoe lava flows.

At Hunter's Point, the top of the andesitic lava flows of Lake Shore Traps are conformably overlain by contact conglomerates of the Copper Harbor Formation (Fig. 15) simple geo map of Copper Harbor and Hunter's point). The strike of bedding is about E-W and dip is about $35^{o}$ to the north (towards the lake). The orientation of the contact is roughly parallel to the orientation of Hunter's Point.

From the Hunter's Point parking lot, follow the walkway to beach towards the west side of the point. As the walkway ends, you will be on outcrops of lava flows of the Lake Shore Traps (Fig. 15). Walking to the east, the beach gives way to a rocky shoreline. In erosional coves, you can see contacts between lava flows, represented by vesicular to amygdalodoidal andesitic lava (top of the lava flow) overlain by massive andesitic lava (massive interior of the overlying lava flow). The massive lava flow interiors within the Lake Shore Traps often retain relict olivine and interstitial glass due to the overall low degree of alteration (weathering and hydrothermal). Highly visible red hematitic bands form circular patterns within the massive interior; this banding is interpreted to be the result of alteration. Secondary minerals filling amygdules include agate, chalcedony, quartz, laumontite, analcite, calcite, and smectite in amygdules; this suite of minerals is equivalent to zeolite facies metamorphism. In contrast, in massive lava flow interiors within the Portage Lake Volcanics the olivine and interstitial glass are completely replaced by Mg-Fe phyllosilicates and amygdule filling minerals are equivalent to higher degree of metamorphism, greenschist facies. The Lake Shore Traps are geographically more distal to the thermal high and increased hydrothermal activity that resulted in the native copper deposits, hence, lower degree and grade of burial metamorphic/hydrothermal alteration.

To the west from the walkway, you can see a rocky point extending towards Lake Superior, the rocks in this point are conglomerates of the Copper Harbor Formation. The sharp contact between the uppermost lava flow of the Lake Shore Traps and the conglomerates can be viewed on the eastern edge of this rocky point. The conglomerate above the contact is dominated by rounded to sub-rounded boulders that are matrix-supported. There are proportionately more basaltic and andesitic clasts in this conglomerate bed than stratigraphically higher elsewhere along the Lake Superior shoreline such as at Stop 7 as these clasts are derived from erosion of the Lake Shore Traps updip towards the highlands on the edge of the rift (the updip rocks are now missing having been removed by erosion). The very poor sorting and fine matrix-supporting the clasts suggest this conglomerate could have been deposited as a debris flow. Sedimentary debris flows are common in alluvial fan depositional environments. The Copper Harbor Formation was deposited in an alluvial fan derived from highlands to the south in the vicinity of Keweenaw Bay.

Additional outcrops of the Copper Harbor Formation can be seen on the far western end of the cobble beach. These outcrops consist of interbedded conglomerates and sandstone that are typical of the formation as a whole. The conglomerates are described at Stop 7. There are several prominent

white-colored calcite -filled fractures (calcite veins) within these outcrops. The calcite veins are northerly oriented consistent with the orientation of faults cutting the Portage Lake Volcanics about 5 km to the south. Calcite veins are a common occurrence in the Copper Harbor Formation and some of them contain native copper such as those described at Stop 6, Great Sand Bay, and at Stop 10, Fort Wilkins.

## Stop 9: Brockway Nose and Brockway Mountain Viewpoints

*<u>Directions:</u> Continue driving 2.7 miles (3.8km) east on M-26 to Brockway Mtn. Drive. Turn right and drive 0.6 miles (0.4km) to Brockway Nose turnoff. [UTM 5257463N 160432304E (NAD27 CONUS)]*

Brockway Mountain Drive intersects M-26 just west of Copper Harbor. After a steep climb upwards there is a pullover at the second hairpin curve which is Brockway Nose viewpoint (Figs. 10 and 15). Brockway Nose provides an excellent view of Copper Harbor and Lake Fanny Hooe (Figure for Hunter's Point). The top of Brockway Mountain is accessed by continuing upwards from Brockway Nose. Brockway Mountain is a conglomerate ridge that reaches and elevation of over 400 m, with excellent views of the ridge and valley topography of the northern shore of the Keweenaw Peninsula.

From Brockway Nose viewpoint, the town of Copper Harbor is the prominent visible feature (Fig. 15). The town of Copper Harbor began as a boom town in 1843, following the discovery of copper in the vicinity. Porter's Island, at the mouth of Copper Harbor on the west side (left) was the site of the first government land office. Hunter's Point is west of Porter's Island. On the east side of the mouth of Copper Harbor, the Copper Harbor Lighthouse, built in 1866, is visible. Lake Fanny Hooe is located south of Copper Harbor. Fort Wilkins is located on the north shore of Lake Fanny Hooe on the thin strip of land between the lake and harbor. It was built in 1844, with the intent to protect the miners from potentially hostile Indians. Fort Wilkins is now a Historic State Park and is discussed more at Stop 10. Nearby, the Estivant Pine is a 2.06 $km^2$ nature sanctuary established in 1973, containing one the last stands of virgin white pines in the Midwest and the last stand in the Upper Peninsula. Some of the trees are up to 600 years old (www.michigannature.org). In 1955, the white pine was designated the state tree of Michigan. Copper Harbor and several other harbors between here and Eagle River are located within the Lake Shore Traps. Dipping massive interior of the basaltic to andesitic lava flows of the Lake Shore Traps occur at the head of the harbors.

From the Brockway Mountain viewpoint there are an excellent $360^o$ views. Underfoot, the Copper Harbor Conglomerate dips about $20^o$ to the north. Near the base of the ridge on the south side, opposite Lake Superior, there is an exposure of a single basaltic lava flow erupted as part of the Lake Shore Traps. With care, southwest of the gift shop at the high point, one can view the dipping conglomerates of the Copper Harbor Formation and see the lava flow near the base of the ridge.

To the west, the Lake Shore Traps form island chains on a prominent ridge in the vicinity of Agate Harbor and Esrey Park. The rocks of the Copper Harbor Formation are found in the drowned

valleys and along the outer ridge jutting into Agate Harbor and associated island chain. The ridges of the Lake Shore Traps and Copper Harbor Formation along the Keweenaw Peninsula's north shore are also the site of numerous shipwrecks.

Lake Bailey (with the small island) and Lake Upsom occupy a topographically low valley on a finer-grained clastic horizon (sandstone and siltstone) within the Copper Harbor Formation which is overall composed of conglomerates.

Just to the south of Lake Bailey, is the ridge of Mt. Lookout, marking the contact between the basal conglomerates of the Copper Harbor formation and the uppermost basalt lava flows of the Portage Lake Volcanics. The inland lake almost directly south, is Lake Medora, and just before the lake is a prominent ridge which marks the stratigraphic position of the Greenstone flow (see Stop 4).

In the distance, farther to the south across Lake Medora, is Mount Bohemia, a dioritic stock-sized intrusion within the lower section of the Portage Lake Volcanics.

To the southwest, a distant ridge is Gratiot Mountain, which is a small shallow rhyolite intrusive body that cuts the Portage Lake Volcanics.

To the east are the communities of Copper Harbor and Lake Fanny Hooe (better viewed from Brockway Nose), both of which occupy the same stratigraphic horizon as Lake Bailey. Just south of Copper Harbor is a golf course that is part of Brockway Mountain lodge. Brockway Mountain lodge was built during the Great Depression in the 1930's by the WPA.

To the north, Lake Superior is the prominent feature. On the skyline 65 km away, is Isle Royale National Park, which can be visible on a clear day. The skyline of Isle Royale is formed by the Greenstone Flow, as it is on the Peninsula. The beds on Isle Royale dip towards the Keweenaw Peninsula forming the Lake Superior "syncline." Viewed from here, the Midcontinent Rift proper extends from the Keweenaw Fault, originally a graben bounding fault on the edge of the rift, just south of Mt. Bohemia to the Isle Royale Fault, also originally a graben bounding fault on the edge of the rift, just northwest of Isle Royale.

Glacial erosion exposed Keweenawan and pre-Keweenawan relatively hard and competent bedrock on the edges of the Midcontinent rift system. Dipping well-cemented conglomerates of the Copper Harbor Formation are exposed at Brockway Mountain and basaltic lava flows of the Portage Lake Volcanics are exposed when viewing south. Both are relatively resistant to glacial erosion. On Isle Royale, on the southeast (Keweenaw side) are exposed the same conglomerates of the Copper Harbor Formation and on the northwest side, there are exposed basaltic lava flows of the Portage Lake Volcanics. In the center of what is now Lake Superior, much less competent, nearly flat lying, very fine sandstone and siltstone of the Freda Formation was at the bedrock surface. The latest glacial advance(s) preferentially eroded out the less competent rocks in the center of the rift, resulting in present day Lake Superior following the horseshoe shape of the MCR. Very large volumes of water filled the basin as a result of melting of the glaciers, turning it into a glacial lake. The Duluth Glacial Lake was the largest of these glacial lakes and only elevations above roughly 400 m (1,300 ft) were emergent. Brockway Mt. and Mt. Bohemia.

## Stop 10: Fort Wilkins Historic State Park

*<u>Directions</u>: Continue driving east on M-26 4.6 miles (7.4 km) until arriving at entrance to Fort Wilkins State Park on right. [UTM 5257338N 160434763E (NAD27 CONUS)]*

The description of this stop is reproduced with modifications from Bornhorst and Barron (2011).

Fort Wilkins was built in 1844 by the U.S. Army to provide order on the Keweenaw frontier and to protect the copper resources during the Civil War (Figs. 10 and 15). The army built 27 structures to house two full strength infantry divisions. After the soldiers were needed in the Mexican War in 1846, the fort was abandoned. Fort Wilkins became a State Park in 1923. During the 1930s under the Work Project Administration, the fort underwent extensive restoration. Many of these structures still survive today and have been either been restored or rebuilt after archeological excavations. Today, the restored buildings are a museum and contain exhibits on the mining history of the area. Fort Wilkins is a popular destination in the summer for recreation and camping.

Considerable exploration activity took place in the immediate vicinity of the fort, and there are shafts and exploration pits between Lake Fanny Hooe and the harbor, mostly from exploration during the period from1843 to 1846. Just north of the park store, several pits provide evidence of early mining activity by European settlers. The Pittsburgh and Boston Mining Company operated here in the 1840's on a vein of native copper within the Copper Harbor Formation; the vein was reported to be up to 0.3 m wide. This venture was not profitable. In 1853 and for several decades thereafter mining activity took place about 4.4 km south of the fort in a series of workings called the Clark Mine. The mineralization at the Clark Mine is hosted in both fissures and basalt flow tops. It consists of prehnite, epidote, analcite, quartz, laumontite, adularia, microcline, chlorite, datolite, calcite and several copper minerals including native copper, chalcocite, cuprite and tenorite. Agates are conspicuous as vesicle fillings in the Copper Harbor area especially in the Lake Shore Traps.

Opposite Fort Wilkins, on the harbor shoreline is a view of the Copper Harbor Lighthouse, one of the first on Lake Superior built in 1866. Near the lighthouse on the Lake Superior shoreline is the famous "green rock". The "green rock" is a vein that was described by Douglass Houghton. Houghton himself may have never really understood the uniqueness of the district. Conventional wisdom at the time led him to the interpretation that the “green rock” was the surficial alteration of a sulfide ore (Krause, 1992). Nevertheless, Houghton had a profound impact in promoting the district. His report to the Michigan legislature started the first major mining rush in North America to the Keweenaw Peninsula where the first economic discovery in 1845 at the Cliff Mine (Stop 11) was followed by many more until mining ceased in 1968. Douglas Houghton drowned in 1845 near Eagle River, MI while leading a geological expedition.

## Stop 11: Cliff Mine Rock Pile

*<u>Directions</u>: Get back onto M26 and continue driving east to stoplight in Copper Harbor. Turn right and drive 21.8 miles(35km) southwest on US41 past Phoenix until Cliff Dr. Turn right on Cliff Dr. and drive 0.4 miles(0.6km) to mine site.[UTM 52247173N 160400875E (NAD27 CONUS)]*

The description of this stop is reproduced with minor modifications from Bornhorst and Barron (2011).

Fissure (vein) deposits were of little importance to the overall copper production from the Keweenaw Peninsula native copper district (Figs 10 and 7). Only a few fissure mines, including the Cliff Mine, were profitable. The Cliff Mine worked the Cliff fissure (vein) from 1845 to 1887 and produced a total of about 38 million lbs of refined copper (Table 2). The Cliff fissure is nearly at right angles to the attitude of bedding and dips steeply to the east. The productive portion of the fissure is under the Greenstone flow. While most of the mineralization was confined to the fissure, some lava flow tops (amygdaloids) cut by the fissure contained native copper. Multiple large masses of native copper, some up to 100 tons, were taken out of the Cliff Mine. Among the fissure deposits, the Cliff Mine produced the most native silver. Minerals other than native copper and native silver include adularia, apophyllite, calcite, chlorastrolite, chlorite, datolite, epidote, laumontite, and prehnite (alphabetical). Many specimens contain multiple minerals and illustrate paragenetic relationships.

Fissures ranges in size from tight cracks to more than 3 m wide. In this part of the native copper district, fissures strike across the lava flows and dip steeply. Fissures formed as tension cracks related to bending of the lava beds, transverse to the axis of the MCR (Butler and Burbank, 1929). The steep ridge near the Cliff rock pile is the Greenstone flow (see also Stop 9). Here it makes up the entire high ridge from bottom to top and with a northward dip of about $25^{o}$. The very thick massive relatively impermeable interior of the Greenstone flow likely played an important role in the localization of native copper. The fissures acted as efficient pathways for fluid movement. On a local scale, fluids migrating upward through these open fractures and were impeded beneath the massive interior of the Greenstone flow and were forced to move laterally into adjacent permeable horizons. In general, flows beneath the thicker section of the Greenstone flow in this area contain more dispersed native copper than elsewhere, but economic deposits are not common.

## Stop 12: Jacobsville Formation M-26 Tamarack

*Directions: Drive 6 miles (9.6km) south west on Cliff Dr. until it intersects with US41/M26. Turn right and continue 5.2 miles (8.3km) to Calumet and turn left at the second stoplight onto Lake Linden Ave/M26 south. Drive 3.9 miles(6.3km) downhill to Lake Linden(blinking light) and turn right on M26. Drive 5.8 miles (9.3km) through Tamarack City to sandstone roadside outcrops. [UTM 5222120N 16038915E (NAD27 CONUS)]*

The description of this stop is reproduced from Bornhorst and Barron (2011).

The Jacobsville Sandstone is a red-bed succession consisting of feldspathic and quartzose sandstones, conglomerates, siltstones, and shales up to 1,000 m thick that were deposited by fluvial processes in a rift-flanking basin (Fig. 10). Overall, there are neither interbedded lava flows nor cross-cutting dikes and, thus, the age of the Jacobsville Sandstone is inferred to be ca. 1,060 to 1,020 Ma. Jacobsville sedimentation was the last Precambrian event associated with the development of the MCR. The Jacobsville Sandstone at this stop displays features characteristic of the unit as a whole. At the northeastern end of the outcrop, reddish shale and red-brown siltstone are exposed at the highway level. They are overlain by two fining-upward sequences of conglomerate and red, red-brown, and white cross-bedded sandstone. The lower conglomeratic bed is planar and can be traced 30 m to the southwest, along with the directly underlying shale and siltstone. Farther to the southwest, the section is almost entirely cross-bedded red sandstone; some beds are contorted and mottled. The sandstone consists of almost equal parts of rounded-to-sub-rounded quartz, feldspars, and lithic fragments. Clasts in the lower conglomerate are predominately sub-angular, and rhyolitic in composition, with subordinate mafic volcanic rocks.

Mining history will be viewed along M-26 from Lake Linden to Mason as an extension of the actual Stop described above, history summarized here from Molloy (2007). Just before leaving Lake Linden on the left is the C&H Mill. The C&H Mill was first built in 1867 with several expansions as milling practices changed and closed in 1956. Like Quincy, C&H also reclaimed copper from the sand tailings. The Houghton County Historical Museum is located on the edge of Lake Linden and exhibits mining and local history. Just outside of Lake Linden on the left are the remains of the Calumet and Hecla (C&H) smelter. C&H was the largest native copper producer in the district. The large building at the north end of the site next to the highway was the C&H mineral storage building where crushed and concentrated copper ore was smelted. It is now occupied by Peninsula Copper Industries which primarily recovers copper from scrap copper such as printed circuit boards to make copper sulfate as a fungicide for the wood preservative industry and to make other specialty copper compounds. About 2 miles (3.2 km) southwest of Lake Linden, is the only remaining steam stamp that was part of the Ahmeek Mill. The steam-driven stamp hammers could deliver about 104 blows per minute and process 7,000 tons of ore a day. About 1.5 miles southwest of the Ahmeek Mill, the Quincy Mining Company built a reclamation plant in 1942 to 1943 to reprocess the stamp sand tailings along Torch Lake, and from 1943 to 1967 recovered approximately 50,000 tons of copper. One of the mining dredges used in the recovery process sank in a storm in 1956 and is located just offshore. The foundations between the road and the dredge are part of the Quincy Mills for processing native copper ore.

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# Field Trip 2

# Caledonia Mine, Keweenaw Peninsula Native Copper District, Ontonagon County, Michigan

Theodore J. Bornhorst
*A.E. Seaman Mineral Museum, Michigan Technological University, 1404 E. Sharon Avenue, Houghton, MI 49931*

Robert J. Barron
*Department of Geological and Mining Engineering and Sciences, Michigan Technological University, 1400 Townsend Drive, Houghton, MI 49931*

Richard C. Whiteman
*Red Metal Minerals, 202 Ontonagon Street, Ontonagon, MI 49953*

*Directions: Leave downtown Houghton and head south on M-26 towards South Range. Stay on M-26(highway turns into M-38 past the Mass City turnoff) for 39 miles past Greenland(Pat's Auto & Sports Center) to Ridge Road. Turn left (south) on Ridge Road approximately 1 mile to Caledonia Rd. Drive southwest 1.7 miles (2.7 km) to the Caledonia Mine. The Caledonia mine is privately owned and permission is required to enter this property [UTM 5179684N 160338022E (NAD27 CONUS)]*

**The Caledonia Mine, surface and underground, is strictly private property. Permission is required to enter the property.**

## Introduction

The Caledonia Mine is part of the Keweenaw Peninsula native copper district of the western Upper Peninsula of Michigan (Fig. 1). The Caledonia Mine is located in the Greenland-Mass subdistrict about 40 km southwest of the Baltic Mine, the southernmost major native copper mine (Fig. 2). In total the mines on the Evergreen succession produced about 73 million lbs of copper at grades ranging from 0.5 to 1.25 % (Weege and Pollock, 1971) and, as compared to total district production of 11,000 million lbs of copper, this subdistrict was a minor producer. The largest producer among the Greenland-Mass subdistrict mines was the Mass Mine which produced about 51 million lb of refined copper from 1851 to 1923 (Butler and Burbank, 1929). The Adventure Mine produced about 11 million lb of copper. Despite its low production of copper, the geologic characteristics of the Greenland-Mass subdistrict deposits are typical of the native copper deposits elsewhere in the Keweenaw Peninsula. Since native copper mining in the Keweenaw Peninsula ceased in 1968, underground access to observe or study the native copper deposits is limited today to four mines, two of which are in the Greenland-Mass subdistrict (Caledonia and Adventure Mines). The Caledonia Mine is one of the few remaining localities where a typical native copper ore body hosted by the top of a basalt lava flow can be observed up close underground.

This field trip guide relies on existing publications by Bornhorst and Whiteman (1992 and 1995) and Bornhorst and Barron (2011). The geologic and human history overview is summarized from Bornhorst and Lankton (2009). Since professional and collector oriented underground field trips to the Caledonia Mine are common, the introductory explanations have been expanded to provide a more readily stand alone field trip guide. Underground, the Caledonia Mine is relatively dry and regular field boots are usually sufficient; hard hats and lights are required. The field trip involves an easy walk underground to observe the character of native copper mineralization in an adit and a drift that parallels the strike of the tabular native copper ore body (lode) that is hosted by the top of the Knowlton lava flow. The cross-cutting adit provides opportunity to observe the Knowlton lode in cross-section as well as to observe underlying lava flows of the Evergreen succession. An optional more difficult segment of the field trip involves climbing up into an underground stope on the Knowlton lode to observe the character of mineralization and to collect specimens. Specimens of native copper and associated minerals from hosted by rocks of the Caledonia Mine can be collected on the rock pile adjacent to the adit. The Caledonia Mine is owned and operated by Red Metal Minerals as an educational facility and to recover specimens for resale in the general public and mineral collector markets.

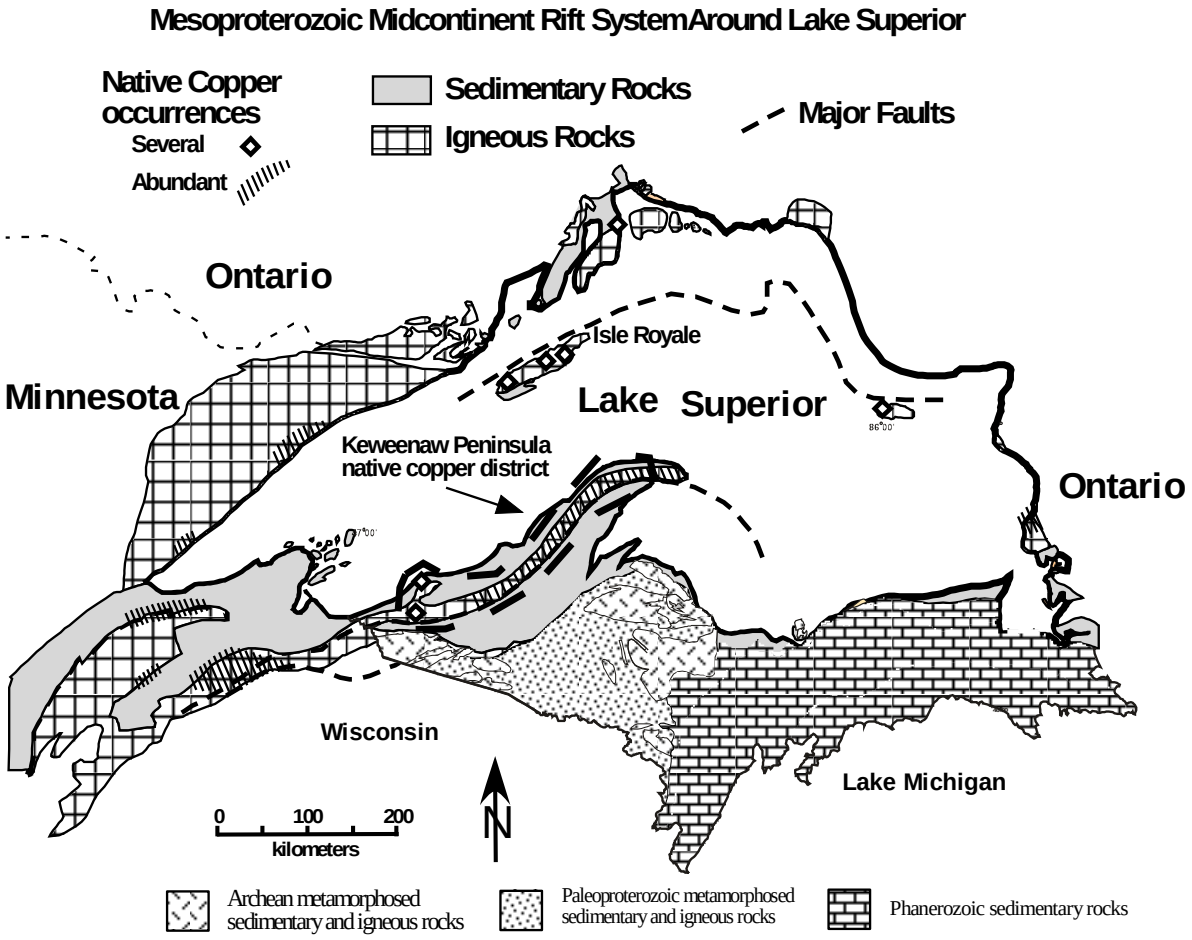



Figure 1: Generalized bedrock geologic map of showing the Keweenaw Peninsula Native Copper District.

**Regional Geologic and Human History Overview**

The Keweenaw Peninsula is home to the largest known accumulation of native copper on the planet termed the Keweenaw Peninsula native copper district. The district is unique in comparison to copper mining districts elsewhere in that native copper comprises nearly all of the metallic minerals in the mined ore bodies. Approximately 11 billion pounds of refined copper from 380 million tons of ore were produced from native copper mines from 1845 to 1968 (Weege and Pollock, 1971). Small quantities of native silver occur with the native copper. Copper sulfides are uncommon in the Keweenaw Peninsula native copper district although chalcocite occurs in veinlets cutting the deposits (White, 1968). Near the tip of the Keweenaw Peninsula, there are several small unmined chalcocite-dominated deposits but their connection with the native copper deposits is uncertain (see Field Trip 3 this volume, Maki and Bornhorst, 1999).

The native copper deposits of the Keweenaw Peninsula are hosted by rocks of the Mesoproterozoic Midcontinent Rift (MCR) (Fig. 1and 2). The MCR is filled with more than 25 km of volcanic rocks and 8 km of clastic sedimentary rocks (Cannon et al., 1989 and 1993). This thick succession of rocks was emplaced between about 1.15 to 1.03 Ga (Cannon et al., 1989; Davis and Paces, 1990; Heaman et al., 2007). Volcanic rocks were erupted on land surface initially over a broad area above a mantle plume and later, were erupted from fissure volcanoes within the normal fault-bounded rift graben. The volcanic rocks erupted during this syn-rift phase of the MCR were predominantly subaerial tholeiitic flood basalt lava flows. The subaerial basalt lava flows have a top which is vesicular (amygdaloid) and/or brecciated (fragmental amygdaloid) underlain by a massive (vesicle-free) interior. The typical flow is 10 to 20 m thick. Minor gravels and sands were deposited on top of the lava flows during hiatuses in volcanic activity and today occur as red conglomerate and sandstone that are interbedded with the lava flows. After active rifting and volcanic activity ended, the rift basin continued to sag. Rivers carried gravels, sands, silts, and muds to fill this sagging rift basin, and with subsequent burial they were lithified into clastic sedimentary rocks which occupy the center portion of the rift today (Merk and Jirsa, 1982). Volcanic rocks crop out around the margin of the rift (Fig. 1).

The last and final phase of the MCR resulted from a regional compressional event due to collision of continental land mass along the eastern edge of North America at that time (Grenville Orogeny, Cannon, 1994). Compression inverted the rift-bounding normal faults into reverse faults as well as folding, faulting, and fracturing rift-filling volcanic and clastic sedimentary rocks. Native copper and related minerals were emplaced during this regional compressional event about 1.06 to 1.04 Ga (Bornhorst, 1997).

For roughly 500 million years, from about 1.0 Ga to 500 Ma, there were no geologic events recorded by rocks of the Upper Peninsula. During this time interval, erosion exposed the native copper deposits to the surface and downward percolating oxidizing groundwaters had access to alter the native copper (Bornhorst and Robinson, 2004). After being buried beneath Phanerozoic sedimentary rocks (500 Ma to 175 Ma) (Catacosinos and others 2001), Pleistocene continental glaciations removed all but a few outliers of these rocks from the Keweenaw Peninsula and exposed the native copper deposits at the surface.

Native people began exploiting native copper by *ca.* 7,000 years ago as the land surface of the Keweenaw Peninsula emerged above the retreating glacial lake levels. At first these prehistoric ancient miners likely found boulders of native copper (locally termed float copper) with their distinctive green weathered crust of malachite among the brown, gray, red, and white rocks. As they discovered the usefulness of native copper, they moved on to mining of bedrock. Because of the scars these early exploits left on the landscape, most mines of the Keweenaw Peninsula were rediscovered later including the Caledonia and those nearby. The first major mining rush in North America was started by Douglass Houghton through his report to the Michigan legislature in 1841. The Cliff Mine became the first profitable mine in the district in 1849. The Minesota Mine, in the southwest cluster near Caledonia Mine (Fig. 3) became profitable soon after the Cliff. In 1880, copper production from native copper mines of the Keweenaw Peninsula accounted for up to 80 % of the nation's copper production. The peak copper production occurred in 1916 at 267 million pounds with mining ending in 1968. The Keweenaw National Historical Park was created in 1992 to preserve and interpret the historical importance of native copper mining to the history of the U.S.

**Native Copper Deposits of the Keweenaw Peninsula**

The pre-mining geologic resource of the Mesoproterozoic Keweenaw Peninsula native copper district totaled about 20 billion lbs of Cu (Bornhorst and Barron, 2011). Most of the native copper in the district is hosted in the permeable and porous brecciated and amygdaloidal lava flow tops (~58.5% of production) and interflow conglomerate-sandstone horizons (~39.5% of production). The ore is "sandwiched" above and below between barren massive basalt that lacks permeability and porosity and is geometrically found in tabular bodies between 3 and 5 m thick that have the same orientation as surrounding host rocks, i.e., stratiform lode. The typical lode has a lateral extent of 1.5 to 11 km and extends down-dip 1.5 to 2.6 km (Butler and Burbank, 1929; White, 1968). Native copper fills open spaces from a few cm across (e.g., vesicle-fillings) to small-to-moderately sized openings (e.g., space between lava flow top breccia fragments or between clasts in conglomerate) that contain native copper masses weighing up to several pounds and rarely weighing tons. A minor amount of native copper was produced, ~ 2%, from high-angle tabular veins that cut across the volcanic-dominated strata.

Native copper is closely associated with over 100 different minerals in the Keweenaw Peninsula although only about 25 of them are common. These minerals fill the same open spaces along with and instead of native copper (Butler and Burbank, 1929; Stoiber and Davidson, 1959; White, 1968). The suite of minerals is similar to those found where rocks have undergone very low to low grade burial metamorphism, < $300^{O}$C. Overall, higher temperature assemblages are spatially associated with the area of native copper deposits where the thermal anomaly was greatest because of focused hydrothermal fluid flow. In areas more distal to the deposits, the open spaces are filled with lower temperature assemblages. At any one location, there is a recognizable sequence in the precipitation of minerals from hydrothermal fluids due to changing hydrothermal fluid temperature and composition. The absolute age of the hydrothermal activity is coincident with the age of the regional compressional event at about 1.06 to 1.04 Ga (Bornhorst et al., 1988). The compressional event provided the plumbing system, faults/fractures, that facilitated movement of the hydrothermal fluids into the sites of future mineable deposits of native copper (Bornhorst, 1997).

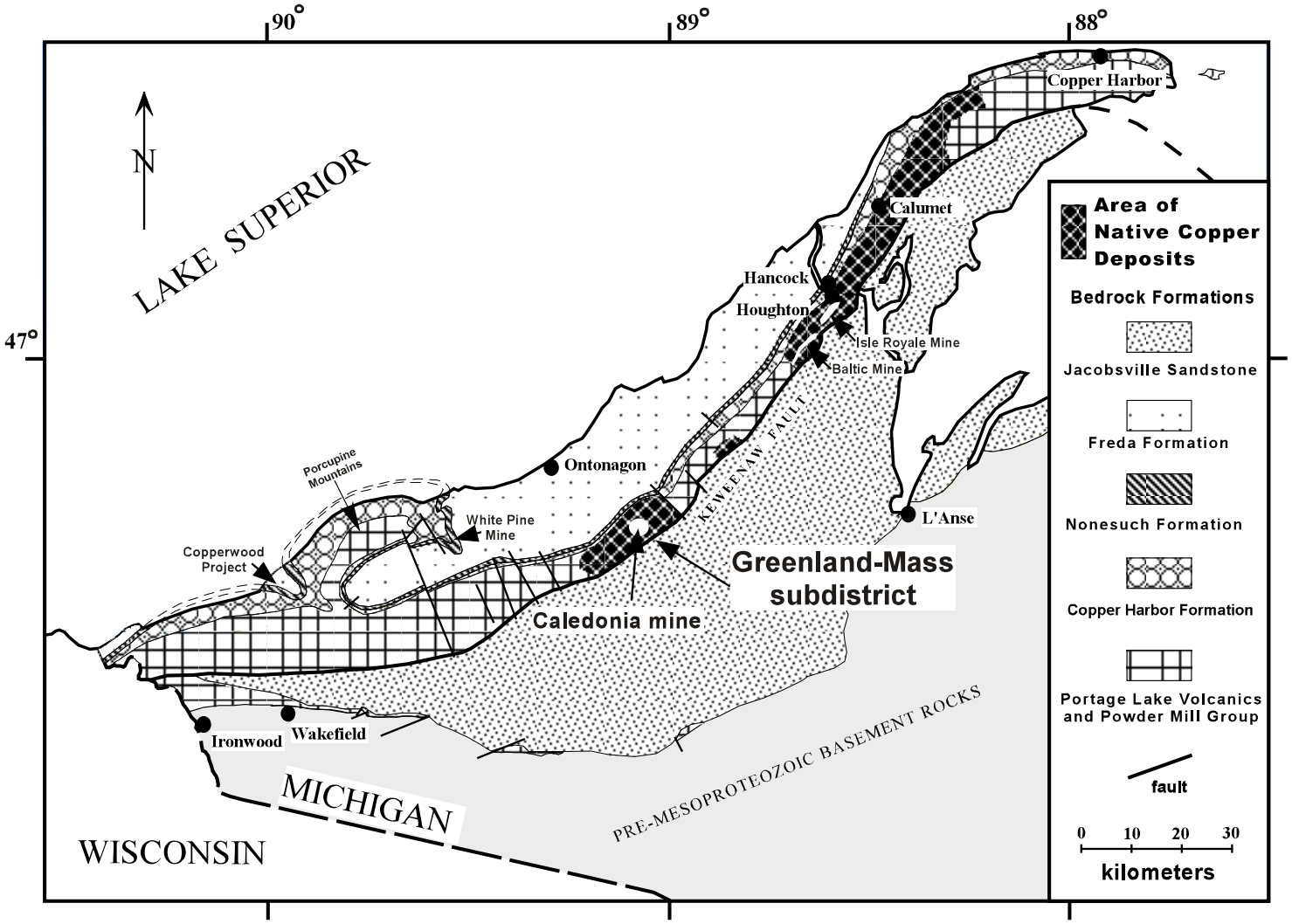



Figure 2: Bedrock geologic map of the western part of the Upper Peninsula of Michigan showing the location of the Greenland-Mass subdistrict and the Caledonia Mine.

The native copper mineralizing event was widespread throughout the exposed MCR (Fig. 1). Burial metamorphism at depth of rocks down-dip from the deposits was the likely source of the mineralizing hydrothermal fluids; small amounts of copper and other constituents were leached from the rift-filling basalt-dominated rocks. Subaerial eruption of the basalt lava flows likely resulted in the degassing of most of the contained sulfur leaving them sulfur poor. Thus, the hydrothermal fluids generated from them were low in sulfur and the movement of these fluids through the same sulfur poor rocks resulted in sites of ore deposition where host rocks were also sulfur poor. This low sulfur environment favored the deposition of native copper rather than copper sulfide. The heating of the volcanic rocks during burial probably reached a maximum millions of years after they were erupted and it was most likely the coincidence of increased fluids generated at this temperature maximum with the regional compressional event which played a critical role in providing the plumbing system necessary for producing the deposits (Bornhorst, 1997).

**Geology of the Evergreen Series**

The native copper mines in the Greenland-Mass subdistrict produced native copper from the tops of rift-filling lava flows that comprise the Evergreen succession within the Portage Lake Volcanics (Fig. 3 and 4). The Evergreen succession of basalt lava flows have a total thickness of about 210 m. The individual copper-rich lava flows within the succession were each informally named. From bottom to top the Evergreen succession consists of: the Evergreen flow: a 3 to 15 m thick plagioclase porphyritic otherwise aphanitic basalt lava flow; the Ogima flow, a 30 to 43 m thick slightly plagioclase glomerophyritic basalt lava flow; the Butler flow, a 15 to 27 m thick plagioclase glomerporphyritic basalt lava flow; and a horizon of thin plagioclase glomeroporphryitic flows 75 to 90 m thick. This latter stratigraphic horizon of multiple flows includes the Mass flow. The South Knowlton flow overlies this horizon and is a plagioclase glomeroporphyritic lava flow up to 15 m thick and at the top of the Evergreen succession is the Knowlton flow, a 9 to 21 m thick plagioclase glomeroporphyritic lava flow (Calumet and Hecla, 1958). The volcanic rocks nearby underlying the Evergreen succession are ophitic and aphanitic basalt lava flows. The nearby overlying volcanic rocks are ophitic basalt lava flows. The Evergreen succession is stratigraphically at the level of the Isle Royale flow near Houghton.

The tops of these flows were productive over a strike length of about 5 km. Of the lava flows in the Evergreen succession, the Butler flow top yielded the most copper followed by the Evergreen and Knowlton flow tops which also yielded significant amounts of copper. Most of the Evergreen succession basalt lava flow tops are brecciated (fragmental amygdaloid) with considerable lateral (along strike) variation in the degree of brecciation and thickness. In some areas, thin lava flows lack brecciated tops and are simply vesicular (amygdaloid). In general, the best grades of copper occur where the brecciated flow top thickens. Secondary minerals in all of the flow tops are quite similar. Quartz, feldspar, pumpellyite, chlorite, calcite, and epidote are abundant minerals filling amygdules and spaces between breccia fragments. There is less abundant native silver, prehnite, datolite, and laumontite.

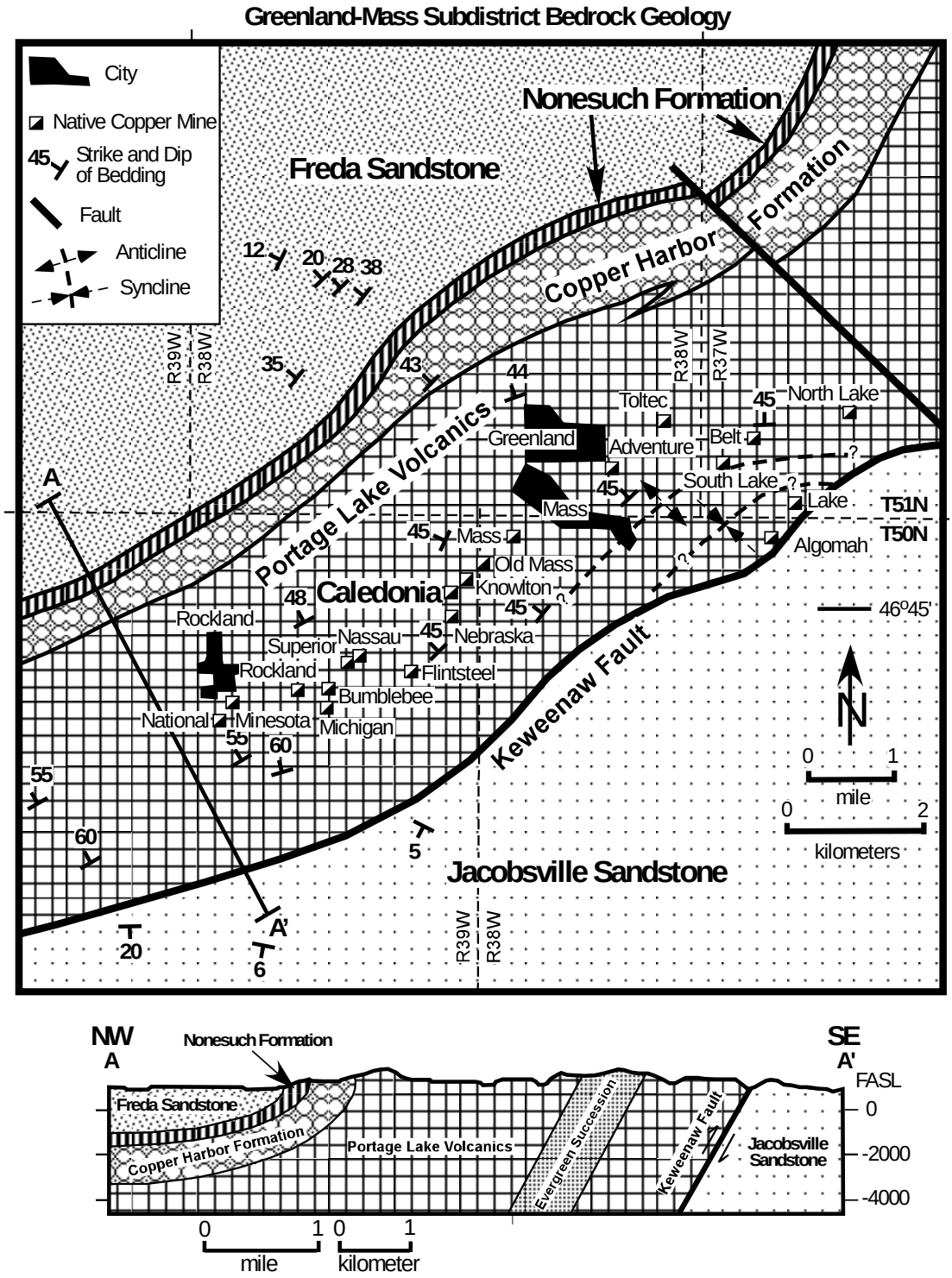



Figure 3: Generalized bedrock geologic map of the Greenland-Mass subdistrict of the Keweenaw Peninsula Native Copper District showing the location of native copper mines. Geologic cross section shows the Evergreen Succession within the Portage Lake Volcanics; a lithostratigraphic column is given in Figure 4. Subdistrict bedrock geology and cross section modified from Whitlow (1974).

The Evergreen succession in the Greenland-Mass subdistrict dips about 45$^{o}$ NW towards Lake Superior and forms a local broad open anticlinal structural bend (Fig. 3). The largest mine, the Mass Mine, occurs near the maximum bend in this anticline. Faults with significant vertical displacement are uncommon as most have displacement of < 1 m. There are multiple veins in tension fractures that cut perpendicular across the lava flows in association with the anticline (Butler and Burbank, 1929). However, some veins are parallel to strike of the lava flows but dip in the opposite direction. In the stratigraphically equivalent Isle Royale lode, Broderick (1931) describes similar strike parallel veins which he interpreted to be feeders of ore fluid into the top of the lava flow.




Figure 4: Stratigraphic position of the Evergreen succession that hosts native copper deposits of the Greenland-Mass subdistrict as compared to lithostratigraphic units of the Keweenaw Peninsula, Michigan. Units from the Powder Mill Group to the Jacobsville are all Mesoproterozoic in age and related to the Midcontinent rift.

**Geology of the Caledonia Mine**

In the context of mines in the Keweenaw Peninsula native copper district, the Caledonia Mine was very small, producing only about 6.8 million lb of refined copper from the top of the Knowlton basalt lava flow (Knowlton lode), the youngest basalt lava flow of the Evergreen Series (Fig. 4). The near horizontal adit of the mine follows approximately along strike of the Knowlton lava flow where it connects with the Knowlton Mine (Fig. 5) and to where it connects to the stopes of the Mass Mine “C” shaft (not shown). At the Mass Mine, the stratigraphically lower Bulter lava flow top was the principal focus of native copper mining. In the Greenland-Mass subdistrict the Butler lava flow top was developed for about 2000 m along strike and to a maximum depth of 300 m along dip. The most abundant secondary minerals in the Butler are quartz and calcite with slightly lesser amounts of K-feldspar and epidote. Prehnite and pumpellyite are usually much less abundant and chlorite is present in amounts < 1 %. The Butler contains a high number of veins, usually they strike subparallel to the strike of the Butler lava flow top and have dips both similar to the dip of bedding and at a high angle to bedding (Butler and Burbank, 1929).

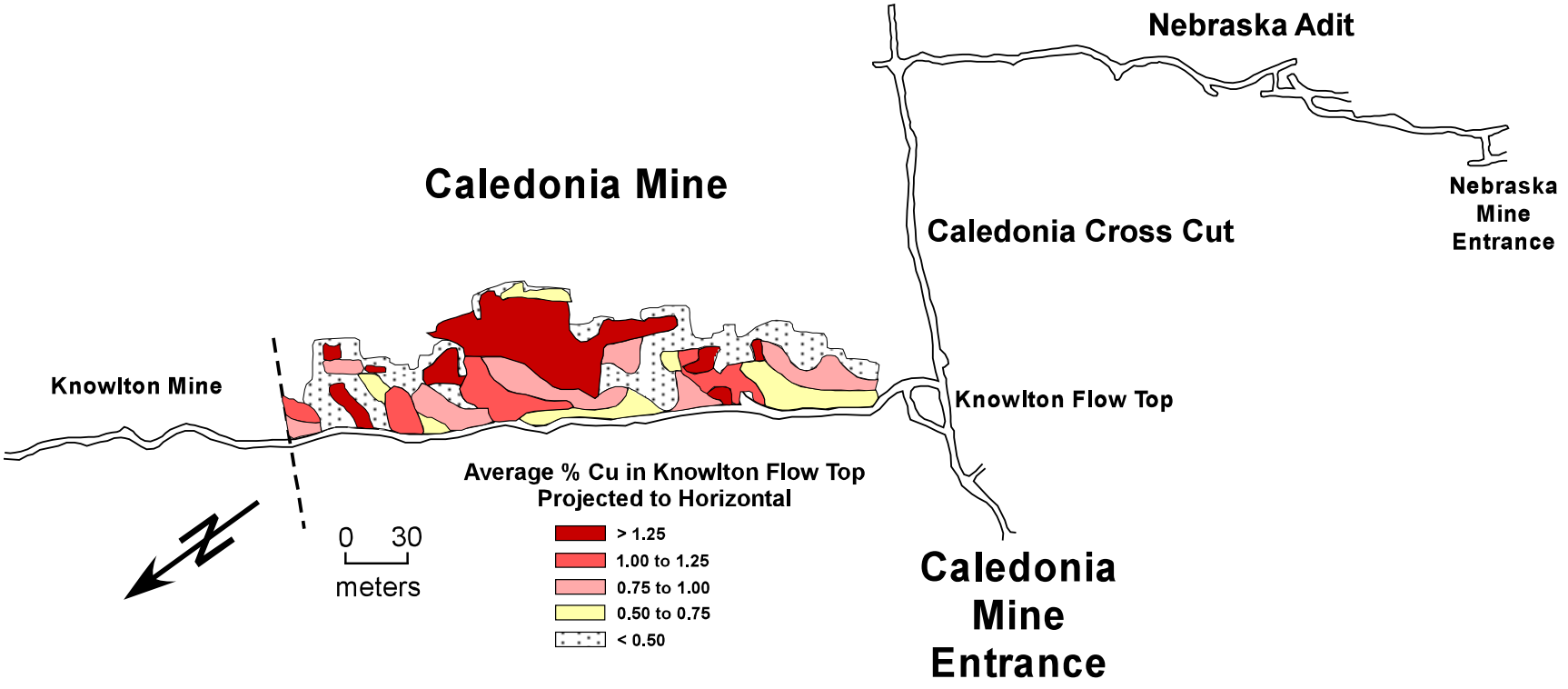



Figure 5: Map of the underground workings at the level of the Caledonia Mine entrance. The average grade of Cu is shown for the mined out areas of the Knowlton flow top. Map and Cu grade from Calumet and Hecla (1958).

In the Greenland-Mass subdistrict, the Knowlton flow top was developed for about 3000 m along strike and to a maximum depth of about 375 m. The Knowlton was the focus of native copper mining at the Caledonia Mine. The Knowlton lava flow top is a brecciated flow top or fragmental amygdaloid. Stopes at Caledonia were raised on the Knowlton lode upwards from the drift toward the topographic high of the Caledonia bluff (Fig. 6). The floor (footwall) of the stopes reflects the original depositional irregularities of the top of the underlying lava flow such as gentle flexures. The average thickness of the Knowlton lava flow top is about 2.5 m but locally it can thicken to around 6 m (Calumet and Hecla, 1958). In general, a thicker flow top results in better ore. While most of the ore occurs in the top of the Knowlton lava flow top, there are pockets of ore that extend into the underlying Knowlton massive flow interior footwall and are closely associated with strike-parallel fractures and veins. The grade of the ore in the flow top may correlate with these footwall pockets of ore; there is an approximate strike parallel (drift parallel) orientation of the grade of the ore (Fig. 5). The workings of the Caledonia Mine provide excellent access to observe the 3-D geometry of the Knowlton lode. An elongated volume of highly epidotized basalt characterizes a footwall ore pocket that was mined out in the 1990s by Red Metal Minerals and will be visited on this field trip.

Veins are of two types. Veins within the Knowlton flow top that extend into underlying Knowlton massive flow interior and contain the same basic minerals as those found in the lode itself (including native copper) are considered synchronous with the native copper deposit at the Caledonia Mine. Native copper occurs as small to large masses in the fractures and seems to be more abundant in the overlying adjacent tabular ore body within the lava flow top. The fractures typically have little or no displacement. Adjacent to the fractures even massive basalt can be highly altered and host native copper. One main-stage vein that has been studied in more detail by Bornhorst strikes subparallel with the strike of the flow top but dips more steeply, about 80$^{o}$NW dip of the vein as compared to 45$^{o}$NW of the lava flow top. This vein has been traced along strike for over 100 m. It extends into the footwall, but is hard to identify as it enters the lode. Within this vein the intensity of alteration varies from slight to very high. Original basalt can be completely converted to a green soft epidote and lesser chlorite rock, or a hard epidote and lesser quartz rock. Overall mineralogy and paragenesis in the vein is similar to the lode. The vein contains pockets of a soft blue-green mineral identified as corrensite by XRD (mixed layered clay mineral with 50/50 chlorite and smectite unit cells stacked in perfect alternation).

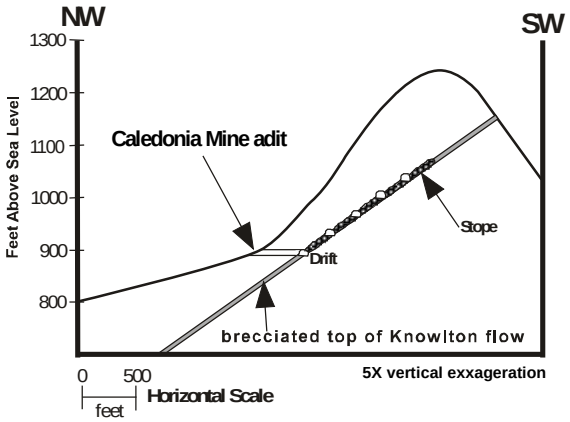



Figure 6: Cross-section sketch of the topography of Caledonia bluff showing the positions of the Caledonia Mine adit and the top of the Knowlton lava flow.

This vein has yielded outstanding museum quality specimens of crystalline native silver (now part of the A.E. Seaman Mineral Museum collection). These native silver specimens were encased in white calcite; the calcite was removed by acid cleaning. This vein also yielded clusters of colorless calcite crystals internally laced with native copper from open vugs. Several masses of native copper weighing over 100 kg and small groups of copper crystals originally encased in white calcite (removed by acid cleaning) have also been recovered during exploration. Some of the copper crystals were coated with very small cubic native silver crystals. There are multiple other veins at Caledonia synchronous with native copper precipitation. One of these veins along the drift that will be visited is notable for hosting datolite; the datolite commonly contains very-fine inclusions of native copper. The occurrence of veins at the Caledonia Mine is quite similar to veins described by Broderick (1931) occurring in the Baltic and Isle Royale Mines near Houghton. At Caledonia, they are interpreted as pathways for ascending hydrothermal fluids and thus, played an important role in the deposition of native copper and associated minerals. At Caledonia, there are veins that crosscut the native copper mineralized lode. These post-copper mineralization veins contain calcite and laumontite and are barren of copper. Several of these post-mineral veins are readily visible along the down-dip side of the Caledonia adit.

At the Caledonia Mine, the most abundant mineral filling amygdules and spaces between fragments in the Knowlton lode is calcite which is closely followed by subequal amounts of quartz, epidote and red K-feldspar. There are lesser amounts of prehnite, pumpellyite and chlorite. Native copper is present in small amounts with an average grade of about 1.2 % Cu. Native silver and datolite are present in much lesser amounts. Least abundant are laumontite, adularia, and corrensite (clay mineral). No major differences exist in the abundance of secondary minerals averaged over the scale of 100's of meters. In contrast, over the scale of a few meters the distribution of secondary minerals is variable to highly variable. While a secondary mineral may be completely absent in one zone and extremely abundant in another, the meter scale variation does display a degree of regularity. For example, within the Knowlton flow top secondary minerals may occur in overlapping bands; the bands are consistent with the progressive filling of open spaces indicated by amygdule paragenesis. The intensity of alteration is highest near both the hanging wall and footwall of the brecciated flow top lode where apparently there was preferential flow of hydrothermal fluids. Distribution of secondary minerals also has a poorly defined correlation with the occurrence of synchronous veins. In general, native copper tends to be more commonly associated with epidote, calcite, and quartz. Rarely is native copper abundant in areas with abundant K-feldspar.

Paragenetically, K-feldspar is an early formed mineral followed by epidote and then datolite, prehnite, pumpellyite, chlorite, calcite and quartz. Native copper is found as inclusions in epidote, calcite, quartz, and datolite. Much of the calcite that is overall synchronous with native copper precipitation does not contain obvious inclusions of native copper. Post-native copper mineralization hydrothermal minerals in veins and open space fillings as coatings on earlier formed minerals include calcite, laumontite, adularia, and corrensite. These likely formed from superposition of later lower temperature hydrothermal fluids on earlier higher temperature formed minerals associated with native copper during collapse of the hydrothermal system. This relationship is found elsewhere in the broader district.

Post-emplacement alteration of hydrothermal minerals is most obvious for native copper. At Caledonia, tenorite and cuprite (Cu oxide) often but not always occurs as a thin coating on native copper that is found in open space fillings. The tenorite and cuprite could have its origin during Precambrian weathering and downward-percolating ancient groundwater leading to supergene alteration of the native copper prior to Phanerozoic marine submergence or during more recently since Pleistocene glaciations eroded and exposed the Caledonia deposits at the surface. Today, only a few fractures within the stopes of Knowlton lode are damp with meteoric groundwater despite the shallow depth, an argument in favor of Precambrian tenorite and cuprite. However, the presence of modern groundwater flow into the mine, such as near the intersection of the cross-cut and the Knowlton drift, suggests Pleistocene age for the tenorite and cuprite cannot be precluded. In addition to tenorite and cuprite, there are occasional copper carbonate minerals (such as malachite), brochantite (hydrated Cu sulfate), atacamite (hydrated Cu chloride) and unknown green to blue-green minerals on native copper surfaces. These may also be supergene in origin. However, there is at least one mineral, gerhardtite (hydrated Cu nitrate) that is the result of post-mining chemical reactions.

An extension of the Caledonia adit cuts across the lava flows towards the Nebraska Mine; the cross-cut was a component of Calumet and Hecla's exploration program (Fig. 5). This cross-cut intersects the South Knowlton lava flow top directly below the base of the Knowlton lava flow. An NSF-sponsored Teachers Earth Science Institute (educating middle and high school teachers about mining) drilled, blasted, and mucked out the South Knowlton lava flow top to its present expanded opening in the early 2000s. Below the South Knowlton, there are several thin basalt lava flows that can be identified before the stratigraphic level of the Butler lava flow. While the Butler lava flow top is readily identified in the cross-cut by abundant amygdules filled with K-feldspar and calcite, it lacks significant native copper here. A small fault can also be seen along the cross-cut. This cross-cut and adit connects the Caledonia Mine to the Nebraska Mine where the Butler lava flow top was a principal target of mining.

**Mining History of the Caledonia Mine**

The Caledonia Mining Company began its operations in 1863 after acquisition of the mining rights of the former Nebraska Company and acquisition of the adjacent Kansas Properties. The workings at that time consisted of a horizontal adit driven about 90 m to the Butler deposit on the west end of the Caledonia bluff and two shafts about 60 m deep (Nebraska Mine) (Fig. 5). A vein showing mineralization was explored on the north side of the bluff by four adits and while the vein proved to contain too little copper, the adits intersected the Knowlton, South Knowlton, Mass, and Butler lava flow tops. About 900,000 lbs of refined was produced from 1863 to 1870 by the mining of native copper at a grade of about 1.25 % Cu from the Knowlton and Butler lava flow tops. Production halted in 1870 when a fire destroyed the processing facility. Subsequently, the Caledonia Mining Company acquired the Flintsteel properties in 1870 and despite investing in a new processing facility, the Caledonia operations closed before significant ore was processed. Captain Martin leased the Caledonia properties and from 1873 to 1881 produced more than 330,000 lbs of copper, including a single mass weighing 80,000 lbs (40 short tons). After the lease expired, mining ceased. In 1901 there was a failed proposal to merge the Caledonia properties with other mineral rights in the Greenland-Mass subdistrict and to construct a processing facility on Lake Superior some distance away. There was no reported mining from the Caledonia properties for 56 years from 1881 to 1937.

The Calumet and Hecla Consolidated Mining Company was the major producer of copper from the mines north of Houghton. Calumet and Hecla did exploration core drilling and reopened the Caledonia adit. They drifted some 600 m along the strike of the Knowlton flow top and estimated the grade of native copper ore to be 1.45 % Cu. The workings from the nearby Nebraska Mine were connected to the Caledonia with a cross-cut and adit. Exploration of the Caledonia Mine by Calumet and Hecla ended in c.a. 1941 as a result of World War II. After World War II, Calumet and Hecla resumed exploration of Caledonia and removed a 200 ton bulk sample from the Knowlton lode in 1950. The sample had a very promising grade of 1.84 % Cu. From 1951 to 1958, Calumet and Hecla produced 5.55 million lbs of refined copper with an average grade of 1.24 %. This program included the stoping on the Knowlton lode visible above the drift today. Subsequently, Copper Range Company acquired the mineral rights at the Caledonia Mine. The Caledonia Mine was a candidate for in-situ leaching of Cu, hence a limited evaluation of the mine was completed by Copper Range and the U.S. Bureau of Mines from 1971 to 1972. The in-situ leaching option was abandoned due to potential problems with groundwater pollution.

In 1985, Red Metal Minerals acquired the mineral rights for the Caledonia Mine and other properties in the Greenland-Mass subdistrict from the Copper Range Company. The Caledonia adit was reopened and reconditioned to undertake a limited program of exploration. Red Metal Minerals mucked out broken rock as well as drilled and blasted new areas to recover native copper and other minerals. Over 28 years from 1985 to present, Red Metal has removed a single mass weighing 3000 lbs. Masses of recovered native copper are sold to the general public and mineral collectors. Red Metal distributes native copper on a wholesale basis to retail outlets that distribute Caledonia native copper around the world. You may find native copper from the Caledonia Mine for sale in unexpected places, even visitor gift shops at other copper mines! A large mass of native copper is on exhibit at the A.E. Seaman Mineral Museum, on the campus of Michigan Tech in Houghton. When the native copper is dispersed through the rock, Red Metal uses this material to prepare decorative bookends and cut slabs. The Caledonia Mine has yielded minerals sought after by mineral collectors such as native copper crystals, native silver crystals, datolite nodules, copper in calcite crystals as well as adularia and epidote. At this time, Red Metal does not undertake in underground drilling and blasting to recover specimens from the mine. Instead, the Caledonia Mine serves as an educational facility, used by Michigan Tech and others, and to recover specimens from already broken rock for resale in the general public and mineral collector markets. Since most of the native copper mines of the Keweenaw Peninsula are closed and flooded, the Caledonia Mine is significant today because it provides rare access to observe the character of native copper mineralization underground in three dimensions.

**Acknowledgements**

We thank Allan Blaske for helpful reviews which improved the quality of this field trip guide.

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# Field Trip 3

# Geology of Silver Mountain, Houghton County, Michigan

Evgeniy Kulakov

Department of Geological and Mining Engineering and Sciences, Michigan Technological University, 1400 Townsend Drive, Houghton, MI 49931.

*Directions: Leave downtown Houghton and head east on Montezuma avenue and College Avenue. Continue on US 41south for 27 miles. In Baraga MI turn right and continue to follow M38 west for 13.5 miles. Make left turn on S Laird road and follow it for 5.7 miles. Take a left turn onto Forest Rd 2276 and continue 1.8 miles. Turn right onto Forest Rd 193/NF-2270. In 0.9 miles make a right turn onto Forest Rd 922 road. Arrive in 0.4 miles. [UTM 5171550N 367079 E (NAD27 CONUS)].*

## Introduction



Figure 1: View from the top of Silver Mountain looking east.

Silver Mountain is located at the south-eastern part of Keweenaw Peninsula, near Sturgeon River Falls, Houghton County, Michigan (sections 1 and 12 of T 49 N, R 36 W, and section 6 of T 49N, R 35 W). A former US Forest Service lookout tower, Silver Mountain is now designated as a scenic view. Silver Mountain is a 100 meter high glacially polished dome-shaped hill with a 360° view (Fig.1) of the surroundings. The low lying areas around Silver Mountain consist of unconsolidated glacial deposits unconformably overlying Mesoproterozoic Jacobsville Sandstone. Fourteen shallow-dipping tholeiitic lava flows are exposed on the top and sides of the mountain.

Silver Mountain was first geologically mentioned by Burt (1849) who reported its location and small amounts of copper sulfides associated with calcite. Foster and Whitney (1851) described Silver Mountain as basaltic flows made up of labradorite and hornblende with scattered nodules of quartz and chalcedony. They interpreted the mountain as igneous rocks that intruded the surrounding sedimentary strata. In the mid 1850s, the National Company drove an approximately 30 m long adit. However, metallic mineralization was insufficient to justify additional mining. Lane (1909) concluded that the rocks of Silver Mountain were typical Keweenawan lava flows.

**Regional Geology**

The Midcontinent Rift (MCR) extends from NE Kansas northward to Lake Superior and through Michigan (Cannon et al., 1989). Rifting began ~ 1.1 Ga during an interval of reversed polarity of geomagnetic field with the oldest erupted material being reversely magnetized. These oldest lava flows include the Siemens Creek Formation of the Powder Mill Group, the lowermost part of North Shore Volcanics, Osler Volcanics, and the lower part of Mamainse Point Formation, located around Lake Superior (Paces and Miller, 1993; Davis and Green, 1997) (Fig.2 and 3). This early stage of magmatism occurred from 1109 to approximately 1105 Ma (Heaman et al., 2007; Paces and Miller, 1993) and was followed by a quiescence period when the geomagnetic field reversed to normal polarity (Davis and Green, 1997).

Magmatism resumed by 1102 Ma (Paces and Miller, 1993) during the normal polarity interval. During this interval, the main stage of the rift-related magmatism was represented by a sequence of approximately 200 lava flows of the Portage Lake Volcanics that erupted within a short interval around 1095 Ma (Davis and Paces, 1990). Magmatism of the Portage Lake Volcanics ended and the rift basin was filled with clastic sedimentary rocks during continued sagging (Bornhorst and Lankton, 2009).

The final phase of the MCR was continental compression at about 1060 Ma related to the Grenville Orogeny which inverted original rift-bounding graben normal faults into high angle reverse faults (Cannon, 1994). During late rift compression, rift-wide burial metamorphic/hydrothermal fluids altered rift-filling rocks and formed the native copper deposits of the Keweenaw Peninsula native copper district (Bornhorst, 1997; Bornhorst and Barron, 2011).

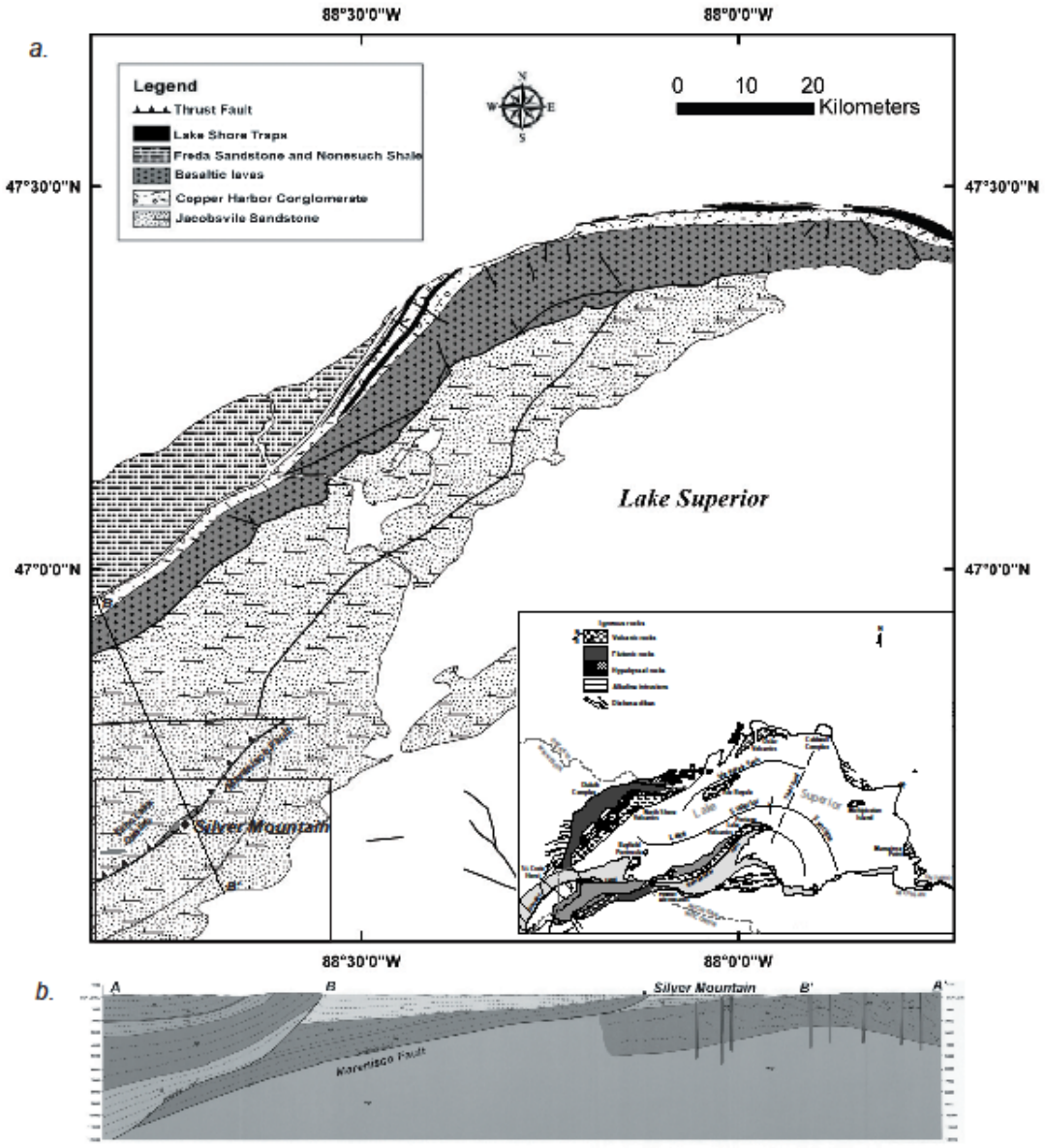



Figure 2**:** a. Generalized geologic map of Keweenaw Peninsula. The inset shows the geology of the Lake Superior segment of the Midcontinent Rift. The inset from Ojakangas et al, (2001); b, Section of Keweenaw Peninsula along the Line A-A' (B- B'part of the section is present on the map). Section is taken and modified from Cannon and Nicholson (2001).

Fourteen tholeiitic lava flows have been recognized at Silver Mountain and that dip NE at about 15$^{o}$. The lava flows of Silver Mountain are characterized by moderate to high magnetic anomalies and overall high positive gravity anomaly (Campbell, 1952). The magnetic and gravity anomalies are similar to those attributable to the Siemens Creek Formation. Based on this geophysical data, the lava flows at Silver Mountains were interpreted as being Keweenawan in age and part of the South Range Traps (termed Siemens Creek Formation today) (Campbell, 1952). However, Silver Mountain is an isolated knob and there is no direct geological contact with the rest of the Powder Mill Group. The mountain is located in the immediate vicinity of the Marenisco fault (Fig.2). Reverse and thrust faulting occurred during the latest compressional stage of the MCR evolution and resulted in uplifting the deeply buried strata to the surface. At Silver Mountain, the Marenisco fault has uplifted the lava flows stratigraphically older than the Portage Lake Volcanics which are exposed in the Keweenaw Peninsula because of uplifting along the Keweenaw fault. The Jacobsville Sandstone surrounds the uplifted basalt flows at Silver Mountain and vicinity. The Jacobsville Sandstone was deposited in a rift-flanking basin that while initiated by, and contemporaneous with compression, its deposition continued after compression. About 10 km west-southwest of Silver Mountain, there are gravity and magnetic anomalies resulting from the Echo Lake Gabbro which was also uplifted during compression along another fault. This geophysical anomaly was drilled in 1994 and confirmed as a layered intrusion (Waggoner, 1994). Additional drilling by Bitterroot Resources identified a 5.5 m thick interval containing 0.5 to1 ppm total Pt + Pd + Au (Cannon and Nicholson, 2001).

There is at least two faults cross-cutting Silver Mountain (Roberts, 1940). One fault can be observed at the adit at the base of Silver Mountain. It strikes N85°E and dips approximately 60° N (Roberts, 1940). The other fault is exposed near the NE side of Silver Mountain. This fault strikes N45°W and dips at about 80° N.

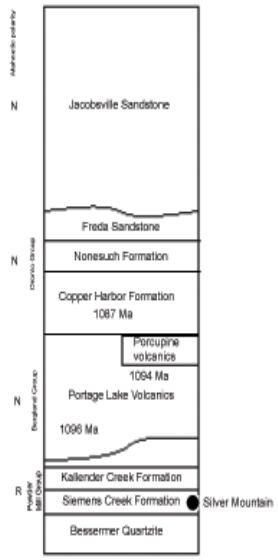



Figure 3: Stratigraphic column of Midcontinent Rift rocks in the western Upper Peninsula of Michigan. The available radiometric ages shown are from Davis and Paces, (1990). ”R” and “N” indicate reversed and normal polarity of remnant magnetization, respectively.

The massive interiors of the lava flows are fine-grained with intergranular texture. The predominant rock forming mineral is plagioclase with laths up to about 2 mm in length with an aspect ratio of around 10:1 . Typically, more equant altered mafic minerals, less than 1 mm across, and opaques, less than 0.2 mm across, fit between the plagioclase laths. In most massive interiors, the mafic minerals are completely pseudomorphically replaced by chlorite, however, in some interiors; patches of original pyroxene have survived a combination of burial metamorphic and hydrothermal alteration. Some plagioclase has altered to sericite. The plagioclase laths and space between them contain irregular patches of calcite. The abundance of calcite is near zero in some flow interiors and much greater in others, but always less than a few percent.

Amygdules are filled with quartz, calcite, chlorite, adularia, sericite, hematite, bornite, and chalcopyrite. Small amounts of copper sulfides are particularly noted in flow tops cropping out at the top of the mountain. The nonmetallic minerals are similar to those found throughout the Keweenaw Peninsula (Butler and Burbank, 1929). However, the occurrence of the copper sulfides chalcopyrite and bornite in amygdules is very uncommon in the Keweenaw Peninsula. Copper sulfides are reported by Robertson (1975) in the tops of lava flows at Mount Bohemia near the tip of the peninsula.

**New Paleomagnetism Data**

Rocks of the MCR are probably among the worlds most extensively studied by paleomagnetic methods (Halls and Pesonen, 1982). Reversed polarity of natural remanence was recently reported in the flows of Silver Mountain (Kulakov et al, 2012). Well-defined characteristic remanent magnetization in samples from 13 flows revealed a paleomagnetic mean direction typical for reversely magnetized Keweenawan rocks (Fig. 4) (Kulakov et al., 2012). The paleomagnetic direction was similar to that found in the Lower North Shore Volcanics that have been dated at 1107.9±0.8 Ma (Davis and Green, 1997) and the Powder Mill Group (Halls and Pesonen, 1982) dated at 1107.3±1.6 Ma (Davis and Green, 1997). Thus, the likely age of the Silver Mountain basalts is 1107 to 1108 Ma and the same as the Siemens Creek Formation, Powder Mill Group.

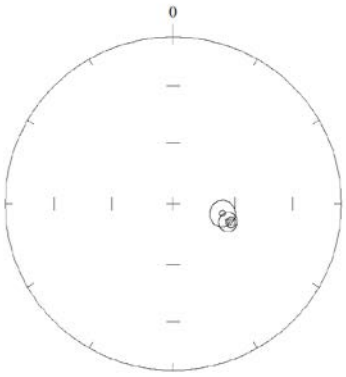



Figure 4: Equal area projection showing the mean paleomagnetic directions for selected reversely magnetized rocks from the MCR. Open square - Silver Mountain (N=13) (Kulakov et al, 2012.); open triangle – Powder Mill Group (N=9) (Palmer and Halls, 1986); open circle – North Shore volcanics (N=21) (Halls and Pesonen, 1982).

## Geochemistry

Major and trace element geochemical analysis were conducted on samples from seven flows (Fig 5.) These samples were characterized by very uniform composition for both major and trace elements. The major and trace element composition of the Silver Mountain flows are very similar to that reported for the Upper Siemens Creek formation and equivalent rocks of the lowermost part of the North Shore Volcanics and Osler Volcanics. These youngest rift-related flows belong to basalt type II of Nicholson et al, (1997). The compositional similarity of the flows of Silver Mountain to this group of basalts, and in particular to the Upper Siemens Creek rocks further confirms the close relationship of the Silver Mountain lava flows to the Powder Mill Group. Nicholson et al. (1997) concluded that these basalts were derived from a mantle plume, but were contaminated by continental lithospheric crust.

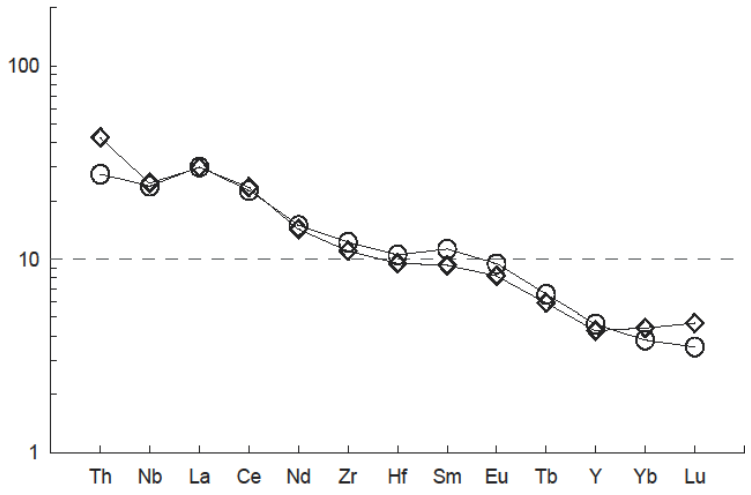



Figure 5: Primitive mantle normalized plot comparing the average trace element composition of rocks of Silver Mountain (N=7) (open squares) and Basalt type II from the Upper Siemens Creek Formation (N=18) (open circle). Siemens Creek Formation data from Nicholson et al. (1997) and Silver Mountain data from Kulakov et al. (2012)

## Field Trip Stop

This field trip provides the opportunity to observe the lava flows at Silver Mountain and to enjoy a scenic view from the top. The trip begins from the parking area at the bottom of Silver Mountain with a short hike up a trail consisting of a combination of rocky terrain and built in stairs to the top.

The Silver Mountain adit is located near the parking area with the 1850s poor rock pile scattered near the entrance. It mainly consists of amygdaloidal basalt from the oldest exposed flow at Silver Mountain. The adit exists because the top flow here was more mineralized adjacent to a fault striking parallel to the adit. Fragments of fault breccia can be observed.

After viewing the mineralized rocks from the adit, the trip will proceed laterally to view the relatively thick lava flows at the base of Silver Mountain. The trip will then continue with a climb to the top of the mountain and involves strenuous physical exertion. Along the way there will be an opportunity to observe the character of the massive interiors and amygdaloidal flow tops of several flows with different thicknesses. The thickest ( ~ 6 m) lava flow at Silver Mountain crops out approximately half way to the top. Towards the top, the thickness of the flows tends to decrease.

At the top, there is an excellent scenic view of the surrounding area. The shallow-dipping lava flows roughly parallel the gentle sloping topography of the top of Silver Mountain making it more difficult to observe contacts between the cross sectional views of the lava flows. The highest point is on the updip side (southwestern side). Proceeding northeast from the top and slightly down the steep side one can observe the amygdaloidal top and underlying massive interior of a flow. The flow top is particularly notable because the amygdules are filled with copper sulfides (chalcopyrite) and calcite.

The trip to the top is not recommended in stormy or wet weather. The glacially polished rocks surfaces at the top can be quite slippery when wet.

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# Field Trip 6

## Geology and Environmental Site Conditions of the Copperwood Deposit, Gogebic County, Michigan

Theodore J. Bornhorst
*A.E. Seaman Mineral Museum, Michigan Technological University, 1404 E. Sharon Avenue, Houghton, MI 49931*

Allan Blaske
*AECOM, 401 S. Washington Square, Suite 100, Lansing, Michigan, 48933*

Dave Anderson and Thomas D. Repaal
*Orvana Resources US Corp., 10199 Lake Road, Ironwood, Michigan 49938*






## Introduction

The Copperwood deposit is a stratiform copper deposit in Gogebic County, Upper Peninsula, Michigan that is hosted by gray to black shales and siltstones filling the Midontinent rift (MCR) (Fig. 1). Copperwood is a reduced-facies or Kuperschiefer-type sedimentary rock-hosted stratiform copper deposit (Bornhorst and Williams (in press)). Mineralization at Copperwood contains 33.2 million short tons of Canadian National Instrument 43-101 compliant Measured and Indicated resources with an average grade of 1.65 % Cu and 4.34 ppm Ag, There are 3.0 million short tons of inferred resources with an average grade of 1.07 % Cu and 2.01 ppm Ag (Ward 2011). The resources are based on a cutoff composite grade of 0.8 % Cu and thickness of 5 ft (1.5 m).

The Porcupine Mountains sedimentary rock-hosted copper district (Bornhorst and Barron, 2011) encompasses the White Pine Mine and the Copperwood deposit (Fig. 1). The White Pine Mine produced approximately 4.5 billion lbs of Cu and 50 million ounces of Ag from 1953 to 1996 with a few interruptions in production. Copperwood was discovered in 1956 (a subsidiary of AMAX) after a USGS publication in Economic Geology (White and Wright, 1954) indicated potential for copper mineralization in the Western Syncline. In addition to Copperwood, the U. S. Metals and Refining Company exploration program discovered 3 lower grade mineralized areas within the Western Syncline with indicated and inferred resources of 1,348 million short tons with an average grade of 1.34 % Cu (Kulla and Thomas, 2011). During1957 to 1958, a 71 m vertical shaft, 635 m of drifting, and 3 small stopes were completed in the higher grade Copperwood deposit (Bornhorst and Williams, in press). A mine was not developed because of presumed ground stability issues that would force excess dilution during mining. Advances in mining technology combined with higher Cu prices make Copperwood an economic deposit today. Orvana Minerals Corp began exploration and environmental baseline studies at Copperwood in 2008. This was quickly followed by the first Canadian National Instrument 43-101 compliant mineral resource reported in 2010 (Kulla and Parker, 2010), prefeasibility in 2011, and feasibility in 2012 (Keane et al., 2012). Copperwood was granted a mining permit by the State of Michigan in 2012 and in February 2013, the Michigan Department of Environmental Quality granted the wetlands permit which is the last major permit needed before construction and production can proceed. Production is projected to begin in the near future.

The descriptions in this field guide are based on Orvana (2011) and Bornhorst and Williams (in press). This field guide is published with permission granted by Orvana Minerals Corporation; however, the content is the sole responsibility of the authors.

### Regional Geologic Setting

The broad 300 km wide MCR in Michigan consists of more than 25 km thick succession of rift-filling tholeiitic flood basalts with minor interbedded red conglomerates and sandstones overlain by 8 km thick succession of rift-filling clastic sedimentary rocks (Cannon, 1993). A rift-flanking basin is filled with 3 km of sandstone. These rocks are make up the Keweenawan Supergroup in Michigan and were deposited from 1.15 to about 1.03 Ga (Fig. 2) (Heaman et al., 2007; Davis and Paces, 1990; Cannon et al., 1989).

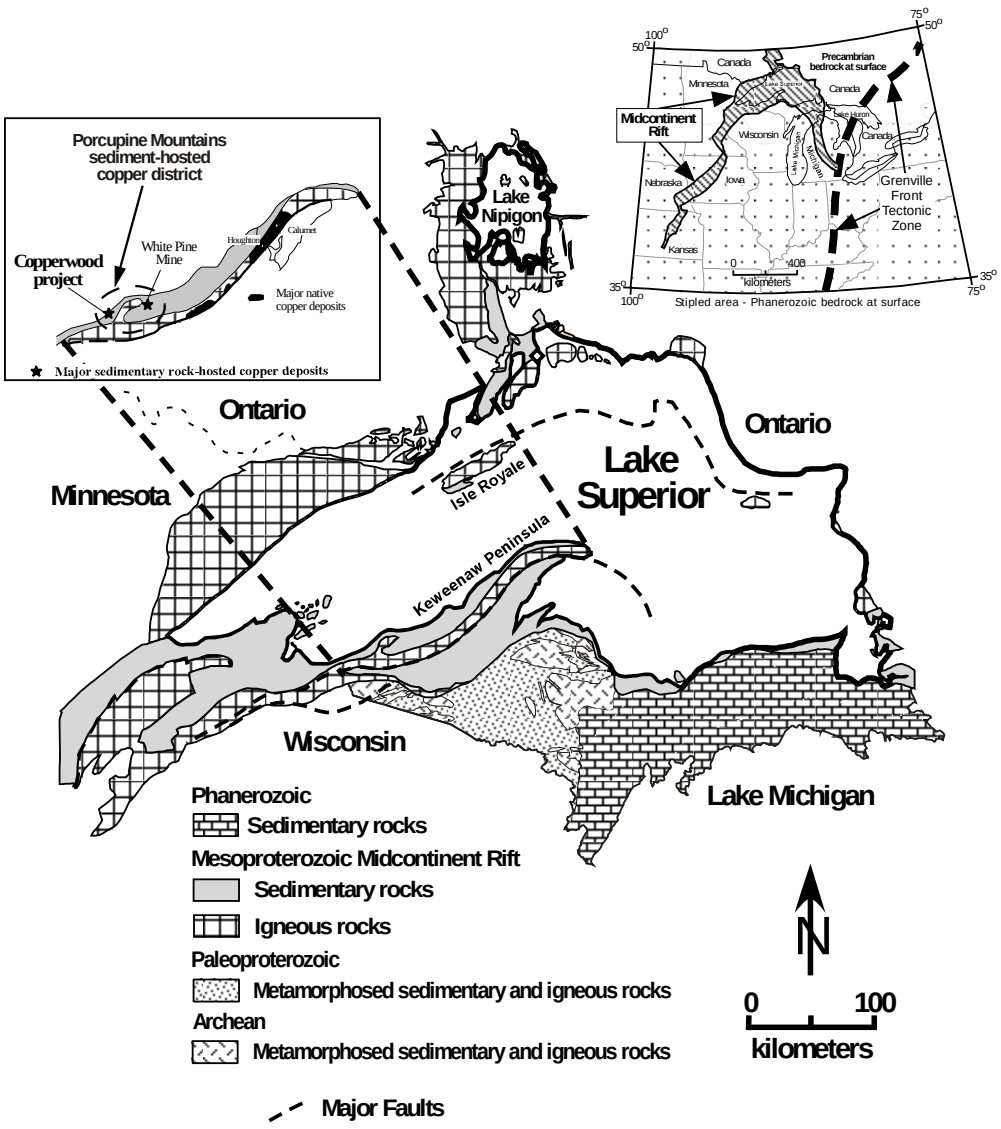



Figure 1: Generalized bedrock geologic map of the Midcontinent Rift. Modified from Bornhorst and Barron (2011).

The bedrock at Copperwood is part of the rift-filling clastic sedimentary rocks. Continental compression occurred at 1.06 Ga in response to the Grenvillian collision along the eastern edge of the North American continent inverting the original graben-bounding faults into reverse thrust faults (Fig. 2) (Cannon, 1994). A syncline at Copperwood is a result of this compressional event.

Erosion followed continental compression from about 1.03 Ga to 0.5 Ga (500 Ma) during which time multiple km of bedrock was removed and likely exposing the Copperwood orebody at the bedrock. This would have allowed a period of downward percolating groundwater into the Copperwood deposit (Bornhorst and Robinson, 2004). Marine submergence beginning 500 Ma buried the Precambrian bedrock under multiple km of Phanerozoic sedimentary rocks; evidence of Phanerozoic rocks is missing at Copperwood.

Pleistocene glaciation over the last 2 million years removed the Phanerozoic rocks overlying Copperwood and once again exposed the orebody at the surface. The retreat of the last glaciers about 10,000 years ago left behind unconsolidated glacial deposits that today overlie the Precambrian bedrock at Copperwood.

Peterson (1985, 1986) determined that four distinct advances of glacial ice occurred in the western Upper Peninsula during the late Wisconsinan time (14,500 to 10,200 years B. P.). Two of these advances and retreats produced the current surficial features at Copperwood. The first advance moved out of the Lake Superior basin south to the Wisconsin-Michigan border and upon retreat, left behind the lower till. The youngest advance occurred approximately 10,200 years B.P., when the glacier overrode previous till deposits at Copperwood. The approximate southern limit of this last advance was approximately 3 km south of Copperwood. Following this final retreat, glacial Lake Duluth covered the Copperwood area. Several Lake Duluth shorelines are evident along the topography at Copperwood.

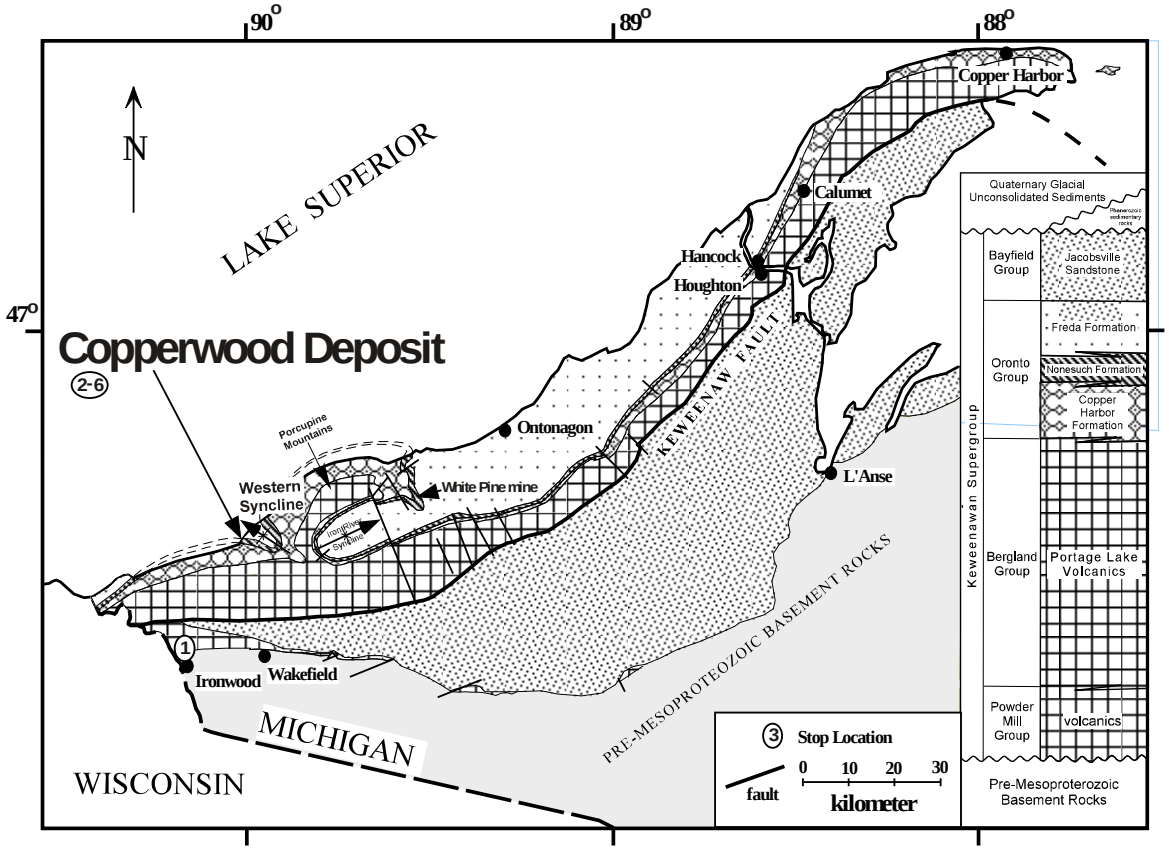



Figure 2: Bedrock geologic map and lithostratigraphic column for the western part of the Upper Peninsula of Michigan. The location of the field trip stops are shown.

## Mesoproterozoic Bedrock Geology

The Copperwood deposit is on the southwest limb of the open shallow-plunging Western Syncline (Fig. 3). The bedrock at Copperwood consists of clastic sedimentary rocks of the Mesoproterozoic Oronto Group (Fig. 2) that strike approximately east-west and dip approximately 10 degrees to the north. The MCR bedrock at Copperwood is overlain by unconsolidated Pleistocene glacial sediments.

The lowermost lithostratigraphic layer at Copperwood is the Copper Harbor Formation that consists of primarily of reddish-sandstone. The Nonesuch Formation overlies and interfingers with the Copper Harbor Formation. It consists of gray-black shale and siltstone to gray-white siltstone to brownish-red siltstone that are subdivided into multiple informal subunits. The Freda Formation gradationally overlies the Nonesuch Formation and consists of brown siltstone at Copperwood.

*Lithostratigraphy*

Copper Harbor Formation. Overall, the Copper Harbor Formation is composed of red-brown conglomerates and sandstones with lesser siltstones in an upward- and basinward-fining sequence. In Michigan, there is a maximum exposed thickness of about 2,000 m (Elmore 1984). The Copper Harbor Formation sedimentary rocks are fluvial and deposited a coalescing alluvial fan environment.

At Copperwood, the Copper Harbor Formation is the oldest lithostratigraphic bedrock formation. It is lithologically dominated by red-brown to white and gray fine- to coarse-grained, arkosic sandstone. The Copper Harbor Formation in one drill hole consisted of 140 m of sandstone, a 1 m thick red, matrix-supported conglomerate, and more sandstone. Outcrops of the Copper Harbor Formation along the southern portions of Namebinag Creek and an unnamed creek indicate an increase in conglomerate facies in the lower portions of the formation. Outcrops along these streams are predominantly conglomerates with lesser amounts of sandstones.

The uppermost few feet of the Copper Harbor Formation intersects in all of the Orvana and legacy exploration drilling at Copperwood. The uppermost Copper Harbor Formation consists of interlaminated red-brown siltstones and shales with occasional beds of very fine-grained sandstones. Uncommonly, there are interbedded, thin beds of dark-gray shales and siltstones less than 1.5 cm below the upper contact indicative of interfingering overlying Nonesuch lithologies. Absent in some holes, is the red-brown siltstone at the top of the Copper Harbor Formation. It can be up to 1 m thick, but is typically less than 30 cm in thickness. Assay data has shown, the uppermost 1 m of the Copper Harbor Formation (red-brown siltstone and sandstone) does not carry significant amounts of copper. There is a dramatic and abrupt change from the reddish Copper Harbor Formation to dark-gray to black shales and siltstones of the overlying Nonesuch Formation. At Copperwood, this abrupt transition defines the change from the Copper Harbor Formation to the Nonesuch Formation.

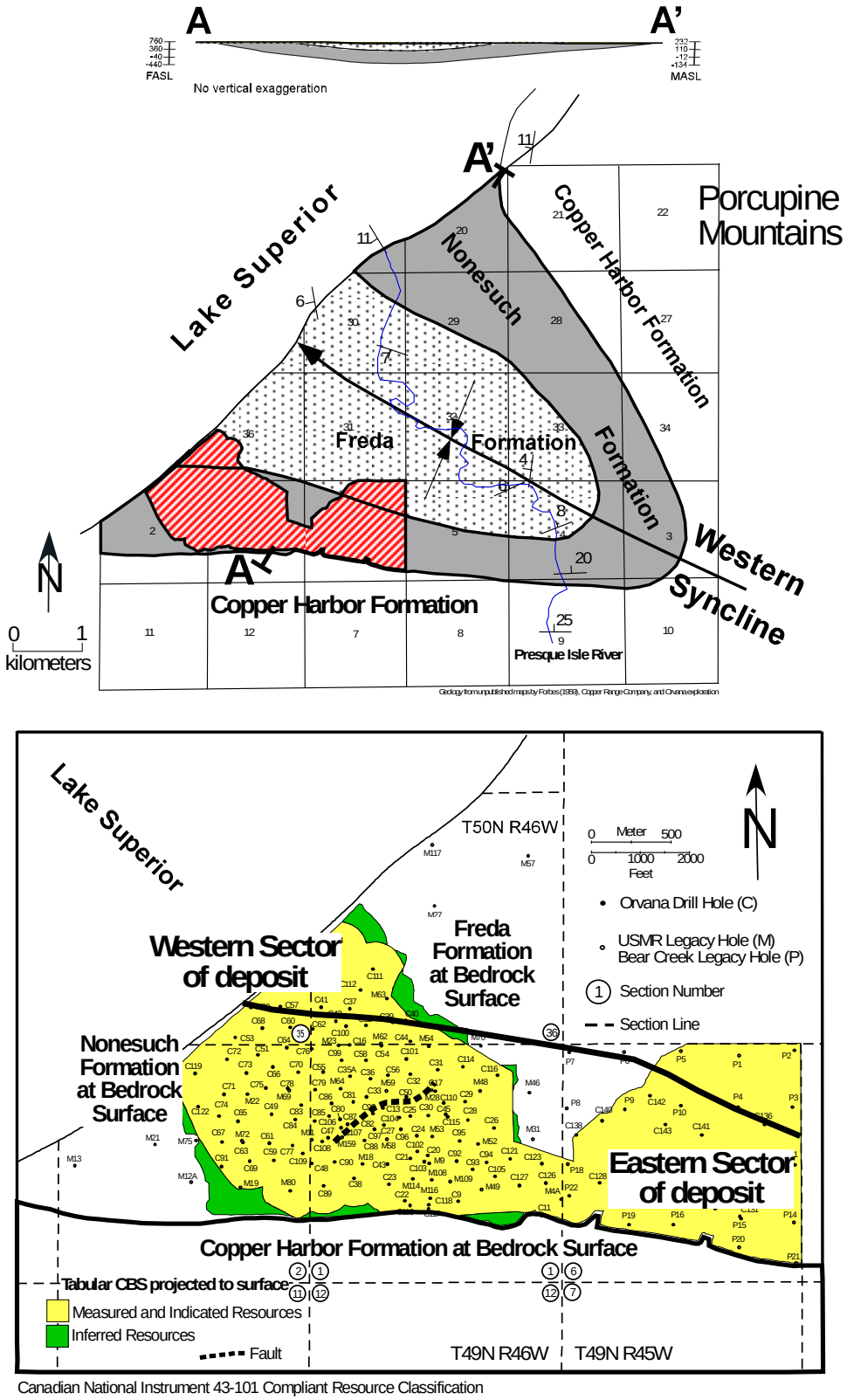



Figure 3: Geologic map of the Western Syncline and Copperwood deposit. Modified from Bornhorst and Williams (in press).

Nonesuch Formation. Overall, the Nonesuch Formation is composed of characteristically black-to-gray -green siltstones, shales, carbonate laminates, and minor sandstones with a maximum thickness of 215 m. Elmore et al. (1989) and Suszek (1997) interpreted the depositional environment of the Nonesuch Formation to be dominantly anoxic lacustrine ranging from marginal lacustrine (sandflat-mudflat) to lacustrine to lacustrine-to-fluvial subenvironments.

At Copperwood, a completed stratigraphic section is exposed in the northeast part of the property, at a thickness of 200 to 215 m. The upper contact is missing due to erosion throughout most of Copperwood property. The formation has been subdivided into multiple informal members on the basis of lithologic variations (Fig. 4). All of these members have remarkable lateral continuity throughout the Copperwood area (Fig. 5).

The Parting Shale member is at the base of the Nonesuch Formation and is further subdivided into units (Fig. 4). The three lower units of the Parting Shale are the host to copper mineralization at Copperwood and together are termed the Copper-Bearing Sequence (CBS).

The lowermost unit of the Parting Shale and CBS is termed Domino after terminology used at the White Pine Mine. Domino averages 1.6 m thick in the western sector and thins to about 60 cm thick in the eastern sector. The overall average thickness is 90 cm with a range from 9 cm to 2.3 m. Domino is characterized by laminated dark-gray to black shales and siltstones. Domino hosts the highest-grade copper at Copperwood. The contact between Domino and the overlying Red Massive unit is sharp and easily recognized in drill core as an abrupt change from the dark gray/black (Domino) to red-brown (Red Massive).

The Red Massive unit of the Parting Shale and medial unit of CBS averages 35 cm thick, ranges in thickness from near zero to 1.2 m, but is usually less than 50 cm thick. It is somewhat thicker in the eastern sector than in the western sector. Red Massive is characterized by massive dark red-brown siltstones with interbedded red-brown, fine-grained sandstones. The contact between Red Massive and the overlying Gray Laminated unit is gradational and is placed where the color changes from reddish gray to gray.

The Gray Laminated unit of the Parting Shale and upper unit of CBS averages 1.1 m thick and ranges in thickness from 50 cm to 4 m. Gray Laminated is characterized by of light-to-medium, gray-to-reddish-gray laminated siltstones; some intervals are massive. The contact between Gray Laminated and the overlying Red Laminated is gradational and is placed where the color changes from gray-dominated to mixed maroon and gray.

The Red Laminated unit of the Parting Shale and hanging wall of the CBS ranges in thickness from 10 cm to 3.4 m and is more typically 1.2 to 1.8 m thick. Red Laminated is characterized by laminated siltstones with bimodal color distribution of maroon/red-brown and gray. Typical Red Laminated has mottled or wavy maroon intervals interspersed with medium-gray to reddish-gray siltstones. The contact between Red Laminated and the overlying Gray Siltstone is gradational.

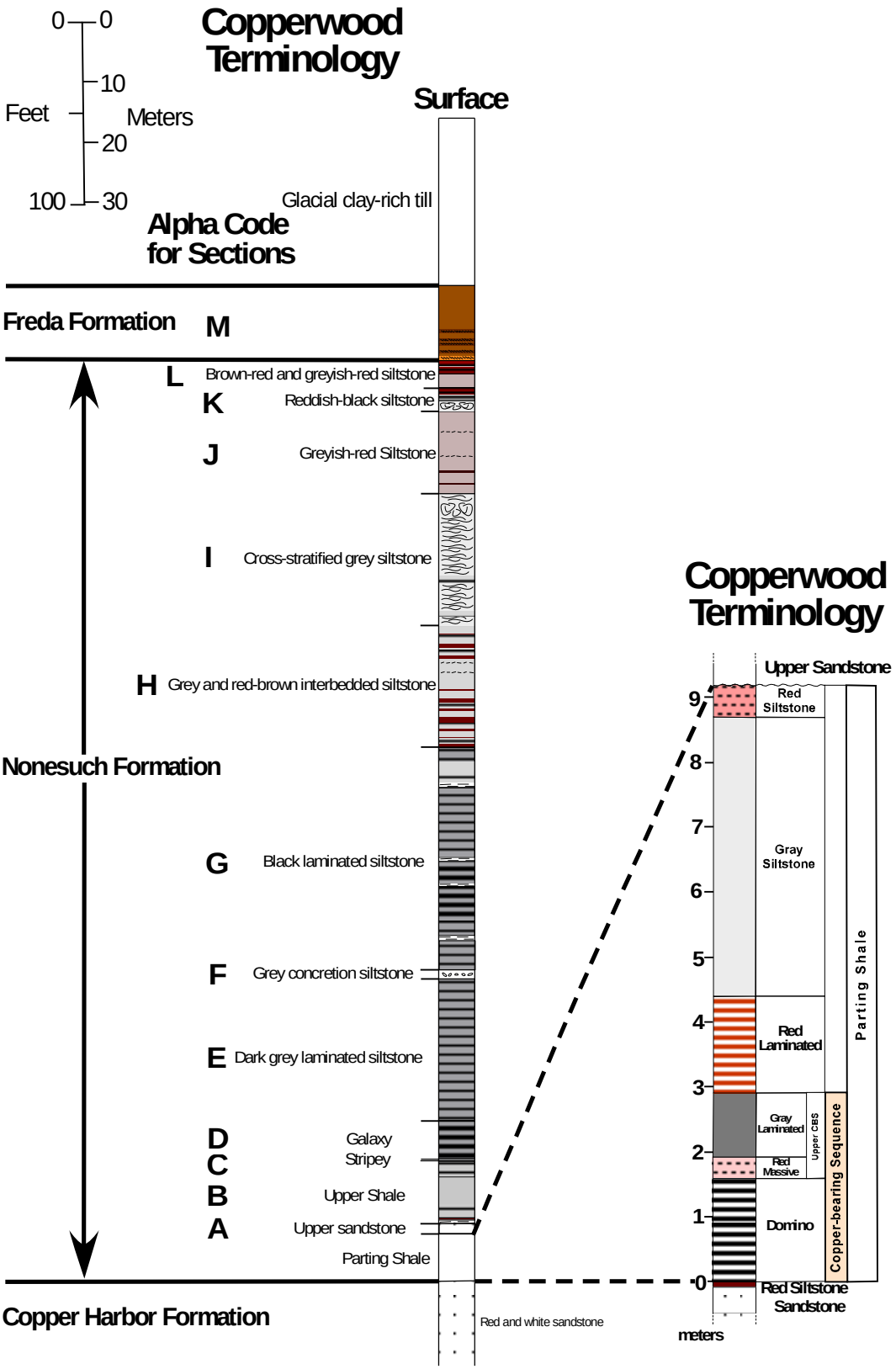



Figure 4: Lithostratigraphic column of Copperwood bedrock units with detail of Parting Shale member. Modified from Bornhorst and Williams (in press).

Freda Formation. The top of the Nonesuch Formation gradually transitions into the Freda Formation over an interval of about 10 m where beds of coarse, light-brown siltstones and massive to cross-bedded, dark reddish-brown siltstones are intercalated with grayish-red siltstones. The contact is placed at the base of a brown to white, cross-bedded siltstone. At Copperwood, the Freda Formation is up to 120 m thick above and consists of brown siltstone. Only the base of the formation occurs at Copperwood as the rest has been removed by erosion.

*Structure*

The structure at Copperwood is simple and consists of bedrock units that dip gently to the north on the southwest limb of the Western Syncline (Fig. 5). Under the unconsolidated glacial sediments, dips for all bedrock units vary from 12° in the south near the subcrop to 8° in the north nearer the synclinal axis. The bottom surface of the CBS approximates a gently curved, dipping plane lacking significant undulations.

One fault has been identified at Copperwood (Fig. 3). This fault is interpreted to be a shallow, north-dipping reverse fault with 3 to 7 m of vertical displacement. Minor fractures with less than 1 inch of displacement were observed in multiple holes and these fractures are typically healed by calcite.

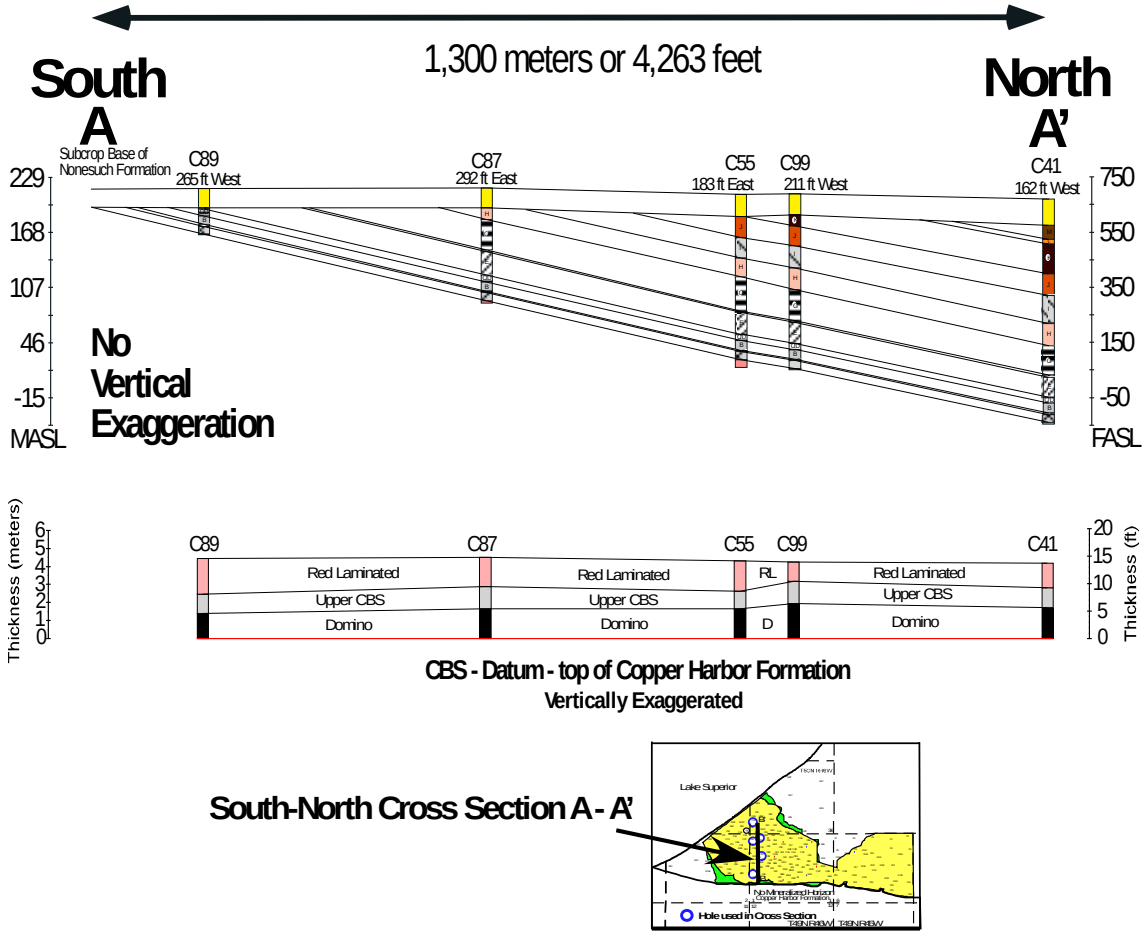



Figure 5: Geologic cross section of the Copperwood deposit. Modified from Bornhorst and Williams (in press).

**Copperwood Deposit**

At Copperwood, copper mineralization is hosted by gray to black shales and siltstones of the CBS (Fig. 4). The footwall consists of red-bed sandstones and minor siltstones of the Copper Harbor Formation and the hanging wall consists of maroon/red-brown to gray siltstones of the Red Laminated unit. Geologic cross sections using the base of the CBS as the horizontal datum demonstrate the high degree of lithologic continuity of the CBS. The copper orebody is a conformable and tabular with an average thickness of 2.5 m. In the western sector the CBS averages 2.9 m thick whereas in the eastern sector it averages 2.1 m thick. The Copperwood deposit contains a total (all categories) undiluted geologic resource of about 1.16 billion lbs of Cu and 4.4 million ounces of Ag.

The deposit (CBS) is characterized by copper in the form of chalcocite with minor amounts of silver. Other copper minerals such as chalcopyrite and bornite occur above the CBS (Bornhorst and Williams, in press). Pyrite is virtually nonexistent in the CBS, but above the Parting Shale, pyrite does occur in low abundance. The CBS is dominantly siltstones that are composed of over 90% silicate minerals (quartz, clinochlore, muscovite, illite, K-feldspar, plagioclase), about 2% calcite, and 3% hematite. Overall, pyrite and other minerals with the potential to generate acids are lacking whereas calcite, which is an acid-neutralizing mineral species, is abundant.

The White Pine Mine produced copper from almost the entire Parting Shale and from the overlying Upper Shale (Fig. 4) (Mauk, 1992; Ensign et al., 1968). Whereas the three layers that compose the CBS at Copperwood are lithologically similar to those at the White Pine Mine, their thicknesses and proportions are not. The total Parting Shale is much thicker at Copperwood and e.g., the Domino within the western sector is typically 2.5 times thicker than at White Pine. At the White Pine Mine, the ore minerals are chalcocite and native copper whereas Copperwood is devoid of native copper. At White Pine, the chalcocite and minor native copper is stratiform ore interpreted as being related to diagenesis (Brown, 1971; Ensign et al., 1968). The native copper represents a second stage of mineralization (Mauk et al., 1992) hosted in faults and fractures and is interpreted as being related to the native copper deposits of the Keweenaw Peninsula (Bornhorst, 1997; see Field Trip 1 this volume). Copperwood lacks the complexity of the White Pine deposit (Bornhorst and Williams, in press).

The White Pine copper deposit straddles a right-lateral strike-slip fault and an anticline (Ensign et al., 1968; Johnson et al., 1995). Thrust faults, strike-slip faults, and normal faults are encountered throughout White Pine and these faults and folds are mostly related to late rift compression. Some clearly compression-related thrust faults host sheets of native copper. The second-stage mineralizing fluids were likely of the same origin as those related to native copper in the Keweenaw Peninsula (Bornhorst, 1997).

The Copperwood deposit is an example of a reduced facies or Kupferschiefer-type sedimentary rock-hosted copper deposit (Bornhorst and Williams, in press). Bornhorst and Williams (in press) proposed that at Copperwood "chalcocite replacement of pyrite in unlithified sediments during diagenesis is a result of emanating upward-focused, compaction-driven, Cu-bearing saline basinal waters whose Cu was leached from the underlying red-bed paleoaquifer." In comparison to most examples of reduced facies or Kupferschiefer-type (Hitzman et al., 2010, 2005; Cox et al., 2003), Copperwood is notable for its simplicity (Bornhorst and Williams, in press).

## Pleistocene Unconsolidated Glacial Deposits

The unconsolidated deposits at Copperwood consist primarily of a reddish-brown glacial till. This glacial till unconformably overlies the bedrock and ranges in thickness from 0 (at outcrops along streambeds south of the subcrop of the Copperwood deposit) to 43 m; average thickness is approximately 25 m (Fig. 6). The top of the bedrock surface is generally smooth. Boulders or otherwise weathered bedrock were encountered in some borings at the bedrock surface, but at most locations the glacial till sits on a scoured bedrock surface. The surface of the bedrock is more or less parallel to the ground surface topography, and slopes toward the north-northwest.

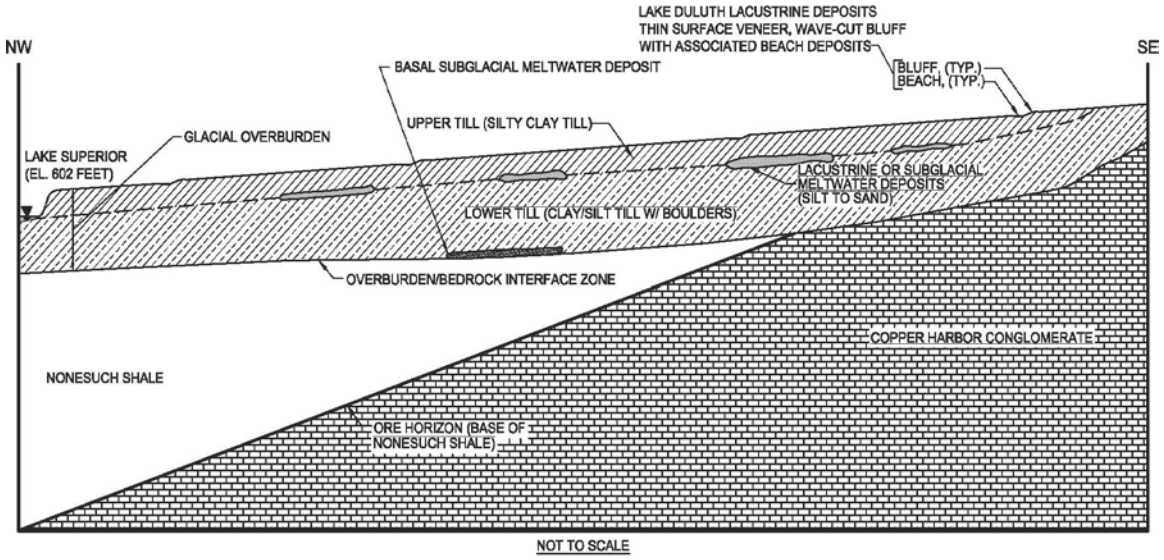



Figure 6: Schematic cross section through the Copperwood project with generalized geology.

The glacial till at Copperwood is a mud-matrix supported diamictite. This diamictite is massive with no stratification, lamination or fining upward or downward and the matrix is a uniform mix of sand, silt, and clay. Grain-size distribution curves are typical of those associated with subglacial tills (Fig. 7). The characteristics lead to the interpretation of the glacial till at Copperwood as a subglacial diamictite (Kemmis, 2008), meaning that it was deposited beneath the glacial ice. The diamictite is dense and overconsolidated, characteristic of subglacial till, but unlike normally consolidated to slightly overconsolidated lacustrine deposits.

The glacial till was described during the soil boring program as variations of a silty clay based on slight variations in the observed portions of silt, clay, and sand. Field classification ranged from silt, clay, silty clay, clayey silt, silty sand, and sandy silt. The till was found to contain trace (1 to 9%) to little (10 to 19%) to some (20 to 34%) amounts of sand and gravel. Soil samples tested for particle size distribution indicated that the average composition of the cohesive (silt/clay) was approximately 50% silt, 20% clay, and 30% sand (Fig. 8). Sample analysis of the till fine fraction (<200 sieve size) by X-ray diffraction indicate that quartz was the major component,

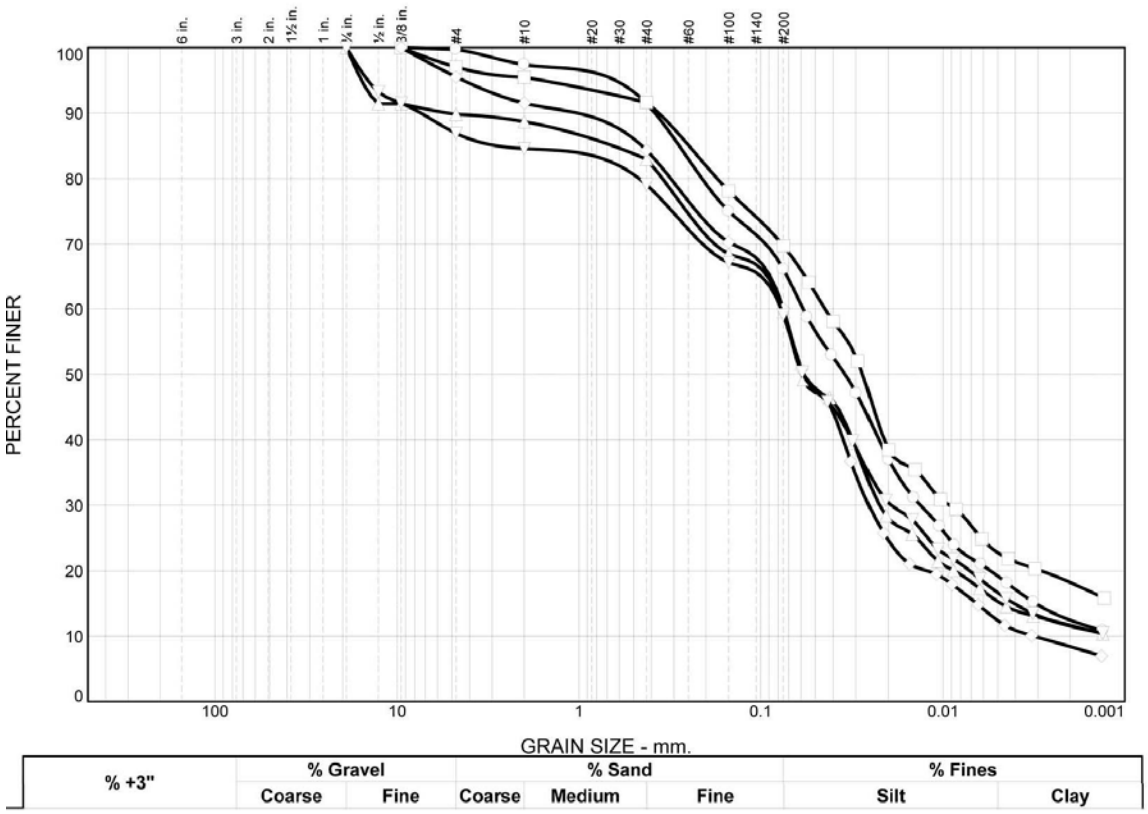



Figure 7: Grain size distribution of glacial till samples from the Copperwood site.

with very little clay minerals (kaolinite, illite, montmorillonite, etc.) present. Analyses also indicated the presence of small amounts of feldspar, micas, calcite, and hematite. Testing using dilute hydrochloric acid indicated the presence of very finely-ground calcite, volumetrically too low to be detected by X-ray diffraction. The gravel/cobble portion was difficult to estimate from the laboratory analyses, due to the mass of the samples subjected to sieve analysis. Some samples contained no gravel-sized fraction, while others contain up to 45% (by dry weight) of gravel (coarse and fine). Inspection of Rotosonic soil cores revealed as much as 10 to 20% coarse fraction (% gravel by volume) in the recovered soil core. Gravel, cobbles and boulders larger than 9 cm (diameter of drill sampling device) were encountered during the drilling as well as during site reconnaissance activities. Large boulders (up to 1 m in diameter) were present within the bluffs along the Lake Superior shoreline. The gravel portion of the till consisted of dark reddish-brown sandstone (between 60% and 70%), diabase (between 8% and 18%), granite/gneiss (between 7% and 13%), basalt/amygdaloid (between 1% and 5%), and other types of rocks (between 2% and 8%). The sandstone and basalt are likely to have been locally derived from the Keweenawan Supergroup bedrock. Beach stones on the Lake Superior shoreline represent a material washed from the entire thickness of the till outcrops along the lakeshore. The predominance of red sandstone (likely Freda Formation) in the till accounts for the overall red-brown color of the glacial material.

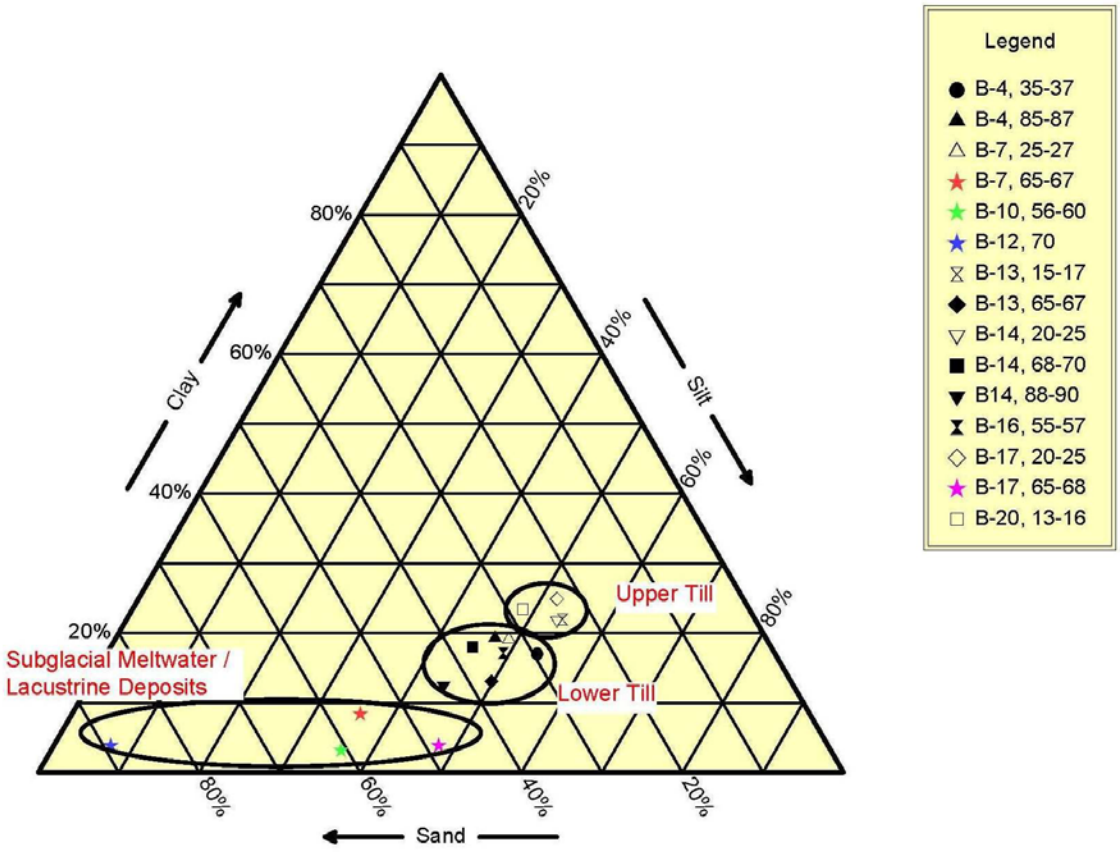



Figure 8: Soil texture plot of glacial deposits from the Copperwood site.

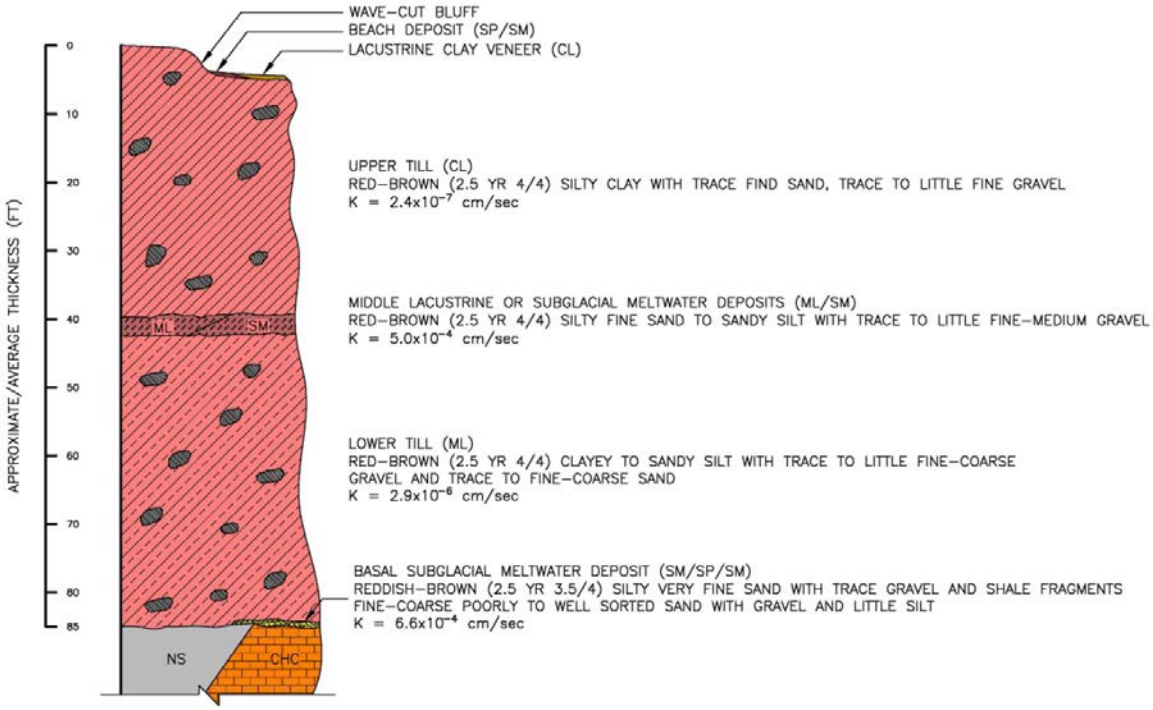



Figure 9: Stratigraphic section of the glacial overburden deposits at the Copperwood site.

The glacial till can be divided into two units (upper till and lower till), based on matrix grain size, amount and size of gravel, and vertical distribution (Fig. 9). In addition, there are thin (< 3 m with an average thickness of < 1 m) and isolated layers of coarser (non-cohesive) sediments throughout Copperwood in various borings that are located predominantly between the upper and lower till units. These generally consisted of fine to medium sand with varying amounts of silt and clay. These granular deposits, when encountered, are not laterally extensive, and for the most part cannot be correlated between adjacent borings. They are interpreted as lacustrine, intra-till sands or subglacial melt water deposits. Additional granular deposits were found in three borings at the base of the overburden, on the top of the bedrock surface.

Peterson (1985) described the glacial deposits of the area as thin (< 10 m) drift over bedrock. Hack (1965) described the Ontonagon Plain (located on the east side of the Porcupine Mountains) to be underlain by reddish-brown glacial lake sediments and till. He described three units within the glacial deposits – the lower, intermediate, and upper units. The lower unit is described as a stony till containing locally derived subangular boulders and fragments. The intermediate unit is described as till and laminated silt and clay in distinct layers that are believed to be lacustrine sediments. This unit is less stony than the lower unit. The upper unit is described as a clayey till that is much less stony than either of the lower units. Thin lacustrine deposits related to glacial Lake Duluth are described as a patchy thin (< 60 cm) overlying veneer of strongly-laminated clay, silt and sand of variable thickness. The three primary glacial units described by Hack (1965) are present at Copperwood: the lower till which is slightly coarser grained with more (and larger) clasts than the upper till, the intermediate unit, less well defined at Copperwood, but composed of the laterally inconsistent layers of silty and sandy sediments, and the upper till. A thin (< 1 m thick) of lacustrine deposits related to Lake Duluth exists at Copperwood. The glacial deposits at Copperwood are composed predominantly of subglacially deposited till that was streamlined by glacial movement into elongated drumlins and flutes. While the flutes or ridges are not obvious within at Copperwood, such features may have been the cause of the distinct modern Copperwood drainage pattern.

## Objectives of Field Trip

This field trip is designed to provide a geologic overview of Copperwood deposit hosted by the Mesoproterozoic MCR bedrock through descriptions and observations of drill core. The environmental site conditions at Copperwood will be observed and discussed onsite, especially those associated with surface and groundwater. The unconsolidated Pleistocene glacial deposits that unconformably overly the deposit will be observed as they play an important role in the environmental site conditions. The environmental site conditions are a critical aspect of permitting a modern mine. The location of Field Trip stops depend on site activities and accessibility. Those described below are based on full access and moderately dry conditions. Participants will be provided a map at the time of the field trip.

***Access to the Copperwood site is strictly forbidden without express permission from Orvana Minerals US Corp. No samples of any kind are allowed to be removed from the Copperwood site during this field trip.***

## STOP 1: Orvana Offices, Ironwood, Michigan

*The first stop of the field trip will be to the offices of Orvana Resources U. S. Corp. in Ironwood. An overview of the Copperwood geology and project will be provided. Core of the bedrock and unconsolidated glacial deposits will be available for inspection and discussion.*

Core drilling of the Copperwood deposit will be available for inspection. Since there is little lateral variation within the Copperwood orebody, core from only a few drill holes are necessary to observe the character of the orebody as a whole. The unconsolidated glacial deposits were sampled using Rotosonic drilling techniques. Core samples of the glacial till will be available for inspection. Slight differences in the matrix composition and gravel content can be observed in the various core samples.

## STOP 2: Copperwood Historic Test Mine Rock Pile and Weather Station

*The Copperwood historic mine rock pile is a result of testing mining in 1957 to 1958. This rock pile was once much larger, as prior to Orvana's activities at Copperwood, this rock was used as fill in wet areas of roadways and along stream crossings. The rock pile was used by Orvana as a staging area for exploration and pre-development activities and for an environmentally focused rock pile study. The rock pile will be removed upon creation of the tailings disposal facility.*

The rocks in this 1957-58 rock pile are dominated by unprocessed ore (CBS), but include hanging wall and footwall rocks as well. None of the ore was processed during this 1957-58 test mining, except for bulk samples delivered to Michigan Tech (then known as Michigan College of Mining and Technology) for process testing. The rock pile has been subjected to slightly more than 50 years of weathering which continues today. There is no evidence of acid drainage from the rock pile. Blocks of black shale, likely Domino, which contains the most copper within the CBS, are present on the surface of the rock pile and can be identified by notable green surface coloration. The green mineral has been identified by X-ray diffraction as malachite (copper carbonate). These blocks readily separate into smaller fragments along bedding planes

and, upon separation, the malachite is only visible for less than 2 cm from the edge of the block even when the bedding planes are visibly moist.

An environmentally focused study was initiated by Orvana to validate bench scale laboratory determined rates of release for chemical constituents and evaluate the long-term environmental impact of weathering of Copperwood unprocessed ore-bearing rock. After 50 years of well aerated, well-drained, and high-infiltration leaching, the rock pile has been and continues to be acid-neutralizing since the precipitation today is acidic. The study of the rock pile will be discussed at this stop.

Immediately to the south of the rock pile a weather station is located in a clearing on the east side of the entrance road. This station was installed in 2008 and used to collect site-specific temperature, wind, precipitation, ground temperature, and air quality information for the Environmental Impact Assessment.

## STOP 3: Monitoring Well Sites

*Monitoring well nests were installed across the site to determine groundwater and aquifer characteristics (flow direction, flow rate, vertical gradients, groundwater chemistry, etc.). As we travel towards the Lake Superior shoreline, several sets of monitoring wells can be observed, and we will stop briefly to discuss the data collected from them.*

At each monitoring well site, a well was installed into the glacial deposits, and a second well installed into the underlying bedrock.

Groundwater elevation data collected from the well network indicates that the groundwater in the glacial deposits, upper portions of the Copper Harbor formation, and the Nonesuch formation flow to the north-northwest, toward Lake Superior, and generally follows the slope of ground surface topography. The indicated groundwater velocity is from 0.63 to nearly 2.8 feet per year (fpy) within the bedrock units and from 0.7 fpy to nearly 1.1 fpy in the glacial deposits.

The highest concentrations of total dissolved solids (TDS) and chloride in the groundwater are located in the top of the Copper Harbor Formation and the bottom of the Nonesuch Formation; concentrations decrease upward through the Nonesuch (Fig. 10). The groundwater in the glacial deposits generally contains much lower TDS concentrations. There is very little connection between groundwater within the various units, except for the uppermost unit in the glacial deposits which illustrates characteristics of recharge from surface water.

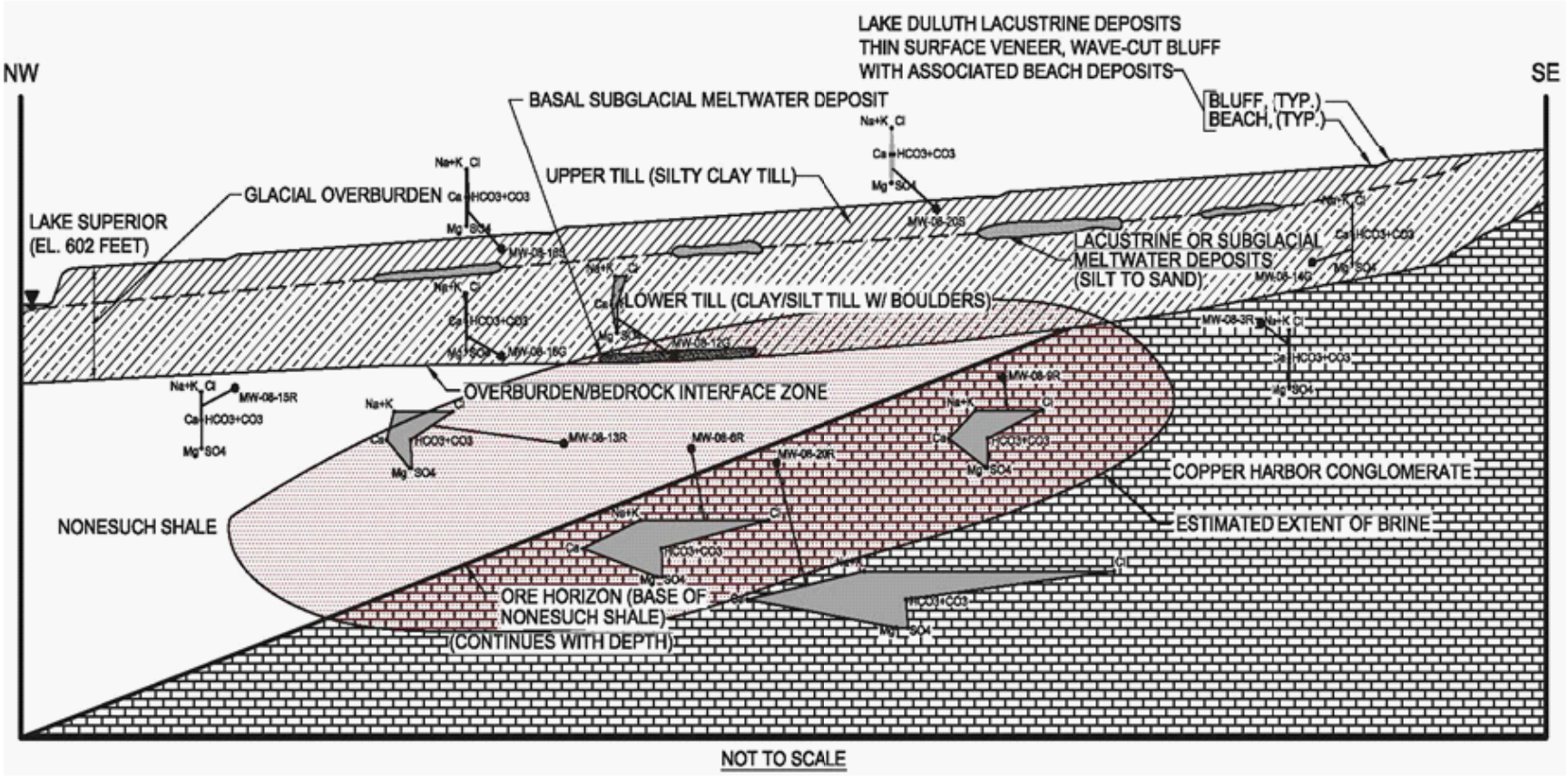



Figure 10: Schematic cross section illustrating the distribution of brine-rich groundwater. Groundwater chemistry plots (Stiff diagrams) indicate the relative cation/anion concentrations.

Groundwater in the uppermost portions (upper 30 feet) of the glacial deposits has relatively low TDS (average of 477 mg/L). It is depleted in sodium, chloride, and sulfate, and is mainly calcium bicarbonate type water. This water type is typical of groundwater near a precipitation-fed recharge zone that has had relatively short contact time with the geologic materials.

Groundwater within the remainder of glacial deposits has an average TDS of 878 mg/L (ranging between 110 mg/L to 7,300 mg/L) and it is depleted in magnesium and sulfate. The ion composition varies from sodium chloride, calcium chloride, sodium bicarbonate/carbonate, to calcium bicarbonate/carbonate water types. These variations in water type and TDS concentrations indicate the groundwater is not in connection with a precipitation-fed recharge zone and lacks connection with or flow path between the overlying uppermost zone. Sodium- and/or chloride-type water typically indicates that groundwater is equilibrating with the surrounding geologic matrix.

For wells screened in the Nonesuch Formation, TDS ranges between 115 mg/L to 34,000 mg/L, with an average of 6,800 mg/L, which is about eight times that for groundwater residing in the overlying glacial deposits. Calcium chloride is the dominant water type, particularly when TDS is elevated. Sodium is also a dominant cation in groundwater at some wells. This groundwater is not near a precipitation-fed recharge zone and is not connected with the overlying glacial deposits.

Wells screened in the Copper Harbor Conglomerate have a range of TDS between 140 mg/L to 66,000 mg/L, with an average of 10,545 mg/L, which is about 1.5 times that for groundwater in the overlying Nonesuch Formation. Groundwater in one well has TDS at 66,000 mg/L, which is greater than the TDS of sea water (35,000 mg/L). The groundwater is calcium-chloride type. In addition to calcium and chloride, sodium and bicarbonate/carbonate ions are dominant in groundwater from wells with lower TDS. This groundwater lacks connection with other groundwater zones beneath the site.

## STOP 4: Surface Water Monitoring Points

*Monitoring of the flow characteristics of the surface water in streams was performed at several locations at Copperwood. As we travel towards the Lake Superior shoreline, streams can be observed, and we will stop briefly to discuss the data collected from them.*

The surficial drainage system at Copperwood is part of the Lake Superior watershed and is composed entirely of small streams, roughly parallel to one another, flowing to the northwest from higher ground towards the south directly into Lake Superior. There are no lakes at Copperwood. Water flow within the streams is flashy and significantly controlled by timing and duration of precipitation. No groundwater contribution has been observed in these streams.

High flow in streams occurs during spring when the snow melts and after significant rain events. Flow increases and decreases quickly during rain events. All of the streams have periods of zero measurable flow either due to dry conditions or freezing in the winter. During the summer between rain events, a slight “trickle” of water can be observed flowing between cobbles in the stream bed which enters and exits multiple isolated pools. The many isolated pools found on all of the streams are also maintained by water flow just beneath the stream-bottom substrate. The

upper reaches of the streams at Copperwood can be classified as ephemeral (flow only during or immediately after periods of precipitation) and the lower reaches as intermittent (flows only during certain times of the year). Perennial streams, which have continuous flow, are not present at the site. Ephemeral streams have no base flow and the stream beds are above the water table. Intermittent streams have base flow for at least some periods of the year. At the Copperwood site, this base flow is extremely small. No springs, seeps, or areas of wetland vegetation were observed that would indicate groundwater discharge to the surface water environment.

Surface water at Copperwood has a neutral to slightly alkaline pH with most values are between 6.5 and 8.0. Lake Superior water is slightly alkaline (average and median pH of about 8). The average and median dissolved oxygen values for surface water (8.3 mg/L and 7.9 mg/L) and Lake Superior water (8.9 mg/L and 8.0 mg/L) indicate the waters are oxidized as is typical for surface water in contact with the atmosphere. The surface water is calcium-bicarbonate type, which is associated with precipitation and little or no contact with soil. The surface water contains TDS at levels of approximately 20% of that in the groundwater in the uppermost glacial till, which is consistent with very little groundwater contribution to surface water.

## STOP 5: Incised Stream Channels

*The streams at the site are deeply incised into the glacial overburden. Along the entrance road, the stream channels are only a few feet deep, yet at the north end near Lake Superior, the channels are as much as 12 m deep. As we travel towards the Lake Superior shoreline, streams can be observed, and we will stop briefly to discuss the characteristic of these channels.*

The incised stream channels are contained within steep-walled valleys. Active erosion is present within the steep-walled portions of the stream valleys. The floors of the valleys are generally flat and range in width between 15 and 60 m. The streams meander within the bottoms of the valleys with significant portions of the streams dammed by beavers creating tiered meadows within the valleys. The upper portions of the stream valleys are generally narrower and shallower than those further downstream. The overall gradient of the streams at Copperwood is approximately 20 m per km.

## STOP 6: Glacial Exposures along Lake Superior Shoreline

*At this stop we will examine the exposure of unconsolidated glacial material in the eroding bluff along the Lake Superior shoreline.*

The current shoreline of Lake Superior is dominated by a steep bluff which rises as much as 15 m above the lake surface. The entire face of the bluff is composed of slumped blocks of silt/clay-rich till. The majority of the bluff is not vegetated, but slumped blocks of soil often contain trees and surface vegetation that were brought down from the top of the bluff; the bluff is experiencing significant erosion.

A beach of mixed sand and cobbles is present at the base of the bluff. The maximum width of this beach is 10 m and in many places is much narrower. The beach does little to protect the base of the bluff from wave action and in many places, wave action reaches across the narrow beach to the base of the bluff.

The upper till unit of the glacial diamictite is exposed on the face of the bluff. The diamictite has a massive structure with no observed stratification or laminations. It is composed of a silty clay matrix with scattered gravel. The gravel is mostly cobbles and pebbles less than 8 cm in diameter, but there are boulders up to one meter in diameter. The cobbles and pebbles on the beach are predominantly red sandstone which is likely from the Freda Formation. The exposed diamictite is interpreted as a subglacial till.

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# Geology of the Keweenawan Supergroup, Porcupine Mountains, Ontonagon and Gogebic Counties, Michigan

**Laurel G. Woodruff[1], William F. Cannon[2], Suzanne W. Nicholson[2], Klaus J. Schulz[2], and Robert Wild[3]**

**[1]U.S. Geological Survey, St. Paul, Minnesota; [2]U.S. Geological Survey, Reston, Virginia; [3]Porcupine Mountain Wilderness State Park, Ontonagon, Michigan**

## Introduction

This field trip examines the geology of the rocks of the Keweenawan Supergroup (1.1 Ga) and related intrusive rocks of the Midcontinent rift system (MRS) exposed in and around the Porcupine Mountains. Most stops on this trip were visited in a previous Institute on Lake Superior Geology field trip guidebook (Cannon and others, 1992). The stop descriptions here are taken largely from that field guide with minor updates, new location maps and photographs. Because of uncertainties of weather, road conditions, and remaining snow pack in early May, the specific stops that we will visit will not be known until the date of the trip. Latitudes and longitudes are from GPS readings using WGS 84 datum.

## General Geology

The 1.1 Ga Midcontinent rift system (MRS) is a prominent 2500 km linear feature on gravity and magnetic maps that extends from Kansas north to Lake Superior and then southeast beneath the Michigan basin to where it is cut off by the Grenville Front near Detroit, Michigan (Fig. 1). The MRS cuts across Early Proterozoic and Archean terranes and is attributed to crustal extension resulting from upwelling and decompression melting of an anomalously hot mantle plume at the base of the continental lithosphere (Nicholson and others, 1997). In the Lake Superior region, nearly complete crustal separation accompanied emplacement of as much as 2 million $km^3$ of extrusive basalt and possibly an equal volume of intrusive rocks from about 1108 Ma to about 1087 Ma (Hutchinson and others, 1990; Nicholson and Shirey, 1990; Allen and others, 1995). In the Lake Nipigon area, intrusive rocks that have been attributed to the Midcontinent Rift event suggest that magmatism may have started as early as 1115 Ma (Easton and others, 2007). The deepest part of the rift subsided along large normal growth faults to a depth of nearly 30 km, accommodating at least 20 km of rift-related volcanic rocks. Following the volcanic phase of rifting, thermal subsidence accommodated deposition of up to 10 km of overlying sedimentary rocks in central rift grabens and flanking basins. Rocks of the MRS in western Lake Superior are known as the Keweenawan Supergroup and contain a remarkably complete record of igneous intrusion, flood basalt volcanism, and clastic sedimentation (Fig. 1).

Along the south shore of western Lake Superior, initial subsidence of the MRS is recorded by deposition of the Bessemer Quartzite (Fig. 2), a blanket of relatively pure, fluvial sandstone as much as 100 m thick (Ojakangas and Morey, 1982). The Bessemer Quartzite is overlain by a great thickness of subaerial flood basalt flows and lesser intermediate and rhyolitic rocks (Powder Mill Group and Portage Lake Volcanics). Conformably overlying the volcanic rocks are fluvial sedimentary rocks (Copper

Harbor Conglomerate and Freda Formation) and lesser lacustrine sedimentary rocks (Nonesuch Formation) which are as much as 8 km thick beneath Lake Superior (Cannon and others, 1993) and at least 5 km thick on shore in the field trip area. Along the south shore of western Lake Superior, post-rift movement along the rift-bounding Keweenaw fault has caused large block rotation so that much of the Keweenawan Supergroup section is steeply to vertically dipping.

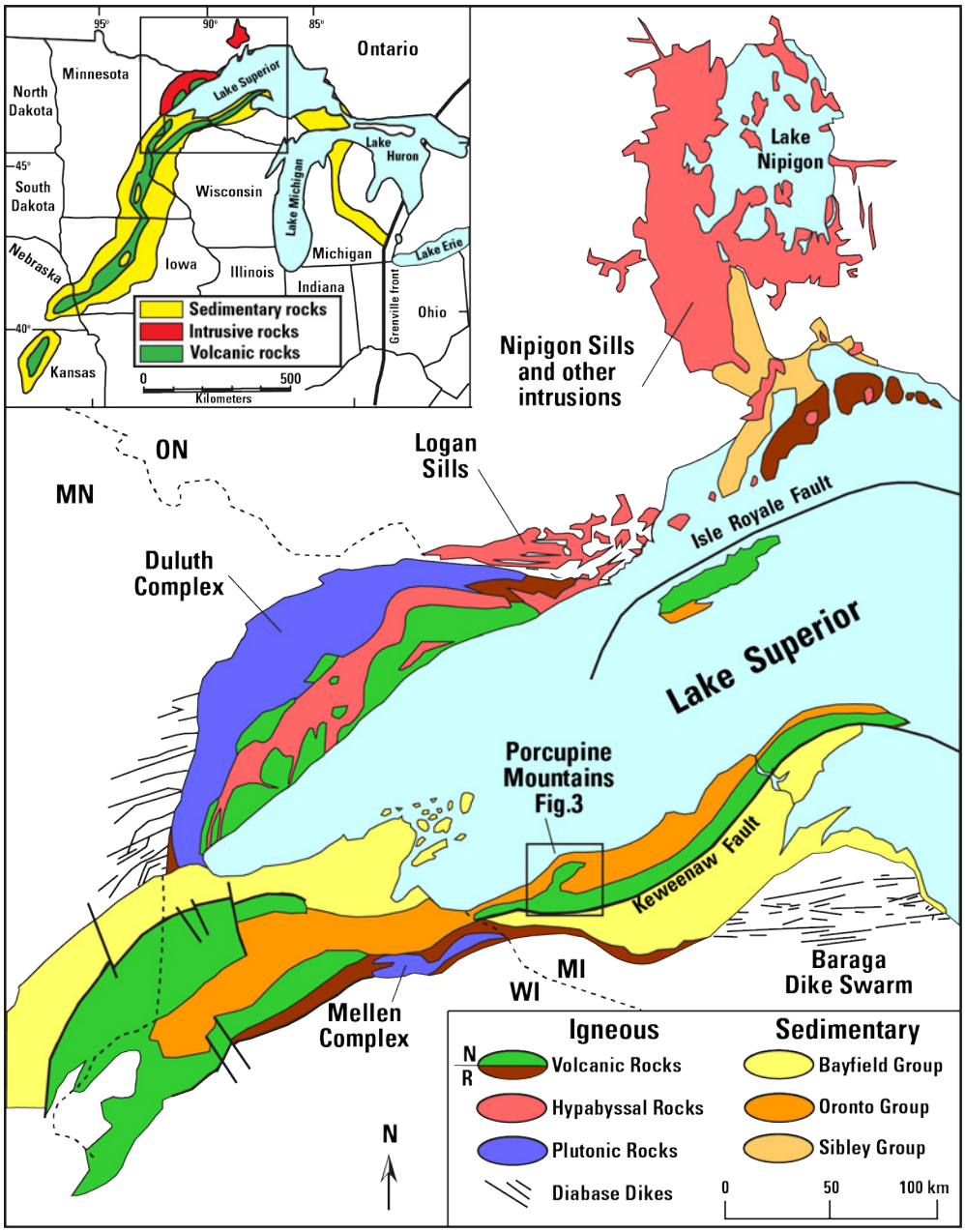



Figure 1. Regional bedrock geologic map of western Lake Superior showing the distribution of rocks related to the Midcontinent rift system (N is magnetically normal; R is magnetically reversed) and area of the geologic map in Fig. 3 (modified after Miller and Chandler, 1997).

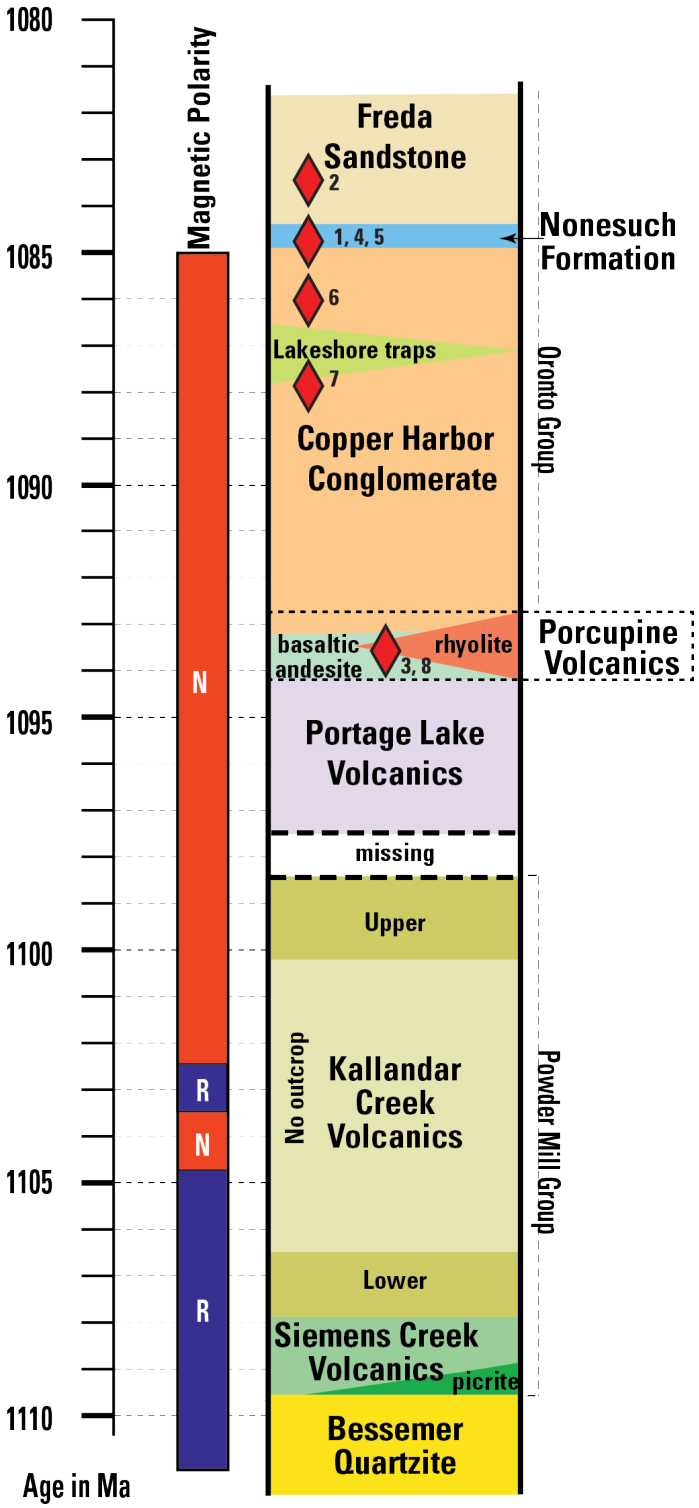



Figure 2. Age in Ma, magnetic polarity (N is normal and R is reversed), and stratigraphic section for rocks of the Keweenawan Supergroup in western Michigan and northeastern Wisconsin. Field trip stop numbers are placed in their relative stratigraphic positions (modified from Zartman and others, 1996).

Volcanic rocks of the Keweenawan Supergroup (Fig. 2) range in composition from olivine tholeiite to rhyolite. By far the dominant rock type is high-Al olivine tholeiite ($Al_20_3$ = 15 to 19 wt. %) followed by lesser high-Fe tholeiite and rocks of intermediate and felsic composition (Green, 1982; Paces, 1988; Nicholson and others, 1997). The basalts commonly are ophitic in texture and the dominant phenocryst is plagioclase. The most primitive Keweenawan basalts are geochemically similar to ocean island basalts and have incompatible trace elements ranging from slightly to strongly enriched compared to depleted or primitive mantle. Radiogenic isotope analyses (Sr, Nd, and Pb) of the main stage high-Al olivine tholeiites suggest that a likely source of the voluminous basalts was a trace element-enriched mantle plume (Paces and Bell, 1989; Nicholson and Shirey, 1990; Nicholson and others, 1997).

Flows near the base of the exposed Portage Lake Volcanics north of the Keweenaw fault were erupted at about 1096 Ma and those near the top of the formation at about 1094 Ma (Davis and Paces, 1990). Thus, the great thickness of Portage Lake Volcanics, at least 8 km in this area, was erupted in only a few million years. Because synchronous volcanism occurred along the entire trend of the rift, the rift system as a whole was producing basalt at a rate unrivalled by any modern analog (Cannon, 1992).

The Porcupine Volcanics (Fig. 2) create much of the topography of the Porcupine Mountains. The Porcupine Volcanics overlie the Portage Lake Volcanics, and represent a volcanic center that became active late in the volcanic history of the region, at about 1093 Ma. As much as 5 km of andesite, rhyolite, and basalt were erupted in a large shield volcano and deposited on top of a flat-lying lava plain composed of Portage Lake Volcanics. The unit has a lateral extent of about 35 km on the present erosion surface. The present arcuate shape of the Porcupine Mountains and the unusual hook-shaped map pattern of the Porcupine Volcanics are partly a reflection of the original shape of the volcanic shield (Fig. 3). Rhyolite near the top of the section has an age of 1093±1.4 Ma (Zartman and others, 1997).

The Porcupine Volcanics consist of a sequence of subaerially deposited (in order of abundance) andesite, basalt, felsite, and quartz-porphyry lava flows, and minor interbedded volcaniclastic lithic sandstone, siltstone, and conglomerate (Hubbard, 1975). The abundance of felsic rocks, most common near the top of the formation, where they occur as both lava flows and domes (stops 3 and 8), and the predominance of andesite over basalt clearly distinguish the Porcupine Volcanics from underlying Portage Lake Volcanics. An abundance of intermediate and felsic volcanic rocks, such as the Porcupine Volcanics, is atypical of the MRS as a whole and is limited to only a few felsic volcanic centers associated with shield volcanoes.

The major element compositions of the basalt, basaltic andesite, and andesite of the Porcupine Volcanics and the Portage Lake Volcanics are similar, but the Porcupine Volcanics are distinctly enriched in light rare earth elements (LREE) and Th compared to the Portage Lake Volcanics (Fig. 4). The two formations differ more significantly in their rhyolite chemistry and mineralogy. The rhyolite that occurs most commonly in the Portage Lake Volcanics is aphyric or may contain sparse quartz phenocrysts. In contrast, numerous rhyolite bodies in the Porcupine Volcanics range from rhyolites that are aphyric to those with abundant quartz and/or feldspar phenocrysts. Rhyolites of the Portage Lake Volcanics on the Keweenaw Peninsula typically have lower abundances of incompatible trace elements

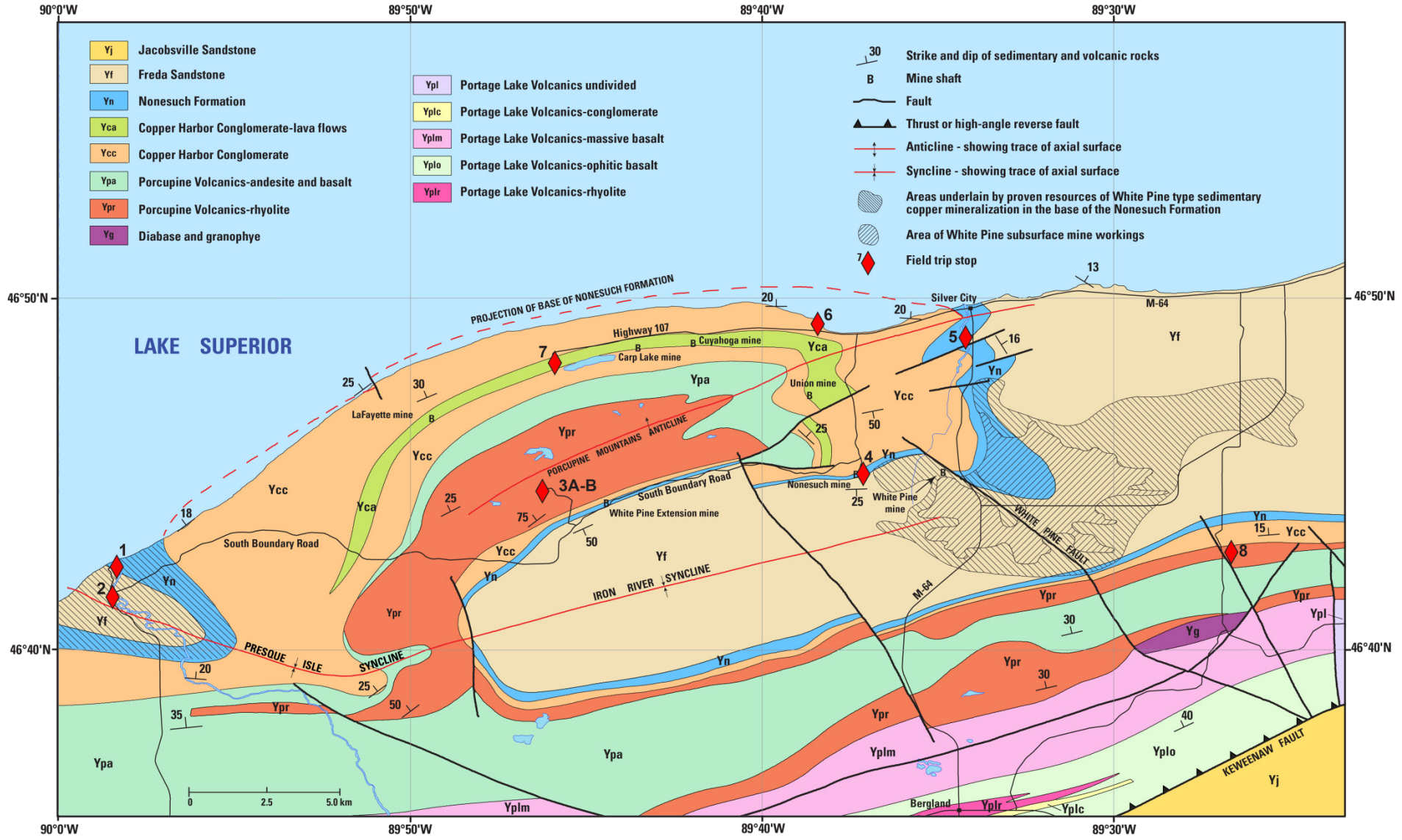



Figure 3. Generalized geologic map of the Porcupine Mountains area, modified from Cannon and others (1995). Field trip stops are shown as red diamonds.

(such as LREE, Zr, Y, Hf, and Th) than rhyolite of the Porcupine Volcanics (Fig. 4*A*). Radiogenic isotope analyses suggest that most Portage Lake rhyolites were derived by partial melting of already erupted Keweenawan basalt with a minor contribution, if any, from older basement, whereas rhyolites of the Porcupine Volcanics have a much larger contribution from older basement.

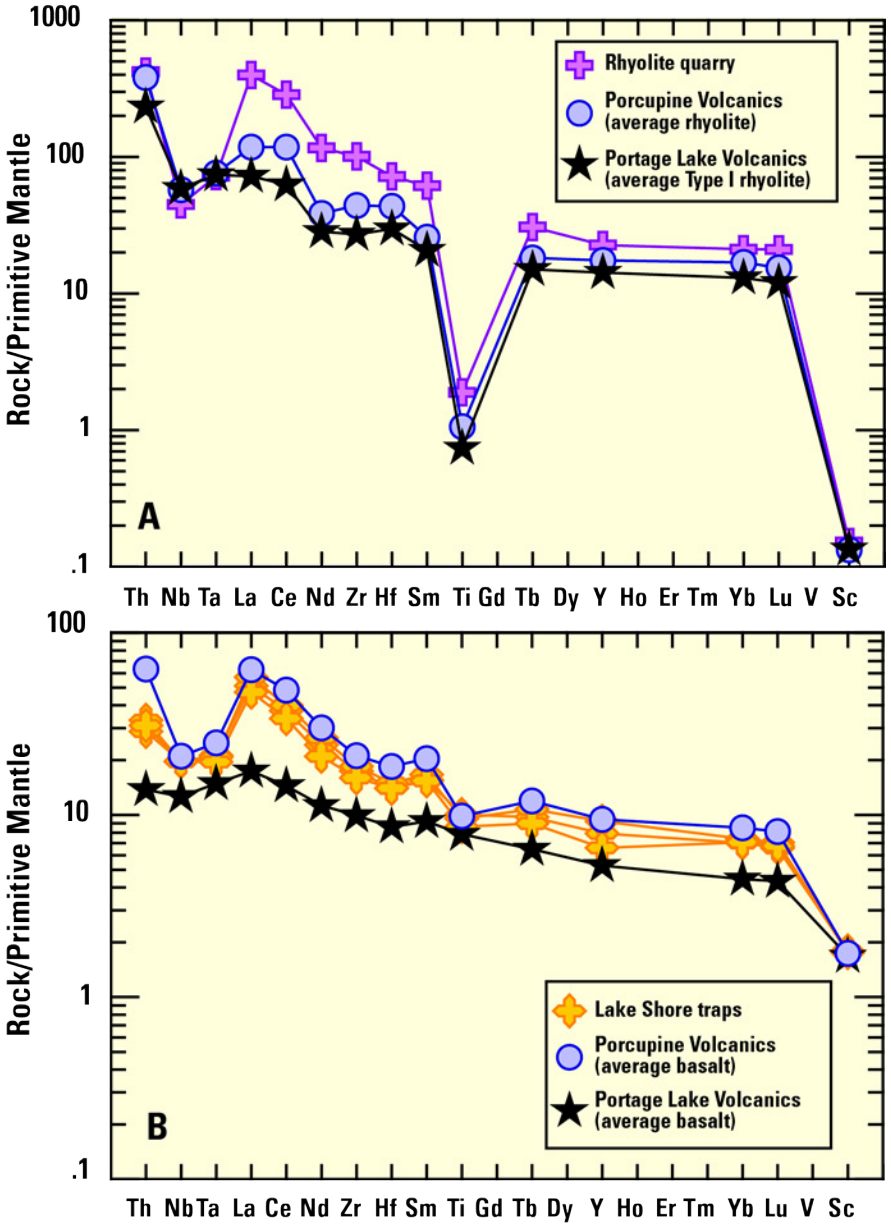



Figure 4. *A*. Spidergram illustrating the average compositions of Portage Lake Volcanics Type 1 rhyolites, Porcupine Volcanics average rhyolite, and a sample of rhyolite from the quarry at stop 8. *B*. Spidergram illustrating the average compositions of Portage Lake Volcanics average basalts, Porcupine Volcanics average basalts, and samples of the Lake Shore traps from the area of stop 7.

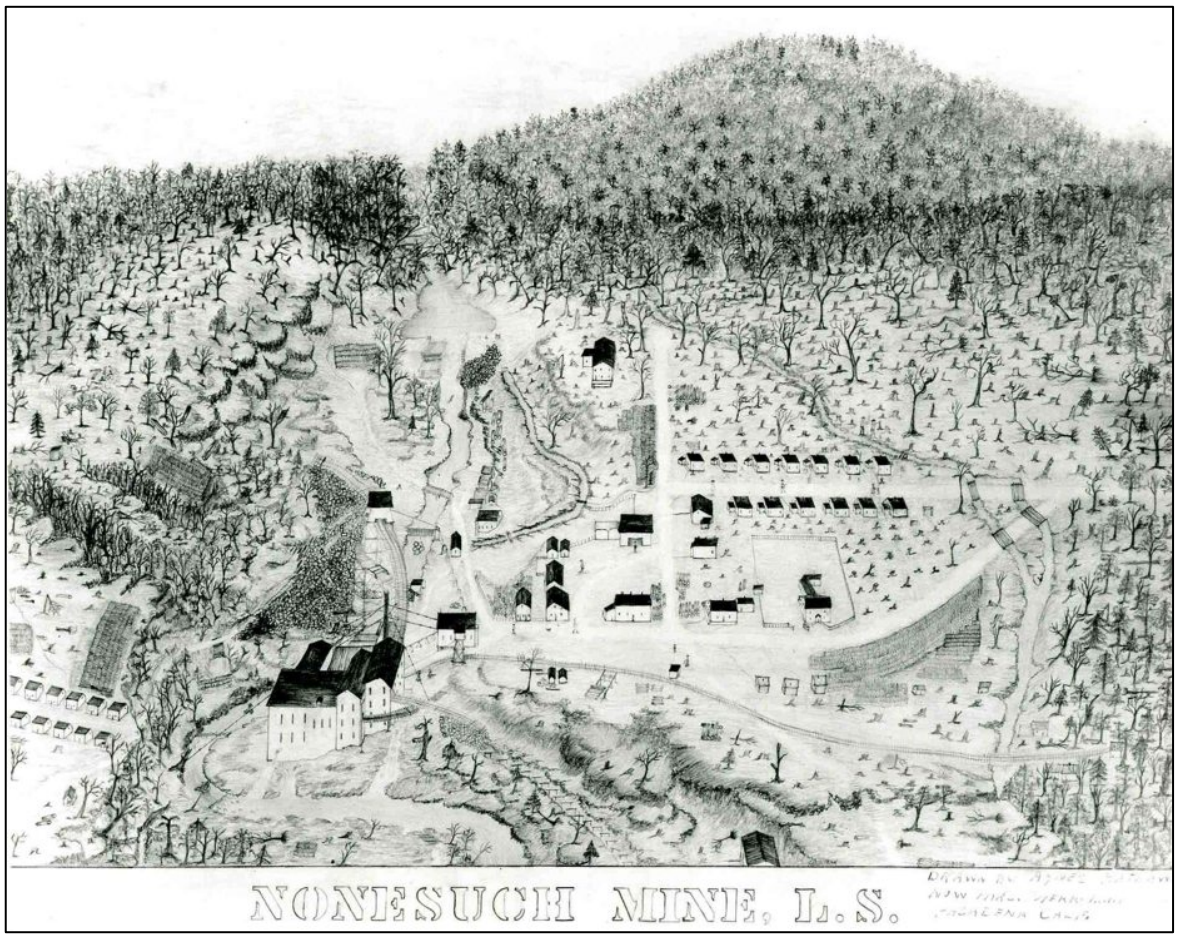
Abrupt changes in thickness of the Porcupine Volcanics are inferred along prominent structural breaks that are especially evident on the aeromagnetic map of the area (King, 1987). These breaks are believed to be synvolcanic normal faults, which outlined a central caldera. A major gravity low, centered just south of the Porcupine Mountains, was interpreted by Klasner (1989) as a large, shallow felsic intrusion. This intrusion may have been the subvolcanic felsic magma chamber that erupted much of the Porcupine Volcanics. Eruption of the Porcupine Volcanics marked an end of major volcanism in the Midcontinent Rift, with further events dominated by fluvial and lesser lacustrine sedimentation.

The Copper Harbor Conglomerate is a sequence of red to brown arkosic conglomerates and sandstones, interpreted as a northward prograding alluvial fan complex (Daniels, 1982). The Copper Harbor crops out along a discontinuous belt from the east end of the Keweenaw Peninsula westward into Wisconsin. There is an inverse relationship between the thickness of the Copper Harbor and the Porcupine Volcanics such that the Copper Harbor thins as the Porcupines Volcanics become thicker (White, 1972; Cannon and Nicholson, 1992). This suggests that the broad shield volcano that formed the Porcupine Mountains was a persistent topographic high during the time of Copper Harbor deposition. In the vicinity of the Porcupine Mountains, true conglomerate is rare; the Copper Harbor typically is a platy, reddish, fine- to medium-grained sandstone/siltstone (stop 6). Within the Copper Harbor, there are up to 31 basaltic andesite to andesite lava flows interspersed with sediment, generally in the upper part of the formation (Fig. 2). These flows, informally known as the Lake Shore traps, have an age of about1087±1.6 Ma (Davis and Paces, 1990) and form the prominent north flank of the Porcupine Mountains (stop 7).

A unit of dark gray shale and siltstone, the Nonesuch Formation (stops 1, 4 and 5), conformably overlies and interfingers with the upper Copper Harbor Conglomerate. The Nonesuch is generally interpreted to have been deposited in a perennial lake located at the toe of a transgressing-regressing alluvial fan complex (Daniels, 1982, Elmore and others, 1988; Suszek, 1997). Basal beds of the Nonesuch and locally the top of the Copper Harbor contain regional-scale low-grade stratiform copper mineralization (White and Wright, 1954). Economic–grade ore bodies are located on the western (Presque Isle) and eastern (White Pine) flanks of the Porcupine Mountains. Copper typically occurs as fine-grained disseminated chalcocite; at White Pine chalcocite is accompanied by minor native copper. The White Pine mine, a large underground mine just east of the Porcupine Mountains, produced more than 1.8 million metric tons of copper from 1955 until the mine closed in 1995. The Copperwood deposit, as the Presque Isle deposit is now known, has just completed the pre-mining permitting process. The historical Nonesuch Mine (stop 4) is an example of the early mining history in the region. Deposition of the Nonesuch Formation was succeeded by a return to fluvial redbed deposition during which at least several kilometers of red to brown sandstone and siltstone of the Freda Sandstone (stops 1 and 2) were deposited. The Freda, although still rich in volcanic detritus, is compositionally more mature than older sandstones, which probably indicates that Archean and Early Proterozoic basement in the source area was exposed by erosion of the overlying Keweenawan basalts.

Structures along much of the MRS are generally simple, consisting of thick monoclinical sections of rock titled toward the rift axis by a combination of rift subsidence, and later compression andrift inversion. Near the Porcupine Mountains, the structure is somewhat more complicated. The

major structure along the south shore of western Lake Superior is the Keweenaw fault, a reverse fault which partly inverted the central graben of the rift. A seismic section just east of the area shows the Keweenaw fault to be a north-dipping listric thrust (Hinze and others, 1990). Thus, in the area of the field trip, all sedimentary and volcanic units are on the upper thrust plate. Several cross faults that probably developed during eruption of the Porcupine Volcanics complicate the geology of the upper plate as do the gentle folds in the Porcupine Mountains, which produce a repetition of stratigraphy. Thrust faulting and folding can be indirectly dated in the interval from approximately 1060 to 1040 Ma (Ruiz and others, 1984; Bornhorst and others, 1988; Cannon and others, 1993). The regional compression that reversed the sense of movement along rift-bounding faults is likely the result of the Grenville orogeny (Cannon, 1994).




The bustling community of Nonesuch, drawn in 1884 by Agnes Hathaway, a miner's daughter. The town grew up around the Nonesuch copper mine that operated sporadically from 1866 until 1912. Now a ghost town, Nonesuch in its heyday once had a post office, school house, lumber mill, boarding house, and baseball team. The site (stop 4) became part of Porcupine Wilderness State Park in 1988. Image provided by Robert Wild.

## Stop 1: Nonesuch Formation and Freda Sandstone at the mouth of the Presque Isle River: 46.7087ºN -89.9734ºW

The upper portion of the Nonesuch Formation and the base of the Freda Sandstone are well exposed in the gorge of the Presque Isle River near its mouth and along the shore of Lake Superior west of the river (Fig. 5). Continuous exposures along the picturesque gorge of the river extend from just upstream of Nawadaha Falls to the lakeshore. Exposures continue in bluffs along the lakeshore for about half a mile west of the river mouth. An examination of exposures near the river mouth and a short distance to the west along the shore require a round trip hike of nearly a mile, mostly on well-maintained trails and stairways.

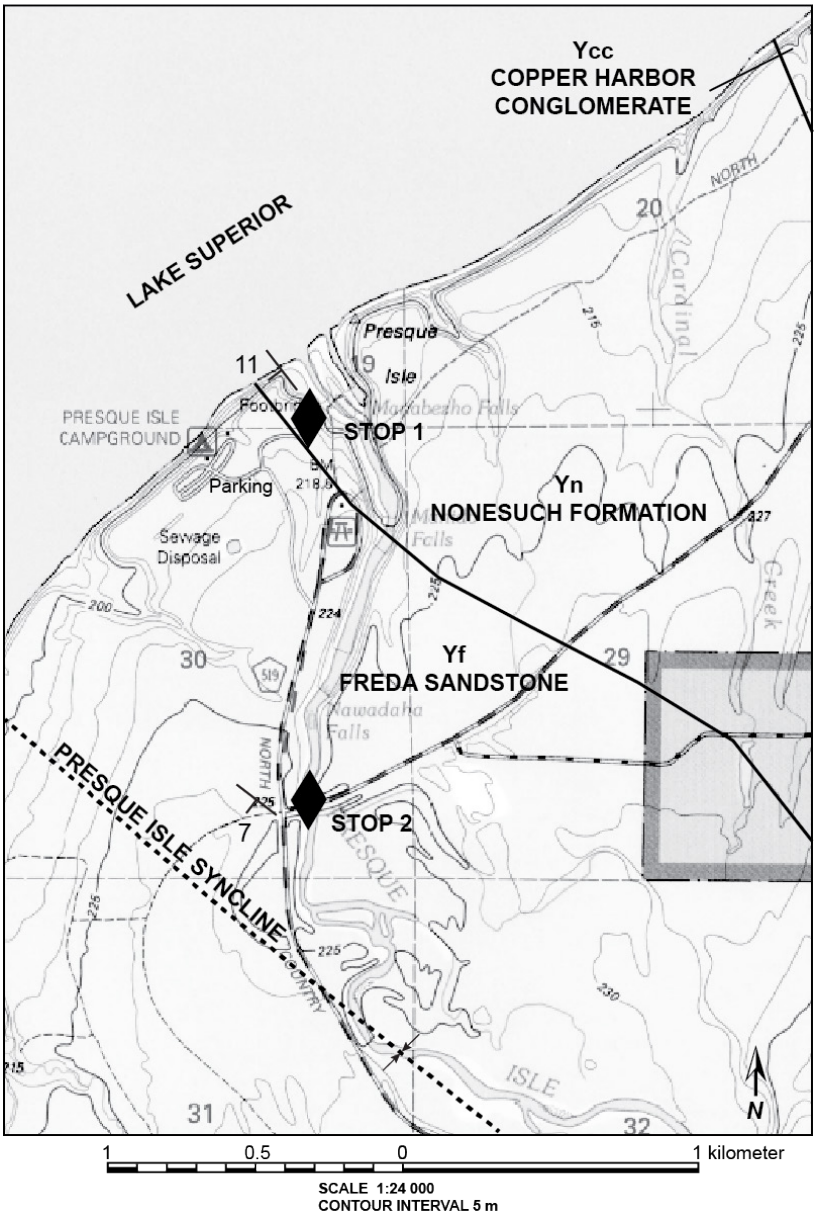



Figure 5. Location and geologic setting for stops 1 and 2. Geology is generalized from Cannon and others (1995). Structures are dotted and contacts are thin lines.

The rocks exposed here are on the northeast limb of the Presque Isle syncline, a gentle northwest-plunging fold. Dips range from nearly flat to about 10º SW. The Nonesuch Formation is distinguished from other sedimentary units of the Keweenawan Supergroup by a predominance of gray, green, or black fine-grained sediments. Lower Keweenawan felsic, intermediate, and mafic volcanic rocks were the major contributor of detritus to the Nonesuch, with contributions from Early Proterozoic and Archean crystalline rocks increasing up section (Suszek, 1997). Many of the units here show trough cross-bedding, symmetrical and asymmetrical ripples, rib and furrow structures, and parting lineations (Fig. 6*A*-*C*). A good example of ball-and-pillow structure, probably indicative of seismically-generated slumping, is seen in a one meter thick bed best exposed on the west bank of the river just upstream from the lower gorge (Fig. 6*B*). Finer-grained rocks include well-laminated shales, which are most abundant lower in the section. The Nonesuch displays coarsening-upward sequences at scales ranging from a few meters to the entire thickness of the unit. On a smaller scale, fining upward sequences are common in units from a few centimeters to a few meters thick. The Nonesuch grades upward to the Freda Sandstone through a zone of dark gray laminated and small-scale cross-bedded siltstone and sandy mudstone are interbedded with medium to coarse-grained reddish brown sandstone. There is a gradual change in oxidation state, and grain-size and bedding thickness both increase, reflecting increased environmental energy (Daniels, 1982).

Although not exposed in this area, the lower part of the Nonesuch section is strongly mineralized with very fine-grained chalcocite. Orvana Mineral Corporation is in the process of developing the Copperwood deposit, with an estimate of total proven and probable reserves of 27.42 at 1.41% Cu and 3.6 ppm Ag for contained metal of 852 million pounds of Cu and 3.2 million ounces of Ag (Bornhorst and Williams, 2011). While the lower, mineralized fine-scale stratigraphy at Copperwood is directly correlative with the stratigraphy at the White Pine mine, the upper stratigraphic sequence of the Nonesuch in the Presque Isle syncline does not correlate as well with the Nonesuch stratigraphy at White Pine. This indicates that at least during early Nonesuch deposition, sedimentary conditions were very uniform over the entire region surrounding the Porcupine Mountains. Copperwood also lacks the structural complexity and hydrothermal overprint characteristic of White Pine. The Copperwood deposit occurs along a single dipping plane and only one fault has been recognized; White Pine is cut by a major strike-slip fault and numerous smaller thrust faults. The widespread chalcocite mineralizing event is hydrothermally overprinted at White Pine by a second stage influx of Cu-bearing fluids that deposited native copper along faults and adjacent parting planes (Mauk and others, 1992).

## Stop 2: Freda Sandstone along the Presque Isle River: 46.6962ºW -89.9744ºN

At this stop, near the axis of the Presque Isle syncline, reddish cross-bedded sandstone typical of the lower part of the Freda Sandstone is exposed (Fig. 5). The gently southwest dipping beds are probably 100 to 200 m above the base of the formation and slightly higher stratigraphically than those at stop 1. These are dominantly lithic somewhat micaceous sandstones. The Freda marks a return to fluvial redbed deposition following the lacustrine deposition of the underlying Nonesuch Formation. The Freda Sandstone is a very thick unit in much of the rift in the western Lake Superior region and volumetrically is the dominant unit of the post-rift sedimentary fill. Thirty kilometers to the west, at the mouth of the



*A*. Potholes along the lower gorge of the Presque Isle River eroded into nearly horizontal shale of the Nonesuch Formation.



*B*. Ball-and-pillow structures. Approximately 1-meter-thick bed of disrupted sediments between laminated siltstone.



*C*. Ripple marks in laminated siltstone.

Figure 6. Sedimentary structures in the Nonesuch Formation exposed in the Presque Isle River during low flow. Photographs by Bill Cannon.

Montreal River, about 3500 m of the upper part of the Freda is exposed in lakeshore bluffs. Seismic sections indicate the Freda is even thicker under the lake. The Freda generally becomes finer-grained and more mature upwards. Daniels (1982) interprets the cyclic sandstone-mudstone sedimentation of the Freda as an alluvial channel-fill sequence.

## Stop 3: Porcupine Volcanics - Rhyolite at Summit Peak and Beaver Creek:

46.7433ºN -89.7711ºW

Summit Peak is the highest point in Porcupine Mountains Park and one of the highest points in the state. The observation tower at the summit provides a panoramic view of the field trip area. To the south, the highlands are underlain by the Portage Lake Volcanics and Porcupine Volcanics along the main monocline of volcanic rocks of the Keweenawan Supergroup. The lowlands immediately to the southeast are underlain by rocks of the Oronto Group in the east-plunging Iron River syncline. Looking east along strike, the stack from the former smelter at White Pine can be seen in the distance. To the north, the interior of the park extends over the rugged topography in the foreground to Lake Superior in the distance. The interior of the park is maintained as a wilderness area with access by hiking only. The park also contains some of the largest stands of virgin timber in Michigan.

Most of the park interior is underlain by a thick unit of rhyolite composed of a series of lava flows and domes typified by the rocks seen at this stop (Fig. 7). There are good exposures of coarse rhyolite breccia present along the trail leading to Summit Peak and at the overlook platform west of the summit. This breccia is probably the carapace of a rhyolite dome (stop 3A). Excellent exposures of typical intermediate and felsic units of the Porcupine Volcanics occur on the north side of the hill (549 m elevation, stop 3B) along the Beaver Creek Trail about 0.5 mi from the Summit Peak parking lot. The units dip to the south and include, in stratigraphic order, sparse outcrops of intermediate to mafic rocks as well as massive aphyric rhyolite in the creek bed. Moving up the slope, these rocks are overlain by a coarse rhyolite breccia or debris flow. The breccia contains clasts ranging in size from nearly a meter to less than 1 cm. The breccia is clast-supported and some clasts are subrounded, whereas others are flow-banded. Overlying the breccia is a medium-grained, vesicular basalt flow. Capping the hill, and overlying the basalt, is an aphanitic massive rhyolite that is microspherulitic with some crackle breccia on the easternmost end. The following latitude and longitude readings are given as a guide to locating the different units in the Beaver Creek section.

1) 46.7409ºN -89.77678ºW: massive aphyric rhyolite in stream bed
2) 46.74067ºN -89.77821ºW: rhyolite breccia; some fragments flow-banded
3) 46.74037ºN -89.77795ºW: vesicular basalt flow
4) 46.73981ºN -89.77702ºW: massive to flow-banded microspherulitic rhyolite; eastern end is crackle breccia
5) 46.74067ºN -89.77821ºW: rhyolite breccia; some fragments flow-banded

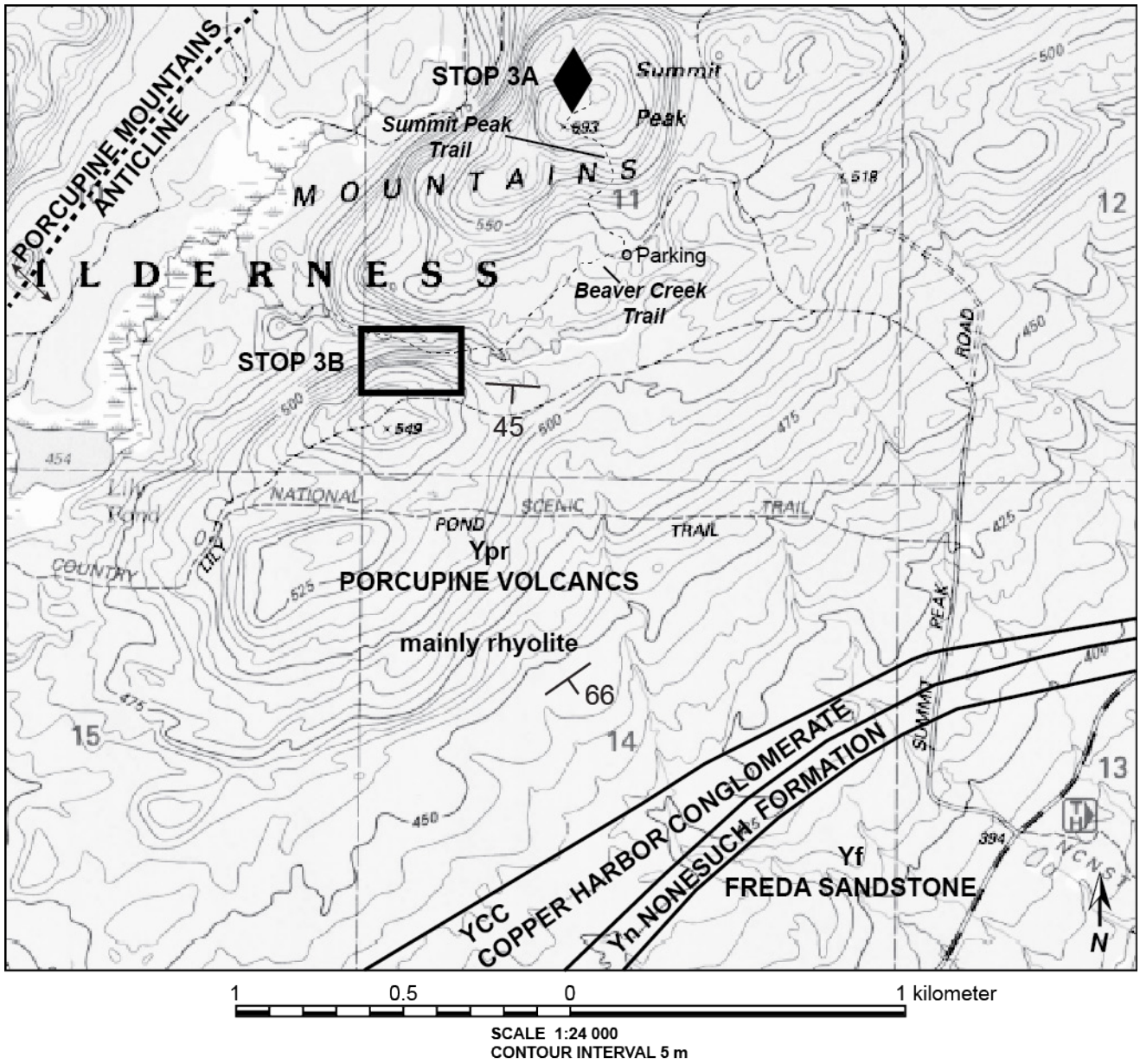



Figure 7. Location and geologic setting for stops 3*A* and *B*. Geology generalized from Cannon and others (1995). Structures are dotted and contacts are thin solid lines. Box indicates the area of multiple units along the Beaver Creek Trail for stop 3*B*.

## Stop 4: Historic Nonesuch Mine Site: 46.760ºN -89.620ºW

The Nonesuch Mine first opened in 1866, extracting finely disseminated native copper from sandstone and shale near the contact between the Copper Harbor Conglomerate and the Nonesuch Formation (Fig. 8). The mine went through a long history of openings and closings. In the 1880's the prospects looked quite encouraging, with an operating stamp mill that eventually produced 110 tons of copper, transported by tram to a dock 5 miles away at Union Bay. Four separate shafts on either side of the river extended to a depth of about 460 feet (Butler and Burbank, 1929). A town site with around 300 people sprang up around the operation. However, the very fine-grained nature of the copper made

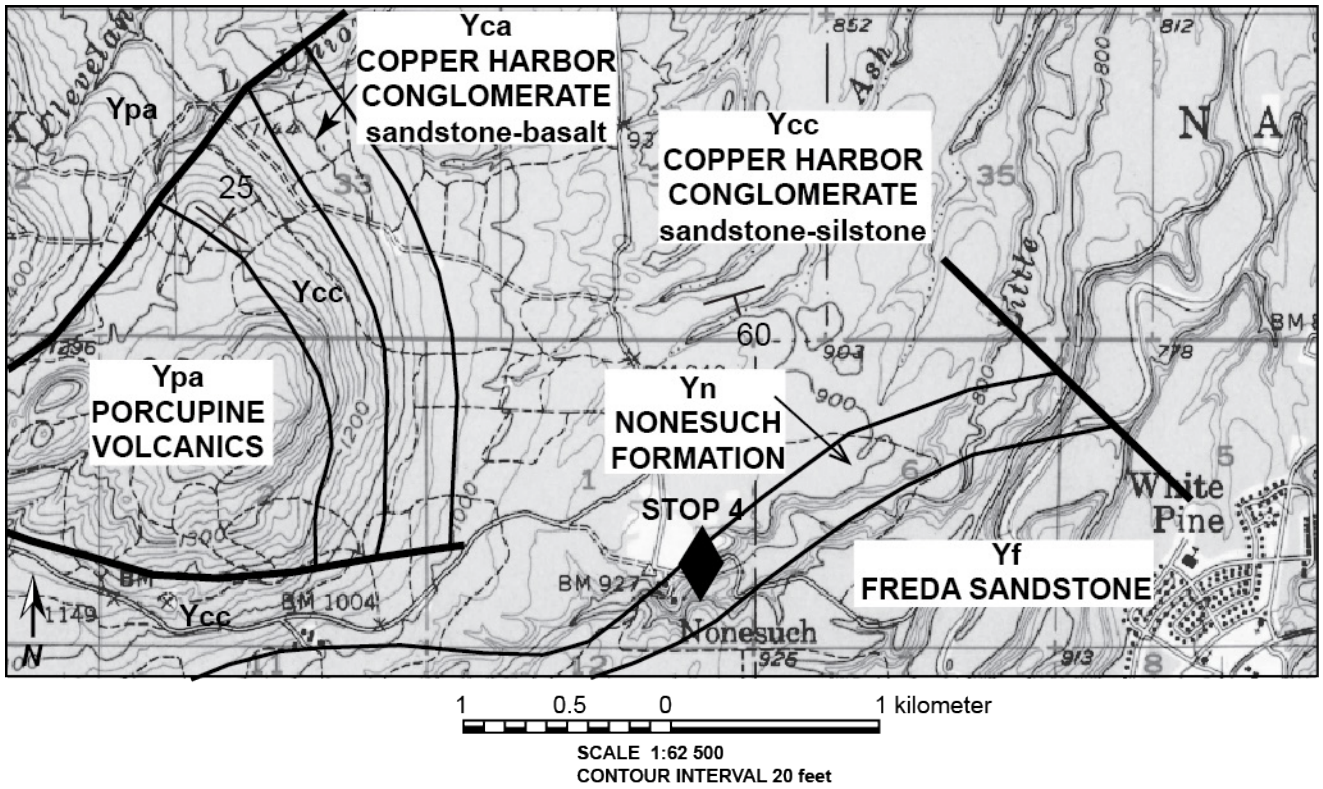



Figure 8. Location and geologic setting for stop 4. Geology generalized from Cannon and others (1995). Contacts are thin lines and faults are thick lines.

extraction difficult. A final attempt with chemical leaching that was successful in small pilot trials proved to be unsuccessful on a large-scale, and the mine closed again late in 1884, with most of the machinery subsequently stripped from the site. Several unsuccessful efforts were made to reopen the mine, one in 1906 and another in 1912, after which the mine closed for good. Total production for the Nonesuch Mine is estimated at 389,000 pounds of copper from 1868-1885 (Butler and Burbank, 1929). The former town site is now listed as a ghost town, with old foundations and lilac and apple trees in a grassy field as the only evidence of the town today.

The contact between the Copper Harbor Conglomerate and base of the Nonesuch Formation is well exposed in small rapids on the Little Iron River where beds dip about 30$^{o}$ to the east (Fig. 9A). The upper Copper Harbor is a fluvial sandstone, well cross-bedded, and contains lenses of conglomerate as much as 0.5 m thick with clasts as large as 15 cm. (Fig. 9B). Above a transition zone less than a meter thick, is thinly laminated shale and siltstone of the basal Nonesuch Formation. This is the horizon from which the ore was mined, although mineralization is not evident in this exposure.



*A*. Laminated gray shale at the base of the Nonesuch Formation a few meters above the contact with the Copper Harbor Conglomerate.



*B*. Conglomerate lens in the uppermost Copper Harbor Conglomerate. The base of the Nonesuch is exposed at the base of the falls (far left).

Figure 9. Nonesuch Formation and Copper Harbor Conglomerate exposed during low flow in the Little Iron River near at the Nonesuch mine site. Photographs by Bill Cannon.

## Stop 5: Nonesuch Formation at Bonanza Falls: 46.8177ºN -89.5701ºW

The most complete exposure of the Nonesuch Formation in the region is along the Big Iron River near Bonanza Falls, although access can be difficult and dangerous at times of high water (Fig. 10). The Nonesuch is exposed nearly continuously in a gently southeast-dipping section from just upstream of Bonanza Falls to the sharp bend in the river near the northeast corner of section 13 (Fig. 11*A*). A detailed measured section is presented by Suszek (1991). The exposed rocks total 226 m of section, which includes nearly the entire Nonesuch, although neither the upper nor lower contact is directly exposed.

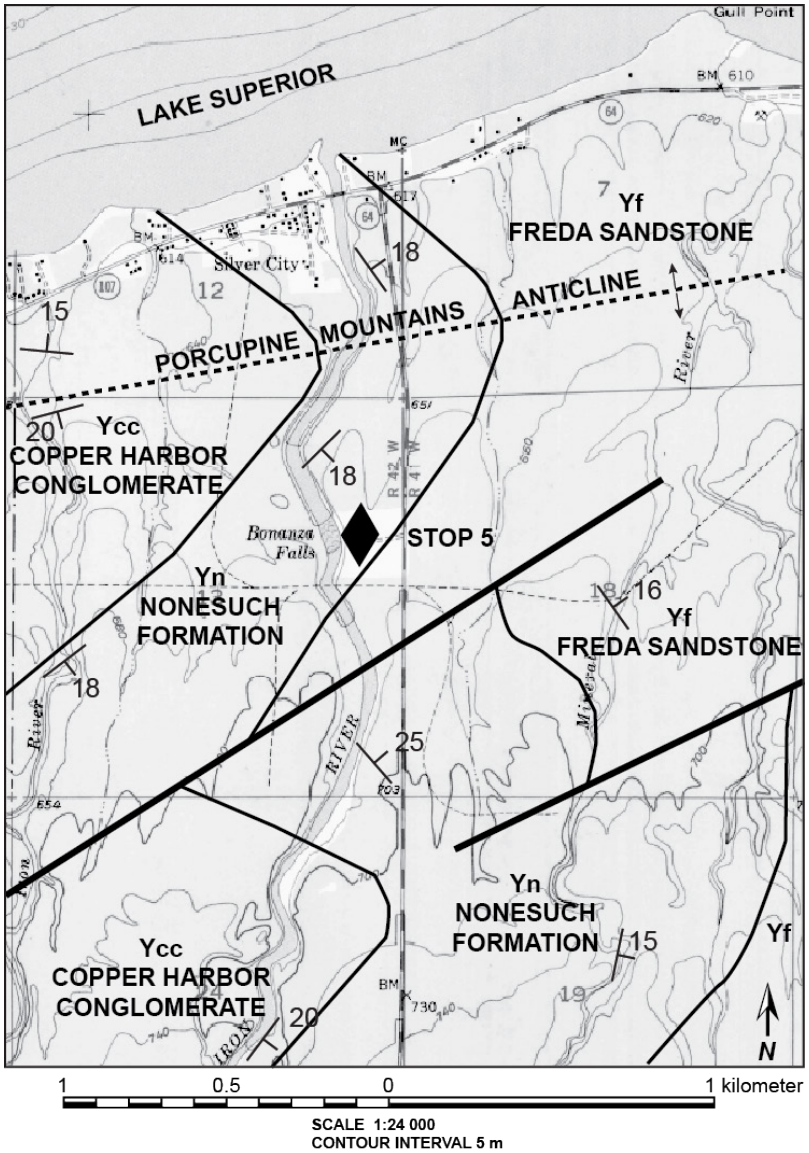



Figure 50. Location and geologic setting for stop 5. Geology generalized from Cannon and others (1995). Structures are dotted lines, contacts are thin lines, and faults are thick lines.



*A*. Exposure of the Nonesuch Formation in the Big Iron River at Bonanza Falls, looking upstream. Beds dip gently upstream.



*B*. Contorted bedding in the Nonesuch at Bonanza Falls, probably generated by seismic liquefaction. Scale card is 85 cm wide.

Figure 6. Nonesuch Formation exposed at Bonanza Falls during low flow in the Big Iron River. Photographs by Bill Cannon.

The Nonesuch Formation in the Big Iron River section is dominantly siltstone and fine-grained sandstone with minor shale. Many rocks have trough cross-bedding, symmetrical and asymmetrical ripples, rib and furrow structures, parting lineations, and soft sediment deformational features (Fig. 11*B*). The finer-grained rocks include well-laminated shale, which is most abundant lower in the section. The shaley units commonly have ball-and-pillow structures and calcareous concretions.

The Nonesuch here displays coarsening-upward sequences at scales ranging from a few meters to the entire thickness of the unit. On a smaller scale, upwardly fining sequences are common in units from a few centimeters to a few meters thick. In the lower 10 m of the section, copper mineralization occurs as concentrations of chalcocite, bornite and malachite along bedding planes. The mineralization is cogenetic with the major copper mineralization in the White Pine mine, where the downdip extension of this unit was mined just to the south and east. A good exposure of the mineralized base of the Nonesuch Formation and the top of the Copper Harbor Conglomerate occurs along the Little Iron River near the center of the SW1/4, Section 13, but requires a walk of about 1 mi south from Highway 107. It is an easy walk along an unmaintained trail on the east bank for those who can spend more time in the area. At this location, remains of early mining efforts for native silver occurs there, as well as 'ore' specimens from old dumps.

Copper was initially introduced to the lower Nonesuch Formation in both the White Pine and Copperwood districts during early diagenesis, probably by upward circulating connate waters which dissolved copper from the underlying redbeds of the Copper Harbor Formation (Swenson and Person, 2000). Chalcocite, largely of microscopic size, formed by the replacement of diagenetic pyrite. A later phase of copper mineralization documented by Mauk and others (1992) in the White Pine mine, introduced native copper mostly along fault zones and adjacent strata. This second stage of native copper mineralization commonly occurs as large thin plates of native copper developed along bedding planes in the Nonesuch. It is locally known as sheet copper and is probably cogenetic with the classic native copper mineralization of the Portage Lake basalts along the Keweenaw Peninsula.

## Stop 6: Upper part of Copper Harbor Conglomerate at Union Bay Campground:

46.8253ºN -89.6418ºW

Good exposures of reddish sandstone containing thin conglomerate beds are abundant along the shore of Lake Superior in the park at the Union Bay Campground (Fig. 12). The Copper Harbor Conglomerate is typically characterized by coarse volcanogenic conglomerate, which forms most of the lower part of the section throughout much of its outcrop belt and which grades up into finer grained sandstone (Elmore, 1984). However, south of the village of Ontonagon, there is a different facies relationship. The lower part of the formation here is mostly sandstone, siltstone, and basalt or andesite lava flows; conglomerate is very subordinate. These rocks underlie the high hills immediately south of Highway 107. A coarse conglomerate facies occurs higher in the section, but forms less than 10 percent of the thickness. The exposures here at Union Bay are near the base of the upper unit and are probably about 1,000 m above the base of the formation. Sandstone layers at Union Bay dip 10-20° to the north. Sandstone is volcanogenic and quartz-poor. There are excellent examples of trough cross-bedding, generally indicating a northeastward current vector. A variety of other sedimentary features including

desiccation cracks, rip-up clasts, oscillation and current ripples, and swash marks are also present (Fig. 13*A*-*D*).

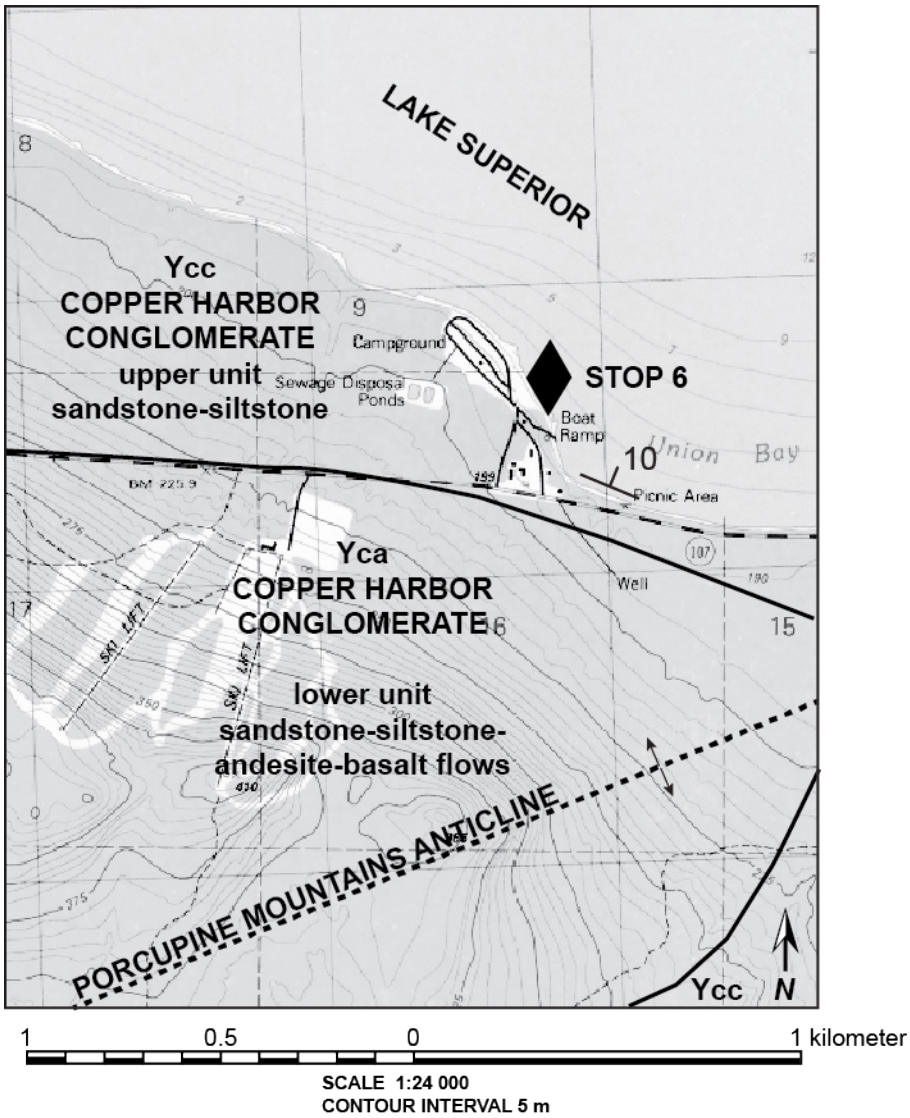
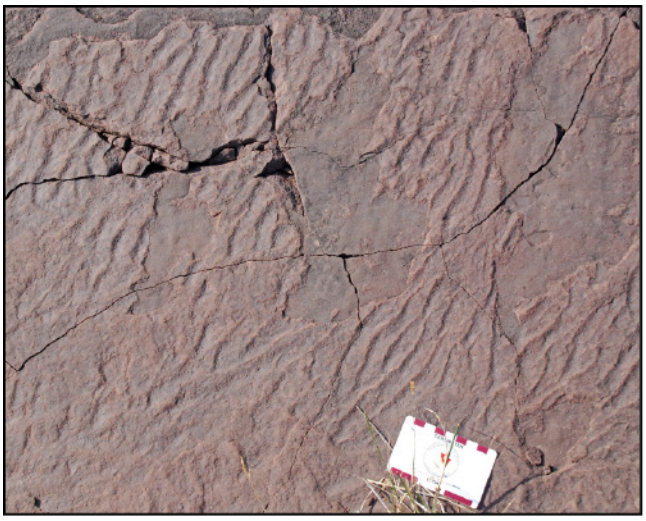


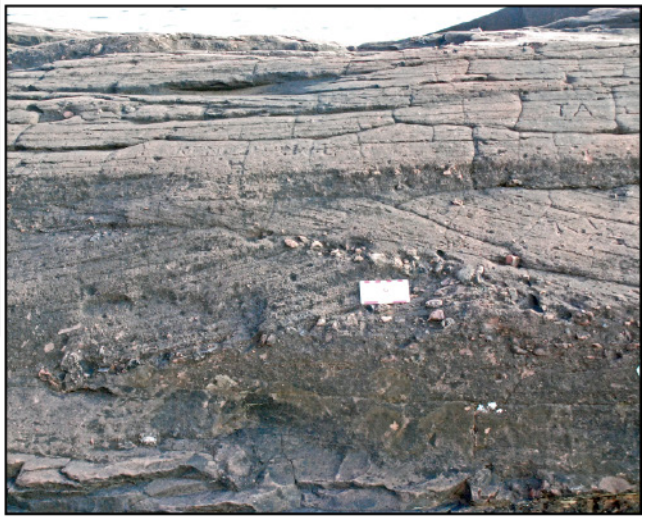

Figure 7. Location and geologic setting for stop 6. Geology generalized from Cannon and others (1995). Structures are dotted and contacts are thin lines.

The exposures of Copper Harbor Conglomerate north of the Porcupine Mountains are the farthest from the source highlands to the south. The relative scarcity of thick coarse conglomerate compared to exposures farther south probably reflects the distal nature of these rocks. These northernmost outcrops are a good representative of much of the Copper Harbor beneath Lake Superior. The rocks at Union Bay show a an irregular coarsening-upward trend as opposed to the fining-upward trend typical of the more proximal parts of the unit. This relationship is consistent with northward prograding alluvial plain deposition.

Several large boulders are distributed along the beach and are composed of conglomerate typical of the lower part of the Copper Harbor elsewhere. In the boulders the conglomerate contains, almost exclusively, clasts of Keweenawan Supergroup volcanic rock types common in the region. An



*A*. Ripple marks



*B*. Cross-bedding



*C*. Mudcracks



*D*. Lineations on bedding surface, possibly indicating current directions. Note variations in direction between various bedding planes.

Figure 13. Sedimentary features in red siltstone and sandstone of the Copper Harbor Conglomerate along the Lake Superior shoreline at Union Bay Campground. Scale card is 85 cm in width. Photographs by Bill Cannon.

interesting question is the source of these boulders. Although the Copper Harbor does contain some conglomerate beds nearby (for example, Fig. 9*B*), none of the streams entering Lake Superior in this area seem capable of transporting such large boulders. Most streams, especially near the lakeshore, have low gradients, and streambeds do not contain such large boulders. These boulders are very likely glacial eratics that have been transported from some distance away. Ice movement was from the northeast, roughly parallel to the present shoreline. The boulders probably came from the Copper Harbor Conglomerate farther east toward the Keweenaw Peninsula where thick conglomerate is common.

## Stop 7: Basalt flows within the Copper Harbor Conglomerate (Lake of the Clouds overlook): 46.8031ºN -89.7641ºW

Along Highway 107 leading to the Lake of the Clouds overlook are several exposures of conglomerate within the Copper Harbor Conglomerate. To the north is a good view of Lake Superior and the lowlands underlain by sedimentary rocks of the Oronto Group. From the overlook parking lot, a short hike leads to the overlook and a spectacular view of Lake of the Clouds and the Porcupine Mountains Wilderness State Park (Figs. 14 and 15).

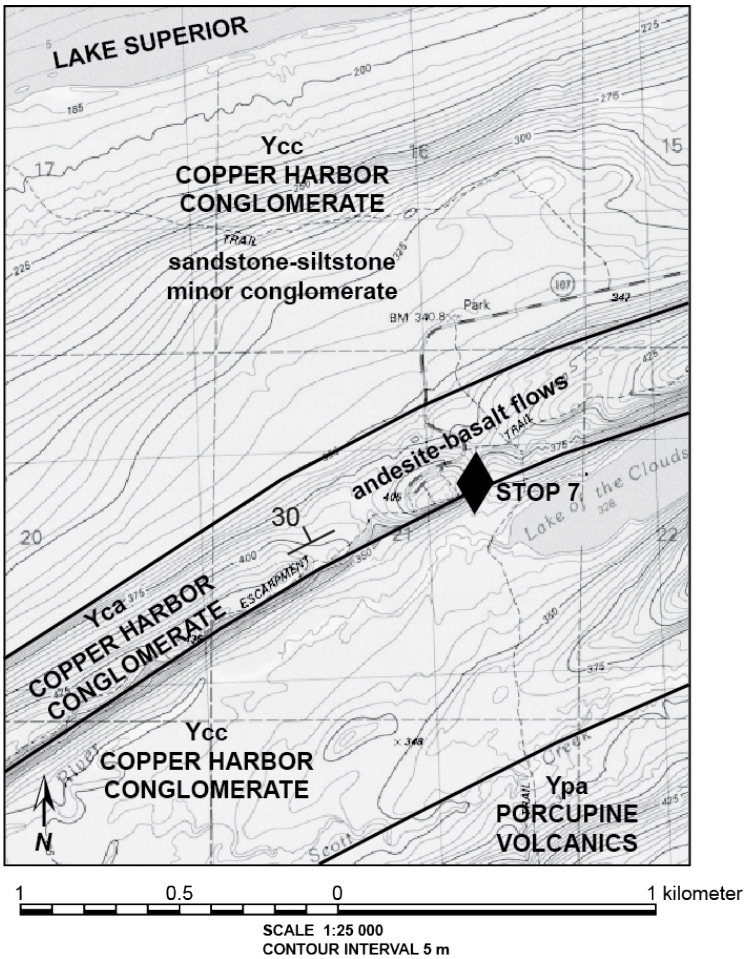



Figure 8. Location and geologic setting for stop 7. Geology generalized from Cannon and others (1995). Contacts are thin lines.



Figure 15. Panoramic view of Lake of the Clouds, Porcupine Mountains Wilderness State Park. Photograph by Bill Cannon.

The overlook is along the south escarpment of a high ridge supported by a series of north-dipping lava flows within the Copper Harbor Conglomerate (Fig. 16). These flows are known as the Lake Shore traps. Comparable flows within the Copper Harbor Conglomerate at the tip of the Keweenaw Peninsula have an age of 1087.2 ± 1.6 Ma (Davis and Paces, 1990). The low area south of the ridge, including Lake of the Clouds, is underlain by sandstone and siltstone and a few basalt flows which constitute the lower part of the Copper Harbor Conglomerate. The higher regions farther south are underlain by volcanic rocks, mostly rhyolite of the Porcupine Volcanics.

Toward the east end of the overlook area, a large glaciated surface shows a series of thin basalt flows, which average a few meters thick. Individual flows can be readily identified by chilled vesicular bases, in places containing inclusions of older flows, and by rubbly or vesicular tops. Abundant epidote alteration and vesicle fillings impart a distinctive greenish cast to flow margins (Fig. 16). Hubbard (1975) described the flows in the Copper Harbor as mostly andesite with minor basalt, but chemical analyses of two samples from this locality indicate that they are basalt, similar in composition to average basalt from the Porcupine Volcanics. An interesting feature of some of these flows is the incorporation of very fine-grained red sediments both in vesicles near their base and in thin fractures extending for a meter or more up into flows. Thin vestiges of these sediments are also along flow contacts. We interpret these sediments as wind-blown dust that accumulated on flow surfaces shortly after eruption. It was still unconsolidated when the next flow was erupted. The soft sediment was then injected upward by the weight of the overlying flow.

Compared to Portage Lake Volcanics, these basalts are enriched in incompatible trace elements and show a distinct negative Nb-Ta anomaly in primitive mantle normalized incompatible trace element patterns similar to the basalts from the Porcupine Volcanics (Fig. 4*B*). This negative anomaly is a likey indication of crustal contribution to the parent magmas.



Figure 16. Contact between two basalt flows at the Lake of the Clouds overlook. The vesicular top of one flow (bottom of photo) has amygdules filled with greenish epidote. The base of the overlying flow has pipe vesicles near the lower contact, grading up to a more massive interior (towards the top of the photo). Reddish fine-grained clastic material occurs in cracks in the upper flow. Scale card is 85 cm wide. Photograph by Bill Cannon.

## Stop 8: Porcupine Volcanics: Rhyolite quarry near White Pine:

46.7155ºN -89.4435ºW

The rocks exposed here are part of a small rhyolite body near the top of the Porcupine Volcanics (Fig. 17). The body probably does not extend much beyond the hill into which the quarry is cut. The quarry provides a cross section through part of a subaerial agglutinate deposit. Agglutinate deposits form at vents by spatter of erupting magma and buildup of mounds of hot, viscous material. The mound of erupted material eventually flows outward under its own weight, resulting in large flow folds such as seen in this quarry. The near-vent nature of this deposit is deduced from the presence of lithic fragments within the rhyolite, large and small contorted flow folds, and stratification of rhyolite (light and dark units) (Fig. 18*A*-*C*). At the edges of agglutinate deposits, flowage typically has homogenized the magma, and, as such, stratification and folding are generally not preserved. Zartman and others (1997) reported an age of 1093.6 +/- 1.8 Ma for rhyolite from this quarry.

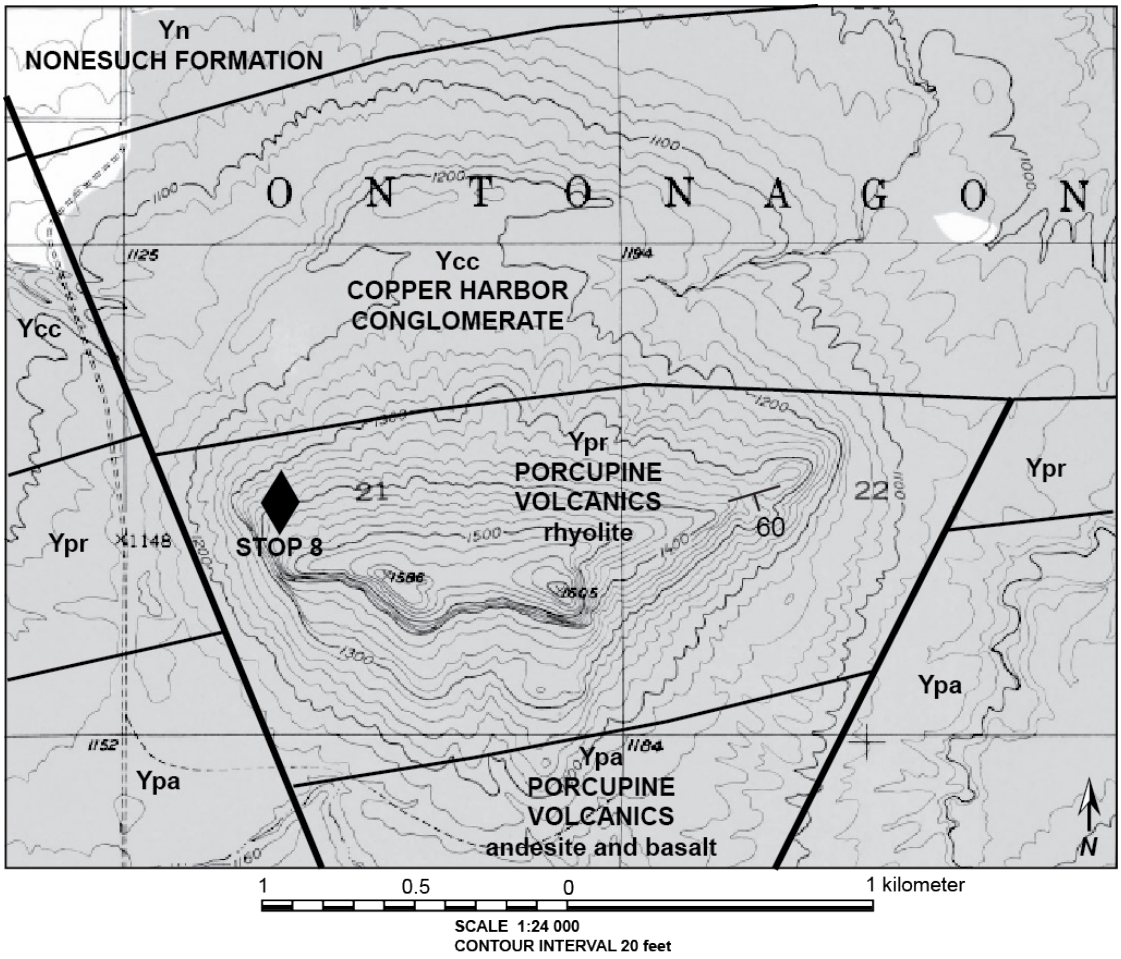



Figure 17. Location and geologic setting for stop 8. Geology generalized from Cannon and others (1995). Contacts are thin lines and faults are thick lines.

This rhyolite contains feldspar phenocrysts, which are typically aligned parallel to the stratification and foliation. Rhyolite of the Porcupine Volcanics is enriched substantially in such incompatible trace elements as Th, Ba, Zr, Hf, and LREE compared to rhyolite in the Portage Lake Volcanics on the Keweenaw Peninsula. Isotopically, rhyolite at this stop has an initial Nd isotopic signature ($\varepsilon_{Nd(1100Ma)}$ about -14) that reflects a substantial contribution from older crustal sources. The Portage Lake rhyolite from the Keweenawan Peninsula ($\varepsilon_{Nd(1100Ma)}$ about 0) is thought to be derived from partially melted early Keweenawan basalts (Nicholson and others, 1997).

*A*. Lithic fragment showing flow contact between light and dark rhyolite.





*B*. Flattened and drawn-out lithic fragments in rhyolite matrix.



*C*. Contoured flow pattern showing contrasting rhyolite melts.

Figure 18. Agglutinate textures in Porcupine Volcanics rhyolite, exposed in quarry blocks. Photographs by Bill Cannon.

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