



**INSTITUTE ON
■ L A K E ■
SUPERIOR
GEOLOGY**

PROCEEDINGS

**38TH ANNUAL MEETING
MAY 6-9, 1992**

HELD AT

HURLEY, WISCONSIN

VOLUME 38

MAY 1992

**PART 2
FIELD TRIP GUIDEBOOK**

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General Chairman: Albert B. Dickas, University of Wisconsin-Superior
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Wisconsin Geological and Natural History Survey

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BRUCE A. BROWN,
WISCONSIN GEOLOGICAL
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PART 2
FIELD TRIP GUIDEBOOK

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**ARCHEAN AND
EARLY PROTEROZOIC
GEOLOGY OF THE
GOGEBIC DISTRICT,
NORTHERN MICHIGAN
AND WISCONSIN**

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WITH A CONTRIBUTION FROM

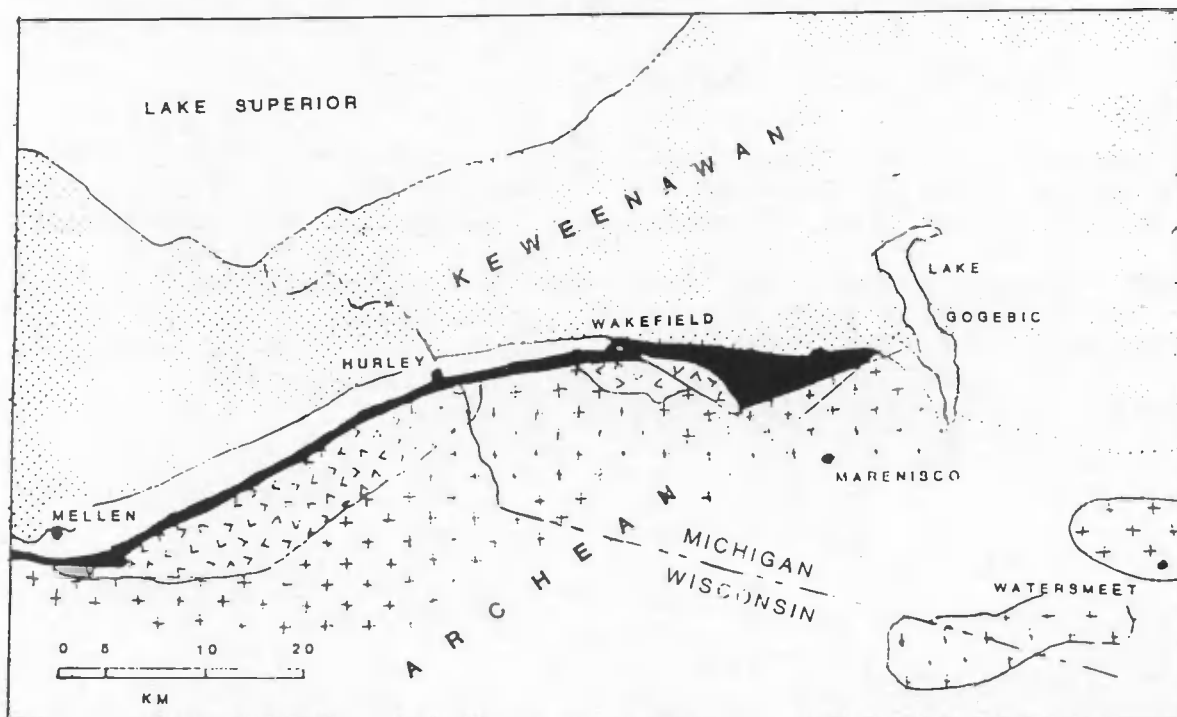
Kathy J. Licht
St. Norbert College

This field trip is the outgrowth of numerous student field trips led by the authors over the past quarter century. Most of the stop localities have been suggested by geologists who have worked in the district. However, some localities are the result of our own studies in the area. One of us (LaBerge) spent several weeks mapping in the area in July, 1991, supported by the U.S. Geological Survey, and Ojakangas worked in this region intermittently with U.S. Geological support from 1978 until 1990.

EARLY PROTEROZOIC GEOLOGY OF THE GOGEBIC DISTRICT NORTHERN MICHIGAN AND WISCONSIN

INTRODUCTION

The Gogebic range extends from Lake Gogebic, Michigan, westward approximately 128 km to near Lake Namekagon, Wisconsin (Figure 1). The major structure in the area is a north-facing monocline that exposes Archean rocks to the south and Middle and Late Proterozoic (Keweenaw Supergroup) rocks to the north.



EXPLANATION

MIDDLE PROTEROZOIC

 Volcanics and sediments

EARLY PROTEROZOIC

 Graywacke-slate
 Iron-Formation

ARCHEAN

 Granitoids
 Volcanics

Figure 1. Generalized map of the Gogebic range, showing the Early Proterozoic rocks sandwiched between Archean rocks on the south and Middle Proterozoic Keweenaw Supergroup rocks on the north. (Adapted from Morey and others, 1982)

Early Proterozoic rocks lie discordantly between these major sequences. As a result of the steep northerly dip of the monocline (a consequence of Keweenaw tilting), the geological map of the district is basically a stratigraphic cross-section.

The general geology of the Gogebic range has been known since the pioneering work of Irving and Van Hise (1892), Van Hise and Leith (1911), Hotchkiss (1919), and Aldrich (1929). Although subsequent studies have done much to "refine" the geological picture -- and much still remains to be done -- most of the early interpretations have withstood the test of time. The broad-scale geology of the eastern Gogebic district is presented on the Iron River 1° x 2° Quadrangle (Cannon, 1986).

STRATIGRAPHY

The Gogebic range contains excellent exposures of rocks that range in age from Archean (about 2,700 Ma) and Early Proterozoic (2,300 - 1,900 Ma), to Middle and Late Proterozoic-"Keweenaw"- (1,200 - ~600 Ma). Stratigraphic and structural relationships of several of the rock sequences can be observed in outcrops. Thus, it is possible to demonstrate the relationship of the rock sequences to each other. The general sequence of rock units in the Gogebic district is shown in Figure 2, a diagrammatic longitudinal section along the Gogebic range.

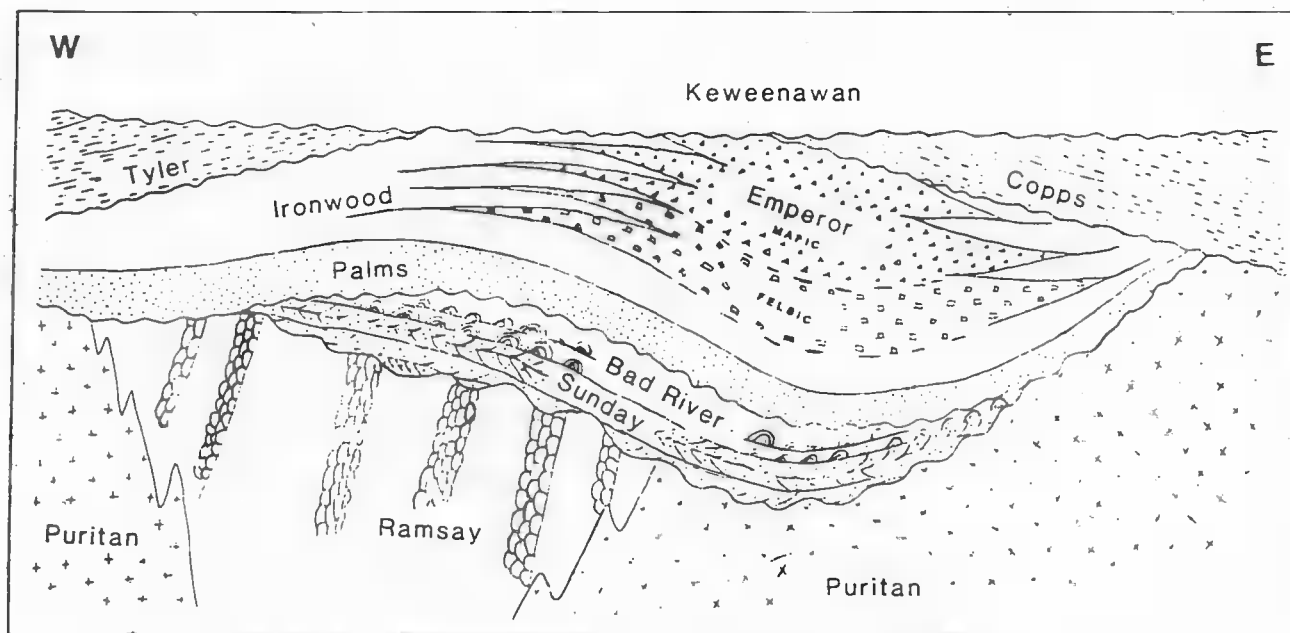


Figure 2. Idealized sketch showing generalized stratigraphic section on the eastern Gogebic range.

ARCHEAN ROCKS

Archean rocks in the Gogebic range are variably deformed and metamorphosed greenstones and granitoid rocks typical of Archean greenstone-granite terranes (or belts), and, indeed, are an extension of the Wawa Greenstone belt of Canada. The district is near the southern margin of the Superior Province of the Canadian Shield. Because this field trip is concerned primarily with the Early Proterozoic rocks, only a brief description of the Archean rocks is presented here.

Ramsay Formation. Archean volcanic rocks are exposed in a belt about 5 km wide and 16 km long in the eastern Gogebic range (Prinz and others, 1975). Schmidt and Hubbard (1972) named volcanic and sedimentary rocks in the Ramsay-Wakefield area as the Ramsay Formation. The Ramsay Formation consists mainly of pillowed, fragmental, and massive basaltic rocks in the eastern part of the area; however, felsic volcanic rocks and meta-graywackes are dominant to the west, near the Wisconsin border. Most of the rocks have been metamorphosed to greenschist facies, with well-preserved pillows and other primary features. Locally, especially adjacent to the Puritan Quartz Monzonite, the rocks have been metamorphosed to amphibolite facies, and few, or no, primary features are preserved. The Ramsay Formation may be the oldest rock unit in the Gogebic range proper, however, older gneisses (to 3,560 Ma) are exposed in the Watersmeet area, some 30 km southeast (Sims and others, 1984).

Puritan Quartz Monzonite. The volcanic-sedimentary sequence of the Ramsay Formation has been intruded by large granitoid masses named the Puritan batholith (Schmidt and Hubbard, 1972). The Puritan Quartz Monzonite exposed in the central and western Gogebic range is a weakly-deformed post-tectonic pluton that ranges in composition from granite to tonalite and has been dated to $2,735 \pm 16$ Ma (Sims and others, 1984).

Granitoid rocks are also extensively exposed in the Marenisco area on the eastern end of the Gogebic range. Parts of the granitoid has prominent gneissic banding and associated amphibolite. The other, more extensive phase, is a weakly deformed medium-grained granite with widely scattered pegmatite bodies (Fritts, 1969; Trent, 1973).

The Archean rocks were eroded and are unconformably overlain by Early Proterozoic rocks of the Marquette Range Supergroup. Because the Gogebic range is a north-dipping monocline, Early Proterozoic rocks rest on Archean rocks to the south along the entire length of the range.

EARLY PROTEROZOIC ROCKS

At the ends of the Gogebic range the Archean rocks are unconformably overlain by rocks of the Chocoday Group of the

Marquette Range Supergroup, Sunday Quartzite and Bad River Dolomite at the eastern end and only Bad River Dolomite at the western end. Chocoday Group rocks are absent from the central part of the range, and evidently were eroded prior to deposition of overlying units.

Sunday Quartzite. The Sunday Quartzite was deposited unconformably on Archean rocks (mostly the Ramsay Formation), however, no regolith on the Archean has been observed. The basal unit is a prominently cross-bedded reddish quartzite. Conglomerate layers with quartz and granite cobbles to about 8 cm are present within the lower part of the formation. Most of the formation is a gray vitreous quartzite in which cross-bedding is common and current ripple marks are well developed in places. According to Schmidt (1973) the Sunday Quartzite is present only on the eastern end of the range, where it has a maximum thickness of about 46 m. One of the type localities of the Sunday Quartzite has been described in the center of the district near the Newport Mine (Van Hise and Leith, 1911) and the problem of this "lost locality" has been discussed by Schmidt (1973).

Bad River Dolomite. The Sunday Quartzite grades upward into the Bad River Dolomite. The transition is marked by interbedded dolomite and quartzite; the dolomite beds are thicker and more abundant and quartzite beds are thinner and less abundant upward. The dolomite weathers to a distinctive tan/brown color and contains abundant layers and irregular patches of gray to black chert. Stromatolitic layers with mounds that range in size from about 5-50 cm in diameter become more abundant upward in the dolomite. Stromatolitic units tend to be prominently silicified. Maximum thickness of the dolomite is about 120 m, but in most areas it is considerably less.

A period of erosion followed deposition of the quartzite and dolomite of the Chocoday Group, removing the Sunday Quartzite and Bad River Dolomite from all but the eastern and western ends of the Gogebic range.

Palms Formation. The Palms Formation is the basal unit of the Menominee Group of the Marquette Range Supergroup. It rests unconformably on the Bad River Dolomite and Sunday Quartzite on the east and west ends of the range, and on the Puritan Quartz Monzonite and Ramsay Formation in the central part of the range. Little or no regolith is developed on the erosion surface.

The Palms Formation is about 146 meters thick and contains a basal mud-rich unit, a central interbedded mud-silt-sand unit, and an upper sand-rich unit (Ojakangas, 1983). A thin (usually less than 3 m thick) conglomerate occurs locally at the base of the formation (Aldrich, 1929). The Palms grades abruptly upward into the Ironwood Iron-formation. Thin (1-5 cm) beds of granular iron-formation are interbedded with quartzite (Ojakangas, 1983) across a thickness of 10 m. The abundance and thickness of beds of iron-formation increases upward with a corresponding decrease in detrital beds. Recent studies by Ojakangas (1983) indicate

that the Palms Formation may have been deposited by tidal currents in a transgressing sea (Figure 3).

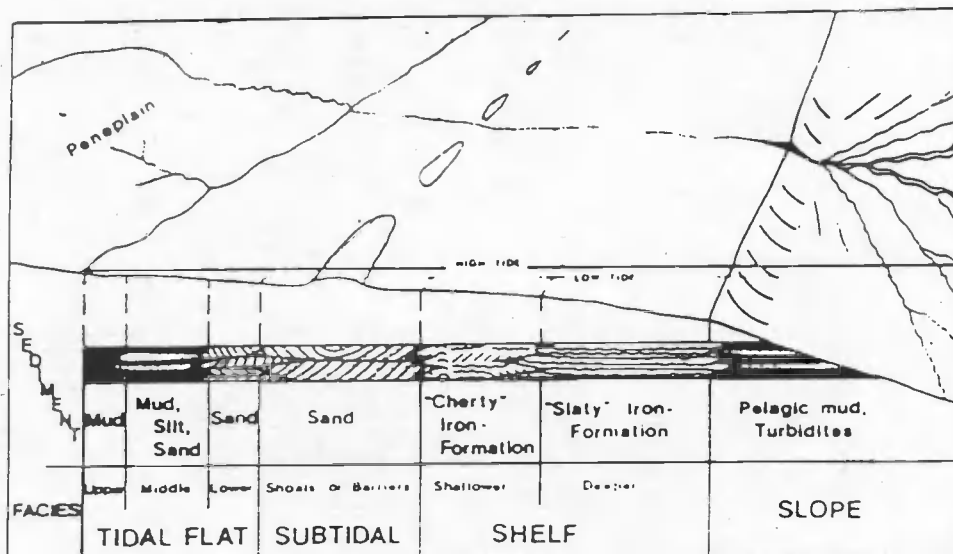


Figure 3. Sedimentational model showing lateral relationships of the siliciclastic tidal facies, the iron-formation facies on the shelf, and the deeper-water turbidite-mud facies. Thicknesses of units not drawn to scale. (From Ojakangas, 1983)

Ironwood Iron-Formation. The transition from the Palms Formation into the Ironwood Iron-Formation records an abrupt change from detrital to chemical sedimentation. Like other iron-formations, the Ironwood contains little detrital material even though it is 150-275 m thick.

Although iron-formation has a simple composition -- chert and iron minerals -- it is extremely varied in appearance. Both the chert and the iron minerals are varied in color. Although it is widely cited as indicating the chemistry of the depositional environment (James, 1954), the mineralogy of iron-formations is the product of the depositional, diagenetic, metamorphic, and to some extent, the weathering environment that the rocks have undergone. Still moderate subsequent changes to the primary mineralogy do not usually reduce our ability to infer primary mineralogy.

Iron-formations have two basic textural types. One is laminated with layers of chert about 5-10 mm thick alternating with layers of iron minerals of similar thickness. This laminated type of deposit is commonly referred to as "slaty" iron-formation (Figure 4). It has a grain size and bedding characteristic of siltstones (LaBerge, 1964, Dimroth, 1968). The other textural type of iron-formation, referred to as "cherty" iron-formation, consists of 1-10 cm thick layers containing fine sand-size (<2 mm) "grains" of chert with variable amounts of iron minerals set in a chert matrix. These layers of "granular" chert are extremely variable in thickness over short distances, commonly forming lenses (Figure 5). The cherty lenses are



Figure 4. Photo of thin-bedded (laminated) iron formation with some thick-bedded (granular) layers. Mt. Whittlesey, near Mellen, WI.



Figure 5. Photo of thick, irregularly bedded, granular iron-formation. Mt. Whittlesey, near Mellen, WI.

typically separated by 0.5-3 cm layers of laminated iron minerals. "Cherty" iron-formations have the grain size and bedding characteristics of sandstones (Mengel, 1963; Dimroth and Chauvel, 1973). The mineralogy of the iron minerals is almost completely independent of the textural varieties of iron-formation. Most geologists agree that granular ("cherty") iron-formation represents shallow-water deposition and laminated iron-formation represents somewhat deeper (quieter) water deposition.

Several iron-formations in the Lake Superior region have been subdivided into "members" on the basis of their bedding (textural) characteristics. Hotchkiss (1919) divided the Ironwood Iron-formation into five members. The ~60 m thick Plymouth Member is an irregularly bedded granular cherty unit with a basal zone containing some detrital quartz and patches of stromatolitic jasper. The overlying 46 m thick Yale Member is mainly an even-bedded (slaty) siderite-chert unit with a basal carbonaceous pyritic slate zone and a layer that is probably tuffaceous. The succeeding Norrie Member is a 60 m thick unit of irregularly bedded cherty iron-formation. The Pence Member is about 45 m of siderite-chert iron-formation, and the uppermost Anvil Member is a 15-60 m thick mixture of slaty and cherty iron-formation. These members are continuous and uniform throughout the 128 km length of the range. However, they are difficult to distinguish east of Wakefield, Michigan, where a thick sequence of volcanic rocks, the Emperor Volcanic Complex (Trent, 1973), is interbedded with the iron-formation.

Iron ores. The iron ores on the Gogebic range were produced by the chemical removal of silica from the iron-formation. Dissolution of the silica required the removal of nearly 50 percent of the original volume of the rock and produced very porous orebodies. The ore consisted mainly of earthy goethite and hematite; however, in open spaces produced by removal of silica, large botryoidal masses of crystalline hematite and goethite were found, as well as local concentrations of psilomelane (romancheite), manganite, rhodochrosite, calcite, manganocalcite, barite, gypsum, and marcasite. The presence of iron ore cobbles in Keweenawan rocks in the district suggests that at least some of the ore was formed prior to deposition of rocks of the Keweenawan Supergroup. Nearly all the mines on the Gogebic range were underground mines (Figures 6 and 7). Mining on the Gogebic range began in 1884 and ended in 1967, when the underground mines could no longer compete with the large open pit taconite mines elsewhere in the world. Ore was taken by rail to Ashland, Wisconsin, and Escanaba, Michigan, from where it was shipped to steel mills in Indiana and Ohio.

Emperor Volcanic Complex. The Emperor Volcanic Complex constitutes a thick pile of volcanic rocks and sills in the sedimentary sequence on the eastern Gogebic range where they reach a thickness of at least 2,000 meters. Although the volcanic rocks in the eastern Gogebic range have been known for 100 years (Irving and Van Hise, 1892), few studies have been undertaken on these rocks. Irving and Van Hise (1892) stated

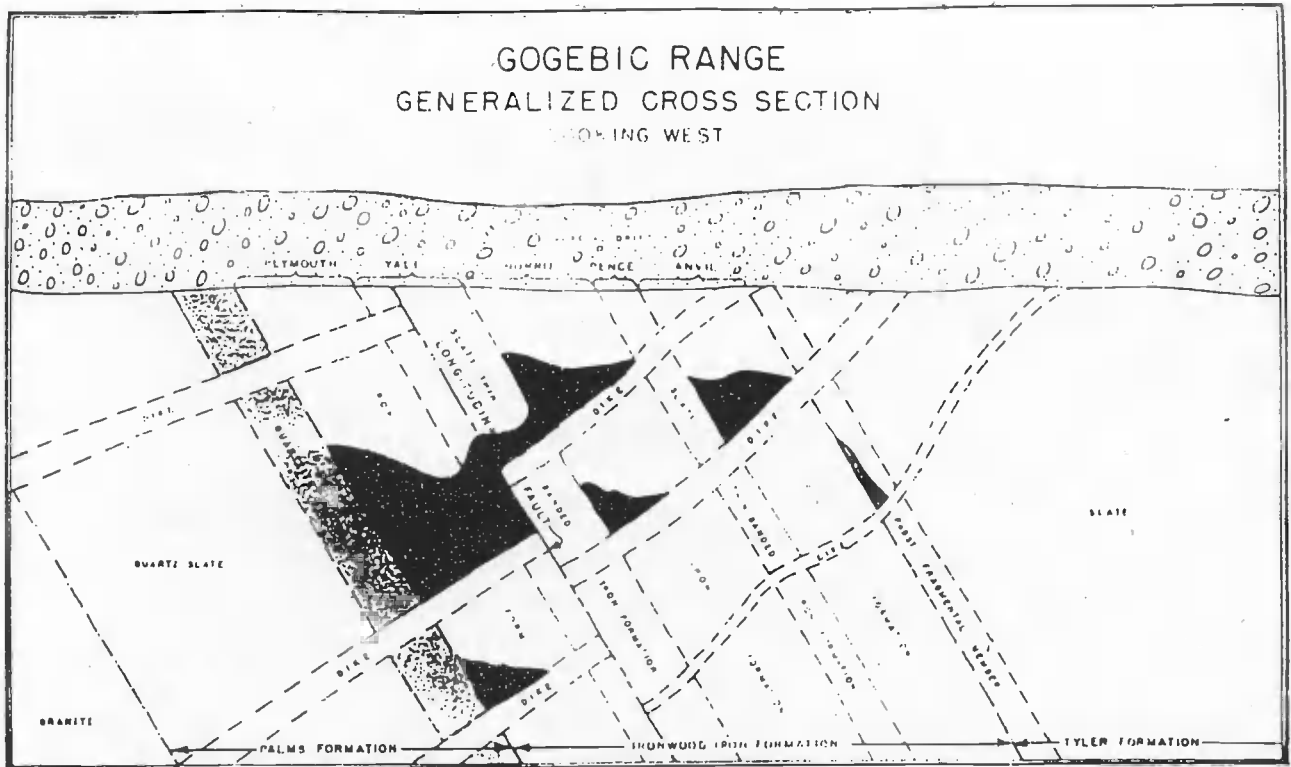


Figure 6. Schematic diagram showing the occurrence of iron orebodies on the Gogebic range. (From Goldich and Marsden, 1956)

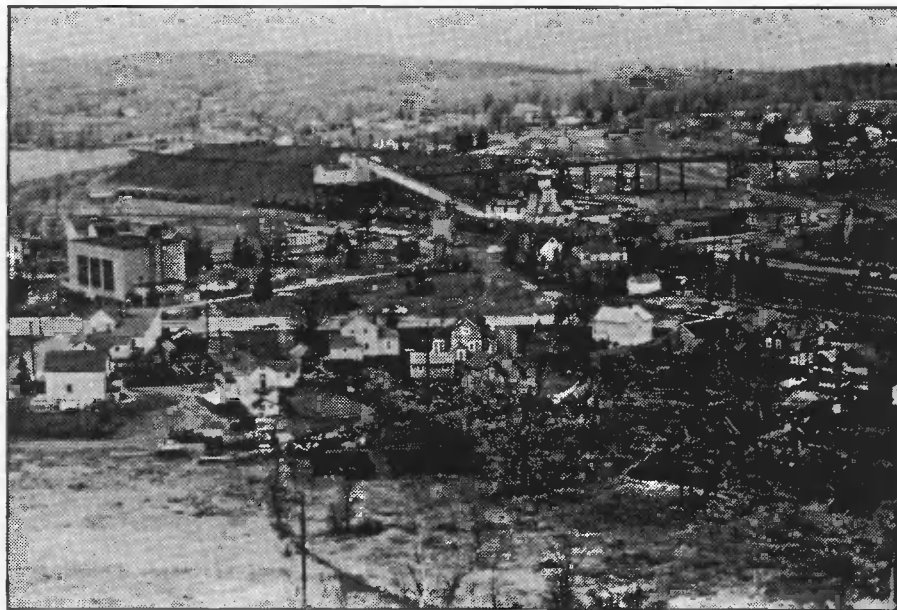


Figure 7. Sunday Lake mine in 1960. Note the stockpile of iron ore in the upper left. Photo taken from the top of Radio Tower Hill In Wakefield, Michigan.

that the volcanic rocks are interbedded with the Ironwood Iron-formation, which has been demonstrated convincingly by subsequent exploration drilling for iron ore in the area. Mapping by Trent (1973) documented the distribution of igneous rocks and showed some of the lithologies present. Dann (1978) showed that the volcanics range in composition from basalt to dacite, and that they have been metamorphosed only to prehnite-pumpellyite, or to lowest greenschist facies. He also elaborated on the possible volcanic setting in which the rocks may have formed. Trace element studies by Schulz (Sims and others, 1990) indicate that the volcanic rocks are rift-related continental tholeiites.

Reconnaissance mapping in 1991 by LaBerge and J. S. Klasner indicates that the Emperor Volcanic Complex consists of a wide variety of mainly subaqueous volcanic rocks and sills. Both mafic and felsic volcanics were recognized. Mafic rocks include sills (such as the Wolf Mountain sill of Trent, 1973) pillowed and massive flows, some with columnar jointing, and extensive hyaloclastites and pillow breccias (Figure 8). Felsic rocks include hyaloclastites, massive felsite breccias, and units comprised of felsite breccia with 2-10 cm clasts in a finer hyaloclastite matrix (Figure 9). The felsite breccia - hyaloclastites may be debris flows from a subaqueous felsic dome. Although there are zones in the Emperor Volcanic Complex with strong foliation, large areas are almost undeformed. Because of the low metamorphic grade and minimal deformation, primary features are well-preserved.

The relationship between volcanism and iron-formation has long been debated, with some advocates (e.g., Van Hise and Leith, 1911) arguing in favor of volcanic contributions of iron, and others (e.g., James, 1954) contending that there is no genetic relationship between volcanism and iron-formation deposition. The presence of a thick volcanic sequence interbedded with the iron-formation in the eastern Gogebic range shows that there certainly was volcanism contemporaneous with iron-formation deposition in the Lake Superior region. Therefore, the Gogebic range is similar to the Hamersley range in Western Australia and the Proterozoic iron-formations in South Africa where numerous volcanic ash layers are interbedded with the iron-formations (LaBerge, 1966a, 1966b).

Tyler and Copps Formations. The Ironwood Iron-Formation, and locally the Emperor Volcanic Complex, are overlain by a 2,000 m thick graywacke-slate sequence. The unit is thickest in the western and eastern ends of the range, and has been completely removed by erosion prior to deposition of the Keweenaw Supergroup in an area near Wakefield, Michigan. Throughout most of the range the unit is called the Tyler Formation, however the correlative(?) portion at the eastern end of the range is called the Copps Formation. The Tyler Formation is correlated with the Michigamme Formation to the east, and with the Thomson, Rabbit Lake, Virginia and Rove Formations to the west and north in Minnesota and Ontario. All these units were probably deposited in the same basin.

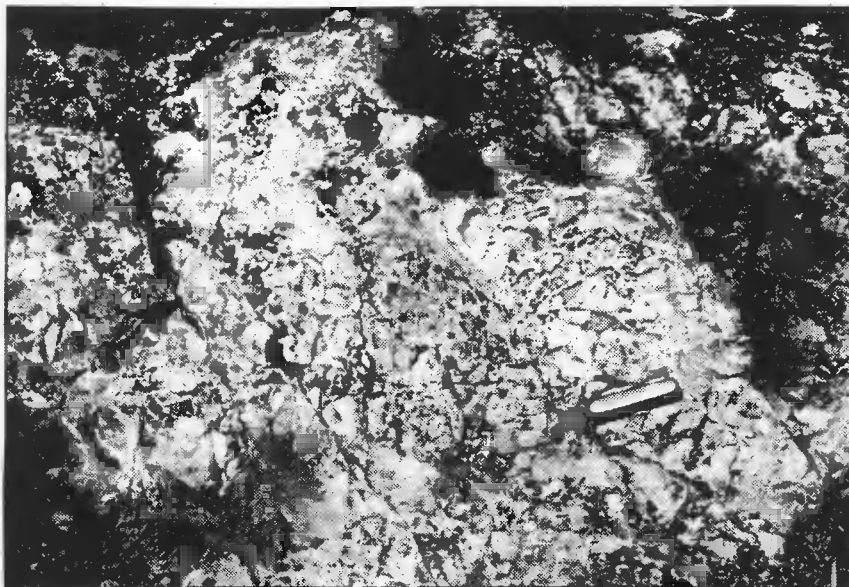


Figure 8. Pillow breccia in Emperor Volcanic Complex. Pillows and pillow fragments in a hyaloclastite matrix. Jackknife is on a pillow.



Figure 9. Felsic phase of Emperor Volcanic Complex. Felsite clasts in a finer hyaloclastite matrix. Jackknife is 8 cm.

The relationship of the Tyler Formation to the underlying iron-formation is unclear. According to Hotchkiss (1919) the basal unit of the Tyler Formation is the Pabst Member, a distinctive unit containing mainly jasper and quartz pebbles. Figure 10, from drill hole data (Atwater, 1938) shows the relationship of the conglomerate to the underlying iron-formation. Hotchkiss (1919) interprets the relationships between the Tyler and the iron-formation to be unconformable. Atwater (1938) did not believe that a major unconformity separated the Ironwood and Tyler. Schmidt and Hubbard (1972) interpreted the formations to be gradational and sedimentation to have been continuous in the central Gogebic. However, Trent (1973) stated that field relationships indicate an unconformity between the Ironwood and Tyler on the eastern Gogebic. Schmidt (1980) summarizes the field evidence pertaining to the Ironwood-Tyler contact as well as the several alternative interpretations.

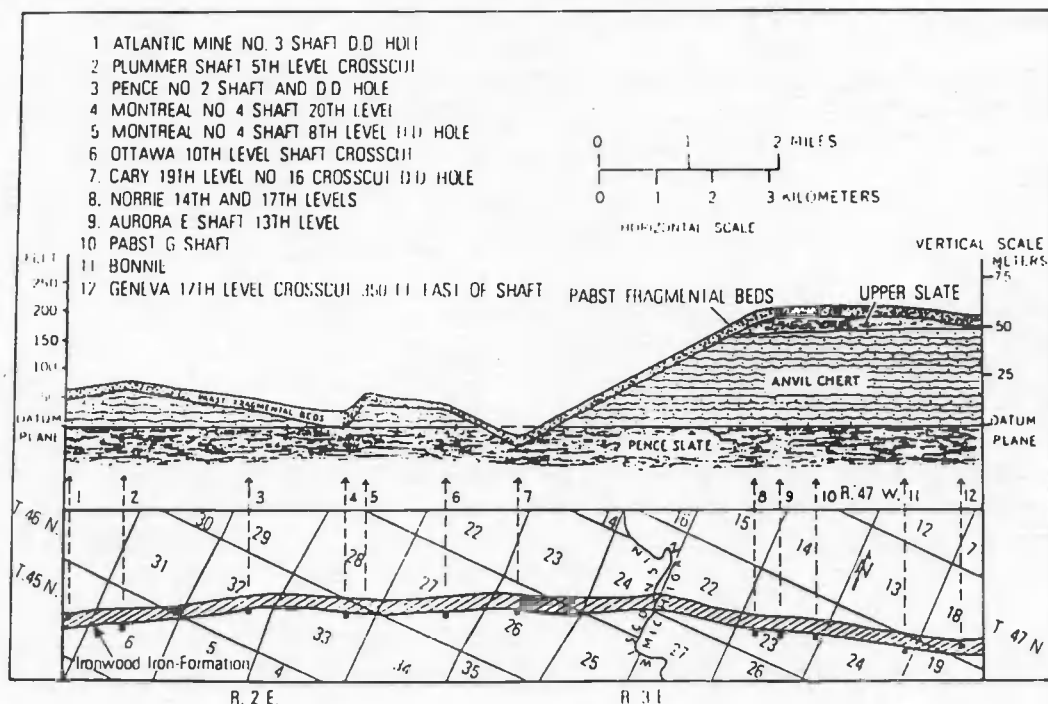


Figure 10. Longitudinal section through part of the Gogebic district, showing one interpretation of relationship between the "Pabst Member" of the Tyler Formation and the Anvil and Pence Members of the Ironwood Iron-Formation. The upper slate is a layer of thin-bedded carbonate iron-formation, at least partly within the Anvil Member. Compiled from measured sections in mines and drill holes. Location of sections given in map of Ironwood Iron-Formation in lower part of diagram. Modified from Atwater (1938, p. 163, fig. 2) (From Schmidt, 1980)

The Tyler Formation consists dominantly of intercalated argillite with lesser amounts of metasiltstone and graywacke. Graded metagraywacke beds containing Bouma sequences indicative of deposition by turbidity currents are common, but structureless beds are even more abundant and a grain-flow mechanism may also have been operative. Sole marks and small-scale cross-bedding indicate a paleocurrent trend from east-southeast to west-northwest (Alwin, 1976).

The "average" graywacke of the Tyler consists of 28 percent micaceous matrix and 72 percent framework grains (Alwin, 1976). Of the framework grains, 73 percent are mono- and polycrystalline quartz and chert, 10 percent are plagioclase (mostly altered plagioclase), and 17 percent are rock fragments, mainly of granitic to quartz dioritic and volcanic origin, but some grains of sedimentary and metamorphic origin also are present. Thus, the "average" graywacke of the Tyler is compositionally submature quartzose lithic graywacke that is texturally immature as defined by angular framework grains and poor sorting.

The paleocurrent directions and the composition of the graywacke suggest that the Tyler Formation was derived from a dominantly "granitic" source to the southeast.

MIDDLE PROTEROZOIC ROCKS

Following the Penokean orogeny, about 1,860 - 1,830 Ma, the Gogebic range and the rest of the Lake Superior region was subjected to erosion for nearly 700 million years. Erosion removed an unknown thickness of rocks in the range, but more rock was removed from some areas than in others. For example, at least 2,000 meters of the Tyler Formation is preserved in the western Gogebic, and a comparable thickness of the Copps Formation is preserved at the eastern end of the range. However, all of the Tyler, and some of the underlying Ironwood Iron-Formation as well, was removed in the eastern Gogebic where the range experienced more folding. The erosion interval presumably ended with the onset of the Middle Proterozoic Keweenawan rifting event.

Bessemer Quartzite. In most of the Gogebic range the Bessemer Quartzite, the basal unit of the Keweenawan Supergroup, unconformably overlies the Tyler Formation. However, locally on the eastern part of the range the basal Keweenawan rests on the Ironwood Iron-Formation or the Emperor Volcanic Complex. The quartzite is approximately 100 m thick, has channel cross-bedding and is a relatively mature quartz sandstone (Ojakangas and Matsch, 1982). The Bessemer Quartzite evidently originated in a fluvial environment in a slight sag that presaged the Keweenawan rifting that developed into the Midcontinent Rift System (Ojakangas and Matsch, 1982).

Powder Mill Group. Overlying the Bessemer Quartzite is a thick sequence of mainly basaltic lava flows that formed during the main phase of Keweenaw rifting in this area. Volcanic activity (and rifting?) began approximately 1,200 Ma ago and produced hundreds of lava flows that have an aggregate thickness of more than 3,000 m. The lowermost basaltic lava flows that overlie the Bessemer Quartzite in the Bessemer, Michigan, area are pillowed, suggesting that they were extruded in water, either a lake or a fluvial system.

Oronto Group. Keweenaw Supergroup volcanic rocks are overlain by a thick sequence of mainly red clastic rocks in the Lake Superior region. Various representatives of this sequence are well exposed north of the Gogebic range. The basal unit of the Oronto Group is the Copper Harbor Conglomerate, which is as thick as 2,000 m. The major lithology is red arkosic sandstone and siltstone with numerous conglomeratic horizons. The Copper Harbor Conglomerate probably represents alluvial fan deposits formed near a rugged source area along the south side of Lake Superior.

The Copper Harbor Conglomerate is overlain by the Nonesuch Formation, a 100 meter thick clastic unit, that is dominantly gray to black carbonaceous siltstone. The Nonesuch was probably deposited in a lake developed on the subsiding pile of volcanic rocks and Copper Harbor sediments (e.g., Suszek, 1991). In addition to the world class copper deposits of the White Pine mine, the Nonesuch Formation also contains one of the oldest known occurrences of petroleum.

The Freda Sandstone of the Oronto Group consists of approximately 3,000 m of red clastics, and conformably overlies the Nonesuch Formation. The Freda strata is believed to consist of fluvial deposits formed during continued subsidence of the Middle Proterozoic rift.

KEWEENAWAN TILTING

Following the onset of deposition of the volcanic and sedimentary rocks along the Keweenaw Midcontinent Rift System, the south side of the rift, including the Gogebic district, was tilted steeply (65-90°) to the north.

TECTONIC SETTING

Early Proterozoic rocks in the Gogebic range unconformably overlie Late Archean greenstone and granite of the Superior Province of the Canadian Shield. Larue (1981, 1983) suggested that the sedimentary rocks of the Lake Superior region (the Marquette Range Supergroup) were deposited in a number of graben-like basins in a platformal environment. Sims and others (1990)

proposed that rocks of the Marquette Range Supergroup formed on a rifted continental margin. Break-up of the Archean craton resulted in the development of oceanic crust off the southern margin of the Superior Province. Therefore, the sedimentary rocks were deposited during a rifting stage that developed into a passive margin. Nd isotope studies by Gerlach and others (1988) indicate that iron-formations in the Lake Superior region were deposited about 2,100 Ma.

The concept of a rifted continental margin in the Lake Superior region is probably best exemplified by the eastern Gogebic range. As discussed earlier, the stratigraphic succession increases considerably in thickness east of Wakefield, Michigan, and igneous rocks (volcanic rocks and sills) become a major component of the succession. The increase in thickness of the sedimentary sequence and the influx of volcanic rocks is suggestive of a half-graben structure with a hinge near Wakefield and a boundary fault (or faults) near Lake Gogebic. Emplacement of basaltic sills (e.g., the Wolf Mountain sill of Trent, 1973) and eruption of continental tholeiites (Sims, and others, 1990) would be characteristic of a rifted continental margin. The presence of the sills and volcanic rocks in the Ironwood Iron-Formation as well as dramatic eastward thickening of the iron-formation indicates that major graben formation and volcanism occurred during deposition of the Ironwood Iron-Formation (Figure 11).

Following the rifting and passive margin stages, a phase of convergent tectonics produced a volcanic island arc (the Wisconsin magmatic terranes) that collided with the Superior craton. Isotopic ages in rocks of the magmatic terrane range from 1,890 Ma to 1,840 Ma (Sims and others), and docking occurred about 1,860 Ma (Sims and others, 1990). Barovich and others (1989) showed that Nd isotopes indicate that much of the graywackes in the Michigamme Formation were derived from an Early Proterozoic source. This suggests that the graywackes were deposited in a foreland basin during docking of the island arc on the margin of the Superior craton.

Deformation and metamorphism of rocks of the Marquette Range Supergroup on the Gogebic range resulted from the collision of the island arc with the margin of the Superior craton. This collision fits very well with a foreland basin model as developed by Hoffman (1987) for the Early Proterozoic rocks that surround the Archean Superior Province. This model was utilized by Southwick and Morey (1991) to explain lithologic and structural relationships in east-central Minnesota. A similar foredeep model, with associated southward subduction and collision of a volcanic arc against the craton to the north was proposed by Ojakangas (in press) to explain the origin of the Michigamme Formation of the Upper Peninsula of Michigan, as well as the origin of the Copps and Tyler formations in the vicinity of the Gogebic range. In this model, the iron-formation was deposited on the peripheral bulge on the north side of the northward-migrating foreland basin, whereas the aforementioned turbidite

sequences were deposited in the deeper axial parts of the basin. This would require that the Ironwood Iron-Formation and the overlying Tyler Formation have a gradational and conformable relationship, and that they were deposited, at least in part, contemporaneously. In this model, axial tholeiitic volcanism (Hoffman, 1988) and hydrothermal activity in the foredeep would have provided the iron and silica that were precipitated as iron-formation on the shelf, as upwelling brought the solutions to a favorable site of deposition.

Note that the foreland basin model provides an alternative interpretation for the Ironwood and the intertonguing Emperor Volcanic Complex compared with a rifted continental margin model. Obviously more work is needed to resolve this problem.

Deformation during the Penokean orogeny may have resulted in reactivation of the normal faults in the graben structures. We suggest that the structure in the eastern Gogebic may represent a half-graben that has been subjected to thrusting during the Penokean orogeny (Figure 12). Perhaps significantly, erosion prior to deposition of the Keweenawan Supergroup cut deepest into the Early Proterozoic sedimentary sequence in the central Gogebic, where the proposed thrust faults are present. Whereas over 2,000 m of the Tyler Formation remain on the western Gogebic in Wisconsin, and a similar thickness of Copps Formation is present at the eastern end of the range, erosion removed all of the Tyler and part of the iron-formation east of Wakefield. Assuming that erosion cut down to a relatively uniform level, then the central Gogebic may have been uplifted at least 2,000 m more than areas to the east and west. An Archean cored gneiss dome is present near Watersmeet about 50 km to the east. That area may represent an even more pronounced uplift. Therefore, rocks on the eastern Gogebic range may record depositional and deformational events of both the rifted margin and convergent tectonics during the Early Proterozoic in the Lake Superior region.

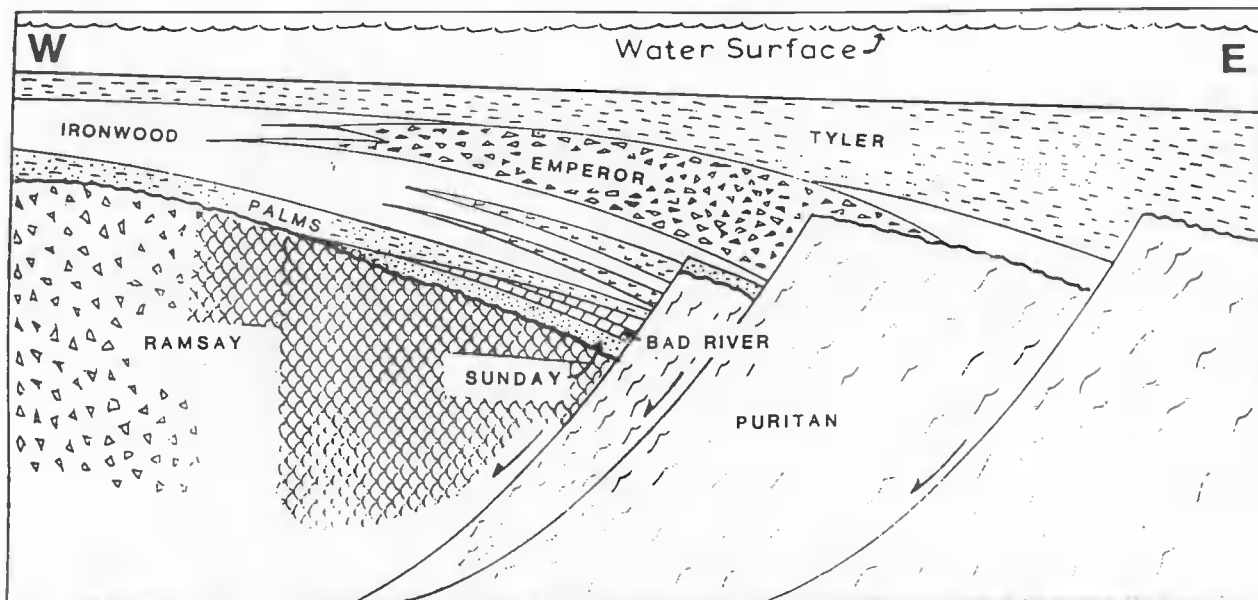


Figure 11. Diagram showing hypothetical pre-Penokean relationships on the eastern Gogebic range.

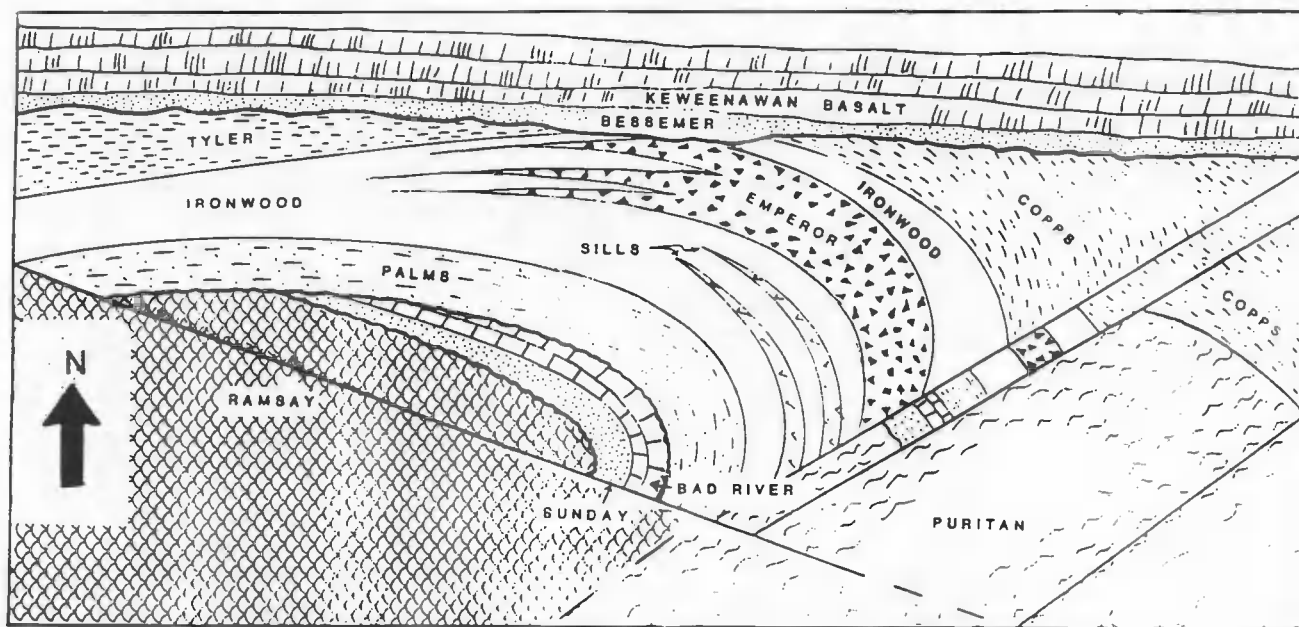


Figure 12. Simplified geologic map of the eastern part of the Gogebic range. (Not to scale)

STOP DESCRIPTIONS

Gogebic Field Trip

Stop 1. Tyler Formation.

Location: Roadcuts at junction of U.S. Hwy. 2 and Hwy. 51 north of Hurley, WI

Rocks exposed at this locality are representative of the upper part of the Tyler Formation, which is about 2,100 meters thick here. It thickens westward to about 2,900 meters in Wisconsin and thins eastward to zero east of Ramsay, where it was entirely removed by erosion prior to deposition of Keweenawan strata. The Tyler is unconformably overlain by the Bessemer Quartzite and a substantial thickness of Keweenawan volcanic rocks.

The relationship of the Tyler to the underlying Ironwood Iron-Formation is uncertain because of poor outcrop. Most geologists (e.g., Atwater, 1938) consider the two formations to be conformable (Schmidt and Hubbard, 1972); however, Aldrich (1929) presents evidence for erosion of the iron-formation and a basal conglomerate at the base of the Tyler. The presence of a graywacke-slate sequence overlying the iron-formation is characteristic throughout the Lake Superior region

As elsewhere on the western Gogebic range, the beds here strike east-northeast, dip 60°-75° northwest, and top to the northwest, part of a large block that was tilted northward during Keweenawan rifting to form a large north-facing monocline. Schmidt and Hubbard (1972) and Klasner and others (1991) pointed out that the cleavage/bedding relationships in these (and other) outcrops are unusual and may be of tectonic significance. As shown in the accompanying diagrams (Figure 13), the graywacke-slate was dipping gently southward prior to Keweenawan time. Because the cleavage dips less steeply than the bedding, rotation of the rocks southward to their pre-Keweenawan sub-horizontal attitude results in a south-dipping cleavage.

Cannon and others (1990) showed that originally vertical Early Proterozoic diabase dikes in the Archean terrane to the south of the Gogebic range now have a southward dip. This dip is explained by a northward rotation along the east-northeast-trending Marenisco fault, a major listric thermal fault with perhaps 10 km of southward thrust motion on it. This motion explains the present steep northward dip of the major monoclinical structure of the Gogebic range.

Alwin (1976) showed that the Tyler consists dominantly of intercalated argillite and slate with lesser amounts of siltstone and graywacke. Graded

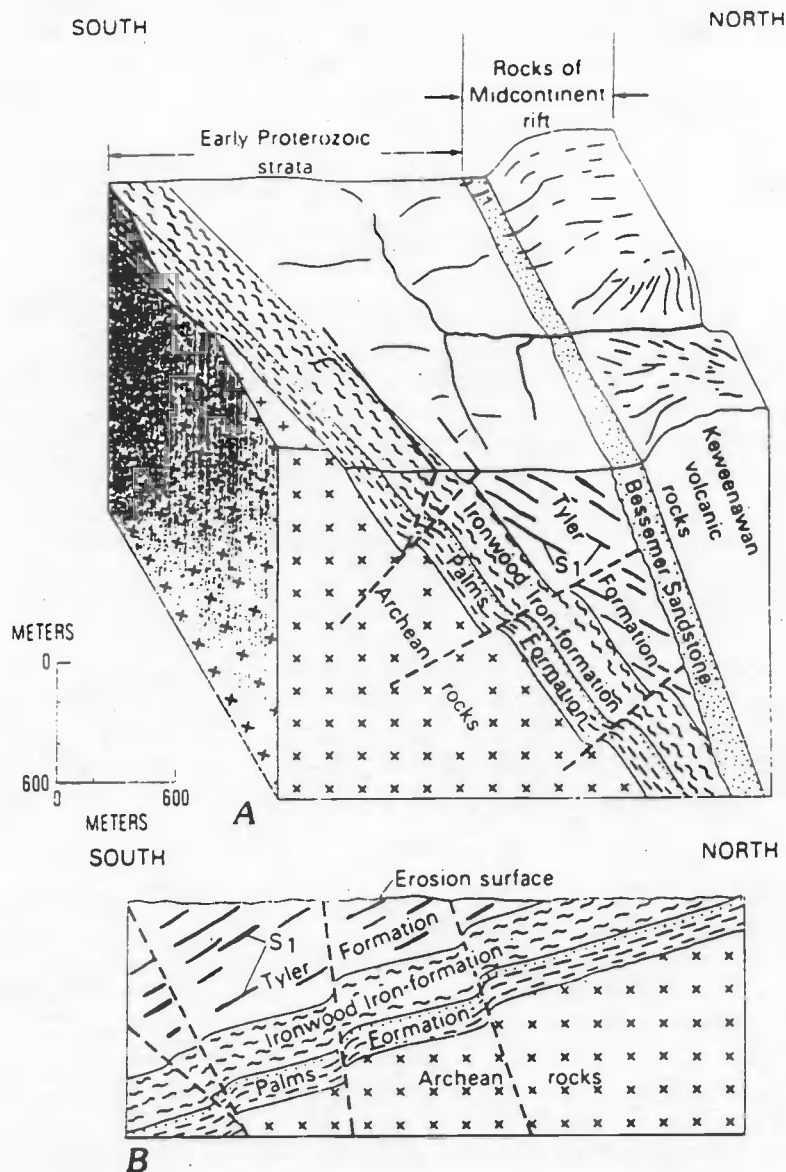


Figure 13.

Generalized block diagram and cross section illustrating structural relationships in Ironwood, Mich.-Hurley, Wis., area. Heavy dashed line, fault. A, Block diagram showing relationship of pre-Keweenaw strata to rocks of Midcontinent rift. Note that S₁ foliation in Tyler Formation dips less steeply than bedding. Modified from Schmidt and Hubbard (1972). B, Reconstructed post-Penokean, pre-Keweenaw position of the strata shown in A. Note gently south dipping S₁ foliation. Modified from Schmidt and Hubbard (1972). (From Klasner and others, 1991)

graywacke beds containing Bouma sequences indicate deposition by turbidity currents. Sole marks and small-scale cross-beds indicate a paleocurrent trend toward the west-northwest. Framework grains in the graywackes are quartz and feldspar and rock fragments of granitic, volcanic, and metamorphic rocks. Therefore, the source area for the Tyler sediments is inferred to be an Archean granitic-metamorphic (cratonic) source area to the southeast. The Tyler was only slightly metamorphosed to sub-greenschist facies during the Penokean orogeny, and the metamorphic grade here was not changed during the Keweenawan rifting.

At this exposure, note graded graywacke beds topping NW, mud-chips in some beds, rare sole marks, loading on some soles, flame structures, and small-scale cross-bedding. The assemblage of sedimentary structures indicates a turbidity current mechanism, and deposition on a submarine fan is likely but not essential. In the fan model the "packets" of graywacke beds would be channel deposits whereas the mudstones would be overbank deposits.

Continue east on U.S. Hwy 2 to Golf Club Road just east of Ironwood. Turn right (south) on Golf Club Road and proceed about 1.1 miles south to large brick buildings of the Newport Mine. Walk to the northeast across the small field to old railroad cuts, that are just north of the main building behind the trees. Caution: Area is now part of a gravel washing operation.

Stop 2. Old Newport mine (Archean/Early Proterozoic Unconformity).

The Palms Formation of the Menominee Group of the Marquette Range Supergroup rests upon the 2,700 m.y. old Puritan Quartz Monzonite (Schmidt, 1976). The Marquette Range Supergroup in the Gogebic range was deposited on a moderately flat erosion surface developed on Archean "granite" and greenstone.

At this locality, the steeply dipping (to the north) unconformable contact between the Palms Formation and Archean granitoid is exposed in the old railroad cut. Small depressions in the surface, as can be seen here, contain laminated cherty material, some of which is granular and somewhat resembles iron-formation. The granules are 0.5 to 1.0 mm in diameter and consist of mixtures of chert, chlorite, and calcite. There are also a few phosphatic argillaceous chips just above the contact. Note that there is no appreciable evidence of weathering of the granite beneath the Palms.

Red argillaceous and silty beds of the lower member of the Palms Formation are exposed here. They contain wavy bedding and small mudcracks, and have been interpreted as upper tidal flat deposits of a transgressing sea (Ojakangas, 1983).

Return to U.S. Hwy. 2, cross it, and continue north approximately 1 mile to rock knob on right, beyond the Kangas farm.

Stop 3. Contact between Bessemer Quartzite and Powder Mill Volcanics

This exposure is approximately two miles north of Stop 2, and the interval between these two stops demonstrates the thickness of the Early Proterozoic sequence in this area. The exposure shows the contact between the basal Keweenaw Bessemer Quartzite and the overlying basalt of the Powder Mill Group. Thus the exposure illustrates the abrupt change from sedimentation of quartzose sandstone to flood basalt volcanism during the Middle Proterozoic in this part of the Lake Superior region. Only the lowermost basalt flows are pillowed, and locally, the soft, silty sediment was squeezed up between rubbly blocks and pillows. Some units of Bessemer-type quartzite do occur higher in the flow sequence. These features indicate that volcanism here was initiated in a sedimentary basin, overwhelmed sedimentation, and resulted in a thick sequence of subaerial flows. Flows just to the north are massive.

Stop 4. Ironwood Iron-Formation and Ironwood-Tyler relationships. (Modified from Schmidt, 1972)

Location: Exposures on the east and west banks of the Black River upstream from Hwy. 2 on the western outskirts of Ramsay, Michigan. (NE $\frac{1}{4}$, SW $\frac{1}{4}$, Sec. 12, T.47N., R.46W.)

The Ironwood Iron-Formation is a major sedimentary Gogebic range, and is the source rock for all of the iron ores mined in the area. Unfortunately, the iron-formation is very poorly exposed on the eastern Gogebic. Excellent exposures of the Ironwood Iron-Formation are present on Mt. Whittlesey, southeast of Mellen, Wisconsin, however, that locality is more than 30 miles away, and will not be visited on this trip.

Schmidt and Hubbard (1972) described several exposures along the Black River at this locality that illustrate different phases of the iron-formation as well as possible stratigraphic relationships between the Ironwood Iron-Formation and the overlying Tyler Formation. They describe thin-bedded chert-carbonate

iron-formation at the top of the Anvil Member of the Ironwood Iron-Formation on the east bank of the river 600 feet upstream from the bridge on the secondary road south of Hwy. 2. Granular jasper is described from test pits in the SE¼, SW¼, Sec. 12, T.47N., R.46W. Schmidt and Hubbard (1972) conclude that the rocks are basically unmetamorphosed examples of the iron-formation. Time limitations will not permit us to visit all of the localities described by Schmidt and Hubbard (1972), and those interested in examining unaltered iron-formation should consult their 1972 guidebook.

Schmidt and Hubbard (1972) also discussed alternative interpretations of the relationship between the Tyler and Ironwood Formations. The contact is exposed on the east bank of the Black River 180 m (600 ft.) upstream from the secondary road, where Schmidt and Hubbard (1972) interpreted the contact to be gradational.

We will examine only the exposure on the west bank of the Black River just south of Hwy. 2 where a lens of carbonate iron-formation and a sulfide unit is present within the lower part of the Tyler Formation. Although exposure is poor, there are outcrops along the stream bank consisting of highly carbonaceous slate, massive and concretionary(?) pyrite/marcasite, and sideritic iron-formation.

Stop 5. Radio Tower Hill in Wakefield.
Palms Formation of Marquette Range Supergroup

Location: The purposes of this stop are to (1) examine an excellent exposure of the Palms Formation and (2) obtain a scenic view of this part of the Gogebic range, of the Archean terrane to the south, and of the Keweenaw terrane to the north.

From the top of the hill, several abandoned iron mines can be seen in the lowland immediately to the north (refer to Figure 7); the ridges in the intermediate distance are underlain by basalt of early Keweenaw (Middle Proterozoic) age. The lowland in the far distance is underlain by the late Keweenaw Jacobsville Sandstone.

The view to the south overlooks Archean terrane consisting of the Ramsay Formation (mafic and felsic metavolcanic rocks) and the Puritan Quartz Monzonite and related rocks (~2,700 m.y.), which constitute a major batholith more than 150 km long.

In this area, the Palms Formation unconformably overlies the Archean Ramsay Formation and is overlain

conformably by the Ironwood Iron-Formation, which underlies the low area north of the hill and also southwest of the hill. Low on the next hill to the east, rare Sunday Quartzite is present. Further east, the Palms unconformably overlies the Bad River Dolomite, and has a thin phosphatic conglomerate at its base. The Palms dips steeply northward, in accord with the regional dip of Proterozoic rocks in the Gogebic range. This regional dip resulted mainly from northward tilting during Keweenawan time, toward the axis of the (Keweenawan) Lake Superior syncline. The Early Proterozoic rocks were not appreciably deformed or metamorphosed during the tectonic event (Penocean orogeny) that closed Early Proterozoic deposition, but a few folds plunging to the east-northeast at 20-25° have been recognized.

Exposures of the Palms Formation at this locality are representative of the formation which extends for 80 km to the west-southwest. Representative rock types well-exposed at the top of the hill include the upper part of the middle member which consists of interbedded argillite, siltstone, and sandstone, and the transition to the massive quartzite of the upper quartzite member. The middle member is thin bedded, with beds generally 2-10 cm thick. In general, beds of buff to pink siltstone and white sandstone alternate with buff, red or green beds of more argillaceous material; coarse-grained, dark-red, hematitic sandstone beds are also present. Most beds are quite continuous across the outcrop but "pinch and swell" irregularities impart a waviness to the generally uniform bedding. Lenticular beds of sandstone, commonly cross-bedded, are common. Flaser bedding is present locally where the mud supply was subordinate to the sand supply. Minor cut-and-fill and soft-sediment deformation structures also are present. Symmetrical but irregular ripple marks and mud cracks, although not visible at this locality, are present on the next hill to the east (Ojakangas, 1983).

Measurement of 42 cross-beds on this hill shows a strong paleocurrent trend to the west with a weaker trend to the east. Nearly 200 measurements in the formation as a whole accentuate this bimodality. Correction for plunge, if indeed the entire formation has a plunge as well as a tilt, would rotate the major and minor modes clockwise, thus relocating the modes to the west-northwest and east-southeast, respectively.

General characteristics of the formation, and especially the bimodal cross-bedding pattern, the bedding styles, and the overall lithologies are highly suggestive of deposition in a tidal environment. Recall that the lower member interpreted as upper tidal

flat, was seen at Stop 2. Here on Radio Tower Hill, the middle member is interpreted as middle tidal flat, and the cross-bedded quartz sand upper member is interpreted as lower tidal flat or subtidal (Ojakangas, 1983).

Thin sections from this locality show that the sandstone beds are feldspathic quartzite. Well-rounded unit quartz grains and feldspar (mostly fresh K-feldspar) grains are the dominant framework constituents, with chert grains a poor third. Silica cement is abundant and illitic clay is present in many samples as minor to abundant matrix. Illite is also the major constituent of the argillaceous beds with chlorite locally prominent. The coarser-grained, hematitic sandstone beds are really thin beds of iron-formation, consisting of granules of hematite, chert, and iron silicates in addition to the common quartz grains.

Deformation of the Palms Formation has not been well studied. Basically, it is part of a large monocline in which folding is rarely seen except at this locality. On the west flank of the hill, several open, round-crested folds exhibit a cleavage that is flatter than the bedding. The extent, origin, and age of these folds and their associated cleavage is not known, but the folds may have formed when the beds were approximately flat-lying, presumably during the Penokean orogeny. Alternatively, the folding might have been caused by drag along the Sunday Lake fault, a northwest-trending fault that passes just west of this hill.

Stop 6. Archean Ramsay Formation (pillowed greenstone)

Location: Roadcut along U.S. Hwy. 2 at eastern city limits of Wakefield, about 2.5 mi. SE of the junction of Highways 2 and 28 in Wakefield.

The Archean Ramsay Formation forms part of the Archean basement upon which the Early Proterozoic rocks were deposited. According to Prinz and others (1975) metavolcanic rocks of Archean age extend eastward for approximately 15 km along the range in the vicinity of Wakefield. They are intruded on the south and west by the Puritan Quartz Monzonite, and are bounded on the east by gneissic rocks of the Puritan Quartz Monzonite. They reported that the eastern two-thirds of the belt of Archean rocks is mainly mafic to intermediate flows and pyroclastic rocks with sparse felsic volcanic rocks. However, intermediate to felsic rocks are dominant in the western part of the belt. The rocks generally strike west to northwest (roughly parallel to

the strike of Early Proterozoic rocks) and dip steeply to the south. Pillows indicate that the lava flows generally face southward.

Prinz and others (1975) suggested that the mafic and the felsic volcanics each have a thickness of approximately 3,000 meters. Metamorphic grade increases from greenschist facies in the west to amphibolite facies along the eastern margin.

Rocks exposed here are typical pillowed mafic volcanic rocks that have been metamorphosed to greenschist facies. They are composed of fine-grained sodic plagioclase, quartz, epidote-zoisite, and chlorite, with some pale amphibole, carbonate and biotite (Prinz and others, 1975). Pillows with tops facing southward are well-exposed on the south side of the highway.

Take U.S. Hwy. 2 east of Wakefield city limits about two miles to Great Lakes Road. Go to compressor station on pipeline and turn left (north) on gravel road. About 0.3 mile in is a "Y"; keep left and go about 1.5 miles to end of road. Walk west on logging road $\pm$ 270 paces, then go north through the woods to the high hill. The outcrop is an escarpment along SW side of hill in the SE $\frac{1}{4}$, Sec. 18, T47N., R.44W., (Wakefield NE 7 $\frac{1}{2}$ Minute Quadrangle.)

Stop 7. Archean-Early Proterozoic unconformity.

The unconformity between the Archean Ramsay greenstone and the Early Proterozoic Sunday Quartzite is exposed along the southwest side of this hill. Note the schistosity in the otherwise massive greenstone, and the very thin basal conglomerate. Just westward from the unconformity, the Sunday Quartzite at first glance appears to be conglomeratic, but it is a thin skin along the face of the outcrop and is probably a fault breccia. This prominent exposure is probably a fault scarp.

Moving along the base of the escarpment, note the cross-bedding (some of it herringbone type, with cross-beds in successive beds oriented at 180° to each other), the numerous layers of mudchip conglomerates, and mud-cracked horizons. These are included in a series of stacked tidal channels to about 2 m thick, each starting with a mud-cracked horizon. These are overlain by a cross-bedded unit that includes herringbone cross-beds, and ends with a parallel-bedded unit. At this exposure, 41 cross-beds yield a broad trimodal (but largely bimodal-bipolar) pattern with the major mode to the northwest and a less prominent mode

to the southeast. All of these features are characteristic of a tidal environment of deposition for the Sunday Quartzite. Thus a sandy tidal environment transgressed onto a peneplained Archean craton.

Continue along escarpment and note the nature of the transition from the quartzite into dolomite. This is the Bad River Dolomite. Note also the abundance of white and black chert within the dolomite, and the presence of stromatolites. No Hammering Please! Because the vertical facies give the horizontal facies relationships, it seems that the carbonate was deposited seaward of the terrigenous sands that are now the Sunday Quartzite, and this carbonate environment then transgressed over the quartz sands as the sea transgressed northward.

Stop 8. Mafic phase, Emperor Volcanics (with contribution by K. Licht)

Location: NE $\frac{1}{4}$, NE $\frac{1}{4}$, Sec. 24, T.47N., R.43W. on Forest Service Roads 523 and then 8640 approximately 6.5 miles north of Marenisco, Michigan.

Rocks exposed on this hill illustrate several varieties of the mafic phase of the Emperor Volcanic Complex. The Emperor Formation is at least 2,000 meters thick in this area but thins rapidly to the west and is not recognized 10 miles to the west in the Wakefield area. The volcanics are truncated on the east by major faults. Therefore, the volcanic rocks form an eastward-thickening wedge interbedded with the Ironwood Iron-Formation and the wedge is truncated by faults as shown by Trent's (1973) mapping.

The Emperor Volcanic Complex consists of a mafic and felsic phase. Dann (1978) showed that the rocks range in composition from basalt to dacite. Mapping by LaBerge and J. S. Klasner in 1991 indicates that the mafic phase Emperor volcanics are the products of mainly subaqueous eruptions, and include massive and pillowed basalts, pillow breccias, hyaloclastites, and debris flows.

Exposures near the base at the north end of the hill are pillowed basalt with little or no interpillow hyaloclastite. Higher on the hill and in the cliffs along the western side of the hill, some exposures are mainly hyaloclastite, others are hyaloclastite containing pillows and pillow fragments, and some exposures are mainly pillows (Figure 14).

Figure 15 is a possible interpretation of the rocks exposed here. Lava fountaining in a shallow



Figure 14. Pillow breccia with pillow fragments in a hyaloclastite matrix. Jackknife is 8 cm long.

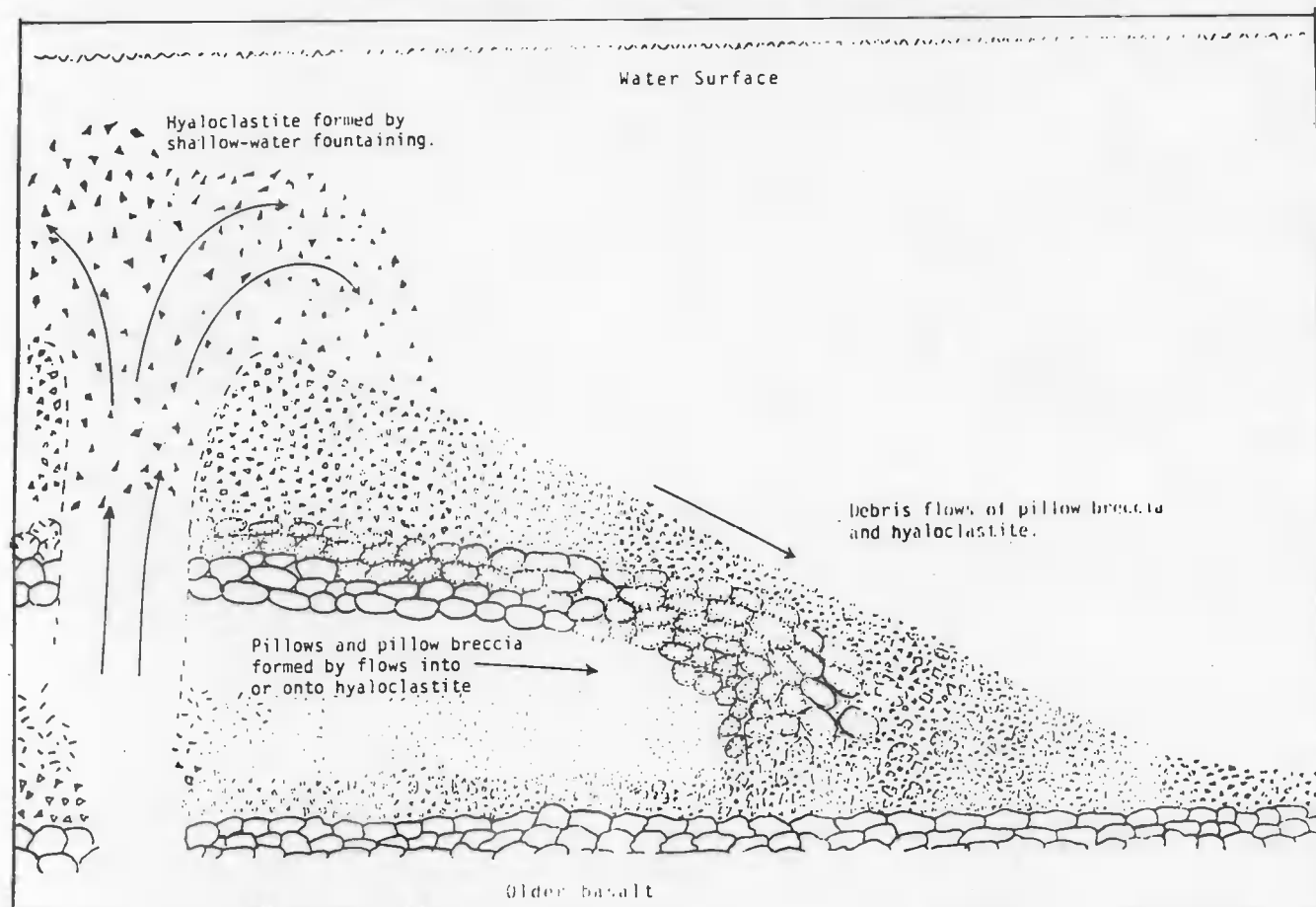
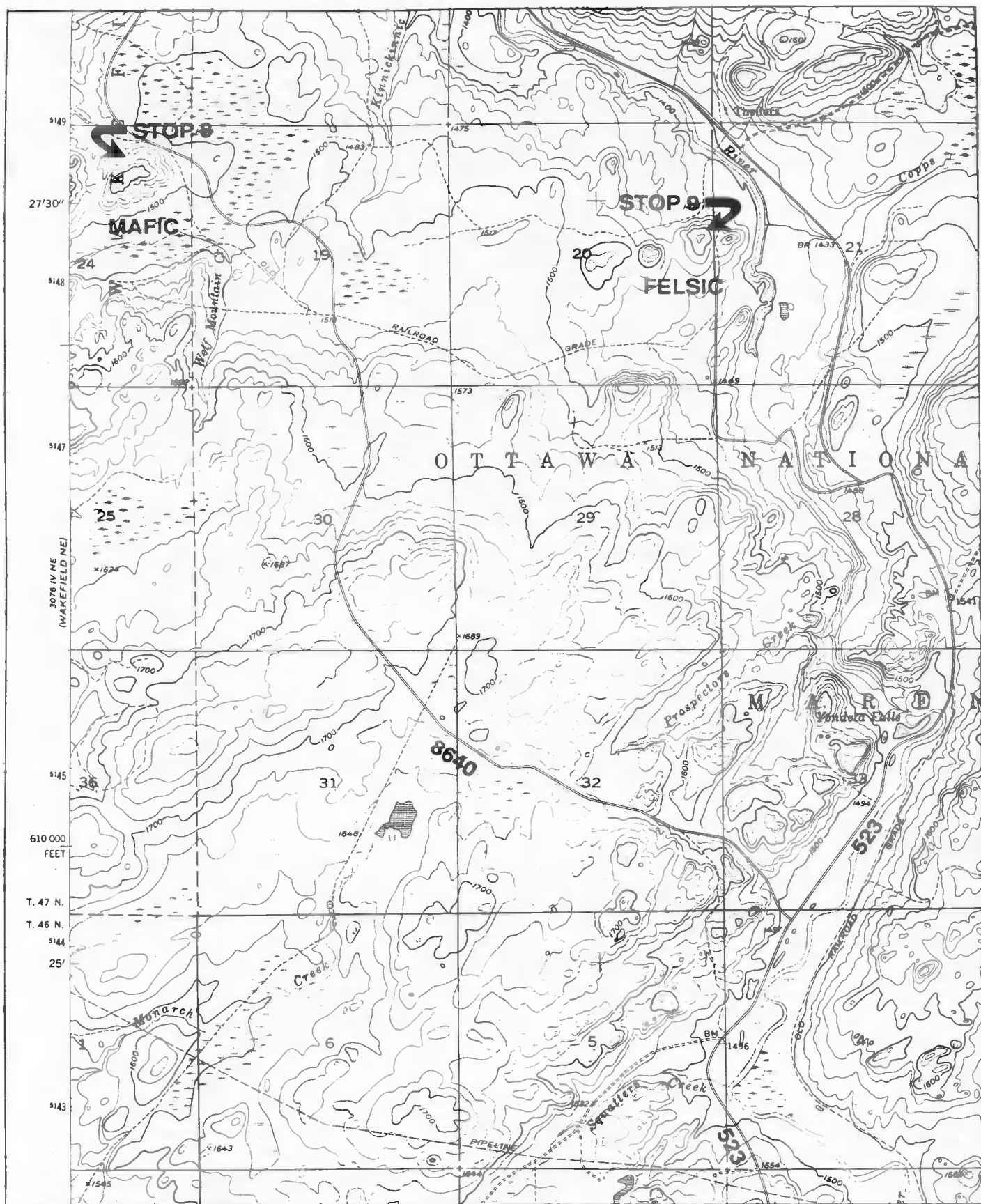


Figure 15. Sketch showing possible environment of formation of basalts exposed at Stop 8.



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subaqueous environment may have produced a hyaloclastite dome. Later eruptions may have resulted in emplacement of pillow lavas onto (or into) the hyaloclastite. Additional hyaloclastite may have formed on the surface of the flow. Pillows and pillow fragments cascading off the flow front may have become mixed with hyaloclastite. Debris flows of hyaloclastite and pillow breccia may also have been formed.

Rocks collected from this stop were analyzed in thin-section and by chemical analysis. Some textures and unique findings will be described in the following paragraphs.

Many pillows within the hyaloclastite contain reaction rims around the outer margins and along quench cracks. The thicknesses of these rims vary from a few millimeters to a few centimeters. The reaction rims are characterized by a fairly uniform highly altered basaltic texture with some remnant tabular plagioclase. Vein fillings, within the reaction rims contain chlorite, recrystallized quartz, and epidote.

Jasper occurs sporadically and tends to be "molded" between the pillows and in fractures within pillows. Source of SiO_2 for the jasper may have been gelatinous ooze on the seafloor at the time of eruption. The source of the Fe was probably the extruded basalt.

The massive basalt on the north end of the hill has an abnormally high SiO_2 content of 62.48%. In thin-section, epidote, recrystallized quartz, chlorite, and calcite are associated with the altered basalt.

The hyaloclastite varies greatly in texture, but varies little in composition. The included basalt fragments are highly altered. The hyaloclastite matrix is composed of plagioclase, calcite, quartz, chlorite, amphibole, and epidote.

Stop 9. Felsic phase, Emperor Volcanics

Location: Outcrop in SE $\frac{1}{4}$ NE $\frac{1}{4}$ Sec. 20, T47N., R.44W near Presque Isle River on Forest Service Road 523 approximately 6 miles north of Marenisco, Michigan.

This exposure contains rocks representative of some varieties of the felsic phase of the Emperor Volcanics. The dominant lithology at this locality consists of relatively angular, tan, 2 to 10 cm, monolithic clasts of microcrystalline felsite in a

finer hyaloclastite matrix. Whereas the small hyaloclastite fragments typically have a whitish alteration rind, the larger felsite clasts generally lack alteration rinds. Prinz and others (1975) stated that the felsite clasts as well as the matrix have a dacitic composition. In zones of deformation the hyaloclastite matrix tends to be strongly foliated, whereas the felsite clasts tend to show little deformation (Figure 16). Some clasts contain abundant perlitic cracks, others are more massive felsite. This lithology of the felsic phase of the Emperor extends westward at least 5 km, where there are extensive exposures in Secs. 14 and 23 north of Wolf Mountain. In addition to the variety of rock exposed here, Sec. 14 north of Wolf Mountain contains extensive brecciated felsite with scattered quartz phenocrysts. Although the rock is extensively brecciated, there appears to have been little or no movement of adjacent blocks with respect to one another. In short, the rock appears to be a felsic dome that has been brecciated.

Surrounding the brecciated felsite are numerous exposures with breccia fragments of felsite in a felsic hyaloclastite matrix (Figure 17). It is suggested that the felsite breccia may be a felsic dome emplaced into felsic hyaloclastite produced by an earlier eruption. Debris flows off the dome may have resulted in more widespread units of the felsite breccia/hyaloclastite as shown in Figure 18.



Figure 16. Photo of sawed slab of Emperor Volcanics from Stop 9. Larger, angular felsite clasts in a finer grained, foliated hyaloclastite matrix. Slab is 18 cm long.

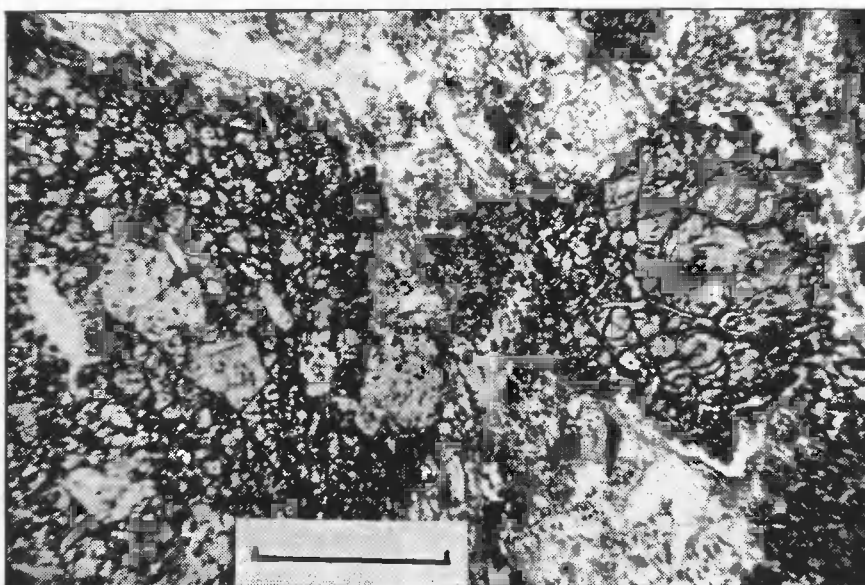


Figure 17. Photomicrograph of hyaloclastite from the felsic phase of the Emperor Volcanics. Note the perlitic cracks in the oval clast on the right side of photo. Bar scale is 2 mm.

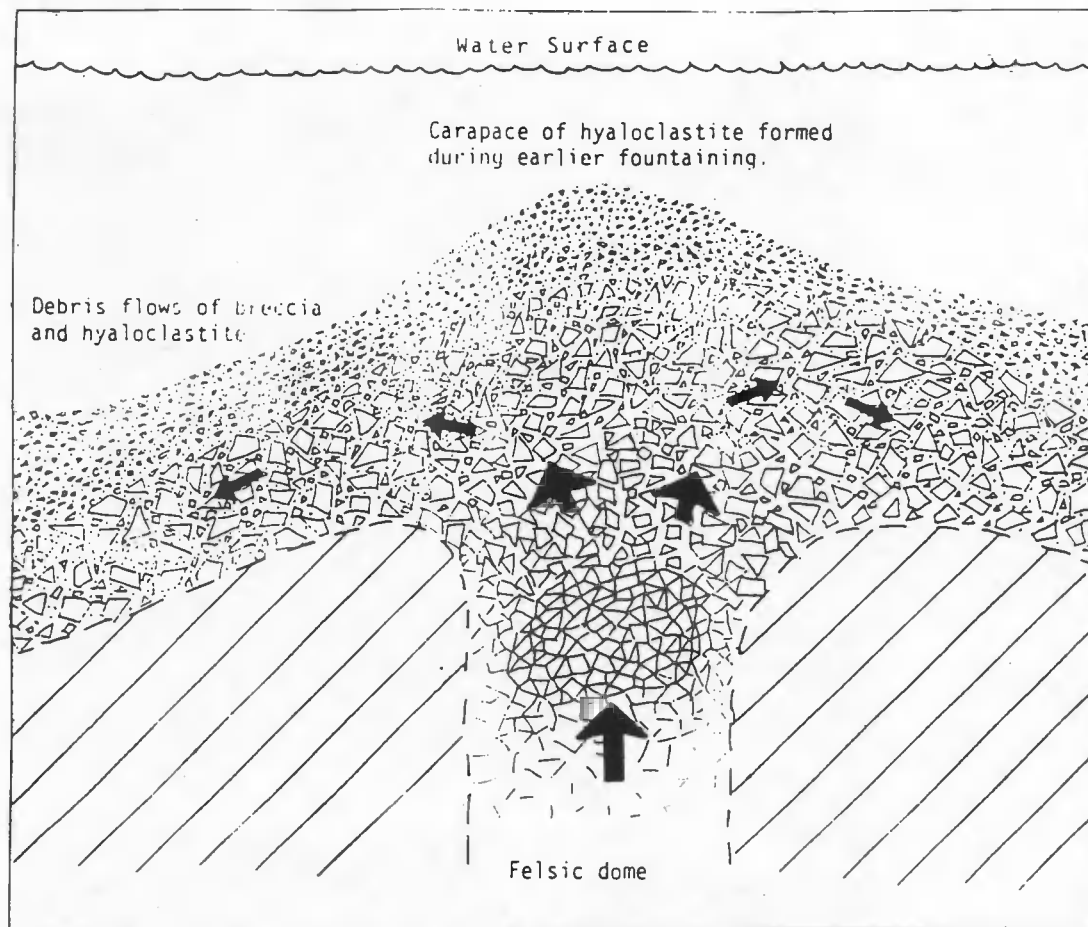


Figure 18. Sketch of possible relationships within the felsic phase of the Emperor Volcanics. Highly fractured felsite blocks become dispersed in finer grained felsic hyaloclastite in debris flows off a subaqueous lava dome.

Stop 10. Archean metagraywacke

Location: Roadcuts on Highway 2, about 10.5 mi SE of Marenisco, SW $\frac{1}{4}$, Sec. 30, T. 46 N., R. 41 W. (Thayer, MI 15' quadrangle).

The metagraywacke in these roadcuts is part of a northeast-trending belt 4 km wide and 20 km long. The rocks are highly folded on tight north-northeast-trending axes, and the unit is probably about 1 km thick. The graywacke is a fine-grained, thin- to thick-bedded rock with a pervasive northeast-trending schistosity (S_2) that is Penokean in age; an older schistosity (S_1) that is Archean in age is nearly parallel to bedding. Graded beds are common, and tops at this outcrop are to the south. Reversals of grading indicate that the fold crests are about 200 m apart.

The rocks were metamorphosed to amphibolite grade during the Archean and are totally recrystallized; most of the garnet was retrograded to chlorite + muscovite + epidote and opaques during the Penokean orogeny (Sims and others, 1984). The nature of the original sediment is unknown because of the thorough recrystallization, but CIA indices (chemical indices of alteration) calculated for the graywackes indicate an original immature detritus that had undergone little if any weathering in the source area and could have been immature volcanoclastics. The area was mapped by Fritts (1969), Trent (1973), and Prinz (1981). Cannon (1986b) produced a structural and tectonic map of the region.

Stop 11. Watersmeet Gneiss (3560 Ma)

Location: Roadcuts on Highway 45, about 4 mi N of the junction of Highways 2 and 45. This junction is on the S edge of the village of Watersmeet. Roadcut location is S edge, NE $\frac{1}{4}$, Sec. 4, T. 46 N., R. 39 W. (Watersmeet, MI, 15' quadrangle).

This exposure is located in the northern part of the Watersmeet dome which has an Archean core about 25 km long in an E-W direction and about 8 km wide in a N-S direction. The dome is considered to be a mantled gneiss dome. Sims (1990) recognizes considerable Early Proterozoic rocks in the core infolded with the Archean. Deformation of the Archean and adjacent Early Proterozoic rocks occurred during the Penokean orogeny. The rocks in the core were metamorphosed to the amphibolite facies, a 15 km-wide zone of epidote amphibolite facies surrounds the core, and this in turn passes outward into greenschist facies. (This is the Watersmeet metamorphic node of James, 1955). The Early Proterozoic rocks to the north of the dome have been overturned toward the northwest. This area was studied by Sims and others (1984) and Peterman and others (1980).

The gray gneiss is a tonalitic augen gneiss about 3560 Ma old (zircon data) (Sims and others, 1984); it is a medium-gray, medium- to coarse-grained, biotite-rich, irregularly layered rock with plagioclase augen. The alternating layers are feldspar-quartz-rich and biotite-rich. A modal analysis showed 48% plagioclase, 30% quartz, 8% K-feldspar, and 13% biotite. A possible protolith was dacitic volcanic rock (Sims and others, 1984). In this exposure, biotite leucogranite veins 2590 Ma old (Rb/Sr and U-Pb methods) cut the gneiss. Secondary whole-rock and mineral isochrons give ages of 1750 to 1800 Ma and are thought to approximate the time span of the Penokean event in this area (Sims and

others, 1984).

Whereas the Archean gneisses of the area may have been folded three times, the cross-cutting Late Archean leucogranite shows only a single foliation, that of the Penokean event.

The gneiss is comparable in age to those of the Minnesota River Valley, about 3500 Ma old: these are parts of the ancient gneiss terrane that is separated from the granite-greenstone terrane to the north by the GLTZ, the Great Lakes Tectonic Zone of Sims and others (1980).

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**KEWEENAWAN
SEDIMENTARY ROCK
OF THE
SOUTH SHORE,
LAKE SUPERIOR**

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INTRODUCTION

The sedimentary and igneous rock composing the Precambrian (Middle Proterozoic, Keweenawan) section of the Lake Superior stratigraphic column have had a long history of study. Almost six decades ago (1933), the geology of the "Lake Superior Region," constituting Excursion C-4, was presented as Guidebook 27 of the XVI International Geological Congress. The field guide for this excursion was prepared by W. O. Hotchkiss of the Michigan College of Mining and Technology (now Michigan Technological University) and was purchased for the sum of 25 cents.

In his introduction to that guidebook, C. K. Leith (University of Wisconsin) stated:

The Lake Superior region has been of special interest to students of pre-Cambrian geology because it presents the largest and most varied pre-Cambrian succession that has been definitely worked out. Its content of valuable iron and copper ores has made possible more intensive and detailed studies than have been accorded to extensive pre-Cambrian areas elsewhere. The pre-Cambrian succession now known (in the Lake Superior area) represents a greater thickness of sediments and a larger time than all the post-Cambrian of North America.

It is interesting to note how Charles Leith, in a later section of this guidebook, presented the Keweenawan sequence, now known as the Midcontinent Rift sequence:

Keweenawan: Next below (the Cambrian) is the nonfossiliferous Keweenawan series, consisting of an immense mass, possibly 5 miles (8 km) thick, of sandstone, with intercalated shales and conglomerates, containing in its lower part large quantities of extrusive lavas and intrusive laccoliths and sills. In degree of metamorphism it is more like the Cambrian than the underlying Huronian series. It has characteristic reddish, yellowish, and purplish colors and carries various evidences that it was essentially a continental deposit under semiarid conditions. Its lower part is tilted in marked unconformity to the Cambrian, but its upper part lies nearly, if not quite, parallel to the Cambrian. Obviously, it was mainly deposited in an independent basin before the incursion of the Upper Cambrian sea. Although the Keweenawan is pre-Cambrian in the sense of preceding the Upper Cambrian transgression, having structural and igneous affiliations with the pre-Cambrian, and being nonfossiliferous, it may be Cambrian in the sense that it was being formed at the same time as Middle and Lower Cambrian sediments in distant Cambrian seas.

Today as we gather for this 38th Institute on Lake Superior Geology fieldtrip, comparisons with conditions in 1933 form distinct contrasts. In 1992, with a singular exception, the copper mines are closed and the massive iron ore fields have fallen under different economic climates. Yet, a new era of evaluation and exploration has begun. The Lake Superior Keweenawan section is widely recognized as one of the finest extant examples of continental rifting. The advent of industrial field evaluation of the hydrocarbon potential of these rift strata, beginning in 1983, has directly and indirectly facilitated the collection of approximately 4200 km of seismic reflection data in the Lake Superior district alone, in addition to supportive magnetic and gravity surveys.

This new period of exploration within the Lake Superior region has altered our interpretation of its geologic development, especially during Keweenawan time. Many of these updated geologic

relationships will be examined during the process of this field trip. Our principal focus will be on the geology and geologic history of the youngest Precambrian rock in the region, the Keweenawan. This rock is a remnant of major continental rifting, initiated circa 1100 Ma. The resultant Midcontinent Rift System trends from the Lake Superior region to as far southwest as central Kansas and as far southeast as southern Ohio.

As of this writing (March 2, 1992) site preparation has been underway for several weeks for the Terra/Patrick #7-22 wildcat borehole, a test of the hydrocarbon potential of the Wisconsin section of the Midcontinent rift of national significance and interest. Located in 22, 47N, 6W, Keystone Township, Bayfield County, Wisconsin, this well will be operated by Terra Energy, Ltd of Traverse City, Michigan, in partnership with Patrick Petroleum Company of Jackson, Michigan, under farmout arrangements with Amoco Production Company (USA), Houston, Texas. The target depth will be 6,000 feet, permitting a test of the oil or gas potential of the Oronto Group sequence of sedimentary rocks. The structure is a large drag-fold anticline with north closure against the Douglas reverse fault. The surface expression of this structure extends over an area of approximately 60 square miles. Available information on this state record depth wildcat will be included in the 38th Institute on Lake Superior Geology Proceedings as an abstract (Albert B. Dickas).

The site of the Terra/Patrick #7-22 borehole is approximately 6 miles north-northeast of an Oronto Group wildcat test borehole announced in 1985 by Amoco Production Company (USA). This borehole, planned for 13, 46N, 7W, and targeted for 15,000 feet, was never drilled for economic and legislative reasons.

GEOLOGY OF THE MIDCONTINENT RIFT SYSTEM ALONG THE SOUTH SHORE OF LAKE SUPERIOR, WISCONSIN AND MICHIGAN

Reprinted with permission from Dickas, Albert B. (editor), 1989, Lake Superior basin segment of the Midcontinent Rift System: Field trip guidebook T344, 28th International Geological Congress, 62 pages.

Introduction

The Midcontinent Rift System has been mapped over an axial length of in excess of 3,300 km. Portions of this Precambrian (ca. 1.1 Ga) extensional structure are contained in the states of Kansas, Nebraska, Iowa, Minnesota, Wisconsin, Michigan, and Ohio. This report presents the geology of this rift system within the geographic region generally defined as northwestern Wisconsin and the adjacent upper peninsula of Michigan.

Wisconsin Section

The Midcontinent Rift sequence of Precambrian sedimentary strata in the western Lake Superior region constitutes some of the oldest known and studied such assemblages in the United States. These units were initially described by Hunt (1873), who named them the Keweenaw Group after the cupriferous strata exposed along the Keweenaw Peninsula of Michigan. Thwaites (1912) divided this section into the Oronto (lower) and Bayfield (upper) groups (Fig. 1). More recently the U. S. and Ontario Geological surveys have been informally using the term Keweenawan Supergroup.

The outcrop belt of the Oronto Group is traced from northwestern Wisconsin east and northeast to the tip of Keweenaw Point, Michigan. The Oronto Group crops out principally along the Ashland syncline (Fig. 2), which is centrally located on the St. Croix horst. The Ashland syncline is asymmetrical as the steeper dips are found within the southern limb. The horst is bounded by the reverse-throw Douglas and Isle Royale faults to the north and the Lake Owen and Keweenaw faults to the south. While seismic information suggests the presence of Oronto components in the flanking basins, deep drilling has yet to confirm this conjecture. However, on the basis of a drag fold exposure of Oronto type strata (Freda Sandstone) in Douglas County, Wisconsin, White (1966) supported the concept of Oronto strata being found north of the Douglas fault. Alternately, White (1966) supported the possibility of post-Oronto (Bayfield Group) inliers being found within the horst position as suggested by heavy mineral analyses conducted by Tyler et al. (1940).

The Oronto Group is 40 percent sandstone and 60 percent shale (Thiel, 1956) and constitutes a maximum 6,100 m thick sequence of reddish conglomerate, sandstone, and mudstone. It shows a general upward increase in maturity, from poorly sorted arkosic to quartzose, relatively well-sorted sedimentary rocks. Much of this group was locally derived through erosion of Middle Keweenawan basalts and transported only a short distance to its depositional basin.

MID-CONTINENT RIFT	1 - Flank
TIME STRATIGRAPHY	2 - Horst
	3 - U. Peninsula
	4 - L. Peninsula

FIGURE 1: Stratigraphic terminology and time relationships of Upper Keweenawan sedimentary rocks along the Midcontinent Rift System trend. After Dickas (1986), reprinted by permission of the American Association of Petroleum Geologists.

In conformable and interfingering contact with the Copper Harbor Conglomerate is the Nonesuch Formation, a hydrocarbon source rock (Dickas, 1991) and the center of industrial interest focused on the Midcontinent rift since the early 1980's. These strata are gray to black, easily distinguishable from enveloping red to brown formations. The grain size ranges from medium sand to clay with shale subordinate to siltstone and sandstone. This unit is the thinnest of the Oronto Group, ranging from at least 140 m in northern Wisconsin to a maximum 200 m in adjacent Michigan. Framework composition emphasizes mafic rock fragments, with analysis showing ranges of 15 to 30 percent (Hite, 1968). Of economic interest is the cupriferous nature of the basal 10 m of the Nonesuch Formation, especially in the White Pine, Michigan area where an estimated 11 million tones of copper have been found (Cox et al., 1973), predominantly in the form of chalcocite and native copper. The variety of sedimentary structures found in this formation have locally been employed as supporting evidences for a variety of depositional environments. The best model appears to be that of reducing conditions within a standing body of water (Hubbard, 1975), distinguished by frequent variations in water depths (Daniels, 1982), influenced by fluvial and deltaic associations (Barghoorn et al., 1965), and marked by periodic salinity levels exceeding that of gypsum precipitation. Recently, however, Hieshima and Pratt (1991) offer evidence supporting a marine embayment environment.

The copper is found in direct lithologic association with finely disseminated carbonaceous matter which averages 0.5% of the formation by volume. Kelly and Nishioka (1985) believe copper precipitation was directly facilitated by the intimate presence of organic material. The organic matter, upon chemical extraction, shows remarkable structural features that have been preserved by a bituminization, rather than a carbonization, process. This material is found both as matrix masses and as recognizable remains, such as bacteria cells, algal-like units and fungal hyphae, with volume distribution being concentrated in the finest grained strata. Overall, individual laminae are alternately rich and poor in organic components. Moore et al. (1969), after thin-section analysis of the organic rich shale layers, concluded these units are characteristic of lake, swamp and tidal flat environments and that the majority of preserved organic material developed in place.

The cupriferous zone is also of interest because, at the White Pine mine, associated vugular porosity and fractures often are filled with small amounts of solid and liquid hydrocarbons. Analysis by Barghoorn et al. (1965) indicates this Nonesuch crude oil contains higher concentrations of alkanes and lower percentages of aromatics than an average crude oil, and resembles the paraffinic crude commonly associated with Paleozoic production in the state of Pennsylvania. Further analysis has indicated the oil is indigenous to its host rock (Eglinton et al., 1964), is of organic origin (Moore et al., 1969), and was formed under mild formational temperature conditions (Barghoorn et al., 1965). A low thermal environment is also supported by Brown (1971), who reported the presence of pink bornite and djurleite in the Nonesuch Formation, minerals which cannot exist above respective ranges of 75 degree C and 95 degree C (167 F and 203 F).

The Nonesuch crude presently represents one of the oldest known liquid hydrocarbon, being approximately 1,046 million years in age. This was determined for the Nonesuch host by using rubidium-strontium ratios. This date is verified by the work of Ruiz et al. (1984), who obtained an age of 1,047 +/-35 million years by Rb-Sr dating of calcite filled veins cutting

across the cupriferous shale of the Nonesuch Formation in the White Pine area. As this calcite contains primary liquid oil inclusions, Kelly and Nishioka (1985) refer to this date as the time of oil entrapment and thus consider it as a minimum age for White Pine hydrocarbon.

The upper unit of the Oronto Group, the Freda Sandstone, lies conformably over the Nonesuch Formation. The nature of the Freda Sandstone upper contact is not known for nowhere is this formation found in exposed (Myers, 1971) or drilled contact with any unit of the Bayfield Group or its Michigan correlative, the Jacobsville Sandstone (Fig. 1). While listed by Hite (1968) as exceeding 3,660 m in thickness, only 1,500 m is found on the north coast of the Keweenaw Peninsula (Halls, 1966), near the type section at Freda, Michigan. The Freda Sandstone is composed of basal, subordinate conglomerate fining-upward to sandstone and siltstone. In color, a repetition of the Copper Harbor red to brown spectrum is observed, interrupted by spotty and laminar unoxidized zones (Daniels, 1982). With an analyzed maturity index of 0.63, a feldspar range of 10 to 24 percent (Hite, 1968), and the grain shape being angular to sub-rounded, these sandstone units are classed as immature. Daniels (1982) considers the average sand unit to be a fine-grained feldspathic lithic arenite. Sedimentary structures include cross-bedding, mud-cracks, graded beds, ripple marks, and a wide variation in paleocurrent indicators. The implied depositional environment ranges from fluvial (Daniels, 1982) to fluvial plain or tidal flat (Hamblin, 1961).

Along the eastern shore of Lake Superior, in the Pt. Mamainse region of Ontario, a 60 m section of gray sandstone is found unconformably overlying Keweenawan basalts. Work by Hamblin (1961) and Annels (1973) allows correlation of these strata, termed the Mica Bay sandstone, to the Freda sandstone and suggests a fluvial environment of deposition.

Overall the Oronto Group can be viewed as a thick sequence of clastic sedimentary rocks that are stratigraphically unified by similar heavy mineral suites, an upward and distal decreasing in grain size, and an upward increase in lithologic maturity. The three formations of this group were derived from a common source of Keweenawan igneous material (Tyler et al., 1940). While each formation is interpreted in light of slightly different depositional environments, Elmore and Daniels (1980) view the group environment as a transgressive-regressive alluvial fan, lacustrine and fluvial system that filled the developing Midcontinent Rift basin during the final stages, and after cessation, of volcanic activities.

The Bayfield Group, wherever studied, is found to be in distinct contrast to the Oronto Group. The Bayfield sequence displays a higher degree of compositional maturity and structurally the bedding is more commonly subhorizontal. In terms of heavy minerals the tourmaline-zircon to epidote ratio is reversed in the Bayfield Group as compared to the Freda and Nonesuch assemblages (Tyler et al., 1940). There is a question as to the type of contact between these groups. Supporting an unconformable contact is the reported change in paleomagnetic pole position (Du Bois, 1962), the change in accessory minerals (Tyler et al., 1940) and the immediate change from steep Oronto dips to low Bayfield dips (Morey and Ojakangas, 1982). In support of a conformable contact, or one of minimal hiatus, is the lack of clasts of Oronto age in the basal Orienta sandstone (Thwaites, 1912), and the nature of the previously mentioned Freda Sandstone drag fold exposure in Douglas County, Wisconsin (Tyler et al., 1940). From a tectonic point of view, Morey and Ojakangas (1982) state the

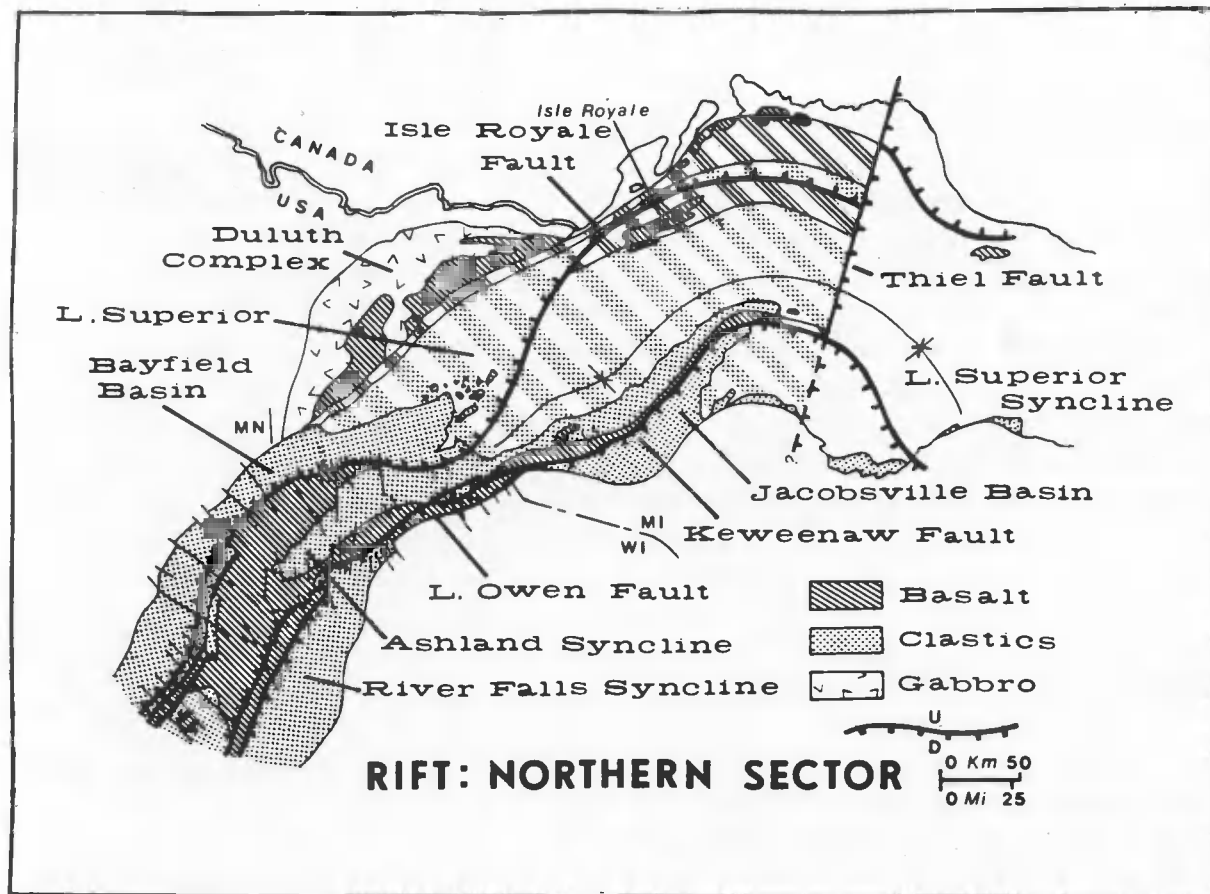


FIGURE 2: Midcontinent Rift System trend and geology from northwestern Minnesota into the Upper Peninsula of Michigan. Projection of Keweenawan geology under western Lake Superior is shown by hachured patterns. From Dickas (1986), reprinted by permission of the American Association of Petroleum Geologists.

change from Oronto to Bayfield sedimentation marks a transition from an extensional regime to one characterized by vertical tectonic processes, i.e., the activation of the St. Croix horst by uplifting of 4,600 m (Thiel, 1956).

The type section for this group is along the southwestern shore of Lake Superior from the Apostle Islands to just east of Superior, Wisconsin. Reported group thicknesses range from a minimal 815 m in outcrop (Craddock, 1972), to a maximum 2,100 m in the subsurface as suggested by seismic review (Mooney et al., 1970).

Because of the relative uniformity in composition, the three Bayfield sandstones can be described as a unit. Thiel (1956) considered the overall composition to be 99% sandstone and 1% shale. Petrographic studied by Myers (1971) showed that quartz constitutes approximately 80% of framework grains. The Orienta and Chequamegon sandstones and siltstones are commonly red in color and feldspathic in mineralogy and contain layers of shale and conglomerate. The Devil's Island sandstone is buff to white in color and so quartzose as to be classified as an almost pure orthoquartzite. An ascending fluvial-lacustrine-fluvial environment of deposition is documented by respectively, trough cross-bedding, ripple marks, and a return to cross-bedding in the Chequamegon Sandstone (Morey and Ojakangas, 1982). Myers (1971), states the

Bayfield Group sedimentary rocks were derived mainly from older Keweenaw sediments and probably represent the reworking of Oronto Group units. Ojakangas and Morey (1982), however, believe the Bayfield Group clastics were derived from an older granitic terrane.

West of the Thiel fault (Fig. 2), the basic structure of Lake Superior is a continuation of northwestern Wisconsin Midcontinent Rift geology. The Ashland syncline is projected offshore as the Lake Superior syncline. Evidence is lent by northwest dips of Oronto Group sedimentary rocks along the Keweenaw Peninsula (Daniels, 1982) and the southeast dips in sedimentary rock of the same age which outcrop on the southwest shore of Isle Royale (Wolff and Huber, 1973). This major syncline was later disrupted by reverse faulting, forming the offshore segment of the St. Croix horst (Fig. 2). Both the Bayfield basin and the St. Croix horst offshore extension are considered to be floored by Middle Keweenaw clastics of the Oronto and/or Bayfield groups (Green, 1982). Thus, the western portion of Lake Superior is generally portrayed to be immediately underlain by Keweenaw Supergroup clastics up to 10 km in thickness (Hinze et al., 1982).

Michigan Section

Keweenaw-age outcrops form the entirety of the Keweenaw Peninsula jutting into south-central Lake Superior (Fig. 2). The keel of this peninsula is formed of Portage Lake volcanics, upon which all Upper Keweenaw sedimentary rocks in the area rest (Fig. 1). Those outcrops north and northwest of the Portage Lake Volcanics form the classic Oronto Group. South and southeast of the Portage Lake volcanic outcrop belt and in contact with the Keweenaw fault is found a thick sequence of northerly dipping clastics named the Jacobsville Sandstone by Lane and Seaman (1907) from outcrops near Jacobsville, located near the southeast base of the Keweenaw Peninsula. By geologic position these outcrops appear to be an easterly extension of the River Falls syncline. However, because the River Falls syncline and the Jacobsville sandstone outcrops are not in direct contact due to the exposure of intervening Archean granitoid rocks, this area of Jacobsville sandstone deposition is herein referred to as the Jacobsville basin (Fig. 2).

In this basin the Jacobsville sandstone has a maximum thickness of 867 m by drill hole measurement (Kalliokoski, 1982), but on the Keweenaw Peninsula proper its maximum thickness has been determined by geophysics to be approximately 3,000 m. The formation is predominately quartzose and is composed of conglomerates, sandstones, siltstones, and shales, generally found in repetitive upward-fining sequences. The conglomerates are generally basal, up to 100 m in thickness and contain clasts composed of iron formation, vein quartz, volcanics, and metamorphic rocks. The sandstones are fine to coarse grained, light reddish to purple to mottled cream-white in color and contain structures ranging from cross-bedding to oscillation and current ripple-marking. Shales and mudstones appear to be in the minority in terms of thickness and are found interbedded with laminated sands. The overall section varies from sub-arkose to a quartz arenite, with plagioclase found more commonly in the younger layers (Kalliokoski, 1982).

The Jacobsville sandstone was derived from highlands situated to the southwest and southeast of the depositional basin. The determined environments of deposition include fluvial,

lacustrine, and alluvial fans. Because the Jacobsville sandstone is nowhere found in direct contact with either the Oronto or Bayfield Group formations, its exact relationship within the Keweenawan Supergroup is in some doubt. As it overlies Portage Lake volcanics and underlies the Late Cambrian Munising sandstone there seems little doubt regarding its Keweenawan age. Considering that Tyler et al. (1940) noted a strong similarity between the Jacobsville Sandstone and the Orienta sandstone heavy mineral suites and that Kalliokoski (1982) related the Jacobsville Sandstone to the Fond du Lac Formation of Minnesota on the basis of sediment, mineral, and clay matrix comparison, most workers today consider the Jacobsville Sandstone to be a Michigan equivalent to the Bayfield Group of neighboring Wisconsin (Fig. 1).

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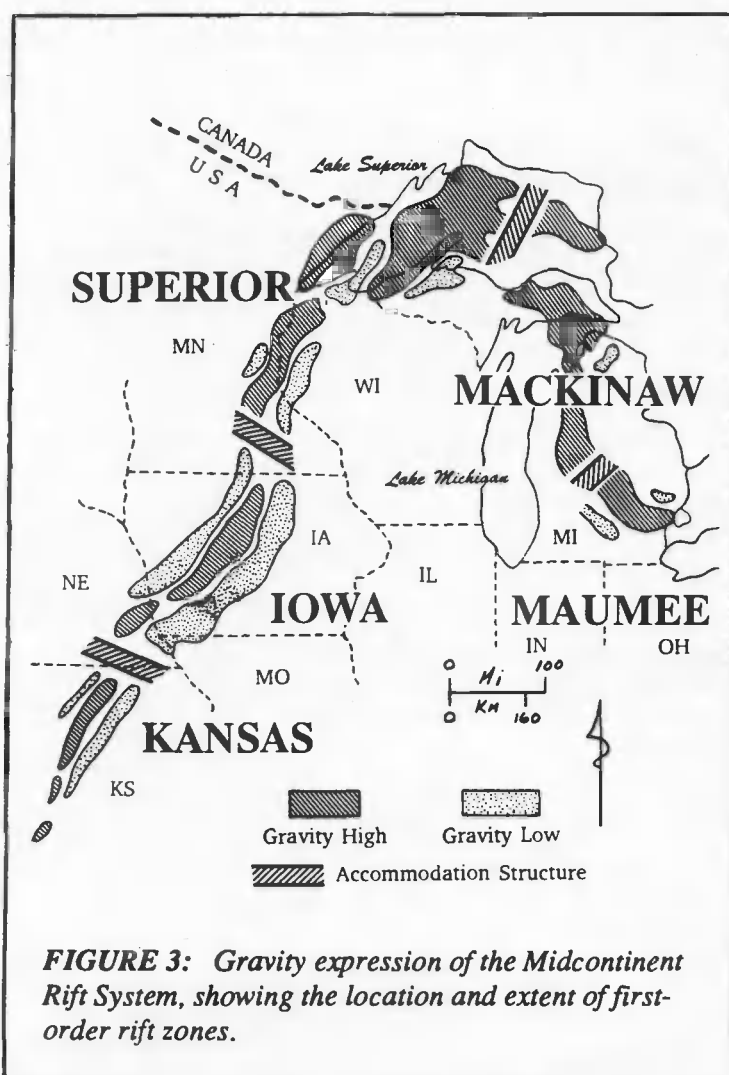
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CENTRAL NORTH AMERICAN CASE FOR SEGMENTED RIFT DEVELOPMENT

Reprinted with permission from Dickas, A. B. and M. G. Mudrey, Jr, 1989, Abstracts, 28th International Geological Congress, vol. 1, p. 1-396 to 1-397.

The Midcontinent Rift System (MRS) is a major Middle Proterozoic intracontinental, thermotectonic structure that has been traced by regional gravity and magnetic data, subsurface drilling, and outcrop control over a length of 3,300 km in the central United States. In the Lake Superior region, the only area in which MRS outcrop is known, this structure was infilled by plateau lavas and sedimentary rock comprising the Keweenaw Supergroup. About 1140 Ma, basalt was extruded along the rift. After cessation of volcanic activity, generally maturing upward, clastic, sedimentary rock was deposited. Geophysical analysis in the 1950's (Thiel, 1956) suggested a broad graben with a central horst separating wedge-shaped thick accumulations of low-density sedimentary rocks. These sedimentary sections were identified by negative gravity anomalies. Since then, models and geologic maps of the rift have been consistent in demonstrating an early developed symmetrical extensional crustal basin filled with extrusive and volcanoclastic rock, subsequently modified by late-phase compressional faulting forming the presently identified central horst (Fig. 2, previous paper).



Recent acquisition of over 4200 km of onshore and offshore reflection seismic data and analysis of deep mineral exploration drill holes from northern Wisconsin suggest a different geometry for the MRS. This geometry, as proposed here, is similar to structures attributed to the Gregory rift in East Africa. Bosworth et al. (1986) suggest the Gregory rift structure formed through the development of deep, crustal, opposing detachment systems. As these detachments intersect at depth, one side locks, resulting in half-graben development. With lateral rift propagation, a subregional series of taphrogeosynclines (half-graben) are formed, with adjacent grabens displaying igneous and sedimentary infill packages of opposing geometry. The initial rift is thus unable to extend laterally in uniform structural patterns but rather forms a linear pattern of individualized, en echelon, asymmetric basins, each

with its own infilling history and geometry and each separated by a type of cross-structure, generally referenced as "accommodation" faulting.

On a regional basis, the final structural configuration of a rift may also be dependent upon total extension developed at right angles to rift propagation. Schuepbach and Vail (1980) suggest that with such differential rift development, and even though structural extension along rift strike may be synchronous, basin formation and filling could be diachronous. Quennell (1984) supported this evolution by mapping the Dead Sea rift as divided into three unique segments, each operating independently in geologic development. Viewed as an entirety, a rift may thus display first-order, regional-size rift segments (Dickas, 1986), representing coeval stages of differing rift maturity. In turn, these segments might be divided into subregional-sized, second-order sub-basins separated by "accommodation" faulting.

Viewed from the perspective of this geometry, the extent of the MRS is here interpreted to be composed of five first-order rift segments, termed "zones" by Rosendahl (1987). Clock-wise from the southwestern end of the MRS, these zones are here named Kansas, Iowa, Superior, Mackinaw, and Maumee (Fig. 3), and are identified on the basis of major interruptions in gravity and magnetic patterns, seismic review, terrane compositions, and borehole analysis.

Locally, and developing as a function of overall rift extension, zones may be further divided into a series of structurally independent basins. These second-order sub-basins, termed "units" by Rosendahl (1987), are separated by "accommodation" faulting. Within the Superior zone of the MRS, four units are now recognized on the basis of seismic interpretation, interrupted geopotential trends, and core analysis. These sub-basins are here named the Chisago, Brule, Ontonagon, and Manitou units (Fig. 4). These units appear to have undergone a similar history of structural development during the early phase of MRS evolution. On the basis of differential listric movements along axial-oriented, rift fault systems, igneous and sedimentation wedges of alternating isopach patterns distinguish juxtaposed zones. Isopach patterns of these wedge geometries differ by being symmetrical parallel, but alternating asymmetrical perpendicular, to the MRS axis.

In light of this suggested model, future stratigraphic and seismic correlations can no longer be interpreted within the framework of a symmetric rift model existing along the entirety of the MRS axis. Instead, such correlations must be dependent upon consideration of isolated igneous and sedimentary packages and isopach geometries created in response to first- and second-order degree of tectonic development. Exploration philosophies, whether they be directed to base-metal or hydrocarbon evaluation, must be constrained so as to recognize related regional and subregional alterations in structural style.

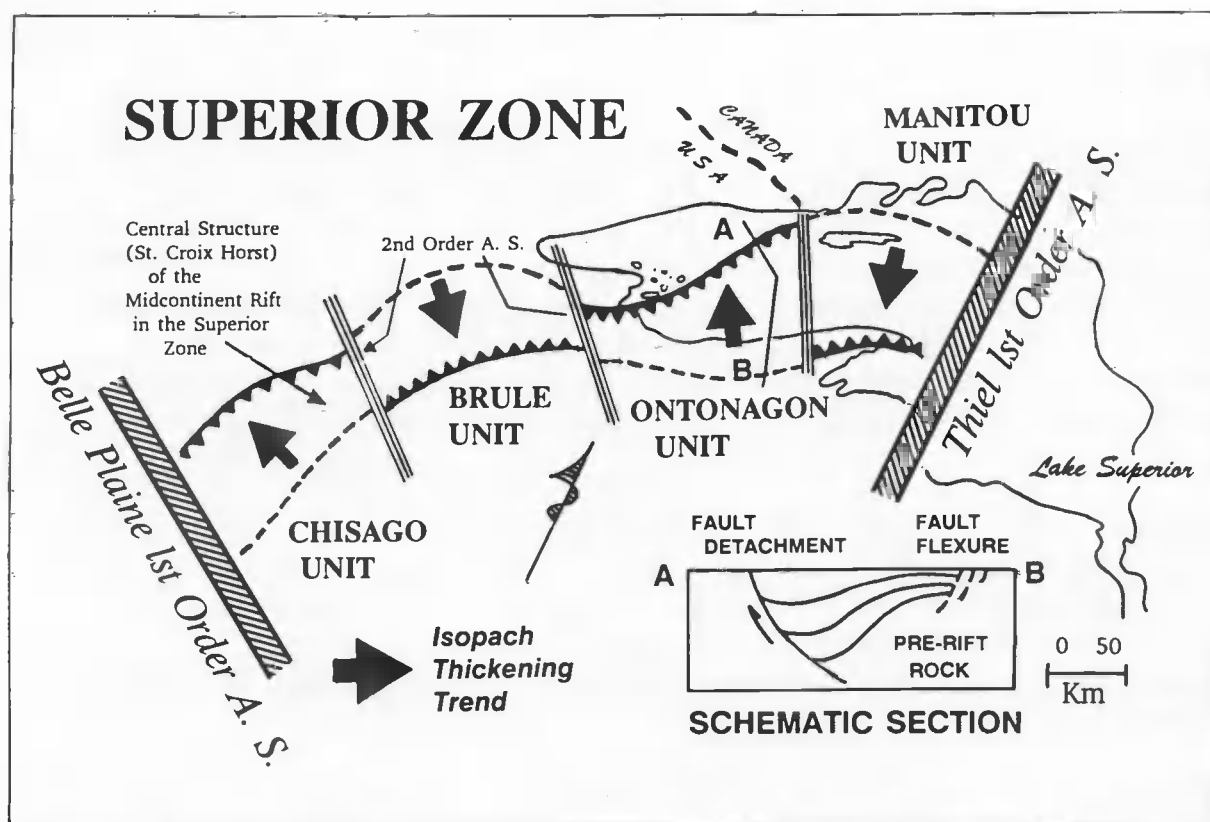


FIGURE 4: Structural rift units composing the Superior rift zone, Midcontinent Rift System, as defined by accommodation structures (A.S.) and isopach thickening trend,

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LAKE SUPERIOR BASIN AS A HYDROCARBON FRONTIER

Modified from a series of requested articles written by Albert Dickas for the daily print media of northwest Wisconsin, January, 1992, during site preparations for the Terra/Patrick #7-22, Bayfield County, Wisconsin exploratory wildcat.

Since the initial discovery of crude oil at a depth of 21.2 m beneath the rolling hills of Titusville, Pennsylvania in 1859, hundreds of thousands of boreholes have sought this natural resource in the United States. Wisconsin has played a small role in this 133 year old "hydrocarbon revolution." Between 1865 and 1992, 49 boreholes have sought economic volumes of oil or gas in the Badger state. Not one of these wells however, has found anything other than fresh and salt water.

The majority of exploration in Wisconsin has taken place close to the Lake Michigan shoreline between Milwaukee and Door County (Fig. 5). Here the rocks form the western edge of the Michigan basin, one of the most prolific producers of hydrocarbon in the United States. Unfortunately, this productivity is basically confined to the deeper portions of the basin, and not along its edges. The remaining area of Wisconsin has long been considered non-productive because of the great geologic age and physical nature of much of its bedrock.

In the early 1980s, pessimism turned to guarded optimism. Petroleum geologists began to recognize that rocks deposited during early chapters of earth history--the so termed Precambrian Eon--possesses, contrary to conventional wisdom, those seven to eight characteristics necessary for the development of an oil or gas field. Studies suggested these characteristics were often contained within rift structures, formed by the dissection of pre-existent continents, much as Africa is presently being torn into portions along the African Rift Zone.

It has been known for 35 years that northwestern Wisconsin occupies a portion of the world-class Midcontinent Rift System, extending 3300 km from Kansas to Ohio by way of Lake Superior. With the recognized association of ancient rift rocks with economic hydrocarbon reserves, that portion of the Midcontinent rift lying between the upper peninsula of Michigan and Kansas was quietly invaded by the United States oil and gas industry. Initial interests included Mobil, Texaco, Standard of California, and Amoco, among others. Field crews began operating in the early 1980's and special "frontier region" analysis teams of geologists, geophysicists, paleontologists, and geochemists were formed. Studies ranged from hand investigation of rock samples to sophisticated computer reviews of newly collected data pertaining to the seismic, magnetic and gravity character of the rift.

In 1984 Texaco made the first move by drilling, in northeast Kansas, their #1 Noel Poersch well to a record state depth of 3444 m. In certain aspects this was a guinea-pig borehole. Because the rift had never been drilled beyond normal water-well depths of several hundreds of feet, industry was probing into the geologic darkness. Expectations were based upon comparison of Kansas data to the near-surface geology of Douglas and Bayfield Counties, Wisconsin, a region where rift rocks are exposed at the surface and thus easy to study. The Texaco drilling plan envisioned several thousands of feet of sandstone and organic shale overlying an even thicker sequence of layered lava. Under ideal conditions either oil or gas would have formed within the organic shales and subsequently moved upward into the sandstones until trapped and concentrated into an economic "pool."

Drilling ceased in early 1985 but detailed results were not released until mid-1988. This three year delay made sense when information became public. The drill sequence appeared reversed when compared to expectations, with lava and related igneous rock overlying thick sandstone layers. Shale was found only in trace amounts and that shale present was not organic. Without organic shale an oil or gas field is not possible. The Texaco well was classed as "dry and abandoned" and exploration geologists began to reinterpret the geologic history of the Midcontinent rift.

Amoco Production Company was the second major player to enter the search for Precambrian-age rift hydrocarbon. Disenchanted by the Kansas results, Amoco geologists chose west-central Iowa as their preferred drilling site. In March of 1987 their #1 Eischeid borehole was set into operation and seven months later reached a state depth record of 5,441 m. This borehole was shrouded in secrecy, protected by fence and guard. Under favorable weather conditions a nearby hill was covered with curious observers and industry scouts and skimpy reports were issued daily by the media. Even after the projected depth was reached Amoco, following the lead established by Texaco in Kansas, maintained secrecy. With the release of information in the spring of 1990 encouragement was given by the fact a normal sequence of rock column was drilled--shales were found and they were organic.

The high risks associated with the oil and gas business is largely attributed to the fact that while the proper conditions for an economic reservoir are often found, the sought after product is often absent. The Iowa venture was high risk in planning and execution and the end result was labeled "dry and abandoned."

The third borehole to test the Midcontinent rift was initiated in late 1987 and completed two months later. The operator was again Amoco Production Company, but now the geographic site was near the upper peninsula community of Munising, Michigan. The final total depth was 2210 m. While complete information has not yet been released, it has been reported that all three of the formations composing the target Oronto Group were drilled--but not in the normal order as known from Wisconsin geology.

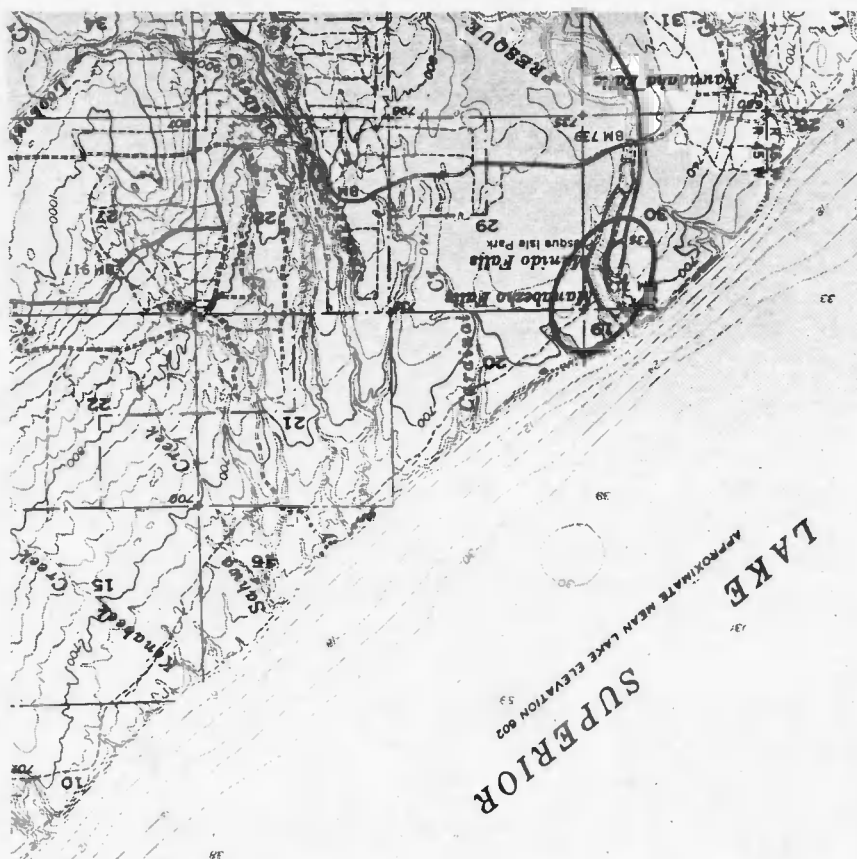
It is interesting to note this borehole is located in a region of historic, little known, exploration for crude oil. In 1865 the Ontonagon Petroleum Company, financed by a capital stock offering of \$500,000, was formed for the purpose of "engaging in and carrying on the business of mining---petroleum." By December of that year the first and only venture of this company was underway, fueled by a rapid increase in the price of oil stock and an outbreak of "oil fever." The final drilling depth was 38 m. While microfilm records indicate "two barrels (of oil) have been hoisted up," there is no record of the Ontonagon Petroleum Company beyond the year of its incorporation.

Since 1929 active underground oil seeps have been reported from the White Pine mine in Ontonagon County, Michigan. Analyses shows this crude to be very similar chemically to high grade oil of Phanerozoic age produced in the state of Pennsylvania. Radiometric age-dating techniques suggest the White Pine product may be the oldest known crude oil in the world. While these volumes are modest, their presence keeps alive the possibility of "black gold" production in northern Wisconsin.

As of early March, 1992, site preparation was underway for the fourth test of the hydrocarbon potential of this rift. Located in Bayfield County, Wisconsin, this wildcat has been programmed as a 1825 m (6000 ft.) test of the Oronto Group in its position overlying the central St. Croix horst. Available information on this Terra/Patrick #7-22 borehole will be presented in the Proceedings volume of this 38th meeting of the Institute on Lake Superior Geology.

Title: Presque Isle State Park - Nonesuch Formation

Location: Mouth of the Presque Isle River, Sec. 19 and 30, T. 50 N., R. 45 W., Gogebic County, Michigan (Thomason 15-minute topographic quadrangle, 1956).



Authors: M.G. Mudrey, Jr. and P.A. Daniels, Jr. (modified from Hite, 1968 and Daniels, 1982)

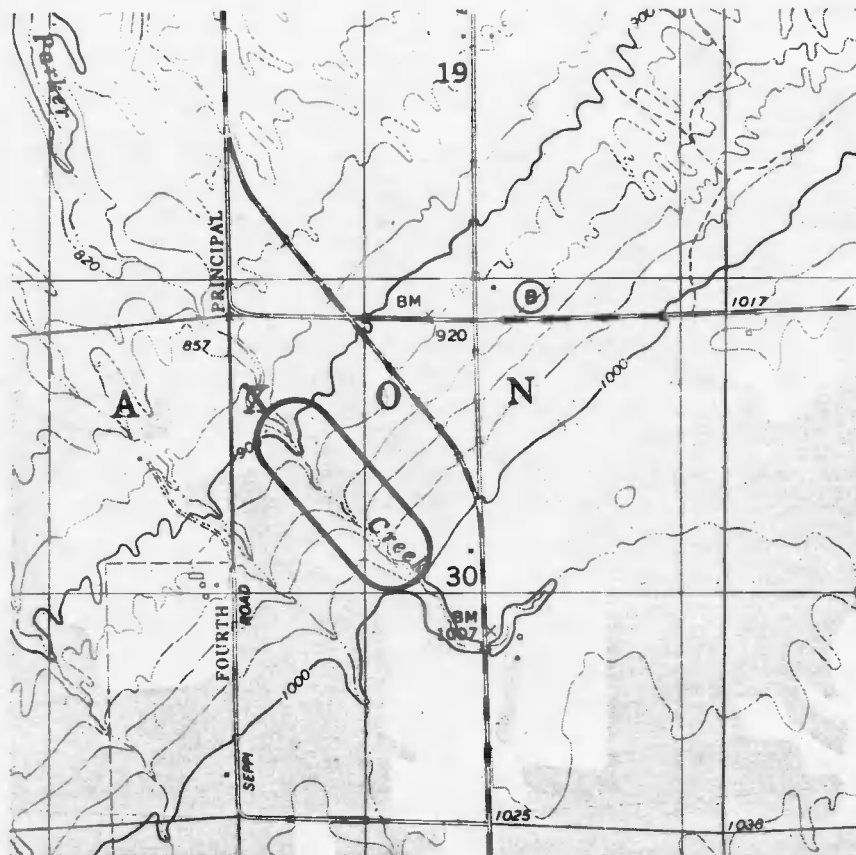
Description: The Nonesuch is more drab in color and generally of greater textural and compositional maturity than the underlying or overlying red bed sequences. Generally, the Nonesuch is a medium gray to black shale to medium sandstone. The finest grained rock is the darkest in color. The bedding ranges from massive-appearing to evenly laminated with many laminae less than 1 mm thick. Thin, rhythmic grain-size alternations are common in microscale, and many show graded bedding and dewatering features such as clastic dikes. The rock is rich in organic material and is petroliferous in the White Pine area. The Nonesuch averages about 180 m in thickness, and generally thins to the southwest. In the White Pine area about 40 km east of Presque Isle State Park, the Nonesuch thickens basinward at more than 4 m/km. Structural attitude varies from overturned in the Porcupine Mountain area to about 10 degrees in the White Pine and Presque Isle River areas, and 75 degrees in Wisconsin.

The Nonesuch possesses a great variety of sedimentary features. Both horizontal stratification and cross stratification occur. Laminated stratification is the most common; lenticular, wavy or irregular, and ripple stratification also occur. Units rarely exceed 60 cm in thickness, and the thicker beds are coarser sandstone and conglomerate. Alternating laminae of fine gray sandstone and black shaley siltstone less than 0.5 cm thick are common. The cross-stratification is of two types, planar cross stratification and rib and furrow. Both of these types are abundant on the Presque Isle River in Michigan. Shrinkage cracks and disturbed bedding occur locally and sometimes abundantly. Sediment and current transport data support a dominant flow regime to the west-southwest during deposition of the Nonesuch in the Presque Isle area.

Discussion: The sedimentary structures indicate that the depositional environment of the Nonesuch Formation was that of a standing body of water, with perhaps significant variation in water depth. Salinity was at least high enough to precipitate gypsum.

The initial formation of this water body could have occurred either due to subsidence along the rift, creating a closed topographic low that would then act as local base level, or perhaps, more likely, the disruption of regional drainage pattern by some type of damming.

Location: Approximately 300 m upstream to the southeast from the northwest corner of section 30. NW1/4, Sec. 30, T. 47 N., R. 1 E., Iron County, Wisconsin (Oronto Bay 7 1/2-minute topographic quadrangle, 1980).



Author: M.G. Mudrey, Jr. (modified from Myers, 1971 and Rosenberry, 1924)

Description: Approximately 600 m of continuous exposure from the last Keweenaw lava flow, through the Copper Harbor and Nonesuch Formations into the lower 300 m of the Freda area exposed along Parker Creek (also known as Davis Creek). Because the upper reaches of the creek tend to run parallel to the bedding, the Copper Harbor appears to be much thicker than normal.

All of the units trend N. 60 degrees E., 85 degrees NW. There are a few, very minor flexures and faults in the wall of the valley, but the section appears to be continuous with no repetition by faulting.

Approximately 200 m of Copper Harbor Formation are exposed in the upper reaches of the creek. The unit appears to consist of five main conglomeritic zones about 30 to 60 m in thickness, interbedded with coarse, sandy

conglomerate. Both lithologies are clast-supported conglomerate. The Copper Harbor is relatively typical, consisting of rounded cobbles of Keweenaw volcanic rock and granite and iron formation from the underlying Early Proterozoic and Archean succession.

About 310 m of Nonesuch Formation is exposed. The Nonesuch consists of several dark colored slate units, interbedded with sandy slate. The lower contact of the Nonesuch and the Copper Harbor is gradational in that the nonconglomerate interbeds become finer grained and more abundant as the upper contact with the Nonesuch is approached. Bedding is clearly defined in the Nonesuch, and local cross bedding and deformation structures can be recognized. The upper contact of the Nonesuch with the Freda is also gradational in that sand beds become more common and slate beds less so.

The lower member of the Freda Formation is a thick bedded, fine- to medium-grained sandstone that is strongly stained with iron. Local pebble zones are common. The middle Member of the Freda is seen only at the northernmost outcrop in the creek. It is much better exposed in Spoon Creek to the west. This member consists of siltstone and very fine-grained sandstone with occasional shale and cross-stratified fine-grained sandstone beds. Bedding is more pronounced in the middle member compared to the lower member. The upper member of the Freda is best seen at the mouth of the Montreal River to the east, particularly at Superior Falls in Michigan. The upper member consists of thick bedding units of micaceous siltstone, very fine-grained, laminated sandstone with abundant scours and current direction indicators, and slightly conglomeritic, fine-grained sandstone with bedding poorly developed. The White River location is probably in the upper member.

Discussion: Clearly, the Oronto Group represents a single depositional episode, the individual formations representing various environments within that episode. All of the units appear to be gradational. The Freda appears to thicken from its type locality near Freda, Michigan, to the Montreal River and west. The Nonesuch, conversely, appears to thin toward the west from Presque Isle State Park. At Copper Falls State Park the dominant shale unit appears to be less than 20 m thick; however, the shale-bearing interval is about 140 m thick.

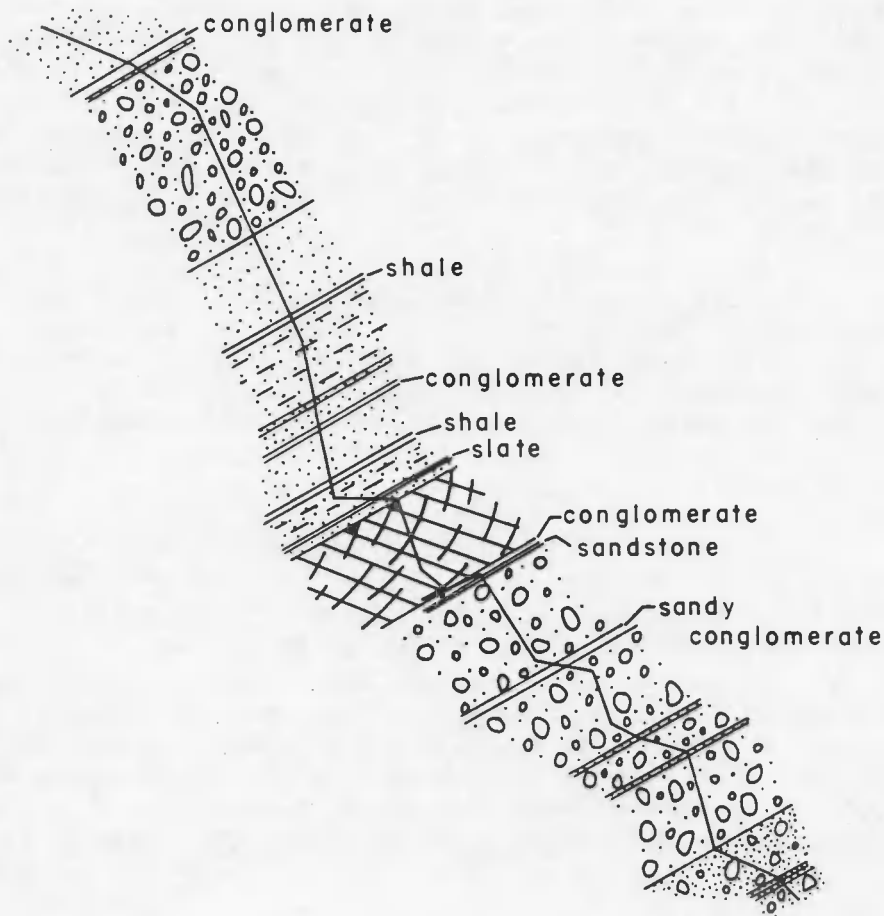
All along the Gogebic Range, the Early and Middle Proterozoic units dip northwest at a steep angle. Reconstruction of the thickness of the section results in an estimate that is too thick, 20 km of section from Hurley to Lake Superior. In Michigan this is easy to understand as the Keweenaw and associated faults clearly repeat parts of the section; however, in Wisconsin because of poor exposure, the faults are not readily recognized. As will be seen and discussed at White River, there is at least one major anticline known in outcrop at Marble Point, and at least one major syncline with the Freda. A more reasonable estimate for the total thickness of the Oronto Group would be 5 km.

NE 25

○

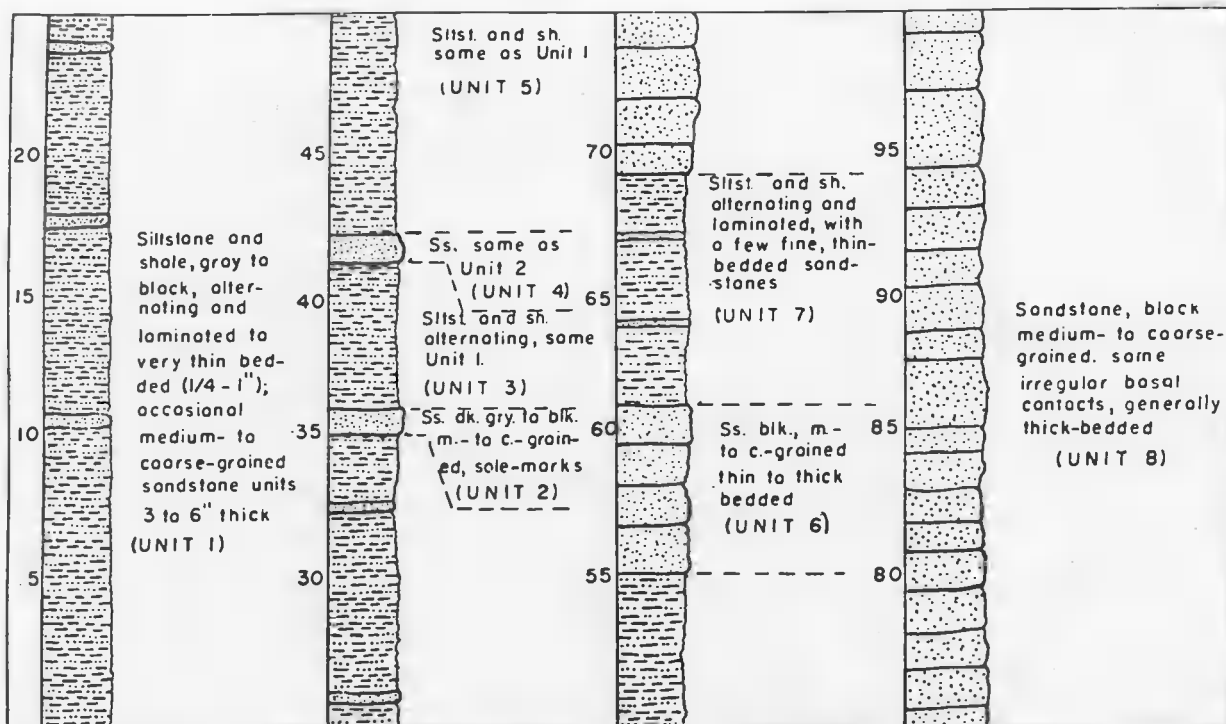
Distance from NE25 to E1/4 25
= 1/2 mile

-  conglomerate
-  sandy conglomerate
-  sandstone
-  sandstone and shale
-  shale
-  slate

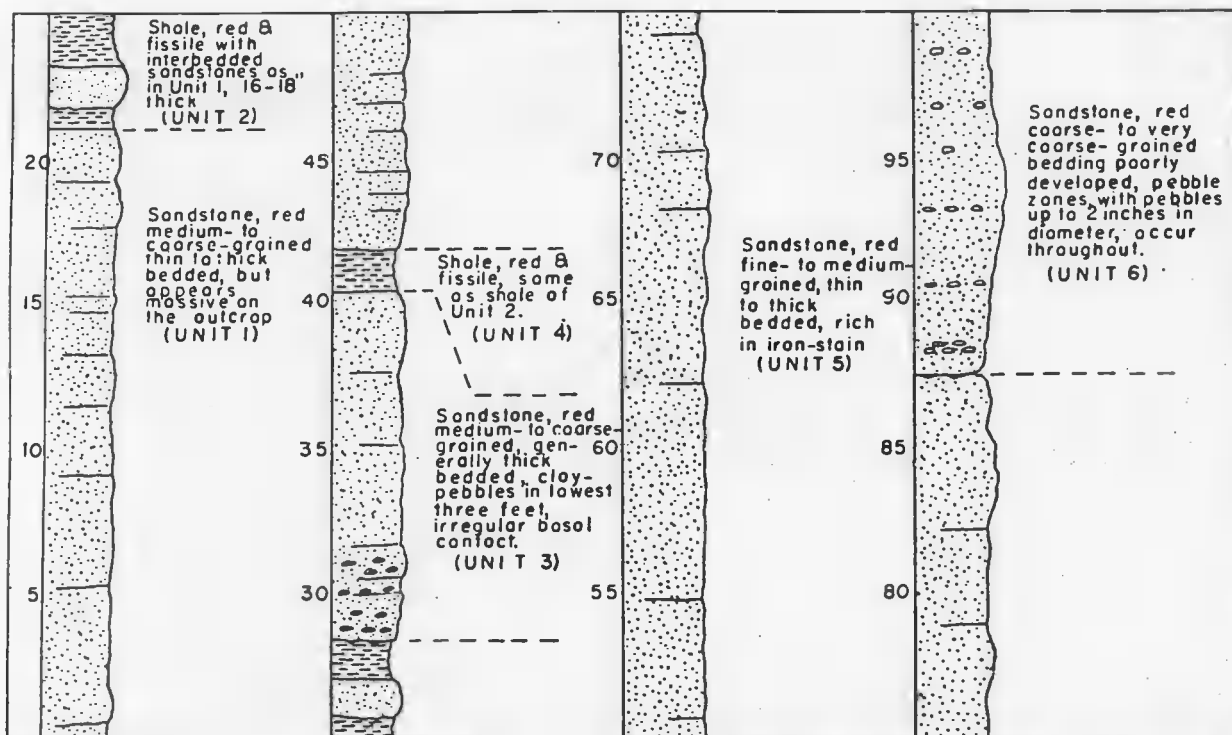


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E 1/4 25

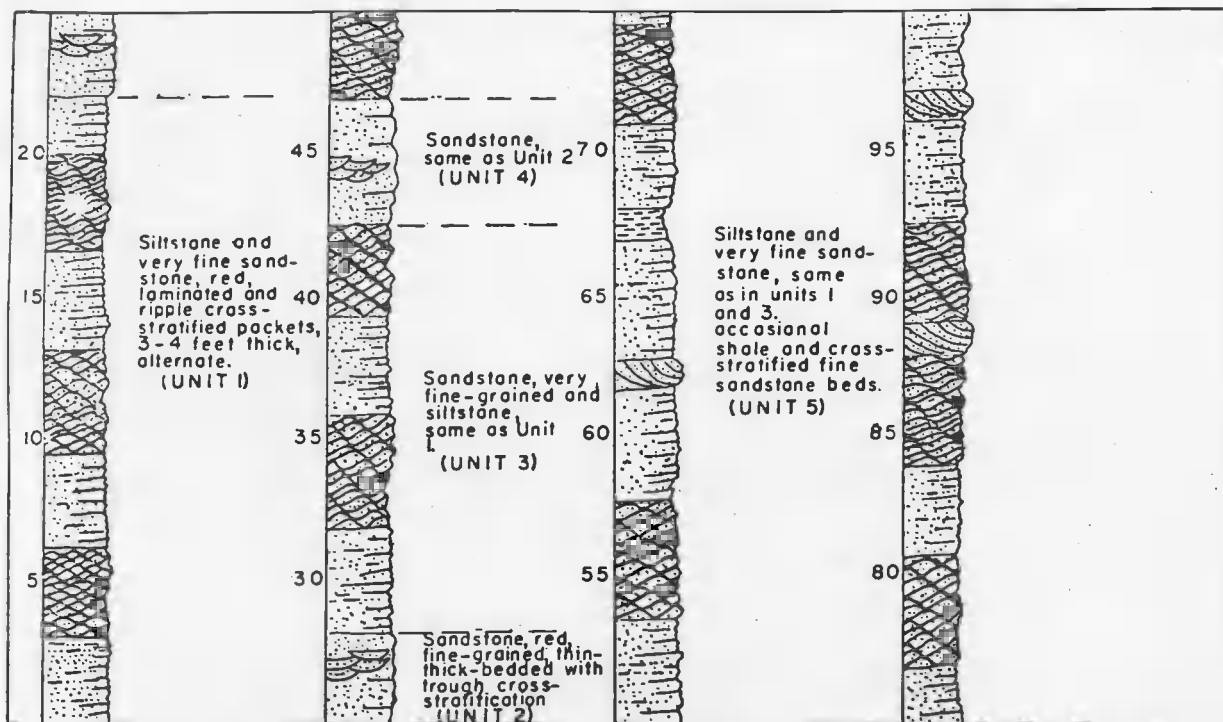
Sketch of geology in Parker Creek from Rosenberry (1924).



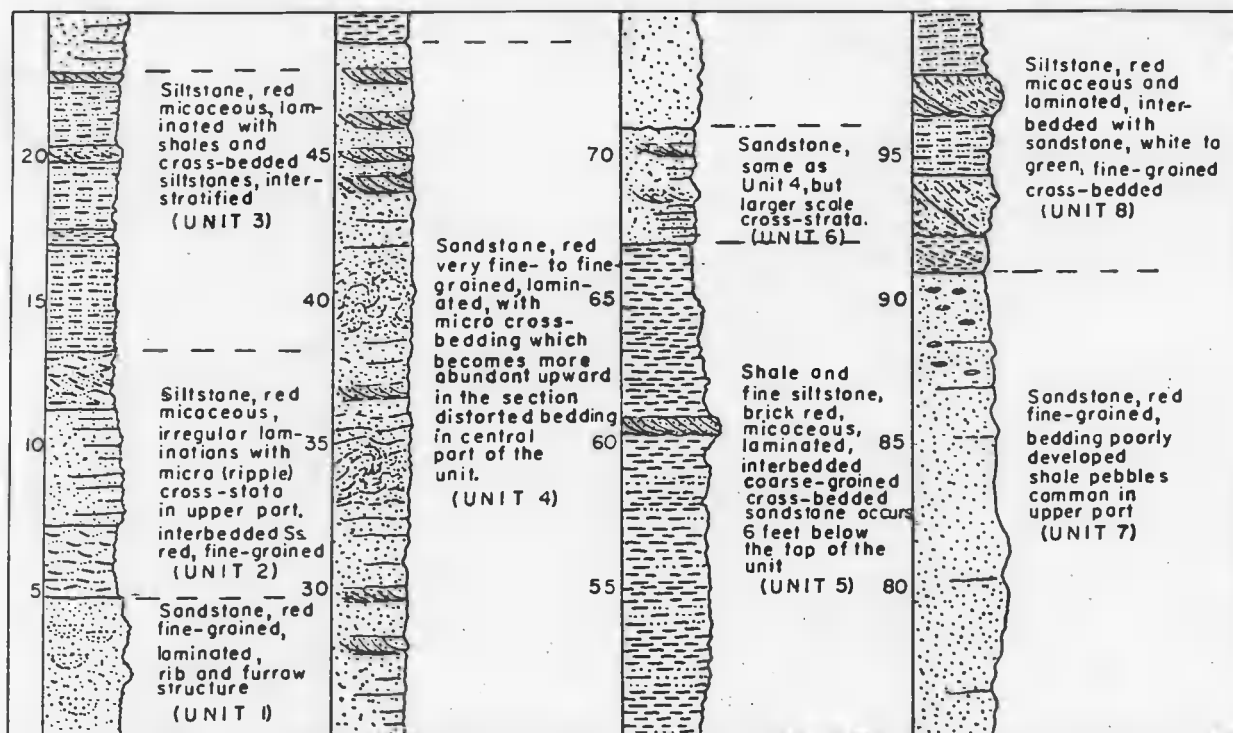
Stratigraphic section showing 100 feet of typical Nonesuch Formation (figure 20 of Hite, 1968).



Stratigraphic section showing 100 feet of typical Lower Freda Formation (figure 30 of Hite, 1968).

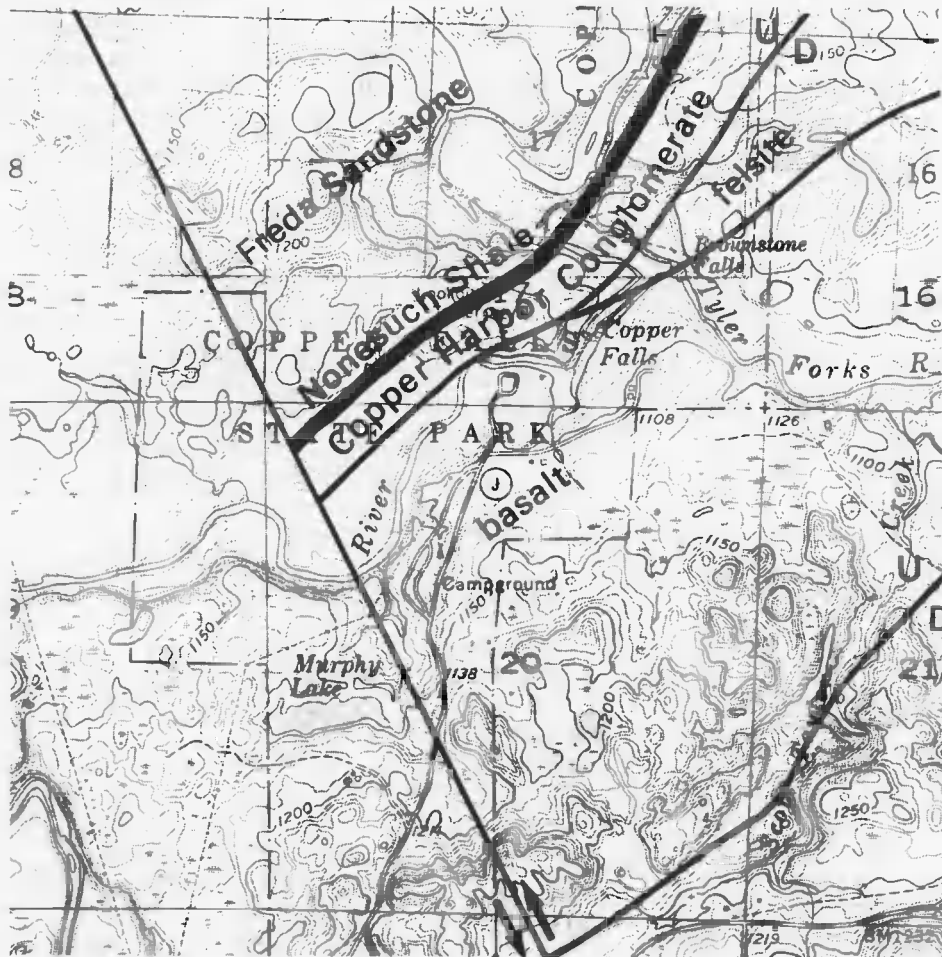


Stratigraphic section showing 100 feet of typical Middle Freda Formation (figure 33 of Hite, 1968).



Stratigraphic section showing 100 feet of typical Upper Freda Formation (figure 42 of Hite, 1968).

Location: Northwest of Mellen on Highway 169. SE1/4, Sec. 17, T. 45 N., R. 2 E., Ashland County, Wisconsin. (Mellen and High Bridge 7 1/2-minute topographic quadrangles, 1967 and 1984, respectively).



Author: M.G. Mudrey, Jr.

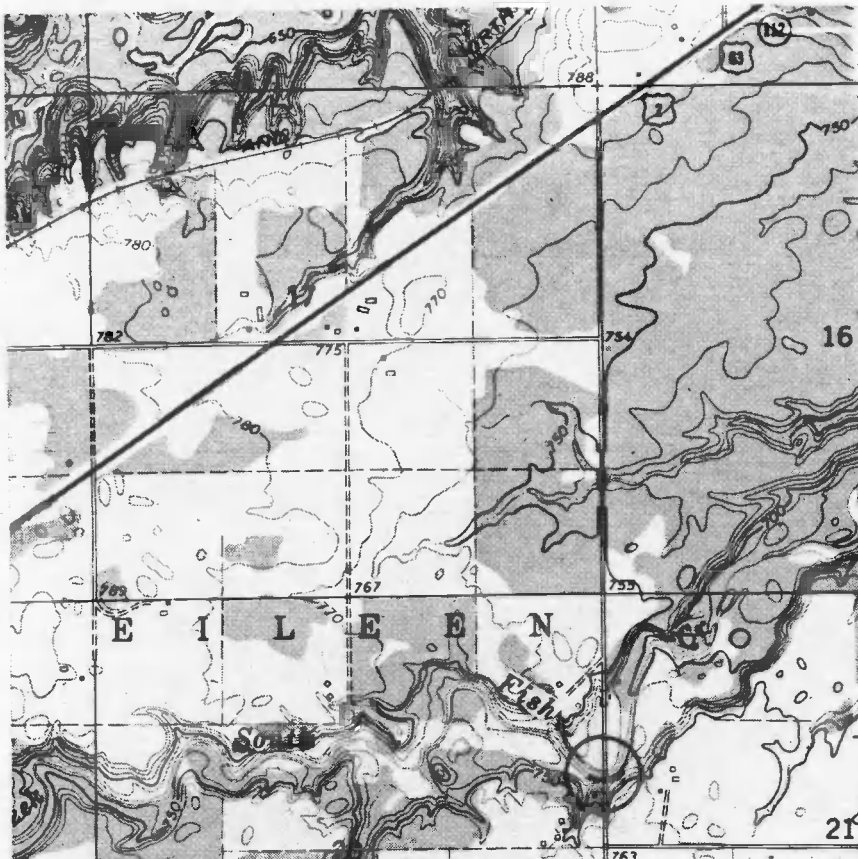
Description: Copper Falls is formed where the Bad River cuts through the resistant ridge of Keweenaw lava flows. Downstream, 0.6 km from Copper Falls at the contact between basalt to the south and rhyolite to the north, the Tyler Forks River joins the Bad River at Brownstone Falls, which occurs at the contact between basalt to the south and rhyolite to the north. The conglomerate of the Copper Harbor Formation is exposed in the lower gorge at Devils Gate about 200 m downstream from Brownstone Falls. The conglomerate is 129 m thick; at the Montreal River the conglomerate is 515 m thick. Northwest of the conglomerate is the black shale of the Nonesuch Formation; northwest of the Nonesuch is the sandstone of the Freda Formation. There is black shale distributed through 141 m of section, but only 20 m are considered Nonesuch Formation. Above the interval with black shale, the sediment is red shaley arkose.

The name Copper Falls comes from a small copper prospect in a small ravine about 200 m south of the concession building. The working was begun by the Ashland Mining Company in August 1864 and closed in February 1866. Copper in a quartz vein in diabase was prospected. Apparently the property was reopened at the turn of the century.

Discussion: Although faults complicate the local geology, the essentially conformable relationship between the Oronto Group and Keweenawan flows can be seen. The lower parts of the Oronto Group, the Copper Harbor and the Nonesuch, are significantly thinner here than at localities to the east.

Title: South Fish Creek

Location: Exposures in banks of South Fish Creek beneath bridge on north-south secondary road 1.2 miles south of U. S. Highway 2 on the east line of the $SE\frac{1}{4}$, $SE\frac{1}{4}$, $NE\frac{1}{4}$, Sec. 20, T.47N., R.5W., Bayfield County (Moquah 7.5 minute topographic quadrangle, 1964).



Author: M. E. Ostrom (modified from Myers, 1971)

Description: Exposures of steeply-dipping Freda Sandstone exhibit the lithologic and mineralogic character of the formation. A description of the strata downstream from the bridge is:

PRECAMBRIAN SYSTEM

Keweenaw Series

Oronto Group

Freda Sandstone Formation (11.0 feet)

- 11.0' Sandstone, grayish red to reddish brown, uniformly fine-grained, hard, cross-bedded with parting lineation. Much leaching. Penecontemporaneous deformation. Current ripple marks found in float.

BASE OF EXPOSURE

Significance: Provides evidence of environmental, geologic and structural history. Examine lithology and mineralogy. What do they signify? What direction is the top of the beds? Measure dip and strike of beds. What do these mean in terms of structural history? From what direction did the sand come? What is the origin of the red color?

References: Thwaites, 1912; Myers, 1971.

Title: Washburn Harbor -- Chequamegon Formation BA 48/04W/33

Location: Outcrop on northside of Washburn Harbor, SW1/4, SE1/4, Sec. 33, T48N, R4W, Bayfield County, Wisconsin (Washburn 7.5-minute quadrangle, topographic, 1975).



Author: M.G. Mudrey, Jr. (1992)

Description: This low outcrop of Chequamegon sandstone of the Bayfield Group consists of thick, massively bedded, fine - to medium-grained, grayish-red, quartzose sandstone. The grains are subangular to subrounded. The bedding surfaces are gritty to coarse grained, but are thin and discontinuous. Some bedding planes display scattered small quartz pebbles. Here, bedding dips gently to the southeast; however, at Big Rock Wayside Park at the east quarter corner of Section 24, T48N, R5W, crossbedded sandstone dips 15 degrees to the northwest, suggesting that small, low-amplitude folds with a northeasterly trend may occur in the region.

Discussion: The Chequamegon sandstone is similar to the underlying Orienta sandstone, which is found in fault contact

with Keweenawan volcanics and Oronto group strata. Bayfield Group strata are not known to have high dips, and on this basis have been distinguished from the upper parts of the underlying Freda sandstone. The Douglas Fault intervenes between this locality and South Fish Creek where the Oronto Group dips south at a high angle to the south. A petroleum test well has been drilled between these two locations immediately adjacent to the Douglas Fault. At the time of this writing, no further information was available.

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**GEOLOGY OF
KEWEENAWAN
SUPERGROUP ROCKS
NEAR THE
PORCUPINE MOUNTAINS,
ONTONAGON & GOGEBIC
COUNTIES, MICHIGAN**

William F. Cannon,
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U.S. Geological Survey,
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INTRODUCTION

This field trip examines the geology of rocks of the Keweenaw Supergroup (1.1 Ga) and related intrusive rocks of the Midcontinent rift system (MRS) in the western part of the northern peninsula of Michigan. The combination of stops includes all formations of the Keweenaw Supergroup in this region. Examination of all described localities requires more than a single day and participants are encouraged to use this guidebook on their own to supplement the localities that will be visited on our one-day trip. Because of uncertainties of weather, road conditions, and remaining snow pack in early May in this region of very heavy snowfall, the stops that we will visit will not be known until the date of the trip. Stops are numbered in stratigraphic order, from oldest to youngest, not in the order in which they will be visited.

GENERAL GEOLOGY

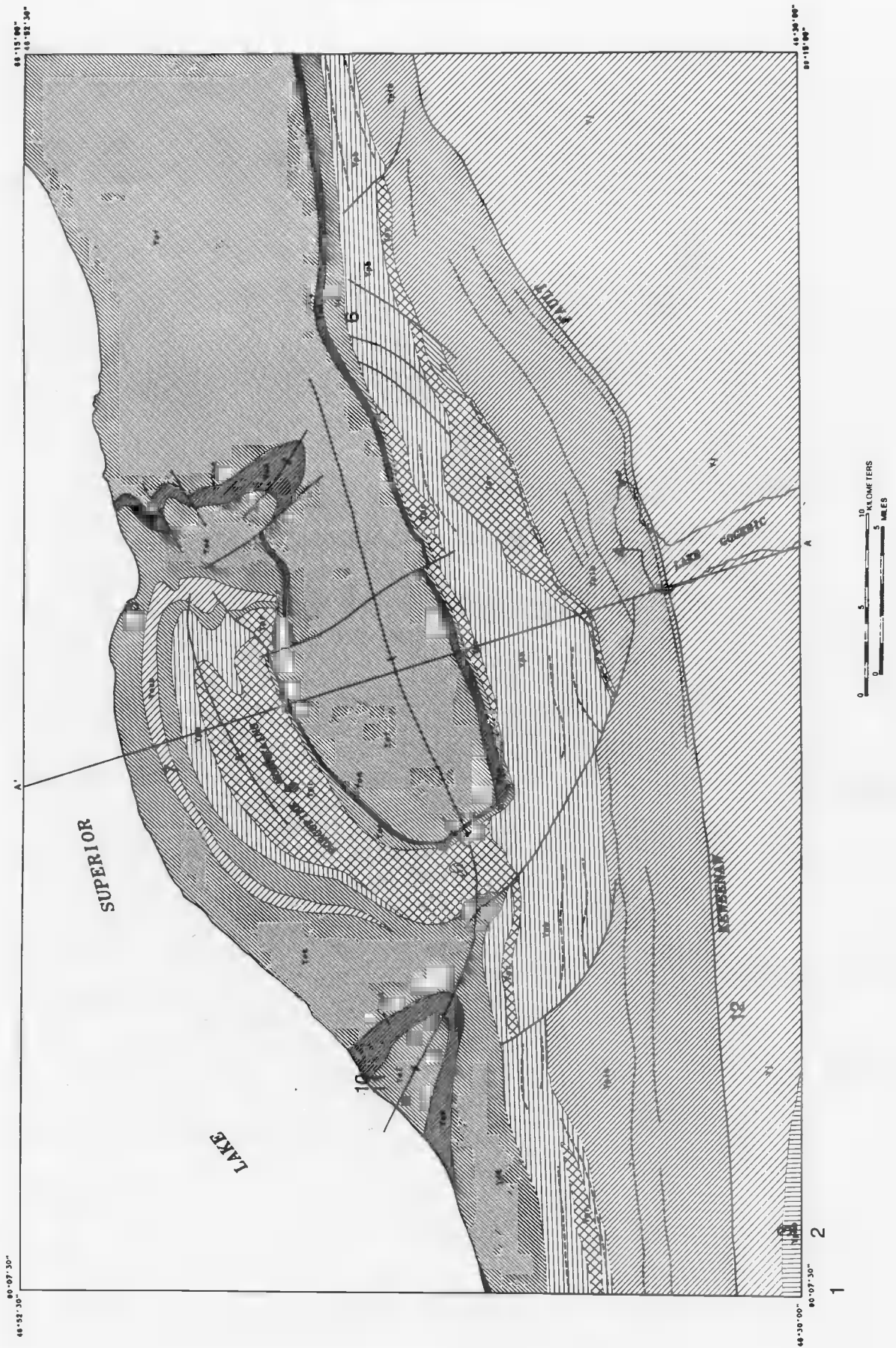
The Keweenaw Supergroup is a very thick sequence of volcanic and sedimentary rocks that was deposited during and shortly after an episode of continental rifting at about 1.1 Ga, when the MRS formed within the Proterozoic craton. The Keweenaw Supergroup was deposited in and marginal to rift graben and in a post-rift thermal basin more or less centered on the axis of the former rift. In the area of the field trip (figs. 1 and 2), the Keweenaw stratigraphic section (fig. 3) consists of a thin basal quartzite (Bessemer Quartzite) that is overlain by a great thickness of subaerial flood basalt flows and lesser andesite and felsite flows (Powder Mill Group and Portage Lake Volcanics). The volcanic section becomes progressively thicker toward the rift axis and exceeds 20 km in places beneath Lake Superior (Behrendt and others, 1988; Cannon and others, 1990).

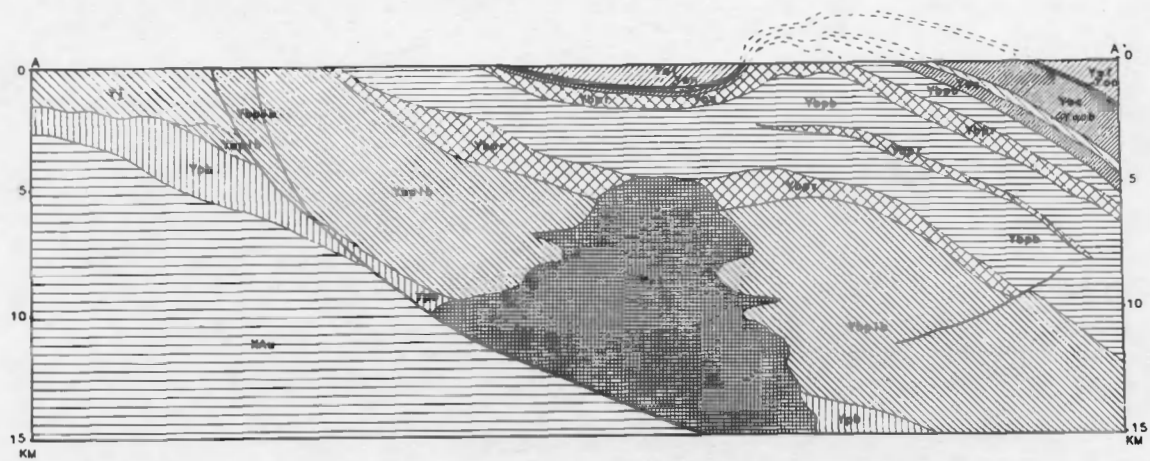
In the vicinity of the Porcupine Mountains, the volcanic sequence includes, at the top, a stratovolcano composed mostly of andesite and rhyolite flows and sub-volcanic felsic intrusions (Porcupine Volcanics). These late intermediate and felsic volcanic rocks are atypical of the MRS as a whole and are limited to a few felsic volcanic centers.

Conformably overlying the volcanic rocks are fluvial sedimentary rocks and lesser lacustrine sedimentary rocks (Oronto Group), which are as much as 8 km thick beneath Lake Superior (Cannon and others, 1990) and at least 5 km thick on shore in the field trip area.

Post-rift reverse faulting has caused block rotation on a very large scale so that the Keweenaw Supergroup section is steeply to vertically dipping. More than 13 km of volcanic rocks and 5 km of sedimentary rocks are exposed in a generally north-facing section. This enormously thick section constituted at least half of the crustal thickness by the close of rifting and thus our field trip provides a traverse through the upper half of the Middle Proterozoic crust.

The initial subsidence of the MRS is recorded by deposition of the Bessemer Quartzite (stop 1), a blanket of relatively pure, fluvial quartzite as much as 100 m thick (Ojakangas and Morey, 1982). It formed in a broad basin, more or less centered on the site of the future deep rift. The age of deposition has an upper bound of about 1109 Ma, the age of the oldest overlying basalt flows (Davis and Sutcliffe, 1985). Following deposition of the Bessemer, the area became volcanically active and a great thickness (as much as 15 km) of basalts and lesser andesites and rhyolites accumulated in only about 15 m.y. The earliest basalts constitute the Siemens Creek Formation of the Powder Mill Group (Hubbard, 1975a) (stops 1 and 2), which lies conformably on the Bessemer Quartzite. Soft sediment deformation of the Bessemer by the overriding basal basalt flow (seen at stop 1) indicates a very short interval between the units. Locally, as at stop 1, the basal flow is pillowed, but all succeeding flows are subaerial.





- EXPLANATION**
- Intrusive rocks**
- Cobble or diabase-grenophyre intrusive complexes
 - Inferred felsic stock
- Volcanic and sedimentary rocks**
- Jacobeville Sandstone- red and brown feldspathic and quartzose sandstone, siltstone, conglomerate
 - Oronto Group
 - Freda Formation- red sandstone and mudstone
 - Nonesuch Shale- gray-black siltstone, shale, fine sandstone
 - Copper Harbor Conglomerate- coarse red-brown conglomerate and coarse sandstone
 - Copper Harbor Conglomerate basalt member
- Bergland Group**
- Porcupine Volcanics rhyolite member- aphyric to quartz and feldspar aphyric rhyolite in subaerial domes, flows and debris flows
 - Porcupine Volcanics basalt and andesite member- thin flows of andesite and minor tholeiitic basalt
 - Portage Lake Volcanics- mostly aphyric tholeiitic basalt flows, minor andesite, rhyolite and interflow conglomerate
- Powdermill Group**
- Powdermill Group (undivided)- Tholeiitic basalt flows, minor andesite and rhyolite
- Lower Proterozoic and Archean rocks (undivided)
- Trend of prominent aeromagnetic lineaments

Figure 2. Generalized geologic map and section of the Porcupine Mountains area showing the location of field trip stops.

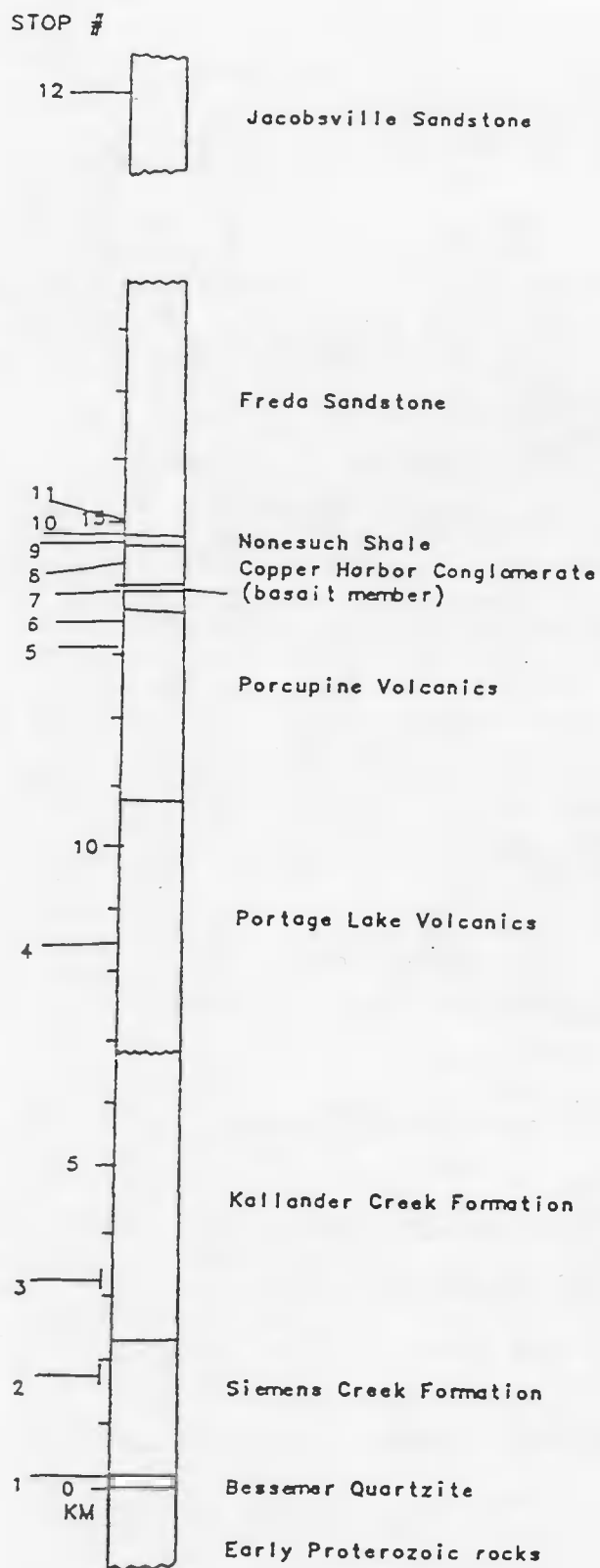


Figure 3. Stratigraphic section of Keweenaw Supergroup rocks in the field trip area showing the stratigraphic position of stops. Thicknesses are approximate average thicknesses for the area. Some formation thicknesses vary significantly across the region.

Overlying the Siemens Creek Formation is a thick section of more felsic rocks (stop 3) composed of rhyodacite, trachyandesite, andesite, and basalt. These rocks constitute the Kallander Creek Formation of the Powder Mill Group (Hubbard, 1975a). The Powder Mill volcanic rocks are distinguished from overlying basalts of similar character largely on the basis of their reversed magnetic polarity as opposed to the normal polarity of younger units such as the Portage Lake Volcanics. In Michigan, the Powder Mill Group crops out in a belt that is separated from the outcrop belt of the overlying Portage Lake Volcanics by the Jacobsville Sandstone, which lies with angular unconformity on Powder Mill rocks. In the subsurface, however, the Powder Mill and Portage Lake probably form a continuous depositional sequence.

Volcanic rocks of the Keweenaw Supergroup range in composition from olivine tholeiite to rhyolite. By far the dominant rock type is high-Al olivine tholeiite ($\text{Al}_2\text{O}_3 = 15\text{--}19 \text{ wt\%}$) followed by lesser high-Fe tholeiite and rocks of intermediate and felsic composition (Green, 1982; Brannon, 1984; Paces, 1988). The olivine basalts commonly are ophitic in texture and the dominant phenocryst is plagioclase. The most primitive Keweenaw basalts are geochemically similar to primitive midocean ridge basalts. However, incompatible trace elements in most Keweenaw basalts are enriched compared to depleted or primitive mantle. Radiogenic isotope analyses (Sr, Nd, and Pb) of the main stage high-Al olivine tholeiites suggest that a likely source of the voluminous basalts is a mantle plume (Paces and Bell, 1989; Nicholson and Shirey, 1990).

Some of the oldest flows, such as those in part of the Siemens Creek Formation, have distinctly different chemical compositions than the younger basalts. These basal rocks are transitional to weakly alkaline olivine basalts characterized by low Al_2O_3 content and clinopyroxene phenocrysts. Locally, in other parts of the rift, basal flows are picritic but such flows have not been identified in this area.

Flows near the base of the exposed Portage Lake Volcanics north of the Keweenaw fault were erupted at about 1096 Ma and those near the top of the formation at about 1094 Ma (Davis and Paces, 1990). Thus, the great thickness of Portage Lake Volcanics, at least 8 km in this area, was erupted in only a few million years. The estimated eruption rate for Portage Lake Volcanics in the western Lake Superior region must have approximated the present eruption rate of the Hawaiian hot spot, the most vigorous volcanic center of the modern earth (Cannon, in press). Because synchronous volcanism occurred along the entire trend of the rift, the rift system as a whole was producing basalt at a rate unrivalled by any modern analog.

The area of the Porcupine Mountains contains a felsic to andesitic volcanic center that became active late in the volcanic history of the region, at about 1090 Ma. The Porcupine Volcanics (previously called the "unnamed formation") (Cannon and Nicholson, 1992) were erupted from that center and accumulated as much as 5 km of andesite, rhyolite, and basalt in a large shield deposited on top of the Portage Lake Volcanics lava field and centered near the Porcupine Mountains. The present arcuate shape of the mountains and the unusual hook-shaped map pattern of the Porcupine Volcanics is partly a reflection of the original shape of the volcanic shield.

The Porcupine Volcanics consists of a sequence of subaerially deposited andesite, basalt, felsite, and quartz-porphyry lava flows, and minor interbedded volcanoclastic lithic sandstone, siltstone, and conglomerate (Hubbard, 1975b). The abundance of felsic rocks and the predominance of andesite over basalt clearly distinguish the Porcupine Volcanics from the underlying Portage Lake Volcanics. Felsite is most common near the top of the formation where it occurs as both flows and domes (stops 5 and 6).

The major element chemistry of basalt, basaltic andesite, and andesite of the Porcupine Volcanics and the Portage Lake Volcanics are very similar, but the Porcupine Volcanics are distinctly enriched in rare earth elements (REE) and Th compared to the Portage Lake Volcanics. The two formations differ more profoundly in their rhyolite chemistry. Rhyolite that occurs most commonly in the Portage Lake Volcanics is aphyric or may contain sparse quartz phenocrysts. Rhyolites of the Portage Lake Volcanics on Keweenaw Peninsula have very low abundance of incompatible trace elements (such as REE, Zr, Y, Hf, and Th), whereas the rhyolite body near Bergland (stop 4), one of the few rhyolite bodies with the Portage Lake Volcanics in the Porcupine Mountains area, has moderate abundances of incompatible trace elements. In contrast, the numerous rhyolite bodies in the Porcupine Volcanics range from rhyolites that are aphyric to those with abundant quartz and/or

The thrust faulting and folding can be indirectly dated in the interval of approximately 1060 to 1040 Ma (Ruiz and others, 1984; Bornhorst and others, 1988; Cannon and others, 1990). The cause of the regional compression is not known with certainty but is likely linked in some manner with the Grenville orogeny, which was in progress at that time.

FIELD TRIP STOPS

Stop 1.--Bessemer Quartzite and basalt at the base of the Powder Mill Group

A low rock knob in a pasture east of the county road (see fig. 4) shows quartzite of the upper part of the Bessemer Quartzite in contact with the overlying basal basalt flow of the Powder Mill Group. The basalt is the lowermost flow of the Siemens Creek Formation (Hubbard, 1975a). The Siemens Creek is composed predominantly of thin flows of basalt and minor andesite. Average thickness of flows is about 3 m, but the basal flow is about 50 m thick at this locality. Here, the basal flow is nonporphyritic and holocrystalline. Unlike some of the basal flows along strike to the west, these flows do not contain clinopyroxene phenocrysts. Chemical analyses (table 1) of the basal flows are similar to the younger high-alumina main-stage basalts of the Portage Lake Volcanics. The basal Siemens Creek flow is pillowed, which is very unusual for basalts of the Keweenawan Supergroup, nearly all of which are subaerial. In places, stringers of the quartzite have been injected up into rubbly material at the base of the flow, indicating that the Bessemer was unconsolidated at the time of eruption.

Note that dips here are about 65° N. Contrast these with steeper to slightly overturned dips at stops 2 and 3, higher in the Powder Mill section. This updip fanning relationship commonly shown by the Powder Mill flows led Hubbard (1975a) to propose that they were deposited in a basin that was centered south (updip) of the present outcrop belt and the volcanics originally thickened in that direction. If that interpretation is correct, there must have been a rapid change in depocenters at the outset of volcanism because most paleocurrent directions in the Bessemer are northerly (Ojakangas and Morey, 1982).

Stop 2.--Basalt flows of Siemens Creek Formation

Roadcuts along Powderhorn Road (fig. 4), which follows the valley of Powder Mill Creek, provide a convenient cross section of the Powder Mill Group in its type area as defined by Hubbard (1975a). At stop 2 typical basalt flows of the Siemens Creek Formation are exposed and at stop 3 the Kallander Creek Formation can be seen. Steeply north-dipping basalt flows of the Siemens Creek Formation are well exposed in outcrops on the hill west of the road and in low roadcuts. The basalt flows are thin, typically a few meters or less thick. Commonly these fine-grained flows have amygdaloidal and rubbly flow tops, and pipe vesicles can be seen locally at the base of some flows. The basalts show patchy alteration to chlorite and have abundant chlorite in the matrix. This relatively high degree of metamorphism is typical of basalts of the Siemens Creek Formation and is one distinction between them and less metamorphosed basalt of the overlying Portage Lake Volcanics. Some flows are porphyritic with small plagioclase phenocrysts.

Stop 3.--Kallander Creek Formation

About 2 km north of stop 2, a road improvement project has produced freshly blasted outcrops in the Kallander Creek Formation. Because all flows are dipping nearly 90° in this area, the drive between stops passes up through about 2 km of section, an impressive pile of basalt, yet this is only 10% of the total thickness of basalt erupted into the deepest parts of the rift. The flows in the Kallander Creek Formation are basaltic to andesitic, more rarely rhyolitic, and generally a few meters thick. Overall, the Kallander Creek is more felsic than the Siemens Creek. At this stop the exposed flows include chloritized amygdular basalt, fine-grained basalt with plagioclase phenocrysts (locally a "daisy stone"), non-porphyritic andesite, andesite with plagioclase phenocrysts, and tuff breccias containing both basalt and andesite fragments. Although rhyolite occurs higher in the formation, none is exposed here. The Kallander Creek flows are generally less metamorphosed than the Siemens Creek flows.

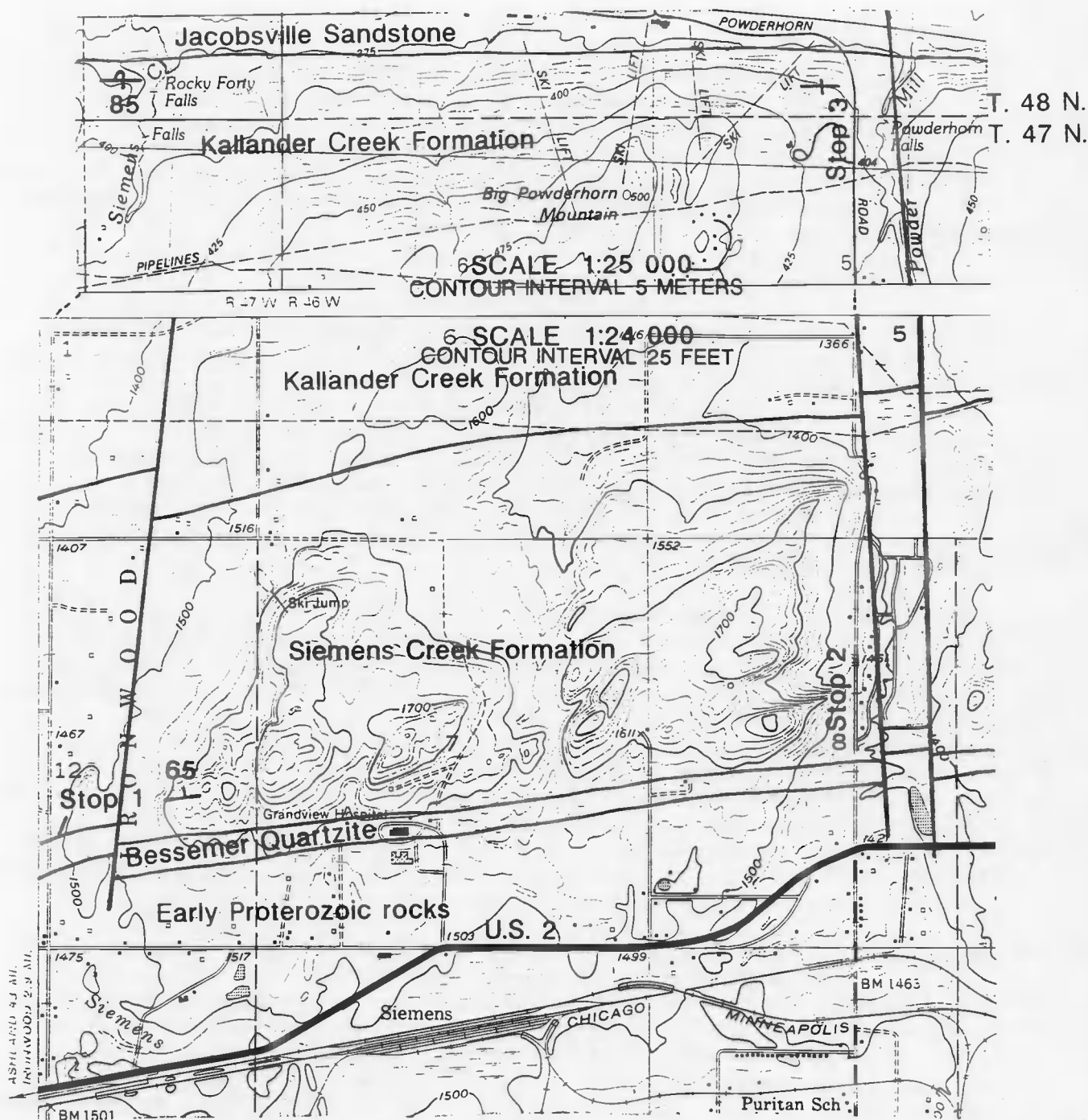


Figure 4. Location and geologic setting for stops 1, 2, and 3. Geology generalized and modified from Hubbard (1975a). Note change in scale between the two adjacent quadrangles.

TABLE 1: CHEMICAL ANALYSES FOR SOME VOLCANIC UNITS IN THE KEWEENAWAN SUPERGROUP

ELEMENT	1 N=2	2 N=4	3 N=5	4 N=1	5 N=2	6 N=11	7 N=8
SiO ₂	52.02	77.23	48.66	71.79	47.60	48.29	50.68
TiO ₂	1.75	0.12	1.69	0.42	2.26	1.70	2.47
Al ₂ O ₃	14.32	12.91	16.51	12.84	17.03	16.99	14.36
Fe ₂ O ₃	1.85	0.60	1.83	1.74	2.04	1.65	2.11
FeO	9.42	1.19	10.38	3.49	12.30	10.90	11.94
MnO	0.17	0.03	0.18	0.10	0.19	0.22	0.23
MgO	6.93	0.30	7.37	1.18	7.60	8.16	5.32
CaO	8.18	1.43	10.59	0.23	5.11	6.50	6.46
Na ₂ O	3.57	1.57	2.29	0.76	4.09	2.86	2.70
K ₂ O	1.57	4.62	0.32	7.38	1.26	2.53	2.89
P ₂ O ₅	0.23	0.01	0.18	0.06	0.53	0.20	0.86
SUM	99.99	100.00	100.00	99.99	100.00	100.00	100.00
Cr	233	2.9	188	2.5	168	159	64
Ni	94	1	204	<2	95.5	188	42
Nb	20	50	8	32	17	8.7	18
Rb	35	212	7	219		61	62
Sr	470	18	252	53		490	407
Zr	123	281	100	1130		126	294
Y	16	112	23	103		26	56
La	19.0	76.9	11.6	273	37.2	11.9	55.7
Sm	4.9	12.9	4.0	27.3	7.1	4.0	11.1
Yb	1.67	10.00	2.08	10.4	3.55	2.12	5.11
Hf	3.12	10.24	2.65	22.2	4.475	2.65	7.10
Ta	1.04	3.92	0.58	2.96	0.86	0.59	1.32
Th	1.94	39.48	1.04	35.70	2.63	1.20	5.94

1 Average analysis of Siemens Creek basalts

2 Average analysis of rhyolite from quarry near town of Bergland

3 Average analysis of basalts of the Portage Lake Volcanics north of crossroads at Merriweather near Bergland

4 Analysis of rhyolite at quarry near White Pine mine

5 Average analysis of basalts at Lake of the Clouds overlook in Porcupine Mountains Wilderness State Park

6 Average analysis of basalts of the Portage Lake Volcanics on Keweenaw Peninsula

7 Average analysis of basalts from the Porcupine Volcanics

Stop 4.--Portage Lake Volcanics

This stop illustrates rhyolite, basalt, and interflow sandstone of the Portage Lake Volcanics. Exposures are in a roadcut and quarry along the north side of Michigan Highway 28 (fig. 5). The prominent south-facing scarp that runs parallel to the highway usually is the geomorphic expression of the Keweenaw fault, but here the fault lies more than a mile to the south, and the Portage Lake Volcanics underlie the lowlands surrounding the north end of Lake Gogebic as well as the highlands north of the scarp.

The Portage Lake Volcanics are a thick pile of dominantly basaltic, subaerial flows, with lesser rhyolites and interflow sedimentary rocks. At this locality, however, rhyolite is the dominant rock type, although good examples of basalt are in the roadcut. Another unusual feature here is the local development of pillows along the toe of a flow where it apparently entered a body of water in which sandstone was being deposited.

The stratigraphy in the quarry area consists of an ophitic basalt flow overlain by a thin interflow sandstone to siltstone which in turn is overlain by an amygdaloidal basalt. The amygdaloidal basalt exhibits well developed pipe vesicles at road level. Rhyolite overlies the amygdaloidal basalt and is well exposed in the quarry floor and walls and is overlain by ophitic basalt flows on the hillside above the quarry. The basalts here are identical compositionally to low TiO_2 basalts on Keweenaw Peninsula (table 1).

It is not clear whether the rhyolite is a single shallow intrusive body or a series of flows. It contains several lenticular masses of basalt and an unusual greenish sandstone. Are these xenoliths in an intrusion or discontinuous interbeds? Several features suggestive of extrusive origin can be noted. Folds are present near the contact of the rhyolite with the underlying basalt. In the upper part of the quarry, the rhyolite shows well-developed columnar jointing, some of which is bent, following a ramp structure in the rhyolite. The possibility of a series of flows is enhanced by the distribution of spherulitic rhyolite. Spherulites are common throughout the quarry but are concentrated near the base of the rhyolite, near the contact with a sedimentary lens, and near the top of the quarry. More massive or stony rhyolite occurs between the spherulitic zones.

Alteration related to copper mineralization is widespread in the quarry. Epidote is disseminated in much of the rhyolite, giving it an unusual greenish color. Secondary copper minerals are abundant on joint surfaces.

Stop 5.--Porcupine Volcanics: Rhyolite at Summit Peak

Summit Peak is the highest point in Porcupine Mountains Park and one of the highest points in the state. The observation tower at the summit provides a panoramic view of the field trip area (fig. 6). To the south, the highlands are underlain by the Portage Lake Volcanics and Porcupine Volcanics along the main monocline of volcanic rocks of the Keweenawan Supergroup. The lowlands immediately to the southeast are underlain by rocks of the Oronto Group in the east-plunging Iron River syncline. Looking east along strike, the smelter at White Pine can be seen in the distance. To the north, the interior of the park extends over the rugged topography in the foreground to Lake Superior in the distance. The interior of the park is maintained as a wilderness area and access is only by hiking. It contains the largest stand of virgin timber in the state.

Most of the park interior is underlain by a thick unit of rhyolite composed of a series of flows and domes typified by the rocks seen at Summit peak. Good exposures of coarse rhyolite breccia are along the trail leading to the summit and at the overlook platform west of the summit. This breccia is probably the carapace of a rhyolite dome.

Excellent exposures of typical intermediate and felsic units of the Porcupine Volcanics can be seen on the north side of the hill (549 m elevation) near the center of Sec. 31. Take the Beaver Creek Trail about 0.5 mi from the Summit Peak parking lot. The units dip to the south and include, in stratigraphic order, sparse outcrops of intermediate to mafic rocks in the creek bed, overlain by a vesicular siliceous andesite, which is in turn overlain by a coarse rhyolite breccia or debris flow. The breccia contains clasts ranging in size from nearly a meter to less than 1 cm. The breccia is clast supported and some clasts are subrounded. Overlying the breccia is a medium-grained basalt flow. Capping the hill, and overlying the basalt, is an aphanitic massive rhyolite that is microspherulitic.

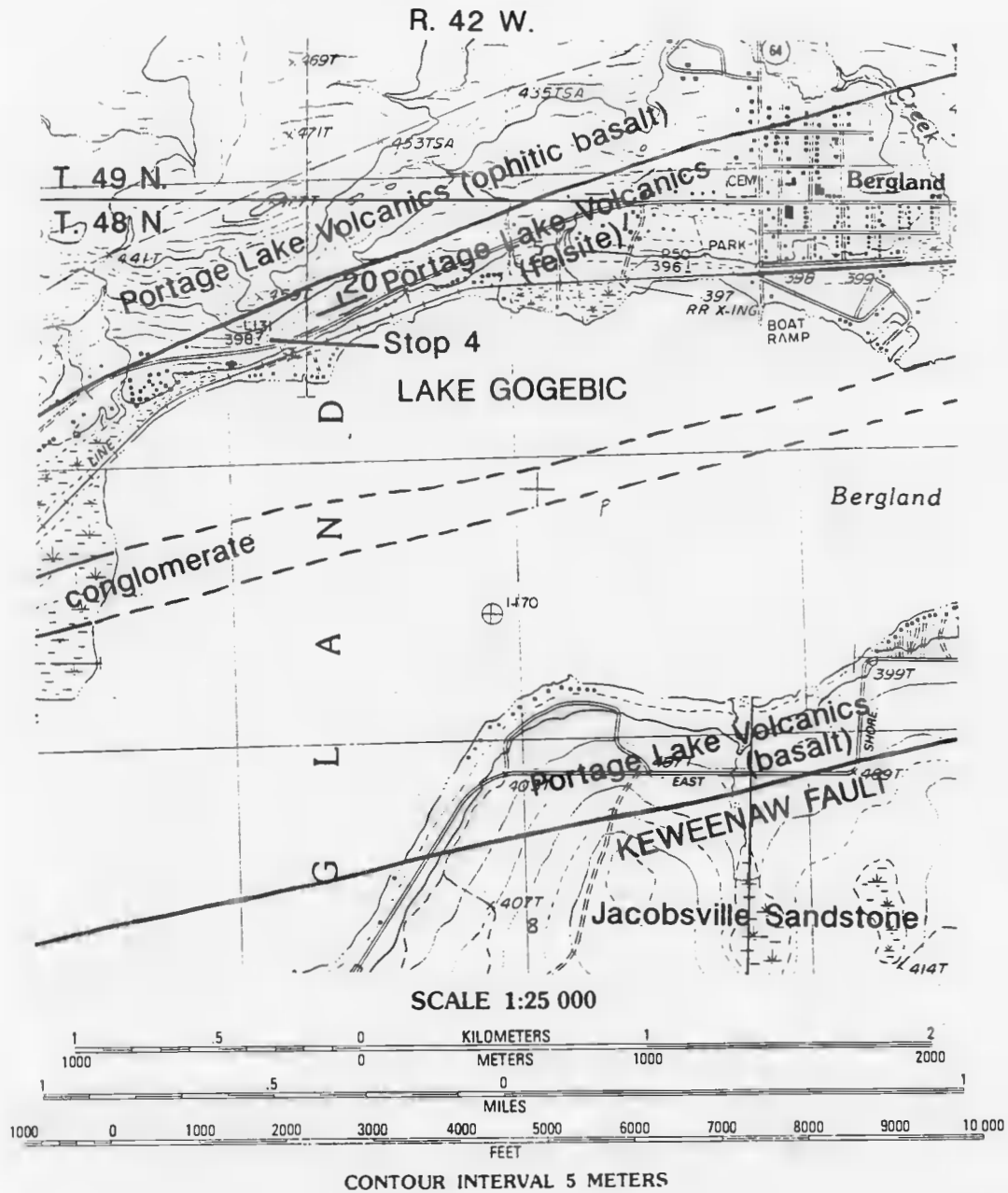


Figure 5. Location and geologic setting of stop 4. Geology generalized from unpublished compilation by Walter White.

Stop 6.--Porcupine Volcanics: Rhyolite quarry near White Pine

The rocks exposed here are part of a small rhyolite body near the top of the Porcupine Volcanics. The body probably does not extend much beyond the hill into which the quarry is cut (fig. 7). The quarry provides a cross section through part of a subaerial agglutinate deposit. Agglutinate deposits form at vents by spatter of erupting magma and buildup of mounds of hot, viscous material. The mound of erupted material eventually flows outward under its own weight resulting in large flow folds such as seen in this quarry. The near-vent nature of this deposit is deduced from the presence of lithic fragments within the rhyolite, large flow folds and stratification of rhyolite (light and dark units). At the edges of agglutinate deposits, flowage typically has homogenized the magma and such stratification and folding are generally not preserved.

The rhyolite here contains feldspar phenocrysts, which are typically aligned parallel to the stratification and foliation. It is enriched substantially in such incompatible trace elements as Zr, Ba, Hf, and the light rare earth elements compared to the rhyolite in the Portage Lake Volcanics near Bergland (table 1). These differences suggest different petrogenesis and sources for the two rhyolites in spite of their physical proximity.

Stop 7.--Basalt flows within the Copper Harbor Conglomerate (Lake of the Clouds overlook)

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

The overlook is along the south escarpment of a high ridge supported by a series of north-dipping lava flows within the Copper Harbor Conglomerate (fig. 8). The low area south of the ridge, including Lake of the Clouds, is underlain by sandstone and siltstone and a few basalt flows which constitute the lower part of the Copper Harbor Conglomerate. The higher regions farther south are underlain by volcanic rocks, mostly rhyolite of the Porcupine Volcanics.

Toward the east end of the overlook area, a large glaciated surface shows a series of thin basalt flows, which average a few meters thick. Individual flows can be readily identified by chilled vesicular bases, in places containing inclusions of older flows, and by rubbly or vesicular tops. Abundant epidote alteration and vesicle fillings also impart a distinctive greenish cast to flow margins. Hubbard (1975b) described the flows in the Copper Harbor as mostly andesite with minor basalt, but chemical analyses of two samples from this locality (table 1) indicate that they are basalt similar to an average basalt from the Porcupine Volcanics. Compared to Portage Lake Volcanics, these basalts are enriched in incompatible trace elements.

Stop 8.--Upper part of Copper Harbor Conglomerate at Union Bay Campground

Good exposures of reddish sandstone containing thin conglomerate beds are abundant along the shore of Lake Superior at this locality (see fig. 9). The Copper Harbor Conglomerate is noted for coarse volcanogenic conglomerates, which form most of the lower part of the section throughout much of its outcrop belt and which grade up into finer grained sandstones. But, north of the Porcupine Mountains there is a different facies relationship. The lower part of the formation is mostly sandstone, siltstone and lava flows; conglomerate is very subordinate. These rocks underlie the high hills immediately south of the highway. The coarse conglomerate facies is less abundant and is higher in the section. The exposures here at Union Bay are near the base of the upper unit and are probably about 1,000 m above the base of the formation. The sandstones at Union Bay dip 10-20° to the north. They are volcanogenic and quartz-poor. They show excellent examples of trough cross bedding, generally indicating a northeastward current vector, and a variety of other sedimentary features including desiccation cracks, rip-up clasts, oscillation and current ripples, and swash marks.

The exposures of Copper Harbor Conglomerate north of the Porcupine Mountains are the farthest removed from the source highlands to the south of any part of the unit on land. The relative scarcity of thick units of coarse conglomerate compared to exposures farther south probably

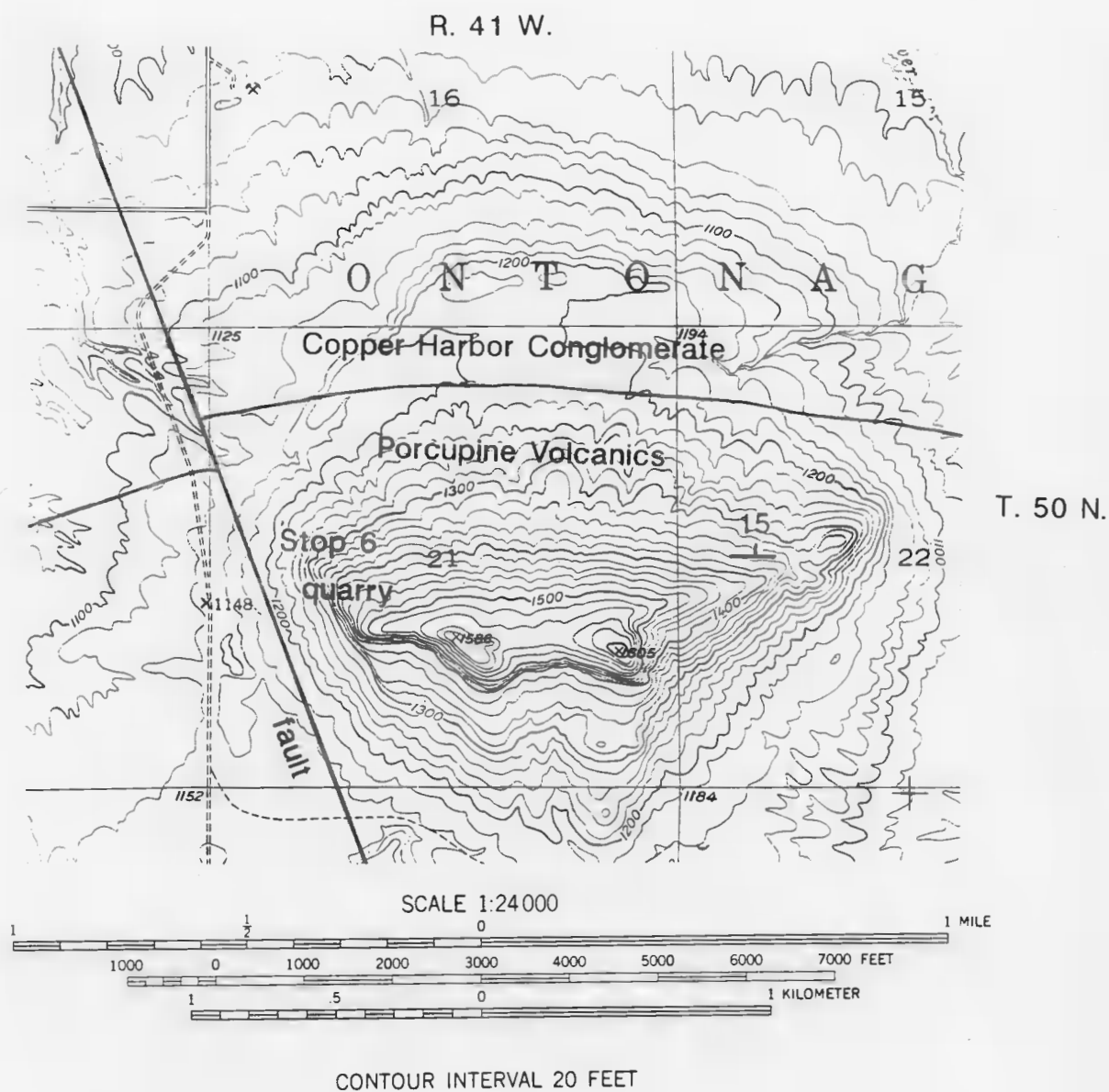


Figure 7. Location and geologic setting of stop 6. Geology generalized from Johnson and White (1969).

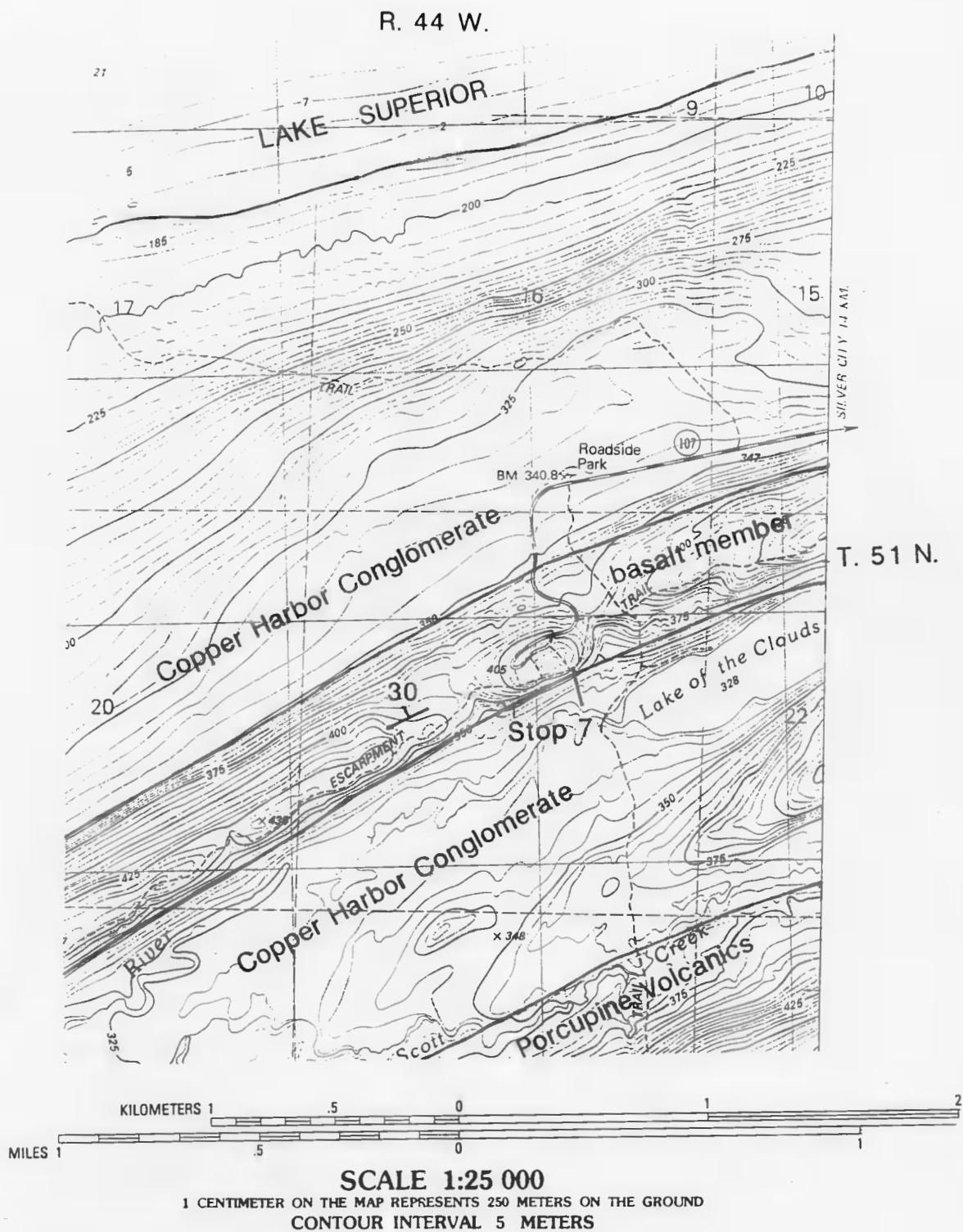


Figure 8. Location and geologic setting of stop 7. Geology generalized from Hubbard (1975b).

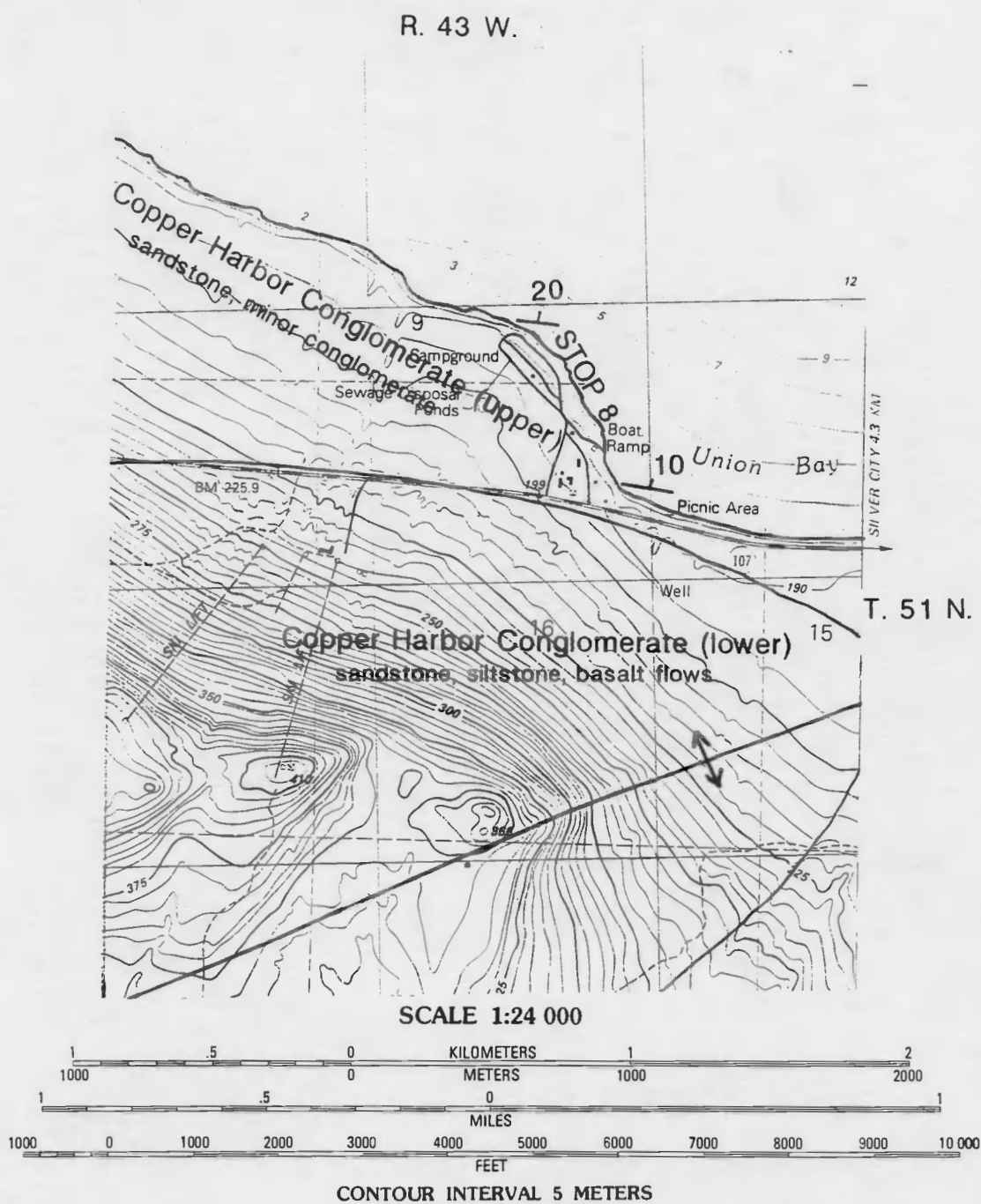


Figure 9. Location and geologic setting of stop 8. Geology generalized from Hubbard (1975b).

reflects the distal nature of these rocks and may be a good representative of much of the Copper Harbor beneath Lake Superior. The unit here shows a crude coarsening-upward trend as opposed to the fining-upward trend typical of the more proximal parts of the unit. This relationship is consistent with northward prograding alluvial plain deposition.

Several large boulders are distributed along the beach and are composed of conglomerate typical of the lower part of the formation elsewhere. These conglomerates contain almost exclusively clasts of Keweenawan Supergroup volcanic rocks types common in the region. An interesting question is the source of these boulders. Although the Copper Harbor does contain some conglomerate beds of this aspect nearby, none of the streams entering Lake Superior near here seem capable of transporting such large boulders. Most have low gradients, especially near the lakeshore, and the streambeds do not contain such large boulders. No outcrops of coarse conglomerate would be expected nearby to the north beneath the lake; so if the boulders are glacial erratics, they must have been transported for a long distance, perhaps from the vicinity of Isle Royale.

Stop 9.--Nonesuch Shale at Bonanza Falls

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

The Nonesuch Shale is distinguished from other sedimentary units of the Keweenawan Supergroup by the predominance of gray, green or black, fine-grained sediments. The Big Iron River section has a predominance of siltstones and fine sandstones over true shales. Many rocks show trough cross bedding, symmetrical and asymmetrical ripples, rib and furrow structures, and parting lineations. The finer-grained rocks include well-laminated shales, which are most abundant lower in the section. The shaley units commonly have ball and pillow structures and calcareous concretions.

The Nonesuch displays coarsening-upward sequences at scales ranging from a few meters to the entire thickness of the unit. On a smaller scale, normally graded sequences are common in units from a few centimeters to a few meters thick.

In the lower 10 m of the section copper mineralization can be seen as concentrations of chalcocite, bornite and malachite along bedding planes. The mineralization is cogenetic with the major copper mineralization at the White Pine Mine where the downdip extension of this unit is mined just to the south and east.

A good exposure of the mineralized base of the Nonesuch Shale and the top of the Copper Harbor Conglomerate is along the Little Iron River near the center of the SW 1/4, Sec. 13, but requires a walk of about 1 mi south from Highway 107. It is an easy walk along an unmaintained trail on the east bank of the river for those who can spend more time in the area. Remains of early mining efforts for native silver can be seen there, as well as "ore" specimens from old dumps.

White Pine Mine

The trip will not visit the White Pine Mine; but because of the importance of this orebody, the following brief summary is provided.

White Pine recovers copper and silver from a very large strataform orebody in the base of the Nonesuch Shale and in places from the upper few meters of the Copper Harbor Conglomerate. Excellent summaries of the geology and origin of the orebody are provided by White and Wright (1966), Ensign and others (1968), White (1971), and Brown (1971). Ore, composed mostly of chalcocite and lesser native copper, grades about 1.1% Cu and 9 g/t Ag over a mining height typically about 5 m. Some beds within that interval consistently contain greater than 3% Cu. Reserves are about 200,000,000 t of extractable ore. Mining of the gently dipping orebody is by room and pillar method. The mine workings underlie about 35 km² of the low country south and east of the town of White Pine.

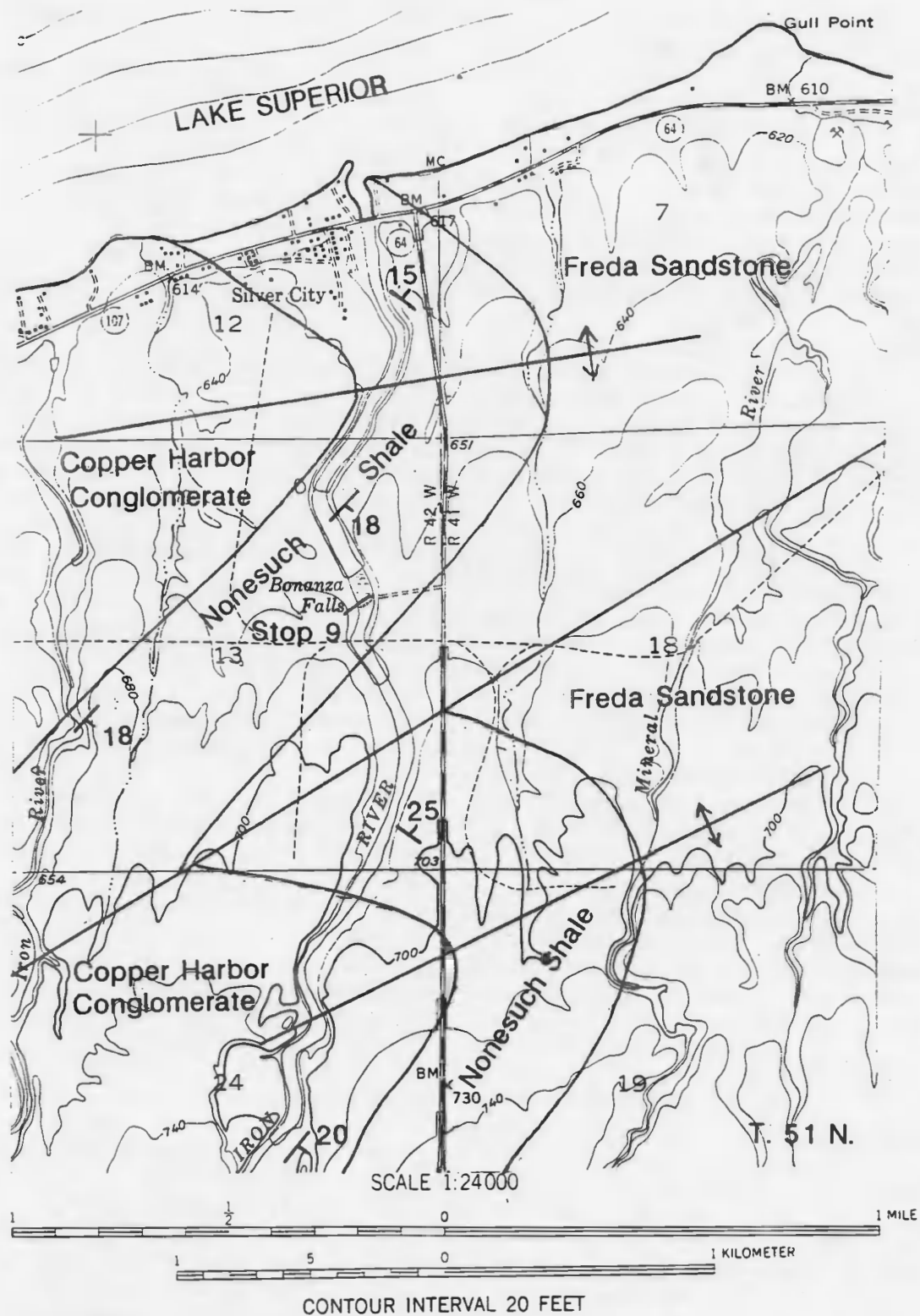


Figure 10. Location and geologic setting of stop 9. Geology generalized from Copper Range Corp./Michigan Geological Survey cooperative mapping (unpublished).

STRATIGRAPHIC SECTION OF NONESUCH SHALE AT BONANZA FALLS
Generalized from Suszek (1991)

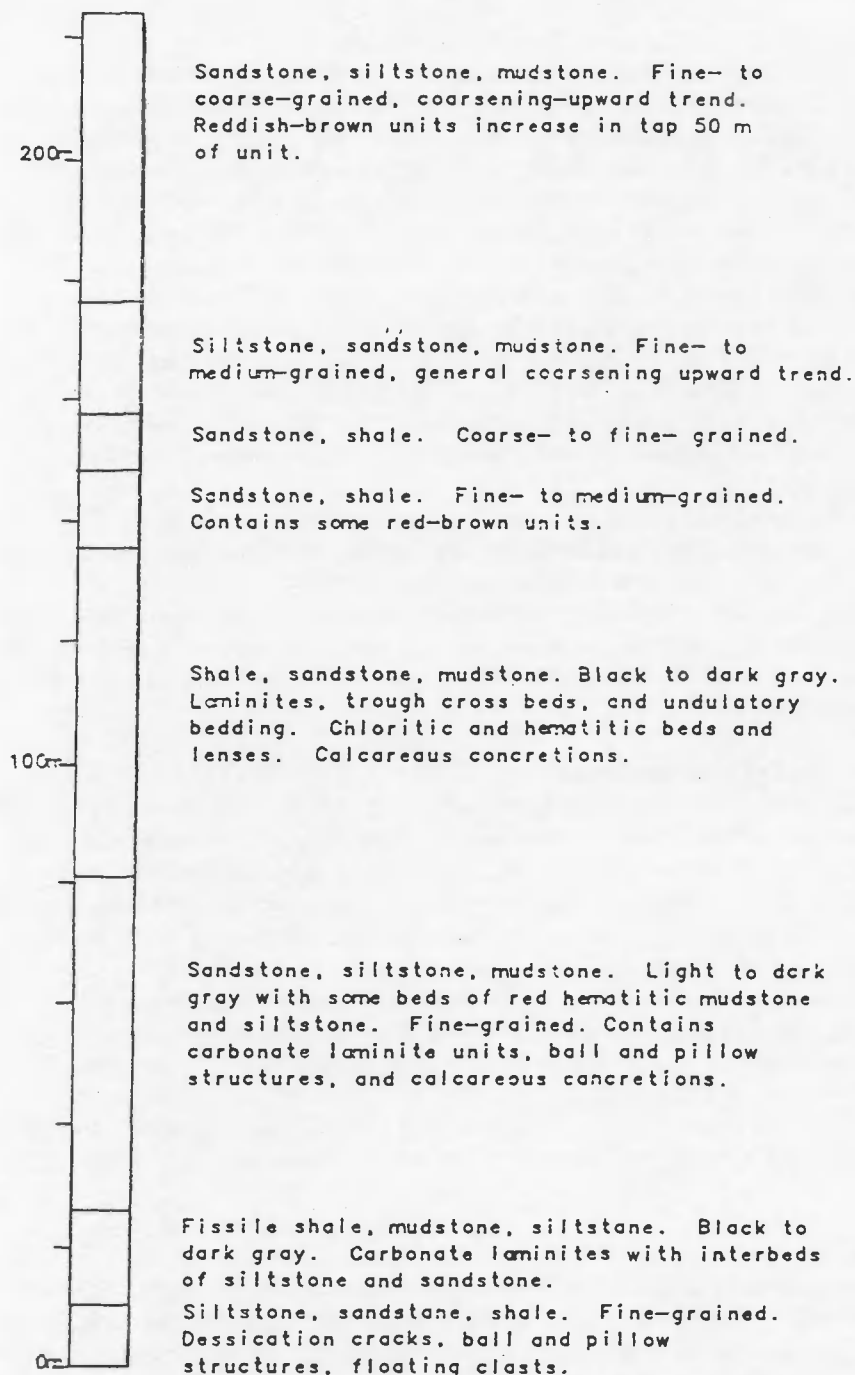


Figure 11. Measured stratigraphic section of Nonesuch Shale along Big Iron River near Bonanza Falls, generalized from Suszek (1991).

Copper was introduced to the Nonesuch Shale mostly during early diagenesis, probably by upward circulating connate water which dissolved copper from the underlying redbeds. Chalcocite, largely of submicroscopic size, formed by the replacement of diagenetic pyrite. A later phase of copper mineralization, documented by Mauk and others (1989) and Mauk and others (in press), introduced native copper, mostly in structurally disturbed zones. This second stage mineralization is probably cogenetic with the classic native copper mineralization of the Keweenaw Supergroup basalts.

Stop 10.--Nonesuch Shale and Freda Sandstone at Presque Isle River

The upper portion of the Nonesuch Shale and the base of the Freda Sandstone are well exposed in the gorge of the Presque Isle River near its mouth and along the shore of Lake Superior west of the river (fig. 12). Continuous exposures along the picturesque gorge of the river extend from just upstream of Nawadaha Falls to the lakeshore. Exposures continue in bluffs along the lakeshore for about half a mile west of the river mouth. In the interest of time, we will limit our examination to exposures near the river mouth and a short distance to the west along the shore. This requires a round trip hike of nearly a mile, mostly on well-maintained trails and stairways.

The rocks exposed here are on the northeast limb of the Presque Isle syncline, a gentle northwest-plunging fold. Dips range from nearly flat to about 10° SW. The Nonesuch Shale here is very similar to that described for stop 9 on the Big Iron River about 30 km to the east. The lower part of the section in this area (not exposed) is also strongly mineralized with copper and contains a fine-scale stratigraphy directly correlatable with the stratigraphy at the White Pine mine, indicating that sedimentary conditions were very uniform over the entire region surrounding the Porcupine Mountains during Nonesuch deposition.

The upper part of the Nonesuch exposed here, as described in more detail by Suszek (1991), consists of fining-upward sequences and laminites interbedded with graded sandstones and siltstones. Most rocks are dark gray with a few reddish lenses and beds.

The Nonesuch grades upward to the Freda Sandstone through a zone in which dark gray laminated and small-scale cross-bedded siltstone and sandy mudstone is interbedded with medium- to coarse-grained reddish brown sandstone. The upper contact of the Nonesuch is placed where the reddish sandstone becomes dominant.

Stop 11.-- Freda Sandstone along Presque Isle River

This stop, near the axis of the Presque Isle syncline (fig. 12), shows reddish cross bedded sandstone typical of the lower part of the Freda Sandstone. The gently southwest dipping beds exposed here are probably 100-200 m above the base of the formation and slightly higher stratigraphically than those seen at stop 10. They are mostly lithic sandstone and are somewhat micaceous in places. The Freda marks the return to fluvial redbed deposition following the lacustrine deposition of the underlying Nonesuch Shale.

The Freda is a very thick unit in much of the rift in the Lake Superior region and is volumetrically the dominant unit of the post-rift sedimentary fill. Along the Montreal River, about 30 km to the west, about 4,000 m of Freda are exposed. Seismic sections indicate that it is even thicker beneath the Lake.

The Freda Sandstone generally becomes finer-grained and more mature upward. Excellent exposures of the upper part of the Freda can be seen near the mouth of the Montreal River, in bluffs along the lakeshore.

Stop 12.-- Jacobsville Sandstone

The roadcut along the west side of the road is a fairly typical exposure of the conglomerate found abundantly in the Jacobsville Sandstone in this region. Although exposures are rare, west of Lake Gogebic the Jacobsville appears to be dominated by conglomerate and coarse sandstone in contrast to areas farther east where it is a submature to mature quartzose, feldspathic, and lithic arenite with minor shale, siltstone, and conglomerate layers (Kalliokoski, 1982).

At this stop the conglomerate strikes east and dips 9° to the north, but only about 500 m to the north it dips steeply to the south (62°) as a result of drag along the Keweenaw fault. The

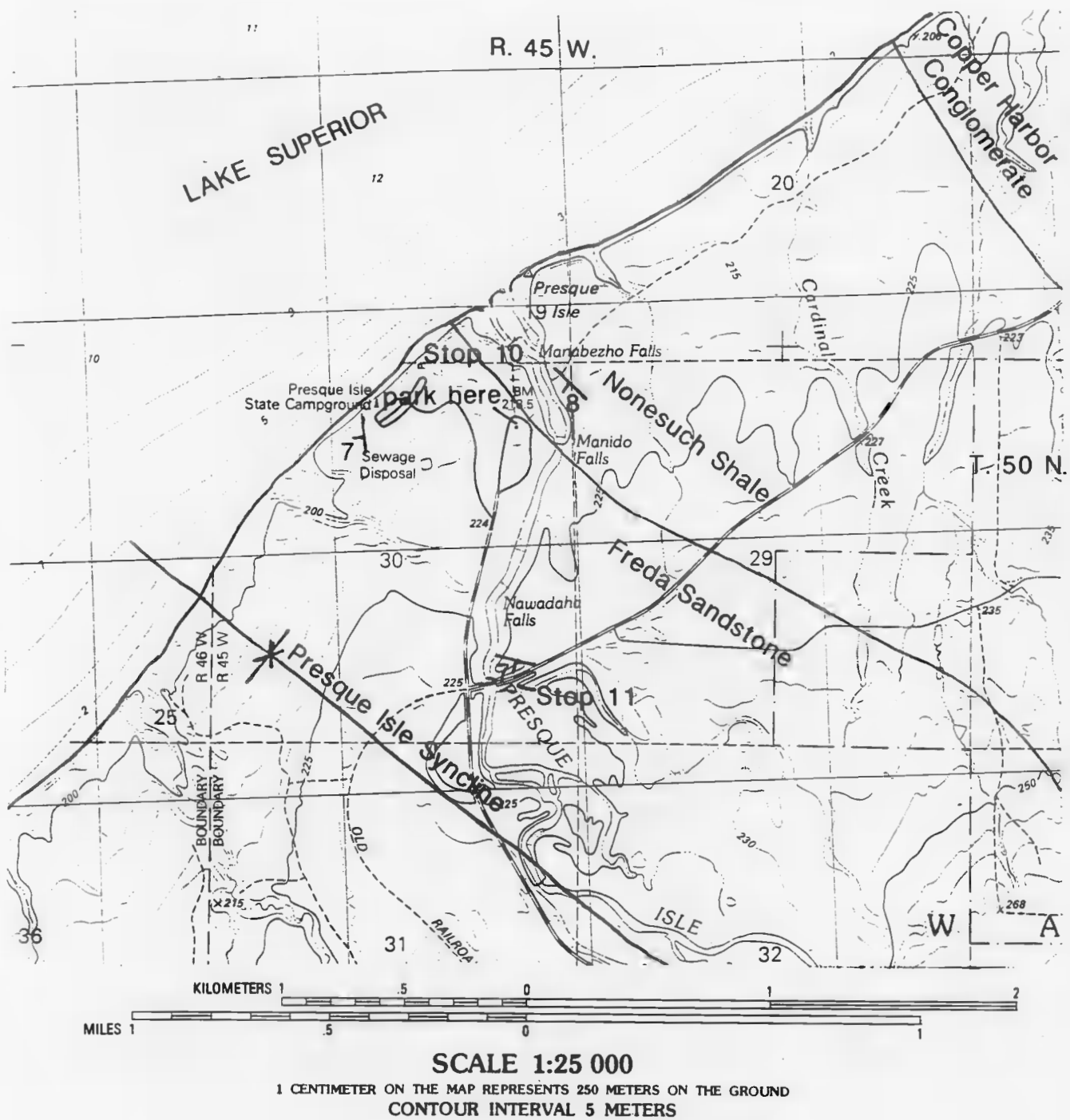


Figure 12. Location and geologic setting of stops 10 and 11. Geology generalized from Copper Range Corp./Michigan Geological Survey cooperative mapping (unpublished).

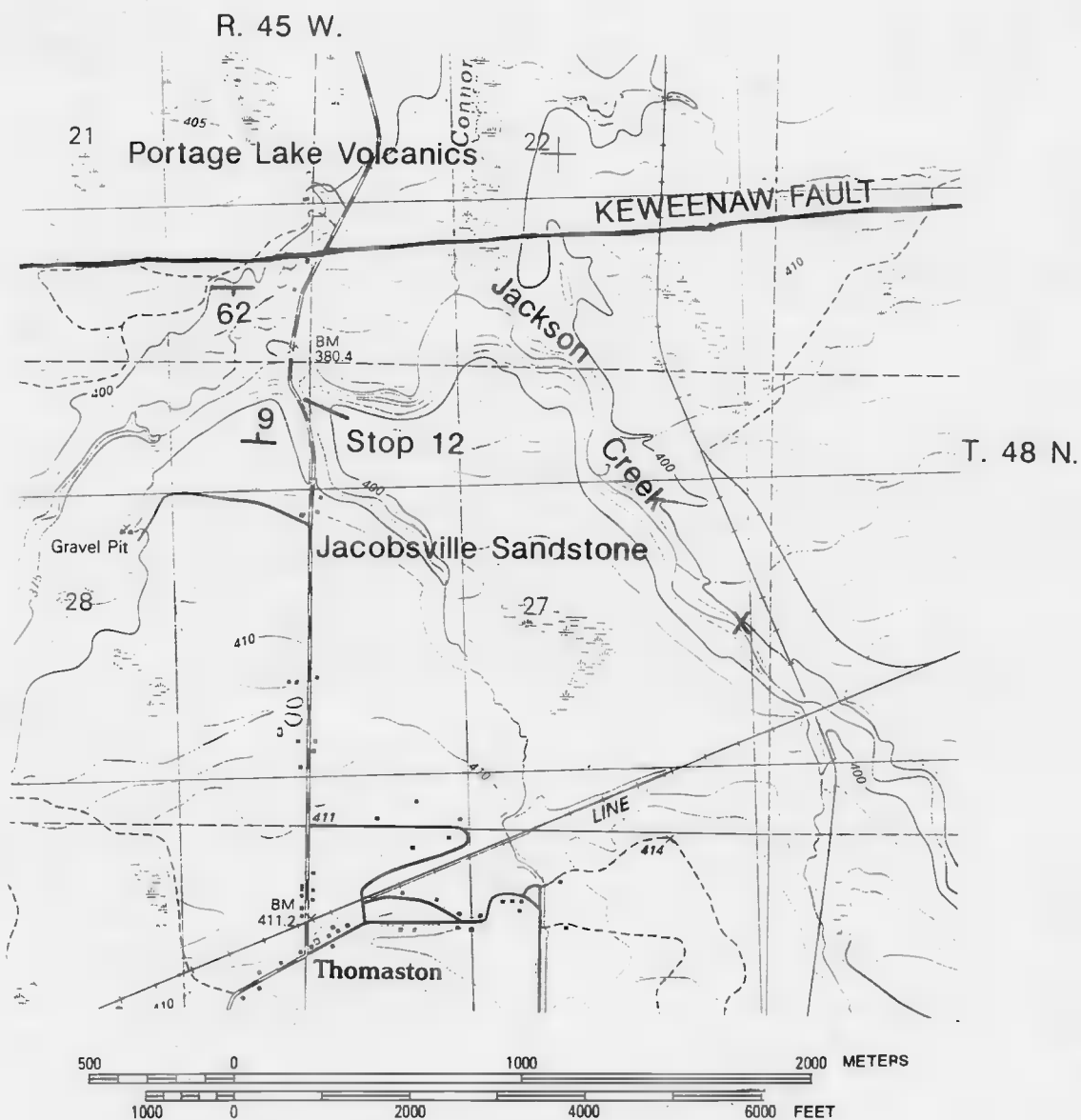


Figure 13. Location and geologic setting of stop 12. Geology generalized from Copper Range Corp./Michigan Geological Survey cooperative mapping (unpublished).

conglomerate is poorly consolidated and weathers to a material that superficially resembles glacial till or outwash, which covers most of the roadcut. From time to time parts of this material slump away to expose less weathered conglomerate, and small exposures of fresh conglomerate are usually exposed in the drainage ditch beside the road.

The conglomerate is clast-supported. Coarse sandy matrix surrounds pebbles. Where weathered, it is generally unconsolidated but on fresher surfaces is fairly well cemented. Clay fills most interstices and coats all grains. It is uncertain whether the clay is a product of diagenetic cement or is an early mechanical infiltration resulting from seepage of muddy water through coarse alluvium.

Clasts are subrounded to well rounded, range in size from 3-10 cm, and consist mostly of iron-formation (63-77%) and quartzite (8-20%) derived from Early Proterozoic rocks such as those now exposed in the Gogebic Range. Small amounts of altered rhyolite (1-5%) are believed to be derived from Keweenawan Supergroup rocks. Vein quartz (9-10%), chert (1%), and other lithologies (1% or less) derived from Archean rocks are minor. Several other conglomerate outcrops to the west have similar compositions, but outcrops less than 2 km to the southeast along Jackson Creek (shown by X on the map) are dominated by vein quartz (73-82%) with lesser amounts of quartzite (11-13%) and iron-formation (9-12%). These clast lithologies suggest that most of the source area to the south had been stripped of Keweenawan volcanics by the time of Jacobsville deposition. Although the surface consisted mostly of Early Proterozoic sedimentary rocks, the presence of large quantities of vein quartz pebbles indicates at least some drainages were eroded down into Archean crystalline rocks.

In a few places clasts found in place appear to be imbricated but otherwise bedding within the conglomerate units is not apparent. Clay and iron coatings cover nearly all clasts so that the various lithologic types are not apparent unless clasts are broken open. Many clasts have a highly shiny surface or patina that looks remarkably like desert varnish and some have distinctly striated surfaces of unknown origin. A few clasts have been found that have shapes and surface textures resembling ventifacts.

In the streambanks east of the road, slightly higher in the section, the conglomerate is inter-layered with coarse sandstone. The sandstone layers are poorly sorted, discontinuous irregular layers approximately 5-30 cm thick. The sandstone consists of subangular to subrounded, medium to coarse sand (.25-2 mm) and gravel (2-30 mm) with larger cobbles scattered throughout. In thin section the sand is framework supported and composed predominantly of quartz and feldspar with minor lithic fragments, biotite, and muscovite. Pores are filled with clay and many grains are coated with iron minerals.

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**GEOLOGY OF THE
GREAT LAKES
TECTONIC ZONE
IN THE
MARQUETTE AREA,
MICHIGAN—
A LATE ARCHEAN
PALEOSUTURE**

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Denver, Colorado

Guide to the geology of the Great Lakes tectonic zone in the Marquette area, Michigan--A Late Archean paleosuture

P.K. Sims and Z.E. Peterman

SUMMARY

The Great Lakes tectonic zone (GLTZ) is an Archean crustal boundary of subcontinental length that separates a greenstone-granite terrane (southern part of Superior province of Canadian Shield) on the north from a partly older gneiss terrane on the south. The GLTZ is generally interpreted as a paleosuture resulting from continent-continent collision. The tectonic zone is covered in most of the Lake Superior region by Proterozoic rocks or Pleistocene glacial deposits, and its position and characteristics were previously determined mainly by geophysical data. Geologic mapping in the Marquette, Michigan area provides for the first time direct observations of the structure.

In the Marquette area, the GLTZ is characterized by a zone of mylonite (orthomylonite) that formed at brittle-ductile transition conditions; this was superposed on previously deformed rocks of both the Archean greenstone-granite terrane and the Archean gneiss terrane. The boundaries of the GLTZ trend about N. 60° W. Foliation in the mylonite strikes about N. 70° W. and dips steeply southwest. A pronounced stretching lineation and tight fold hinges plunge about 42° S. 43° E. The attitude of the stretching lineation (line of tectonic transport), together with asymmetric structures indicative of movement sense, indicate that collision at this locality was oblique. This resulted in dextral-thrust shear along the boundary, northwestward vergence, and probable overriding of the greenstone-granite terrane by the gneiss terrane. Transmittal of the dextral shear stress across a large area of the greenstone-granite crust (Superior province) to the north may have formed the nearly east-west foliation, upright folds, and northwest- to east-west-trending dextral faults and shear zones at least as far north as the Quetico fault, in southern Ontario, a distance of about 250 km. These structures in Superior province rocks are superposed on older recumbent folds.

As a whole, the GLTZ is characterized by systematic angular bends that alternately trend west-northwestward, as in the Marquette area, and northeastward. This zigzag pattern probably reflects relict irregularities in the continental margin (Superior province) composed of greenstone-granite crust. Late Archean convergence along this margin resulted in a variable trajectory of stress into the greenstone-granite crust and probably in along-strike diachroneity of orogeny. The major deformation resulted from oblique compression at promontories, which acted as buttresses against which compressive stress was directed into the crust. In addition to the dominant foliation, major brittle-ductile to brittle strike-slip faults also formed, such as the Vermilion fault system in northern Minnesota and the Quetico and Rainy Lake-Seine River faults in southern Ontario; these resulted from a more brittle continuum of the transcurrent shear caused by collision along the GLTZ.

INTRODUCTION

The Great Lakes tectonic zone (GLTZ) is an Archean crustal boundary more than 1,000 km long that separates a greenstone-granite terrane (southern part of Superior province) on the north from a gneiss terrane on the south (Sims and others, 1980; Sims and Peterman, 1981; Peterman, 1979). The GLTZ is covered throughout most of the Great Lakes region by younger

Proterozoic rocks or Pleistocene glacial deposits, but recently it has been delineated and studied in outcrop in an area south of Marquette, Michigan (fig. 1).

The boundary was first recognized in Minnesota (Sims and Morey, 1973; Morey and Sims, 1976) from regional geologic relations, which indicated that the two adjoining basement terranes had different geologic histories and probably had evolved separately. Regional magnetic and gravity data were utilized to determine the position of the boundary. Later (Sims, 1980), the boundary was approximately delineated in the western part of Upper Michigan (Sims, 1980; Sims and others, 1984) and northwestern Wisconsin (Sims and others, 1985), east of the Middle Proterozoic Midcontinent rift system, and it was inferred by indirect evidence to extend eastward through the Sudbury structure, where it is truncated by the Middle Proterozoic Grenville tectonic zone (Sims and others, 1980).

Recent geologic mapping in the Sands and Palmer 7½-minute quadrangles, Michigan (fig. 1), previously mapped by Gair and Thaden (1968) and Gair (1975), has delineated this Archean boundary in outcrop for the first time (Sims, 1991). It is exposed on the south side of the Early Proterozoic Marquette synclinorium, and its northwestern projection into the trough coincides with a major Early Proterozoic fault, the Richmond fault. In the Marquette area, the GLTZ is a mylonite zone about 2 km wide that is overprinted on rocks of both the greenstone-granite terrane and the Archean gneiss terrane. In this area, the GLTZ is interpreted as a continent-continent collision zone. The collision was oblique, resulting in dextral wrench shear on the N. 60° W.-trending boundary and northwestward vergence of the gneiss terrane against the greenstone-granite terrane.

The purpose of this field guide is to examine the exposed GLTZ in the context of the regional geology, to discuss genetic relationships between convergence along the boundary and structural features in the Archean rocks to the north, and to present a new interpretation of the evolution of the GLTZ throughout the Lake Superior region.

GEOLOGIC SETTING

The Great Lakes tectonic zone (GLTZ) is moderately well exposed in the Sands and Palmer 7½-minute quadrangles in Upper Michigan (fig. 1). It separates the two distinctive Archean terranes in the area. The northern greenstone-granite terrane is composed largely of Late Archean granitoid rocks, with subordinate approximately coeval metavolcanic and metasedimentary rocks of greenstone affinity. The layered rocks and most of the granitoid rocks were metamorphosed (mainly to greenschist facies) and deformed during Late Archean orogeny. The southern Archean gneiss terrane is composed mainly of layered gneiss, migmatite, and amphibolite--rocks that are distinct from those in the greenstone-granite terrane. Except for late-tectonic to post-tectonic, generally small(?) granitoid bodies, the rocks of the terrane are metamorphosed, mainly to amphibolite facies. The rocks exposed within the two terranes in Upper Michigan are closely similar to those in Minnesota (Morey and Sims, 1976; Sims and others, 1980), thus establishing the identity of the GLTZ in the Marquette area.

The Archean rocks in Michigan are overlain in the Marquette synclinorium, the Republic trough, and the Dead River, Clark Creek, and Baraga basins by shelf and foredeep deposits of the Early Proterozoic Marquette Range Supergroup (fig. 1; Cannon and Gair, 1970; Barovich and others, 1989).

Van Hise and Bayley (1897) introduced the names "northern complex" for the Archean rocks north of the Marquette synclinorium and "southern complex" for the Archean rocks south of the synclinorium.

A Late Archean age for the GLTZ is now established by regional geologic relationships in north-central United States. The Archean rocks in the greenstone-granite terrane in northern Minnesota (Hudleston and others, 1988) and northernmost Michigan (fig. 1; Sims, 1991) are characterized mainly by ductile and brittle structures formed in response to dextral shear, which accords with the deformation pattern observed in exposures of the GLTZ south of Marquette. These structures include a generally west-trending steep foliation and upright folds, widespread Z-shaped folds, and northwest- to west-trending dextral faults, indicative of dextral shear; the structures are superposed on recumbent folds (Johnson and Bornhorst, 1991). In contrast, Early Proterozoic Penokean deformation in the Marquette area had little effect on the Archean basement (Cambray, 1984). Cambray (1984) proposed that the Penokean deformation was produced by horizontal compression that was transmitted from the basement to the folded cover rocks by narrow ductile shears in the basement. Readjustment of rigid basement blocks along these shears utilized old weaknesses, which resulted in some shortening but not folding of the basement rocks. For the Early Proterozoic Marquette synclinorium, Cambray (1984) proposed a nearly north-south compressional axis, which initially produced reverse dip slip on faults bounding the syncline and compressed the Early Proterozoic sedimentary rocks within the syncline into west-trending folds. Subsequently, resistance to this movement within the trough resulted in sinistral strike slip motion and the development of F_2 folds with northwest-trending axial surfaces and variable plunge.

Gair and Thaden (1968) applied the name "Compeau Creek Gneiss" to both the foliated granitoid rocks of the Archean greenstone-granite terrane (northern complex) and the layered gneisses and massive intrusions of the gneiss terrane (southern complex), and subsequent investigators extended this terminology to the western parts of the southern complex (Cannon and Simmons, 1973). To apply this name to rocks in both Archean terranes, however, is inappropriate, because the two terranes consist of distinctive rock types of different origins. Accordingly, in this guide informal lithologic names are used to describe the crystalline rocks in the two terranes.

ARCHEAN GREENSTONE-GRANITE TERRANE

The greenstone-granite terrane in the Marquette area (fig. 1) consists of a several thousand-meter-thick succession of metamorphosed subaqueous mafic to felsic flows, pyroclastic rocks, and volcanogenic sedimentary rocks that is intruded by small bodies of gabbro and ultramafic rocks and by large plutons of granitoid rocks (Bornhorst, 1988; Johnson and Bornhorst, 1991). The metamorphosed volcanic rocks have been named the Ishpeming greenstone belt (Morgan and DeCristoforo, 1980); they represent the southwestern extension of the Wawa (Shebandowan) subprovince of the Superior province of Canada (Card and Ciesielski, 1986). Felsic volcanic rocks adjacent to an ultramafic body north of Ishpeming are host to the gold deposits of the Ropes mine (Brozdowski and others, 1986; Brozdowski, 1988, 1989), and the ultramafic body hosts additional mineral prospects (Bodwell, 1972). Foliated tonalite from the northern complex (Hammond, 1978) in the greenstone-granite terrane, has a U-Pb zircon age of $2,703 \pm 16$ Ma (recalculated by Zell E. Peterman), and associated rhyolite has a U-Pb zircon age of $2,780 \pm 69$ Ma (recalculated by Zell E. Peterman). These ages are consistent with more precise U-Pb ages in the Wawa subprovince in adjacent Canada (Corfu and Stott, 1986).

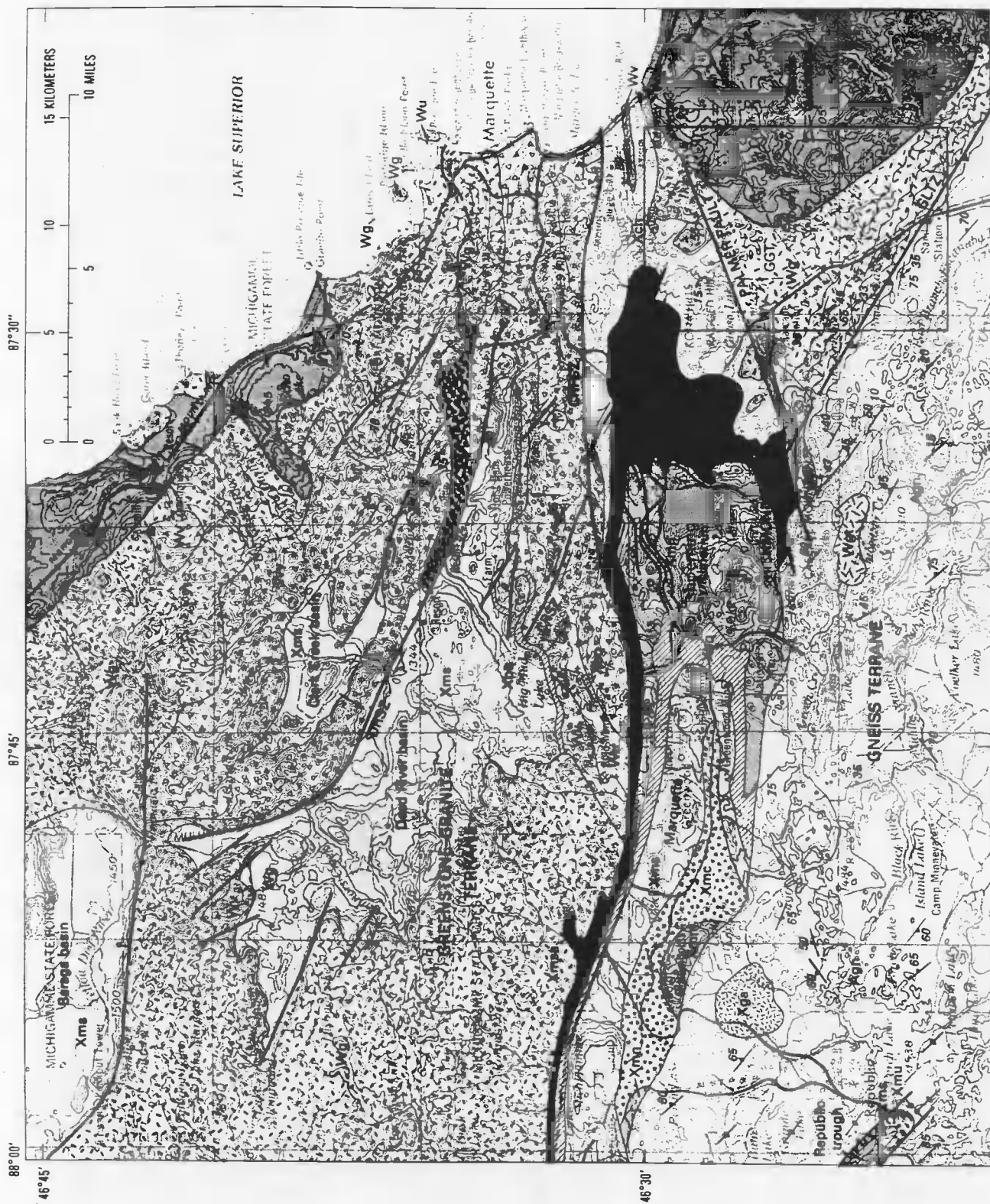
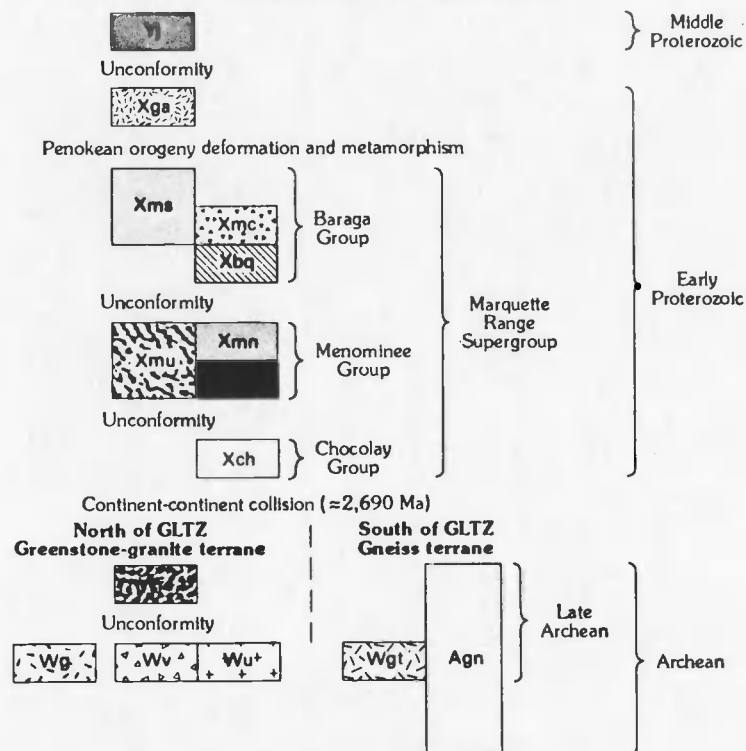


Figure 1. Geologic map of part of Marquette 1°2' quadrangle, Michigan. After Sims (1991). Palmer 7½-minute quadrangle is immediately west of Sands quadrangle. Marquette 7½-minute quadrangle is immediately north of Sands quadrangle.

CORRELATION OF MAP UNITS



DESCRIPTION OF MAP UNITS

	Middle Proterozoic Jacobsville Sandstone
	Early Proterozoic Alkali granite ($1,733 \pm 25$ Ma)
	Marquette Range Supergroup
	Michigamme Formation—Dominantly slaty rocks in lower part
	Clarksburg Volcanics Member of Michigamme Formation
	Goodrich Quartzite
	Menominee Group, undivided, in Republic trough
	Negaunee Iron-formation
	Siamo Slate and Ajibik Quartzite, undivided
	Chocolay Group, undivided

Archean

Greenstone-granite terrane

Reany Creek Formation

Granitoid rocks—Dominantly granodiorite and tonalite but including granite; generally foliated

Mafic to felsic flows and pyroclastic rocks and volcanogenic sedimentary rocks

Ultramafic rocks

Gneiss terrane

Granite near Tilden (Hammond, 1978)

Gneiss, migmatite, and amphibolite—Includes foliated and massive granite

Contact

High-angle fault—Bar and ball on downthrown side

Transcurrent fault—Showing relative horizontal movement

Thrust fault—Sawteeth on upthrown side

Strike and dip of foliation

Inclined

Vertical

Bearing and plunge of lineation—May be combined with foliation symbol

Direction of stratigraphic tops

Bearing and plunge of minor fold

Mylonite—Dominantly orthomylonite

CRSZ Carp River shear zone

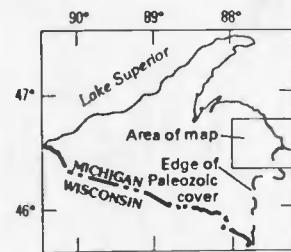
CRFSZ Carp River Falls shear zone

DRSZ Dead River shear zone

GLTZ Great Lakes tectonic zone

GGT Greenstone-granite terrane

Ropes mine



INDEX MAP

Granitoid Rocks

The Late Archean granitoid rocks in the greenstone-granite terrane (fig. 1) are dominantly pink to grayish pink, generally medium grained, porphyritic, foliated, and homogeneous tonalite or granodiorite (Gair and Thaden, 1968, p. 18-23). They contain xenoliths and schlieren of biotite schist and amphibolite. In the Sands quadrangle (fig. 1), weakly foliated granite is also a common rock type (table 1), but its age relation to the tonalite-granodiorite was not determined. The bimodal composition of the granitoid rocks suggests, however, that the granite represents a discrete magmatic event.

The granitoid rocks of the greenstone-granite terrane in the Marquette area exposed on both sides of the Marquette synclinorium are strongly altered (table 1). Plagioclase (oligoclase) is saussuritized and albite twinning is largely obscured, and biotite is largely changed to chlorite and opaque oxides (locally rutile). The rocks are highly fractured and fracture surfaces are coated by chlorite and other propylitic minerals. Along the southern margin of the Marquette synclinorium, the granitoid rocks are exceptionally rich in quartz (table 1). The granitoid rocks in the Sands quadrangle also are protomylonites. The protomylonite is characterized by recrystallization of quartz to fine grain sizes and localized shear-induced recrystallization of plagioclase and potassium feldspar to fine-grained polycrystalline aggregates (type 1P, 11P, 1M, and 11M structures of Hanmer, 1982).

The relative homogeneity of the foliated granitoid rocks, the occurrence of sharp-walled xenoliths, and the general absence of a lithologic layering that could represent relict sedimentary or volcanic layering suggest that the granitoid rocks of the greenstone-granite terrane are of magmatic origin. Their pervasive foliation is attributed to deformation subsequent to primary crystallization, as discussed following.

Structure

The rocks in the greenstone-granite terrane (northern complex), on the north side of the GLTZ (fig. 1), record early recumbent folding (F_1) of metavolcanic rocks of the Ishpeming greenstone belt. Superposed deformation (D_2) produced northwest- to west-trending upright, upward- and downward-facing folds (F_2) that are Z-shaped in plan view (Johnson and Bornhorst, 1991). An axial plane foliation was developed during F_2 . The Z-symmetry of the F_2 folds is consistent with their development in a deformation regime with a dextral shear component. Associated granitoid rocks also were deformed by D_2 . Younger, northwest- to west-trending faults, some of which have demonstrable dextral movement, transect and offset the folded rocks. Commonly, these faults separate domains of volcanic rock that have opposite stratigraphic facing directions.

Foliation and lithologic layering in the vicinity of the Ropes mine (R, near center of fig. 1) is puzzling with respect to the dominant regional structure. Here, foliation and lithologic layering strike about N. 70° E. and are nearly vertical (Brozdowski, 1988, p. A-44). The published geologic map that includes the Ropes mine area (Negaunee SW quadrangle; Clark and others, 1975) also shows a steep (stretch?) lineation that plunges southeastward in unit Wkf of the Late Archean Kitchi Schist (included in unit Wv on fig. 1). Possibly, the northeast-trending foliation in the Ropes mine area represents the east-northeast-trending limb of a large-scale D_2 Z-fold.

Table 1. Approximate modal content of granitoid rocks in Archean greenstone-granite terrane of the Marquette area

[Tr., trace; blank, absent]

Rock Constituent	Sample number																		
	136R	177A	177B	179A	179B	182A	182B	183A	186A	199	200	202	208	211A	136-1	203	1	2	3
Plagioclase	37	35	60	66	54.5	48	52.5	38	54.5	36	56.5	60	39	41	27.5	33	61.2	58.5	50.6
Quartz	30	32.5	24	21.5	28.5	45	36	34	24	27.5	37.5	28	30	27	39	28	26	22.2	37.5
Potassium feldspar	30	31	10	5.5	2.5	0	0.5	24.5	14.5	31	3	1	30	24	32	37.5	3.8	12.8	1.4
Biotite	Tr.	Tr.	Tr.	Tr.	10	Tr.	Tr.	2.5	7	5	Tr.	10	0	Tr.	Tr.	Tr.	0.2	Tr.	1
Chlorite	3	1.3	6	7	Tr.	6	6.5	Tr.	Tr.	Tr.	3	Tr.	Tr.	8	1.5	0.5	3.6	2.5	1.8
Epidote					4	Tr.		Tr.	Tr.								0.6	Tr.	Tr.
Sphene					Tr.												0	0	
Opaque oxides	Tr.	Tr.	Tr.	Tr.		Tr.	Tr.	Tr.		Tr.		Tr.		Tr.	Tr.	Tr.	0.2	Tr.	Tr.
Accessory minerals	Tr.	0.2	Tr.	Tr.	0.5		0.5	1	Tr.	0.5	Tr.	1	1	Tr.	Tr.	1	Tr.	Tr.	Tr.
Muscovite							4			Tr.	Tr.						2.5	1.3	7

SAMPLE DESCRIPTIONS

- 136R. Pink, fine- to medium-grained, slightly porphyritic, altered granite; contains wispy inclusions of biotite schist. Sands quadrangle, 1,800 ft W., 675 ft N. of the SE. corner of sec. 3, T. 47 N., R. 25 W.
- 177A. Pink, medium-grained, porphyritic, massive, altered granite; cut by chlorite-coated fractures. Sands quadrangle, 775 ft S., 2,150 ft E. of the NW. corner of sec. 16, T. 47 N., R. 25 W.
- 177B. Pink, fine-grained, porphyritic, massive, altered granite. Sands quadrangle. Same locality as 177A.
- 179A. Pinkish-gray, medium-grained, porphyritic, altered tonalite; contains oriented layers of biotite schist. Palmer quadrangle, 650 ft S., 250 ft W. of the NE. corner of sec. 26, T. 47 N., R. 26 W.
- 179B. Gray, medium-grained, foliated, altered tonalite. Biotite is fresh. Palmer quadrangle. Same locality as 179A.
- 182A. Pinkish-gray, medium-grained altered tonalite gneiss. Sands quadrangle, 1,350 ft E., 100 ft N. of the SW. corner of sec. 31, T. 47 N., R. 25 W.

- 182B. Pinkish-gray to medium-gray, altered tonalite gneiss. Sands quadrangle. Same location as 182A.
- 183A. Pinkish-gray, fine- to medium-grained granite gneiss. Sands quadrangle, 675 ft N., 300 ft E. of the SW. corner of sec. 31, T. 47 N., R. 25 W.
- 186A. Pinkish-gray, medium-grained, altered granodiorite gneiss. Palmer quadrangle, 2,500 ft N., 2,200 ft E. of the SW. corner of sec. 36, T. 47 N., R. 26 W.
199. Pinkish-gray, medium-grained granite gneiss. Sands quadrangle, 1,000 ft S., 2,000 ft W. of the NE. corner of sec. 16, T. 46 N., R. 25 W.
200. Pinkish-gray, fine- to medium-grained altered tonalite gneiss. Sands quadrangle, 775 ft N., 2,075 ft E. of the SW. corner of sec. 9, T. 46 N., R. 25 W.
202. Gray, fine-grained biotite tonalite gneiss cut by chlorite-coated fractures. Sands quadrangle, 850 ft N., 2,350 E. of the SW. corner of sec. 5, T. 46 N., R. 25 W.
208. Pink, medium-grained, leucocratic granite gneiss. Sands quadrangle, 2,150 ft N., 2,000 ft W. of the SE. corner of sec. 8, T. 46 N., R. 25 W.
- 211A. Pink, medium-grained, altered granite gneiss. Sands quadrangle, 2,550 ft S., 150 ft W. of the SE. corner of sec. 7, T. 46 N., R. 25 W.
- 136-1. Pink, medium- to coarse-grained, foliated granite gneiss. Fractures coated by chlorite. Sands quadrangle. Same location as 136R.
203. Pink, medium-grained, leucocratic granite gneiss. Sands quadrangle, 200 ft N., 2,300 ft W. of the SE. corner of sec. 5, T. 46 N., R. 25 W.
1. Average of 14 samples of foliated tonalite. Marquette quadrangle, north of the Marquette synclinorium (Gair and Thaden, 1968, table 6, no. 1).
 2. Average of five samples of foliated granodiorite. Marquette quadrangle, north of the Marquette synclinorium (Gair and Thaden, 1968, table 6, no. 9).
 3. Average of 11 samples of quartz-rich, foliated tonalite. Sands quadrangle, south of the Marquette synclinorium (Gair and Thaden, 1968, table 6, no. 6).

The steep, northwest- to west-trending shear zones and faults in the greenstone-granite terrane, as exposed in the northern complex, are tens of meters to a few hundred meters wide. They are highly schistose zones characterized by an intense, close-spaced foliation, a steep stretching(?) lineation, and strong retrograde alteration (Johnson and Bornhorst, 1991). The faults are of both ductile and brittle-ductile types. Major structures include the Dead River shear zone (fig. 1, DRSZ) (Puffett, 1974), which forms the northern boundary of the Dead River basin; the Carp River shear zone (CRSZ), east-northeast of the Ropes mine, which separates two different blocks of Archean volcanic rocks; and the Carp River Falls shear zone (CRFSZ), which forms the northern margin of the Marquette synclinorium in this area. The Carp River Falls shear zone is reported to be cut by relatively undeformed mafic dikes of presumed Late Archean age, indicating an Archean age for the shear zone (Baxter and Bornhorst, 1988).

In the area north of the Palmer fault (fig. 1), foliated and fractured Late Archean granitoid rocks form partly fault-bounded domes surrounded by Early Proterozoic rocks of the Marquette Range Supergroup. The large Archean granitoid body in the Sugarloaf Mountain area, 8 km north of Marquette, is also a dome; foliation in adjacent metavolcanic rocks (Puffett, 1974) dips gently to moderately away from the granite contact.

Late-Tectonic Conglomerate (Archean)

The youngest Archean unit in the greenstone-granite terrane is the Reany Creek Formation (Puffett, 1969; 1974). On the basis of a reinterpretation of age and structural relationships, the Reany Creek Formation is no longer included as part of the Chocoy Group or the Marquette Range Supergroup (Sims, 1991). It is a heterogeneous body of conglomerate, arkose, chloritic slate, graywacke, and boulder-bearing slate. It transects structures in the older volcanic rocks, is less deformed than the metamorphosed volcanic rocks, and is bounded along its southern margin by the northwest-trending Dead River shear zone (fig. 1). Its age has been uncertain (see Puffett, 1974), but its penetrative foliation, asymmetry of basin fill, and relationship to the Dead River shear zone strongly suggest that it developed concurrently with dextral shear along the Dead River shear zone. It is similar to "Timiskaming-type" sequences, such as the Seine Group, in northern Ontario, now commonly interpreted as forming in Archean analogs to modern pull-apart basins (Poulsen, 1986).

ARCHEAN GNEISS TERRANE

The Archean gneiss terrane in the Marquette area (fig. 1) constitutes the greater part of the southern complex, as defined by Van Hise and Bayley (1897). It consists of gneiss, migmatite, and amphibolite, substantial amounts of deformed and undeformed granite pegmatite, and massive to weakly foliated granite plutons. Cannon and Simmons (1973) have described the general rock types in much of the southern complex. One rock has been dated as Late Archean. A sample of gray gneiss (called Compeau Creek Gneiss by Gair, 1975) collected in the SE $\frac{1}{4}$ SW $\frac{1}{4}$ of section 36, T. 47 N., R. 27 W. (Hammond, 1978) has a U-Pb zircon age of $2,779 \pm 21$ Ma and a lower intercept age of 802 ± 76 Ma (recalculated by Zell E. Peterman).

Gneiss and Associated Granitoid Rocks

Compositionally layered, medium-grained gneiss and migmatite are the dominant rock types in the gneiss terrane in the Marquette area. Layered felsic gneisses ranging in composition from tonalite to granite predominate. Smaller amounts of massive to layered amphibolite are

intercalated with the felsic gneiss, but amphibolite constitutes layers several tens of meters thick at places, as can be seen on the geologic map of the Palmer 7½-minute quadrangle (Gair, 1975, pl. 1), which is immediately west of the Sands quadrangle. The felsic gneisses are gray to pinkish gray; typically, compositional layering is expressed by different proportions of the major silicate minerals, as for example, (1) plagioclase-quartz-biotite-microperthite, (2) microperthite-quartz-plagioclase-biotite, and (3) biotite-quartz-plagioclase-microperthite. Textural differences at places emphasize the compositional layering. Pink aplitic granite and granite pegmatite commonly transect the gneiss and amphibolite and locally form migmatite. Metasedimentary rocks such as iron-formation, form layers in the felsic gneisses in the vicinity of the Republic trough (Cannon and Simmons, 1973); they were not observed in the Palmer and Sands quadrangles, in the Marquette area.

Pinkish-gray to pink, medium-grained, massive to weakly foliated, homogeneous granite (table 2) intrudes the gneisses in several places. Hammond (1978) delineated a body of massive granite about 4 km² in areal extent south of Ishpeming (unit Wgt, fig. 1), which he informally called the "Tilden granite." It is a gray to pink, medium-grained, locally porphyritic, and massive granite that locally contains oriented xenoliths of mafic gneiss. It is cut by pink pegmatite and is highly fractured. The fractures have slickensided surfaces and a thin coating of chlorite and other propylitic alteration minerals. U-Pb isotopic data on a sample of the granite near Tilden (Hammond, 1978) indicate an age of $2,585 \pm 15$ Ma (R.E. Zartman, oral commun., 1991).

A nearly circular body of alkali granite (unit Xga, fig. 1) about 2 km in diameter occurs about 3 km south of Humboldt (Schulz and others, 1988). The granite is light red to brick red, generally massive, fine- to medium-grained, and equigranular to hypidiomorphic granular. The granite is similar compositionally to Sn-W mineralized alkali feldspar granites of the Arabian Shield (Jackson and Ramsay, 1986) and the Nigerian younger granite province (Kinnaid and others, 1985). The granite has a Rb-Sr whole-rock age of $1,733 \pm 25$ Ma, which is interpreted as a crystallization age (Zell E. Peterman, written commun., 1988); it is a post-tectonic intrusion.

Structure

Archean gneisses in the gneiss terrane (southern complex) form a northwest-trending antiformal structure that closes to the west and is overlapped by Paleozoic rocks to the east (Sims, 1992). An infolded belt of Early Proterozoic (Marquette Range Supergroup) rocks indents the Archean fold nose in the Republic trough (fig. 1). In the area west of the Republic trough, Taylor (1967) determined two principal phases of deformation: (1) early, probably flat-lying folds with axial planes trending northeastward, and (2) younger upright folds with steep northwest-trending axial surfaces. The younger folds mainly control the distribution of the rock units.

In the Marquette area, early gently inclined to recumbent folds that trend northwestward and plunge gently northward (fig. 2) are the dominant structure in the gneisses. These folds deform an older foliation (S_1). Boudinage accompanied the folding; the boudins plunge subparallel to gently inclined fold hinges. A later steep, northwest-oriented foliation, that is axial planar to northwest-trending folds, occurs at least locally, but it does not significantly affect the distribution of the rock units. These younger structures will be examined at stop 3.

Table 2. Approximate modal content of granitoid rocks in the Archean gneiss terrane of the Marquette area

[Tr., trace; blank, absent]

Rock Constituent	Sample No.				
	153A	146-88	1	226A	226B
Plagioclase	34.7	33.5	25	26.3	39.3
Quartz	28	23.8	25	38.2	32.7
Potassium feldspar	28	34.2	40	33.0	25.0
Biotite	9	8.0	3	2.2	3
Chlorite		Tr.	Tr.	Tr.	Tr.
Muscovite	Tr.	Tr.	2		Tr.
Epidote		Tr.			
Sphene			Tr.		
Opaque oxides	Tr.		Tr.		
Accessory minerals	0.3	0.5	Tr.	0.3	Tr.

SAMPLE DESCRIPTIONS

- 153A. Pinkish-gray, medium-grained, foliated granite, SE $\frac{1}{4}$ NE $\frac{1}{4}$ of section 7, T. 46 N., R. 26 W. Biotite is weakly altered.
- 146-88. "Tilden granite" of Hammond (1978). Quartz and biotite are recrystallized in shears; biotite is slightly altered to chlorite.
1. "Tilden granite" of Hammond (1978, p. 63). Potassium-feldspar is microperthite. Plagioclase has concentric zoning. Highly fractured.
- 226A. Light-gray, medium-grained, foliated granite. Cut by fractures. Biotite is highly altered to chlorite. Quarry, SE $\frac{1}{4}$ SW $\frac{1}{2}$ of section 21, T. 46 N., R. 26 W.
- 226B. Pale-reddish-brown, medium-grained foliated granite. Cut by shears, some with mylonite. Biotite is highly altered to chlorite and calcite. Same locality as 226A.

GREAT LAKES TECTONIC ZONE

The GLTZ is characterized in the Marquette area by a mylonite zone about 2 km wide that has been superposed on previously deformed layered rocks and granitoid rocks of the northern Archean greenstone-granite terrane and previously deformed rocks of the southern Archean gneiss terrane. Therefore, the mylonite overprints rocks of the two terranes. The great width of this shear zone and the preponderance of mylonite distinguish it from the other, much narrower shear zones and faults in the region. The mylonite abruptly grades northward into protomylonite and highly fractured and altered rocks along the northeast wall of the GLTZ (fig. 3).

Boundaries (walls) of the GLTZ are subparallel and trend N. 55°-60° W., subparallel to the contact between the Archean greenstone-granite and gneiss terranes (fig. 3). Foliation in the mylonite has an average attitude of N. 70° W., 75° SW (fig. 4). Thus, the foliation strikes 10°-15° more westward than the trend of the shear zone walls. The angular relationship between the orientation of the foliation and shear zone boundaries suggests dextral wrench shear along the walls (Simpson, 1986, fig. 2b). A pronounced rodding (stretching) lineation in the mylonite has an average plunge of 42° and an average bearing of S. 43° E.

Hinges of tight to open folds plunge subparallel to the stretching lineation (fig. 4). The folds have both symmetrical and asymmetrical patterns. Z-shaped folds are most common, but S-shaped folds coexist and both may occur in the same outcrop area, as on the west shore of Powell Lake (fig. 3). Nearly pervasive lichen covers outcrops and prevented determination of

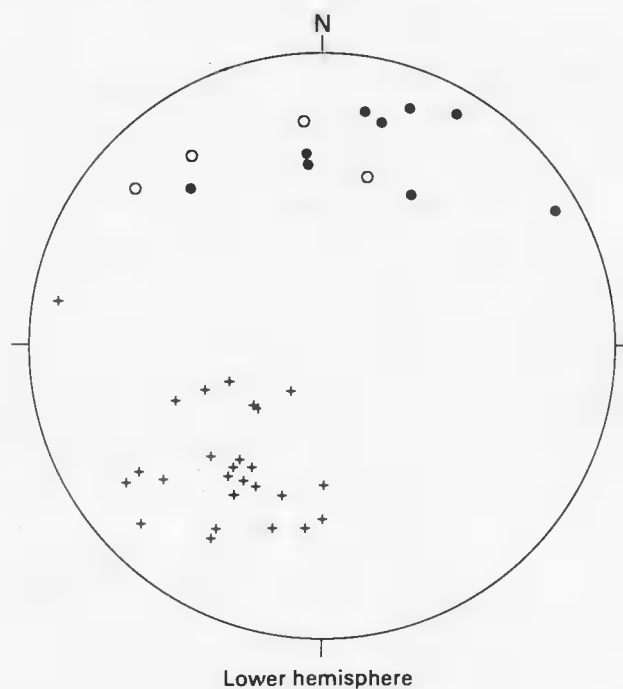


Figure 2. Equal-area projection of poles to foliation, lineations, and fold axes in gneisses of Archean gneiss terrane.

fold symmetries over large parts of the GLTZ. Ridley and Casey (1989) have demonstrated that symmetrical folds, with axes subparallel to the maximum extension direction (X finite strain axis), are produced by wrench shear, whereas asymmetric folds, with axes close to the extension direction, form in a regime of combined wrench and thrust shear. In wrench shear, because the fold axis does not lie initially within the slip plane of shear, the axis rotates toward the direction of shear displacement and maximum extension with the addition of progressive strain. In combined wrench and thrust shear, at high strain the fold axial plane and both limbs approach parallelism with the shear plane. Ridley (1986) has further shown that dextral wrench shear and sinistral wrench shear can result from strain inhomogeneity in the same shear-parallel extension within a single deformation regime.

Mylonite

The mylonite in the GLTZ is mainly orthomylonite, as defined by Wise and others (1984), inasmuch as surviving megacrysts compose 10-20 percent of the rock. In the terminology of Hanmer (1987), the mylonites are mainly "heteroclastic" because the porphyroclasts have variable size ranges, but include "homoclastic" mylonite having more uniform textural characteristics. Ultramylonite is absent except on a scale of a few centimeters.

The mylonite typically has a pronounced planar foliation that differs markedly from the more irregular foliation in rocks outside the GLTZ. In this respect, the mylonite fits the pattern of most mylonite zones in comprising a "straight zone." Quartz typically is recrystallized into "ribbon quartz," yielding a pronounced rodding (stretch) lineation. The relatively stiff minerals, plagioclase and potassium feldspar, are generally recrystallized to finer grain sizes, with the development of core-mantle structures (White, 1976) or type 1P and 1M structures (Hanmer, 1982); these structures form oriented aggregates of quartz and feldspar that produce a prominent stretch lineation. Accompanying biotite is mainly recrystallized in planar or irregular shears. The quartz-poor amphibolite in the GLTZ has a conspicuous lineation given by elongate aggregates of actinolite and chlorite. The mafic mineral assemblages in rocks in the GLTZ are upper greenschist facies, indicating retrogressive alteration of amphibolite to actinolite schist.

Interpretation

The Great Lakes tectonic zone has been interpreted (Gibbs and others, 1984) as a paleosuture resulting from continent-continent collision that juxtaposed the Archean gneiss and greenstone-granite terranes. In the Marquette area, the stretching lineation, which represents the line of tectonic transport (Schackleton and Ries, 1984), indicates that collision was oblique, resulting in dextral wrench-thrust shear along the N. 60° W.-trending boundary (paleosuture). Asymmetric structures, mainly microscopic, indicate northwestward vergence and probable overriding of the Archean greenstone-granite terrane by the Archean gneiss terrane.

The oblique collision would be expected to produce dextral shear across a large region north of the GLTZ (fig. 5). The extent of the area affected by this dextral transcurrent shear is not definitely known, however, because dextral shear was the dominant mechanism of deformation throughout most of the Superior province (Card, in press).

Structures in the greenstone-granite rocks of northern Minnesota (Wawa subprovince) are remarkably similar to those in northern Michigan. In northern Minnesota, deformed and metamorphosed volcanic and sedimentary rocks of the Vermilion district (Sims, 1976; Sims and

Figure 3--Continued

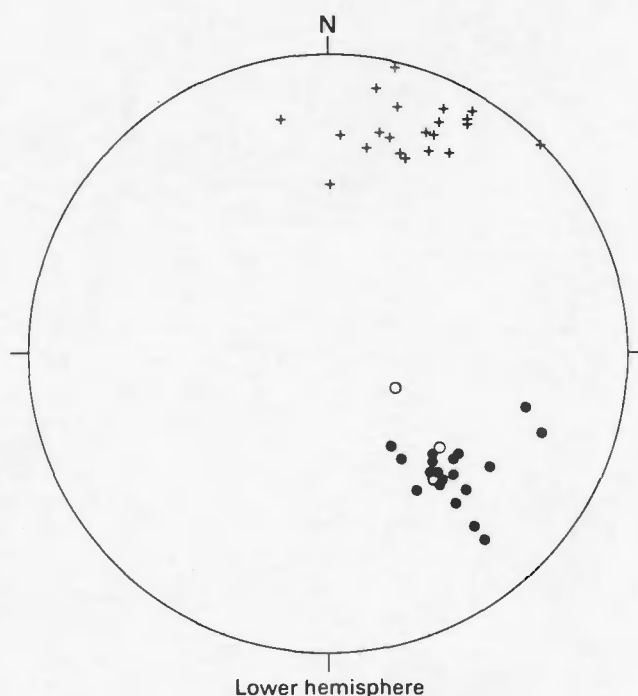
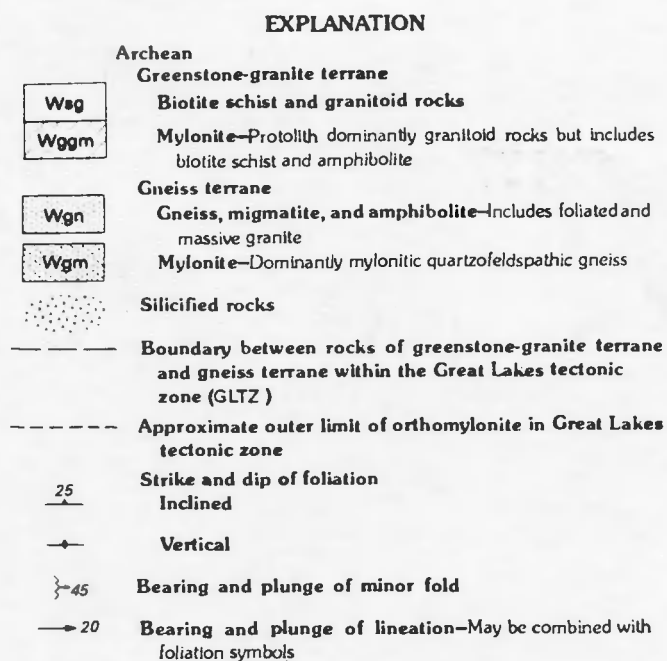


Figure 4. Equal-area projection of poles to foliation (crosses), stretching lineation (dots), and fold hinges (open circles) in mylonite of the Great Lakes tectonic zone, Sands and Palmer 7½-minute quadrangles, Marquette area, Michigan.

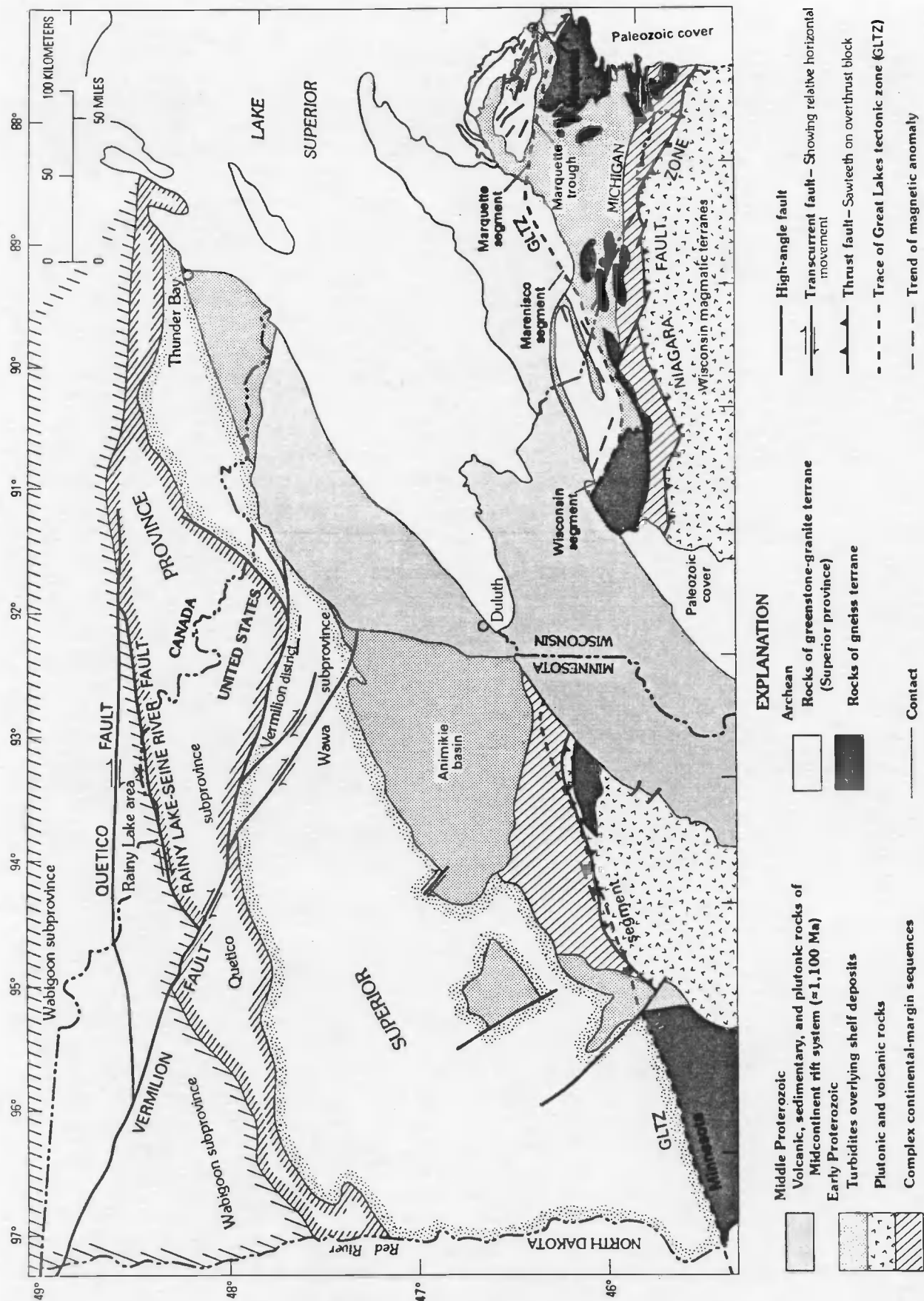


Figure 5. Simplified tectonic map of the Lake Superior region showing Great Lakes tectonic zone and adjacent Archean terranes. After Sims (1991). Terminology of Early Proterozoic rocks modified from Southwick and Morey (1991).

Southwick, 1985) compose an east-trending belt between higher grade rocks of the Late Archean Vermilion Granitic Complex (Quetico subprovince, fig. 5) (Southwick, 1972) and the Giants Range batholith to the south. The measured strain, a cleavage, upright folds, and a mineral lineation in this belt have been attributed to the "main" phase of deformation (D_2) that followed an early nappe-forming event (D_1) (Bauer, 1985). The nappes show little evidence of a penetrative fabric (Hudleston, 1976). Hudleston and others (1988) attributed the (D_2) deformation to regional dextral transpression, as the strain pattern requires a northeast-southwest component of shortening in addition to shear. They further proposed that major dextral faults, such as the Vermilion fault (fig. 5), are later, more brittle, expressions of this shear regime. They concluded that the D_2 transpressive deformation resulted from oblique compression between the two more rigid crustal blocks to the north (Quetico subprovince) and south (Giants Range batholith). A similar tectonic regime has been recognized in the Rainy Lake area (fig. 5; Poulsen and others, 1980; Day and Sims, 1984; Wood, 1980), where early recumbent folding was followed by upright folding and dextral strike-slip faulting.

Recent precise isotopic analyses of zircon, titanite, and rutile from the Rainy Lake area, Canada (fig. 5; Davis and others, 1989), which lies between the Quetico and Wabigoon subprovinces, have provided time constraints on these structural events. The major deformation, including nappe emplacement, thrusting, and local doming, took place between 2,696 Ma and 2,692 Ma; this deformation was followed shortly by transcurrent faulting and simultaneous deposition of conglomerate/arenite (Timiskaming-type rocks, as in Seine Group), which occurred in the interval 2,692-2,686 Ma. Late (Algonian) granitic plutons were emplaced about 2,686 Ma, although some are older. In the Wawa subprovince, west of Thunder Bay (fig. 5), Corfu and Stott (1986) found that the D_1 deformation occurred during or before the intrusion of the Shebandowan Lake pluton at $2,696 \pm 2$ Ma. Deformation D_2 in this area occurred between $2,689 \pm 3/-2$ Ma and $2,684 \pm 6/-3$ Ma, similar to the age suggested by Davis and others (1989) for D_2 in the Rainy Lake wrench fault zone. These ages are compatible with the less precise isotopic ages on rocks in northern Minnesota and Michigan (Peterman, 1979), and it seems probable that the rocks and structures throughout the Wawa and Quetico subprovinces are approximately coeval (Percival, 1989). Although convergence along the GLTZ undoubtedly was diachronous, collision probably occurred in the approximate interval 2,692-2,686 Ma (Davis and others, 1989), to yield the transcurrent faults and Timiskaming-type rocks.

Kinematic Analysis

The attitude of the stretching lineation (line of tectonic transport) in the mylonite exposed south of Marquette together with asymmetric meso- and micro-structures that reveal sense of movement, indicate that the oblique collision resulted in: (1) dextral-thrust shear along the GLTZ, and (2) northwestward vergence and probable overriding of the Archean greenstone-granite terrane by the Archean gneiss terrane. Kinematic indicators--rotated mica grains within narrow compositional layers, asymmetric porphyroclasts with tails (σ type; Simpson, 1986), and asymmetric microfolds in mylonitic layering--demonstrate northwestward vergence. This information implies southward subduction of the Archean greenstone-granite terrane (Wawa subprovince) beneath the Archean gneiss terrane (fig. 1).

Evolution

The northwest direction of tectonic transport during suturing of the Archean terranes, as ascertained from the Marquette area, provides a means for determining the evolution of the GLTZ and the variable trajectory of stress into the Superior province crust.

The GLTZ in the Lake Superior region is characterized by systematic angular bends that alternately trend northeastward and west-northwestward (fig. 5). Presumably this zigzag pattern reflects relict irregularities in the margin of the Archean greenstone-granite terrane (or Superior province) crust, which was a continental margin before the convergence and collision with the southern Archean gneiss terrane.

The northeast-trending and west-northwest-trending segments of the GLTZ have different structural styles. Deformation along the northwest-trending segments of the GLTZ, as particularly shown by data from the Marquette segment, was principally caused by dextral transpression resulting from oblique collision. Transmittal of this transcurrent shear into rocks north of the GLTZ yielded a widespread, pervasive west-northwest- to west-striking foliation, subparallel upright folds, and northwest- to west-trending dextral faults and shear zones in the Archean greenstone-granite terrane.

The similarly-oriented northwest-trending segment of the GLTZ in northwestern Wisconsin has many structural features in common with the Marquette segment. Foliation and upright folds in low amphibolite-facies rocks of the Archean greenstone-granite terrane (unit Wga, fig. 4, Sims and others, 1985) strike west-northwest and mineral lineations and fold hinges mainly plunge gently southeast. The boundary between the two terranes is not exposed because of a glacial cover, but is presumed to lie along the south edge of unit Wga. Numerous northwest-trending dextral faults, some of which reactivated in Early Proterozoic time, have been mapped in the area (Sims and others, 1985; fig. 1).

Collision along the northeast-trending segments of the GLTZ, on the other hand, produced northeast-trending structures of apparently more restricted areal extent. In the northeast-trending Marenisco segment (fig. 5; Sims and others, 1984), the boundary is covered by Early Proterozoic sedimentary and volcanic rocks, however, lithologic layering and foliation in rocks of the adjacent Archean greenstone-granite terrane near the boundary trend northeastward and are deformed into upright, moderately tight northeast-trending folds that plunge 45°-50° SW. These structures are presumably subparallel to the covered Archean GLTZ boundary. Archean metamorphism has been overprinted by Early Proterozoic Penokean nodal metamorphism centered on the Watersmeet dome (Sims and others, 1985; Sims, 1990); the presence of relict garnet at a few places in the Archean rocks of the greenstone-granite terrane near the boundary suggests that these rocks were metamorphosed to at least upper greenschist facies in Archean time. Similarly, north-verging Penokean deformation in the boundary zone overprinted Archean structures (Sims and others, 1984). An axial plane S_2 (Penokean) penetrative cleavage, striking northeast and dipping 45°-70° SE., was superposed on the previously folded rocks. Apparently the Archean rocks were not refolded, however, as a result of the Penokean deformation.

In the northeast-trending Minnesota segment of the GLTZ (fig. 5), neither the terrane boundary nor adjacent Archean rocks on either side are exposed. These rocks are covered in west-central Minnesota by thick Quaternary glacial deposits and in central Minnesota by Early Proterozoic sedimentary and volcanic rocks of the Animikie basin (Southwick and others, 1988). The GLTZ

has been investigated, however, by a detailed aeromagnetic survey, by computer-generated mapping of the second vertical derivative of the gravity field, and by shallow test drilling (see Southwick and Sims, in press); the boundary has been located rather accurately on the basis of these data. The drilling has shown that the rocks on the northwest side are volcanogenic sedimentary and mafic to intermediate volcanic rocks, which are metamorphosed to upper greenschist facies and intruded by Archean tonalite (Southwick and Chandler, 1983). These rocks are typical of the Archean greenstone-granite terrane in exposed parts of the Lake Superior region. A seismic reflection profile in central Minnesota acquired by COCORP (Consortium for Continental Reflection Profiling) has been interpreted to indicate that the GLTZ in this area is a shallow ($\sim 30^\circ$) north-dipping tectonic feature (Gibbs and others, 1984). In east-central Minnesota, the GLTZ is covered by Proterozoic rocks of the Animikie basin. The structural style in the Proterozoic cover indicates north-verging tectonism (Southwick and others, 1988), and as in the Marenisco segment, the Archean crustal boundary had a role in defining Penokean deformation.

Deformation along both of the northeast-trending (Minnesota and Marenisco) segments of the GLTZ resulted mainly from northwest-southeast crustal shortening, probably dominantly by flattening strain. The direction of tectonic transport during convergence was virtually perpendicular to the juncture of the two terranes at these localities.

The origin of the zigzag pattern of the south edge of the Superior province, now marked by the GLTZ, is uncertain. The prevailing thought is that the Wawa subprovince is one of a sequence of stacked island arcs that formed progressively from north to south above north-dipping subduction zones as the continental mass to the south of the GLTZ (i.e., Archean gneiss terrane) migrated to the north (Card, 1990). With this interpretation, possible modern analogs of the Superior province are the convergent-plate boundaries of the western Pacific, as for example those of the Indonesian region (Hamilton, 1979).

The physical resemblance of the southern margin of the Superior province to the Appalachian-Ouachita Paleozoic orogenic belt (Thomas, 1977), however, suggests a possible alternative interpretation for the origin of the Superior margin. In this interpretation, the Superior margin was a rifted continental margin. Two interpretations have been made for the origin of the Paleozoic continental margin: (1) rift segments were offset by transform faults, as suggested by Thomas (1977, 1983), or (2) intersections between active rift arms at triple junctions (Rankin, 1976). Of these two suggestions, the rift-transform mechanism seems the more likely, with the Minnesota and Marenisco segments being the rifted segments (fig. 5) and the northwestern Wisconsin and Marquette segments being highly modified transform faults. Regardless of the mechanism by which the zigzag Archean continental margin originated, the subsequent trace of the orogenic belt was probably inherited from the shape of the earlier margin.

CONCLUDING REMARKS

Convergence along the irregularly shaped margin of the Archean greenstone-granite terrane (GLTZ) resulted in a variable trajectory of stress into the continental crust and probably in along-strike diachroneity of orogeny. Structural data from the Marquette area, in particular, as well as elsewhere along the GLTZ, suggest that the major direction of tectonic transport was northwestward. Accordingly, promontories such as those along the concave parts of the Marquette and Wisconsin segments of the GLTZ (fig. 5), where the zone bends from a northeast orientation to a northwest one, must have projected as buttresses against which

compressive stress was directed into the continental crust. Oblique compression at these points produced dextral shear across the region north of the suture, probably at least as far northward as the Quetico fault (fig. 5), a distance of about 250 km. This shear imposed a roughly east-west, steep structural fabric on the rocks and, also as a late, more brittle expression of the shear regime (Hudleston and others, 1988), the northwest- to west-trending dextral transcurrent faults.

Along the Marenisco and Minnesota segments of the GLTZ (fig. 5), where convergence was more perpendicular to the ancient continental margin, a northeast-trending structural fabric was imposed on the rocks immediately cratonward from the suture.

We suggest that the main structural fabric (D_2) in rocks of the Archean greenstone-granite terrane in the north-central United States (Wawa and Quetico subprovinces; fig. 5) resulted from the collision along the GLTZ. The predominance of orthomylonite rather than ultramylonite and the nearly pervasive retrogressive alteration (greenschist facies) in rocks of the greenstone-granite terrane suggest that the exposed collision zone was developed at a moderately shallow crustal level (at brittle-ductile transition conditions). As discussed in a previous report (Sims and others, 1980), the Archean structures in this regime also played a strong role in subsequent tectonism, especially in the Early Proterozoic north-verging deformation.

We further suggest that the late-tectonic granite bodies in the Archean gneiss terrane are possibly related to the collision along the GLTZ and presumed southward subduction. The available age data on these granites are compatible with a presumed 2.69 Ga age for the collision. The "Tilden granite" of Hammond (1978) in Michigan has a Late Archean age ($2,585 \pm 15$ Ma), although both the U-Pb and Rb-Sr systems are disturbed. In the Minnesota River valley, in southwestern Minnesota, a large pluton of late-tectonic granite (Sacred Heart Granite) has a Pb-Pb age of about 2,605 Ma (Doe and Delevaux, 1980) and a Rb-Sr age of about 2.7 Ga (Goldich and others, 1970). Doe and Delevaux (1980) have shown that ^{207}Pb - ^{204}Pb values in the Sacred Heart Granite are characteristic of ensialic environments, as contrasted with the ensimatic (arc) granitoid bodies in the Superior province (greenstone-granite terrane). The ensialic environment indicates that the Archean gneiss terrane had been cratonized prior to emplacement of the Sacred Heart Granite. Precise ages are required to test the hypothesis that the Late Archean granites south of the GLTZ were indeed formed during continent-continent collision.

Cumulative data on the Archean Superior province (see Hoffman, 1989, for review) indicate that it consists of generally east trending belts of island arc and related rocks that were assembled progressively from north to south (Card, in press), before finally colliding with the Archean gneiss terrane (continent) on the south at about 2,690 Ma. This pattern of accretion as well as the tectonic style, is not unlike that in modern plate-tectonic regimes, indicating that plate-tectonic mechanisms existed in the Archean as well as in the Proterozoic and Phanerozoic.

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Field stops and road log
Field stops shown on figure 6

Miles	
0	Depart from Ramada Inn in downtown Marquette on Michigan highway 28 (M28) and proceed west.
3.8-6.3	Numerous outcrops of pillow basalt along M28. Basalt comprises a major part of Archean greenstone-granite terrane (Wawa subprovince) in the northern complex of Marquette district.
	Early Proterozoic rocks of Marquette syncline lie to the south of highway.
7.2	Turn left (south) from M28 onto M35.
10.8	M35 intersects Marquette County road 480 (Co. 480). Continue south on M35.
11.2	Outcrops of Early Proterozoic Negaunee Iron-formation.
14.1	High dumps at 12 o'clock, Empire iron mine waste rock. Pass through Palmer.
16.9	Turn right (west) on County highway 565.
18.1	At road fork turn right onto County highway 476.
19.5	Stop 1. Roadcut in "Tilden granite" (Hammond, 1978).
	The "Tilden granite" is a pinkish-gray to pink, medium-grained, locally porphyritic, massive to weakly foliated, homogeneous granite that locally contains oriented xenoliths of mafic gneiss. It intrudes gneisses and in turn is cut by pink pegmatite. The granite is highly fractured; the fractures have slickensided surfaces and a thin coating of chlorite and other propylitic alteration minerals. The granite has a U-Pb concordia intercept age of $2,585 \pm 15$ Ma (R.E. Zartman, oral commun., 1991). Chemically, the granite has characteristics of post-orogenic, within plate granite on discrimination diagrams (T. Bornhorst, written commun., 1990), which is consistent with its post-tectonic field relationships.
	Continue southwest on Co. 476.
19.7	Turn around and return to M35.
22.4	Junction of Co. 476 and M35. Proceed southeast on M35.
23	Stop 2. High roadcut on M35 in Archean gneiss (of Archean gneiss terrane). Road shoulder is narrow. Be careful.
	Excellent outcrop of layered gneiss (of Archean gneiss terrane) on the northeast side of road. Amphibolite and quartzofeldspathic gneiss are interlayered, and are cut by red pegmatite. Pegmatite and aplite are boudined and brecciated; boudins

plunge parallel to recumbent fold axes. Tight folds with axial planar foliation plunge about 25° with a strike of N. 5° W., as does a mineral lineation.

This outcrop is typical of the gneiss of the Archean gneiss terrane, but amphibolite is uncommonly abundant here. Tight recumbent folds that plunge gently northward are the dominant structures in this part of the gneiss terrane. Return to vehicle and continue southeast on M35.

23.7 M35 intersects private road (Cleveland Cliffs Iron Company).

24.2 Stop 3. Low outcrop on northeast side of road in grassy area. Outcrop has been covered by loose sand and is lichen free.

This outcrop contains northwest-trending folds with an axial planar foliation visible on polished surfaces that strikes N. 65° W., and dips 60° NE. This structure is superposed on recumbently folded gneiss, and therefore is younger. Thick layers of felsic gneiss retain a dominant older foliation that strikes N. 25° W.

The steep northwest-trending foliation and associated folds are parallel to the trend of the antiformal southern complex (of Marquette district). Although minor in this area, upright northwest-trending folds are the dominant structures in the Republic area, 30 km to the west.

Continue southeast on M35.

24.3 Turn around at Community Club and proceed northwest on M35 toward Palmer.

24.9 M35 intersects private road (CCI). Continue north on M35 to intersection with Co. 480.

32.3 Intersection of M35 and Co. 480. Turn right (east) on Co. 480.

35.6 Turn right (south) on secondary road. Roads for remainder of field excursion are unmarked; accordingly, we have arbitrarily assigned letter designations A through L (see fig. 3) to specific segments as an aid in location. The main road into area (road A) is a graded sandy dirt road; other roads followed on the trip are not graded and may be rutted or soft, requiring some care in driving.

39.6 Road A crosses Chicago and Northwestern railway near Gentian.

39.7 Take right road fork (B) and continue southwest on B.

40.6 Junction with east-west road (C). Proceed west on C for 0.3 mi.

40.9 Stop 4. Rock knob north side of road C.

This outcrop is typical of the Archean greenstone-granite terrane. It consists of granite gneiss (or foliated granite) that contains lenses of biotitic and chloritic amphibolite. These rocks are cut by granite pegmatite and aplite, which crosscut the foliation in the granitoid rocks. Note that crosscutting aplite and pegmatite are not sheared out. Foliation strikes N. 70° W. and dips 65° SW.

Although this outcrop lies near the north margin of the Great Lakes tectonic zone and although it has a foliation subparallel to that in the GLTZ, the outcrop and those to the west-northwest are excluded from the GLTZ because the crosscutting pegmatite and aplite are not sheared out and the rocks are protomylonite. The northern boundary is placed at the abrupt transition from "straight-banded" mylonite in the GLTZ to rocks that retain recognizable igneous structures, such as here. The rocks to the north of the GLTZ are protomylonite.

Return to vehicle and continue west on C for 0.2 mile.

41.1 Intersection of roads C and D. Turn left on D and continue for 0.4 mi.

41.5 Stop 5. Outcrops of milky quartz and altered Archean rocks about 50 yards west of road D.

Milky quartz and silicified country rock, such as those exposed here, characterize the hills to the west. These outcrops comprise the most eastern exposed part of a quartz lense, oriented about east-west, that is about 1.4 km long. The quartz body may occupy a major tension fissure zone related to dextral wrench shear within the GLTZ.

Turn around and proceed north on D to intersection with C.

Then follow C and D to road fork at Gentian crossing.

42.8 Turn sharply right from road B onto road E; continue south for 1.2 mi.

44 Road fork; continue south on road F.

44.35 Road fork; continue south on right fork (G), then turn left at road fork onto H and continue east on H to locked gate.

45.0 Park and walk on private road east and northeast to west shore of Powell Lake (distance 0.3 mi). Walk is easy.

Stop 6. This is an exceptionally good series of outcrops of mylonite within the GLTZ because they are periodically washed clean. The rock is dominantly granite gneiss containing pods of migmatized amphibolite and hornblende schist. The protolith is granite and amphibolite of the Archean greenstone-granite terrane.

Foliation in the gneiss strikes about N. 75° W. and dips 75° SW., but is variable. A conspicuous stretching lineation plunges uniformly about 55° and bears S. 45° E. Both dextral and sinistral asymmetric folds plunge subparallel to the lineation; dextral folds predominate. Porphyroclasts are common, but they do not clearly have internal monoclinic symmetry and elongate tails, necessary to determine movement sense (Simpson, 1986).

Return to parked vehicle at locked gate. Proceed west on road H to road fork (distance 0.25 mi), then turn right on H to junction of H with J (distance 0.15 mi).

- 45.4 Junction of roads H and J. Turn sharply left on J and continue for 0.2 mi.
- 45.6 Road fork. Continue on right fork (K) for 0.1 mi.
- 45.7 Park and walk on trail to northwest, a distance of about 750 ft, to outcrops.

Stop 7. Low outcrops of pink mylonitic gneiss with lenses of biotite schist. Protolith is considered to be felsic gneiss of Archean gneiss terrane. Foliation strikes N. 75° W. and dips 65° SW. A quartz rodding lineation plunges 50° and bears S. 45° E. Note abundant ribbon quartz. This outcrop is in the southern part of the GLTZ.

Return to road. Turn around and follow roads K and J to road fork with H.

Proceed south on H to junction (1.5 mi).

- 46.2 Continue south on L.
- 47.3 Road L intersects power line. Park and walk 750 ft southwest to low outcrops along power line.

Stop 8. Outcrops comprise pink granite gneiss with amphibolite layers and lenses, cut by pink granite pegmatite. Foliation strikes N. 40° W., and dips 80° NE. Rocks are not sheared. Two parallel diabase dikes cut the Archean gneiss.

These outcrops are within the Archean gneiss terrane, a distance of about 0.65 km south of the south boundary of the GLTZ. Thus, we have crossed from the Archean greenstone-granite terrane, through the GLTZ, into the Archean gneiss terrane to the south of the GLTZ.

End of field trip. Suggest returning to Co. 480 via dirt roads L, G, F, E, and A.

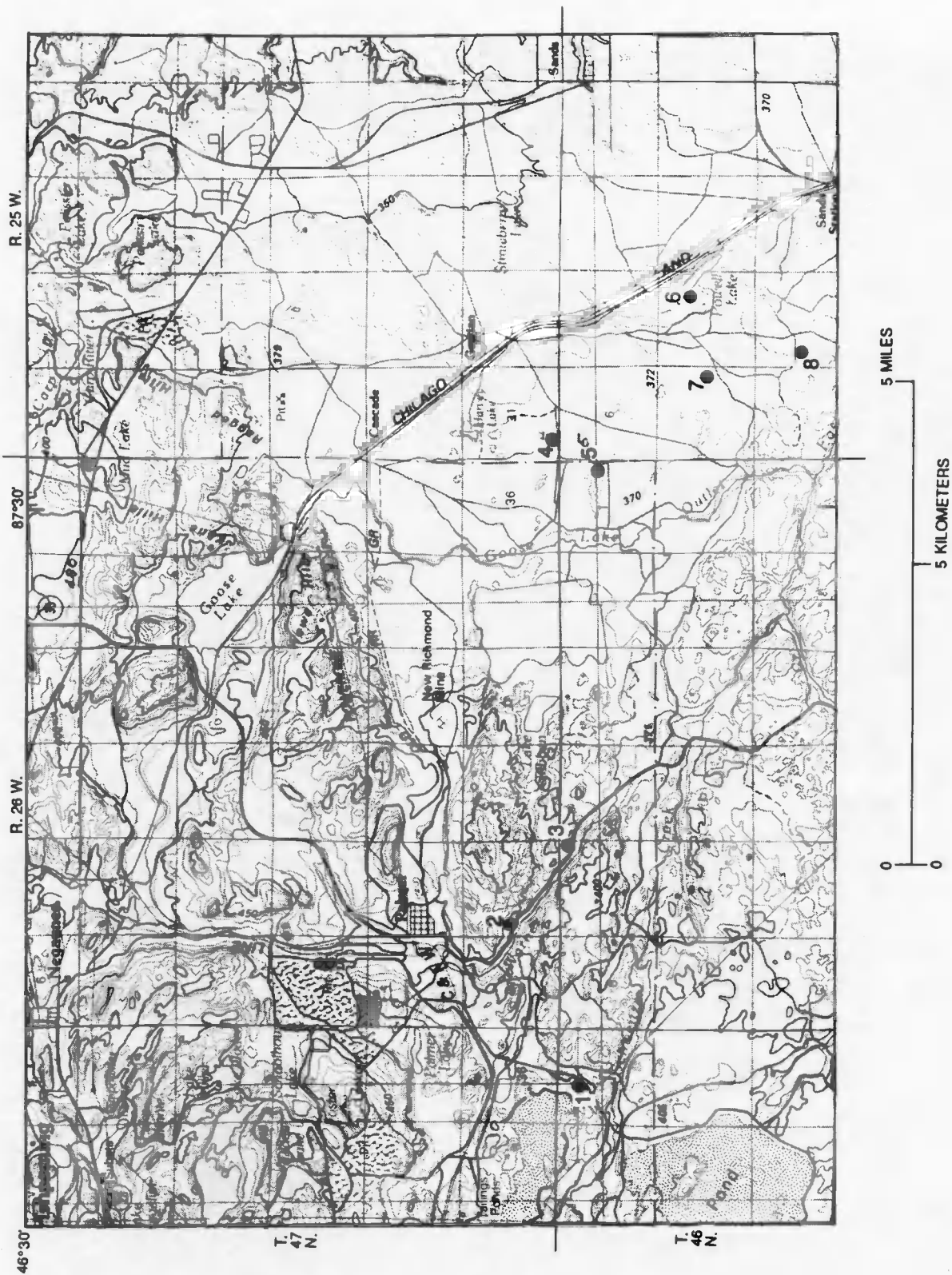


Figure 6. Field trip stops. Marquette and most of Michigan highway 28 are just north of the map area.

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