70th Annual Meeting

Institute on Lake Superior Geology

Houghton, Michigan May 15-18, 2024



Proceedings Volume 70
Part 2 - Field Trip Guidebook



70th Annual Meeting
Institute on Lake Superior Geology

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Sponsored by:

A. E. Seaman Mineral Museum
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Theodore J. Bornhorst, Erika Vye, Patrice Cobin, and James DeGraff

Proceedings Volume 70
Part 2: Field Trip Guidebook

Compiled by Patrice F. Cobin and Theodore J. Bornhorst

Cover Photo: The only known color photograph of in situ colorless calcite crystals with inclusions of native copper. Vug is about 15 cm across and 30 cm deep; located at the top of the Knowlton basalt lava flow at the 4th level, 850 ft stope, of the Caledonia Mine, Michigan. Photo taken in 1994 soon after the vug was blasted open. Native copper in the calcite crystals has not been visibly altered despite being about 1 billion years old.

Photograph by Theodore J. Bornhorst

70th Institute on Lake Superior Geology

Volume 70 consists of:

Part 1: Program and Abstracts

Part 2: Field Trip Guidebook

- **Trip 1:** *Mesoproterozoic Midcontinent Rift-filling Strata and Native Copper Deposits of the Keweenaw Peninsula, Michigan*
- **Trip 2:** *Mining History and Geology of the Quincy Mine, Keweenaw Peninsula Native Copper District, Michigan*
- **Trip 3:** *Geoheritage of Buffalo Reef: Industrial Impact on Land, Culture, and Fish Sovereignty*
- **Trip 4:** *Keweenaw Fault System Geometry and Kinematics: Clues to Its Nature and Origin*
- **Trip 5:** *Geology and History of a Native Copper Mine: Adventure Mine, Ontonagon County, Michigan*
- **Trip 6:** *Southern Complex Granitoids, Gneisses, and Migmatites: New Data, Discoveries, and Perspectives*
- **Trip 7:** *Landslides on the Ontonagon River at Military Hill*

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Part 2: Field Trip Guidebook

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Field Trip 1

Mesoproterozoic Midcontinent Rift-filling Strata and Native Copper Deposits of the Keweenaw Peninsula, Michigan

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Introduction

The geology of the far western Upper Peninsula of Michigan consists of three temporally distinct episodes. During the Mesoproterozoic, about 1.1 Ga, up to 30 km of Keweenaw Supergroup volcanics and clastic sediments filled an intracratonic rift, the Midcontinent Rift (MCR) (Figures 1 and 2) (Cannon et al., 1989). After about 500 million years of erosion, the MCR rocks were buried by Phanerozoic sedimentary rocks from about 500 Ma to 175 Ma (Catacosinos et al., 2001). There are no exposures of rocks in the interval between 175 Ma to about 2.5 Ma (Velbel, 2009). Pleistocene continental glaciations, beginning about 2 million years ago, removed the Phanerozoic rocks from the Keweenaw Peninsula leaving only a few Phanerozoic outliers. About 10,000 years ago glaciers retreated from the Lake Superior basin and left behind a variety of unconsolidated clastic sediments. The geologic evolution of the far western Upper Peninsula is illustrated in Figure 3.

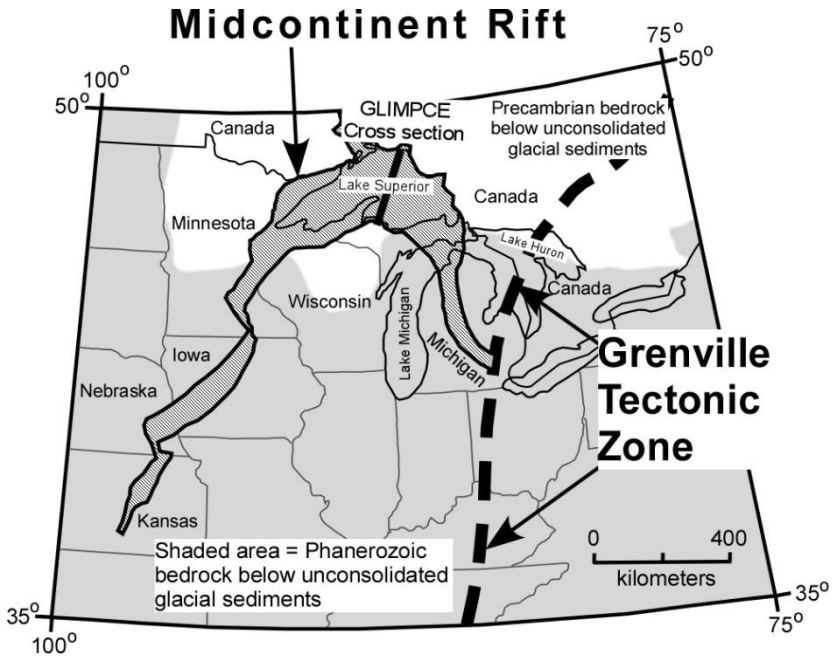


Figure 1: Generalized geologic map of the Midcontinent Rift showing Grenville tectonic zone with interpretative cross-section in Figure 2. Modified from Bornhorst and Barron (2013).

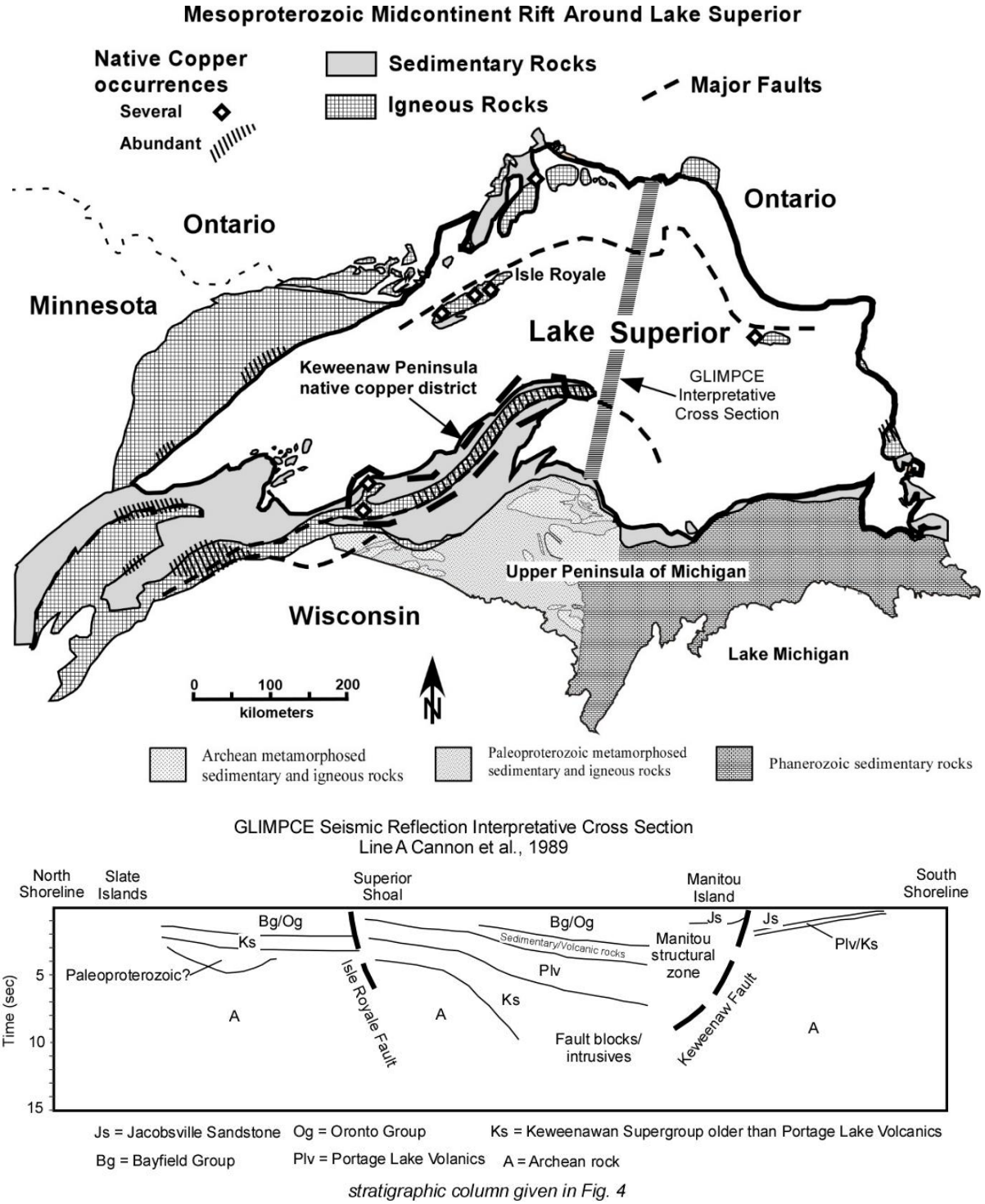


Figure 2: Generalized bedrock map showing the exposed rocks of the Midcontinent Rift around Lake Superior, the native copper occurrences, and the bedrock of the Upper Peninsula of Michigan modified from Bornhorst and Barron (2013). Interpretative cross section from Cannon et al. (1989).

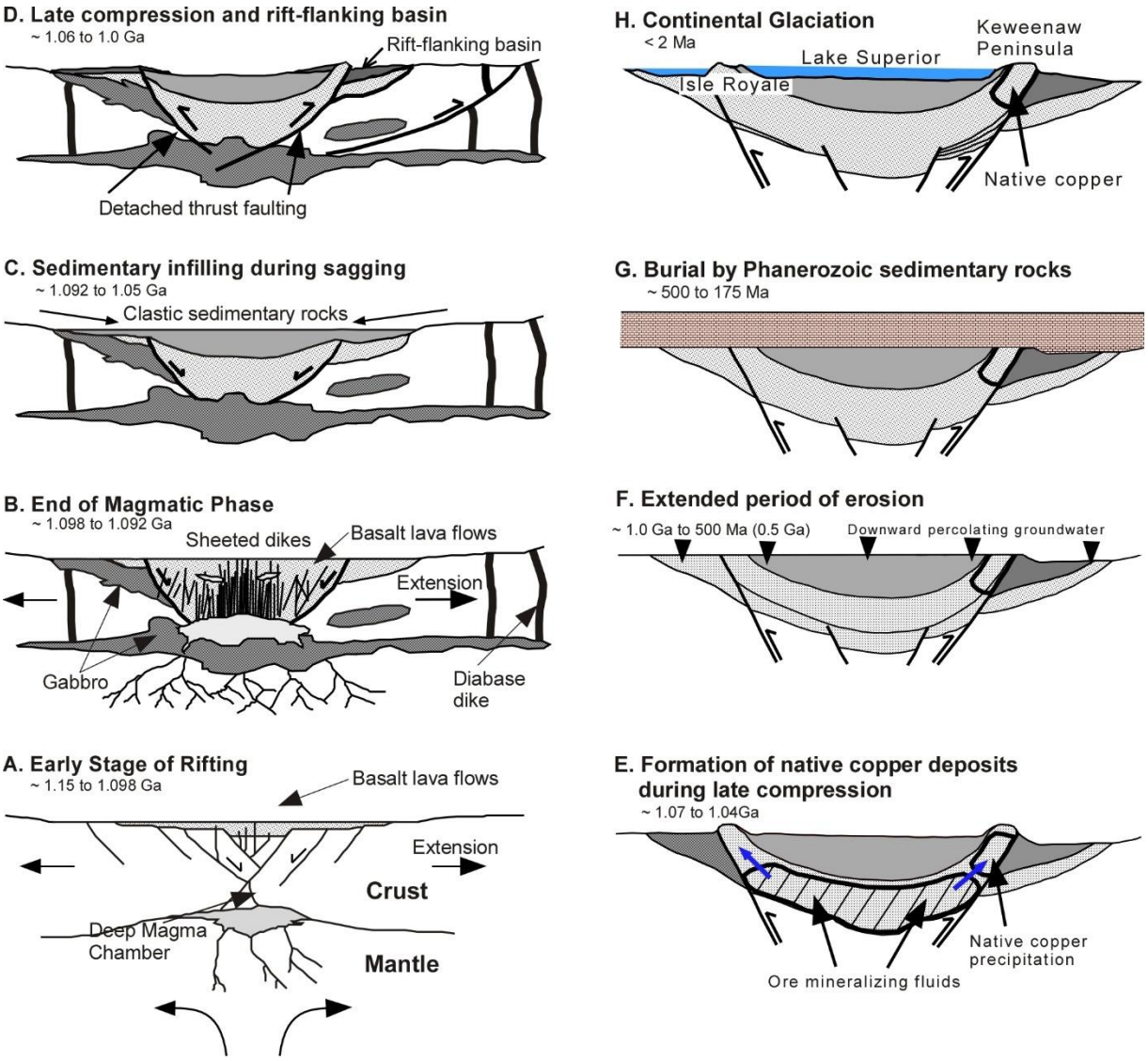


Figure 3: Cartoon NW to SE cross sections from Minnesota (left) to the Upper Peninsula (right) illustrating the progressive geologic evolution. Modified from Bornhorst and Lankton (2009).

In the strictest sense, the geographic area of the Keweenaw Peninsula proper extends from L'Anse northwest to Lake Superior perpendicular to strike of the strata, however, the term Keweenaw Peninsula has also been applied to the area containing MCR rocks farther to the south from L'Anse to the White Pine area (Figure 4 and 5). The geologic descriptions in this field trip guide are mostly restricted to the Keweenaw Peninsula proper. The descriptions provided here were modified from a combination of Bornhorst and Barron (2011, 2013), Bornhorst and Lankton (2009), Bornhorst and Rose (1994) and Bornhorst et al. (1983). These sources are mostly used here without specific citation or quotation.

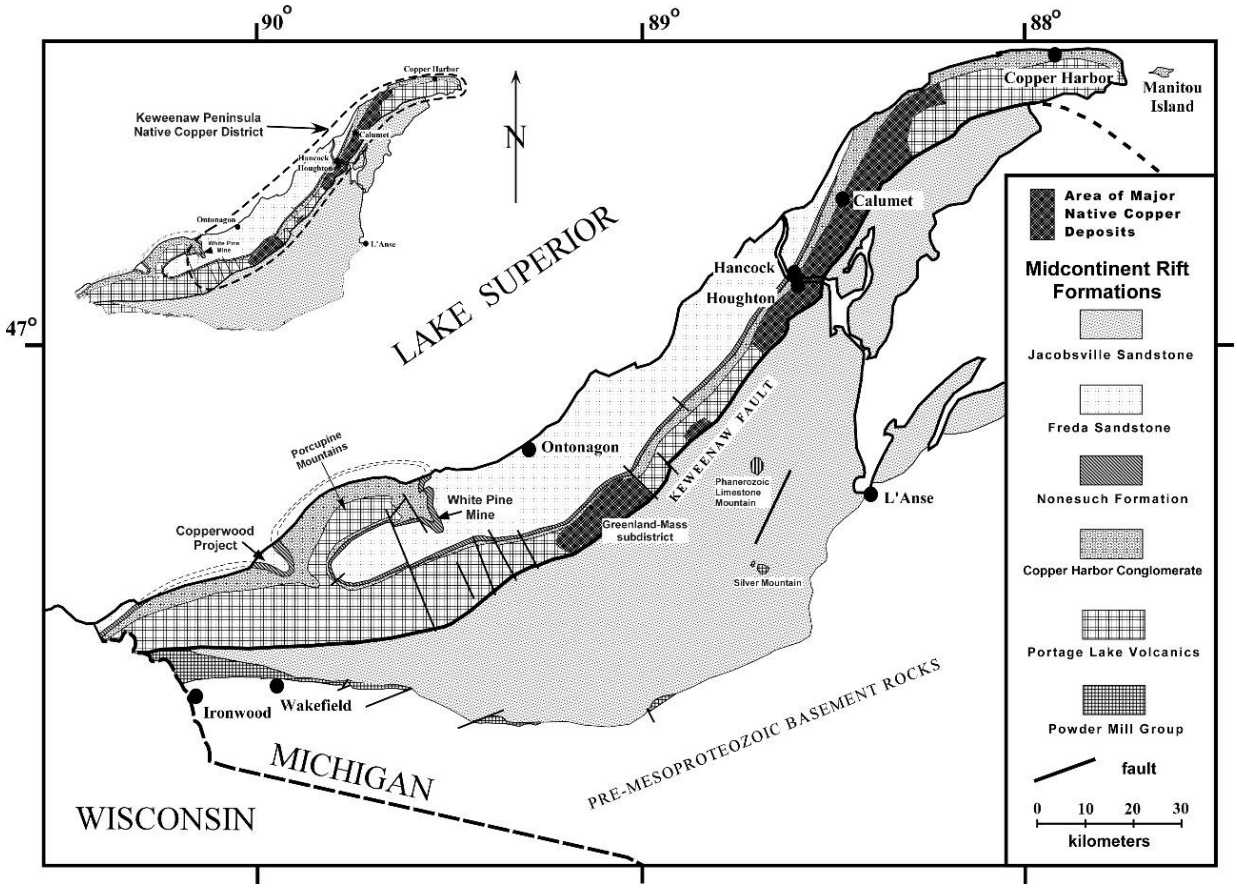


Figure 4: Bedrock geologic may of the western Upper Peninsula of Michigan showing area of the Keweenaw Peninsula native copper district (from Bornhorst and Barron, 2013).

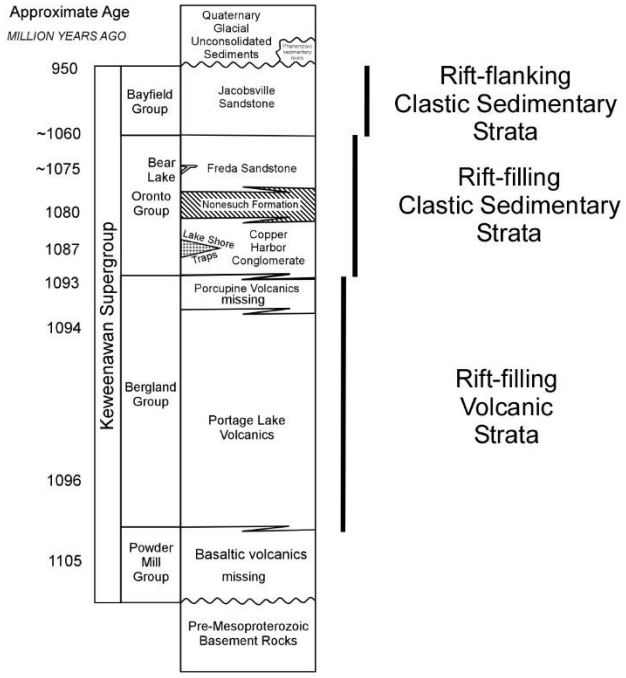


Figure 5: Stratigraphic column the Keweenaw Peninsula, Michigan.

Midcontinent Rift Strata

The Keweenaw Peninsula is located on the southern margin of the Lake Superior segment of the MCR (Figures 1and 2). The rock units that are associated with the MCR have been termed the Keweenawan Supergroup (Figure 5). These rocks were deposited about 1.1 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 (Figures 2 and 3) (Hinze et al., 1990; Cannon et al., 1989).

The MCR geology of the Keweenaw Peninsula can be divided into northwest-dipping, rift-filling volcanic and clastic sedimentary rocks under the central highlands and northwest flank of the Keweenaw Peninsula (Figure 4) and flat to low-dipping, rift-flanking clastic sedimentary rocks located on the southeast side. The Keweenaw Fault separates the rift-filling and rift-flanking strata. The rift-filling strata are subdivided into volcanic-dominated and clastic sedimentary rock-dominated lithologies. (Figure 5).

Portage Lake Volcanics

The Portage Lake Volcanics of the Keweenaw Peninsula (Figures 4 and 5) are 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 which are located stratigraphically near the base of the exposed formation. Interflow reddish-colored conglomerate and sandstone layers are less than 5 % by volume and are stratigraphically scattered throughout the Portage Lake Volcanics although 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 near 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 et al., 1997 and reference therein). Much of the Portage Lake Volcanics erupted over 2 to 3 million years from 1,096.2+/-1.8 (Copper City flow, Figure 6) to 1,094.0+/-1.5 (Greenstone flow, Figure 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. Magmatic differentiation after eruption is especially significant in 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 of the Portage Lake Volcanics is cyclical with minor and major cycles superimposed on an overall trend. The basalt magmas were derived by partial melting of sub-continental upper mantle with an overall compositional trend towards younger more primitive basalt compositions as a result of less crustal contamination (Paces, 1988; Paces and Bell, 1989). The repeated magmatism at the rift axis and progressive crustal thinning provided pathways for magma with less extended contact with crustal rocks. 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 enriched plume-related mantle (Nicholson et al., 1997; Nicholson and Shirey, 1990; Paces and Bell, 1989).

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. There is one thin hyaloclastic unit in the upper part of the formation (Johnson, 1985). 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 (Figure 6). The Greenstone flow has been correlated down dip across the Lake Superior syncline to Isle Royale (Longo, 1982; Huber, 1975).

Table 1: Average 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

The uppermost 5 to 20% of the tops of most individual lava flows are vesicular with between 5 and 50% vesicles (White, 1968). 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 are largely filled with secondary minerals, except for the stratigraphically uppermost lava flows; the filled vesicles are amygdules. Thus, local terminology is to call lava flows with vesicle-only tops, amygdaloids and those with brecciated tops fragmental amygdaloids.

There are 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 into the exposed Portage Lake Volcanics. The rhyolitic volcanic setting is analogous to the shield-type central volcanoes of Iceland (Nicholson, 1991).

Interflow clastic sedimentary rocks layers of the Portage Lake Volcanics are recognized as informal members since they are important stratigraphic markers in an otherwise monotonous succession of basalt lava flows. Many of them are given informal names (Figure 6). A few of them can be traced along strike for large distances, up to 90 km. These interflow sedimentary rock layers consist of red-colored conglomerates with lesser amounts of interbedded sandstone and occasional significant amounts of siltstone and shale. These informal members range in thickness from a few cm up to about 40 m (Merk and Jirsa, 1982; White, 1968; Butler and Burbank, 1927). The typical conglomerate is characterized by sub-rounded to angular pebbles in a sandy matrix. Clast size varies from granules 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, as 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 margins of the MCR toward the center of the rift basin (now under Lake Superior) (White, 1968).

Copper Harbor Conglomerate

The Copper Harbor Conglomerate is the oldest formation of rift-filling clastic sedimentary rocks and conformably overlies and interfingers with the top of the Portage Lake Volcanics (Figures 4 and 5). It consists of red-brown clastic sedimentary rocks with a maximum exposed thickness 2,000 m. Conglomerates and sandstones are the dominant lithologies in the Copper Harbor Conglomerate. 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 coalescing alluvial fans (Elmore, 1984). The diamictites are debris flow in origin. Sandstone interbeds are more common in the upper 2/3 of the formation. Sandstones are predominantly red-brown, subangular-to-

angular lithic graywackes with volcanic lithic fragments. The sandstones exhibit current-ripples, trough-cross beds, current and parting lineations, and reduction spots. Abundant calcite cement in select conglomerate and coarse sandstone layers was probably deposited as vadose carbonate or caliche (Kalliokoski, 1986). Thin red-colored siltstone and shale interbeds have desiccation cracks and are interpreted as periodic drying of the surface. In the Copper Harbor area, there are also laminated cryptoalgal carbonate beds and ooid lenses. 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 stromatolites (genus *Colleria*).

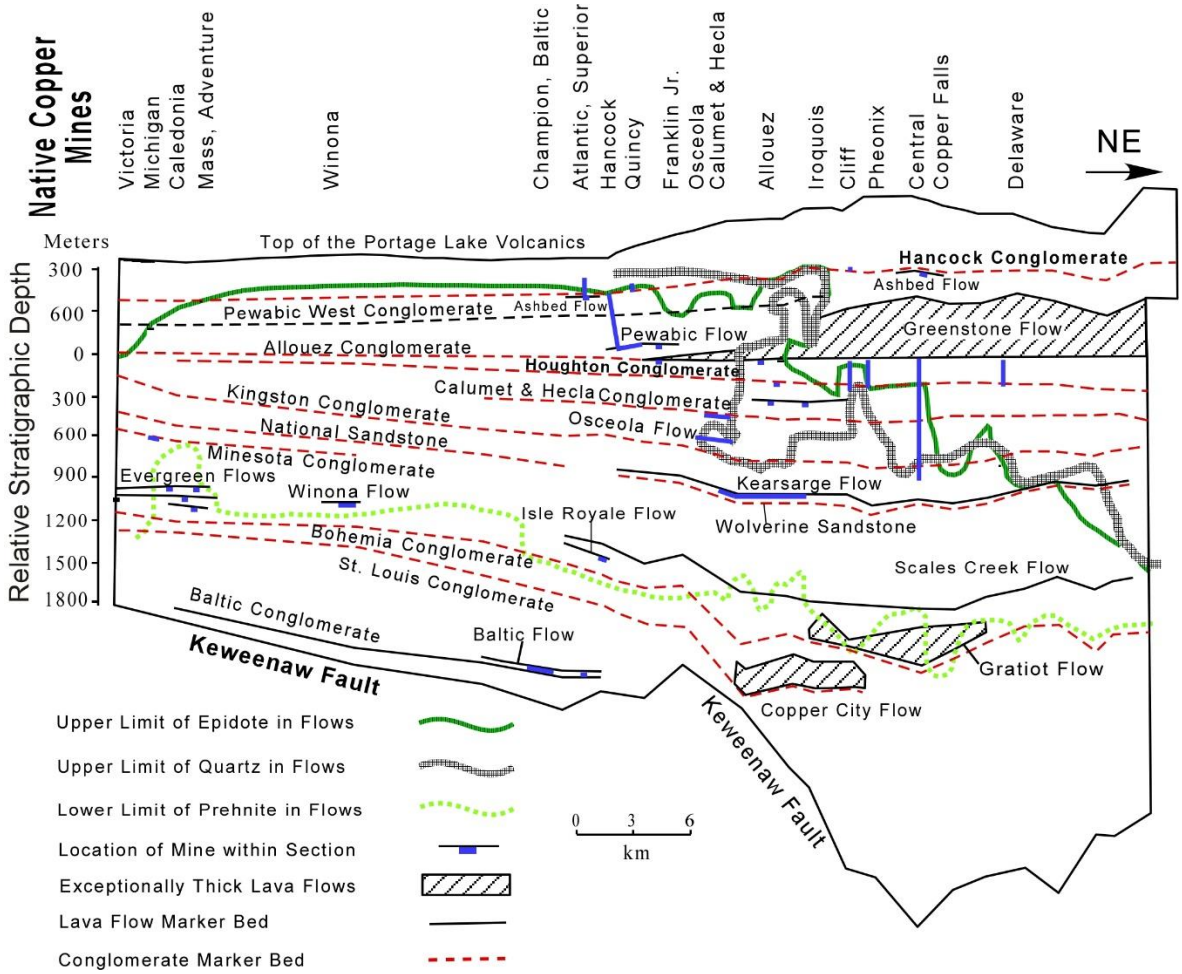


Figure 6: Generalized stratigraphy of the Portage Lake Volcanics in a strike parallel (longitudinal) section. Modified from Stoiber and Davidson (1959). Figure 4 shows location of Greenland-Mass subdistrict (Michigan, Caledonia, Mass, Adventure Mines).

For decades The Copper Harbor Conglomerate CHC and overlying Nonesuch Formation (Figure 5) have been interpreted by many geologists as non-marine. Elmore (1984) interpreted the environment as a prograding coalescing non-marine 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). However, a number of sedimentological features could be interpreted as either non-marine or marine and thus, non-marine interpretations often relied on other evidence (Jones et al., 2020). In the stromatolite interval, Jones et al. (2020) cite bimodal (herring-bone) transport

directions indicated by ripple marks that are mud draped and reactivated as evidence of a shallow marine environment. Hummocky cross stratification suggests waves on a marine shelf generated by storms (Jones et al., 2020). Periodic to rhythmic sedimentological features are indicative of "cyclical periodicity" of tidal deposition on a marine shoreline and are among evidence cited by Jones et al. (2020). Jones et al. (2020) conclude that the Copper Harbor Conglomerate and overlying Nonesuch Formation were deposited in a "braided fluvial-evaporitic shoreline-marine embayment" rather than fluvial-non-marine lacustrine setting. Geochemical evidence provided by Stüeken et al. (2020) also supports a marine estuary. The climate was probably arid with flashy seasonal streams. The highlands to the southeast from which the Copper Harbor Conglomerate was derived are now buried under the Jacobsville Sandstone.

The Copper Harbor Conglomerate in the Keweenaw Peninsula includes a succession of subaerially deposited lava flows. Lane (1911) used the name, Lake Shore Traps, for this informal member (Figure 5). This member is well exposed near the tip of the Keweenaw Peninsula where the unit is composed of 31 lava flows and one interflow conglomerate with a maximum thickness of about 600 m (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 passive subsidence with rift-filling clastic sedimentation and little to no magmatism except for the Lake Shore Traps. These subaerial lava flows range from Fe-rich olivine tholeiitic basalt at the base to Fe-rich olivine-bearing tholeiitic basaltic andesites and tholeiitic andesites and are likely a shield volcano. Geochemical data are best 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 Conglomerate (Figures 4 and 5). It consists of dominantly black-to-gray-to-green to red-gray siltstone and shale with a maximum thickness 240 m. Bornhorst and Williams (2013) provide a stratigraphic column of the entire Nonesuch Formation just south of the Porcupine Mountains State Park from exploration drilling. Exposures of the Nonesuch Formation in the Keweenaw Peninsula proper are limited with the best exposure at the Hancock campground and boat launch on M-203 (Stop 14). There are excellent exposures of the Nonesuch Formation along the Big Iron and Presque Isle rivers in the White Pine area (Woodruff et al., 2013).

In areas with thicker stratigraphic section, 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, 2013). Well-laminated to massive black to dark-gray siltstone and shale are the dominant lithologies near the base of the Nonesuch Formation. The base of the Nonesuch Formation hosted 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, 2013; Williams and Bornhorst, 2023). A thin carbonate laminate yielded a Pb-Pb isochron age of 1,081 ± 9 Ma (Ohr,1993). The environmental setting of the Nonesuch is described above under the Copper Harbor Conglomerate.

Freda Sandstone

In the Keweenaw Peninsula the Freda Sandstone is the youngest rift-filling clastic sedimentary rock formations (Figures 4 and 5). The contact between the lower most Freda Sandstone and Nonesuch Formation is gradational (Bornhorst and Williams, 2013). The exposed thickness is greater than 3,700 m, with the top of the formation submerged beneath Lake Superior. The Freda Sandstone is generally poorly exposed except along the Lake Superior shoreline. This field guide provides an optional stop 13 at the McLain State Park where exposures of the Freda are visible during times of low levels of Lake Superior. Angular and tabular specimens are obtainable at the beach from outcrops just offshore. The last MCR magmatism was Bear Lake, an intrusive-extrusive dome of alkaline trachyandesite was emplaced near the middle of the exposed Freda Sandstone (Kulakov et al., 2018).

Red-brown fine to very-fine sandstone, siltstone, and mudstone are the dominant lithologies in the Freda Sandstone. Fining-upward sequences occur on the scale of a few meters. The Freda Sandstone was deposited in an environment characterized by shallow meandering streams (Daniels, 1982). Based on regional correlations the Freda was likely deposited between 1,080 to 1,060 Ma.

Jacobsville Sandstone

The Jacobsville Sandstone was deposited in a rift-flanking basin (Figure 3D) and is outside the scope of this field guide. Its stratigraphic relationship with other formations is not determined. It occurs in a contiguous geographic region bound on the northwest side by the Keweenaw Fault and on the southeast by an unconformable contact with Paleoproterozoic and Archean basement rocks (Figure 4). The Jacobsville Sandstone is estimated to be more than 2,900 m thick and the top is not exposed (Kalliokoski, 1982). 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). The Jacobsville Sandstone was deposited in an environment characterized by shallow meandering streams (Kalliokoski, 1988).

Faults, Folds, Fractures

The last episode of the Midcontinent Rift was characterized by post-rift compressional inversion that facilitated hydrothermal formation of native copper deposits (Woodruff et al., 2020; Bornhorst, 1997). This compression transformed original normal faults into reverse faults, reactivated other extensional rift-related faults/fractures, and produced new compression-only faults/fractures and folds. Rather than being an inverted rift-related normal fault, the Keweenaw fault was likely a detached thrust (DeGraff and Carter, 2023). 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 (Figure 1) beginning as early as 1.08 Ga (Cannon, 1994; Cannon and Hinze, 1992; Hoffman, 1989). Final inversion of the MCR by compression during Grenville orogeny occurred between 1,010-980 Ma (Hodgin et al., 2022).

The Mesoproterozoic Midcontinent Rift-filling strata of the Keweenaw Peninsula dip moderately toward the center of the rift with the angle of dip increasing toward Keweenaw fault where the stratigraphic base is truncated (Figure 7). The dip of the strata is interpreted as a combination of syn-depositional downwarpage and structural tilting in response to reverse faulting caused by regional continental compression (Woodruff et al., 2020; Cannon, 1994).

There are many faults/fractures in the Mesoproterozoic rocks of the Keweenaw Peninsula. Some of these were exclusively formed during extension of the Midcontinent Rift when graben-bounding normal faulting was prominent along the margin (Figure 3). However, most faults/fractures were likely either reactivated by or directly produced by the regional compressional event. The Keweenaw Fault strikes and dips more or less parallel to the bedding of the truncated Portage Lake Volcanics (Figure 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). It is a detached thrust fault related to regional compression. 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. However, the rocks within and adjacent to the fault are altered by late-stage hydrothermal fluids.

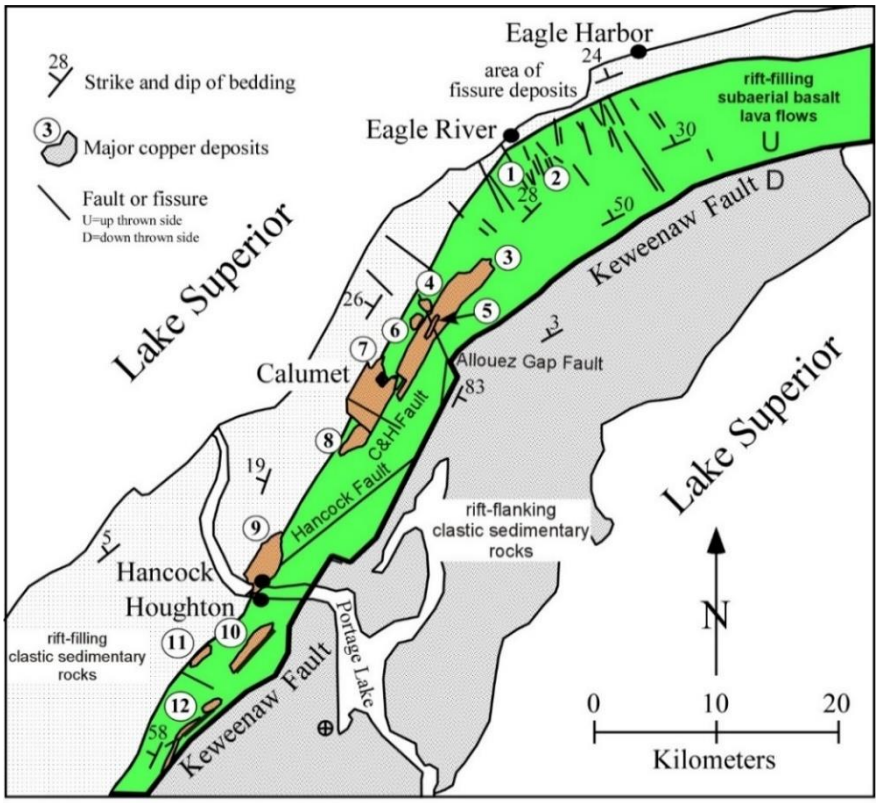


Figure 7: Simplified geologic map showing the location of the major deposits within the Keweenaw Peninsula native copper district, Michigan. Table 2 provides the names and production for the numbered deposits. The areas shown on the map are the mined out down-dip portion projected to the surface. All 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 see 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**	

Several faults occur oblique to the strike of bedding. In the Eagle River area, fault-controlled native copper veins are common in association with high-angle faults whose displacement is from 0 to 200 m, (Figure 7; see also Figure 19; Butler and Burbank, 1929). The Allouez Gap fault (Figure 7) bisects the largest lava flow top hosted native copper deposit in the district (see Figure 16) and was likely a significant conduit for native copper mineralizing hydrothermal fluids (Bornhorst, 1997). Correspondence between the thickness of the Kearsarge lava flow and the Allouez Gap fault suggests this fault was active during deposition of the Portage Lake Volcanics. It is interpreted as having been reactivated during regional compression.

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 as well as those in the Greenland-Mass subdistrict (Broderick, 1931). Faulting occurred before, during, and after deposition of native copper and its associated alteration minerals based on fault brecciated and recemented alteration minerals. There is a close relationship between faulting/fracturing produced by or reactivated by compression and native copper deposits. The compressional structures acted as pathways for mineralizing hydrothermal fluids (Bornhorst 1997).

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 late regional compression (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 ~20 billion lbs of copper (Bornhorst and Barron, 2011). Small quantities of native silver are temporally and spatially associated with the native copper. The major ore producing horizons are located in a 45 km-long belt in the Keweenaw Peninsula (Figures 4 and 7) and in a subdistrict to the southwest. Native copper and silver were the only economic metallic minerals and were co-precipitated with a suite of nonmetallic alteration minerals (Figure 8). Sulfide minerals, such as chalcocite, are uncommon in native copper deposits and when present only occur in trace amounts. Sulfide minerals occur in late-stage veins (Figure 8). 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 (Woodruff et al., 2020; 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 hosted by the Portage Lake Volcanics where there is 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 which 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/ 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 (Table 2).

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 barren massive basalt of the same flow as the mineralized flow top and hanging wall interior of 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 amounts of copper. As brecciated/fragmental amygdaloidal transitions 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 a 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 elongated, 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 on the incline below the surface (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 production of refined copper were hosted by them. These deposits were 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 overlying lava flow. The porosity of underlying brecciated/fragmental amygdaloid lava flow top is often greatly decreased by silt and sand filling the primary open space between fragments of the flow top hence, the originally porous flow top underlying a conglomerate bed acts more like an aquiclude in the paleohydrologic hydrothermal system. Native copper tends to be concentrated along specific stratigraphic bands within the conglomerate that are 0.5 to 5 m thick (Weege et al., 1972).

The Calumet and Hecla Conglomerate was by far the largest single native copper deposit in the district producing 4.2 billion lbs. as compared to the next largest deposit, the Kearsarge flow top which produced 2.3 billion lbs. from a fragmental amygdaloid (Figure 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; 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 in the clastic sedimentary host rock 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. However, overall, the vein deposits are of slight economic importance in the district. 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 anastomosing filled open spaces. While brecciation within vein deposits is common, gouge is not present (Butler and Burbank, 1929). The lava flow tops and conglomerates adjacent to the vein 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 occurs as both finely disseminated and as masses weighing many tons. The grade of native copper in the veins has the nugget-effect making determination of grade difficult. Several small vein deposits are localized just beneath the thickest basalt flow in the district, the Greenstone flow. For these veins the 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 faults (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 Figure 16). 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 also likely important for upward transport of ore fluids. 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 (Bornhorst, 1997).

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 area of major native copper deposits rocks were less altered at lower temperature. The intensity and degree of alteration also varies as a function of position within lava flows; the massive interiors of lava flows are 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 some flows, but secondary minerals exist in the massive interiors of all flows regardless of their thickness. While the thicker massive interiors of lava flows contain secondary minerals, their original igneous geochemical composition is often only slightly or essentially not modified by secondary hydrothermal processes.

There are more than 50 different secondary alteration minerals in the Keweenaw Peninsula; most of them are related to hydrothermal processes and some are related to supergene processes. Only about 20 alteration minerals are major to less common minerals (Figure 8). 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 of 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).

There is a district wide temporal (paragenetic) and spatial variation in the assemblage of alteration minerals which was first well described by Butler and Burbank (1929) and later summarized by White (1968). Recently Bodden et al. (2022) have refined the paragenetic and spatial variation of the hydrothermal minerals (excluding igneous and supergene related minerals) (Figure 8). The hydrothermal alteration minerals can be subdivided into main-stage which paragenetically overlap with the precipitation of native copper (Figure 8). While district-wide there is a well-defined mineral paragenesis, individual deposits may not exactly follow the district-wide timing of precipitation (compare Figure 8 to Stop 5). The main-stage is interpreted by Bodden et al. (2022) as formed during a continuous hydrothermal event.

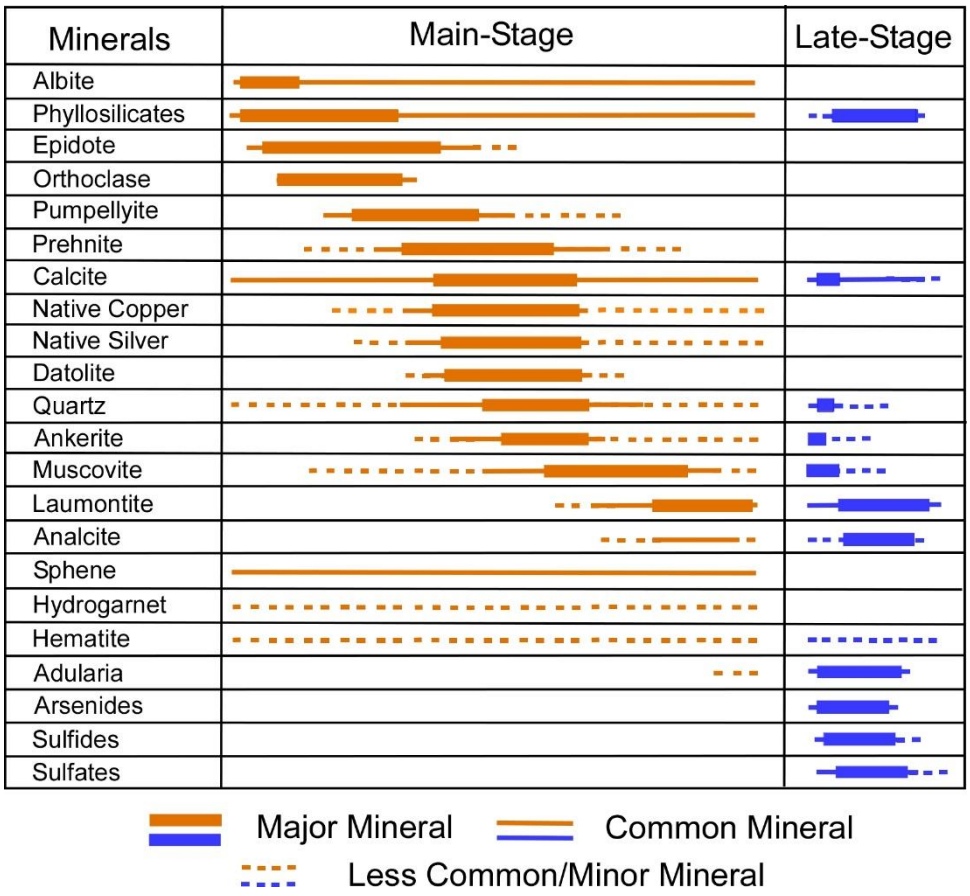


Figure 8: Paragenesis and relative abundance of secondary hydrothermal alteration minerals in the Keweenaw Peninsula native copper district. After Bodden et al. (2022).

The late-stage minerals are widespread but volumetrically minor. They commonly occur in small veins/fractures which cross-cut the main stage minerals or as coatings on main-stage vug filling minerals. Late-stage alteration minerals are notably more abundant near the Keweenaw fault. The suite of late-stage minerals are distinguishable by the occurrence of sulfur-bearing minerals, sulfides and sulfates, and by an assemblage of lower temperature minerals in areas where they overprint an assemblage of main-stage minerals formed at higher temperatures. The timing of the late-stage hydrothermal event is uncertain. There could have been no time break or a major time break between the main-stage and late-stage hydrothermal events. Bodden et al. (2022) suggested that the main-stage and late-stage hydrothermal events are practically continuous with each other.

Main-stage alteration minerals are spatially zoned perpendicular to stratigraphic strike as demonstrated for the Calumet area of the district (Figure 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 (Stop 5). The alteration mineral zones of the Portage Lake Volcanics are similar to the North Shore Volcanic Group of Minnesota (Schmidt and Robinson, 1997). Bodden et al. (2022) mapped the occurrence of

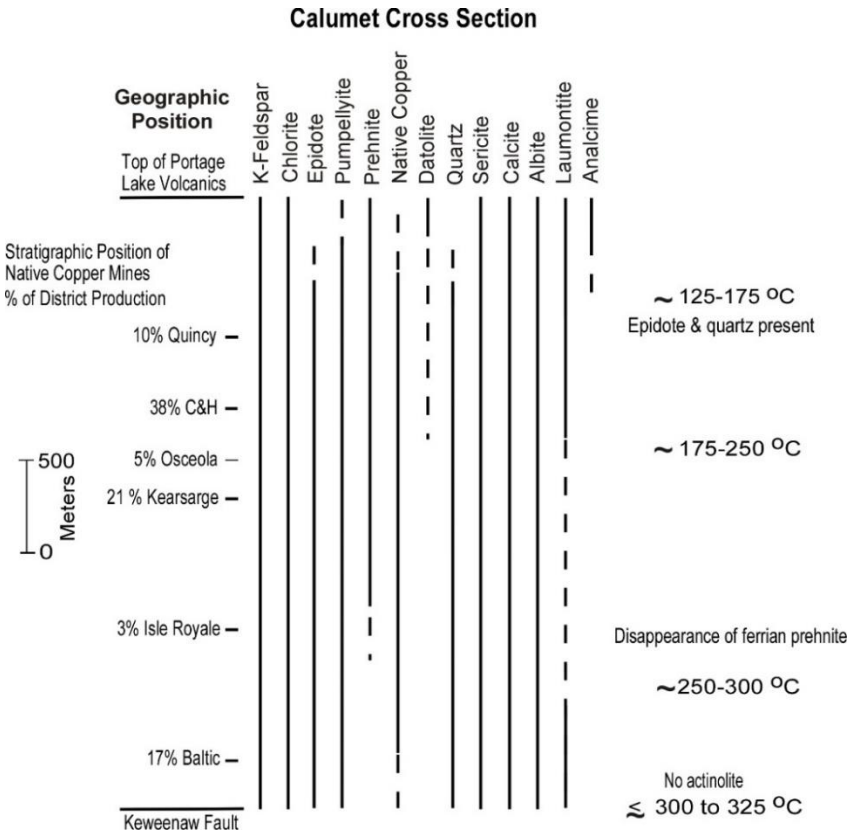


Figure 9: Distribution of prominent secondary hydrothermal alteration minerals in the Portage Lake Volcanics in a cross-section in vicinity of Calumet at the center of the major deposits of the Keweenaw Peninsula native copper district modified from Bornhorst and Rose (1994).

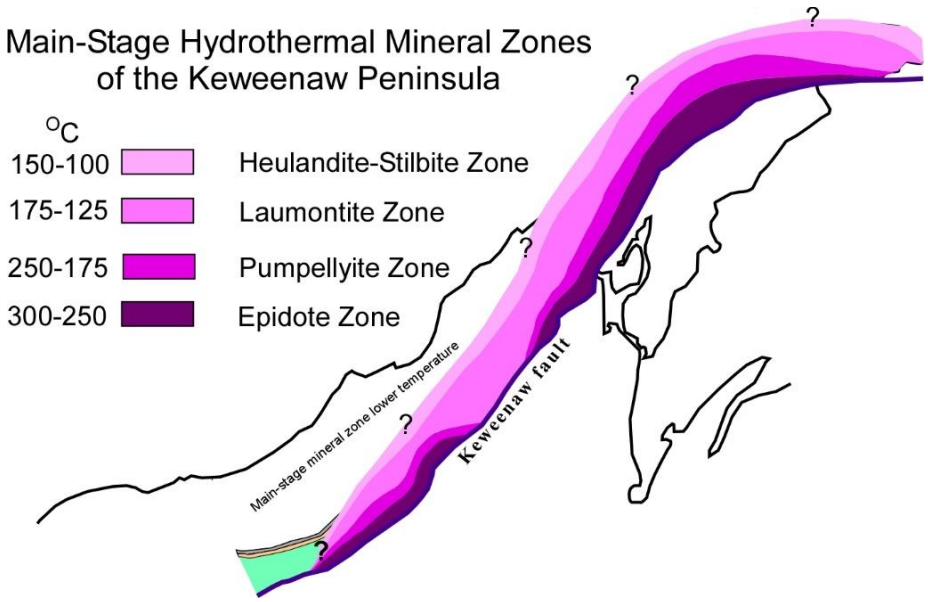


Figure 10: Main-stage hybrid metamorphogenic hydrothermal mineral zones of the Keweenaw Peninsula (modified from Bodden et al., 2022)

alteration minerals of the Keweenaw Peninsula into those zones used for the North Shore Volcanic Group (Figure 10; Schmidt and Robinson, 1997). These zones can be equated to the temperatures of mineral formation (Bodden et al., 2022). The spatial zoning of alteration minerals is consistent with a thermal high associated with the major native copper deposits (compare Figure 4 and 10). The mineral zones dip more gently towards Lake Superior than the strata, implying that the strata were at least somewhat tilted prior to main-stage hydrothermal alteration (Livnat, 1983; Broderick, 1929).

Native copper mineralization is younger than the Copper Harbor Conglomerate, which hosts rare veins of calcite and native copper (see Stop 7). White (1968) interpreted the age of native copper mineralization as after the deposition of parts or all of the Freda Sandstone. The Bear Lake igneous body within the Freda Sandstone is native copper mineralized. Minor amounts of native copper occur within the lower beds of the Jacobsville Sandstone near Rice Lake (Calumet and Hecla unpublished drill core log). 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 1050 Ma (+/- ~ 20 Ma) (Bornhorst et al., 1988). This age is consistent with the approximate age of 1060+/-20 Ma for regional continental compression that caused thrust faulting along the Keweenaw Fault (Cannon et al., 1993). Thus, the age of main-stage hydrothermal alteration is about 1060 to 1050 Ma contemporaneous with regional continental compression and some 30 million years after eruption of the Portage Lake Volcanics.

Genesis of the Main-Stage Native Copper Deposits

This section is summarized from Bornhorst and Mathur (2017) and Bodden et al. (2022) and illustrated in Figure 11.

Native copper occurs throughout the MCR in Wisconsin, Minnesota, and Ontario (Figure 2). It formed during a regional hydrothermal event from about 1060 to 1050 Ma (Bornhorst et al., 1988). The regional Cu-bearing hydrothermal fluids are best explained as generated during burial metamorphism of rift-filling basalts with temperatures reaching a thermal maximum approximately 30 million years after the end (~1085 Ma) of widespread rift magmatism. The suite of main-stage hydrothermal minerals precipitated, except native copper, (Figure 7 and 8) is similar to those found elsewhere rocks have undergone very low to low grade burial metamorphism at less than about 300^{O}C. The coincidence of regional continental compression with a burial thermal maximum (Woodruff et al., 1995) 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 burial metamorphogenic hydrothermal ore fluids, copper was leached at depth from the rift-filling basalt strata (Bornhorst and Mathur, 2017, 2018). More than sufficient amount of copper was available to have been leached from the buried rift-filling basalts.

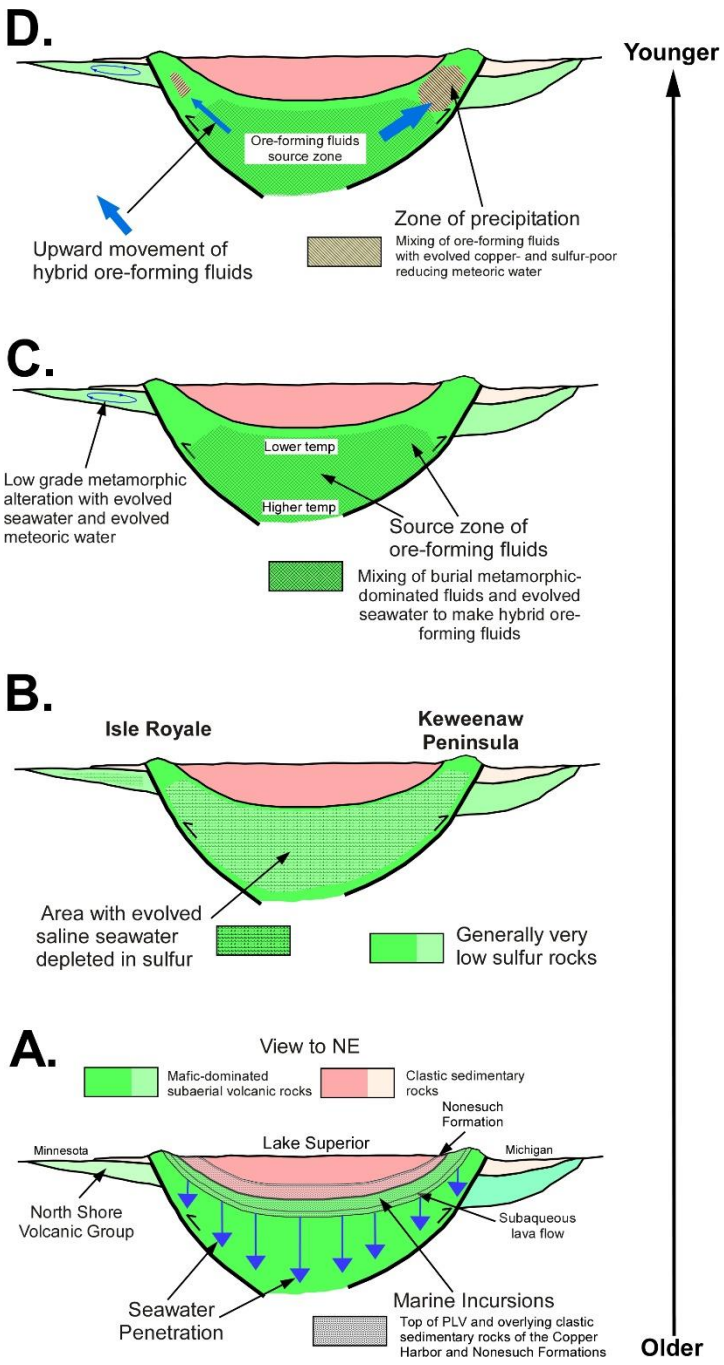


Figure 11: Cartoon cross sections showing conceptual genetic model of the native copper deposits of the Keweenaw Peninsula formed at about 1070 to 1040 million years ago. Modified from Bodden et al. (2022). A. Marine incursions and seawater penetration during deposition of volcanic and sedimentary rocks in MCR. B. Area prior to burial metamorphism with sulfur depleted evolved seawater providing salinity for ore-forming fluids. C. Burial metamorphism with generation of burial metamorphic hydrothermal fluids. Mixing the burial metamorphic fluids with evolved seawater produces copper-bearing hybrid metamorphogenic ore-forming hydrothermal fluids. D. Precipitation of main-stage minerals as a result of mixing of the ore-forming fluids with meteoric water, decreasing temperature, and water-rock reactions.

The rift-filling volcanic rocks were low in sulfur when they erupted and the little available sulfur was degassed prior to solidification (Bodden et al., 2022; Bornhorst and Mathur, 2017). These low-sulfur rift-filling volcanic rocks were buried into the source zone where burial metamorphogenic hydrothermal fluids were generated (Figure 3 and 11C). Since the very low sulfur rift-filling basalts in the source zone were the same as those in the fluid pathways to the zone of precipitation, the fluids remained sulfur poor. Since the native copper ore host rocks were again the same rift-filling very low sulfur volcanic rocks, the burial metamorphogenic hydrothermal fluids remained depleted in sulfur. The lack of sulfur resulted in the precipitation of native copper rather than copper sulfides.

While the metamorphogenic hydrothermal fluids lacked sulfur, several studies (Kelly, 2022; Kelly, 2020; Püschner, 2001; Brown, 2006; Livnat, 1983; Jolly, 1974) have suggested that the main-stage hydrothermal fluids had at least moderate degree of salinity. The possible sources of salinity were evaluated by Bornhorst and Mathur (2017), Bornhorst (2021), and Bodden et al. (2022). Viable sources of salinity for the hydrothermal fluids need to also satisfy the constraint of very low sulfur. Bornhorst (2021) hypothesized that evolved formation water derived from sulfur depleted seawater could have been the source of salinity. As seawater penetrates mid-ocean ridge basalts it is heated, reacts with the host basalt, and as a result of precipitation of minerals it becomes depleted in sulfur. If the sulfur content of the Mesoproterozoic seawater was low in sulfur, as proposed by Blattlet et al. (2022), then less depletion of sulfur would have been needed.

During deposition of the youngest Portage Lake Volcanics and overlying Copper Harbor and Nonesuch formations there were probable incursions of an arm of the sea into the rift for a significant amount of time. This could have resulted in seawater deeply penetrating into the underlying rift-filling volcanic rocks (Figure 11A; Bornhorst, 2021). During burial, the rift-filling volcanic and clastic sedimentary rocks and contained seawater was progressively heated and thereby evolved to be depleted in sulfur (Figure 11B). Continued heating during burial resulted in burial metamorphic hydrothermal fluids which then thoroughly mixed with the evolved seawater to form a hybrid metamorphogenic-dominated ore-forming fluid (Figure 11C; Bodden et al., 2022). These main-stage ore-forming hydrothermal fluids moved upwards from the source zone through the same very sulfur poor strata as in the source rocks (Figure 11D). As they moved upwards they cooled, interacted with host rocks, and in the relatively shallow zone of precipitation they variably mixed with sulfur-poor, low salinity, reduced meteoric water (Figure 11D; Bodden et. al, 2022). These processes resulted in precipitation of native copper and main-stage hydrothermal minerals. Higher temperature main-stage mineral assemblages are spatially associated with the area of native copper deposits where the thermal anomaly was greatest because of focused hydrothermal fluids (Figure 10). The possible depth of the zone of precipitation is poorly estimated with a best guess at this time of 10 to 15 km (Kelly et al., 2022; Kelly, 2020). Within the native copper district, the suite of main-stage minerals, including native copper, is followed by late-stage minerals precipitated at lower temperatures than the main-stage hydrothermal fluids coincident with the native copper district.

Late-Stage Hydrothermal Minerals

The suite of late-stage minerals is widespread and similar throughout the Keweenaw Peninsula. The late-stage suite is readily distinguished in the main area of the native copper district since late-stage minerals are lower temperature (100 to 150^{O}C) than the main-stage minerals in district itself. However, outside of the native copper district where main-stage minerals are expected to be formed at lower temperature, the main-stage and late stage are indistinguishable. Bodden et al. (2022) suggested that late-stage fluids are a variable mixture of hybrid metamorphogenic hydrothermal fluids, meteoric water, and shallow seawater, the latter being a source of sulfur in the late-stage fluids.

Phanerozoic

The Keweenaw Peninsula was subjected to a 500-million-year period of erosion, from about 1 Ga to 0.5 Ga (500 Ma) and multiple kilometers of rock were eroded exposing the native copper deposits at the surface (Figure 3). Downward percolating groundwaters supergene altered native copper and produced a suite of including cuprite, tenorite, malachite, and chrysocolla (Bornhorst and Robinson, 2004). The rocks of the Keweenaw Peninsula were subsequently buried by Paleozoic sedimentary rocks associated with the Michigan basin beginning about 500 Ma (Figure 3) and ending Precambrian supergene alteration.

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 Paleozoic outliers slightly south of the Keweenaw Peninsula (Figure 3). The last glacial episode exposed the native copper deposits at roughly the same erosional level as at 500 Ma or the end of the Precambrian (Bornhorst and Robinson, 2004). 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 submerged under 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.

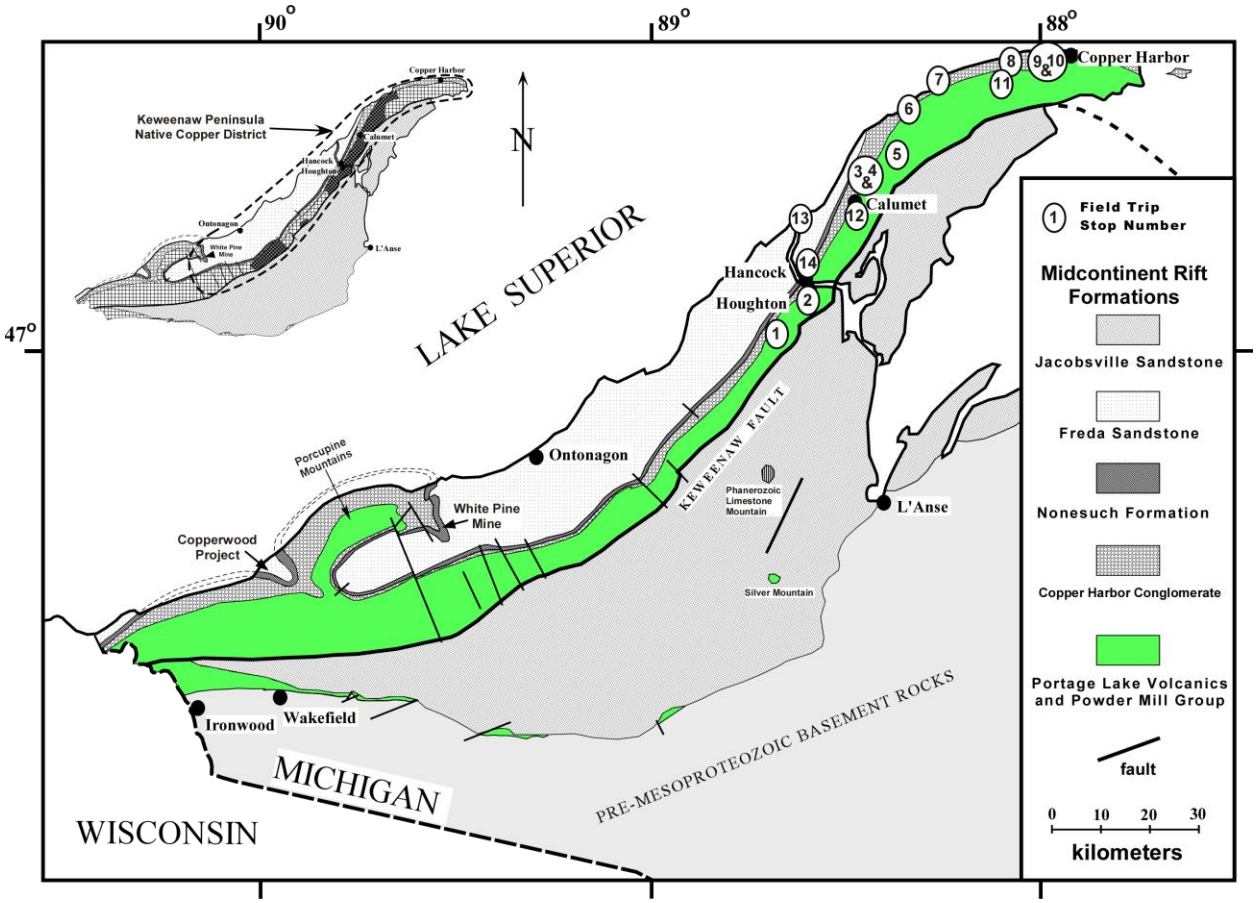


Figure 12: Geologic map of the far western part of the Upper Peninsula of Michigan showing field trip stops.

Objectives of Field Trip

This field trip is designed to provide an overview of the Mesoproterozoic Midcontinent Rift-filling strata and native copper deposits of the Keweenaw Peninsula (Figure 12). There are four rift-filling formations: Portage Lake Volcanics, Copper Harbor Conglomerate, Nonesuch Formation, and Freda Sandstone. The Nonesuch and Freda formations are poorly exposed in the Keweenaw Peninsula thus, only two optional stops are included in this field guide. The Jacobsville Sandstone is a rift-flanking formation and is outside the scope of this field trip. The rift-filling strata are overlain by unconsolidated Pleistocene glacial sediments. There is one glacial related stop.

IMPORTANT NOTE TO ALL READERS:

Many of the field trip stop descriptions and significant parts of the introductory geologic overview have been previously published especially in other Institute on Lake Superior guidebooks e.g., Bornhorst and Barron (2013) and the extensive guides by Bornhorst and Rose (1994) and Bornhorst et al. (1883). The stop descriptions in this field guide, as compared to previously published guides range from exact wording to significantly modified wording without specific citation. Two of the stops in this field guide, Stops 2 and 8, were not visited by previous field guides involving Bornhorst. These stop descriptions are new.

Stop 1: Subaerial basalt lava flow cross-section at South Range Quarry

Latitude: 47.07750N; Longitude: -88.64240W

Directions: Drive west through downtown Houghton on US-41 to south M26. Drive about 4.5 miles to unmarked road on west side of M-26 just before church on south side of unmarked road. Proceed on road to tree line. Walk NE uphill to quarry.

THIS STOP IS ON PRIVATE PROPERTY. PLEASE GET PERMISSION TO ENTER PROPERTY.

Volcanic textures and structures typical of moderate-to-thick subaerial lava flows within the Portage Lake Volcanics are well exposed in this old quarry (Figure 13). As one traverses up the hill to the quarry along the rubbly path there is a low-profile exposure of a 4 m thick interflow conglomerate bed which is also exposed laterally along the SE slope face of this knob. The conglomerate layer is stratigraphically the National Sandstone member and is approximately near the middle of the exposed Portage Lake Volcanics stratigraphic section (Figure 6). The conglomerate is overlain by an 18 m thick lava flow (A).

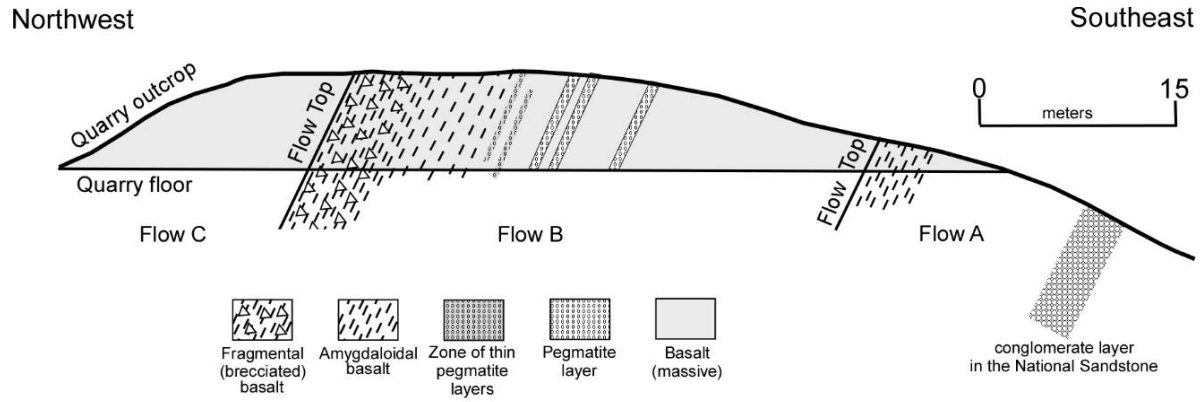


Figure 13: Geologic cross section of the South Range Quarry (modified from White, 1971).

Laterally continuous interflow sedimentary beds provide critical stratigraphic markers within the Portage Lake Volcanics, an otherwise monotonous volcanic pile with many laterally discontinuous lava flows. The sedimentary unit exposed below the quarry has been correlated with the National Sandstone, a marker bed in the Mass-Rockland area (Figure 6). At South Range Quarry, the National Sandstone is a massively bedded, pebble-cobble framework conglomerate, composed of silicic with subordinate mafic volcanic clasts that are subangular to subrounded, within a matrix of poorly-sorted medium-to-coarse sand of similar composition.

The Portage Lake Volcanics basalts in this part of the stratigraphic section are mainly olivine tholeiites and erupted as subaerial lava sheets. The principal lava flow exposed in the quarry walls illustrates many of the volcanological features of Portage Lake Volcanics lava flows. The top and bottom of the South Range Quarry lava flow (B) are exposed. Flow B was deposited directly on top of Flow A and consists of aphanitic chilled basalt. The base of Flow B occurs where amygdules disappear abruptly in the top of the underlying flow.

The upper surface of the main flow (B) was brecciated slightly by movement of lava after the formation of an upper crust. The flow top breccia (locally termed fragmental amygdaloid) is laterally discontinuous. The fragmental amygdaloid rapidly grades downward to an unbrecciated, highly amygdaloidal (vesicular) flow top. Note the variation in vesicle size and shape downward in the flow. There are numerous layers of flattened amygdules (vesicles) in the flow top with their orientation parallel to the top and bottom of the flow (B). The orientation of these layers may represent laminar flow planes within the flow top. The section was tilted after emplacement.

Slow cooling of the lava flow caused solidification toward the flow interior at a rate which allowed development of subophitic to ophitic textures (ophitic texture denoted by 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 (B). Before final solidification, small amounts of volatile-rich, differentiated residual liquid were likely injected into thin discontinuous layers and lenses (tabular openings produced during cooling). Most of these pegmatoid layers are subparallel to the bottom and top surfaces of the flow.

A typical pegmatoid zone consists of a 5 to 10 cm border zone at the top and bottom which is composed of a medium-to-coarse grained aggregate of Ab-rich plagioclase, prisms of Fe-rich clinopyroxene and abundant Fe-Ti oxides, as well as accessory minerals such as apatite and zircon (Cornwall, 1951c). The cores of the pegmatoid zones are 5 cm to 1.2 m thick consisting of a green vesicular basaltic rock. Zircons extracted from pegmatoids within thick Portage Lake Volcanics basalt flows have yielded high-precision U-Pb dates (e.g., Davis and Paces, 1990).

There are thin tabular layers and flattened amygdules (vesicles) in the top of the flow and in the massive interior of the flow (B) that are composed of a brown and sometimes green fine-grained material that have been described interpreted by White (1971) as detrital material. Alternatively, this green-to-red cherty rock, could be simply alteration minerals (quartz, prehnite, epidote, and pumpellyite) filling fractures and nearly complete pseudomorphic replacement of basalt. Numerous pegmatoid layers are exposed in the quarry walls, as well as in the glacially-polished surfaces above and to the north of the quarry.

The effects of regional hydrothermal alteration can be observed within the vesicular flow top and pegmatoid zones. Vesicles are filled with a variety of secondary minerals including quartz, epidote (olive green), prehnite (waxy light green), calcite, pumpellyite (pale bluish green), chlorite (dark green to black) and traces of native copper (pinkish color). Pseudomorphic replacement of basalt by fine-grained secondary minerals is most intense where permeability was highest. The massive interior of the flow is only a little visibly altered, however plagioclase is altered to albite and pyroxene is altered to chlorite. In the vicinity of selected fractures there can be intense epidote or prehnite alteration. The massive interior was a relatively impermeable horizon in the paleohydrothermal system. Fracturing during late compression integrated the system and provided limited pathways for upward movement of ore fluids.

Stop 2: Subaerial basalt lava flow, eastbound US-41 Houghton

Latitude: 47.12144N; Longitude: -88.56474W

Directions: Drive west through downtown Houghton and loop around (Yooper Loop) to head east for 0.8 miles along US-41 or Montezuma Avenue (one-way two-lane highway). Stay left (north side) as though going back through downtown and turn into unmarked gravel lot just before large outcrops on left (north). Use sidewalk to walk to outcrop. BE CAREFUL OF TRAFFIC.

The rock cut at the southeast end of downtown Houghton provides an excellent example of the characteristics of Portage Lake Volcanics subaerial basalt lava flows (Figure 14). There is a sidewalk providing access to the Stop 2 south-facing road cut. While there is also a sidewalk on the other side of the road crossing the road is discouraged. These other outcrops can be accessed by parking uphill from them on the other side of the road. There are also north-facing exposures of this same stratigraphic interval on west bound US-41 (Shelden Avenue) east of the Houghton U.S. Post Office. The exposures at Stop 2 are located stratigraphically above the Scales Creek flow and below the Kearsarge flow, slightly closer to the Scales Creek flow (Figure 6). The lava flows of the exposure and vicinity strike approximately N30^{o}E and dip toward the center of the rift (Lake Superior) at about 55^{o} northwest (White, 1956).

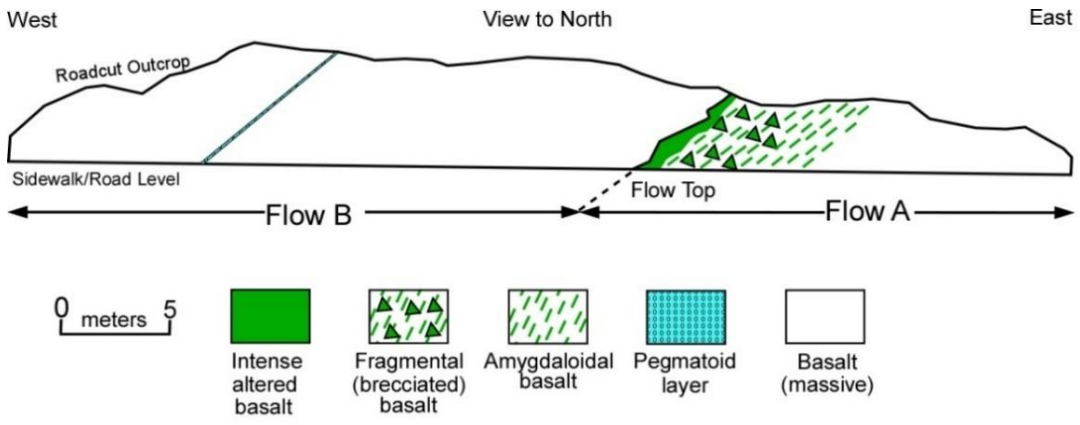


Figure 14: Geologic cross section of road cut eastbound US-41, southeast end of downtown Houghton.

There are parts of two lava flows exposed at Stop 2 (Figure 14). The base of the stratigraphically lower of these two lava flows (A) is not exposed on the east end of the Stop 2 exposures. It is also not exposed on the other side of the road. The top of Flow A is well exposed and is overlain by massive basalt of Flow B. The contact itself is denoted by an abrupt change from underlying highly altered greenish, slightly brecciated amygdaloidal basalt of Flow A (about 0.5 m thick) that is overlain by massive dark grey to black basalt of Flow B. This greenish zone of the top of Flow A is underlain by about 3.5 m of amygdaloidal basalt with slightly fragmental (brecciated) basalt, also Flow A. The slightly fragmental basalt is gradational downward (east) towards the center of Flow A where the abundance of amygdules (filled vesicles) is sufficient to call the rock amygdaloidal basalt. Amygdaloidal basalt lacking fragments is about 4 m thick. There is an arbitrary boundary where the

abundance of amygdules is too low to call the rock amygdaloidal, although it contains some amygdules. The abundance of amygdules progressively decreases towards the center of Flow A, the porous and permeable top of Flow A is about 8 m thick. Flow A is greater than 14 m thick as its base is not exposed.

Notably, flattened amygdules (vesicles) occur along planar layers in the flow A top with their orientation roughly parallel to the top of the flow. Larger flattened amygdules are about 2 by 2 by 1 cm and between them, appearing to connect larger amygdules, are much smaller amygdules 1 to 2 mm thick. The orientation of these layers may represent laminar flow planes within the flow top. Layers of flattened amygdules are also numerous at Stop 1.

Flow A is overlain by Flow B and the top of Flow B is not exposed at Stop 2. However, its top is poorly exposed on the other (south) side of the road. Flow B is about 30 m thick. There is a pegmatoid layer in the massive basalt interior of Flow B (Figure 14). Flow B is thicker than the average flow. The pegmatoid is distinguished as notably amygdaloidal. Pegmatoids are discussed further at Stop 1.

The effects of regional hydrothermal alteration can be observed within the top of Flow A and the pegmatoid layer in Flow B. The massive basalt interior of Flow A and B are much less altered than the flow top. However, the primary magmatic plagioclase in the massive basalt has been replaced by albite and the primary mafic minerals are replaced by chlorite, pumpellyite, and iron oxides. The intensely altered basalt at the very top of Flow A is largely replaced by hydrothermal alteration minerals including epidote, prehnite, pumpellyite, quartz, chlorite, calcite, and trace native copper. Amygdules are frequently filled with colorless to white quartz. a mixture of prehnite pumpellyite and quartz, a mixture of pumpellyite, a mixture of quartz with lesser calcite, only quartz, and only pink inclusions of native copper in milky or colorless quartz. There is visible native copper in some amygdules.

The massive interior of the flow is much less altered than the flow top and represents a relatively impermeable horizon in the paleohydrothermal system. In contrast, the flow top was a pathway for movement of hydrothermal fluids.

Stop 3: Overview at Bumbletown Hill

Latitude: 47.290100N; Longitude: -88.417250W

Directions: From Portage Lift Bridge follow US-41 towards Copper Harbor and proceed through Calumet towards Allouez for about 15.5 miles to Bumbletown Road. Turn left (west) and proceed one mile up to the top of Bumbletown Hill via Cedar Street. Walk around outside of communication tower fence to get excellent views as described below.

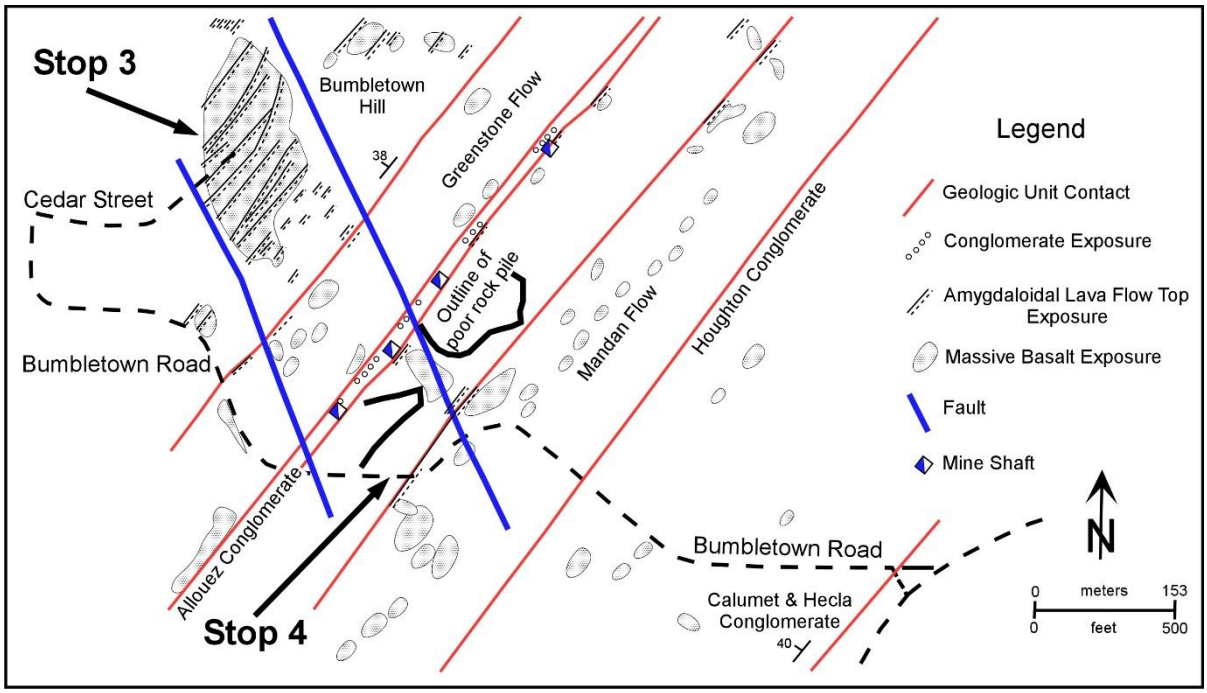


Figure 15: Geologic sketch map of Bumbletown Hill modified from White (1971).

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. From the top of Bumbletown Hill the land slopes very gradually to the northwest toward Lake Superior. This slope is similar throughout much of the northwestern side of the Keweenaw Peninsula. The area is underlain mainly by conglomerates and sandstones of the Copper Harbor Conglomerate dipping at about 20 to 30 degrees NW. 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. Next to the Keweenaw fault beds of the Jacobsville Sandstone can be steeply dipping.

Looking northeast along the strike of the Portage Lake Volcanics, one can see the cuesta form of the ridge underlain by the Greenstone flow. At Bumbletown Hill, the Greenstone flow is only 75 m thick (Figure 15), but the flow thickens abruptly to more than 400 m near this end of the cuesta ridge. The Greenstone flow dips northward at about 25^{o} toward the center of the Lake Superior. It can be traced along much of the Keweenaw Peninsula (Figure 6) 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 to 1,500 km^3 (Longo, 1983; White, 1960). The Greenstone Flow is an enormous lava flow. It is possible that rather than having been a lava flow the Greenstone Flow was a lava lake. Regardless, the Greenstone Flow perhaps is the greatest single continuous outpouring of lava on Earth.

Very slow solidification of this great mass of magma allowed extensive *in-situ* magmatic differentiation (Cornwall (1951a, 1951b). Magmatic differentiation resulted in a massive, ophitic (lath-shaped plagioclase surrounded by large irregular masses of clinopyroxene) base of the flow; an overlying zone of intercalated subophitic and pegmatoidal layers; an upper ophitic zone; and a fine-grained, vesicular flow top. The lower ophitic zone experienced rates of undercooling low enough to allow growth of clinopyroxene oikocrysts up to 5 cm in diameter.

The geochemical composition of the Greenstone Flow magma is more evolved than typical olivine tholeiites; which constitute the greatest volume of the Portage Lake Volcanics. Primitive olivine tholeiite and quartz tholeiite occur between the Greenstone Flow and the top of the Portage Lake Volcanics. Generally, magmas of the Portage Lake Volcanics become more primitive and less crustal contamination with time during the development of the Midcontinent Rift (Paces, 1988). At Bumbletown Hill, the Greenstone Flow is only 75 m thick and is composed of a thick amygdaloidal flow top with some exposures of fine-grained columnar basalt.

To the left of the cuesta ridge the rocks consist of the top of the Portage Lake Volcanics and the bottom of the Copper Harbor Conglomerate. To the right of the ridge, the more distant hills are formed by lava flows near the base of the Portage Lake Volcanics.

Bumbletown Hill is located on the southwest side of Allouez Gap, a NW- to SE-trending valley (see Figure 7). 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 (Figure 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 (Figure 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 Figure 16), which was the second largest copper producer in the native copper district. There was a readily visible line of poor rock piles, a little more than 1,500 m southeast of Bumbletown Hill, from the many mines which were producing native copper from the Kearsarge deposit. Many of these piles are now gone as they have been crushed for aggregate. The line is still visible in the fall when leaves are not on the trees. About 1,200 m N65^{o}E of the hilltop, the Houghton conglomerate and the stratigraphically lower Iroquois flow produced 33 million pounds of copper. East of Bumbletown Hill but no longer visible is the Kingston Mine, one of the most recent native copper mines to open and last to close. It was discovered by exploration near the Allouez Gap fault. It only produced 20 million pounds of copper from 1963 to 1968.

Stop 4: Interflow Conglomerate at Bumbletown Hill

Latitude: 47.287136N; Longitude: -88.415365W

Directions: From top of Bumbletown Hill turn around and head downhill 0.5 miles to pull over on left just before dirt road.

Specimens of Allouez Conglomerate are scattered about the southeast flank of base of Bumbletown Hill. The remnants of Allouez Conglomerate poor rock piles are private property (Figure 15). Near this pullover there is the opportunity to collect specimens of the Allouez Conglomerate.

The Allouez Conglomerate is one of a small number of interflow clastic sedimentary horizons within the Portage Lake Volcanics and is visible on the lower slopes southeast of Bumbletown Hill. Conglomerate layers within the Portage Lake Volcanics are important stratigraphic marker horizons (Figure 6). Correlation of basaltic lava flows along strike would be difficult without clastic sedimentary marker beds deposited during periodic waning of volcanism. The Allouez Conglomerate can be traced more than 120 km along strike from Mass to Delaware (Figure 6). The Allouez conglomerate is just below the Greenstone flow. At Bumbletown Hill the Allouez Conglomerate was mined for native copper (Figure 15) albeit it only yielded about 75 million lbs of refined copper (Table 2). Elsewhere the Allouez Conglomerate has yielded additional native copper.

The Allouez Conglomerate consists of mostly red-colored conglomerate with lesser amounts of sandstone and siltstone. The largest clasts 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 by White (1971) gave the following results: 16% basalt, mostly amygdaloidal; 36% quartz porphyritic rhyolite; 11%, feldspar porphyritic rhyolite; and 37% felsic granophyre. The Houghton Conglomerate is almost entirely clasts of quartz porphyry as is the Kingston Conglomerate. There are conglomerates within the lower Portage Lake Volcanics near the tip of the Keweenaw Peninsula whose clasts are clearly sourced from an extrusive dome of rhyolite. This could be the explanation for the uniformity of clasts in the Houghton and Kingston Conglomerates. In contrast, the heterogeneity of the Allouez Conglomerate clasts suggests a less restricted source area (White, 1971).

Evidence of native copper mineralization can be seen in some rocks of the Allouez Conglomerate at nearby 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.” Chalcocite is a product of late- stage hydrothermal fluids.

Supergene alteration resulting from the downward percolation of groundwater is not common at depth in most the native copper deposits. Supergene alteration products are likely common when native copper ore bodies are at or very near the surface. At this stop, supergene alteration minerals are common including chrysocolla, malachite, and cuprite. The depth of occurrence is unknown although given the widespread distribution in rocks of the poor rock piles it seems likely at least some supergene alteration occurred at depth as documented elsewhere. The occurrence of supergene alteration minerals at depth in the native copper mines was used by Bornhorst and Robinson (2004) to hypothesize that at least some supergene alteration was Precambrian in age.

Stop 5: Seneca Mine Rock Pile

Latitude: 47.311915N; Longitude: -88.365818W

Directions: At the intersection of Bumbletown Road and US-41, turn left (northeast). Drive about 2.9 miles almost through Mohawk to 1st Street when you will turn left (northwest). Proceed on 1st Street/Seneca Location Road 0.3 miles to gated road. Walk to rock piles.

THIS STOP IS ON PRIVATE PROPERTY. PLEASE GET PERMISSION TO ENTER PROPERTY.

The Kearsarge lode was worked by the Seneca Mine, one of multiple mines which produced native copper from the top of the Kearsarge lava flow over a strike length of more than 12 km and down-dip as much as 2,500 m (Figure 16). 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 Calumet & Hecla 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° NW (Figures 6 and 16). It lies directly above the Wolverine Sandstone (Figure 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 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 are variable, but they can make up a large percentage and phenocrysts are up to 2.5 cm in length. This zone is probably the result *in situ* floating of plagioclase during surface crystallization of the flow. The phenocrysts likely formed in a shallow magma chamber. Specimens with abundant plagioclase phenocrysts are common on this rock pile.

The basalt of 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 minerals are replaced by alteration minerals to varying degrees.

The amygdule and interfragment spaces are filled with (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 (Figure 17). A zonal stratabound arrangement of amygdule minerals in the Kearsarge deposit is seen in the Ahmeek Shaft No. 3 (Figure 18). The zoning may be explained by deposition of secondary minerals from a hydrothermal solution moving along a permeable channel. 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 is consistent with the paragenetic relationships seen in individual samples from the rock pile. 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 (Figure 16), 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 mineral zones.

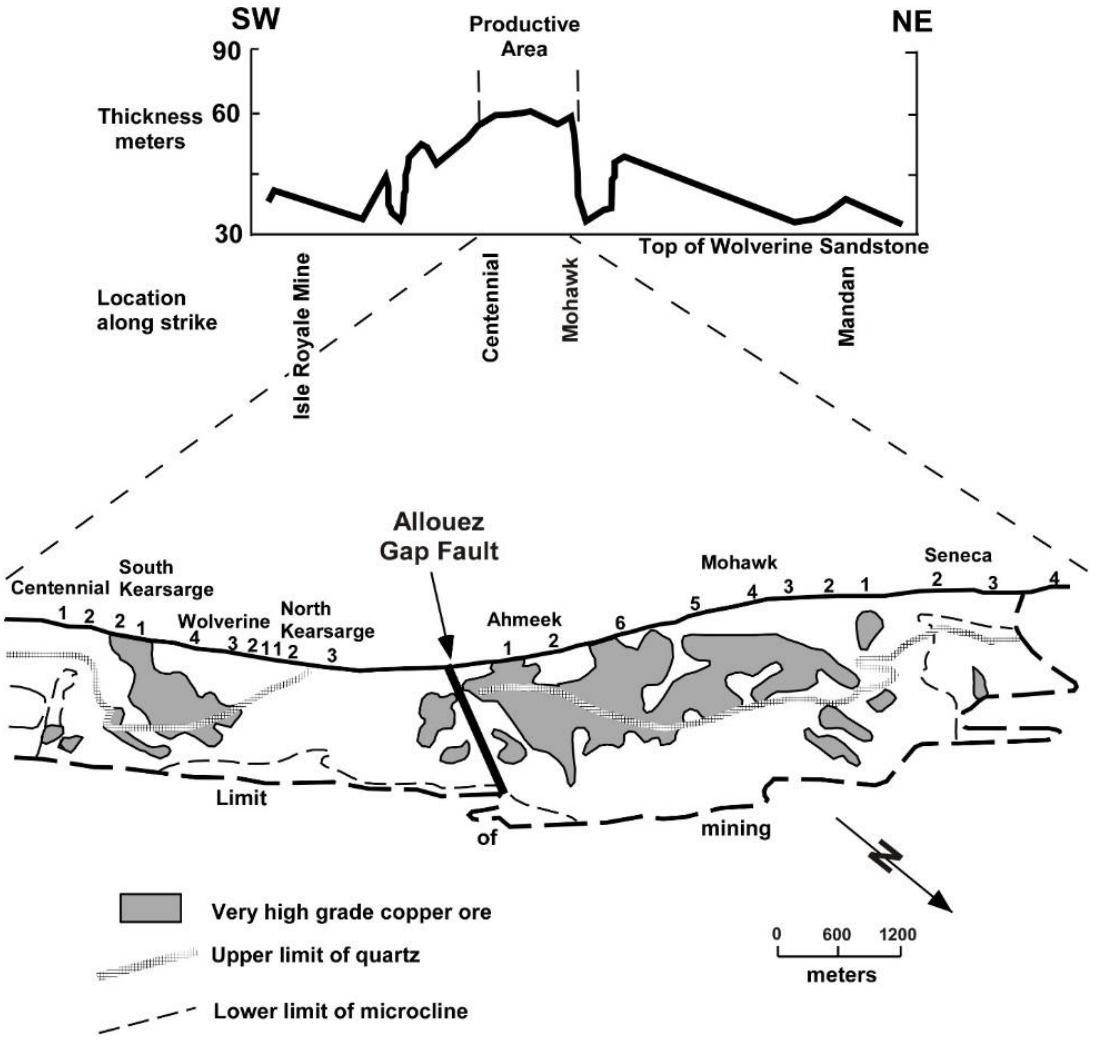


Figure 16: Thickness of the Kearsarge lava flow showing the productive area to be the thickest in the top diagram modified from Butler and Burbank (1929). The Kearsarge flow top ore body is bisected by the Allouez Gap fault. Bottom diagram shows strike parallel down-dip projection to vertical showing 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 of the line (lower) and K-feldspar is absent on the down-dip side lower line shown. The Kearsarge flow dips about 35 to 40^{o} NW. Mine names and shaft numbers are noted.

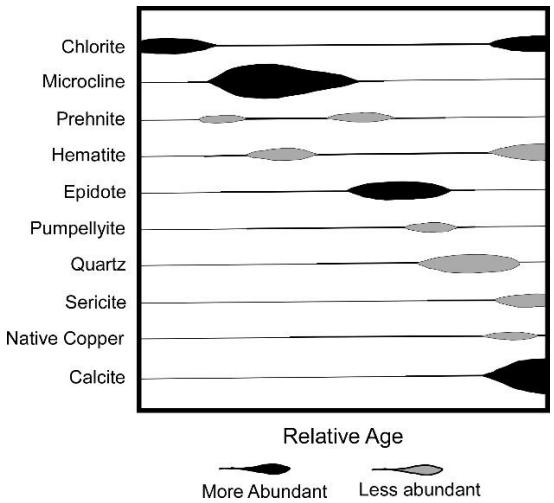
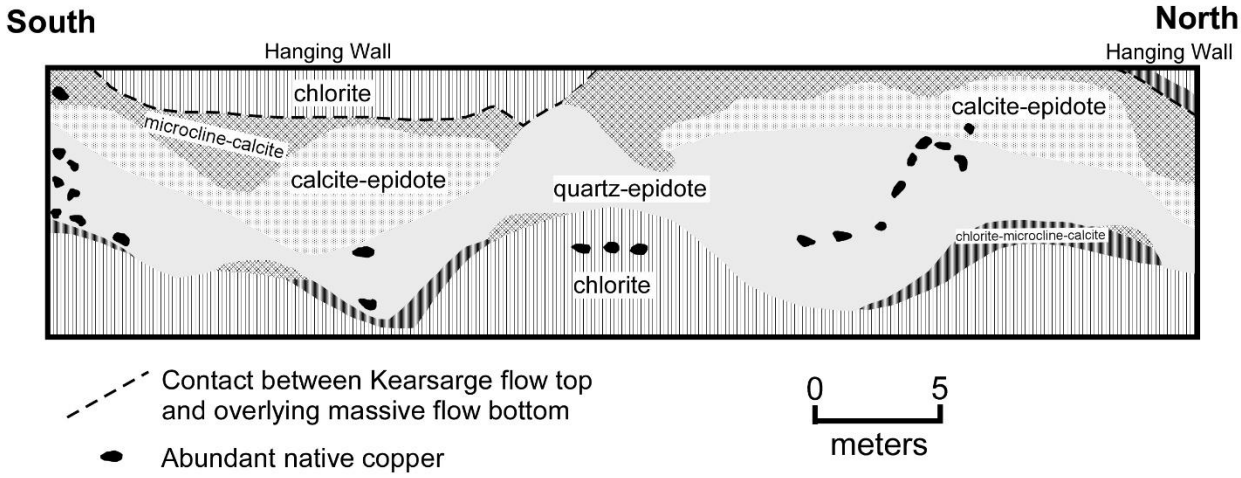


Figure 17: Paragenesis of secondary hydrothermal alteration minerals in the Kearsarge deposit at the Wolverine No. 2 Mine.



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 18: 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.

The Allouez Gap Fault bisects the thickest segment of the Kearsarge Flow along its 55 km strike length (Figure 16). 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). The latter may be late-stage. 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 with 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% (microcline); pink feldspar (adularia) 15%; epidote, 17%; prehnite, trace; pumpellyite, trace and quartz, trace. Many specimens contain multiple minerals and illustrate paragenetic relationships.

Stop 6: Eagle River Falls

Latitude: 47°24'44.9N; *Longitude:* - 88°17'47.3W

Directions: Return to US-41 from Seneca Mine and turn left (northeast) continuing 7.2 miles to junction of US-41 and M-26. Turn left (north) on M-26 towards Eagle River and proceed 2.3 miles across bridge and immediately right after the bridge into the parking lot.

The waterfalls of Eagle River are near the contact between the top of the Portage Lake Volcanics and the base of the Copper Harbor Conglomerate (Figure 5). The contact dips about 30º 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 of the peninsula. The tholeiitic basalt subaerial lava flows just below the contact are pahoehoe type with a ropy upper surface. The orientation of the ropes indicates that the flow 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 Conglomerate. The Copper Harbor Conglomerate consists of red-brown rhyolite-pebble conglomerate but includes many sandstones 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 Conglomerate. The environmental setting of the Copper Harbor Conglomerate is discussed further at Stop 8.

This contact marks an abrupt change in the geologic evolution of the Midcontinent rift. Below this contact the dominant strata is a very thick succession of subaerial basalt lava flows erupted from fissure vents under Lake Superior that filled the progressively extended and down dropped rift basin

during active Midcontinent rifting. Below this contact the Portage Lake Volcanics consists of more than 200 individual lava flows with a cumulative exposed thickness of about 5,000 m; the base is fault truncated (Figure 4 and 5). Abruptly above the contact lava flows are strikingly absent and clastic sedimentary rocks are the dominant strata deposited in a sagging basin after active extension ended. The clastic sedimentary strata above this contact consists mostly of conglomerate and sandstone with a cumulative exposed thickness of more than 5,700 m; the top is not exposed (Figure 4 and 5). While generally absent, a thin package of mafic to intermediate volcanic rocks (Lake Shore Traps) deposited as a shield volcano interfingers with clastic sedimentary rocks of the Copper Harbor Conglomerate. The very last magmatic activity in the MCR is the lone Bear Lake alkaline igneous body near the middle of the clastic sedimentary strata.

Stop 7: Great Sand Bay

Latitude: 47. 446140*N; Longitude: -88.* 216411*W*

Directions: Continue driving northeast (right from parking area) on M-26 for 4.5 miles until the Great Sand Bay paved pullover with overview and beach access.

The Great Sand Bay overlook provides a beautiful view of Lake Superior (Figure 19). 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. 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 Conglomerate. In the Keweenaw Peninsula the Lake Shore Traps are interbedded near the middle of the Copper Harbor Conglomerate (see Stop 9). The massive interiors of these lava flows are more resistant to erosion than the underlying and overlying conglomerates and sandstones of the Copper Harbor Conglomerate. As a result, harbors such as 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 crosscut the Eagle River shoals located about 0.5 to 1 km offshore. To date, there have been a total of 36 underwater copper veins discovered from the eastern tip of Great Sand Bay (visible at this stop) to Eagle River, about 3.2 km west. Some of these submerged veins likely connect with veins recognized from on land exposures (Figure 19). The native copper that is naturally on the bottom lands of Lake Superior are grouped as "lake copper." The largest lake copper specimen ever recovered underwater was a massive 19-ton unattached copper slab in July of 2001. It was recovered from one of these vein deposits north of Jacobs Creek in about 9 m of water. This large underwater native copper vein is on display at the A.E. Seaman Mineral Museum in the outside copper pavilion.

Many of the submerged veins 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 the 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 associated with well-defined copper crystal specimens.

Bornhorst and Barron (2017) provide additional information about the Guiness World record tabular 19-ton native copper mass.

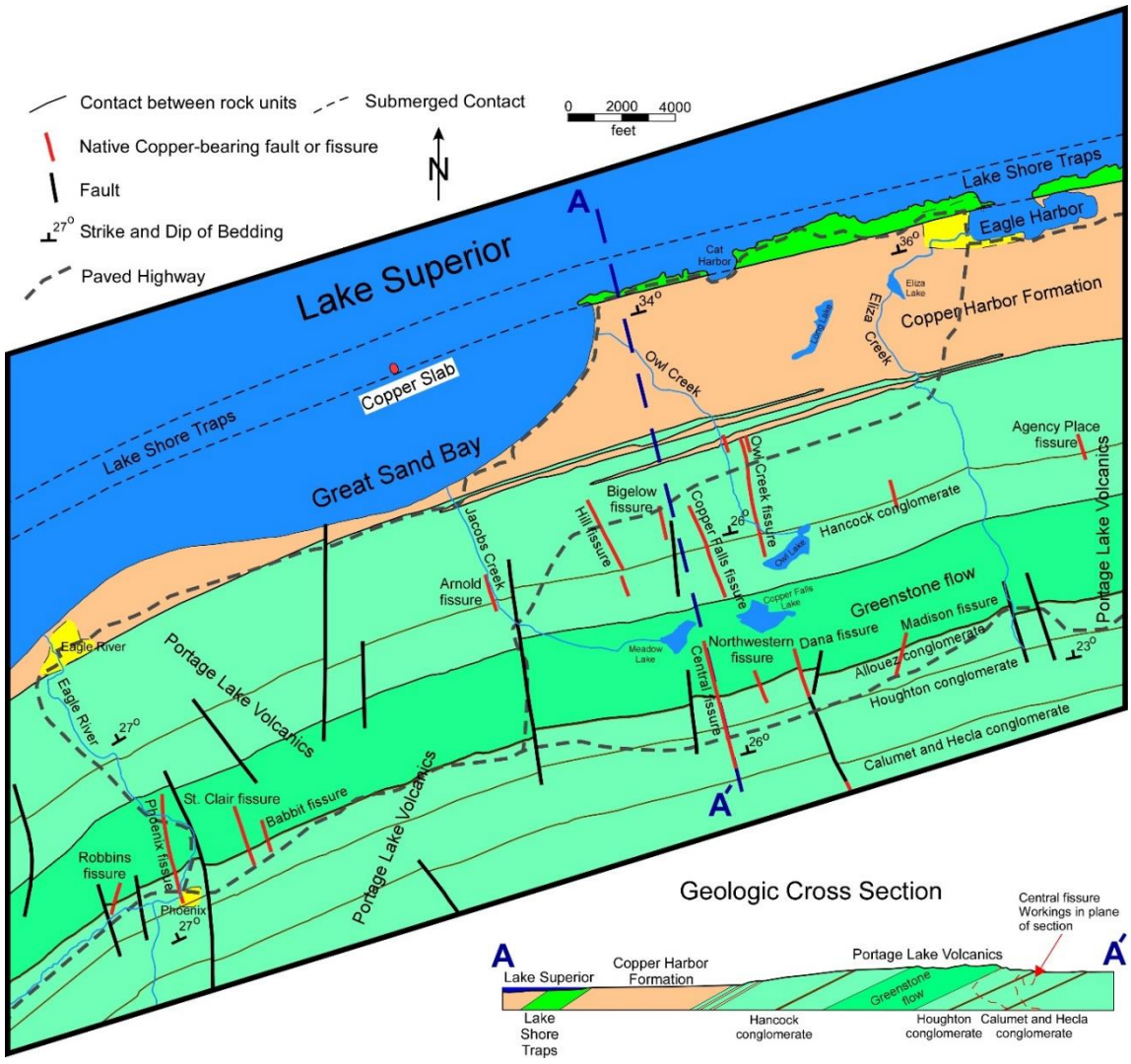


Figure 19. Geologic and location map of the 19-ton submerged native copper vein recovered from Great Sand Bay, Keweenaw Peninsula, Michigan (from Bornhorst and Barron, 2017).

Stop 8: Copper Harbor Conglomerate, J. & M. Lizzadro Lakeshore Preserve

By Daniel J. Lizzadro-McPherson

Latitude: 47. 479128N; Longitude: -87.979576W

Directions: Continue east-northeast on M-26 from Great Sand Bay to Eagle Harbor. Continue on M-26 towards Copper Harbor for 9.2 miles (14.8km) until the sign for the preserve appears on the lakeward side of the road. Park at the pull-off on the north-side of M-26 for beach access.

PLEASE DO NOT DAMAGE THE OUTCROP. NO HAMMERS ALLOWED. COLLECTION OF BEACH ROCK ONLY.

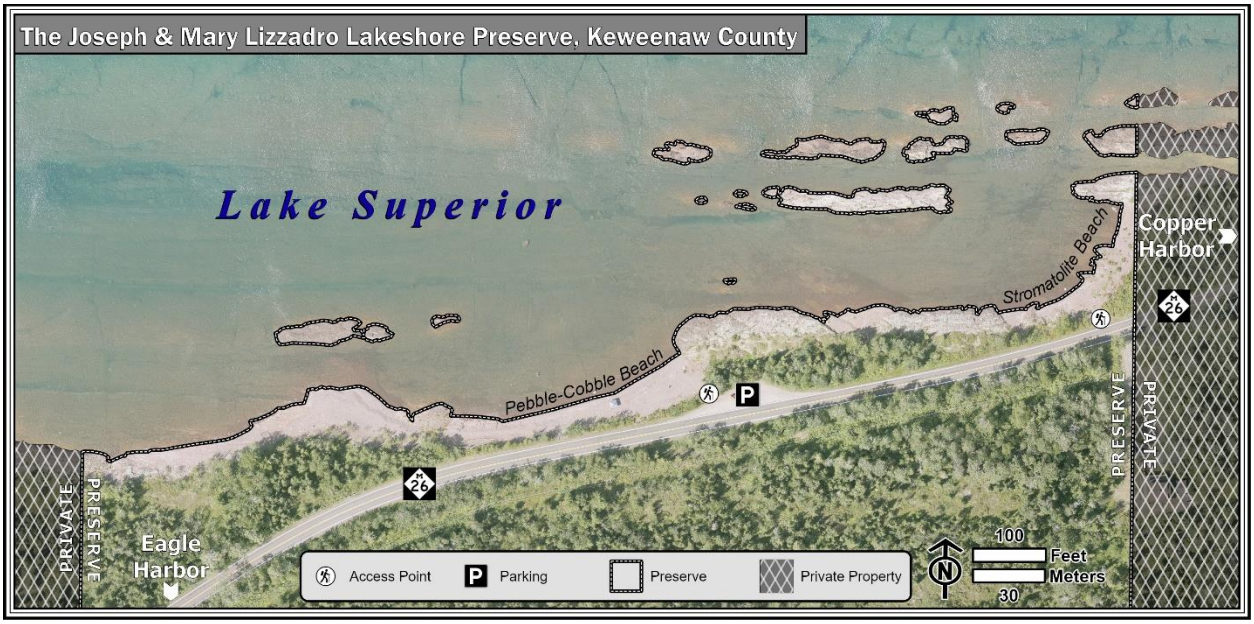


Figure 20: Overview of the J. & M. Lizzadro Lakeshore Preserve, which encompasses the entire shoreline between private properties (shaded, cross-hatched areas) along M26.

The Joseph & Mary Lizzadro Lakeshore Preserve is located near the northern-most point on the Keweenaw Peninsula and is host to prominent outcrops of Copper Harbor Conglomerate, colorful cobble-pebble beaches, and iconic views of Lake Superior. This site is notable due to the diverse facies of CHC, rare occurrences of stromatolites (genus *Colleria*) and raindrop impressions found among the wave-washed bedrock exposures. Established as a preserve in 2003 by the Houghton Keweenaw Conservation District and the Keweenaw Land Trust, the Lizzadro Lakeshore Preserve protects over 640-feet of undeveloped shoreline and several small islands (Figure 20). The preserve showcases one of only three A-ranked Michigan occurrences of the globally rare *imperilied plant* community-type (G2) bedrock beaches. Thanks to a generous donation by Gina Nicholas, the preserve was named in honor of my grandparents, Joseph & Mary Lizzadro, with the hope that this significant Geoheritage site remains protected for many future generations to enjoy.

The bedrock geology of the preserve consists of red-colored clastic sedimentary rocks grouped under the stratigraphic formation of the Copper Harbor Conglomerate (Figure 5). The Copper Harbor Conglomerate. ranges from 490 m thick near the Wisconsin border, to about 1,310 m thick on the Keweenaw Peninsula, reaching a maximum exposed thickness of 2,000 m on the shores of Isle Royale. Overall, the Copper Harbor Conglomerate is generally a medium reddish-brown colored wedge of fluvial siliciclastic conglomerates and sandstones that rapidly filled-in the rift basin as volcanism waned and subsequently terminated (Daniels, 1982; Cannon and Nicholson, 2000; Woodruff et al., 2020). The base of the Copper Harbor Conglomerate. locally interfingers with the uppermost subaerial basaltic lava flows of the Portage Lake Volcanics (Figure 5). Near the stratigraphic lower-third of the Copper Harbor Conglomerate there is a succession of basaltic to intermediate subaerial lava flows, informally named the Lakeshore Traps (Figure 5). Outcrops of Copper Harbor Conglomerate. exposed along the Lake Superior shoreline at the Lizzadro Lakeshore Preserve (Figure 21) are stratigraphically above Lakeshore Traps (Cornwall, 1954).

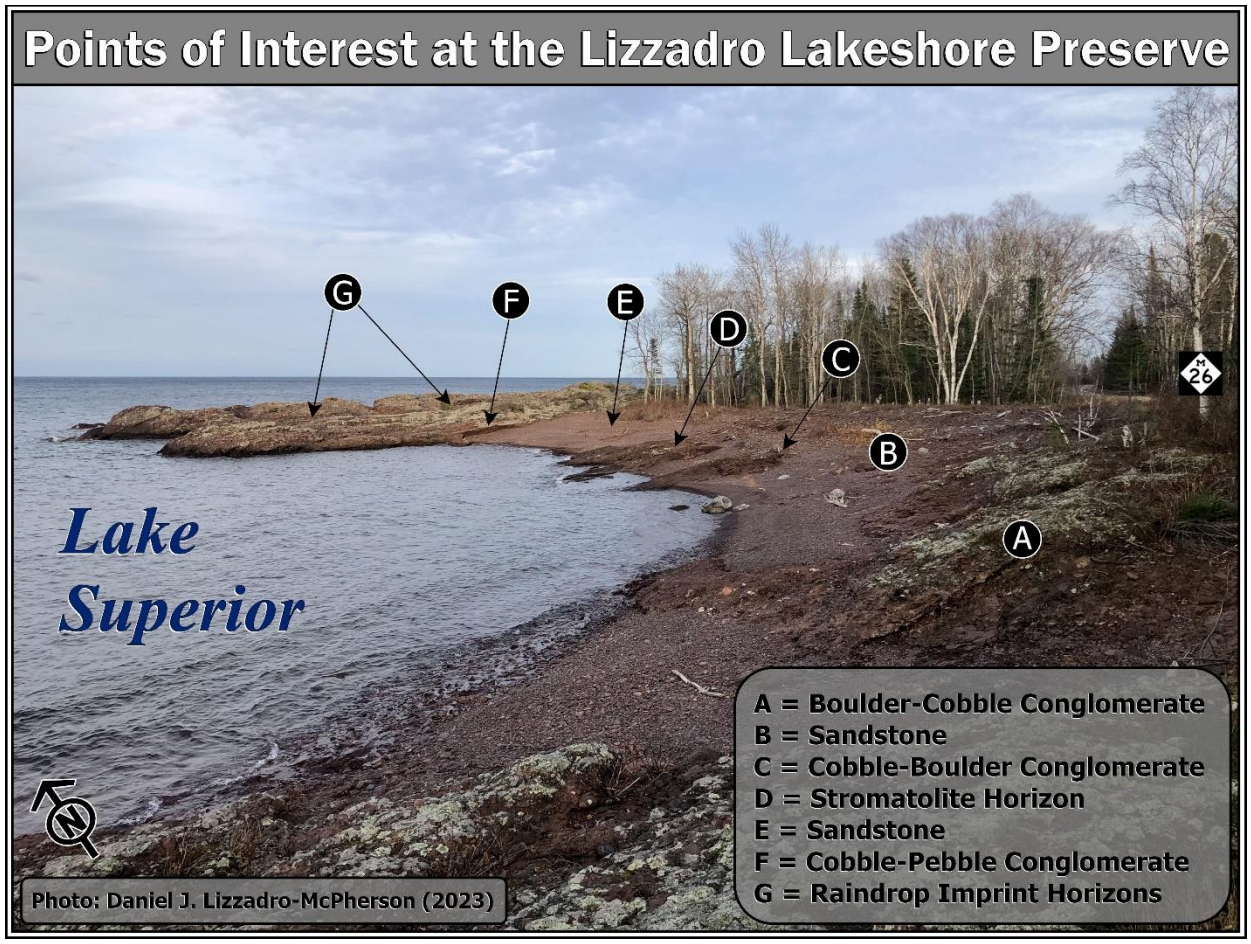


Figure 11: Overview of geologic features present at the J. & M. Lizzadro Lakeshore Preserve.

Nearby at well-studied Dan's Point, outcrops of Copper Harbor Conglomerate exhibit lithologic facies that are characteristic of the upper two-thirds of the formation. The conglomeratic facies (Figures 21 and 22: Points A, C, and F) are primarily clast-supported and comprised of rounded to well-rounded poorly sorted clasts, consisting of a 2:1 silicic-to-mafic volcanic rock fragment ratio with minor pyroclastic, plutonic, and metamorphic rocks (Elmore, 1984). The conglomerate matrix is comprised of coarse sand-sized subangular grains cemented with carbonate and iron oxides. The sandstone facies (Figures 21 and 22: Points B and E) are predominantly subangular to angular, lithic graywackes which exhibit a number of sedimentary structures including: current-ripples, cross beds, and parting lineations. Many of the coarser sandstones and conglomerates in the upper section have calcite-rich cement in the matrix, consistent with a vadose or semiarid, caliche-style environment (Kalliokoski, 1986). Two noteworthy outcrop features include: 1) a thin continuous zone of laminated cryptoalgal carbonate where laterally-linked stromatolite are draped over cobbles and over contorted sandy siltstone layers (Figures 21 and 22: Point D); and 2) raindrop imprints preserved in fine- to medium-grained sandstone lenses (Figure 21: Point G). At the time of deposition, the region was nearly equatorial in geographic position and the climate was likely arid with seasonal rainfall patterns conducive to flashfloods and the development of vadose carbonate (Elmore and Vander Voo, 1982; Kalliokoski, 1986). The raindrop imprints in sandstone (Figures 21 and 22: Point G) support this paleoclimate assumption.

Details surrounding the origin and depositional setting of the Copper Harbor Conglomerate and overlying Nonesuch Formation (Figure 5) have recently been reevaluated. The long-held inference of the origin for these two formations has been interpreted as a non-marine, fluvial-to-lacustrine couplet (White and Wright, 1960; Elmore, 1983, 1984; Ojakangas et al., 2001). The Copper Harbor Conglomerate closely resembles modern-day examples of coalescing fluvial and prograding alluvial fan deposits with varying facies that exhibit proximal-to-distal braided streams, sheet flooding and sand flat features (Elmore, 1984). Isolated cryptoalgal carbonate and ooid lenses are interpreted by Elmore (1983) to have formed in shallow, medial fan lakes and possibly in abandoned or low-water stream channels with limited sediment input (Elmore, 1983). However, marine sedimentological features often resemble and can be easily mistaken for non-marine features unless contextualized with additional evidence, such as isotopic geochemistry.

New research presents evidence for a shallow-marine estuarine origin for at least the upper-third of Copper Harbor Conglomerate and overlying Nonesuch Formation based on new sedimentological observations (Jones et al., 2023) and geochemical analyses (Stüeken et al., 2020; Jones et al., 2020). Sedimentological observations include periodic to rhythmic flaser-wavy-linsen-pinstripe bedding, superimposed sets of ripple cross-laminations with bimodal (herring-bone) sediment transport directions, desiccation cracks and hummocky cross-stratification (Jones et al., 2023). Periodic to rhythmic textures are indicative of tidal-influenced marine depositional settings. Geochemical evidence indicates that gypsum evaporite fabrics have a marine sulfur isotopic composition (Stueken et al., 2020) and that pseudomorphs after gypsum formed in a saline-to-brackish waterbody (Jones et al., 2020). Both studies conclude that the upper section of the Copper Harbor Conglomerate and overlying Nonesuch formation were deposited in a braided fluvial-evaporitic, sabkha-like, tidally-influenced shallow marine embayment rather than fluvial-lacustrine non-marine depositional setting.

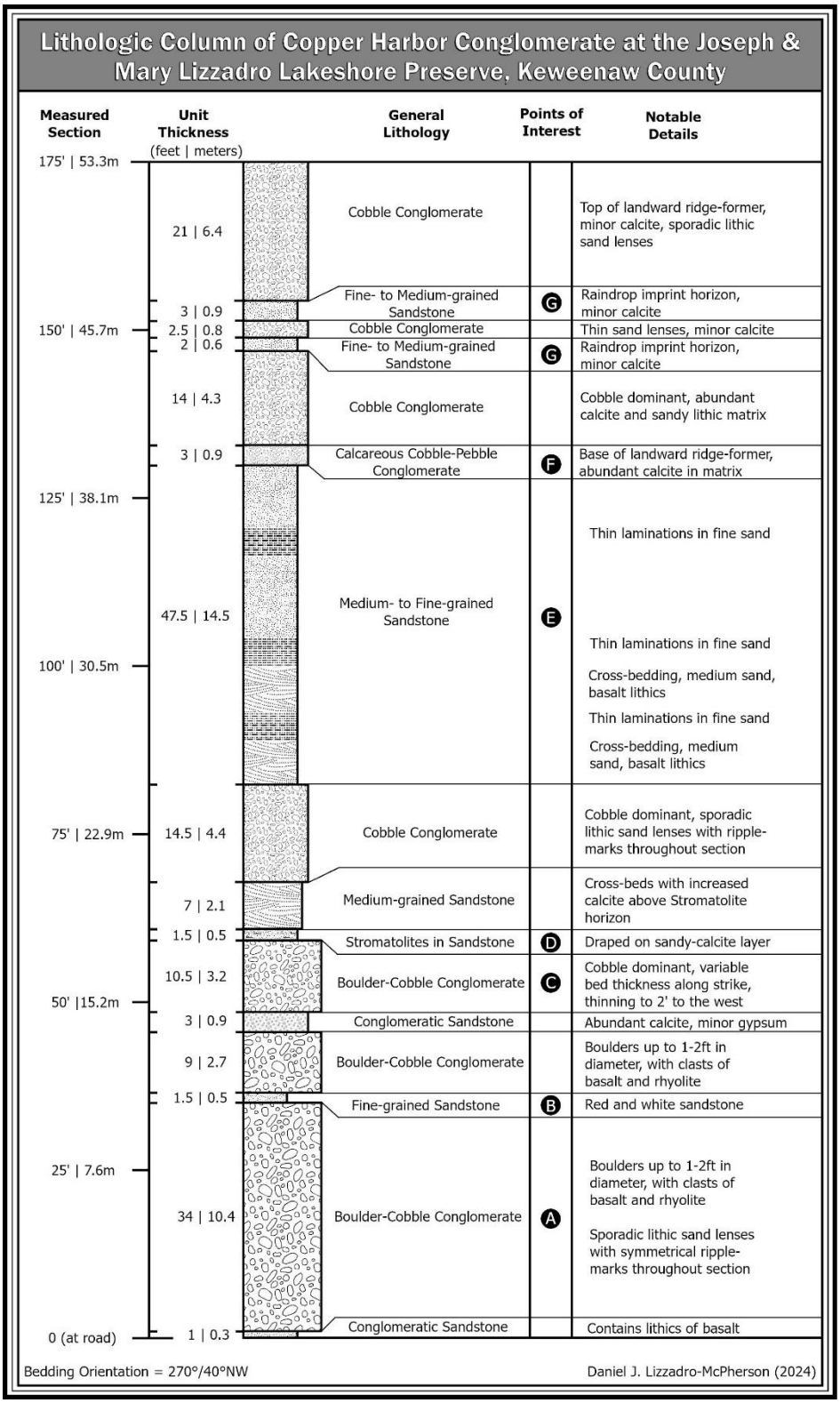


Figure 22: Lithologic column of measured section at the Lizzadro Lakeshore Preserve (data collected by Lizzadro-McPherson, Bornhorst, and Vye (2023). Bedding about E-W strike and 35 to 40° N dip.

Stop 9: Copper Harbor Conglomerate and interbedded Lakeshore Traps at Hunter's Point Park

Latitude: 47.474355N; Longitude: -87.899199W

<u>Directions:</u> Continue driving east on M-26 for 3.7 miles to North Coast Road. Turn left (northwest) on North Coast Road and proceed 0.3 miles to Harbor Coast Lane. Turn right and drive 0.3 miles to the parking area for Hunter's Point Park at end of the road.

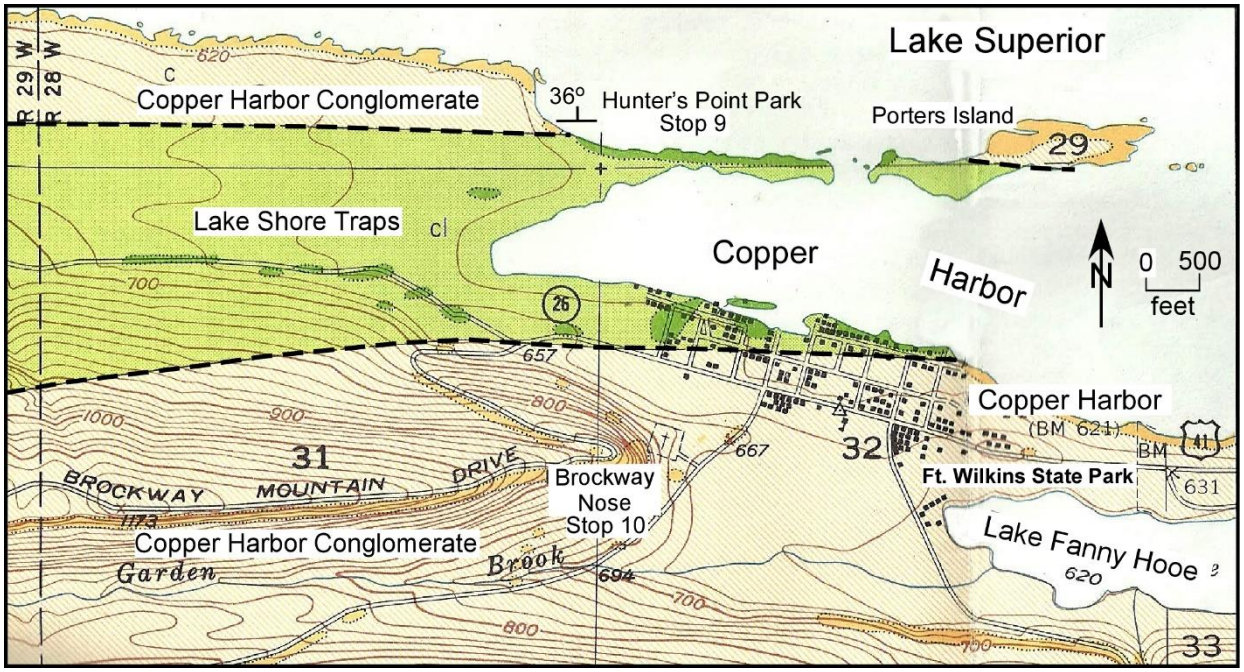


Figure 23: Geologic map of the Copper Harbor area taken directly from Cornwall (1955) showing the location of Hunter's Point (Stop 9) and Brockway Nose (part of Stop 10). Geology from Cornwall (1954).

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 (Figure 23). Prior to becoming an official park the point was a popular hiking destination for visitors. The landowners subdivided the area for residential housing which would have restricted public access without its conversion into a park. The origin of the name 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 Conglomerate is overall composed of volcanogenic clastic sedimentary rocks, dominantly conglomerates with lesser sandstone, siltstone, and shale such as observed at Stop 8. 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 Conglomerate 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 (Figure 23). The Lake Shore Traps consist of Fe-rich olivine tholeiite, basaltic andesite, and andesite lava that were erupted during a time of waning volcanism within the MCR at 1087.2 +/- 1.6 Ma (Davis and Paces, 1990). 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 conglomerates of the Copper Harbor Conglomerate (Figure 23). The strike of bedding is about E-W and dip is about 36º 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. 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 flow top overlain by 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 (hydrothermal and weathering). 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.

Secondary minerals filling amygdules include agate, chalcedony, quartz, laumontite, analcite, calcite, and smectite. 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. Highly visible red hematitic bands form circular patterns within the massive interior; this banding is interpreted to be the result of alteration related to weathering rather than hydrothermal fluids.

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 Conglomerate. 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 8 as these clasts are likely derived from erosion of strat equivalent to the Lake Shore Traps updip towards the highlands of the Keweenaw Peninsula 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 Conglomerate was deposited in an alluvial fan derived from highlands to the southeast in the vicinity of Keweenaw Bay.

Additional outcrops of the Copper Harbor Conglomerate can be seen on the far western end of the pebble to cobble beach. These outcrops consist of interbedded conglomerates and sandstone that are typical of the formation as a whole. These conglomerates are similar to those described at Stop 8. 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 Conglomerate and some of them contain native copper such as those described at Stop 7, Great Sand Bay.

Stop 10: Overview at Brockway Nose

Latitude: 47.467061N; Longitude: *-87.898581W*

Directions: Return from Hunter's Point to M-26. Turn left on M-26 towards Copper Harbor and drive 0.3 miles (0.5km) to Brockway Mountain Drive. Turn right, uphill, on Brockway and proceed 0.6 miles (1.0km) to Brockway Nose pullover on the left.

Brockway Nose provides an excellent view of Copper Harbor and Lake Fanny Hooe (Figure 23). Copper Harbor and several other harbors between here and Eagle River have the Lake Shore Traps at the harbor entrance as the dipping massive interiors of these basaltic to andesitic lava flows are relatively more resistant to erosion. From Brockway Nose viewpoint, the town of Copper Harbor is a prominent visible feature. The town of Copper Harbor began as a boom town in 1843, following the nearby discovery of native copper. Porter's Island, at the mouth of Copper Harbor on the west side of the harbor's Lake Superior entrance (left) was the site of the first government land office. Hunter's Point is west of Porter's Island (Figure 23).

On the east side of the mouth of Copper Harbor, the Copper Harbor Lighthouse, built in 1866, is visible. 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. The vein contained native copper and secondary copper alteration minerals. This location and others in the Keweenaw Peninsula became the foundation of the geological investigations of Douglass Houghton. Houghton's report to the Michigan legislature that sparked the first major mining rush in North America to the Keweenaw Peninsula.

Lake Fanny Hooe is located southeast 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. Nearby, the Estivant Pines is a 0.8 mi^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.

Stop 11: Overview at Brockway Mountain

Latitude: 47.464260N; Longitude: -87.969506W

Directions: Continue uphill and towards Brockway Mountain for 3.4 miles (5.5km).

The top of Brockway Mountain is accessed by continuing upwards from Brockway Nose. Brockway Mountain is a conglomerate ridge that reaches an elevation of over 400 m, with excellent views of the ridge and valley topography of the northern shore of the Keweenaw Peninsula. At Brockway Mountain, the Lake Superior shoreline is oriented about east-west

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, at the southwest end of the rock wall, one can view the dipping conglomerates of the Copper Harbor Conglomerates and see the lava flow near the base of the ridge.

To the west, the Lake Shore Traps form island chains and a prominent ridge in the vicinity of Agate Harbor and Esrey Park. The ridges of the Lake Shore Traps and Copper Harbor Conglomerate along the Keweenaw Peninsula's north shore are also the site of numerous shipwrecks. Lake Bailey (with the small island) and Lake Upson occupy a topographically low valley underlain by a finer-grained clastic horizon (sandstone and siltstone) within the Copper Harbor Conglomerate which was easier to erode by the glaciers than 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 Conglomerate and the uppermost basalt lava flows of the Portage Lake Volcanics. This contact was viewed at Stop 6. 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 as also seen at Stop 3.

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 not easily viewed from Brockway Mountain (better viewed from Brockway Nose). 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 roughly 50 miles (80km) 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, near 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 on the other side of Lake Superior, just northwest of Isle Royale.

Glacial erosion exposed Keweenawan and pre-Keweenawan relatively hard and competent bedrock on the edges of the MCR. 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 1.1 billion year old MCR. 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 Duluth Glacial Lake was the largest of these glacial lakes and only elevations above roughly 400 m (1,300 ft) were emergent such as here at Brockway Mountain and the visible Mt. Bohemia.

Stop 12: Float Copper at US-41 Calumet

Latitude: 47.241989N; Longitude: -88.448427W

Directions: Follow US-41 from Copper Harbor to Calumet. Across the street from the headquarters of the Keweenaw National Historical Park. Google Maps shows the float copper as "Float copper memorial."



Figure 24: Float copper exhibit along US-41 near headquarters of the Keweenaw National Historical Park.

A glacially transported native copper mass is on exhibit at this stop (Figure 24). It weighs 4,263 kg (9,392 lbs) and was found about 7 miles SW of Calumet in less than three feet of surficial sediment/soil. Native copper deposits of the Keweenaw Peninsula were exposed at the bedrock surface at the time of Pleistocene glaciations. The glacial ice entrained masses laying at the surface from previous erosion and plucked masses of malleable native copper from the tabular lodes and veins/fissures. These originally irregular masses were largely cleaned of other minerals and were smoothed and flattened by abrasion from other rocks carried by the glacial ice. The native copper masses were "floating" in the glacial ice, hence locally called “float” copper. When the glaciers retreated about 10,000 years ago, unconsolidated rock debris was left behind by the melting ice as deposits of gravel, sand, and clay. Masses of native copper were scattered among the other sediments carried by the glacier. 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 strata exposed in the Keweenaw Peninsula. Most of the large float copper masses did not move far from their bedrock source in the Keweenaw Peninsula, but smaller masses have been transported quite far and have been found southward in Lower Michigan, Indiana, and Illinois (Bornhorst, 2017). The largest known float copper was discovered in the early 2000s and weighed about 35 tons (~70,000 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 world’s largest existing float copper weighs 26.6 tons (53,200 lbs) was discovered on Quincy Mine claims near Hancock. It is exhibited at a museum in China. The famous Ontonagon boulder was a 1,700 kg (3,708 lbs) float copper mass much smaller than the float copper on exhibit at this stop. The Ontonagon boulder was visited by numerous early European explorers. After Michigan became a territory, Henry Rowe Schoolcraft led an expedition in 1820 with a special goal of seeing the Ontonagon boulder. In 1831, Douglass Houghton accompanied Schoolcraft and visited the boulder too. Pieces of native copper were hacked off of the boulder by Houghton and one of these pieces is part of the University of Michigan mineral collection held by the A. E. Seaman Mineral Museum under the Michigan Mineral Alliance. The Ontonagon boulder was removed from the Keweenaw Peninsula to the nation’s capital in 1843 and is now part of the National Museum of Natural History, Smithsonian Institution’s collection.

Float copper masses were altered by oxygenated groundwater and precipitation since the glaciers retreated. Many masses likely had smoothed fresh copper surfaces as abrasion in the glacial ice cleaned off and smoothed the surfaces. The alteration of these surfaces would have occurred in the last 10,000 years. This supergene alteration produced a surface coating on the native copper consisting of varying amounts of 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, $(Cu_3(CO_3)_2(OH)_2)$ (Figure 24). When small 10s of cm sized masses of float copper are cut, the typical surface alteration is often less than several mm thick. While native copper is not stable in contact with water at typical oxidizing surface conditions, the coating of copper oxides, in particular cuprite, inhibits surface oxidation and thereby protects the native copper from extensive alteration. An open access article provides more about float copper (Bornhorst, 2017).

The basalt mine rock buildings are part of the Keweenaw National Historical Park. They were once all part of the Calumet and Hecla Mining Company (Bornhorst and Molloy, 2017). The Calumet and Hecla Mining Company was incorporated in 1871 as a consolidation of the Calumet

(formed in 1865), Hecla (formed in 1866), Portland, and Scott Mining Companies. The buildings are built, as are many of the buildings, of local materials, including rock from the Calumet & Hecla Mining Company mines. The national 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 past 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 immigrated to the United States. Older mining districts, such as the Keweenaw Peninsula, 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 this stop whereas now, fewer people live in all of Houghton County.

Behind the float copper stands the statue of Alexander Agassiz. Alexander was the son of famous Harvard biologist Louis Agassiz. Alexander was the president of the Calumet & Hecla Mining Company for over 40 years. The statue was moved here from its previous location near Agassiz Park near downtown Calumet in the 1960s. It now stands in front of the Keweenaw History Center, the location of the archives of the Keweenaw National Historical Park. This building was the Calumet & Hecla Library. It is said that at one time this library had more volumes in its collection than the Michigan State Library. Built in 1898, it served as an employee library and bathhouse. The baths were in the basement, until a new bathhouse was constructed in 1911 allowing the basement to be remodeled into additional library space.

Stop 13: Freda Sandstone at McLain State Park

Latitude: 47.238371N; Longitude: -88.613116W

Directions: Follow US-41 to M-203 to McLain State Park. There is an entrance fee. From the pay station turn right towards the campground. Park near the large open area and walk towards the covered shelter and gazebo. Walk down a sandy slope from the gazebo to the lake shore and look around for blocks and slabs of red-colored sandstone.

The Freda Sandstone is generally poorly exposed in the Keweenaw Peninsula except for numerous cliff exposures along the shore of Lake Superior southwest of this stop. At this stop, depending on the level of Lake Superior, slabs and blocks of Fred Sandstone can be found along the beach. If the lake level is low enough, then at the shoreline the bedrock of the Freda Sandstone is partially exposed in shallow water and just off shore. The Freda consists of red-colored fine sandstone and siltstone. The red is interrupted by whiteish reduced zones and spots. There are occasional outcrops of Freda Sandstone landward of the Lake Superior shoreline in the area north and south of Portage Channel which are shown on U.S. Geological Survey geologic quadrangle maps and generally located along creeks (Cornwall and Wright, 1956). There are good exposures of Freda Sandstone on the Lake shore between McLain State Park and Porcupine Mountains State Park. The bedding of the Freda at McLain State Park dips about 5^{O}W as compared to the underlying Nonesuch at Hancock Campground and boat launch (Stop 14) where it dips 25^{O}W. This shallowing of dip up-section is typical of the rift-filling strata, and is mostly due to syn-depositional down warping of the rift-filling strata. The Freda Sandstone is generally fine sandstone which is interpreted to have been deposited in a shallow fluvial environment.

There are multiple excellent well-described stops in the vicinity of White Pine, Michigan to examine the Freda Sandstone and these stops are well described by Woodruff et al. (2013)**.**

Stop 14: Nonesuch Formation at Hancock Campground and Boat Launch

Latitude: 47.133755N; *Longitude:* -88.620581W

Directions: Follow US-41 to M-203 to Hancock Boat Launch and Campground. Drive towards the boat launch and park. At the shoreline you will find a small outcrop of Nonesuch Formation. Walk towards the tree area approximately perpendicular to the Portage Canal shore line and boat launch. About 150 ft from the pavement, you will find the long ago abandoned rock quarry.

The Nonesuch Formation is generally poorly exposed in the Keweenaw Peninsula. At this stop the Nonesuch Formation crops out around the margin of a historic rock quarry which is northeast of the Hancock boat launch. Here the Nonesuch Formation is a fine- to-medium grained, gray-to reddish brown sandstone with subordinate interbedded reddish-brown laminated siltstone and shale. The attitude of bedding is about $N30^{O}E$ and $25^{O}W$ (Cornwall and Wright, 1956).

Overall, the Nonesuch Formation consists primarily of siltstone and shale with subordinate amounts of sandstone. At Hancock campground area the formation is coarser grained since this locality is on the northern fringe of the depositional basin centered some 60 km southwest of this stop near White Pine, Michigan. The Nonesuch can be distinguished from the formations below and above by its generally grayish color. Most Nonesuch is a ripple, laminated siltstone with reddish-gray partings. Siltstones and sandstones of the Nonesuch are composed of around 30 to 40 % rock fragments and 60 to 70 % mineral grains. The rock fragments are mostly volcanic with a 2:1 ratio of mafic-to-silicic + intermediate composition (Daniels, 1982).

There are multiple excellent well-described stops in the vicinity of White Pine, Michigan to examine the Nonesuch Formation and these stops are well described by Woodruff et al. (2013)**.**

Acknowledgments

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Field Trip 2

Mining History and Geology of the Quincy Mine, Keweenaw Peninsula Native Copper District, Michigan

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Objectives of Field Trip

The historic Quincy Mine was the fourth largest copper mine in the Keweenaw Peninsula native copper district. From 1851 to 1967, mining and processing of native copper ore produced ca. 488 million kilograms of refined copper via several shafts sunk along the top of the Pewabic basaltic lava flows. The Quincy Mine is interpreted and made accessible to the public by the Quincy Mine Hoist Association. The field excursion include tours of: the Quincy No. 2 shaft-rockhouse, the largest steam-driven mine hoist in the world, geology exposed along 675 m of an adit that intersects the historic Quincy Mine workings at the No. 5 shaft on the 7th level (107 ft below surface), the Quincy smelter works on the shore of Portage Lake, and two sites on Torch Lake for extraction of native copper from mined ore. Dress appropriately for the underground portion of the tour where the temperature averages 7^{o} C (45 ^{o} F). Walk-in is on a wet surface and will access areas not normally visited by the public. Hard hats and lights will be provided.

Brief history of the Quincy Mine

Douglass Houghton's copper report in 1841 sparked the beginning of the first mining rush of North America to the Keweenaw Peninsula. This human event has its true beginning in the origin of the rocks and the native copper they host. The human story begins with native people's exploitation of native copper and their introduction of native copper to European explorers. Houghton's copper report led to human migration to the Keweenaw Peninsula, the discovery of the first native copper mine in 1845, the Cliff Mine, and the beginning of modern mining. The mining district quickly developed into the most significant copper district in the United States with peak copper production between 1890 and 1915. The district began to decline from 1920 to the end of mining in 1968 and the decline was accompanied by human exodus; the local population continues to decline today. The Keweenaw Peninsula exhibits the classic boom and bust cycle that is often associated with mining. In 1992, a national park was created to preserve and interpret the rich human and geologic history of the now dormant native copper district.

The Quincy Mining Company was established in 1846 and was termed “Old Reliable” for its reliable payment of dividends to its shareholders. From 1866 to 1890 the Quincy Mine area developed into a robust mine. The Quincy Mine was one of the first mines in the Keweenaw Peninsula to evaluate their operations to maximize efficiency and hence, maximize profits. In 1866 they were the third mine to install a man-engine. They experimented with Burleigh power drills in 1868, but the large and cumbersome drills proved a failure. They built a rock house complete with rock crushers in 1873 which eliminated the time and labor involved in the old kiln houses. In 1879, black powder was replaced by dynamite and most hand drilling was eliminated when they adopted Rand Company's "Little Giant" power drills. While these innovations were adapted to increase efficiency, other changes were made to solve problems. The Quincy Mine could not find enough water in the abandoned shafts for its steam plant near the mine up on the top of Quincy hill which led to installation of a pumping station to pump water uphill from Portage Lake. In 1887, they closed their stamp mill on Portage Lake as the tailings they deposited into the lake began to hinder navigation. Quincy proceeded to build a new mill on Torch Lake. By building a new mill they were able to adopt steam-powered stamps rather than the old Cornish drop stamp technology (Lankton and Hyde, 1982). Quincy was able to continue to adapt because they had sufficient copper ore to mine.

The Quincy Mining Company began to rebuild itself during the boom years for the whole Keweenaw Peninsula native copper district from 1890 to 1915. It acquired more powerful engines and hoists and began to raise two skips at a time in balance with each other. In 1891, they built the photogenic and functional No. 6 combined shaft and rockhouse. The man-engine reached its maximum depth in 1892 and as a result Quincy began using man-cars to raise and lower workers using the same hoist they used for ore skips. They built their own smelter on Portage Lake on the old Pewabic Mill site in 1898. In 1905 they began to replace their 2-cylinder engines with 4-cylinder engines to get better use of the steam and reduce coal consumption.

Underground mining was changing too. Miners switched from candles to paraffin-based lamps in 1896-97 and to the less-smoky calcium carbide lamps in 1912-14. Miners began using machines to cut up masses in 1906 rather than drills. Quincy became the first mine in the district to modernize tramming and hauling when they installed a haulage system with a GE battery-storage locomotive in 1901. Quincy designed and patented their own side-dumping tramcars. By 1903 Quincy had 15 electric locomotives. They experimented without success with power shovels underground. In 1905, they reduced the wait time to dump cars underground by digging 500-ton underground storage bins into the hanging wall above the inclined shafts and thereby were able to hoist 25 % more ore to the surface. In 1910, Quincy operated 160 two-man drills, each of which weighed 245 pounds. The lighter one-man drills significantly decreased the cost of mining but this new technology left a miner alone underground which became one issue of the 1913 strike (Lankton, 1991).

The Keweenaw Peninsula native copper mining district never fully recovered after the 1913 strike. The Quincy Mining Company continued to move forward after the strike, and they installed the world's largest steam hoist. Quincy constructed the No. 2 hoist house in 1918 and began operating the world’s largest steam hoist in 1920 as the shaft reached an incline depth of 7750 (2360 m) feet. In 1920 Quincy increased efficiency of processing and smelting by adding

Wilfley tables to their stamp mill and adding a casting wheel at the smelter to eliminate hand pouring of ingots. As the underground workings went deeper, mining became much more difficult. In 1926, they had to install additional pumps to reduce the time used waiting for hoisting water-bailing skips rather than rock skips. In 1927, there was a major underground fire that impacted production. As the mine went deeper, they had to install fans to reduce the temperature at the bottom where miners still sometimes had to work at temperatures of 98^{o}F (Lankton and Hyde, 1982).

The Great Depression closed most Keweenaw mines. The Quincy Mine closed in 1930. The Calumet & Hecla Mining Company closed most of its mines except for a lucky few that operated on reduced shifts. The Quincy Mining Company reopened the No. 6 and No. 8 shafts on a limited basis in 1937. It wasn't until World War II and the advent of price controls that a larger production schedule was resumed. District-wide production resumed on a broader scale when price controls were lifted on August 31, 1945 and Quincy permanently ended its underground mining. The reclamation mill continued to operate and make a profit. In 1948, on its 100th anniversary, Quincy paid a dividend of 25 cents per share due only to the copper produced by the reclamation from tailings in Torch Lake plant (Lankton and Hyde, 1992). The Quincy smelter reopened in the mid-1950s after being closed for about 15 years as Quincy could no longer send reclamation concentrates to the Calumet & Hecla Mining Company smelter. The Quincy smelter remained open until the reclamation plant closed in 1967. The mines were allowed to begin flooding in 1970 (Thurner, 1994). The Keweenaw Peninsula native copper mining district has remained dormant in the 54 years to follow except for several episodes of exploration. Highland Copper Company, from 2011 to 2015, has been the latest explorer to attempt to reopen native copper mines in the Keweenaw Peninsula.

The economies of mining districts are typically sustained by one principal industry, mining. When the mines are profitable and expanding the local communities also do well and when the mines suffer decline the local communities also suffer decline. This creates the boom to bust mining cycle. The local population follows this boom-and-bust trend. During the boom of mining Calumet was a vibrant city and since the mines have closed it has contracted. The last value obtained from the mining industry is selling the useful equipment to other mines followed by the dismantling of buildings to sell for scrap. Evidence of mining such as shaft-rockhouses and industrial buildings disappeared too. The surface rock piles left from mining once dotted the landscape of the Keweenaw Peninsula, but over time they too are disappearing as they are an inexpensive source of crushed stone for roads and other purposes. In the late 1980s Calumet community leaders envisioned that the past might be the key to the future of Calumet and they sought development of historical tourism. This led to creation of the Keweenaw National Historical Park in 1992. The park consists of limited park owned lands/structures and multiple Keweenaw Heritage Sites which are public, private, and non-profit. Together the park and its cooperative sites preserve and interpret the mining history of the Keweenaw Peninsula and supports historical tourism.

The brief history of the Quincy Mine was slightly modified from text written by Larry Molloy, published by Bornhorst and Molloy (2016), and republished by Bornhorst (2022).

Geologic Overview of the Quincy Mine

Readers are referred to Bornhorst (this volume) and Bornhorst and Lankton (2009) for the geologic setting of the Quincy Mine and its native copper deposits hosted by the Portage Lake Volcanics. This overview of the geology of the Quincy Mine is from Bornhorst and McDowell (1992), Bornhorst et al. (1986), and Butler and Burbank (1929).

The Portage Lake Volcanics comprise several hundred subaerial lava flows erupted within the Midcontinent Rift of North America about 1.1 billion years ago. There are occasional interbedded sedimentary rock layers dominated by conglomerate (Bornhorst, this volume). The native copper ores at the Quincy Mine occur in tabular bodies hosted by the originally porous and permeable tops of subaerial basalt lava flows. Ore-forming hydrothermal fluids precipitated native copper in the open spaces some 30 million years after the lava flows were erupted.

About 50 lava flows of the Portage Lake Volcanics are exposed along the adit, twelve of them beneath the interflow sedimentary layer termed the Allouez conglomerate. The Allouez conglomerate and overlying Greenstone flow can be traced from the Houghton-Hancock area to the tip of the Keweenaw Peninsula (Bornhorst, this volume). A clay gouge along a bedding plane fault with undetermined slip occurs at the top of the Allouez conglomerate (Bornhorst and McDowell, 1992). Similar bedding plane faults are common at the tops and bases of conglomerate layers throughout the district.

Native copper ore at the Quincy Mine is hosted by a group of relatively thin porous tops of lava flows that are difficult to correlate laterally without mapping the flows in detail (Butler and Burbank, 1929). The productive tops of the lava flows at Quincy, termed the Pewabic amygdaloids, are not brecciated as is common in the larger flow-top mines in the district. Cavernous zones within the Pewabic flows began as open spaces and subsequently were filled with hydrothermal minerals described above. There is much variability within the tabular lode from well to poorly banded and from high-grade of copper ore to practically barren poor rock (Butler and Burbank, 1929).

Stratigraphically, about 12 lava flows with a cumulative thickness of about 100 m occur between the Allouez conglomerate and the overlying Pewabic flows which are the host rocks for the native copper deposits at the Quincy Mine. However, at this location along the adit, the Pewabic flows are not mineralized as they are on the northwest side of the Hancock fault. There are about 14 additional lava flows until the Hancock fault is reached. The Hancock fault is marked by a distinctive clay gouge, almost pure corrensite, and a green corrensite-rich brecciated mineralized zone adjacent to the gouge (Bornhorst and McDowell, 1992). The Hancock Mine produced native copper from a mineralized segment of the Hancock fault, suggesting that it may have been a feeder of ore-forming hydrothermal fluid into void spaces in the tops of the Pewabic flows (Bornhorst and McDowell, 1992).

The amygdaloidal flow tops exposed by the adit are filled with a number of hydrothermal alteration minerals. Butler and Burbank (1929) describe the main-stage hydrothermal minerals at the Quincy Mine: quartz and calcite are abundant throughout the lode, commonly as euhedral crystals in open cavities; pumpellyite is less abundant but is present throughout the lode; epidote is less abundant than

pumpellyite but is a common hydrothermal mineral; chlorite is particularly abundant in amygdules near the bases of Pewabic lava flows and is locally replaced by quartz and calcite; prehnite is present but not common; and datolite is present only in upper levels of the mine. The Pewabic lode of the Quincy Mine is notable for spectacular euhedral calcite with visibly unaltered pink to rose colored inclusions of native copper, which are highly sought by mineral collectors. Native silver is closely associated with native copper, although abundance is low. Laumontite is sparse and associated with small fissures. Thus, it may be a late-stage hydrothermal mineral rather than main-stage (Bornhorst, this volume). Butler and Burbank described early pumpellyite and epidote followed by quartz, calcite, and native copper. Bumgarner (1980) described amygdules indicating that prehnite and chlorite were early hydrothermal minerals followed by quartz, then chlorite, and lastly calcite.

Overview from native copper ore to copper products

At the Quincy Mine and elsewhere in the Keweenaw Peninsula native copper was extracted from tabular ore bodies by underground mining methods. Quincy Mine yielded about 42,870,000 tons of ore (1 ton = 2,000 lbs) with recovered refined copper totaling 1,077,000,000 lbs at an average grade of 1.26 % copper per ton of ore. The amount of silver in the native copper ore can be estimated from incomplete production statistics in Butler and Burbank (1929) to be roughly 0.2 oz of silver per ton of ore.

The ore is blasted underground into small enough size to be able to be hoisted to the surface. Rock lacking sufficient native copper was sent to the poor rock pile or into abandoned underground mine openings. Prior to being sent to the mineral processing plant to recover native copper, the broken ore is sized by a slatted grating, “grizzly”. The broken ore passing through the grating (most of the ore) is sent to the mineral processing plant and those fragments too large are broken further at the surface near the shaft rock-house with a steam driven hammer. One reason a fragment could be too large is because it is mostly native copper. Native copper is malleable and large masses are not readily fragmented by underground blasting. Once most of the rock was removed from the larger fragments, the copper-dominant fragments were put into barrels (barrel copper) or, if they were too big, they were put onto a flat rail car and shipped directly to the smelter instead of to the mineral processing plant.

The ore from the Quincy Mine was processed to separate native copper from the barren host rock and barren minerals in order to produce a product sufficiently enriched in copper to be smelted. The ruins of the Quincy processing plant will be visited at Stop 4. The first essential step in processing was crushing the ore into sand and smaller sized fragments using stamp mills. The crushing aims to produce fragments which are mostly native copper or mostly rock and thereby liberates the native copper from the rock.

Because native copper is much denser than the host rock (about 3 times denser than the barren host rock), fragments of native copper can be separated from the barren host rock using water and gravity methods such as jigging. Hence, the mineral processing plant usually was constructed near a body of water. The mineral processing plant produced a “concentrate” which was composed of sand-sized and finer native copper with some fragments or partial fragments of barren rock because no separation method is able to completely separate every particle. The ore

mined at Quincy averaged about 25 lbs of copper in a ton of ore. The copper concentrate was more than 50 % copper, hence most of the sand and smaller size fragments from the crusher ended up being waste, which still contained some copper. This waste, called tailings, was transported by water slurry and dumped into lakes, especially Torch Lake.

The amount of native copper in a sand-sized fragment may have been so low that it was correctly separated into the tailings. In other fragments there was enough copper in them but they were incorrectly separated into the tailings. For example, the copper could have been finer grain size than the fragment and it was diluted by barren host rock and minerals. By crushing such fragments to a finer grain size, more copper could have been liberated from the barren host material. Quincy Mine kept track of the lost copper as it dumped tailings into Torch Lake. They later went back and recovered much of the lost copper by reprocessing the tailings using newer more efficient technology. The mineral processing plant and recovery of lost copper are discussed at Stop 4.

The copper concentrate from the processing mill was shipped to the Quincy Smelter Works which is discussed further at Stop 2. The smelting and refining process resulted in solid Quincy copper ingots of up to 99.8 % pure copper. The copper ingots transported to markets by ships. The copper mined and processed by Quincy was sufficiently pure to be fabricated into a variety of usable copper products such as electrical wire.

Copper ore to copper products text from previously published field trip guide by Bornhorst (2022) with modification.

Field Trip Stops

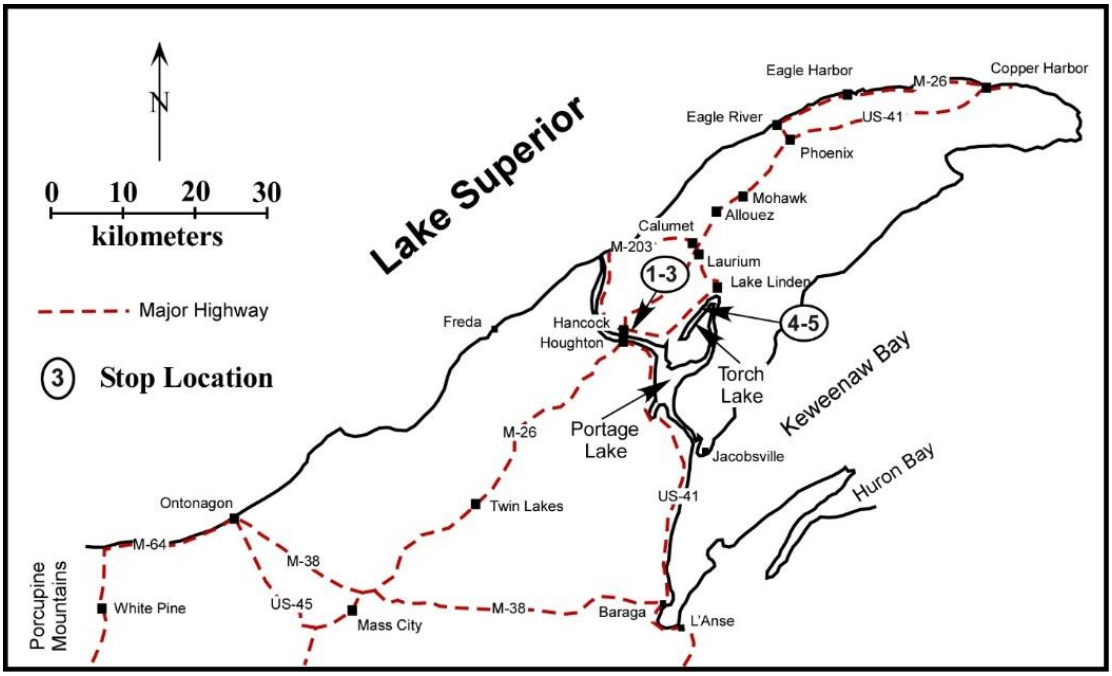


Figure 1: Map showing approximate stop locations of Field Trip 2 to the Quincy Mine.

Stop 1: Quincy Mine, Keweenaw Heritage Site of the Keweenaw National Historical Park

Latitude: 47.137137N*; Longitude:* -88.574875W

Directions: From Michigan Tech drive west on US-41 (left from parking lot by the MUB) and continue through downtown Houghton across the Portage Lake lift bridge through downtown Hancock and uphill to the Quincy Mine No. 2 shaft rock-house turning right into the parking lot just past the shaft rock-house.

The Quincy Mining Company was incorporated in 1846 and operated until 1967. Quincy mined underground from nine shafts on the Pewabic flow top and there was an industrial complex associated with mining (Figure 2). Throughout its history, Quincy Mining Company paid dividends on such a regular basis it was nicknamed "Old Reliable". Quincy Mining Company produced a total of 1.08 billion pounds of refined copper and approximately 100 million oz of silver from approximately 43 million tons of ore at an average grade of 25.1 lbs. of copper per 2000 lb. ton of ore (including copper reclaimed from Quincy tailings). The Quincy Mine ranks as the fourth largest mine in the native copper district. In 1921 the No. 2 shaft was the world's deepest. The Nordberg steam hoist is the world's largest.

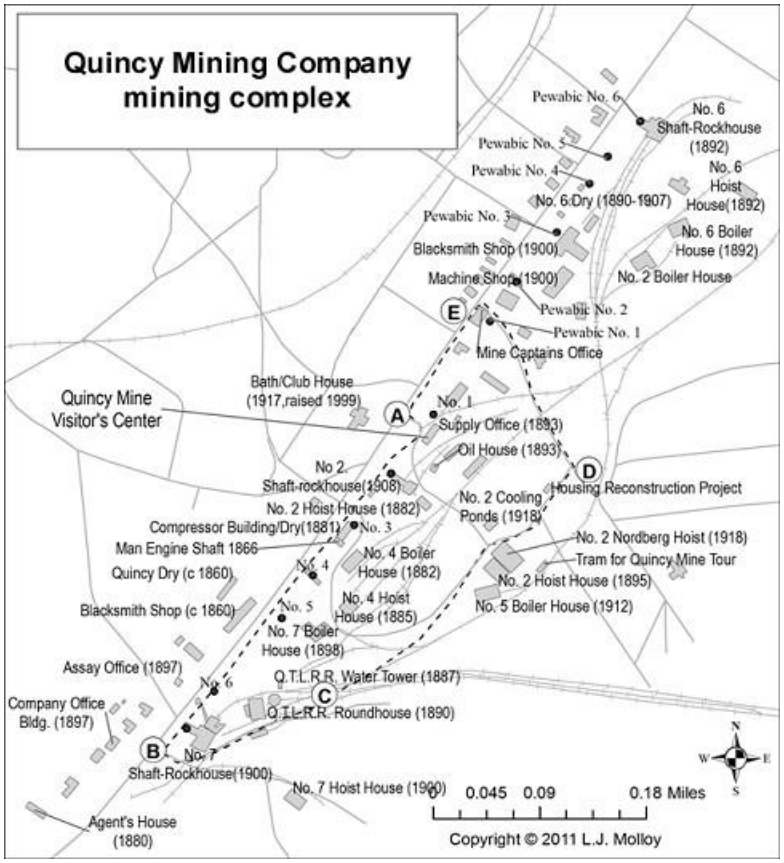


Figure 2: The Quincy Mining Company complex ca. 1900. The No. 2 shaft-rockhouse is near the center of the drawing. From Molloy (2011) with permission.

The No. 2 shaft of the Quincy Mine opened in 1858. At the beginning of mining a simple house was built over the shaft. By 1892, Quincy introduced the concept of hoisting the ore and doing initial crushing and sorting of the ore in the same building, a shaft-rock house. The current Quincy No. 2 shaft rock house was built in 1908 (Figure 3). The Quincy No. 2 shaft rock house is 147 feet (45 m) tall and the angle on the side of the building facing US-41 is at the dip angle of the native copper deposit. Behind the shaft rock house, there are two of the original eight pulley stands and stanchions that were used to support a steel cable extending to the No. 2 hoist house built in 1919. Mining at the No. 2 shaft ended in 1931.

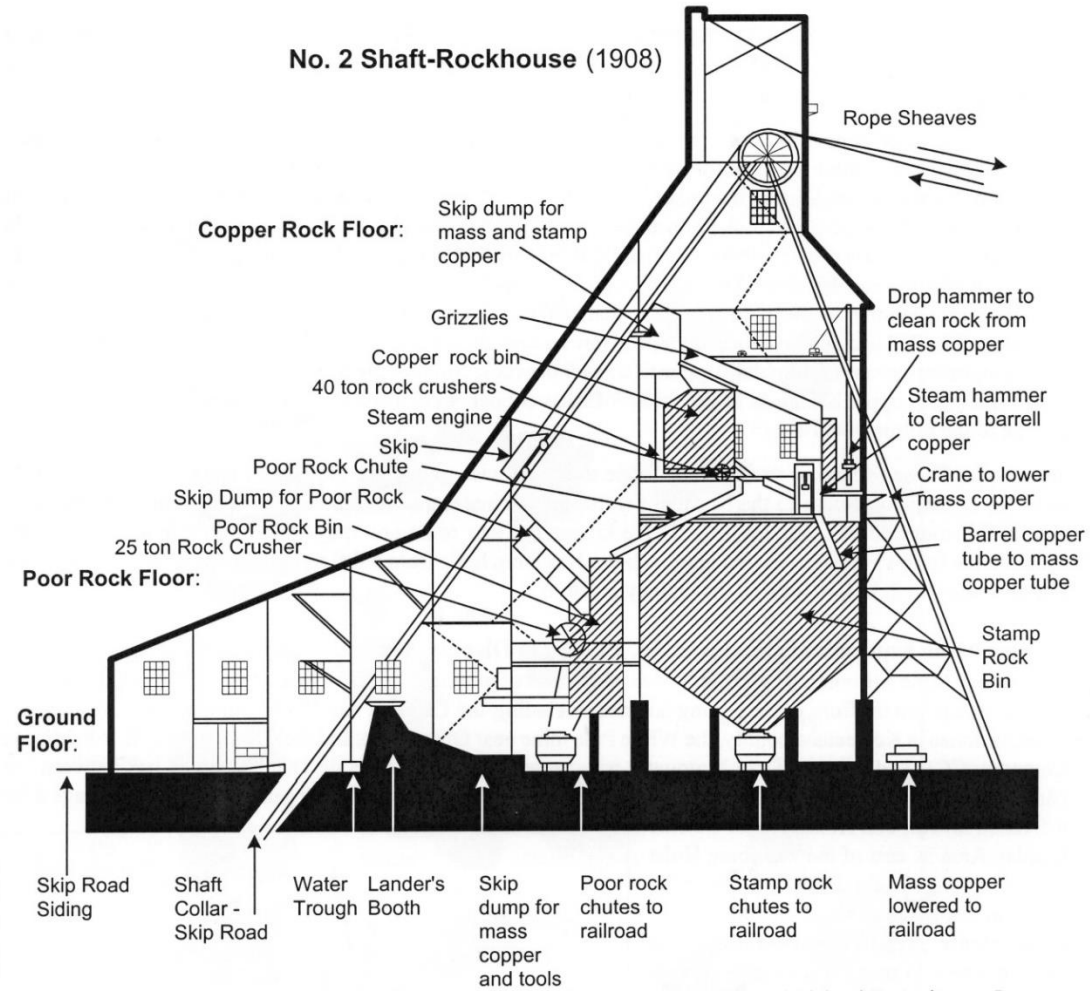


Figure 3: Quincy Mine No. 2 shaft-rockhouse. This drawing is based on Historic American Engineering Record drawing, MI- 2,19/34, Durward W. Potter, Jr., 1978 and Richard K. Anderson, Jr., 1979. From Molloy (2007).

By 1917, the No. 2 shaft had reached such great depths that the hoist engine housed in the 1895 hoist house was no longer adequate. The Quincy Mine needed a large and faster hoist to continue its production. In 1918, the No. 2 hoist house was constructed but World War I delayed

delivery of a new hoist until 1919. The Nordberg hoist began operating in 1920 as the shaft reached an incline depth of 7750 (2360 m) feet. The Nordberg hoist consists of four cross-compound steam engines that work as one (Figure 4a and 4b). The new hoist could move an ore skip carrying 10 tons of rock (13 tons total weight) up at 3200 feet per minute (36 miles per hour) and was more energy efficient than the hoist it replaced. The Nordberg hoist, the world's largest, operated 24 hours per day for 11 years until mining ended in 1931; to a depth of over 9000 feet (2743 m) on the incline (Molloy, 2007).

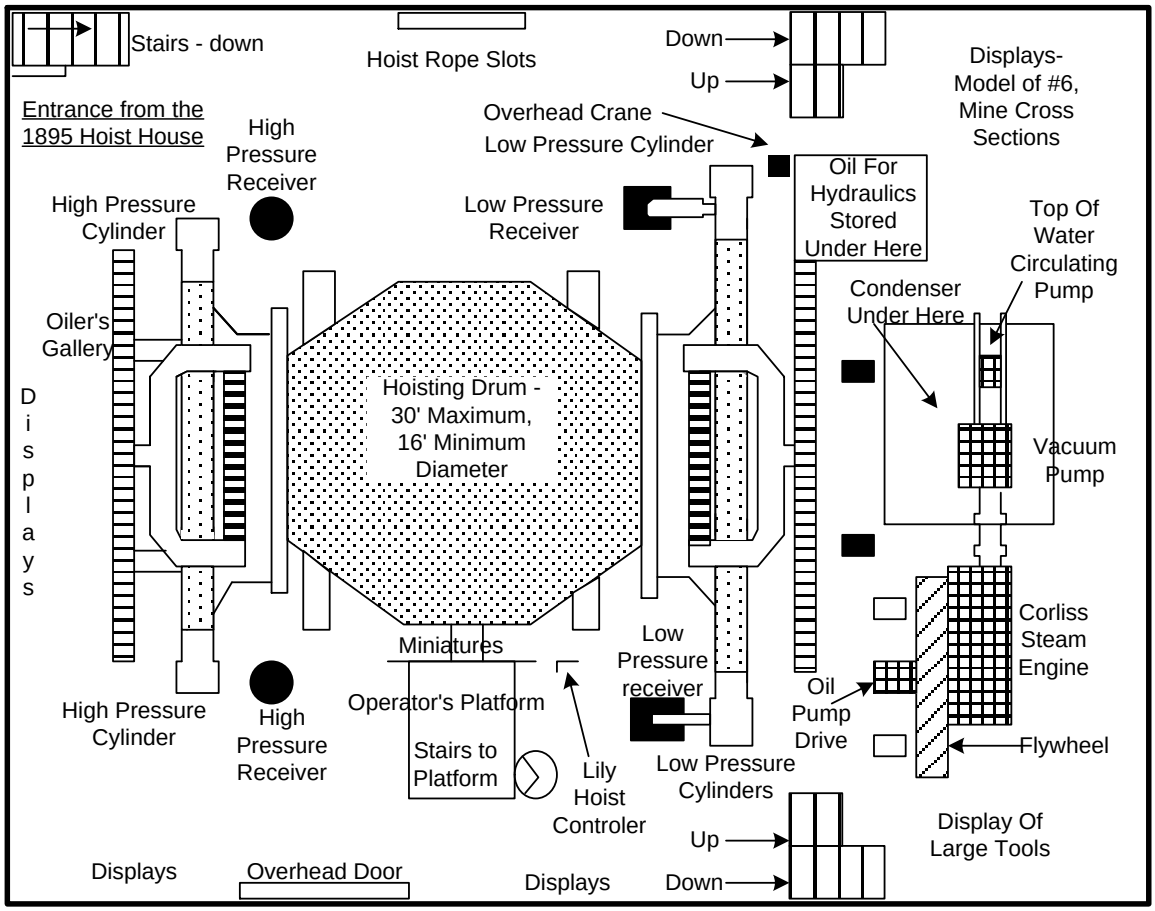


Figure 4a: Quincy Mining Company No. 2 shaft Nordberg hoist diagram. This drawing is based on HAER drawing, MI-2,14/34, Durward W. Potter, Jr., 1978. From Molloy (2007)

A steam engine functions by using alternating intake and exhaust valves to allow steam to enter a chamber, expand inside of the chamber, and use the force of the expansion to push a piston as the steam expands. The double-acting pistons used here can be pushed both up and down in the cylinder by the expanding steam. The entire engine occupies 60 feet by 54 feet of floor space and is 60 feet tall. It is a cross-compound steam engine, an engine where steam is used twice. This common technique accounts for the “choo-choo-choo-choo” sound one hears near steam powered trains.

Today the hoist remains an engineering marvel and is still the world’s largest steam mine hoist.

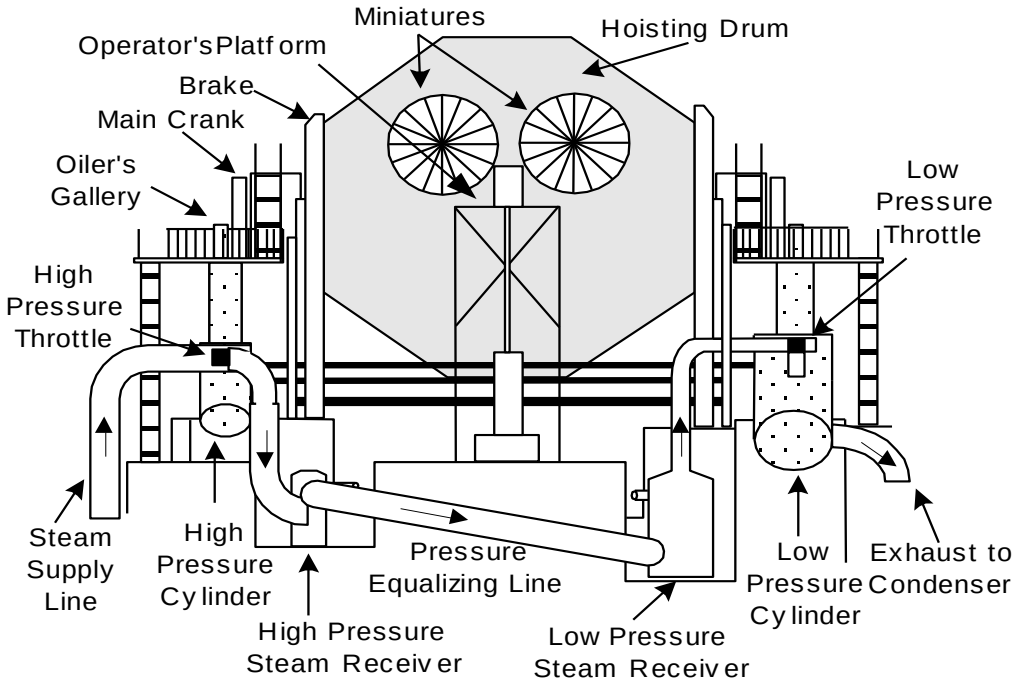


Figure 4b: This diagram illustrates the major features of the hoist and traces the flow of steam through the hoist. It is based on HAER drawing, MI-2,13/34, Jon R. Carter, 1978. From Molloy (2007).

At the Quincy Mine and elsewhere in the Keweenaw Peninsula native copper was extracted from tabular ore bodies by underground mining methods. The ore is blasted underground into small enough size to be able to be hoisted to the surface. Prior to being sent to the mineral processing plant to recover native copper, the broken ore is sized by a slatted grating, "grizzly." Those fragments too large are broken further by steam hammers. The native copper-dominant fragments were put into barrels (barrel copper) or, if they were too big, they were put onto flat rail cars. The barrel copper and larger masses were shipped directly to the smelter located downhill from the shaft rock-house. Next to the Quincy Mine No. 2 shaft rock house is a specimen of mass copper weighing hundreds of lbs. that would have been shipped directly to the smelter.

Text from previously published field trip guide by Bornhorst (2022) with limited modification.

Stop 2: Quincy Smelter Works

Latitude: 47.126688N*; Longitude:* -88.565290W

Directions: From Stop 1 turn left from Quincy Mine returning downhill to Hancock, just before the Portage Lake lift bridge follow M-26 through underpass and from bridge continue 0.4 miles (0.65 km) to the Quincy Smelter Works.

The Quincy Smelter was constructed in 1898 and initially consisted of four or five reverberatory furnaces until switching to two larger furnaces in 1920. The purpose of the smelter was to produce bars of copper with as few as possible impurities. Copper with low enough impurities was ready to be made into copper products.

The copper concentrate from the mill, barrel copper, and larger masses are melted, turned from solid to liquid in a furnace. Smelting is melting at temperatures higher than the melting point. By being higher than the melting point the liquid metal separates from the liquid silicate rock. Since the density of metallic copper is much higher than other components in the liquid rock, the liquid copper sinks to the bottom of the furnace while the other components, molten slag (waste liquid) float to the top. Men skimmed off the molten slag and it was dumped on a pile where it was left to cool into glass and minerals. Across from the parking lot opposite the agent's house/office is a pile of solidified slag.

The molten copper was tapped from the bottom of the furnace into a refining furnace where it was rabbled and poled. Rabbling (stirring) agitates the molten copper and allows the introduction of small amounts of air which oxidizes the impurities (copper has less tendency to oxidize than impurities). The oxidized impurities are lighter than liquid copper and thus, rise to the surface and are skimmed off the molten copper. After the impurities are skimmed off of the molten copper, green sapling poles are inserted into the molten copper (poling). Poling introduces carbon (wood) which reduces the amount of copper oxides in the molten copper. The refined copper was ladled and poured into molds and sprayed with water to cool it. The solid copper shapes were then dumped into a tank of water to complete cooling. Quincy cast copper ingots of several shapes and sizes (bars, wedges, and cakes). Refined copper from Quincy was up to 99.8 % copper which was sufficient for fabrication of copper products in the early 1900s. Some copper ingots had to undergo further refining. Today, the copper would undergo electrolytic refining to make it more than 99.99 % pure copper.

Quincy constructed the smelter to be able to refine and ship its own copper. It also accepted custom work from neighboring mining operations. While underground mining ended in 1945, the Quincy smelter remained open until 1971. After closure of the underground mine, Quincy was actively recovering copper from reprocessing of tailings from Torch Lake. Fortunately, the smelter site remained intact after closing and is now open for guided tours. In 2014, the Keweenaw National Historical Park Advisory Commission acquired the smelter from Franklin Township (Figure 5).

Text from previously published field trip guide by Bornhorst (2022) with limited modification.



Figure 5: The historic Quincy Smelter Works. View July 2022 towards the north from Houghton waterfront.

Stop 3: Quincy Mine Adit to 7th Level Underground Workings

Latitude: 47.130383°N*; Longitude:* 88.573824°W

Directions: Turn left onto M-26 heading west and drive 0.5 mi (0.8 km) to its junction with US-41 north of the Portage Lake lift bridge. At the stop sign, merge onto US-41 and continue west for two blocks (~0.1 mi, ~0.15 km) to Dunstan Street on the right before a BP gas station. Turn right onto Dunstan and drive uphill for three blocks on the right (~0.15 mi, ~0.25 km) to Mason Avenue. Turn right onto Mason and drive east about 0.2 mi (0.3 km) to near a sharp right-hand curve. Just before the curve, turn left to enter an unpaved driveway that leads to the adit.

The Quincy adit was first opened in 1892 by the Quincy Mining Company and subsequently widened in the 1970s by Michigan Tech mining engineering faculty and students for an underground educational and research facility. The adit enters the south side of Quincy hill at a point located ~690 m SSE of the iconic Quincy No. 2 shaft house along US-41 at the top of the hill. From the portal, the adit follows an azimuth of 331° for 675 m, where it intersects the 7th level workings at the Quincy No. 5 shaft (Bumgarner, 1980). The average strike of PLV layers here is 213° (right-hand rule), which means that the adit cuts across them in a direction 30° clockwise from the dip direction. Approximately fifty lava flows and one conglomerate layer, dipping on average 50° NW, are exposed along the adit (Bumgarner, 1980; Bornhorst et al, 1986). Geologic sites of interest in the adit and side passages are described as sites 3A to 3E in the following text and shown in Figure 6.

Underground workings of the Quincy Mine were developed along a series of parallel, relatively thin, lava flow tops, referred to collectively as the Pewabic flow top within the Portage Lake Volcanics. The mine was developed to a vertical depth of 1897 m (6225 ft) comprising 92 levels. The ore body decreases in dip from 54º at the surface to 32º at the bottom levels. Pewabic flows are characterized by large cavernous zones up to 1.5 m thick, interpreted by Butler and Burbank (1929) as coalescing vesicles and large gas cavities. Alternatively, the large cavities could have

been caves or lava tubes formed by lava drainage beneath a solidified flow surface. Openings were connected up to 100s m along formation strike and, where especially well developed, they may host 2 to 10 high-grade copper zones. An interconnected series of such openings would have provided continuous flow paths for ore-forming hydrothermal solutions. Several prominent and steeply dipping veins extend throughout the mine. They probably helped to integrate the hydrothermal system as their mineralogy is similar to that of the hydrothermal mineral assemblage in the flow tops.

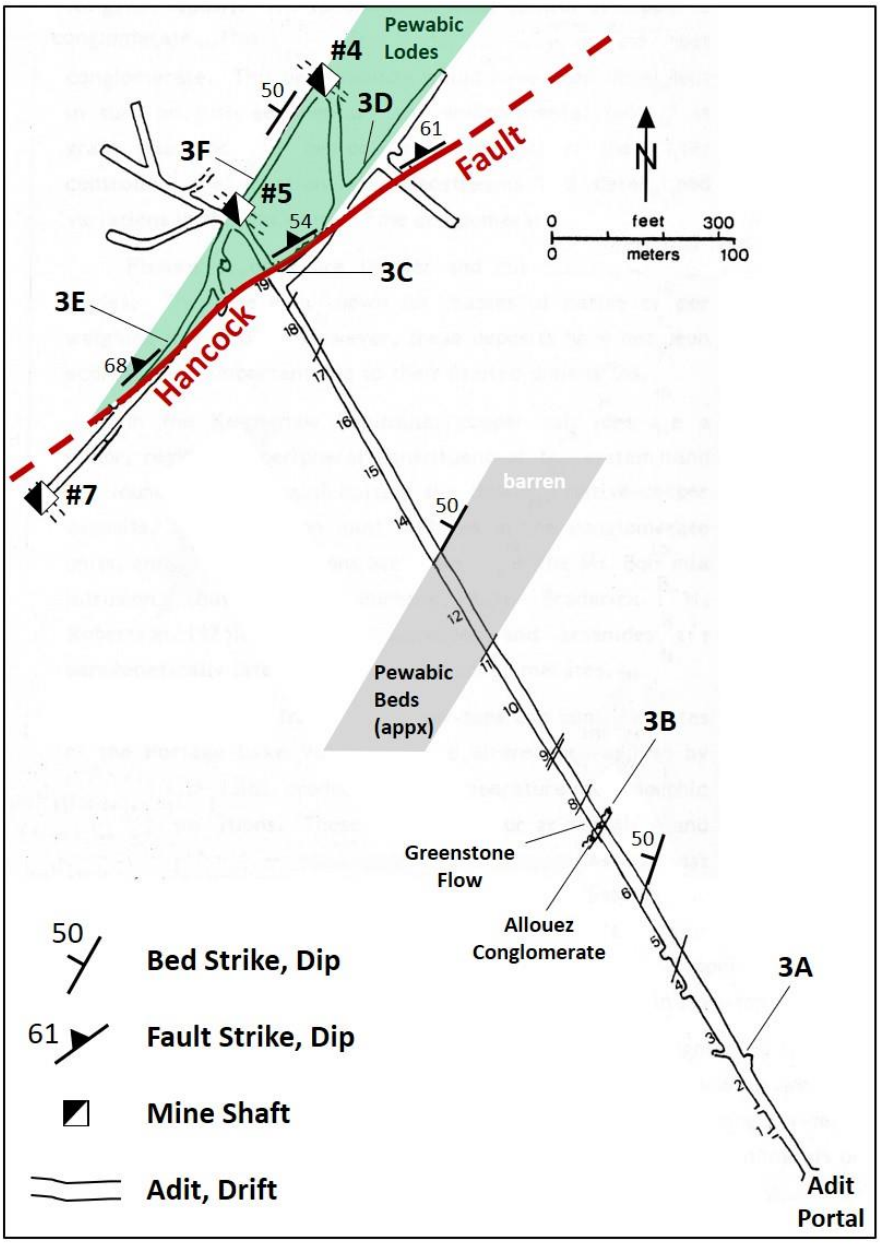


Figure 6: Map of Quincy adit and 7th level mine workings (modified from Bumgarner, 1980). Six geological sites of interest are labeled 3A through 3F. Numbers 1 – 19 along adit are distances in hundreds of feet.

Site 3A: Contact between two basaltic lava flows. [~75 m from portal]

The contact between the two lava flows at this site shows many characteristics that are common in the Portage Lake Volcanics of the Keweenaw Peninsula. Contacts between successive subaerial lava flows are recognized by textural and color differences between the top of the older layer and base of the younger one, as well as by the geometry of their boundary. Massive flow interiors grade into margins with finer grain size and abundant vesicles often arranged in bands parallel to flow contacts. Flow tops have the highest abundance of vesicles. Upper flow surfaces were exposed to the atmosphere and escaping gases during and after emplacement, and later to hydrothermal fluids permeating along flow tops that filled many voids (amygdules) with an assemblage of minerals (Butler and Burbank, 1929; Stoiber and Davidson, 1959; White, 1968; Bornhorst, 1997), thus modifying their color as seen here.

Looking at the adit's northeast wall (right side walking in), the top of the older flow toward the southeast is highly vesicular and has a lighter greenish tone relative to the younger to the northwest. In this case, the distinctive lighter tone and greenish tinge of the flow-top rock results from secondary epidote, pumpellyite, and various copper minerals, which are not present in the base of the overlying flow. The adit cuts nearly perpendicular across the contact between the flows, such that their boundary can be traced from one side of the adit across its ceiling to the other side. This exposure is nearly ideal for measuring strike and dip of the contact because it is exposed in 3-D over a sufficient distance to allow visual averaging of irregularities that are typical of flow contacts.

Site 3B: Allouez conglomerate and the Greenstone flow. [~290 m from portal]

The Allouez conglomerate and overlying Greenstone flow are well correlated stratigraphic units along the strike of the historic copper mining district (Butler and Burbank, 1929; Stoiber and Davidson, 1959). They can be traced from near the tip of the Keweenaw Peninsula to southwest of the Quincy Mine, a strike length of 80 kilometers. In the Quincy adit, the Allouez conglomerate is a 2-m-thick layer and the overlying Greenstone flow is only 10-15 m thick, far less than its maximum thickness elsewhere. The conglomerate here is typical of most conglomerates interbedded with mafic lava flows of the Portage Lake Volcanics. It is largely clast-supported and contains rounded to subrounded cobbles and pebbles of felsic volcanic rocks and subordinate mafic rocks. At this site, the bedding surface between conglomerate and basalt is marked by a clay seam, or "fluccan" following Cornish mining terminology (Hubbard, 1898). Hubbard (1898) systematically documented such clay seams in mines accessible at the time and he noted that they typically follow contacts between interflow sedimentary layers and lava flows, but also occur at contacts between flows. The clay seams are commonly associated with polished surfaces, fault gouge, brecciation and alteration of adjacent units, and secondary mineralization in fractured zones generally less than a couple of meters thick. As noted by Hubbard (1898), these phenomena indicate that many layer boundaries in the Portage Lake Volcanics have slipped during one or more deformation events. Such surfaces are essentially layer-parallel faults that slipped because of their weaker mechanical strength relative to the layers themselves.

Site 3C: Hancock fault at the adit and in drift to the northeast. [~570 m from portal]

The Hancock fault is a major splay of the Keweenaw fault system that intersects the main Keweenaw fault about 14.5 km northeast of here and downdip at a depth of around 3 km (Cannon and Nicholson, 2001; DeGraff and Carter, 2023). Over its 18-km length, the Hancock fault strikes 234° (right-hand rule) and cuts obliquely across the exposed Portage Lake Volcanics in a direction that is ~25° clockwise from formation strike and dips northwest at a steeper angle than layering. It cuts the Pewabic flows of the Quincy and Hancock mines.

The Hancock fault had a strong influence on the distribution of copper in these two mines as follows:

1. A portion of the fault in the Hancock Mine had copper ore that was exploited;
2. Copper is also abundant along the Hancock fault in the Quincy Mine;
3. Copper mineralization in Pewabic beds of the Quincy Mine is restricted to the hanging wall northwest of the fault.

The Hancock fault has been proposed as an important pathway for ore-forming fluids to access porous zones at the Quincy and Hancock Mines (Bornhorst et al., 1986).

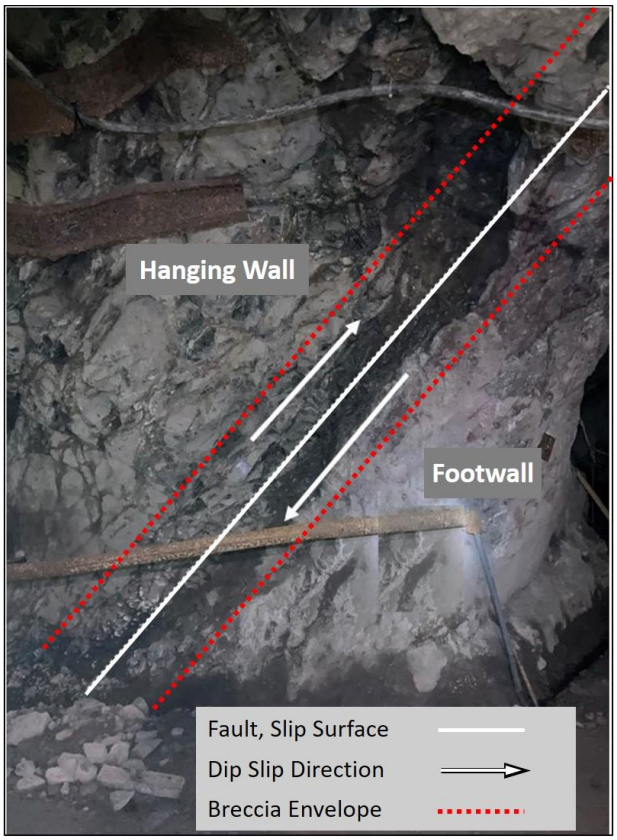


Figure 7: Cross-sectional view of the Hancock fault looking northeast at the wall of the adit. Height of the view is about two meters. The slip surface is marked by a clay seam.

The Hancock fault is best exposed and investigated in the Quincy Mine workings, where it is cut by the adit and drifts leading northeast and southwest from the adit (Figure 7). At this site, the Hancock fault is observed crossing the adit and following the northeast drift for about 50 meters, so that measuring its orientation is easy. At the adit, the fault strikes 239° and dips 54° NW, in contrast to stratigraphic layering that strikes 213° and dips a bit less than layering. The fault zone here has a thin (~3 cm) medial clay seam, marking the main slip surface, that is contained within a breccia envelope of disaggregated fragments. Outside of the breccia envelope, country rock is highly fractured near the fault but the rock is still intact. As the fault is followed northeastward along the drifts orientation varies and the thicknesses of its clay seam and surrounding breccia increase and decrease and are not always symmetrically arranged.

Site 3D: Ropy pahoehoe at the top of a Pewabic flow in the first northeast drift. [~570 m from portal; ~75 m along drift]

Farther along the same drift at Site 3D is a good exposure of the surface of one of the Pewabic flows (Figure 8). Like subaerial mafic lava flows elsewhere, the tops of Portage Lake Volcanics flows show characteristics of two fundamental types – a smooth or ropy geometry (pahoehoe, locally termed amygdaloid) and a rough blocky geometry (aa, locally termed fragmental amygdaloid). Most Portage Lake Volcanics lava flows have smooth flow tops and yet ropy pahoehoe is rarely observed, probably because plan views of flow surfaces are uncommon and because of degradation of flow surfaces between successive extrusive events. The ropy pahoehoe observed on the northwest side of the drift is well preserved. The view here is upward at the base of an overlying flow, and so the pahoehoe feature is really a mold of the upper surface of the underlying flow top that has been removed along the drift. The curved geometry of pahoehoe ropes results from local movement of partly solidified lava crust that forms arcuate patterns that are convex in the local direction of flow (Fink and Fletcher, 1978; Self et al., 1998). The local flow direction indicated by the convexity of the ropes seen here is updip and toward the southwest, i.e. from the center towards the edge of the rift.

Most flow-top native copper ore bodies are hosted by brecciated flow tops because their relatively large volume of open space was well connected and provided relatively easy movement of ore-forming fluids (Butler and Burbank, 1929; White, 1968; Bornhorst, 1997). Flows with smooth to ropy pahoehoe tops are usually not favorable because vesicles have limited volume of not well connected open space which hindered movement of ore-forming fluids. However, Pewabic flows with pahoehoe tops had economic grades of copper mineralization because they had large connected openings up to 1.5 m wide and extending up to 100s m along strike. Butler and Burbank (1929) proposed that the exceptional Pewabic lode was the result of an extremely gaseous flow that produced a great abundance of vesicles that coalesced to produce large connected voids. An alternative hypothesis is that the large openings resulted from lava draining from pools and tubes beneath a solidified flow surface.

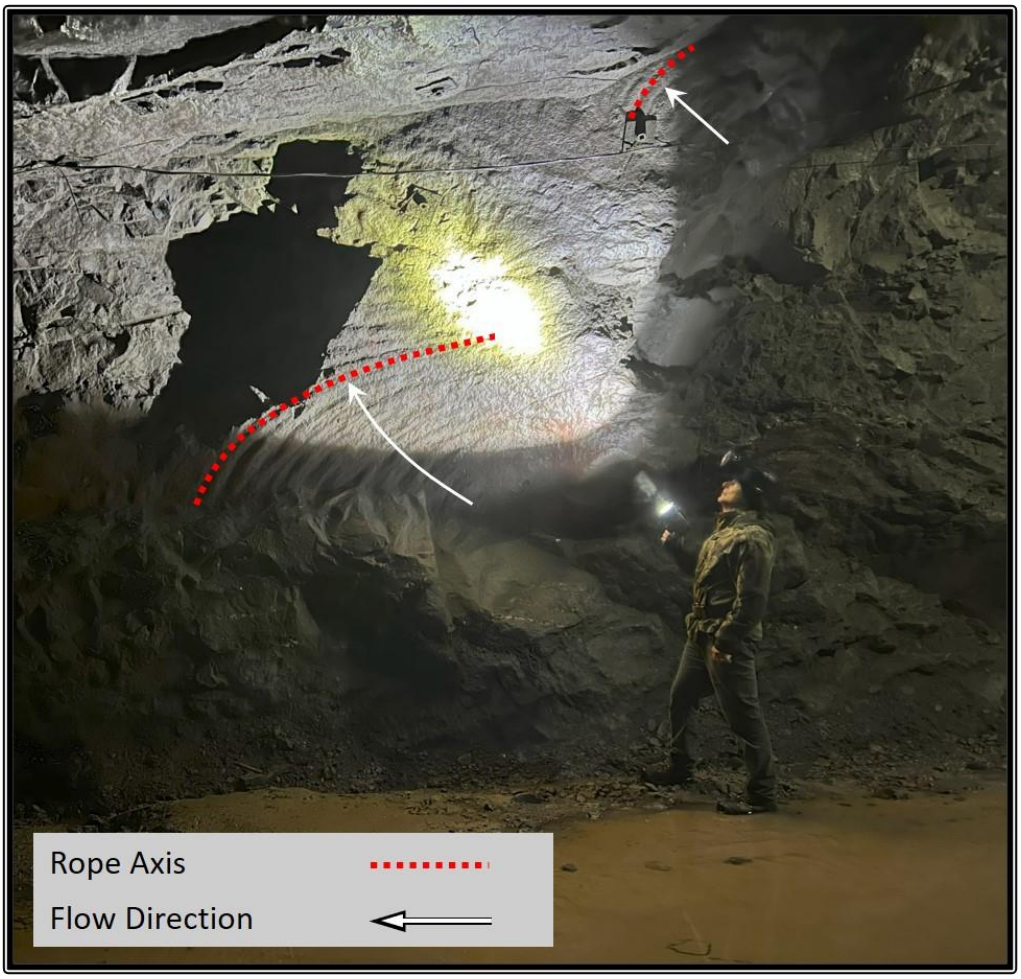


Figure 8: Ropy pahoehoe in the hanging wall of the first drift northeast from the adit. View is to the northwest. The two highlighted patterns indicate flow updip and toward the southwest. Height of the view is about six meters.

Site 3E: Hancock fault in drift to the southwest. [~675 m from portal; ~55 m along drift]

This site provides another opportunity to examine the Hancock fault where it follows much of the drift leading southwest from the No. 5 shaft to the No. 7 shaft (Figure 6). From Site 3C at the adit, the fault angles across the corner of unmined rock between the adit and southwest drift and is next exposed on the southeast wall of the drift at Site 3E. From here, the fault trace rises up the southeast wall, gradually angles across the ceiling, and descends the northwest wall where it enters the drift's hanging wall before reaching the No. 7 shaft.

Native copper is frequently found in the fault's hanging wall up to the breccia zone but has not been found in the footwall. Here, the "medial" clay seam is generally thicker than at Site 3C but its thickness varies considerably along the fault trace, as does the thickness of the disaggregated breccia envelope. It is difficult to estimate the thickness of the still-intact zone of fractured rock

that encloses the breccia zone because, as seen at other faults examined in detail (Caine et al., 1996; Caine and Forster, 1999), its intensity gradually decreases away from the fault to a background value for the system.

Study of exposures of the Hancock fault and nearby satellite faults in the Quincy Mine is needed to better understand the slip characteristics and wall-rock modification of the Hancock Fault. This could also help understand the influence of the Hancock fault on copper mineralization. Fault slip is being investigated by the measurement and analysis of slip indicators – slickenlines and steps – on the surfaces of the many small faults associated with the Hancock fault. Observations of how rocks along the Hancock fault were modified physically and chemically have been mostly qualitative so far, though with some exceptions (Bornhorst and McDowell, 1992; Langfield et al., 2023).

Site 3F: Hanging wall of Pewabic flow in drift northeast of No. 5 shaft. [~675 m from portal; ~30 m along drift]

Along the drift leading northeast from the No. 5 shaft to the No. 4 shaft (Figure 6), the hanging wall of the drift and attached stope exhibit slickenlines directed parallel to dip of the lava flows. One may need to search a little and use oblique lighting to see these fault surface features, which manifest layer-parallel dip slip on the boundary between two lava flows. Another surface with similarly oriented slickenlines occurs in a parallel drift a few tens of meters northwest beyond the hanging wall of this drift. As discussed at Site 3B, Hubbard (1898) documented 12 layer boundaries with layer-parallel slip, but this surface and the other northwest of here were not among them. He mentioned polishing and slickensides on such surfaces but did not specify their orientation. It is likely that such layer-parallel slip is far more common than has been observed because, like the ropy pahoehoe texture, it can only be observed where an ideal exposure permits viewing. The documented layer-parallel slip within the PLV section is strong evidence for a detached style of thrusting, which was used recently to model the cross-sectional geometry of the Keweenaw and Hancock faults (DeGraff and Carter, 2023).

Stop 4: Quincy Mining Company Processing Plant and Dredge

Latitude: 47.146295N*; Longitude:* -88.460474W,

Directions: From Quincy Smelter Works turn right (east) and continue 5.7 miles (9.2 km) to ruins of the Quincy processing plant on left carefully pulling into open lot on west side of ruins. The Quincy Mine dredge No. 2 is located on the edge of Torch Lake on the other side of the road.

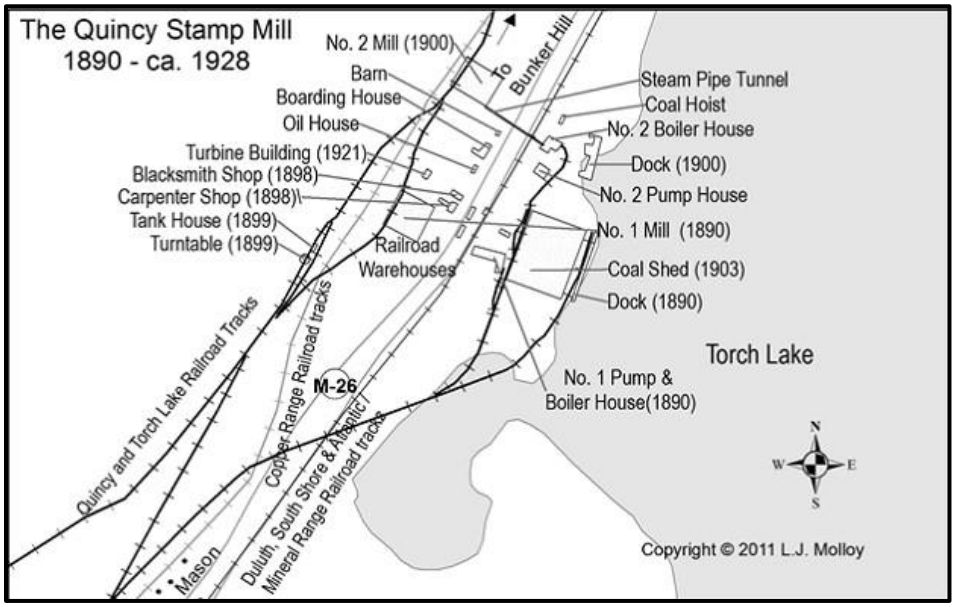


Figure 9: The Quincy Mining Company mills, 1890-1928. Star shows the location of the Quincy dredge No. 2. From Molloy (2011) with permission.

The Quincy Mining Company had to move their mill from its Portage Lake site below Quincy Hill because tailings deposited in the lake were beginning to hinder navigation. Construction began on the Quincy mill at this site in 1888 (Figure 9). Mill No. 1 began with three rock crushing stamps and two additional were added in 1892. The building closest to the dredge on the north side of the road contained the 1890 mill which was modified over time. The square building adjacent to it was a turbine building that was built in 1921. As production increased, Stamp Mill No. 2 was built to the north of the No. 1 mill in 1900 and had three stamps.

Underground mining activities at the Quincy Mine focused on mining ore (rock with sufficient recoverable copper to make a profit). Inevitably the mine also produced waste (uneconomic) rock along with the ore rock. Some of this waste rock can be seen today throughout the Keweenaw Peninsula as poor rock piles. Some of the waste rock was used underground as fill for already mined out stopes. The broken ore from underground blasting that is small enough to pass through a coarse grating was sent by train to the processing mill.

The purpose of the mill was to remove as much rock as possible from the native copper producing a product where the percent of contained copper is up to 20 times higher than the percent of copper in the ore. The copper-rich product is sand to smaller sized fragments of native copper mixed with similar sized rock and mineral fragments (termed “copper concentrate”). The inefficiencies in separating native copper from rock and minerals fragment results in some amount of waste rock and minerals in the copper concentrate. At the end of processing in the mill, most of the rock and minerals end up as fine sand to sand of waste containing only a small amount of copper (termed tailings).

The Quincy Mill used several techniques to separate the copper from the rock and minerals. The essential first step in milling begins with crushing the ore rock to a small enough size to liberate the unwanted rock and minerals from the native copper. The ore rock and minerals containing disseminated copper of various sizes was crushed by steam stamps. Much of the rock was liberated from the native copper by crushing the broken ore fragments derived from mining to fragments 0.0165 to 0.188 inches in diameter (fine to very fine sand size). Larger masses of native copper were removed prior to crushing at the mine and sent directly to the smelter. Since the density of native copper is much higher than the liberated particles of rock and mineral, gravity and water methods could be used to separate the waste rock and minerals (tailings) from the copper. Early mines used jigging to concentrate the native copper from the tailings. Jigging was a well-developed mineral separation technology prior to the first mining of native copper in the Keweenaw Peninsula beginning in 1845. Jigging is accomplished by placing the sand sized particles on a screen where pulsating water allows the heavier copper particles to settle while the lighter particles rise to the top and overflow the screen or are skimmed off the top as waste tailings. Jigging resulted in a “concentrate” of sand sized native copper particles with some rock and minerals that were not copper-bearing (tailings), about 50 % copper and the rest waste rock and mineral. The concentrate was sent to the smelter to be turned into nearly pure copper.

The waste tailings were dumped into nearby Torch Lake. However, using jigging up to 25 % of the native copper was incorrectly classified as waste (tailings) and dumped the copper was lost as the tailings were dumped into nearby Torch Lake. Over time new technologies were introduced into the mills by Quincy to increase efficiency and loose less copper into the tailings. Quincy used froth flotation in its processing circuit as early as 1920s. Violent agitation of copper and rock/mineral particles in an oily water along with frothing agents and other chemicals cause bubbles to rise to the surface with copper particles attached to their surface. The bubbles and copper particles were captured from the top of the flotation tank while the waster rock/mineral (tailings) sank to the bottom. and was slurried to Torch Lake. The heavier copper particles were carried upward by the floatation bubbles rather than sinking due to gravity by jigging. The mineral processing circuit could begin with jigging and with the tailings after jigging sent to floatation cells to recover more copper. Later Wilfley tables were added to the mineral processing circuit to help minimize loss of copper from finer grain sizes. A Wilfley table utilizes gravity and water to separate denser particles. A shaking ribbed table with a film of water flowing along the long axis results in the higher density copper particles concentrating in beds behind the riffles.

The Quincy Mining Company knew that the tailings contained a lot of copper that they were unable to recover with technology available at that time. The company had considerable foresight and kept detailed maps each year of the copper content of the tailings that were deposited into the lake. As technology improved, they were able to reclaim copper from the tailings at a profit. Quincy built a special reclamation processing mill near here in 1942-43 for $1.2 million. The main building, 124'x255', had six Harding ball mills to grind the tailings even finer than the stamp mills in order to release more fine particles of copper from the rock/mineral and facilitate addition of Wilfley tables and flotation cells to mineral processing circuit.
Across the road is Quincy Mining Company dredge No. 2 on the shore of Torch Lake. Torch Lake was filled with several 100 millions of tons of tailings since not only did Quincy Mine operate a mill on its margin so did many other companies, especially Calumet and Hecla Mining Company. The tailings were sucked up via the dredge and sent to the reclamation mill to recover more copper from the tailings. The reprocessed tailings were redeposited back into Torch Lake. This dredge was built in 1913 by Calumet & Hecla Mining Company. In 1951, the Quincy Mining Company purchased the dredge and it became known as Quincy Dredge No. 2. It could process over 10,000 tons of tailings per day and it had 141 ft suction pipe that could work 115 ft below the surface of the lake (Figure 10).

Tailings were conveyed to the mill via a tube held up with pontoons. Quincy recovered 100 million lbs. of copper from tailings in Torch Lake from 1943 until it closed in 1967 or about 10 % of total production. In the 1800s and early 1900s, depositing 100s of millions of tons of tailings into Torch Lake was acceptable practice and the environmental consequences were not considered. Today, the environmental impact of tailings must be carefully considered to obtain a permit, "social license" to operate a mine. The mining companies did not just put tailings into Torch Lake, they also used it to dispose of other waste such as that from electrical systems and barrels filled with chemicals. The early processing plants used only water to separate the copper from the rock. Later floatation cells were used in processing ore from the mine and in processing reclaimed copper from the tailings. The floatation cells used chemicals in water. These chemicals along with the tailings were put into Torch Lake. Discovery of fish with tumors in Torch Lake led to its being designated an U.S. Environmental Protection Agency Superfund site. Some of these chemicals were biodegradable and are no longer present in the Torch Lake water but would have been present decades ago, thereby could have readily caused tumors in older fish. There are far too many tons of tailings in Torch Lake to remove them and thus, the mitigation strategy is to simply cover them with soil. Fortunately, the copper ores of the Keweenaw Peninsula lack pyrite that is known under certain environmental settings to produce acid drainage and in turn can result in significant environmental impact. The lack of acid generating potential and the otherwise mostly inert minerals in the waste rock has greatly lessened the environmental impact of the tailings themselves. Today crushed poor rock from the mines is used directly as aggregate and is also crushed for use as aggregate.

Text from previously published field trip guide by Bornhorst (2022) with limited modification.

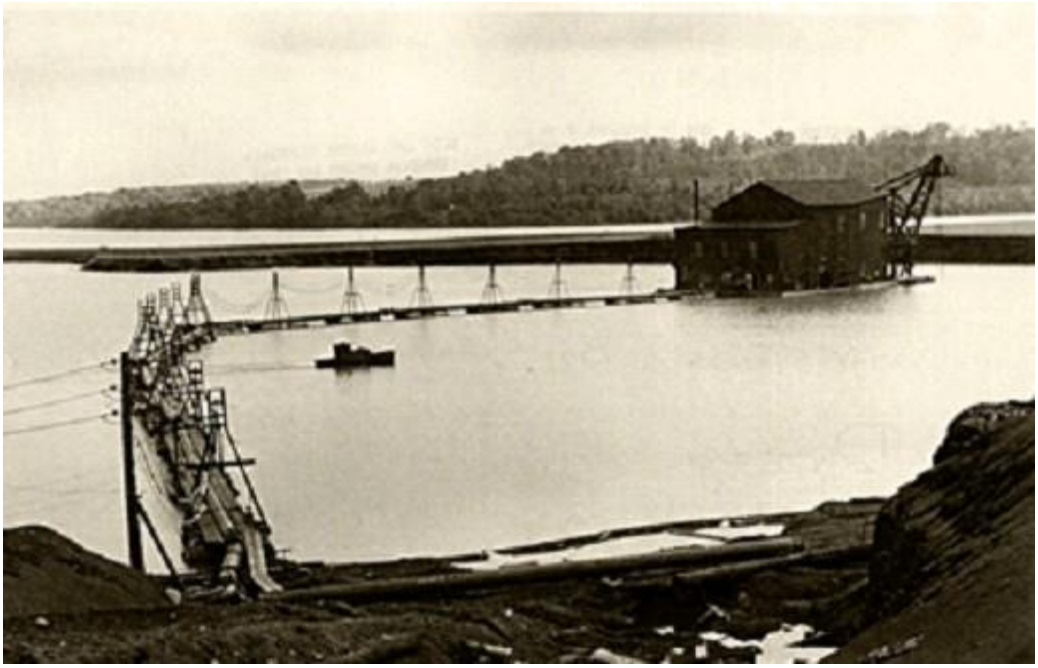


Figure 10: Quincy Mining Company dredge No. 2 working on Torch Lake, ca. 1920s. From Molloy (2011).

Stop 5: Ahmeek Mining Company Stamp Mill

Latitude: 47.168633N*; Longitude:* -88.435810W

Directions: From Quincy Mill Ruins and Dredge carefully pull out of gravel lot and continue east on M-26 for 1.9 miles (3 km) to stamp mills of the Ahmeek Mining Company.

This is the only remaining steam-powered stamp in the Keweenaw Peninsula. The Ahmeek Mining Company had four of these Nordberg compound steam-powered stamps installed in 1910, and four additional stamps were added in 1914. Rail cars brought rock to the mill using the trestle, located above the trees across the street from the stamp. The compound-expansion nature of the stamps represented a major improvement in processing of copper ore. The stamp could strike approximately 104 24-inch blows per minute. The mill could process approximately 7,000 tons of ore in a 24-hour period.

This is not part of the Quincy Mine but we stop here to see a steam-powered stamp used for crushing the native copper ore.

Text from previously published field trip guide by Bornhorst (2022) with limited modification.

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Field Trip 3

Geoheritage of Buffalo Reef: Industrial Impact on Land, Culture, and Fish Sovereignty

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INTRODUCTION

Buffalo Reef is a geoheritage site with scientific, educational, cultural, and aesthetic value. This field trip explores the relationship between geology, mining waste, and culture of Buffalo Reef - a 2,200-acre natural cobble feature of Lake Superior's lakebed southeast of the Keweenaw Peninsula and about 20 miles northeast of Houghton. Finely crushed waste rock - stamp sand - from copper ore milling operations at the community of Gay have been moved by currents along the shoreline of the Keweenaw Peninsula to Big Traverse Bay, thus covering this highly productive spawning ground for lake trout and whitefish. This has negative implications for commercial fisheries, local economies, subsistence uses, and the spiritual, physical, and cultural well-being of tribal nations that identify as fishing people. Further, this has adverse impacts on tribes' ability to exercise their treaty rights to fish in this area (Buffalo Reef Task Force, 2024).

Geoheritage is a nascent yet evolving field in the United States that considers the protection, interpretation, and management of geologic features with significant scientific, educational, cultural, or aesthetic value (Brocx & Semeniuk, 2007; Geological Society of America, 2017; National Park Service & American Geosciences Institute, 2015; Reynard & Brilha, 2017). More

distinctively, geoheritage emphasizes the importance of the varied personal values people have for geologic features and explores the wide-ranging *relationships* people have with landscape. The rich geodiversity of the Keweenaw has fostered relationships for millennia and has imbued our place with significant sites that provide opportunities to broaden both Earth science and cultural literacy (Rose & Vye with Martin, 2017; Vye, 2016). The rich geosites of the Keweenaw provide an accessible platform for people to learn about deep time, Earth's dynamic processes, and importantly, the diverse relationships and reciprocity people have with our geologic underpinnings.

The Keweenaw is renowned for Earth's largest native copper deposits that became the first great copper mining district in the United States. From 1845 to 1968, over 11 billion pounds of refined copper were produced in Keweenaw mines, making it a cornerstone of the American economy in the 19th and 20th centuries (Bornhorst & Lankton, 2009; Bornhorst & Barron, 2011). Interpretations and public educational programming depicting the relationship between people and geology have largely focused on stories and heritage associated with the European diaspora that fueled the Copper Boom of 1845-1968 and how copper's prolific use for transatlantic cables, telegraphs, electricity, and the auto industry ultimately modernized the country (Bornhorst & Lankton, 2009). Less interpreted, yet central to our history, is how geology has shaped, and continues to shape, the ancestral, traditional, and contemporary lands, waters, and livelihoods of the Anishinaabeg- the Three Fires Confederacy of Ojibwe, Odawa, and Potawatomi peoples and their many more-than-human relatives. As such, this field trip visits three stops (*Figure 1*) to explore mining impacts on culture, subsistence uses, and the wellness of all beings.

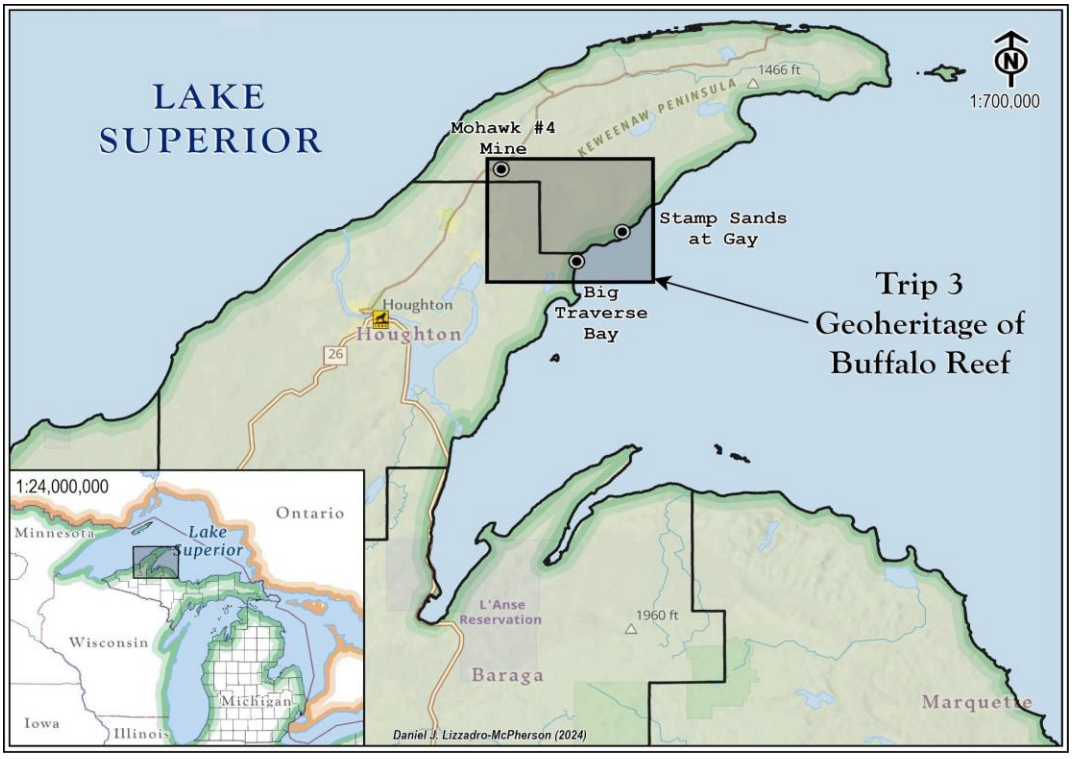


Figure 1: Map of field trip site locations, courtesy of D.J. Lizzadro-McPherson, MTU

***Stop 1 - Mohawk, MI*:** The trip begins at the poor rock pile where the Mohawk No. 4 mine shaft once stood - *a native copper host rock from which some of the Gay stamp sands were generated.*

Stop 2 - Gay, MI: Participants will then travel to the town of Gay to *walk the stamp sands* and learn how Tribal, State, Federal, and academic partnerships are collaborating to *mitigate environmental damage and ultimately restore Buffalo Reef to the ecological resource that has sustained both tribal and non-tribal communities for generations.*

Stop 3 - Big Traverse Bay & Buffalo Reef, MI: From Gay, we will travel to Big Traverse Bay to learn more about *why shoreline and habitat restoration efforts are necessary and how stakeholders and rights holders are working together to address this environmental justice issue in our community*. Participants will benefit from visiting Buffalo Reef aboard Michigan Tech's Research Vessel Agassiz. This part of the trip enables participants to compare healthy parts of the reef with areas that have been inundated with stamp sands.

STOP 1 - MOHAWK, MI

At this site, we will learn more about the source rock for the Gay stamp sands after a brief overview of the geological history of the Keweenaw region. Participants will have time to explore the poor rock pile before departing for Gay.

Directions: Leave downtown Houghton and head north on U.S. 41 towards Calumet. Stay on U.S. 41 until you reach Mohawk; turn right on 6th street, just past E Phoenix Street there is a place to park on the left-hand side of the road. This space provides access to the Mohawk No. 4 poor rock pile (Figure 2).

Coordinates: 47.301203, -88.363369

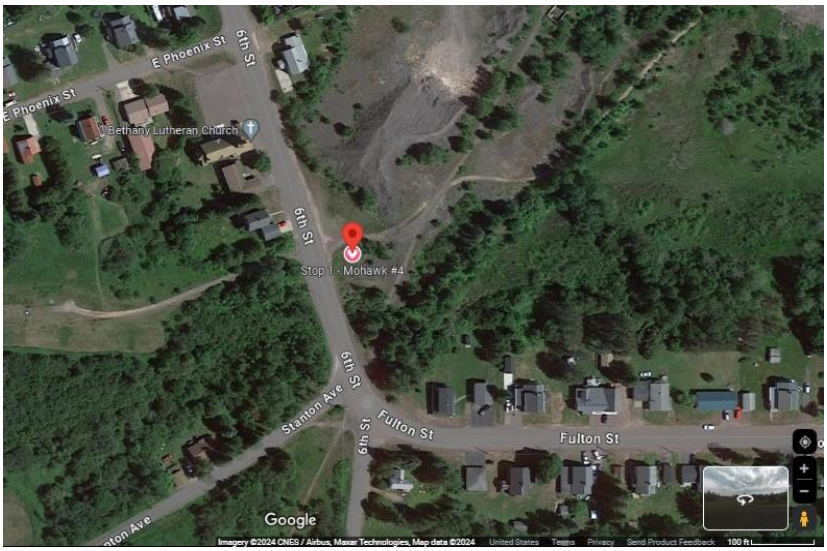


Figure 2: Parking location to visit Mohawk No.4

Geological Origin Story

Michigan's Keweenaw Peninsula has a rich and globally significant geodiversity in tandem with a fascinating cultural story. This is the site of the largest native copper dominated deposits known on Earth, the oldest metal workings in the Western Hemisphere, and a recent diaspora of

European cultures that flocked to the region for copper mining in the late 1800s – this has shaped an entangled mosaic of cultural, mining, and industrial heritage.

Importantly, the Keweenaw Peninsula offers an important window to Earth's past, exposing the heart of the Mesoproterozoic Mid-Continent Rift. Located on Lake Superior in Michigan's Upper Peninsula (*Figure 3*); the abundant geodiversity sites are the result of a flood basalt sequence comprised of hundreds of voluminous lava flows, interbedded and covered by a redbed clastic sedimentary rocks. An upwelling of heat and magma from a hot spot initiated great lava flows erupting from the rifting of supercontinent Rodinia. The rift created a ~3000 km U-shaped feature in the center of North America that extends from Kansas, through what is now Lake Superior and terminated in what is now Michigan (Cannon, 1994, Cannon and Nicholson, 2001, Stein et al., 2015).

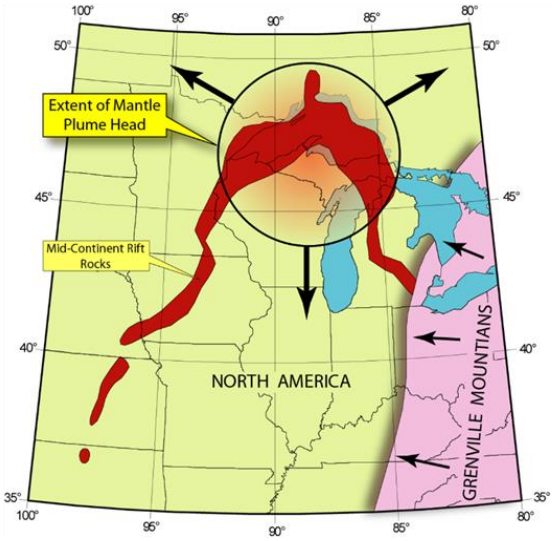


Figure 3: Extent of rifting associated with the Mid-Continent Rift (K. Schulz, USGS)

Some of the largest lava flows on Earth erupted and ponded in the Mid-Continent rift over many centuries (Huber, 1983). During quiet times, red-brown conglomerates and sandstone were deposited between flows in high-energy alluvial fans (Elmore, 1984). The interbedded lava flows and sedimentary layers were normally faulted resulting in a syncline feature that extends from Isle Royale to the Keweenaw Peninsula, now the basin for Lake Superior (*Figure 4*). Copper was brought to the surface by hydrothermal systems mineralizing permeable layers such as the amygdaloids of lava flow tops and the cracks and crevices of conglomerate rocks (Bornhorst & Barron, 2011, Nicholson et al, 1997, Cannon, 1984).

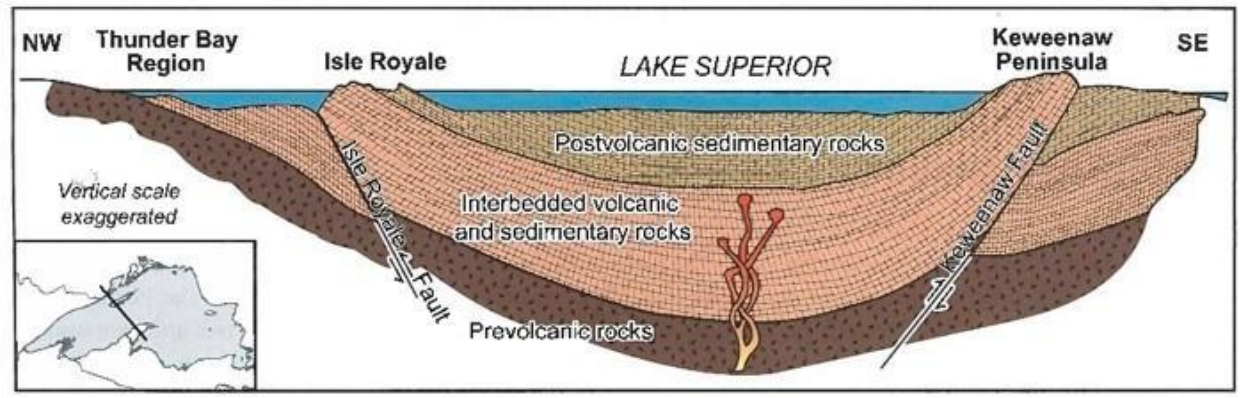


Figure 4: Cartoon depicting interbedded lava flows and minor sedimentary rocks (Huber, 1983.

The Keweenaw is noted for mining nearly 9000 years ago (Martin, 1995, 1999). The valued red metal has since been traded up and down the Mississippi by the Ojibwa caretakers of this landscape and used in ceremony today. From 1845 to 1968, ~11 billion pounds of refined copper were produced from Keweenaw Peninsula mines making it one of the cornerstones of the American economy and the first great metal mining district in the United States (Bornhorst and Barron, 2011, Bornhorst and Lankton, 2009). The region has been extensively mapped and researched since the mid-1800s because of the discovery of copper and the subsequent mining boom. This field trip explores the impacts related to one of the many mining sites in operation from the late 1800s to mid-1900s in the Keweenaw region - *Mohawk Mine, a native copper host rock from which some of the Gay stamp sands were derived.*

Mohawk No. 4

This field stop explores the site of the Mohawk mine, specifically the poor rock pile associated with the Mohawk No. 4 shaft (*Figure 5*). Rock from this site is the main source for the Gay stamp sands. Following the discovery of copper on the property in 1896 by lumberman Ernest Koch, the Mohawk Mining Company was founded. It was incorporated in 1898 and lasted until 1932 (John Stanton as president, later replaced by Joseph Gay). In 1923 the Mohawk Mining Company took over the neighboring Wolverine Copper Mining Company and the Michigan Copper Mining Company. In 1934 the company was purchased by the Copper Range Company (Molloy, 2008). The Mohawk Mine had six shafts numbered 1 through 6 running from north to south along what is now US 41. The was a profitable mine paying out over $15 million in shareholder dividends between 1906 and 1932 (Clark, 1978).



Figure 5: A group of Keweenaw youth explore the poor rock pile at Mohawk No. 4 (photo credit Erika Vye)

The No. 4 shaft of the Mohawk Mine was constructed at this site in 1901 and stayed open until the mine closed in 1932 (*Figure 6*). In the early days of operation, the shaft reached a depth of 501 feet. Technological advances in 1904 led to equipping the No. 1, 2, and 4 shafts with Nordberg Conical Drum Hoists (*Figure 7*) enabling depths of up to 6000 feet. With this technology, the shaft reached a depth of 900 feet by 1906, 1,175 feet by 1908, and by 1922 a final depth of 2,832 feet. In 1914, the No. 4 shaft was producing between 450 and 500 tons of ore per day. Mining continued from the No. 4 shaft until 1924, with a brief lull in operations until it was reopened in 1926, eventually closing in 1932 (Clark, 1978).



Figure 6: No 4 shaft in Mohawk, MI (photo courtesy of the MTU Archives)

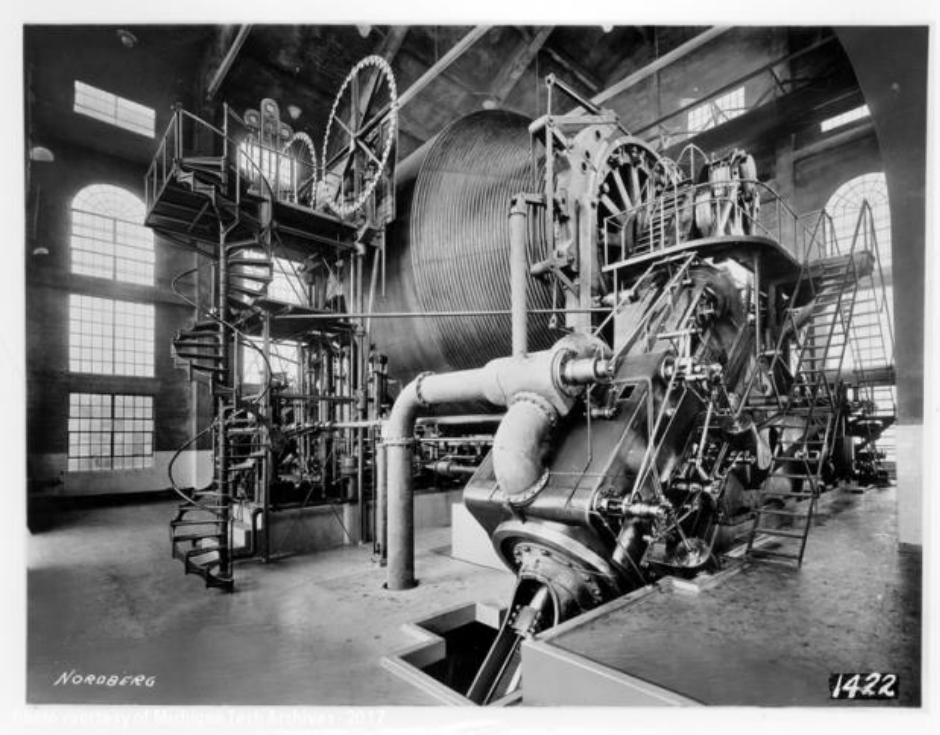


Figure 7: Image of Nordberg hoist; the world's largest Nordberg hoist can be visited at the Quincy Mine in Hancock, MI, pictured here (photo courtesy of the MTU Archives)

Ore Deposit and Significant Minerals

The Kearsarge lode at Mohawk is a native copper-dominated deposit with minor native silver hosted by basalt lava flows of the Portage Lake Volcanics (about 1.1 billion years old). The strike of the Kearsarge orebody is N45E, dips at 35W, and is 2.9 meters thick, 1,490 meters wide, with a total length of 3,250 meters spanning an area of 250 hectares. The productive part of the fragmental amygdaloidal lode was richer at greater depths than nearer to the surface where the lode consisted of more massive basalt. The Kearsarge lode was the second largest in the district and averaged 6 to 13 feet thick (Butler & Burbank, 1929; White et al, 1953; and USGS, 2005).

The fragmental amygdaloid ore of the Mohawk mines is associated with arsenic-rich minerals, more common than in most of the other Keweenaw deposits. The Mohawk mine is well known among mineral collectors for its large occurrence of *mohawkite* - a mixture of algodonite, domeykite, and copper (Moore, 1971). Mohawkite was first found on the property in 1901, on the first level north of the No. 1 shaft. In addition to copper, the mine also produced a small amount of native silver (*Figure 8*).

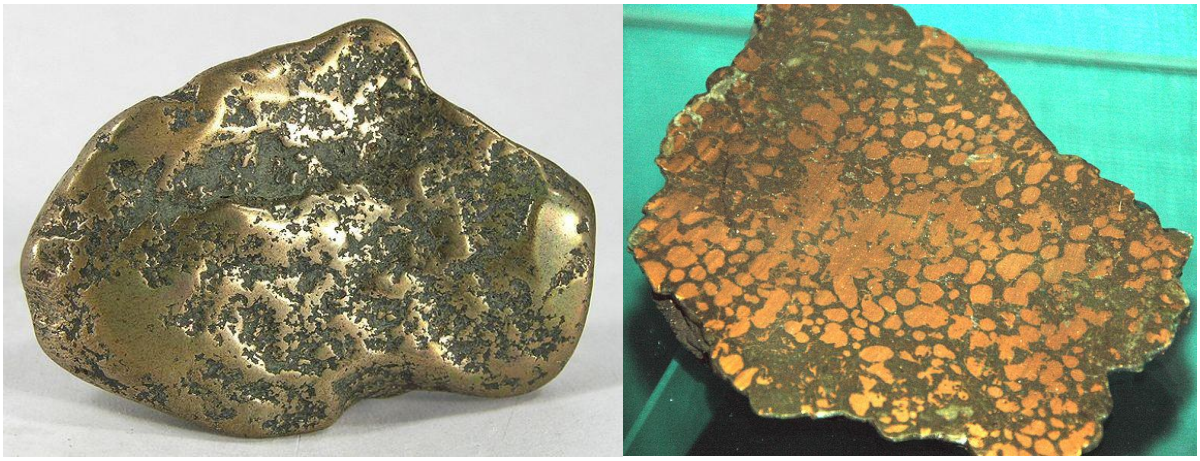


Figure 8: Left -Polished mohawkite, a rare mixture of copper and copper arsenides, is named after the Mohawk-Ahmeek area of the Keweenaw (Photo by Robert M. Lavinsky); Right – native copper in amygdaloidal basalt.

Geologic features of note include the occurrence of the Mohawkite Fissure that crossed the Kearsarge lode. From 1900-1901, 105 metric tons of mohawkite ore was produced from the Mohawk Mine at the No. 1 shaft. From 1902 through 1925 the No. 1 shaft produced about 117,000 metric tons of refined copper. Additional veins were discovered south of the No. 2 shaft in 1901 resulting in 230,000 pounds of mohawkite (Butler and Burbank, 1929).

Mohawkite proved to be challenging in the Keweenaw as smelters were not able to process it on account of the high arsenic levels and the deadly fumes produced. A unique smelter was built in Hackensack Meadows, New Jersey for the specific purpose of processing the ore; it became

operational in 1901. The mohawkite ore contained mostly copper and arsenic, it also contained small amounts of nickel and cobalt, as well as about 20 ounces of silver per ton of ore. (Stevens, 1902)

Copper-bearing rocks were transported 11 miles by rail to the town of Gay for milling near the mouth of the Tobacco River on Traverse Bay. This is the next stop on the field trip.

STOP 2 – GAY, MI

This part of the field trip begins at the *Gay Sands* sign just outside of the town of Gay. This stop is intended to engage participants in a discussion of how the stamp sands were generated and to learn how Tribal, State, Federal, and academic partnerships are collaborating to mitigate environmental damage and ultimately restore Buffalo Reef to the ecological resource that has sustained both tribal and non-tribal communities for generations. Participants will have an opportunity to walk the stamp sands (accessed just south of the *Gay Sands* sign): note that at times large trucks are moving through this area; always exercise caution and stay together.

Directions: *From Mohawk, travel to the town of Gay; follow the Mohawk-Gay Road. When you arrive in Gay, turn right on Main Street, then left on 2nd Street (which becomes the Gay Lac la Belle Rd), park on the right-hand side of the road beside the smokestack and local signage that shares information about the Gay stamp sands (Figure 9).*

Coordinates: 47.226766, -88.161933



Figure 9: Left - Gay Sands signage; Right – Gay mill smokestack (photo credit Erika Vye)

How were the stamp sands created?

Liberating the copper from the host rock required a large water source - in this instance, Lake Superior. In 1900, the Wolverine mine and mill opened in Gay. This piqued the interests of New York financiers of the Central, Atlantic, and Baltic mining companies and shortly thereafter the Mohawk mines and mills were opened. The two mills were built in Gay by the Mohawk and Wolverine mining companies enabling the processing of copper ore. A railroad was built to

transport the copper-bearing rocks from Mohawk to Gay; poor rock was left at the mine sites in Mohawk remaining in piles today. A dock was built nearby to import coal to power operations and to load copper on ships to transport to other places (Keweenaw County Historical Society). *Note:* A QR is included in the resources section of this guide leading to the recently developed *Gay Mills Exhibit* created by the Keweenaw County Historical Society.

An unusual landscape & vastly different soundscape

When visiting this site take note of what you hear; try to imagine what the soundscape of this place might have been like in the early 1900s. At that time, the stamps at both mills were running constantly. Imagine the ground shaking with over 70 stamps per minute crushing pieces of rock brought up from underground to free the copper from the matrix. The next step required washing the material to separate the copper and produce a copper-rich mineral concentrate to then be sent to the smelter and formed into copper ingots (Lankton, 2005). The material was brought to the top of the mill with a railway line. To process the copper and create the concentrate, water was pumped from Lake Superior, and the tailings and waste slurry were dumped back into the lake with conveyors. The length continued to grow as the tailings built up along the shoreline (Keweenaw County Historical Society). *Figure 10* shows an image of this process on the left, with what remains of this structure today. *Figure 11* illustrates the milling process.



Figure 10: Left - Conveyor sending tailings slurry back to the lake (Photo credit: MTU Archives) during milling operations in Gay; Right - wood beams left from conveyor structure as sands are continually swept away by wind and currents over time (Photo by Erika Vye)

What is happening to the sands?

From 1900-1930 the mills flushed over 22.7 million metric tons of tailings to Lake Superior leaving a large bank of black sand, the consistency of kitty litter, along the shore. This finely crushed waste rock has moved, and continues to move, along the southeast shoreline of the Keweenaw Peninsula to Grand Traverse Bay Harbor threatening the nearby Buffalo Reef. Since the 1930s Lake Superior's fierce storms and strong currents have eroded the stamp sand bank, pushing the tailings into the lake and gradually moving them southward along the shoreline to

the Traverse River. These stamp sands contain high amounts of copper and cover 1,426 acres of shoreline and lakebed. Currently, 30% of the reef has been impacted by stamp sands; modeling predicts that by 2025 stamp sands will impact 60% of the reef. The next stop focuses on the importance of understanding the impacts of these predictions and what is happening for restoration. Before traveling to the last stop we will have lunch at the Gay Park. This site also hosts the Gay Historic School and Museum open from 1 PM to 4 PM on Wednesday and Saturday.

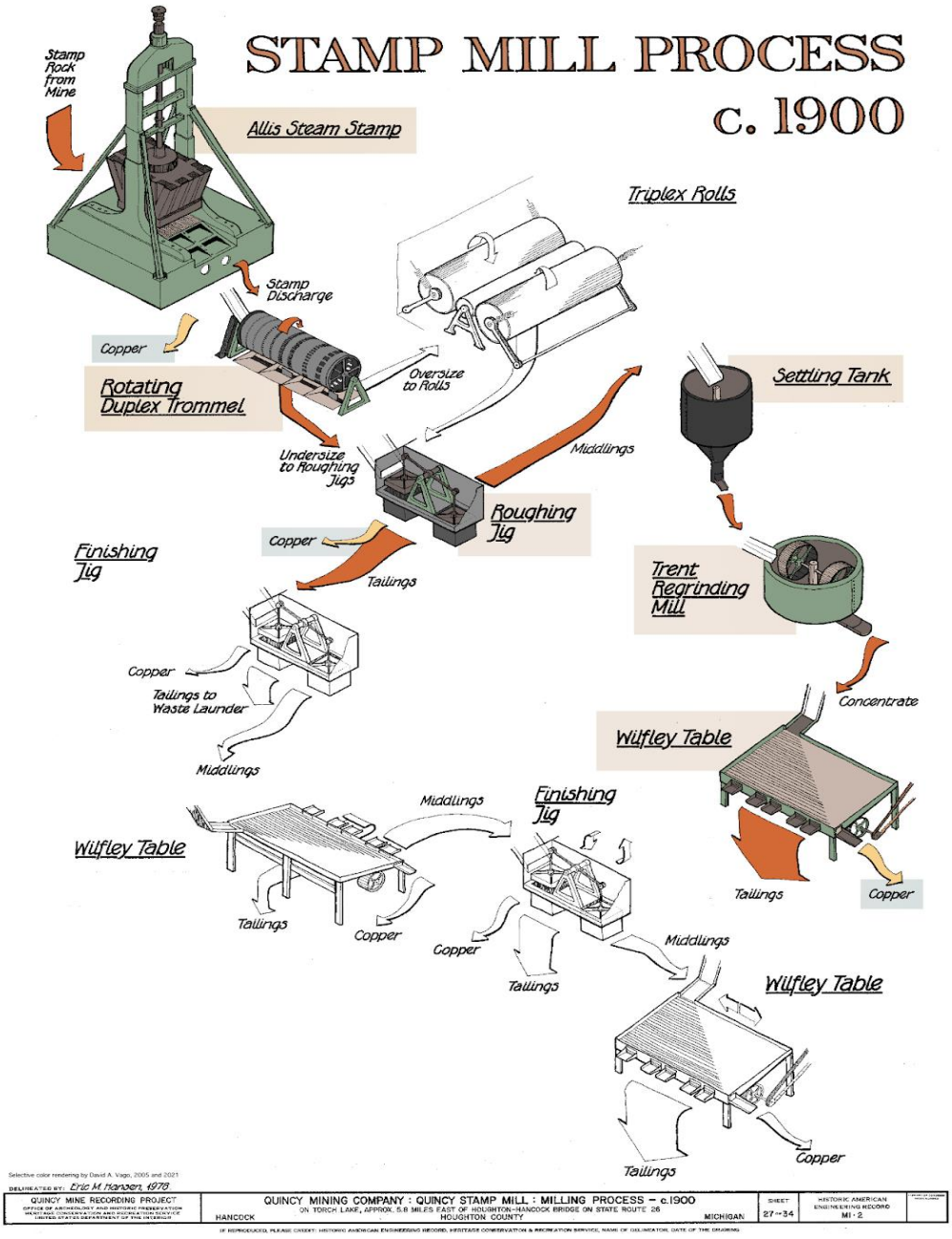


Figure 11: This image describes how copper was processed in the mills - first by gravity and large amounts of water through steam stamps, the rock was funneled into a trommel that sorted and classified the pieces by size. A jig then separated the copper from the mine rock. Next, a series of vibrating Wilfley tables separated the tailings from the small-sized copper. This image was recently modified for a Keweenaw County Historical Society exhibit on the Gay Mills (a link for the exhibit is found in the Resources section of this guide). Original delineation by the Historic American Engineering Record, Heritage Conservation & Recreation Service, Eric M. Hansen, 1978. Modification by David A. Vago, MTU, for the Houghton County Historical Society, 2005; updated 2021.

STOP 3 - BIG TRAVERSE BAY & BUFFALO REEF. MI

At this site, we will explore how the Gay sands are impacting life and culture in the region, learn about remediation efforts, and how we can share in educating ourselves and others about this environmental justice issue.

Directions: From the Gay Park by the Historic School and Museum travel left on Lake Street, then right on 1st Street (becomes the Lake Linden-Gay Rd). Turn left onto Rice Lake Rd, which becomes the Big Traverse Rd. Turn left on the Traverse River Rd, please drive slowly along this road as there are families and small children frequently playing here. This takes approximately 15 min from Gay to Big Traverse. Note that the dock is on the other side of the harbor; to get there simply head back to Big Traverse Rd, then turn left. The dock is about ¼ mile ahead.

Coordinates: Big Traverse Bay 47.189857, -88.236644, Dock to board Agassiz 47.190769, -88.237154

What is Buffalo Reef?

Buffalo Reef is a natural cobble feature in Lake Superior, located just off the eastern edge of the Keweenaw Peninsula in the U.P. of Michigan (*Figure 12*). The reef has historically maintained an invaluable spawning habitat for fish species such as lake trout and lake whitefish. Buffalo Reef is estimated to supply 33% of the tribal harvest of lake trout and lake whitefish from Michigan waters of Lake Superior (i.e. 136,375 pounds of whitefish and 61,830 pounds of lake trout average annual yield 2001-2016). Additionally, if the stamp sands migrate south of Grand Traverse Harbor, they will threaten the undisturbed native sand that serves as habitat for juvenile whitefish (Kerfoot et al, 2012 & 2021). Juvenile whitefish produced on Buffalo Reef migrate south of

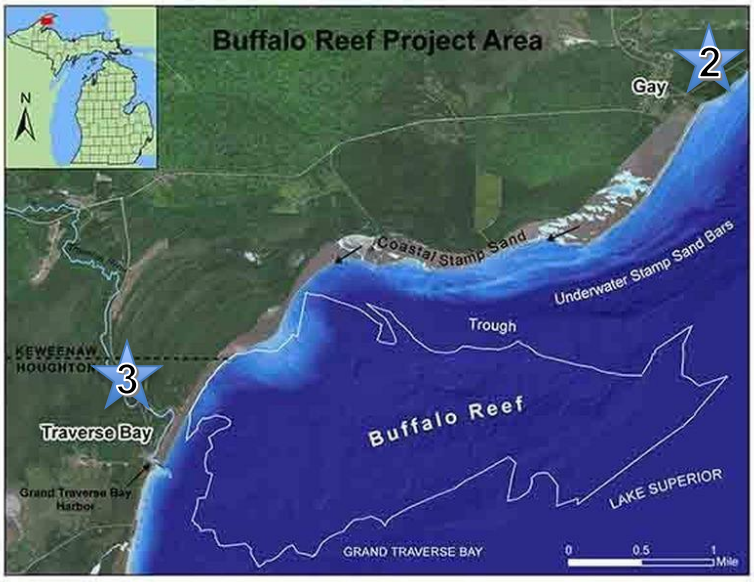


Figure 11: Location of Buffalo Reef, also not the locations of both Field Trip Stop 2 and 3, marked by stars.

Grand Traverse Harbor to sandy, shallow-water habitat where they feed before migrating to the deeper water they inhabit as adults. Without this habitat, whitefish recruitment in the vicinity of Buffalo Reef would be greatly diminished. Whitefish provide the bulk of the commercial, cultural, and spiritual value for the tribal communities that use this resource (BRTF, 2024).

How are the stamp sands impacting Buffalo Reef?

Most of the cobbles are glacial rocks, scattered around the Jacobsville sandstone bedrock highs of the reef. During spawning, fish drop eggs into the crevices between rocks. Cobbles are coated with a natural organic film, the basis of a food web for feeding fish. Stamp sands move into the northern cobble field of Buffalo Reef burying cobbles and killing living communities on rocks along the leading edge of the sands. (*Figure 13*). These impacts to fish habitat create a risk for a decline in commercially and culturally important fish keynote species. Importantly, a decline in species impacts the ability to exercise treaty rights. The sands also have high enough copper concentration (about two-eighths of a percent) to be toxic to fish and other organisms that live in Grand Traverse Bay. There are additional impacts to human health through fish consumption. Other concerns include the impacts of the copper-toxic sands on coastal wetlands - medicine chests to the KBIC - as the material migrates up streams thereby affecting plant and amphibian habitat (Kerfoot et al, 2012, 2021, *in prep*).

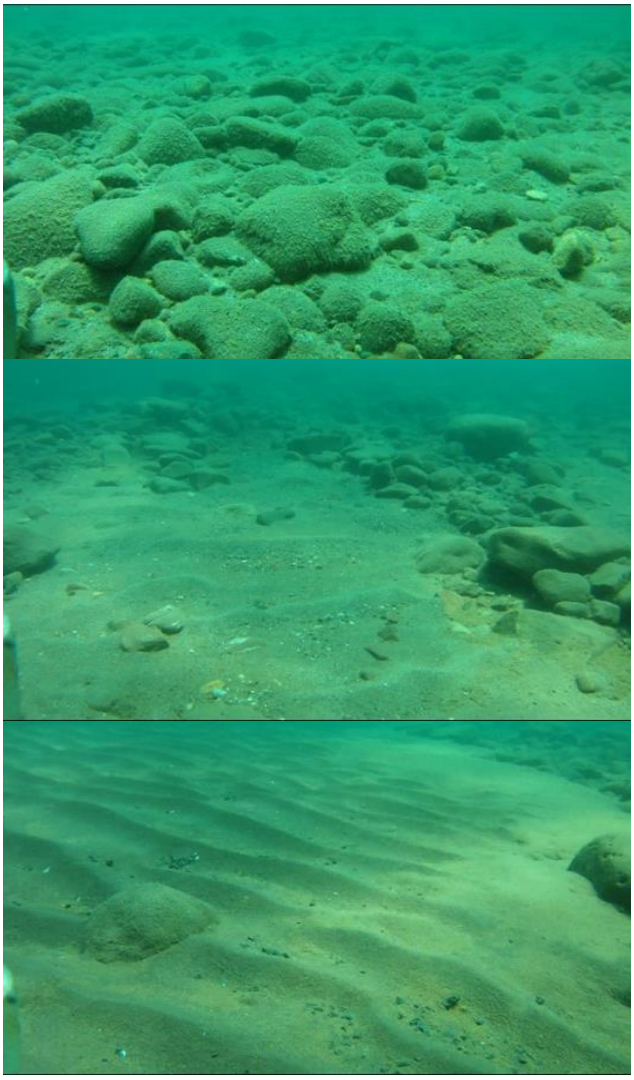


Figure 12: Healthy reef habitat progressively inundated with stamp sands until covered completely (Photos by Charlie Kerfoot)

The sands also impact the aesthetics and sense of place inspired by the landscape; the aerial photos in *Figure 14* illustrate the movement of the stamp sands inundating white sand beaches along the south shore of the Keweenaw Peninsula. The stamp sands also create challenges for people coming and going from the Big Traverse Bay Harbor; the sands have repeatedly blocked the harbor stopping boat traffic. For example, in 2015, Big Traverse Bay Harbor was dredged removing 4,500 cubic yards of stamp sand. Dredging happens repeatedly here, without this continued maintenance there is a great risk of harming the region's fishing industry. As such, local stakeholders are working with the Army Corps of Engineers on a long-term solution.



Figure 14: Aerial views of the entry of Traverse River into Lake Superior. The regular movement of stamp sands over the top of the breakwater requires annual or biennial dredging to keep the small Traverse River Harbor, home of a commercial fishing fleet, open to entry. Note the black sands on the left to the north of the channel compared to the white (natural) sands to the right. Left photo credit: Neil Harri; Right photo credit: Charlie Kerfoot

Impacts on Fish Sovereignty

Buffalo Reef lies within the ceded-territory homelands established by the Treaty of 1842 where 11 Lake Superior Bands of Ojibwa retain rights and responsibilities to harvest fish from the Michigan waters of Lake Superior. Buffalo Reef has always been considered a culturally significant harvesting ground for local communities. Fishing is the strand of the cultural core that ties history to the present day and the future; it is a vital part of the foundation for cultural beliefs and values, traditional lifeways, and even individual identity (Gagnon, 2018; KBIC, 2017).

The KBIC identify as fishing people and have always had a strong focus on cultivating and protecting relationships within ecosystems that support healthy food sovereignty initiatives within the community (*Figure 15*). The Anishinaabeg teachings and ways of life emphasize that landscapes and waters are an intricate system of diverse relationships, and interconnected rights and responsibilities rooted in an acknowledgment and understanding that humans and nature are relatives. These sentient elements have been honored in ceremony since time immemorial and these important teachings



Figure 14: Fresh catch on Keweenaw Bay

and traditions continue today. These relationships are founded in the long-standing nation-to-nation agreements between the Anishinaabe Ojibwa and all orders of creation from rock, water, fire, and wind; the physical world of sun, stars, moon and earth; plant beings; animal beings; and human beings - all rooted in the First Treaty with *Gichi Manidoo* (the Creator), also known as Sacred Law, Original Instructions, and Natural Law (Johnston, 1976; Keweenaw Bay Indian Community, 2021). Today, Tribal, State, Federal, and Academic partnerships are combining efforts to mitigate damages and ultimately restore Buffalo Reef as the ecological resource that has sustained both tribal and non-tribal communities for generations.

Buffalo Reef Remediation Efforts

This is a complex environmental justice issue requiring local, regional, and state stakeholders and rights holders to work together on research and restoration efforts for Buffalo Reef. It was tribal fishermen who first alerted our communities of this issue, underscoring the vital need to bridge knowledges for a holistic understanding of what is happening within our landscape and to recognize the gifts that different knowledges share. Research over the past years has included: a) modeling wave action, currents, and how the sands are migrating across the coastal shelf and along the shoreline; b) use of LiDAR and sonar assessments to understand the extent of the sands underneath water; c) submerging remotely operated underwater vehicles to take pictures of the boulder and cobble fields to assess the quality of the remaining fish habitat; d) researching environmental impacts on the benthic and fish communities; e) exploring considerations for building a revetment to hold back stamp sands from further entering the bay; f) conducting archaeology to understand industrial heritage, mining legacies and Native American uses of the area prior to colonization; and f) and meeting with community members to understand social and economic impacts on the communities involved.

After years of observation, scientific study, and dredging at Big Traverse, the EPA funded a feasibility study for a long-term solution. In 2017 the USEPA endorsed the formation of a Buffalo Reef Task Force (BRTF) comprised of multiple state, federal, and tribal agencies. In addition, several academic institutions and private entities have joined the team, recognizing that this issue is larger than any single entity can accomplish on its own.

The Buffalo Reef Final Alternatives Analysis Report (Report) was recently on public notice from January 30 to March 1, 2024. This report provides an overview of efforts made by the BRTF - *"in 2019, the BRTF identified 13 potential alternatives to remediate the stamp sands and restore the habitat. The alternatives were screened based on constructability, operation and maintenance requirements, environmental impact, ecological sustainability, initial costs and legacy costs, regulatory requirements, public input and opinion, time needed for implementation, impact to local populations, and potential for beneficial use of the stamp sands".*

Three alternatives were selected for further consideration that included the dredging and disposal of stamp sands in the following locations: 1) White Pine Mine, a closed tailings basin; 2) a Lakeside Placement site; and 3) an Upland Placement site. In 2022, a public meeting was held to discuss the one alternative that would be feasible - the removal of the stamp sands to a regulated and lined landfill to be constructed near Gay. These remediation efforts require land ownership,

both at the shore for a proposed jetty to impede migration of the stamp sands and an upland placement area to move the mining waste.

The Report indicates that the BRTF's preferred remedial alternative and "Potential Plan" identifies the "Upland Placement Alternative" as the BRTF's preferred remedial alternative (*Figure 16*). The study compares three scales of implementation for the Upland Alternative are compared in the report. *"The scale of implementation differs based on the volume of stamp sands to be dredged and disposed of during the implementation phase. Additionally, the study provides an assessment of the impact the volume of stamp sands removed during implementation has on the cost and duration of the operation, maintenance, repair, replacement, and rehabilitation phase."*

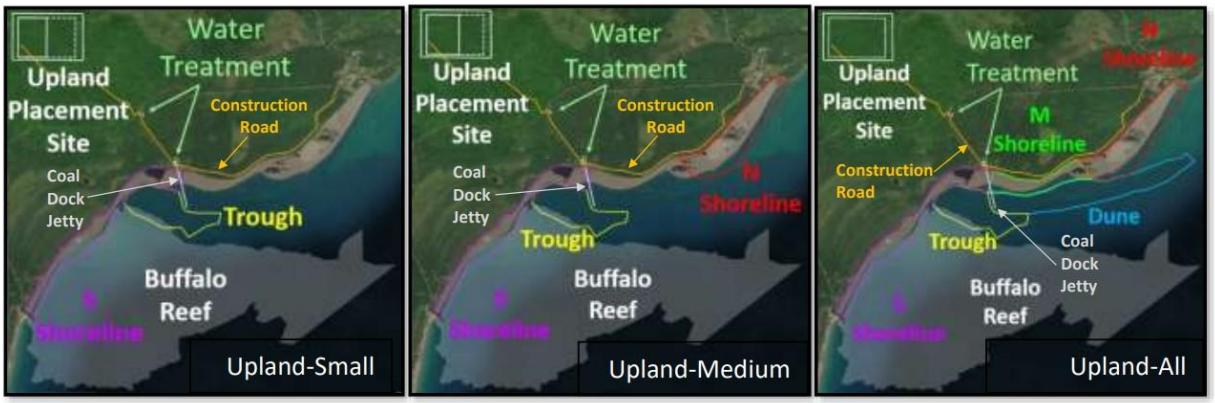


Figure 16: Upland Placement Alternative: Small, Medium, All Dredging Scales represented (Left to Right). S Shoreline – South Shoreline; M Shoreline – Middle Shoreline; and N Shoreline – North Shoreline. The Upland Placement footprint required to dispose of the volume of stamp sands removed during project implementation varies per dredging scale. In the placement site measure, the solid white line represents the estimated placement site footprint to contain the stamp sands dredged during the project implementation phase. The dashed white line represents the estimated placement site footprint to contain stamp sands dredged during the OMRR&R project phase (sourced from page ES-5 of the Buffalo Reef Final Alternatives Analysis Report).

RETURN TO HOUGHTON

Billion-year-old geologic processes brought copper to the surface where people could find it - a gift from the deep Earth. Relationships with the red metal have varied over millennia resulting in different values and actions within our shared lands and waters. Further efforts are required by all stakeholders and rights holders to communicate and elevate the importance of this environmental issue with decision-makers, government officials, and environmental advocacy groups. This field trip is meant to deepen our understanding of these relationships and to reflect on the following questions:

What are the relationships between people, Earth, and Buffalo Reef? Why does this matter?
How have people impacted Buffalo Reef in the past, and how will they in the future?
What can you do to help with the remediation efforts of this place?

<u>Directions:</u> From Big Traverse Bay Harbor, head north on S Big Traverse Rd and continue along Rice Lake Rd. Take M-26 S to US-41 S in Houghton.

EDUCATIONAL RESOURCES

Access the following websites and videos by following the QR codes.

Saving Buffalo Reef Website (DNR)

Saving Buffalo Reef Video (GLIFWC)

Buffalo Reef Restoration (Great Lakes Now, Ep. 1006 Segment 3)

Gay Milling at Gay - A Lake Superior Story -

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Field Trip 4

Keweenaw Fault System Geometry and Kinematics: Clues to Its Nature and Origin

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Objectives

This field trip is designed to demonstrate four fundamental attributes of the Keweenaw fault system along the Keweenaw Peninsula: (1) geometry of the fault network and individual faults; (2) style of deformation of hanging-wall and footwall strata; (3) zonation of deformed rocks in fault zones; and (3) kinematic slip indicators on individual faults.

Stops in this guide were chosen and grouped into four areas to illustrate the segmented geometry of the Keweenaw fault system and to point out fault blocks defined by intersections between faults in three main directions. Kinematic slip indicators observed at some stops manifest a preponderance of oblique slip with components of dextral strike slip and northwest-side-up reverse slip. Based on this information and fault network geometry, the overall nature of the fault system is interpreted to be transpressional and consistent with forcing by the Grenville orogeny.

Introduction

The Keweenaw fault extends southwest from the tip of the Keweenaw Peninsula in Michigan's Upper Peninsula to near Ashland, Wisconsin (Fig. 1), a distance of about 250 kilometers. It is one of two faults along the south shore of western Lake Superior with reverse movement that juxtaposes Mesoproterozoic volcanic layers against Meso-Neoproterozoic clastic sedimentary strata, the other being the Douglas fault in Wisconsin and Minnesota. Irving and Chamberlin (1885) first established the existence of the Keweenaw fault by combining detailed observations in outcrops and trenches with astute reasoning to settle a long-running debate about the nature of the contact between volcanic layers to the northwest and sandstone strata to the southeast. Prior to their work, Wadsworth (1884) had argued that the sandstone was older and dipped northwest beneath the volcanic layers based on their similar dip at some locations along their contact. Afterward, this idea was resurrected intermittently until the mid-1970s, when deep drilling for oil and gas established the younger age of sandstone in the Keweenaw fault's footwall once and for all.

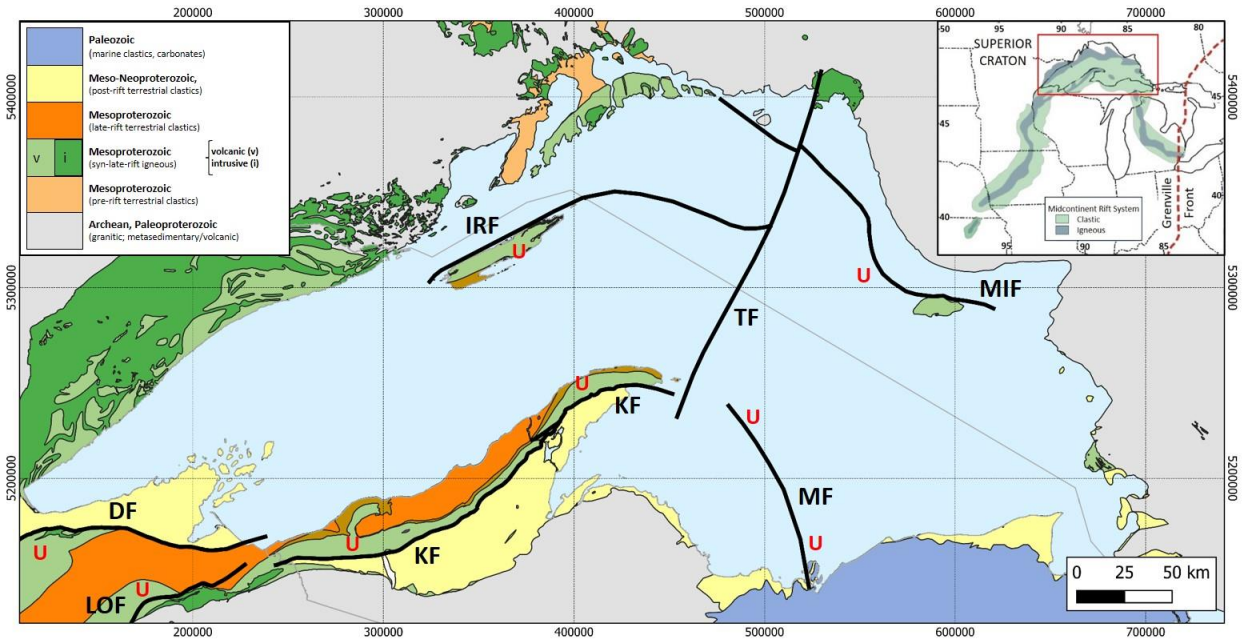


Figure 1: Mesoproterozoic Keweenawan Supergroup of the Midcontinent Rift System (inset map modified from Stein) in the Lake Superior region. For sources of geologic units and faults, see DeGraff and Carter (2023). KF–Keweenaw fault; LOF–Lake Owen fault, DF–Douglas fault; IRF–Isle Royale fault, MIF–Michipicoten Island fault, TF–Thiel fault; MF–Munising fault (provisionally named). Red "U" on upthrown sides of faults.

Tectonic Evolution

The Keweenaw fault and related major faults along the southern edge of the western Lake Superior basin cut rocks of the Midcontinent Rift System (Fig. 1) and have slipped several kilometers in a reverse sense (Cannon and Nicholson, 2000, 2001; DeGraff and Carter, 2023). Ideas to explain the origin and evolution of the Keweenaw fault differ in relation to the roles of rifting and orogenesis on fault initiation and movement over time. The Midcontinent Rift System (MRS), formed by extension of Laurentian crust in Mesoproterozoic time, was accompanied by voluminous mafic volcanism followed by a period of crustal sag with siliciclastic sediment filling the resulting basin (Cannon et al., 1989; Cannon, 1992; Hinze et al., 1990; Stein et al., 2015). Woodruff et al. (2020) defined stages of magmatism and tectonism in MRS evolution as follows: (1) Plateau Stage with widespread volcanism and distributed minor extension (c.1112 to c.1105 Ma); (2) Rift Stage when volcanism and extension reached maximum intensity along a central subsiding rift basin (c.1102 to c.1090 Ma); (3) Late-Rift Stage with declining volcanism transitioning to sedimentary infill of the rift basin (c.1090 to c.1083 Ma); and (4) Post-Rift Stage with sedimentary infill of a sag basin (c.1083 to c. 1060 Ma). Prior to recognition of the MRS, the Keweenaw fault was interpreted as a thrust that formed during an unspecified compressional event following eruption of lava flows and deposition of siliciclastic strata in the Lake Superior basin (Irving and Chamberlin, 1885; Butler and Burbank, 1929, White, 1968). After recognition of the rift, a new model for formation and evolution of the Keweenaw fault and others around the Lake Superior basin proposed that they formed as normal faults during stages 1 and 2, and then were inverted as reverse faults by post-rift Grenville compression (Cannon et al., 1989; Hinze et al., 1990; Cannon, 1994), i.e. the fifth

Compressional Stage of Woodruff et al. (2020). A recent model, discussed below in relation to the inversion model, is that the Keweenaw fault formed by thrusting during post-rift Grenville compression and did not exist as a normal fault during MRS extension or sag (DeGraff and Carter, 2023). In some ways, the latter model reverts to earlier ideas about the Keweenaw fault.

Ideas about the timing of reverse slip on the Keweenaw fault have evolved from the original ideas of Irving and Chamberlin (1885), who inferred two episodes of movement – one prior to deposition of the footwall sandstone and a second afterward. Early thrusting elevated the region northwest of the fault, creating a highland of mafic volcanic layers that shed debris into the tectonic basin southeast of the fault. A second episode of thrusting occurred after most of the sandstone deposition based on deformed footwall strata with steep to overturned attitudes toward the southeast. Later workers have suggested quasi-continuous fault slip that began before Jacobsville deposition and continued during most or all of its deposition (Cannon and Nicholson, 2000), and others have argued for a late reactivation of the fault system during one or more phases of the Appalachian orogeny (Craddock et al., 1997).

Stratigraphy

Extension during MRS evolution followed by compression during the Grenville Orogeny produced a broad syncline with strata dipping toward an axis beneath Lake Superior (Fig. 1). The related rocks in the vicinity of the Keweenaw Peninsula are well described by Cannon and Nicholson (2000, 2001), whose stratigraphic column and nomenclature are adopted here (Fig. 2). Here, the main geologic units of the Keweenawan Supergroup are from oldest to youngest: (1) Siemens Creek Volcanics and Kallander Creek Volcanics (Powder Mill Group) with basaltic to andesitic lava flows; (2) Portage Lake Volcanics (Bergland Group) mostly composed of basaltic to andesitic flows with lesser conglomerate and sandstone layers; (3) Oronto Group strata beginning with the Copper Harbor Conglomerate and locally interbedded basalt-andesite flows, continuing with the Nonesuch Formation composed of siltstone and shale, and ending with the Freda Sandstone composed of lithic sandstone and siltstone; and (4) Jacobsville Sandstone mostly composed of quartzose to subarkosic sandstone and generally correlated with the Bayfield Group in Wisconsin, USA. Archean granite and gneiss of the Superior craton, with a Paleoproterozoic cover of graywacke and slate (Michigamme Formation), lie beneath the Keweenawan rocks south of Keweenaw Bay.

The following descriptions focus on the two geologic units that are juxtaposed along the Keweenaw fault system, namely the Portage Lake Volcanics in the hanging wall and the younger Jacobsville Sandstone in the footwall (Fig. 3). Both units are stratified and have compositional and textural variations between layers that likely influenced their mechanical behavior, which in turn influenced fault initiation and propagation. A more complete description of other Precambrian units along the Keweenaw Peninsula is provided elsewhere in this field trip volume (see Field Trip 1 in this volume).

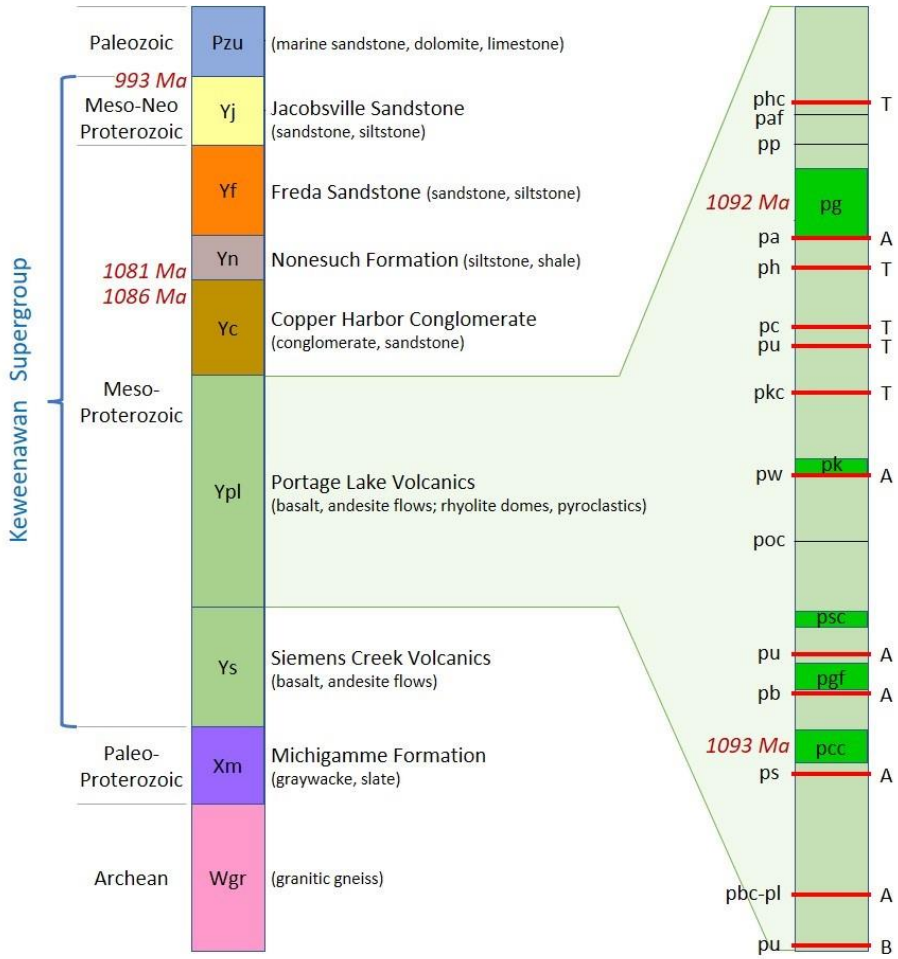


Figure 2: Left. Regional stratigraphy following Cannon and Nicholson (2001). For sources of radiometric ages in red italics, see DeGraff and Carter (2023). Right. Stratigraphy of the Portage Lake Volcanics from USGS bedrock geology maps of the Keweenaw Peninsula. Intense green units are major lava flows: pcc–Copper City; pgf–Gratiot; psc–Scales Creek; pk–Kearsarge; pg–Greenstone. Letter codes on left side are sedimentary layers, except for one: pu–unnamed conglomerate (cgl), pbc-pl–Baltic-Lac La Belle cgl; ps–St. Louis cgl; pb–Bohemia cgl; poc–Old Colony sandstone; pw–Wolverine sandstone; pkc–Kingston cgl; pc–Calumet and Hecla cgl; ph–Houghton cgl; pa–Allouez cgl; pp–Pewabic West cgl; paf–ashbed flow top; phc–Hancock cgl. Bold red bars mark units with layer-parallel slip observed in mines and trenches. Letters on right side indicate where slip occurred (T–top; B–bottom; A–top and bottom).

The Portage Lake Volcanics (PLV) exposed along the Keweenaw Peninsula is truncated on the southeast by the Keweenaw fault system, leaving an undetermined thickness of PLV in the footwall beneath Jacobsville Sandstone. The PLV section in the fault's hanging wall is estimated to be 3000 to 5000 m thick and to contain about 300 flows (Cannon and Nicholson, 2000). It mostly consists of subaerial basaltic flows with less abundant andesitic flows, rhyolitic to dacitic extrusive domes, and associated pyroclastic layers (Butler and Burbank, 1929; White, 1968; Cannon and Nicholson, 2000, 2001). Six named basalt flows are sufficiently traceable along strike due to their

great thickness and distinctive texture to serve as marker units (Fig. 2). The thickest of these, the Greenstone flow, in the upper part of the section is traceable for 80 kilometers in outcrop and drill holes and has a precise radiometric age date of 1091.6 ± 1.3 Ma (Swanson-Hysell et al., 2019; cf. 1094.0 ± 1.5 Ma, Davis and Paces, 1990). The next thickest flow, the Copper City flow, in the lower part of the section is traceable for 25 kilometers and is precisely dated at 1093.4 ± 1.4 Ma (Swanson-Hysell et al., 2019; cf. 1096.2 ± 1.8 Ma of Davis and Paces, 1990).

About 3% of the PLV section consists of conglomerate and sandstone deposited between some of the lava flows (Butler and Burbank, 1929; White, 1952; Merk and Jirsa, 1982). Eleven named interflow units (Fig. 2) are traceable along strike for considerable distances up to 90 km (Bornhorst and Barron, 2011), and thus are useful for correlation. Interflow sedimentary layers mostly consist of clast-supported, pebble-to-cobble conglomerate with a gravelly to sandy matrix, which is well indurated and tends to form prominent strike ridges. Conglomerate layers are coarsely bedded and poorly stratified, except where sandy lenses or conglomeratic sandstone occur. Less common are sandstone and siltstone that may occur as interbeds within a conglomerate layer or locally may make up an entire interflow layer. Interflow sedimentary layers may be up to 40 m thick and locally may pinch to near zero thickness (Butler and Burbank, 1929; Merk and Jirsa, 1982; White, 1968). Clasts and matrix grains in the interflow layers were derived from volcanic rocks of the rift basin, with felsic material generally being far more abundant than expected from the small proportion of felsic volcanic rocks that make up the PLV section (< 1% according to Nicholson, 1992). These so-called felsic conglomerates may occur by themselves or in association with mafic conglomerates, in which case the felsic layer tends to lie atop the mafic layer lying atop the underlying lava flow (Butler and Burbank, 1929).

Stops of this field trip are all within the lower part of the PLV section exposed in the fault's hanging wall. For the first half of the trip, the thick Copper City flow is an important stratigraphic reference that be traced for 25 km along strike. Another important stratigraphic reference with a greater strike extent is the St. Louis conglomerate (#6 on USGS bedrock geology maps), which lies at the base of the Copper City flow in many places but elsewhere is separated from it by a thinner basalt flow. For the second half of the trip, the key stratigraphic reference is the Lac La Belle conglomerate (#3 on USGS maps), which lies 580 m to 1050 m stratigraphically below the St. Louis conglomerate. Other lava flows and interflow sedimentary layers in the lower part of the PLV have not yet been correlated well enough to be useful as regional stratigraphic references. A few intrusions cut the lowermost PLV layers along the Keweenaw Peninsula, such as the Mt. Bohemia syenodiorite stock (Cornwall, 1954a) and scattered dikes largely of intermediate composition (Robertson, 1975).

Jacobsville Sandstone (JS) in the Keweenaw fault's footwall fills a sub-basin that extends southeast beneath Keweenaw Bay to an onlap with Paleoproterozoic metasedimentary rocks, northeast to near Stannard Rock, and southwest to Lake Gogebic (Fig. 1). The unit may reach to 2-3 kilometers in depth based on gravity modeling and early seismic reflection data (Bacon, 1966; Aho, 1969; Kalliokoski, 1982). An exploratory drill hole (Mayflower #41) through the fault near Calumet penetrated JS to a total depth of 803 m (White, 1985), and a deeper drill hole 8.6 km southeast of the fault (Rice Lake #1) reached a total depth in sandstone at 1106 m (Keweenaw NHP, 2016). Regionally correlated units are not well defined for the unit because of lateral facies variability and lack of marker beds. However, Hamblin (1958) recognized the following

succession: (1) basal “conglomerate facies” with locally-sourced clasts deposited adjacent to topographic highs; (2) “lenticular sandstone” facies dominated by planar and trough cross-bedding; (3) “massive sandstone” facies with laterally persistent layers and occasional planar cross-bedding; and (4) “red siltstone” facies with thinly bedded, fine-grained, siliciclastic layers and local trough cross-bedding. Based on this facies succession, Hamblin (1958) interpreted Jacobsville depositional environments as transitioning from (1) alluvial-fluvial at the base, (2) to dominantly fluvial in the lower cross-bedded part of the unit, (3) to dominantly lacustrine in the upper massive part of the unit, and (4) to variably lacustrine and fluvial in the uppermost thinly bedded part of the unit. He inferred that the general sequence of facies and depositional environments was time transgressive, representing both a vertical time sequence at any given point as well as a basinward change of depositional environment at any given time.

The JS section along the Keweenaw Peninsula consists of siliciclastic strata that generally conform to the facies of Hamblin (1958) but with some variations. Proximal to the fault system, the lower parts of exposed sections commonly have brownish conglomerate layers consisting of subangular volcanic clasts chaotically dispersed in a poorly indurated muddy matrix, and interbedded with soft, fissile, reddish-brown siltstone and shale (Irving and Chamberlin, 1885; Hamblin, 1958; DeGraff, 1976; Brojanigo, 1984). In such sections with muddy conglomerate and shaly strata, interbeds of whitish to orangish to pinkish, fine- to medium-grained sandstone occur as isolated thin beds near the base and become more abundant and thicker upward in the section until muddy strata are rare. The quartzose to subarkosic sandstones are locally conglomeratic and often cross-bedded, which is typical of most JS strata exposed away from the fault system to the southeast. Sand-prone JS strata also occur near the fault system where the muddy conglomerate and shaly facies is absent. Brojanigo (1984) interpreted the muddy conglomerate layers as debris flows derived from an upland of PLV rocks to the northwest, and the more mature sandy strata as fluvial deposits derived from older quartzo-feldspathic rocks to the south. We interpret this bimodal alluvial-fluvial assemblage to be basal Jacobsville, whose PLV-derived detritus is evidence of prior reverse slip on the Keweenaw fault system that elevated the northwest hanging wall and allowed volcanics there to be weathered, eroded, and transported southeastward into the JS sub-basin.

The time span of JS deposition has been difficult to determine due to a lack of fossils and igneous rocks interbedded with or crosscutting Jacobsville strata (Kalliokoski, 1982). Deposition probably did not occur prior to reverse slip on the Keweenaw fault system (KFS) because conglomerate low in the JS section near the KFS contains PLV detritus sourced from an elevated region to the northwest (Brojanigo, 1984; Cannon and Nicholson, 2000). The earliest slip on the fault system is generally taken to be ~1060 Ma based on Rb-Sr ages of gangue minerals in fractures associated with native copper mineralization of the Keweenaw Peninsula (Bornhorst et al., 1988) and reset biotite ages in Archean granite gneiss uplifted along related faults west of Lake Gogebic (Cannon et al., 1993). Later reverse slip on the Keweenaw fault that deformed adjacent Jacobsville strata is relatively well constrained at 985 ± 30 Ma by a U-Pb date for late-kinematic vein calcite in the fault zone, in combination with a maximum depositional age of ~993 Ma for deformed strata high in the Jacobsville section (Hodgin et al., 2022). Most Jacobsville deposition probably occurred in the period 1060-985 Ma, although somewhat younger layers could have been deposited in the region (Craddock et al., 2013; Malone et al., 2016).

Keweenaw Fault System – Key Observations and Concepts

The Keweenaw fault has been described as a strike fault because of the general parallelism between its map trace and PLV layers in its hanging wall (Fig. 3). Butler and Burbank (1929) also noted a remarkable similarity in dip of hanging-wall layers and the fault, such that PLV layers change dip along strike by about the same amount and in the same sense as changes in the underlying fault surface. Where differences in dip exist, the fault cuts 5° to 17° more steeply and upward across PLV layers in a southeasterly direction, i.e. in the thrusting direction (DeGraff and Carter, 2023). Related to these surface observations, Hubbard (1898) had systematically catalogued in mine workings many layer-parallel faults along contacts between volcanic and interflow sedimentary layers (Fig. 2). These bedding-plane faults are manifested by fault gouge, polished surfaces, breccia with alteration, and secondary mineralization in fractured zones up to 8 m thick (65% ≤ 1.5 m). Hubbard's observations indicate that PLV layer boundaries were relatively weak and became detached during one or more deformation episodes. Furthermore, his observations along the Keweenaw Peninsula mining district northeast of Portage Lake document layer-parallel slip throughout the PLV section.

The surface and subsurface evidence of layer-parallel slip within the PLV section is consistent with knowledge about PLV stratigraphy. The lava flows have variations in texture and mineralogy across their thickness that likely influence their mechanical properties. Massive holocrystalline flow interiors grade into margins with finer grain size and abundant vesicles and amygdules, often arranged in bands parallel to the contacts. Flow tops have the highest abundance of amygdules, are commonly brecciated, and often have been modified by weathering between eruptions or by mineralizing fluids that later migrated along these once-permeable zones (Butler and Burbank, 1929; Stoiber and Davidson, 1959; White, 1968; Bornhorst, 1997). Where one flow lies directly atop another, a significant mechanical contrast exists across their comparatively weak contact. Contrasts in layer characteristics and properties are even more significant at contacts between interflow sedimentary layers and lava flows. Altogether, the evidence just summarized implies a detached style of formation for the Keweenaw fault and related splays (DeGraff and Carter, 2023).

Although the Keweenaw fault has been described as a high-angle reverse fault, its northwesterly dip at the surface varies from as low as 20° up to 70° (Butler and Burbank, 1929). Cross-section construction along a transect with abundant constraining data from mining operations and surface mapping implies that the fault is nearly horizontal northwest of its surface trace at Hungarian Falls (DeGraff and Carter, 2023, their Fig. 6), one of the stops on this trip (see Stop 2-1, Fig. 8). Several cross-sections on USGS bedrock geology map sheets from the 1950s show the fault dipping ≤ 25° or show nearly horizontal PLV layers near surface that imply a similarly-dipping fault below (e.g., Laurium, Ahmeek, Mohawk, and Bruneau Creek quadrangles). Thus, the fault and overlying PLV strata dip between nearly horizontal up to 25° NW at several locations and, there, the hanging-wall PLV block has overridden footwall JS strata by as much as 2.5 kilometers.

Between 2017 and 2022, three mapping projects funded by the USGS EDMAP program have refined fault and stratal geometries along the Keweenaw fault, and the results indicate that what was largely considered to be a single fault is better characterized as a fault system (Tyrrell, 2019; Mueller, 2021; Lizzadro-McPherson, 2023; Gamet, 2023, Langfield, 2024). The Keweenaw fault

system (KFS), as used here, refers to an ensemble of fault segments whose collective motion has a significant component of dextral strike slip in addition to the long-recognized component of reverse slip with northwest side up. The fault system consists of segments that define three main directional sets: (1) a dominant set that defines the trend of the fault system and locally separates more steeply dipping PLV layers to the northwest from less steeply dipping PLV layers to the southeast; (2) splay faults that are angled 15-30° clockwise from set 1; and (3) connector faults angled 35-75° counterclockwise from set 1 that join footwall splays to the main fault trend (Fig. 3). The three directional fault sets maintain these angular relationships among each other as the curved KFS changes direction from a 35° azimuth near Houghton to a 95° azimuth near the tip of the peninsula. The interconnected nature of the three fault sets defines fault blocks with long dimensions parallel to the local trend of the KFS.

Fault-slip indicators measured on ~400 fault surfaces show that the collective ratio of reverse slip to dextral slip on the KFS varies from 1:1 near Houghton to 1:2 or more near the tip of the peninsula. In other words, strike slip is twice as large as dip slip on the KFS near the tip of the peninsula where fault azimuth is 60° clockwise from its azimuth near Houghton (Tyrrell, 2019; Mueller, 2021; Lizzadro-McPherson, 2023; Gamet, 2023; Langfield, 2024). This change in the ratio of strike slip to dip slip is expected as the fault's azimuth approaches the estimated 105° azimuth of maximum compressive stress that is attributed to the Grenville orogeny. The left-stepping arrangement of footwall splay faults and the fault blocks they help to define is consistent with a transpressional fault system as indicated by the fault-slip data. The relatively short, north- to northeast-trending, connector faults (set 3) are associated with similar-trending fold axes at relatively high angles to the estimated maximum compression direction and shortening direction. The connector faults are inferred to have mostly reverse slip with west side up and to accommodate northeast to east transport of footwall fault blocks associated with dextral transpressional slip of the entire KFS.

Why concern ourselves with the geometric details and timing of slip on the KFS? From a science perspective, the arrangement of fault segments in the system along with fault-slip indicators like slickenlines provide clues to the mechanics of fault initiation and kinematics of fault slip. Improved understanding of these aspects of the KFS should apply to similar faults around Lake Superior and help to understand their causative tectonic events and stress regimes. From a practical perspective, faulting along the Keweenaw Peninsula and on strike to the southwest provided pathways for upward migration of mineralizing fluids (Bornhorst, 1997), leading to copper deposits that once supported a thriving mining industry and are still prospective today. Native copper is commonly found along major and minor faults in the mining district. The Hancock fault cutting the Quincy and Hancock mines is an example of a major fault with copper mineralization concentrated in its hanging wall. Another example is the Allouez Gap fault that bisects the Kearsarge flow-top copper deposit, the largest of this type in the district (Bornhorst, 1997). Faults likely provided pathways for upward-moving ore fluids into vesicular flow tops at the Baltic and Isle Royale deposits and others in the Greenland-Mass subdistrict (Broderick, 1931). The earliest native copper deposits exploited in the district east of Eagle River were along subvertical fracture zones and minor transverse faults with displacements less than 200 m (Butler and Burbank, 1929). Therefore, a better understanding of fault geometry and timing may provide insights about the distribution of known deposits and about the possible presence of undiscovered deposits.

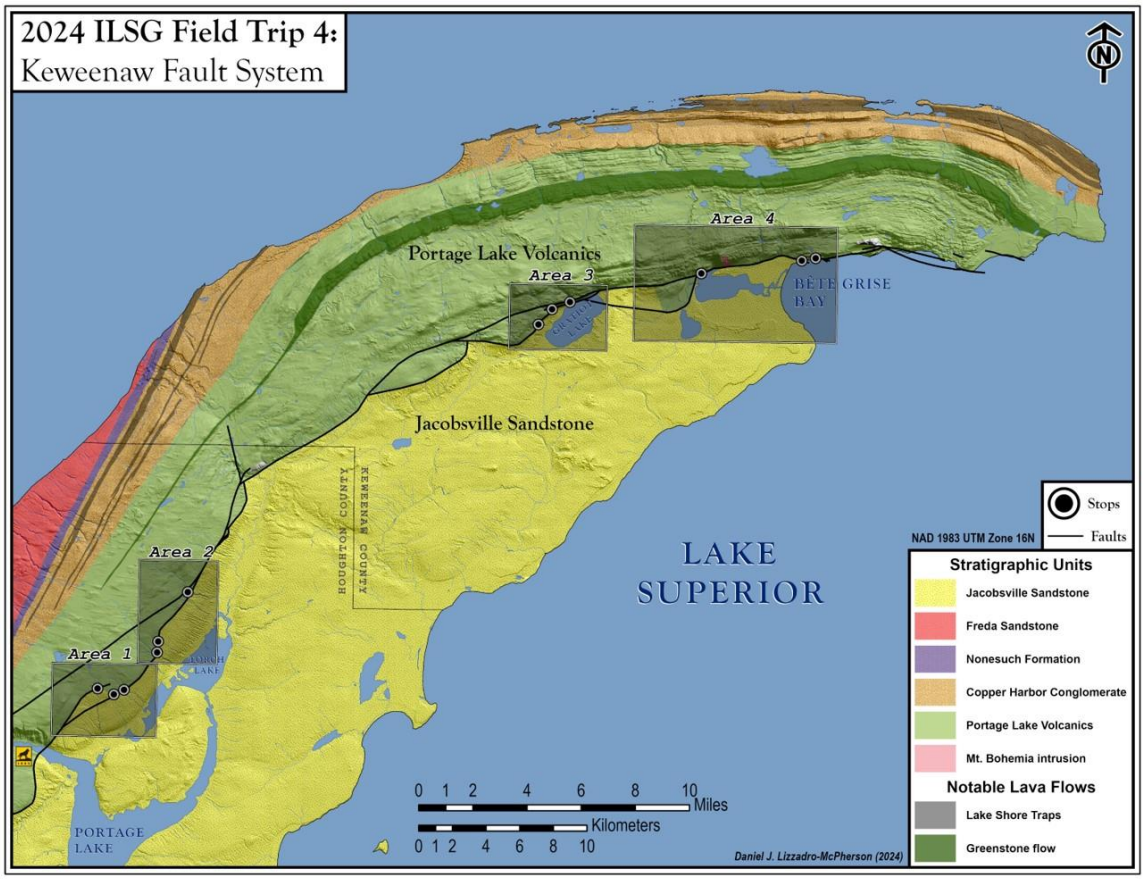


Figure 3: Geologic map of the Keweenaw Peninsula showing the Keweenaw fault system north of Portage Lake and field trip stops within the four focus areas. Faults are indicated by curved black lines.

Field Trip Stops

The field trip stops described below are grouped into four areas to illustrate themes related to fault system geometry, stratal relationships and deformation, fault blocks, deformation fabric in fault zones, and slip kinematics. Each area has an introductory overview of the key themes and nature of the stops, followed by individual stop descriptions.

Reported strike and dip values follow a right-hand-rule convention. The dip value is followed by letters indicating the cardinal direction of dip, which is redundant but makes for clarity.

About half of the stops in this guide are on public land. Those on private land are specified in the stop descriptions as requiring permission from the owners to access.

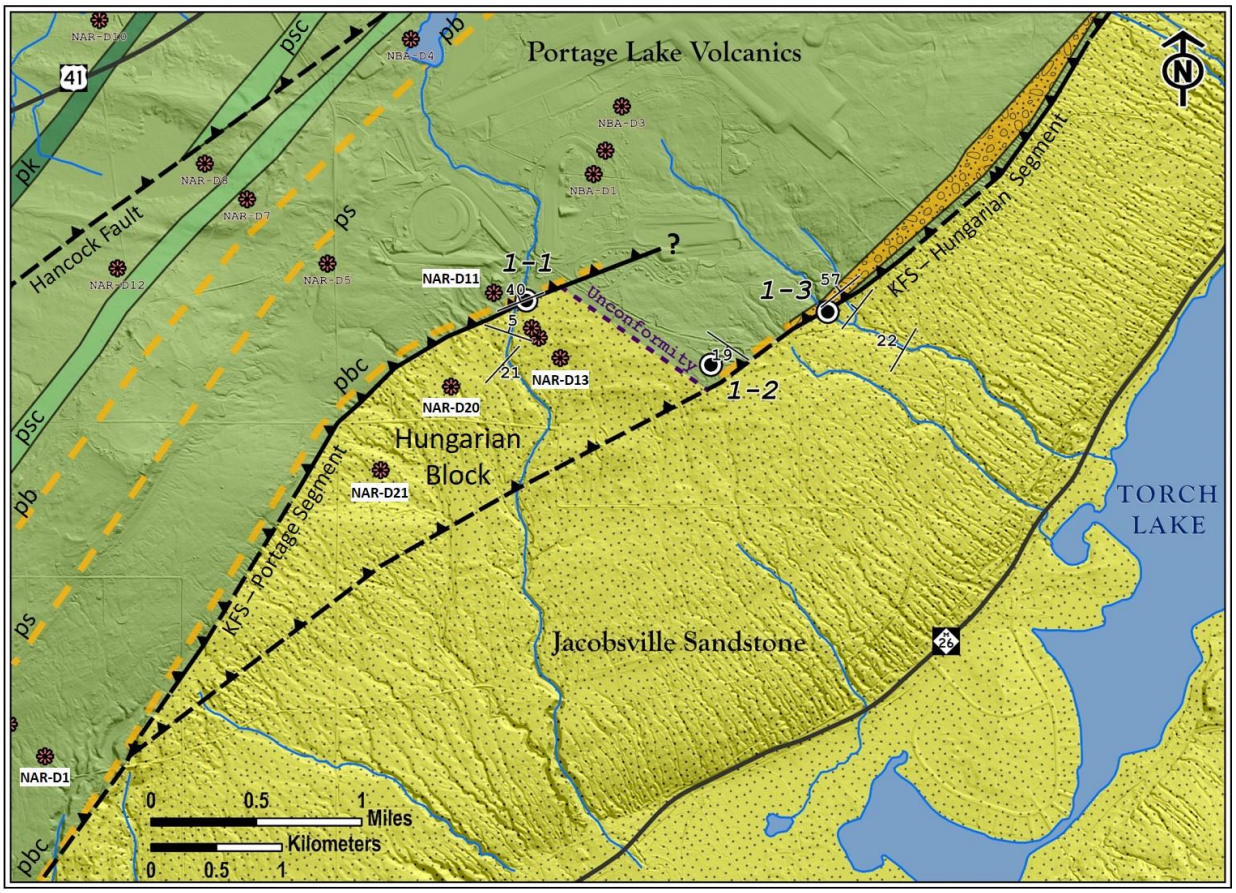
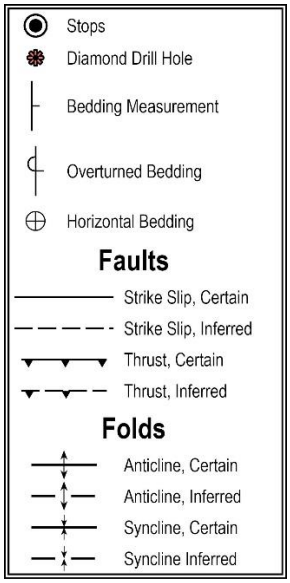


Figure 4: Geologic map of Area 1 with the Keweenaw fault system (KFS) and field trip stops 1-1 to 1-3. Geologic unit codes: pbc – Baltic conglomerate, ps – St. Louis conglomerate, pb – Bohemia conglomerate, psc – Scales Creek flow, pk – Kearsarge flow with Wolverine sandstone at its base. Teeth along reverse faults are on upthrust sides.



Area 1: Hungarian Fault Block – Southwest End

STOPS IN THIS AREA ARE ON PRIVATE LAND AND REQUIRE PERMISSION TO ACCESS.

The three stops in Area 1 are within the vehicle testing grounds of Michigan Tech's Keweenaw Research Center (KRC), under lease from the Houghton County Memorial Airport (Fig. 4). From west to east, the stops are on the main Keweenaw fault zone (1-1), in a basalt quarry within a fault-bounded block (1-2), and on a footwall splay fault (1-3) that passes through Hungarian Falls to the northeast in Area 2. Within the fault-bounded block, an unconformity between the PLV and JS is interpreted to dip shallowly southwest. These three stops illustrate an en echelon, overlapping, fault geometry that define a fault block whose long dimension is parallel to the trend of the Keweenaw fault system (KFS). Recent mapping shows that such fault patterns and fault-bounded blocks are repeated along the KFS north of Portage Lake (Fig. 3). A similar configuration of faults with the enclosed block having a west-dipping PLV-JS unconformity occurs in Area 4 (Figs. 3 and 13).

Stop 1-1: Gooseneck Creek fault exposure

Directions: From the Portage Lake lift bridge between Houghton and Hancock, follow US-41 north for 6.7 mi (10.8 km) to Airpark Boulevard on the right. Turn right toward Houghton County Memorial Airport and drive 0.5 mi (0.8 km) to the Keweenaw Research Center (KRC) on the right. Enter the KRC parking area and wait in vehicles. We may be escorted by KRC staff for 1.5 mi (2.4 km) to the south edge of their vehicle testing grounds, where we will park and walk. [Lat: 47° 9.259'N | Lon: 88° 29.895'W]

The fault exposure on Gooseneck Creek is part of the main Keweenaw fault zone that is traced by means of outcrops, drill holes, and water wells along a line bearing 36° from the Michigan Tech campus to the intersection of Airport Park and Forsman roads, about one kilometer southwest of this stop (Fig. 4). Northeast from that intersection, the fault's map trace curves eastward so that here it trends 72°, as do the hanging-wall PLV strata. This stop is the most northeasterly site where this segment of the KFS can be observed because the flat upland to the northeast, where the airport is located, has little to no bedrock exposure.

Hubbard (1898) seems to be the first geologist to describe this site. While his report puts the outcrop about 80 m south of its actual location, his geologic description is very similar to what was observed and measured during the 2021-2022 EdMap project (Langfield, 2024).

"A conglomerate here, underlain by trap, strikes N. 72° E., and dips northerly 44°, the trap being in contact on the south with the sandstone, which is much broken and disturbed but appears to dip rather flat to the N. E." (Hubbard, 1898)

The fault's main slip surface is not exposed but its position is determined to within a couple of meters by the proximity of hanging-wall PLV outcrops to footwall JS outcrops. PLV strata in the hanging wall, striking 252° and dipping 40-44° NW, change stratigraphically upward from amygdaloidal basalt at the fault at creek level to a felsic pebble conglomerate, and then to a series of basaltic flows for as far as outcrop exists to the north and northwest. The stratigraphic

position of the conglomerate layer suggests that it is the Baltic (#3) conglomerate shown on the USGS Hancock and Chassell bedrock geology maps (Cornwall, 1956a; White, 1956). Abundant fractures within the basalt flows are not obviously systematic, although detailed work might reveal dominant sets. Jacobsville strata in the footwall commonly appear massive due to being thickly bedded to locally cross-bedded, which makes their attitude difficult to measure. In general, JS strata dip gently both toward and away from the fault, suggesting an open anticlinal structure with an axis that trends approximately east-west (Fig. 4).

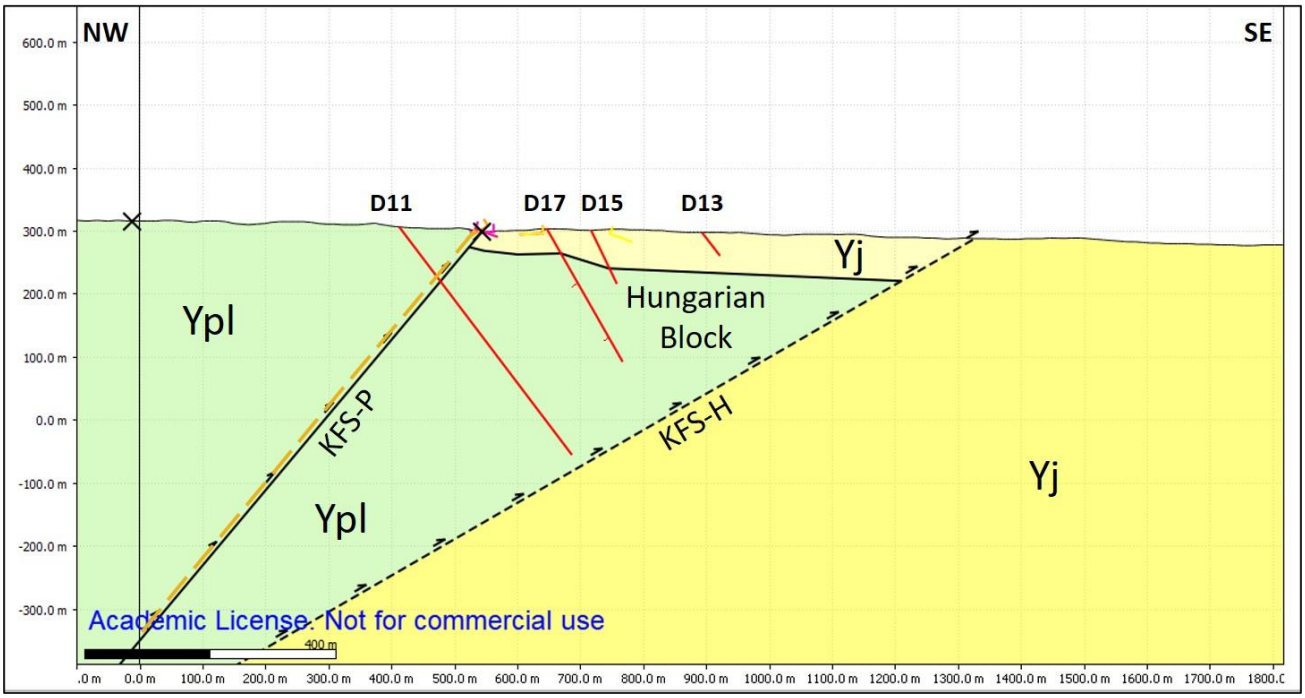


Figure 5: Gooseneck Creek cross-section in progress using outcrop and drill hole data (Langfield, 2024). Red lines below topography are New Arcadian drill hole trajectories. Colors and letter codes of geologic units are explained in Figure 2. KFS-P = Keweenaw fault system - Portage segment; KFS-H = Keweenaw fault system - Hungarian segment. Dashed orange line above KFS-P marks what may be the Baltic (#3) conglomerate.

Diamond drill holes (DDH) of the Calumet and Hecla Consolidated Copper Company provide data that are critical for the interpretation of this site, situated on a NW-trending line of four New Arcadian holes drilled in 1911-1912 (Fig. 4). Southeast of here, three drill holes penetrated the following beneath glacial overburden (hole depth converted to vertical depth): DDH #17 about 140 m SE cut 46 m of JS sandstone followed by 109 m of PLV basalt; DDH #15 about 205 m SE cut 69 m of JS sandstone and conglomerate followed by 14 m of PLV basalt; DDH #13 about 355 m SE cut 37 m of JS sandstone (Fig 5). Southeast of the fault, therefore, drilling reveals a veneer of JS strata less than 70 m thick overlying PLV basaltic lava flows. About 135 m northwest of the fault, DDH #11 inclines 52° toward this site and reaches a total depth of 457 m. It cuts a single felsic conglomerate layer between depths of 93 m and 109 m (16 m apparent thickness) that we correlate to the 6-m-thick conglomerate layer seen here in outcrop, which yields an average stratal dip of 53° NW. Below the conglomerate, DDH #11 penetrated ~2

meters of sludge that we interpret as fault gouge, and then continued through basaltic flows to its bottom at a vertical depth of 363 m below the DDH #17 surface location to the southeast. The position of the fault below the conglomerate in DDH #11 and not far below the conglomerate at the surface indicates that the fault dips nearly parallel to hanging-wall PLV strata. While at the surface the fault juxtaposes PLV strata against JS strata in the footwall, at depth in DDH #11 the fault's hanging wall and footwall both have PLV strata, which is consistent with the thin veneer of JS strata penetrated by DDH #17 and DDH #15 southeast of here.

Stop 1-2: Basalt quarry of the Keweenaw Research Center

Directions: Return to vehicles and drive north 0.15 mi (0.25 km) to KRC perimeter road. Turn right and drive southeast for 0.75 mi (1.2 km) to a basalt quarry on the right. Pull off road into a gravel terrace on the right and park. [Lat: 47° 9.144'N | Lon: 88° 29.023'W]

The KRC basalt quarry was opened in 2018-2019 to provide materials for expansion of their vehicle testing facilities. The quarry lies between the fault segment just visited at Gooseneck Creek and a parallel segment to be visited at the next stop (Fig. 4). The quarry exposes portions of at least two basalt lava flows that dip shallowly southwest. The shallow dip is roughly manifested by subhorizontal benches of the quarry and can be measured on the first dry bench west of the quarry pond, where the upper surface of a lava flow has been exposed. Inspection of the irregular subhorizontal surface reveals isolated patches of sediment whose stratification yields an average strike of 125° and dip of 19° SW.

The shallow southwesterly dip of PLV strata is important to the understanding of structural geometry. Early geologists noted that older PLV strata near the Keweenaw fault and its splays commonly have anomalous orientations relative to younger PLV strata away from the fault zone, which have a well-defined regional trend (Hubbard, 1898; Butler and Burbank, 1929). In most of the PLV section, strata generally strike northeast to east and dip 35° – 55° northwest to north. At this stop and for ~3 kilometers north and northeast, PLV strata dip less than 25° in various directions, including counter-regional to the southeast. Such anomalies are clues to the structural configuration of the area along the Keweenaw fault system.

Specific to this stop, the shallow southwesterly dip of PLV stratal is consistent with data from DDHs #17 and #15, where a thin veneer of JS strata overlies PLV basaltic rocks. The nature of the PLV-JS contact is not described in the core logs, but the normal stratigraphic sequence indicates that it is an unconformity. The unconformity dips shallowly southwest to south based on the two cited DDHs and DDH #20 located ~510 m southwest, which penetrates the unconformity ~35 m lower relative to a common datum. Projecting the unconformity updip brings it to the surface southwest of this stop. Therefore, we interpret that basalt at the quarry roughly correlates with basalt below JS strata in the three DDHs by passing beneath an erosional wedge of JS strata on a SW-dipping unconformity (Langfield, 2024). This new interpretation differs from one involving a transverse fault shown on USGS bedrock geology maps for the Hancock and Laurium quadrangles (Cornwall and Wright, 1956a, 1956b).

Stop 1-3: East branch of Quincy Creek fault exposure

Directions: Return to vehicles and drive east on the perimeter road for 0.25 mi (0.4 km) to an intersection with a gravel road on the left. Turn left and drive ~0.35 mi (~0.6 km) to a clearing on the right. Turn into the clearing and park. [Lat: 47° 9.282'N | Lon: 88° 28.548'W]

The east and west branches of Quincy Creek cross a segment of the Keweenaw fault system that is about 750 meters southeast of the fault segment at Gooseneck Creek and subparallel to it. Both branches of Quincy Creek constrain the position of the fault to within 8-10 m by the proximity of hanging-wall PLV outcrops to footwall JS outcrops, but the fault zone itself is not exposed. Outcrops along the west branch of Quincy Creek are the most southwesterly constraint on the position of this fault segment because of glacial deposits that cover bedrock southwest of here. However, we interpret the fault to continue southwesterly to an acute intersection with the fault segment seen on Gooseneck Creek at a point about halfway to Portage Lake (Fig. 4). Northeast from this stop, the fault segment is easily traced by means of outcrops and water wells along an azimuth of 42° for a distance of 3 km to the upper Hungarian Falls in Area 2. We will focus on the east branch of Quincy Creek because it provides better exposures near the fault.

Again, Hubbard (1898) appears to have been the first geologist to describe outcrops at this site as well as other outcrops in stream valleys to the northeast that cross the fault line. Although his report puts the outcrop 150-175 meters north-northeast of its actual location, his geologic description of part of the outcrop is very similar to what was observed and measured during the 2021-2022 EdMap project (Langfield, 2024).

"In Sec. 22 . . . occur outcrops of sandstone and of a conglomerate with a very sandy matrix. The pebbles in the conglomerate are subangular and some of them are of quartz porphyry. The dip is about 50°-54° N. W., strike about N. 45°-50° E." (Hubbard, 1898)

The outcrop just described begins 60 meters downstream from the gravel access road and extends another 30 m downstream. The sedimentary layers here are well indurated and well stratified, having an average strike of 225° and dip of 35° NW. A reddish-brown sublithic sandstone is the dominant rock type along the semi-continuously exposed section in the creek bed, with subordinate conglomeratic sandstone and pebble-to-granule conglomerate that ranges from matrix-supported to clast-supported. Clasts are mostly subrounded to subangular and have a variety of compositions but are dominantly felsic. Whereas Hubbard (1898) thought that these were JS strata, we interpret them as PLV strata because of their induration, lithic nature, and attitude that are similar to other PLV sedimentary units in the area and differ from JS sandstone strata to be seen downstream.

The south end of the 30-m extent of indurated sandstone and lesser conglomerate coincides with a sharp deflection in the creek by ~20 meters east before the creek resumes its southerly course at a sharp right bend. Downstream from this point for ~30 meters, intermittently exposed sandstone and minor conglomerate differ in many aspects from the sedimentary strata upstream of the creek's deflection. The sandstone is lighter toned and locally streaked off-white to beige, is less

indurated and commonly friable, and has a more quartzose composition. The pebble-to-granule conglomerate layer in the first outcrop downstream from the creek's deflection has a muddy to silty matrix, is poorly indurated, and has subrounded to rounded clasts with a variety of compositions. These characteristics of the sandstone and one conglomerate layer are typical of Jacobsville strata observed elsewhere in fault juxtaposition with hanging-wall PLV strata.

The sharp deflection in the creek is interpreted, therefore, as marking a fault contact between PLV sedimentary strata upstream and JS strata downstream. This interpretation is supported by structural observations. Strata in the first outcrop south of the creek's deflection strike 220° and dip 65° NW (overturned), but become less steeply dipping over a short distance as the creek is followed downstream to the south. These stratal attitudes and their significant change going downstream from the creek's deflection contrast with stratal orientation and its consistency over a similar distance upstream of the deflection. About 15 meters south of the creek's deflection, a small fault striking 225° and dipping 80° N cuts thickly bedded JS strata in a larger outcrop along the west bank of the creek. South of the fault, deformation bands in the sandstone are expressed as quasi-linear to broadly sinuous ridges on the outcrop surface. Deformation bands are cataclastic shear zones that develop during compression of partly indurated, porous, clastic material (Fossen et al., 2007). They are generally more cemented than the host material, accounting for their raised relief on erosional surfaces, and have not been observed in PLV sedimentary strata. Farther downstream, JS strata are abundantly displayed for a distance of 1.4 km, nearly down to the derelict Quincy Mining Company stamp mill along state highway M-26. Over that distance, JS strata generally dip less than 15° NW and display a few broad open folds with NNE-trending axes.

Area 2: Hungarian Fault Block – Northeast End

The first two stops in Area 2 are on the same fault segment seen at Quincy Creek (2-1) and on its curved portion (2-2) that connects back to the main Keweenaw fault zone to the north (Fig. 6). A PLV sedimentary layer, usually conglomerate but locally sandstone, is traceable in the hanging wall from Quincy Creek through the east branch of Dover Creek near Hungarian Falls, where it begins a smooth northward curve from 42° to 345° azimuth, a change of over 55°. From this curved stratal geometry in map view, we infer a single smoothly curved fault along the southeast and east edges of the Hungarian fault block. The long straight part of the fault is the block's southeast edge that parallels the overall KFS, whereas the short curved part is the block's east edge and connects the fault segment back to the main Keweenaw fault zone. This geometry and a northwesterly decrease in stratal dip based on drill hole data define a single scoop-shaped fault rather than an intersection of two distinct faults. Stop 2-2 illustrates a common dynamic of the KFS – northeast and east edges of fault-bounded blocks were thrust eastward along west-dipping reverse faults. The third stop in Area 2 is on the main Keweenaw fault zone (2-3), which aligns with its counterpart in Area 1 (Fig. 3). The relationship between the main fault zone and the connector fault may involve the connector fault terminating against a main fault zone that continues to the southwest (Fig. 6 at "?"). Another option is that the hanging-wall sedimentary layer is continuous from Stop 2-2 to Stop 2-3 and is draped over a lateral ramp in the KFS that steps up in stratigraphy to the northeast. The stratigraphic relationships across the Hancock fault are discussed in the Stop 2-3 description.

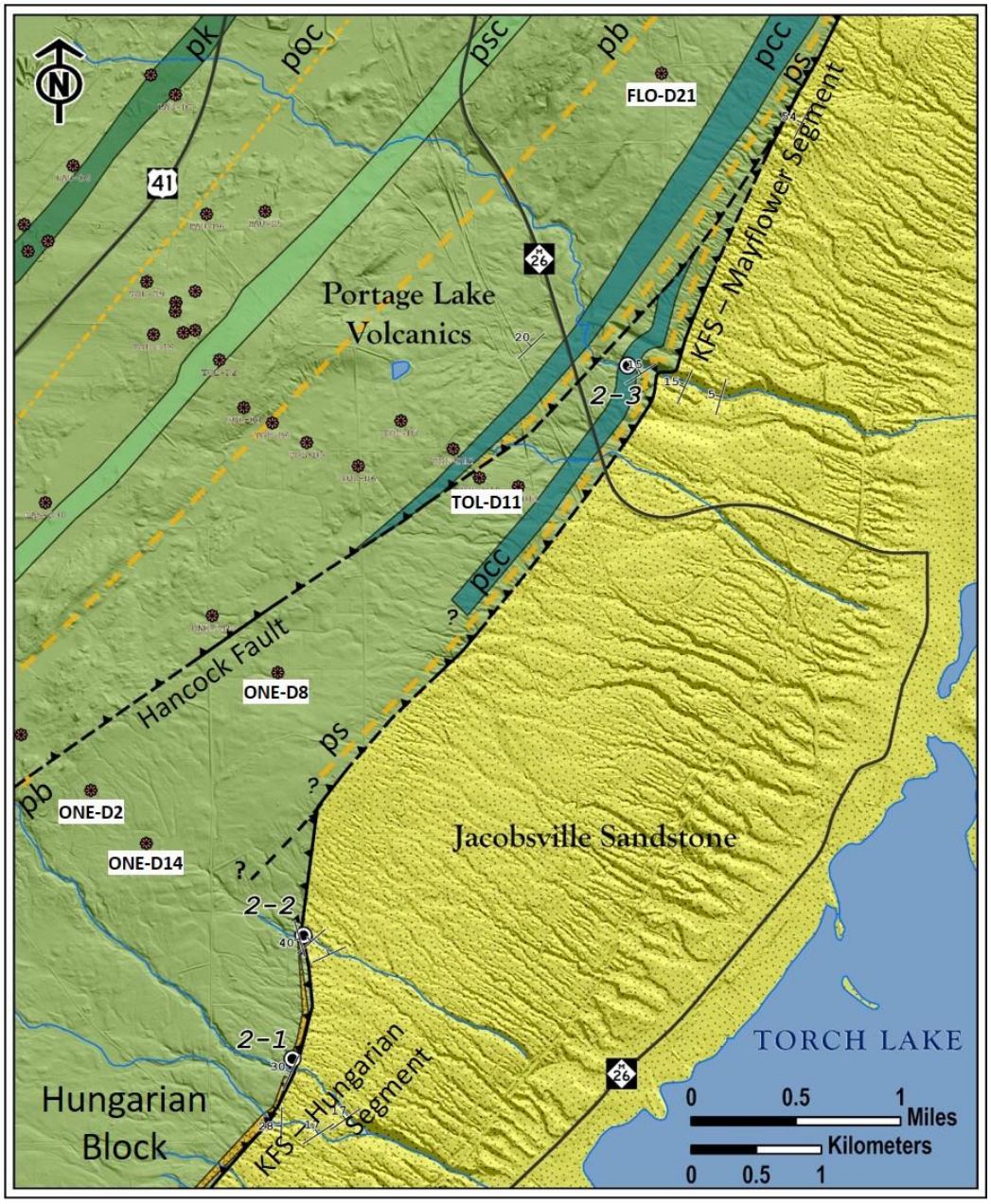


Figure 6: Geologic map of Area 2 with the Keweenaw fault system (KFS) and field trip stops 2-1 to2-3. Geologic unit codes: ps – St. Louis conglomerate, pcc – Copper City flow; pb – Bohemia conglomerate, psc – Scales Creek flow, poc – Old Colony sandstone; pk – Kearsarge flow with Wolverine sandstone at its base. Teeth along reverse faults are on upthrust sides. See Figures 4 or 13 for symbology.

Stop 2-1: Dover Creek fault exposure at Upper Hungarian Falls

Directions: Return back to the KRC perimeter road, then turn right and follow it back to the KRC entrance on Airpark Boulevard. Turn left and drive 0.5 mi (0.8 km) to US-41. Turn right and drive 1.6 mi (2.5 km) to Oneco Road. Turn right onto Oneco and drive 3.1 mi (5.0 km) to Amygdaloid Street. Turn left and follow Amygdaloid and Tamarack Hill Road for 0.3 mi (0.5 km) to M-26. Turn left and drive 0.35 mi (0.56 km) to 6th Street in Hubbell. Turn left and then, after the second house on the left, veer left onto Golf Course Road going uphill. Drive 0.5 mi (0.8 km) to a rutted gravel road on the left. Depending on conditions, we may turn left and take the gravel road. Otherwise, park along Golf Course Road and walk to the stop at Upper Hungarian Falls. [Lat: 47° 10.440'N | Lon: 88° 27.086'W]

The fault segment crossed by Dover Creek is near the northeast end of the segment seen at Quincy Creek, and is where the fault's surface trace begins to curve northward (Fig. 6). Dover Creek has three sets of falls that are worthwhile visiting. The two downstream falls have larger drops entirely over Jacobsville Sandstone. We will visit the upper set of falls at the fault contact between PLV strata upstream in the hanging wall and JS strata in the footwall. This site has been visited by geologists since the mid-1800s and is considered a classic exposure of the Keweenaw fault (original sense). The most insightful description by early geologists is in the work of Irving and Chamberlin (1885), whose diligent field work, keen observations, and logical reasoning convinced the geological community of the time that the contact between PLV strata and JS strata was a large fault. Prior to their work, some geologists argued that JS strata lay stratigraphically below PLV strata based in part on their similar dip here and at Houghton-Douglass Falls on Hammell Creek (Stop 2-3). Of historical interest, Roland Duer Irving is this year's nominee for recognition by the ILSG as a Pioneer of Lake Superior Geology, and Thomas Chrowder Chamberlin is another well-known geologist of his time, famous for his classic 1890 work "The method of multiple working hypotheses" published in Science. Both were contemporaries of John Wesley Powell, a U.S. Army officer during the American Civil War, famed explorer of the American west, and second director of the U.S. Geological Survey from 1881–1894.

The faulted relationship between PLV and JS strata at this stop was revealed by excavations made by a "force of miners" hired to trench across the PLV-JS contact at three locations on the southwest valley wall (Fig. 7). The trenches exposed a fault zone dipping 30-35° NW along the contact, which has the following internal zonation from hanging wall to footwall, as summarized from Irving and Chamberlin (1885) and converted to true thickness.

1. Trap [basalt]: highly fractured but in place. (2.57 m)
2. Trap debris: disintegrated basaltic fragments in a lumpy crudely laminated clay, having a transitional boundary with zone 1. (0.39 m)
3. Clay: red and "shaly" with light grayish-green spots, some sandy seams, and occasional lumps of disintegrated trap. (0.13 m)
4. Trap debris: similar to zone 2 but with more clayey material, whose dark color contrasts with adjacent red clay. (0.13 m)

5. Trap debris mixed with red clay: trap debris is similar to zone 4 but red clay and minor sand gives zone 5 a reddish tint. (0.17 m)
6. Sandstone: light reddish and quartzose. (0.51 m)

Footwall JS strata near the fault contact dip 10° or less, but at the fault contact they are bent downward to be nearly parallel with the fault surface. The trench observations indicate that the fault propagated upward across subhorizontal JS strata and the overriding PLV strata crushed and abraded the truncated edges of JS strata.

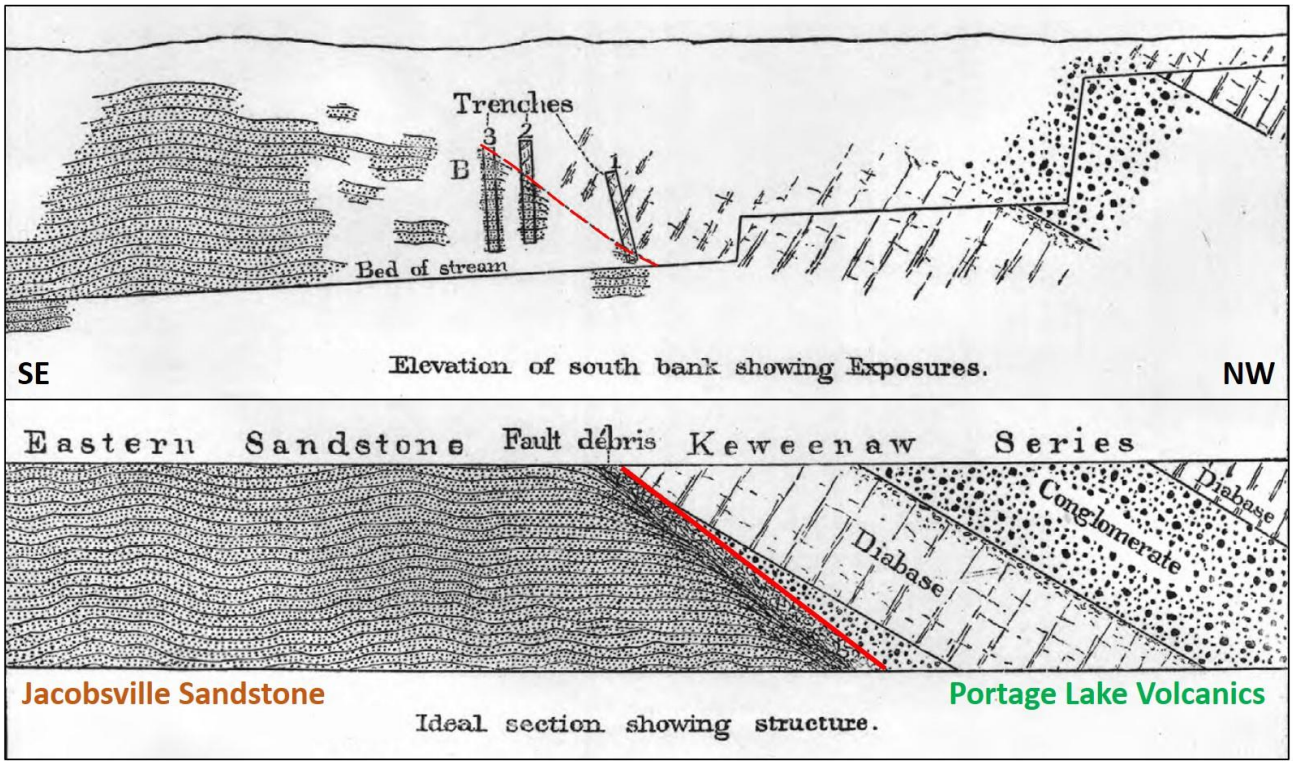


Figure 7: Trenches at upper Hungarian Falls that exposed relationships across the fault, marked by the red lines (Irving and Chamberlin, 1885).

Downstream from the trenches, JS strata generally dip 10-20° NW toward the fault but exhibit a few broad open folds similar to what is observed downstream along Quincy Creek. Upstream from the trenches, PLV strata in the hanging wall begin with a fractured basalt flow at the minor falls below the main falls, and then progress stratigraphically upward to a felsic cobble-pebble conglomerate at the main falls, followed by a series of basaltic flows intermittently exposed for over two kilometers upstream. Stratal dip decreases upstream from 25-30° NW at the main falls to flat-lying and then to 15° SE to define an open syncline, which is succeeded upstream by an open anticline before reaching the Hancock fault (Fig. 8). A well-laminated, 1-m-thick sandstone layer at the base of the 6-m-thick conglomerate layer provides a confident formation strike of 205° and dip of 30° NW, which is essentially parallel to the underlying fault surface. This geometry indicates that the hanging-wall PLV section at this location became detached along

layering somewhere downdip to the northwest and was thrust southeast and upward to its current position along a ramp that cuts upward across younger layers.

The fault segment exposed at this stop is penetrated by the vertical Oneco #9 DDH located 1.7 km west-northwest, at a depth of 556 m, which gives an average fault dip over this distance of 19° NW after accounting for topography (Fig. 8; DeGraff and Carter, 2023). Because the fault dips 30-35° NW at the surface, it must dip less than 19° NW over some portion of its trajectory between here and the Oneco #9 DDH. That is, its dip must shallow going in a northwesterly direction similar to the shallowing of PLV stratal dips observed upstream along Dover Creek, which is characteristic of a detached style of faulting with layer-parallel slip.

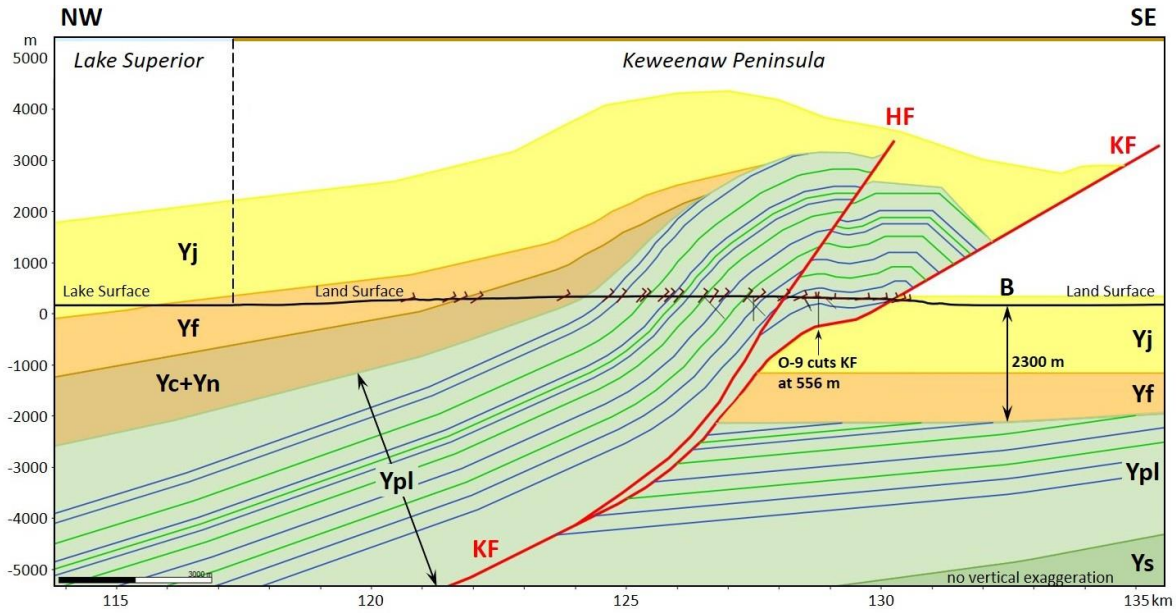


Figure 8: Cross-section along Dover Creek based on outcrop and drill hole data (DeGraff and Carter, 2023). Long bar inclination of L-shaped symbols at land surface shows apparent dip. Thin black lines below topography are drill hole trajectories. Colors and letter codes of geologic units are explained in Figure 2. KF–Keweenaw fault; HF–Hancock fault; B–Bacon (1966) seismic experiment.

Stop 2-2: Beaudoin Creek fault exposure

<u>Directions</u>: Return to Golf Course Road and turn left to go uphill. Drive 0.5 mi (0.8 km) on Golf Course Road to Beaudoin Creek. We will park along the west shoulder of the road north of the creek culvert. This will require turning vehicles around at the first driveway north of the creek. [Lat: 47° 10.806'N | Lon: 88° 27.049'W]

THIS STOP IS ON PRIVATE PROPERTY. PERMISSION IS REQUIRED TO OBTAIN ACCESS.

The small creek crossed by Golf Course Road is unofficially called Beaudoin Creek after the property owner, who also owns the Wood'n Spoon specialty food and gift shop in Mohawk, MI. We will walk upstream to the west for 300 meters to outcrops first described by Hubbard (1898), who traced the PLV-JS fault contact throughout this area and noted remarkable changes in its strike. Between Quincy Creek and the west branch of Dover creek (not visited), the Hungarian fault segment has an azimuth of 42°, but it begins to curve northward at the east branch of Dover Creek, where it has an azimuth of 25° (Fig. 6). The gradual change in fault direction continues from the last stop to this one such that here the fault trace and hanging-wall PLV strata have an azimuth of 345°, which completes a total change in fault strike of 55-60° along a smooth arc. The NNW-trending portion of the fault segment extends more than a kilometer north to a point along another smooth curve of the fault toward the northeast, which re-aligns the fault with the trend of the Portage fault segment. It is unlikely that the position of the second broad curve in the fault north of here is a coincidence, and more likely that it manifests in some way a continuation of the KFS-Portage segment seen at Stop 1-1.

Along Beaudoin creek east of the fault, JS strata are intermittently exposed over a distance of 170 meters and mostly consist of yellowish quartzose sandstone with occasional layers of reddish siltstone and conglomerate. Over most of this distance, JS strata dip less than 15° W toward the fault, but within 20 meters of the fault they dip more steeply at 70° E to vertical. The fault contact between PLV basaltic rock to the west and JS sandstone to the east is located to within a meter by the exposures, but the fault zone is not well exposed due to the degraded basaltic rock in the hanging wall. PLV stratigraphy here is very similar to that observed at Hungarian Falls, beginning with a basaltic lava flow of low relief that extends upstream to a small pond below a 6-m-tall waterfall. The waterfall is over a NNW-trending ridge of felsic cobble-pebble conglomerate ~10 m thick that has a meter-thick basal layer of siltstone to fine-grained sandstone, as seen at the previous stop. West of the conglomerate ridge, a series of basaltic flows crops out along the creek bed for another 200 meters. A reliable measurement of PLV strata orientation from the basal unit of the conglomerate layer gives a strike of 170° and dip of 45° W, which probably is also the attitude of the fault surface by analogy with Hungarian Falls and based on the interpretation of a detached fault system.

The smooth curve of the fault segment and subparallel PLV strata that are traceable from southwest of Stop 2-1 to north of Stop 2-2, combined with the decrease in dip of the fault and PLV strata toward the Oneco #9 DDH, imply a curved fault surface in three dimensions. A number of geometries involving smaller faults and folds are possible, but the overall geometry implies a larger scoop-shaped fault surface that plunges between southwest and west. Further work to integrate surface mapping with subsurface DDH data are needed to fully define the fault and stratal geometries in this area. For now, we interpret the curved fault segment as part of the Hungarian segment of the KFS, which defines the long southeast edge and shorter east edge of the Hungarian fault block (Fig. 6) with relatively shallow dipping PLV strata.

Stop 2-3: Hammell Creek fault exposure at Houghton-Douglass Falls

<u>Directions</u>: Drive southeast on Golf Course Road 1.0 mi (1.6 km) to M-26. Turn left and drive 2.6 mi (4.2 km) to 10th Street in Lake Linden, where M-26 makes a 90° left turn. Turn left and drive 2.0 mi (3.2 km) to a parking area on the right. Turn in and park. [Lat: 47° 12.413'N | Lon: 88° 25.611'W]

Houghton-Douglass Falls (a.k.a. Douglass Houghton Falls) was purchased by the State of Michigan in 2018 and, as of this writing, is being converted into a day park with walking trails and signage. The work is not yet complete and access is still somewhat limited but allowed. We will view the site from overlooks at the top of the steep canyon walls and will not descend to the fault contact near the base of the falls due to time constraints. The total vertical drop of 34 m (110 feet) makes Houghton-Douglass Falls the tallest in Michigan.

This stop is another classic exposure of the Keweenaw fault (original sense) that has been investigated by geologists since the mid-1800s. The trail to the overlook area crosses the Hancock fault, not yet found in outcrop, and PLV strata in the Keweenaw fault's hanging wall. The Hancock fault has been traced 16 km on an azimuth of 55° by means of drill holes from the Hancock and Quincy mines to an intersection with the Keweenaw fault one kilometer northeast of the falls (Fig. 3). The Hancock fault seems to have had an important role in the occurrence and distribution of native copper at those two mines (Bornhorst et al., 1986; Field Trip 2 in this volume). Copper mineralization at the base of Houghton-Douglass Falls next to the Keweenaw fault was investigated in an adit opened by the Douglass Houghton Mining Company, organized in 1845 (Stephens, 1902), and recently observed in veins (Gamet, 2023).

The geometry of two PLV layers near the intersection of the Hancock and Keweenaw faults is critical to understanding geologic relationships at Houghton-Douglass Falls (Fig. 6). The Laurium bedrock geology map shows the St. Louis (#6) conglomerate and overlying Copper City flow in the hanging walls of the Hancock and Keweenaw faults north and west of their intersection but not to the southwest in the acute fault wedge containing Houghton-Douglass Falls (Cornwall and Wright, 1956b). The two units are easily recognized and correlated in drill holes and outcrops from many kilometers northeast of Calumet-Laurium in a southwesterly direction to the Hancock fault. The St. Louis conglomerate is a felsic, pebble-cobble, clast-supported conglomerate with locally significant lithic and conglomeratic sandstone that is associated with rhyolitic rocks along strike (Hubbard, 1898; White et al., 1953; Nicholson, 1992; Gamet, 2023). The Copper City flow is recognized for its anomalous thickness of 180 m at a point 7 kilometers to the northeast, coarse grain size, and pegmatitic segregations that have been precisely dated (Fig. 2). Recent mapping aided by drill hole data has allowed these two units to be correlated across the Hancock fault into the acute fault wedge, where they are identified in outcrop at Houghton-Douglass Falls (Fig. 6).

The main drop at Houghton-Douglass Falls is over the Copper City flow, which extends from upstream of the waterfall down to a ledge ~10 meters above its base. From the north side of the gorge, a planar fabric with meter-scale spacing over much of the flow thickness dips gently

upstream. Observations at the top of the waterfall suggest that this fabric results from amygdule layers that dip 20-25° NW. Below the main waterfall on the ledge above the base, a 3-m-thick sedimentary layer, first noted by Hubbard (1898), consists of a felsic pebble-granule conglomerate and coarse lithic sandstone whose layering provides a reliable strike of 225° and dip of 20° NW, i.e. parallel to amygdule layers at the top of the waterfall. This is interpreted to be the St. Louis (#6) conglomerate below the anomalously thick Copper City flow. Both layers project northeast along strike to intersect the Hancock fault at a point where their map offsets across the fault match the offsets of PLV layers previously correlated across the fault (Fig. 6). Below the St. Louis conglomerate is a basalt lava flow that becomes increasingly fractured at the foot of the falls, at which point sheared basaltic rock is observed along the south face of the gorge.

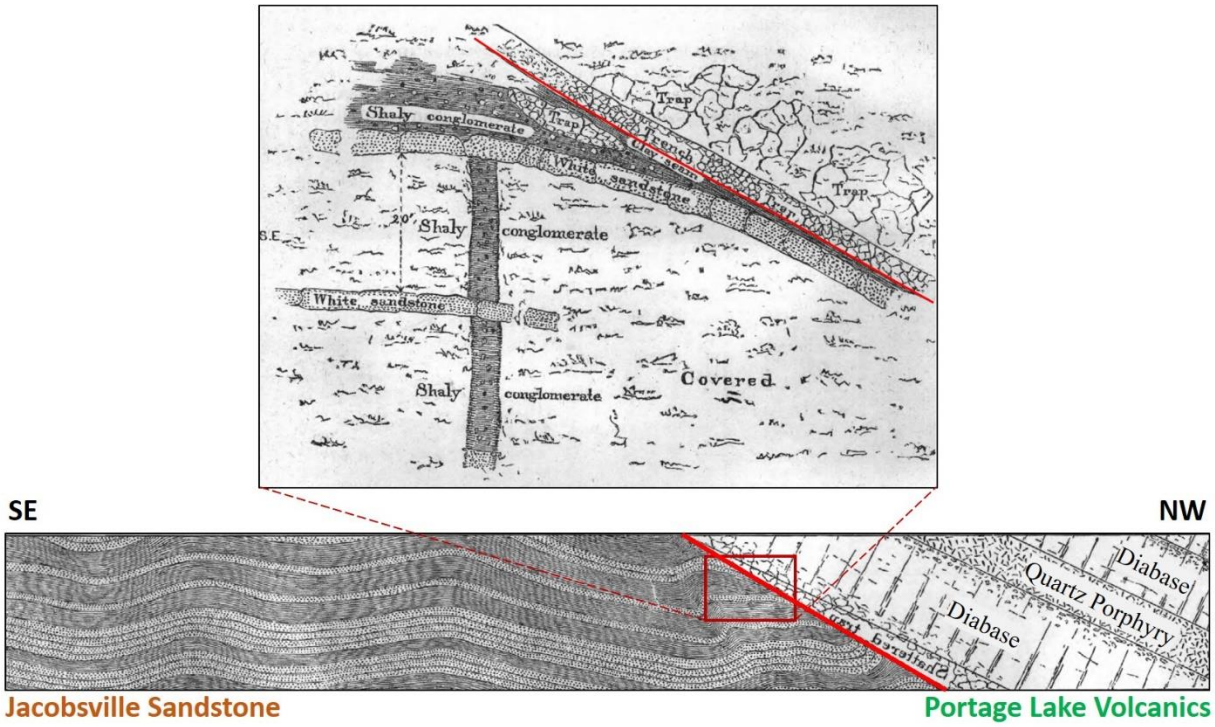


Figure 9: Trenches below Houghton-Douglass Falls that exposed relationships across the fault, marked by the bold red line (Irving and Chamberlin, 1885).

Irving and Chamberlin (1885) focused their investigation of Houghton-Douglass Falls along the bottom and walls of the gorge below the falls, again combining careful field observations with trenching across the fault surface. Two trenches running up and down the south valley wall and a third trench along the fault surface (Fig. 9) revealed similar relationships to those observed in the trenches at Hungarian Falls but with some important differences. Basaltic rock in the hanging wall has a clay seam at the fault surface that transitions upward to a clayey breccia and then to fractured but intact basalt as seen in the Hungarian trenches. However, footwall JS strata here are generally poorly indurated, red, shaly conglomerate with subordinate whitish quartzose sandstone layers, which is the opposite relationship of dominant to subordinate lithologies at Hungarian Falls. The downward flexing of subhorizontal JS strata to dips of 20-30° close to the fault surface is again

observed, but the flexing is spread over a larger horizontal distance than at Hungarian Falls. The wider zone of flexure here may result from the poor induration and ductility of the red shaly conglomerate here relative to the moderately indurated sandstone strata at Hungarian Falls.

Irving and Chamberlin (1885) estimated a fault dip of 25-30° NW based on their excavations, which is similar to their estimate of PLV stratal dip of 25° NW in the hanging wall. Recent field work yielded a fault dip of 20° NW based on a three-point method using surveyed points across the gorge and by visually siting upward along the fault surface (Gamet, 2023). This result for fault dip matches the PLV stratal dip measured on the St. Louis conglomerate and on amygdule layers in the Copper City flow. While earlier and recent dip values are slightly different, both studies agree that hanging-wall PLV strata are essentially parallel to the underlying fault surface, which is indicative of a detached thrust system. A recent cross-section through Houghton-Douglass Falls uses concepts related to detached thrusting, conservation of volume, and ductile portions of the JS section to model fault geometry and deformation of strata in the hanging wall and footwall of the fault system (Fig. 10).

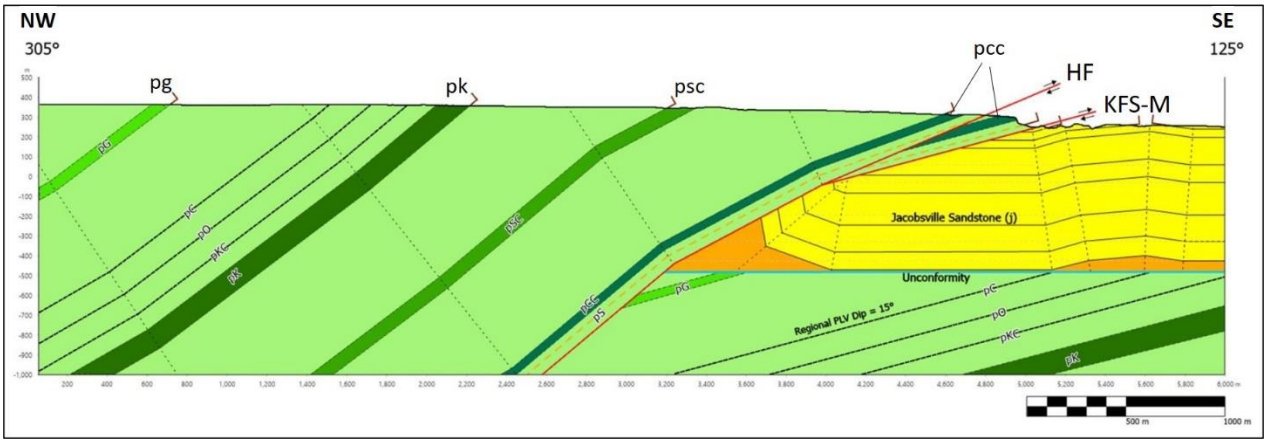


Figure 10: Cross-section along Hammell Creek at Houghton-Douglass Falls based on outcrop data and drill hole correlations (Gamet, 2023). Geologic unit codes: pcc – Copper City flow; psc – Scales Creek flow, pk – Kearsarge flow; pg – Greenstone flow. KFS-M = Keweenaw fault system - Mayflower segment; HF = Hancock fault.

Area 3: Snake Creek Fault Block – Keweenaw Fault Zone at Lake Gratiot

The three stops in Area 3 (Fig. 11) are on the east boundary fault of the Snake Creek block (3-1), on the main Keweenaw fault zone north of Lake Gratiot (3-2), and at the intersection of these two fault trends (3-3). The Keweenaw fault zone roughly parallels the north edge of Lake Gratiot and, relative to Areas 1 and 2, it trends more easterly with an azimuth of 72° and dips more steeply northwest. Other fault segments in Area 3 are similarly oriented clockwise relative to their counterparts in Areas 1 and 2. South of the main Keweenaw fault zone, two fault-bounded blocks have long dimensions oriented parallel to the KFS (Fig.3). West of Lake Gratiot, the Snake Creek block is bounded on the southeast by a NE-trending connector fault, along which PLV layers are

thrust southeast over JS strata (Fig. 11). Deformed hanging-wall PLV strata and footwall JS strata have NE-trending fold axes parallel to their fault contact. Fold axes in the hanging wall plunge 23° with an azimuth of 44° (Mueller, 2021). The southern boundary fault of the Snake Creek block is an ESE-trending footwall splay of the main Keweenaw fault zone (Fig. 3) that juxtaposes PLV strata on the north against JS strata to the south. Similar to the northeast end of the Hungarian block (Fig. 6), this footwall fault splay appears to curve north and merge seamlessly into the connector fault along the southeast edge of the Snake Creek block.

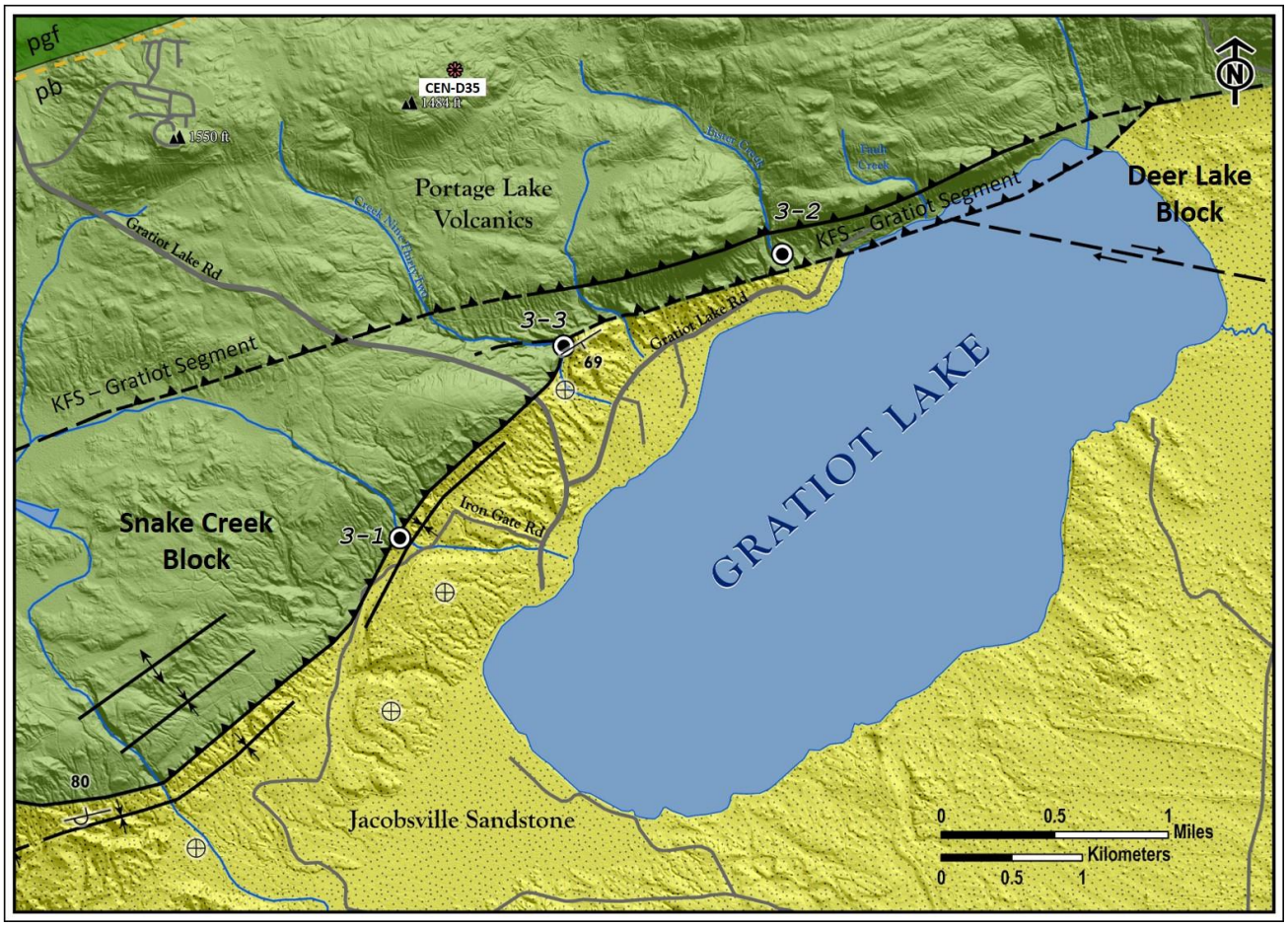


Figure 11: Geologic map of Area 3 with the Keweenaw fault system (KFS) and field trip stops 3-1 to 3-3. Geologic unit codes: pb – Bohemia conglomerate, pgf – Gratiot flow. Teeth along reverse faults are on upthrust sides. See Figures 4 or 13 for symbology.

The main Keweenaw fault zone in this area is well exposed along several creeks that empty into Lake Gratiot and, based on outcrop relationships and cross-section models, it consists of two parallel branches (Fig. 11). The northern branch has a well-developed gouge and breccia zone that is at least 20 m wide in places and perhaps as much as 45 m wide. This branch juxtaposes PLV basaltic flows on the north against presumably younger basaltic flows on the south, whereas the southern branch juxtaposes basaltic flows on its north against JS strata on the south. The double fault zone north of Lake Gratiot continues in a west-southwest direction past the termination of the

connector fault intersecting from the southwest and forms the northern boundary of the Snake Creek block. This main fault zone trajectory differs from the 1950s USGS bedrock geology maps and was first proposed by Cannon and Nicholson (2001) in their map compilation. We have adjusted the position of their proposed fault based on a combination of outcrop relationships, drill hole data, and topographic features, and we consider evidence for its existence to be compelling.

Stop 3-1: Unnamed creek fault exposure along Iron Gate Road

Directions: Continue driving north on M-26 for 2.0 mi (3.2 km) to Hecla Street. Turn right and drive through historic downtown Laurium for four blocks (0.5 mi, 0.8 km) to 1st Street, a.k.a. School Street. Turn left and drive 0.35 mi (0.56 km) to US-41. Turn right and drive 17.6 mi (28.3 km) past the road to Eagle Harbor to historic Central Location. Turn right onto Gratiot Lake Road and drive ~4.8 mi (~7.7 km) to unpaved Iron Gate Road on the right. Turn right and drive 0.5 mi (0.8 km) to an unimproved dirt road on the right. Turn into the side road and park. [Lat: 47° 21.222'N | Lon: 88° 9.340'W]

This stop will be a quick show-and-tell to explain fault and stratal geometries at the east edge of the Snake Creek fault block that extends 5 kilometers to the west (Fig. 11). A short walk northwest along the two-track road leads to a 3-m-tall outcrop of JS in the creek northeast of the road. This outcrop is much larger than other JS outcrops sometimes exposed for 35 meters upstream in the creek bed, depending on spring run-off. About 80 meters upstream is the first PLV basalt outcrop where the creek emerges from the upland to the northwest. The boundary between the upland with PLV bedrock and the lower flatter area to the southeast with JS bedrock has a trend of 42° and it extends ~3 kilometers from the southeastern rounded corner of the Snake Creek fault block to the main Keweenaw fault zone at Nine Thirty Two Creek (Stop 3-3). Several small creeks flowing southeast from the upland toward Lake Gratiot cross this boundary and expose bedrock, constraining the position of the geologic contact but not exposing it.

The sandstone strata here strike 36° and dip 63° SE in the inferred direction of stratigraphic up, indicating that they have been rotated down to the southeast (Lizzadro-McPherson, 2023). Similar orientations of JS strata are noted at the mouths of other creek valleys where they emerge from the upland. Scattered JS outcrops to the southeast have nearly flat-lying strata, indicating that tilting of JS strata is negligible beyond 100 to 150 meters from the PLV-JS contact. In the upland northwest of the contact, fractured PLV strata with veins of secondary minerals generally do not present good opportunities to determine strike and dip. Where possible to measure, consistent northeasterly strikes with dips both to the northwest and southeast define an anticline-syncline pair with NE-trending fold axes that parallel the PLV-JS contact.

The relationships along the east edge of the Snake Creek block are evidence of a NE-trending fault along which PLV strata to the northwest were thrust over JS strata to the southeast. Similar to the curved fault at the northeast end of the Hungarian block, the fault at the east end of the Snake Creek block also has a curved geometry where it wraps around the block's southeast corner and gradually changes direction by 55° to a west-northwesterly trend. Along the fault's

NE-trending section, its position and nature in the latest map (Fig. 11) do not differ much from what is shown on the Bruneau Creek bedrock geology map (Wright and Cornwall, 1954). However, the earlier map labels this fault the "Keweenaw fault," whereas we interpret it to be part of a curved splay in the footwall of the main Keweenaw fault that lies to the north, which we will visit at the next stop.

Stop 3-2: Main Keweenaw fault zone at Eister Creek and Falls

<u>Directions:</u> Return 0.5 mi (0.8 km) on Iron Gate Road back to Gratiot Lake Road. Turn left and drive 0.2 mi (0.3 km) to East Gratiot Lake Road. Turn right and drive 1.0 mi (1.6 km) to a parking area along the road. [Lat: 47° 21.975'N | Lon: 88° 7.967'W]

Eister Creek near the falls provides excellent exposures across the northern branch of the main Keweenaw fault zone where PLV strata in the hanging wall are juxtaposed against presumably younger PLV strata in the footwall (Figs. 11 and 12). Jacobsville strata that commonly occur in the footwall of the main fault zone are not present here, though they may constitute bedrock south of the southern branch of the main fault zone. Elsewhere nearby in the footwall, the JS unit has thicknesses ranging up to 130 m confirmed in outcrop and greater than 135 m in a water well northwest of Lake Gratiot. In general, the thickness of the JS unit in this area appears to be much less than to the southwest near Houghton, where the unit is known to be at least 785 m thick at the fault system and at least 1,100 m thick away from it to the southeast, but could be 2,000 to 3,000 m thick based on geophysical data.

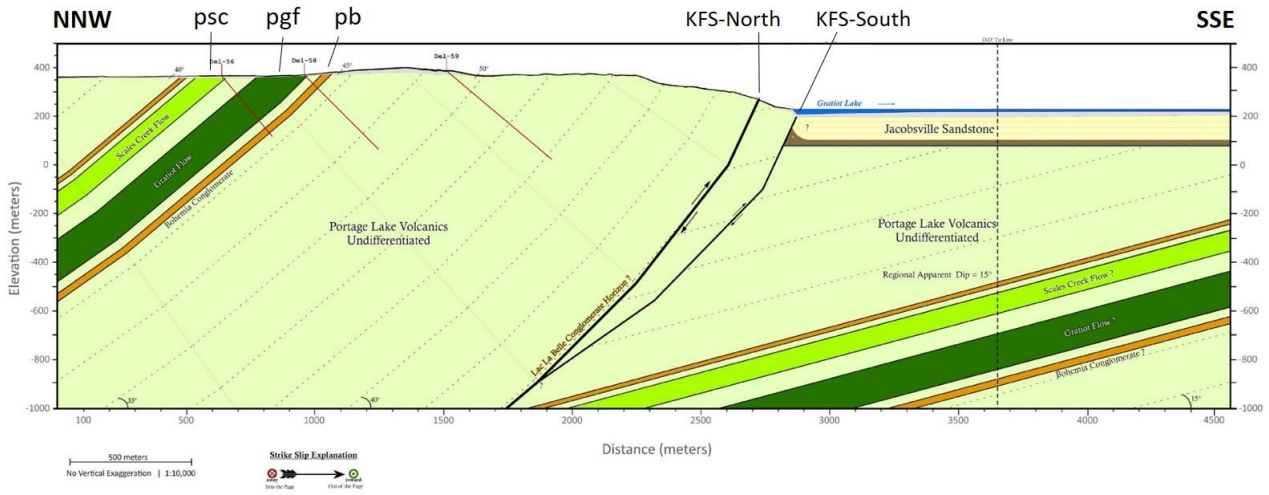


Figure 12: Cross-section along Fault Creek east of Eister Creek (Lizzadro-McPherson, 2023). Geologic unit codes: pb – Bohemia conglomerate; pgf – Gratiot flow; psc – Scales Creek flow. KFS = Keweenaw fault system.

Walking north up Eister Creek from the parking area, scattered outcrops of PLV basaltic lava are first encountered at the mouth of the incised creek valley. About 80 meters into the narrow gorge

where it becomes deeper, the basaltic rocks become more fractured and are locally sheared, brecciated, and gougy. Over the next 40-50 m upstream to the base of Eister Falls, fracture intensity increases and cataclastic shearing becomes more prominent until the rock is essentially a breccia cut through by shear zones. The area around the base of the falls is at the core of the main Keweenaw fault zone that is up to 45 meters wide along the creek, depending on how its northern and southern edges are defined. A rock wall perhaps 4 meters wide that projects from the eastern wall of the gorge consists of fault breccia and gouge that is better indurated that the surrounding material, which is also brecciated and gougy. This ridge is inclined steeply upstream, and is taken to define the orientation of the fault zone as striking 255° and dipping 76° N. Upstream from the projecting wall of breccia, the long gradual rise of Eister Falls exposes highly fractured basaltic rocks with some shear zones, but the rock mass is mostly intact and unlike the completely brecciated fault core. The fault-core relationships seen at Eister Creek are even better displayed along another creek located ~800 m east-northeast of here at a locality informally named "Fault creek" (Fig. 12).

Stop 3-3: Nine Thirty Two Creek

Directions: Return 1.0 mi (1.6 km) on East Gratiot Lake Road to Gratiot Lake Road. Turn right and drive 0.4 mi (0.6 km) to a narrow driveway on the right. Either turn into the driveway and park where possible or park along the right shoulder of the paved road. [Lat: 47° 21.711'N | Lon: 88° 8.730'W]

THIS STOP IS ON PRIVATE PROPERTY. PERMISSION IS REQUIRED TO OBTAIN ACCESS.

A walk of about 300 meters down the overgrown old road to Lake Gratiot leads to a hairpin turn in Nine Thirty Two Creek that marks the intersection of two fault trends (Fig. 11). The south branch of the main Keweenaw fault zone generally follows the creek valley upstream from the hairpin turn toward the west and, in the opposite direction, it follows an ENE-trending path to the north shore of Lake Gratiot. The thrust that defines the eastern edge of the Snake Creek block follows the western side of the creek downstream from the hairpin turn and terminates northward against the main Keweenaw fault zone.

The Keweenaw fault zone upstream of the hairpin turn is manifested by basaltic rocks that are fractured, brecciated, and altered for a few hundred meters to the west. Its east-northeast path is marked by a linear depression that connects to the neighboring creek valley where altered basalt occurs in a cutbank. PLV strata north of the Keweenaw fault zone generally strike east-west and dip moderately north based on two nearby creek traverses. These hanging-wall strata are juxtaposed against presumably younger footwall PLV strata west of the creek's hairpin turn and against footwall JS strata east of it (Fig. 11). This change in juxtaposition of geologic units along the Keweenaw fault zone occurs because the thrust fault intersecting it on the footwall side raises PLV strata on the west over JS strata to the east.

South of the creek's hairpin turn, JS strata crop out for 235 meters along the creek bed and valley walls and consist of yellowish to reddish, medium-grained, quartzose sandstone that is poorly

laminated along with subordinate reddish, muddy-to-silty, clast-supported conglomerate that is poorly indurated. The exposed JS strata are generally subhorizontal but their dip increases to 69° SE with a strike of 58° within 30 meters of the hairpin turn. The rotation of JS strata downward to the southeast is consistent with components of reverse slip on the intersecting faults that contain the sandstone within their obtuse angle (Fig. 11). Above the sandstone outcrops to the west, the upland caprock is a layer of trachyandesite overlying a basaltic layer, whose contact dips about 10° NE. This stratal orientation may represent a plunging fold axis that is parallel to fold axes near the southeast corner of the Snake Creek block.

Some important questions arise from the geometric relationships observed at the intersection of the two faults. For example, what happens to the main Keweenaw fault zone west and south of this fault intersection? Wright and Cornwall (1954) show the Keweenaw fault coming from Eister Creek as bending southwest to follow the PLV-JS fault contact discussed at Stop 3-1. They also show a splay of the Keweenaw fault continuing along Nine Thirty Two Creek as far as Gratiot Lake Road. Cannon and Nicholson (2001) continued this splay fault in a broad arc parallel to regional strike to a reconnection with the Keweenaw fault of Wright and Cornwall (1954) about 6 kilometers east of Mohawk. Based on new mapping and integration of DDH data, we have modified the trajectory of the fault extension proposed by Cannon and Nicholson (2001) and we propose that this is actually the main Keweenaw fault zone. The corollary to this interpretation is that the thrust fault along the east edge of the Snake Creek block is a footwall splay of the main fault zone. In other words, we think that the main Keweenaw fault zone is not always the one that juxtaposes PLV strata in the hanging wall against JS strata in the footwall.

Area 4: Keweenaw Fault Zone with Footwall Splays, Lac La Belle to Bête Grise

The three stops in Area 4 are along part of the KFS that changes direction from a 72° azimuth to nearly east-west (Fig. 3). The first stop is in the deformed hanging wall of the main Keweenaw fault zone (4-1) that runs along the northwest edge of Lac La Belle, past the southern base of Mt. Bohemia, and along most of the paved road to Bête Grise Bay east of here (Fig. 13). In a westerly direction, the main fault zone forms the northern boundary of the Deer Lake block, whose long dimension is again parallel to the KFS. The Deer Lake block is limited on the south by a footwall splay that diverges from the main fault zone north of Lake Gratiot and juxtaposes PLV basaltic flows on the north against nearly vertical JS strata at the Little Gratiot River (Lizzadro-McPherson, 2023; DeGraff, 1976). Based on diamond drill hole data, the eastern edge of this fault-bounded block is a connector fault where PLV strata are thrust eastward over JS strata, whereas the block's western edge has a thin cover of Jacobsville Sandstone unconformably overlying weathered basaltic rock. Field relationships and magnetic data suggest that the footwall fault splay curves north and merges into the connector fault along the east edge of the Deer Lake block.

The other two stops in Area 4 are at historic localities on the shore of Bête Grise Bay that Irving and Chamberlin (1885) investigated as part of their USGS Bulletin 23 titled "Observations on the junction between the eastern sandstone and the Keweenaw series on Keweenaw Point, Lake Superior". The Bête Grise block is defined by the main Keweenaw fault zone on the north, by a footwall splay that diverges from it onshore and passes offshore while remaining visible in shallow

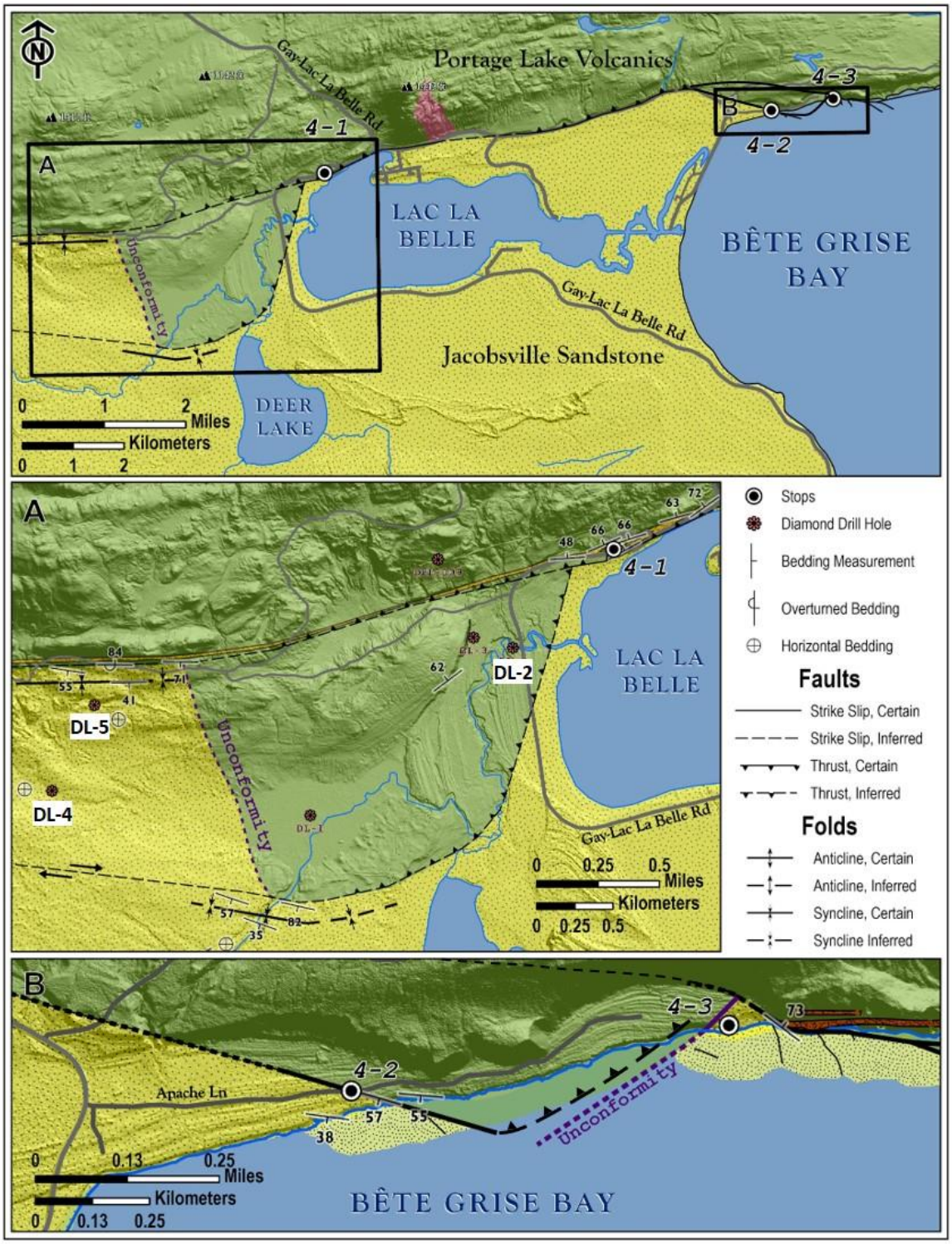


Figure 13: Geologic map of Area 4 with the Keweenaw fault system and field trip stops 4-1 to 4-3. Teeth along reverse faults are on upthrust sides.

water (4-2), and by an inferred thrust along the east edge of the block (4-3). The thrust is inferred from a NE-trending belt of intense fracturing and veining in altered basaltic rock that follows the shoreline parallel to the strike of the PLV-JS unconformity exposed to the east along the shore. Thrusting of hanging-wall PLV strata toward the southeast has tilted footwall PLV strata and JS strata above the unconformity 50° SE.

Stop 4-1: Haven Falls

Directions: Get back onto Gratiot Lake Road and drive 4.3 mi (6.9 km) back to US-41. Turn right and drive 5.7 mi (9.2 km) to Lac La Belle Road. Turn right and drive 4.2 mi (6.8 km) to a Y-intersection. Veer right and then through the sharp right curve for 0.5 mi (0.8 km) to Haven Falls Park on the right. Enter the park and park. [Lat: 47° 22.913'N | 88° 1.716'W]

The small but beautiful park at Haven Falls spans a terrace formed by a previous higher stand of Lake Superior that was about 10 meters above the current lake level. The cliff of felsic conglomerate over which Haven Creek flows was probably a shoreline cliff at the time of that higher lake level. This stop is in the proximal hanging wall of the main Keweenaw fault zone (Fig. 13) and its stratigraphy is exposed almost continuously from south of Lac La Belle Road, which is privately owned, to well upstream of the falls. About 650 meters east of here and south of the main fault zone, an east-directed thrust fault, cored by the Deer Lake #2 DDH, defines the east edge of the Deer Lake block (Fig. 13A). Slip on this thrust fault has pushed PLV basaltic lavas on the west up and over JS strata to the east (Lizzadro-McPherson, 2023) in a manner similar to what occurs at the east edge of the Snake Creek block. The Haven Falls stop is, therefore, analogous to the Eister Creek stop except that the fault core here is not as well exposed south of the paved road (cf. Figs. 12 and 14).

The stratigraphic sequence in the hanging wall begins with a highly fractured, veined, and locally brecciated and sheared lava flow that crops out on both sides of the paved road, but please stay on the north side to respect the private property on the south side. This strongly deformed lava flow lies along a topographic step up from the lowland at the lakeshore to the old lake terrace, and it probably marks the northern edge of the core of the fault zone. Upstream from this first lava flow, a 12-m-wide belt of conglomerate is followed by a 30-m-wide belt of ophitic basalt that locally is highly fractured, veined, and sheared. The felsic cobble-pebble conglomerate at the main falls is largely clast-supported and has a massive appearance. It extends for a horizontal distance of 16 meters along the creek to well above the top of the falls, where it is brecciated along a fault zone that cuts slightly up section toward the east. A reliable formation strike of 245° and dip of 65° NW comes from a sandstone layer above the falls near the northern edge of the second conglomerate layer. The two conglomerate layers here are collectively known as the Lac La Belle conglomerate (Cornwall, 1954a) and have been tentatively correlated with the Baltic (#3) conglomerate near Houghton. Although there is considerable uncertainty about this correlation due to the distance involved, the Lac La Belle conglomerate lies well below the St. Louis (#6) conglomerate seen at many of the earlier stops of this field trip. This means that the Keweenaw fault system here cuts the PLV section at a significantly deeper level than near Calumet, Laurium, and Lake Linden.

The stratigraphic units at the falls have been traced by detailed mapping along strike in both directions. About 350 meters east-northeast, the two conglomerate layers merge into one layer where the intervening basalt flow pinches out, but otherwise the layers can be traced continuously along strike. The only structural complications observed along strike are a few faults that cut at acute angles upward across layers toward the east and have small reverse offsets, similar to the fault at the top of Haven Falls. Detailed mapping in 2019-2020 did not find evidence of four transverse faults shown on the Delaware bedrock geology map as offsetting the Lac La Belle conglomerate, neither in terms of offsets nor enhanced fracturing (Lizzadro-McPherson, 2023). In fact, the kinematics of slip along the main Keweenaw fault zone would argue for a system of smaller subparallel faults rather than a series of faults normal to the main fault zone.

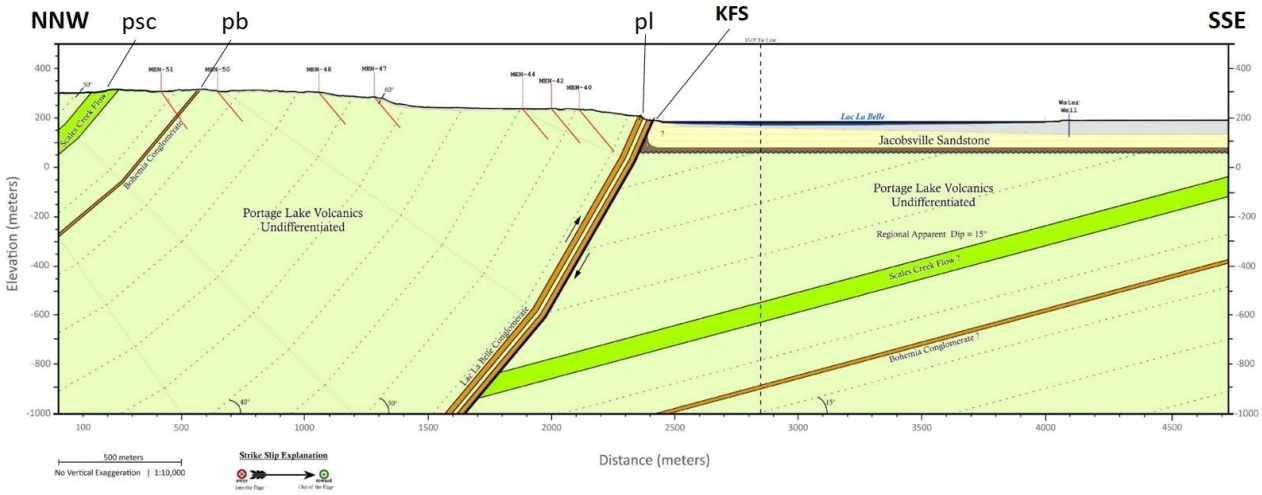


Figure 14: Cross-section along Haven Creek at the falls (Lizzadro-McPherson, 2023). Geologic unit codes: pb – Bohemia conglomerate; psc – Scales Creek flow. KFS = Keweenaw fault system.

THE NEXT TWO STOPS ARE ALONG APACHE LANE, WHICH IS PRIVATELY OWNED AS ARE ALL PROPERTIES ALONG IT. PERMISSION IS REQUIRED TO OBTAIN ACCESS.

Stop 4-2: Bête Grise Shoreline, Irving & Chamberlin Historic Site

Directions: Exit Haven Falls Park and drive east back to the stop sign at the Y-intersection. Turn sharply right onto Bête Grise Road and drive 3.2 mi (5.2 km) to Apache Lane on the left. [Geology Note: at 1.9 mi / 3.1 km along BG Road, an outcrop to the north is the location of the dated calcite vein in the fault zone.] Turn left, drive 0.5 mi (0.8 km), and then park along the right side of the road. [Lat: 47° 23.345'N | Lon: 87° 56.905'W]

We will walk down a moderately steep slope to the shore, where Irving and Chamberlin (1885) used a “force of miners” to strip the shoreline bare in order to better expose the contact between PLV layers on the north and JS strata to the south (Figs. 13B and 15). This is probably not a field

practice we could get away with today even if the site was not privately owned. Depending on the lake level, we may be able to see the contact between highly fractured, veined, and partly altered basaltic rock at the base of the shoreline scarp and JS strata that strike 105° and dip 55° S. An aerial image of the shoreline at this stop shows a sharply defined line with an azimuth of 105° that separates uniformly dark-toned PLV basaltic rocks on the lake bottom and along the shore to the east from variably lighter-toned layers of JS strata to the south and west (Fig. 16). This is the fault line that may be observed onshore or can be closely constrained by nearby outcrops. The aerial image shows many small faults as darker lines and narrow zones that cut JS strata at angles approaching 90° and offset strata by less than a meter or two. Near the fault line, splay faults break the JS unit into blocks up to 30 meters long parallel to the fault line and 4 meters wide.

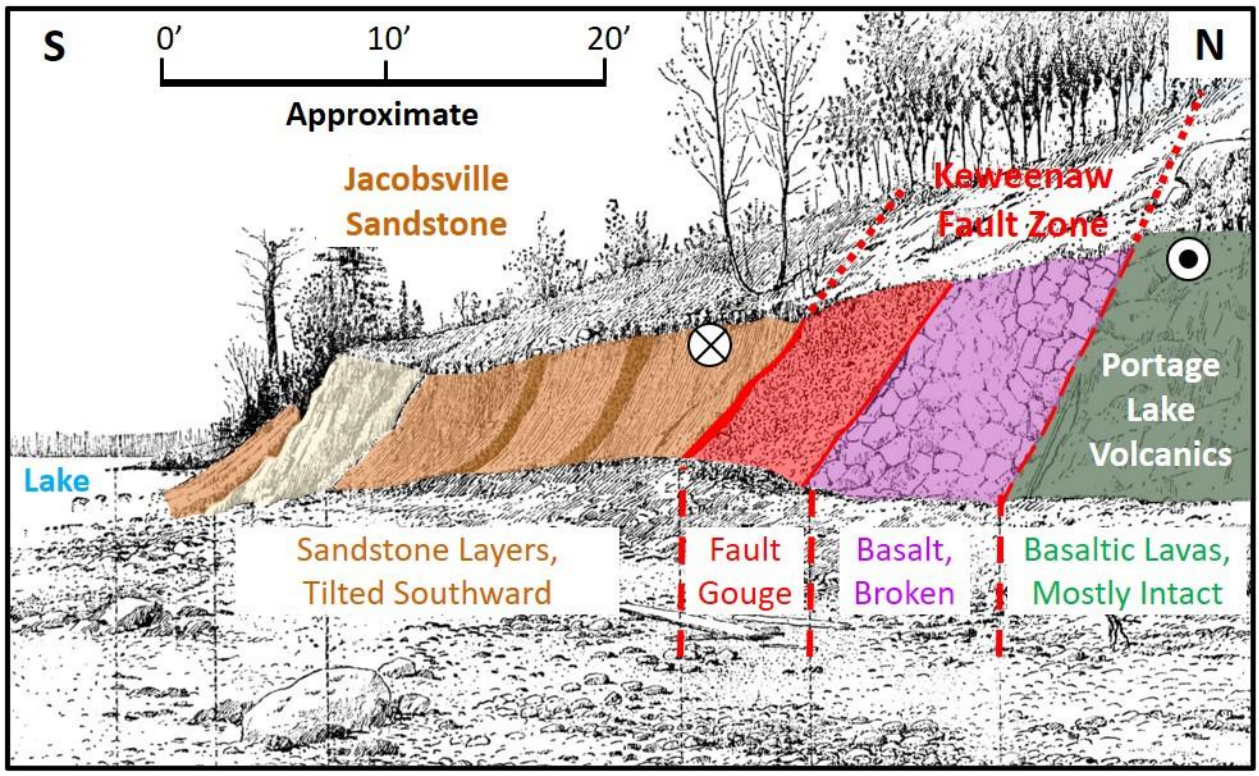


Figure 15: Cross-sectional view of a segment of the KFS exposed by excavation along the Bête Grise Bay shoreline. Fault strike = 100°, dip = 55° S (Irving and Chamberlin, 1885). Circle with black dot indicates movement toward viewer; circle with cross indicates movement away.

Onshore north of the fault line, PLV strata are highly fractured, veined, and sheared for at least 55 meters along the shoreline to the east, which is about 25 meters perpendicular to the fault. Basaltic outcrops along Apache Lane and to the north do not exhibit such deformation. Following the shore to the west and south, JS strata change in terms of both facies and structural orientation. The basal part of the section consists of reddish, thin-bedded, siltstone and mudstone with minor fine-grained sandstone (6-7 m true thickness), followed upward by a lighter-toned package of silty to fine-grained sandstone with minor silty pebble conglomerate that forms a resistant ridge (~2 m), and then an interval of reddish siltstone and muddy conglomerate (~4 m).

From this point southward and up section, JS strata tend to be sand-prone but alternate between: (1) pinkish to orange, fine to medium grained, quartzose sandstone; (2) red, thinly bedded, siltstone to mudstone; and (3) muddy to silty, poorly indurated, conglomerate layers. In summary, the JS unit tends to clean upward from a silt- and mud-dominated basal section with conglomeratic units near the fault to a quartzose sand-prone section away from the fault. Along with these stratigraphic changes, the dip of JS strata decrease from 55° S near the fault to about 20° S at a perpendicular distance of 60 meters from the fault, and presumably becomes subhorizontal not much further to the south.

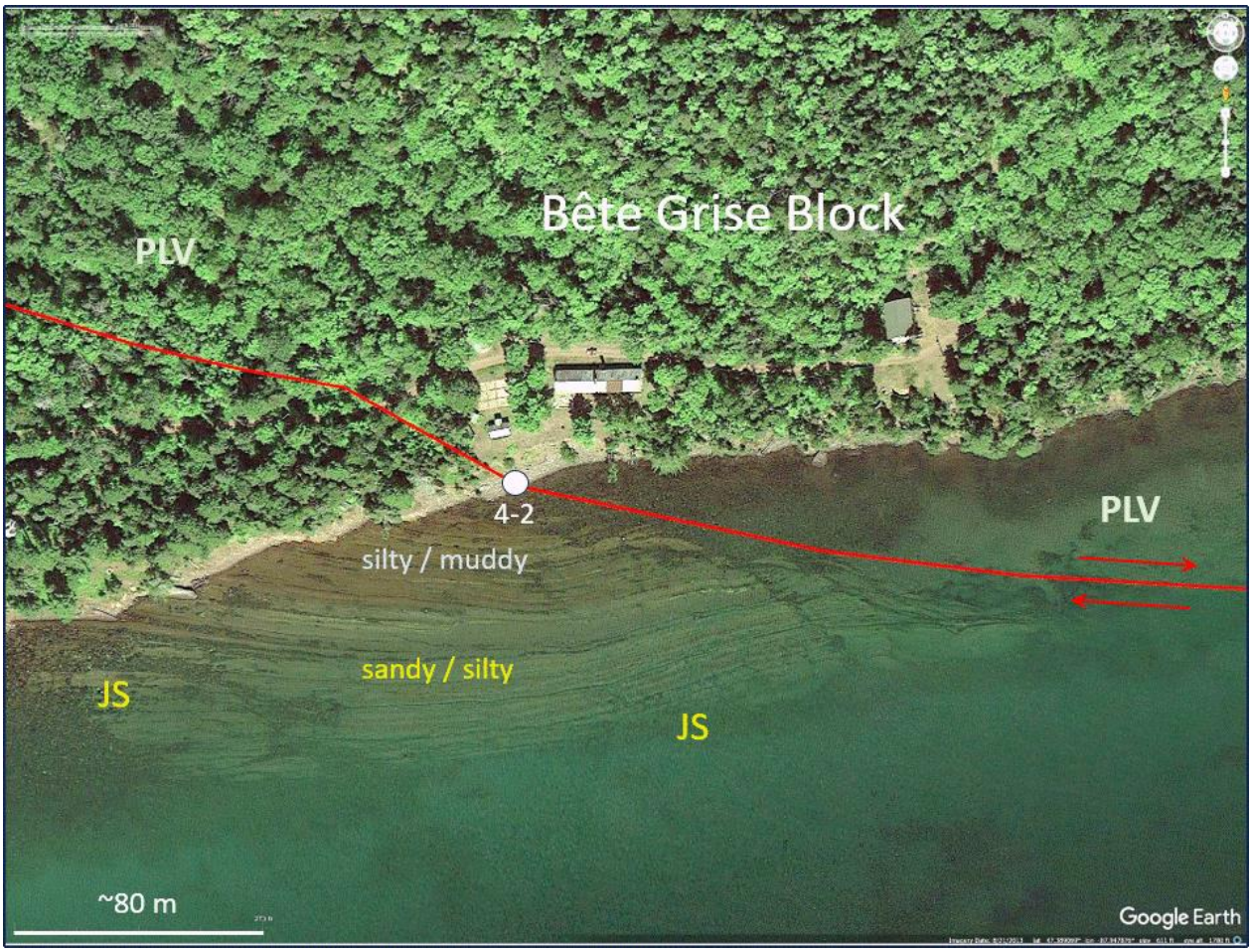


Figure 16: Aerial image of Stop 4-2 showing the south boundary fault of the Bête Grise block. The fault is interpreted to have dominant strike slip. Stratigraphic up in the JS is to the south.

It is clear that movement along the fault has juxtaposed older PLV strata to the north against younger JS strata to the south and that the north side has a component of upward movement, as noted elsewhere. However, what is the nature of this fault? The work by Irving and Chamberlin (1885) and their team exposed a fault surface that dips about 55° S, essentially parallel to adjacent JS strata (Fig. 15). Based on textbook definitions, this would be a normal fault with younger JS strata in the hanging wall to the south above older PLV strata in the footwall to the north. Something about this interpretation seems paradoxical, however, because at previous stops

the main fault zone dipped north or northwest and also because the tectonic setting of the Keweenaw fault system was compressional.

Part of the paradox results from textbook definitions of reverse and normal faults that are based on idealized planar fault surfaces that have mostly dip slip. In the case of curved or corrugated fault surfaces with mostly strike slip, the textbook nomenclature for dip-slip faults may lead to confusion. Depending on the portion of a corrugated fault surface that is exposed, a mostly strike-slip fault with a smaller dip-slip component may exhibit an apparent normal or reverse component of dip-slip even though the actual dip-slip component is the same everywhere along the fault. The Keweenaw fault system in this area trends nearly east-west and is dominated by right-lateral strike slip with lesser north-side-up dip slip (2:1 ratio of strike-to-dip slip), and we infer that motion on this fault is dominantly right-lateral strike slip. The fault forms the southern edge of a fault-bounded block (Fig. 13B) whose east side will be visited next.

Stop 4-3: Bête Grise Shore, PLV-JS Unconformity and Fault

Directions: Continue driving east on Apache Lane for 0.4 mi (0.6 km) to the end of the road. Park where space allows. [Lat: 47° 23.437'N | Lon: 87° 56.350'W]

We will walk 100 meters east and 50 meters south to access the shoreline, where geologists traveling along the coast by boat in the mid-1800s described impressive layers of sandstone on the rocky bottom of Lake Superior (Figs. 13B and 17). This area was examined later by none other than Irving and Chamberlin (1885) and then by Cornwall (1954b). The aerial image of the shoreline and offshore region reveals what the early explorers reported, a set of well-defined parallel layers that curve sharply from NE-trending layers on the western side to EW-trending layers along the shore to the east. The shoreline stop is at the western edge of the JS strata where an unconformity between PLV and JS strata is tilted about 50° SE.

Depending on water level and shoreline erosion, we may be able to see the unconformity between saprolitized PLV basaltic lava to the northwest and JS strata to the southeast. If the saprolite is exposed, please do not disturb it by digging, picking at it, or walking on it. The basaltic protolith has been completely converted to clay minerals and still retains its original textures, including whitish veins that are approximately normal to the tilted unconformity. Northwest of the unconformity, i.e. deeper below the paleosurface, PLV basaltic rocks do not exhibit such alteration. Saprolitic basaltic rock that retains original textures, such as ophitic texture, has been observed elsewhere in the area where the PLV-JS unconformity is inferred, such as the east end of the Deer Lake block (Lizzadro-McPherson, 2023; DeGraff, 1976).

East of the unconformity along the shore, a recessive basal part of the JS section consists of thinly bedded, reddish siltstone and mudstone with minor interbedded fine-grained sandstone. The recessive strata here strike northeast and dip moderately southeast over a distance of 25 meters to the first small resistant sandstone layer (1-2 m thick). Next in the section is another recessive interval (6-7 m wide) of reddish siltstone with interbeds of poorly indurated muddy conglomerate, followed by a larger ridge of resistant sandstone that begins a sequence of

alternating resistant sandstone beds, recessive reddish siltstone, and reddish poorly indurated conglomerate. Thus, the basal JS section here is very similar to the section at Stop 4-2. The main difference is in their structural attitude, which differs in strike by 55-60°. At Stop 4-2, the dip direction of JS strata is 195° and away from the inferred strike-slip fault on the south side of the Bête Grise block, whereas the dip azimuth of JS strata on the east side of the fault block is 138° and away from the tilted PLV-JS unconformity (Fig. 13B). We infer that southeast tilting of the PLV-JS unconformity resulted from southeast thrusting along the east edge of the Bête Grise fault block, based in part on intense fracturing observed along the shoreline northwest of the unconformity at stop 4-3.

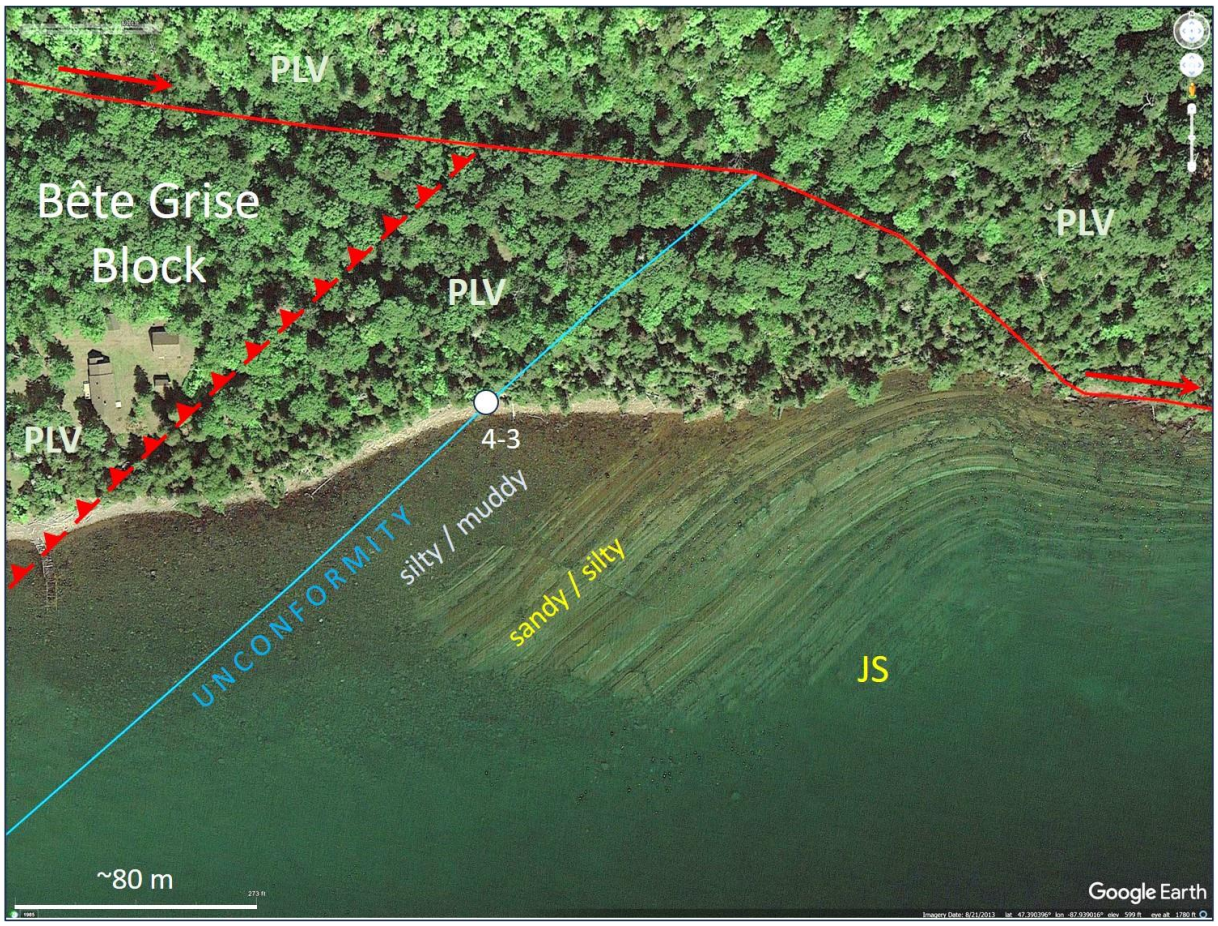


Figure 17: Aerial image of Stop 4-3 showing the north and east boundary faults of the Bête Grise block. The east boundary fault is interpreted to have dominant dip slip with west side thrust eastward, whereas the north boundary fault is interpreted to have dominant strike slip. Stratigraphic up in the JS is to the southeast and south.

If time permits, we will ascend the shoreline scarp to the flat bench above and walk another 100 meters to reach the eastern edge of JS outcrop along the shore. Here, JS strata strike nearly east-west and are vertical to slightly overturned to the south, forming narrow ridges and eroded furrows and clefts due to differential erosion of the resistant and recessive layers (Fig. 17). Near the eastern edge of JS outcrop, vertical JS strata are flanked on the north by PLV strata that begin

with a felsic conglomerate and continue upslope with PLV basaltic lavas. The contact between PLV strata on the north and vertical JS strata to the south is a major fault that intersects the shoreline, where a fault breccia is exposed several meters east of the last onshore JS outcrop. From this point, the fault turns eastward and runs along the shoreline for 200 meters before continuing offshore and splitting into two branches.

This is the last stop and we hope that you had a good experience that will help you to understand other fault systems. Thank you for your participation!

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Field Trip 5

Geology and History of a Native Copper Mine: Adventure Mine, Ontonagon County, Michigan

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[Latitude: 46.777224; Longitude: -89.081906]

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). Stay on M-38 towards Ontonagon. Follow signs to Adventure Mine. ***The Adventure Mine is privately owned and operated as a publicly available mine tour***

Introduction

The historic Adventure Mine is part of the Greenland-Mass subdistrict of the Keweenaw Peninsula native copper district of the western Upper Peninsula of Michigan (Figures 1 and 2). The Adventure Mining Company began mining native copper in 1850. It permanently ceased mining in 1917 at the time when most small mines of the subdistrict ceased mining operations. From the 1980s to today, the mining activities at the Adventure Mine have yielded specimens of massive native copper and copper crystals for purchase by tourists and mineral collectors.

The Adventure Mine is a Keweenaw Heritage Site of the Keweenaw National Historical Park. It is located in the Greenland-Mass subdistrict about 40 km southwest of the Baltic Mine in Painesdale, the southernmost major native copper mine in the district (Figure 2). The subdistrict yielded about 85 million lbs (39 million kg) of refined copper at grades ranging from 0.5 to 1.25 percent (Butler and Burbank, 1929; Weege and Pollock, 1971).

The Adventure Mine was the second largest producer in the subdistrict, yielding about 11 million lbs. (5 million kg) of refined copper from the tops of five different basaltic lava flows (Butler and Burbank, 1929).

Despite the relatively low production of copper from the Greenland-Mass subdistrict (~0.8 % of the total district production of ~ 5 billion kg (11 billion lbs), the geologic characteristics of the Greenland-Mass subdistrict deposits are typical of the native copper deposits elsewhere in the Keweenaw Peninsula (e.g., Stoiber and Davidson, 1959; Butler and Burbank, 1929; etc.). Since

mining in the Keweenaw Peninsula native copper district ceased in 1968, underground access to observe or study the native copper deposits has been limited. Currently there is access to only two mines in the main area of the district (Quincy and Delaware) and two mines in the Greenland-Mass subdistrict (Adventure and Caledonia). These mines are operated as underground experiences for tourists, except for Caledonia.

This field trip guide relies on existing publications by Bornhorst and Whiteman (1995); Bornhorst et al. (2013); Bornhorst and Barron (2011); and Butler and Burbank (1929). The overview of the geologic and part of the human history is summarized from Bornhorst and Lankton (2009), Bornhorst and Mathur (2016), and Bodden et al. (2022). Brandon Erickson prepared a brief history of the Adventure Mine which is included in this guide.

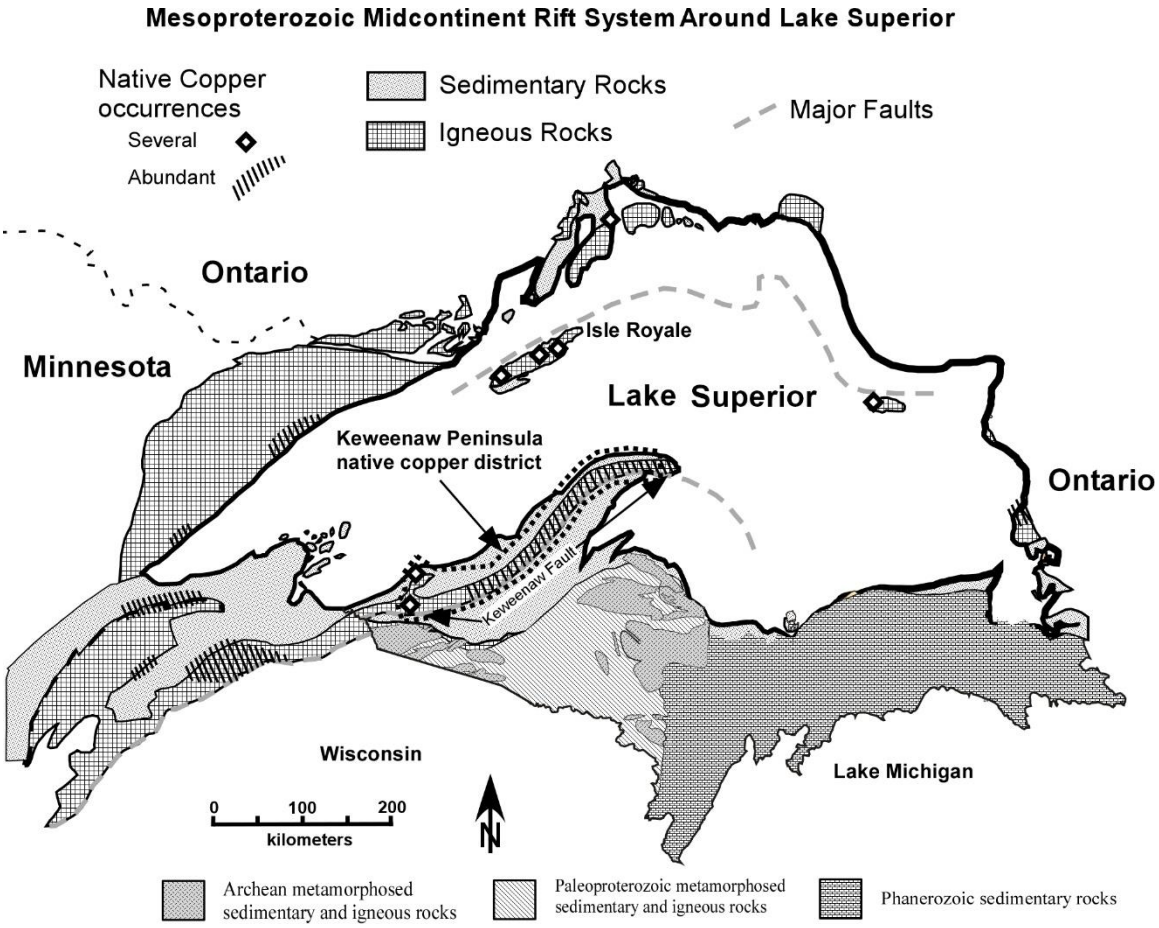


Figure 1: Simplified bedrock geology of the Mesoproterozoic Midcontinent Rift around Lake Superior. Modified after Bornhorst et al., 2013)

Overview of Geologic History

The largest known accumulation of native copper in the world is the Keweenaw Peninsula native copper district (Figure 2). In comparison to other copper mining districts where major copper ore

minerals are sulfides, nearly all of the copper in the district occurs as native copper. About 5 billion kg (11 billion lbs) of refined copper were extracted from about 380 million tons of ore between 1845 and 1968 via underground mines (Weege and Pollock, 1971). Small quantities of native silver occur with the native copper. An estimate using incomplete records suggests the amount of silver was between .05 to .5 oz per ton of ore. Native copper and silver were co-precipitated.

The native copper deposits are hosted by the Mesoproterozoic Midcontinent Rift (MCR) (Figures 1 and 2). More than 25 km of volcanic rocks and 8 km of clastic sedimentary rocks fill the center of the MCR (Cannon et al., 1989 and 1993). These rift-filling rocks were emplaced between about 1.15 to ~1 Ga (Cannon et al., 1989; Davis and Paces, 1990; Kulakov et al., 2018; Heaman et al., 2007).

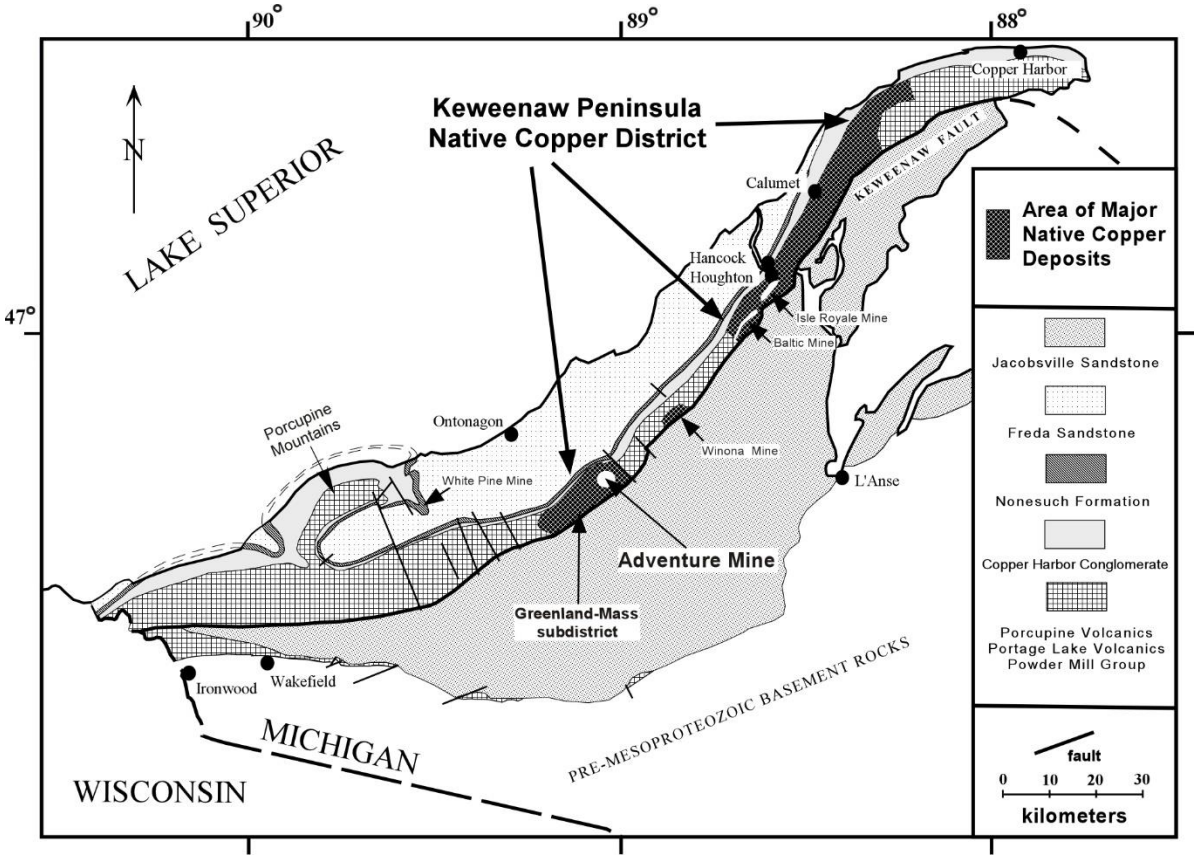


Figure 2: Simplified geologic map of the Keweenaw Peninsula and vicinity modified from Bornhorst et al. (2013). At the White Pine mine "Keweenawan" native copper cuts across diagenetic the shale-hosted chalcocite deposit.

Early eruptions of MCR basaltic lava flows were scattered on a broad land area above a developing mantle plume. These early eruptions were followed by many eruptions from fissure vents concentrated in the center part of the MCR (now buried under the center of Lake Superior). These eruptions were dominated by fissure volcanoes along linear faults. The Portage Lake Volcanics and Porcupine Volcanics are rift-filling basaltic volcanic rocks erupted about 1.1 billion years ago

during the active rifting of the MCR (Figure 3). The MCR was bounded on the edges by down dropped normal faults resulting in the MCR being a faulted basin. The basin progressively dropped down by stretching and by magma erupted at the surface during active rifting.

Eruptions of basalts of the Portage Lake Volcanics were on land surface (subaerial). Subaerially erupted basalt lava flows have either a vesicular or brecciated and vesicular top (pahoehoe or aa lava flow). Mineral filled vesicles are termed amygdules and flow tops dominated by vesicles are termed amygdaloids and those dominated by breccia clasts of amygdaloidal basalt are termed fragmental amygdaloids. The top of a subaerial lava flow is underlain by a massive (relatively vesicle-free) basalt. Massive basalt flow interior in thinner flows is fine grained and in thicker flows it is coarse-grained (ophitic). Thin flows can be vesicular throughout with vesicles more abundant at the top. The typical flow is 10 to 20 m thick. Eruptions of basaltic lavas were cyclical and during eruptive hiatuses minor gravel and sand were deposited on top and infiltrated in the tops of occasional lava flows. These clastic sedimentary layers are overlain by basalt lava flows. A desert environment 1.1 billion years ago resulted in red coloration of these clastic sedimentary rocks.

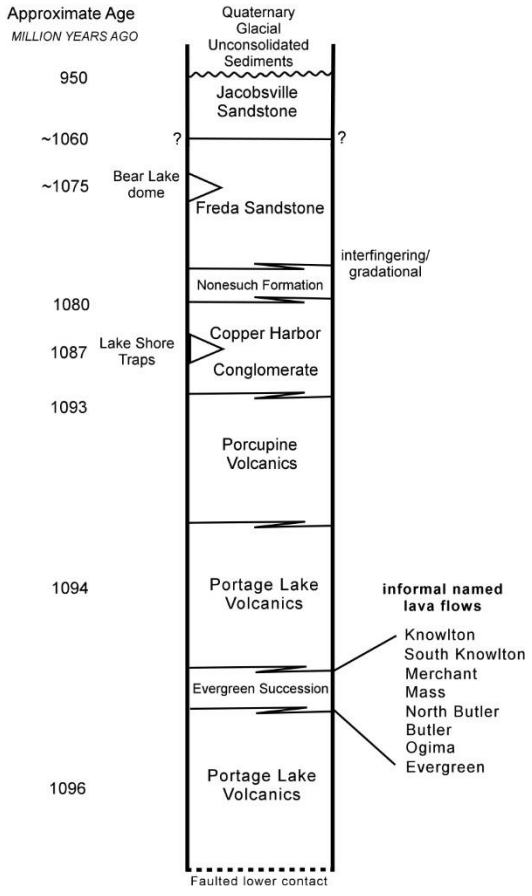


Figure 3: Stratigraphic column for the Adventure Mine region with approximate ages.

Active rifting and basaltic volcanic activity ended over a short period of time, but the rift basin continued to passively sag and was progressively filled with clastic sedimentary rocks from the Copper Harbor Conglomerate to the Freda Sandstone (Figure 3). In the Adventure Mine region, the

rift-filling volcanic rocks (Porcupine Volcanics and Portage Lake Volcanics, Figure 3) were first covered by red-colored gravels and sands (Copper Harbor Conglomerate. Overlying the Copper Harbor Formation are black- to gray-colored muds and silts (Nonesuch Formation). Lastly the rift was filled with a thick section of red-colored fine sandstones (Freda Sandstone). Today, the rocks of the Keweenaw Peninsula and Adventure Mine region span the edge of the MCR, and consist of a thick section of rift-filling subaerial basaltic lava flows overlain by a thick section of rift-filling clastic sedimentary rocks (Figures 1, 2, and 3).

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 resulted in reverse and thrust faults as well as folding and fracturing of rift-filling volcanic and clastic sedimentary rocks. Native copper and related minerals were emplaced during this regional compressional event (Bornhorst, 1997). From the beginning of regional compression at about 1.06 to 1 billion years ago the Jacobsville Sandstone was deposited in a rift-flanking basin until about 1 billion years ago (Figure 2).

There were no recorded geologic events by rocks of Michigan's Upper Peninsula from about 1.0 billion years ago to 500 million years ago. Erosion likely exposed the native copper deposits at the surface and downward percolating oxidizing groundwaters had access to alter the native copper (Bornhorst and Robinson, 2004). The MCR rocks were buried by Phanerozoic rocks deposited from 500 to 175 million years ago (Catacosinos and others 2001). The native copper deposits of the Keweenaw Peninsula were again at the surface after erosion of overlying rocks by Pleistocene glaciers over the last 2.5 million years.

Native Copper Deposits of the Keweenaw Peninsula

The cumulative pre-mining geologic copper resource of the Keweenaw Peninsula native copper district totaled about 9 billion kg of copper (20 billion lbs; Bornhorst and Barron, 2011). About ½ of the geologic resource was recovered. Speculative concentrations of copper in rocks are even greater. These concentrations have not been mined for various reasons such as too low of grade or too deep below the surface.

Permeable and porous primary geologic settings that have sufficient open spaces that were sufficiently connected with each other facilitated the movement of ore-forming hot waters (hydrothermal fluids) from which native copper and other minerals were precipitated. Compression of the MCR strata integrated the primary permeability and porosity of the hydrothermal plumbing system with compression generated faults/fractures (Bornhorst, 1997). The absolute age of the main-stage of hydrothermal activity coincides with the age of the regional compressional event at about 1.06 to 1.04 billion years ago (Bornhorst et al., 1988). The permeable and porous tops of amygdaloidals and fragmental amygdaloids hosted ~58.5% of produced native copper. Horizons of conglomerate and sandstone between lava flows, which are also permeable and porous, hosted ~39.5% of produced native copper. Native copper ore bodies are "sandwiched" on the bottom side by the massive basalt interior of the flow whose top hosts the native copper ore body. On the top side the ore bodies are sandwiched by massive basalt of the overlying lava flow. The sandwiched ore-bodies are geometrically approximately tabular

(called lode) with a thickness between 3 and 5 m and the same orientation as surrounding host lava flows. The typical lode extends down-dip 1.5 to 2.6 km and has a lateral extent of 1.5 to 11 km (Butler and Burbank, 1929; White, 1968). Open spaces in amygdaloidal lava flow tops (vesicles) and sandstones/conglomerates are typically up to a cm across. They are typically filled dominantly by gangue minerals and lesser native copper. Less frequently the entire open space was filled with masses of native copper. Open spaces between breccia fragments in the top of a lava flow (fragmental amygdaloid) or between clasts in conglomerate will tend to have larger masses of native copper and can weigh up to several lbs, to tens of lbs to hundreds of lbs. and rarely weighing tons.

A minor amount of the total produced native copper, ~ 2%, was from sub vertical tabular open spaces (veins when filled with minerals) that follow faults and fissures that perpendicularly cut across the volcanic-dominated strata. Ore-forming hydrothermal fluids readily moved along faults and fissures since they have a relatively large amount of interconnected open-space; these ore-bodies are also tabular lodes. Since the size of open space is large the corresponding size of masses of native copper can also be large weighing multiple tons with the largest masses being several hundred tons.

Native copper is closely associated with about 22 common and many more uncommon minerals (Bodden et al., 2022; Butler and Burbank, 1929; White, 1968). These minerals fill the same open spaces along with and instead of native copper. The suite of minerals is similar to those found where rocks have undergone very low to low grade burial metamorphism at less than about < 300^{O}C (Bodden et al., 2022). Thermal modeling suggests that peak burial metamorphic conditions at depth were between 400 to 500^{o}C (Woodruff, 1995). Burial metamorphic processes resulted in ore-forming hydrothermal fluids carrying copper leached from the tops of buried rift-filling basalt lava flows. Batches of metamorphogenic-dominated hydrothermal fluid generated over time were similar to one another.

The rift-filling volcanic rocks were very low in sulfur when they erupted, and the little contained sulfur degassed into the atmosphere. During subsequent deposition of rift-filling clastic sedimentary rocks there was an incursion of seawater into the rift for a significant amount of time. This resulted in seawater deeply penetrating into the underlying rift-filling volcanic rocks (Figure 4A). During burial, the rift-filling volcanic and clastic sedimentary rocks were progressively heated and during initial heating the seawater evolved to be depleted in sulfur similar to expelled modern sea floor hydrothermal fluids (Figure 4B). Continued heating during burial resulted in burial metamorphic-dominated hydrothermal fluids with copper leached from the rift-filling volcanic rocks. These fluids were well mixed with the evolved seawater resulting in hybrid metamorphic-dominated ore-forming fluids (Figure 4C). These main-stage hybrid metamorphic-dominated ore-forming hydrothermal fluids moved upwards from the source zone through the same very sulfur poor strata as in the source rocks (Figure 4D). As they moved upwards they cooled, interacted with host rocks, and in the relatively shallow zone of precipitation they variably mixed with sulfur-poor reduced meteoric water (Figure 4D; Bodden et. al, 2022). These processes resulted in precipitation of native copper and main-stage hydrothermal minerals. Higher temperature main-stage mineral assemblages are spatially associated with the area of native copper deposits where the thermal anomaly was greatest

because of focused hydrothermal fluids. Within the native copper district, the suite of main-stage minerals that is associated with native copper and is followed by late-stage minerals precipitated at lower temperature than the main-stage.

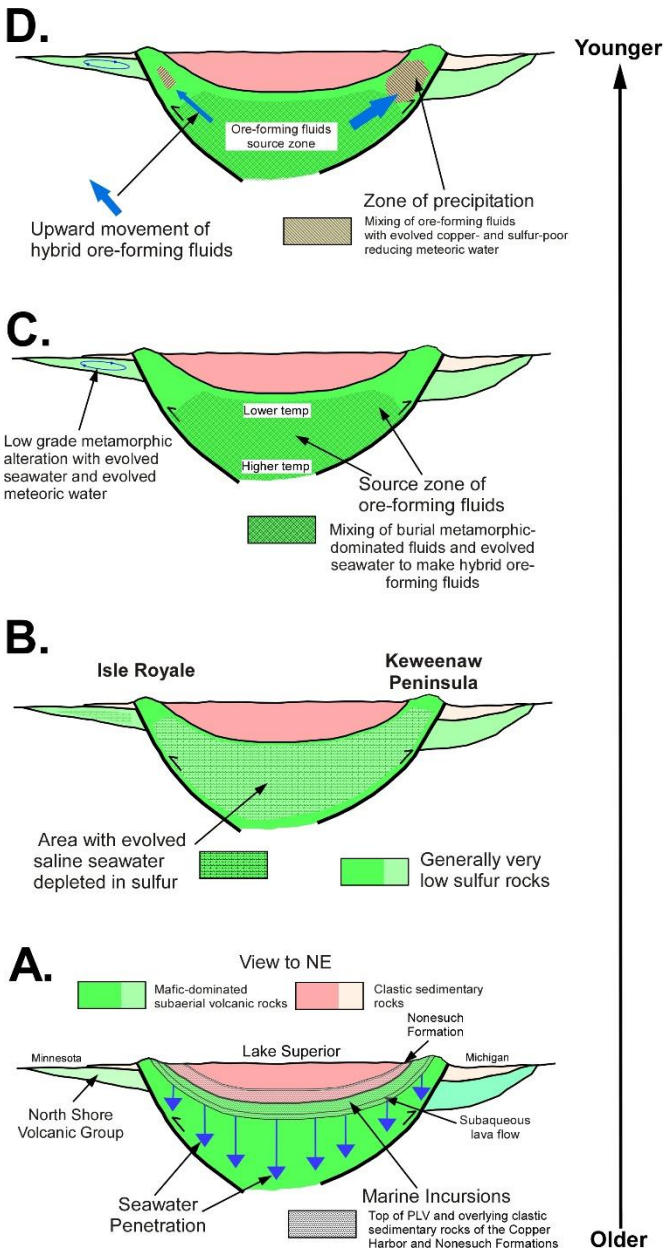


Figure 4: Cartoon cross sections showing conceptual genetic model of the native copper deposits of the Keweenaw Peninsula formed at about 1060 to 1040 million years ago. Modified from Bodden et al. (2022). A. Marine incursions and seawater penetration during deposition of volcanic and sedimentary rocks in MCR. B. Area prior to burial metamorphism with sulfur depleted evolved seawater providing salinity for ore-forming fluids. C. Burial metamorphic fluids mixing with evolved seawater produce hybrid ore-forming fluids. D. Precipitation of main-stage minerals, including native copper, as a result of mixing of ore-forming fluids with meteoric water, decreasing temperature, and water-rock reactions.

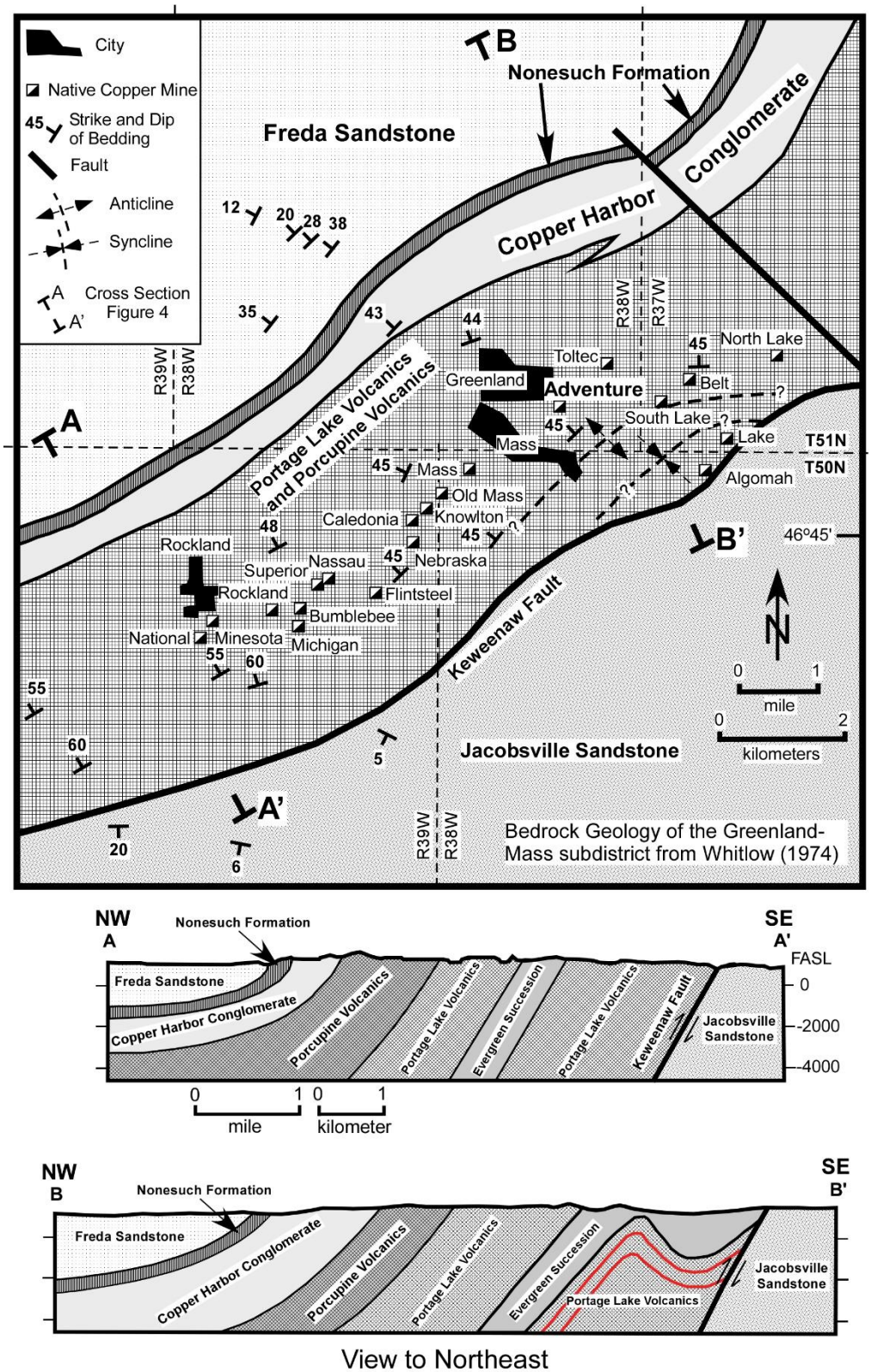


Figure 5: Bedrock geologic map and cross sections of the Greenland-Mass subdistrict of the Keweenaw Peninsula Native Copper District. Modified from Whitlow (1974).

The Evergreen Succession

Butler and Burbank (1929) recognized the Evergreen lava flow and a succeeding number of lava flows of the Portage Lake Volcanics as having distinctive lithologies and hosting the native copper deposits in the Greenland-Mass subdistrict (Figure 3 and 5). These are informally termed the Evergreen Succession (Figure 3). The Evergreen Succession is stratigraphically about 150 m (500 ft) above the Bohemia (No. 8) conglomerate (Butler and Burbank, 1929).

The Evergreen Succession basaltic lava flows are slightly more intermediate in composition than other lava flows within the PLV (Butler and Burbank, 1929). They are characterized by porphyritic or glomerporphyritic texture although thicker flows are ophitic. It is difficult to correlate individual lava flow with one another except in developed areas for mining of native copper where individual lava flow can be traced along strike from mine to mine. In the Greenland-Mass subdistrict most of the native copper was produced from the tops of lava flows of the Evergreen Succession. Those flow tops hosting native copper are generally fragmental amygdaloidal lodes with the best areas for native copper being where the flow top is thicker. Thin amygdaloidal only flow tops or those with areas of massive basalt mixed in the flow top are typically lower grade. The Evergreen Succession is at a similar stratigraphic position as those lava flows developed at the Isle Royale Mine to the north of the subdistrict in the Houghton area of the main district (Butler and Burbank, 1929). The Evergreen flows were also developed for native copper in the Winona area in the middle between the main district and the subdistrict.

The individual copper-rich lava flows within the Evergreen succession were each informally named (Figure 3). The Evergreen flow is a 3 to 15 m thick plagioclase porphyritic lava flow. The Ogima flow is a 30 to 43 m thick slightly plagioclase glomerophyritic basalt lava flow. The Butler flow is a 15 to 27 m thick plagioclase glomerporphyritic basalt lava flow. The Mass and Merchant flows are up to about 25 m thick. The South Knowlton flow is up to 15 m and is a plagioclase glomeroporphyritic basalt. At the top of the Evergreen succession is the Knowlton flow which is a 9 to 21 m thick plagioclase glomeroporphyritic basalt. Between the Butler and Knowlton flows there are a number of thin flows of plagioclase glomeroporphryitic basalt with total thickness of 75 to 90 m thick (Calumet and Hecla, 1958).

The tops of the Evergreen lava flows were productive over a strike length of about 5 km. Native copper mined from the Evergreen Succession was extracted from many different mines with some of them connecting with others. The Butler flow top yielded the most copper followed by the Evergreen and Knowlton flow tops which also yielded significant amounts of copper. Vesicle- and inter-fragment void-fillings consist of quartz, calcite, K-feldspar, epidote, prehnite, pumpellyite, and chlorite (Table 1). Less abundant main-stage minerals are native copper, native silver, and datolite. Laumontite and adularia are common late-stage minerals.

Table 1: Percent amygdule-filling minerals estimated from rock piles adjacent to mines/shafts of the Greenland-mass subdistrict. Unpublished data by Stoiber and Davidson 1959).

	% Amygdule-Filling Mineral						
	Quartz	Calcite	Red K-Feldspar	Epidote	Prehnite	Pumpellyite	Chlorite
Mine/Shaft							
Adventure #1	52	5	6	26	2	9	tr
Adventure #2	30	20	5	40	0	1	4
Adventure #3	11	13	30	37	5	1	3
Adventure #4	36	27	0	2	24	10	1
National #2	22	17	3	5	53	0	0
Old Mass	17	21	40	15	4	2	1
Mass C	22	10	45	19	0	3	1
Mass 1 &2	55	23	0	22	0	trace	trace
Mass B	19	22	40	18	0	trace	1
Mass A	45	16	trace	35	trace	3	1
Michigan	63	27	trace	5	5	0	0
Michigan	21	30	35	3	10	0	0
Flintsteel #1	18	52	18	5	0	trace	7
Flintsteel #2	23	45	18	5	0	0	9
Butler	20	45	20	5	0	10	0
Knowlton	30	38	15	6	8	0	8

The Evergreen Succession in the Greenland-Mass subdistrict dips about 45º NW and forms a local broad open anticline (Fig. 6). The largest mine, the Mass Mine, occurs near the maximum bend in this anticline. Most faults have displacement of < 1 m while those faults with significant vertical displacement are uncommon. There are multiple veins in tension fractures in the area of maximum bend that cut perpendicular across the lava flows (Butler and Burbank, 1929). There are some veins that are parallel to the 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-forming fluid into the top of the lava flow.

The Adventure Mine

The Adventure Mine was very small, producing only about 5 million kg (11 million lbs) of refined copper, in context of all mines in the district which produced about 5 billion kg (11 billion lbs. Most of the production of native copper from the Adventure Mine came from the top of the Knowlton basalt lava flow (Knowlton lode; Figure 6 and 7). There was also significant

production from the Butler lode and minor production from the Evergreen, Ogima, and Merchant lodes (Figure 6 and 7).

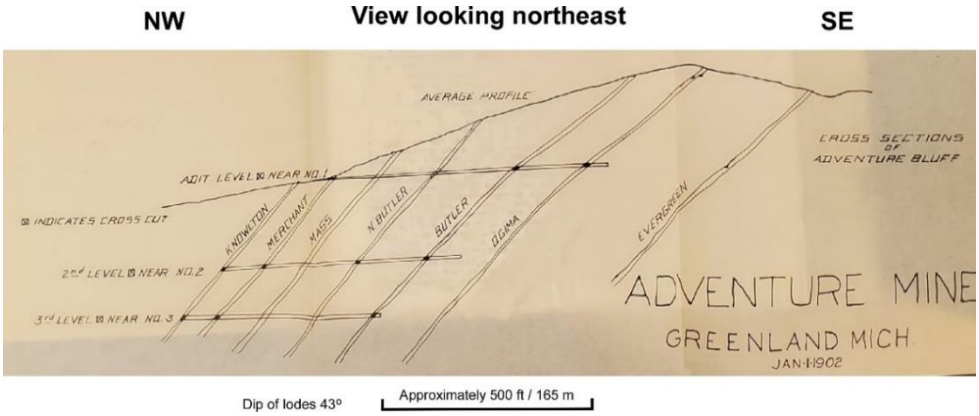


Figure 6: Historic 1902 sketch cross section showing the lodes of the Evergreen succession at the Adventure Mine. Adits perpendicular to strike of the lava flows are shown and drifts parallel to the strike are indicated by black squares. The field excursion utilizes the adit near the No. 1 shaft.

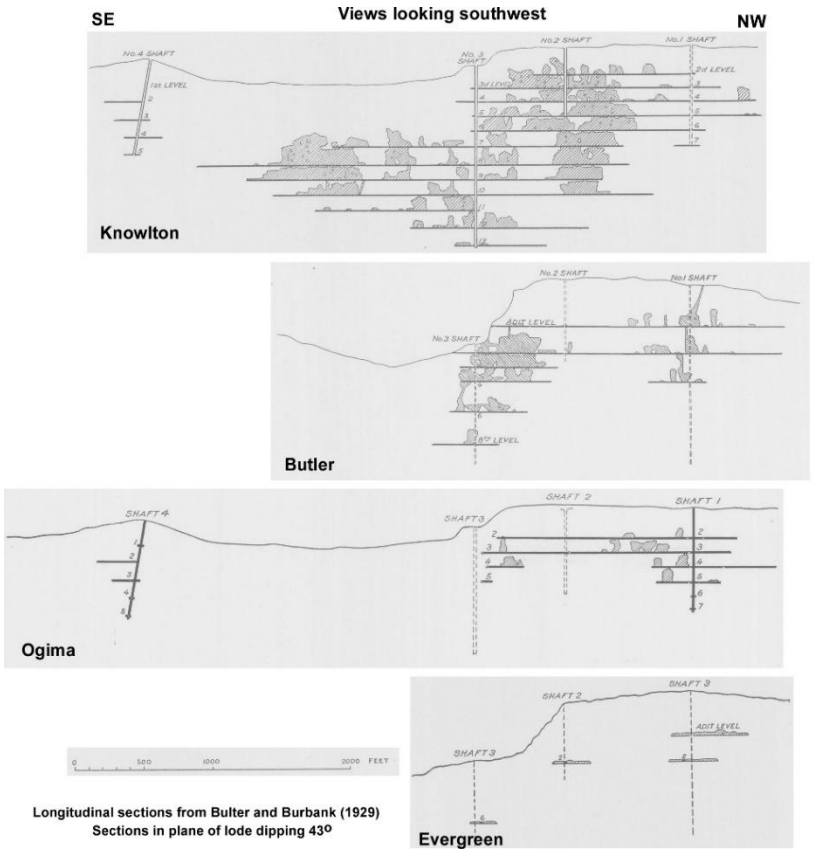


Figure 7: Longitudinal sections (parallel to strike) of the Evergreen succession native copper lodes showing underground workings (openings) for the Adventure Mine shafts #1 to #4. Section from Butler and Burbank (1929).

The Knowlton was the focus of native copper mining at the Adventure Mine. The Knowlton lava flow top is a fragmental amygdaloid. In the subdistrict, the Knowlton flow top was developed for about 3000 m along strike and to a maximum depth of about 375 m. At the nearby Mass Mine (Fig 5), the Bulter lava flow top was the principal focus of native copper mining. It was the second focus of mining at the Adventure Mine. In the subdistrict the Butler lava flow top was developed for about 2000 m along strike and to a maximum depth of 300 m down dip. The most abundant secondary minerals in the Butler are quartz and calcite with slightly lesser amounts of K-feldspar and epidote (Table 1). Prehnite and pumpellyite are usually much less abundant and chlorite is present in amounts < 1 %. The Butler contains a high number of veins. Usually, the veins 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). 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 which were likely feeders of hydrothermal fluids (Bornhorst et al, 2013).

At the Adventure Mine, on average the most abundant main-stage minerals filling amygdules and spaces between fragments is quartz which is closely followed by epidote and then calcite and red K-feldspar (Table 1). There are lesser amounts of prehnite, pumpellyite, and chlorite. Native copper is present in small amounts with average grades of between 0.5 to 1.25 % copper with native copper associated epidote, quartz, and calcite. Native silver and datolite are present in much lesser amounts. Least abundant are the late-stage hydrothermal minerals precipitated after native copper that occur in open space fillings as coatings on earlier formed minerals; late-stage minerals include calcite, laumontite, and adularia and in cross cutting fractures and veins.

Alteration of hydrothermal mineral is most obvious for native copper. 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 likely formed by downward-percolating groundwater when the native copper deposits were sufficiently near the surface (supergene alteration) as they are today. In addition to tenorite and cuprite, there are occasional copper carbonate minerals (such as malachite), brochantite (hydrated Cu sulfate) and atacamite (hydrated Cu chloride). These are likely to be supergene in origin. At least one mineral, gerhardtite (hydrated Cu nitrate) is the result of chemical reactions involving explosives.

At the Adventure Mine a near horizontal cross-cut adit beginning at Shaft No. 2 connects to the near horizontal Butler drift (Figure 8). To the southeast the Butler drift daylights at the Overview Entrance/Exit. To the northwest the Butler drift connects with the Shaft No. 1 cross-cut adit and with the cross-cut adit to the Ogima lode where there is a large, in place mass, of native copper (Figure 8).

The underground of the Adventure Mine is at a stable temperature of about 6^{o}C and is relatively dry and regular field shoes are usually sufficient; hard hats and lights are required and provided by Adventure Mining Company (tour operator). The field trip involves an easy walk

underground to observe the character of native copper mineralization in a horizontal adit that cross cuts the lava flows and a horizontal drift that parallels the strike of the top of the Butler lava flow which hosts a tabular native copper ore body (lode) (Figure 8). The character of native copper mineralization is readily observable in adits, drifts, and stopes (Figure 9).

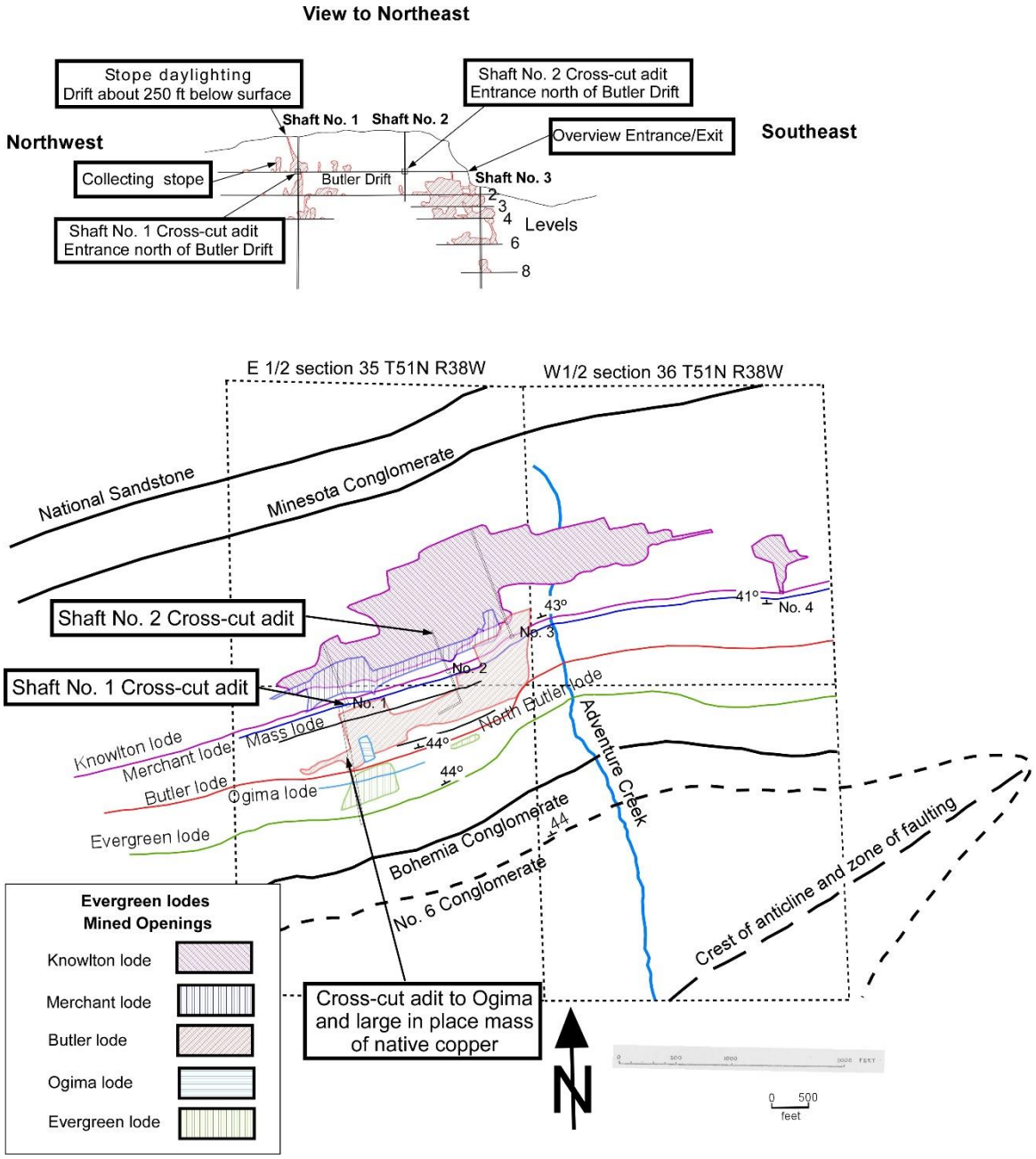


Figure 8: A. Longitudinal section of the Butler lode at the Adventure mine showing workings/openings projected to the vertical and locations for the field trip. B. Geologic map of the Adventure Mine showing workings/openings projected to the horizontal and locations for the field trip. Modified from Butler and Burbank (1929).

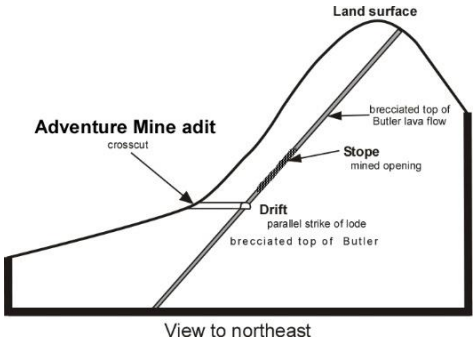


Figure 9: Cross section sketch of the topography at the Adventure Mine showing top of the Butler lava flow and the Shaft No. 1 crosscut adit.

Overview of the Human History

As the land surface of the Keweenaw Peninsula emerged above the progressively retreating glacial lake levels by *ca.* 7,000 years ago native people took an interest in native copper since its malleability facilitated making tools. At first native peoples likely found boulders of native copper deposited from the glaciers (locally termed float copper) along with gravels and sands derived from erosion of local bedrock and bedrock north of Lake Superior in Canada. These boulders of native copper have a distinctive weathered surface crust of malachite, a green copper-bearing mineral. The green color would have made the relatively infrequent float copper boulders stand out among the other brown, gray, red, and white rocks. The float copper would have also been much heavier than other rocks of the same size. The native people shaped the native copper into tools and decorations. After they depleted the float copper boulders on the surface, they needed a new source of native copper to be able to continue making and trading these items. The native peoples likely found native copper in bedrock because of the green coloration as compared to black- or red-colored host rocks and then became prehistoric miners. There are many shallow mine pits throughout the Greenland-Mass subdistrict.

Early European explorers were shown specimens of float copper by native inhabitants which created interest in the Keweenaw Peninsula. In 1841, Douglass Houghton's report to the Michigan legislature (Michigan's first state geologist) sparked the first major mining rush in North America. The first significant discovery of native copper was in 1845 at the Cliff Mine which in 1849 became the first profitable native copper mine in the Keweenaw Peninsula native copper district. There were only a few profitable native copper mines from 1845 to the early 1860s. The Minesota Mine, in the Greenland-Mass subdistrict (Figure 6) southwest of the Adventure Mine, became profitable a few years after the Cliff Mine. Many discoveries led to the opening of many mines in the early 1860s. But by 1880, the fate of most of these mines was the same as described below. Initial excitement of possible riches from mining copper was promoted by discoveries of mass copper that implied high grade ore (Figure 10C). Unfortunately, the existence of masses of copper did not necessarily indicate high grade ore. These masses represent the sampling problem termed the "nugget effect." When the deposit formed there was a clustering of copper in distinct parts of the ore body into a "nugget". If the mass of copper was missed during exploration the estimate of the grade of the ore body could be far too low and if a



Figure 10A: View of Adventure Mine No.3 shaft-rockhouse soon after the mine was permanently closed in 1917.
Photo from: Michigan Technological University Copper Country Archives Negative MS042-093-Z566B.



Figure 10B: View of Adventure Mine No.3 shaft-rockhouse on left approaching on Adventure Mine Road and looking southwest towards the shaft-rockhouse, ca 1917.
Photo from: Michigan Technological University Copper Country Archives Negative 00632.



Figure 4C: Barrel and mass native copper from the Adventure Mine ca. 1897-1898.
Photo from: Michigan Technological University Copper Country Archives.

Figure 10: Historic photos from the Adventure Mine.

mass is found the grade can be far too high. Mining decisions, such as putting in a shaft or constructing an oversized mill, made from discovery of masses of copper can be costly mistakes especially when funds are limited. Many of the mining projects were underfunded making it difficult to succeed and as a result of lack of funds the operations had to frequently close and in many cases the company merged with another company if they could convince shareholders of potential to discover a large ore body with high grade. There were also frequent shareholder assessments. Rather than paying a dividend to each shareholder from excess funds (profits) an

assessment is the opposite and required each shareholder to pay the company a fee for each share. The fate of the Adventure Mine briefly described below follows this fate and was permanently closed and abandoned by 1917.

Adventure Mine History
(modified with permission from text provided by Brandon Erickson)

In 1848 the Adventure Mining Company began exploration for native copper in the Greenland-Mass subdistrict. Exploration activities were focused on a topographic bluff where several different lodes were exposed at the surface. By 1850, the Butler lode appeared to have best potential and the first production of native copper began in 1850. Despite the initial promise the mine struggled to turn a profit as the Butler lode was very rich in copper in some areas and in other areas it was barren. By 1855, the Adventure Mining Company itself ceased mining and to survive in 1855 the company introduced a tributing system. Under tributing, miners would receive a percentage of copper profits instead of a daily wage. In 1856, during financial troubles, a water-powered stamp mill was built along nearby Adventure Creek. By the start of the Civil War in 1861, the richest known ore shoots had been mined out and in 1864 the mine was sold to new investors.

The new Adventure Copper Company explored the Butler lode on the eastern side of the bluff but also started an exploration crosscut on the northern slope. The purpose of this adit was to cut across all the lodes. Today's mine tours enter the mine using this adit (Figure 8, Shaft No. 1 crosscut.) By 1869, the adit had reached the Butler lode and the miners commenced drifting along the lode. Several years later this zone proved rich enough to warrant the sinking of a shaft from the top of the bluff, which is seen on today's tours as the "skylight stope." (Figure 8). Mining was centered around this area until an economic downturn in 1877 forced the mine to cut costs and once again only support a small handful of tribute miners.

In 1890, the tribute miners uncovered the Knowlton lode, which, unlike the others, cropped out at the base of the bluff. The deeping of shaft No. 1 started immediately and soon reached sufficient depth to begin drifting along the Knowlton lode. The Knowlton lode was very rich in stamp rock (fine sand sized copper disseminated throughout the ore body and lacking the nugget effect). The high-grade copper ore incentivized the miners to continue sinking the shaft to 200 feet deep. At these depths, work was severely hindered by the lack of more modern equipment, and without financial backing the Knowlton efforts ceased by 1893.

Adventure Mine was revived again in 1898, as the Adventure Consolidated Mining Company and listed on the Boston stock exchange. This new company was backed by 2.5 million dollars perhaps prompted by "riches" from mass copper (Figure 10C). The new investment at Adventure Mine was sufficient to build a company town, put in a railroad spur, purchase modern drills, construct and equip a state-of-the-art stamp mill, and install an electric tram line. The company first dewatered the No. 1 shaft and started sinking No. 2 shaft. The promising initial results prompted the company to start a new fully modern third No. 3 shaft (Figure 10A and 10B). This shaft is near the present-day parking lot. By 1903, the No. 1 shaft production had declined and was abandoned at a depth of 700 feet. In response, the company started a fourth shaft on the

eastern limit of their lands (Figure 8). This shaft was a failure and was abandoned after a short time. Shipments of ore from the No. 3 shaft (Figure 10A and 10B) to the stamp mill declined in tonnage and grade. The mine was forced to use diamond drilling to explore for a new ore shoot.

In 1909, they found a series of promising new lodes, but to reach them a vertical shaft would need to be sunk. This endeavor around 1910 was Adventure Mines's " last hope." The Adventure Consolidated Mining Company diverted all resources to sink a shaft to 1,500 feet and explored several different lodes along the way. Unfortunately, this exploration was a failure as the ore bodies were not large enough or with high enough grade to make them profitable. The Adventure Mine was more or less abandoned in 1910 but one last attempt was made in 1916 when copper prices increased. The No. 3 Shaft (Figure 10A and 10B) was dewatered down to 700 feet and soon after they were shipping 300 tons of concentrate per day to the smelter. This effort was short-lived and on October 27, 1917 the company ceased all mining and processing activities.

Keweenaw National Historical Park

The historical significance of the Keweenaw Peninsula native copper district can readily be deduced from the fact that 80% of the new copper for the entire United States was produced from the district in 1880. This was the peak of significance of native copper mining in the district. By 1900, the Keweenaw Peninsula produced only 25 % of the United States new copper. However, absolute copper production from the district peaked in 1916 at an annual production of 121 million kg (267 million pounds). Mining ended in the Keweenaw Peninsula native copper district in 1968. The Adventure Mine peaked in production of copper between 1902 and 1907 at a total of about 3.9 million kg (8.5 million lbs.).

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. The national park visitors center in Calumet provides an excellent overview of the historical significance of the district. Keweenaw Heritage Sites are affiliated with and support activities of the national park. The A.E. Seaman Mineral Museum and Quincy Mine are heritage sites and support the activities of the national park. The Adventure mine is also a Keweenaw Heritage Site and welcomes tourists and visitors seeking an underground mining experience.

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Field Trip 6

Southern Complex Granitoids, Gneisses, and Migmatites: New Data, Discoveries, and Perspectives

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Introduction

The Superior Province is part of the Archean Canadian Shield in North America and represents one of the oldest and most stable cratonic regions on Earth, encompassing parts of Canada and the United States, including Michigan, Wisconsin, and Minnesota. The Archean craton in Northern Michigan is divided into a Northern Complex and Southern Complex, which are separated by the Great Lakes Tectonic Zone (GLTZ) (Morey and Sims, 1976; Sims et al., 1980). The northern portion of the Southern Complex consists of classic 'dome-and-keel' structures characterized by domes of Archean basement surrounded by keels of Paleoproterozoic Marquette Range Supergroup lithologies (Annhaeusser et al., 1969). The Archean rocks include a complex assemblage of granitoids, gneisses, and migmatites intruded by numerous mafic dikes and/or sills. Our recent research on the area has revealed new aspects of the igneous and metamorphic evolution that have improved our understanding of the assembly of this large igneous-metamorphic complex through the Archean-Proterozoic transition, but numerous questions regarding the origin evolution of these rocks remain unresolved. This field excursion will include the exploration of a number of different terrains representative of the magmatic, metamorphic, and structural evolution of the Southern Complex and the overlying metasedimentary rocks; highlighting new discoveries while at the same time providing an opportunity to investigate still unresolved questions regarding the geologic evolution of the region.

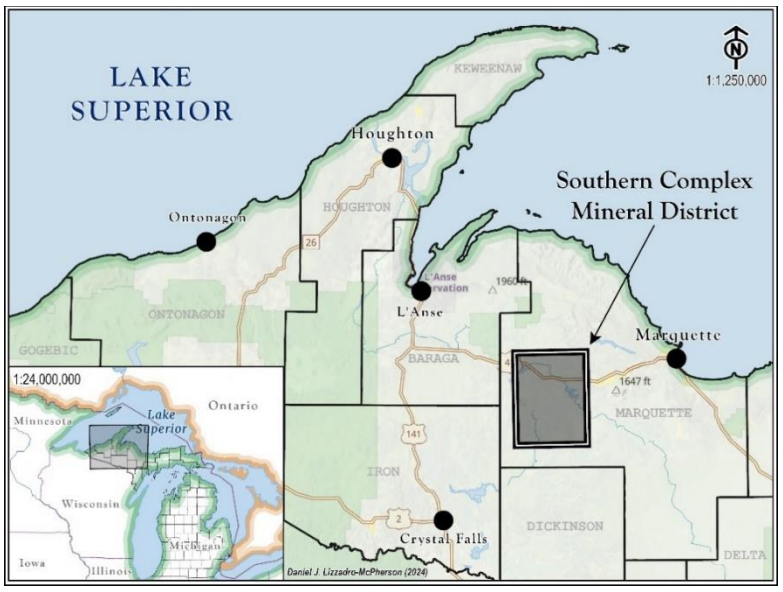


***Figure 1**. Regional map outlining the location of the Southern Complex Mineral District near Marquette, Michigan.*

Evolution of the Superior Craton

Mesoarchean

The Mesoarchean evolution of the Superior Craton in North America encompasses a critical period of crustal evolution between 3.2 to 2.8 billion years ago. During this time, significant tectonic and magmatic processes shaped the early Earth's crust and laid the foundation for the stable continental core we recognize today. In the early Mesoarchean, around 3.2 billion years ago, smaller continental fragments began to accrete and amalgamate due to tectonic processes related to subduction and associated magmatic activity (Percival et al., 2012; Thurston et al., 2008; Wyman, 2010). These proto-continents served as the building blocks for the Superior Craton (Percival et al., 2012). Intense magmatic activity occurred, leading to the generation and growth of the continental crust within the Superior Craton, as magma intruded into and solidified within the existing crust (King et al., 1998). Greenstone belts, characterized by volcanic and sedimentary rocks, also began to form during this period. These belts, such as the Abitibi and Wawa greenstone belts of Canada (Thurston, 2002) and the Ishpeming greenstone belt of the Upper Peninsula, Michigan, USA (Bornhorst and Johnson, 1993), are important features of the Superior Craton and provide insights into early Earth processes, including volcanic activity and the nature of oceanic environments. The rocks of the Superior Craton underwent significant metamorphism and deformation during the Mesoarchean and high temperatures and pressures caused by tectonic activity led to the development of foliations and other prominent structural features in the rocks.

Neoarchean

During the Neoarchean Eon, which lasted from approximately 2.8 to 2.5 billion years ago, the crust of the Superior Craton underwent further significant geological evolution. This involved the continued accretion of smaller continental blocks and terranes through tectonic processes such as subduction, collision, and magmatic activity (Mole et al., 2021). In particular, the Minnesotan orogeny occurred around 2.7 to 2.6 billion years ago. This was a period of intense tectonic activity characterized by the collision and amalgamation of various smaller continental blocks and island arcs, leading to the formation of a larger continental mass (Schmitz et al., 2018). The collisional process resulted in the growth of the Superior Province and the formation of the granite-greenstone terranes that comprise much of the region. The amalgamation of these crustal fragments contributed to the expansion and stabilization of the Superior Craton. Neoarchean rocks include extensive granitic intrusions, which formed through the partial melting of existing crustal rocks or through the emplacement of mantle-derived magmas (Mole et al., 2021). These granitic intrusions contributed to the growth of the continental crust and are often associated with mineralization and hydrothermal activity (Mole et al., 2021). Greenstone belts, characterized by volcanic and sedimentary rocks, continued to develop during the Neoarchean within the Superior Craton (Polat et al, 1998; Polat and Kerrich, 2000). These belts represent ancient oceanic crust and island arc environments, and they are interspersed with granitic intrusions. Metamorphic processes affected both the greenstone belts and the granitic intrusions within the craton. Hydrothermal activity continued to play a significant role in the formation of mineral deposits within the Superior Craton during the Neoarchean. Ore deposits such as gold, iron, and copper formed in association with granitic intrusions, greenstone belts, and hydrothermal alteration zones, contributing to the economic significance of the region (Mole et al., 2022).

Southern Complex, Marquette District: Compeau Creek and Bell Creek batholith

The southern complex is located south of the Marquette synclinorium and is dominated by the Archean Bell Creek batholith, which consists primarily of coarse-grained megacrystic, high-K igneous rocks with minor amounts of mafic gneiss and metasedimentary rock layers typically found concordant with the foliation of the gneiss. Bell Creek includes trondhjemite-tonalite-granodiorite (TTG), granites, gneisses, and migmatites. Migmatite comprises only a small portion of the complex, distributed at irregular intervals throughout the region and is assigned to a unit referred to as Compeau Creek.

The age and origin of the Southern complex of the Marquette region has been debated for decades. It was originally thought to be genetically related to similar lithologies found in the nearby Northern Complex (Cannon and Simmons, 1973; Van Schmus and Woolsey, 1975). However, the Southern Complex is separated from the Northern complex by the Great Lakes Tectonic Zone (GLTZ), which is a continental scale suture/fault zone (Morey and Sims, 1976; Sims et al., 1991). The only age information available before our study of the Bell Creek batholith was obtained by Tinkham (1997) from a single zircon with a U-Pb age of ~2.61 Ga. This period marks the onset of the Archean-Proterozoic transition, which is associated with a shift from the production of dominantly mantle-derived magmas that differentiated to form new continental crust to the early stages of significant recycling of crustal material (Taylor & McLennan, 1995; Valley, 2005). This is a crucial period in Earth's history due to a decrease in global heat flow, and potentially the onset of 'modern style' subduction (Brown et al., 2020) and an increased number of sedimentary environments (Taylor and McLennan, 1995).

The origin of trondjemite-tonalite-granodiorite (TTG) and high-K granites during the Archean Eon, including those produced in the Bell Creek batholith, is closely linked to the evolution of continental crust. Several models have been proposed to explain the formation of TTGs during the Archean: 1) *Partial melting of pre-existing continental crust.* During the Archean, the Earth's crust was thicker and more mafic (rich in magnesium and iron) compared to modern crust (Tang et al., 2016). As a result, when portions of this crust were subjected to high temperatures and pressures, particularly in subduction zones or during collisional events, they could undergo partial melting to produce high potassium granites. 2) *Partial melting of the mantle.* In this scenario, mantle-derived melts ascend through the crust, assimilating and interacting with continental crust along the way. These interactions can lead to the enrichment of potassium and other incompatible elements in the melts, ultimately resulting in the formation of high potassium granites. 3) *Derivation from hybrid sources.* It is also possible that the formation of high potassium granites during the Archean involved a combination of crustal and mantle processes. This hybrid model suggests that both the crust and mantle contributed to the source materials for the granites, with melting and mixing occurring at various depths within the Earth's crust.

New petrogenetic insights on the origin of Bell Creek batholith

Our new bulk-rock major and trace element data and U-Pb zircon dates (including oxygen and Lu-Hf isotopes) provide insight into several aspects of the origin of these rocks. First, extensive U-Pb zircon dating of Bell Creek and Compeau Creek rocks from several recent Michigan Tech geology graduate student studies (Table 1; Petryk, 2019 and Barth, 2023) indicates that the bulk of the magmas were emplaced during a single tectonic event between ~2.4 to 2.6 Ga and is attributed to the collision of the Paleoarchean Minnesota River Valley Terrane (MRVT). Second, bulk-rock

major and trace element compositions of the Compeau Creek migmatite/gneiss and Bell Creek granitoids/gneisses are consistent with formation in a continental arc type tectonic environment (Figure 2).

Table 1. *Age summary for Bell Creek granitoids (data from Petryk, 2019)*

Bell Creek granitoid subgroups	Sample	U-Pb crystallization age (Ga)	Inherited grains (Ga)	Metamorphic grains (Ga)
Normal'	CLG-14B	2.42 ±0.042	2.8-3.6	2.3
	CCG-12A	2.43 ±0.160	2.7	2.0
	BCG-4A	2.51 ±0.021	2.8-3.9 *(4.2)*	
	BCG-1A	2.58 ±0.056	2.7-2.8	2.3
	BCG-8B	2.54 ±0.040	3.1	
Foliated	CCG-9D	2.59 ±0.028	3.3-3.5	
	BCG-7C	2.61 ±0.047	2.7-3.3	2.3, 2.4
	CCG-9E	2.56 ±0.038	2.7-2.8	2.3, 2.4
Fine-grained	CCG-6A	2.53 ±0.070	-	2.3, 2.4
Clotted	BCG-8A	2.55 ±0.017	2.7-2.9	2.2, 2.3
	CCG-1A	2.55 ±0.033	2.7-3.2	1.8, 2.3

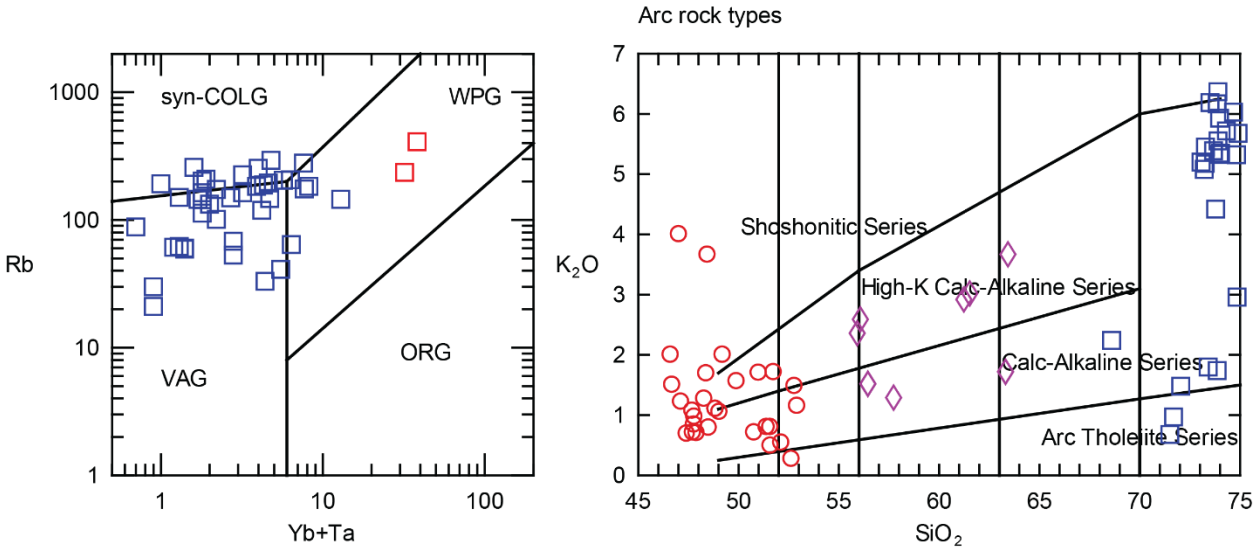


Figure 2. *Left: Tectonic discrimination diagram showing a volcanic arc to syn-collision tectonic setting for Bell Creek granitoids and gneissic rocks (blue) and two younger alkaline granites consistent that formed within-plate following the emplacement of the Bell Creek batholith. Right: Arc rock discrimination diagram showing the high-K nature of most of the Bell Creek and Compeau Creek rocks. Note that the mafic rocks plotted here represent dikes/sills of varying and relatively unknown age.*

Third, the oxygen isotopic composition of zircon from the granitoids is dominated by mantle-like values (5.3±0.3) indicating additional of a significant amount of juvenile magma to the crust; however, the hafnium isotopic compositions of the same zircons are heterogeneous and have model ages that indicate the involvement of older Mesoarchean crustal lithologies at the source and during emplacement in the mid- to upper-crust. Our current interpretation is that the granitoids, gneisses, and migmatites were formed through a hybrid process, initially by partial melting of an isotopically heterogeneous Mesoarchean lower crust followed by assimilation of mid- to upper-crustal lithologies. Continental arc subduction produces TTGs when high heat flow facilitates partial melting of lower crust and potentially later as the magma reaches the level of final emplacement in the mid- to upper-crust. Interestingly, the high oxygen isotopes ($\delta^{18}O > 6‰$) and εHf isotopes, with a range between -2 and -24, together indicate a potentially greater role for older Mesoarchean crust in the formation of the Southern Complex magmas than what has been found in the Canadian portion of the Superior craton. Therefore, it appears as if the evolution of the southernmost region of the Superior Craton involved a much greater contribution from crustal lithologies than what has currently been found in the well-studied Canadian segment of the Superior Craton (Mole et al., 2019).

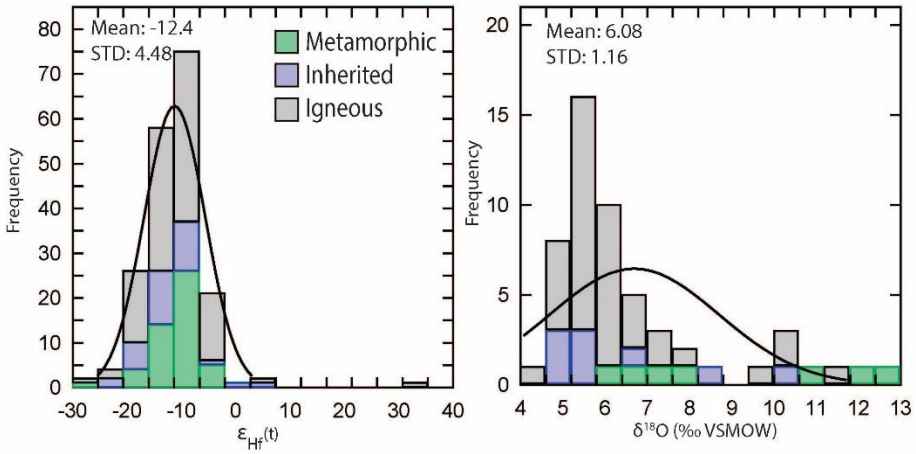


Figure 3. *Hf-O isotopic compositions of zircon from Bell Creek TTGs*

Mid- to upper-crustal level mixing and assimilation

The Archean granitoids of the Bell Creek batholith also display field evidence of assimilation and mafic-felsic magma mixing at the level of emplacement. Small (up to a cm or two) xenoliths that appear as clots throughout the granitoids in the area consist of biotite, chlorite, garnet and quartz indicative of assimilation of pelitic crust. However, basement crustal lithologies of this type are apparently not well exposed. In our investigation we have identified several outcrops that have quartzite and schist in contact with, or within, gneissic or granitoid host rocks, but the ages have yet to be determined (in progress). These metasedimentary rocks outcrop within the igneous intrusions and are considered to be the best possible candidates for remnants of Archean supracrustal material that was assimilated into the felsic magmas. In addition, our new O and Hf isotope data from single, inherited zircons indicate incorporation of Mesoarchean age crustal lithologies (Table 1 and Figure 3). Major and trace element data reveal a slightly peraluminous

character (Figure 4), high-K (Figure 2), and enrichment of incompatible trace elements (Figure 2) and are best explained as reflecting the assimilation of metasedimentary crustal lithologies.

The mafic intrusions also show evidence of felsic 'blebs' and complex interactions with the host granitoids throughout the region and are interpreted to reflect magma mixing, which would indicate that at least some of them are sills rather than dikes. However, later generations of mafic intrusions clearly crosscut the dominant foliation, have sheared boundaries, and/or sharp contacts with the host rock and represent at least four to five distinct episodes of magmatism related to younger events unrelated to the granitoids.

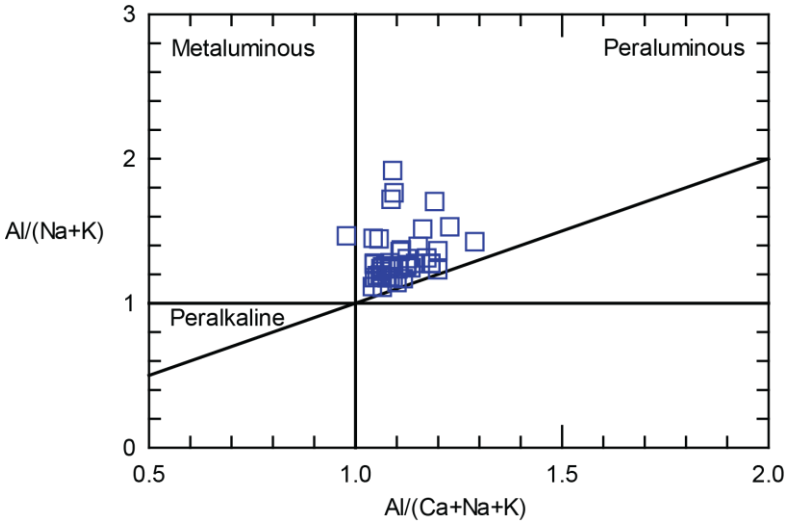


Figure 4. *Shand's index for peraluminosity of Compeau Creek and Bell Creek granitoid, gneissic and migmatitic rocks.*

Paleoproterozoic

During the Paleoproterozoic Eon (roughly 2.5 to 1.6 billion years ago), the Superior Craton underwent significant geological evolution, marked by a series of tectonic, magmatic, and metamorphic events. The early Paleoproterozoic was dominated by the development of rift basins during the breakup of the Superia supercraton during rifting that began ~2.1 Ga, which separated the Wyoming Province from the Superior Province (Drenth et al., 2021). These rift basins accumulated thick sequences of sedimentary rocks, including sandstones, shales, and iron formations that are today known in the Lake Superior region as the Marquette Range Supergroup and Huronian Supergroup (Ojakangas et al., 2001). The Penokean Orogeny, which occurred around 1.85 to 1.75 billion years ago, resulted from the collision of the Superior Craton with other continental blocks that contributed to the growth of Laurentia, leading to crustal thickening, mountain-building, and metamorphism (Schulz and Cannon, 2007). Hydrothermal activity during the Paleoproterozoic played a significant role in the formation of mineral deposits that include iron-rich sedimentary deposits, such as banded iron formations (BIFs), which were deposited during this time, along with important mineral deposits such as iron, copper, and gold (DeMatties, 2022). Following the main tectonic events of the Paleoproterozoic, the Superior Craton experienced episodes of post-orogenic magmatism, marked by the emplacement of large igneous provinces and granitic intrusions.

Field Trip Objectives

This field trip is designed to provide a geologic overview of the formation of the Neoarchean Bell Creek batholith including associated metasedimentary rocks and the Paleoproterozoic metasedimentary rocks that filled the deep basins.

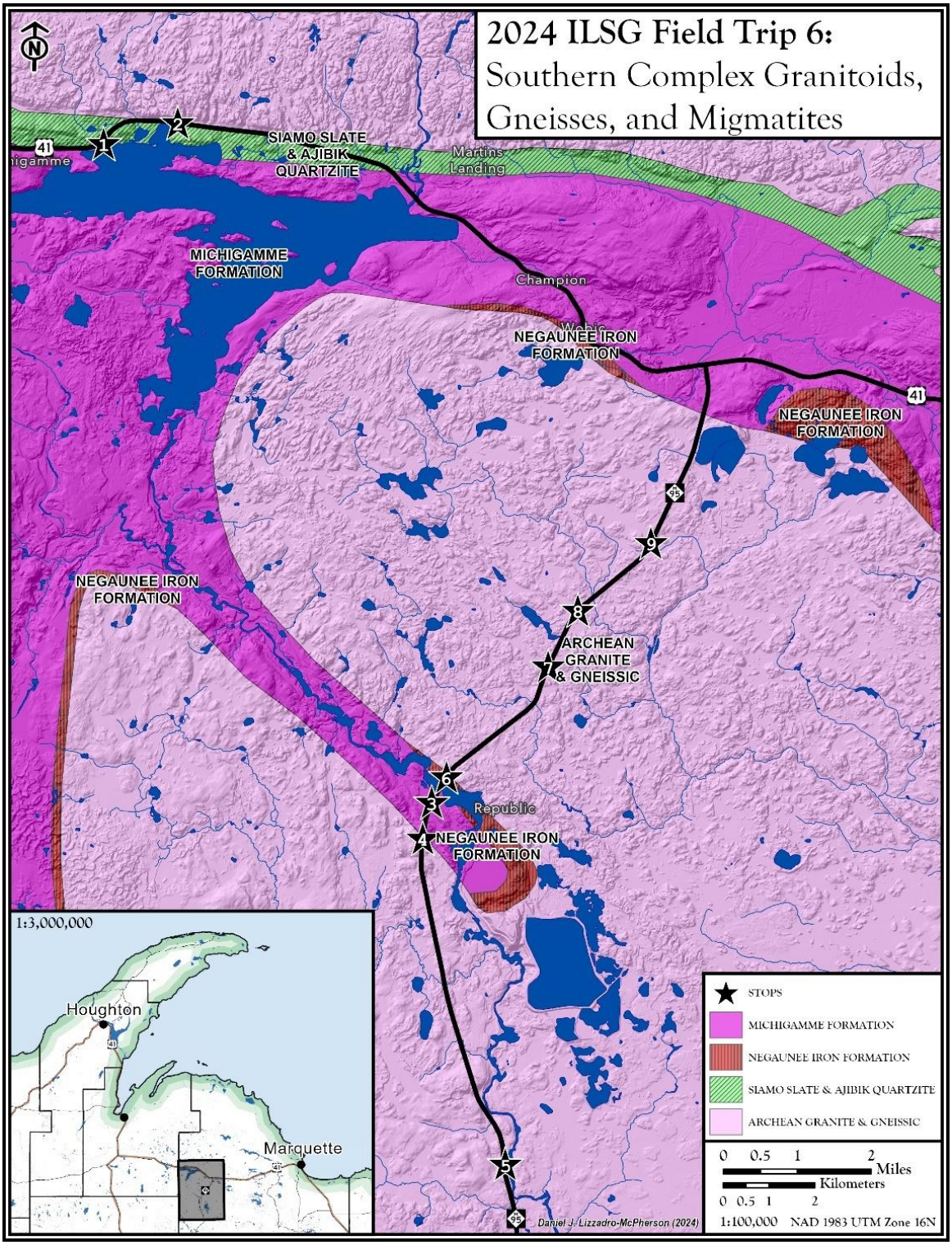


Figure 5. *Generalized geologic map of the Southern Complex Mineral District.*

Stop 1: Paleoproterozoic Negaunee Iron Formation (FR/Xn) and mafic dike

Directions: Leaving Michigan Technological University drive south along US-41 ~63 miles to the Michigamme Roadside Park.
N 46° 32’ 20”
W88° 5’ 31”
The outcrop is prominently exposed along the northern side of US-41 across from the Michigamme Roadside Park, which overlooks Lake Michigamme to the south.

Paleoproterozoic Negaunee Iron Formation that is weakly magnetic containing hematite-goethite with banded chert as fine laminations. The formation is part of a westward plunging syncline conformably overlying Siamo Slate or Ajibik quartzite with a gradational contact and lies unconformably over the Archean basement complex (Gair and Thaden, 1968). This unit includes sideritic slates, grunerite-magnetite-schists; ferruginous slates, ferruginous cherts and jaspilite (Van Hise and Bayley, 1895; Cannon and Gair, 1970). Here, the iron formation is dipping ~65° to the SW and is in sharp contact with the adjacent massive, medium-grained to porphyritic amphibolite dike. A younger, near vertical dike can be observed on the east side of the outcrop in sharp contact with the older amphibolite dike/sill.

Stop 2: Paleoproterozoic Ajibik quartzite (Xa)

Directions: Heading east along US-41 ~0.5 miles exposure of Ajibik quartzite outcrops along the northern edge of the road.
N 46° 32’ 35”
W88° 4’ 16”

The Ajibik consists of basal conglomerates, slates, and graywackes that grade into the overlying quartzite. At this location, the small exposure of the Ajibik is massive, thick bedded white to buff orthoquartzite. Some relic cross-bedding may be present, but clear evidence of ripple marks or other features reflecting the original depositional environment are not present. A mafic dike cross-cuts the quartzite roughly NW-SE and is highly sheared along a sharp contact. Bedding is difficult to identify, but it appears to be dipping ~60° to the SW.

Stop 3: Paleoproterozoic Michigamme formation (Xms)

Directions: Head east on US-41 S toward Orange Rd. 8.4 miles. Turn right onto M-95 S, and drive ~7.3 miles, crossing the Michigamme River basin to destination.
N 46° 24’ 37”
W87° 59’ 45”

This exposure is of the lower slate member, which includes laminated iron-rich rock consisting of biotite-garnet-cummingtonite-quartz schist with thin beds of quartzite. Minor fold axes are prominent here as chevron folds plunging to the NW and reflect the regional fold axis orientation. Boudinage, which form when single, competent layers are stretched into separate pieces through plastic and/or brittle deformation mechanisms and reflect the presence of minor quartzite layers or lenses within the schist.

Stop 4: Archean Bell Creek batholith granitoid (Wbcf)

Directions: Head southwest on M-95 N ~0.5 miles to destination.
N 46° 24’ 13”
W87° 59’ 52”

The most common form of Bell Creek batholith is exposed here as a medium to coarse-grained granitoid that is locally porphyritic with sparse megacrysts of alkali-feldspar (up to several cms). Minerals include alkali-feldspar, oligoclase, biotite, oxides with apatite and zircon as common accessory phases. The abundance of mafic xenoliths (clots) (Figure 6) within the granitoids is correlated with the degree of peraluminosity of the host rock. Some minor pegmatitic veins are also present. Small mafic injections are in sharp contact with the granitoid indicating emplacement that post-dates the main magmatic episode. There is some weak alignment of alkali-feldspar megacrysts that can be best observed on the top of the outcrop.



Figure 6. *Left: Mafic clot containing Fe-chlorite or biotite, K-feldspar, and garnet (almandine). Garnet-biotite geothermometry yields an equilibration temperature range between 450°C and 550° C. Right: Example of mafic xenoliths clustered within granitoid.*

Stop 5: Undivided granitic rocks (Wgu)

Directions: Head south on M-95 toward Co Rd 601 ~4.8 miles to destination. Outcrop is located just south of the city park along the Michigamme river.
N 46° 20’ 13”
W087° 58’ 22”

A metasedimentary sequence is exposed here, which doesn’t appear on the regional geologic map. Blonde and gray/black feldspathic quartzite forms alternating beds with gradational contacts dipping to the north (Figure 7). The blonde quartzite is dominated by quartz, with minor amounts of feldspar and sparse garnet (1-2mm). The gray/black quartzite has similar proportions of quartz and feldspar to the blonde quartzite but includes muscovite, biotite and oxides nearing the abundance of feldspar. Complex interaction with mafic dikes/sills also appear in this outcrop with minor ptygmatic folding and mafic enclaves throughout. Vertical dike contacts are sharp, whereas an apparently older generation of mafic intrusions have more complex contact margins with the host rock. Near vertical mafic dikes on both ends of the outcrop have sharp contacts with the host rock penetrating locally as splays or fingers.



Figure 7. *Metasedimentary sequence dominated by blonde quartzite and a grey to black quartzite.*

Lunch Stop: Leif Erickson Roadside Park along the Michigamme River

Stop 6: Archean Bell Creek batholith and Compeau Creek gneiss/migmatite (Wbcf/Wccg)

Directions: Drive north along M-95 towards Welsh's Rd. ~5.7 miles to destination. Outcrop exposed prominently on the east and west sides of the road.
N 46° 24' 54"
W87° 59' 31"

The contact between Compeau Creek to the south and Bell Creek batholith to the north is clearly exposed at this outcrop. A zircon U-Pb date of 2426±160 Ma was obtained for the Bell Creek granitoid at this location. The granitoid is fine- to medium-grained, typically equigranular and has an apparently gradational contact with the adjacent migmatite/gneiss. Migmatite/gneiss with ptygmatic folding is best exposed in the outcrop on the west side of the road, but the highly deformed continuation of this rock type is prominently exposed with mafic intrusions on the east side of the road.

Stop 7: Archean Bell Creek batholith (Wbcg)

Directions: Drive north along M-95 ~2.1 miles to destination.
N 46° 26' 16"
W87° 57' 48"

Exposure of Bell Creek megacrystic granitoid similar to Stop 4. Minerals include alkali-feldspar, oligoclase, biotite, oxides and apatite with apatite and zircon as common accessory phases. A strong foliation is not apparent here, but the mafic xenoliths (clots) are well represented as mm- to cm-sized dark spots typically clustered and randomly oriented. Pegmatitic veins are common and vary in width up to tens of centimeters.

Stop 8: Archean Migmatite (Wbsc); Compeau Creek?

Directions: Drive north along M-95 ~0.8 miles to destination.
N 46° 26' 53.7"
W087° 57' 20.0"

This exposure is likely the Compeau Creek quartofeldspathic migmatite/gneiss dipping to the south. The highly deformed migmatite/gneiss includes numerous mafic inclusions and enclaves (Figure 8). However, it is unclear how much of the mafic-felsic segregations are representative of the initial magma mixing followed by deformation or melanosome-leucosome complementary rocks derived by melting. Ptygmatic folding is prominently displayed and likely represents felsic segregations that have buckled during deformation. A zircon U-Pb date of 2525±70 Ma was obtained for the gneiss, which overlaps in time with the dates obtained for other Bell Creek granitoids along the M-95 corridor.



Figure 8. *Near vertical mafic dike cross-cutting migmatite with chaotic mixture of mafic-felsic components.*

Stop 9: Archean Compeau Creek migmatite/gneiss (Wccg)

Directions: Drive north along M-95 ~1.4 miles to destination.
N46° 27'42"
W87° 56' 05"

This outcrop is mapped as Compeau Creek migmatite/gneiss that is presumed to be older than the adjacent Bell Creek granitoid/gneiss. A zircon U-Pb date for the migmatite is slightly older (2633±46 Ma) than the pink, clotted granite, which has been dated to 2550±33 Ma. Inherited zircon grains range in age from 2717 to 3200 Ma. The Compeau Creek gneiss/migmatite on the northern end of this outcrop appears to include a sliver of what might be Archean quartzite (Figure 9); note that the Paleoproterozoic Goodrich quartzite is mapped directly to the north and would, therefore, be in direct contact with the migmatite/gneiss. There are numerous mafic dikes/sills that manifest as either single generation dikes that clearly cross-cut the existing foliation or more complex dismembered dikes (possibly sills) associated with mixing at the time of felsic magma emplacement (Figure 10). Inclusions of felsic material (either bulk rock or individual feldspar crystals) can be found in some of the composite dikes/sills indicative of magma mixing that may have occurred contemporaneously between the mafic magma and a liquid dominant felsic host magma. The foliation is steeply south-dipping (~70°).



Figure 9. *Band of quartzite concordant with the foliation of gneiss/migmatite fabric.*

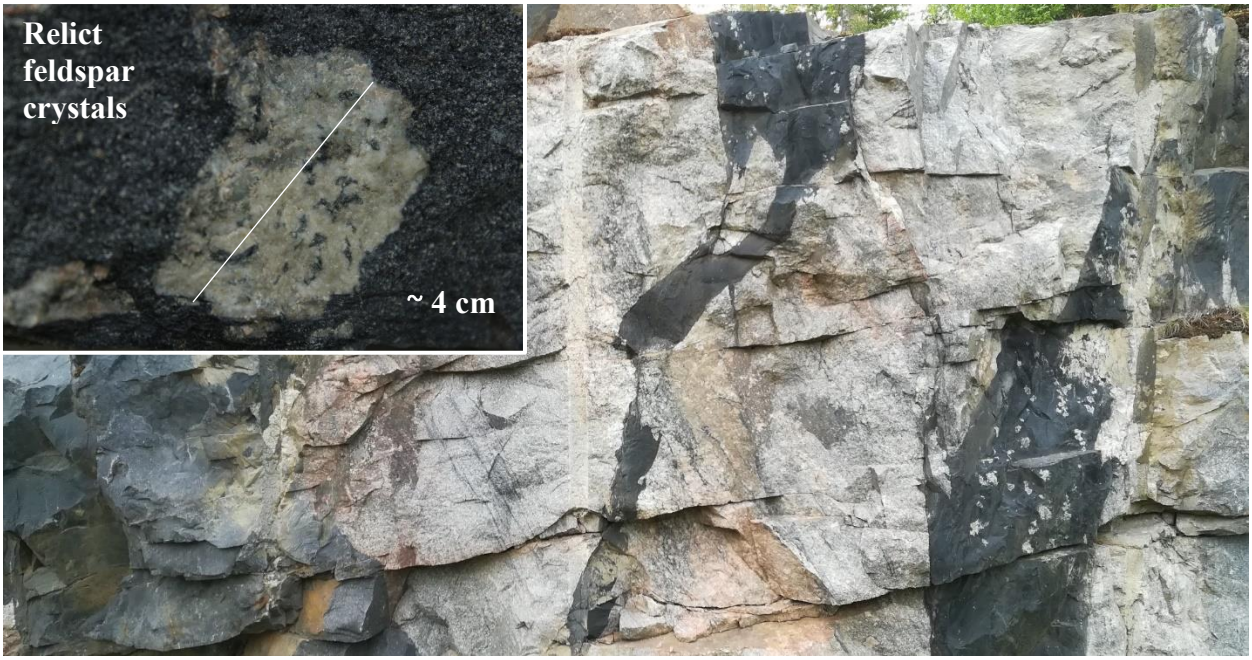


Figure 10. *Mafic intrusion with felsic xenocrysts/xenoliths interpreted to have likely formed through mixing with the host magma during emplacement. This particular intrusion has a Nd model age of ~2.57 Ga, which is within error of the age obtained for the Bell Creek gneiss at this location.*

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Field Trip 7

Landslides on the Ontonagon River at Military Hill

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Figure 1: 2003 landslide east of US-45 on the East Branch of the Ontonagon River.

Introduction

An impressive sight, at least to geologists and engineers, are the landslides along US-45 as it descends into the Ontonagon River Valley at Military Hill. US-45 follows the trail used by Indigenous peoples such as the Menominee, Dakota and Anishinaabe (Ojibwe/Chippewa) tribes, and later fur traders making their way to and from to the mouth of the Ontonagon River at Lake Superior. In 1844, U.S. Secretary of War William Wilkins, proposed a road between Fort Howard in Green Bay, WI to Fort Wilkins just north of Copper Harbor, MI to create a supply route to the Army posts and to improve access to frontier communities such as Copper Harbor and other neighboring mine sites. Congress, however, did not fund the road. In 1862, with British troops in nearby Ontario and the prospect of Great Britian entering the war on the side of the confederacy, the road became a national security issue. According to a Michigan historian, Le Roy Barnett,

> "*More than half the copper used in the United States came from mines along the proposed frontier passage. As a report of the U.S. Senate Committee on Military Affairs noted, "In case of hostilities with Great Britain, and a descent upon this important portion of our lake coast, there would be no means of affording succor without a road [to the region], and these valuable deposits of copper and other ores might be lost to us. Senator Jacob Howard pressed the point, explaining that if the Soo Canal were "seized and closed ... there would be no means of getting into the [Keweenaw] country or out of the country, either with troops or munitions of war or without them, except by means of some such road as this." (The Free Library, 2024)*

On March 3, 1863, Abraham Lincoln signed an Act of Congress authorizing construction of the military road. Road construction was started in 1863 but was not completed until 1872. Following construction, the road was used primarily for timber and other commercial interests until the railroad network in the western Upper Peninsula became operational a few years later. Again, according to LeRoy Barnett,

> *"Despite the years of work put into the route, it never became the avenue of settlement and commerce its supporters anticipated. As legislators in the early 1860s had feared, railroads--built shortly after the road's completion--soon siphoned traffic. Furthermore, parts of the road were poorly maintained. "There was no surfacing on most of the road," wrote Upper Peninsula author/historian Knox Jamison of a trip in 1915, "and if it had rained recently, you had better not try it." Getting up and down the Military Hill--a massive body of slippery, wet clay through which the Ontonagon River had cut a 300-foot gorge--took him at least two hours*."

In 1923, the Michigan Highway Department added Military Road into the state's transportation network as a section of State Highway M-26 but did not appreciably improve the road, leaving it with a gravel surface and in general following the route's topography. In 1957, the Michigan State Highway Department added this section of M-26 to the federal highway system as US-45.

Becoming a federal highway requires the road to meet both state and federal road design standards, such as the highway having a maximum grade of less than 8%. This required significant excavation of the glacial clay slopes and most likely the start of the Military Hill landslides. Construction was completed in 1959 with a concrete pavement and a grade meeting state and federal specifications.

While the Military Hill clay soils are known for their difficult road construction issues, the highway department most likely did not fully appreciate the difficulty with the excavation and long-term behavior of the slopes. An investigation, either upstream or downstream of the US-45 river crossing, for example, would have shown many old and new landslides not to mention the red silt laden river itself that never changes throughout the seasons. The landslides are in the glacial lacustrine and till sediments in the Ontonagon Basin formed during the area's deglaciation. One of the more recent landslides near the river crossing, which occurred in 2003, is shown in Figure 1, is just upstream of US-45 and one of the sites we will visit. Figure 2 shows the US-45-Ontonagon River crossing and the location of the landslide north of the confluence of the East and Middle Branches of the Ontonagon River. The steep slopes and the meandering of the river are evident. Figure 2 also shows the stops that we will be making at Military Hill.

The clays in the Military Hill area were deposited in the final stages of the Late Wisconsin glacial period, estimated to be at its maximum at about 26,000 BP with deglaciation ending about 9,500 years BP when the Superior Lobe of the Laurentide Ice Sheet retreated into the Superior Basin for the last time (Attig et al., 1985). Fortunately, the northern Great Lakes region is known as the "*type area*" for the stratigraphic subdivisions of the late-glacial period of North America so the area has been investigated in some detail (Evenson et al., 1976). The first and the most detailed investigation of the glacial sediments in the Ontonagon area was conducted between 1905 and 1919 by Frank Leverett who identified the area's main glacial features including Glacial Lake Ontonagon and Duluth (Leverett, 1928). Later, Hatch (1965) investigated the postglacial drainage evolution and stream geometry in the Ontonagon area while detailed mapping was conducted by Peterson (1985 and 1986).

The area's prominent feature is the Copper Range, a topographic ridge formed by the Portage Lake Volcanics. The Copper Range was influential in the formation of the Ontonagon River Basin shown in Figure 3 and Figure 4. As shown in these figures, the Ontonagon River System converges at a gap along the Copper Range. According to Leverett (1928), the Copper Range formed an ice-margin boundary during the last couple of glacial advances resulting in the formation of a series of proglacial lakes shown in Figure 5. The proglacial lakes Ontonagon, Ashland, Brule and Duluth formed along this ice-margin with drainage flowing westward to the St. Croix River. As isostatic rebound occurred and the Superior Lobe retreated into the Superior Basin, the drainage merged through a gap in the Copper Range. Leverett notes that it was probable that Lake Goebic was in a pre-glacial valley that extended northward to Lake Superior but was prevented from draining into Lake Superior by a moraine that filled the gap. Thus, the West and South Branch of the Ontonagon River flowed eastward along the Copper Range to where it drained into the present Ontonagon River.

Another distinct feature of the area is a series of parallel rivers that formed between Copper Range and Lake Superior as shown in Figure 6 and 7. Hatch (1965) investigated the drainage system indicating that "*it was strongly grooved by glacial flutings parallel to the direction of ice motion. In places the grooves are buried by lacustrine sediments of glacial Lake Duluth*." It is possible that the Ontonagon River, as it began flowing through the Copper Range gap at Military Hill, followed one of the parallel drainages. Due to a greater volume of water, however, it has significantly cut through the Lake Ontonagon and Duluth sediments as shown in Figure 6.

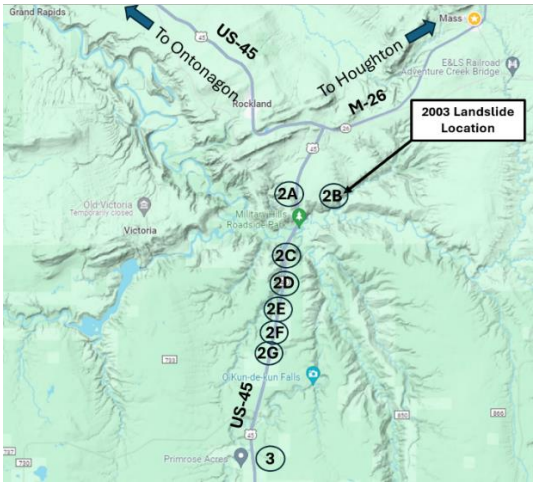


Figure 2: Military Hill along US-45 showing Field Trip Stops (Google Maps, 2024).



Figure 3: Ontonagon River Basin location (USACE, 2010).

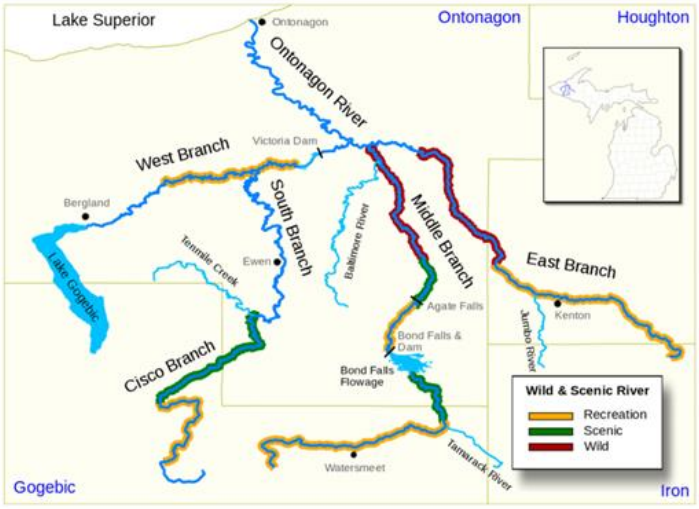


Figure 4: Ontonagon drainage basin showing river recreation, science, and wild designations (Ontonagon, 2023).

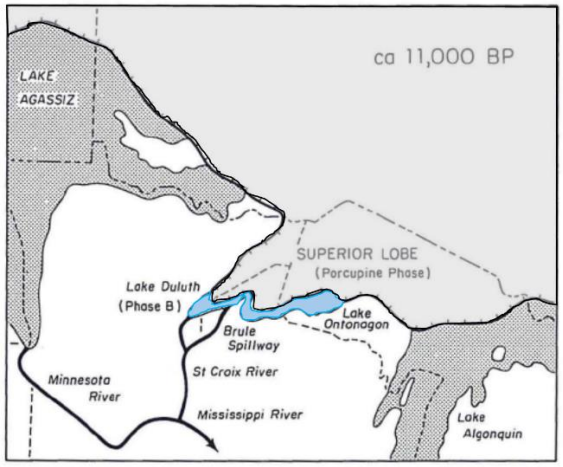


Figure 5: Development of proglacial lakes along the Superior Lob (Farran and Drexler, 1985).

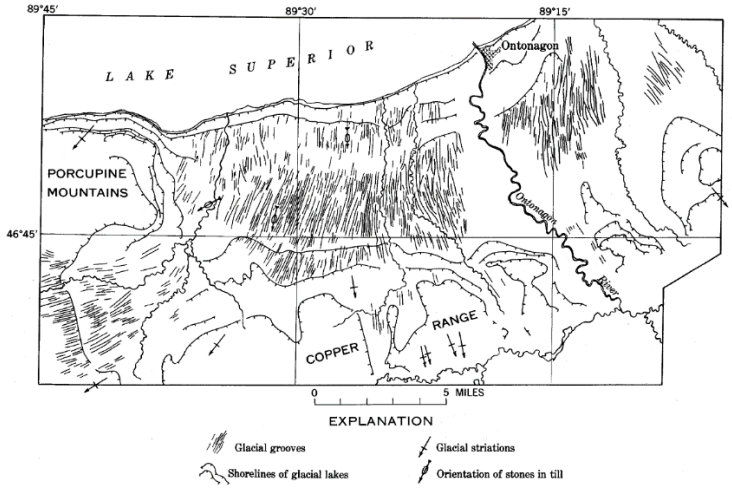


Figure 6: Parallel drainage system on the Ontonagon Plain by glacial grooving (Hatch, 1965).



Figure 7: Aerial photo of the parallel river systems on the Ontonagon till plain.

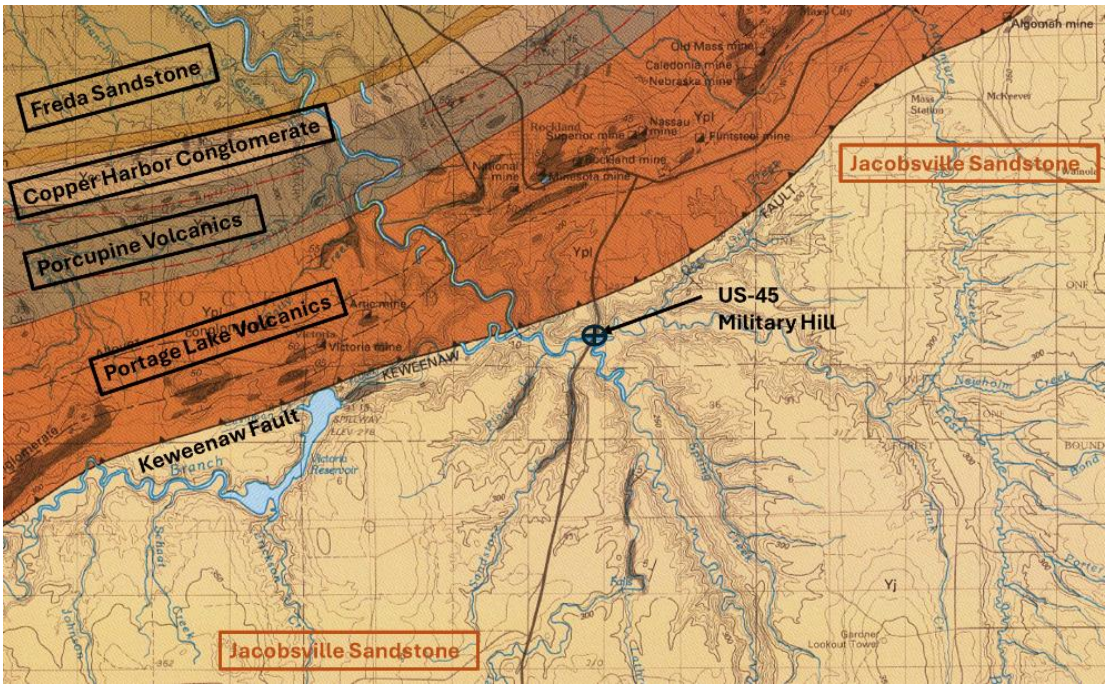


Figure 8: Bedrock geology of the Military Hill area (Cannon et al., 1995).

Background Information

Bedrock Geology

The landslides at Military Hill are located over the Jacobsville Sandstone, just south of the Keweenaw Fault, as shown in Figure 8.

Ontonagon River Watershed Overview

The Ontonagon River Watershed, at 1,348 sq. miles, is the second largest watershed entering Lake Superior behind the St. Louis watershed west of Duluth, MN, at 3,584 sq. miles. What makes the Ontonagon River watershed distinctive, however, is the large amount of red sediments that continually flows into Lake Superior. In 1987, the National Geography Magazine, ran an article on "The Great Lakes Trouble Waters" showing an aerial view of the sediment discharge from the Ontonagon River into Lake Superior with the caption, "*Each year millions of tons of sediment – like this red clay silt spilling into Lake Superior from Michigan's Ontonagon River – enters the lakes from tributary stream. Many contaminated with agricultural chemicals and industrial wastes (National Geographic, 1987).*" The cover and photo of the Ontonagon River are shown in Figure 9. In the early 2000s, the US Corps of Engineers (USACE, 2010) conducted a 516e sediment study on the Ontonagon River to determine the source of this large quantity of sediments to estimate future dredging requirements. After its investigation, the USACE reported:

> "*The Ontonagon River watershed is primarily undeveloped and consists of forested land uses with little urbanization or agriculture. Much of the watershed experiences significant erosion of the incised valley walls due to the highly erodible soils associated with the lacustrine geology of the lower reaches. Based on the comparison of historic and present-day river morphology it is concluded that the valley walls and river banks have been a large contributor of sediment long before logging or other anthropogenic disturbances were present in the watershed. Therefore, the primary contributor of sediment yield in the watershed is natural processes, rather than anthropogenic alterations. Based on this conclusion, a quantification of geologic time scale sediment yields was conducted.*"

The report went on the state:

> *There is evidence that the deep valleys and steep walls that exist today are due to natural sediment transport processes and landscape scale geomorphic evolution that have carved out the valleys and deepened the Ontonagon River and its tributaries. Frequent examples of mass wasting of the valley walls exist throughout the downstream reaches of the Ontonagon River and a flat terrace with steep valley walls adjacent to a meandering channel with a narrow beltwidth are typical in the lacustrine areas of the watershed…..Moreover, an average sediment yield of 2.4 million tons per year is more than an order of magnitude greater than current sediment yields of similar watershed sizes in the Great Lakes."*

The USACE report did not investigate which branch of the river produced the greatest amount of

sediments. Based on a study by Weidner et al., (2019), however, it would appear the East and Middle Branches produce the most sediment based on the frequency of observable landslides. This would be constitent with Peterson's 1985 mapping showing that the East and Middle Branches are mostly located in the Glacial Lake Ontonagon sediments as seen in Figure 10, which is a portion of Peterson's glacial geology map. The length of the East and Middle Branches of the Ontonagon River can also be seen in the soils map from the USDA Natural Resources Conservation Service landform map shown in Figure 11 where the East and Middle Branch extend to the southern end of the lake sediments.



Figure 9: July 1987 issue of the National Geographic Magazine (1987).

Review of Landslide Activity on the Ontonagon River

Travelers along US-45 can view a landslide just to the west of the US-45 – Ontonagon River crossing. A 1,000-foot walk upstream on the East branch of the Ontonagon River will provide the opportunity to inspect a large-scale landslide that occurred in 2003. If you continue to walk upstream from the 2003 landslide, about every bend in the river will show landslide activity, either current or past as shown in Figure 12.

Weidner et al., (2019) studied the landslide activity in the Military Hill area (Figure 13a) for the development of a landslide susceptibility map based on riverbank erosion-triggering. Additional landslide studies were conducted by Koons (1965), Dyl (1979), and Smith (2012), all master theses or reports at Michigan Tech. The Weidner et al., study utilized aerial and satellite imagery from the United States Geological Survey (USGS) EarthExplorer web tool for the years 1992 through 2016 identifying 21 landslides. The landslide locations in the study area are shown in Figure 13(a).

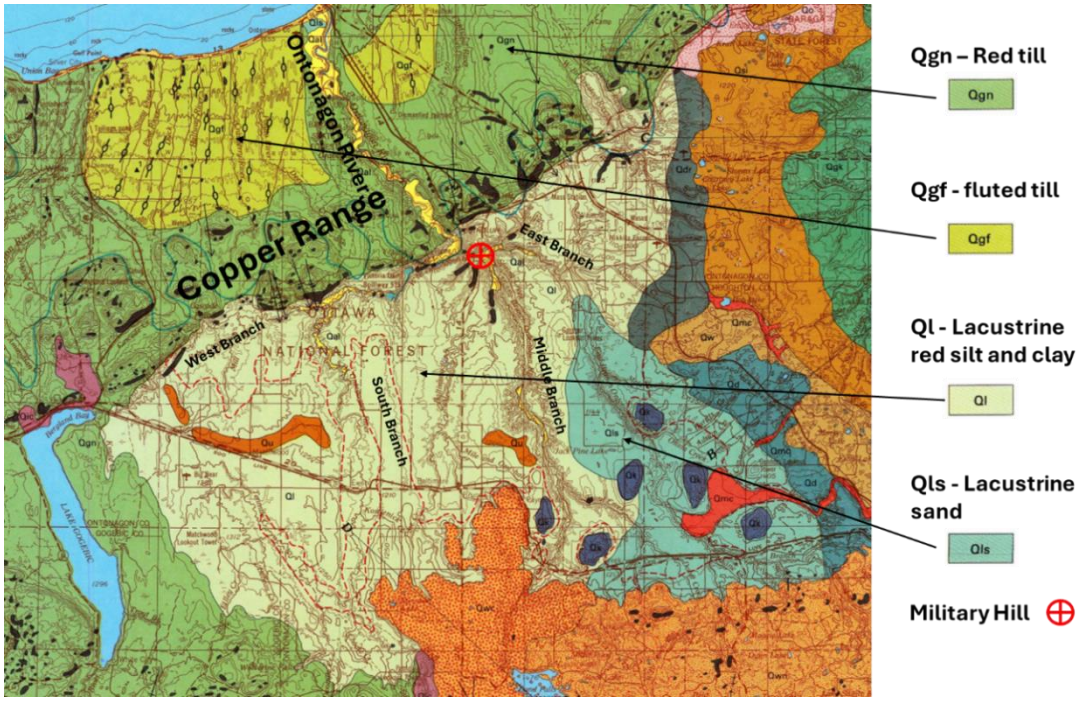


Figure 10: A portion of the USGS map of the glacial history of Iron River 1° x 2° Quadrangle (Peterson, 1985).

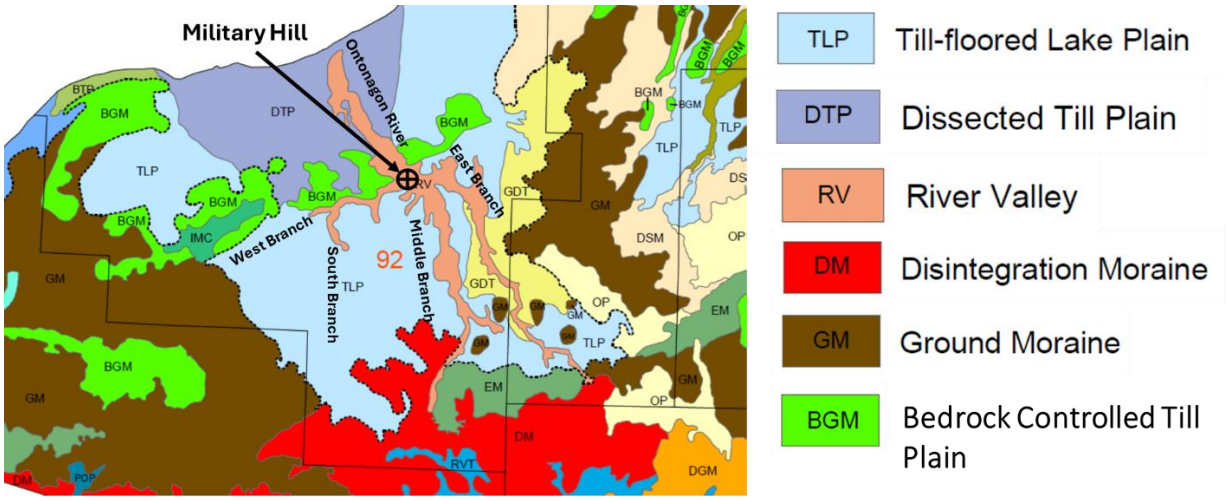


Figure 11: Landforms map of the study area adapted from Jerome (2006), showing the main soil regimes.

River hydraulic data were obtained from a USGS gauging station downstream. The USGS Scoops3D limit-equilibrium analysis software was used to develop a "*factor of safety*" map of the study area, which is shown in Figure 13(c). A significant portion of the Ontonagon River slopes have a factor of safety less than one, which is supported by the observable landslide activity shown in Figure 13(b) through the lacustrine sediments of Glacial Lake Ontonagon and Duluth.



Figure 12: Landslide activity upstream of US-45 on the East Branch of the Ontonagon River (Google Map, 2024).

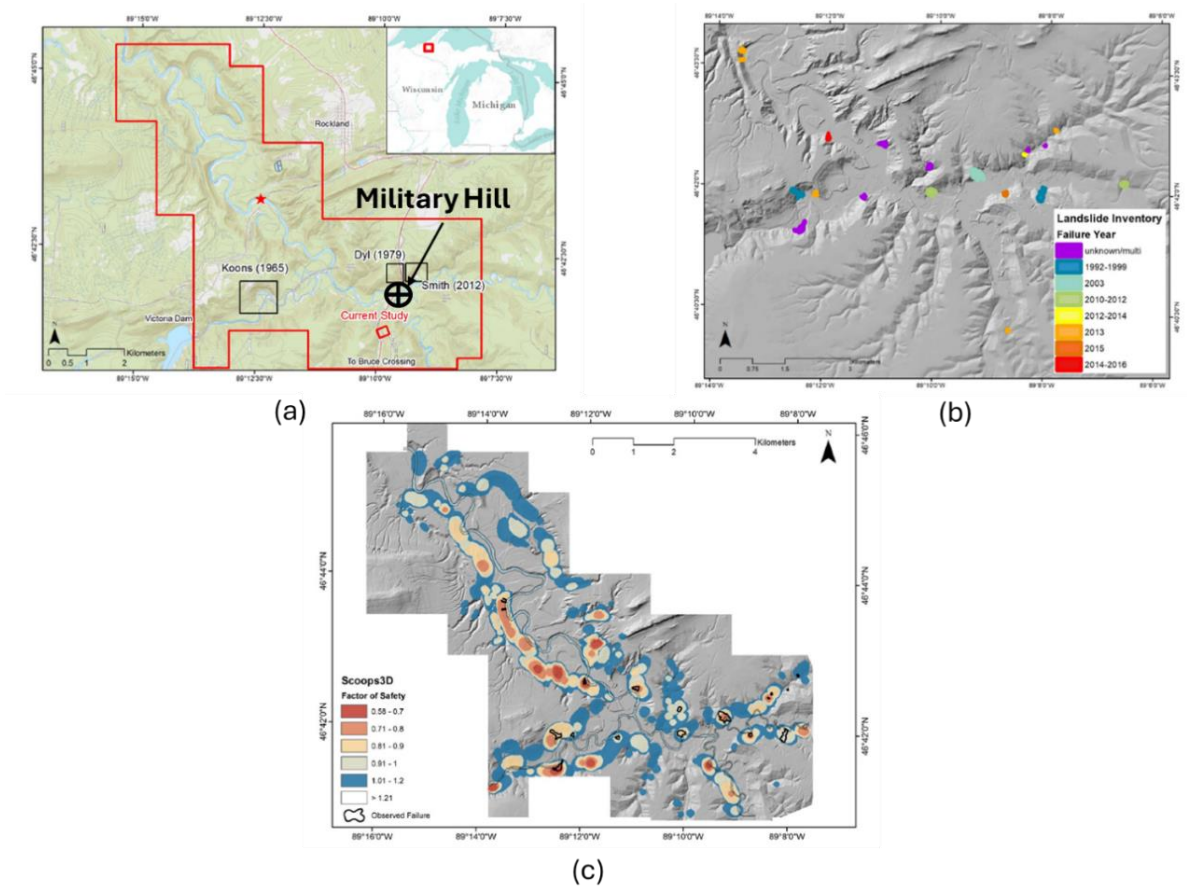


Figure 13: Weidner et al., Military Hill landslide investigation, (a) study area, (b) observed landslides between 1992 and 2016, and (c) Scoops3D factor of safety assessment.

Objectives of the field trip:

The objective of the field trip is to investigate landslides in the sediments deposited in the former Glacial Lakes Ontonagon and Duluth. In general, the sediments consist of lacustrine sediments deposited in Glacial Lake Ontonagon and later in Lake Duluth which overlay older till deposits. The first stop, however, will be at Quincy Hill, north of Hancock, MI to observe glacial grooving and striations in the Portage Lake Volcanics indicating the direction of glacial lobe movement across the Keweenaw Peninsula. The second and third stops will be in the US-45 Military Hill area. A summary of the stops are as follows:

- **Stop 1:** Top of Quincy Hill, north of Hancock, to view glacial grooving in the Portage Lake Volcanics indicating the direction of Keweenaw Bay Lobe movement westward into the Ontonagon Lobe that formed Lake Ontonagon.
- **Stop 2A:** The north side of Military Hill to observe the effects of vegetation on limiting landslide development.
- **Stop 2B:** US-45 Military Hills Roadside Park: 2003 Large landslide on the East Branch of the Ontonagon River
- **Stop 2C:** Lower Military Hill Erosion with Slope Movement
- **Stop 2D:** Middle Military Hill with Partial Vegetation and Some Slope Movement
- **Stop 2E:** Middle Military Hill with More Vegetation and Limited Slope Movement
- **Stop 2F:** Middle Military Hill with Some Vegetation and Varved Clay Slope Movement
- **Stop 2G:** Upper Military Hill with Active Slope Movement on both Sides of Highway in non-varved clay
- **Stop 3:** Slope movement two miles south of Military Hill, across from Primrose Acres, where a recently excavated slope on the east side of US-45 has been moving for about eight years.

Stop 1: Glacial Grooves at the Quincy Hill Historic Park Lookout

Directions: *From Michigan Tech drive west through Houghton on US-41 and cross the Houghton-Hancock Bridge, staying on US-41 going into Hancock. In Hancock, go straight uphill, passing Hancock's main street, onto East White Street. East White Street will take you to US-41 bypassing downtown Hancock. At the US-41 stop, take a right turn, going uphill to the Quincy Mine Hoist. Directly across from the entrance from the Quincy Mine Hoist take a left turn onto No. 2 Road. Follow No. 2 Road a short distance until you come to a two-track road on your left that takes you to the Keweenaw National Historic Park's "Quincy Mine Dryhouse Ruins" parking lot. The glacial grooves are a short distance from the parking lot. From Michigan Tech to the parking lot is 3.6 miles.*
Lat: 47.135904°, Lon: -88.577986°



Figure 14: View, looking west, of glacial grooves sculptured into the Portage Lake Volcanics at the Keweenaw National Historic Park.

Glacial grooving and striations are common along the Keweenaw Peninsula as the glaciers moved over the Portage Lava Volcanics, which were resistant to glacial erosion. It is generally assumed glaciers came from northeastern Canada moving south to southwest but are surprised to see grooves heading due west as shown in Figure 14. The reason for this direction is the movement of the Keweenaw Bay Lobe filled Keweenaw Bay and then moved west, south, and east as shown in Figure 15. The Keweenaw Bay Lobe then was stopped by the Ontonagon Lobe and to the west by the Michigamme Lobe to the east.

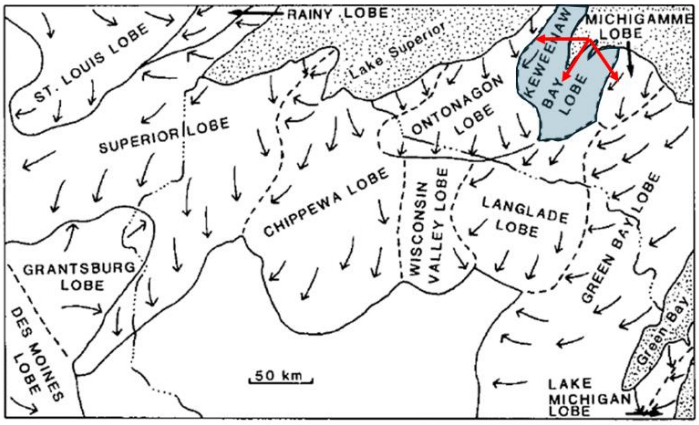


Figure 15: Location of glacial lobes in the western Lake Superior Basin (from Attig et al., 2013).

Stop 2: Military Hill – Seven Stops – Safety instruction will be provided at the site

Directions: *From Michigan Tech drive west through Houghton on US-41 to M-26. At the Houghton-Hancock Bridge go straight through the intersection to access M-26. Stay on M-26. Going southwest 37.8 miles to the M-26/M-38 intersection. Take a left turn, staying on M-26 through Mass City, to the M-26/US-45 intersection, 5.5 miles. Take a left turn at the M-26/US-45 intersection headed south 1.8 miles to Stop 2A.*
Lat: 46.705726°, Lon: -89.159581°

Stop 2A: North US-45 - Slope with Vegetation and Small Slope Movement

In the Military Hill area, the north side slopes required less excavation than on the south side and therefore there is less slope movement. In addition, where the slopes were excavated, they tend to be more vegetated as seen in Figure 16. This is due (possibly) to the north facing slope losing the spring snow before the south facing slopes and thus having longer growing season.



(a) (b)

Figure 16: Stop 1 (a) descending Military Hill going south and (b) vegetated slopes with some slope movement.

Stop 2B: US-45 Military Hill Roadside Park: 2003 Large landslide on the East Branch of the Ontonagon River

Directions*: From Stop 2A to Stop 2B go 0.4 miles south on US-45 to the Military Hills Roadside Park parking lot on the east side of US-45. Starting at the rest area facilities (outhouses) on the northside of the parking lot, walk northeast through the woods to the landslide. There is no path, but the woods are relatively easy to walk through. The 2003 large scale landslide is approximately 1,400 ft (411 m) northeast of the parking lot as shown in Figure 17. There is a smaller landslide north of the path, which has been active for many years, but is difficult to access. The slope debris, below the landslide, has a thick undergrowth, which is difficult to walk through.*

Military Hills Roadside Park: Lat: 46.699650°, Lon: -89.158627°
Military Hill 2003 Landslide: Lat: 46.703234°, Lon: -89.154132°

In 2003, a large-scale landslide occurred on the East Branch of the Ontonagon River as shown in Figure 1 and Figure 17. The landslide's stratigraphy consisted of lacustrine varved clay over a clean alluvial sand that grades downward into a red silty sand, silt and then clay till as illustrated in Figure 18 and Figure 19. While the failure mechanism is unknown, it is speculated that the spring runoff and possibly high-water table caused the alluvial sand to liquefy causing the massive landslide. As the liquefied sand lost strength, it started to flow outward into the river channel as shown in Figure 20(b) and (c). During the slope's collapse, a sliding plane formed in the varved clay as shown in Figure 20(a) allowing the varved clay to fail over the alluvial sand. Liquefaction boils are shown in Figure 20(b) and (d).

It will take about 20 minutes to walk to this landslide through the woods from the roadside park. While there is no path, the woods are fairly open and easy to walk in. However, the landslide itself is more difficult due to the landslide debris and the vegetation that has developed over the years. Caution must be observed when walking over this site and it is highly recommended that the landslide slopes are not accessed.



Figure 17: Stop 2B Military Hills Roadside Park and path to the 2003 landslide.

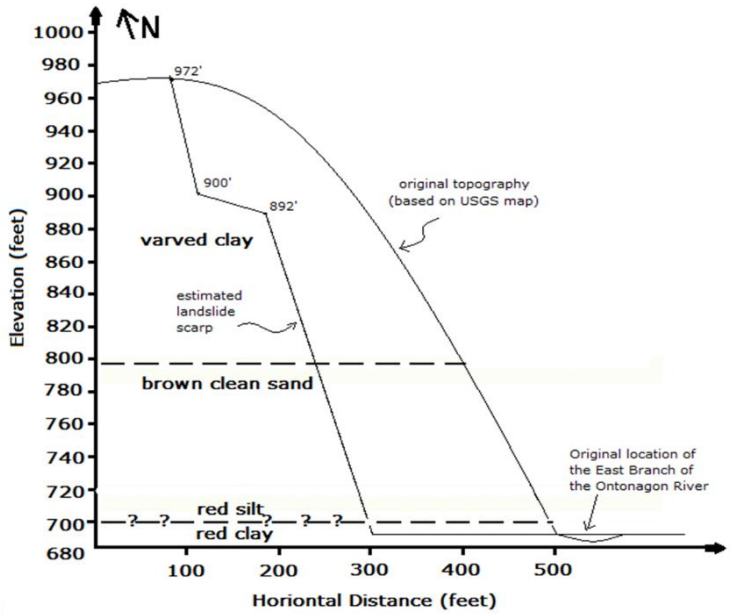


Figure 18: Assumed stratigraphy of the 2003 landslide (Smith, 2012).

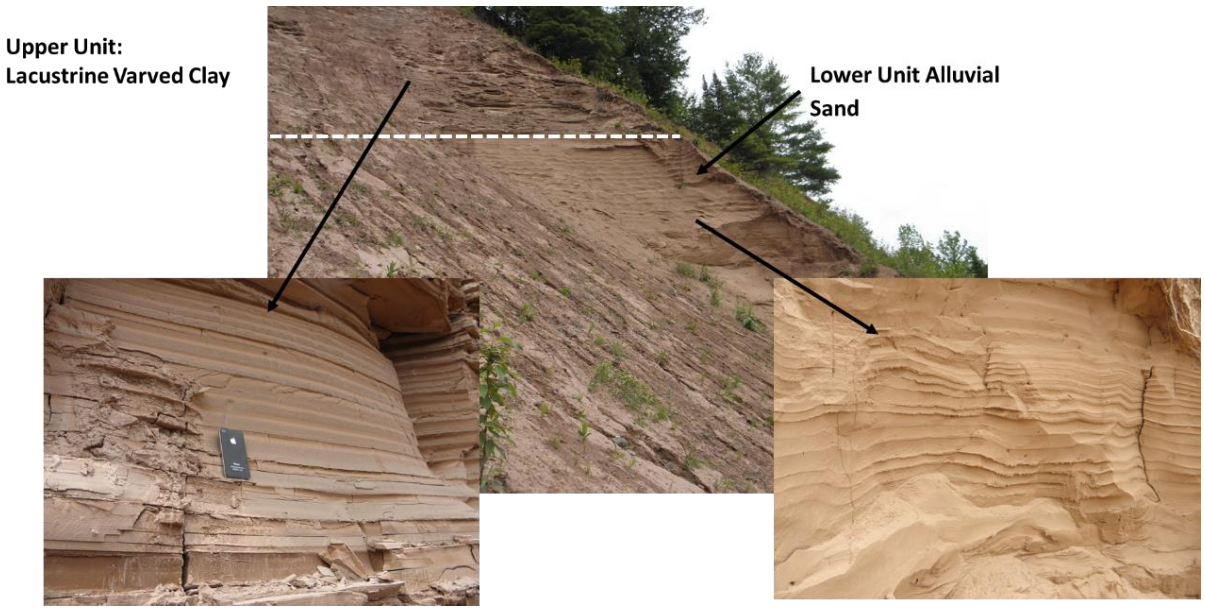


Figure 19: 2003 landslide illustrating the varved clay directly over an alluvial sand.

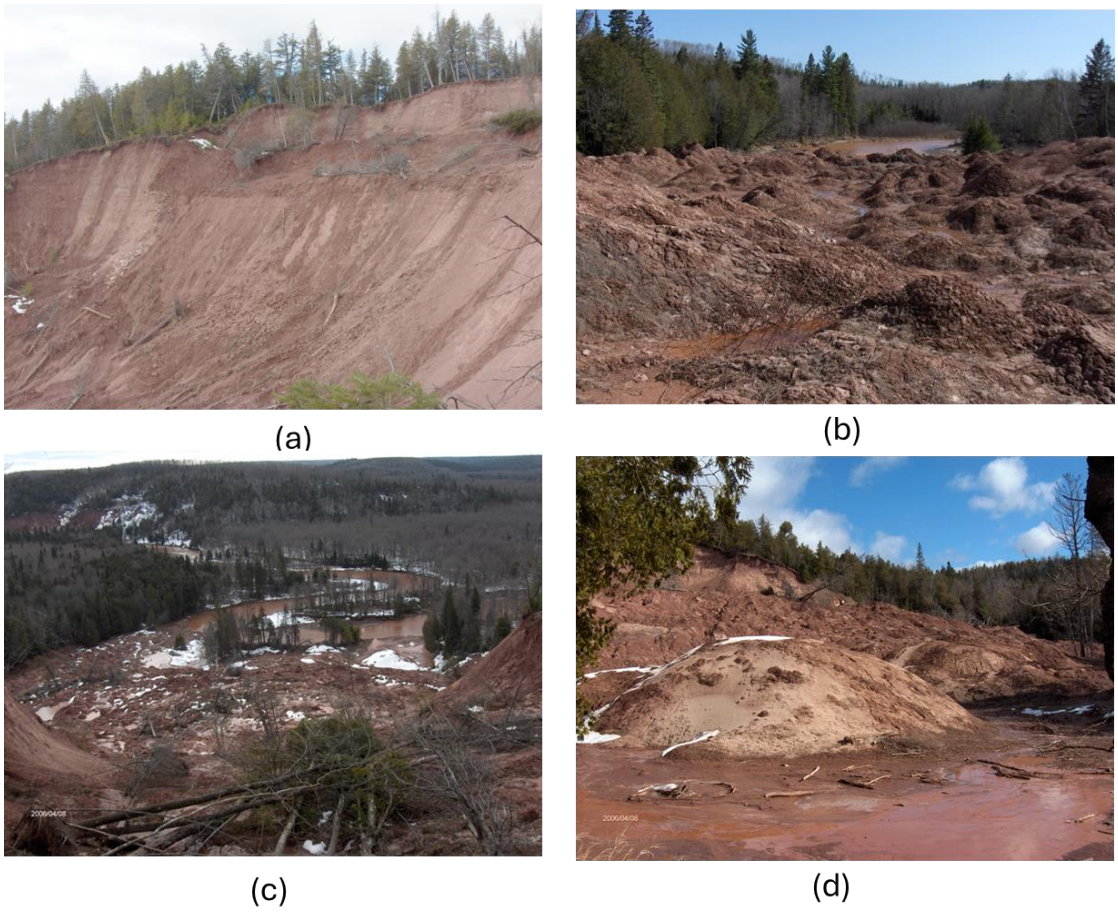


Figure 20: Landslide illustrating (a) a sliding plane on the varved clay, (b) liquefaction boils in the varved clay, (c) the flow debris into the river's channel, and (d) a large sand boil.

Stop 2C: Lower Military Hill Erosion with Slope Movement

Directions: *From Stop 2B go 0.8 miles south on US-45 across the US-45 bridge to Stop 2C, which is the start of the lower landslides with significant erosion.*
Lat: 46.690087°, Lon: -89.165513°

At the following five stops we will travel up Military Hill's southside. In general, the soils traveling up the south side of Military Hill are like the sediments seen at Stop 2B except for the alluvial sands. The varved clays overlie a sandy brown silt, which have significant erosion (Figure 21). The base of the varved clay can be seen moving over the sandy silt. As we moved up US-45 at Stop 2C a short distance, MDOT has been attempting to stabilize a small landslide just below the base of the varved clay as shown in Figure 22. The clay content appears to be higher in this area. On the west side of US-45 you can observe the steep US-45 embankment that descends down to Sandstone Creek, exposing the Jacobsville Sandstone.



Figure 21 Stop 2C showing the contact between the varved clay and a brown/red sandy silt, which is eroding.



Figure 22 Stop 2C small landslide.



Figure 23 Stop 2C showing the US-45 embankment that descends to Sandstone Creek on the Jacobsville Sandstone.

Stop 2D: Middle Military Hill with Partial Vegetation and Some Slope Movement

Directions: *From Stop 2C go 0.23 miles south on US-45 Stop 2D, which is at the middle landslides.*
Lat: 46.686869°, Lon: -89.164320°

Stop 2D is a short distance up US-45 and is in the same sediments as Stop 2C, which also are sloping uphill as seen in Figure 24. Movement of the varved clay base can be seen in the upper portion of Figure 24.



Figure 24 Stop 2D varve clay movement over the lower sandy silt but without the erosion seen at Stop 2C.

Stop 2E: Middle Military Hill with More Vegetation and Limited Slope Movement

Directions: *From Stop 2D go 0.30 miles south on US-45 Stop 2E, an excavated slope that is now partially vegetated.*
Lat: 46.682564°, Lon: -89.164857°

At Stop 2E, the slope has much more vegetation with limited apparent slope movement.



Figure 25 Stop 2E, slope with more vegetation and limited slope movement.

Stop 2F: Middle Military Hill with Some Vegetation and Slope Movement

Directions: *From Stop 2E go 0.34 miles south on US-45 Stop 2F, a slope with vegetation that had was excavated for the construction of US-45.*
Lat: 46.677892°, Lon -89.166195°

At Stop 2F, there is again less vegetation but more slope movement Figure 26. The clay soils at this elevation are not varved.



Figure 26: Stop 2F showing partial vegetation and upper slope movement.

Stop 2G: Upper Military Hill with Active Slope Movement on both Sides of Highway

Directions: *From Stop 2F go 0.23 miles south on US-45 Stop 2G, a slope with active slope movement.*
Lat: 46.674640°, Lon -89.167126°

At Stop 2G, landslides occur on both sides of the highway as shown in Figure 27. Figure 27(a) shows slope movement on the west side of US-45, which has less movement than on the eastside of US-45 shown in Figure 27 (b), (c) and (d). These landslides have been moving for many years and appear to be in a non-varved clay. While limited soil investigation has been conducted in these sediments, it appears that at the higher elevations in the Military Hill area, a non-varved lacustrine soil overlies the varved clay soils indicating that it possible that this sequence might indicate when Glacial Lake Duluth and Lake Ontonagon combined. An interesting feature of Glacial Lake Ontonagon is that it is at a much higher elevation at 1,320 feet than Glacial Lake Duluth. Isostatic rebound would have had to occur for the lakes to combine.

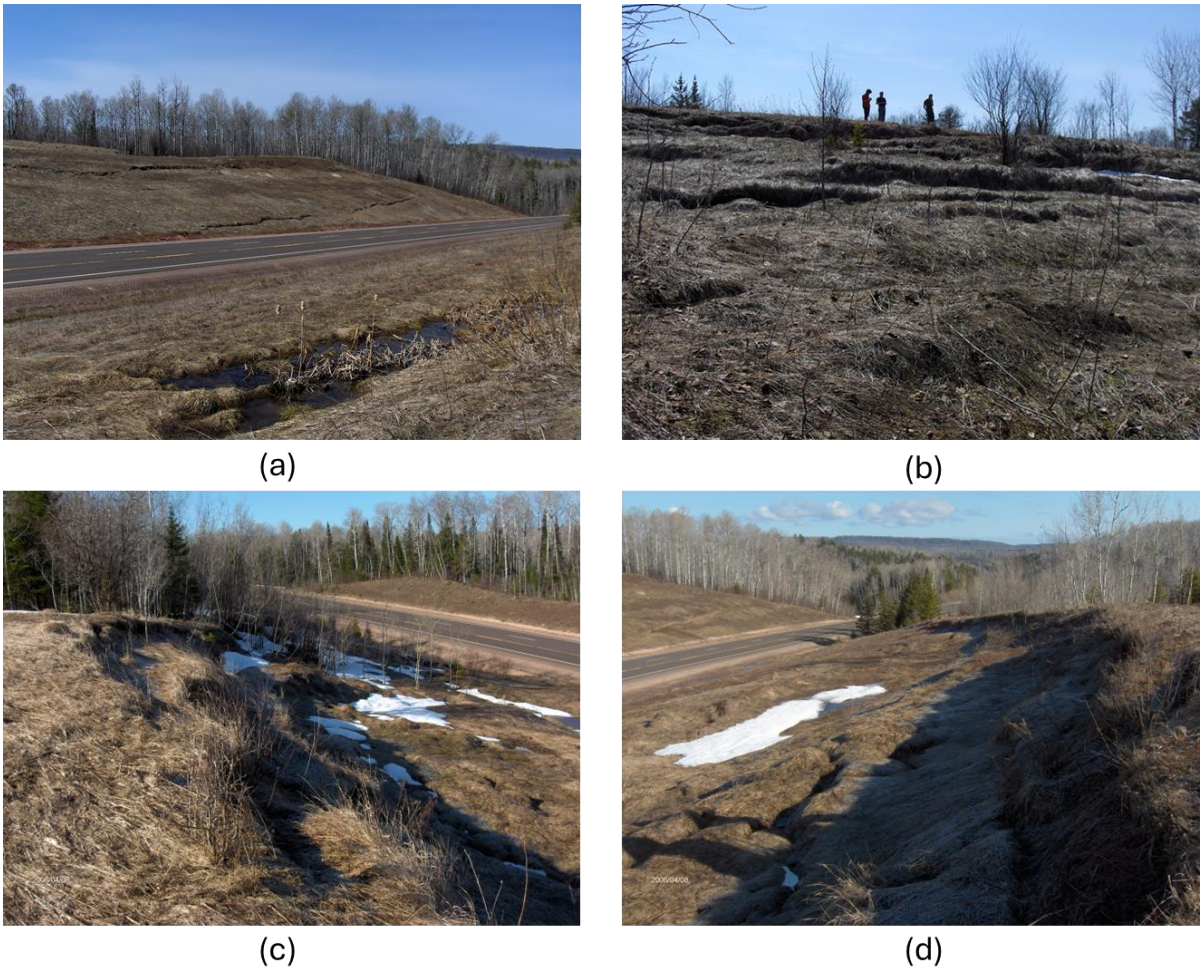


Figure 27 Stop 2G near the top of Military Hill showing slope movements on both sides of the highway with (a) west side, (b) top of east side, (c) south section of slope movement on east side, and (d) north portion of slope movement on the east side.

Stop 3: US-45 Recent Excavation and Resulting Slope Movement

Directions: *From Stop 2G go 3.4 miles south on US-45 Stop 3, a slope with active slope movement.*
Lat: 46.627240°, Lon -89.178520°

Stop 3 is at a recent location where MDOT excavated a natural slope to improve drainage along the eastside of US-45. The natural slope was excavated at a 2H:1V angle, the same slope as the highway embankment as can be seen in Figure 28. Soon after excavation, however, the slope started to move and has been moving ever since. Figure 29 was taken one year after the photo in Figure 26. The US-45 embankment was constructed with the local clay soil but was compacted, whereas the natural slope was not. The clay soils at this site and at Stop 2G have relatively high plasticity and are at the interface of the low (CL) and high (CH) plasticity soils defined by the Unified Soil Classification System (USCS) with moisture contents in the 20 to 25% range. As we saw at Stop 2F, the lacustrine clay soils are not stable at the angles at which they were excavated, unless compacted.



Figure 28 Stop 3 showing a recent excavation into the clay soil at a 2(H):1(V) angle.



Figure 29 Stop 3 - excavated slope one year after the photo in Figure 26 was taken.

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