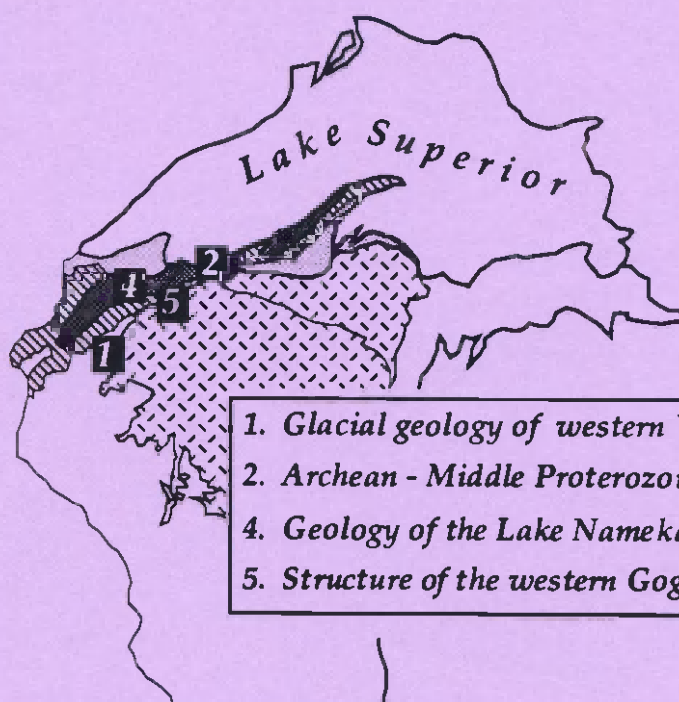




INSTITUTE ON LAKE SUPERIOR GEOLOGY

CABLE, WISCONSIN

MAY 15 - 19, 1996



1. *Glacial geology of western Wisconsin*
2. *Archean - Middle Proterozoic transect*
4. *Geology of the Lake Namekagon region*
5. *Structure of the western Gogebic Range*

PROCEEDINGS VOLUME 42

PART 3 - FIELD TRIP GUIDEBOOK



INSTITUTE ON LAKE SUPERIOR GEOLOGY

42ND ANNUAL MEETING

MAY 15 - 19, 1996

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VOLUME 42

PART 3 -- FIELD TRIP GUIDEBOOK

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FIELD TRIP #1: GLACIAL GEOLOGY OF WESTERN WISCONSIN

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FIELD TRIP #2: GEOLOGY OF THE MONTREAL RIVER MONOCLINE:
A TRAVERSE THROUGH 25 KM OF THE CRUST

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FIELD TRIP #4: EARLY TO MIDDLE PROTEROZOIC GEOLOGY OF THE
LAKE NAMEKAGON REGION

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FIELD TRIP #5: LAKE NAMEKAGON AND PENOKEE GAP AREAS, WEST
GOGBIC RANGE, WISCONSIN

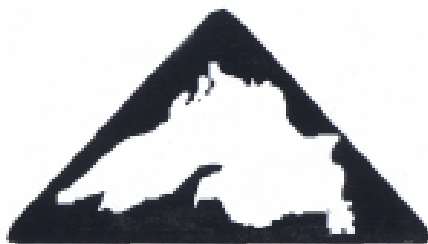
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GUIDE FOR FIELD TRIP # 1

GLACIAL GEOLOGY OF WESTERN WISCONSIN

FIELD TRIP #1

GLACIAL GEOLOGY OF WESTERN WISCONSIN

by

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I. Introduction

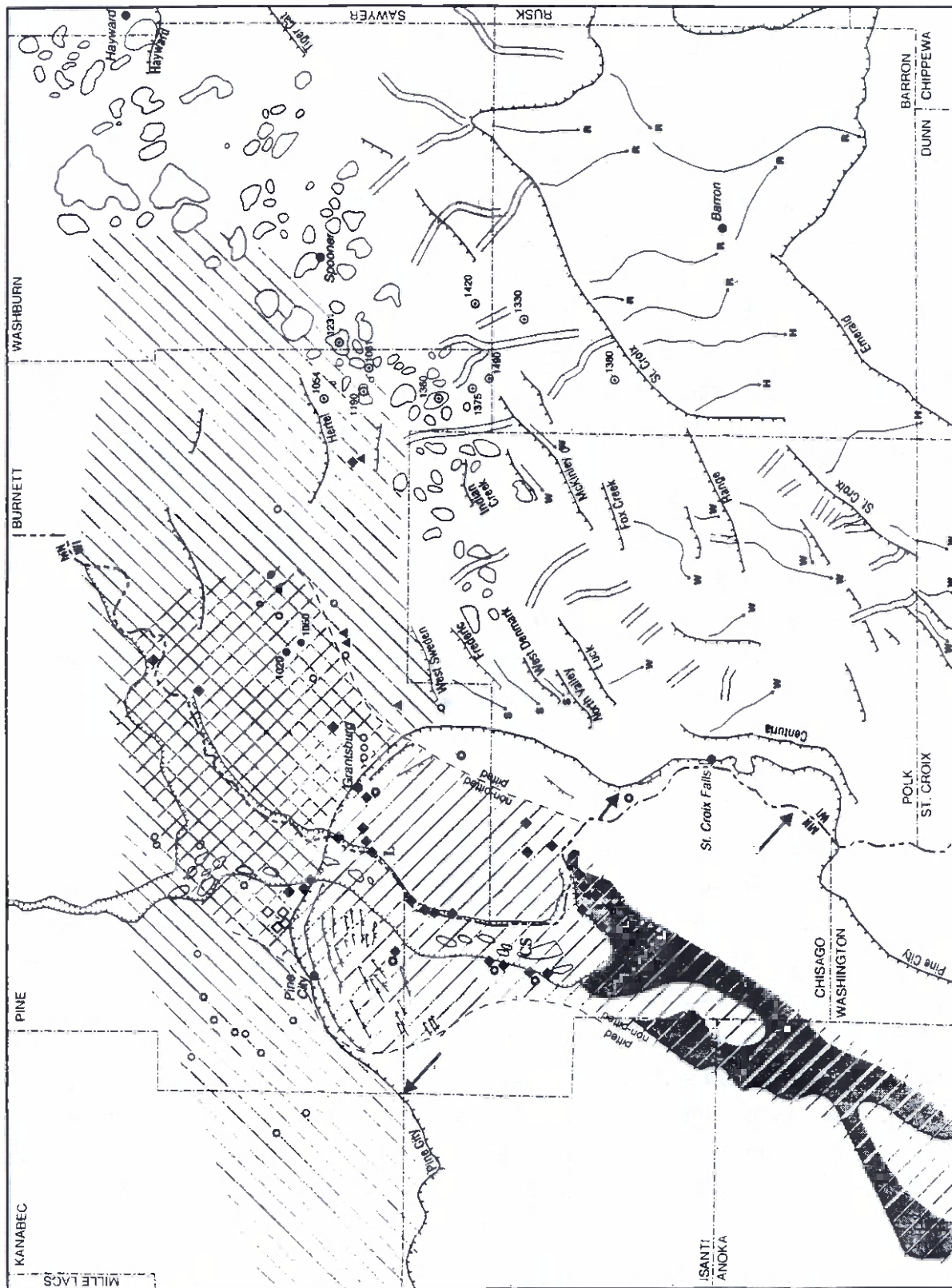
The purpose of this ILSG field trip is to get a general overview of the landforms, sediments, and geologic history of the Pleistocene features of northwest Wisconsin. To that end, the field trip has been designed to show exposures and outcrops that reveal the different types of deposits and landforms throughout the region. Though not bound in this booklet, I will hand out and refer to a publication called "Glaciation of Wisconsin" by Clayton and others (1991). Figures 1 through 5 are the figures from the "Glaciation of Wisconsin" brochure; the figures included within this report start with Figure 6.

Certainly, one of the goals of a regional understanding of glacial deposits is to be able to predict, or at least expect, surface-material variations that in part determine the character of human interaction with the landscape (groundwater resources, geologic engineering character of sediment, etc.). Additionally, glacial sediments and landforms provide the primary record for the study of the nature and history of former ice sheets. Studies of these features, along with the use of modern glacial environments as analogs, provide a foundation upon which one can reconstruct glacial history.

II. Ice lobes, phases, stratigraphy, and brief glacial history of western Wisconsin

Northwestern Wisconsin was glaciated several times during the Pleistocene, but the topography and surficial sediment are associated primarily with advances of the Superior Lobe and Grantsburg Sublobe (an offshoot of the Des Moines Lobe) during the last part of the Wisconsinan Glaciation (Fig. 1-4).

The earliest known glacial advance was during the Reeve Phase of the Des Moines Lobe, which occurred probably over 730,000 years ago and deposited gray, loam till of the Pierce Formation (Fig. 1-3, 1-5). This material is deeply buried throughout the region and occurs only in patches; some surface exposures occur in southern St. Croix County and Pierce County. The till is deeply weathered, even



- distribution of braked and brown
 varved clay exposed beneath
 Helgesen and Lindholm (1977)
 extent of glacial Lake Lind
 extent of glacial Lake Grantsburg
 ice-flow direction of Grantsburg
 Sublobe inferred from 14 fabric
 low-relief Grantsburg Sublobe
 moraine
 Kettle River and St. Croix channel
 with bars on the Chongestane
 surface
 tunnel channel
 boundary between pitted and non-
 pitted topography
 dominant meltwater drainage route
 Red Cedar River
 Hay River
 Willow River
 flows to the southwest
 ice-margin position
 county boundary
 Spooner Hills
 outcrop of glacial Lake Grantsburg
 varved clay
 outcrop of glacial Lake Grantsburg
 varved clay exposed beneath
 Trade River Bl.
 outcrop of glacial Lake Lind
 varved clay
 outcrop of glacial Lake Lind
 near-shore sed.
 location and elevation of glacial
 Lake Grantsburg wave-
 washed gravel
 selected spot elevations
 outcrop of silver Superior Lobe
 town



0 5 10 km

Figure 1-6.-- Portion of northwestern Wisconsin and adjoining Minnesota showing features associated with the most-recent glacial events. Spot elevations are included to show that elevations increase from central Burnett County to the St. Croix ice margin. Based on their relative position, the Centuria, Luck, Fox Creek, and McKinley ice margins are correlated with the Tiger Cat ice margin of Clayton (1984). Tiger Cat and Hayward ice margins from Clayton (1984) and Attig and others (1985). Emerald, St. Croix, Early Chippewa, and Late Chippewa ice margins from Johnson (1986). Pine City ice margin in Minnesota from Cooper (1935) and Hobbs and Goebel (1982). Location of glacial Lake Grantsburg and glacial Lake Lind varved clay in Minnesota from Chris Hemstad and myself (unpublished data) and Gary Meyer (personal communication). Only tunnel channels in Wisconsin are shown. Extent of the glacial Lake Grantsburg in Minnesota is in part from Cooper (1935). CS (in eastern Minnesota) = location of Camp Sunrise. L = townsite of Lind, Wisconsin. Till fabric measurements in Minnesota from Chernicoff (1983).

where overlain by younger material.

The Pierce till is overlain in places by reddish-brown, sandy loam till and outwash of the River Falls Formation, which was deposited during the Baldwin Phase of the Superior Lobe and the Dallas Phase of the Chippewa Lobe probably over 130,000 years ago (Fig. 1-3, 1-5). This material is weathered, as shown by clay-mineral alterations and magnetite weathering (Johnson, 1986) and thick soil profiles (Baker and others, 1983). I interpret this weathering to have likely occurred during the Sangamon Interglacial interval.

The Superior Lobe and Chippewa advances that occurred during the last part of the Wisconsin Glaciation began with the Emerald Phase and early Chippewa Phase (Fig. 1-1, 1-6). Deposits from these phases occur at the surface underneath the gently rolling topography of the southern Barron County, southeastern Polk County, and central St. Croix County.

Part-way through the late Wisconsin Glaciation, the Superior Lobe changed behavior and began to form a different suite of landforms and sediment types. These are associated with more than 10 former ice-margin positions in western Wisconsin (Fig. 1-6, 1-7), beginning with the St. Croix Phase. During these events, the Superior and Chippewa Lobes left behind reddish-brown sandy loam till and outwash of the Copper Falls Formation, as well as striations, grooves, outwash plains, outwash fans, tunnel channels, eskers, hummocks, ice-walled-lake plains, ice-dammed-lake plains, end moraines, and the Spooner Hills of northeastern Polk County, southeastern Burnett County, and central Washburn County. The Superior Lobe ice-margin positions (Fig. 1-6) can be traced for several kilometers laterally and represent ice-margin stillstands, ice-margin advances, or sedimentologic events that imprint the location of the ice margin (for example, the tunnel-valley/outwash-fan pairs clearly indicate the position of the ice margin but do not require an ice-margin stillstand or readvance).

As the Superior Lobe retreated, glacial Lake Lind formed in the preglacial St. Croix River valley (Fig. 1-7A) and lasted about 1000 years based on the number of varved silt and clay couplets found in lake sediment (Sunrise Member of the Copper Falls Formation).

The Grantsburg Sublobe, an offshoot of the Des Moines Lobe (Fig. 1-7C, 1-7D), advanced during the Pine City Phase across a fluvial plain that developed on top of glacial Lake Lind sediments (Fig. 1-7B). The Grantsburg Sublobe deposited gray, loam till of the Trade River Formation and left behind end moraines, till plains, ice-marginal channels, and ice-dammed-lake plains. Glacial Lake Grantsburg was

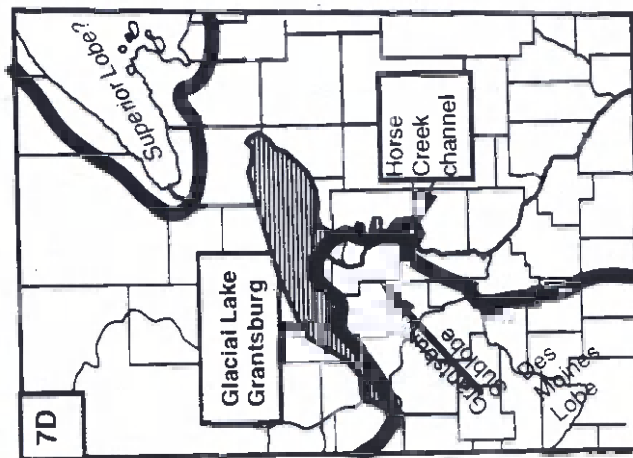
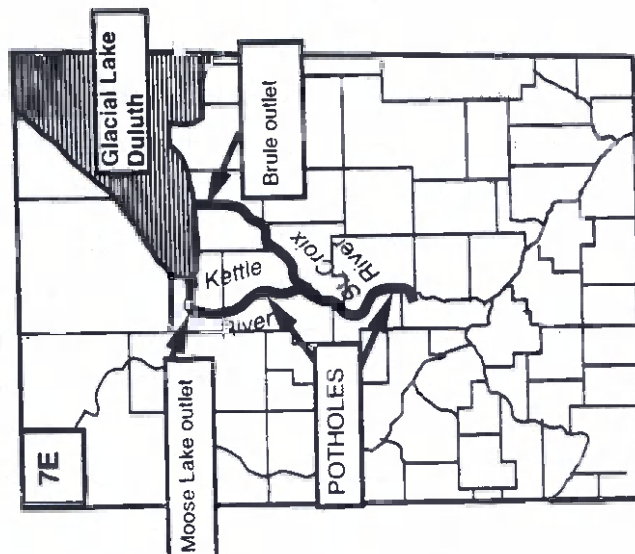
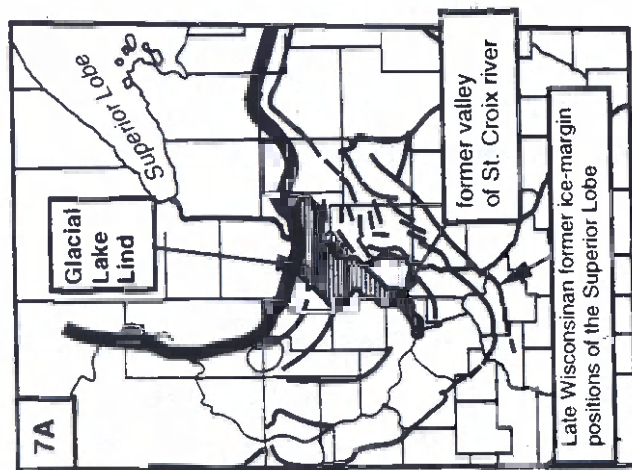
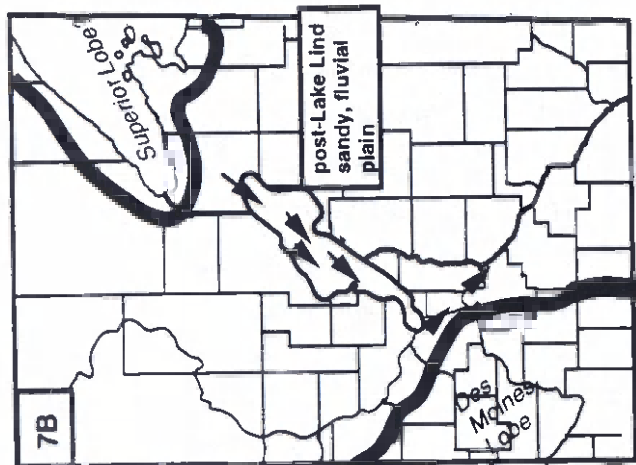
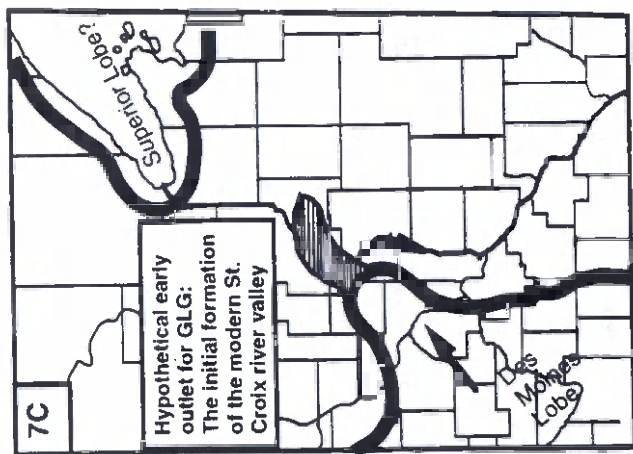


Figure 1-7.—Five sketch maps of the field-trip area showing selected Late Wisconsin events. 7A—Glacial Lake Lind formed in the lowland of the former St. Croix River. The Superior Lobe at this time is shown by the heaviest line; the medium lines mark former Late Wisconsin ice-margin positions of the Superior Lobe. 7B—Glacial Lake Lind filled in with sand deposited in prograding deltas and fluvial plains. 7C—The Grantsburg Sublobe advanced across the sand plain, dammed the St. Croix River, and formed glacial Lake Grantsburg. The initial outlet of glacial Lake Grantsburg would have formed the modern St. Croix River valley in the Taylor's Falls area. 7D—The Grantsburg Sublobe continued to advance to the Pine City margin and the glacial Lake Grantsburg expanded. The Superior Lobe may have still been in the Duluth drainage basin at this time. 7E—Glacial Lake Duluth drained through the study area along the St. Croix River.

dammed in front of the advancing ice and lasted approximately 100 years, based on the number of varved silt and clay couplets of the Falun Member of the Trade River Formation (see varve localities in Fig. 1-6). The location of the lake's outlet is unknown.

Numerous drainage events from ice-dammed Lake Superior (glacial Lake Duluth, Fig. 1-7E) caused the St. Croix River to incise its valley and created the famous potholes in the basalt around St. Croix Falls. An interruption in these drainage events allowed a large sand surface to develop in northwestern Polk County and in Burnett County. Dunes later developed on this surface.

III. Ages of late Wisconsin glacial events

The date of the late Wisconsin glacial events of northwestern Wisconsin is uncertain. A radiocarbon date of 20,500 years B.P. $\pm$ 400 (I-5443) (Wright, 1972; Wright and others, 1973) was obtained from disseminated organic matter in lake sediments at Wolf Creek in central Minnesota. This lake is behind the St. Croix moraine and would not have been ice-free until the ice margin had retreated 60 km from the maximum limit, possibly post-dating the St. Croix Phase maximum by hundreds of years. However, there is a possibility of contamination of the sediment by old carbon, and the date may not be reliable. Clayton and Moran (1982), Mickelson and others (1983), and Johnson (1986) suggest a somewhat younger date for the St. Croix Phase, approximately 18,000 to 15,000 years B.P. Their chronologies are based on dates from wood in other regions, where there is less chance of contamination. However, little wood is associated with the deposits of the St. Croix Phase. The Emerald Phase (the pre-St. Croix Phase, Late-Wisconsin advance in western Wisconsin) is considered by Johnson (1986; in press) to have occurred around 20,000 to 25,000 B.P.

The dates of the phases that occurred after the St. Croix Phase are unclear with the exception of the Marquette Advance, which is dated around 9,900 B.P. (Clayton, 1984).

IV. Landforms of western Wisconsin

The landscape created during overall retreat of the Superior Lobe in western Wisconsin consists predominantly of hummock tracts, ice-walled-lake plains, pitted-outwash plains, regions of collapsed outwash, and rolling to streamlined till uplands. Tunnel valleys, eskers, and the Spooner Hills are also present. Pleistocene deposits are generally 30 to 45 m thick (up to 85 m in places). Precambrian basalt

crops out as striated and grooved knobs in the western and northern parts of Polk County, Barron quartzite in eastern Barron County, and Cambrian sandstones in southern Barron County. Drainage of meltwater during retreat was generally to the south during deglaciation (see arrows, Fig. 1-6).

In the following section, I discuss the major landform types in the region, and describe what each landform type signifies.

Moraines.--There are very few moraines, in the classic sense, in western Wisconsin. The best-developed are along the Pine City Phase margin of the Grantsburg Sublobe, but only in places (Fig. 1-6). Few morainic ridges occur at Superior Lobe ice-margin positions. The term 'morainic ridge' is used for relatively narrow features deposited at the margin of the ice. They contain little hummocky topography, and are composed of various types of till and ice-contact sand and gravel. Much of the length of the "St. Croix moraine" is not characterized by morainic ridges, but consist of hummock tracts, outwash plains, and glacial thrust systems.

Hummocks and hummock tracts (Stop 1-3).--I use the term 'hummock' to mean a hill or knob, irregular to circular to oval in plan, occurring in groups, and associated with irregular to circular to oval hollows, depressions, or wetlands. This term is used regardless of the type of sediment composing the hummock. A variety of terms has been applied to this type of landform in glaciated regions, including dead-ice moraine, hummocky moraine, stagnation moraine, collapsed outwash, kame, and kame complex. I use the term 'hummock tract' to describe a distinct region of hummocks (in the same way that the term 'drumlin field' describes a group of drumlins).

Individual hummocks within the field-trip area are 5 to 20 m high and 25 to 400 m in diameter with a mean height of 10 m and width of 130 m. Hummocks are irregularly spaced with respect to each other but show a frequency of one hummock per 250 to 500 meters. There are 15 to 40 hummocks/km². Hummock slope ranges from 2 to 14 degrees.

Hummock tracts (Fig. 1-8) lie in a zone parallel to and up to 20 km behind the St. Croix and other Superior Lobe ice-margin positions; hummocks are absent along certain lengths of former ice-margin positions (Fig. 1-8). The hummock tracts do not form a continuous band but are separated from each other by outwash plains, regions of collapsed-outwash, tunnel valleys, and rolling, upland till surfaces. Hummock tracts in the field-trip area commonly form upland areas that have elevations above adjacent outwash surfaces. Outwash surfaces that lie adjacent to the hummock tracts are graded to nearby ice-margin positions a few kilometers

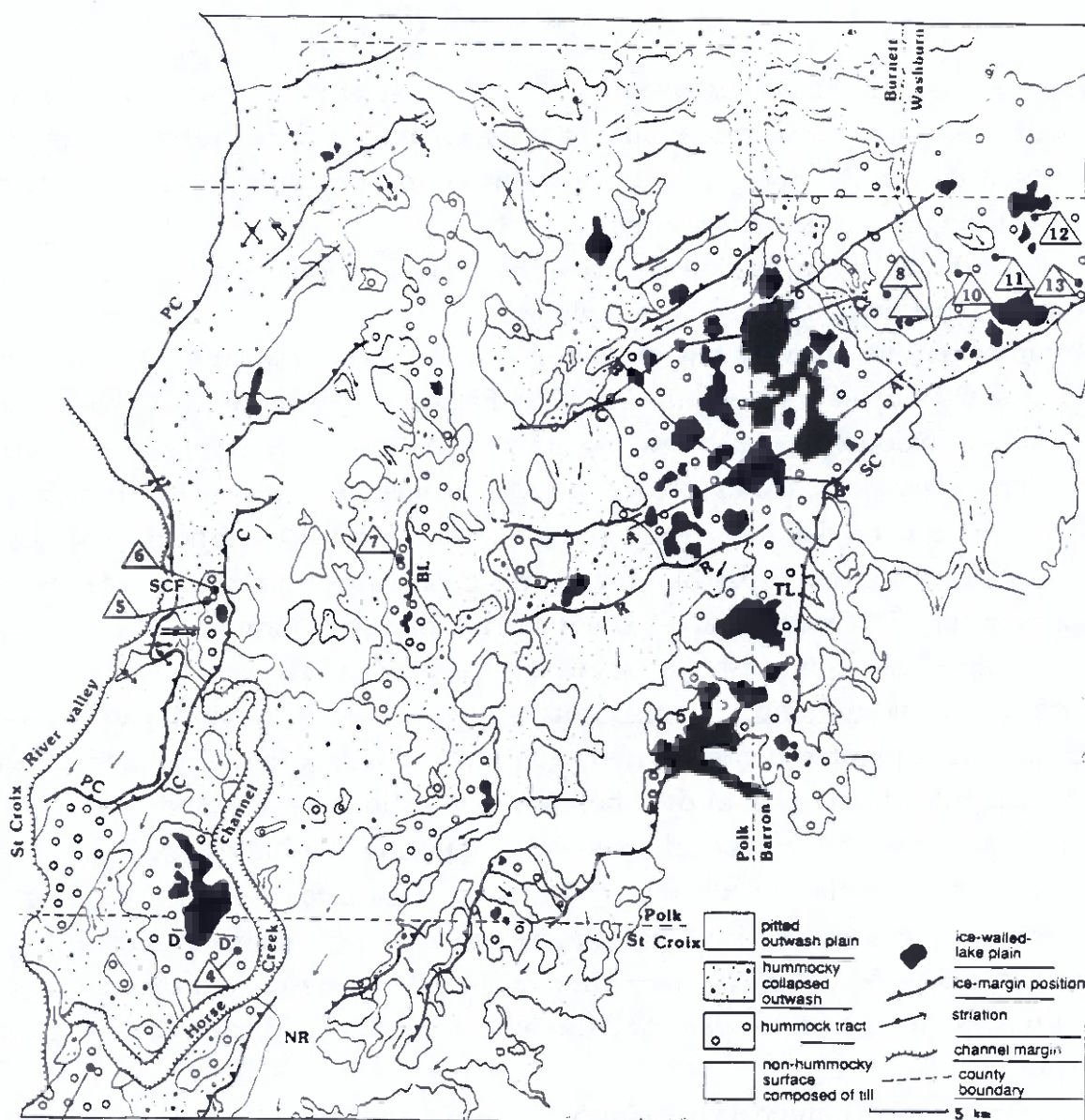


Figure 1-8.--A part of western Wisconsin showing the field-trip area (black in inset) and geomorphic features. Areas left blank on the map include predominantly rolling to dendritically-drained surfaces underlain by till (with a few basalt knobs in the western and northwestern parts of the study area). NR=city of New Richmond, SCF=city of St. Croix Falls, TL=city of Turtle Lake, C=Centuria ice-margin position, BL=Balsam lake ice-margin position, R=Range ice-margin position, PC=Pine City ice-margin position, SC=late St. Croix ice-margin position, GS=Grantsburg Sublobe. Dashed line in inset is trace of the Superior Lobe ice-margin limit where covered during the Pine City Phase of the Grantsburg Sublobe; arrows in inset show general ice-flow direction.

further up ice to the north or west. In several locations, the outwash plains located between hummock tracts are collapsed into tunnel valleys that had been filled with thick, stagnant ice. In part, the tunnel valleys may have controlled the location of outwash plains by localizing meltwater drainage in supraglacial troughs that formed as ice flowed into the quiescent tunnels. Additionally, tunnel-valley formation may have affected location of hummock tracts and development of hummocks by locally removing basal and englacial debris from overlying glacial ice as the tunnel formed.

The origin of hummocks has been interpreted in many different ways. I have found at least twelve different models:

- 1- supraglacial reworking and redistribution of debris released supraglacially from stagnant ice (Boulton, 1967, 1972; Clayton, 1967; Clayton and Moran, 1974; Paul, 1983).
- 2- slumping of supraglacial debris into crevasses (Tanner, 1914);
- 3- squeezing of subglacial debris into basal crevasses or hollows under stagnant ice (Hoppe, 1952; Stalker, 1960).
- 4- molding of basal till under active ice as a transition form between drumlins and Rogen moraine (Aario, 1977).
- 5- patchy deposition of lodgement till under active ice (Menzies, 1982);
- 6- melting of simple or stacked layers of debris-rich ice producing melt-out till hummocks (Möller, 1987).
- 7- deformation of supraglacial and englacial debris by rising diapirs of clean ice (Minell, 1979).
- 8- rising diapirs of subglacial debris (Zelcs, 1993).
- 9- patchy formation of ground ice beneath stagnant, cold-based glaciers (Aario, 1992).
- 10-melting of debris-rich, stagnant ice containing extensive karst tunnels (Kemmis and others, 1994).
- 11-partial erosion of till plains by subglacial meltwater (Rains and others, 1993).
- 12-Extrusion of proglacial sediment by glacier-forced groundwater flow (Boulton and Caban, 1995).

The hummocks in the field-trip area are considered to have formed as stagnant-ice features (#'s 1 and 6) because of their close association with ice-walled-lake plains. The hummocks formed from the melting of stagnant, debris-rich ice. The resulting topography formed as a result of collapse and differential settling as the melting, buried, debris-rich ice lowered the overlying supraglacial debris. Numerous exposures show the presence of faults, folds, and steeply tilted beds

within the hummocks. These features are best explained as being caused by collapse due to the melting of buried, stagnant ice.

However, much of the till in the hummocks is interpreted to be melt-out till. The interpretation of melt-out till in many of the hummocks is based primarily on the strong pebble fabric parallel to regional ice flow (Fig. 1-9). I interpret the diamicton with strong fabric as likely being melt-out till (rather than lodgement till or deforming-bed sediment) for three reasons. (1) The diamicton contains faults and subtle stratification that have been related in the literature to melt-out till. The thin sand lenses present in the diamicton at the St. Croix Falls dump and speedway sites are similar to lenses described in diamicton by Möller (1987). Möller uses these lenses as arguing for a melt-out origin. (2) In over half of the samples with ice-flow-parallel fabric, the fabric is stronger than those reported for lodgement till. (3) Melt-out till requires stagnant ice for formation, and stagnant ice is indicated by the hummocks themselves. Flow till does occur in some exposures, but it is not as common as the till I interpret as melt-out till. Flow till has poor fabric development and is interbedded with fluvial sediment.

In terms of glaciologic implications, hummocks have been used to imply surging conditions leaving large hummock tracts (Wright, 1980; Johnson and Savina, 1987) or that the hummocks indicate extensive freezing-on of material under non-temperate glaciers (Boulton, 1972; Paul, 1983; Möller, 1987; Sollid and Sørbel, 1988; Attig and others, 1989; Mooers, 1990) and therefore indicate cold conditions.

The ice-walled-lake plains and melt-out-till hummocks of western Wisconsin represent the coarse-grained end of a continuum with the stagnant ice landscape developed in clayey till that was described in North Dakota by Clayton (1967). In general, I suggest that sandier glacial debris tends to yield a greater proportion of melt-out till in hummocks, and more clay-rich glacial debris tends to produce a greater proportion of flow till and debris-flow sediment in hummocks.

Ice-walled-lake plains (Stop 1-4).--There are over 40 ice-walled-lake plains in the field-trip area (Fig. 1-8), and nearly all of them occur in hummock tracts, suggesting that their origin is related to the origin of the hummocks. Ice-walled-lake plains form initially as supraglacial lakes on stagnant, debris-rich ice. The lake water melts through the thin ice, and sediment is deposited in the lake from adjacent debris-rich ice. Upon melting of stagnant ice, the sediment left behind forms a smooth, dish-shaped plateau in the midst of a hummock tract.

Ice-walled-lake plains in the field-trip area range in size from 1 to 13 km²,

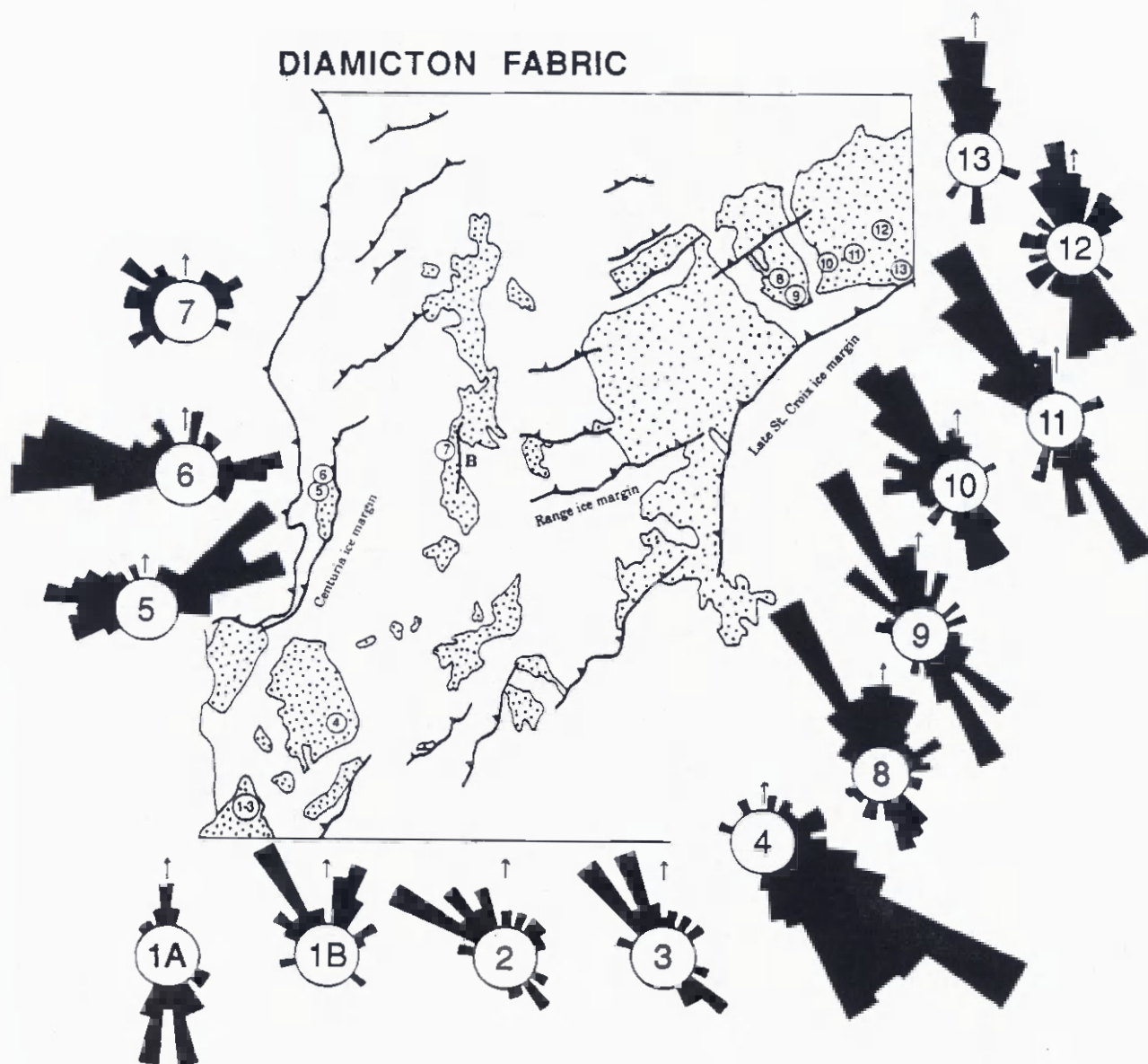


Figure 1-9.--Map of the study area showing rose diagrams of fabric in diamicton. Hachured lines = ice-margin positions (B = Balsam Lake ice-margin position); stippled areas = hummock tracts; circles = site locations; arrows = north. Fabric measurements consist of 25 pebbles except site 4 (75 pebbles combined from three places in the outcrop); sites 5 and 6 (50 pebbles combined from two places in each outcrop); sites 8, 9, 10, 11, and 12 (50 pebbles). Diamicton at site 4 contains steeply dipping stratification showing that fabric has been rotated. Rotation of stratification to 0° rotates dominant fabric mode to $N70^{\circ}W$.

with a typical size around 2 km². Ice-walled-lake plains commonly have a rim around their margin that stands 3 to 10 m above the center of the lake plain. Stream- and wave-sorted sand and gravel occur along the margins of the ice-walled lakes. In places, the rim sediment is interbedded with till that flowed from adjacent ice into the lake. Coarse sediment is restricted to the margins of the lake plain; much of the interior portions are underlain by silt, with as much as 20 m underlying the larger plains.

The sediment in the ice-walled-lake plains was deposited after active ice left the area. This is known because some ice-walled-lake plains cross known ice-margin positions, signifying that the ice that walled the lake came from two separate advances.

Tunnel valleys and eskers (Stops 1-1, 1-2, 1-6, and 1-8).--Tunnel valleys are river valleys believed to have been carved by subglacial streams that became incised into the glacier bed. These valleys were first described in Denmark and later recognized in northern Germany. Numerous subglacial river channels associated with late Wisconsinan Glaciation in the Upper Midwest have been described by Mickelson and others (1983) and Attig and others (1989) as being formed by rapid or catastrophic drainage. Therefore, these authors have proposed the term tunnel channel for these features, suggesting that their formation occurred under bank-full conditions.

Tunnel valleys are a common landform in western Wisconsin, the greatest number is associated with the St. Croix Phase ice margin. Half of the tunnel valleys in western Wisconsin contain eskers (Fig. 1-6).

In western Wisconsin, it is possible to link together tunnel valley segments into a network with some valley systems as long as 40 km. However, breaks in the tunnel-valley sequences, development of outwash plains of different ages graded to different parts of the same tunnel-valley sequence, and the presence of the best-developed tunnel valleys next to former ice-margin positions, suggest that the tunnel valleys in western Wisconsin were formed as relatively short segments near former ice-margin positions during overall retreat.

About half of the tunnel valleys in western Wisconsin also contain central eskers. The association of eskers within the tunnel channels indicates two levels of discharge during subglacial drainage. A large discharge is shown by the large size of the tunnel channel as well as the fact that it is an erosional feature. It is suggested that the tunnel channel represents an initial, short-lived burst of subglacial meltwater, perhaps under hydraulic head built up behind a frozen margin (Wright, 1973; Attig and others, 1989). As discharge abates, the subglacial meltwater pathway

is maintained at a smaller size. This smaller tube later becomes filled with stream sediment and becomes an esker upon melting of the surrounding ice.

Eskers that do not occur in tunnel channels are also present in western Wisconsin. The largest occurs in northwest Polk County and has its terminus at the apex of the large outwash fan upon which Frederic lies. The elevation of this esker rises from 1080 feet near West Sweden to 1200 feet at its terminus, indicating that the sediment in this esker was deposited as water flowed up regional slope. Other prominent eskers include one, locally referred to as "The Hogsback," at the St. Croix County line, and one revealed in a large collapse pit 3 km north of Centuria.

Outwash plains (Stops 1-1 and 1-2).--Outwash plains cover much of western Wisconsin and constitute one of the most common landforms remaining from the Wisconsin Glaciation (Fig. 1-6). These plains were built by meltwater streams emanating from the melting Superior Lobe.

Outwash plains are clearly associated with former ice-margin positions, and the position of former Superior Lobe ice margins are interpreted in part by the sharp northern and western up-slope limits of the outwash plains (Fig. 1-5). Near former ice-margin positions, the slope of the outwash plains is relatively steep ranging from 2.0 to 4.0 m/km, but may be locally as high as 8.5 m/km. In more distal positions, the slope is around 1.0 m/km.

A significant amount if not most of the outwash in the outwash plains was derived from tunnels underneath the Superior Lobe discharging meltwater and sediment. In several places in the county, the outwash plain has a fan shape at former ice-margin positions whose apex is coincident with the mouths of a tunnel valley or an esker. It is clear in these cases that the tunnel system underneath the ice, now indicated by the tunnel valleys and eskers, was the source for the sediment that composes the outwash plain.

Spooner Hills (Stop 1-10).--The Spooner Hills refer to a line of hills of high relief that contain unlithified sediment (Fig. 1-7). These hills are distinct from any other landform in the field trip area and represent a landform type not commonly described in glaciated regions. I refer to these hills as the Spooner Hills after the village of Spooner, Wisconsin, which lies in the center of the band of hills.

The Spooner Hills range in area from 1 to 5 km² and exhibit relief as high as 60 m. The Spooner Hills are roughly equidimensional, but a number of them are elongate to the southeast parallel to the direction of regional ice flow, suggesting that they may have been somewhat streamlined. Hill-top elevations in the Spooner Hills generally increase from the northwest to the southeast (Fig. 1-7).

Based on this information, the following hypothesis is offered as an explanation for the formation of the Spooner Hills. Because the hills contain a variety of sediment types representing a variety of depositional environments, and because the sediment may predate the most recent glaciation, it is likely that the Spooner Hills are erosional features. Several lines of evidence suggest that the erosion occurred subglacially. The Spooner Hills are commonly covered with the till of the most recent glaciation and are somewhat streamlined in a direction parallel to the regional ice-flow direction of the most recent glaciation. In addition, the orientation of the band of the Spooner Hills parallels the St. Croix ice margin suggesting that they are related. The inter-hill valleys form a branching network that, in some places, connects up with prominent tunnel valleys. It is suggested that these valleys were excavated by subglacial meltwater and the tunnel valleys represent the outlets for meltwater and sediment. However, this connection cannot be shown convincingly along the entire Spooner Hills chain. The water and sediment was released at the glacier margin and the sediment was deposited proglacially in large outwash fans and outwash plains. In this scenario, the Spooner Hills would have formed when the ice was at the St. Croix ice margin or the Tiger Cat-McKinley ice margin (Fig. 1-6).

Two published descriptions of landforms elsewhere in the Midwest similar to the Spooner Hills support the suggestion that subglacial meltwater erosion may have created them. Clayton (1986) describes till ridges in Portage County, Wisconsin, that have similar characteristics to the Spooner Hills. He suggests that they may have been cut by subglacial meltwater. In the Puget Lowland, Washington, hills occur that have a size and shape similar to the Spooner Hills (Booth and Hallet, 1993). The hills in the Puget Lowland occur among channels that have been interpreted as cut by subglacial meltwater.

V. Types of till discussed on this field trip.

The interpretation of the glacial landscape requires that the sediment composing landforms be interpreted as well. The following list describes the basic characteristics of lodgement till, melt-out till, flow till, and deforming bed sediment. All of these sediment types may look similar (massive, poorly sorted, pebbly), but subtle differences allow interpretation in some places.

Lodgement till.--Lodgement till is deposited by plastering of glacial debris from the sliding base of a moving glacier by pressure melting and/or other mechanical processes (Dreimanis, 1989), and is characterized by:

1. Laterally and vertically consistent orientation of fabric elements, structures, and surface marks on clasts.
 - a. parallel striae on lodged clasts
 - b. parallel a-axes of pebbles
 - c. sediment smudges, thrust structures, shear planes, and fissility showing shear
2. Comminution producing bi-modal grain-size with terminal modes in silt size.
3. Materials of local derivation nearer the base of the till.
4. Overconsolidated.
5. Occurs in ground moraine, flutes, drumlins, proximal sides of end moraines.

Melt-out till.-- Melt-out till is deposited by a slow release of glacial debris from ice that is not sliding or deforming internally (Dreimanis, 1989), and is characterized by:

1. Debris banding from ice preserved but with gradational contacts and flattening of structures. Melt-out till may be massive.
2. Clasts of unlithified material may be present.
3. Clasts aligned parallel to glacier flow, but may be transverse. Dip of pebbles is up-ice but lessened by loss of ice.
4. Grain-size may be coarser than lodgement till, and grain-size distribution may be more variable.
5. Striated clasts present, but faceted clasts are rare.
6. Draping of layers over clasts as interstitial ice is lost.
7. Winnowing of fine particles and redeposition, perhaps as coats around clasts.
8. Lenses and veins of sorted sediment cased in till. Sorted sediment deposited in englacial tubes.
9. Normal consolidation.
10. Associated with stagnant ice landforms.

Flow till.--Flow tills (broadly speaking) may derive from any glacial debris upon its release from glacier ice or from a freshly deposited till, in direct association with glacier ice. The redeposition is accomplished by gravitational slope process, mainly by gravity flow, or by squeeze flow, and it may take place ice-marginally, supraglacially, or subglacially, and subaerially or subaquatically (Dreimanis, 1989), and is characterized by:

1. Lithology (color, grain-size, mineralogy) inherited from parent ice (some sorting during flow).
2. Common stratification and interbedding with water-sorted sediments.
3. Depending on grain size and water content, sorting, grading (inverse or normal) may occur.

4. Fabric determined by local stresses (local slopes) and stresses are low, producing random to strong fabric with maxima parallel to transverse stress orientation (not ice flow).

5. Normally consolidated.

Deforming bed sediment.--Deforming bed sediment forms under a sliding, wet-based glacier. The glacier's basal shear stress is transmitted to the bed, causing bed deformation. The thickness of this deforming layer is hotly debated among glaciologists and glacial geologists: is it centimeters or meters? Deforming bed sediment ostensibly has the same characteristics as lodgement till, although some authors (i.e., Hart, 1994) suggest that deforming bed sediment has a weak pebble fabric.

VI. Contrasts in glacial landforms

One of themes of this field trip is to point out and discuss the variations in morphology of glacial deposits in the terminus region of the Superior Lobe.

<u>Emerald Phase</u>	<u>St. Croix Phase and later phases</u>
Few hummocks	Many hummocks and ice-walled-lake plains behind multiple ice-margin positions
Few collapse depressions	Many collapse depressions
No tunnel valleys or eskers	Many tunnel valleys, some with eskers, as well as the Spooner Hills
No mappable outwash plains	Extensive outwash plains
Integrated, dendritic drainage	Deranged drainage
	Thrust features (?)
No drumlins(?)	No drumlins

The landforms left behind during the Emerald Phase of the Superior Lobe are distinct from those formed during the St. Croix and later phases. Features formed by widespread stagnant ice (hummocks and ice-walled-lake plains) and extensive discharge of subglacial meltwater (tunnel valleys, eskers, and outwash plains) dominate the landscape left by the St. Croix Phase, but are absent or rare in the landscape left by the Emerald Phase. I interpret these phases to have occurred during the last part of the Wisconsinan Glaciation and that the difference in the landform suites is due to a change in character of the Superior Lobe at the beginning of the St. Croix Phase. I suggest that the Superior Lobe changed from a cold, non-

surging glacier during the Emerald Phase, to a less cold, surging glacier during the St. Croix Phase.

The landforms of the Emerald Phase (listed above), the thin to patchy cover of the till deposited during the Emerald Phase, and the presence of well-developed permafrost features suggest that climate during the Emerald Phase was colder than during the St. Croix Phase. The absence of tunnel valleys indicates less subglacial meltwater present during this phase, perhaps because cold conditions prevented meltwater from draining through the glacier to the bed. Meltwater formed predominantly supraglacially flowed, for the most part, over clean ice and had a low sediment load. Ice flowed into pre-Emerald-Phase river valleys, which later became collapse depressions when warming allowed buried ice to melt. Following retreat of the ice margin, permafrost conditions helped to allow rapid development of a dendritic drainage network and likely much surface till was eroded. In part, the drainage network may have been inherited from the pre-Emerald-Phase surface because it would not have been deeply buried by Poskin till. (Though I suggest that these features indicate cold conditions, they may also imply that the Emerald Phase did not last long.)

During the St. Croix Phase, large-scale subglacial drainage of the Superior Lobe began and formed tunnel valleys and eskers. As mentioned earlier, much of the meltwater stream sediment in the outwash plains of western Wisconsin can be traced to tunnel-valley mouths. Additionally, the Spooner Hills are thought to be the result of erosion by subglacial meltwater. The increase in subglacial meltwater features is accompanied by the presence of widespread stagnant ice features.

Johnson and Savina (1987) suggest that the Superior Lobe surged during the St. Croix and later phases, and Clayton and others (1985) suggest that many lobes of the Laurentide Ice Sheet surged. Increased subglacial discharge of meltwater is associated with surging (Kamb and others, 1985), and surging can leave large amounts of stagnant ice that melts to form hummocks (Wright, 1980). Clapperton (1975) suggests that large amounts of debris are frozen on to the base of the glacier during surging. A thick, basal, debris-rich layer in a mass of stagnant ice would produce high-relief hummocks like those in Polk County. Surging is a periodic behavior of glaciers and the size and geometry of the tunnel valleys indicate that they operated only for a short time, and that they were not all operating at the same time (Attig and other, 1989; Patterson, personal communication).

On the other hand, the hummocks may simply indicate cold conditions rather than surging. Many authors suggest that hummocks are produced by non-

temperate glaciers (glaciers with ice colder than 0°C) (Boulton, 1972; Paul, 1983; Möller, 1987; Sollid and Sørbel, 1988; Attig and others, 1989; Mooers, 1990). These authors describe the process of meltwater generation where the bed is at 0°C and the subsequent freezing-on of debris at the base of the glacier as the meltwater moves subglacially towards the margin and encounters ice colder than 0°C. This process produces a thick, basal, debris-rich zone. The margin of the cold glacier stagnates during melting and the melting of the debris-rich ice produces hummocky topography. Evidence of permafrost in western Wisconsin suggests that the hummocks may simply indicate a non-temperate glacier, and not necessarily a surging glacier.

It is not clear in western Wisconsin if the water that cut the tunnel valleys was melted from the bed or from the surface of the glacier. Wright (1973) states that tunnel valleys formed under the Superior Lobe because subglacial meltwater was held by a frozen margin of the glacier. Mooers (1989) argues that the Superior Lobe was temperate and that the tunnel valleys indicate meltwater from the surface of the glacier reaching the glacier bed. Gustavson and Boothroyd (1987) document a surficial source for subglacial meltwater on the temperate Malaspina Glacier, Alaska. The presence of ice-wedge casts behind the St. Croix Phase ice-margin limit in Wisconsin suggests that the ice surface would have been well below 0°C and that little surface meltwater would make it to the glacier bed. However, cold conditions were present during the Emerald Phase when no or few tunnel valleys were formed.

I suggest that the climate supported permafrost conditions during both the Emerald and St. Croix Phases, but was gradually getting warmer. Ice-wedge casts outside the Emerald Phase ice-margin limit are numerous and well-developed (Johnson, 1986), whereas those behind the St. Croix Phase ice-margin limit are few and less well developed. During the St. Croix Phase, even though permafrost conditions were present, surface meltwater began to find pathways to the glacier bed. The presence of increased water at the bed caused the lobe to surge, tunnel valleys to form, and left widespread stagnant ice.

VII. Bibliographic information on the glacial geology of western Wisconsin

The St. Croix moraine was named by Berkey (1897) for a north-south band of hummocks occurring immediately east of the town of St. Croix Falls, Wisconsin. However, the first comprehensive mapping and description of the St. Croix moraine and associated deposits was that of Strong (1880) and T.C. Chamberlin

(1883). Leverett (1932) transferred the name to apply to the hummocky deposits that marked the edge of the Superior Lobe in western Wisconsin. Wright and Ruhe (1965) called the event that made these features the St. Croix Phase.

Along the St. Croix Phase deposits in Wisconsin, a number of geologists have made several somewhat dissimilar interpretations of the position and nature of the St. Croix ice-margin position (Chamberlin, 1905, 1910; Leverett, 1932; Mathieson, 1940; Black, 1959; Berg, 1960; Wright, 1972; Wright and others, 1973; Johnson, 1986; Johnson and Savina, 1987; Mooers, 1989). Refinements in the stratigraphy have been made by Baker and others (1983), Mickelson and others (1984), Johnson (1986; in press), and Attig and others (1988). The history of the glacier-dammed lakes (glacial Lake Lind and glacial Lake Grantsburg) occupying the St. Croix, Clam, and Yellow River drainage basins has been discussed by several workers (Berkey, 1905; Hansell, 1930; Burkhead, 1931; Butz, 1931; Cooper, 1935; Johnson, 1992, 1994; Addis and others, 1996). A regional understanding of the details of the last glacial period has been enhanced by large-scale geologic mapping of Pleistocene deposits in the Lake Superior region (Clayton, 1984), the Twin Cities region (Meyer, 1985), Barron County, Wisconsin (Johnson, 1986), Hennepin County, Minnesota (Meyer and Hobbs, 1989), Dakota County, Minnesota (Hobbs and others, 1990), Washington County, Minnesota (Meyer and others, 1990), Ramsey County, Minnesota (Patterson, 1992), and Polk County, Wisconsin (Johnson, in press).

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IX. Field trip sites.

Mileage from Cable to Stop 1-1: 60 miles

Stop 1-1 and Stop 1-2

St. Croix Phase outwash plain, distal site and St. Croix Phase outwash fan, proximal, ice-marginal site

Quadrangle: Haugen Quadrangle, Wisconsin, 7.5 minute series, sec. 18, T. 36 N., R. 11 W., sec. 13, T. 36 N., R. 12 W.

Significance: These stops show the character of sediment deposited by braided, outwash streams, both immediately next to a former ice-margin position, as well as further away.

Description: Superior Lobe outwash is probably the most common sediment facies western Wisconsin and consists primarily of sand, slightly gravelly sand, and gravelly sand, with much of the gravel in the pebble fraction. Stream sediment deposited within 1 km of former ice-margin positions is commonly much coarser and contains cobbly gravel with occasional boulders. This coarser facies of Copper Falls stream sediment is crudely horizontally bedded and is more poorly sorted than stream sediment deposited in more distal positions. Further away from former ice-margin positions, the stream sediment is dominated by sand in horizontal and cross-bedded layers.

Gravel in the outwash, as well as Superior Lobe till, is composed primarily of basalt, gabbro, granite, gneiss, rhyolite, and sandstone. This assemblage is typical of sediment derived from the Lake Superior basin.

The proximal site is located on an outwash fan associated with the tunnel valley in which Bear Lake lies. It is suggested that all the outwash in this outwash plain emanated from the tunnel-valley mouth at Haugen.

Questions: Are there any features present here that suggest rapid deposition? What would a catastrophic deposit look like?

Mileage from Stop 1-2 to Stop 1-3: 15 miles

Stop 1-3

Barron County hummock sites

Quadrangle: Lower Vermillion Lake Quadrangle, Wisconsin, 7.5-minute Series, T. 36 N., R. 13 W.

Significance: (1) Discussion of the occurrence of melt-out till in hummocks and (2) contrast between the landforms of the Emerald and St. Croix Phases.

Description: Uniform diamicton is exposed in five hummocks in northwestern Barron County (sites 8, 9, 10, 12, 13, Figs. 1-8, 1-9). In all cases, the exposures are small and only uniform till is revealed. The pebble fabric at these sites is strong ($S_1=0.72, 0.80, 0.74, 0.72, 0.85$) and similar to those reported by Lawson (1979) for melt-out till and by Dowdeswell and Sharp (1986) for melt-out till and undeformed lodgement till. Additionally, the major mode in each fabric is parallel to the regional ice-flow direction as determined by local striations and the orientation of the nearby ice-margin position (Fig. 1-9). At site 12 (Fig. 1-8), a drill hole revealed 20 m of diamicton with uniform grain size and uniform relative magnetic susceptibility beneath the base of the surface exposure. Though these sites were not well-enough exposed to investigate other structures and contacts, the strongly oriented pebble fabric parallel to the regional ice-flow direction indicates the diamicton is likely melt-out till.

As discussed in the text, features formed by widespread stagnant ice (hummocks and ice-walled-lake plains) and extensive discharge of subglacial meltwater (tunnel valleys, eskers, and outwash plains) dominant the landscape left by the St. Croix Phase, but are absent or rare in the landscape left by the Emerald Phase. I have interpreted this change to represent a change in ice dynamics, from a cold, non-surging glacier to a less-cold, surging glacier. See text for discussion.

Questions: Is it possible for supraglacial melt-out till to survive collapse? Is it merely chance that till fabrics in hummocks are strong and parallel to regional ice-flow direction? Do the changes seen from the Emerald to St. Croix Phases represent changes in ice dynamics? or merely differences in erosion (and thus age)? Is it possible for supraglacial meltwater to penetrate through the thickness of the glacier if permafrost conditions are present?

Mileage from Stop 1-3 to Stop 1-4: 16 miles

Stop 1-4

McKinley ice-walled-lake plain

Quadrangle: McKinley, Wisconsin, 7.5 minute series, sec. 31, T. 36 N., R. 14 W.

Significance: One of the largest ice-walled-lake plains in western Wisconsin.

Description: The McKinley ice-walled-lake plain is about 7 km² and is one of the largest ice-walled-lake plains in western Wisconsin (Fig. 1-8). About 40 ice-walled-lake plains occur in the terminus region of the Superior Lobe in western Wisconsin, and another 20 to 30 occur in the continuation of these deposits in Minnesota. Nearly all of them occur in hummocky tracts, suggesting that their origin is related

to the origin of the hummocks. Ice-walled-lake plains in the study area range in size from 1 to 13 km², with a typical size around 2 km². Ice-walled-lake plains commonly have a rim around their margin that stands 3 to 10 m above the center of the lake plain. Stream- and wave-sorted sand and gravel occur along the margins of the ice-walled lakes. In places, the rim sediment is interbedded with till that flowed from adjacent ice into the lake. Coarse sediment is restricted to the margins of the lake plain; much of the interior portions are underlain by silt. Drill holes at Stop 5 encountered 20 m of silt overlying Superior Lobe till. The ice-walled-lake plains occur at elevations similar to those of the surrounding hummocks. The rim ridges of the ice-walled-lake plains are approximately the same elevation or slightly lower than the tops of hummocks adjacent to the lake plain.

Ice-walled-lake sediment described elsewhere may occur in hummock tracts, but within the Southeastern Sector, apparently ice was thin enough that the open water was able to melt entirely through the stagnant ice, leaving nicely preserved ice-walled-lake plains.

Questions: What is the relationship of permafrost to the ice-walled-lake plains? Do the ice-walled-lake plains indicate that the hummocks must be stagnant ice features? What was the source of sediment for the ice-walled-lake plains? What do they indicate about sediment distribution in the ice?

Mileage from Stop 1-4 to Stop 1-5 (optional): 10 miles

Mileage from Stop 1-4 to Stop 1-6: 16 miles

Stop 1-5 (optional)

Turtle Lake ice-walled-lake plain delta

Quadrangle: Clayton, Wisconsin, 7.5 minute series, sec. 6, T. 33 N., R. 14 W.

Significance: An exposure of a delta that prograded into the Turtle Lake ice-walled-lake plain.

Mileage from Stop 1-5 to Stop 1-6: 23 miles

Stop 1-6

Eskedahl tunnel valley and esker

Quadrangle: Big Round Lake, Wisconsin, 7.5 minute series, sec. 28, T. 36 N., R. 16 W.

Significance: A good example of a tunnel valley with an esker

Description: The Eskedahl tunnel valley and esker is 25 km long and 0.5 to 0.8 km wide. The relief of the valley from the bottom of one of the lakes to the

surrounding plain is about 25 m; the top of the esker is almost everywhere several meters below the top of the surrounding plain.

Many tunnel valleys end in an outwash fan, but the Eskedahl tunnel valley and esker ends in Big Round Lake with no apparent outwash fan. Tunnel valleys that later become deeply buried by outwash tend to lose their straight shape and collapse into bulbous, elongate lakes, like Bone Lake or Big Round Lake. Very coarse outwash occurs at the head of the Fox Creek outwash plain, and it is likely that the tunnel valley fed the Fox Creek outwash plain.

Only a couple of road cuts occur in the esker here and little is revealed about the internal composition of this esker. One outcrop consists of predominantly sand.

Questions: Does the geometry of the tunnel valley and esker indicate the levels of discharge? Or is the valley cut by a meandering tunnel with the esker marking 'the last' position of the tunnel? What was the sources of water? Does this geometry require a frozen toe?

Mileage from Stop 1-6 to Stop 1-7: 13 miles

Stop 1-7

Erickson striations and grooves

Quadrangle: Milltown, Wisconsin, 7.5 minute series, SW1/4 SW1/4 sec. 11, T. 36 N., R. 18 W.

Significance: Striations and Nye channels cut into basalt that indicate two ice-flow directions and two different glacier-bed conditions.

Description: At this site, broad, shallow grooves are present that are oriented NE-SW. These are cut by younger striations that show ice-flow to the SE (Fig. 1-10). Numerous outcrops of basalt in this area show a similar pattern, with the grooves, interpreted to be Nye channels, consistently showing ice-flow to the SW. The grooves resemble those described by Walder and Hallet (1979) and are likely cut by subglacial meltwater. The striations within the grooves were cut by ice flowing into the grooves.

We postulate that the grooves form during surges when the Superior Lobe exhibited plug flow and flowed directly to the SW. The excess amount of subglacial water present during surging would have cut the grooves. When surging stopped, a normal ice profile developed and ice flowed laterally towards the margins (to the SE).

Questions: Can ice form these grooves or is water necessary? Is it possible that the grooves are pre-Late-Wisconsinan?

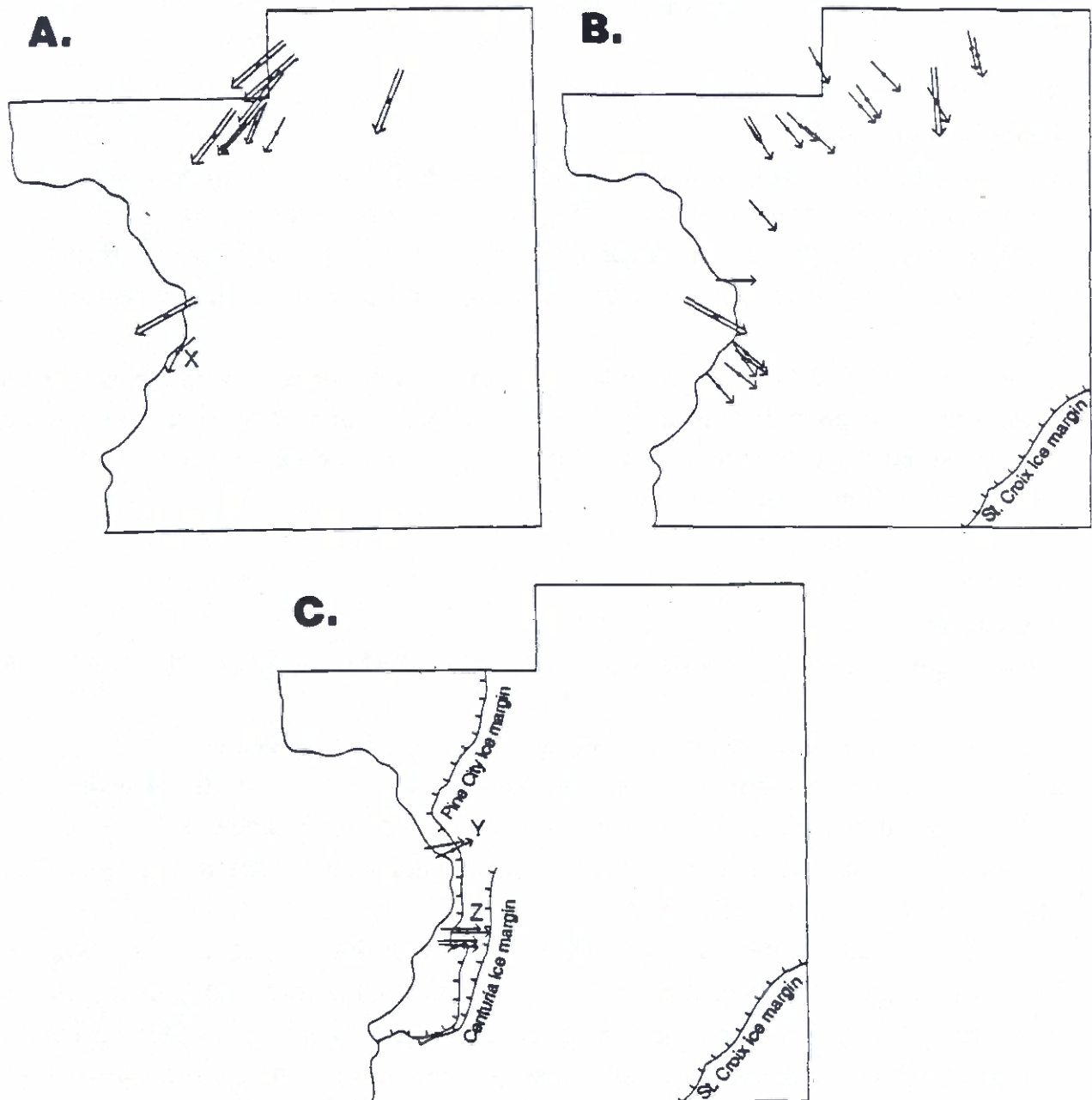


Figure 1-10.-- Striations and grooves in Polk County. Double arrows = grooves; single arrows = striations. A: Orientation of grooves and striations showing ice flow to the southwest. In most places, cross-cutting relationships were apparent and always show striations and grooves in this orientation to be older than those shown in B. Striations at X are striations and crag-and-tail striations atop Eagle Peak and Observation Rock in Interstate State Park. B: Prominent southeast striations with some grooves. C: Striations showing easterly flow and Pine City and Centuria ice-margin positions. The group at Z are interpreted to have been cut by the Superior Lobe but later than striations at the same site shown in B. See text for explanation. The group at Y was interpreted by Chamberlin (1905) to have been formed by the Grantsburg Sublobe.

Mileage from Stop 1-7 to Stop 1-8: 9 miles

Stop 1-8

Frederic esker and fan

Quadrangle: Frederic, Wisconsin, 7.5 minute series, sec. 21, T. 37 N., R. 17 W.

Significance: A prominent esker feeding an ice-marginal outwash fan.

Description: This pit contains a complicated sedimentology with large variations in grain size. Collapse features, antiformal bedding, and possible injection features are present.

Questions: If this esker is surrounded by collapse hummocks, how is the esker form preserved? Is it possible that the antiformal bedding is primary? What does this esker indicate about the nature of the Superior Lobe? What do eskers mean?

Mileage from Stop 1-8 to Stop 1-9: 20 miles

Stop 1-9

Searles varve site

Quadrangle: Hertel, Wisconsin, 7.5 minute series, SW1/4 NE1/4 sec 15, T. 38 N., R. 15 W.

Significance: Proglacial lake sediments and melting-out of stagnant ice

Description: This sandpit contains a red, sandy Superior Lobe till that is overlain by a coarsening upward sequence of varved lake sediment. This sequence is overlain by fluvial pebbly sand that is part of an outwash plain that grades to the Hertel ice-margin position 3 km to the north.

The Superior Lobe till, occurring at the base of the pit and in a large knob in the center of the pit, is finer than Superior Lobe till further south. It is interpreted to represent till deposited during a readvance of the Superior Lobe during overall retreat at which time the proglacial lake sediment was incorporated into the ice. Several outcrops of finer Superior Lobe till occur throughout the region (triangles in Fig. 1-6) and suggest that this till may be associated with an ice advance that reached the northernmost part of Polk County.

As ice retreated from an unnamed ice margin 3 km to the south (Fig. 1-6), a proglacial lake fronted the ice. The pit contains a wonderful coarsening upward sequence of varved lake sediment. It is not clear if this lake sediment belongs to glacial Lake Lind, glacial Lake Grantsburg, or some other ice-marginal lake. If it is glacial Lake Grantsburg, the sediment contains no sediment derived from the Grantsburg Sublobe.

The varved section contains 47 winter clay layers 2 to 10 mm thick. The summer layers thicken and coarsen upward, from silty layers less than 1 cm thick, to fine-sand, ripple-bedded layers up to 40 cm thick. The varved section is overlain by horizontally laminated sand with laminations of segregated heavy minerals (beach sand?), which is in turn overlain by pebbly sand, interpreted to be fluvial sand.

The fluvial sand underlies an outwash plain that grades to the Hertel ice-margin position (Fig. 1-6). This means that roughly 50 years passed from the uncovering of this spot to the filling in of the lake by sand derived from the nearby Superior Lobe. This suggests an ice retreat rate of 80 m/yr.

The lake sediment is draped over the till knob, but faults and steeply dipping beds show that a large part of the relationship is due to collapse, presumably by melting of buried ice. An angular unconformity in the varve sequence indicates that collapse occurred progressively while the lake was in existence. The cranberry bogs to the south of the pit occur in a larger collapse pit that indicates melting of buried ice continued after the lake was filled in.

A possible ice-wedge cast was found at this site in 1990 in the fluvial gravel.

Questions: What is the role of lakes in hastening the melting of buried ice? There is a lot of detail in this pit, what else can we learn from this sequence?

Mileage from Stop 1-9 to Stop 1-10: 10 miles

Stop 1-10

The Spooner Hills

Quadrangle: Timberland, Wisconsin, 7.5 minute series, sec. 4, T. 37 N., R. 14 W.

Significance: A stop in the midst of the Spooner Hills, suggested to have been cut by subglacial meltwater.

Description: As discussed in the text, the Spooner Hills is a name given to this band of hills that lie 15 to 25 km behind the St. Croix Phase ice-margin position. Some key observations include:

- the hills have an area from 1 to 5 km²,
- relief as high as 60 m,
- a number of them are elongate to the southeast parallel to the direction of regional ice flow,
- hill-top elevations generally increase from the northwest to the southeast,
- inter-hill valleys tend to align with tunnel-valleys in places,
- the hills are composed of a variety of unlithified materials.

The following hypothesis is offered as an explanation for the formation of the

Spooner Hills. Because the hills contain a variety of sediment types representing a variety of depositional environments, and because the sediment may predate the most recent glaciation, it is likely that the Spooner Hills are erosional features. Several lines of evidence suggest that the erosion occurred subglacially. The Spooner Hills are commonly covered with the till of the most recent glaciation and are somewhat streamlined in a direction parallel to the regional ice-flow direction of the most recent glaciation. In addition, the orientation of the band of the Spooner Hills parallels the St. Croix ice margin suggesting that they are related. The inter-hill valleys form a branching network that, in some places, connects up with prominent tunnel valleys. It is suggested that these valleys were excavated by subglacial meltwater and the tunnel valleys represent the outlets for meltwater and sediment. However, this connection cannot be shown convincingly along the entire Spooner Hills chain. The water and sediment was released at the glacier margin and the sediment was deposited proglacially in large outwash fans and outwash plains. In this scenario, the Spooner Hills would have formed when the ice was at the St. Croix ice margin or the Tiger Cat-McKinley ice margin (Fig. 1-6).

We suggest that this type of landform may be more common than presently recognized in glacial landscapes.

Questions: What are the critical questions to ask to ascertain the origin of these hills? What is the source of the tremendous amount of water needed to cut these hills? What are other possible mechanisms for their origin?

Mileage from Stop 1-10 to Cable: 45 miles



**OVERVIEW OF THE BEDROCK GEOLOGY
OF THE CABLE - HURLEY AREA, WISCONSIN**

OVERVIEW OF THE BEDROCK GEOLOGY OF THE CABLE-HURLEY AREA, WISCONSIN

by

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The area encompassing Field Trips 2, 4, and 5 is underlain by bedrock ranging in age from Late Archean to Middle Proterozoic. Rocks record three principal cycles of deposition and deformation: 1) submarine volcanism and sedimentation in Late Archean time, followed by major deformation and granite emplacement; 2) epicontinental marine sedimentation in Early Proterozoic time, followed by weak to moderate faulting and folding during the Penokean Orogeny; and 3) extension, basaltic volcanism and related intrusions, and continental sedimentation in the Midcontinent Rift System in Middle Proterozoic time, followed by inversion and thrust faulting. Figure 1 shows the general geology of the region, and Figure 2 indicates the approximate location of stops for trips 2, 4, and 5.

GEOLOGY OF ARCHEAN ROCKS

Volcanic and sedimentary rocks of late Archean age are exposed in a westward-tapering wedge from Hurley to near Mellen, Wisconsin, forming the basement directly beneath Early Proterozoic rocks. The volcanic rocks grade southward and westward into gneiss and migmatite in a border zone of the Puritan Batholith, a large granitic intrusion emplaced at about 2750 Ma (Sims and others, 1977). West of Mellen, volcanic rocks are unknown and Early Proterozoic rocks rest directly on gneiss and granite. Archean rocks are not a focus of the field trips but will be seen at one stop on Trip 2 (Stop 2-10).

Details of Archean geology are imperfectly known because of very sparse exposures. The most complete exposures are in the areas underlain by volcanic rocks between Hurley and Upson. These volcanic rocks consist mostly of basalt, andesite, and rhyolite; dacite breccia (Stop 2-10) is the most common rock type directly beneath the Middle Proterozoic rocks between Hurley and Upson (Greathead, 1975). The volcanic rocks form a section at least 6 km thick, and are overturned so that they face northwestward and dip steeply to moderately southeastward.

GEOLOGY OF EARLY PROTEROZOIC ROCKS

Early Proterozoic rocks include strata of the Marquette Range Supergroup along the Gogebic Iron Range. These rocks are the focus of Field Trip 5, and will also be seen on parts of Trips 2 and 4. The continuity of Early Proterozoic strata across the area is broken by the Mineral Lake intrusion, which truncates the sequence from near Mineral Lake westward to the Marengo River, a distance of about 10 km.

Area east of Mineral Lake intrusion: East of Mineral Lake, the Early Proterozoic sequence displays the classical Gogebic Range stratigraphy with a basal argillite-quartzite unit (Palms Formation) overlain successively by banded iron-formation (Ironwood Iron-formation) and a shale-graywacke turbidite unit (Tyler Formation). From about Ballou Creek westward a thin, possibly discontinuous, unit of dolomite and chert breccia of the Bad River Dolomite lies between the Palms Formation and underlying Archean basement rocks.

The Early Proterozoic section is 2-3 km thick and records sedimentation successively on a stable craton, passive margin, and compressional foreland. The oldest strata, represented by the Bad River Dolomite, were deposited unconformably on Archean rocks, apparently in a stable cratonic setting. The Bad River is a cherty dolomite, commonly metamorphosed to tremolitic marble. It also includes a distinctive chert breccia (Stops 2-9 and 2-10) that constitutes most of the unit from near the Bad River eastward for about 10 km to near Ballou Creek. The breccia appears to be a variably reworked residuum of chert nodules and beds left by solution of the dolomite during the weathering interval, prior to deposition of the unconformably overlying Palms Formation.

The principal iron-bearing sequence is the Menominee Group, consisting of the Palms Formation and Ironwood Iron-formation. We interpret the Menominee Group to have been deposited on a south-facing passive margin. The Palms Formation (Stops 2-9 and 2-10) consists of banded argillite near the base, and grades upward into thick-bedded quartzite near the top. It lies unconformably on Archean rocks and the Bad River Dolomite. At its upper contact the Palms grades upward over a meter or two into ferruginous beds of the Ironwood Iron-formation. The Ironwood contains numerous lithologic types of iron-formation. East of Upson, Wisconsin, it consists of five laterally continuous members that are alternating units of thin, even-bedded, carbonate iron-formation and thick, wavy-bedded, hematitic iron-formation (Hotchkiss, 1919; Huber, 1959). West of Upson (Stops 2-8 and 5-B), these lithologies are also present, but laterally

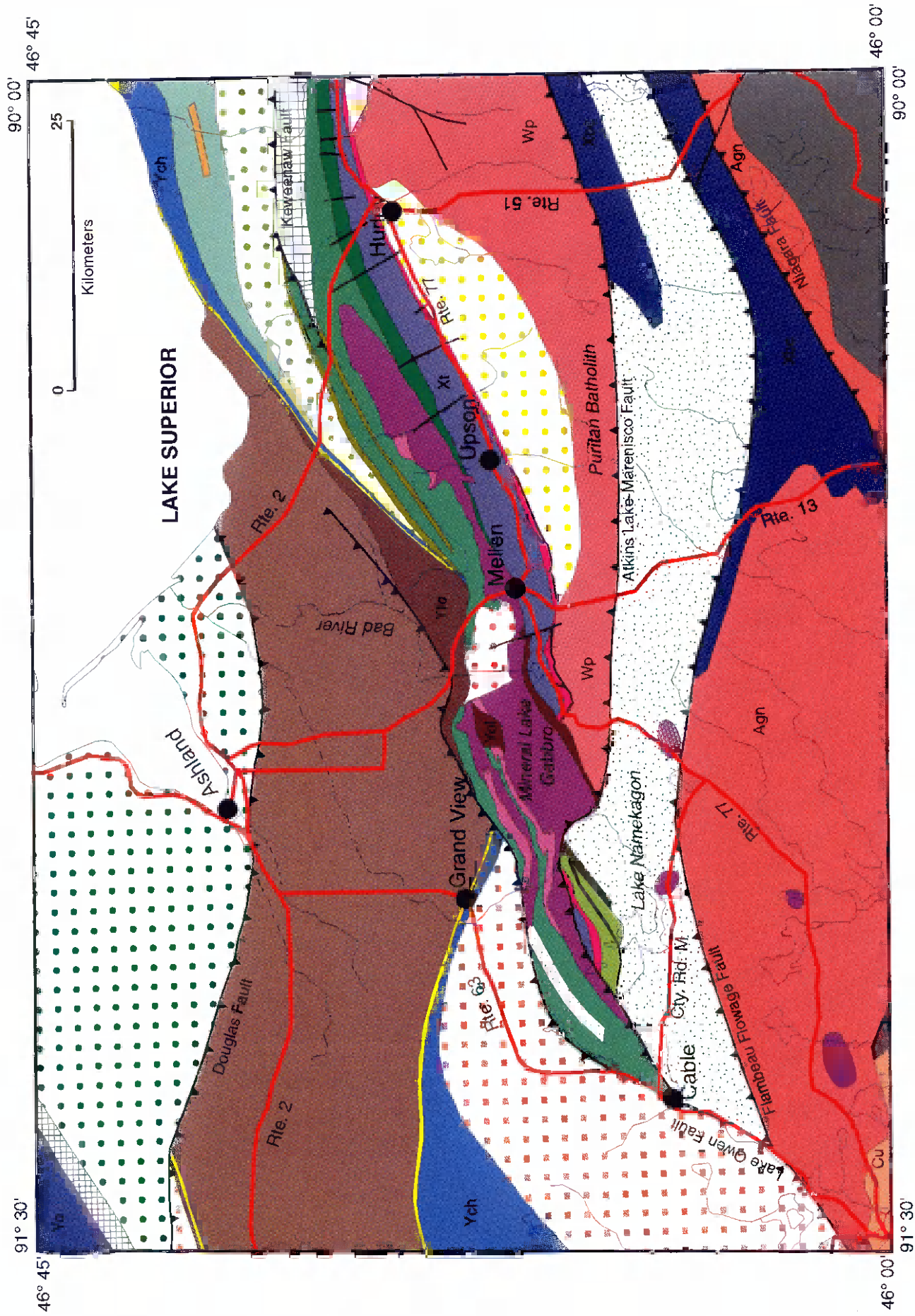


Figure 1: Geologic map of part of northern Wisconsin and Michigan showing region of field trips 2, 4, and 5.

EXPLANATION

	Cambrian sandstone		Porcupine Volcanics rhyolite		Early Proterozoic
	Middle Proterozoic		Porcupine Volcanics andesite and basalt		
	Jacobsville Sandstone		Portage Lake Volcanics		
	Chequamegon Sandstone		Kallander Creek Volcanics rhyolite		
	Devils Island Sandstone		Kallander Creek Volcanics conglomerate		
	Orienta Sandstone		Kallander Creek Volcanics andesite and basalt		
	Freda Sandstone		Granite		
	Freda Sandstone conglomerate facies		Granophyre		
	Nonesuch Formation		Anorthosite and gabbro		
	Copper Harbor Conglomerate		Olivine gabbro and serpentinite		
	Chengwatana Volcanics		Siemens Creek Volcanics		

continuous members are more difficult to define, in part because of primary sedimentary changes, but also because of complications induced by deformation and metamorphism.

The Tyler Formation (Stops 2-7 and 5-B) overlies the Ironwood and is a thick succession of graywacke and slate deposited in a foreland basin during accretion of island arc terranes to the south at about 1850 Ma. We believe that the basal contact with the Ironwood Iron-formation is most likely a low-angle unconformity, although the nature of the contact has been a point of controversy for many years.

Early Proterozoic rocks are essentially unmetamorphosed east of Upson, but metamorphic grade is progressively higher towards the west where assemblages include garnet and a variety of amphiboles, including abundant grunerite. Near the contact with the Mineral Lake intrusion, assemblages include fayalite, orthopyroxene, and garnet in iron-formation. The metamorphism was static and caused by Middle Proterozoic thermal events.

Early Proterozoic rocks dip steeply to vertical nearly everywhere along the Gogebic Iron Range. This attitude results mostly from Middle Proterozoic deformation, discussed below, but we have distinguished a component of relatively strong Early Proterozoic deformation as well. The Early Proterozoic deformation is part of the Penokean orogeny, caused by collision of a series of island arcs with the southern edge of the Superior craton at about 1850 Ma (Sims and others, 1989). The area lies north of the suture and spans the transition from a foreland fold and thrust belt to the essentially undeformed craton.

From Hurley to Upson, Penokean deformation is nearly nonexistent. A weak slaty cleavage is developed locally in the Tyler Formation but no folds are known. Westward from Upson, compressional deformation intensifies and is expressed both by thrust faults and folds (Stops 2-10 and 5-B). On Mt. Whittlesey, the Ironwood Iron-formation is structurally thickened both by folding and repetition by thrust faults. Farther west, near Mineral Lake, large, nearly isoclinal folds were developed. These folds originally were upright but now are recumbent as a result of Middle Proterozoic tilting.

From about Ballou Creek westward there appears to be a detachment fault at or very near the base of the Early Proterozoic section (Stop 2-10) so that Early Proterozoic rocks were decoupled from Archean rocks during Penokean deformation. The amount of structural shortening in the Early Proterozoic strata is unmatched in the Archean basement.



Figure 2. Part of the Ironwood 10x20 quadrangle showing the approximate locations of stops for field trips 2, 4, and 5. See individual guides for more detailed locations.

Area west of Mineral Lake intrusion: West of the Mineral Lake intrusion, Early Proterozoic strata are present from the Marengo River to the north side of Lake Namekagon. These strata share many similarities with their equivalents to the east but also display some important differences.

The Early Proterozoic succession includes an approximately 300-m thick section of Bad River Dolomite, exposed near the Marengo River south of Grand View, Wisconsin. It is typically a gray dolomitic marble with abundant lenses and beds of chert, partly replaced by tremolite, and lies unconformably on Archean rocks. This is the thickest section of Chocolay Group rocks preserved on the Gogebic Iron Range. The overlying Palms Formation is much like that known to the east, consisting of laminated argillite near the base and thick-bedded quartzite near the top.

The Ironwood Iron-formation (Stops 4-1 and 5-A) lacks wavy-bedded hematitic iron-formation that are common to the east. However, it does contain units of laminated cherty magnetite iron-formation. The Ironwood grades laterally to the west into non-cherty silicate iron-formation, magnetic and non-magnetic argillite, and black pyritic shale. Large sills of metadiabase (Stops 4-1 and 5-B) also occur in the iron-formation, in contrast to areas farther east where mafic intrusions are much less abundant.

Two unusual breccia units (Stops 4-1 and 5-B), each as much as 20 m thick, are exposed in the iron-formation near Atkins Lake. They have not been described prior to 1995 (Feher and Flood, 1995). The breccias consist of angular to subrounded clasts as much as 1 m across composed of banded siliceous sedimentary rocks in a matrix of dark gray-green massive rock. The matrix is clearly igneous and has a basaltic composition but whether the units are extrusive or intrusive is uncertain. In either case, the presence of these breccias indicates that igneous activity occurred very close in time to deposition of the iron-formation in the western part of the Gogebic Iron Range.

The situation is similar to that observed in the eastern part of the Gogebic Iron Range in Michigan, where the Emperor Volcanic Complex is interlayered with the Ironwood Iron-formation. Apparently quiescent, platform-like conditions, expressed in the lateral continuity of lithologic units of the Menominee Group in the central part of the Gogebic Range, change into more volcanically-active environments to both the east and west. Lateral change from cherty iron-formation common in the central part of the range, to non-cherty silicate iron-formation and ferruginous argillite in the west, also suggests proximity to a clastic sediment source that diluted the chemically precipitated components of the iron-formation. The increased detrital component in the Ironwood, along with contemporaneous volcanism, at both the east and west end of the range, may indicate transition to a region of extension and rift-basin development that marked

continental breakup at the outset of the Penokean orogenic cycle.

The Tyler Formation is unknown west of the Mineral Lake intrusion but is probably present, at least locally, beneath low, swampy country north of the Ironwood outcrop belt. Magnetic traverses across these areas show extremely flat profiles characteristic of the Tyler elsewhere.

The western termination of strata of the Gogebic Iron Range is in the area between Lake Namekagon and Cable, where the Atkins Lake fault, a Middle Proterozoic structure discussed below, cuts down section and truncates Early Proterozoic strata against Middle Proterozoic volcanic rocks. Outcrops are extremely sparse in that area, but the fault trace can be located using magnetic data.

GEOLOGY OF MIDDLE PROTEROZOIC ROCKS

Middle Proterozoic rocks consist of a thick sequence of continental volcanic rocks overlain by red fluvial clastic rocks and lesser lacustrine deposits, that together make up the Keweenawan Supergroup. All are related to the Midcontinent Rift System that developed at about 1100 Ma. The rift was tectonically inverted shortly after it formed. In the Hurley-Cable area, the inversion resulted in southward-directed thrusting and northward tilting of a crustal-scale block, so that the Middle Proterozoic rocks now lie in a northward-dipping monoclinial sequence, the Montreal River monocline (Cannon and others, 1993b). The rocks in the monocline provide a complete history of the rift including extension and continental volcanism, post-rift thermal subsidence and sedimentation, and later inversion resulting from regional compression.

Extension and volcanism: The lower part of the Keweenawan Supergroup is composed of volcanic and lesser sedimentary rocks that formed during the extensional phase of the rift and accumulated to a thickness of as much as 12 km in this area, but were as much as 20 km thick along the rift axis to the north. The lowest unit is the Bessemer Quartzite (Stop 2-6), chiefly gray to white, thin-bedded quartzite, mostly of lacustrine origin (Ojakangas and Morey, 1982). A massive basal conglomerate is made up mostly of clasts of Early Proterozoic and Archean rocks typical of those exposed south of the Bessemer outcrop belt. The Bessemer appears to have formed in a broad basin that contained a shallow lake. This basin is apparently the first manifestation of extension related to the Midcontinent rift, and formed immediately prior to the initiation of continental flood-basalt volcanism, which was soon to dominate the region.

The oldest volcanic sequence is the 1.3-km thick Siemens Creek Volcanics (Stops

2-6, 4-3, and 4-7). These volcanic rocks have not been dated in this area, but are probably equivalent to the Osler Volcanics on the north shore of Lake Superior where eruption began at about 1109 Ma (Davis and Sutcliffe, 1985). We have made an informal subdivision of the Siemens Creek Volcanics into a thin, discontinuous lower unit and a much thicker upper unit. Flows in the lower unit (Stop 2-6) are geographically restricted basalts that have small clinopyroxene phenocrysts, a rarity in Keweenawan volcanic rocks, and we recognize a distinctive chemical composition (low Al_2O_3 , high Cr, Ni, MgO, and steep REE patterns), indicating that the flows reflect partial melts from a deep, enriched mantle source with little or no fractionation prior to eruption. The upper part of the Siemens Creek Volcanics (Stops 4-3 and 4-7) comprises widespread flood basalts of a more evolved composition, ranging from basalt to andesite. Their chemical composition indicates a larger degree of partial melting of a mantle source, modified by fractional crystallization during ponding within or at the base of the lithosphere prior to eruption. The basal 1-3 flows of the Siemens Creek Volcanics, erupted into the lake into which the Bessemer Quartzite was being deposited, typically are pillow basalt and pillow breccia. The Bessemer was unconsolidated at the time of eruption, and Bessemer-like sediments can be found in places as interflow accumulations above the lowermost flow. All remaining overlying flows appear to be subaerial based on lack of pillow structures.

The Kallander Creek Volcanics (1.2-4.5 km thick) overlie the Siemens Creek Volcanics and consists of a much higher proportion of andesite and rhyolite flows than are present in the Siemens Creek. We recognize two informal units in the Kallander Creek Volcanics. A lower unit, about half of the total stratigraphic thickness, is mostly andesite and basalt (Stop 4-6), and appears to record continuation of the widespread continental flood basalt eruption present in the Siemens Creek Volcanics, whereas the upper part is more localized and most probably represents a central volcano near Mellen. The upper part of the section contains abundant rhyolite and conglomerate (Stops 2-4 and 4-8) in addition to basalt, andesite, and other rocks of intermediate composition (Stop 2-5). The lower unit is very similar to the lower Siemens Creek basalts (small degree partial melts erupted from deep in the mantle). However, our analyses show that the basalts in the lower Kallander Creek are more strongly fractionated (high TiO_2 and FeO, low MgO), and lack clinopyroxene as a phenocryst phase. A rhyolite near the top of the lower unit was erupted about 1107 Ma (Davis and others, 1995).

In contrast to the lower unit, the upper Kallander Creek unit contains abundant rhyolite and conglomerate in addition to basalt and andesite. Chemical compositions of

mafic and intermediate rocks indicate that parent magmas were generated from large-degree partial melts of a mantle source, and that substantial fractional crystallization took place within or near the base of the crust. Rhyolites were generated, at least in part, by partial melting of crustal material. A rhyolite that forms the uppermost unit of the Kallander Creek Volcanics (Stop 2-4) was erupted at about 1099 Ma (Zartman and others, 1995). The Mellen Complex (Stops 4-4 and 4-5), an intrusive suite of gabbroic rocks, granophyre, and granite, was emplaced into the Kallander Creek Volcanics and older units at about 1102 Ma (Cannon and others, 1993a) and probably represents a shallow magma reservoir that fed eruptions from the central volcano.

The area near Mellen remained a topographic high after formation of the central volcano and no further volcanic rocks were deposited. To both the east and west, additional flood basalt volcanism deposited a great thickness of volcanic rocks, mostly confined to the central graben of the rift. These volcanic rocks are represented by the Portage Lake Volcanics to the east and the Chengwatana Volcanics (Stops 4-9 and 4-10) to the west. The two units are quite similar chemically, each dominated by basalt with lesser intermediate and felsic rocks. In each, the basalts can be divided into two major groups based on whether the TiO_2 content is greater or less than 2.5 wt % (Nicholson and Shirey, 1990). The low- TiO_2 basalts are the most abundant and represent widespread flood basalt volcanism: they are typically ophitic in texture and commonly form thicker flows than the high- TiO_2 basalts. The high- TiO_2 basalts form thinner and more localized flows, and commonly are interlayered with more intermediate and felsic rocks in the section. The main phase of rift extension and volcanism continued for about 5 m.y., until about 1094 Ma (Davis and Paces, 1990).

The youngest volcanic rocks in the area are basalt and andesite flows of the Porcupine Volcanics (Stop 2-3) that make up a central volcanic complex centered to the east of the Hurley-Cable area, near the Porcupine Mountains in Michigan. Outcrops of the Porcupine Volcanics in the Hurley-Cable area comprise basalt and andesite flows and interflow conglomerate and sandstone. A rhyolite near the top of the unit was erupted at about 1094 Ma (Zartman and others, 1995).

Thermal subsidence and sedimentation: Following eruption of the Portage Lake and Chengwatana Volcanics, and coincident with development of the Porcupine Volcanics, the tectonic regime changed. Earlier extension-driven subsidence and related volcanism was replaced by a regime in which extension rapidly waned and further subsidence was caused by cooling and contraction of the hot root zone of the rift. The thermally subsiding basin was generally centered over the previous rift basin but was much

broad. This basin was filled with red clastic fluvial rocks and lesser lake deposits. As much as 7 km of sedimentary rocks accumulated in the deepest part of the basin.

The oldest sediments, comprising the Copper Harbor Conglomerate (Stop 2-2) and perhaps the overlying lake beds of the Nonesuch Formation (stop 2-2), were deposited in a basin transitional between the earlier extensional rift basin and the successor thermally subsiding basin. A regional topographic high centered near Mellen affected the distribution and lithology of nearby sedimentary units. The Copper Harbor Conglomerate and Nonesuch Formation both pinch out against this highland so that the overlying Freda Sandstone (Stop 2-1) locally lies unconformably on volcanic and intrusive rocks. The Freda Sandstone near Mellen is atypically conglomeratic, reflecting proximity of the nearby highland that served as a source of coarse clastic debris.

Rift inversion: Toward the close of the thermal subsidence phase and coincident with late phases of sedimentation, a period of regional compression began, that caused inversion of the rift structure. In the Cable-Hurley area, compression resulted in a series of major listric thrust faults along which previously deeply buried parts of the rift were thrust southward and the upper plates of the faults were tilted northward. This thrusting event is the major controlling factor in the present distribution and structural attitude of the rocks of the region.

The principal fault in the area is the Marenisco fault and its western extension, the Atkins Lake fault. From east to west across the area, the Marenisco fault gradually rises in the stratigraphic section. In the east, the fault trace is about 10 km south of the Early Proterozoic-Archean contact so that a large thickness of Archean rocks and the entire Early Proterozoic and Middle Proterozoic section is in the upper plate above the fault. The importance of the Marenisco fault as a Middle Proterozoic structure was not recognized until 1992, when Rb-Sr biotite ages demonstrated that the upper plate had been uplifted and cooled at about 1060 Ma (Cannon and others, 1993b). In the eastern part of the area, thrusting and northward rotation of the upper plate has resulted in exposure of a cross section of nearly vertically dipping strata about 20 km thick as well as about 10 km of underlying Archean basement rocks. This structure, the Montreal River monocline, exposes about 3/4 of the crust as it existed just after rifting and is the focus of Field Trip 2.

To the west, the Marenisco fault gradually rises to higher structural levels in Archean rocks until, along the Marengo River, it cuts upward across Early Proterozoic strata. From there westward, the fault has thrust Middle Proterozoic volcanic rocks southward over Early Proterozoic and Archean rocks. The fault is traced mostly by

aeromagnetic data, but extensive exposures of intensely sheared and mylonitized rocks can be seen along the valley of the Marengo River south of Grand View, Wisconsin.

A second important fault is the Lake Owen fault. Southwestward from Cable, this fault separates Middle Proterozoic volcanic rocks of the Chengwatana Volcanics from Archean rocks. Northeastward from Cable the fault passes up section in the footwall, so that the Chengwatana Volcanics are juxtaposed against successively younger rocks. Eventually, northeast of Mellen, the fault passes into the Freda Sandstone, where its expression is obscure and it appears to die out.

The Keweenaw fault, a major reverse fault to the east of the area, also dies out within the volcanic section northeast of Mellen.

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GUIDE FOR FIELD TRIP # 2

**GEOLOGY OF THE MONTREAL RIVER MONOCLINE:
A TRAVERSE THROUGH 25 KM OF THE CRUST**

FIELD TRIP #2
GEOLOGY OF THE MONTREAL RIVER MONOCLINE:
A TRAVERSE THROUGH 25 KM OF THE CRUST

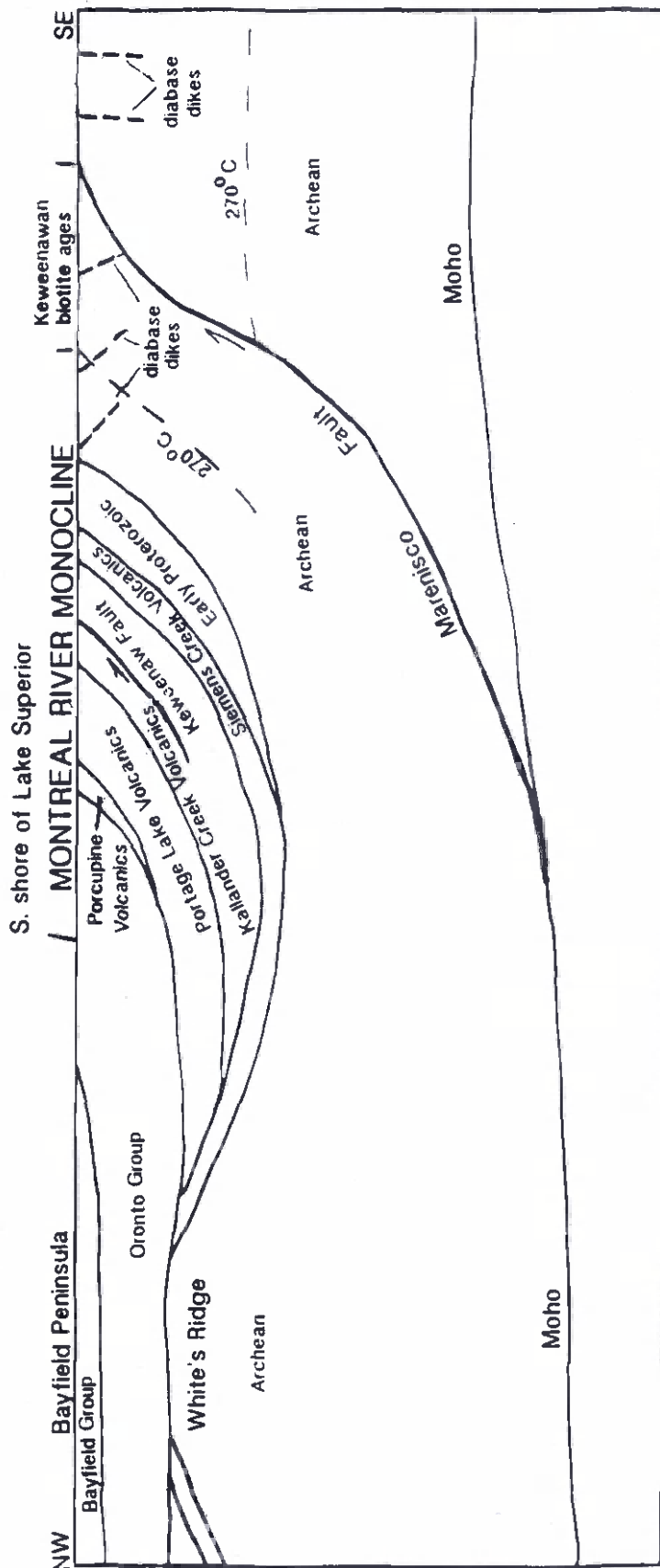
by
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U. S. Geological Survey, MS 954, Reston, VA 22092

South of Lake Superior, near the Wisconsin-Michigan border, a remarkable cross section displays what had been the upper 35 km of the crust. This section exposed as a sequence of nearly vertically dipping and north-facing volcanic and sedimentary units of Early and Middle Proterozoic age and underlying Archean rocks (Figures 2-1 and 2-2). This structure, named the Montreal River monocline by Cannon and others (1993), resulted from southward-directed thrusting on a northward-dipping listric thrust of crustal scale, the Marenisco fault. The age of thrusting is known to be about 1040-1060 Ma, the date of reset biotite Rb-Sr ages in Archean rocks that were upthrust and cooled on the upper plate above the fault (Cannon and others, 1993). The trip begins at the mouth of the Montreal River where extensive exposures of vertically dipping Freda Sandstone, the youngest unit exposed in the monocline, form bluffs along the Lake Superior shore. The traverse progresses southward for about 25 km through successively older and steeply dipping to vertical strata of the Keweenawan Supergroup, the Early Proterozoic Marquette Range Supergroup, and into Late Archean metavolcanic rocks.

STOP 2-1. Freda Sandstone at the mouth of the Montreal River (Fig. 2-3)

The Freda Sandstone is the youngest unit exposed in the Montreal River monocline but probably was overlain originally by as much as a few kilometers of continental sandstones deposited late in the history of the rift. The exposures seen here, representing the middle to upper parts of the Freda, may have been buried by as much as 3-5 km under younger parts of the Freda and younger sediments of the Bayfield Group. The sandstone is well exposed in high continuous bluffs along the Lake Superior shoreline. It is mostly a well bedded and commonly crossbedded feldspathic sandstone. Dips are essentially vertical. Volcanic rock fragments are a major constituent, indicating that rift-derived volcanic rocks composed much of the source area. Muscovite is also a common detrital mineral in many layers indicating that a basement terrane was also exposed that contributed a substantial part of the detritus.

Figure 2-1. Schematic cross section of the southern limb of the Midcontinent rift showing the regional structural context of the Montreal River monocline. The section extends approximately from the Bayfield Peninsula to Mercer, Wisc., a distance of roughly 100 km. To the NW a prominent basement high, White's Ridge, limited the depositional extent of much of the Keweenaw volcanic section and remained a positive topographic feature until the close of the extensional phase of the rift. As more widespread thermal subsidence supplanted extension, the ridge was buried beneath the Oronto and Bayfield Groups, which remain nearly flatlying. Southeastward from White's Ridge to the present outcrop belt of the volcanics, the volcanic section becomes progressively thicker. This geometry suggests that the deepest part of the rift in this area may actually have been to the south of the preserved volcanic rocks, in an area since uplifted and deeply eroded. The present steeply northward-dipping attitude of the rocks in the Montreal River monocline is believed to be the result of crustal-scale ramping on the Marenisco fault. The Marenisco fault is interpreted to penetrate the entire crust and to have thrust rift units and their Early Proterozoic and Archean basement southward. The listric geometry of the fault surface caused the upper plate to tilt northward, after which erosion has exposed the present crustal-scale cross section. The involvement of Archean rocks in the monocline structure is shown by the southerly dip of originally vertical diabase dikes and by Middle Proterozoic Rb-Sr ages of biotite, which mark time of uplift and cooling. The 270°C isotherm, the Rb-Sr blocking temperature of biotite, was located by determining Rb-Sr closure ages of biotite in Archean rocks. Rocks north of the isotherm (originally above the isotherm) remained cooler than 270°C throughout Keweenaw burial and retained ages reflecting Early Proterozoic and Archean thermal conditions. Rocks between the isotherm and the Marenisco fault were heated above 270°C by Keweenaw burial and Rb-Sr biotite ages were reset. In this area biotite ages range from about 1040-1060 Ma and date the time of uplift, erosion, and cooling caused by thrusting along the Marenisco fault. South of the fault, Middle Proterozoic burial was not deep and biotite ages again reflect Early Proterozoic thermal events.



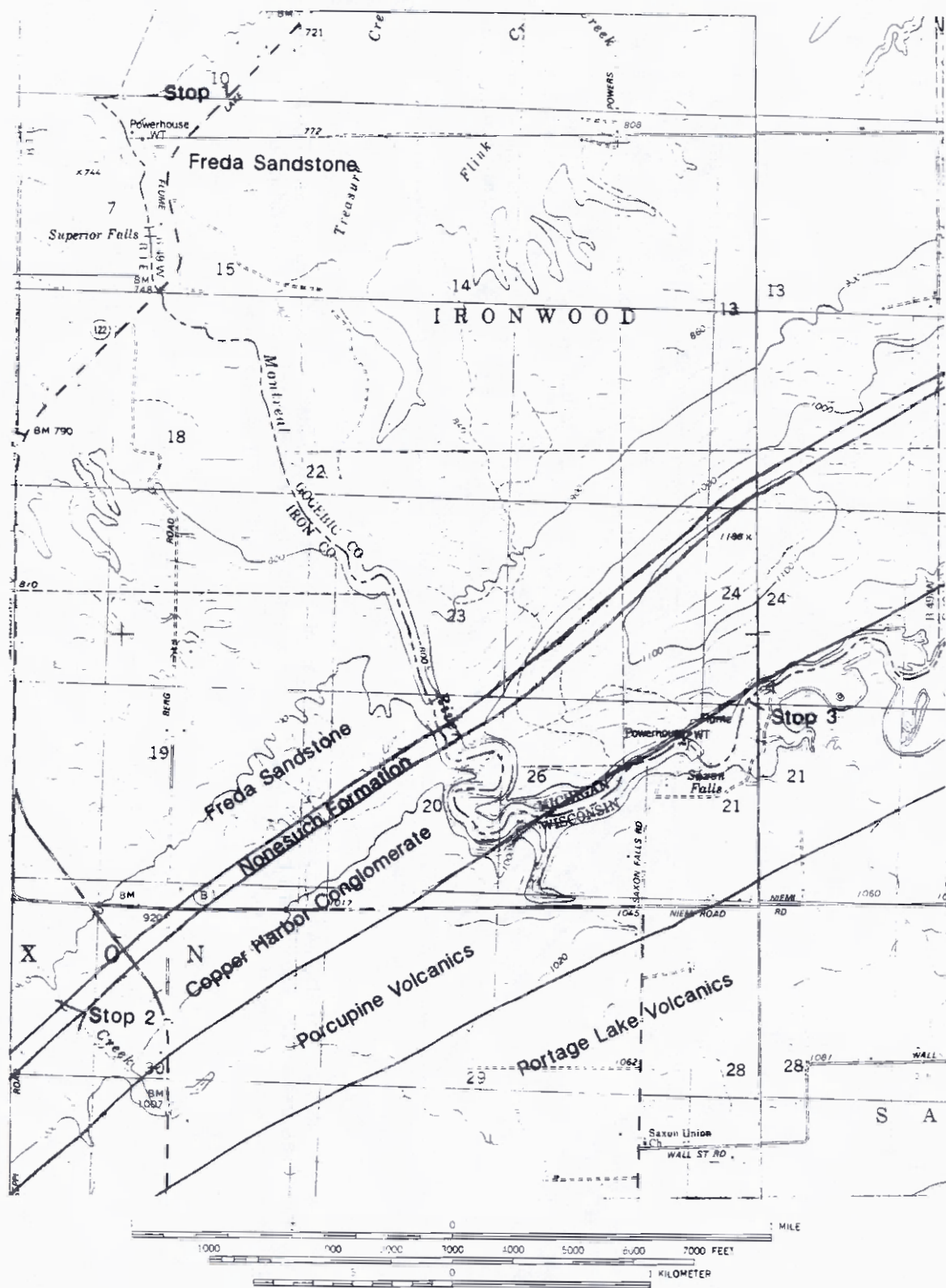


Figure 2-3. Parts of the Oronto Bay and Little Girls Point 7 1/2' quadrangles showing the general geology and location of field trip stops 1, 2, and 3.

An interesting secondary structural element is the presence of numerous nearly flatlying or very gently north-dipping faults. These faults invariably have only a few centimeters to about a meter of offset in which the upper plate has moved relatively north, as expressed by drag on bedding and in some places by kinked, rather than broken, beds. Such small faults are common in exposures along the Montreal River. In the Copper Harbor Conglomerate, they cut clasts that can be identified on opposite sides of the fault giving precise displacements. The age and cause of these small faults are unknown.

STOP 2-2. Middle part of Oronto Group along Parker Creek (Fig. 2-3)

The top of the Copper Harbor Conglomerate, the Nonesuch Formation, and the lower part of the Freda Sandstone can be seen along the incised bed of Parker Creek. The Parker Creek section is the most complete exposure of the Nonesuch and adjacent units in the Lake Superior region. Beds dip from 75°N to vertical. Detailed stratigraphy of the Parker Creek section was described by Suszek (1991) on whose work the following generalized description is based. The upper part of the Copper Harbor Conglomerate consists of massive to crudely bedded clast-supported conglomerate with cross-bedded, pebbly sandstone interbeds. Clasts consist mostly of basalt and rhyolite with lesser granite and gabbro, all probably derived from slightly older Keweenawan rocks. Banded iron-formation, jasper, quartzite, and chert clasts are a minor component but are widespread, indicating that pre-Keweenawan basement was also exposed.

The Nonesuch Formation is about 130 m thick and comprises of mostly gray to pale brown siltstone, sandstone, and minor shale. It is generally thinly bedded to laminated, but thicker massive beds also occur. Small-scale trough crossbedding is common. There is an overall coarsening-upward trend to the formation, although many individual units display fining-upward textures. The fine-grained rocks of the Nonesuch Formation lie on coarse conglomerate of the Copper Harbor. A transitional bed, one meter thick, contains reddish-brown, trough-crossbedded sandstone at the base and grades upward into gray-brown calcareous siltstone. The upper contact of the Nonesuch with the Freda Sandstone is gradational over about 20 m, in which gray siltstone and shale are interbedded with reddish-brown, medium to coarse sandstone.

Additional exposures of the upper part of the Copper Harbor Conglomerate can be seen in low roadcuts along Highway 122 in the NW 1/4 of Sec. 30, T. 47 N., R. 1 E.

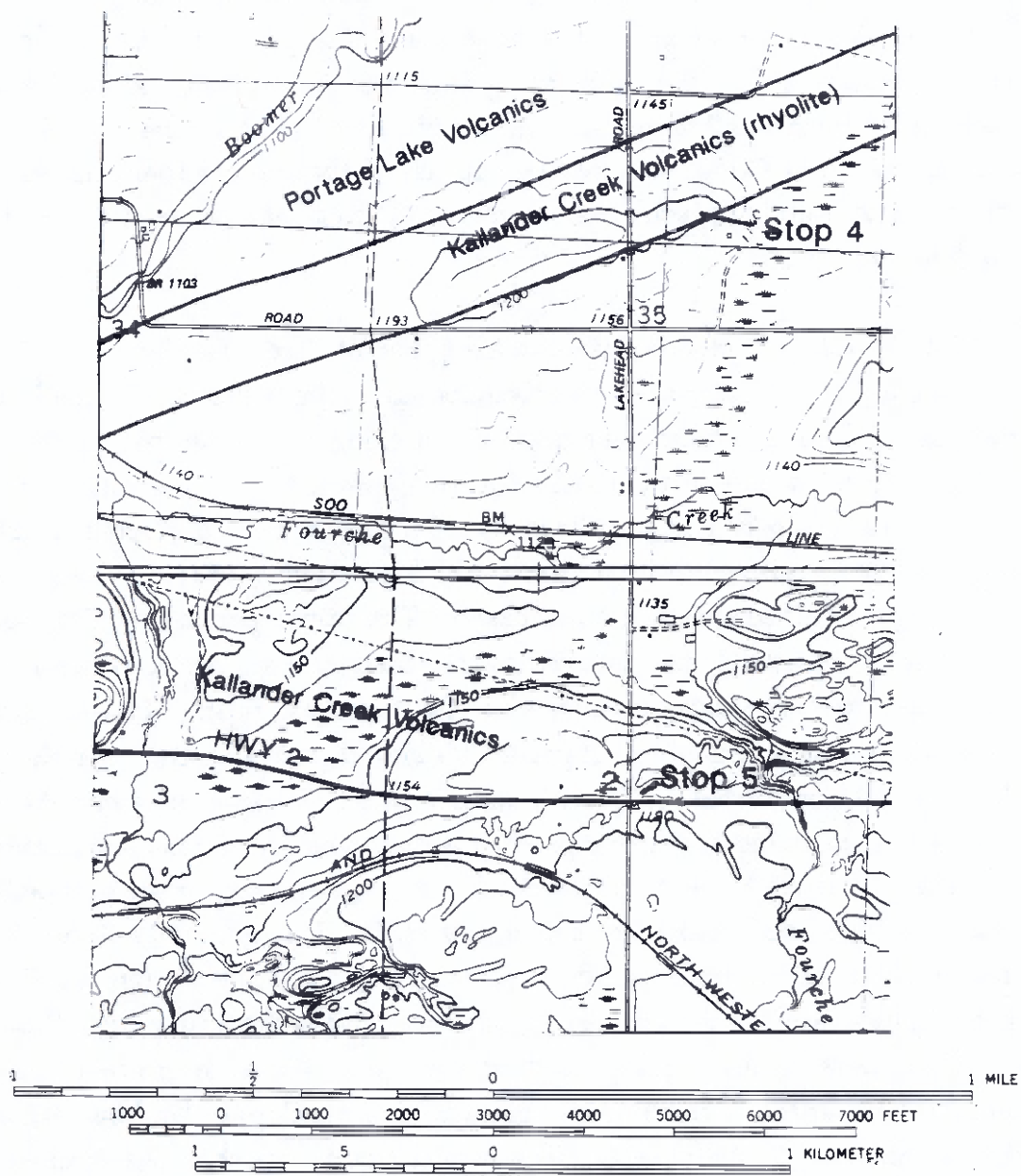


Figure 2-4. Parts of the Little Girls Point and Iron Belt 7 1/2' quadrangles showing the general geology and location of field trip stops 4 and 5.

STOP 2-3. Top of Porcupine Volcanics at Saxon Falls (Fig. 2-3)

Basalt flows and interflow conglomerate and sandstone of the uppermost part of the Porcupine Volcanics are exposed along the Montreal River below the dam at Saxon Falls flowage. The uppermost exposed unit is thin-bedded red sandstone and siltstone, which is underlain by a massive basalt flow about 100 m thick. This flow is underlain by about 30 m of interbedded conglomerate and sandstone. The lowest exposed unit is another basalt flow with an amygdular, partly brecciated top. About the upper half meter of the flow shows typical rubbly flow-top breccia, and the upper three meters are coarsely amygdular. The most abundant amygdule-filling mineral is thomsonite, which occurs in masses of coarse radiating blades in the larger amygdules. The predominance of this low temperature zeolite indicates that rocks at this level of burial were only weakly heated before uplift.

STOP 2-4. Rhyolite at top of Kallander Creek Volcanics (Fig. 2-4)

The uppermost unit of the Kallander Creek Volcanics is a thick, mostly massive, quartz- and plagioclase-phyric rhyolite. We informally designate this unit the Sheep Farm rhyolite and this stop is our informal type locality. The reason for the name is obvious. The unit is at least 300 m thick and appears to be the product of a single eruptive event. It can be traced from this locality westward for about 30 km to Copper Falls State Park near Mellen, where it is the resistant rock that holds up Brownstone Falls on the Tyler Forks River. At that locality, we have located by field measurement the horizon of the principal magnetic reversal recorded in the Keweenaw section. The reversal lies a few flows beneath the rhyolite. The rhyolite has been dated at 1099 Ma (Zartman and others, 1995), so the age of the magnetic reversal must also be approximately 1099 Ma. At this stop, the exposures reveal the basal part of the flow, which contains numerous blocks of rhyolite up to about 20 cm across. These blocks have a slightly different texture and phenocryst content than the host rhyolite.

STOP 2-5. Dacite in Kallander Creek Volcanics (Fig. 2-4)

The upper member of the Kallander Creek Volcanics contains abundant rocks of intermediate composition in addition to basalt and rhyolite. We have proposed that the upper part of the Kallander Creek Volcanics represents the partly eroded remains of a broad central volcano, and that the Mellen Complex is the crystallized remnants of a magma chamber that was intruded into the volcanic edifice late in its eruptive history (Cannon and others, 1993). At this locality our data show that the rock knob in the gravel pit is a fine-grained reddish-brown dacite ($\text{SiO}_2=64\text{--}66$ wt%; $\text{TiO}_2=1.26$ wt%;

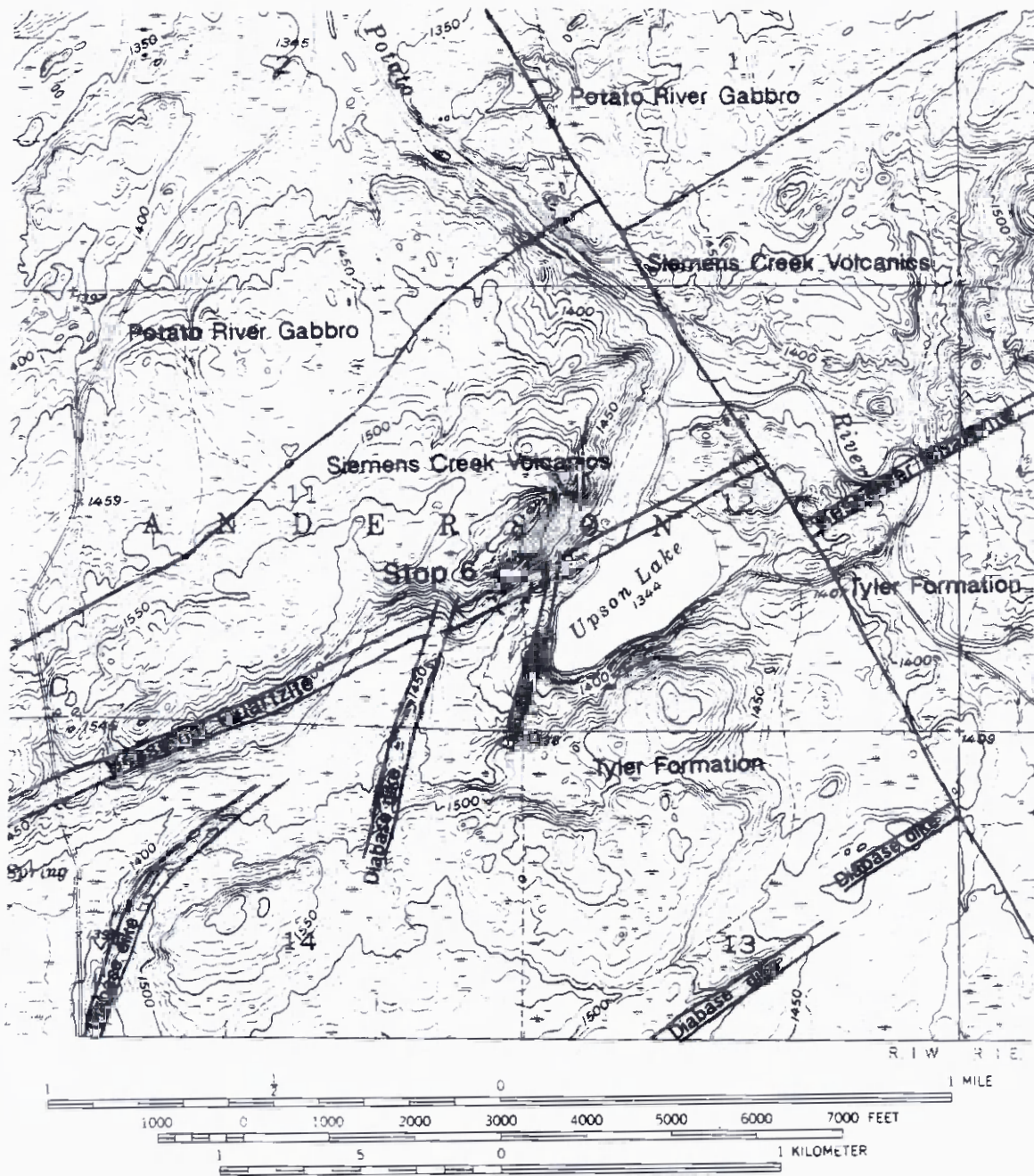


Figure 2-5. Part of the Saxon 7 1/2' quadrangle showing the general geology and location of field trip stop 6.

STOP 2-3. Top of Porcupine Volcanics at Saxon Falls (Fig. 2-3)

Basalt flows and interflow conglomerate and sandstone of the uppermost part of the Porcupine Volcanics are exposed along the Montreal River below the dam at Saxon Falls flowage. The uppermost exposed unit is thin-bedded red sandstone and siltstone, which is underlain by a massive basalt flow about 100 m thick. This flow is underlain by about 30 m of interbedded conglomerate and sandstone. The lowest exposed unit is another basalt flow with an amygdular, partly brecciated top. About the upper half meter of the flow shows typical rubbly flow-top breccia, and the upper three meters are coarsely amygdular. The most abundant amygdule-filling mineral is thomsonite, which occurs in masses of coarse radiating blades in the larger amygdules. The predominance of this low temperature zeolite indicates that rocks at this level of burial were only weakly heated before uplift.

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$Ce_N/Yb_N=4.5$) that contains small plagioclase phenocrysts. The rock has a strong primary foliation, perhaps indicating that it is the basal zone of a flow, and weathers into platy fragments parallel to the nearly vertical foliation. Small outcrops nearby show a coarser, more massive dacite. The dacite is overlain by a basalt ($SiO_2=50.24$ wt%; $TiO_2=3.78$ wt%; $Ce_N/Yb_N=3.7$) that is exposed in low outcrops along the power line north of the gravel pit.

STOP 2-6. Upson Lake area--Siemens Creek Volcanics, Bessemer Quartzite, and Tyler Formation (Fig. 2-5)

At this locality the base of the Keweenaw section is exposed in contact with the underlying Tyler Formation. At the top of the prominent bluff north of Upson Lake the lowermost basalt flows of the Siemens Creek Volcanics are exposed. They are pillowed and brecciated as a result of eruption into a shallow lake in which the Bessemer Quartzite had been deposited. The flows have a distinctive chemical composition marked by low Al_2O_3 (8-12 wt%), and high Cr, Ni, and MgO. In addition, rare earth element (REE) patterns are quite steep (Ce_N/Yb_N = about 15) and heavy REEs have low abundances. These chemical characteristics indicate that the magmas formed by a small degree of partial melting of a deep, enriched mantle source (postulated to be a mantle plume), and that the magma erupted directly to the surface without undergoing significant crystallization and fractionation. Flows of this chemical character can be recognized in the field by the presence of small phenocrysts of clinopyroxene. These clinopyroxene-phyric flows have been traced from near Bessemer, Michigan, to just west of this locality. They apparently represent a unique magma composition that was erupted only at the very outset of volcanism. Identical flows have been recognized at the base of the North Shore Volcanics in Minnesota, both at Ely's Peak and at Pigeon Point and Grand Portage (Green, 1977).

An interflow sedimentary bed is exposed between the first and second flows. The sediments are, in part, an algal laminite and are now substantially metamorphosed so that original calcareous layers are a fine-grained mat of wollastonite. The metamorphism was caused, at least in part, by emplacement of the Potato River Gabbro about 1 km up-section.

The Siemens Creek Volcanics are underlain by the Bessemer Quartzite, the oldest unit in the Keweenaw Supergroup. Scattered exposures of white to gray laminated quartzite are on the slope below the basalt exposures. Passing down-section, the Bessemer contains basal beds of coarse conglomerate composed largely of Early

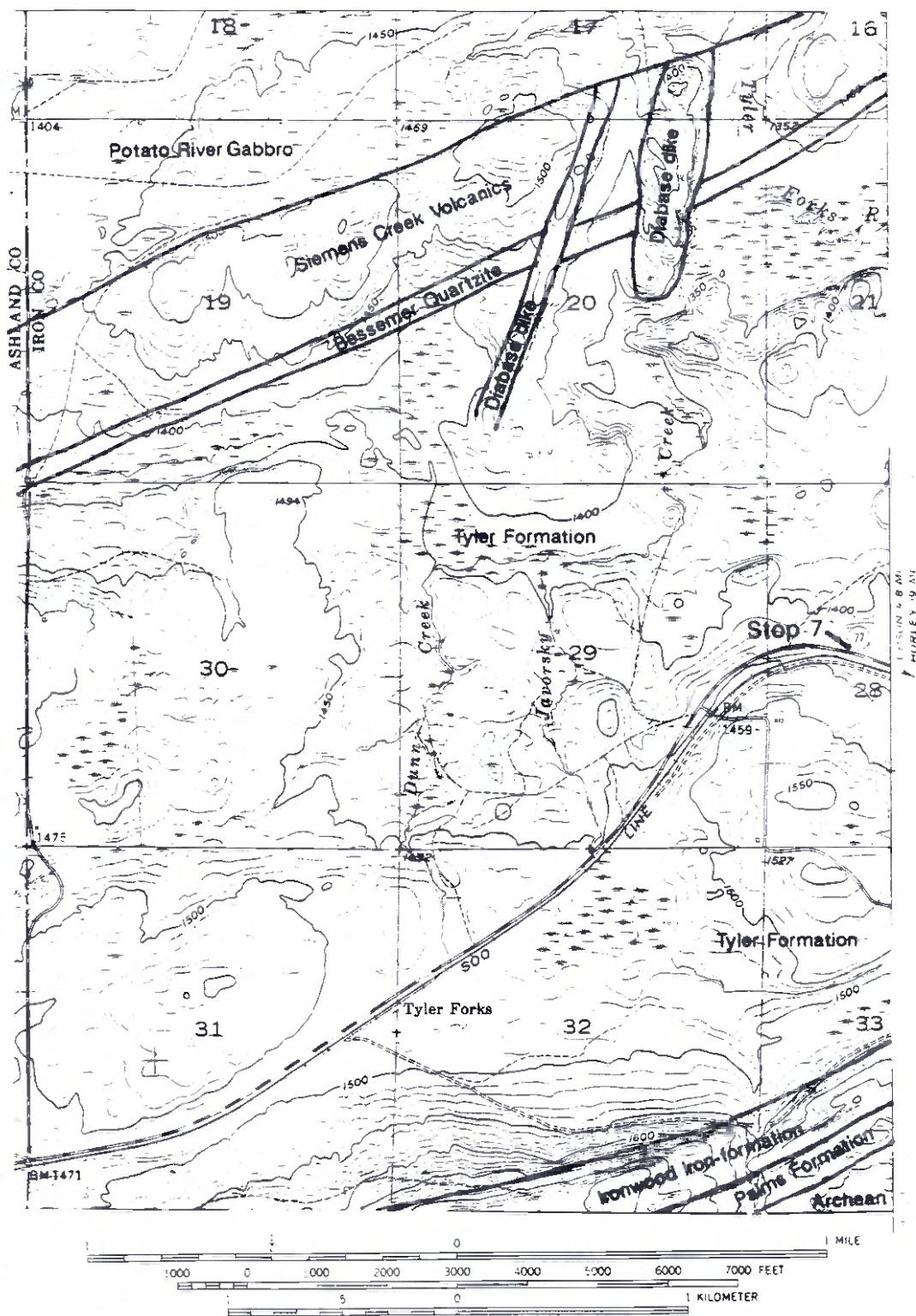


Figure 2-6. Part of the Mt. Whittlesey 7 1/2' quadrangle showing the general geology and location of field trip stop 7.

Proterozoic metasedimentary clasts, including iron-formation. Small garnet porphyroblasts are common in the matrix.

A small exposure near the base of the slope shows an angular unconformity between basal conglomerate of the Bessemer and thin-bedded argillite of the Tyler Formation. This is the only exposure known to us where the unconformable relationship between the Keweenawan Supergroup and the Marquette Range Supergroup can be seen.

STOP 2-7. Tyler Formation (Fig. 2-6)

This roadcut exposes typical thin-bedded graywacke and argillite of the Early Proterozoic Tyler Formation. Although dips are nearly vertical, the lack of penetrative fabric indicates little structural effect of the Penokean orogeny on the Tyler in this area. The highest grade metamorphic mineral is biotite, which is common throughout the matrix.

STOP 2-8. Ironwood Iron-formation on Mt. Whittlesey (Fig. 2-7)

The prominent ridge of Mt. Whittlesey is held up by the Ironwood Iron-formation. An unusually great thickness of iron-formation is present here owing to structural repetition by thrust faults and folds produced during the Penokean orogeny. Figure 2-7 shows our recent interpretation of the geology of the Mt. Whittlesey area that will be seen at Stops 2-8, 2-9, and 2-10. Mt. Whittlesey lies at the easternmost extent of the Penokean (Early Proterozoic) foreland fold and thrust belt. At Mt. Whittlesey, a series of thrusts repeats the Early Proterozoic section, having detached it from Archean rocks along a basal decollement. The thrusts all rise in the section to the east along a series of lateral ramps. Northward tilting of the region in Middle Proterozoic time results in the current map pattern that provides an unusual cross-sectional view of the thrust complex. To understand the Penokean structural geometry the map pattern should be viewed as a longitudinal cross section along the original trend of the thrust belt, so that the upper plates of originally nearly flat-lying faults would have moved northward away from the viewer. Because of later tilting, these original thrusts now mimic the geometry of normal faults--that is, the upper plates appear to have moved down to the north. The ramps that rise to the east are lateral ramps along which the plates have undergone transcurrent movement rather than overthrust movement. Displacement along the basal decollement diminishes eastward as splays from the decollement successively rise in the section. East of Stop 2-10 the Early Proterozoic section lies unconformably on Archean rocks.

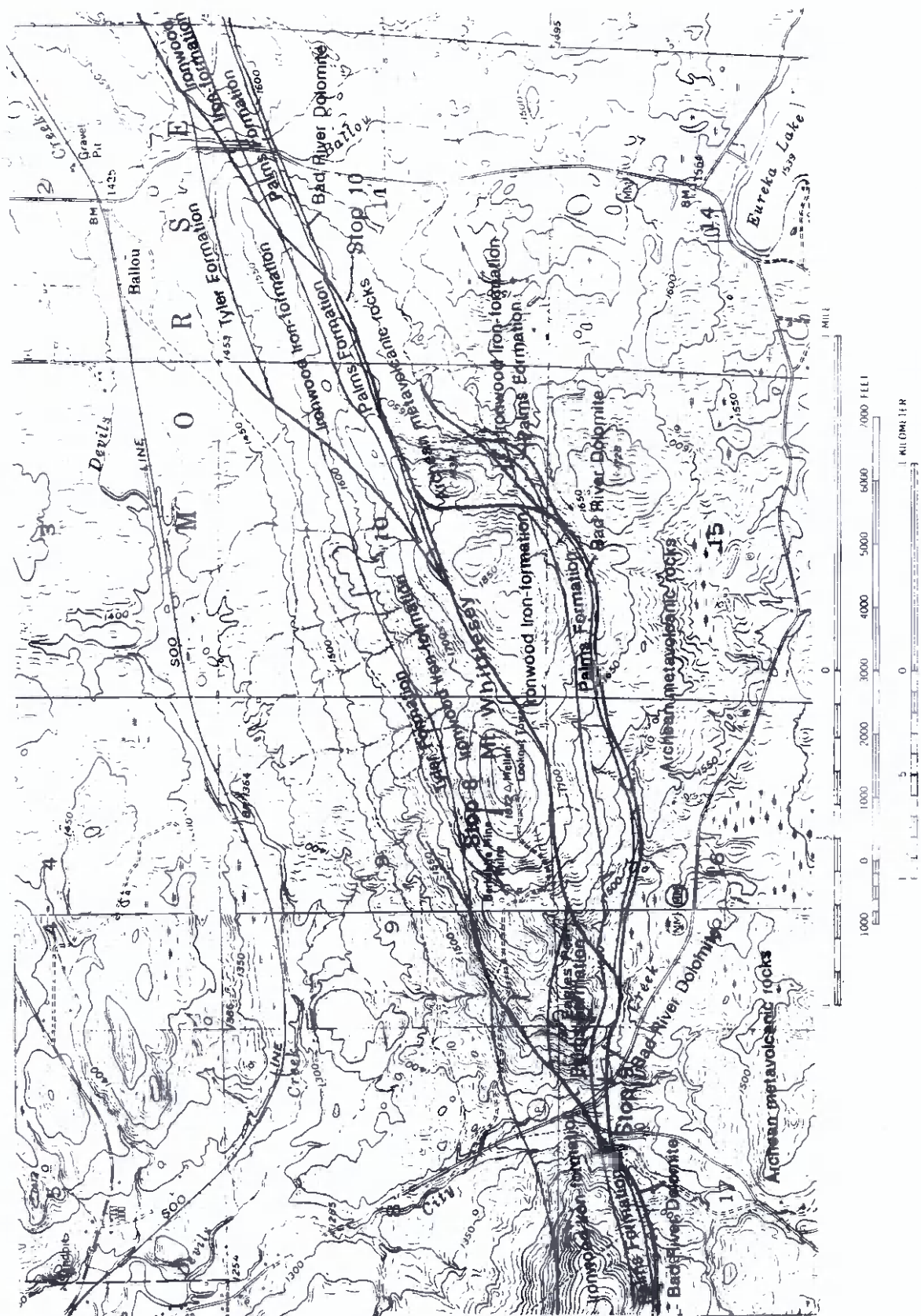


Figure 2-7. Part of the Mt. Whittlesey and Mellen 7 1/2' quadrangles showing general geology and location of field trip stops 8, 9, and 10.

At Stop 2-8 excellent exposures of the Ironwood Iron-formation occur on the north face of the ridge near the summit, and along a railroad grade near the Berkshire mine ruins. At the former locality, a large area was cleaned of overburden in the 1920s in anticipation of open pit mining. Only a small pit was developed near the base of cleared area; the remainder provides an excellent, glacially polished, exposure in which many details of sedimentary features are extraordinarily well displayed. At the Berkshire mine ruins an abandoned railroad grade provides a continuous cross section of several hundred meters of section and a tight fold is exposed in the cut.

The iron-formation contains several lithologic variations. The classic five-member internal stratigraphy defined to the east (Hotchkiss, 1919; Huber, 1959) consists of alternating members of even-bedded, carbonate iron-formation (slaty iron-formation) and irregularly-bedded hematitic jasper. This stratigraphy becomes more complex and irregular westward along the iron range; we have not been able to apply it west of Mt. Whittlesey. The iron-formation exposed here characteristically has many lithologic variations on a small scale so that individual exposures may show a complete range of lithologies interbedded on a scale of meters or less.

STOP 2-9. Palms Formation and Bad River Dolomite at Eagles Peak (Fig. 2-7)

At the southwest end of Mt. Whittlesey a prominent bluff, Eagles Peak, is held up by laminated argillite typical of the lower part of the Palms Formation. At the westernmost tip of the bluff the unconformable contact between the Palms and underlying Bad River Dolomite is well exposed. The Palms can be seen to be draped into shallow depressions along the top of the Bad River, which also has a weathered and rusty zone for about half a meter below the contact. The rusty color appears to be caused by recent oxidation of small amounts of pyrite along the contact.

The Bad River Dolomite at this locality is a chert breccia composed of angular to subrounded fragments of gray chert cemented mostly by a second generation of chert. Locally, well-rounded quartz grains partly fill interstices between breccia fragments, and magnetite is locally abundant as euhedral grains in the matrix. At other localities nearby, magnetite was sufficiently abundant to have promoted the sinking of prospect shafts in the early days of iron exploration. These are present generally immediately below the contact with the Palms Formation. This distinctive chert breccia has been traced as a mappable unit for about 12 km along strike. Locally it is underlain by marble and tremolitic marble more typical of the Bad River Dolomite in this region. We believe that the breccia is a residual deposit, formed by solution of an originally thick section of cherty dolomite during the weathering interval prior to Palms deposition.

The residue of insoluble chert was variably reworked into its present character and cemented by secondary silica. Magnetite was apparently introduced hydrothermally sometime after deposition of the Palms Formation.

STOP 2-10. Contact of Early Proterozoic strata and Archean metavolcanic rocks on the southeast flank of Mt. Whittlesey (Fig. 2-7)

Low outcrops near the base of a prominent ridge supported by the Palms Formation reveal the basal contact of Early Proterozoic strata with underlying dacite breccia of Late Archean age. The breccia has a penetrative structural fabric, best expressed by the prominent elongation of clasts. Long axes plunge moderately southward, but must have plunged moderately to steeply northward prior to tilting of the rocks into the monoclinial structure. The basal Early Proterozoic unit is chert breccia of the Bad River Dolomite, the same unit seen at stop 2-9. Along the contact the upper meter or two of the dacite breccia is strongly sheared parallel to the contact, and the Archean fabric typically is completely overprinted by this secondary fabric. Shearing is weak to absent in the chert breccia.

To our knowledge these outcrops are the only ones in the western part of the Gogebic Range where a basal detachment between the Early Proterozoic strata and Archean basement rocks can be directly observed. We believe the Early Proterozoic strata were thrust northward along a nearly flatlying basal contact during the Penokean orogeny. Because of later tilting of the Montreal River monocline, the present apparent displacement reflects a down-to-the-north normal fault. This fault can be traced westward, and about 1 km to the west of these exposures it passes up-section so that the Bad River-Palms-Ironwood sequence of the upper plate is juxtaposed over Ironwood Iron-formation in the lower plate, producing the unusually wide outcrop belt of Ironwood along higher parts of Mt. Whittlesey. Just east of this stop the basal thrust appears to pass upward into the Early Proterozoic strata. Exposures near Ballou Creek show that the base of the Early Proterozoic rocks lies unconformably on Archean volcanic rocks without an intervening shear zone.

The Archean rocks seen at Stop 2-10 are the structurally lowest parts of the monocline that are seen on the field trip. But the trace of the Marenisco fault, the master thrust that formed the Montreal River Monocline, lies about 8 km to the south. Thus, all of the stratigraphic section to the north that is seen on the trip, plus an additional 8 km of Archean rocks to the south, form the upper plate of the Marenisco fault, a coherent structural unit that was uplifted and tilted late in the Middle Proterozoic history of the region.

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GUIDE FOR FIELD TRIP # 4

**EARLY TO MIDDLE PROTEROZOIC GEOLOGY OF
THE LAKE NAMEKAGON REGION**

FIELD TRIP #4
EARLY TO MIDDLE PROTEROZOIC GEOLOGY OF
THE LAKE NAMEKAGON REGION

by

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This field trip offers new interpretations of the structure, geochemistry, and stratigraphy of the Lake Namekagon region based on recent studies by the U. S. Geological Survey. The trip will emphasize the rocks of the Powder Mill Group (Siemens Creek Volcanics and Kallander Creek Volcanics) and Chengwatana Volcanics of the Keweenawan Supergroup, and the cogenetic Mellen intrusive complex, all rocks related to the 1.1 Ga Midcontinent rift. Stratigraphic and geochronologic data from the different rock units of the Midcontinent rift in this region suggest a complicated history of volcanic eruption and magma emplacement. Early extrusive volcanics, flood basalts of the Siemens Creek Volcanics and perhaps the lower Kallander Creek Volcanics, erupted from a number of different centers in a generally subsiding basin throughout the western Lake Superior region. Development of a large shield volcano complex, possibly centered near the present location of Mellen, Wisconsin, was accompanied by intrusion of multiple magma chambers in the volcanic shield. This produced the Mellen intrusive complex and eruption of volcanic equivalents, the upper Kallander Creek Volcanics (Cannon and others, 1993). With a decline in shield volcano activity, flood-basalt eruption from within the central graben of the rift became dominant, forming a thick series of basalt flows, the Chengwatana Volcanics. Sedimentary rocks deposited during the post-extensional phase of the rift are covered by Pleistocene glacio-fluvial and lake deposits and are not exposed in this area. Post-rift reverse faulting along the Lake Owen fault and Marenisco-Atkins Lake fault during late Keweenawan time rotated all units to the northwest, tilting them to nearly vertical in some areas. As a result, a south-to-north traverse through the rift lithologies in the area north of Lake Namekagon provides a cross section both through time, from older to younger, and of crustal depth, from relatively deep to shallow.

The first stop will be in an Early Proterozoic sequence of iron and silica-rich argillites of the Ironwood Iron-formation and a metadiabase sill. This stop is notable for a spectacular exposure of unusual igneous breccia units (Figs. 4-1 and 4-2). All

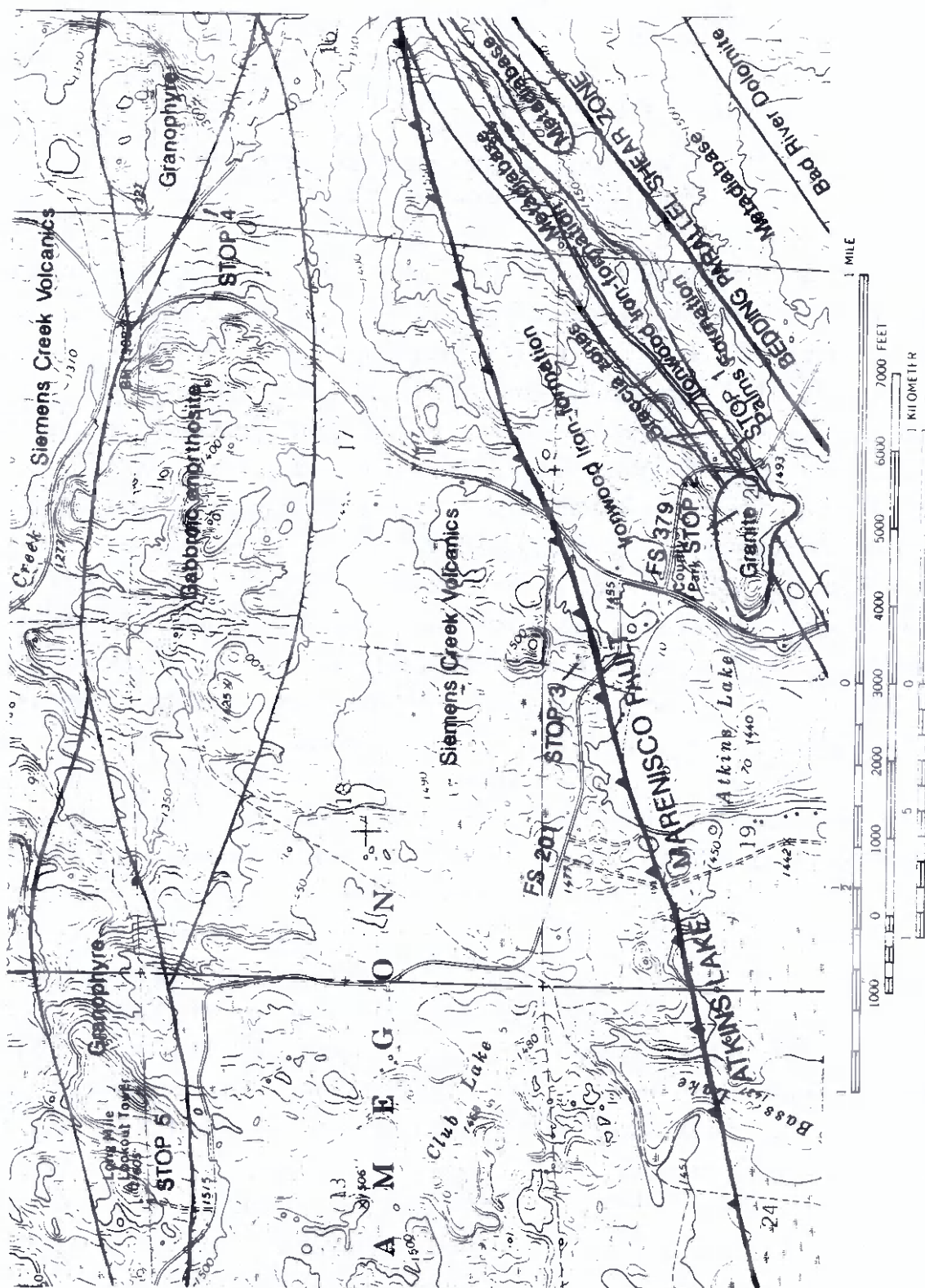


Figure 4-1. Part of the Grand View 7 1/2' quadrangle showing general geology and locations of field trip stops 1, 2, 3, 4, and 5.

subsequent stops will examine rocks of the Middle Proterozoic Midcontinent rift, including the Mellen intrusive complex (Stops 4-2, 4-4, and 4-5: Fig. 4-1) and volcanic rocks of the Powder Mill Group (Stops 4-3, 4-6, 4-7, and 4-8: Figs. 4-1 and 4-3) and Chengwatana Volcanics (Stops 4-9 and 4-10: Figs. 4-3 and 4-4)).

STOP 1. Early Proterozoic debris flow and metadiabase (Figs. 4-1 and 4-2).

The topographic highland extending northeastward from Forest Service Road 379 has numerous exposures of Ironwood Iron-formation and metadiabase sills typical of the westernmost extent of the Gogebic Iron Range. Near the junction with Forest Service Road 201, just east of Atkins Lake, two unusual breccia units are also exposed. The breccia units contain angular to subrounded clasts, up to 1 m in length, of metamorphosed, banded siliceous sedimentary rocks in a matrix of dark gray-green massive rock. Clasts mainly are siliceous argillite, a very few of which are magnetic. Most are tabular with angular corners, but some clasts are bent or folded, suggesting that they were not fully lithified when incorporated in the matrix. Clasts consist of alternating layers of recrystallized quartz, fine-grained chlorite, biotite or actinolite, garnet, and generally minor amounts of magnetite (Feher and Flood, 1995). The matrix is a fine-grained mafic rock, composed of fine- to medium-grained amphibole, garnet, plagioclase, and magnetite, with conspicuous amygdules in places. Metamorphic recrystallization has destroyed much of the primary texture, although relict plagioclase phenocrysts are present locally. Preliminary geochemical data indicate a tholeiitic composition for the matrix (Feher and Flood, 1995). Between the breccia units is a massive metadiabase sill about 50 m thick. The top of the sill is chilled against the base of the upper breccia unit. The base of the sill is separated from the top of the lower breccia unit by a thin selvage of magnetic iron-formation. The breccia units and interlayered sill can be traced for about 1 km along strike, but likely extend farther into areas with poor exposure.

The breccia units may be extrusive basalt flows or ash that incorporated clasts during eruption. Features suggestive of an extrusive origin include: 1) perfectly concordant contacts between breccias and thinly laminated sediments and a lack of any observed cross-cutting relationships; 2) strong lithologic contrast between the breccias and undoubtedly intrusive metadiabase exposed in the same outcrop; 3) common amygdular textures in the breccias that are absent in nearby metadiabase; and 4) rare areas of rubbly matrix that resemble pyroclastic textures.

Alternatively the breccias may be related to the mafic sill, which could have been emplaced at a shallow level into wet, semiconsolidated sediments. Evidence consistent

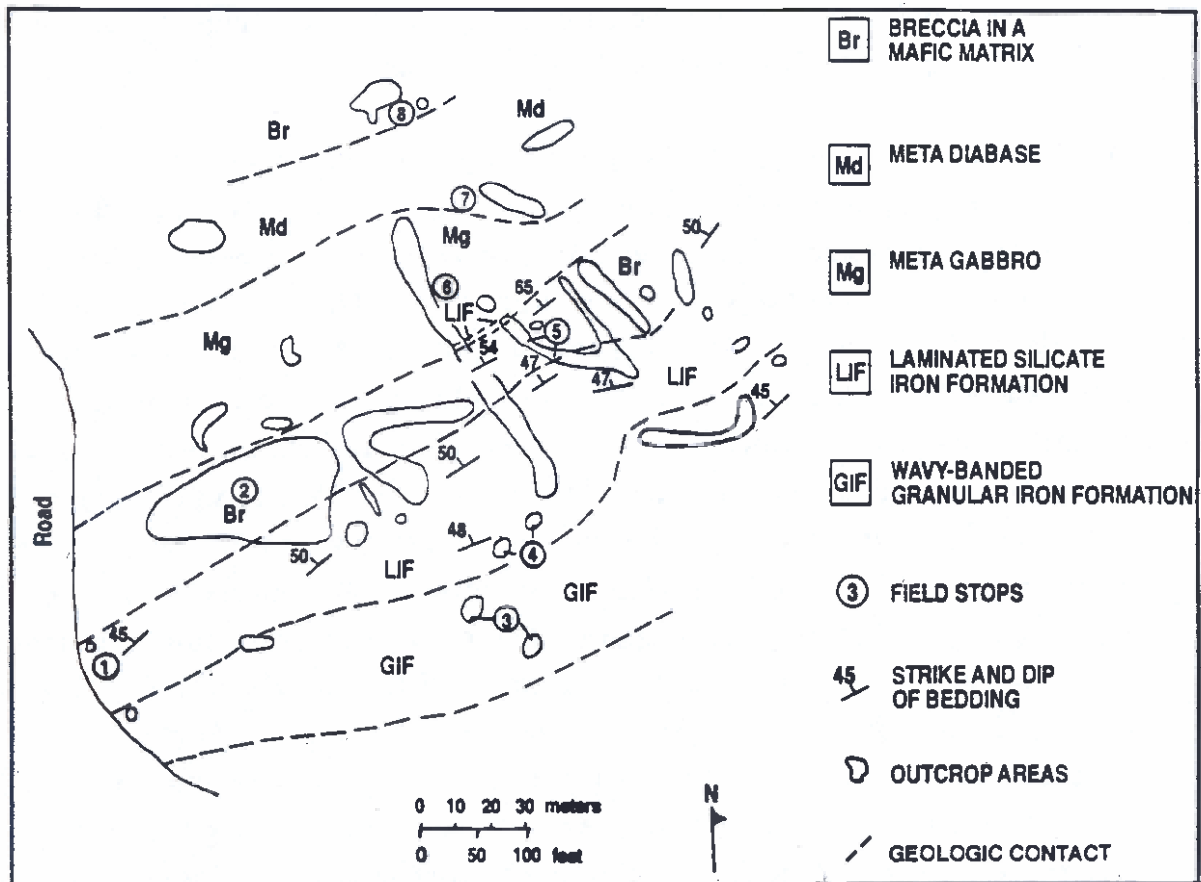


Fig. 4-2. Detailed map of the Atkins Lake area showing outcrops and geologic relationships among iron-formation, breccia units, diabase and gabbro. Unpublished map prepared by Klasner and LaBerge.

with an intrusive origin includes: 1) occurrence of a breccia zone both above and below the metadiabase sill (although the lower zone is separated from the sill by a thin seam of iron-formation); and 2) incorporation into the breccias of tabular (initially lithified) and bent (initially unlithified) blocks of the same composition.

Neither of the above interpretations seems to explain all observed relationships, and the origin of the breccia units thus remains enigmatic. An especially problematic feature is that nearly all of the clasts are exotic and are not composed of immediately surrounding iron-formation, but most closely resemble banded argillite of the lower part of the underlying Palms Formation.

STOP 2: Granite of the Mellen intrusive complex (Fig. 4-1).

The metadiabase and iron-formation seen in Stop 4-1 are intruded by several small granitic dikes. These dikes probably are offshoots of a small granite body exposed in several outcrops east of Atkins Lake, in the open woods south of Forest Service Road 379. The granite is pink and white, with a medium- to coarse-grained equigranular to porphyritic texture. Principal minerals are potassium feldspar, plagioclase, biotite, and quartz. The granite intrudes Early Proterozoic metasediments of the Ironwood Iron-formation; the lack of metamorphic effects and shearing within the granite suggests it is Middle Proterozoic in age. The granite is very similar in appearance to the well-known Mellen granite, a phase of the Mellen intrusive complex, which is locally exposed over a large area more than 20 km to the northeast of this locality.

In addition to this small granite body, the Mellen intrusive complex is made up of two large, layered gabbroic bodies, the Potato River intrusion and the Mineral Lake intrusion (Stop 4-4), and several related igneous bodies including the Mellen granite, the peridotitic Rearing Pond intrusion, and a series of gabbro and granophyre sills (Stop 4-5). A granophyre of the Mineral Lake intrusion has a U-Pb zircon age of 1102.0 ± 2.8 Ma (Zartman and others, 1995). The Mellen granite, which intrudes and brecciates gabbro of the Mineral Lake intrusion, has an indistinguishable U-Pb zircon age (within error) of 1100.9 ± 1.4 Ma (Zartman and others, 1995).

STOP 3. Siemens Creek Volcanics -- pyroxene hornfels and granophyre (Fig. 4-1).

The Siemens Creek Volcanics, the lowermost unit of the Powder Mill Group, is the oldest volcanic unit in the Midcontinent rift in northern Wisconsin. It consists of a sequence of flood basalts erupted as relatively thin flows. The Marenisco-Atkins Lake fault bounds the Keweenawan rocks in this area with Keweenawan rocks thrust

extension of the larger intrusion to the east. The rock consists of medium- to coarse-grained plagioclase, pyroxene, and magnetite, although minerals in the gabbro typically are altered. Because pyroxene and magnetite are more resistant than plagioclase to weathering, a hackley surface has developed on the outcrop, with knobs of mafic minerals prominent.

STOP 5. Granophyre of Long Mile Lookout (Fig. 4-1).

The upper part of the Mineral Lake intrusion becomes increasingly Fe-rich with the development of ferrogabbro and granophyre. Seifert and others (1992) determined that felsic intrusions associated with the Mineral Lake gabbroic anorthosite formed from evolved mantle magmas that intruded along older, mafic intrusive bodies and early volcanics, with further crystal fractionation occurring in place. The granophyre at this stop is an offshoot from the upper part of the Mineral Lake intrusion, one of several thin granophyric sills that interfinger with gabbroic sills in the western part of the Mellen intrusive complex. Granophyre exposed in the small hill of Long Mile Lookout typically is pink, fine- to medium-grained, and contains small phenocrysts of plagioclase in a graphic-textured matrix of potassium feldspar and quartz. The granophyre typically weathers brick red because of the presence of hematite.

STOP 6. Basalt and brecciated flow top, lower Kallander Creek Volcanics (Fig. 4-3).

The Kallander Creek Volcanics consists of a sequence of basalt, andesite, and rhyolite. The lower Kallander Creek Volcanics consists of flood basalts, similar to the Siemens Creek Volcanics, but the upper Kallander Creek may be an extrusive equivalent to the Mellen intrusive complex. A wedge of the Kallander Creek Volcanics, striking northeast-southwest and dipping to the northwest, occurs between the Marinesco-Atkins Lake fault to the south, the Lake Owen fault to the north, and the Mineral Lake intrusion to the east. Largely unaltered volcanics of the Kallander Creek can be traced into thin bands of metavolcanic rocks within the Mineral Lake intrusion, caught up between thin sills of gabbro and granophyre, attesting to a complicated history of volcanic eruption and magma emplacement.

At this stop a basalt flow near the base of the Kallander Creek Volcanics is exposed, including a thick flow top-breccia. The basalt varies from massive to sparsely amygdular, and is metamorphosed to an assemblage of hornblende-epidote-biotite-chlorite-plagioclase. The flow-top breccia consists of angular fragments of highly amygdular basalt. Breccia fragments are mostly a few centimeters in diameter but range up to about 1 meter. Amygdules mainly comprise quartz and epidote with minor

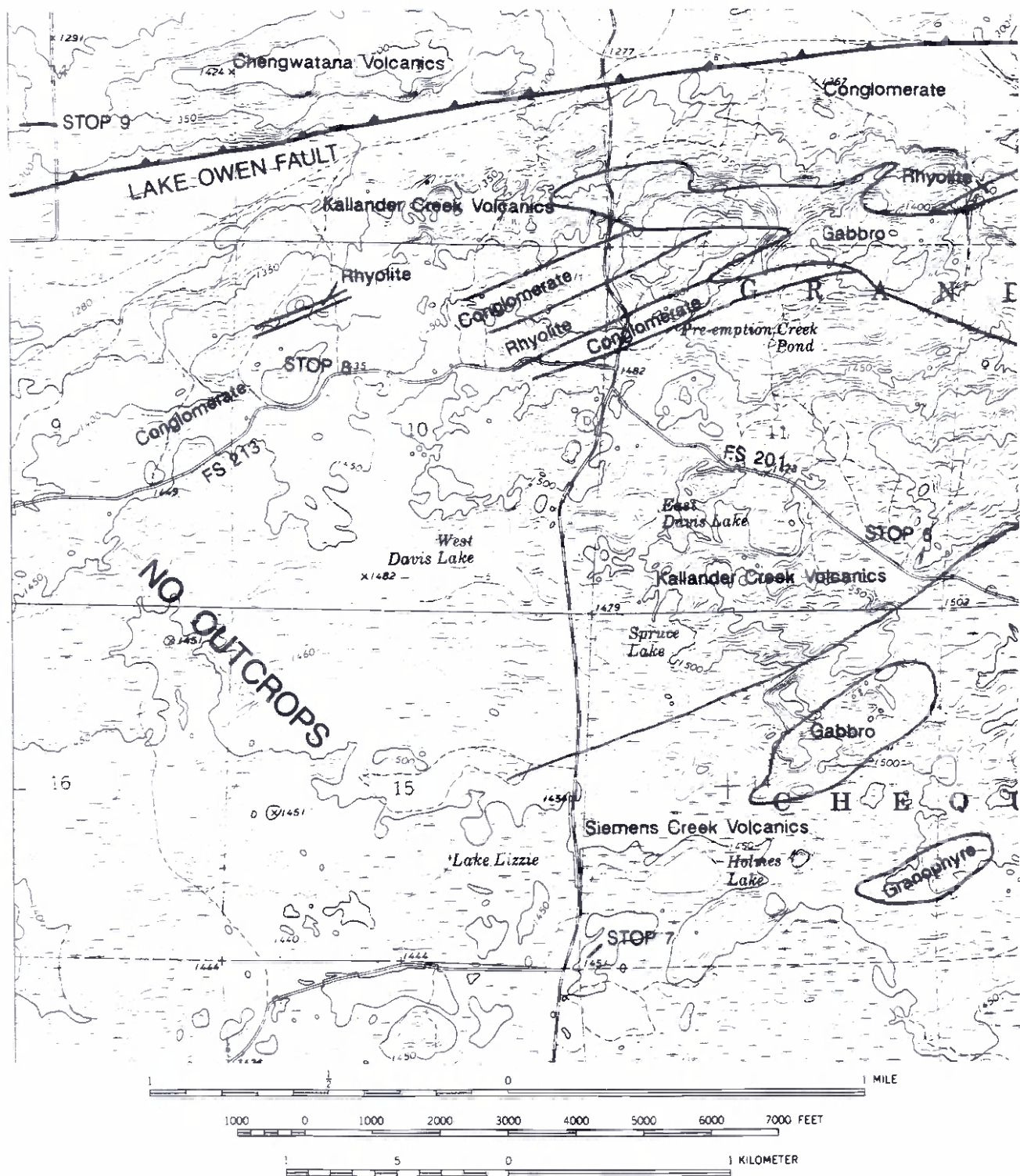


Figure 4-3. Part of the Grand View 7 1/2' quadrangle showing general geology and the locations of field trip stops 6, 7, 8, and 9.

K-feldspar and very sparse sulfide minerals. Locally, vesicles are exceptionally large, as much as 10 cm, and some are incompletely filled with secondary minerals, leaving open spaces.

The matrix of the flow top breccia is a very fine-grained clastic rock, now metamorphosed to chert-like material, although clastic textures are readily seen in thin section. Locally the matrix is finely laminated parallel to the orientation of the flow, even though the matrix is only thin seams deposited between the clast-supported breccia. In places (best seen in other outcrops) the laminations are highly contorted.

We interpret this silty matrix to be wind-blown dust that rapidly settled onto basalt flows while they were still molten and moving, becoming incorporated into open spaces in flow top breccias. Under some climatic conditions the sudden eruption of flood basalt may have generated intense wind and dust storms, which quickly deposited a layer of dust over the erupting flow.

A thin basalt dike cuts the basalt flow near the northeast end of the outcrop. Field measurements of magnetic polarity indicate that the basalt flow is reversely polarized, consistent with results for the Kallander Creek Volcanics elsewhere. The dike has normal polarity, and possibly was emplaced during the later eruption of the normally polarized Chengwatana Volcanics.

STOP 7: Picrite in the Siemens Creek Volcanics (Fig. 4-3).

This stop is out of stratigraphic sequence, but the significance of the outcrops makes it notable. The small outcrop just off County Road D, east of Lake Knotting, shows basalt flows within the lowest exposed unit of the Siemens Creek Volcanics in the region. Immediately to the south, Keweenawan intrusive rocks separate the flows from the Marenisco-Atkins Lake fault. A sample of these flows collected from this outcrop is picritic ($\text{MgO} = 16.9 \text{ wt } \%$; $\text{Mg\#} = 0.72$) and to date represents the only Keweenawan picrite identified on the south shore of western Lake Superior. Typically the lower member of the Siemens Creek Formation has substantially lower Al_2O_3 (8 to $<13 \text{ wt } \%$) and steeper rare earth element patterns ($\text{Ce}_\text{N}/\text{Yb}_\text{N} \approx 15$) than any other Keweenawan basalts. This picrite is different in composition from the well-documented picrites in the Keweenawan section at Mamainse Point in eastern Lake Superior (Berg and Klewin, 1988). Chemical modelling suggests that the picritic composition sampled at this outcrop represents an uncontaminated small-degree partial melt of the mantle plume postulated to have produced the voluminous magmatism associated with the Midcontinent rift.

STOP 8. Conglomerate and rhyolite of the Kallander Creek Volcanics at 'Davis Hill' (Fig. 4-3).

The middle portion of the Kallander Creek Volcanics in the Lake Namekagon area mainly comprised conglomerate and rhyolite with subordinate amounts of basalt. Exposures at this stop illustrate the lithologic variations of the unit. From Forest Service Road 213, walk northward over the prominent hill (informally known as Davis Hill in older reports) immediately north of the road. The best exposures of conglomerate are high on the western slope of the hill and along the north slope. Outcrops are poorly sorted and crudely bedded volcanogenic conglomerate containing minor sandstone interbeds. Clasts of cobble to boulder size are mostly rounded to subrounded rhyolite, and lesser basalt and andesite. Some clasts consist of basalt containing large pink phenocrysts of plagioclase; this lithology is fairly common in Kallander Creek flows, but is very rare elsewhere in the Keweenaw section. It appears, therefore, that the conglomerate formed through erosion of the slightly older lower part of the Kallander Creek Volcanics. About a kilometer to the east, this same conglomerate unit is cut by a tongue of the Mineral Lake gabbro that was emplaced at 1102.0 ± 2.8 Ma (Zartman and others, 1995), which places a minimum age on deposition of the conglomerate. Cement in the conglomerate is mostly a mixture of epidote and quartz, suggesting that relatively high-temperature fluids permeated this part of the section. A little copper staining can be seen rarely, but we have not identified any primary copper minerals.

Continue walking northward for a few hundred meters to a prominent bedrock knob. Along the south face of this knob is exposed a rhyolite unit about 20 m thick. The rhyolite is purplish, plagioclase phyric, and mostly massive. In a few places near the base of the hill, faint outlines of flattened pumiceous fragments can be seen, indicating that the rhyolite probably is a welded tuff. The remainder of the knob contains good exposures of typical conglomerate.

From the top of the knob a panoramic view to the north can be seen. The prominent east-west ridge about a kilometer distant is underlain by ophitic basalt flows of the Chengwatana Volcanics, which will be examined at the next stop. The Lake Owen fault lies just south of the ridge and has thrust the Chengwatana Volcanics southward over the Kallander Creek Volcanics.

STOP 9. Amygdaloidal ophitic basalt flow in Chengwatana Volcanics (Fig. 4-3).

The Chengwatana Volcanics records the main stage of flood-basalt volcanism of the Midcontinent rift in northern Wisconsin. Basalts erupted as a thick series of subaerial flows into a subsiding central graben. Reverse faulting during a regional

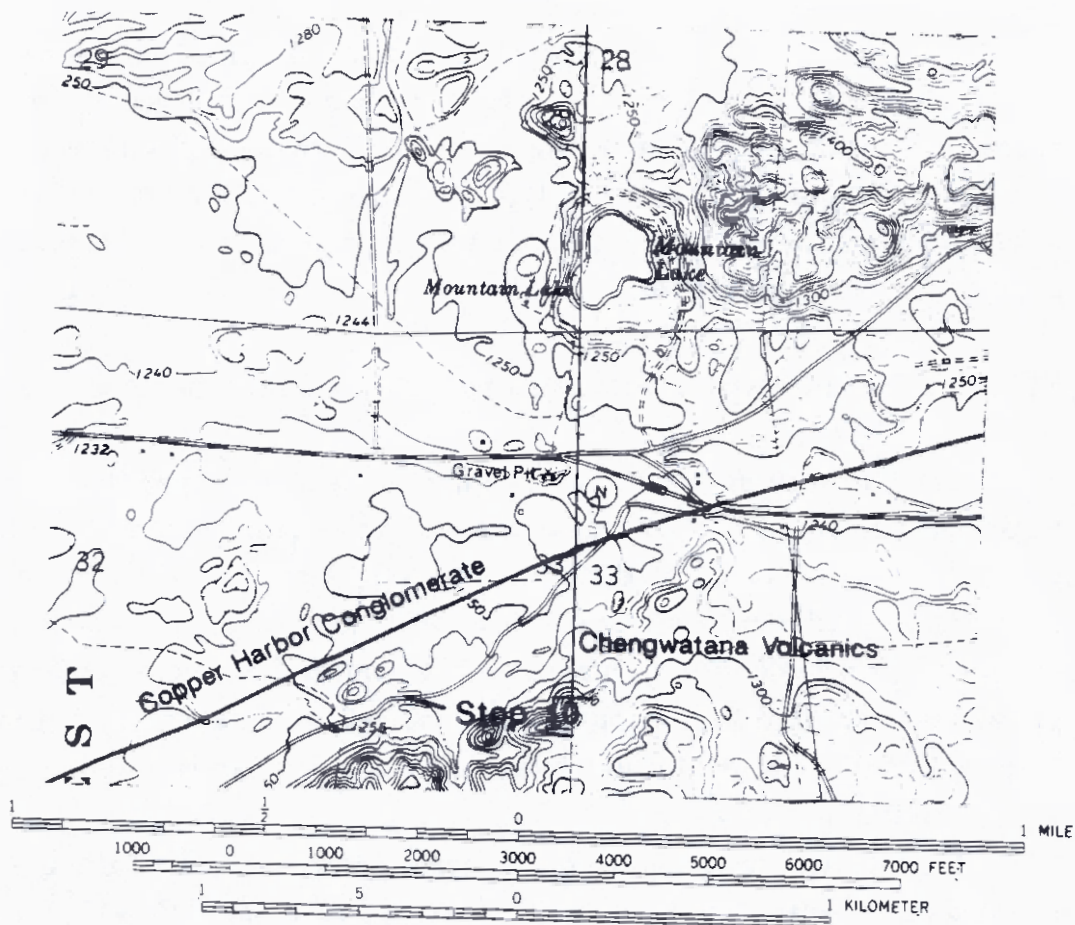


Figure 4-4. Parts of the Drummond and Upper Eau Claire Lake 7 1/2' quadrangles showing general geology and the location of field trip stop 10.

compressional event inverted the graben, forming the St. Croix horst, which contains the Ashland Syncline. On the north limb of the horst, the Chengwatana Volcanics are thrust along the Douglas fault over sediments of the younger Bayfield Group, while along the southern limb the Lake Owen fault thrusts younger Chengwatana Volcanics over older Kallander Creek Volcanics.

At this stop flows of the upper Chengwatana Volcanics are exposed along a steep south-facing bluff. The Lake Owen fault, probably at the base of the bluff, cuts up section so that the lower Chengwatana Volcanics are not exposed here. The flows are thick, monotonous ophitic basalts, typical of this part of the unit. Several amygdular flow tops are well exposed. Amygdules comprise a variety of secondary minerals, including quartz and epidote.

STOP 10. Ophitic basalt and interflow conglomerate near the top of the Chengwatana Volcanics (Fig. 4-4).

These exposures show what is certainly one of the upper flows, and may be the uppermost flow, of the Chengwatana Volcanics and an interflow conglomerate. Immediately north of the road are low outcrops of coarse volcanogenic conglomerate and interbedded sandstone that dip 20 - 30° to the north. The conglomerate is composed principally of rhyolite clasts. Cement in the conglomerate and sandstone varies from calcite to epidote plus quartz, indicating that even in this stratigraphically high part of the volcanic section, relatively high-temperature alteration has produced quartz-epidote assemblages.

Immediately to the north of the conglomerate outcrops, along a low ridge, is exposed coarse-grained ophitic basalt. This is typical of many flows in the upper part of the Chengwatana Volcanics where thick ophitic flows predominate. Rocks of a similar type are exposed in a series of low outcrops for about 100 m farther north. Whether these outcrops are the same flow or additional flows is unknown. Magnetic maps suggest that the contact with the overlying Copper Harbor Conglomerate lies just north of these outcrops, and that the featureless sand plain to the north is underlain by the Copper Harbor .

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GUIDE FOR FIELD TRIP # 5

**LAKE NAMEKAGON AND PENOKEE GAP AREAS,
WEST GOGEBIC RANGE, WISCONSIN**

FIELD TRIP #5
LAKE NAMEKAGON AND PENOKEE GAP AREAS,
WEST GOGEBIC RANGE, WISCONSIN

by

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INTRODUCTION

This field guide focuses on two regions of the Western Gogebic Range: Atkins Lake in the Lake Namekagon area and Penokee Gap near Mellen, Wisconsin (Overview Fig. 1). The Gogebic Range is approximately 120 kilometers long and consists mainly of a northwest-dipping sequence of Early Proterozoic iron-formation, interbedded volcanic rocks and detrital sediments sandwiched between Archean crystalline basement rocks on the southeast and Middle Proterozoic (Keweenawan) volcanic and sedimentary strata of the Midcontinent Rift on the northwest.

The section of the Gogebic Range that lies west of Mineral Lake provides the westernmost exposures of Early Proterozoic rocks of the Marquette Range Supergroup (Cannon and Gair, 1970). Unlike the eastern end of the Gogebic Range where the Sunday Lake Quartzite underlies the dolomite, the Bad River Dolomite at the western end of the range rests directly on Archean gneisses and volcanic rocks. In the Atkins Lake area, the dolomite is a 300 m thick gray dolomitic marble with abundant lenses and beds of chert. The dolomite thins towards the east and is discontinuously exposed. Near Ballou Creek (east of Mellen) it is mostly a variegated chert breccia cemented by chert and (or) quartz, locally with lenses of dolomite near the base. This lithology is interpreted as a residuum of chert nodules and beds formed as the dolomite was dissolved during the weathering interval prior to deposition of the unconformably overlying Palms Formation.

Locally, magnetite and hematite concentrations are present at the top of the chert breccia. Exploration shafts were sunk on some of these magnetite-rich zones near Penokee Gap and on the eastern flank of Mt. Whittlesey during early exploration for iron ores near the turn of the century. The occurrence of these magnetite concentrations in conglomerate at the top of the Bad River Dolomite suggests that they are paleo-placer deposits produced by streams flowing on the pre-

Palms erosion surface. Alternatively, they may be of hydrothermal or metamorphic origin.

The Palms Formation, which is well exposed at Penokee Gap and at Atkins Lake in the Lake Namekagon area, is the basal unit of the Menominee Group and unconformably overlies the Bad River Dolomite. It is comprised of a thinly bedded argillaceous lower unit and a thickly bedded, relatively pure, quartzite upper unit. Ojakangas (1983) similarly described the Palms elsewhere on the Gogebic Range and suggested that the Palms Formation was deposited in a tidal-flat environment in a transgressing sea.

Near Penokee Gap, the uppermost Palms Formation grades across 1 to 2 m into the overlying Ironwood Iron-formation. Here, the base of the Ironwood has thin (3 m) conglomerate zone. Elsewhere a one meter thick zone of stromatolitic jasper occurs at the base of the iron-formation. Detailed mapping and drill core analysis by U.S. Steel geologists in the late 1950's, based on work of Hotchkiss (1919), subdivided the Ironwood Iron-formation eastward from Penokee Gap into five units. Although our mapping indicates that alternating units of granular, wavy bedded and laminated iron-formation are present west of Penokee Gap, it also indicates that, except for the first few hills of iron-formation west of the gap, the five-fold division of U. S. Steel does not extend westward. In the Lake Namekagon area the Ironwood Iron-formation contains interbedded units of granular, wavy-bedded and laminated cherty magnetic iron-formation, similar to exposures elsewhere along the Gogebic Range. The iron-formation, however, is interbedded with and grades laterally into magnetic and non-magnetic argillite, as well as mafic igneous rocks. These igneous rocks appear to be large sills of metadiabase with textures suggesting emplacement into semi-consolidated sediments (or basalt flows interbedded with the sediments). Drill cores in the Trapper Lake area, just north of Lake Namekagon, intersected granular and laminated magnetic iron-formation with thin interbeds of argillite, and at least 60 m of black pyritic argillite, several metadiabase sills, and several thin dikes of granite and pyroxenite. Thus, the Ironwood Iron-formation grades westward into a deeper water, anoxic, detritus-rich environment with penecontemporaneous igneous activity. This is in marked contrast to the central part of the Gogebic Range, where detrital rocks are essentially lacking in the iron-formation, and igneous rocks are scarce.

Recent work on the east end (LaBerge and Ojakangas, 1992; Anderson and Klasner, 1993; LaBerge and Klasner, 1994; Klasner and LaBerge, 1994A; Klasner et al., in press) and in the central part and west end (Klasner and LaBerge, 1994B; LaBerge

et al., 1995) of the Gogebic Range shows that, in contrast to the central part, strata on either end of the range thicken abruptly, are structurally more complex, and contain volcanic strata interbedded with Ironwood Iron-formation. On the Eastern Gogebic Range abrupt lateral changes in stratigraphy and structure within the Ironwood Iron-formation and interbedded volcanic rocks are interpreted to indicate the development of grabens in an extensional tectonic environment (Klasner and LaBerge, 1994A; LaBerge and Klasner, 1994). Likewise, on the western end of the range, deposition of detrital sediments (mainly argillites) and contemporaneous igneous activity also suggest a more tectonically active environment. We suggest that these stratigraphic and structural relationships indicate the development of rift basins at both ends of the Gogebic Range in a divergent tectonic regime during deposition of platform sediments, including iron-formation.

The Tyler Formation overlies the Ironwood Iron-formation, and at Penokee Gap consists of about 800 m of cyclically bedded black slate and siltstone. We believe that the Tyler is unconformable on the Ironwood, as is the stratigraphically equivalent Copps Formation in the Eastern Gogebic Range. This is supported by an outcrop of conglomeratic, sandy graywacke that lies near the base of the Tyler Formation about 600 m west of Penokee Gap along the north slope of the ridge of iron-formation. Other remnants of Tyler exposed along the contact with the underlying iron formation at Penokee Gap are intensely sheared and consist primarily of fine-grained quartz in an anastomosing matrix of opaque carbonaceous material. In an exposure about 2 km west of Penokee Gap, near the contact with the Ironwood Iron-formation, the Tyler is a black, carbonaceous slate. In thin section the carbonaceous slate is quartz-rich with a strongly foliated biotitic matrix in which superimposed concentrically-zoned cordierite porphyroblasts give it a spotty appearance. The cordierite porphyroblasts are hexagonal in cross section, about 2 mm in length and 1 mm wide and display remnants of aligned mica flakes in their cores.

Farther north, in a rock cut along the railroad tracks that extend through Penokee Gap, the Tyler Formation consists of beds of slate and siltstone that show subtle grading from quartz-rich silt at the base to more clay-rich units at the top. The slate is carbonaceous with aligned flakes of biotite, sericite and chlorite, and it also contains porphyroblasts of cordierite. Although igneous activity was contemporaneous with iron-formation deposition on both ends of the Gogebic Range, we found no Early Proterozoic igneous rocks within or cutting the overlying Tyler Formation that lies along most of the Gogebic Range. We interpret the lack of

igneous rocks within the Tyler to indicate a significant gap between deposition of the Ironwood and Tyler Formations.

In addition to cordierite within the Tyler, there are numerous other minerals that indicate that the Early Proterozoic strata near Penokee Gap were metamorphosed. Foliation within the Tyler and Palms Formations, expressed by aligned mica and chlorite, along with other deformational features in the Early Proterozoic strata, indicate that there was a regional deformational and metamorphic event in the Penokee Gap region. Other porphyroblasts in the Early Proterozoic strata, along with the cordierite, however, suggest the occurrence of a later metamorphic event. These include garnet within the iron-formation and skeletal garnets in the Tyler near its contact with the iron-formation, non-aligned chlorite porphyroblasts locally within the Palms that overprint the regional foliation, and possibly the euhedral magnetite crystals along the contact between the Palms and Bad River Formation. Although the origin of these metamorphic minerals has not been studied in detail, the fact that some of them overprint the regional deformational fabric clearly indicates that they postdate it. The most obvious possibility is that they were caused by thermal metamorphism associated with intrusion of the Mineral Lake Gabbro that lies about 2 km to the north of Penokee Gap, but further study of metamorphism in this region is needed to fully confirm this.

NOTE: Field stops at both the Atkins Lake and Penokee Gap involve walking over steep, hilly terrane, especially at Penokee Gap. Also, a walk of approximately four miles is involved in the Penokee Gap part of this field trip.

PART A: LAKE NAMEKAGON AREA (ATKINS LAKE)

Numerous exposures of the Ironwood Iron-formation and the Palms Formation are present in a 4,000 m long northeast-trending zone from Atkins Lake to the Marengo River, where the Early Proterozoic units are truncated by the Marenisco-Atkins Lake fault (Fig. 5-1). The Palms Formation consists of an upper massive quartzite member and a lower argillaceous quartzite member, similar to exposures elsewhere on the Gogebic Range. (Note: Due to time limitations, we will not visit exposures of the Palms Formation on this trip). The Ironwood Iron-formation consists of magnetic wavy-bedded and laminated cherty iron-formation and argillite, some of which is also magnetic. Mafic sills or flows lie in contact with the iron-formation. Unusual breccia units, the focus of field stops in this area, are associated with one of the sills.

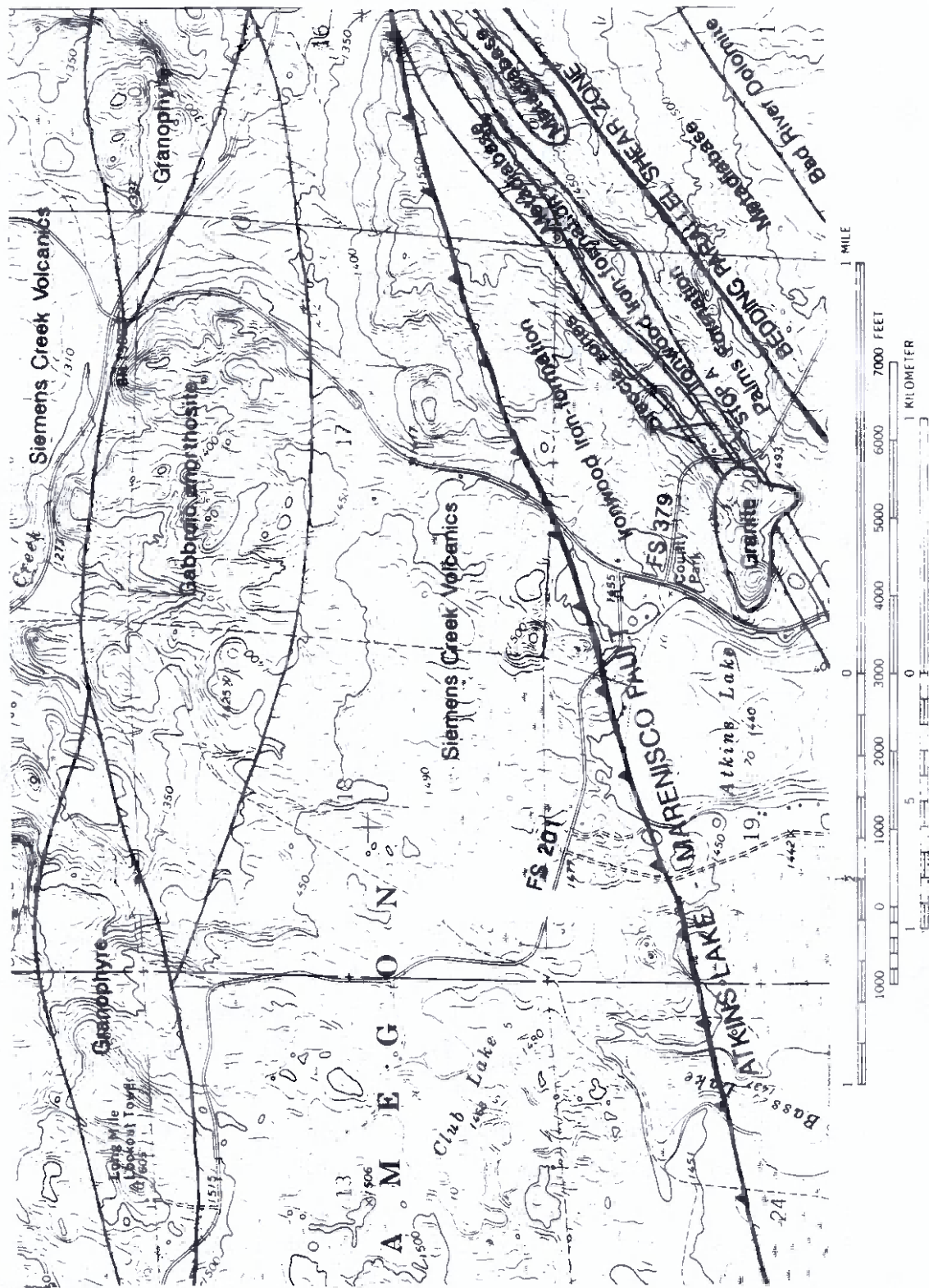


Figure 5-1. Part of the Grand View 7 1/2' quadrangle showing general geology and location of field trip stop A.

The breccia units contain angular to sub-rounded clasts, up to several meters in length, of banded siliceous sedimentary rocks in a matrix of fine-grained, massive igneous rock. The igneous rock is locally fine-grained near the margins where the breccia occurs but coarse-grained in the interior of the igneous body, where it ranges from fine-grained diabase to medium-grained gabbro (Feher and Flood, 1995). However, the mineralogy and, to some extent, the textures have been modified by Middle Proterozoic (Keweenaw) metamorphism. Clasts within the breccia are mainly siliceous argillite; a few are magnetic. Most clasts are long, tabular bodies. Bent or folded clasts are not uncommon, suggesting that the sediments were only semi-lithified when incorporated in the breccia. The argillite consists of alternating layers of recrystallized quartz, fine-grained chlorite, biotite or actinolite, garnet, and variable amounts of magnetite (Feher and Flood, 1995). The matrix is a fine-grained mafic rock with conspicuous vesicles around some clasts. The matrix is composed of medium-grained amphiboles and plagioclase laths, probably a relict ophitic texture (Feher and Flood, 1995). Plagioclase phenocrysts are present locally. Preliminary geochemical data indicates a tholeiitic composition for the matrix (Feher and Flood, 1995).

The origin of the breccia is problematic. The breccia is not a debris flow because the matrix is igneous and not fragmental: the clasts appear to have been incorporated into a magma. The presence of angular, and in places deformed, blocks of country rock within a magma argues for intrusion into semi-consolidated sediments at a shallow depth. The presence of breccia at both the upper and lower margins of the igneous body argues for a common origin for both breccia units, most likely the intrusion of a sill into wet, unconsolidated sediments, forming, perhaps, a peperite-like body.

Alternatively, the breccia and mafic rocks may represent an extrusive basaltic flow that incorporated blocks of the sediments over which it flowed, with subsequent concordant intrusion of a mafic sill separating the basalt flow into two parts. Features arguing for this origin are: 1) perfectly concordant contacts with thinly laminated sediments exposed at several outcrops, and the lack of any observed cross-cutting contacts; and 2) marked lithologic contrast between the breccia and undoubtedly intrusive sills of coarse-grained, massive, inclusion-free metadiabase within the same outcrops.

Regardless of whether the breccias are the result of an extrusive or a very shallow intrusive igneous event, they indicate that igneous activity was occurring contemporaneously with iron-formation deposition in the western part of the

Gogebic Range. The situation is similar to that of the eastern part of the Gogebic Range where the Emperor Volcanic Complex is interlayered with the Ironwood Iron-formation (LaBerge and Klasner, 1994). In both areas platformal sediments, including the Bad River Dolomite, the Palms Quartzite, and the Sunday Lake Quartzite in the east, are similar to equivalent rocks in the central part of the Gogebic Range. Igneous activity, however, becomes a significant feature during the time of deposition of the Ironwood Iron-formation on both ends of the Gogebic Range. Whereas igneous activity in the central part of the range is limited to deposition of a few minor ash beds (R. G. Schmidt, personal communication, 1995), igneous deposits on either end of the Gogebic Range indicate a far more volcanically active environment in these regions.

ATKINS LAKE FIELD STOPS (see Figure 5-2)

Our mapping of the sill and breccia units near Atkins Lake (Fig. 5-3) reveals a 70 m thick igneous body with breccia units up to 20 m thick along both the upper and lower margins.

STOP 5A-1. This outcrop consists of layered magnetic, greenish, actinolitic iron-formation with metadiabase, and a breccia zone with clasts of iron-formation in a fine-grained metadiabase matrix. Bedding orientations at this outcrop are consistent with orientations of bedding elsewhere in the area indicating that the outcrop is in place and not a raft of iron-formation entrapped in the igneous rock.

Walk northeast about 60 m to the top of a hill.

STOP 5A-2. At this location, large thin slabs of argillite lie in a matrix of locally vesicular metadiabase. The slabs have various orientations: some are contorted and folded, and one is rolled into a ball with a central layered portion surrounded by a concentric zone of smaller clasts in a metadiabase matrix. Slabs range from a meter or more in length down to a centimeter or less.

Walk east along the crest of the hill about 90 m to a north-northwest-trending valley. When in the valley follow the edge of the hill toward the south-southwest for about 60 m.

STOP 5A-3. Strongly magnetic, wavy-bedded, cherty iron-formation. Lenticular chert beds are up to 15 cm thick. This bedding style is typical of wavy-bedded iron-formation farther east along the Gogebic Range.

Walk north-northeast along edge of hill about 25 m.

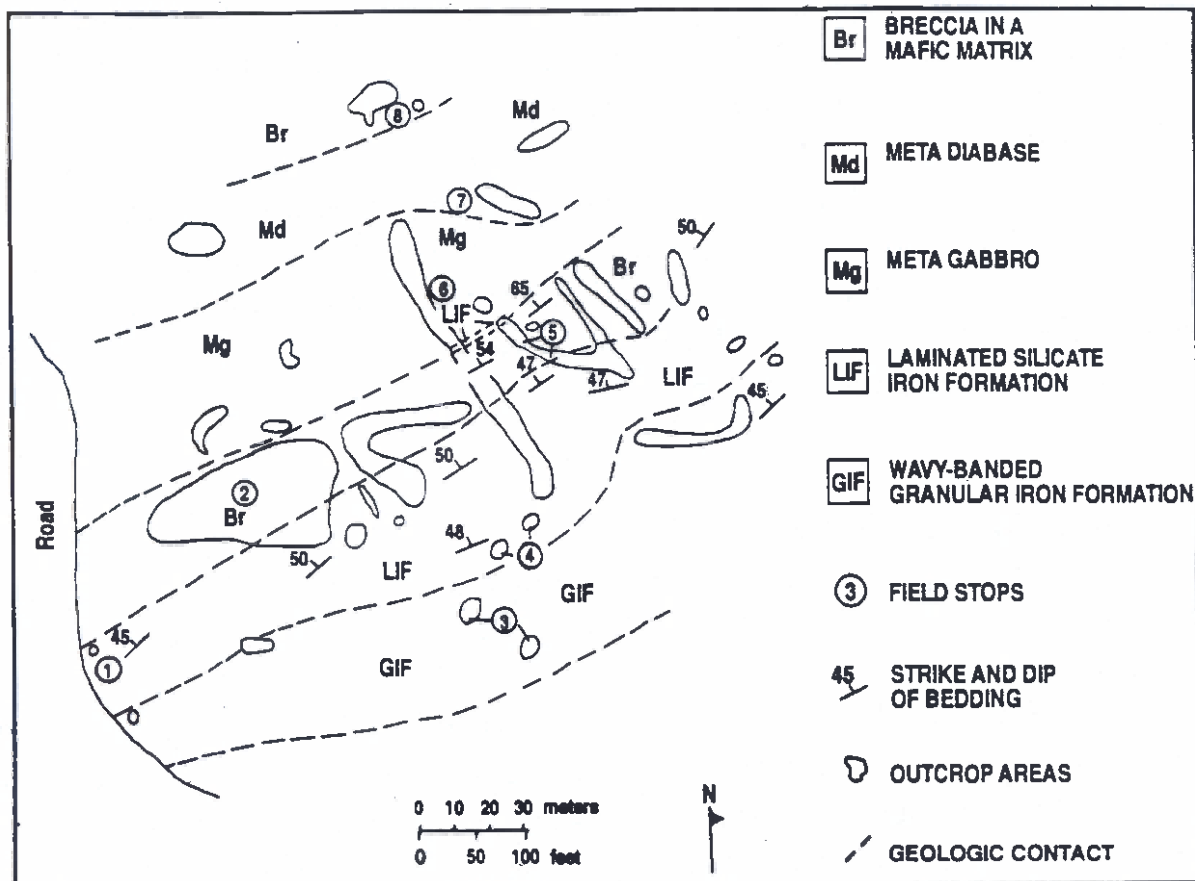


Fig. 5-2. Detailed map of the Atkins Lake area showing outcrops and geologic relationships among iron-formation, breccia units, diabase and gabbro. Unpublished map prepared by Klasner and LaBerge.

STOP 5A-4. Thin-bedded, non-magnetic chert-argillite or argillaceous iron-formation. Some cherty layers appear boudinaged. The argillite unit is about 50 m thick and grades into a more argillaceous unit with less cherty layers to the north.

Walk northeast about 55 m.

STOP 5A-5. The top of the argillaceous iron-formation layer (LIF on Fig. 5-2) is in sharp contact with the overlying breccia unit, in which argillite clasts are surrounded by a metabasalt matrix. Clasts in the breccia tend to become larger higher in the section (northward from the contact). Remnants of what appears to be a thin slab of iron-formation lie along the top of the breccia unit (Br on Fig. 5-2) at both stops 5 and 6.

Walk east-southeast about 10 m.

STOP 5A-6. Observe the southeast continuation of the thin slab of iron-formation that lies along the contact between the breccia on the southwest and massive metagabbro on the northeast.

Walk northeast about 50 m.

STOP 5A-7. Massive metadiabase/metagabbro.

Walk northwest about 35 m.

STOP 5A-8. Breccia. Clasts of mainly argillite in a vesicular metabasalt matrix. Argillite clasts are contorted.

PART B: PENOKEE GAP

Penokee Gap lies about 3.2 km southeast of Mellen, Wisconsin. It is a water gap in the Gogebic Range formed by the Bad River as it cuts across the Range. Easiest access to Penokee Gap is by a 1.6 km walk south along the railroad tracks that follow the Bad River.

The Penokee Gap area itself and the area just west of the gap (Fig. 5-3) provide abundant data on local stratigraphy and structure of the Gogebic Range. It is important to note, however, that the entire region was tilted steeply northward during the closing phase of Middle Proterozoic tectonism so that all Early Proterozoic and Archean structural orientations are not original orientations.

Strata at Penokee Gap consist of Archean gneiss overlain by Early Proterozoic Bad River Dolomite, Palms Formation with a thin-bedded, intensely sheared lower unit and a massive quartzitic upper unit, Ironwood Iron-formation and cyclically-bedded Tyler Formation, mainly consisting of slates and siltstones. A thin Middle Proterozoic (Keweenaw) diabase dike intrudes the Tyler Formation. More

TABLE 1: Summary of structural components and our interpretation of the tectonic evolution at Penokee Gap.

<u>EVENT</u>	<u>STRUCTURAL ELEMENTS</u>
D_A - Late Archean tectonism	S_A - Archean gneissic foliation S_S - Discrete, narrow shear zones parallel to gneissic foliation.
D₀ - Divergent tectonism associated with rifting of Archean continent. Formation of passive continental margin.	S₀ - Flat-lying beds unconformably deposited on passive continental margin in the Penokee Gap area.
D₁ - Convergent tectonism causing thin-skinned, north-verging folding and thrusting. Regional greenschist-grade metamorphism.	F₁ - Folds in S ₀ with L ₁ fold axes. S₁ - Axial planar foliation. L_B - Long axes of boudins.
D₂ - Continued north-verging thin-skinned deformation coaxial with D ₁ .	S₂ - Spaced fracture foliation. F₂ - Kink folds primarily in S ₁ with L ₂ fold axes.
D_M - Middle Proterozoic rifting of the continental crust and associated deposition of sediments and igneous activity.	Formation of the Midcontinent Rift. Intrusion of Mineral Lake Gabbro and, possibly, thermal metamorphism to form porphyroblasts at Penokee Gap.
D_K - Northward tilting of Archean and Early Proterozoic crust along listric thrusts during compressional phase of the Middle Proterozoic Midcontinent Rift.	Tilting resulted in formation of a northwest-dipping monoclinial sequence of Early and Middle Proterozoic strata and formation of north-northwest-striking, near vertical faults, including the formation of S _K joints and fractures.

information on the Early Proterozoic stratigraphy at Penokee Gap is covered above in the introduction. Table 1 and the discussions below summarize the structural and stratigraphic features observed in the Penokee Gap area and our interpretation of the tectonic events that formed them.

Late Archean deformation (D_A) resulted in the formation of the now near-vertical, easterly-striking gneissic foliation (S_A on Fig. 5-3) and parallel shear zones (S_S , not shown on Fig. 5-3). This structural fabric is not present in any of the overlying Early Proterozoic strata that were unconformably deposited on it. The time of formation of the shear zones is not known: the shears may have formed during the subsequent convergent phase of Early Proterozoic tectonism, but they have a different orientation than structural features in the Early Proterozoic strata.

Initial rifting of the Archean crust D_0 was accompanied by unconformable deposition of Early Proterozoic strata (S_0) on Archean gneisses. The rift event was followed by convergent tectonism. At Penokee Gap, two phases (D_1 and D_2) of deformation have been recognized in the Early Proterozoic rocks. D_1 involved thin-skinned, north-verging overthrusting that detached Early Proterozoic strata from the underlying Archean crust, folded it, and formed multiple bedding-parallel shear zones within the Early Proterozoic strata. Along the Gogebic Range, west of Upson, Wisconsin, numerous upward-ramping thrust faults (Cannon, personal communication, 1994; Klasner and LaBerge, 1994B) detach Early Proterozoic strata from the underlying Archean basement. Evidence for D_1 deformation is abundant in Early Proterozoic strata near Penokee Gap. In the iron-formation, folds (F_1) in bedding (S_0) define gently northwest-plunging fold axes (L_1 -- see Figs. 5-3, 5-4A and 4B). A field sketch of a recumbent fold in iron-formation is shown in Figure 5-5. Similarly, northwest - plunging folds occur in the Tyler east of the Penokee Gap fault as shown by the best great circle fit of poles to bedding (Fig. 5-4C). Penetrative axial planar foliation (S_1) to the F_1 folds in the Tyler Formation tends to be flat-lying (Fig. 5-4D). In the iron-formation S_1 foliation occurs as a fracture cleavage (see Fig. 5-5). Prominent shear zones with penetrative S_1 foliation occur at the base of the Palms and top of the Ironwood Iron-formation along the contact between the Tyler and Ironwood strata (Fig. 5-3). Near its lower contact, the Palms is thinly-bedded and units that contain fine-grained quartz in a sericite matrix are intercalated with finer grained, sericite-rich layers with ultra-fine quartz grains. Both the quartz-rich and sericite-rich layers are foliated and constitute a broad zone of shearing within the basal Palms: however, the more strongly-foliated, sericite-rich layers with ultra fine quartz are more intensely sheared. The S_1 shear foliation generally dips

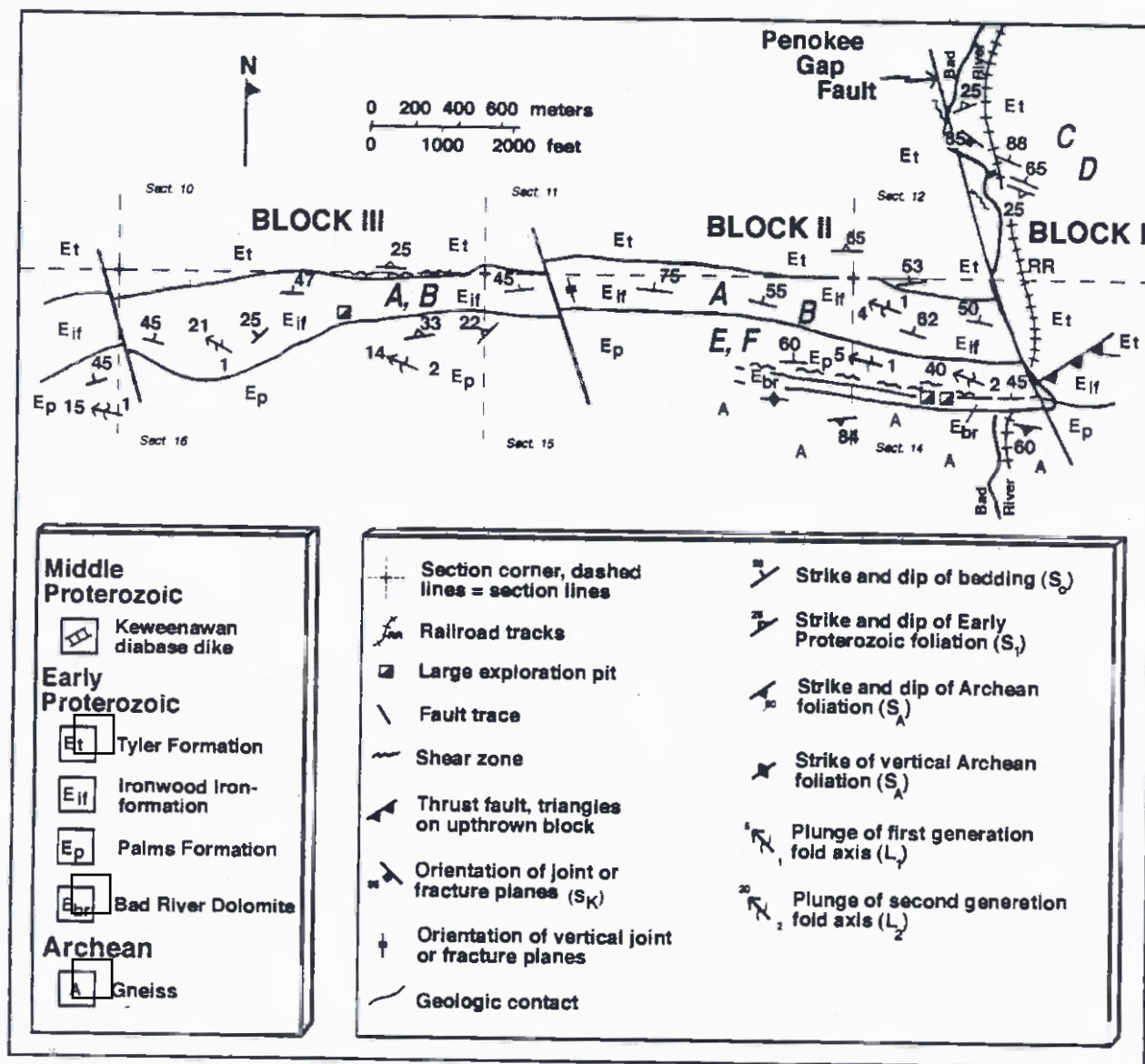


Fig. 5-3: Geologic sketch map of Penokee Gap area. Blocks I, II, and III represent segments of the Gogebic range near Penokee Gap that are delineated by northwest-trending faults. Capital letters indicate general location of data for stereoplots in Fig. 5-4.

northwest and strikes northeast (Fig. 5-4E), and constitutes the zone along which Early Proterozoic strata are detached from Archean basement. Likewise, a shear along the contact between the Ironwood and Tyler Formations constitutes another zone of detachment of the Tyler from underlying Ironwood Iron-formation. We consider that both formation of the shear zones, with accompanying foliation, and folding of the Tyler and older strata, with accompanying axial planar foliation, were part of the same fold-thrust event (D_1). That is, they were formed contemporaneously, and we therefore label both the shear foliation and the axial planar foliation as S_1 .

D_2 deformation was characterized by continued north-verging thrusting that resulted in further deformation along the S_1 shear zones in pre-Tyler Early Proterozoic strata. Within the lower Palms, prominent F_2 kink folds in S_1 shear foliation plunge gently northwest (Fig. 5-4F) and are prime examples of D_2 deformation. Although F_2 kink folds are not present in the Tyler, S_1 foliation in these strata tends to be flat-lying (Fig. 5-4D) compared to the generally northwest dipping S_1 foliation in pre-Tyler strata (Fig. 5-4E), and S_1 is folded about a nearly northwest-trending fold axis that is roughly coaxial with F_2 fold axes (Fig. 5-4F) in the sheared Palms Formation. Thus, we think that the Tyler was also affected by D_2 deformation.

Sense of structural vergence in the lower Palms shear zone is shown in Fig. 5-6A, B and C. Northward structural vergence along a small, south-dipping thrust fault (see Fig. 5-6A) in which the hanging wall moves toward the north as shown by a telescoped tabular layer (Fig. 5-6B) and by the sigmoidal curvature and offset of beds across the thrust fault (Fig. 5-6C).

Sense of structural vergence in the Tyler Formation at Penokee Gap is also toward the north where vertical to overturned graded beds of siltstone and slate stratigraphically face northward (even after post-Early Proterozoic (Keweenaw) rotation is removed). Although such sense of folding and rotation of bedding is not proof of overall structural vergence, the northward sense of rotation in Tyler strata agrees with kinematic indicators in the older strata, suggesting that D_1 and D_2 represent a continuum of northerly-directed overthrusting associated with the convergent phase of Early Proterozoic tectonism in the Penokee Gap area.

Evidence for the formation of the Middle Proterozoic Midcontinent Rift (labeled D_M in Table 1) lies mainly north of the Gogebic Range. Nevertheless, this event affected Early Proterozoic and Archean strata at Penokee Gap. For example, the final phase of tectonic evolution of the Midcontinent Rift involved a

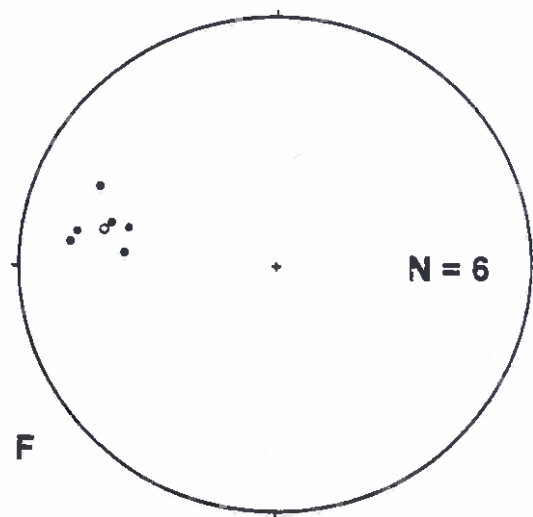
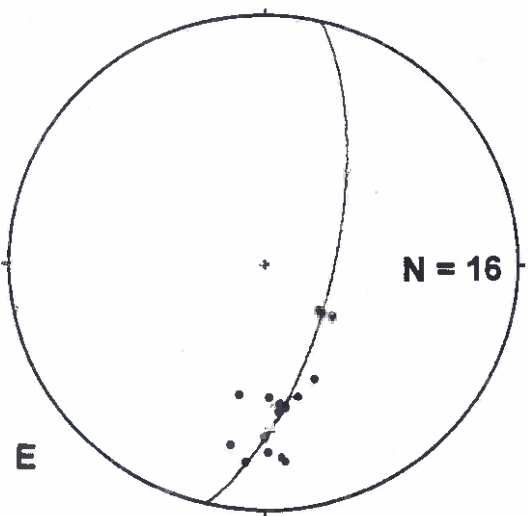
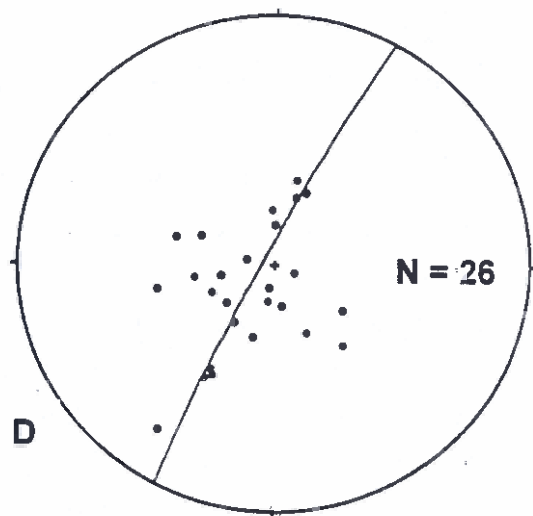
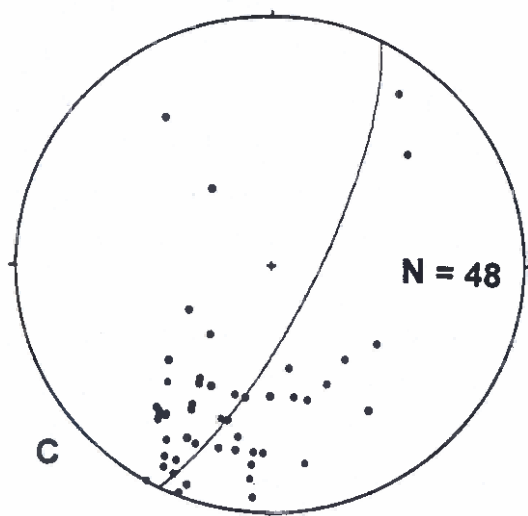
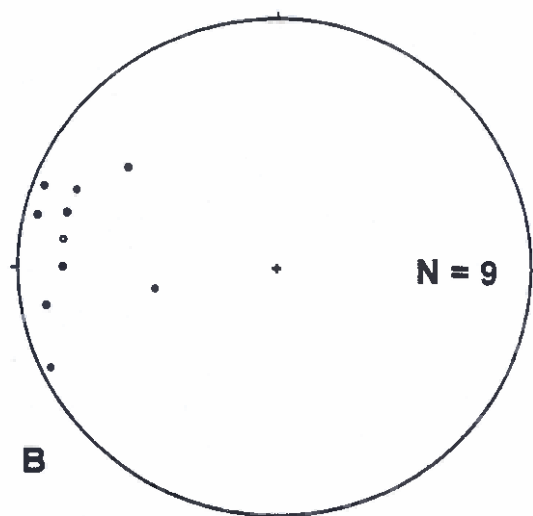
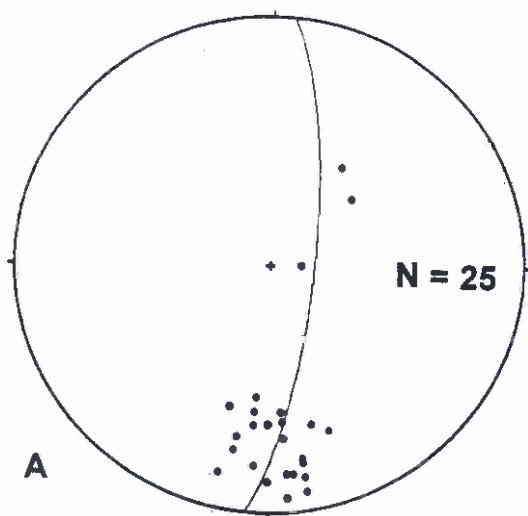


Fig. 5-4: Lower hemisphere, equal area stereo plots of structural features observed in the Penokee Gap area. Stereoplots are keyed to Fig. 5-3 by capital letters and according to blocks shown on Fig. 5-3. N = number of observations.

- A. Poles to bedding (S_0), primarily from iron-formation in blocks II and III. Best fit great circle defines an F_2 fold with an axis that plunges 14° , N 85° W.
- B. Orientation of L_2 axes of folds in bedding (S_0), primarily from iron-formation, in blocks II and III. Solid dots are field measurements, best fit orientation (open circle) defines a fold axis that plunges 17° , N 82° W.
- C. Poles to bedding (S_0) in block I, primarily from the Tyler Formation. Best-fit great circle defines an F_1 fold axis that plunges 15° , N 64° W.
- D. Poles to S_1 foliation in the Tyler Formation. Best-fit great circle defines an F_2 fold axis that plunges 04° , S 62° E.
- E. Poles to S_1 shear foliation in the lower Palms Formation of block II. Best-fit great circle defines the orientation F_2 fold that plunges 22° , N 77° W.
- F. Orientation of L_2 kink folds in S_1 shear foliation in the lower Palms, block II. Solid dots are field measurements, best-fit orientation (open circle) defines a fold axis that plunges 31° , N 75° W.

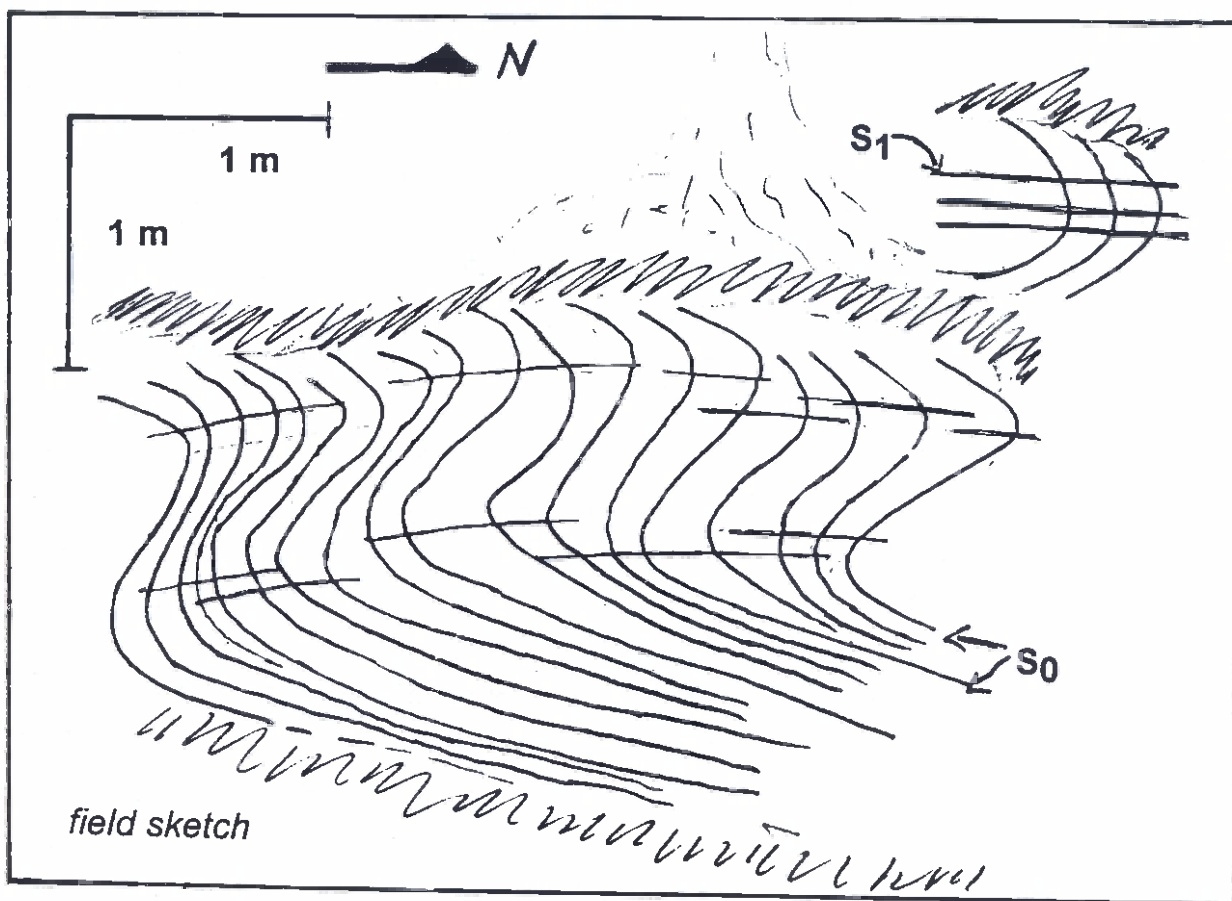


Fig. 5-5: Field sketch of recumbent fold in Ironwood Iron-Formation just west of Penokee Gap. S_0 = bedding. S_1 = axial planar fracture foliation.

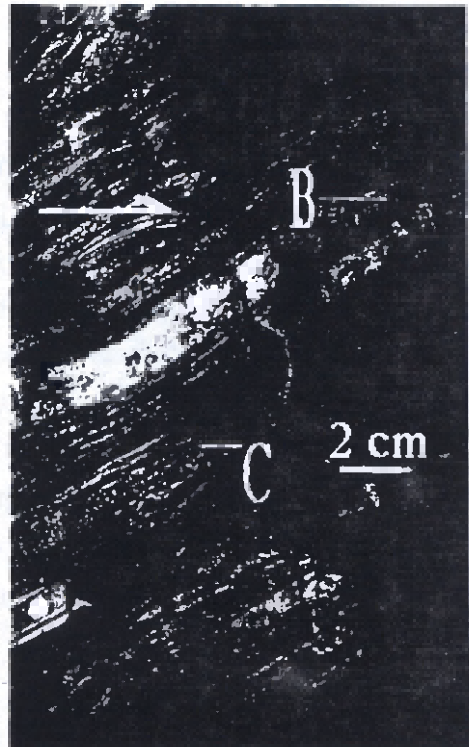
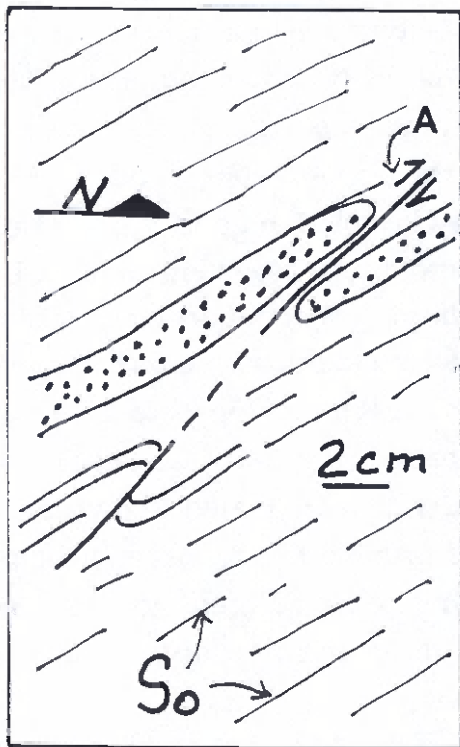


Fig. 5-6: Sketch and photograph from within S_1 shear zone in Palms Formation showing a small, north-verging (hanging wall toward the north) thrust fault.

A. Hand sketch of thrust fault and shortened layering.

B. Telescoped layer in photograph.

C. Sigmoidally folded and offset bed in photograph.

compressional stage (D_K on Table 1) that affected Early Proterozoic and Archean strata at Penokee Gap. During D_K deformation, rigid blocks of Keweenawan and Early Proterozoic strata, along with underlying Archean gneiss were tilted northward along north-dipping listric thrust faults (Cannon and others, 1993). This northward tilting created the north-dipping monoclinial sequence of Early Proterozoic strata that characterize most of the Gogebic Range today. In addition, our mapping at Penokee Gap and along the Gogebic Range to the west of the gap indicates that the rotated blocks are divided into discrete segments (blocks I, II and III on Fig. 5-3) that are bounded by steeply-dipping, north-northwest-striking fault. Besides the stratigraphic offsets across the inferred faults, numerous steeply-dipping, northwest-striking joints or closely spaced fractures (S_K) are also considered structural expressions of this phase of deformation (Fig. 5-3). The Penokee Gap fault is perhaps the most prominent of these faults, and it created a zone of weakness along which the Bad River ultimately flowed forming the Penokee water gap in the Gogebic Range. Steeply northwest-plunging folds adjacent to these faults are observed in outcrops of the Tyler and were most likely formed by drag along the faults.

Formation of the Midcontinent Rift may have also resulted in metamorphism of Early Proterozoic strata. As noted above, geologic evolution of the Gogebic Range near Penokee Gap involved at least two metamorphic events. The first was a dynamic, low-grade event that accompanied regional D_1 deformation. It is characterized by formation of platy minerals of sericite, biotite, and chlorite that lie along S_1 foliation. The second metamorphic event was most likely related to intrusion of the Middle Proterozoic Mineral Lake Gabbro, which lies about 1.6 km north of Penokee Gap. This event is characterized by a suite of metamorphic minerals including cordierite, garnet, chlorite, magnetite, and amphibole, some of which clearly overprint the D_1 deformational fabric and, thus, postdate it. Metamorphism, however, has not been studied in detail in this area and is worthy of further study.

PENOKEE GAP FIELD STOPS (see Fig. 5-7)

Walk south from County Road GG at Foster Junction along railroad tracks for about 1.3 km to the first large rock cut (Stop 1 on Fig. 5-7). Although most field stops will be in stratigraphic order from oldest to youngest, the first stop will be in the Tyler Formation, the youngest Early Proterozoic unit encountered on the walk southward along the railroad tracks to the older strata.

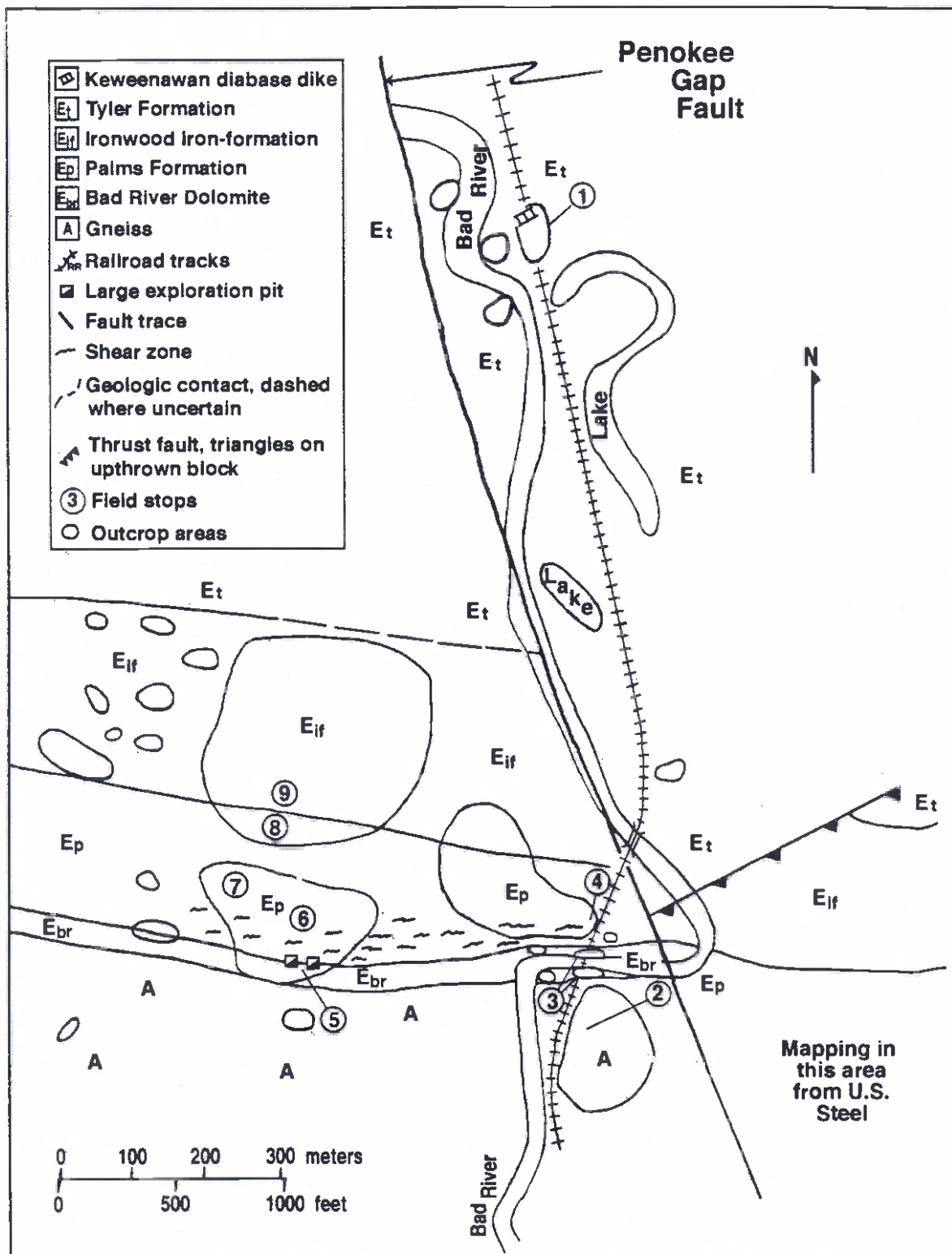


Fig. 5-7: Geologic field map of the Penokee Gap area showing location of field stops.

STOP 5B-1. Tyler Formation and Keweenaw diabase dike: SW 1/4, NE 1/4,
SW 1/4, S11, T44N, R3W

The Tyler Formation is exposed for roughly 100 m in a rock cut along the railroad tracks and consists of multiple beds of black carbonaceous slate and siltstone, subtly graded with stratigraphic tops toward the north. The Tyler strata consist of mostly silt-sized quartz grains in a carbonaceous matrix of biotite, chlorite and sericite. Small (~1 cm-diameter) cordierite porphyroblasts are dispersed throughout the siltstone/slate.

Bedding (S_0) in the Tyler at the northernmost end of the rock cut (east side of the railroad tracks) is overturned toward the north, but, progressing southward, most bedding dips steeply northwest and strikes northeast as shown in Fig. 5-4C. S_1 foliation lies at a large angle to bedding, is generally low-dipping to flat-lying, and is folded around a fold axis (L_2) that plunges gently to the southeast as shown by the great circle on Fig. 5-4D.

Isolated zones of steeply-dipping, north-northwesterly-trending fracture cleavage with steeply plunging folds (possibly drag folds) are found in outcrops especially along the west side of the railroad tracks. These, and other northwest-trending joints and fracture zones in the region, are considered part of a family of north-northwest faults in the Penokey Gap area that divide the Gogebic Range into individual segments (blocks I, II, and III on Fig. 5-3). They are interpreted as Middle Proterozoic SK structures formed during the final closing phase of formation of the Midcontinent Rift.

A 3 m-wide undeformed Keweenaw diabase dike is exposed at the north end of the rock cut on the west side of the railroad tracks.

Walk further south along the railroad tracks to the first rock cut on the east side of the tracks, just past the second bridge.

STOP 5B-2. Archean gneiss: SE 1/4, NW 1/4, S14, T44N, R3W

Amphibolitic to granitoid gneiss. The Archean gneiss forms the basement upon which overlying Early Proterozoic strata were deposited. The west-northwesterly-striking gneissic foliation dips steeply south. Parallel shear zones several centimeters wide occur in the gneiss. Although the strike of the gneissic and shear foliations is roughly parallel to the strike of Early Proterozoic structural features, there are no structures in the overlying Early Proterozoic strata that have similar dips. Other than postdating the gneissic foliation, the age of shear formation is not known. The shears may be Early Proterozoic in age.

Examine outcrops along both banks of Bad River, just beneath railroad bridge.

STOP 5B-3. Bad River Dolomite: NE 1/4, SE 1/4, NW 1/4, S14, T44N, R3W

Exposures on the south side of the Bad River are tan-colored dolomite with chert lenses and layers. Bedding is oriented N80°W, 45°N. Note that the dolomite rests directly on Archean gneiss. On the eastern end of the Range, the Sunday Quartzite occurs between the Archean basement and the Bad River Dolomite.

At this locality the Bad River Dolomite is about 20 m thick. The basal portion is mainly dolomite that has vague cross-bedding and possible stromatolites suggestive of shallow water deposition. The upper part of the Bad River Dolomite (on the north side of the river) is a variegated chert and chert breccia. This lithology is typical of exposures for about 15 km in the western part of the Gogebic Range. We interpret the cherty upper portion of the Bad River Dolomite to represent a residual concentration of chert layers and nodules formed by solution of dolomite during erosion that preceded deposition of the overlying Palms Formation.

We will examine another exposure of the cherty portion of the Bad River Dolomite about 515 m to the west at a later stop.

Walk north from the railroad bridge to the first large outcrop on the west side of the tracks.

STOP 5B-4. Palms Formation: SE 1/4, NW 1/4, NW 1/4, S14, T44N, R3W

Alternating beds (S_0) of silty to finer-grained, more argillaceous strata are present here. Thicker-bedded, more psammitic (silty) strata on the south end grade into finer-grained, more mica-rich strata on the north end of the outcrop, indicating that stratigraphic facing is to the north. Both the silty layers and mica-rich layers are penetratively foliated, but the argillaceous layers have less quartz as well as finer-grained quartz in an intensely-foliated sericite matrix. Penetrative foliation (S_1) is generally oriented N75°E, 60°NW, parallel to bedding. There are also numerous F_2 kink folds in S_1 foliation with L_2 axes that plunge about 16°, N87°W. These penetratively-foliated strata are considered part of a D_1 shear zone that was subsequently deformed by a D_2 deformational event to form the F_2 kink folds. We will take a closer look at the style of deformation in this shear zone to the west at Stops 6 and 7.

Walk west from the railroad bridge for approximately 515 m to a large talus slope along the steep south flank of an easterly trending hill. Climb the talus slope to two exploration pits near the top of the steep slope.

Note: This walk is strenuous, through thick brush and along steep hillsides with loose rocks that may roll down the hillside if disturbed. BE CAREFUL!

STOP 5B-5. Bad River Dolomite: SE 1/4, NW 1/4, NW 1/4, S14, T44N, R3W

The contact between the Bad River Dolomite and the Palms Formation is exposed in two exploration pits at this locality. Other exploration pits at this stratigraphic horizon are present on the eastern flank of Mt. Whittlesey, approximately 9 km to the east. Concentrations of magnetite are localized in conglomeratic zones at the top of the cherty breccia portion of the Bad River Dolomite. The magnetite is in a sandy matrix that surrounds angular and rounded chert fragments up to at least 70 cm long. Magnetite-rich zones do not extend up into the overlying Palms Formation. Therefore the magnetite concentrations appear to be pre-Palms. Along strike from the magnetite occurrence, the Bad River Dolomite consists of a spectacular breccia with some rounded and angular blocks of chert in a reddish hematitic matrix.

The local concentrations of magnetite in conglomeratic rocks at an unconformity argue for a paleo-placer origin along a braided stream course. However, much of the magnetite is euhedral, showing no indication of rounding that might be expected in a placer deposit. The euhedral magnetite may indicate that these are hydrothermal deposits, but no hydrothermal bleaching of surrounding rocks is evident, nor are any hydrothermal gangue minerals, such as quartz crystals, present. Alternatively, the euhedral magnetite crystals may be metamorphic in origin. Clearly, the origin of these peculiar magnetite occurrences is problematic. A mining company has analyzed these conglomerates and report that there are no precious metals in them.

Walk north-northeast about 80 m to the crest and north flank of the hill.

STOPS 5B-6 and 5B-7. Sheared, thin-bedded Palms Formation: SE 1/4, NW 1/4, NW 1/4, S14, T44N, R3W.

The purpose of these two stops is to illustrate our interpretation of the kinematics of deformation in a D_1 shear zone. Lithology at this stop is similar to that at Stop 4 where layers of silty Palms are intercalated with finer-grained mica-rich layers, both of which have penetrative S_1 foliation. At Stop 6 numerous small boudins are formed in thin quartz-rich layers that have orientations close to the orientation of S_1 foliation planes. Also, several F_2 kink folds with L_2 fold axes deform the S_1 shear fabric at this stop. At Stop 7, about 70 m northwest of the area of Stop 6, there is a small, south-dipping D_1 thrust fault (Fig. 5-6A) with telescoped layers (Fig. 5-6B) and a sigmoidal drag fold with offset beds (Fig. 5-6C).

The three structural elements: S_1 foliation, the boudinaged quartz layers and the small thrust fault provide information to interpret the kinematics of the shear

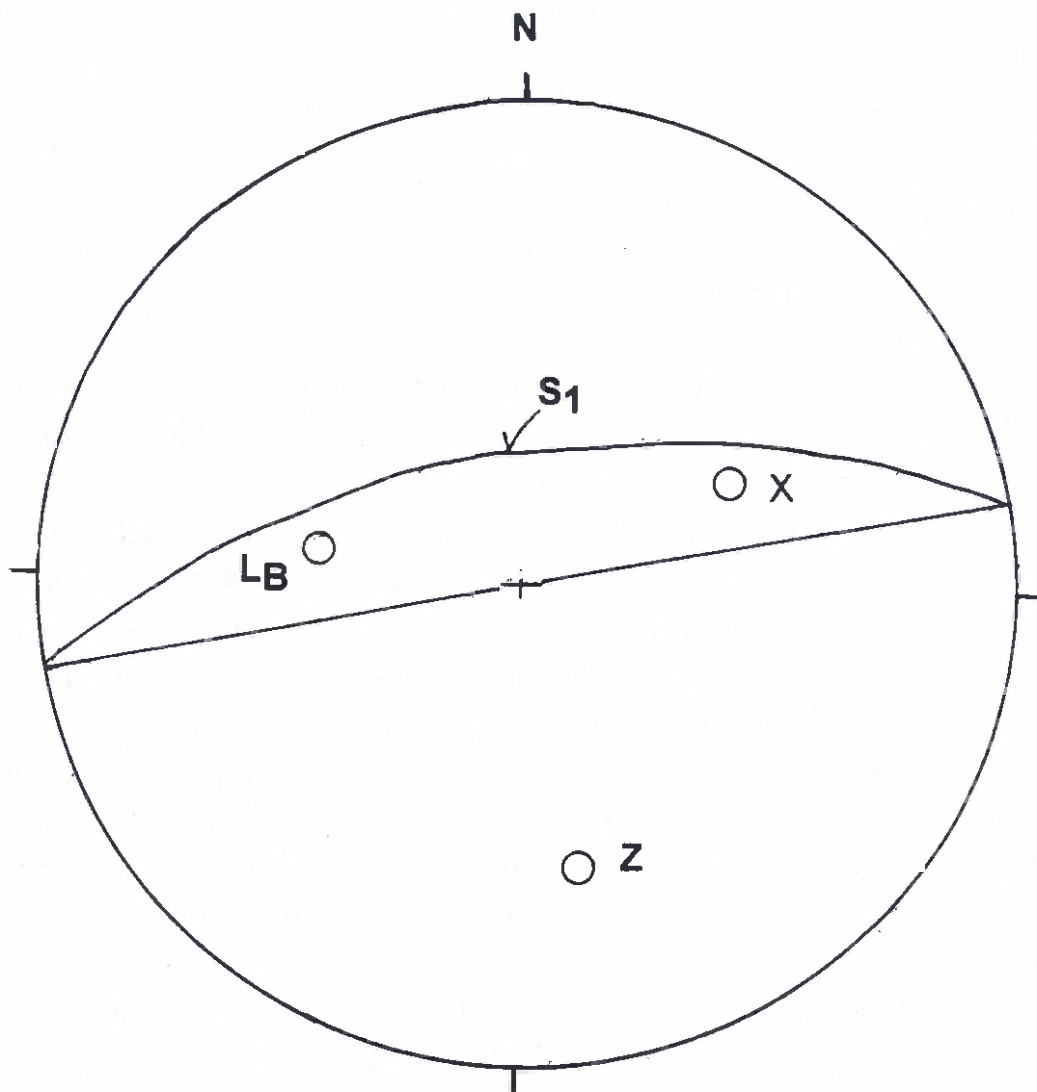


Fig. 5-8: Lower hemisphere, equal angle stereoplote showing the orientations of S_1 foliation (great circle), orientation of the long axes of boudins (L_B), orientation of the axis of stretching (X), and orientation of the axis of shortening (Z) in the sheared Palms Formation at Stops 5B-6 and 5B-7. See text for discussion.

zone. They are all considered to be D_1 structures. Penetrative S_1 foliation formed as a result of simple shear along an inferred, now steeply north-dipping to vertical shear zone that dips more steeply than S_1 foliation. Sense of movement in the shear couple must have been down to the north to form S_1 foliation. The general orientation of S_1 , based on field measurements, is shown by the great circle on the stereographic plot in Fig. 5-8. The boudins formed by stretching perpendicular to their long axes (L_B). L_B axes lie roughly in the S_1 plane and plunge about 45° toward the northwest. The orientation of the axis of stretching (X) in the shear zone can be estimated from the orientation of the long axes of the boudins. The stretch axis (X) also lies in the S_1 plane at right angles to the long axes of the boudins L_B , as shown on Fig. 5-8. That is the rocks in the shear zone were stretched along an axis (X) that plunges toward the northeast as shown on Fig. 5-8.

Finally, the small thrust formed in a region where S_1 foliation did not develop. The thrust lies parallel to the axis of maximum shortening (Z on Fig. 5-8). That is, it developed at right angles to S_1 foliation as shown on Fig. 5-8. Essentially, the small thrust accommodated shortening in rocks where S_1 foliation could not develop because of their composition (see the photograph, Fig. 5-6).

The absence of similar structures in the underlying Archean basement requires that Early Proterozoic strata must be detached from them. Thus, we interpret the intensely foliated strata in the lower part of the Palms Formation to represent a ductile shear zone that detached the Early Proterozoic from the underlying Archean gneiss.

Walk north, across a narrow valley for about 70 m to the next steep hill.

STOP 5B-8. Uppermost Palms Formation: SE 1/4, NW 1/4, NW 1/4, S14, T44N, R3W.

This stop illustrates the massive, thickly bedded quartzite typical of the upper part of the Palms. Both this massively bedded unit and the lower thin-bedded argillaceous units continue westward at least to the Atkins Lake area, where outcrop dies out.

The Palms Formation generally strikes $N80^\circ W$ and dips $65^\circ NE$ in this area, shows cross-bedding, and is a clean orthoquartzite. Westward from Penokee Gap the quartzite is a prominent ridge-former with numerous, excellent exposures.

Because of the exceedingly steep slopes and loose rocks it is dangerous for a large group to climb to the top of the bluff to view the quartzite-iron-formation contact.

Walk north up the relatively gentle slope just west of the quartzite bluff for

about 30 m.

STOP 5B-9. Ironwood Iron-formation: NW 1/4, NW 1/4, S14, T44N, R3W.

The Ironwood Iron-formation conformably overlies the Palms Formation from which it grades through a transition of several beds of granular iron-formation in the upper Palms. Locally, the base of the Ironwood has a thin (3 m) conglomerate zone, elsewhere, locally a thick zone of stromatolitic jasper occurs at the base of the iron-formation.

The Ironwood Iron-formation eastward from Penokee Gap was subdivided into five units based on bedding style and mineralogy (Hotchkiss, 1919). Our mapping west of Penokee Gap shows that the iron-formation consists of alternating units of granular, wavy-bedded iron-formation and laminated iron-formation. However, except in the first few hills of iron-formation west of Penokee Gap, the five-fold field division of Hotchkiss and U. S. Steel does not extend westward along the Gogebic Range.

The basal 75 m of the iron-formation is mainly a granular, wavy bedded unit with irregular lenticular chert beds separated by magnetite-rich layers that branch and recombine around the cherty lenses. This unit may represent the Plymouth member of Hotchkiss (1919). Overlying this wavy-bedded unit is about 30 m of mainly laminated, magnetic iron-formation with scattered intraformational breccia units. This lithology may represent the Norrie and Pence members of Hotchkiss (1919). The uppermost exposures are laminated, weakly to non-magnetic iron-formation. Some layers contain rounded elongate chert concretions in a somewhat argillaceous matrix, possible representing the Anvill member of Hotchkiss (1919). Near the northernmost exposures of the Ironwood Iron-formation, the recumbent fold (Fig. 5-5) mentioned above is present, and the contact with the Tyler Formation lies along the north flank of the hill supported by the iron-formation. There are no exposures of the Tyler in this region, however.

END OF FIELD STOPS

Walk east-southeast for about 820 m to railroad tracks along the Bad River, and then about 2 km north on railroad tracks to vehicles.

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