



64th Annual Meeting
Institute on
Lake Superior Geology
Iron Mountain, Michigan
May 15-18, 2018



1. Cambrian sandstone on Vulcan Iron-formation,
Groveland Mine, Felch, MI
2. Pillowed basalt in Hemlock Formation, Way Dam, MI
3. Athelstane Quartz Monzonite at Dave's Falls, WI
4. Quinnesec iron mine, Menominee Range, MI

Proceedings Volume 64
Part 2: Field Trip Guidebooks



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Proceedings Volume 64

Part 2: Field Trip Guidebooks

Compiled by William F. Cannon

Cover Photos: 1. Sandstone of Munising Formation (Upper Cambrian) lying on the Vulcan Iron-formation at Groveland Iron Mine. Seen on trip 1. Photo by William Cannon. 2. Pillowed basalt of the Hemlock Formation at Way Dam, Michigan. Seen on Trip 2. Photo by Thomas Waggoner., 3. Dave's Falls on the Pike River near Amberg, Wisconsin. Bedrock is the Athelstane Quartz Monzonite cut by diabase dikes. Seen on trip 4. Photo by William Cannon. 4. Quinnesec Iron Mine on the Menominee Iron Range, Michigan. Seen on Trip3. Photo by William Cannon.

64TH INSTITUTE ON LAKE SUPERIOR GEOLOGY

VOLUME 64 CONSISTS OF:

PART 1: PROGRAM AND ABSTRACTS

PART 2: FIELD TRIP GUIDEBOOK

TRIP 1: ARCHEAN AND PALEOPROTEROZOIC GEOLOGY OF THE FELCH DISTRICT,
CENTRAL DICKINSON COUNTY, MICHIGAN

TRIP 2: GEOLOGY OF THE HEMLOCK FORMATION

TRIP 3: GEOLOGY AND IRON ORES OF THE MENOMINEE IRON RANGE, DICKINSON
COUNTY, MICHIGAN

TRIP 4: GRANITOID ROCKS OF THE PEMBINE-WAUSUA TERRANE IN NORTHEASTERN
WISCONSIN

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

*Authors, 2018, abstract title, 64th Institute on Lake Superior Geology Proceedings, v. 64,
Part 1, Field Trip Guidebook, p. xx.*

Proceedings Volume 64, Part 1: Program and Abstracts, and Part 2: Field Trip Guidebook are published by the 64th Institute on Lake Superior Geology and distributed by the Institute Secretary:

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Some figures in this volume were submitted by authors in color, but are printed grayscale to conserve printing costs. Full color imagery will appear in the digital version of the volume when it is available on-line at:

<http://www.lakesuperiorgeology.org>

ISSN 1042-9964

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FIELD TRIP 1

Tuesday May 15, 2018

ARCHEAN AND PALEOPROTEROZOIC GEOLOGY OF THE FELCH DISTRICT, CENTRAL DICKINSON COUNTY, MICHIGAN

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INTRODUCTION

This trip examines the stratigraphy, structure, and economic geology of Precambrian rocks near the town of Felch in central Dickinson County of northern Michigan. The location of the area is shown on Figure 1 relative to other well-documented structures and iron ranges of the region. Precambrian rocks range in age from Archean to Paleoproterozoic, and outliers of Cambrian sandstones are also widespread. Relationships along the basal Cambrian unconformity are included in the trip. Much of the interest in the area, both geologically and economically, has been focused on the Felch trough (Figure 2) where Paleoproterozoic rocks of the Chocoy and Menominee Groups form a complex syncline that is infolded and infaulted with Archean gneisses. Minor amounts of iron ore were produced in the early history of the area and a major concentrating-grade mine, the Groveland, was active in the 1960's and 1970's.

The geology of the area was mapped and described in detail by the U.S. Geological Survey in 1950's and published in 1961 (James et al., 1961). The information in that report is the basis for much of the descriptive material in this guide. A few more recent studies have added significant additional data and geochronological constraints that clarify some of the relationships between various rock units and the events that formed them. An underlying theme of the trip is that a substantial amount of additional research is warranted to fully understand this complex region and that relatively abundant outcrops, along with newly acquired high-resolution geophysical data, make this a very attractive region for a wide variety of research.

Stratigraphy. The long-accepted stratigraphic sequence of Archean and Paleoproterozoic rocks that was defined by James et al. (1961) requires some modification based on more recent radiometric age determinations. Some of these are discussed in more detail below in individual stop descriptions. A tentative correlation chart in Figure 3 reflects these suggested changes, but additional work seems warranted to clarify some aspects before proposing formal changes to stratigraphic nomenclature and correlations.

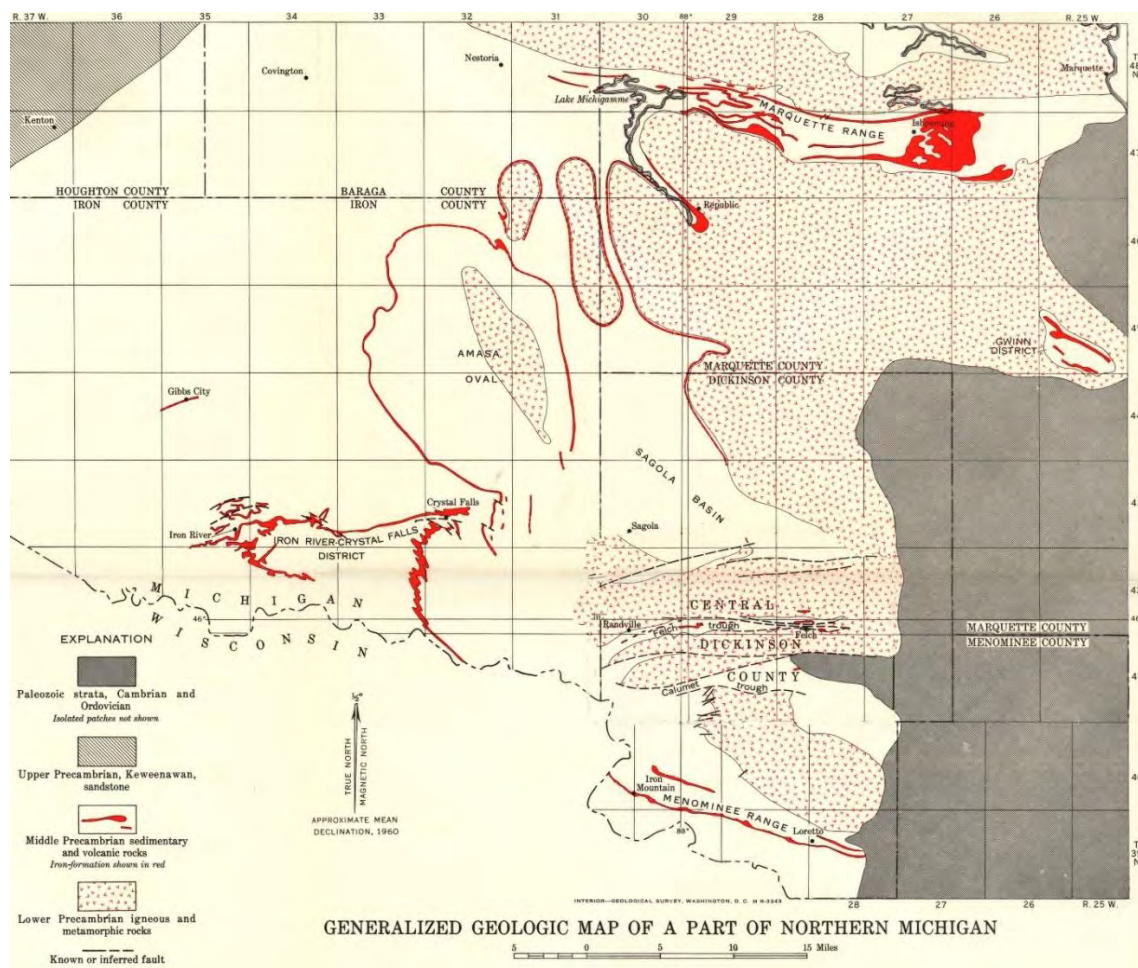


Figure 1. Generalized geologic map of part of the Upper Peninsula of Michigan showing the location of central Dickinson County in relation to major structures and iron ranges of the region. Modified slightly from James et al. (1961, plate 1).

The oldest rocks of the region are complex gneisses of the Carney Lake Gneiss and other probably correlative units in structurally separated fault blocks. They have long been considered Archean. Recent geochronology of samples collected south of the fieldtrip area using the USGS/Stanford Sensitive High Resolution Ion Microprobe (SHRIMP) produced U-Pb data on zircons that confirm an Archean age (Ayuso et al., 2017; 2018). Two samples were collected for radiometric dating from the southern half of the Carney Lake complex: 1) sample 1 is from a granitic K-feldspar-bearing gneiss that is locally pegmatitic; 2) sample 2 is from a banded and folded gray to red granitic gneiss. Abundant zircons (70-200 μ) were obtained from sample 1 that range from anhedral to subhedral, contain complex igneous and irregular growth zoning, and multiple growth rims; these zircons have irregular to pyramidal overgrowths. The zircons from sample 2 range from slightly rounded to subhedral and are otherwise mostly similar to zircons from sample 1. A total of 129 cores and rims were analyzed. Individual zircons have older ages near their cores (mostly discordant) and younger ages near their rims (Figure 4A).

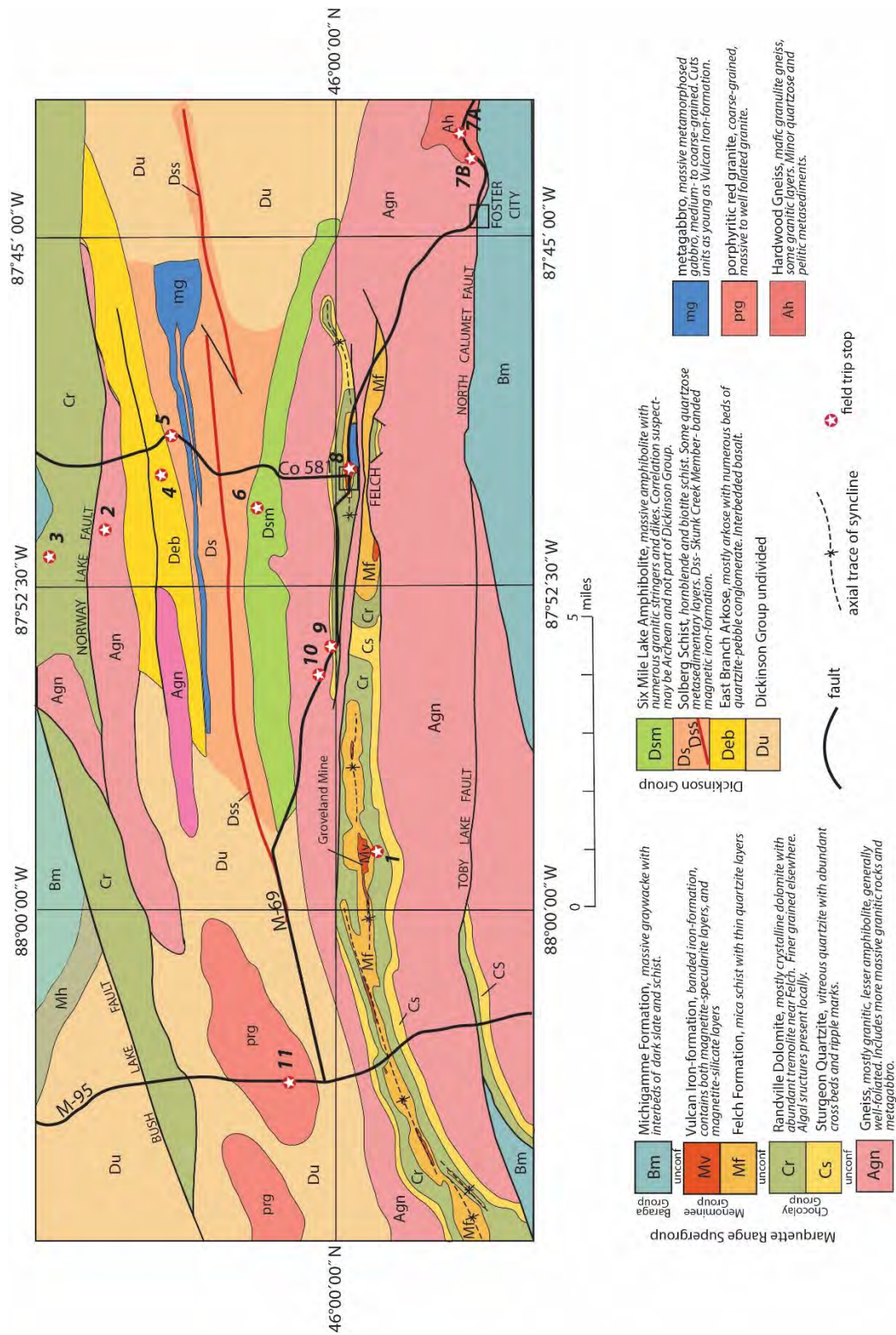


Figure 2. Precambrian geology near Felch, Michigan showing locations for Field trip 1. Generalized slightly from James et al. (1961). Numerous outliers of Munising Formation (upper Cambrian) are not shown.

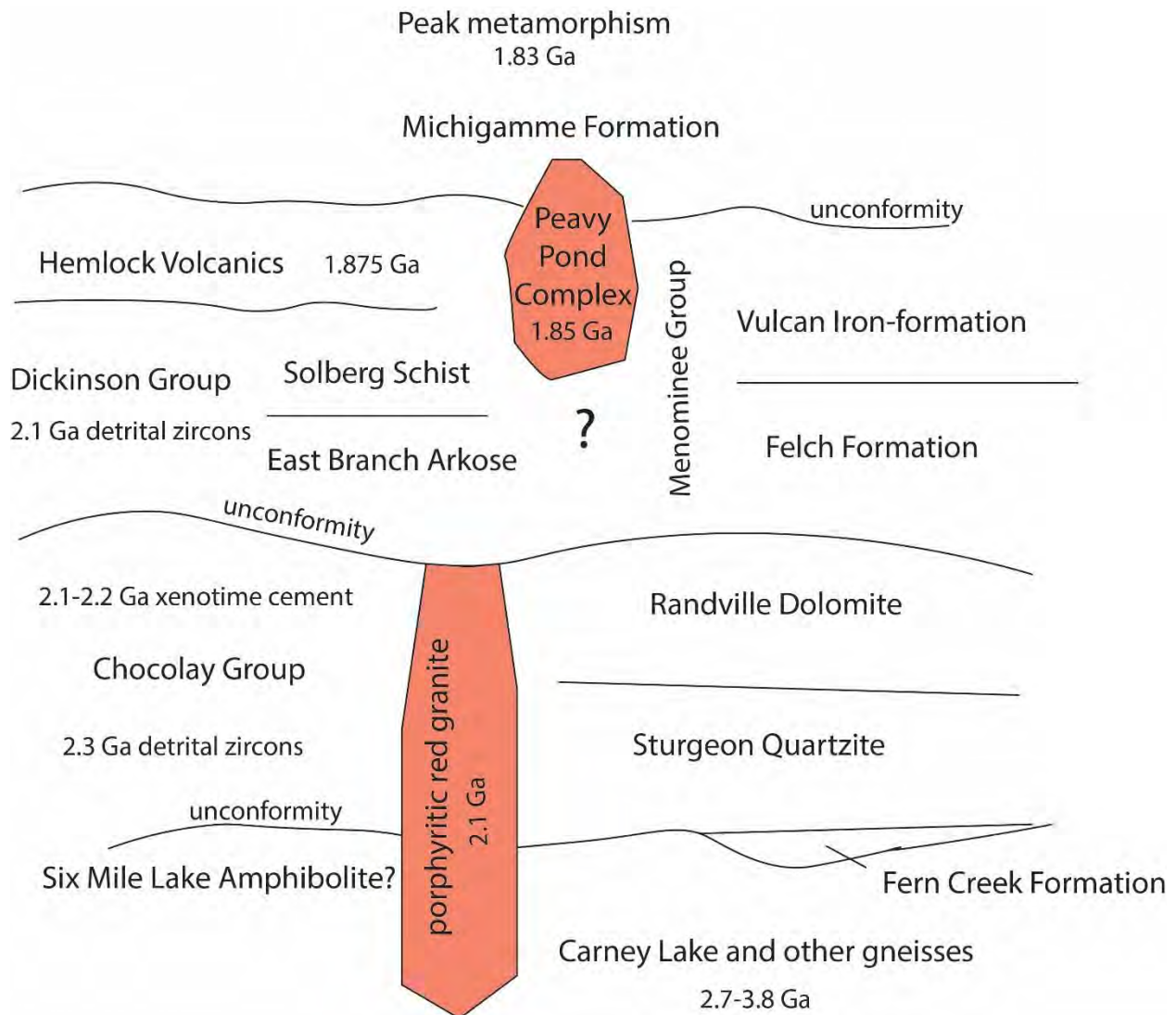


Figure 3. Proposed correlation chart for the area of Field trip 1. Modifications from previous correlations are based on recent radiometric determinations that provide direct ages for some units and place constraints on the ages of others.

On a concordia diagram, U-Pb data plot as clusters of data points ranging from concordant to discordant and suggest several chords and intercepts that are common to both samples from the Carney Lake Gneiss (Figure 4B). That study identified cores of individual zircons as old as 3.8 Ga. The most common age for individual zircons and for rims on older grains is about 2.75 Ga and records a younger major event in the late Archean.

James et al. (1961) defined the Dickinson Group, consisting of the East Branch Arkose, Solberg Schist, and Six Mile Lake Amphibolite, as Archean based on field relationships between the Six Mile Lake Amphibolite, the upper formation of the group, and presumed Archean gneisses to the south. However, U-Pb dating of detrital zircons in the East Branch Arkose has shown that it must be 2.1 Ga or younger (Craddock et al., 2013). We suggest that the Six Mile Lake Amphibolite is also Paleoproterozoic because of chemical similarities between it and other Paleoproterozoic mafic rocks. The Archean age proposed by James et al. (1961) was based on field relationships

between amphibolites and Archean granites and gneisses that require further evaluation. The Solberg Schist is intruded by a gabbro that has the distinctive composition of the Six Mile Lake Amphibolite (see Stop 6 for discussion). James et al. (1961) described the contact between the East Branch Arkose and Solberg Schist as transitional. Further study of the Dickinson Group is clearly required to resolve the age relationships of its three formations.

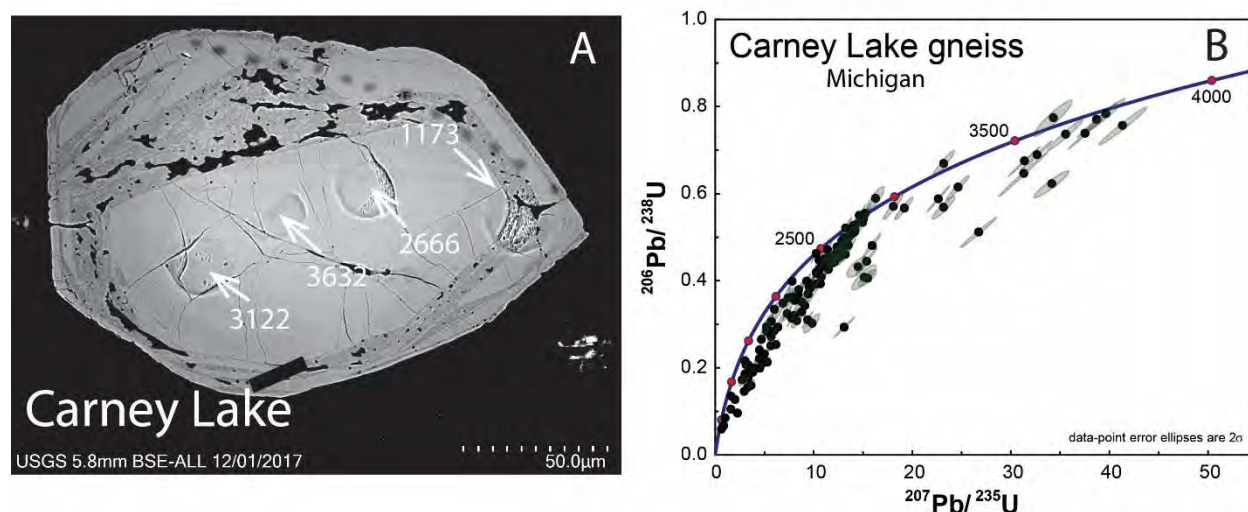


Figure 4. A- BSE (back scatter electron) image of a zircon from the Carney Lake Gneiss showing ages of four analyzed spots. B- Concordia diagram for 129 spot analyses from zircons in the Carney Lake Gneiss.

A distinctive granite in the area referred to informally as the “porphyritic red granite” by James et al. (1961), has now been dated using SHRIMP U-Pb on clear and pale brown zircons at 2.1 Ga (Ayuso et al., 2018; and discussion of Stop 11). The porphyritic red granite provides a local source for the relatively common detrital zircons of that age found in the East Branch Arkose of the Dickinson Group (Craddock et al., 2013). These zircons also suggest an erosional interval after 2.1 Ga to unroof the granite prior to deposition of the Dickinson Group. The Dickinson Group is overlain by the Hemlock Volcanics northwest of the field trip area. Those volcanics are dated at 1.875 Ga (Schneider et al., 2002) and establish a minimum age for the Dickinson Group.

The Chocoma Group, consisting of the glaciogenic Fern Creek Formation, Sturgeon Quartzite, and Randville Dolomite, is well constrained to have been deposited between 2.3 Ga, the age of detrital zircons within it, and 2.1-2.2 Ga, the age of xenotime cement in the sediments (Vallini et al., 2006).

The Menominee Group, including the classic stratigraphy of the Vulcan Iron-formation and underlying Felch Formation, is nowhere in contact with the Dickinson Group although the two are not far separated in the Felch area (see Figure 2). They may be largely time equivalent in spite of rather pronounced lithologic differences.

The Michigamme Formation lies at the top of the stratigraphic succession across the region and has been widely interpreted to have an unconformable contact with the underlying units. The Michigamme and underlying Hemlock Volcanics are intruded by the Peavy Pond Complex to the

northwest of the field trip area. A recent SHRIMP U-Pb date on honey-colored euhedral to subhedral zircons for the Peavy Pond Complex yielded an age of 1.85 Ga (Figure 5) (Ayuso et al., this volume). The results place a minimum age on at least the lower part of the Michigamme Formation. This presents something of an enigma in that elsewhere in the Upper Peninsula the Sudbury impact layer, which was deposited at 1.85 Ga, lies at or near the base of the Michigamme Formation (Cannon et al., 2010). It seems, therefore, that rocks assigned to the Michigamme Formation in the Felch area could be substantially older than the Michigamme Formation in much of the larger region of the Upper Peninsula.

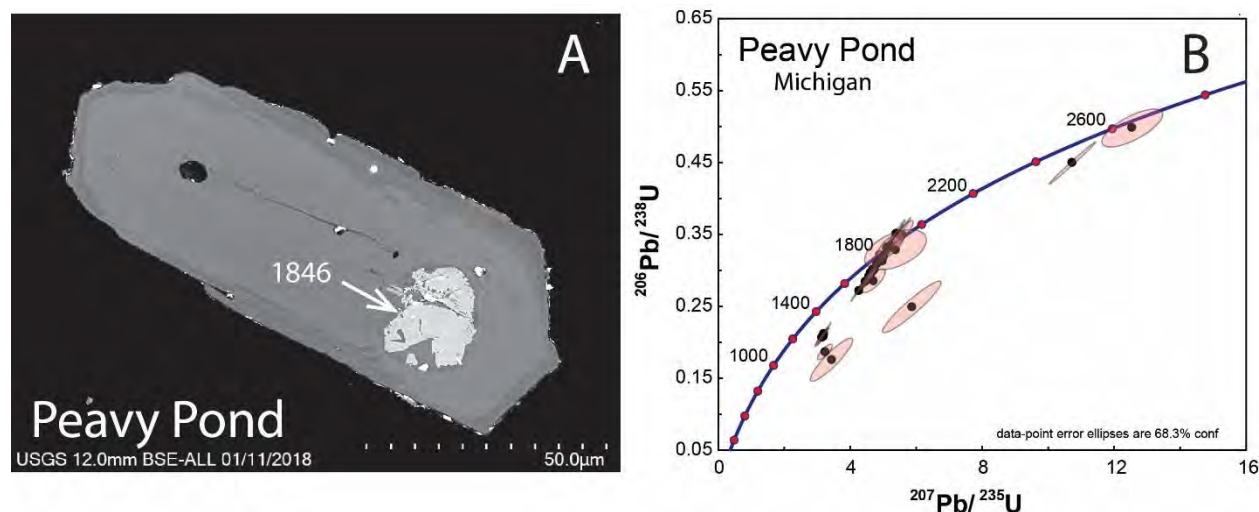


Figure 5. A- BSE (back scatter electron) image of a zircon from the Peavy Pond Complex showing the SHRIMP U-Pb age of an analyzed spot. Bright spots are residual gold coating from SHRIMP analyses. B- Concordia diagram for 32 spot analyses of zircons from the Peavy Pond Complex.

Tectonics. Rocks in the Felch region record multiple orogenic events from 3.8 Ga to younger than 1.83 Ga. Archean rocks are very complexly deformed and metamorphosed by at least two Archean orogenies, one at 3.8 Ga and a later 2.75 Ga event. They also record one or more deformational and metamorphic events in the Paleoproterozoic. Although the work of James et al. (1961) mapped and described much of the Archean of the area, detailed structural studies have not been done, and much remains to be learned about the sequence of rock-forming and deformational events recorded in these complex gneisses.

More study has been devoted to the tectonics of Paleoproterozoic rocks and at least an outline of their tectonic history is known, although much additional detail seems likely to be decipherable with further examination. Only two published structural studies based on substantial new field data have been published since 1961. Ueng and Larue (1988) defined four tectonic terranes and six phases of deformation within Paleoproterozoic strata. Klasner and Sims (1993) proposed a somewhat different sequence and suggested that the set of faults that dominate the map pattern near Felch were backthrusts formed in later phases of the Penokean orogeny at about 1.85 Ga. Both of those studies were conducted prior to significant geochronologic constraints determined over the past two decades. They ascribed all deformational events to the Penokean orogeny,

which by that definition was a prolonged and complex event including both lateral and horizontal shortening phases.

More recently, several geochronologic studies show that the tectonic sequence is much more complex and that the region bears the imprint of two other orogenies after the Penokean. A sequence of Archean-cored gneiss domes surrounded by Paleoproterozoic sedimentary and volcanic rocks across central Minnesota, northern Wisconsin, and northern Michigan, once thought to be late Penokean structures, was documented to have formed in mid geon 17 (Schneider et al., 2004). This gneiss dome corridor was interpreted to have been related to subduction during the Yavapai orogeny at about 1.75 Ga. Archean masses near Felch are part of the gneiss dome corridor and, although now dismembered by later faulting, record that Yavapai basement doming event rather than Penokean deformation.

A distinctive aspect of the Felch region is a set of easterly and ENE-trending faults that cut all other structures (see Figure 2). They result in a sequence of fault slices that dominate the map pattern of the region. A comparable array of such faults is not known elsewhere in the region. This faulted domain is bounded on the north by the Bush Lake fault, across which there is a sharp discordance in structural trends, nearly 90° in places. The age of the faulting can be constrained by the fact that it offsets the 1.85 Ga Peavy Pond Complex to the west of the field trip area. It also has been shown to offset metamorphic isograds (Attoh and Klasner, 1989) of the Peavy node where peak metamorphism has been dated at 1.83 Ga (Schneider et al., 2004; Holm et al., 2007). Attoh and Klasner (1989) also documented a change in metamorphic pressures across the Bush Lake fault with peak pressures of 4.8 kbar to the south and 3.3 kbar to the north. This suggests that fault motion was south side up with vertical displacement on the order of five kilometers. This set of faults also partly dismembers the gneiss dome structures related to the mid-geon 17 Yavapai orogeny and suggests that they are post-Yavapai structures. An imprint of the geon 16 Mazatzal orogeny in this region was documented by Romano et al. (2000), who showed a heating event of 300-350°C across the Felch region in late geon 16. Whether this heating was accompanied by deformation has not been determined in the Felch region, but to the south, in Wisconsin, strong deformation of Baraboo interval quartzites shows conclusively that strong Mazatzal-aged deformation was widespread. It seems reasonable, with our present level of understanding of the tectonics of the Felch region, that the easterly-trending set of faults is the northernmost manifestation of Mazatzal deformation.

Metamorphism. In a classic work on regional metamorphism, James (1955) identified three metamorphic nodes within Paleoproterozoic rocks in the Upper Peninsula of Michigan. One of those, the Peavy node, encompasses the area of this field trip. Most stops lie within the staurolite zone and the effects of recrystallization are obvious in outcrop. Since James' work, radiometric dating and more modern metamorphic petrology studies have further defined the character of the metamorphism of the region. The age of peak metamorphism of the Felch node has been determined as circa 1.83 Ga from dates derived from metamorphic monazite (Schneider et al., 2004; Holm et al., 2007) measured in Archean gneiss near Foster City. Peak temperature and pressure of metamorphism were about 600-650°C and 5 kbars (Attoh and Klasner, 1988). A younger amphibolite facies metamorphism was documented at 1.78-1.74 Ga related to Yavapai accretion and gneiss doming (Holm et al., 2007).

Much older metamorphism, probably culminating at about 2.75 Ga, is evident within Archean rocks, many of which are gneisses and migmatites. The Hardwood Gneiss, seen at Stop 7,

records granulite facies conditions with estimated pressures of 8.2-11.6 kb and temperatures of ~770°C for an initial event, and conditions of 6.0-10.1 kb and temperatures of 610-740°C for a second event inferred to be Paleoproterozoic but “pre-Penokean” (Peterson and Geiger (1990).

Economic geology. Iron, mostly related to the Vulcan Iron-formation, has been the principal commodity of interest in the Felch region. Early exploration and attempts to develop mines are summarized by James et al. (1961). The first indication of ore in the Felch trough was a description of a ridge of high grade iron ore in eastern Iron County by Jacob Houghton and reported to William Burt, a government land surveyor, in Marquette County in 1846 and was recorded in Senate Documents for the 31st Congress (Jackson, 1849). At what later became the Groveland mine a veneer of oxidized Vulcan Iron-formation was found on the south side of the ridge that drew the attention of early developers. But once operations started stripping a pit, the veneer was found to be only several feet thick and did not extend to depth as originally thought. The first development was an economic disaster as monies had been invested in town sites and the extension of a railroad to the Escanaba iron furnaces.

Being not far from more prospective areas such as the Menominee Range to the south, the Felch area was heavily prospected and numerous attempts at mining occurred between 1880 and 1913. Only four mines of any significance were developed in the Felch area and produced a total of about 625 million tons of ore, mostly of low grade. These direct-shipping ores were soft masses of hematite and goethite that were found directly beneath the unconformity between the Vulcan Iron-formation and overlying Cambrian sandstones. They are widely accepted to be paleosupergene deposits formed by late Precambrian and/or early Cambrian weathering. The Felch region differs from much more productive nearby regions, such as the Menominee Range and Iron River-Crystal Falls district, in being substantially metamorphosed. Rocks, including the Vulcan Iron-formation, are coarse-grained as a result of metamorphic recrystallization and thus are less susceptible to the paleo-weathering than comparable iron-formations in other nearby districts where large paleosupergene ore bodies were formed.

With the advent of large-scale concentrating and pelletizing technology in the 1950's, portions of the Vulcan Iron-formation became targets for concentrating-grade ore production. The Groveland area proved to have sufficient tonnage of iron-formation accessible by open-pit mining to allow development of the Groveland mine, a significant iron-producer in the 1960's and 1970's.

The potential for mineral production in this area has dropped in recent decades as exploration projects, such as for uranium and diamonds, have failed to find deposits of current economic viability. There are active gravel pit operations, reprocessing of Groveland Mine waste rock dumps for crushed stone, and small quarries in the Randville Dolomite for decorative stone. This limited activity is the current extent of mineral resource exploitation in central Dickinson County

FIELD TRIP STOPS

Stop 1. Groveland mine. (45.988°N, 87.981°W) The geology within the long-abandoned open pit of the Groveland mine is not accessible for field trips because of flooding, slumping of pitwalls, and safety concerns of current owners. The mine property is fenced and is not accessible to the public. We have been given access to the property to examine material on large waste-rock piles that show good examples of the various lithologies of the Vulcan Iron-formation, and provide views into parts of the flooded pit. Figure 9 shows the present surficial character of the mine area and the location of the waste piles available for observation and sampling.



Figure 6. The Groveland Mine has a long history that began in the late 1880's and continued into the 1980's. This view, probably from the 1970's, is looking south with the plant in the upper central part of the air photo. Source M. A. Hanna Company. Archives of the Michigan Department of Environmental Quality.

History. Mining of iron ore at Groveland began as an underground operation in 1891 on outcrop of the Vulcan Iron-formation identified in 1846 by assistants to William Burt, the government land surveyor. The operation was abandoned after only a few years of operation due to the lack

of direct shipping ore. The mine was reopened in 1901 and mined for four years by Corrigan, McKinney & Company. In 1907 the mine was again started up by the Groveland Mining Company, and had production through a 294 foot deep, three compartment shaft with levels at 70, 140, and 210 feet. Iron content remained a problem and production ceased in 1913. It was reported that the last shipment had unacceptable iron content and was dumped into Lake Erie. Several companies gained ownership of the properties and in 1926 test pits and trenches were completed by an independent developer, Mr. R. M. Adams.

In 1948 the properties were consolidated through leases by M. A. Hanna Co. and in 1951 the Groveland became their first taconite project. A pilot project ran for six months to develop grinding and concentration processes of the jasper ores, but it took seven years to develop a viable process to treat the complex ore, which is unique because of its mineralogy and metamorphism. The ore is very coarse grained and was defined by the operators as three types; magnetite, magnetite silicate, and specular hematite (Figure 7). In 1957 construction of the concentrating and pelletizing plant began and in 1959 the plant became operational with an output of 700,000 tons of concentrate annually. The mine became the second concentrating grade (taconite) operation in Michigan, following closely the opening of the Humbolt mine on the Marquette range. In 1963 a traveling grate pellet plant was completed with a capacity of 1.25 million tons per year. The \$35 million expansion also included a concentrator upgrade and production was increased to 1.6 million tons annually. In 1968 a fourth line was added and resulted in an annual capacity of 2 million tons. 1977 was a record year with the output of iron concentrate reaching 2.1 million tons. Overall, the expenditures totaled \$70 million, and employed 530 resulting in an investment of \$132,000 per employee. Annual payroll was \$12.5 million and taxes provided \$1.332 million of revenue to the state and local governments. Operating services and supplies were \$31.5 million for the local economy. In 1980 the mine closed after producing about 36 million tons of pelletized iron concentrate. Portions of the fresh water ponds have been developed into recreational fishing areas for the public.

Geology. The most complete geologic description of the Groveland deposit is by Cumberlidge and Stone (1964), two geologists with M.A. Hanna Mining Company, and was based on extended observations as the present pit was developed. They showed that the ore body formed the keel of a complex doubly plunging syncline that was overturned to the south. The thickest extent of the Vulcan Iron-formation was along the south limb as shown in the plan map and cross section, (Figure 8). The north overturned limb was capped by Cambrian sandstone prior to stripping, and initial underground development was probably in the iron-formation subcrop below the sandstone contact. The Vulcan Iron-formation is divided into three informal members at the Groveland mine (Figure 8): 1) the lower Vulcan consisting mostly of hematitic jasper, 2) the middle Vulcan is dominantly even-bedded magnetite-silicate iron-formation with lesser hematitic jasper, and 3) the upper Vulcan Iron-formation is uniformly bedded magnetite-silicate iron-formation. The mine is located in the staurolite zone of regional metamorphism, and iron silicates in the Vulcan Iron-formation are commonly coarse-grained reflecting that intense recrystallization. The most common silicate minerals identified by Cumberlidge and Stone (1964) are Ca-Mg hornblende, tremolite, and actinolite. Pyroxene, biotite, and cummingtonite are less abundant and garnet is rare. Some representative photographs of the geology of the mine area is shown in Figure 7.

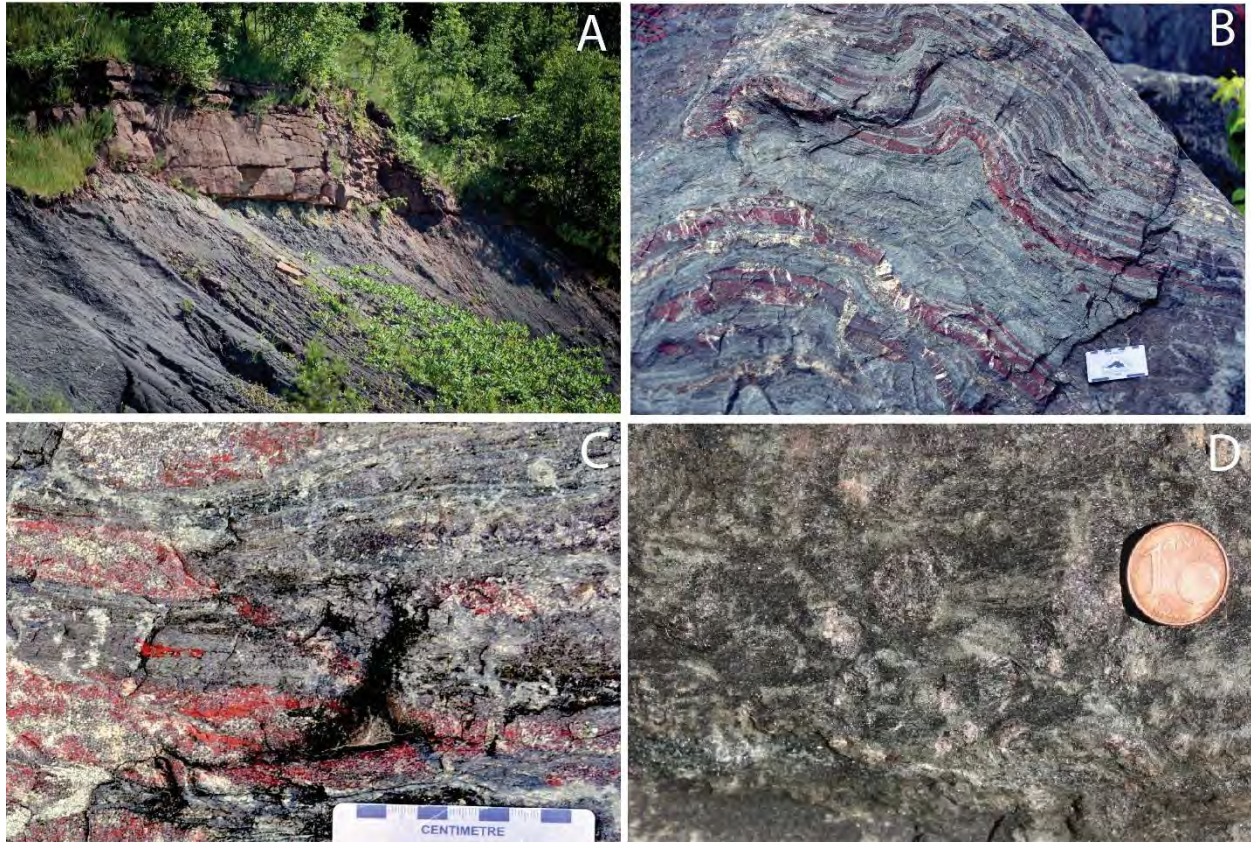


Figure 7. Photographs from the Groveland mine. A- Unconformity between flat-lying sandstone of the Munising Formation and weathered Vulcan Iron-formation on the north wall of the Groveland pit. B- Folded jasper-specularite iron-formation, C-Lenticular beds of oolitic jasper with specularite interbeds, D-Silicate-magnetite iron-formation with large sheaves of amphibole.

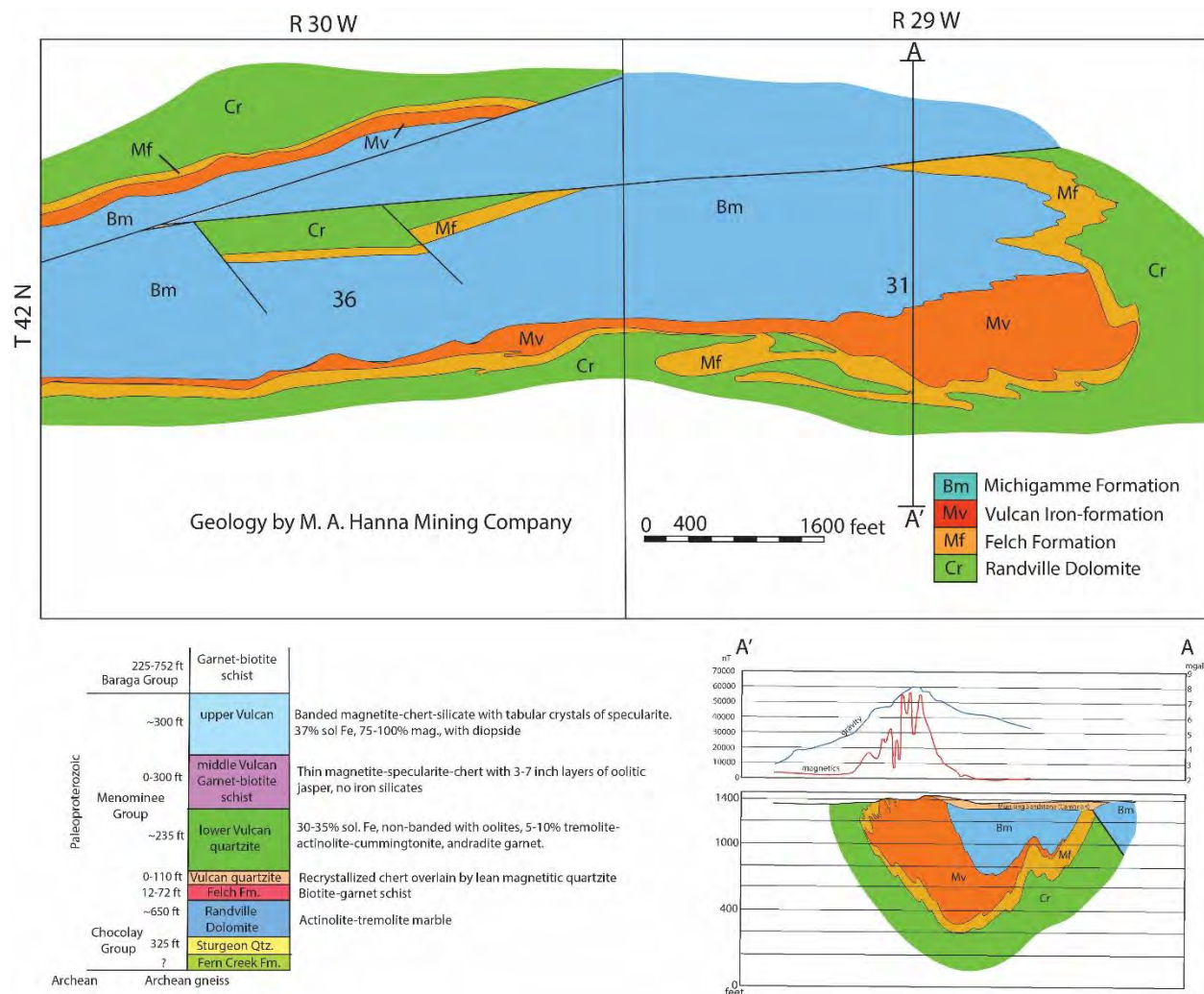


Figure 8. Geologic map and cross section (with magnetic and gravity profiles) of the Groveland mine area. The Groveland pit was developed in the thickest part of the Vulcan Iron-formation in the eastern part of the map. Source: M.A. Hanna Mining Company from archives of the Michigan Department of Environmental Quality. Stratigraphic section summarizes descriptions of the Vulcan Iron-formation and related units as reported in Cumberlidge and Stone (1964) provided by Thomas Waggoner (personal communication). Unit names are informal mine terminology.

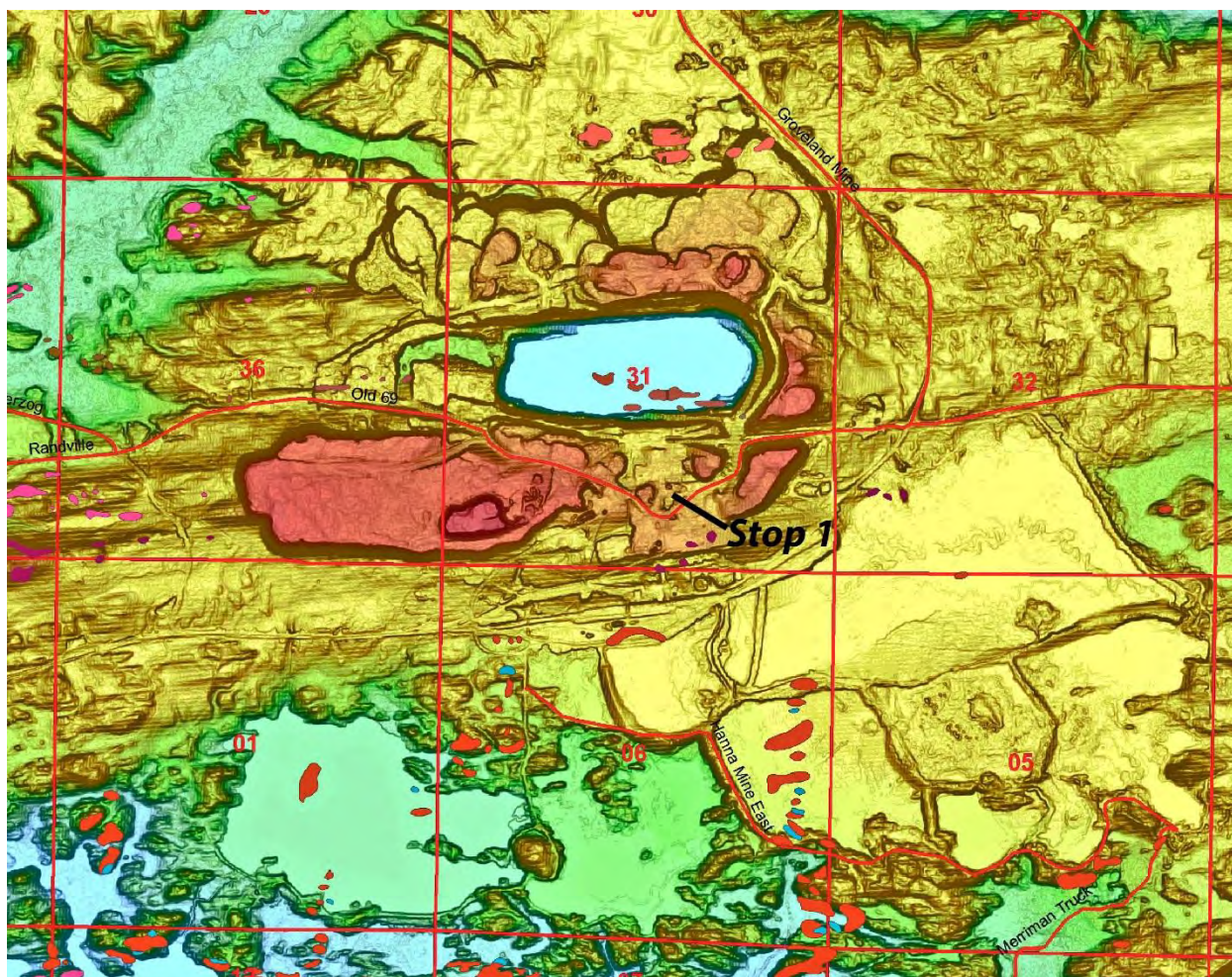


Figure 9. False color LiDAR image of the area of the Groveland mine showing the location of Stop 1 and outcrops as mapped by James et al. (1961) prior to mine development.

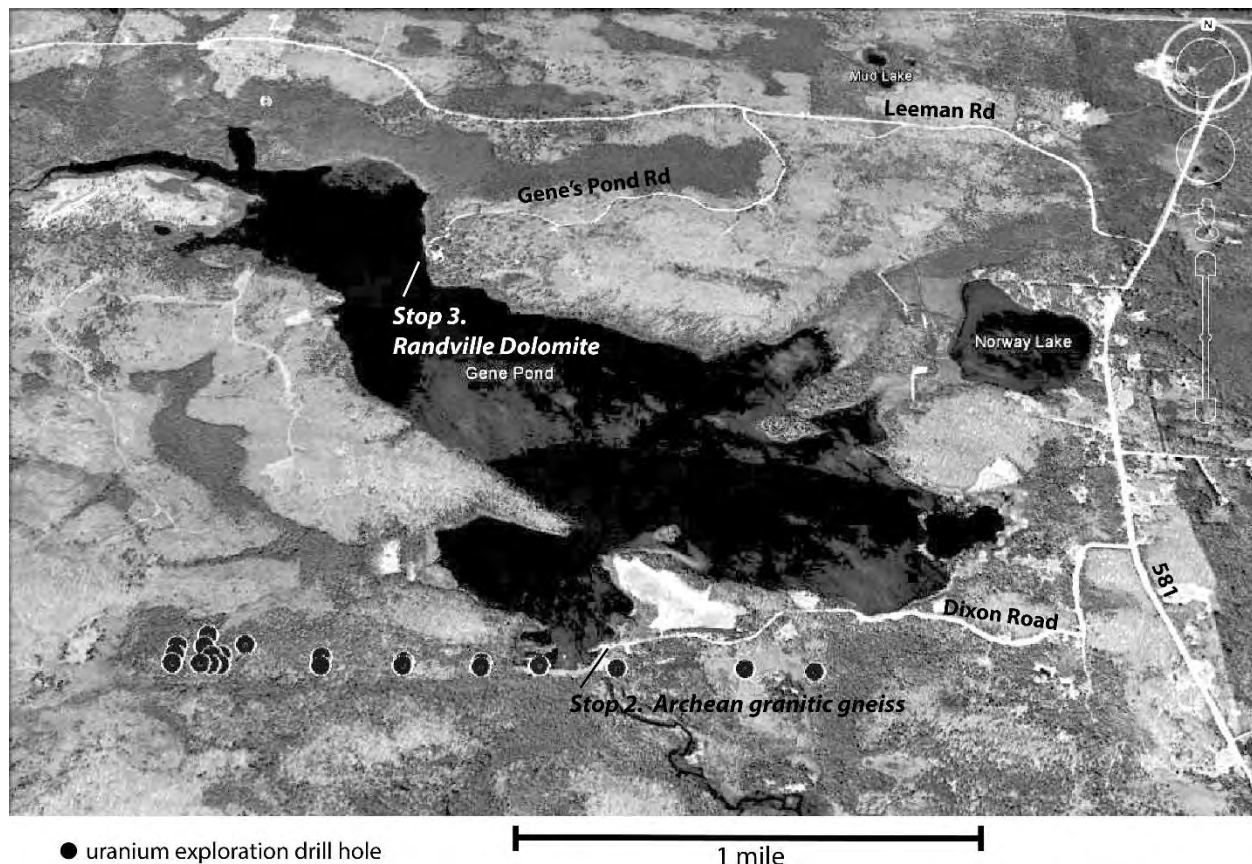


Figure 10. Location of Stops 2 and 3 and the location of holes drilled for uranium exploration near Stop 2. Image from Google Earth.

Stop 2. Archean gneiss at Gene's Pond. (46.058° N, 87.855° W) At the boat launch site at the west end of Dixon Road are several outcrops of Archean granitic gneiss and mafic dikes that cut the gneiss. The granitic rocks are mostly plagioclase porphyritic rocks with a moderately to well developed, nearly vertical shear foliation (Figure 11A). The mafic dikes cut the foliation. They appear largely massive and undeformed in outcrop but are thoroughly amphibolitized and the amphiboles show a weak alignment in thin section (Figure 11B). Based on landforms, we interpret that there are outliers of Munising Formation (Upper Cambrian sandstone) both east and west of this locality, and further interpret that the present land surface here is very nearly the exhumed unconformity at the base of the Cambrian. The unusual reddish hue of much of the granite may be a reflection of weathering or alteration along the unconformity (see Figure 11A and 12).

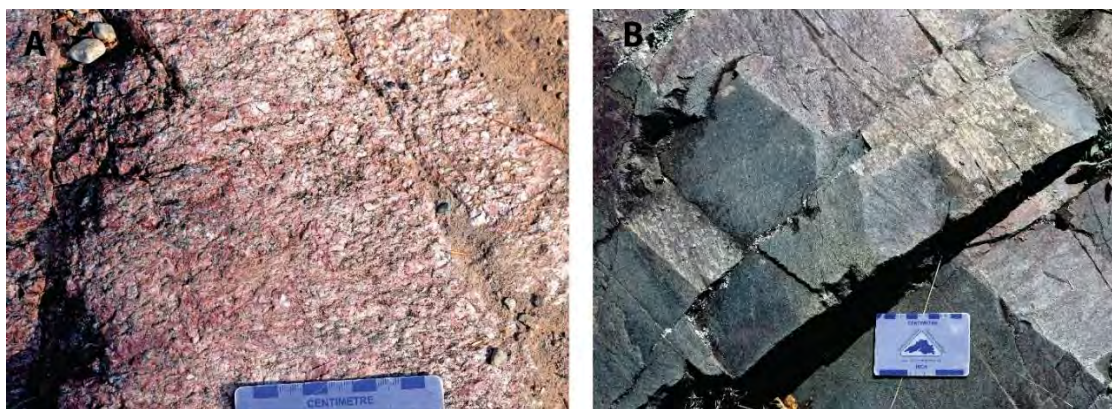


Figure 11. A- Moderately sheared Archean granite. B- Massive mafic dike with blocky fracture.

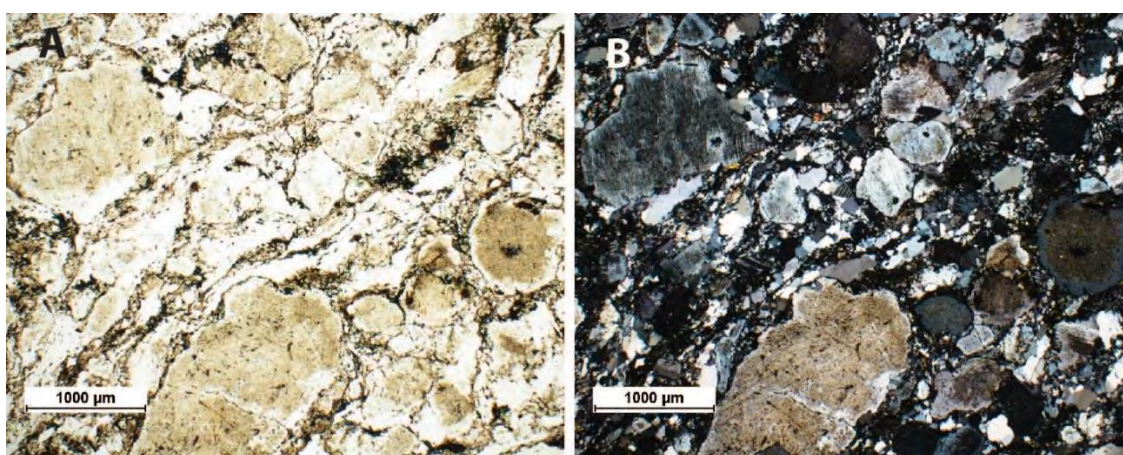


Figure 12. Photomicrographs of sheared granitic rock at Stop 2. Rock is mostly plagioclase with moderately developed cataclastic textures. Nearly all plagioclase grains are stained with submicroscopic hematite (?). A- Plane polarized light, B- Crossed Nichols.

A rather extensive exploration effort for unconformity type uranium deposits was undertaken here in the early 1980's by Minatome Corporation (Hunter, 1986; Lehman, 1987). This included twenty shallow drill holes shown on Figure 10, one of which was only a few tens of meters south of these outcrops. Uranium, occurring as pitchblende, was found as open-space fillings along with calcite, hematite, and minor chlorite. The drilling defined an E-W brittle fault that dips about 60° north and is subparallel to an older mylonitic fault. The mineralized assemblage heals breccias that are most common in the hanging wall of the brittle fault. The drill holes were located to test the depth extent of surface radioactive occurrences but found that no mineralization extended more than 85 m below the surface and that the surface extent of mineralization of individual occurrences was about of equal vertical dimensions. U-Pb dating of the pitchblende (Lehman, 1987) yielded a range of results, all of which were of Paleozoic or younger age. A likely conclusion is that the mineralization formed in Paleozoic or younger times just below the unconformable contact of Cambrian sandstone and Archean gneiss. This exploration and its results have never been described in detail, but drill logs and core for most of the holes are available for study at the Michigan Geological Sample Repository in Harvey, Michigan.

Stop 3. Randville Dolomite at Gene's Pond. (46.072° N, 87.866° W.)

A small lakeside outcrop just south of the boat launch at Gene's Pond public access site displays many of the typical features of the Randville Dolomite in the northern part of the Felch area. In much of the Felch area strong metamorphism has converted the Randville Dolomite into coarsely crystalline white to gray marble. Much of the primary structure is obliterated. However, Stop 3 lies north of the highest grade metamorphic zones and metamorphic recrystallization is only minor with abundant primary features preserved. Here, the Randville underlies a large area between the Bush Lake fault on the north and the Norway Lake fault on the south. It has been described in some detail by Clark (1961, Chapter C of James et al., 1961). Unfortunately, the more illustrative Randville outcrops described in detail by Clark are not easily accessible for field trips. The small outcrop at Stop 3 is shallowly dipping and well bedded dolomite with undulose, generally upward-domed, bedding that is likely stromatolite structures (Figure 13)



Figure 13. Randville Dolomite at Stop 3 showing undulose bedding, probably reflecting weakly developed stromatolitic mounds.

The general description of the Randville provided by Clark (1961, p. 107-109) is “*The Randville dolomite is apparently divisible into three members: (1) an upper member and (2) a lower member of dolomite with minor interbedded slate, separated by (3) a slate member with minor interbedded dolomite.... The total thickness is more than 800 feet in the Norway Lake area. Neither the top nor the bottom of the formation is exposed, and the character of the rocks that immediately underlie and overlie the Randville dolomite is not known. Most of the dolomite*

is massive to thin bedded, and stromatolites (algal structures) and intraformational conglomerate are common. The dolomite is light gray to red on fresh surfaces and weathers white to light brown. It has a fine sugary texture. Grains of quartz sand, most of which show undulatory extinction, are abundant in some beds and in some places comprise more than 50 percent of the rock. No oolites were found.

The stromatolites, in sections normal to bedding planes, are concentrically banded structures with domal or columnar form, and in sections parallel to bedding planes they are concentrically banded elliptical forms. Most are 1 to 3 inches in diameter and 2 to 6 inches high. The banding of the stromatolites is convex upward, providing a reliable criterion for tops of beds. Where the structures are partly replaced by chert the forms are accentuated on weathered surfaces.

Intraformational conglomerates or breccias are present Most of the pebbles are dolomite, but a few pebbles of dolomitic slate occur. The pebbles are 1 to 4 inches in diameter and are well rounded. No strong dimensional orientation is evident. The matrix is dolomite with intermixed coarse quartz sand."

The slates, as described by Clark (1961) are dark gray to gray-green and are composed largely of sericite with some quartz, chlorite, and microcline. Graded beds are common.

Stops 4, 5, and 6. The Dickinson Group

The three formations originally defined as the Dickinson Group (James et al., 1961), 1- the East Branch Arkose (Stop 4), 2- the Solberg Schist (Stop 5), and 3- the Six Mile Lake Amphibolite (Stop 6) will be examined from north to south, the originally interpreted stratigraphic order.

Stratigraphy and age- The Dickinson Group was originally defined by James et al. (1961) to include three formations, from presumed oldest to youngest, the East Branch Arkose, the Solberg Schist, and the Six Mile Lake Amphibolite, which they interpreted to be in conformable contact with each other. The East Branch Arkose is a sequence of arkosic conglomerate and sandstone with interbedded mafic volcanic rocks. The Solberg Schist consists of finer clastic and volcanic rocks with at least one interbedded banded iron-formation, the Skunk Creek Member. The Six Mile Lake Amphibolite was interpreted to be highly metamorphosed mafic volcanic rocks with abundant granitic intrusions. James et al. (1961) provided detailed geologic maps and descriptions of each unit. This discussion is based largely on their work. Each unit has a maximum thickness of 2,000 to 4,000 feet, and James et al. (1961) state that a total thickness of 10,000 to 12,000 feet for the group is indicated.

This original work was done before radiometric ages were available so relative ages were based on field relationships. The supposed lower formation of the group, the East Branch Arkose, seems clearly to lie unconformably on Archean gneisses to the north as indicated by numerous gneissic pebbles in conglomerates of the East Branch Arkose. To the south, James et al. (1961) believed that the Six Mile Lake Amphibolite was intruded and altered by a large batholith of Archean age, so ascribed a late Archean age to the group.

More recently, the acquisition of radiometric data has modified the permissible age range for the Dickinson Group. All units of the group were recrystallized during regional metamorphism related to the Peavy metamorphic node (James, 1955). That metamorphism has been dated at

approximately 1.83 Ga, the age of metamorphic growth of monazite (Holm et al., 2007), thus providing a minimum age for the group. Detrital zircons reportedly from the East Branch Arkose show a spectrum of ages (Craddock, et al, 2013) that includes a strong, well defined, peak at approximately 2.1 Ma which would establish a maximum age for the East Branch. Unfortunately the coordinates given by Craddock et al. (2013) for the sample correspond to a roadcut of Solberg Schist, not East Branch Arkose, putting some uncertainty on interpretation. But if the Solberg is conformable with the East Branch, as interpreted by James et al. (1961) the results still provide a constraint for the group as a whole even if the sample is from the Solberg. Thus, the available radiometric age constraints place the East Branch Arkose and possibly Solberg Schist sedimentation and volcanism within the range of 2.1 to 1.83 Ga, similar to other Paleoproterozoic strata of the region.

We suggest a reinterpretation of the Dickinson Group in which it is entirely Paleoproterozoic. The age of the Six Mile Lake Amphibolite is problematic in this interpretation in that James et al. (1961) described a southward transition of amphibolites into gneisses that are clearly of Archean age. Whether these amphibolites are a part of the Six Mile Lake or some older sequence is not clear at present and requires further evaluation.

Lithology- The following lithologic descriptions are summarized entirely from the detailed descriptions provided in James et al. (1961).

East Branch Arkose: As described by James et al., (1961. p. 13-14), “The formation consists of thick-bedded arkose with many beds of coarse conglomerate, interbedded with metamorphosed tuffs and basic volcanic flows. The conglomerates, though not the dominant rock type, are the most striking feature of the formation. The beds typically are 10 to 30 feet in thickness. The pebbles in the conglomerate have been drawn out into lenses that, on a horizontal surface across the nearly vertical beds, have a length-to-width ratio of about 3:1. In most parts of the area this shearing is parallel to bedding, that is, eastward but in a few places it is at an angle. Linear structure is not pronounced; most of the flattened pebbles have a length in vertical section about equal to that in horizontal section. In a few places a nearly vertical linear structure marked by grooving of the pebbles is evident. Vitreous quartzite is the dominant rock type among the pebbles, with granite gneiss, slate or schist, and quartz being of lesser abundance. Some of the quartzite pebbles show well-defined bedding. The arkose is pink to gray, massive, and abundantly cross-bedded. In many outcrops and in hand specimens it closely resembles a granite gneiss, but the well-defined crossbedding is complete proof of its sedimentary origin. In the more southerly outcrops of the East Branch arkose, dark-gray fine-grained tuffs are interbedded with the arkose; the rock consists principally of quartz and untwinned feldspar, with scattered grains of epidote, biotite, and carbonate. Rounded grains of opalescent quartz are present in some layers. Metabasalt flows are not uncommon. The rock is black and hornblende-rich. In the outcrops in the NW sec. 17, T. 42 N., R. 28 W., the originally scoriaceous top of one of these flows can be seen. Some of the metamorphosed flows are moderately magnetic and give rise to the aeromagnetic anomalies shown on the general map of the district.”

Solberg Schist- James and et al. (1961, p. 17-18) describe the Solberg Schist as follows. “The Solberg schist lies immediately south of the East Branch arkose. A considerable amount of interbedding of arkose and schist is evident in the outcrops, so that location of the contact between the units is somewhat arbitrary. . . . The more northerly exposures of the unit consist

chiefly of dark fine-grained hornblende and biotite schists. Locally, muscovite is an abundant constituent. Some outcrops show a banding, essentially parallel to the foliation, which may represent original layering. In one place, near the south edge of sec. 13, T. 42 N., R. 29 W., the schist is interbedded with coarse clastic material similar to rock of the East Branch arkose. The more southerly exposures of the Solberg schist consist of quartz-mica schist, parts of which might be better termed micaceous quartzite. This rock is exposed in the north part of sec. 24, T. 42 N., R. 29 W., and secs. 21 and 2, T. 42 N., R. 28 W. In general, the rock is massive, gray, and well banded. The banding, which consists of alternation of quartzitic and micaceous layers, almost certainly represents original bedding. Linear structure is strongly developed in all the schists, especially in the hornblendic varieties. It is marked by orientation of hornblende needles and by biotite. The lineation is in the plane of foliation and in general plunges eastward at a low angle."

The Skunk Creek member of the Solberg schist is a bed of iron-formation. This bed gives rise to a very strong magnetic anomaly, by means of which the iron-formation can be traced in a belt across most of the map area. . . . The Skunk Creek member has been penetrated by drill holes in several places. It is chiefly from this drill core that information concerning the lithologic character has been obtained, although none of the holes has cut the entire unit. The distinctive part of the formation is a thin-banded rock consisting of alternating layers of granular quartz (probably originally chert), magnetite, and various mixtures of hornblende, biotite, grunerite, garnet, and epidote. This material grades into biotite-hornblende schist, containing magnetite-rich layers, by interbedding at both the upper and lower contacts. The iron-formation is cut by many thin dikes of coarse pegmatite; garnet, and tourmaline are commonly developed near the contacts. The thickness of the Skunk Creek member is somewhat uncertain because of the small amount of data available, but it is about 100 feet."

Six Mile Lake Amphibolite- James et al. (1961, p. 18-19) described the Six Mile Lake Amphibolite as follows. *"In outcrops the amphibolite is a dark almost black massive fine- to medium-grained rock in which hornblende is the major constituent. In thin section feldspar (andesine or oligoclase) is abundant; in hand specimens it is less noticeable but gives a faint salt-and-pepper appearance to the rock, especially on surfaces broken across the foliation. Compositional layering is evident in some places, but in general the rock is homogeneous. The more southerly outcrops approach banded gneiss in character. Foliation parallel to the compositional layering is generally present, but may be subordinate to a strong linear structure that characterizes the rock. The lineation, which is marked by orientation of hornblende needles, plunges eastward at a low angle. In almost every outcrop the amphibolite is cut by dikes, pods, or irregular bodies of younger pegmatite."*

Structure- The Dickinson Group appears to form a thick south-facing monocline that dips vertically to steeply southward. Virtually all the numerous stratigraphic top determinations from cross bedding in the East Branch are south facing and show no indication of fold repetition (James et al., 1961). Such determinations are lacking in the Solberg Schist, but based on the presumed conformable relationship to the East Branch, at least the lower portions of the Solberg seem likely to be south-facing. A series of magnetic anomalies, most notably that produced the Skunk Creek Member, can be followed for many kilometers as markedly straight traces,

indicating that large scale folds within the Solberg are lacking on the scale of the James et al. study and that it, like the East Branch, is entirely south-facing.

On a smaller scale, a strong schistosity is developed in both units and outcrop-scale folds with shallow plunges are fairly common. Whether these features reflect a much larger structure, only partly preserved, or are the largest-scale structures that formed as the units were rotated to their near-vertical orientation is a matter for speculation with our present understanding of the region.

Correlation and tectonic setting of deposition- The radiometrically indicated age range of the Dickinson Group suggests that it is a possible time-correlative of the Menominee and/or Baraga Groups exposed nearby and across the southern Lake Superior region. Volcanic rocks interlayered with iron-formations of the Menominee Group were formed at 1.875 Ga (Schneider et al., 2002) and a diabase sill intruded into iron-formation in the Marquette iron range has been dated at about 1.890 Ga (Peitrzak-Renaud and Davis, 2014). The ejecta layer from the Sudbury impact at 1.850 Ga lies near the upper contact of the Menominee Group and the base of the overlying Baraga Group (Cannon et al., 2010). Post tectonic granite plutons that intruded the Baraga Group at 1.833 Ga (Schneider et al., 2001) provide a minimum age. It is possible, therefore, that the Dickinson Group was also deposited during or slightly before the Menominee-Baraga sequence, but its lithology and apparent tectonic setting are unmatched by any other sequence in the region.

The combined stratigraphic thickness of 2-3 kilometers of predominantly clastic sediments with interlayers of mafic volcanic rocks argues for deposition in a rapidly subsiding basin in which a generally fining-upward sediment sequence was deposited. The East Branch Arkose, based on lithology and bedding features, seems clearly to be fluvial. Yet, the occurrence of banded cherty iron-formation in the Solberg Schist argues that the basin evolved into marine conditions by the later parts of deposition of that formation. It seems reasonable that the Dickinson Group was deposited in an extensional rift basin in the back-arc basin phase of the Penokean orogeny at about 1875 Ma as defined by Schulz and Cannon (2007).

Stop 4. East Branch Arkose. (46.044° N, 87.840° W) Good exposures of the East Branch Arkose can be seen just west of Spring Hill Road. Please respect private property immediately south of the area examined by this trip. The area was mapped in detail (James et al., 1961) and a portion of the map is reproduced in Figure 14. All lithologies are exposed here including coarse, cross-bedded arkose (Figure 15A), quartzite pebble conglomerate (Figure 15b, c, d) and massive basalt. They can be seen in sequence along a short south-to-north transect. All units have near-vertical dips and face south. Slightly to the west, a set of diabase dikes cuts the units at a low angle. Of interest in the conglomerates is the great preponderance of pebbles of white to pink quartzite, a small percentage of which have well preserved bedding. The only known source for such clasts in the region is the Sturgeon Quartzite, which is the basal unit of the Paleoproterozoic sequence in most of the area. Pebbles of granite are common but make up only a small percentage of the pebble-sized clasts. They were used by James et al. (1961) as further evidence of an Archean source, but considering the new age for the porphyritic red granite of 2.1 Ga, the clasts could be from that unit rather than the Archean. Quartzite pebbles typically have flattened shapes that are more likely to be original shapes rather than caused by deformational flattening. A suggestion of imbricate structures can be seen locally. The abundant sand-sized grains of K-

feldspar in the arkosic beds further indicate a granitic (Archean or porphyritic red granite?) source for much of the formation.

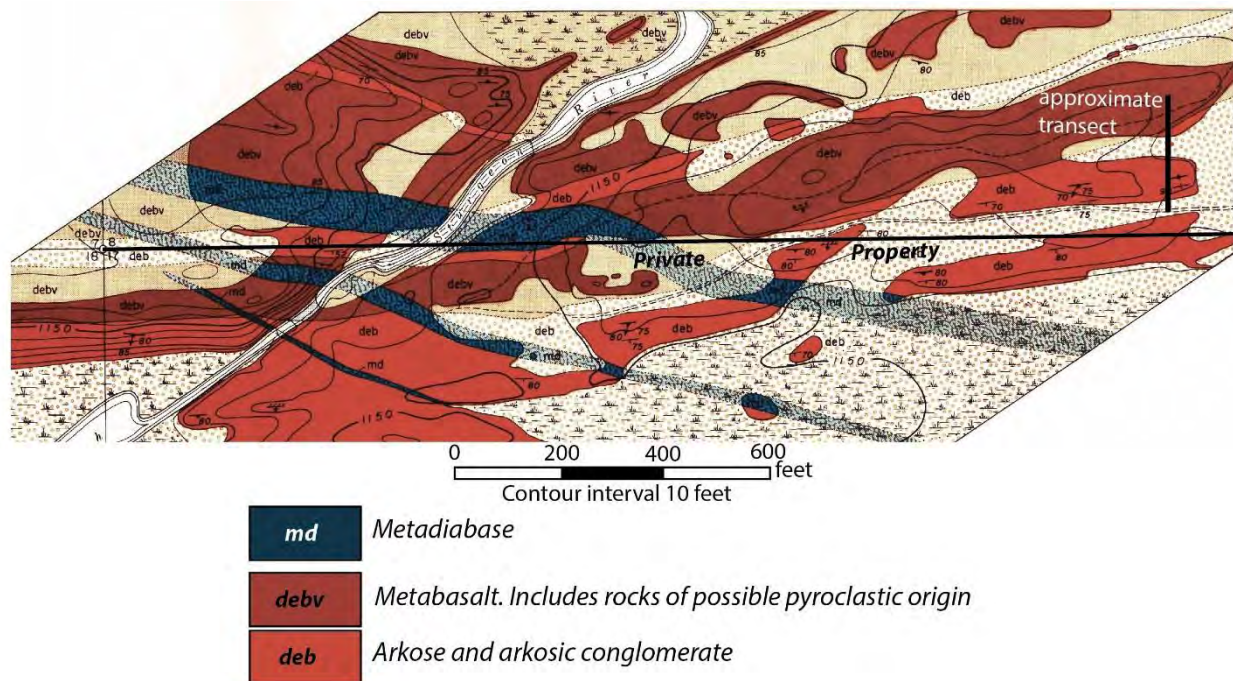


Figure 14. Map of the outcrop area of East Branch Arkose from James et al., (1961). Darker shades are areas with abundant outcrops; lighter shades are covered. The approximate transect for the field trip is shown near the east edge of the area and is about 150 meters west of Spring Hill Road.

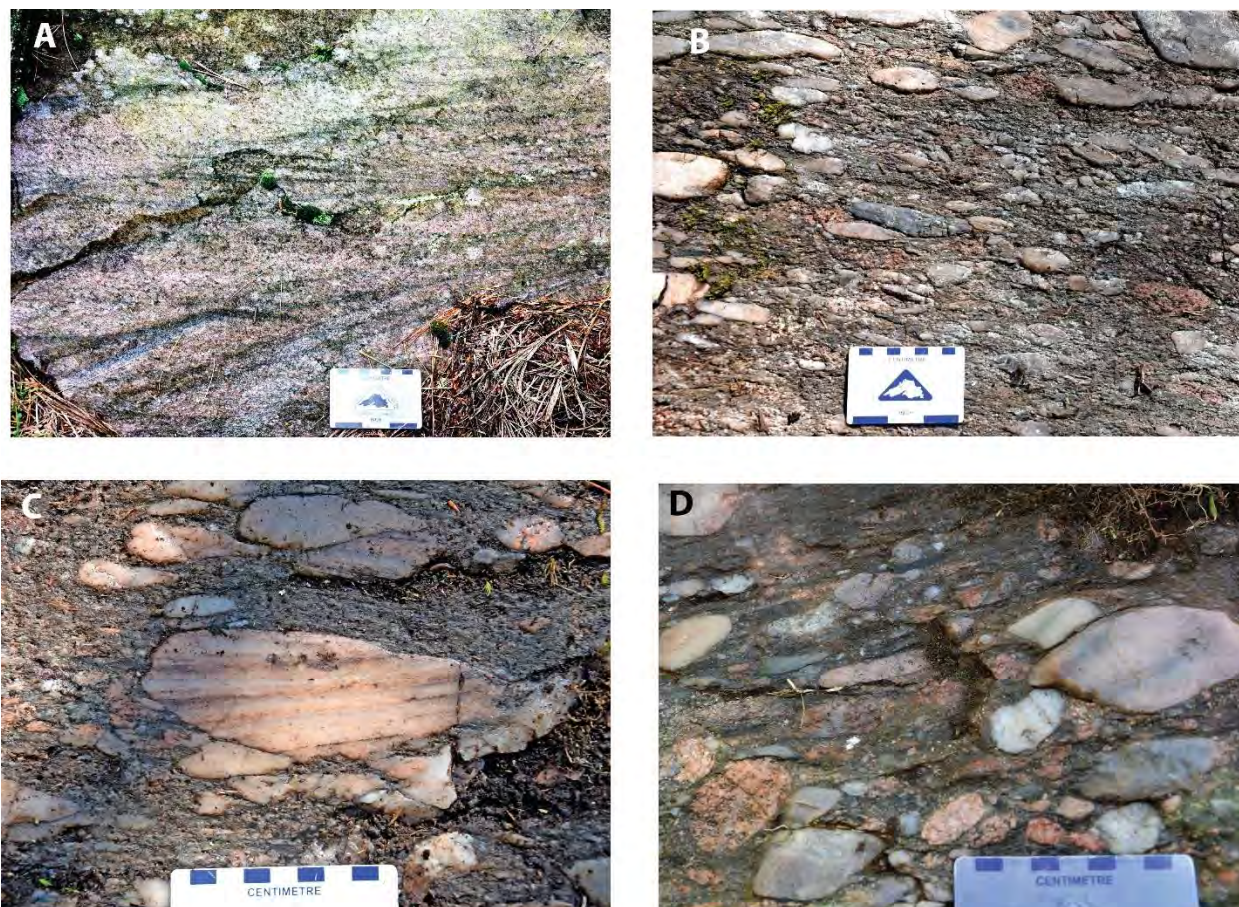


Figure 15. Photographs of the East Branch Arkose. A- Cross-bedded coarse-grained arkose. B- Conglomerate consisting mostly of quartzite pebbles. Imbricate pebbles indicate current flow from right to left. C- Conglomerate with clast of quartzite with relict bedding. D- Close-up of conglomerate containing both quartzite and granite pebbles.

Two samples of the basalt (amphibolite) collected along strike have very similar tholeiitic basalt compositions ($\sim 7\%$ MgO, $\sim 50\%$ SiO₂) with intermediate TiO₂ (~ 1.4 - 1.9%) and FeO_{total} ($\sim 12.5\%$). The trace elements are characterized by moderately enriched light rare earth elements (REE) and no negative Nb-Ta anomaly when normalized to primitive mantle (Figure 16). A sample of one of the dikes cutting the Archean granitic gneiss at Stop 2 has a trace element content identical to the basalt in the East Branch Arkose except for an enrichment in Th (Figure 16B); it also has higher SiO₂ ($\sim 54\%$) than the basalt and may have been contaminated by the granitic gneiss. An amphibolite sampled in a road cut on the west side of Felch also has a composition very similar to the basalt in the arkose (Figure 16). The East Branch mafic rocks are very similar in composition to basalts found in continental rifts.

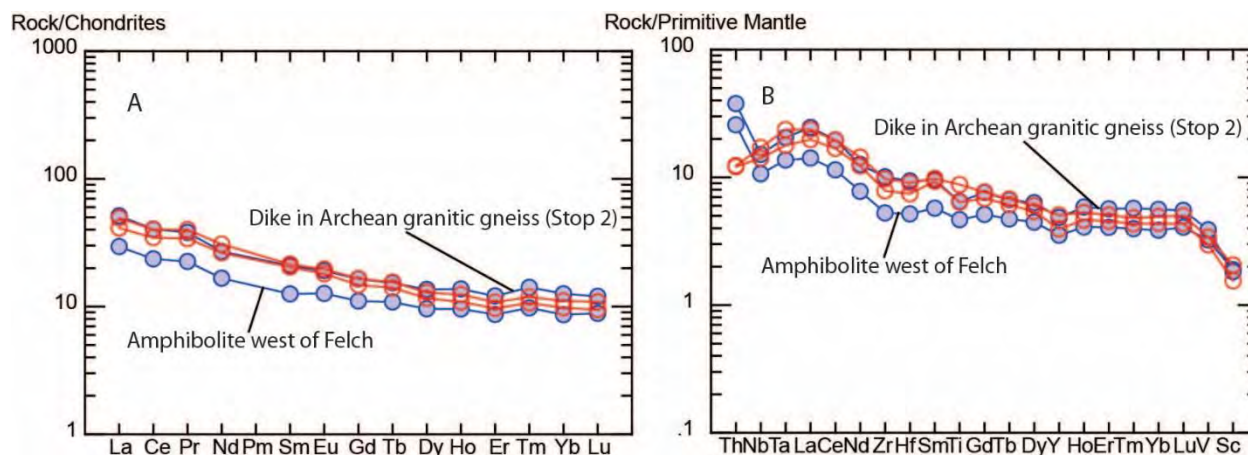


Figure 16. Chondrite normalized REE plot (A) and primitive mantle normalized trace element plot (B) of basalt in the East Branch Arkose, a metadiabase dike in Archean granitic gneiss at Stop 2, and an amphibolite (sill?) from a road cut on the west side of Felch.

Stop 5. Solberg Schist. (46.04°N, 87.82°W) This roadside outcrop on west side of County Road 581 is representative of much of the Solberg Schist. The rock is interlayered amphibolite and biotite-garnet schist with a prominent near-vertical foliation. The unusual ribbed appearance of the outcrop surface may reflect subtle bedding although lithologic changes across the ribs are not obvious (Figure 17B). If they are bedding, the outcrop displays a westward-plunging antiform. Radiometric data on detrital zircons are reported by Craddock et al. (2013) from a location whose coordinates correspond to this outcrop. The spectrum of ages ranges from about 3.8 Ga to about 2.1 Ga. The significant peak of ages at 2.1 Ga places a maximum age on the unit. Unfortunately the sample was described as East Branch Arkose by Craddock et al. (2013), so there is some uncertainty as to where the dated sample was collected and what it represents. This outcrop, and many others, show intense, small-scale deformation structures and commonly near-vertical foliation and bedding. Although these suggest that the unit is complexly deformed and may include significant repetition of stratigraphy, the disposition of the Skunk Creek Member is enigmatic in that it has a nearly straight outcrop trace for more than 20 kilometers (see Figure 2) and shows no indication of fold repetition.

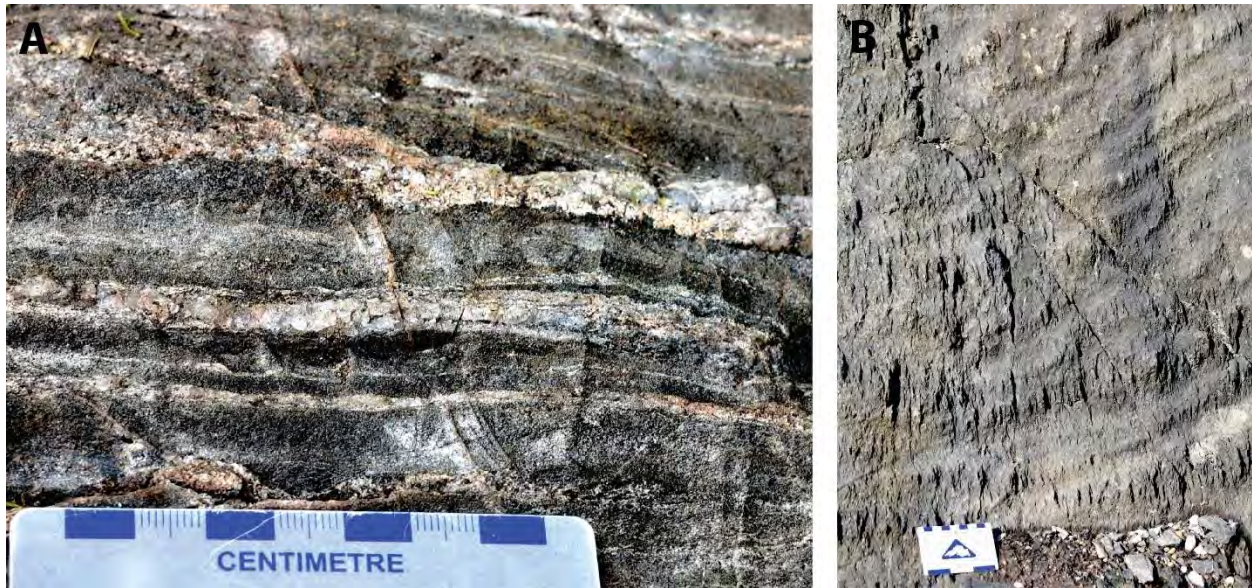


Figure 17. Outcrop photographs of the Solberg Schist. A-Well-bedded schist with interlayers of micaceous schist and more quartzose layers. Thin stringers of granite in upper half. B- Shallowly-dipping beds (?) cut by vertical foliation.

Compositionally, the Solberg mafic schist ranges from basalt (~6-12% MgO, ~45-50% SiO₂) to andesite (~4% MgO, ~56% SiO₂) with relatively high TiO₂ (~1.7-2.4%) and FeO_{total} (~13-16%). Samples have steep light REE-enriched chondrite normalized patterns and negative Nb-Ta anomalies when normalized to primitive mantle (Figure 18). They are compositionally distinct from the amphibolites in the East Branch Arkose and the Six Mile Lake Amphibolites.

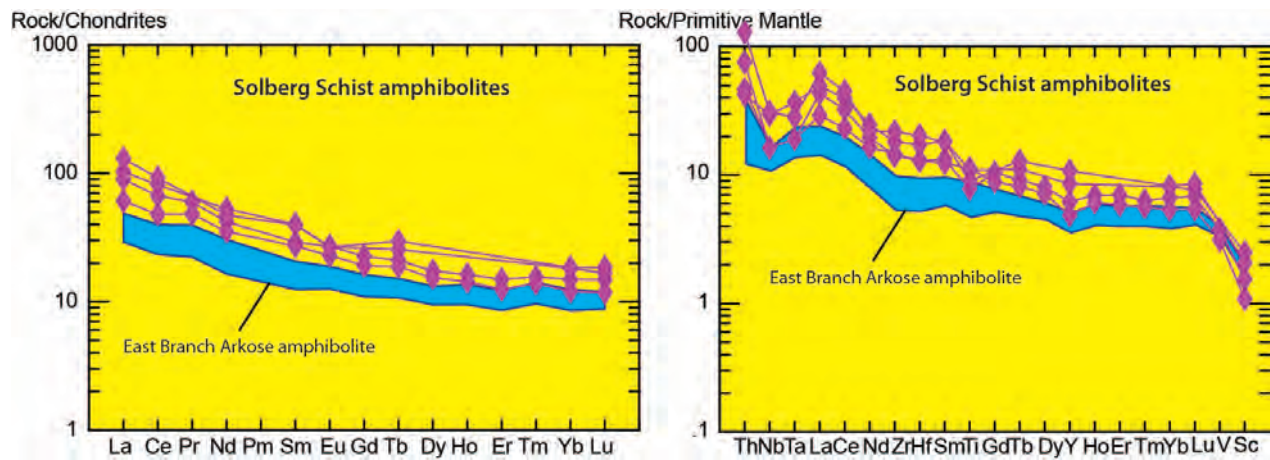


Figure 18. Chondrite normalized REE plot (A) and primitive mantle normalized trace element plot (B) for samples of the Solberg Schist. Field for basalt of the East Branch Arkose and related amphibolites (shown in Figure 16) shown for comparison.

Stop 6. Six Mile Lake Amphibolite. (46.02°N, 87.846°W) A small exposure on the west side of Wickman's Marsh Road is typical of the Six Mile Lake Amphibolite. Here it is a rather uniform schistose amphibolite with small deformed granitic stringers. It is cut by two undeformed pegmatite dikes (Figure 19) The Six Mile Lake Amphibolite was originally described and named by James et al. (1961), as summarized above, who placed it as the uppermost formation in the Dickinson Group and ascribed an Archean age. As also discussed above, that is now in question because of 2.1 Ga detrital zircons in the lower part of the group. At least a portion of the rocks included in the Six Mile Lake by James et al. (1961) are intruded by granitic rocks of Archean age and appear to grade southward into banded gneiss the makes up much of the Archean in that area. Whether these latter amphibolites are truly a part of the Six Mile Lake or are an older amphibolite unit that lies adjacent to the Six Mile Lake is a subject for further evaluation.

Compositionally the Six Mile Lake Amphibolite is a tholeiitic basalt characterized by low TiO_2 (<1.5 wt. %) and trace element content. Unlike most of the amphibolites in the region, which are characterized by enriched light REE and negative Nb and Ta anomalies when normalized to primitive mantle, the Six Mile Lake Amphibolite has a flat chondrite normalized REE pattern and no Nb and Ta anomalies (Figure 20). It should be noted that the large metagabbro body in the Solberg Schist has a similar composition to that of the Six Mile Lake Amphibolite (Figure 20). Amphibolites from the Carney Lake Gneiss complex and the Hardwood mafic gneiss also have compositions similar to the Six Mile Lake Amphibolite (Figure 20).



Figure 19. Six Mile Lake Amphibolite. A.- Hornblende schist with granitic stringers. B.- Schist cut by pegmatite dikes.

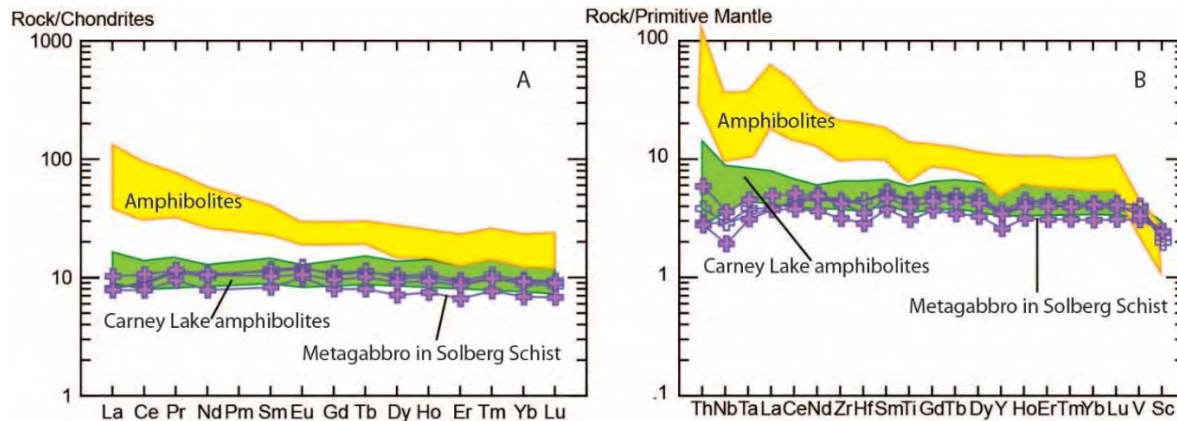


Figure 20. Chondrite normalized REE plot (A) and primitive mantle normalized trace element plot (B) for samples of Six Mile Lake Amphibolite and related rocks. Fields for some other amphibolites from the region shown for comparison

Stop 7A. Mafic granulite of the Hardwood Gneiss. (45.969°N, 87.711°W) Roadcuts on both the north and south sides of Highway M-69 are good examples of typical granulite of the Hardwood Gneiss. The rocks consist of various assemblages of hornblende, pyroxene, plagioclase, microcline, and garnet. They are very strongly foliated (Figure 21A) and, in places, show prominent compositional layering (Figure 21B). The Hardwood Gneiss was recognized and named by James et al. (1961) as an unusual unit of very highly metamorphosed rocks that are exposed over an area of only about 5 square kilometers at the eastern edge of their study area (see Figure 2). To the south, the Hardwood is in fault contact with the Paleoproterozoic Michigamme Formation. To the west and north it is in contact with Archean granite and granitic gneisses, but the nature of the contact is not known. The area of exposure is bounded on the east by Cambrian and younger strata. It is quite possible that the Hardwood underlies a substantially larger area beneath that cover. The general structure of the Hardwood in the area of exposure is the keel of a gently east-plunging synform defined by the gneissic foliation, so it is likely that the gneisses continue eastward in the pre-Cambrian basement.

James et al. (1961, p.22-23) described the Hardwood as follows “*In general, the gneiss is strongly layered, with individual layers ranging from a fraction of an inch to a few feet in thickness. The dominant rock type is a dark-medium-grained gneiss composed of hornblende, plagioclase, and pyroxene. Interlayered with this rock are beds of dark gneiss but of different grain size, beds of dark vitreous-lustered rock with alternating light and dark laminae, garnet-quartz-mica schist, and light-colored rock that resembles quartzite. Some of the layers are rich in magnetite. The gneiss appears to have been originally a volcanic sequence, at least in part tuffaceous, with some inter-bedded sedimentary rocks, intruded by gabbro sills. The rocks have been dynamothermally metamorphosed under conditions that resulted in the alteration of most of the original pyroxene in the igneous facies to hornblende and garnet, and the development of mica, hornblende, and plagioclase in rocks that appear to have originally been acidic volcanics. The Hardwood gneiss, as seen in the exposures, is folded along axes that plunge eastward at low angles, and the general structure appears to be an eastward plunging syncline.*”

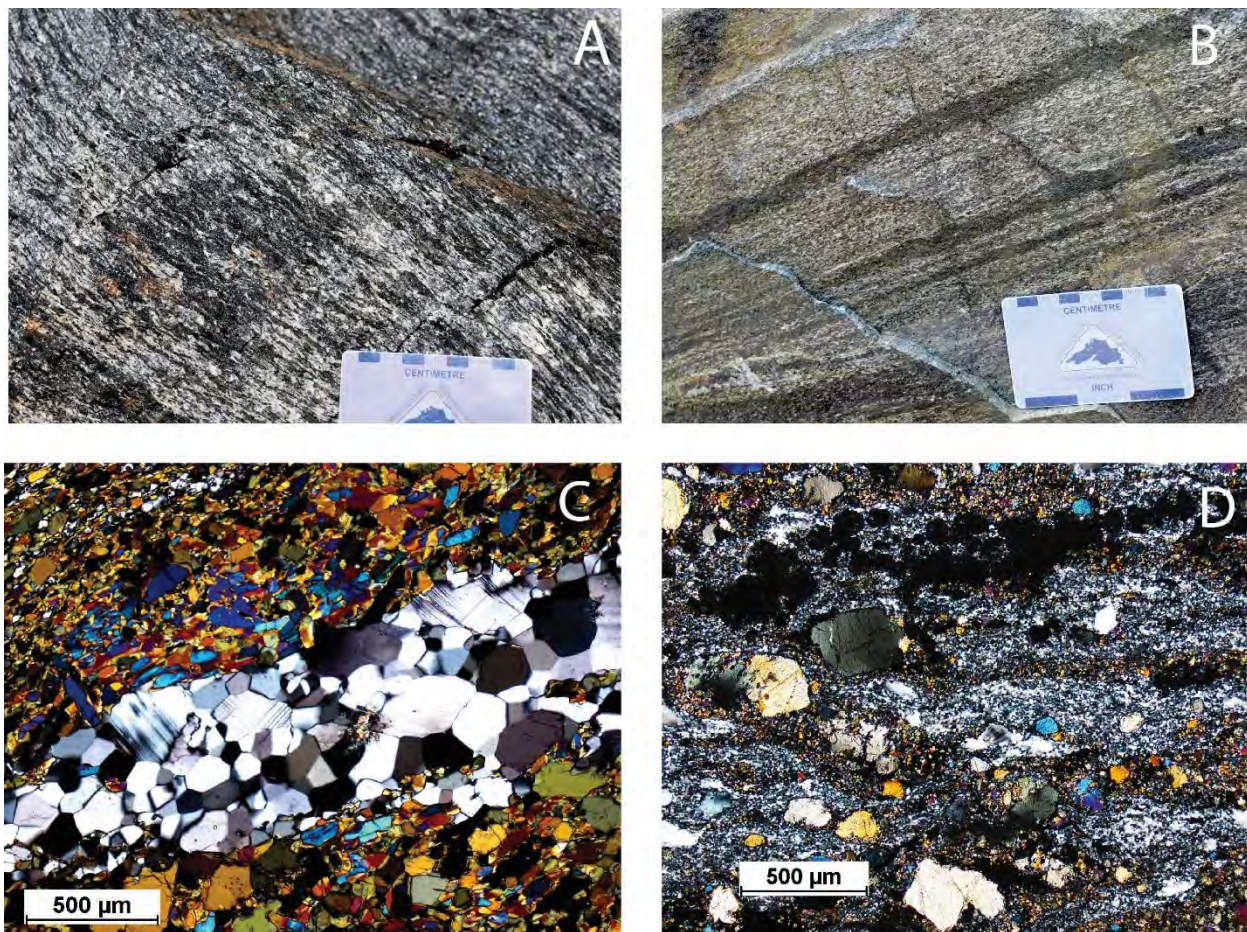


Figure 21. Hardwood Gneiss at stop 7A. A- Typical mafic granulite with strong, somewhat anastomosing, foliation. B- Straight-banded granulite gneiss. Compositional layering is expressed as variations in plagioclase:mafic mineral ratio. C and D are photomicrographs of two contrasting textures. C- Granoblastic textured interlayered quartz-microcline-plagioclase rock (center) and hornblende with minor (ortho?)pyroxene. D- Foliated rock with pyroxene (larger light grains) and small garnets (dark layer near top) in quartz-sericite matrix.

Additional lithologies that have been included in the Hardwood gneiss are metasediments (Stop 7B), including quartzite and pelitic schist. The schists include garnet-biotite-sillimanite-kyanite mineral assemblages.

Chemical analyses of mafic rocks in the Hardwood Gneiss are generally similar to those of the Six Mile Lake Amphibolite with similar low TiO_2 and trace element content (Figure 22). One mafic gneiss sample has enriched light REE and a large negative Nb-Ta anomaly is likely the result of contamination by felsic crustal rocks.

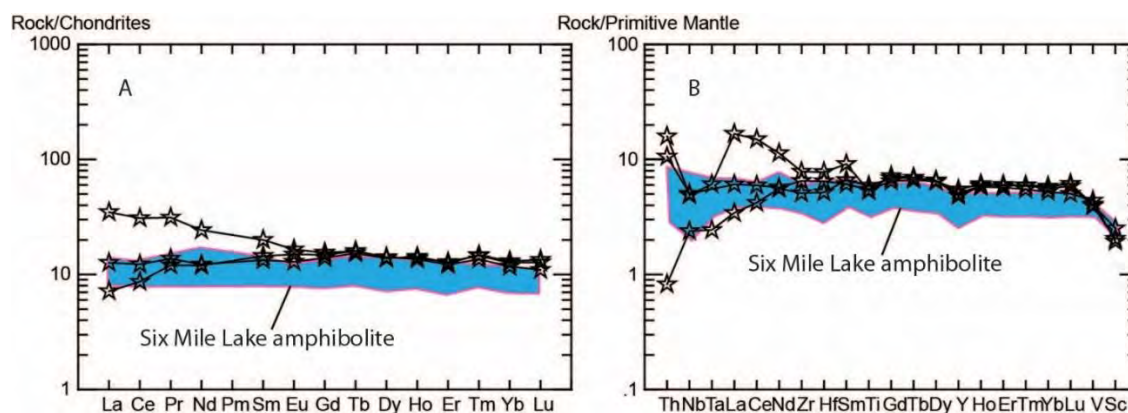


Figure 22. Chondrite normalized REE plot (A) and primitive mantle normalized trace element plot (B) for mafic gneiss samples from the Hardwood Gneiss. Field for Six Mile Lake Amphibolite shown for comparison.

Metamorphism of the Hardwood Gneiss was studied by Peterson and Geiger (1990) who determined conditions of its metamorphism based on mineral assemblages. They defined two distinct episodes of metamorphism. Geothermobarometry indicates conditions of 8.2-11.6 kbar and ~770°C for the earliest event, and conditions of 6.0-10.1 kbar and 610-740°C for the second. They proposed that the older event was Archean and contemporaneous with a high-grade metamorphic event recorded in the Minnesota River Valley. They interpreted the younger event as probably Paleoproterozoic and pre-Penokean, with metamorphic conditions more intense than those generally ascribed to the Penokean orogeny in Michigan. Although they recognized both events in the typical layered gneisses, only the second event is recorded by the metasedimentary units. These extremely high metamorphic pressures are unique in the southern Lake Superior region and are comparable to or slightly higher than those of the Kapuskasing structure in Ontario. For comparison, the metamorphic pressures interpreted for the Peavy node, the nearest Paleoproterozoic metamorphic node, are less than 5 kbars (Attoh and Klasner, 1989). The metamorphic conditions recorded by the Hardwood Gneiss are equivalent to temperatures and pressures of the lowermost crust.

The Hardwood Gneiss has been widely accepted to be of Archean age and to be part of the gneiss complex that forms the basement for the Paleoproterozoic sedimentary and volcanic sequence, although no radiometric ages have been previously determined. Our new SHRIMP U-Pb zircon data reveal a group of concordant to nearly concordant zircon spot analyses at ca. 2750-2500 Ma (Ayuso et al., 2018) documenting a Neoproterozoic component (Figure 23). A second group of spot analyses (Figure 23) document a younger period of zircon growth ca. 1900 to 2200. A thermal event of this age has not been recognized previously in the region. Metamorphism of the Peavy node, which encompasses the area of the Hardwood Gneiss, has been dated at ca. 1830 Ma at outcrops within a few kilometers west of the Hardwood (Schneider et al., 2004; Holm et al., 2007) using Pb-Pb ages of monazite. Furthermore, metamorphic pressures of the Peavy node are estimated at ca. 5 kb (Attoh and Klasner, 1988) in contrast to the much higher pressures determined for the Hardwood (Peterson and Geiger (1990)). An additional significant feature of the Hardwood Gneiss is that we have detected no Eoarchean components, unlike the nearby Carney Lake Gneiss, which has yielded numerous Eoarchean spot analyses. Thus, the Hardwood Gneiss is presently enigmatic in terms of its parentage, metamorphic history, and kinematics of emplacement relative to surrounding rock units.

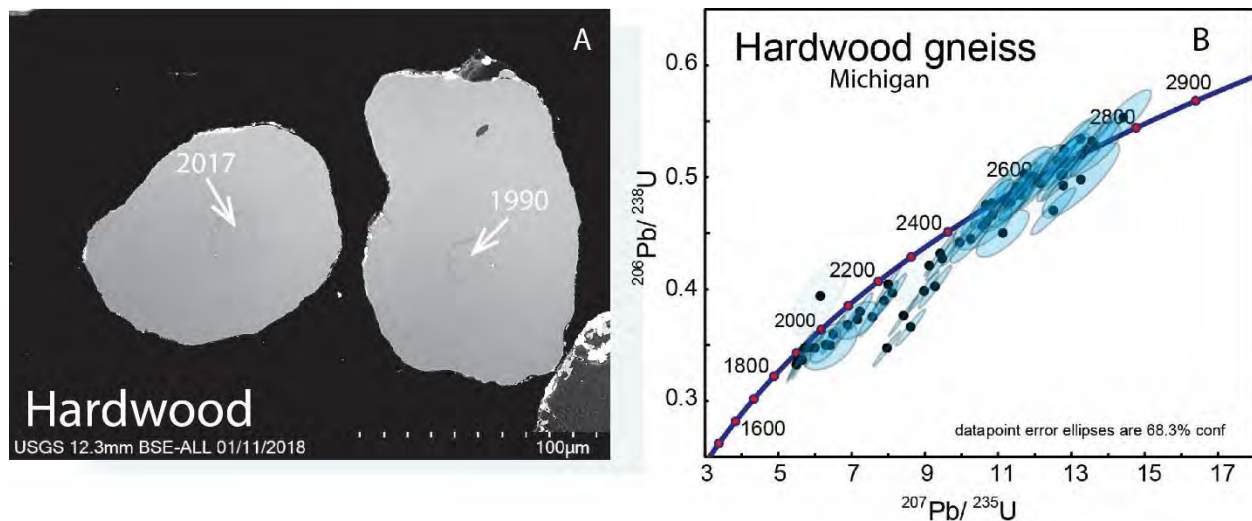


Figure 23. A- BSE (back-scatter electron) image of two anhedral zircons from the Hardwood Gneiss showing SHRIMP U-Pb analyzed dates for two spots. B- Concordia diagram for 56 spot analyses for the Hardwood Gneiss.

Stop 7B. Pelitic schist of the Hardwood Gneiss. (45.967°N, 87.729°W) A low roadcut on the west side of Swan Peterson Road, just north of its intersection with M-69 is garnet- biotite-sillimanite-kyanite schist. These are among the structurally (stratigraphically?) lowest parts of the Hardwood Gneiss. Garnet porphyroblasts are common (Figure 24A). A strongly developed foliation dips about 30 degrees east. White elongate masses (Figure 24B) consist of quartz and both kyanite and sillimanite. They are elongated within the foliation but have slightly flattened shapes in cross section (Figure 24B). They define a prominent lineation that plunges about 35° east.

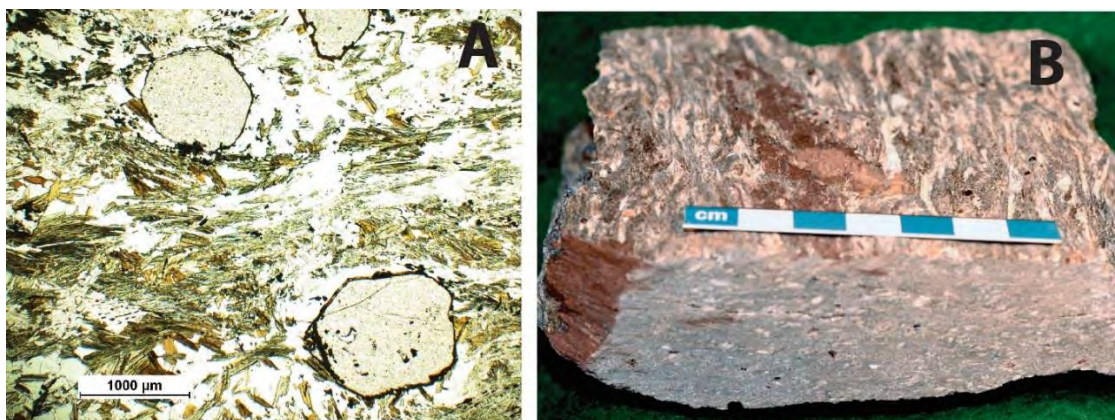


Figure 24. A- Photomicrograph of biotite schist with porphyroblasts of euhedral to subhedral garnet. B- Hand sample of schist. Upper surface is the schistose parting and shows elongate somewhat contorted white masses of quartz and aluminosilicates, both sillimanite (fibrolite) and kyanite. Cut front face shows that these are somewhat flattened, rod-like masses that define a prominent lineation along fold axes.

Stop 8. Cambrian/Paleoproterozoic unconformity. (45.997°N, 87.841°W) Roadcuts on the north side of Highway M-69 west of the village of Felch expose the unconformity between the Munising Formation (Cambrian) and Paleoproterozoic strata of the Vulcan Iron-formation and Randville Dolomite. Along most of the roadcut nearly flat-lying sandstone of the Munising Formation is at the base of the exposures (Figure 25A), but the higher hills to the north are underlain by Paleoproterozoic rocks indicating that substantial topography existed along the unconformity.

Near the west end of the outcrop (Figure 25B) the unconformity is a steeply eastward dipping contact across which the flat-lying Munising abuts against a topographic knob composed of clasts of Vulcan Iron-formation, some of which have indications of secondary iron enrichment (Figure 25C). Munising sand fills the voids in the rubble. The material appears to have been a loosely packed pile of iron-formation rubble at the time of Cambrian transgression. Voids were filled with sand that infiltrated the pile. The angular shape of clasts indicates that material was not transported any substantial distance. The material may have been talus accumulated along the base of a south-facing slope.

Farther east along the roadcut the lower part of the exposures are again angular rubble of both Vulcan Iron-formation and Randville Dolomite with voids filled with Munising sand. Exposures higher on the hill just to the north are entirely the Paleoproterozoic units indicating that the rubble is a thin carapace of material lying against a steep south-facing slope that existed during Cambrian transgression. In some places the rubble has intervals of well-bedded sandstone within it (Figure 25D); an indication that the rubble was being delivered to the site during sandstone deposition. These relationships suggest that the rubble of Paleoproterozoic rocks was a talus deposit along the base of a marine shoreline cliff during deposition of the Munising Formation.

The relationships seen here between Precambrian rocks and the Cambrian sandstone emphasize that considerable topography existed on the surface over which the Cambrian seas transgressed across the area. Similar relationships are seen on Field Trip 3 along the Menominee Range. The relationships also suggest that outliers of Cambrian sandstone may be much more widespread under low areas than has been shown on previous maps, and that outcrops of Precambrian rocks on the present land surface were likely only at shallow depth beneath the unconformity before being exposed by younger erosion. Much of the topography presently seen over Precambrian areas is likely to be largely relict topography of the Cambrian landscape.



Figure 25. A- Flat-lying red sandstone of the Munising Formation (foreground) passing laterally into rubble of Vulcan Iron-formation at far left. View looking northwest. B- Steeply dipping unconformity shown by dashed line between Munising Formation (right) and iron-formation rubble (left). C- Clast of specularite-jasper iron-formation within rubble surrounded by Munising sand. D- Rubble of iron-formation and dolomite with thin, nearly horizontal interbeds of Munising Formation.

Stop 9. Randville Dolomite and post-Cambrian breccia. Along highway M-69, near the intersection with County Road 11, a roadcut on the north side of the highway exposes coarsely recrystallized carbonate and quartz. We interpret this to be the Randville Dolomite which, in previous mapping (James et al., 1961), was traced to within about 500 meters of this newer roadcut. An unusual feature of this exposure is a zone of breccia about 10 meters wide composed of fragments of both the Randville and of the Munising Formation. The Munising Formation occurs as coherent clasts of red sandstone (Figure 26), not unconsolidated sediments as at Stop 8, indicating that the breccia formed sometime after deposition and lithification of the Munising Formation. The Randville Dolomite occurs as angular clasts as large as about a meter diameter. The cause of this late brecciation is not clear. One possibility is that it is a karst collapse breccia formed by solution of the Randville. The Munising fragments must have been transported downward at least a few meters assuming that the Cambrian unconformity was originally slightly above the present land surface. Another possibility is that the breccia is related

to kimberlite intrusion, although we have found no indication of igneous rocks in the breccia matrix. Post-Ordovician kimberlites are known at numerous localities in the region and most have clasts of Ordovician carbonates that have fallen downward in the pipes. (See discussion in Field Trip 2 of this volume).

Other areas of disturbed Cambrian sediments have been described in the area (James et al., 1961). The Munising has been observed with dips as high as 60° locally. Most of these disturbed areas are along Precambrian faults and were interpreted to be caused by Paleozoic or later reactivation of those faults. The faults were inferred and, in one instance documented, to have normal offset. As mapped by James et al. (1961), a fault does pass about 100 meters south of this exposure so there is some possibility that the brecciation seen here is simply related to reactivation of that fault.



Figure 26. Sandstone of the Munising Formation with steep eastward dip in carbonate breccia.

Stop 10. Banded gray gneiss. (46.004°N, 87.900°W) The term “banded gray gneiss” was applied informally to a belt of Archean rocks that occur mostly immediately north of the Felch trough. Roadcuts on both sides of Highway M-69 display representative lithologies of this unit. (Note that the present position of M-69 is substantially different from that shown in James et al, (1961) because of relocation related to the Groveland mine development.)

This exposure is migmatitic gneiss composed mostly of mafic material with minor granitic stringers (Figure 27A). Most layering is nearly vertical and numerous tight isoclinal folds are present (Figure 27B). An unusual feature is a large inclusion of coarse-grained plagioclase-hornblende gneiss shown on the south cut. The inclusion has a strong foliation that lies at a distinctly lower angle than that of the surrounding gneiss.

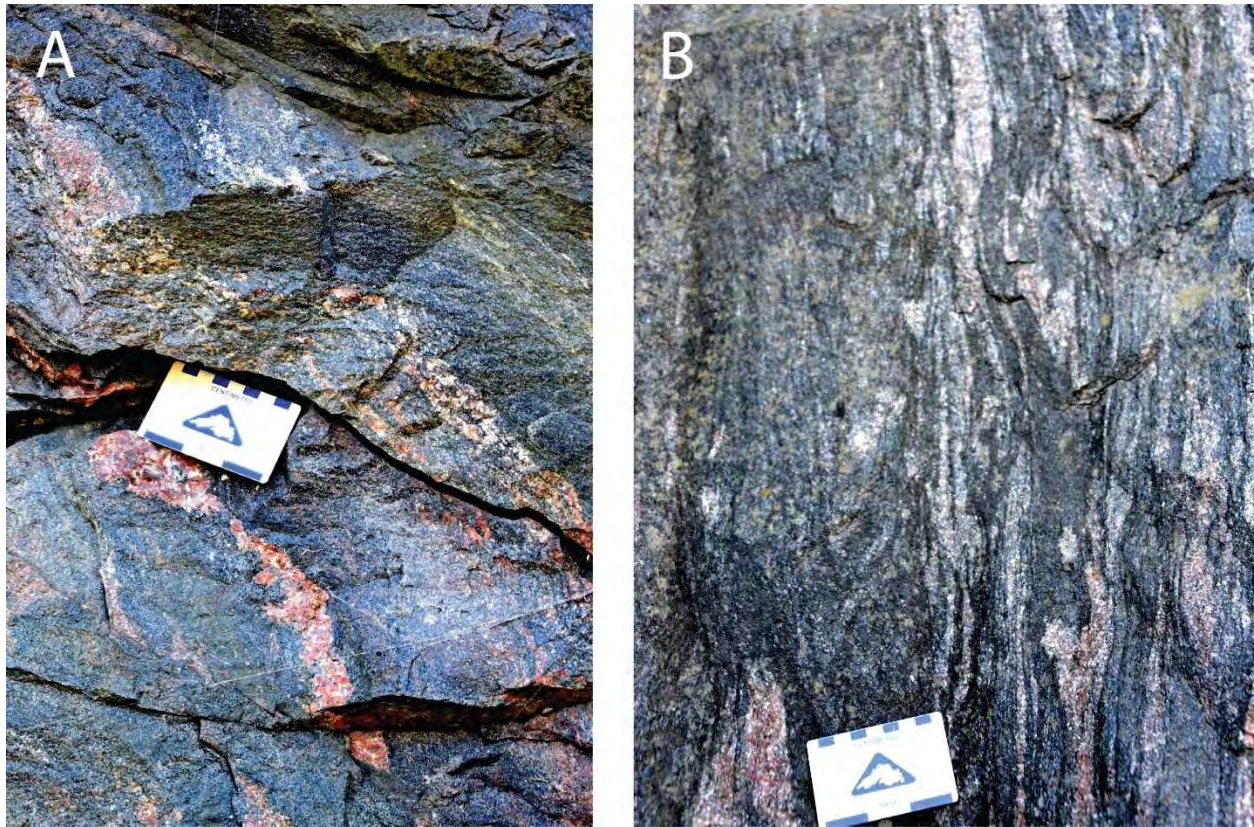


Figure 27. A- Typical amphibolitic gneiss with numerous granitic stringers. Many stringers are undeformed and are apparently post- or late-tectonic injections or segregations. B- Tightly folded migmatitic gneiss.

The general description of the banded gray gneiss in this vicinity, given by James et al. (1961, p. 121-122) is as follows:

“Most of the gneiss is light gray, or alternating light and dark gray, and is thinly layered. The rock typically contains thin rather discontinuous biotitic and hornblendic layers, not more than a millimeter thick, alternating with quartz-feldspar layers several millimeters thick. Both on fresh breaks and on weathered surfaces the gneiss is somewhat mottled as a result of segregation of feldspar into patches a few millimeters across and 5 or 6 mm long. Near some dikes of granite pegmatite, the gneiss contains layers a few millimeters wide composed of pinkish feldspar, some of which cut across the gneissic layering.

In some places, particularly near the southern margin of the gray gneiss belt, the gneiss contains lenses and pods of amphibolite as much as 100 feet thick and a quarter of a mile or more long. The light-colored layers of the gneiss are composed chiefly of feldspar, quartz, and biotite. The texture is exceedingly irregular and the abundance of the various constituents varies widely from place to place. The larger grains quartz, plagioclase, and microcline vary in size from about 0.5 mm to 1.5 mm. The microcline, which usually occurs as smaller crystals than the other constituents, is unaltered. The plagioclase in two specimens is oligoclase, whereas in a third it is probably albite. Other minerals, present in small quantities, are muscovite, chlorite, epidote, zircon, leucoxene, apatite, and iron oxides. The amphibolite that makes up the dark layers and pods in the banded gray gneiss is virtually identical to the Six-Mile Lake amphibolite previously

described. The rock is medium grained, with strong preferred orientation of the minerals. Hornblende, plagioclase, and quartz are the chief constituents."

An analysis of a sample of the gray gneiss from this road cut indicates it is basaltic with ~50% SiO₂, ~5% MgO, and relatively high TiO₂ (~1.9%) and FeO_{total} (~14.5%). The trace elements are characterized by enriched light REE and negative Nb-Ta anomaly when normalized to primitive mantle (Figure 28). The mafic gray gneiss composition overlaps with that of the Solberg Schist (Figure 28) although the Solberg is likely of Paleoproterozoic age.

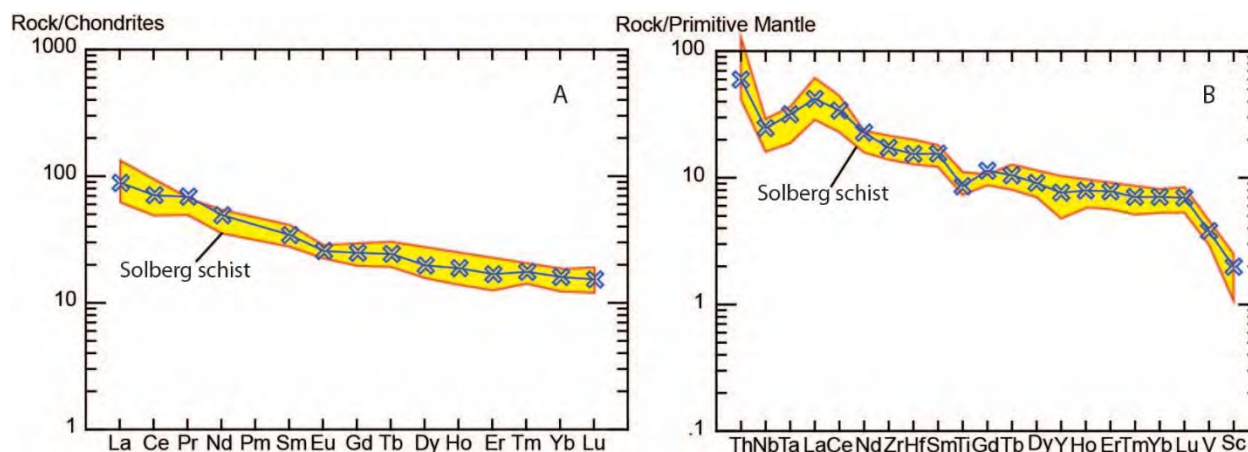


Figure 28. Chondrite normalized REE plot (A) and primitive mantle normalized trace element plot (B) for mafic gray gneiss. Field for Solberg Schist is shown for comparison.

Stop 11. Porphyritic red granite. (46.009°N, 88.061°W) Roadcuts on both the east and west sides of Highway M-95 are examples of typical porphyritic red granite, which was recognized as a map unit by James et al. (1961). The porphyritic red granite is a ferroan potassium-rich granite with A-type within-plate chemical characteristics. As mapped, it occurs in two elliptical bodies as shown on Figures 2 and 29. However, outcrops in the area are very sparse and the margins of the granite are poorly constrained. Likewise, its relationship to surrounding units is not clear. The surrounding area was designated “Dickinson group undivided” based largely on the westward extension of magnetic anomalies, mostly in the Solberg Schist, from areas of better exposure to the east. Because the magnetic anomaly produced by the Skunk Creek Member, a medial bed in the Solberg Schist, passes to the south, it is likely that the porphyritic red granite is surrounded by the lower parts of the Dickinson Group. Whether the granite bodies are intrusive into the Dickinson Group or are domes of pre-Dickinson basement is not clear from available exposures. The foliation is approximately parallel to the inferred margins of the granite bodies and dips steeply outward from their centers indicating that they are domal structures.

As described by James et al. (1961), “*The rock is generally homogeneous and coarse grained. Inclusions are rare, but dark schlieren and compositional layering locally are present. . . . Lenticular to tabular pink feldspars about half an inch long are abundant and impart a porphyritic appearance to the rock. Some of the feldspars are euhedral, but most are in fact augen, and on horizontal surfaces a foliation produced by oriented feldspars is faint to distinct. In vertical sections the structure is easily seen. In most exposures, steeply plunging lineation,*

marked by orientation of both microcline augen and biotite, is well developed, and in places it is the dominant structure.

The feldspars form about two-thirds of the rock; quartz and biotite make up the remaining one-third. Quartz itself makes up 10 to 20 percent of the rock. The cores of many of the microcline phenocrysts are white or colorless and are mantled with feldspar that is salmon pink or red. In thin section the microcline crystals have indefinite borders against a finer grained aggregate of microcline, oligoclase that is reddish and kaolinized, and quartz. The relationships suggest granulation of the borders of original tabular microclines followed by recrystallization of the granulated material.”

The preferred interpretation of James et al. (1961) was that the granite is pre-Dickinson group and therefore likely to be of Archean age, based on their assignment of an Archean age for the Dickinson Group. A new radiometric age (SHRIMP U-Pb data for zircon) determined for the granite (Ayuso et al., 2018) is ca. 2.099 Ga. (Figure 30). This well constrained age is rather surprising in that no other granites of comparable age are known in the region. Granites of this age may provide a local source for the rather abundant 2.1 Ga detrital zircons reported from the nearby Dickinson Group (Craddock et al., 2013). The date further strongly suggests that the porphyritic red granite is the basement on which the Dickinson Group was deposited and was uplifted in domal structures after Dickinson deposition.

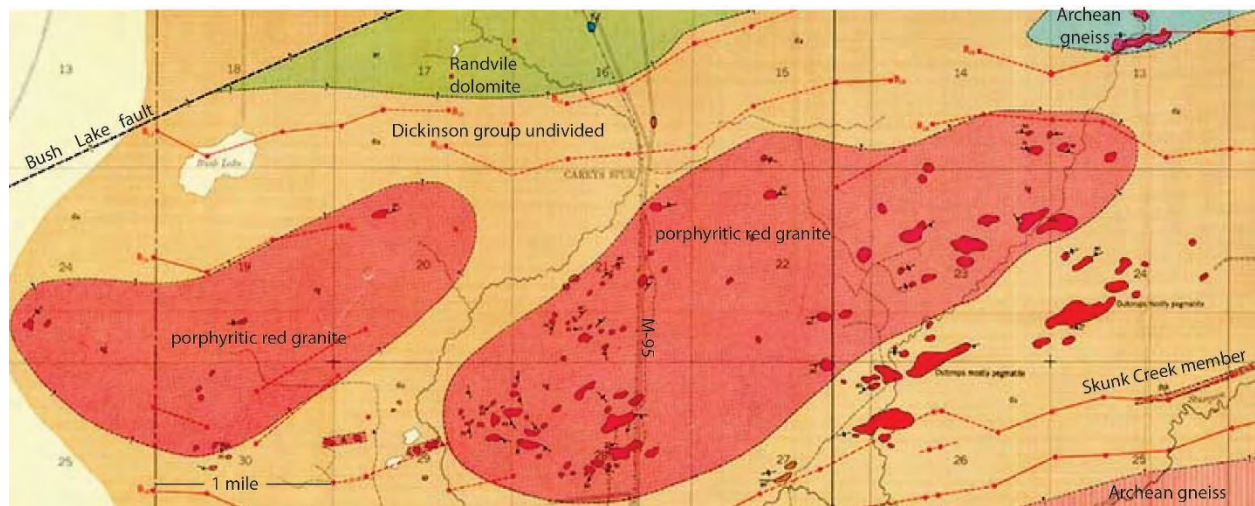


Figure 29. A portion of Plate 2 west from James et al. (1961) showing two bodies of porphyritic red granite occurring as cores of domes surrounded by undivided strata of the Dickinson Group. Red dots are aeromagnetic anomalies and red lines are inferred connections of magnetic beds between measurement points, solid where probable, dashed where uncertain.

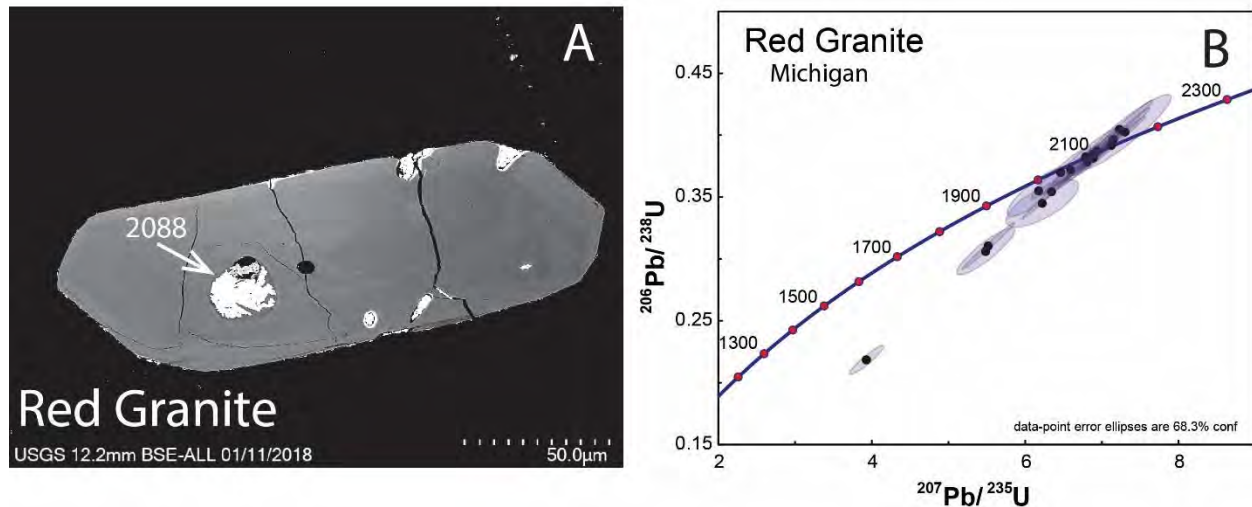


Figure 30. A- BSE (back scatter electron) image of zircon grain from the porphyritic red granite showing age of spot analyzed by SHRIMP. B- Concordia diagram for 18 spot analyses of zircons from the porphyritic red granite.

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FIELD TRIP 2

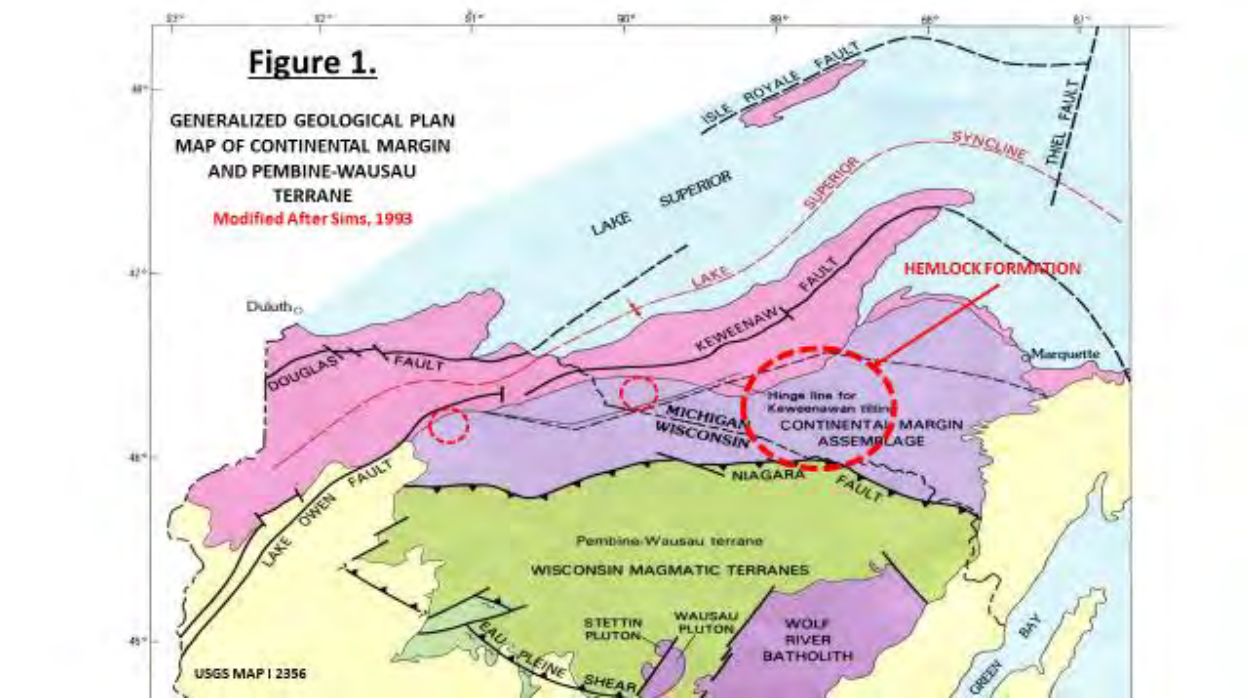
Tuesday May 15, 2018

GEOLOGY OF THE HEMLOCK FORMATION

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INTRODUCTION

The one day field trip will make eleven stops to examine the major rock types that make up or impact the 1,874 Ma Paleoproterozoic Hemlock Formation (Figure 1). The ~30,000 foot thickness of the primarily tholeiitic basalt is particularly rich in iron oxides and could easily have provided both the iron and silica incorporated in major portions of the Lake Superior type iron formations. The first stop will examine the Lake Ellen Kimberlite which is the only easily accessible kimberlite in the Upper Peninsula Kimberlite District and it also intrudes the Hemlock Formation. The differentiated West Kiernan sill will be visited including the ultramafic lower unit and the base metal rich differentiate near the base of the gabbro unit. The upper transition will be examined where plagioclase levels approach 80% and contains significant titaniferous magnetite, stilpnomelane and apatite. Other stops will illustrate rhyolite, volcanic conglomerates, amygdaloidal and pillowed basalts. Also, one of the stops will examine the Mansfield iron member which is one of several areal restricted iron formations present in the Hemlock.



General Geology

The ~2.7 Ma Margeson Creek Gneiss in the center of the Amasa Uplift is overlain by the Randville Dolomite of the Chocoday Group and is equivalent to the Kona Dolomite on the Marquette Range and the Badriver Dolomite on the Gogebic Range. Dating by Vallini and others (2006) indicates the age of the Sturgeon quartzite under the Randville dolomite in the Iron Mountain area is 2.3-2.2 Ga.

The Randville dolomite directly underlies the Hemlock Formation and exhibits several lithologies. The most pervasive lithologic type is a sandy dolomite while a coarse orange arkose, not present in the Randville equivalents elsewhere, is common at the southern portion of the Amasa Uplift. The age of the Randville is at least 300 million years older than the 1874 Ma Hemlock Formation (Schneider and others, 2002). The basal conglomerate and quartzite units present in the Chocoday Group of the Marquette Range and Menominee Range are absent around the Amasa Uplift.

At the south end of the Amasa Uplift the Randville is approximately 1800 feet thick and consists of sandy dolomite, feldspathic quartzite, arkose and sericite argillite. Some authors have postulated that the Randville has pinched out on the north end (Cannon and others, 1976; Foote, 1981). However, subsequent drilling in section 28, T. 46 N., R. 33 W indicates the dolomite is present on the north end, but that same drilling was not extensive enough to identify lithologic types or thickness. Overburden depths on the north end of the Uplift can vary from 50 feet to over 560 feet with no reported outcrops.

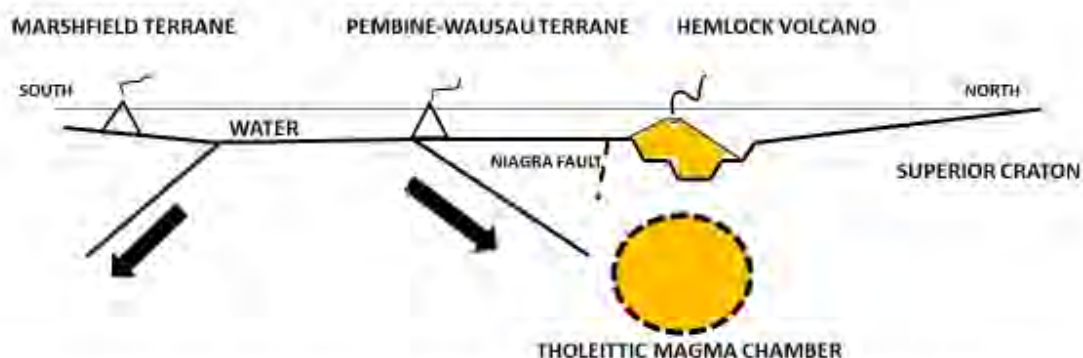


Figure 2. SKETCH OF HEMLOCK VOLCANIC EMPLACEMENT IN EXTENSIONAL BASINS ALONG THE SOUTHERN MARGIN OF THE SUPERIOR CRATON @~1874 Ma

During the early Penokian orogeny the Pembine-Wausau terrane (Figure 2) was already accreted to the Superior Province by 1875 Ma (Figure 2). Back-arc extensional basins south of the Niagara fault contains numerous volcanogenic massive sulfide deposits (VMS). Rifting on the continental margin produced a basin(s) into which banded iron formations formed coeval with volcanism. The age of banded iron formation deposits and the intra-arc rift massive sulfide deposits in the

Pembin-Wausau terrane are both ~1874 Ma. One of the largest volcanic centers is the Hemlock Formation, located in Iron County, Michigan. The volcanic pile achieved a thickness in excess of 30,000 feet west of the Amasa Uplift and thins away in all directions (Figure 3). It is estimated the footprint exceeds 2,800 square miles.

Geophysics

A total field positive magnetic anomaly around the west side of the Amasa Uplift is caused by increased magnetite content in the upper 6000 feet of the Hemlock basalts (Figure 3). The Amasa Iron-formation is essentially non-magnetic as primary iron minerals have all been oxidized to hematite and in some cases enriched to iron ore that was mined early in the last century. Major conductivity zones are associated with graphitic slates internal to the volcanics and the overlying graphitic Michigamme slates. The presence of disseminated sulfides in the lower portion of the normal gabbro of the West Kiernan sill produce minor EM anomalies.

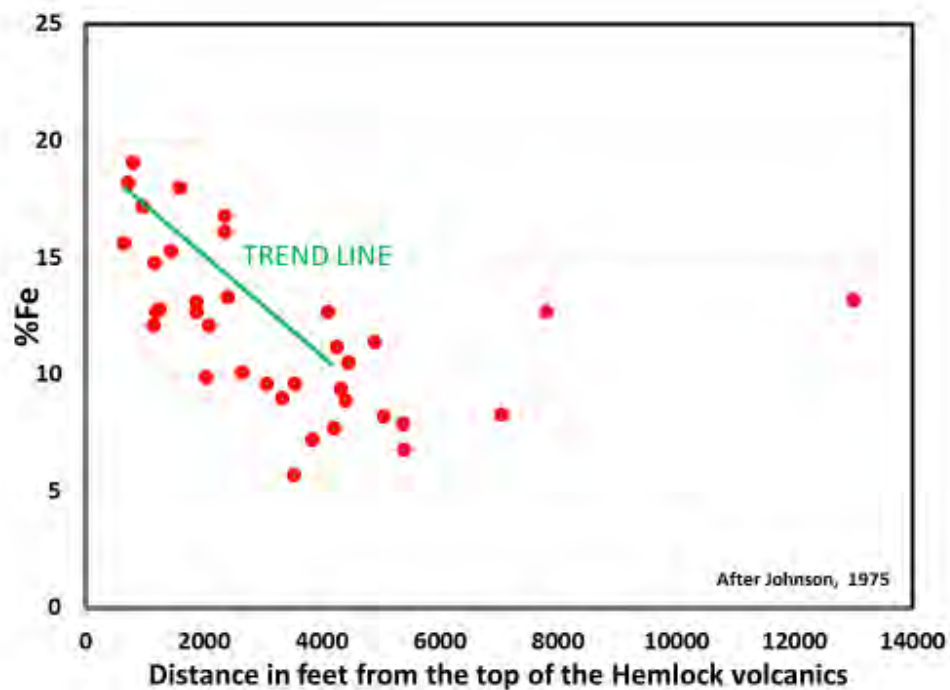


Figure 3. Increase in iron content in basalt near the top of the Hemlock Formation

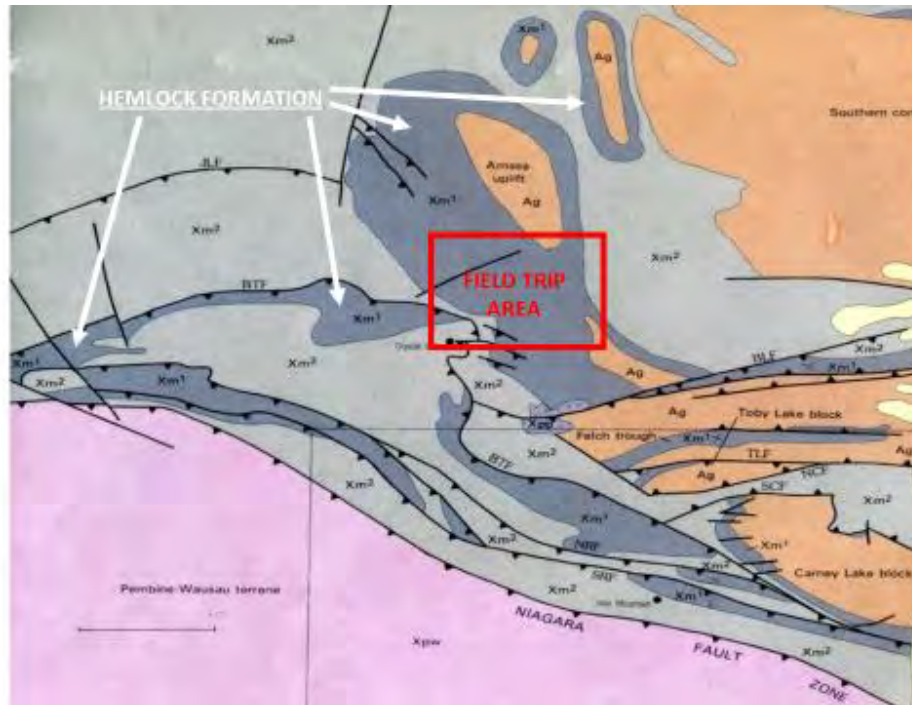


Figure 4. Plan geology map showing the sub crop of the Hemlock Formation (Xm²) and the later thrust faulting. Field trip area is defined by the red box. (After USGS Map I-2356)

Generally the exposed Archean portion of the Amasa Uplift exhibits low gravity readings. Most of the Hemlock sub crop area exhibits a muted gravity contrast. Positive gravity values are associated with the Riverton Iron-formation in the Iron River-Crystal Falls allochthon and an area northwest of the Amasa Uplift. A Bouguer ground gravity map with limited field stations of the Amasa Uplift area show an increase in the gravity field toward areas of thicker basalts of the Hemlock Formation, especially where the increased magnetite content increased the specific gravity of the rock. After deposition of the Michigamme Formation the continued northern push by the Wisconsin Magmatic terrane produced a number of east-west thrust fault panels replicating the stratigraphy in each panel (Figure 4).

Hemlock Formation

The Hemlock Formation was named (Clements, 1899) for volcanic rocks found near the Hemlock River west of the Amasa Uplift. The Hemlock belong to the Menominee Group of the Marquette Range Supergroup. The volcanics and sediments were deposited subaqueously and are believed to be terrigenous sedimentary sourced (Johnson, 1975; Dann, 1978; Ueng, 1987; and Beck, 1991). Beck (1991) described the volcanics as continental flood basalts.

Principle lithologies listed in order of decreasing abundance are: basalt, volcanoclastics (referred to as agglomerates, hyaloclastites and breccias), rhyolite, graphitic slates and small iron formation members (one, the Bird, is an iron oxide chert and the other, the Mansfield, is a carbonate chert).

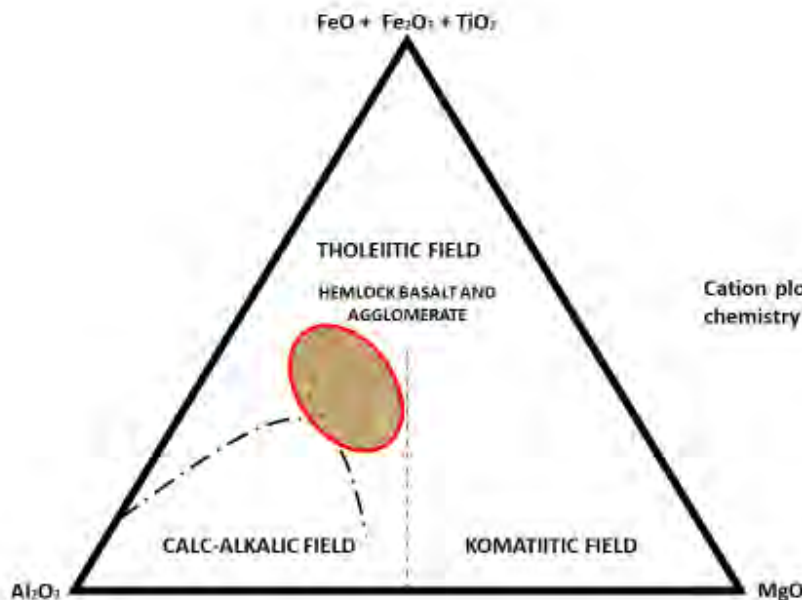


Figure 5.
Cation plot diagram showing the Hemlock
chemistry in the tholeiitic field

Modified after Foote, 1981

The base of the Hemlock Formation rests unconformably on the Randville Dolomite except in the vicinity of Michigamme Mountain where the base of the Hemlock rests on a small deposit of unique quartzite composed of angular quartz and chert clasts along with minor massive chert. Portions of the quartzite have been replaced by magnetite and specularite while some of the chert contains secondary specularite that replaced the chert yielding a crude banding appearance.

The Hemlock tholeiitic basalt magma (Figure 5) by its reduced nature, concentrated iron in the residual magma, principally as magnetite. Replacement iron oxides are found in the small clastic unit located on the southeast portion of the Amasa Uplift at the base of the Hemlock Formation.

Iron oxide concentrations, principally magnetite, are found sporadically in the basaltic portion of the Hemlock and in the upper ~6000 feet of the basalts. Following active volcanism banded iron formations were formed (Amasa [west side of Uplift] and Fence River [east side of the Uplift]).

There are a number of intrusive sills found within the Hemlock Formation. One, the West Kiernan sill, is approximately thirteen miles long and one mile thick as measured at outcrop. It is differentiated into four distinct units plus a chilled diabase contact zone. Relatively thin, but internally layered sills like the West Kiernan sill, consisting of basal peridotite overlain by pyroxenite, gabbro, and granophyre, are unusual in that injection of relatively crystal-free magma does not appear to form well layered, well-differentiated intrusions (March, 2006). For example, the large Sudbury impact melt sheet (3 km x 200 km; volume of ~30,000 km³) shows no sign of layering and very little sign of differentiation while the more than 300 m thick Palisades sill consists only of an olivine-enriched lower layer (~3 m thick), diabase, ferrodiabase, and granophyre (Walker, 1969). However, similarly layered and differentiated sills like the West Kiernan sill have been described from other Precambrian terranes including the Archean Vermilion district in Minnesota (Schulz, 1982), the Abitibi greenstone belt in Ontario (MacRae, 1969), the Eastern Goldfields region in Western Australia (Williams and Halberg, 1973) and the

Barberton Mountain Land in South Africa (Anhaeusser, 1985). For the Archean examples, the layered sills appear to reflect crystallization from an iron-rich picrite magma (Schulz, 1982). Whether the West Kiernan sill is also related to a high-iron picrite magma is unclear.

Until the late 1980's the volcanic extrusives (Badwater Greenstone) around the Iron River-Crystal Falls District were considered to be a distinct and separate mafic extrusive unit from the Hemlock. Several papers suggested that Badwater Greenstone was actually the same unit as the Hemlock based on similarity of rock types, textures and geochemistry (Dann, 1978, 1979).

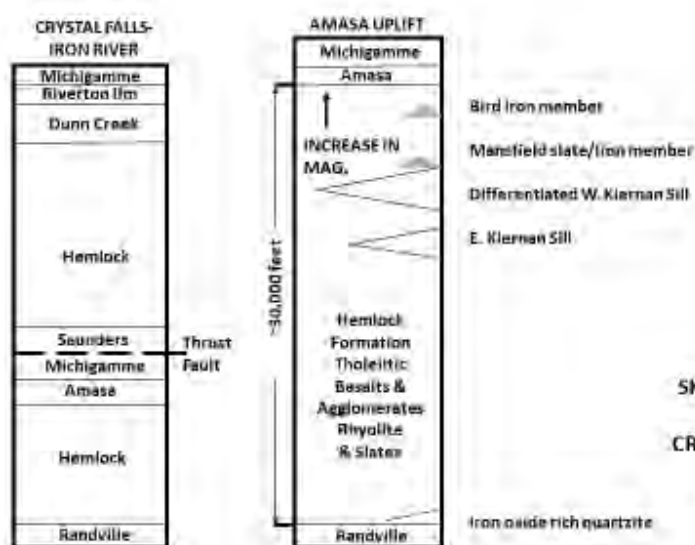


Figure 6.

SKETCH OF THE STRATIGRAPHIC
COLUMNS FOR THE
CRYSTAL FALLS-IRON RIVER AREA
AND
THE SW AMASA UPLIFT
NO SCALE

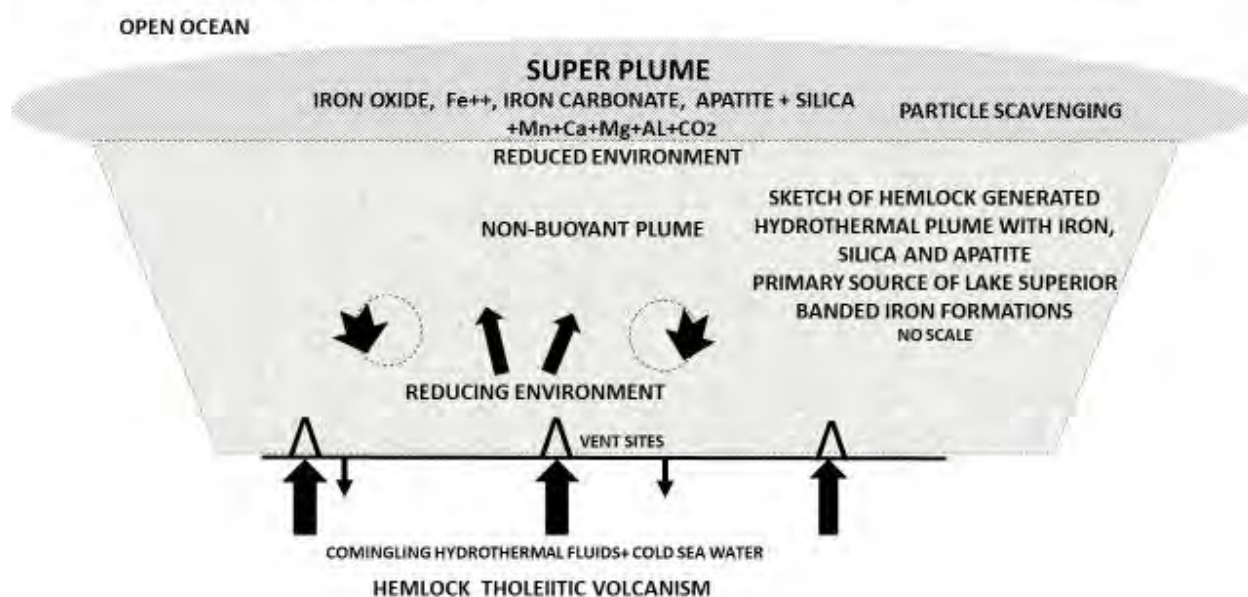
In the mid-1980's several USGS personnel were looking south across the Paint River and noticed a pink colored rock beneath the greenstone and identified it as the Saunders Formation (Randville equivalent). This suggested that the entire Iron River-Crystal Falls district could be an allochthon (detachment fault) that had been forced northward up and over the underlying younger Michigamme Formation. In addition it was recognized that the Paint River Group was not a younger sequence than the Baraga Group but rather a fault repetition with the slates over the Riverton Iron Formation equivalent to the Michigamme slates (Figure 6). Regional metamorphism produced during the Penokean orogeny in the area of the field trip falls either in the chlorite or biotite isograd (Weir, 1986).

Hemlock Formation as a Flux Source for portions of Iron Formations in the Lake Superior District

Intermittently, during and at cessation of active volcanism the magma source continued to yield significant quantities of iron, silica, phosphorus and carbon dioxide forming a large volume super plume (Figure 7) containing both instantly crystalized iron oxides and amorphous silica along with significant soluble Fe^{++} that spread in the ocean over thousands of square kilometers. Both

magnetite and specularite were produced at the vent site(s) under equilibrium conditions that allowed instant crystallization, much like the sulfide particulate matter associated with black smoker vents. Very fine specular hematite is found at the core of much of the magnetite in banded iron formations worldwide (Han, 1966, 1978, 1988). The iron oxides (i.e. microplaty-specular hematite and magnetite) iron carbonate and amorphous silica settled to the sea floor (possibly on seasonal changes) forming alternating bands of chert and iron minerals with thickness based on flux available from the source, current directions and saturation conditions at the deposition site. Near shore banded iron formations exhibit granular textures, cross bedding and occasional reef development of stromatolites while deeper deposits are lithic without oolitic or granular layers or significant traces of biogenic activity.

Figure 7. LATE HEMLOCK FLUX EMANATIONS FEEDING BANDED IRON FORMATIONS



Ancient VMS (e.g. New Brunswick #3 and 6; Austin Brook; Manitowadge, Ontario; Lokken, Norway; Bending Lake, Ontario) and SED-EX (e.g. Little Commonwealth and Dunkel exploration, Wisconsin) deposits, with associated banded iron formations, have modern day analogues (without banded iron formations) at spreading plate centers where iron is precipitated as an oxyhydroxide ostensibly converted to hematite and magnetite at a later period. Oxyhydroxides do not convert to specularite or magnetite under normal natural conditions. Current sea floor discharge systems yield small sea floor deposits from small plumes under predominantly oxidizing conditions. Large systems capable of producing sufficient silica and iron for the creation of banded iron formations would require a supersized plume extending over large areas. Also, these large systems would create a significant reduced environment where early formed iron compound particles can scavenge additional iron. Sea floor ventings studied since discovery in the 1970's are useful in observing the mechanisms in play in the formation of iron oxides and silica near the vent sites. Lake Nyos in Cameroon (Ozawa, 2016) illustrates the formation of crystalline siderite from a vent source high in Fe^{+2} and CO_2 under a slightly acid environment. This would suggest that a plume environment at the bottom of the sea can quickly

precipitate iron carbonate along with amorphous silica which can separate into chert and iron rich layers (Krapez, 2006).

Conditions impacting the type and nature of the banded iron deposits produced include:

- Plume buoyancy.
- Hydrothermal water temperature, pH and Eh.
- Amount of iron and silica flux including CH₄, CO₂, H₂S and fluorine.
- Discharge conditions including sea water temperature, salinity and pressure (depth).
- Discharge and plume environment, oxidation and reduction conditions, plume size/thickness.
- Size of the igneous flux source
- Definable presence of iron and silica available to produce enough flux to form iron formations.

FIELD TRIP STOPS

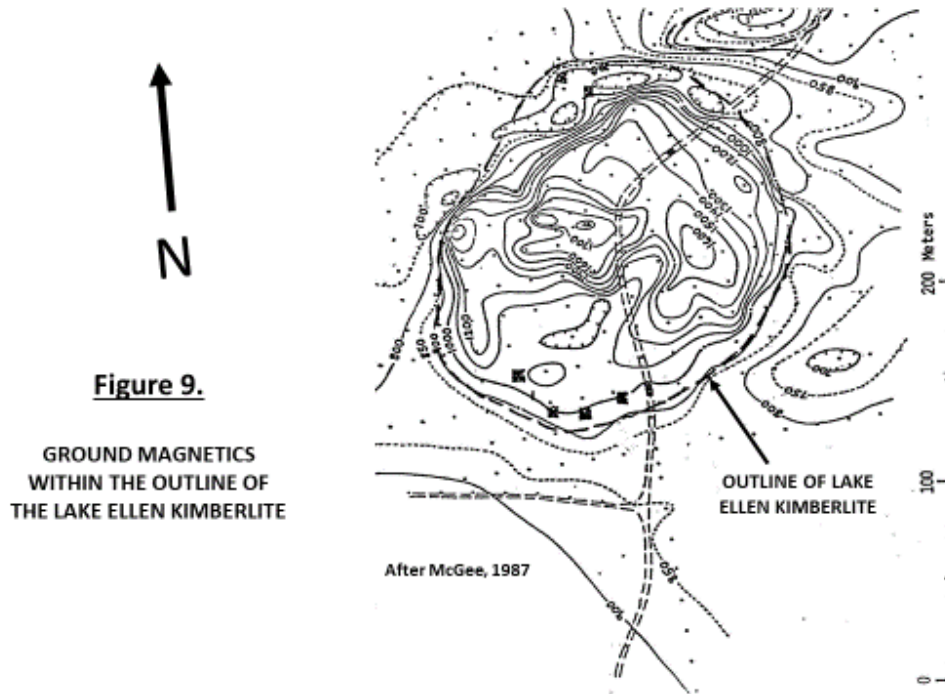
Ten of the eleven stops are shown in Figure 8 below.



Figure 8. Hemlock field trip stops 1-7 and 9-11 Kiernan and Lake Mary Quadrangle

Stop 1. Lake Ellen Kimberlite

UTM: 46° 10.478 N 88° 10.587 W, SWSW Sec. 27, 44-31



Gair (1956) in USGS Bulletin 1044, plate 2 noted a small magnetic anomaly in the SWSW of Section 27, T. 44N, R. 31 W. and made an observation that the magnetic high was caused by magnetite in the glacial till. In 1971 William Spence and Klaus Schulz discovered the Lake Ellen kimberlite (also referred to as Site 10) under thin soil cover while conducting base metal exploration for industry. Cannon (1981) wrote a report on the occurrence. Crystal Exploration acquired control of the property and conducted geophysics (Figure 9), drilling, trenching, bench testing and analytical work. This activity spurred other companies to conduct air and ground magnetics, soil and stream surveys, land leasing and follow-up drilling.

Shapes of kimberlites are usually round or elliptical and can cover an area from a few acres and up to 200+ acres. The Lake Ellen kimberlite is approximately 600 feet in diameter (Figure 9). Most of the kimberlites in the district have a similar physical appearance (Figure 10). Depth of burial and degree of weathering can make geophysical prospecting difficult. The Lake Ellen pipe has a variable magnetic signature. Unweathered kimberlites contain minor magnetite which can be oxidized to hematite near surface and lose their magnetic properties. Where weathering produced clays or where epiclastic units are present the electromagnetic techniques can be an effective location tool. Inhomogeneity can mask certain parts of a single pipe. In addition to geophysical techniques, soil or stream sampling can be an effective method of discovery because the Lake Ellen is overlain by a thin soil cover. An excavation pit used for both bulk sampling and woods road construction constitutes the focus of the first stop.

The known kimberlite district extends from the Michigamme Reservoir in the northwest to Powers in the southeast, a distance of 62 miles. The oval outline of the kimberlite district has a width of approximately 19 miles (Figure 12).



Figure 12. Plan map of the Michigan Upper Peninsula Kimberlite District.

A subset of purple chromium rich garnet xenocrysts can be used to project the potential of any kimberlite to contain diamonds. Orange, red-orange and light orange garnets are from eclogite xenoliths while purple, red and pink garnets fall within peridotite affinity. The Lake Ellen kimberlite has very few indicator garnets, 3 of a population of 178 (McGee, 1988) or 1.7% indicating a lower probability of containing diamonds. A 180 short ton bulk sample produced four diamonds which in quality and number make the Lake Ellen kimberlite uneconomic. None of the discovered kimberlites in the field have economic concentrations of diamonds.

Xenoliths and xenocryst compositions indicate the diatreme originated in the upper mantle. Based on data obtained from garnets the calculated equilibrium temperatures range from 950°-1100° C. The kimberlite originated from approximately 140-160 km below the surface (Griffin, 2004).

Age dating of Site 73 Kimberlite emplacement yielded a zircon age of 155 Ma while a K-Ar determination on phlogopite yielded a 190 Ma age indicating a Jurassic Period (138-205 Ma) emplacement. Granulite whole rock data suggest two age groups with affinities to tholeiitic basalts. Trace element analyses suggest one group is of Archean derivation while a second group is aligned with the Keweenaw extrusive rocks of the Mid-Continent Rift (Zartman and others, 2012).



Figure 13.
PLAN MAP OF STOPS
2a THROUGH 2d.

Stop 2a. Basal Hemlock Formation mineralized quartzite (Figure 13)

UTM: 46° 9.545N 88° 11.087W, NENE Sec. 4, 43-31

Stop 2 will concentrate on Michigamme Mountain, a local topographic high (Figure 13). The principle rock at this location is a unique mineralized quartzite of limited areal extent. Gair (1956) named the quartzite Goodrich with the younger Hemlock overlying the quartzite. We now know that the age of the Hemlock (1.84 Ga) matches the age of the major iron formations in the Lake Superior region. The quartzite is composed of sub-angular quartz and granular fragments of chert. The quartzite overlies a thin massive chert. A later hydrothermal influx of specularite (Figure 14) and magnetite (Figure 15) associated with potassic and silica enrichment replacing some of the quartz and chert in the quartzite. Subsequent surface supergene oxidation converted some of the magnetite to martite. Soluble iron values are generally below 20% but can exceed 50%. At this location we are standing on an east-west structural anticline that may possibly be a chert dome.

Stop 2b. Specularite chert from test pits.

UTM: 46° 9.583N 88° 11.115W, SESE Sec. 33, 44-31

In the saddle between the two mineralized quartzite topo highs is a massive chert with secondary specularite (Fig. 13). You will note quite a bit of cross cutting specular hematite suggesting the iron oxide was emplaced in the chert at a later time. From Stops 2a & b you can see the highest elevation on Michigamme Mountain 80 yards to the southwest. The topographic high outcrop is quartzite with only minor magnetite and the occasional quartz vein containing micaceous selvages along the contacts



Figure 14. Specularite chert from saddle on Michigamme Mountain



Figure 15. Martite after magnetite with quartz/adularia vein from Michigamme Mountain.

Stop 2c. Hydrothermally enhanced rhyolite

UTM: 46° 9.581N 88° 11.116W, SWSE Sec. 33, 44-31

While ascending Michigamme Mountain note the occasional black rock both to the north and underfoot. This is the same magnetite, now martite, quartzite to be seen at Stop 2a.

The quartzite is overlain by a rhyolite flow representing the basal Hemlock Formation present along the eastern outcrop area of the Amasa Uplift. The rhyolite has quartz eyes and is typically pink on outcrop and orange on fresh surface (Figure 16 and 17). The orange color is primarily due to the addition of potassium feldspar. Chemistry of most of the rhyolite in the Hemlock shows the difference in K-spar (Table 1). Slight additional silica is provided by numerous random quartz veins that can be found in both the rhyolite (Figure 17) and underlying quartzite. Some of the veins in the quartzite contain selvages of micaceous hematite. It is suspected the potassium and silica were introduced at the time both magnetite and specularite replaced portions of the chert and quartzite at the base of the Hemlock. Further support of this timing is noted by the absence of alteration of overlying tuff and agglomerate at this stop.



Figure 16. Rhyolite outcrop showing quartz veins



Figure 17. Enhanced alkali/silica rhyolite porphyry

Oxide	Alt rhyolite*¹	Unaltered rhy.*²
SiO ₂	73.5	72.7
Al ₂ O ₃	12.2	10.1
Fe ₂ O ₃	2.3	1.3
FeO	.22	3.1
MgO	.31	1.8
CaO	.21	1.4
Na ₂ O	.23	1.0
K ₂ O	9.85	3.9
TiO ₂	.99	.47
P ₂ O ₅	.08	.08
	99.9	95.9

Table 1.

Comparison of altered and unaltered rhyolite within the Hemlock Formation.

*1 USGS Bull. 1044 p. 54

*2 USGS Bull. 1226 p. 21

Potassium values are twice those of other Hemlock rhyolite flows (see Table 1). Gair and Wier (1956, p. 53) “The acid volcanic rocks in the western part of the quadrangle (Kiernan-Sec. 36, T. 44 N., R. 32 W; Sec. 5, T. 43 N., R. 31 W) is much fresher and poorer in feldspar than the rock in the vicinity of Michigamme Mountain”. Table 1 shows the chemical difference between fresh and altered rhyolite. Chemistry of most of the Hemlock rhyolites matches that of the granophyre of the West Kiernan Sill. Ueng (1988) suggested the rhyolite formed in the magma chamber by crystal fractionation similar to the West Kiernan Sill differentiation that produced the granophyre. It can be speculated that the iron oxide addition to the underlying quartzite and chert occurred after extrusion of the rhyolite and prior to the next basaltic eruption. The overlying mafic tuff and agglomerate do not show any alteration beyond the generation of chlorite common to most of the Hemlock Formation indicating the oxide event occurred before emplacement of the basaltic rocks. On our trip returning to the vehicle, we will examine test pit material that represents fine specularite replacing the host quartzite.

Stop 2d. Agglomerate (optional)

UTM: 46° 9.692N 88° 10.985W, SESE Sec. 33, 44-31

A significant portion of the Hemlock is composed of fragmental tholeiitic basalt variously referred to as agglomerate, breccia or conglomerates (Figures 18-19) depending on visual degree of sorting or angularity. This stop has been highly weathered and is covered by moss and lichens. A more photogenic opportunity will be afforded at Stop 8.



Figure 18. Fine grained volcanic conglomerate.



Figure 19. Coarse grained agglomerate (hyaloclastites) breccia.

Stop 3. East Kiernan Sill (optional)

UTM: 46° 8.828N 88° 12.568W, SWSE Sec. 5, 43-31

The East Kiernan sill is a much smaller intrusion than the Western Kiernan sill. It is primarily an undifferentiated gabbro containing 50-80% prismatic hornblende after augite, albite, titaniferous magnetite, and apatite with less than 5% quartz. The plagioclase has been saussuritized and some hornblende converted to chlorite while the oxides have been converted to titanite and rutile. A number of gabbro samples of both Kiernan sills have been analyzed for PGEs (Bornhorst, 1990). None of the samples show values above background.

Stop 4a. Mansfield Mine Location Monument

UTM: 46° 6.835 N 88° 13.076 W, NWNW Sec. 20, 43-31

In 1889 the Mansfield Mining Co. took a lease on the Mansfield natural iron ore deposit from J.M. Longyear. Bessemer iron ore (<.045% P) was mined from six levels down to 435 feet below surface. Most of the workings were under the Michigamme River.



Figure 20. Sign and plaque commemorating the Mansfield Mine Site disaster of 1893

On the evening of Sept. 27, 1893 a lower level pillar gave way causing the rock above to cave to surface flooding the mine workings with water from the Michigamme River. Twenty seven miners lost their lives in an instant. In 1897 the DeSoto Iron Co. of Springfield, IL bought the mine, redirected the river channel and reopened the mine. The Oliver Mining Co. (now USX) operated the mine from 1911 until the exhaustion of iron ore in 1913 at which time the workings had reached the 17th level about 1480 feet below the collar. A total of 1,462,504 long tons (LT) were mined between 1890 and 1913. In 1983 the mine site was designated the Mansfield Mine Location Historic District. A plaque with the names of the 27 men who died in the disaster marks the site (Figure 20). A few restored buildings mark the site of the mining community.

Stop 4b. Mansfield Iron bearing slate member and agglomerate/volcaniclastics.

UTM: 46° 6.945' N 88° 13.185' W, NWNW Sec. 20, 43-31

The vertical #2 shaft went through fine volcanic conglomerate of the Hemlock and can be found on the mine dump. The slate portion is also represented on the dump (Figure 21) along with the iron formation. The Mansfield iron member has a limited areal extent of about 4 miles on strike. It was originally a siderite chert with soluble irons ranging from 10-47.3% that averaged ~25%. Near the contact with the intrusive West Kiernan Sill the Mansfield iron member was metamorphosed to a magnetite stilpnomelane chert. At the Mansfield Mine location the original siderite chert was far enough from the sill contact to avoid metamorphic alteration, but it did experience supergene oxidation and enrichment (Figure 22) with ore grades averaging about 52.8% natural iron with less than .045% P making it Bessemer type ore.



Figure 21. Graphitic slate with pyrite cubes



Figure 22. Direct shipping hematite ore with gypsum(white arrow).

Stop 5. Steeply dipping Hemlock pillowed and massive basalt-Hemlock Falls Dam Site.

UTM: 46° 7.840 N 88° 13.504 W, SESE Sec. 7, 43-31

The bulk of the Hemlock Formation is composed of massive, amygdaloidal and pillowed basalts and volcaniclastic rocks of the same composition. On the western edge of the Hemlock Falls Dam abutment is a large outcrop of pillow basalt dipping steeply to the west. Directly overlying the pillowed basalt is massive basalt as seen on the left side of Figure 23.



Figure 23. Steeply dipping tholeiitic pillowed basalts at the west abutment of the Hemlock Falls Dam site.

Stop 6. Amygdaloidal basalt of the Hemlock Formation

UTM: 46° 6.461 N 88°13.931 W, NWNW Sec. 20, 43-31

On the north side of the County Road is a low outcrop of broken amygdaloidal basalt porphyry (Figure 24) with abundant gas vesicles filled with secondary albite quartz and carbonate. Some parts of the basalt contain a few per cent sulfides, mostly pyrite. Amygdaloidal basalts are not as abundant as massive basalts or agglomerates.

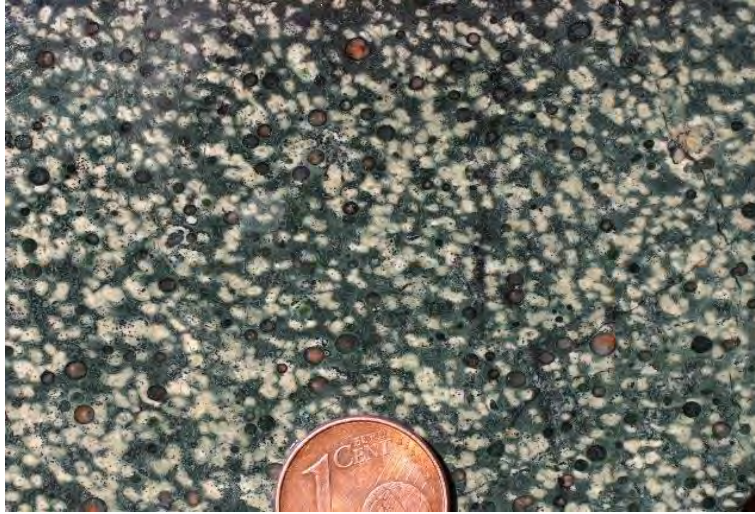


Figure 24. Polished surface of amygdaloidal basalt at Stop 6

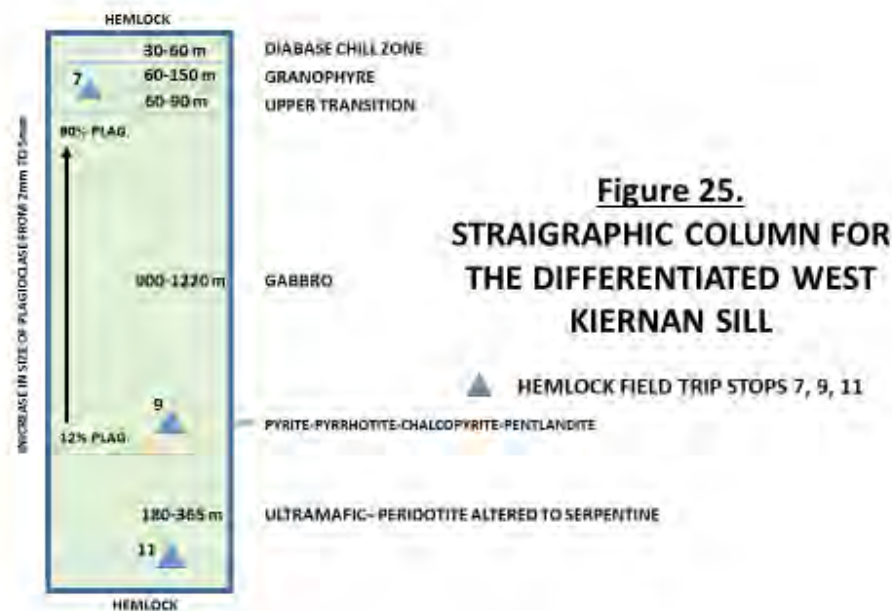
DIFFERENTIATED WEST KIERNAN SILL

The West Kiernan Sill is approximately 13 miles long and averages 1.1 miles in thickness. It is designated a sill in that most contacts are concordant with the Hemlock units. Regional metamorphism for the entire sill is in the greenschist metamorphic zone.

The sill can be divided into four distinct rock units: a basal serpentinitized peridotite (Stop 11), a thick normal gabbro, a transition gabbro and, occasionally, an upper granophyre (Figure 25). A chill zone in the contact area is usually a diabase. The basal peridotite is composed of serpentine with aggregates of tremolite, talc and magnetite with minor amounts of carbonate and chlorite.

Overlying the serpentine is a 900-1200 m thick normal gabbro. An increase in size and amount of plagioclase is noted from the base to the top of the unit (Figure 25). Alteration stilpnomelane is noted in the upper portion of the unit and becomes common in the overlying transition gabbro. The transition (Stop 7b) zone contains spotty concentrations of titaniferous magnetite that show up as magnetic bullseyes on airborne magnetic survey maps.

The granophyre looks like a medium grained granite with minor or missing mafic minerals. Minerals present include albite, oligoclase, microcline and orthoclase with occasional needle of magnetite. Calcite is also pervasive. Geochemical analyses of the sill units are given in Table 2 along with analyses of the Hemlock basalt.



	1	2	3	4	5	7	8
Oxide	Peridotite* ¹	Gabbro* ¹	Transition Gabbro* ¹	High Iron* ²	Granophyre	Hemlock Basalt* ³	Hemlock Agglomerate* ⁴
SiO ₂	39.01	44.94	57.2	34.1	69.12	48.2	47.6
Al ₂ O ₃	6.56	19.70	10.8	3.4	12.83	14.6	14.4
Fe ₂ O ₃	11.49	1.72	8.8	13.1	.78	6.7	5.4
FeO	5.30	7.40	7.5	22.3	5.85	7.8	7.6
MgO	23.84	8.91	2.7	8.5	1.38	5.9	6.6
CaO	3.57	9.22	4.8	7.6	.70	5.8	7.8
Na ₂ O	--	1.94	2.8	.13	1.72	3.03	3.29
K ₂ O	.02	.36	1.1	.10	3.44	.92	.70
TiO ₂	2.04	.73	2.9	5.42	.73	2.21	2.15
P ₂ O ₅	.19	.09	.76	.38	.11	.30	.19
MnO	.11	.13	.26	.38	.05	.17	.17
CO ₂	.08	.45	.37	-	.70	.71	.70
H ₂ O	7.10	4.56	3.90	3.0	2.25	2.64	2.57
Total	99.97	100.26	98.90	98.40	99.75	98.98	99.17

*¹ Bayley, 1959, Amer. Jour. of Sci. v. 257, p. 428 (col. 1-3)

*² Wier, 1967, USGS Bull. 1226, p. 35 (NW ¼ Sec. 12, 43-32)

*³ Foose, 1981, N=7

*⁴ Foose, 1981, N=10 agglomerate/volcaniclastics

Table 2. West Kiernan Sill-Whole rock analyses

Based on the similarities in geochemistry between the West Kiernan and the Hemlock extrusives (Ueng, 1987) concluded both major igneous rocks were co-magmatic in origin. He further stated crystal fractionation in the magma chamber was responsible for the occasional rhyolite flows. Xenoliths of Hemlock are present in both the East and West sills (Bayley, 1959; Wier, 1967).

Stops 7a-b-c will allow us to examine the transition gabbro (7c) containing abundant magnetite and the overlying granophyre (7a). Whole rock assays for these outcrops are shown on Table 3. Note the major increase in silica, sodium and potassium coupled with a significant reduction in iron oxides and titanium in the granophyre. The two rock types represent a significant chemical change during late stage differentiation.

OXIDE	A*¹	B*²
SiO ₂	34.1	51.0
Al ₂ O ₃	3.4	16.3
Fe ₂ O ₃	13.1	1.4
FeO	22.3	9.2
MgO	8.5	8.5
CaO	7.6	10.1
Na ₂ O	.13	1.83
K ₂ O	.10	.68
H ₂ O	3.0	-
TiO ₂	5.42	.85
P ₂ O ₅	.38	.05
CO ₂	<.05	-
MnO	.38	.14
S	.38	-

*¹ test pit with magnetite in transition gabbro.

USGS Bulletin 1226, p. 35

*² Fox M.S. thesis sample A5P outcrop near railroad cut.

Table 3. Whole rock analyses upper Kiernan Sill transition gabbro near granophyre contact (Stop 7).

Stop 7a. West Kiernan Sill granophyre

UTM: 46° 8.545 N 88° 15.742 W, NWNW Sec. 12, 43-32

The outcrop is typical of the granophyre phase of the West Kiernan Sill. Very little ferromagnesium mineralization is present and it does resemble a granite.

Stop 7b. Coarse grained transition gabbro

UTM: same as 7a.

This location is very coarse grained and contains around 80% plagioclase, typical of the upper part of the transition gabbro. The apatite is of the fluorine variety.

Stop 7c. Transition gabbro with magnetite

UTM: 46° 8.508 N 88° 15.717 W, NWNW Sec.12, 43-32

These test pits are near the top of the transition gabbro where abundant titaniferous magnetite is common. Additional minerals include stilpnomelane/bannisterite, ilmenite, biotite, augite,

feldspar, chlorite and ferrohornblende (Figure 26-27). Ueng (1988) speculated that magnetite, ilmenite and apatite crystallized during fractionation in the slowly cooled sill.



Figure 26. Skeletal titaniferous magnetite in transition gabbro.



Figure 27. Gabbro porphyry with high titaniferous magnetite content.

Stop 8. Top of Hemlock with increased magnetite

UTM: 46° 14.923' N 88° 26.206' W, Sec. 4, 44-33

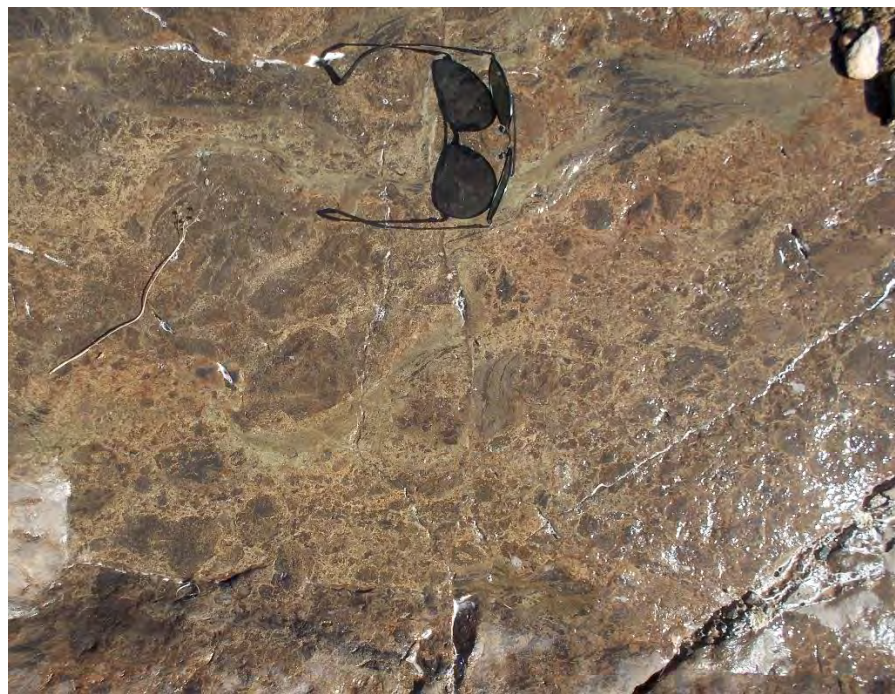


Figure 28. Hemlock agglomerate with iron oxide values above 20% mostly as magnetite.

Stop 8 represents the upper 6000 feet of the Hemlock. Fine grained pillow and fragmental tholeiitic basalts (Figure 28) predominate and contain up to 25% FeO with most of the iron over 12% FeO as magnetite.

Stop 9. West Kiernan Sill-copper/nickel gabbro

UTM: 46° 4.923 N 88° 12.644 W, SENW Sec. 32, 43-31

The majority of the sill consists of a normal gabbro. Plagioclase and augite were the major minerals in the rock. The augite has been altered to hornblende. At this stop the gabbro outcrop contains disseminated pyrrhotite, chalcopyrite and pentlandite (Figure 29). The outcrop exhibits a typical northern climate gossan with shallow oxidation representing low grade disseminated sulfides. A grab sample from this outcrop assayed 0.35% copper/nickel.

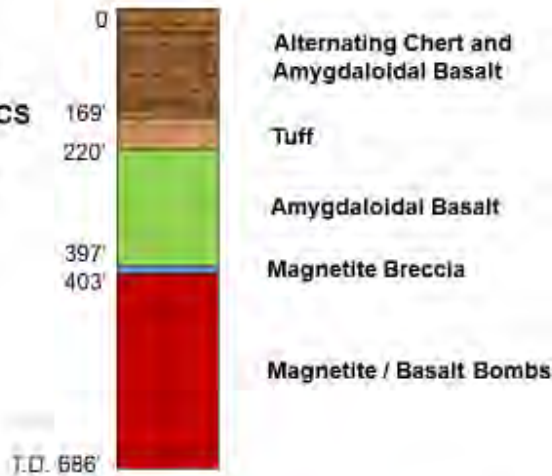


Figure 29. Polished slab showing Chalcopyrite-pentlandite-pyrrhotite gabbro from the West Kiernan sill.

Stop 10. Magnetite bombs in the Hemlock Formation UTM: 46° 5.191' N 88° 12.015' W

In 2005 Prime Meridian Resources drilled a coincident TEM and magnetic anomaly target in the Hemlock Formation east of the Michigamme River and just north of highway M-69. The magnetic anomaly is due to concentrations of magnetite and the conductor anomaly is due to increased amounts of pyrite and chalcopyrite in some parts of the magnetite rich zone. The hole identified as KS-102-2 penetrated alternating chert, tuffs and amygdaloidal basalts bottoming in magnetite rich basalts resembling volcanic ejecta or bombs (Figure 30).

Figure 30.
HEMLOCK VOLCANICS
DDH KS 102-2
IRON COUNTY,
MICHIGAN



Logged by D. Diskin

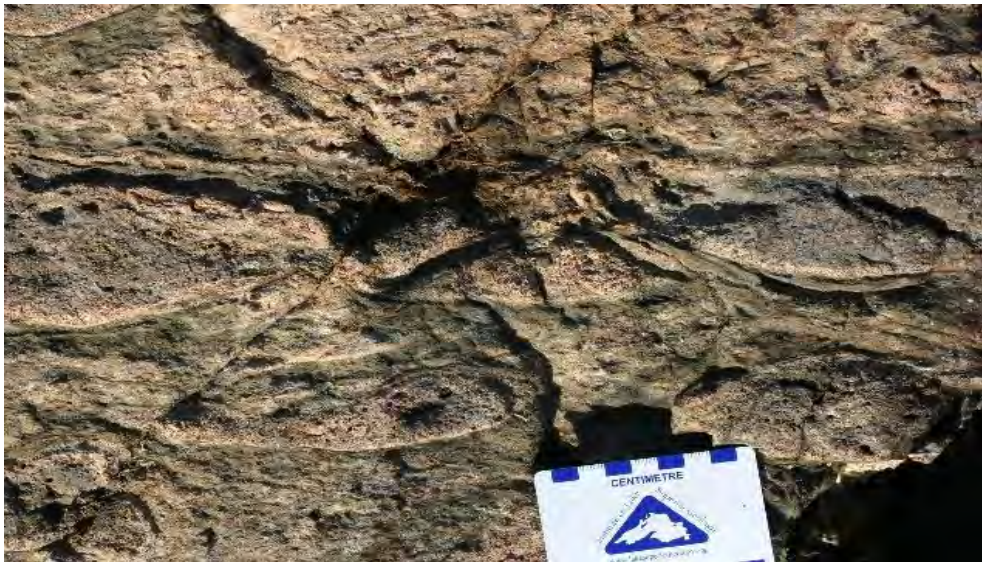


Figure 31. Photo of the outcrop of magnetite bombs with the up direction to the top. Field trip participants will examine diamond drill core representing the siliceous zone and the magnetite rich zone at the bottom of the hole.

The best outcrop example is approximately 1 mile round trip from this location, however, the outcrop is now encrusted in lichen obscuring the salient features seen in the Figure 31. We will examine core from DDH KS-102-2 which exhibits excellent features of the magnetite bombs (Figure 32).



Figure 32. Diamond drill core showing magnetite bombs with albite and calcite filling amygdules.



Figure 33. Polished bomb showing amygdules and groundmass magnetite and titanite.

Individual fragments range in size from 5-7 inches long and 1-2 inches thick and contain variable magnetite concentrations in the groundmass ranging from 15-70% (Figure 33). The groundmass consists of albite, titanite and magnetite. The amygdules have been filled with albite, epidote, calcite and ilmenite. Both the magnetite and titanite are absent from the vesicles supporting the concept that the magnetite was primary and was not involved with later hydrothermal overprint. The size of the magnetite is bimodal with two distinct morphologies present. The first are the small euhedral individual crystals of magnetite while the second type can best be described as aggregates of anhedral magnetite. The distribution of the magnetite in the groundmass is not uniform. In some areas the amount of magnetite approaches 70%. It is suggested the crystallization had occurred by the time it was expelled from the vent. The magnetite was observed to increase in size and amount near the edges of vesicles and between large vesicles. The fine interstitial material between the bombs is devoid of magnetite and titanite indicating a different source for the fine grained interstitial material.

It has been suggested that what appears to be bombs are in reality very small flattened pillows and not bombs at all. This interpretation is a possibility, however, the lack of magnetite in the fine grained material between fragments would be difficult to reconcile pillow creation.

One more significant fact is the magnetite is rather pure and does not contain any titanium suggesting that element separation occurred in the magma chamber. Ueng (1988) also described rhyolite bombs in the central portion of the Hemlock Formation.

Stop 11. Serpentinized peridotite at the base of the West Kiernan sill

UTM: 46° 4.894 N 88° 11.832 W, SWNW Sec. 33, 43-31

Stop 11 represents the basal peridotite of the West Kiernan Sill that has undergone extensive serpentinization. The roadside outcrop on the north side of M-69 exhibits fuzzy outlines of altered phenocrysts of pyroxene. Ueng (1988) did not observe any relic olivine. The process of serpentinization creates abundant secondary magnetite as evidenced by the strong magnetic attraction exhibited by the outcrop. Additional alteration minerals include tremolite, talc,

carbonate, chlorite and some actinolite. Chemistry of the ultramafic rocks are shown in **Table 2** where the MgO values are about 23.8%.

Acknowledgements

I would like to thank Klaus Schulz and Bill Cannon for their constructive conversations and comments on the guide content. Appreciation is expressed to two other reviewers for their helpful review. I would also like to thank Stacy Saari for work on several illustrations. Appreciation is extended to the multiple private land owners for allowing access to several of the field trip stops. During the poster session a display of North American kimberlite samples from the Doug Duskin collection will be on display for comparison with the Lake Ellen kimberlite visited on this field trip. The bibliography at the end of the field trip guide contains many more references than cited in the text to aid anyone who wishes to delve further into the subject in greater detail.

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FIELD TRIP 3

Friday, May 18, 2018

GEOLOGY AND IRON ORES OF THE MENOMINEE IRON RANGE, DICKINSON COUNTY, MICHIGAN

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INTRODUCTION

The Menominee Iron Range was visited on a previous ILSG field trip in 2003. This trip differs substantially from that previous trip and visits mostly new localities with only three exceptions. The following introductory material is reproduced in large part from the 2003 Guidebook (LaBerge et al., 2003). The trip visits most of the stratigraphic units of the area including the Carney Lake Gneiss, the Archean basement on which Paleoproterozoic sedimentary rocks were deposited. The emphasis of this trip is the Paleoproterozoic section and most stratigraphic units will be seen. Several stops are focused on the Vulcan Iron-formation, the principal iron-bearing unit of the Range. Both the unaltered formation and the secondary ores that formed within it will be seen.

The Menominee Iron Range, one of the principal iron producing districts of the Lake Superior region, produced about 260 million tons of high grade iron ore between 1873 and 1946, but has been inactive since. The ores were secondary concentrations of iron oxides and hydroxides within the Paleoproterozoic Vulcan Iron-formation. The ores are generally believed to be paleosupergene concentrations that formed on the Cambrian surface and were covered by late Cambrian sandstone of the Munising Formation. The range also lies very near the Niagara fault zone, the paleosuture between the Superior craton and the accreted Pembine-Wausau arc terrane of northern Wisconsin, and bears the imprint of the strong deformation produced during the accretion.

Stratigraphy. The presently accepted stratigraphic terminology for the Menominee Iron Range was developed by Bayley et al. (1966) and modified only slightly since. The stratigraphic relationships are shown in Figure 2, which is modified from Bayley et al. (1966) to reflect changes in terminology and radiometric age determination since that publication. Precambrian rocks range in age from Archean to Mesoproterozoic. They are capped by Late Cambrian sandstones, which occur as numerous outliers and underlie most of the higher ridges along the Range

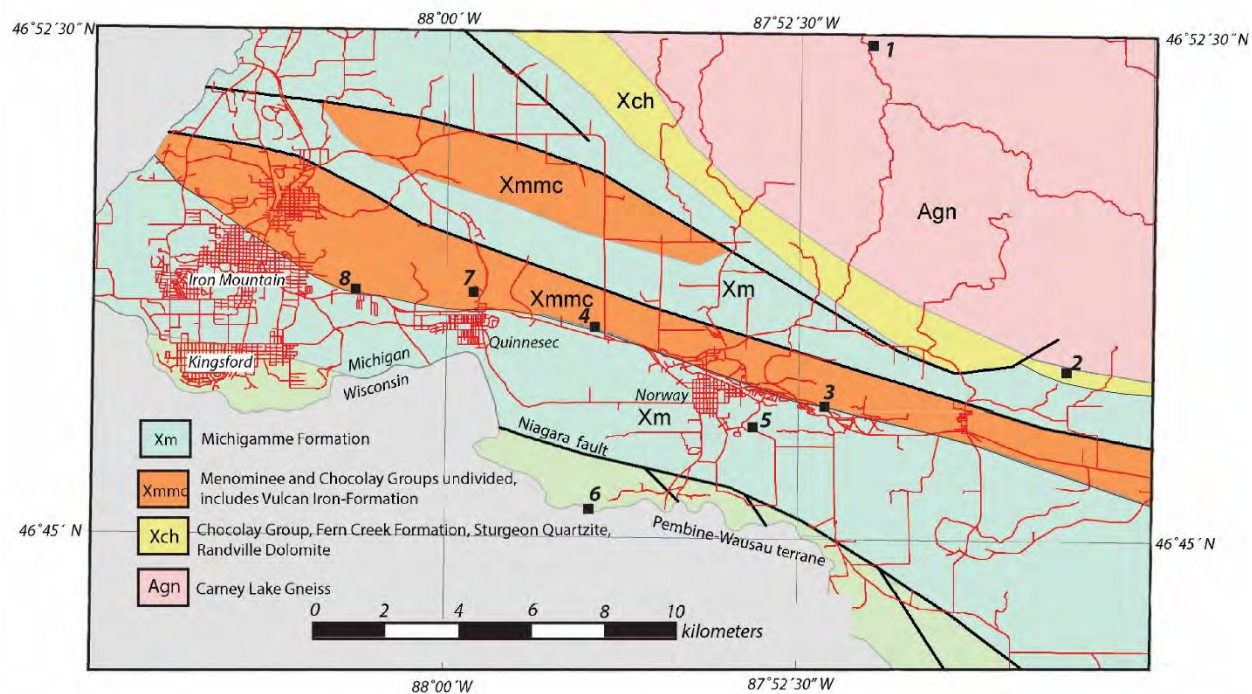


Figure 1. Generalized geologic map of the Menominee Iron Range showing the location of the field trip stops.

Precambrian	Paleozoic	Late Cambrian	Trempealeau Formation, sandstone unconformity Munising Formation, sandstone		
	Mesoproterozoic	Keweenawan	diabase dikes, 1.1 Ga		
	Paleoproterozoic	Intrusive rocks	pegmatite, aplite, quartz dikes and veins, metagabbro dikes and sills		
		Marquette Range Supergroup	Paint River Group, not represented (now believed equivalent to Menominee Group)		
			Baraga Group	Michigamme Formation	
			Menominee Group	unconformity ~1.85 Ga	Loretto Slate Member Curry Iron-bearing Member Brier Slate Member Traders Iron-bearing Member
				Vulcan Iron-formation	
				Felch Formation	
Archean			unconformity Chocolay Group 2.2-2.3 Ga	Randville Dolomite Sturgeon Quartzite Fern Creek Formation	-unconformity?
			Carney Lake Gneiss, complex granitic and mafic gneisses, younger granitic and mafic intrusions. Mostly 2.5-2.7 Ga, but has vestiges of 3.8 Ga gneisses.		

Figure 2. Sequence of formations in the Menominee Iron Range. Modified from Bayley et al. (1966, table 6).

The following summary draws heavily on previous descriptions by Bayley et al. (1966), who provided the most recent comprehensive study of the Range, and LaBerge et al. (2003), who prepared the 2001 ILSG fieldtrip guidebook for the Range. The Precambrian stratigraphic units can be divided into four principal sequences ranging from Archean to Mesoproterozoic. Radiometric age determinations conducted since the previous ILSG guidebook provide new clarity on absolute ages of these sequences.

Archean- Archean rocks form the basement on which the Paleoproterozoic rocks of the Menominee Iron Range were deposited. They underlie a large area north of the Range and are known as the Carney Lake Gneiss. The Carney Lake is a complex, multilithic assemblage of mostly granitic and lesser mafic rocks, which are described more fully in the description of Stop 1. The most recent rock-forming period of the complex was at about 2.75 Ga, but recent age determinations (Ayuso et al., 2017; this volume) have identified cores of zircon grains as old as about 3.8 Ga indicating a very extended history within these rocks.

Paleoproterozoic- The Paleoproterozoic rocks of the Menominee Iron Range are entirely sedimentary and together make up the Marquette Range Supergroup. Three individual groups are present; from oldest to youngest they are the Chocolay, Menominee, and Baraga Groups. Each group is separated by unconformities and the Supergroup, as well, lies unconformably on the Archean Carney Lake Gneiss.

Chocolay Group- The Fern Creek Formation, Sturgeon Quartzite, and Randville Dolomite make up the Chocolay Group. The basal formation, the Fern Creek Formation, is a glaciogenic unit that is preserved only sporadically in the area. Stop 2 is one of the best localities to study it. The stop has been well described by R.J Ojakangas (LaBerge et al., 2003). The medial unit of the Chocolay Group, the Sturgeon Quartzite (see Stop 2 description), forms a continuous blanket of orthoquartzite throughout the area and lies directly on the Carney Lake Gneiss in areas devoid of the Fern Creek. A sericitic unit at the top of the Fern Creek (see Stop 2 description) may be a paleosol indicating a period of weathering between deposition of the Fern Creek and the Sturgeon. The Randville Dolomite, the uppermost unit of the Chocolay Group, is an extensively exposed shallow-water carbonate unit containing numerous stromatolite horizons and other indications of shallow or intertidal deposition. Radiometric ages of detrital zircons and authigenic xenotime have bracketed the depositional age of the Chocolay Group between 2.2 and 2.3 Ga and support its correlation with lithologically similar units in the Huronian Supergroup of Ontario (Vallini et al., 2006).

Menominee Group- The Menominee Group is composed of the basal Felch Formation and overlying Vulcan Iron-formation, the major iron-bearing member of the Menominee Iron Range. The group lies unconformably on the Chocolay Group, although it is structurally concordant with the older rocks in most places. Bayley et al. (1966) describe several localities where there is an angular discordance between the Felch and Vulcan; where there is a basal conglomerate in the Felch Formation it is composed in large part of clasts of the Randville. The absolute age of the Menominee Group in this area has not been determined but regional relationships provide some constraints. To the northwest, the Hemlock Volcanics, part of the Menominee Group and interlayered with iron-formation probably approximately coeval with the Vulcan, have been dated at 1.87 Ga (Schneider et al., 2002). That date indicates that more than 300 million years likely separate the end of Chocolay Group deposition and the deposition of the Menominee Group.

Felch Formation- The Felch Formation is a sericitic slate and quartzite unit that overlies the Randville Dolomite. It consists of thin-bedded sericitic slate and phyllite and intercalated thin-bedded quartzite, with the quartzite layers being more prevalent near the top of the formation (Bayley et al., 1966). It is about 100 feet thick on the south range but is as much as 500 feet thick on the north range. Bayley et al. (1966) considered the Felch Formation to be correlative with the Ajibik Quartzite and Siamo Slate of the Marquette district and the Palms Formation of the Gogebic district. The Felch Formation is conformable and gradational with the overlying Vulcan Iron-formation.

Vulcan Iron-formation- The Vulcan Iron-formation is the major iron-bearing unit of the Menominee district. It is well known from numerous mines and drill holes, but generally is not well exposed in natural outcrops. The iron-formation is divided into four units, two composed mainly of granular iron-formation and two composed of slate and slaty iron-formation. In succeeding order the units are the Traders Iron-bearing Member, the Brier Slate, the Curry Iron-bearing Member, and the Loretto Slate. They are described in detail by Bayley et al. (1966). The Traders and Curry Members contain layers of granular jasper alternating with layers of magnetite and hematite. The Brier and Loretto Members are mainly laminated siliceous iron-rich slate, which locally contains laminae of detrital quartz, feldspar, micas, zircon, and tourmaline. According to Dutton (1958), the iron-formation is about 1,000 feet thick, of which about 730 feet is ferruginous slate (Brier Slate - 330 feet, Loretto Slate - 400 feet) and 270 feet is granular iron-formation (Traders - 100 feet, Curry - 170 feet). The stratigraphy within the Vulcan is seen well at Stops 3, 7, and 8.

A detailed stratigraphic section (from oldest to youngest) from the Curry Mine located between the towns of Vulcan and Norway shows that above the Randville Dolomite there is a 20 foot conglomerate consisting of novaculite (dense, fine-grained siliceous rock resembling chert) boulders and smaller angular fragments cemented by ferruginous silica. A similar rock is seen at Stop 8. A fault breccia occurs in two zones bordering a dolomitic slate, over a 25-foot interval. Above the fault breccia is 26-foot interval of vitreous quartzite and then 69 feet of Felch Formation that is divided into several horizons including quartz slate, blocky green slate, massive brown chert, blocky brown talcose slate, shaly brown talcose slate, and topped by the "Trader's quartzite" (informal terminology). The Vulcan Iron-formation lies on top of the quartzite and is 111 feet thick with several horizons noted; ferruginous slaty iron-formation, wavy-bedded red cherty iron-formation, massive wavy iron-formation with brownish red chert lenses, even-bedded iron-formation with brown chert beds, and the uppermost unit is a massive brown granular chert horizon. The Brier slate is in fault contact with the Traders Iron-bearing Member. It is grey to brown (oxidized) laminated slate, 100 feet thick that is also in fault contact with the Curry Iron-bearing Member. The Curry is 158 feet thick with a thin basal slaty phase and thick even-bedded, dark reddish purple granular chert with specular hematite laminae and in cross fractures. The Loretto Slate is the next formation horizon at about 45 feet thick and bounded by sheared contacts. It is a dark brown, thinly laminated, blocky ferruginous slate. The "Hanbury slate" (Michigamme Formation) overlies the Loretto and in the 5th level of the Curry mine consists 405 feet of ferruginous mottled red and white slate, then a greenish grey thinly laminated slate with a high chlorite content, and finally a soft pyritic black carbonaceous slate with graphite on shear planes. This stratigraphic section from the Curry Mine is the most complete sequence known for the Range and was developed by Penn Iron Mining Company geologists.

Baraga Group- The Baraga Group consists of a single unit, the Michigamme Formation. The belts underlain by the Michigamme Formation are very poorly exposed, which accounts, at least in part, for the lack of detailed mapping of what may well be otherwise discernible map units. According to Bayley et al. (1966), the Michigamme Formation consists chiefly of slate, especially quartzose, micaceous, and graphitic varieties, subgraywacke, quartzite, conglomerate, dolomite, dolomitic quartzite, and some iron-formation. More recent exploration drilling also has identified units of mafic volcanic rocks. An unconformity between the Michigamme and underlying Vulcan Iron-formation is indicated by the presence of basal a conglomerate, reported from a few localities, that contains clasts of iron-formation and other Menominee and Chocoday Group lithologies, and by regional truncation of pre-Baraga Group units beneath the basal Michigamme units. Although the Michigamme Formation lies on the Vulcan Iron-formation along both the north and south ranges, the Vulcan is largely absent to the north. The stratigraphic section bounding the Archean Carney Lake Gneiss consists of only the Chocoday and Baraga Groups, with the Menominee Group absent except for the extreme eastern end of the area. These relationships suggest that a topographic high existed to the north of the Menominee Range during or shortly after the time of Menominee Group deposition.

Mesoproterozoic- The only rocks of Mesoproterozoic age are thin dikes of unmetamorphosed and undeformed diabase of probable Keweenawan age. They are known mostly where they cut the Carney Lake Gneiss. Typical dikes are only a meter or two wide and commonly have chilled margins against the rocks into which they are intruded.

Structure. The Menominee iron district (Figure 3) is a south-facing homocline of Paleoproterozoic strata in which stratigraphic repetitions are created by two major faults and by folding internal to fault slices (Bayley et al., 1966). The faults cut the folds longitudinally, approximately along the fold axes, repeating the Paleoproterozoic sequence three times. The structural elements of the Range are shown in Figure 3, reproduced from Bayley et al. (1966, figure 22). The 3-D geometry is shown in Figure 4, reproduced from Bayley et al. (1966, figure 23). On the north, the Carney Lake Gneiss forms the core of a broad anticlinal structure (Figure 3). The Paleoproterozoic strata lie unconformably on the gneiss and dip steeply to the south or are overturned (as at Stop 2) and dip steeply north and face south. Interestingly, on this northernmost sequence of strata the Menominee Group, including the Vulcan Iron-formation, is absent and the Michigamme Formation lies directly on the upper unit of the Chocoday Group. This suggests that there was uplift in the area of the Carney Lake Gneiss concurrent with or shortly after deposition of the Menominee Group, creating a topographic high. In the south, the Paleoproterozoic strata are repeated twice by major faults to form the two ranges of the district. These faults were named the North Range fault and South Range fault by Bayley et al. (1966). The faults have steep dips at the present level of exposure and consistently show southside-up displacement. More recent interpretations (e.g., Sims and Schulz, 1993) consider them to have been thrust faults, which were steepened by continued shortening of the thrust panels. The rocks in the hanging wall (south side) of these faults have no indications of Archean basement rocks, in contrast to the area immediately to the north where the Carney Lake Gneiss is an integral part of the structure. The north range and south range panels may be allochthons detached from basement and thrust northward over the more autochthonous sequence of the northern part of the district. The Menominee Range is bounded on the south by the Niagara fault, along which it is in contact with volcanic rocks of the Wisconsin magmatic terranes.

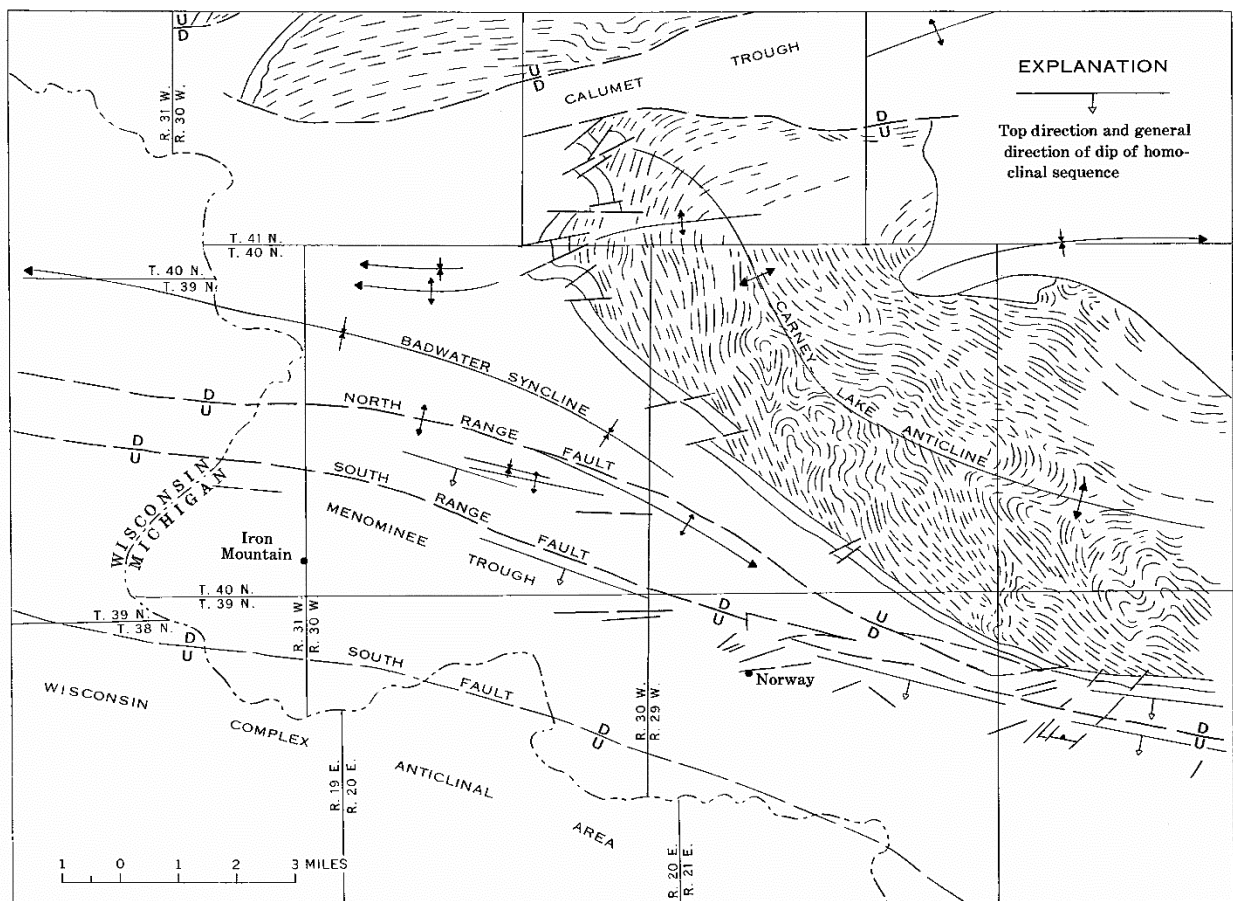


Figure 3. Structural elements of the Menominee Iron Range from Bayley et al. (1966, figure 22). The “south fault” is now referred to as the Niagara fault and is recognized as the suture between continental margin assemblages to the north and the accreted Wisconsin Magmatic Terranes to the south.

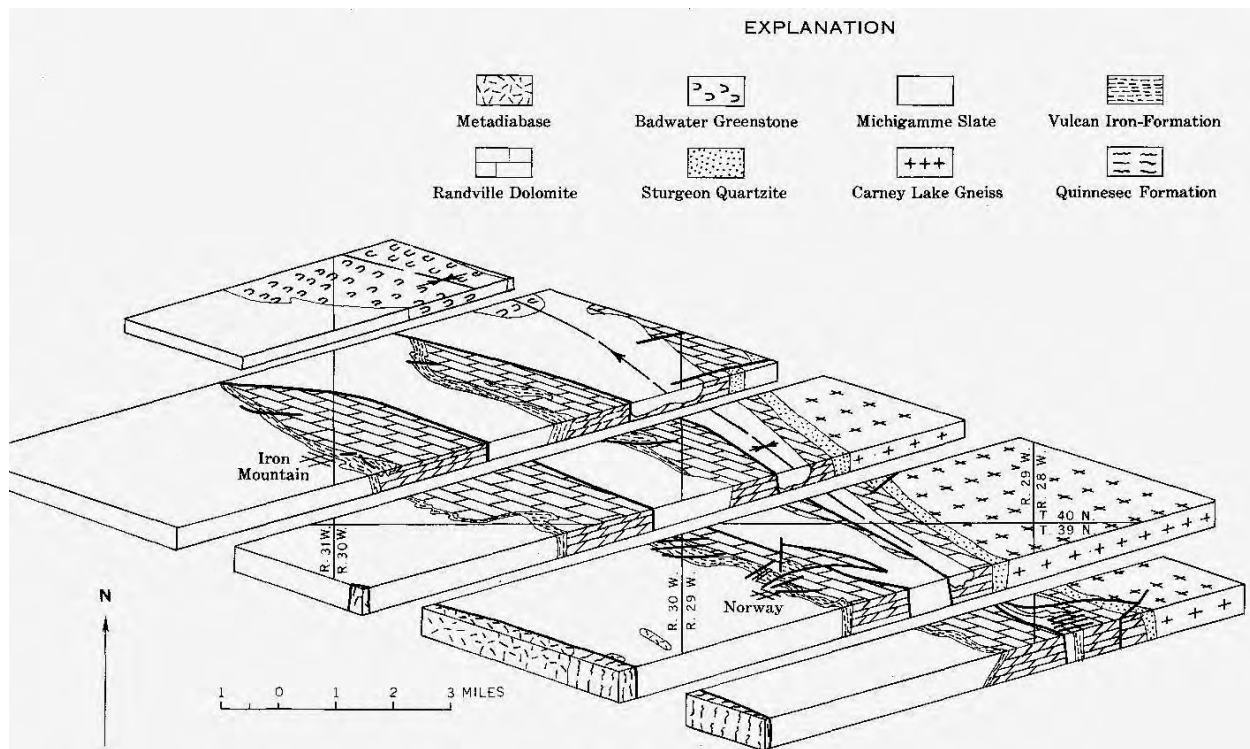


Figure 4. Block diagram showing distribution of stratigraphic units of the Menominee Iron Range, from Bayley et al. (1966, figure 23).

Iron deposits

Iron ore was discovered in the Menominee district in 1848 by two explorers, J.W. Foster and S.W. Hill, according to Winchell (1895). However, iron mining did not begin until 1870, when N.P. Saxton started digging pits and trenches on the site of the Breene Mine, with the first ore being shipped in 1873 (Bayley et al., 1966). All the major mines had been opened by 1878. Production continued until 1946, with a total production from the district of approximately 85,000,000 tons (Bayley et al., 1966). Seven mines produced nearly 77,000,000 tons of ore, with a majority of the production from the district coming from three mines, the Chapin (27,500,000 tons), the Penn (21,700,000 tons) and the Aragon (11,200,000 tons) (Dutton, 1958). Production from the Chapin Mine ended in 1934 with a major collapse of the workings. The subsidence from this collapse formed the lake on the north side of Iron Mountain. A causeway across the lake now carries the traffic on Highways US 2 and US 141. Ore from the district was hauled by rail to Escanaba, Michigan; from there it was carried by boat to steel mills on the lower Great Lakes. The majority of the ore shipped from the district was high-grade (>50 % Fe) natural iron ore. Some 'siliceous hematite' ore (40-50% Fe) was produced from the Millie Pit and the Traders Pit. The Traders ore was used at one time for an experimental project at the Ardis Furnace where an attempt was made to high grade the ore utilizing high temperature roasting. The ruins of the Ardis Furnace are now on the National Register of Historic Places and can be visited near downtown Ironwood. The last of the mines in the area was the Groveland Mine in the Felch trough that operated until the early 1980's using beneficiation methods to process both hematite and magnetite ore with complex iron silicates to produce a concentrate.

Although the iron-formation in the Menominee district was studied as a possible source of beneficiating ore ("taconite ore"), no commercial operation has been undertaken. In the early 1950's the Oliver Mining Company completed extensive diamond drilling and evaluation of underground ore reserves in the Vulcan – Norway portion of the Range and had designed open pit operations to extract ore from the Curry and Traders Iron-bearing Members. The area included the Curry, Brier Hill, and Aragon shafts which were the deepest mines on the Range at over 2000 feet.

FIELD TRIP STOPS

Stop 1: Carney Lake Gneiss (45.873°N, 87.86°W)

The Carney Lake Gneiss occurs north of the Menominee Iron Range and forms the Archean basement on which the Paleoproterozoic strata of the Range were deposited. The Carney Lake Gneiss was defined and described by Bayley et al. (1966) who mapped the unit in some detail and published maps and lithologic descriptions of it, but did not attempt to decipher its obviously highly complex internal history. The general descriptions of the Carney Lake below are extracted from that publication.

According to Bayley et al. (1966, p. 20-29) *"Granitic gneiss constitutes about 85 percent of the Carney Lake Gneiss; of the remainder, about 5 percent is granodiorite and syenite dikes, and about 10 percent is inclusions of older rock. The gneiss is not uniform in composition or appearance, but varies from a gray plagioclase-biotite gneiss to red microcline-biotite gneiss. For the purpose of discussion these types will be designated gray gneiss, composite gneiss, and red gneiss, respectively."*

"The gray gneiss probably constitutes about 25 percent of the complex collectively called the Carney Lake Gneiss. It is most abundant in the northern half of the complex where it contains many amphibolite inclusions. In thin sections the gray gneiss shows abundant plagioclase, quartz, and biotite. The foliation is well shown by aligned biotite, and also by the plagioclase and quartz, which are arranged in subparallel elongated grains and in lenticles. Cataclastic structures are common."

"The composite gneiss constitutes at least 70 percent of the complex. It is present almost everywhere, but is more abundant in the southern half of the area, where it contains minor patches of red gneiss and many inclusions of biotite schist. ... The grain size of the composite gneiss ranges from medium to coarse. The gneiss is streaky and consists of red and gray elements, the red parts composed of pink microcline and quartz, the gray parts chiefly of plagioclase and biotite, which are the same minerals that constitute the gray gneiss. At some places the red part forms patches and streaks within the gray, at others the gray is enveloped by the red, and at still others the two elements form alternating layers. The red part commonly occurs as veins or layers of coarse pegmatite which cut across the foliation or bifurcate and join other layers. Here and there veins and layers of the red material swell and form large pods of pegmatite which fade transitionally into the gray rock. Pegmatite pods may also pinch and swell along the strike of the foliation of the gneiss. Locally they stop abruptly against the gray rock, only to appear again further along the strike. ... The red gneiss is medium grained and weakly

foliated. Fresh specimens are pink or red, whereas weathered specimens are brownish pink. Like the composite gneiss, the red gneiss consists of two components of different age, an older part composed of extensively altered plagioclase, and a younger part that consists of quartz, microcline, muscovite, and minor amounts of albite, but the red gneiss generally contains more quartz and microcline and less biotite and plagioclase than the other gneisses.”

“Granodiorite occurs as rare dikes that are most abundant in the southern half of the complex and are more likely to be closely associated with the composite gneiss than the other types. The granodiorite is massive, equigranular, pink, medium to fine grained, and brownish-pink weathering.”

“Inclusions in the Carney Lake Gneiss constitute less than 10 percent of the complex, but bear importantly on the character and origin of the gneiss. They consist of amphibolite, biotite schist, and metasedimentary rocks. The inclusions are clearly older than the gneiss and may represent (1) engulfed parts of the pre-gneiss Dickinson Group which occurs to the north of the complex consists, in part, of a metamorphosed series of basic tuffs, graywacke-type deposits, and basaltic flows (James et al., 1961), or (2) engulfed parts of the Quinnebec Formation which occurs to the south of the complex and consists of metavolcanic rocks, greenstone, amphibolite, and schist.” (Note: Radiometric ages determined since the Bayley et al. (1966) report indicate that both the Dickinson Group and Quinnebec Formation are younger than the Carney Lake Gneiss.)

“The field relations show that (1) the gray gneiss grades into composite gneiss, into inclusions of amphibolite, and, more rarely, into inclusions of biotite schist, (2) this gneiss exhibits sharp contacts against the inclusions and appears as dikes in them, (3) the composite gneiss grades into red gneiss and into biotite schist, and (4) both the composite gneiss and the red gneiss contain inclusions of biotite schist and occur as dikes and stringers in some of the inclusions. Further, the gneisses are cut by red granodiorite dikes; one of these dikes, in turn, is cut by a late middle Precambrian metadiabase dike, and another metadiabase dike contains an inclusion of granodiorite. The relations of the syenite, which is known only in the southeast corner of the complex, and the granodiorite dikes are not clear, but the lack of foliation of the granodiorite dikes and the slight foliation of the syenite may indicate that the syenite was emplaced before the granodiorite dikes.”

The descriptions in Bayley et al. (1966) and the relatively cursory examination that we have so far conducted in the Carney Lake Gneiss make it obvious that these rocks contain a very extended and complex history. In particular, our recent documentation of zircon grains with cores as old as 3.8 Ga (Ayuso et al, 2017; this volume) indicates that these rocks are undoubtedly part of the Gneiss Terrane defined by Sims et al. (1980) and contain vestiges of Eoarchean crust.

Geochronology of samples of the Carney Lake Gneiss done using the USGS/Stanford Sensitive High-Resolution Ion Microprobe (SHRIMP) produced U-Pb data on zircons that confirms an Archean age (Ayuso et al., 2017; this volume). Two samples were collected for radiometric dating from the southern half of the complex: 1) sample 1 is from a granitic K-feldspar-bearing gneiss that is locally pegmatitic; 2) sample 2 is from a banded and folded gray to red granitic gneiss. Abundant zircons (70-200 μ) were obtained from sample 1 that range from anhedral to subhedral, contain complex igneous and irregular growth zoning, and multiple growth rims;

these zircons have irregular to pyramidal overgrowths. The zircons from sample 2 range from slightly rounded to subhedral and are otherwise mostly similar to zircons from sample 1. One hundred and twenty nine analyses of cores and rims were obtained. Individual zircons have older ages near their cores (mostly discordant) and younger ages near their rims. On a concordia diagram (Figure 5), U-Pb data plot as clusters of data points ranging from concordant to discordant and suggest several chords and intercepts that are common to both samples from the Carney Lake Gneiss (Figure 5). That study identified cores of individual zircons as old as 3.8 Ga. The most common age for individual zircons and for rims on older grains is about 2.75 Ga and records a younger major event in the late Archean (Figure 6).

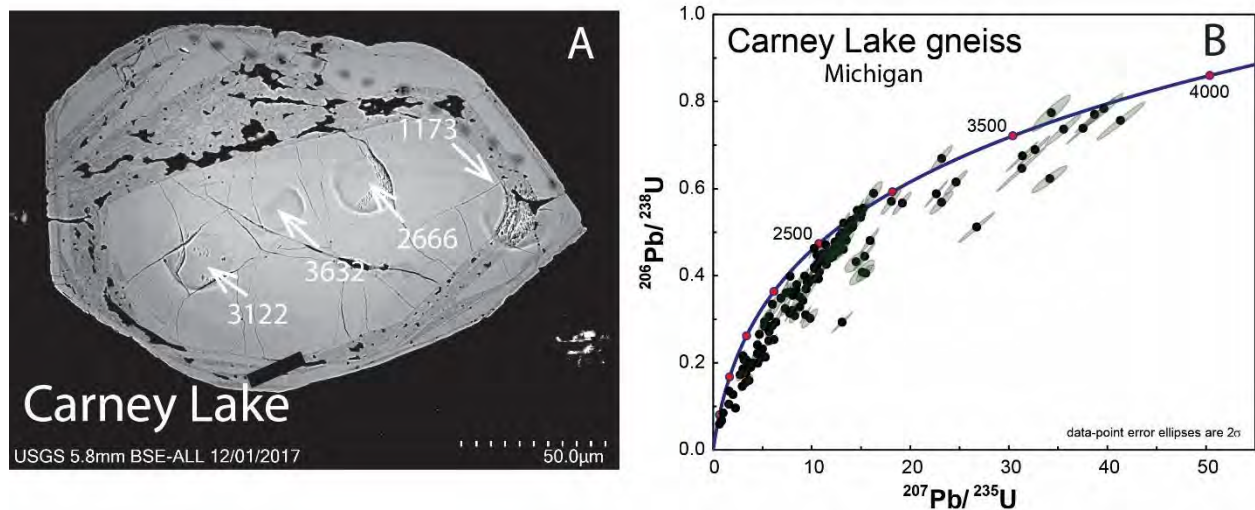


Figure 5. A-BSE (back scatter electron) image of a zircon from the Carney Lake Gneiss showing ages of four analyzed spots. B- Concordia diagram for 129 spot analyses from zircons in the Carney Lake Gneiss.

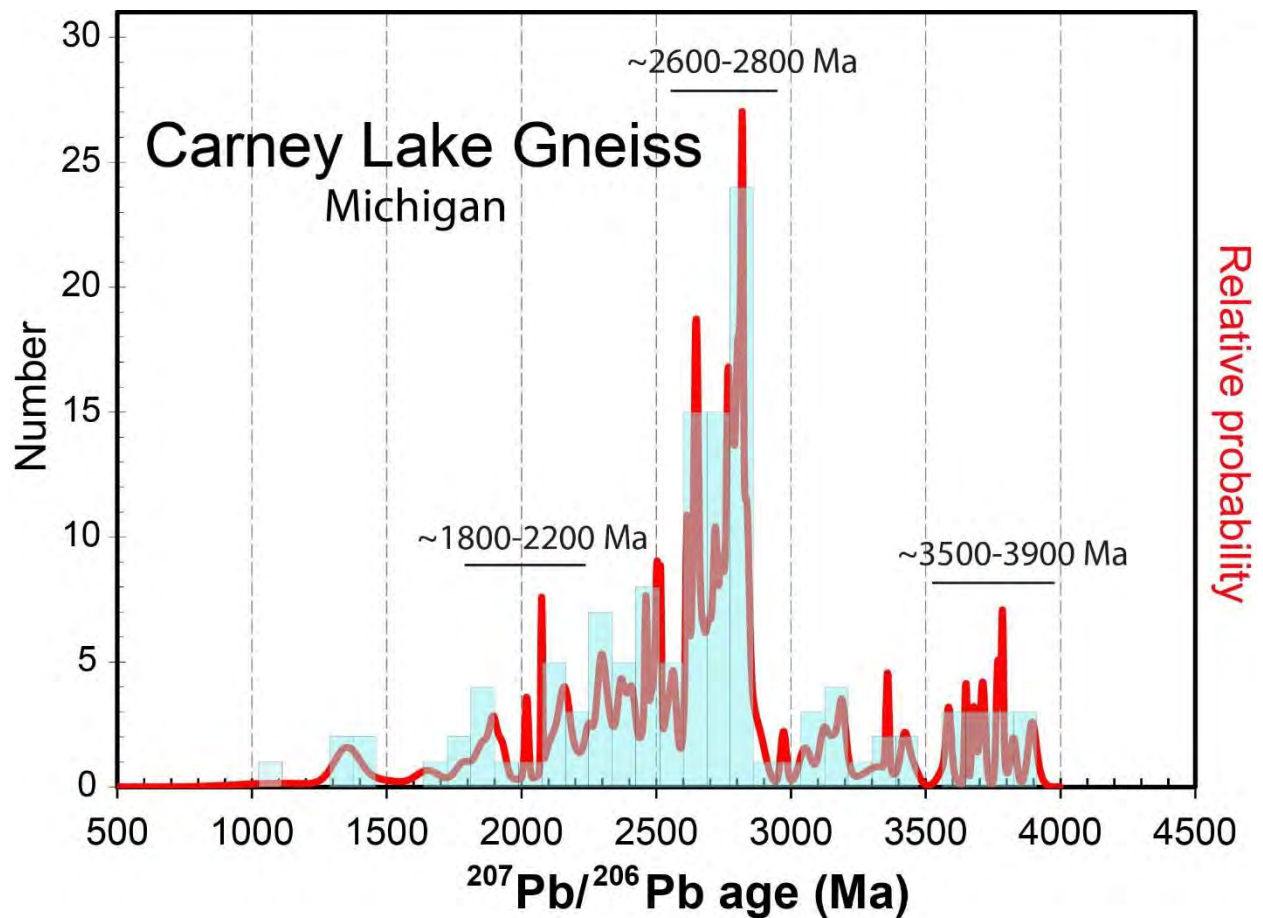


Figure 6. Histogram showing the distribution of individual SHRIMP spot analyses of zircons from the Carney Lake Gneiss. Shaded areas are number of analyses. Solid line is relative probability.

The gross structure within the Carney Lake Gneiss as mapped by Bayley et al. (1966) is a dome elongated in an east-west direction as defined by foliation and compositional layering of the gneisses. However, the age of the doming event is uncertain. Basal Paleoproterozoic strata surrounding the dome are generally steeply dipping, in part overturned, and mostly concordant with the contact with the Carney Lake, indicating that much of the doming post-dates deposition of those strata that are as young as about 1850 Ma. Thus, much of the internal structure of the Carney Lake likely has a strong Paleoproterozoic imprint superimposed on a complex set of Archean structures.

A series of outcrops along a powerline east of Norway Truck Road provides a good example of various lithologies and structural complexity of the Carney Lake Gneiss. Figure 7 shows locations for outcrops and locations of photos in Figure 8. In general, the gneisses are more amphibolitic to the west and become progressively more granitic to the east, although a great deal of finer-scale complexity also is seen here. A few thin (1-3 m) mafic dikes that are younger than the complex gneissic structure can also be seen.

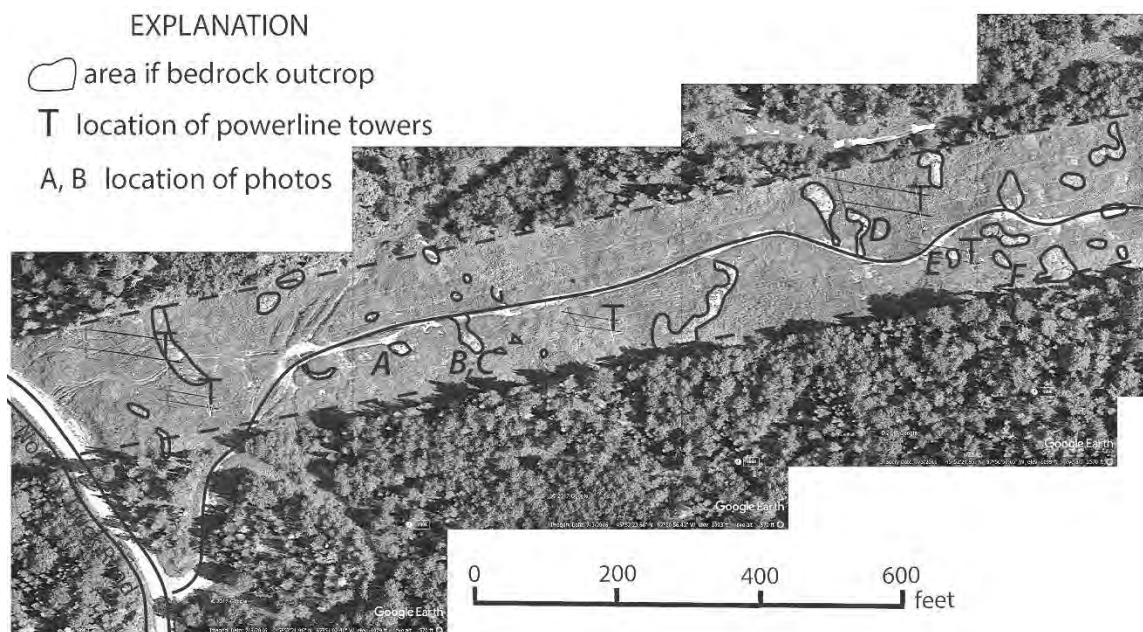


Figure 7. Map showing location of outcrops of Carney Lake Gneiss along a powerline and location of photographs shown in Figure 8.

A



B



C



D



E



F



Figure 8. Photographs of Carney Lake Gneiss along powerline. Locations shown on Figure 7.

A-Amphibolitic gneiss cut by weakly deformed pegmatites. B-Contorted granitic gneiss with amphibolite inclusions. C-Amphibolitic gneiss with granitic stringers. D-Granitic gneiss with

moderately dipping foliation. E-Foliated amphibolitic gneiss cut by undeformed pegmatite. F-Banded gneiss with intrafolial isoclinal folds.

Stop 2: Sturgeon River locality. Archean basement, Fern Creek Formation, and Sturgeon Quartzite. (45.784°N, 87.789°W)

(This description is modified slightly from a previous field trip guide (LaBerge et al., 2003) that was written by Richard Ojakangas. This locality also has been modified in recent years by removal of a previous hydroelectric dam and draining of the impoundment. As a result, some additional outcrops have been created, especially of the Carney Lake Gneiss, but these are not the focus at this stop. The following description refers to the dam for locational purposes and is still useful in that vestiges of the dam can still be seen.)

The Fern Creek Formation and Sturgeon Quartzite are the lower two formations of the Chocoday Group. The group is discontinuous and has been recognized in parts of the Marquette Iron Range and Gogebic Iron Range as well as here in the Menominee Range. The age of the group is bracketed between about 2.2 and 2.3 Ga based on ages of detrital zircon grains and hydrothermal xenotime (Vallini et al., 2006). The group appears to be equivalent to lithologically similar formations in the Huronian Supergroup in Ontario. Although the Sturgeon Quartzite is essentially continuous along the Menominee Range, the Fern Creek is preserved only locally, one of the best exposures being at this stop. Sericitic sediments near the top of the Fern Creek have been interpreted to be reworked paleosols formed prior to deposition of the Sturgeon Quartzite as discussed below (originally in Ojakangas' stop description). If true, there is a disconformity between the Fern Creek and Sturgeon, which may account for the very limited preservation of the Fern Creek.

Here the Sturgeon River has cut a deep gorge through the Sturgeon Quartzite; the formation was named for this locality. This small area has been well studied, especially because of the presence of the Archean-Paleoproterozoic contact at the dam. The area has been described by Credner (1869), Brooks (1873), Rominger (1881), Irving (1890), Bayley (1904), Lamey (1937), Pettijohn (1943), and Trow (1948).

Substop 1. Walk past the gate to the end of the road at the powerhouse and dam. We will traverse back up the road to the vehicles, thus observing the rock units in stratigraphic sequence. The dam was constructed on Sturgeon River Falls, which was held up by a thick mafic dike that can be seen in the woods off the east end of the dam. The unconformity between the Archean Carney Lake Gneiss and the Paleoproterozoic Fern Creek Formation can be seen in a small ground-level exposure adjacent to the dam (Figure 9). The lowest bed in the Fern Creek is a diamictite at this spot, whereas a short distance to the west on the river bottom by the power station, the lowest unit is arkosic sandstone with rare oversized stones.



Figure 9. Unconformity at Sturgeon Dam. Hammer head rests on Archean Carney Lake Gneiss and hammer handle is on basal diamictite of the Fern Creek Formation. Nearby in the river bottom, the basal unit of the Fern Creek is arkosic sandstone with rare dropstones, illustrated in Figure 11.

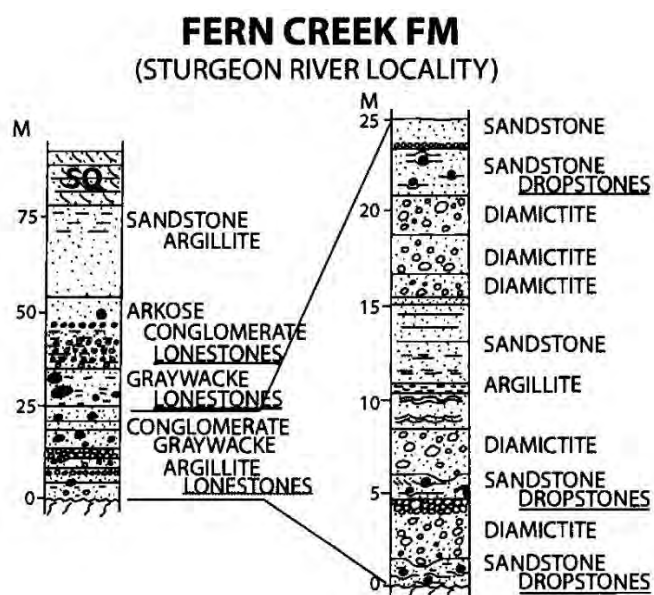


Figure 10. Stratigraphic column at the Sturgeon River locality. SQ at the top of the column designates Sturgeon Quartzite.

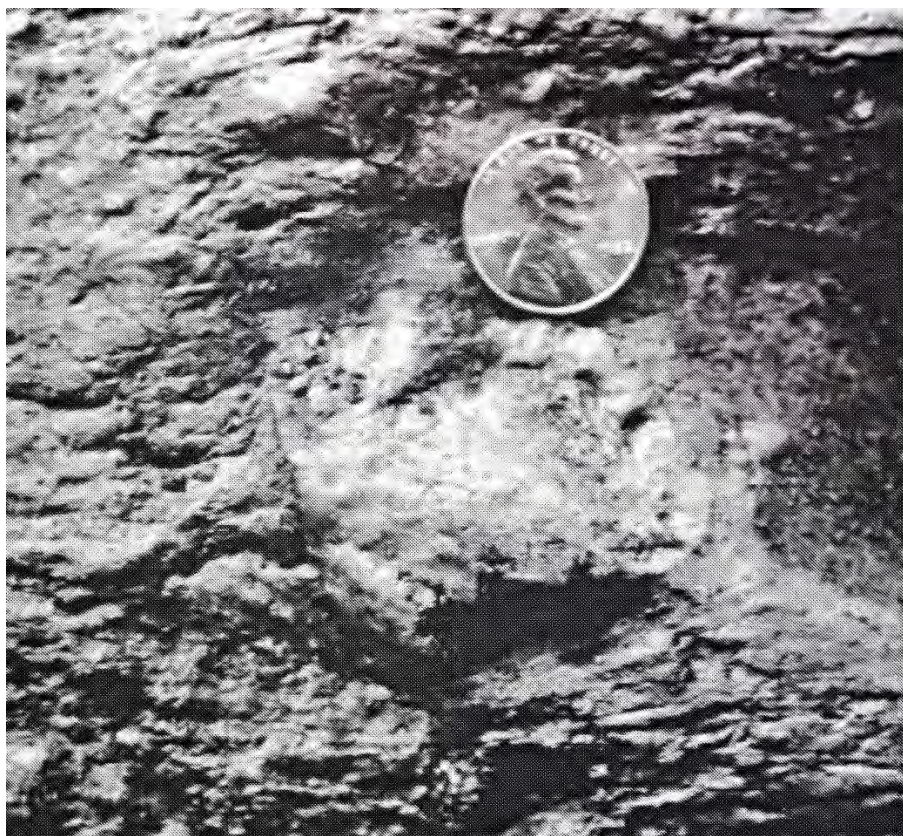


Figure 11. Granitic dropstone in lowest sandstone of the stratigraphic column. Note that the stone has pierced and bowed down the underlying strata.

Figure 10 is a measured column of the Fern Creek Formation. The lower 25 m are well exposed when there is no water in the channel. Note that this portion of the formation consists of five beds of diamictite (matrix-supported conglomerate) as thick as 2.5 m, three arkosic sandstone beds as thick as 2.6 m with rare oversized stones, stacked arkosic sandstone beds with minor intercalated siltstone and argillite laminae, an argillite bed 4.5 cm thick, and a 15 cm conglomerate.

Interestingly, the well-exposed section seen in the river bottom is not found on the west bank of the river; only 1½ m of conglomeratic rock is present there. Apparently, the more complete section is preserved in a topographic low on the Archean surface. However, faulting may be a factor as well, for weathered pyrite is present along a fault between the Archean basement and the Fern Creek west of the powerhouse.

The middle 25 m of the Fern Creek Formation is relatively poorly exposed; Figure 10 shows this portion consisting of conglomerate, graywacke sandstone with oversized stones, and arkosic sandstone with oversized stones.

Interpretation: This is a glaciogenic sequence. The diamictites may be thin tills deposited beneath glacial ice, but more likely are debris flow deposits as suggested by one diamictite bed that grades upward into sandstone. Some of the conglomeratic beds are difficult to clearly classify as

either matrix-supported or clast-supported. One 20 cm bed at the 15 m level in the section is graded from medium sand to clay, suggestive of a turbidity current mechanism. Several of the oversized stones in the sandstone and greywacke beds show either a bowing down of the underlying laminae or an actual penetration, indicating that the stones were dropped into the basin from above and are indeed dropstones. Other limestones may be dropstones, too, but clear evidence is lacking. The likely mechanism for deposition of dropstones is release from melting icebergs or from a floating glacier.

Substop 2: The 25 m section between 50 and 75 m on Figure 10 is intermittently exposed on the west bank of the river, but this area is usually inaccessible because of high water. It includes beds of sericitic quartzite interbedded with sericite schist. The sericitic nature of this interval is illustrated by a small road-level outcrop between the road and the river just north of the quartzite ridge. This is a sericitic quartz pebble conglomerate with sericite clay chips, some reddish rather than yellow-green in color.

Interpretation: This sericitic portion of the column is interpreted as a reworked paleosol that formed on the Fern Creek Formation during a warm climatic period that followed glaciation. Trow (1948) first suggested that this might be a paleosol.

Substop 3: Sturgeon Quartzite ridge. Note that the bedding is slightly overturned towards the south, and that cross-bedding indicates that stratigraphic tops are to the south. Cross-bedding is of both trough and planar types. According to Trow (1948), the general cross-bedding indicates a paleocurrent trend from the northwest toward the southeast. Since the original field trip guidebook was prepared (LaBerge et al., 2003), geochronological studies (Vallini et al., 2006) have constrained the age of the Sturgeon Quartzite, and by inference of the underlying Fern Creek Formation. Most detrital zircons have ages between 2.5 and 2.7 Ga, but there is also a well-defined cluster of ages at about 2.3 Ga, thus providing a maximum age of deposition. Xenotime overgrowths on zircon grains are as young as 2.1 Ga and define a minimum age. These ages are consistent with age ranges determined for equivalent units in the Marquette and Gogebic Iron Ranges.

Interpretation: Abundant asymmetrical ripple marks have low ripple indices (wave length/ripple height) indicative of deposition by water rather than by wind. The beds are generally of even thickness, indicative of a shallow marine rather than a fluvial environment of deposition.

Stop 3. Underground tour of the Iron Mountain Iron Mine. (45.782°N, 87.864°W) The Iron Mountain Iron Mine has been in operation as a tourist locality for 60 years and thousands of people have enjoyed this historic site (Figure 12). The #2 adit is one of three exploration tunnels driven perpendicular to the strike of the Vulcan Iron-formation in search of high-grade (>50% Fe) ore in the early 1870's along the south side of Brier Hill (Figure 13). The #1 adit was driven north to intercept at depth high grade ore that cropped out at the surface to the west of #2 adit. The first iron ore production came from this operation which was a combination of open pit and underground mining. The exploration then turned east along strike to explore the Traders Iron-bearing Member of the Vulcan Iron-formation, but lost the member due to longitudinal faulting. The #2 adit was completed and crossed several faults and duplication of beds to find the Traders Iron-bearing Member 1,000 feet into the hillside. (Figures 14, 15). Further exploration sited the #3 adit a half a mile to the east and intercepted the complete section of "Hanbury Slate"

(Michigamme Formation of present usage), Curry Iron-bearing Member, Brier Slate Member, and Traders Iron-bearing Member. The #3 East Vulcan Shaft was located based on finding high grade ore at this locality.



Figure 12. The gateway to paradise. Portal to thousand-foot-long adit of the Iron Mountain Iron Mine.

The tour will allow for the observation of a complete stratigraphic section from the “Hanbury Slate” (Michigamme Formation) at the entrance of the adit, through the Loretto Slate, Curry Iron-bearing Member into the Brier Slate Member (duplicated by folding and faulting), then through the Traders Iron-bearing Member and terminating at the “Footwall Slate”, which here is a breccia. These explorations were all drilled with hand-held drill bits and sledge hammers. Black powder was used for blasting and all material was hand loaded into tram cars and pulled with mules.

The tunnel next was turned west to follow the contact of the Traders Iron-bearing Member and the footwall, but only goes about 70 feet before it intercepted another fault. The iron formation is brecciated and it is believed that the tunnel is near the subcrop of the formation with the overlying Cambrian sandstone because much of it is filled with friable sand through this section. The tunnel meanders a little and gains some elevation while intercepting brecciated and shallow dipping broken banded iron-formation as it comes to an intersection where tunnels go off in three directions (Figure 14).

The tunnels to the north and west explore a wedge of Traders Iron-bearing Member and outlined a block showing low angle dips to the south. At this point a small ore body was found and extracted leaving a stope with broken slabs of rock. Along the south edge is a decline filled with spring water that wraps around a pillar. The stope connects to a shaft about 30 feet deep that was used to determine if more ore occurred at depth. The west tunnel explores the Traders Iron-bearing Member about 500 feet along the East Vulcan longitudinal fault and terminates at the zone of caving to the west. The south tunnel crosses the strike of the Traders Iron-bearing Member and Brier Slate Member again and intercepts another fault where high grade ore is located. This passage was developed from the Old Central Shaft which was sited near the #1 adit. The extraction of ore at this location using sublevel caving methods left this large void (stope) as the ore was removed from below and the waste rock allowed to settle and partially prop up the workings.

Several items should be noted: In the large stope, the unconformity between the iron formation, which strikes N 75° W and dips 70° SW, and the horizontal sandstone is striking. The ground water rain can be heard when it is quiet. The mined bottom of this ore block is about 600 feet to the southeast where it is cut by the large longitudinal fault and offset to the south. At that point in the operations, a 'New' Central shaft (Figure 18) was completed and used to extract ore to the 1200 foot level and develop ore to 1,600 feet toward East Vulcan #4 shaft. That was the extent of mining at the end of WWII when it became too costly to mine these orebodies without significant upgrades to equipment. Mines to the west: the West Vulcan, Curry, Brier Hill, and Aragon, were connected by tunnels to this property and went to depths of 2,400 feet.



Figure 13. Plat of the Central Vulcan are in 1938. Geological interpretation and mine sites including subsidence area.

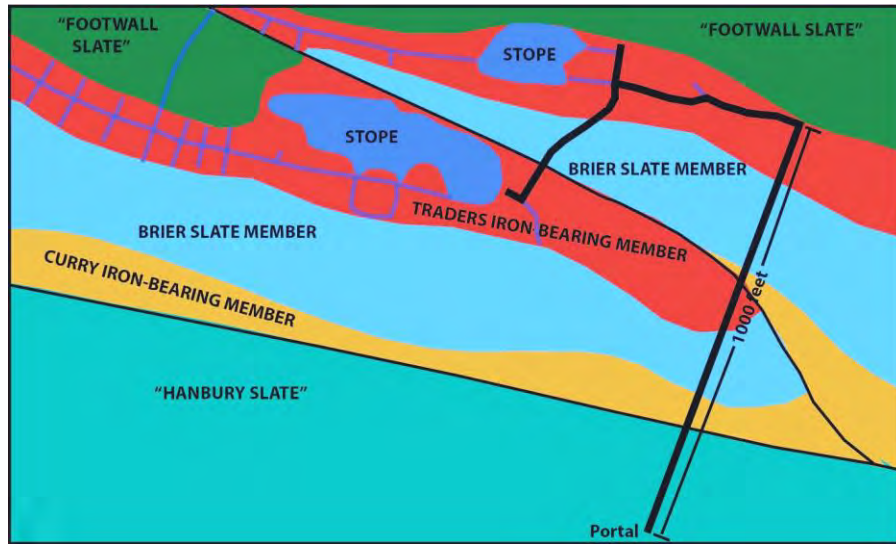


Figure 14. Geological interpretation of the Iron Mountain Iron Mine adit area at the ‘Tunnel Level’, which correlates to the 1st Level of the old Central Vulcan Mine. Based on a blueprint from the Penn Iron Company, circa 1900 (see Figure 15). Walking tour, shown in heavy black line begins at the portal.

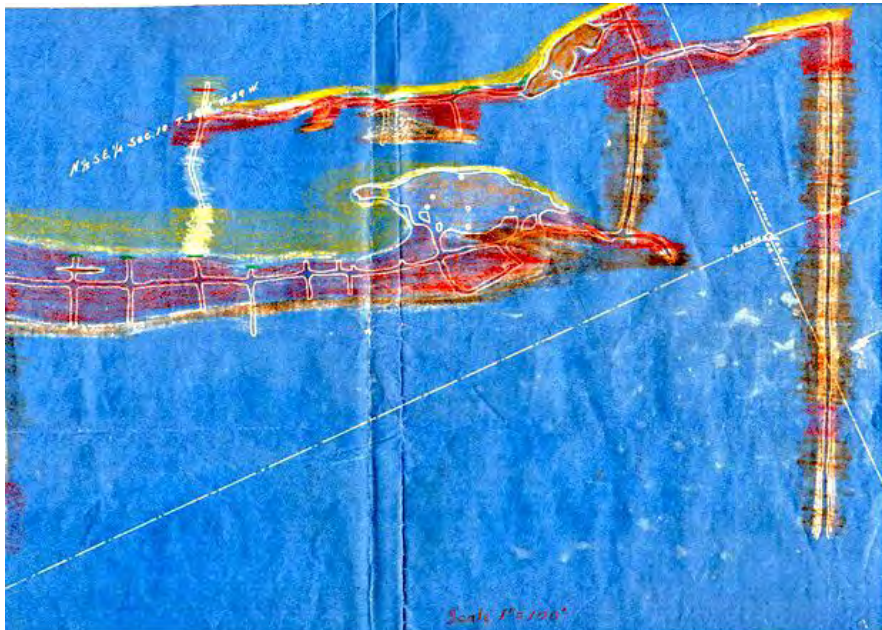


Figure 15. Blueprint of Penn Mining Company adit #2. The initial straight section of the adit trending N 20° E, is 1,000 feet long for a scale reference. Note the offset of the Traders Iron-Bearing Member on this image and Figure 14 and the location of the ore relative to fault geometry. When compared to other mines in the Range, this structure displays east dipping fault control where all mines to the west have west-dipping ore bodies.



Figure 16. Traders Iron-bearing Member, shattered martite-jasper displaying 30° dip to the south into the wall. A carbide light for scale.



Figure 17. Quartz druze on Jasper Iron Formation from the Central Vulcan Mine. Specimen was collected from a retaining wall on the north side of the west parking lot at the shaft. The wall contained several specimens of trace minerals that cement brecciated martite ore including calcite and pyrite. Only the quartz specimens survived the weathering since closure in 1946. The minerals occurred in “water courses”, in the ore bodies, usually in vertical channels as described by local miners.



Figure 18. The New Central Shaft, Vulcan, Michigan. View looking west. Old Central shaft, circa 1877, was due north 1,500 feet' (500m) and the underground workings connected at the 6th and 9th levels between the shafts.

Stop 4. Randville Dolomite. (45.806°N, 87.951°W)

The Randville Dolomite, the youngest formation of the Chocoy Group, is exposed extensively in the region. It is well described by Bayley et al. (1966, p. 35) from which the following description is excerpted.

“Massive clastic dolomite makes up a large part of the Randville Dolomite and is closely associated with thick- and thin-bedded sandy dolomite, dolomitic and quartzose slate and phyllite, and pebbly dolomite conglomerate. Thick beds of nearly pure crystalline dolomite are present in some areas and probably make up an important part of the formation. A most distinctive rock type in the formation shows algal structures (stromatolites). These are domical, 1-3 inches high, 3-12 inches in diameter, and composed of nested laminae of pure dolomite. The algal structures occur nearly every place in the district where the dolomite is exposed. They form reefs as much as 50 feet thick and of great but undetermined linear extent. They are also present in the Randville Dolomite of central Dickinson County (James et al., 1961) and in the Kona Dolomite of the Marquette district. As pointed out by James, stromatolite structures are also reported in nearly all dolomite of late Precambrian age-in the western United

States and Canada, Australia, South Africa, and Fennoscandia-and most geologists now accept the view that they represent fossil algal colonies. In the mapped area the algal dolomite is usually associated with thin-bedded sandy and conglomeratic dolomite of shallow-water deposition. This general association may be best observed in the outcrop area southeast of Lake Antoine, where algal dolomite, ripple-marked sandy dolomite, and thin dolomite beds showing mud cracks occur together.”

At Stop 4, an active gravel/stone operation, recent activities have exposed conglomeratic dolomite. The rock is a poorly sorted, thick-bedded, intraformational conglomerate composed almost entirely of clasts of dolomite as much as about 10 cm diameter. Clasts visible in hand specimen range down to sand-sized grains. All clasts are composed of very fine-grained gray to pinkish dolomite. The matrix is somewhat darker, coarser-grained dolomite. The rock appears to be an intraformational conglomerate and we have seen no exotic clasts that would indicate clastic input for a distant source. We also have not seen any clearly biogenic features here. The total thickness of this conglomeratic unit was not exposed in September, 2017, but it appears to be at least 10 meters thick. Figure 19 illustrates the typical lithology.

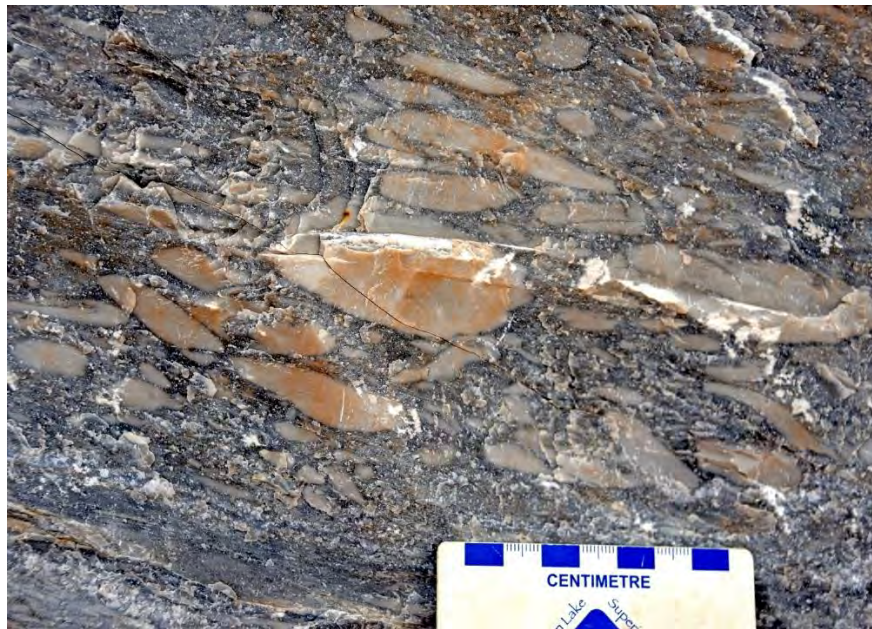


Figure 19. Conglomerate composed entirely of clasts of Randville Dolomite.

Stop 5. Michigamme Slate. (45.777° N, 87.889° W)

Brickyard Road, Norway Michigan. Exposures are on private property to which we have been granted access for this field trip. The bedrock ridge south of Hanbury Lake in sections 15 and 16, T. 39 N., R 29 W. contains the most extensive exposures of the Michigamme Slate on the Menominee Range. These rocks have been informally referred to as the Hanbury Slate in some early reports, but were renamed the Michigamme Slate by Bayley et al. (1966).

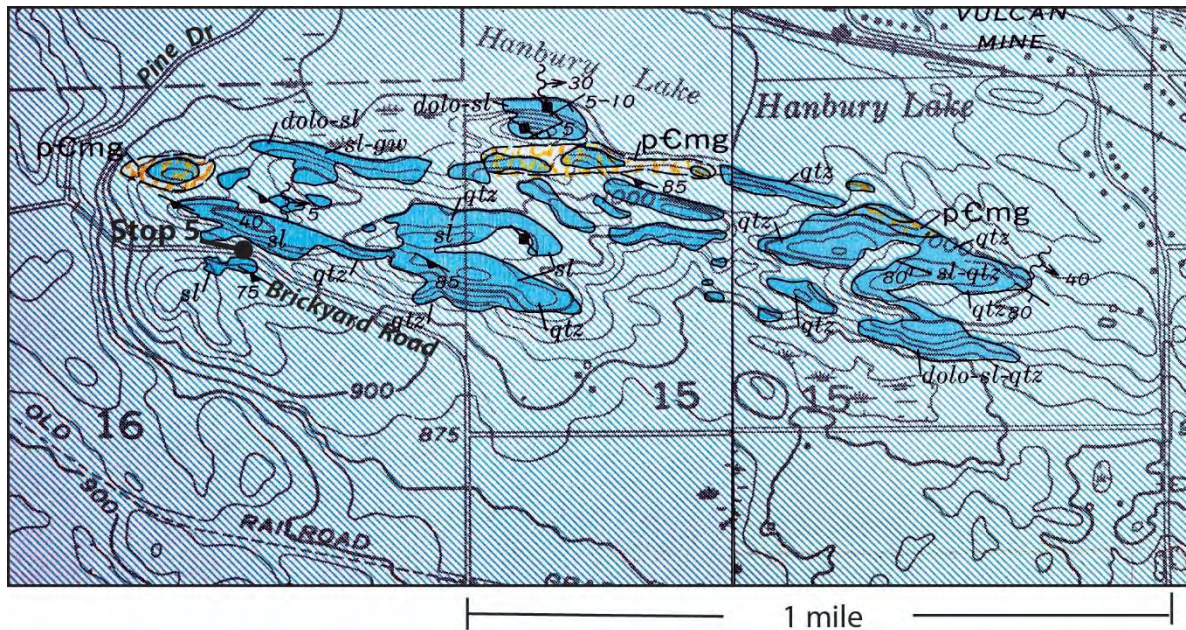


Figure 20. Geologic map of the Michigamme Slate in the Hanbury Lake area. The figure shows parts of plates 2 and 3 of USGS Professional Paper 513 (Bayley et al., 1966). The area is entirely underlain by Michigamme Slate except for a few bodies of intrusive metagabbro (pCmg). Areas of outcrop are shown by darker shade. Various lithologies of the Michigamme are indicated as sl-slate, qtz-quartzite, dolo-dolomite, gw-graywacke.

Stop 5 is near the west end of the ridge where lithologies change from highly sheared greywacke slate to dolomitic slate intruded by metagabbro. Folds indicate a duplication of strata that impacts true thickness estimates of the Michigamme Slate in the southern portion of the Range. Complex folding of the sediments is evident in outcrop along with a significant change in lithology which includes greywacke, quartzite, carbonate, and pyritic carbonaceous shale. Folds have vertical to steep southward-dipping axial planes as indicated by the prominent foliation, and plunge from 30-40° to the east. In USGS Monograph XLVI (Bayley, 1904), a detailed description of the area describes both large (meters) to small (centimeters) scale folding that exhibits strike and dips normal to the regional strike of the range (N 75° W). On the west and north ends the area the slate is cut by metadiabase dikes. These dikes are likely the same age as those identified in the mine workings at the Penn Mines Central Shaft, and at the Cyclops and Norway mines open pits on Norway Hill.

Descriptions and discussion of the area from USGS Professional Paper 513 (Bayley et al., 1966, p. 60) follows: *Dolomitic rock.- Dolomitic quartzite, dolomitic shale, and dolomite occur chiefly in the broad belt of outcrops south of Hanbury Lake. Dolomitic quartzite occurs south of Hanbury Lake only, where it is associated with dolomitic slate and dolomite and with intrusive metagabbro. The quartzite beds appear to be confined to the eastern three-fourths of the group of outcrops south of the lake, probably because the quartzite beds lens out to the west or are doubled back in a fold. Numerous minor folds in the slate show small areas where the beds strike north across the overall northwest foliation, and folding is thus indicated as the more likely cause of the limited distribution of the quartzite.*

The dolomitic quartzite is dark grey or, if encrusted with limonite, brown. The beds are 1-10 feet thick and commonly show quartz-filled cross-fractures that do not enter the adjacent slate beds. A distinctive characteristic of the quartzite is the presence of chips of black slate as much as 6 inches long in most beds. The rock is made up of about equal parts of well-rounded and well-sorted quartz grains and dolomite, and trace amounts of carbonaceous dust. The quartz grains all show undulatory extinction when viewed under the microscope, a feature probably inherited from the source rocks inasmuch as the quartzite does not appear to be deformed internally.

The only exposed dolomite in the formation is confined to a belt of outcrops trending northwest from south of Hanbury Lake. The northwestern most rocks on the belt are dolomitic slates which outcrop in secs. 4 and 5, T. 39 N., R. 30 W. South of Hanbury Lake the dolomite is light colored, banded, and somewhat slaty: it occupies the north part of the group of outcrops. The best exposed rock is at the lakeshore. The beds are folded, and in the northern-most outcrop the general strike is nearly normal to the trend of the outcrop belt; The dips are low to the southeast, because these beds are on the crest or in the trough of a minor fold. The lesser plications and folds on the beds plunge at low angles, less than 30 degrees SE. West of Hanbury Lake, in the south parts of secs. 7 and 8, T. 39 N., R. 29 W., are outcrops of siliceous and dolomitic grey slate and rather thick-bedded siliceous gray dolomite. A rind of limonite that coats the exposed surfaces of the dolomite indicates that the carbonate is probably ferruginous, an observation previously made by W. S. Bayley, who reported the chemical analysis shown in Table 29. On the assumption that all the iron, magnesia, and lime form carbonates, W. S. Bayley gives the composition of the carbonate as about 9 percent FeCO_3 , 41 percent MgCO_3 , and 50 percent CaCO_3 .

Stop 6. Niagara fault splay at Piers Gorge. (45.759°N, 87.942°W) (text reproduced from 2003 ILSG guidebook. Laberge et al., 2003)

Rocks exposed along the Menominee River at Piers Gorge are almost certainly a branch of the Niagara fault zone and represent one of the few exposures of the fault zone. This location is about one kilometer south of the mapped trace of the Niagara fault. The hill lying north of the gorge, but still south of the mapped fault, is underlain by metagabbro that is much less deformed than the rocks in the gorge. These relationships indicate that strain along the fault zone was distributed very heterogeneously and concentrated in discrete zones of very high strain surrounding islands of weakly deformed rocks. The rocks in the gorge are highly foliated and lineated quartz-sericite schists and chloritic schists, probably developed from felsic and mafic volcanic rocks. Felsic and mafic volcanic rocks with only weak foliation, along with mafic sills with little internal deformation, are exposed on both sides of this strongly foliated zone. Metagraywacke of the Marquette Range Supergroup is exposed in Norway, about 2 miles north of this locality, and volcanic and plutonic rocks of the Wisconsin magmatic terranes are exposed along the Menominee River in this area. The foliation here strikes N 80-85° W and dips 80-85° N. and has a stretch lineation that plunges 60-65°, N 85° W.

As the recognized boundary between the dominantly sedimentary rocks of the Marquette Range Supergroup to the north and the Wisconsin magmatic terranes to the south, the Niagara fault zone is commonly referred to as a suture. However, it lacks some features

(such as a *mélange*) that are typical of suture zones. Geophysical evidence (Attoh and Klasner, 1989; and LaBerge and Klasner, 2001) suggests that thinned continental crust of the Superior craton has been overridden by the Wisconsin magmatic terranes, and extends in the subsurface for 10-50 miles south of the Niagara fault zone. If this is the case, the Niagara fault zone may be the frontal thrust on which oceanic rocks of the Wisconsin magmatic terranes overrode the continent margin assemblage of the Marquette Range Supergroup. Continued compression of the suture zone resulted in the steepening of the thrust surfaces into their present, nearly vertical orientation.



Figure 21. Schist in Niagara fault zone at Piers Gorge. A- Highly foliated schist along north bank of the Menominee River. B- Nearly vertical foliation with modern slump toward river producing a spurious shallower foliation. Both photos from 2003 ILSG field trip.

Stop 7. Quinnesec Mine (45.810°N, 87.991°W)

The abandoned workings of the Quinnesec mine (known locally as the Devil's Icebox) are mainly in the Traders Iron-bearing Member of the Vulcan Iron-formation. The property is fenced and accessible by arrangement with the property owner. The mine lies on the overturned north limb of a second-order syncline (Figure 22). The Precambrian strata at the mine dip about 60° north, but face southward, inasmuch as the Brier Slate Member of the Vulcan is along the south side of the excavated approach to the mine, and the Felch Formation is along the north wall of the workings. Cross-sections through the Vivian Mine immediately to the west of Stop 7 show the geometry of the westward plunging folds and the overturned structure we see at the mine exposure. The ore is specular hematite and jasper and the brecciated subcrop was considered an ore where it has been reworked in a shoreline environment during the Cambrian Period.

The mine workings provide an exceptional view of the unconformity and basal Cambrian sandstone overlying the mine workings along the north side of the hill (Figure 24 A, B). The basal portion of the sandstone contains numerous angular slabs of oxidized iron-formation, iron ore, and slate in a sandy matrix (Figure 24 C). Clearly, this area was a small island as the Cambrian sea advanced over the area. Cross sections (such as Figure 23) also show the steep local relief that existed on the Precambrian erosional surface. The complex folding and duplication of beds made for a more resistant area of iron-formation that likely led to the development of a topographic high. The clasts of iron ore in the basal conglomerate also indicate that the ore here was formed before the Cambrian sea covered the area.

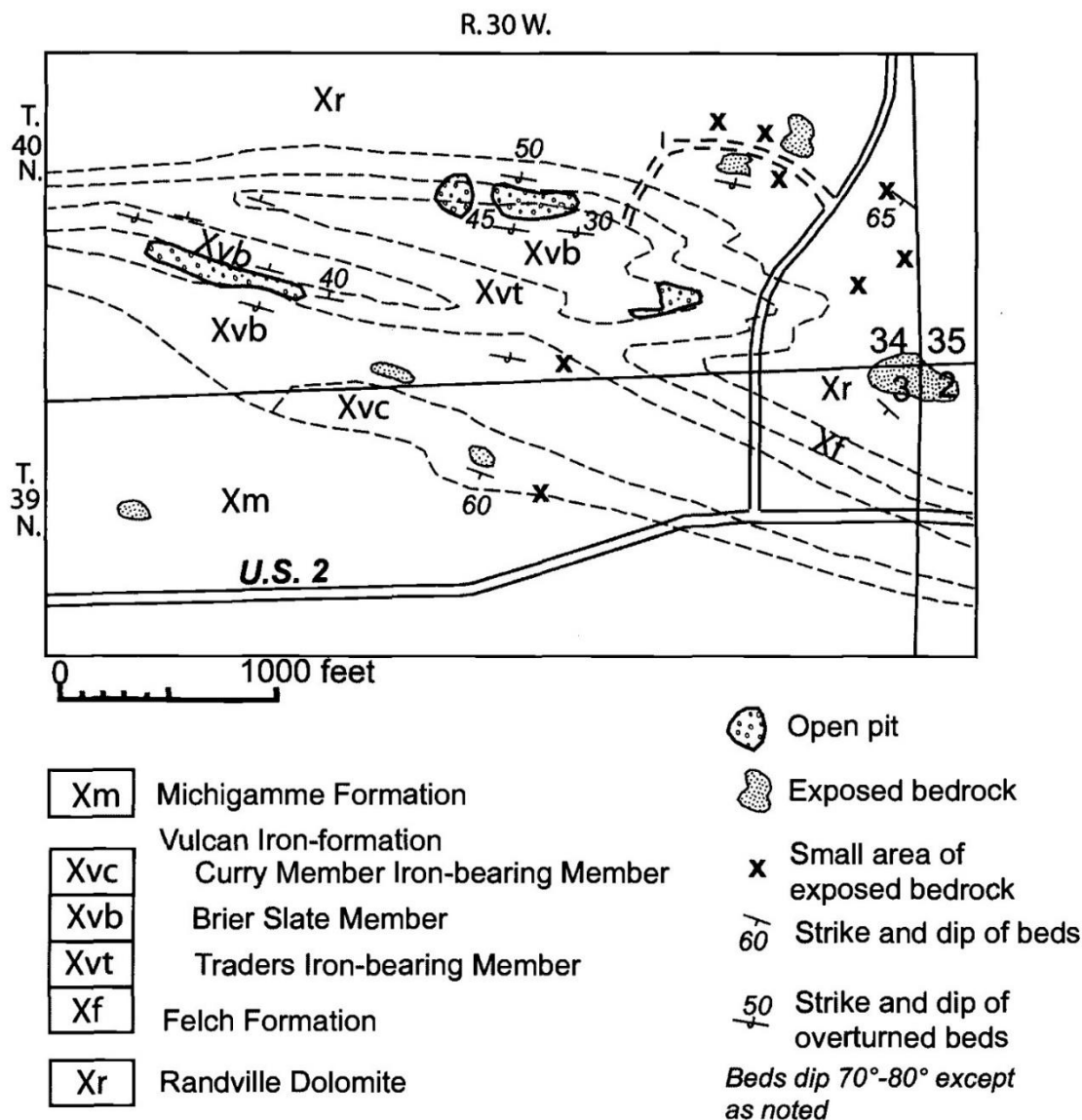


Figure 22. Geologic map of the Quinnesec Mine and vicinity showing that the workings were developed in the overturned northern limb of a small syncline. From LaBerge et al. (2003, based on mapping by Bayley et al. (1966)).

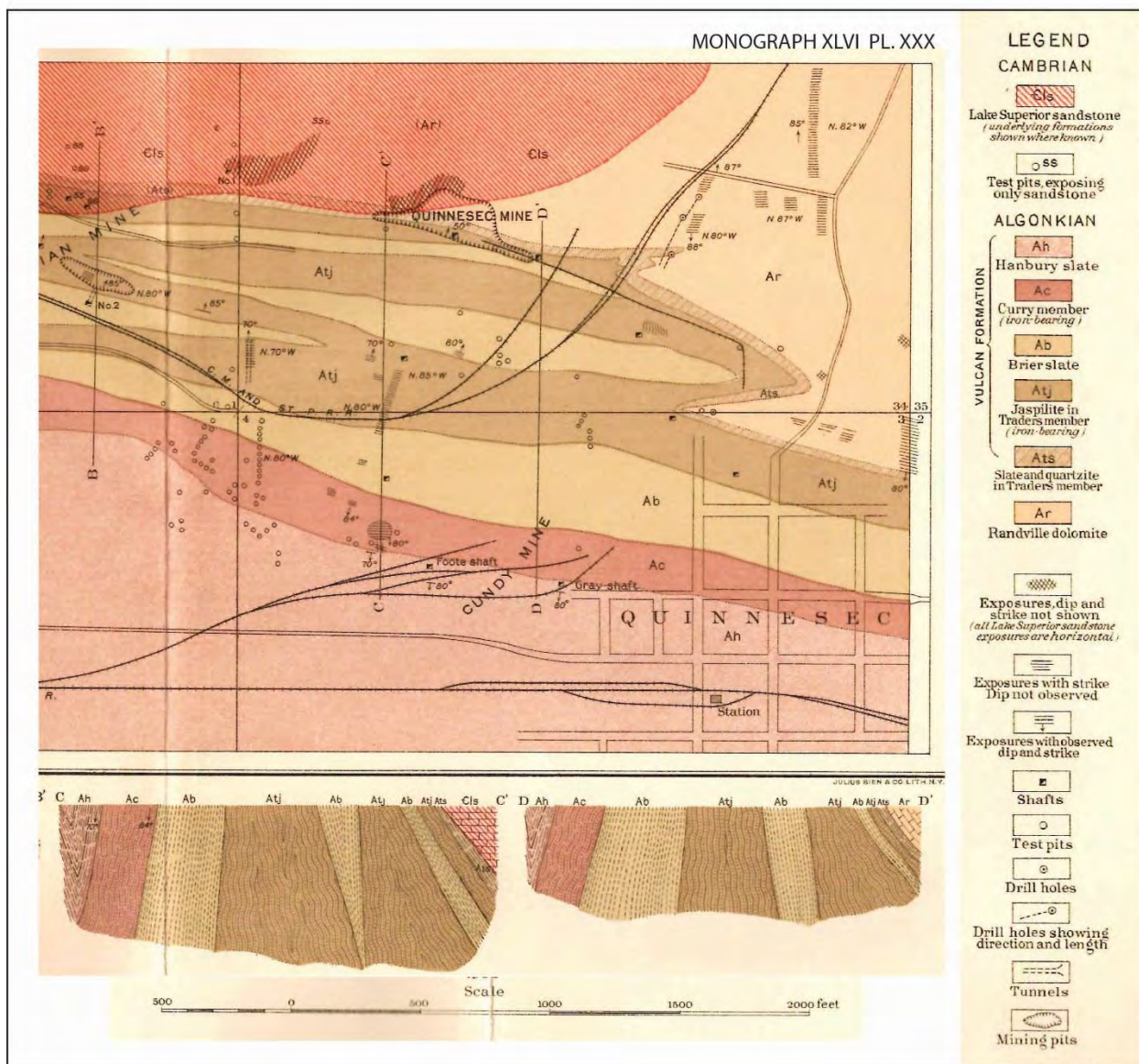


Figure 23. A portion of plate XXX from U.S. Geological Survey Monograph XLVI (Bayley, 1904).

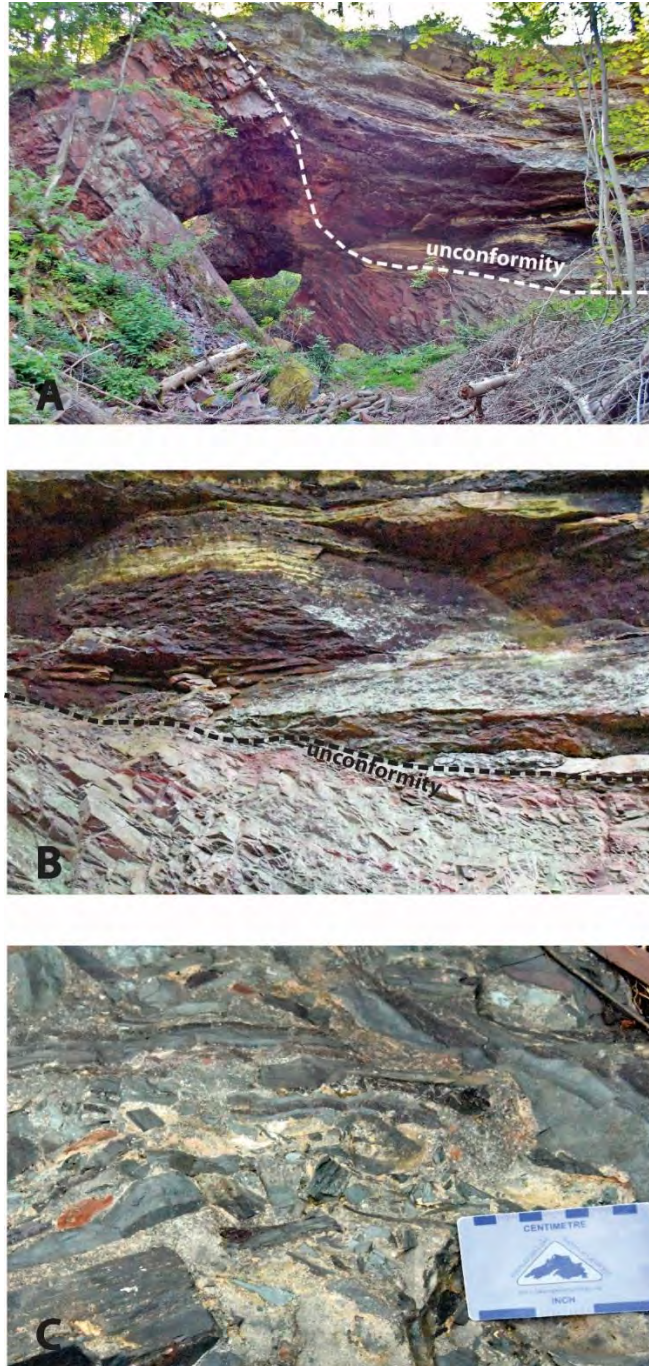


Figure 24. Photographs of the Quinnesec Mine workings. A- View looking west into the mine workings. The Traders Iron-bearing Member of the Vulcan Iron-formation dips steeply north and faces south. The unconformity with the overlying Munising Sandstone (Late Cambrian) is well exposed and shows steep topography that existed on the Precambrian units during Cambrian marine transgression. Photo by Thomas Waggoner. B- Close-up view of the unconformity. C- Basal breccia of the Munising Sandstone, probably talus deposited at the base of the paleoescarpment formed by the Vulcan. Large clasts are entirely iron-formation, many of which show secondary iron enrichment. Lighter matrix is quartzose sand.

Stop 8. Keel Ridge area (45.810°N, 88.028°W)

The area of the previous Keel Ridge mine is now operated for crushed stone as well as being excavated for future business development. The Keel Ridge mine was one of the earliest mines opened on the Menominee Range in 1880, but only produced until 1899 with a total production of 93,101 tons. The mine was located just to the northwest of the large stripped area we will examine at this locality. The stratigraphic section exposed is from the Randville Dolomite on the north to the Michigamme Slate (“Hanbury Slate”) on the south and includes an excellent cross section of various members of the Vulcan Iron-formation. Although the area is easily accessible from U.S. 2 and the various units of the Vulcan Iron-formation can be observed and sampled, it is private property and should not be entered without permission of the owner.



Figure 25. The Keel Ridge mine area. Geologic units are as indicated in USGS Professional Paper 513, Plate 1 (Bayley et al., 1966). Excavation reveals the northwest-striking formations that include the upper part of the Felch Formation (“Traders Slate”), Traders Iron-bearing Member, Brier Slate, and Michigamme Slate (“Hanbury Slate”). The formations face to the south and dip to the north at 80°- 90°.



Figure 26. View looking east on east side of exposure showing the upper part of the Felch Formation (“Traders Slate”) and conformable contact with the Traders Iron-bearing Member of the Vulcan Iron-formation. Dips are vertical to overturned toward the south.

The northernmost exposures are of the Randville Dolomite, which here consists of a breccia of angular siliceous material (chert?) in a siliceous carbonate matrix. This may be a residual accumulation of chert nodules and beds formed by solution of the carbonates of the Randville on the long-lived erosion surface, now expressed as the unconformity between the Randville and the Felch Formation.

The next outcrop to the south is the Felch Formation (“Traders Slate”) which is a rare exposure of this unit in the entire range. Bayley (1904) referred to the distinctive sericitic slate or quartzite found at the top of the Felch Formation as either the “Traders Quartzite” or the “Traders Slate” depending on the predominant lithology. The formation is described in USGS Professional Paper 513 (Bayley et al., 1966, p. 38) as:

“Lithology of the Felch Formation is remarkably uniform throughout the length of the south iron range, but variable along the north range. On the south range the formation is about 100 feet thick and consists of thin-bedded sericite slate and phyllite, and intercalated thin-bedded quartzite. The quartzite layers appear to be prevalent in the upper part of the formation, and a thin (4 in. to 3 ft.) key bed of dark ferruginous quartzite, the so-called “Traders quartzite,” is commonly present near the top of the formation. The fine-grained clastic rocks which make up the major part of the formation on the south iron range include slate, phyllite, siltstone, and schist. All these rocks show minor differences imposed during deposition and modifications imposed by later deformation and low-rank metamorphism, but they bear a close outward resemblance to one another and show a common mineralogy.

They are predominantly thin-bedded rocks, the layers commonly less than 1 cm thick, and most show bedding-plane fissility. Cleavage surfaces of phyllite are lustrous and spangled with tiny plates of white mica. On fresh surfaces the rocks are gray to greenish gray, but where weathered they may be pale green, red, or light buff, or almost white and mottled with red; color banding is not conspicuous. The chief mineral components of all fine-grained types are pale green sericite and quartz; minor components are feldspar, chlorite, biotite, hematite, and magnetite. Most of the rock layers are composed of about equal parts of the two chief components, but layers composed predominately of one or the other are common. Medium to coarse well-rounded grains of quartz and potassic feldspar, commonly visible to the unaided eye, are scattered throughout many specimens of the slaty rock and form wafer-thin discontinuous quartzite stringers between the slaty layers. These latter characteristics of the slate are useful but not infallible criteria for identifying Felch strata in the field. The prevailing texture of the rocks is micro-schistose. In some specimens the quartz grains as well as the sericitic groundmass are elongated in the plane of schistosity, which at most places parallels the bedding.”

Several tens of feet are exposed at this location including the sharp conformable contact with the overlying Traders Iron-bearing Member.

The two members of the Vulcan Iron-formation that were most significant economically are the Traders and Curry Iron-bearing Members. They were described in USGS Professional Paper 513 (Bayley et al., 1966, p. 43) as follows:

“The rocks of the Traders and Curry members are iron formation, which has been defined by James (1954, p. 239) as “a chemical sediment, typically in bedded or laminated, containing 15% or more of a layer of sedimentary origin, commonly but not necessarily containing layers of chert.”

The iron-formation of the Vulcan is thin bedded in commonly laminated, but it does not display uniformity in the thickness of the beds. Individual beds generally range from 1 mm to 30 cm in thickness. As a rule, beds of granular jasper alternate with beds composed chiefly of oxides of iron, principally hematite Fe_2O_3 (69.94% iron), and a lesser amount of magnetite Fe_3O_4 (72.4% iron). Almost all of the iron-rich layers contain a small amount of crystalline quartz, and at some places dolomitic carbonate and chlorite as well.

The iron-formation usually is dark. Viewed from a distance it commonly appears dark gray or reddish-brown, but at close range it appears as a medley of deep red or maroon, metallic gray, and black. If much oxidized, hues of orange and red are dominant. Most jasper beds are maroon (liver colored) or red. They are generally thicker than adjacent iron rich beds and most are uniformly straight bedded, but irregular beds and lenticular beds are common.

The jasper beds are composed chiefly of red jasper granules, specular hematite, magnetite, and metachert (a fine-grained mosaic of crystalline quartz). The granular character of most jasper beds can be seen by the unaided eye, but a wetted surface and a hand lens are helpful. In their primary state the jasper granules are a mixture of amorphous silica and red iron oxide. In their primary state the jasper granules are a mixture of amorphous silica and red iron oxide (fig. 13). In their characteristic crystallized state, the iron oxide is specular hematite, magnetite, or both, and the silica is crystalline quartz (fig 14). Most jasper beds contain, in addition to jasper granules, oololiths which are made up of concentric layers of red amorphous hematite and silica

about a nucleus of quartz or jasper. Re-crystallized oololiths form the same products as the granules. The granules, in shape, size, and appearance, resemble the greenalite granules, that are so characteristic of the Biwabik Iron-Formation of the Mesabi Range. They may represent the analog of the greenalite granules, formed under oxidizing conditions."

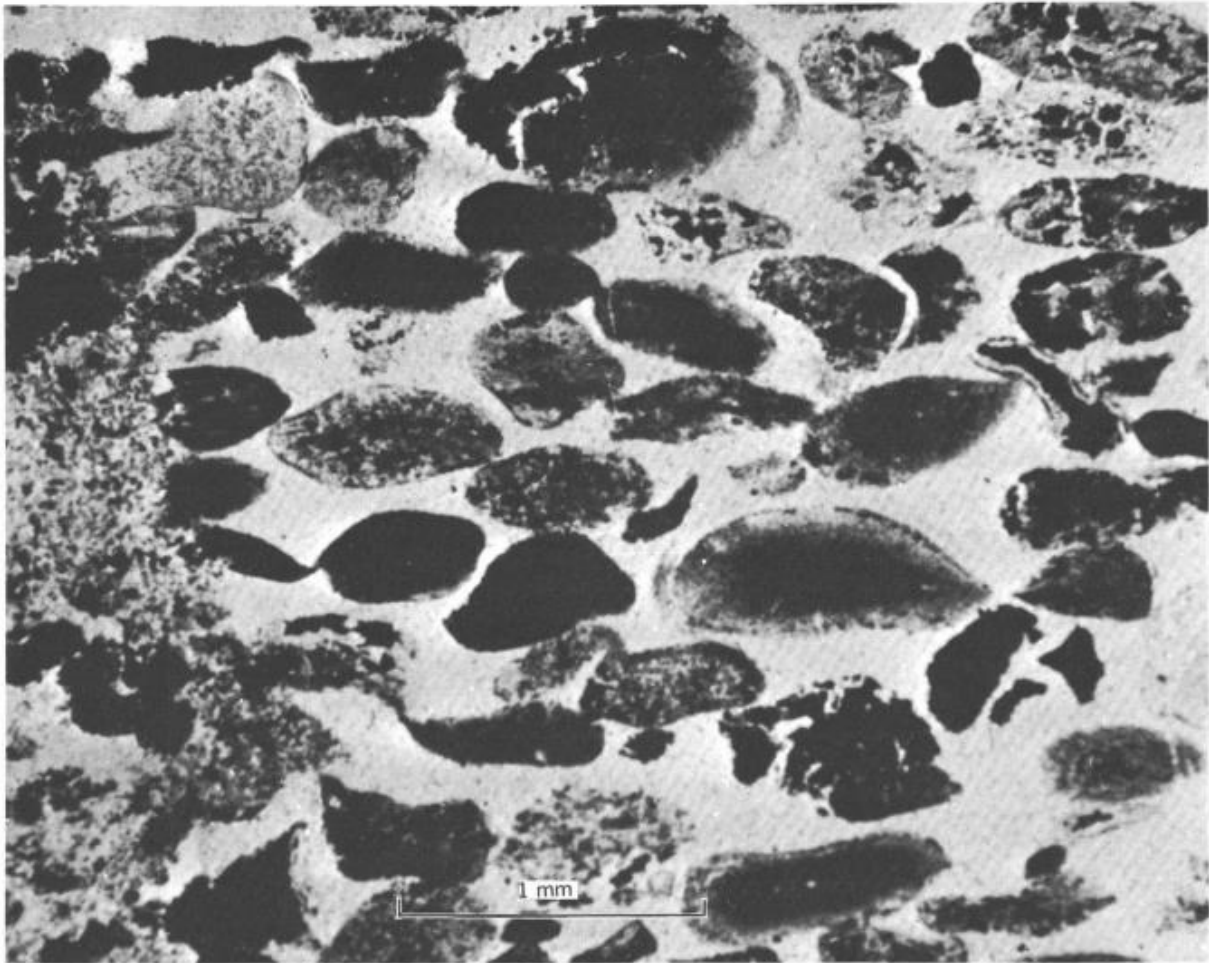


Figure 27. Figure 13 from Bayley et al., 1966, p. 44. *"Photomicrograph showing jasper granules in a metachert matrix. The very dense granules are bright-red noncrystalline jasper. The gray (salt-and-pepper) granules show an early stage of crystallization-segregation of iron oxide as hematite or magnetite from silica. Ordinary light."*

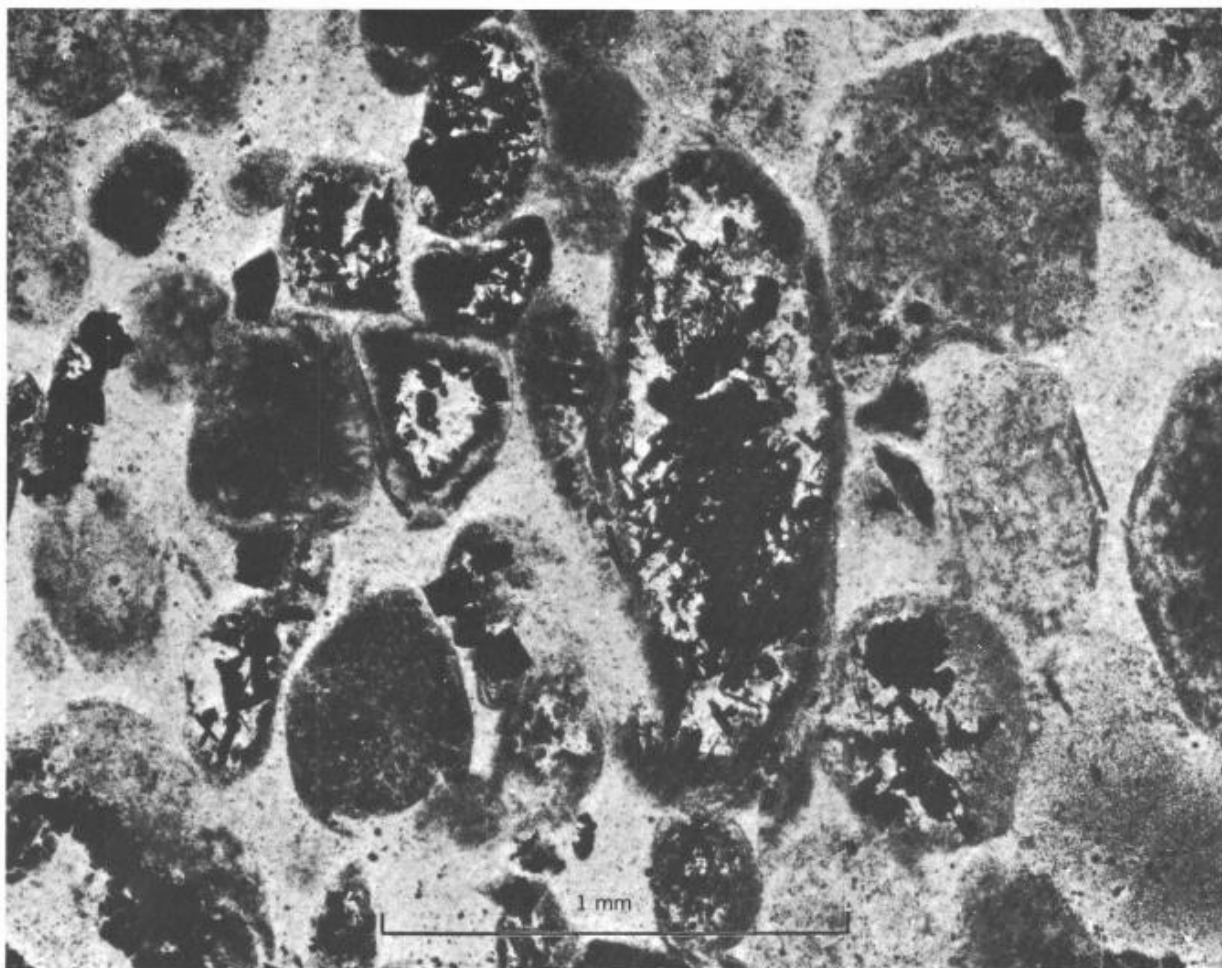


Figure 28. Figure 14 from Bayley et al., 1966, p. 45. *“Photomicrograph showing several types of crystallization of jasper granules. Quartz, white; specular hematite and magnetite, gray or black. Note the wide variation in the iron to silica ratios from one granule to another. The black spicules are specular hematite, and the square to rectangular sections are magnetite. Most granules pictured contain both iron minerals. The mottled granule (top center) represents an early stage of crystallization and segregation. The shapes of the iron oxide segregations suggest incipient specular hematite. Plane-polarized light.”*

The Brier Slate is in fault contact with the Traders Iron-bearing Member and is oxidized at this location. The Curry Iron-bearing Member is missing either because of faulting or non-deposition. This interpretation is based on the formations exposed in the underground workings of the Keel Ridge mine. The Michigamme Slate (“Hanbury Slate”) is in fault contact with the Brier Slate and is best exposed on the west side of the excavation. The outcrop shows significant shearing and oxidation of the slate with probable duplication of section to the south

Presented next are detailed descriptions of the Traders Iron-bearing Member, the Curry Iron-bearing Member, and the Brier Slate by Oliver Mining Company geologists from crosscuts in the mines east of this locality. Comments in their archived records state that the direct shipping ores (+50% Fe) from the various mines could be identified by their physical and mineralogical characteristics.

Traders Iron-bearing Member. A detailed description of the Traders Iron-bearing Member from the Penn Mining Co. to the east shows a measured section of 112', less than the average 132' measured in sections in the south range by the USGS. An abrupt conformable contact with the underlying "Traders Quartzite" occurs in the mine.

Basal 3 feet: Ferruginous slate or slaty iron-formation consisting of a micaceous, medium-grained, even-bedded unit, with a high silica content slate. Iron content, 26-32%. Earthy hematite occurs with specular hematite scattered throughout.

Next 12 feet: Wavy-bedded jasper iron-formation with fine-grained red (liver colored) chert bands. Beds consist of thin even slaty specular hematite laminae (not readily cleavable), and long narrow (1/4" – 4"), lenses of fine grained to saccharoidal red (liver colored) chert. Grades gradually to a granular chert phase (next layer), by change of color and finer grain size. Iron content 33-40%.

Next 18 feet: Massive wavy bedded iron formation with reddish brown granular chert lenses. A fairly massively, bedded jasper. Thin laminae of slaty specular hematite (not readily cleavable) occur with heavy lenses of reddish-brown saccharoidal to granular chert. Top grades into overlying even-bedded iron-formation by decrease in the amount of granular chert. Bottom grades into red chert phase by a gradual change of color and becomes finer grained while the specular hematite background remains the same. Iron content is 34-40%.

Next 64 feet: Even-bedded iron-formation with dark brown, fine-grained chert. (and occasional granular chert lenses). Even-bedded cherty iron-formation composed of thin (1/64" to 1/4") laminae of slaty specularite with narrow (1/4") lenticular bands of dark-brown, fine-grained chert. Near the base occur heavier bands of the dark brown, fine-grained chert up to 2 inches in thickness which disappear toward the top and bedding there becomes uniformly thin and quite even. One section shows one of these heavy fine-grained chert bands. Another section shows a typical thin, even-bedding of the higher phase. Occasional heavy lenses of dark, reddish-brown, granular chert occur frequently near the base where this unit grades into the granular chert below but they become rarer in the upper portion. Iron content is 34-43%.

Next 15 feet: Massive dark granular chert. Massive, irregularly bedded, lean, granular chert, dark-brownish-purple in carbide light. Contains very little slaty ferruginous matter and the chert is all granular with many red jasper granules. Specularite occurs in thin irregular veinlets through the body of the chert. Iron content 25-33%.

Curry Iron-bearing Member. A measured section from this locality is 158 feet in fault contact with Brier Slate.

Basal 14 feet: Slaty basal phase: Even-bedded blocky, dark-brown, siliceous, slaty ferruginous rock, containing very little free chert. Laminations are 1/8 inch- to 1/2 inch-thick and consist of slaty brown, hematite, becoming bluer with increase of specularite toward the top. What free chert exists is purplish granular Curry-type. This horizon seems favored for ore concentration. Iron content is 32-33%.

Next 144 feet: Cherty phase: It is a heavy bedded, straight-bedded blocky specular cherty iron-formation with groups of thin, even, rich specular laminae alternating with

irregular lenses (1/4 inch to 4 inches thick) of dark reddish-purple granular chert, shot through with the regular veinlets and mottles of specularite. The chert is invariably granular and dark reddish purple in color. Iron content is 32-43%.

Brier Slate. The Brier Slate separates the Traders and Curry Iron-bearing Members in this mine cross-section. The slate is 104 feet thick and displays contacts that are faulted with both iron-formations. The Brier Slate is a soft, fine to medium grained, thinly laminated, blocky, ferruginous slate. The color is very dark with a white streak where it is unoxidized, but it is generally oxidized to a chocolate brown color with a dull red streak. Bedding laminae are thin but prominent, especially near the bottom, and grain size varies by laminae, with the coarser grains in the heavier laminae. A coarse phase occurs near the middle. Concentration may produce higher iron analysis locally, but never an ore body. Iron content averages 23%, varying between 15 to 30%.

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FIELD TRIP 4

Friday May 18, 2018

GRANITOID ROCKS OF THE PEMBINE-WAUSAU TERRANE IN NORTHEASTERN WISCONSIN

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With a contribution from Marcia Bjornerud, Lawrence University

INTRODUCTION

This trip examines granitoid rocks of the Pembine-Wausau terrane that are exposed in northeastern Wisconsin. The Pembine-Wausau terrane is one of the Paleoproterozoic magmatic arcs that comprise the internal domain of the Penokean orogen (Figure 1; Schulz and Cannon, 2007). The terrane consists of mafic to felsic volcanic rocks ranging from tholeiitic to calc-alkaline in composition, subordinate sedimentary rocks, and granitoid intrusive rocks of largely calc-alkaline affinity. It was accreted to the southern margin of the Archean Superior craton beginning about 1,875 Ma along a paleosuture now marked by a major ductile deformation zone, the Niagara fault zone (Sims et al., 1985).

The rocks in northeastern Wisconsin have been key to understanding the stratigraphic and tectonic evolution of the Pembine-Wausau terrane and the nature of the Penokean orogeny. Rocks in the area are fairly well exposed, especially compared to other areas in northern Wisconsin. Granitoid rocks constitute nearly half of the outcropping rocks of the area and are mainly granodiorite and tonalite, but include gabbro, diorite, and granite (Sims and Schulz, 1993). An older suite, ranging in age from about 1,890–1,870 Ma, is dominantly calcic to calc-alkaline and appears to be cogenetic with the volcanic arc magmatism, while younger, 1,860–1,840 Ma, calc-alkaline to alkaline plutons are broadly contemporaneous with collision of the Pembine-Wausau terrane with the Superior craton margin (Sims et al., 1992). Younger post-tectonic intrusions, emplaced at about 1,835 and 1,760 Ma, consist of alkali-feldspar granite suites (Sims et al., 1993).

In the area of the field trip (Figure 2) most of the exposed granitoid rocks are part of the Dunbar dome, an irregular, asymmetrical structure ~470 km² in area, composed of gneiss, migmatite, amphibolite, and foliated to unfoliated granitoid rocks mantled by steeply dipping sedimentary and volcanic rocks of Paleoproterozoic age (Sims et al., 1992). The Dunbar dome is one of several roughly correlative domes in northern Wisconsin (Morey et al., 1982) which have less well-exposed gneiss and granitoid rocks with comparable isotopic ages (Sims and Peterman, 1980; Sims et al., 1989) and chemical compositions (Sims et al., 1993).

In the Dunbar dome, compositionally varied gneisses, assigned by Cain (1964) to the Dunbar Gneiss, are intruded by five major plutons named the Marinette Quartz Diorite, Newingham

Tonalite, Hoskin Lake Granite, Spikehorn Creek Granite, and Bush Lake Granite (Sims et al., 1992). This field trip will examine exposures of each of the major units of the Dunbar dome except the Bush Lake Granite. Detailed descriptions of these major units including their petrography, structure, geochemistry, age, and petrogenesis are given in Sims et al., 1984; 1985; and 1992 (the 1984 reference is available for free download from the ILSG website <http://www.lakesuperiorgeology.org/>; the 1992 reference is available for free download from the USGS website <https://www.usgs.gov/products/publications/official-usgs-publications>). In addition to the granitoid units of the Dunbar dome, the field trip will examine exposures of the Twelve Foot Falls Quartz Diorite, a subvolcanic intrusion that is comagmatic with calc-alkaline volcanic rocks in the Quinnesec Formation (Sims et al., 1992) as well as the post-tectonic (~1,835 Ma) Athelstane Quartz Monzonite and Yavapai-age (~1,750 Ma) Amberg Granite which occur south of the Dunbar dome (Sims, 1990; Sims et al., 1993).

Sims et al. (1992) concluded that the granitoid rocks within and outside the Dunbar dome were derived from different sources based on their contrasting chemistry, and presumably were developed in different tectonic environments, and were subsequently superposed tectonically. They attributed the Newingham Tonalite and Twelve Foot Falls Quartz Diorite to the subduction processes that formed the volcanic arc represented by the Quinnesec volcanic rocks. In contrast, the intrusions within the dome, which range from syn- to post-tectonic, were attributed to melting of continental lithosphere during collision of the arc with the continental margin of the Superior craton. However, these conclusions need to be reevaluated in light of acquired Nd isotope data (Van Wyck and Johnson, 1997; Schulz and Ayuso, 1998) and new understanding of the processes that produce granitoid rocks in orogenic belts (Hildebrand and Whalen, 2017).

Although the granitoid rocks of the Dunbar dome range in composition from calcic tonalite to calc-alkaline granodiorite to alkali-calcic quartz diorite, they have a surprisingly small range of enriched ϵ_{Nd} values centered on 0 and depleted mantle model ages of ~2.0 to 2.2 Ga. The Dunbar Gneiss has the most negative $\epsilon_{\text{Nd}}(1,860)$ values of -2.1 to -3.4 and the Hoskin Lake Granite the most positive $\epsilon_{\text{Nd}}(1,835)$ value of +1.71; the Marinette Quartz Diorite and Newingham Tonalite have similar $\epsilon_{\text{Nd}}(1,860)$ values near 0 (+0.12 and +0.39, respectively). Only the Twelve Foot Falls Quartz Diorite, which is a subvolcanic intrusion, has a strongly positive $\epsilon_{\text{Nd}}(1,900)$ value of +4.54 indicating derivation from a long-term light rare earth element (REE) depleted source. The narrow range of enriched ϵ_{Nd} values for the Dunbar dome granitoids is unlikely to be the result of crustal contamination as it would be highly fortuitous for granitoids of such varying chemistry to all have similar degrees of crustal contamination. Instead, the narrow range in ϵ_{Nd} values is more likely a characteristic of the source from which the granitoids were derived. In addition, the data indicate that the Newingham Tonalite is likely not a syn-volcanic intrusion, as suggested by Sims et al. (1992), but rather is a syn-collisional intrusion and part of the Dunbar dome suite.

A characteristic feature of the Dunbar dome granitoids is that they are relatively enriched in Ba, K, Nb, Rb, Sr, Ta, and Th, and have steep, light REE-enriched patterns (Sims et al., 1992). Recently, Hildebrand and Whalen (2017) examined the geochemistry of a number of major

Cordilleran-type batholiths including the Sierra Nevada, Peninsular Range, Idaho-Montana, and Cascades-Coast Plutonic Complex of North America among others. They noted a clear compositional distinction between plutons generated as syn-volcanic intrusions during subduction and those emplaced as syn- to post-tectonic intrusions during slab failure (breakoff). In particular, they showed that magmas generated during slab failure have relatively high Nb/Y, Sr/Y, and Sm/Yb ratios (Figure 3) as well as evolved radiogenic isotopes. These characteristics are shown by intrusive rocks ranging from gabbro to granite and calcic to alkaline in composition (Hildebrand and Whalen, 2017). They concluded that the distinctive whole-rock geochemistry, as well as radiogenic and stable isotope compositions, of slab failure magmas involve only minor amounts of crustal contamination and are derived mainly from plagioclase-absent melting of garnet-bearing rocks in the mantle (for example, garnet pyroxenite, eclogite, and/or subcrustal lithosphere). As seen in Figure 3, the Dunbar dome granitoids, including the Newingham Tonalite, plot in the fields defined by Hildebrand and Whalen (2017) for slab failure magmas.

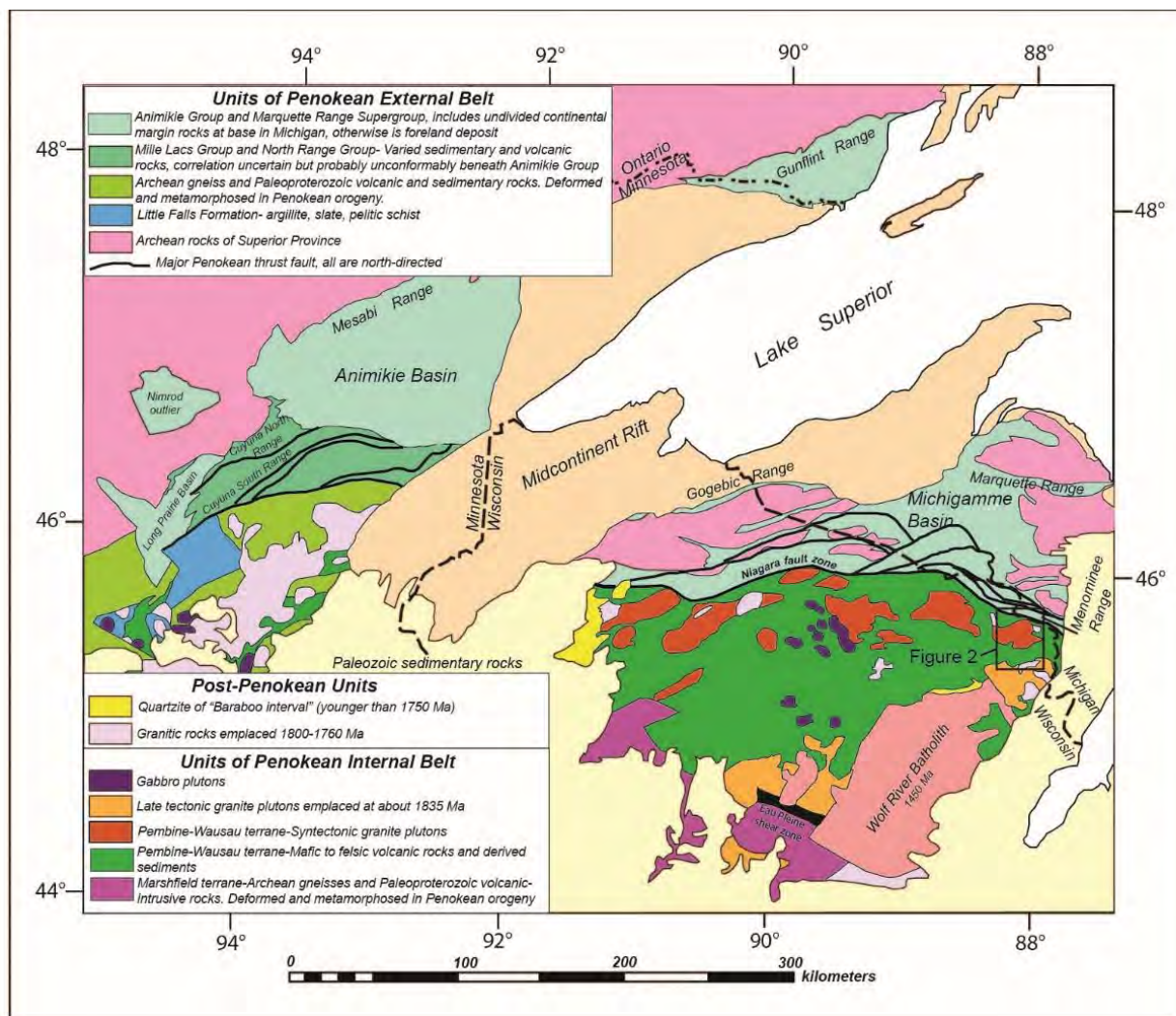


Figure 1. Generalized geologic map of the Penokean orogen in the Lake Superior region showing approximate location of Figure 2 (from Schulz and Cannon, 2007).

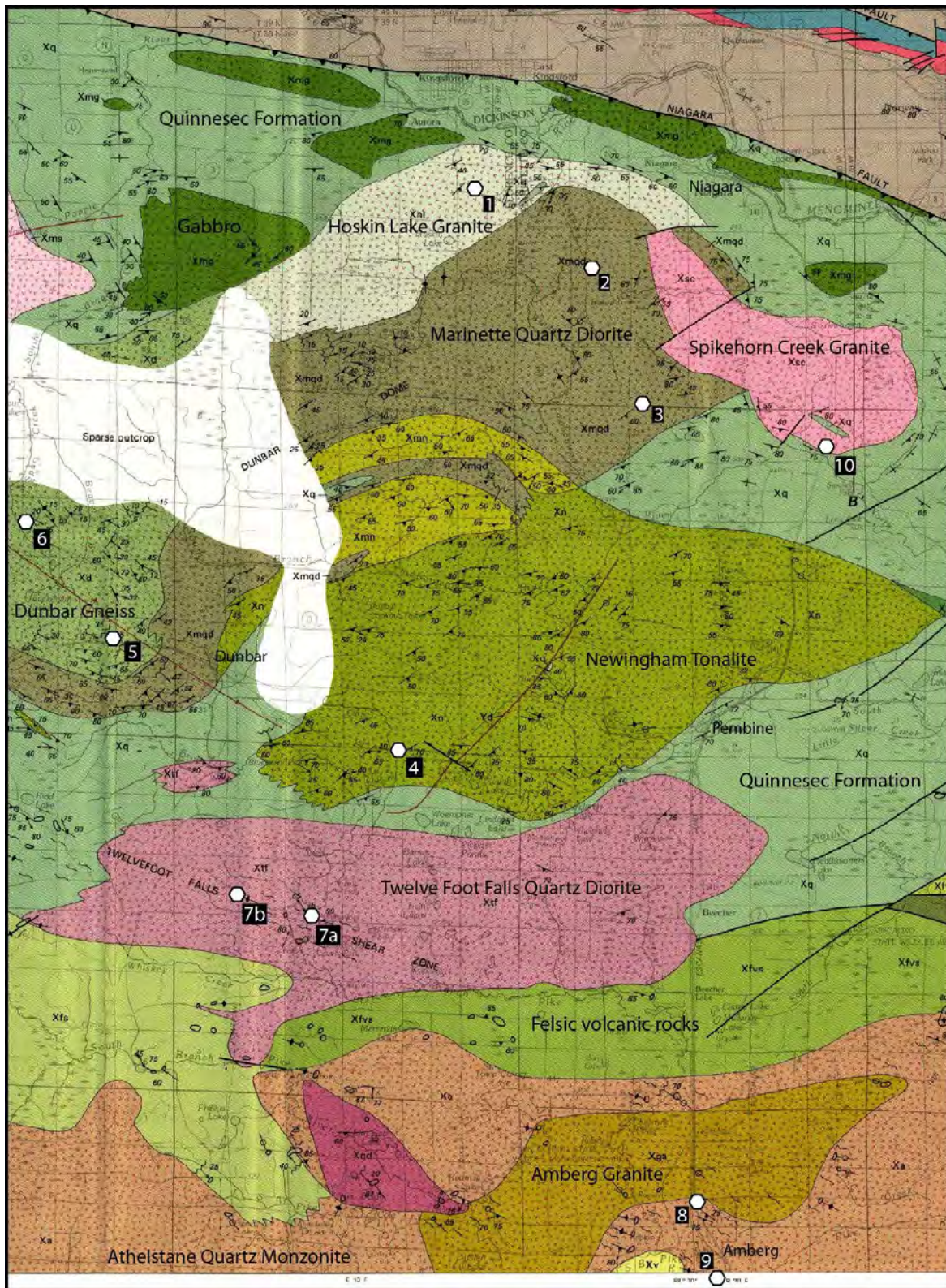


Figure 2. Geologic map for a portion of northeast Wisconsin showing the locations of the field trip stops (from Sims and Schulz, 1993).

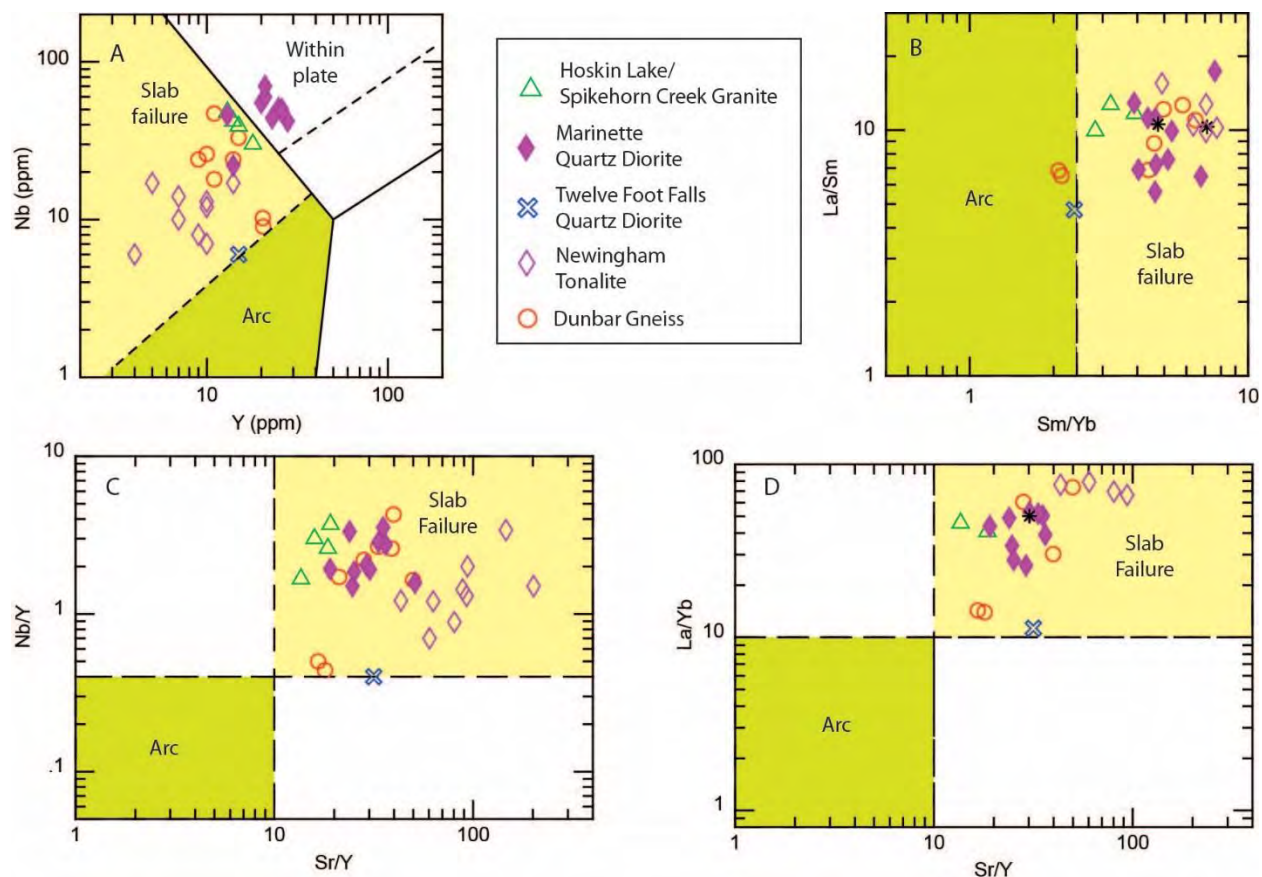


Figure 3. Plots of Nb vs Y (A), La/Sm vs Sm/Yb (B), Nb/Y vs Sr/Y (C), and La/Yb vs Sr/Y (D) for samples from the Dunbar dome and Twelve Foot Falls Quartz Diorite. Fields after Hildebrand and Whalen (2017).

FIELD TRIP STOPS

Stop 1. Hoskin Lake Granite (Outcrop on north side of County road N; 45.764° N., 88.071° W.; Note that the outcrop is on private property and permission from the owner is required)

The Hoskin Lake Granite is an arcuate, convex-northward body of granite on the north margin of the Dunbar dome characterized by (1) pink to gray, medium- to coarse-grained inequigranular granite with large, oriented, tabular potassium feldspar crystals (Figure 4), (2) abundant inclusions of mafic-intermediate volcanic rocks of the Quinnesec Formation, and (3) late, euhedral crystals of potassium feldspar that lie athwart to an older foliation and, at least locally, transect centimeter-thin quartz veins. As noted by Cain (1964), the southern margin of the granite appears to be gradational into rocks assigned to the Marinette Quartz Diorite and evidence for K-metasomatism along the border of the two units is compelling. To the east, the Hoskin Lake Granite appears to grade into the post-tectonic Spikehorn Creek Granite. Although different in appearance, the two granites have similar compositions (Sims et al., 1992). Excellent descriptions of the Hoskin Lake Granite are given in Bayley et al. (1966) and Sims et al. (1992).

An Nd isotope analysis of one sample of the Hoskin Lake Granite gave a $\epsilon_{Nd}(1,835) = +1.71$

with a depleted mantle model age of 1.99 Ga (Schulz and Ayuso, 1998), suggesting derivation from a slightly light REE depleted source.



Figure 4. Hoskin Lake Granite.

Stop 2. Marinette Quartz Diorite (Railroad cut on County road O; 45.747° N., 88.033° W.) **Note:** Access to this railroad grade is strictly prohibited without prior approval of the owner.

The railroad cut shows metamorphosed Marinette Quartz Diorite cut by dikes of Hoskin Lake Granite and leucogranite. The Marinette Quartz Diorite is a large, layered sill-like intrusive body that was emplaced in the contact zone between the Dunbar Gneiss and the Quinnesec Formation and the intrusive Newingham Tonalite. It is dominantly composed of quartz diorite and diorite with moderately high biotite (~10 to 20%), hornblende (trace to 30%), and sphene (trace to 6%) contents. In the north-central part of the Dunbar dome, the rocks contain variable amounts of potassium feldspar and are interpreted as hybrid rocks reflecting post-crystallization K-metasomatism (Sims et al., 1992). In the eastern part of the body, which is relatively unmetamorphosed, the Marinette Quartz Diorite is a dark gray to black, medium-grained, hypidiomorphic granular rock with well- to ill-defined layering and generally lacks a penetrative foliation; rare clinopyroxene occurs as relicts in the cores of some hornblende crystals. South of Dunbar, the Marinette Quartz Diorite is medium to dark gray, mesocratic, layered quartz diorite and diorite cut by abundant granite pegmatite and aplite dikes (Figure 5). Based on the presence of mineralogical layering and geochemistry, the Marinette Quartz Diorite is interpreted as dominantly cumulate rocks derived from an alkaline mafic to intermediate magma with within plate-syn-collisional compositional characteristics (Sims et al., 1992). Uranium-lead zircon dating of the Marinette Quartz Diorite gives an age of $1,862 \pm 15$ Ma, which overlaps the age of the Dunbar Gneiss that it intrudes (Sims et al., 1992). Neodymium isotope analyses of the Marinette Quartz Diorite show $\epsilon_{\text{Nd}}(1,860) = +0.7$ to $+0.12$ with a depleted mantle model age of ~2.1 Ga (Barovich et al., 1989; Schulz and Ayuso, 1998), suggesting derivation from an enriched source.



Figure 5. Marinette Quartz Diorite south of Dunbar cut by granite pegmatite dikes.

Stop 3. Marinette Quartz Diorite (Brown Spur road going east from County road O; 45.713° N., 88.011° W.)

Although there appears to be no actual outcrop of Marinette Quartz Diorite at this stop, there are numerous angular boulders of black, medium- to coarse-grained diorite to quartz diorite typical of the eastern, less metamorphosed part of the intrusion.

Stop 4. Newingham Tonalite (At the intersection of Highway 8 and 1 Mile Road, north side of road; 45.628° N., 88.096° W.)

The Newingham Tonalite forms a large body, ~75 km² in area, that intrudes the volcanic rocks of the Quinnesec Formation along the southeast margin of the Dunbar dome (Figure 2). The contact zone is at least 100 m wide, and consists of interlayered tonalite bodies and generally angular amphibolite (metabasalt) inclusions. Near the contact with the Marinette Quartz Diorite, the Newingham Tonalite has been largely converted to granodiorite or granite by post-crystallization addition of potassium feldspar forming a hybrid, megacrystic facies of the Newingham Tonalite (Sims et al., 1992). The Newingham Tonalite is mostly a uniform light gray, medium-grained, slightly porphyritic rock (Figure 6) that generally has a good secondary foliation except in the eastern portion north of Pembine. It is locally cut by dikes of slightly porphyritic tonalite and, occasionally, granite pegmatite. The Newingham Tonalite has the compositional characteristics of high Al₂O₃-type tonalite-trondhjemite suites including high Al₂O₃ (>15 wt.%) and Sr (>600

ppm), low K and Rb, and steep REE patterns depleted in heavy REE ($\text{La/Yb} = 60\text{--}90$) (Sims et al., 1992). Uranium-lead zircon dating of the Newingham Tonalite gives an imprecise age of $1,861 \pm 40$ Ma (Sims et al., 1992). Neodymium isotope analysis of one sample of the Newingham Tonalite gave a $\epsilon\text{Nd}(1,860 \text{ Ma}) = +0.39$ with a depleted mantle model age of 2.09 Ga (Schulz and Ayuso, 1998), suggesting derivation from an enriched source or contamination by older crustal rocks.



Figure 6. Newingham Tonalite.

Stop 5. Dunbar Gneiss (Intersection of Highway 8 with County road U, west side of road; 45.655° N. , 88.199° W.)

Exposed here is a low outcrop of mainly megacrystic granite gneiss that contains rafts of amphibolite and is intruded by granite pegmatite. The gneiss has a pervasive steeply dipping N. 50° W. foliation. The Dunbar Gneiss generally consists predominantly of gray biotite gneiss, which is layered at scales ranging from a few centimeters to several meters reflecting differences in the amount and kind of major minerals as well as differences in grain size (Figure 7). Granite pegmatite and aplite intrude the Dunbar Gneiss particularly in the western part of the dome and can compose more than 50 percent of the outcrop. As described by Sims et al. (1992, p. 7). . . .”*The biotite gneisses are mylonitic rocks. They have a dominant xenomorphic granular (granoblastic) texture and a penetrative foliation expressed by oriented biotite and, less commonly, by elongate and flattened aggregates of quartz and plagioclase that are generally subparallel to compositional layering....The biotite gneisses have a moderate range in composition from layer to layer; their average and modal composition is granodiorite. Plagioclase (calcic oligoclase-andesine) and quartz are the principal minerals, potassium feldspar varies from 0 to about 30 percent, and biotite generally composes from 10 to 20 percent*

of the rocks. Myrmekite and myrmekitic plagioclase are abundant....Sphene (titanite) is the principal accessory mineral and comprises as much as 2 percent of the rock."

The Dunbar Gneiss is calc-alkaline in composition with intermediate SiO_2 contents (~62 to 72 wt.%), moderately high Al_2O_3 (~14 to 17.5 wt.%), and high K_2O (2.4 to 5.1 wt.%). It also is enriched in Ba, Nb, Rb, light REE, Ta, and Th, and has steep REE patterns with depleted heavy REE (Sims et al., 1992). A sample of Dunbar Gneiss just north of this outcrop gave a U-Pb zircon concordia upper intercept age of $1,862 \pm 5$ Ma (Sims et al., 1992). Neodymium isotope analyses by Barovich et al. (1989), Schulz and Ayuso (1998), and Van Wyck and Johnson (1997) gave very similar results with $\epsilon\text{Nd}(1,860) = -2.1$ to -3.4 and depleted mantle model ages of 2.20 to 2.41 Ga. The isotope data suggest the protolith of the Dunbar gneiss was derived from an enriched source.



Figure 7. Dunbar Gneiss and granite pegmatite.

Stop 6. Dunbar Gneiss (Spur Lake Road going west from County road U; 45.684° N., 88.232° W.)

The exposures on the east side of the road consist of compositionally layered biotite gneiss and lesser amphibolite intruded by megacrystic biotite gneiss, granite pegmatite, and aplite. All rocks are deformed and have a vertical N. 50 - 55° W. foliation.

Stop 7. Twelve Foot Falls Quartz Diorite and mylonite, ultramylonite, and pseudotachylyte along the Twelve Foot Falls shear zone

The Twelve Foot Falls Quartz Diorite is an elongate, east-west trending body that intrudes and locally contains inclusions of metavolcanic rocks of the Quinnesec Formation south of the Dunbar dome (Figure 2). The quartz diorite is generally massive in the eastern part, but becomes foliated towards the west. At the type locality at Twelve Foot Falls on the north branch of the Pike River, the quartz diorite has been intensely sheared by the Twelve Foot Falls shear zone and is mainly a mylonitic gneiss. As described by Sims et al. (1992, p. 43).*...."Outside the shear zone, the quartz diorite is medium to coarse grained and is characterized by subhedral plagioclase (sodic andesine) crystals as much as 1 cm long, smaller subhedral hornblende*

crystals, in part pseudomorphic after pyroxene, and anhedral crystals of blue quartz as much as 1 cm in diameter. Microcline locally occurs as a late interstitial mineral. The primary texture is hypidiomorphic granular, but finer grained secondary textures are superposed on it at many places. Characteristically, the rock is considerably retrograded: plagioclase is partly to largely altered to epidote and albite, and hornblende is partly altered to biotite, epidote, and chlorite. Other alteration minerals are sphene, opaque oxides, and calcite.”

The quartz diorite has a calc-alkaline andesite ($\text{SiO}_2 = 57 \text{ wt.}\%$) composition with low TiO_2 (0.44 wt.%) and high field strength element contents (Sims et al., 1992). It is similar in composition to andesite volcanic rocks in the Quinnebec Formation and is interpreted to be a subvolcanic intrusion. A sample dated by Schulz and Schneider (2005) gave a U-Pb zircon concordia upper intercept age of $1,889 \pm 6 \text{ Ma}$ age. This age shows that the volcanic rocks of the Quinnebec Formation are significantly older than the intrusive rocks of the Dunbar dome and places a minimum age on the Pembine ophiolite present within the Quinnebec Formation (Schulz and Schneider, 2005). An Nd isotope analysis of the quartz diorite gave a $\epsilon\text{Nd}(1,900) = +4.54$ with a depleted mantle model age of 1.87 Ga (Schulz and Ayuso, 1998) indicating a depleted source and no crustal contamination. The strongly positive ϵNd value for the quartz diorite is similar to that determined for the mafic volcanic rocks of the Quinnebec Formation (Beck and Murthy, 1991).

(Material below and Stops 7a and 7b contributed by Marcia Bjornerud, Lawrence University)

The Twelve Foot Falls shear zone (Sims, 1990) can be traced for at least 20 km along strike in northwest Marinette County, Wisconsin, from Twelve Foot Falls County Park on the north branch of the Pike River to just south of Kidd Lake. The timing of displacement on the shear zone is only broadly constrained; the shear zone transects the Twelve Foot Falls Quartz Diorite ($1,889 \pm 6 \text{ Ma}$) as well as the metavolcanic Quinnebec Formation, and it lies immediately south of the $1,862 \pm 5 \text{ Ma}$ Dunbar Gneiss (Sims et al., 1992). The vertical to steeply northeast-dipping foliation and mylonite bands in the Twelve Foot Falls Quartz Diorite are broadly parallel to the foliation in the southern part of the Dunbar dome, but the Twelve Foot Falls shear zone does not cut through the dome itself. Sims (1990) suggested that the northern part of the Amberg Granite (U-Pb zircon age $1,752 \pm 8 \text{ Ma}$) also is transected by the shear zone; if so, the zone would have developed during or after the Yavapai orogenic cycle. The sense of slip also is poorly constrained; a weak down-dip lineation points to dip-slip motion, but it is not clear whether the slip sense was reverse – which would suggest activity synchronous and sympathetic with convergence on the Niagara Fault – or normal, which would indicate slip related to late-orogenic relaxation.

Stop 7a: Twelve Foot Falls County Park (45.579° N. , 88.137° W.)

Note that Marinette County parks require an entrance fee; use self-service registration box in main parking area.

The main falls are visible from the parking area across a small pool in the Pike River. Follow the narrow foot path north of the picnic area to reach the outcrop adjacent to the falls, where the Twelve Foot Falls Quartz Diorite is well-exposed. The rock there is strongly foliated and locally mylonitized, and both the foliation and mylonitic fabric are defined by bands of quartz and feldspar alternating with aligned hornblende crystals (partly regressed to chlorite), indicating that

the overall schistosity and localized zones of high strain formed at peak metamorphic (amphibolite facies) conditions.

Depending on water levels, there is another area of exposed rock about 220 m downstream from Twelve Foot Falls, just above Eight Foot Falls. There, dark, branching discordant veins 0.3-0.5 cm wide and 10-15 cm long cut across the foliation in the host rock (Figure 8). In thin section, the veins are found to contain a mesh of fine retrograded hornblende (?) crystals with high aspect ratio, arranged with no preferred orientation in a non-crystalline matrix that is dark in plane polarized light. These macro- and micro-scale characteristics suggest that the veins represent devitrified pseudotachylyte injection veins – frictional melt generated on a fault plane during seismic slip and injected as ‘hydro’-fractures into the surrounding rock (Nadziejka and Bjørnerud, 2014; Larson and Bjørnerud, 2017). Significantly, the pseudotachylyte material can be seen in both outcrop and thin section to have been cut by, and in places incorporated into, the mylonitic bands. This indicates that brittle seismic failure occurred at least once while the rocks were still at depths and temperatures where crystal plastic deformation was predominant. We have also found small amounts of pseudotachylyte at Eighteen Foot Falls, about 1 km upstream from Twelve Foot Falls, and at Dave’s Falls near Amberg.

Thin sections of specimens from Eight Foot Falls also show mutually cross-cutting relationships between plastically deformed quartz veins and pseudotachylyte (Figure 9). This indicates that brittle tensile fracture and fluid flow occurred in alternation with seismic failure and ductile deformation. Some of the quartz veins contain significant amounts of pyrite. In addition, the foliation is in places transected by discontinuous cm-long veins in which hornblende and quartz occur as fibrous crystals perpendicular to the walls. The crystals have growth bands and fluid inclusion planes oriented parallel to the vein walls. These features suggest that the veins formed by the ‘crack-seal’ mechanism, in which cyclic fluid pressure variations cause hydrofracturing and incremental mineral growth. Crack-seal veins are most commonly found in the shallow upper crust; the fact that hornblende is one of the vein-filling minerals indicates that in this case, the process occurred at greater depth and higher temperatures.

In combination, these observations provide an exceptional glimpse into the complex interplay of deformation mechanisms and fluid flow in the middle crust during an orogenic event. Large earthquake ruptures apparently penetrated downward into rocks that were otherwise at temperatures high enough for full crystal plasticity. Such mutually cross-cutting relationships between mylonites and pseudotachylytes have been reported from only a small number of sites around the world (Sibson and Toy, 2006). Strain incompatibilities related to these ruptures may have caused dilatancy and large fluid pressure gradients that led to the formation of quartz-pyrite and quartz-hornblende veins.



Figure 8. Outcrop photos of pseudotachylyte and mylonite in the Twelve Foot Falls Quartz Diorite at Eight Foot Falls. Brunton compass and pencil indicate scale. White arrows show places where pseudotachylyte has been offset along the mylonitic foliation. Although the apparent offset is left lateral, lack of three-dimensional exposure makes true slip vector difficult to determine.

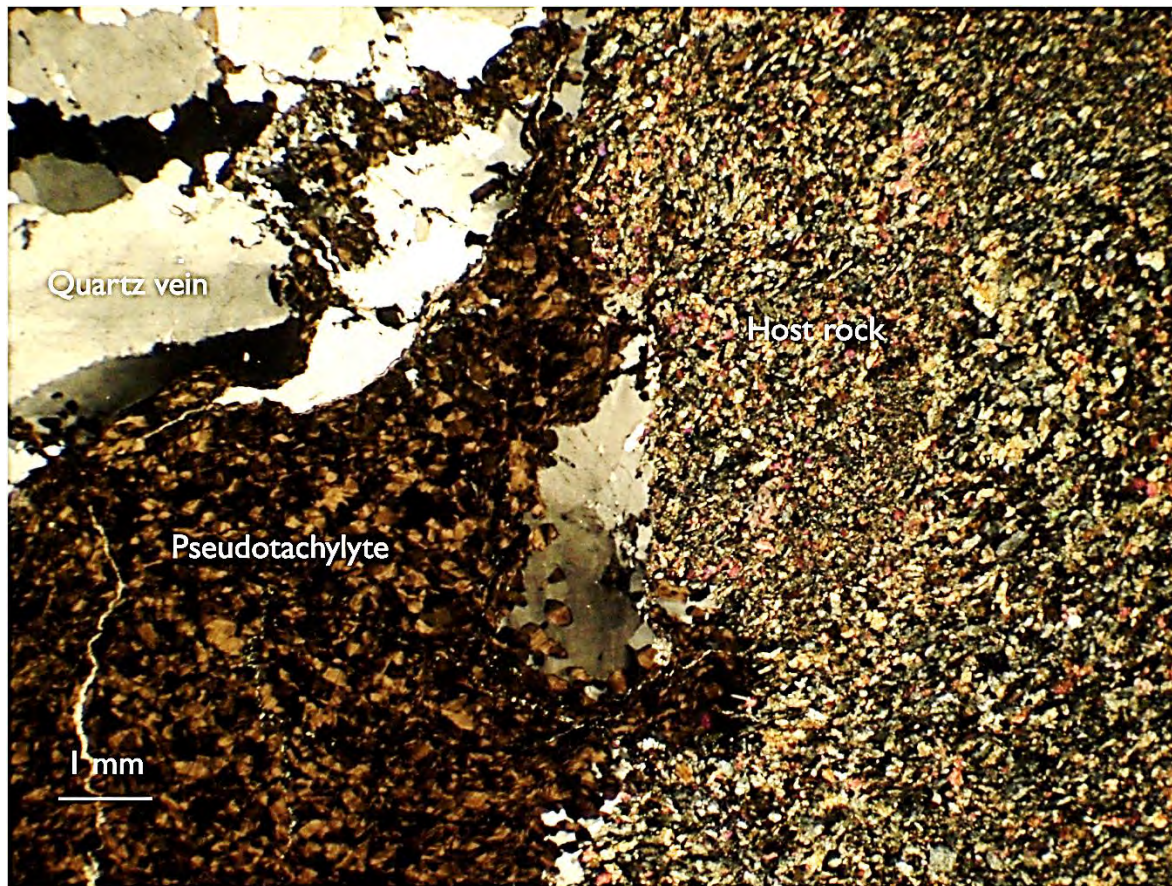


Figure 9. Photomicrograph (cross-polarized light) of finely recrystallized quartz diorite host rock cut by a quartz vein that was in turn intruded by pseudotachylite (altered entirely to clinochlore). Scalloped edge of quartz grain in middle of image is consistent with melting. Vein quartz shows undulatory extinction, indicating that plastic deformation followed or alternated with brittle fracture.

Stop 7b: Powerline exposure of Twelve Foot Falls Shear Zone (From the intersection of Twelve Foot Falls Road and Forest Road 513, drive 1.3 km (0.8 miles) west on 513 to an open area where a major powerline crosses the road; 45.584° N., 88.156° W.)

This site provides a glimpse of the strain heterogeneity typical of the Twelve Foot Falls shear zone. The textural character of the Twelve Foot Falls Quartz Diorite ranges from igneous to ultramylonitic, in some cases over distances of centimeters. The foliation and mylonitic bands dip about 80° NE. In places, a weak down-dip mineral lineation is discernible. In thin section, feldspar porphyroclasts show extremely long tails – suggesting very high shear strains – but no consistent asymmetry that would allow the sense of shear to be determined unambiguously.

Stop 8. Athelstane Quartz Monzonite and Amberg Granite (U.S. Highway 141 and Black Sam Road just north of Amberg; 45.517° N., 87.996° W.)

This pavement outcrop consists of Athelstane Quartz Monzonite cut by dikes of Amberg Granite (Figure 10; Medaris et al., 1973). The Athelstane Quartz Monzonite intrudes felsic volcanic rocks (Beecher Formation) north of this stop and extends for considerable distance both to the west and south (Sims, 1990). It is a pink medium- to coarse-grained granite to granodiorite with

allotriomorphic granular texture and 5 to 10 percent biotite and(or) hornblende (Sims et.al., 1993). Typically, the quartz monzonite has a clotty appearance due to the interstitial nature of the mafic minerals (Figure 10). Small metavolcanic inclusions are present locally. The Athelstane Quartz Monzonite is mildly peraluminous and has high SiO_2 (68–77 wt.%), intermediate Al_2O_3 (12–15 wt.%), K_2O greater than Na_2O , and enrichment in iron ($\text{FeOt}/(\text{FeOt} + \text{MgO}) = \sim 0.9$) (Sims et al., 1993). Samples plot in within-plate and syn-collisional fields on trace element tectonic discriminant diagrams (Sims et al., 1993). A U-Pb zircon age of $1,835 \pm 15$ My was determined on a sample from a large quarry south of this stop (Sims, 1990). This age overlaps with that determined for the Spikehorn Creek Granite on the northeast side of the Dunbar dome (Sims et al., 1992). Barovich et al. (1989) determined a $\epsilon\text{Nd}(1,835) = +1.1$ with a depleted mantle model age of 2.07 Ga on a sample of the quartz monzonite. The positive ϵNd value is similar to that determined for the Hoskin Lake Granite in the Dunbar dome (see Stop 1) and suggests derivation from a similarly light REE depleted source.

The Amberg Granite, seen here as dikes cutting the Athelstane Quartz Monzonite, is gray, fine- to medium-grained, with a hypidiomorphic granular texture and has biotite as the major ferromagnesian phase. It also occurs in at least three intrusive bodies within the Athelstane Quartz Monzonite (Sims, 1990). It has a U-Pb zircon age of $1,752 \pm 8$ Ma (Sims, 1990); it is one of a number of small plutons of this age found across northern Wisconsin (Sims et.al., 1993). These $\sim 1,750$ Ma plutons are coeval with the anorogenic granite-rhyolite terrane in south-central Wisconsin (Anderson et al., 1980; Smith, 1983). A sample of the Amberg Granite has an $\epsilon\text{Nd}(1,750 \text{ Ma}) = -0.91$ and a depleted mantle model age of 2.17 Ga (Schulz and Ayuso, 1998), suggesting derivation from an enriched source.



Figure 10. Athelstane Quartz Monzonite cut by dikes of Amberg Granite.

Stop 9. Athelstane Quartz Monzonite and mafic dikes (Dave's Falls County Park just south of Amberg; 45.497° N., 87.989° W.)

Excellent exposures of the Athelstane Quartz Monzonite occur on both sides of the Pike River. The Athelstane is cut by mafic dikes (~3 to 20 m wide) that strike about north-south. The dikes, at least three of which are exposed in the park, weather recessively relative to the Athelstane (Figure 11). The dikes have an andesitic composition (SiO_2 ~53 to 55 wt.%) with low MgO (~2.5 to 3.5 wt.%) and TiO_2 (~1.8 wt.%) contents, enriched light REE chondrite normalized patterns, and negative Nb-Ta and Ti anomalies on a primitive mantle normalized trace element plot (Figure 12). Dikes with similar composition have been observed cutting outcrops of the Twelve Foot Falls Quartz Diorite and in drill core cutting the Back Forty massive sulfide deposit in Michigan (Schulz, unpublished data). The dikes are post-1,835 Ma and pre-Keweenaw in age, but their actual age is not known.



Figure 11. Mafic dike (in valley, looking north) cutting the Athelstane Quartz Monzonite.

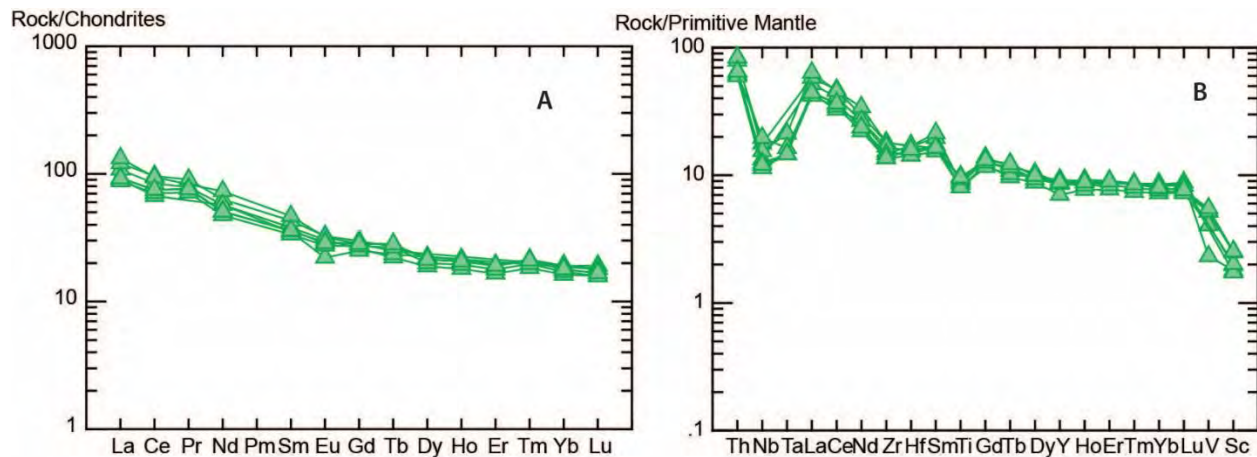


Figure 12. Chondrite normalized rare earth element plot (A) and primitive mantle normalized trace element plot (B) for andesite dikes from northeastern Wisconsin (Schulz, unpublished data).

Stop 10. Spikehorn Creek Granite (East side of U.S. Highway 8 at intersection with Morgan Park Road; 45.703° N., 87.981° W.)

The road cut shows the Spikehorn Creek Granite with metabasalt inclusions (Figure 13). As described in Sims et al. (1992, p. 27-28)... *"The Spikehorn Creek Granite is a gray to pinkish-gray, fine- to medium-grained rock containing scattered anhedral potassium feldspar grains as much as 2 cm in diameter. It is generally massive, but locally (especially near the margins of the body), it bears a mylonitic foliation expressed mainly by recrystallized quartz leaves and oriented biotite....The granite has sharp intrusive contacts against the Quinnesec volcanic rocks and the Marinette Quartz Diorite, and small ramifying dikes intrude these rocks for distances as much as 400 m from the contact....The Spikehorn Creek Granite (of the Niagara lobe) and the Hoskin Lake Granite are compositionally similar except that the Hoskin Lake has slightly higher K₂O content."*



Figure 13. Spikehorn Creek Granite with angular metabasalt inclusions.

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