FIELD TRIP 8

Saturday, May 21

GEOLOGY OF THE PRE-CRETACEOUS WEATHERING PROFILE AND CRETACEOUS STRATA OF THE MINNESOTA AND COTTONWOOD RIVER VALLEYS

Leaders

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INTRODUCTION

Exposures created by the Minnesota River Valley and mining in that area provide opportunities to view products of the intense pre-Cretaceous chemical weathering episode, and the Upper Cretaceous rocks that immediately overlie them.

The intense weathering that predated the deposition of the Cretaceous strata acted on this part of the North American craton and produced a peneplain mostly covered by chemically mature residuum. The little relief that remained was mostly the result of unequal resistance to weathering related to parent rock composition or hydrologic properties. The relative resistance of minerals to chemical weathering is documented by the Goldich Mineral Stability Sequence (Goldich, 1938). Quartz is most resistant. The feldspars, some of which weather easily, break down and form kaolin clay. The resistant quartz, and the weathering product kaolin, dominate the mineralogy of the Cretaceous strata that overlie the weathered surface.

The Cretaceous strata that occur here were deposited on the eastern, shallow side of a foreland basin occupied by the Western Interior Seaway. The stable nature of the craton affords little accommodation for overlying sediment, and the strata we see likely represent a minor fraction of the sediment that was deposited and then mostly eroded by fluctuating sea levels and migrating shorelines. The Cretaceous strata represent interdistributary, distributary, and lacustrine brackish and freshwater environments of a fluvially-dominated delta plain and contain many rock and fossil types in a small area.

The weathering products and the sedimentary rocks are mined for use in brick-making and cement. Two of the clay mines we will visit represent the only non-aggregate and non-iron-ore mines developed in Minnesota since 1995.

In the vicinity of New Ulm and Courtland it is possible to see exposures that span over 2.7 billion years of geologic time. They include: 1. Minnesota's southeasternmost outcrops of Archean rock (granite), 2. Proterozoic (approximately 1.5 billion years) Sioux Quartzite, 3. Cambrian rocks (the Eau Claire Formation), 4. Cretaceous sediments, and 5. Quaternary outwash and till.

Note to participants: Our field trip stops will require walking along (or in) creeks and muddy locations (clay mines), so a good pair of water-resistant boots is recommended. Because three of our stops are mining operations, please be aware of your surroundings and use caution. Steep walls, water-filled pits, falling rocks, and loose, easily dislodged materials are potential hazards to you and others around you. Uneven ground and thick undergrowth will be encountered at the other stops, not to mention walking impediments such as domestic livestock and their output.

FIELD TRIP STOPS

DIRECTIONS: The mine entrance for Stop 8-1 is located about 0.5 mile east of Brown County Road 8 on Brown County Road 10.

STOP 8-1

Frohip Clay Mine (Northern Con-Agg Company)—an exposure of chemical weathering products of the Archean Morton Gneiss (Fig. 8.1)

Location: T. 112 N., R. 33 W., sec. 33, SW, NE
Morgan NE quadrangle

Description: The Frohip Mine exposes the products of intense chemical weathering of the Archean Morton Gneiss. The Morton Gneiss is a complex and severely deformed gneiss with four major components:

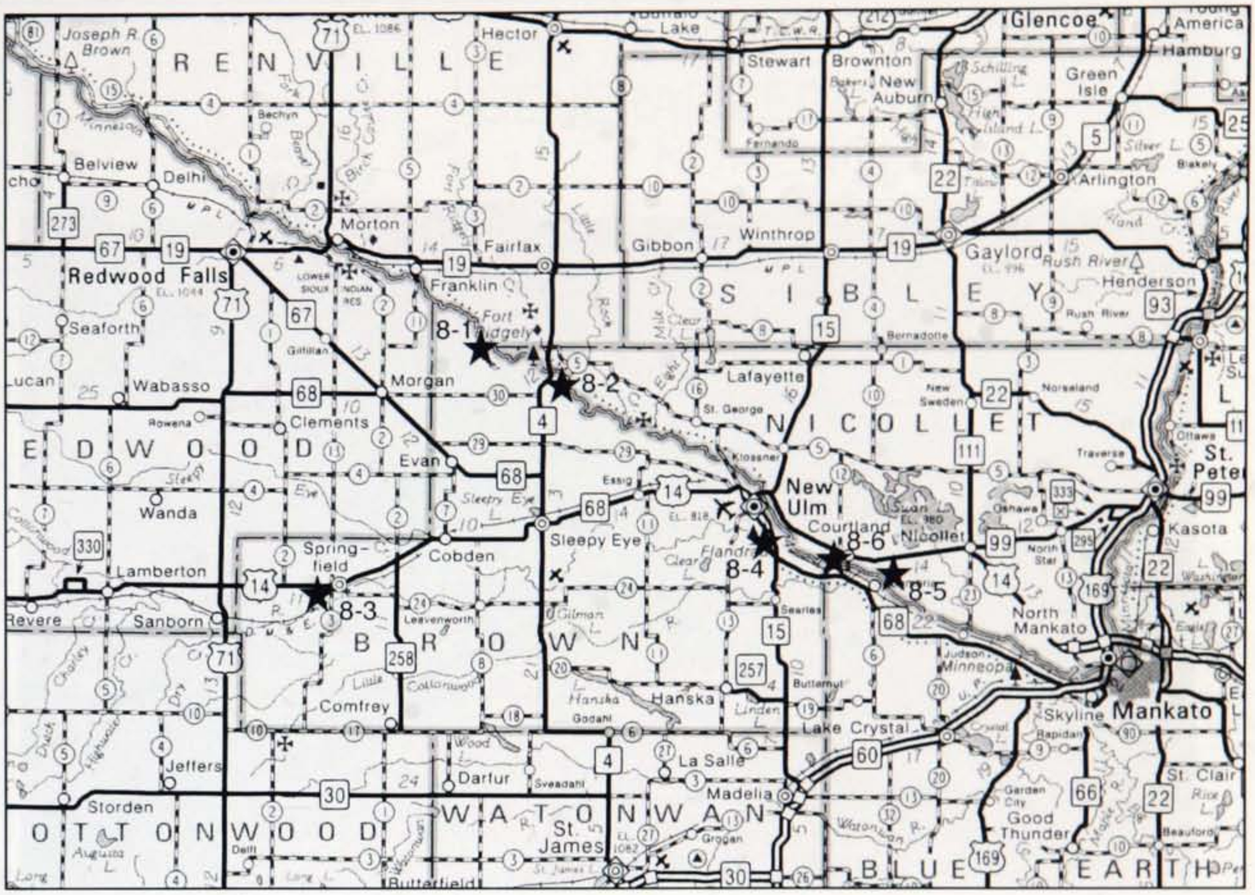

Figure 8.1. Map of field trip stops.

1. Gray tonalite that dates as old as 3,600 m.y.
2. Blocks and lenses of amphibolite (previously dikes)
3. Pink pegmatitic gneiss emplaced before major deformation (3,050 m.y.)
4. Pink granite, relatively undeformed (2,600 m.y.)

Rock in this complex weathers unevenly and produces a residuum of mixed composition. Relict gneissic textures can be seen in the residuum, and in the northwest corner of the pit a large, dark, amphibolite raft (weathered to chlorite-rich clay) can be seen (Fig. 8.2). A natural gamma log of a well across the road from the pit suggests that the weathering may extend as deep as 190 feet.

Most of the primary kaolin removed from this mine is low-grade, and is shipped to Iowa to make portland cement. The clay is attractive for cement making because its low alkali content minimizes the generation of highly alkaline waste products.

High-grade kaolin like that found in Georgia (white, bright, and containing very little grit) can be used in paper-making as filler, or as a coating for glossy paper. Kaolin is also used as filler for rubber, paint, pharmaceuticals, livestock feed, and in ceramics. Some of the better quality Minnesota River Valley kaolins have been tested for their paper-grade potential. With processing, these clays can be upgraded. However, substitute products such as precipitated calcium carbonate have taken market share from filler-grade kaolin.

NEXT: The mine entrance for Stop 8-2 is located about 0.25 mile east of State Highway 4 on Brown County Road 10.

STOP 8-2

Highway 4 Clay Mine (Minnesota Valley Minerals Company)—an exposure of weathering residuum and overlying Cretaceous strata

Location: T. 111 N., R. 32 W., sec. 17, NW, NE
Sleepy Eye NW quadrangle

Figure 8.2. The Frohip Clay Mine exposes products of chemical weathering of the Morton Gneiss.

Description: The exposure at the Highway 4 Mine changes with mining stages and water levels, but can include the top of the weathering profile on the Morton Gneiss, Late Cretaceous sandstone and claystone, and overlying glacial till and outwash.

The Cretaceous strata have been assigned to the Nishnabotna and Woodbury Members of the Dakota Formation by Toth (1996; Fig. 8.3). He divided these rocks into an upper argillaceous and organic-rich unit, and a lower arenaceous unit with abundant kaolin fragments and rip-ups. Sedimentary structures suggest a fluvial environment of deposition. Hajek (2002) interpreted alternating sandstone and claystone packages with original dip and carbonaceous drapes on foresets as an example of inclined heterolithic stratification (IHS). IHS is thought to be the result of large-scale bedform migration and the carbonaceous drapes are the result of a tidal influence on the river system (Thomas and others, 1987). Fossilized leaves, twigs, and pinecones have been recovered here.

The material mined here is used in the manufacturing of bricks at the Ochs Brick and Tile Company in Springfield. Some of the clay is also used for artwork.

NEXT: The mine entrance for Stop 8-3 is located about 1.25 miles south of U.S. Highway 14 and 1.5 miles west of the city of Springfield.

STOP 8-3

Springfield Mine (Ochs Brick and Tile Company)—an exposure of interdistributary deposits (Fig. 8.4)

Location: T. 109 N., R. 35 W., sec. 26, SE, NE

Sanborn NE quadrangle

Description: Early investigations of clays mined by Ochs Brick and Tile Company at a mine in Springfield were performed by Grout and Soper (1919). Subsequent investigations of the mine we are visiting today by Sloan (1964) and Hauck and others (1990) focused on the geology of this deposit. The deposit, first developed for clay production in the 1940s and 1950s, contains marine and non-marine Late Cretaceous shales, siltstones, and sandstones. The clays from this and other deposits in the

HWY 4 (BROICH) PROPERTY GENERALIZED STRATIGRAPHIC NOMENCLATURE

						UNIT	DESCRIPTION	APPROX. THICK.
PLEISTOCENE	(undivided)			Glacial and post-glacial deposits (0 to 150 ft)	BP			
CRETACEOUS	DAKOTA FM.	WOODBURY MEMBER	Non-marine sand/silt/clay	Organic sediments (10 to 20 ft)	B3	B3-3	Dark gray/black plastic shale	2-5 ft
						B3-2	Gray silt/shale +/- Fe-staining	2-5 ft
						B3-1	Light gray silt and very fine sand; can have a thin, very organic/lignitic base.	5-10 ft.
		NISHNABOTNA MEMBER	Argillaceous	Secondary kaolinitic sediments (20 to 30 ft)	B2	$B2\text{-}3_2$	Clay-rich kaolinitic shale (steel blue), +/- Fe	2-5 ft.
						$B2\text{-}3_1$	Silty light gray/mauve shale (blocky white)	2-5 ft.
			Arenaceous			? ↓ B2-2	Cross-bedded clays/silts/and sands (channel sequences) w/ abundant lignitic "trash"; channel bases lignitic/organic. Arenaceous > Argillaceous	6-10 ft.
						B2-1	Sandy sediments, w/ clay clast rip-ups, coarse quartz fragments scattered throughout, and some Fe staining. Basal secondary unit; arenaceous.	6-9 ft.
	MORTON GNEISS	SAPROLITE		Primary kaolin (> 75 ft in DH BRS96-04)	B1	B1-1	Upper primary; well-weathered. Faint gneissic textures; > 20% clay-sized, low alk.	15 ft.
						B1-2	Gneissic textures more prominent; < 20% clay-sized. Alkali content jumps (> 1%)	10 ft.
						B1-3 ↓ etc.	Coarsening w/ depth, gneissic textures increasingly prominent; feldspars fresher. Overall color darkening (greenish gray).	50+ ft.
ARCHEAN		BEDROCK		Crystalline bedrock	B0	n/a		

Figure 8.3. Stratigraphic nomenclature for the Highway 4 Clay Mine.

Springfield area have been used by Ochs since they were discovered by Adolph Casimir Ochs along the Cottonwood River in 1891 (Ochs Brick Company, 2005).

The Springfield Mine exposes up to 60 feet of sandstone, siltstone, and mudstone (Fig. 8.5). These sediments are commonly found in coarsening-upward sequences a few meters thick topped by a thin sandstone. Sideritic ironstone nodules and thin siderite-cemented beds are common. Brackish and freshwater fauna and abundant plant debris are present. This facies is thought to represent deposition in a normally quiet interdistributary bay environment where flooding and sedimentation caused by overbank and crevasse-splay events are also common (Fig. 8.6). The coarsening sediment size is associated with channel migration toward the site, and the abrupt return to finer-grained sediment is related to avulsion of the channel upstream. As with most Cretaceous strata in Minnesota, erosion and reworking are important components of the depositional history.

The uppermost strata at the mine contain shark teeth and fish vertebrae and may mark the transgression of marine conditions into the area.

Ochs is Minnesota's only brick producer, and makes over 60 million bricks annually. Ochs uses clay from its Springfield Mine and from two Minnesota Valley Minerals Company mines (the Highway 4 Mine of Stop 8-2, and another mine in Courtland at which we may make a brief stop—Stop 8-6). These three clay sources, combined with a state-of-the-art kiln allow Ochs to produce a range of brick colors (from dark brown to red to buff) that can satisfy many architectural needs and clients.

Figure 8.4. The Springfield Mine of the Ochs Brick and Tile Company.

A tour of the brick plant has been arranged, and will be followed by lunch.

NEXT: The parking area for Stop 8-4 is about 0.25 mile south of the Cottonwood River bridge on State Highway 15.

STOP 8-4

New Ulm—Cretaceous strata exposed along the Cottonwood River and its tributaries

Location: T. 109 N., R. 30 W., sec. 4, NW, NE

New Ulm quadrangle

Description: Exposures of Cretaceous strata along the Cottonwood River between Springfield and New Ulm show a transition from organic-rich, gray mudstones and siltstones similar to those at Ochs, to massive sandstones with gray clay interbeds (Fig. 8.7). Just west of New Ulm, the sandstones overlie pisolitic kaolin and kaolinitic sandstone, which in turn overlie bluish-green clays and silts. The pisolitic kaolin and kaolinitic sandstone appear to be transitional into the underlying bluish-green sediment, suggesting that the bluish sediments are a residuum of weathered Paleozoic rocks. The pisolitic interval represents development of a paleosol at the top of the reworked residuum. Most of the kaolin in the sediments was likely derived from weathering of igneous and metamorphic rocks (probably to the north and west), rather than the siliciclastic and carbonate Paleozoic rocks. These light-colored secondary kaolinitic sediments suggest that this portion of the Cottonwood River valley hosts clays and sediments that are older than many of the Late Cretaceous sediments we saw exposed further to the west near Springfield, and are part of a paleochannel system that carried a greater proportion of organic-poor materials derived from the developing kaolinitic saprolith. Sedimentary features, including cross bedding and foreset beds (inclined layers of a cross-bedded unit, specifically on the frontal slope of a delta), within the stratigraphically lower secondary kaolinitic sediments suggest a general west to east depositional direction, whereas the overlying Cretaceous sediments near New Ulm indicate an overall east to west depositional direction (Zanko and others, 1998).

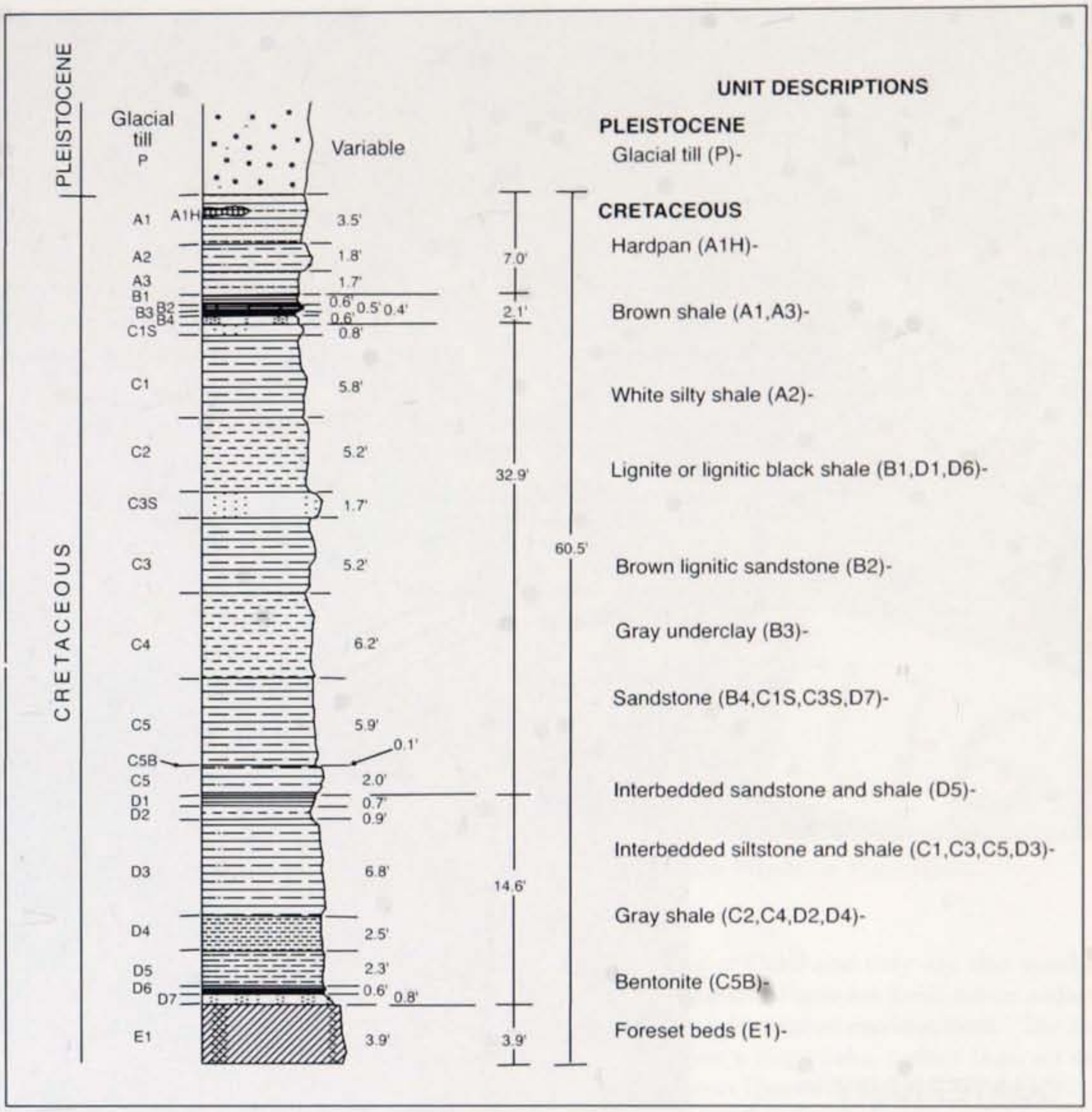

Figure 8.5. Stratigraphic column for the Springfield Mine exposure.

In this deeply cut valley of the Cottonwood River, the sandstone and claystone are well exposed along tributary creeks. Note the yellow sulfur staining. If the weather conditions and sunlight are right, gypsum crystals can be seen within the sediments, indicative of a non-freshwater environment of deposition, or of post-depositional incursion of marine waters associated with rising sea levels.

NEXT: From downtown Courtland go 0.5 mile east on Highway 14, then southeast on Nicollet County Road 25 approximately 0.45 mile, then south 0.55 mile and southeast slightly more than 2.0 miles on the unnamed dirt road to the creek crossing.

STOP 8-5

Courtland, Minnesota—Virgil Bruns property: Cretaceous sandstone, Ostrander Member of the Windrow Formation

Location: T. 109 N., R. 29 W., sec. 14, SE, SW Cambria quadrangle

Description: At this final stop, one of the area's best exposures of the Ostrander Member of the Windrow Formation (Fig. 8.8) is accessible in a deeply cut stream gully and tributary to the Minnesota River. The uppermost portion is well-cemented, conglomeritic, and contains highly polished quartz, chert pebbles, and kaolin aggregates indicative of a chemically weathered source area. The lower portion

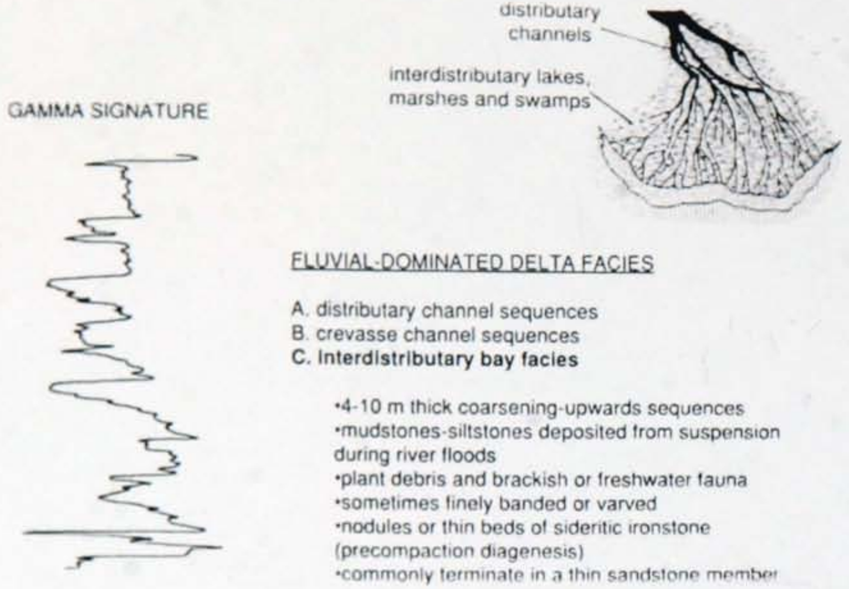

Figure 8.6. Depositional environment, sedimentary features, and natural gamma expression of the interdistributary bay facies of a fluvial-dominated delta system.

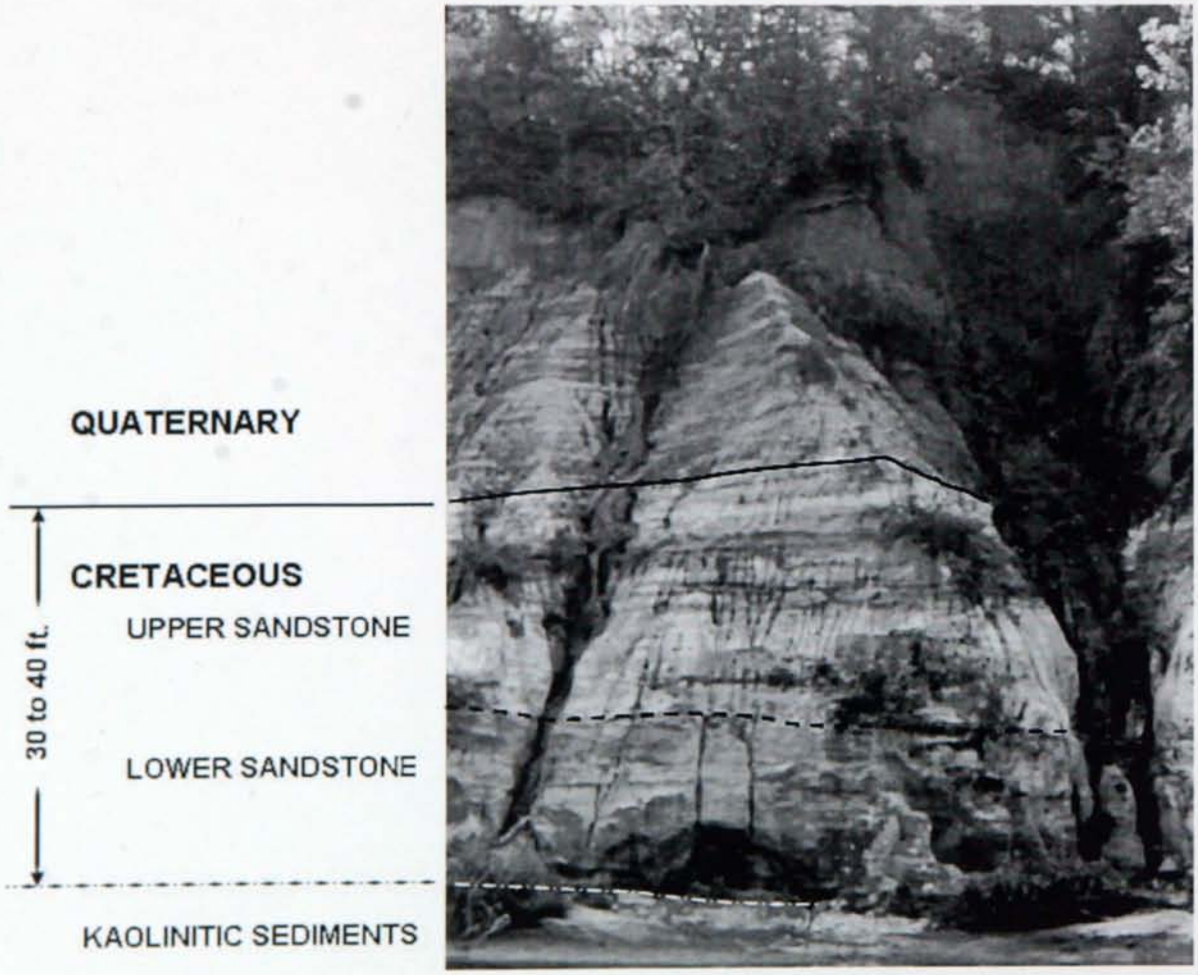

Figure 8.7. Typical exposure of Cretaceous strata in the New Ulm area.

Figure 8.8. The conglomerate facies of the Ostrander Member of the Windrow Formation.

is sandstone and contains excellent examples of large-scale foreset beds, cross-bedding, and Liesegang banding. These strata are illustrative of a higher energy depositional environment than we have seen thus far and the size of the bedforms suggests a distributary channel environment. The bedforms and the presence of chert indicate an east-to-west flow direction.

NEXT: From Courtland, go west 2.45 miles on Highway 14, and then south 0.5 mile on the mine road.

STOP 8-6 (optional)

Courtland Clay Mine (Minnesota Valley Minerals Company)—exposures of Cretaceous lacustrine and fluvial deposits

Location: T. 109 N., R. 30 W., sec. 1, NE, SW
New Ulm quadrangle

Description: Terrace deposits of glacial River Warren overlie clay-rich Cretaceous strata and weathering residuum developed on Cambrian rocks. The materials mined here are an excellent component for brick-making at Ochs and they are also used as art clays for ceramics. These are fresh-water sediments, deposited in a lacustrine environment. The lack of roots suggests a large lake, rather than an oxbow type. The Sioux Quartzite forms a topographic high within a mile of this site and differential weathering of the Sioux Quartzite and the Cambrian rocks that underlie the Cretaceous strata may be responsible for creating the basin in which these rocks were deposited. Leaf fossils are abundant, further suggesting a low energy environment of deposition. Importantly, fossil evidence for some of the earliest angiosperms (flowering plants) has been found at this location by Dr. David Dilcher of the Florida Museum of Natural History. At the base of the Cretaceous section is a thin, sandy horizon of coarse-grained quartz fragments and aggregates of white kaolin.

Below the Cretaceous deposits are sediments presumed to be Cambrian in age, with an uppermost aqua-colored sandy clay that grades into a red and green carbonate-bearing shale. This entire package most likely represents chemically weathered units of the Eau Claire Formation. These units are exposed in the surrounding area in stream valleys to the northwest, east, and south of the mine. Even though the Eau Claire Formation shows considerable lateral extent in the area, the presence of carbonates makes the units unsuitable for brick making.

The CK2 and CK1 units referred to in the stratigraphic column (Fig. 8.9) are used for brick-making. The CK2 unit fires to a lighter color than the underlying CK1 unit. CK1 can be further divided into an upper component (CK1-U) that contains relatively round Fe-rich concretions and numerous fossil leaf impressions, and a lower component (CK1-L) that contains relatively flat concretions and fewer leaf impressions.

The lowest Cretaceous unit CK0 is composed of interbedded Cretaceous clays and sands that commonly grade to basal quartzose sands. Fossil flora imprints are also found along silty/sandy partings in the bedding. It is thickest where the clay-rich (CK2 and CK1) units are thinnest, and vice versa. It appears to occur as a depositionally distinct unit rather than being transitional into CK2 and CK1. The CK0 unit is commonly underlain by a thin (<2 centimeters) layer of lignitic "trash" and/or a thin (<5 centimeters) conglomeritic layer that contains 0.5 to 1.0 centimeter quartz fragments suspended in a matrix of light gray sand and clay. The base of the conglomerate layer marks the Cretaceous boundary at the Courtland property.

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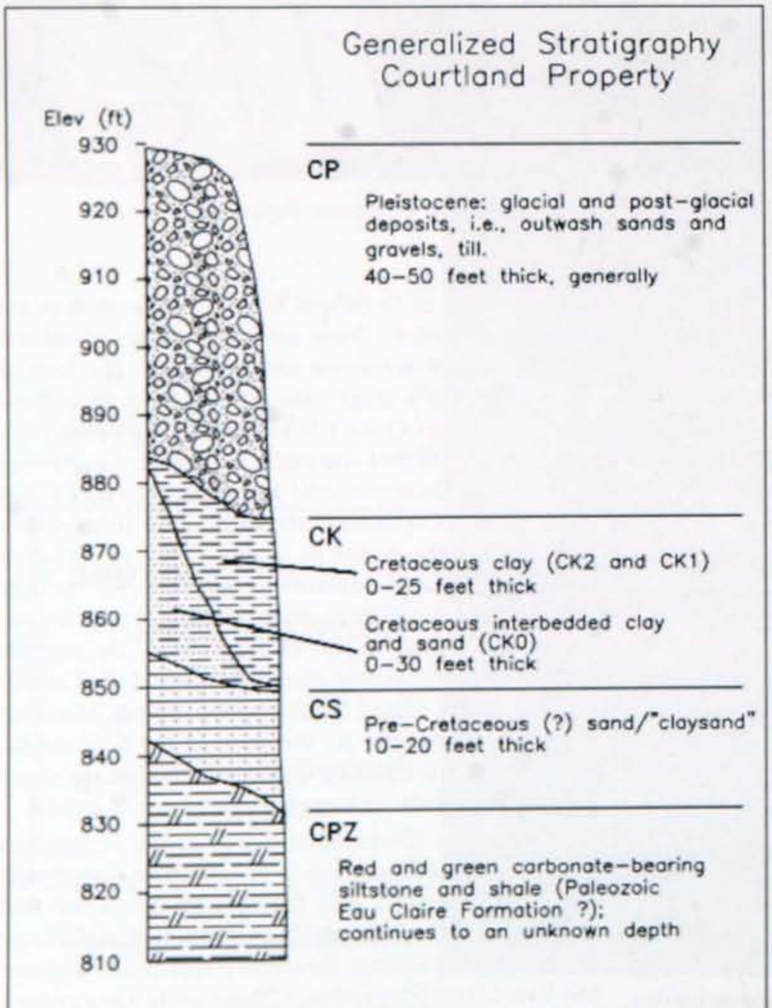

Figure 8.9. Generalized stratigraphy for the Courtland property.

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FIELD TRIP 9

Saturday, May 21 – Sunday, May 22

ARCHITECTURE OF AN ARCHEAN GREENSTONE BELT: STRATIGRAPHY, STRUCTURE, AND MINERALIZATION

Leaders

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INTRODUCTION

Archean greenstone belts are one of the world's premier geologic settings for hosting a variety of economically important mineral-deposit types. These deposits include high-grade iron ore, lode gold, volcanogenic massive sulfide, komatiite associated nickel, magnesite, and a number of others. The origin of these deposits is intrinsically linked to the architecture of the greenstone belt, namely the interrelationships between stratigraphy, structural setting, and multiple generations of hydrothermal fluids.

The Vermilion district of northeastern Minnesota contains one of the classic granite-greenstone terranes in the United States. This district comprises the south-central part of the Wawa subprovince of the Superior Province of the Canadian Shield, and has been broadly correlated with the Saganagons Assemblage of the Wawa subprovince in northwestern Ontario (Peterson and others, 2001; Peterson and Patelke, 2003). In Canada, the Wawa subprovince hosts numerous lode gold (for example the Hemlo and Renabie districts) and volcanic-hosted massive sulfide ore bodies (for example the Winston Lake, Willroy, Big Nama Creek, Willecho, and Geco deposits; Fyon and others, 1992). The Vermilion district is well known for its numerous, previously mined, massive hematitic iron-ore deposits. These iron deposits were discovered in the early 1880s, and virtually all subsequent exploration efforts in the region were targeted on similar iron-formation hosted hematite deposits. However, the discovery of world-class ore deposits in Ontario (the Kidd Creek volcanic-hosted massive sulfide deposit in 1964 and the Hemlo gold deposit in 1980) led to short periods of both base metal and gold mineral exploration in the Vermilion district. To date, no lode gold and/or volcanic-hosted massive sulfide ore bodies have been discovered in the Vermilion district, although abundant evidence exists that future exploration may result in the discovery of economically important deposits.

GREENSTONE BELTS

A strong debate continues on the origin, development, and architecture of Archean greenstone belts, particularly with regard to the roles of subduction, plume magmatism, rifting, diapirism, and autochthonous vs. allochthonous development (for example De Wit, 1998; Hamilton, 2003). Studies in the Superior and Slave Provinces of Canada indicate that strongly contrasting tectonic styles may have been in operation at the same time. For example, at circa (ca.) 2.7 Ga, large diapiric batholiths and synclinal greenstone keels may suggest that diapirism was an important tectonic process in the Slave Province (Bleeker, 2002). Stott (1997) proposed that the linear distribution of belts suggests that accretionary tectonics (such as plate tectonics) may have dominated in the Superior Province. Neither theory precludes the other, and in developing models for Archean tectonic evolution, no one model will be equally applicable to all areas. Hoffman (1990) presented a model of greenstone-belt formation via arc-trench progradation as an application of the principle of lateral and temporal equivalence, also known to sedimentologists as "Walther's Law." In this model (Fig. 9.1), accretion of the overriding plate in a subduction zone involves scraping of material off the downgoing plate and arc magmatism. The off-scraped material consists of sediment and the tops of igneous bathymetric highs (such as island arcs, remnant arcs, seamounts, oceanic island chains, submarine plateaus, fracture zones, and microcontinents).

We hope to portray our present understanding of the Vermilion district in the context of its geologic architecture—highlighting the interrelationships between stratigraphy, structure, and mineralization.

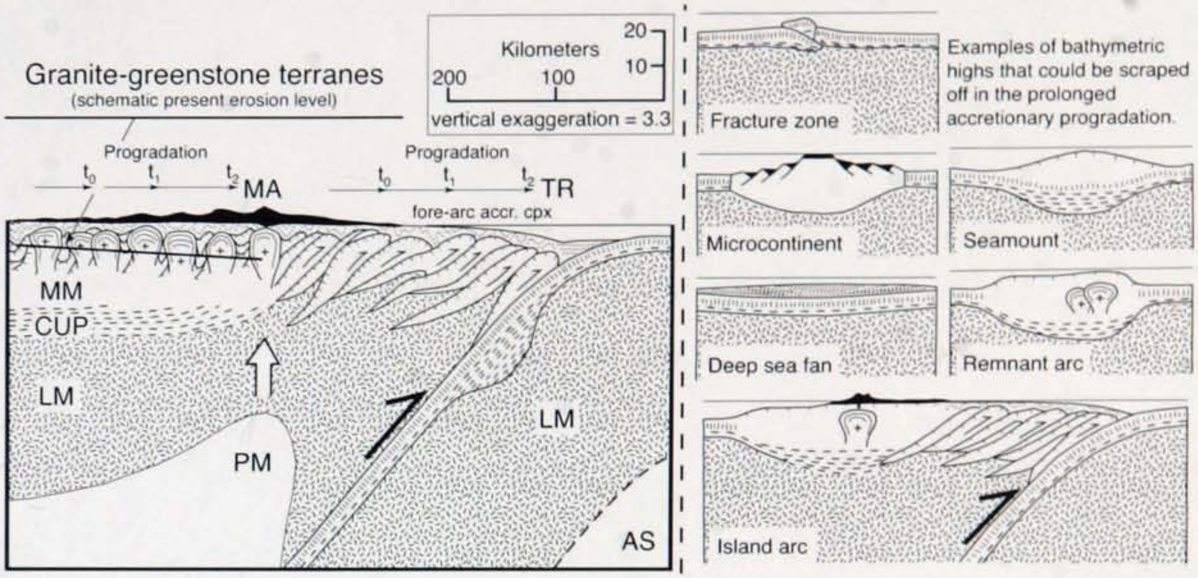


Figure 9.1. Prograding arc-trench model for the generation of granite-greenstone terranes. On the left is an idealized cross-section of an arc-trench system: AS, asthenosphere; CUP, cumulate underplating; LM, lithospheric mantle; MA, magmatic arc; MM, magma melting and mixing; PM, zone of partial melting; TR, subduction trench. Fore-arc accretion is achieved by scraping sediment and topographic highs off of the downgoing plate. Examples of bathymetric highs that could be scraped off and eventually form greenstone belts are depicted on the right; modified from Hoffman (1990).

The trip will revisit outcrops on which many historical discussions, a few of them heated, pertaining to the regional geologic setting of the Archean occurred. Many of the stop descriptions in this field trip have been modified from the Field Trip Guidebook of the 50th Annual Meeting of the Institute on Lake Superior Geology in 2004. In particular, volcanology and hydrothermal alteration within the Lower member of the Ely Greenstone was described in detail by Hudak and others (2004), gold mineralization north of the Mud Creek shear zone was described in detail by Peterson and Patelke (2004b), and classic outcrops in northeastern Minnesota were described by Jirsa and others (2004). Readers of this guidebook should review these documents for detailed descriptions.

REGIONAL GEOLOGICAL SETTING

Supracrustal rocks in the Vermilion district consist of volcanic-dominated stratigraphic sequences of the Wawa subprovince of the Superior Province of the Canadian Shield. Rocks of the Wawa subprovince in northern Minnesota are divided on the basis of stratigraphic and structural setting into the Soudan belt to the south, and the Newton belt to the north (Jirsa and others, 1992; Southwick and others, 1998). The boundary between these contrasting structural panels can be traced geophysically across the width of Minnesota, and was designated informally as the Leech Lake structural discontinuity (Jirsa and others, 1992). In the region west and north of the Soudan Mine, the Leech Lake structural discontinuity occurs along the Mud Creek shear zone (Hudleston and others, 1988), small segments of the Vermilion and Wolf Lake faults (Sims and Southwick, 1985), and the Bear River fault (Jirsa and others, 1992). The Soudan belt contains large, broad folds involving calc-alkalic and tholeiitic volcanic strata overlain by, and locally interdigitated with, turbiditic rocks. In contrast, the Newton belt consists of elongate, northeast-trending, and mostly northward-younging volcanic and volcaniclastic sequences. Volcanic rocks of the Newton belt differ from those of the Soudan belt in containing locally abundant komatiitic flows and peridotitic sills. The two belts are fault-bounded, and the relationship between stratigraphic units within each belt is largely conformable, although faults obscure contacts locally. In its eastern extension, the Soudan belt is continuous with the Saganagons assemblage in Ontario and terminates against the Saganaga pluton and Northern Light Gneiss. The Newton belt extends discontinuously eastward into the Shebandowan District of Ontario to form the Greenwater and Burchell assemblages. Intrusive rocks in both belts vary from gabbroic and felsic porphyries demonstrably related to volcanism, to large plutons emplaced post-tectonically. Both districts contain unconformable, Timiskaming-type sequences composed of calc-alkalic volcanic rocks, conglomerates, and finer-grained sedimentary rocks. A simplified regional geologic map of the Neoarchean terranes of northeastern Minnesota and adjacent Ontario is presented in Figure 9.2.

Lithostratigraphic units in the western Vermilion district include: 1. The Lower member (Lower Ely), Soudan iron-formation member, and Upper member (Upper Ely) of the Ely Greenstone, the Lake Vermilion Formation (including the informally named Britt and Gafvert Lake sequences), and the Knife Lake Group of the Soudan belt; 2. The Bass Lake sequence (Peterson and Jirsa, 1999) and the Newton Lake Formation of the Newton belt; and 3. Syn- to post-tectonic granitoid intrusions of the Giants Range batholith, and a suite of post-tectonic alkalic stocks and plutons. Contacts between the different units are typically conformable, although considerable overlap in time and space is documented between volcanic and sedimentary sequences (Southwick, 1993). Rock types associated with the lithostratigraphic units in the area are presented in Table 9.1.

STRUCTURAL GEOLOGY

Periods of generally north–south directed compression resulted in three major regional deformation events in the Neoarchean terranes of northern Minnesota. The earliest deformation event (D_1) produced broad, locally recumbent folds within the Soudan belt and major fault zones throughout the region. In the Newton belt, D_1 was accommodated by thrust imbrication of large crustal blocks, resulting in mainly northward stratigraphic facing. Field relationships indicate that uplift, faulting, and the deposition of Timiskaming-type clastic sedimentary sequences in local fault-bounded basins occurred late in D_1 deformation (Jirsa, 2000). A large, map-scale structure related to D_1 deformation in the western Vermilion district is the Tower–Soudan anticline, which is a west-plunging anticline within which the axis and plunge changes orientation along strike from nearly vertical in basalt, to shallow northeast-plunging in the western sedimentary rocks. Axial-planar cleavage associated with this early fold typically is lacking, although Hooper and Ojakangas (1971), Hudleston (1976), Bauer (1985), and Jirsa and others (1992) have described early cleavage (S_1) locally.

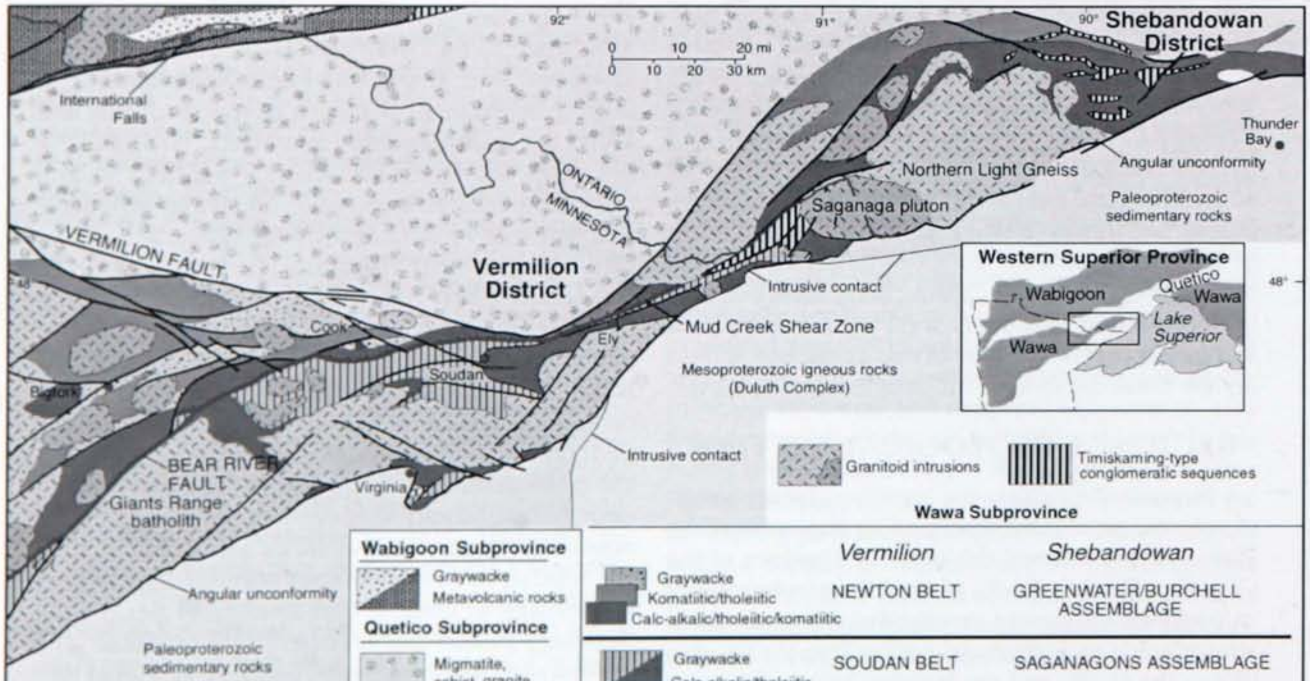


Figure 9.2. Simplified correlation map of Neoarchean assemblages across the U.S.–Canada border (modified from Peterson and others, 2001). Inset illustrates major subprovinces of the southwestern Superior Province.

Table 9.1. Lithostratigraphic units within the western Vermilion district (from Peterson and Jirsa, 1999).

Intrusive rocks	
Late intrusions	Plutons and stocks of syenite, monzonite, diorite, and lamprophyre
Vermilion Granitic Complex	Granite, schist, amphibolite, and schist-rich migmatite
Giants Range batholith	Granite, granodiorite, monzodiorite, and schist-rich migmatite
Supracrustal rocks	
Newton belt	
Newton Lake Formation	Tholeiitic and komatiitic basalt flows, intrusions, and clastic strata
Bass Lake sequence	Tholeiitic basalt lava flows, iron-formation, and felsic porphyries
Soudan belt	
Knife Lake Group	Graywacke, slate, conglomerate, and sheared equivalents
Lake Vermilion Formation	Graywacke, slate, dacitic tuff, and minor conglomerate
Gafvert Lake sequence	Dacitic to trachyandesitic lava flows, tuffs, and intrusions
Britt sequence	Tholeiitic basalt lava flows
Upper Ely Greenstone	Tholeiitic basalt lava flows and iron-formation
Soudan iron-formation	Layered cherty iron-formation, epiclastic rocks, and tuff
Lower Ely Greenstone	Calc-alkalic and tholeiitic basalt-rhyolite lava flows, tuffs, epiclastic rocks, and minor iron-formations

A second deformation event (D_2) associated with synchronous regional metamorphism resulted in foliation development and structures having largely dextral asymmetry. D_2 is constrained in the Vermilion district to the time period 2,674 to 2,685 Ma (Boerboom and Zartman, 1993), and between about 2,680 and 2,685 Ma in the Shebandowan district (Corfu and Stott, 1998). Because D_2 deformation affected all of the supracrustal rocks in the area and is reasonably constrained by geochronology, the regional foliation (S_2) can be used in the field to temporally relate other structural, intrusive, and deformation events. The relationship between S_2 fabric and shear structures indicates that most shearing occurred relatively late in the D_2 event. Major shearing that produced the Mud Creek and related shear zones is attributed to the late stages of D_2 dextral transpression.

Structures related to the third deformation event (D_3) include abundant northeast- and northwest-trending faults that dissect the stratigraphic assemblages. Named structures related to D_3 include the northeast-trending Waasa and Camp Rivard faults east of the Soudan Mine area, and the west–northwest-trending, crustal-scale Vermilion and related faults that form the Wawa–Quetico subprovince boundary.

ECONOMIC GEOLOGY

Since the mid-1860s, numerous mineral exploration programs have been conducted in the Vermilion district. Most of these exploration programs focused on identifying minable deposits of massive hematitic iron-ores, such as those mined between 1883 and 1962 in the Soudan iron-formation at the Soudan Mine. During the 1980s and early 1990s, subeconomic lode-gold mineralization was discovered in close proximity to the east–west-trending Murray shear zone, which dissects the Lower Ely member, and is in close proximity to the Mud Creek shear zone, which separates the Soudan belt from the Newton belt to the north. Four volcanic-hosted massive sulfide prospects occur within the Lower member of the Ely Greenstone, and occur in close proximity up-section from a semiconformable quartz-epidote alteration zone that extends for at least 19 kilometers along strike in the north limb of the Tower–Soudan anticline (Peterson, 2001). These volcanic-hosted massive sulfide prospects include the Skeleton Lake prospect (drilled by Exxon, 1972), the Eagles Nest prospect (drilled by Newmont, 1988), the Fivemile Lake prospect (drilled by Teck, 1994), and the Purvis Road prospect (drilled by Rendrag, 1999). Recent studies of these three types of mineral deposits in the Vermilion district link stratigraphy and structure, and thus help unravel the architecture of the greenstone belt. Brief descriptions of hematite, volcanic-hosted massive sulfide, and lode-gold mineralization in the field trip area are presented below.

Origin of massive hematite from algoma-type iron-formation

Most iron ores mined today comprise the iron oxide minerals magnetite, Fe_3O_4 (72 percent iron); hematite, Fe_2O_3 (70 percent iron); goethite, $Fe_2O_{3(s)} \cdot H_2O$, (63 percent iron); and limonite, a mixture of hydrated iron oxides (up to 60 percent iron). The world's most important iron-ore resources occur in iron-rich sedimentary rocks (20 to 40 percent iron) known as banded iron-formations, which occur on all continents and are almost exclusively of Precambrian age. In many iron-mining districts, such as the Mesabi range of northern Minnesota, the banded iron-formations are mined as iron ore with the iron content concentrated into pellets (~65 percent iron) in large on-site facilities. In other districts, such as the historic Vermilion range of the Soudan Mine area and the Hamersley district, western Australia, the banded iron-formations are the source rocks for large, natural high-grade concentrations of iron that typically occur as bodies of massive hematite and/or hematite-goethite with less than 60 percent iron.

The origin of these important natural concentrations of iron minerals remains highly debated. The iron atoms in hematite are all Fe^{3+}, whereas in magnetite they are comprised of two Fe^{3+} and one Fe^{2+} atoms. Therefore, the transformations of magnetite to hematite, or hematite to magnetite, in Fe-conservative systems is always a redox reaction, with the Fe^{2+} atoms in magnetite oxidized to Fe^{3+} atoms, or the Fe^{3+} atoms in hematite reduced to Fe^{2+} atoms, by reactions such as:

$$2Fe_3O_{4(mt)} + 1/2O_{2(g)} \rightarrow 3Fe_2O_{3(hm)} \qquad (1)$$

and

$$3Fe_2O_{3(hm)} + H_{2(g)} \rightarrow 2Fe_3O_{4(mt)} + H_2O \qquad (2)$$

In the past, the study of the transformation of magnetite to hematite (1) and conversely hematite to magnetite (2) in natural systems has largely focused on these reactions, which require either an oxidizing or reducing agent and an occurrence under specific redox environments. Because virtually all of the known concentrations of high-grade iron ores are hematite-dominant, the exploration for such deposits has concentrated on supergene enrichment of magnetite-rich banded iron-formations. In this model, magnetite-rich banded iron-formations are uplifted and subjected to weathering under oxygenated conditions to form goethite-rich ores, and are subsequently buried and metamorphosed to hematite-rich ores (Morris, 1985). Problems with this model have recently been discussed by Ohmoto (2003) for the Tom Price Mine of the Hamersley district, and are applicable to the origin of the massive hematite ores of the Soudan Mine. Ohmoto (2003) has proposed an alternative mechanism for the transformation of magnetite-rich iron-formations to massive hematite ores by the acid-base reaction:

$$Fe_3O_{4(mt)} + 2H^+ \rightarrow Fe_2O_{3(hm)} + Fe^{2+} + H_2O \qquad (3)$$

Similar to most acid-base reactions, reaction (3) would be most efficient at high temperatures, and such hydrothermal fluids are capable of leaching silica as well as Fe^{2+} from magnetite. In addition, the conversion of magnetite to hematite by reaction (3) produces a volume decrease of 32 percent, greatly increasing permeability of the rocks, which would facilitate further water-rock reactions and enhance conversion of banded chert-magnetite to massive hematite.

Volcanic-hosted massive sulfide-associated volcanic and hydrothermal alteration processes in the Lower member of the Ely Greenstone

Geologic mapping by Peterson (2001) has indicated the presence of a regional, semiconformable, quartz-epidote alteration zone extending for at least 19 kilometers along strike within the Lower Ely member along the north limb of the Tower–Soudan anticline. This type of alteration is a common feature in many Archean volcanic-hosted massive sulfide camps (such as the Noranda [Gibson, 1989] and Snow Lake [Skirrow and Franklin, 1994]), and is attributed to silica- and calcium-dumping that occurs in the deep, sub-seafloor as downwelling hydrothermal fluids are heated to temperatures in excess of 350° C (Franklin, 1986, 1993). Semiconformable alteration zones associated with volcanic-hosted massive sulfide systems are generally much larger in area than their associated mineralization, and therefore provide exploration geologists regional areas in which to concentrate more detailed, follow-up field mapping, geochemical studies, and geophysical surveys for identifying volcanic-hosted massive sulfide targets.

The composition and distribution of hydrothermal alteration mineral assemblages in the Lower Ely member are similar to that described in major lava-flow dominated volcanic-hosted massive sulfide mining districts worldwide (such as the Noranda Camp in Quebec; Morton and Franklin, 1987; Franklin, 1996; Gibson and others, 1999; Hudak and Morton, 1999; Peterson, 2001; Hudak and others, 2002a, b).

Results of recent studies in the Vermilion district indicated that not only are the compositions and geometries of the regional alteration mineral assemblages identical to those present in many lava-flow dominated massive sulfide mining districts, but also that detailed alteration mineral chemistries (Hocker and others, 2003) are consistent with those associated with the volcanic-hosted massive sulfide ore deposits in these mining camps. These two observations suggest that the processes that formed the alteration mineral assemblages in the Lower Ely member were similar to those that formed equivalent alteration zones in well-established volcanic-hosted massive sulfide mining camps.

A general genetic model for the formation of volcanic-hosted massive sulfide deposits and associated hydrothermal alteration zones, as presented by Franklin and others (1998), requires convective metalliferous hydrothermal fluid generation in the sub-seafloor environment via heating of down-welling seawater and leaching of metals from the enclosing volcanic and sedimentary strata (Fig. 9.3). The size of a convective hydrothermal system is a function of the abundance of heat in the upper two kilometers of the sub-seafloor crust (Franklin, 1996; Franklin and others, 1998). The intrusion of hypabyssal synvolcanic dikes and/or sills into the shallow sub-seafloor may vigorously enhance the dynamics of convective hydrothermal cells (Campbell and others, 1981). On reaching a critical reaction temperature of ~350° C, sustained acid pH in the hydrothermal fluid (evolved fluid) is achieved, and metals are leached from the rocks into the evolved fluid via primary mineral breakdown by calcium metasomatism, silicification, and hydrolysis reactions (Seyfried and others, 1999). In basalt-dominated systems (such as that in the Lower Ely member), leaching-related alteration of mafic "source" zones (lower semi-conformable alteration) forms a mineral assemblage composed of albite-epidote-zoisite/clinozoisite-actinolite-quartz. These zones are variably metal-depleted, and are characterized by patchy silicification and epidotization associated with areas metasomatically enriched in silica and calcium.

In lava-flow dominated stratigraphic sequences, regionally confined discordant "pipe-like," and more regionally extensive "semiconformable" alteration zones are present (Morton and Franklin, 1987). The "pipe-like" semi-conformable alteration zones are closely associated with zones of cross-stratal permeability (for example synvolcanic fault zones), and are characterized by well-defined, vertically extensive alteration zones containing anomalous abundances of sericite, chlorite (both Fe- and Mg-rich varieties), actinolite/ferroactinolite, quartz, pyrite, and locally, chalcopyrite and/or pyrrhotite. Semiconformable alteration zones extend for several kilometers to tens of kilometers in the rocks stratigraphically beneath and adjacent to volcanic-hosted massive sulfide mineralized horizons (Santaguida and others, 2002a, b). In mafic-dominated volcanic environments, such alteration typically is associated with regional zones of spilitization (an alteration assemblage composed of albite + quartz + Mg-rich chlorite ± sericite), silicification (quartz ± albite), and epidote-quartz alteration (epidote + quartz ± actinolite ± carbonate; Morton and Franklin, 1987; Gibson and others, 1999; Santaguida and others, 2002a, b). Regional semiconformable alteration zones in felsic rocks in volcanic-hosted massive sulfide producing camps such as Noranda (Quebec) or Sturgeon Lake (Ontario) typically comprise extensive zones of spilitization, silicification, and sericitization (sericite + quartz ± Mg-rich chlorite; Morton and Franklin, 1987; Gibson and others, 1999).

Both discordant and semiconformable alteration zones have been discovered in the Vermilion district (Table 9.1), and have been described by Hudak and Morton (1999), Odette and others (2001b), Peterson (2001), and Hudak and others (2002b). Semiconformable alteration zones in the Lower Ely member are dominated by mineral assemblages containing various proportions of quartz, epidote, zoisite/clinozoisite, Fe-chlorite, Mg-chlorite, actinolite, ferroactinolite, sericite/pyrophyllite, and albite. Odette and others (2001a, b) and Hudak and others (2002b) have shown via mass balance analysis that semiconformable quartz + epidote ± actinolite ± albite ± chlorite alteration mineral assemblages in the Fivemile Lake area are metasomatically enriched in calcium and silica, and depleted in base metals (copper and zinc) by 50 to 90 percent. Pipe-like, northeast-trending, disconformable alteration zones in the Lower Ely member are largely composed of iron-rich chlorite, sericite/pyrophyllite, actinolite and/or ferroactinolite. Pipe-like alteration zones that have been mapped up-section have to date not led to the discovery of economically significant volcanic-hosted massive sulfide deposits, but have been instrumental in locating potential base-metal sulfide-bearing stratigraphic horizons and localized chemical exhalites.

Lode-gold ore deposit model and gold in the Vermilion district

The brief description of Archean lode-gold deposits that follows is presented as both a basic reference and also to highlight the important features

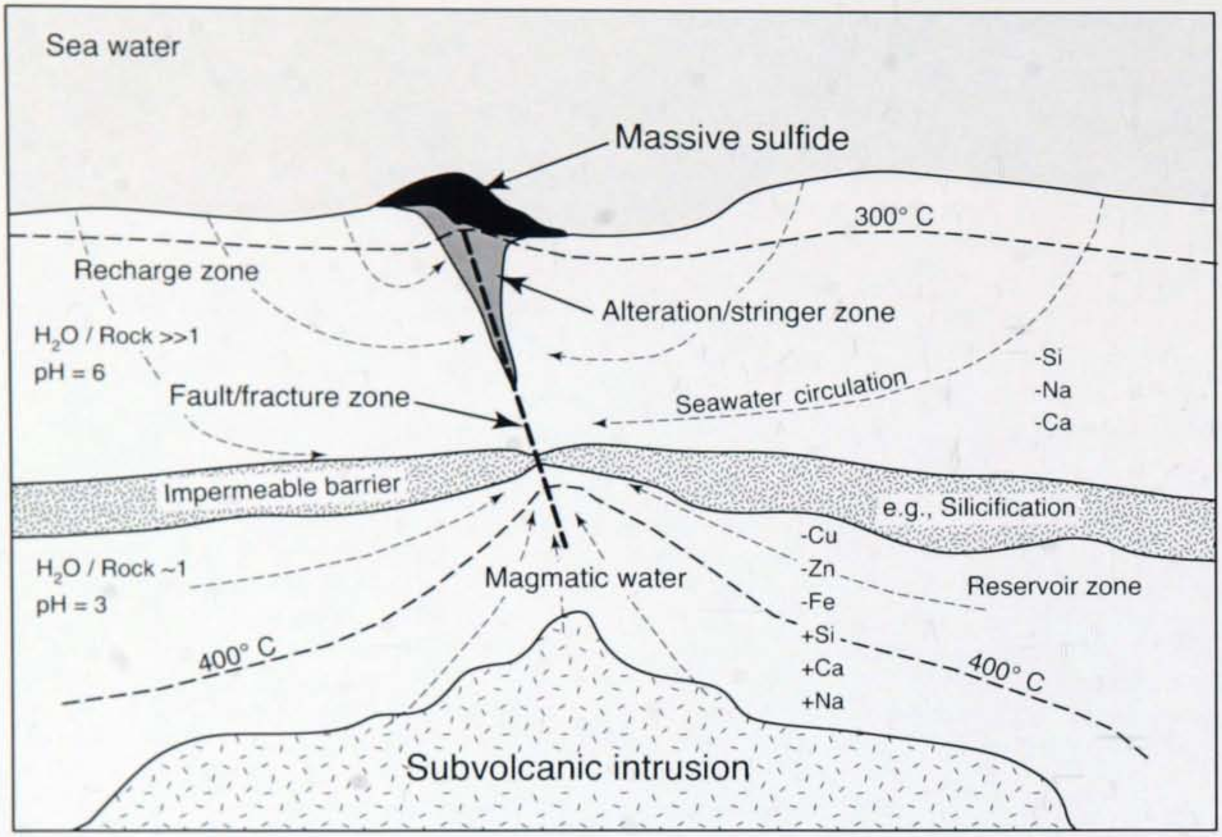


Figure 9.3. Simplified schematic model of a convective hydrothermal system associated with the formation of Noranda-type (Morton and Franklin, 1987) or lava-flow dominated-type (Gibson and others, 1999) volcanic-hosted massive sulfide deposits (modified from Franklin, 1996).

of the model that will be seen during the field trip. Archean lode-gold deposits are one category of ore deposit classified as mesothermal lode-gold deposits (Hodgson, 1993). This deposit type has also been called orogenic gold (Groves and others, 2000), greenstone gold (Robert and others, 1991), Archean lode gold, mesothermal gold-quartz veins, shear-hosted gold, low-sulfide gold-quartz veins (Berger, 1986b), lode gold, Mother Lode veins (Bohlke and Kistler, 1986), and iron-formation-hosted gold deposits (Rye and Rye, 1974; Fripp, 1976; Thorpe and Franklin, 1984; Berger, 1986a; Kerswill, 1993; Vielreicher and others, 1994; McMillan, 1996).

Whatever the name, they are a widespread group of epigenetic ore deposits that have formed in similar settings throughout geologic time. In general, the deposits form during compressional or transpressional deformation at convergent plate margins in accretionary or collisional orogens (Fig. 9.4). They form over a large crustal depth range (2 to 20 kilometers) from deep-seated, low-salinity H_2O-$CO_2 \pm CH_4 \pm N_2$ ore fluids, with gold transported as reduced sulfur complexes. The ore fluids are generated during lower crustal metamorphism from dehydration reactions. Regional structures provide the main control on distribution of lode-gold deposits and mining camps. In many terranes, first-order faults or shear zones appear to have controlled regional fluid flow, with greatest ore-fluid fluxes in and adjacent to subsidiary faults, shear zones, and/or large folds. Highly competent and/or chemically reactive rocks are the most common hosts to the larger deposits. Gold deposition occurs late during the evolutionary history of the host terranes, normally within D_3 or D_4 in a D_1 to D_4 deformation sequence. Absolute ages of mineralization support their late-kinematic timing, and in general, suggest that deposits formed diachronously toward the end of the evolutionary history of hosting orogens.

The late timing of lode-gold deposits is critical to geology-based exploration methods, and hence mineral potential evaluations for these deposits. The late timing is important because of the present structural geometry of the deposits, the mining

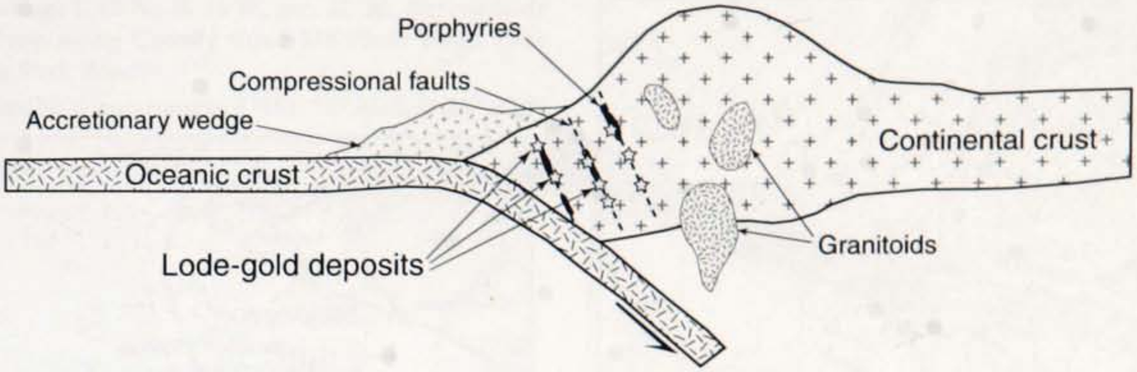


Figure 9.4. Generalized tectonic model for the formation of mesothermal gold deposits, after Groves and others (2000).

camps, and the fact that the enclosing geologic terranes are essentially all similar to the structural geometry during gold mineralization. Therefore, the interpretation of bedrock geologic maps and cross-sections can be used to discern the physical conditions that existed at the time of ore deposition. Exploration for mesothermal lode-gold deposits should incorporate various aspects of the ore deposit model into criteria that can vector into the most favorable areas for hosting such mineralization. The most fundamental characteristic of this class of deposit is the spatial association of the deposits to regional structures. Zones of widespread carbonate alteration (adjacent to regional structures) should be identified and used to focus subsequent exploration. Within carbonate alteration zones, gold is typically only in areas containing quartz veins, silicification, and/or sericite alteration (with or without sulfides). Two general structural controls on the orientation of lode-gold ore veins include deflections and curvatures of shear zones, and where high-strain zones intersect favorable geologic elements (Poulsen and Robert, 1989).

A widespread area of gold mineralization occurs in numerous prospects east of Lake Vermilion, within the Vermilion greenstone belt of northeast Minnesota. The mineralization occurs in rocks of the Neoarchean (~2.7 Ga) Bass Lake sequence (Peterson and Jirsa, 1999) of the Wawa subprovince of the Canadian Shield. This zone of abundant gold mineralization is bounded to the south by the Mud Creek shear zone and to the north by the Vermilion fault. The main access to these prospects is along Mud Creek Road (St. Louis County Road 38). A brief period of mineral exploration for lode-gold deposits in this immediate area of the Vermilion district occurred in the mid 1980s to early 1990s. These programs typically consisted of grid-based geologic mapping, bedrock sampling, ground geophysics, and the completion of soil geochemical surveys. Conversations with many of the people involved in gold exploration programs in the immediate field trip area (centered on T. 62 N., R. 14 W., sec. 6), and compilation of all exploration data from the district as a whole (data from the terminated lease files of the Minnesota Department of Natural Resources), has led to the conclusion that interpretation of linear structural elements exposed in outcrops were not used in designing exploratory drilling plans. Therefore, many of the prospects discovered as a result of these exploration programs remain untested by drilling.

FIELD TRIP STOPS (Fig. 9.5)

The small map insets showing stop locations are taken from U.S.G.S. 7.5-minute topographic quadrangles listed with each stop. The first day of the field trip will include stops located between the towns of Tower and Ely (Fig. 9.6).

DIRECTIONS: From Minneapolis, our route north to the Vermilion district will via Interstate 35, State Highway 33, U.S. Highway 53, and State Highway 169 to the junction with State Highway 135 on the west side of the town of Tower. From the junction of 169/135, drive east approximately 9.5 miles to St. Louis County Road 128. Turn right and continue approximately 1.6 miles to an outcrop on the east side of the roadway.

STOP 9-1

Silicified Fivemile Lake sequence pillow lavas/ regional semiconformable alteration

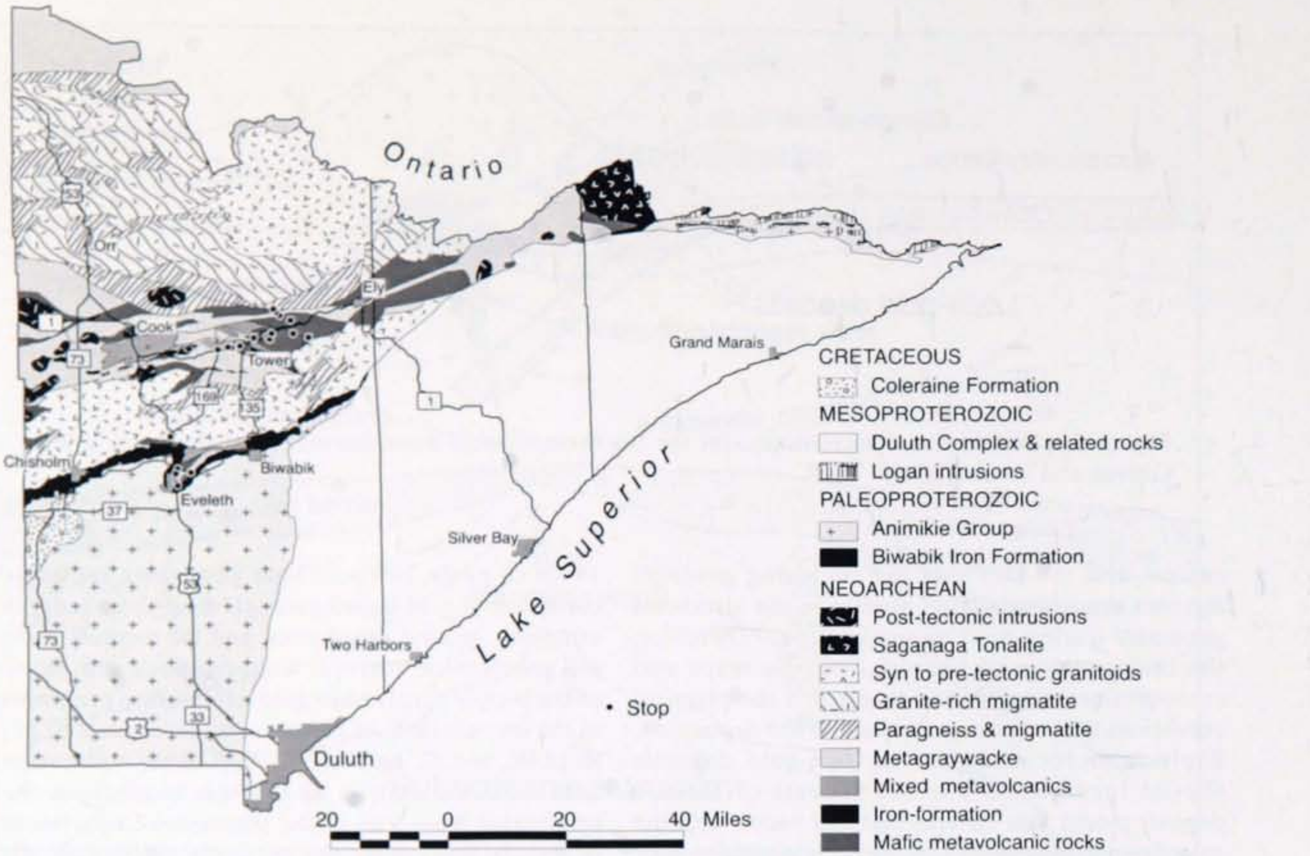


Figure 9.5. Generalized map depicting the major roads, field trip stops, and simplified regional geology of the Arrowhead region of northeastern Minnesota.

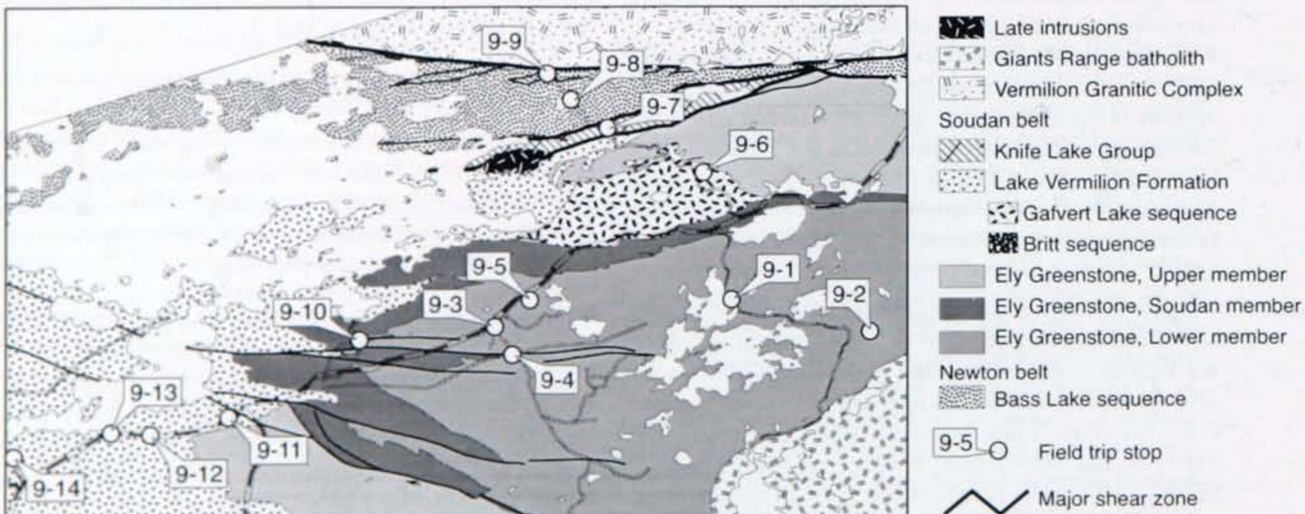


Figure 9.6. Simplified geologic map of the field trip stops in the Vermilion district. Geology modified from Peterson and Jirsa (1999).

Location: T. 62 N., R. 14 W., sec. 22, SE, SE; roadside outcrop along County Road 128 (Bear Head Lake State Park Road)

Eagles Nest quadrangle; UTM: 567,810E/5,297,800N

Description: At this location we can observe part of the regionally extensive quartz-epidote alteration zone. The outcrop contains relatively undeformed bun- and mattress-shaped pillows. Interpillow hyaloclastite zones are generally pale to dark green in color, and are chlorite and/or actinolite-rich. Minor red-brown staining locally occurs in these zones, and is indicative of the presence of trace amounts of pyrite and/or chalcopyrite. Pillow selvedges commonly contain up to 10 percent round to oval, pipe-like quartz-epidote and/or actinolite chlorite amygdules. The cores of the pillows are typically pale green-gray in color due to nearly wholesale replacement of the original igneous minerals by quartz and epidote. This quartz-epidote alteration is typical for much of the Lower member of the Ely Greenstone, and is one of the most important components of possible volcanic-hosted massive sulfide exploration in the district (leaching of copper and zinc) out of a large volume of rock due to hydrothermal alteration.

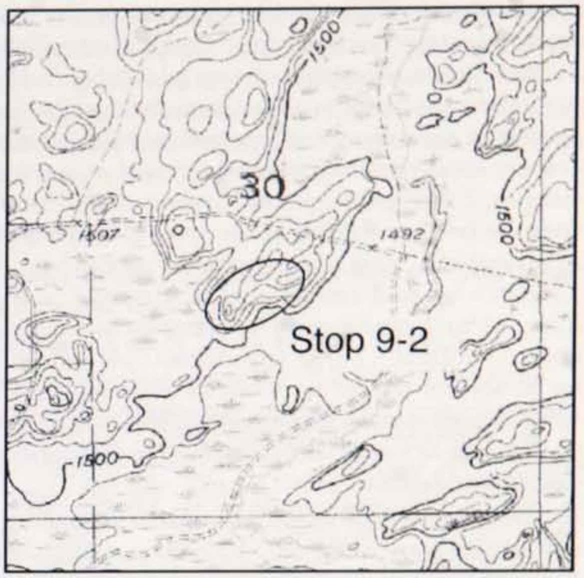
NEXT: Continue on County Road 128 southeast for approximately 2.2 miles to the Purvis Forest Management Road. Turn left and continue about 0.4 mile to a logging road on the right side (south). Follow the logging road approximately 0.15 mile to a series of outcrops.

STOP 9-2

Xenolithic hornblende diorite, Purvis pluton

Location: T. 62 N., R. 13 W., sec. 30, NE, SW

Eagles Nest quadrangle; UTM: 571,805E/5,296855N



Description: The Purvis pluton is an east–west-trending, moderate-sized (~3 cubic kilometers), sill-like multiphase dioritic to tonalitic intrusion with a strike length of 5.7 kilometers and a thickness that ranges from 100 to 1,200 meters (Peterson, 2001). This intrusion occurs in the lower stratigraphic section of the north limb of the Tower–Soudan anticline (Peterson and Jirsa, 1999; Jirsa and others, 2001). Recent work by Drexler and others (2004) indicated that the intrusion has several phases including: 1. Xenolithic hornblende diorite, 2. Xenolithic hornblende tonalite, 3. Xenolithic leucotonalite, 4. Leucotonalite and trondhjemite, and 5. Leucotonalite dikes.

Detailed field mapping by Hovis (2001), Peterson (2001), and Drexler and others (2004) suggested that the Purvis pluton is a synvolcanic intrusion based on the following characteristics: 1. It lacks a contact metamorphic aureole, 2. Its uppermost contact is proximally associated with intense, semiconformable quartz + epidote alteration zones, 3. D_2 deformation fabrics occur in both the intrusion and the surrounding country rocks, and 4. Early xenolithic diorite phases are cross-cut by thin, commonly D_2-deformed dikes of younger tonalite and trondhjemite phases. Galley (2002, 2003) indicated that these characteristics are key features of synvolcanic intrusions temporally associated with the genesis of many Precambrian volcanic-hosted massive sulfide deposits. Peterson (2001) has suggested that the Purvis pluton may have been the heat source that drove hydrothermal systems that produced the Eagles Nest and Purvis Road volcanic-hosted massive sulfide prospects.

This locale offers an opportunity to investigate the xenolithic hornblende diorite phase of the

Purvis pluton. The outcrop adjacent to the road predominately contains four types of xenoliths: 1. Dark green xenoliths of amygdaloidal (5 to 8 percent) basalt-andesite pillow lavas that locally have preserved selvedges and interpillow hyaloclastite, and are locally contact metamorphosed along their margins, 2. Pale green epidote + quartz-altered basalt-andesite lava xenoliths that are up to 15 centimeters in diameter, 3. Rare, coarse-grained gabbro/diorite xenoliths up to 3 centimeters in diameter, and 4. Rare, <1 to 2 centimeters in diameter subangular chert xenoliths. Large iron-formation xenoliths up to several meters in diameter and amphibolite xenoliths up to several centimeters in diameter may be observed in xenolithic hornblende tonalite outcrops east of this location.

Iron-formation xenoliths present in outcrops east of here were likely derived from iron-formation horizons that occur immediately southwest of Purvis Lake. Basalt and altered basalt fragments also were derived from the surrounding Lower Ely member. The presence of epidote-quartz altered mafic xenoliths suggests that this phase of the Purvis pluton stoped its way upward into an earlier-formed proximal zone of quartz-epidote alteration formed from high-temperature seawater-rock interaction (for example Galley, 2003). Amphibolite xenoliths are believed to represent contact metamorphosed basalt fragments based on petrographic similarities (Drexler and others, 2004). Drexler and others (2004) have shown that coarse-grained gabbro/diorite fragments likely represent xenoliths of the earliest phases of the pluton.

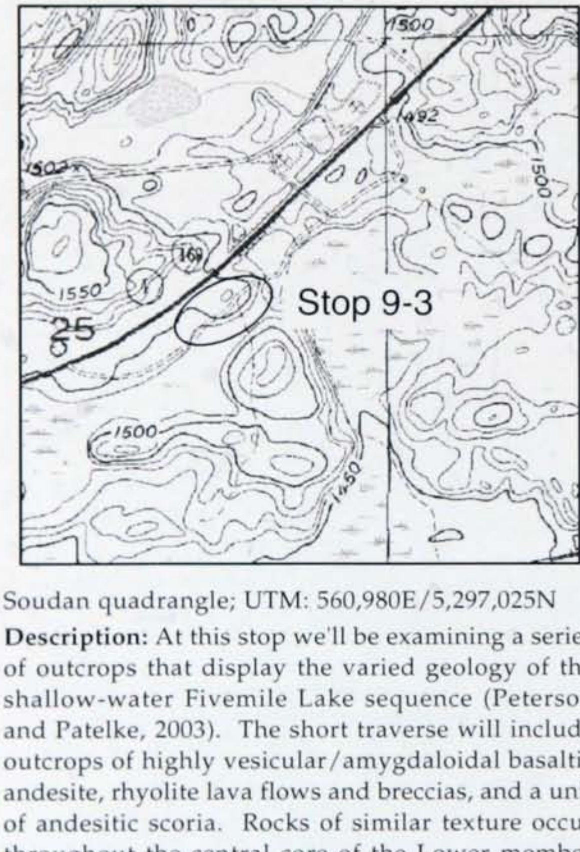
Studies of ancient volcanic-hosted massive sulfide deposits have documented that the deposits commonly occur in depressions on the paleo-seafloor (third-order basins), but modern deposits on the seafloor are found on high-standing structures, such as ridges. These differences are probably more apparent than real, in that the modern deposits are generally confined to the axial graben, or depression, of what otherwise are high-standing structures. In addition, both ancient and modern deposits occur in areas of anomalously high heat flow, generally linked to synvolcanic intrusions beneath the hydrothermal systems. The recent mapping (Peterson, 2001; Drexler and others, 2004) in the Purvis Road area has shown the presence of all of the attributes of typical volcanic-hosted massive sulfide-forming hydrothermal systems. These attributes include a synvolcanic intrusive heat source (the Purvis pluton), a paleotopographic high-standing structure, volcanic-hosted massive sulfide-style alteration mineral assemblages, and the presence of copper and zinc-rich massive sulfide (recent logging in this area has exposed numerous angular boulders of massive sulfide in the basal till).

NEXT: Return to Highway 169, turn left (west) and travel approximately 4.4 miles to the junction with Murray Forest Management Road. Turn left (south) and travel approximately 0.2 mile to the south-curving bend in the road. Walk along the old logging road to the west.

STOP 9-3

Shallow-water volcanic rocks of the Fivemile Lake sequence: andesite, rhyolite, and scoria

Location: T. 62 N., R. 15 W., sec. 25, SW, NE



Soudan quadrangle; UTM: 560,980E/5,297,025N

Description: At this stop we'll be examining a series of outcrops that display the varied geology of the shallow-water Fivemile Lake sequence (Peterson and Patelke, 2003). The short traverse will include outcrops of highly vesicular/amygdaloidal basaltic andesite, rhyolite lava flows and breccias, and a unit of andesitic scoria. Rocks of similar texture occur throughout the central core of the Lower member of the Ely Greenstone.

NEXT: Continue south along Murray Forest Management Road for approximately 0.6 mile to the junction with the old DM & IR rail line. Walk west along the rail line for 50 meters to the outcrops on the north and south side of the rail line.

STOP 9-4

Gold prospect along the Murray shear zone

Location: T. 62 N., R. 14 W., sec. 30, SW, SW

Soudan quadrangle; UTM: 561,490E/5,296,205N

Description: This is one of the few gold prospects in the Lower member of the Ely Greenstone. South of the old rail line, intense shearing associated with the north edge of the Murray shear zone culminated with the formation of chlorite-ankerite-sericite schists. In the mid-1980s, Newmont Exploration discovered lode-gold mineralization along the northern margin of the Murray shear zone. Gold mineralization in this area, named the Murray prospect by Newmont, is associated with quartz-carbonate-pyrite-galena-tetrahedrite veins in strongly sheared and carbonatized rocks. Newmont reported values up to 12.5 parts per million gold during the course of their exploration.

An estimate of the amount of displacement within the panel of rocks bounded by the Murray shear zone is given in Table 9.2 (Peterson and Patelke, 2003). These values were calculated geometrically by using the average plunge of measured lineations (71°) and two measured lines of possible correlative stratigraphy offset by the bounding shear zones. The calculated total displacement values (net slip) are quite large (up to 13.8 kilometers, or 42,000 feet of net slip), but the displaced rocks would still fall within the range of depth generally associated with greenschist facies metamorphism.

Table 9.2. Calculated displacement along the Murray shear zone, in kilometers.

Lineation plunge	Strike slip	Dip slip	Net slip
71°	4.5	13.1	13.8
71°	3.0	8.7	9.2

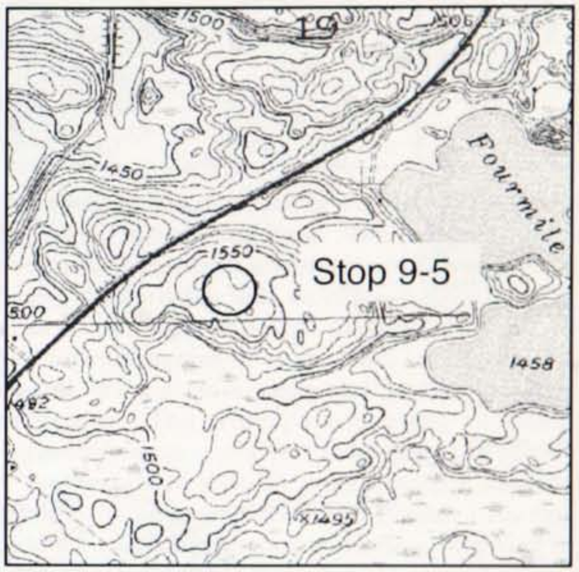
NEXT: Return to Highway 169 via the Murray Forest Management Road. Turn right on Highway 169 and travel approximately 0.7 mile to the junction of a logging/gravel road on the south side of the highway. Walk up the road to the southwest to the very large outcrop on the top of the hill.

STOP 9-5

Central Basalt sequence sheet flows, pillow lavas, and perlitic hyaloclastite

Location: T. 62 N., R. 14 W., sec. 19, SE, SW

Soudan quadrangle; UTM: 562,000E/5,297,800N



Description: The Central Basalt sequence (Peterson and Patelke, 2003) comprises a steeply north-dipping (75° to vertical), north-facing sequence of sparsely amygdaloidal pillowed and massive lava flows of basalt composition that are believed to be correlative with the tholeiitic Armstrong Lake volcanic sequence mapped in the Eagles Nest quadrangle (Jirsa and others, 2001). Relative to massive and pillowed basalt and andesite flows in the Fivemile Lake sequence, Central Basalt sequence lava flows are notably less amygdaloidal and lack multiple pillow rind structures. In addition, the Central Basalt sequence lacks the thick sequences of scoriaceous basalt-andesite lapilli tuffs that are commonly interstratified with lava flows in the Fivemile Lake sequence. These characteristics of the Central Basalt sequence indicate eruption and deposition in a deeper submarine environment than the stratigraphically older Fivemile Lake sequence, and suggest overall increasing water depth during the temporal development of the Lower Ely Greenstone.

The outcrop comprises two east–southeast-striking massive basalt flows, ranging from at least 5 to 9 meters in thickness, that are separated by a 10-meter-thick flow unit comprising pillows and pillow lobes (Fig. 9.7). Flow 1, at the southern part of the outcrop, is composed of a pale to dark green, faintly feldspar-phyric (~10 percent 0.5 to 1 millimeter laths), sparsely amygdaloidal, basalt sheet flow that locally exhibits tortoise-shell jointing formed in response to contraction during cooling. The uppermost 10 to 40 centimeters of the coherent part of Flow 1 is generally silicified and epidotized. Petrographic observations indicate that this section of the flow also contains up to 70 percent <0.1 centimeter round spherulites. An irregular contact occurs between the coherent basalt flow and an overlying 1- to 2-meter-thick unit of dark

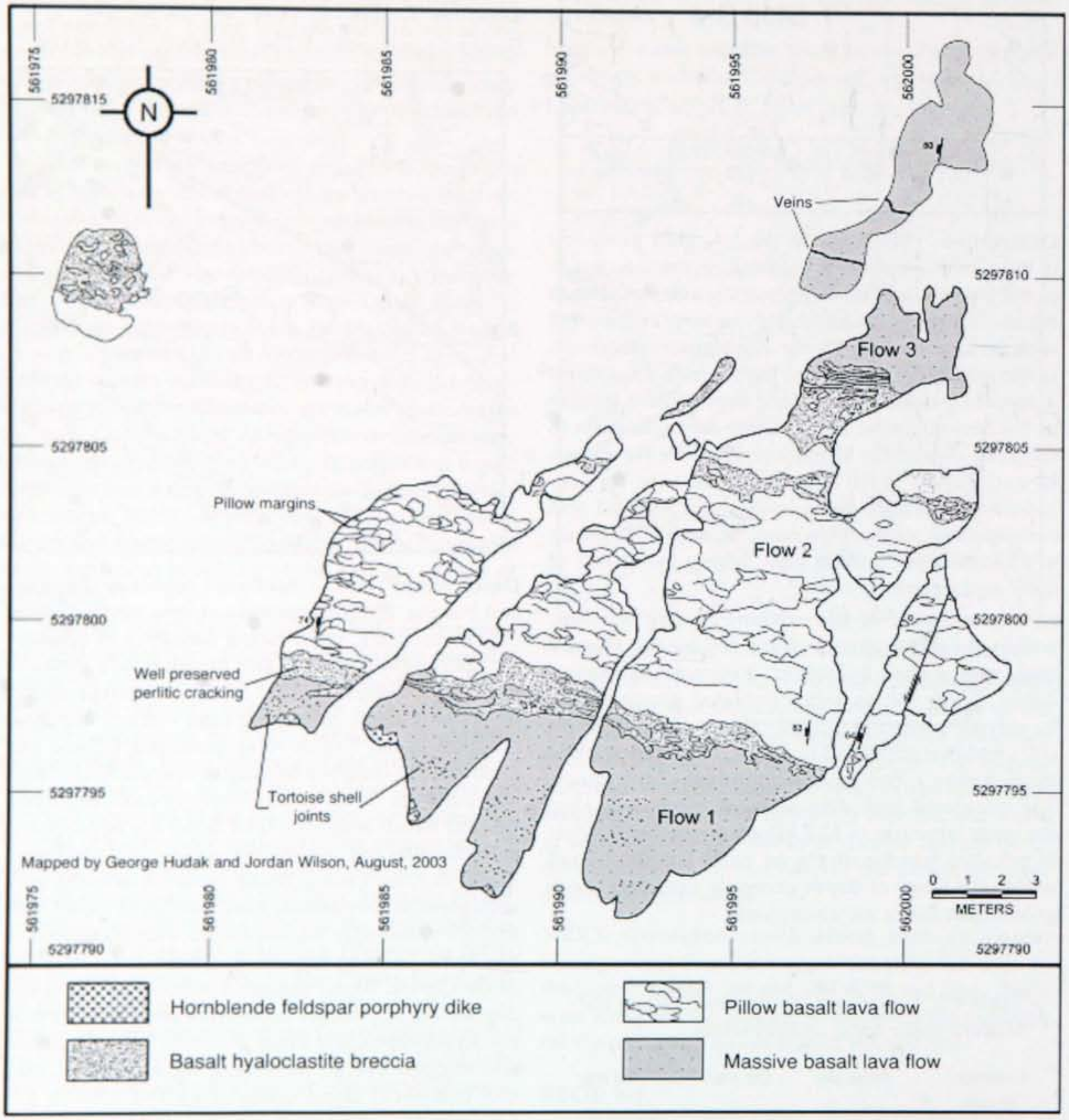


Figure 9.7. Detailed geologic map of sheet flows, pillow lavas, hyaloclastite, and associated "self-peperite." Map from Hudak and others (2004).

green, exceptionally well-preserved perlitic in situ hyaloclastite and associate self-peperite (Batiza and White, 2000). The hyaloclastite formed from non-explosive fracturing of the basalt glass developed on the flow top due to quenching by water, whereas the perlite formed following deposition by hydration of volcanic glass. An irregular contact occurs between the hyaloclastite and Flow 2, which is composed of north-facing mattress- to bun-shaped pillow lavas and pillow lobes with numerous "neck and knob" structures. Individual pillows have well developed perlitic hyaloclastite margins that range from 1 to 4 centimeters in width. Pillow buds indicate propagation from east to west, suggesting the volcanic vent was located east of this location. The coherent pillows and lobes are overlain by up to 2.5 meters of hyaloclastite breccia that contains 20 to 40 percent subround to subangular pale gray-green basalt lapilli in a jigsaw puzzle-fit dark green perlitic hyaloclastite matrix.

The upper contact of Flow 2 and the overlying basalt sheet flow (Flow 3) is irregular, and is marked by thin (1 to 8 centimeters thick), sheetlike basalt fragments that are up to 1.6 meters in length. These fragments locally appear to be isoclinally folded about an east–west-trending fold hinge. Although the genesis of this structure is currently not well understood, it may be due to syneruptive deformation of either thin slabs of hot, basal flow margin crust from the overlying flow, or thin injections of basalt magma into the hyaloclastite from either the underlying pillows or the overlying sheet flow. Flow 3 comprises an at least 10-meter-thick, pale green-gray, slightly feldspar-phyric, sparsely amygdaloidal sheet flow. Steep, north–northeast-trending, west-dipping D_3 joints are well developed in this unit, as are lens-shaped psuedo-pillows that are up to 50 centimeters in diameter.

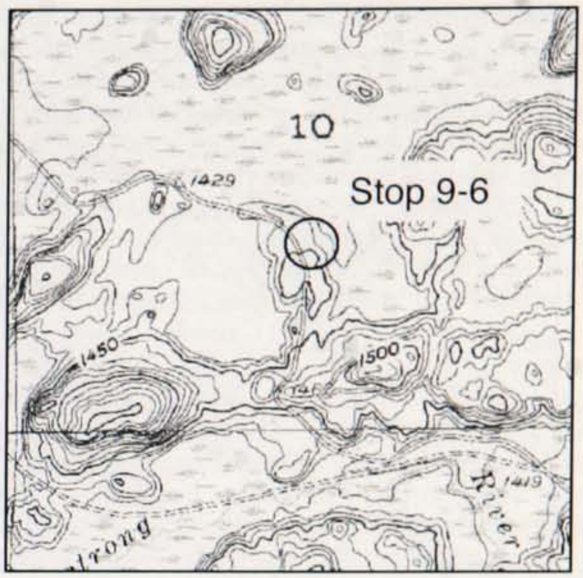
NEXT: Return to Highway 169, turn right and travel approximately 3.1 miles to the junction of Mud Creek Road (St. Louis County Road 38). Turn left (north) on 38 and travel approximately 1.5 miles to a series of low-lying outcrops on the east side of the road.

STOP 9-6

Fragmental rocks of the Gafvert Lake sequence

Location: T. 62 N., R. 14 W., sec. 10, NE, SW

Eagles Nest quadrangle; UTM: 567,000E/ 5,301,490N



Description: The informally named Gafvert Lake sequence (Peterson and Jirsa, 1999) is interpreted to represent an Archean stratovolcano of andesitic to dacitic composition that stratigraphically overlies rocks of the Ely Greenstone. The complex includes lava flows, fragmental rocks (tuff, lapilli tuff, tuff breccia, debris flow deposits) and porphyritic intrusions. The widespread nature of dacitic fragmental rocks of Gafvert Lake sequence affinity in the Vermilion district indicates that repeated episodes of explosive volcanism (Crater Lake-type caldera formation) occurred in the area. Capping the central portion of the Gafvert Lake sequence are a number of thick, massive pyrite horizons that have metal signatures associated with volcanic-hosted massive sulfide, epithermal, and biologic affinity (Peterson, 2001). We will examine a series of outcrops of fragmental dacitic deposits located on the east side of Mud Creek Road.

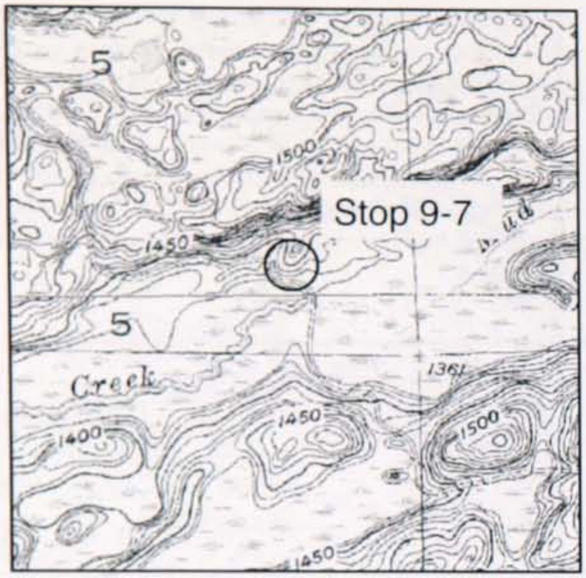
NEXT: Continue northwest on 38 for approximately 2.1 miles to the boat landing parking area along Mud Creek. Walk approximately 100 meters north along the road to the outcrop along the east side of the roadway.

STOP 9-7

Mud Creek shear zone

Location: T. 62 N., R. 14 W., sec. 5, SE, SE; outcrop just northwest of Mud Creek near the road

Chad Lake quadrangle; UTM: 564,230E/5,302,800N



Description: The regional scale Mud Creek shear zone occupies the east–northeast-trending valley of Mud Creek, which is clearly visible at this location. This shear zone separates rocks of the Newton Belt (here the Bass Lake sequence) to the north and rocks of the Soudan belt (Gafvert Lake sequence and the Upper Ely Greenstone) to the south. Development of this shear zone is a product of largely dextral transpressive deformation that has been partitioned into discrete zones, presumably late in D_2 deformation. It is generally believed that gold-bearing mineralization was introduced during these later deformation events, and the Mud Creek shear zone and environs continue to attract considerable attention as a gold target. The Mud Creek shear zone is analogous with major faults (Destor-Porcupine fault) and "breaks" (Cadillac-Larder Lake break) of major lode-gold mining districts in Canada. Historic gold assays taken from rocks of the shear zone itself are essentially devoid of gold, as is the case for most major structures within Archean lode-gold mining camps. This series of outcrops is located within the northern margin of the internal highly strained zone of the shear, and include outcrops of: 1. Ankerite-sericite-quartz-green mica-pyrite schist with quartz and tourmaline knots, and 2. Highly folded and compositionally banded phyllites with quartz veins. The protolith for these rocks is unknown, because of the intense deformation, but could be any of several rock types in the region, including quartzofeldspathic porphyry, metavolcanic rock, or graywacke.

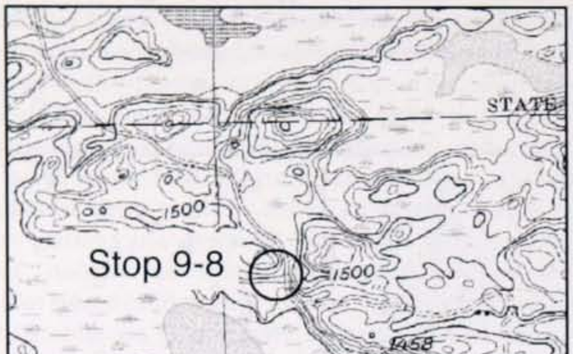
NEXT: Drive northwest along County Road 38 approximately 1.1 miles to the widened portion of the road.

STOP 9-8

Sheared quartz-feldspar porphyry, basal till, and detailed mapping interpretations

Location: T. 62 N., R. 14 W., sec. 5, SW, NW

Chad Lake quadrangle; UTM: 563,190E/5,303,615N



Description: East of Lake Vermilion, the geology of the Bass Lake sequence is dominated by six basic rock types that include: 1. Tholeiitic pillowed basalt flows interpreted to have formed in a deep-water setting based on volcanic textures, 2. Gabbro sills interpreted as synvolcanic in age due to their stratigraphic continuity and similar deformation as the enclosing pillowed basalts, 3. Felsic porphyries (feldspar porphyry and quartz-feldspar porphyry) interpreted to have intruded during late stages of D_2 deformation based on field relationships and geochronology (quartz-feldspar porphyry from the Pac Man Pond prospect returned a $^{207}Pb/^{206}Pb$ age of 2683.0 ± 1.4 Ma; Peterson and others, 2001), 4. Algoma-type iron-formation, 5. Thinly-bedded argillite and siltstone, and 6. Sheared rocks, which are dominated by chlorite-rich schist, phyllite, and phyllonite. In addition, localized areas of fragmental felsic volcanic rocks occur stratigraphically below distinct iron-formation horizons.

In the last twenty years, numerous gold prospects have been discovered in the eastern portion of the sequence. These prospects generally fall into one of three categories: 1. Auriferous quartz-carbonate-pyrite veins and sulfidized zones in iron-formation, 2. Auriferous quartz-sericite-ankerite-pyrite schists, or 3. Felsic intrusive-hosted auriferous quartz veins and stockworks. All of the prospects are found within areas of moderate to strong iron-carbonate alteration, with the best mineralization commonly found within sericitic alteration zones. Numerous equigranular and porphyritic felsic intrusions occur within the areas of alteration and gold mineralization, and are a good guide for locating mineralized structures. The gold mineralization is generally related to deformation in subsidiary structures associated with movement along the D_2 Mud Creek shear zone.

Widening of the roadbed of County Road 38 in 2003 exposed a number of new outcrops and cuts into the basal till in this area. Detailed geologic mapping of gold prospects north of the Mud Creek shear zone by Peterson and Patelke (2004a) included mapping these new exposures of the Bass Lake sequence. For this stop, we will traverse along County Road 38 and look at these new exposures.

NEXT: Continue northwest along County Road 38 approximately 0.55 mile to a small, yellowish outcrop on the east side of the road.

STOP 9-9

The Kerr McGee gold prospect

Location: T. 63 N., R. 14 W., sec. 31, SE, SE

Chad Lake quadrangle; UTM: 562,480E/5,304,610N

Description: The Kerr McGee gold prospect is hosted within an extensive zone of highly strained rocks, interpreted to be a subsidiary structure associated with the Mud Creek shear zone. Moderate- to high-grade gold mineralization at the Kerr McGee prospect occurs within multiple thin (0.2 to 2.0 meters) zones of quartz-sericite-ankerite-pyrite ± green mica ± tourmaline schist hosted by an extensive zone of essentially gold-barren, chlorite-rich schist. Thin and probably boudined iron-formation horizons occur locally in the chlorite-rich schist, and locally are strongly mineralized in this area. Mineralized zones locally contain extensive foliation and shear parallel quartz, ankerite, and/or quartz-ankerite veins, and may widen in zones of silicification. The style of gold mineralization exposed in the Kerr McGee prospect is similar to both the Clear Cut (~0.5 mile west) and Railroad Zone (1.5 miles east) prospects. In fact, the sericitic zone that hosts the mineralization may have continuity to both of these other prospects.

Three-dimensional visualization (Fig. 9.8) of the detailed lithological and structural mapping by Peterson and Patelke (2004a) within the Kerr McGee prospect area revealed important information that can be used to design drilling plans that significantly increase the chance of intersecting gold mineralization exposed in outcrop at the surface. For example, drill hole RC-3, which is located 100 meters east of the main gold showing on the eastern side of this knob, was drilled due north (at a dip of 45°) and targeted to intersect the mineralization exposed in outcrop at the Kerr McGee showing. Chevron Resources drilled this hole in 1987, at the western boundary of their lease property (the prospect was then held by Kerr McGee). Detailed structural mapping in these outcrops revealed that the rocks within the mineralized zone have moderate to strong elongation and intersection (foliation and shear planes) lineations trending 60° and dipping northeast at 72°. The best interpretation of the down-dip orientation of the mineralized zone is this lineation trend and plunge, and drill hole RC-3 never intersected the mineralized zone.

NEXT: Return to Highway 169 via County Road 38. Turn right (west) and travel approximately 7.3 miles to the junction of Jasper Road in the town of Soudan. Turn right on Jasper Road and follow it to the T-junction (~0.5 mile). Turn right, go up the hill, park at the mine buildings, and walk about 150 feet north and uphill to an outcrop on the right.

STOP 9-10

No hammering please!

Archean Soudan iron-formation member of the Ely Greenstone

Location: T. 62 N., R. 15 W., sec. 27, NE, NE; Soudan Underground Mine State Park

Soudan quadrangle; UTM: 557,120E/5,296,660N

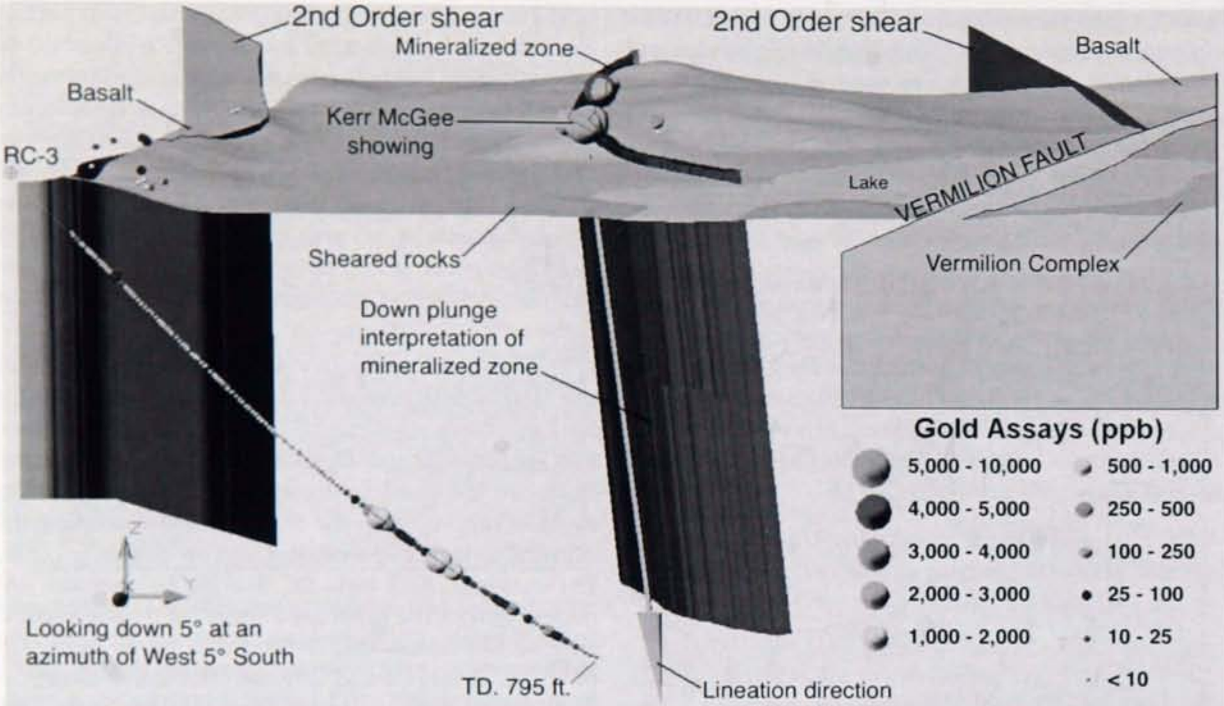


Figure 9.8. Three-dimensional view of the relationship between structural boundaries, the mineralized zone exposed on the surface at the Kerr McGee prospect, and drill hole RC-3. Upward extension to the surface of the two anomalous zones (greater than 1,000 parts per billion gold) intersected in hole RC-3 would place these zones in the black spruce and cedar swamp located south–southeast of the prospect.

Description: This classic exposure of the Soudan iron-formation member of the Ely Greenstone lies on the north limb of the Tower–Soudan anticline, and at the stratigraphic top of the volcanic sequences known collectively as the Lower member of the Ely Greenstone. The outcrop displays two generations of tight folding in delicate laminae of chert (creamy white), chert-hematite jasper (red), and magnetite-chert (black to silver-colored). The second generation of folds (F_2) is tectonic in origin, having subvertical axial surfaces that trend east, and steeply plunging axes. Most display Z-asymmetry. The earlier folds (F_{0-1}) appear to have been sharply refolded to produce complex interference patterns. Lundy (1985) studied folding at this locality and concluded that some of the apparent interference structures are the product of early-formed sheath folds that did not involve refolding by D_2. The F_1 structures are predominantly intrafolial and exhibit a great variety of style and orientation, implying they formed by layer-parallel, soft-sediment slumping.

It is interesting to observe the rhythmic microlaminae (1 millimeter or so thick) in various cherty beds exposed here and speculate about the paleoenvironment—that is, whether these represent daily heating/cooling, tidal, climatic, annual, or some other repetitive influence in the depositional environment. What is known about units of iron-formation in the Ely Greenstone, of which there are many, is that deposition occurred during periods of relative volcanic and tectonic quiescence by the slow subaqueous "rain" of chemical precipitates.

The deep excavations in this area are the early workings of the Soudan Iron Mine, the first in Minnesota. The mine produced about 16 metric tons of high-grade hematite ore (60 to 63 percent iron content) from 1884 until 1962, when the land was deeded to the State of Minnesota and converted to a state park. Most of the production came from underground workings that began here in 1900, and which now can be visited on guided tours. The mine also houses an underground physics research facility at 2,340 feet below the surface. A massive expansion of that facility is under consideration to create a national underground laboratory at considerably greater depths (Peterson and Patelke, 2003).

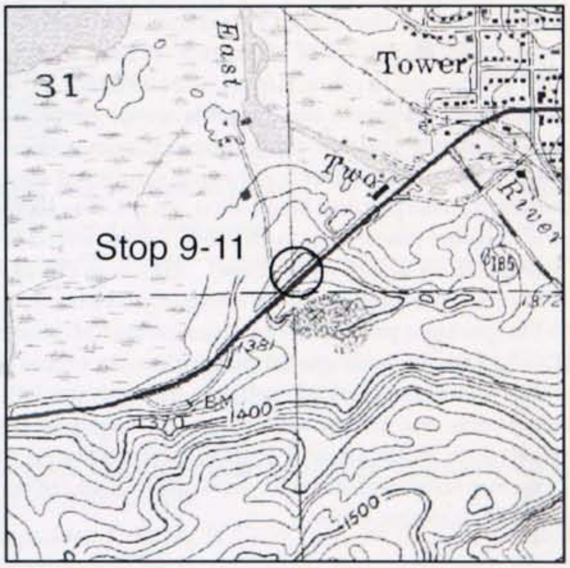
NEXT: Return to Highway 169 and turn right. Follow 169 through the town of Tower to the large outcrops immediately west of town (approximately 3.1 miles).

STOP 9-11

Archean fragmental volcanic rocks

Location: T. 62 N., R. 15 W., sec. 32, SW, SW; Highway 169 road cut, west edge of Tower

Tower quadrangle; UTM: 553,380E/5,294,430N



Description: This outcrop consists of fragmental, variably reworked volcanic conglomerate and tuffaceous rocks of the Gafvert Lake sequence of the Lake Vermilion Formation. The rock is composed of about 85 to 95 percent dacitic detritus, 3 to 5 percent gray clasts of graywacke, slate, and basaltic andesite, and a small percentage of magnetic and sulfidic fragments. Fragments range in size from a few millimeters to 20 centimeters. The generally poorly developed sorting and bedding, together with varied clast composition, implies a debris-flow origin. Compare these rocks with those of Stop 9-6.

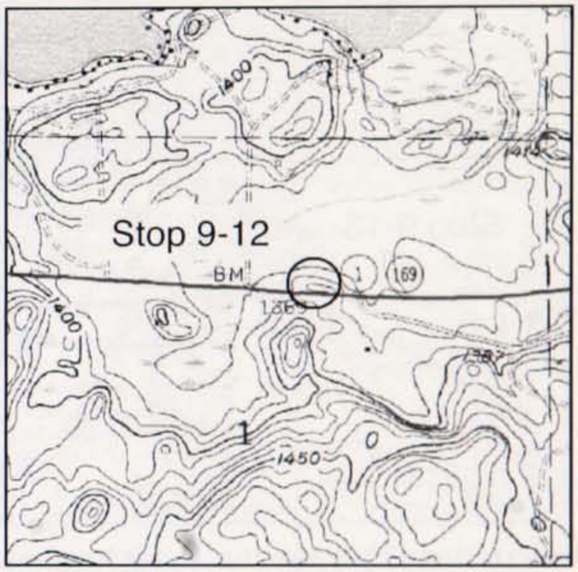
NEXT: Continue west on Highway 169 approximately 1.5 miles to a road cut.

STOP 9-12

Archean dacitic tuff/Paleoproterozoic or Mesoproterozoic diabase dike

Location: T. 61 N., R. 16 W., sec. 1, SW, NE; Highway 169 road cut

Tower quadrangle; UTM: 551,160E/5,293,960N



Description: These road cuts expose outcrops of white, dacitic, tuffaceous, sedimentary rock, a component of the Lake Vermilion Formation. Regionally, the formation consists of all compositional gradations between what appears to be first-cycle tuff, tuffaceous graywacke, and mixed-source graywacke, interbedded on all scales. In a general way, the tuffaceous component increases in proportion to the east toward the Tower–Soudan anticline. The presumed source of the dacitic volcanic detritus exposed in this area is stratovolcanos of Gafvert Lake affinity, which overlie the composite volcanic shield complex of the Ely Greenstone. Ring plains and irregular basins composed of detritus shed from the high-standing volcanic complex are now represented by the Lake Vermilion Formation.

The northeast-trending, steeply dipping, 7-meter-wide diabase dike that cuts tuffaceous rocks has been the source of considerable debate. Its petrographic (olivine-bearing) and geochemical (silica undersaturated) composition is similar to

Mesoproterozoic dikes (Schmitz, 1994); yet it lies nearly along strike with, though east of, dikes of the Paleoproterozoic Kenora–Kabetogama dike swarm.

NEXT: Continue west on Highway 169 approximately 0.7 mile to the junction with St. Louis County Road 526.

STOP 9-13

Multiply folded Archean graywacke

Location: T. 61 N., R. 16 W., sec. 2, NE, NE; south side of Highway 169 just east of County Road 526

Tower quadrangle; UTM: 550,050E/5,294,000N

Description: This outcrop at the road and several smaller ones in the bush nearby show the superposition of two generations of folds in thin-bedded, well-graded graywacke of the Lake Vermilion Formation. The second-generation folds (F_2) are associated with a regional axial plane cleavage in which sedimentary clasts are flattened. The earlier F_1 folds have no associated cleavage and tend to be erratic in form, trend, and distribution. Folds display "eye" and "mushroom" shapes that locally are interpreted to be sheath folds (Hudleston and others, 1987). These characteristics are consistent with deformation of poorly lithified sediment. The superposition of deformation events is manifest in the transection of F_1 folds by cleavage related to D_2. In this area and to the west, one can find anticlinal synclines and synclinal anticlines, indicating stratigraphic inversion prior to D_2 folding.

NEXT: Continue approximately 1.7 miles to the west to the junction of St. Louis County Road 77. Turn right on 77 and follow for approximately 0.6 mile to the outcrop on the left side of the road on the north side of Vermilion Dam.

STOP 9-14

Archean graywacke at Pike River Dam

Location: T. 61 N., R. 16 W., sec. 3, NW, SW; west side of County Road 77, on the north side of the river

Note that Fortune Bay Casino—the overnight hotel—lies to the north off of County Road 77.

Tower quadrangle; UTM: 547,300E/5,293,340N

Description: One of the truly classic outcrops of graywacke of the Lake Vermilion Formation is beautifully exposed at this stop. Prior to about the 1950s, no depositional mechanism could satisfactorily explain the coincidence in graywacke of: 1. Coarse-grained sand derived from a source many kilometers distant and having an altered clayey matrix, 2. Interbedded black slate, and 3. The lack of evidence for reworking in shallow water (indicative of deposition below wave base). This was changed when the concept of turbidity currents was introduced to the geologic profession by Kuenen and Migliorini (1950). Despite widespread publication on turbidites in more modern geologic settings through the 1950s and 1960s, the facies model was not refined and applied to Archean and Proterozoic strata in the Lake Superior region until somewhat later (Morey, 1965; Ojakangas, 1966).

END OF DAY 1

NEXT: Return to Highway 169 and turn west (south). Travel south on Highway 169 approximately 28 miles to the junction with Highway 53; follow 53 south approximately 0.7 mile to the Laurentian Divide wayside rest stop (Fig. 9.9).

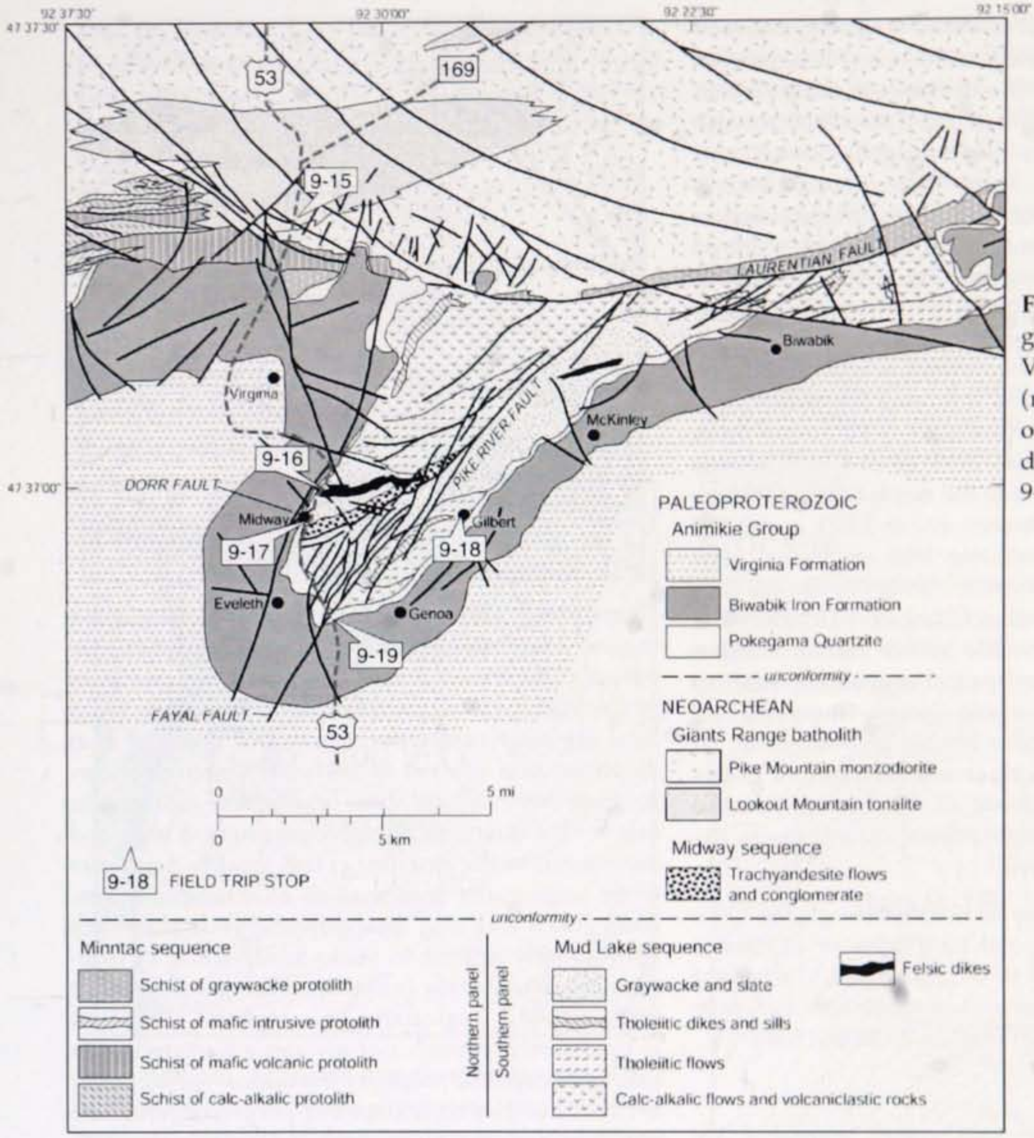


Figure 9.9. Generalized geologic map of the Virginia horn area (modified from Jirsa and others, 1998) showing details of field trip Stops 9-15 to 9-19.

STOP 9-15

Archean Giants Range batholith at "Confusion Hill," Laurentian Divide

Location: T. 59 N., R. 17 W., sec. 19, SE, SE; wayside off Highway 53

Virginia quadrangle; UTM: 534,337E/5,269,458N

Description: Exposed near this wayside and in road cuts on both sides of the highway is an array of variably layered intrusions having both tonalitic (white) and dioritic (black) compositions. A cursory look shows intrusive relationships that

conclusively demonstrate that diorite was emplaced into tonalite at one locality, and at another, tonalite was emplaced into diorite. In detail, all compositions intermediate between the two end members are also present locally. Although the dioritic component is abundant here, the bulk of the mapped unit is tonalitic. Emplacement of this unit, now known as the Lookout Mountain tonalite, probably involved some degree of magma mingling. Dikes of tonalite that cut the adjacent high-grade supracrustal rocks of the Minntac sequence contain metamorphic fabrics, yet little evidence of metamorphic origin can be seen in the interior of the body, implying it is syntectonic with respect to D_2 deformation. U-Pb zircon dates (Boerboom and Zartman, 1993) of two components of the batholith exposed to the north bracket the age of D_2 deformation between about 2,674 and 2,682 Ma. Exposures at Confusion Hill are a small part of the Giants Range batholith, which forms the core bedrock of the Laurentian (drainage) Divide. The batholith is a 40-mile-wide belt of intrusions that can be traced on geophysical maps and outcrop east to the Mesoproterozoic Duluth Complex, and west beyond the western border of Minnesota. It separates Archean supracrustal sequences in the Virginia horn from those of the Tower–Soudan area—making stratigraphic correlation between the two districts speculative.

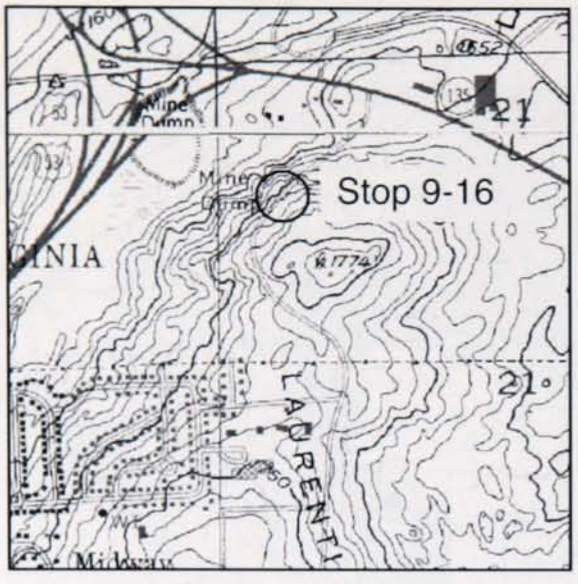
NEXT: Follow Highway 53 south through the town of Virginia. Take the exit for Highway 135 (east) approximately 0.5 mile to Bourgin Road. Turn right (south) on Bourgin Road and continue about 0.4 mile to a large cut on the left (east) side of the road.

STOP 9-16

Archean graywacke and slate, intruded by quartzofeldspathic porphyry

Location: T. 58 N., R. 17 W., sec. 21, SW, SW; road cuts on east side of the Bourgin Road

Eveleth quadrangle; UTM: 536,311E/5,260,659N



Description: Outcrops along this side of the road expose quartzofeldspathic porphyry intruded into variably deformed graywacke, siltstone, and slate of the Mud Lake sequence. The sedimentary rocks here are moderately deformed, but much of that deformation is inferred to predate the main cleavage-forming event D_2, and some may be soft-sediment in origin. The quartzofeldspathic porphyry is large and continuous to the east, but at this locality it appears to be segmented into a zone of multiple dikes. Both graywacke and quartzofeldspathic porphyry are intensely altered to some combination of iron-carbonate minerals (ankerite, ferroan dolomite) and sericite. Regionally, this style of alteration is commonly, though not always associated with quartzofeldspathic porphyry intrusions—presumably because the quartzofeldspathic porphyry remained more structurally rigid than the enclosing sedimentary rocks during the shear-related deformation event that accompanied alteration late in D_2. Most gold mineralization in the area is closely allied to this alteration, yet this outcrop is surprisingly barren.

One of the earliest gold discoveries in Minnesota was made by J.W. Gruner (Grout, 1937) in a railroad cut not far from Stop 9-16. The cut exposes graywacke intruded by quartzofeldspathic porphyry, having visible gold associated with small quartz veins. Despite several episodes of mineral exploration in this area (most notably the Newmont Exploration in the 1980s), no economic gold deposits have been discovered.

NEXT: Follow Bourgin Road to the south and west to a frontage road on the east side of Highway 53. Turn north (right) on the frontage road and travel about 0.2 mile to the first road to the right, turn uphill and continue to #7 Mesabi Lane.

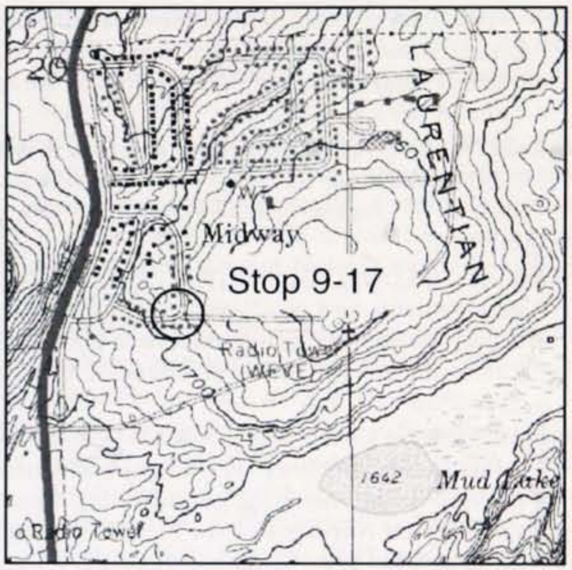
STOP 9-17

Private property! Permission must be obtained before entering!

Archean conglomerate

Location: T. 58 N., R. 17 W., sec. 20, SW, SE, 7 Mesabi Lane; town of Midway

Eveleth quadrangle; UTM: 535,713E/5,259,459N



Description: Archean conglomerate and lithic sandstone that form the driveway are part of the northeast-trending Midway sequence, which contains these strata types locally interbedded with subaerially deposited, calc-alkalic (trachyandesitic) volcanic rocks. The sequence is inferred to have formed after earliest deformation (D_1) of the enclosing graywacke and basaltic rocks of the Mud Lake sequence, but before the cleavage-forming D_2 deformation that affected both sequences. The conglomerate contains clasts of basalt, graywacke, porphyritic trachyandesite, and quartzofeldspathic porphyry. This provenance indicates that the older Archean rocks of the Mud Lake sequence were intruded by quartzofeldspathic porphyry, deformed, and uplifted to provide detritus to what was probably a successor or "pull-apart" basin developed along a major structure now occupied by the Pike River fault zone.

Midway sequence conglomerate has previously been interpreted as a basal sediment (Sutton, 1963) and as a proximal turbidite fan deposit (Levy, 1991), depositionally transitional with graywacke and slate of the Mud Lake sequence. Subsequent investigation (Jirsa, 2000) indicated that the conglomerate is part of a Timiskaming-type clastic and volcanic sequence that unconformably overlies the older volcanic strata. Deposition of the Midway sequence required uplift, subaerial erosion, continental volcanism, and deposition in isolated basins along a major structural break.

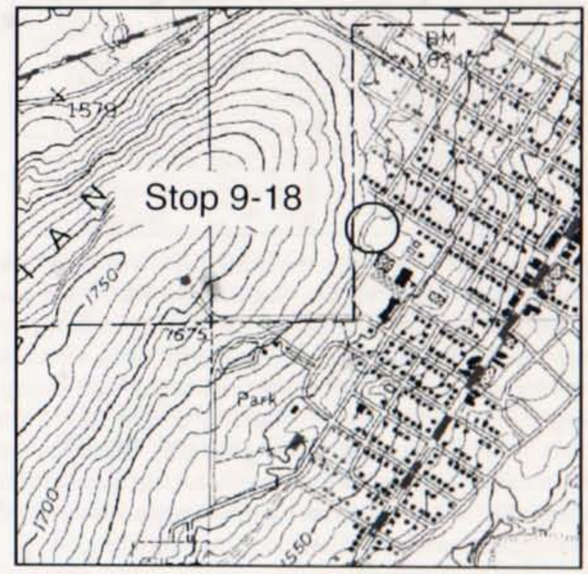
NEXT: Return to Highway 135. Turn right and follow 135 approximately 2.4 miles to a residential street on the northwest side of the town of Gilbert. Turn right and go 5 blocks and park by Gilbert Junior High School.

STOP 9-18

Archean pillowed and massive greenstone

Location: T. 58 N., R. 17 W., sec. 23, NW, SE, SW; north edge of athletic fields, Gilbert Junior High School

Gilbert quadrangle; UTM: 539,820E/5,259,750N



Description: This outcrop of pillowed and massive basalt is part of the Archean Mud Lake sequence, metamorphosed to low greenschist-grade. Pillow shapes indicate stratigraphic facing is to the northwest, which places this outcrop on the south side of a major D_1 structure known as the Mud Lake syncline. Note also the presence locally of fractures filled with reddish jasper, presumably deposited in depressions on the rock surface by overstepping of Paleoproterozoic seas during deposition of the Biwabik Iron Formation.

NEXT: Follow residential roads to State Highway 37 in the center of Gilbert. Turn right and travel on

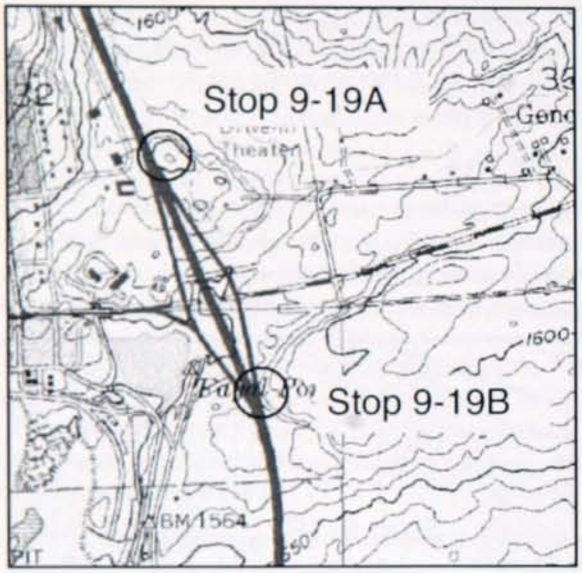
37 approximately 3.3 miles to the on/off ramps onto Highway 53.

STOP 9-19

Paleoproterozoic Pokegama Quartzite (A) and Biwabik Iron Formation (B)

Location: T. 58 N., R. 17 W., sec. 32, SE, SE, and adjacent, junction of Highways 37 and 53

Eveleth quadrangle; UTM: Scattered outcrops extend from 535,956E/5,256,913N on the north (Stop 9-19A), to 536,263E/5,256,200N on the south (Stop 9-19B)



Description 9-19A: Unconformably overlying the Neoarchean rocks of the Virginia horn area are the Animikie Group sediments of Paleoproterozoic age. Coarse grain size and massive beds as thick as 1.5 meters characterize this outcrop of the sandy, upper member of the Pokegama Quartzite. Thin beds of shale and siltstone separate the massive beds. Ojakangas (1993) interpreted the deposition of this facies to be within a high-energy, lower tidal or subtidal environment.

Description 9-19B: This exposure of gently southeast-dipping strata is part of the Lower Cherty member of the Biwabik Iron Formation. It overlies and is generally in transition with the Pokegama Quartzite at Stop 9-19A. Notice that both formations have sandy textures and cross-bedding, implying a moderately high-energy depositional environment. The most significant difference between these two units is the abrupt change in sediment source from the extrabasinal quartz grains in the Pokegama Quartzite, to recycled, chemically precipitated chert in the Biwabik Iron Formation. Measurements of cross-bedding in the iron-formation are bimodal, implying deposition in a tidally influenced marine environment (Ojakangas, 1993).

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FIELD TRIP 10

Saturday, May 21 – Sunday, May 22

THE WESTERN MARGIN OF THE KEWEENAWAN MIDCONTINENT RIFT SYSTEM: GEOLOGIC HIGHLIGHTS OF ARCHEAN, PALEOPROTEROZOIC, MESOPROTEROZOIC, AND PALEOZOIC BEDROCK IN EASTERN MINNESOTA AND NORTHWESTERN WISCONSIN

Leaders
Terry Boerboom, Minnesota Geological Survey
Daniel Holm, Kent State University
Laurel Woodruff and Bill Cannon, U.S. Geological Survey
Karl Wirth, Macalester College

INTRODUCTION

This field trip will span a wide variety of rock types and ages, ranging from Neoarchean granitic gneiss to Paleozoic sandstone, that occur adjacent to, within, or on top of the Mesoproterozoic Midcontinent rift system. Unconformities beneath rock units of Paleoproterozoic, Mesoproterozoic, and Paleozoic age will be examined. Trip localities include a mixture of places rarely visited as well as some popular field trip destinations.

The first day we will visit progressively younger rocks, beginning in Archean gneiss, continue through Paleoproterozoic metasedimentary and metavolcanic rocks, and end with the Mesoproterozoic sedimentary, volcanic, and intrusive rocks of the Keweenawan Supergroup. The second day will focus on sedimentary and volcanic rocks of the Keweenawan Supergroup, basal Paleozoic strata, and end by examining Quaternary scour features formed in Mesoproterozoic bedrock. The following discussion gives a brief summary of the geologic setting of the area in and adjacent to the Midcontinent rift system.

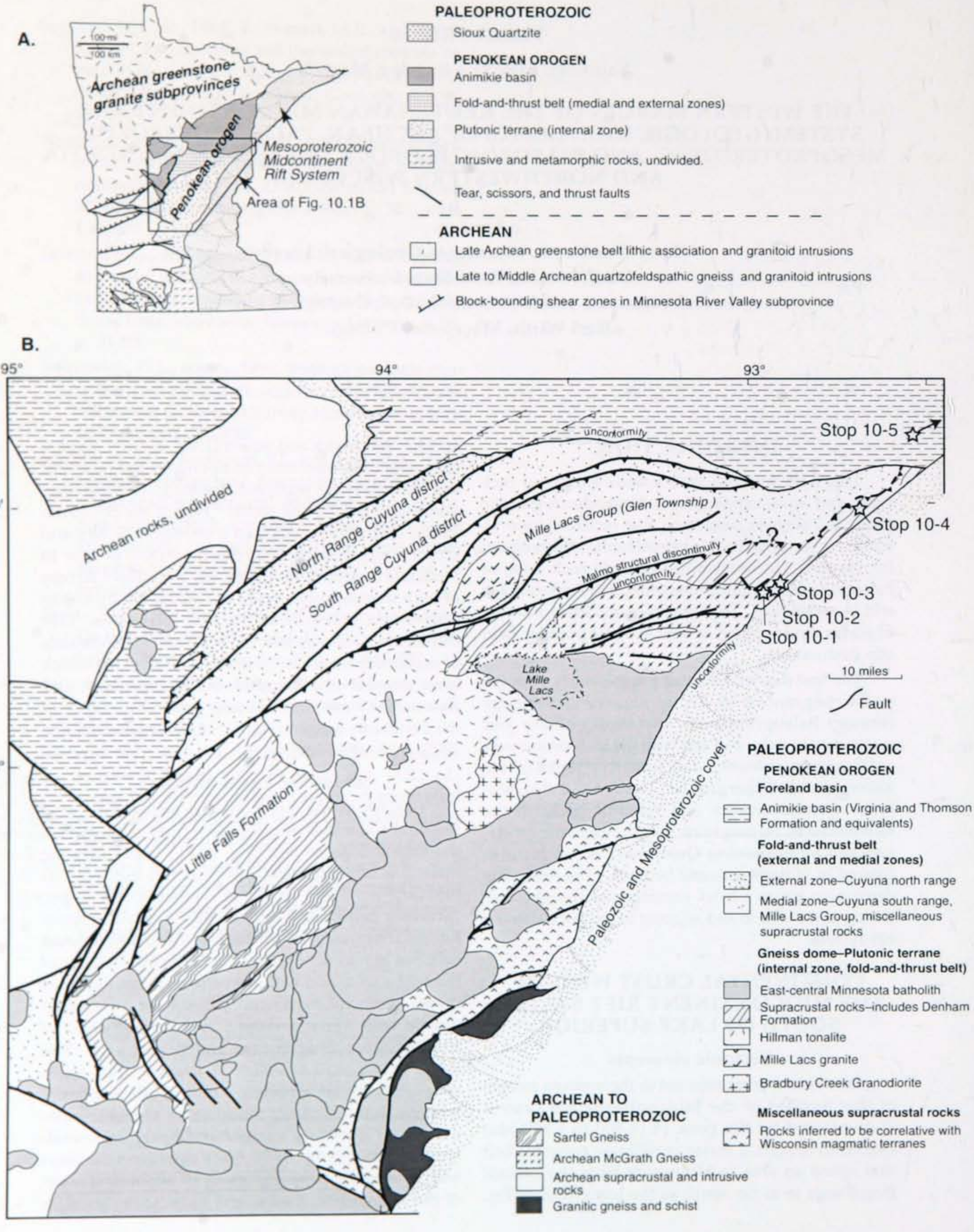
CONTINENTAL CRUST WEST OF THE MIDCONTINENT RIFT SYSTEM SOUTH OF LAKE SUPERIOR

Tectonic elements

Continental crust adjacent to the western margin of this portion of the Midcontinent rift system is dominated by the geon 18 (1,800 to 1,900 Ma) Penokean orogen, a major curvilinear orogenic belt that spans an area in Minnesota from the Mesabi Iron Range to as far south as the Iowa border (Fig. 10.1A). Southwick and others (1988) initially divided the orogen into two main components: the Animikie basin to the north (external foredeep), and the fold-and-thrust belt to the south. Within the fold-and-thrust belt, they recognized a stratigraphically and structurally complex northern terrane of low to moderate metamorphic grade, and a southern terrane of high metamorphic grade characterized by gneiss domes and voluminous granitic intrusions. Our research now shows that the southern zone consists of metamorphic rocks (gneisses and schists), which were pervasively intruded, metamorphosed, and then rapidly exhumed during geon 17 collapse of the Penokean orogen (Fig. 10.1B; Holm and others, 2005; also see Field Trip 4).

Animikie basin foredeep

The Animikie basin is conceptualized as a foreland basin that developed in front of (north of) the fold-and-thrust belt (Fig. 10.2). It includes the main depositional basin as well as two outliers that probably represent erosional remnants of a larger, formerly continuous mass. The Animikie basin formed after initial deformation of the fold-and-thrust belt, but in part synchronous with thrust stacking of the fold-and-thrust belt onto the Archean Superior craton. Strata of the Animikie basin unconformably overlie both Archean crust and deformed rocks of the Penokean orogen fold-and-thrust belt. The strata at the northern edge of the Animikie basin are essentially undeformed, with a gentle south dip, whereas at the southern margin they are folded into a series of generally upright and open, east–west-trending folds (Stop 10-5). Slaty cleavage associated with folding first appears about 15 kilometers south of the Mesabi Iron Range, and increases in strength


A.
100 mi
100 km
Archean greenstone-granite subprovinces
Penokean orogen
Mesoproterozoic Midcontinent Rift System
Area of Fig. 10.1B
PALEOPROTEROZOIC
Sioux Quartzite
PENOKEAN OROGEN
Animikie basin
Fold-and-thrust belt (medial and external zones)
Plutonic terrane (internal zone)
Intrusive and metamorphic rocks, undivided.
Tear, scissors, and thrust faults
ARCHEAN
Late Archean greenstone belt lithic association and granitoid intrusions
Late to Middle Archean quartzofeldspathic gneiss and granitoid intrusions
Block-bounding shear zones in Minnesota River Valley subprovince
B.
95°
94°
93°
46°30'
46°
45°30'
Stop 10-5
Stop 10-4
Stop 10-3
Stop 10-2
Stop 10-1
unconformity
North Range Cuyuna district
South Range Cuyuna district
Mille Lacs Group (Glen Township)
Archean rocks, undivided
Malmo structural discontinuity
Lake Mille Lacs
Little Falls Formation
Paleozoic and Mesoproterozoic cover
10 miles
Fault
PALEOPROTEROZOIC
PENOKEAN OROGEN
Foreland basin
Animikie basin (Virginia and Thomson Formation and equivalents)
Fold-and-thrust belt (external and medial zones)
External zone–Cuyuna north range
Medial zone–Cuyuna south range, Mille Lacs Group, miscellaneous supracrustal rocks
Gneiss dome–Plutonic terrane (internal zone, fold-and-thrust belt)
East-central Minnesota batholith
Supracrustal rocks–includes Denham Formation
Hillman tonalite
Mille Lacs granite
Bradbury Creek Granodiorite
Miscellaneous supracrustal rocks
Rocks inferred to be correlative with Wisconsin magmatic terranes
ARCHEAN TO PALEOPROTEROZOIC
Sartel Gneiss
Archean McGrath Gneiss
Archean supracrustal and intrusive rocks
Granitic gneiss and schist

southward in conjunction with progressively tighter folds (Southwick and others, 1998).

The northern margin of the Animikie basin is marked by a thin basal quartzite (the Pokegama Quartzite) that onlaps circa (ca.) 2,700 Ma Archean granite-greenstone terrane to the north (Fig. 10.2). The Pokegama Quartzite is overlain by the Biwabik Iron Formation, which in turn is overlain by the Virginia Formation in stratigraphic continuity. On a regional scale, the Virginia Formation becomes thicker and more coarse-grained to the south, and merges into the Thomson Formation (Stop 10-5). The Animikie Group unconformably overlies dikes of the Kenora–Kabetogama dike swarm (Southwick and Day, 1983), which have been dated at ~2,120 to 2,067 Ma (Schmitz and others, 1995; Schmitz, unpub. data).

Fold-and-thrust belt

The fold-and-thrust belt (Fig. 10.2) shows decreasingly lower metamorphic grades and structural complexity (indicating decreasing depth of tectonic burial) from south to north. It contains a complex assemblage of thrust-stacked, folded and faulted metasedimentary and metavolcanic rocks that include associated hypabyssal mafic sills, which overall are of low to moderate metamorphic grade. The medial zone (Fig. 10.1) includes the South range of the Cuyuna district and a broad area that extends from the Moose Lake area southwest as far as Todd County, including the Mille Lacs Group (Morey, 1978). Iron-formations in the medial zone are closely associated with mafic volcanic rocks and euxinic shale. The external zone, which forms the northwesternmost panel and includes the Cuyuna district North range (Fig. 10.1), is composed of folded and weakly metamorphosed strata composed dominantly of sedimentary rocks including the Mahnomen, Trommald, and Rabbit Lake Formations. Iron-formations in the external zone (mainly the Trommald Iron Formation) are associated with fine-grained argillaceous rocks. Aeromagnetic data and geophysical models clearly indicate that tightly folded strata of the Cuyuna North range district continue to the east beneath unconformably overlying strata of the Animikie Group. This relationship has been verified by geophysical models (Carlson, 1985).

Gneiss dome–plutonic terrane

This region (the southern internal zone of Southwick and others, 1988) contains relatively high-grade schistose rocks (such as the Little Falls Formation and Denham Formation; Stops 10-2B and 10-2C) but is dominated by a series of overlapping granitoid intrusions recently termed the east-central Minnesota batholith (Fig. 10.1; see Field Trip 4 in this volume; for example Holm and others, 2005; Chandler and others, in press). The internal zone also includes the Neoarchean McGrath Gneiss (Stop 10-1A), a moderately foliated, metamorphosed, porphyritic granite that yields a U-Pb zircon age of 2,557 ± 15 Ma (Holm and others, 2005). Although originally deformed and metamorphosed during the

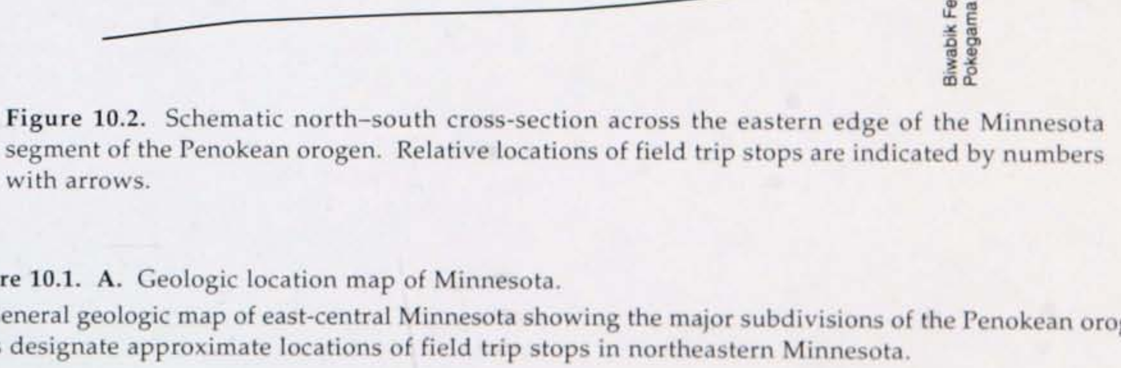


Figure 10.2. Schematic north–south cross-section across the eastern edge of the Minnesota segment of the Penokean orogen. Relative locations of field trip stops are indicated by numbers with arrows.

Figure 10.1. A. Geologic location map of Minnesota.

B. General geologic map of east-central Minnesota showing the major subdivisions of the Penokean orogen. Stars designate approximate locations of field trip stops in northeastern Minnesota.

Penokean orogeny, this area was strongly overprinted by geon 17 plutonism and metamorphism prior to its exhumation.

Malmo structural discontinuity

The gneiss dome–plutonic terrane is separated from the Penokean fold-and-thrust belt by the Malmo structural discontinuity (originally termed the Malmo thrust), inferred to be a major, south-dipping fault (Wunderman and Young, 1987). The Malmo structural discontinuity (Figs. 10.1, 10.2) separates relatively low-grade metamorphic rocks to the north from relatively high-grade metamorphic and igneous rocks to the south. To the west, the Malmo structural discontinuity is clearly marked by the sharp truncation of linear aeromagnetic anomalies associated with thin iron-formations in the Cuyuna South range and surrounding area, indicating that the "internal" zone containing the Little Falls panel was thrust west–northwest over lower-grade rocks of the medial zone. West of Mille Lacs Lake, the Malmo structural discontinuity juxtaposes post-Penokean plutons to the south against older metamorphic rocks to the north. Also, recent metamorphic geochronology indicates that the ages of metamorphism differ across the western Malmo structural discontinuity (geon 18 to the north and geon 17 to the south). These data all imply that the Malmo structural discontinuity is a post-Penokean geon 17 structure along which the gneiss dome–plutonic terrane (or the Penokean orogen internal zone) was uplifted and subsequently exhumed.

East of Mille Lacs Lake, the location of the Malmo structural discontinuity is more obscure due to the lack of contrast in geophysical data. Historically, the eastern end of the Malmo structural discontinuity has been placed at the northern margin of the McGrath Gneiss. However, more recent mapping (for example Boerboom and others, 1999) indicated that the Malmo structural discontinuity may lie a considerable distance north of the McGrath Gneiss. The graywacke unit (Stop 10-3), which overlies the Denham Formation (Stop 10-2), contains many of the same features as the Little Falls Formation, including similar metamorphic grade and abundant calcareous concretions, and if this correlation is correct, the Malmo structural discontinuity must lie north of the Denham Formation. An alternative explanation may be that vertical displacement along the Malmo structural discontinuity is less in the east than in the west, where exhumation was greater. This explanation is supported by the fact that geon 17 metamorphism is not recorded in metasedimentary rocks north of the McGrath Gneiss dome (McKenzie, 2004). Overall, when the east-central Minnesota batholith intruded, the region east of Mille Lacs Lake was likely at a shallower crustal level than the region west of Mille Lacs Lake. Differential vertical exhumation along the Malmo structural discontinuity (greatest in the west and less in the east) might explain both the different geophysical nature of the Malmo structural discontinuity and the different metamorphic age pattern across it from west to east.

LITHOLOGIC UNITS

McGrath Gneiss

The Neoarchean McGrath Gneiss is exposed in several localities between the Denham area and Mille Lacs Lake to the west. Although this unit locally contains layered gneissic structure, it is best characterized as a metamorphically foliated porphyritic granite. The 2,557 Ma McGrath Gneiss is only slightly younger than the undeformed 2,600 Ma Sacred Heart Granite of the Minnesota River Valley in southwestern Minnesota (Doe and Delevaux, 1991). The strong sub-solidus fabric that exists in the McGrath Gneiss is absent in the nearby 2,009 Ma Mille Lacs Granite (Holm and others, 2005). This suggests that the fabric of the McGrath Gneiss was developed prior to the Penokean orogeny, perhaps during continental rifting, which began about 2,100 Ma and ultimately led to the formation of a south-facing continental margin in this region during Paleoproterozoic time (Roscoe and Card, 1993). The discrete shear bands and recrystallization present in the Mille Lacs Granite (Boerboom and others, 1999) and the steep folding of the McGrath Gneissic fabric around sub-vertical, east–west-oriented axial planes are almost certainly the result of Penokean deformation (Holm and others, 1988). Holm and Lux (1996) interpreted the basement-cover contact of the McGrath Gneiss dome to be a detachment fault because of an apparent 50 m.y. cooling age difference between basement and cover rocks. However, more recent detailed argon ion laser data show that the basement and cover rocks have essentially the same lower temperature thermal history (Fig. 10.3; McKenzie, 2004). Also, as discussed under Stop 10-1, field and petrographic evidence (Boerboom and others, 1999) indicates that the McGrath Gneiss had a saprolite developed on it at the time of deposition of the overlying Denham Formation. Together, these data indicate that the boundary is simply a domed, nonconformable contact.

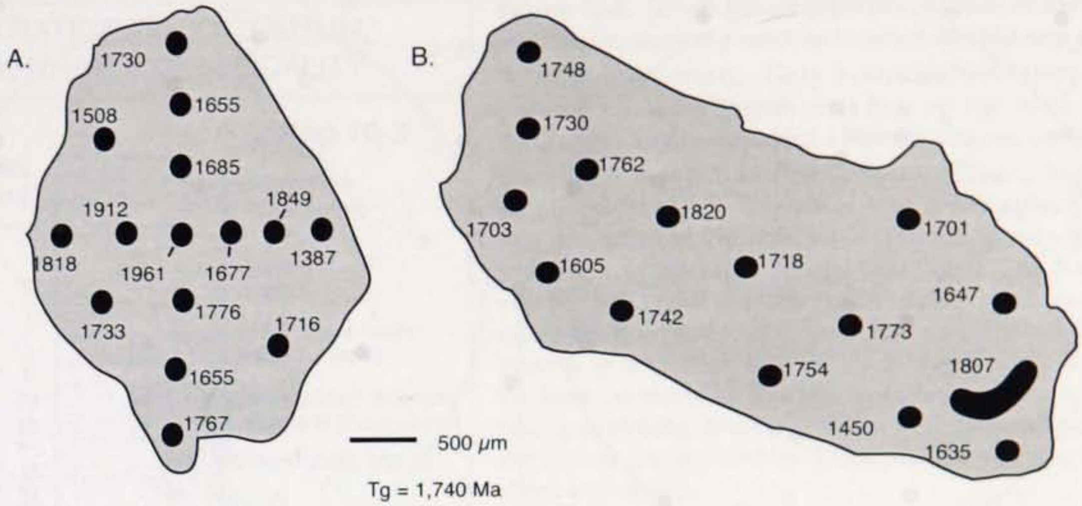


Figure 10.3. Spot age data (in Ma) of muscovite grains from argon ion laser data (McKenzie, 2004).

A. McGrath Gneiss basement near Stop 10-1.

B. Metapelite cover rocks at Stop 10-3.

Denham Formation

The Denham Formation (Morey, 1978) is composed of a heterogeneous mixture of interbedded pelitic to arenitic arkosic sedimentary rocks, calc-alkaline pillow basalt, dolostone, and fragmental volcanic rocks. This sequence unconformably overlies the Archean McGrath Gneiss (Stop 10-1), and has been subjected to amphibolite-facies metamorphism. Drill holes (cores and cuttings) from north and east of the Denham locality (Stop 10-2) show that the Denham Formation is transitional upward into graywacke (for example Stop 10-3), with the transition marked by a 1-meter-thick zone of graphitic argillite. Aeromagnetic data indicate that the Denham Formation extends westward along the north edge of the McGrath Gneiss to Mille Lacs Lake. Incomplete drilling records and corroborating aeromagnetic anomalies indicate that remnants of the Denham Formation are present within the McGrath Gneiss as infolded remnants.

The age of the Denham Formation is constrained by the underlying McGrath Gneiss (2,557 ± 15 Ma; Holm and others, 2005). The Denham Formation and overlying pelitic schist and metagraywacke (Stop 10-3) are part of the Mille Lacs Group (Morey, 1978), which is cut by a granite stock that yields a U-Pb zircon age of 2,009 Ma (Holm and others, 2005). Volcanic rocks in the Denham Formation have provided a Sm-Nd isochron age of 2,197 ± 39 Ma (Beck, 1988). Thus, the age of the Denham Formation is constrained to be somewhere between 2,550 to 2,009 Ma. Metamorphic monazite from pelitic schist at Stop 10-3 yields age domains (from 85 spot analyses) at 1,840 and 1,800 Ma (McKenzie, 2004).

Rocks of the Denham Formation have undergone regional amphibolite-grade metamorphism and at least two periods of deformation attributed to the Penokean orogeny. The first deformation event was synchronous with metamorphism to the garnet zone of the amphibolite facies (Holm, 1986). It produced an early S_1 foliation that is typically bedding-parallel, and a locally strong, shallowly plunging, stretching lineation (Fig. 10.4A). S_1 foliation and S_0 bedding were subsequently folded along steeply dipping axes (Fig. 10.4B), concurrent with or followed by peak metamorphism that produced staurolite. In the Denham valley (Stop 10-2), the stratigraphic sequence dips variably to the north, having local F_2 folds with overturned limbs. North of the valley (Stop 10-3), bedding and S_1 in graywacke are nearly horizontal, and are deformed into open F_2 folds with local crenulation features. Farther north the bedding dips south, defining a broad, regional-scale, upright F_2 syncline.

Despite deformation and metamorphism, the stratigraphy of the Denham Formation forms a coherent package that is shown schematically in Figures 10.5 and 10.6. Omitting the prefix "meta" for clarity, the base of the Denham Formation consists of interbedded siltstone and cross-stratified pebble conglomerate. This is overlain by coarse-grained and locally conglomeratic arkose that apparently pinches out laterally. The primary clastic grains in

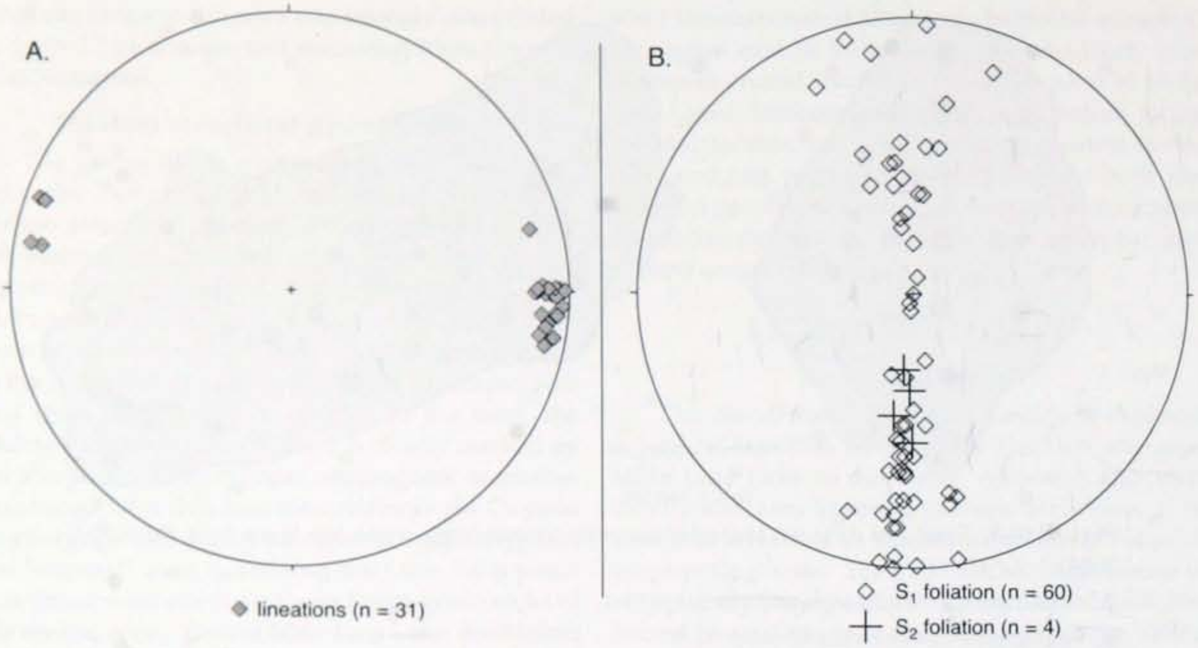


Figure 10.4. Equal area, lower hemisphere projections of structural elements of the Denham Formation and overlying metagraywacke unit to the north.

A. Measured lineations.

B. Poles to measured foliations. Open diamonds represent the main, early foliation, which has been refolded by a second deformation that produced only a weak north-dipping foliation (crosses).

the dolomitic arkose are well preserved owing to the abundance of dolomite in the matrix, which absorbed most of the deformation. The arkose is interbedded with amygdaloidal basalt flows that contain well-preserved primary macroscopic flow features. The flows grade stratigraphically upward (northward) from massive bases to pillowed interiors to fragmental upper portions. The volcanic flows are thickest at the eastern outcrop limit, where at least four flows of nearly 1,000 feet total thickness were recognized, and thin westward to two flows of 300 feet total thickness. An upper sequence of mafic fragmental volcanic rocks nearly merges to the east with the pillow basalts (Fig. 10.6), possibly implying that later stages of volcanic activity were more explosive in nature and that both volcanic horizons emanated from a common center to the east, which may presently be buried beneath the Mesoproterozoic Fond du Lac Formation. Arkosic and pelitic strata (now staurolite-garnet-mica schist) between the flows and fragmental volcanic sequence pinch out to the east where the flow package thickens, and are not present in drill holes to the north and east of the Denham valley. The northernmost outcrops in the valley consist of very pure dolomite, now marble, which contains ptygmatically folded and strongly lineated quartz veins. Drill cores show that the dolomitic marble is at least 500 feet thick, and is overlain by graywacke that is exposed discontinuously to the north for some distance (for example Stop 10-3). A thin layer of graphitic argillite marks the contact between the dolomite and overlying graywacke.

Field and petrographic observations imply that clastic detritus in the Denham Formation was derived in large part from a weathering residuum on the subjacent McGrath Gneiss. Near the contact with the Denham Formation, the McGrath Gneiss grades abruptly from granite gneiss containing quartz, orthoclase, plagioclase, and biotite, to strongly foliated, quartz- and sericite-rich schist that contains orthoclase, but no plagioclase. The arkosic parts of the Denham Formation similarly lack plagioclase, and are composed of quartz and orthoclase grains, together with small clasts of granitic gneiss. Studies of saprolite developed beneath Cretaceous sedimentary rocks on Precambrian crystalline rocks in southwestern Minnesota may provide an analog (Setterholm and others, 1989). These studies demonstrated that plagioclase is one of the first minerals to alter to kaolinitic clay during the weathering process, and that orthoclase and quartz are the most resistant to weathering. The basal

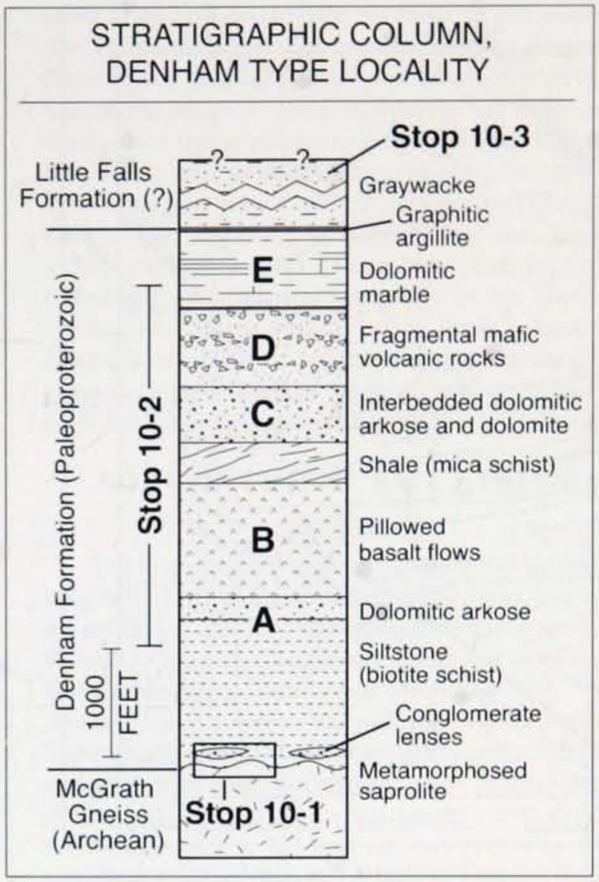


Figure 10.5. Schematic stratigraphic section of the Denham Formation, showing relative location of field trip stops.

A. (Southernmost). Gray biotite schist interpreted as metamorphosed siltstone (unit Epds on Fig. 10.6), overlain by brownish-tan weathered dolomitic arkose (unit Epda).

B. Pillow basalt flows (unit Epdb). Classic basalt flow sequences, with massive bases that grade upward (north) into amygdaloidal pillow basalt, which in turn grades to coarsely fragmental or blocky flow tops. The fragmental and amygdaloidal character of the flows indicates eruption into a shallow-water environment. Two distinct flows are present in the valley, but in hills to the east where the unit thickens, as many as four flow sequences were identified. Strong stretching lineation evident in fragmental portions of the flow is accentuated by differential weathering between clasts and matrix.

Outcrop gap is inferred to represent garnet-staurolite-mica schist that is exposed both east and west of this gap; unit Epdg.

C. Dolomitic arkose (unit Epda). Weathered reddish-tan color is due to abundant dolomite in the matrix. Coarse sand-sized detrital grains of orthoclase and quartz, and rare cobbles identical to the McGrath Gneiss are present. The original clastic grains in the arkosic rocks are well preserved despite metamorphism, largely because strain has been partitioned into the comparatively more ductile dolomitic matrix. As discussed at the McGrath Gneiss stop (10-1), the arkosic rocks contain almost no plagioclase, presumably due to sediment derivation from a weathered McGrath Gneiss source.

D. Fragmental volcanic rocks (unit Epdv). Dark green amphibolitic clasts and scattered clasts of white cherty material are strongly flattened and lineated. On a map view, this fragmental unit nearly merges to the east with the underlying pillow basalt, possibly indicating a common volcanic source located to the east beneath the Keweenawan Fond du Lac Formation.

E. Dolomitic marble (unit Epdm). Tan where weathered, dark gray where fresh. Drill holes (cores and cuttings) to the east intercept as much as 500 feet of massive dolomite and show that the marble changes from gray to white with depth.

Cretaceous strata locally consist of reworked saprolite, including beds of cross-stratified sandstone and nearly pure kaolinitic shale. Exposures of basal Cretaceous sedimentary rocks locally contain detrital orthoclase and quartz derived by slight reworking of weathered granite grus, set in a carbonate matrix. We infer that similar processes occurred in Paleoproterozoic time by erosion and reworking of weathered McGrath Gneiss into beds of calcareous arkose and kaolinitic shale. These were subsequently metamorphosed to produce recrystallized dolomitic arkose and staurolite-garnet-sericite schist. Slight weathering of orthoclase may have liberated potassium for the inferred conversion of kaolinite to sericite during metamorphism.

The Denham Formation is interpreted to represent a rift-margin assemblage deposited during the Paleoproterozoic era that is genetically similar to, and perhaps temporally equivalent to, the Chocolay Group in Michigan. In this setting, the McGrath Gneiss was part of the continental margin that was weathered and eroded to provide detritus to an evolving rift basin undergoing active, shallow-water volcanism. Interbedded arkose and dolomite higher in the stratigraphic section represent foundering of the shelf and deepening water, possibly by subsidence of localized grabens. The paucity of clastic material in the upper, dolomite-dominated part of the sequence indicates that deposition of coarse-grained detritus

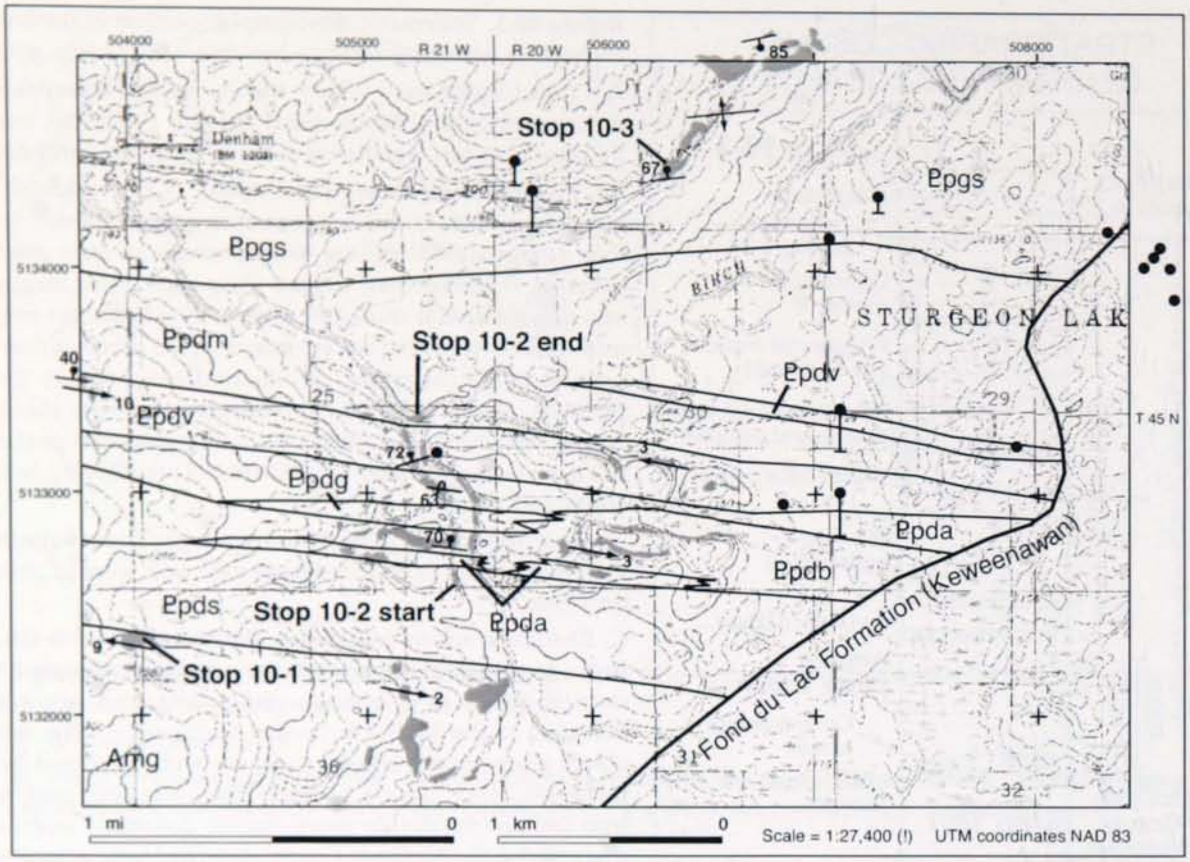


Figure 10.6. Bedrock geologic map of the area around the type locality of the Denham Formation, showing stops. Black circles are locations of exploratory drill holes and gray areas are mapped outcrops. McGrath Gneiss—Amg; Denham Formation—Epds (metasiltstone), Epda (meta-dolomitic arkose), Epdb (metabasalt), Epdg (metapelite), Epdv (meta-fragmental mafic volcanic rocks), Epdm (dolomitic marble), Epgs (metagraywacke).

was restricted to the shallow, nearshore environment adjacent to the McGrath Gneiss. The stratigraphically upward sedimentological gradation of dolomite to graywacke indicates further deepening water and associated turbidite deposition. The deformation of the Denham Formation is inferred to be the product of basin closure during the Penokean orogeny.

Thomson Formation

The Thomson Formation (Stop 10-5) is situated at the southeastern margin of the Animikie basin (Figs. 10.1 and 10.2). It is correlative with the Virginia Formation, which is exposed along the Mesabi Iron Range at the northern margin of the Animikie basin. As reported by Morey (1978), the type locality of the Thomson Formation is composed of 34 to 46 percent graywacke, 27 to 39 percent siltstone, and 27 percent slate. The Thomson Formation has been metamorphosed to the greenschist facies, and is folded into a series of open, east-trending folds with axes that dip vertical to steeply south and plunge shallowly to the east and west. The Thomson Formation is cut by a series of northeast-trending diabase dikes that are presumed to be associated with the Midcontinent rift system. See the introduction for Field Trip 2 for more information about the dikes.

The Thomson Formation formerly included the higher-grade schistose metagraywacke that lies to the south, such as those visited in Stops 10-3 and 10-4. This correlation inferred that the metamorphic grade increased transitionally to the south, from greenschist-facies near Thomson Dam to amphibolite-grade near Moose Lake and Denham. However, Holst (1984) recognized a "southern structural terrane" in which metagraywacke and related metasedimentary rocks contain an early, near bedding-parallel, S_1 foliation and recumbent isoclinal folds. This early fabric is refolded along open and upright folds having east–west axial traces that plunge gently both east and west. The second generation of folds has the same style and orientation as those to the north in the Thomson Formation proper. In the Thomson Formation, folding has produced slaty cleavage, and in the southern terrane, folding has crenulated the S_1 cleavage to produce an S_2 cleavage. Based on this work, and subsequent mapping (for example Southwick and others, 1988; Boerboom and Chandler, 2001), the "southern structural terrane" is now interpreted to be part of the Penokean fold-and-thrust belt, unconformably overlain by the Thomson Formation. Furthermore, aeromagnetic data and lithologic attributes imply that the amphibolite-grade graywacke association to the south may be correlative with the Little Falls Formation to the southwest.

MIDCONTINENT RIFT SYSTEM

Overview

The Midcontinent rift system is a 2,500-kilometer-long belt of volcanic and sedimentary rocks that form a curvilinear feature extending northeastward from Kansas to the Lake Superior region, and continuing southeastward across lower Michigan (Fig. 10.7). Rocks in the rift system are well exposed in the Lake Superior region and their continuation to the southeast and southwest beneath Paleozoic cover rocks is clearly defined by gravity, magnetic, seismic, and drilling information. The rift and its exceptionally thick sequence of flood basalts formed at about 1,100 Ma during a period of lithospheric extension and copious basalt generation interpreted to have been caused by arrival of a new mantle plume at the base of the lithosphere (Hutchinson and others, 1990; Nicholson and Shirey, 1990; Cannon and Hinze, 1992; Nicholson and others, 1997). Nearly complete crustal separation in the Lake Superior region generated as much as 2 million cubic kilometers of basalt and possibly an equal volume of intrusive rocks in a relatively short period of 15 million years (Hutchinson and others, 1990; Nicholson and Shirey, 1990; Cannon and Hinze, 1992; Allen and others, 1995). Thermal subsidence following volcanism resulted in deposition of up to 10 kilometers of sediments in many parts of the central rift graben. The Midcontinent rift system transects a variety of older rocks that range from Archean to Paleoproterozoic in age (for example Stops 10-1 through 10-5). South of Lake Superior, the Midcontinent rift system is covered by Paleozoic sedimentary rocks (for example Stop 10-12; Fig. 10.7). Exposed rocks in western Lake Superior contain a remarkably complete record of igneous intrusions, flood basalt volcanism, and clastic sedimentation, and the evolution of the rift is well constrained by chemical and isotopic analyses, high precision U-Pb zircon dates, and comprehensive seismic imaging, both on-land and beneath Lake Superior. Comprehensive studies of the Midcontinent rift system have been published over the years, including Wold and Hinze (1982), Ojakangas and others (1997), a special volume of the Canadian Journal of Earth Sciences (v. 34, 1997), and a number of guidebooks, including Miller and others (1995a, b) and Wirth and others (1998).

St. Croix horst

Early extensional faults that formed the margins of subsiding, volcanic-filled grabens in the Midcontinent rift system were reactivated into reverse faults by late-stage compressional stresses that may have been the far-field products of convergence within the Grenville Province to the east (Cannon, 1994). This reversed displacement created a system of uplifted horsts along much of the length of the rift. In Wisconsin the central portion of the Midcontinent rift system was uplifted along a system of paired, reverse listric faults to produce the St. Croix horst, bounded on the west by the Douglas and Pine faults, and to the east by the Hastings-Atkins Lake and Cottage Grove-Lake Owen faults (Fig. 10.7). Originally, the horst may have been an asymmetric graben, or half graben. The Lake Owen fault was a major growth fault on the southeast side of the graben and the volcanic fill thickens toward and terminates against the fault as indicated by seismic and gravity data. The Douglas fault on the northwest margin is not clearly a growth feature and may be a thrust formed during rift inversion (Nicholson and Cannon, 2003). Gravity and seismic data indicate that the St. Croix horst contains a 10- to 20-kilometer-thick sequence of volcanic and sedimentary rocks (Chandler and others, 1989; Allen and others, 1997); this contrasts with the more than 30-kilometer section of basalts and sediments imaged beneath Lake Superior (Cannon and others, 1989) and indicates a greater degree of uplift and erosion of the St. Croix horst than elsewhere in the rift. The metamorphic grade of volcanic rocks in the St. Croix horst, particularly greenschist facies assemblages in volcanic rocks near the Clam Falls-Taylors Falls region (Fig. 10.7; Wirth and others, 1997), is higher than that typically noted near Lake Superior, and consistent with more uplift and exposure of rocks that once were deeply buried.

Outcrops within the St. Croix horst are sparse, but detailed geochemical and geophysical data define at least two volcanic units within the central graben. The older rocks are the Chengwatana Volcanics, composed of basalt, minor andesite, and rare rhyolite flows (Fig. 10.8). Flow thicknesses range from 3 to 100 meters thick. At least 6 interflow conglomerate units have

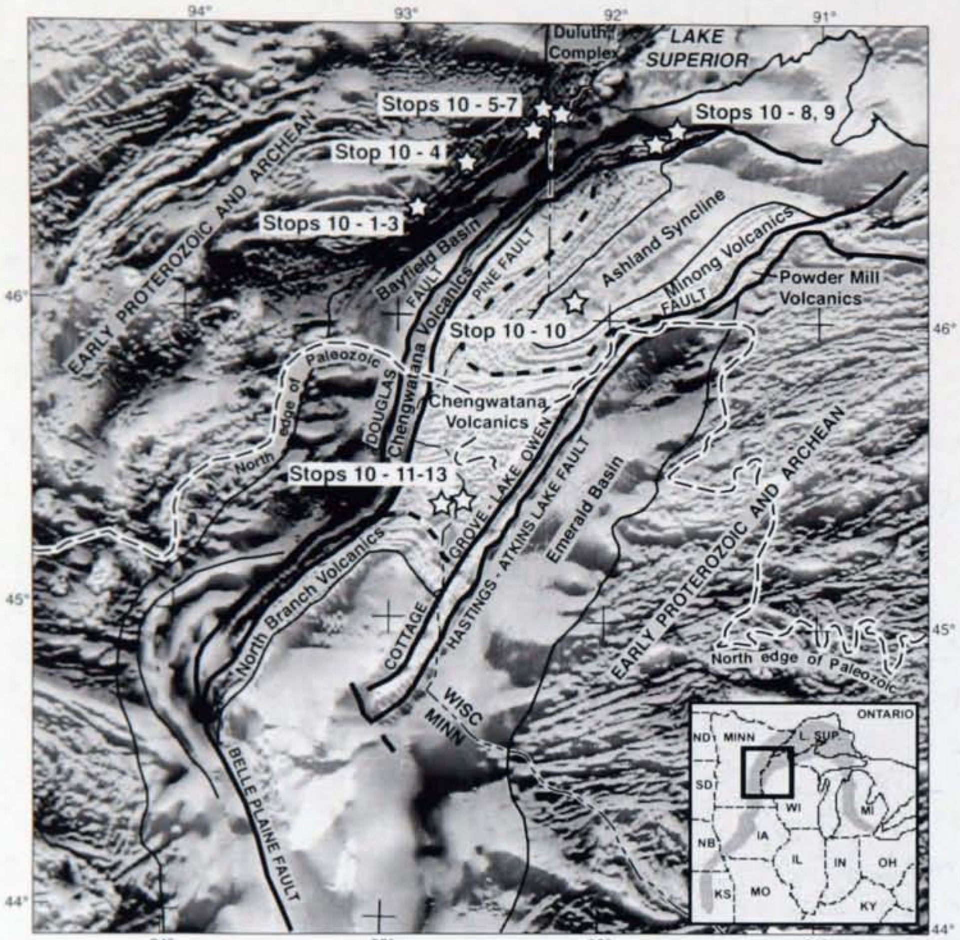


Figure 10.7. Aeromagnetic anomaly map of the St. Croix horst region in Minnesota and Wisconsin, showing subdivisions of the St. Croix horst volcanic sequences into Chengwatana and Minong Volcanics. General stop locations are indicated by stars. Inset map shows the general extent of the Midcontinent rift system, and the location of the larger figure. Modified from Cannon and others (2001).

been identified. In western Wisconsin, a rhyolite flow in the middle of the Chengwatana Volcanics is dated at 1,101.8 ± 6.7 Ma (Fig. 10.8; C.E. Isachsen, unpub. data, 2003). Two other rhyolite flows near the base of the exposed flow sequence of the Clam Falls region yielded a zircon U-Pb age of 1,102 ± 5 Ma (Wirth and Gehrels, 1998).

The younger Minong Volcanics are a sequence of low-TiO_2 flood basalts that overlie the Chengwatana Volcanics along a low angle disconformity apparent from aeromagnetic data (Fig. 10.7). The lower part of the Minong Volcanics on the southeast limb is mostly high-titanium basalt flows interlayered with andesite and rhyolite. The abundant high-TiO_2 basalt and a quartz-phyric rhyolite in the upper part of the Minong Volcanics, dated at 1,094.6 ± 2.1 Ma (Zartman and others, 1997), may indicate the presence of a localized magmatic center. The intrusive Amnicon Complex, located on the western side of the horst, may represent a second magmatic center that was coeval with the Duluth Complex to the northwest and the Mellen Complex to the southeast.

Volcanic rocks in the central horst are overlain by clastic sedimentary rocks of the Oronto Group. Only the lowermost unit, the Copper Harbor Formation, is preserved within the horst where as much as 2 kilometers of sandstone and conglomerate lie along the axis of the syncline. Along the margins of the horst, volcanic rocks are juxtaposed with clastic sediments of the Bayfield Group and equivalent sandstones. Flanking basins on either side of the horst contain some 4 to 5 kilometers of sedimentary

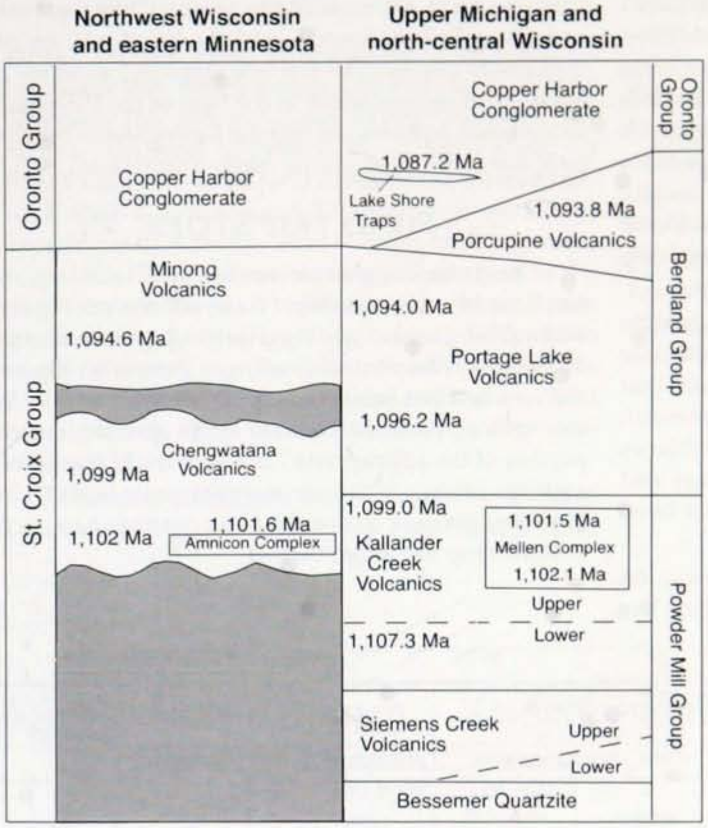


Figure 10.8. Stratigraphic column for Keweenawan volcanic rocks on the south shore of western Lake Superior. Modified from Nicholson and others (2001) by Suzanne Nicholson; see Nicholson and others (2001) for source of dates.

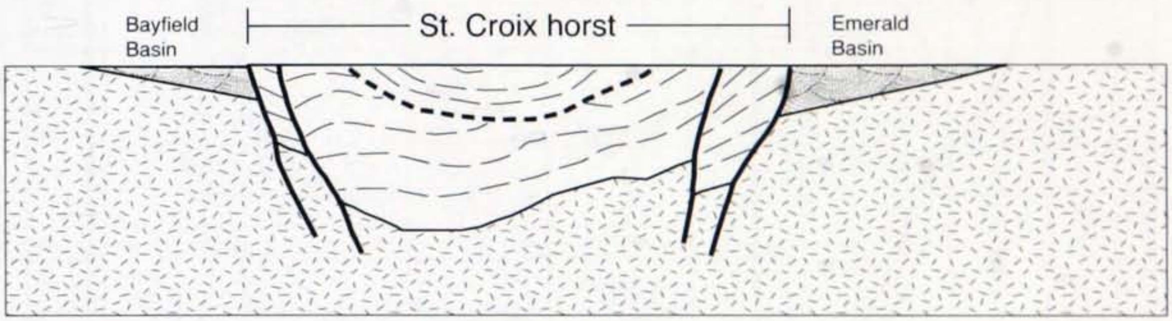


Figure 10.9. Schematic cross-section model of the Midcontinent rift system perpendicular to the St. Croix horst, across the southern tip of the Ashland Syncline. Basalt flows are designated by dashes. Modified from Cannon and others (2001).

fill (Fig. 10.9). The Bayfield basin, which includes the Fond du Lac Formation and Hinckley Sandstone in Minnesota and the Orienta, Devil's Island, and Chequamegon Sandstones of the Bayfield Group in Wisconsin, follows the western margin of the horst, and the Emerald Basin (Allen and others, 1997) lies along the eastern margin of the horst. Aeromagnetic and drilling data show that the southern part of the Chengwatana volcanic group between the Pine and Douglas faults (Fig. 10.7) is overlain by a substantially thick vestige of mineralogically immature sedimentary rocks of uncertain correlation that pre-date the reverse faults.

CAMBRIAN STRATA

Overview

During the late part of the Cambrian period, southeastern Minnesota was covered by an incursion of shallow epicontinental seas that transgressed onto a land surface of generally low relief. However, prominent knobs formed by resistant bedrock were present at a few localities, such as in the Taylors Falls–St. Croix Falls area (Stop 10-12). These bedrock highs acted as local sources for cobbles and boulders shed off of islands. Conglomeratic facies, locally known as the "Mill Street conglomerate," are found from the top of the Eau Claire Formation through the base of the Franconia Formation (Fig. 10.10).

The Mill Street conglomerate is known to contain an unusual assemblage of trilobites, inarticulate brachiopods, and monoplacophoran mollusks that represent an intertidal, near-shore environment; however, a detailed study of the fossil assemblages that would help to more precisely assign age and formation to the sedimentary strata has not been undertaken.

The second day of this field trip will focus on varied aspects of the St. Croix horst, including the Douglas fault, volcanic rocks in the Chengwatana and Clam Falls Volcanics, and the basal portion of the Oronto Group. The last two stops will examine a basalt-clast conglomerate at the base of the Paleozoic section, and potholes cut into the Keweenawan basalts by Pleistocene post-glacial meltwater floods.

FIELD TRIP STOPS

The general geologic setting and locations of stops in Minnesota (day 1) are shown on Figure 10.1. The general geologic setting and locations of stops in Wisconsin (day 2) are shown on Figure 10.7. A detailed location map is included with each stop description, with a base made at scale from a portion of the appropriate 7.5' quadrangle map. The regional setting of the various rock units is given in the introduction, and only facts pertinent to each specific stop are given below.

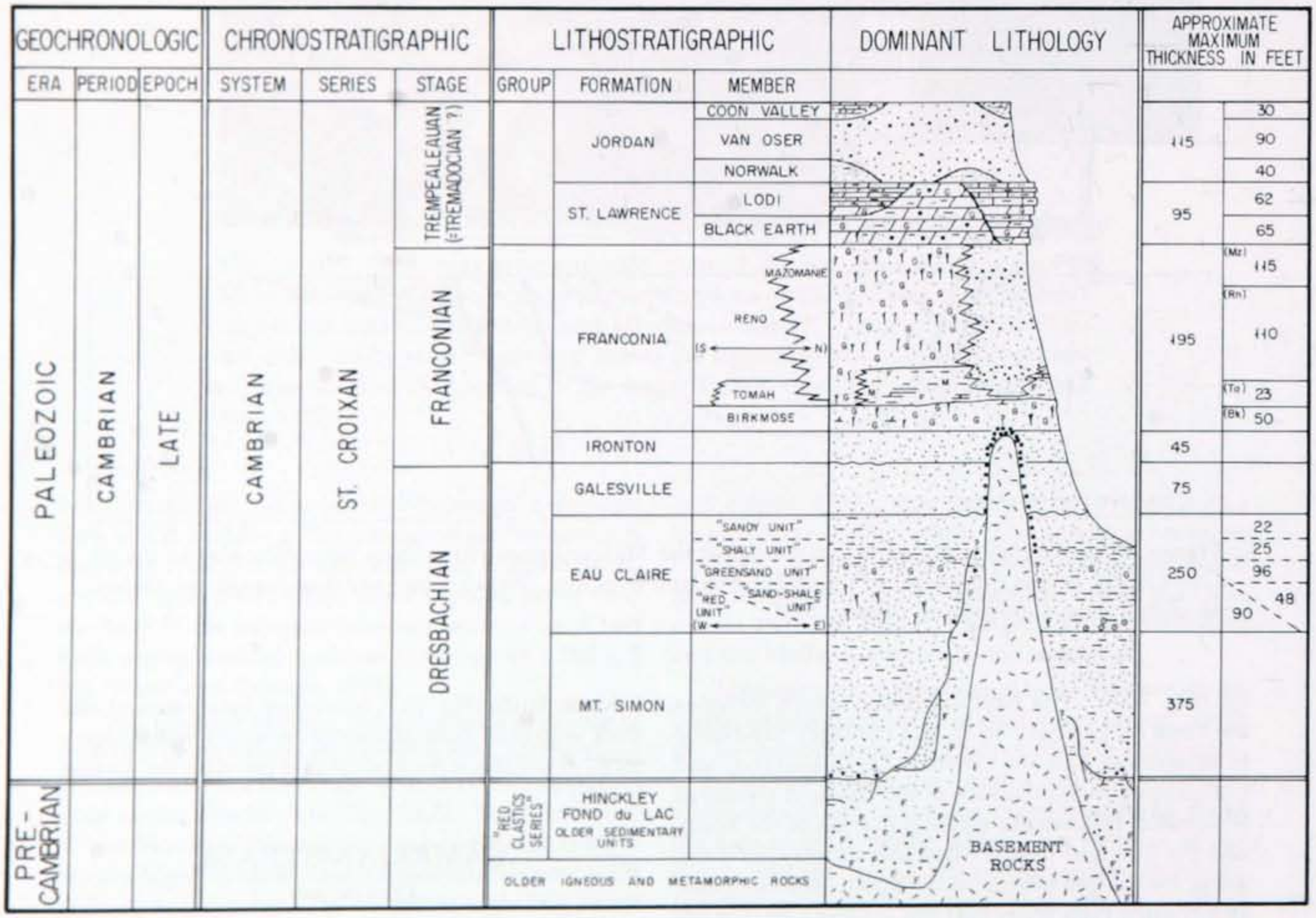


Figure 10.10. Stratigraphic column of Upper Cambrian rock units in Minnesota (from Mossler, 1987). Conglomeratic facies surrounding bedrock highs are shown by black dots.

DAY 1

DIRECTIONS: From Minneapolis drive north on Interstate 35W, which merges into I-35, to the Willow River exit (no. 205, a distance of approximately 105 miles). Go west to Willow River, turn north on Pine County Road 61 for 3 miles to the intersection with Pine County Road 52. Go west on 52 for approximately 6.2 miles (road changes from pavement to gravel). The road makes a right angle turn to the north toward the town of Denham, but instead turn south and walk or drive on the small dirt path to a gate. Proceed past the gate into an open pasture until the road crosses a gentle rise and the west end of an outcrop (Stop 10-2A). Outcrop is 0.65 mile south of 52.

Alternate directions are to go to the town of Denham, drive 0.7 mile south on Pine County Road 52, and continue south on a dirt track after 52 makes a 90° bend to the east.

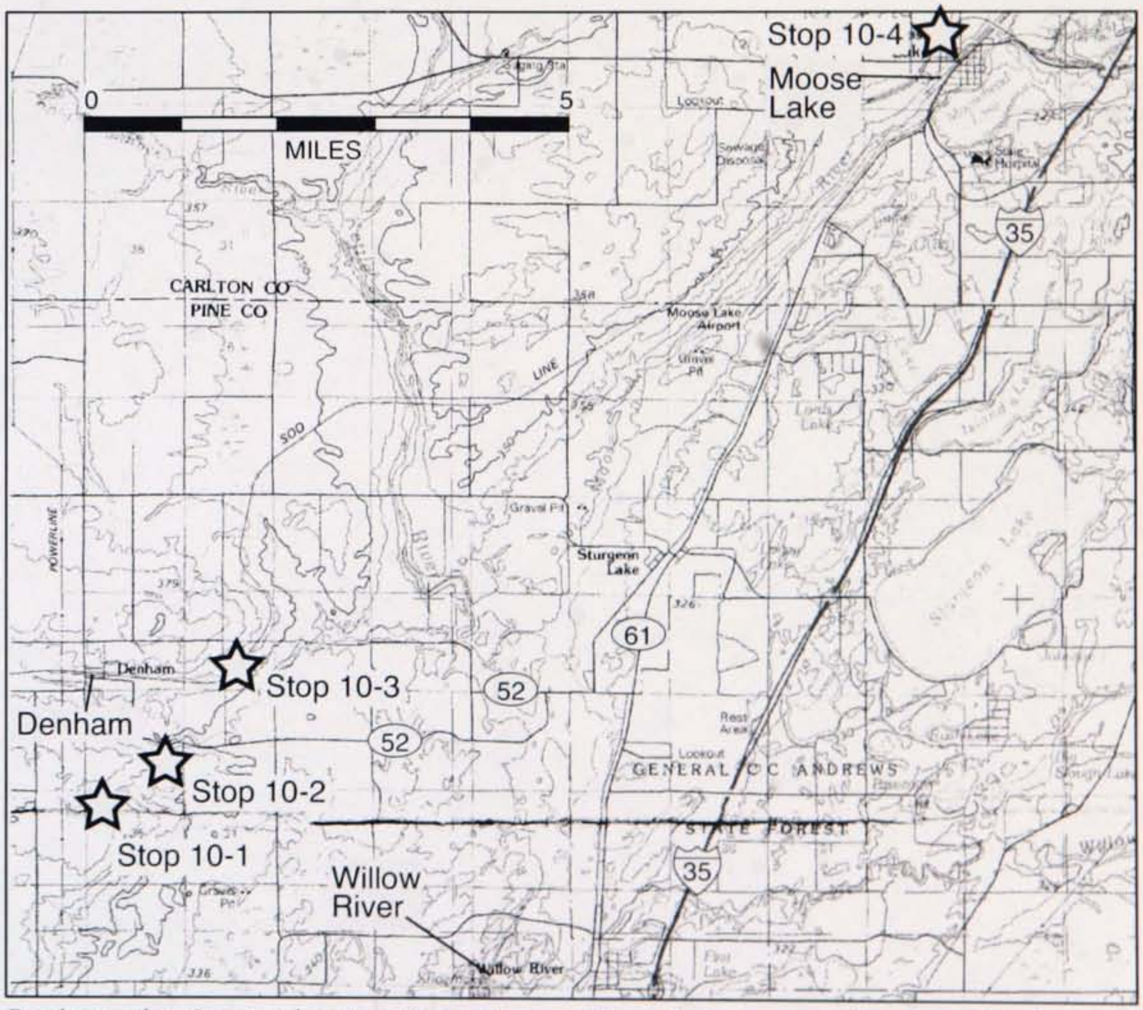
STOP 10-1

Private property! Permission must be obtained before entering!

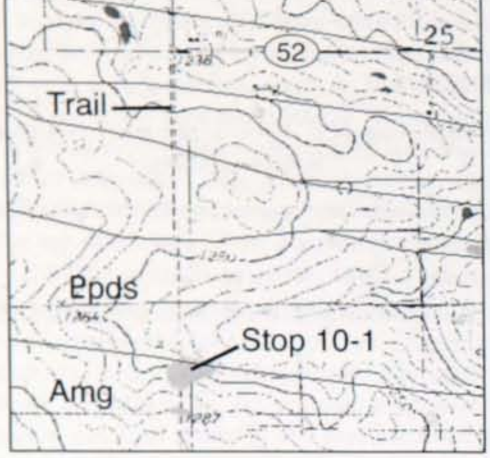
McGrath Gneiss and contact with overlying Denham Formation

Location: T. 45 N., R. 21 W., sec. 36

Denham quadrangle; UTM: 503,979E/5,132,367N



Road map showing stop locations in the Denham–Moose Lake area; southern leg of day 1.



Highlights: Archean McGrath Gneiss and metamorphosed saprolite (Amg), conglomerate at the base of the Paleoproterozoic Denham Formation (Ꞓpds)

Description: Stops 10-1, 10-2, and 10-3 are shown in apparent stratigraphic context on Figure 10.5. These stops will examine in order the north-facing transition from the Neoarchean McGrath Gneiss (Stop 10-1), to the overlying Paleoproterozoic Denham Formation (Stop 10-2), and graywacke that stratigraphically overlies the Denham Formation (Stop 10-3).

At this locality, the northernmost outcrops of the McGrath Gneiss grade abruptly from the typical mineral assemblage of quartz, orthoclase, plagioclase, and biotite to strongly foliated, quartz- and sericite-rich schist that contains coarse-grained relict orthoclase, but no plagioclase. The latter assemblage, present just below the base of the Denham Formation, may be the product of metamorphism of partially weathered granite, in which the plagioclase feldspar had weathered to clay, but quartz and potassium feldspar remained fresh.

Just north of the McGrath Gneiss is a thin layer of poorly exposed biotite schist of sedimentary protolith, which contains irregular lenses of coarse-grained conglomerate with abundant pebbles of quartz and pink orthoclase, but no plagioclase (unit Ꞓpds on Fig. 10.6). Similarly, the clastic component of the arkosic beds higher up in the stratigraphy of the Denham Formation (Stop 10-2B) is comprised almost entirely of quartz and orthoclase feldspar with essentially no plagioclase feldspar. Together, these observations are consistent with our preliminary interpretation that the McGrath Gneiss was weathered to the point that plagioclase had altered to clay minerals, and this weathered residuum acted as a source of detritus for the overlying Denham Formation.

NEXT: Drive east 0.75 mile on County Road 52 to a small gated trail to the south. Walk south down the trail for approximately 0.5 mile, following the edge of a ravine to the southernmost outcrop on west side of the ravine (see location map). This stop will traverse from south to north over outcrops located along the west edge of the ravine. There are also outcrops on the east side of the ravine that show parts of the stratigraphy not exposed on the western valley traverse.

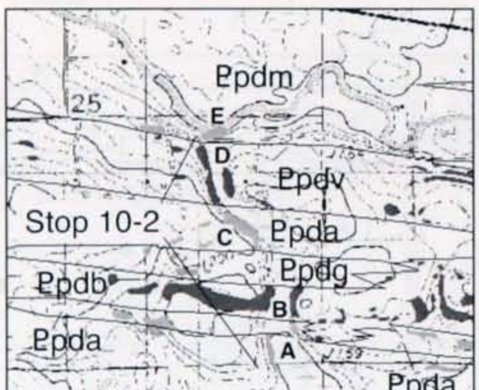
STOP 10-2

Private property! Permission must be obtained before entering!

Denham Formation

Location: T. 45 N., R. 21 W., sec. 25, SE

Denham quadrangle; UTM Start: 505,406E/5,132,575N; UTM End: 505,208E/5,133,330N



Highlights: Interbedded biotite schist (siltstone), pillow basalt, dolomitic arkose, and dolomitic marble (Fig. 10.5)

Description: From bottom to top, the Denham Formation consists of fine-grained metamorphosed siltstone (Ꞓpds), dolomite-cemented arkosic arenite with rare cobbles of McGrath Gneiss (Ꞓpda), pillowed basalt flows (amphibolite; Ꞓpdb), shale (staurolite-garnet-mica schist; Ꞓpdg), interbedded dolomitic arkose and dolostone (Ꞓpda), fragmental basaltic volcanic rocks (amphibolite; Ꞓpdv), and at the top of the section, pure dolostone (marble; Ꞓpdm). See Figure 10.5 for the stratigraphic section of the Denham Formation.

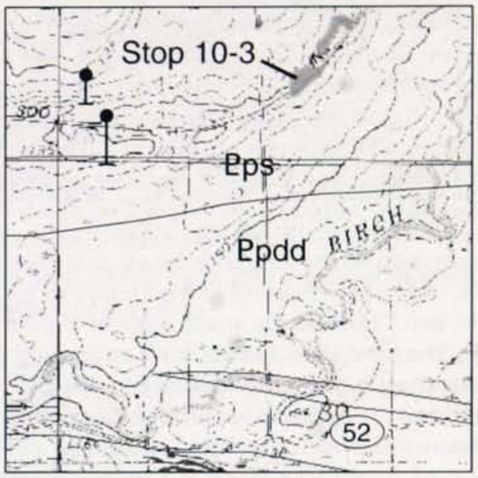
NEXT: Drive east approximately 0.75 mile on County Road 52 to a "T" intersection with a gravel road to the north. Go north on the gravel road 0.65 mile to a junction with the Soo Line Trail—a former railroad grade now utilized as an ATV trail. Park at the trail junction and walk northeast on the trail about 500 feet to outcrops and low road cuts.

STOP 10-3

Metagraywacke and argillite

Location: T. 45 N., R. 20 W., sec. 19, SE

Denham quadrangle; UTM: 506,343E/5,134,441N



Highlights: Graded beds, metamorphic assemblage

Description: The railroad cuts and flat outcrops consist of metagraywacke with pelitic beds. Pelitic units are micaceous and contain small garnet and staurolite crystals, although the latter are more abundant further to the northeast.

Abundant 1-millimeter-diameter garnets are synkinematic with respect to S_1, and have a well-developed internal schistosity defined by quartz and ilmenite inclusions. At this locality, inclusion-rich cores are surrounded by haloes of inclusion-free garnet, and staurolite has overgrown both the schistosity and the crenulation cleavage. Garnet rim analyses give final equilibration temperatures of 520 to 590° C and a pressure of ca. 6 kbar. Chemically homogeneous monazite grains at this locality give a prominent metamorphic age domain of ca. 1,800 Ma, and a less prominent age domain of 1,840 Ma (McKenzie, 2004).

In the series of outcrops to the northeast of this stop, bedding and S_1 cleavage are gently folded along a series of shallow, east-plunging, northeast-striking, F_2 fold axes similar in orientation to those in the Denham Formation. A series of 10- to 30-centimeter-thick graded beds and scoured crossbedding give a sense of younging to the north.

Drill cores from approximately 1 mile southeast of this stop show that this graywacke conformably overlies the marble to the south. The transition between the graywacke and marble is marked by a thin (1 meter) interval of carbonaceous argillite.

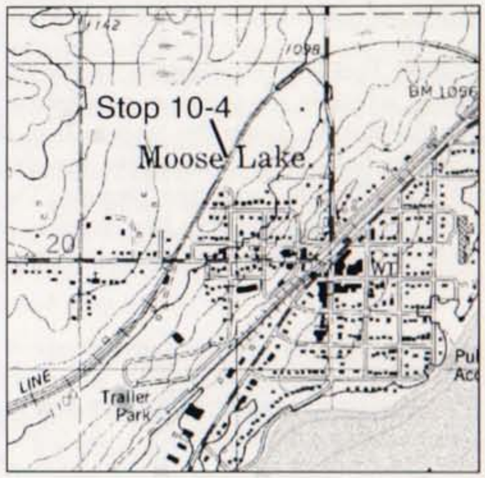
NEXT: Take County Road 52 back east to County Road 61. Turn left (north) on County Road 61 and drive approximately 7.5 miles to the junction with State Highway 27 at a stop light in Moose Lake. Turn left (west) on Highway 27 and follow it a short distance to the Soo Line trail; just before the trail, turn left and park in the parking lot. Walk north along the Soo Line trail for approximately 0.2 mile to outcrops along the trail.

STOP 10-4

Schist at Moose Lake

Location: T. 46 N., R. 19 W., sec. 20, NE

Moose Lake quadrangle; UTM Start: 517,815E/5,144,635N; UTM End: 518,205E/5,145,203N



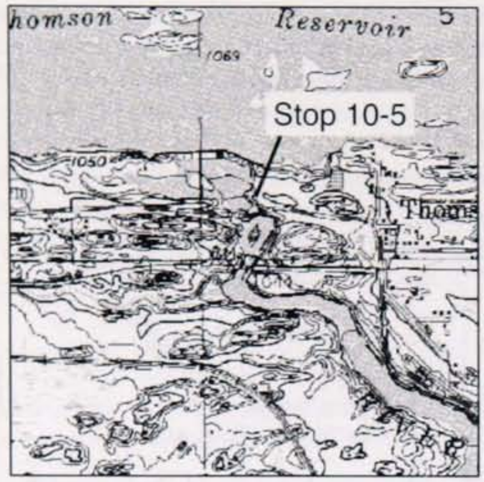
Highlights: Garnet-grade metagraywacke and pelitic schist; folded bedding and S_1 foliation

Description: This pelitic schist is part of the same belt of rocks as the last stop, but here it is north of the staurolite isograd (McSwiggen, 1987). A well-developed, bedding-parallel, moderately south-dipping S_1 foliation is folded by steeply to moderately south-dipping F_2 folds, and an axial-planar crenulation cleavage has developed parallel to the F_2 fold axes.

Internal inclusion trails in small (0.5-millimeter-diameter) garnets show as much as 160° of rotation. These garnets give final equilibration temperatures of 440 to 500° C. The age of metamorphism here is ca. 1,830 Ma based on monazite ages obtained from a similar grade rock south of this locality.

An east–northeast-trending diabase dike of presumed Mesoproterozoic age may be visible toward the north end of this set of outcrops, but the exposure has become quite overgrown.

NEXT: Drive back east to County Road 61. Turn left (north) on 61 and follow the signs to take Highway 27 east to Interstate 35 north. Go approximately 19 miles northeast on I-35 to the Carlton/Highway 210 exit (no. 235). Go east on State Highway 210 through the town of Carlton. Cross the St. Louis River approximately 1 mile east of Carlton. Turn left into the parking area just east of the St. Louis River.



STOP 10-5

Slate and graywacke of the Thomson Formation; Keweenawan diabase dikes

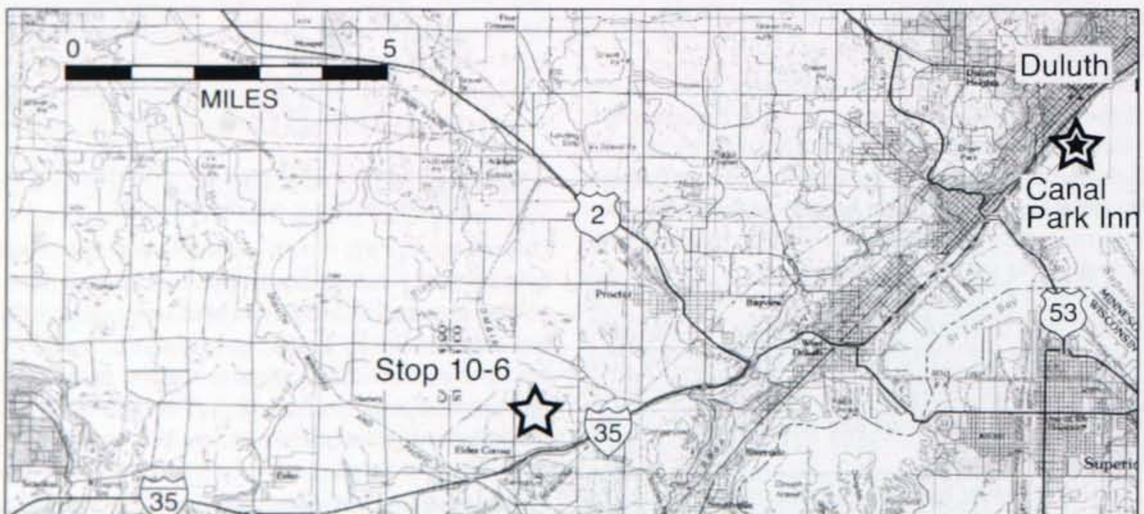
Location: T. 48 N., R. 16 W., sec. 5, SW

Cloquet quadrangle; UTM Start: 546,612E/ 5,168,119N

Highlights: Variety of sedimentary structures, folds, and diabase dikes

Description: This is the type locality for the Thomson Formation, which is the southern equivalent of the Virginia Formation, along the Mesabi Iron Range. The Virginia Formation overlies the Biwabik Iron Formation, and collectively the Virginia/Thomson Formations cover a broad area (the Animikie basin) that extends south from the Mesabi Iron Range to a few miles south of this locality (Fig. 10.1A). Deformation of the Animikie strata increases to the south, in closer proximity to the Penokean fold-and-thrust belt. Here at the south edge of the Animikie basin, the strata are folded into a series of gently east- and west-plunging, open symmetric to locally asymmetric, slightly overturned folds with near-vertical axial-planar cleavage.

Several northeast-trending diabase dikes of Mesoproterozoic age cut the Thomson Formation. These dikes are part of the Carlton County dike swarm, which is one of a number of dike swarms



Road map showing stop locations in the Carlton–Duluth area; northern leg of day 1.

that flank the Midcontinent rift system. The dikes here were emplaced along northeast-trending joints in the graywacke, and they exhibit well-developed, subhorizontal, columnar cooling joints and have chilled margins. The approximately 2-meter-wide dike just below the parking area is reversely polarized, and like most of the dikes in the Carlton County swarm, has a composition of high Fe-Ti continental tholeiite (Green and others, 1987). Overall, dikes of the Carlton swarm range from a few centimeters to greater than 60 meters in width. Narrow contact metamorphic albite-epidote hornfels is found adjacent to the contacts of only the thicker dikes. Green and others (1987) summarized the geologic, geochemical, and paleomagnetic data on the dikes and demonstrated that these swarms are most likely Mesoproterozoic in age.

See Stop 2-19 of Field Trip 2 for a more thorough description of this stop.

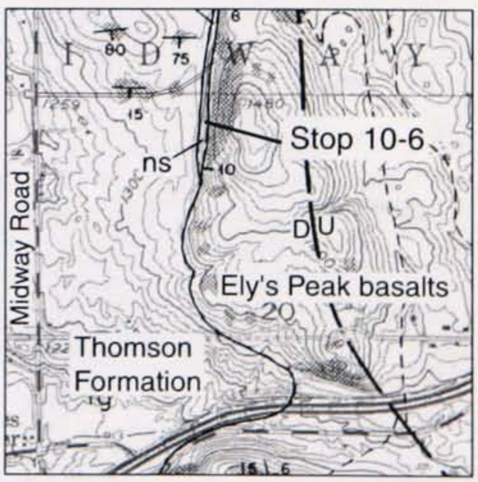
NEXT: Return to Carlton; in Carlton turn north on State Highway 45 to I-35. Go east (north toward Duluth) on I-35 approximately 7 miles to Midway Road (St. Louis County Road 13). Turn north on Midway Road and continue 0.8 mile to a road to the east. Turn right on this road and travel approximately 0.2 mile, park and walk east to the base of the bluff.

STOP 10-6

Nopeming Sandstone and basal Keweenawan basalt flows (unconformity between Mesoproterozoic and Paleoproterozoic rocks)

Location: T. 49 N., R. 15 W., sec. 17, SE, SW

Esko quadrangle; UTM Start: 555,580E/5,174,380N



Highlights: Paleo/Mesoproterozic unconformity, quartzite, pillowed basalt flows

Description: This area shows the unconformable relationships between the Paleoproterozoic Thomson Formation and overlying Mesoproterozoic sedimentary and volcanic rocks of the Keweenawan Supergroup. Outcrops of steeply dipping Thomson Formation rocks are present in the lower flats west of the sharp knob. At a higher elevation at the base of the bluff to the east are outcrops of essentially undeformed, gently east-dipping Keweenawan sedimentary and volcanic rocks. Although the actual Paleoproterozoic/Mesoproterozoic unconformity is not exposed, it can be inferred that this is an angular unconformity representing a time gap of about 800 million years. The geologic contacts shown on the inset map above are from Kilburg and Morey (1977).

The Keweenawan section here consists of the Nopeming formation (unit ns) overlain by the Ely's Peak basalts. The Nopeming formation is approximately 30 feet thick, and is composed mainly of interbedded quartz-rich conglomerate and quartzite, with some beds of siltstone near the top showing soft sediment deformation structures. The overlying Ely's Peak basalts—the lowest volcanic rocks in the North Shore Volcanic Group—are reversely polarized, and contain vague pillow structures that indicate volcanic eruption into a subaqueous environment.

See Stop 2-18 of Field Trip 2 for more information about this stop.

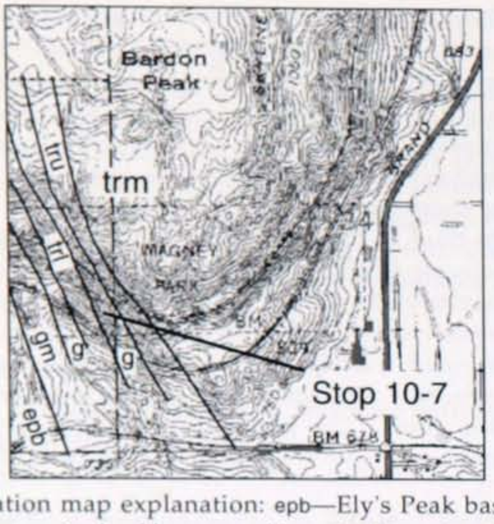
NEXT: Drive back south on County Road 13. Continue over I-35 to the junction with St. Louis County Road 3. Turn left (southeast) on County Road 3 and go approximately 0.5 mile and turn left (east) onto Skyline Parkway. Follow Skyline Parkway for 2.4 miles to a parking area and walk downhill to railroad tracks and proceed west on the tracks to outcrops.

STOP 10-7

Bardon Peak

Location: T. 49 N., R. 15 W., sec. 34, SW, and sec. 33, SE

Esko quadrangle; UTM Start: 555,580E/5,174,380N



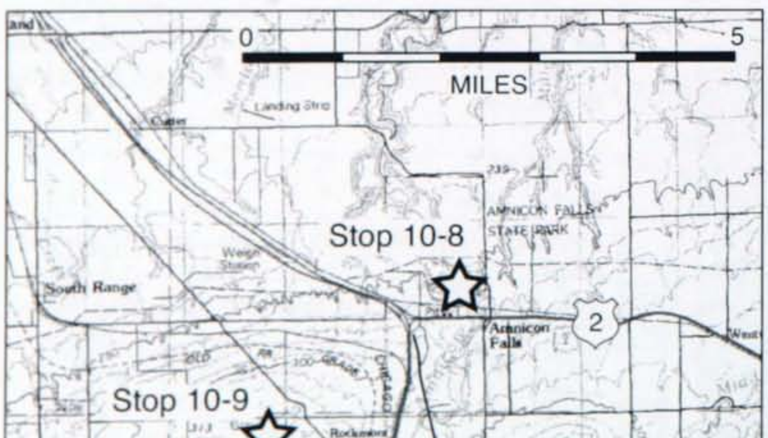
Location map explanation: epb—Ely's Peak basalts; gm—marginal gabbro; g—gabbro; trl—lower troctolite; tru—upper troctolite; trm—main troctolite.

Highlights: Basal contact and lower troctolite zone of the Duluth Complex

Description: This stop is near the base of the Duluth Complex and is in the vicinity of the transition between the basal contact zone and the troctolite zone of the Duluth Complex (Miller and others, 1995b). Roadside outcrops and nearby knolls expose layered gabbro, troctolite, and peridotite bodies. The stop is located approximately 500 meters east of the steeply-dipping (45° east) basal contact of the Duluth Complex. This contact can be observed to the west along the railroad tracks that are below the parking area for this stop. Beneath the Duluth Complex are shallow-dipping (15° east) flows of Ely's Peak basalt, observed at the previous stop. Here the basalt is recrystallized to hornfels.

The lower part of the Duluth layered series consists of interlayered gabbro and troctolite that are cross cut by peridotitic bodies (Ross, 1985; Miller and others, 1993). Gabbroic and troctolitic rocks display a variety of types and scales of layering, ranging from macrolayering (meter scale) of rock types to centimeter-scale isomodal layering. Mineral analyses reveal cryptic compositional differences in olivine between melatroctolite-dunite and gabbro. Some layers pinch out along strike and may record trough layering. The layered rocks near the basal contact of the Duluth layered series are cross-cut by small bodies of coarse-grained, biotitic oxide dunite to peridotite. Combined, these features are interpreted to record the early crystallization history of the Duluth layered series chamber, including processes of turbulent convection, magma reinjection, and volatile fluxing from the footwall (Miller and others, 1995b).

To the south is a view of the drowned estuary of the St. Louis River. During Late Wisconsin glaciation, water levels in glacial Lake Duluth rose to more than 165 meters above current lake levels when eastward drainages were dammed by retreating ice. During this time (approximately 12,000 years ago), glacial meltwater drained southward through the Brule and St. Croix Rivers (Stops 10-12 and 10-13). After retreat of the glaciers, an eastward drainage was established through the Straits of Mackinac. Uplift in



Road map showing stop locations in the Amnicon Falls area; northern leg of day 2.

the northeast, due to glacial rebound, is causing Lake Superior to tilt to the south, resulting in drowning of the mouth of the St. Louis River. Water levels in this part of the lake are estimated to be rising at a rate of 15 centimeters per century (Ojakangas and Matsch, 1982).

END OF DAY 1

OVERNIGHT IN DULUTH

DAY 2

NEXT: From Canal Park Inn in Duluth, get on Interstate 35 and head south approximately 3 miles to U.S. Highway 53 south, via exit 255B (left exit). Follow Highway 53/U.S. Highway 2 for approximately 13 miles, then split off east on Highway 2 (Highway 53 veers south at this point). Proceed east on Highway 2 for approximately 0.8 mile, turn left (north) on Douglas County Road U. The entrance to Amnicon Falls State Park is a short distance north on the left (west) side of County Road U.

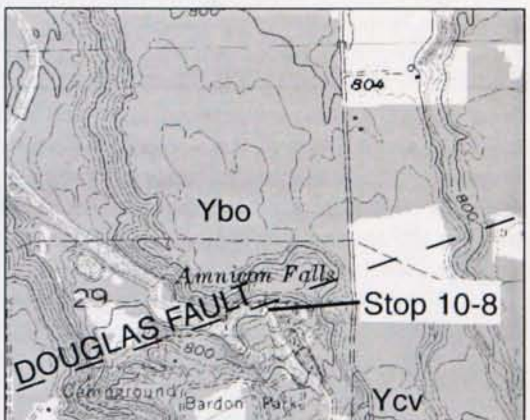
STOP 10-8

No hammering or collecting please!

Douglas fault, Chengwatana Volcanics, and Orienta Sandstone of the Bayfield Group

Location: T. 48 N., R. 12 W., sec. 29

South Range quadrangle; UTM: 584,890E/ 5,162,537N



Highlights: Douglas fault, Chengwatana Volcanics (Ycv), Orienta Sandstone (Ybo)

Description: This stop reveals one of the last major tectonic events related to the Midcontinent rift. In this area of northern Wisconsin, the central part of the rift consists of fault-bounded blocks of volcanic rocks, chiefly basalts with minor andesites and rhyolites, flanked by half-graben basins filled with younger clastic sediments (Chandler and others, 1989). The rift-bounding faults, including the Douglas fault on the northern limb of the rift, and the Lake Owen fault along the southern limb of the rift (Fig. 10.1), are steeply dipping, reverse faults with basalts on the upthrown side. The Lake Owen fault was a major growth fault on the southeast side of the graben during the extensional phase of the rift, which lasted from about 1,100 Ma to about 1,094 Ma. A compressional event dated at approximately 1,060 Ma (Cannon, 1994) reversed the sense of motion along the Lake Owen fault, resulting in the development of a central horst. The Douglas fault on the northwest side of the horst is not clearly a growth feature and may be simply a thrust formed during rift inversion (Nicholson and Cannon, 2003). Thrust displacement on the Douglas fault must be 20 kilometers or more because it juxtaposes the base of a thick volcanic sequence, the Chengwatana Volcanics, over the younger Orienta Sandstone of the Bayfield Group.

The Bayfield Group is divided into three units. The lowest member is the arkosic Orienta Sandstone, overlain by the quartzose Devils Island Sandstone, and the uppermost unit, the feldspathic Chequamegon Sandstone. These three units represent waning sedimentation during the last stages of thermal subsidence in the region of the Midcontinent rift. All three members of the Bayfield Group were deposited in fluvial or lacustrine environments. The Orienta and Chequamegon members were deposited by northeastward-flowing braided streams. The middle unit, the Devils Island Sandstone, represents deposition across sand flats that were intermittently covered by shallow ponded water. Each preceding unit may have provided material for successive units, resulting in nearly pure quartz sandstones.

The Douglas fault is exposed at three separate waterfalls within the park. The falls developed where the Amnicon River pours over massive, erosion-resistant basalt, forming plunge pools in softer sandstone. The best exposure of the fault is at the base of the Upper Falls, just west of the parking lot, where basalt of the Chengwatana Volcanics, dipping between 30° and 40° southeast, overlies the younger Orienta Sandstone. The fault plane is defined by a 3-meter-thick zone of fault gouge and cemented breccia. The Orienta Sandstone is nearly vertical near the fault, but the dip decreases away from the fault, becoming nearly horizontal within 30 meters. The Orienta Sandstone exposed here is characterized by a layer of conglomerate near the falls, and thin, flat-lying layers displaying cross-beds that are best seen from the covered bridge crossing the river. The Douglas fault is also exposed in the Now and Then Falls to the east of the parking area, and at Snake Pit Falls on the western arm of the Amnicon River. Separate basalt flows within the Chengwatana Volcanics can be distinguished by the presence of amygdaloidal minerals in flow tops with massive interiors along the southern shore of the island formed by the two branches of the Amnicon River.

The Chengwatana Volcanics in this area are dominated by localized high-titanium basalts with accompanying intermediate and felsic volcanic rocks. This association has been postulated elsewhere in the rift to be related to central volcanic complexes, sites of prolonged shallow magma chamber development and accompanying volcanism. The Amnicon gabbroic and granophyric intrusive complex cuts the base of the Chengwatana Volcanics and most likely represents the magma chamber for a central volcanic complex. The extensive granophyre in the Amnicon pluton is identical chemically to an overlying rhyolite flow. This rhyolite flow has a preliminary date of about 1,102 Ma (Nicholson and others, 2004).

NEXT: From Amnicon Falls State Park, return to Highway 2 and go west. Just past the junction with Highway 53 turn south on Douglas County Road C, and then west on Douglas County Road E for about 1.5 miles. Turn west onto Bayfield Road and proceed for 0.2 mile. Pull into an abandoned quarry entrance, park, and walk south along the dirt road to quarry.

STOP 10-9

Private property! Permission must be obtained before entering!

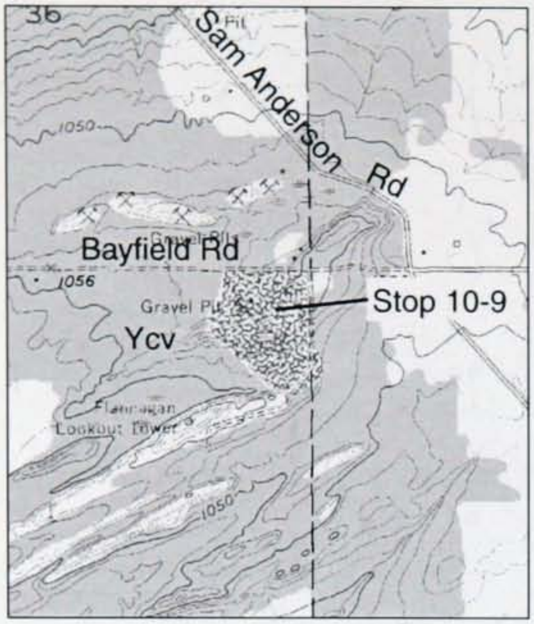
South Range Quarry

Location: T. 47 N., R. 13 W., sec. 21, NE

South Range quadrangle; UTM Start: 581,741E/ 5,159,927N

Highlights: Amygdule minerals in Chengwatana Volcanics (Ycv), basalt flow features and alteration

Description: A quarry cut into the south face of a basalt ridge in the Chengwatana Volcanics exposes a northeast-striking, south-dipping sequence of basalt flows, with massive interiors and amygdaloidal and brecciated flow tops. The flows have sparse red plagioclase crystals in a fine-grained matrix of augite and plagioclase, commonly altered to



secondary chlorite and epidote. Epidotization of brecciated flow tops imparts a distinctly greenish color in contrast to black, less altered flow interiors. Flow tops have a complex sequence of amygdule mineralization that includes the presence of copper sulfides, dominantly chalcopyrite, with minor bornite, and pyrrhotite.

Copper mineralization in the Midcontinent rift is dominated by native copper in basalts and interflow sediments in the Portage Lake Volcanics in Michigan, or by copper sulfides in reduced facies of the Nonesuch Formation at the White Pine deposit, Michigan. Copper sulfides in basalt of the Midcontinent rift are rare; the only other notable occurrence is minor chalcocite mineralization restricted to a small area of the Keweenaw Peninsula, Michigan. The presence of copper sulfides in amygdules in this quarry as well as several other quarries in the area may have important implications for regional metallogeny of the rift. The style of mineralization in these basalt quarries is analogous to native copper mineralization in Michigan in that it is associated with alteration of permeable regions in the flow tops, but it is mineralogically distinct. The occurrence of chalcopyrite rather than native copper or chalcocite requires high sulfur activity in the mineralizing fluid; an anomalous feature for the typically sulfur-poor, subaerial erupted basalts of the Midcontinent rift. Sulfur isotope values for chalcopyrite are near zero, suggesting an igneous source for sulfur.

In this quarry, as well in other basalt quarries in the region, copper sulfides occur with other minerals in amygdules, which are sparse in flow interiors and more abundant in flow tops. Other amygdaloidal minerals include calcite, chlorite, epidote, hematite, microcline, prehnite, and quartz. Several generations of quartz can be found including white or clear quartz, blue quartz, and tan to pink agate (Cordua, 1989). Chalcopyrite appears to be most abundant in the lower parts of altered flow tops, where epidote alteration is less severe, rather than in the more thoroughly altered rock of the upper flow tops. This relationship can be seen especially well near the east face of the quarry, where a mass of strongly epidotized rock contains little or no sulfide minerals, but is surrounded by a halo of mineralization about one meter wide.

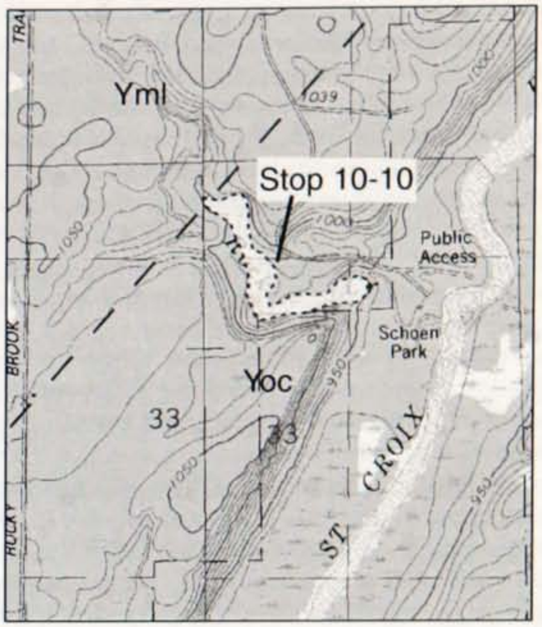
NEXT: Return to Highway 53 and head south approximately 29 miles, past Solon Springs, to the junction with Douglas County Road T. Go west on County Road T for about 16 miles. Two miles after crossing the St. Croix River, turn south on Rocky Brook Trail. Follow the road south and west for 4 miles to the junction with Schoen Road, which leads to Schoen Park and a public access site on the St. Croix River. Proceed south on Schoen Road for 0.7 mile (0.3 mile from the end at the river) to a small pull-off on the south side of the road. Walk south about 50 meters to the north bank of Rock Creek, and follow the creek bed toward the St. Croix River.

STOP 10-10

Copper Harbor Formation

Location: T. 43 N., R. 14 W., sec. 33, NE

Scovils Lake quadrangle; UTM Start: 567,742E/ 5,113,243N



Highlights: Conglomerate (Yoc), sandstone, siltstone, Minong Volcanics (Yml)

Description: Sandstone, siltstone, and conglomerate of the Copper Harbor Formation are exposed extensively along Rock Creek upstream (west) from the Schoen Park public access site on the St. Croix River. This site is on the north limb of the Ashland Syncline, and the exposures are near the base of the Copper Harbor Formation. The Copper Harbor Formation is the oldest member of the Oronto Group, which also includes the Nonesuch and the Freda Formations. Rocks of the Oronto Group are a volcanic-clastic sequence deposited during the transition from an earlier extensional rift basin to a successor thermal

Road map showing the location of Stop 10-10; central leg of day 2.

subsidence basin that was generally centered on the older rift basin but was much broader. In other areas of the rift, the waning volcanic activity in the rift is marked by thin basalt flows intercalated into the lower part of the Copper Harbor Formation. Rocks of the Copper Harbor Formation are typically volcanic clast-supported conglomerate facies that fine upward and basinward into sandstones and siltstones (Daniels, 1982). The depositional environment for conglomerates in the Copper Harbor Formation is interpreted to be prograding alluvial fan complexes that developed along the margins of the rift basin, with sandstone and siltstone facies representing floodplain deposits laid down well away from the basin margin.

At this stop, the westernmost outcrops of the Copper Harbor Formation are coarse-grained conglomerate with basalt and felsic clasts in a matrix cemented with silica and calcite. Basalt clasts can have amygdules filled with chalcedony. Bedding dips gently (~15°) to the southeast. Following the outcrop downstream moves up in the stratigraphic section. Going up-section, the percent of conglomerate diminishes in favor of sandstone and silt—the typical progression of fining upward in the Copper Harbor Formation. The easternmost outcrops exposed in a high cut bank of Rock Creek are the highest in the stratigraphic section, with interbeds of rare conglomerate and finer-grained sandstone and siltstone in which bedding units are typically about a meter thick. The finer-grained sediments split into thin bedding slabs. Most of the Copper Harbor Formation in the Ashland syncline is composed of finer-grained clastic rocks.

NEXT: Return up Schoen Road to Rocky Brook Trail. Turn left (west) on Rocky Brook Trail and follow it west and south around a series of curves to State Highway 35 (see map on page 201). Turn left (south on Highway 35) and stay on Highway 35 for approximately 60 miles until it intersects U.S. Highway 8. Turn right (west) on Highway 8 and go approximately 4 miles and turn south back onto Highway 35, to the entrance of Wisconsin Interstate State Park on the right. Drive into the park, check in at the park office, and obtain a park map. Access to Stop 10-11 (Eagle Peak) is from a hiking trail leading west from the Pines group camp, on the Eagle Peak trail.

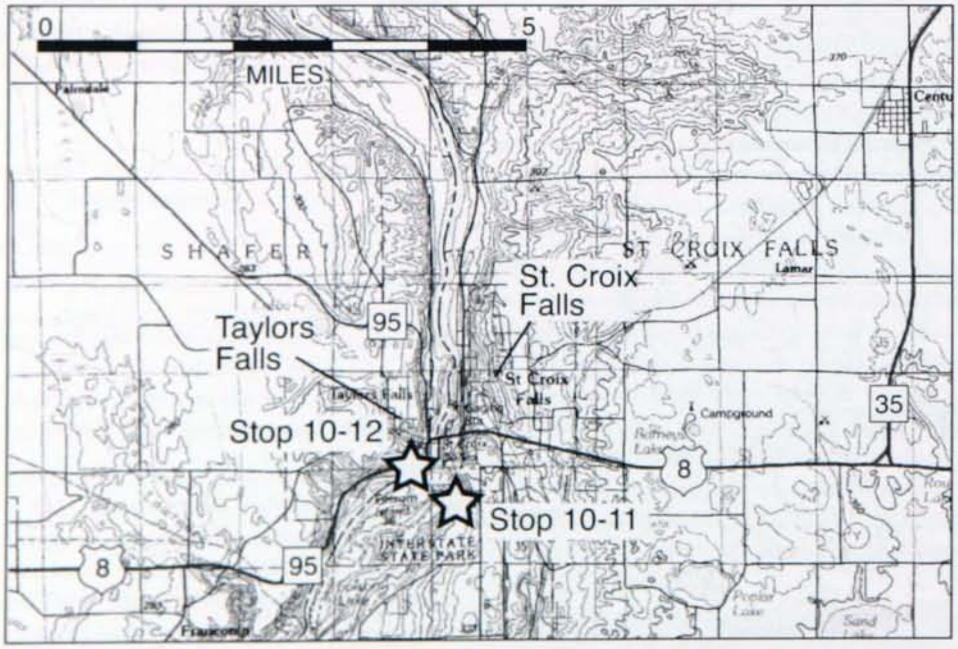
STOP 10-11

No hammering or collecting please!

Eagle's Peak Basalts

Location: T. 34 N., R. 19 W., sec. 36, SE

St. Croix Dalles, WI-MN quadrangle; UTM Start: 526,622E/5,026,004N



Road map showing the location of the Interstate Park area; southern leg of day 2.

Highlights: Mesoproterozoic (Keweenawan) plagioclase porphyritic lava flow (Chengwatana Volcanics in the Clam Falls area), scenic overview of Cambrian paleotopography, and Quaternary glacial features

Description: At this locality, a plagioclase porphyritic volcanic flow can be observed. This flow is near the southern limit of the St. Croix horst and is among the southernmost surface exposures of Keweenawan volcanic rocks. An excellent view of the "stair-step" topography that is characteristic of the Midcontinent rift can be observed from the summit of Eagle Peak. The west slope of Eagle Peak is a dip slope that dips gently (~15°) to the west and is sub-parallel with the top of a plagioclase porphyritic basalt flow. The steeper east side exposes a cross-section of the flow.

Also visible from Eagle Peak is the distribution of Cambrian sedimentary rocks. The scenic dalles of the St. Croix River record final downcutting of the river into basalt after excavation of the softer overlying sediments. Erosional remnants of Cambrian sediment are exposed along many of the steep basaltic hillslopes in the area (for example Stop 10-12). An erosional scarp formed of Cambrian sediment is visible to the east and north of Eagle Peak. Other features related to the Quaternary history of the dalles on the St. Croix are also visible from Eagle Peak, including a possible plunge pool or whirlpool (Lake of the Dalles), glacial striations, and chattermarks.

In the Taylors Falls region, the Eagle Peak flows (Cordua, 1989) form a distinctive marker unit that consists of at least three flows with a composite thickness up to 60 meters. Only one flow is exposed at Eagle Peak. This flow is characterized by large (less than 6 centimeters long) plagioclase phenocrysts in a fine-grained matrix consisting of epidote, chlorite, actinolite, and albite. Amygdule fillings consist of quartz, epidote, potassium feldspar, and chlorite. The pressure and temperature conditions of metamorphism, as constrained by mineral reactions and the sodium content of actinolite, are estimated to be approximately 0.25 GPa (~7.5 kilometer depth) and 350° C, consistent with estimates of the uplift of the St. Croix horst from geophysical evidence (Allen and others, 1997).

The ages of volcanic flows in the Taylors Falls region are relatively poorly constrained. Zircons from a thick flow exposed in a nearby quarry (approximately 4 kilometers east of Eagle Peak) yielded a U-Pb age of 1,099 ± 2 Ma (M. Schmitz, unpub. data). This age is consistent with the largely normal magnetic polarities of the rocks in this region (Kean and others, 1997). In comparison, zircons from two rhyolite flows near Clam Falls yielded ages of 1,102 ± 5 Ma (Wirth and Gehrels, 1998). These relations suggest that the Chengwatana Volcanics (Cannon and others, 2001) in the Taylors Falls–Clam Falls region likely correlate with the upper Kallander Creek Volcanics or lower Portage Lake Volcanics of northwestern Wisconsin (Fig. 10.8; Nicholson and others, 1997).

Volcanic flows of the Taylors Falls region are almost without exception high-alumina and high-iron tholeiitic basalts. Rare rhyolite flows are exposed near Clam Falls (approximately 40 kilometers north of this stop), but are not observed in the Taylors Falls region. Most flows in the southern St. Croix horst are moderately evolved (Mg# = 0.37 to 0.58) and are variably enriched in the incompatible trace elements (for example LREE, thorium). All of the flows are characterized by pronounced depletions of tantalum and niobium, relative to thorium and lanthanum. Initial epsilon neodymium values of the basalts and rhyolites typically range from -1.5 to -3.0. Together, the geochemical data suggest that the flows are the result of variable crystal fractionation and contamination of plume-derived melts (Wirth and others, 1997).

NEXT: Return to the parking area at the group campground. Continue southeast along Silverbrook Trail for approximately 400 meters. Enter the low ravine on the south side of the trail.

STOP 10-12

No hammering or collecting please!

Paleozoic conglomerate

Location: T. 34 N., R. 19 W., sec. 36, SE

St. Croix Dalles, WI-MN quadrangle; UTM Start: 527,029E/5,025,776N

Highlights: Unconformity between Upper Cambrian conglomerate and sandstone with Mesoproterozoic volcanic rocks

Description: After crossing a low marshy area, Skyline Trail enters onto flows of the Trap Rock Alley unit (Cordua, 1989). The floor of the ravine marks the approximate contact between Mesoproterozoic volcanic rock and Upper Cambrian conglomerate and sandstone. The conglomerates and sandstones exposed along the south wall of the ravine were deposited against steeply sloping surfaces of the volcanics and are tentatively assigned to the Mazomanie Formation of the Tunnel City Group (Mazomanie Member of the Franconia Formation in Minnesota (Mossler, 1987). The basalt boulder conglomerate (Mill Street conglomerate of Berkey, 1897) grades vertically and horizontally into pebble conglomerate and sandstone. The basalt clasts are weathered and are generally characterized by low sphericity. Interclast sandstone is poorly bedded. The Mill Street conglomerate contains an unusual assemblage of trilobites, inarticulate brachiopods, and monoplacophoran molluscs representing an intertidal, rocky shoreline environment (Berkey, 1898; Webers, 1972; Cavaleri and others, 1987).

This exposure is one of a few in North America where conglomerate marks the base of the Upper Cambrian era. Other exposures occur in the Baraboo region, where Upper Cambrian conglomerate overlies Proterozoic quartzite. The conglomerates mark a second Late Cambrian transgression of shallow epicontinental seas. At that time, most of the midcontinent region was characterized by low relief. However, on the basis of water-well data, the Upper Cambrian–Middle Proterozoic unconformity in the Taylors Falls region shows up to 30 meters relief. It is envisioned that the coarse-grained boulder conglomerates record deposition in high-energy shoreline environments during the Upper Cambrian. The interclast quartz sands likely originated from distant sources.

NEXT: Return to Highway 8. Turn left (west) and cross the St. Croix River into Minnesota. Immediately after crossing the bridge, turn left at the traffic light into the Minnesota Interstate State Park.

STOP 10-13

No hammering or collecting please!

Potholes in Interstate Park

Location: T. 34 N., R. 18 W., sec. 30, SW

St. Croix Dalles, WI-MN quadrangle; UTM Start: 527,323E/5,027,344N

Highlights: Potholes formed in the Clam Falls Volcanics

Description: This stop provides an opportunity to examine spectacular potholes formed in flows of the Dresser and Trap Rock Alley units (Clam Falls Volcanics). The base of the Trap Rock Alley Flow is exposed along the west side of Trap Rock Alley; the Dresser Flows are exposed to the east. The top of the Dresser Flow contains many amygdules and is extensively epidotized. The flows dip 15° west.

More than 80 potholes are present in this area of the park (Glacial Gardens). The potholes range from decimeter depressions to giant "kettles" that are up to 20 meters deep and 6 meters in diameter. The potholes along the trail are 7.5 to 18 meters above the current river level; others have been found as much as 34 meters above river level. Today, many of the larger potholes are partly filled with silt, mud, peat, and grind stones.

The potholes at this locality formed during a period of high discharge near the end of the Wisconsin glaciation (Ojakangas and Matsch, 1982). Lake levels in glacial Lake Duluth overflowed to the south through the Brule and St. Croix Rivers when ice dammed the Straits of Mackinac. The "Dalles of the St. Croix" mark a nickpoint where floodwaters flowed from basalt onto less resistant Cambrian sandstone and shale. Fast-moving currents and a steep gradient in this region likely contributed to the formation of the many potholes.

NEXT: To return to the Twin Cities follow Highway 8 west to Interstate 35, a distance of approximately 20 miles. Go south on Interstate 35. After the town of Forest Lake the freeway splits into I-35E and I-35W. Take I-35W to go to Minneapolis and the hotel, or I-35E to go to St. Paul.

END OF TRIP

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FIELD TRIP 11

Saturday, May 21 – Sunday, May 22

GEOLOGY AND SEDIMENTOLOGY OF THE PALEOPROTEROZOIC ANIMIKIE GROUP: THE POKEGAMA FORMATION, THE BIWABIK IRON FORMATION, AND VIRGINIA FORMATION OF THE EASTERN MESABI IRON RANGE, AND THE THOMSON FORMATION NEAR DULUTH, NORTHEASTERN MINNESOTA

Leaders

Richard W. Ojakangas, Professor Emeritus, University of Minnesota Duluth
Mark J. Severson, Natural Resources Research Institute,
Peter K. Jongewaard, United Taconite Mining Company
John L. Arola, Ispat Inland Mining Company
Joel Thomas Evers, Retired–LTV Mining Company
Douglas G. Halverson, Northshore Mining
G.B. Morey, Chief Geologist Emeritus, Minnesota Geological Survey
T.B. Holst, University of Minnesota Duluth

INTRODUCTION

Iron-formation was described as early as 1866 by Henry Eames on what was to become the Mesabi Iron Range. Several attempts were made by individuals to find ore on the Mesabi range on their way north to the iron mines of the Vermilion range (Soudan to Ely, Minnesota); however, it was not until November 16, 1890 that the first rich iron ore on the Mesabi range was discovered by the Merritt brothers near what is now Mountain Iron, Minnesota. In 1892, the first shipment from this mine was 4,245 tons of ore (White, 1954). Exploration for iron ore ensued and within the next few years, most of the productive parts of the Mesabi range were discovered.

The Mesabi Iron Range is the largest iron range in the United States and is one of the largest in the world. It is 0.25 to 3.0 miles wide and 120 miles long (Fig. 11.1). The Biwabik Iron Formation, as thick as 750 feet, in general dips gently to the southeast at an angle of about 7° to 15°. The iron-formation, called taconite, typically contains 30 to 40 percent iron and 40 to 50 percent SiO_2, plus other components (Morey, 1992). In numerous places along the length of the range, silica was leached out, thereby enriching the iron content to over 55 percent. These pockets became the high-grade natural ore mines; there were more than 500 individual mines prior to merging into larger mines as the ore between adjacent properties was removed. These were very important in making the United States an industrial giant, and were instrumental in providing raw material for World Wars I and II. As the high-grade ore was depleted, the taconite process was developed. In 1967, taconite production exceeded natural ore production. Currently, six taconite plants are in production (Fig. 11.2).

The name of Biwabik Iron Formation was chosen by Van Hise and Leith (1901, p. 356), "...because the word Biwabik is the Chippewa word for a piece or fragment of iron." The word taconite is also used in discussions pertaining to hard, unoxidized portions of the iron-formation. H.V. Winchell (1882, p. 135) originally called portions of the Biwabik Iron Formation "taconyte" because he thought the rocks correlated with lower Cambrian rocks in the Taconic Mountains in northern New England. Since that time, many geologists have used taconite in their descriptions of the iron-formation and it has thus become firmly established. Perhaps a more proper definition for taconite is an economic term for iron-formation from which iron can be profitably extracted after fine-grinding, followed by magnetic separation and pelletizing (Morey, 1993).

REGIONAL GEOLOGY

The peneplaned Archean craton in the Lake Superior region formed a platform upon which a Paleoproterozoic continental margin assemblage was deposited in Minnesota, Michigan, and Wisconsin.

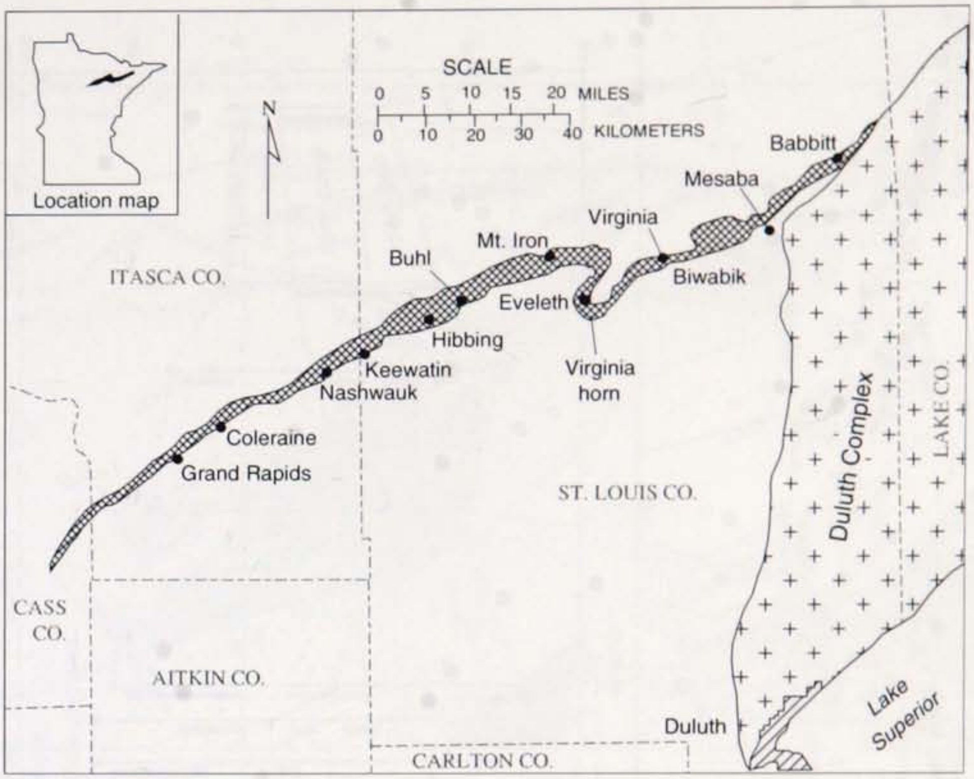


Figure 11.1. Generalized map of the Mesabi Iron Range (cross-hatched). Note the Duluth Complex on the east side.

Extension resulted in localized rifts that received thicker accumulations of sediments and volcanic rocks than did adjacent parts of the platform. Seas transgressed onto the continent one or more times and an ocean basin opened south of present-day Lake Superior. Island arcs that formed during southward subduction collided with the craton margin to the north as the ocean basin closed. A remnant of this oceanic crust is poorly preserved as a dismembered ophiolite sequence in Wisconsin (Schulz, 1987, 2003). The arc volcanics are preserved as the Wisconsin magmatic terranes. The collision resulted in a fold-and-thrust belt known as the Penokean orogen. To the north of the fold-and-thrust belt, a northward-migrating foreland basin—the Animikie basin—developed as the stacked thrusts weighed down the crust (Fig. 11.3). Thick turbidite successions were deposited along the basin axis, and terrigenous clastics and Lake Superior-type iron-formation were deposited on the shelf along the northern margin (the foreland or peripheral bulge) of the basin. See Ojakangas and others (2001) and Severson and others (2003) for more detailed summaries on Paleoproterozoic basin development in the Lake Superior region.

The development of the Midcontinent Rift System at 1.1 Ga severed the basin into northwestern and southeastern segments (Fig. 11.3). If the Midcontinent Rift System rocks are removed from the geologic map, the different portions of the Animikie basin become contiguous and the fold-and-thrust belt rocks of Minnesota, Wisconsin, and Michigan become continuous (Fig. 11.4).

Figure 11.5 is an interpretive cross-section of the Animikie basin during its formative stages, with sediments derived from the Archean basement to the north and from the fold-and-thrust belt to the south.

The Paleoproterozoic supracrustal rocks in the northwestern segment, including east-central and northeastern Minnesota and the adjoining part of Ontario, are for the most part poorly exposed. However, mining of iron ore on the Mesabi and Cuyuna ranges and continued mining of taconite on the Mesabi range have resulted in excellent artificial exposures and an abundance of drill hole information. Geophysical surveys and stratigraphic test drilling by the Minnesota Geological Survey also have been major sources of information (for example Southwick and others, 1988).

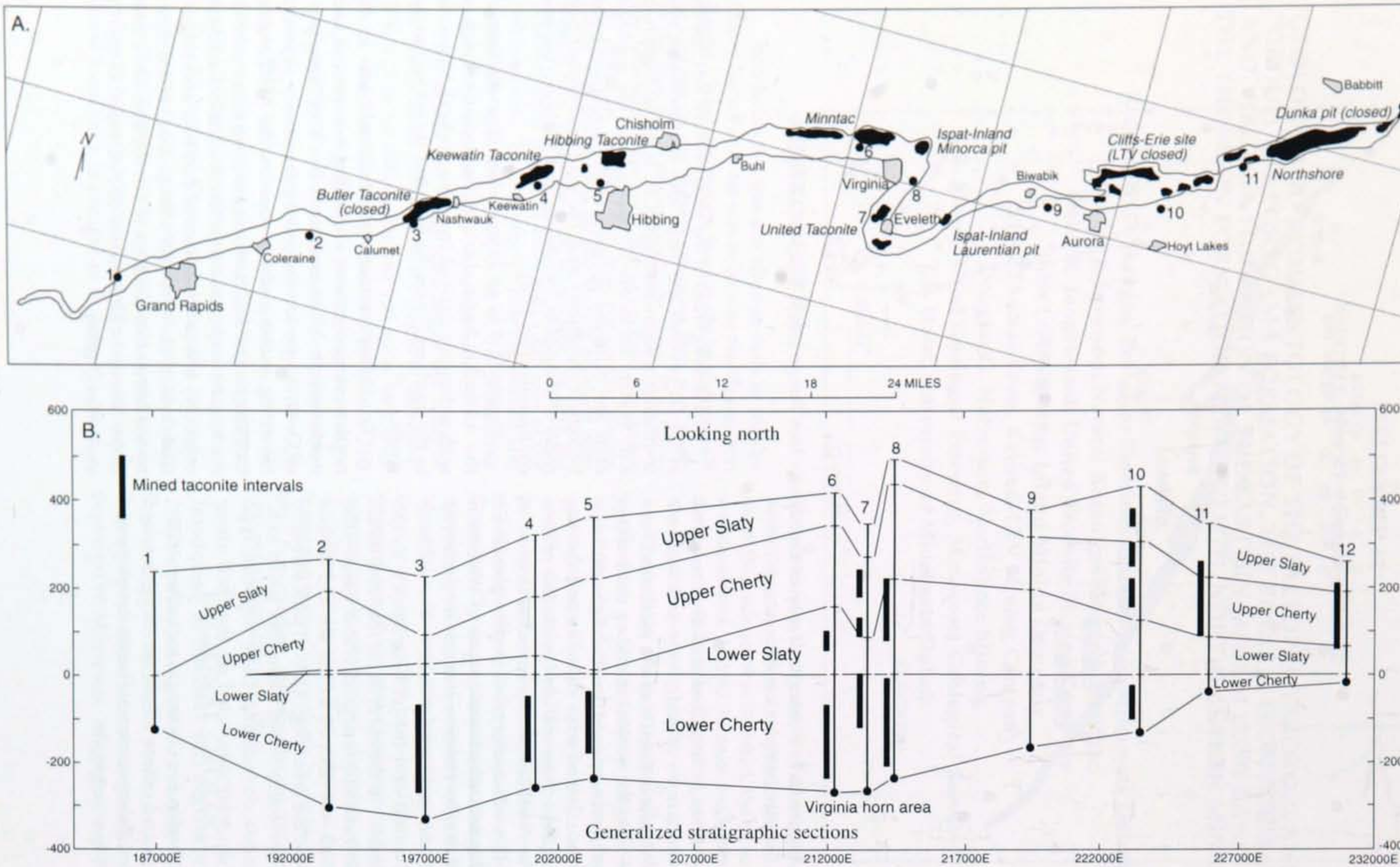


Figure 11.2. Generalized map of the Mesabi Iron Range.

A. Aerial distribution of taconite pits (black) and cities.

B. A longitudinal section of the Biwabik Iron Formation showing mined taconite intervals as black columns adjacent to sections (compiled by H. Djerlev, 1993; modified from Morey, 2003).

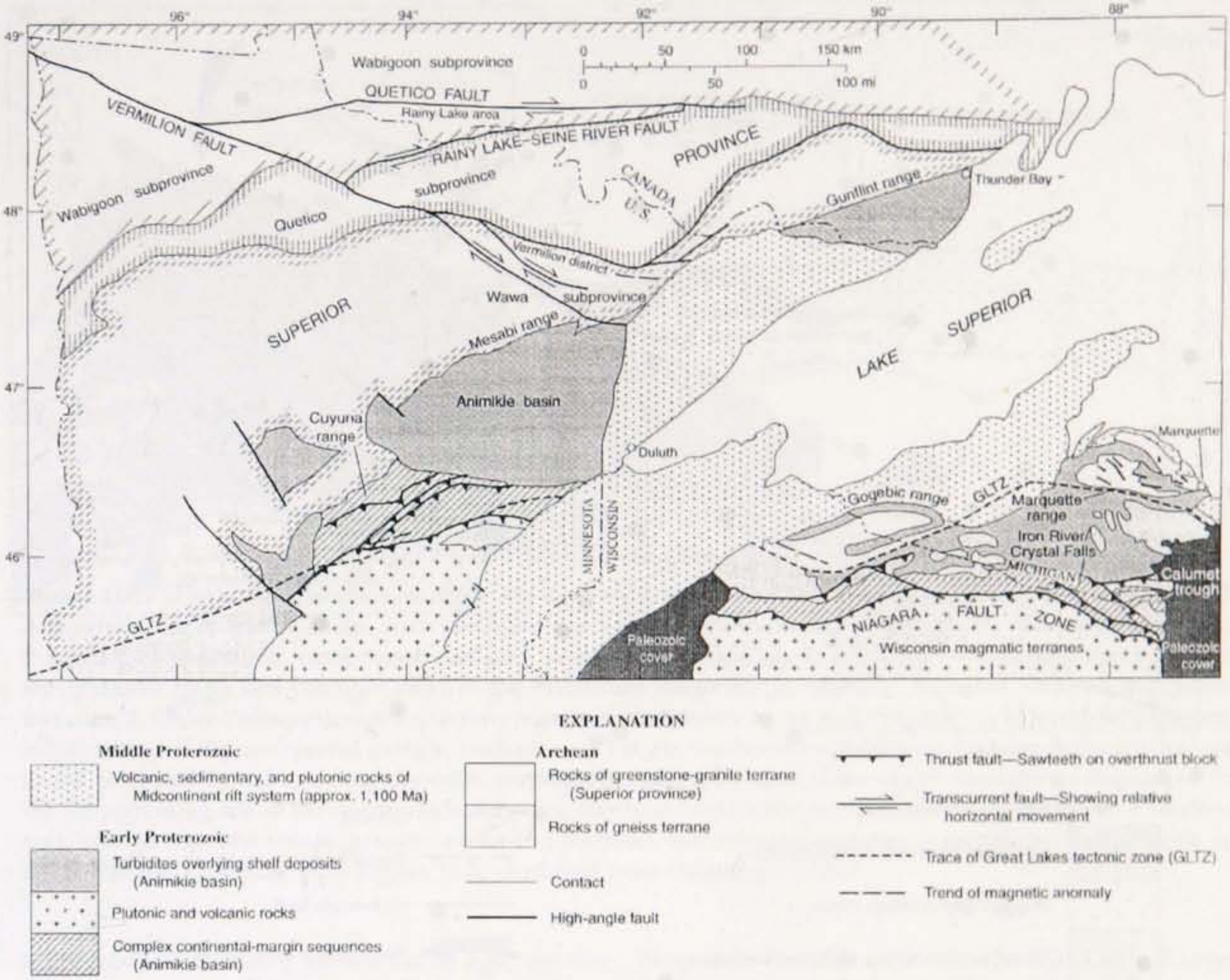


Figure 11.3. Generalized geologic map showing the distribution of Precambrian rocks and structural elements of the Lake Superior region; modified from Ojakangas (1994).

Animikie Group

The Animikie Group unconformably overlies the Mille Lacs and North Range Groups to the south and the Archean basement to the north (see Fig. 11.6; Southwick and Morey, 1991). Magnetic data show North Range structures are present beneath Animikie strata to the east of the exposed North Range Group (Chandler, 1993).

The group consists of three conformable major formations on both the Mesabi and Gunflint ranges. The respective units on the two ranges are the Pokegama Formation and the Kakabeka Quartzite (the lowest units), the Biwabik and Gunflint Iron Formations (the middle units) and the Virginia and Rove Formations (the upper units, composed of graywacke and shale). The Thomson Formation in the northern part of east-central Minnesota is correlative with the Virginia and Rove Formations. The Biwabik and Gunflint Iron Formations are on strike with each other and were probably continuous prior to the intrusion of the Duluth Complex at about 1,100 Ma.

In the model presented here, the Animikie Group in Minnesota and Ontario on the Mesabi and Gunflint ranges and the Baraga Group of Michigan and Wisconsin on the Gogebic range were both deposited in the Animikie foreland basin. The basal units comprised of siliciclastic sediment derived from the Archean basement, and the overlying iron-formation, were deposited in a shallow sea on the northern edge (the peripheral bulge or foreland) of the northward-migrating Animikie basin (for example Ojakangas, 1994). Additional details are provided below in the section titled "Environments of deposition, Animikie Group."

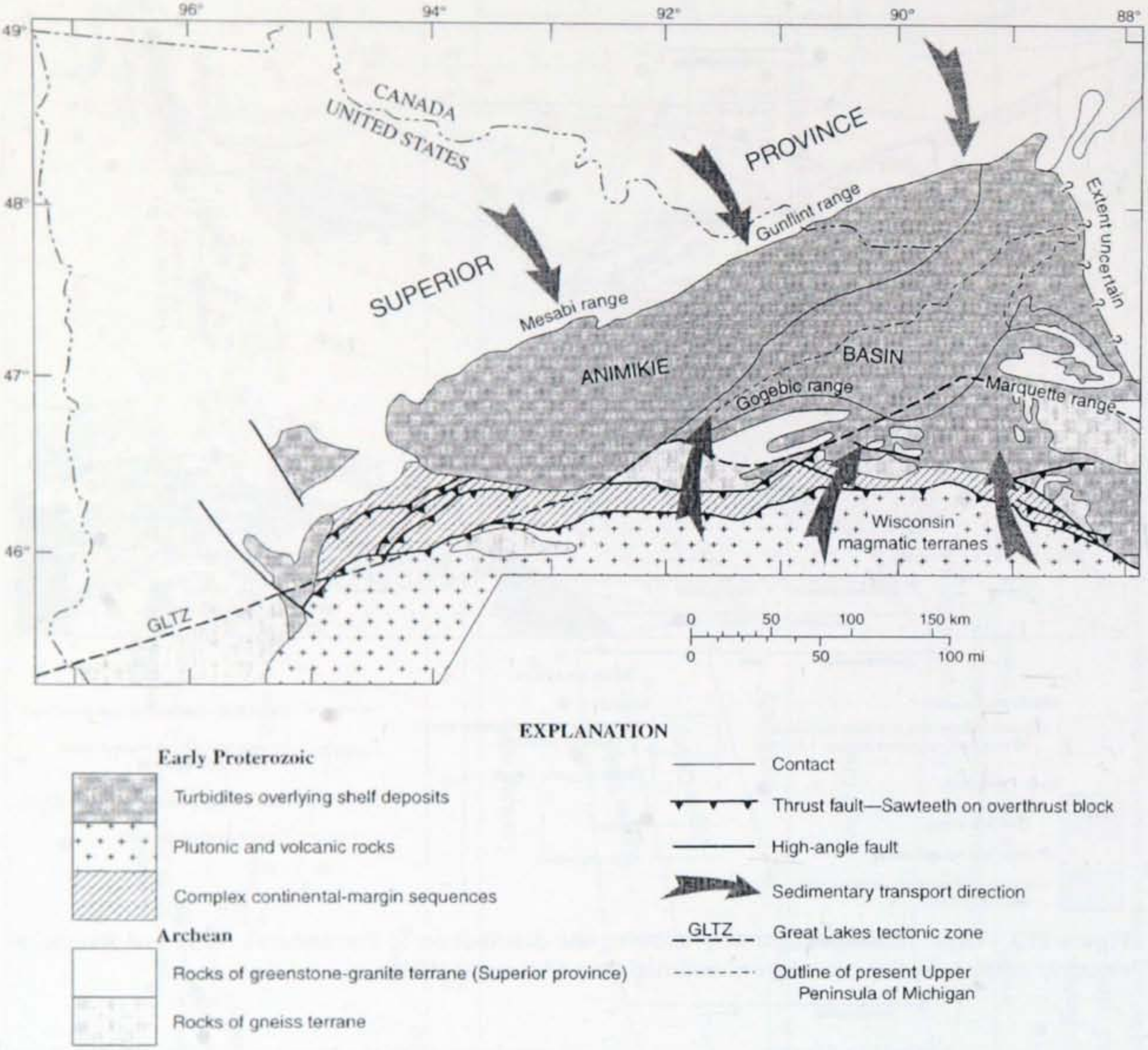


Figure 11.4. Schematic hypothesized paleogeography at the time of sedimentation of the Animikie Group turbidites that overlie shelf deposits in the Animikie basin. The rocks of the 1,100 Ma Midcontinent Rift System have been removed from the map, and Michigan and Wisconsin are thus positioned 60 miles closer to Minnesota and Ontario than they were after the formation of the Midcontinent Rift System. Arrows denote generalized transportation directions of sediment from major source areas. Compare with Figure 11.3; modified from Ojakangas (1994).

The siliciclastic and iron-formation units are exposed on the Gogebic range of northern Michigan and Wisconsin (the Palms Quartzite and Ironwood Iron Formation), on the Mesabi range of northern Minnesota (the Pokegama Formation and the Biwabik Iron Formation), and on the Gunflint range of northeast Minnesota and Ontario (the Kakabeka Quartzite and the Gunflint Iron Formation), and are lithostratigraphic equivalents. They probably were continuous from south to north prior to development of the Midcontinent Rift System in Mesoproterozoic time. A consequence of this model is that they are diachronous, with the units in Michigan and Wisconsin (located about 60 miles to the south of the Mesabi range during deposition) thus somewhat older than those in Minnesota and Ontario. The thickest and uppermost units in the basin, essentially lithostratigraphic correlatives

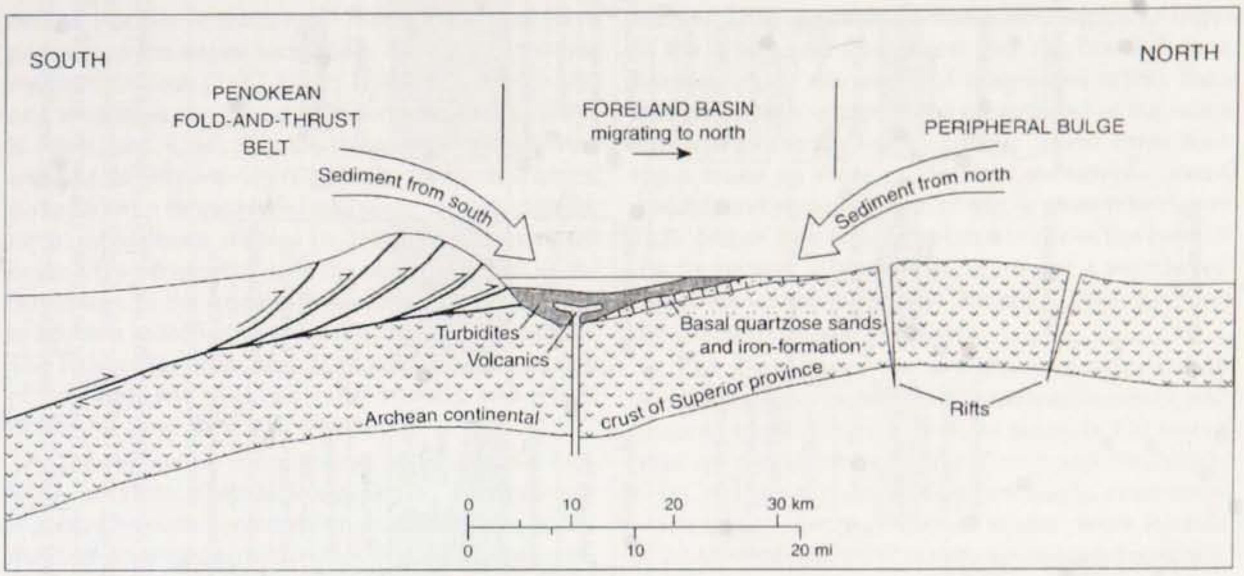


Figure 11.5. Schematic cross-section depicting deposition of the Animikie Group turbidites that overlie shelf deposits in the Animikie basin, with sediment derived from both the north and south. The southern area, the fold-and-thrust belt, comprises a complex assemblage including: 1. Accreted Paleoproterozoic volcanic and plutonic rocks and volcanic rocks of the Wisconsin magmatic terranes; 2. Accreted Archean miniplate terranes; 3. Older Paleoproterozoic passive-margin sedimentary rocks and volcanic rocks produced during initial rifting of the continental margin, both scraped off the southward-subducting Archean Superior craton; and 4. Recycled initial foredeep deposits, possibly including basal shallow-water sandstones deposited in the transgressing sea of the northward-migrating foreland basin. The peripheral bulge comprises a source-rock assemblage of Archean granitic rocks and Archean volcanic-sedimentary (greenstone) belts. Scale is approximate. Compare with Figure 11.4; modified from Ojakangas (1994).

but probably differing somewhat in age, are the Michigamme, Tyler, and Copps Formations of the southeastern segment and the Thomson, Virginia, and Rove Formations of the northwestern segment. These are typical turbidite-shale (flysch) sequences, with graded beds and intercalated muddy "rain-out" sediment (Fig. 11.6).

Ages

Along the Mesabi range, the Pokegama Formation rests unconformably on diabase dikes of the Kenora–Kabetogama dike swarm that give a Rb-Sr isochron age of 2,125 ± 45 Ma (Southwick and Day, 1983; Beck, 1988), and this provides a maximum age for deposition of the Pokegama Formation. A minimum age of 1,930 ± 25 Ma (Pb/Pb) for the Pokegama Formation was obtained by Hemming and others (1990) from quartz veins that cut the Pokegama Formation. A U/Pb age on euhedral zircons from an ash layer in the lower Gunflint Iron Formation of Ontario is 1,878 ± 2 Ma (Fralick and Kissin, 1998; Fralick and others, 2002). A similar age of 1,874 ± 9 Ma was obtained on zircon from rhyolite in the Hemlock Formation that is adjacent to (and is possibly interlayered with) the Negaunee Iron Formation in the Marquette Range Supergroup of Michigan (Schneider and others, 2002). A zircon age from an ash layer near the base of the Virginia Formation is 1,850 Ma (Hemming and others, 1996), and an age of 1,821 ± 16 Ma has been obtained from an ash layer in the Rove Formation about 70 meters above the Gunflint Iron Formation (Kissin and others, 2003). Several of these ages are shown on Figure 11.6.

Pokegama Formation

This formation has long been called the Pokegama Quartzite, but because it contains appreciable argillite and siltstone, the name Pokegama Formation is more appropriate. It has been studied by several workers since it was named by Winchell (1893) for exposures at the western end of the Mesabi Iron Range. Much of the previous work has been summarized by Morey (1972, 1973, 2003).

Few natural exposures exist, as thick glacial drift generally covers the formation. Outcrops, road cuts, and mine cuts occur at a few places along the length of the range, but most exposures are in the

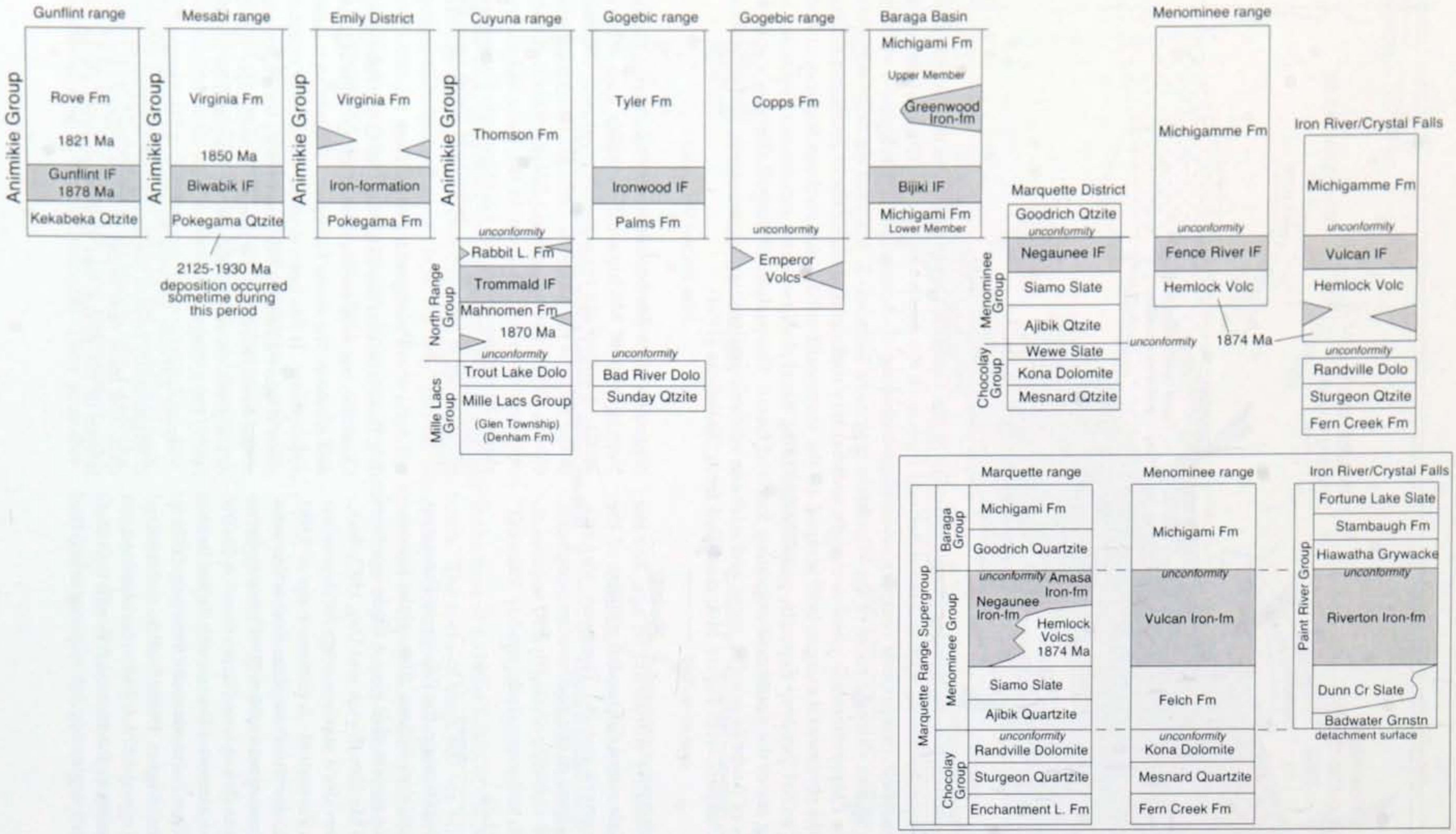


Figure 11.6. Generalized correlation chart of Paleoproterozoic strata in the Lake Superior region (after Morey and Southwick, 1995). Note that recently obtained age dates are shown for the Gunflint Iron Formation, Mahnomen Formation, Hemlock Volcanics, and the Rove and Virginia Formations. Also included is a correlation chart (lower right corner) of strata in Menominee, Iron River–Crystal Falls and surrounding terranes (LaBerge and others, 2003—includes changes to previous usage not yet officially adopted by the U.S.G.S.). Iron-formations are shaded; modified from Severson and others (2003).

central portion of the range. A few drill holes have penetrated the entire formation. One is located just south of Eveleth (T. 57 N., R. 17 W., sec. 5, NE, NE) and another is southwest of Mountain Iron (T. 58 N., R. 18 W., sec. 8, SE, SE); the thicknesses are 167 feet and 85 feet, respectively (Fig. 11.7). Other drill cores, some recently rediscovered and some recently drilled, have not yet been studied in detail. Numerous drill holes have penetrated only the upper few feet of the formation, as the drilling was generally undertaken in relation to iron ore exploration and development. The Pokegama Formation is thin at the eastern end of the range and thickens to the western end where it may be more than 300 feet thick.

The formation is composed of three main rock types—argillite, siltstone, and quartzite. The quartzite is generally silica-cemented quartz sandstone, and is therefore an orthoquartzite rather than a metaquartzite. Morey (2003) determined that mineralogical changes in the Pokegama Formation and the Biwabik Iron Formation are the result of diagenesis rather than metamorphism, except at the eastern end of the range adjacent to the Duluth Complex. These three rock types make up three gradational members—lower, middle, and upper—respectively, as shown in Figure 11.7. Minor thin conglomerates occur at the base of the formation, and seem to represent a weathered residuum on the surface of Archean rocks, perhaps reworked by fluvial processes.

The Pokegama Formation unconformably overlies Archean metavolcanic, metasedimentary, and plutonic rocks. There may be as much as 100 feet of relief on the Archean surface (Grout and Broderick, 1919), but the surface was, nevertheless, essentially a peneplain. Some Archean "knobs" were islands when the Pokegama Formation was being deposited,

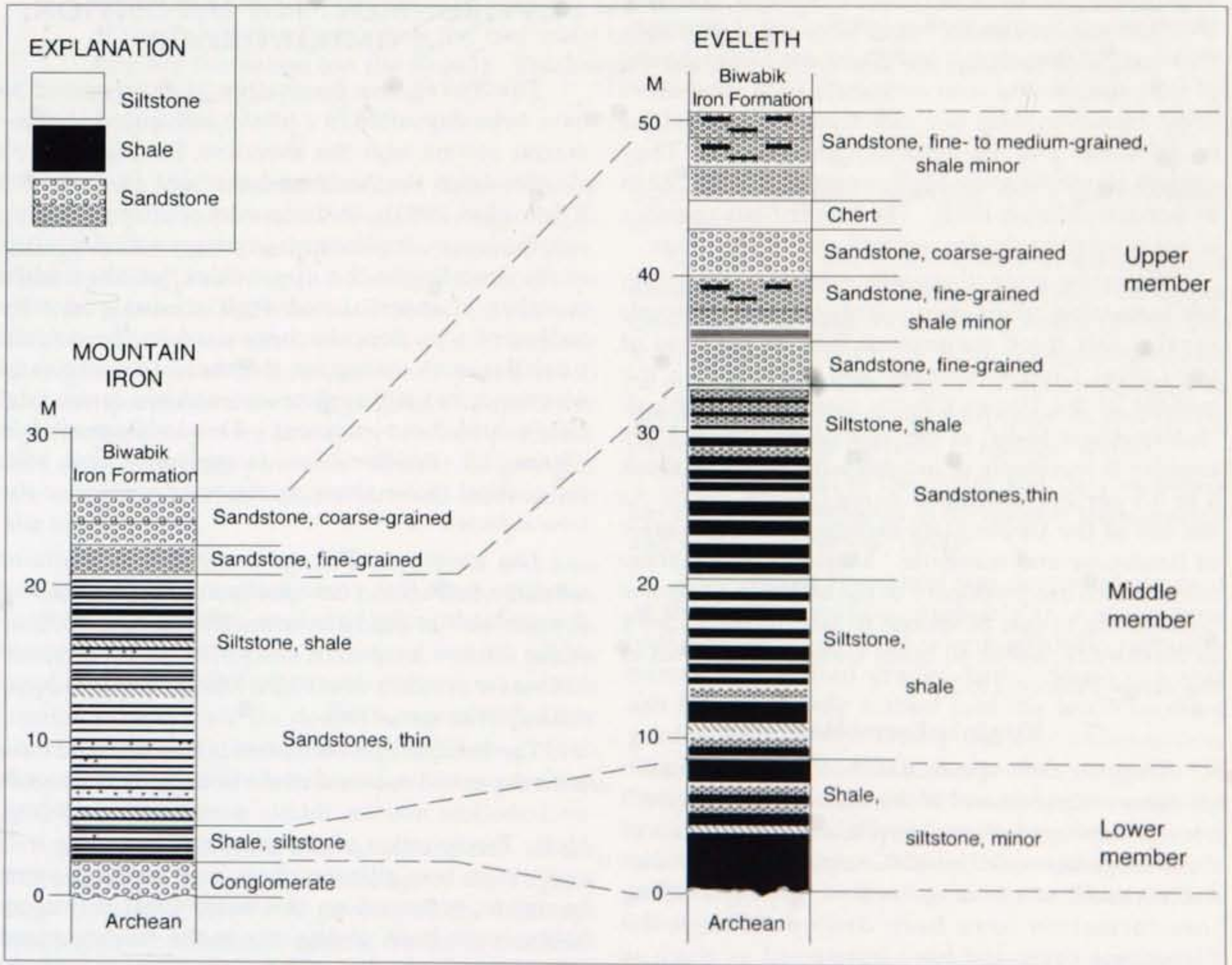


Figure 11.7. Measured sections from two drill holes that penetrate the entire Pokegama Formation. Dark shading represents shale, thin blank units represent siltstone, the slanted pattern represents sandstone and siltstone, and the dotted pattern represents sandstone. Modified from Ojakangas (1983).

and are present in the wooded areas between Eveleth and Virginia where they have been re-exhumed. The Pokegama Formation–Biwabik Iron Formation contact is gradational, with some cherty horizons in the upper Pokegama Formation and some sand grains of quartz in the lowest bed of the Biwabik Iron Formation. Various geologists have placed the contact at different stratigraphic levels.

Biwabik Iron Formation

This is one of the world's major iron-formations, and the largest in the United States. The formation is 200 to 750 feet thick and consists of four divisions as defined by Wolff (1917). These lithostratigraphic units, now informal members, are from the bottom up, the Lower Cherty, the Lower Slaty, the Upper Cherty, and the Upper Slaty (these are miners' terms, and do not indicate metamorphism; Fig. 11.2). The cherty members are dominantly granular (sand-textured), thick-bedded (several inches to a few feet), and are largely composed of chert and iron oxides. The slaty members are dominantly fine-grained (mud-textured), thin-bedded (less than 1 inch), and composed mostly of iron silicate and iron carbonate with local chert beds. However, these two rock types are interbedded on all scales and are generally gradational. They contain about the same high quantities of silica, 42 to 47 percent (Morey, 1992). The Lower Slaty member is not present at the far western end of the range.

There are some diagnostic marker units within the formation. Two stromatolite-bearing intervals several feet thick are present, one at the base of the Lower Cherty member and the other in the middle of the Upper Cherty member. The black "Intermediate Slate" at the base of the Lower Slaty member is reportedly an ash-fall tuff containing about 4 to 5.5 percent aluminum oxide (Morey, 1992). At the top of the Upper Slaty member are several feet of limestone and dolomite. Most of these marker units, which are prominent in the eastern and central parts of the range, pinch out to zero in the vicinity of Nashwauk, about 40 miles from the west end of the range (Morey, 1992).

Virginia Formation

There are rare exposures of the Virginia Formation in mines at the east end of the Mesabi range where it has been metamorphosed by the mafic intrusions of the Mesoproterozoic Duluth Complex. Several holes drilled south of the range to study the underlying iron-formation have been drilled through the Pleistocene cover and have intersected as much as 1,443 feet of the preserved lower part of the formation (Lucente and Morey, 1983).

The lower portion of the formation in the drill holes is dominantly black shale. The upper portion of the drill core, while still dominantly shale, contains beds of siltstone and fine-grained feldspathic graywacke comprising thickening- and coarsening-upward turbidite sequences. Ash-fall tuff, cherty sideritic iron-formation, chert, and limestone are minor rock types low in the formation. The contact with the underlying Biwabik Iron Formation is gradational. The clastic rocks were largely derived from the Archean rocks to the north, with some contributions from lower Proterozoic rocks to the south (Lucente and Morey, 1983).

The Virginia Formation is correlated with the Thomson Formation (Morey and Ojakangas, 1970) that is exposed 60 miles to the south in the vicinity of Carlton and Cloquet, and also with the Rove Formation in northeast Minnesota and adjacent Ontario (Morey, 1967).

ENVIRONMENTS OF DEPOSITION, ANIMIKIE GROUP

The Pokegama Formation is interpreted to have been deposited in a tidally influenced shallow marine setting near the shoreline, having received clastics from the Archean basement to the north (Ojakangas, 1983). In this model of a transgressing sea, the lower (argillaceous) member was deposited at the shoreline in the upper tidal flat, the middle member of intercalated argillaceous and silty sediment was deposited seaward in the middle tidal flat, and the upper member of quartz sand was deposited still further seaward in a lower tidal flat/subtidal environment. This is illustrated in Figure 11.8. Walther's Law is applicable here, with the vertical facies showing the relationships of the lateral facies.

The lowermost Pokegama Formation contains siltstone beds that contain alternating thicker and thinner laminae that have been interpreted as evidence of the diurnal inequality, and are being investigated further for possible clues to the Paleoproterozoic lunar orbit (Ojakangas, 1996).

The Biwabik Iron Formation is interpreted to have been deposited seaward of the Pokegama Formation on a shallow marine, tidally dominated shelf (Fig. 11.8). Precipitation of iron minerals including iron carbonate, iron silicate, chert, and perhaps some hematite, occurred on the outer shelf in waters below wave base, giving rise to the mud-textured (slaty) iron-formation. These minerals were likely related to upwelling waters from the deeper part of the basin.

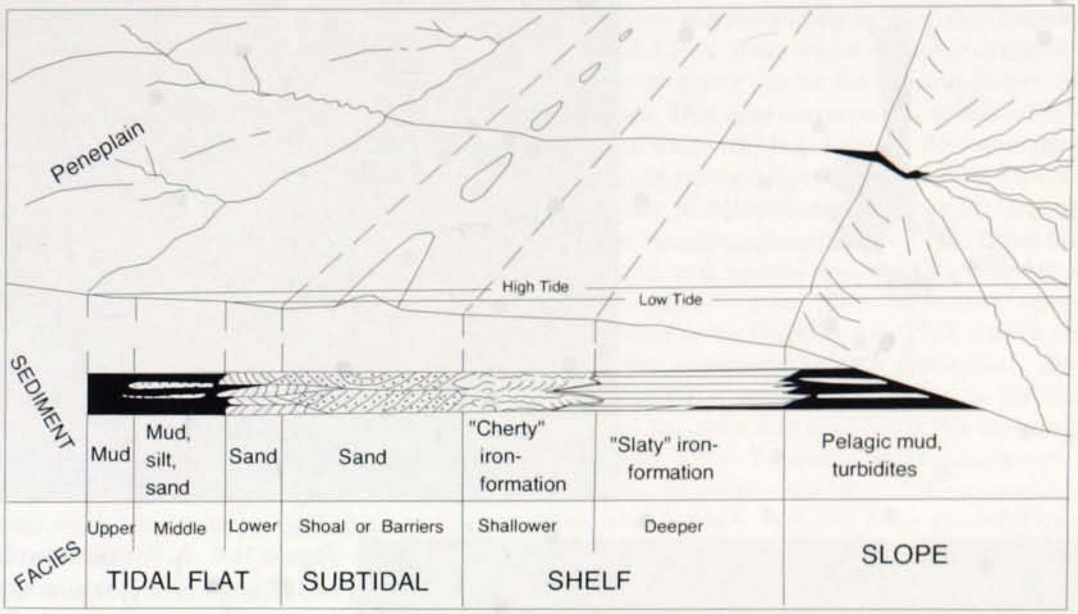


Figure 11.8. Sedimentation model showing lateral relationships of the siliciclastic tidal facies of the Pokegama Formation, the two main facies of the Biwabik Iron Formation, and the Virginia Formation (on the slope?). Thicknesses and geography area not to scale; modified from Ojakangas (1983).

The two sand-textured members (Lower Cherty and Upper Cherty) formed in a shallow-water, high-energy environment, as indicated by stromatolites, cross-bedding, and rounded (locally oolitic) grains of iron minerals and chert. Shoreward-moving tidal currents (flood tides) and/or storms may have disrupted the mud-textured sediment (precipitates) and transported sand-sized aggregates into shallower water where they were altered by seafloor processes and early diagenetic processes. Thus these granules are interpreted as "intraclasts" derived from within the basin.

Shallow channels up to a mile wide and tens of feet deep were cut into the Lower Slaty member and filled with sand-textured grains of iron minerals and chert in the Virginia horn area. These grains apparently were derived from shallow water and carried seaward into the deeper water environment in which the iron minerals were precipitating. Ebb-flow tidal currents are interpreted as the erosion and transportation agent.

A plot of 102 cross-bed measurements in the Minorca Mine on the northeast edge of the Virginia horn (Fig. 11.2) shows 90 percent of the readings making a very prominent mode to the north–northeast and a minor, broader mode to the south (Fig. 11.9). This distribution is interpreted as the product of a strong flood tide toward the paleogeographically determined northern shoreline and a much weaker ebb tide.

A study of the orientations of stromatolite mounds in the stromatolite horizon within the Upper Cherty member was conducted by Boerst (1999). His map is presented in Figure 11.10. A paleocurrent plot of mound elongation (Fig. 11.10) is interpreted as the result of shore-normal tidal currents and shore-parallel longshore currents in shallow water.

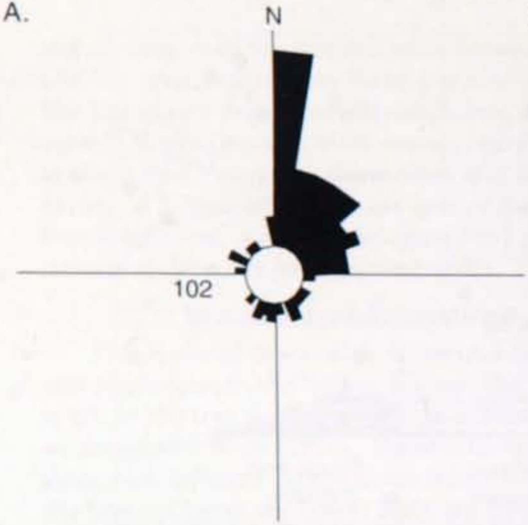
The repetition of the cherty and slaty members has long been interpreted as the result of transgression and regression (White, 1954).

The Virginia Formation was deposited seaward of the iron-formation, probably in a slope-type environment (Fig. 11.8) where episodic turbidity currents deposited graded beds. Some volcanic ash falls evidently settled into the basin forming graded beds with a totally volcanic composition. The dominance of black, fissile shale suggests the "raining out" of clay (such as settling through the water column) and deposition in deep, anoxic water below the wave base. Minor, thin, sandstone lenses were deposited by bottom currents (Lucente and Morey, 1983).

MINERALOGY

The detailed origins of the iron minerals are exceedingly complex and are beyond the scope of this

A.



B.

C.

Figure 11.9. **A.** Paleocurrent rose diagram of 102 cross-bed measurements from the Lower Cherty member in the Minorca Mine.

B. Photo of cross-bedding in the Minorca Mine.

C. Photo of herringbone cross-beds in the Minorca Mine.

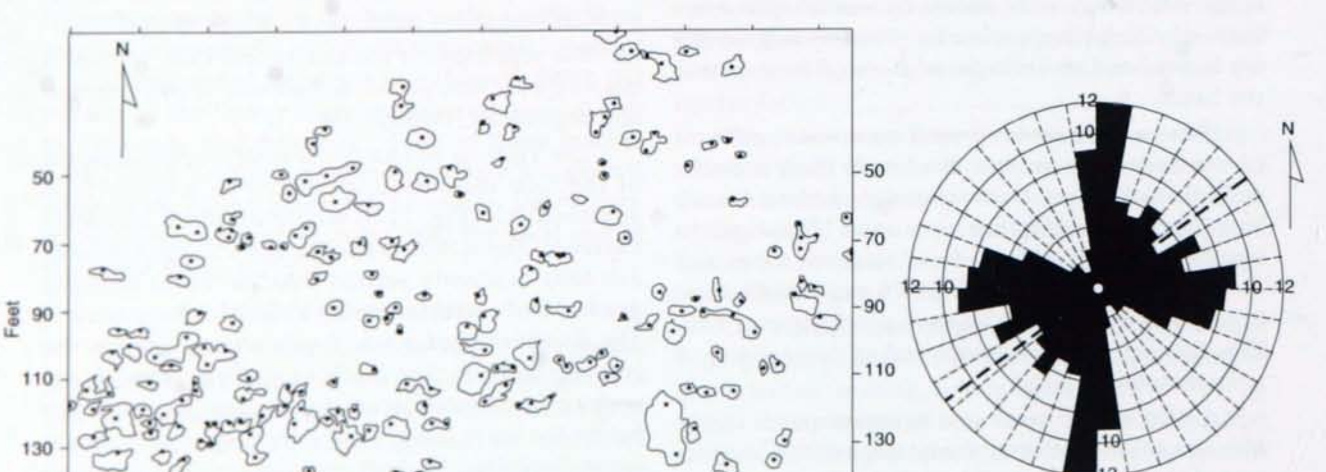


Figure 11.10. Mapped stromatolite mounds in the algal submember (I submember) in the Upper Cherty member of the LTV 2E pit. The rose diagram represents the elongation of the mounds, with each elongate mound plotted on both sides of the rose diagram. From Boerst (1999).

introduction. In brief, Eh and pH are major controls on the stability of the various iron minerals in both the depositional and diagenetic environments, and in the easternmost Mesabi range, in the metamorphic environment as well. Recrystallization and replacement of the granules during diagenesis has been extensive, and probably consisted of a number of discrete events.

Earlier work on the oxidized taconites of the western Mesabi range was accomplished by Bleifuss (1964). He showed that late hematite was developed by the oxidation and pseudomorphic replacement of magnetite octahedra, that layers of goethite were precipitated from solutions likely derived from the oxidation of siderite, and that some goethite formed by the oxidation of acicular iron silicate minerals. Additional work was done by Ojakangas in Zanko and others (2003).

All of the magnetite grains are euhedral and are interpreted as late diagenetic in origin. Some of the hematite inclusions and crystals in magnetite are similar to those illustrated by Han (1982). He proposed that much of the magnetite formed by the replacement of, and overgrowth on, pre-existing hematite that served as nuclei. Han further suggested that ionic diffusion of ferrous iron was a key process in the formation of the magnetite. Organic carbon may have acted as a reductant in this process.

The nature of the major hydrologic events that removed 40 to 60 percent of the silica and oxidized the iron minerals, thus forming the high-grade (natural) ore bodies, has long been debated. Were they descending, cool, meteoric waters or ascending hydrothermal waters related to igneous activity? Did this occur during the Cretaceous (the age of conglomerates composed of clasts of high-grade hematite), or prior to that time? Morey (1999) provided an excellent review of the arguments. He then proposed that a large-scale, topography-driven, hydrothermal ground-water system moved waters northward through the sands of the underlying Pokegama Formation, from the vicinity of the regional Penokean orogenic uplift in northern Wisconsin and east-central Minnesota, 40 to 80 miles to the south.

PRODUCTION FIGURES—IRON ORE AND TACONITE

The annual amounts of direct-shipped ore and taconite produced from the Mesabi Iron Range are shown in Figure 11.11. Production and shipping of direct ore started in 1892 and rose steadily until 1953 when a maximum 76 million tons were produced in one year (note the precipitous drop in direct ore production corresponding to the Great Depression). At around 1955, there was a dramatic decrease in the amount of direct ore as the various mines became depleted. This also corresponds to the initial start-up of taconite mining, using a concentrating and pelletizing method developed by E.W. Davis of the University of Minnesota. Reserve Mining opened the first taconite operations in 1955 (Peter Mitchell Mine) and was shortly followed by Erie Mining in 1957 (the old LTV site). Six more taconite operations were added in the 1960s, and by 1967, annual taconite production exceeded direct ore production. The mid-1980s marked a serious depression in the iron ore and steel industry that resulted in the closure of one operation (Butler Taconite) and the bankruptcies of two other taconite producers. More recently, LTV Steel and Eveleth Taconite have closed; Evtac has since reopened as United Taconite.

WHAT'S IN A NAME? (THOSE CONFUSING IRON-FORMATION SUBMEMBERS!)

The four-fold stratigraphy of Lower and Upper Cherty and Lower and Upper Slaty members (Wolff, 1917) is still used at each of currently operating (and inactive) taconite mines on the Mesabi Iron Range. However, each of the mining companies further subdivides the Biwabik Iron Formation into several submembers based on bedding types (Fig. 11.12) and mineral assemblages. It is at this point that the Biwabik Iron Formation stratigraphy becomes very complicated and at times confusing. This is mainly due to the following reasons:

- There are localized lateral facies changes between mines (and even within a single mine). Some mines reconcile these differences by splitting out numerous submembers (each with a distinct bedding type, texture, ore grade, and/or mineral assemblage), whereas other mines lump many of these same differences within a single submember.
- There are significant lateral facies changes over several miles between mines. For example, a particular horizon may be massive-bedded at one location but is regular-bedded a few miles away. This is particularly troublesome within the Upper Cherty member in the western two thirds of the Mesabi range.
- Not all mines use the same numbering system - some use abbreviations (for example LC—Lower Cherty member) followed by a number (as in LC-5 at the top of the Lower Cherty member). However, other mines use an alphabet system,

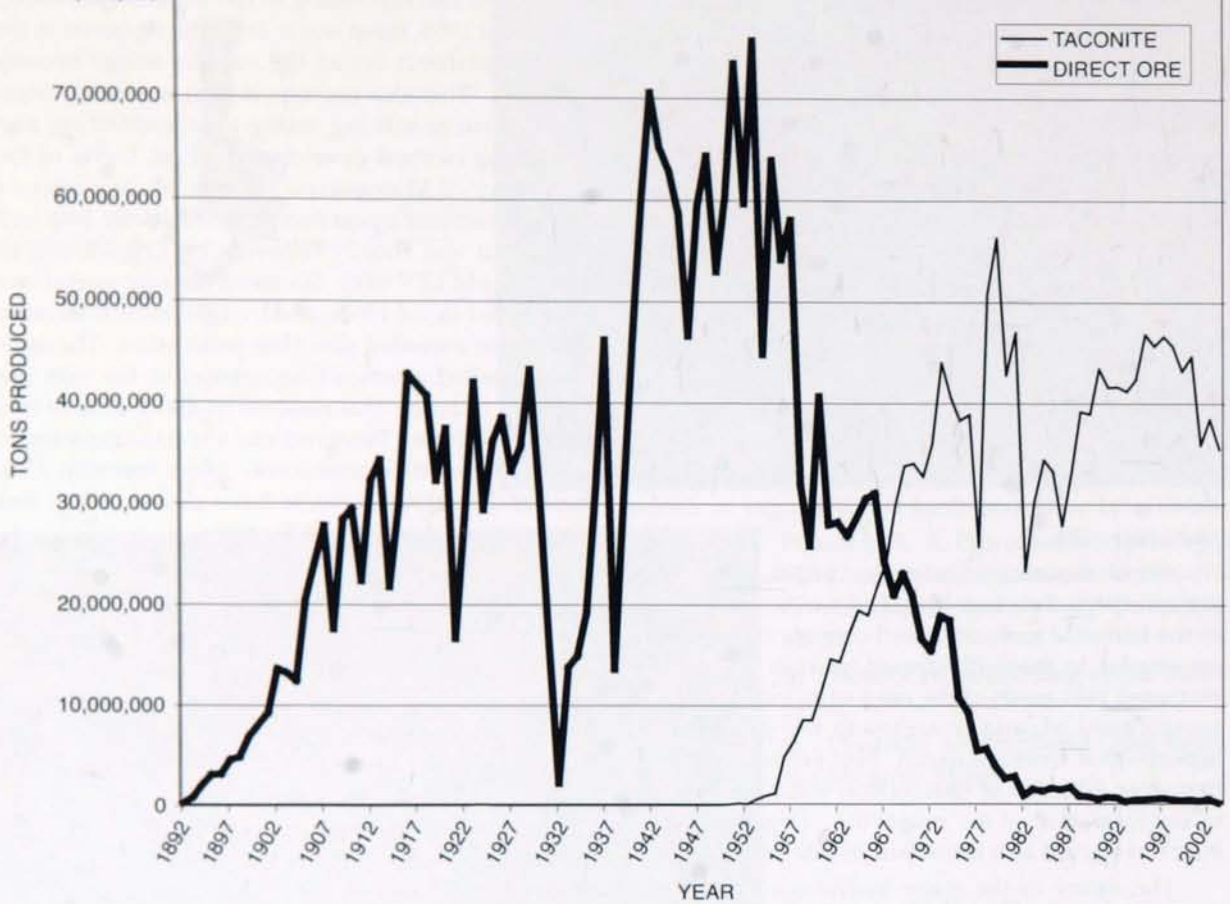


Figure 11.11. Annual production figures for direct ore (includes all forms of direct ore) and taconite for the period from 1892 to 2003 from the Mesabi Iron Range. Data and graph from James Sellner, Minnesota Department of Natural Resources, Division of Lands and Minerals, Hibbing, Minnesota.

devised by Gundersen and Schwartz (1962), starting with the A submember at the top of the Upper Slaty member (in this system the top of the Lower Cherty member corresponds to the R submember). And further still, another mine refers to the Lower Cherty member as the number 1 unit and subdivides it into eight submembers, with 1-8 at the top of the Lower Cherty member.

- Some mines label downward in their numbering system, whereas other mines label upward in their numbering system.

The submember nomenclature that is used at each of the mines is summarized in Figure 11.13. It can readily be seen on this summary chart that any particular submember name changes nomenclature from one mine to the next. This is because there are few good marker horizons within the Biwabik Iron Formation, and even these can exhibit gradual lateral facies changes or pinch-and-swell relationships to each other. A few of the potential marker horizons within the Biwabik Iron Formation are presented below.

- *Top contact of the Biwabik Iron Formation with the Virginia Formation*—In the eastern half of the Mesabi Iron Range a carbonate horizon is present at the very top of the Upper Slaty member and the contact between the Biwabik Iron Formation and Virginia Formation is easily recognized (Gruner, 1924). However, to the west of Hibbing, the carbonate layer is absent and lenses of thin-bedded iron carbonate iron-formation are present in the Virginia Formation, and the top of the Biwabik Iron Formation is not easily discerned.
- *Submember I/Algal unit*—A thin unit containing algal stromatolites and jasper-bearing intraformational conglomerate is present near

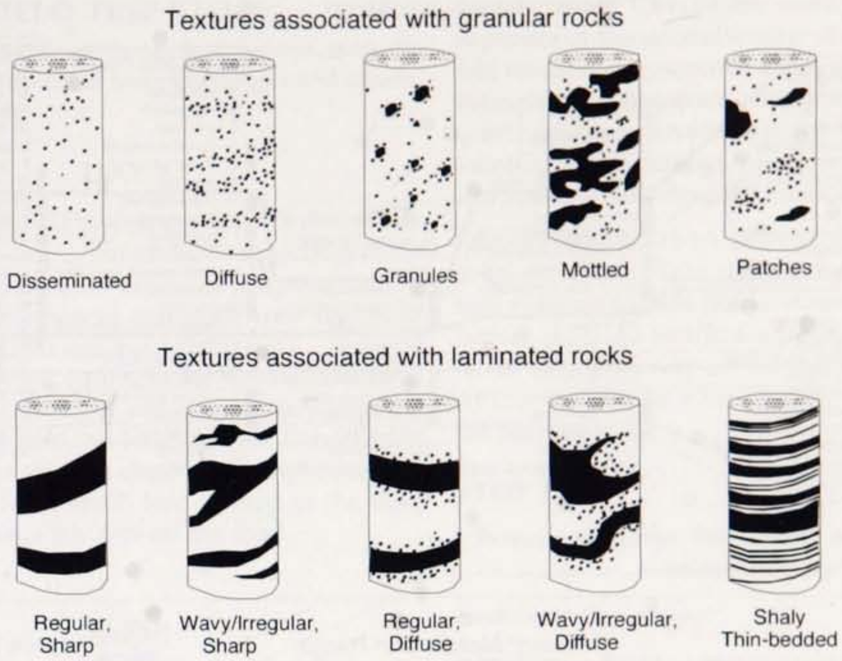


Figure 11.12. Textural characteristics of the Biwabik Iron Formation (from a classification scheme developed by geologists of the Hanna Mining Company; modified from Pfleider and others, 1968).

the top of the Upper Cherty member. This submember is easily recognized but is not present west of Hibbing.

- *Lower Slaty member*—The Lower Slaty member has a very well-defined "Intermediate Slate" (also referred to as the Q submember or Paint Rock in Figure 11.13) at its base that is characterized by a black, carbon-rich, thin-bedded, slaty unit that commonly contains pyrite. This unit is readily evident at all of the mines on the Mesabi range. However, the upper contact of the Lower Slaty member "...is indefinite and a gradual change to other slaty phases takes place." This "...makes the dividing line between the two [Upper Cherty and Lower Slaty member] somewhat arbitrary." (Gruner, 1924, p. 20). The upper contact of the Lower Slaty member is particularly troublesome in the Virginia horn area. Gruner (1946, p. 45) included lenses of cherty and wavy-bedded taconite (referred to as the Interbedded Chert—IBC unit at Minntac; Fig. 11.13) in the Lower Slaty member, whereas White (1954) included these same units in the overlying Upper Cherty member.
- *Base of the Biwabik Iron Formation*—The base of the Lower Cherty member is generally characterized by thin-bedded iron-formation (also called the "red basal unit") with localized algal stromatolite and basal conglomerate horizons. However, at many localities the base of the Biwabik Iron Formation exhibits a gradational contact with the underlying Pokegama Formation. In the Virginia horn area, the base of the Biwabik Iron Formation contains an iron-bearing sandstone (White, 1954) that some mines include with the iron-formation, whereas others lump this type of material with the Pokegama Formation.

From the above description it is readily evident that there are few good marker horizons within the iron-formation; even the upper and lower contacts of the iron-formation are gradational and subject to various interpretations. The "Intermediate Slate" and the algal horizon in the Upper Cherty member are the only easily recognizable marker units. However, even using these horizons as markers, one can see from Figure 11.13 that there are problems.

Clearly, much additional work needs to be done to understand how submembers at one mine correlate with submembers at an adjacent mine. These types of studies could inevitably be important in determining why ore grades, and waste rock characteristics, change between mines and even within a single mine. For example, some of the best ore-grade taconite corresponds to the

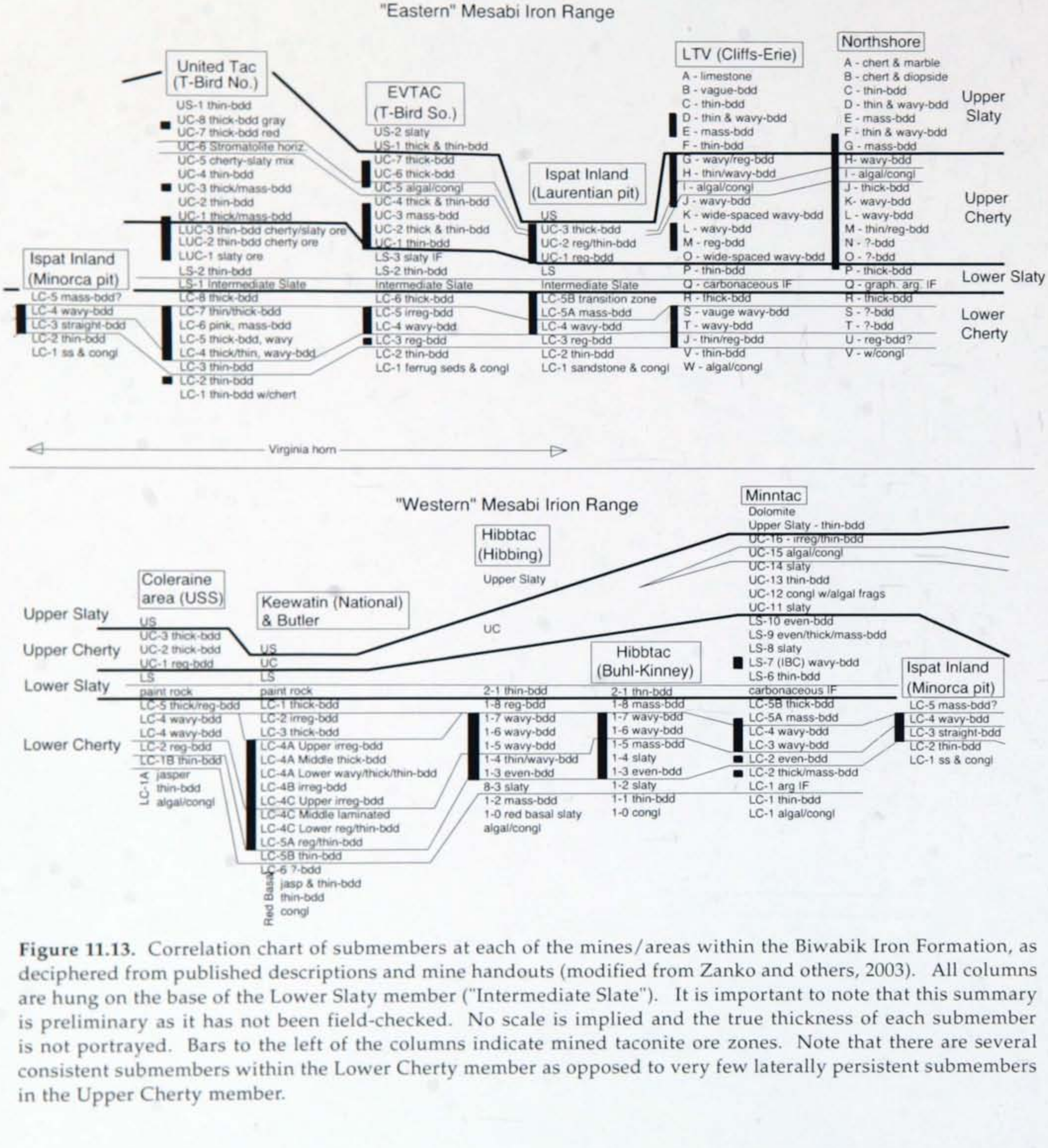


Figure 11.13. Correlation chart of submembers at each of the mines/areas within the Biwabik Iron Formation, as deciphered from published descriptions and mine handouts (modified from Zanko and others, 2003). All columns are hung on the base of the Lower Slaty member ("Intermediate Slate"). It is important to note that this summary is preliminary as it has not been field-checked. No scale is implied and the true thickness of each submember is not portrayed. Bars to the left of the columns indicate mined taconite ore zones. Note that there are several consistent submembers within the Lower Cherty member as opposed to very few laterally persistent submembers in the Upper Cherty member.

wavy-bedded or irregular-bedded taconite present in both the Lower or Upper Cherty members. The sedimentary environment that produced this type of taconite ore has not been fully documented nor have any recent detailed sedimentological studies been attempted. A better understanding of the various sedimentological textures in the Biwabik Iron Formation could ultimately lead to an increased ability to better predict changes in ore grades as they relate to facies changes.

FIELD TRIP STOPS

Note that many of these descriptions, and the text above, are modified from Ojakangas and others, (2004; Fig. 11.14).

DAY 1

DIRECTIONS: On U.S. Highway 53 in Eveleth, there is a stoplight at Grant/Industrial Avenue. Just north of this intersection there are some low road cuts of Archean metagraywacke and slate; note that both bedding and cleavage are subvertical. Proceed northward past the stoplight for about 0.4 mile to a gas station. Turn right and immediately turn left on the frontage road (Midway Drive). Drive past a church and past the first street on the right (Mesabi Lane). Watch for a small, low outcrop in the trees on the right, just a few feet off the road.

STOP 11-1A

No hammering please!

Pokegama/Archean unconformity

Location: T. 58 N., R. 17 W., sec. 20, NW, SW, SE

Eveleth quadrangle; UTM: 535,445E/5,259,520N

Description: This is the only easily accessible exposure of the unconformity. A close examination will reveal the presence of a thin smear of Pokegama Formation conglomerate composed of schist and vein quartz clasts upon Archean schist with a near-vertical foliation. Note that the foliated schist clasts are flat and have a subhorizontal orientation.

NEXT: From Stop 11-1A, drive a long block northward to Merritt Drive. Turn right. About 100 feet farther, turn right on Mesabi Drive. About another 100 feet farther, there is a fork in the street. Take the left fork to the first driveway on the left (#7 Mesabi Lane). At this location is a broad, flat, rock exposure on the driveway.

STOP 11-1B

Private property! Permission must be obtained before entering!

No hammering please!

Jaspillite on Archean metaconglomerate

Location: T. 58 N., R. 17 W., sec. 20, SW, SE (7 Mesabi Lane)

Eveleth quadrangle: UTM: 535,610E/5,259,460N

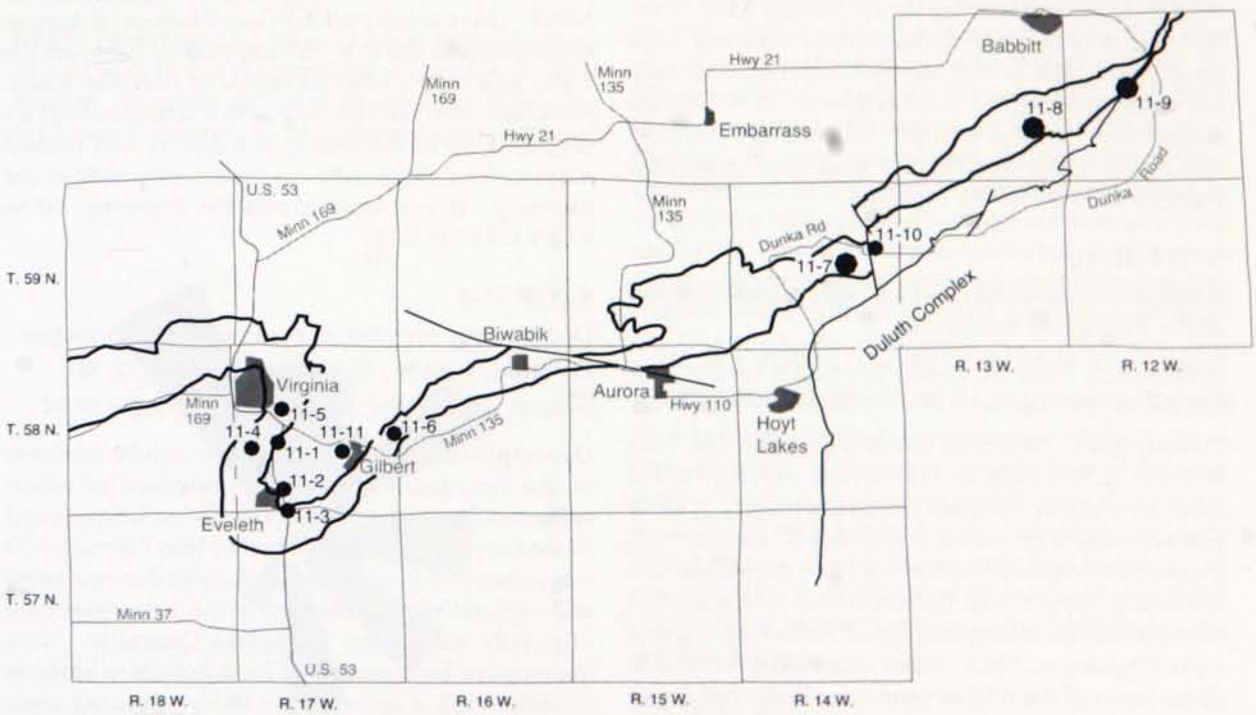


Figure 11.14. Locations of stops that will be visited on the field trip. Generalized contacts of the Biwabik Iron Formation for the eastern half of the Mesabi Iron Range, and Duluth Complex are shown. Note that in order to observe Stops 11-1, 11-2, and 11-3 in stratigraphic order, we will be back-tracking for short distances.

Description: This is an excellent exposure of jaspillite resting unconformably on subvertical Archean volcanogenic metaconglomerate and metasandstone. The outcrop is 30 to 40 feet topographically higher than Stop 11-1A, and presumably was that much higher when the jaspillite was deposited. Is it an erosional remnant of basal Biwabik Iron Formation or is it a local chemical precipitate at the base of the Pokegama Formation? The Archean conglomerate and lithic sandstone that form the driveway here are part of the northeast-trending Midway sequence, containing these strata types locally interbedded with subaerially deposited, calc-alkalic (trachyandesitic) volcanic rocks (Jirsa and others, 2004). The sequence is inferred to have formed after earliest deformation (D_1) of the enclosing graywacke and basaltic rocks of the Mud Lake sequence, but before the cleavage-forming D_2 deformation that affected both sequences. The conglomerate contains clasts of basalt, graywacke, porphyritic trachyandesite, and quartzofeldspathic porphyry. This provenance indicates that the older Archean rocks of the Mud Lake sequence were intruded by quartzofeldspathic porphyry, deformed, and uplifted, to provide detritus to what was probably a successor or "pull-apart" basin developed along a major structure now occupied by the Pike River fault zone (Jirsa, 2000).

NEXT: Continue driving down Mesabi Lane to the frontage road (Midway Drive). Turn right and drive past Stop 11-1A to the second stop sign and turn left to the intersection with Highway 53. Cross the northbound lane and turn left. Drive several hundred feet south to the middle of a long road cut on the right (west) side of the highway.

STOP 11-1C

Argillaceous lower member and silty middle member of the Pokegama Formation

Location: T. 58 N., R. 17 W., sec. 20, NW, SW, SE

Eveleth quadrangle; UTM: 535,380E/5,259,555N

Description: This road cut is about 500 feet long and 5 to 10 feet high, and is the only exposure of the *lower argillaceous member*. It consists largely of shale and siltstone with minor fine-grained sandstone. It has been interpreted as having been deposited in a low-energy upper tidal flat environment in a sea that transgressed onto the peneplaned surface of Archean rocks (Ojakangas, 1983). Minor channeling is common at the bases of the thicker sandstone beds, and at one spot, 0.5 meter of section has been eroded. Small-scale cross-bedding is present in some siltstone beds, and elongated sole marks are visible on the bottoms of some sandstone and siltstone beds. A total of 57 of these paleocurrent indicators show that the currents were generally oriented in a north–south direction, parallel to the presumed paleoshoreline that was located immediately to the east. A few concretions as large as 6 inches in diameter are present, as is soft-sediment deformation. Hemming and others (1991) illustrated the soft-sediment folding and interpreted it as evidence that the Animikie basin was tectonically active during deposition of the Pokegama Formation and the Biwabik Iron Formation. Alternatively, it is interpreted herein as soft-sediment slumping into tidal channels. Fine laminations and sequences of laminations in the sandstone beds have been interpreted as tidal rhythmites (Ojakangas, 1996).

Above this road cut, in the brush and trees on the hillside, the *middle silty member* is exposed in artificial cuts along an ATV trail. A short walk to the south provides easy access to the trail (rather than clambering uphill through the brush). Blocks of the more massive *upper sandy member* are present higher on the hillside above the trail.

Note: The middle silty member of the Pokegama Formation was once poorly exposed in the flat area across the highway from the U.S. Hockey Hall of Fame. However, the best part of the poor exposure has since been covered by a frontage road.

NEXT: Drive south past the Grant/Industrial Avenue stoplight, past the U.S. Hockey Hall of Fame on the right, and stop along the highway near the Rustic Rock Inn. You are now opposite a large road cut on the east side of the four-lane highway and parked adjacent to a small road cut on the west side of the highway. If you walk across the highway, do so VERY CAREFULLY!

STOP 11-2

Upper sandy member of the Pokegama Formation

Location: T. 58 N., R. 17 W., sec. 32, E 1/2, SE

Eveleth quadrangle; UTM: 536,055E/5,256,775N

Description: This is the upper sandy member of the Pokegama Formation, composed of silica-cemented quartz sandstone. It is the rock type found immediately beneath the Biwabik Iron Formation; it was penetrated by countless drill holes during mining and exploration, and resulted in the formation being originally named the Pokegama Quartzite. Note the massive beds separated by thin beds of shale or siltstone. Silica cementation likely obscured some original cross-bedding. This member was interpreted by Ojakangas (1983) as having been deposited in a high-energy, lower tidal or subtidal environment.

NEXT: Proceed south on Highway 53 for 0.6 mile (passing an overpass and a long road cut of iron-formation on the right) to a left-turn lane. Make a U-turn and proceed north on the highway for about 0.2 mile to a road cut on the off-ramp for State Highway 37 (right side of highway).

STOP 11-3

Lower Cherty member, Biwabik Iron Formation
Location: T. 57 N., R. 17 W., sec. 5, E 1/2, NE, NE
Eveleth quadrangle; UTM: 536,230E/5,256,285N

Description: Note that this member of the Biwabik Iron Formation overlies the sandy member of the Pokegama Formation of the last stop, and that both units dip gently to the southeast. Observe the thick wavy bedding, the trough cross-bedding, and the sandy texture of the iron minerals and chert grains. The cross-beds are best observed in this eastern road cut, but are also present at both ends of the longer cut on the other side of the highway. Cross-bedding measurements (Fig. 11.15), although not definitive, are suggestive of a tidally-influenced marine environment (also see Figure 11.9). This, coupled with Walther's law of succession of sedimentary facies (the facies observed vertically are also similarly related laterally), places the deposition of the iron-formation seaward of the Pokegama Formation.

NEXT: From Stop 11-3, proceed north along Highway 53 past Stop 11-2 to the stoplight at Grant/Industrial Avenue. Turn left on Grant Avenue and follow the road past a cemetery to the entrance road to United Taconite (U-Tac). Stops 11-4A and 11-4B are at the Thunderbird North Mine of United Taconite (formerly Eveleth Taconite, or Evtac). Simplified stratigraphic submembers of the iron-formation, according to terminology used by Evtac, are portrayed in the two left-hand columns of Figure 11.16.

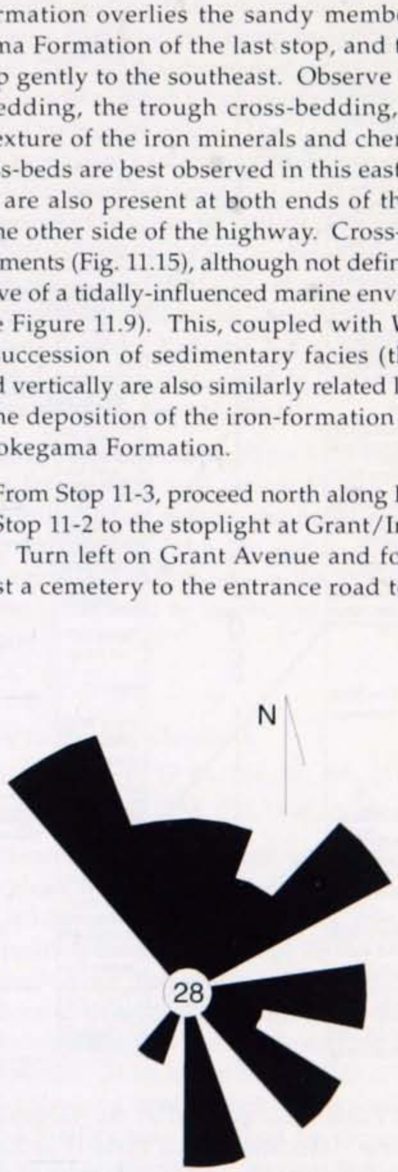


Figure 11.15. Rose diagram of 28 cross-bedded measurements, mostly from the road cut of Stop 11-3, with a few from nearby exposures.

STOP 11-4A

Submember LC-3 near the base of the Lower Cherty member; United Taconite's Thunderbird North Mine

Location: T. 58 N., R. 17 W., sec. 20; northeast end of the Thunderbird North Mine
Eveleth quadrangle; UTM: 535,290E/5,260,255N

Description: The LC-3 submember (new U-Tac terminology or "footwall ore" in the old Evtac terminology) is characterized by regular-bedded and wavy-bedded cherty taconite wherein most of the magnetite is contained within dark-colored wavy (irregular) beds that locally show good pinch-and-swell relationships. This is the lowest unit mined at the Thunderbird North Mine and contains approximately 20 to 23 percent mag-Fe and 2 percent Davis tube silica.

The Auburn fault can also be viewed at this locality. The fault zone consists of a 20-foot-wide zone of broken-up and highly oxidized vuggy rock. To the northwest is the Auburn Mine that was originally developed as an underground mine by the Minnesota Iron Company between 1894 and 1902. It was reopened as an open pit by the Oliver Iron Mining Company (a subsidiary of United States Steel Corporation) in 1951 and was essentially closed in the 1960s; scram operations continued intermittently until 1999. The mined ore was located along a main ore trough that coincides with the northwest-trending Auburn fault. Mined material was obtained from the Upper Cherty and Lower Slaty members, and locally from the Lower Cherty member. Jim Small of Edward Kramer and Sons (unpub. data) reported that he was unable to find any evidence of the Auburn fault anywhere on the newly-scraped floor of the mine during the final scramming period prior to allowing the pit to fill with water. United Taconite is currently filling the mine with waste rock and Highway 53 may eventually be relocated to pass over this filled site, to allow taconite mining to the northeast of the Auburn Mine.

STOP 11-4B

Submembers LC-2 through LC-8 of the Lower Cherty member and the contact with the Lower Slaty member

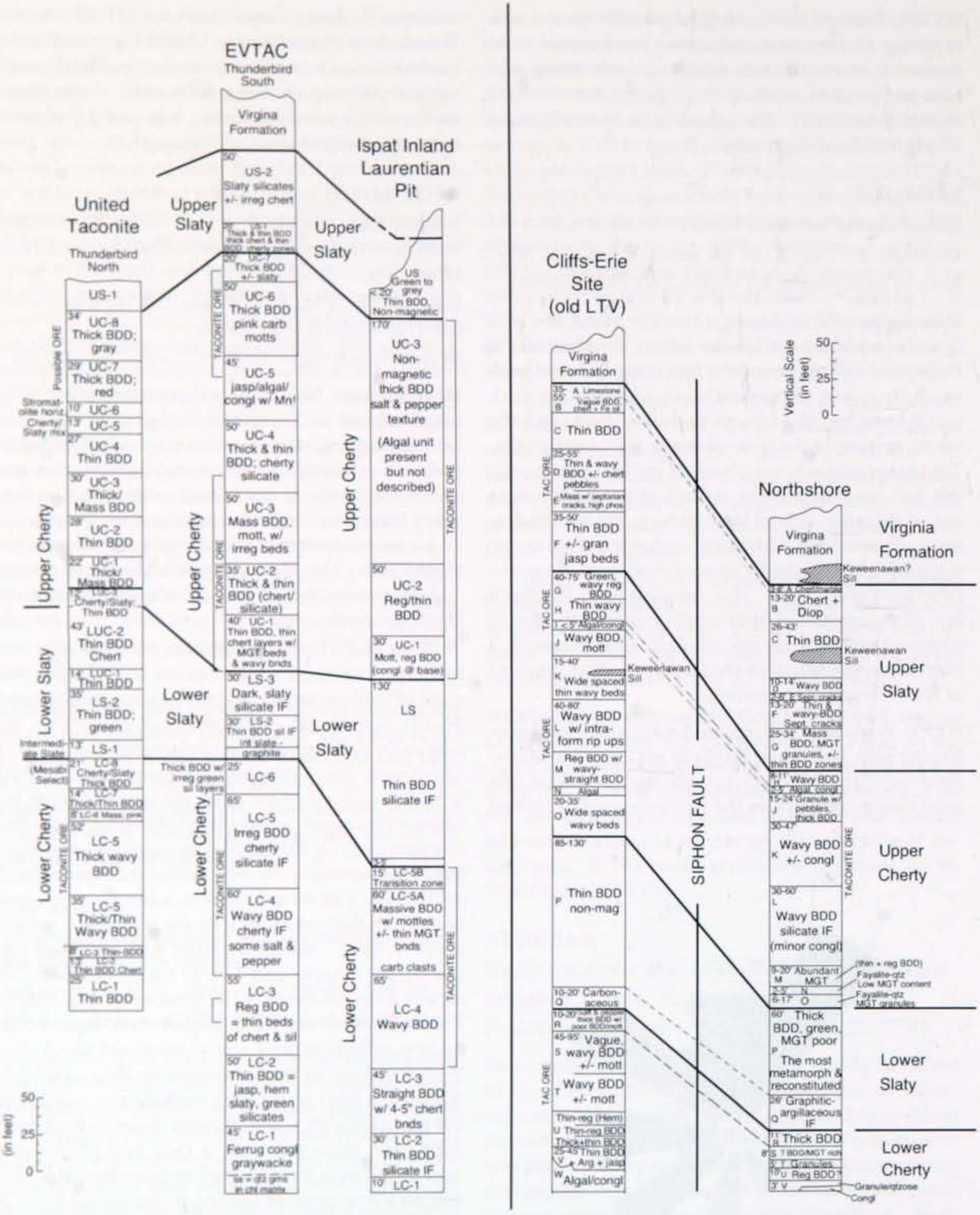


Figure 11.16. Submember nomenclature for the Biwabik Iron Formation as used at the various taconite mines and idled mine sites that will be visited on this field trip. Modified from Plate II in Zanko and others (2003). Mined horizons are shown by a bar on the side of the column. Note that the nomenclature portrayed here is preliminary in nature and future correlations of the various submembers used at the different mines will need field checking and modification.

Location: T. 58 N., R. 17 W., sec. 19; north-central side of the Thunderbird North Mine

Eveleth quadrangle; UTM: 533,915E/5,259,540N

Description: About 150 vertical feet, or most of the Lower Cherty member can be viewed at this site. Most of the units are thick-bedded and constitute taconite ore. Toward the bottom of the vertical section is the LC-4 submember that exemplifies typical wavy-bedded taconite characteristic of much of the taconite ore of the Mesabi range. Within the LC-4 submember the magnetite is in: 1. The wavy beds, 2. Disseminated throughout the cherty bands, and 3. Within mottles that are generally less than 1 centimeter in diameter and cored by iron-carbonate (siderite). The LC-4 submember is 40- to 45-feet-thick and contains 20 to 27 percent mag-Fe and 1 to 2 percent Davis tube silica. The LC-4 submember is easily recognized in drill core due to the wavy beds and a salt-and-pepper texture that is defined by disseminated magnetite. Waste materials in the vertical section include the LC-3 submember (referred to as slaty waste) and the LC-8 submember (used as aggregate and referred to as "Mesabi Select"). The upper contact of the Lower Cherty member with thin-bedded and iron-poor rocks of the Lower Slaty member can also be seen at this site.

NEXT: From United Taconite, drive back to Highway 53. Turn left (north) and follow Highway 53 for about 1.5 miles to just beyond the point where State Highway 135 merges from the right. A large sign marks the turnoff to Mineview in the Sky Overlook. Follow the small road to its end on top of a waste rock dump.

STOP 11-5

Mineview in the Sky Overlook

Location: T. 58 N., R. 17 W., sec. 17, NE, NW, SE

Eveleth quadrangle; UTM: 535,710E/5,261,650N

Description: The open pit below the overlook is the Rouchleau Mine that connects northward with a dozen additional mines. Ore was mined from this composite pit from 1893 until 1997. More than 300 million tons of iron ore and 250 million tons of waste material (lean rock plus glacial overburden) were removed from this manmade canyon that is as deep as 450 feet. Most of the high-grade ore formed along a 4-mile-long, north–northwest-trending fault, along which silica was removed by ground water (meteoric or hydrothermal?), leaving the low-grade iron-formation (about 30 percent iron) enriched to 50 to 55 percent iron. The Rouchleau Mine produced 52 million tons of this high-grade "natural ore" from 1920 to 1976 (Fig. 11.17).

This area is located on the "Virginia horn," a large, Z-shaped bend in the otherwise straight east–northeast-trending Mesabi range (Figure 11.14). The town of Virginia is visible to the north-northwest. Six miles to the northwest on the horizon is United States Steel's Minntac taconite operation. Steam visible to the northeast is from the Ispat Inland taconite plant. To the west is United Taconite's Thunderbird North Mine.

NEXT: From the Mineview Overlook we will proceed to the Laurentian Mine of Ispat Inland by driving south on Highway 53 (less than 0.5 mile), and then taking Highway 135 east about 3 miles to the town of Gilbert. At the junction of Highway 135 and 37, turn left and proceed north to the gated entrance of the Laurentian Mine. The stratigraphic terminology of submembers in the Laurentian Mine is portrayed on Figure 11.16.

STOP 11-6A

Lower Slaty member and the "Intermediate Slate" at the base of the Lower Slaty member, Ispat Inland's (Mittal Steel Company) Laurentian Mine

Location: T. 58 N., R. 17 W., sec. 24, NW, SE, NE; Laurentian Mine

Gilbert quadrangle; UTM: 542,315E/5,260,700N to 5,424,85E/5,260,840N

Description: The entire stratigraphic section of the Lower Slaty member (130-feet-thick) can be viewed at this impressive face within the Laurentian Mine. The "Intermediate Slate," at the base of the Lower Slaty member, is characterized by thin-bedded, black, organic-rich slate that locally exhibits bright, shiny graphitic surfaces with bedding-parallel slickensides. Pyrite is common to this submember and is present as both disseminated fine- to medium-grained cubes and as thin disks (marcasite?) along bedding planes. All of the Lower Slaty member constitutes waste rock.

Note that the name "slate" has been applied to all thin-bedded rocks in the Biwabik Iron Formation, but this term is a misnomer, because these rocks are essentially unmetamorphosed and do not have the cleavage of a true slate but merely a parting parallel to bedding (White, 1954). Morey (1992, 1993) reported that the "Intermediate Slate" possibly contains an ash-fall component. It has the highest Al_2O_3 content of any analyzed sample of the Biwabik Iron Formation.

STOP 11-6B

UC-1 submember at the base of the Upper Cherty member

Location: T. 58 N., R. 17 W., sec. 24, SE, NE, NE; Laurentian Mine

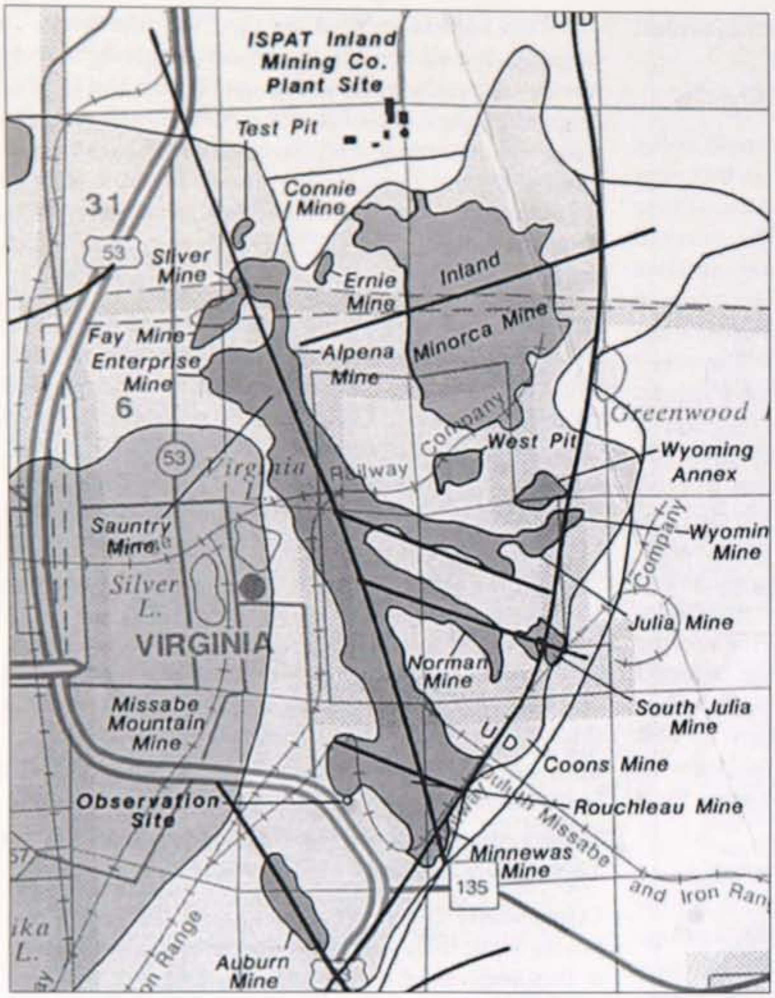


Figure 11.17. General distribution of natural-ore zones (dark shaded) along a series of northwest-trending faults in the Rouchleau Mine and adjacent mines (from Meineke and others, 1999).

Gilbert quadrangle; UTM: 542,545E/5,260,790N

Description: The UC-1 submember is a regular-bedded (1 to 2 inches straight/even beds) granular cherty unit that contains magnetite-rich bands, pink Fe-carbonate mottles, and local cross-bedding. This material constitutes ore in that it contains 16 to 40 percent mag-Fe (the mag-Fe is extremely variable in this unit) and 2 to 4 percent Davis tube silica. Localized pods of very iron-rich taconite (up to 40 percent mag-Fe) were present but have been recently mined out.

NEXT: Return to the mine entrance, turn left on Highway 135 and proceed east past Biwabik for about 1.5 miles to St. Louis County Road 138. Turn left and continue north 3.5 miles to Giants Ridge resort.

OVERNIGHT AT GIANTS RIDGE RESORT

DAY 2

NEXT: Return south to Highway 135 and proceed east about 2 miles, where 135 heads north (left turn). Go about 6 miles to north (gated) entrance to of the Cliffs-Erie site (old LTV mine). Continue down a private drive to the guard shack near the office buildings of the Cliffs-Erie site. After receiving permission to enter the property, go straight and follow this road about 3.5 miles to a T-intersection with Dunka Road (private company road). Follow Dunka Road to Stop 11-7. The stratigraphic nomenclature for the Cliffs-Erie site is presented in Figure 11.16.

STOP 11-7

Algal submember (I submember) near the top of the Upper Cherty member, Cliffs-Erie site

Location: T. 59 N., R. 14 W., sec. 23, N 1/2, NW; Cliffs-Erie site (old LTV Pit 2E). Access to this site is via Dunka Road, which is a private mining company road.

Allen quadrangle; UTM: 568,202E/5,2706,25N

Description: Algal structures were first described by Leith (1903) as "contorted bedding." Grout and Broderick (1919) were the first who assigned an organic origin to them. The algal submember within the Upper Cherty member consists of mounds of fossilized algal colonies that are separated by jasper-bearing intraformational conglomerate; the thickness is 2 to 20 feet. This horizon occurs only in the eastern half of the range (not present west of Hibbing) and is only sporadically present between Hibbing and Chisholm.

This locality is an excellent place to view a nearly horizontal portion of the iron-formation that contains abundant individual mounds of algal stromatolites. Stripping of glacial overburden in this area once revealed a dip slope the size of a football field that contained stromatolite mounds (Graber, 1993). Figure 11.10 illustrates a large portion of that exposure that has since been mined. The present stop is at an area located several hundred feet west of that site. Internally, the mounds are characterized by many individual, columnar, finger-like structures that are convex upward. The mounds protrude up through a thin veneer of the overlying thin- to wavy-bedded H submember. Measurements on a nearby mine face in this horizon showed that all the columnar stromatolites were inclined at 30° to the vertical; unfortunately, that site has also been removed by mining.

Stromatolite samples may be collected at the extreme eastern edge of this exposure. Also at this locality, the J and H submembers rarely contain anthraxolite, which is an organic bitumen containing 95 percent or more carbon that is black with a vitreous luster and conchoidal fracture and resembles obsidian (Morey, 1994). Morey (1994) reported that anthraxolite is present throughout the iron-formation but is most common beneath the carbon-rich "Intermediate Slate." Furthermore, he suggested it formed via a mechanism of concentrating carbon from a mass-kill phenomenon, followed by later migration of a carbon-rich liquid to form the anthraxolite.

NEXT: Proceed eastward about 14 miles down Dunka Road (through a locked gate) to the Peter Mitchell Mine operated by Northshore Mining. The stratigraphic nomenclature of submembers in the Peter Mitchell Mine is portrayed on Figure 11.16.

STOP 11-8

Submembers C, D, F, and G (Upper Slaty member), H and I (Upper Cherty member) at Northshore Mining's Peter Mitchell Mine

Location: T. 60 N., R. 13 W., sec. 26, S 1/2, NE; several subunits of the Upper Cherty member can be viewed within short walking distances at this stop

Babbitt quadrangle

Description:

Submembers G (wavy-bedded taconite ore), H (wavy-bedded taconite ore), and I (stromatolite-bearing unit) at UTM: 578,533/5,278,414N

The I submember is present at the base of the exposure and is overlain by the wavy-bedded H submember. At the top of H is a 0.5- to 1.0-foot-thick intraformational conglomerate that separates the H and G subunits (the G submember is actually present within the Lower Slaty member according to Gundersen and Schwartz, 1962). Note that both H and G subunits constitute taconite ore and both are characterized by wavy bedding.

Submember F with small septaria-like structures at UTM: 578,857E/5,278,440N

The F submember is thin- to wavy-bedded and locally contains small septaria-like structures that consist of whitish quartz-filled subvertical fractures in the granular cherty layers. Even though the F submember contains appreciable magnetite, it is classed as waste material because the magnetite is too fine to be economically concentrated.

Submember F with small septaria-like structures and submember G with minor garnets (optional) at UTM: 578,841/5,278,503N

Submember C (with Keweenawan sill) and submember D (optional) at UTM: 578,730E/5,278,206N

Both the C and D submembers are thin-bedded units of the Upper Slaty member; however, the D submember is different in that it contains slightly thicker beds and lenses of chert. The "contact" between the C and D submembers is exposed on this bench. A 20-foot-thick Keweenawan sill is also present at this stop (see below for geologic description) and is positioned approximately in the middle of the C submember.

STOP 11-9

Submembers A, B, and C within the Upper Slaty member, partially-melted Virginia Formation, two Keweenawan sills, and base of the Duluth Complex

Location: T. 60 N., R. 12 W., sec. 16, W 1/2, SW; Peter Mitchell Mine. Several subunits of the Upper Slaty member, the Virginia Formation, and three Keweenawan (1,100 Ma) intrusives can be viewed within short walking distances at this stop (see descriptions and locations listed below).

Babbitt quadrangle

Description:

Keweenawan Sill ("BIFSill") within the C submember at UTM: 584,051/5,280,958N

A 2- to 18-foot-thick sill is present in the middle of the C submember at the Peter Mitchell and Dunka Pit areas, and within the J submember in the LTV 2E pit. The sill is generally fine- to medium-grained with locally very coarse-grained plagioclase phenocrysts and vertical columnar jointing. A granoblastic texture is evident in thin-section, indicating that the sill was emplaced in the early Keweenawan and was later metamorphosed by intrusion of the Duluth Complex. Hauck and others (1997) noted that this sill is chemically similar to the Logan sills to the northeast in the Rove Formation, and have informally called this sill a "Logan-type" sill.

C submember at UTM: 584,129E/5,281,033N

The C submember is dominated by well-laminated, thin-bedded, slaty iron-formation containing magnetite, fayalite, ferrohypersthene, and chert.

Submembers A, B, and C at UTM: 584,193E/5,28,1142N

At the very top of the Biwabik Iron Formation is a 2- to 6-foot-thick chert and marble unit (A submember) that corresponds to the carbonate horizon that is present in only the eastern half of the Mesabi range. This unit is locally absent in some areas (non-depositional unconformity) and extremely thick in other areas. The B submember is characterized by alternating chert and diopside bands up to one foot thick; marble layers are locally present. In some areas at this stop, pink granophyric veins locally cut the B submember. These veins exhibit pinch-and-swell relationships in that the veins thicken within the diopside bands and pinch in the chert bands.

Keweenawan Sill ("VIRGSil") at the base of the Virginia Formation at UTM: 584,226E/5,281,159N

At the very base of the Virginia Formation is a 2- to 100-foot-thick sill that consists of a fine-grained, granoblastic rock with varying amounts of plagioclase, clinopyroxene, orthopyroxene, hornblende, olivine, and biotite. The informal term of "Cr-bearing sill" was first used by Hauck and others (1997) to highlight the relatively high chromium contents (600 to 1,200 parts per million) that are characteristic of this sill. This sill exhibits two varieties: 1. A fine-grained, massive, gray-colored unit (this exposure) that is extremely difficult to distinguish from the hornfelsed Virginia Formation, and 2. A medium- to coarse-grained unit that is olivine- and/or hornblende-rich and is easily recognized.

Partially-melted Virginia Formation in close proximity to the Duluth Complex at UTM: 584,261E/5,280,908N

In close proximity to the Duluth Complex, the well-bedded sediments of the Virginia Formation are typically transformed into a rock that at first appearance looks like an intrusive rock due to the presence of randomly oriented biotite. This rock is informally referred to as the "recrystallized unit," but is more properly classed as a diatexite (Sawyer, 1999). During emplacement of the Duluth Complex, the sediments of the Virginia Formation were heated, generating 20 to 40 percent pervasive partial melts, that enabled these rocks to literally flow in response to stresses that were applied during emplacement. All bedding planes are obliterated and what remains is a medium-grained recrystallized rock that contains plagioclase, cordierite, orthopyroxene, and randomly oriented biotite. Within this recrystallized matrix are blocks/boudins of more structurally competent siltstone and calc-silicate hornfels (originally limey layers).

Basal contact of the Duluth Complex at UTM: 584,266E/5,280,874N

At this locality the basal contact of the South Kawishiwi intrusion is irregular with localized "fingers" of the footwall Virginia Formation protruding upward into the intrusive rocks. Rocks of the South Kawishiwi intrusion consist of weakly to moderately mineralized, fine- to medium-grained, ophitic augite troctolite to olivine gabbro. Copper-nickel values are unknown for this exposure.

NEXT: Leave the Peter Mitchell Mine and drive west on Dunka Road back toward the general vicinity of Stop 11-7. The next stop is located immediately east of the Siphon fault.

STOP 11-10

No hammering please!

Virginia Formation near Siphon fault, Cliffs-Erie Site

Location: T. 59 N., R. 14 W., sec. 26, SE, SE, NE

Allen quadrangle; UTM: 569,505E/5,271,610N

Description: This is the only natural exposure of the Virginia Formation on the Mesabi Iron Range. Unfortunately, it is only a few feet thick. A total of 1,443 feet of the formation is present in drill cores from holes drilled south of the range. Note the graded beds, mud chips, concretions, and loading at the bases of these beds. The bedding is near vertical in this location due to proximity to the north-trending Siphon fault—an inferred growth fault (Graber, 1993) wherein the iron-formation decreases in thickness to the east (across the fault) by about 100 feet.

NEXT: Drive west to Gilbert via Dunka Road and Highway 135. From the main street in Gilbert, drive north (uphill) for a few blocks on Wisconsin Avenue to the north side of the athletic field behind the Eveleth-Gilbert Junior High School.

STOP 11-11

Archean pillowed greenstone, Gilbert

Location: T. 58 N., R. 17 W., sec. 23, NW, SE, SW

Gilbert quadrangle; UTM: 539,820E/5,259,750N

Description: This exposure of pillowed and massive metabasalt is part of the Archean Mud Lake sequence, metamorphosed to low greenschist grade. Pillow shapes indicate stratigraphic tops to the northeast, which places this outcrop on the south side of a major D_1 structure known as the Mud Lake syncline (Jirsa and others, 1998; Jirsa and Boerboom, 2003). Note also the presence of a few local fractures filled with red jasper, likely deposited in depressions on the rock surface by overstepping of the Paleoproterozoic sea during deposition of the Biwabik Iron Formation. There is no Pokegama Formation at this locality.

NEXT: Drive to Eveleth on Highway 37 and then to Cloquet via Highway 53 and State Highway 33. Follow Highway 33 through Cloquet to Interstate 35. Head south for about 2 miles to exit 235. Turn left (east) on State Highway 210 and follow it for about 3 miles to the stop sign in Carlton. Continue straight on Highway 210 for 1 mile to a bridge over the St. Louis River. Cross the bridge and park in the lot on the left.

STOP 11-12

Thomson Formation, equivalent of the Virginia Formation, at Thomson Dam

Location: T. 48 N., R. 16 W., sec. 5, SW, SW

Cloquet quadrangle; UTM: 545,610E/5,168,100N

Description: The Virginia and Thomson Formations (and the Rove Formation to the northeast in Minnesota and Ontario and the Michigamme Formation in Wisconsin and Michigan) were deposited in the Animikie basin. See Figures 11.4 and 11.5 for the regional perspective of deposition in a foreland basin to the north of the Penokean fold-and-thrust belt. See Morey and Ojakangas (1970) for details on the sedimentology of the Thomson Formation.

The Thomson Formation is best exposed at this type locality in the valley of the St. Louis River (Fig. 11.18). Jay Cooke State Park encompasses most of the exposures along 8 miles of the river valley. The rapids and gorges make this a world-class kayaking locality. Native Americans and later the voyageurs knew this stretch of river well, for it necessitated a 7-mile portage ("The Grand Portage of the St. Louis") that took 3 days if the travelers were laden with furs and supplies.

The original graywacke, siltstone, and mudstone beds were metamorphosed under lower greenschist conditions and deformed into broad and open folds (Fig. 11.18) with subvertical cleavage and subhorizontal lineations during the Penokean orogeny (circa 1,850 Ma?) as described by Holst (1984) and others. Sedimentary structures, including small-scale cross bedding, loading features (soft-sediment deformation), and sole marks, are characteristic of deposition by turbidity currents. Also present are abundant diagenetic concretions rotated into the plane of cleavage.

The cross bedding in siltstone beds and in the upper parts of graded graywacke beds indicates flow toward the south. This is interpreted as the result of paleocurrent flow down a southerly sloping paleoslope. In contrast, the sole marks indicate a dominant east–west trend of longitudinal flow, perhaps along the axis of the basin (Morey and Ojakangas, 1970).

The mafic dikes visible here are part of a major swarm of northeast-trending dikes that were feeders to overlying Keweenawan lava flows (part of the 1,100 Ma North Shore Volcanic Group) that have long since been eroded away (Green, 1972). The nearest flows today are present about 8 miles to the east of this locality. A vertical dike about 10 feet wide on the east bank of the channel north of the bridge forms a slight topographic low. Excellent horizontal cooling columns are present in this dike. At low-water stage, a dike 1 to 2 feet wide with excellent chilled glassy margins is present about midway between the highway bridge and the dam on the highest rock exposures.

When the water level is low, excellent exposures of the graywacke, siltstone, and mudstone (slate) are also found between the bridge and the dam. Note that layers of aligned concretions give the orientation of the bedding in the rather massive slate. A few sole marks are present in the artificial channel that crosses the main exposure between the bridge and the dam. At high-water stage, the best available exposure is the high road cut just west of the bridge. A thick quartz pod is present just north of the bridge on its western end.

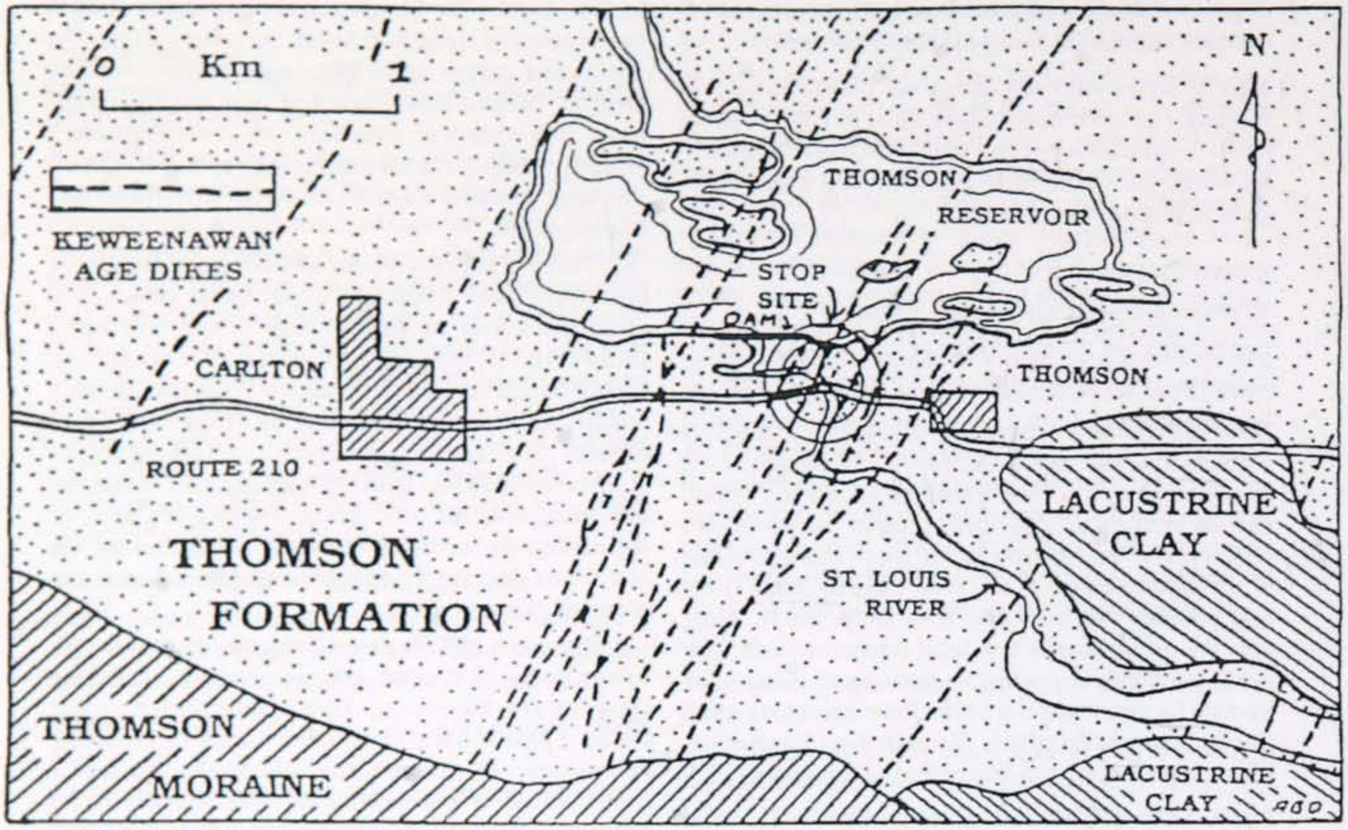


Figure 11.18. Geologic map of Thomson Dam area (from Wright and others, 1970).

To the south of the highway, opposite the parking lot, is a short path that leads to a viewpoint from which to observe the anticline illustrated in Figure 11.19. One can continue southward from this viewpoint, by walking in the pine trees along the top of the river channel for a few hundred feet, finally reaching the old railroad trestle that is now part of a long bicycle route that extends from Duluth to Hinckley. Before reaching the trestle, a large pothole is observable on the south end of an island across the main river channel. To the east of the trestle is a high railroad cut in slate, cut by a 3-foot-wide mafic dike. Looking southward from the trestle, one can see the 19-foot-thick, black, muddy Otter Creek member, a marker unit in the Thomson Formation utilized to ascertain that the total exposed thickness of the section in this area is only about 3,000 feet (Wright and others, 1970).

About 100 feet west of the trestle is a path on the right side of the bicycle route that leads back to the west end of the highway bridge, with the part of the path closest to the highway passing through a topographic notch that was formed by the erosion of the same thick mafic dike that is exposed north of the bridge.

NEXT: Return from Stop 11-12 to Carlton on Highway 210 and continue west almost to Interstate 35. A few hundred feet before the I-35 overpass, turn south on Carlton County Road 61. Follow this for about 2.8 miles, turn left on Gillogly Road, and follow it for 0.6 mile to a low outcrop on the left (east) side of the road.

STOP 11-13

Multiple deformation in graded graywacke beds of the Mille Lacs Group, indicating the existence of northward-directed nappes during the Penokean orogeny

Location: T. 48 N, R. 17 W., sec. 21, SW, NW
Iverson quadrangle; UTM: 537,370E/5,164,070N

Description: This location is a few miles south of the Thomson Formation of the previous stop. The Thomson Formation is situated in the "northern structural terrane" that displays foliation evidence for only one period of folding. This locality is

Figure 11.19. Anticline at Thomson Dam site. Drawn by Wendell Wilson, 1968.

in the "southern structural terrane" and displays evidence for two periods of folding and foliation development.

Holst (1984) argued that the earlier period of folding in the southern terrane also involved the emplacement of nappes. The evidence he cited for northward-directed nappes in the southern terrane included lithologic differences between the two terranes, with higher metamorphic grades to the south toward the main fold-and-thrust belt; also see Morey and Southwick (1984) and Southwick and others (1988). Other evidence includes the pervasive nature of the S_1 foliation, as shown in this exposure. Facing directions of F_2 folds also indicate that a very large area of the southern terrane is on the upper limb of an F_1 fold (Fig. 11.20). The refraction pattern of the S_1 foliation pattern in the region that includes this outcrop also suggests the existence of northward-directed nappes in the southern terrane, as explained below.

Several graded beds at this locality strike east–west and dip to the north. In the fine-grained tops of these beds, a cleavage (S_1) can be observed. It is very gently folded, with horizontal axes trending east–west. If the cleavage is traced toward the bottom of a bed, it is seen to change its orientation markedly, and it becomes a spaced cleavage that dips moderately toward the south. A crenulation cleavage (S_2), vertical or very steeply dipping to the south, can be seen in the upper portion of the beds. A line drawing of these relationships is shown in Figure 11.21. This S_2 cleavage has the same orientation as the cleavage in the Thomson Formation to the north, and is interpreted as an equivalent cleavage formed during the later stages of the Penokean orogeny.

NEXT: Backtrack to Highway 210 and I-35. Drive south on I-35 to Minneapolis.

END OF TRIP

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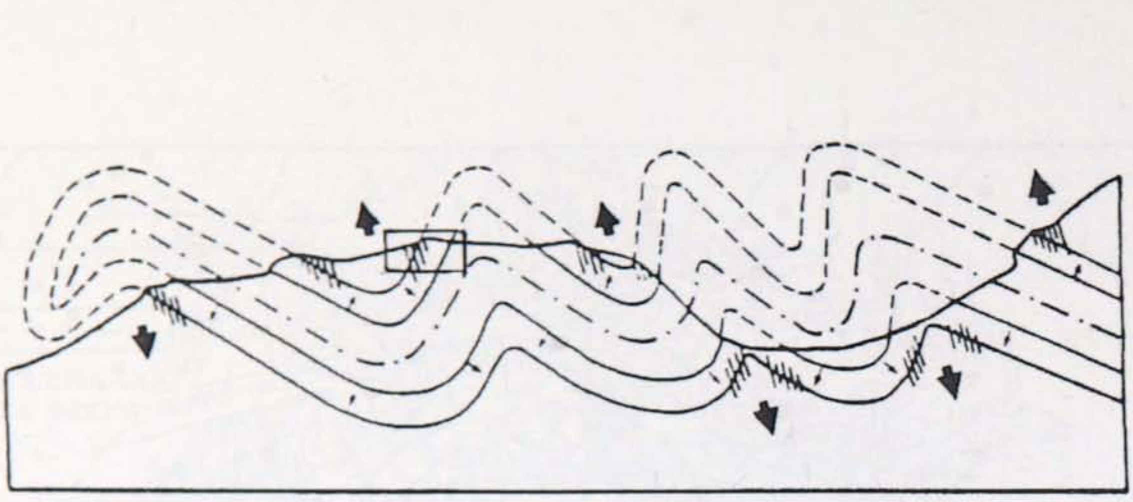

Figure 11.20. Illustration of facing direction of F_2 folds (after Borradaile, 1976). The dashed-dot line is the axial plane of an F_1 fold. Foliation shown diagrammatically is S_2 axial planar crenulation cleavage. Small arrows indicate stratigraphic tops, large arrows indicate facing directions of F_2 folds. Box indicates interpreted position of outcrop at Stop 11-13.

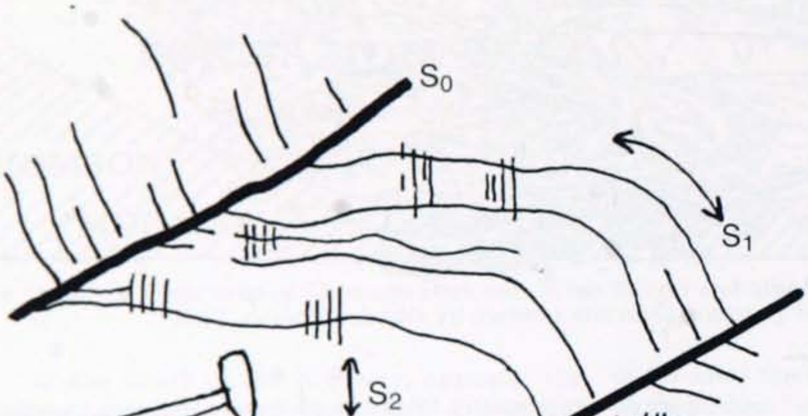

Figure 11.21. Line drawing (modified from Holst, 1992) of geometrical relations of bedding (S_0), early foliation (S_1), and crenulation cleavage (S_2) at the Gillogly Road outcrop (Stop 11-13). Hammer handle is 40 centimeters long.

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FIELD TRIP 12

Saturday, May 21 – Sunday, May 22

PRE-WISCONSINAN AND WISCONSINAN GLACIAL STRATIGRAPHY, HISTORY, AND LANDSCAPE EVOLUTION, WESTERN WISCONSIN

Leaders

Kent M. Syverson, University of Wisconsin–Eau Claire
Robert W. Baker, University of Wisconsin–River Falls
Steven Kostka, University of Wisconsin–Madison
Mark D. Johnson, Göteborg University, Sweden

INTRODUCTION

Evidence for several glaciations and different styles of landform development is observed in western Wisconsin. Eroded and weathered till outcrops provide clues to glacial activity prior to the Wisconsinan glaciation. Relatively unmodified till sheets and landforms document events that occurred during the Wisconsinan glaciation. During the past thirty years, much research has been conducted on the glacial stratigraphy and geomorphic evolution of western Wisconsin. The goal of this field trip is to revisit glacial lithostratigraphic units that have long been recognized in western Wisconsin (Mickelson and others, 1984; Attig and others, 1988) and re-evaluate them in the light of more recent discoveries. In addition, changes in landform assemblages will be used to evaluate different processes that were operating at the southern margin of the Laurentide Ice Sheet.

BEDROCK GEOLOGY OF WESTERN WISCONSIN

Pleistocene sediments in western Wisconsin cover a deeply incised bedrock surface with more than 150 meters of relief in places. The oldest Precambrian rock units are exposed in the Chippewa River valley. These rock units include 1,850-million-year-old gneiss, schist, amphibolite, and breccia that have been deformed and intruded by granitic dikes (Myers and others, 1980; Holm and others, 1998a). These are unconformably overlain by the Flambeau and Barron Quartzites that form resistant highlands in Chippewa, Barron, and Rusk Counties. These highlands influenced Pleistocene ice flow and glacial landform development. The Flambeau and Barron Quartzites were deposited in a shallow sea or in braided streams during the "Baraboo Interval" between 1,750 and 1,630 Ma (Holm and others, 1998b; Medaris and Dott, 2001) and metamorphosed between 1,650 and 1,630 Ma (Holm and others, 1998b; Romano and others, 2000). The Midcontinent rift system formed approximately 1,100 Ma, and much basalt was extruded along the rift axis that roughly coincides with the St. Croix River valley along the Minnesota–Wisconsin border (Van Schmus and others, 1982).

Precambrian rock units in western Wisconsin are unconformably overlain by Paleozoic rock. These Paleozoic rock units represent multiple fluctuations in sea level on the North American continent during the earliest part of the Paleozoic era. Sandstone and shaly sandstone dominate the Cambrian strata (for example the Mt. Simon, Eau Claire, Wonewoc, and Jordan Formations; Havholm, 1998), and to the west, these are overlain by dolomitic Ordovician rock units (for example the Ancel and Prairie du Chien Groups; Mudrey and others, 1987; Brown, 1988). Dendritic stream valleys deeply dissect the entire Paleozoic section.

PLEISTOCENE GEOLOGY

Large continental glaciers advanced across western Wisconsin numerous times during the Pleistocene epoch, and perhaps as early as the latest Pliocene (Attig and Muldoon, 1989). Although Wisconsin lies well north of the maximum extent of Quaternary glaciation, the Driftless Area of southwestern Wisconsin remained unglaciated even though areas to the south were glaciated several times (Fig. 12.1; Chamberlin and Salisbury, 1885; Hobbs, 1999; Cutler and others, 2001). The oldest glaciations occurred more than 130,000 years ago, long before

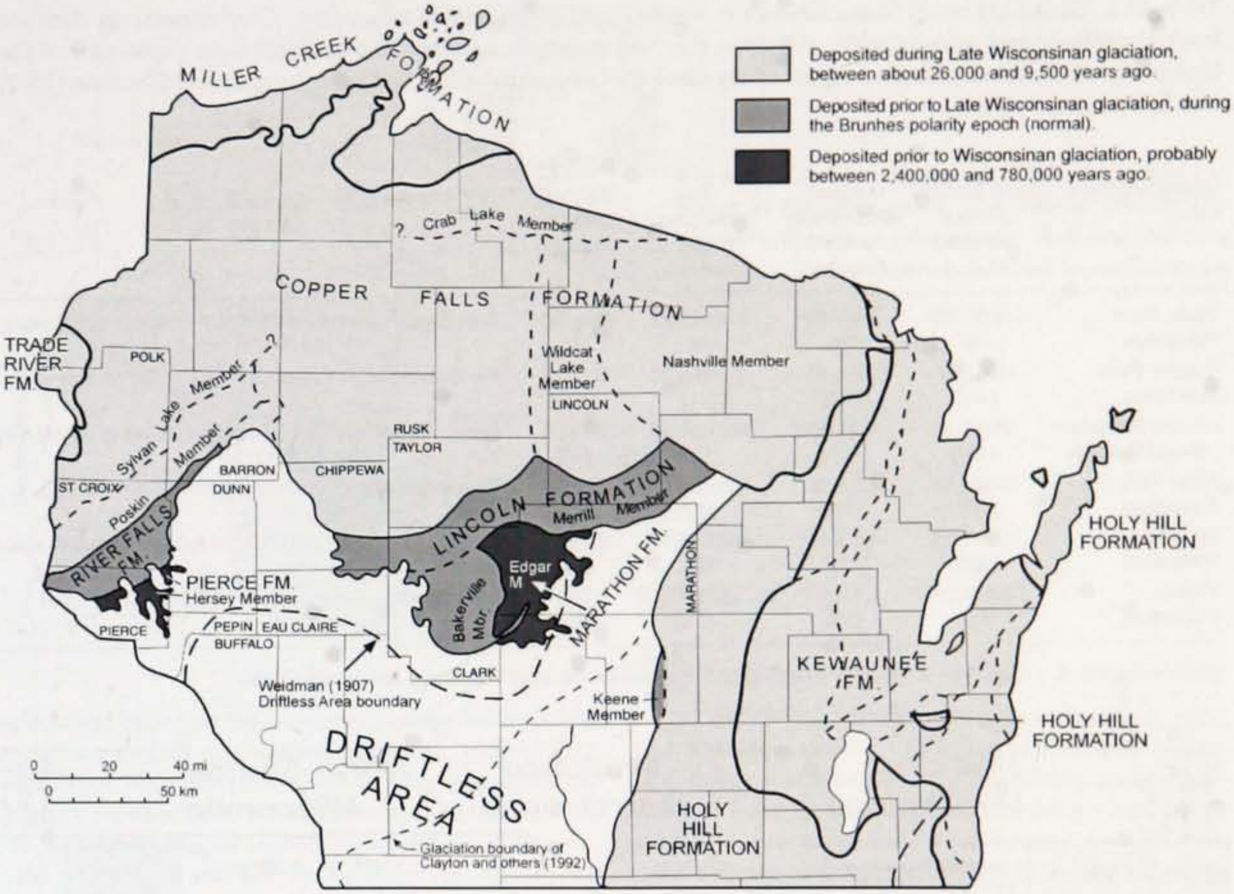


Figure 12.1. Pleistocene lithostratigraphic units for till in northern Wisconsin. Short dashed lines in the glaciated areas designate member boundaries; modified from Clayton and others (1992) using information from Syverson and Colgan (2004).

the Late Wisconsinan glaciation, and several till units were deposited during these glaciations (Fig. 12.2; Table 12.1). Glaciers last advanced into Wisconsin during the Late Wisconsinan glaciation (26,000 to 9,500 years ago; Fig. 12.1). All ages in this paper are reported in uncalibrated radiocarbon years unless stated otherwise. Late Wisconsinan glacial deposits still exhibit glacial landforms and continuous till sheets that aid interpretations.

Characteristics of glacial sediment in western Wisconsin have been influenced greatly by the rock units over which the glacier flowed. Ice from the Keewatin ice dome to the northwest (including the Des Moines lobe; Fig. 12.3) deposited silt-rich, calcareous tills. Ice from the Labradoran ice dome to the northeast flowed out of the Superior lowland (including the Superior and Chippewa lobes; Fig. 12.3) and deposited reddish-brown, sandy tills containing Precambrian basalt, banded iron-formation, and red sandstone.

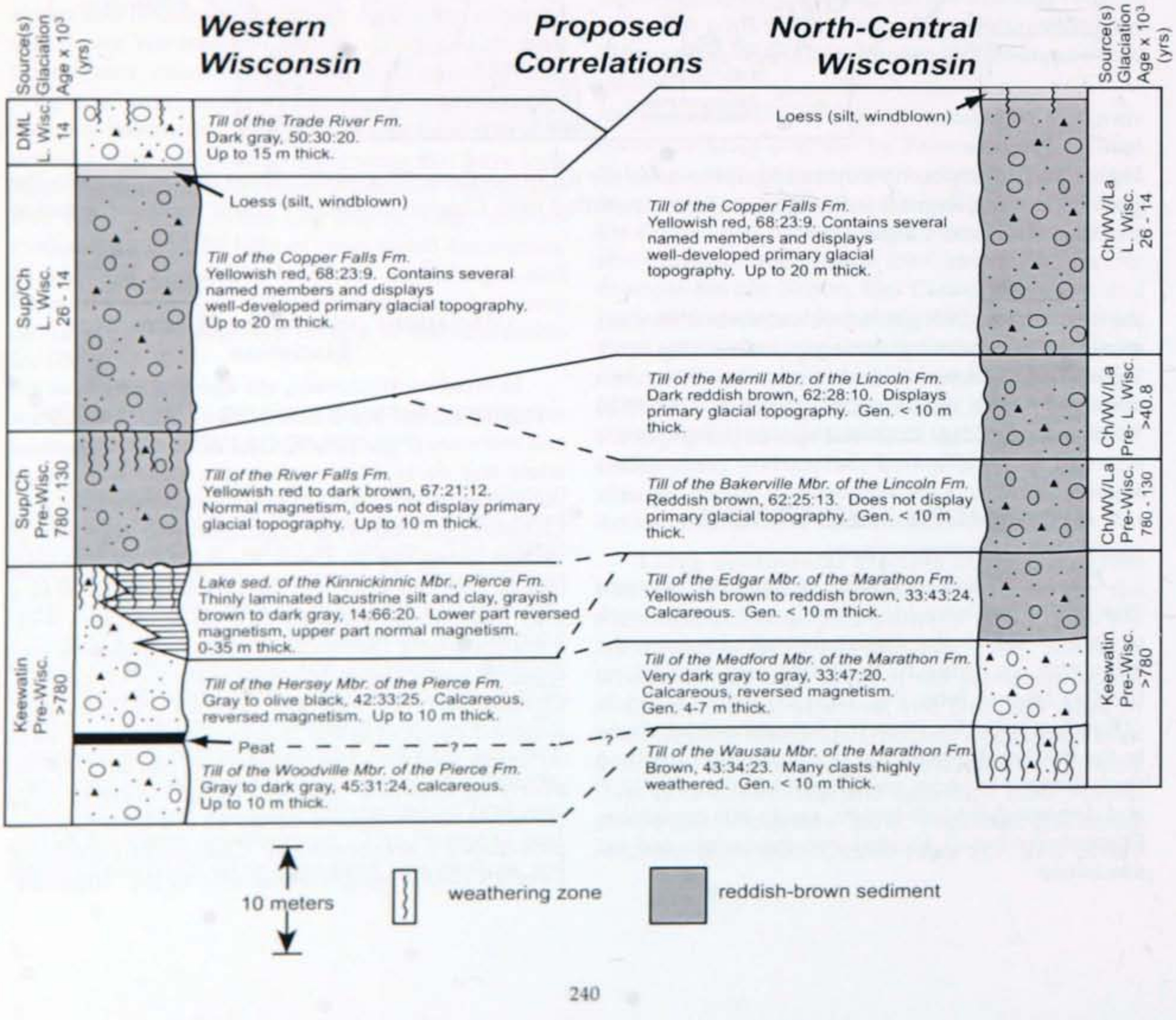
Glaciations prior to the Wisconsinan glaciation

In western Wisconsin, the age and origin of till units that extend south beyond the Late Wisconsinan end moraines (Figs. 12.1, 12.4) have been controversial since the early 1880s when the first work was published on this area. In northwest Wisconsin, R.T. Chamberlin (1905, 1910) described two till units in valleys along the St. Croix River. The oldest unit (called the "old gray" till by Leverett, 1932) was a gray, calcareous, and clay-rich till associated with a northwesterly (Keewatin) source. A paleosol and erosional surface were described in the top of this unit. Chamberlin (1910) and later Leverett (1932) initially assigned this unit to the "Kansan" glaciation (a term no longer used for a pre-Wisconsinan glaciation). The overlying unit (the "old red" till of Leverett, 1932) contained reddish-brown, sandy till of a Lake Superior provenance (Labradoran ice). Chamberlin (1910) and Leverett (1932) assigned this unit to the "Illinoian"

Table 12.1. Summary of till characteristics in western and north-central Wisconsin. Clay mineralogy data are from Thornburg and others (2000). All grain size and magnetic susceptibility analyses were performed at the University of Wisconsin–Madison Quaternary Geology Laboratory. Modified from Treague and Syverson (2002) and Syverson (in press).

Lithostratigraphic unit	Sand:Silt:Clay percent (number)	Magnetic susceptibility SI units (number)	Clay mineralogy[1] K:I:S:V % (number)	V:K ratio ±std dev (number)	Munsell color
Trade River Formation	49:31:20 (34)	2.0×10^{-3} (34)	6:43:39:12 (4)	2.0±1.0 (4)	Dark brown to yellow-brown (7.5YR 4/4 to 10YR 4-5/4)
Copper Falls Formation	68:23:9 (485)	2.9×10^{-3} (446)	6:38:35:21 (11)	5.1±3.2 (11)	Reddish-brown to brown (5YR 4/4 to 7.5YR 4/4)
Lincoln Formation Merrill Member	55:31:14 (151)	2.3×10^{-3} (135)	6:31:38:25 (13)	4.7±1.9 (13)	Reddish-brown to strong brown (5YR 4/3 to 7.5YR 4/6)
River Falls Formation	67:21:12 (88)	1.5×10^{-3} (75)	17:29:36:18 (27)	1.3±1.1 (27)	Reddish-brown to yellow-red (5YR 4/4-6)
Marathon Formation	40:41:19 (359)	1.5×10^{-3} (352)	8:21:54:17 (9)	2.3±1.0 (9)	Yellow-brown to gray (10YR 5/6 to 2.5Y 5/1) and reddish-brown (5YR 4/4)
Pierce Formation	39:37:24 (41)	1.1×10^{-3} (38)	23:25:40:12 (13)	0.5±0.1 (13)	Dark gray to yellow-brown (10YR 4/1 to 10YR 5/4)

[1]Clay minerals: K = kaolinite, I = illite, S = smectite, and V = vermiculite, from Thornburg and others (2000)



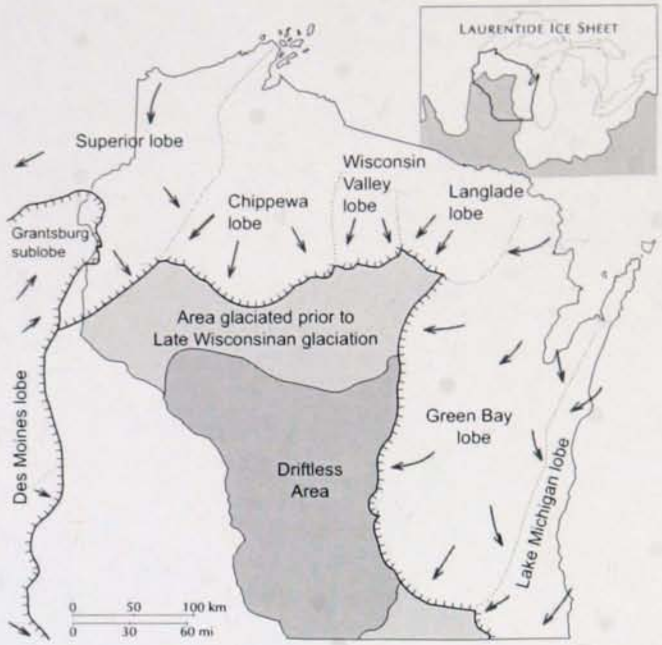


Figure 12.3. Ice lobes of the Laurentide Ice Sheet at the maximum flow extent during the Late Wisconsinan glaciation. Arrows indicate direction of ice flow; modified from Johnson (2000).

glaciation based on the erosional surface below it and a similar degree of weathering on the "old red" till and other "Illinoian" tills in Iowa. Disputes about the history represented by the different units have arisen because most glacial landforms have been removed, and weathering and erosion have extensively altered the surficial expression of the resulting till units. These till units will be examined on day 1 of the field trip and are summarized below.

Deposition of the Pierce Formation

Till of the Pierce Formation is the oldest glacial unit in western Wisconsin (Figs. 12.1, 12.2; Baker and others, 1983; Mickelson and others, 1984; Johnson, 1986; Attig and others, 1988). Pierce Formation tills probably represent ice advances from a Keewatin source during at least two phases of glaciation (Baker and others, 1983; Mickelson and others, 1984; Johnson, 1986; Attig and others, 1988). This ice crossed Cretaceous shale and deeply weathered crystalline rock, both sources of the clay mineral kaolinite (Morey and Setterholm, 1997), as well as Paleozoic limestone. Thus, till of the Pierce Formation is gray to brown, calcareous (where unleached), silty, and has an elevated kaolinite percentage (Table 12.1; Baker and others, 1983; Thornburg and others, 2000). Hinke (2003) reported laboratory hydraulic conductivity values for Pierce Formation till in St. Croix County ranging from 10^{-6} to 10^{-8} centimeters per second. Pierce Formation till is deeply weathered, yellowish brown (10YR 5/4) where oxidized, with soil profiles up to 2.9 meters thick and leaching depths of 3.5 meters (Mickelson and others, 1984). Till of the Woodville Member of the Pierce Formation marks the first ice advance recognized (Fig. 12.2). Lake sediment of the Eau Galle Member of the Pierce Formation is present below the Woodville till, and peat and wood are found above the Woodville till at the type section (Attig and others, 1988, p. 8, 11).

Keewatin ice flowed southeast across the Mississippi River during the later Reeve Phase, deposited till of the Hersey Member of the Pierce Formation above the Woodville till in western Wisconsin, and dammed the major southwesterly flowing tributaries of the Mississippi River such as the Chippewa and Red Cedar Rivers (Fig. 12.5;

Figure 12.2. Glacial lithostratigraphy in western and north-central Wisconsin. Vertical scale is only approximate, and mean grain size is reported as sand:silt:clay percentages. Sediment deposited by the following lobes (or from the following regions): Ch, Chippewa lobe; DML, Des Moines lobe; Keewatin, derived from the Keewatin ice dome to the northwest; La, Langlade lobe; Sup, Superior lobe; WV, Wisconsin Valley lobe. Proposed ages and correlations for glacial units are also shown. Ages for Wisconsinan units are in ^{14}C yr B.P., and ages from earlier events are in calendar years. Modified from Syverson and Colgan (2004).

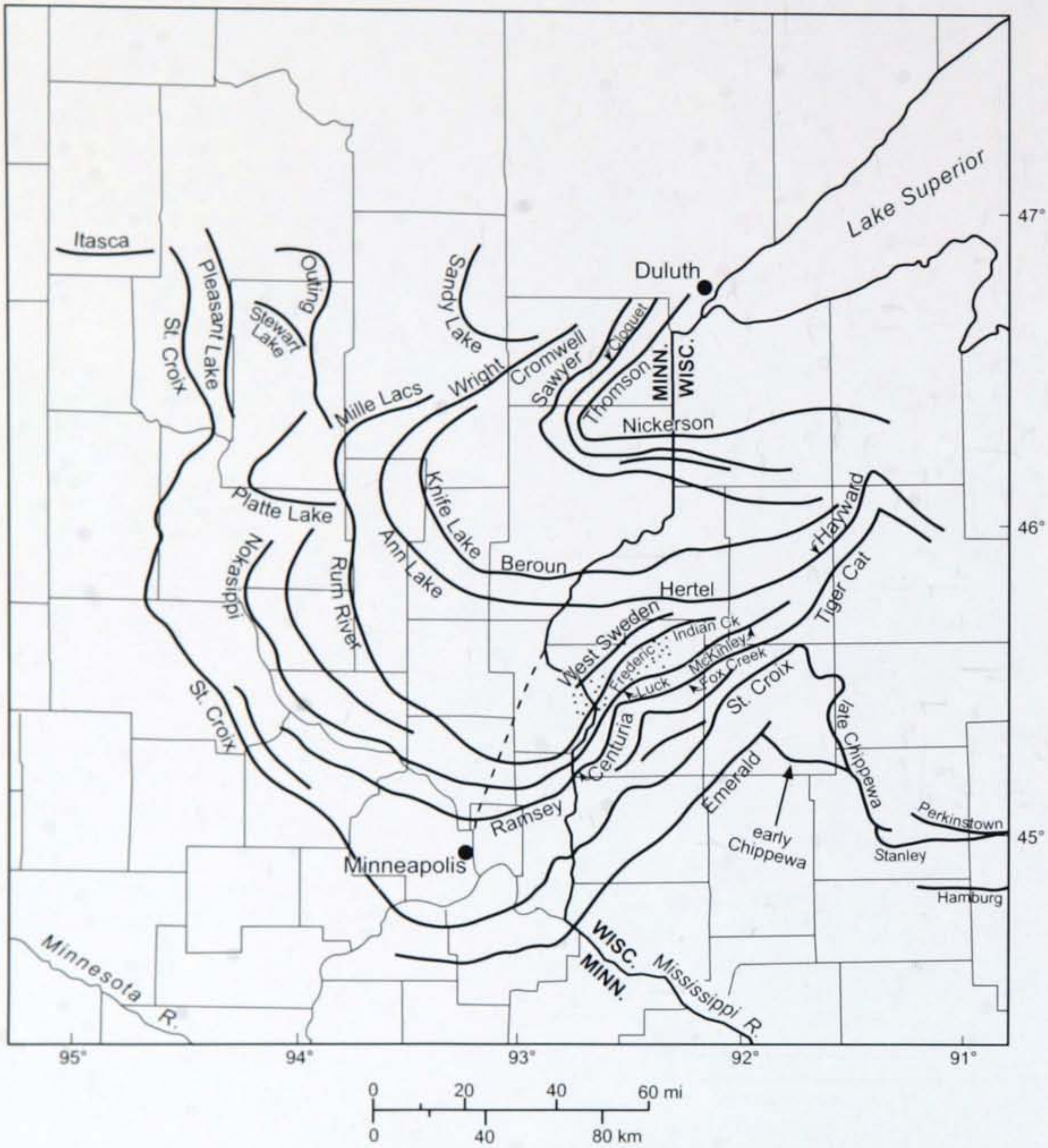


Figure 12.4. Hypothetical correlation of Late Wisconsinan ice margins of the Superior lobe in Wisconsin and Minnesota and a portion of the Chippewa lobe. The dashed line represents the former path of the St. Croix River prior to Late Wisconsinan glaciation. The stippled area is an outcrop of basalt hills in Polk County, and one can see that this high area influenced the shape of the southern margin of the Superior lobe. Note also that the retreat rate was much higher in the west than in the east. Modified from Johnson and Mooers (1998) using information from Syverson and Colgan (2004).

Baker and others, 1983; Johnson, 1986). The maximum easterly extent of this ice is poorly constrained (to be discussed later). As the ice margin retreated westward from its maximum position, the resulting lakes lengthened and formed an interconnecting network that covered more than 6,000 square kilometers and extended at least tens of kilometers east of the modern Mississippi River valley. More than 150 meters of thinly laminated glaciolacustrine sediment, the Kinnickinnic Member of the Pierce

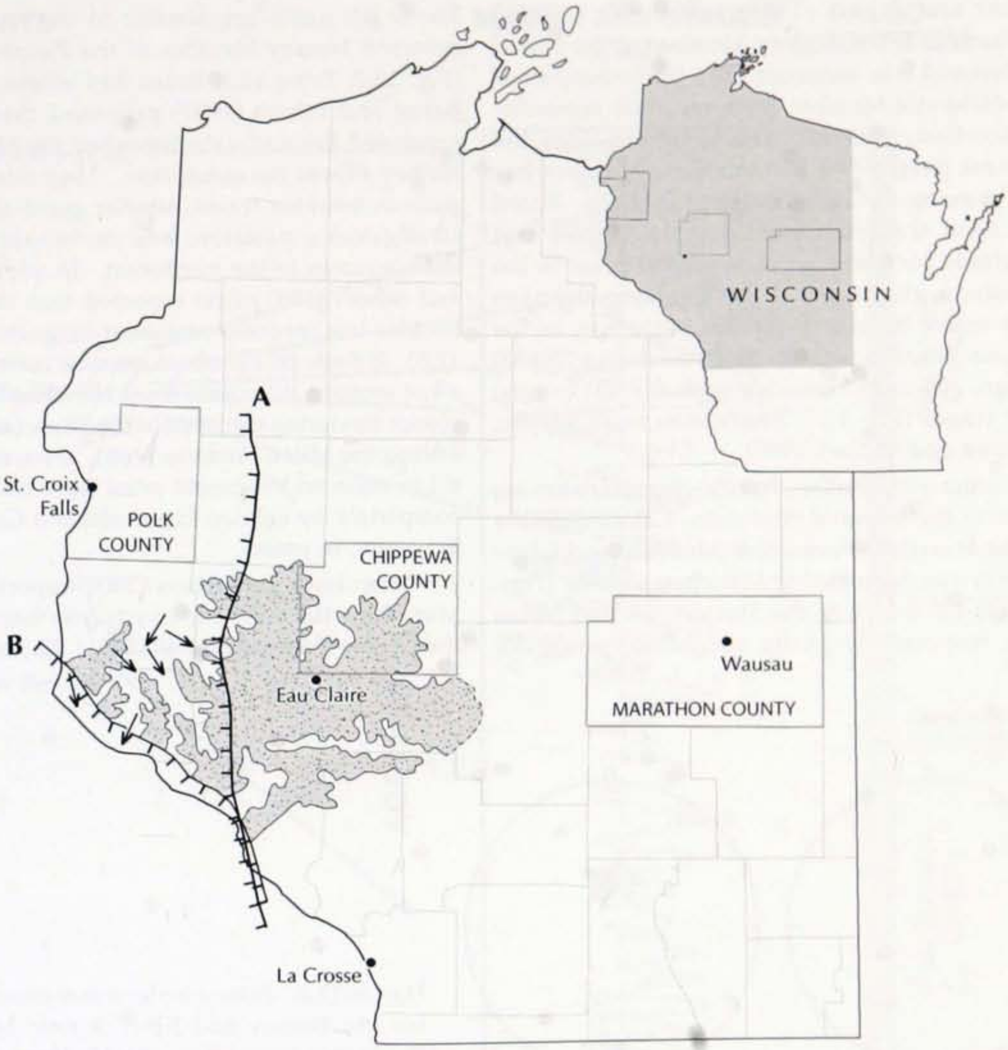


Figure 12.5. Ice-margin positions (hachured lines), ice-flow directions from pebble fabrics (arrows), and ice-dammed lakes (stippled areas) associated with Keewatin ice flowing from the northwest during the Reeve Phase (modified from Baker and others, 1983, and Johnson, 2000). The Kinnickinnic Member of the Pierce Formation was deposited in the ice-dammed lakes. Pebble-fabric and lake-distribution data are from Baker and others (1983); ice margin A is from Baker and others (1983) and Johnson (1986). Ice margin B may represent the approximate location of the ice margin when much of the Kinnickinnic Member was deposited (Baker and others, 1983). Prior to deposition of the Kinnickinnic Member, this ice may have extended as far east as Marathon County (Baker and others, 1987; Syverson and Colgan, 2004).

Formation, accumulated in these lakes in places (Fig. 12.2). In Chippewa County the Kinnickinnic Member is up to 35 meters thick. The Kinnickinnic Member contains thinly laminated, calcareous, dark gray (10YR 4/1) silt and clay (mean sand:silt:clay percentages 14:66:20) with soil profile development generally less than 2 meters thick (Baker and others, 1983). Kinnickinnic Member lake sediment interfingers with and also overlies the Hersey Member. Based on counting varves, Baker (1984a) estimated that the lakes existed for more than 1,200 years. This low-hydraulic-conductivity material in western Wisconsin is an important aquitard and has been a source of clayey material for bricks and landfill liners (Gunderson and Syverson, 1994).

Baker and others (1983) and Baker (1984b) stated that till of the Hersey Member of the Pierce Formation and lake sediment from the lower part of the Kinnickinnic Member have reversed remanent magnetization, whereas lake sediment from the uppermost part of the Kinnickinnic Member has normal remanent magnetization (Fig. 12.6). Based on this data, they suggested that the Hersey and Kinnickinnic Members were deposited prior to the Wisconsinan glaciation at the Emperor–Brunhes polarity epoch boundary 460,000 years ago, or the Matuyama–Brunhes polarity epoch boundary 780,000 years ago, and could coincide with several oxygen isotope stages (Fig. 12.7; Shackleton and Opdyke, 1973; Lowe and Walker, 1997).

Till units with similar physical characteristics are observed in north-central Wisconsin. Calcareous, silty till of the Marathon Formation (Medford and Edgar Members) was deposited in Marathon County (Figs. 12.1, 12.2, 12.5) during the Stetsonville and Milan Phases, respectively (Attig and Muldoon, 1989). These till units are similar to the magnetically reversed Hersey Member of the Pierce Formation (Fig. 12.2; Table 12.1; Baker and others, 1983), and Baker and others (1987) proposed that markedly expanded Keewatin ice deposited the Medford and Hersey tills at the same time. They cited evidence such as boulder trains, similar grain size, similar stratigraphic position, and carbonate and black shale sources to the northwest. In addition, Baker and others (1987) first reported that the Medford till also has reversed remanent magnetization (Fig. 12.6). If these till members are time correlative, then all of western Wisconsin must have been covered by glacier ice during the Stetsonville Phase (and probably during the Milan Phase as well). If so, much of this till in western Wisconsin must have been removed completely by erosion (Syverson and Colgan, 2004; Syverson, in press).

Thornburg and others (2000) reported that the Marathon till contains much less kaolinite than the Pierce Formation till (Table 12.1). Syverson

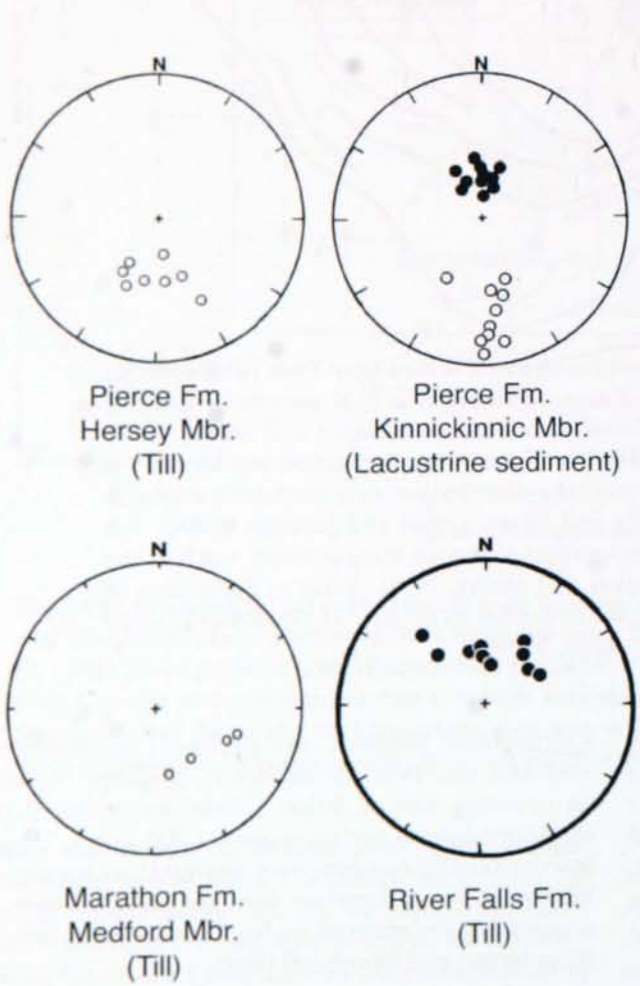


Figure 12.6. Paleomagnetic remanent directions for the Hersey and Kinnickinnic Members of the Pierce Formation, the Medford Member of the Marathon Formation, and the River Falls Formation. Projection is equal area. Solid dots represent lower hemisphere projections (normal magnetic polarity) and the open dots represent upper hemisphere projections (reversed polarity). Hersey and Medford Member tills display reversed polarity, and Kinnickinnic Member lake sediment contains a reversed to normal magnetic sequence; modified from Baker and others (1983).

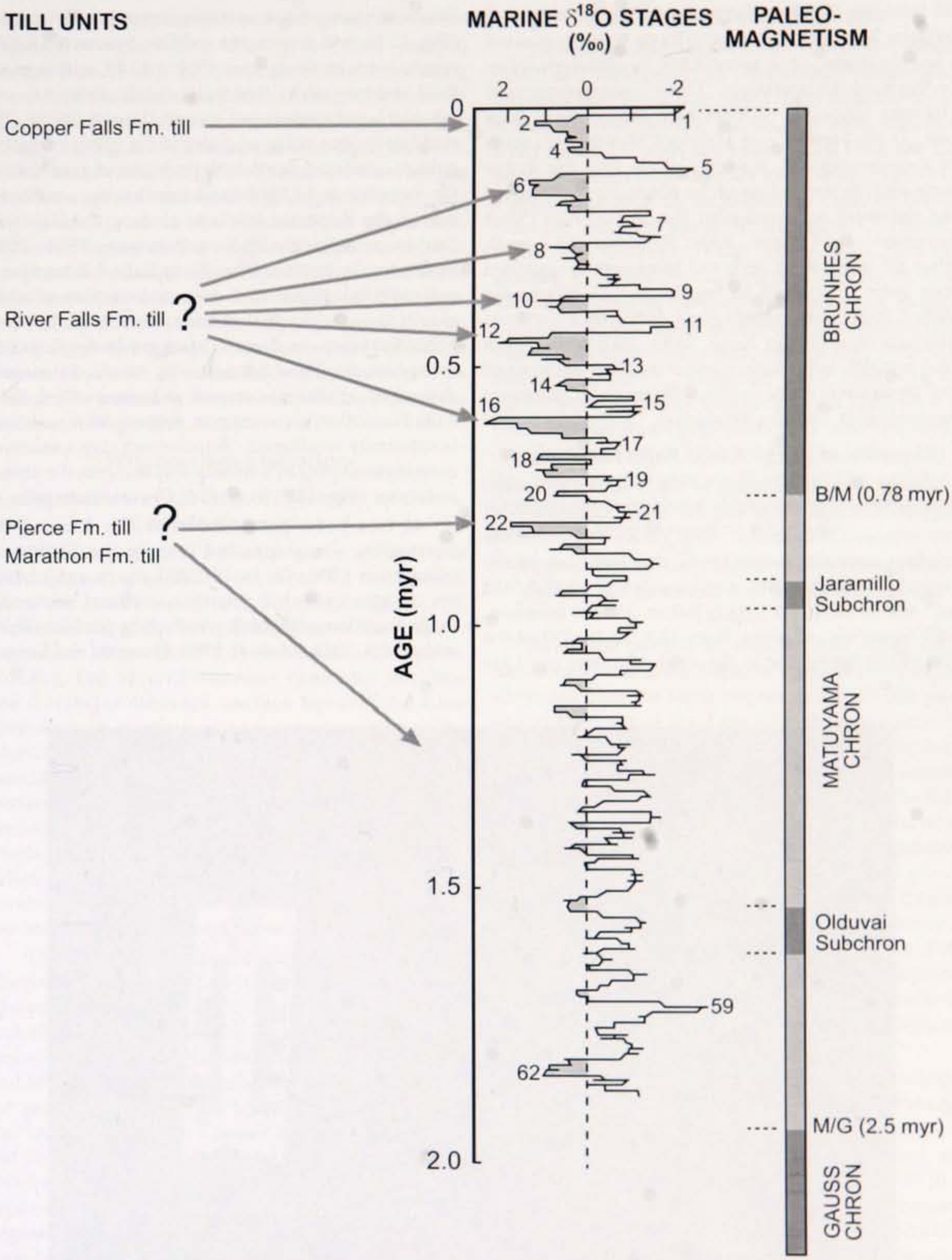


Figure 12.7. Oxygen isotope curve showing stages associated with glaciations (even numbers, shaded), interglacials (odd numbers), and the associated paleomagnetic chrons. Pierce and Marathon Formation tills have reversed polarity and probably were deposited sometime prior to stage 20. The River Falls Formation till could have been deposited during any number of glaciations during the Brunhes Chron (normal polarity); modified from Rutter (1992).

and Johnson (2001) proposed that the Hersey and Medford Member tills might have been deposited at approximately the same time by different lobes flowing from the northwest. These lobes incorporated different amounts of kaolinite along their flow lines. If the Hersey and Medford Member till units are time-equivalent, it seems likely that the Reeve Phase and the deposition of the Kinnickinnic Member lake sediment occurred after the Stetsonville Phase (Syverson and Colgan, 2004; Syverson, in press). Other till units with reversed remanent magnetism have been described in south-central Wisconsin (Miller, 2000), Iowa (Boellstorff, 1978), and northern Missouri (Rovey and Kean, 1996, 2001), but much work remains to be done in order to better understand how those units relate to tills with reversed remanent magnetism in western Wisconsin.

Deposition of the River Falls Formation

Extensive weathering of the Pierce Formation suggests that much time elapsed before ice readvanced into western Wisconsin. The Pierce Formation is overlain unconformably by till and outwash of the River Falls Formation, the equivalent of the "old red" till of Leverett (1932; Baker, 1984b; Johnson, 1986; Syverson, in press; Figs. 12.1, 12.2). Till of the River Falls Formation is the most extensive pre-Late Wisconsinan surficial unit in westernmost Wisconsin (Fig. 12.1), and it contains reddish-brown (5YR 4/4) sandy loam to sandy clay (Table 12.1), with a mean field and laboratory hydraulic conductivity value of 6.5×10^{-3} centimeters per second (Hinke, 2003). The reddish-brown color and abundant clasts of basalt, gabbro, and red sandstone indicate deposition by the Superior and Chippewa lobes flowing southward out of the Superior lowland during the Baldwin, Dallas, and Foster Phases (Johnson, 1986, 2000; Syverson, in press). The River Falls till surface is extensively eroded and does not display original glacial topography (Baker and others, 1983; Johnson, 1986; Syverson, in press). Soil profile development extends to depths of 2.8 meters in the till. Proximally deposited, meltwater stream sediment of the River Falls Formation is common in western Wisconsin and is extremely weathered. Soil-derived clay sometimes extends to depths of 5 meters and cements the stream sediment (Fig. 12.8; Stop 12-6; Syverson, in press).

It has been proposed that the River Falls Formation was deposited during the "Illinoian" glaciation (300,000 to 130,000 years ago) based on its stratigraphic position, normal remanent magnetization, and thick weathering profiles (Baker and others, 1983; Johnson, 1986; Syverson and Colgan,

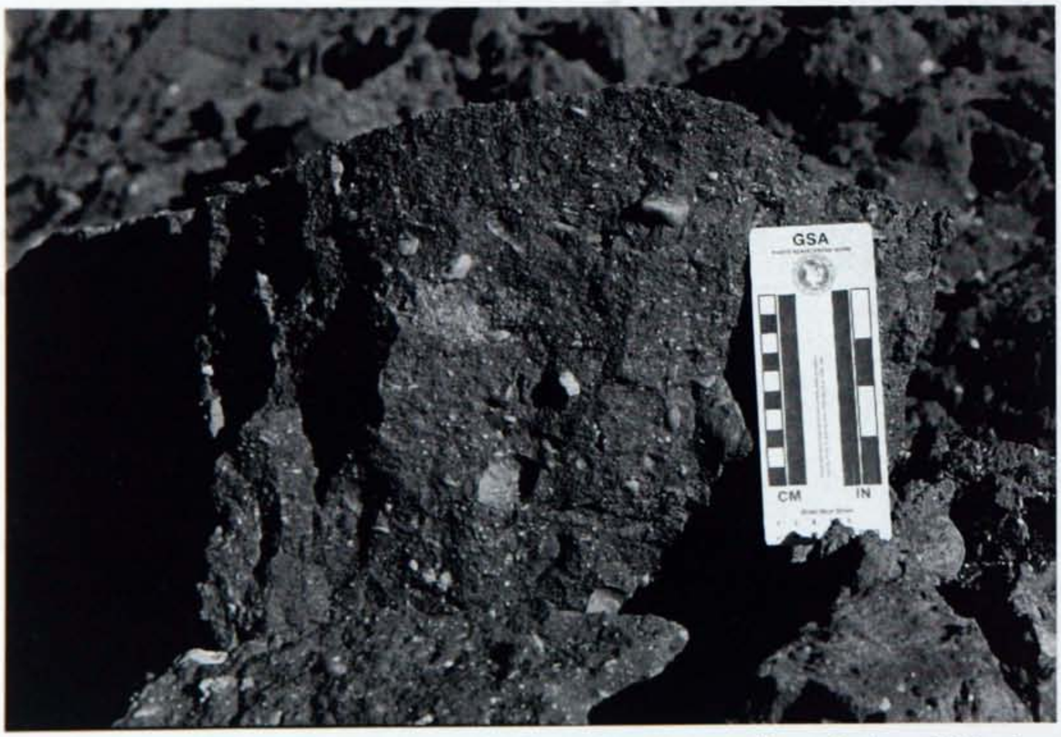


Figure 12.8. Clay cementation of River Falls Formation stream sediment in the soil B horizon at Stop 12-6. From Syverson (in press).

2004). However, marine oxygen isotope records indicate two glaciations during that time interval (stages 6 and 8; Fig. 12.7) and several others during the rest of the most recent normal magnetic polarity epoch, the Brunhes (even-numbered stages between 10 and 20; Fig. 12.7; Shackleton and Opdyke, 1973; Lowe and Walker, 1997). The thick weathering profiles seen on this field trip argue strongly for deposition prior to the Wisconsinan glaciation. However, based on the oxygen isotope curve, it is not possible to assign a more specific age to the River Falls Formation sediments with confidence (Syverson, in press). Johnson (1986) and Syverson and Colgan (2004) suggested that till of the River Falls Formation is correlative with till of the Bakerville Member of the Lincoln Formation in Marathon and Clark Counties to the east (Figs. 12.1, 12.2).

Wisconsinan glaciation

Deposition of the Merrill Member of the Lincoln Formation

The final glacial advance prior to the Late Wisconsinan glaciation occurred as ice flowed from the north out of the Superior region and deposited till of the Merrill Member of the Lincoln Formation (Figs. 12.1, 12.2). In eastern Chippewa County, the Merrill Member typically overlies the Cambrian bedrock surface beyond the Late Wisconsinan maximum ice-margin position. The Merrill Member contains reddish-brown (5YR 4/4), noncalcareous sandy loam till (Table 12.1). The till surface has little to moderate stream incision, lacks well-developed surficial weathering horizons, and displays some streamlined glacial landforms and low-relief, hummocky topography. The mean hydraulic conductivity value for the till is 2.2×10^{-4} centimeters per second (Muldoon and others, 1988).

The Merrill Member was deposited during the Hamburg Phase (Fig. 12.4; Attig and Muldoon, 1989). The presence of original glacial topography and the lack of extensive weathering suggest that the till was deposited during the Wisconsinan glaciation. Stewart and Mickelson (1976) presented clay mineral evidence for greater weathering of Merrill Member till than the Late Wisconsinan Copper Falls Formation till, but Thornburg and others (2000) could not reproduce this trend to the west. Stewart and Mickelson (1976) reported an age greater than 40,800 ^{14}C yr B.P. for organic material overlying till of the Merrill Member. Based on these considerations, it has been proposed that the Merrill Member was deposited during the early part of the Wisconsinan glaciation (Ham and Attig, 1997; Syverson and Colgan, 2004; Syverson, in press).

Late Wisconsinan glaciation

In western and north-central Wisconsin, Wooster (1882) and Weidman (1907) described younger glacial deposits that were associated with hummocky end moraines. Two young glacial tills were identified: a red sandy till with abundant clasts from the Lake Superior basin (Labradoran ice dome), overlain by a gray, calcareous till deposited by a glacier flowing from the west (the Keewatin ice dome; Chamberlin, 1905, 1910; Leverett, 1932). These units have been studied in detail by numerous scientists to verify the conclusions of the early geologists (for example Clayton, 1984; Mickelson and others, 1984; Johnson, 1986, 2000; Attig and others, 1988; Attig, 1993; Ham and Attig, 1997). Detailed field mapping has identified numerous glacial phases indicated by Late Wisconsinan end moraines and other ice-marginal landforms (Fig. 12.4). The remainder of this overview summarizes the current knowledge of Late Wisconsinan glacial stratigraphy, geomorphology, and chronology in western Wisconsin.

Deposition of the Copper Falls Formation during the Emerald and Early Chippewa Phases

After the period of erosion and weathering mentioned above, the Superior and Chippewa lobes advanced from the north approximately 26,000 years ago at the start of the Late Wisconsinan glaciation (Fig. 12.3; Black, 1976). Sediment from this advance is included in the Copper Falls Formation and consists of till, outwash, and lacustrine sediment. The till is reddish-brown, sandy loam, and contains abundant "black and red" lithologies—Keweenawan sandstone, rhyolite, and Keweenawan basalt as well as granite and gneiss (Table 12.1). In eastern Barron County, south of the Blue Hills, red quartzite is common in the till (Mickelson and others, 1984; Johnson, 1986, 2000; Attig and others, 1988). Hinke (2003) reported a mean field and laboratory hydraulic conductivity value for this till in St. Croix County of approximately 10^{-3} centimeters per second.

The St. Croix and Chippewa moraines of western Wisconsin do not represent the maximum ice-margin position during the Late Wisconsinan glaciation. It is clear based on geomorphology, mineralogy, and stratigraphy that these two lobes advanced 10 to 15 kilometers farther south prior to a retreat to the more prominent moraines. This more extensive position, first recognized by Mathiesen (1940), was reached during the Emerald and Early Chippewa Phases of the Superior and Chippewa lobes. At this time, ice extended to a point south of Barron and Polk Counties (Figs. 12.4, 12.9). This more extensive advance may

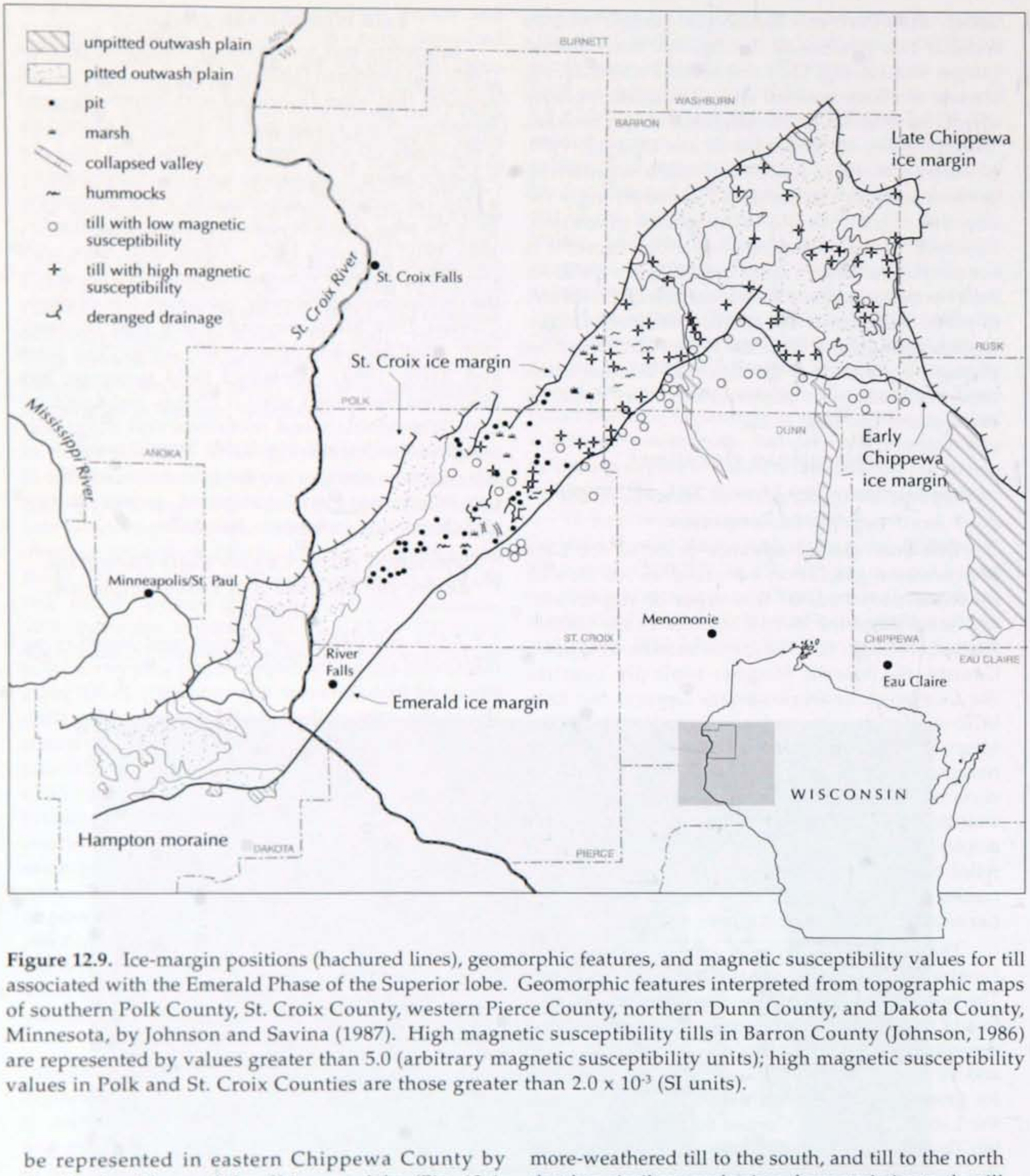


Figure 12.9. Ice-margin positions (hachured lines), geomorphic features, and magnetic susceptibility values for till associated with the Emerald Phase of the Superior lobe. Geomorphic features interpreted from topographic maps of southern Polk County, St. Croix County, western Pierce County, northern Dunn County, and Dakota County, Minnesota, by Johnson and Savina (1987). High magnetic susceptibility tills in Barron County (Johnson, 1986) are represented by values greater than 5.0 (arbitrary magnetic susceptibility units); high magnetic susceptibility values in Polk and St. Croix Counties are those greater than 2.0×10^{-3} (SI units).

be represented in eastern Chippewa County by the Stanley Phase of the Chippewa lobe (Fig. 12.4; Syverson, in press).

Though the landscape immediately south of the prominent moraines is gently rolling and at first glance similar to the regions underlain by River Falls till, clay mineralogy and magnetic susceptibility of the surface till show a distinct boundary between more-weathered till to the south, and till to the north that has similar weathering characteristics to the till in the moraines (Fig. 12.9; Johnson, 1986; Thornburg and others, 2000). Johnson (1986) concluded that the maximum Late Wisconsinan ice-margin position was south of the prominent end moraines based on these differences in lithology (ascribed to weathering) and geomorphology (most notably pitted outwash plains that grade to the St. Croix and Chippewa moraines; Fig. 12.9).

The landforms from the Emerald Phase of the Superior lobe and Early Chippewa Phase of the Chippewa lobe are distinct from those formed during the St. Croix and later phases. Hummocks, ice-walled-lake plains, tunnel channels, eskers, and outwash plains dominate the landscape formed during the later St. Croix and Late Chippewa Phases, but are absent or rare in the landscape formed during the Emerald and Early Chippewa Phases. Johnson (2000) suggested that the Superior lobe changed from a cold, non-surging glacier during the Emerald Phase, to a warmer, surging glacier during the St. Croix Phase. The landforms of the Emerald Phase, the thin to patchy cover of the Poskin Member till (Copper Falls Formation till deposited during the Emerald Phase), and the presence of well developed permafrost features suggest that the climate was colder during the Emerald Phase than during the St. Croix Phase. The absence of tunnel channels indicates that less subglacial meltwater was present during this phase, perhaps because cold conditions prevented meltwater from draining through the glacier to the bed. Meltwater mainly formed supraglacially, flowed for the most part over clean ice, and had a low sediment load. Ice flowed into pre-Emerald Phase valleys that later became collapse depressions when warming allowed buried ice to melt. Following retreat of the ice margin, permafrost conditions enhanced rapid development of a dendritic drainage network, and it is likely that much till was eroded.

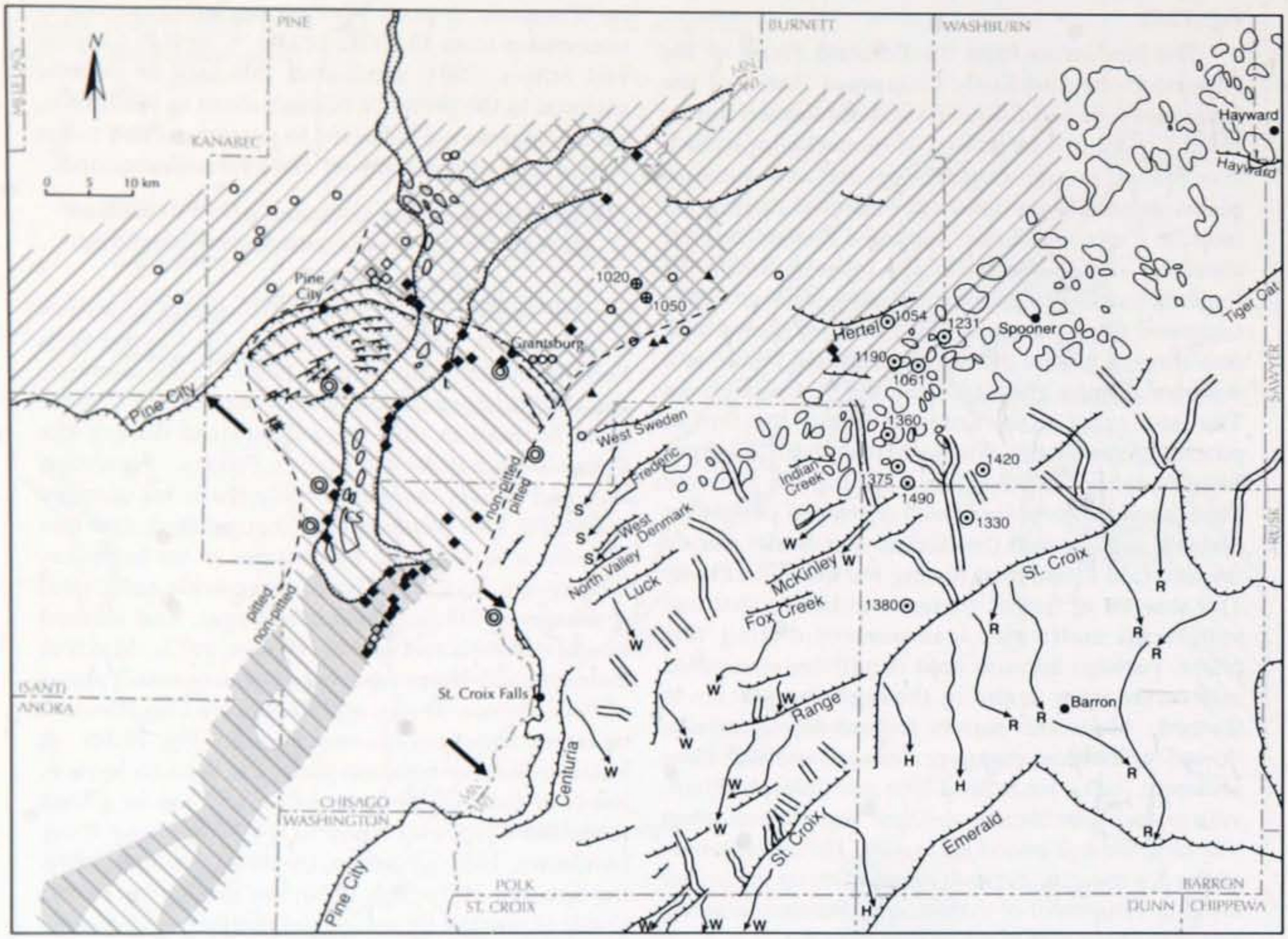
The age of the Emerald and Early Chippewa Phases is not known, but it is certainly Late Wisconsinan, probably between 20,000 and 25,000 ^{14}C yr B.P. Ice during the Late Wisconsinan glaciation crossed the drainage divide south of Lake Superior by 26,000 ^{14}C yr B.P. This is based on spruce wood in western Wisconsin buried by 60 meters of glacial stream sediment (26,060 ± 800 ^{14}C yr B.P.; Black, 1976; Attig and others, 1985). However, age control in Wisconsin is poor because organic material is uncommon from 13,000 to 26,000 ^{14}C yr B.P. Clayton and others (2001) attributed this lack of organic material to the presence of permafrost in Wisconsin, yet many areas not subjected to permafrost also suffer from a lack of radiocarbon dates from this period.

Deposition of the Copper Falls Formation during the St. Croix and Late Chippewa Phases

Till in the St. Croix[1] and Chippewa moraines is included in several members of the Copper Falls Formation (Attig and others, 1988). These members are almost identical to the members of the Copper Falls Formation that were deposited during the Emerald and Early Chippewa Phases. However, as noted above, the geomorphic character changes markedly, and Johnson (2000) suggested that this difference was caused by changes in ice behavior. During the St. Croix Phase, large-scale subglacial drainage of the Superior lobe began and formed tunnel channels and eskers (Wright, 1973). Much of the meltwater-stream sediment in the outwash plains of Polk, Barron, Dunn, and Chippewa Counties can be traced to tunnel-channel mouths (Fig. 12.10). A broad, extensive outwash plain, the Wissota terrace, heads at the Chippewa moraine and can be traced along the Chippewa River to the Mississippi River (Andrews, 1965; Syverson, in press). Additionally, the Spooner Hills (Fig. 12.10) are thought to be the result of erosion by subglacial meltwater (Johnson, 1999).

The increase in subglacial meltwater features is accompanied by the presence of widespread stagnant ice features including broad belts of hummocks dotted with ice-walled-lake plains. Johnson and others (1995) stated that although hummocks clearly seem to be collapse features, in western Wisconsin they are often composed of uniform till bearing a strong pebble

[1]The use of the name "St. Croix" in association with glacial features in western Wisconsin was originally applied to a band of hummocky topography running north–south immediately east of the town of St. Croix Falls in Polk County. This was called the "St. Croix moraine" by Berkey (1897). R.T. Chamberlin (1905) used the term "St. Croix moraine" in the same sense as Berkey and recognized several other features in Polk County that he called moraines, including the "Alden moraine." The name was first applied to landforms farther to the southeast by Leverett (1932), and it is this hummocky region that geologists generally refer to as the "St. Croix moraine." Although Leverett recognized Chamberlin's other moraines (such as the Alden), he did not recognize the "St. Croix moraine" in the sense defined by Berkey (1897) and used by Chamberlin (1905). Leverett gave no explanation for the name change. Wright (1973) referred to the ice advance that made this landscape the "St. Croix Phase." The ice-margin position that was represented by Berkey's and Chamberlin's original "St. Croix moraine" was called the Centuria Phase by Johnson (2000).



EXPLANATION OF MAP SYMBOLS

- Distribution of buried red and brown laminated lake sediment of Helgesen and Lindholm (1977)
- Extent of glacial Lake Lind
- Extent of glacial Lake Grantsburg
- Ice-flow direction of Grantsburg sublobe inferred from till fabric
- Low-relief Grantsburg sublobe moraines
- Kettle River and St. Croix channel with bars on the Chengwatana surface
- Tunnel channel
- Boundary between pitted and non-pitted topography
- Dominant meltwater drainage route
  - R Red Cedar River
  - H Hay River
  - W Willow River
  - S flows to the southwest
- Ice-margin position
- Spooner Hills
- Outcrop of glacial Lake Grantsburg varved clay
- Outcrop of glacial Lake Grantsburg varved clay exposed beneath Trade River till
- Outcrop of glacial Lake Lind varved clay
- Outcrop of glacial Lake Lind near-shore sediment
- 1050 Location and elevation of glacial Lake Grantsburg wave-washed gravel
- 1490 Selected spot elevations
- Outcrop of silty Copper Falls till
- Town

fabric. They concluded that this till is likely melt-out till, too sandy and well-drained to flow during collapse following melting of the underlying ice.

Clayton and others (1985) suggested that many lobes of the Laurentide Ice Sheet surged during the Late Wisconsinan glaciation, and Johnson and Savina (1987) agreed that the landforms of the Superior lobe support this idea. Increased subglacial discharge of meltwater is associated with surging (Kamb and others, 1985), and surging can produce large tracts of stagnant ice that melt to form hummocks (Wright, 1980). Clapperton (1975) suggested that large amounts of debris are frozen to the base of the glacier during a surge. A thick, basal, debris-rich layer in a mass of stagnant ice could produce high-relief hummocks like those in Polk, Barron, St. Croix, and Chippewa Counties.

A similar surging mechanism is implied by the landforms of the Chippewa lobe. The Chippewa lobe flowed southeasterly during the Perkinstown Phase but then flowed southwesterly (a change of 90°) during the Late Chippewa Phase to build the high-relief hummocky Chippewa Moraine in northwestern Chippewa County (Fig. 12.11; Waggoner and others, 2001; Syverson, in press). Ham and Attig (1997) described these events in the Chippewa and Wisconsin Valley lobes in Wisconsin and proposed that they represent glacier surges that contributed to moraine formation by carrying much debris to the glacier surface (see Stop 12-10 for more detailed discussion).

The scenario outlined above for the St. Croix and Late Chippewa Phases suggests that the moraines do not represent long periods of climatic equilibrium,

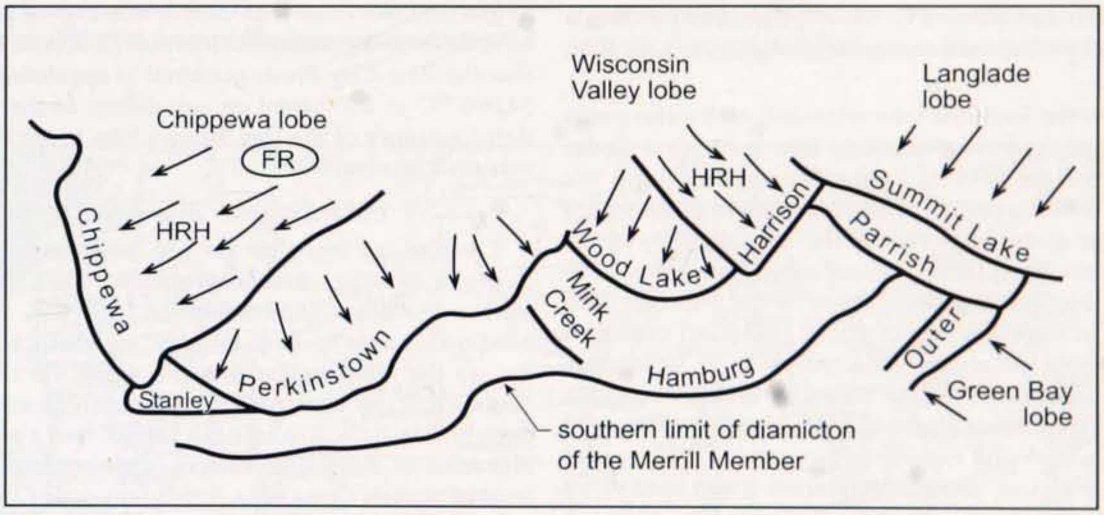


Figure 12.11. Ice-flow directions and moraines of the Wisconsinan glaciation. The high-relief, hummocky (HRH) Chippewa and Harrison moraines formed after the Chippewa and Wisconsin Valley lobes underwent major changes in ice-flow direction. These major changes of flow direction are thought to represent ice-surging events that contributed to the development of the high-relief moraines. LRH= low-relief, hummocky moraine; FR=Flambeau Ridge. Modified from Attig and others (1998) and Waggoner and others (2001).

Figure 12.10. Part of northwestern Wisconsin and adjoining Minnesota showing features associated with the most recent glacial events. Spot elevations (in feet) are included to show that elevations increase from central Burnett County to the St. Croix ice margin. Emerald, St. Croix, Early Chippewa, and Late Chippewa ice margins are from Johnson (1986). The Pine City ice margin in Minnesota is from Cooper (1935) and Hobbs and Goebel (1982). Locations of glacial Lake Grantsburg and glacial Lake Lind varved clay in Minnesota are from Chris Hemstad, M.D. Johnson, and Gary Meyer (unpub. data). Only tunnel channels in Wisconsin are shown. The extent of glacial Lake Grantsburg in Minnesota is in part from Cooper (1935). Till fabric measurements in Minnesota are from Chernicoff (1983). Modified from Johnson (2000). CS (in eastern Minnesota) = location of Camp Sunrise. L = Lind, Wisconsin.

but merely the outer boundary of short-lived surge events. In fact, the St. Croix ice-margin position actually consists of several penecontemporaneous ice-margin positions (Fig. 12.10), suggesting that the bands of hummocks in the St. Croix moraine may be the products of several surge events. The age of these advances is not known, but Johnson (2000) suggested they occurred between 15,000 and 18,500 ^{14}C yr B.P.

The history of the Superior lobe after the St. Croix and Late Chippewa Phases is a sequence of overall retreat punctuated with numerous readvances. This is also true for the Chippewa lobe, but to a lesser extent. These readvances left several landforms that indicate distinct ice-margin positions in western Wisconsin (Figs. 12.4, 12.10). Numerous ice margins are marked by outwash heads, bands of hummocks, and tunnel-channel mouths. These discrete landforms are often difficult to connect laterally in order to identify continuous, isochronous ice-margin positions, but Johnson and Mooers (1998) attempted to produce a map showing continuous ice-margin positions (Fig. 12.4).

As the Superior lobe retreated, meltwater paths changed from southeasterly (along the Red Cedar and Willow Rivers) to southwesterly (along the St. Croix River). Apparently, stagnant ice and a change in drainage caused the former valley of the St. Croix River (shown as a dashed line in Fig. 12.4) to be blocked, and the retreat of the Superior lobe allowed the expansion of glacial Lake Lind in eastern Minnesota and western Wisconsin (Fig. 12.10; Johnson and others, 1999). Glacial Lake Lind sediment consists of reddish-brown, varved sand, silt, and clay up to 30 meters thick and represents more than 1,000 years of sedimentation. Johnson and others (1999) used varve correlations to show that the Superior lobe retreated at a rate of 150 to 200 meters per year during the lake's existence. Johnson (2000) has included this sediment in the Sunrise Member of the Copper Falls Formation. The age of glacial Lake Lind is unknown, but it likely existed between 14,000 and 18,000 ^{14}C yr B.P. Glacial Lake Lind gradually filled with deltaic sediment derived from the retreating Superior lobe, and it was completely filled with sediment prior to the advance of the Grantsburg sublobe and prior to the retreat of the Superior lobe into the Lake Superior basin.

Deposition of the Trade River Formation

The Grantsburg sublobe of the Des Moines lobe advanced from the west-southwest into western Wisconsin during the Pine City Phase and deposited till of the Trade River Formation (Figs. 12.3, 12.10; Wright, 1972; Wright and others, 1973; Chernicoff, 1983; Johnson, 2000). This ice occupied an area previously covered by the Superior lobe, including much of the filled-in glacial Lake Lind basin. Trade River Formation till is gray, loamy, and calcareous, similar to the tills of the pre-Wisconsinan Pierce and Marathon Formations, but Trade River Formation till exhibits a low degree of weathering and erosional modification (Table 12.1; Johnson, 2000). The Pine City Phase ice margin dammed the St. Croix River drainage along the Minnesota and Wisconsin border and glacial Lake Grantsburg formed (Fig. 12.10). Varve counts imply that the ice advanced into glacial Lake Grantsburg at rates of 5 to 7 kilometers per year. Varves deposited during the main phase of glacial Lake Grantsburg suggest that this lake lasted only about 100 years (Johnson and Hemstad, 1998; Johnson, 2000). A flat, poorly drained lake plain marks the area once covered by glacial Lake Grantsburg, but this surface is best thought of as the delta-filling surface of glacial Lake Lind with only a thin cap of glacial Lake Grantsburg sediment. Johnson (2000) estimated that the Pine City Phase occurred at approximately 14,000 ^{14}C yr B.P. based on correlation to the well-dated advance of the Des Moines lobe to the Bemis margin in central Iowa.

Lake Superior diversion

As the ice margins wasted north across the Superior drainage divide, proglacial lakes formed within the deep Superior basin (Clayton, 1984). Modern outlets toward the east were blocked by ice, so the lakes drained to the south via the St. Croix River (for example glacial Lakes Nemadji and Duluth). Initially, glacial Lake Duluth had a surface elevation of 327 to 330 meters (approximately the level of Skyline Drive in Duluth, Minnesota; Clayton, 1984; Farrand and Drexler, 1985). Glacial Lake Duluth drained into the St. Croix River system through an outlet near Moose Lake, Minnesota. A later, slightly lower outlet near Brule, Wisconsin supplied water to the headwaters of the St. Croix River. Water may have occupied both outlets when the level of glacial Lake Duluth was at its highest (Farrand and Drexler, 1985).

The large volume of water diverted into the St. Croix River system incised the deep gorge of the St. Croix River that serves as the boundary between Minnesota and Wisconsin. This event also formed large potholes in the Precambrian basalt well above the modern river level at Interstate State Park in Wisconsin and Minnesota (Black, 1974; Clayton, 1984; Johnson, 2000). Johnson (2000) suggested that this valley incision occurred approximately 9,000 to 12,000 ^{14}C yr B.P.

Fine-grained sediment was deposited in the lakes within the Superior basin. The last glacial advance into northwestern Wisconsin (the Lakeview Phase of Clayton, 1984) eroded the silt- and clay-rich lake sediment and deposited stone-poor till units that contain 70 to 90 percent silt and clay (Miller Creek Formation in northern Wisconsin; Mickelson and others, 1984). Black (1976) reported wood dates from red, clay-rich till of this event in northern Wisconsin (9,730 ± 140 ^{14}C yr B.P. and 10,100 ± 100 ^{14}C yr B.P.). Clayton (1984) correlated this 9,900 ^{14}C yr B.P. advance with the Marquette Phase in the Upper Peninsula of Michigan and estimated that the glacier margin wasted out of Wisconsin for the last time by 9,500 ^{14}C yr B.P.

FIELD TRIP STOPS (Fig. 12.12)

DIRECTIONS: Cross the St. Croix River at Stillwater, Minnesota onto Highway 35/64. Veer right onto St. Croix County Road E. The route will cross State Highway 35 in the town of Houlton approximately 0.2 mile east of the bridge. Proceed along County Road E for an additional 3.1 miles. Turn left (north) onto Valley View Trail and follow it for 2.8 miles to the Bass Lake Cheese factory. Note the knob and kettle topography along this stretch. Approximately 1.2 miles after turning onto Valley View Trail, the road crosses over what was described by Hinke (2003) as part of a tunnel valley system associated with the Sylvan Lake Member of the Copper Falls Formation. Turn left onto 60th Street. Drive 1 mile north to where the road turns right and becomes 150th Avenue. Continue along 150th Avenue for 0.2 mile. The road then turns to the left (north) and becomes 63rd Street. Proceed along 63rd Street for 1 mile and turn left onto St. Croix County Road I. Follow County Road I north for 2.1 miles to the intersection with State Highways 35/64 in the town of Somerset. Turn right (east) onto Highway 35/64. Follow Highway 35/64 for 0.1 mile and turn left (north) onto State Highway 35. Continue on Highway 35 for 6.9 miles and turn right (east) onto 10th Avenue. Drive east for 3.7 miles. Stop 12-1 is a small exposure on the left (north) side

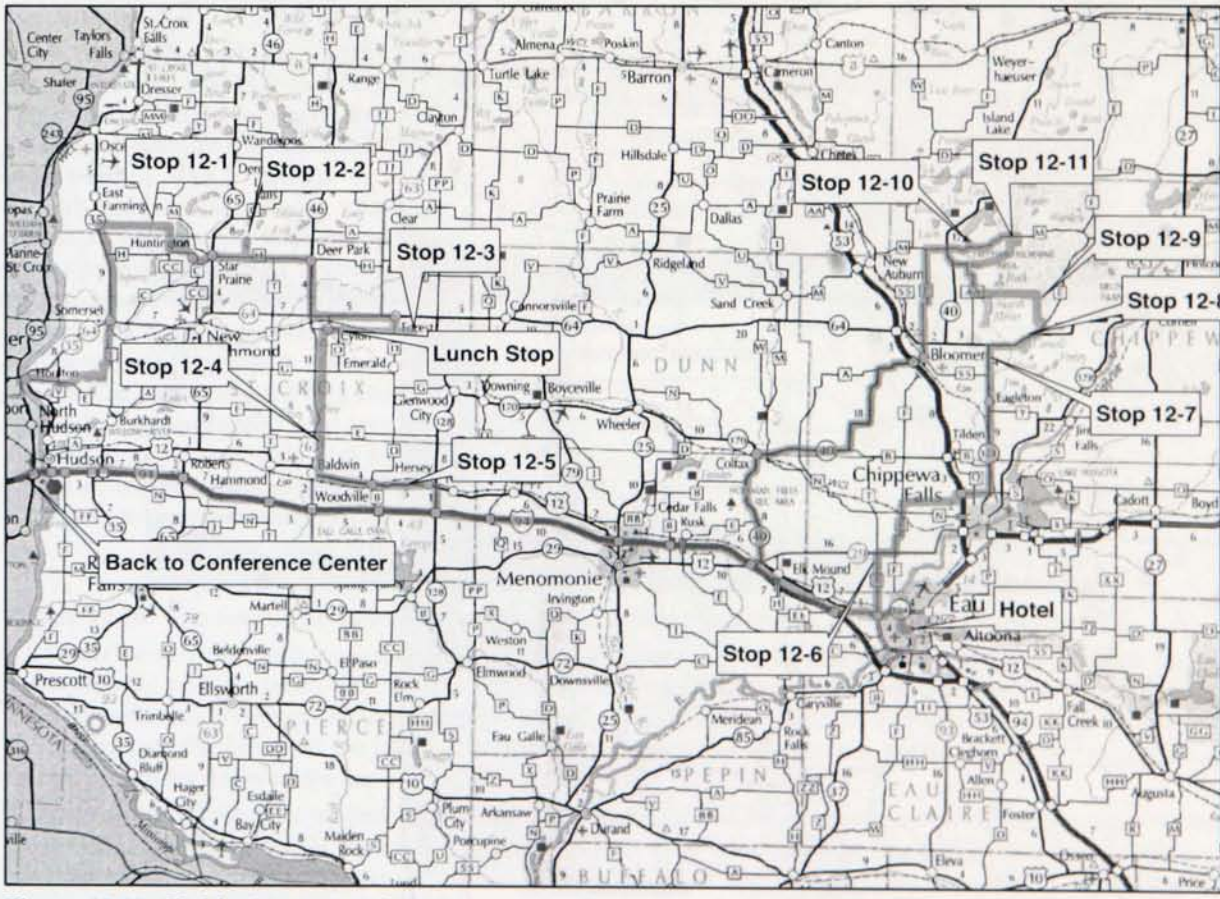


Figure 12.12. Field trip route and stops.

of the road. This stop is located approximately 0.1 mile east of the farm at 2267 10th Avenue.

STOP 12-1

Cedar Lake ice-walled-lake plain

Location: T. 32 N., R. 18 W., sec. 29, SW, SE

New Richmond North quadrangle; UTM: 529,850E/5,007,800N

Description: The Cedar Lake ice-walled-lake plain is the largest ice-walled-lake plain in western Wisconsin at almost 13 square kilometers. Figure 12.13 shows typical features of what Clayton and Cherry (1967) called an unstable-environment ice-walled-lake plain (Fig. 12.14). The Cedar Lake ice-walled-lake plain has a well-defined rim 10 meters higher than the plain's center. At this stop we see clearly that the rim is composed of sorted sand and gravel in well defined foresets (Fig. 12.15). The presence of this coarse-grained, water-sorted sediment and the orientation of the foresets strongly suggest that the sediment source was the adjacent stagnant-ice surface. A stream system must have developed on the thick supraglacial sediment, and these streams transported sediment to the ice-walled lake.

Figure 12.13. Parts of New Richmond North, Nye, Osceola, and Somerset North quadrangles, Wisconsin and Minnesota showing the Cedar Lake ice-walled-lake plain line surrounded by hummocky topography. Contour interval is 10 feet, except in the southwestern part, where it is 20 feet. Hummocks in this region contain Copper Falls Formation till, gravity-flow deposits, outwash, and lake sediment. DH = location of drill hole mentioned in the text. The topographic profile in Figure 12.16 trends east–west through this drill hole.

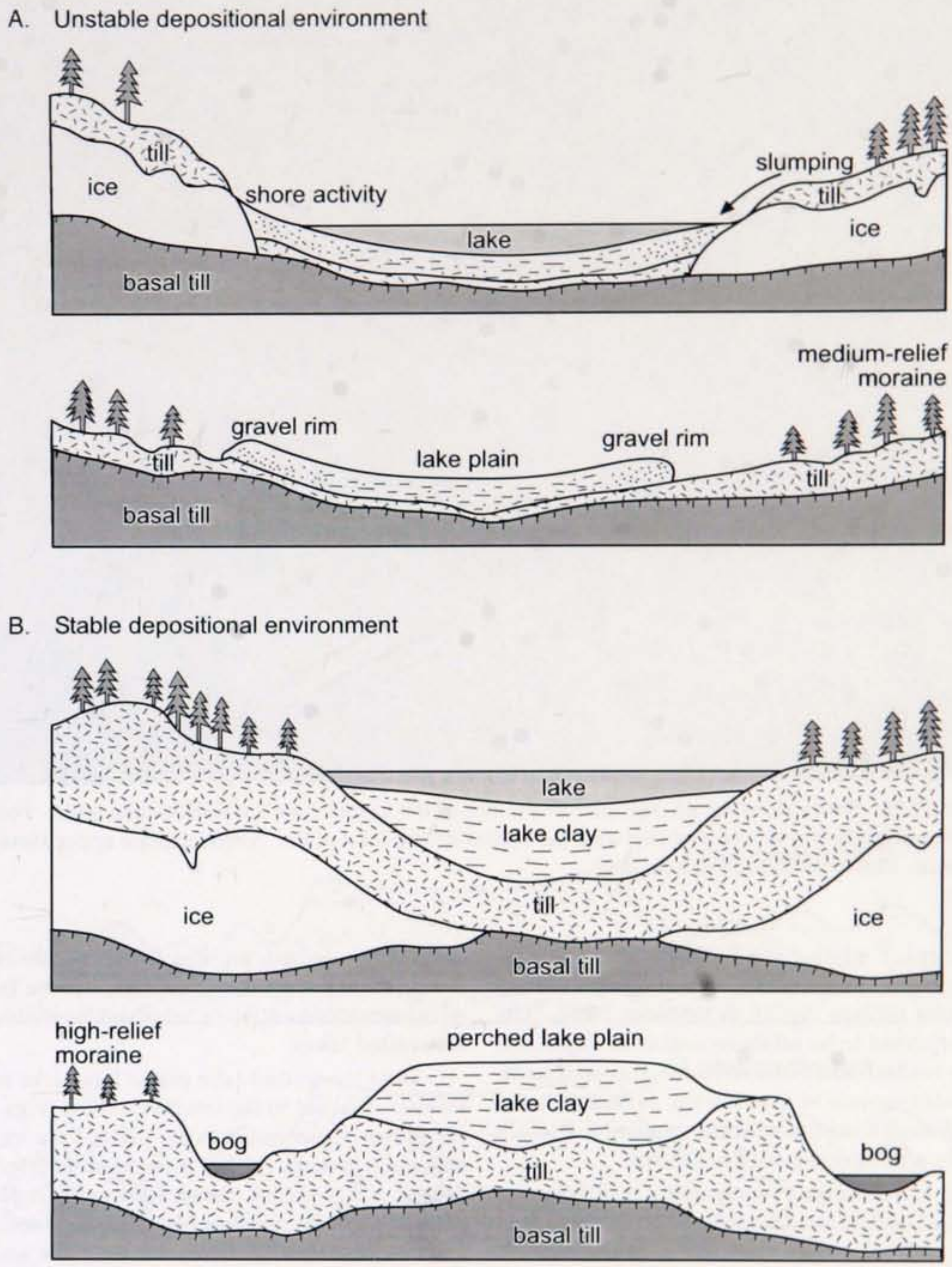


Figure 12.14. The formation of ice-walled-lake plains (modified from Clayton and Cherry, 1967).

A. Unstable-environment ice-walled-lake plain. Sediment is thin on the ice surrounding the lake, so ice melts rapidly and produces a lot of water that carries large quantities of sediment into the lake. Coarse-grained sediment is deposited near the shore in deltas, and silt and clay are deposited in more quiet-water conditions near the center of the lake. After the ice melts and the lake drains, gravel rim ridges may remain around the outer parts of the ice-walled-lake plain.

B. Stable-environment ice-walled-lake plain. Thick sediment on the ice prevents the ice confining the lake from melting quickly. Little meltwater is produced, so little coarse-grained sediment is transported into the lake. Rather, silt and clay are deposited in the long-lived lake. When the surrounding ice melts, the former lake plain remains perched high in the landscape.

Figure 12.15. Photo showing an exposure in the rim of the Cedar Lake ice-walled-lake plain. Foreset beds are slightly deformed and dip into the center of the lake plain. Vertical scale approximately 4 meters. Photo by Robert Baker, 2004.

The center of the ice-walled-lake plain is underlain by 20 meters of brown silt and overlies Copper Falls till (see Fig. 10 in Johnson, 2000). The silt is interpreted to be offshore sediment deposited in the ice-walled lake. As with the rim sediment, the sediment source is interpreted as supraglacial stream sediment from the adjacent stagnant ice, which necessarily was covered with sediment.

Figure 12.16 shows a topographic profile across the southern end of the Cedar Lake ice-walled-lake plain (Fig. 12.13). Note that the rims are similar in elevation to surrounding hummocks, and also that the thickness of the offshore sediment (dashed pattern) is much greater than the relief of the ice-walled-lake plain. The geomorphology, stratigraphy, and sedimentology indicate to us that the lake was surrounded (walled) by stagnant ice. Because of the existence of permafrost features in the region, it is also likely that the lake existed while permafrost conditions existed. This means that the melting of stagnant ice was significantly checked until the permafrost melted, an idea presented by Ham and Attig (1996a). This also indicates that the formation of hummocks took place after sedimentation in the ice-walled lakes.

Most ice-walled-lake plains have lake sediment thinner or equal to the landform relief (Figs. 12.17A, B). However, sediment in the Cedar Lake ice-walled-lake plain is much thicker than the relief (Fig. 12.17C). This is a significant observation with regards to a current controversy about hummock genesis. Some authors (Eyles and others, 1999; Boone and Eyles, 2001) suggested that most hummocks, and even those features called ice-walled-lake plains, are the result of subglacial squeezing under the stagnant ice of a glacier that has surged. They would argue, for example, that the type of ice-walled-lake plain seen in Figure 12.17A is a product of upward squeezing of soft till into an ice-walled lake and only a small amount of lake sediment would accumulate on top of this squeezed sediment; similar stratigraphy could also be achieved by an ice-walled lake initially filling

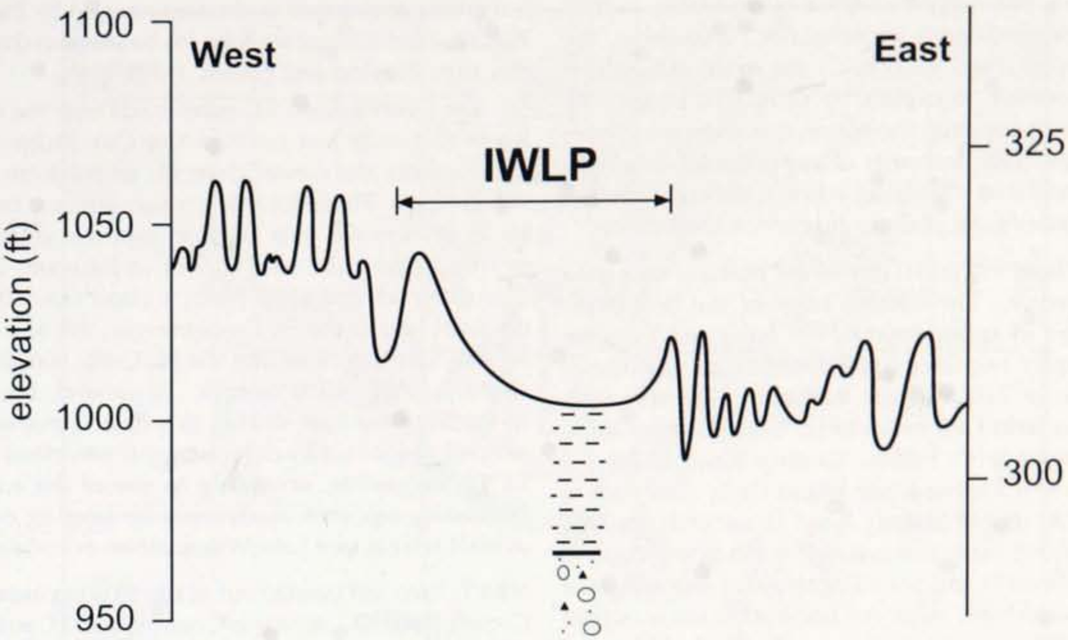


Figure 12.16. Topographic profile across the southern end of the Cedar Lake ice-walled-lake plain showing bordering hummocks and drill hole information. Location runs east–west through the drill hole (marked DH) in Figure 12.13 (but the cross section line is not shown on that figure). Note that the ice-walled-lake plain rims are similar in elevation to surrounding hummocks, and also that the thickness of the offshore sediment (dashed pattern) is much greater than the relief of the ice-walled-lake plain. Compare with Figure 12.17.

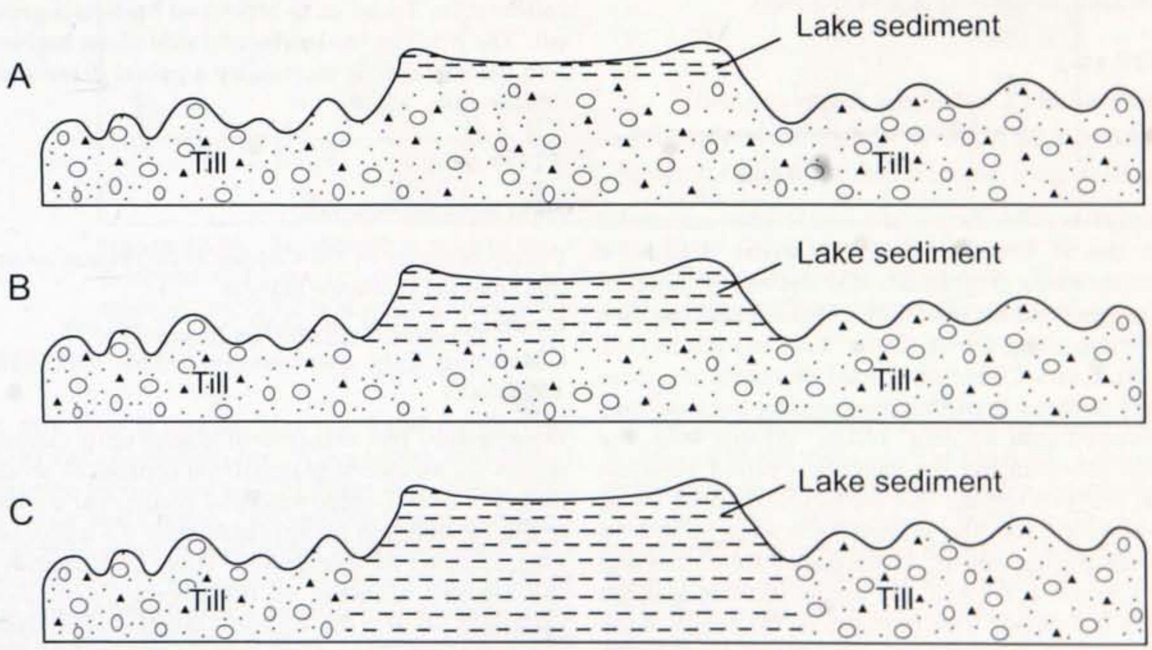


Figure 12.17. Ice-walled-lake plains with varying thicknesses of lake sediment. Types A and B are quite common and have been reported widely in Wisconsin, the Dakotas, southern Canada, and Europe (Johnson and Clayton, 2003). The Cedar Lake ice-walled-lake plain is an example of type C. See text for explanation.

with flow till, followed by a more stable period when lake sediment accumulated. However, the other types (Figs. 12.17B, C) are more difficult, if not impossible, to explain by subglacial squeezing. This in turn supports the notion that there must have been significant amounts of supraglacial debris on the ice, and that squeezing was not the main process in ice-walled-lake plain or hummock formation.

NEXT: Turn left (east) out of the parking area onto 10th Avenue. The eastern edge of the lake plain is crossed in approximately 0.9 mile, and then the topography becomes quite hummocky. Continue 0.5 mile to 210th Street and turn right (south). Continue south 1.7 miles to St. Croix County Road H and turn left. Follow County Road H for 3.2 miles to Star Prairie River Island Park. The park is on the left side of County Road H just after the Star Prairie Trout Farm and just before the intersection of County Road H and State Highway 65. We will have a snack stop here. After our break, turn left (east) out of the parking lot onto County Road H. This road follows Highway 65 for slightly less than 0.1 mile before County Road H splits (left) from Highway 65. Continue east on County Road H for 2.5 miles to St. Croix County Road CC and turn left (north). Drive north on County Road CC for 1.8 miles. There is a U.S. Fish and Wildlife Service parking area on the right side (east) of the highway just north of the small lake on the east side of the road.

STOP 12-2

County Road CC esker and tunnel channel

Location: T. 32 N., R. 17 W., sec. 33, NW, NE

Deer Park quadrangle; UTM: 541,000E/5,007,600N

Description: This stop is at a scenic esker associated with the St. Croix Phase. The esker is about 6 kilometers long (Fig. 12.18), and numerous collapse-pits suggest that the esker lies within a tunnel channel. Elsewhere along the St. Croix Phase margin in Polk and St. Croix Counties, tunnel channels are more clearly defined than this one, mainly because they are eroded into till (Fig. 12.10). At this stop, the region surrounding the esker is a pitted outwash plain, and this implies that the tunnel channel, esker, and stagnant ice that existed here were buried by outwash during retreat. When buried ice melted and this outwash collapsed, the tunnel channel was not clearly expressed. This seems to be the origin of the many elongate lakes or strings of lakes surrounded by outwash in these counties (for example Wapagasett Lake, Bone Lake, Long Lake, and Deer Lake in Polk County). Interestingly, tunnel channels are common in the area covered by the Superior lobe, but rather uncommon in the area covered by the Late Wisconsinan Chippewa lobe (to be seen on day 2 of this trip; Clayton and others, 1999).

The County Road CC esker starts near the Apple River and ends just northeast of Oak Ridge Lake. Many eskers and tunnel channels with eskers along the St. Croix Phase ice-margin position are fronted by large outwash fans at theirs mouths, such as at nearby Clear Lake (see Fig. 18 in Johnson, 2000). The esker at this stop lacks a clear fan. It also does not end at the St. Croix margin, but at a point several kilometers behind the St. Croix margin at a slightly younger ice margin. In general, the edge of the Superior lobe during this time is marked by several discontinuous ice-margin positions (Fig. 12.10) suggesting, according to one of the authors (Johnson), repeated readvances by surging during overall retreat (see Late Wisconsinan overview).

NEXT: Turn left (south) out of the parking area onto County Road CC, return to County Road H, and turn left (east). Follow County Road H for 6.2 miles to State Highway 46 in the town of Deer Park and turn right (south). Drive south 3.9 miles on Highway 46 to State Highway 64 and turn left (east). Proceed east on Highway 64 for 6 miles to the town of Forest, turn right (south) on St. Croix County Road D. Drive south on County Road D for 1 mile. Turn left (east) onto St. Croix County Road S (180th Avenue) and follow it for 1.4 miles to Milestone Materials gravel pit. The pit is on the left (north) side of the highway and the entrance is marked by a gravel drive and a yellow gate.

STOP 12-3

Hard hats are required

Poskin Member of the Copper Falls Formation and the Emerald Phase ice margin

Location: T. 31 N., R. 15 W., sec. 33, SE, SW

Glenwood City quadrangle; UTM: 560,310E/4,997,060N

Description: The sequence of glacial units exposed within the Milestone Materials pit consists of several meters of cross-bedded sand and gravel near the base of the pit overlain by approximately 0.5 meter of a gray, compact till (Fig. 12.19). These units represent the Hersey Member of the Pierce Formation, deposited prior to the Wisconsinan glaciation based on reversed remanent magnetism (Figs. 12.2, 12.6). Hersey Member till is unconformably overlain by a horizontally bedded sand unit of the River Falls Formation and approximately 4 meters of massive River Falls till at the surface (Fig. 12.19). The goal

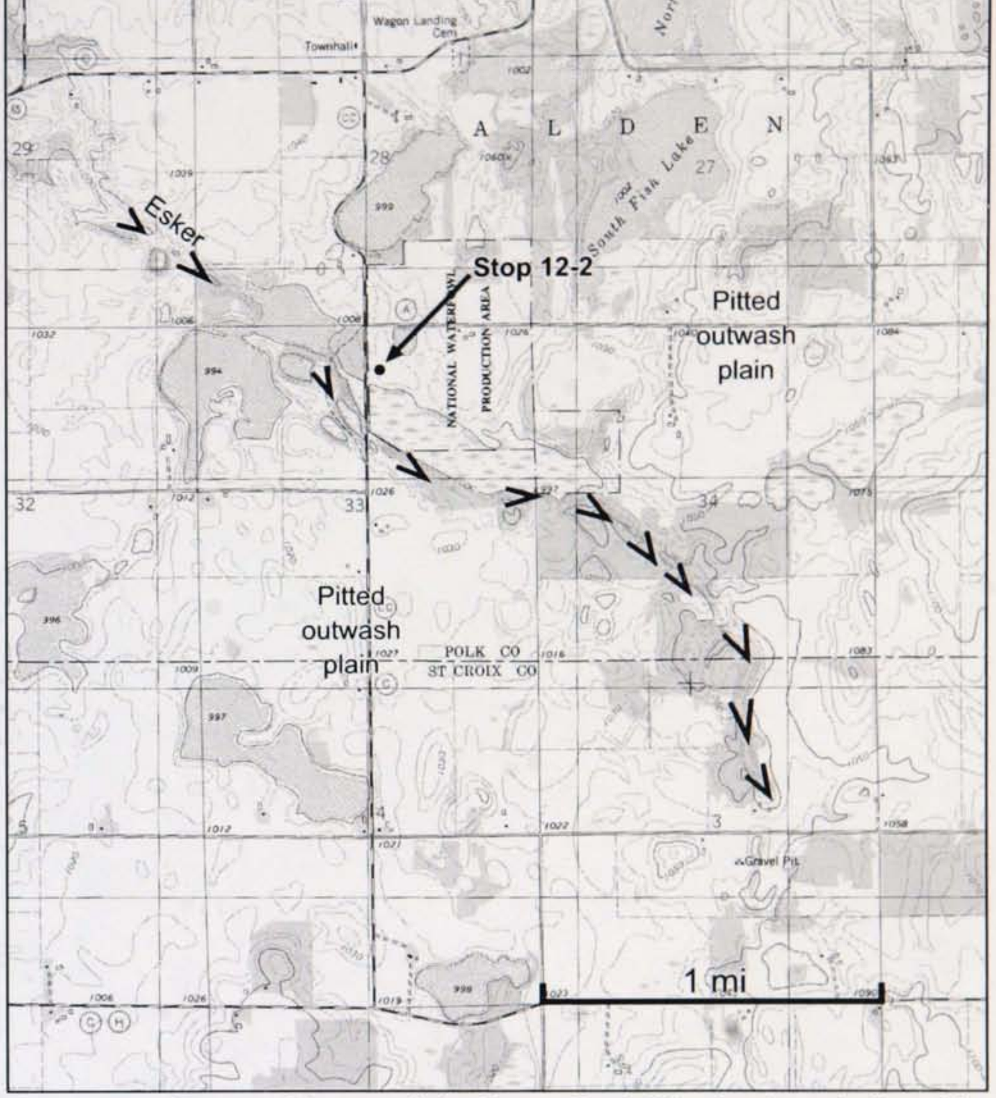


Figure 12.18. Topographic map of the County Road CC esker (crest indicated by symbols) and tunnel channel taken from the Deer Park quadrangle, Wisconsin. This esker at Stop 12-2 is about 6 kilometers long, and it lies within a tunnel channel that is obscured by collapsed outwash.

of this stop is to compare the Superior-derived River Falls Formation (pre-Wisconsinan, considered Illinoian by workers such as Baker and others, 1983; Syverson and Colgan, 2004) and the Superior-derived Poskin Member of the Copper Falls Formation (Late Wisconsinan). In eastern St. Croix County, till of the Poskin Member is a reddish-brown sandy loam that averages 69 percent sand, 22 percent silt, and 9 percent clay, and till of the River Falls Formation is also a sandy loam that averages 65 percent sand, 23 percent silt, and 12 percent clay.

Because the tills have similar texture and color, it is difficult to tell them apart in the field. Laboratory techniques are required to make such a determination. Johnson (1986, 2000) noted that the younger, less weathered Poskin Member till has a higher average magnetic susceptibility than the River Falls Formation till in Barron and St. Croix Counties. This is important because there is no well defined moraine marking the extent of the ice that deposited the Poskin Member. Johnson (2000) observed geomorphic evidence in St. Croix County, including deranged drainage, hummocks, and collapsed valleys, that correlated

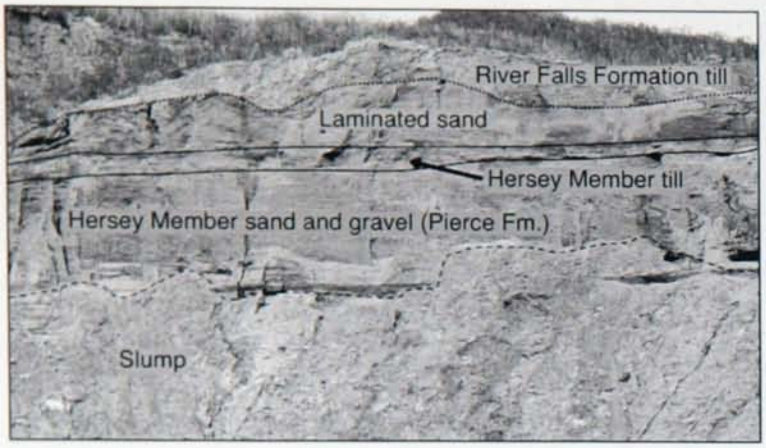


Figure 12.19. Exposure along the northeastern wall of the pit at Stop 12-3. Hersey Member of the Pierce Formation outwash and till is unconformably overlain by outwash sand and till of the River Falls Formation. Vertical scale approximately 10 meters.

with high magnetic susceptibility values for the local till (Fig. 12.9). Johnson (2000) used these data to interpret the extent of the Poskin Member of the Copper Falls Formation (and therefore the extent of ice during the Emerald Phase). Kostka (unpub. data) has noted low kaolinite:illite ratios in till samples collected to the north and west of the Emerald Phase margin. Low kaolinite:illite ratios in the Superior-derived tills of western and north-central Wisconsin have been interpreted as representing a shorter period of weathering, and hence a younger age (Stewart and Mickelson, 1976; Johnson, 1986; Thornburg and others, 2000; Syverson and Colgan, 2004).

Kostka (unpub. data) investigated domestic well construction reports associated with two small plains west of Emerald, Wisconsin. Four well logs from the easternmost plain, the Emerald plain, indicate clay accumulations between 50 and 80 feet thick, and Kostka (unpub. data) interpreted this clay as ice-dammed-lake sediment from the Emerald Phase. Cuttings from a solid-stem boring in the Emerald plain contained approximately 93 percent silt and clay and had what appeared to be broken pieces of fine-grained laminations (possibly varves). Along the western edge of the Emerald plain is an area of hummocks that Johnson (2000) cited as evidence for the Emerald Phase (Fig. 12.9). Ice at this position would have prevented water from flowing northwest through the modern drainage now called Hutton Creek. Downstream (northwest) from the Emerald plain is another small plain near the town of Cylon (the Cylon plain) where five well logs indicate accumulations of silt and clay that range from 61 to 125 feet thick. As the ice retreated toward the northwest, water would have been dammed in front of the ice in this low area as well. The presence of ice-dammed lakes in this area lends strong support for the Emerald Phase as proposed by Johnson (2000).

NEXT: Turn right (west) out of Milestone Materials and drive 6 miles to the town of Cylon. Lunch at the town park. After lunch, proceed 0.3 mile west on County Road S (180th Avenue) to 220th Street and turn left (south). Drive 8.0 miles to the intersection with St. Croix County Road E. Cross County Road E—the entrance to the Kusilek gravel quarry is on the right (west) side of 220th Street approximately 0.1 mile past County Road E. The south wall of the pit is an exposure of the River Falls Formation.

STOP 12-4

River Falls Formation type locality; the Joseph Kusilek gravel quarry

Location: T. 29 N., R. 16 W., sec. 18, SE, NE, NE Emerald quadrangle; UTM: 550,350E/4,983,550N

Description: This stop, the type locality for the River Falls Formation (Mickelson and others, 1984), has been an active gravel operation for at least 30 years. Although the exposure has changed considerably during this time, the geology has remained both complex and intriguing. The north-facing quarry wall exposes a complex assortment of loess, till, and deformed sand and gravel of the River Falls Formation overlying stratified sand and gravel of the Pierce Formation.

Most of the till of the River Falls Formation is rather massive basal (lodgement) till. However, the upper portion of the unit in parts of St. Croix County is weakly stratified, contains discontinuous lenses of deformed sand and gravel, and is probably supraglacial in origin. The color of the River Falls till varies vertically within the weathering profile from yellowish-red (5YR 4/6) in the argillic horizon to reddish-brown (5YR 4/4) in the C horizon. The River Falls Formation is deeply weathered with solum thicknesses up to 2.8 meters. The unweathered till matrix is sandy clay loam averaging 60 percent sand, 15 percent silt, and 25 percent clay. Pebble lithologies average 64 percent igneous, 11 percent metamorphic, and 25 percent sedimentary rock types.

In the eastern portion of the exposure at this stop, approximately 1 to 3 meters of massive till overlies about 1 meter of weakly bedded sediment, possibly basal melt-out till or a lens of ice-contact-stratified sediment (Fig. 12.20). Farther to the west, the River Falls Formation till overlies about 2 to 3 meters of River Falls Formation stream sediment cut by faults (Fig. 12.21). Along the western portion of the exposure, thin River Falls Formation till overlies up to 5 meters of Pierce Formation sand and gravel (outwash). Pierce Formation outwash is distinguished from the sand and gravel of the River Falls Formation on the basis of color and clast lithologies. Pierce Formation stream sediment color is lighter (light gray, 2.5Y 7/1 to 10YR 7/1) than the redder River Falls sand and gravel (pink to reddish-yellow, 5YR 7/3-6) and contains abundant limestone, shale, and ironstone concretions, and the River Falls sediment is rich in basalt, gabbro, and rhyolite (Fig. 12.22; Jensema, 1987).

Several questions to be answered are whether the deformation structures seen at this stop are the result of collapse from the melting of underlying buried ice

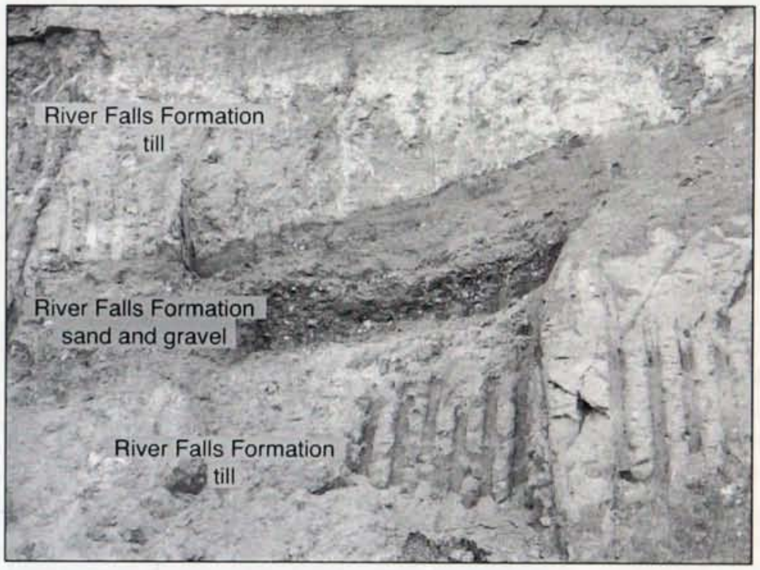


Figure 12.20. Lens of River Falls Formation sand and gravel (approximately 1 meter thick) within River Falls Formation till that is exposed at Stop 12-4.

Figure 12.21. Deformed River Falls Formation sand and gravel, approximately 2.5 meters thick, underlying River Falls Formation till at Stop 12-4.

Figure 12.22. Pierce Formation sand and gravel underlying 1 to 4 meters of River Falls till at Stop 12-4. See stop description for possible origins of the faulting.

(ice-contact-stratified sediment) or are the result of deformation as ice overrode the area and deposited the River Falls Formation till (deformed outwash).

NEXT: Turn right (south) out of the gravel pit and continue on 220th Street to U.S. Highway 12 (2.7 miles); turn left (east). Drive east on Highway 12 for 6.4 miles. There is a driveway on the left (north) side of the highway at fire number 2868. Entrance to the Cook quarry is via a private drive.

STOP 12-5

Entrance is via private property! Permission must be obtained before entering!

Hersey Member of the Pierce Formation type locality; the Rose Ann and Rodney Cook gravel quarry

Location: T. 29 N., R. 15 W., sec. 29, SW, SW, SE Wilson quadrangle; UTM: 561,150E/4,978,600N

Description: This stop is the type locality for the Hersey Member of the Pierce Formation (Mickelson and others, 1984). Although this quarry has not been operational for more than 25 years, the till exposure along the southern wall is in very good condition. The matrix texture of the unweathered till at this location is a strongly calcareous loam averaging 42 percent sand, 33 percent silt, and 25 percent clay. The weathered Hersey Member till is typically loam to clay loam. The color ranges from yellowish-brown (10YR 5/6) in the weathered zone to dark gray (10YR 4/1) in the C horizon (Fig. 12.23). Part of the very dark color of the C horizon here is due to the presence of 9 percent finely disseminated organic matter (Baker, 1984b).

Figure 12.23. Portion of the south wall at Stop 12-5 showing about 5 meters of both oxidized and unoxidized till of the Hersey Member of the Pierce Formation.

Sand fraction, clay mineralogy, and pebble-fabric data all suggest that the Hersey Member was deposited by an ice advance from a northwestern (Keewatin) source (Fig. 12.5; Baker and others, 1983). This was an extensive advance that reached north-central Wisconsin and where the correlative till of the Pierce Formation, the Medford Member of the Marathon Formation, was deposited (Fig. 12.2; Baker and others, 1987; Syverson and Colgan, 2004). Paleomagnetic analyses of the Hersey and Medford tills show that they are reversely magnetized and that the Kinnickinnic glaciolacustrine sediments were deposited during a period of time that spanned a transition from reversed to normal polarity (Fig. 12.6; Baker and others, 1982, 1983, 1987). It is not clear whether the magnetic transition represented by the glaciolacustrine sediments is the Matuyama–Brunhes transition, which has been redated at 780 ka (Izett and Obradovich, 1994; Singer and others, 1999), or the Matuyama–Reunion transition dated at 2.1 Ma (Baker and others, 1987). For comparison, the till of the River Falls Formation was deposited during the Brunhes normal-polarity epoch and is therefore younger than 780 ka (Figs. 12.6, 12.7).

NEXT: Exit the drive to the left (east) following Highway 12 for 1.5 miles to State Highway 128 and turn right (south). Follow Highway 128 south for 1.5 miles and turn left (east) onto Interstate 94. Continue approximately 6 miles east on I-94 and begin the steep descent into the Red Cedar River drainage basin. This steep scarp marks the easternmost edge of the Ordovician carbonate rocks in this area. Continue east on I-94 for 24 miles to the Highway 12/124 exit and travel east. Continue east and south for approximately 7 miles on Highway 12/124, Highway 12, and Truax Boulevard (Business Route for Highway 12) to the Ramada Inn Convention Center, downtown Eau Claire, 205 S. Barstow Street.

END OF DAY 1

OVERNIGHT IN EAU CLAIRE

NEXT: Return to the intersection of Highway 12 (Clairemont Avenue) with Highway 124 on the northwest side of Eau Claire. Continue straight (directly north) across the 12/124 intersection and the road becomes Eau Claire County Road T. Drive 3.5 miles and turn left (west) onto 30th Avenue in Chippewa County. Drive 0.25 mile west on 30th Avenue and turn left (south) into the building center parking lot. We will drive 0.5 mile beyond the gate into the Menard pit operated by American Materials.

STOP 12-6

Hard hats are required

Weathered River Falls Formation stream sediment, Menard pit

Location: T. 28 N., R. 10 W., sec. 26, SW, NE

Albertville quadrangle; UTM: 614,100E/4,970,300N

Description: Outcrops of stratified sandy gravel and sand up to 12 meters thick have been exposed continuously in this pit for the last ten years. In the northern part of the pit, 4 to 6 meters of yellowish-red (5YR 4/6) sandy gravel is exposed. The uppermost 3 to 5 meters contain massive clayey sandy gravel with clay concentrations up to 28 percent. The sediment is cemented in places by clay that Syverson (in press) interpreted as weathering-derived (Fig. 12.8). One section of the northern face exposes 1 meter of silty sand with contorted bedding. Although the silty sediment looks similar to Hersey Member till, gravel lithologies throughout the entire section are Superior-lobe lithologies (Keweenawan basalt, gabbro, rhyolite, and Lake Superior agates). Because the silty sand is above Superior lithologies, this material is interpreted as River Falls Formation lake sediment that has been cryoturbated. Ice-wedge casts indicative of permafrost are quite common in western Wisconsin (Black, 1965; Johnson, 1986; Holmes and Syverson, 1997; Clayton and others, 2001). In the past, ice-wedge casts have been observed in the southern exposure of the pit, but no good examples were visible when this guidebook was written.

Soil-derived clay near the surface reduces the permeability of the stream sediment and makes the resulting soils seem till-like. Jakel and Dahl (1989) mapped some soil series in western Chippewa County that were interpreted as forming in "till" parent material. However, most of these soils developed in River Falls Formation stream sediment and grade downward into highly permeable stream sediment. The thickness of the weathering horizon is variable and seems to be a function of the amount of erosion, but in northern Chippewa County these upland stream sediment units appear less weathered. The River Falls Formation might represent several different glaciations.

The clay-enriched sediment at the surface grades downward into horizontally bedded, permeable sandy gravel and sand that contains rocks up to 50 centimeters in diameter. Some of the boulders are shaly, fine-grained sandstone that could not survive long transport distances. This suggests that Eau Claire Formation bedrock is present relatively close to the surface near this pit. This stream sediment is up to 35 meters thick in the southwestern part of Chippewa County (Albertville area), but in many areas this unit is discontinuous and only 3 to 5 meters thick above the Cambrian bedrock. This material is an important source of aggregate in the region, but clay accumulations make it less useful (higher plasticity) than younger glacial stream sediment in Chippewa County.

River Falls Formation stream sediment is easily distinguished from stream sediment of the Copper Falls Formation by its thick clay-enriched zone, yellowish-red, oxidized color, and a higher concentration of cobbles than distal Copper Falls Formation stream sediment. In addition, the landscape underlain by River Falls stream sediment is more eroded and does not preserve the original outwash plain surface.

The extensive weathering zone and the eroded land surface suggest that the outwash was deposited prior to the Wisconsinan glaciation. Syverson (2004, in press) mapped this outwash as part of the River Falls Formation. The presence of agates in this proximal outwash suggests a strong Superior lobe contribution to this stream sediment, as opposed to a Chippewa lobe influence (Syverson, 2004). If so, ice of the Superior lobe must have covered most of Dunn County, and the adjacent Chippewa lobe reached as far south as the Foster area in southern Eau Claire County (Bement and Syverson, 1995; Syverson, 2004, in press). Syverson (2004, in press) proposed that the extensive River Falls Formation stream sediment may have been deposited in an interlobate junction or a reentrant between the Superior and Chippewa lobes during retreat from their maximum ice-margin positions.

NEXT: Drive 0.5 mile north on the pit access road to 30th Avenue and turn right (east). Drive 0.25 mile on 30th Avenue and turn left (north) on County Road T. Follow County Road T for 1 mile and turn right on State Highway 29 (please note—this is the old two-lane highway, not the four-lane highway that is under construction). Continue east on Highway 29 for 1.5 miles and turn left (north) on Chippewa County Road F; follow it for 4.5 miles and turn right (east) on Chippewa County Road S. Travel east on County Road S for 6.7 miles and turn left (north) on State Highway 124; continue north for 9.9 miles and turn right onto a pit access road (Bischel pit, Milestone Materials). Drive into the pit and park.

STOP 12-7

Hard hats are required

Proximal Copper Falls Formation stream sediment, Bischel pit

Location: T. 30 N., R. 8 W., sec. 5, NW, SW

Bloomer quadrangle; UTM: 627,050E/4,996,200N

Description: This pit displays up to 10 meters of proximal Copper Falls Formation glacial outwash deposited during the Late Wisconsinan glaciation (15,000 to 20,000 years ago). The Chippewa lobe initially extended approximately 400 meters to the west of the Chippewa moraine during Late Wisconsinan time, and then wasted back to the eastern margin of this pit. Hummocky stream and glacial sediment within the Chippewa moraine is located directly east of this pit. Thus, here the prominent Chippewa moraine, formed during the Late Chippewa Phase, coincides with the maximum extent of the Chippewa lobe during the Late Wisconsinan glaciation. This is unlike the area seen yesterday in St. Croix County, where ice extended tens of kilometers beyond the outermost well developed Late Wisconsinan moraine. The large bedrock highland to the southeast (T. 30 N., R. 8 W., secs. 32, 33, 34) might have controlled the maximum extent of the ice, or potentially the ice surged beyond previous ice-margin positions during the Late Chippewa Phase.

Sediment in this pit is typical for proximal glacial outwash of the Copper Falls Formation. In the northeasternmost part of the pit, brown (7.5YR hues), clast-supported gravel and sandy gravel up to 10 meters thick are exposed (Fig. 12.24). Cobbles and boulders up to 40 centimeters in diameter are abundant, and the sand was largely removed in the high-energy fluvial environment. In fact, so much sand was transported away by stream action that the gravel operator at this site initially had to import sand to obtain the proper rock-to-sand proportions for the hot mix facility! A fining-upward sequence in the uppermost 1.5 meters of the northern pit face (Fig. 12.24) represents waning water-flow conditions when most buried ice had melted in the Chippewa moraine region, perhaps thousands of years after the active glacier margin had wasted toward the north and the permafrost conditions ended (Florin and Wright, 1969; Ham and Attig, 1996a). Evidence for ice-contact sedimentation was visible in the easternmost wall of the pit (collapsed beds filled with horizontally bedded to massive sandy gravel), but this area is now slumped.

Newer parts of the pit to the west are changing rapidly at this time. Generally, 4 to 7 meters of clast-supported cobble gravel and sandy gravel are interbedded with gravelly sand and sand. The outwash rapidly fines away from the former ice

Figure 12.24. Coarse-grained proximal stream sediment of the Copper Falls Formation at Stop 12-7. The gravel and sandy gravel fine upward and were deposited within 100 meters of the Chippewa moraine by water flowing from right to left. This gravel-rich sediment is an ideal source of commercial aggregate. Shovel and clipboard at base of outcrop for scale (arrow). From Syverson (in press).

margin, so ice-contact faces have been major gravel exploration targets along this section of Highway 124.

The brown color and lack of pedogenic clay in this Copper Falls Formation outwash is a marked contrast to the yellowish-red, extremely weathered outwash of the River Falls Formation observed at Stop 12-6. The Copper Falls Formation outwash here marks the start of the Wissota terrace, the highest outwash terrace that extends all the way from the Chippewa moraine to the Mississippi River (Andrews, 1965). The Wissota terrace is a prominent feature in Chippewa County. The Chippewa River and its tributaries aggraded as the amount of sediment supplied to the system increased during the Late Wisconsinan glaciation. Syverson (in press) reported that even tributary valleys that start in unglaciated drainage basins are graded to the Wissota terrace level, and these unglaciated valleys do not contain evidence for backflooding. Syverson (in press) proposed that enhanced erosion rates associated with permafrost conditions (Clayton and others, 2001) might have allowed the unglaciated drainage basins to aggrade at rapid rates and remain graded to the Wissota surface.

NEXT: Return to Highway 124 and turn right (north); follow 124 for 0.7 mile, cross the Late Wisconsinan ice maximum position, and turn right (east) on State Highway 64. Travel east through the Chippewa moraine on Highway 64 for 3.3 miles to a point 0.5 mile west of the intersection with Chippewa County Road E. We will stop on the shoulder and persons wishing to photograph the ice-walled-lake plain rim ridge must cross to the north side of Highway 64 very carefully—this is a busy road. The best photograph opportunity is toward the northeast where a barn can be used for scale (Fig. 12.25).

STOP 12-8

Ice-walled-lake plain rim ridge (brief photo stop)

Location: T. 31 N., R. 8 W., sec. 35, SE, NW

Bob Lake quadrangle; UTM: 631,300E/4,998,550N

Description: The relatively flat surface at this stop is an unstable-environment ice-walled-lake plain (Clayton and Cherry, 1967). The former ice-walled lake was surrounded by stagnant ice overlain by relatively thin sediment (Fig. 12.14). Ice beneath the thin sediment melted quickly because it was poorly insulated. In addition, Ham and Attig (1996a, 1997) proposed that unstable-environment ice-walled-lake plains formed after the end of permafrost (approximately 13,000 years ago) when the ice could melt more rapidly. The rapidly melting ice created an unstable, dynamic environment where the ice-walled-lake plains and surrounding hummocks formed low in the landscape. The melting produced numerous streams and gravity flows that transported much coarse- and fine-grained sediment into the lakes. Coarse-grained sediment was deposited in deltas around the outer margins of the lake (Fig. 12.14). These deltas are commonly parts of rim ridges observed around the outer lake-plain margins following lake drainage.

This photograph stop highlights a prominent ice-walled-lake plain rim ridge to the north (Fig. 12.25). It would be difficult to maintain such a steep slope if the sediment was slumping or flowing into the lake. Thus, this sharp-crested ice-walled-lake plain rim ridge probably formed after the lake drained,

Figure 12.25. Unstable-environment ice-walled-lake plain (foreground) with sharp-crested rim ridge in the background at Stop 12-8. Photograph looking northeast from State Highway 64 near Himple Lake (T. 31 N., R. 8 W., sec. 35, SE, NW, Bob Lake quadrangle). The ice-walled lake probably had drained before sediment from the ice surface slumped onto the lake plain and formed the steep rim ridge; from Syverson (in press).

and would contain poorly sorted sediment that slid off the glacier surface.

NEXT: Continue east on Highway 64 for 0.5 mile and turn left (north) on County Road E. Continue northeast on E for 0.4 mile and turn left (north) on 180th Street (Bob Lake Road). Continue on this road as it makes a 90° turn to the west (called 205th Avenue) and another 90° turn to the north (175th Street/Bob Lake Road). After traveling 2.2 miles from County Road E, turn left (west) onto a gravel pit road immediately west of Bob Lake.

STOP 12-9

Ice-walled-lake plain sedimentology, Bob Lake

Location: T. 31 N., R. 8 W., sec. 14, SE, SW

Bob Lake quadrangle; UTM: 631,360E/5,002,300N

Description: This pit exposes several different types of sediment deposited in an unstable-environment ice-walled-lake plain (Clayton and Cherry, 1967; Syverson, in press). In the southern part of the pit, 5.5 meters of deltaic sand to sandy gravel are exposed (Fig. 12.26). The lowermost 2 meters contain sand and sandy gravel foreset beds that dip at a 160° azimuth. These foreset beds display Type A ripple-drift cross lamination. This is overlain by horizontal topset beds 3.5 meters thick that coarsen upward into pebble to cobble gravel. Clasts up to 25 centimeters in diameter are present. Farther to the northwest along the southern face of the pit, the sediment is dominated by fine- to medium-grained sand interbedded with sandy silt layers 1 to 3 centimeters thick. Cut-and-fill structures and draped lamination are commonly observed in this material. This is overlain by sandy pebble gravel. This sediment sequence represents a delta prograding into the ice-walled lake (Syverson, in press).

The northern part of the pit exhibits an 8- to 10-meter-high vertical exposure of reddish-brown (5YR 4/4), sandy loam till of the Copper Falls Formation. The diamicton is quite uniform and typical of the Copper Falls Formation till that covers much of northwest Wisconsin (Table 12.1). The till contains abundant banded iron-formation and volcanic rocks

Figure 12.26. Deltaic foreset bedding in the Bob Lake ice-walled-lake plain at Stop 12-9. Foresets dip southward into the former ice-walled lake. Field notebook and shovel indicated for scale (see arrow).

from the Midcontinent rift region, but Lake Superior agates are not as abundant as in the River Falls Formation sediment in Chippewa County. The till is also very similar to sediment found in hummocks within western Wisconsin.

The topographic reversal process has been proposed as a mechanism to form hummocks as sediment is transported laterally from high areas on the ice surface into adjacent low-lying areas on the ice surface (Clayton, 1967; Johnson and Clayton, 2003). Supraglacial processes such as debris flows, stream flow, and water ponding should cause bedding and grain-size variations vertically and laterally, as well as color differences within hummocks (Clayton and Cherry, 1967; Benn, 1992). Many hummocks in western Wisconsin do contain extremely variable sediment assemblages. However, Ham and Attig (1993, 1996b, 1997) and Johnson and others (1995) have found surprisingly uniform sediment in vertical bore holes through hummocks in Wisconsin. Johnson and others (1995) also observed strong pebble orientations in diamicton within hummocks in Polk and Barron Counties (western Wisconsin), a characteristic more typically associated with deposition directly from glacier ice.

Johnson and others (1995) proposed that the uniform till in hummocks is supraglacial meltout till formed beneath a thick insulating layer of supraglacial sediment. Johnson and others (1995) stated that the thick layer of supraglacial sediment slowed ice ablation to a point where little water was produced, so the supraglacial sediment contained less water and was more resistant to mass movements. Thus, sediment slowly melting out at the ice surface inherited massive, uniform sediment characteristics and pebble fabrics typical of sedimentation directly by glacier ice.

NEXT: Return to 175th Street and turn left (north). Continue 0.6 mile and then the road makes 90° turn to the left (west) and becomes 225th Avenue. Continue west on 225th Avenue for 0.8 mile where the road curves to the north around Big Buck Lake. A large, unvegetated, stable-environment ice-walled-lake plain is located directly north of the road. Continue west on 225th Avenue for 2.6 miles and cross O'Neil Creek. After crossing the creek, we start traveling across the pitted outwash plain and outwash plain that formed at the same time as the Chippewa moraine. Continue west on 225th Avenue for 1.0 mile and turn right (north) on Chippewa County Road AA. Follow County Road AA around numerous turns for 2.5 miles and make a 90° turn to the left (west). At this point there is an old gravel pit south of the road and a more recent gravel pit to the north of the road. These pits mined glacial outwash of the Copper Falls Formation deposited adjacent to the Chippewa moraine. Continue west on County Road AA for 0.5 mile and observe the low-relief, hummocky Chippewa moraine to the north (wooded, trends to the northwest away from the road) and the outwash plain (farm fields) that slope away from the moraine. The outwash plain gradient increases toward the moraine. Turn right (north) on State Highway 40. Follow Highway 40 for 1.7 miles. The highway follows the outermost margin of the Chippewa moraine (right side) and the outwash plain (left side). Turn right (east) on Chippewa County Road M. Follow County Road M for 1.8 miles through the hummocky Chippewa moraine. Turn left (north) at the sign for the Chippewa Moraine Ice Age National Scientific Reserve Visitor's Center. Follow the road to the visitor's center for 0.2 mile to the crest of the hill and park.

STOP 12-10

Formation of the high-relief Chippewa moraine

Location: T. 32 N., R. 8 W., sec. 30, SW, SW

Marsh-Miller Lake quadrangle; UTM: 624,500E/ 5,008,780N

Description: The Chippewa moraine is a prominent landform in Barron, Rusk, and Chippewa Counties, parts of which were described by Mathiesen (1940), who called it the "Inner Morainic System;" Black (1974), who called it the "Bloomer moraine;" and Cahow (1976), Johnson (1986), and Syverson (1998a, b; in press). The view south from the Chippewa Moraine Visitor's Center overlooks a kettle lake (South Shattuck Lake) and high-relief hummocks within the Chippewa moraine. High hills underlain by Cambrian rock are visible on the horizon to the southwest.

The Chippewa moraine is dominated by a 16-kilometer-wide, triangular tract of high-relief hummocks and irregularly shaped lakes located southwest of Flambeau Ridge (Fig. 12.27). The hummocks are circular to elongate, 15 to 30 meters high, 100 to 300 meters in diameter, and spaced 200 to 300 meters apart. The most prominent hills in the Chippewa moraine (such as Baldy Mountain and the Chippewa Moraine Visitor's Center hill) are circular to oval, 700 to 1,000 meters in diameter, 40 to 50 meters high, and are underlain by lake sediment (see map in Syverson and others, 1995). The well for the Chippewa Moraine Visitor's Center penetrated 86 meters of sediment, much of it fine-grained and not useful for supplying water, and it did not encounter bedrock (see Fig. C-6 in Syverson, 1998b). The visitor's center hill is a stable-environment ice-walled-lake plain that has two crests (Syverson and others, 1995; Syverson, in press). This suggests that two ice-walled lakes coalesced as the ice walls confining the lakes slowly melted back, and then more lacustrine sediment was deposited at lower levels within the new ice-walled lake. Attig (1993, p. 14) observed coalesced ice-walled-lake plains in Taylor County.

The Stanley and Perkinstown moraines in eastern Chippewa County are slightly older and lower relief than the Chippewa moraine (Fig. 12.4; Syverson, in press). The Perkinstown moraine, named by Attig (1993, p. 20) in Taylor County, extends westward from Taylor County toward Cornell. It is a low- to moderate-relief (8 to 20 meters) hummocky zone that is up to 9 kilometers wide in a north–south direction. The southerly to southwesterly flowing Chippewa lobe formed the Perkinstown moraine during the Perkinstown Phase (Syverson, in press). This was followed by the Late Chippewa Phase when the Chippewa lobe flow direction changed markedly toward the southwest, truncated older morphologic features associated with the Perkinstown Phase to the east, and formed the high-relief Chippewa moraine.

The difference in morphology between the older, low- to moderate-relief (8 to 20 meters) Perkinstown moraine and the younger, high-relief (15 to 30 meters) Chippewa moraine is striking. Such hummocky moraines are thought to require thick sediment on the ice surface (Gravenor and Kupsch, 1959; Clayton, 1967; Mickelson and others, 1983; Lagerbäck, 1988; Sollid and Sorbel, 1988; Johnson and others, 1995; Ham and Attig, 1996a, 1997; Colgan and others, 2003; Johnson and Clayton, 2003). If the relief of a hummocky moraine is similar to the original sediment thickness (Clayton, 1967, p. 38), then supraglacial sediment at the Late Chippewa ice margin was generally two to three times thicker than supraglacial sediment at the Perkinstown ice margin.

Several mechanisms have been proposed for the accumulation of thick supraglacial sediment in midcontinental areas, as summarized by Johnson and Clayton (2003). These include compression near the ice margin as thinning ice slows (Fig. 12.28; Paterson, 1994, p. 253), compressive ice flow caused by a glacier flowing over permafrost (Attig and others, 1989; Clayton and others, 2001; Johnson and Clayton,

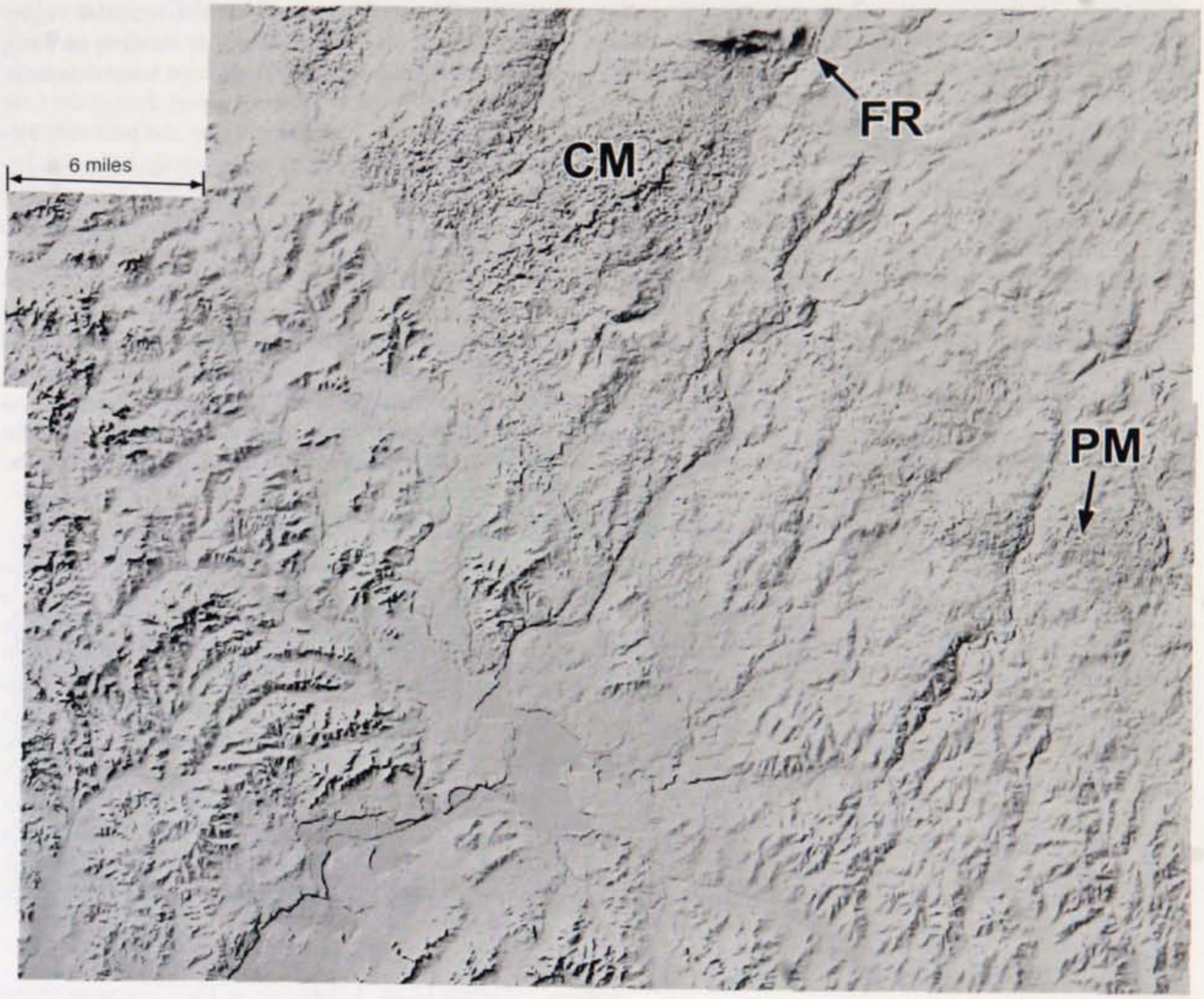


Figure 12.27. Shaded-relief digital elevation map of Chippewa County. The highest-relief portion of the Chippewa moraine (CM) is directly downflow (southwest) of the Flambeau Ridge (FR), the high quartzite knob. The Perkinstown moraine (PM) is indicated, and the incised Paleozoic bedrock surface is clearly visible in the western part of the county; modified from Syverson (in press).

2003), and enhanced compression at the margin of a surging glacier (Hambrey and others, 1996). The Perkinstown and Chippewa moraines formed at approximately the same latitude during permafrost conditions, although at different times, so it seems likely that many of these processes could have been at work in both areas.

The widest, highest-relief portion of the Chippewa moraine is located southwest (directly downflow) from Flambeau Ridge, a 150-meter-high Precambrian quartzite ridge (Fig. 12.27; Syverson, in press). This area also coincides with the thickest glacial sediment in the county (greater than 60 meters; Lippelt, 1988) and a reentrant in the outermost part of the Chippewa moraine (Cahow, 1976; Syverson, in press). Based on these relationships, it seems likely that the high-relief Chippewa moraine is somehow related to Flambeau Ridge (Cahow, 1976; Waggoner and others, 2001). Waggoner and others (2001) and Syverson (in press) proposed that Flambeau Ridge formed a prominent obstacle to ice flow that slowed the ice, enhanced compressive ice flow (Paterson, 1994, p. 254) and the transportation of sediment into the ice, and eventually resulted in thicker-than-normal sediment accumulations on the ice surface. As Flambeau Ridge melted out of the ice, large-scale ice stagnation occurred in the Chippewa moraine region downflow from the obstruction. The thick sediment insulated the underlying ice, and eventually the active Chippewa lobe separated from the stagnant ice mass southwest of Flambeau Ridge (Fig. 12.28). Outwash plains formed sloping away in all directions as this ice mass melted and the hummocky moraine topography developed. Cahow (1976, p. 41, 173) proposed that this section of the Chippewa moraine formed as an interlobate moraine downflow from Flambeau Ridge. However, Waggoner and others (2001) noted that hummocks in this area are largely made of till-like sediment, not the glacial stream sediment that is common in interlobate moraines (Mickelson and Syverson, 1997).

The high-relief Chippewa moraine also might have been formed by different ice-flow conditions than those operating during the Perkinstown Phase of the Chippewa lobe. Attig and others (1998) proposed that the Perkinstown moraine formed during non-surging conditions as ice flowed south–southeast toward the margin (Fig. 12.11). Truncated ice-flow indicators in the middle of the Chippewa lobe region show a marked change in flow direction from southeasterly to southwesterly during the Late Chippewa Phase. Such major changes in ice-flow direction during the last part of the Wisconsinan glaciation have been attributed to surging events (Ham and Attig, 1996b; Attig and Ham, 1997; Attig and others, 1998; Johnson, 2000). Attig and others (1998) recognized that the high-relief Chippewa and Harrison moraines are associated with major, late-stage changes in flow direction of the Chippewa and Wisconsin Valley lobes, respectively (Fig. 12.11). According to Attig and others (1998), the Chippewa lobe surged rapidly toward the southwest over permafrost during the Late Chippewa Phase. The ice froze to the bed near the margin and sliding stopped. This would have quickly reduced ice velocity, strengthened the compressive flow regime, enhanced upward ice flow and erosion, and produced thick sediment accumulations on the ice surface (Johnson and others, 1995; Clayton and others, 2001). The sharp, rather linear eastern boundary of the Chippewa moraine suggests that stagnant, debris-covered ice from the Perkinstown Phase might have prevented ice from flowing into the area east of the Chippewa River during the Late Chippewa Phase, as suggested by Attig (1993) and Syverson (in press).

NEXT: Retrace our path to County Road M and turn left (east) on M. Continue east for 2.7 miles to the intersection with 152nd Street. This intersection is on a circular, unstable-environment ice-walled-lake plain with rim ridges on the western and northeastern edges. Continue east on County Road M for 1.6 miles and turn right (south) on 167th Street. Continue south on 167th Street for 0.9 mile and turn right (west) on 260th Avenue. Drive west on 260th Avenue for 0.8 mile. As the road ascends a steep hill and curves toward the left, we are climbing the ice-contact face for the Plummer Lake ice-walled-lake plain. At the flat crest of the hill, a grassy parking lot for the Ice

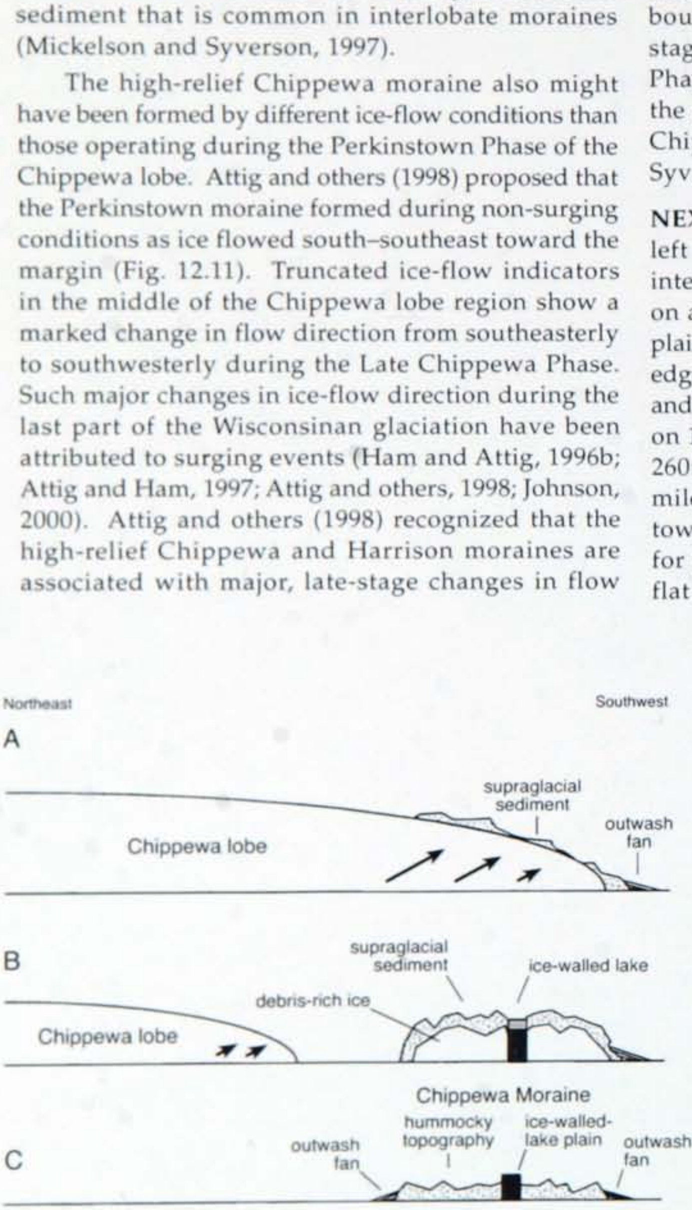


Figure 12.28. Formation of the Chippewa moraine; modified from Ham and Attig (1997).

A. The Chippewa lobe margin remains in the same position for a period of time. Compressive ice flow occurs as ice slows at the margin. This carries sediment from the base of the glacier into the ice where the sediment later melts out at the ice surface. Meltwater deposits sand and gravel in the outwash plain sloping away from the ice margin.

B. Debris-covered ice melts slowly, stagnates, and eventually separates from the active ice of the Chippewa lobe. Low areas fill with water (ice-walled lakes) or sediment. Ham and Attig (1997) proposed that this occurred during permafrost conditions.

C. Stagnant ice melts, perhaps over a period of several thousand years, after permafrost conditions end. Meltwater stream sediment is deposited in outwash plains sloping away from the moraine in all directions. Hummocky topography and ice-walled-lake plains mark the moraine.

Age National Scenic Trail is located on the right (north) side of the road. The entrance is narrow and obscured by dense brush on either side.

STOP 12-11

Ice-walled-lake plain sedimentology and morphology, Plummer Lake

Location: T. 32 N., R. 8 W., sec. 27, SW, SW

Bob Lake quadrangle; UTM: 629,300E/5,008,500N

Description: This stop examines the geomorphology and sedimentology of a classic unstable-environment ice-walled-lake plain (Figs. 12.14A, 12.29; Clayton and Cherry, 1967). The flat, elevated nature of the Plummer Lake ice-walled-lake plain is visible from the southwest shore of Dumke Lake. The oval plain is approximately 1.1 kilometers long, 0.8 kilometer wide, and rises approximately 24 to 31 meters (80 to 100 feet) above the surrounding kettle lakes (Fig. 12.29). The ice-walled-lake plain contains sediment of the Copper Falls Formation. Hand borings around the outside of this ice-walled-lake plain reveal poorly sorted, silty, gravelly sand interbedded with silty fine-grained sand. This sediment was deposited in a nearshore lacustrine environment with fluctuating energy levels. A drill hole near the center of the plain penetrated 23 meters of laminated silt and silt loam (Figs. 12.29, 12.30, 12.31; Syverson, 2000, in press). This drill hole, as well as domestic well-logs in Chippewa County, suggest that ice-walled-lake plain sediment is approximately as thick as the lake plain is high (Fig. 12.17B). The linear zone of hummocky sandy gravel (stream sediment) to the northeast of the ice-walled-lake plain may mark an englacial outlet for the former ice-walled lake (Fig. 12.29; Syverson, in press).

The Plummer Lake ice-walled-lake plain is incised by sinuous, 3- to 6-meter-deep, v-shaped stream valleys that head at the top of the ice-contact face overlooking Dumke and Plummer Lakes (Cahow, 1976; Syverson, in press). Several unstable-environment ice-walled-lake plains in Chippewa County are incised by similar channels (Cahow, 1976, Fig. 25, pl. 1; Syverson, in press). These formed soon after the ice-walled lakes drained and water flowed directly from the surrounding ice masses onto the exposed ice-walled-lake-plain sediment and rapidly eroded the material.

NEXT: Turn right and continue west on 260th Avenue for 0.8 mile to the intersection with 160th Street. A nice view of the high, flat, Plummer Lake ice-walled-lake plain is visible northeast of Dumke Lake. Turn left on 255th Avenue. Drive southwest for 2.5 miles through the hummocky Chippewa moraine. Turn left (south) on 137th Street. Continue 0.5 mile south on 137th Street and turn right (west) on 245th Avenue. Drive west for 0.75 mile to County Road AA and continue straight (west) on County Road AA. This brings us back to the Chippewa moraine boundary that we saw previously and we will briefly retrace our route. Continue west on County Road AA for 3.0 miles and turn left (south) on County Road F. This

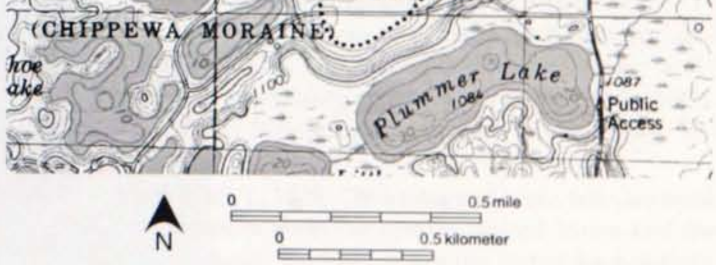


Figure 12.29. Part of the Bob Lake quadrangle showing the Plummer Lake unstable-environment ice-walled-lake plain. The outermost edge of the ice-walled-lake plain is shown by dots. Stop 12-11 coincides with a drill hole location (marked by the star) where drilling penetrated 23 meters of lake sediment over gravity-flow sediment (see sediment log in Figure 12.30). The dashed line marks several sinuous, dry stream channels that head at the top of a steep ice-contact face. These channels formed as meltwater flowed directly off glacier ice onto the lake plain soon after lake drainage. Arrowheads northeast of the ice-walled-lake plain indicate the start of a linear zone of hummocky sandy gravel (stream sediment) that may mark an outlet for the former ice-walled lake. Contour interval 10 feet (3 meters), original scale 1:24,000.

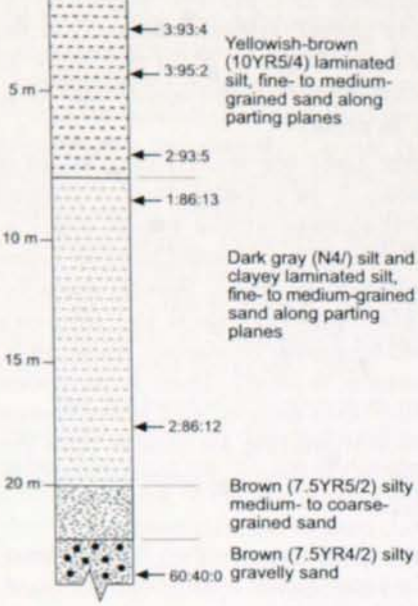


Figure 12.30. Offshore sediment log for the Plummer Lake ice-walled-lake plain, Chippewa Moraine Ice Age National Scientific Reserve, Wisconsin (see Fig. 12.29 for location). Sand:silt:clay ratios are indicated at the appropriate sample depths. The laminated silt and clay settled from suspension in the ice-walled lake, and the two lower units may represent gravity-flow sediment. The ice-walled-lake plain rises 25 to 34 meters above surrounding kettles, so the lake sediment is nearly as thick as the landform is high; from Syverson (in press).

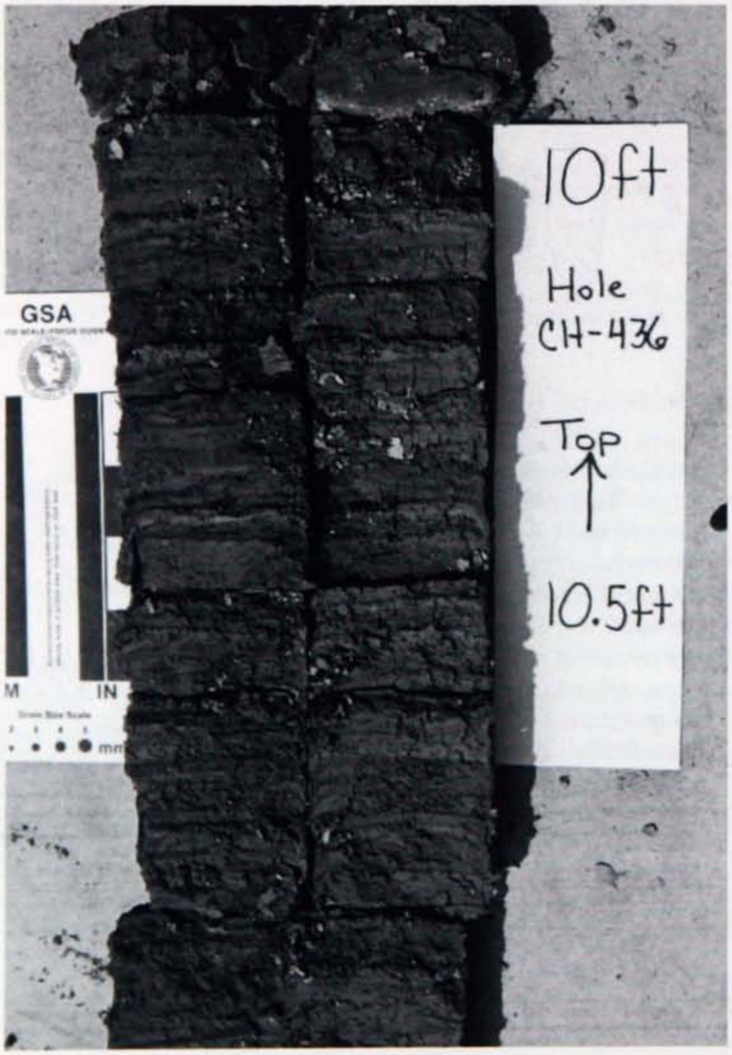


Figure 12.31. Laminated silt and silt loam deposited in an offshore environment, Plummer Lake ice-walled-lake plain, Chippewa Moraine Ice Age National Scientific Reserve (drill hole depth 10 feet [3 meters]), see Fig. 12.29 for location). From Syverson (in press).

highland has thin (3 to 4 meters) River Falls Formation till over Cambrian sandstone. Continue south on County Road F for 6.1 miles to the intersection with State Highways 40 and 64. Drive straight across the intersection (south) onto Highway 40. Follow Highway 40 for 2.0 miles through the city of Bloomer and cross U.S. Highway 53. Continue west and south on Highway 40 for 26 miles through Colfax to the intersection with I-94 east of Menomonie. Take I-94 west to the Twin Cities.

END OF TRIP

ACKNOWLEDGMENT

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