# The Gogebic Iron Range—A Sample of the Northern Margin of the Penokean Fold and Thrust Belt

Professional Paper 1730

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**U.S. Department of the Interior**
DIRK KEMPTHORNE, Secretary

**U.S. Geological Survey**
Mark D. Myers, Director

U.S. Geological Survey, Reston, Virginia: 2008

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# The Gogebic Iron Range—A Sample of the Northern Margin of the Penokean Fold and Thrust Belt

By William F. Cannon,[1] Gene L. LaBerge,[2] John S. Klasner,[3] and Klaus J. Schulz[1]

## Abstract

The Gogebic iron range is an elongate belt of Paleoproterozoic strata extending from the west shore of Lake Gogebic in the upper peninsula of Michigan for about 125 km westward into northern Wisconsin. It is one of six major informally named iron ranges in the Lake Superior region and produced about 325 million tons of direct-shipping ore between 1887 and 1967. A significant resource of concentrating-grade ore remains in the western and eastern parts of the range.

The iron range forms a broad, gently southward-opening arc where the central part of the range exposes rocks that were deposited somewhat north of the eastern and western parts. A fundamental boundary marking both the tectonic setting of deposition and the later deformation within the Penokean orogen lies fortuitously in an east-west direction along the range so that the central part of the range preserves sediments deposited north of that boundary, whereas the eastern and western parts of the range were deposited south of the boundary. Thus, the central part of the range provides a record of sedimentation and very mild deformation in a part of the Penokean orogen farthest from the interior of the orogen to the south. The eastern and western parts of the range, in contrast, exhibit a depositional and deformational style typical of parts closer to the interior of the orogen. A second fortuitous feature of the iron range is that the entire area was tilted from 40° to 90° northward by Mesoproterozoic deformation so that the map view offers an oblique cross section of the Paleoproterozoic sedimentary sequence and structures. Together, these features make the Gogebic iron range a unique area in which to observe (1) the lateral transition from deposition on a stable platform to deposition in a tectonically and volcanically active region, and (2) the transition from essentially undeformed Paleoproterozoic strata to their folded and faulted equivalents.

Paleoproterozoic strata in the Gogebic iron range are part of the Marquette Range Supergroup. They were deposited unconformably on Neoarchean rocks consisting of a diverse volcanic suite (the Ramsay Formation) which was intruded by granitic rocks of the Puritan Quartz Monzonite. The Marquette Range Supergroup in this region consists of a basal sequence of orthoquartzite (Sunday Quartzite) and dolomite (Bad River Dolomite), both of which are part of the Chocolay Group. The group is preserved only in the eastern and western parts of the range but was probably present throughout before the erosion interval that separated it from the overlying Menominee Group. The Menominee Group consists of basal clastic rocks (Palms Formation) that grade upward into the Ironwood Iron-Formation, which is the principal iron-bearing unit of the range. The Ironwood interfingers with the Emperor Volcanic Complex in the eastern part of the range and with volcanic rocks and gabbro in the western part of the range. The Ironwood is overlain unconformably by the Tyler Formation in the central and western parts of the range and by the Tyler's equivalent, the Copps Formation, in the eastern part of the range.

Strata in the central part of the iron range are entirely sedimentary. Deposition occurred in a relatively stable tectonic setting, at least until the deposition of the Tyler Formation. The Tyler consists largely of turbidites deposited in a foreland basin in advance of accreting volcanic arcs to the south. Penokean deformation in the central part of the range was very minor; the evidence of deformation consists of steep faults with small offsets and a few bedding-parallel faults that also have small offsets and that are recognized only in mine workings. In both the eastern and western parts of the iron range, abrupt facies changes mark a passage into a more tectonically and volcanically active belt. These relationships are especially well displayed in the east where a graben, the Presque Isle trough, began to subside during deposition of the Ironwood Iron-Formation. The thickness of the Ironwood increases into the graben and its internal stratigraphy also changes. The most prominent changes in the graben are the presence of a thick volcanic unit, the Emperor Volcanic Complex of the Menominee Group, and comagmatic gabbro sills that interfinger with the Ironwood. In the western part of the range, volcanic rocks and comagmatic gabbro sills are also present in the Ironwood, but a graben that is equivalent to the Presque Isle trough is not evident.

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Penokean structures are well developed in both the eastern and western parts of the iron range. They consist of folds ranging from outcrop to regional scale and thrust faults which, in places, either repeated the section or detached it from Neoarchean basement. The sharp transition from the little-deformed central part of the range to the more intensely deformed eastern and western parts coincides closely with the earlier developed transition from the stable sedimentary setting in the central part to the tectonically active sedimentation in the east and west parts. The extensional structures that formed during sedimentation may have helped to control the extent of later Penokean compressional structures.

# Introduction

The Gogebic iron range is one of several geologically classic and economically significant iron ranges of the Lake Superior region (fig. 1). The Gogebic iron range is an arcuate east-west-trending belt in northern Michigan and Wisconsin that spans the northern margin of the Penokean fold-and-thrust belt and thus includes rocks deformed by the 1880–1830 Ma Penokean orogeny as well as rocks that remained largely undeformed. Paleoproterozoic metasedimentary and metavolcanic rocks, known collectively as the Marquette Range Supergroup, make up the iron range, but bounding Neoarchean basement rocks to the south and Mesoproterozoic rocks of the Midcontinent Rift on the north are also discussed.

The Gogebic iron range has been studied and described in several generations of investigations, largely because of its former importance as a major iron-mining district. Studies of the entire range include Irving and Van Hise (1892), Allen and Barrett (1915) and Aldrich (1929). Numerous more local studies followed these early investigations. The intent of this paper is to summarize previous studies and add new information based on studies by the authors between 1990 and 1996, particularly in the eastern and western parts of the range; and to present a new stratigraphic and structural synthesis for the Paleoproterozoic strata and the Penokean foreland tectonic events that deformed them. Special emphasis is placed on the stratigraphic and structural relationships across the Penokean deformational boundary. The manuscript was prepared in 2003 and in the intervening period until publication in 2008 several advances in understanding the regional setting of the Gogebic iron range have been made, which would modify some of the interpretive parts of this paper. But details of relationships within the range are not substantially changed since 2003.

The Paleoproterozoic strata of the Marquette Range Supergroup are at least partly correlative with the Huronian Supergroup of Ontario and the Animikie Group and Mille Lacs Group of Minnesota. Together, these sequences constitute an assemblage of sedimentary and lesser volcanic rocks deposited on Neoarchean basement rocks along the south-facing continental margin of the Superior Province (fig. 1). They record the transition from stable cratonic sedimentation in the oldest units, through extensional continental margin sedimentation, to deposition in a foreland basin during accretion of volcanic arcs in the youngest units (see Morey, 1996, for a summary of Paleoproterozoic stratigraphy of the region). All of these rocks now lie within or immediately north of the fold-and-thrust belt of the Penokean orogen, which formed north of a suture marking the boundary of continental-margin rocks on the north with exotic volcanic terranes on the south.

In this paper, the Gogebic iron range is divided into three segments and each is discussed separately. The eastern segment extends from the easternmost exposures of the Copps Formation along the western shore of Lake Gogebic, westward to the Sunday Lake fault near Wakefield, Mich. (pl. 1*B*). The central segment, which includes all of the productive iron-ore district, continues from the Sunday Lake fault westward to Tyler Forks River near Upson, Wis. The western segment extends to the westernmost exposures of Paleoproterozoic strata where they are truncated by the Atkins Lake-Marenisco fault (pl. 1*A, B*). This report summarizes the geology of the central part of the iron range based mostly on previous studies, and presents new data and interpretations for the eastern and western parts of the range by combining previously published information with new field investigations conducted from 1990 to 1996. This report also documents that substantial changes in both sedimentation and structure occurred over distances of only a few kilometers along strike. In particular, sedimentation of the major iron-bearing section, including the Ironwood Iron-Formation, is shown to span a tectonic boundary and to pass laterally through an interfingering relationship into volcanic rocks of the Emperor Volcanic Complex in the east.

Because the Gogebic iron range lies obliquely across the transition from essentially undeformed rocks on the north to substantially deformed rocks to the south, it provides the opportunity to document the nature of the early deformational phases of the orogeny.

The iron range is part of the Michigamme subterrane of the Penokean fold-and-thrust belt terrane, and is the northernmost exposed subterrane of the continental margin assemblage. Rocks of the Michigamme subterrane were deformed largely by thin-skinned tectonism, which did not greatly affect the Neoarchean basement. Where the Neoarchean basement was involved in Penokean deformation, it was by movement of fault blocks, rather than by penetrative deformation. The range is located within a very foreshortened part of the fold-and-thrust belt. The range lies only 50 to 60 kilometers (km) north of the Niagara fault, which is the suture between the continental margin and the accreted arcs of the Wisconsin magmatic terranes to the south. South of the Gogebic iron range, Paleoproterozoic strata become much more intensely deformed and metamorphosed in the subterranes located successively closer to the suture zone.

The Gogebic iron range presents a unique opportunity to examine lithologic and structural changes and interpret processes and events across a Paleoproterozoic boundary. These processes range from stable cratonic sedimentation in

**Figure 1.**  Generalized geologic map of the upper Great Lakes region, U.S. and Canada, showing the location of the Gogebic iron range and other Paleoproterozoic iron ranges of the Lake Superior region.

the central part of the range to more tectonically and volcanically active sedimentologic processes in the eastern and western parts of the range. This same boundary later formed a tectonic front separating the rather intense Penokean deformation in the eastern and western parts of the range from the nearly undeformed rocks in the central part of the range. The present orientation of the Gogebic iron range also provides a unique view of Penokean deformation because of Mesoproterozoic monoclinal tilting. Crustal-scale listric thrusting in the interval 1,100 to 1,000 million years ago (Ma) tilted the iron range from 40° to 90° northward (Cannon and others, 1993). Subsequent erosion resulted in the current map pattern, which is essentially a longitudinal cross section of the range as it existed after Penokean deformation.

# General Geology

The Gogebic iron range is an arcuate belt of Paleoproterozoic sedimentary and volcanic rocks of the Marquette Range Supergroup extending from the western shore of Lake Gogebic in the Upper Peninsula of Michigan, westward for about 125 km into northern Wisconsin (fig. 1; pl. 1*B*). The range varies from less than 1 km to as much as 5 km wide and is bounded by Neoarchean granitic and metavolcanic rocks on the south and Mesoproterozoic volcanic and intrusive rocks on the north. In gross form, the range is a north-facing monocline dipping from about 40° to nearly 90°. In more detail, faulting and folding repeat the stratigraphic section in places, particularly in the eastern and western segments. For discussion purposes, the range consists of three segments. The central segment is characterized by well-defined and laterally continuous stratigraphic units, particularly within the Ironwood Iron-Formation, and by a lack of intense Penokean deformational features. In the eastern segment, the detailed internal stratigraphy of the Ironwood is lacking, significant changes in stratigraphic thickness occur over short intervals, the Ironwood interfingers with volcanic rocks of the Emperor Volcanic Complex, and sills of diabase are abundant. Penokean deformational features are also well developed and include folds and low-angle faults that have repeated the stratigraphic section. The western part of the range lies west of Tyler Forks River (pl. 1*A*). In this part of the range, the internal stratigraphy of the Ironwood is obscure, folding and faulting are intense in many areas, and large sills of diabase within the Ironwood and Palms Formations are common in the westernmost portions of the area. The stratigraphic relations within the Marquette Range Supergroup and its overlying and underlying units are shown in a schematic stratigraphic section in figure 2.

## Neoarchean Rocks

Granitic and volcanic rocks of Neoarchean age bound the Paleoproterozoic sequence of the Gogebic iron range on the south and constitute the basement on which it was deposited

unconformably. The Neoarchean rocks were best described by Schmidt (1976) and Greathead (1975) for part of the central Gogebic iron range and by Sims and others (1977) for the eastern part of the area. They were also mapped in varying levels of detail by Trent (1973), Prinz and Hubbard (1975), and Klasner and others (1998). Except in the easternmost part of the area, a scarcity of outcrops has hampered detailed examinations of the Neoarchean rocks.

A variety of volcanic rock is present and collectively forms the Neoarchean Ramsay Formation as defined by Schmidt (1976). The Ramsay includes thick sequences of pillowed to massive metabasalt (greenstone), particularly in the east. This metabasalt sequence is uniformly south-facing, as indicated by the abundant pillows, and dips steeply. When the Mesoproterozoic northward tilting is removed, the metabasalt sequence is restored to a north-dipping and downward-facing section that apparently was part of the overturned limb of a recumbent fold that formed during the Neoarchean (Klasner and others, 1998). Other units of the Ramsay Formation include fragmental volcanic rocks of mafic to felsic composition, volcanogenic graywackes, and minor argillite. Between Hurley and Upson, Wis., the volcanic rocks are at least 6 km thick and range in composition from basalt to rhyolite; dacite breccia is the most common rock directly beneath the Paleoproterozoic unconformity (Greathead, 1975). In most areas, the Ramsay has undergone only greenschist-facies metamorphism; however, locally, especially near the contact with the Puritan batholith, the Ramsay has been metamorphosed to amphibolite facies.

Granitic rocks of the Neoarchean Puritan Quartz Monzonite (Schmidt, 1976) form a batholith that intrudes the Ramsay Formation. The Puritan Quartz Monzonite (pl. 1*B*) is mostly massive to porphyritic, locally weakly foliated, and contains abundant pegmatitic segregations. A Rb-Sr isochron age of 2,710±140 Ma was determined by Sims and others (1977) for massive granitic rocks of the Puritan Quartz Monzonite. Strongly gneissic rocks are common, especially near the adjacent volcanic rocks of the Ramsay Formation, and have been given local names in several places (Fritts, 1969; Schmidt, 1976). The locally gneissic units are not distinguished from the more massive phases of the batholith on illustrations in this chapter because outcrops are insufficient to do so in most of the area and because the emphasis of the report is on Paleoproterozoic features.

## Paleoproterozoic Rocks

Paleoproterozoic sedimentary and lesser volcanic rocks constitute the Marquette Range Supergroup, which includes the Ironwood Iron-Formation, the principal iron-bearing strata of the Gogebic iron range. The Marquette Range Supergroup is composed of three groups that are separated from each other by regional unconformities (fig. 2). From oldest to youngest they are: (1) the Chocolay Group, composed of the basal Sunday Quartzite and overlying Bad River Dolomite; (2) the

Menominee Group, composed of the Palms Formation and the overlying Ironwood Iron-Formation, as well as the Emperor Volcanic Complex; and (3) the Baraga Group, composed of the Tyler Formation in the western and central part of the iron range and the laterally equivalent Copps Formation in the eastern part of the iron range. Paleoproterozoic intrusive rocks (diabase and gabbro dikes and sills) are also common throughout the iron range. Dikes are abundant in the Neoarchean rocks and constitute two swarms trending northeast and northwest. Dikes are also common in the Chocolay and Menominee Groups. These latter dikes were identified mostly in extensive underground mine workings and are very seldom seen in natural exposures. Large diabase sills occur mostly in the eastern and western extremes of the range and mainly intrude the Ironwood Iron-Formation and Emperor Volcanic Complex.
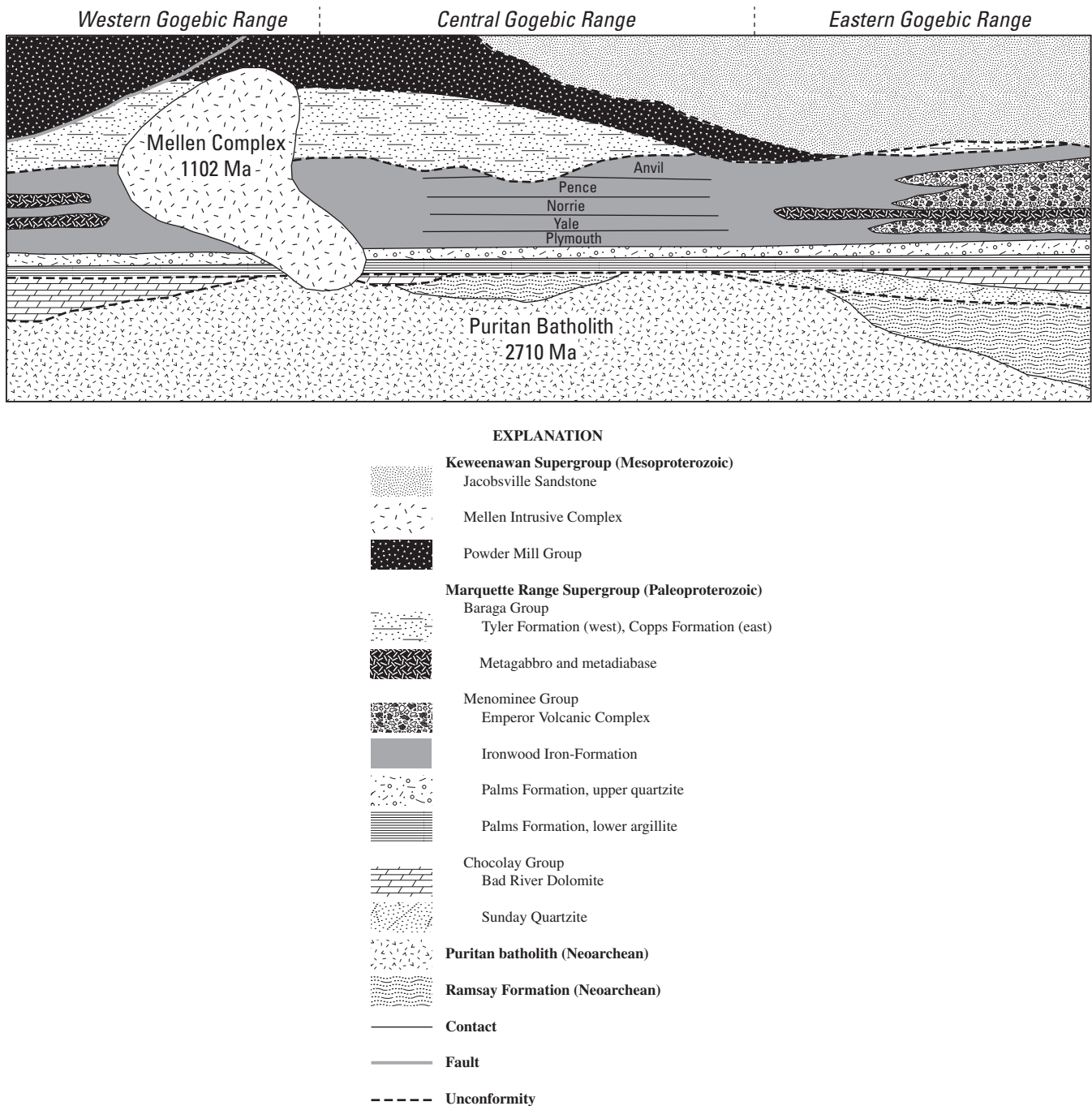


**Figure 2.**   Schematic stratigraphic section along approximately 150 km of the Gogebic iron range illustrating the stratigraphic relationships within the Menominee Group, and the relation of the Menominee Group to other stratigraphic sequences in the region. The base of the Menominee Group is used as a horizontal datum. Five members of the Ironwood Iron-Formation are shown in the central Gogebic iron range.

Detailed descriptions of the Marquette Range Supergroup strata are given in papers by Irving and Van Hise (1892), Van Hise and Leith (1911), Allen and Barrett (1915), Hotchkiss (1919), Aldrich (1929), Huber (1959), Alwin (1976), and Schmidt (1980). The following brief descriptions are based largely on these earlier reports and are supplemented by our own observations. This report emphasizes features that help explain the contrast between the depositional and tectonic styles of the central part of the range and those of the eastern and western parts.

## Chocolay Group

The Chocolay Group is absent from the central part of the Gogebic iron range but is present as more or less continuous strata in both the eastern and western parts. The group includes the basal Sunday Quartzite, an orthoquartzite that grades upward into dolomite and dolomitic marble of the Bad River Dolomite.

### Sunday Quartzite

In the eastern part of the Gogebic iron range, the Sunday Quartzite is the basal Paleoproterozoic unit and forms an apparently continuous belt about 9 km long (Klasner and others, 1998). The eastern part of this outcrop belt is shown in figure 3. The lower part of the Sunday is a reddish, prominently crossbedded orthoquartzite that contains conglomeratic layers of quartz and granitoid cobbles as large as 8 centimeters (cm) diameter near its base. The rest of the formation is a gray, vitreous quartzite with common crossbeds and local current ripple marks. Schmidt (1980) reported that the Sunday is as much as 46 meters (m) thick, and grades upward into the Bad River Dolomite. Perhaps the best exposure of the Sunday Quartzite is on the southwest side of the hill in the SW¼ sec. 18, T. 47 N., R. 44 W. (Klasner and others, 1998), west of the area shown in figure 3. Some of the crossbeds are of the herringbone type, where successive crossbeds are oriented 180° to each other. These units, along with numerous mud-chip conglomerates and mud-cracked units, form a series of stacked deposits that are each about 2 m thick. Each sequence starts with a mud-cracked unit, is overlain by a herringbone-cross-bedded unit, and terminates with a parallel-bedded unit. All of these features suggest that the Sunday Quartzite was deposited in a tidal environment (Ojakangas, 1983). The Sunday Quartzite is not known in the central and western parts of the Gogebic iron range where either the Bad River Dolomite or Palms Formation lies directly on Neoarchean basement rocks instead.

### Bad River Dolomite

The Bad River Dolomite is conformable with and gradational into the Sunday Quartzite in the eastern Gogebic iron range. The transition is marked by several meters of interbedded quartzite and dolomite; dolomite beds become thicker and more abundant upward. The Bad River contains beds and irregular patches of gray to black chert and stromatolitic layers and mounds that range from 5 to 50 cm across. In the eastern Gogebic iron range, the Bad River weathers to a distinctive brown color, suggesting that the dolomite contains a significant amount of iron. Schmidt (1980) reported that the dolomite reaches a maximum thickness of about 120 m in the eastern Gogebic iron range. The formation thins westward and was either eroded or was not deposited from the Wakefield, Mich. area westward for a distance of about 75 km.

The Bad River reappears in the western Gogebic iron range, west of the Tyler Forks River in Wisconsin (pl. 1*A*), as apparently isolated lenses of dolomite resting on Neoarchean basement; most are too small to show at the current map scale. A unit as much as several tens of meters thick of variegated chert breccia overlies Neoarchean basement from the Tyler Forks River for about 20 km westward (pl. 1*A*). The unit is interpreted to be a residual deposit of chert fragments resulting from the weathering and dissolution of the cherty dolomite and the accumulation and slight reworking of the residuum. Although this unit is included in the Palms Formation (described below) in this report, its presence indicates that the Bad River was once extensive in the western part of the Gogebic iron range and its absence results from weathering and erosion prior to deposition of the Palms Formation.

Near Mineral Lake (pl. 1*A*) and near Atkins Lake, Wis., at the far western end of the range, the Bad River reaches a thickness of about 300 m (Cannon and others, 1996). Near Atkins Lake the Bad River consists of gray, dolomitic marble containing abundant lenses and layers of chert. Stromatolitic beds alternate with thinly laminated or massive dolomite. Most of the stromatolites exposed in the abandoned quarry in the NW¼ sec. 22, T. 44 N., R. 5 W. (pl. 1*A*) are low mounds less than 25 cm across. However, a large glacial boulder several kilometers to the southwest contains silicified stromatolites nearly a meter across (fig. 4).

The basal portion of the Bad River crops out in the SE ¼ sec. 15, T. 44 N., R. 5 W. (pl. 1*A*). Here, the Bad River is an arkosic quartz-pebble conglomerate that rests directly on Neoarchean granitic rocks. The conglomerate grades up through about 10 m into arkosic dolomite and eventually into the cherty dolomite that constitutes the bulk of the unit.

Metamorphism in the Atkins Lake area resulted in extensive development of actinolite and tremolite from the reaction between the dolomite and chert. The metamorphism occurred within the contact aureole of the Mesoproterozoic gabbro of the Mineral Lake intrusion.
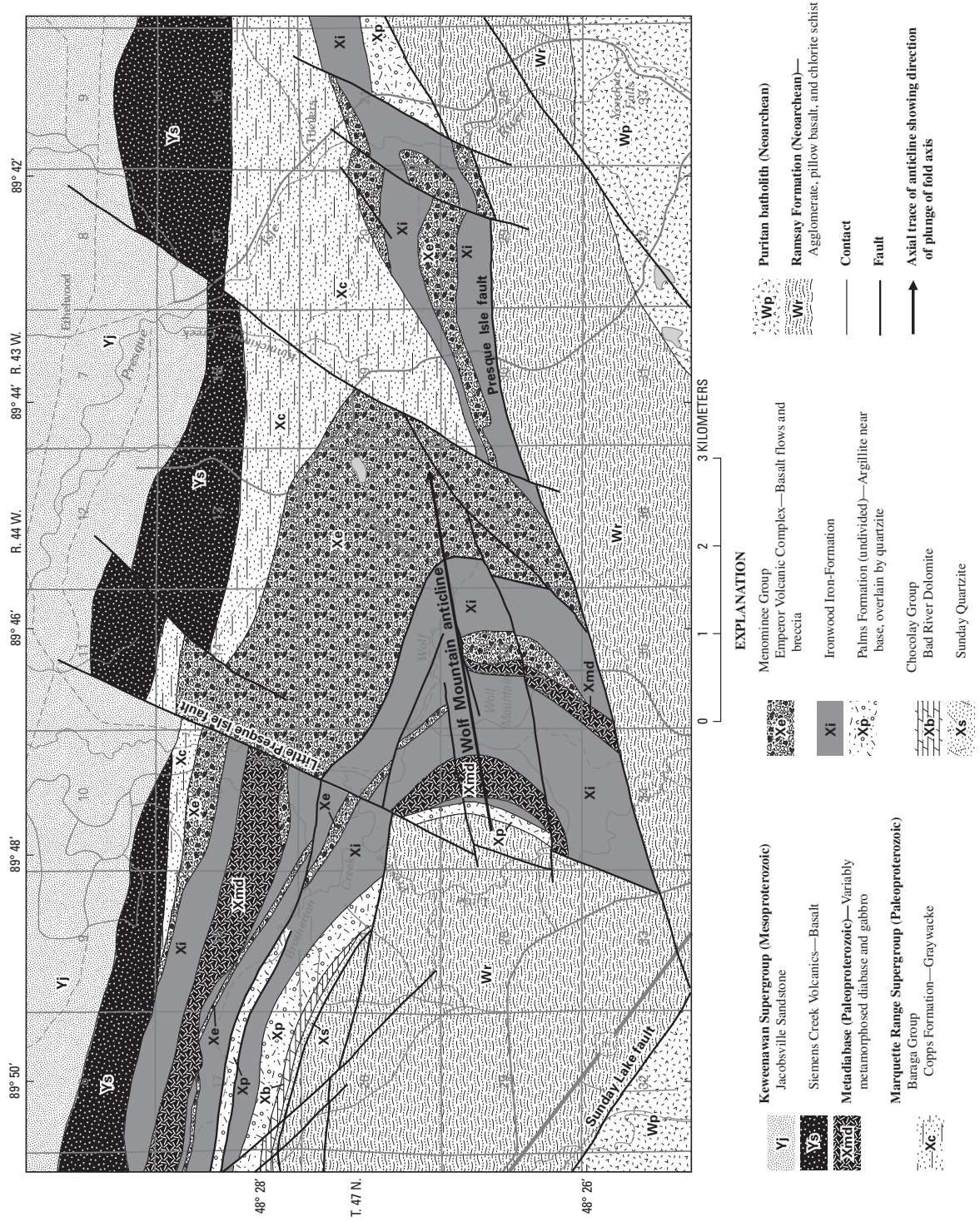
**Figure 3.**    Geologic map of part of the eastern Gogebic Range. Generalized from Klasner and others (1998).
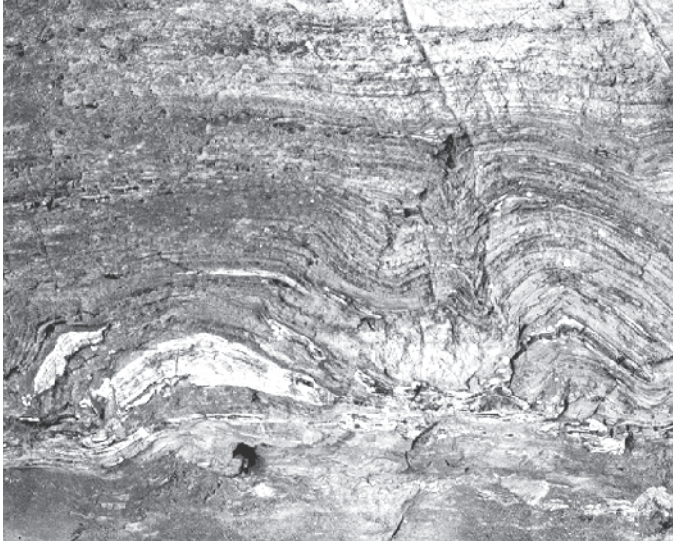
**Figure 4.**    Stromatolites in Bad River Dolomite. Photograph is of a large boulder in NE¼, sec. 5. T. 43 N., R. 5 W. Field of view about 0.5 m across.

## Menominee Group

Strata of the Menominee Group are present throughout the western and central parts of the Gogebic iron range and are absent only in the eastern half of the eastern segment. The group consists of the Palms Formation and the conformably overlying Ironwood Iron-Formation. In the eastern part of the range, volcanic rocks of the Emperor Volcanic Complex, which interfinger with the Ironwood, are included in the Menominee Group.

### Palms Formation

The Palms Formation unconformably overlies Chocolay Group rocks on the eastern and western ends of the Gogebic iron range, and overlies Neoarchean rocks in the central part of the range. We have shown only two map units within the Palms on plate 1*A* although a regionally consistent internal stratigraphy is present. The principal part of the Palms Formation is a coarsening upward succession of argillite near the base (the "quartz slate" unit of older literature) grading to quartzite near the top (Aldrich, 1929). Ojakangas (1983) refined the description, showing that the Palms has a lower argillaceous unit, a central mud-silt-sand unit, and an upper sand-rich unit; he also stated that the Palms displays a bedding pattern and style that suggest deposition in a tidal environment. These distinctive lithologic subdivisions of the Palms are readily recognizable throughout the length of the Gogebic iron range suggesting that the depositional environment of the Palms was exceptionally uniform over the region.

We have defined an additional unit at the base of the Palms that occurs throughout a 15-km-long belt in the central part of the range, from sec. 1, T. 44 N., R. 2 W., to sec. 16, T. 44 N., R. 3 W., and in much of the eastern part of the range. This basal unit consists of a breccia of variegated angular to

subrounded clasts of chert as much as 40 cm long in a siliceous matrix. Rounded quartz pebbles and sand grains are present locally in the matrix between the chert clasts. This rock type is interpreted to be a residuum of chert nodules and segmented layers, originally contained in the Bad River Dolomite; the residuum formed as a result of prolonged weathering and dissolution of dolomite prior to deposition of the overlying Menominee Group sediments. In the east, the basal breccia overlies the Bad River Dolomite. In the central part of the iron range, a thin unit of dolomite locally underlies the breccia, but most commonly the breccia directly overlies Neoarchean basement rocks.

Magnetite and hematite concentrations are present locally at the top of the basal chert breccia unit. Shallow exploration shafts were excavated into some of these magnetite-rich zones during the early days of exploration for iron ore in the region. Several such exploration shafts are on the ridge extending east from Mount Whittlesey in Wisconsin; several others are located 1 km west of where the Bad River transects the unit (known informally as Penokee Gap). The presence of these magnetite concentrations within sandy chert breccia units on an erosion surface suggests that they are paleo-placer deposits and that their intermittent occurrence indicates deposition in streams. Chemical analyses of several of the deposits failed to show any heavy minerals other than magnetite. The magnetite presumably was derived from Neoarchean rocks exposed during the erosion interval. The probable paleo-placer deposits support the interpretation of Klasner and LaBerge (1996) that the chert breccia is a lag accumulation on an erosion surface. Although the regional unconformity developed on the Bad River Dolomite has been known since the work of Irving and Van Hise (1892), a recent interpretation by Sims and others (1989) suggests that the erosion was related to slight rotation of blocks of Neoarchean basement along the Penokean continental margin.

### Ironwood Iron-Formation

The Palms Formation grades upward over an interval of several meters into the Ironwood Iron-Formation, marking a change from clastic to chemical sedimentation. The Ironwood, like iron-formations elsewhere in the Lake Superior region, has two distinct lithologies and bedding styles, as described by Huber (1959): (1) "wavy-bedded ferruginous chert" consisting of irregular beds and lenses of granular or oolitic chert (a few millimeters to several centimeters thick) which are separated by thin laminae of more evenly bedded, iron-rich material; the iron-rich laminae consist largely of iron oxides, silicates (mostly minnesotaite and stilpnomelane), and carbonate; and (2) "even-bedded ferruginous slate" in which the beds are extremely regular and finely laminated. The chert is generally dense and flinty. These two lithologies are generally thought to reflect different depths of water in which the iron-formation accumulated; the wavy-bedded iron-formation was deposited in shallow (agitated) water, and the laminated iron-formation was deposited in deeper (quieter) water.

Based on a preponderance of one or the other of these different bedding styles, Hotchkiss (1919) divided the Ironwood within the previously productive central part of the iron range into five members. These are, from the base upward, the Plymouth, Yale, Norrie, Pence, and Anvil Members (fig. 2). The Plymouth Member is 40 to 46 m thick and consists mainly of irregularly bedded, granular iron-formation containing thin intraformational conglomerates. Stromatolitic units are present locally. The overlying Yale Member is 14 to 22 m thick, has a 3-m-thick basal unit of pyritic shale (suggesting periodic anoxic conditions), and consists mainly of laminated chert-carbonate iron-formation and some granular or intraformational conglomerate units. Overlying the Yale Member is the Norrie Member which is 34 to 36 m thick and consists mainly of irregularly bedded, granular iron-formation. The 30- to 37-m-thick Pence Member consists mainly of laminated chert-carbonate-silicate iron-formation containing some irregularly bedded, granular units. The uppermost unit, the Anvil Member, is a 16-m-thick, magnetite-rich, laminated chert-carbonate iron-formation containing some irregularly bedded, granular horizons. Additional details of the stratigraphic subdivisions of the Ironwood were provided by Huber (1959).

The central part of the range has few natural outcrops so our understanding of it depends largely on past studies such as by Hotchkiss (1919), Aldrich (1929), and Huber (1959), that were conducted when extensive underground mining and exploration drilling was taking place, and from the more recent summary by Schmidt (1980). Few new observations of the Ironwood were made in the central part of the iron range during this study. The older data show that the five members of the Ironwood are persistent in character, although of variable thickness, from near Wakefield, Mich., westward for about 70 km to the Tyler Forks River west of Upson, Wis. Except for the presence of some ferruginous argillite horizons and pyritic shale units, the Ironwood contains little detrital or volcanic material throughout this belt. It should be noted that many authors of early reports on the Gogebic iron range used the term "slate" for thinly laminated or fissile, fine-grained rocks, even where cleavage is completely lacking. The common usage of slate as a lithologic term for much of the Gogebic iron range is misleading because most of these rocks do not possess a penetrative deformational fabric, but rather are simply finely laminated sedimentary rocks.

East of Wakefield, Mich., the stratigraphy of the Ironwood changes abruptly. A unit of ferruginous argillite from 30 to 50 m thick that appears within the lower part of the Ironwood immediately east of the Sunday Lake fault indicates the deposition of clastic material that is absent in the central part of the range (Huber, 1959). More strikingly, the Emperor Volcanic Complex (Trent, 1976) is interfingered with the iron-formation in and near the Presque Isle trough. In addition to the volcanic rocks, the Ironwood Iron-Formation contains significant detrital material in the eastern Gogebic iron range. Allen and Barrett (1915) reported that an argillite unit up to 147 m thick is within the Ironwood in secs. 16, 17 and 18, T. 47 N., R. 44 W. An argillite unit of unknown thickness inter-

bedded within the Ironwood is also exposed beneath a mafic sill on the west side of the hill in the NW¼ sec. 26, T. 47 N., R. 44 W.

Both the abrupt changes in the stratigraphy of the Ironwood east of Wakefield, and the rift-related volcanic activity in a fault-bounded basin that included a significant influx of detrital material document a tectonically unstable environment during deposition of the Ironwood in that area. Inasmuch as the underlying Palms Formation retains its internal lithologic subdivisions throughout the area, the rifting, subsidence, and associated volcanism apparently did not begin until the time of deposition of the Ironwood Iron-Formation.

## Relation between the Ironwood Iron-Formation and the Emperor Volcanic Complex

Volcanic rocks and mafic sills of the Emperor Volcanic Complex (Trent, 1976) are a significant component of the stratigraphic section in the eastern Gogebic iron range. Furthermore, the various subdivisions of the Ironwood, developed so prominently in the central part of the range, are not as clearly recognized to the east. Even at mines such as the Sunday Lake and Vicar within a few kilometers east of the Sunday Lake fault, the correlation of the internal stratigraphy of the Ironwood was controversial among mining geologists (Huber, 1959). Volcanic rocks of the Emperor Volcanic Complex intertongue with the Ironwood and provide proof that the Ironwood was deposited at the same time that volcanism was occurring in this region. The Little Presque Isle fault (fig. 3) marks a prominent change in the lithology of the Ironwood and in the thickness of the Emperor. West of the fault, the Ironwood is mostly irregularly bedded, granular iron-formation typical of shallow-water deposits. East of the fault, the Ironwood is mainly the laminated variety typical of deeper water deposits. The Little Presque Isle fault also marks a prominent change in the distribution and thickness of the Emperor Volcanic Complex. West of the fault, the Emperor consists of a mafic sill and several thin basalt flows interbedded with the Ironwood. East of the fault, the Emperor thickens abruptly to as much as 2,000 m and is composed dominantly of hyaloclastites, pillow breccias, and related debris flows (fig. 5) as well as lesser pillowed basalts, crackle breccias, and massive basalt.

Exposures on the hill in the NE¼ sec. 24, T. 47 N., R. 44 W. (fig. 4), are representative examples of the rock types in the Emperor. A zone containing 2- to 3-m-long pillows is overlain by a debris flow that has a matrix of 0.5- to 2.0-cm-long hyaloclastite fragments in which both pillow fragments and complete pillows (commonly containing quench cracks) are suspended. Scattered pods and patches of quartz and jasper are also present in the debris flows. Many of the pillow fragments and most of the hyaloclastite fragments are composed primarily of zoisite. Larger pillow fragments may have a zoisite rind and an epidote core, which suggests leaching of iron from the fragments. The deposits are interpreted to be debris that flowed off a subaqueous lava dome (or domes). The abrupt change in stratigraphy across the Little Presque Isle fault sug-

gests that it was a growth fault along the northwest side of a graben that formed during Ironwood deposition. Farther east, the Emperor thins toward the Presque Isle fault, which suggests that the fault bounds the southeastern side of the graben (fig. 3) (Klasner and others, 1998) that is herein named the Presque Isle trough.

Chemically, the Emperor consists of rift-related tholeiite (Sims and other, 1989) and mafic rocks, all of which form the vast majority of the volcanic sequence; however, subordinate amounts of volcanic rocks composed mainly of fine sericite with well-preserved shard structures and quartz phenocrysts suggest that felsic volcanic rocks are at least a minor component. Dan (1978) and Prinz (1981) report finding rocks as felsic in composition as dacite in the Emperor. More details of the chemistry of the Emperor Volcanic Complex are given below in the section on "Paleoproterozoic Intrusive Rocks."

## Stratigraphic Relations in the Menominee Group and Emperor Volcanic Complex

Both the abrupt changes in the stratigraphy of the Ironwood Iron-Formation east of Wakefield, Mich., and the rift-related volcanic activity in a fault-bounded basin that includes a significant influx of detrital material document a tectonically unstable environment of deposition of the Ironwood in that area. Because the underlying Palms Formation retains its internal lithologic subdivisions throughout the area, the rifting, subsidence, and associated volcanism apparently did not begin until the time of Ironwood deposition.

Similar stratigraphic relations within the Ironwood Iron-Formation are present on the western end of the Gogebic iron range. Details of the geology of the western part of the range are shown on plate 1A. The five-fold internal stratigraphy of



**A**
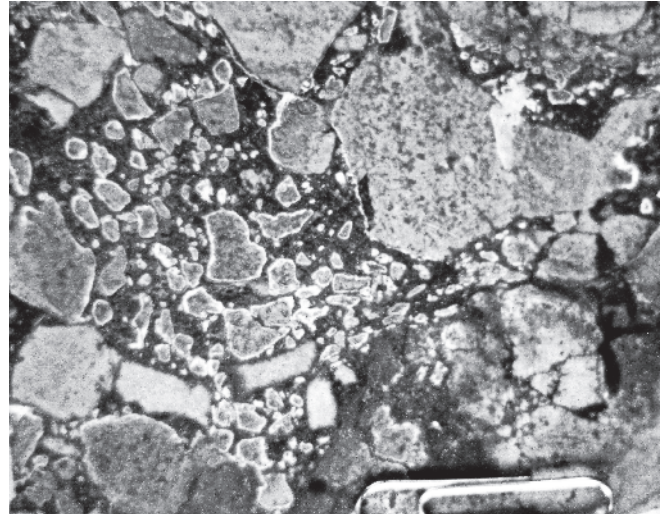
**B**
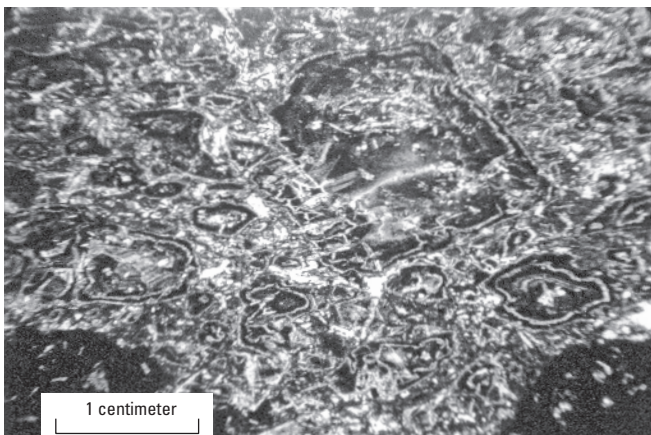
**C**
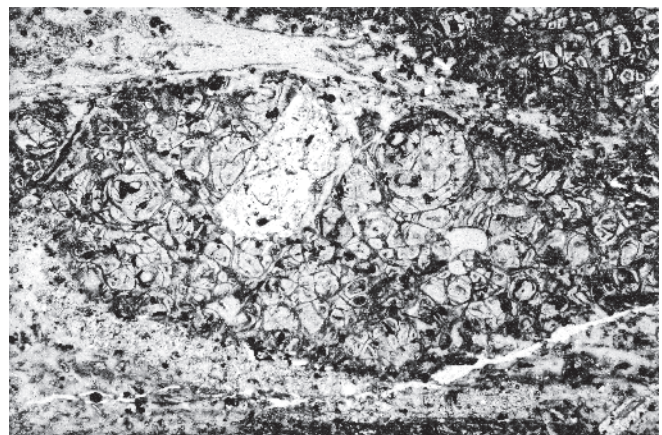
**D**

**Figure 5.**   Examples of typical rocks types of the Emperor Volcanic Complex. *A*, Debris flow containing pillow breccia fragments in a finer-grained hyaloclastite matrix from outcrop near Wolf Mountain. *B*, Polished surface of debris flow breccia. Paper clip is 3 cm long. *C*, Photomicrograph of hyaloclastite breccia.*D*. Photomicrograph showing perlitic cracks within a clast contained in debris flow.

the Ironwood in the central part of the range cannot be traced west of the village of Tyler Forks in sec. 32, T. 45 N., R. 1 W. The difficulty in tracing members of the Ironwood may result (in part) from an increased metamorphic grade toward the west within the contact aureole of the Mellen Intrusive Complex (Mesoproterozoic), and (in part) from an increasing complexity of internal folding and faulting in the Ironwood. In this study the various members could not be recognized in the western segment, even in the least metamorphosed and least deformed sections. The five-fold subdivision of the central part of the range probably does, indeed, lose its definition toward the west, perhaps as a result of somewhat more variable depositional environments in the western part of the range. Although the detailed studies conducted by U.S. Steel Corporation in the 1950s distinguished internal units in the Ironwood as far west as Penokee Gap on the Bad River (summarized by Marsden, 1978), the definition of the units was based, in part, on metallurgical characteristics. They did not correlate entirely with the classic five-member subdivision of the central part of the range.

Outcrops in the NE¼ sec. 20, T. 44 N., R. 5 W., in the Atkins Lake area (pl. 1*A*) contain both the laminated and granular varieties of the Ironwood Iron-Formation, which overlie at least 15 m of thick-bedded quartzite that is typical of the upper Palms Formation. The change in lithology is further displayed by laminated argillite as much as 15 m thick that forms interbeds in the lower part of the Ironwood. The argillite is succeeded by a unit of amygdaloidal basalt about 6 m thick containing numerous tabular to round, randomly oriented blocks of argillite and iron-formation as much as several meters long. Some of the fragments are relatively rounded and are internally intensely deformed; others are sharply angular and have little or no internal deformation. The basalt matrix adjacent to some of the larger fragments is discolored and contains more vesicles than is typical, which suggests that the basalt may have chilled against the fragments. A meter-thick zone of laminated, magnetic iron-formation (which may be a large slab included in the basalt) separates the basalt from an overlying, 10- to 20-m- thick metadiabase sill. The upper margin of the sill is chilled against a breccia unit containing chert-rich argillite clasts in an amygdaloidal basalt matrix similar to the unit beneath the sill. Local alteration rinds that are suggestive of pillows or pillow breccia are present in the basalt. Outcrops in sec. 20, T. 44 N., R. 5 W., have zones of hyaloclastite-bearing basalt.

At least three metadiabase sills are present in the Ironwood Iron-Formation in the Atkins Lake area. Their intrusion into the semi-consolidated argillite and iron-formation produced peperite-like features caused by the interaction of the basaltic magma with the wet sediments (Klasner and LaBerge, 1996). The peperite textures suggest that the sill emplacement was nearly contemporaneous with the iron-formation deposition. Several drill cores from near Trapper Lake (sec. 26, T. 44 N., R. 6 W.) (pl. 1*A*) about 5 km west along strike from Atkins Lake, intersected laminated and irregularly bedded iron-formation and interbedded turbidites, a 70-m-thick graphitic

argillite, and several metadiabase sills that had highly brecciated border zones similar to those exposed near Atkins Lake.

Collectively, these data suggest that the depositional environment of the Ironwood changed westward from a stable platform environment into an unstable basin. Igneous activity, turbidites, and at least periodic stagnant anoxic conditions accompanied iron-formation deposition. A comparison of the lithologic changes with those of the eastern Gogebic iron range suggests that the igneous activity and turbidites in the west may also have been a response to the formation of fault-bounded depositional basins. The presence of such structures, however, is not obvious within the present exposures. The Palms Formation retains its lithologic units and thickness both in the eastern Gogebic iron range and throughout the area, indicating that the unstable conditions and volcanism developed contemporaneously with the deposition of the Ironwood Iron-Formation.

Although volcanism that is contemporaneous with the deposition of iron-formation is well represented on the eastern and western ends of the range, volcanism has not been reported in the central part of the range where extensive mine openings and drilling have exposed essentially continuous sections of the Ironwood Iron-Formation. However, Paleoproterozoic metadiabase dikes and sills are common in the central part of the range where numerous dikes have cut the iron-formation at nearly right angles to bedding. Less commonly, sills were emplaced along the bedding planes. These rocks are discussed in more detail below in the section on Paleoproterozoic intrusive rocks. The dikes were, in fact, important features in controlling the location and distribution of iron- ore bodies in the productive central part of the range. The age of these dikes is not known other than that they are somewhat younger than the Ironwood Iron-Formation and possibly older than the Tyler Formation. These dikes may have been feeders for volcanic rocks higher in the section, but if so, such volcanic rocks have not been recognized. The dikes do not seem to be present in the overlying Tyler Formation. Therefore, the available evidence suggests that there was igneous activity over much of the length of the Gogebic iron range more or less at the same time as iron-formation deposition.

## Baraga Group

The Baraga Group, which contains the youngest Paleoproterozoic rocks in the Gogebic iron range, is a thick sequence of dominantly turbiditic rocks. These rocks overlie the Menominee Group and display a contact that varies from a definite unconformity in the east to a more problematic, but probably a low-angle unconformity or disconformity in the central and western parts of the range. This greywacke and slate sequence was assigned different names in various areas of Wisconsin and Michigan. In the central and western parts of the Gogebic iron range, it was named the Tyler Formation by Van Hise (1901). Rocks that are very similar lithologically and stratigraphically in the eastern Gogebic range were named Copps Formation by Allen and Barrett (1915). Although there

is little doubt that these two units are correlative, they are physically separated by only a few kilometers across an area where the Baraga Group has been removed by erosion, and thus they continue to maintain separate names. The sequence is thickest in the eastern and western parts of the range, but was completely removed by erosion prior to the deposition of the Keweenawan Supergroup (Mesoproterozoic) near Wakefield, Mich. (pl. 1B). The Tyler and Copps Formations are generally correlated with the Michigamme Formation, which is widely distributed in other parts of northern Michigan.

## Copps Formation

The Copps Formation crops out from near Wakefield, Mich., eastward to Lake Gogebic (fig. 3, pl. 1B). The Copps overlies the Ironwood Iron-Formation and Emperor Volcanic Complex in the western part of its outcrop area, and overlies Neoarchean granitoid rocks to the east toward Lake Gogebic. The Copps clearly overlies, with an unconformable contact, the older rocks in the eastern part of the Gogebic iron range, as indicated by the progressive eastward truncation of pre-Copps units (Klasner and others, 1998).

The basal part of the Copps is a brown-weathering quartzite about 5 m thick that contains several conglomeratic zones. Conglomerates near the Presque Isle River in sec. 20, T. 47 N, R. 43 W. (fig. 4) contain abundant chert and jasper clasts, along with some clasts of volcanic rocks, all of which are similar to the rocks of the underlying Ironwood Iron-Formation and Emperor Volcanic Complex. To the east, in secs. 23 and 24, T. 47 N., R. 43 W., where the Copps directly overlies Neoarchean granitoid rocks, the basal conglomerate contains abundant granitoid boulders as large as 40 cm in diameter.

The basal quartzite unit is overlain by a thick greywacke and slate sequence containing abundant graded beds. Sand-sized grains in some graded units consist of quartz, plagioclase, microcline, recrystallized chert, and sericitized volcanic rocks, all of which indicate a source area containing granitoid, sedimentary, and volcanic rocks. The coarser parts of the graded beds also commonly contain carbonate concretions. Individual turbidite units, which are generally no more than 30 cm thick, locally contain conglomeratic zones and are consistently north-facing.

## Tyler Formation

The Tyler Formation is present from Wakefield, Mich., westward (pl. 1B), and its preserved thickness increases to more than 3,000 m in Wisconsin. The Tyler is primarily a turbidite succession of intercalated graywacke and black slate, argillite, and chert containing abundant Bouma sequences (Alwin, 1976). Grains in the graywacke suggest a source area with a significant granitic component. Paleocurrent indicators suggest transport to the northwest (Alwin, 1976). The lower part of the Tyler, including as much as 100 m of strata known mainly from exploration drill holes and mine workings, consists of sideritic black slate and ferruginous conglomerates.

The nature and exact stratigraphic position of the Tyler's contact with the underlying Ironwood Iron-Formation has been in doubt for many years and various authors have argued for either a conformable or unconformable contact. Because the only exposures of the basal part of the Tyler were in mine workings (now long-abandoned) and in exploration drill holes, there has been no recent evidence to resolve this issue. A ferruginous (sideritic) shale unit which is as much as 90 m thick and contains as much as 25 percent iron, is present in the lower part of the Tyler Formation (Hotchkiss, 1919). Atwater (1938) stated that the ferruginous shale unit is present throughout most or all of the Gogebic iron range, and that a conglomeratic quartzite (the Pabst Member described later) forms the basal part of the Tyler Formation in the eastern part of the range. Siderite-chert iron-formation has also been widely reported in the basal 100 m of the Tyler (Hotchkiss, 1919).

## Pabst Member

Van Hise (1901) named the Tyler Formation and considered it to conformably overlie the Ironwood. However, Hotchkiss (1919, p. 18) suggested that the Pabst Member, a conglomerate with pebbles of iron oxide, quartz and jasper, was the basal unit of the Tyler Formation, and that the Pabst was deposited unconformably over the Ironwood. Hotchkiss' data came mainly from exposures in mine workings and drill cores in the productive central part of the Gogebic iron range. Aldrich (1929, p. 165–166) concurred with Irving and Van Hise (1892) that the Tyler conformably overlies the Ironwood. Atwater (1938) examined the relationship between the Ironwood and the Pabst Member as it is exposed in a number of mines between Iron Belt, Wis., and Wakefield, Mich., and concluded that an erosional surface existed between the Ironwood and Tyler Formations. His diagram (Atwater, 1938, p. 163), used as a basis for figure 6, shows that a substantial part of the upper Ironwood was eroded prior to deposition of the Pabst. Descriptions of the Pabst Member from Hotchkiss (1919) and Atwater (1938) suggest that it is similar to the conglomeratic quartzite that forms the base of the Copps Formation in the eastern part of the Gogebic iron range; however, Aldrich (1929) reported that a shaley unit forms the base of the Tyler, and that it underlies the ferruginous shale unit on the western end of the range, where an unconformity between the Ironwood and Tyler is uncertain. Finally, Schmidt and Hubbard (1972) and Schmidt (1980), based on a review of this older data, concluded that the contact between the Ironwood and Tyler Formations is most likely gradational and that sedimentation was continuous in the Bessemer, Mich. area. They state that the Ironwood grades up into a shale and that, in the Bessemer area, the Pabst Member occurs above the shale.

Based solely on the evidence of past reports and on regional stratigraphic relationships, we believe that the base of the Tyler Formation should be placed at the base of the Pabst Member and that a low-angle to disconformable contact exists between the Pabst and Ironwood. The findings of Atwater (1938) that the Pabst truncates the internal units of the Iron-
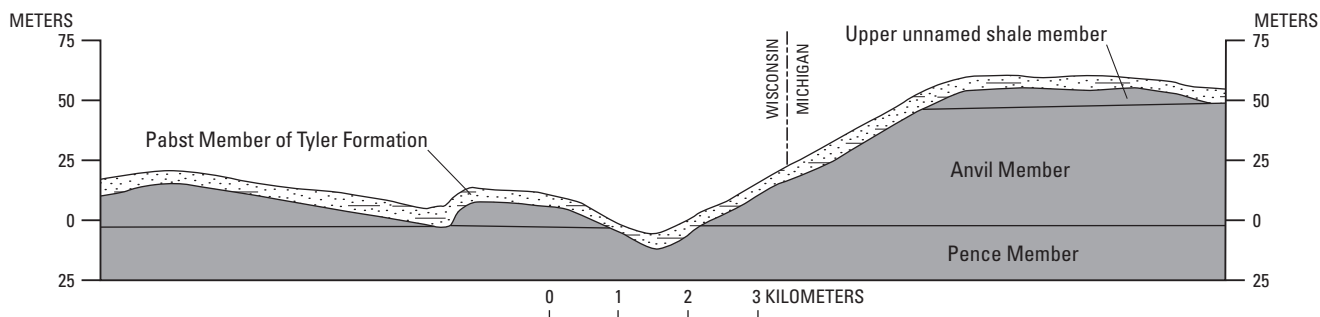
**Figure 6.** Schematic longitudinal section of part of the central Gogebic iron range illustrating the crosscutting relations between the Pabst Member of the Tyler Formation and underlying members of the Ironwood Iron-Formation. Simplified from Atwater (1937).

wood at a low angle seems to be compelling evidence of the unconformable relationship. The Ironwood may indeed have graded up into a shale as suggested by Schmidt and Hubbard (1972) and Schmidt (1980), but this shale is preserved only locally, as at Bessemer, where pre-Pabst erosion did not cut down into the Ironwood Iron-Formation. The interpretation that the Pabst is the basal conglomerate of the Baraga Group and that the group is unconformable on the underlying Ironwood Iron-Formation is fully consistent with the relation of the Copps to older units in the eastern part of the iron range and with relationships elsewhere in northern Michigan where an unconformity between the Baraga Group and older rocks is well established. We propose, therefore, to establish the Pabst Conglomerate as the basal member of the Tyler Formation. Further, we consider the relatively local units of black sideritic shale that underlie the Pabst Member as upper beds of the Ironwood Iron-Formation and mark a transition from the relatively shallow-water, oxygenated depositional environment of the Anvil Member to euxinic conditions marking the close of the iron-formation deposition in this area.

## Age and Correlation of the Marquette Range Supergroup

### Stratigraphic Correlations within the Marquette Range Supergroup

The various lithologic units that compose the Marquette Range Supergroup in the Gogebic iron range are part of a more geographically extensive rock sequence that encompasses the other iron ranges of northern Michigan and Wisconsin (Marquette, Menominee, Iron River-Crystal Falls iron ranges) and the intervening areas of non-iron-bearing strata. Modern stratigraphic correlation for these rocks dates from the work of James (1958) who introduced and defined the group names still in use (Chocolay, Menominee, and Baraga Groups for the Gogebic iron range). James (1958) also introduced the term "Animikie Series" for the entire Paleoproterozoic sequence, implying an approximate correlation with the Animikie Group of Minnesota. Cannon and Gair (1970)

introduced the name "Marquette Range Supergroup" in place of "Animikie Series" in compliance with the North American Stratigraphic Code (several groups constitute a supergroup, not a series, and "Animikie" was already established as a group name).

With regard to the Gogebic iron range, until recently, the only modification of the stratigraphic terminology defined by James (1958) was the formal designation of Emperor Volcanic Complex by Trent (1976). The Emperor, as therein defined, is a volcanic unit of formation rank and is part of the Menominee Group. Trent recognized its essential contemporaneous and intertonguing relation with the Ironwood Iron-Formation. Recently, an alternative correlation of some units in the Gogebic iron range was proposed by Morey (1996) and Ojakangas and others (2001). These authors proposed to reclassify the Palms Formation, Ironwood Iron-Formation, and Emperor Volcanic Complex as part of the Baraga Group rather than accepting their long-accepted placement in the Menominee Group. The correlation that they proposed accepted the Chocolay Group, as previously defined, to include the Sunday Quartzite and Bad River Dolomite. They proposed that the Menominee Group is completely absent in the Gogebic iron range. They thus correlated the Palms Formation with units such as the Goodrich Quartzite in the Marquette iron range and the Ironwood Iron-Formation with the Bijiki Iron-Formation Member of the Michigamme Formation in the western Marquette iron range.

We do not accept this new interpretation for several reasons. Although Morey (1996) and Ojakangas and others (2001) acknowledge that an unconformity separates the Copps Formation from the Emperor Volcanic Complex in the eastern Gogebic iron range, they also accept, without question, a conformable contact between the Tyler and the Ironwood farther west as proposed by some previous authors. As discussed previously, the nature of that contact is still not firmly established because exposures are no longer accessible; however, we believe that a preponderance of evidence favors an unconformable contact between the basal part of the Tyler (its Pabst Member) and the underlying Ironwood Iron-Formation, especially based on the evidence that the Pabst transects stratigraphic units in the Ironwood. Morey (1996) and Ojakan-
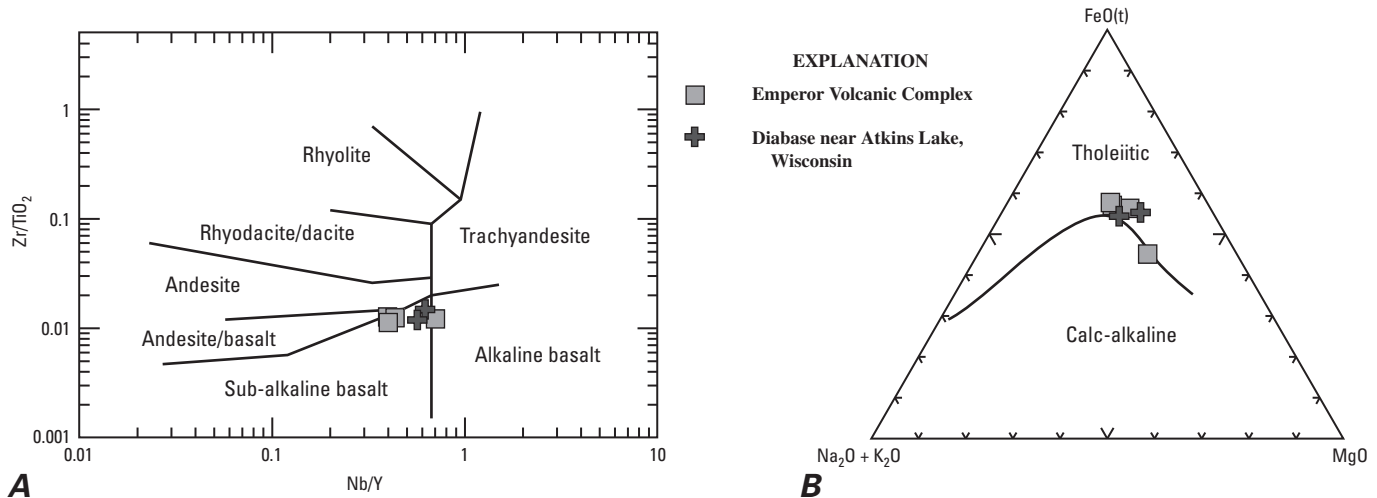
**Figure 7.** Geochemical discrimination diagrams showing the subalkaline tholeittic character of the mafic rocks from the Emperor Volcanic Complex (solid squares) and diabase near Atkins Lake. See table 1 for details of each sample. Boundaries in *A* from Winchester and Floyd (1977). Tholietic-calc-alkaline boundary in *B* after Irvine and Baragar (1971).

gas and others (2001) cite the presumed conformable contact as the principal evidence that the Palms, Ironwood, and Tyler units are a continuously deposited sequence and are therefore correlative with the Goodrich, Michigamme, and Bijiki succession of the Marquette iron range.

We believe that the clear evidence of an angular unconformity at the base of the Copps Formation, and at least the possibility of an unconformity at the base of the Tyler Formation both argue against the new correlation by Morey (1996) and Ojakangas and other (2001). In the western Marquette iron range, the base of the Baraga Group is represented by the Goodrich Quartzite, which includes a substantial thickness of conglomerate containing iron-formation detritus. This conglomerate is therefore equivalent to the basal conglomerate of the Copps Formation described above and to the Pabst Member of the Tyler Formation as well.

The Goodrich Quartzite grades upward into graphitic and pyritic black slate of the lower part of the Michigamme Formation. The overlying Bijiki Iron-Formation Member of the Michigamme Formation is a thin (100 m or less) carbonate and silicate iron-formation of local importance in the western part of the Marquette iron range and can be traced for about 15 km along strike (Cannon and Klasner, 1977, 1978). The Bijiki is evenly bedded and apparently was deposited in deep and (or) quiet water. The unit is overlain by additional pyritic and graphitic black slate of the Michigamme. The Bijiki is evidently a product of local iron-rich deposition in a stagnant basin and thus is quite distinct from the Ironwood Iron-Formation in which abundant shallow-water and well-oxygenated facies are present. The Bijiki is, therefore, much more akin to the sideritic shale units that overlie the Pabst Member in the lower part of the Tyler Formation. Drilling records indicate the sideritic shale of the lower Tyler is much more extensive than the Bijiki, but the sideritic shale was never distinguished

as a mappable unit because of a lack of outcrops. Also, there is clear evidence that strong tectonic and volcanic activity accompanied the deposition of the Ironwood in contrast to the quiet conditions indicated for the Bijiki. Although the true correlation may inevitably rely on future precise geochronological data, the evidence in hand does not warrant the substantial recorrelation proposed by Morey (1996) and Ojakangas and others (2001). Rather, the available evidence substantially supports the long-accepted correlations that we presented above.

### Absolute Age of the Marquette Range Supergroup

Precise, radiometrically determined ages for deposition of the Marquette Range Supergroup were not available until recent work by Schneider and others (2002). Until then, the supergroup was restricted to the approximate interval between 2.6 Ga (the age of youngest basement rocks) and 1.85 Ga (the approximate age of deformation and metamorphism). Schneider and others (2002) reported an age of 1,874±9 Ma based on sensitive high-resolution ion microprobe (SHRIMP) U-Pb dating of zircons from the Hemlock Formation, a unit about 50 km east of the eastern part of the Gogebic iron range. The Hemlock is a lateral equivalent of the Negaunee Iron-Formation and is also probably equivalent to the Emperor Volcanic Complex. A SHRIMP U-Pb zircon age of 1,878±2 Ma for a thin volcanic unit in the Gunflint Iron-Formation in Ontario was determined by Fralick and others (1998, 2002). The Gunflint is widely accepted as a correlative of the Ironwood Iron-Formation of the Gogebic iron range. These two ages, although determined for rocks somewhat removed from the Gogebic range, provide a good approximation for the age of deposition of the Menominee Group. Compressional deformation of the Marquette Range Supergroup probably began by at least 1,850 Ma and a post-tectonic pluton was dated at

1,835±9 Ma (Schneider and others, 2002). The age range of Baraga Group deposition is therefore constrained to the interval from about 1,875 to 1,850 Ma.

The age of the Chocolay Group remains more problematic. Chocolay rocks are preserved only sporadically within the southern Lake Superior region and lie beneath a regional unconformity marking the base of the Menominee Group. The Chocolay may, therefore, be substantially older than the Menominee Group. Numerous authors have suggested that the Chocolay is correlative with the upper parts of the Huronian Supergroup in Ontario, with which it shares substantial lithologic similarity, particularly the common occurrence of glaciogenic units (not present in the Gogebic iron range) and thick orthoquartzite sequences (Puffett, 1969; Young, 1970, 1973, 1983; Ojakangas, 1984, 1985, 1988; Roscoe and Card, 1993). If that correlation is correct, the Chocolay Group must be older than 2,200 Ma, which is the age of the Nipissing Diabase that cuts the Huronian Supergroup (Corfu and Andrews, 1986).

## Paleoproterozoic Intrusive Rocks

Paleoproterozoic dikes and sills of diabase and gabbro cut both the Marquette Range Supergroup and the Archean basement rocks. Sills intrusive into the Ironwood Iron-Formation are found on both the eastern and western ends of the Gogebic iron range. In the east, several large gabbroic sills (as much as 300 m thick) are concordant with the Ironwood and have been folded with it. Likewise, in the west, at least three sills have intruded the Ironwood. In the intervening central part of the range, including the productive mining district, sills are sparse but diabase dikes are very common and were probably intruded during the same magmatic event as the sills. Unpublished mine maps and sections show approximately 100 mapped dikes cutting the Ironwood Iron-Formation between Wakefield, Mich., and Iron Belt, Wis. The dikes cut the Ironwood bedding at approximately right angles and form both northeast- and northwest-trending sets. Similar dikes intruded the Neoarchean rocks south of the Gogebic iron range and were mapped in detail where outcrops are adequate (Trent, 1973; Prinz and Hubbard, 1975; Klasner and others, 1998). These dikes occur in two sets that intersect at a nearly right angle; one set trends east-northeast and the other northnorthwest, and all generally dip moderately to the south. The attitude of the dikes has been affected by the northward tilting of the region during events related to the Midcontinent Rift at about 1.1 Ga (Cannon and others, 1993); before the rifting event, the dikes were essentially vertical.

Paleoproterozoic dikes are very abundant in rocks of the Menominee Group and Neoarchean basement, but they are apparently absent in the Tyler and Copps Formations. Areas underlain by the Tyler and Copps have only scattered outcrops and there is a possibility that dikes are present (but concealed). However, some of those areas have outcrop density comparable to areas of Neoarchean rocks where dikes can be readily mapped. Diabase dikes are typically more resistant to weathering than the graywacke and shale units of the Tyler and Copps and would likely form ridges if present. For example, topographic ridges are commonly formed by Mesoproterozoic dikes where they have cut the Tyler Formation such as sec. 20, T. 45 N., R. 1 W. (pl. 1A). Thus, the mafic intrusive activity most likely occurred shortly after deposition of the Menominee Group or possibly concurrent with it, and the igneous rocks, like their sedimentary host rocks, probably are truncated by the unconformity between the top of the Menominee Group and the base of the Baraga Group. Because the sills are approximately coextensive with the Emperor Volcanic Complex in the east and are also located near volcanic units in the west, the intrusive and volcanic rocks are probably comagmatic.

Four samples of mafic volcanic rocks from the Emperor Volcanic Complex in the eastern Gogebic iron range were analyzed for major and trace elements. In addition, two samples from a mafic sill that intruded the Ironwood Iron-Formation near Atkins Lake, Wis., at the western end of the Gogebic iron range were analyzed. These geochemical data are presented in table 1.

All samples are subalkaline tholeiitic basalts (fig. 7). Sample EMP 7 is less evolved than the other samples as it has a higher Mg number (58.7), lower FeOt and $TiO_2$ concentrations, and higher Cr and Ni contents (table 1). The other samples have very similar evolved compositions: low Mg numbers (48.6–40.2), high FeOt (11.76–13.72 weight percent), and $TiO_2$ (1.31–1.87 weight percent). As a group, the samples define a trend of increasing FeOt, $TiO_2$, and incompatible trace elements (for example, Zr, Nb, Y, La) and decreasing Mg numbers typical of tholeiitic basalt suites. Decreasing Cr and Ni and decreasing Mg numbers are consistent with fractionation of olivine and clinopyroxene.

Chondrite-normalized rare-earth-element (REE) plots indicate that all samples have been enriched in light REE, having abundances from 30 to 75 times chondritic abundances and heavy REE from 9 to 15 times chondritic abundances (fig. 8A). None of the samples has a europium anomaly, which suggests limited plagioclase fractionation. Chondrite-normalized La/Yb ratios range from 3.3 to 6.6, generally increasing with decreasing Mg number (table 1). Compared to the composition of primitive mantle, these basalts have variably negative Nb, Ta, Ti, and a positive Th anomaly (fig. 8B). The compositional similarity between the samples of Emperor basalt and samples of diabase from the sill near Atkins Lake supports the correlation of these mafic igneous rocks.

The variation in the La/Yb ratio shown by the basalt samples is greater than would result from simple crystal fractionation (Ueng and others, 1988). The variation suggests the possible involvement of more complex magmatic processes such as crustal assimilation and fractional crystallization. The role of that process in the evolution of the Emperor basalts is supported by (1) the general correlation of the increasing La/Yb ratio with the decreasing Mg number, and (2) the variably negative Nb, Ta, Ti, and positive Th anomalies on the primitive-mantle-normalized plot (fig. 8B) (Ueng and others, 1988).

**Table 1.**   Whole-rock chemical analyses of Paleoproterozoic mafic rocks from the Emperor Volcanic Complex and diabase near Atkins Lake.

[All measurements in weight percent unless otherwise noted. Major elements recalculated to 100-percent volatile free. Samples 1–4 from Emperor Volcanic Complex; samples 5 and 6 from diabase near Atkins Lake. Fe0t is total iron as Fe0. LOI, loss on ignition at 900°C. NA, not analyzed]

| Sample number[1] | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| Field number | EMP-3-84 | EMP-6-84 | EMP-7-84 | MI-99-5 | 98STO-309 | 98STO-310 |
| Major element oxides, in weight percent | | | | | | |
| $SiO_2$ | 50.7 | 51.63 | 51.26 | 51.72 | 51.14 | 49.54 |
| $Al_2O_3$ | 15.49 | 14.74 | 15.81 | 15.72 | 15.26 | 14.85 |
| FeOt | 12.91 | 13.72 | 10.42 | 11.76 | 13.17 | 13.27 |
| MgO | 5.11 | 5.18 | 8.31 | 5.55 | 6.13 | 7.05 |
| CaO | 8.59 | 7.56 | 8.49 | 10.00 | 6.85 | 9.26 |
| $Na_2O$ | 3.15 | 3.71 | 3.71 | 2.42 | 3.48 | 2.75 |
| $K_2O$ | 1.48 | 1.15 | 0.65 | 1.16 | 1.42 | 0.92 |
| $TiO_2$ | 2.04 | 1.83 | 1.06 | 1.31 | 1.99 | 1.87 |
| $P_2O_5$ | 0.22 | 0.22 | 0.18 | 0.17 | 0.28 | 0.24 |
| MnO | 0.31 | 0.25 | 0.11 | 0.20 | 0.28 | 0.26 |
| Total | 100 | 100 | 100 | 100 | 100 | 100 |
| Volatiles, in weight percent | | | | | | |
| $LOI/H_2O+$ | 1.79 | 1.84 | 2.56 | 1.82 | 1.7 | 1.6 |
| $H_2O-$ | NA | NA | NA | NA | 0.1 | 0.1 |
| S | NA | NA | NA | NA | 0.03 | 0.04 |
| Trace elements (ppm) | | | | | | |
| Sc | 36.1 | 40.1 | 37.3 | 38 | 44 | 40 |
| V | NA | NA | NA | 278 | 279 | 217 |
| Co | 44 | 48 | 48 | 46 | 48 | 54 |
| Cr | 58 | 49 | 364 | 81 | 174 | 164 |
| Ni | 55 | 58 | 118 | 122 | 68 | 118 |
| Rb | 13.5 | 26.7 | 15 | 31 | 46.2 | 31.7 |
| Sr | 357 | 177 | 310 | 246 | 240 | 330 |
| Cs | NA | 0.25 | 0.33 | 0.30 | 1.42 | 1.74 |
| Ba | 512 | 546 | 200 | 398 | 796 | 338 |
| Y | 27 | 30 | 15 | 24 | 29 | 26.5 |
| Zr | 149 | 137 | 72 | 100 | 177 | 133 |
| Nb | 19 | 13 | 6 | 9.5 | 18 | 15 |
| Hf | 3.97 | 3.78 | 1.84 | 2.85 | 3.51 | 3.07 |
| Ta | 1.59 | 1.12 | 0.435 | 1 | 1.01 | 0.939 |
| Th | 4.07 | 5.34 | 1.5 | 15 | 2.72 | 2.35 |
| U | 0.83 | 1.04 | 0.25 | 1.77 | 0.39 | 0.45 |
| Zn | 96 | 90 | 82 | 69 | 112 | 106 |
| Cu | 128 | 122 | 113 | 125 | 192 | 185 |
| La | 24.6 | 23.8 | 9.75 | 24.00 | 19.6 | 16.85 |
| Ce | 48 | 45.5 | 18.6 | 45.25 | 38.9 | 34.25 |

**Table 1.**    Whole-rock chemical analyses of Paleoproterozoic mafic rocks from the Emperor Volcanic Complex and diabase near Atkins Lake.—Continued

[All measurements in weight percent unless otherwise noted. Major elements recalculated to 100-percent volatile free. Samples 1–4 from Emperor Volcanic Complex; samples 5 and 6 from diabase near Atkins Lake. Fe0t is total iron as Fe0. LOI, loss on ignition at 900°C. NA, not analyzed]

| Sample number[1] | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| Field number | EMP-3-84 | EMP-6-84 | EMP-7-84 | MI-99-5 | 98STO-309 | 98STO-310 |
| Trace elements (ppm)—Continued | | | | | | |
| Pr | NA | NA | NA | 5.01 | NA | NA |
| Nd | 25 | 21.5 | 9 | 20.65 | 20 | 18.4 |
| Sm | 6.13 | 5.65 | 2.77 | 4.91 | 5.53 | 4.97 |
| Eu | 1.84 | 1.79 | 0.94 | 1.65 | 1.62 | 1.51 |
| Gd | NA | NA | NA | 4.81 | 5.37 | 4.87 |
| Tb | 1.011 | 0.975 | 0.44 | 0.860 | 0.899 | 0.785 |
| Dy | NA | NA | NA | 4.89 | NA | NA |
| Ho | NA | NA | NA | 0.99 | NA | NA |
| Er | NA | NA | NA | 2.86 | NA | NA |
| Tm | NA | NA | NA | 0.401 | 0.525 | 0.497 |
| Yb | 3.26 | 3.22 | 1.97 | 2.4 | 3.45 | 3.04 |
| Lu | 0.457 | 0.493 | 0.273 | 0.4 | 0.494 | 0.447 |
| Element ratios | | | | | | |
| [La/Yb]cn[2] | 5.0 | 4.9 | 3.3 | 6.6 | 3.8 | 3.7 |
| La/Ta | 15.5 | 21.3 | 22.4 | 31.4 | 19.4 | 17.9 |
| Th/Ta | 2.6 | 4.8 | 3.4 | 20.0 | 2.7 | 2.5 |
| Zr/Y | 5.5 | 4.6 | 4.8 | 4.2 | 6.1 | 5.0 |
| Zr/Nb | 7.8 | 10.5 | 12.0 | 10.5 | 9.8 | 8.9 |
| Nb[3] | -31.7 | -62.7 | -15.6 | -188.0 | -10.8 | -10.8 |
| Ta[4] | -6.5 | -43.9 | -11.0 | -165.2 | -11.8 | -6.7 |
| F/F+M[5] | 0.72 | 0.73 | 0.56 | 0.68 | 0.68 | 0.65 |
| Mg#[6] | 41.4 | 40.2 | 58.7 | 45.7 | 45.3 | 48.6 |

[1]Samples 1, 2, and 3 were analyzed by the U.S. Geological Survey (USGS) using wavelength-dispersive x-ray flrouescence spectroscopy for all ten major oxides: J. Tagart, A. Bartel, and D. Siems analysts: energy dispersive x-ray fluorescence spectrometry for Rb, Sr, Ba, Zr, Y, and Cu: J. Jackson, analyst: and instrumental neutron activation analysis (INAA) for all other elements including the rare-earth elements: J. Mee, analyst. Sample 4 was analyzed by XRAL Laboratories, Don Mills, Ontario using inductively coupled plasma-atomic emission spectroscopy (ICP-MS) for all ten major elements and V, Y, Cu, and Nb: all other elements, including the rare-earth elements, were analyzed by the USGS using INAA: C. Palmer, analyst. Samples 5 and 6 were analyzed by Actlabs Laboratories, Ltd., Ancaster, Ontario, using ICP-MS and a lithium metaborate/tetraborate fusion method.

1. EMP-3-84        Massive basalt    SW¼SE¼ sec. 20, T.47N., R.43W.
2. EMP-6-84        Diabase          NW¼NE¼ sec 26, T.47N., R.44W.
3. EMP-7-84        Pillow basalt    NW¼SE¼ sec. 23, T.47N., R.44W.
4. MI-95-9         Gabbro           NE¼NW¼ sec. 26, T.47N., R.44W
5. 98-STO-309      Diabase          SW¼NE¼ sec. 20, T.44N., R.5W.
6. 98_STO-310      Diabase          SW¼NE¼ sec. 20, T.44N., R.5W.

[2] [La/Yb]cn = La/Yb chondrite-normalized ratio.

[3] Nb = Nb/Chon – (Th/Chon + La/Chon)/2.

[4] Ta = Ta/Chon – (Th/Chom + La/Chon)/2

[5] F/F+M = FeOt/FeOt + MgO.

[6] Mg# = Mg/Mg + total Fe.

Continental crust is characterized by prominent negative Nb, Ta, Ti, and positive Th anomalies (Taylor and McLennan, 1985); therefore, the assimilation of continental crust during the fractional crystallization of the basalt may account for the variable inheritance of the crustal compositional anomalies (De Paolo, 1981). Ueng and others (1988) showed that crustal assimilation and fractional crystallization processes were important in the evolution of other Paleoproterozoic basalts within the Marquette Range Supergroup in northern Michigan, including the Hemlock Formation and Badwater Greenstone. Crustal assimilation for the Hemlock basalts is further supported by Nd isotope data (Beck and Murthy, 1991).
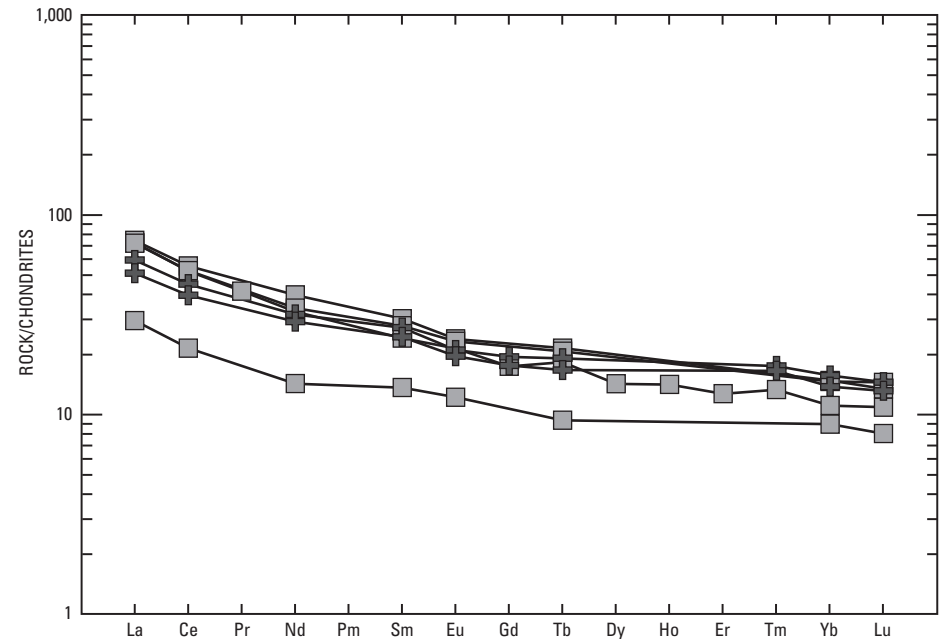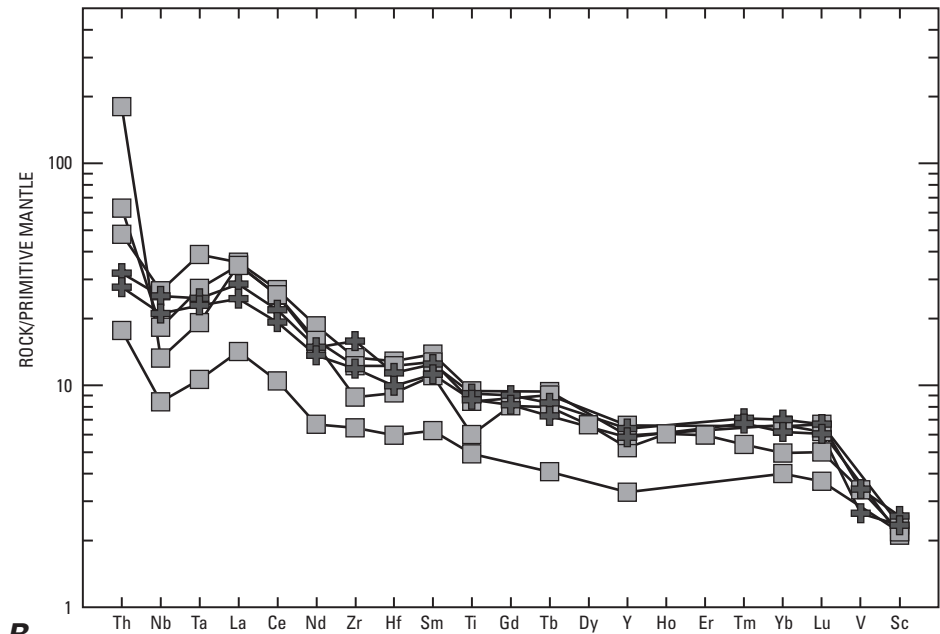


**Figure 8.** Chondrite-normalized rare earth element plot (*A*) and primitive-mantle-normalized multi-element plot (*B*) for mafic rocks from the Emperor Volcanic Complex and diabase near Atkins Lake. Symbols as in figure 7. Chondrite normalizing values from Nakamura (1974); primitive-mantle normalizing values from Kerrich and Wyman (1997).

## Mesoproterozoic Rocks

The Paleoproterozoic strata of the Gogebic iron range are overlain by a great thickness of subaerial volcanic and sedimentary rocks of the Keweenawan Supergroup, which was deposited at about 1.1 Ga in the Midcontinent Rift. This report will not discuss these rocks in detail, but will highlight some features that are pertinent to the geology of the Gogebic iron range. For detailed information on the geology of the Keweenawan rocks in the area, see recent maps by Cannon and others (1995) and Cannon and others (1996).

The basal unit of the Keweenawan Supergroup in the western part of the Gogebic iron range consists of quartzite and conglomerate of the Bessemer Quartzite (pl. 1*A*). The Bessemer pinches out eastward and is absent east of Ramsay, Mich., where basalt flows of the Siemens Creek Volcanics overlie rocks of the Marquette Range Supergroup. The unconformity between the Marquette Range Supergroup and Keweenawan Supergroup marks a hiatus of more than 700 million years (m.y.). There is only slight angular discordance between the two supergroups, however, especially in the central part of the iron range. The discordance indicates that (1) little Penokean deformation occurred in the central part of the iron range and (2) the present steep northward dips of rocks of the Marquette Range Supergroup resulted from Mesoproterozoic deformation rather than from Penokean deformation. In the far western part of the iron range, Mesoproterozoic rocks were thrust southward along the Atkins Lake-Marenisco fault over the Marquette Range Supergroup rocks so that west of sec. 14, T. 44 N., R. 5 W., the contact appears to be entirely a fault (pl. 1*A,B*).

Mesoproterozoic intrusive rocks that were intruded at approximately 1.1 Ga also have some importance in studies of the Paleoproterozoic strata. Gabbro of the Mineral Lake intrusion truncates the Paleoproterozoic strata in sec. 14, T. 44 N., R. 4 W. (pl. 1*A*). Paleoproterozoic rocks are absent for about 9 km to the west but then appear again in sec. 13, T. 44 N., R. 5 W. and continue for about 20 km farther west before being truncated by the Atkins Lake-Marenisco fault. The most pervasive effect of the Mesoproterozoic intrusions was the formation of a broad contact-metamorphic aureole that altered the character of the Ironwood Iron-Formation from about the Tyler Forks River to the west. Metamorphism of the Ironwood becomes increasingly intense westward. In the Mount Whittlesey area, iron-rich amphiboles are common in more reduced parts of the Ironwood. Near the contact with the Mineral Lake intrusion southeast of Mineral Lake, the Ironwood contains iron-rich orthopyroxenes and local fayalite. The increasing magnetite content of the Ironwood west of Tyler Forks River is also an expression of increasing metamorphic grade. Rocks in the western part of the Gogebic iron range west of the Mineral Lake intrusion, have also been substantially metamorphosed. The Ironwood is typically an iron-amphibole-garnet-magnetite rock in that area.

# Structural Geology

The principal purpose of this section is to describe the development of structures in the Paleoproterozoic rocks, which vary from large-scale folds and faults to fabric elements observed in outcrop. We have divided the Gogebic iron range into eastern, central, and western sections, as defined above; in this part of the report, we describe the structural development of each section. We also compare and contrast the structural history of the three sections in order to establish the changes that occurred from the very weakly deformed Penokean foreland, characterized by the central part of the range, to the more intensely deformed eastern and western sections of the range. The structural history of the region began with the intense Neoarchean deformation of the basement rocks south of the Gogebic iron range at roughly 2.7 Ga, and ended with the large-scale northward tilting of the entire region in the closing stages of the Midcontinent rifting at about 1.1 Ga. These phases of deformation are discussed only briefly to emphasize the aspects that are pertinent to understanding the geology of Paleoproterozoic rocks and structures.

A summary of the structural history of the region is given in table 2. Alpha-numeric designations used both in the text and table 2 are as follows: (1) D designates deformational events with the numeric subscripts indicating the order of formation ($D_1$, $D_2$, and so on); (2) S designates planar structures such as bedding ($S_0$), first-generation foliation ($S_1$), second-generation foliation ($S_2$), and so on; (3) L designates linear structures, such as first-phase fold axis (or $S_0/S_1$ intersection lineation) ($L_1$), second-phase fold axis ($L_2$), and stretch lineation ($L_S$); (4) F designates folds in order of their formation ($F_1$, $F_2$, and so on); $L_A$ and $S_A$ indicate Archean lineations and foliations, respectively; and $L_P$ and $S_P$ indicate Proterozoic lineations and foliations, respectively.

## Mesoproterozoic Structures

Mesoproterozoic structures are discussed first in this report because their geometric effects on the Paleoproterozoic rocks must be removed in order to decipher the geometry of the Penokean deformation. Mesoproterozoic strata of the Keweenawan Supergroup north of the Gogebic iron range dip north at angles ranging from as low as 30º to nearly vertical. Although these strata unconformably overlie strata of the Marquette Range Supergroup, they are nearly concordant structurally, the contact being generally a low-angle unconformity. It has been clear since at least the work of Schmidt and Hubbard (1972) that the present northward dips of the Marquette Range Supergroup strata result from the same Mesoproterozoic monoclinal tilting that affected the Keweenawan strata. Schmidt and Hubbard (1972) further proposed that, at the time of deposition of the Keweenawan rocks, the Marquette Range Supergroup strata dipped gently south because they now gen-

**Table 2.**    Sequence of tectonic events in the Gogebic iron range.

[D, deformational events with the numeric subscripts indicating the order of formation ($D_1$, $D_2$, and so on); S, planar structures such as bedding ($S_0$), first-generation foliation ($S_1$), second-generation foliation ($S_2$), and so on; L, linear structures, such as first-phase fold axis (or $S_0/S_1$ intersection lineation) ($L_1$), second-phase fold axis ($L_2$) and stretch lineation; F, folds in order of their formation ($F_1$, $F_2$, and so on); $L_A$ and $S_A$, Archean lineations and foliations, respectively; $L_P$ and $S_P$, Proterozoic lineations and foliations, respectively]

| Event | Description | Resulting Feature |
| --- | --- | --- |
| Mesoproterozoic | | |
| Block rotation ($D_4$) | Northward rotation of rigid blocks in upper plate of listric thrusts | North-facing monocline, small-offset cross faults. |
| Rifting and volcanism | Formation of continental rift basins | Midcontinent Rift and thick volcanic and sedimentary fill (Keweenawan Supergroup). |
| Uplift | Uplift and peneplanation of Paleoproterozoic strata of Penokean fold-and-thrust belt | Unconformity at top of Marquette Range Supergroup. |
| Paleoproterozoic | | |
| Block faulting ($D_3$) | Reactivation of Presque Isle fault and Little Presque Isle fault and related faults. Formation of Sunday Lake fault and related faults | Sets of steep faults cutting Midcontinent Rift system, spaced shear zones in Archean rocks. |
| Compressional deformation | | |
| $D_2$ | Continued thin-skinned coaxial deformation of Paleoproterozoic strata, including folding of $D_1$ structures. Folding of $D_1$ thrust faults into east-plunging anticline | $L_2$ folds, $S_2$ foliation, Wolf Mountain anticline. |
| $D_1$ | Detachment of Paleoproterozoic strata from Archean basement along north-verging thrusts, folding of bedding | $L_1$ fold axes, $S_1$ foliation, detachment and thrust faults. |
| Flexural subsidence | Deposition of turbidites in advance of northward-migrating fold-and-thrust belt | Thick sequence of turbidites of Tyler Formation. |
| Passage of foreland bulge | Uplift and erosion of Menominee Group rocks | Unconformity at top of Menominee Group. |
| Extensional deformation | Formation of ancestral Presque Isle trough, deposition of iron-formation and volcanic rocks in subsiding trough | Presque Isle trough and thick sequence of Ironwood Iron-Formation and Emperor Volcanic Complex. |
| Flexural subsidence | Formation of shallow-marine stable platform, deposition of argillite and quartzite | Transgressive argillite to quartzite sequence of Palms Formation. |
| Uplift and erosion | Deep weathering and erosion of Chocolay Group and Archean basement rocks | Unconformity at base of Menominee Group. |
| Subsidence of platform | Shallow-marine sedimentation on platform | Quartzites and carbonates of Chocolay Group. |
| Uplift and erosion | Peneplanation of Archean fold belt | Profound unconformity at base of Paleoproterozoic. |
| Neoarchean | | |
| Regional deformation | Regional-scale nappes, major faults | Thick sequences of partly overturned volcanic rocks. |
| Island-arc formation | Eruption of volcanic sequences and emplacement of granitic plutons | Volcanic rocks of Ramsay Formation and emplacement of Puritan batholith. |

erally dip more gently north than do the Keweenawan rocks. Restoring the basal Keweenawan strata to the horizontal by rotation to the south rotates the Marquette Range Supergroup strata to a south dip. Figure 9, modified slightly from Schmidt and Hubbard (1972), shows these relations.

The term "Montreal River monocline" was introduced by Cannon and others (1993) for the remarkably thick succession of steeply north-dipping rocks that crops out from the Lake Superior shoreline near the mouth of the Montreal River at the Wisconsin-Michigan border, southward for nearly 40 km to the Atkins Lake-Marenisco fault. The monocline has exposed a section of rocks at least 30 km thick containing no apparent structural repetition. The exposed strata include all of the Keweenawan Supergroup units, all of the Marquette Range Supergroup units in the Gogebic iron range, and about 20 km of Neoarchean basement rocks. Cannon and others (1993) used structural data based on the present orientation of originally vertical diabase dikes and geochronologic evidence to date and map the extent of uplift accompanying the rotation. They proposed that the Montreal River monocline, including all of the Gogebic iron range, is the rotated upper plate of a listric thrust fault of near-crustal dimensions whose surface trace is the Atkins Lake-Marenisco fault. The fault formed at about 1.06 Ga as indicated by Rb-Sr cooling ages of biotite from rocks in the rotated and uplifted upper plate (Cannon and others, 1993), and is roughly coincident with other major thrust faults responsible for partially inverting the deep central graben of the Midcontinent Rift.

Penetrative deformation apparently did not occur during this Mesoproterozoic thrusting and rotation. Deformation appears to have been limited solely to the rotation of rigid blocks. Faults that have small offsets nearly perpendicular to the bedding of the Marquette Range Supergroup rocks were mapped along much of the Gogebic iron range (for instance, Cannon and others, 1995, 1996). Many of these faults can be projected northward to align with faults in the lower units of the Keweenawan Supergroup. The offset of units ranges from negligible to a few hundred meters. The faults are both left lateral and right lateral. These faults most likely formed during the major Mesoproterozoic thrust faulting and rotation, possibly to accommodate changes in the shape of the rigid upper plate of the listric thrust. Also, the contact between the Mesoproterozoic and Paleoproterozoic strata has been offset by these faults (pl. 1B), which indicates that the Little Presque Isle and parallel faults in the Presque Isle trough were reactivated at this time.

## Atkins Lake-Marenisco Fault

The Atkins Lake-Marenisco fault is one of two major faults identified by Cannon and others (1993) as Mesoproterozoic thrust faults, the other being the Pelton Creek fault south of the study area. The fault, as defined here, is a composite of two long-recognized faults in different parts of the area, which, until recent field work, had not been correlated because much of their extent is in areas of very limited bedrock expo-

sure where their character and location were poorly known. The Marenisco fault was named by Sims (1992) in northern Michigan near the eastern end of the Gogebic iron range. The Marenisco fault juxtaposed Neoarchean granitic rocks on the north with Paleoproterozoic metasedimentary rocks of the Marquette Range Supergroup on the south, and was interpreted originally to be a north-verging Penokean thrust fault. There are no exposures of bedrock in or near the fault trace (pl. 1); its location and nature were interpreted largely from aeromagnetic patterns.

A fault contact between Mesoproterozoic volcanic and intrusive rocks on the north and Paleoproterozoic and Neoarchean rocks on the south had been known for many years. This fault was named the "Crystal Lake-Atkins Lake fault" by Aldrich (1929), who recognized that it was the structural equivalent and possibly the direct westward extension of the Keweenaw fault in northern Michigan. The name is simplified here to Atkins Lake fault because the only exposures of the fault are north and east of Atkins Lake in northern Wisconsin (pl. 1A,B). Aldrich (1929) correctly surmised that it is a major thrust fault along which Mesoproterozoic volcanic and intrusive rocks were thrust southward over Paleoproterozoic strata of the Marquette Range Supergroup. Mapping for this study shows that the Atkins Lake fault is a westward extension of the Marenisco fault, although it is at a different stratigraphic position than the Marenisco fault to the east.

The Atkins Lake-Marenisco fault zone is well exposed in the gorge of the Marengo River in secs. 14 and 15, T. 44 N., R. 5 W. (pl. 1A). The fault zone is best exposed from the prominent waterfalls in SW¼ sec. 14 for about 500 m along the river to the west. In this area, bedrock bluffs along the south side of the river consist of Neoarchean granitic rocks and the Paleoproterozoic Bad River Dolomite. There is no trace of fault-related deformation detectable in the outcrops.

Along the north side of the river, exposures of a distinctive fault rock, apparently a finely crushed Neoarchean granite, extend from the river bank for several tens of meters to the north. This rock appears massive in outcrop, is dark gray on fresh surfaces and weathers to a pinkish-gray color. Thin sections show that it is composed of a mixture of very angular fragments of quartz and plagioclase in a finely comminuted matrix of quartz, plagioclase, chlorite and sericite (fig. 10). Chlorite is ubiquitous and has a prominent preferred orientation; however, because chlorite composes only a small percentage of the rock, it is not abundant enough to impart a visible fabric except by microscopic examination (fig. 10). The rock is overlain on the north by strongly metamorphosed basalt of the Mesoproterozoic Siemens Creek Volcanics. Additional exposures of the rock occur sporadically along the north side of the river valley for about 1 km farther west to near the center of sec. 15.

Strongly sheared Archean granite in fault contact with the Ironwood Iron-Formation can also be seen sporadically in outcrops for a few hundred meters east of the waterfalls. Small outcrops of highly sheared metabasalt of the Siemens Creek Volcanics may be seen in NW¼ sec. 20, T. 44 N.,

R. 5 W. (plate 1*A*), immediately north of Atkins Lake, and are interpreted to mark the Atkins Lake fault. These are the only exposures of the Atkins Lake-Marenisco fault along its entire surface trace of about 150 km.

The present interpretation of the Atkins Lake-Marenisco fault is that it is a Mesoproterozoic thrust fault, as proposed by Cannon and others (1993), but its geometry in the western part of the Gogebic range is substantially different than shown in that report. Recent mapping for the present study suggests that the thrust fault, which was deep within Neoarchean rocks in the central part of the area, passes upward along a lateral ramp to truncate the Mineral Lake intrusion (part of the Mellen Intrusive Complex) on the west and eventually cuts upward through the Paleoproterozoic section of the Marquette Range Supergroup (pl. 1*A,B*). In the western part of the area, this fault bounds the south side of Mesoproterozoic rocks and has emplaced them southward over Paleoproterozoic rocks.

## Paleoproterozoic Structures

Structures of Paleoproterozoic age include extensional features that formed at the same time as the Marquette Range Supergroup sedimentation, and contractional features that formed during the ensuing Penokean orogeny. The following sections present (1) details of the Paleoproterozoic structural evolution, as observed in some key areas of the Gogebic iron range, and (2) a study of the contrast between the central portion of the range (described first), where Penokean deformation was minimal, and the more intensely deformed eastern and western parts of the range. For each area, the present orientation of structural elements is described followed by an interpretation of the approximate geometry of the structures prior to Mesoproterozoic rotation which was derived by restoring the local basal Mesoproterozoic strata to a horizontal orientation. For the western Gogebic iron range, mostly new
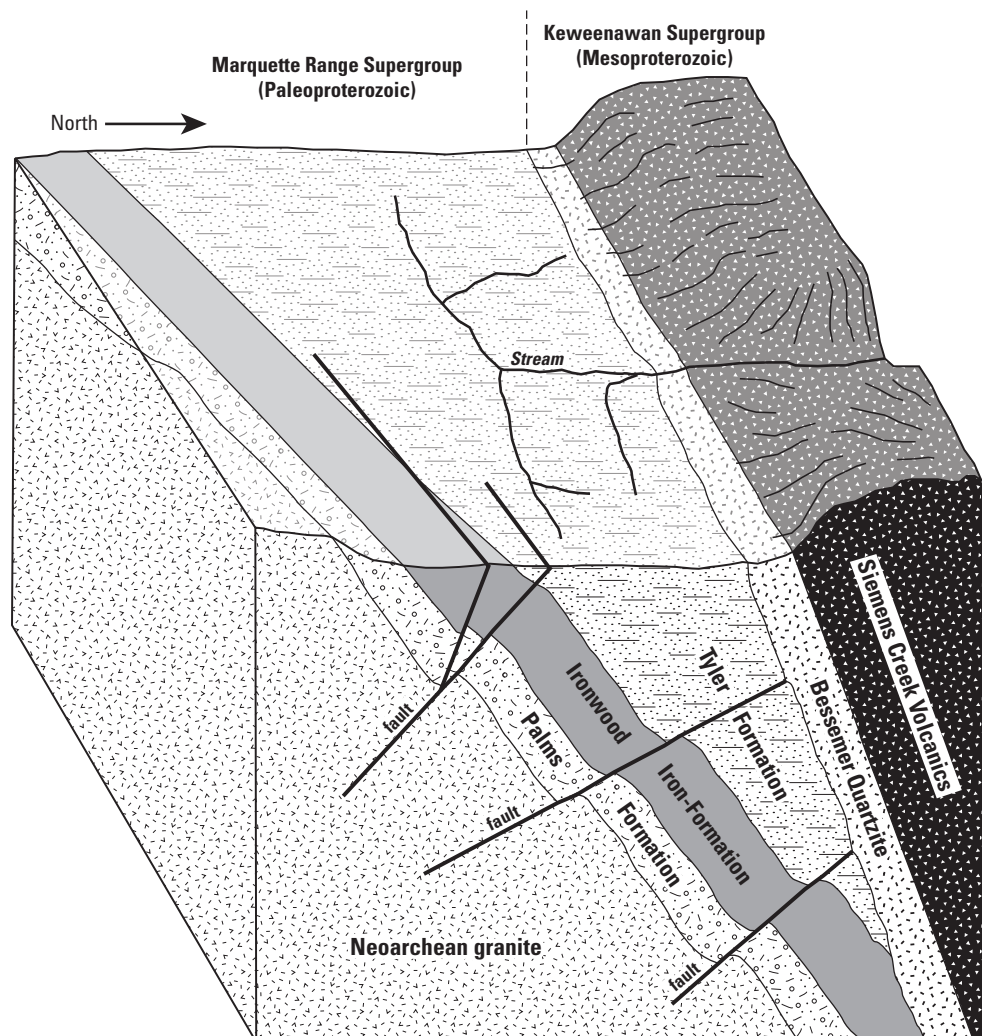


**Figure 9.**  Block diagram illustrating the geometric relationship between Neoarchean, Paleoproterozoic and Mesoproterozoic rocks in the central part of the Gogebic iron range. Simplified from Schmidt and Hubbard (1972).

observations made during the current investigation are presented; these descriptions also include details of stratigraphy and lithology.

## Central Gogebic Iron Range

The central part of the Gogebic iron range extends from the Sunday Lake fault near Wakefield, Mich. to the Tyler Forks River, west of Upson, Wis. (pl. 1*B*). This part of the iron range accounted for a great majority of the previous iron-ore production of the range and was mined for nearly its entire 40-km strike length. Studies conducted during the height of mining activity, such as Aldrich (1929), Hotchkiss (1919),



**Figure 10.** Photomicrographs of rocks from the Atkins Lake-Marenisco fault zone on the north side of the Marengo River, SW¼, Sec. 14, T. 44 N., R. 5 W. *A*, Angular to subrounded grains of quartz and plagioclase in very fine-grained matrix of chlorite, sericite, quartz, and plagioclase. Plane polarized light. *B*, Material similar to *A* under crossed nichols.

and Allen and Barrett (1915), provided many details of the geology of the central part of the district. Hotchkiss (1919) based his studies largely on observations in the extensive underground mine workings and, therefore, provided the best documentation of the structures of the central part of the range where outcrops are rare. Much of the following is based on descriptions found in Hotchkiss (1919), along with more recent observations, only a few of which were the result of the present study.

The central Gogebic iron range, commonly referred to in older reports as the "central monocline," consists of a steeply northward-dipping succession of Paleoproterozoic strata that are virtually devoid of internal folds and only rarely exhibit a penetrative deformational fabric. Paleoproterozoic structures are principally faults of four types, as defined by Hotchkiss (1919). Their definitions are paraphrased as follows:

1. Transverse faults—Faults striking nearly perpendicular to a rock formation and nearly vertical in dip.

2. "Eureka"-type faults—Faults striking nearly parallel to the strike of a rock formation, or parallel to eastward-pitching dikes, and nearly perpendicular to the beds.

3. Reverse faults—The Sunday Lake fault is the only one of its type.

4. Bedding faults—Faults parallel to the beds of the Iron-wood Iron-Formation.

Three of these types cannot be recognized from outcrop information because of the relatively small displacements and very sparse bedrock exposures, but they were well documented in the mines and were important in localizing iron ores.

Figure 11, simplified from figure 21 of Hotchkiss (1919), shows the geometric relationships of fault types 1, 2, and 4, above. Because faults of similar small offsets and of the same trend as type 1 faults cut the lower units of the Keweenawan Supergroup along the north edge of the Gogebic iron range, it is possible that type 1 faults are Mesoproterozoic structures and were not formed during the Penokean deformation. They do, however, clearly offset bedding faults, which we interpret as Penokean compressional features; therefore, if they are Penokean structures, then they formed late in the deformational history. Type 2 faults offset bedding faults and are in turn offset by transverse faults; therefore, they are intermediate in age between faults of definite Penokean age and faults of possible Mesoproterozoic age. The tectonic setting of the type 2 faults is not clearly established. When Mesoproterozoic tilting is restored by removing the present dip of Mesoproterozoic strata, type 2 faults are nearly vertical.

Bedding faults (type 4), which have displacements of at least 300 m, were well known in mine workings and were best described by Hotchkiss (1919). Bedding faults were identified in at least four stratigraphic horizons in the Ironwood Iron-Formation, but the principal fault of this type is confined to the Yale Member of the Ironwood and was identified over the entire strike length of the central Gogebic iron range. In
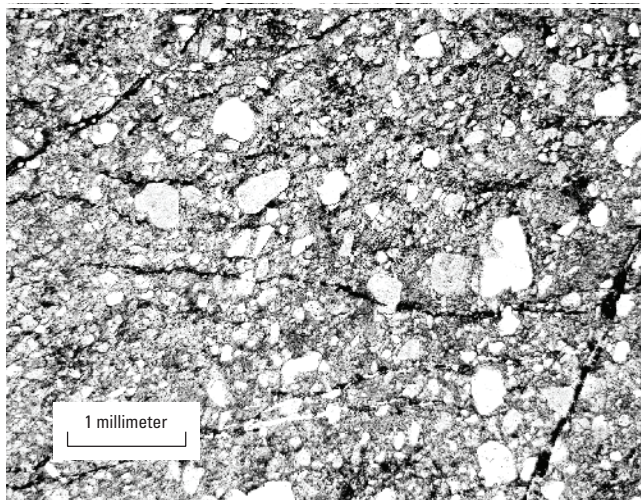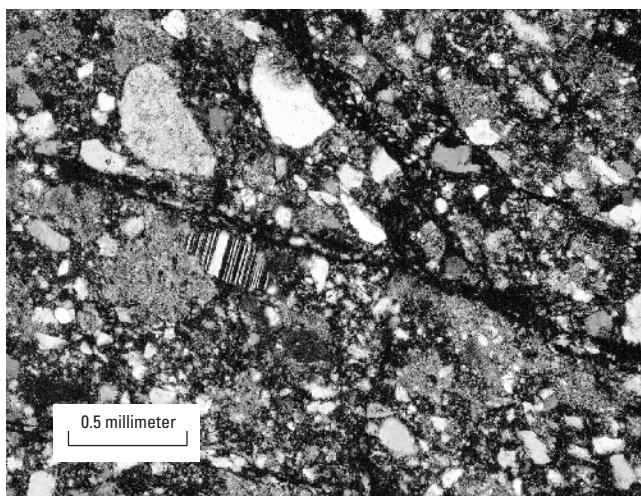
places, this fault constitutes a single surface, but more commonly the displacement was distributed between several surfaces, all within the Yale Member. The faults have offset the bedding-normal diabase dikes, which were used as markers to determine displacement. Some displacements were summarized for various mines by Hotchkiss (1919). Offsets everywhere have a sense of movement on north side of the fault to the east. Because the offsets are determined for planar features, there is no unique solution for total displacement. Total displacement may be much larger than the values stated by Hotchkiss (1919). Unfortunately, there are no recorded observations of kinematic indicators in the fault zones. When Mesoproterozoic tilting is restored, these bedding faults dip gently south and were apparently low-angle thrust faults. If these faults are Penokean compressional features, then they are most likely north-directed thrust faults.

The total number of bedding faults in parts of the stratigraphic section that are not exposed in mine workings cannot be determined. The relatively small, bedding-parallel displacements, coupled with the very sparse outcrops in the region, make these faults nearly impossible to detect except in man-made exposures. They may be common and could account, in total, for a considerable amount of Penokean tectonic transport.

Some evidence indicates contraction and tectonic shortening within the Tyler Formation, although megascopic folds are apparently absent. Several outcrops of thin-bedded shale and sandstone of the Tyler Formation are located in NE¼ sec. 28 and NW¼ sec. 29, T. 46. N., R. 2 E. near Hurley, Wis. Bedding ($S_0$) in these outcrops is oriented N. 61° E., 80° N. (stereoplot O on pl. 1*B*). Crossbeds in the sandstone show that the steeply northwest-dipping beds face toward the northwest. $S_1$ cleavage is oriented N. 65° E., 60° N., which is nearly parallel to the strike of bedding, but less steeply dipping ($S_1$ on stereoplot E, pl. 1*B*). Cleavage-bedding intersections trend generally N. 65° E. as shown by the diamonds ($L_1$) on stereoplot N (pl. 1*B*). Rare kink bands deform both cleavage and bedding, indicating that Tyler rocks here were affected by two phases of Penokean deformation. Basal Keweenawan strata only about 100 m to the north dip about 80° N. Restoring the Tyler Formation to its pre-Mesoproterozoic orientation indicates that bedding was horizontal and cleavage dipped about 20° S. There is no evidence of folds within the Tyler Formation in the central Gogebic iron range; all of the reported top determinations indicate that north-facing strata and dips are similar throughout the unit. The bedding-cleavage intersections are unusual in that they are not parallel to mappable fold axes. The well-developed cleavage in these outcrops might be caused by tectonic compression during north-directed thrusting on concealed bedding faults, along which upper plates successively overrode lower plates and thus caused successively greater shortening upsection. These outcrops with unusually well-developed cleavage are within the highest stratigraphic levels exposed in the Tyler Formation, which suggests that penetrative deformational features may become more intensely developed at higher structural levels. On a somewhat broader

scale, bedding and cleavage in the Tyler Formation between Hurley and Upson have a similar orientation, as illustrated in stereoplots C and D, respectively, on plate 1*B*.

In summary, the central Gogebic iron range was subjected to only minor deformation during the Penokean orogeny. The rocks of the Marquette Range Supergroup were apparently tilted as much as 20° S. Faulting occurred in two or three phases. First, bedding-parallel thrust faulting produced a small northward transport of the upper plates of the thrust faults. Later, steep faults cut these thrust faults. No folds are known in the central Gogebic iron range, except within a few meters of faults, where drag folds having amplitudes and wavelengths of a meter or less have been reported. In spite of the apparent lack of folds, some rocks high in the stratigraphic section have a well-developed slatey cleavage that is apparently related to movements during thrust faulting.

## Eastern Gogebic Iron Range

The eastern Gogebic iron range extends from the Sunday Lake fault near Wakefield, Mich., eastward for about 35 km to the western shore of Lake Gogebic (pl. 1*B*). This section contains the most intensely and complexly deformed rocks in the Gogebic iron range. The descriptions and interpretations presented here result from several studies by others during the past several decades, as well as our own studies in the 1990s, part of which were presented in Klasner and others (1998). Unlike other parts of the Gogebic iron range, the results of deformation here can be divided into structures that formed during both extensional and contractional phases.

### Extensional Phase

A prominent graben, the Little Presque Isle trough (Klasner and others, 1998), forms a triangular area in the central portion of the eastern Gogebic iron range (fig. 3). This graben, bounded by the Presque Isle fault on the southeast and the Little Presque Isle fault on the west, was subsiding during the deposition of the Ironwood Iron-Formation and the eruption of the Emperor Volcanic Complex; thus, it controlled the distribution and thickness of both of those units. Because of the Mesoproterozoic rotation of the area, the map pattern provides an oblique cross section of the now-deformed graben and the rocks that filled it. The principal evidence that both the Presque Isle and Little Presque Isle faults were active during sedimentation is (1) the drastic change in thickness of most units of the Menominee Group across the Little Presque Isle fault, and (2) the apparent termination of units against the Presque Isle fault.

The oldest unit in the graben, the Paleoproterozoic Palms Formation, has a poorly constrained thickness that is based on only a few drill holes within and outside of the graben. The thickness of the Palms does not seem to be strongly influenced by the bounding faults, so the graben may not have been active at the time of Palms deposition. The thickness of the Ironwood Iron-Formation likewise is difficult to determine with preci-
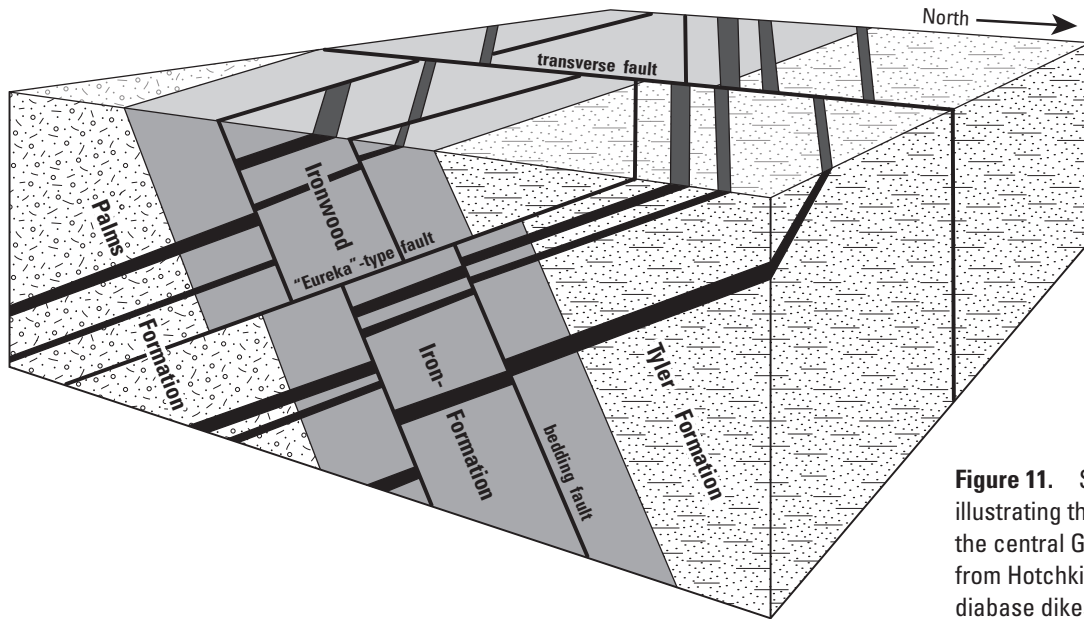
**Figure 11.** Schematic block diagram illustrating the geometry of fault sets in the central Gogebic iron range. Simplified from Hotchkiss (1919). Black units are diabase dikes.

sion because of structural complications, including stratigraphic repetitions by faulting; however, the Ironwood has an approximate thickness of 1,000 m within the graben and 650 m outside of the graben, west of the Little Presque Isle fault. Both estimates were made from cross sections by Klasner and others (1998) and Prinz (1967), each of whom constructed them based on outcrops and drill hole records.

The most dramatic change in thickness is exhibited by the Emperor Volcanic Complex. Immediately west of the Little Presque Isle fault, outside the graben, the Emperor is approximately 300 m thick; however, immediately east of the fault, within the graben, it is about 1,500 m thick (Klasner and others, 1998). Thus, within the graben the Menominee Group is about 1.5 km thicker than immediately outside of the graben. This marked change in thickness resulted from subsidence of the graben during the deposition of the Ironwood Iron-Formation and the Emperor Volcanic Complex. Later contractional deformation (discussed below) modified the graben substantially, but the relationships discussed above provide evidence for its formation during the extensional deformation that was coincident with Menominee Group sedimentation. In that regard, the graben formed in a manner similar to other long-known syndepositional structures in the region such as the Marquette trough (Larue and Sloss, 1980).

### Contractional Phase

At least two phases of thin-skinned deformation ($D_1$ and $D_2$) occurred within the contractional regime in the eastern Gogebic iron range. Both were thrusting and folding events in which Paleoproterozoic strata were detached either from underlying strata or from Neoarchean basement rocks. Slices of Neoarchean basement rocks were also incorporated into some thrust slices. In addition, later reverse faults during a

later deformation event ($D_3$) involved both Paleoproterozoic and Neoarchean rocks. The Sunday Lake fault (type 3 of Hotchkiss, 1919) is the most prominent of these.

Thin-skinned ($D_1$) deformation in the monocline west of Little Presque Isle fault resulted in the folding of bedding ($S_0$) in Paleoproterozoic strata. The $L_1$ fold axes plunge gently east (stereoplot F on pl. 1*B*). An axial-planar foliation ($S_1$) was also formed during this phase (stereoplot G on pl. 1*B*). The results of $D_1$ deformation can be seen in the westernmost outcrops of penetratively deformed Paleoproterozoic rocks on a hill, locally known as "Radio Tower Hill," in Wakefield, Mich. (SW¼ SE¼, sec. 10, T. 47 N., R. 45 W.). This hill is about 2 km east of the Sunday Lake fault. Scour channels and crossbeds in thinly bedded argillaceous siltstone of the Palms Formation indicate that stratigraphic tops are toward the north. Bedding ($S_0$) was folded about gently east-plunging fold axes ($L_1$ on stereoplot E, pl. 1*B*). Axial-planar $S_1$ cleavage is oriented N. 57° W., 35° NE. No $D_2$ deformation elements were observed at this location, but they become prominent in outcrops farther east.

$D_2$ folds were not observed either in rocks of the underlying Ramsay Formation or in the batholith of Puritan Quartz Monzonite, both of which constitute the Neoarchean basement in this part of the Gogebic iron range. A stereoplot of layering in the Ramsay (stereoplot M on pl. 1*B*) shows that it consistently strikes northwest and dips southwest, which indicates that the Ramsay was not folded during the Penokean events. Likewise, foliation within the Puritan batholith strikes east-northeast and dips steeply (stereoplot L on pl. 1*B*), in contrast to trends of Penokean deformation in the overlying strata.

The fault that detached much of the Paleoproterozoic strata from Neoarchean basement is exposed in a large outcrop in the SE¼ sec. 18, T. 47 N., R. 44 W. (Klasner and others, 1998). The unconformity between the Ramsay Formation and

the Sunday Quartzite can be seen in this outcrop. Crossbeds in the Sunday Quartzite indicate that it faces north. A mylonitic shear zone parallel to bedding is present near the top of the Sunday Quartzite. Structural vergence in the shear zone was to the north, which suggests that the Paleoproterozoic rocks were detached from the underlying Neoarchean rocks (and the basal part of the Sunday in this location) on this originally gently dipping shear zone.

D$_2$ deformation west of the Little Presque Isle fault resulted in the folding of S$_1$ axial-planar cleavage about L$_2$ fold axes that are coaxial with the D$_1$ fold axis (stereoplot G on pl. 1$B$). Although D$_2$ structures formed after D$_1$ folding, they do not necessarily record a discrete event; instead, D$_1$ and D$_2$ structures likely were formed in a continuum of north-verging, convergent deformation.

The same sequence of deformation is present in rocks east of the Little Presque Isle fault, within the Presque Isle trough. Stereoplot H (pl. 1$B$) shows that bedding was folded about gently east-plunging L$_1$ fold axes in a manner similar to that shown in stereoplot F (pl. 1$B$) west of the Little Presque Isle fault. S$_1$ axial-planar foliation related to F$_1$ folds was, in turn, folded about L$_2$ fold axes that are coaxial with L$_1$ (stereoplot I on pl. 1$B$), which is the same as that shown west of the fault. The gently plunging axis of the Wolf Mountain anticline (fig. 3) is parallel to the L$_1$ and L$_2$ fold axes found both east and west of the Little Presque Isle fault. Also, shear zones and faults of the D$_1$ phase were folded about the anticline (Klasner and others, 1998), which indicates that the anticline is a D$_2$ structure. Such structures are not found in the Neoarchean basement rocks, indicating that rocks in the Wolf Mountain anticline were detached from Neoarchean basement.

A final deformation event (D$_3$) in the eastern Gogebic iron range resulted in several sets of faults along which Paleoproterozoic and Neoarchean rocks were broken into blocks. Some of these faults are certainly of Paleoproterozoic age and, therefore, formed during the Penokean orogeny;, many of the faults, including the Little Presque Isle fault, caused an offset of the basal units of the Keweenawan Supergroup (Mesoproterozoic), which clearly indicates that the faults were reactivated during the Mesoproterozoic deformation. The Little Presque Isle fault produced nearly 500 m of offset of the surface trace of the base of the Siemens Creek Volcanics (Mesoproterozoic). This Mesoproterozoic fault movement may also be responsible for the observed relationships farther south along the fault, but Mesoproteozoic deformation is difficult to isolate from earlier deformation in those areas.

The Sunday Lake fault (pl. 1$B$) is the most prominent of the D$_3$ faults. It can be traced from well within Neoarchean basement in the eastern part of the Gogebic iron range, through the Marquette Range Supergroup to the west, to a point where it is truncated by the unconformity at the base of the Mesoproterozoic rocks near Wakefield, Mich. The fault resulted in about 2 km of left-lateral offset of surface traces of Paleoproterozoic rocks, but did not offset the basal units of the Siemens Creek Volcanics. The Sunday Lake fault was apparently a major fault that formed late during the Penokean

orogeny. A few outcrops of Neoarchean rocks along the fault trace have a vertical to steeply south-dipping foliation. Studies of the fault during mining indicated that it is nearly vertical (Hotchkiss, 1919). If 65º of Mesoproterozoic rotation is removed, the Penokean orientation of the fault is a north-striking and moderately east-dipping fault plane in which the upper plate had a significant westward reverse component of displacement, but for which total displacement is unconstrained. Elsewhere in the Penokean orogen, a deformational sequence that began with thin-skinned deformation and culminated in basement-block uplift along reverse faults has been proposed (Cannon, 1973; Klasner, 1978). Faults that bound basement blocks have diverse orientations and are commonly highly divergent from trends of older contractional structures. The Sunday Lake fault is, therefore, one of that family of late-tectonic reverse faults and the deformational sequence observed in the eastern Gogebic iron range is similar to that established elsewhere in northern Michigan. The Mesoproterozoic rotation of the eastern Gogebic iron range has provided a unique cross-sectional view of structures that can be seen elsewhere only in plan view.

Restoration of the Presque Isle trough to its pre-Keweenawan orientation provides a view of the trough, its bounding faults, and its contained structures from a different perspective than the current cross-sectional view seen in figure 3. To restore the trough to its pre-Keweenawan orientation, the structures were rotated 50º S. around an east-oriented axis (fig. 12). In this process, the fold axes' orientations were changed only slightly because they were nearly parallel to the axis of rotation. Likewise, the strike of the Presque Isle fault changed only slightly because it is nearly parallel to the rotation axis. The fault's dip, however, changed from 60º SE. to about 80º NW. (compare stereoplot A and stereoplot B on fig. 12). In the same manner, the strike of the gently southeast-dipping Little Presque Isle fault changed from N. 20º E. to N. 65º E. and its southeast dip steepened to 60º (compare stereoplot C to stereoplot D on fig. 12); thus, when the Mesoproterozoic rotation was removed, the Little Presque Isle fault and the Presque Isle fault diverge only slightly in strike and define a gently east-northeast-plunging trough bounded by steep faults (see E on fig. 12) which is similar to the west-plunging Marquette trough to the east.

Mesoproterozoic rotation of the trough has some other implications. The Little Presque Isle fault is one of a family of faults that crosscut the Presque Isle trough (fig. 3). The faults extend northward, offsetting the unconformity at the base of the Mesoproterozoic rocks, indicating that the Little Presque Isle fault set was reactivated during the Mesoproterozoic deformation; therefore, the Mesoproterozoic reactivation of these faults probably occurred during south-directed thrusting on the Atkins Lake-Marenisco fault. The steep faults may have formed in order to accommodate the changing shape of the upper plate of this regional listric thrust fault.

The easternmost extent of our studies included outcrops of the Copps Formation near Lake Gogebic. As shown in plate 1$B$ and fig. 3, the unconformable base of the Copps Forma-

tion truncated older parts of the Marquette Range Super-group. Eastward from sec. 23, T. 47 N., R. 43 W., the Copps lies directly on Neorchean granitic rocks. Rocks in Gogebic County Park near the south end of Lake Gogebic (NW¼,SW¼ sec. 3, T. 40 N., R. 42 W.) contain many structural elements which are summarized here for comparison with those described in the Gogebic iron range farther to the west. At the park, the Copps Formation consists of medium- to fine-grained graywacke with graded beds, refracted cleavage, and stretched concretions, all of which provide information on the structural history of the easternmost part of the Gogebic iron range.

The initial phase of deformation resulted in $F_1$ folds that plunge 50º N., 75º E. within the Copps. As shown in stereo-plot J of plate 1B, poles to bedding ($S_0$) indicate little evidence of small-scale folding and only two $L_1$ fold axes were observed; however, there is abundant evidence for folding of the $S_1$ cleavage that is axial planar to $F_1$ folds. $S_1$ is folded about $L_2$ fold axes that plunge 50º at S. 60º E. (stereoplot K on pl. 1B), and $S_2$ axial-planar foliation is also folded, both of which indicate that there were additional phases of deformation in these rocks. The long axes of the stretched concretions within the Copps plunge 59º, S. 24º E; thus, the Copps at the easternmost end of the Gogebic iron range was deformed at least twice about east-trending fold axes. When corrected for Keweenawan rotation, both the $L_1$ and $L_2$ fold axes plunge moderately toward the northeast and roughly parallel to the $L_1$ and $L_2$ folds in the Presque Isle trough, which suggests that both areas were subjected to the same deformational events.

## Summary of Eastern Gogebic Iron Range Structure

The eastern Gogebic iron range was substantially more tectonically active during the Penokean orogeny than was the central part of the iron range. The eastern Gogebic iron range is unique in that it preserves clear evidence of extensional deformation that coincided with deposition of the Menominee Group. Growth faults bounding a graben (the Presque Isle trough) exerted a strong influence on the character and thickness of the Ironwood Iron-Formation and Emperor Volcanic Complex. Contractional deformation was also pronounced and consisted of an early, thin-skinned phase that resulted in folds and cleavage that were geometrically uniform throughout the entire eastern part of the iron range, but were more widely and intensely developed to the east. The contractional deformation was absent in Neoarchean basement rocks; the fabrics of the Neoarchean rocks still reflect the Neoarchean deformation, thereby attesting to the thin-skinned character of this phase of deformation. A final Paleoproterozoic deformational phase resulted in reverse faults and the uplift of the fault blocks. The Sunday Lake fault is the most prominent of these faults and appears to have thrust Neoarchean and Paleoproterozoic rocks to the west or northwest.

After the Paleoproterozoic deformation event, the entire eastern Gogoebic iron range and the adjacent Neoarchean basement were tilted northward as rigid blocks at about 1,100 Ma (Cannon and others, 1993). This resulted in the present map pattern, which may be interpreted as an oblique cross section of the Paleoproterozoic structures (such as the Presque Isle trough). The uplift also reactivated the Paleoproterozoic faults of the Little Presque Isle fault set.
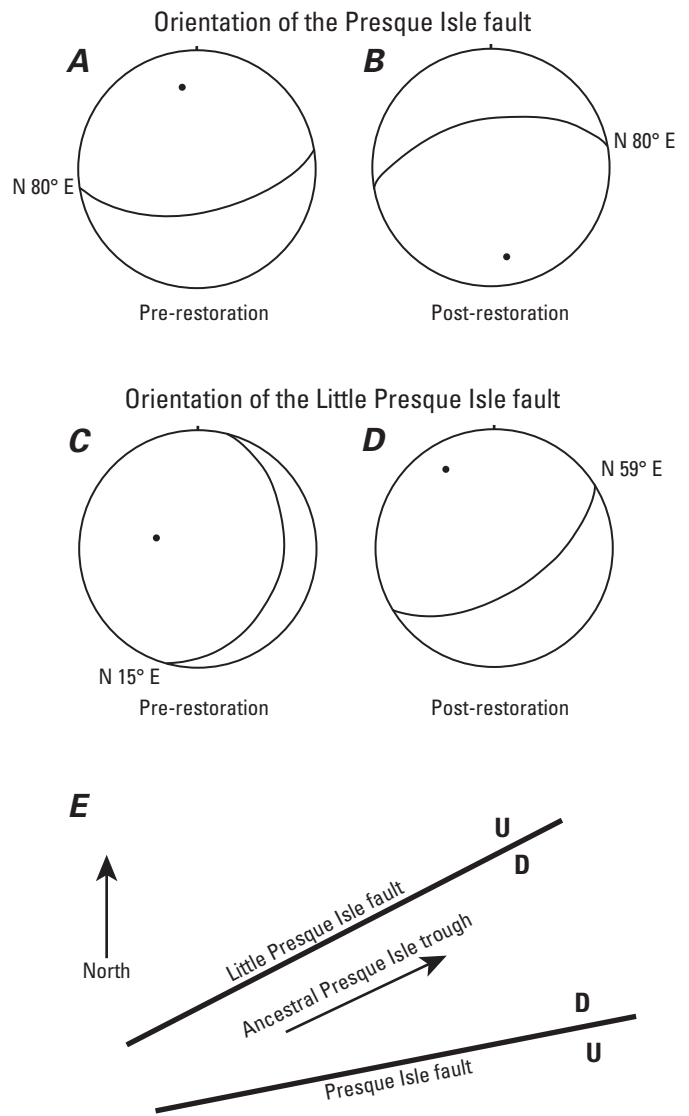


**Figure 12.**    Diagram illustrating the restoration of the Paleoproterozoic geometry of the Presque Isle trough constructed by compensating for the Mesoproterozoic northward rotation. The Presque Isle trough in Paleoproterozoic time was an east-west-trending graben bounded by normal faults, the Presque Isle and Little Presque Isle faults.

## Western Gogebic Iron Range

The western Gogebic iron range extends from Tyler Forks River for about 55 km westward to the point where the Marquette Range Supergroup rocks have been truncated by the Atkins Lake-Marenisco fault. The Tyler Forks River was chosen as the eastern boundary for the western Gogebic iron range because (1) it is approximately the western limit to which the classical internal stratigraphy of the central Gogebic iron range has been traced with confidence, and (2) it is the approximate eastern limit of intense Penokean deformation in the Marquette Range Supergroup.

The western Gogebic iron range differs from the central Gogebic iron range both in lithologic character (described above) and in structure. Compared to the central and eastern parts of the range, the western Gogebic iron range has been much less studied, with the exception of unpublished mining company studies. The most recent comprehensive description was by Aldrich (1929). As a result, we examined the western part of the range in some detail from 1990 to 1994, during which time we compiled maps at 1:24,000 scale and revised some aspects of the previously published geology. This work was aided by access to very detailed data gathered in the late 1950s by United States Steel Corporation (USS) as part of a detailed economic evaluation of iron resources in the area from Tyler Forks River westward to the Bad River at Penokee Gap. That work included 1:2,400-scale geologic mapping, ground magnetic surveys, and drilling. Some of that has been generalized in order to incorporate it into this report. The new 1:24,000-scale detailed map (pl. 1A) of the western Gogebic iron range forms the basis for the following descriptions of the structural history of this part of the range.

The principal new finding of this study is the recognition of faults that are here interpreted as thrust faults that formed during the Penokean orogeny. These faults are nearly parallel to bedding in the Marquette Range Supergroup rocks and lie both at the base of the section where they form basal décollements on the Neoarchean basement, and within the lowermost few hundred meters of the Marquette Range Supergroup rocks. These faults, in places, splay up in the section and result in the repetition of units such as is seen on Mount Whittlesey. Because of northward rotation of the western Gogebic iron range during the Mesoproterozoic, both the Marquette Range Supergroup rocks and the faults within them now have steep north dips. Folds of outcrop scale and larger are also common in parts of the western Gogebic iron range and are especially prominent near Mineral Lake, where a combination of upright folds formed during the Penokean orogeny and Mesoproterozoic rotation produced the present geometry of recumbent folds. The following paragraphs describe these structures from the east, near Tyler Forks River, to the west, ending where the Marquette Range Supergroup rocks were truncated by the Atkins Lake-Marenisco fault.

Immediately west of Tyler Forks River tight chevron folds were observed in a few outcrops of the Ironwood Iron-Formation, which indicates Penokean deformation; Aldrich

(1919) described tightly folded rocks in this same area. Most folds have axial planes inclined moderately to the south. The Marquette Range Supergroup rocks appear to be autochthonous over Neoarchean basement. Outcrops of the lower part of the Palms Formation, which is probably only a few meters above the basal contact with Neoarchean basement, show no tectonic fabric. Within a few hundred meters east of the map area of plate 1A, a well-exposed unconformity was described by Aldrich (1929) in which tens of meters of paleotopographic relief was observed on the upper surface of the Neoarchean rocks. The basal beds of the Palms Formation fill the topographic depressions on that surface. The basal décollement appears to be absent this far east in spite of its clear presence only a few kilometers to the west, as described below.

We have inferred that a basal décollement extends eastward from the Mount Whittlesey area, where it is well exposed, to approximately NE¼ sec. 11, T. 44 N., R. 2 W., a few hundred meters east of its easternmost exposure. The eastern extent of the detachment is not well constrained by outcrop information because there are no additional exposures of the basal contact of the Marquette Range Supergroup for several kilometers to the east. Westward from Ballou Creek, there are abundant outcrops in the Mount Whittlesey area. The detailed studies by USS guided our study of that area and were reinterpreted in several areas in order to produce the map on plate 1A. The geology in this area is of sufficient complexity that, even with very detailed maps, magnetic surveys, and drilling by USS, and the recent work for this report, the structure is not unequivocally delineated.

Mount Whittlesey is the easternmost intensely deformed area of the western Gogebic iron range and thus provides an example of the nature of the leading edge of the Penokean foreland deformation belt. The area presents a structural pattern in which the Marquette Range Supergroup rocks were detached from Neoarchean basement. Thrust faults splayed up from the basal décollement and produced structural repetitions of the Ironwood Iron-Formation. Folds of various scales are also present. The map pattern yields an oblique cross section of the Penokean structures which approximately parallels the trend of those structures. Cross section B–B′ on plate 1A shows a transverse view of the structures, both as presently configured (section A) and with Mesoproterozoic rotation removed to approximate the original Penokean geometry (section B).

The basal décollement is best seen in exposures in the NW¼ sec. 11, T. 44 N., R. 2 W., where the basal breccia unit of the Palms Formation overlies dacite breccia of the Ramsay Formation. The dacite breccia has a strong tectonic fabric that resulted from Neoarchean deformation. Breccia fragments were highly stretched and define a lineation plunging 20° to 30° NW. A weak to moderate foliation strikes generally WNW. and dips about 50° S. When corrected for Mesoproterozoic rotation, the dips of the lineations and foliation become steeper and, therefore, consistent with structures in other Neoarchean rocks in the region. Near their contacts with the Marquette Range Supergroup, a prominent shear fabric is present in both

the Neoarchean volcanic rocks and in the Palms Formation. This foliation is in the lowermost few meters of the Palms, but extends for 100 m or more into the Ramsay Formation. The foliation strikes about N. 80° E., dips from 20° to 75° N., and, in some outcrops, overprints south-dipping Neoarchean foliation. One additional small outcrop in the NW¼ sec. 16, T. 44 N., R. 2 W. shows a basal shear zone in which the Bad River Dolomite, a unit too small to show on plate 1*A*, has highly sheared and silicified rocks near its base; the shearing and silicification presumably formed during thrusting on a basal detachment zone.

Several additional thrust faults account for much of the structural complexity of the Mount Whittlesey area. The thrusting appears to have cut Neoarchean rocks, at least locally, such as in SE¼ sec. 10, T. 45 N., R. 2 W., where a sliver of Palms Formation and Ironwood Iron-Formation is structurally overlain by volcanic rocks of the Ramsay Formation. A thrust fault that splayed from the basal décollement on the south side of Eagles Peak migrated up section to the east and repeated the Ironwood Iron-Formation; this repetition accounts for the unusual apparent thickness of the Ironwood on Mount Whittlesey. Still farther east, that thrust fault again became the basal thrust. The existence of these faults was documented best by USS mapping that showed offsets of units within the Ironwood Iron-Formation. These internal units are not shown on plate 1*A*.

Folds are common in the Ironwood Iron-Formation on Mount Whittlesey. For instance, the hill containing the eastern part of the lowermost thrust panel in the SW¼ sec. 10, T. 44 N., R. 2 W., is underlain by Ironwood that has been folded into a west-plunging syncline. Additional folds that have wavelengths and amplitudes of roughly 10 m are well exposed in an abandoned railroad cut at the Berkshire mine in the SE¼SW¼ sec. 9, T. 44 N., R. 2 W. The detailed maps by USS indicate several areas where large-scale folds were inferred within individual lithologic units of the Ironwood, but the reason for this interpretation is not obvious from the information examined for this report; therefore, the folds are not shown on plate 1*A*. Folds at outcrop scale are also widely distributed. Stereoplot Q (pl. 1*B*) of poles to bedding ($S_0$) shows that beds were folded about $L_1$ axes that plunge gently westward. Axial-planar $S_1$ foliation was, in turn, folded about $L_2$ fold axes that plunge gently at S. 72° W., as shown on stereoplot P (pl. 1*B*).

Restoring the structures to their Paleoproterozoic orientation by removing 70° of Mesoproterozoic rotation produced relatively little change in the fold-axis orientations because they plunge roughly parallel to the axis of rotation. Corrected $L_1$ fold axes plunge gently toward the east, whereas $L_2$ axes plunge gently southwest. At the eastern end of Mount Whittlesey, the basal shear zone strikes about N. 70° E. and dips 60° to 75° NW., and bedding in the overlying Paleoproterozoic rocks dip 45° to 67° NW. When corrected for Keweenawan rotation, the shear zone becomes nearly horizontal and bedding in the overlying Paleoproterozoic rocks dips gently southward.

West from Mount Whittlesey, the map pattern of the Marquette Range Supergroup rocks shows a north-facing monocline that is similar to the structure of the central Gogebic iron range, yet folds and faults are widespread although not of sufficient magnitude to have repeated map units. In the Penokee Gap area, where the Bad River transects the range and for about 5 km farther west, mapping is based on both the USS data, which extends westward through sec. 14, and our own observations farther west. A description of this area was included in a field guide (Klasner and LaBerge, 1996) and much of the following is an excerpt from that guide. The north-facing monocline of Paleoproterozoic rocks is evident in stereoplot B (pl. 1*B*). The Paleoproterozoic rocks have a continuous basal unit of chert breccia in the Palms Formation. Small remnants of the Bad River Dolomite occur locally, such as along the Bad River in SW¼ sec. 14, T. 44 N., R. 3 W., but are too small to show at the scale of plate 1*A*. The remainder of the rocks consist of the typical sequence of Palms Formation overlain by the Ironwood Iron-Formation and Tyler Formation.

Sparse outcrops of Neoarchean rocks south of the range have a nearly vertical, east-striking foliation and parallel shear zones. Neither of these structural elements was observed in the overlying Paleoproterozoic rocks; therefore, they are the result of Neoarchean deformation. Two closely related phases of Penokean deformation were recognized in the Penokee Gap area ($D_1$ and $D_2$). $D_1$ resulted in thin-skinned, north-verging thrusting that (1) detached the Paleoproterozoic strata from underlying Neoarchean basement rocks, or (2) displaced Paleoproterozoic rocks on shear zones approximately parallel to bedding. $D_1$ deformational features are widespread in the Penokee Gap area. They include folds ($F_1$) in bedding ($S_0$) about gently northwest-plunging axes in the Ironwood Iron-Formation and Palms Formation (stereoplot A on pl. 1*B*). Similar folds as well as axial-plane cleavage are locally present in the Tyler Formation. Fracture cleavage formed locally in the Ironwood Iron-Formation (stereoplot B on pl. 1*B*).

Prominent shear zones shown by $S_1$ mylonitic foliation are present in the Palms Formation and at the contact between the Ironwood Iron-Formation and Tyler Formation. The Palms is a thin-bedded, quartzose argillite immediately above its basal breccia member. The unit contains a mylonitic foliation approximately parallel to bedding, which is apparently a zone of concentrated shear across which overlying units were detached from the basal Paleoproterozoic and Neoarchean basement rocks. A similar shear zone lies along the base of the Tyler Formation. The shear zone in the Palms Formation verges northward, as indicated by the displacement and folding of individual beds across the mylonitic foliation. Northward vergence in the Tyler Formation is also indicated by locally observed small-scale folds.

$D_2$ deformation was characterized by the continued formation of north-verging thrust faults and the continued deformation along the $S_1$ shear zones. Within the Palms Formation, prominent $F_2$ kink folds in $S_1$ mylonitic foliation plunge gently to the northwest (stereoplot B on pl. 1*B*) and

are prime examples of the $D_2$ deformation. $F_2$ kink folds were not observed in the Tyler Formation, but $S_1$ foliation appears to have been folded about northwest-plunging axes that are mostly parallel to $F_2$ axes (stereoplot B on pl. 1*B*).

When corrected for the Mesoproterozoic rotation, the Paleoproterozoic rocks dip gently to moderately to the south. The shear zones in the Palms and Tyler also dip to the south and clearly seem to have originally been north-verging thrust faults. Both $L_1$ and $L_2$ folds plunge gently east to slightly southeast, roughly parallel to the $L_1$ and $L_2$ folds in the eastern Gogebic iron range.

Still farther west, the area south and east of Mineral Lake (pl. 1*A*) contains the most intensely folded Paleoproterozoic rocks in the western Gogebic iron range. Outcrops of the Ironwood Iron-Formation and Palms Formation contain many northwest-plunging folds that have wavelengths of a few hundred to about five hundred meters. Smaller-scale folds are also common at outcrop scale. Fold axes, as measured in outcrop-scale folds, bear roughly N. 45° W. and plunge 45° NW. Axial surfaces have moderate to shallow southwesterly dips. Figure 13, for instance, shows a tight fold in the Ironwood Iron-Formation, which is typical of the general geometry of folds in this area. Axial surfaces are inclined to the south and anticlines have gently south-dipping south limbs and steep to overturned north limbs. The folds are apparently rootless and the Paleoproterozoic rocks are apparently detached from the Neoarchean basement rocks. A basal detachment is not exposed here, but some of the rocks in the southernmost outcrops of the Palms Formation, such as at the southeast corner of sec. 18, T. 44 N., R. 3 W., have a strong foliation not generally common in the unit, which suggests that there is probably a shear zone immediately south of these outcrops. A splay from the basal detachment has thrust Palms Formation over the Ironwood Iron-formation in the SE¼ sec. 8, T. 44 N., R. 3 W. The only outcrop of Neoarchean rocks mapped in the area is in the NW¼ sec. 24, T. 44 N., R. 4 W. Here, the rock is a massive granitoid of the Puritan batholith and is only about 100 m from the probable basal detachment. The lack of a Penokean fabric and the proximity of this outcrop to intensely folded Marquette Range Supergroup rocks further indicate the thin-skinned nature of the deformation. When the effects of Mesoproterozoic rotation are removed, the folds are upright, the basal detachment is subhorizontal (pl. 1*A*, cross section A–A′), and fold axes have shallow plunges in a more westerly direction than their present orientation.

An unusual feature of the Mineral Lake area is a thick section of the Bad River Dolomite found in the NW¼ sec. 24, T. 44 N., R. 4 W., which appears to be several hundred meters thick. The rock consists of cherty, dolomitic marble that is typical of the Bad River elsewhere in the Gogebic iron range. This section is the only known preserved Bad River for many kilometers along the strike of the range and appears to have been overridden by the basal detachment of the main part of the range. This section of Bad River Dolomite may be a fault block that was dropped down early in the history of the area, before the Bad River-Palms erosion interval, and thus was

preserved because of its structural position. If that is the case, the Bad River must have been an extensive unit in the area and was removed by pre-Palms dissolution and erosion. Tight folds are present in some outcrops and have a geometry similar to that seen to the north in the Ironwood, which suggests that the Bad River was folded during the same events as the overlying rocks.

The Marquette Range Supergroup was truncated on the west by the base of the Mineral Lake intrusion (a large mafic pluton). Metamorphism of the country rock was intense and, within a few hundred meters of the contact, the Ironwood was converted to a coarse-grained granoblastic rock composed of quartz (recrystallized chert), magnetite, orthopyroxene, and local fayalite. Near the intrusive contact, bedding of the Ironwood strikes to the northeast, nearly parallel to the base of the intrusion, which suggests that local deformation resulted from emplacement of the pluton.

Gabbro of the Mineral Lake intrusion truncated the Marquette Range Supergroup rocks in the NW¼ sec. 14, T. 44 N., R. 4 W., and the strata are absent for about 10 km to the west where they reappear, first as an enclave within the Mineral Lake intrusion, and finally as a continuous section in the footwall of the Atkins Lake-Marenisco fault.

Some important relationships are shown in outcrops along the Marengo River in secs. 14, 15, and 16, T. 44 N., R. 5 W. The geology of the Marengo River area was discussed above in the section on Mesoproterozoic structures because of the excellent exposures of the Atkins Lake-Marenisco fault in the area. The following paragraphs explain some additional aspects of the Paleoproterozoic geology.

The Marengo River area differs from other parts of the Gogebic iron range in being mostly in the footwall of the Atkins Lake-Marenisco fault; however, it is also in the hanging wall of the related Pelton Creek fault. Because of this relationship, the area apparently was rotated to the northwest somewhat independently of the rocks in other parts of the iron range. Paleoproterozoic rocks are preserved in two areas. An



**Figure 13.**    Small scale recumbent fold in Ironwood Iron-Formation in NE¼, sec. 24, T. 44 N., R. 4 W. View looking west.

enclave of strongly metamorphosed Paleoproterozoic rocks consisting of the Ironwood Iron-Formation and unnamed gabbro was mapped within the Mesoproterozoic Mineral Lake intrusion. The enclave underlies a triangular area in secs. 13, 14, 23, and 24, T. 44 N., R. 5 W. and is truncated on the west by the Atkins Lake-Marenisco fault. The Ironwood is mostly an evenly-bedded cherty, iron-amphibole-rich rock containing variable amounts of magnetite. No deformational features were observed in the relatively few outcrops of the iron-formation, all of which contain beds that dip moderately to steeply to the north-northwest.

The Paleoproterozoic rocks south of the Atkins Lake-Marenisco fault form a northwest-dipping monocline above Neoarchean granitic rocks. The basal unit of the Marquette Range Supergroup (the Bad River Dolomite) lies unconformably on the granite. The basal unconformity is well exposed near the southeast corner of sec. 15, T. 44 N., R. 5 W. At that location, neither the dolomite nor the granite show tectonic fabric. The dolomite generally does not show deformation fabric in the widespread outcrops and in a quarry to the southwest. These observations indicate that the Bad River was essentially unaffected by Penokean deformation and that the 45° to 55° dips are a result of Mesoproterozoic rotation. A Mesoproterozoic diabase dike near the quarry in sec. 22, T. 44 N., R. 5 W., dips 50° S., which indicates substantial Mesoproterozoic rotation of this structural block. Although there is no unique solution for the amount and direction of rotation, it seems likely that the Bad River Dolomite was nearly flat-lying prior to the Mesoproterozoic deformation.

Evidence for substantial Penokean deformation appears higher in the stratigraphic section. A broad zone of intense shearing is well exposed in the argillite near the base of the Palms Formation that is parallel to its contact with an underlying sill of metamorphosed gabbro in the SW¼ sec. 15, T. 44 N., R. 5 W. (pl. 1A). The stratigraphically lowermost 50 m of the shear zone largely obliterates the bedding, except for the most quartzose of the layers, which are tightly folded. Those folds were largely dismembered by attenuation of their limbs. The intensity of deformation diminishes upsection, but the effects of shearing are seen for at least an additional 100 m. Small folds indicate a northward vergence. The shearing is roughly parallel to bedding in the Palms and is generally parallel to the strike of the rock units in the area. The shear zone, therefore, appears to be a major detachment surface along which the Palms Formation, Ironwood Iron-Formation, and metagabbro sills were thrust over undeformed Bad River Dolomite, and would have been a low-angle detachment prior to Mesoproterozoic tilting.

The Palms Formation and Ironwood Iron-Formation overlying the shear zone constitute a northwest-facing monocline, but show small folds in some outcrops. The few observed fold axes plunge at moderate angles to the northwest and restore to shallowly plunging folds when the Mesoproterozoic rotation is removed. The general trend of Paleoproterozoic units is interrupted by a plug of massive, pink, coarse-grained granite in sec. 20, T. 44 N., R. 5 W. The granite does not show deforma-

tional fabric and is lithologically identical to granitic phases of the Mesoproterozoic Mellen Intrusive Complex; therefore, we interpret it to be a Mesoproterozoic intrusion.

The westward continuation and eventual termination of the Marquette Range Supergroup west of the Marengo River area is poorly defined. The westernmost bedrock exposures of the strata are in sec. 26, T. 44 N., R. 6 W. The Ironwood Iron-Formation in that area consists of thin-bedded, slightly cherty rock composed mostly of iron-amphiboles, and only minor amounts of magnetite. Drill holes in the area also indicated similar weakly magnetic iron-formation interlayered with black shale. Additionally, metagabbro sill crops out in sec. 26. We have attempted to trace the Ironwood westward by magnetic surveys (including ground traverses) but the anomalies are very weak and indicate that the magnetite content continues to decline to the west. Our interpretation is that the Marquette Range Supergroup rocks were truncated against the Atkins Lake-Marenisco fault in secs. 33 and 34, T. 44 N., R. 6 W. The westward extension of the Bad River Dolomite is even more problematic. There are no exposures of it west of the abandoned quarry in sec. 22, T. 44 N., R. 5 W. so the projected continuation of it to the west and its truncation by the Atkins Lake-Marenisco fault is speculative.

# Economic Geology

The Gogebic iron range was a major iron ore producer for 90 years from 1877 to 1967. During that period, about 325,091,000 tons of ore were shipped from about 40 mines between Upson, Wis., on the west and Wakefield, Mich., on the east. This 40-km-long belt was mined nearly continuously, mostly from underground workings. The Gogebic iron range ranks third nationally in natural ore production behind the Mesabi iron range in Minnesota and the Marquette iron range in northern Michigan. Ore bodies are masses of nearly pure iron oxides and hydroxides that formed as secondary concentrations (so-called "natural ores") within the Ironwood Iron-Formation. The term "natural ore" distinguishes these deposits from lower-grade concentrating ores (commonly called "taconite") which must be concentrated in order to produce a useable product. Taconite has never been produced from the Gogebic iron range, but a detailed assessment of its western part from Upson to Mineral Lake indicates that this 35-km-long belt contains a resource of about 3.7 billion tons of material (Marsden, 1978) that was deemed to be economically mineable at the time of that assessment, making the western Gogebic iron range one of the largest undeveloped taconite resources in the United States.

## Natural Iron Ores

Although the presence of iron-bearing rocks was known as early as 1848 from observations by linear surveyors near Upson, Wis., iron ore was not discovered until many years

later when a trapper discovered ore beneath an overturned tree at what became the Colby mine near Bessemer, Mich. (Reed, 1991). The first mining commenced in 1884 and within four years annual shipments from the range exceeded one million tons. Detailed examination of the district began shortly afterward by the U.S. Geological Survey and culminated with a report by Irving and Van Hise (1892) that presented a remarkably complete and accurate model of the nature, control, and origin of the natural ores. Although refined by later works such as Hotchkiss (1919), the model was not substantially modified for more than a century and withstood perhaps the ultimate test—it guided further exploration that led to the discovery of hundreds of millions of tons of ore.

Natural ores occur exclusively in the Ironwood Iron-Formation as irregular masses of soft, earthy iron oxides and hydroxides. Iron-carbonate and iron-silicate minerals within the Ironwood oxidized and chert was replaced with iron hydroxides and oxides. In some ores, relict bedding was preserved. The ore bodies are mainly localized along the keels of plunging structural troughs, which are most commonly formed by the intersection of diabase dikes with the top of the Palms Formation or with the top of shaley units within the Ironwood. Figure 14 is simplified from Hotchkiss (1919, fig. 26) and summarizes the essential features of the geometry of the ore bodies. Iron ore extends from the surface to depths in excess of 1.5 km; the greatest depth of mineralization has not been determined.

The origin of ore as a product of a structurally focused flow of deeply circulating, oxygenated ground water that both oxidized the original iron minerals and replaced the

chert with iron minerals was first proposed by Irving and Van Hise (1892) and is still the accepted theory of the ore genesis mechanism. More than 70 years of mining and study of the Gogebic iron range has not produced any convincing evidence to contradict their original hypothesis.

The age of mineralization is quite unconstrained. Irving and Van Hise (1892) correctly surmised that ores formed entirely after the rocks were in their present structural orientation so that dike-bed intersections formed plunging troughs. The structural tilting of the range occurred at about 1,060 Ma (Cannon and others, 1993), thus placing a lower limit on the age of mineralization. A similar constraint is suggested by the distribution of ore relative to the effects of Mesoproterozoic metamorphism. A relationship between metamorphic recrystallization and ore formation was noted by James (1955). Natural ores in all of the iron ranges in the Lake Superior region are restricted to areas of very low-grade metamorphism where the original very fine grain size was preserved. Even modest metamorphic recrystallization inhibited ore formation, apparently because the increase in grain size caused by metamorphic recrystallization diminished the grain surface area-to-mass ratio of individual iron minerals and made oxidation and dissolution less efficient. In the Marquette iron range, for instance, natural ores are confined to the eastern part of the range, which is in the chlorite zone of regional metamorphism, but they are absent elsewhere in the range, where, although geologically similar, the rocks are metamorphosed to biotite and higher grades (James, 1955). A similar relationship is present in the Gogebic iron range. The western limit of natural ores near Upson, Wis., coincides with the appearance
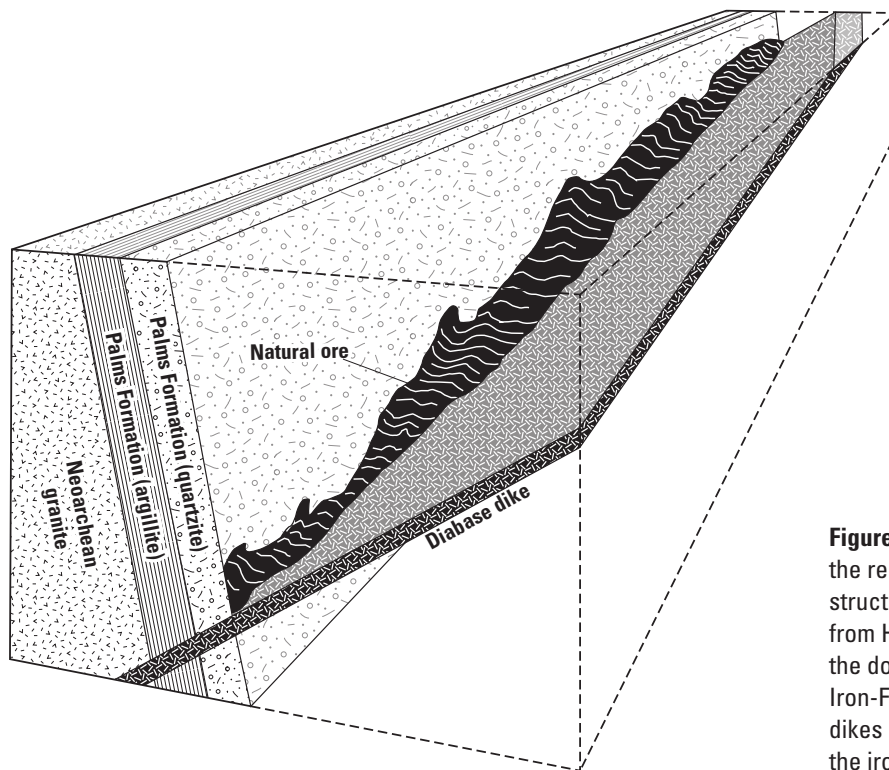


**Figure 14.** Schematic block diagram showing the relation of natural ores to the stratigraphy and structure of the central Gogebic iron range. Simplified from Hotchkiss (1919). Ore bodies were localized in the downward-closing structures in the Ironwood Iron-Formation formed by intersection of diabase dikes and impermeable shale beds beneath or within the iron-formation.

of metamorphic recrystallization in the contact metamorphic aureole of the Mellen Intrusive Complex. Natural ores are absent in the remaining western part of the range. The age of the Mellen Intrusive Complex is about 1,100 Ma (Cannon and others, 1996), which indicates that ore formation must have been younger, but there is virtually no upper limit on the age of mineralization.

Some aspects of ore formation are still poorly understood. There has been little modern interest in studying these ores, because the ore bodies have been largely depleted and are otherwise of little commercial interest. In addition, access to ore bodies is no longer available. Some interesting remaining questions are as follows:

1.  Under what topographic, tectonic, and climatic conditions can vigorously circulating ground-water systems reach depths of more than one kilometer?

2.  What was the chemistry of ground water that allowed large amounts of iron to be dissolved, transported, and redeposited under apparently strongly oxidizing conditions in which ferric iron is generally virtually insoluble?

Recently, Morey (1999) proposed that the high-grade iron ores of the Mesabi iron range in Minnesota, which are in some respects similar to ores on the Gogebic iron range, may have formed as a result of large-scale lateral movement of heated formation waters during Penokean deformation. According to his theory, the Mesabi ores were formed in Paleoproterozoic time and their origin is not related to the present erosion surface. The Gogebic iron range ores clearly seem to have been controlled by the Mesoproterozoic structures and metamorphism, and are therefore unlikely to have formed during Paleoproterozoic fluid flow as suggested for the Mesabi iron range.

Iron ores in all of the mining districts in northern Michigan and Wisconsin contain a suite of accessory minerals including adularia, barite, calcite, manganite, psilomelane, and rhodochrosite. These minerals typically occur in cavities within the hematite and goethite or limonite ores. The accessory minerals locally cement fragments of the iron ore. Their nature and occurrence suggest that they formed by the same processes that produced the iron ore. These accessory minerals have the potential to provide information as to the age and temperature of formation of the ores; however, no systematic study of these minerals has been undertaken.

## Taconite

Taconite has not been mined on the Gogebic iron range, but a large taconite resource has been identified in the western part of the range. The resource was quantified by Marsden (1978) in a study for the U.S. Bureau of Mines. Much of Marsden's work was based on information gathered by U.S. Steel Corporation during the late 1950s (see previous discussion on the structure of the western Gogebic iron range), and on the projection of that information into less studied areas. Marsden (1978) estimated that a 35-km-long belt of taconite extended from Upson to Mineral Lake in Wisconsin. He divided the Ironwood Iron-Formation into five units that are somewhat different from the classic members established farther to the east. Based on metallurgical tests that indicated the percentage of recoverable iron as magnetite, he identified two continuous ore zones separated by a layer of waste rock. The lower ore zone included most of the Plymouth Member and the upper ore zone included mostly the Pence and Norrie Members. Marsden then did an economic simulation of the amount of material that could be mined and concentrated at a profit using economic assumptions from Mesabi iron range taconite operations. The principal economic constraint was the depth to which open-pit mining could proceed in these steeply dipping units before the cost of stripping the hanging-wall waste rock made deeper mining uneconomic. Marsden calculated that 3,711,000,000 tons of taconite could be mined profitably with prices, costs, and technologies of the mid-1970s. This resource constitutes one of the largest undeveloped iron resources of the Lake Superior region. Economically recoverable amounts of ore may be substantially different from Marsden's estimate of 25 years ago and could be either larger or smaller depending on the interplay of value and cost of production. The previous estimates did not consider environmental and aesthetic factors of large open-pit mines and related processing facilities, which would undoubtedly be major factors affecting the future of this resource.

Parts of the Ironwood Iron-Formation in the eastern Gogebic iron range are also considered to be a taconite resource, but are much less surely quantified. The rather broad outcrop belts of mostly moderately dipping magnetic iron-formation from sec. 17, T. 47 N., R. 44 W., to sec. 22, T. 47 N., R. 43 W., seem to offer some possibility of future economic mining. Metallurgical data on recoverable iron is not available for these deposits so the percentage of the iron-formation that might contain acceptable grades of iron recoverable by magnetic separation cannot be estimated.

# Synopsis

The Gogebic iron range is an integral part of the Paleo-proterozoic continental margin assemblage, which was deposited on the south-facing margin of a craton composed of Archean crystalline rocks. The iron range formed near the inner edge of the foreland fold-and-thrust belt of the Penokean orogeny in which the rocks of the Marquette Range Supergroup were deposited and soon after were deformed. The rocks of the Gogebic iron range provide the most complete and best-studied record of events in this part of the orogen.

## Sedimentation of the Marquette Range Supergroup

### Environmental Setting of Deposition

#### Chocolay Group

The earliest recorded deposits of the Marquette Range Supergroup are clean quartz sands of the Sunday Quartzite, which is the basal unit of the Chocolay Group. Sedimentary structures indicate deposition in shallow water, at least partly in a tidal environment. The Sunday Quartzite, known only in the eastern part of the Gogebic iron range, becomes thinner toward the west and pinches out near the town of Wakefield, Mich. The extent to which this pinch-out is erosional rather than depositional is not known. The overlying Palms Formation lies unconformably on the Sunday so it is likely that the Sunday originally extended somewhat farther west than its current outcrop limit. The Sunday Quartzite grades upward into the Bad River Dolomite of the Chocolay Group, which indicates continuous deposition during which the clastic quartz supply was lost. The thick, partly stromatolitic, carbonate succession of the Bad River accumulated in shallow water. The Bad River is as much as 300 m thick in both the eastern and western parts of the iron range. In the eastern part of the iron range the Bad River always overlies the Sunday Quartzite, but in the west it directly overlies Neoarchean basement rocks and has only a few meters of arkosic dolomite at its base. The clean quartz sand of the Sunday Quartzite appears to be completely absent in the western part of the iron range. In the central part of the range, the Bad River is present only in a few isolated patches; however, the extensive chert breccia unit of the basal Palms Formation, which is interpreted to be a residuum of chert produced by dissolution of the Bad River, indicates that the Bad River Dolomite was probably continuous along most or all of the Gogebic iron range but was removed by erosion before the Palms Formation was deposited. In summary, the quartz sands of the Sunday Quartzite probably had a depositional pinch-out somewhere west of Wakefield, Mich. The Bad River Dolomite overstepped the pinch-out and was deposited as a continuous carbonate bank over the entire Gogebic iron range.

The time span between deposition of the Bad River Dolomite and the unconformably overlying Palms Formation is not well constrained, but may be substantial. Many authors have proposed that the Chocolay Group correlates with the upper part of the Huronian Supergroup of Ontario (Puffett, 1969; Young, 1970; Ojakangas, 1982, 1985). If that is correct, then the Chocolay Group must be older than the 2,200 Ma Nipissing Diabase (Corfu and Andrews, 1986), which intruded rocks of the Huronian Supergroup. Recent age determinations for the Menominee Group and equivalent rocks (Fralik and others, 2002; Schneider and others, 2002) indicate and age of roughly 1,870 Ma. Thus, an erosion interval of more than 300 my might separate the two groups.

### Menominee Group

The deposition of the Menominee Group, which is the principal iron-bearing group of the region, began with fine-grained, laminated muds of the lower Palms Formation. In the central part of the Gogebic iron range, the paleosurface was mantled with a residuum of chert fragments that formed by the dissolution of the Bad River Dolomite during the post-Chocolay Group erosion interval. This residuum was slightly reworked and mixed with quartz sand during the Palms transgression to form the basal chert breccia unit. Where the chert breccia is absent, the laminated argillite appears to lie directly on older rocks, which indicates that the base of the transgressive sequence formed in a relatively low-energy environment in which finely laminated muds were able to accumulate. The lower argillite grades upward into a medial unit of somewhat coarser grained, siliceous mudstone and sandstone, which in turn grades upward to thick-bedded quartzite (Ojakangas, 1983). Although the total thickness of the Palms is somewhat variable along the length of the iron range, this pattern of coarsening-upward sediments is developed everywhere. The uniformity of the internal stratigraphy suggests that the Palms was deposited under stable conditions with no local influence of either a tectonically induced subsidence or uplift.

The upper quartzite of the Palms grades upward into the Ironwood Iron-Formation. The transition zone is between 1 and 2 m thick and records the transition from the deposition of clean quartz sand of the Palms Formation to the chemical sedimentation of the Ironwood Iron-Formation. The Ironwood records the earliest basin instability in both the eastern and western parts of the Gogebic iron range. The central part of the iron range remained as a stable and quiescent depositional environment throughout Ironwood deposition. Water depth (or wave-base depth) varied enough to produce frequent alternations of quiet-water and agitated-water bottom conditions, but there is no evidence of either volcanism or syndepositional faulting anywhere in the central Gogebic iron range. In contrast, intrusive and extrusive igneous rocks are common in the eastern and western part of the iron range. Volcanic rocks are interbedded with the Ironwood and gabbro sills are common. The Presque Isle trough (discussed above) in the eastern part of the iron range, is a graben that formed during Ironwood

deposition and controlled the thickness and character of the Ironwood and the interlayered Emperor Volcanic Complex (Klasner and others, 1998). The iron-rich depositional setting in which most of Ironwood accumulated eventually was transformed into a basin in which pyritic black shales were deposited. The black shales are preserved in some areas of the central part of the range and in the westernmost part of the range; however, they may have been more widespread but were removed by erosion before deposition of the Baraga Group.

The entire Menominee Group appears to be a transgressive sequence in which a barrier bar and back-bar lagoon transgressed over the low-relief surface developed on Neoarchean rocks and strata of the Chocolay Group (Ojakangas, 1983). The fine-grained, laminated sediments of the lower part of the Palms Formation, which commonly overlie older rocks without a basal conglomerate phase, indicate that the low-energy environment of a lagoon was the initial depositional setting of the transgression. Further transgression superposed the coarser quartzose sediments of the barrier bar over the lagoonal accumulations as shown by the coarsening-upward character of the Palms. The iron-rich facies that constitutes the Ironwood Iron-Formation apparently existed immediately seaward of the barrier bar and spanned the depth of wave base; therefore, the Ironwood was deposited directly on the coarse quartzose sand of the barrier bar. A euxinic basin, in which the pyritic black shale accumulated, appears to have been still farther seaward. With the continued transgression, the black shale was superposed over the iron-rich part of the Ironwood, which resulted in pyritic carbonaceous rocks in the upper part of the Ironwood and lower part of the Tyler Formation.

## Baraga Group

The Baraga Group, consisting of the Tyler Formation in the west and the stratigraphically equivalent Copps Formation in the east, overlies the Menominee Group with a low-angle unconformable contact. The base of the Copps Formation is a conglomerate composed mostly of locally derived fragments of Neoarchean rocks and iron-formation. Likewise, the Pabst Member of the Tyler Formation is composed mostly of fragments of the underlying Ironwood Iron-Formation. The time span of the unconformity is not known, but it was sufficiently long enough to have allowed the Ironwood to become lithified before it was eroded and also long enough to have allowed erosion to completely remove the Chocolay and Menominee Groups in the eastern part of the Gogebic iron range. Roughly 100 m of carbonaceous pyritic shale and cherty iron-carbonate-bearing iron-formation overlies the basal conglomerates. Most of the Copps and the Tyler is composed of graywacke and related clastic rocks that were deposited as a submarine fan. Detailed sedimentologic studies by Alwin (1976) established the turbidity flow character of this 3,000-m-thick section and documented a southerly source for the clastic material. Much of the detritus appears to have been derived from Archean crystalline rocks, which indicates that a significant

upland of Archean rocks existed south of the Gogebic iron range at this time. At present, Archean crust extends only from the study area to the Niagara fault, about 50 km south of the Gogebic iron range, beyond which the crust consists entirely of Paleoproterozoic volcanic-arc rocks. Penokean tectonism may have greatly foreshortened this part of the continental margin.

## Tectonic Setting of Deposition

The tectonic setting of deposition of the Marquette Range Supergroup has been discussed extensively (for instance, Larue and Sloss, 1980; Larue, 1981; Larue, 1983; Cambray, 1978; Hoffman, 1987). The interpretation that deposition was in a foredeep basin, as proposed by Hoffman (1987), is consistent with much of the observational evidence (particularly for rocks of the Menominee and Baraga Groups) derived from the Gogebic iron range and other iron ranges of the region.

### Chocolay Group

The interpretation that deposition of the orthoquartzites and dolomites of the Chocolay Group was on a stable platform or passive margin also has been widely accepted. Questions remain regarding the relation of the platform and passive-margin sequence to the subsequent depositional, volcanic, and deformational events. There are no radiometric constraints on the age of deposition of the Chocolay Group other than the age of the Neoarchean basement and the 1,870 Ma age for the unconformably overlying Menominee Group. A definitive solution to the age problem is not yet at hand, but the commonly proposed correlation of the Chocolay Group with parts of the Huronian Supergroup in Ontario is likely to be correct. This correlation requires the Chocolay Group to have been deposited before 2,200 Ma, which is a minimum age for the Huronian Supergroup (Corfu and Andrews, 1986). If the Chocolay Group is, in fact, older than 2,200 Ma, and if it formed during the continental breakup leading to the eventual Penokean collision at 1,870 Ma, then the passive-margin phase of the Penokean orogenic cycle must have persisted for more than 300 my.

### Menominee Group

Radiometric ages for the Menominee Group (Schneider and others, 2002) indicate that the Menominee was deposited contemporaneously with the development and accretion of volcanic arcs in the Wisconsin magmatic terranes to the south. That relationship strongly suggests a foredeep depositional setting for the Menominee. The depositional model of Hoffman (1987) proposed that iron-formations were deposited on the outer ramps of foredeep basins in areas where normal faulting (created by crustal flexure) is common. With time, migration of the foredeep basin toward the continent superposed turbidite deposits of the axial zone of the foredeep basin over the iron-bearing sequences that were deposited on

the outer ramp. Observations from the Gogebic iron range indicate that the Palms Formation was deposited in water that was shallow enough to record tidal fluctuations throughout its stratigraphic extent (Ojakangas, 1983). The overlying Ironwood Iron-Formation (at least in the central part of the range) also was deposited in water that was shallow enough to be periodically above the wave base. There are no indications that water depths were substantially below the wave base in any of the Menominee Group rocks in the central part of the Gogebic iron range. On the eastern and western ends of the range, the Palms Formation maintains its characteristic internal stratigraphy and evidence of a tidal depositional setting; however, the Ironwood Iron-Formation was deposited (at least in part) in subsiding fault-bounded troughs and was interlayered with volcanic rocks.

The Menominee Group in the central part of the Gogebic iron range appears to have been deposited in the most distal parts of the foredeep basin on the slowly and rather uniformly subsiding crust. Water depths were always shallow and fluctuated from tidal to somewhat below the wave base. Rocks on the eastern and western ends of the range appear to have been deposited in deeper water in the proximal parts of the foredeep basin, where more intense crustal flexing resulted in syndepositional normal faulting. A modified foredeep basin model was proposed by Schneider and others (2002) in which oblique collision of arc terranes accentuated crustal extension in the continental margin. This additional extension, when superposed on the extensional features that might be expected solely from orthogonal collision, might account for the well-developed depositional grabens and the abundant volcanic and mafic intrusive rocks immediately adjacent to tectonically stable parts of the basin.

The uniformly shallow-water depositional setting for the central part of the Gogebic iron range suggests deposition in the most continentward part of the foredeep basin, possibly north of a flexural forebulge (outer arch of Hoffman's (1987) model). If so, the northward migration of the forebulge may account for the unconformity between the Menominee and Baraga Groups.

## Baraga Group

Within its lowest 100 m, the Baraga Group passes from a basal conglomerate, through euxinic black shale and cherty carbonate iron-formation, into deposits of a turbidite fan. The Baraga Group records the deposition of outer ramp facies (consisting of shales and iron-formation) followed by the deposition of a thick turbidite succession (Tyler and Copps Formations) in the axial zone of the foredeep basin. The Tyler is composed mostly of granitic detritus from a highland of Archean basement rocks that existed at that time to the south of the Gogebic iron range, and was deposited by northward-flowing turbidity currents (Alwin, 1976). Tectonic foreshortening of the continental margin is suggested by the roughly 50 km distance between the Gogebic iron range and the margin of accreted arc terranes to the south. The foreshortening

may have resulted in telescoping of the outermost extensions of continental crust during collision, which then elevated the area and allowed for eventual erosional destruction of thrust sheets that consisted of Archean basement rocks.

## Volcanism

The chemistry of volcanic rocks also provides information on the tectonic environment in which the Marquette Range Supergroup was deposited. The Emperor Volcanic Complex is one of several volcanic units interbedded in the Marquette Range Supergroup. Other Paleoproterozoic volcanic units include the Clarksburg Volcanics Member of the Michigamme Formation, and the Hemlock Formation, and Badwater Greenstone. Detailed geochemistry, however, is available only for the Hemlock Formation and Badwater Greenstone (Ueng and others, 1988; Schulz, unpub. data [1984]). The composition of basalts from the Emperor Volcanic Complex compares very well with the composition of rocks from the Hemlock Formation and Badwater Greenstone. Both the Hemlock and the Badwater are predominantly subalkaline tholeiitic basalts that are characterized by moderately enriched light REE and other incompatible elements, and by variably negative Nb and Ta and positive Th anomalies (fig. 15). Unlike the Emperor Volcanic Complex and Badwater Greenstone, the Hemlock Formation is bimodal in that it locally contains abundant rhyolite of largely crustal origin (Schneider and others, 2002). The presence of crustally derived rhyolites in the Hemlock Formation is further evidence for the interaction between basaltic magmas and Archean basement during the evolution of the Menominee Group.

Determining the tectonic affinity of volcanic suites based on geochemical tectonic discrimination diagrams must be done with caution and in conjunction with knowledge of the geology of the area. This is particularly true when considering possible continental basalts as they often do not plot as expected in the within-plate basalt fields on commonly used tectonic discrimination diagrams (Holm, 1982; Wang and Glover, 1992). This results from the fact that some continental basalts show variable levels of relative depletion in high-field-strength elements (HFSE), particularly Nb, Ta and Ti, which form the basis of discrimination in many tectonic discrimination diagrams (Wang and Glover, 1992). The relative depletion of HFSE in some continental basalts, while still a subject of controversy, is typically attributed (1) to the derivation of the basalts from HFSE-depleted subcrustal lithosphere, and (or) (2) result from crustal contamination (Norry and Fitton, 1983; Dupuy and Dostal, 1984; Carlson, 1991).

The composition of the Emperor basalts, as well as that of the Hemlock and Badwater, resembles continental tholeiitic basalts; all three plot mainly in within-plate basalt fields on tectonic discrimination diagrams (fig. 16; see also Ueng and others, 1988). On some tectonic discrimination diagrams, the Emperor and related Paleoproterozoic basalts extend into arc-related basalt fields (for example, fig. 16*D*) because of the variable degrees of crustal contamination, which results in the

relative depletion in Nb and Ta and an enrichment in incompatible elements such as Th and La .

The interpretation of the Emperor and related Paleoproterozoic basalts as continental tholeiites is compatible with the geology of the Menominee Group. The sequence lacks arc-related rocks, which typically include (1) calc-alkaline andesites, and (2) large volumes of reworked volcaniclastic material that are typically part of volcanic suites in island-arc terranes. The Emperor and related basalts are either interbedded with or are laterally equivalent to relatively thick accumulations of fine-grained epiclastic and chemical sedimentary rocks, which is an association that suggests deposition in extensional basins far from an active magmatic arc. Schulz and others (1993) proposed that the Menominee Group and its contained volcanic rocks were deposited during the active rifting phase of the evolution of the Paleoproterozoic continental margin in the Lake Superior region. A more recent study of rhyolite from the Hemlock Formation yielded a U-Pb zircon age of 1,874±9 Ma (Schneider and others, 2002). This date suggests that (1) the Hemlock Formation and correlative volcanic units of the Menominee Group were coeval with arc-related volcanic and plutonic rocks that are now preserved in the accreted terranes to the south in Wisconsin (Sims and others, 1989), and (2) that the Menominee Group is in a second-order basin in a foredeep related to oblique subduction and subsequent collision along the Superior craton margin (Schneider and others, 2002).

A possible analogue of the proposed Menominee Group setting is the Late Silurian to Early Devonian rift that was caused by dextral transpression in the northern Appalachians (Dostal and others, 1993; Malo and others, 1995). Volcanic rocks of mostly Middle Silurian to Early Devonian age locally constitute a significant part of the supracrustal sequences that unconformably overlie Ordovician and older rocks that were deformed during the Ordovician Taconian orogeny (Keppie, 1992). The volcanic sequences are bimodal in that they contain (1) continental tholeiite to transitional basalts and (2) rhyolites that were derived by crustal anatexis (Dostal and others, 1989, 1993; Van Wagoner and others, 2002). Basalts from the Late Silurian to Early Devonian rift basins in the northern Appalachians have a composition that is similar to that of the Paleoproterozoic basalts of the Emperor Volcanic Complex; both types exhibit variably negative Nb and Ta anomalies (fig. 17).

The sedimentologic, structural, and geochemical evidence favors the interpretation that the Menominee and Baraga Groups were deposited along a convergent margin that lay north of the accreting arc terranes that are now prevalent south of the Niagara fault. Subsidence and uplift were largely flexural responses of the southern edge of the Archean craton that was being overridden by volcanic arcs. Extensional features such as the Presque Isle trough, which were important in localizing volcanic rocks and controlling thickness and facies variations in sedimentary rocks, were second-order extensional features caused by the oblique collision of the southern edge of the craton with the volcanic arcs of the Wisconsin magmatic terranes.



**Figure 15.** Primitive-mantle-normalized multi-element plot showing the compositional similarity of mafic rocks from the Emperor Volcanic Complex and diabase near Atkins Lake (symbols same as in figure 7) with basalts of the Hemlock Formation and Badwater Greenstone (shaded field). Hemlock data from Ueng and others (1988); Badwater data from Schulz (unpublished data). Primitive-mantle normalizing values from Kerrich and Wyman (1997).
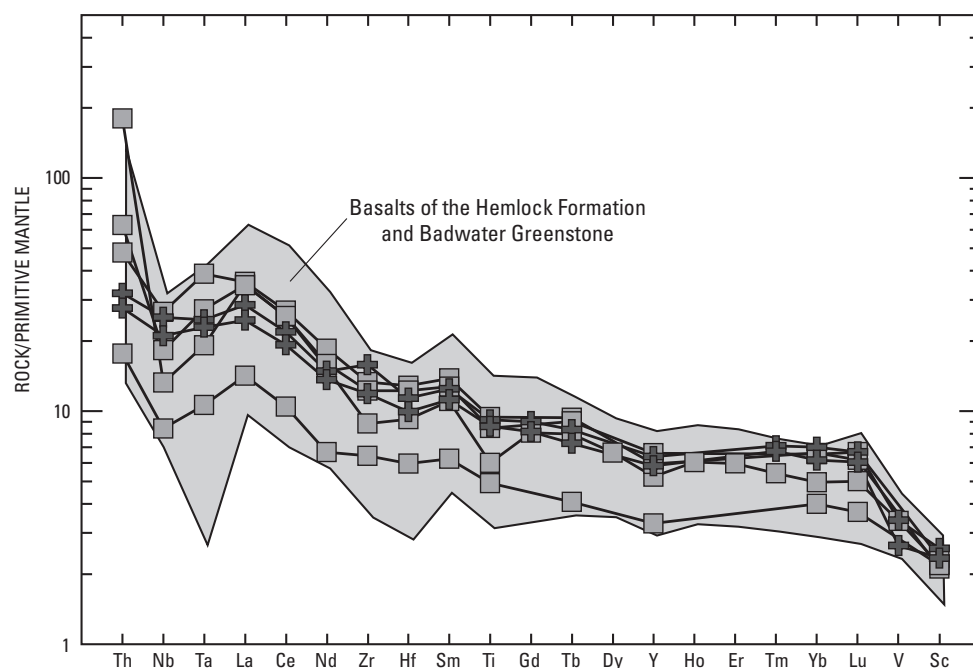
## Penokean Deformation

The intensity of Penokean folding and faulting varied substantially from the central part of the Gogebic iron range to both the eastern and western parts. In the central part of the range, folds are absent except for small folds near faults. Faults formed in at least three sets. Early, bedding-parallel faults were well documented in mine workings, but are no longer accessible for study. The best developed of these faults was traced for the entire length of the productive part of the iron range. The amount of displacement was not well determined but was probably small, in the range of hundreds of meters. These faults were most logically north-directed thrust faults, but kinematic indicators from the fault zones have not been described. Many more such faults may occur in other parts of the central part of the range but have not been detected. Slatey

cleavage that formed locally in the argillaceous beds of the Tyler Formation (especially in higher parts of the section) suggests that a significant northward transport may have occurred. Two sets of steep faults have offset the bedding-parallel faults. Their contribution to Penokean deformation is not completely resolved; at least some of the faults were reactivated during Mesoproterozoic time or later.

In the eastern part of the range, the effects of Penokean deformation appear abruptly at the Sunday Lake fault near Wakefield, Mich. This reverse fault resulted in about 2 km of offset of the bedrock units and was truncated by the unconformity at the base of the Mesoproterozoic strata, indicating that it is clearly a Penokean structure. East of the Sunday Lake fault, Penokean folds and faults are widespread. Folds plunge to the east or west when restored to their Paleoproterozoic orientation. Faults were of sufficient magnitude to produce
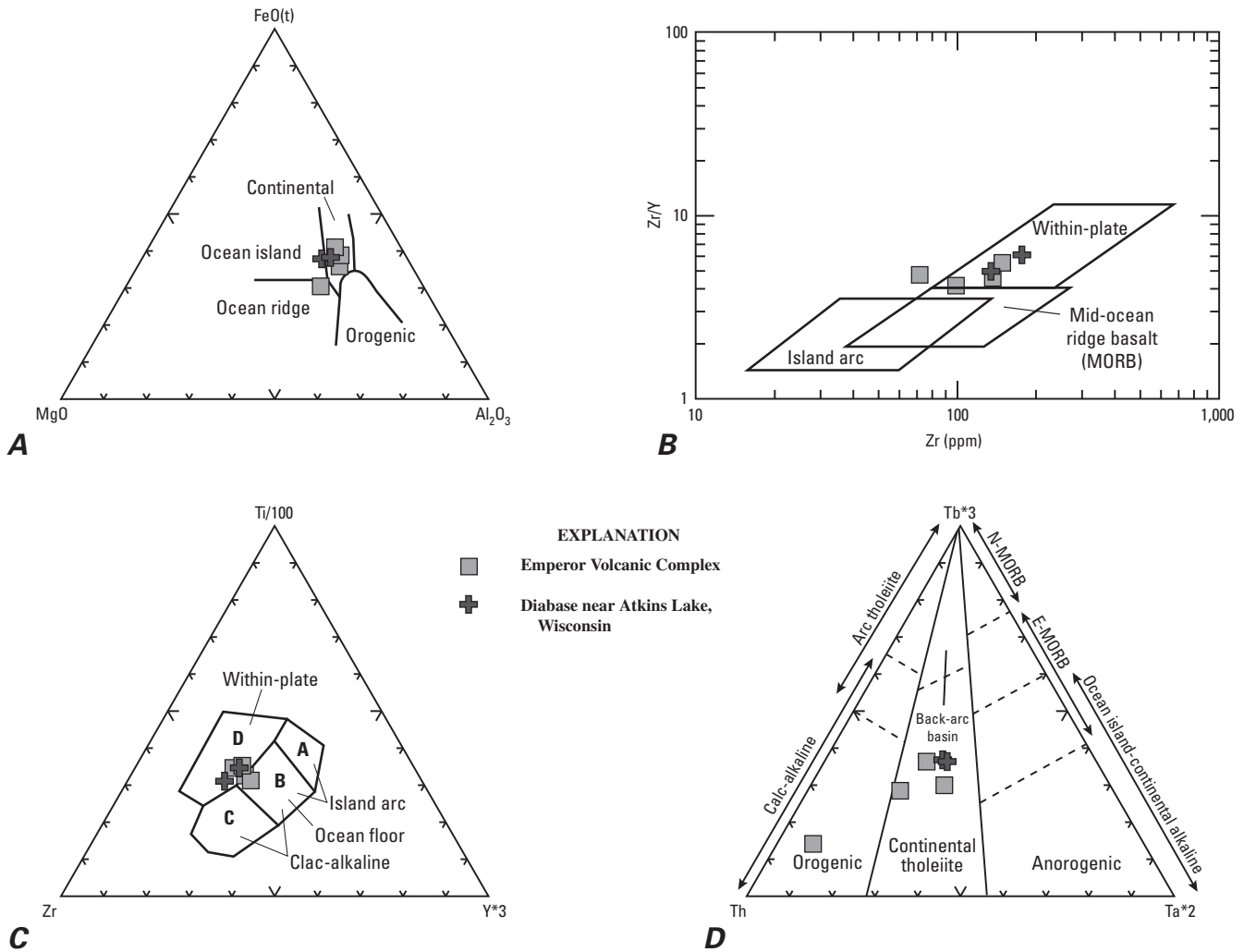


**Figure 16.**    Tectonic discrimination diagrams showing the within-plate affinity of the mafic rocks of the Emperor Volcanic Complex and diabase from near Atkins Lake. (*A*) from Pearce and other (1977). (*B*) from Pearce and Norry (1979), (*C*) from Pearce and Cann (1973), and (*D*) from Cabanixs and Thiéblemont (1988).

structural repetitions within the iron-bearing sequence. The relations here show that there was a sharp transition from very weakly deformed rocks of the central part of the range to much more intensely folded and faulted rocks of the eastern part of the range.

In the western part of the range, a similar sharp transition occurs between the nonfolded and only moderately faulted rocks as far west as the Tyler Forks River and the tightly folded and thrust-faulted rocks farther west. The structural pattern, as shown in map view, is deceptively simple west of the Tyler Forks River. Mesoproterozoic block rotation of the Penokean fold-and-thrust belt resulted in a map view that approximates a cross section drawn parallel to fold axes. This geometry inherently does not indicate the true geometry of the folds. When the map pattern is visualized without the Meso-proterozoic rotation, it reveals tight folds and a significant repetition of rocks by thrust faults. As in the east, restoring structures to their Paleoproterozoic orientations indicates that fold axes plunge gently to the east or west. Thrust faults are largely parallel to bedding, although numerous ramps also were mapped.

The Paleoproterozoic rocks of the Gogebic iron range illustrate significant facies changes across a boundary that represents a change from relatively stable deformational conditions (the central part of the range) to more tectonically active conditions (eastern and western parts of the range). This same boundary was also a deformational front during Penokean contractional deformation, which suggests that Penokean deformation was controlled in part by older structures that formed during crustal extension or earlier.

Finally, Paleoproterozoic and Neoarchean rocks were tilted northward as rigid blocks within the upper plate of a crustal-scale listric thrust fault at about 1,100 Ma (Cannon and others, 1993). Thus, the entire Gogebic iron range, in present map view, represents a longitudinal cross section of the Penokean fold belt, which provides a unique view of Paleoproterozoic sedimentation and deformation within the Penokean orogen.
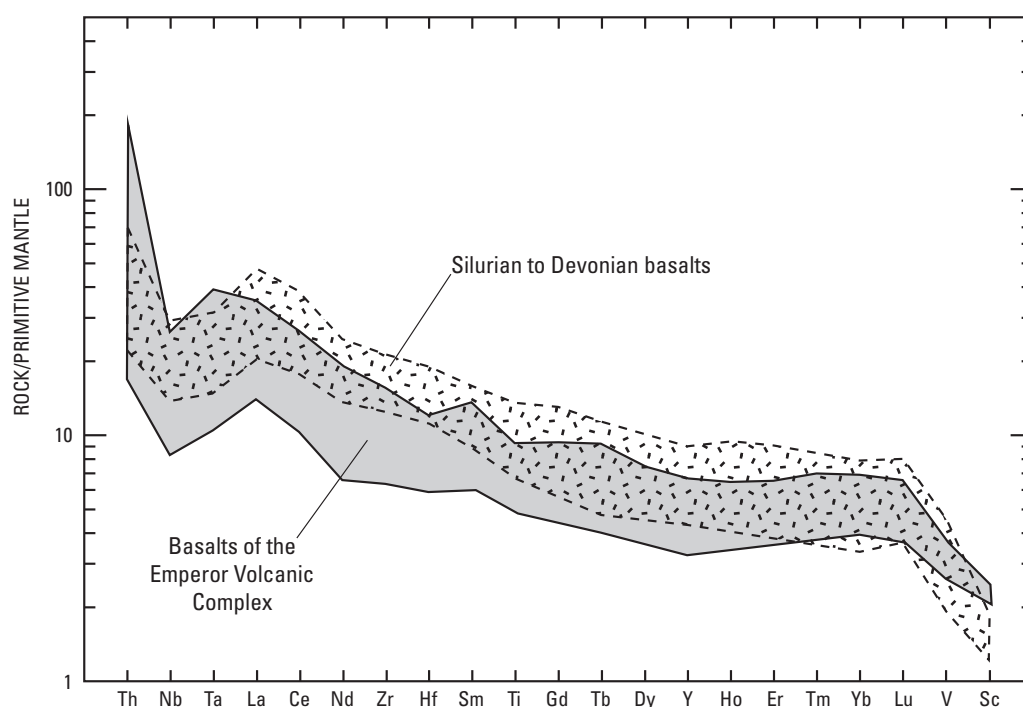


**Figure 17.** Primitive-mantle-normalized multi-element plot showing the compositional similarity of rift-related Late Silurian to Early Devonian basalts from the northern Appalachians (symbols) and the mafic rocks from the Emperor Volcanic Complex and diabase near Atkins Lake (shaded field). Data for Late Silurian to Early Devonian basalts from Dostal and others (1989, 1993). Primitive-mantle normalizing values from Kerrich and Wyman (1997).

# Acknowledgments

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