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GUIDEBOOK 21



FIELD TRIP GUIDEBOOK FOR SELECTED GEOLOGY IN MINNESOTA AND WISCONSIN

Lori Robinson, Editor



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FIELD TRIP GUIDEBOOK FOR SELECTED GEOLOGY IN MINNESOTA AND WISCONSIN

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On the cover: Upper: Ordovician fossils collected near the Mississippi River in St. Paul, Minnesota (Field Trips 1 and 6), photo by Lori Robinson
Lower: Neoarchean Soudan Iron Formation (Field Trip 2), photo by Mark Jirsa

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GENERAL INFORMATION

Participants should be prepared for weather of any type. Dry, mild weather will be the goal, but bring rain gear and a dry pair of socks. Insect repellent should be close at hand, too.

UTM coordinates are provided in NAD 83, zone 15. Section subdivisions read from smallest to largest quarter; for example, "NW, SE, SW" should be read "NW quarter of the SE quarter of the SW quarter."

All quadrangles listed refer to the U.S. Geological Survey 7.5-minute series.

ACKNOWLEDGMENTS

The editor thanks Philip Heywood for his assistance with figure preparation and Lori Day for her assistance word processing.

NOTE ON MEASUREMENTS USED IN THIS GUIDE

Although the metric system is preferred in scientific writing, certain measurements are still routinely made in English customary units; for example, distances on land are measured in miles and depths in drill holes are measured in feet. Both metric and English customary values appear in this guidebook. To assist readers, conversion factors for some of the common units of measure are provided below.

English units to metric units:

To convert from	to	multiply by
inch	millimeter	25.40
inch	centimeter	2.450
foot	meter	0.3048
mile	kilometer	1.6093

Metric units to English units:

To convert from	to	multiply by
millimeter	inch	0.03937
centimeter	inch	0.3937
meter	foot	3.2808
kilometer	mile	0.6214

FIELD TRIP 1

Saturday, May 14

THE GEOLOGY OF THE MISSISSIPPI RIVER VALLEY—TWIN CITIES REGION: USING AN URBAN RIVER FOR INQUIRY-BASED EARTH SCIENCE EDUCATION

Leaders

Karen Campbell, National Center for Earth Surface Dynamics
Kent Kirkby, University of Minnesota

INTRODUCTION

The Mississippi River is an easily recognized feature of maps of North America, and stories about it fill American history books. Many students learn in school that the river begins as a trickle small enough to cross in one step in northern Minnesota, and ends as a wide and mighty body of water at a delta in the Gulf of Mexico. Some may learn that along the way, it drains all or parts of 31 states and 2,350 square miles. Many are familiar with colorful tales of western expansion across the river and the romantic era of steamboat travel as immortalized by Mark Twain. Less familiar is the river's geologic history, which in many ways is an even more dramatic one, in which natural physical forces created vast changes in the landscape. Exciting chapters in this geologic story unfolded in the Twin Cities area and can still be read from the record preserved in the local bedrock.

Three main bedrock units are exposed in the Mississippi River valley in the Twin Cities area. The oldest is the St. Peter Sandstone, composed almost exclusively of quartz grains, very poorly cemented together. This sandstone is quite resistant to chemical weathering, but easily eroded by physical processes. A clay layer, called the Glenwood Formation, overlies the St. Peter Sandstone, and is easily erodible by both chemical and physical processes. When exposed, this layer is often covered by vegetation, even when the other units are not, and can therefore be difficult to see. The "top" layer, located above the Glenwood Formation, is the Platteville Formation, a limestone containing many plant and animal fossils and traces of the burrows and tracks of animal pathways. The Platteville Formation is more resistant than either of the other units exposed in the valley, but contains many vertical cracks or "joints." As the river flows over the Platteville Formation, water cuts down through these joints. The softer layers below the Platteville Formation are then eroded by the water. With its supporting base gone, large chunks of the Platteville Formation break away, waterfalls develop, and the process of cutting deep gorges speeds up (Fig. 1.1).

These three bedrock units formed during the Ordovician period, some 460 million years ago. They were deposited by seas that covered the area, the sequence of sand, clay, and carbonate materials being deposited as the sea level slowly rose over time. Fossil plants and animals found in these sedimentary rocks give us clues to the climate; Minnesota lay in a warm equatorial region during this period. As time passed, successive layers of ocean deposits buried one another, the rocks cementing together under the accumulated weight. These three units represent just part of a thick sequence of nearly flat-lying layers of alternating sandstones, limestones, dolostones, and shales that resulted, blanketing much of what we now think of as the Midwest.

Why then did early explorers write of mountains on the upper reaches of the Mississippi River? Long after the depositional events that shaped these sedimentary rocks occurred, recurring massive glaciers covered much of North America. As the glaciers began to melt, shrinking northward, vast quantities of meltwater deepened existing river channels and cut new ones. This process occurred several times. About 11,000 years ago, as the last glaciers retreated north into Canada and Minnesota became mostly free of ice, the force of meltwater from the north helped the Mississippi carve "mountains" out of the nearly flat sediments representing the floors of ancient oceans.

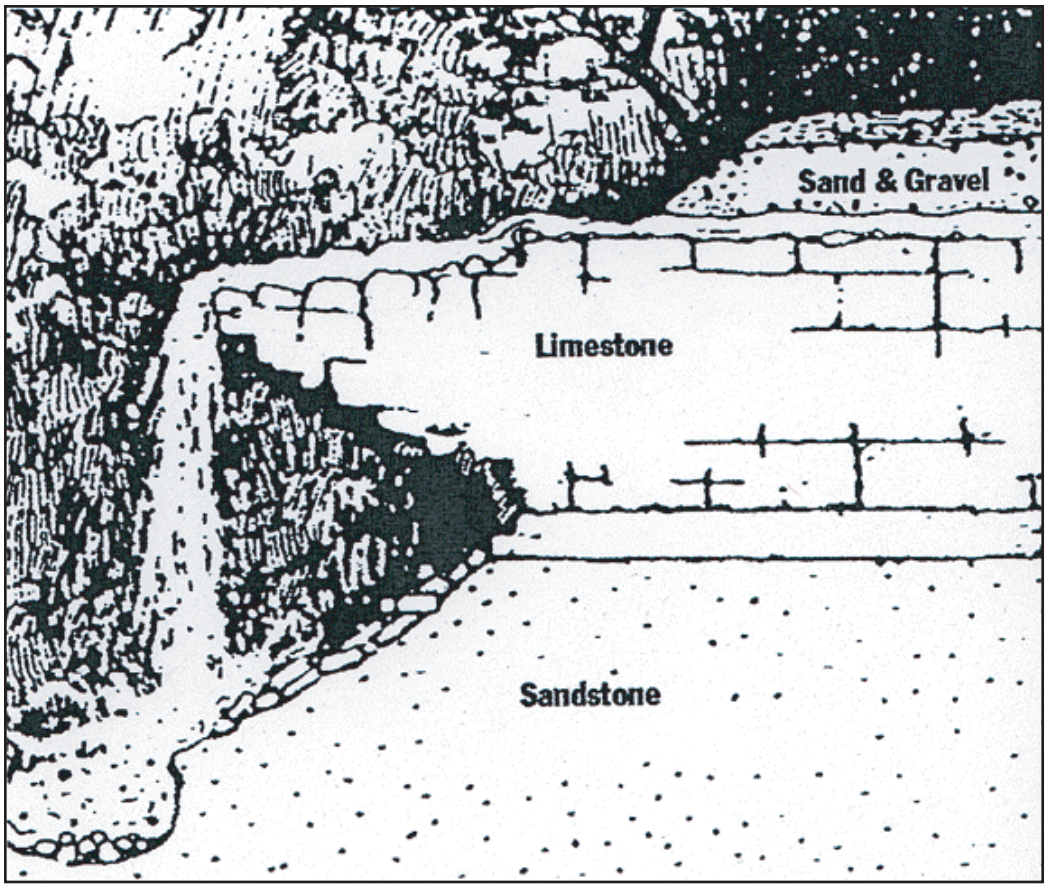


Figure 1.1. Typical development of a waterfall similar to those we will see today (Institute for Minnesota Archaeology, 1999).

FIELD TRIP STOPS (Fig. 1.2)

This field trip is designed as an inquiry activity, exploring the clues in the landscape and the rocks that can reveal the river's history and that of the rocks it flows over and through. Therefore, in the log that follows, sample questions for exploring that landscape are included.

DIRECTIONS: Begin at Mounds Park Overlook, Earl Street at Mounds Boulevard in St Paul.

STOP 1-1

Mounds Park Overlook

Location: Earl Street at Mounds Boulevard, St. Paul

Description: This stop offers a panoramic view of the Mississippi River as it flows through St. Paul. Note the size and shape of the river valley. Think about whether the river we see today could have carved this valley.

NEXT: From Stop 1-1, take Mounds Boulevard north to Kellogg Boulevard, drive west to the Wabasha Street Bridge (Stop 1-2).

STOP 1-2

Wabasha Street Bridge

Location: Kellogg Boulevard and Wabasha Street North, St. Paul

Description: This stop offers another view of the river valley from above. Again, note the size and shape of the valley and think about whether today's river could have carved such a wide channel. Look below the bridge to see all three bedrock units exposed.

STOP 1-2A (Optional)

Science Museum of Minnesota

Location: 120 Kellogg Boulevard, St. Paul

Description: The Science Museum's main floor contains several interesting galleries related to the river. Inside the front door, in the lobby area, the

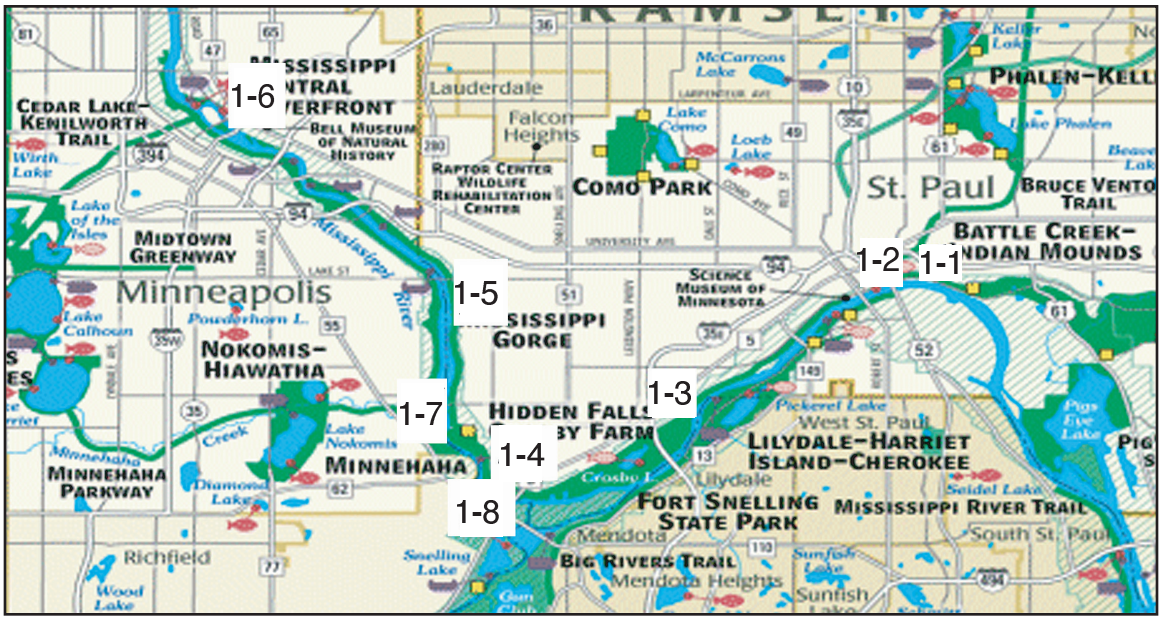


Figure 1.2. Map of field trip stops (Metropolitan Council, 2003).

Mississippi National River and Recreation Area maintains a free visitors' center with exhibits, videos, and interpreters. The center's entrance features a floor-mounted aerial photo of the Mississippi River flowing through the Twin Cities. Inside the Museum, the Mississippi River Gallery overlooks the river and offers the visitor many ways to explore the human and natural history of the river. From the Elements Cafe terrace at the top level of the building, another good vista of the river valley is visible. Finally, on the river side of building, on level 1, the Science Museum's outdoor Big Back Yard, open May through September, offers multiple opportunities to explore natural and engineered river landscapes through interactive exhibits, maps, and miniature golf.

NEXT: Continue along Kellogg Boulevard, turning left onto Eagle Cliff Road just west of the Science Museum of Minnesota. Turn right onto Shepard Road at the traffic light and proceed to Crosby Regional Park. Turn into the park for Stops 1-3 and 1-4 (Fig. 1.3).

STOP 1-3

Crosby Regional Park

Location: Shepard Road at Davern Street; enter the park, follow the road down the hill to the left and park. Walk back up the road to the cave.

Description: At this stop, we can examine the St. Peter Sandstone and look for fallen blocks of Platteville Formation limestone. Walk into the entrance to the cave. Caves have served many purposes along the bluffs in the Twin Cities region. Are the caves a natural feature?

STOP 1-4

Two Rivers Overlook

Location: The west end of Crosby Road, in Crosby Park

Description: This stop offers a view from above of the spot at which the Mississippi River, flowing from the north, joins the Minnesota River, flowing from the northwest. This is a good spot to begin looking for differences in the two river valleys, as we now begin following the Mississippi northward.

STOP 1-4A (Optional)

Ford Dam

Location: East River Road, just south of the Ford Parkway bridge

Description: Finished in 1917, Lock and Dam #1 on the Mississippi River, known locally as the "Ford Dam," provides hydroelectric power to the Ford Motor Company situated on the St. Paul side of the River. The lock is part of the navigation system that allows barges and pleasure boats to travel the relatively shallow Mississippi River north to Minneapolis. The Ford Motor Company chose this site for another reason in addition to the hydroelectric power potential of the river. What commercial uses

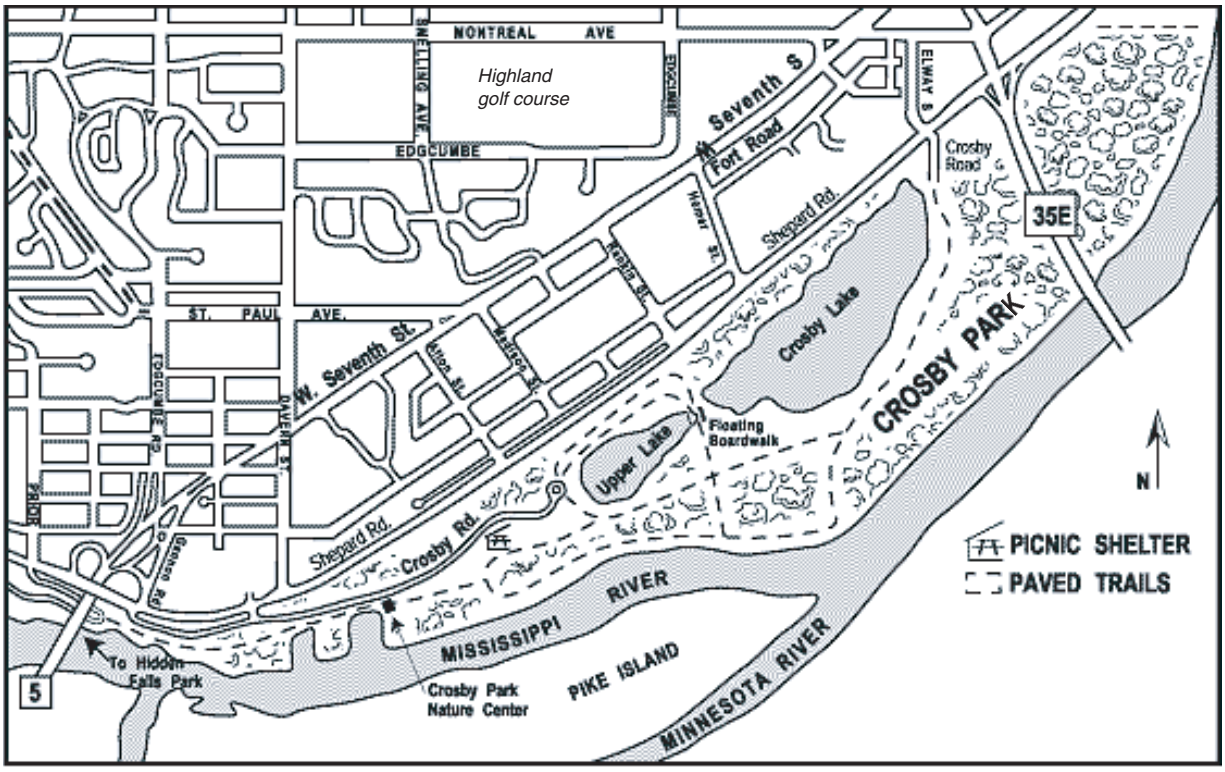


Figure 1.3. Map of Crosby Regional Park (St. Paul Parks and Recreation, 2005).

might an automotive company have for the local bedrock?

NEXT: From Crosby Park, continue briefly on Shepard Road to East River Road and turn right onto East River Road. Depending upon the time of year (summer vegetation hides many of the outcrops), many turnouts along the way offer opportunities to view the river, including one at the Ford Dam (or Lock and Dam #1, described above). At Summit Avenue, a small park is the location of Stop 1-5.

STOP 1-5

Summit Avenue Monument

Location: Summit Avenue and East River Road, St. Paul

Description: Below the monument, look for the contacts between the three rock units in this area. Small fossils can often be found in the soft banks directly north of the monument. Can you tell what rock unit these fossils are from? Typical fossils found here include crinoids and bryozoans. Do they resemble any life forms we find in the river today? Looking across the river, can you conclude anything about the rock formations on the other side from the shape of the bank or any other observations you can make?

NEXT: From here, follow East River Road until it ends at the University of Minnesota. Follow 14th Street north to Fourth Street SE. Turn left, following 4th Street SE to 3rd Avenue SE. Turn left on 3rd Avenue SE to St. Anthony Falls Laboratory at the base of 3rd Avenue SE (Stop 1-6).

STOP 1-6

St. Anthony Falls Laboratory

Location: 2 3rd Avenue SE, Minneapolis. From the parking lot, walk across the channel of water toward the laboratory building; at the small parking lot there, stop to observe the river.

Description: The retreat of glacial River Warren Falls, fed by meltwater from glacial Lake Agassiz to the north and west, rapidly cut through the local bedrock, splitting in two when it encountered the Mississippi River at Fort Snelling (Fig. 1.4). It retreated rapidly up the gorge now visible in Minneapolis. By the time Father Hennepin described the falls the Native inhabitants had known as HahaTanka (Dakota) or Kakabika (Ojibwe) in the 1600s, and renamed them the Falls of St. Anthony for his patron saint, Anthony of Padua, the falls were slightly south of where they are today (Fig. 1.5). They retreated another short distance to their current location before early settlers

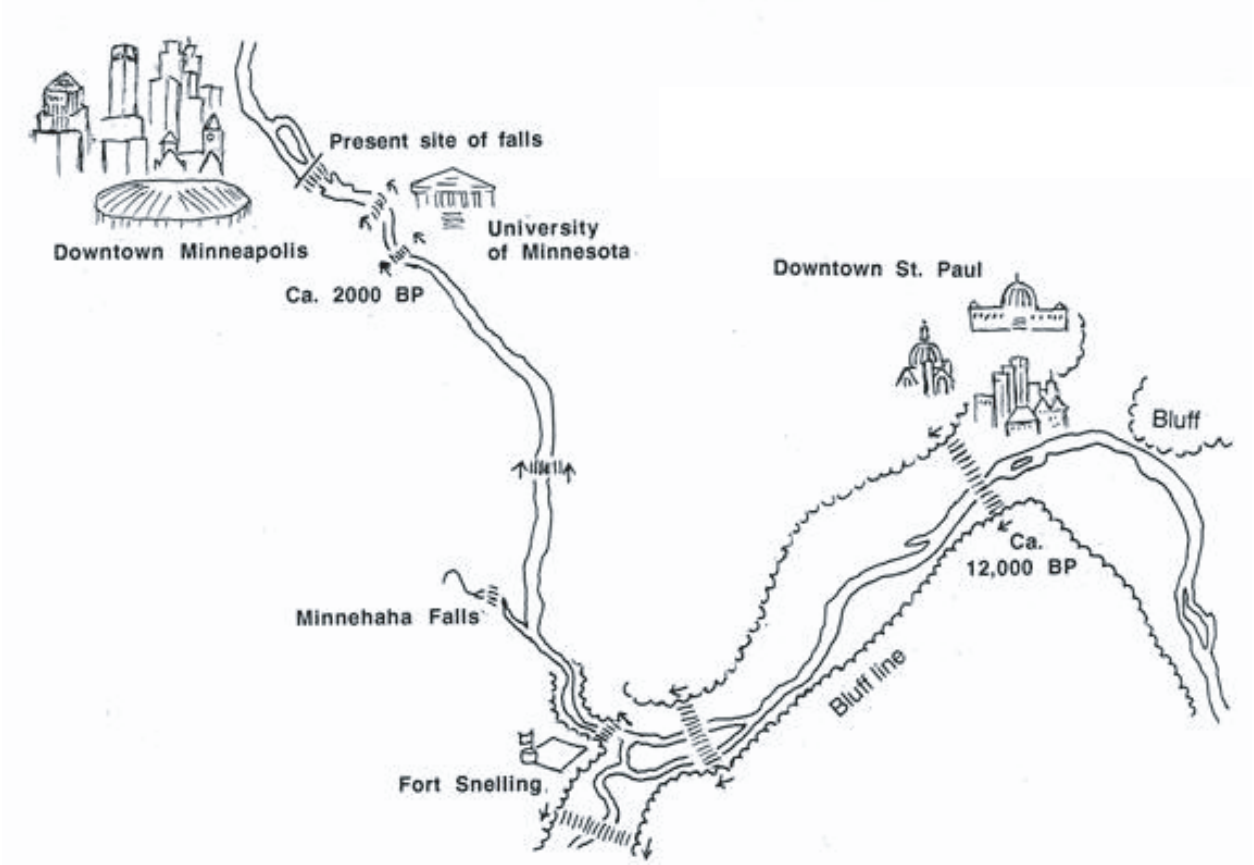


Figure 1.4. Illustration showing the retreat of glacial River Warren (Institute for Minnesota Archaeology, 1999).



Figure 1.5. The Falls of St. Anthony, painted by Henry Lewis as it looked in the 1840s (Institute for Minnesota Archaeology, 1999).

built various structures to stabilize the falls and utilize their power for lumber and flour milling, and later, hydroelectric power. The current concrete apron and lock were finished in the 1960s.

Look at the Falls of St. Anthony, as seen from the laboratory. Is this a natural waterfall? Why or why not? How high is the falls? Has it always been here? Will it stay here? Why is Minneapolis called the "Mill City"? What is the purpose of the channel of water we crossed to access the view? What are the walled, grassy areas between the parking area and the falls? Why is there a laboratory here? Is the bedrock here similar to what we have seen so far today?

NEXT: From St. Anthony Falls Laboratory, go north on 3rd Avenue SE to University Avenue, turn right (east) to 14th Street, then right into campus and rejoin East River Road. Travel south to the Franklin Avenue Bridge. Turn right and cross the bridge and continue south on West River Road to 46th Avenue. At 46th, follow signs to Minnehaha Park (Stop 1-7). From the parking lot, walk down to the pavilion, then down the stairs to the waterfall. From there, follow the footpath to the first bridge. Cross the bridge and observe the valley-like area in front of you.

STOP 1-7

Minnehaha Falls (Fig. 1.6)

Location: 4801 Minnehaha Avenue, Minneapolis

Description: As you walk down to the waterfall, note the rocks in this wall. Are they from this area? Why

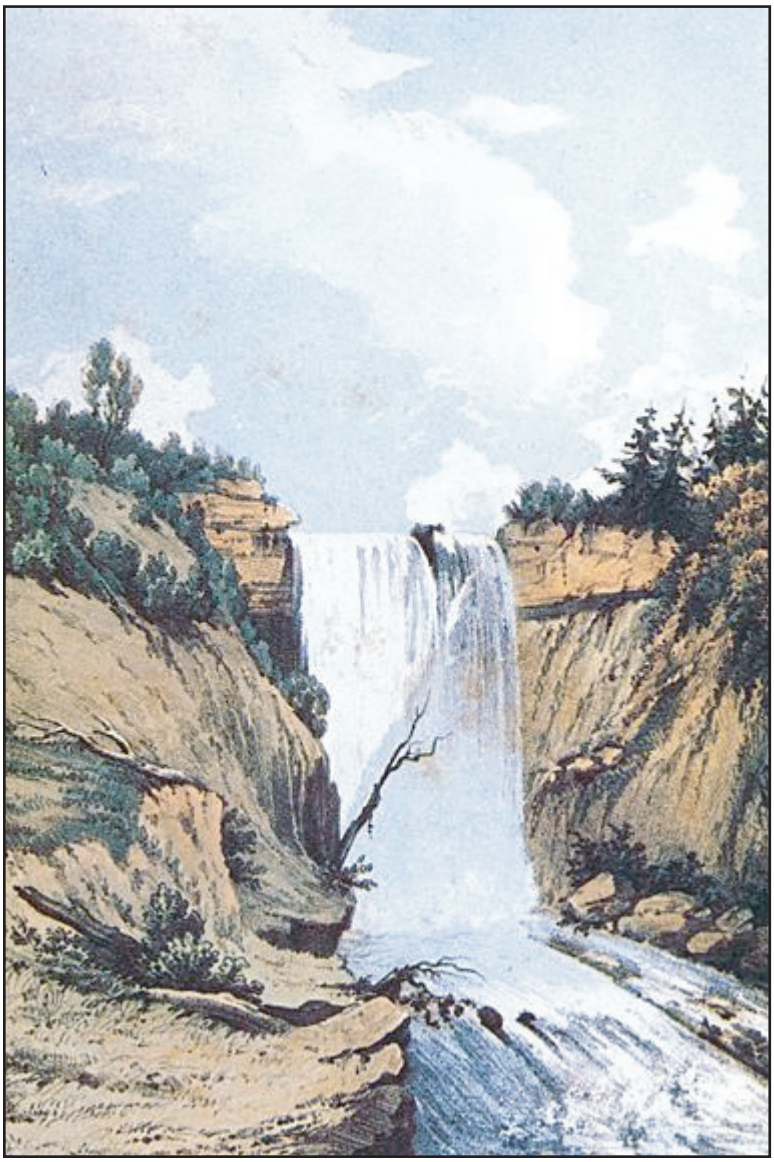


Figure 1.6. Minnehaha Falls as painted by Henry Lewis circa 1840 (Institute for Minnesota Archaeology, 1999).

or why not? How might they have gotten here, if they are local? Looking at the waterfall, what can you conclude about the rocks it exposes? Have we seen them before? As you continue along the path, stop to note the contacts between rock layers. What rocks do we see exposed in the riverbed? After you cross the footbridge, spend some time observing the grassy valley. How did this form? What is the hill to the east?

NEXT: Climb the stairs back up the bank and return to the parking lot. Take 46th Avenue west to Hiawatha Avenue/State Highway 55. Turn left (south) on Highway 55 and follow it as it turns eastward; watch for the entrance to Fort Snelling State Park at State Highway 5 and Post Road. Enter the park and follow the park road to the visitor center at the end of the road.

STOP 1-8

Fort Snelling State Park

Location: Highway 5 and Post Road in St. Paul

Description: If time allows, a hike up the trail toward Minnehaha Park offers additional opportunities to examine the local bedrock closely. A trail onto Pike Island leads to the confluence of the Minnesota and Mississippi Rivers, the location viewed in Stop 1-4. Inside the visitors' center are several dioramas of the area and an explanation of the retreating waterfall.

REFERENCES

Metropolitan Council, 2003, Regional parks–North East map: <http://www.metrocouncil.org/parks/map2htm>.

Institute for Minnesota Archaeology, 1999, From site to story: The upper Mississippi's buried past: <http://www.fromsitetostory.org/tcm/tcmintro.asp>.

St. Paul Parks and Recreation, 2005, Crosby Regional Park: <http://www.stpaul.gov/depts/parks/userguide/crosby.html>.

FIELD TRIP 2

Tuesday, May 17 – Wednesday, May 18

CLASSIC PRECAMBRIAN GEOLOGY OF NORTHEAST MINNESOTA

Leaders

Mark A. Jirsa and James D. Miller, Jr., Minnesota Geological Survey

INTRODUCTION

This field trip presents the great diversity of Precambrian rock types in northeastern Minnesota using some of the most illustrative and accessible outcrops. The field stops include rocks ranging in age from Archean to Mesoproterozoic (Fig. 2.1 and Table 2.1). For the most part, the stops are presented in geographic rather than geochronologic order. Temporal settings can be deduced from the various figures, descriptions, and Table 2.1.

Three topical areas are represented:

- *Virginia horn*—Stops 2-1 to 2-7. Archean greenstone-granite terrane and Paleoproterozoic iron-formation and associated strata of the Animikie Group.
- *Tower–Soudan*—Stops 2-8 to 2-10. Archean graywacke and iron-formation.
- *North Shore and Duluth*—Stops 2-11 to 2-19. Mesoproterozoic volcanic, sedimentary, and intrusive rocks of the Midcontinent rift, and Paleoproterozoic deformed graywacke-slate of the Animikie Group.

Although we consider many of these to be "classic" outcrops, it should also be noted that—because of access and location—these are only *some* of the classics; many instructive outcrops are

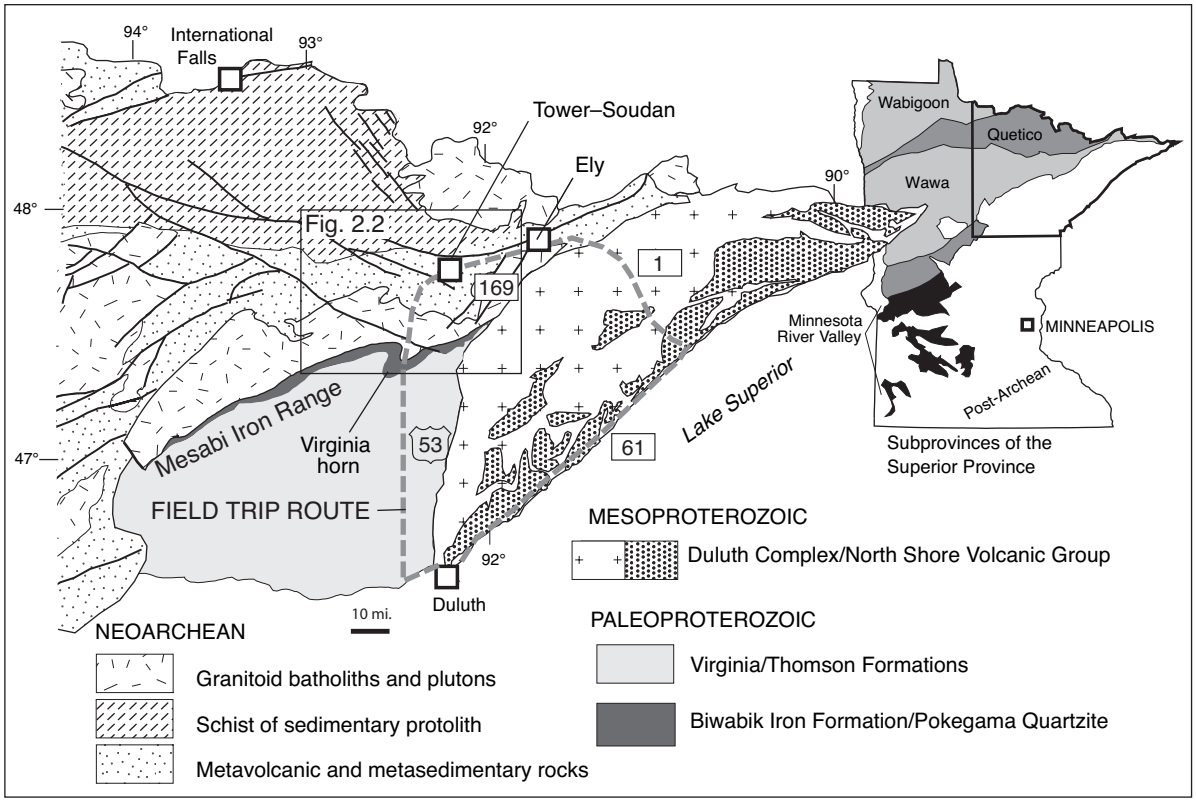


Figure 2.1. Generalized geologic map of northeastern Minnesota showing field trip route.

Table 2.1. Stratigraphic and temporal classification of Precambrian rocks in northern Minnesota (modified from Sims, 1972). Ages are referenced in the text. Unconformities are represented by both dashed and solid horizontal lines. Numbered circles denote field trip stops in their approximate chronostratigraphic order.

FORMATION/GROUP/SUCCESSION/SEQUENCE		INTRUSIONS	EVENT/SETTING	ERA
				900 Ma
Keweenawan Supergroup	Hinckley Sandstone Fond du Lac Sandstone North Shore Volcanic Group (11) (12) (14) (15) Puckwunge/Nopeming sandstones (18)	Duluth Complex (16) (17) Beaver Bay Complex (13) Diabase dikes (19)	**Midcontinent rift**	MESOPROTEROIZOIC
				1,600 Ma
Animikie Group	Virginia/Thomson/Rove Formations (19) Biwabik/Gunflint Iron Formations (~1,880 Ma) (6) Pokegama Quartzite (5) (4)	K-K diabase dikes	**Penokean orogeny**	PALEOPROTEROZOIC
				2,500 Ma
Superior Province — Quetico subprovince		Vermilion Granitic Complex; Giants Range batholith (7)	**D_3 Deformation** **D_2 Deformation** (~2,680 Ma)	
Superior Province — Wawa subprovince	(3) Knife Lake Group/Midway/Ogishke conglomerate (< 2,690 Ma) Newton Lake Formation/Bass Lake sequence Lake Vermilion Formation (8) (9) (1) (2) Mud Lake sequence/Ely Greenstone (~2,720 Ma) (Upper, Soudan iron-formation, and Lower members) (10)		**D_1 Deformation**	NEOARCHEAN

comparatively less accessible. It is likely that we have omitted some favorites, for which we apologize; however, other fine outcrop descriptions can be found in field trip guidebooks in the accompanying references, including guides of the Institute on Lake Superior Geology (notably those of year 2004), the Minnesota Geological Survey Guidebook Series, and the Geological Society of America guidebooks (for example Biggs, 1987).

FIELD TRIP STOPS

In the field trip locations given below, the small map insets that accompany each description are taken from the U.S. Geological Survey 7.5-minute topographic quadrangles listed with each stop.

VIRGINIA HORN AREA
Stops 2-1 to 2-7
Archean greenstone-granite and Paleoproterozoic Animikie Group

Geologic setting

The term "Virginia horn" is applied to an area near the town of Virginia where the generally east-trending, Paleoproterozoic Biwabik Iron Formation makes an abrupt bend to the southwest, creating a horn shape (Figs. 2.2 and 2.3). The iron-formation unconformably overlies Neoarchean bedrock exposed within an uplifted, wedge-shaped block.

The Archean rocks are part of the Wawa subprovince of the Superior Province, and are similar in most respects to other greenstone-granite terranes of the subprovince. Geochronologic data are sparse; however, the similarity with other terranes implies depositional and intrusive ages on the order of 2,720 to 2,670 Ma. The supracrustal rocks are subdivided into northern and southern panels on the basis of metamorphic grade and deformation style. The northern panel, immediately south of the Giants Range batholith, contains intensely lineated, amphibolite-grade schist having volcanic, intrusive, and clastic protoliths (Minntac sequence). The southern panel contains lithologically and stratigraphically similar rocks that were metamorphosed to much lower grades, ranging from prehnite-pumpellyite to low greenschist (Mud Lake sequence). The two panels are separated by the east-trending, post-metamorphic, Laurentian fault. The metamorphic cleavage-forming event in both panels was the second (D_2) of three

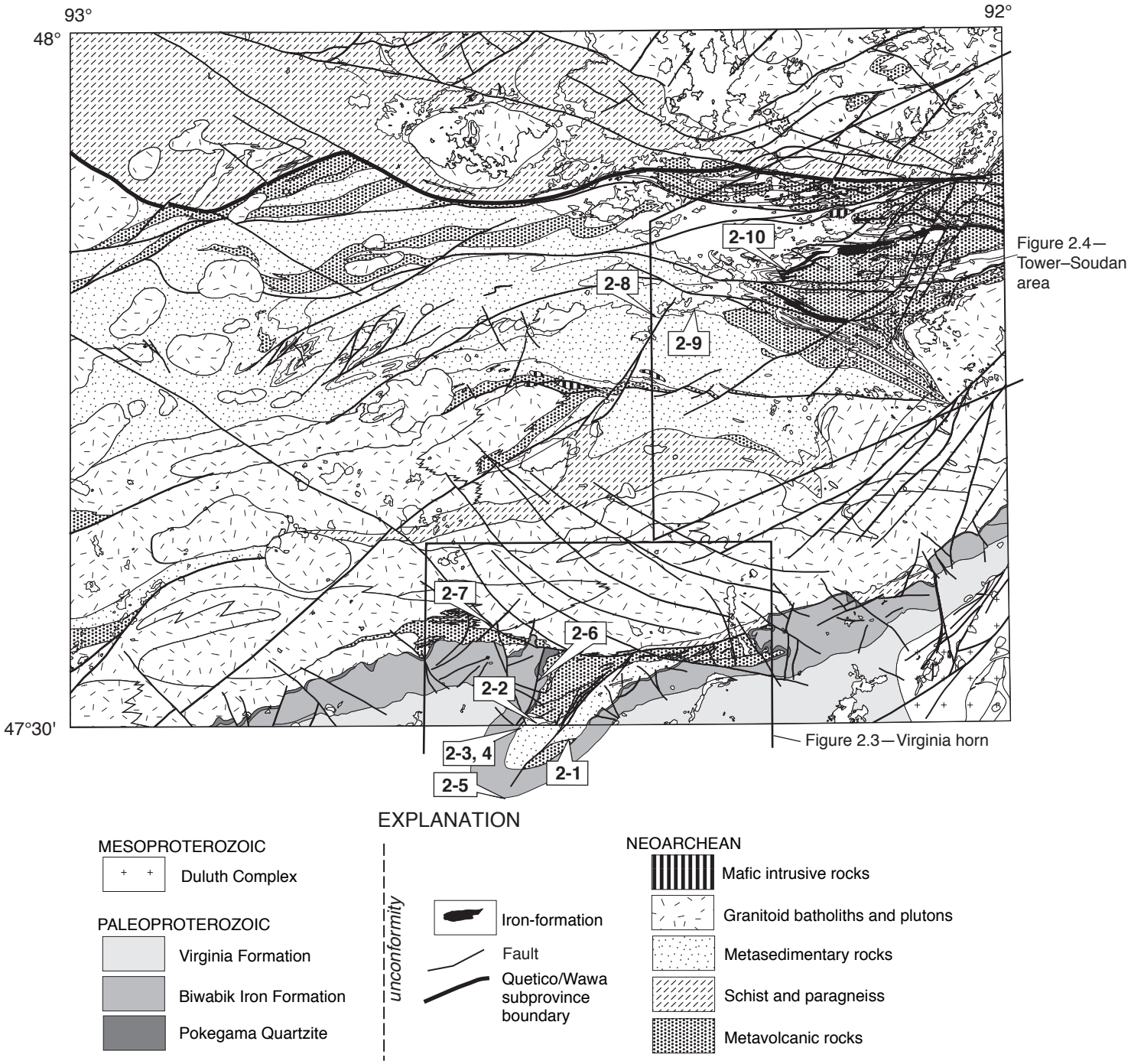


Figure 2.2. Simplified geologic map of the Vermilion Lake 30' x 60' quadrangle (modified from Jirsa and Boerboom, 2003a) showing the location of field trip Stops 2-1 to 2-10. Outlines show location of additional figures.

deformations—no metamorphic effects are recognized from the other two deformation events. The first (D_1) involved upright folding, soft-sediment deformation, and complex faulting. Strata of the southern panel form a broad, southwest-plunging, D_1 syncline that is cored by graywacke, slate, and minor felsic tuff, and has outer limbs of calc-alkalic and tholeiitic volcanic strata. All of these rocks were cut by variably porphyritic, felsic intrusions prior to D_2. The syncline is bisected by a fault- and unconformity-bounded, alluvial fan-fluvial-volcanic succession inferred to represent largely subaerial deposition in localized pull-apart basins (the Midway sequence). The third deformation event (D_3) produced localized semi-brittle

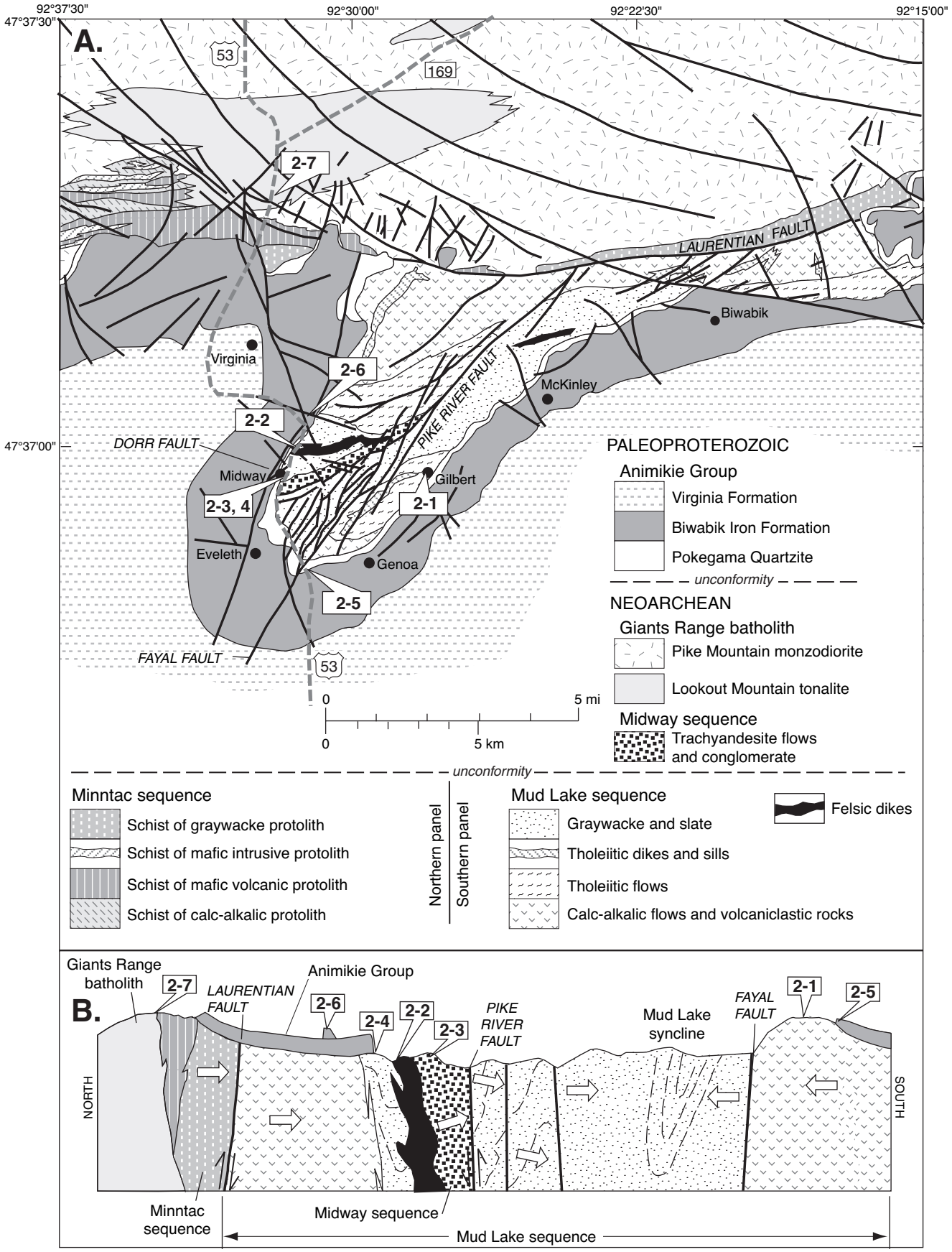


Figure 2.3. Generalized geologic map (A) and schematic cross-section (B) of the Virginia horn area (modified from Jirsa and Boerboom, 2003b) showing details of field trip Stops 2-1 to 2-7. Block arrows indicate general direction of stratigraphic facing.

crenulation of D_1 and D_2 structures, and selective reactivation of earlier-formed faults.

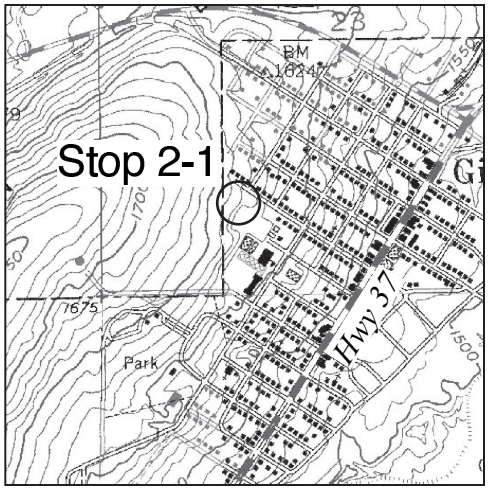
The Paleoproterozoic strata exposed in the Virginia horn are part of the Animikie Group, a tripartite sequence of sedimentary rocks, including basal quartzite and siltstone (Pokegama Quartzite), medial iron-bearing strata (Biwabik Iron Formation), and upper graywacke and shale of turbidite origin (Virginia Formation). The sediments were deposited during the compressional phase of the Penokean orogen in a northward-migrating, foredeep basin. A depositional age for the group can be inferred from interbedded volcanic tuff in the apparently equivalent Gunflint Iron Formation, which produced a U-Pb zircon date of approximately 1,878 Ma (Fralick and others, 2002). The belt of exposure forming the Mesabi range defines a monocline striking east–northeast and dipping shallowly (0° to 12°) southward. The exception to this trend is in the Virginia horn, where strike varies from north–south to northeast, and dips as great as 25° are recorded. This paired syncline-anticline is inferred to be related to uplift of a horst, now manifest in the Archean core of the structure, which formed by a combination of folding and faulting (Morey, 2003). Offset along the Laurentian fault, which was south-side down after the D_2 metamorphic and deformation event, was subsequently reactivated to produce north-side down movement during and after deposition of the Animikie Group.

STOP 2-1

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; off Wisconsin Avenue, 4 blocks northwest of State Highway 37 in Gilbert

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.

Discussion: Detailed structural study by Jirsa and others (1998) and Jirsa and Boerboom (2003b) demonstrated that tholeiitic (Stop 2-1) and calc-alkalic volcanic rocks and tholeiitic intrusions are conformably overlain by graywacke and slate (Stop 2-2). In detail, the succession forms a broad, twice-deformed syncline that has been segmented by faults of several generations.

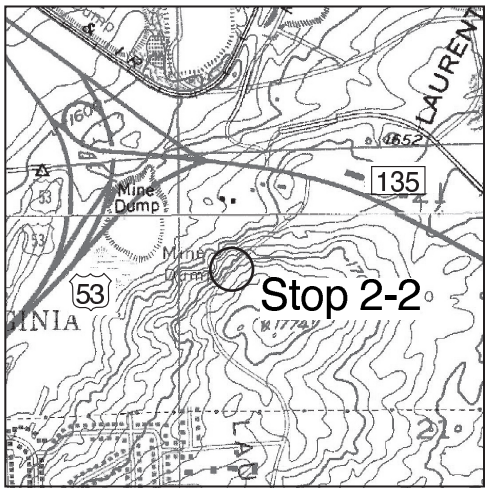
NEXT: Return to Highway 37, travel northeast to State Highway 135. Follow 135 west approximately 2.5 miles to Bourgin Road. Turn left (south) on Bourgin Road and continue about 0.4 mile to a large cut on the left (east) side of the road (Stop 2-2).

STOP 2-2

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 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.

Discussion: One of the earliest gold discoveries in Minnesota was made by J.W. Gruner (Grout, 1937) in a railroad cut not far from Stop 2-2. The cut exposes graywacke intruded by quartzofeldspathic porphyry, having visible gold associated with small quartz veins. Despite several episodes of prospecting and systematic study of the region, no economic gold deposits have been discovered.

NEXT: Follow Bourgin Road south and west to a frontage road on the east side of U.S. Highway 53. Turn north (right) on the frontage road and travel about 0.2 mile to the first road on the right, turn uphill and continue to #7 Mesabi Lane (Stop 2-3).

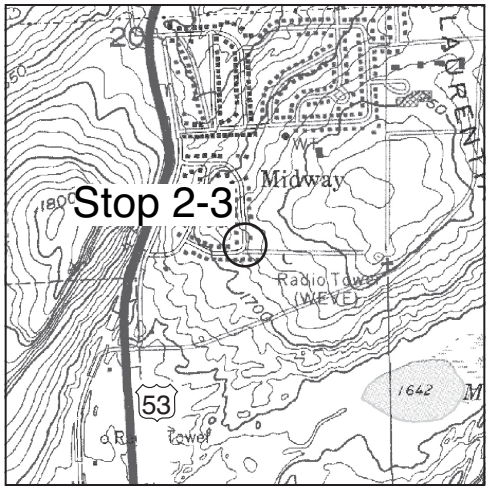
STOP 2-3

Private property! Permission must be obtained before entering!

Archean conglomerate

Location: T. 58 N., R. 17 W., sec. 20, SW, SE; No. 7 Mesabi Lane; town of Midway

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



Description: The 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, quartzofeldspathic porphyry, and porphyritic trachyandesite. This provenance indicates that the older Archean rocks of the Mud Lake sequence were intruded by quartzofeldspathic porphyry, deformed, uplifted, and eroded 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. Also note the presence of remnant skins of red jasper as at Stop 2-1, indicating that this outcrop surface represents the paleo-seafloor during deposition of the Paleoproterozoic Biwabik Iron Formation.

Discussion: The Midway sequence is similar in many respects to parts of the Knife Lake Group near Ely (Jirsa and Miller, 2004), and the Seine Group near International Falls (Jirsa, 2000). All of these have attributes of the so-called Timiskaming-type sequences, which include the namesake Timiskaming Group near Timmins, Ontario and the Shebandowan Group near Thunder Bay, Ontario. The Timiskaming-type rocks are inferred to represent successor-basin deposits that post-date deposition of the older greenstones in which they occur by as much as 30 million years (Corfu and Stott, 1998).

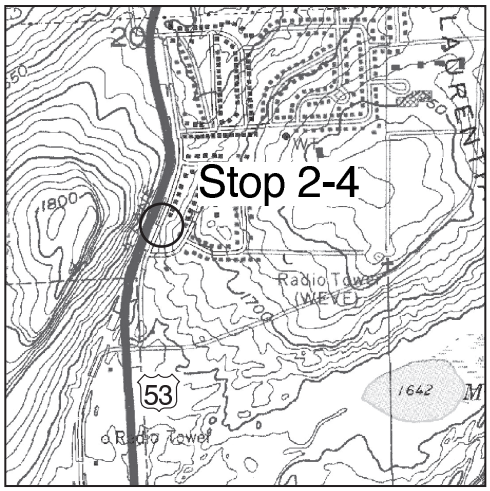
NEXT: Return to the frontage road paralleling Highway 53; turn north and travel about 500 feet to a low outcrop (Stop 2-4) on the east side of the road.

STOP 2-4

Archean/Paleoproterozoic unconformity

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

Eveleth quadrangle; UTM: 535,441E/5,259,520N



Description: In the brush on the east side of the frontage road is a small exposure of Archean conglomerate that has nearly vertical foliation and lithologic content much like that at Stop 2-3, capped by a thin and discontinuous "skin" of subhorizontally foliated conglomerate. This thin skin of conglomerate represents basal deposition of the Paleoproterozoic Pokegama Quartzite, the lowest of a tripartite sequence of formations that constitute the Animikie Group. Age dates are somewhat speculative; however, the Archean rocks of the various sequences in the Virginia horn probably are about 2.7 Ga (for example Peterson and others, 2001), and this part of the Animikie Group is inferred to be about 1.8 Ga (via Fralick and others, 2002). Thus, this unconformity represents a geologic hiatus of approximately 900 million years—almost twice that of all Phanerozoic time.

The large road cut visible on the west side of Highway 53 is the argillaceous lower member of the Pokegama Quartzite containing shale, siltstone, and sandstone. Minor channeling is common at the base of thicker sandstone beds, and small-scale cross-bedding occurs in some siltstone beds. The existence of bimodal/bipolar paleocurrent directions has been interpreted by Ojakangas (1993) to indicate deposition in a low-energy, upper tidal flat environment in a sea that transgressed the deeply eroded Archean surface. Local soft-sediment deformation may be the product of syn-depositional tectonism, or alternatively may represent localized collapse near tidal channels.

Discussion: Work by a number of authors (notably Southwick and others, 1988) demonstrated that the Animikie Group was deposited along the leading edge of a foredeep that transgressed north over the Archean craton during the Penokean orogeny. In detail, deposition of basal quartzite (Pokegama Quartzite), medial chemical (Biwabik Iron Formation), and upper turbiditic (Virginia Formation) sediments represents a transgression of shore-face and near-shore, shelf, and slope environments, respectively (Ojakangas, 1993).

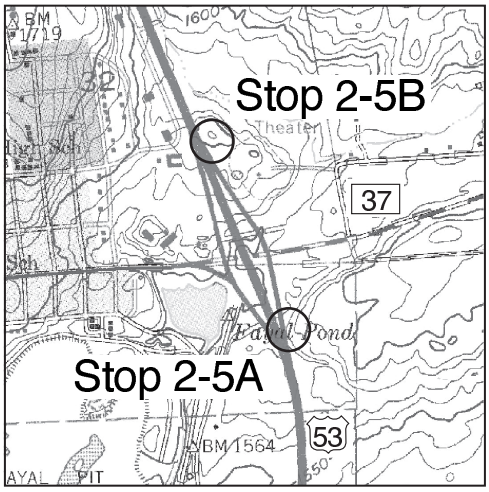
NEXT: Travel south on Highway 53 about 2 miles; drive south of the junction of Highways 53 and 37 and turn around; proceed north to a large outcrop on the east side of Highway 53 just south of State Highway 37 (Stop 2-5A).

STOP 2-5

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 2-5B), to 536,263E/5,256,200N on the south (Stop 2-5A)



Description 2-5A: This exposure of gently south-dipping strata is part of the Lower Cherty member of the Biwabik Iron Formation. It lies nearly at the crest-line of the anticline that forms half of the Virginia horn structure. The iron-formation overlies and generally grades into the Pokegama Quartzite exposed at Stop 2-5B. Notice that both formations have sandy textures and cross-bedding, implying a moderately high-energy depositional environment. The most significant difference between these two formations 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 and bipolar, implying deposition in a tidally influenced marine environment (Ojakangas, 1993).

Discussion: The Biwabik Iron Formation is generally divided into 4 units, informally known as the Lower

cherty, Upper cherty, Lower slaty, and Upper slaty members. Although these are convenient names, they are somewhat misleading. The cherty units are granular beds of recycled chemical precipitates including chert, iron-oxides, iron-carbonates, and iron-silicates. They are interbedded on all scales with "slaty" units of fine-grained, laminated iron-silicates and iron-carbonates. In most of the Mesabi Iron Range, the iron-formation and associated strata were not significantly metamorphosed, and much of the textural and mineralogic attributes are products of diagenesis and subsequent fluid movement. As a result, no slaty cleavage exists, and thus the term “slate” is applied only as a convenient field identifier.

NEXT: Continue north on Highway 53 to just past Highway 37; stop at a large outcrop on the east side of Highway 53 (Stop 2-5B).

Description 2-5B: This is the sandy, upper member of the Pokegama Quartzite. It is characterized by coarse grain size and massive beds as thick as 1.5 meters, separated by thin beds of shale and siltstone. Ojakangas (1993) interpreted the deposition of this facies as within a high-energy, lower tidal or subtidal environment.

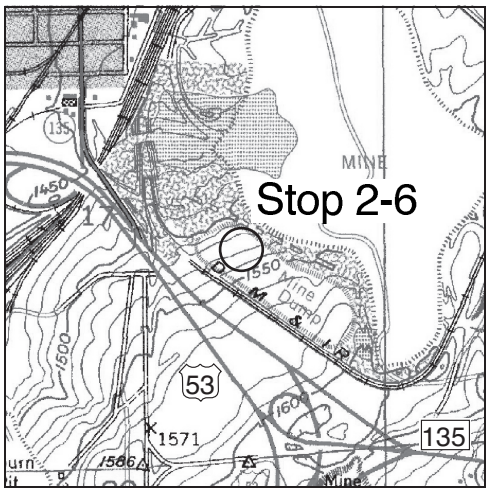
NEXT: Drive north on Highway 53 approximately 4 miles to the entrance to the Mineview overlook (Stop 2-6), just northwest of the junction of Highways 53 and 135. Follow the driveway to the overlook.

STOP 2-6

Mineview in the Sky Overlook; Paleoproterozoic iron-formation

Location: T. 58 N., R. 17 W., sec. 17, NW, SE. North of the junction of Highways 53 and 135

Virginia quadrangle; UTM: Top of overlook approximately 535,710E/5,261,650N



Description: North from this overlook is a 3-mile-long complex of abandoned mining properties, known collectively as the Rouchleau Mine, all developed within the Paleoproterozoic Biwabik Iron Formation. Actually, there were 15 separately named mines within view that shipped ore during the period from 1893 to 1986. All of them, and nearly 400 more along the 150-mile-long Mesabi Iron Range, extracted oxidized (hematite- or goethite-rich) and leached iron-formation generally referred to as "natural ore." Iron-formation at this point lies on the north-trending limb separating the syncline to our west and the anticline to the east. The natural ore deposits here are localized along a set of fault zones (Fig. 2.3) that presumably provided the plumbing system for fluids that first oxidized the formation and produced permeability, and secondly, leached silica from the porous zones. The natural ores mined here contained as much as 50 percent iron and less than 10 percent silica. Since the 1950s, the "ore of choice" has changed. The mammoth open-pit mine in the distance to the northwest, U.S. Steel Company's Minntac mine, is developed in unoxidized, magnetite-rich ore containing about 30 percent iron, and 50 percent silica. This type of ore at Minntac, and five other open-pit mines currently in operation along the range, is the source of the iron concentrate known as taconite. The name taconite has also been applied generally to unoxidized iron-formation containing sufficient iron to be mined for a profit using today's technology.

Discussion: Nearly 70 percent of the 3.6 billion metric tons of iron ore produced on the Mesabi Iron Range between the years 1892 and about 2000 was extracted as natural ore. Although it has been generally accepted that these ores formed along fracture, bedding, and fault planes by processes of oxidation and leaching, the source of altering solutions has been the subject of considerable debate among economic geologists for nearly 70 years. Much of the literature and geologic observations on the issue were reviewed in Morey (2003). Many writers supported the concept of descending meteoric waters to account for the dissolution of silica and oxidation of iron minerals. Others, including Gruner (1930), believed the geologic features were better explained by ascending hydrothermal solutions. Gruner's theory failed to gain common acceptance, in part because no driving mechanism for such a hydrothermal system could be envisioned. The integration of Animikie Group strata into the tectonic context of the Penokean orogen in east-central Minnesota revived the theory of hydrothermal fluid flow within the Pokegama Quartzite and ultimately the iron-formation, as part

of a continent-scale, gravity-driven, ground-water system (Morey, 1999). The debate continues, fueled in part by the observation that most of the alteration occurs near the present land surface.

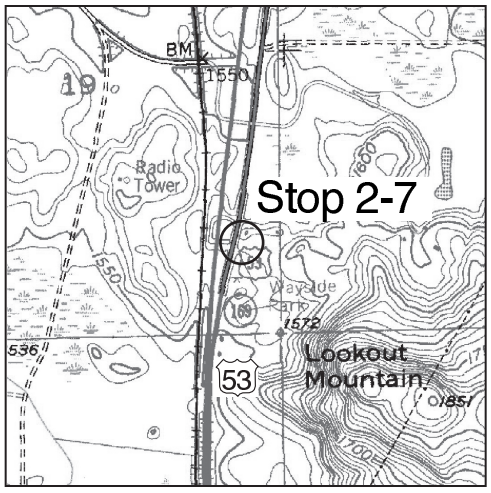
NEXT: Return to Highway 53 and travel north through the city of Virginia to a wayside rest area approximately 2 miles north of town (Stop 2-7).

STOP 2-7

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: 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 in the near absence of high-precision geochronologic data.

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 high-grade 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.

NEXT: Travel north on Highway 53 to the junction with State Highway 169; follow the latter to the northeast approximately 28 miles to St. Louis County Road 77. Turn left (northwest) on 77 and proceed about 0.5 mile to Pike River (Stop 2-8).

TOWER–SOUDAN AREA
Stops 2-8 to 2-10
Vermilion District

Geologic setting

Like the Virginia horn, the Tower–Soudan area contains exposures of Archean greenstone and granite of the Wawa subprovince. The two belts are separated by the Giants Range batholith (Figs. 2.2, 2.4), obfuscating their correlation. Rocks in the two areas are stratigraphically and geochemically similar, yet vastly different in structural style. The predominant structural feature here is a broad regional fold known as the Tower–Soudan anticline. Hudleston (1976) recognized that the anticline formed in the comparatively brittle volcanic strata during D_1 deformation. To the west of the anticline, in a more ductile package of graywacke and slate of the Virginia Formation, D_1 is manifest in a series of large thrust nappes containing large sections of down-facing strata. In both areas, strata are overprinted by D_2 deformation that produced regional metamorphism to greenschist and amphibolite grades, and structures that include local interference folds and a well developed rock cleavage. In addition to structural mapping (for example Hudleston, 1976; Hudleston and others, 1988; Schultz-Ela and Hudleston, 1991; Jirsa and others, 1992; Jirsa and Boerboom, 2003b), the study of geochemistry (for example Arth and Hanson, 1975; Schulz, 1980; Southwick and others, 1998) and U-Pb zircon geochronology (for example Boerboom and Zartman, 1993; Peterson and others, 2001) in Minnesota and adjacent Ontario are beginning to resolve some of the temporal issues. Much work remains in this endeavor.

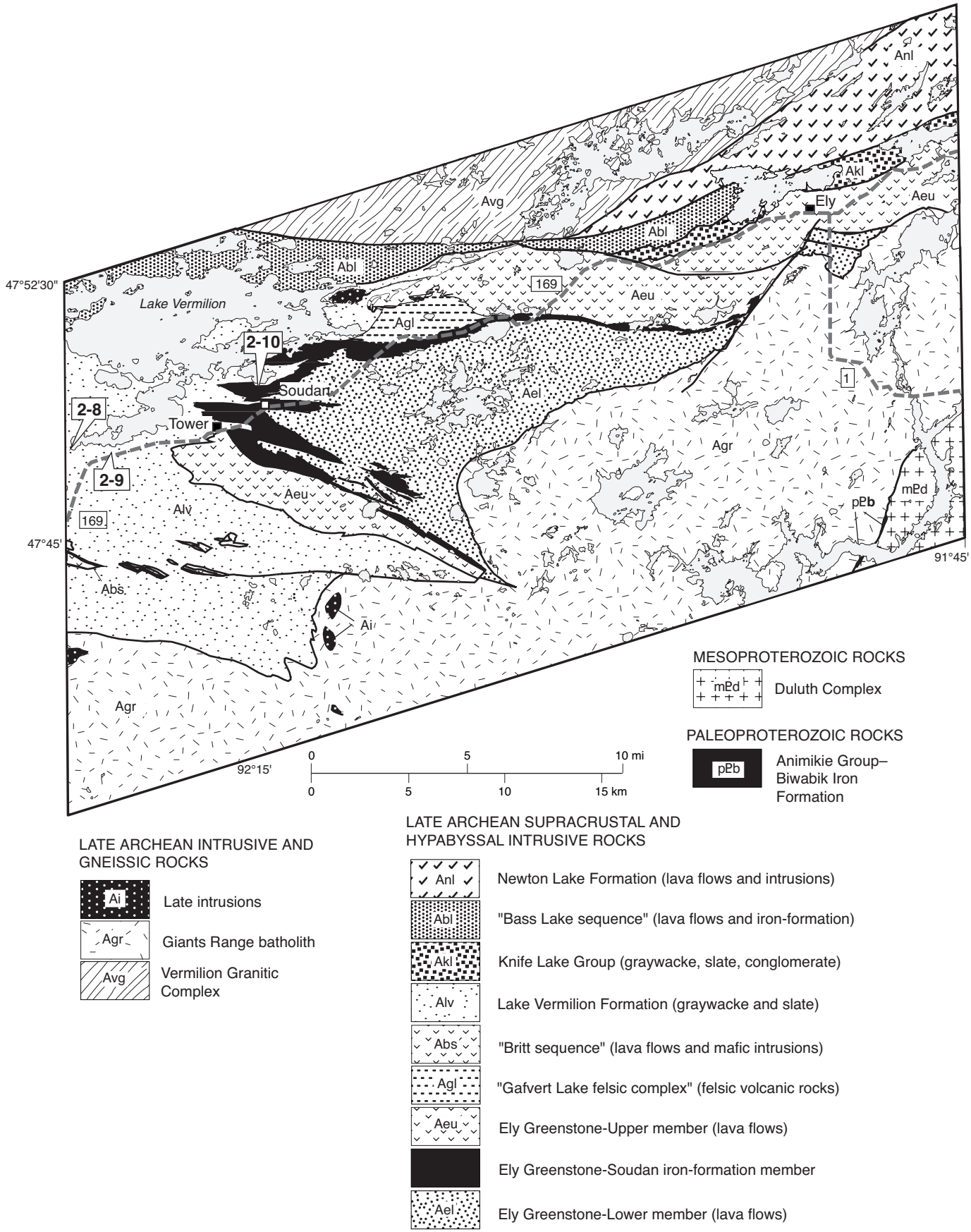


Figure 2.4. Geologic map of the western Vermilion district (modified from Peterson and Jirsa, 1999) including the Tower–Soudan and Ely areas, and showing details of field trip Stops 2-8 to 2-10.

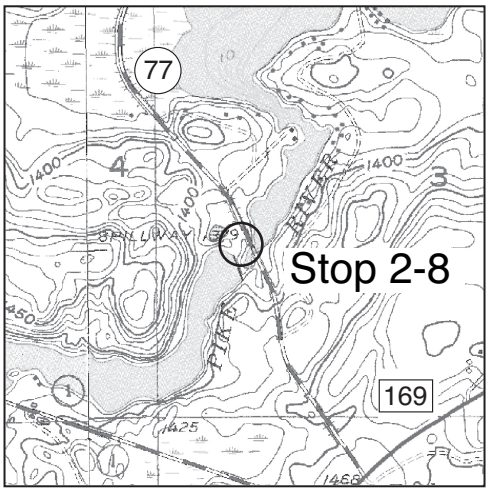
STOP 2-8

No hammering please!

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

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



Description: This glacially scoured outcrop exposes a nearly perfect cross-section of straight-bedded, variably graded, feldspathic graywacke and black slate. The feldspar-rich, dacitic composition of sandy textured beds is presumed to represent derivation from the Gafvert Lake felsic volcanic sequence exposed to the east in the Soudan area. Regionally, a series of outcrops from Gafvert Lake westward shows an irregular transition from proximal, possibly subaerial deposition on the east, to distal submarine turbiditic fan deposition to the west (Jirsa and others, 1993). The beds contain numerous "soft-sediment" deformation features including load structures, flames, intrafolial slump folds, and possibly some of the cross-stratal faulting. Bedding is nearly vertical and graded beds indicate stratigraphic younging to the south. This topping direction, and the presence of a weak D_2 cleavage that is left of bedding, indicate westward structural facing in the cleavage; consistent with a position on the south limb of a large, south-overturned, regional, D_1 fold structure—the western extension of the Tower–Soudan Anticline. Northeast-trending kink bands, fault zones, and raised quartz veins traversing the outcrop are assigned to the latest deformation event, D_3.

Discussion: The stratigraphic transition described above can be seen in viewing two outcrops not described in detail here. One is a road cut containing dacitic tuff cut by a Mesoproterozoic diabasic dike that lies 2.4 miles east of the junction of Highway 169 and County Road 77 (UTM: 551,160E/5,293,960N). More proximal sulfidic lapilli tuff and conglomerate are exposed in a road cut just west of the village of Tower (UTM: 553,380E/5,294,430N). Strata at both locations are considered part of the Lake Vermilion Formation, though detritus clearly was derived from felsic volcanic rocks of the Gafvert Lake sequence.

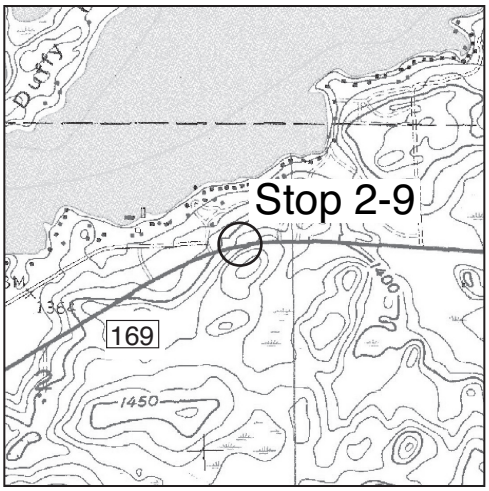
NEXT: Return to Highway 169 and turn east. Continue approximately 1.7 miles to just east of the junction of 169 and St. Louis County Road 526.

STOP 2-9

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 roadside and several smaller ones in the brush 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: Travel east on Highway 169, through the city of Tower, and continue east to the town of Soudan. Follow signs to the left toward Soudan Underground Mine State Park for approximately 0.3 mile—at this point, the state park entrance will be on your left. Do not follow the park signs, but rather continue on the roadway turning toward the right and follow the road to a junction about 0.4 mile to the east. Turn left and follow this road for about 0.5 mile, more or less straight past the Soudan Fire Station and up the hill toward the back side of the state park and Stuntz Bay. Disembark at the mine buildings and walk about 150 feet north and uphill to an outcrop on the right.

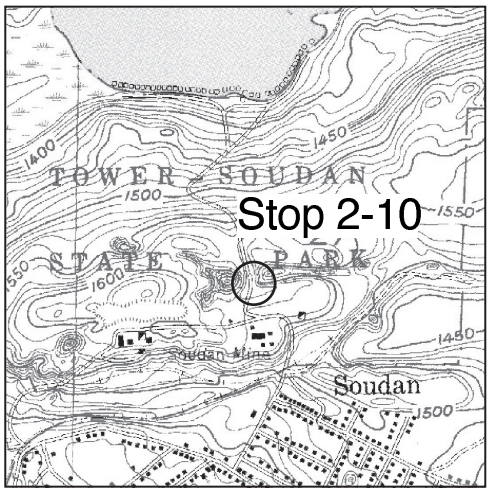
STOP 2-10

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



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. In many respects, this outcrop is a microcosm of the regional deformation. Two generations of tight folding are preserved in delicate laminae of chert (creamy white), chert-hematite jasper (red), and hematite-magnetite-chert (black to metallic gray). 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 instead the product of early-formed (F_{0-1}) sheath folding (Fig. 2.5). The F_1 structures are predominantly intrafolial and exhibit a great variety of style and orientation, implying that they formed by layer-parallel, soft-sediment slumping.

It is interesting to observe the rhythmic microlaminae (1 millimeter or so thick) 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.

Discussion: 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).

END OF DAY 1

OVERNIGHT IN TOWER

NEXT: From Tower, drive east on Highway 169 to Ely; at the east end of town, turn south on State Highway 1 and proceed about 50 miles to Finland. Turn left on Lake County Road 6, then at the junction with State Highway 61, turn left and proceed about 15 miles to the entrance to Sugarloaf Cove Interpretive Center. From the parking area, walk the trail past the interpretive center to the beach and then east to Sugarloaf Point. From Tower, the total distance is about 100 miles.

DAY 2
North Shore and Duluth Area
Midcontinent rift (Stops 2-11 to 2-18) and Animikie Group (Stops 2-18 and 2-19)



Figure 2.5. Detailed map of Soudan iron-formation outcrop showing bedding trajectories and several generations of folds and faults (from Lundy, 1985).

Geologic setting

Magmatic activity related to the 1.1 Ga Midcontinent Rift produced a thick edifice of lava flows and subvolcanic intrusions that are exposed along Minnesota's North Shore of Lake Superior. The Midcontinent rift formed along a 2,000-kilometer-long arcuate graben that extended in two arms from the Lake Superior region—one to the southwest to Kansas; the other to the southeast to Lower Michigan. The rifting is thought to have been influenced, if not caused by the impact of a starting mantle plume beneath the Lake Superior region about 1,109 Ma (Nicholson and others, 1997). Magmatic activity persisted over a 23 million year period in the Lake Superior region and infilled the rift graben with as much as 20 kilometers of subaerial lava flows and associated intrusions. The volcanic activity combined with extensive magmatic underplating of the lower crust resulted in much of the original continental crust in the western Lake Superior area being completely replaced by juvenile mafic crust (Allen and others, 1997). With the cessation of magmatic activity, continued downwarping of the rift resulted in the burial of igneous rocks by clastic sediments (the Oronto and Bayfield Group). During this sedimentation phase, regional compression caused a reversed motion on many of the originally normal graben-bounding faults and locally resulted in the formation of a central horst. The cause of this compression is attributed to the far-field affects of the Grenville orogeny, which was contemporaneously occurring along the eastern margin of North America (Cannon, 1994).

The geology of northeastern Minnesota (Fig. 2.6) is dominated by volcanic rocks and comagmatic intrusive rock formed along the northwestern flank of the Midcontinent rift. The volcanic rocks, termed the North Shore Volcanic Group, are mostly exposed along the shoreline where they comprise a volcanic edifice about 10 kilometers thick (Green, 1972). The flows are predominantly tholeiitic to subalkaline basalts, but also include intermediate and felsic flows and fluvial interflow sedimentary rocks. All lavas except for a few of the basal flows were erupted subaerially and most have the sheetlike form of flood basalts. The flows are gently tilted toward the axis of the rift in an arcuate pattern such that the top of the volcanic pile is situated midway along the shore near Tofte, and the base of the pile is exposed near Grand Portage and Duluth (Fig. 2.6). This part of the trip will traverse the southwestern limb of the North Shore Volcanic Group from Tofte to Duluth.

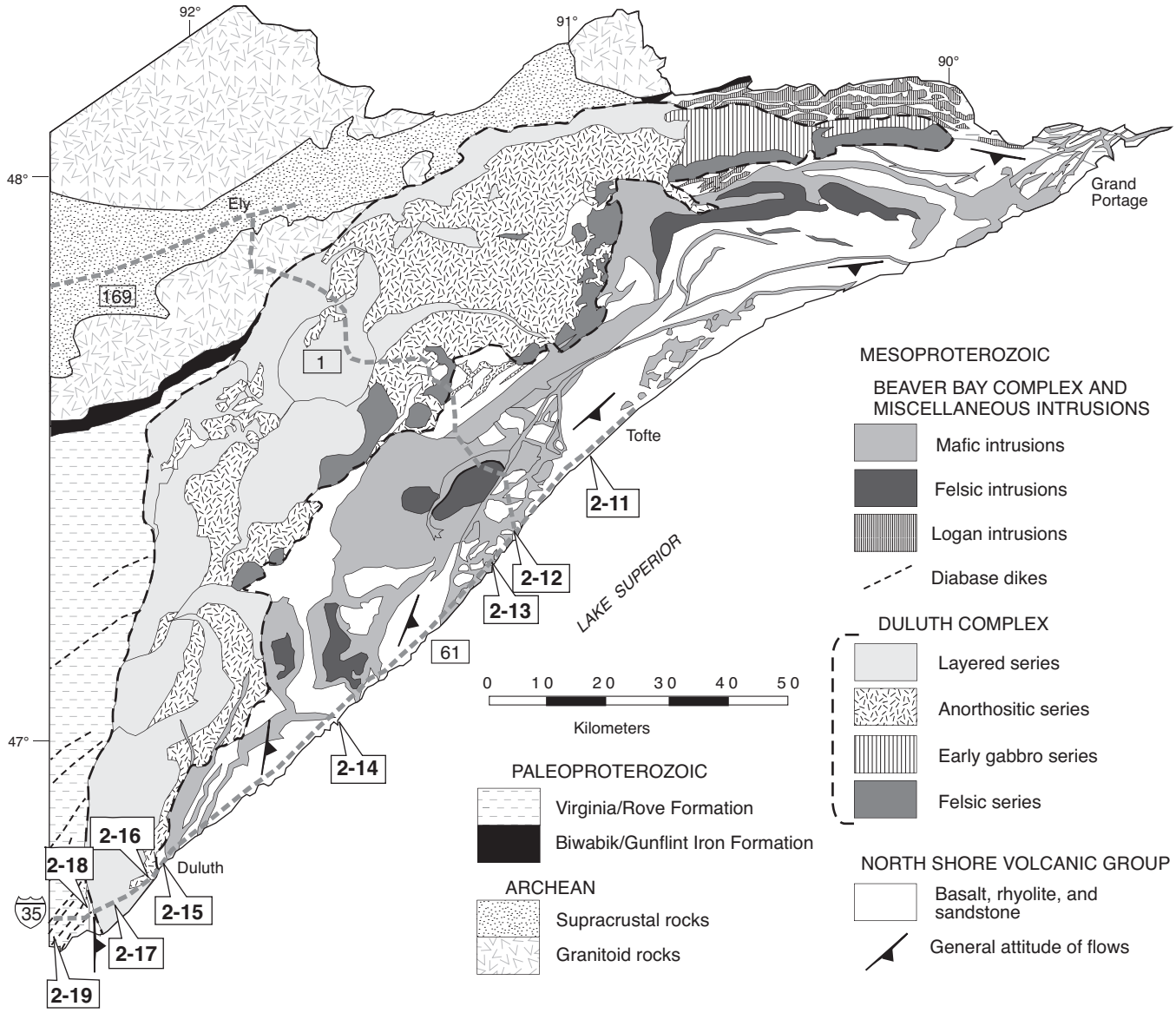


Figure 2.6. Geologic map of northeastern Minnesota (modified from Miller and others, 2002) showing field trip Stops 2-11 to 2-19 on the North Shore and in the Duluth area.

As seen in Figure 2.6, intrusive rocks dominate the geology related to the Midcontinent rift in northeastern Minnesota. The main mass of intrusions extending in an arcuate swath from Duluth north and eastward along the international border form the Duluth Complex. This complex is composed of multiple intrusions of layered gabbro and troctolite, gabbroic anorthosite, and granophyre, which were emplaced as sheetlike bodies into the basal part of the North Shore Volcanic Group (Miller and others, 2002). Duluth Complex intrusions span a similar range of U-Pb ages to what are found in the overlying North Shore Volcanic Group (1,108 to 1,096 Ma; Paces and Miller, 1993), which implies that they likely acted as staging chambers to volcanic eruptions. Intrusions of a similar variety of rock types emplaced structurally higher in the volcanic pile are termed the Beaver Bay Complex. Whereas the deeper Duluth Complex intrusions commonly show cumulate textures and evidence of magmatic differentiation by fractional crystallization, the shallower and generally thinner Beaver Bay Complex intrusions tend not to show signs of internal differentiation. The main exception to this is the well-differentiated Sonju Lake intrusion near Finland, Minnesota (Miller and others, 2002). We will visit intrusions associated with both the Beaver Bay Complex and the Duluth Complex.

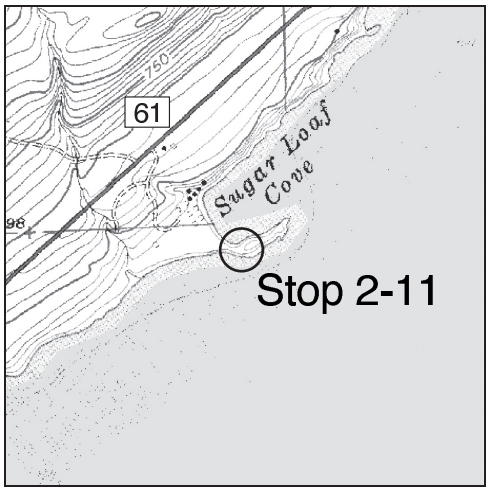
STOP 2-11

No hammering please!

Olivine tholeiitic basalt flows at Sugarloaf Cove (North Shore Volcanic Group)

Location: T. 58 N., R. 5 W., sec. 29, NE, NE

Little Marais quadrangle; UTM: 652,020E/5,261,120N



Note: This stop is on property of the Sugarloaf Interpretive Center Association (a non-profit) and the State of Minnesota. The beach and point are part of a Scientific and Natural Area (Minnesota Department of Natural Resources), established to preserve these features. Please do not use hammers, or remove samples of any kind.

Description: The lava flows exposed in this area are part of the Schroeder–Lutsen basalts, the uppermost stratigraphic unit of the North Shore Volcanic Group. Flows range from several tens of meters thick to thin flow units of a meter or less. Massive interiors are ophitic; each "ophite" is made of an augite oikocryst enclosing many small plagioclase tablets. Abundant tiny olivine grains occur with plagioclase between the ophites. Pipe amygdules are common in the flow bases, and small vesicle cylinders occur in the massive interior of the thick flow that caps the sequence (the Sugarloaf) on the point. These are all pahoehoe flows, and ropy surfaces are well displayed, as are columnar joints and red sandstone-filled fractures ("clastic dikes") in the interiors. The variable orientation of the flow tops highlights the topography developed on even very fluid lava flows such as olivine tholeiites.

Discussion: These olivine tholeiite flows are typical of the volcanic interval called the Schroeder–Lutsen basalts, which form the upper part of the North Shore Volcanic Group. The occurrence of such primitive compositions (Fig. 2.7) near the top of the volcanic pile highlights the fact that North Shore Volcanic Group basalts generally become more primitive (or less evolved) up section. This has been commonly interpreted as indicating a shorter crustal residence time of mantle-derived magmas as the rift evolved and conduits became better established.

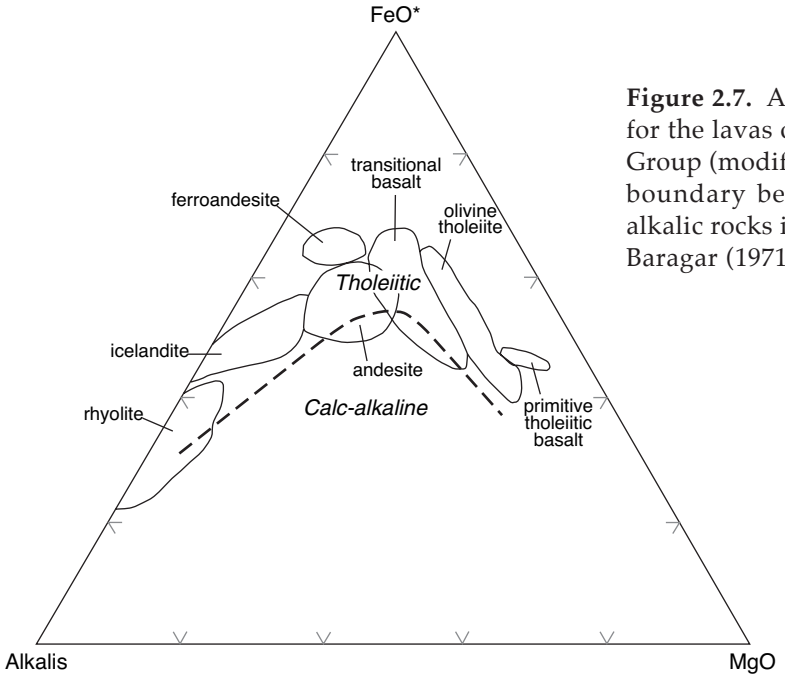


Figure 2.7. AFM compositional diagram for the lavas of the North Shore Volcanic Group (modified from Green, 1982). The boundary between tholeiitic and calc-alkalic rocks is modified from Irvine and Baragar (1971). FeO* = FeO + 0.9Fe_2O_3.

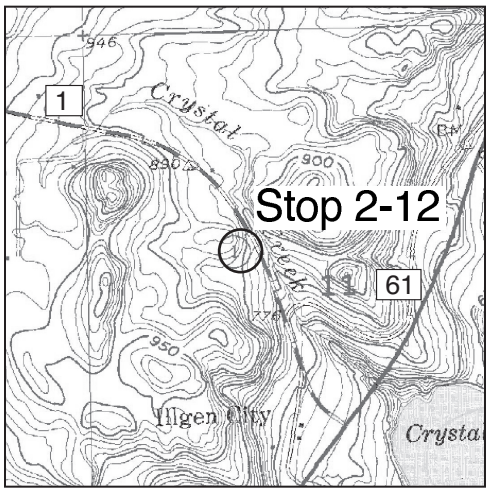
NEXT: Return southwest on Highway 61 approximately 15 miles to the junction with State Highway 1 at Illgen City. Proceed 0.2 mile on 1 and park on the east shoulder. Road cuts are on the west side of the road.

STOP 2-12

Palisade porphyritic rhyolite flow (North Shore Volcanic Group)

Location: T. 56 N., R. 7 W., sec. 11, NE, SW; 0.2 mile north of Highway 61 on Highway 1

Illgen City quadrangle; UTM: 636,600E/5,245,850N



Description: Road cuts on the west side of Highway 1 expose the upper zone of the Palisade rhyolite, a very large (greater than 95 meters) and extensive porphyritic flow that is part of the North Shore Volcanic Group. Much of the flow, seen for example on Palisade Head and Shovel Point and the Highway 61 cut just southwest of the junction with Highway 1, has the appearance of a normal lava flow, but features here at the top, and at the inaccessible base of the cliff at Palisade Head, show clearly that it was erupted explosively. The hot ash settled and welded, and the resulting hot, devolatilized mass flowed as "reconstituted lava" or rheoignimbrite. This road cut in the rapidly chilled top part of the flow is made of flow-brecciated, flow-folded welded tuff; microscopic study reveals deformed shards and fiamme in some samples (more obvious shards are visible in the chilled welded tuff at the flow's base). Pneumatolytic or hydrothermal alteration has kaolinized much of the feldspar at this locality. In exposures of the massive flow interior, flow-aligned platy quartz after tridymite is abundant in the groundmass, implying a high temperature of eruption (greater than 870° C).

Exposures of the massive porphyritic interior of the Palisade rhyolite flow can be observed on Shovel Point and the High Falls of the Baptism River in Tettegouche State Park, just south of here, and two miles to the south on Palisade Head.

Discussion: Rhyolite flows comprise a significant portion of the North Shore Volcanic Group—approximately 25 percent of the northeast limb and approximately 10 percent of the southwest limb (Green, 1972). Field studies and petrography of this and other large felsic flows has indicated that they were erupted at unusually high temperatures for rhyolite, were highly mobile, and that many are rheoignimbrites (Green and Fitz, 1993). Sm/Nd analyses (Vervoort and Green, 1997) imply that most rhyolites were produced by partial melting of the Archean basement. Much of this melting likely resulted from the extensive amount of magmatic underplating of the lower crust by mantle derived magmas that is implied by gravity data (Behrendt and others, 1990).

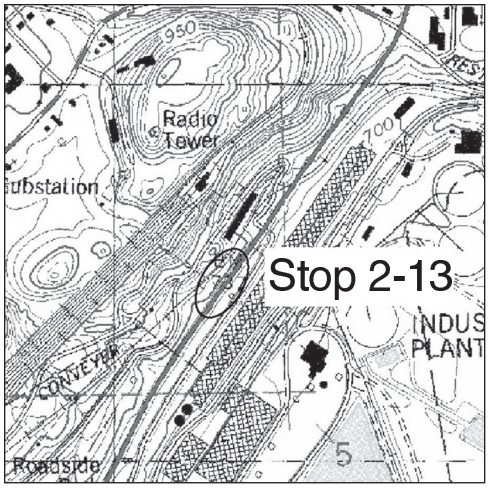
NEXT: Return to Highway 61. Turn southwest (right) on 61 and drive approximately 4.4 miles to a pullout area on the southeast side of the highway adjacent to the gated entrance to the North Shore Taconite Plant. Walk across the highway to a road cut on the northwest side of the road.

STOP 2-13

Beaver River diabase with anorthosite inclusions at Silver Bay (Beaver Bay Complex)

Location: T. 56 N., R. 7 W., sec. 32, SW, SW; road cut on Highway 61 across from the North Shore Taconite Plant Entrance

Silver Bay quadrangle; UTM: 631,300E/5,238,735N



Description: In the road cut, fine-grained, dark-colored diabase hosts several decameter-sized blocks of coarse-grained, pale green-tinted anorthosite. The diabase is part of an aereally extensive network of

dikes and sills that make up the Beaver River diabase—one of the younger intrusive phases of the Beaver Bay Complex. This intrusive unit, which holds up most of the high ground in this area of the North Shore, can be traced from Split Rock Point (15 kilometers to the southwest) to just west of Grand Marais (80 kilometers to the northeast). The diabase is typically a fine-grained ophitic olivine gabbro, but grades into coarser and more subophitic to intergranular gabbro in the medial portions of thicker sheets (up to 150 meters). The exposure here is located within 20 meters of the lower contact of a diabase sheet and thus is rather fine-grained. One of the most distinctive characteristics of the Beaver River diabase is that it commonly hosts large (as much as several hundred meters in diameter), rounded to angular inclusions of nearly pure anorthosite. The anorthosite is typically composed of 90 to 99 percent coarse-grained bytownitic to labradoritic plagioclase, with minor amounts of Mg-rich olivine, orthopyroxene, and less commonly clinopyroxene. These inclusions, which locally are brecciated and recrystallized (Morrison and others, 1983), are particularly common in the lower margins of larger diabase sheets. The contacts between the inclusion and the diabase are sharp, but show no signs of chilling. Both the diabase and the anorthosite inclusions in this exposure are cut by late-stage pinkish veins of alkali feldspar, zeolite, and calcite.

Although not the most spectacular exposure of anorthosite occurrences, we have chosen this stop because of its easy access. Some other locations along the North Shore where anorthosite inclusions in diabase can be viewed are Carlton Peak near Tofte, the inland areas of Tettegouche State Park, and Split Rock Lighthouse State Park. The Carlton Peak occurrence is the largest inclusion known, measuring over 500 meters across. The Split Rock lighthouse location has been described in many other guidebooks (including Miller and others, 1987, Stop 1; Miller, 1995, Stop 2-1; Boerboom and others, 2004, Stop 2-8; Jirsa and others, 2004, Stop 5-21). Jirsa and others (2004) also gave a brief history of the various ideas proposed over the past 150 years for the origin of this distinctive rock.

Discussion: This and other anorthosite inclusions hosted in the Beaver River diabase are different from most anorthositic-series rocks of the Duluth Complex (to be viewed at Stops 2-16 and 2-17); most anorthositic-series rocks are rarely layered, more compositionally evolved, and rarely contain cumulus hypersthene. These inclusions are similar, however, to some anorthosite inclusions within the anorthositic series. The highly disordered structural state of plagioclase (Miller, 1986) and the absence of any discernible chill of the diabase against the anorthosite indicate that these inclusions were derived from a middle to lower crustal source. Isotopic and trace-element compositions of these crustal xenoliths imply that they may be pre-Keweenawan in age (Morrison and others, 1983), but the data are ambiguous. Alternatively, if a plagioclase crystal mush origin for Duluth Complex anorthositic rocks is correct (Miller and Weiblen, 1990), a corollary of such a model is that significant amounts of Keweenawan anorthosite, generated by plagioclase flotation under high pressure, should have formed in the deep crust prior to BBC magmatism at 1,096 Ma. Under deep crustal conditions, such plagioclase cumulates would probably be distinctive in texture and composition from their shallow crustal counterparts. Moreover, the ambiguous isotopic compositions of the inclusions may indicate that anorthosite-forming Keweenawan magmas were contaminated by older crust, rather than older anorthosite being contaminated by interaction with Keweenawan magmas, as concluded by Morrison and others (1983).

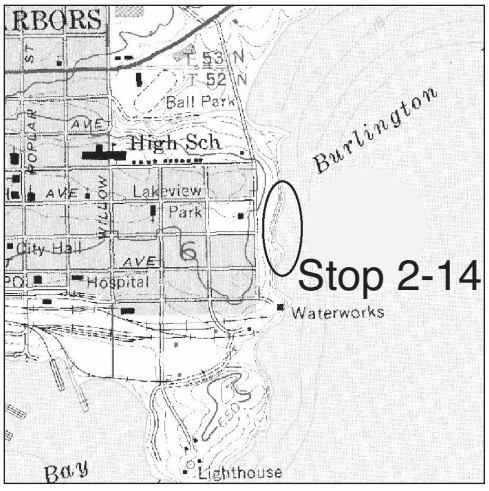
NEXT: Return to Highway 61, turn southwest (left), and continue approximately 29 miles to Two Harbors. At the north end of town, turn left on 1st Street and continue south to the crest of the hill adjacent to the picnic grounds. Walk east to the shoreline.

STOP 2-14

Quartz tholeiite lava flows at Two Harbors Town Park (North Shore Volcanic Group)

Location: T. 52 N., R. 10 W., sec. 6, NE ; shoreline exposures east of city park

Two Harbors quadrangle; UTM: 601,851E/5,208,673N to 601,847E/5,208,132N



Description: Along this section of the shoreline, a sequence of several southeast-dipping quartz tholeiite basalt flows are exposed. These are part of the medial section of the southwestern limb of the North Shore Volcanic Group. In the easily accessible exposures due east of the picnic shelter, the main shoreline shelf is composed of a fine-grained, fresh, intergranular basalt that locally contains vugs and amygdules of agate and crystalline quartz. In the cliff face (rock is unstable, please do not climb), this mildly vuggy basalt abruptly transitions into a flaggy jointed, very amygdaloidal flow top rich in laumontite and calcite. The top of the cliff face is capped by the massive base of another quartz tholeiite flow, which as can be seen in the fallen blocks, displays oxidation lamination, a feature common to basalts containing 50 to 54 percent SiO_2. Scattered amygdules in this overlying flow commonly contain dark green chlorite and gray agate. The contact between the two flows is sharp and locally marked by thin lenses of red siltstone.

South along the shoreline over a point of land, the flow top of the upper oxidation-laminated flow can be observed to abruptly transition into a thick, amygdaloidal, AA-type, autobrecciated flow top. The amygdaloidal basalt blocks seem to be embedded in a matrix of pink laumontite and calcite. This flow top is again capped by a massive overlying flow along a sharp contact.

Note the paleo-topographic control preserved in the flow zonation. Each rocky point along this traverse corresponds to the massive, more resistant base of a flow, whereas each shoreline recess is formed in a rubbly, easily eroded, flow top. This is typical of the entire North Shore where gently tilted flows strike oblique to the trend of the shoreline.

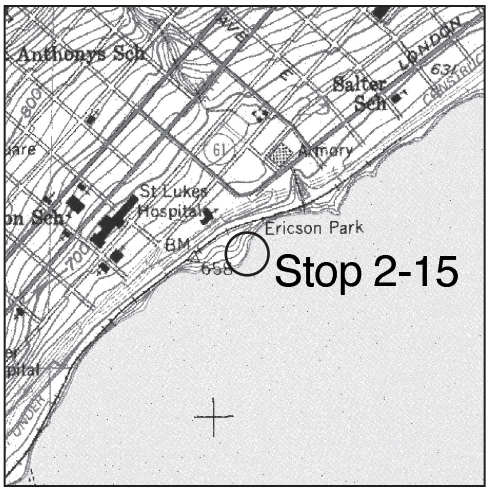
NEXT: Continue southwest on Highway 61 for approximately 30 miles. Near Duluth, Highway 61 becomes London Road; continue southwest on London Road (do NOT turn onto Interstate 35). Continue approximately 1.25 miles to the parking area for Leif Erikson Park near 12th Avenue East. Walk into the park and to the shoreline.

STOP 2-15

Interflow sedimentary rocks at Leif Erikson Park (North Shore Volcanic Group)

Location: T. 50 N., R. 14 W., sec. 23, SW

Duluth quadrangle; UTM: 570,050E/5,182,950N



Description: This location exposes an estimated 110 feet of interflow sandstone, lying on an eroded basalt flow of the gently southeast-dipping Leif Erikson Park lavas. The sandstone is fine- to medium-grained, moderately well sorted, and derived almost totally from underlying and adjacent lava flows—no extrabasinal detritus has been detected (Jirsa, 1984). The presence of planar-tabular and trough cross-bedding, together with the lenticular distribution of interflow units, implies the strata represent occasional stream flow deposits in depressions on the evolving lava surfaces. This is one of many interflow sedimentary units, most of which are "red bed-like" in appearance. By contrast, this interflow has a very distinct green color imparted by the abundance of metamorphic epidote.

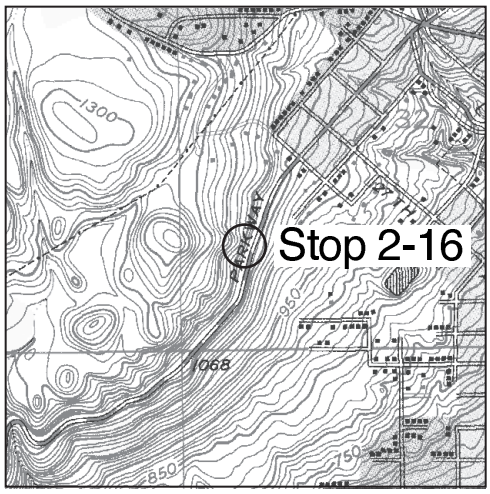
Discussion: Because these sedimentary strata lie in a stratigraphic interval that should have been buried beneath about 7 kilometers of volcanic rocks, it is possible that it has experienced burial metamorphism to nearly greenschist facies (Schmidt and Robinson, 1997). However, lavas just to the northeast of here are only metamorphosed to zeolite grade. Therefore, it seems more likely that the elevated grade observed in this section is a thermal metamorphic affect imposed by the 5-kilometer-thick Duluth Complex, which is just 750 meters below this horizon, and the 500-meter-thick Endion diabase sill, which is 50 meters above this level.

NEXT: Follow signs to Interstate 35, travel southwest on 35 to U.S. Highway 53. Turn north on Highway 53 (Piedmont Avenue) and continue about 1 mile to the intersection with Skyline Parkway. Turn left (southwest) on Skyline Parkway and travel 0.4 mile to the crossing with 28th Ave. W.; continue 0.2 mile farther southwest on Skyline Parkway to an outcrop on the right (Stop 2-16).

STOP 2-16

Layered series "chill," granophyre, and anorthositic series on Skyline Parkway (Duluth Complex)

Location: T. 50 N., R. 14 W., sec. 32, SW, SW; Skyline Parkway about 1/5 mile southwest of 28th Ave. W. Duluth Heights quadrangle; UTM: 564,880E/ 5,179,640N



Description: Exposed at the northeast end of this road cut is a fine-grained mafic rock with intermingled granophyre that together cut coarse-grained olivine gabbroic anorthosite. The gabbroic anorthosite observed here and along much of Skyline Parkway above West Duluth is typical of what is termed the anorthositic series of the Duluth Complex. The fine-grained mafic rock exposed here can be traced up over the ridge to the west where it merges into the upper contact zone of the layered series (Fig. 2.8). As will be observed at the next stop (2-17), the layered series at Duluth is a well-differentiated layered intrusion composed of troctolitic to gabbroic cumulates.

Discussion: Since early mapping by Grout in the Duluth area in 1911, the relationships displayed by this rather innocuous exposure have been critical in interpreting the intrusive history of the Duluth Complex. Grout (1918) and later Taylor (1964) saw this exposure as evidence that the anorthositic series had cooled considerably when the layered series was intruded, and thus was considerably older. This paradigm was accepted by all subsequent workers on the Duluth Complex through the 1990s. Consequently, it was surprising when high-resolution U-Pb ages (Paces and Miller, 1993) revealed that the anorthositic series and the layered series were virtually identical in age (within 0.5 Ma relative to the 23 Ma span of Midcontinent Rift magmatic activity). This precipitated a major change in the perception of the intrusive relationships between these two series here and throughout the Duluth Complex (Miller, 1992).

A closer look at this layered series "chill" reveals that it is not a thermal quench of the layered series at Duluth at all. This rock type is found at the contact with the anorthositic series throughout most of this area and has a remarkably homogeneous, evolved composition (MgO/(MgO + FeO) = ~37; see Table 2 in Miller and Ripley, 1996). In thin section, it is a subprismatic biotitic oxide ferrodiorite. Phase equilibrium modeling of its composition indicates that it should contain evolved compositions of augite, ilmenite and plagioclase—phases that comprise gabbroic cumulates found in the cyclic zone and gabbro zones of the layered series at Duluth. In sum, this rock is too evolved to have produced the entire cumulate pile of the layered series. Rather than this being a thermal quench, Miller and Ripley (1996) interpreted this rock to represent decompression quenching of an evolved, water-saturated, layered-series magma during venting at a time when the cyclic zone was crystallizing. Decompression of less than water saturated magma will result in superheating and a suspension of crystallization (or at least a significant change in phase equilibrium). By contrast, decompression under water saturated conditions will cause supercooling and quenching.

The lobate contacts between the irregular masses of medium-grained granophyre and the fine-grained ferrodiorite host clearly give the appearance of two-magma mixing. Grout (1918) suggested that these two magmas formed by silicate liquid immiscibility of hydrous mafic magmas in the roof zone of the Duluth Complex. The concept of silicate liquid immiscibility, which had come to be totally disparaged during Grout's day, has regained some respectability as a plausible, albeit uncommon, petrologic process (Roedder, 1974). Nevertheless, alternative explanations for this compositional dichotomy can be suggested. One is that these felsic magmas were derived from anatectic melting of various inclusions carried into the layered-series chamber. Because of their high silica content and low density, these felsic melts did not readily mix or assimilate with mafic melt, but rather rose to the roof zone where they ponded beneath the anorthositic-series cupola. During magma venting from the chamber, the felsic melts became entrained and irregularly mixed with the mafic magmas. Because decompression under water saturated conditions would cause rapid crystallization of the mafic magma, the felsic melt, which became irregularly entrapped in the quenched mafic host, would have cooled to a medium-grained texture slower.

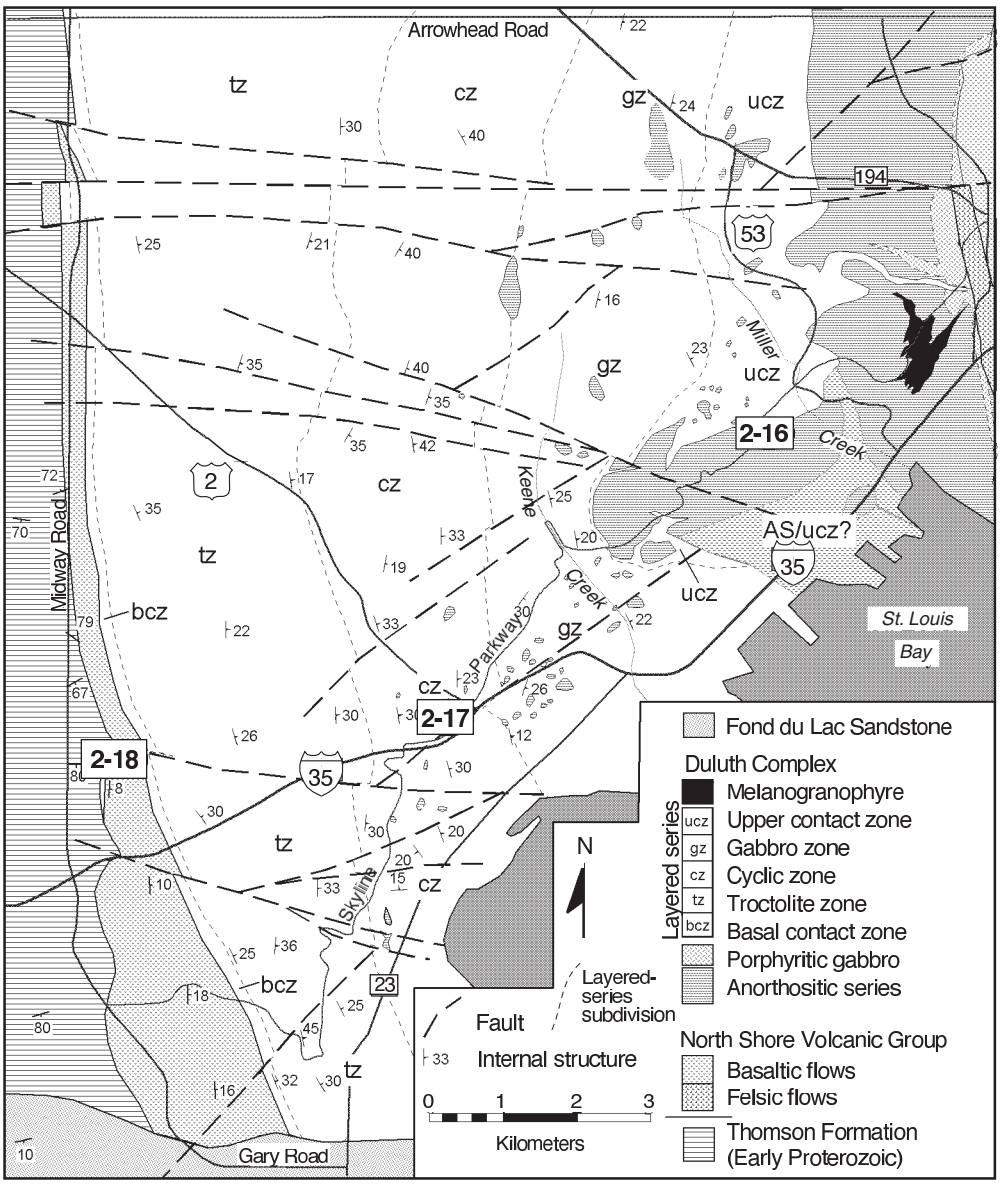


Figure 2.8. Geology of the Duluth Complex in the Duluth area showing the locations of Stops 2-16, 2-17, and 2-18 (modified from Miller, 1995).

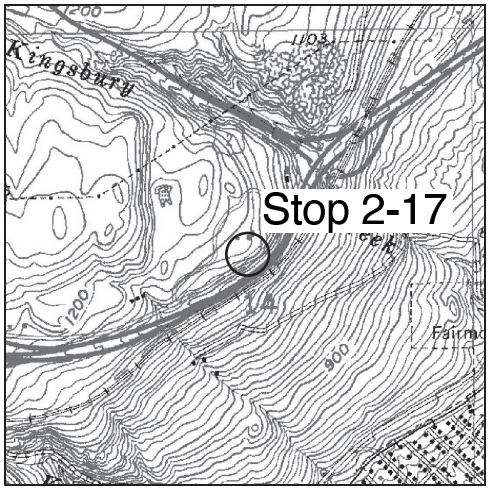
NEXT: Continue southwest along Skyline Parkway to the Thomson Hill rest area. Park in the rest area and walk to road cuts west of the parking area.

STOP 2-17

Layered gabbro cumulates and anorthositic-series inclusions at Thomson Hill rest area (Duluth Complex)

Location: T. 49 N., R. 15 W., sec. 14, SE, NW; road cut adjacent to the parking lot for the Thomson Hill rest area on Skyline Parkway

West Duluth quadrangle; UTM: 560,700E/ 5,175,380N



Description: In the road cut north of the rest area parking lot is an exposure of variably layered, well-foliated olivine oxide gabbro hosting several meter-sized, flattened inclusions of olivine anorthosite. The gabbro is a four-phase cumulate of plagioclase, augite, Fe-Ti oxide, and olivine. This cumulate typifies the upper gabbroic parts of troctolite → gabbro macrocycles that characterize the medial cyclic zone of the layered series at Duluth (Fig. 2.7).

Discussion: In this exposure, the anorthositic inclusions clearly have a pancake shape that is conformable to layering and foliation of the host gabbro. This may represent the original shape of the inclusions or it may indicate compaction of partially molten blocks. In either case, the occurrence of anorthositic inclusions in layered-series rocks is a ubiquitous feature throughout the Duluth Complex. That the opposite relationship is rarely if ever observed has reinforced the long-standing interpretation that the anorthositic series is older than the layered series. However, as noted at the last stop, high precision U-Pb dating (Paces and Miller, 1993) has shown that this age difference is less than 1 million years relative to the 23-million-year magmatic history of the Midcontinent Rift.

Gabbroic anorthosite inclusions are common in the gabbroic cumulates of the layered series at Duluth, but are not present in the troctolitic (plagioclase + olivine) cumulates. One possible explanation for this empirical relationship is that fractional crystallization of an olivine-plagioclase assemblage will lead to a gradual increase in the density of the differentiated magma, whereas crystallization of a gabbroic mineral assemblage involving Fe-Ti oxide will lead to a decrease in magma density, thus allowing detached blocks from the anorthositic-series hanging wall to settle to the magma chamber floor.

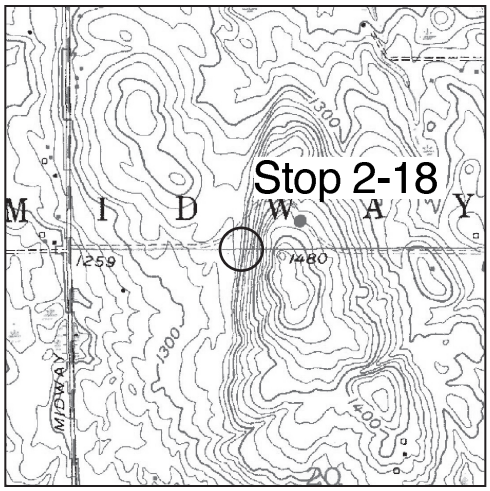
NEXT: Return to Interstate 35 and travel southwest approximately 3 miles to St. Louis County Road 13 (Midway Road). Turn north on Midway Road and continue 0.8 mile to a road to the east (right). Follow this road east approximately 0.2 mile, disembark and walk east to the base of the bluff.

STOP 2-18

Basal North Shore Volcanic Group basalt flow, Nopeming Sandstone, and Thomson Formation, (Mesoproterozoic/Paleoproterozoic unconformity)

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

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



Description: This area contains exposures showing the unconformable relationship between the Paleoproterozoic Thomson Formation and the base of Mesoproterozoic (Keweenawan) volcanic and sedimentary rocks. Scattered outcrops in the brushy area just west of the bluff expose folded and metamorphosed feldspathic graywacke, siltstone, and mudstone or slate of the Thomson Formation. By contrast, the Keweenawan strata exposed in the bluff have been little affected by tectonism since deposition, as shown by their gentle dip to the east. The unconformity is not exposed, but inferring from the two sets of exposures, it appears to be an angular one. It represents a nearly 800-million-year hiatus, based on an age of roughly 1,108 Ma for lower reversed polarity lavas and intrusions elsewhere in the rift (Davis and Green, 1997), and a date of about 1,880 Ma for tuff interbedded with iron-formation (Fralick and others, 2002) in the Animikie Group, of which the Thomson Formation is a component.

The Keweenawan stratigraphic section in this area consists of a basal sedimentary unit, the Nopeming formation, which is conformably overlain by the Ely's Peak basalts. The basalts comprise the oldest volcanic strata of the North Shore Volcanic Group. The Nopeming formation consists of approximately 30 feet (10 meters) of interbedded conglomerate and quartz arenite, with minor siltstone beds in the uppermost part of the unit that is exposed here. Much of the sandstone is medium- to coarse-grained, well-sorted and rounded, quartz arenite. The uppermost, silty parts of the Nopeming formation contain load casts and other structures indicating soft-sediment deformation. The overlying Ely's Peak basalts locally contain pillow structures, indicating subaqueous deposition. Pillow structures are only observed in the lowest lava flows throughout the rift. Interestingly, a nearly identical stratigraphic sequence is observed

at the northeastern flank of the North Shore Volcanic Group as well as at the base of the rift sequence in the Michigan–Wisconsin border area. Together, these sequences indicate that a broad downwarping of the upper crust preceded the onset of volcanism in the western Lake Superior region.

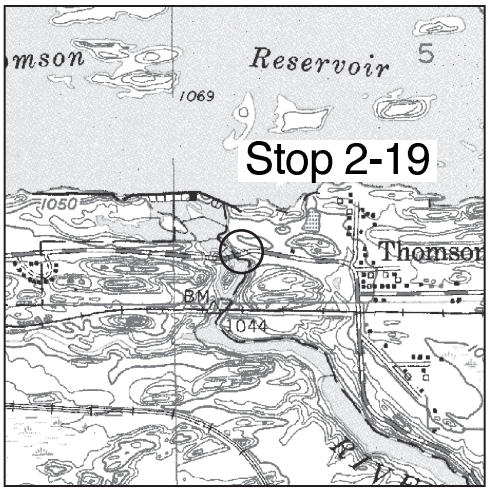
NEXT: Return to Interstate 35 and head south (actually west). Take the Esko/Thomson exit (St. Louis County Road 98/Carlton County Road 1) and head south (left) toward Thomson. At the junction with State Highway 210, turn right and follow it several blocks to a pull out next to the bridge over the St. Louis River (Fig. 2.9).

STOP 2-19

Deformed Thomson Formation (Paleoproterozoic) and Keweenawan diabase dikes

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

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



Description: This is the type locality of the Paleoproterozoic Thomson Formation, represented by approximately 650 feet of strata exposed between the dam north of Highway 210 and the railroad bridge south of it. The Thomson Formation is part of the Animikie Group and considered temporally equivalent to the Virginia Formation that is rarely exposed along the Mesabi Iron Range (Stops 2-4, 2-5, and 2-6). Exposures here contain about equal proportions of graywacke, siltstone, and slate, metamorphosed to the greenschist facies. The formation contains abundant carbonate-rich concretions that locally are useful for delineating master bedding in otherwise massive graywacke beds. Graywacke units range in thickness

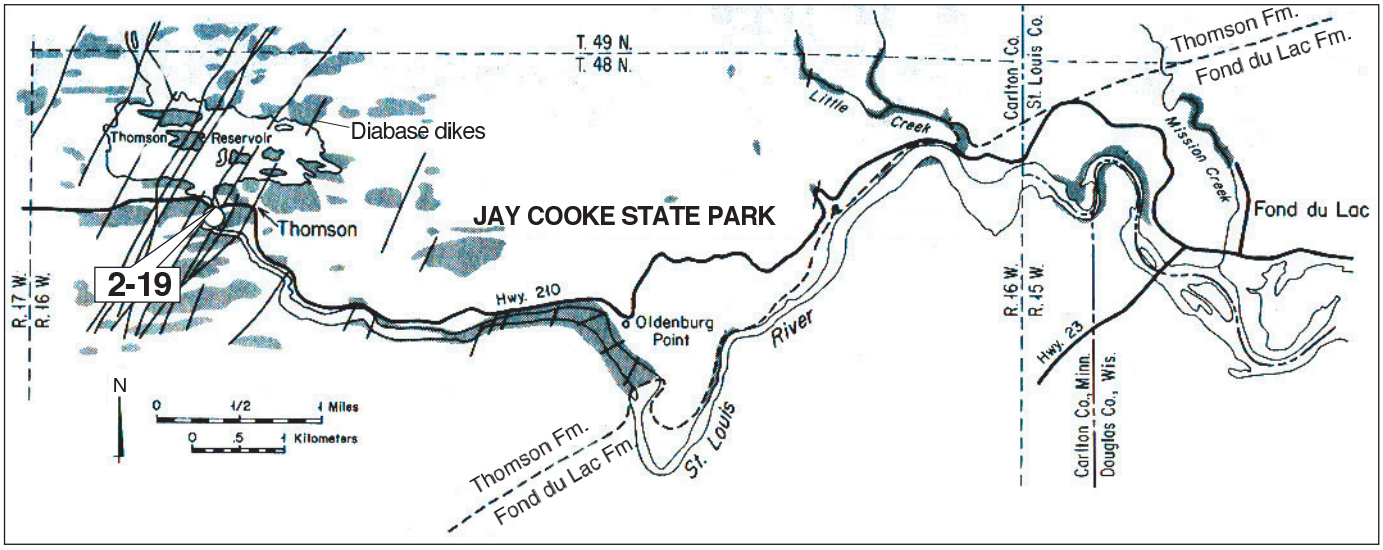


Figure 2.9. Geology and location of the stop in the Fond du Lac, Jay Cooke State Park, and Thomson Dam areas (modified from Jirsa and Morey, 1987).

from 1 inch (2 centimeters) to 14 feet (4 meters), and commonly display sedimentary structures indicative of turbidite deposition, including graded bedding, cross-bedding, sole marks, flute casts, and flame and ball structures (Morey and Ojakangas, 1970). Cross-bedding indicates flow to the south, though the trends of other structures imply more diverse and localized paleoslope directions.

In contrast to the gently dipping Animikie strata along the Mesabi range, the Thomson Formation at this locality is deformed into gentle to open folds at varied scales, presumably related to the Penokean orogeny (Holst, 1984). The fold axes trend east, have vertical to steep south dips, and plunge gently east and west. A well developed axial-planar cleavage is present in the slaty beds, and concretions and mud chips are flattened in the plane of cleavage. The cleavage is deformed locally by kink bands. Quartz veins ranging in width from several centimeters to 3 meters are common in the formation. One of the largest, just north of the Highway 210 bridge, occupies an extensional fracture near the trough of a large syncline. Smaller veins are more contorted and presumably were folded along with the adjacent rock.

Several dikes of ophitic microgabbro ranging in width from a few inches to 200 feet (65 meters) form a dike swarm that occupies northeast-trending joints (Fig. 2.8). The dikes are generally fine- to medium-grained, have chilled, fine-grained margins, and display columnar jointing. Their precise age is unknown; however, their northeast trend, composition, and lack of metamorphism imply they are Mesoproterozoic. Known as the Carlton County dike swarm, the dikes show reversed magnetic polarity and have a paleopole consistent with the 1,108 to 1,105 Ma, early magmatic pulse of the Midcontinent rift (Green and others, 1987).

END OF TRIP

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FIELD TRIP 3

Tuesday, May 17 – Wednesday, May 18

DEPOSITS AND LANDFORMS IN THE REGION GLACIATED BY THE ST. LOUIS SUBLOBE

Leaders

Alan R. Knaeble and Gary N. Meyer, Minnesota Geological Survey
Lisa M. Marlow, U.S. Department of Agriculture—Natural Resource Conservation Service
Phillip C. Larson and Howard D. Mooers, University of Minnesota Duluth

INTRODUCTION

Quaternary deposits in the region glaciated by the St. Louis sublobe have been studied for over a century (Fig. 3.1). Numerous, sometimes contrasting interpretations have been put forth trying to piece together the glacial history of this region. This field trip will review and discuss past research and present new ideas based on recent field mapping. The first day of the field trip will be led by staff members of the Minnesota Geological Survey and will focus on sediments and geomorphology along the western end of the St. Louis sublobe. Over the past five years, the Minnesota Geological Survey has completed the Crow Wing County atlas (Knaeble and Meyer, 2004; Knaeble and others, 2004) and three U.S.G.S 7.5-minute quadrangle (Brainerd, Gull Lake, and Baxter; Hobbs, 2001a, b; Knaeble, 2001) mapping projects that provide a large influx of new observations and analyzed samples. These new data are the primary basis for interpretations suggested on day one. The second day of the field trip will be led by current and former staff from the University of Minnesota Duluth and will examine deposits and landforms located in the central and eastern areas of the St. Louis sublobe. Based on their recent research, they suggest a revised age date for the advance of St. Louis-sublobe ice and for the formation of lobe related glacial lakes—glacial Lakes Aitkin and Upham I and II. These two research groups have added a significant base of information about the glacial geology of the region, but due to the large area covered by the St. Louis sublobe and the complex regional stratigraphy, there is a lot more work to be done before the region's glacial history is thoroughly understood.

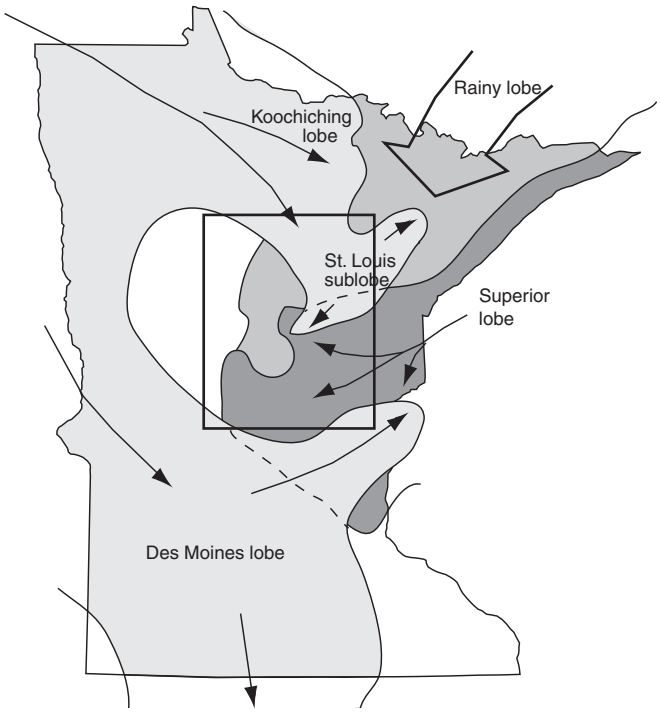


Figure 3.1. Ice-lobe phases with approximate flow direction and ice-marginal extent of each lobe (modified from Knaeble and Meyer, 2004). Box shows the day one field trip area. During the last glaciation, the Late Wisconsinan, the north-central region of Minnesota was covered by ice of the Rainy lobe, which moved into the area from the northeast. Following the retreat of the Rainy lobe, ice of the Superior lobe advanced into the region from the east, covering the eastern and southern portion of the area. Finally, after Superior-lobe ice retreated, ice of the St. Louis sublobe entered the area from the northwest.

DAY 1—DEPOSITS ALONG THE WEST END OF THE ST. LOUIS SUBLOBE

Alan Knaeble and Gary Meyer

INTRODUCTION TO DAY 1

In this field trip description, Quaternary unit names that have been formalized according to the rules of the North American Stratigraphic Code will be used along with commonly used informal names. A committee is in the process of reviewing all of Minnesota's Quaternary stratigraphic unit names and a forthcoming Minnesota Geological Survey Report of Investigations publication will summarize the committee's additions and alterations to these units. Some of these new and revised unit names will be introduced here.

PREVIOUS RESEARCH

Glacial sediment associated with distinct ice lobes was recognized by Upham (1896-1899) and Winchell (1901) in north-central and northeastern Minnesota in and around the region considered to have been covered by the St. Louis sublobe (Fig. 3.1). Leverett (1932) described these drift deposits in detail and then differentiated them on the basis of color, texture, lithology, stratigraphic position, and geomorphology. He observed that the youngest diamicton in the area was a thin, gray till everywhere except south of the Mesabi Iron Range, where it was gray-brown and red, the reddish color interpreted to be incorporation of oxidized iron-formation bedrock (Fig. 3.2). Leverett noted its maximum extent and suggested a name, the St. Louis sublobe. He also recognized the lobe-associated lacustrine sediments of glacial Lakes Aitkin and Upham. His research identified shoreline

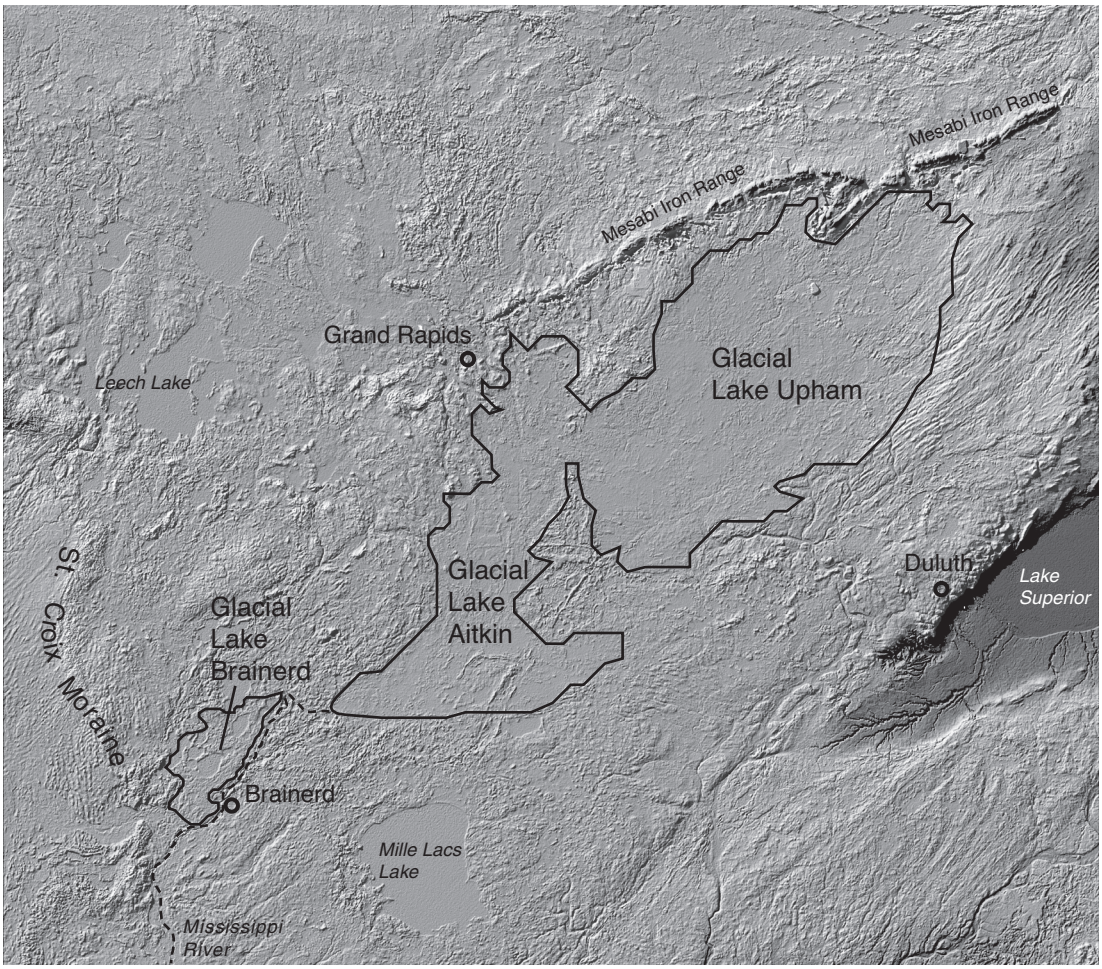


Figure 3.2. Hill-shade relief for the land surface for north-central Minnesota, showing glacial lake basins, geomorphic features, drainage paths (dashed lines), and a few modern-day lakes and cities.

deposits, delineated the extent of the lake basins, and proposed that the lakes drained down the Mississippi River channel. Cooper (1935, 1938) documented the existence of dunes around Brainerd and east of the Mississippi River.

Later studies by Wright (1954, 1955, 1956) hypothesized that the red deposits observed by Leverett south of the Mesabi Iron Range in the glacial Lakes Upham and Aitkin basin (Fig. 3.2) could be attributed to a Superior-lobe ice advance (Valdors phase), a northward protrusion of the ice lobe that formed the Mille Lacs moraine, which took place prior to the advance of the St. Louis sublobe into that region. He concluded that there was no evidence validating the idea that the incorporation of iron from the bedrock was responsible for the red color of the till south of the Mesabi range, as Leverett had proposed. Schneider (1961) studied older sediments and St. Louis-sublobe deposits along the western margin of the sublobe. He also examined drainage patterns in that region. Later studies by Baker (1964) and Wright (1964, 1972; Wright and Ruhe, 1965; Wright and Watts, 1969; Wright and others, 1973) caused Wright to alter his original hypothesis and he postulated that the red till deposits south of the Mesabi range were indeed St. Louis-sublobe deposits as Leverett originally suggested. The revised interpretation proposed that St. Louis-sublobe ice advanced into the glacial Lakes Aitkin-Upham basin from the northwest, spread to the northeast and southwest, and in places incorporated red glacial Lakes Upham I and Aitkin I sediment into the ice, turning the St. Louis-sublobe sediment-load red. Geomorphic features and exposures of interbedded red and gray-brown till provided evidence for this interpretation. Winter's (1971; Winter and others, 1973) conclusions supported Baker and Wright's idea, but he questioned whether the gray-brown and red tills were deposited by ice from the same northwest source area because they have differences in color, particle size, and percentage of northwest indicator clasts.

Subsequent work by Hobbs (1983; Hobbs and Goebel, 1982; Lehr and Hobbs, 1992) added detail about the extent of St. Louis-sublobe deposits and investigated post-glacial drainage of glacial Lakes Aitkin and Upham. Mooers (1988, 1990a, b) constructed a glacial history and investigated ice dynamics of the Rainy lobe—the ice lobe occupying and then retreating from the area prior to the advance of the St. Louis sublobe. Meyer (1989, 1993, 1997) researched and mapped the St. Louis-sublobe deposits north of the Mesabi Iron Range and the Grand Rapids area. He suggested the name "Koochiching lobe" for the ice that covered this region and restricted the name St. Louis sublobe for ice that extended south of the Mesabi Iron Range and occupied the glacial Lakes Aitkin I and Upham I basin (Figs. 3.1, 3.2). Anderson (1998) mapped a portion of the Mille Lacs moraine on the southwest side of Mille Lacs Lake. Recent mapping by Knaeble and Meyer (2004) in Crow Wing County around the western end of the St. Louis sublobe defined the boundaries and sediment composition of glacial Lake Brainerd and redefined stratigraphic relationships in the area. Their conclusions, based on lithology, stratigraphic position, and geomorphology, proposed that the red till north, west, and south of the glacial Lake Aitkin basin is associated with the Superior-lobe ice advance that formed the Mille Lacs moraine, and that the overlying brown till encircling the "snout" of the glacial Lake Aitkin basin was deposited by St. Louis-sublobe ice. Gray glacial Lake Aitkin I sediments containing a few thin red layers commonly separate the two till units in places.

Marlow and others (2004) and Marlow (2004) recently added detail to the drainage history particularly in the eastern region occupied by the St. Louis sublobe. They also proposed an earlier date than previous research for the advance of the St. Louis sublobe based on drainage, geomorphic, and stratigraphic relationships. Recent mapping by Meyer and others (2004) in Itasca County and mapping by Jennings (unpub. data) along the Mesabi Iron Range added further perspective to the region's complex history.

SUMMARY OF REGIONAL GLACIAL GEOLOGY

This summary will provide detail about the complex sequence of glacial materials and depositional environments that will be seen on the first day of the field trip in the western region of the St. Louis sublobe, the area west and north of Mille Lacs Lake (Fig. 3.3). Sediment and landforms record multiple ice advances and retreats from the major source areas in northern Canada—the Keewatin ice center northwest of Minnesota, and the Labrador ice center northeast of Minnesota (Fig. 3.4; Meyer and Knaeble, 1996). Keewatin ice advanced into northwest Minnesota along the present-day Winnipeg lowlands, and ice from the Labrador center advanced into northeast Minnesota through the Lake Superior basin (Wright, 1972). Labrador-sourced ice also entered the area from a more northerly direction after traversing the highlands of the Canadian shield north of the Lake Superior basin. Four primary source areas: Riding

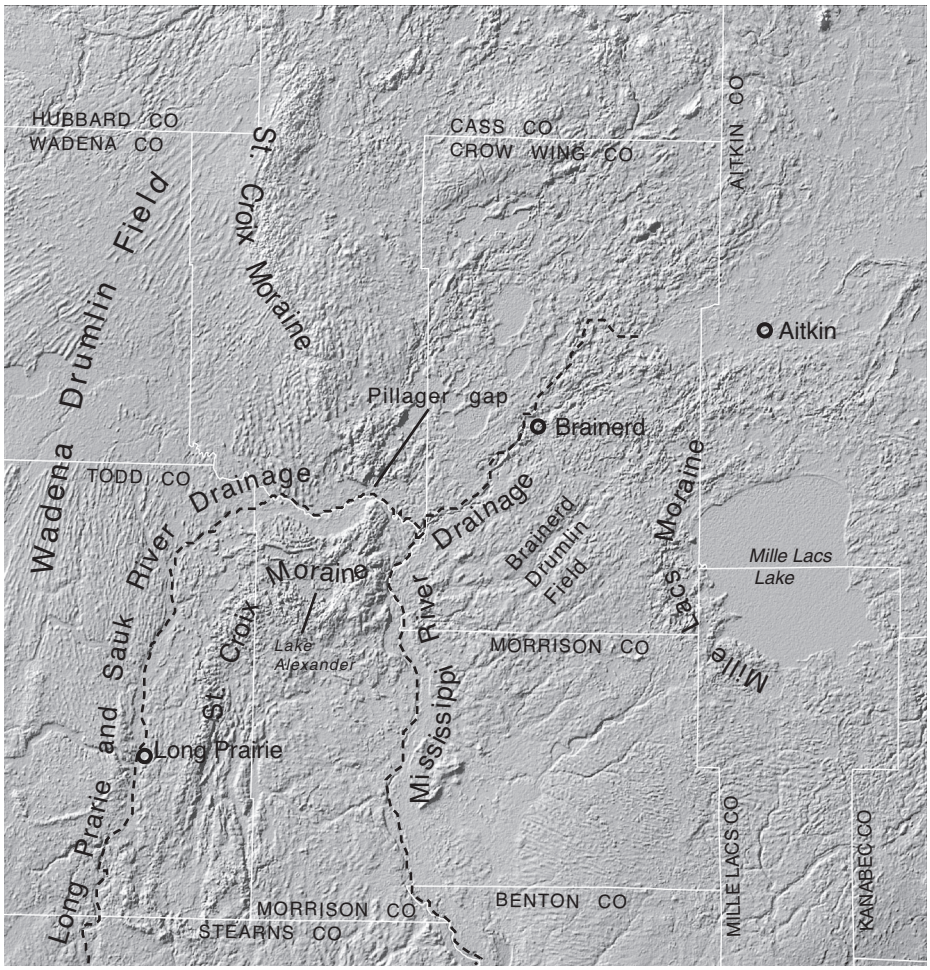


Figure 3.3. Hill-shade relief for the land surface for north-central Minnesota, showing the region of the field trip, geomorphic features, drainage paths (dashed lines), county boundaries, and a few modern-day lakes and cities.

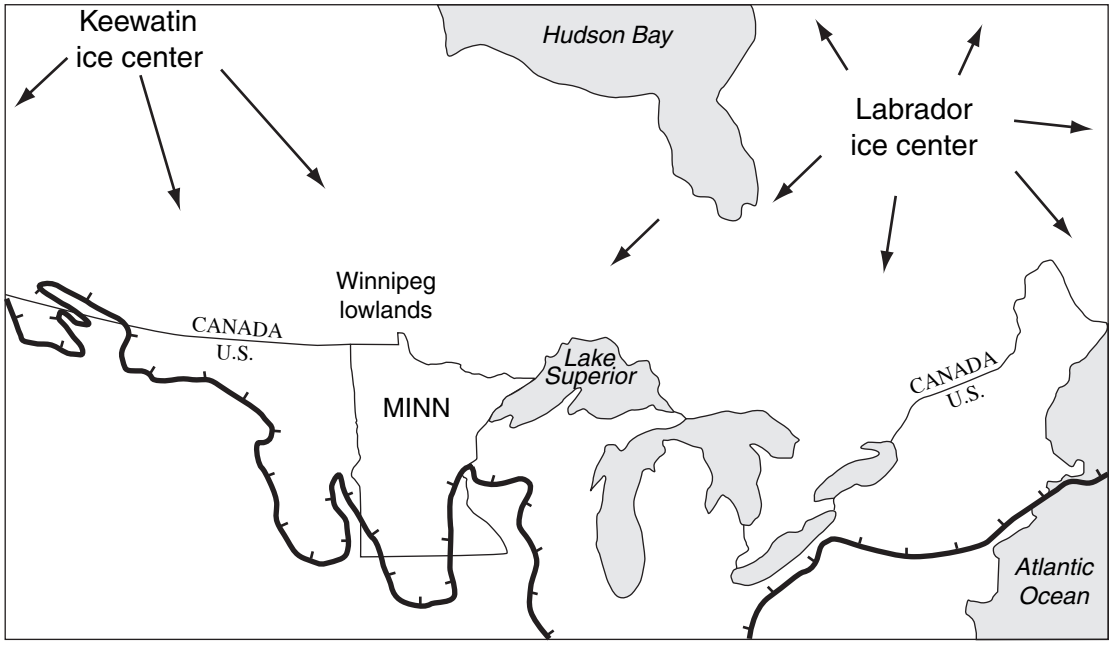


Figure 3.4. Location of Laurentide Ice Sheet accumulation centers. Includes ice lobe flow directions and Late Wisconsinan ice margins (hachured line) at approximately 14,000 years ago (from Knaeble and others, 2004; modified from Lusardi, 1998).

Mountain, Winnipeg, Rainy, and Superior have been defined for Minnesota glacial sediment, based on distinctive rock types found in their respective deposits (Fig. 3.5). Surface deposits across the area were laid down during or after the last glaciation, the Late Wisconsinan, which lasted from about 35,000 to 10,000 years ago (Wright, 1972; Hobbs and Goebel, 1982; Knaeble and others, 2004). Beneath the Late Wisconsinan glacial materials, older Pleistocene drift is present in places where it has not been removed by erosion.

Pre-Wisconsinan stratigraphy

Precambrian bedrock that varies from Archean (approximately 2,700 Ma in age) to Paleoproterozoic (approximately 2,200 to 1,800 Ma in age) underlies the glacial drift of this area (Boerboom and Chandler, 2004). In places where erosion has not removed the upper weathered surface of the bedrock, a clayey saprolith remains. Remnant marine deposits of siltstone and shale related to the Cretaceous interior seaway may be found scattered around the area overlying the Precambrian bedrock (Boerboom and Chandler, 2004).

Knaeble and Meyer (2004) identified pre-Wisconsinan drift deposits associated with the Winnipeg, Rainy, and Superior source areas (Fig. 3.6) from outcrop, Giddings probe auger borings, and rotary-sonic core samples. These deposits correlate with till of sequence x and sequence w units of central Minnesota (Fig. 3.7). Pre-Wisconsinan drift is deeply buried by Late Wisconsinan materials, except for a few places where it is at or near the surface. This is probably due to erosion of younger materials or glacial tectonic thrusting.

Late Wisconsinan stratigraphy

Rainy lobe

Labrador-center ice, which advanced to the southwest and passed over the highlands northwest of the Superior basin, deposited materials associated with the Rainy lobe. In this summary, Rainy-lobe ice that passed north of the Mesabi Iron Range (Fig. 3.8) is considered the Wadena lobe (Wright, 1957, 1962, 1972), a sublobe of the Rainy lobe. The Brainerd lobe of Schneider (1961) is considered the sublobe of the Rainy lobe that passed south of the Mesabi Iron Range.

Wadena lobe

Even though deposits of the Wadena lobe are at the surface in the Wadena drumlin field in

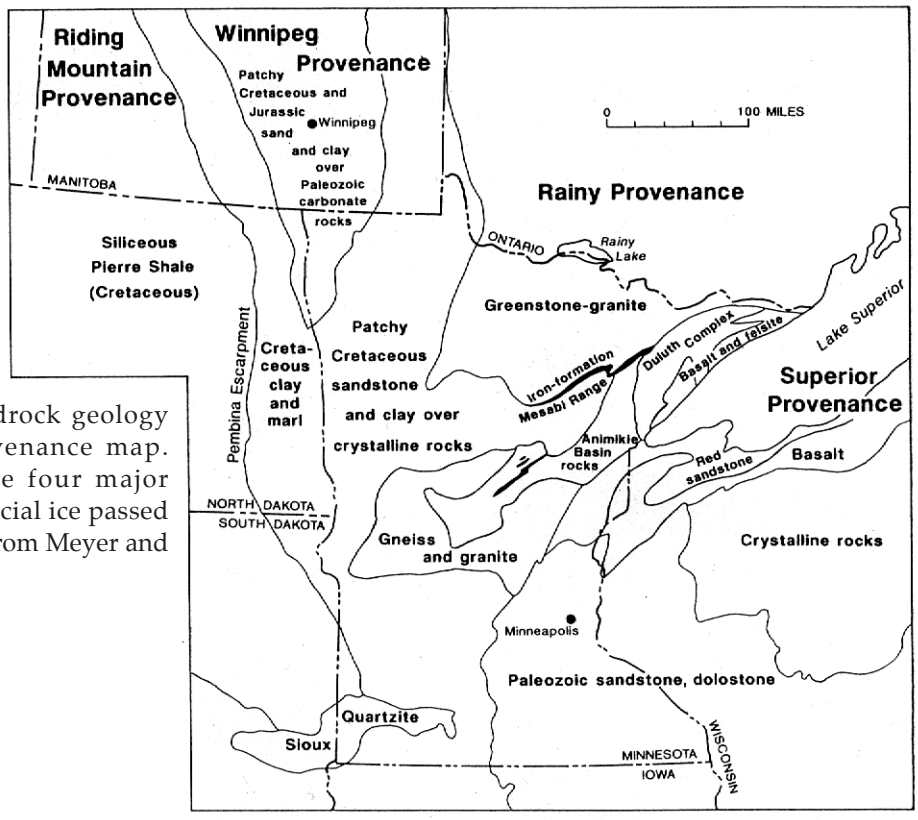


Figure 3.5. Simplified bedrock geology and source-material provenance map. Shows the location of the four major provenances over which glacial ice passed before entering Minnesota (from Meyer and Knaeble, 1996).

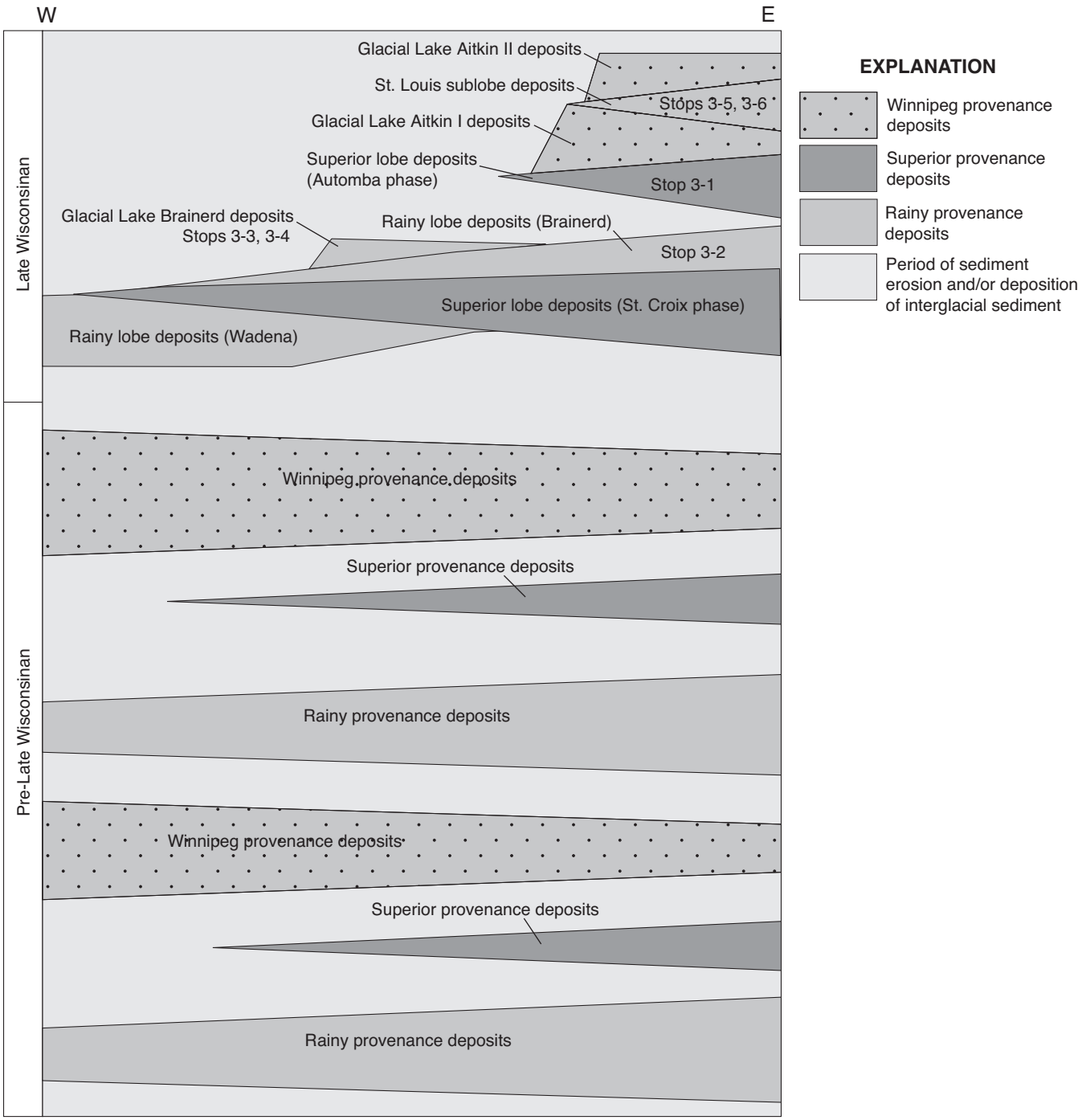


Figure 3.6. Diagram showing relative age and deposit locations (across the region of the field trip) related to the Late Wisconsinan ice lobes, their associated glacial lakes, and the pre-Late Wisconsinan ice lobe provenances. The age column and deposit drawings are not to scale (modified from Knaeble and Meyer, 2004).

central Minnesota, these materials are found almost exclusively in the subsurface east of the St. Croix moraine, where they have been covered by younger Brainerd-lobe and Superior-lobe drift (Fig. 3.3). Knaeble and Meyer (2004) did not encounter any Wadena-lobe till, even in the subsurface, east of the Mississippi River in Crow Wing County. Wadena-lobe till is characteristically yellow-brown where oxidized, gray where unoxidized, and a sandy loam with an average matrix texture of 60 percent

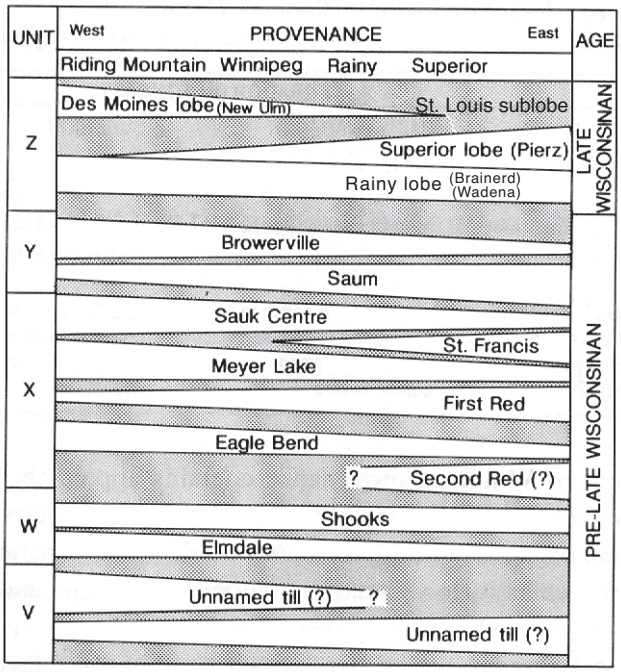


Figure 3.7. Diagram showing the relative timing, bedrock source-material provenance, and relative extent of ice advances that deposited till in central Minnesota (modified from Meyer and Knaeble, 1996).

sand, 28 percent silt, and 12 percent clay (Mooers, 1988, p. 64-65; Meyer and Knaeble, 1996; Harris and others, 1999; Meyer and others, 2001, unpub. data). Other distinguishing characteristics are the virtual absence of Cretaceous shale and dark limestone, and a Paleozoic carbonate clast content ranging from 10 to 25 percent. Three recent studies that analyzed the very coarse-grained (1 to 2 millimeters) sand fraction show an average of 17 percent carbonate for 28 samples (Meyer and Knaeble, 1996), 14 percent carbonate for 118 samples (Meyer and others, 2001; unpub. data), and 18 percent carbonate for 140 samples (Harris and others, 1999).

Brainerd lobe

In contrast to the Wadena lobe, Brainerd-lobe deposits are at the surface east of the lobe's western ice margin, the St. Croix moraine. The nose of the Brainerd lobe appears to have extended south of the Pillager gap to the southeast side of Lake Alexander, and then it curved back to the east in the vicinity of the Crow Wing–Morrison County lines (Mooers, 1988). Its deposits extend eastward over the western half

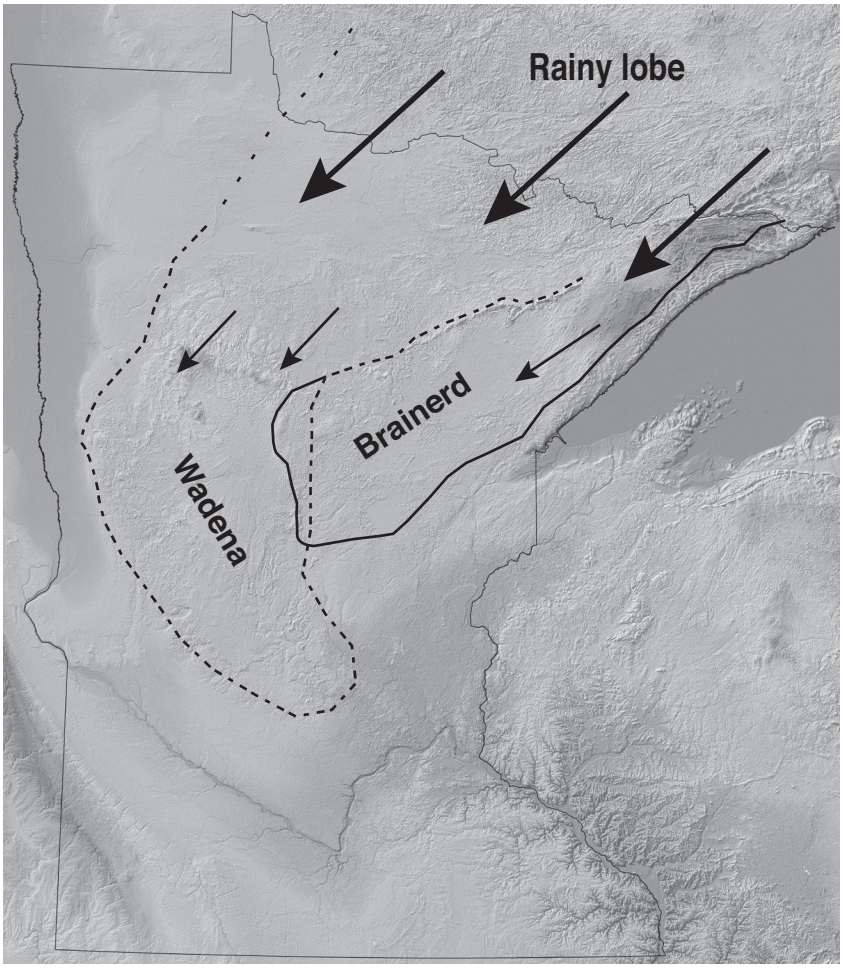


Figure 3.8. Hill-shade relief for the land surface for Minnesota, showing Rainy-lobe flow direction, and the flow direction and approximate extent of its two sublobes, the Wadena and the Brainerd lobes.

of Crow Wing County, where they form the Brainerd drumlin field, a series of recessional moraines, and numerous outwash plains (Fig. 3.3; Schneider, 1961; Mooers, 1988; Hobbs, 2001a, b; Knaeble 2001; Knaeble and others, 2004). Brainerd-lobe materials are covered by younger Superior-lobe and St. Louis-sublobe deposits north of Mille Lacs Lake and are beneath Superior-lobe deposits on the west side of Mille Lacs Lake where they form the core of the Mille Lacs moraine (Mooers, 1988; Anderson, 1998; Knaeble and Meyer, 2004; Knaeble and others, 2004).

Brainerd-lobe till (to be formally named the Long Lake member of the Brainerd formation) is a brown (commonly 7.5YR 5/4 but ranging to 10YR 5/4) color where oxidized and is gray where unoxidized. In places near the Cuyuna Iron Range, the color may be as red as 5YR 5/4 due to incorporation of red iron ore present in the bedrock (Knaeble and others, 2004). The till is characteristically a sandy loam with an average matrix texture of 67 percent sand, 22 percent silt, and 11 percent clay for 194 samples (Knaeble and Meyer, 2004). The till is commonly leached of carbonate to a depth of 10 feet (3.1 meters) or more. Thus, analyses of the 1 to 2 millimeter sand fraction of these samples showed only trace amounts of carbonate. Deep, unleached samples of Brainerd-lobe till contained 3 to 4 percent carbonate (Knaeble and Meyer, 2004). Sand-grain analyses for 164 of the matrix texture samples referred to above found that of the Precambrian 1 to 2 millimeter grains, 15 percent were dark (mafic-igneous and metamorphic rocks) and 7 percent were red (volcanics, felsites, iron-formation, agate, or sandstones; Knaeble and Meyer, 2004). An exposure of Brainerd-lobe drift is visible at Stop 3-2.

Glacial Lake Brainerd

While the Superior lobe and the Brainerd lobe were at the St. Croix moraine, drainage along the western ice margin flowed south in the same valleys that the present day Long Prairie and Sauk Rivers now occupy (Fig. 3.3). As Brainerd-lobe ice retreated to the northeast, it formed a series of recessional moraines (Mooers 1988; Knaeble and others, 2004). As the ice continued to retreat, its meltwater eroded and carried sediment that eventually buried substantial portions of these moraines. Meltwater from the Brainerd lobe ponded behind the St. Croix moraine northeast of the Pillager gap and formed glacial Lake Brainerd (Goldstein, 1985; Mooers, 1988). Glacial Lake Brainerd sediments are to be formally named the Glacial Lake Brainerd member of the Brainerd formation. Recent mapping in Crow Wing County indicated that the lake was in existence for at least 100 years (based on varves and other accumulated sediments), had stagnant ice buried beneath its sediments, and its maximum extent was approximately 150 square miles (388 square kilometers; Fig. 3.9; Knaeble and Meyer, 2004; Knaeble and others, 2004). A layer commonly greater that 20 feet (6 meters) thick of fine- to medium-grained sand covers the entire lake basin except for areas in the northeast corner where meltwater streams deposited sand and gravel as they entered the lake. In places, the fine-grained sand was reworked into dunes by wind after the lake had drained. Silt and clay, locally varved, are found in places beneath the uniform sand layer, particularly from the city of Brainerd northward to Merrifield, and then east as far as the Mississippi River (Fig. 3.9). The lake may have drained to the west through the Pillager gap, or possibly south down the Mississippi River, if the river's drainage had opened up by this time (Mooers, 1988). Glacial Lake Brainerd sediment is exposed at Stops 3-3 and 3-4.

Superior lobe

Ice of the St. Croix phase of the Superior lobe deposited Cromwell Formation drift (Wright, 1972), which formed the southern segment (from the Pillager gap south) of the St. Croix moraine. The Superior and Brainerd lobes were apparently confluent in the vicinity of the Crow Wing–Morrison County line and east to the southern end of Mille Lacs Lake (Fig. 3.3; Knaeble and others, 2004). As both lobes retreated to the east, the Brainerd lobe left a series of recessional moraines, one of which formed the core of the Mille Lacs moraine (Mooers, 1988; Anderson, 1998; Knaeble and others, 2004). The ice masses continued to retreat further east and a large pro-glacial lake formed in and around the basin that now holds Mille Lacs Lake. The Superior lobe then re-advanced during the Automba phase (Wright, 1972) across the basin incorporating fine-grained pro-glacial lake sediment into the ice and subsequently depositing this sediment on top of the Brainerd-lobe materials that formed the core of the Mille Lacs moraine (Mooers, 1988; Anderson, 1998; Knaeble and Meyer, 2004; Knaeble and others, 2004). At this time, Brainerd-lobe ice must have remained further to the northeast because the margin of the Superior-lobe deposits extends north and then forms an arc, which passes through the Cuyuna iron-mining district, east of Mission Lake, and south of Emily before exiting the eastern side of Crow Wing County (Fig. 3.10; Knaeble and Meyer, 2004; Knaeble and others, 2004). This interpretation is based on the textural and lithologic data described below and on description and sample evidence from 15 drill holes located around the rim of the glacial Lake Aitkin

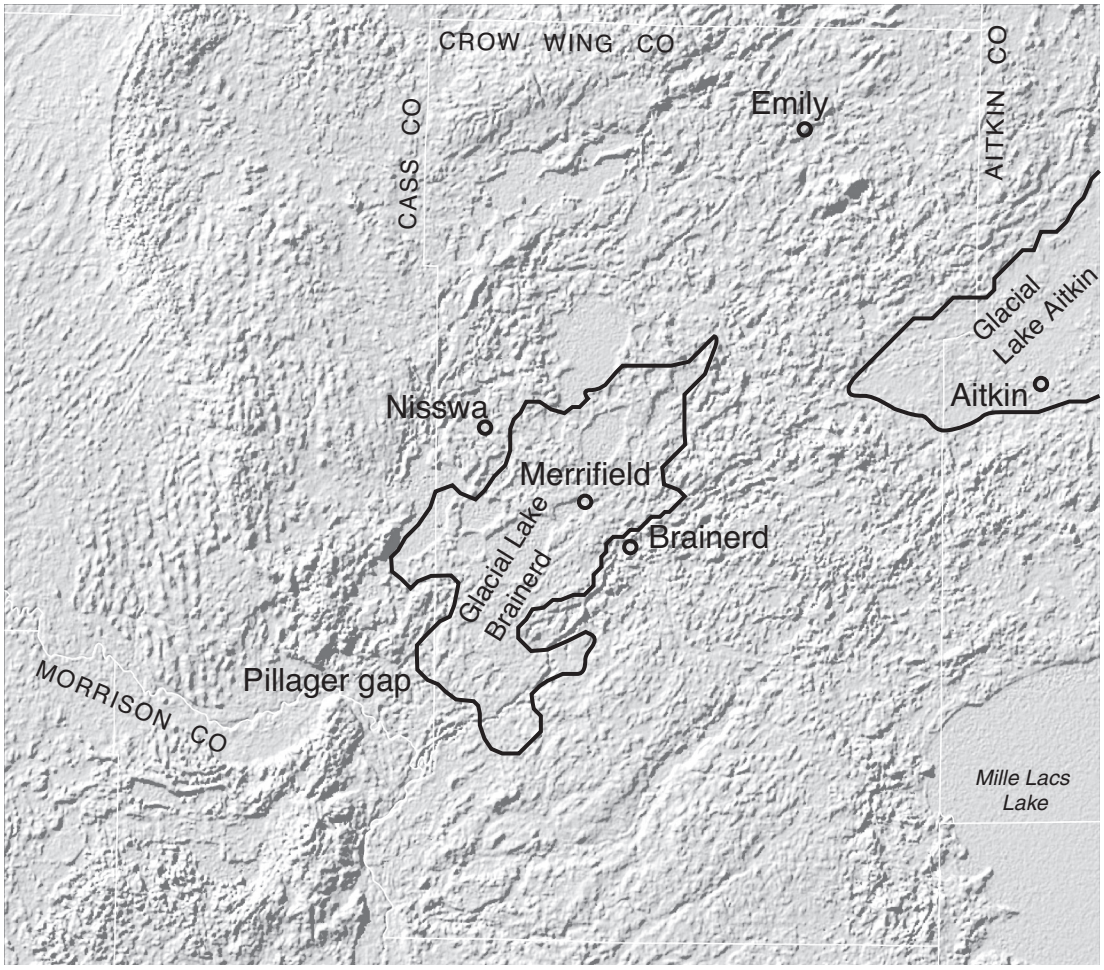


Figure 3.9. Hill-shade relief for the land surface for north-central Minnesota, showing glacial Lakes Aitkin and Brainerd, the Pillager gap, and modern-day counties, cities, and lakes.

basin, where brown St. Louis-sublobe till overlies red Superior-lobe till. This interpretation differs from that of Hobbs and Goebel (1982), which attributed the red drift north of Mille Lacs Lake to the St. Louis sublobe. Further east, the extent of Superior-lobe deposits has yet to be determined. Superior-lobe deposits along the rim and in the glacial Lake Aitkin basin have been covered by younger St. Louis-sublobe and glacial Lakes Aitkin I and II materials.

The fine-grained till of the Automba phase has been informally named the Garrison member of the Cromwell formation (Knaeble and others, 2004). Garrison member till deposits of the Superior lobe are reddish-brown (5YR 5/4) in color where oxidized, dark gray (5YR 4/1) where unoxidized, and generally have high silt content. They average a loam matrix texture of 37 percent sand, 43 percent silt, and 20 percent clay for 184 samples (Knaeble and Meyer, 2004). Although individual samples may be highly variable, most are fine-grained, apparently due to incorporation of silty, pro-glacial lake sediment. The depth of carbonate leaching varies, ranging from 3 to 9 feet (1 to 3 meters). Due to its fine-grained texture, the Garrison till is not as deeply leached as the Brainerd-lobe till (Knaeble and Meyer, 2004). Contrary to earlier interpretations (Hobbs and Goebel, 1982), lithologic characteristics clearly define the Garrison member as having a Superior provenance. Analysis of 127 samples found that the Precambrian 1 to 2 millimeter grains averaged 30 percent dark (mafic-igneous and metamorphic rocks) and 14 percent red (volcanics, felsites, iron-formation, agate, or sandstones; Knaeble and Meyer, 2004). Exposures of this till are visible at Stops 3-1, 3-7, and 3-8.

Meltwater from the southern flank of the Superior lobe flowed from the Mille Lacs moraine ice margin to the southwest through the Brainerd drumlin field via the Nokasippi River and other channels into the

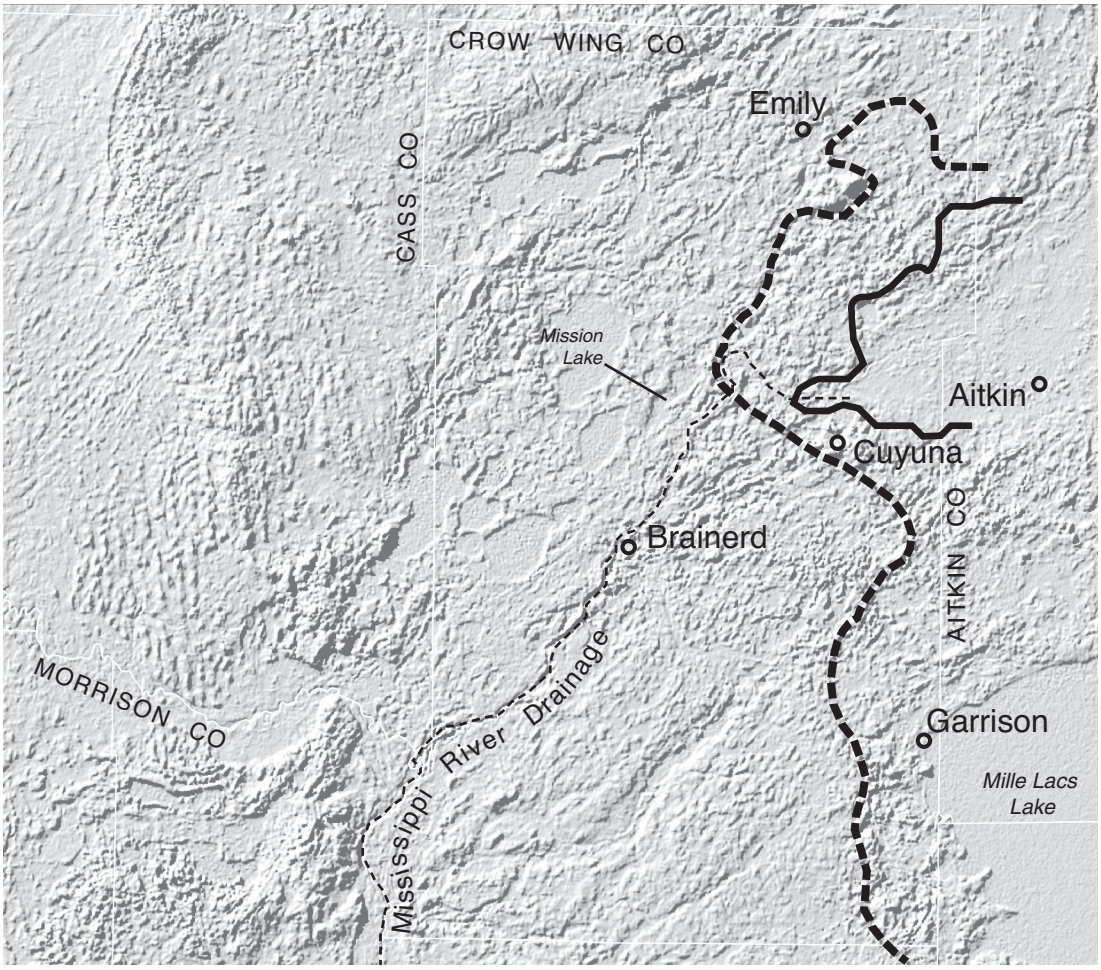


Figure 3.10. Hill-shade relief for the land surface for north-central Minnesota, showing the approximate ice margin of the Automba phase of the Superior lobe (heavy dashed line) and St. Louis sublobe (heavy black line), Mississippi River drainage (dashed line), and modern-day counties, cities, and lakes.

south-flowing Mississippi River (Fig. 3.3; Hobbs and Goebel, 1982; Knaeble and others, 2004). Drainage evidence is not as obvious for the northwestern flank of the Superior-lobe ice margin. Its meltwater may have entered glacial Lake Brainerd, if the lake was still in existence, because evidence of red varved sediments exists along the eastern side of the lake basin. If glacial Lake Brainerd had already drained, meltwater most likely flowed down the Mississippi River; however, from the city of Brainerd northward the Mississippi River has a small channel compared to that of the Nokasippi River (Knaeble and others, 2004). Outwash sediments of the Brainerd lobe are lithologically similar to those of the Superior lobe, making differentiation between the two difficult.

Glacial Lake Aitkin I

Glacial Lake Aitkin I formed as ice of the Automba phase of the Superior lobe melted and withdrew from the region (Figs. 3.2, 3.9). The lake may have remained in existence until the St. Louis sublobe advanced into the area. It became a proglacial lake of the St. Louis sublobe as the ice spread across the glacial Lakes Aitkin–Upham basin. The maximum extent of the lake is difficult to determine because its deposits are buried beneath sediments of the St. Louis sublobe and glacial Lake Aitkin II (Fig. 3.10). Auger samples indicate that glacial Lake Aitkin I sediments were deposited around the rim of the glacial Lake Aitkin II basin at higher elevations than glacial Lake Aitkin II sediments, suggesting that in this area glacial Lake Aitkin I was more extensive than glacial Lake Aitkin II (Knaeble and Meyer, 2004). Sediments associated with the southwestern end (the city of Aitkin westward) of this glacial lake were predominantly gray silt and clay where seen in outcrop and in boring samples. Laminations or

varves are present in places. Thin, red, clayey silt layers are interbedded with gray lacustrine sediment at some sites, but they represent only a minor fraction of the total lake deposits. Some uniform fine-grained sand sequences were also observed (Knaeble and Meyer, 2004). These sediments are to be formally named the Glacial Lake Aitkin I member of the Aitkin formation. From these observations it appears that much of the lake sediment on this end of the lake basin was derived from sediment carried by St. Louis-sublobe ice and that contributions of red Superior-lobe sediment were less significant. Stop 3-5 is an exposure of glacial Lake Aitkin I deposits, which are overlain by St. Louis-sublobe till.

Des Moines lobe

Des Moines-lobe ice did not cover the area but drainage from its melting ice flowed east through the Pillager gap, eroding and depositing sediment in the Crow Wing and Mississippi River channels.

St. Louis sublobe

The St. Louis sublobe, the southeastern protrusion of the Koochiching lobe (a tongue of ice that extended eastward from the Winnipeg lowlands across northern Minnesota), was the last ice advance to enter the region (Fig. 3.1). The ice passed over the Mesabi Iron Range near Grand Rapids and then fanned out to the northeast and southwest, advancing into glacial Lakes Aitkin I and Upham I. The maximum extent of the ice covered these glacial-lake basins and then lapped up onto the surrounding terrain (Fig. 3.10). Fifteen auger sites at the southwest corner of the sublobe detected brown St. Louis-sublobe till overlying red Superior-lobe till, and seven of these sites had glacial Lake Aitkin I sediments separating the tills (Knaeble and Meyer, 2004; Knaeble and others, 2004). St. Louis-sublobe deposits near the ice margin indicate by color and lithology that underlying red Superior-lobe drift was incorporated into the ice and mixed, to varying degrees, with St. Louis-sublobe materials, particularly in the basal portion of the unit.

St. Louis-sublobe till has been informally named the Nelson Lake member of the Aitkin formation and in the region west of Aitkin is characteristically yellow-brown to brown (10YR 5/4 to 7YR 5/4) where oxidized and dark gray (10YR 4/1) where unoxidized. The till has an average matrix texture of 24 percent sand, 39 percent silt, and 37 percent clay for 80 samples (Knaeble and Meyer, 2004). Exposures commonly had some pebbles scattered throughout the upper portion of the unit, but the clasts diminished with depth as the till appeared to grade into sediment of glacial Lake Aitkin I. Apparently as the ice was advancing into glacial Lake Aitkin I it was incorporating the underlying fine-grained lake sediments. As these materials mixed with those in the ice, the resulting till deposits became rich in silt and clay with a diluted pebble population. The till is commonly leached of carbonate to a depth of 3 to 7 feet (1 to 2.1 meters). Analysis of the 1 to 2 millimeter sand fraction of 70 samples from Crow Wing County yielded an average of 7 percent carbonate. Some individual samples had a carbonate content close to 20 percent. Gray, siliceous, Pierre shale was absent or found in only trace amounts. Agate and iron-formation along with red volcanic and sandstone grains averaged only 2 percent of the Precambrian grains (Knaeble and Meyer, 2004). St. Louis-sublobe till exposures will be seen at Stops 3-5 and 3-6.

A second suite of 6 till samples collected near and east of Aitkin suggests the possibility that a second ice advance or phase of the St. Louis sublobe overlies the St. Louis-sublobe materials described above. This till is gray with a loam texture. In the 1 to 2 millimeter grain size it averages a carbonate content of 15 percent and gray silicious shale content of 32 percent (Knaeble and Meyer, 2004). This unit is found at the surface or overlying brown, shale-poor, St. Louis-sublobe till. Schneider suggested this was a distinct unit in his unpublished research during the 1950s. His descriptions and some of his samples, collected at that time and stored, were used to map Crow Wing County and decipher the glacial geology of the region. Further detailed work needs to be done to determine if this lithologically distinct unit is a separate, younger phase of the St. Louis sublobe or if it is a product of lithologic differentiation due to separate ice-steam flow.

Glacial Lake Aitkin II

Melting St. Louis-sublobe ice left little evidence of outwash deposits or ice marginal meltwater channels. As the sublobe stagnated its meltwater ponded, forming glacial Lake Aitkin II. Early drainage for this lake was probably along a series of ice marginal lakes, which emptied into glacial Lake Upham II, and subsequently to outlets farther to the east. As differential isostatic rebound uplifted the northern region of the basin more rapidly than the southern region, drainage shifted to the western margin where the final outlet became the southwest flowing Mississippi River (Hobbs, 1983; Marlow and others, 2004). The channel of the Mississippi River is very narrow where it exits the glacial Lake Aitkin basin, suggesting that glacial Lake Aitkin II drainage was in its final stages when the river became an outlet (Knaeble and others, 2004).

A subtle topographic break at an elevation of 1,225 feet delineates a beach ridge visible in places on both the north and south sides of the basin in Crow Wing County (Fig. 3.10). Glacial Lake Aitkin II sediments (to be formally named the Glacial Lake Aitkin II member of the Aitkin formation) in this area consist chiefly of well-sorted fine-grained sand, but also include coarser-grained sand or silt and clay. Some of this fine-grained sand has been blown into dunes. The thickness of these lake deposits was commonly 15 feet (5 meters) or less, but a few sites were as thick as 30 feet (10 meters). These lake sediments overlay St. Louis-sublobe till deposits (Knaeble and Meyer, 2004; Knaeble and others, 2004).

ACKNOWLEDGMENTS

Figures and tables were designed by Rich Lively and Barbara Lusardi. Thanks are extended to all the land owners and gravel pit owners who allowed access to their property.

FIELD TRIP STOPS (Fig. 3.11)

DIRECTIONS: Leave the Radisson Hotel in Minneapolis; west ~0.3 miles on Washington Avenue to Interstate 35W. Go North ~10 miles to U.S. Highway 10. West ~22 miles to U.S. Highway 169. North ~68 miles to Mille Lacs County Road 25. West ~3 miles (County Road 25 changes to County Road 2 in Crow Wing County) to Crow Wing County Road 138. South ~0.5 miles on township road.

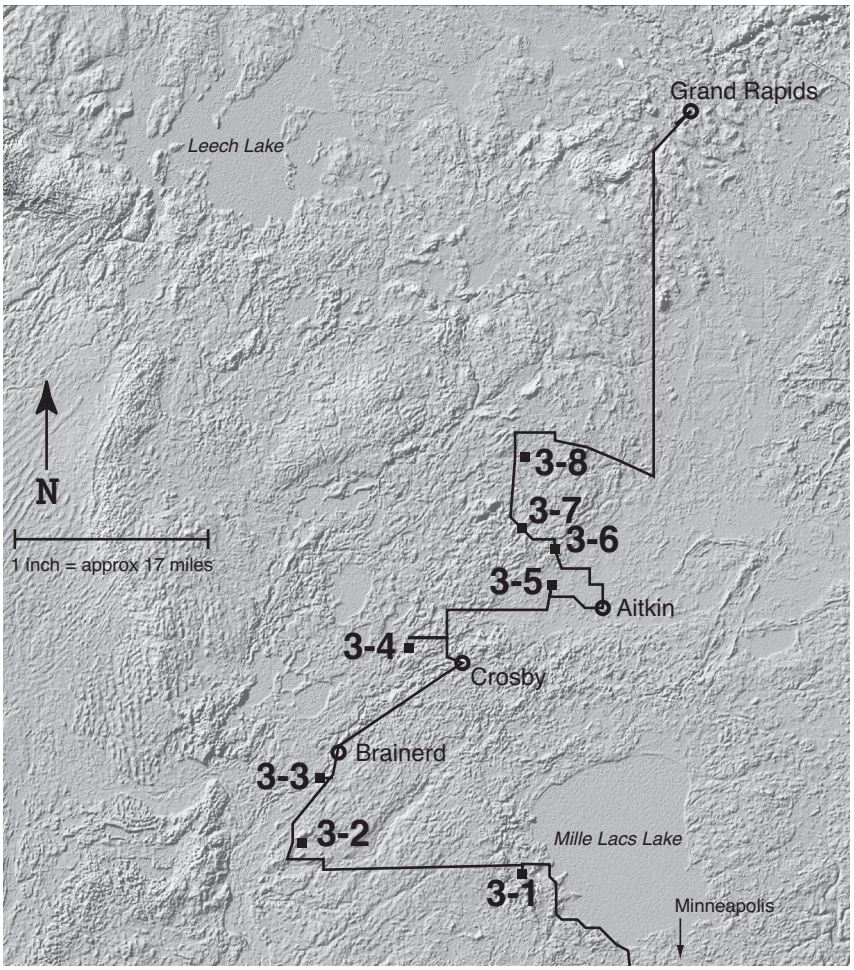


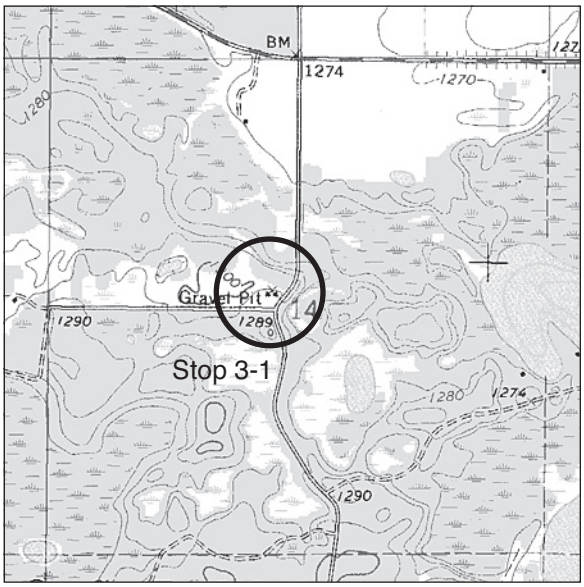
Figure 3.11. Map of the field trip route.

STOP 3-1

Lingwall pit—Superior-lobe deposits (till [Automba phase] and sand/gravel) in the Mille Lacs moraine

Location: T. 43 N., R. 28 W., sec. 15, center

Vineland quadrangle



Description: The Lingwall pit exposes sand and gravel and till of the Garrison member of the Cromwell formation of Superior provenance. Reddish brown, fine-grained silty till up to 9 feet (3 meters) thick caps reddish sand and gravel, cut in places by "clastic dikes" of similar reddish brown silty till. The till was apparently squeezed up from below through cracks in the frozen sand and gravel, under pressure from overriding glacial ice. The till exposed in the pit has a fine-grained texture, specifically a high silt content, and for that reason it is interpreted to consist primarily of reworked lake sediment, from a proglacial lake basin now occupied by Mille Lacs Lake, that was incorporated into the Superior-lobe ice as it readvanced during the Automba phase, depositing materials on the Mille Lacs moraine. A sample of the capping till was found to have a matrix texture of 15 percent sand, 71 percent silt, and 14 percent clay. Its 1 to 2 millimeter crystalline grains are 42 percent light, 38 percent dark, and 20 percent red. A sample of the dike till was found to have a matrix texture of 13 percent sand, 65 percent silt, and 22 percent clay. Its crystalline grains are 37 percent light, 40 percent dark, and 23 percent red.

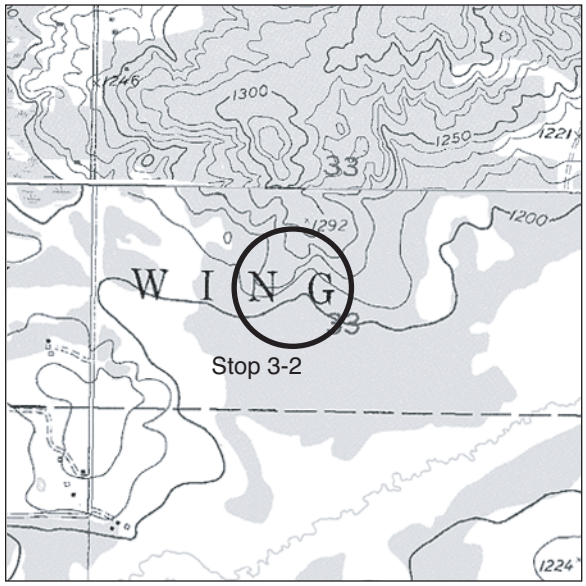
NEXT: North ~0.5 mile on township road to County Road 2. West ~17 miles to Crow Wing County Road 45. North 1.5 miles to Crow Wing County Road 131. West ~4 miles to township road. North ~1.5 miles to a pit road entrance on the east side of the road.

STOP 3-2

Derosier pit—Rainy-lobe deposits (Brainerd formation)

Location: T. 44 N., R. 31 W., sec. 33, SW

Fort Ripley quadrangle



Description: This 40- to 50-foot-high (12 to 15 meters) pit exposure is cut into the side of a large hill and has approximately a 300-foot-wide (90 meters) south face. The eastern end of this face exposes 20 feet (6 meters) of brown sandy till, over sand and gravel, over approximately another 20 feet (6 meters) of brown sandy till. The till and sand/gravel layers vary in thickness across the exposure. Six samples were collected, three from each till layer, and analyzed. The upper till layer samples averaged a matrix texture of 73 percent sand, 17 percent silt, and 10 percent clay. The lithology of the 1 to 2 millimeter grains had no carbonate, except for the lowest sample near the base of the unit, which had 2 percent carbonate (the upper two samples were leached) or shale, and the crystalline subtotal averaged 76 percent light, 16 percent dark, and 8 percent red. The lower till layer samples averaged a matrix texture of 75 percent sand, 15 percent silt, and 10 percent clay; the lithology of the 1 to 2 millimeter grains averaged 100 percent crystalline, 0 percent carbonate, and 0 percent shale (all samples were leached), and the crystalline subtotal averaged 80 percent light, 13 percent dark, and 7 percent red. These test results typify Brainerd formation till—a sandy texture, deep leaching, mid-teen percentage of crystalline darks, and crystalline reds at 7 to 8 percent. Till contacts with the sand and gravel layers are sharp.

This exposure is located on the side of a hill, which is a part of a topographic high interpreted to be a remnant of a recessional moraine of the Brainerd lobe. This site lies in the southern portion of the Brainerd drumlin field (Fig. 3.3). Both till layers are interpreted to be part of the Long Lake member of the Brainerd formation, deposits of the Brainerd lobe (Fig. 3.8). It appears that the deposition of the lower till was followed by northeastward ice retreat during which outwash sand and gravel were deposited, and finally the ice re-advanced, resulting in the deposition of the upper till. The till at this site is an excellent example of the pebbly, sandy, and deeply leached Long Lake member of the Brainerd formation. Its dark and red crystalline fraction average percentages of 15 percent and 7 percent, respectively, are typical for Brainerd formation till.

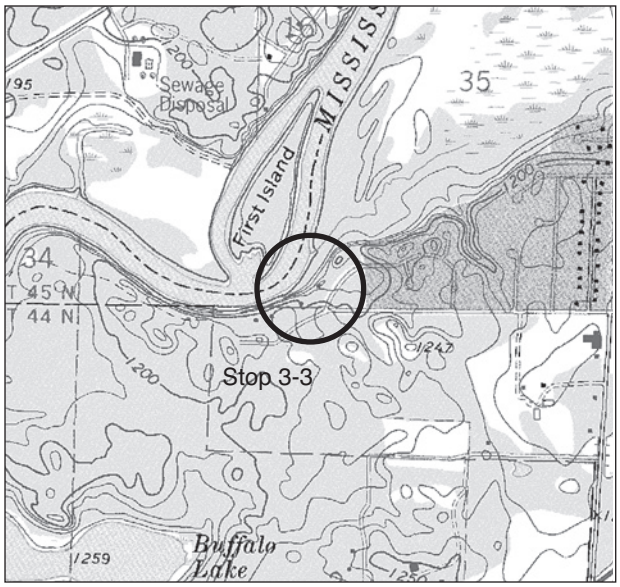
NEXT: North ~3 miles to State Highway 371. Northeast ~4 miles to Buffalo Hills Lane. West ~1 mile to dunes on the south side of the road.

STOP 3-3

South Brainerd dune field—eolian sand dune deposits—reworked glacial Lake Brainerd sands

Location: T. 44 N., R. 31 W., sec. 2, NW, NW, NE

Brainerd quadrangle



Description: The well-sorted, fine- to medium-grained sand at this site is interpreted as an eolian dune deposit based on local landforms. Sand dunes in this area were recognized and mapped by Hobbs (2001a, b) and Knaeble (2001). This uniform sand is the prevailing sediment found at the land surface over almost the entire glacial Lake Brainerd basin, an area of approximately 150 square miles (388 square kilometers) to the north and west of Brainerd (Fig 3.9; Knaeble and Meyer, 2004; Knaeble and others, 2004). Much of this sand may be lacustrine sediment in places, but it is difficult to differentiate between eolian (in this case reworked lake sand) and lacustrine sediments unless landforms, like dunes and blowouts, are present.

Glacial Lake Brainerd formed at the margin of the Brainerd lobe as the ice retreated to the northeast. Meltwater from this ice flowed into the lake and deposited layers of coarser-grained sand and gravel in the northeastern corner of the lake basin. In this area there is evidence (as seen at Stop 3-4) that subaqueous slope failure created turbidite flows, which disturbed other bedded lake sediments as they spread across the basin floor. Visible beach or strandline deposits were not detected, but channels, which probably flowed into the lake, end abruptly at an elevation of about 1,225 feet in the northwestern corner of the lake basin, between Pelican Lake and the city of Nisswa (Fig. 3.9). Dune formation most likely occurred after the lake drained but before vegetation stabilized the landscape, a period when wind was able to rework the lacustrine surface sand.

NEXT: East ~1 mile to Highway 371. North on Highway 371. In Brainerd take State Highway 210 north/northeast to Crosby. Look for signs for Veteran's Park in Crosby.

Lunch in Crosby at Veterans Park on Serpent Lake

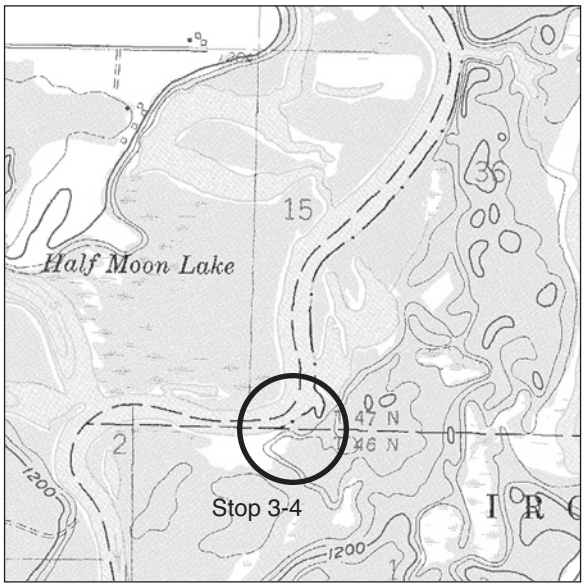
NEXT: West ~1 mile to Crow Wing County Road 30. North and west ~3 miles to Crow Wing County Road 34. South and west ~1.5 miles to Trommald. West on township road located at the northwest corner of Trommald for ~2.5 miles to an ATV access road on the south side of the township road. Hike to the river.

STOP 3-4

Mississippi River varve cut—glacial lake Brainerd varves and underflow fan

Location: T. 46 N., R. 30 W., sec. 1, NW, NE, NW

Trommald quadrangle



Description: At this site, cut-bank erosion by the Mississippi River has exposed a 25-foot-high (8 meters) sequence of glacial Lake Brainerd sediments. Approximately 50 to 75 varves composed of red clay and gray-brown silt extend from beneath the river water line to about halfway up the exposure. Varve thickness varies from less than an inch (1 centimeter) to as thick as 5 inches (13 centimeters). The red clay layers vary in thickness from a few millimeters to about 1 inch (2.5 centimeters). Near the top of the varved section, silt beds and fine-grained sand become the dominant component of the varves. The upper half of the outcrop is composed of (going up the outcrop) interbeds of silt with sand and gravel, gravelly sand, very fine- to fine-grained sand with ripples and cross beds, sand and gravel, and a silt cap. The top of the varved section and the 3 feet (1 meter) immediately above it contain a sediment sequence interpreted to be a subaqueous slope failure deposit. Here beds coarsen upwards containing sand layers, which also coarsen upwards and in places have an erosional lower contact. Deformation and faulting are also evident. One interpretation of this sequence is the following: glacial Lake Brainerd waters initially deposited fine-grained varved sediments in some areas of the lake basin. As meltwater streams dumped sand and gravel into the northeast corner of the lake, a prograding delta advanced out into the lake basin. Slope failure from this sediment fan sent coarser-grained turbidite flows out onto the distal varved sediments. Sediment layers coarsened upwards until sand and gravel were the chief sediments being deposited. When meltwater input diminished, the coarse-grained sediment source waned and the fine-grained sand that covers most of the glacial Lake Brainerd basin was deposited—near the top of this sequence.

Red-colored clay varves were found at a few other sites on the east side of the glacial Lake Brainerd basin, but throughout the rest of the basin the clay varves were gray or gray-brown in color (Fig. 3.9). A question arises as to the source of the red sediment. Is the red sediment a product of meltwater from the Brainerd lobe, which may have incorporated sufficient quantities of red iron-ore bedrock from the nearby Cuyuna Iron Range? Was glacial Lake Brainerd still in existence and receiving meltwater sediment from the Superior-lobe as ice advanced to its maximum margin just east of the lake basin?

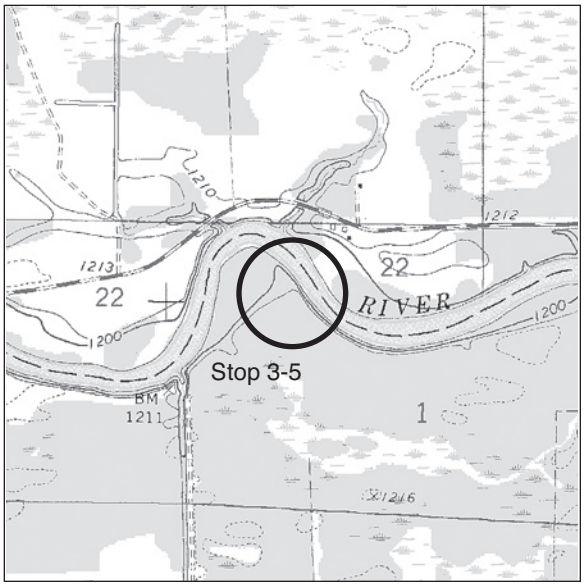
NEXT: East on township road for ~2.5 miles to County Road 34. East and north ~1.5 miles to County Road 30. North and east ~10.5 miles to Crow Wing County Road 32. North for 2 miles to township road. Continue north for ~1.5 miles on the township road and a path along the ditch. Walk to the river.

STOP 3-5

Rinta's Mississippi River cut—sand of glacial Lake Aitkin II /St. Louis-sublobe till/glacial Lake Aitkin I

Location: T. 47 N., R. 28 W., sec. 1, NW

Iron Hub quadrangle



Description: The cut exposes about 11 feet (3.4 meters) of yellowish brown (10YR 5/4), loam to silt loam till, over 3 feet (0.9 meter) of yellowish brown silt to clayey silt with disturbed bedding, over about 4 feet (1.2 meters) of unbedded, gray silty clay with brown mottles, over about 4 feet (1.2 meters) of ripple cross-bedded, fine-grained sand to the river level. The till averages about 7 percent carbonate clasts with only

a trace amount of red clasts, indicating that it is part of the Aitkin formation. The till is interpreted to be a deposit of the St. Louis sublobe. The sediment below the till is interpreted to be lake sediment deposited in glacial Lake Aitkin I (Figs. 3.9, 3.10). This exposure is similar to other observed exposures where the contact between St. Louis sublobe till and glacial Lake Aitkin I sediments is gradational, making it difficult to determine the base of the till or the top of the lacustrine sediment.

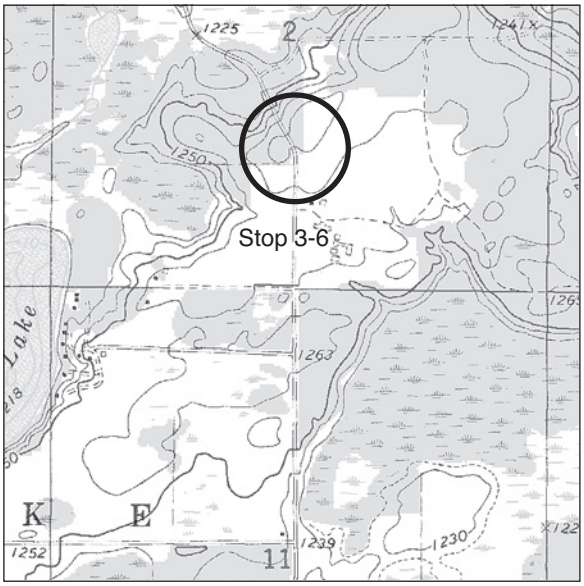
NEXT: South ~1.5 miles to County Road 32. East ~6.5 miles (County Road 32 changes to Aitkin County Road 15) to Aitkin County Road 22 in Aitkin. North and west ~9 miles (County Road 22 changes to Crow Wing County Road 11) to Crow Wing County Road 105. North ~2 miles to an exposure on the east side at the curve in the road.

STOP 3-6

Highway 105 road cut—St. Louis-sublobe till

Location: T. 136 N., R. 25 W., sec. 2, NE, NE, SW

Iron Hub quadrangle



Description: This road cut exposure located on higher ground on the north rim of the glacial Lake Aitkin basin is another example of till deposited by the St. Louis sublobe. The brown 10YR color and the presence of carbonate clasts differentiate this till from Superior-lobe till, which is at the surface just 3 miles (5 kilometers) northwest of this site, beyond the St. Louis-sublobe ice margin. Analysis of samples from this outcrop and from a 30-foot-deep (9 meters) auger boring 300 feet (91 meters) southeast of the outcrop show approximately 15 feet (4.6 meters) of silty clayey till with some pebbles, which overlies about 15 feet (4.6 meters) of glacial Lake Aitkin I silty clay sediment. Pebbles in the till diminish with depth as the till appears to grade into the underlying lake sediment. There is no distinct contact. The clay loam texture of the till and the fact that the coarse-grained sand fraction had no gray silicious shale, averaged 6 percent carbonate, and had very few red clasts, indicates that this is a St. Louis-sublobe deposit (Table 3.1). The bottom 6 inches (15 centimeters) of the boring penetrated fine- to coarse-grained sand, which may indicate a contact with the underlying Superior-lobe drift.

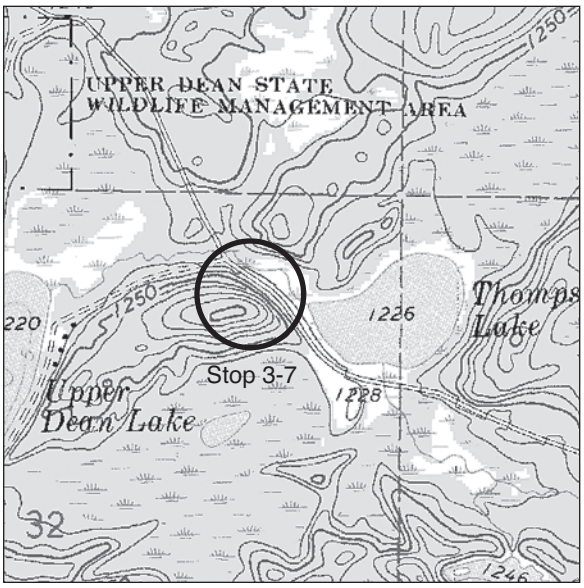
NEXT: North and west on County Road 105 for ~3.5 miles to an exposure on the south side of the road.

STOP 3-7

Thompson Lake road cut—Superior-lobe till and bedded sediments

Location: T. 137 N., R. 25 W., sec. 32, SW, NE, NE

Ross Lake quadrangle



Description: Exposed in this road cut is red silty Superior-lobe till truncated on the southeast side by layered sand and gravel over lacustrine fine-grained sand, silt, and clay. Two till samples at this site averaged 60 percent silt and the coarse-grained sand fraction of the crystalline grains averaged 34 percent dark and 17 percent red clasts. The till was slightly calcareous below 10 feet (3 meters).

The western extent of the St. Louis sublobe ice margin was crossed about halfway between this site and the last site (Fig. 3.10). There were 15 sites around the rim and in the glacial Lake Aitkin basin where yellow-brown to brown carbonate-bearing till with few red clasts overlies reddish-brown,

Table 3.1. Physical characteristics of glacial deposits in north-central Minnesota.

	NORTHWEST		NORTHEAST
PROVENANCE	WINNIPEG	RAINY	SUPERIOR
Lobe	St. Louis	Rainy	Superior
TILL TEXTURE	Loamy to clayey	Sandy	Silty to loamy
TILL COLOR			
Oxidized	Yellow-brown to brown	Brown	Red-brown
Unoxidized	Gray to dark gray	Gray to brown-gray	Gray to red-gray
PEBBLE TYPE			
Carbonate	Uncommon	Absent to rare	Absent to rare
Gray-green rock	Uncommon to common	Uncommon to common	Common to abundant
Red felsite	Absent to uncommon	Uncommon	Common
Gray shale	Absent to rare	Absent	Absent

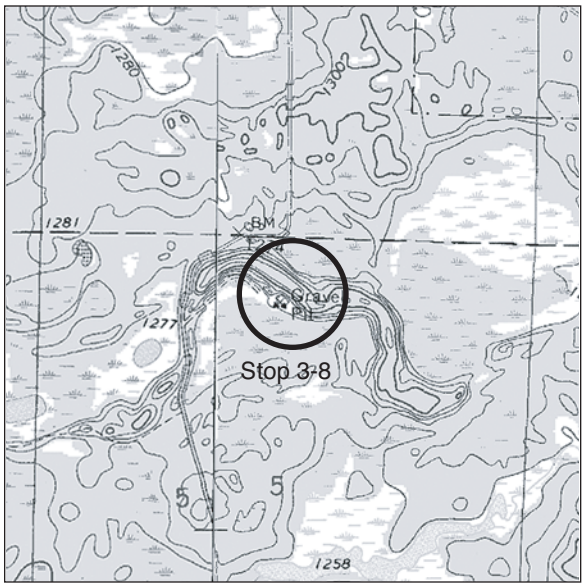
carbonate-lacking till with red clasts. Glacial Lake Aitkin I sediment is present between these tills at 7 of those sites. This stratigraphic evidence along with distinct color and lithologic differences (Stops 3-6 and 3-7 are examples) suggests that the two till deposits are from separate ice lobes, the St. Louis sublobe and the Superior lobe. Most interpretations over the last 40 years have considered these two tills to be St. Louis-sublobe deposits, the color difference being explained by incorporation of red glacial Lake Aitkin sediments.

NEXT: Continue northwest ~1.5 miles on County Road 105 to Crow Wing County Road 36. North ~1 mile to Crow Wing County Road 106. North ~3.5 miles to the pit entrance on the east side of the road.

STOP 3-8

County gravel pit—esker with Superior-lobe till capping sand and gravel

Location: T. 137 N., R. 25 W., sec. 5, NE, NE, NW Ross Lake quadrangle



Description: Today's final stop is at a county gravel pit operated in an esker that runs sinuously from east to west. The core of the esker is composed of approximately 30 feet (9 meters) of mostly horizontally bedded sand and sand/gravel. There is some cross bedding, a few thin silt lenses, and no

obvious collapse features. The clast lithology was estimated to be approximately 35 percent dark and 10 percent red. Overlying the esker core is red-brown, noncalcareous, silty till (near its base the till contains more sand) that varies in thickness from 3 to 10 feet (1 to 3 meters). The silty texture and a lithology of the crystalline clasts calculated at 27 percent dark and 13 percent red, suggests that this till is related to Superior-lobe ice. This deposit has been interpreted to be part of a 3-mile-long (5 kilometers) Superior-lobe esker, which terminates at the lobe's ice margin to the west. If this interpretation is correct, is the till on top of the esker from a Superior-lobe re-advance or could it be supraglacial till deposited on top of the sand and gravel as the overlying ice melted? Or could this be a Brainerd-lobe esker that was overridden by younger Superior-lobe ice?

NEXT: Continue north and east ~2.5 miles to Crow Wing County Road 1. East ~15 miles on County Road 1 (changes to Aitkin County Road 3) to Highway 169. North ~38 miles on Highway 169 to Grand Rapids to the motel.

END DAY 1

OVERNIGHT IN GRAND RAPIDS

DAY 2—DEPOSITS ALONG THE CENTRAL AND EASTERN PARTS OF THE ST. LOUIS SUBLOBE

Lisa Marlow, Phillip Larson, and Howard Mooers

INTRODUCTION TO DAY 2

Many deposits and landforms related to the advance of the St. Louis sublobe are found in and around the glacial Lakes Aitkin and Upham basin. Recent investigations (Marlow, 2004; Marlow and others, 2004) into the glacial geology of northeastern Minnesota have considerably revised the glacial chronology. This part of the field trip examines features (landforms and stratigraphy) of the glacial Lakes Aitkin and Upham basin that record a complex sequence of events related to the advance and retreat of the Laurentide Ice Sheet (Fig. 3.12). Deglaciation and formation of glacial Lakes Aitkin and Upham were followed by drainage of the lakes and eolian activity on exposed lacustrine sediments.

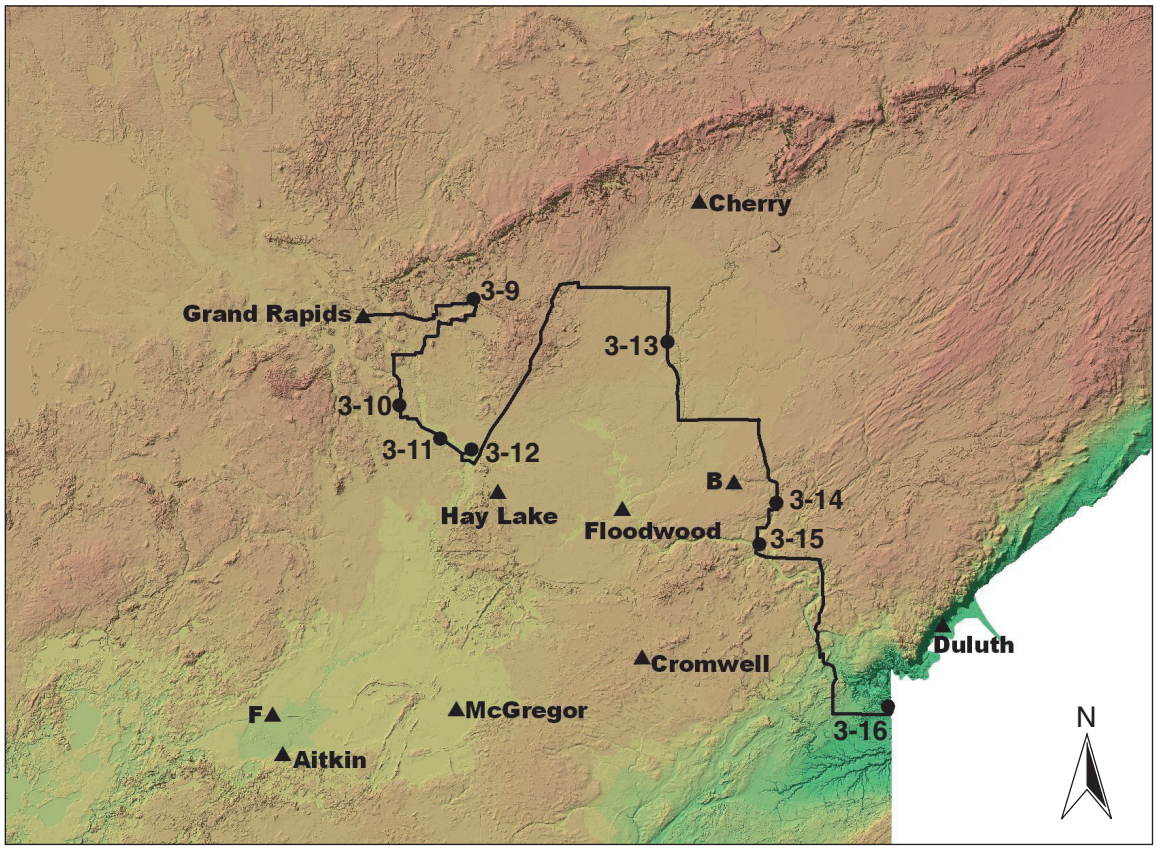
Glacial Lakes Aitkin and Upham occupied a basin bounded by the Giants Range to the north and moraines of the Superior and Rainy lobes and St. Louis sublobe to the south, east, and west. As the Rainy lobe retreated during Late Wisconsinan time, meltwater ponded in front of the ice forming proglacial Lakes Aitkin and Upham I. As the Rainy lobe continued to retreat northeastward across the basin to a margin roughly coincident with the Giants Range, the St. Louis sublobe advanced into the basin (Fig. 3.12). Stagnation and melting of the St. Louis sublobe resulted in formation of glacial Lakes Aitkin and Upham II (Hobbs, 1983). Continued retreat of the Rainy lobe northward from the Giants Range resulted in formation of glacial Lakes Norwood and Koochiching, both of which drained southward into glacial Lakes Aitkin and Upham II. The evolution of the lakes by a combination of downcutting of the outlets and isostatic rebound is recorded by a series of beaches, wave-cut scarps, and multiple outlets.

Dune formation in the glacial Lakes Aitkin and Upham II basin was strongly controlled by sediment availability. Dune clusters show a strong correlation with the presence of underlying fine-grained nearshore lacustrine sand. Dune formation was likely episodic, coinciding with periods of rapid lake-level lowering and exposure of nearshore sands. Cessation of eolian activity resulted from more gradual stabilization of dunefields. Peaks in eolian activity are indicated at 9.8, 9.3 and 7.4 kyr B.P. (all dates given in C^{14} years B.P.) by the magnetic susceptibility record of Hay Lake, a small lake near the glacial Lake Upham II shoreline (Figs. 3.13, 3.14). These peaks in eolian activity may coincide with episodes of drainage of glacial Lakes Upham II and Aitkin II.

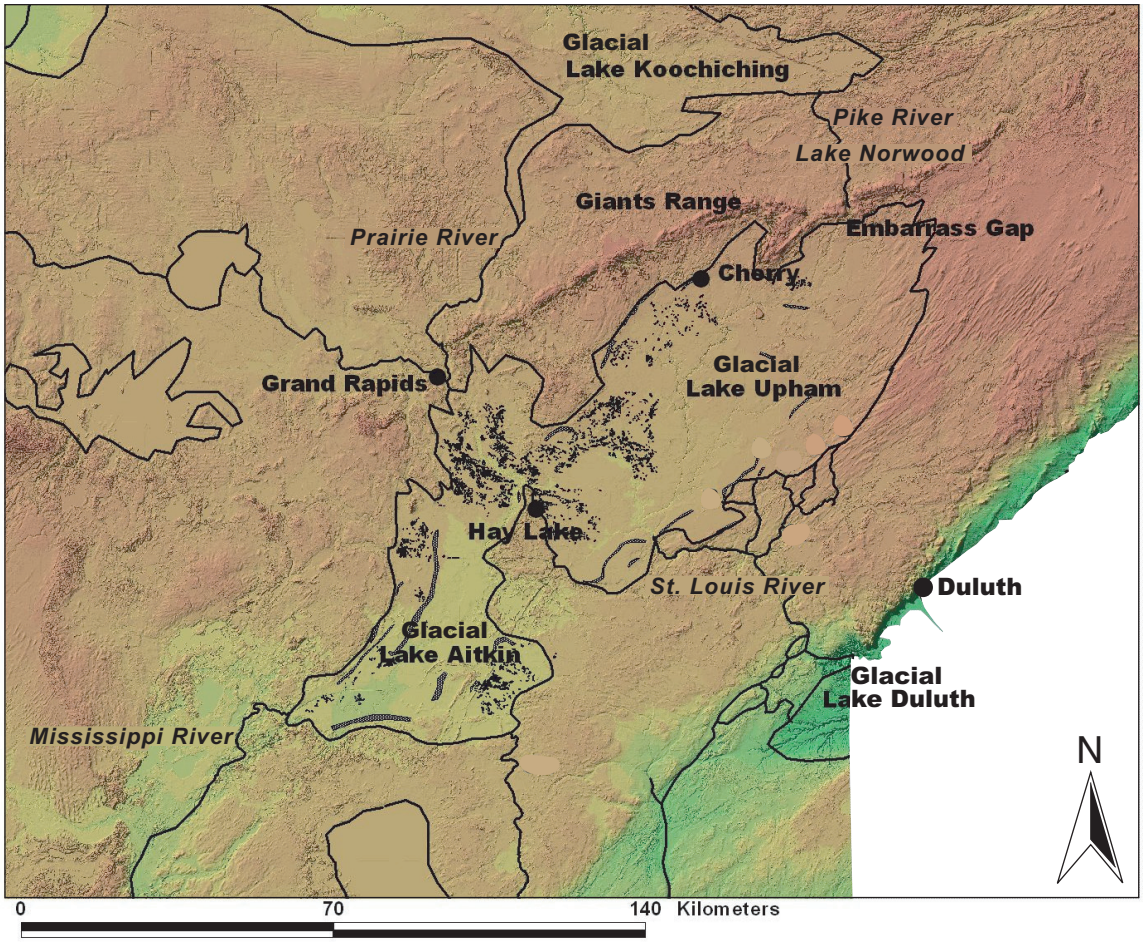
The dunefields in the glacial Lakes Aitkin and Upham II basin display similarity to other Quaternary eolian deposits throughout Minnesota and the Midwest (Figs. 3.15, 3.16). Despite over a century of research on Minnesota's glacial landscape, these paleodunefields have received relatively little attention. Although mentioned by Upham (1896), Winchell (1896), Elftman (1898), Hall and Sardeson (1898), Leverett and Sardeson (1917, 1919), and Leverett (1932), the first comprehensive work on Minnesota's eolian landscape was that of Cooper (1935, 1938).

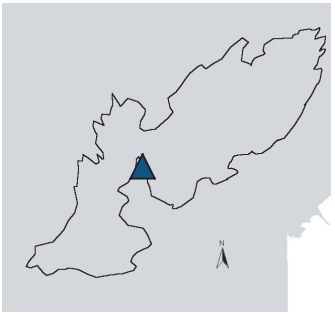
More recently, Grigal and others (1976) examined a dunefield exposed along the southeast shore of Lake Winnibigoshish (Fig. 3.16). They described several buried soil horizons, which yielded radiocarbon dates in the range of 7,910 to at least 5,040 years B.P., suggesting episodic dune formation during the mid-Holocene. The formation of these dunes was related to lake-level instability and episodic

A.


Cherry
Grand Rapids
3-9
3-13
3-10
3-11
3-12
B
3-14
Hay Lake
Floodwood
3-15
Duluth
Cromwell
F
McGregor
3-16
Aitkin
N

B.


Glacial
Lake Koochiching
Pike River
Lake Norwood
Giants Range
Embarrass Gap
Prairie River
Cherry
Glacial
Lake Upham
Grand Rapids
Hay Lake
St. Louis River
Duluth
Glacial
Lake Aitkin
Glacial
Lake Duluth
Mississippi River
N
0
70
140 Kilometers



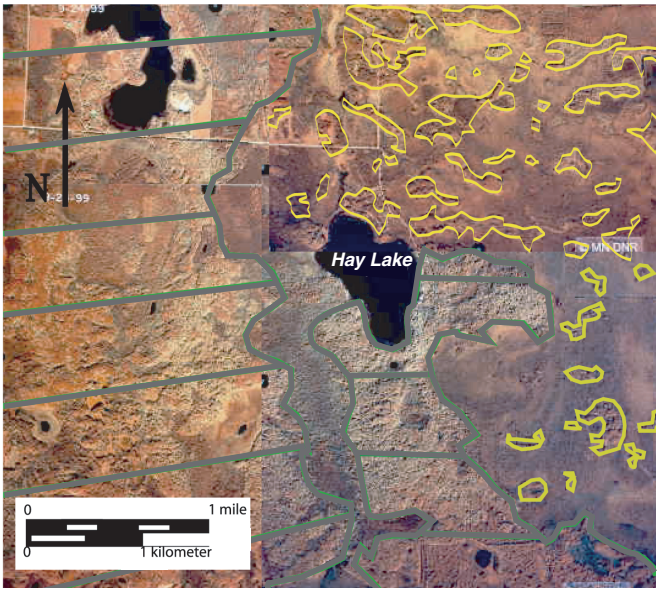


Figure 3.13. Location of Hay Lake. The lined region indicates the Lake Upham upland area; white polygons are dunes (Vanduse quadrangle).

exposure of littoral areas to wind erosion during the mid-Holocene (Larson and Mooers, 2003).

The mid-Holocene climate in the upper Midwest was considerably warmer and drier than the modern climate (Webb and others, 1983; Bartlein and others, 1984; Dean and others, 1984, 1996; COHMAP members, 1988). The warmer, drier conditions resulted in an eastward shift of the prairie ecotone (Fig. 3.16). Lower ground-water levels were accompanied by lake-level lowering, particularly in ground-water-dominant closed-basin lakes and ponds, and increased fire frequency and vegetational stress locally resulted in landscape destabilization.

Grigal and others (1976) speculated that the dunes of the Anoka Sand Plain may be mid-Holocene in age, an idea Keen and Shane (1990) explored in detail in their paleoenvironmental investigation of Ann Lake (Fig. 3.16). They concluded that there was a relatively continuous record of eolian activity throughout the mid-Holocene, a reversal of the ideas of earlier workers (Hall and Sardeson, 1898; Cooper, 1935). Keen and Shane's (1990) whole-core magnetic susceptibility record from Ann Lake records increased silt and fine-grained sand input to the lake from about 8.0 until 4.5 kyr B.P., with discrete peaks in clastic input at circa (ca.) 7.4, 5.8, and 4.9 kyr B.P. However, it is difficult to distinguish whether this sediment was exported to the lake directly from active dunes on the shore, eroded from the shoreline due to lake level instability, or deposited by suspended airborne sediment. Despite these ambiguities, they concluded that the increased clastic input to Ann Lake was due to landscape destabilization and widespread dune activity on the Anoka Sand Plain.

The distinction between lake level lowering and landscape destabilization as triggers of eolian activity is subtle but important. Although the prairie ecotone shifted eastward during the mid-Holocene, this in itself did not create an environment conducive to widespread landscape destabilization. At present, in the central Dakotas there are numerous locations with ample sediment availability, but no widespread dune formation because of the stabilizing vegetation of the prairie environment. Landscape destabilization and dune formation in the existing prairie environment occurs by excessive drying such as in the Nebraska Sand Hills (Winspear and Pye, 1996). Viewed in this context, the Keen and Shane (1990) and Dean (1997)

Figure 3.12. A. Road Log with numbered field trip stops, geographic markers, and discussion sites: F: Farnham site, B: Baker Spider Creek site, and Hay Lake. Image from 30-meter Digital Elevation Model of Minnesota.

B. The glacial Lakes Aitkin and Upham basin. Beaches are noted by shaded polygons; dunes appear as black dots in the basin; boundaries and channels are noted as black lines. Field trip stops will illustrate examples of the following: beach, dune, underflow fan, rhythmites, alluvial fan, glacial Lake Upham sediments in Rainy-lobe outwash, and a wave-washed esker. Map is modified from Hobbs (1983).

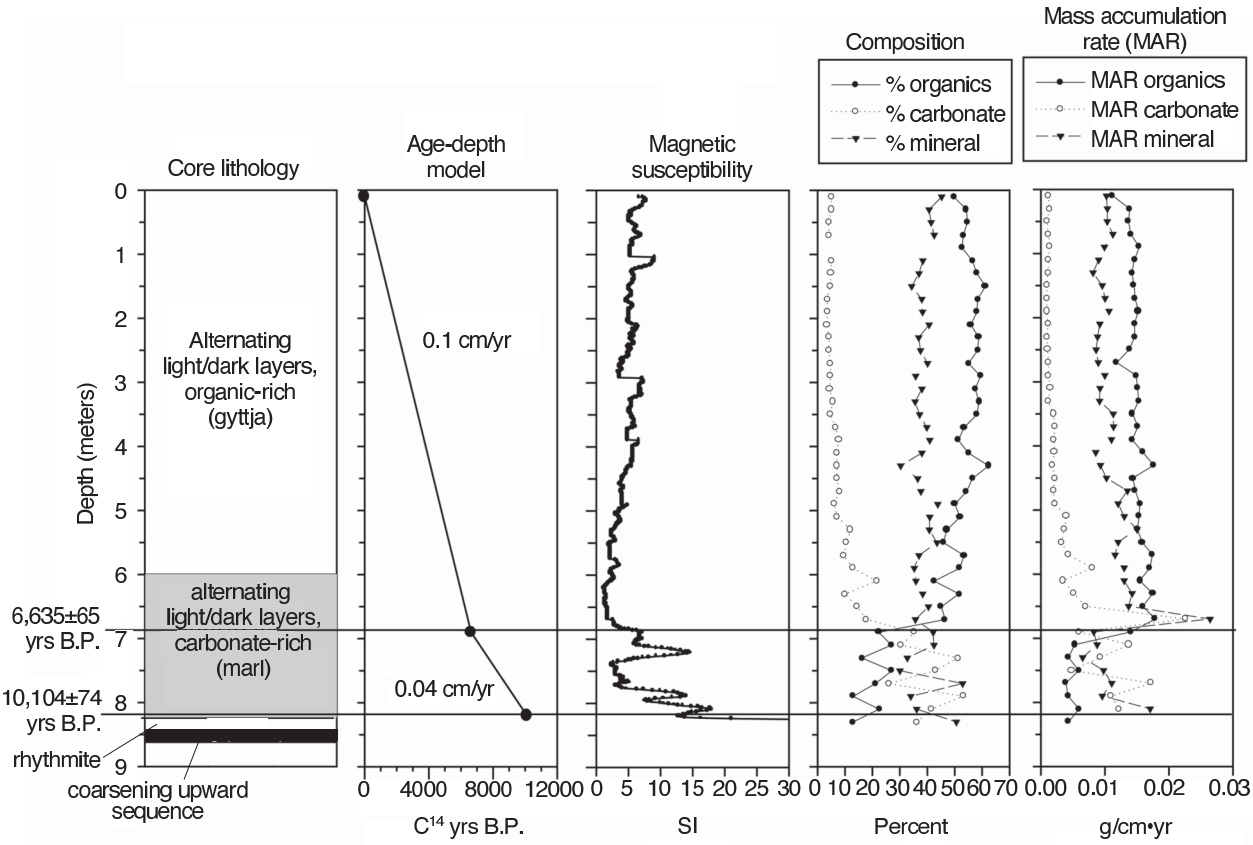


Figure 3.14. Hay Lake sediment core (8.57 meters).



Figure 3.15. North American dune (black) and loess (gray) distribution (Brady and Weil, 2000).

magnetic susceptibility records may merely be a record of dryer, dustier conditions during the mid-Holocene rather than pervasive local eolian activity; Grigal and others (1976) obtained dates from paleosols buried by eolian events.

Today the glacial Lakes Aitkin and Upham basin is a patchwork of small "islands" occupied by upland vegetation formed on stabilized dunes. The "islands" are interrupted by vast areas of peatland developed on the poorly drained, low-relief lakebed composed of fine-grained lacustrine sediment. Locally, scattered topographic highs are underlain by older glacial sediments predating formation of the lakes.

STOP 3-9

Goodland esker

Location: T. 55 N., R. 23 W., sec. 9, SE, NE Calumet quadrangle

Description: This stop is a gravel pit exposure in the Goodland esker, one of the most prominent glacial landforms in Itasca County. The location of the esker and details of its formation place important constraints on the timing of the St. Louis-sublobe advance and deposition of the Alborn drift (Fig. 3.17).

The morphology of the esker system is best described by the model of Shreve (1985). The proximal (with respect to the Giants Range) segment is an ~600-meter-wide, 10-kilometer-long channel incised into bedrock and older drift (tunnel valley), occupied by a

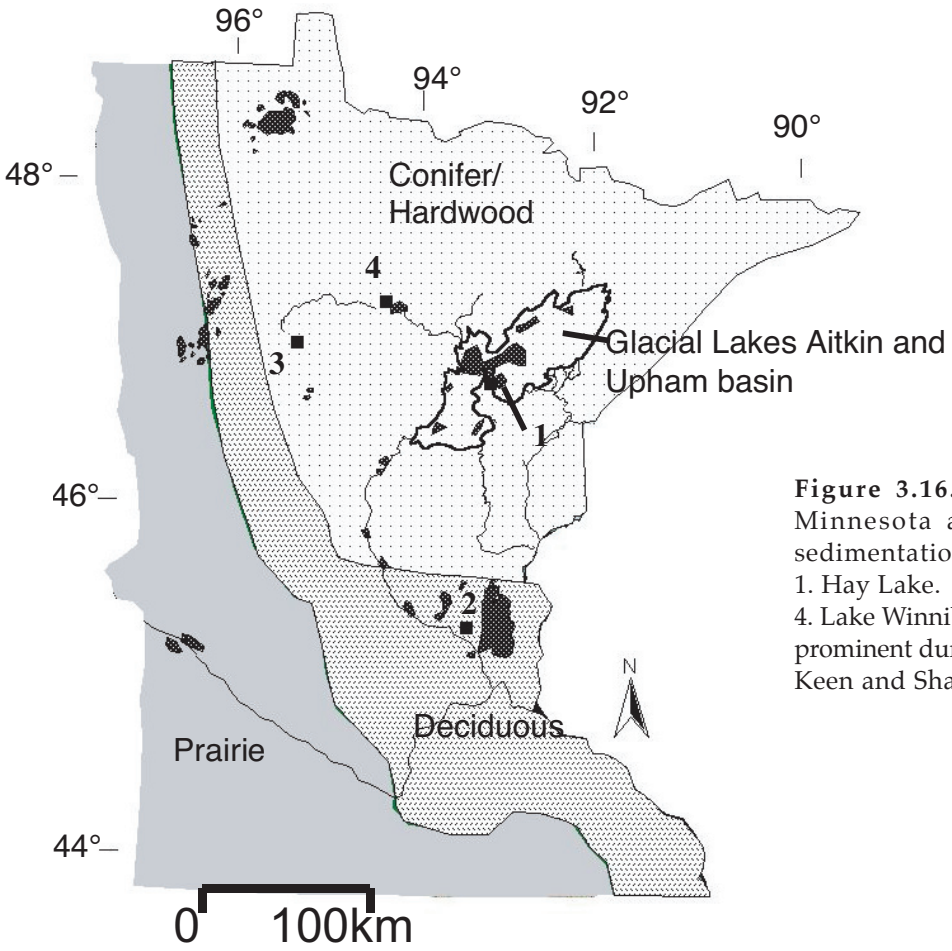


Figure 3.16. Location of sites in Minnesota associated with eolian sedimentation referenced in the text. 1. Hay Lake. 2. Lake Ann. 3. Elk Lake. 4. Lake Winnibigoshish. Filled areas are prominent dune colonies (modified from Keen and Shane, 1990).

multi-crested esker. The medial segment consists of a single, broad-crested esker segment deposited in a 10-kilometer-long channel incised into the overlying ice. The distal component of the esker system is a supraglacial fan complex.

The base of the subglacial channel drops from 428 meters at the crest of the Giants Range to less than 400 meters at a point 6 kilometers downstream, then rises to 409 meters over the next 4 kilometers. Downstream of this point, the elevation of the top of the broad-crested esker rises 19 meters over 10 kilometers, from 434 to 455 meters at the apex of the distal fan.

The transition from a multi-crested to broad-crested esker is apparently triggered by an increase in the adverse slope, up which the subglacial channel flowed. Multi-crested eskers form in areas of moderate upgradient flow as the channel tends to migrate laterally, rather than upward, in the ice, and broad-crested eskers form in areas of steeper upgradient flow due to freezing of the tunnel walls, a process favoring low, wide tunnel geometries (Shreve, 1985).

Lithologies in the multi-crested segment esker are dominantly Archean granitics and greenstones transported from north of the Giants Range. The broad-crested segment has a markedly different lithologic assemblage than the multi-crested segment. In addition to Archean lithologies, an abundance of Paleoproterozoic Animikie basin lithologies are present including quartzite, sulfidic mudstones, iron-formation, and pisolitic lateritic iron-formation. The abundance of locally-derived Animikean lithologies indicates material in the broad-crested segment was deposited as incision of the N-channel was taking place upstream. The broad-crested segment is therefore older than the multi-crested segment. The wide range of mean sizes of the various Animikean lithologies attests to the strong control of primary rock characteristics on particle size. Note in particular the abundance of Pokegama Quartzite as large (~1 meter) boulders and its paucity in smaller size fractions.

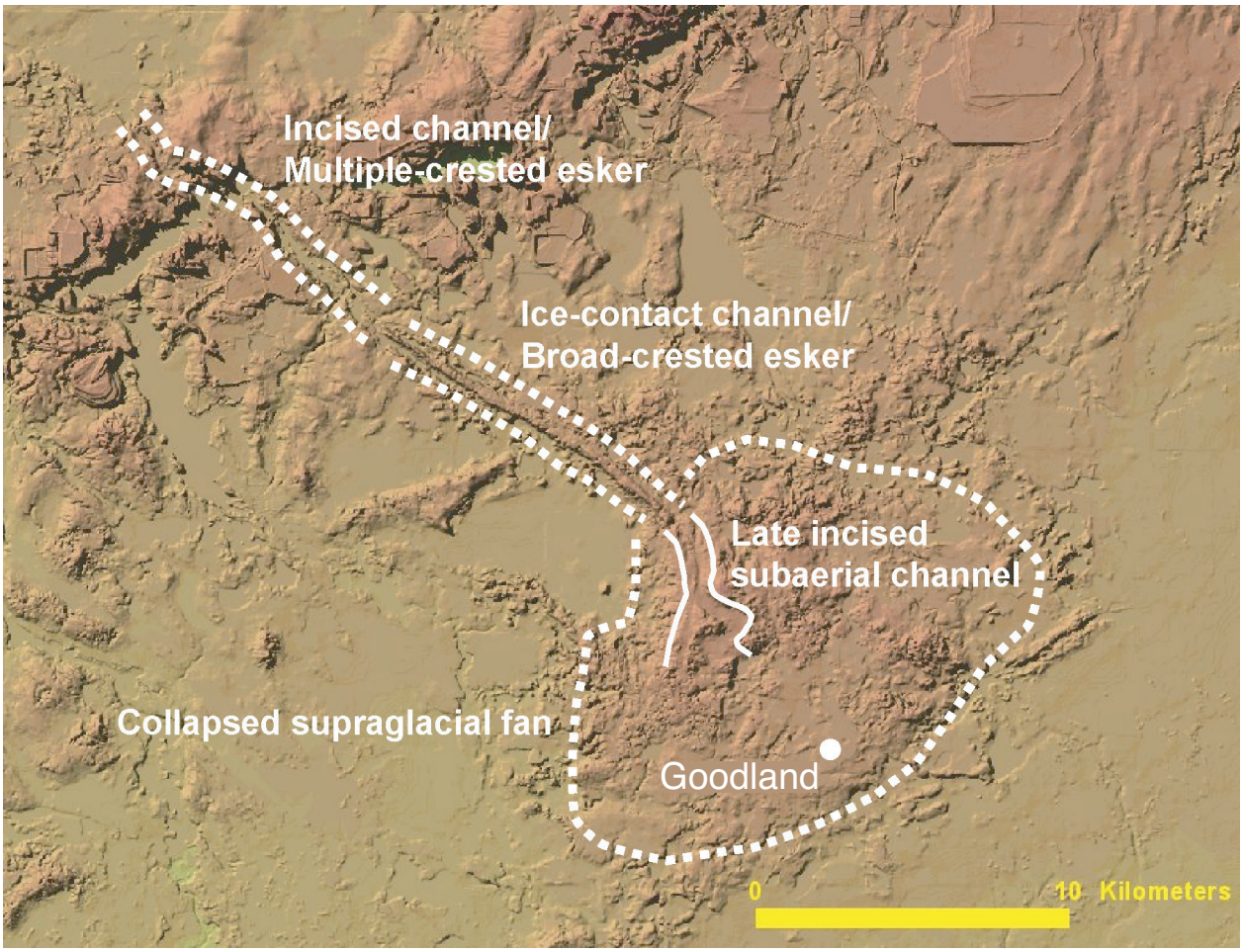


Figure 3.17. Main features of the Goodland esker system (image from 30-meter DEM of Minnesota).

Deposition of the distal fan occurred throughout the period the esker system was active. The initial phase was characterized by deposition of a supraglacial fan on stagnant ice. Fan deposits are identifiable up to 10 kilometers from the fan apex, and cover in excess of 100 square kilometers. The maximum elevation on the fan complex is about 477 meters, indicating at least 20 meters of ice was present above the outlet of the englacial channel at the fan apex. During later stages of the esker system, underlying stagnant ice melted, collapsing the earliest deposited fan sediment. The final phase of fan formation was characterized by incision of the collapsed fan head down to about 450 meters.

The subglacial drainage system that deposited the esker was probably short-lived. It could not have formed before the St. Louis sublobe occupied the area around Grand Rapids, blocking the natural southwesterly flow of meltwater from the watershed of the present-day Prairie River. Similarly, it could not have persisted after St. Louis-sublobe ice wasted away and the Prairie River began flowing into glacial Lake Aitkin II. Consequently, it may have been active a few hundred years at most. Despite its short duration, the large size of the esker system—it is one of the largest in Minnesota—attests to an enormous discharge of meltwater responsible for its formation. Meltwater was gathered from an interlobate zone that existed to the north of the Giants Range between the St. Louis sublobe and the Rainy lobe. This interlobate region may have drained an area well in excess of 1,000 square kilometers of the ice sheet.

The elevation difference between the fan apex (477 meters) and the surrounding (subglacial) landscape (400 meters) indicate that a continuous cover of Rainy-lobe ice at least 75 meters thick was present south of the Giants Range at the time of esker formation and advance of the St. Louis sublobe. This argues strongly against the postulated ice-free zone between the northern margin of the St. Louis sublobe

and the southern margin of the Rainy lobe. Previous workers have postulated a relatively late advance for the St. Louis sublobe (ca. 11.7 kyr B.P.), correlating it with the Vermilion phase (ca. 12.0 kyr B.P.), or an even later phase, of the Rainy lobe (Wright, 1972; Hobbs, 1983). However, the presence of active Rainy-lobe ice just north of the Giants Range at the time of the advance of the St. Louis sublobe and esker formation provides further evidence that the St. Louis sublobe advanced at a significantly earlier date.

STOP 3-10

Dune in glacial Lake Aitkin basin

Location: T. 53 N., R. 24 W., sec. 7, SW, NW

Split Hand Lake quadrangle

Description: Exposed in the sand pit is a small (less than 2 meter) dune composed of characteristic 4ϕ sand (Fig. 3.18). The 4ϕ mean grain size of the dunes sampled in these areas in the basin match that of sediments from the underflow fan sediments that will be seen at Stop 3-11. These sediments have been mapped in the Aitkin and Itasca County soil surveys (Nyberg, 1987, 1999) as the Zimmerman, Cowhorn, and Wawina soil series, all composed of fine- to very fine-grained sands of lacustrine origin. Most of the sand exposed in the pit is massive and structureless. However, relict cross bedding can sometimes be observed in small patches near the base of the sequence. Primary sedimentary structures in the upper portion of the exposure have been obliterated by bio and crioturbation.

Clusters of eolian dunes are widely distributed throughout the glacial Lakes Aitkin and Upham basin (Fig. 3.19). They formed as glacial Lakes Aitkin and Upham II incrementally drained, exposing areas of littoral sediment to wind erosion. Dune heights range from 1 to 5 meters and are composed of characteristic fine- to very fine-grained sand (Marlow, 2004). They occur as parabolic or longitudinal dunes, but are commonly distorted in morphology as a result of forming in a variable hydrologic environment (Bagnold, 1941). There are a large number of longitudinal dunes and elongate clusters of dunes oriented in a northwest–southeast direction, suggesting they formed under prevailing northwest winds (Fig. 3.20). Bagnold's (1941, p. 281) description of dunes indicated that they "...tend to occur in belts or chains, whose direction coincides with that of the resultant long-period sand vector Q," with Q being sand flow because of the sum of the strong and gentle wind directions, and the width of the belt at right angles to Q.



Figure 3.18. Dune exposure at Stop 3-10.

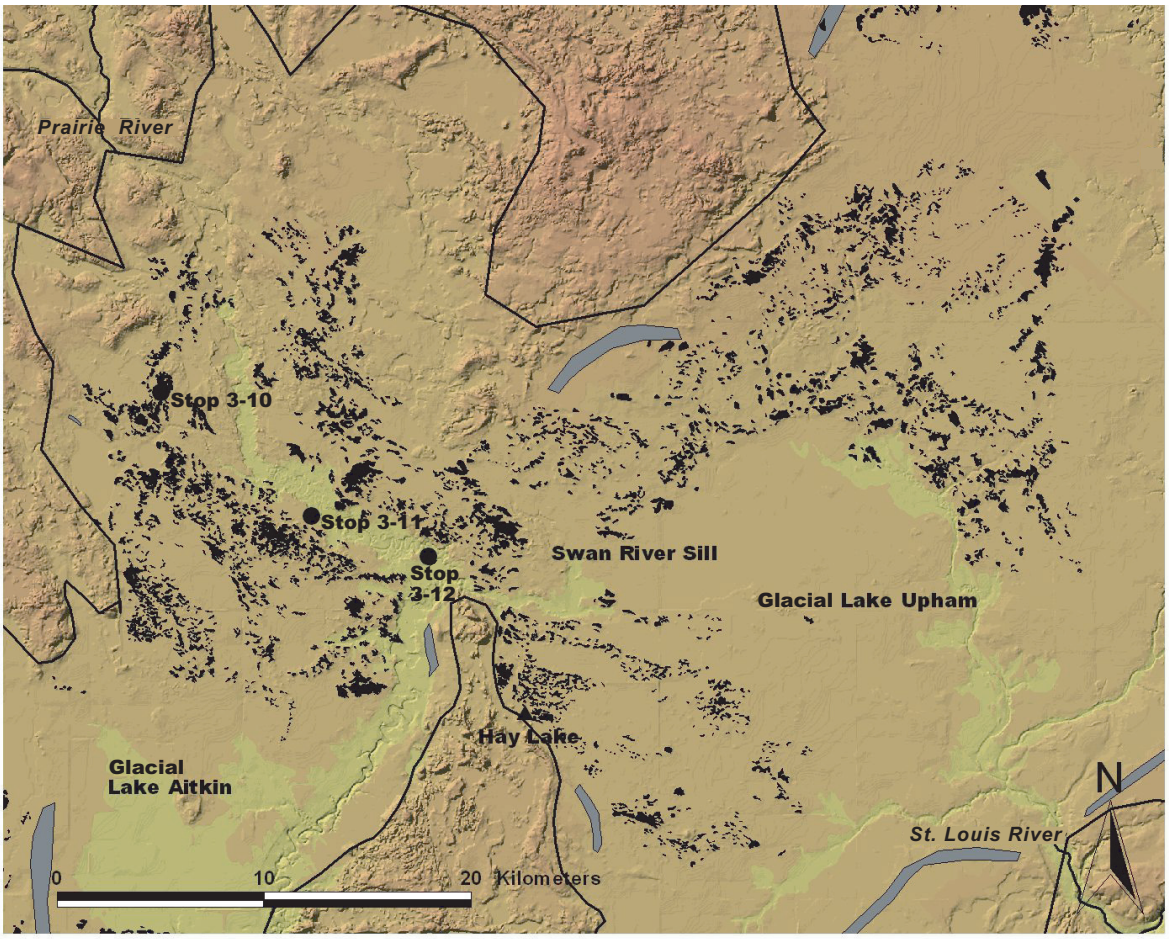


Figure 3.19. Trend in dune clusters (black) indicating a northwest–southeast wind direction. Elongate shaded polygons are beaches.

STOP 3-11

Prairie River underflow fan in glacial Lake Aitkin II basin

Location: T. 53 N., R. 24 W., sec. 35, NE, NE Jacobson quadrangle

Description: Glacial Lakes Aitkin and Upham II had two successive major meltwater inlets, the Embarrass Gap and the Prairie River. Inflow of the Prairie River into glacial Lake Aitkin II resulted in deposition of a large underflow fan extending 50 kilometers from its apex (Hobbs, 1983). The fan was later incised by the Mississippi River.

The cutbank at this stop exposes ~9 meters of sandy underflow fan sediments overlying lacustrine clay (Fig. 3.21). Visible within the underflow sediments are several Bouma sequences.

Outflow from glacial Lake Koochiching into glacial Lake Upham II through the Embarrass Gap ceased when a lower outlet opened north of Grand Rapids because of collapse of stagnant ice-cored terrain. Meltwater then flowed down the Prairie River into glacial Lake Aitkin II, lowering the level of glacial Lake Koochiching from 427 to 411 meters (1,400 to 1,350 feet; Hobbs, 1983). The Prairie River entered glacial Lake Aitkin II at an elevation of 396 meters (1,300 feet; Fig. 3.12). Although it is not known exactly when the Embarrass Gap was abandoned, it must have occurred prior to a 10.2 kyr B.P. transition from predominantly clastic to organic lake sedimentation in Sabin Lake located in the Embarrass Gap (Björk, 1988; Lehr and Hobbs, 1992). Fenton (1983) suggested that the Prairie River outlet was initiated between 12.3 and 10.8 kyr B.P. based on the chronology of the Lake Agassiz basin to the west. Clayton (1983) suggested 11.5 kyr B.P. for the inception of the Prairie River outlet.

The Prairie River inlet was abandoned once glacial Lake Koochiching began to flow into glacial Lake Climax. However, glacial Lakes Aitkin and

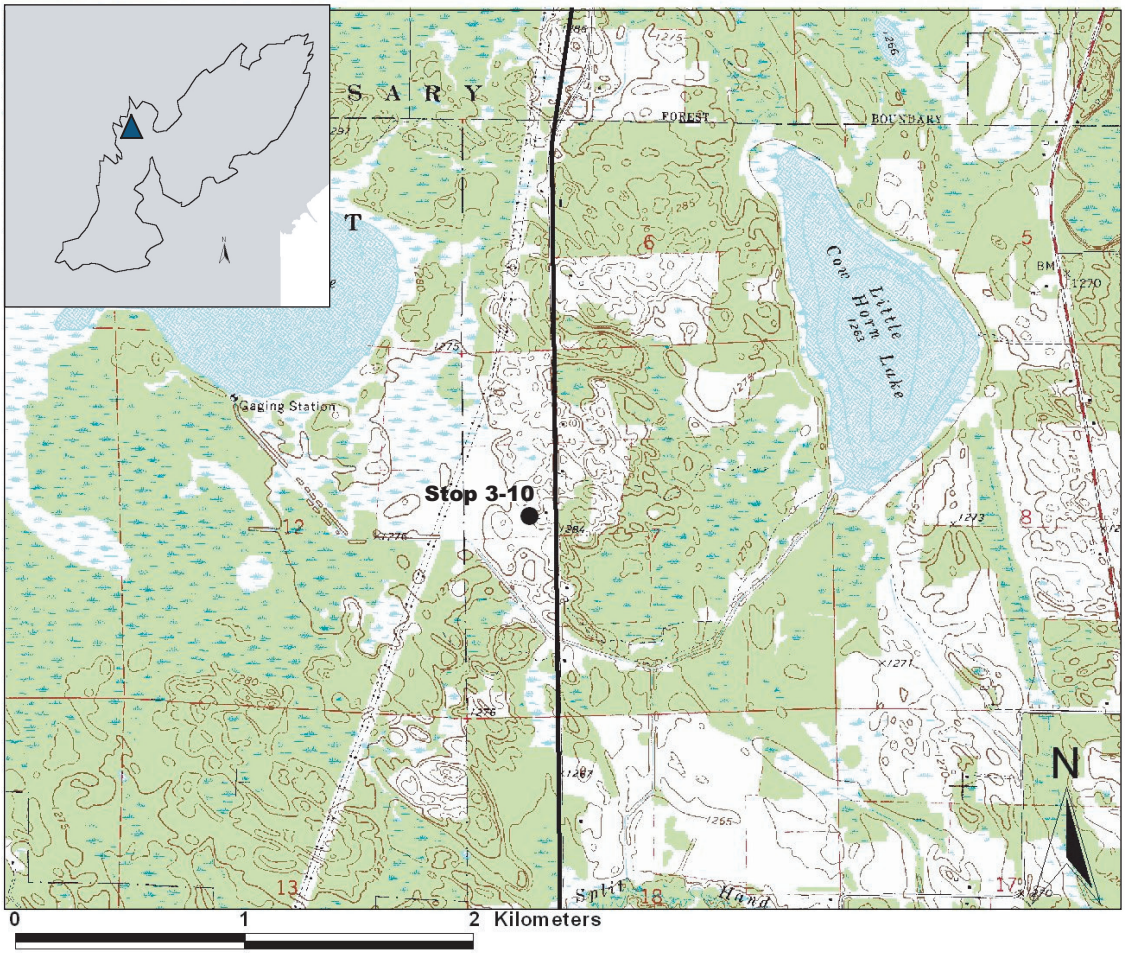


Figure 3.20. Topographic map of distorted dunes in the northern end of the glacial Lake Aitkin basin (see inset for location) showing a combination of crescentic and longitudinal dunes.

Upham II persisted after abandonment on the last meltwater inlet. Glacial Lake Upham finally drained as the St. Louis River outlet was incised to its modern level. Hobbs (1983) indicated that water could have still been flowing into glacial Lake Aitkin after drainage of glacial Lake Upham by way of the Mississippi River, which has a prominent terrace at 389 meters (1,275 feet).

Farnham and others (1964) obtained a date of 11,635 yr B.P. on a paleosol in the southwestern part of the glacial Lake Aitkin II basin. This paleosol is overlain by thin marl and 3 feet of clay, indicating glacial Lake Aitkin was in existence well after this time. A radiocarbon date of 10,000 yr B.P. obtained from marl from the Ball Bluff quadrangle in the northeastern part of the glacial Lake Aitkin basin (Hobbs, 1983) suggests the lake persisted at least until this time.

There are two alternative explanations for how the southwestern portion of the glacial Lake Aitkin II basin was successively drained and re-inundated.

The close similarity between the presumed initiation of the Prairie River inlet at 11.5 kyr B.P. (Clayton, 1983) and the re-inundation of the southwestern part of the glacial Lake Aitkin II basin at 11,635 yr B.P. suggests a causal relationship. Glacial Lake Aitkin II may have assumed a lower, stable level by ca. 11.6 kyr B.P. The relatively rapid lowering of glacial Lake Koochiching would have sent a surge of meltwater into the lake, resulting in a rise in lake level even if only temporarily. The lacustrine sediment overlying the paleosol dated by Farnham and others (1964) was thus derived from the Prairie River inlet.

The alternate explanation is that much of the glacial Lake Aitkin basin was drained through

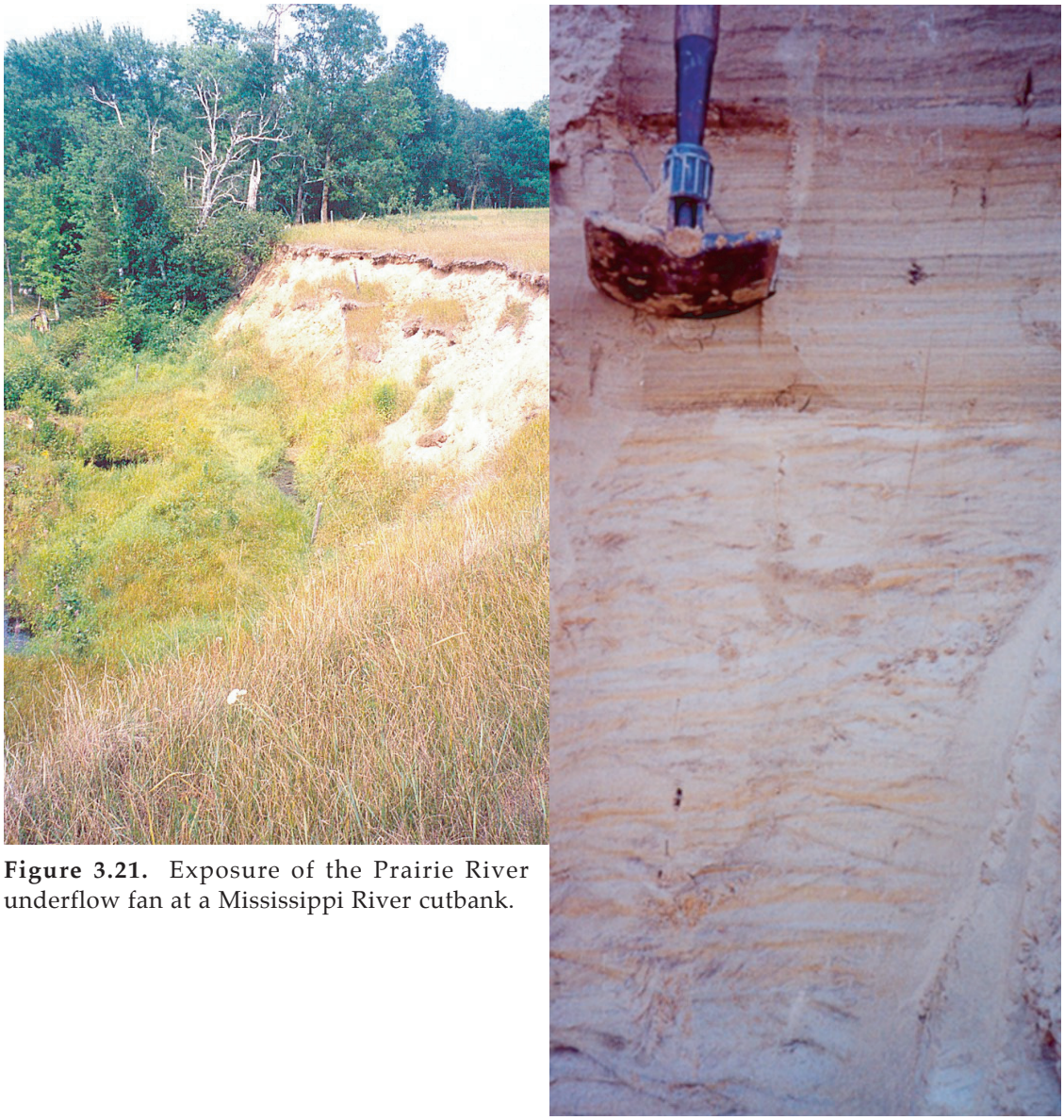


Figure 3.21. Exposure of the Prairie River underflow fan at a Mississippi River cutbank.

glacial Lake Upham II by downcutting of the St. Louis River outlet. The Swan River Sill (Fig. 3.20) controlled the water level in the remaining lake. As isostatic rebound of the basin progressed, the sill on the northeastern side of the basin rose at a higher rate than the southwestern portion of the basin. Lake level transgression resulted in re-inundation of the lake bottom exposed by drainage of glacial Lake Upham.

STOP 3-12

Rhythmites overlain by underflow sands in the glacial Lake Aitkin II basin

Location: T. 52 N., R. 23 W., sec. 4, SW, NW

Jacobson quadrangle

Description: This is an interesting site that may document a dramatic inception of the Prairie River inlet into a basin that was previously depositing clays. Clay that contains rhythmites is overlain by underflow sands followed by a layer of coarser-grained cross-bedded sands that are down-dipping basinward and overlain by more underflow sands (Fig. 3.22).

STOP 3-13

Rainy-lobe esker in glacial Lake Upham basin overlain by wave-washed St. Louis-sublobe till

Location: T. 54 N., R. 20 W., sec. 1, NE, NE

Toivola quadrangle



Figure 3.22. Underflow fan sedimentation interrupted by cross-bedded coarser-grained sands.

Description: There are many landforms within the glacial Lakes Aitkin and Upham II basin that predate development of the lakes. This exposure is an example of an esker deposited during retreat of the Rainy lobe that was later modified by waves (Fig. 3.23). This esker and others like it became wave-washed "islands" once glacial Lakes Aitkin and Upham I and II formed.

Exposed at the base of the sequence are coarse-grained gravels. The gravels contain abundant northeast-provenance material and locally derived mudstone and shale from the Paleoproterozoic Virginia Formation and perhaps younger Cretaceous strata. They were deposited by a beaded esker system during retreat of the Rainy lobe. Overlying the gravels is a fine-grained till deposited by the St. Louis sublobe. On top of the till is a sequence of nearshore sands and gravels. These presumably eroded from that portion of the esker rising above the level of glacial Lake Upham II; the strandline formed at about 397 meters (1,200 feet) elevation. The uppermost portion of the sequence is a blanket of eolian sediment exported from the surrounding lake plain to the upland after final drainage of glacial Lake Upham.

STOP 3-14

Johnson pit—Rainy-lobe outwash overlain by glacial Lake Upham I sediments and St. Louis-sublobe till

Location: T. 51 N., R. 18 W., sec. 1, NE, NE

Alborn quadrangle

Description: The glacial Lakes Aitkin and Upham basin was occupied by glacial lakes on two separate occasions during the Late Wisconsin glaciation. The retreat of the Rainy and Superior lobes from their maximum positions formed a series of ice-cored recessional moraines bordering the lake basins to the south and west, leading to ponding of water and formation of glacial Lakes Aitkin and Upham I (Wright, 1972; Hobbs, 1983; Lehr and Hobbs, 1992). The initial retreat of the Rainy lobe from the Mille Lacs and Outing moraines (Mooers, 1988) led to the formation of glacial Lake Aitkin I (Fig. 3.24). Continued retreat of the Rainy lobe from the Sandy Lake moraine led to the formation of glacial Lake

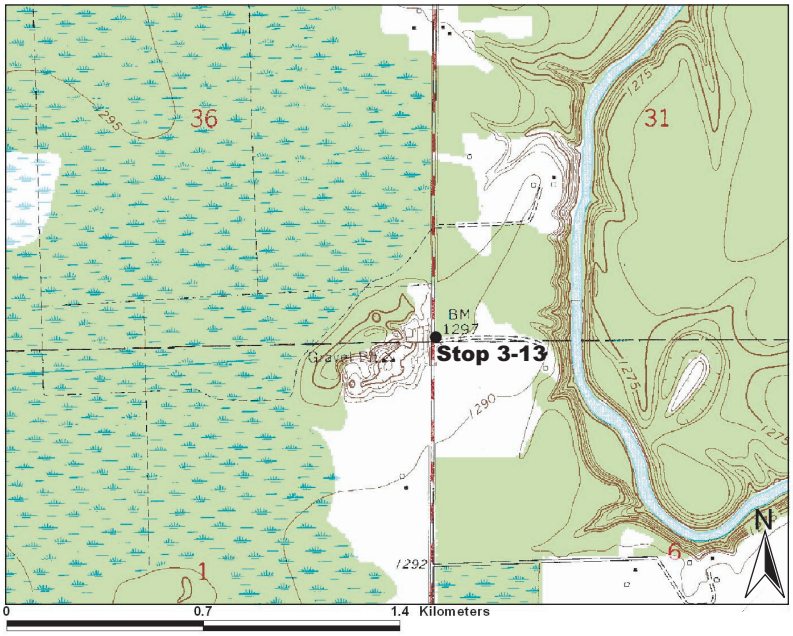


Figure 3.23. Topographic map of a wave-washed esker island in glacial Lake Upham II.

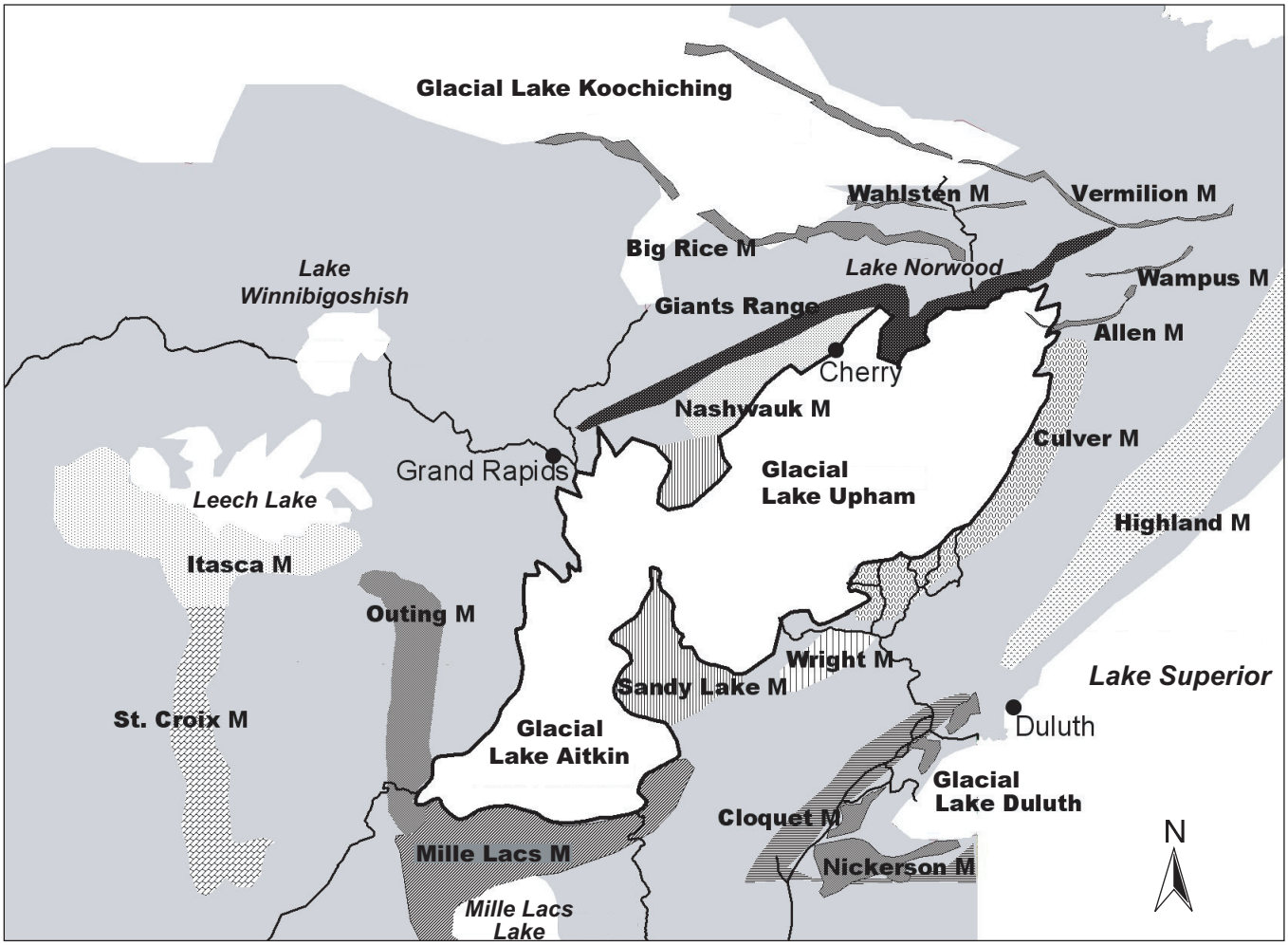


Figure 3.24. Prominent moraines (M) in the glacial Lakes Aitkin and Upham basin with some modern features; modified from Mooers (1988) and Lehr and Hobbs (1992).



Figure 3.25. Rainy-lobe outwash overlain by glacial Lake Upham I sediment in St. Louis sublobe till.

Upham I. Although the exact time of formation of glacial Lakes Aitkin and Upham I is not known, it occurred after the Rainy lobe retreated from the St. Croix moraine about 15.5 kyr B.P. (Clayton and Moran, 1982; Mooers and Lehr, 1997), and before the St. Louis sublobe advanced into the glacial Lakes Aitkin and Upham basin from the northwest.

Rainy-lobe outwash is exposed at the base of the sequence. The upper portion of the sequence is a St. Louis-sublobe till. Between the outwash and till are elongate slabs of fine-grained lacustrine sediment derived from glacial Lake Upham I. This lacustrine sediment was eroded from deeper water and thrust onto the outwash during the advance of the St. Louis sublobe (Fig. 3.25).

STOP 3-15

Artichoke/St. Louis River terrace with loess

Location: T. 51 N., R. 18 W., sec. 26, SW, SW Brookston quadrangle

Description: Following stagnation of St. Louis-sublobe ice, meltwater flowed through a combination of surface and englacial channels toward the ice margin, then along the ice margin through a chain of small proglacial lakes.

Among the earliest and highest outlets in the glacial Lake Upham basin from which these small lakes drained were the Us-Kab-Wan-Ka River (427 meters; 1,400 feet) and Chicken Creek (434 meters; 1,425 feet) channels. These channels drained to the southeast along the margin of the Superior lobe, eventually reaching the St. Croix River (Hobbs, 1983). Proglacial lakes and subglacial meltwater in the glacial Lake Aitkin basin drained southward through the Snake channel (381 meters; 1,250 feet) to the St. Croix River and perhaps southwest to the Mississippi River (378 meters; 1,240 feet; Hobbs, 1983). All of these outlets lie above the highest main beaches of glacial Lakes Aitkin and Upham II, and the DEM correction for isostatic rebound indicates ice must have been present in the basin for them to function (Marlow, 2004). Therefore, they were likely associated with small proglacial lakes rather than the main stages of the lakes (Fig. 3.26).

After the St. Louis-sublobe advance, the margin of the Rainy lobe retreated north of the Giants Range resulting in the ponding of water between the retreating ice and the Giants Range, forming glacial Lake Norwood (Fig. 3.27; Winchell, 1901; Hobbs, 1983). Glacial Lake Norwood drained south through the Embarrass Gap, entering glacial Lake Upham II at an elevation of 436 meters (1,450 feet; Hobbs, 1983). This was the first major meltwater inlet to glacial Lakes Aitkin and Upham II. During this time a significant amount of stagnant ice remained in the basin, and it is possible that glacial Lakes Aitkin and Upham were separated by stagnant ice along the Swan River Sill (Fig. 3.27).

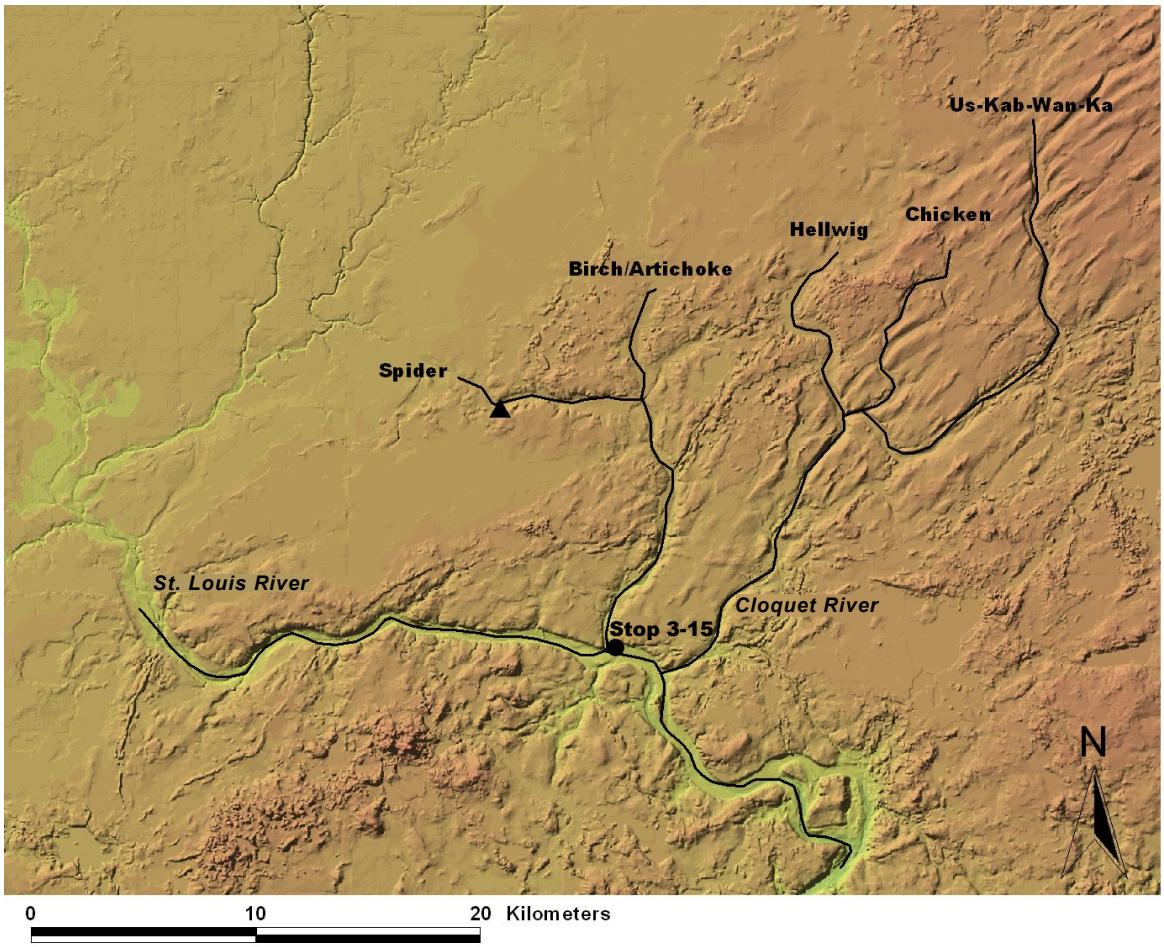


Figure 3.26. Location of early outlets to glacial Lake Upham; triangle marks the location of the core taken by Baker (1965) containing marl.

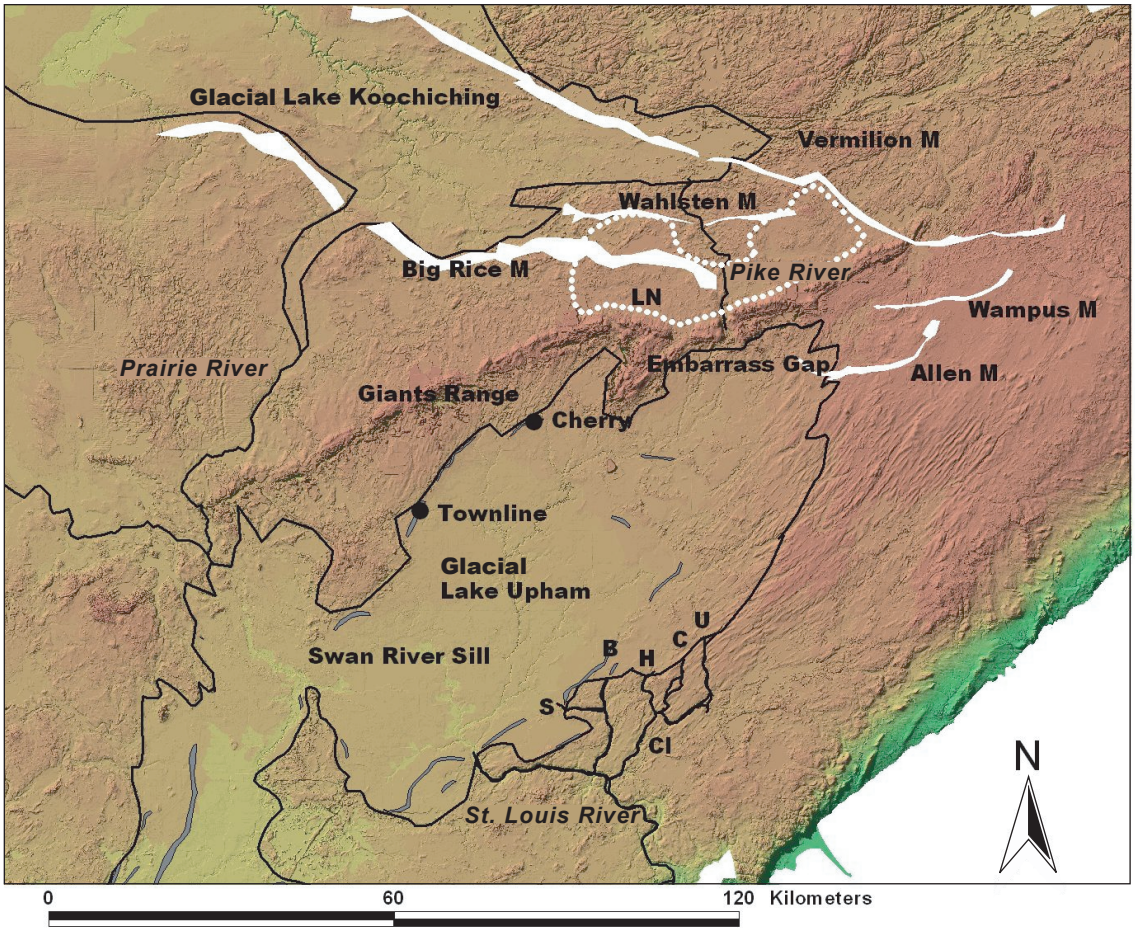


Figure 3.27. Significant features associated with the recession of the Rainy lobe after the St. Louis-sublobe advance. The extent of glacial Lake Norwood (LN) is indicated by a dotted line. White polygons are recessional moraines.

As the Rainy lobe continued to retreat northward, glacial Lake Norwood increased in extent. Downcutting in the Embarrass Gap lowered the level first to 436 meters (1,430 feet) and eventually to 427 meters (1,400 feet). This lower stage is known as glacial Lake Koochiching (Leverett, 1932; Nikiforoff, 1947; Hobbs, 1983; Lehr and Hobbs, 1992).

While glacial Lake Norwood was in existence, the upper strandlines (411 to 421 meters; 1,350 to 1,380 feet) in the extreme northeastern part of glacial Lake Upham II formed. These levels of glacial Lake Upham likely correspond to outlets at Hellwig Creek (409 meters; 1,330 feet), Birch/Artichoke (406 meters; 1,320 feet), and Spider Creek (400 meters; 1,300 feet). A combination of isostatic rebound and downcutting eventually led to successive abandonment of the Hellwig Creek, Birch/Artichoke, and Spider Creek outlets and lake level drop.

Baker (1965) reported a bulk radiocarbon date of 13,000 ± 400 yr B.P. from a sequence of lacustrine marl (sample W-1234) within the Spider Creek outlet (Fig. 3.26). The marl must post-date the cessation of drainage through the channel because marl formation requires shallow, still water. Baker (1965) expressed concern that this date was too old due to possible contamination by lignite. However, this date is consistent with the other evidence presented in this field trip description. The Spider Creek date places the minimum age of glacial Lakes Aitkin and Upham II, and therefore the maximum limit of the St. Louis sublobe, prior to 13.0 kyr B.P.

After the time of upper beach formation, glacial Lake Upham experienced a drop in water level documented at Townline beach (411 meters; 1,350 feet) through the use of Ground Penetrating Radar (GPR; Figs. 3.27, 3.28). The GPR results are interpreted as a down stepping of shoreline deposits (Fig. 3.29). The progradation of the bedform resulted in a constructional shoreline and indicates regression of the lake. The regression is an example of Forced

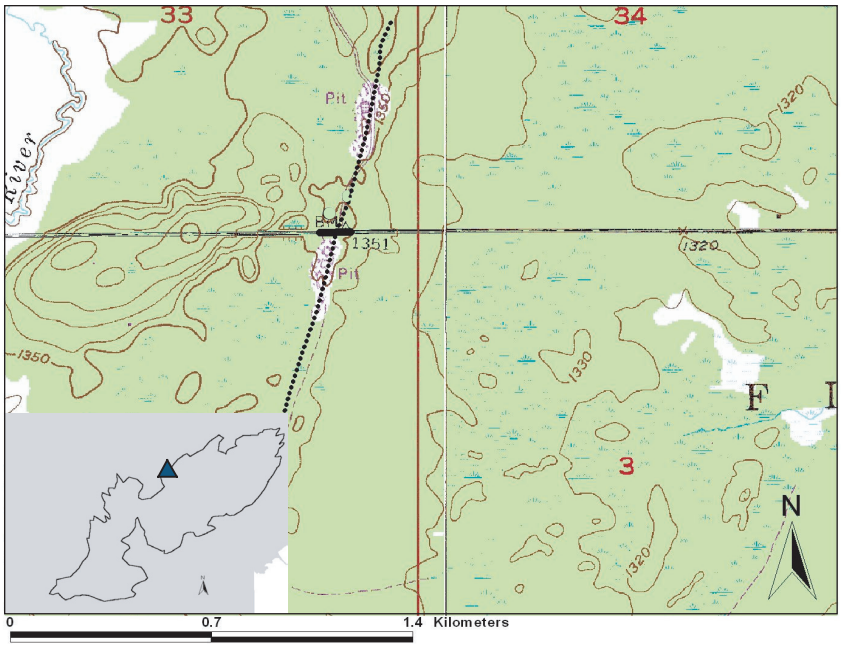


Figure 3.28. Topographic map of the shoreline of glacial Lake Upham II along Townline Road south of Hibbing. Shoreline is indicated by a black dotted line. The heavy solid line indicates where GPR data were collected (Silica and Riley quadrangles). See inset for location.

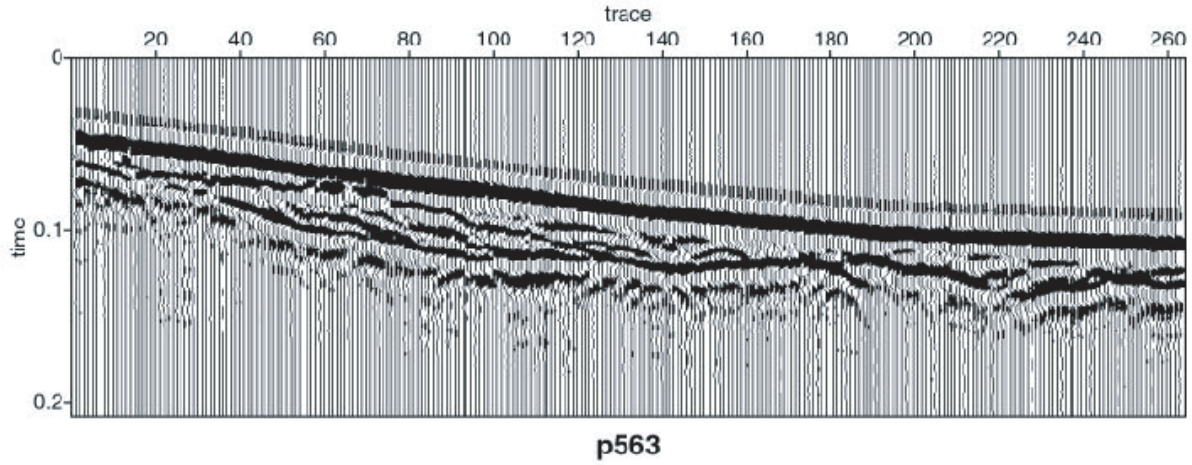


Figure 3.29. Ground Penetrating Radar profile of the glacial Lake Upham II shoreline. Vertical data represent ~9 meters; horizontal data ~121 meters (data processing by Nigel Wattrus at Large Lakes Observatory, University of Minnesota Duluth).

Regression as discussed by Posamentier and Allen (1999). Forced regression takes place when there is a relative sea-level fall that progressively exposes the sea (or lake) floor, thereby causing the shoreline to migrate seaward (Posamentier and Allen, 1999).

At Stop 3-15, the intersection of the Artichoke and St. Louis Rivers, there is a prominent terrace at 375 meters (1,230 feet) that extends 1 kilometer (0.6 mile) across (Fig. 3.30). This is a thick deposit of terrace gravels. The cavities between gravels are filled with loess. The loess likely originated from the glacial Lake Upham basin immediately post-drainage when it was susceptible to eolian activity (Fig. 3.31).

STOP 3-16

Glacial Lake Duluth underflow fan

Location: T. 48 N., R. 15 W., sec. 30, NW, SW

Frogner quadrangle

Description: This location preserves the record of several glacial events that affected the western Lake Superior region and provides insight into their

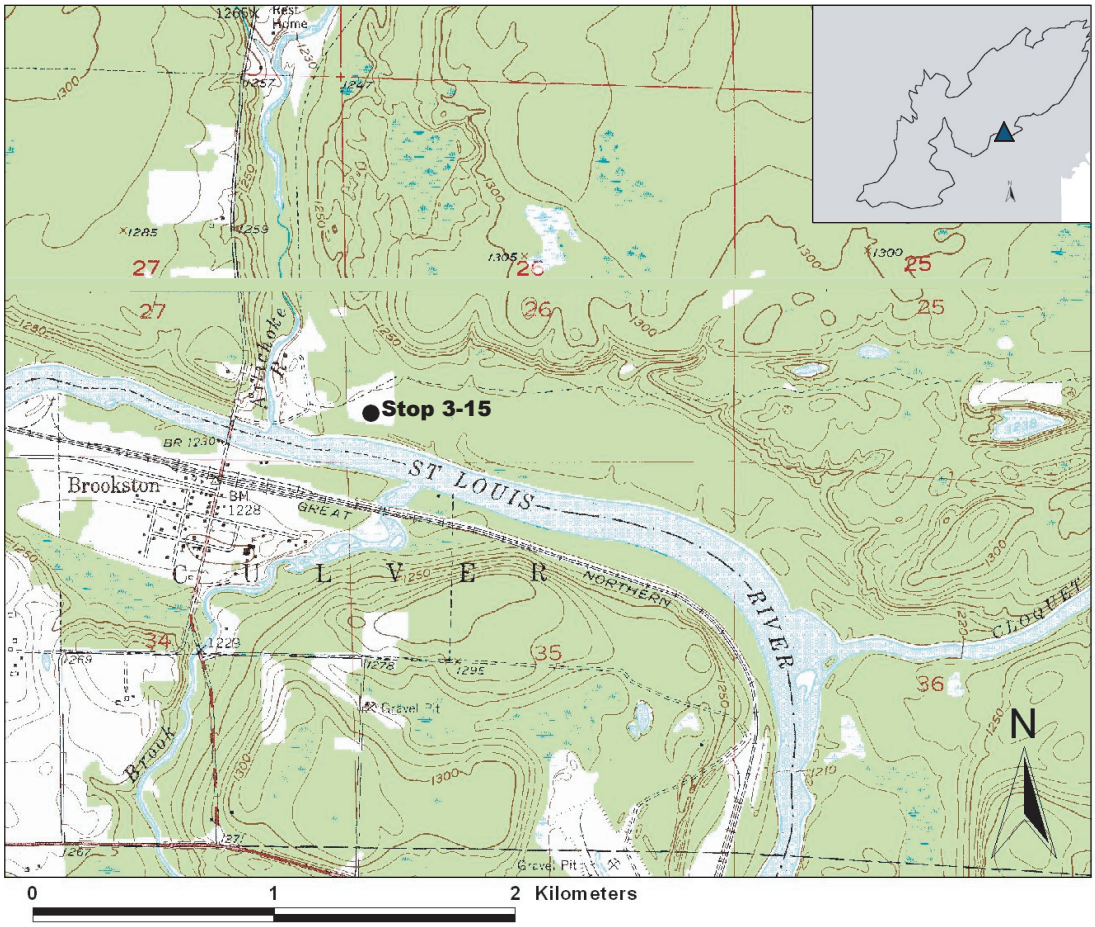


Figure 3.30. Topographic map of the Artichoke/St. Louis River terrace.

relative chronologies. Here, red glacial Lake Duluth clays are overlain by a subaqueous fan deposited by meltwater from glacial Lakes Aitkin and Upham II (Fig. 3.32). These in turn are overlain by Superior-lobe till deposited by the Marquette advance. These relationships indicate meltwater drained from glacial Lake Upham into glacial Lake Duluth (Table 3.2; Wright, 1972; Wright and others, 1973; Clayton and Moran, 1982; Mooers and Lehr, 1997).

During the last phase of glacial Lakes Aitkin and Upham II, the two lakes were thought to have separated and drained through different outlets; glacial Lake Upham through the St. Louis River and glacial Lake Aitkin through the Mississippi River (Hobbs, 1983). A digital elevation model (DEM) reconstruction correcting the basin for isostatic rebound confirms the separation of the two lakes (Marlow, 2004). The Mississippi River outlet to glacial Lake Aitkin is a relatively narrow passage through collapsed ice-cored terrain. It is unlikely this underdeveloped outlet carried significant meltwater. Final drainage of glacial Lake Aitkin therefore occurred after significant meltwater contributions to the lake basins had ceased.

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Figure 3.31. Terrace gravels with loess.

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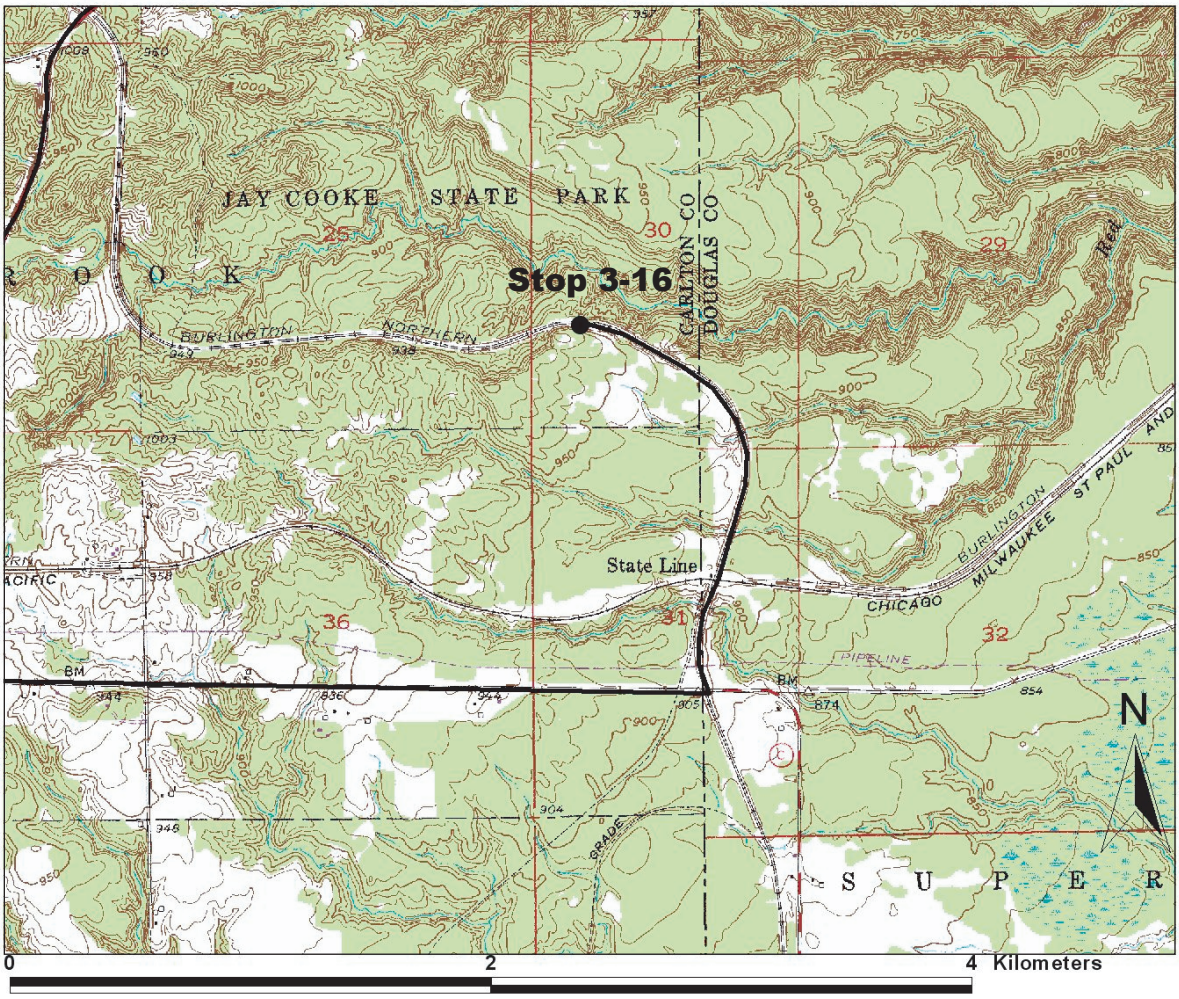


Figure 3.32. Topographic map of the underflow fan at the mouth of the St. Louis River drainage in the glacial Lake Duluth basin. The sediments were deposited by meltwater from glacial Lake Upham II.

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Table 3.2. The glacial chronology of northeastern Minnesota outlined below resulted from a compilation of several resources (Wright, 1972; Clayton and Moran, 1982; Clayton, 1983; Fenton, 1983; Hobbs, 1983; Björk, 1990; Lehr and Hobbs, 1992; Mooers and Lehr, 1997; Marlow, 2004; Larson, unpub. data).

Phase	Moraine	Lobe	Glacial lakes	Other events	C^{14} date (kyr B.P.)
Marquette		Superior	Aitkin II / Upham II / Duluth	Drainage of Aitkin II	9.9
				Drainage of Upham II	
				Cessation of meltwater inflow to Aitkin and Upham II	~11.7
Vermilion	Vermilion	Rainy		L. Duluth underflow fan	12.0
				Inflow from Prairie River	
Big Rice	Big Rice/Wampus	Rainy			
				St. Louis River outlet	
				Inflow from Embarrass Gap	
Nickerson	Nickerson	Superior		Us-Kab-Wan-Ka/Chicken/ Hellwig/Birch/Spider outlets	
				Formation of Goodland esker	
Northof-nashwauk	Allen	Rainy			
Alborn	Culver	St. Louis			
Split Rock	Cloquet	Superior	Aitkin I / Upham I		~13.0
	Sandy Lake	Rainy			
Automba	Outing	Rainy			
St. Croix	St. Croix	Rainy			~15.5-16

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FIELD TRIP 4

Wednesday, May 18

GRANITES OF THE EAST-CENTRAL MINNESOTA BATHOLITH

Leaders

Terry Boerboom, Minnesota Geological Survey
Daniel Holm, Kent State University
Randy Van Schmus, University of Kansas

INTRODUCTION

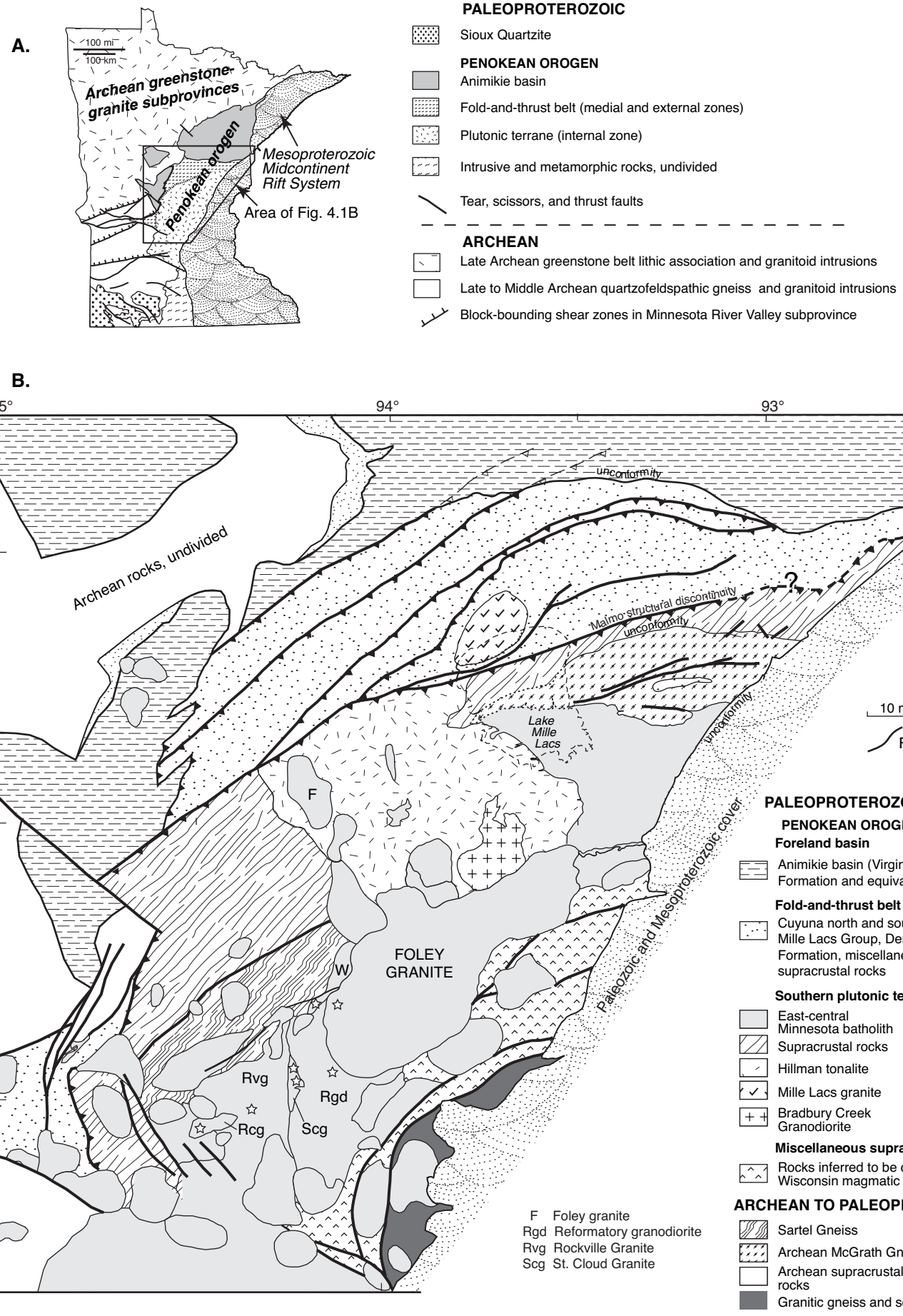
The 1,870 to 1,830 Ma Penokean orogenic rocks of the North American midcontinent are part of a vast belt of juvenile crust accreted onto the southern margin of Laurentia-Baltic continent during late Paleoproterozoic time. Subsequently, the Penokean orogenic belt twice formed the source region for a generation of crustal-melt batholiths: first at 1,790 to 1,770 Ma in east-central Minnesota (Holm and others, 2005), and then again at 1,470 Ma in central Wisconsin (Van Schmus and others, 1975; DeWane and Van Schmus, 2003). The driving forces for magma genesis in ancient Precambrian terranes are difficult to assess. For instance, the generation of dominantly felsic post-Penokean geon 17 magmatism has historically been attributed to thermal weakening of overthickened crust (Windley, 1993; Holm and Lux, 1996), to plumes (Hoffman, 1989), and to abrupt mantle lithospheric delamination (Holm, 1999; Boerboom and Holm, 2000). Most recently, detailed U-Pb zircon age data have led us to suggest that geon 17 magmatism in the southern Lake Superior region was driven by northwest directed subduction beneath the newly accreted Penokean terrane (Holm and others, 2005). On this field trip, we examine the dominantly granitoid rocks that make up the east-central Minnesota batholith in the St. Cloud District. Although volumetrically small, mafic magmas were also apparently generated and commingled with the more abundant felsic magmas during the entire circa (ca.) 20 m.y. interval of batholith generation. Their existence attests to an important mantle contribution in the process of magma genesis. The time span of magmatism and its location on the continental edge of an active plate margin suggest the batholith represents an ancient exhumed continental magmatic arc analogous to many Mesozoic Pacific-rim continental arcs.

This field trip highlights the results of a five-year collaborative study by the authors to better understand the important role of post-Penokean magmatism in the growth and stabilization of this part of the craton.

In addition to their geologic importance, the granitoid rocks of the St. Cloud district have been an important source of building materials. The area has a long quarrying history that peaked between 1930 and 1940 when 18 different quarry companies were operating (Thiel and Dutton, 1935). Currently, rocks of the east-central Minnesota batholith host five active dimension-stone quarries and several crushed-rock aggregate quarries that are essential building blocks to the Twin Cities transportation and building infrastructure.

EAST-CENTRAL MINNESOTA BATHOLITH

The Penokean orogen in east-central Minnesota (Fig. 4.1) consists of a northern foreland basin (the Animikie basin), a medial fold-and-thrust belt, and a southern plutonic terrane (referred to as the internal zone by Southwick and others, 1998). Prior to our recent geochronologic work, the only published U-Pb age data on Paleoproterozoic plutons in east-central Minnesota were a 1,982 ± 2 Ma age on a sheared granite intrusive into the Archean McGrath gneiss and an 1,869 ± 5 Ma date on the deformed Bradbury Creek Granodiorite (Goldich and Fischer, 1986). The Hillman tonalite has a weak to moderate regional northeast fabric, and in places is somewhat migmatitic, with abundant enclaves of high-grade schist. Once interpreted as Archean in age, mapping by Boerboom and others (1999) showed that the Hillman tonalite intrudes metamorphosed Paleoproterozoic country rock (the Little Falls Formation), the Penokean Bradbury Creek Granodiorite, and deformed tonalitic


A.
100 mi
100 km
Archean greenstone-granite subprovinces
Penokean orogen
Mesoproterozoic Midcontinent Rift System
Area of Fig. 4.1B
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'
Archean rocks, undivided
unconformity
Malmo structural discontinuity
unconformity
unconformity
?
Lake Mille Lacs
F
FOLEY GRANITE
W
Rvg
Rgd
Rcg
Scg
Paleozoic and Mesoproterozoic cover
10 miles
Fault
PALEOPROTEROZOIC
PENOKEAN OROGEN
Foreland basin
Animikie basin (Virginia and Thomson Formation and equivalents)
Fold-and-thrust belt
Cuyuna north and south ranges, Mille Lacs Group, Denham Formation, miscellaneous supracrustal rocks
Southern plutonic terrane
East-central Minnesota batholith
Supracrustal rocks
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
F Foley granite
Rgd Reformatory granodiorite
Rvg Rockville Granite
Scg St. Cloud Granite

gneisses that occur in close proximity to the Bradbury Creek Granodiorite.

Recent field, geophysical, and drill core studies from the Minnesota plutonic terrane indicate that it comprises abundant late- to post-tectonic plutons that form a physically continuous area of bedrock over 7,000 square kilometers (Fig. 4.1). Three recent geologic maps that cover the entire plutonic terrane (Boerboom and others, 1995, 1999; Jirsa and others, 1995) categorize the undeformed intrusions into four separate groups: 1. Granodioritic to dioritic rocks, 2. Gray granites, 3. Red granite plutons, dikes, and sills, and 4. Late mafic plugs and diabasic to quartzofeldspathic dikes.

Geologic descriptions of select east-central Minnesota batholith intrusions

The southern plutonic terrane of the Penokean orogen is dominated by the east-central Minnesota batholith, which is composed of some twenty separate intrusions that range from mafic to dominantly felsic-intermediate composition, mostly emplaced between approximately 1,787 to 1,772 Ma (Table 4.1). This field trip will examine several of the post-Penokean intrusions in the southern part of the east-central Minnesota batholith, namely the Richmond, Rockville, Foley, and St. Cloud (red) granites, the Reformatory granodiorite, the Watab diorite, and northeast-trending diabase dikes that post-date the granitoid intrusions. A more thorough description of many of these units, including modal analyses and chemical compositions, is available in Boerboom and Holm (2000).

Richmond granite

The Richmond granite is a dark green to greenish-pink, hypersthene-bearing hornblende-biotite granite with charnockitic affinity, characterized by carlsbad-twinned K-feldspar phenocrysts that range from 2 to 6 centimeters in length, commonly with rapakivi texture (Stop 4-1) that define a consistent east–northeast-trending trachytoid magmatic texture (Fig. 4.2). Quartz typically forms anhedral-interstitial, monocrystalline grains up to 1 centimeter across; however, locally the quartz has been flattened into elongate ribbons that define local meter-scale shear bands parallel to the primary trachytic fabric. The timing of these localized shear bands is not known, but speculatively they may be the product of syn-emplacement deformation of an almost completely crystallized, semi-plastic magma. A volumetrically small part of the Richmond granite is composed of pink, quartz-rich, feldspar-phyric leucogranite and associated aplite.

The dominant mafic minerals in the Richmond granite are reddish-brown biotite that commonly forms symplectic intergrowths with quartz, and dark green to brown hornblende that typically forms subpoikilitic grains rimming pyroxene. Hypersthene is locally abundant, especially in the dark green, quartz-poor phases, and it is typically rimmed or replaced by shaggy overgrowths of fine-grained actinolite intergrown with secondary opaque oxides, or more rarely by "iddingsite." Apatite, ilmenite, and zircon form abundant accessory phases. Yellow garnet is rarely present. Myrmekitic plagioclase-quartz intergrowths are common at the margins of microcline phenocrysts.

The Richmond granite contains scattered enclaves of medium-grained ortho- and clinopyroxene-bearing ferromonzodiorite. These enclaves are similar in lithology and texture to an area of outcrops within the Richmond granite of dark gray, medium-grained, weakly porphyritic and trachytic, apatitic hornblende-biotite-quartz ferromonzodiorite. Only one xenolith of country rock, gray tonalitic gneiss at least 2 meters in size, was noted in this granite.

Outcrops of hornblende ferrogabbronorite with inclusions of noritic anorthosite are located less than one mile northwest of the westernmost Richmond granite outcrops. These outcrops are part of a larger geophysically defined composite intrusion that is composed mostly of mafic rocks and includes the Richmond granite.

The Richmond granite may be analogous to ~1,000 Ma porphyritic intrusive charnockites from the Prince Charles Mountains, east Antarctica (Zhao and others, 1997). These plutons range from 1 to 110 square kilometers, are dominated by simply twinned K-feldspar phenocrysts up to 7 centimeters long, and contain plagioclase, orthopyroxene, clinopyroxene, biotite, and quartz, with accessory apatite, ilmenite, and magnetite. The Richmond granite has chemical

Figure 4.1. A. Simplified Precambrian bedrock map of Minnesota showing the location and extent of the Penokean orogen in Minnesota.

B. Simplified Precambrian bedrock geologic map of east-central Minnesota showing the location of the east-central Minnesota batholith within the Penokean orogen. Stars designate approximate locations of stops shown on Figure 4.10.

Table 4.1. Summary of new U-Pb and Ar-Ar isotopic results from east-central Minnesota.

Field trip stop	Rock unit	U-Pb age (Ma)	Ar-Ar age (Ma)
4-1	Richmond granite	1,772 ± 3	———
4-2	Rockville Granite	1,780 ± 7	1,739 ± 13 (bio)
4-2		———	1,739 ± 19 (hbd)
4-3,4-4	St. Cloud Granite	1,779 ± 5	1,742 ± 13 (bio)
4-3,4-4	Reformatory granodiorite	1,783 ± 11	1,760 ± 6 (bio)
4-3,4-4		———	1,788 ± 3 (hbd)
4-4	QFP dike	1,174 ± 7	———
4-5	Watab diorite	1,780 ± 6	1,768 ± 8 (hbd)
4-6	Foley granite	1,779 ± 4	———

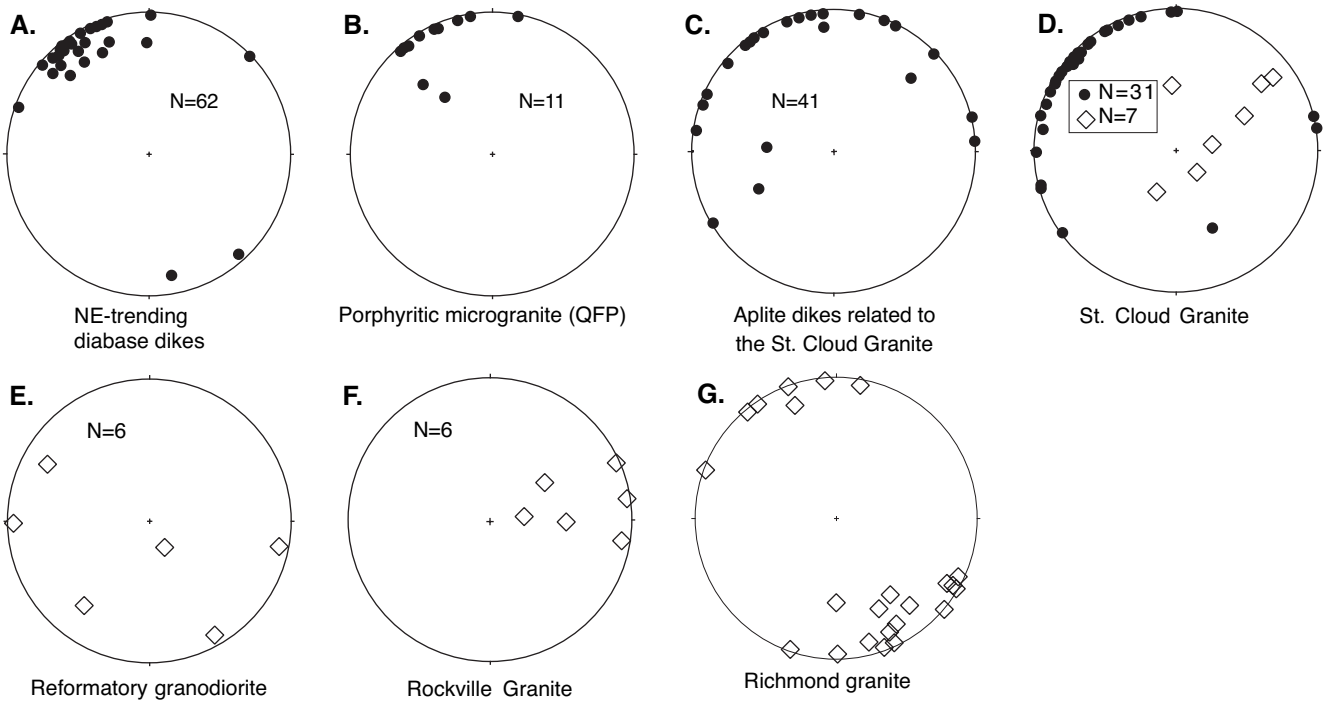


Figure 4.2. Stereographic projections showing measured orientations of dikes and sills (A-D, black circles), and trachytoid fabrics (D-G, diamonds). All data are projected into the lower hemisphere as poles to planes.

and mineralogical traits similar to those in Antarctica, which were emplaced into granulite-grade metamorphic rocks immediately after the peak of metamorphism.

Rockville Granite

The Rockville Granite is presently quarried for dimension stone at three separate locations, and is marketed by the Cold Spring Granite Company as Rockville White (Stop 4-2A), Rockville Beige (Stop 4-2B), and Diamond Pink.

The Rockville Granite is a very coarse-grained, pinkish- to grayish-white, weakly rapakivi-textured granite characterized by 1- to 3-centimeter microcline phenocrysts as well as smaller oligoclase (Johnson,

1978) phenocrysts in a groundmass of quartz, feldspar, biotite, and hornblende, with accessory zircon, apatite, titanite, epidote, and allanite. Poikilitic yellow garnet was noted in one sample. The Rockville Granite locally displays a weak north-striking, west-dipping trachytoid fabric (Fig. 4.2A). Semi-brittle shear bands with quartz ribbons, similar to those in the Richmond granite, were noted in large boulders but not in outcrop.

The Rockville Granite contains scattered mafic enclaves that vary from coarse-grained, black, hornblende-biotite diorite to fine-grained, gray, hornblende-biotite quartz monzonite; both have a pristine trachytoid igneous texture. The diorite commonly contains microcline megacrysts, whereas the monzonite contains scattered plagioclase phenocrysts. Round, 3- to 15-centimeter diameter enclaves of microcline-phyric diorite were noted in a quarry just south of the town of Rockville (Fig. 4.3A; Stop 4-2A). In contrast, a 5- to 30-centimeter-thick dioritic dike with a similar texture cuts the Rockville Granite in a quarry to the northeast. This dike has indistinct margins, and contains xenoliths of white, equigranular granite that are compositionally similar to the Rockville Granite (Fig. 4.3B).

As discussed below, these enclaves, the microcline phenocrysts in the enclaves, and the dike of similar lithology are indicative of the commingling of mafic and felsic magmas during magma ascent and emplacement.

Reformatory granodiorite

The Reformatory granodiorite, named after the Minnesota State Reformatory (or St. Cloud State Reformatory) that is built from this rock (Stop 4-7), is presently extracted for dimension stone from only one quarry near St. Cloud, marketed as Charcoal Gray. However, there are numerous inactive pits from which this rock was once quarried.

The Reformatory granodiorite is a dark to pinkish-gray, medium-grained, weakly porphyritic biotite-hornblende granodiorite. It is characterized by small, gray, tabular oligoclase to andesine phenocrysts (Johnson, 1978) as much as 5 millimeters long, in a groundmass of plagioclase, microcline, hornblende, and biotite, with accessory pyroxene, titanite, opaque Fe-Ti oxides, and apatite. Weak, generally subhorizontal trachytoid fabric can be observed locally in the granodiorite (Fig. 4.2). Ovoid, dark gray enclaves ranging from 1 centimeter to rarely 3 meters in diameter are present in varied proportions throughout the Reformatory granodiorite. These enclaves were previously interpreted as inclusions of amphibolitic country rock, but their mineralogy

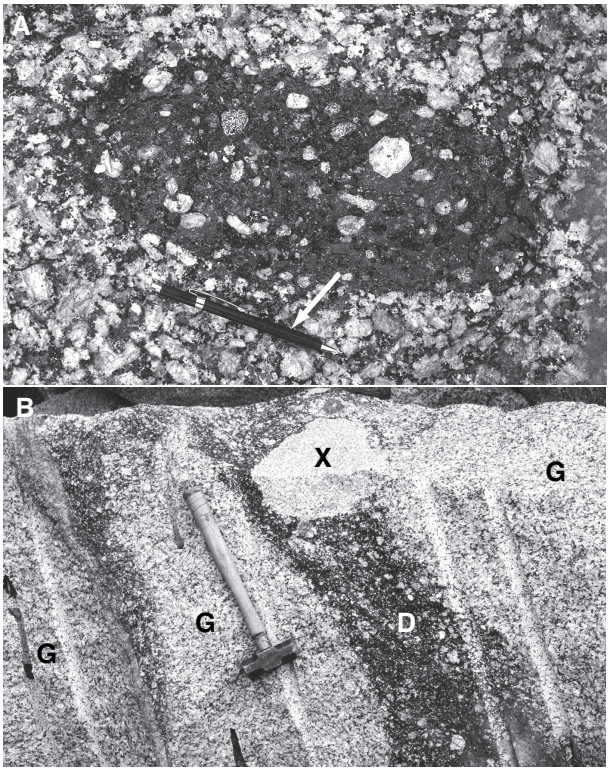


Figure 4.3. A. K-feldspar and plagioclase phenocrysts in a mafic dioritic enclave in the Rockville Granite, from the quarry at Stop 4-2A. Arrow points to a pencil for scale.

B. Rockville Granite (G) cut by a biotite-hornblende quartz diorite dike (D), and containing a xenolith (X) of equigranular white granite. The composition and texture of the dike (D) is similar to the enclave in Stop 4-2A.

and pristine trachytoid igneous fabric are inconsistent with such an origin. We interpret the enclaves to represent mafic magma incorporated within granitic melts because of magma mingling processes.

Metamorphic country-rock xenoliths in the Reformatory granodiorite are rare and range in size from less than 50 centimeters to as much as several meters. The xenoliths consist mostly of high-grade quartzofeldspathic gneiss that contains abundant cordierite and variable amounts of garnet, sillimanite, and magnetite.

St. Cloud Granite

The St. Cloud Granite was once quarried extensively, and hence can be readily observed in abandoned quarries such as those in the SW 1/4 of section 20, and the SE 1/4 of section 19, T. 124 N., R. 28 W. (Stops 4-3, 4-4). Production started in the

early 1900s, peaked around 1940, and ceased in the mid-1960s.

The St. Cloud Granite is a coarse-grained, dusky red colored, hornblende granite, with a distinctive mottled pink and black color, and includes a minor proportion of pink aplitic granite differentiate. The main minerals are microcline, plagioclase (albite to oligoclase; Johnson, 1978), quartz, hornblende, and biotite, with accessory Fe-Ti oxides, apatite, zircon, titanite, and epidote. A weak porphyritic texture is defined by pink microcline phenocrysts that are only slightly larger than the other minerals. A vague trachytoid fabric, defined by aligned feldspar and hornblende crystals, is common. This fabric parallels the margins of the dikes and sills of the St. Cloud Granite (Fig. 4.2). The St. Cloud Granite typically contains abundant angular xenoliths of Reformatory granodiorite and rare xenoliths of garnet amphibolite.

On the basis of recent mapping and geophysical interpretations, the St. Cloud Granite is thought to have been emplaced along the contact between the Rockville Granite and Reformatory granodiorite. Although on published geologic maps the St. Cloud Granite is shown as a continuous body, in reality it is composed of an intertwined mixture of sills and dikes of varied thickness that intrude the Reformatory granodiorite. In places along its western margin near the Rockville Granite, the texture coarsens to the point where it resembles the Rockville Granite, and a cogenetic relationship between those two cannot be ruled out. The St. Cloud Granite may also be related to the much larger Foley granite batholith (Stop 4-6). U-Pb ages for the Foley, St. Cloud, and Rockville granites are very similar (Table 4.1).

Watab diorite

The Watab diorite (Stop 4-5) forms an irregularly-shaped, north–northeast-trending body adjacent to the western edge of the Foley batholith. The bulk of the Watab diorite is dark gray, fine-grained, apatitic quartz-hornblende diorite, but pinkish-green granodioritic phases are common, as are felsic clots and net-vein-like segregations and dikes. The diorite contains minor pyroxene, but most original pyroxene has been converted to a retrograde assemblage of hornblende and biotite, either by deuteric alteration or retrogressive thermal metamorphism related to emplacement of the surrounding intrusions.

Inclusions with a texture and mineralogy like the Watab diorite are common in the Reformatory granodiorite, but locally the Watab diorite textures grade into textures similar to the Reformatory granodiorite, possibly indicating a cogenetic relationship between the two such as a border phase effect. However, the apparent field relationships between the Watab diorite and the Reformatory granodiorite conflict with the apparent U-Pb ages (Table 4.1). This might be explained by the considerable overlap in the error margins for both, or else that the Watab diorite-like inclusions in the Reformatory granodiorite are of a similar but unrelated rock type.

Foley granite

The Foley granite is similar to the St. Cloud Granite, but more pale pink in color and lower in mafic content. The extent of the Foley granite is well defined by aeromagnetic data, which show that it covers a broad area (Fig. 4.1). Geophysical models (Chandler and others, in press) indicate that the Foley batholith forms a thin, lens-like body that is approximately 1.5 kilometers thick, underlain by the Reformatory granodiorite on the south and the Hillman tonalite on the north.

Quartz-feldspar porphyry dikes

Northeast-trending dikes (Fig. 4.2) of porphyritic microgranite, herein referred to as quartz-feldspar porphyry dikes, cut the St. Cloud, Foley, and Rockville granites and Reformatory granodiorite. None of these dikes were noted in the Richmond granite, consistent with it having the youngest U-Pb zircon age of the east-central Minnesota batholith (Table 4.1). The quartz-feldspar porphyry dikes are typically 2 to 4 meters wide, but one is at least 30 meters wide. An abandoned quarry (Crystal Gray Quarry) just southwest of the intersection of Interstate 94 and the Sauk River, is in a porphyritic granite that has phenocrysts with the same type of zoning and shape as those in the smaller quartz-feldspar porphyry dikes. Although the "Crystal Gray" has a more coarse-grained groundmass than the typical quartz-feldspar porphyry dike, it may represent either an unusually wide (greater than 100 meters) quartz-feldspar porphyry dike, or a small stock equivalent to the quartz-feldspar porphyry dikes.

The quartz-feldspar porphyry dikes have chilled, locally flow-banded margins and are texturally similar to volcanic rhyolite (Fig. 4.4). They contain 3 to 20 percent euhedral phenocrysts of microperthite up to 10 millimeters in size, and 3 to 15 percent euhedral to slightly resorbed quartz phenocrysts up to 4 millimeters in diameter. The groundmass consists of fine-grained, anhedral-granular to spherulitic quartz and feldspar with varied amounts of fine-grained biotite, hornblende, actinolite, and chlorite,

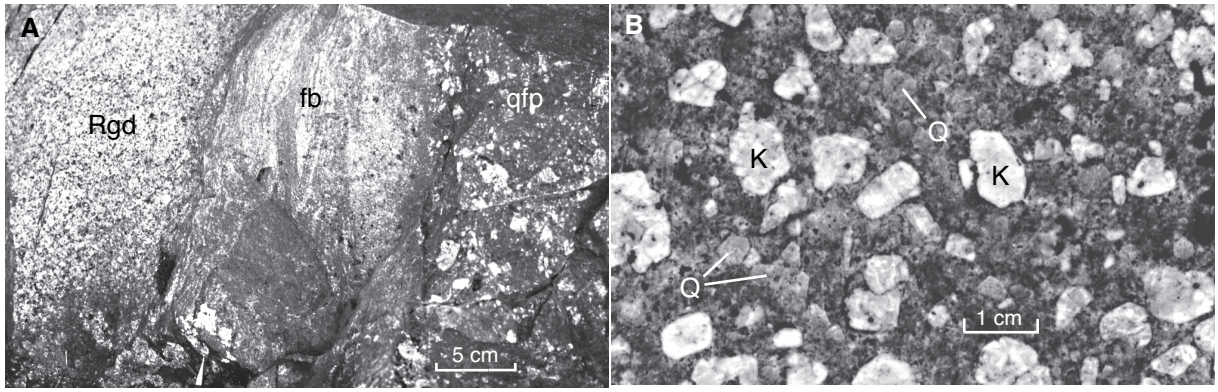


Figure 4.4. A. Porphyritic microgranite (quartz-feldspar porphyry) dike with flow-banded chilled margin (fb) that cuts the Reformatory granodiorite (Rgd).

B. Texture of the quartz-feldspar porphyry on a sawn face, shown actual size. K-feldspar phenocrysts (K) and quartz phenocrysts (Q) are flagged. Quartz phenocrysts are difficult to distinguish because of lack of contrast between them and the pinkish-gray matrix.

and accessory fluorite, apatite, Fe-Ti oxides, epidote, titanite, zircon, and allanite.

The quartz-feldspar porphyry dikes occur in close proximity to the northeast-trending diabase dikes, but due to their similar orientation, crosscutting relationships are rarely observed. At one location a quartz-feldspar porphyry dike is apparently cut by northeast-trending diabase, at another they are apparently comagmatic (see the description of diabase dikes below). At the north wall of the Martin-Marietta Aggregates quarry, a quartz-feldspar porphyry dike clearly intrudes and occupies the center of a diabase dike, but this diabase has a different mineralogy and texture than the other diabase dikes, as described below.

Northeast-trending diabase dikes

These dikes are exposed in several outcrops and old quarries in the St. Cloud area. They are known to cut all the granitoid rocks visited on this field trip, except that none were observed in the Richmond granite. They are mostly 1 to 3 meters thick, but locally as much as 8 meters thick, are vertical to steeply south dipping (Fig. 4.5; Stop 4-3), and generally show only a vague, normally-polarized aeromagnetic expression. They have chilled margins and follow preexisting fractures in the granites they intrude, and rarely contain slivers of granite xenoliths. The dikes have been weakly metamorphosed or deuterically altered, with primary pyroxene now replaced by a mixture of actinolite and chlorite; however, the rock retains its primary igneous texture and there is no metamorphic fabric. Some small diabase dikes (less than 50 centimeters) contain abundant xenocrysts of clear quartz and

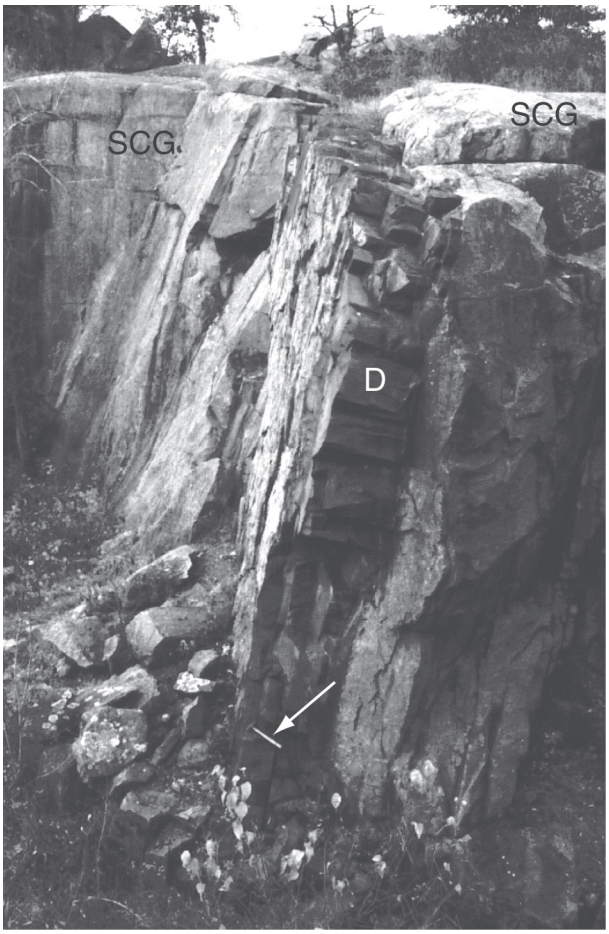


Figure 4.5. Typical diabase dike (D) exposed by quarrying operations in the SE 1/4 sec. 19, T. 124 N., R. 28 W. (Stop 4-3B). View is to the west–southwest. The dike strikes roughly N. 50° E., dips steeply south, and here intrudes the St. Cloud Granite (SCG). The hammer (arrow) shown near the lower part of the photograph has a 40-centimeter-long handle.

turbid feldspar as well as small xenoliths of granitic rock fragments. These xenocrysts show a great deal of resorption and are typically concentrated by flow segregation into the centers of the dikes. The relationship between these xenocrystic dikes and the main northeast-trending diabase dike swarm is not known, but they may represent the early onset of this episode of mafic magmatism. K-Ar and Ar-Ar whole rock ages ranging from 1,164 to 1,570 Ma are likely to be only minimum age estimates (Hanson, 1968; Boerboom and Holm, 2000). Horan and others (1987) reported that Pb isotope data are consistent with an age of about 1,800 Ma.

One northeast-trending diabase dike is sharply chilled against the Rockville granite on one margin, but is not chilled against the contact with a quartz-feldspar porphyry dike on the other. The diabase adjacent to the quartz-feldspar porphyry dike contains xenocrysts of quartz and feldspar derived from the quartz-feldspar porphyry dike. These features indicate a possible cogenetic relationship between the diabase and the quartz-feldspar porphyry dike.

Another northeast-trending diabase dike present along the north wall of the Martin-Marietta Aggregates quarry may represent a separate, earlier generation of dikes. This dike has ophitic and trachytoid textures, and contains small phenocrysts of plagioclase, chromite, and altered olivine, with groundmass pyroxene variably replaced by fibrous amphibole. It is chilled against the Reformatory granodiorite wall rocks, but is itself clearly intruded by a quartz-feldspar porphyry dike. The relationship of this dike to the main diabase dike set is not known, but again may indicate close temporal relationships between mafic and felsic dike sets.

Northwest-trending diabase and ferrodiorite dikes

This dike set will not be visited by this field trip, but is worth noting here. The northwest-trending diabase dikes are inferred from geophysical data to be as much as 100 meters wide, and can be easily traced by their prominent, reversely polarized aeromagnetic signature. Aeromagnetic data indicate that this dike set cuts all of the components of the east-central Minnesota batholith, including the Richmond granite, and they may represent the youngest of the diabase dike swarms in this area.

A small exposure in northwestern Stearns County in the town of Sauk Centre is of the same lithology as a short drill core that is located along strike over the same linear reverse magnetic anomaly. Both of these rocks are dark green, medium-grained, deuterically altered, apatitic quartz-hornblende ferrodiorite that contains both magmatic hornblende and fibrous uralitic amphibole after primary pyroxene. They contain abundant oxides, including titanomagnetite, in which the ilmenite has been variably replaced by fine-grained titanite and/or leucoxene that encloses skeletal (111) magnetite lamellae. The ferrodiorite in outcrop displays a weak trachytoid igneous fabric that strikes N. 60° W., parallel to the magnetic anomaly beneath the outcrop.

Two other northwest-trending diabase dikes are exposed in the St. Cloud area, one of which is chilled against and crosscuts a northeast-trending diabase dike. Both of these dikes are fine-grained and have chilled margins, and compared to the other diabases are less altered and metamorphosed, more strongly magnetic (magnetite and pyrrhotite), and contain fresh clinopyroxene. One of these northwest-trending dikes was sampled and measured for remanant magnetism, and is normally polarized, thus these may not be representative of the geophysically prominent, reversely polarized dike set.

Mafic microgranular enclaves: evidence for magma mingling

The textures, occurrence, geochemistry, and possible origins of mafic microgranular enclaves were summarized by Didier (1973) and Didier and Barbarin (1991). Mafic microgranular enclaves are defined as enclaves of granitoid igneous rocks that typically have a grain size finer than the host granitic rock. Mafic microgranular enclaves are most commonly ovoid in shape, centimeter to decimeter in size, with sharp contacts against the granitic host. Larger minerals such as K-feldspar phenocrysts commonly cross the boundary between the host and mafic microgranular enclaves. Typical mafic microgranular enclaves have zoning and quench textures indicative of undercooling of the mafic magma against the relatively cool granitic host. They contain plagioclase laths aligned by magmatic flow in a groundmass of poikilitic quartz and K-feldspar (Vernon, 1991), and have igneous textures that are distinct from the typical foliated or granoblastic metamorphic rocks. K-feldspar megacrysts are common in mafic microgranular enclaves, and are attributed to the passage of preexisting feldspar phenocrysts from the felsic magma into the more mafic melts that form the enclaves (Didier and Barbarin, 1991). Rapakivi textures can also be attributed to magma mixing (Hibbard, 1981, 1991).

The textural and mineralogical attributes of mafic microgranular enclaves in the Rockville Granite and Reformatory granodiorite are similar to features attributed elsewhere to the commingling of

mafic and felsic magmas. The mafic enclaves in the Reformatory granodiorite and Rockville Granite are ovoid in shape, with quenched, trachytoid plagioclase and clinopyroxene crystals set in a groundmass of poikilitic quartz and K-feldspar. In the Rockville Granite, commingling of mafic and felsic magmas during magma ascent is indicated by the presence of coarse dioritic phases that occur as both porphyritic enclaves and dikes (for example Fig. 4.3), and fine-grained mafic microgranular enclaves that show pristine trachytoid igneous textures.

Origin of mafic and felsic magmas

The mingling of mafic and felsic magmas is most likely in regions where crustal melts are generated by underplating of continental crust by mafic, mantle-derived magma (Cruden and others, 1995, and references therein). Crustal melts generated by the underlying mafic magma will accumulate at the base of the crust, and if sufficient quantities are generated, the melt may become fully mobilized, ascend rapidly, and entrain some of the underlying mafic magma as it rises. Mafic magma underplating may take place in a variety of tectonic settings (for example Turner and others, 1992; Cruden and others, 1995; Holm, 1999). During underplating, increased thermal input at the base of the crust produces a granitic melt that commingles, in various proportions, with the underlying mafic, mantle-derived magma. The degree of mixing and interaction between the two magmas is dependent upon several factors, including relative differences in viscosity, density, thickness, temperature, and composition, as well as degree of crystallization within the mafic underplate, the ponded felsic melt, and the overlying crust. The rate of magma ascent also affects the extent of interaction between the two magma types (Cruden and others, 1995). Several lines of evidence indicate that at least some of the Paleoproterozoic granites of the St. Cloud district may have formed by crustal underplating by high-temperature mantle material:

1. Based on zircon and apatite saturation thermometry techniques, the granite melts are inferred by Spencer (1987) to have formed at high temperatures (800 to 1,000° C). Spencer (1987) also suggested that the high-temperature conditions, as well as isotopic similarities between the diabasic dikes and granitic intrusions, indicate the interaction of mantle and crustal sources.
2. The presence of mafic microgranular enclaves and the textures of these enclaves, and rapakivi textures in the host granites, are features attributed elsewhere to magma mingling processes (Didier and Barbarin, 1991) that occur as a result of mafic underplating.
3. The close and probable cogenetic association of the porphyritic microgranite dikes with diabase dikes may reflect bimodal magma generation, a common feature of post-orogenic magmatic suites attributed to extension associated with mantle lithospheric thinning (Turner and others, 1992).

To date, no geochemical study has been undertaken on the relationship of the mafic microgranular enclaves and the host granitoid rocks of the east-central Minnesota batholith .

Weathering profiles developed on Precambrian bedrock

Drill core studies and examination of exposures at the edges of some quarries reveal that the Precambrian basement rocks here, as typical in all of Minnesota, have a variably thick and variably preserved section of weathered bedrock. This interval of chemical weathering is known from drill core and outcrop in Minnesota to be formed beneath Cretaceous sedimentary rocks, likely during a time of subtropical climate. Exposures in this area and in southwestern Minnesota show that weathering is concentrated along more permeable fractured zones in the bedrock. As weathering progressed along vertical and horizontal fractures, the weathered zoned merged together and enveloped knots of unweathered bedrock (Fig. 4.6A). These remnants of unweathered rock within the saprolite, or corestones, are prominent features of the Rockville Beige quarry (Fig. 4.6A; Stop 4-2B).

The weathered cap on crystalline basement rocks has been variably eroded by glaciation. In places where the entire weathering residuum is preserved (such as beneath Cretaceous strata), the uppermost weathered section consists of hard, pisolitic, Fe- and Al-rich material that progresses down into clay-rich saprolite that typically retains the original texture of the rock despite alteration. The contact between weathering residuum and fresh rock varies from sharp to gradational, depending on the nature of the bedrock (such as granite vs. schist). A good example of the sharp and undulating nature of the weathered/fresh rock surface can be seen just north of the quarry at Stop 4-2B across the Sauk River. There, erosion by glacial meltwater (vs. erosion by ice contact) in the present-day Sauk River channel has eroded away the saprolite, leaving behind a highly irregular, lumpy outcrop surface that mimics the weathered/unweathered rock interface in the Rockville Granite. In the Rockville area, many of the corestones formerly

A.

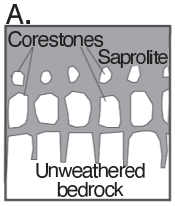


Figure 4.6. A. Diagram illustrating the process of formation of corestones by progressive weathering along fractures.

B. Rockville Beige quarry (Stop 4-2B) showing corestones (C) in saprolite (Sap) on top of unweathered Rockville Granite (Rvg).

B.



enveloped in saprolite have been moved slightly by glacial transport, forming ready-made, well-rounded "glacial erratics" that have been moved only a very short distance.

Geochronology

Holm and others (2005) reported U-Pb zircon age data from 26 plutons and dikes in the plutonic terrane of east-central Minnesota. Of these, 20 separate, undeformed intrusive bodies yield precise ages between 1,787 and 1,772 Ma. Six plutons gave pre-batholithic ages of 2,009 Ma (Mille Lacs granite), 1,877 and 1,858 Ma (Bradbury Creek and tonalitic gneiss), and ca. 1,800 Ma (late tonalite phase of the Hillman migmatite). Similar post-tectonic granites in Wisconsin, reanalyzed in their study, gave consistently younger ages ranging from 1,759 to 1,746 Ma. Their study showed that late to post-Penokean magmatism occurred at ca. 1,800, 1,775, and 1,750 Ma, and generally migrated southeastward across the newly accreted Penokean terrane (Fig. 4.7). These data show that previous references to "1,760 Ma" magmatism and a long-lived magmatic hiatus (between 1,830 and 1,760 Ma) are no longer valid.

Metamorphism of country rock, depth of emplacement, and exhumation history

The plutonic terrane exposes some of the deepest levels of the Penokean orogen regionally (Holm and Selverstone, 1990). Amphibolite-facies Paleoproterozoic rocks from this area were metamorphosed both during and after the Penokean orogeny. The age of the overprinting metamorphism is constrained by $^{40}Ar/^{39}Ar$ and U-Pb metamorphic monazite ages (Holm and Lux, 1996; Holm and others, 1998; McKenzie, 2004). With the exception of a few ca. 1,700 Ma mica dates from the McGrath Gneiss dome, mica and hornblende dates from plutons and country rock in the northern part of the plutonic zone are ca. 1,780 to 1,740 Ma (Holm and Lux, 1996; Holm and others, 1998). In situ metamorphic monazite ages from Paleoproterozoic and Archean country rock reveal a profound ca. 1,770 Ma thermal imprint associated with intrusion of the east-central Minnesota batholith and a secondary ca. 1,800 Ma thermal imprint. The considerable distance of some of these samples from the western edge of the exposed batholith and the absence of Penokean metamorphic

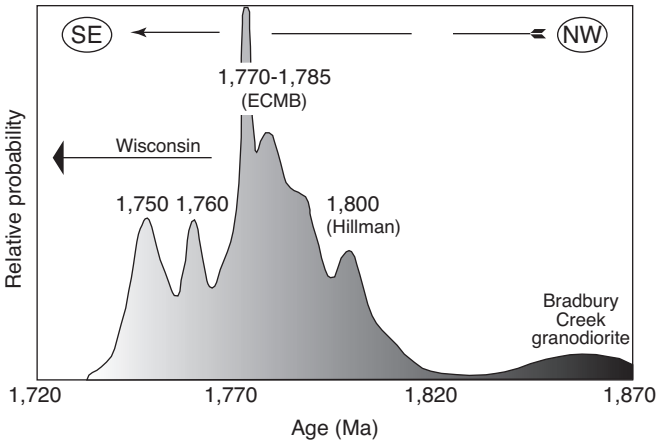


Figure 4.7. Plot of new U-Pb zircon ages for plutonic rocks across the Penokean orogen.

ages suggests that the 1,770 Ma thermal pulse must have been dramatic (McKenzie, 2004).

Application of the Aluminum-in-Hornblende (AH) igneous barometer on phases of the east-central Minnesota batholith indicate paleodepth estimates that increase from approximately 13 kilometers in the north (Holm and others, 1998) to approximately 18 kilometers in the southeast near St. Cloud (Fig. 4.8; Schweitzer and others, 2001). The data from each area are internally consistent for different magmatic phases, implying that very little uplift occurred during emplacement of the east-central Minnesota batholith. This implies that very little uplift occurred during emplacement of the east-central Minnesota batholith. The bulk of post-tectonic exhumation appears to have occurred immediately after intrusion ended. Post-emplacement exhumation is consistent with cooling ages being 10 to 20 m.y. younger than crystallization ages from the batholith.

Tectonic setting

The age of the east-central Minnesota batholith and younger geon 17 plutonic and volcanic rocks in Wisconsin falls within the time span of magmatism in the Yavapai province, northern Central Plains orogen (Van Schmus and others, 1993). We suggest

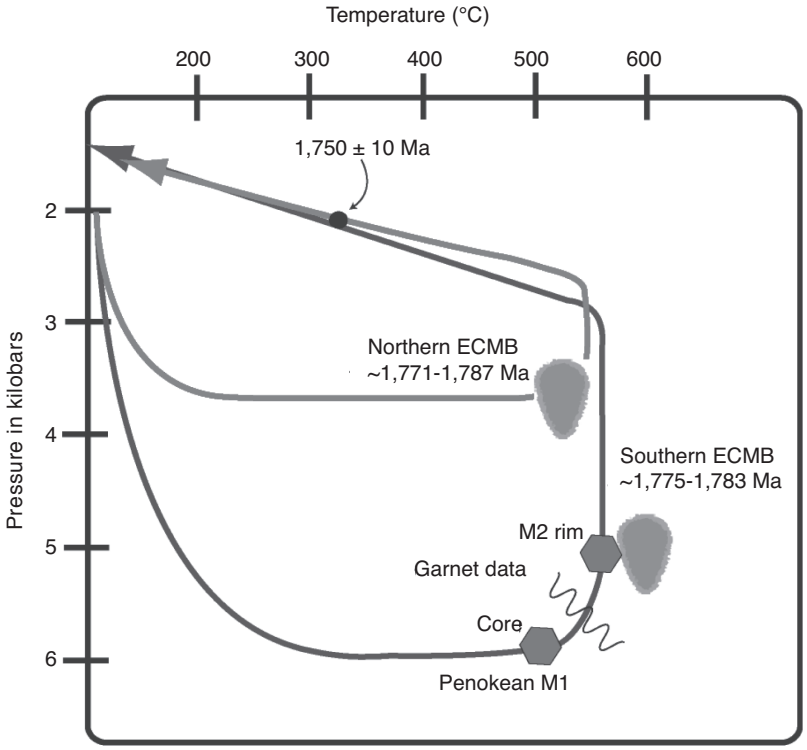


Figure 4.8. Comparison of P-T-t paths for Early Proterozoic rocks near St. Cloud (southern) and those several miles to the northeast (northern). The black circle indicates cooling through closure temperature of biotite. Pressure estimates were taken from Holm and others (1998).

geon 17 magmatism in the southern Lake Superior region was driven by northwest-directed subduction beneath the newly accreted Penokean terrane. The time span of bimodal magmatism and its location on the continental edge of an active plate margin suggest the east-central Minnesota batholith represents an ancient exhumed continental magmatic arc analogous to many Mesozoic Pacific-rim continental arcs. Underplating of continental crust by mafic, mantle-derived magma is an important mechanism by which mafic and felsic magmas can be simultaneously generated and mingled (Cruden and others, 1995). Dewatering of a subducting plate hydrates the warm overlying mantle, triggering partial melting and generation of mafic magmas. Underplating at the base of the overlying crust can, in turn, create felsic crustal melts that ascend rapidly, entraining some of the underlying mafic magma as they rise. The magmas become emplaced when they reach neutral buoyancy or when they are unable to penetrate the high strength lid of the upper crust.

In the scenario presented here (Fig. 4.9), arc magmatism occurred above a northwest-dipping subduction zone following Penokean accretion. New U-Pb zircon ages indicate that similar post-Penokean plutons in Wisconsin are younger than the east-central Minnesota batholith. Migration of arc magmatism to the southeast (Figs. 4.7, 4.9) into Wisconsin at ca. 1,760 to 1,750 Ma is consistent with steepening of the northwest-dipping subducting plate (Holm and others, 2005). Slab rollback would have left the overthickened Penokean orogen unsupported and thus would have facilitated collapse of the Penokean crust and widespread metamorphism. Intrusion and exhumation of the east-central Minnesota batholith was probably an important step in the stabilization history of the newly accreted Penokean crust (Holm and others, 2005).

Contrasting Proterozoic batholiths in the Lake Superior region

We now know that the Penokean orogenic belt was intruded by two batholiths over a 300 m.y. period. Thermochronology and thermal modeling of country rock around the 1,470 Ma Wolf River batholith in southeastern Wisconsin show that the Wolf River batholith had a limited influence on its surrounding country rock (Holm and others, 2004). In contrast, the ~1,770 Ma monazite ages from country rock in east-central Minnesota clearly show that the east-central Minnesota batholith had a dramatic thermal influence. New U-Pb age data show that the east-central Minnesota batholith intruded over an approximately 20 m.y. age range (from 1,790 to 1,770 Ma; Keatts and others, 2004; Holm and others, 2005). In contrast, DeWane and Van Schmus (2003) showed that nine different phases of the Wolf River batholith are all 1,470 Ma. The short-lived magmatic event in Wisconsin may partly explain the limited thermal extent of country-rock heating. Barometric data also show that the two batholiths were emplaced at significantly different depths (Anderson and others, 1980; Schweitzer and others, 2001). The east-central Minnesota batholith was a deep (mid-

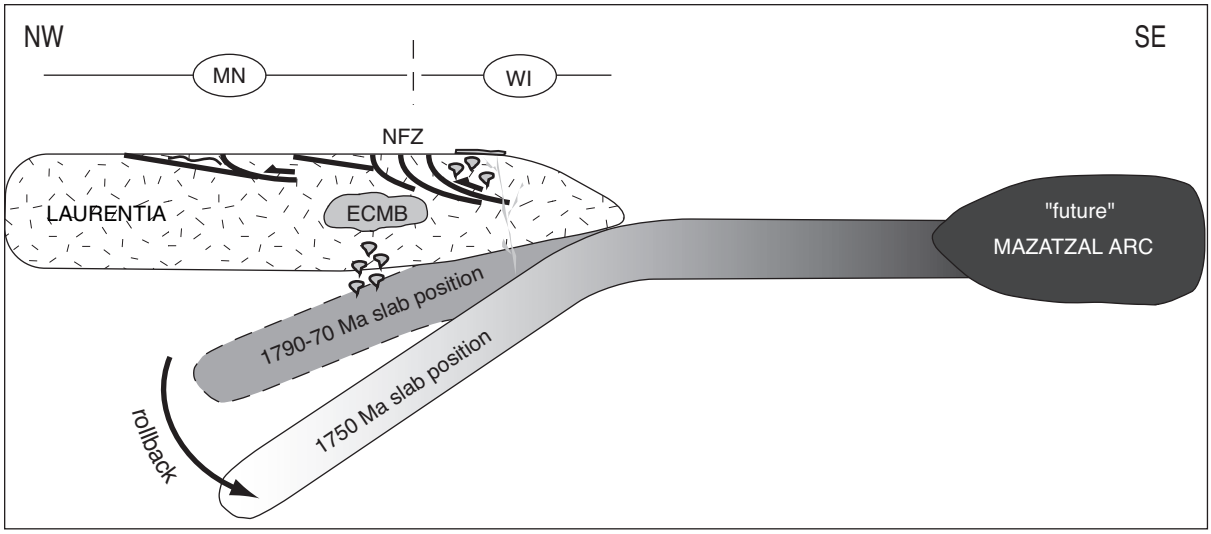


Figure 4.9. Subduction model proposed to explain magmatic arc age progression across the Penokean orogen. South-dipping thrust faults such as the Niagara fault zone in northern Wisconsin (NFZ) are Penokean in age. ECMB—east-central Minnesota batholith.

crustal) magma body that intruded over an extended period of time, thereby thermally affecting county rock for a considerable distance. In contrast, the Wolf River batholith was a shallow (upper-crustal) magma body that intruded over a short period of time, thereby having only a limited thermal effect on its country rock. These differences are likely to reflect the different tectonic setting associated with their emplacement.

FIELD TRIP STOPS

The general locations of all stops are shown on Figure 4.10. A more detailed location map is included with each stop description, with a base made at scale from a portion of the appropriate U.S.G.S. 7.5-minute quadrangle map that should allow the return visitor to accurately locate the features for each stop. The regional setting of the various rock units are given in the introduction, and only facts pertinent to each specific stop are given below.

DIRECTIONS: From Minneapolis, drive west on Interstate 94 approximately 65 miles. Exit on State Highway 23 (exit 164) toward Cold Spring/Rockville. Drive west on Highway 23 for 12 miles to the town of Cold Spring (note: Highway 23 now passes south of the town of Rockville, not through it as shown on the map). Continue west on Highway 23, and turn left

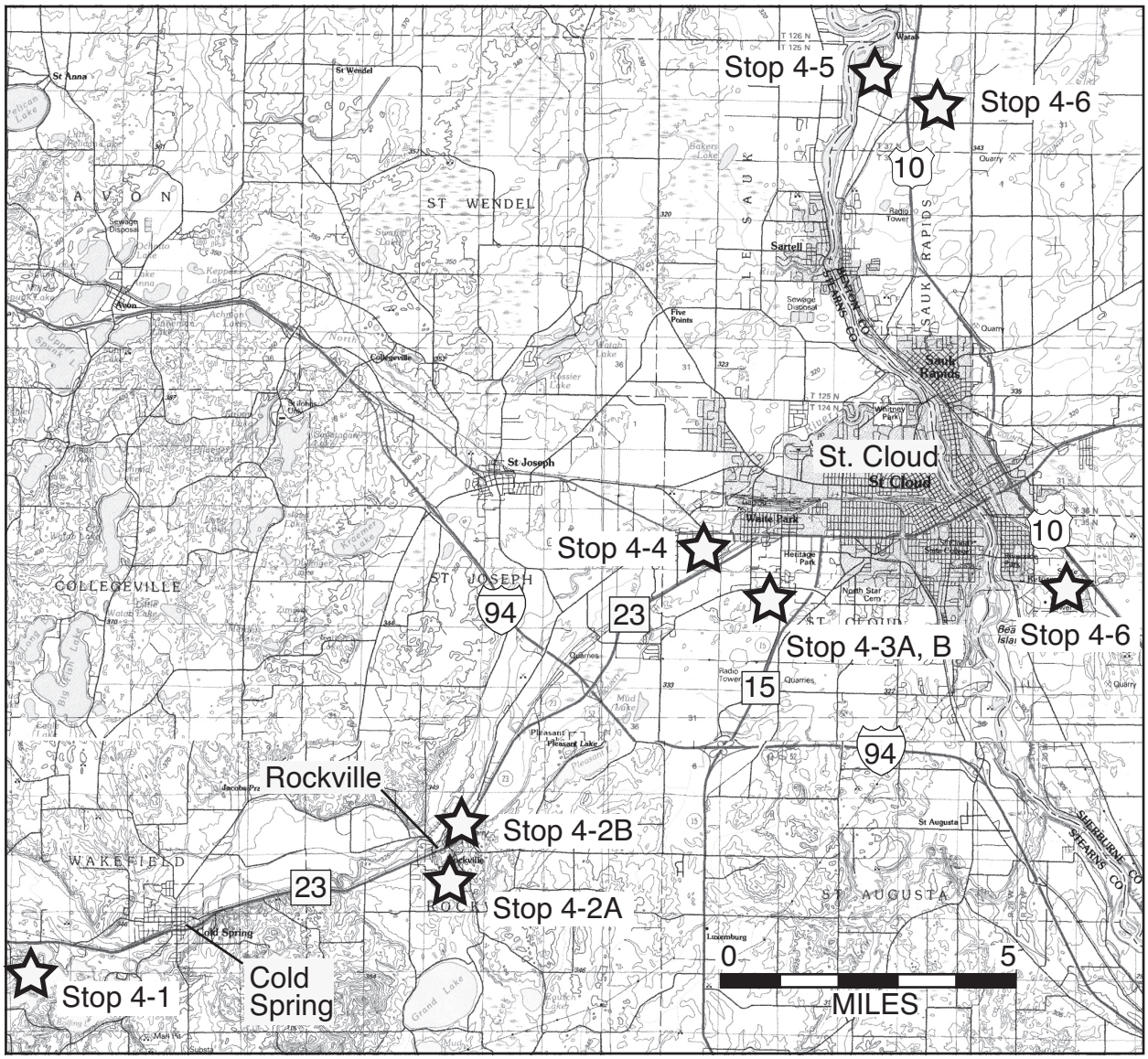


Figure 4.10. General locations of field stops for Field Trip 4.

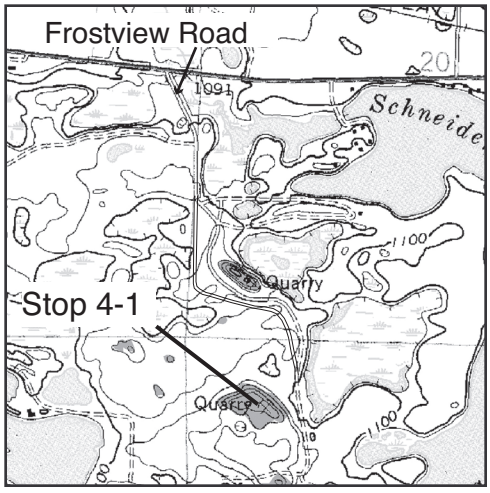
(south) on Frostview Road, which is approximately 3 miles west of the stoplight in Cold Spring. Drive approximately 0.8 mile to 21735 Frostview Road.

STOP 4-1

Private property! Permission must be obtained before entering!

Charnockitic Richmond granite

Location: T. 123 N., R. 30 W., sec. 29, NW Cold Spring quadrangle; UTM: 383,992E/ 5,032,883N



Highlights: Paleoproterozoic Richmond granite—Charnockitic, rapakivi granite with prominent trachytoid K-feldspar phenocrysts.

Description: The Richmond granite yields the youngest U-Pb zircon date of all the intrusions in the east-central Minnesota batholith (Table 4.1; 1, 772 ± 3 Ma; Holm and others, 2005).

Key features of Stop 4-1:

- Dark green charnockitic granite. This color phase grades into more typical pink-tinted Richmond granite, possibly due to oxidation introduced along tight fracture planes. At first glance, the green phase looks like gabbro, but if you trace across the gradation to the pinker phase, you will see that it actually contains abundant quartz and K-feldspar.
- Euhedral, carlsbad-twinned K-feldspar (microperthite) phenocrysts. Some of these are mantled by a thin rim of plagioclase, giving rise to rapakivi texture—best observed on vertical, slightly weathered faces, where plagioclase is bleached white.
- Magmatic-trachytoid alignment of feldspar phenocrysts, oriented N. 45°-60° E., ± vertical. This fabric orientation is similar to that in other outcrops of the Richmond granite.
- Small mafic enclaves. From a petrographic standpoint, these enclaves are mineralogically identical to a fine-grained, trachytoid, apatitic, quartz-orthopyroxene-clinopyroxene monzodiorite that occurs in several outcrops 0.25 mile to the west of this location. This diorite contains abundant zircon and small phenocrysts of both plagioclase and K-feldspar, and locally grades into a coarser-grained phase that is compositionally and texturally similar to the Richmond granite as exposed here. The diorite is inferred to be a phase of the Richmond granite, and the small enclaves scattered throughout the Richmond granite are interpreted as cognate xenoliths of this phase.

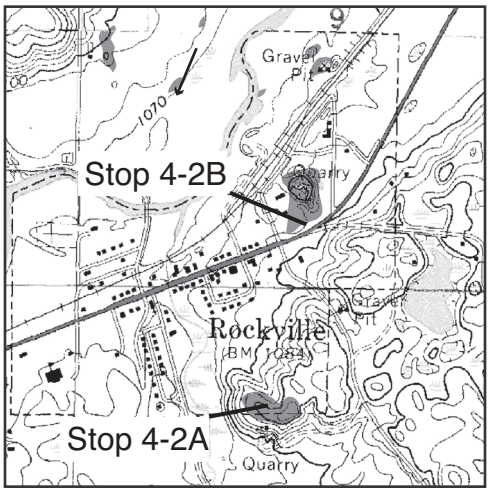
NEXT: Drive back north on Frostview Road to Highway 23. Go east on Highway 23 approximately 5 miles to the town of Rockville and turn into Rockville. Turn east on the main street through Rockville (old Highway 23) and turn right (south) at Pine Street at the east edge of town, past a large gray house made of Rockville Granite. Proceed up Pine Street to the quarry entrance on the right.

STOP 4-2A

Private property! Permission must be obtained before entering!

Rockville Granite "Rockville White" Quarry

Location: T. 123 N., R. 29 W., sec. 16, NW Rockville quadrangle; UTM Start: 395,660E/ 5,035,875N



Highlights: Rockville Granite—porphyritic, weakly rapakivi biotite-hornblende granite; dimension-stone quarry.

Description: This is the "Rockville White" quarry, one of three quarries operated by the Cold Spring Granite Company that are developed in the Rockville Granite. The granite from the other quarries is marketed as "Rockville Beige" and "Diamond Pink." The "Beige" quarry is located approximately 0.5 mile north of this quarry, and the "Diamond Pink" is located about 2.5 miles to the north–northeast. Due to safety concerns, we will not be able to enter the quarry, but will have ample opportunity to view the Rockville Granite up close in the numerous quarried blocks stacked on the premises.

This quarry utilizes the drive-in quarry method, a system that replaces the traditional derrick method of stone removal. The rock is removed in a series of benches, from which stone is removed as a series of large "loaves" that vary in size from 15 to 25 feet high by 20 feet wide by 30 to 150 feet long. These loaves are subsequently drilled off into smaller blocks that are optimally 5.5 x 5.5 x 10.5 feet in size, that weigh approximately 25,000 kilograms (approximately 28 tons). This size is small enough to transport by flatbed trailer without a special permit, and allows for minimum wastage during processing. This quarry and the one at Stop 4-2B are quite productive because the rock is very massive and contains only widely-spaced fractures, which allows for the removal of large blocks of rock and very little waste.

There are many blocks of Rockville Granite stacked near the quarry viewing area. The following textures can be seen in the blocks:

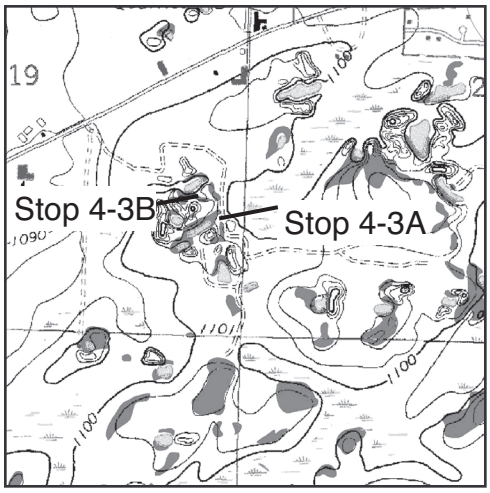
1. Porphyritic and weakly rapakivi-textured feldspar.
2. Mafic enclaves, some containing feldspar phenocrysts. The enclaves are interpreted as the product of the mingling of mafic and felsic magmas.
3. Weak trachytoid texture defined by aligned feldspar phenocrysts. This is typical of the "gray granites."

NEXT: Drive back to old Highway 23. Turn right (east), drive a few hundred feet, and stop adjacent to another large quarry just north of the highway. Look north into the quarry from the edge of the road.

STOP 4-2B

Rockville Granite "Rockville Beige" quarry

Location: T. 123 N., R. 29 W., sec. 9, SE Rockville quadrangle; UTM: 395,753E/5,036,465N



Highlights: Dimension-stone quarry, weathered residuum, corestones.

Description: This quarry was opened in 1895 (?), and has since produced dimension stone for building projects all across the U.S. This quarry was converted from derrick to drive-in a few years ago. The rock is identical to that at Stop 4-2A, except that the K-feldspar is pinker in color, giving rise to the "beige" color of the rock.

Of particular interest from the highway overview at this stop are the prominent corestones present within the upper weathered portion of the rock. Here there are many glacial "erratics" that are clearly corestones that have been moved only a very short distance. The corestones have been utilized as artistic pieces in which designs are polished into the rough, chemically etched surface of the rock.

NEXT: Return to Highway 23 and go northeast toward St. Cloud. Approximately 0.9 mile after Highway 23 crosses Interstate 94, turn right on Bel Clare Drive and proceed east 0.8 mile to a "T" intersection with Stearns County Highway 137. Turn left (northeast) on 137 and drive about one mile to the well-marked entrance to Stearns County Quarry Park on the right. Just before the quarry park road on the left are large earthen berms that surround the south pit of the Martin-Marietta Aggregate quarry. Enter Quarry Park and proceed to the parking area. Walk south on the trail at the south edge of the parking lot for approximately 700 feet. Turn right onto a short switchback trail to a long northeast-oriented quarry (quarry 13 on the park map). See Figures 4.11 and 4.12 for help navigating.

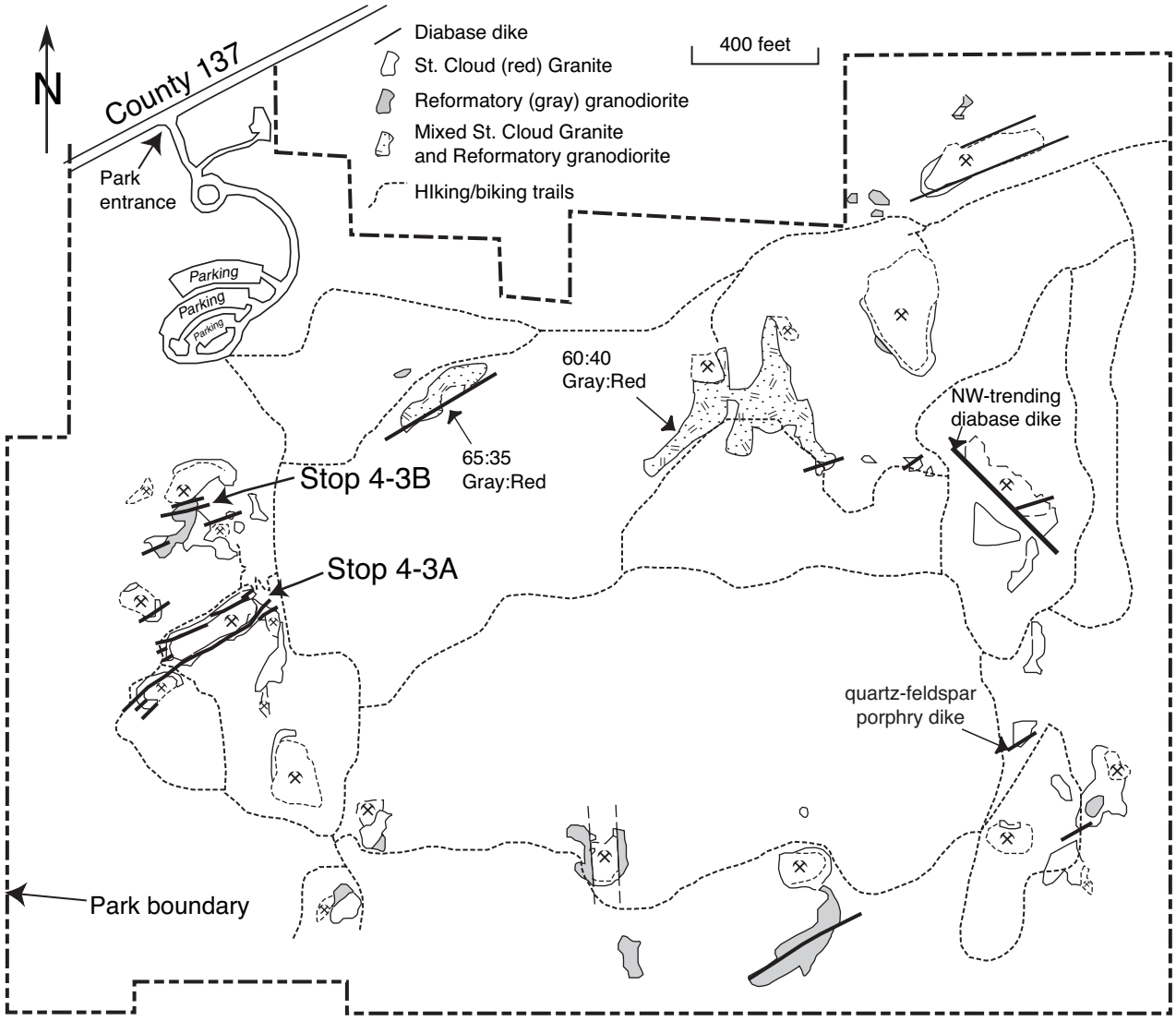


Figure 4.11. Sketch map of Stearns County Quarry Park showing general rock types, quarry locations, and trails.

STOP 4-3

This area contains several inactive quarries (Fig. 4.11) that are now all part of a unique county park. At least seven different quarry companies were once active in what is now the park area. All the operations here quarried the red St. Cloud Granite. Production started in the early 1900s, peaked at about 1940, and ceased in the mid-1950s. The stone they produced was marketed under a variety of "red" names including "North Star Red," "Mahogany Red," and "Rose Red." Most of the pits range from 60 to 100 feet in depth (Thiel and Dutton, 1935).

The park provides excellent exposures that show the relationship between the St. Cloud Granite and the Reformatory granodiorite, as well as northeast-trending diabase dikes, and one northwest-trending diabase dike (Fig. 4.11). Due to time constraints on this trip, we will only visit two quarry areas near the parking lot (Fig. 4.12).

STOP 4-3A

Stearns County Quarry Park—Reformatory granodiorite and St. Cloud Granite

Location: T. 124 N., R. 28 W., sec. 19, SE St. Cloud quadrangle; UTM 4-3A: 403,084E/5,036,465N; UTM 4-3B: 403,044E/5,042,950N

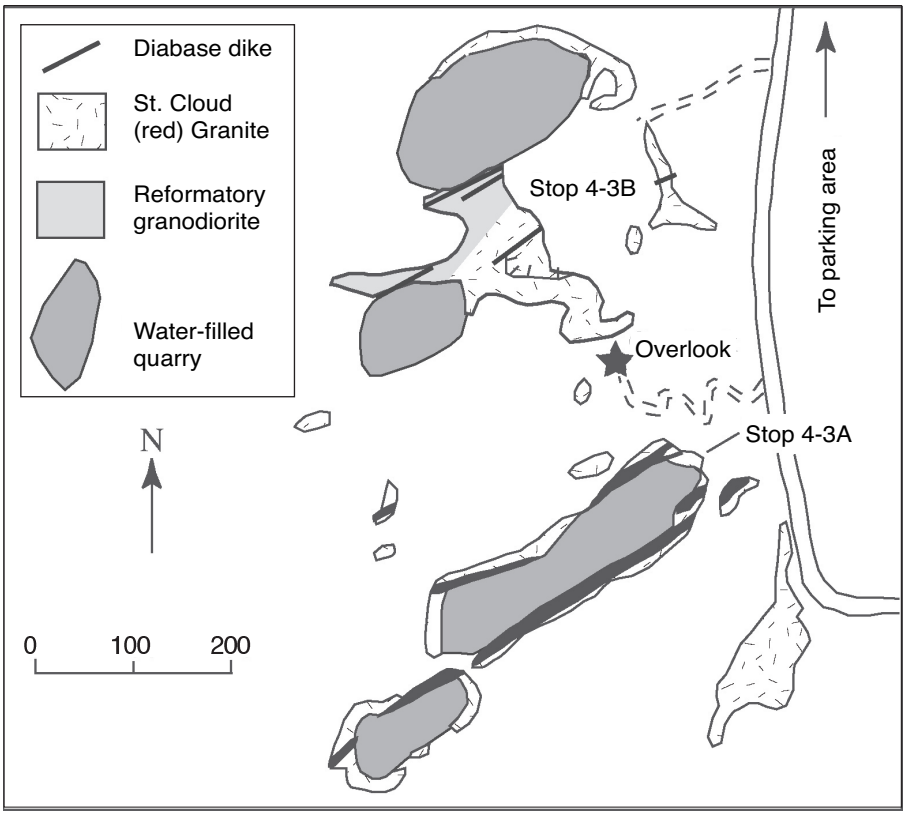
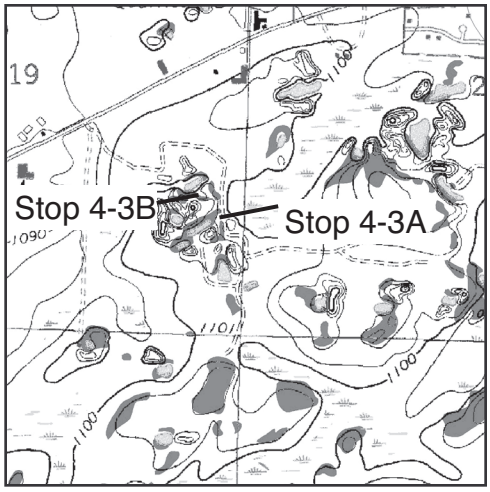


Figure 4.12. Detail of Stops 4-3A and 4-3B.



Highlights: St. Cloud Granite, Reformatory granodiorite, northeast-trending diabase dikes, quarries.

Description: This quarry contains excellent exposures of one of the many northeast-trending diabase dikes that cut the Penokean granitoid rocks in east-central Minnesota. As typical in this area, the quarry walls are edged by diabase dikes.

At the north side of the quarry (east end) is a meter-wide diabase dike that splits and rejoins along its length, with small, 4 to 10 centimeter, north–south "jogs" at the northeast end. There are no obvious cross-fractures in the granite that might have controlled these offsets, making them somewhat enigmatic. The dike margins are chilled, and small granite xenoliths and slivers in various stages of assimilation are included in the diabase. Small, partly assimilated xenoliths tend to be concentrated near the center of the dike, presumably caused by flow entrainment. Small, approximately 0.5-centimeter-diameter amygdules of fibrous actinolitic amphibole can be located with some searching.

At the south side of the quarry are two sub-parallel, northeast-trending diabase dikes that are approximately 3 meters thick—the maximum thickness observed in this dike set. The dikes are cut by thin, white, siliceous, aplitic-textured dikelets that emanate perpendicular to the dike margins and pinch out in the dike interiors. The veinlets—interpreted as partial melts of the granite wallrock—are rarely present in dikes that are less than 2.5 meters thick, presumably because the narrower dikes were not capable of heating the wallrock to temperatures sufficient for melt generation.

A small, highly polished outcrop near the east end (south side) of the pit provides a good surface to

observe the mineralogy of the granite. The St. Cloud Granite has a distinctive black and white mottling to its red color. The base red color is imparted by microcline, the black by hornblende and minor biotite, and the white by saussuritized plagioclase.

Glacial features are well displayed at the south side of the pit. These include highly polished outcrops, glacial striations, and differential scouring of the diabase/granite contact. Most of the subsurface bedrock in Minnesota is covered by variable thicknesses of weathered bedrock residuum (saprolite) that has been stripped away from bedrock topographic highs by glacial scouring. Saprolite developed on granite is typically pale gray-green. Many of the active quarries in this area have exposures of this saprolite along the margins of the pits (such as the Rockville quarry, Stop 4-2B).

An interesting hydrologic observation here is the approximately 10-foot difference in water level between this pit and the small one to the south. The two quarries are separated by little more than the width of the diabase dikes, which apparently act as a dam to ground-water movement.

NEXT: Return to the trail and walk back toward the parking lot approximately 300 feet, turn left up a small trail/road and continue west to the edge of a small quarry (Stop 4-3B).

STOP 4-3B

Three-dimensional view of diabase and quartz-epidote shears in the St. Cloud Granite.

Description: Directly west, across the rubble in the pit, is an excellent exposure of a meter-thick diabase dike that strikes northeast and dips steeply south, cutting the St. Cloud Granite. The dike was sampled for both paleomagnetic study and Ar/Ar dating. Like the dikes at Stop 4-3A, this is typical of the diabase dikes in this area, with chilled margins and scattered centimeter-scale xenoliths of granite wallrock. The granite adjacent to the dike contains numerous thin, northeast-trending, quartz- and epidote-lined, semiductile shear bands that are clearly truncated by and probably in part occupied by the diabase dike. Two other diabase dikes, up to 2 meters thick, are also visible here. One "lines" the south wall of the larger abandoned quarry to the north.

Note the variety of rock types represented in the waste rubble blocks here. Most are indigenous red St. Cloud Granite and gray Reformatory granodiorite, which is the dominant type of bedrock under the city of St. Cloud. The Reformatory granodiorite is presently quarried as dimension stone at a location just south of the city of St. Cloud, less than two miles east–southeast of here. A mixture of Reformatory gray/St. Cloud red rock is quarried and crushed for aggregate just north across the highway, by the Martin-Marietta Company. The trails in this park are made of aggregate produced from that quarry. The maroon-colored blocks here are Archean granites imported from quarries near Milbank, South Dakota. Blocks of grayish-pink, complexly banded gneiss are quarried from the Archean Morton Gneiss in the Minnesota River Valley. Several other blocks of pink granite and pale pink gneiss are from unknown locations outside Minnesota.

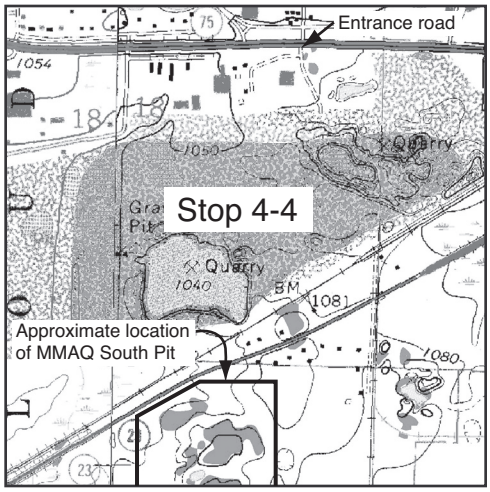
NEXT: Leave the park and turn right on County Road 137. Proceed approximately 0.7 mile to 10th Avenue. Turn left (north) on 10th Avenue and proceed approximately 0.9 mile to Stearns County Road 75/Division Street. Turn left (west) on 75, travel approximately 0.6 mile, and turn left into the Martin-Marietta Aggregates Quarry—the road is marked by a business sign.

STOP 4-4

Martin-Marietta Aggregates Quarry

Location: T. 124 N., R. 28 W., sec. 18, S 1/2

St. Cloud quadrangle; UTM Start: 403,462E/5,045,175N (entrance road to quarry)



Highlights: Crushed rock aggregates quarry, St. Cloud Granite, Reformatory granodiorite, weathering profile, diabase dikes, loose blocks of quartz-feldspar porphyry dike

Description: The Martin-Marietta rock quarry produces around 1.2 million tons per year of crushed rock. Mining and crushing operations run about 7 months per year, and stockpiled product is shipped year around. This quarry utilizes a mixture of rock

types, mainly the Reformatory granodiorite and the St. Cloud Granite, and included diabase and quartz-feldspar porphyry dikes. Unlike dimension-stone quarries, joints and fractures in the rock are not detrimental to operations. Quarry methods involve drilling and blasting, primary crushing in a gyratory crusher, and secondary crushing to various sizes. The primary use of the product is railroad ballast and road construction, as well as other uses such as breakwaters, decorative stone, driveway material, and turkey grit. The aggregate is a high-quality product with ideal physical and chemical properties to meet state and federal road construction standards.

A substantial fraction of the rock in this quarry consists of fine-grained gray granodiorite interpreted by earlier workers as inclusions of metamorphosed and metasomatically-altered country rock xenoliths in the Reformatory granodiorite. Although there are undoubtedly some small xenoliths, textural and geochemical evidence indicates that the bulk of this gray material is more likely a fine-grained border phase to the Reformatory granodiorite, possibly similar to the Watab diorite.

In proper light conditions, a cross-section is visible on the far west end of the pit that shows fresh rock transitioning upward into weathered bedrock residuum (saprolite), overlain by glacial deposits. Diabase dikes are visible on the east and west walls and possibly the floor of the pit as thin, black, vertical bands. A 3- to 4-meter-thick dike of quartz-feldspar porphyry is located directly beneath the observation deck (but not visible from above). There are some blocks of the quartz-feldspar porphyry dike lying along the trail to the observation deck. These dikes have the same orientation and occur in a similar manner to the northeast-trending diabase dikes, and are slightly earlier in timing of emplacement.

Figure 4.13 shows an outcrop/geologic map of an area to the south of here, across Highway 23, that was mapped prior to being opened as a quarry. Two abandoned (now overrun) dimension-stone quarries located on that property utilized the St. Cloud Granite, as at Stop 4-3. This figure demonstrates the complex relationships of the St. Cloud Granite/Reformatory granodiorite, in which the red St. Cloud Granite was emplaced into the gray Reformatory granodiorite as a series of mostly northeast-trending dikes and shallow-dipping sills.

At the time this guide was written it was not known whether we will have the opportunity to enter the quarry. If we can gain access into the pit, we will be able to get a first-hand look at the diversity of rock types here.

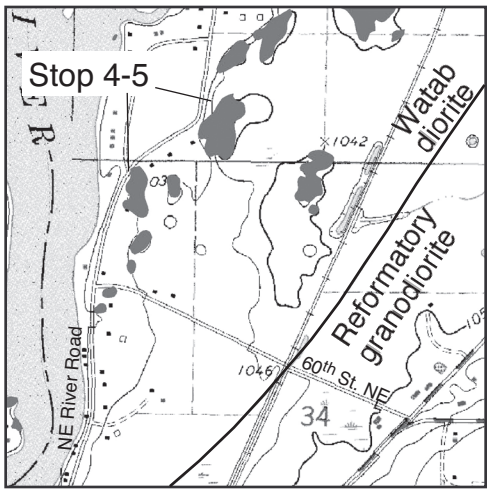
NEXT: Drive east on County Road 75 1.5 miles to the intersection with State Highway 15. Turn left (north) on 15 and stay on it to U.S. Highway 10, a distance of approximately 5 miles. Go north on Highway 10 for 2.5 miles to the 60th Street/Frost Road NW intersection and turn left (west) on Frost Road. Follow Frost Road to a "T" intersection with NE River Road. Turn right on NE River Road and drive 0.4 mile to outcrops on the right (east) side of the road.

STOP 4-5

Watab diorite

Location: T. 37 N., R. 31 W., sec. 27, SW and sec. 34, NW

Little Rock Lake quadrangle; UTM: 407,261E/ 5,057,354N



Highlights: Watab diorite and related felsic phases, comagmatic intrusive relationships

Description: The Watab diorite is generally a massive to weakly foliated, dark grayish- to pinkish-green, fine-grained, locally plagioclase-phyric, apatitic hornblende-quartz diorite to granodiorite. The diorite contains a retrograde assemblage of hornblende and biotite after earlier pyroxene, a product of either deuteric alteration or retrogressive thermal metamorphism related to emplacement of the surrounding granitoid rocks. Inclusions of texturally and mineralogically similar rocks are present throughout the Reformatory granodiorite, and textural gradations observed elsewhere indicate that the Watab diorite may be related to the Reformatory granodiorite, possibly as an early border phase.

The diorite contains local coarse-grained, somewhat appinitic felsic segregations, and is cut by irregular dikes of fine-grained, porphyritic and trachytic biotite granite that is similar in texture and

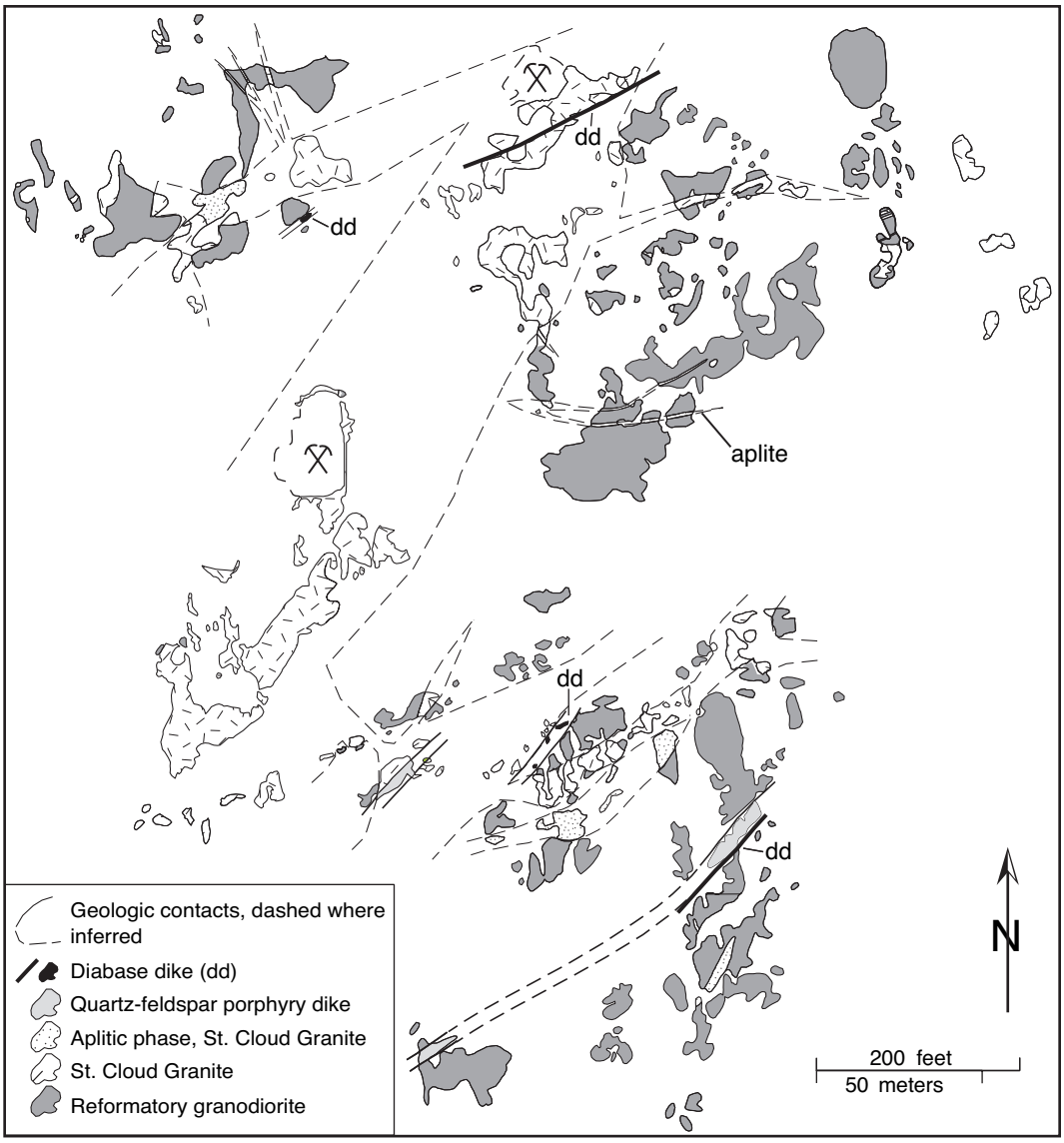


Figure 4.13. Pre-quarry geologic outcrop map of the area now occupied by the Martin-Marietta quarry south of Highway 23; modified from Boerboom and Holm (2000).

mineralogy to the more felsic phases of the diorite, possibly indicating a cogenetic relationship. The felsic dikes commonly contain rounded, irregularly-shaped xenoliths of Watab diorite (Fig. 4.14).

Local pegmatite observed in the diorite varies from boxy, northeast-trending dikes to net-vein segregations and clots with feldspar crystals that have nucleated perpendicular to the walls.

NEXT: Return east to Highway 10 by reversing route. Cross Highway 10 on 60th Street NE/Benton County Road 33. Drive approximately 0.9 mile east on 60th Street to a small abandoned quarry on the left (north) in a pasture.

STOP 4-6

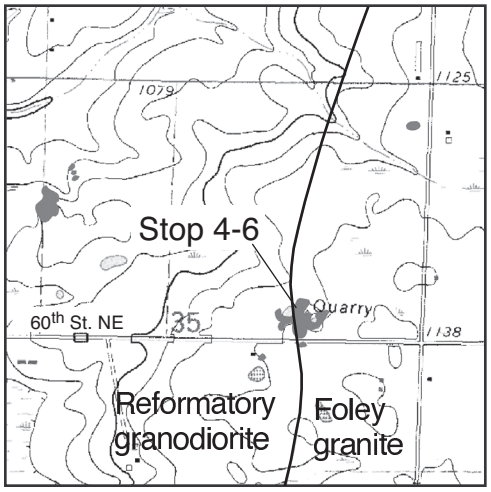
Private property! Permission must be obtained before entering!

Foley granite

Location: T. 37 N., R. 3 1 W., sec. 35, NE Little Rock Lake quadrangle; UTM: 409,646E/ 5,056,378N



Figure 4.14. Irregular felsic dike containing irregular to rounded xenoliths of Watab diorite.



Highlights: Foley granite, contact with Reformatory granodiorite, quartz-feldspar porphyry

Description: This stop shows a contact between the Foley granite and the Reformatory granodiorite. The Foley granite (1,779 ± 4 Ma) is identical in age to the St. Cloud Granite (1,779 ± 5 Ma), and the two are considered to be related. The Foley granite batholith forms a continuous mass, unlike the St. Cloud Granite, which occurs as a series of irregular dikes and sills. Geophysical models (Chandler and others, in press) imply that the western margin of the Foley pluton forms a "puddle"—a thin, lensoid-shaped pluton that is approximately 1.2 kilometers thick.

The Foley granite is a pale pink to salmon-colored, hornblende-biotite granite, as seen in the small eastern quarry at this stop. At the small western quarry the contact between the Foley granite and the Reformatory granodiorite is exposed, which has a strike and dip of N. 10° W., 62° W. The contact is marked by a 1.5-meter-wide chill zone in the Foley granite that contains trachytically aligned, tabular feldspar phenocrysts up to 1 centimeter in length.

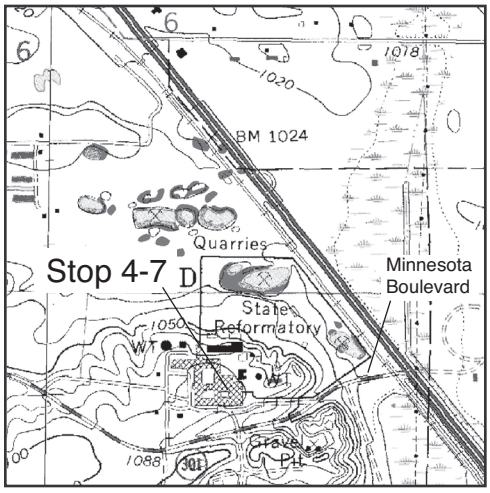
NEXT: Return west to Highway 10. Travel south on Highway 10 toward St. Cloud. To go to optional Stop 4-7, turn right (west) onto Minnesota Boulevard, which is 2.1 miles past the intersection of Highway 10 and County Road 23. The prison wall can be seen from Highway 10 without turning off. After turning west on Minnesota Boulevard, the prison can be seen immediately on the right.

STOP 4-7 (optional drive-by view)

The State Reformatory after which Reformatory granodiorite is named

Location: T. 35 N., R. 30 W., sec. 7, N 1/2

Cable quadrangle; UTM: 412,786E/5,043,720N



Highlights: Prison buildings and wall made of gray Reformatory granodiorite, architecture.

Description: This "stop" shows some architectural applications of the Reformatory granodiorite, known in the dimension-stone industry as "charcoal gray." All the stone for the prison was quarried within or just north of the prison walls.

Bowles (1918, p. 120-121) reported about the Reformatory granodiorite and is summarized here. Prison inmates started excavating rock in small quarries both inside and outside the current prison walls in 1889, producing both dimension stone and crushed rock, of which 200 to 250 carloads per year were sold to the state through at least the year 1918. Starting in about 1902, building stone was quarried for use in constructing the reformatory buildings and walls (Fig. 4.15). The wall, which is 4 feet thick at the base, 2.5 feet thick at the top, and 22.5 feet high, was over three-quarters finished by 1913. The total finished length is approximately 1 mile (Thiel and Dutton, 1935). The octagonal water tower, 30 feet across at the base and 115 feet high, was completed in 1912. Construction of the buildings was apparently still in progress as late as 1935, when Thiel and Dutton (1935) reported that all the stone being produced from the quarries was being used for construction of the buildings inside the wall.

NEXT: To return to the Twin Cities, go back to Highway 10 and continue east to the town of Clear Lake. At Clear Lake turn right (south) at the stoplight and continue to Interstate 94; follow 94 east to Minneapolis.

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Figure 4.15. Photograph of a quarry inside prison walls at the time of active quarrying. From Bowles (1918).

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FIELD TRIP 5

Wednesday, May 18

SINKHOLE ANATOMY 101

Leaders

Dr. E. Calvin Alexander, Jr., University of Minnesota
Howard C. Hobbs and Robert G. Tipping, Minnesota Geological Survey

INTRODUCTION

There are about 12,000 mapped karst sinkholes in Minnesota, most of which are in Fillmore County (Gao and others, 2002). This may represent half of the sinkholes currently visible (Wittuhn and Alexander, 1995). Fresh sinkholes open or reopen at a rate of 1 to 2 percent of the total per year (Dalgleish and Alexander, 1984). Most sinkholes are filled—slowly by natural process or rapidly by humans. Karst processes have been episodically active on these rocks essentially since their deposition. There are many more filled sinkholes than there are currently visible sinkholes. Many of the sinkholes show evidence of multiple openings and fillings.

Sinkhole excavation provides an opportunity to view sediments over a broad age range. Furthermore, excavation leads to insights into the depositional history of the area in addition to identifying pathways of surface drainage to bedrock and possible access to cave passages. Our host, John Ackerman, will provide the "Cave Finder" track hoe and the sites. Bring your sample bags, sampling equipment, cameras, field books, and any other paraphernalia you use. The goal will be to trench through a sinkhole or two as deeply as the track hoe can reach and record and sample whatever materials and structures we find. Depending on what we find, we may decide to compile a joint report on our day.

THE SITE—BACKGROUND AND PREVIOUS EXCAVATIONS

The sinkhole excavation will take place at the Karst Preserve of John Ackerman near Spring Valley in Fillmore County, Minnesota (Fig. 5.1). Mr. Ackerman has been a caver since he was young. In 1987, along with fellow Minnesota Speleological Society member Dave Gerboth, he began to actively pursue access to a former show cave, known as Spring Valley Caverns. In 1990, Mr. Ackerman purchased approximately 180 acres from the landowner, along with underground protective cave rights under the remaining unsold farmland. Explorations since that time have expanded the mapped cave passage from 1 to 5.5 miles (Ackerman, 2001).

Based on his experience of the past 15 years at the site, Mr. Ackerman has found that sinkholes in this setting typically identify the intersection of several passageways; by excavating a sinkhole, he can create a new access point to a cave that may or may not be connected to previously mapped passageways. Mr. Ackerman has a great appreciation for karst processes and the geologic history of his site, along with an awareness of the susceptibility of caves and aquifers to activities at the land surface. He has actively engaged the academic community in his work, and several workshops for regulatory personnel have been held on the Karst Preserve. As a result, an excavation at this site provides an ideal opportunity for science and education in addition to cave access.

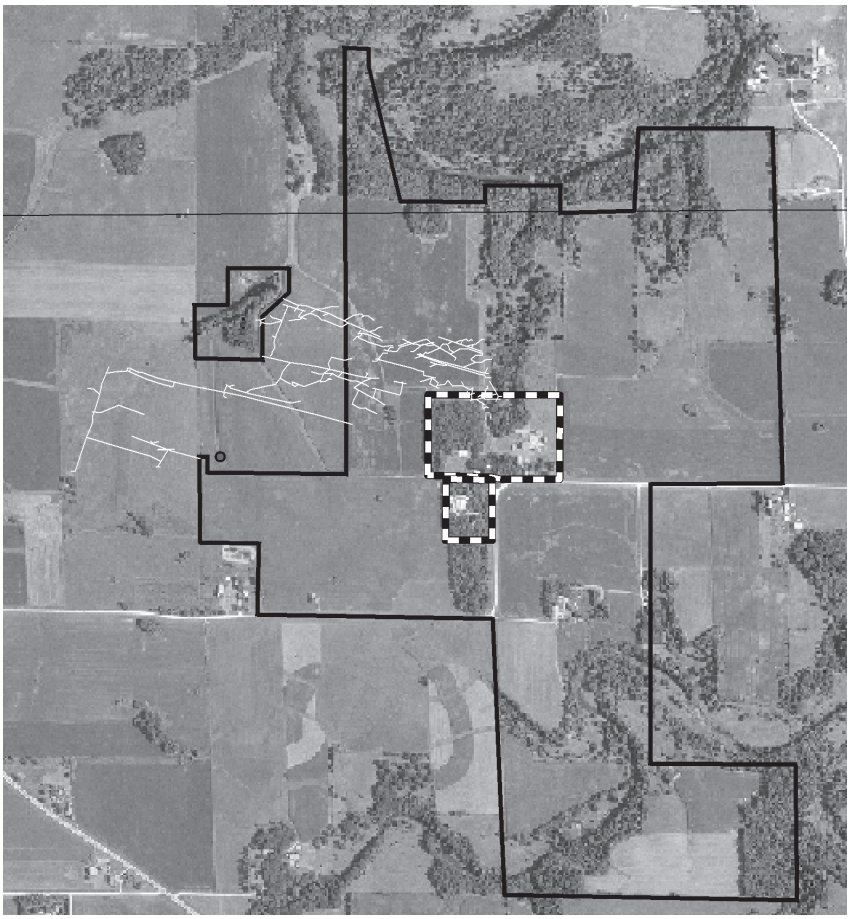
Potential excavation sites are identified by sinkhole locations. These sinkholes are commonly actively removing sediment, forming collapses a meter or more in diameter (Fig. 5.2A). Excavation is performed with a backhoe, commonly to the full extent of the backhoe arm (Fig. 5.2C). The excavation of one sinkhole on the Karst Preserve, MN23:D0074, revealed the following sequence of sediments from the land surface to the competent bedrock surface: "A" horizon top soil, loess, laminated blue clay, boulder pavement, glacial till, and weathered bedrock (Fig. 5.3). In other excavations we have also seen human trash and several different colors of sand bodies.

SUMMARY OF REGIONAL GEOLOGY

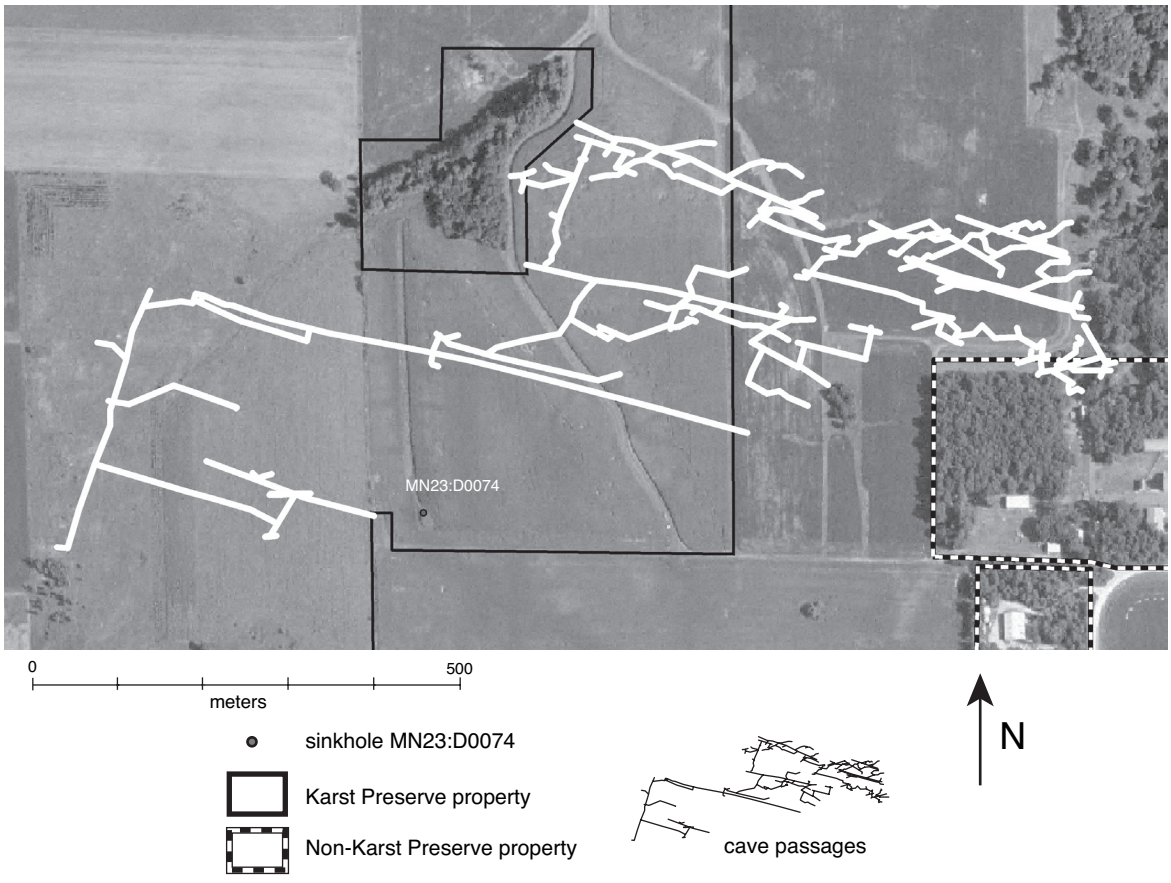
Unconsolidated deposits (Tertiary through Holocene)

During the Tertiary period, Fillmore County was subjected to a long period of subarial weathering.

A.



B.


MN23:D0074
0
500
meters
sinkhole MN23:D0074
Karst Preserve property
Non-Karst Preserve property
cave passages
N

The resulting karst landscape was largely removed by the Pleistocene continental glaciers (for example Hedges and Alexander, 1984 and references therein; Williams, 1997). Remnants of the Tertiary record may include small, isolated patches of sands and clays, polished quartz pebbles (Ostrander gravels), terra rosa sinkhole and joint fillings, and iron deposits, although none have been confidently dated as Tertiary in age. These occur between strata known to be Paleozoic or Cretaceous in age and Pleistocene deposits related to glaciation. During the Pleistocene, continental glaciers covered Fillmore County at least twice and probably several times. As the ice sheets advanced across the county they bulldozed off much of the epikarst produced during subarial weathering and filled the active sinkholes with debris (for example Anthony and Granger, 2003). As the glaciers melted back they left behind both till and eolian loess and sand deposits. The most recent glacial advance across Fillmore County is undated but is interpreted to be pre-Illinoian in age, although Illinoian and Wisconsinan loess and sand deposits may be present. There has been a substantial amount of solution weathering of the top several feet of limestone below the last glaciated surface. The result is a tan to brown material that looks like the unaltered bedrock but that can be excavated with a backhoe.

Finally, during the Holocene Epoch human impacts have begun. No pre-European settlement sites have yet been found in Minnesota sinkholes, but such artifacts probably exist. European settlers in the 19th and 20th centuries viewed sinkholes as an impediment to farming and filled them with whatever unwanted materials they had. This is the history through which we will be excavating.

Description of bedrock in the vicinity of Spring Valley Caverns

Mapped cave passages of Spring Valley Caverns span the Upper Ordovician Dubuque Formation and underlying Stewartville Formation (Figs. 5.4, 5.5). These rocks are near the axis of a thick section of Paleozoic rocks deposited in a broad topographical lowland called the Hollandale Embayment (Mossler, 1998). Downwarping of the embayment, beginning during Ordovician time, accommodated up to 1,500 feet of Paleozoic rocks. These rocks are relatively flat lying, and are for the most part structurally undeformed. Cambrian strata are dominantly siliciclastic, whereas Ordovician and Devonian strata are dominantly carbonate and shale. There have been three major sea incursions in Minnesota, corresponding to the Sauk, (Late Cambrian to Early Ordovician), Tippicanoe (Late Ordovician), and Kaskaskia (Middle Devonian) sequences of Sloss (1963). Extensive weathering prior to each major incursion contributed to brecciation, fracturing, and karstification. Rock and hydrogeologic descriptions (Mossler, 1998; Runkel and others, 2003) of the Maquoketa and Dubuque Formations, along with the overlying Middle Devonian Spillville Formation and underlying Upper Ordovician Stewartville Formation of the Galena Group are included below.

Spillville Formation (Middle Devonian)

The Spillville Formation is a finely crystalline, light brown to grayish-orange dolostone. It has abundant fossil-moldic porosity, with dogtooth calcite spar in larger vugs. The Spillville Formation is very thick to thin bedded. The basal few feet are sandy to silty and may contain minor amounts of shale or shaly limestone. The middle to upper parts of the formation contain thin-bedded, light olive-gray, dense dolomitic limestone. Unit thickness is as much as 78 feet in eastern Mower County.

Secondary porosity and permeability in the form of fractures dominate the upper part of the unit, and fossil-moldic, vuggy porosity is prominent in the lower part. A basal shaly dolostone of low permeability may function as a confining unit locally, as studies suggest in northern Iowa (Witzke and Bunker, 1985). Gamma logs indicate that this basal shaly layer is discontinuous.

Maquoketa Formation (Upper Ordovician)

The Maquoketa Formation is a mostly silty, fine to medium crystalline, light olive-gray, slightly fossiliferous limestone that is interbedded with gray to brownish-gray, unfossiliferous and shaly dolostone. It includes scattered beds of chocolate brown to light olive-gray shale and very light gray dense limestone. The Maquoketa Formation includes scattered light

Figure 5.1. Maps showing the current extent of John Ackerman's Cave Farm property, along with the approximate location of the cave passages of Spring Valley Caverns.

A. The Cave Farm sits between surface drainage ways to the north and south.

B. Blow-up of property in the area of mapped cave passages of Spring Valley Caverns. Location of a previously excavated sinkhole MN23:D0074 (Figs. 5.2, 5.3) is shown.

A.



B.



C.



Figure 5.2. Sequence of photos showing the excavation of sinkhole MN23:D0074.

A. Natural collapse prior to excavation. Thick loess and the current drainage hole are shown.

B. Measurement and preliminary survey; backhoe to be used is visible in the background.

C. Sinkhole excavation nearing the bedrock surface. Person and backhoe shovel for scale. Location of sinkhole is shown in Figure 5.1. Cross section is shown in Figure 5.3. Photos by Dr. Toby Dogwiler, Winona State University, Department of Geoscience.

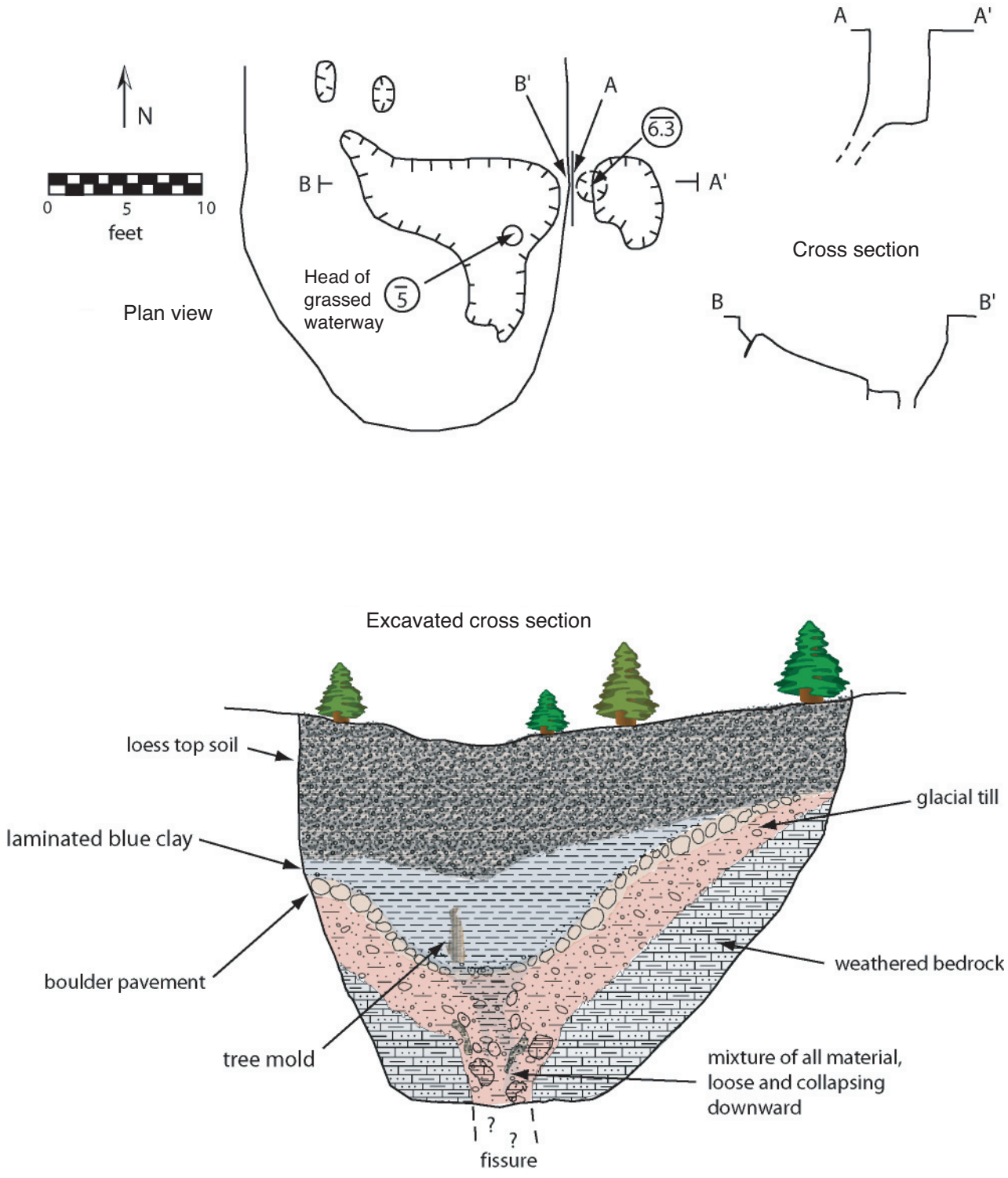


Figure 5.3. Cross section and plan view of sinkhole MN23:D0074. The sequence at the site includes "A" horizon top soil, thick loess, laminated blue clay, boulder pavement, till, and weathered bedrock. A black, funnel-shaped column of "A" horizon material from the land surface to the bedrock fissure is commonly found, and can help guide the sinkhole excavation.

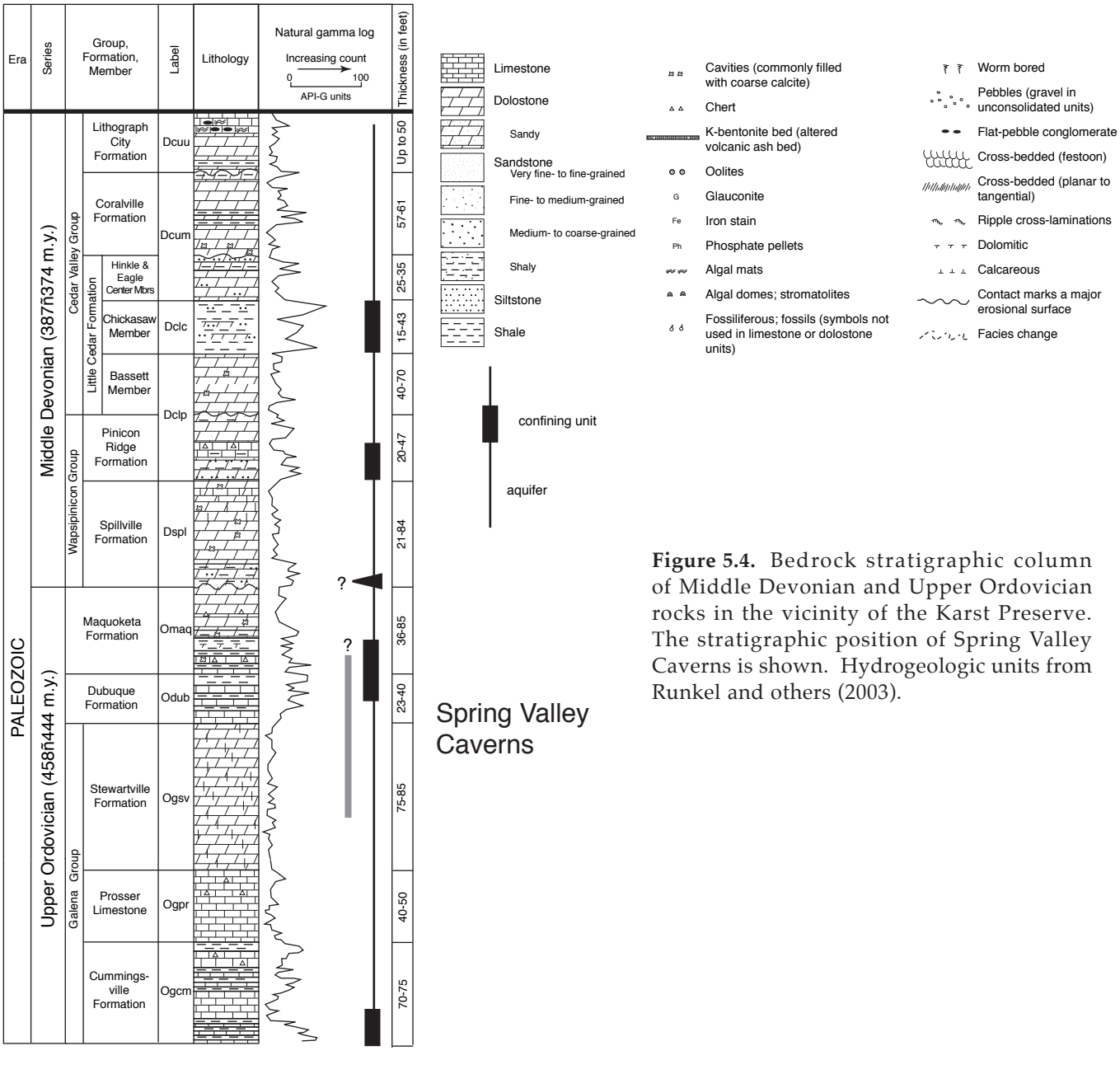


Figure 5.4. Bedrock stratigraphic column of Middle Devonian and Upper Ordovician rocks in the vicinity of the Karst Preserve. The stratigraphic position of Spring Valley Caverns is shown. Hydrogeologic units from Runkel and others (2003).

olive-gray chert nodules, and vugs that are lined with drusy quartz. Principal fossils are crinoids and brachiopods. Unit thickness is 65 to 85 feet.

The dolostone in the upper part may have enhanced permeability owing to fracturing and brecciation that are related to the development of the Middle Devonian unconformity on Ordovician Rocks (Witzke and Bunker, 1985). The lower part of the Maquoketa Formation contains numerous, thin shale beds similar to those that occur in the underlying Dubuque Formation. Hydraulically, the lower part of the Maquoketa and Dubuque Formations together can act as a confining unit where not breached by vertical fractures and/or solution cavities.

Dubuque Formation (Upper Ordovician)

The Dubuque Formation is a light olive-gray limestone with minor dolostone. The limestone is interbedded with thin to medium beds of light olive-gray to light gray calcareous shale. The rock is fossiliferous, with abundant crinoid debris and brachiopods. Formation thickness is up to 35 feet. The shaly upper part is grouped with the lower Maquoketa Formation as a confining unit (Runkel and others, 2003).

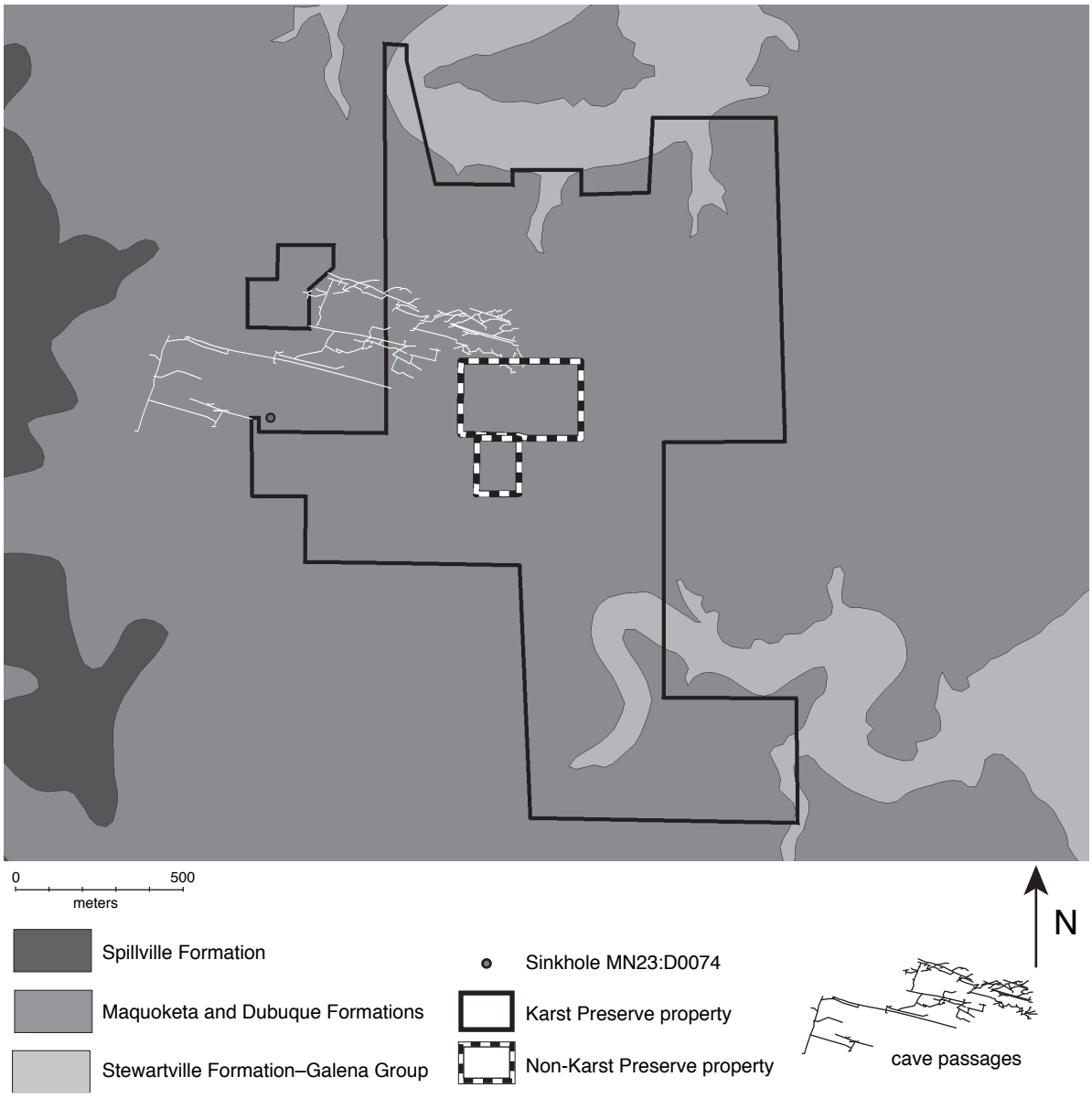


Figure 5.5. Bedrock geologic map of the Karst Preserve property, along with approximate location of the cave passages. Downcutting of the bedrock surface to the Upper Ordovician Stewartville Formation–Galena Group in surface drainageways to the north and south is shown. The Karst Preserve is just east of the Middle Devonian Spillville Formation extent.

Stewartville Formation of the Galena Group (Upper Ordovician)

The Stewartville Formation is a mottled, yellowish-gray and pale olive-gray dense limestone and finely crystalline dolostone. There are few fossils compared to the overlying Dubuque Formation. Unit thickness is 75 to 85 feet.

ACKNOWLEDGMENTS

Special thanks are due to our host, John Ackerman. John is not only hosting us on his Karst Preserve, but also providing and operating the Cave Finder track hoe. The Minnesota Karst Preserve presents a unique opportunity for the discovery, exploration and scientific study of Minnesota's karst features.

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FIELD TRIP 6

Wednesday, May 18

LATE ORDOVICIAN STRATIGRAPHY AND PALEONTOLOGY OF THE TWIN CITIES BASIN

FIELD TRIP 7

Saturday, May 21

LATE ORDOVICIAN LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY OF THE SOUTHERN MARGIN OF THE TWIN CITIES BASIN

Leaders

Robert E. Sloan, Professor Emeritus, University of Minnesota
Michael D. Middleton, University of Wisconsin-River Falls
Gerald F. Webers, Macalester College

The following information describes the tectonics, biostratigraphy, and lithostratigraphy of the Late Ordovician period in Minnesota and describes the history of the fossils you will be collecting today. This description was composed by Robert Sloan. More information on these rocks and fossils is available in Sloan (2005).

INTRODUCTION

The Deicke K-bentonite, by correlation with the T-3 ash bed of Tennessee, is 454.2 Ma in age (Kunk and Sutter, 1984; Kolata and others, 1986; Huff and Kolata, 1990; Huff and others, 1996). The Deicke K-bentonite is a major provincial extinction event. At the species level, the conodont extinction was 10 percent, the brachiopod extinction was 39 percent, the gastropod extinction was 80 percent, trilobite extinction was 90 percent, ostracode extinction was 50 percent, and the echinoderm extinction was 100 percent. This appears to be the Turinian–Chatfieldian stage boundary. The level of generic extinction is much lower. A second major extinction at a level of about 90 percent, took place during the lower part of the Stewartville Dolomite (Sinsinawa strata) during about 0.8 m.y., as a result of shoaling from 50 meters depth to about 5 meters depth.

The duration of the upper Mississippi River valley Late Ordovician (Chatfieldian to Cincinnatian stages) column is 9 m.y. The absolute standing crop of plankton of these seas ranged from 0.1 to 12.5 kilograms per square meter, based on the quantitative plankton kill beneath the Deicke ash fall and the conodont abundance index measured for the entire section. This index (conodonts per 100 grams of sediment) in turn correlates precisely with the inferred height of the Transcontinental Arch. Depths of deposition of these rocks varied from 2 to 50 meters (6 to 160 feet); bottom slopes were negligible throughout the region. Abundant clams, bryozoa, and *Isotelus* represent the shallow end, *Ischadites* and *Dolichoharpes* the deeper end of this spectrum. Rate of deposition averaged 17.6 millimeters per 1,000 years.

Differences between the Illinois and Minnesota classifications of Mohawkian and Cincinnatian rocks are due to tectonic reasons dating from the Proterozoic era. The Transcontinental Arch in Minnesota is the 1.6 b.y. Penokean Mountains of central Minnesota, also known as the Great Lakes Tectonic Zone, a suture between two Archean terranes. Southeastern Minnesota Paleozoic rocks were continuously deposited in the 1.0 b.y. Keweenawan rift valley/Midcontinent Geophysical Anomaly that persistently subsided throughout the Late Ordovician. Rocks in Illinois were deposited on the flanks of the Wisconsin Arch and have minor unconformities due to eustatic sea level fluctuations and local uplifts that coincide in time with pulses of uplift of the Great Lakes Tectonic Zone. Detailed bed tracing over the 540 kilometers between St. Paul, Minnesota and LaSalle, Illinois permits precise extension of local zones based on conodonts, ostracods, bryozoa,

brachiopods, mollusks, echinoderms, and trilobites over the entire region. By virtue of Sweet's (1984) composite standard section (CSS), this zonal synthesis can be can be extended to the rest of the continent with a high degree of precision. Croixan sandstones stripped off the Great Lakes Tectonic Zone were the source of the Turinian stage St. Peter Sandstone and basal Winnipeg Formation. The Decorah Shale and Cummingsville Formation were formed from sediments eroded from the underlying Proterozoic shales and were deposited 40 percent faster in the Twin Cities than at the Iowa/Minnesota border. A 300-meter uplift of the Great Lakes Tectonic Zone at the beginning of the Chatfieldian stage produced the Decorah Shale–Cummingsville Formation wedge in Iowa and Minnesota simultaneously with a 0.8 m.y. unconformity in northern Illinois, and the upper part of the Winnipeg Formation in the Williston Basin. The Prosser Limestone to Fort Atkinson Formation correlate precisely with the Red River Dolomite of the Williston Basin. The Transcontinental Arch was completely submerged during the deposition of the Prosser Limestone and Stewartville Formation, but was again uplifted to form the source of the siliciclastics of the Dubuque Formation and at least some of those of the Maquoketa Formation. During this uplift, the 0.09 m.y. unconformity beneath the depauperate zone was produced in Illinois and Iowa.

LATE ORDOVICIAN ROCK UNITS OF MINNESOTA AND ILLINOIS

Until recently, the St. Peter Sandstone through the Galena Group rocks were called Middle Ordovician in age. As a result of a 1997 decision by the Ordovician subcommittee of the International Union of Geological Sciences, these rocks are now called the Late Ordovician, with the base of the Late Ordovician placed at the middle of the Chazyan stage of the Whiterockian Series. They last from 458 million years ago to the time of lowered sea level due to the latest Ordovician glaciation in what is now North Africa, at about 445 million years ago.

The Illinois Galena Group members and Platteville Group formations are traceable time/rock units bounded by traceable beds, and recognizable over large areas. However, due to facies changes, they are not the first-order mapping units in Minnesota. The Late Ordovician mappable units in Minnesota are the St. Peter Sandstone, Glenwood Formation, Platteville Formation, Decorah Shale, Cummingsville shaly Limestone, Prosser Limestone, Stewartville Formation, Dubuque shaly limestone, and Maquoketa shaly dolomite. Figure 6.1 shows the names and relationships of upper Mississippi River valley Late Ordovician rocks.

There are substantial reasons for the differences between the Illinois and Minnesota classifications. In Illinois, the Platteville Formation and Galena Group form a monotonous sequence of carbonates 100 meters (330 feet) thick. Determination of stratigraphic position in this sequence requires the definition of many time/rock units differentiated on the basis of key horizons of ash beds, variations in clastic content, hardgrounds, small disconformities, and a few distinctive fossils. In Minnesota, on the other hand, the proximity of the Transcontinental Arch introduced more siliciclastic sediment to the carbonate sequence. The proportion of clastics to carbonate varies much more in the 200 kilometers (125 miles) from St. Paul to Iowa, than from northern Iowa to southern Illinois (Fig. 6.1). In Minnesota, rocks were deposited in a tectonic basin; no unconformities exist in the Minnesota sequence, and the thickest pure carbonate sequence is only 40 meters (130 feet) thick. The rock units are alternately defined as carbonate and siliciclastic and are very obvious.

In Minnesota, the rock-unit boundaries associated with the Decorah Shale wedge are diachronous, and for practical purposes we need to use strict rock classifications. Such reasons for geologic mapping as highway aggregate, agricultural limestone, and water-well drilling require that rock names be restricted to a single lithic type as much as possible. The Illinois members can be traced into Minnesota as time/rock units, but do not always remain in the same formation. For example, the Rivoli Member is in the Prosser Limestone at the town of Cummingsville, but due to a facies change involving the increase of clastics, is at the top of the Cummingsville Formation at Cannon Falls, 75 kilometers (50 miles) to the northwest, but it is still recognizable by bed tracing. I suggest that if someone wishes to use a name outside of the region in which it is defined, quotation marks should be used. The Dunleith Formation of Iowa and Illinois is not defined in Minnesota, but its equivalents can be recognized in three different formations: the Decorah Shale, Cummingsville shaly Limestone, and the Prosser Limestone. If the name Dunleith must be used in Minnesota, it should be surrounded by quotation marks. Similarly, the Cummingsville Formation is not defined in Iowa, though it can be recognized.

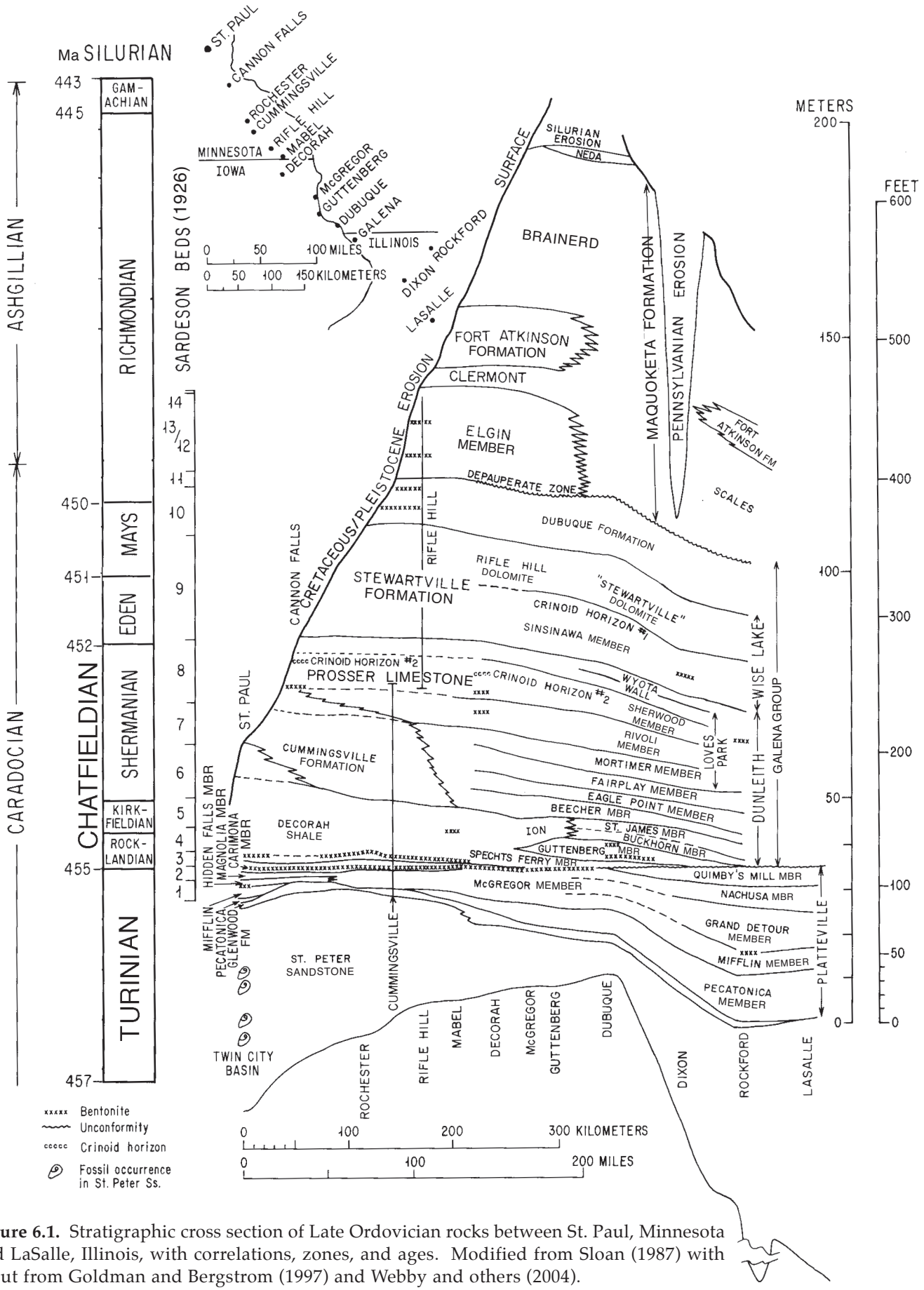


Figure 6.1. Stratigraphic cross section of Late Ordovician rocks between St. Paul, Minnesota and LaSalle, Illinois, with correlations, zones, and ages. Modified from Sloan (1987) with input from Goldman and Bergstrom (1997) and Webby and others (2004).

DETAILED BED TRACING IN THE LATE ORDOVICIAN OF THE UPPER MISSISSIPPI RIVER VALLEY

In Minnesota, detailed bed tracing over long distances began with the tracing of what are now the Deicke and Millbrig K-bentonites by Sardeson (1926). Weiss and Bell (1956) traced these in detail and added the tracing of individual "corrosion zones," later called "corrosion surfaces," still later recognized as hardgrounds. Sloan (1972) added the tracing of individual clay partings and coarse-grained calcarenites, based on 75 detailed measured sections at 5 kilometers (3 miles) spacing. His students added insoluble residue curves and conodont abundance logs. In Illinois, Templeton and Willman (1963) traced many features of this type but came to some erroneous conclusions that were corrected in Willman and Kolata (1978) because sections were too far apart. Levorson and Gerk (1972) and Levorson and others (1979, 1987) traced the Templeton and Willman units into Minnesota.

BIOZONATION

Sardeson (1927) defined 14 beds (Fig. 6.1) that were Oppel zones and were useful for their time. He continued to use these beds as collecting intervals throughout his career. However, bulk collection of fossils from the Decorah Shale, especially in St. Paul, and of acid-insoluble fossils, especially conodonts, from the carbonate units, offers the possibility of further refinement. The durations of species differ in different parts of the section, as can be seen from the durations of the Sardeson beds. Early in the Caradoc transgression during deposition of the Platteville Formation and Decorah Shale, zonal durations are short. Later, during the deposition of the Prosser Limestone and its equivalent the Dunleith Formation, durations increase. Most species transitions show punctuation, such as extinction and replacement, but some show gradualism, phyletic speciation, or transformation. Webers' (1966) and Rice's (1987) studies showed that, as could be expected, not all species in a Sardeson bed began or ended at the same time.

THE TWIN CITIES BASIN

The Twin Cities basin is a local basin on the Hollandale embayment and the Keweenawan rift valley. It is bounded on the east by the Hudson–Afton horst, and on the southwest by the Belle Plaine fault. That fault is a Keweenawan transform fault that displaces the axis of the rift to the east of Cannon Falls. The center of the basin is located on the Mississippi River across from the University of Minnesota Twin Cities campus under Fairview Riverside Hospital, about 2.3 miles northwest of Stop 6-1 of Field Trip 6. The Platteville Formation limestone is thickest there and thins in all directions toward the margin of the basin. There is a 150-foot sill bordering the southeast side of the basin, an extension of the Hudson–Afton horst. Cannon Falls (Stop 7-1 of Field Trip 7) is located on this sill. It is not surprising that the Millbrig K-bentonite is washed out at Cannon Falls and nearby localities on this sill. All formation thicknesses, sedimentary facies, and biofacies changes show that all these structures were continuously growing during deposition of the Cambrian and Ordovician periods. Wave erosion on the sea floor would periodically wash out sediment over structural highs, in this case the Millbrig K-bentonite. Figure 6.2 is a structural contour map on the Cambrian–Ordovician contact, all the relief on this map is post-Cambrian. This suggests that the 6-mile thickness of 1.1 b.y. Keweenawan basalts in the rift valley were still cooling and subsiding as late as the Late Ordovician, if not later.

VOLCANIC ASHES

The present southeast coast of the Ordovician North American Plate was the site of a subduction zone and much volcanism during Late Ordovician time. Thirteen of these eruptions were large enough to leave significant ash beds in the upper Mississippi River valley. They occur in three clusters. The first is the Rocklandian–early Kirkfieldian substage cluster of the Deicke, Millbrig, Elkport, and Dickeyville K-bentonites, in less than a million years. The next cluster of the Calmar, Conover, Nasset, Haldane, and Dygerts K-bentonites occurs in the late Shermanian Rivoli and Sherwood Members of the Dunleith Formation and Prosser Limestone, and in the basal Edenian Sinsinawa Member of the Stewartville and Wise Lake Formations within 1.5 million years, but separated from the earlier cluster by 2 million years. The third cluster of four unnamed ashes occurs in the late Edenian Dubuque Formation and Elgin Member of the Maquoketa Formation in Minnesota. The lower two, at least, are feldspathized shales (Weiss, 1957). A few other ashes are known from isolated localities. This cluster needs much more study than it has received.

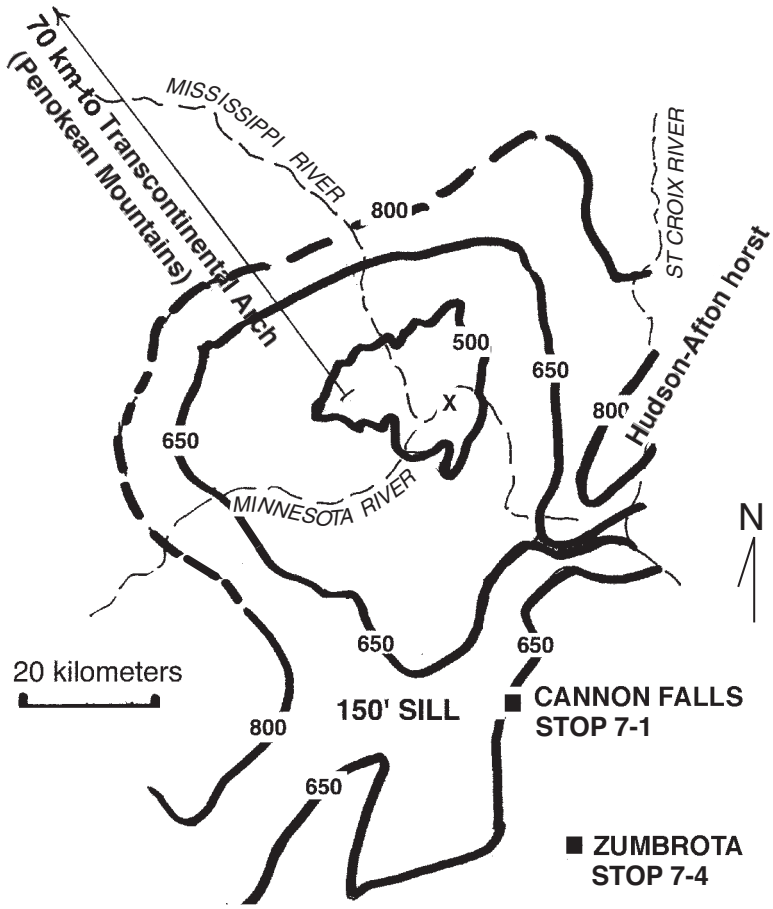


Figure 6.2. Structural contour map of the Twin Cities basin, based on the Jordan Sandstone–Oneota Dolomite contact (approximately the Cambrian–Ordovician contact). Contour interval is 150 feet, the locality of Lilydale Park is marked by an X, Cannon Falls and Zumbrota stops (Stops 7-1 and 7-4, respectively) are marked by squares.

THE DEICKE K-BENTONITE, THE ASSOCIATED EXTINCTION, AND THE CHATFIELDIAN RECOVERY

Work by Sardeson (1926), DeMott (1987), and Rice (1987) defined a major extinction event at the Deicke K-bentonite (Fig. 6.3). This ash bed represents, as closely as I can tell, the Turinian–Chatfieldian (old Blackriveran–Trentonian) stage boundary. The total volume of the ash bed is greater than 1,000 cubic kilometers (one cubic township; Huff and others, 1996). This is more than twice the volume of the 1915 Tambora eruption (the largest historic eruption), 10 times larger than that of Mt. Mazama, which produced Crater Lake, and 400 times larger than the 1980 Mt. St. Helens eruption.

Sloan (1987, 1988, 1992) has demonstrated that all organisms in the upper Mississippi River valley were killed by the 1,000 cubic kilometer Deicke ash fall, and that the replacing fauna is significantly different. The Deicke extinction appears to have extended to the limits of the Transcontinental Arch, composed of a narrow band of old Precambrian mountains reaching from the Adirondacks of New York, swinging north of the Great Lakes, and southwest to Colorado (Fig. 6.3). The fauna immediately beneath the Deicke ash fall is a typical Turinian fauna, the fauna immediately above the Deicke ash fall is the typical Chatfieldian species. The extinction is 82 percent (215 out of 262 species) at the species level, and 36 percent at the generic level. It is the largest single North American extinction event between the end of the Sunwaptan stage of the Cambrian and the end of the Ordovician. In addition, there are 17 "Lazarus" species that became locally extinct only to return in 0.5 to 2 million years. Analysis of the Deicke ash fall by Dokken (1987) and observations by Sloan show no animals dug out from under the ash bed.

Two of 21 natural or residual form species of conodonts (*Scyphiodus primus* and *Polyplacognathus ramosa*) go extinct at this horizon (Webers, 1966) for a 10 percent extinction. Of 18 brachiopods, 7 go extinct (*Pionodema conradi, Campylorthis deflecta, Strophomena plattinensis, Trigrammaria winchelli, Pseudolingula eva,* and *Orbiculoidea lamellosa*; see also Rice, 1987) for an extinction rate of 39 percent; DeMott's (1987) study

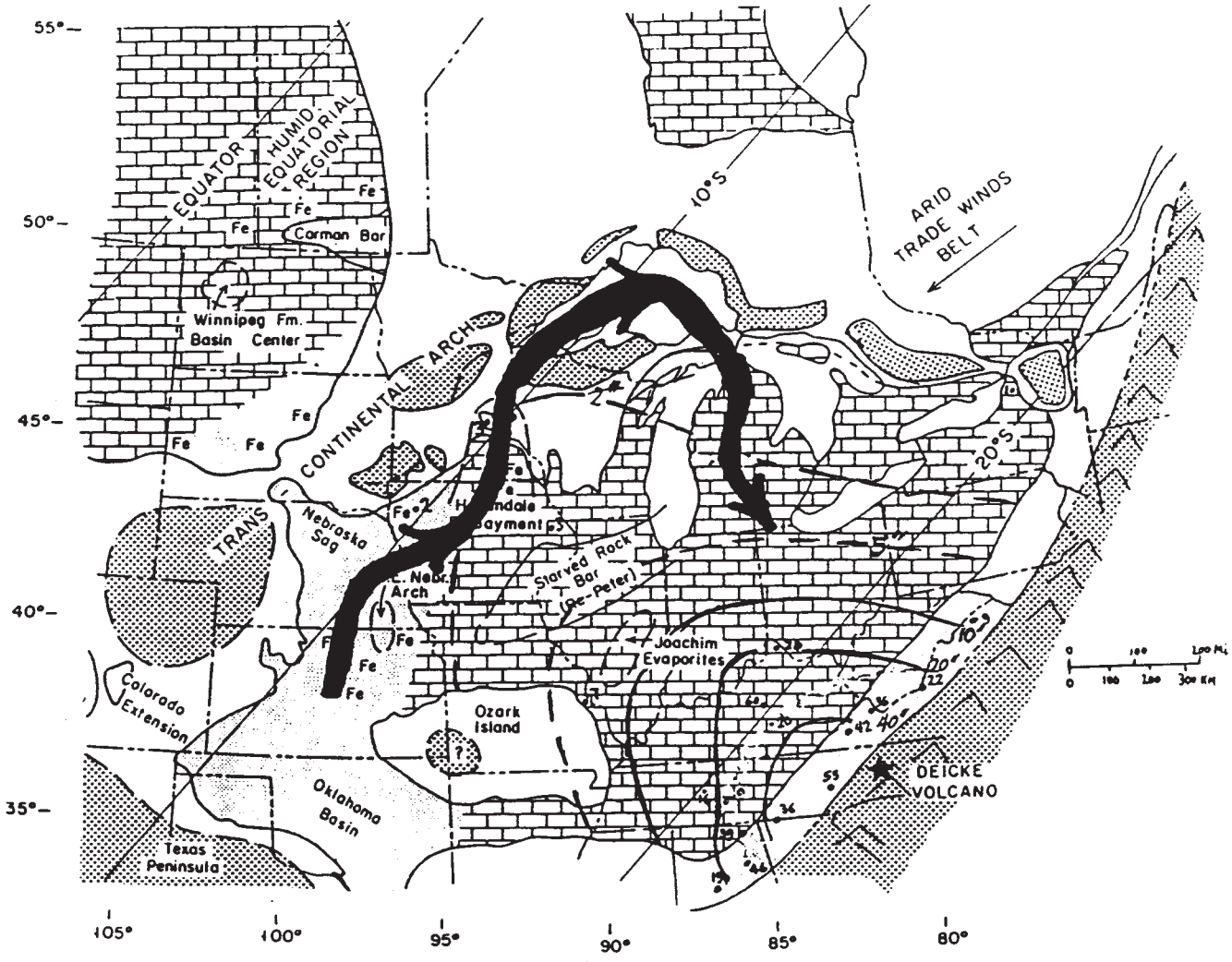


Figure 6.3. Late Ordovician paleogeography of the eastern half of the United States and Canada showing the Transcontinental Arch. Heavy stipple is persistent highlands, light stipple is siliciclastic dominated. Heavy arrow is the axis of the 1.1 Gy Keweenawan Rift, suspected route of post Deicke event repopulation. Isopach contours in inches of post-compaction thickness of the Deicke K-bentonite, star is the suspected location of the Deicke volcano. Modified from Witzke (1980) and Sloan (1987).

of trilobites showed 9 of 10 species go extinct at this horizon for an extinction rate of 90 percent. Among gastropods, 48 of 60 species go extinct at this horizon, for an extinction rate of 80 percent (Sloan, 1987). All the echinoderm species in the Platteville Formation terminate at the Deicke K-bentonite (Kolata and others, 1987). It is possible that there are multiple extinctions near this horizon due to the cluster of four major Rocklandian ashes (Kolata and others, 1986), but the Deicke ash fall is clearly the most important: in more than 100 exposures, no evidence of a benthic organism digging out of this ash bed has been observed. No other single horizon in this sequence of rocks has as many extinctions as this event.

Other species invaded the depopulated area, but not all at the same time. *Isotelus gigas* and *Eomonorachus intermedius* were the pioneer invading trilobites. These are the only Chatfieldian trilobites with planktonic larvae. The rest of the Guttenberg trilobites came in roughly a quarter million years later. Similarly some brachiopods took longer to repopulate the area than others; eight species present in Sardeson's (1926) beds 2 and 4 or 5 are totally missing from bed 3. This difference in times of migration back into the area may be related

to differences in modes of larval distribution, in particular, residence time in the plankton.

The differing proportions of extinction between major classes of invertebrates across the Deicke event can tell us something about the relative sizes of geographic ranges. The radius from the center of the volcano to the Twin Cities basin is approximately 750 miles (1,200 kilometers). The half circle over which the Deicke ash caused total extinction in the North American Plate has an area of 1,000,000 miles in round numbers (2,500,000 square kilometers). That figure is the outer limit of geographic range between those that went extinct and those that did not. All the echinoderms apparently had geographic ranges smaller than that area. Only 10 percent of the trilobites (*Sceptaspis lincolnensis* is the exception) had ranges larger than that figure, as did 20 percent of the gastropods.

As is usual for mass extinctions, the faunal recovery from the extinction was slow. A detailed bed by bed analysis in Minnesota shows a logistic curve of first occurrence starting with only nine species of macrofossils immediately above the Deicke ash fall, and increasing to 67 taxa by the time of the succeeding Millbrig K-bentonite, about 250,000 years later (Fig. 6.4). As a result, Rocklandian faunas were very patchy, dominated by very few species. The Millbrig eruption, approximately 90 percent of the thickness of the Deicke but composed of 3 closely spaced major events rather than one as in the Deicke, had no apparent effect on the fauna; no major extinction of taxa took place. The recovery of taxonomic diversity continued for the next 4 million years to the peak diversity at the time of the highest Caradocian (Shermanian) sea level. By late Shermanian time (near the end of the Chatfieldian), some 3 million years after the Deicke event, the total number of species present was nearly 300, and no single species of megafossil is likely to dominate a single collection. Species durations are shortest in the Rocklandian substage, longer in the Kirkfieldian substage, and longest in the Shermanian substage, indicating rapid coevolution into new communities, which became more stable as they became more complex in ecological relationships.

The following analysis is based on the collections listed in Weiss (1953), and the detailed bed tracing, museum collections, and observations by Sloan, Levorson, and Gerk (unpub. data).

Species terminating at the Deicke ash fall in the upper Mississippi River valley include five brachiopods, *Pionodema conradi, Campylorthis deflecta, Strophomena plattinensis, Trigrammaria winchelli,* and

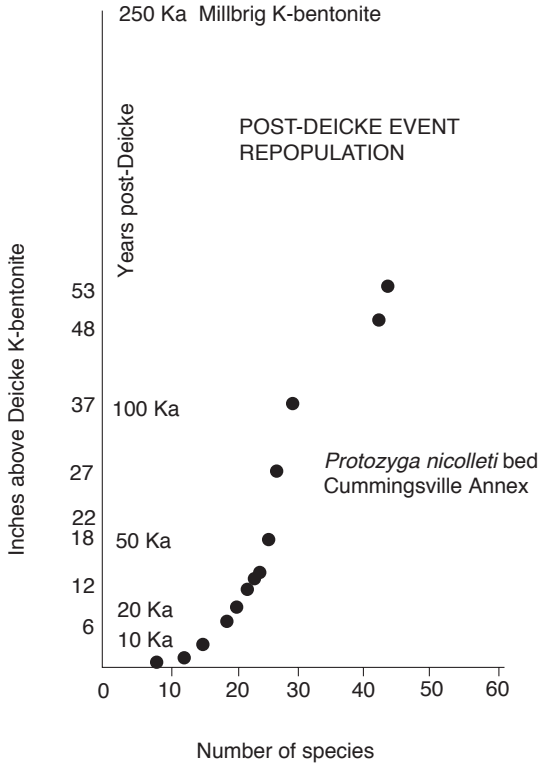


Figure 6.4. Plot showing the increasingly varied fauna of the Spechts Ferry Member part of the basal Chatfieldian stage, to show the rate of repopulation after the Deicke extinction. Data from Weiss (1953), Sloan (1972, unpub. data), and Rice (1987), all projected into the Cummingsville Annex section above the Deicke K-bentonite.

Orbiculoidea lamellosa; 25 of 30 species of clams; 12 named species of trilobites: *Basiliella barrandi, Anataphrus minnesotensis, Thaleops ovatus, Bumastoides milleri, Raymondites longispinus, Dolichoharpes reticulatus, Ceraurinella scofieldi, Ceraurinella templetoni, Gabriceraurus mifflinensis, Encrinuroides rarus, Cybeloides cymelia,* and *Calyptaulax plattevillensis,* all echinoderm species, all cephalopod species, 6 of 12 species of ostracodes (Swain, 1996), and in fossils not recently reviewed, 48 of 60 species of snails.

The term "Lazarus species" for species that go extinct locally only to return a significant time later seems appropriate here. The number of Lazarus species is significantly smaller than those that became extinct or those that were introduced during the repopulation. In the Midwest they include the brachiopods *Acanthocrania setigera, Acanthocrania granulosa, Skenidioides anthonense, Hesperorthis tricenaria, Oepikina minnesotensis,* and *Glyptorthis bellarugosa,* the trilobite *Sceptaspis lincolnensis,* and

the gastropods *Bucania emmonsi, Tetranota bidorsata, Lophospira serratula, Liospira abrupta, L. angustata,* and *Trochonema umbilicatum.* Most other surviving species returned sufficiently rapidly so as not to be termed Lazarus.

The replacing basal Chatfieldian and Rocklandian post-Deicke event fauna includes the following species not recorded earlier: among bryozoa, the most useful is *Prasopora simulatrix,* 18 other species are recorded in Karklins (1987). In brachiopods, 5 species appear at or shortly after the Deicke event. These are *Pionodema subaequata, P. circularis, Zygospira lebanonensis, Rhynchotrema ainsliei,* and *Sowerbyella curdsvillensis (= S. punctostriatus).* Among the trilobites, the immigrants are *Isotelus gigas, I. cf. walcotti, Ectenaspis homalonotoides, Eomonorachus intermedius, Bumastoides porrectus, Raymondites spiniger, Ceraurus plattinensis, Gabriceraurus dentatus,* and *Encrinuroides vigilans,* and somewhat higher, *Dolichoharpes ottawaensis, Ceraurus pleurexanthemus,* and *Flexicalymene senaria.* A very large ostracode fauna apparently migrates into the region when the Decorah Shale begins, very shortly after the Deicke event (Swain, 1996).

These species had ranges that extended out from under the Deicke ash fall, and could reimmigrate from refugia shortly after the ash settled. In the upper Mississippi River valley, 4 species of brachiopods are included in this category; other species with ranges beyond the Deicke ash fall include 8 species of ostracodes, 5 species of clams, 12 species of bryozoa, 7 species of snails, 5 nominal species of clams, and the conodonts.

Not counting the ostracodes, echinoderms, and cephalopods, which have a very high turnover rate, the total is 101 species extinct at the Deicke event, 17 Lazarus species, 46 survivors, and 39 immigrants. This is a very high turnover of the fauna in a very short time, much higher than at any other single level in the entire Mohawkian era.

THE ORDER OF APPEARANCE OF THE CHATFIELDIAN FAUNA

On the basis of Weiss' (1953) 22 measured sections of the Carimona Member and the basal Decorah Shale, the data in Rice (1987), and my own observations, pooled by detailed bed tracing of the measured sections, the order of appearance and development of the basal Chatfieldian fauna is analyzed here (Fig. 6.4). Horizons of first appearance and last appearance of fossils above the Millbrig K-bentonite are listed in the graphic sections of Figures 6.5 and 6.6. The list of species that occur between the Deicke and the Millbrig ash falls and their horizon of first appearance in inches (and calculated years) above the Deicke ash fall projected into the Cummingsville Annex section is as follows:

Inches above the Deicke ash fall	Species	Total number of species
0" —0 years		8
	Arenicolites	
	Bifungites	
	Chondrites	
	Planolites	
	Thallassinoides	
	Doleroides pervetus	
	Pionodema subaequata	
	Endoceras proteiforme	
1" —2,800 years		12
	Isotelus gigas	
	Eomonorachus intermedius	
	Protozyga nicolleti	
	Schmidtella umbonata	
3" —8,400 years		15
	Skenidioides anthonense	
	Hormotoma gracilis	
	Bucania emmonsi	
6" —17,000 years		18
	Rafinesquina trentonensis	
	Liospira abrupta	
	Lophospira oweni	
8" —22,000 years		20
	Pseudolingula iowensis	
	Richardsonoceras sp.	
12" —34,000 years		22
	Conularia trentonensis	
	Primitia sp.	
13" —36,000 years		23
	Loxobucania emmonsi	
14" —39,000 years		24
	Strophomena filitexta	
18" —50,000 years		
	Sowerbyella curdsvillensis	
22" —62,000 years		25
	Prasopora simulatrix	
27" —75,000 years		
Horizon of Cummingsville Annex Special Collection		
	Metaspyroceras wisconsinense	26
37" —104,000 years		29

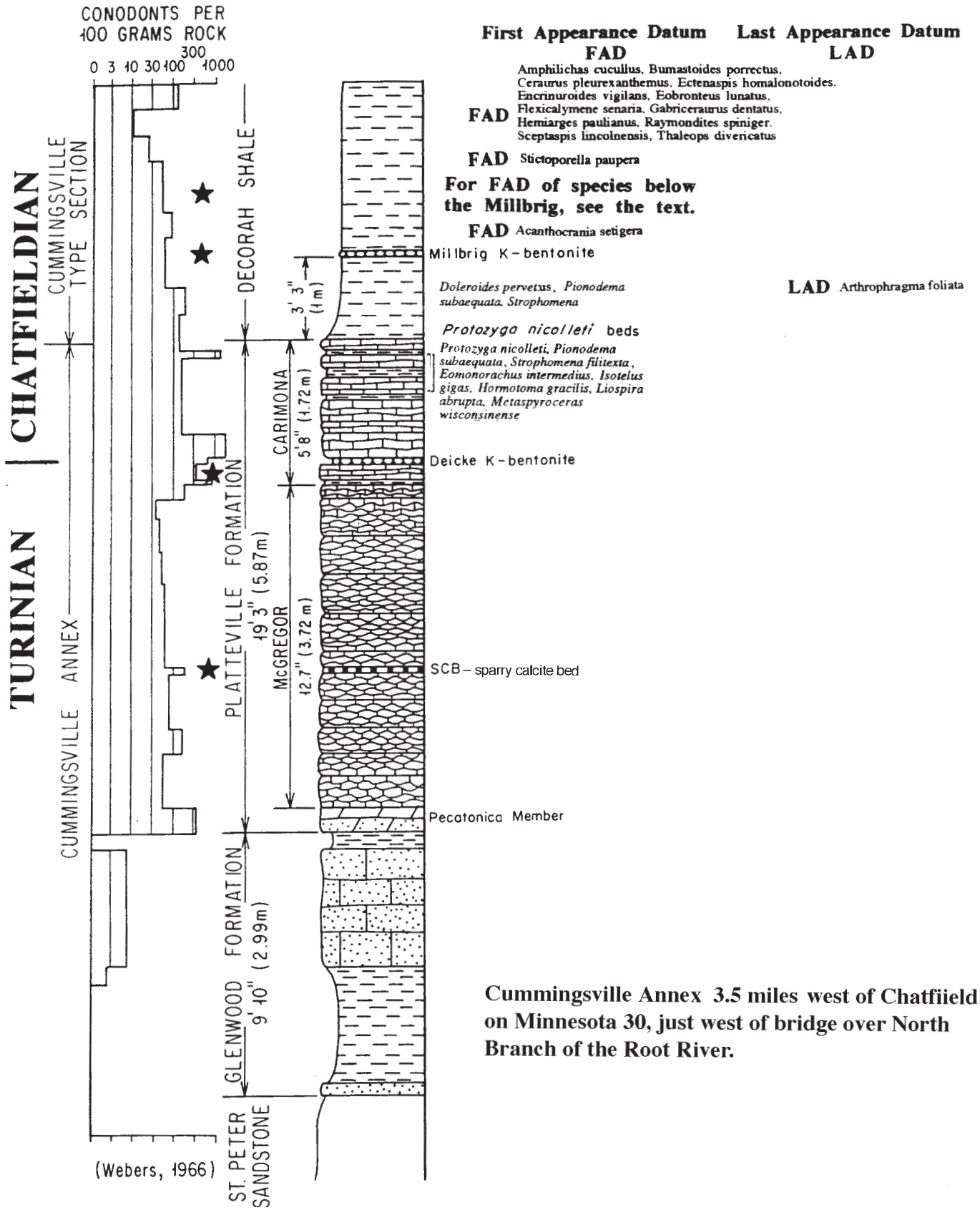


Figure 6.5. Detailed stratigraphy and first and last occurrences of fossils from the Cummingsville Annex, and Cummingsville sections, the type section of the Chatfieldian stage. Originally drawn by Levorson and Gerk (unpub. data), modified and extended from Sloan (1987). The star represents beds with the best preserved and abundant conodonts. With Figure 6.6, it represents the Late Ordovician section of Minnesota. There is about a meter of rock not represented between the two sections. F = Weiss' (1953) fossil collection; R = *Receptaculites*

Figure 6.5. Continued

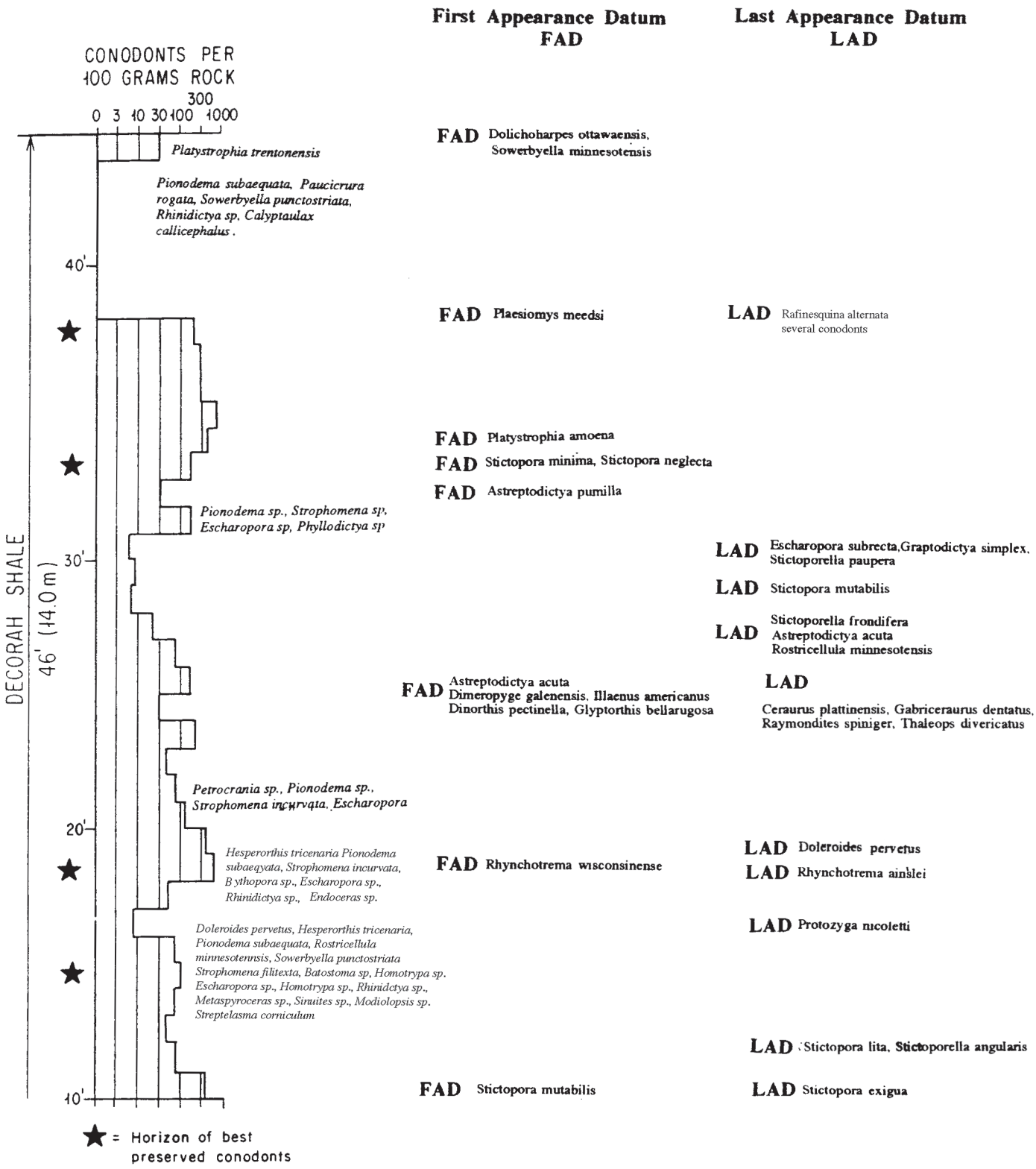


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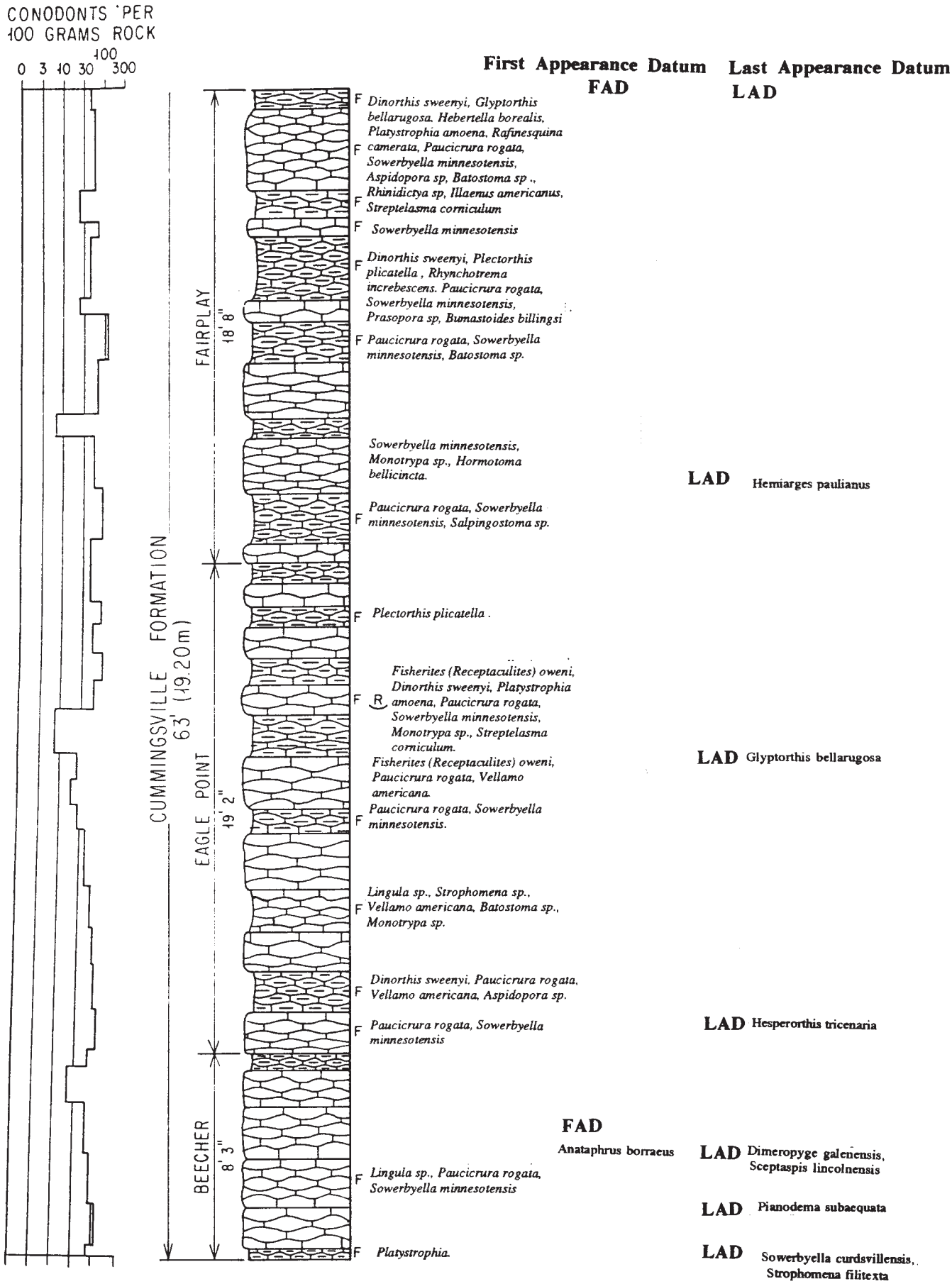
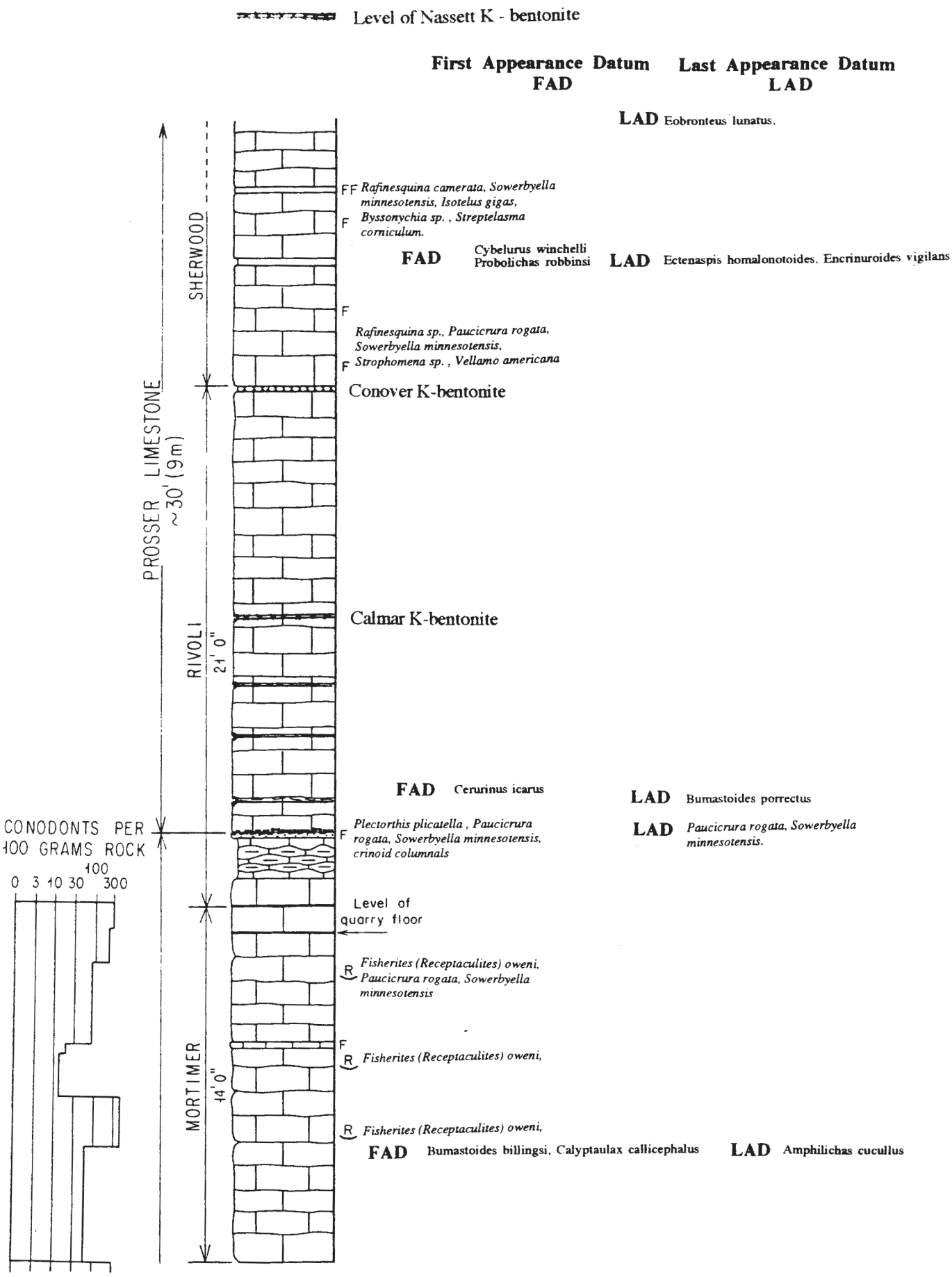


Figure 6.5. Continued



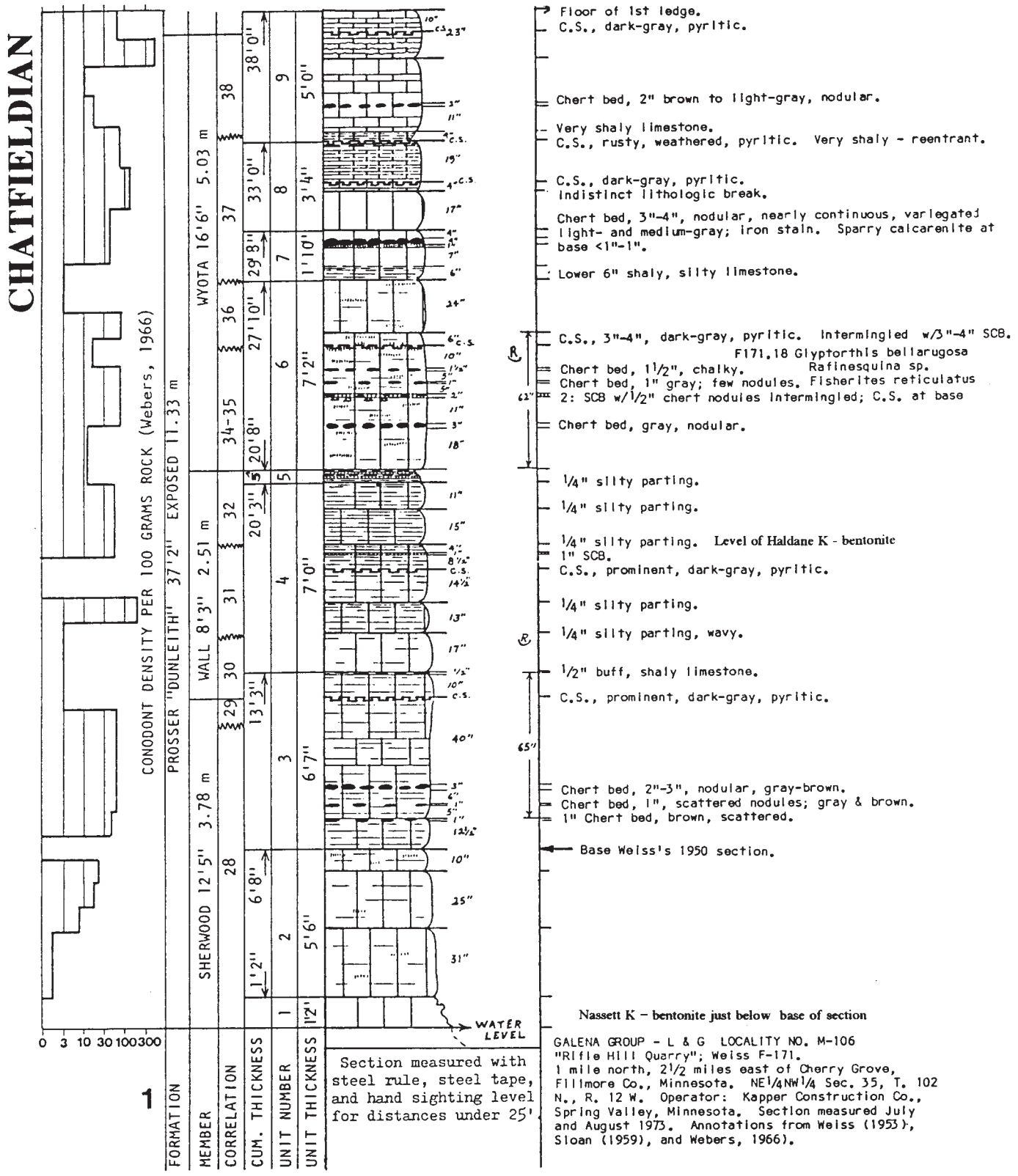


Figure 6.6. Detailed stratigraphy and first and last occurrences of fossils from the Rifle Hill section. Originally drawn by Levorson and Gerk (unpub. data), modified and extended from Sloan (1987).

Figure 6.6. Continued

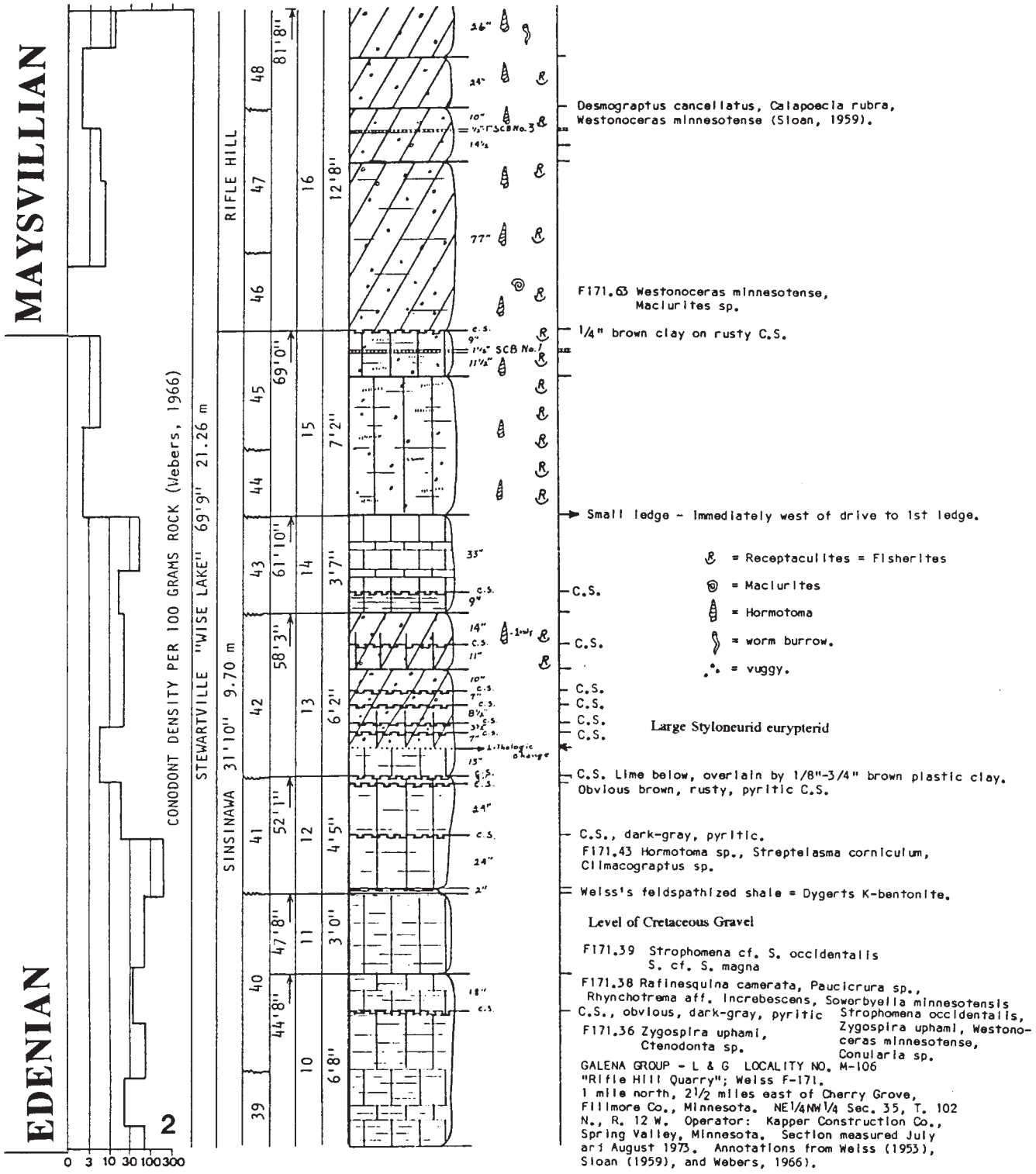


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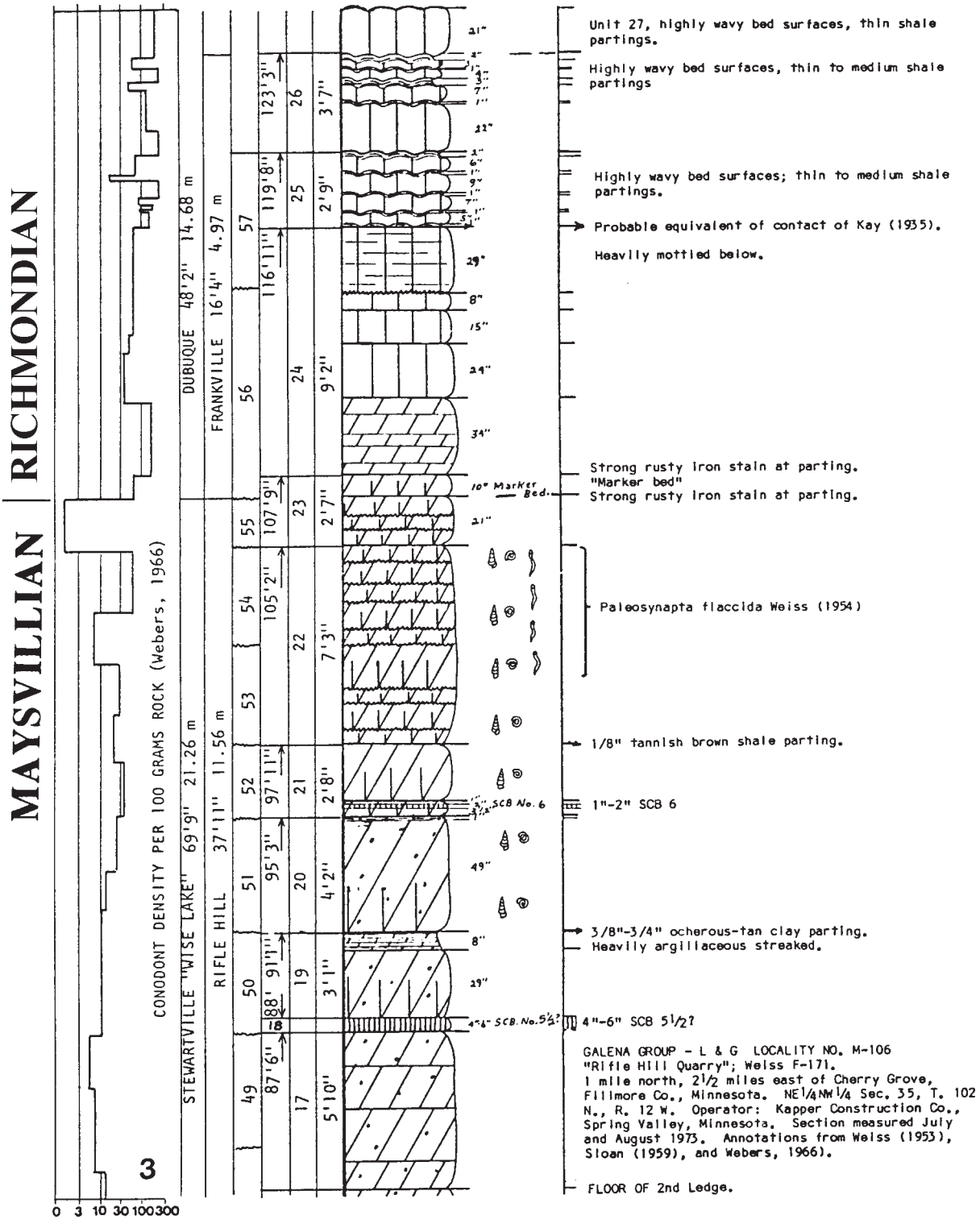
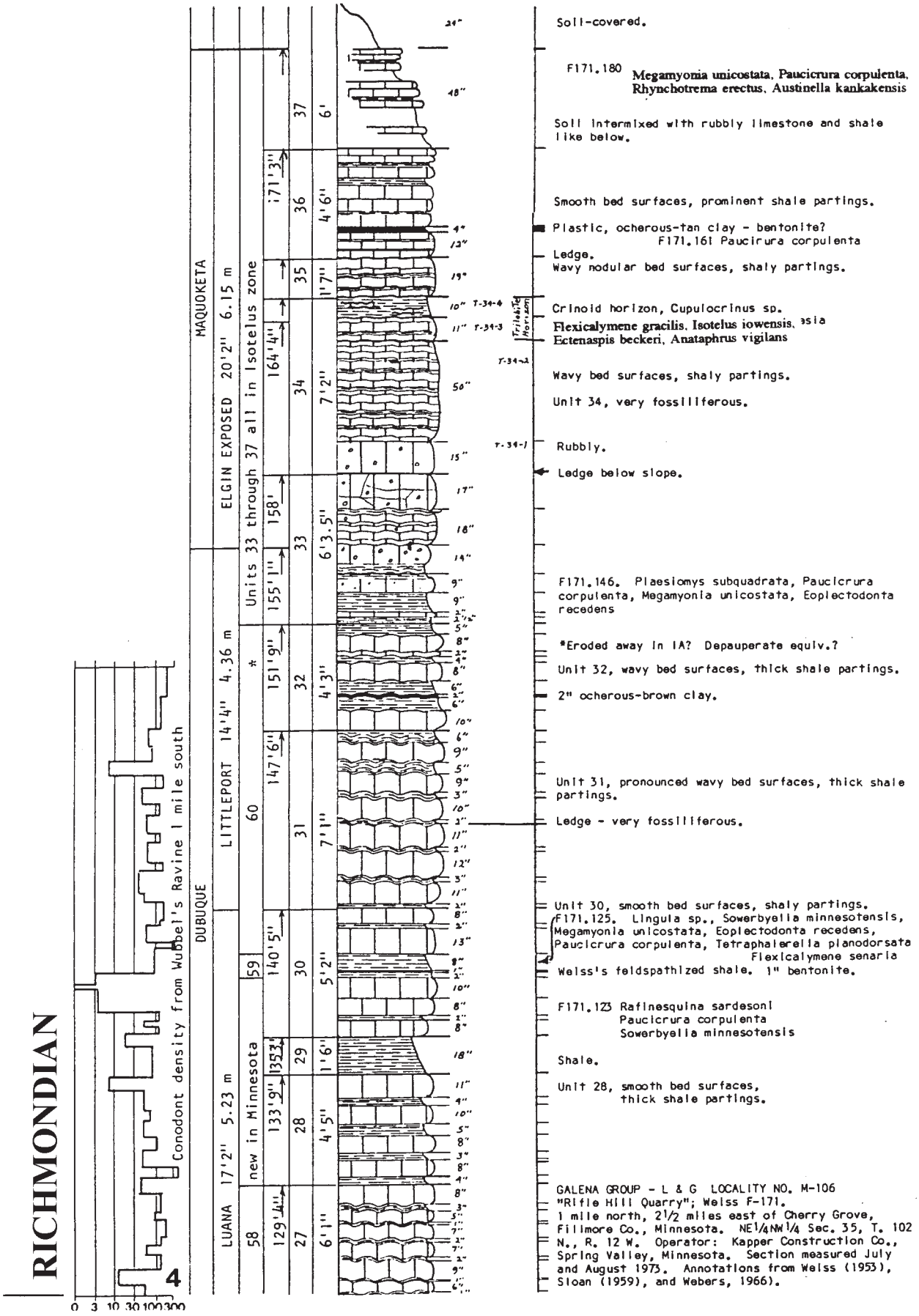


Figure 6.6. Continued



Rostricellula minnesotensis
Platystrophia trentonensis
Ceraurus plattinensis

48" —134,000 years 42

Streptelasma corniculum
Holtedahlina emaciata
Oepikina minnesotensis
Paucicrura rogata
Hesperorthis tricenaria
Petrocrania halli
Bellerophon sp.
Tetranota sexcarinata
Stictoporella angularis
Stictopora lita
Stictopora exigua
Graptodictya simplex
Escharopora subrectus

53" —158,000 years 44

Bellimurina charlottae
Zygospira lebanonensis

54" to 92" —250,000 years 54

Arthrophragma foliata
Stictoporella frondifera
Phyllodictya sp.
Rhinidictya sp.
Fascifera "hamburgensis" (= Diorthelasma weissi; Rice, 1987*)*
Oepikina transitionalis
Platystrophia transitionalis
Holtedahlina emaciata
Oepikina inquassa
Plaesiomys meedsi

****Millbrig K-bentonite****

CHARACTERISTICS OF EARLY IMMIGRANTS

The major feature that determined which animals reached the upper Mississippi River valley first was apparently the time their larvae spent in plankton. As stated above, the first two trilobites to reach southern Minnesota after the Deicke eruption, *Isotelus gigas* and *Eomonorachus intermedius*, were the only Chatfieldian trilobite genera that had planktonic larvae, with a major metamorphosis from the protaspis to meraspid stages, all the others had direct development. The other trilobites took much longer to reach Minnesota. *Ceraurus plattinensis* is the next to occur at 37 inches above the Deicke event, or about 104,000 years, a rate of spread of 25 miles per year.

The brachiopods that appeared early had larvae with long planktonic life. At a 30-day planktonic life, as in many modern brachiopods, it would only take about 40 generations to reach Minnesota from the end of the Transcontinental Arch in Oklahoma. Similarly, although not listed here, the bryozoa with their planktonic larvae arrived early.

The various worms and/or arthropods responsible for the trace fossils *Arenicolites, Bifungites, Chondrites, Planolites,* and *Thallassinoides* all arrived immediately after the Deicke eruption, and presumably had planktonic larvae.

RAREFACTION CURVES AROUND THE DEICKE EXTINCTION

Rarefaction curves are curves showing the percent of individuals of species in a fauna ranked by order of abundance. Normal communities are not dominated by a single species, and populations routinely show a curve with slow exponential dropoff of abundance; about 30 to 40 species can be found in a sample of 1,000 individuals. In communities severely limited in number of species for any reason (in modern communities typical examples are tundra forms, the arctic, or islands), a community can be dominated by one or two species. Four rarefaction curves appear in Figure 6.7, only one of which is shortly after the Deicke extinction (Fig. 6.7B). Of the four, that is the only unusual example of a rarefaction curve.

The Mifflin Limestone rarefaction curve (Fig. 6.7A) is based on a 1977 study by Bretsky and others (1977) in which they collected hundreds of thousands of specimens from 80 measured sections in the upper Mississippi River valley from St. Paul, Minnesota to Rockford, Illinois. They found 170 species, of which 35 made up more than 2.5 percent present in at least one locality. The figures are the percent of the total fauna of these 35 most common species. The temporal duration of the samples is about 0.3 million years, the average age is about 0.5 million years before the Deicke K-bentonite. This interval represents the fauna of the terminal Turinian stage. There are few monospecific fossil horizons. Of these 35 species, 23 disappear at the Deicke K-bentonite, and 14 went extinct. Seven of the 23 were early reimmigrants and two were Lazarus species reappearing significantly later; 10 are uncertain due to incomplete identification.

The Cummingsville Annex *Protozyga nicolleti* bed collection (Fig. 6.7B) is from a catastrophic kill by a submarine slump. The sample was collected 0.5 meter above the Deicke K-bentonite, in the basal Rocklandian substage of the Chatfieldian stage. It

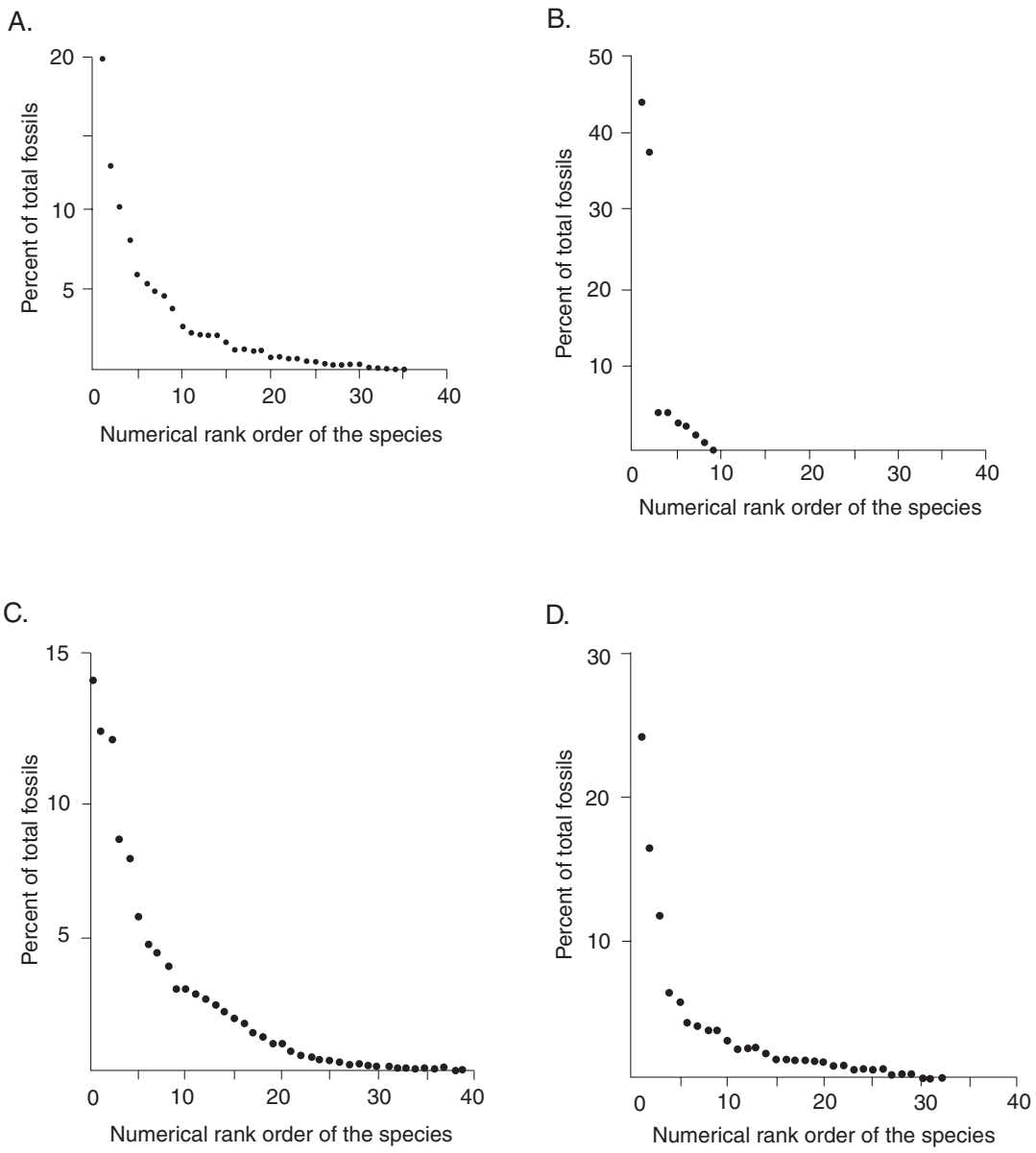


Figure 6.7. Four rarefaction curves to show the effects of the Deicke extinction.

A. Late Turinian rarefaction curve of the Mifflin Member of the Platteville Formation, based on data (many thousands of fossils) from Bretsky and others (1977).

B. Earliest Chatfieldian stage rarefaction curve of 850 fossils from the Carimona Member of the Platteville Formation, 27 inches above the Deicke K-bentonite at the Cummingsville Annex—an unusual rarefaction curve.

C. Rarefaction curve based on a large (5,185 fossils), random washed sample from the upper Decorah Shale 20 meters above the Deicke K-bentonite at Lilydale Park (Stop 6-2).

D. Rarefaction curve based on 1,596 fossils from the mass kill in the *Scalenocystites* bed in the Sherwood Member of the Prosser Limestone at Wagner Hill (Stop 7-2).

is 454 million years old, and includes 850 specimens on a 0.5 square meter bedding plane. Eight species are known, which represents 30 percent of the total of 26 species known from this horizon (a figure based on 28 sampled sections). This is a very unusual rarefaction curve. This fauna represents a restricted fauna of pioneers in the early stages of reconstructing a diverse balanced community following the Deicke extinction. Early Rocklandian substage (basal Chatfieldian) faunas were poor in numbers of species and rich in numbers of individuals of each species. I infer that all these species had long lived planktonic larvae, or were nektonic. Certainly the two trilobites had planktonic protaspids. The few species that had reached the upper Mississippi River valley are present in great numbers in most localities, and there are many monospecific fossil horizons in the Carimona Member of the Platteville Formation and low horizons of the Decorah Shale.

Below is the number of specimens of each species collected from the *protozyga nicolleti* bed at the Cummingsville annex.

Species	Number of specimens
Protozyga nicolleti	372
Pionodema subaequata	319
Strophomena filitexta	40
Eomonorachus intermedius	39
Isotelus gigas	30
Hormotoma gracilis	26
Liospira abrupta	15
Metaspyroceras wisconsinense	9

The next study (Fig. 6.7C) was on a collection of fossils made by David Lazarus (unpub. data) in 1978 from the 19.8- to 20.1-meter level of the Decorah Shale in the abandoned brickyard quarry in Lilydale Regional Park in St. Paul, Minnesota. He made this collection by dispersing the clay from a large bulk sample of the shale by alternately soaking it in kerosene and water. The entire collection weighs 8.2 pounds, and came from a 100-pound sample of sediment. The sample is from the portion of the Decorah Shale belonging to the latest Kirkfieldian stage and was deposited about 1.6 million years after the Deicke extinction event. Identification of the fossils was by Kent F. Adamson (1993). Forty species are represented in this collection, which represents about 16 percent of all the roughly 250 species of fossils known in this horizon. There are few monospecific fossil horizons. Recovery from the Deicke extinction is very well advanced.

The final curve in Figure 6.7 (Fig. 6.7D) is based on a collection of a catastrophic kill made by a submarine slump. It comes from the Wagner Quarry near Cannon Falls in Goodhue County, Minnesota, in the *Scalenocystites* bed of the Sherwood Member of the Prosser Limestone. The collection was made by amateur collectors and was analyzed by David DesAutels (1978). It is described more fully in Stop 7-2.

Isotelus gigas is the most common Chatfieldian asaphid in the upper Mississippi River valley. It appears immediately above the Deicke interval, at the base of the first bed of the Carimona Member. Its pygidium is parabolic in shape. It clearly undergoes a steady increase in maximum size throughout the Chatfieldian and Edenian stages. The Rocklandian substage data are based on over 120 specimens and has a maximum length of 200 millimeters, the Kirkfieldian substage data are based on about 40 specimens, maximum size is 250 millimeters. The late Shermanian substage data are based on about 30 specimens, with a maximum length of 450 millimeters. The earliest Edenian stage data (Sinsinawa member) are based on only 4 specimens, the smallest of which would have come from a trilobite 500 millimeters long.

RADIOACTIVE DATING AND CHRONOLOGY

The key to understanding the absolute ages, rates of sedimentation, rates of evolution and extinction, and synchrony of tectonic events and other processes of short duration lies in a synthesis of data of many types. Although we have known of the K-bentonites since Sardeson's pioneer work beginning in 1924, and our knowledge of details of distribution and correlation has increased steadily, the major contributions to this study have been summarized in Huff and others (1996) and Haynes (1994). They are now directly traceable by fingerprinting techniques for distances of 1,600 kilometers (1,000 miles), as well as by classic bed-tracing techniques.

The K-bentonites are igneous rocks that are interbedded within the normal marine sediments and altered to varying degrees by diagenetic processes. Direct dating of these beds has been done by fission-track dates and by argon ^{40}argon/^{39}argon techniques (Kunk and Sutter, 1984). The fission-track methods have given a standard deviation of between 10 and 16 million years. The argon/argon techniques produce a standard deviation of about 3 million years and show promise of solving much of the internal inconsistency of the fission-track dates.

Kunk and Sutter (1984) dated the T-3 bentonite of Tennessee as 454.2 Ma on the basis of biotite. The mean age of their whole Rocklandian substage cluster of dates is 454.0 Ma; considering the number of samples dated (56), the standard error of the mean is much smaller than 3 m.y. Kolata and others (1987) precisely correlated the Deicke K-bentonite of the upper Mississippi valley with the T-3 bentonite of Tennessee on the basis of rare element fingerprinting. There is no doubt that the Deicke K-bentonite is the same as the T-3 bentonite.

DURATION OF UNCONFORMITIES

At Rockford, Illinois, there is a significant unconformity between the Quimby's Mill K-bentonite (just earlier than the Deicke K-bentonite) and the top 0.36 meter (14 inches) of the Guttenberg Member. The missing interval is 14 CSS units; the duration of the unconformity is thus approximately 0.8 m.y. The missing interval is the Rocklandian substage and part of the Kirkfieldian substage. This unconformity decreases to the northwest, and disappears at Guttenberg, Iowa. In Minnesota, deposition of the Dubuque Formation and the Maquoketa Group was continuous, but in Iowa southeast of Decorah, they are separated by the hardground and associated sediments known as the Depauperate zone. The duration of the unconformity represented by the hardground is difficult to interpret, but Levorson and Gerk (1972) estimated that the Rifle Hill section (their "M-106") has about 1 meter of rock that is missing in Iowa that is 1.6 CSS units, or approximately 0.09 m.y.

The duration of the unconformity between the St. Peter Sandstone and the underlying Shakopee Formation of latest Canadian age (about 485 Ma) is approximately 25 m.y., about the duration of the entire Late Ordovician (the combined Mohawkian and Cincinnatian stages).

The many corrosion surfaces–corrosion zones–hardgrounds in the carbonates do not appear to represent a significant amount of time, individually or collectively.

PLANKTON STANDING CROP

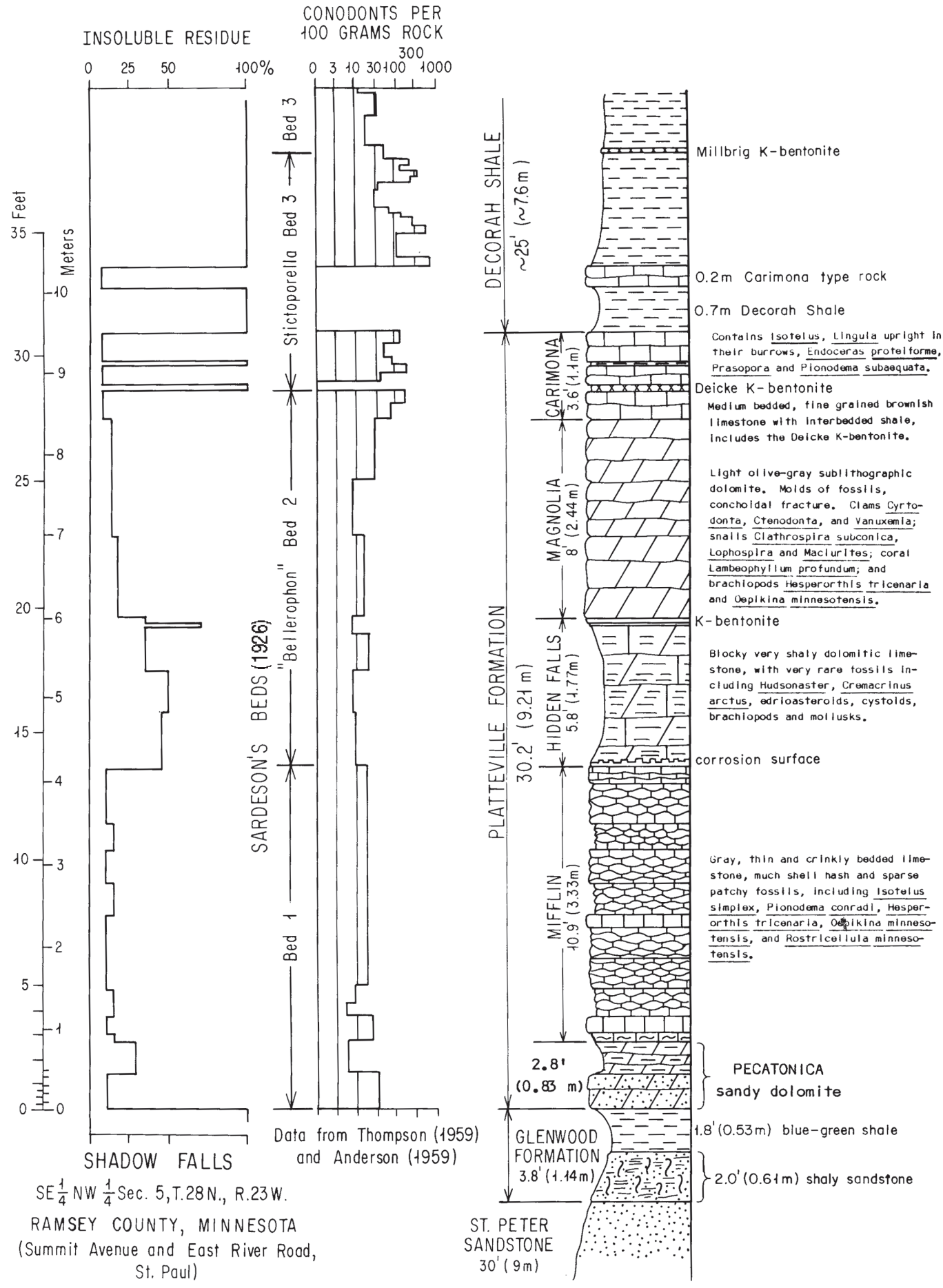
Primary productivity and standing crop of plankton in the sea is usually determined by concentrations of nitrates and phosphates, which are the limiting nutrients in Liebigs' "Law of the Minimum." It is usually measured in grams of wet plankton per square meter. The Diecke K-bentonite provides a spot estimate of primary productivity because it caused a quantitative plankton kill and the organic matter is still preserved as petroliferous shale. I sampled the bottom one centimeter of the Deicke K-bentonite at Cannon Falls, Minnesota several years ago. Dr. Francis Ting determined the total nitrogen in an aliquot, and with this information I calculated the equivalent amount of wet organic material per square meter. The result was 2.5 kilograms per square meter, close to that of modern, productive, shallow, tropical seas.

Conodont abundance in conodonts per 100 grams of sediment appears to be the best measure of local primary productivity. The conodont animal was nektonic, and fed directly on zooplankton and phytoplankton. It seems reasonable to infer that conodont abundance was directly proportional to the plankton abundance. We have routinely calculated conodont abundance for each sample studied for conodonts since 1959. Logs of this statistic are shown with the graphic sections in Figures 6.5, 6.6, and 6.8. The range of values is from 0 to 1,500 conodonts per 100 grams of rock. The conodont density in the beds surrounding the Diecke K-bentonite at Cannon Falls was 300, implying that the total range of productivity in absolute values was from 0.1 to 12.5 kilograms per square meter. In general, the lowest values occur in the Stewartville Formation, and the highest values occur in the Carimona Member and Spechts Ferry Member (Figs. 6.5, 6.6). These figures relate closely to the uplift and exposure of the nearby Transcontinental Arch.

TECTONICS AND SEDIMENTATION

Minnesota Ordovician sediments were deposited in a large bay (the Hollandale embayment) between the Transcontinental Arch in central Minnesota, the Wisconsin Dome and Arch to the east, and the Sioux Quartzite islands to the west. On the northern side of the Transcontinental Arch lies the Williston basin of the Dakotas, Manitoba, and Saskatchewan. The rocks are generally similar on the two sides of the Transcontinental Arch but have differing names. The Winnipeg Formation of the Williston basin in

Figure 6.8. Stratigraphic section of the Glenwood and Platteville Formations, and Decorah Shale at Shadow Falls at the intersection of East River Road and Summit Avenue in St. Paul, Minnesota, with conodont and insoluble residue logs.


INSOLUBLE RESIDUE
0 25 50 100%
CONODONTS PER 100 GRAMS ROCK
0 3 10 30 100 300 1000
Feet
Meters
35
30
25
20
15
10
5
0
10
9
8
7
6
5
4
3
2
1
0
Bed 3
Stictoporella Bed 3
"Bellerophon" Bed 2
SARDESON'S BEDS (1926)
Bed 1
SHADOW FALLS
SE¼ NW¼ Sec. 5, T.28N., R.23W.
RAMSEY COUNTY, MINNESOTA
(Summit Avenue and East River Road, St. Paul)
Data from Thompson (1959) and Anderson (1959)
DECORAH SHALE
~25' (~7.6m)
PLATTEVILLE FORMATION
30.2' (9.21m)
CARIMONA 3.6' (1.1m)
MAGNOLIA 8' (2.44m)
HIDDEN FALLS 5.8' (1.77m)
MIFFLIN 10.9' (3.33m)
2.8' (0.83 m)
GLENWOOD FORMATION 3.8' (1.14m)
ST. PETER SANDSTONE 30' (9m)
Millbrig K-bentonite
0.2m Carimona type rock
0.7m Decorah Shale
Contains Isotelus, Lingula upright in their burrows, Endoceras proteiforme, Prasopora and Pionodema subaequata.
Deicke K-bentonite
Medium bedded, fine grained brownish limestone with interbedded shale, includes the Deicke K-bentonite.
Light olive-gray sublithographic dolomite. Molds of fossils, concholdal fracture. Clams Cyrtodonta, Ctenodonta, and Vanuxemia; snails Clathrospira subconica, Lophospira and Maclurites; coral Lambeophyllum profundum; and brachiopods Hesperorthis tricenaria and Oepikina minnesotensis.
K-bentonite
Blocky very shaly dolomitic limestone, with very rare fossils including Hudsonaster, Cremacrinus arctus, edrioasteroids, cystoids, brachiopods and mollusks.
corrosion surface
Gray, thin and crinkly bedded limestone, much shell hash and sparse patchy fossils, including Isotelus simplex, Pionodema conradi, Hesperorthis tricenaria, Oepikina minnesotensis, and Rostricellula minnesotensis.
PECATONICA sandy dolomite
1.8' (0.53m) blue-green shale
2.0' (0.61m) shaly sandstone

the Minnesota subsurface and Manitoba outcrop is shown by its conodont fauna to be the same age as the interval from the St. Peter Sandstone to the Cummingsville Formation (Bayer, 1959). The overlying Red River Dolomite corresponds to the Prosser to Fort Atkinson Limestones on the basis of the typical upper Red River coral *Bighornia* in the Fort Atkinson Limestone of Iowa, and the Stony Mountain Shale corresponds to the Brainard Shale and Neda Formation. The sand-poor Winnipeg Formation corresponds to the clastic-rich first half of the Middle Ordovician upper Mississippi River valley sequence. It appears that most of the sand on the Transcontinental Arch went to the south side to produce the St. Peter Sandstone.

The density of hardgrounds or corrosion surfaces is another measure of the tectonic influences on regional rates of sedimentation. They represent brief interruptions of deposition. The Waupon core on the eastern edge of the Wisconsin Arch has 122 hardgrounds in 127 feet of Galena Group rocks from the base to the Sinsinawa Member (Mudrey, 1997, p. 99). Levorson and others' (1987) general section for northeast Iowa, farther from the Transcontinental Arch but not over the Keweenawan rift, has 67 hardgrounds for the 241 feet of the same interval, and at Rifle Hill, relatively close to the axis of the Keweenawan rift, there are only 16 hardgrounds in the Wall to Sinsinawa interval of 57 feet. At Cummingsville there are only 3 in the Cummingsville Formation, and none in the Rivoli and Sherwood Members of the Prosser Limestone for a total of 19 for the same interval.

The Transcontinental Arch is composed of Precambrian rocks that were involved is several tectonic events. The Arch is mainly the Great Lakes Tectonic Zone (Sims and others, 1980); most of the historic earthquakes in Minnesota are along this structure. The Great Lakes Tectonic Zone stands above the older Archean crust in Minnesota, under Cretaceous and Pleistocene sediments. The topographic high is about 100 meters (300 feet) above the adjacent Archean rocks, and the width is about 80 kilometers (50 miles). The Arch has a strike of S. 70° W. and is composed of granite plutons and folded and thrust Animikian sediments, mostly shales, of varying metamorphic grade up to the garnet and staurolite zone. The Great Lakes Tectonic Zone can be traced through northern Wisconsin and Michigan, into Ontario, to Sudbury. To the southwest, the extension may be the Colorado Lineament (Sims and others, 1980). The Great Lakes Tectonic Zone was offset by the development of the Keweenawan rift valley about 1.1 b.y. The Keweenawan rift crossed the Great Lakes Tectonic Zone at an angle of about 20°, the total offset was about 160 kilometers (100 miles). In Wisconsin and northern Michigan, the Great Lakes Tectonic Zone is on the south side of Lake Superior, in Minnesota it is northwest of the extension of the lake. The axis of the Hollandale embayment and the origin of the persistently depressed nature of the basin is the Keweenawan Midcontinent rift. The crust in the rift is dominantly basalt to a depth of 10 kilometers (6 miles), and is interbedded near the top with red stream sandstone and shale. This thick pile of basalt is the cause of the Midcontinent geophysical anomaly, in which the mass of the basalt increases the local acceleration of gravity to the highest values in the continent. Despite the fact that the rift was over 500 million years old during Late Ordovician time, it was still subsiding at a slow pace. Minnesota Ordovician rocks are superimposed on this major crustal feature. Northern Illinois Ordovician rocks, however, were deposited over the south-trending Wisconsin Arch, which behaved as a positive feature.

During Caradocian time, a global transgression of sea level occurred, restoring the sea level to the former high level reached during the Ibexian stage. Locally, the first sign of this is the spread of the marine St. Peter Sandstone over the karst surface developed during the long Whiterockian unconformity. The eustatic sea level rose until it reached the highest level during the deposition of the Prosser Limestone, during the Shermanian substage, then sank a little, and finally dropped precipitously during the end of Ashgillian time (Brenchly and Newall, 1984), apparently under glacio-eustatic control at that time. Basement subsidence under the developing pile of carbonates apparently determined the observed rates of deposition.

The apparent source of the St. Peter Sandstone was the Late Cambrian Croixan sandstone sequence that covered the Transcontinental Arch. These sands still occur as close to the axis of the Great Lakes Tectonic Zone as 50 kilometers (35 miles). The volume of the St. Peter Sandstone in the Hollandale embayment is approximately equal to the restored volume of Croixan sediments stripped off the south half of the arch. The sands of the north half went to the Williston Basin. With the sandstones removed, the bedrock consisted of the pre-Mt. Simon Sandstone residuum (Morey, 1972), and the underlying Animikian slates and schists. The Glenwood Formation has a much higher percentage of garnets than the underlying St. Peter Sandstone; small, sand-size garnets are common in these metamorphic rocks near the crest of the arch.

The Mifflin Member of the Platteville Formation reflects deepening of the late Ordovician sea to about 50 meters. A pulse of uplift in the central Minnesota source caused the deposition of the Hidden Falls Member, briefly increasing the clay enough to spread out to the margins of the Twin Cities artesian basin, a basin separated from the rest of the Hollandale embayment by a transform fault offset of the Keweenawan trough to the southeast. At least one basement fault was active during this time. The clay component of the Hidden Falls Member spread into the trough on the eastern side of the Red Wing–Rochester anticline.

At the beginning of the Rocklandian substage/basal Chatfieldian stage, a major pulse of uplift produced a much greater supply of clay to the Hollandale embayment. The clay of the Decorah Shale reached all the way to northern Illinois, a distance in outcrop at right angles to the source of 370 kilometers (230 miles). The maximum thickness of the Decorah Shale wedge is 27 meters in St. Paul, Minnesota. Restoring the clay of the Decorah Shale wedge back to the crest and southern half of the Transcontinental Arch required an uplift of 300 meters (1,000 feet) to produce the shale. This uplift is exactly correlative with the 0.8-m.y. unconformity in northern Illinois on the Wisconsin Arch, and the Deicke–Dickeyville cluster of large ash beds, and also with the depression of the Twin Cities basin marked by the increase in the Deicke–Millbrig interval and the highest local productivity as shown by the conodont density curves, presumably the result of nutrients being eroded off the arch. Very slightly later (Sardeson's [1926] bed 4), another small basement fault in Fillmore and Houston Counties with a total throw of 15 meters (50 feet) thinned the Decorah Shale at Spring Grove underpass to almost half the regional value.

This unique concurrence of unusual events could be coincidence, but the precision of correlation is too great. It appears to be a case of internal deformation in the North American Plate produced by an increase in the rate of subduction or by a collision of another continental plate on our east coast. The Keweenawan trough subsided so that the Diecke–Millbrig interval is 2 to 3 meters (8 to 11 feet) thick; at the Wisconsin Arch at Rockford, Illinois, about 10 meters (30 feet) was eroded or never deposited. With the gradual erosion of the Transcontinental Arch, the width of the clay belt steadily narrowed. The arch was submerged during the deposition of the Prosser Limestone and Stewartville Formation of the Galena Group. Some Platteville and Glenwood Formation rocks were eroded during the deposition of the later Decorah Shale; some weathered and reworked conodonts belonging to taxa otherwise found only below the Deicke interval were found offshore. The timing of pulses of uplift of the Transcontinental Arch can be deduced from the sawtooth profile of the Cummingsville Formation by converting the thickness of pairs of pure and shaly carbonate units in the Cummingsville Formation to time. The pulses appear to have occurred at intervals ranging from 35,000 years to 70,000 years. Mazzulo's (Mazzulo and Ehrlich, 1987) short cycles of irregular and rounded St. Peter Sandstone sand grains appear to be of similar frequency. There is a longer cycle as well, which can be seen in Mazzulo's 10 to 15 foot cycles of roundness, and in the cycles of conodont density in the Cummingsville Formation section, which vary from 8 to 15 feet; the duration of this longer cycle varies from 0.20 to 0.37 m.y., with a mean value of 0.30 m.y.

During the early Richmondian stage, near the end of the Caradocian-equivalent strata, another uplift of the arch took place. This was minor compared to the Decorah Shale uplift, but was nonetheless significant. The shale of the Dubuque Formation came from central Minnesota, and thins toward Illinois. This was noted by Sardeson (1927), and has been documented by Levorson and others (1979). Minnesota has a continuous sequence of beds across the Dubuque Formation–Maquoketa Formation contact, but Iowa and Illinois, outside the Keweenawan trough, instead have the Depauperate zone and the associated hardground, an unconformity that has a calculated duration of 0.09 m.y.

An additional feature of paleogeography is the Middle Proterozoic Sioux Quartzite high, which has wave-cut benches on it at a present elevation of 460 meters (1,500 feet) and reaches a peak elevation of 550 meters (1,800 feet). In the absence of Ordovician rocks, we cannot tell whether these benches are Ordovician or Cretaceous in age. These hills were islands in the Ordovician seas, and served to reduce the flow of sand to the area south of them; the St. Peter Sandstone is conspicuously thinner in northwestern Iowa than in Minnesota.

CHATFIELDIAN STAGE BIOFACIES AND DEPTH OF DEPOSITION

The depth of deposition of these rocks was always shallow, although there are no signs of emergence, at least in Minnesota. Green algae (Chlorophyta) including *Vermiporella*, *Fisherites* (=*Receptaculites*), and *Ischadites* are major carbonate producers in these rocks. Because the red light, on which green

algae depend for photosynthesis, is extinguished at shallower depths than other wavelengths, these algae are usually restricted to depths less than 50 meters (Ginsburg and others, 1972). *Ischadites* is most common in the Prosser Limestone, which represents the deepest facies of these rocks. I infer that the depth range of the Minnesota Late Ordovician sea was from sublittoral to 50 meters (160 feet).

Chatterton and Ludvigsen (1976) developed a classification of trilobite biofacies that they interpreted as a bathymetric gradient. From shallowest to deepest these are: 1. *Bathyurus* biofacies, 2. *Isotelus* biofacies, 3. *Calyptaulax-Ceraurinella* biofacies, typified by those genera and *Sphaeroexochus* and *Cybeloides*, and 4. *Dimeropyge* biofacies, typified by that genus and *Dolichoharpes Bathyurus extans,* and is present in both the Magnolia and Carimona Members (below the Deicke K-bentonite), which represents the shallowest facies recorded. *Isotelus* is present throughout all the units of the section, but is most common in the Pecatonica Member, Carimona Member, Decorah Shale, Dubuque and Maquoketa Formations. The next deeper facies, typified by *Calyptaulax* and *Cybeloides*, is more typical of the Mifflin Member, Cummingsville Formation, and Prosser Limestone. *Dolichoharpes* and *Sphaeroexochus* represent the deepest part of the third facies, transitional to the fourth facies, and occur only in the Prosser Limestone in Minnesota, and in the Mifflin Member and Prosser Limestone in Illinois.

Cisne and others (1984) investigated the faunas of the Platteville Formation, Decorah Shale, and lower Dunleith Formation at Guttenberg, Iowa, and the Platteville Formation and lower Decorah Shale at St. Paul, Minnesota, from the standpoint of depth estimates inferred from variations in biofacies. The extremes of their biofacial components were bryozoans and *Doleroides* for the shallower component, and *Paucicrura* and *Sowerbyella* for the deeper component. They inferred a near zero depth for the St. Paul section, and a 5-meter depth for the basal Decorah Shale at Guttenberg, Iowa. They show a shoaling during the Spechts Ferry Member deposition, deepening during the Guttenberg Member deposition, shoaling again after the Guttenberg Member deposition, and redeepening during the upper Ion Member and lower Dunleith Formation (Cummingsville equivalent) deposition. Trends agree in the two sections after a correction for a miscorrelation of the St. Paul section inherited from Templeton and Willman's (1963, Fig. 24) miscorrelation of the Millbrig K-bentonite in St. Paul with the Elkport K-bentonite of Guttenberg.

Megaripples with a wavelength of 60 centimeters, and amplitude of 5 centimeters occur in the Carimona Member and throughout the Decorah Shale in Minnesota. These have usually been assumed to represent shoaling, with storm wave removal of fine-grained sediment, and reshaping of coarse-grained bottom debris, mostly shells.

The ratio of benthic mollusks to brachiopods appears to represent a depth gradient, with the benthic mollusks a shallower and articulate brachiopods a deeper facies. This would be best measured by absolute counts of specimens, but the trend can also be measured by numbers of taxa. The simple ratio of brachiopods divided by the sum of clams and snails agrees with the other estimates. The Stewartville Formation and St. Peter Sandstone have nearly identical lowest ratios of 1:16 and 1:15; the others are more balanced.

To combine all the types of data, I suggest the following: the shallowest depth recorded in the Minnesota section is the brassy oolite horizon at 19 meters (63 feet) deep, in addition to the two horizons of gypsum crystals in the Decorah Shale of St. Paul; nothing else approached the water surface. The deepest facies represented is the *Ischadites* acme zone in the Sherwood Member of the Prosser Limestone (Dunleith Formation in Iowa and Illinois); I arbitrarily place this at 50 meters depth on the basis of modern green algae distribution and the extinction of red light with depth. I interpret the lingulid-*Isotelus* facies of the Carimona Member and Dubuque Formation to be relatively shallow, from 10 to 25 meters, but always below low tide because there are no signs of stromatolites, raindrop craters, or mudcracks, which are well preserved locally in the Canadian Shakopee dolomite. I interpret the mollusc-rich horizons of the Magnolia Member and Stewartville Formation to be very shallow, 10 meters or less in depth, but below low tide. *Fisherites* appear to extend deeper than *Maclurites*, perhaps to 20 meters. I interpret the bryozoan-*Doleroides* community of Cisne to be about 10 meters depth or less, the *Sowerbyella*- (or *Thaerodonta) Paucicrura* community I interpret to be about 15 to 25 meters depth. The intermediate trilobite community of *Calyptaulax-Cybeloides-Ceraurus* I interpret to be of 20 to 50 meters depth, with *Sphaerexochus* and *Dolichoharpes* representing the deeper part of that facies.

This synthesis of depth facies produces patterns of depth changes consistent with those based on wavelength of ripples and other sedimentary structures. This review is supplemented by one based primarily on the 68 samples collected by William F. Rice from the 27-meter-thick section of the Decorah Shale at Lilydale Park in St. Paul. Bryozoa and mollusks were counted by Chris Schneider, Sloan counted the trilobites. This study

augmented the collections of Levorson and Gerk, Sardeson, and Malcolm P. Weiss (private collections). Modern bivalves are dominant in very shallow water, as are inarticulate brachiopods and bryozoa. Archaeogastropods are predominantly herbivorous and hence are restricted to very shallow water. On the other hand, articulate brachiopods typically occur in deeper water. The ratio of benthic mollusks to articulate brachiopods represents a depth gradient, with the benthic mollusks a shallower and articulate brachiopods a deeper facies. The simple ratio of articulate brachiopods divided by the sum of clams and snails agrees with other estimates. *Pionodema* is the most common brachiopod in the shallower, lower Decorah Shale. *Sowerbyella* dominates in the upper, deeper part. *Paucicrura* is common in the deeper, upper part of the Decorah Shale. The two common *Zygospira* species are most abundant in the deeper-water facies of the upper Decorah Shale. *Rafinesquina* is a deeper water fossil. There is an inverse relation between shallow water bryozoa and deeper water articulate brachiopods. The results are summarized in Figure 6.9.

Clearly, the regional bottom slope was very flat, with relief of 10 to 20 meters (30 to 60 feet) or less in 500 kilometers (300 miles). A considerable amount of carbonate mud must have been put into suspension during temporary lowering of the wave base during major storms, only to settle out after the storm subsided.

The Twin Cities basin is a local basin in the Midcontinent Gravity High, the late Precambrian 1 b.y. Keweenawan rift. Contours in Figure 6.2 are on the Jordan Sandstone–Oneota Dolomite contact, approximately equal to the Cambrian/Ordovician boundary. The thickest Decorah Shale interval that remains is at Lilydale Park, only 70 kilometers from its source, the Transcontinental Arch at St. Cloud, Minnesota.

RATES OF DEPOSITION

The rate of deposition of the Galena Group is remarkably uniform from St. Paul, Minnesota, to Dubuque, Iowa. The thickness of the Dubuque-through-Beecher interval at Dubuque is 92 percent of the thickness of the same interval at Cummingsville, Minnesota and Rifle Hill, Iowa. Most of the difference is taken up by the increase in rate of deposition of shale to the northwest, and by the small unconformity of the Depauperate zone. The Decorah Shale was deposited 40 percent faster in St. Paul than in Fillmore County, near the Iowa border (Rice, 1987). However, as closely as I can determine, the carbonate of the Platteville Formation was deposited at the same rate in the two areas.

In northern Illinois, the rate of deposition of the Platteville Formation was significantly faster than that of the Galena Group or the Platteville Formation in Iowa and Minnesota. DeMott (1987) found no significant change in trilobites that cannot be explained by deeper water in northern Illinois than in Minnesota. The top of the Pecatonica Member appears to be the same age in Illinois as it is in the Twin Cities basin in Minnesota. The Mifflin Member in both places represents the deepest deposition, presumably under the control of eustatic sea level changes. On that basis, the Platteville Formation was deposited about 3 times faster in Illinois near Rockford, than in Minnesota or Iowa.

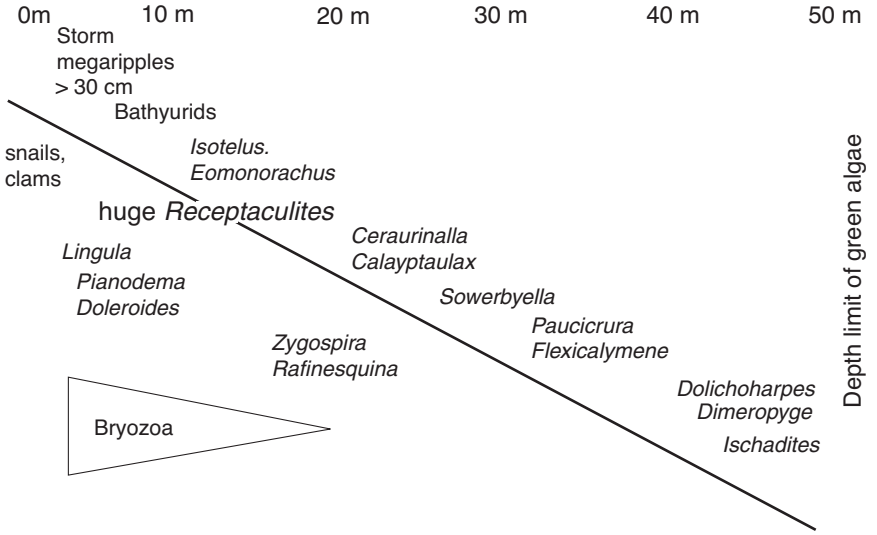


Figure 6.9. An estimate of Chatfieldian paleoecology, with genera plotted at what appears to be their optimum depth. Based on unpublished data collected by Sloan, Schneider, Rice, and DesAutels.

Based on the Cummingsville and Rifle Hill sections between the Deicke K-bentonite and the top of the Dubuque Formation, the mean rate of deposition is 90.85 meters in 7.2 m.y., or 17.619 meters per 1 m.y., or 17.6 millimeters per thousand years. Depth changes are small compared to total sediment accumulation, and so the rate of deposition most likely was determined primarily by the rate of crustal subsidence. Wave dispersal of algal micrite is a likely way of ensuring uniform carbonate deposition.

Because the Glenwood Formation in Minnesota is laterally equivalent to the Mifflin Member, the Pecatonica Member, and the top of the St. Peter Sandstone, it is clear that no significant regional unconformity separates the Ancell Group (St. Peter Sandstone + Glenwood Formation) from the Platteville Group of Illinois.

DURATIONS OF THE RANGES OF SPECIES

The 18 Decorah Shale brachiopods from the Lilydale Park exposure that have their entire range within the exposed rocks have a mean vertical range of 13.08 meters (42.9 feet), equivalent to an average range of 9.4 meters (30.8 feet) at Cummingsville (Rice, 1987). This converts to an average duration of the range of brachiopod species of 0.6 m.y. during the Rocklandian stage.

Sardeson's (1926) beds themselves differ in duration. The Platteville Formation and Spechts Ferry Member beds (1, 2, and 3) average 0.16 m.y. in duration; Decorah beds 4 and 5 are 0.4 m.y. each; Dunleith Formation or Cummingsville Formation and Prosser Limestone beds 6, 7, and 8 average 0.5 m.y. each; bed 9, the Stewartville Formation, is 1.2 m.y. long; and bed 10, the Dubuque Formation, is 0.6 m.y. Except for bed 10, there is a more or less steady slowing of rates of extinction and replacement or evolution. This is logical because one would expect high rates of evolution and short temporal ranges early in a major transgression such as the Caradocian epoch, followed by declining rates of evolution and lengthening temporal ranges as the new communities became stabilized.

FIELD TRIP STOPS

FIELD TRIP 6

DIRECTIONS: Leave the Radisson Hotel and turn left (east) on Washington Avenue to Oak Street (stoplight). Turn right onto Oak and travel four blocks to East River Road. Turn left onto east River Road and proceed southeast for 2.9 miles to the intersection with Summit Avenue in St. Paul. Pull into the turnaround at the base of the monument and park.

STOP 6-1

Shadow falls (synonymous locality names quoted in the literature are Finn's Glen and Summit Avenue; Fig. 6.8)

Location: T. 28 N., R. 23 W., sec. 5, SE, NW; Summit Avenue and East River Road

Description: Follow the path down over glacial till to the Decorah Shale (the lower 6+ meters [20+ feet] is exposed) and the Platteville Formation. This is the best exposed and most accessible exposure of the upper St. Peter Sandstone, Glenwood Formation, Platteville Limestone, and lower Decorah Shale in the Twin Cities. All the units can be seen easily, and there is enough room to spread out. This was our standard experimental Platteville section: whenever we got a bright idea about a new kind of study, this is where we came first. The center of the Twin Cities basin is located only a couple miles away (Fig. 6.10), near the hotel.

A condensed section is as follows, also see the graphic section (Fig. 6.8).

- **6+ meters (20+ feet)** of Decorah Shale, with the 3-centimeter (1 inch) Millbrig K-bentonite clearly visible 2.2 meters (7.3 feet) above the top of the Carimona Member, and 2.9 meters (9.4 feet) above the Deicke K-bentonite. Sardeson's (1926) bed 3 is the interval between the Deicke K-bentonite and the Millbrig K-bentonite; this is the Spechts Ferry Member equivalent, or the *Stictoporella* bed. The conodont density within this unit is as high as 300 per 100 grams of rock. The higher beds here are within Sardeson's bed 4, the *Stictopora* or *Rhinidictya* beds, in which the conodont density is 30 per 100 grams. The type of unusual crinoid *Cremacrinus punctatus* Ulrich came from (bed 4) here.
- **1.1 meters (3.6 feet)** Carimona Limestone Member of the Platteville Formation. The Carimona Member consists of richly fossiliferous massive beds of limestone. *Isotelus* is common in this unit, suggesting that this represents a shallower facies than the other Platteville Formation members. The 10 centimeter thick Deicke K-bentonite is 0.36 meter (1.2 feet) above the base of the member here, always in a deep slot due to geologists digging for samples. The fall of the Deicke ash killed everything in eastern North America. Many species such as the conodont *Scyphiodus primus* terminate at this horizon. The conodont density is

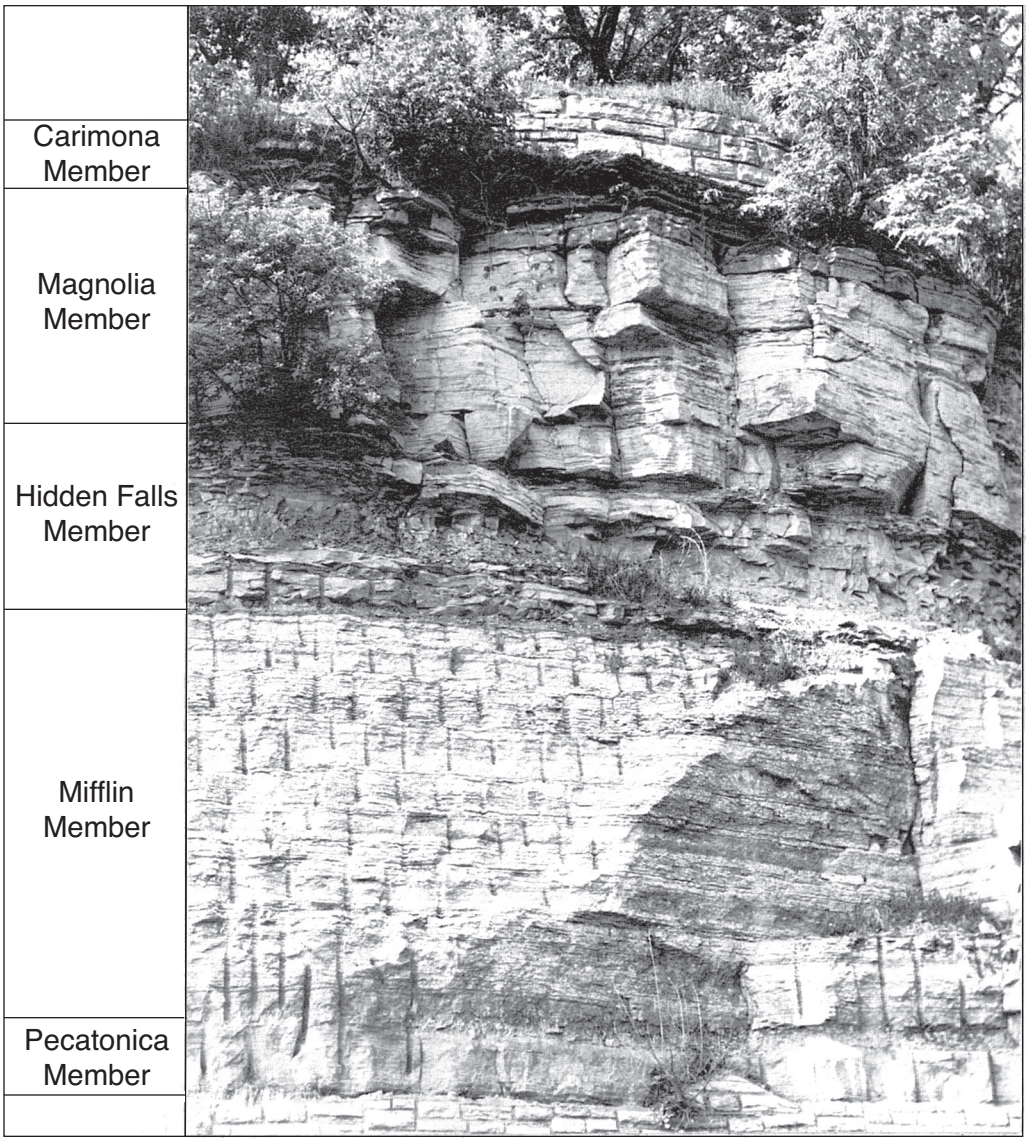


Figure 6.10. Former cliff exposure of the Platteville Formation below Fairview Riverside Hospital as seen in 1958. This is at the exact center of the Twin Cities basin. Minneapolis ground water flows along the top of the Deicke K-bentonite and forms a year round spring that flows down this face. As a result, the road below it is commonly covered with ice in winter. Today the cliff is covered with a 6-inch thick coat of moss. Photo by G.R. Ford.

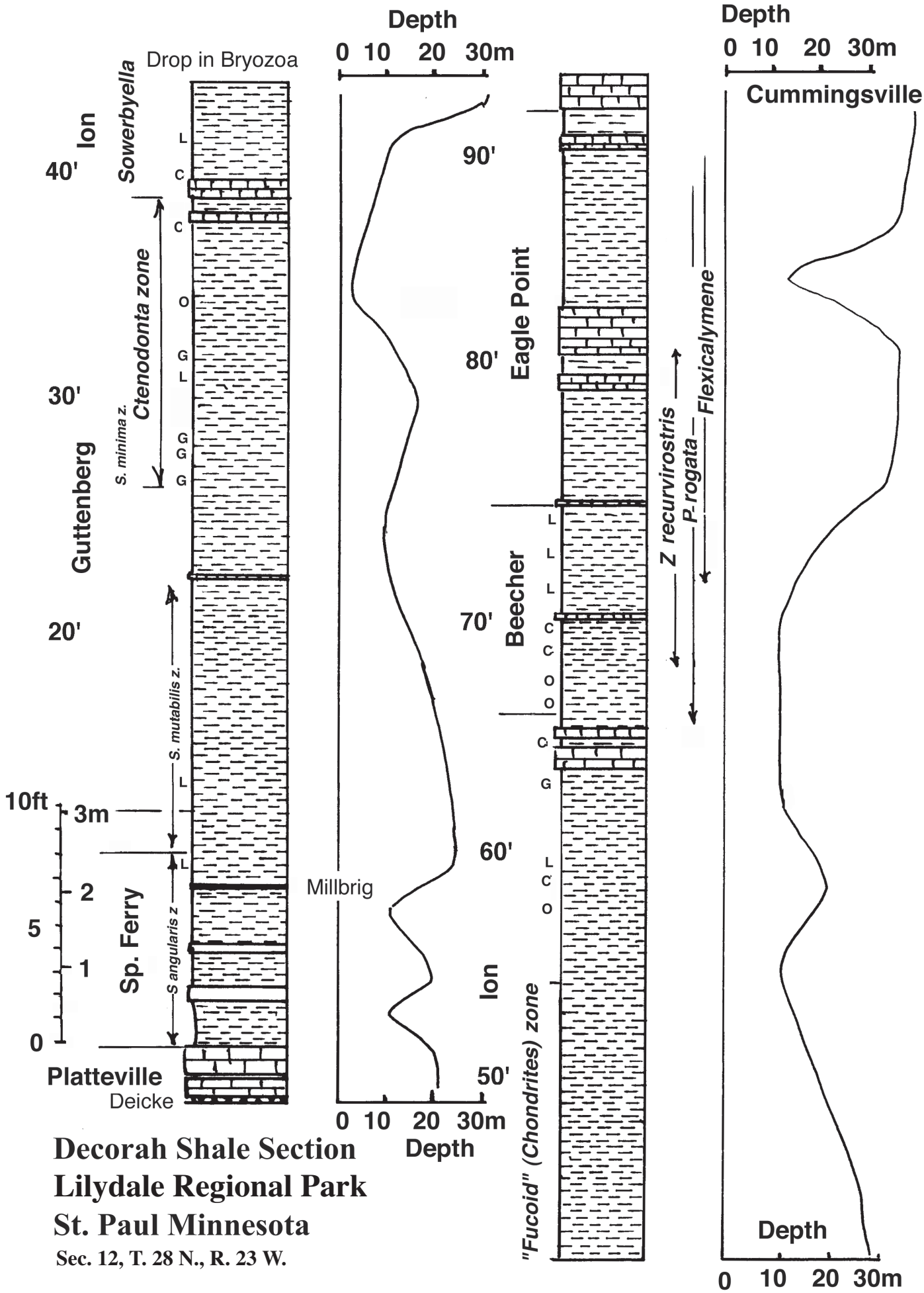
about 100 per 100 grams of rock in the limestone beds. As best as we can determine, based on trilobites and conodonts, this represents the boundary between the Turinian (old Blackriveran) and the Chatfieldian (old Trentonian) stages. Please note the interbedding of Decorah Shale and Carimona Member lithologies near the contact. There is a 0.2 meter (0.7 foot) bed of Carimona Member-type rock separated from the rest of the Carimona Member by 0.7 meter (2.3 feet) of typical Decorah Shale. In the Decorah Shale here (Stop 7-3), this bed is considered part of the Platteville Formation. Minnesota practice is to draw the boundary between the Platteville Formation and Decorah Shale at the point where the Decorah Shale interbeds are thinner than the Carimona Member interbeds. This means that the formation boundary is diachronous, but always within the interval between the Deicke and Millbrig K-bentonites. The base of the Carimona

Member is similarly diachronous, and depends on local lithofacies, depths, and biofacies. The lower Carimona Member is laterally equivalent to the Quimby's Mill Member of Illinois, as well as to parts of the Magnolia and McGregor Members of Minnesota; the upper Carimona member is laterally equivalent to the lower Spechts Ferry Member of Illinois.

- **2.44 meters (8 feet)** dolomite Magnolia Member of the Platteville Formation. This rock has a conchoidal fracture, is sublithographic in character, and contains a rich fauna dominated by the clams *Cyrtodonta, Ctenodonta,* and *Vanuxemia*, the snails *Clathrospira subconica, Maclurites,* and *Lophospira*, monoplacs, the coral *Lambeophyllum profundum,* and the brachiopods *Hesperorthis tricenaria* and *Rafinesquina minnesotensis*. All the fossils are preserved as dolomite rhomb-lined molds. The conodont density is about 10 per 100 grams of rock. Please do your collecting along the path to the north along the valley back to Shadow Falls. The Magnolia Member apparently correlates with the Nachusa Member of Illinois, although it resembles the Quimby's Mill Member more closely in lithology.
- **1.77 meters (5.8 feet)** Hidden Falls shaly dolomite Member of the Platteville Formation. This interval originated from an early uplift of the Transcontinental Arch just before the larger uplift that produced the Decorah Shale. The insoluble content of this member reaches as high as 45 percent. Its main use is for filling holes. A 2-centimeter-thick (1 inch) orange clay layer at the top is probably an unnamed K-bentonite. Fossil occurrences are patchy, but where present, very rich. Sardeson (unpub. data) mined out a spot in this unit in the former Johnson Street Quarry in Minneapolis (now filled with garbage and covered by Interstate 35W) that produced about 20 specimens of the starfish *Protopalaeaster narrawayi*, several specimens of the crinoid *Cremacrinus arctus,* edrioasteroids, cystoids, brachiopods, bryozoa, mollusks, and graptolites. The conodont density is about 10 per 100 grams of rock. The Hidden Falls Member apparently correlates with the Grand Detour Member of Illinois, which is also more shaly than adjacent rocks, although not as shaly as this. Sardeson's (1926) bed 2 includes the Hidden Falls and Magnolia Members.
- **2.62 meters (8.5 feet)** of the limestone Mifflin Member of the Platteville Formation. This crinkly bedded unit has much shell hash in it and spotty horizons of fossils. It resembles the McGregor Member south of the Twin City Basin, and is correlative with the Mifflin Member of Illinois. Many old foundations in the Twin Cities were built of quarry blocks of Mifflin Member limestone.
- **0.58 meter (1.9 feet)** Pecatonica Member of the Platteville Formation. Templeton and Willman (1963) named this local unit the Hennepin Member of the Pecatonica Formation, which seems unnecessary. They only used the name for parts of two other sections, and then for rocks quite unlike those here. The conodont density is about 30 per 100 grams of rock. This is the equivalent of part of the Glenwood Formation at Stop 7-3, and the equivalent of the top of the Pecatonica Member at Stops 7-3 and 7-5. Here it is a dolomite, with many small, rounded phosphate nodules of collophane, and some floating sand grains of St. Peter Sandstone type, clearly blown in by the wind. The basal Platteville Formation is commonly sandy. Sardeson's (1926) bed 1 includes the Pecatonica and Mifflin Members.
- **0.67 meter, (2.2 feet)** of blue-green Glenwood Formation, containing the *Chirognathus* zone conodonts and a few small macrofossils, underlain by 1.5 meters (5 feet) of shaly sand, with much bioturbation. This is a hybrid of Glenwood Formation and St. Peter Sandstone, usually included in the Glenwood Formation.
- **5.4 meters (18 feet)** of exposed typical upper St. Peter Sandstone, measured from the river level.

NEXT: After finishing this stop at noon, we will eat lunch here. Return to the vehicles by 1:00 PM. Proceed east on Summit Avenue to the St. Paul Cathedral and turn right onto Kellogg Avenue for 1 mile. Turn right on Wabasha Street, across from the St.

Figure 6.11. Stratigraphic section of Lilydale Park in St. Paul, Minnesota, on the south side of the Mississippi River in T. 28 N., R. 23 W., sec. 12. Modified from Rice (1987), with additions by Chris Schneider and Sloan. Bryozoan zones are from Karklins (1969), brachiopod zones are by Rice (1987), trilobite zones, depth inferences, and Galena Group member equivalents are by Sloan, *Ctenodonta* and *Chondrites* zones are by Schneider. Abbreviations: C—megarippled coquina, G—gypsum, L—lingulids, O—oolites, S—stictoporella, Z—Zygospira, P—Paucicrura.



Decorah Shale Section
Lilydale Regional Park
St. Paul Minnesota
Sec. 12, T. 28 N., R. 23 W.

Paul Radisson Hotel, and cross the bridge to Fillmore Street (0.4 mile). Turn right on Fillmore Street, which becomes Water Street. Continue on Water Street for 1.5 miles to the gate of Lilydale Regional Park, which is the former Twin City Brick Company quarry and clay pit. This locality is controlled by the St. Paul Parks Department and permission is required to collect fossils.

STOP 6-2

Lilydale Park

Location: T. 28 N., R. 23 W., sec. 12, SE (Fig. 6.11)

Description: This is the richest Decorah Shale collecting area in the Twin Cities and is also the thickest known section of Decorah Shale. Use caution in the big open pit, as this is where J.S. Templeton died in 1953 when he was hit by a slab of Cummingsville limestone that fell from the top of the pit.

Our collecting will be in a gully in the extreme SE 1/4 of the section, it exposes nearly the entire section of the 27.2 meters (89.2 feet) of the Decorah Shale and the bottom 0.45 meter (18 inches) of the Cummingsville Formation shaly limestone, with *Fisherites* (=*Receptaculites*). The lower part of the Cummingsville Formation type-section is here replaced by the upper part of the Decorah Shale. The Platteville Formation is also exposed and good collecting can be had in the loose blocks near the base of the section. By this time you should be able to recognize to which of the five Platteville Formation members a particular block belongs, or ask one of us. These rocks have been zoned in detail for ostracods (Swain and others, 1961; Swain and Cornell, 1987), bryozoa (Karklins, 1969), brachiopods (Rice, 1987), and trilobites. We anticipate that more zonation will be done here. Rice's samples were further analyzed by Chris Schneider and Sloan for molluscs, bryozoa, trilobites, correlations with the Galena Group members, and depth interpretation. The graphic section (Fig. 6.11) at the same scale as the others in this field trip description is based on Rice's data and summarizes all the information on the former brickyard. For a more detailed section see Rice (1987).

In addition to being the thickest Decorah Shale section known, it was deposited 1.4 times as fast as the Decorah Shale in Fillmore County (on the basis of Rice's 1987 Shaw analysis comparing this section with Weiss' [1953] samples from Fillmore County), and is the closest Decorah Shale section to the central Minnesota 1.6 b.y. Penokean Mountains, the Transcontinental Arch of many authors.

We will collect at this locality until 4:30 PM or when people get tired, whichever comes first. We will then return to the Radisson University Hotel.

FIELD TRIP STOPS

FIELD TRIP 7

DIRECTIONS: From the Radisson hotel, turn left onto Washington Avenue, follow several blocks to Huron Avenue and take a right. Follow the signs for Interstate 94 eastbound and continue on I-94 to St. Paul. Exit on U.S. Highway 52 south. Drive south on 52 approximately 60 miles to the intersection with State Highway 19. Turn east (left) for 0.75 mile to the intersection with Main Street in Cannon Falls and turn south on Goodhue County Road 24 (old Highway 52), drive 0.5 mile to the top of the hill, noting the St. Peter Sandstone outcrop by the side of the road. Turn left onto Goodhue County Road 25 for 0.4 mile to an old Platteville Formation quarry on the north side of the road. There are new exposures of St. Peter Sandstone, Glenwood Formation, and basal Platteville Formation along County Road 25 before the old quarry. There is an opening in the divided highway at a new road just past the old quarry where we will do a U turn and park on the north side so people can walk down the bike path to the quarry.

STOP 7-1

Cannon Falls (Fig. 7.1)

Description: Quarry and cuts along Goodhue County Road 25 on the south edge of Cannon Falls and southeastward toward White Rock. The base of the section is in a Platteville Formation quarry 0.4 mile from State Highway 20, and the top is in the Galena Group about 1.5 miles farther along the road. The Platteville Formation limestone here is quite thin (about 15 feet) and this appears to be the result of the loss of the Pecatonica Member and lower part of the McGregor Member. The Hidden Falls Member here is gradational with the underlying Glenwood Formation and very little typical McGregor Member rock is present. The Carimona Member is present in full thickness.

Decorah Shale—Basal 10 feet or so exposed above the quarry. No Millbrig K-bentonite is exposed in this area, it was apparently washed out. That is not surprising because Cannon Falls is on the sill separating the Twin Cities Basin from the rest of the Keweenawan rift valley (Fig. 6.2). The base of Decorah Shale is interbedded with the underlying Carimona Member limestone.

Platteville Formation, Carimona Member—6'11".

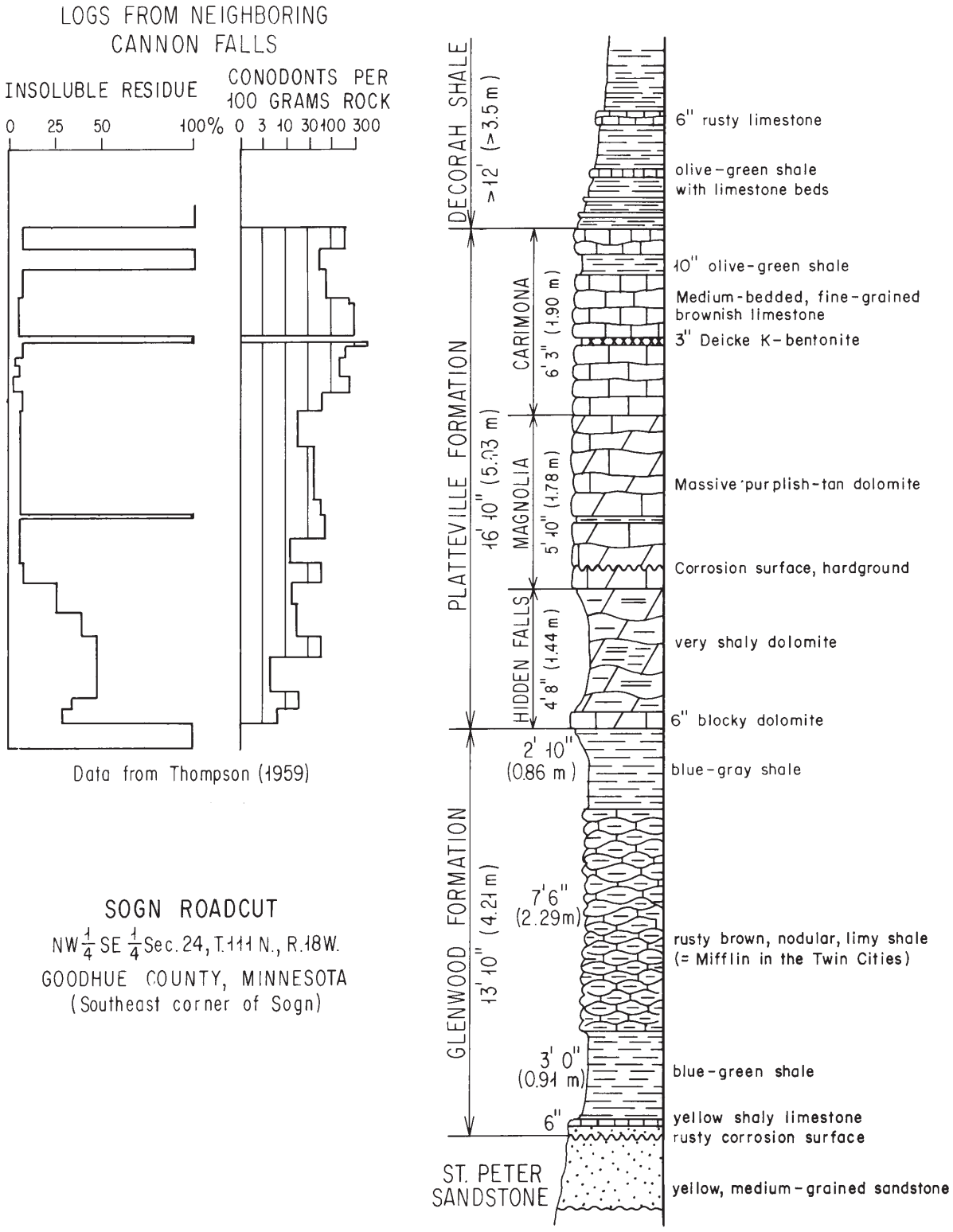


Figure 7.1. Detailed measured section of the Sogn outcrop (Stop 7-3) and the conodont density plot of the very similar Cannon Falls Quarry (Stop 7-1).

9" Carimona Member limestone bed contains *Pseudolingula evae* upright in vertical burrows on 6" spacing.

7" Shale

1"11" Carimona Member limestone in two beds, limestone and dolomitic limestone, yellowish brown to gray, medium bedded, hackly fracture. 10-inch limestone unit above bentonite contains large endoceroids and other mollusks.

5" Deicke K-bentonite, blue shale where unweathered, orange where weathered. Basal 1 inch is petroliferous shale (a quantitative plankton kill).

3'3" Carimona Member limestone as above. Thin fucoidal "oil shale" seams at base and 1 foot higher. Visible between 4.5 and 5.5 feet above base.

6'5" McGregor/Magnolia Member limestone, medium bedded, crinkly bedded. Corrosion surface one foot above base. The base of the quarry is in Hidden Falls Member/Glenwood Formation rocks.

Glenwood Formation and/or Hidden Falls Member. Argillaceous dolomitic limestone in the upper part, plastic and arenaceous green shale in the lower part; all poorly exposed and/or slumped. Much better exposed at Sogn (Stop 7-3).

If you continue along County Road 25 for 0.9 mile, there is a very fossiliferous exposure of Decorah Shale in the ditch on the right. This exhibits about 15 feet of relatively unoxidized Decorah Shale. The material consists of green-to-gray, highly fossiliferous shale. Brassy oolites are common. Thin limestone coquinas are present. The shale is exposed on both sides of the road, although the most productive collecting will be found on the south side. Rains have washed out much of the finer-grained material, leaving a lag concentration of fossils on the surface. The Decorah Shale is about 60 feet thick at Cannon Falls, Minnesota and thins to the south and southeast. The clastic material of this unit is derived from a nearby source area to the northwest associated with an uplift of the 1.6 b.y. Penokean Mountains, the local name for the Transcontinental Arch. The closest exposure of the Penokean Mountains is at St. Cloud, Minnesota, about 90 miles away. Bottom communities dominated by filter-feeders flourished in the warm, shallow, marine seas. The abundant macrofauna of the Decorah Shale is dominated by bryozoans and articulate brachiopods. Tabulate and horn corals, cephalopods, gastropods, pelecypods, and trilobites are also common. Microfauna includes abundant ostracodes and conodonts, with scolecodonts and chitinozoans also represented.

Galena Group—Cummingsville Formation and Prosser Limestone—Approximately 55 feet exposed below drift. Limestone and argillaceous limestone, yellowish or brownish gray, thin and nodular bedded, occasional lenticular or persistent medium beds, minor greenish shale units in the lower 12 feet. Fossiliferous throughout: *Ischadites* at base, *Vellamo* and *Platystrophia* 6 and 25 feet above base, "*Receptaculites,*" *Rafinesquina, Resserella*, and *Sowerbyella* abundant 8 to 11 feet above base.

NEXT: Return to Highway 20, turn left (south) and drive 1 mile to the intersection with Highway 52, continue south on 52 for 4.2 miles. At the top of the hill turn left onto Wagner Hill Way and follow to Wagner Quarry. Levorson and others' (1987) graphic section is a combination of the road cut on the east side of Highway 52 and the quarry.

STOP 7-2

Private property! Permission must be obtained before entering!

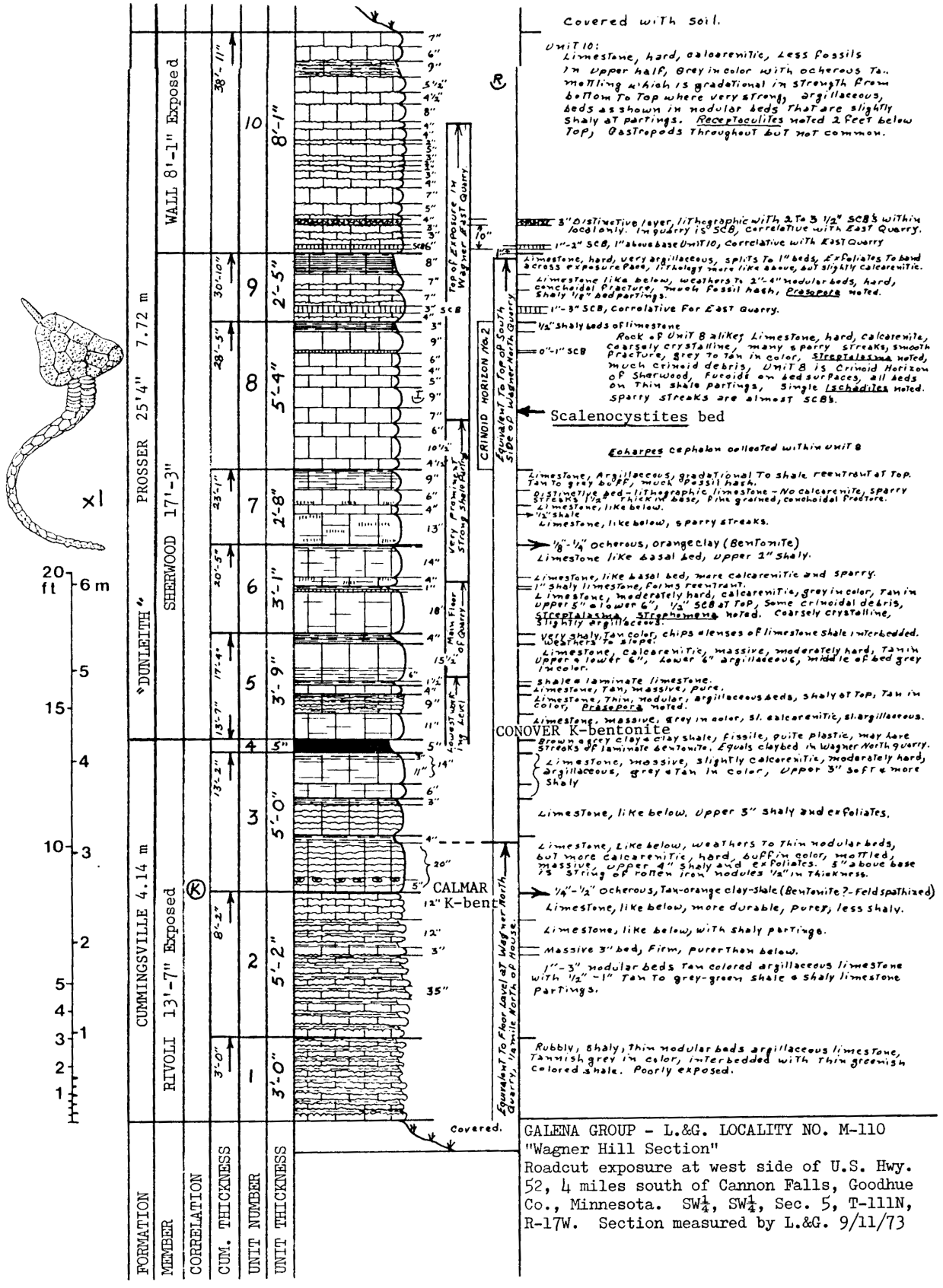
The Wagner Quarry cystoid bed (Fig. 7.2)

Description: Darren Anderson (phone 507-263-3526) is the current owner of the Wagner Quarry. The quarry is still active, but not on weekends. Mr. Anderson requested that collectors use hard hats and sign a hold harmless agreement. They have an even larger quarry about 4 miles to the south.

This stop is based on the measured section of Levorson and others (1987) and on the M.S. thesis of DesAutels (1978). The Wagner Quarry cystoid bed is a distinctive mass kill layer about 3 centimeters thick that occurs 3.43 meters (11 feet, 3 inches) above the base of the Prosser Limestone and the base of the Sherwood equivalent. The quarries and road cuts are in T. 111 N., R. 17 W., secs. 5 and 8, Leon township, 4 miles south of Cannon Falls, Goodhue County, Minnesota.

The kill resulted from a rapid mass burial in carbonate mud to a depth of 3 centimeters, perhaps as a result of a storm wave elsewhere reaching

Figure 7.2. The Wagner Hill road cut section, Stop 7-2. Measured by Cal Levorson and Art Gerk on U.S. Highway 52, with minor additions by Sloan. The upper part is identical to the section in the east Wagner Quarry, where the *Scalenocystites* bed in Levorson and Gerk's (1972) crinoid bed #2 in the Sherwood Member of the Prosser Limestone was collected.



GALENA GROUP - L.&G. LOCALITY NO. M-110
"Wagner Hill Section"
Roadcut exposure at west side of U.S. Hwy. 52, 4 miles south of Cannon Falls, Goodhue Co., Minnesota. SW$\frac{1}{4}$, SW$\frac{1}{4}$, Sec. 5, T-111N, R-17W. Section measured by L.&G. 9/11/73

Table 7.1. Species list rank ordered by abundance.

Platystrophia "biforata"	384
Zygospira recurvirostris	257
Prasopora insularis	182
Pleurocystites squamosus	94
Strophomena trilobata	84
Plaesiomys subquadrata	61
Scalenocystites strimplei	58
Streptelasma corniculum	54
Dolichoharpes ottowaensis	52
Strophomena "incurvata"	41
Stictopora mutabilis	33
Illaenus americanus	32
Maclurites crassa	32
Fusispira inflata	28
Nematopora conferta	21
Raphistomina lapicida	20
Eomonorachus intermedius	19
Trochonema umbilicatum	19
Michelinoceras beltrami	19
Paucicrura rogata	18
Metaspyroceras lesueuri	14
Hormotoma gracilis	13
Zygospira modesta	11
Ischadites iowensis	10
Hudsonaster narrawayi	10
Ceraurus pleurexanthemus	10
Isotelus gigas	5
Glyptocrinus sp.	5
Cremacrinus sp.	4
Ophiletina angularis	3
Ceraurinus cf. *icarus*	1
Hormotoma subangulata	1
Cyclocystoides halli	1
TOTAL 33 species	1,596

bottom. DesAutels (1978) identified, measured, and counted the total macrofossil assemblage of 1,596 individuals of 33 species from a total of 6.86 square meters (74 square feet) of this bed (Table 7.1). The slabs were sold after analysis. Preparation was by an S.S. White Airbrasive machine. This represents the largest unbiased sample of a Prosser Limestone fauna known to the authors, although it does not show the full Late Ordovician fauna in Minnesota: some 330 species are known (Webers, 1972). The present fauna is in the *Ischadites iowensis* zone and consists of 55 percent brachiopods, 15 percent bryozoa, 10 percent echinoderms, 7 percent gastropods, 6 percent trilobites, 2 percent cephalopods and 3 percent corals. The benthic species are slightly dwarfed, reaching on the average about 60 percent of their maximum size elsewhere. Bivalves are totally absent and no conodonts were found.

The fossils were collected in the more eastern of Mr. Anderson's two quarries, in T. 111 N., R. 17 W., sec. 8, NE, NW. Levorson and Gerk (unpub. data) have measured a detailed section, their "M-110," the Wagner Hill section, along Highway 52, in T. 111 N., R. 17 W., sec. 5, SW, SW. The Decorah Shale in the vicinity of Cannon Falls is about 18.5 meters (61 feet) thick, and the lower part of the type Cummingsville Formation has increased in mud content to become the upper part of the Decorah Shale. Similarly, the lower part of the type Prosser Limestone, the equivalent of the Rivoli Member of the Illinois and Iowa classification, has greatly increased in mud content and must be classed as part of the Cummingsville Formation here. The horizon of the contact has shifted upward about 6 meters (20 feet). These exposures are the most northern outcrops yet studied of the Prosser Limestone, which does not extend more than 2 miles north of here. The horizon is high in the Shermanian stage; the estimated age is 450.0 Ma, and it is part of Levorson and Gerk's (1972) crinoid horizon no. 2.

NEXT: Continue south on Highway 52 3.5 miles to Goodhue County Road 9, turn west (right) and follow 3.5 miles to the town of Sogn, turn left (south) on Goodhue County Road 14 and continue 0.1 mile to a road cut.

STOP 7-3

Sogn road cut

Location: T. 111 N., R. 17 W., sec. 24, NE, SE (Fig. 7.1)

Description: The Sogn road cut shows an entire section of Glenwood Formation rocks and the thinnest section of the Platteville Formation known. It is basically similar to Stop 7-1 except that the St. Peter Sandstone and Glenwood Formation are visible here. The lower three members of the Platteville Formation in the Twin Cities are here represented by part of the Glenwood Formation. As is usually the case, the beds of Carimona Member rocks just above and below the Deicke K-bentonite are particularly rich in conodonts, here about 300 conodonts per

100 grams of limestone. The Deicke interval is the upper limit of the conodont *Scyphiodus primus Stauffer*, the type and other rare specimens from the overlying Decorah Shale appear all to be reworked. As closely as we can tell, the Deicke interval is also the Turinian–Chatfieldian stage boundary; it is a major extinction event. Seven miles north at the Cannon Falls quarry, a nitrogen analysis of the 1-centimeter-thick petroliferous shale at the base of the bentonite indicates a quantitative plankton kill. The Millbrig K-bentonite (Spechts Ferry Member) from the lower Decorah Shale is always missing in this county, though it is usually present throughout the upper Mississippi River valley.

We have not sampled it here, but try washing the Glenwood Formation rocks for the *Chirognathus* conodont fauna.

NEXT: Retrace our path 3.5 miles to Highway 52. Hader, Minnesota is a town with an excellent Prosser Limestone quarry just east of the highway; however, we will not be stopping there today. Travel southeast 8.5 miles to a steep grade descending into the valley of the North fork of the Zumbro River over slopes of Prosser Limestone, Cummingsville Formation, Decorah Shale, Platteville Formation, Glenwood Formation, and St. Peter Sandstone. At the base of the grade turn west (right) onto a gravel road just north of the bridge over the North fork of the Zumbro River. The gravel road is Sherwood Trail, opposite Goodhue County Road 7, which turns east off Highway 52 at the same location. It is marked by a sign for the "Shades of Sherwood Campground." Continue 1.6 miles west to a newly widened road to the right (just before a small house on the right and a campground on the left).

STOP 7-4

Private property! Permission must be obtained before entering!

Zumbrota Clay Pit

Location: T. 110 N., R. 16 W., sec. 21 (Fig. 7.3)

Description: This pit was formerly mined for brick clay and is now abandoned. The landowner is Greg Peterson and we appreciate his permission to visit the outcrop. One of Mike Middleton's former students, Chad Johnson, measured the section at Zumbrota when it was better exposed (Chad Johnson, unpub. data).

The access road to the pit climbs up out of the valley through poor exposures of the St. Peter Sandstone, Glenwood Formation, and Platteville Formation (25.5 feet thick here) before reaching the Decorah Shale. The Carimona Member of the Platteville Formation has lingulids in life position here as it has in Cannon Falls and at a quarry outside of River Falls, Wisconsin.

The Deicke K-bentonite formerly was exposed 2.5 feet below the top of the Platteville Formation. There are the usual two beds of Carimona Member type rock in the lower Decorah Shale, the lowest one lifted 2.5 feet above the top of the Platteville Formation. The Millbrig K-bentonite was formerly visible in the ditch alongside the access road, 4.2 feet above the Platteville Formation contact. Unfortunately, the road was widened in 2004, covering the limited exposures of the lower 15 feet of the Decorah Shale. However, that work did create a new road cut to the east of the clay pit, where 21 feet of the upper Decorah Shale is now exposed (Fig. 7.3). The presence of the Millbrig K-bentonite implies we are now off the sill forming the southeast edge of the Twin Cities Basin (Fig. 6.2).

The old clay pit exposes 43 feet of the Decorah Shale from 16 to 59 feet above the base. It is dominantly gray-green fissile shale, but also includes thin, lenticular, coquinoid limestones throughout. Body fossils can be found in the shale, and a burrowed zone occurs on a ledge halfway up the slope. An isolated *Isorophus*-like edrioasteroid was found in the shale near the base of the exposures.

The limestones are particularly interesting. Three horizons of megaripples occur in the pit. They are composed of convex-upward lenses of shell-hash limestone with an average wavelength of 1.6 feet and a thickness of 1 to 2 inches. Many are size-sorted, with the largest fossils in the troughs, and elongate fossils oriented across the crests. Fossils found in these as well as the other limestones include abundant bryozoans, articulate brachiopods, gastropods, and crinoids, with trilobites and cephalopods fairly common, and other groups represented in lower numbers. The most unusual finds from these limestones have been articulated crinoids and cystoids. Slabs including well-preserved *Paleocrinus angulatus*, *Periglyptocrinus spinuliferus*, *Cremacrinus punctatus and Pycnocrinus,* as well as the cystoid *Pleurocystites,* have been found over the years. Apparently the sea floor was subjected to occasional winnowing by currents or storm waves, sometimes producing megaripples, sometimes burying crinoids under piles of shell hash.

NEXT: Follow the access road further along to the left around the hill to encounter another exposure. Here a level terrace was bulldozed to expose fossils

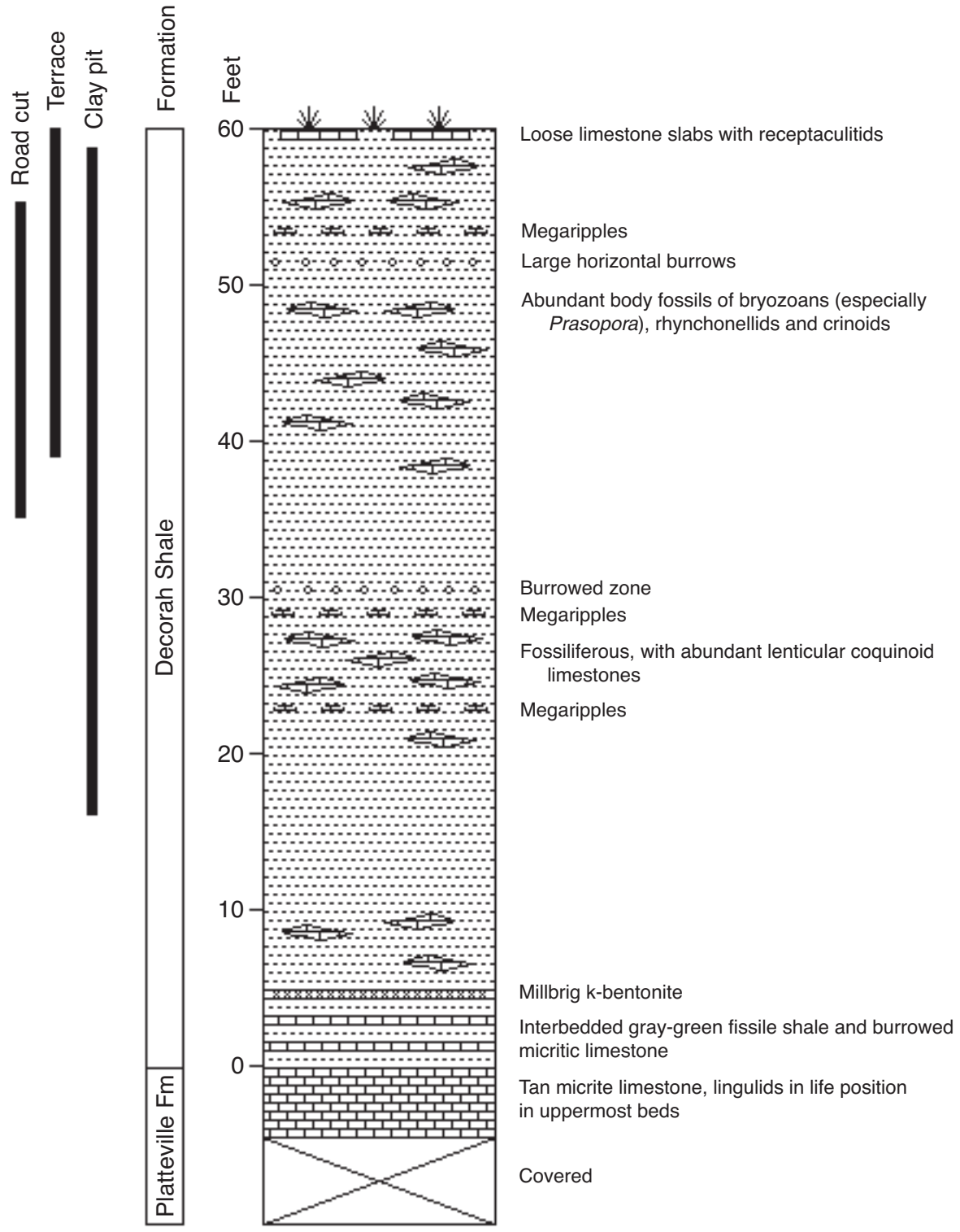


Figure 7.3. Composite stratigraphic section of the Barrclay pit in the Decorah Shale near Zumbrota, T. 110 N., R. 6 W., sec. 21. Exposed intervals are shown at left. Measured by Johnson (unpub. data, 1995) and Middleton (unpub. data, 2004).

for collectors staying at the campground below. The top 21 feet of the Decorah Shale is exposed, capped by a bulldozed pile including fossiliferous limestone slabs containing receptaculitids, presumably representing basal Cummingsville Formation strata. Over the years, the terrace has eroded, and a fossil lag has accumulated on the surface. Abundant rhynchonellid brachiopods, crinoid columnals, and bryozoans including *Prasopora* litter the surface. An interesting zone of large, horizontally-branching *Chondrites* burrows can be seen just above the lag, at the 51-foot level within the formation.

Mining history

This site was originally called Barrclay, for Ed Barr, who bought the property in 1911. He realized the Decorah Shale might have economic value as a source of clay, lying so close to the Chicago, Milwaukee, St. Paul and Pacific Railroad, which ran up the Zumbro River Valley on the south bank. The Barr Clay Products Company began operations in 1911, building a plant on the site for the purpose of manufacturing brick and tile. A number of buildings were involved in the production, and three stacks surrounded by four kilns each were laid out to the west of the plant. Clay excavated from the pit above the plant by a steam shovel on rails was carried by narrow gauge railroad down to a grinding and screening house, the remains of which can still be seen along the access road. From there it went through the production plant to the kilns. Coal for the kilns was brought in by rail, and a series of fans in underground tunnels provided the draft. After firing, the bricks and tile were taken to a spur line of the railroad built right up to the site. At the peak of production around 1920, the plant employed 100 workers, mostly Norwegian immigrants. Eventually Barrclay boasted a boarding house, general store, and a railroad depot, along with lodging for the workers. Operations continued sporadically under various owners until 1938, when mining was shut down permanently. Most of the buildings were removed over the years, only two of the stacks and the clay pit remain to attest to Barrclay's former mining history (information provided by Greg Peterson, unpub. data, 2004).

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FIELD TRIP 8

Saturday, May 21

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

Leaders

Larry Zanko and John Heine, Natural Resources Research Institute
Dale Setterholm, Minnesota Geological Survey

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:

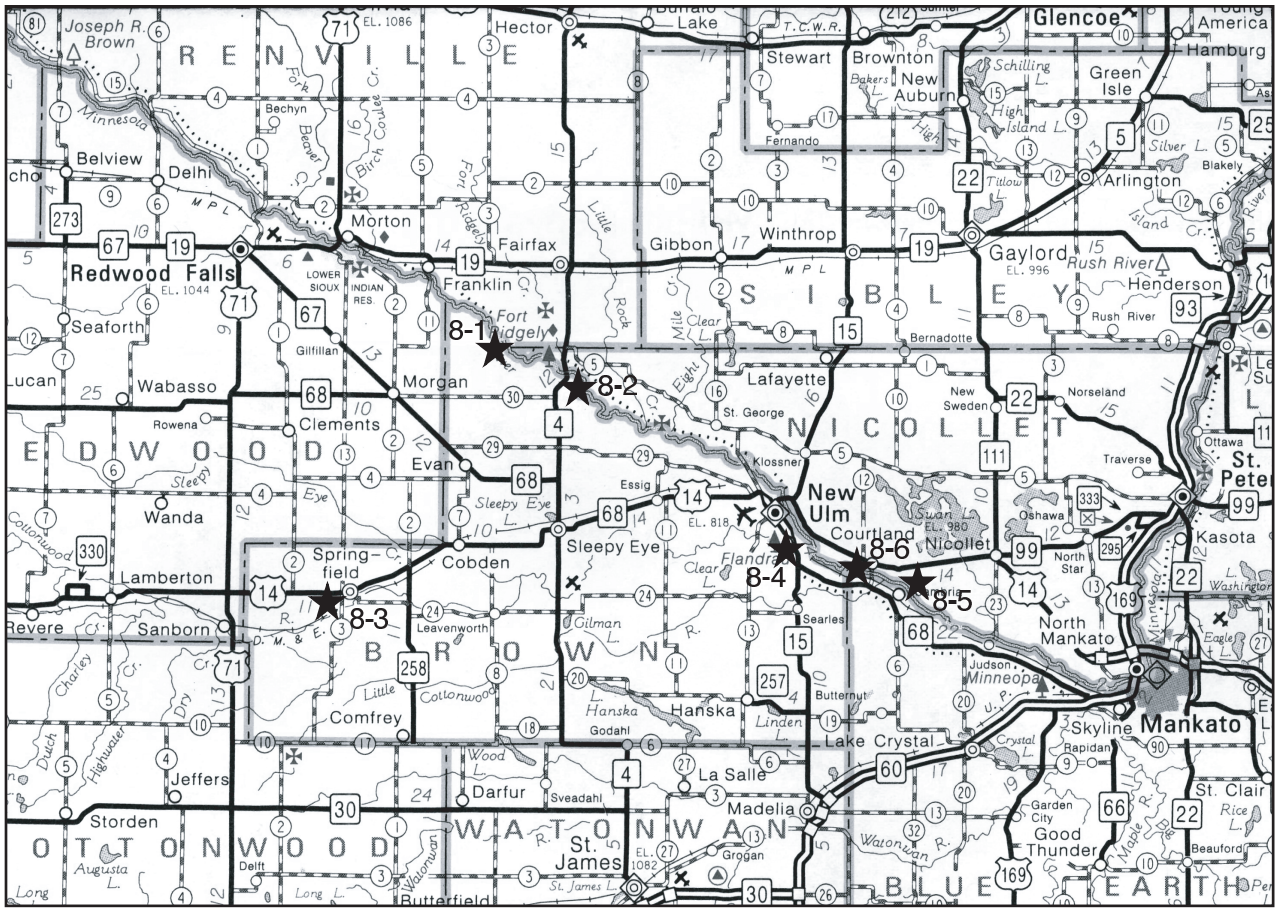


Figure 8.1. Map of field trip stops.

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

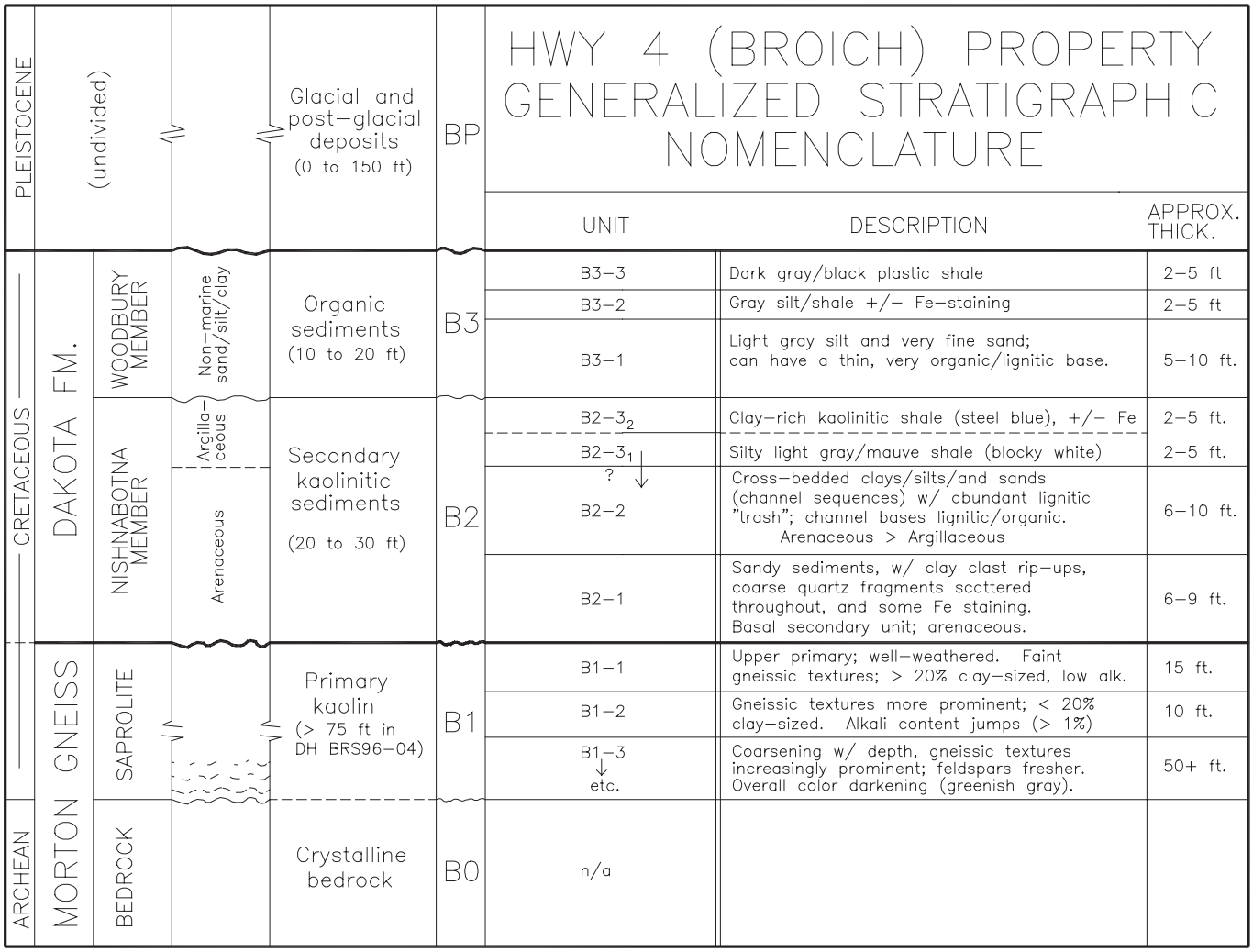


Figure 8.3. Stratigraphic nomenclature for the Highway 4 Clay Mine.

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).

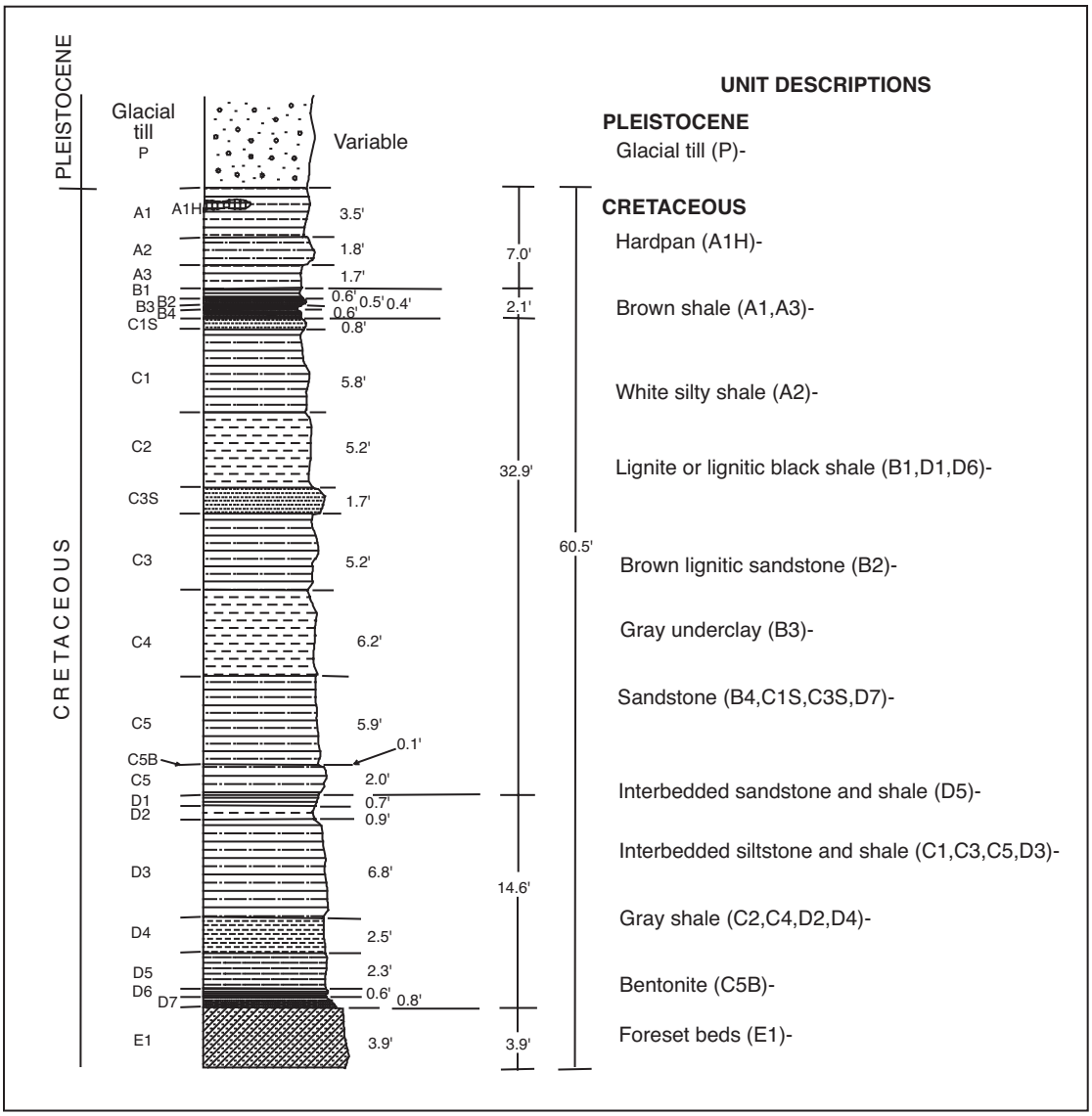


Figure 8.5. Stratigraphic column for the Springfield Mine exposure.

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

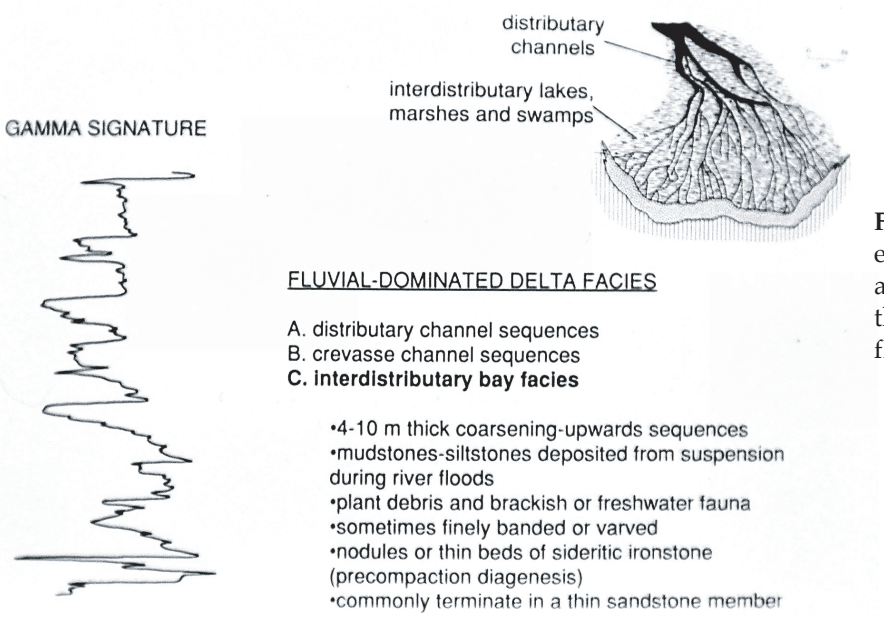


Figure 8.6. Depositional environment, sedimentary features, and natural gamma expression of the interdistributary bay facies of a fluvial-dominated delta system.

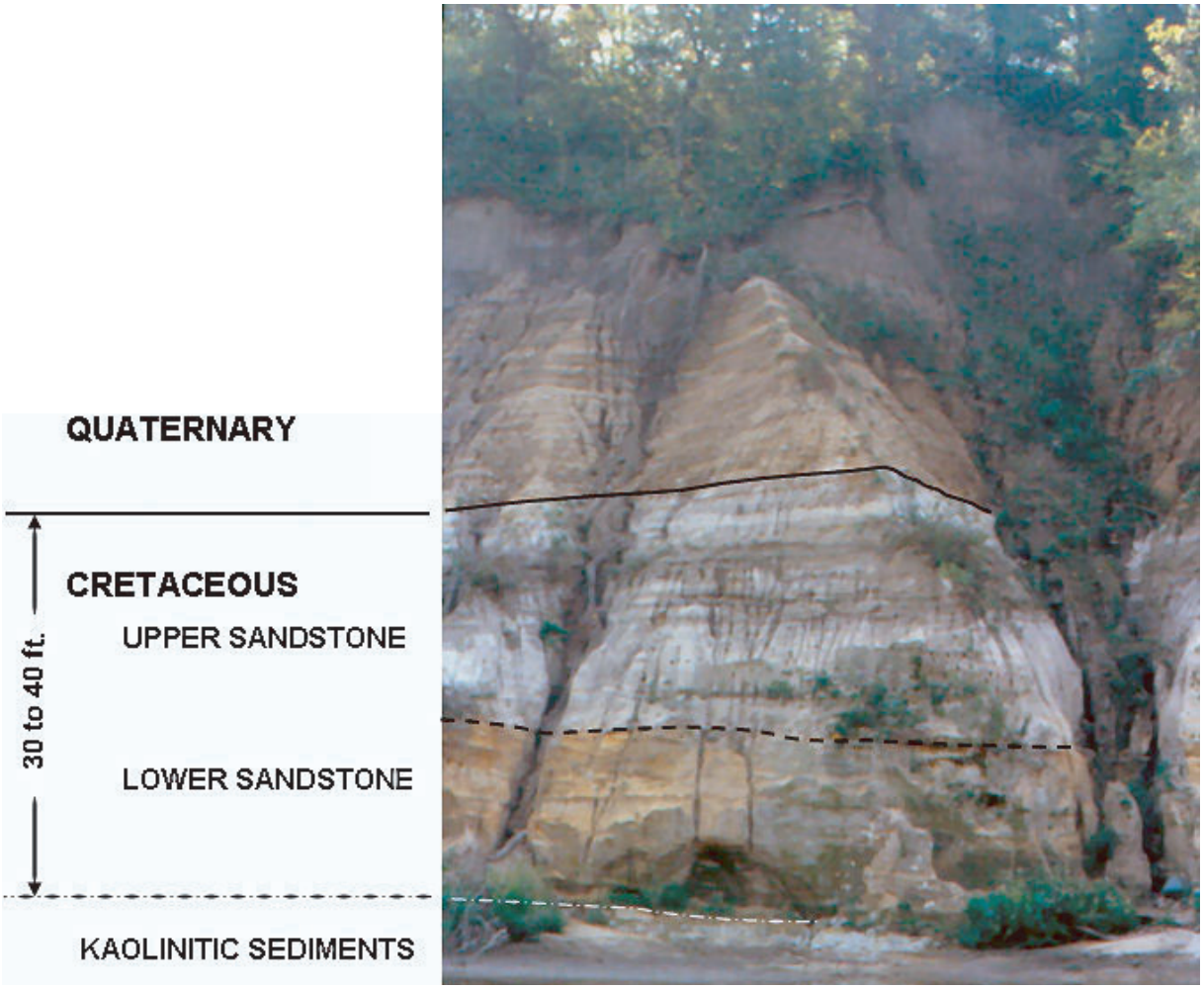


Figure 8.7. Typical exposure of Cretaceous strata in the New Ulm area.

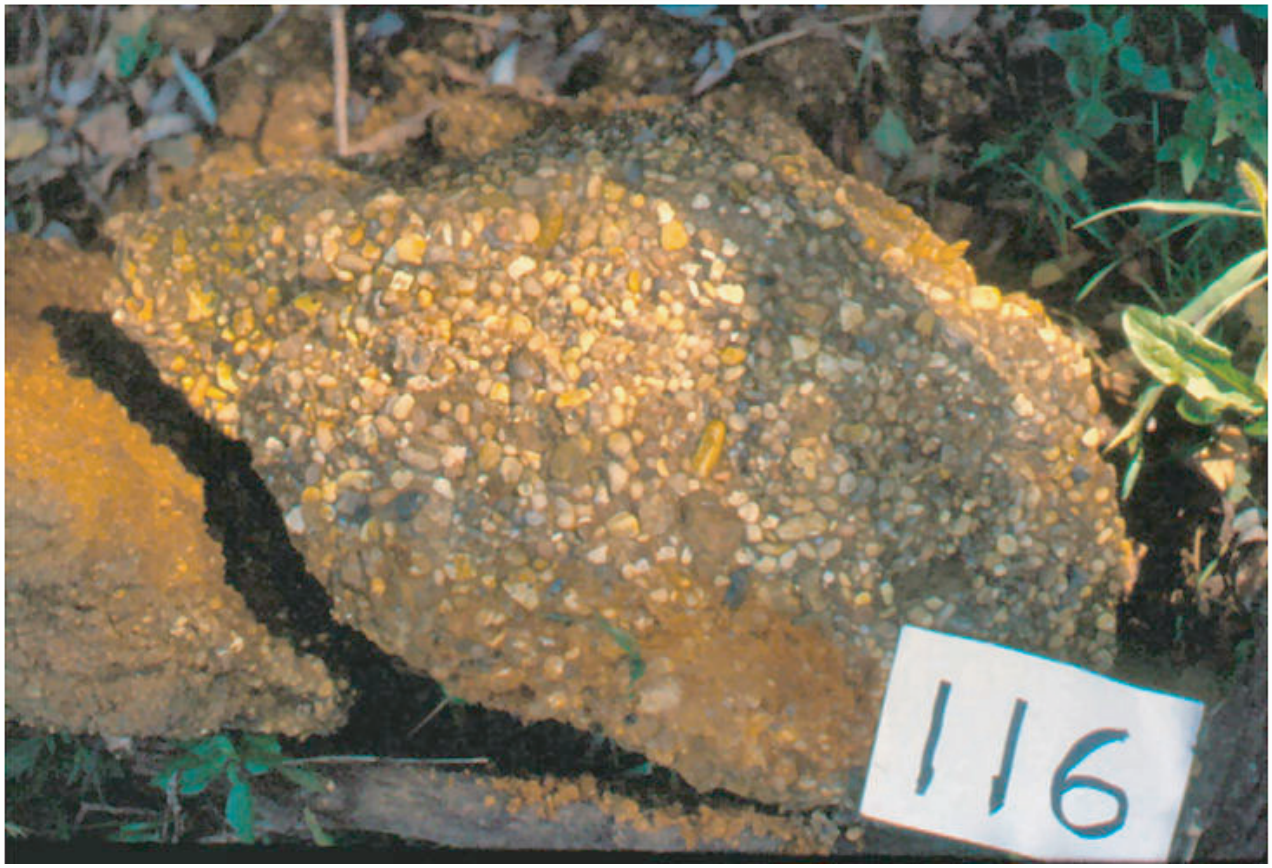


Figure 8.8. The conglomerate facies of the Ostrander Member of the Windrow Formation.

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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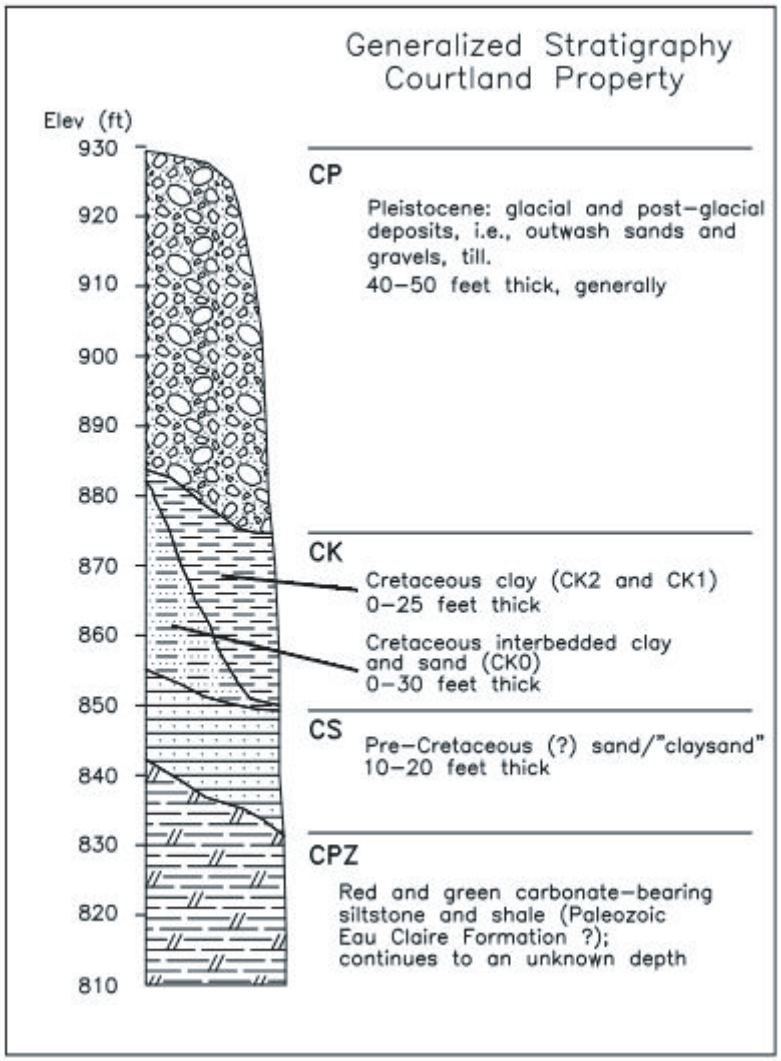


Figure 8.9. Generalized stratigraphy for the Courtland property.

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FIELD TRIP 9

Saturday, May 21 – Sunday, May 22

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

Leaders

Dean M. Peterson, Natural Resources Research Institute
Mark A. Jirsa, Minnesota Geological Survey
George J. Hudak, University of Wisconsin Oshkosh

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.

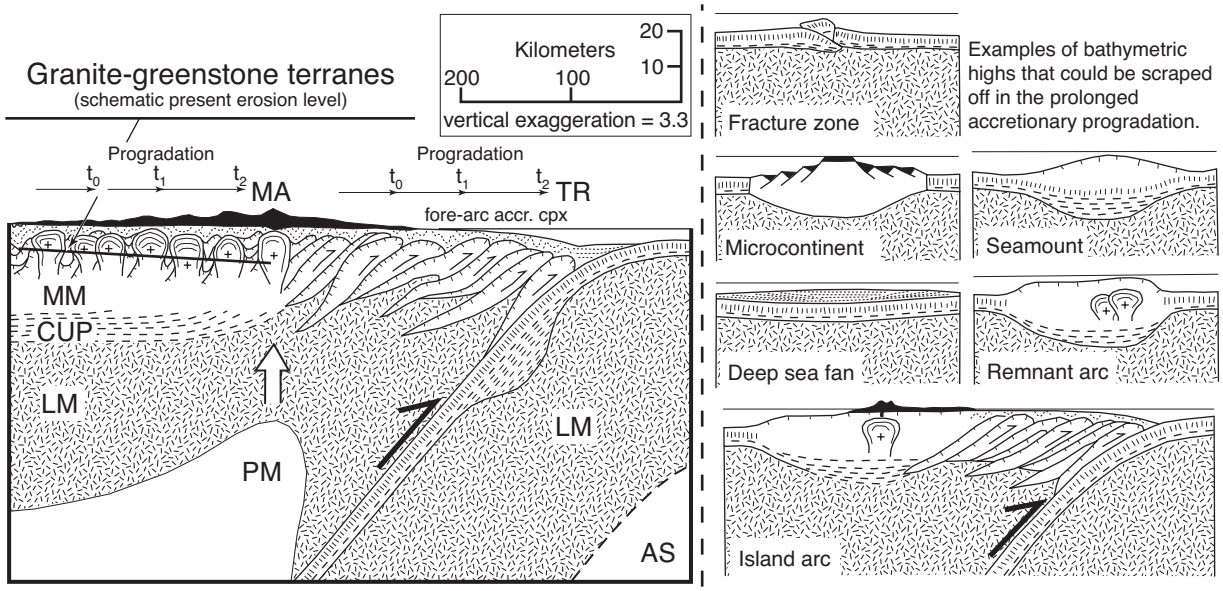


Figure 9.1. Prograding arc-trench model for the generation of granite-greenstone terranes. On the left is an idealized cross-section of an arc-trench system: AS, asthenosphere; CUP, cumulate underplating; LM, lithospheric mantle; MA, magmatic arc; MM, magma melting and mixing; PM, zone of partial melting; 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.

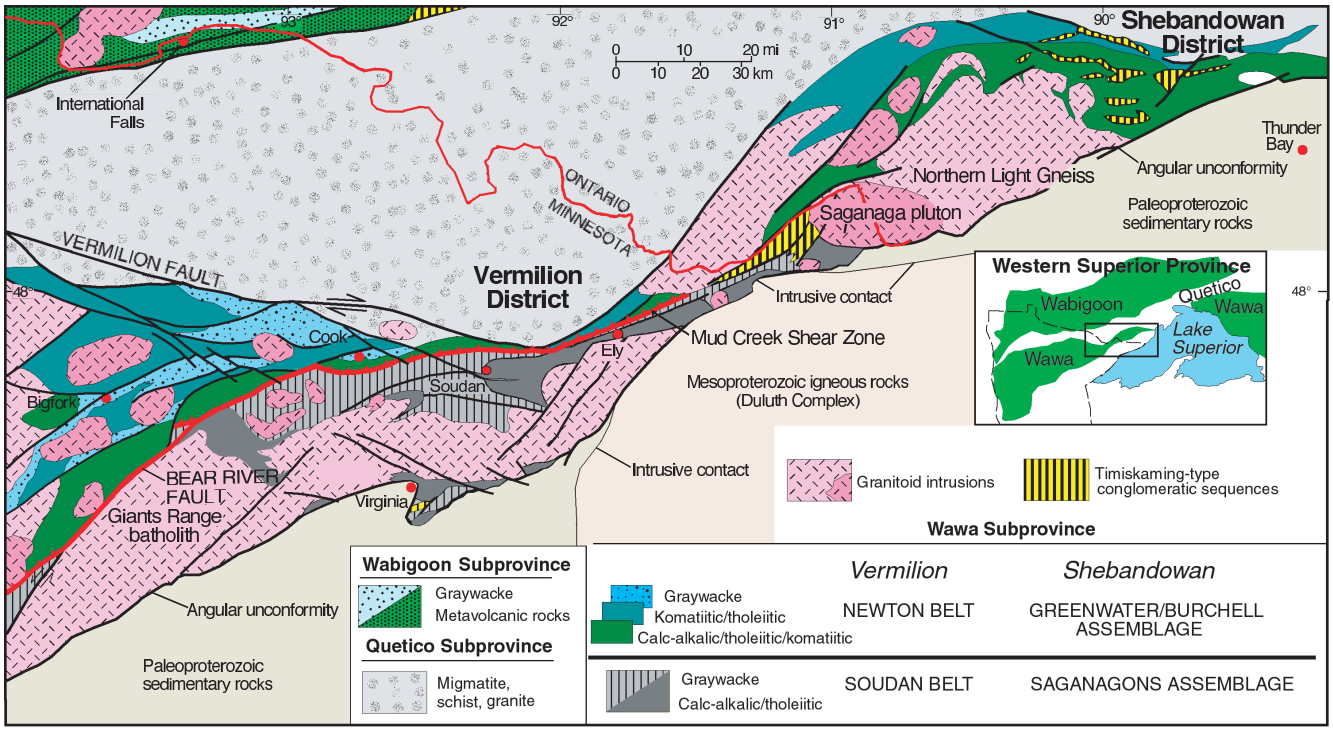


Figure 9.2. Simplified correlation map of Neoarchean assemblages across the U.S.–Canada border (modified from Peterson and others, 2001). Inset illustrates major subprovinces of the southwestern Superior Province.

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, $Fe_2O_{3(s)} * H_2O$, (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:

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

and

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

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:

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

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

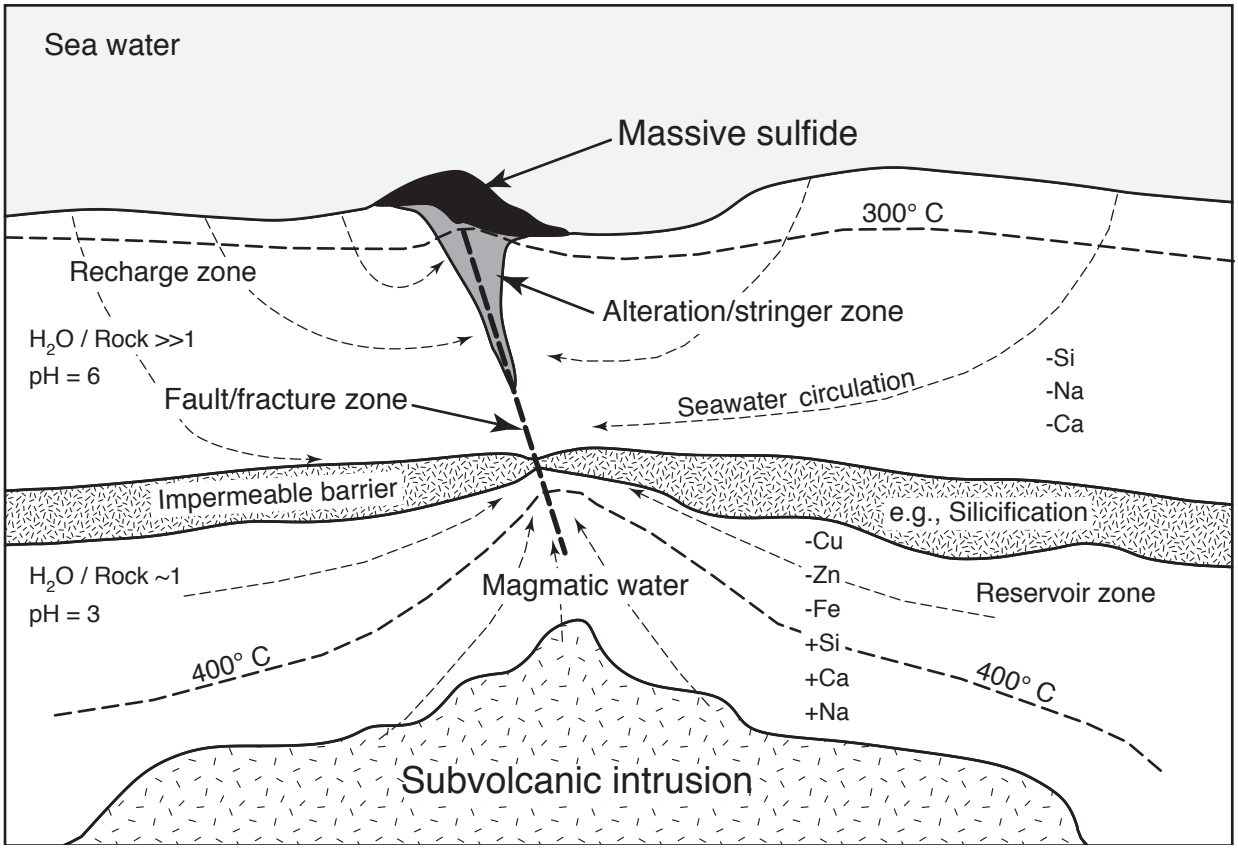


Figure 9.3. Simplified schematic model of a convective hydrothermal system associated with the formation of Noranda-type (Morton and Franklin, 1987) or lava-flow dominated-type (Gibson and others, 1999) volcanic-hosted massive sulfide deposits (modified from Franklin, 1996).

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

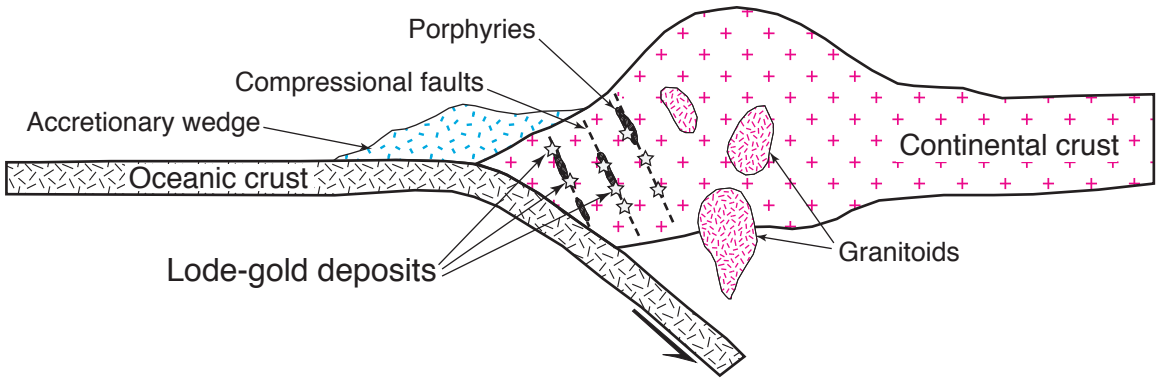


Figure 9.4. Generalized tectonic model for the formation of mesothermal gold deposits, after Groves and others (2000).

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

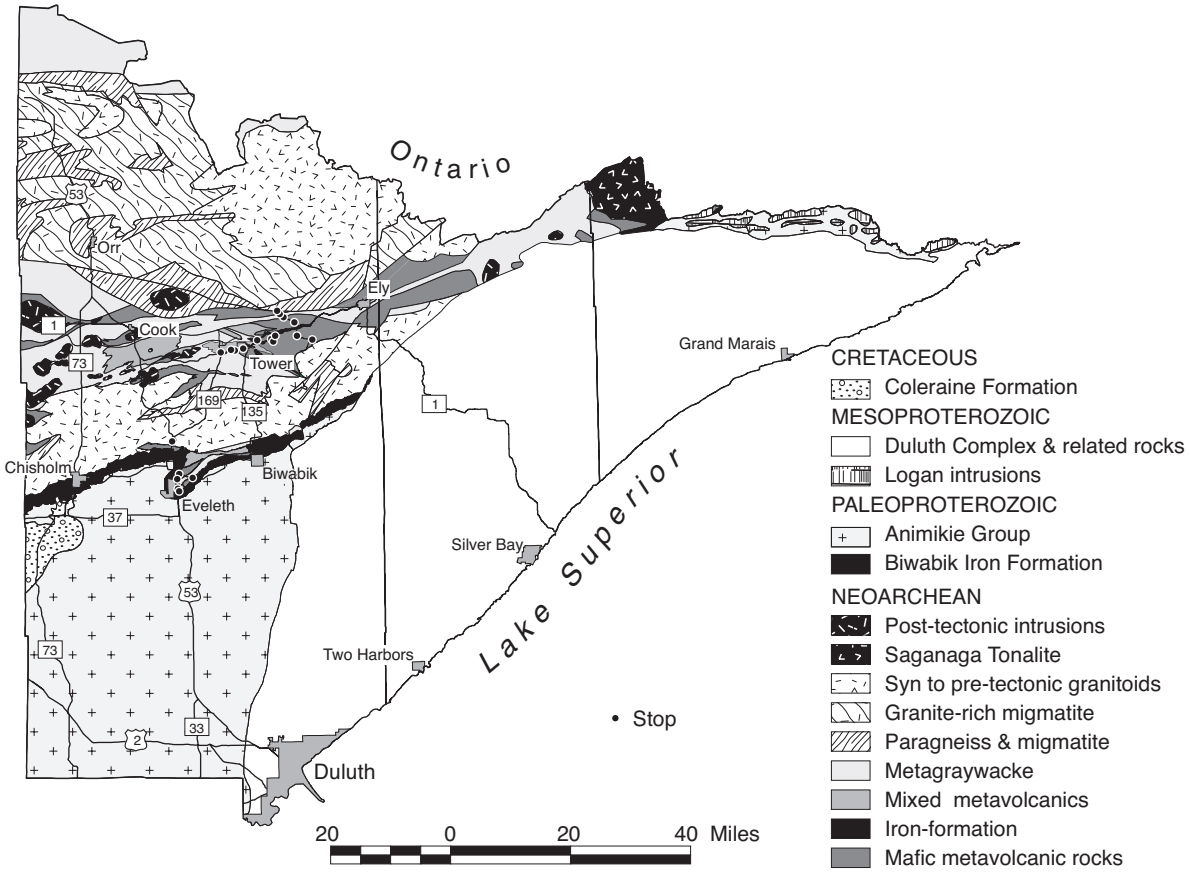


Figure 9.5. Generalized map depicting the major roads, field trip stops, and simplified regional geology of the Arrowhead region of northeastern Minnesota.

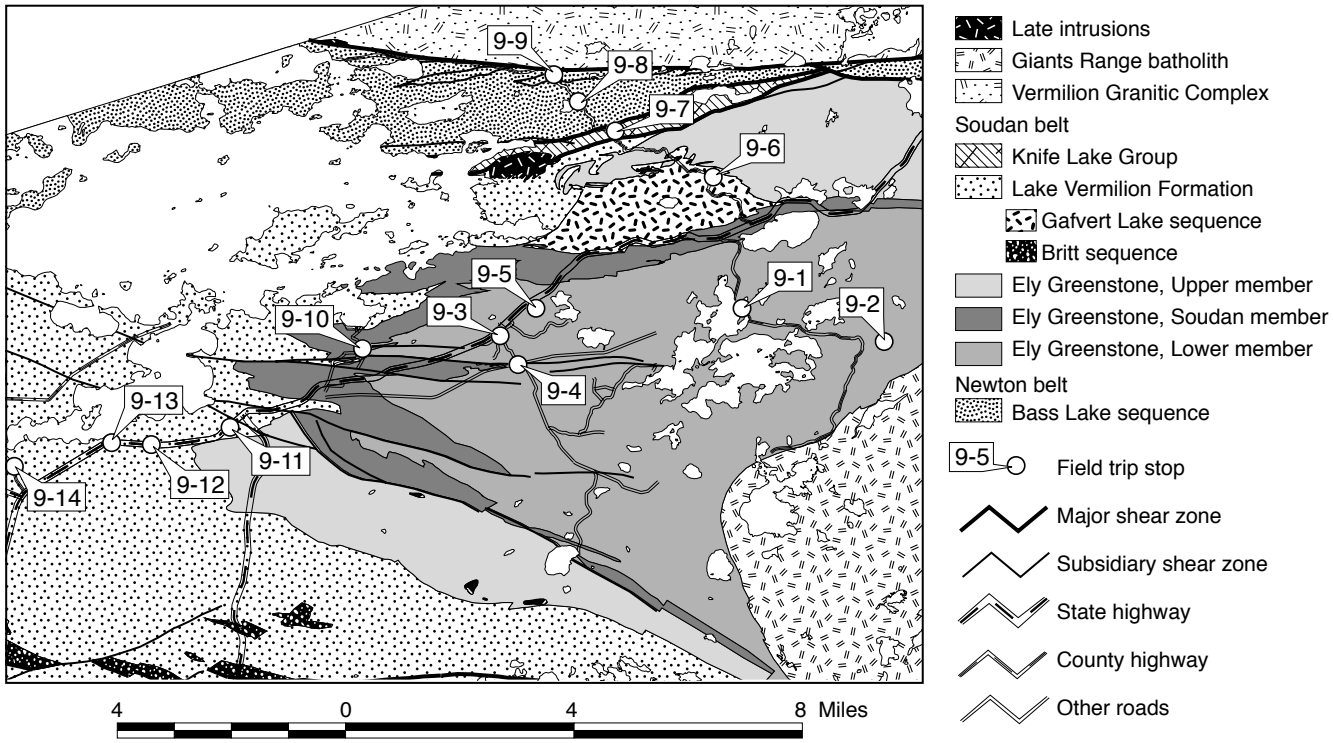
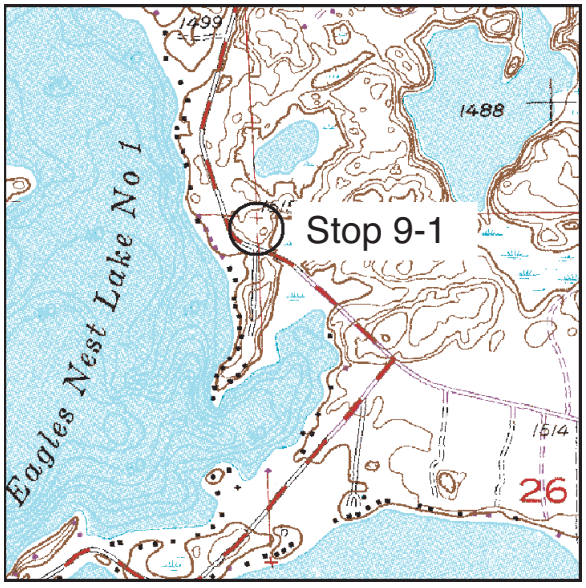


Figure 9.6. Simplified geologic map of the field trip stops in the Vermilion district. Geology modified from Peterson and Jirsa (1999).

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.

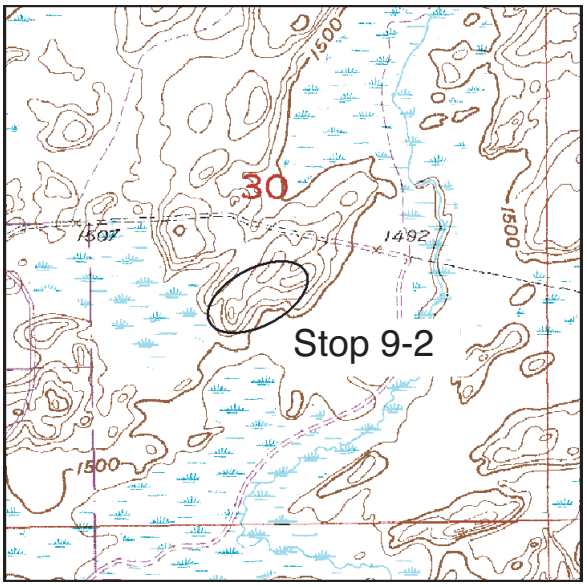
NEXT: Continue on County Road 128 southeast for approximately 2.2 miles to the Purvis Forest Management Road. Turn left and continue about 0.4 mile to a logging road on the right side (south). Follow the logging road approximately 0.15 mile to a series of outcrops.

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.

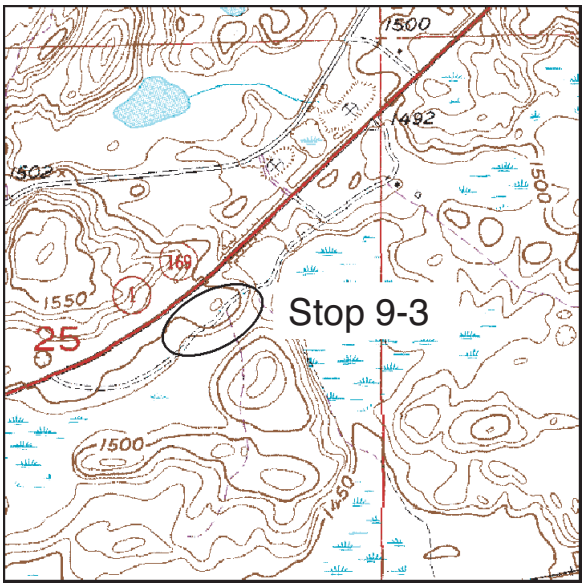
Studies of ancient volcanic-hosted massive sulfide deposits have documented that the deposits commonly occur in depressions on the paleo-seafloor (third-order basins), but modern deposits on the seafloor are found on high-standing structures, such as ridges. These differences are probably more apparent 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.

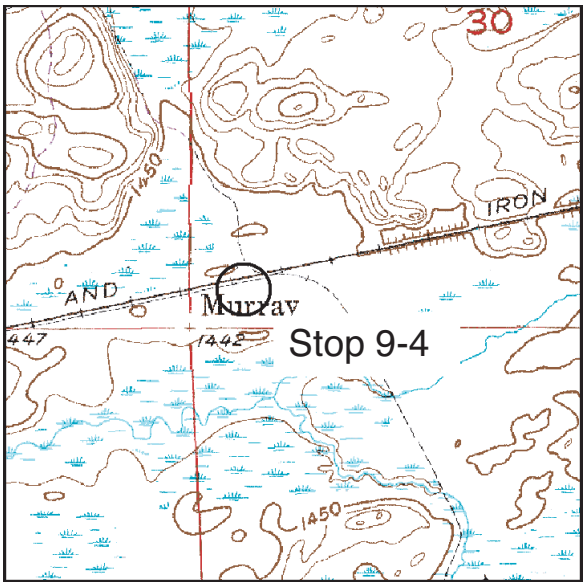
NEXT: Continue south along Murray Forest Management Road for approximately 0.6 mile to the junction with the old DM & IR rail line. Walk west along the rail line for 50 meters to the outcrops on the north and south side of the rail line.

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.

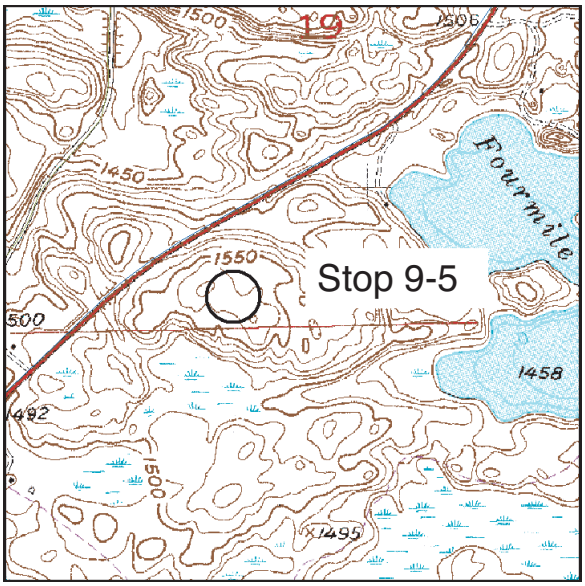
Lineation plunge	Strike slip	Dip slip	Net slip
71°	4.5	13.1	13.8
71°	3.0	8.7	9.2

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

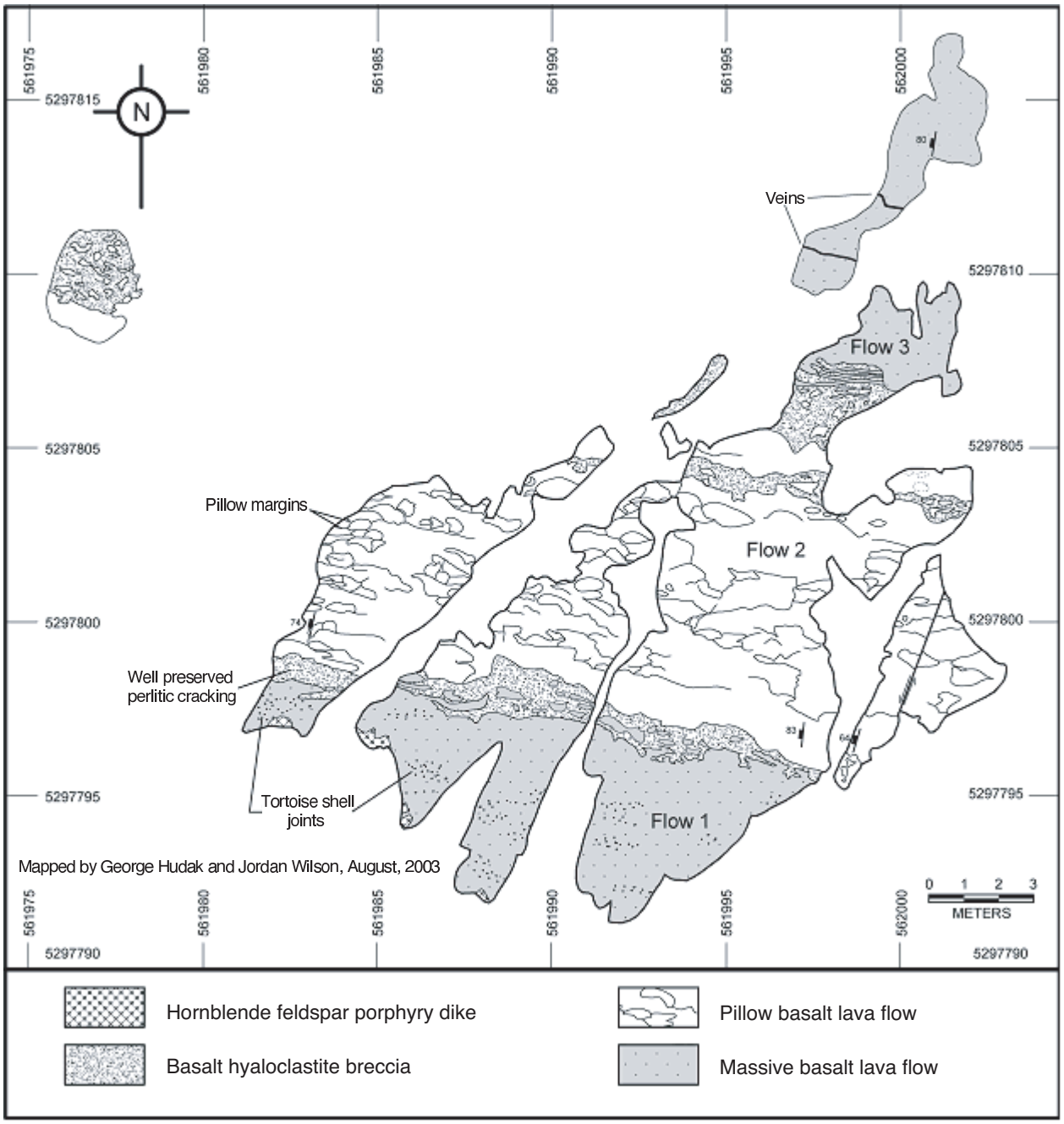


Figure 9.7. Detailed geologic map of sheet flows, pillow lavas, hyaloclastite, and associated "self-peperite." Map from Hudak and others (2004).

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.

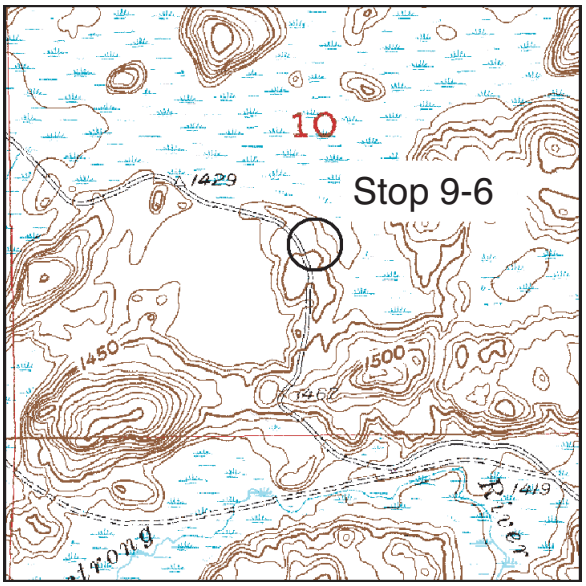
NEXT: Return to Highway 169, turn right and travel approximately 3.1 miles to the junction of Mud Creek Road (St. Louis County Road 38). Turn left (north) on 38 and travel approximately 1.5 miles to a series of low-lying outcrops on the east side of the road.

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.

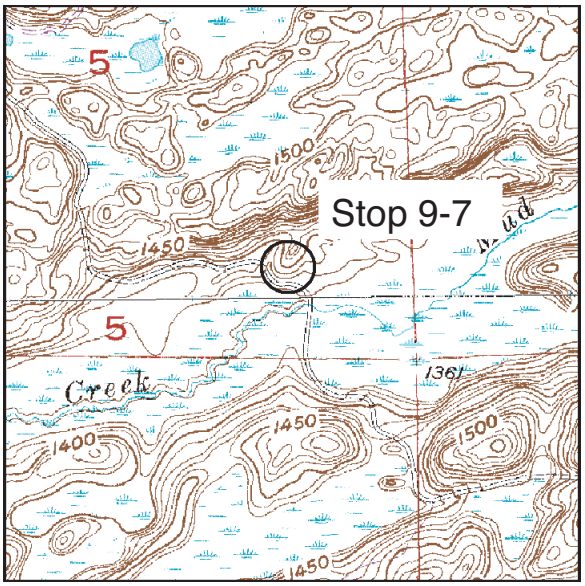
NEXT: Continue northwest on 38 for approximately 2.1 miles to the boat landing parking area along Mud Creek. Walk approximately 100 meters north along the road to the outcrop along the east side of the roadway.

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.

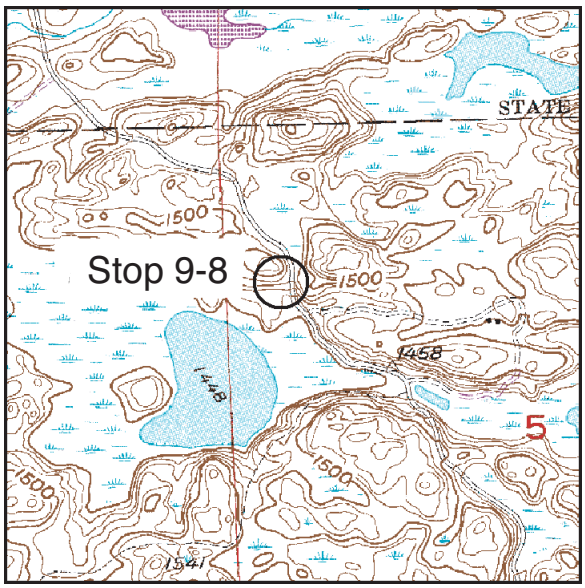
NEXT: Drive northwest along County Road 38 approximately 1.1 miles to the widened portion of the road.

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.

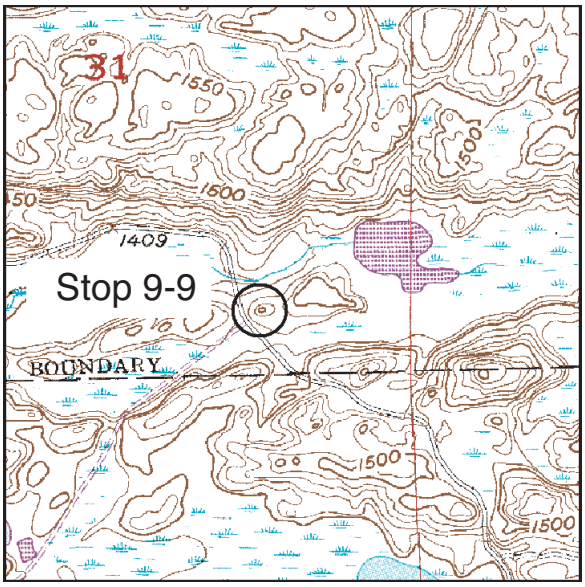
NEXT: Continue northwest along County Road 38 approximately 0.55 mile to a small, yellowish outcrop on the east side of the road.

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

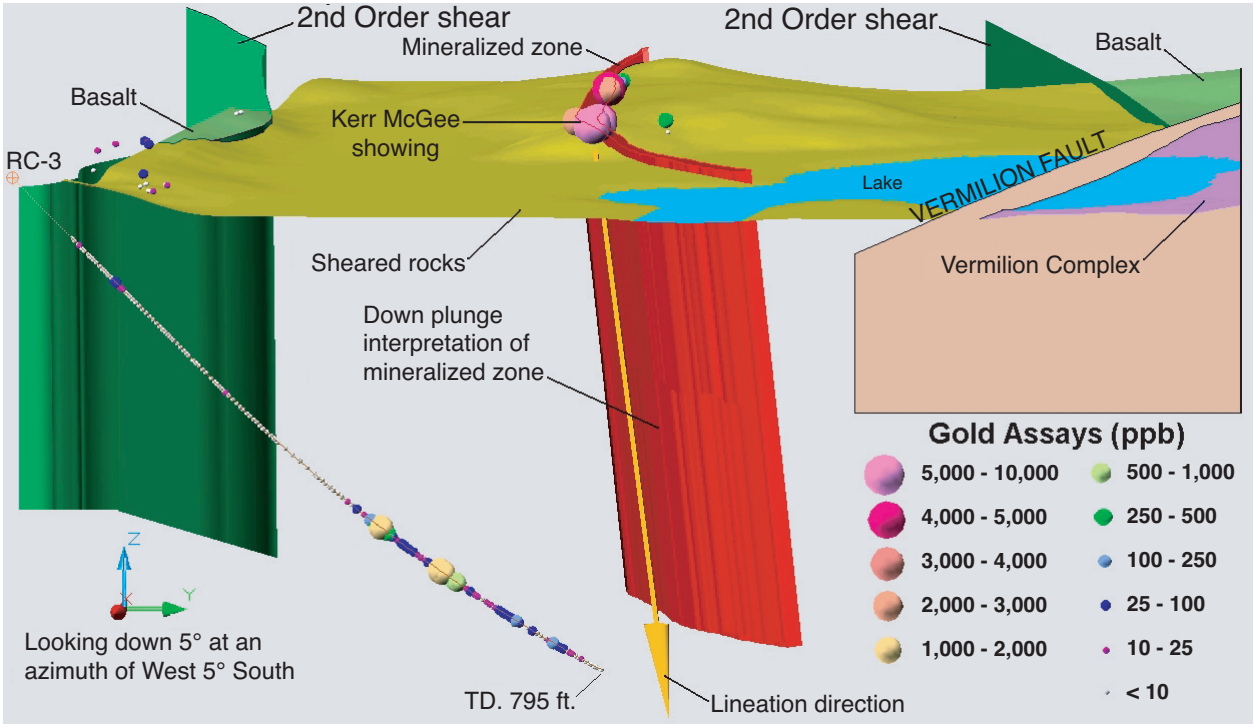
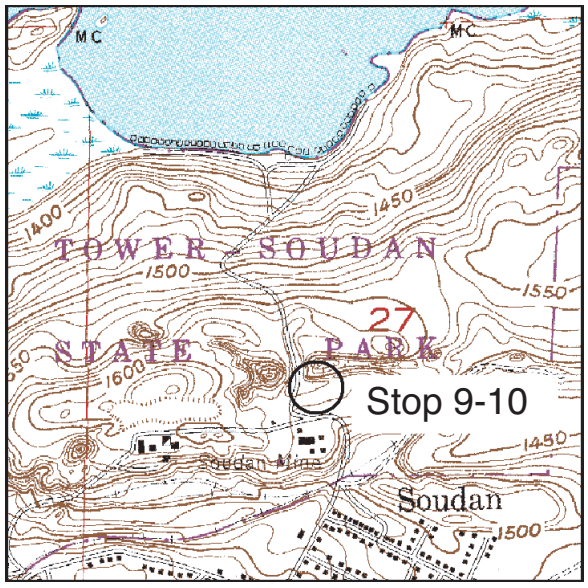


Figure 9.8. Three-dimensional view of the relationship between structural boundaries, the mineralized zone exposed on the surface at the Kerr McGee prospect, and drill hole RC-3. Upward extension to the surface of the two anomalous zones (greater than 1,000 parts per billion gold) intersected in hole 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 ($F_{0\text{-}1}$) 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).

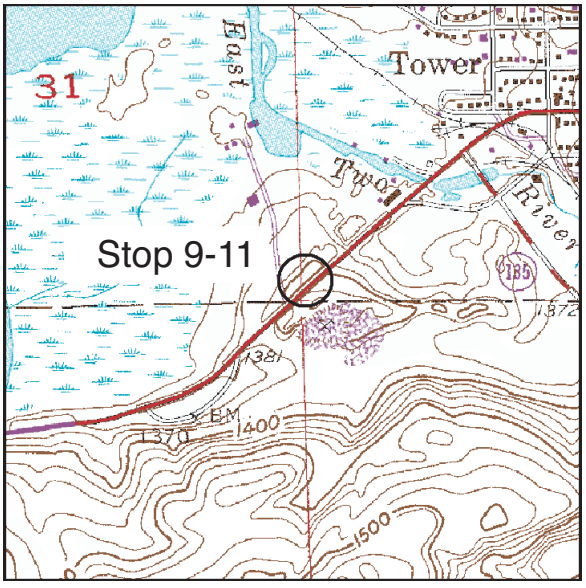
NEXT: Return to Highway 169 and turn right. Follow 169 through the town of Tower to the large outcrops immediately west of town (approximately 3.1 miles).

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.

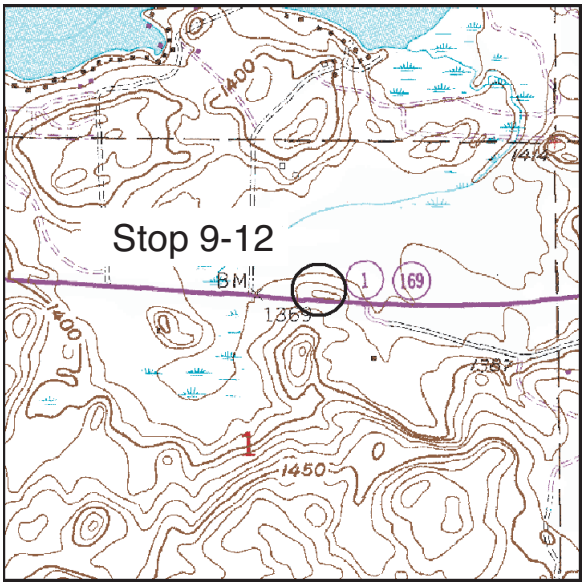
NEXT: Continue west on Highway 169 approximately 1.5 miles to a road cut.

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.

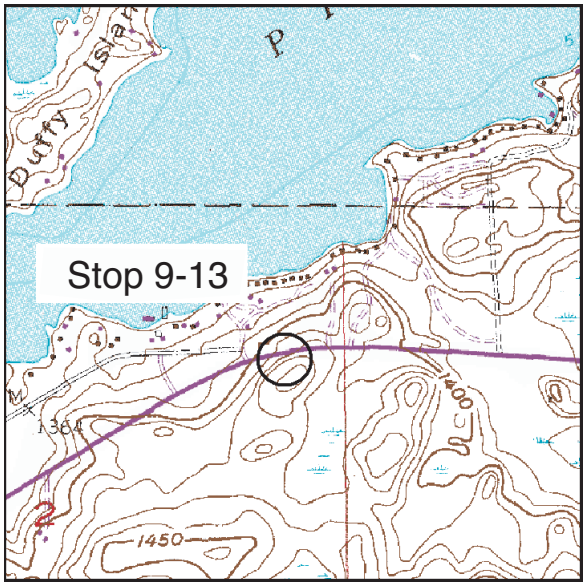
NEXT: Continue west on Highway 169 approximately 0.7 mile to the junction with St. Louis County Road 526.

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.

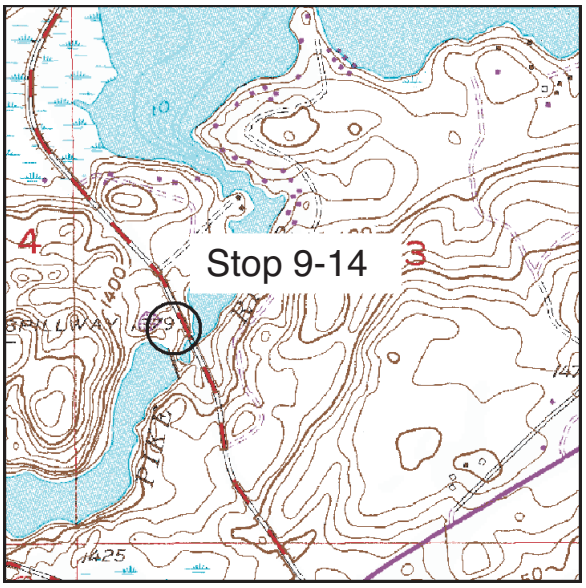
STOP 9-14

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).

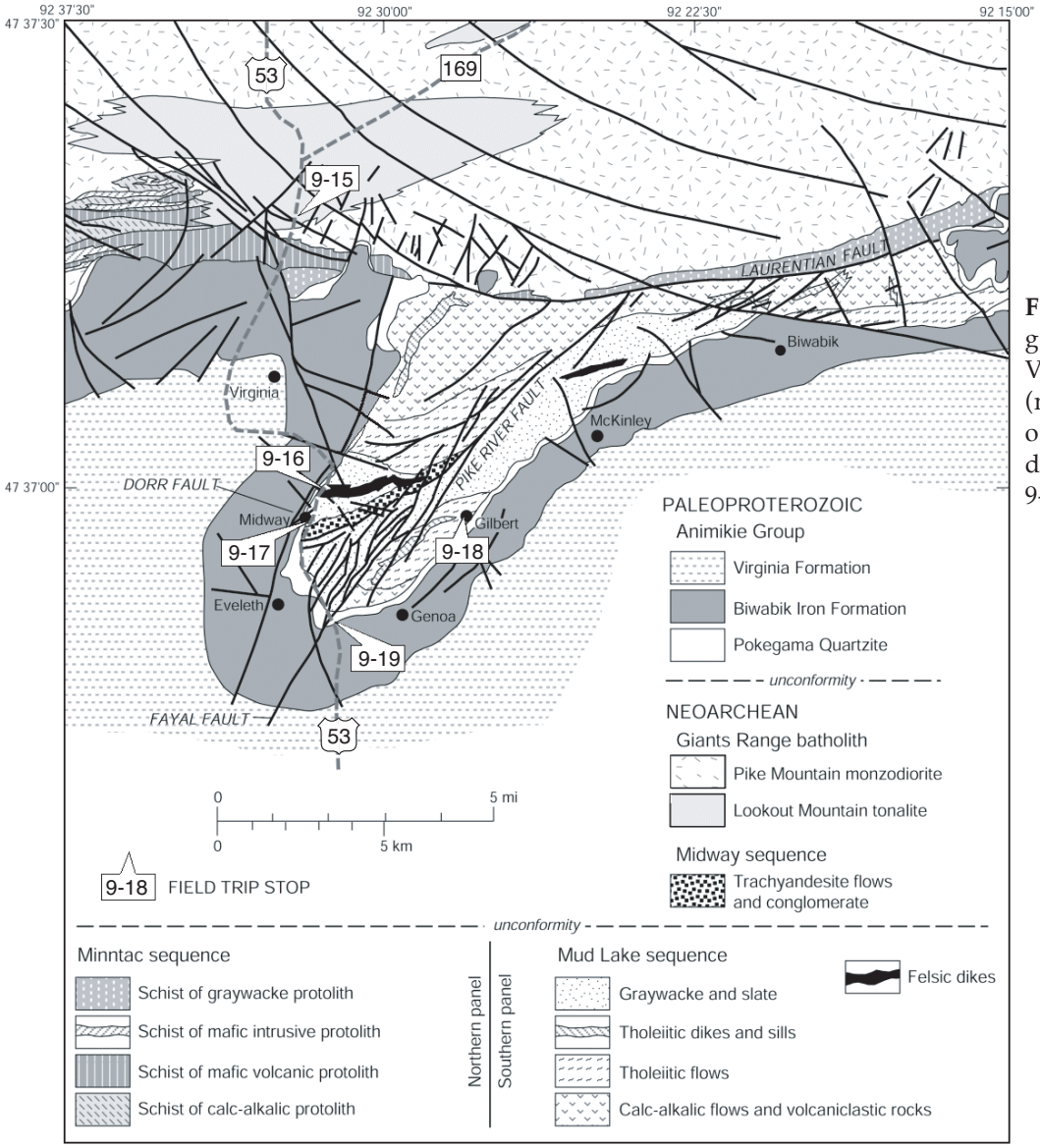


Figure 9.9. Generalized geologic map of the Virginia horn area (modified from Jirsa and others, 1998) showing details of field trip Stops 9-15 to 9-19.

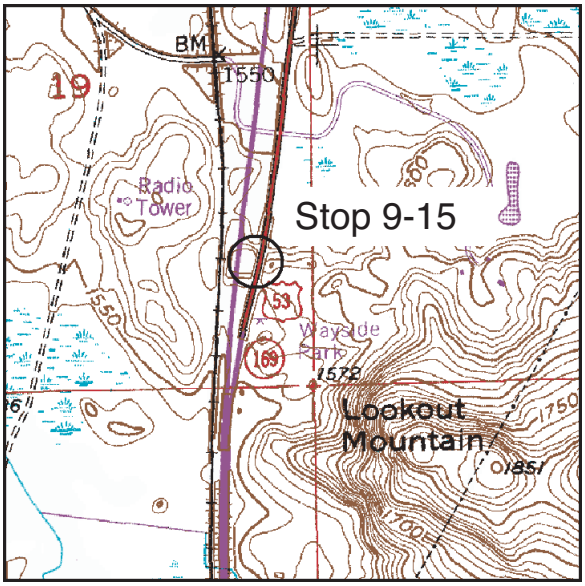
STOP 9-15

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.

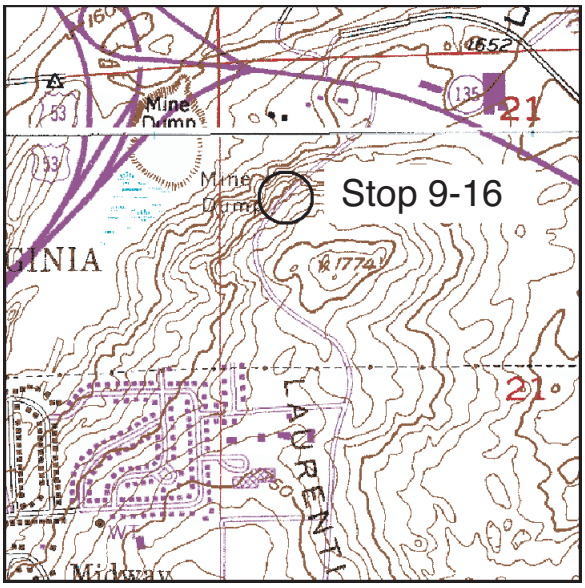
NEXT: Follow Highway 53 south through the town of Virginia. Take the exit for Highway 135 (east) approximately 0.5 mile to Bourgin Road. Turn right (south) on Bourgin Road and continue about 0.4 mile to a large cut on the left (east) side of the road.

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.

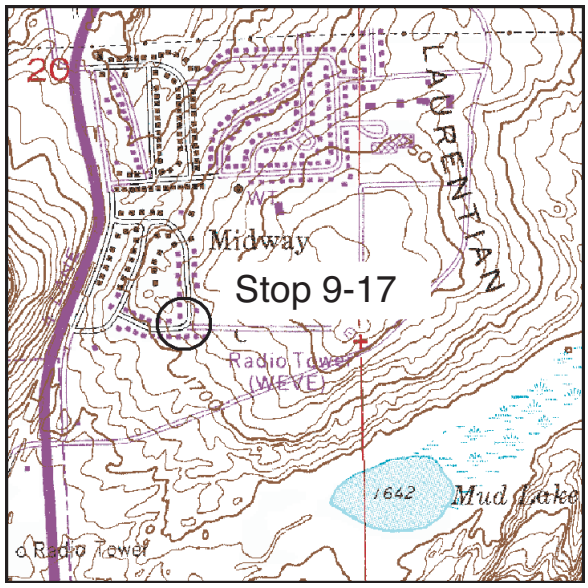
STOP 9-17

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.

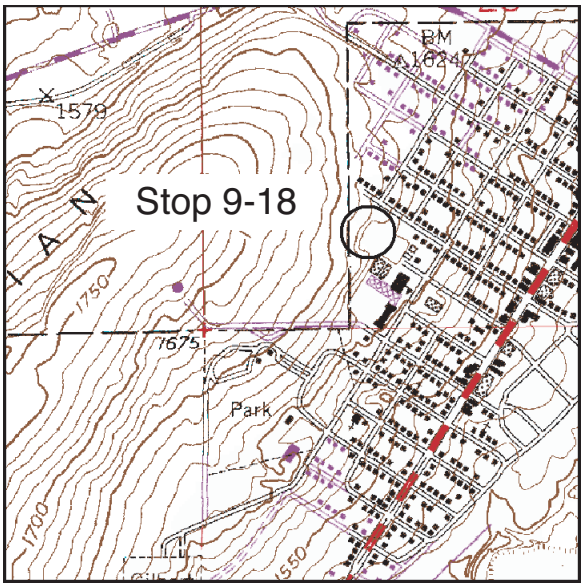
NEXT: Return to Highway 135. Turn right and follow 135 approximately 2.4 miles to a residential street on the northwest side of the town of Gilbert. Turn right and go 5 blocks and park by Gilbert Junior High School.

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

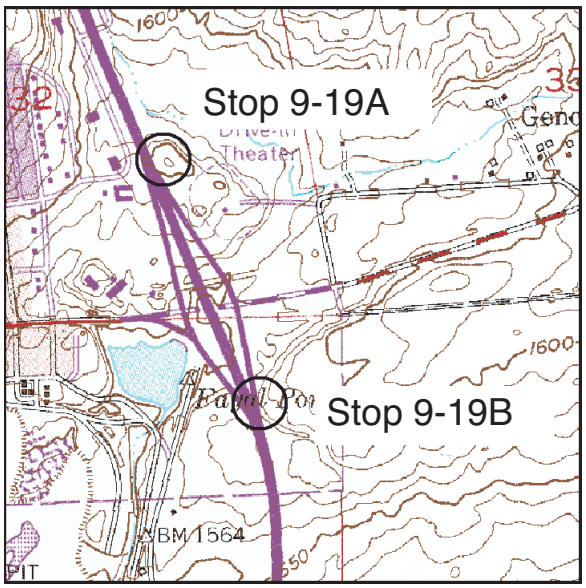
37 approximately 3.3 miles to the on/off ramps onto Highway 53.

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.

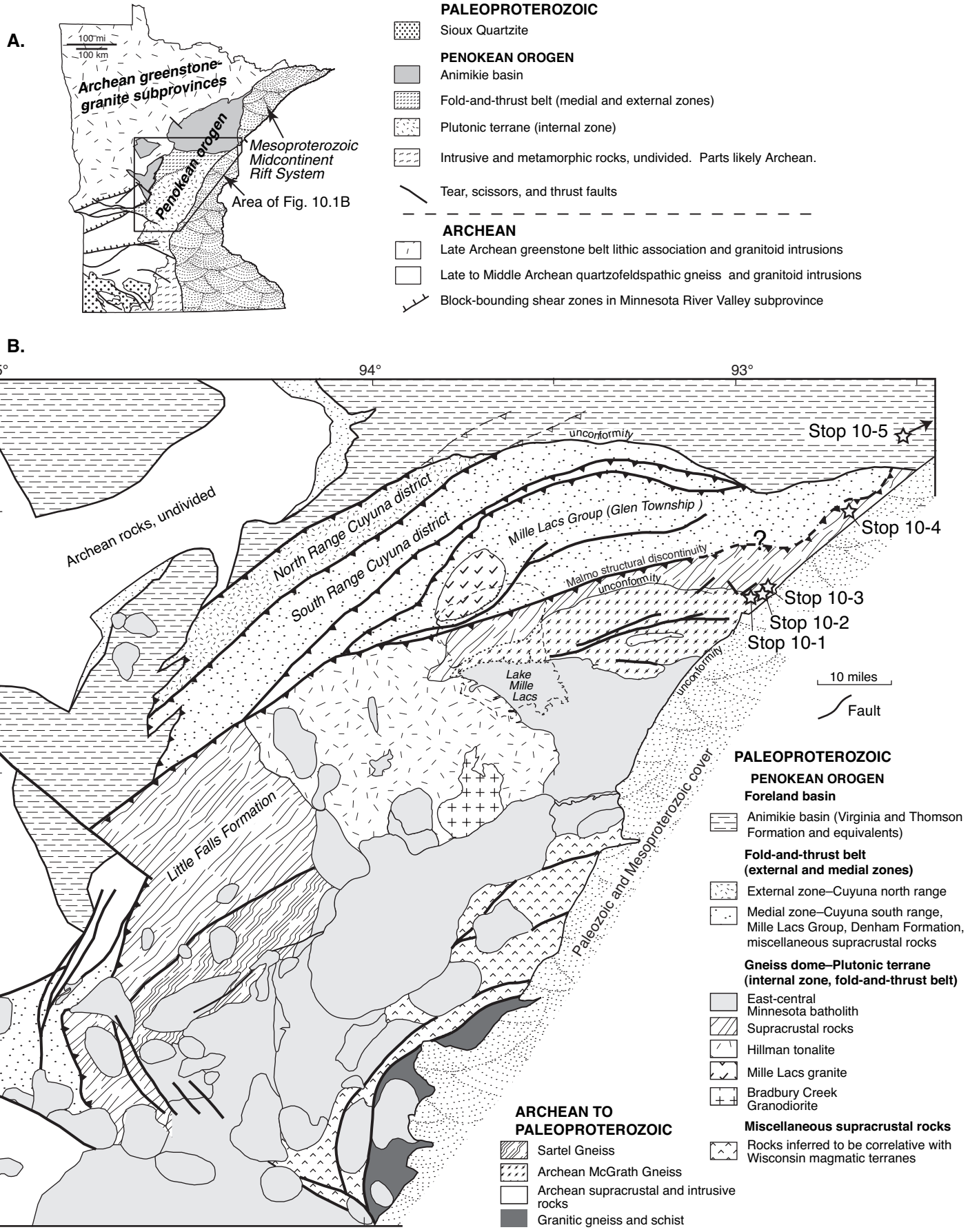
CONTINENTAL CRUST WEST OF THE MIDCONTINENT RIFT SYSTEM SOUTH OF LAKE SUPERIOR

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. Parts likely Archean.
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'
Archean rocks, undivided
North Range Cuyuna district
South Range Cuyuna district
unconformity
Mille Lacs Group (Glen Township)
Malmo structural discontinuity
unconformity
Stop 10-5
Stop 10-4
Stop 10-3
Stop 10-2
Stop 10-1
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, Denham Formation, miscellaneous supracrustal rocks
Gneiss dome–Plutonic terrane (internal zone, fold-and-thrust belt)
East-central Minnesota batholith
Supracrustal rocks
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

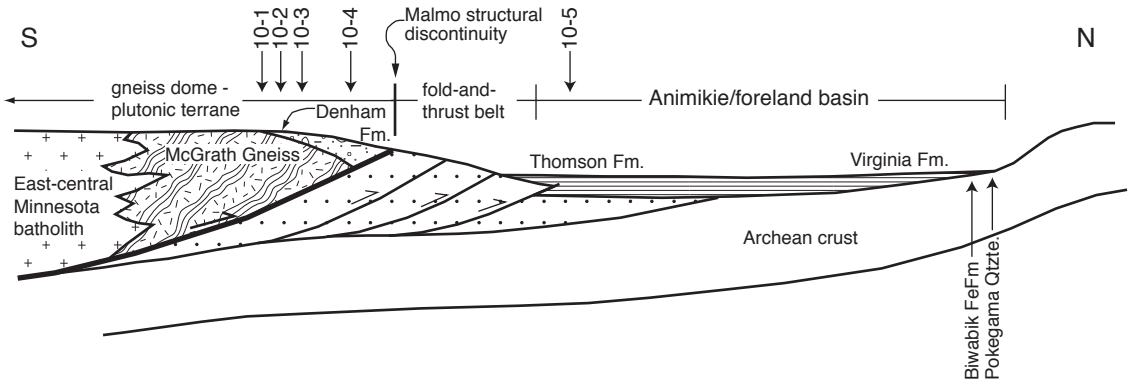


Figure 10.2. Schematic north–south cross-section across the eastern edge of the Minnesota segment of the Penokean orogen. Relative locations of field trip stops are indicated by numbers with arrows.

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.

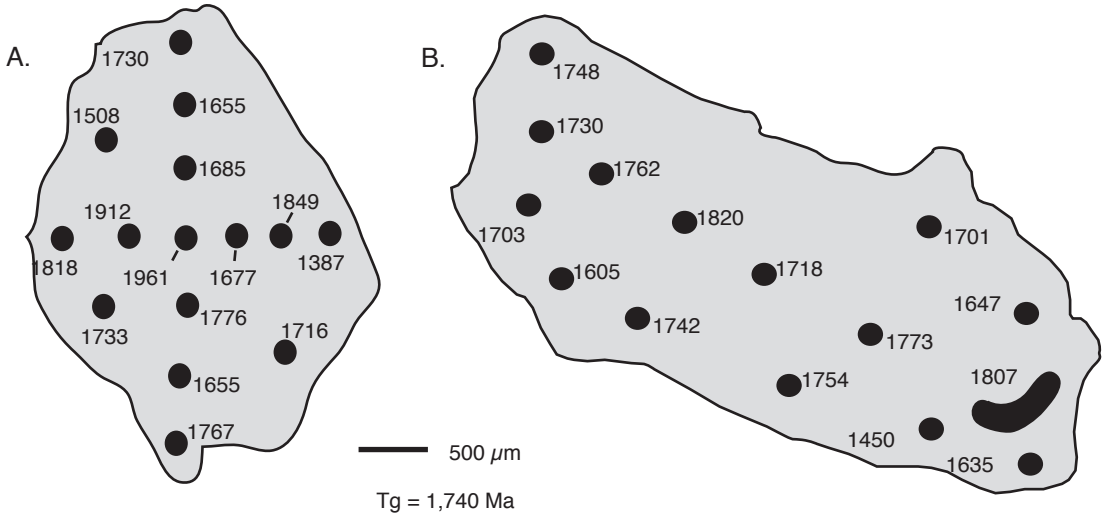


Figure 10.3. Spot age data (in Ma) of muscovite grains from argon ion laser data (McKenzie, 2004).

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

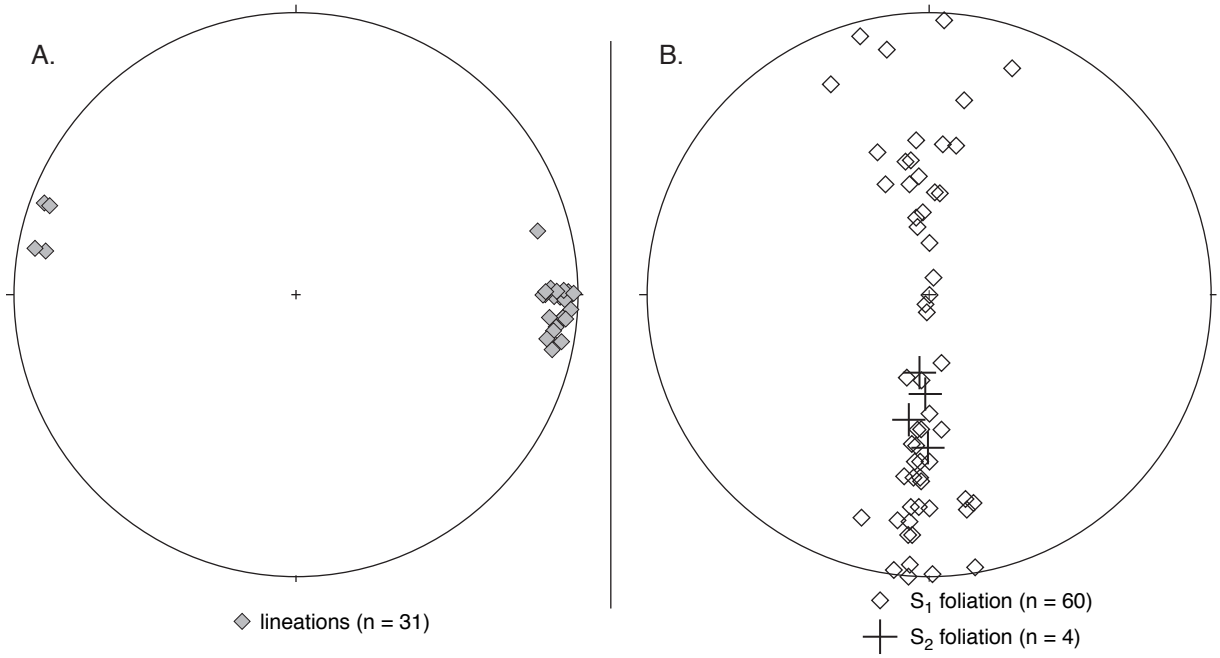


Figure 10.4. Equal area, lower hemisphere projections of structural elements of the Denham Formation and overlying metagraywacke unit to the north.

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

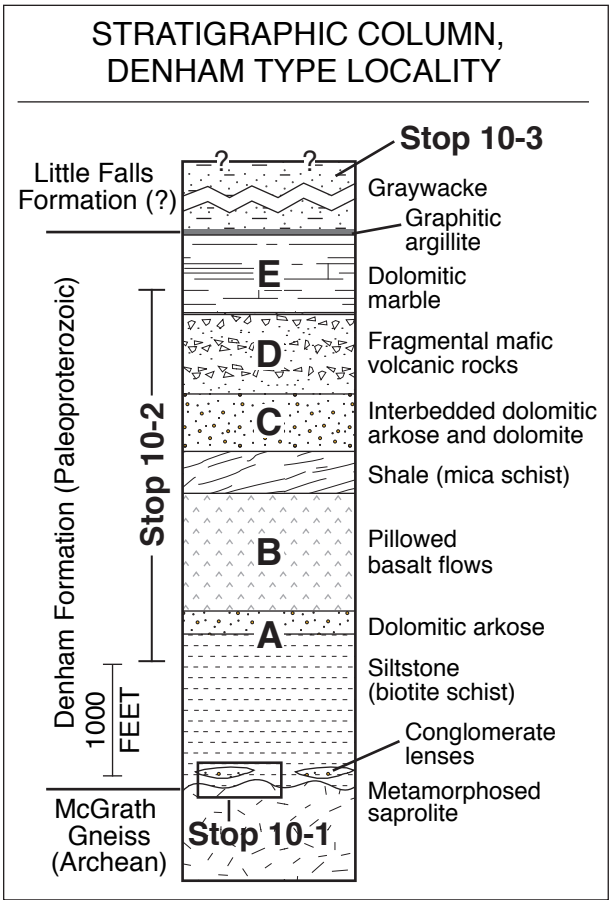


Figure 10.5. Schematic stratigraphic section of the Denham Formation, showing relative location of field trip stops.

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

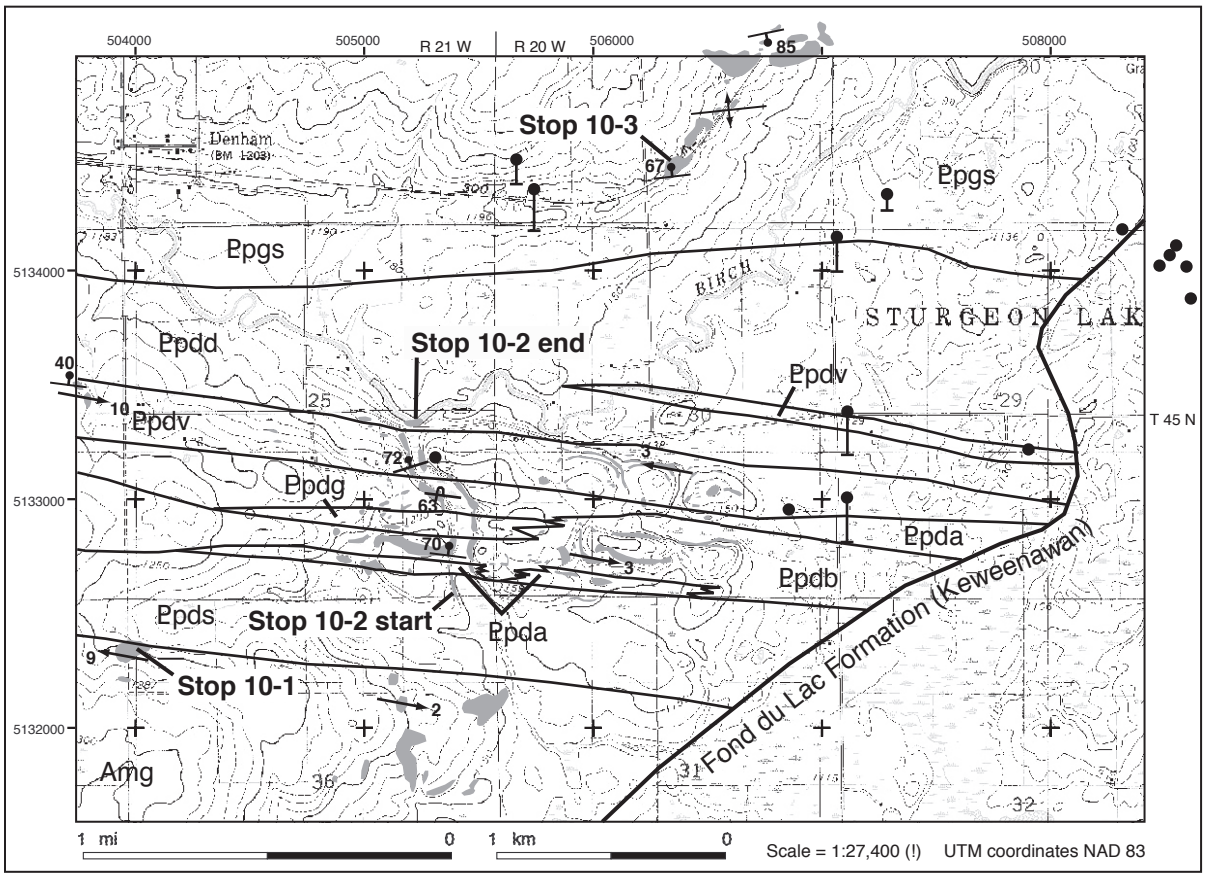


Figure 10.6. Bedrock geologic map of the area around the type locality of the Denham Formation, showing stops. Black circles are locations of exploratory drill holes and gray areas are mapped outcrops. McGrath Gneiss—Amg; Denham Formation—Epds (metasiltstone), Epda (meta-dolomitic arkose), Epdb (metabasalt),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

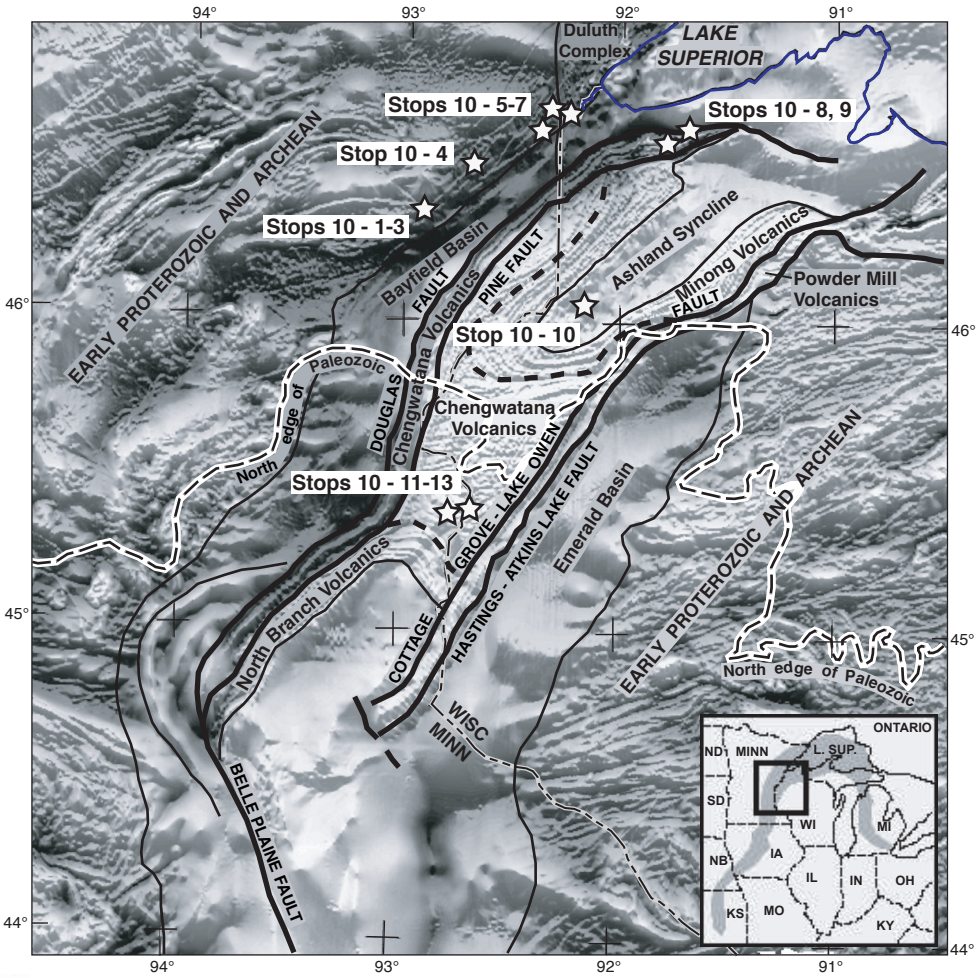


Figure 10.7. Aeromagnetic anomaly map of the St. Croix horst region in Minnesota and Wisconsin, showing subdivisions of the St. Croix horst volcanic sequences into Chengwatana and Minong Volcanics. General stop locations are indicated by stars. Inset map shows the general extent of the Midcontinent rift 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

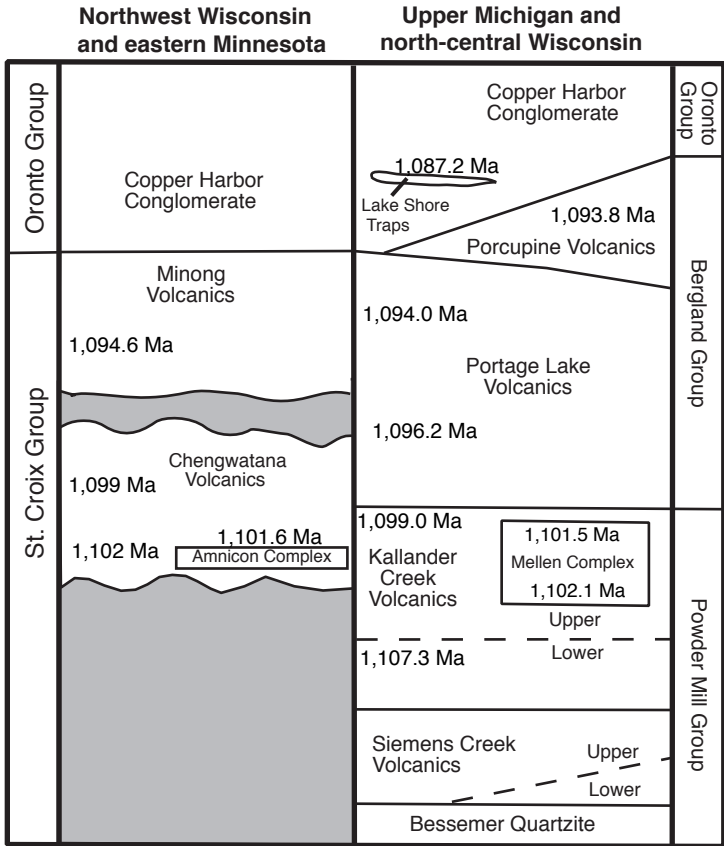


Figure 10.8. Stratigraphic column for Keweenawan volcanic rocks on the south shore of western Lake Superior. Modified from Nicholson and others (2001) by Suzanne Nicholson; see Nicholson and others (2001) for source of dates.

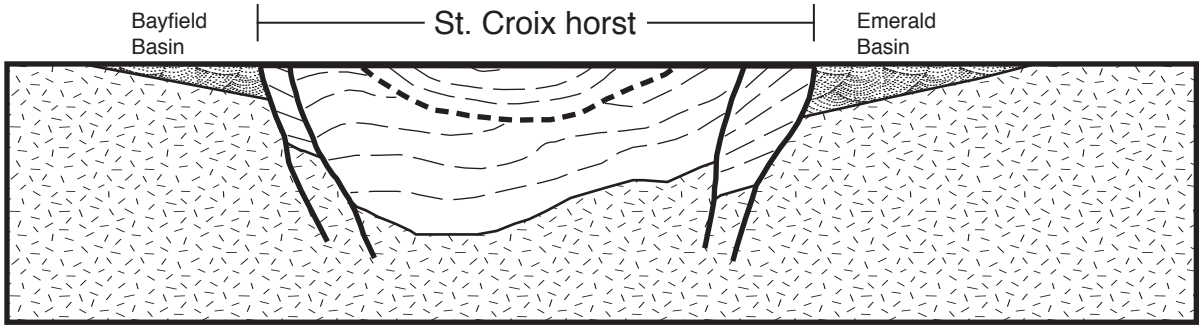


Figure 10.9. Schematic cross-section model of the Midcontinent rift system perpendicular to the St. Croix horst, across the southern tip of the Ashland Syncline. Basalt flows are designated by dashes. Modified from Cannon and others (2001).

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.

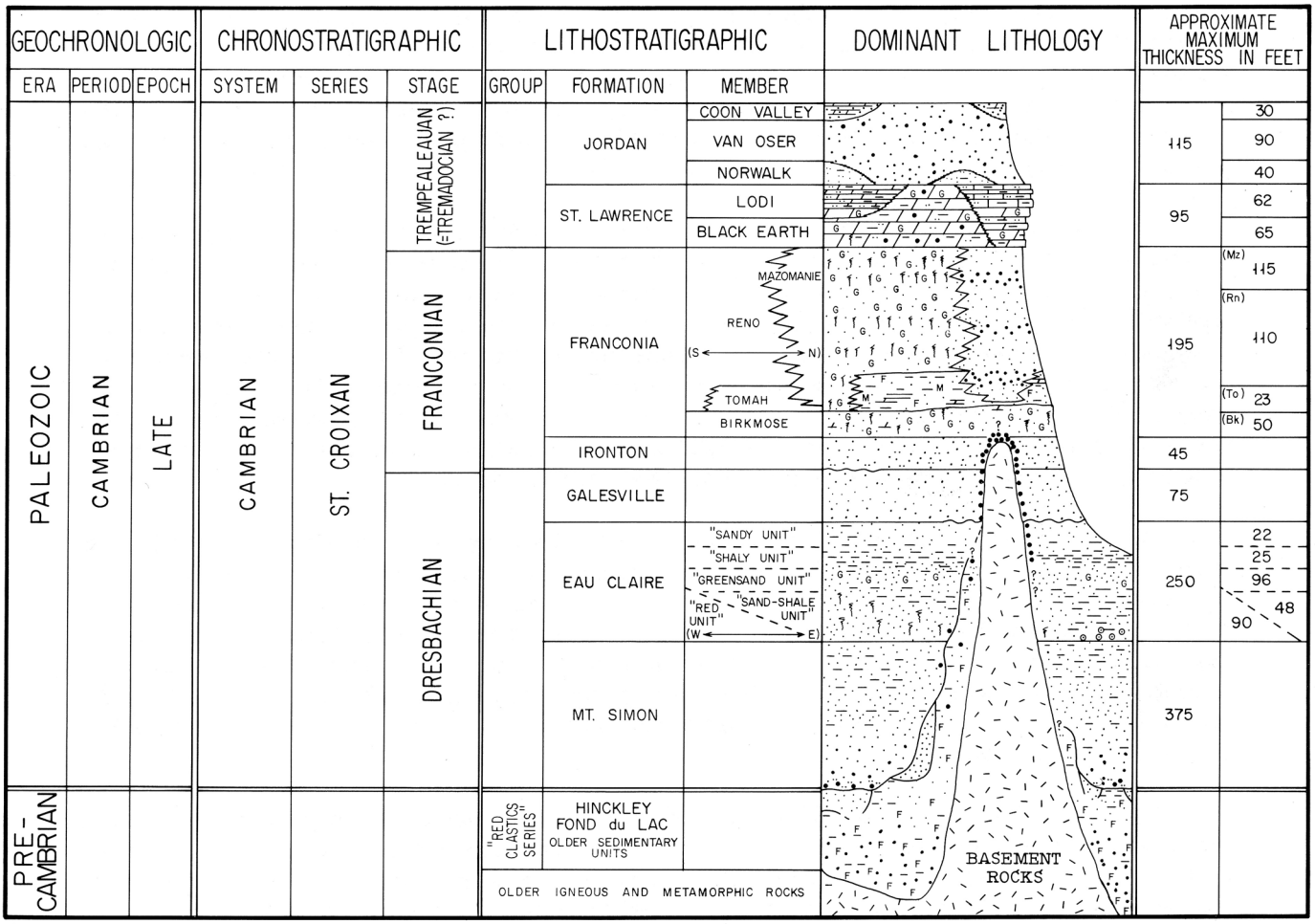


Figure 10.10. Stratigraphic column of Upper Cambrian rock units in Minnesota (from Mossler, 1987). Conglomeratic facies surrounding bedrock highs are shown by black dots.

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.

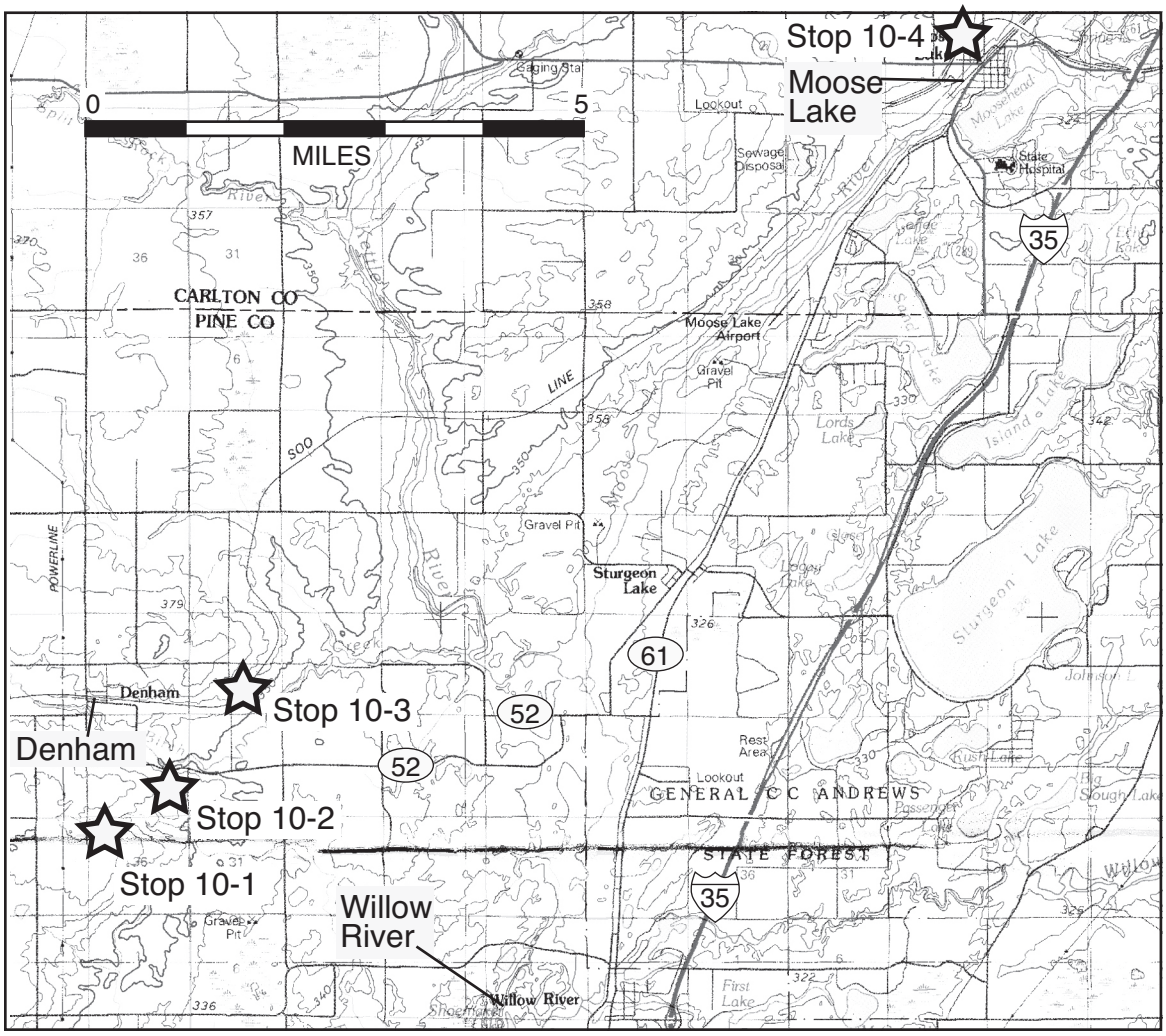
STOP 10-1

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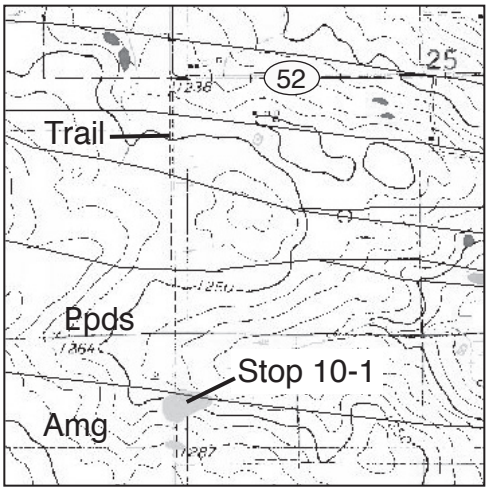
McGrath Gneiss and contact with overlying Denham Formation

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 (Ppds)

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 Ppds 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.

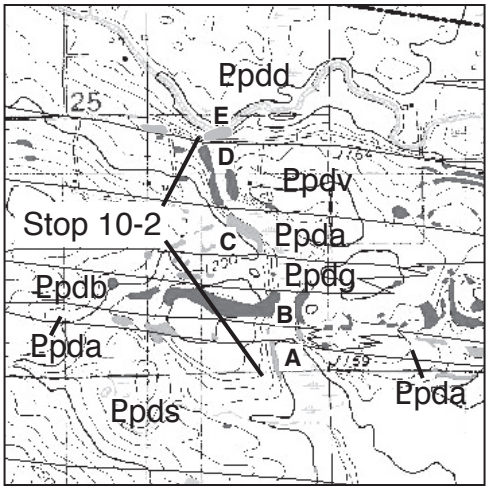
STOP 10-2

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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 (Ppds), dolomite-cemented arkosic arenite with rare cobbles of McGrath Gneiss (Ppda), pillowed basalt flows (amphibolite; Ppdb), shale (staurolite-garnet-mica schist; Ppdg), interbedded dolomitic arkose and dolostone (Ppda), fragmental basaltic volcanic rocks (amphibolite; Ppdv), and at the top of the section, pure dolostone (marble; Ppdm). 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

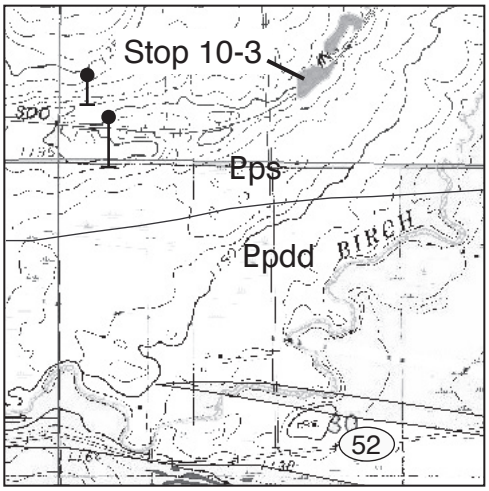
grade now utilized as an ATV trail. Park at the trail junction and walk northeast on the trail about 500 feet to outcrops and low road cuts.

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.

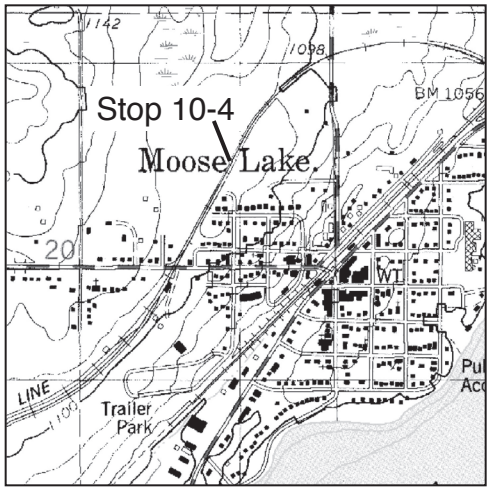
NEXT: Take County Road 52 back east to County Road 61. Turn left (north) on County Road 61 and drive approximately 7.5 miles to the junction with State Highway 27 at a stop light in Moose Lake. Turn left (west) on Highway 27 and follow it a short distance to the Soo Line trail; just before the trail,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

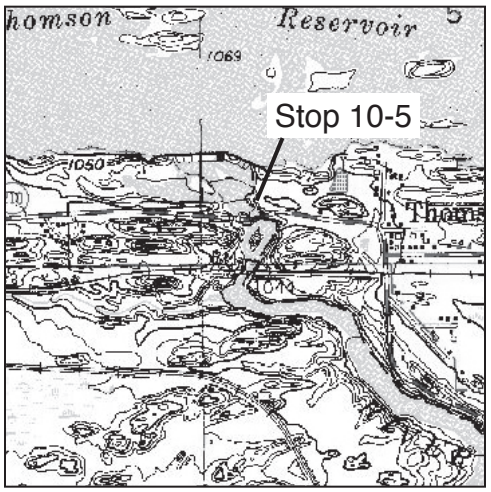
Slate and graywacke of the Thomson Formation; Keweenawan diabase dikes

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

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

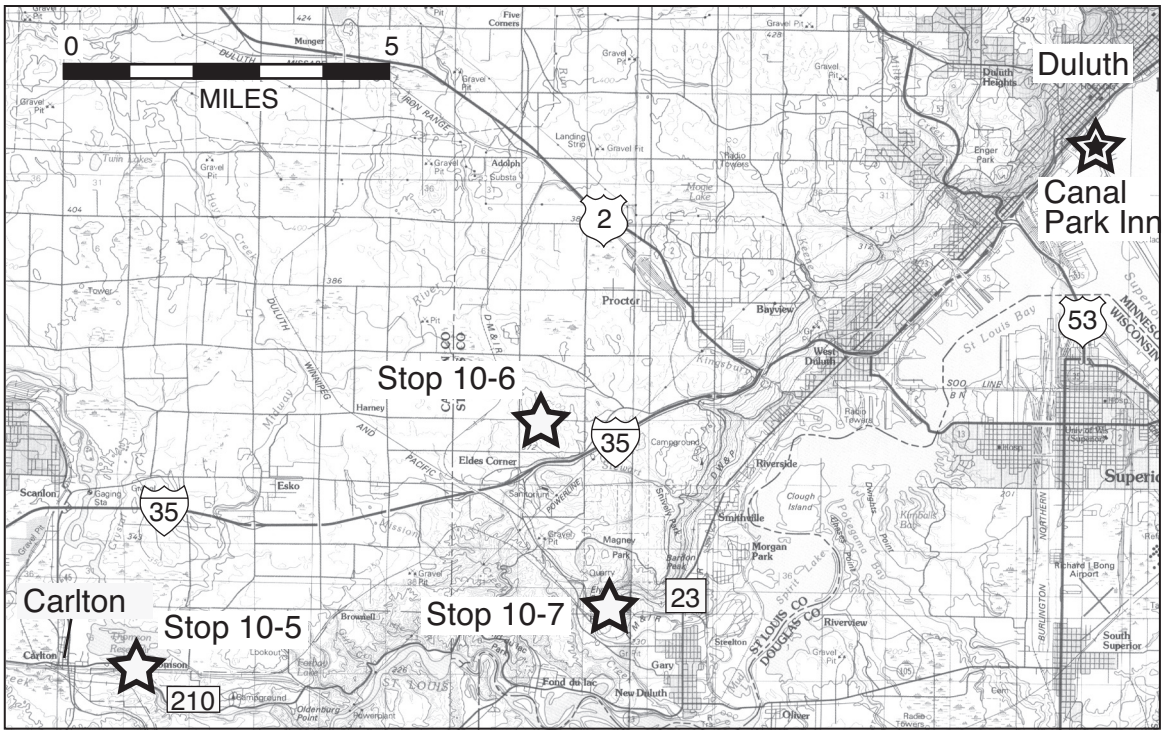
Highlights: Variety of sedimentary structures, folds, and diabase dikes

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.

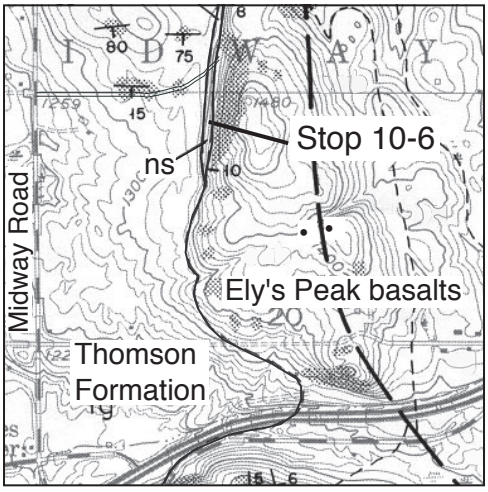
NEXT: Return to Carlton; in Carlton turn north on State Highway 45 to I-35. Go east (north toward Duluth) on I-35 approximately 7 miles to Midway Road (St. Louis County Road 13). Turn north on Midway Road and continue 0.8 mile to a road to the east. Turn right on this road and travel approximately 0.2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

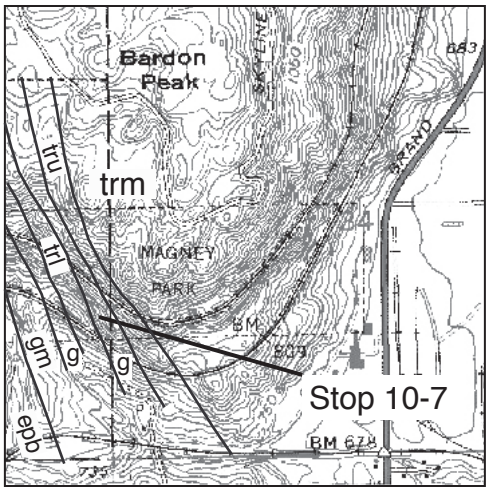
NEXT: Drive back south on County Road 13. Continue over I-35 to the junction with St. Louis County Road 3. Turn left (southeast) on County Road 3 and go approximately 0.5 mile and turn left (east) onto Skyline Parkway. Follow Skyline Parkway for 2.4 miles to a parking area and walk downhill to railroad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



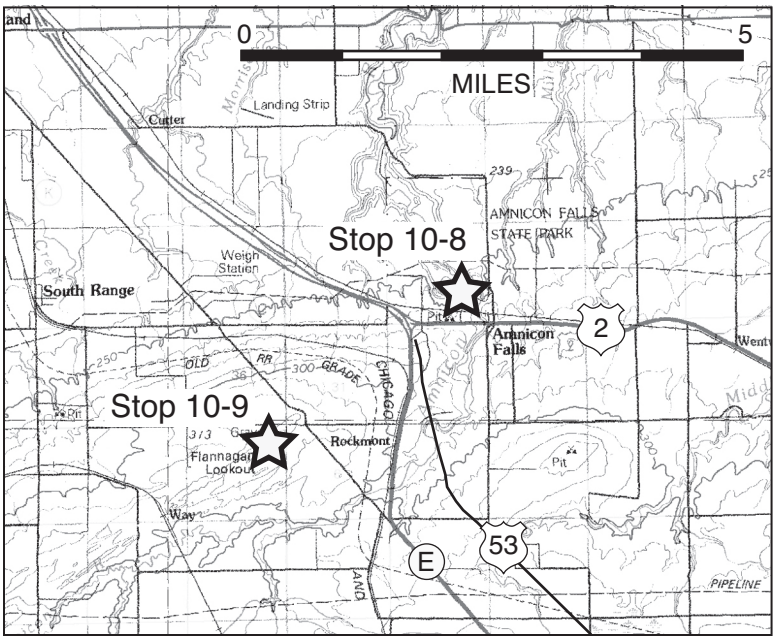
Location map explanation: epb—Ely's Peak basalts; gm—marginal gabbro; g—gabbro; trl—lower troctolite; tru—upper troctolite; trm—main troctolite.

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.

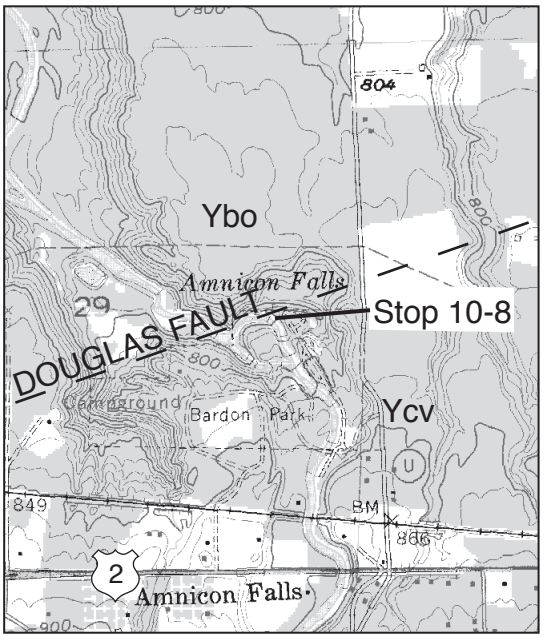
STOP 10-8

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!

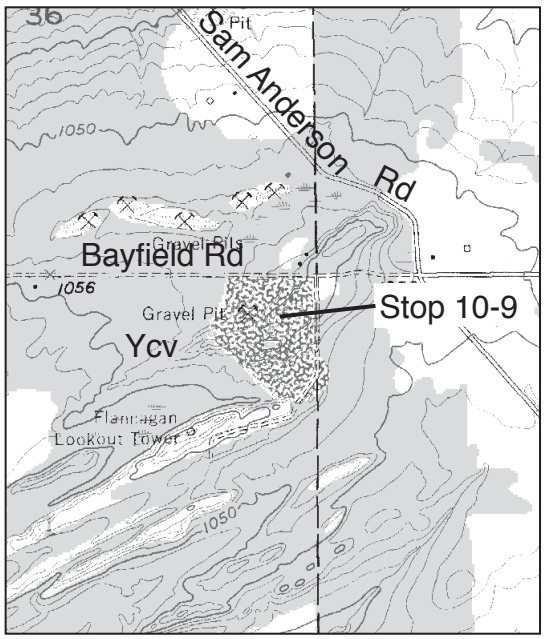
South Range Quarry

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.

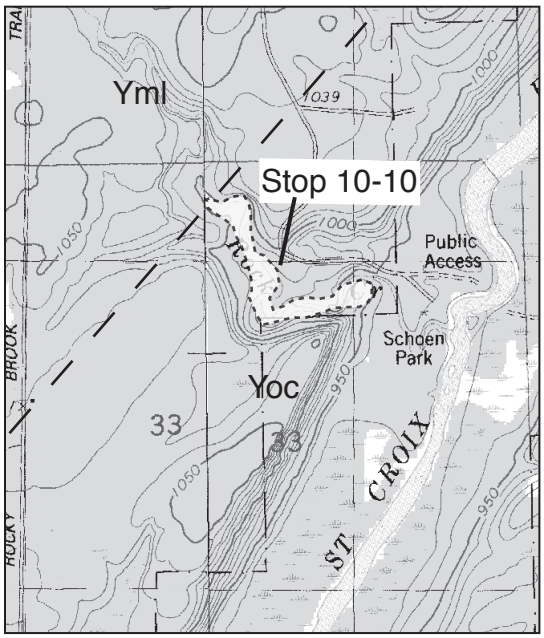
NEXT: Return to Highway 53 and head south approximately 29 miles, past Solon Springs, to the junction with Douglas County Road T. Go west on County Road T for about 16 miles. Two miles after crossing the St. Croix River, turn south on Rocky Brook Trail. Follow the road south and west for 4 miles to the 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

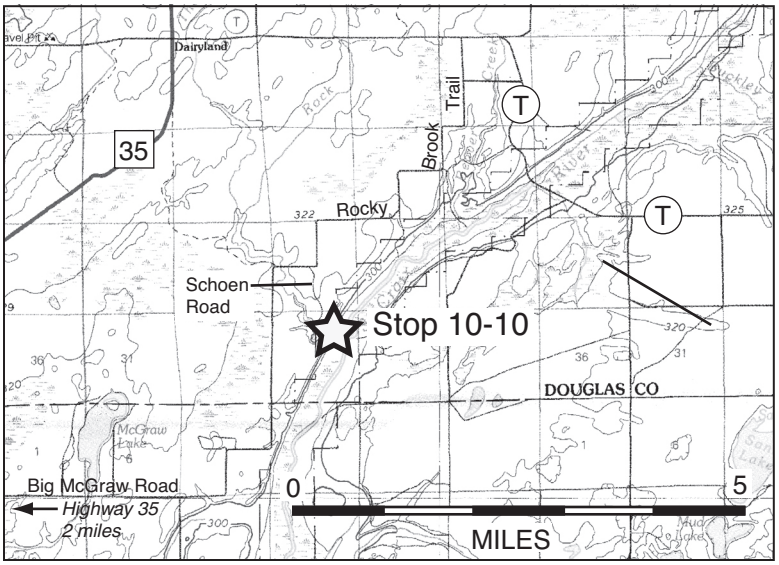
Location: T. 43 N., R. 14 W., sec. 33, NE

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.

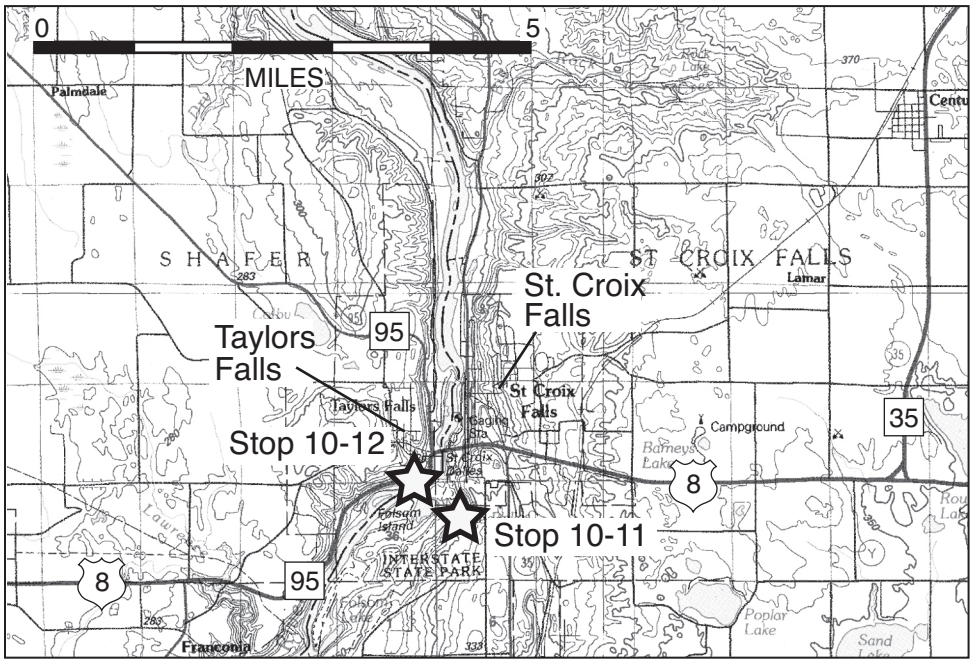
STOP 10-11

No hammering or collecting please!

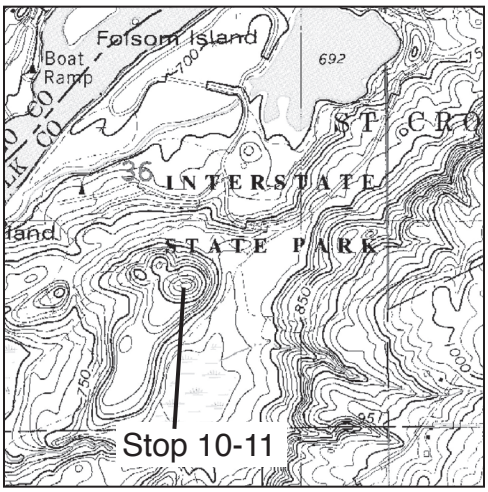
Eagle's Peak Basalts

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.

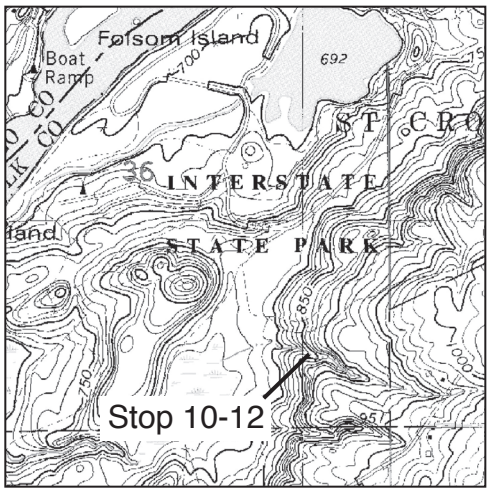
STOP 10-12

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.

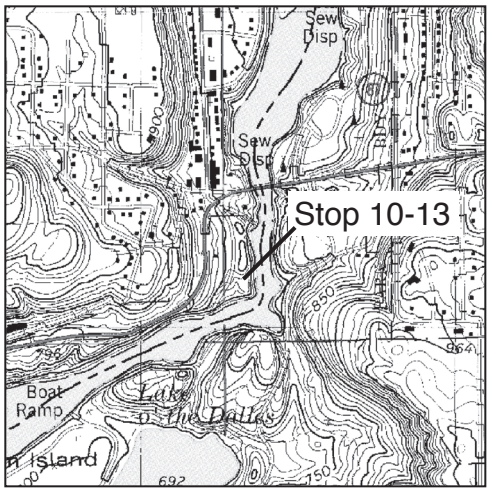
STOP 10-13

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.

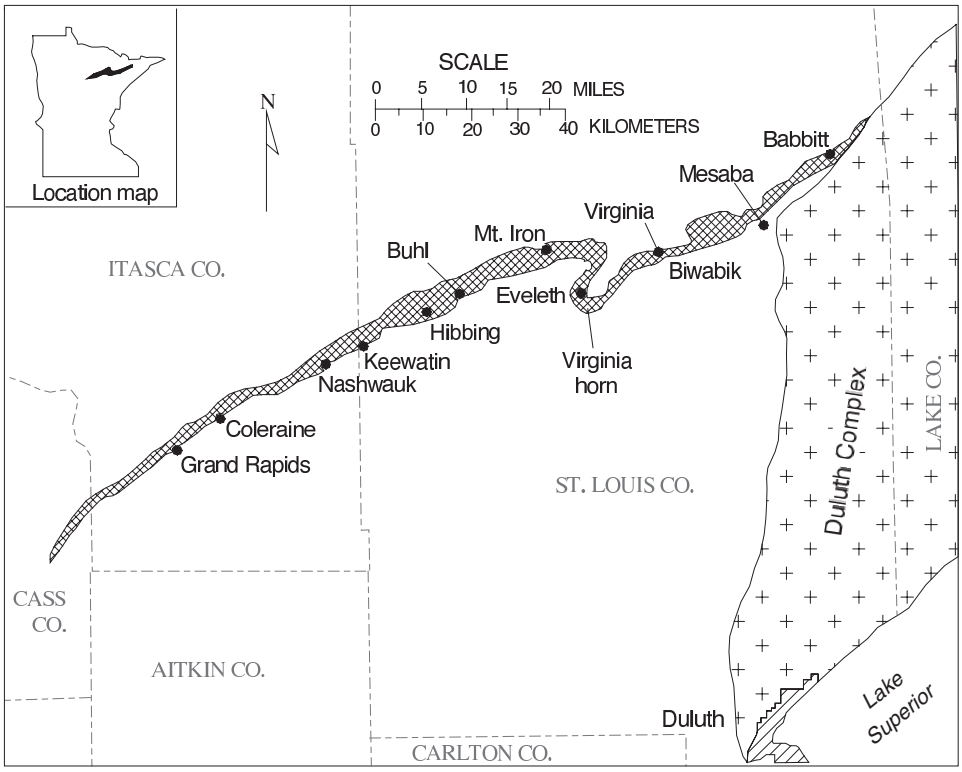


Figure 11.1. Generalized map of the Mesabi Iron Range (cross-hatched). Note the Duluth Complex on the east side.

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).

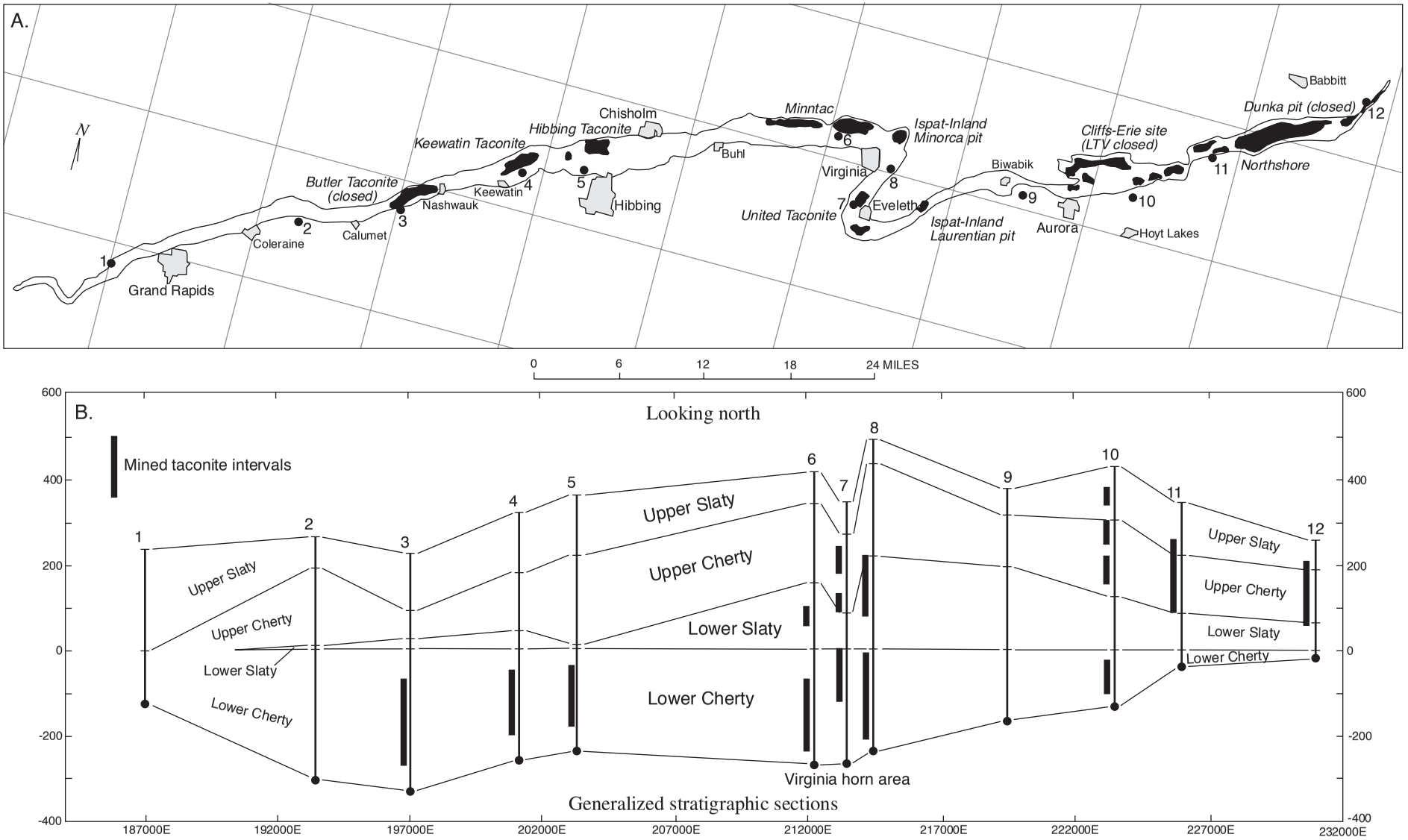


Figure 11.2. Generalized map of the Mesabi Iron Range.

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).

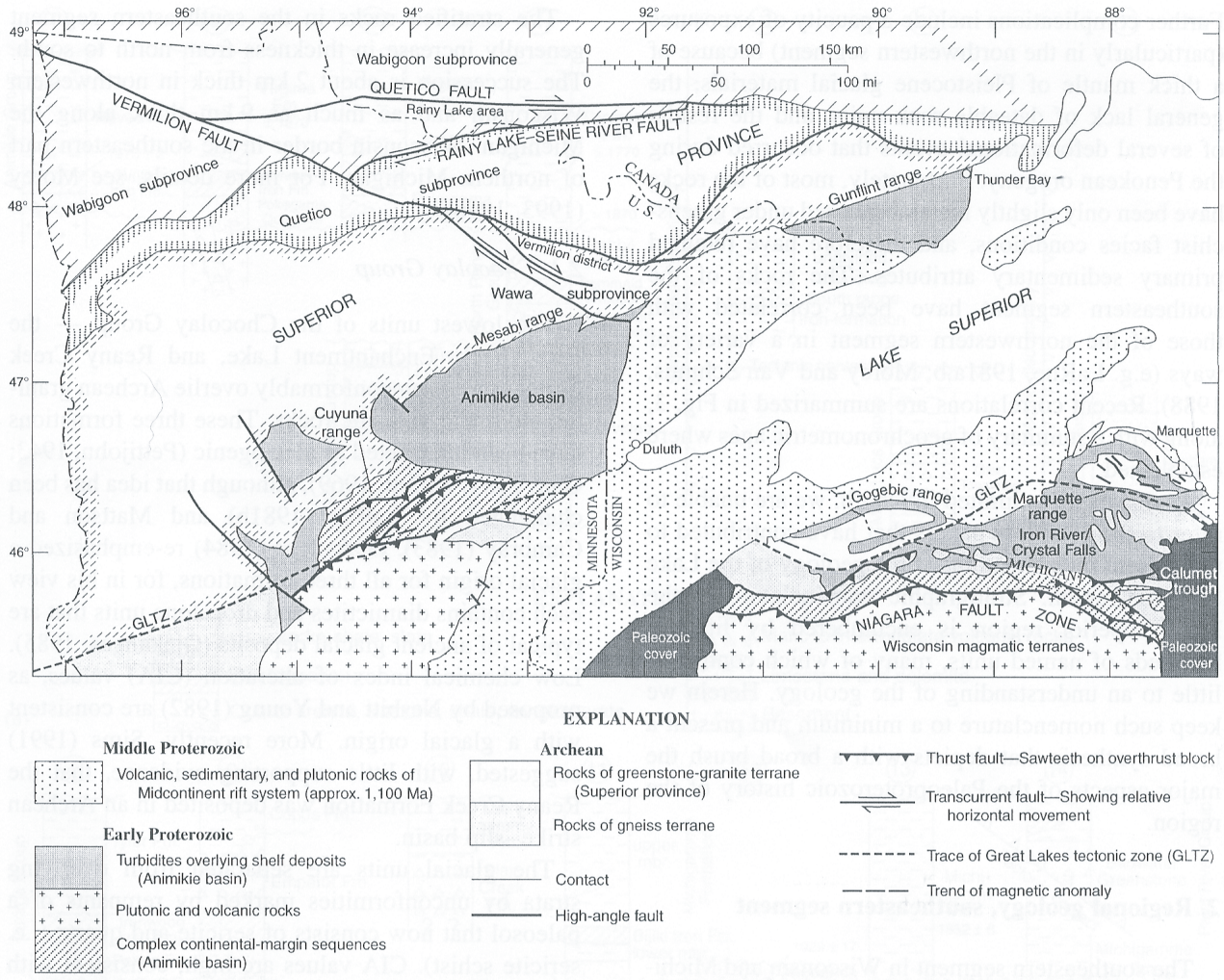


Figure 11.3. Generalized geologic map showing the distribution of Precambrian rocks and structural elements of the Lake Superior region; modified from Ojakangas (1994).

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."

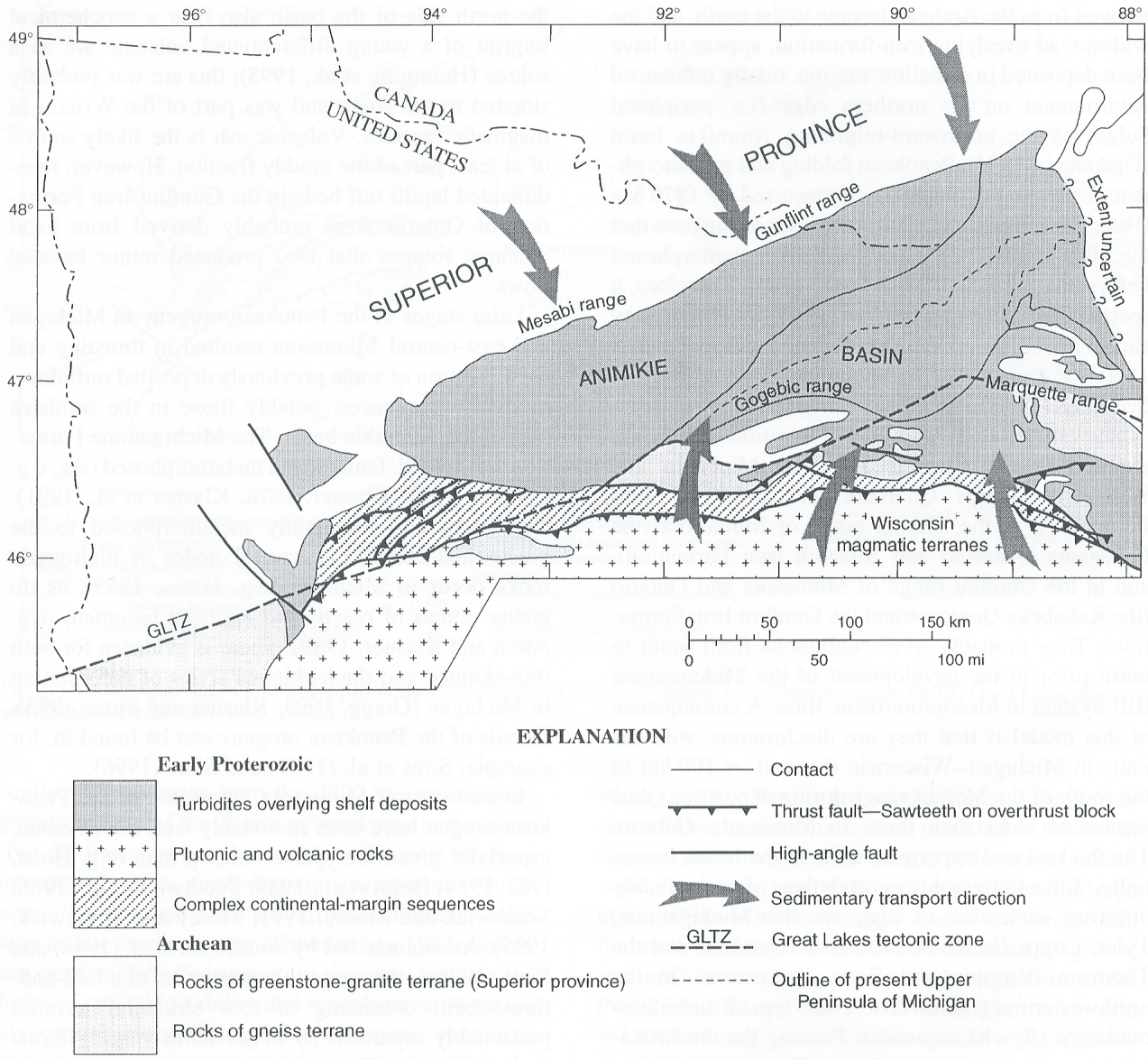


Figure 11.4. Schematic hypothesized paleogeography at the time of sedimentation of the Animikie Group turbidites that overlie shelf deposits in the Animikie basin. The rocks of the 1,100 Ma Midcontinent Rift System have been removed from the map, and Michigan and Wisconsin are thus positioned 60 miles 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

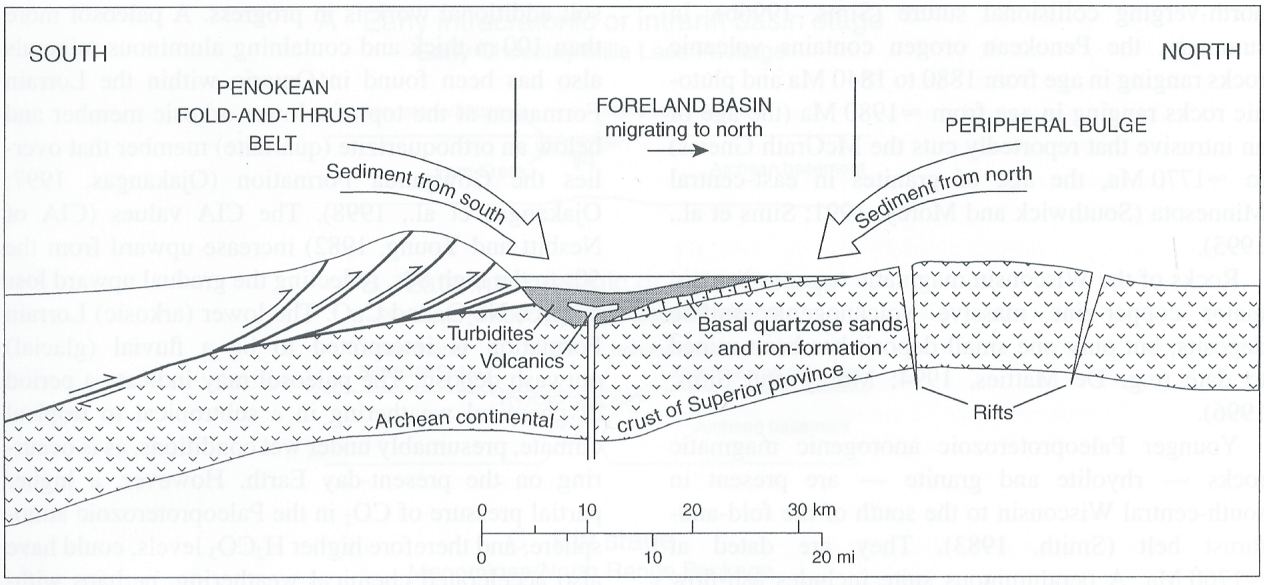


Figure 11.5. Schematic cross-section depicting deposition of the Animikie Group turbidites that overlie shelf deposits in the Animikie basin, with sediment derived from both the north and south. The southern area, the fold-and-thrust belt, comprises a complex assemblage including: 1. Accreted Paleoproterozoic 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

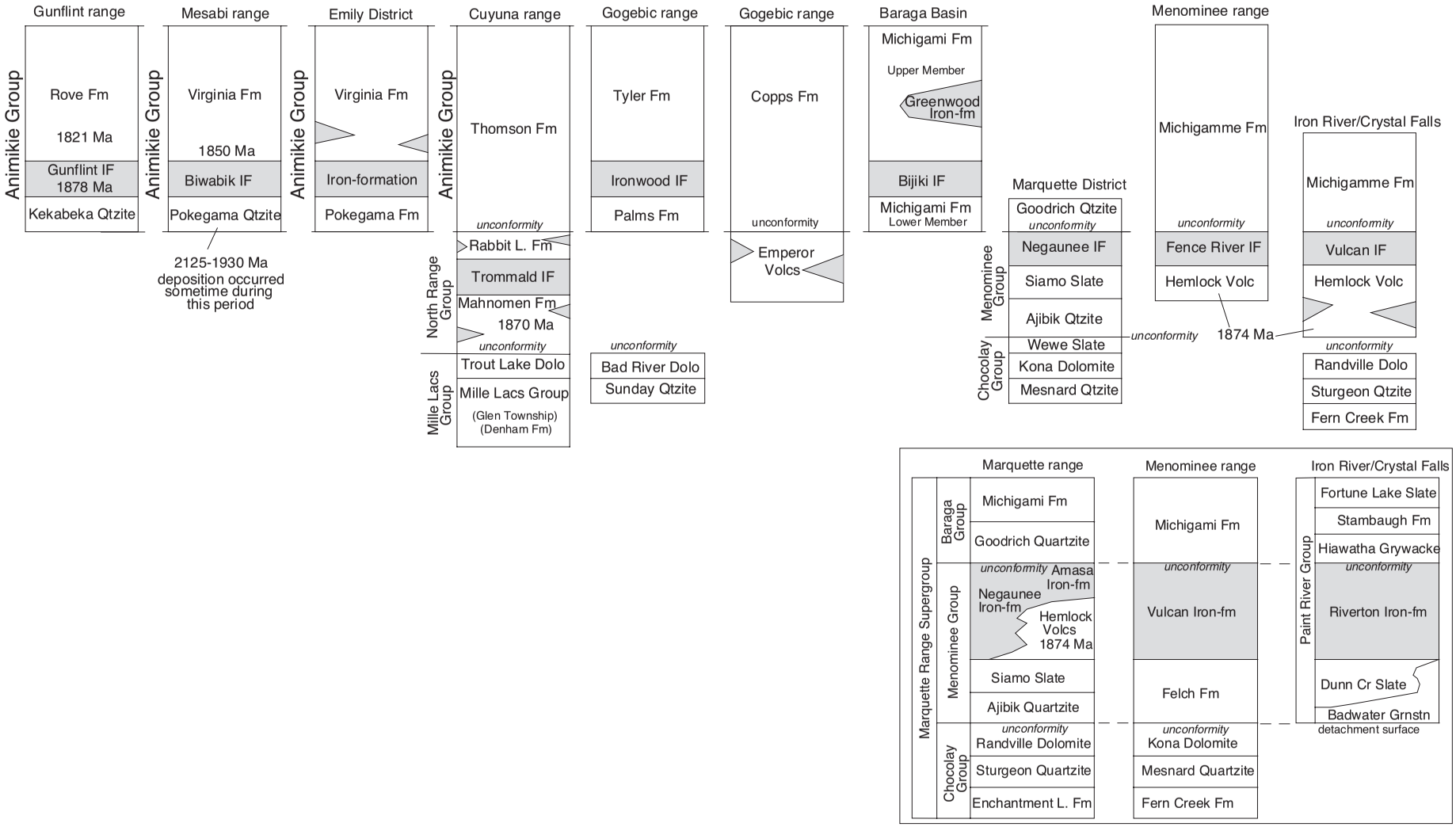


Figure 11.6. Generalized correlation chart of Paleoproterozoic strata in the Lake Superior region (after Morey and Southwick, 1995). Note that recently obtained age dates are shown for the Gunflint Iron Formation, Mahnomen Formation, Hemlock Volcanics, and the Rove and Virginia Formations. Also included 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,

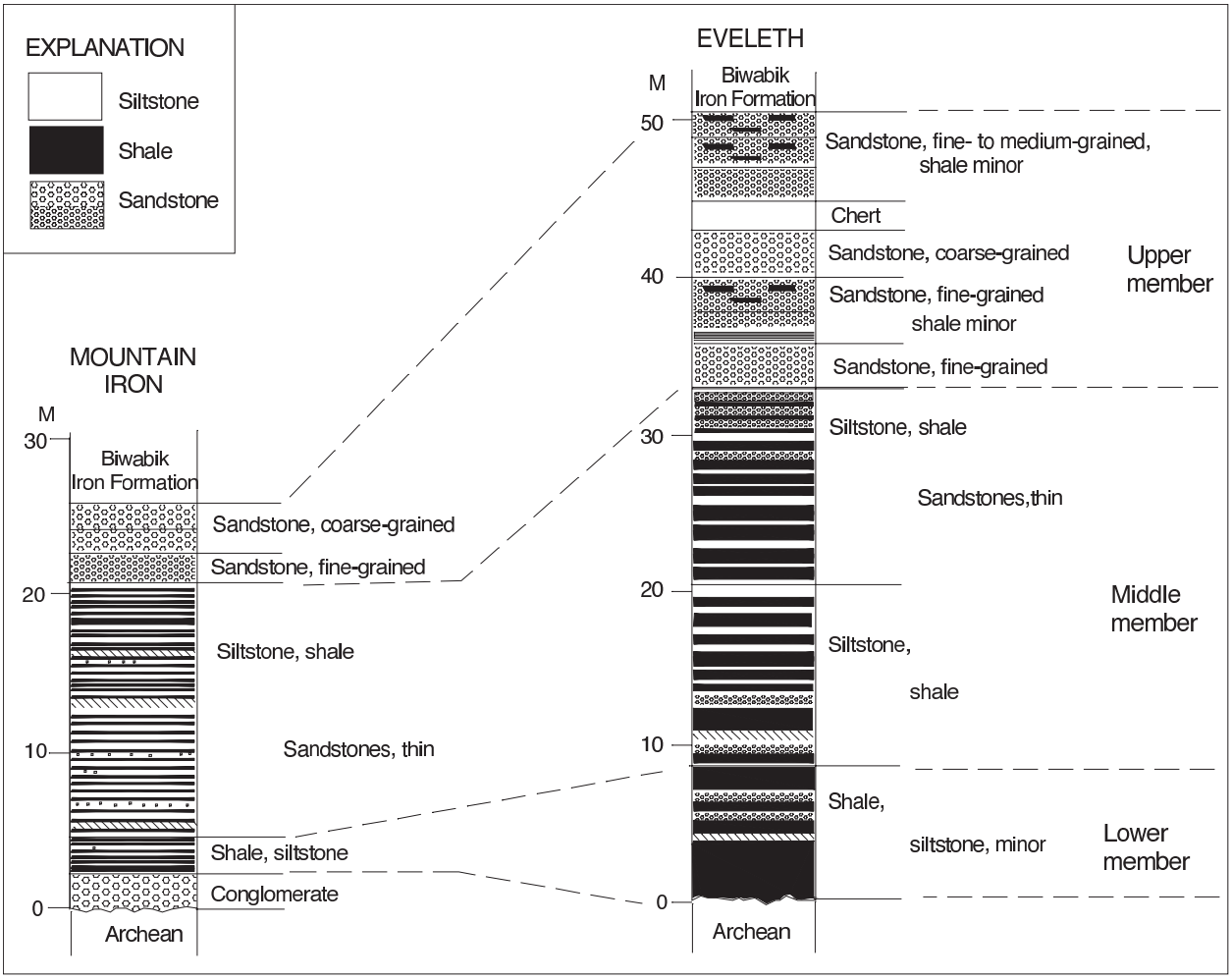


Figure 11.7. Measured sections from two drill holes that penetrate the entire Pokegama Formation. Dark shading represents shale, thin blank units represent siltstone, the slanted pattern represents sandstone and siltstone, and the dotted pattern represents sandstone. Modified from Ojakangas (1983).

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.

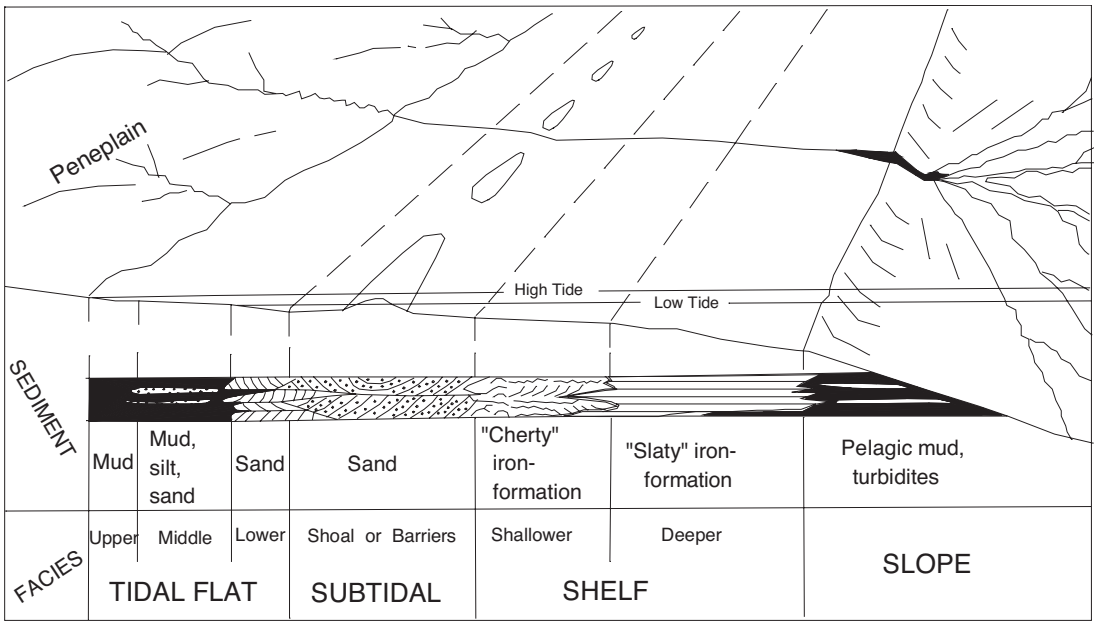


Figure 11.8. Sedimentation model showing lateral relationships of the siliciclastic tidal facies of the Pokegama Formation, the two main facies of the Biwabik Iron Formation, and the Virginia Formation (on the slope?). Thicknesses and geography area not to scale; modified from Ojakangas (1983).

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

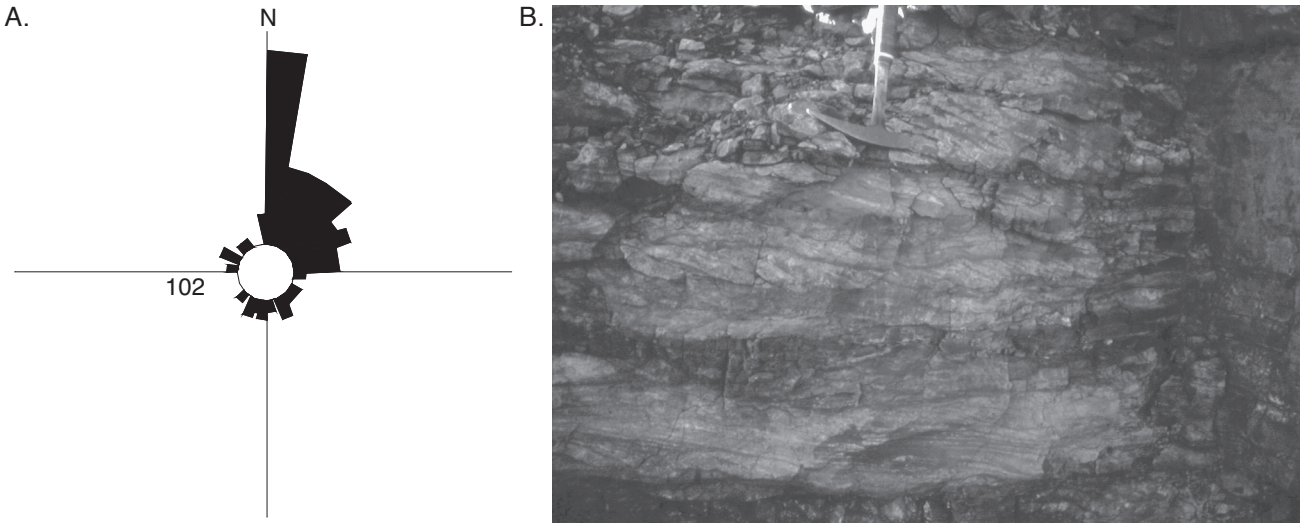


Figure 11.9. **A.** Paleocurrent rose diagram of 102 cross-bed measurements from the Lower Cherty member in the Minorca Mine.

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

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

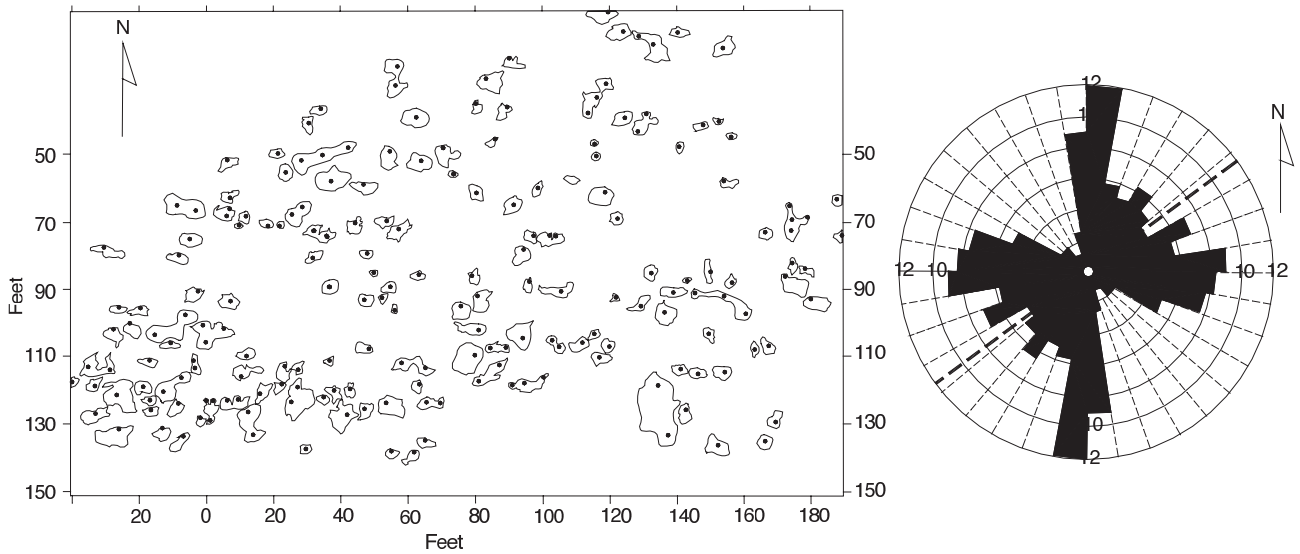


Figure 11.10. Mapped stromatolite mounds in the algal submember (I submember) in the Upper Cherty member of the LTV 2E pit. The rose diagram represents the elongation of the mounds, with each elongate mound plotted on both sides of the rose diagram. From Boerst (1999).

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,

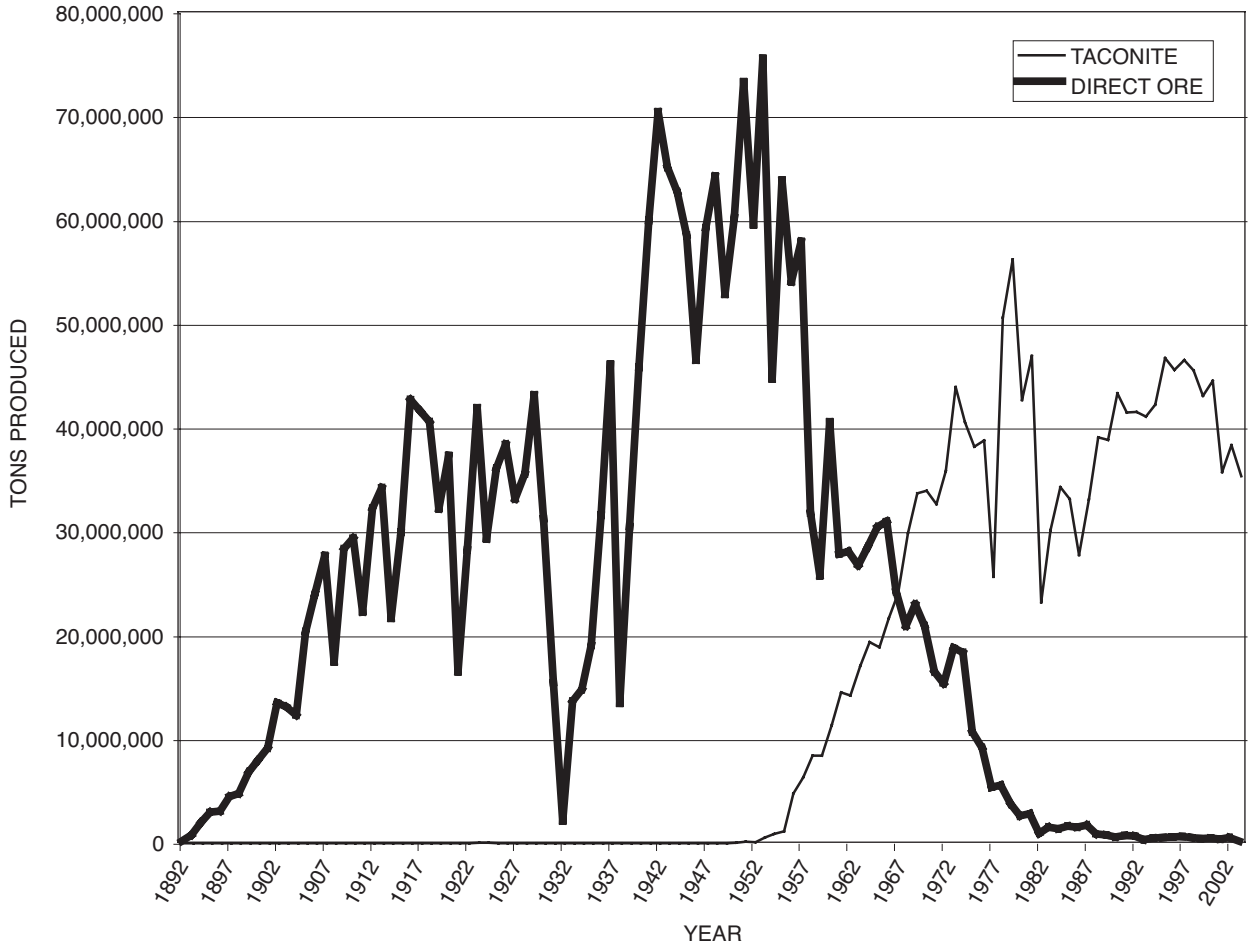


Figure 11.11. Annual production figures for direct ore (includes all forms of direct ore) and taconite for the period from 1892 to 2003 from the Mesabi Iron Range. Data and graph from James Sellner, Minnesota Department of Natural Resources, Division of Lands and Minerals, Hibbing, Minnesota.

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

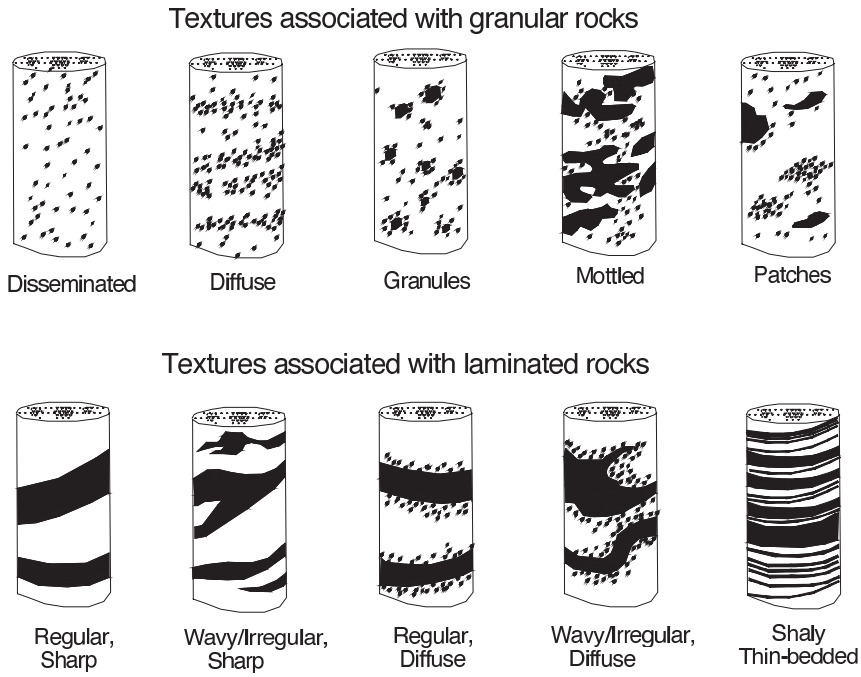


Figure 11.12. Textural characteristics of the Biwabik Iron Formation (from a classification scheme developed by geologists of the Hanna Mining Company; modified from Pfleider and others, 1968).

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

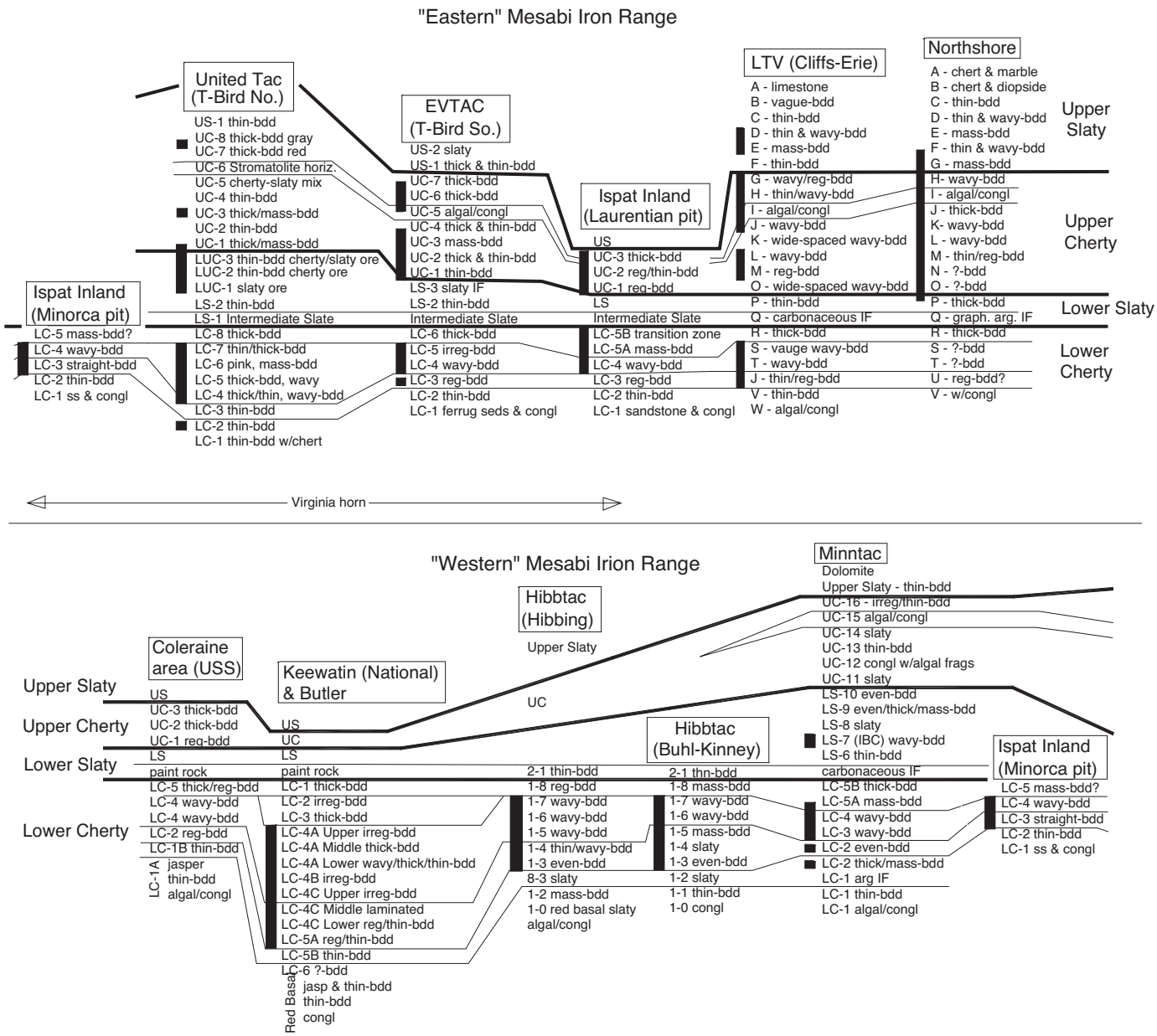


Figure 11.13. Correlation chart of submembers at each of the mines/areas within the Biwabik Iron Formation, as deciphered from published descriptions and mine handouts (modified from Zanko and others, 2003). All columns are hung on the base of the Lower Slaty member ("Intermediate Slate"). It is important 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

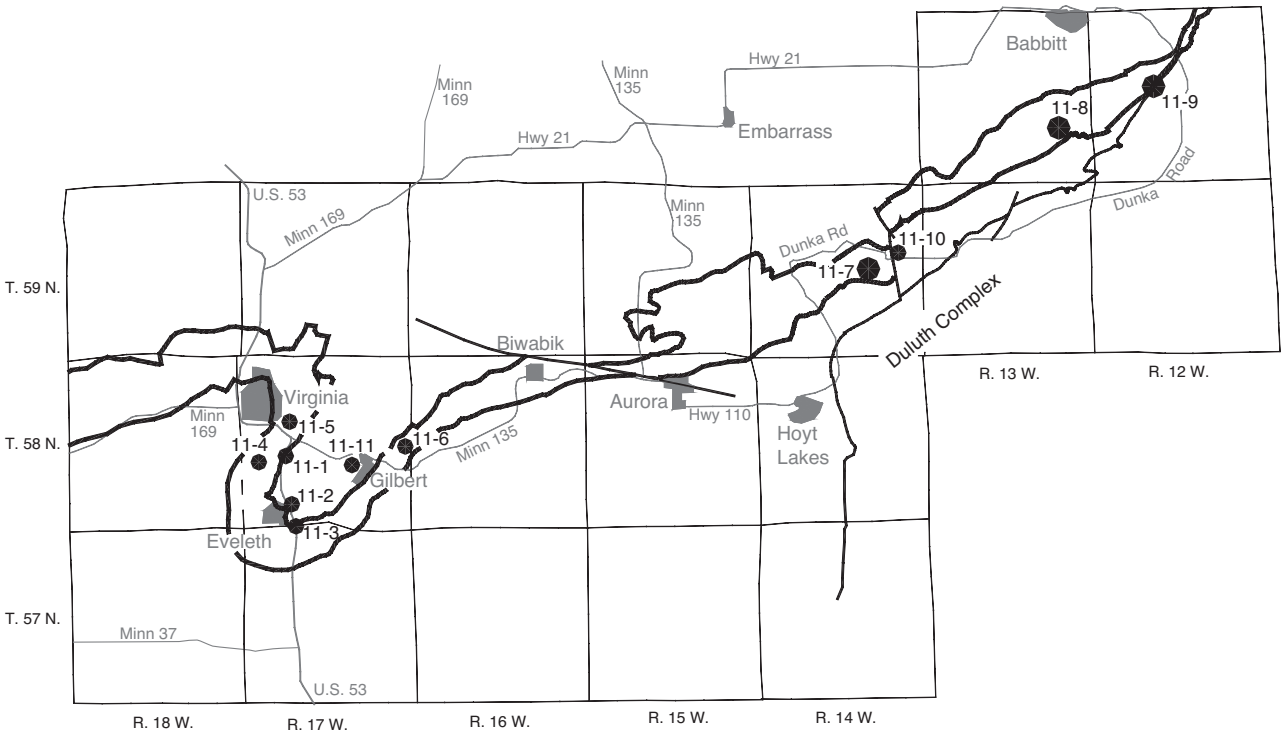


Figure 11.14. Locations of stops that will be visited on the field trip. Generalized contacts of the Biwabik Iron Formation for the eastern half of the Mesabi Iron Range, and Duluth Complex are shown. Note that in order to observe Stops 11-1, 11-2, and 11-3 in stratigraphic order, we will be back-tracking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

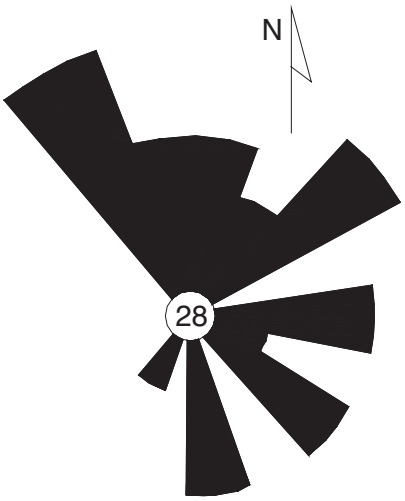


Figure 11.15. Rose diagram of 28 cross-bedded measurements, mostly from the road cut of Stop 11-3, with a few from nearby exposures.

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

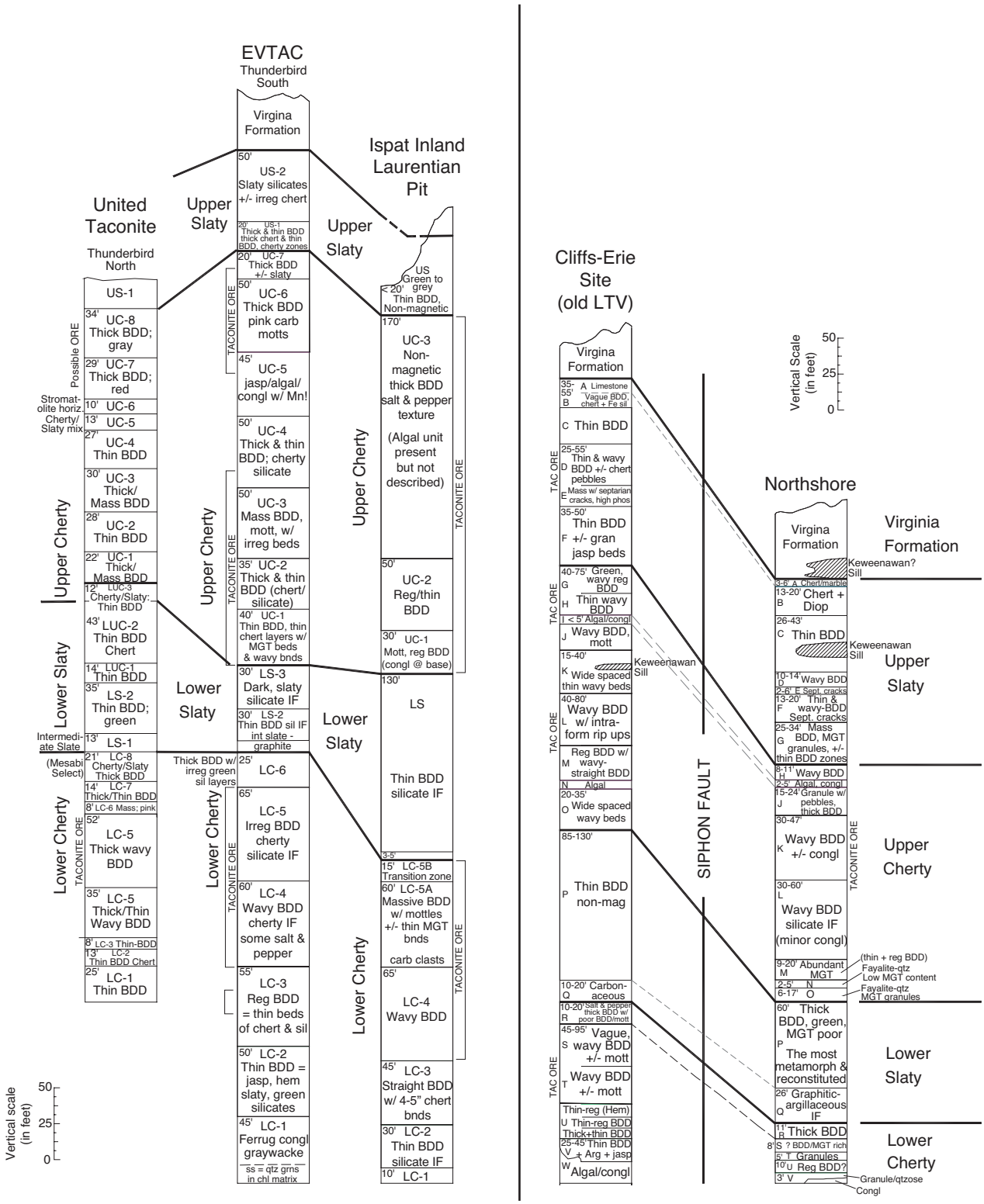


Figure 11.16. Submember nomenclature for the Biwabik Iron Formation as used at the various taconite mines and idled mine sites that will be visited on this field trip. Modified from Plate II in Zanko and others (2003). Mined horizons are shown by a bar on the side of the column. Note that the nomenclature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

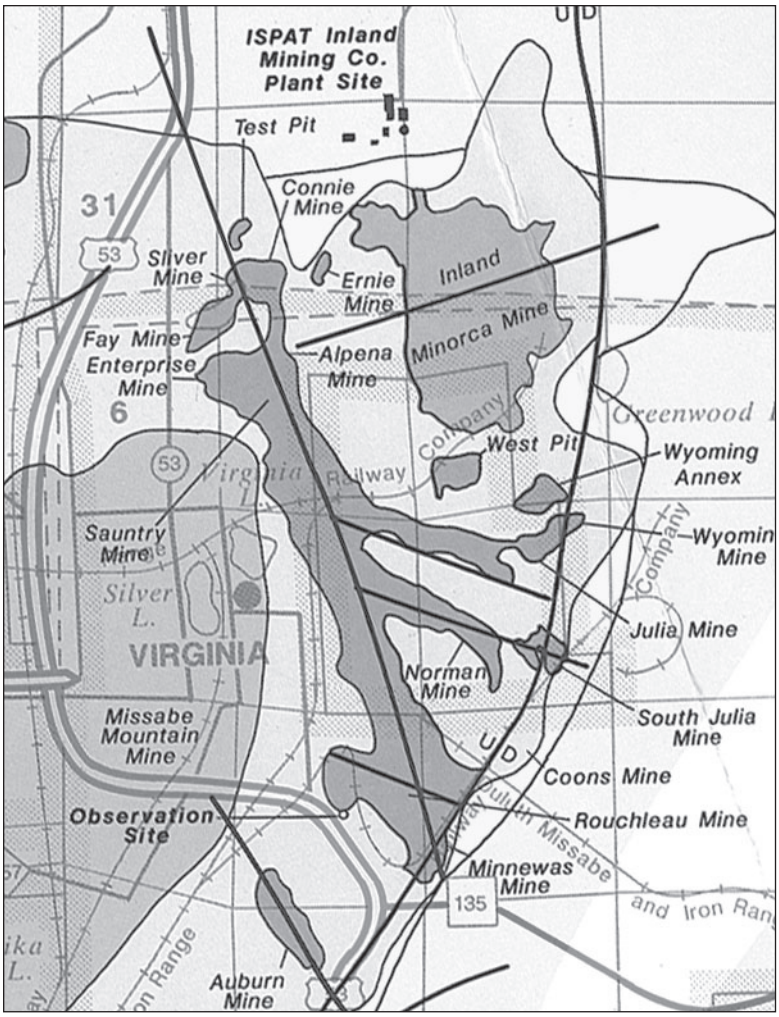


Figure 11.17. General distribution of natural-ore zones (dark shaded) along a series of northwest-trending faults in the Rouchleau Mine and adjacent mines (from Meineke and others, 1999).

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.

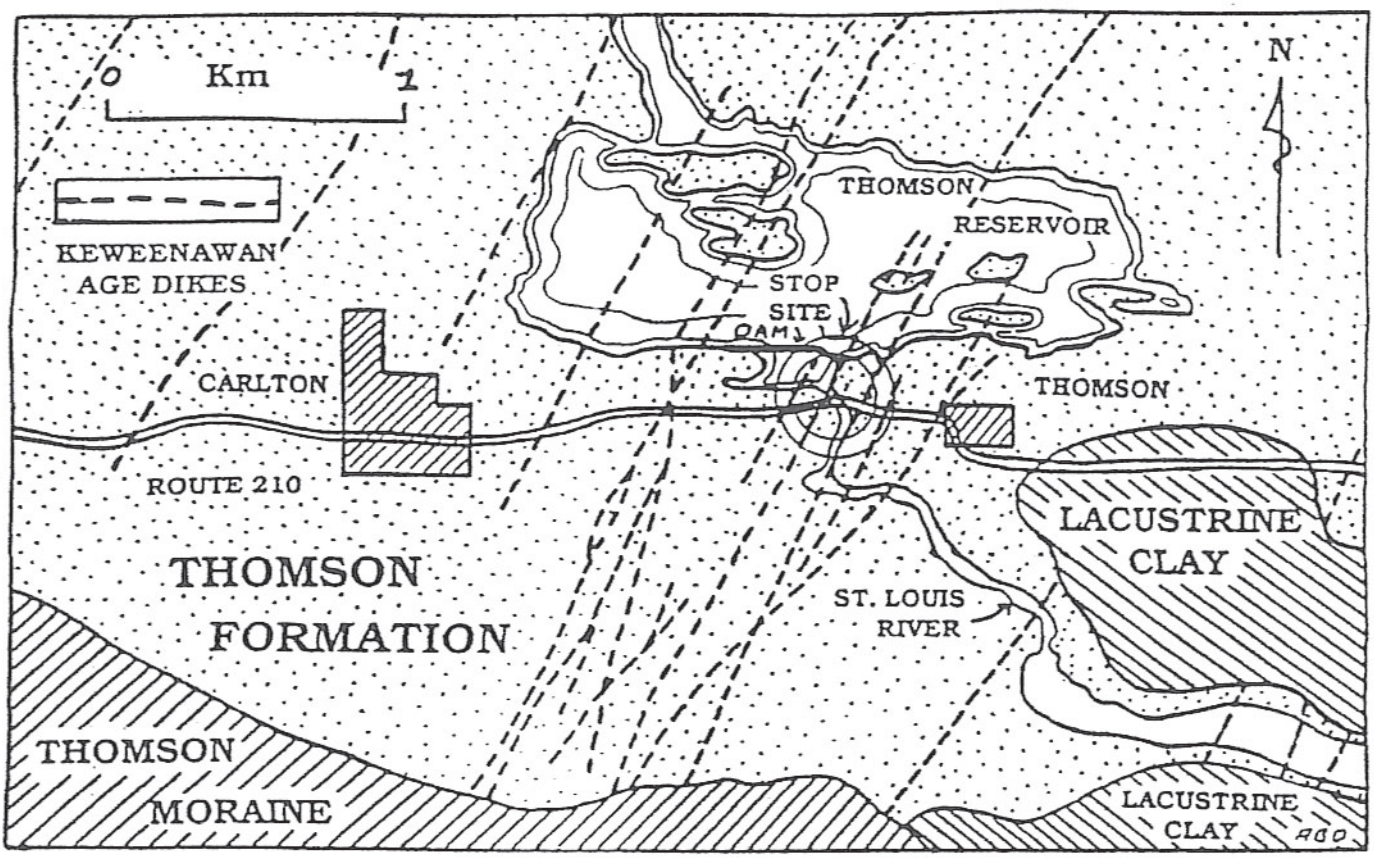


Figure 11.18. Geologic map of Thomson Dam area (from Wright and others, 1970).

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

REFERENCES

Beck, J.W., 1988, Implications for Early Proterozoic tectonics and the origin of continental flood basalts, based on combined trace element and neodymium/strontium isotopic studies of mafic igneous rocks of the Penokean Lake Superior

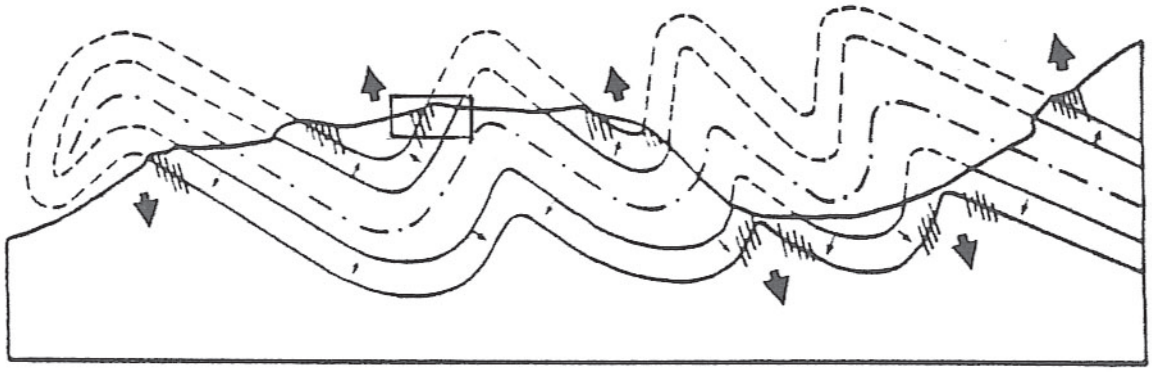


Figure 11.20. Illustration of facing direction of F_2 folds (after Borradaile, 1976). The dashed-dot line is the axial plane of an F_1 fold. Foliation shown diagrammatically is S_2 axial planar crenulation cleavage. Small arrows indicate stratigraphic tops, large arrows indicate facing directions of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

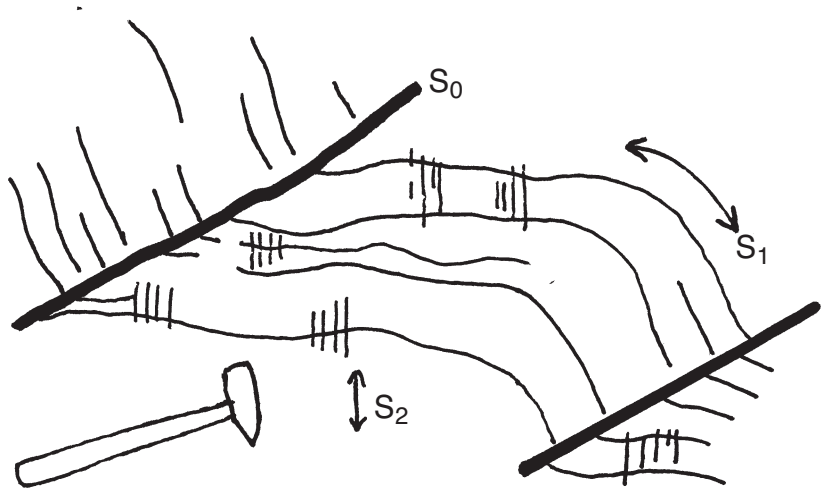


Figure 11.21. Line drawing (modified from Holst, 1992) of geometrical relations of bedding (S_0), early foliation (S_1), and crenulation cleavage (S_2) at the Gillogly Road outcrop (Stop 11-13). Hammer handle is 40 centimeters long.

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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

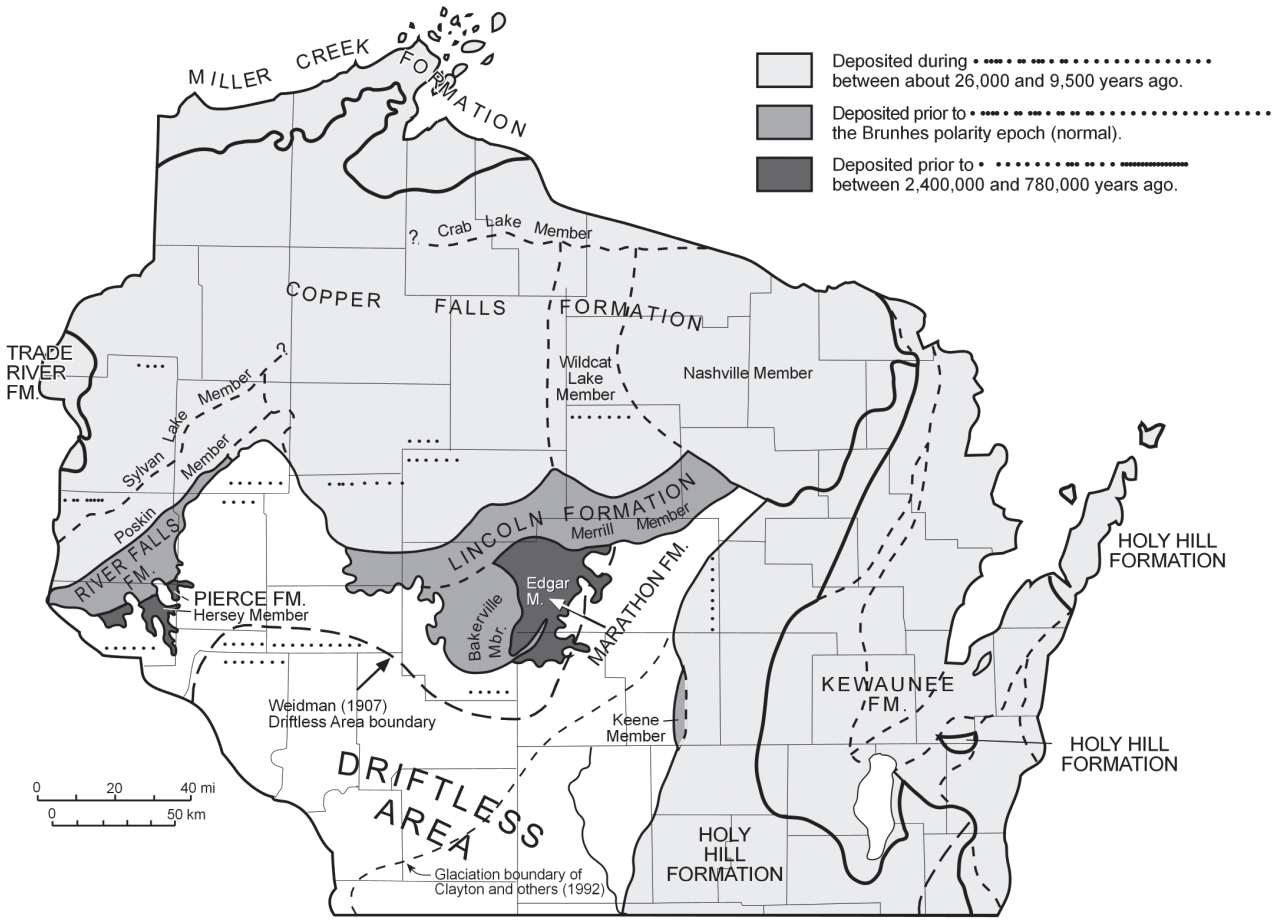


Figure 12.1. Pleistocene lithostratigraphic units for till in northern Wisconsin. Short dashed lines in the glaciated areas designate member boundaries; modified from Clayton and others (1992) using information from Syverson and Colgan (2004).

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.

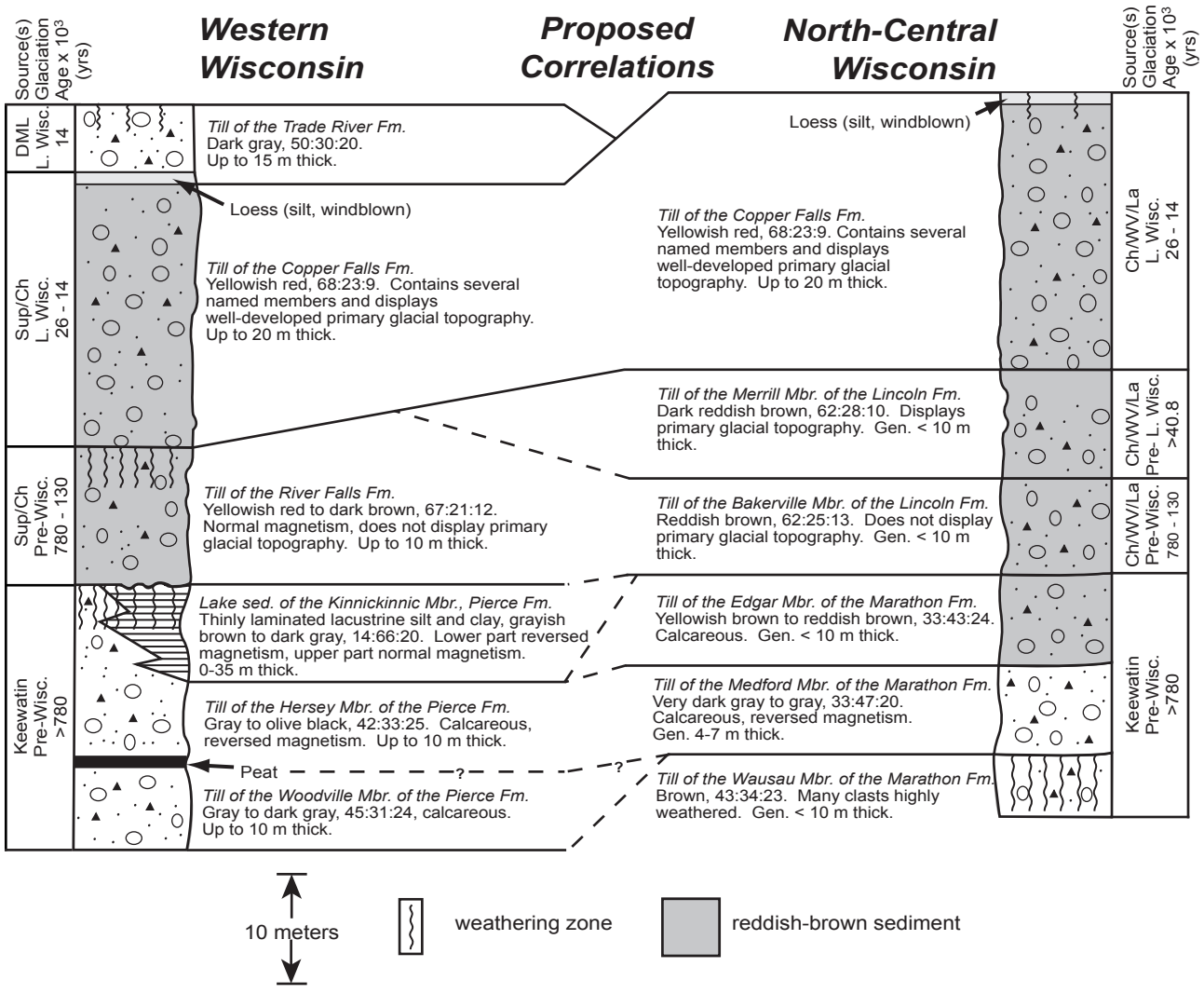
Glaciations prior to the Wisconsinan glaciation

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 x 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 x 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 x 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 x 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 x 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 x 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)



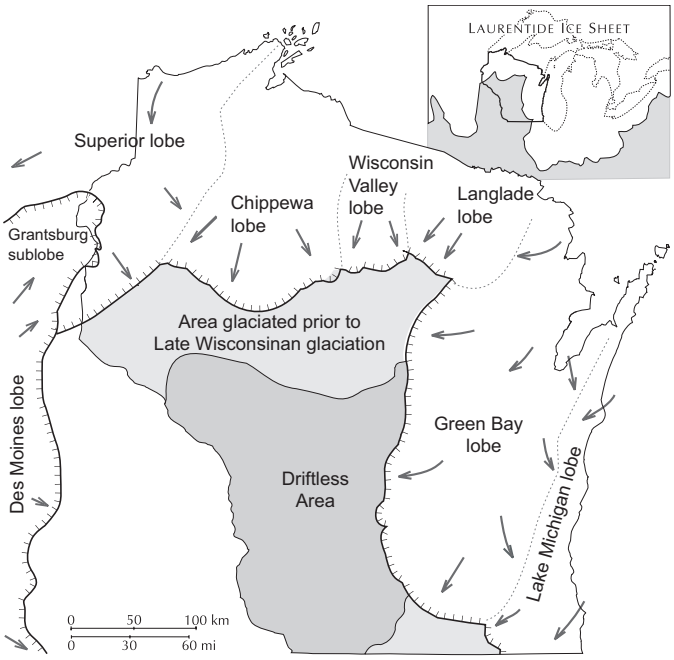


Figure 12.3. Ice lobes of the Laurentide Ice Sheet at the maximum flow extent during the Late Wisconsinan glaciation. Arrows indicate direction of ice flow; modified from Johnson (2000).

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).

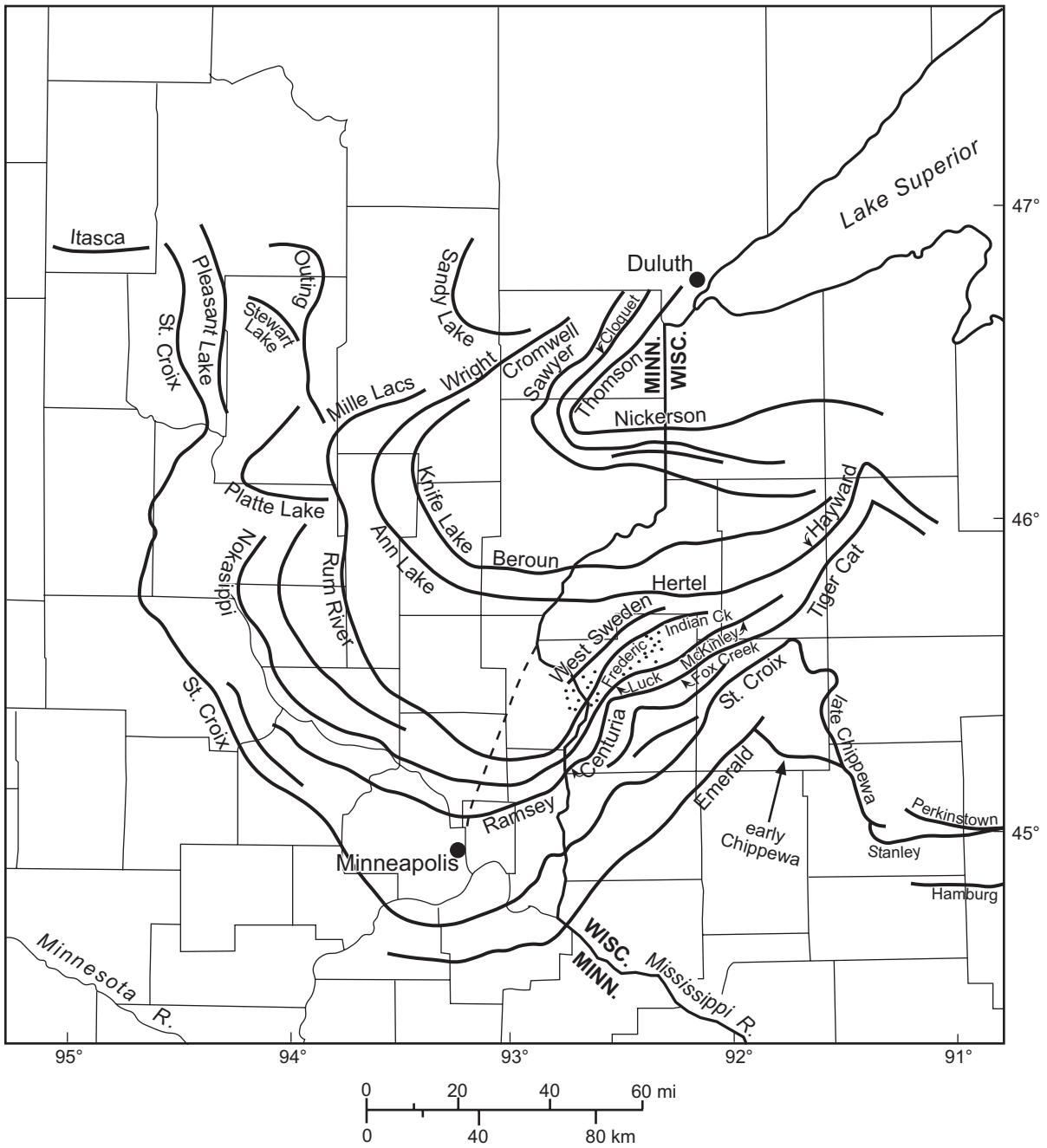


Figure 12.4. Hypothetical correlation of Late Wisconsinan ice margins of the Superior lobe in Wisconsin and Minnesota and a portion of the Chippewa lobe. The dashed line represents the former path of the St. Croix River prior to Late Wisconsinan glaciation. The stippled area is an outcrop of basalt hills 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

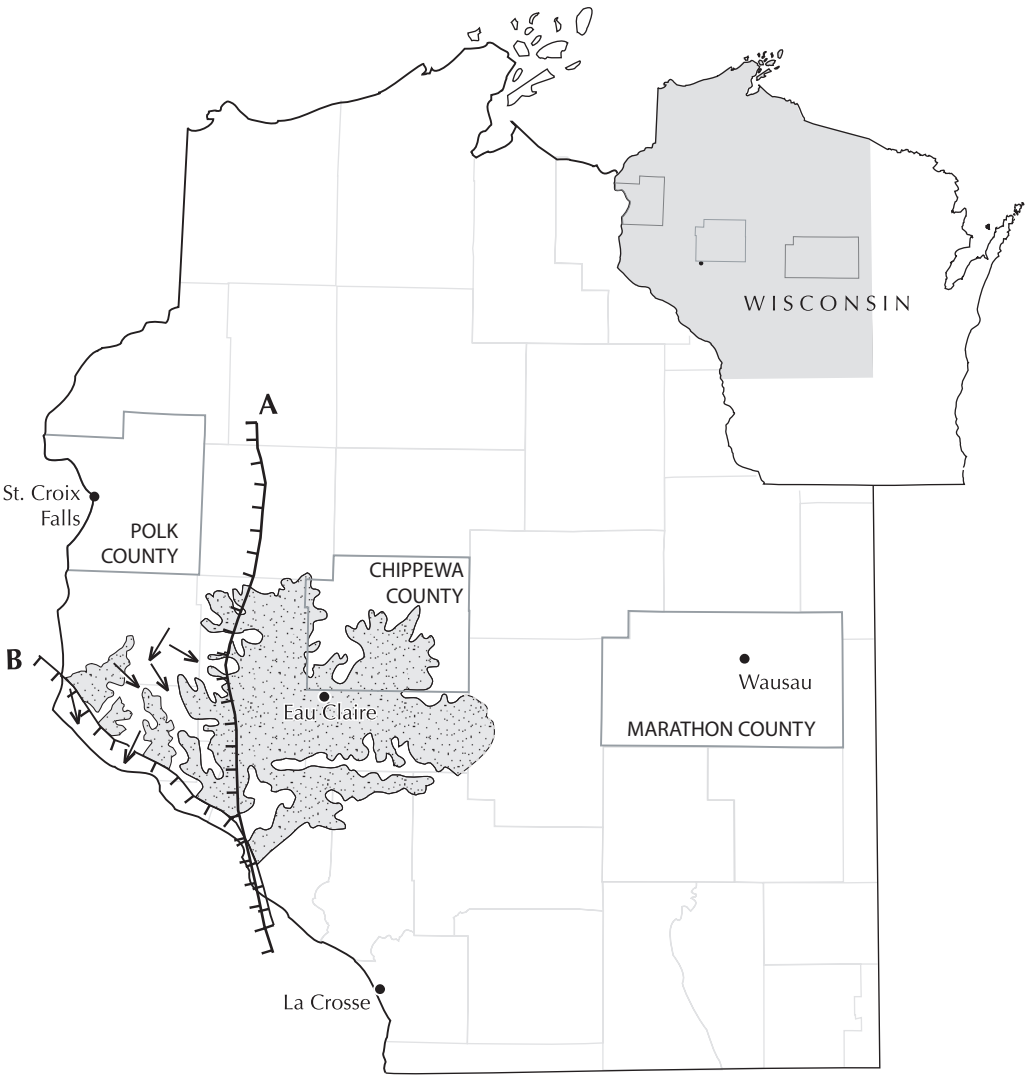


Figure 12.5. Ice-margin positions (hachured lines), ice-flow directions from pebble fabrics (arrows), and ice-dammed lakes (stippled areas) associated with Keewatin ice flowing from the northwest during the Reeve Phase (modified from Baker and others, 1983, and Johnson, 2000). The Kinnickinnic Member 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

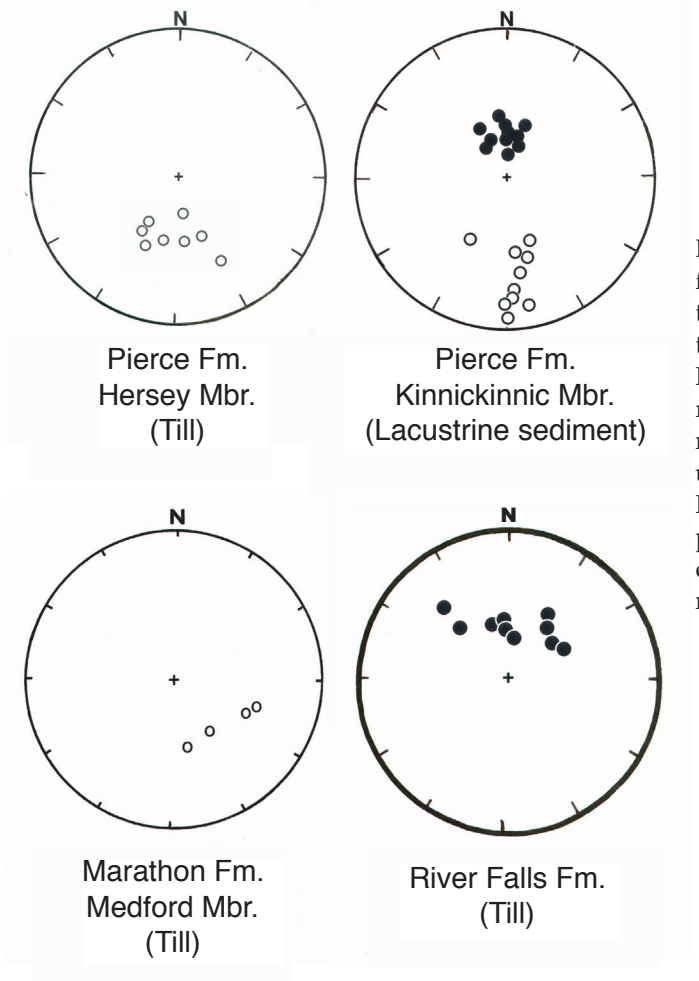


Figure 12.6. Paleomagnetic remanent directions for the Hersey and Kinnickinnic Members of the Pierce Formation, the Medford Member of the Marathon Formation, and the River Falls Formation. Projection is equal area. Solid dots represent lower hemisphere projections (normal magnetic polarity) and the open 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).

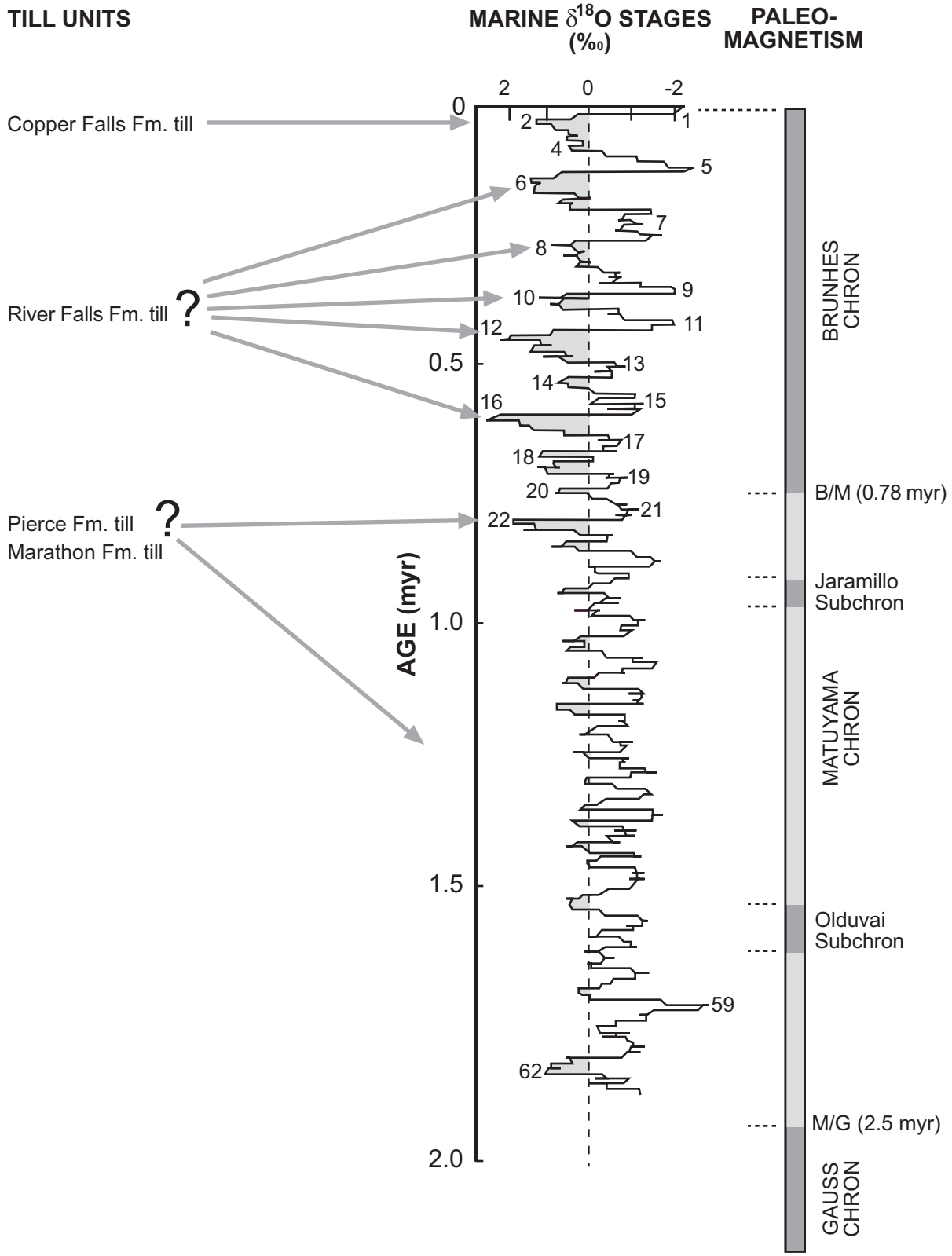


Figure 12.7. Oxygen isotope curve showing stages associated with glaciations (even numbers, shaded), interglacials (odd numbers), and the associated paleomagnetic chrons. Pierce and Marathon Formation tills have reversed polarity and probably were deposited sometime prior to stage 20. The River Falls 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,

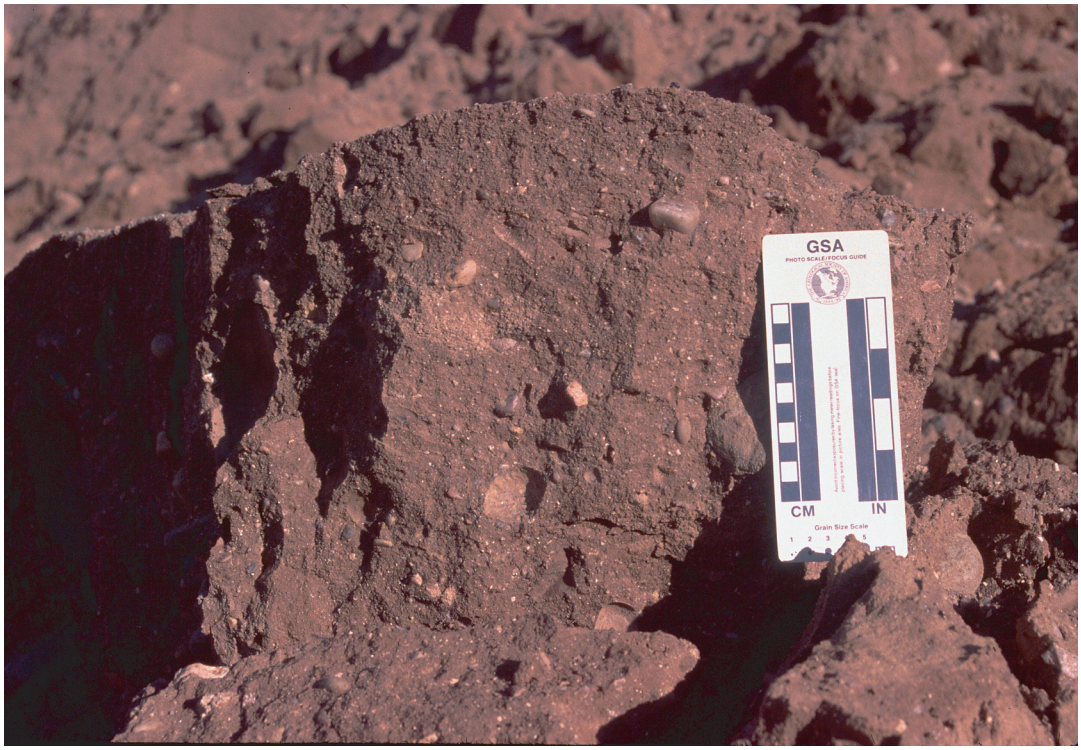


Figure 12.8. Clay cementation of River Falls Formation stream sediment in the soil B horizon at Stop 12-6. From Syverson (in press).

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

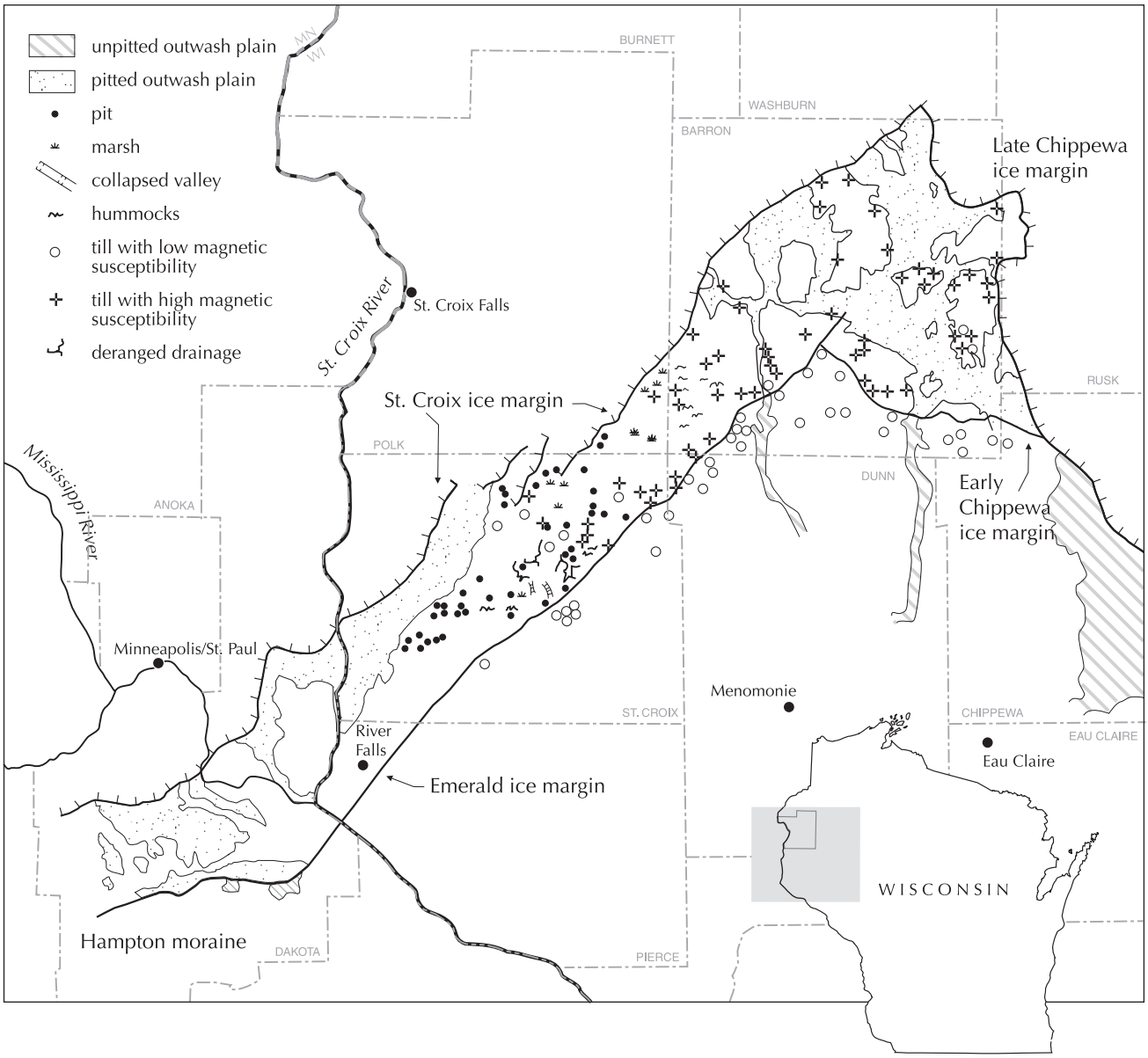


Figure 12.9. Ice-margin positions (hachured lines), geomorphic features, and magnetic susceptibility values for till associated with the Emerald Phase of the Superior lobe. Geomorphic features interpreted from topographic maps of southern Polk County, St. Croix County, western Pierce County, northern 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 x 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.

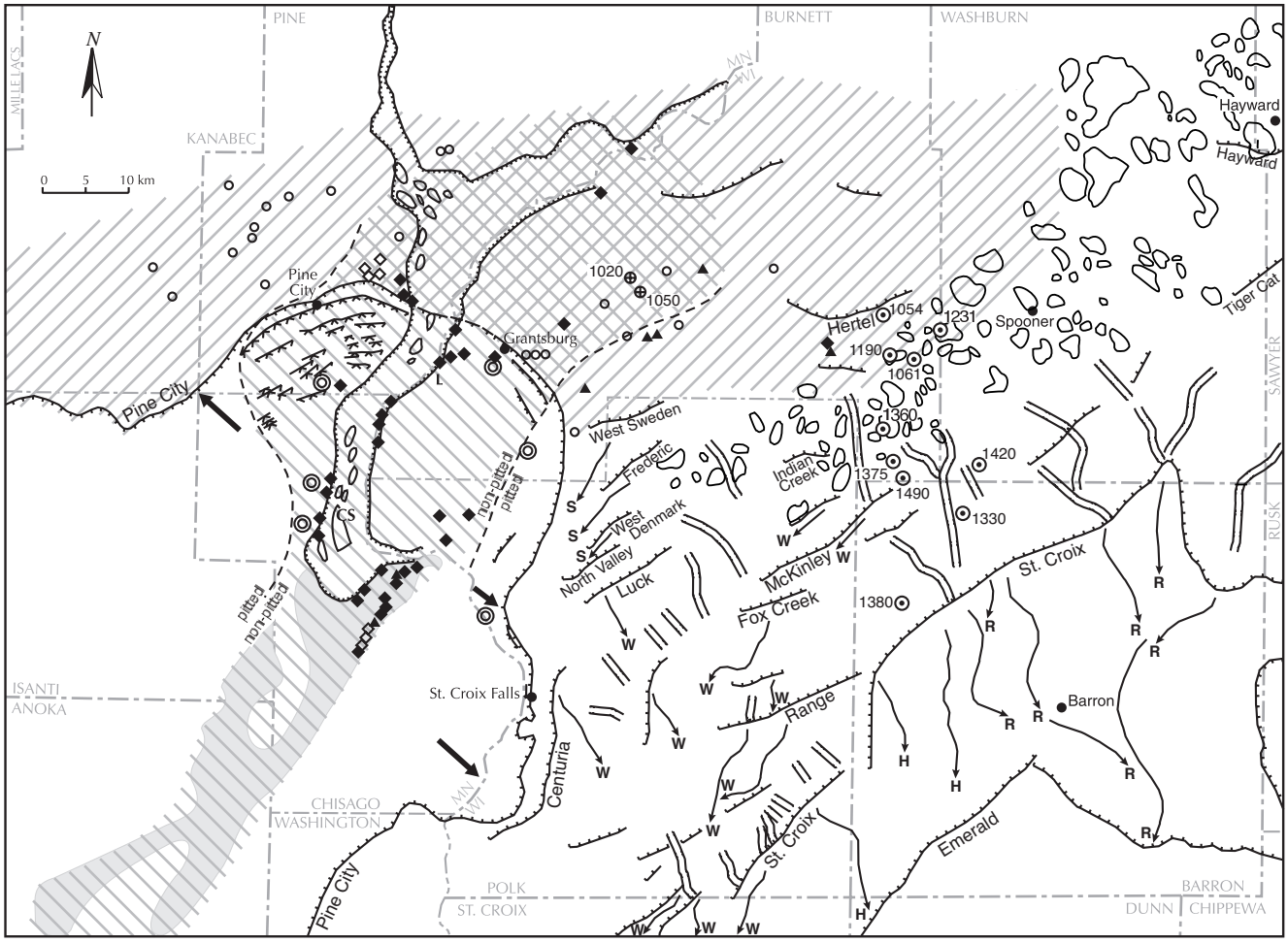
The age of the Emerald and Early Chippewa Phases is not known, but it is certainly Late Wisconsinan, probably between 20,000 and 25,000 ^{14}C yr B.P. Ice during the Late Wisconsinan glaciation crossed the drainage divide south of Lake Superior by 26,000 ^{14}C yr B.P. This is based on spruce wood in 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,

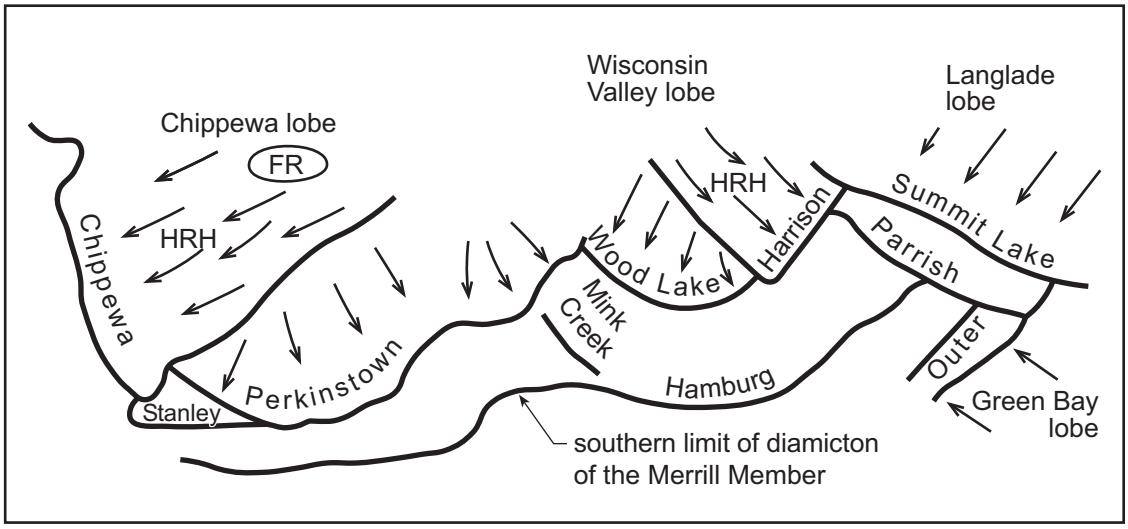


Figure 12.11. Ice-flow directions and moraines of the Wisconsinan glaciation. The high-relief, hummocky (HRH) Chippewa and Harrison moraines formed after the Chippewa and Wisconsin Valley lobes underwent major changes in ice-flow direction. These major changes of flow direction are thought to represent 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

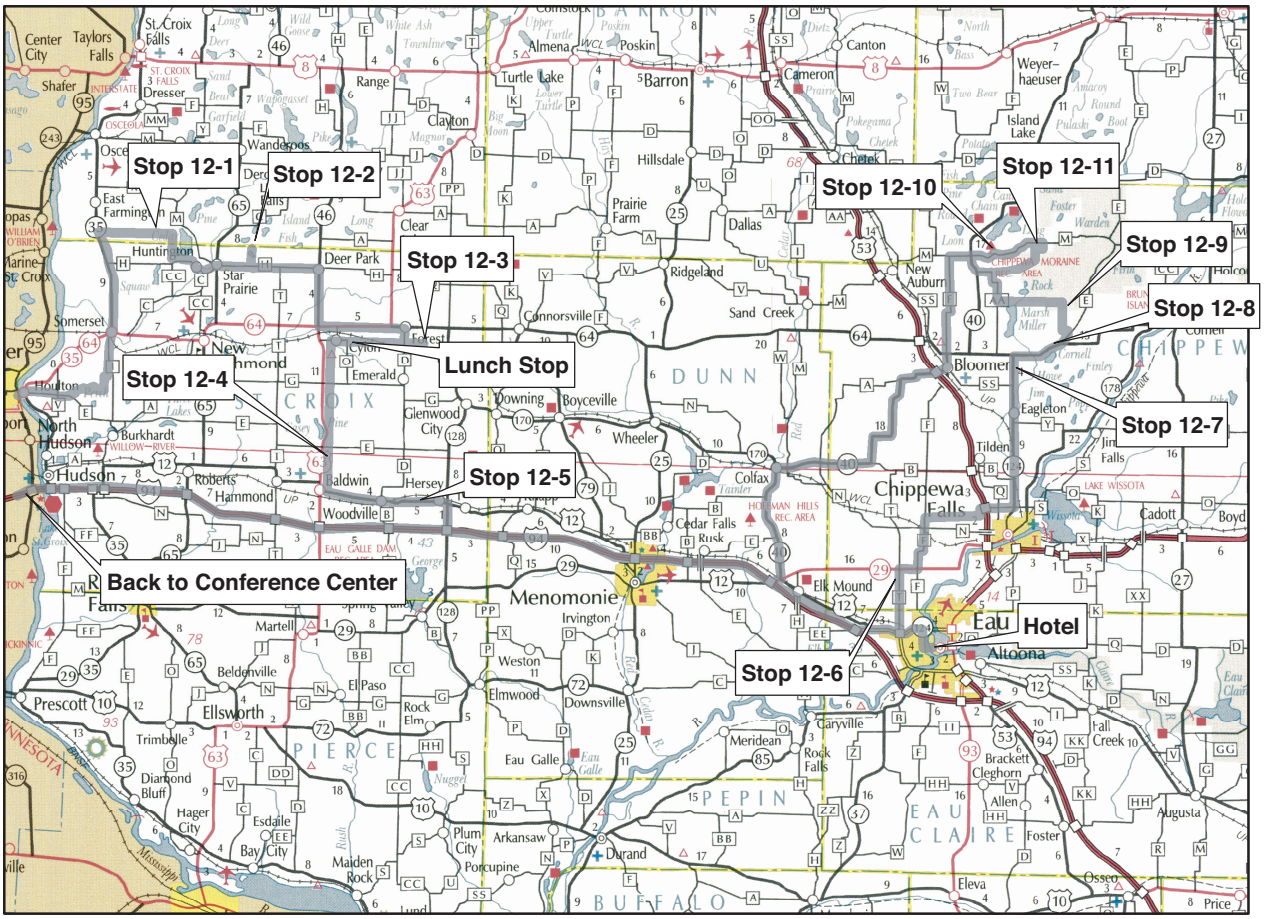


Figure 12.12. Field trip route and stops.

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.

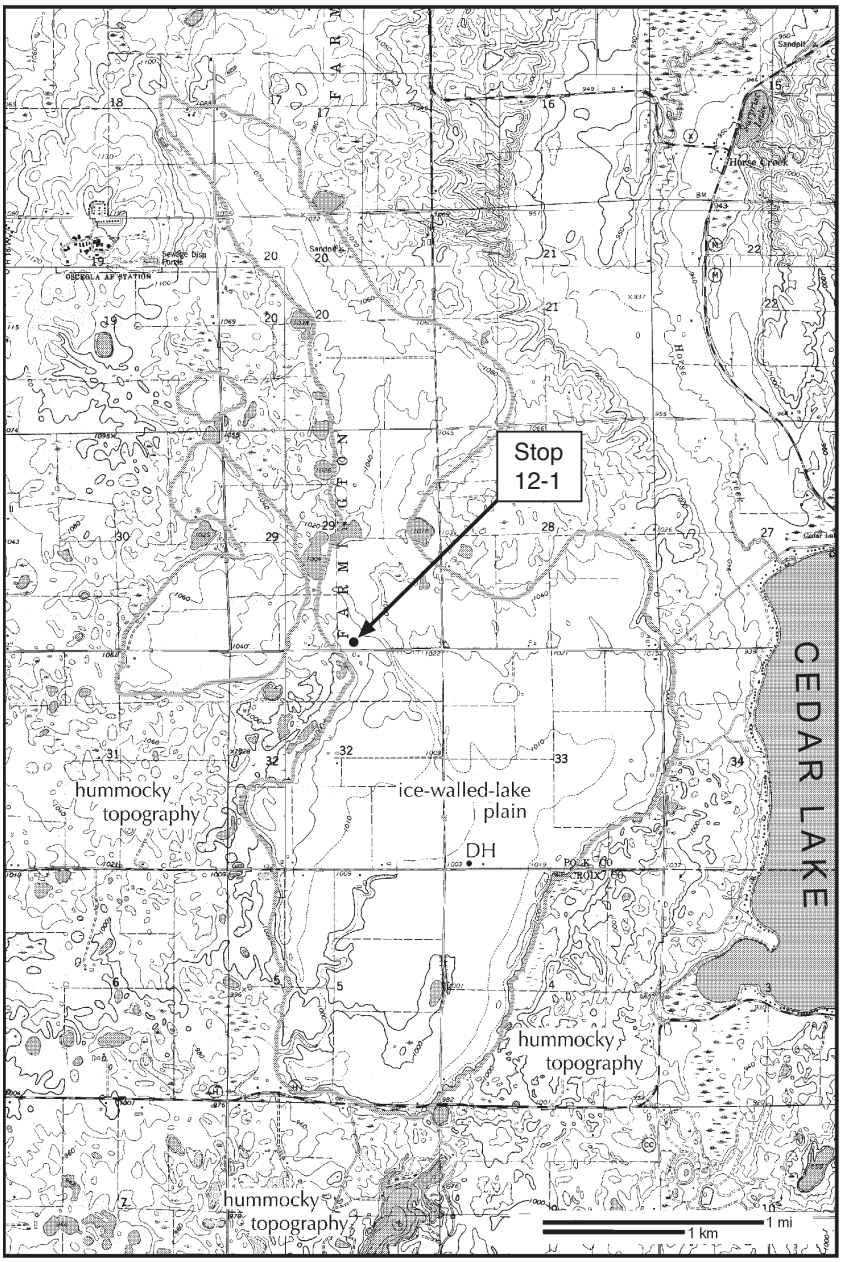


Figure 12.13. Parts of New Richmond North, Nye, Osceola, and Somerset North quadrangles, Wisconsin and Minnesota showing the Cedar Lake ice-walled-lake plain line surrounded by hummocky topography. Contour interval is 10 feet, except in the southwestern part, where it is 20 feet. Hummocks in this region 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.

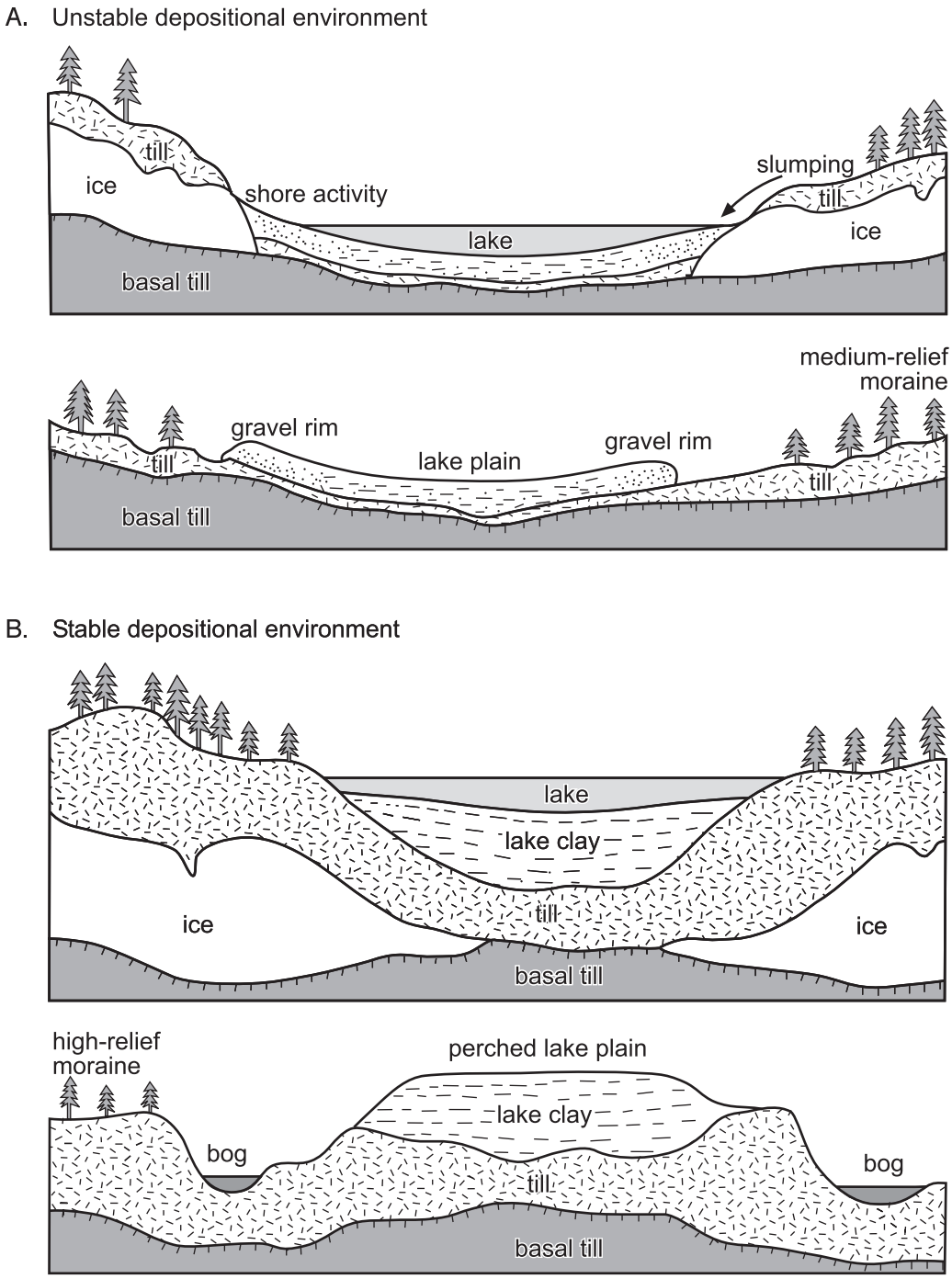


Figure 12.14. The formation of ice-walled-lake plains (modified from Clayton and Cherry, 1967).

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

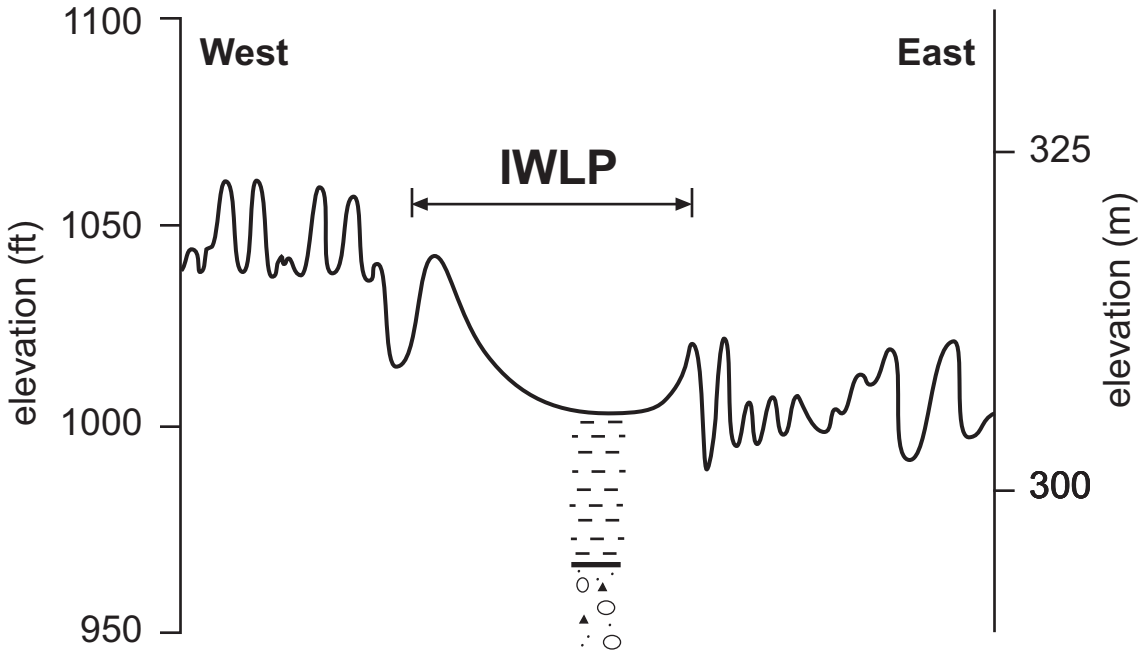


Figure 12.16. Topographic profile across the southern end of the Cedar Lake ice-walled-lake plain showing bordering hummocks and drill hole information. Location runs east–west through the drill hole (marked DH) in Figure 12.13 (but the cross section line is not shown on that figure). Note that the ice-walled-lake 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.

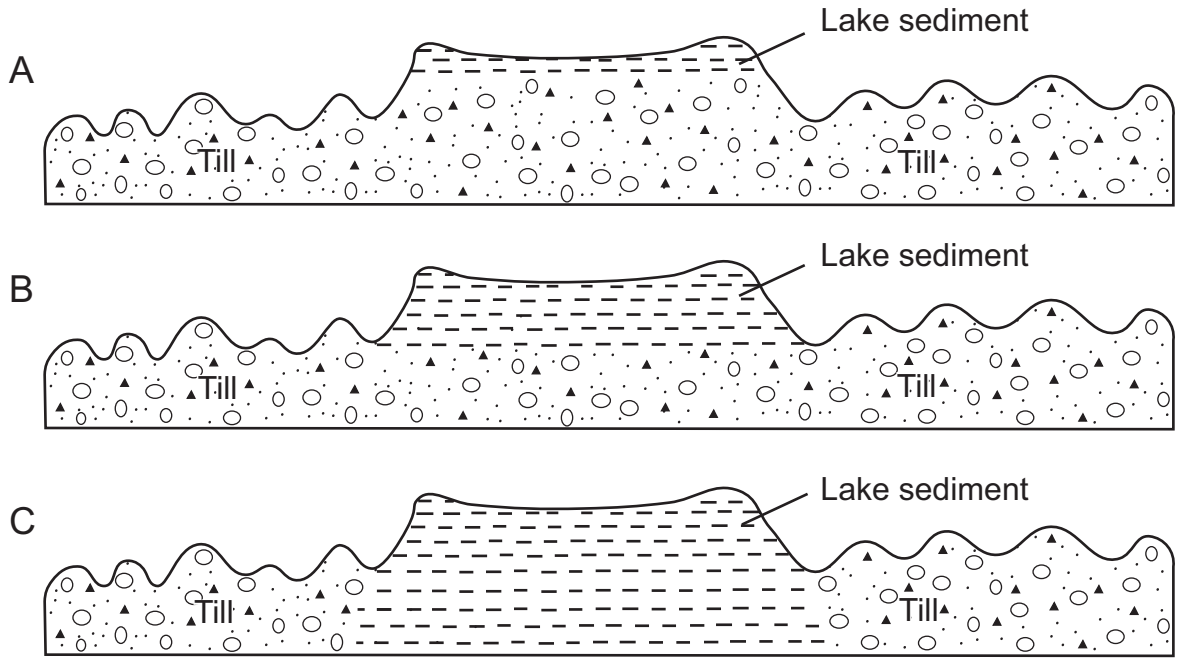


Figure 12.17. Ice-walled-lake plains with varying thicknesses of lake sediment. Types A and B are quite common and have been reported widely in Wisconsin, the Dakotas, southern Canada, and Europe (Johnson and Clayton, 2003). The Cedar Lake ice-walled-lake plain is an example of type C. See text for explanation.

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

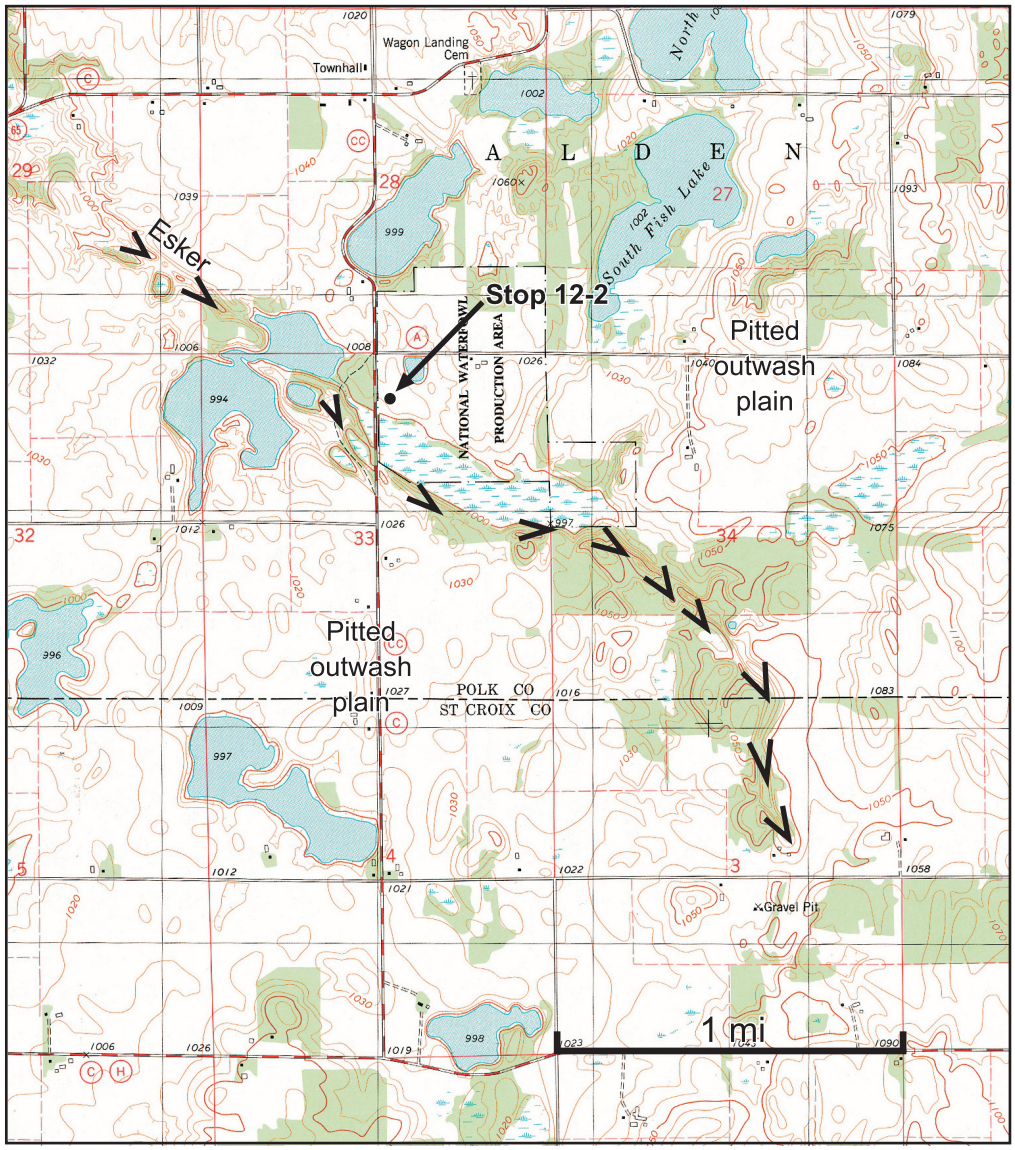


Figure 12.18. Topographic map of the County Road CC esker (crest indicated by symbols) and tunnel channel taken from the Deer Park quadrangle, Wisconsin. This esker at Stop 12-2 is about 6 kilometers long, and it lies within a tunnel channel that is obscured by collapsed outwash.

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

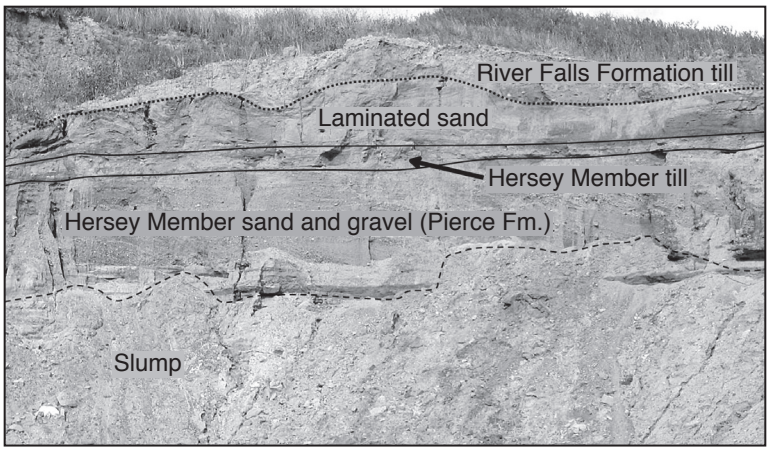


Figure 12.19. Exposure along the northeastern wall of the pit at Stop 12-3. Hersey Member of the Pierce Formation outwash and till is unconformably overlain by outwash sand and till of the River Falls Formation. Vertical scale approximately 10 meters.

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

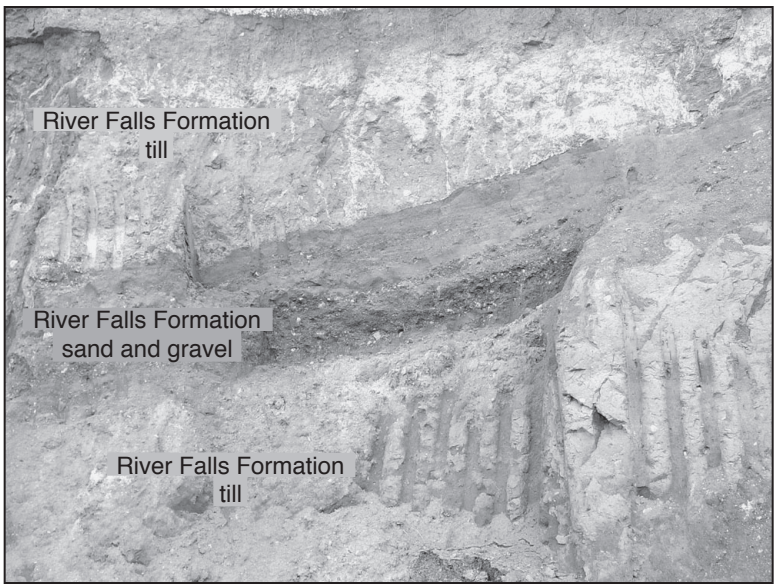


Figure 12.20. Lens of River Falls Formation sand and gravel (approximately 1 meter thick) within River Falls Formation till that is exposed at Stop 12-4.

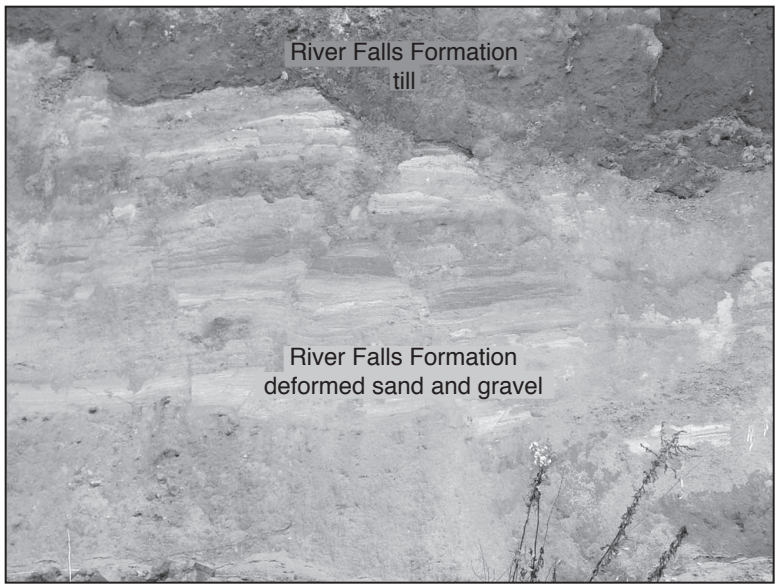


Figure 12.21. Deformed River Falls Formation sand and gravel, approximately 2.5 meters thick, underlying River Falls Formation till at Stop 12-4.

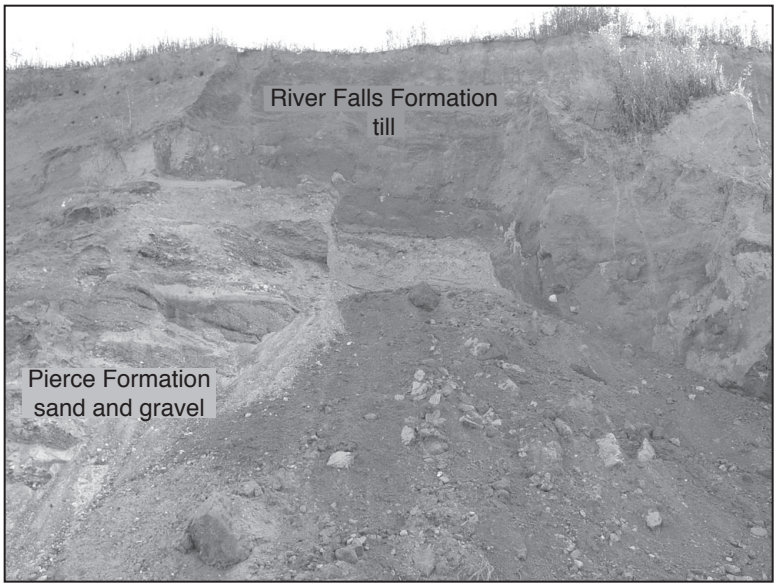


Figure 12.22. Pierce Formation sand and gravel underlying 1 to 4 meters of River Falls till at Stop 12-4. See stop description for possible origins of the faulting.

(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

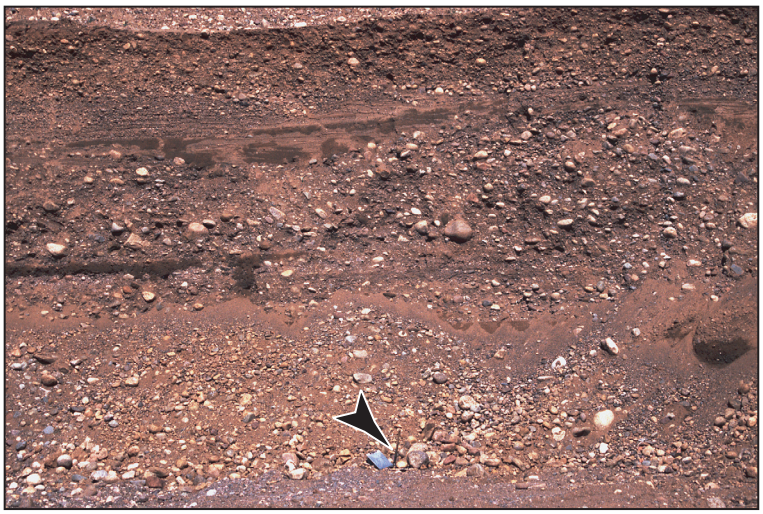


Figure 12.24. Coarse-grained proximal stream sediment of the Copper Falls Formation at Stop 12-7. The gravel and sandy gravel fine upward and were deposited within 100 meters of the Chippewa moraine by water flowing from right to left. This gravel-rich sediment is an ideal source of commercial aggregate. 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,

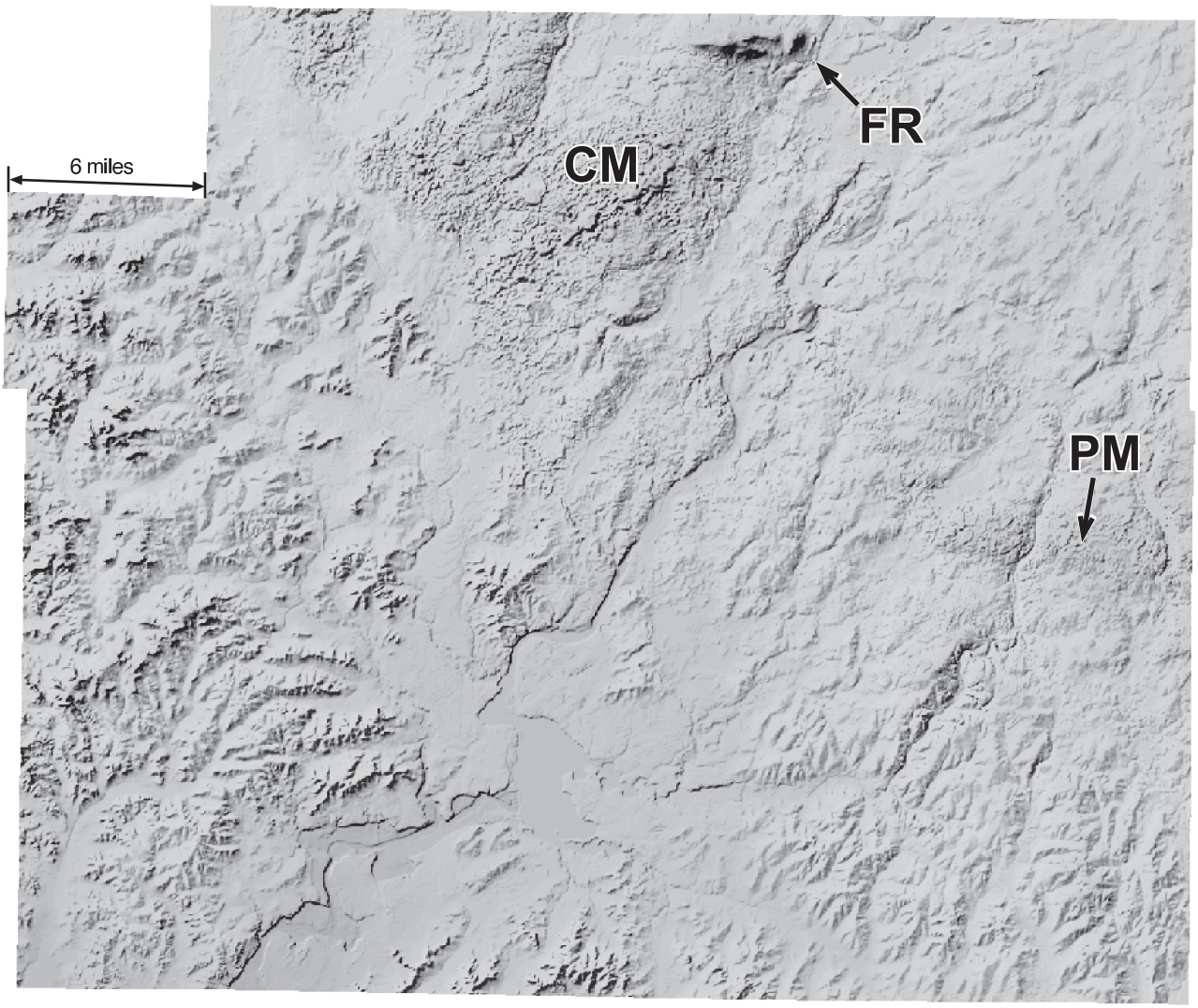


Figure 12.27. Shaded-relief digital elevation map of Chippewa County. The highest-relief portion of the Chippewa moraine (CM) is directly downflow (southwest) of the Flambeau Ridge (FR), the high quartzite knob. The Perkinstown moraine (PM) is indicated, and the incised Paleozoic bedrock surface is clearly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

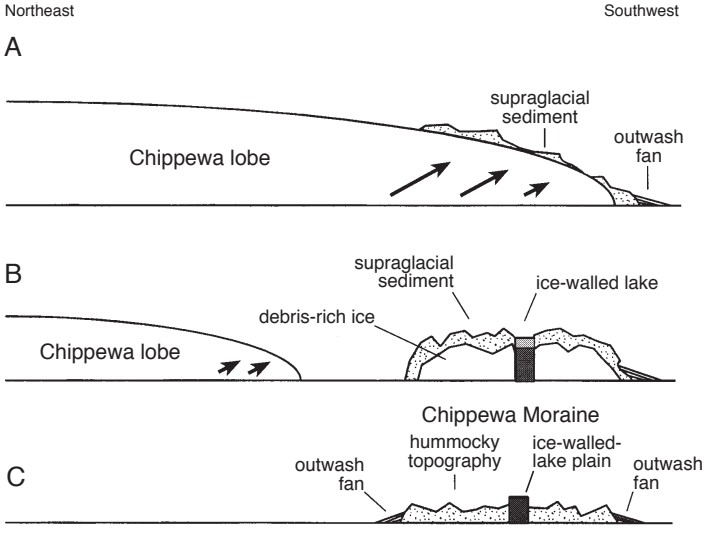


Figure 12.28. Formation of the Chippewa moraine; modified from Ham and Attig (1997).

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

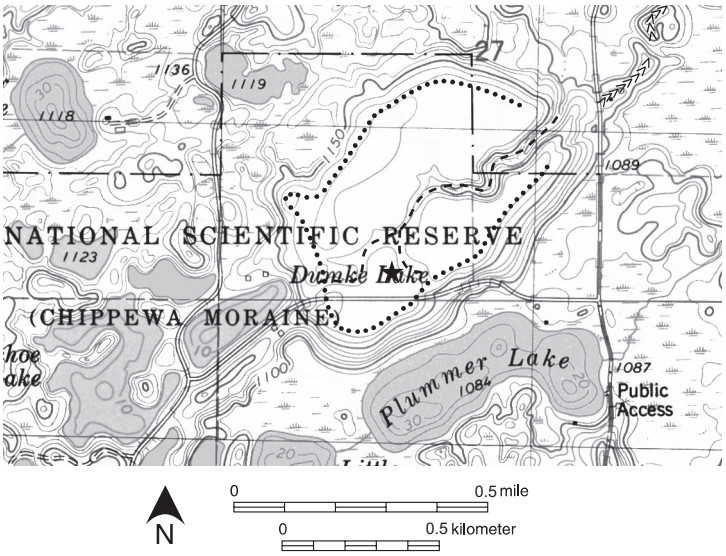


Figure 12.29. Part of the Bob Lake quadrangle showing the Plummer Lake unstable-environment ice-walled-lake plain. The outermost edge of the ice-walled-lake plain is shown by dots. Stop 12-11 coincides with a drill hole location (marked by the star) where drilling penetrated 23 meters of lake sediment 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.

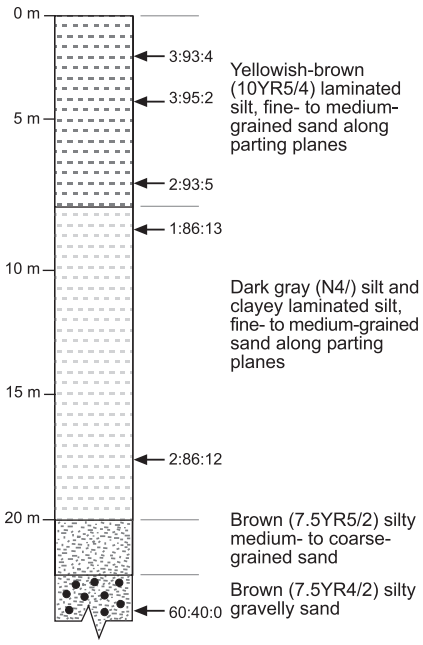


Figure 12.30. Offshore sediment log for the Plummer Lake ice-walled-lake plain, Chippewa Moraine Ice Age National Scientific Reserve, Wisconsin (see Fig. 12.29 for location). Sand:silt:clay ratios are indicated at the appropriate sample depths. The laminated silt and clay settled from suspension in the 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).

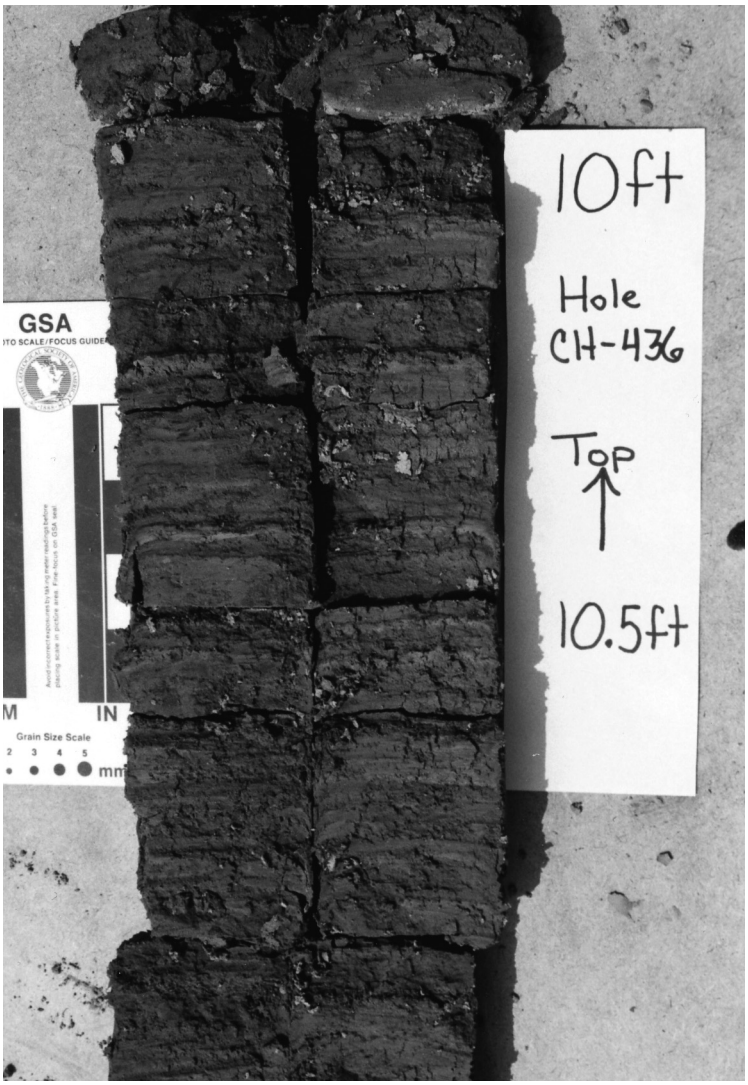


Figure 12.31. Laminated silt and silt loam deposited in an offshore environment, Plummer Lake ice-walled-lake plain, Chippewa Moraine Ice Age National Scientific Reserve (drill hole depth 10 feet [3 meters]), see Fig. 12.29 for location). From Syverson (in press).

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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