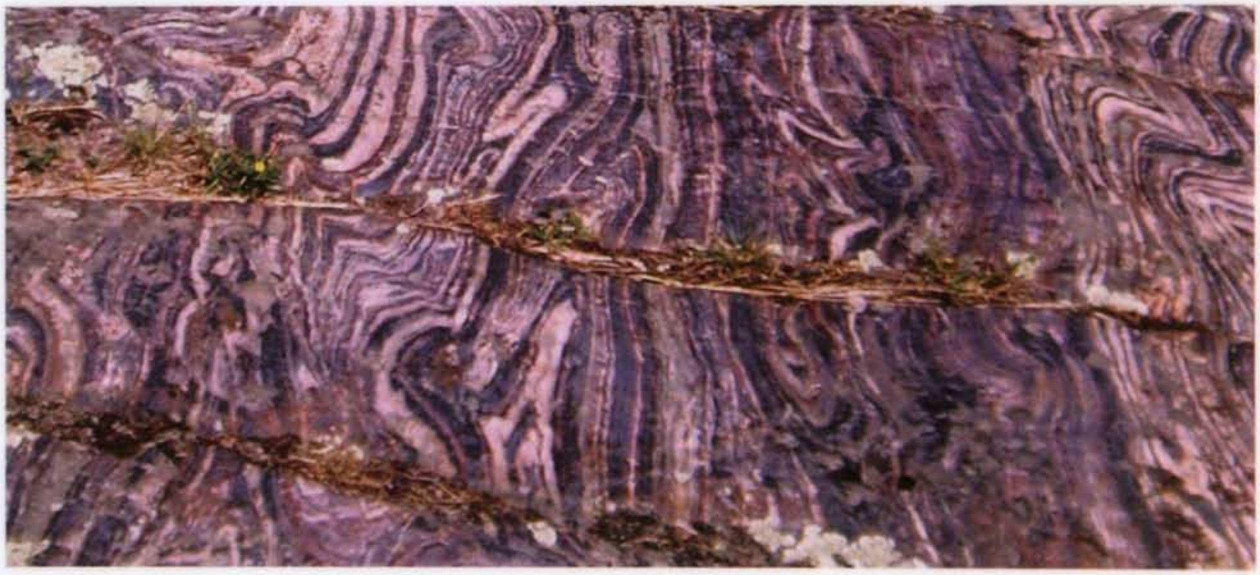
MINNESOTA GEOLOGICAL SURVEY

Harvey Thorleifson, Director

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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The Minnesota Geological Survey has not reviewed all field trip descriptions in this guidebook for quality of scientific content. The views and conclusions contained in each section are those of the author(s). Editorial review was conducted by the Minnesota Geological Survey.

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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.

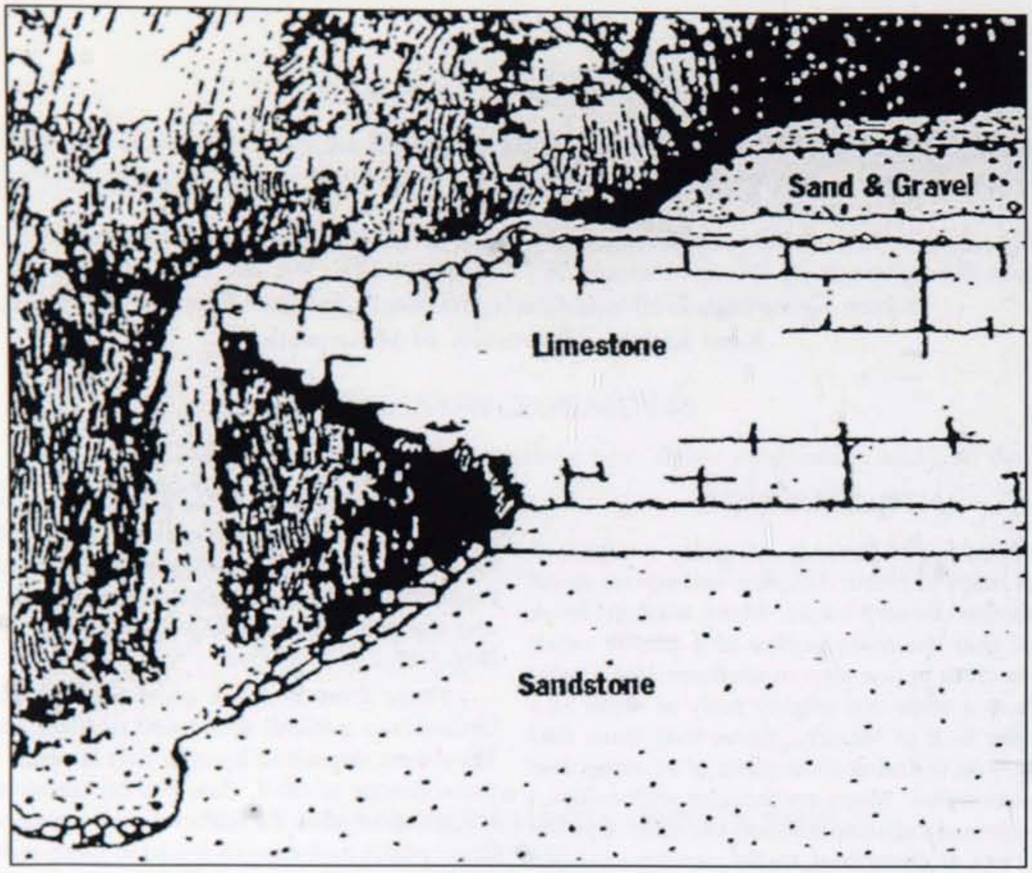


Figure 1.1. Typical development of a waterfall similar to those we will see today (Institute for Minnesota Archaeology, 1999).

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

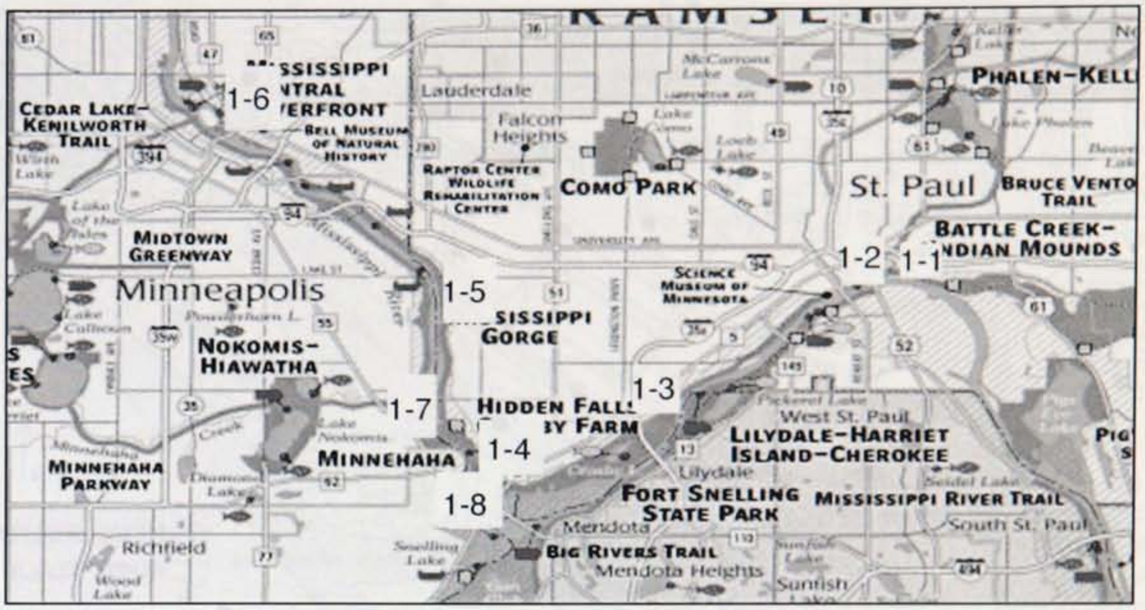


Figure 1.2. Map of field trip stops (Metropolitan Council, 2003).

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

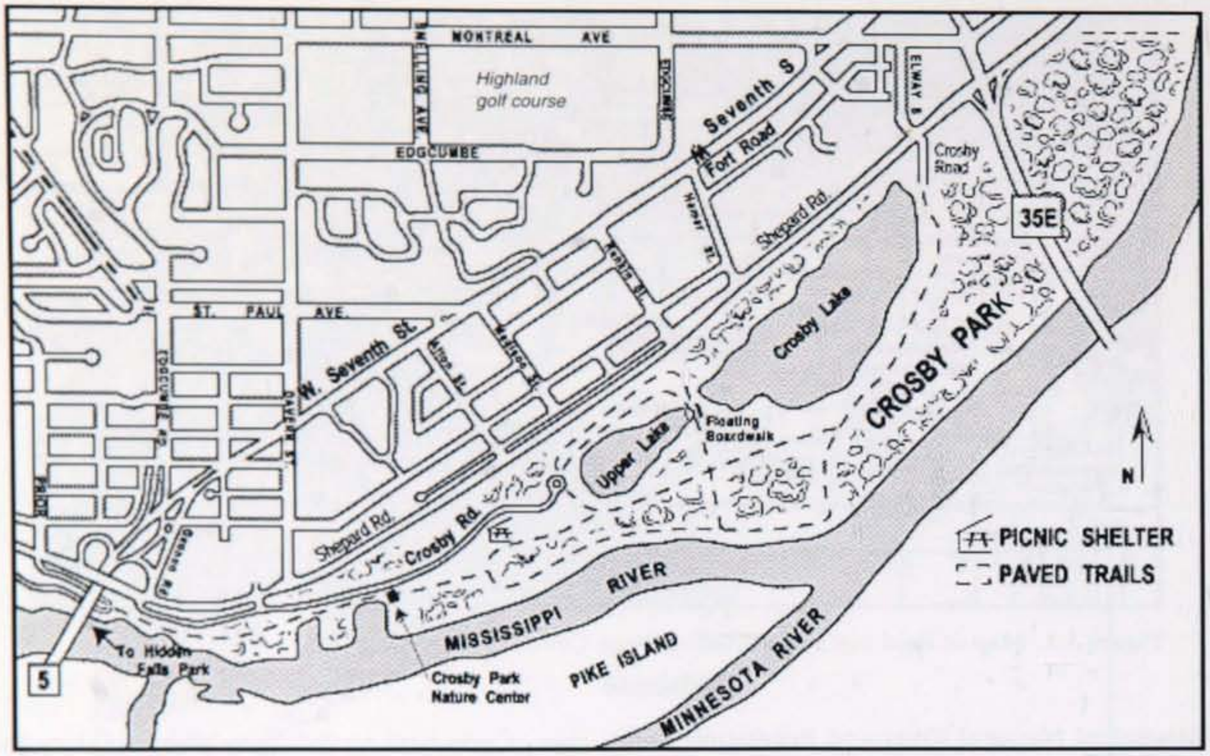


Figure 1.3. Map of Crosby Regional Park (St. Paul Parks and Recreation, 2005).

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

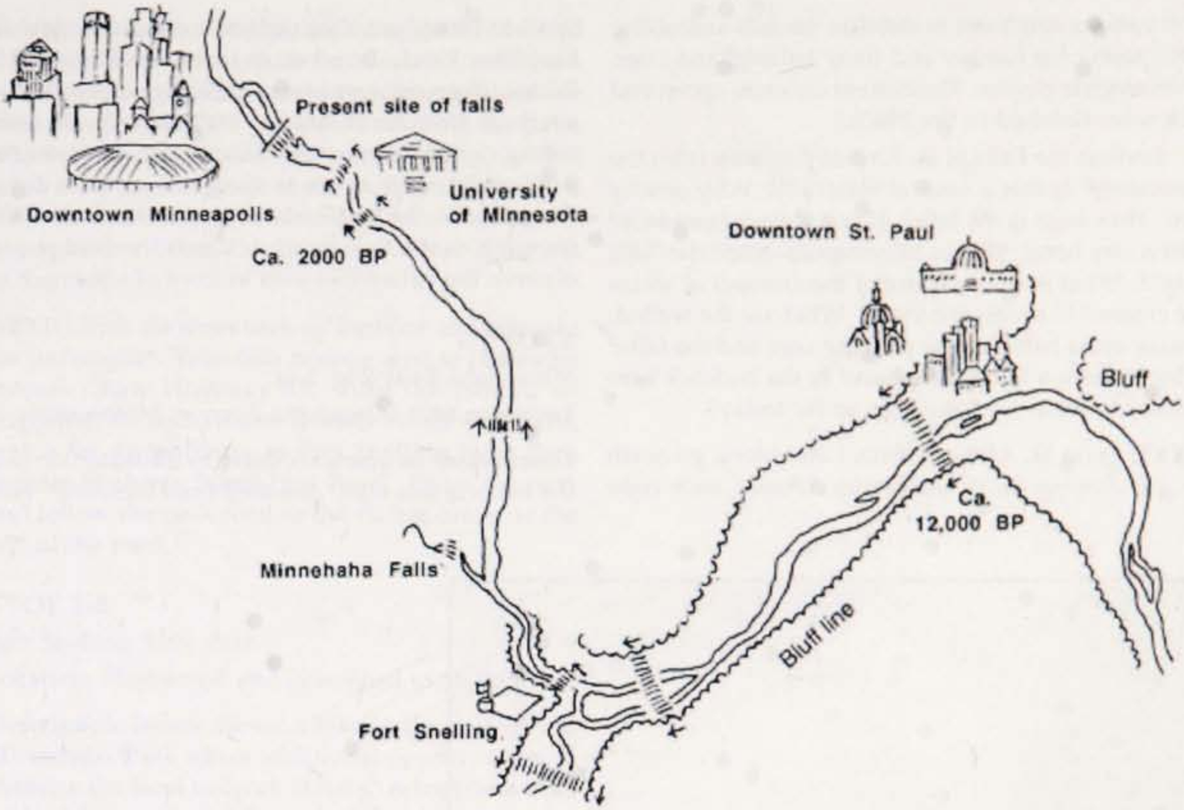


Figure 1.4. Illustration showing the retreat of glacial River Warren (Institute for Minnesota Archaeology, 1999).

Figure 1.5. The Falls of St. Anthony, painted by Henry Lewis as it looked in the 1940s (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

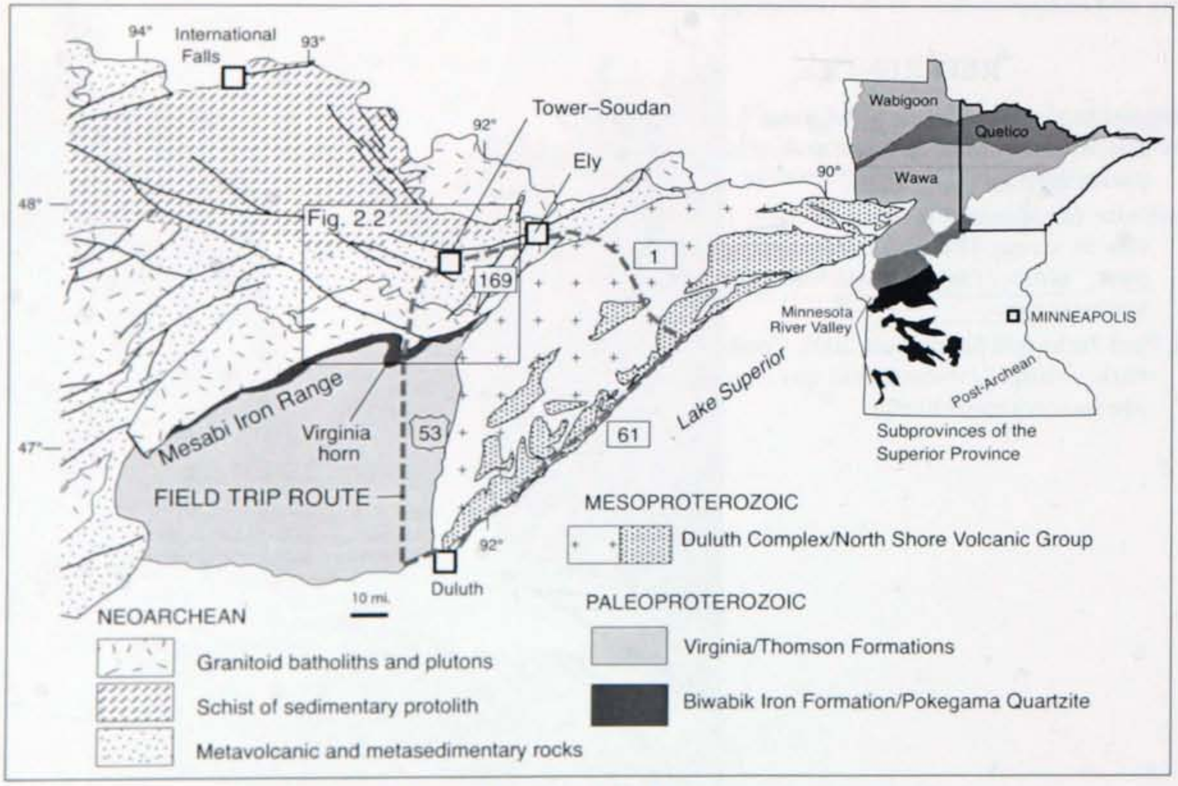


Figure 2.1. Generalized geologic map of northeastern Minnesota showing field trip route.

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		Midcontinent rift	MESOPROTEROZOIC
	– – – – – – – – –			
	North Shore Volcanic Group (11) (12) (14) (15)	Duluth Complex (16) (17)		
		Beaver Bay Complex (13)		
	Puckwunge/Nopeming sandstones	Diabase dikes (19)		
	(18)			1,600 Ma
Animikie Group	Virginia/Thomson/Rove Formations (19)		Penokean orogeny	
	Biwabik/Gunflint Iron Formations (~1,880 Ma) (6)			PALEOPROTEROZOIC
	(5)			
	Pokegama Quartzite	K-K diabase dikes		
	(4)			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)			
	– – – – – – – – – – – – – – – – – – –		D_1 Deformation	
	Newton Lake Formation/Bass Lake sequence			NEOARCHEAN
	Lake Vermilion Formation (8) (9)			
	(1) (2)			
	Mud Lake sequence/Ely Greenstone (~2,720 Ma) (Upper, Soudan iron-formation, and Lower members) (10)			

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

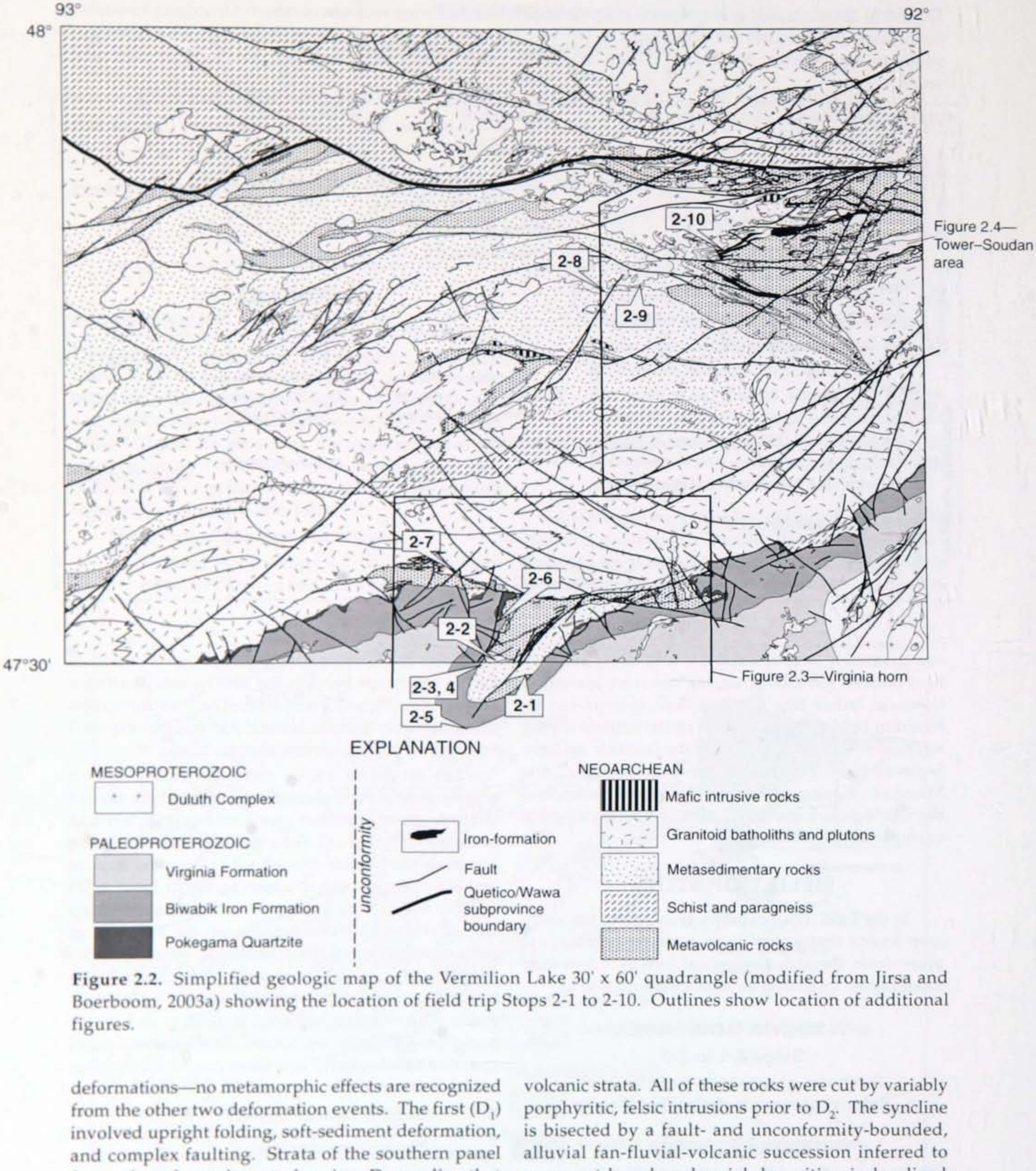


Figure 2.2. Simplified geologic map of the Vermilion Lake 30' x 60' quadrangle (modified from Jirsa and Boerboom, 2003a) showing the location of field trip Stops 2-1 to 2-10. Outlines show location of additional figures.

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

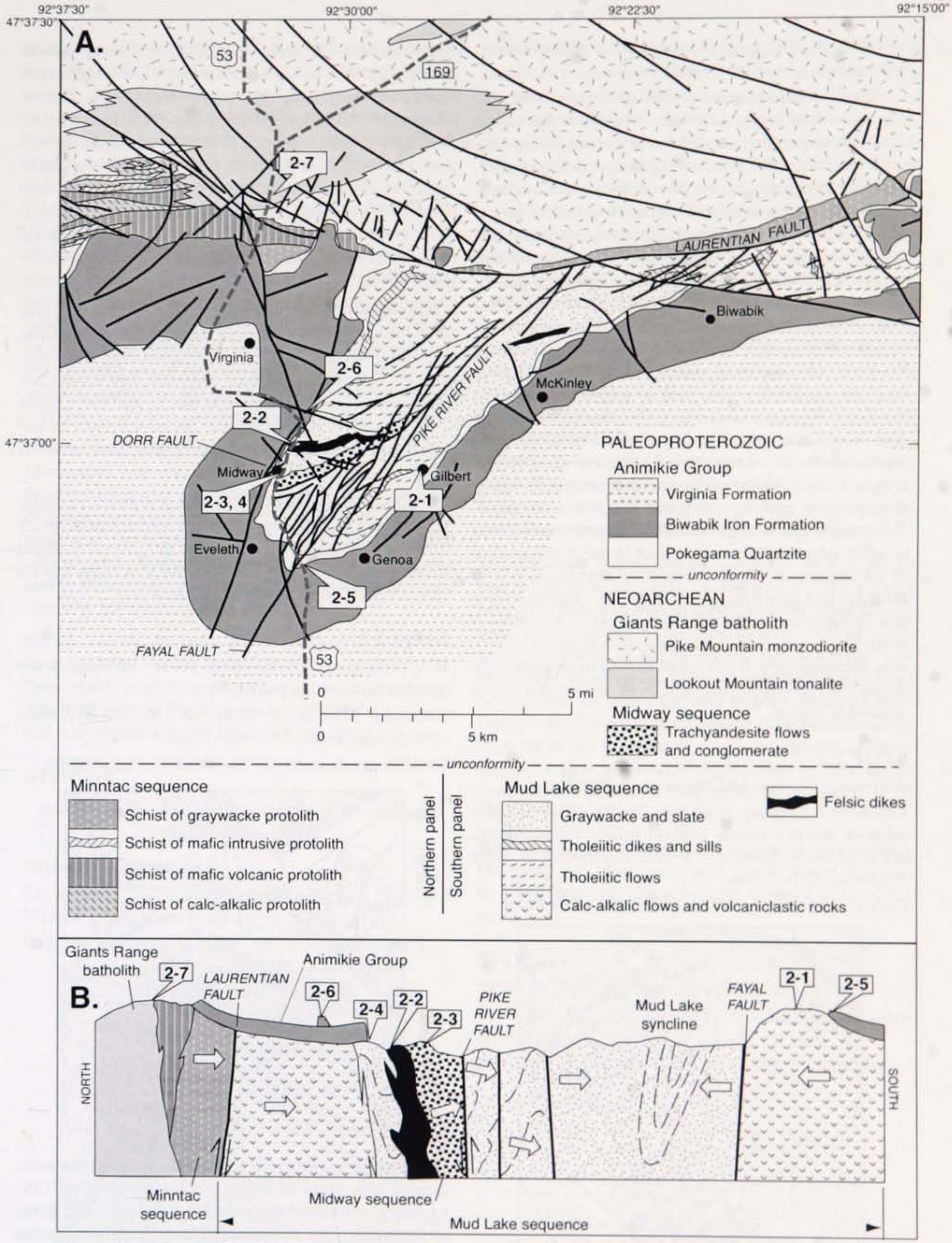


Figure 2.3. Generalized geologic map (A) and schematic cross-section (B) of the Virginia horn area (modified from Jirsa and Boerboom, 2003b) showing details of field trip Stops 2-1 to 2-7. Block arrows indicate general direction of stratigraphic facing.

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

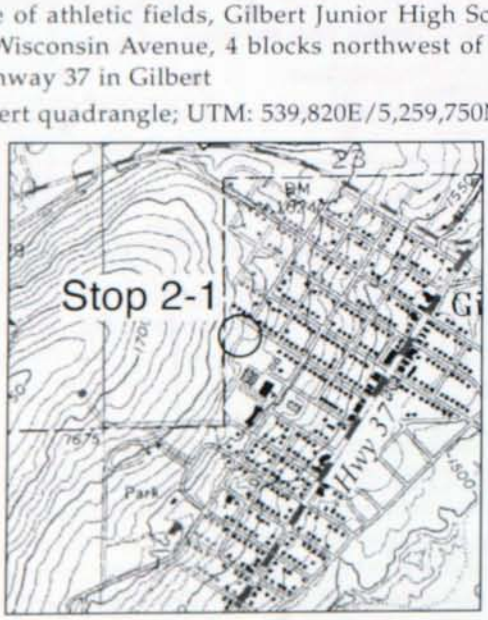
The Paleoproterozoic strata exposed in the Virginia horn are part of the Animikie Group, a tripartite sequence of sedimentary rocks, including basal quartzite and siltstone (Pokegama Quartzite), medial iron-bearing strata (Biwabik Iron Formation), and upper graywacke and shale of turbidite origin (Virginia 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.

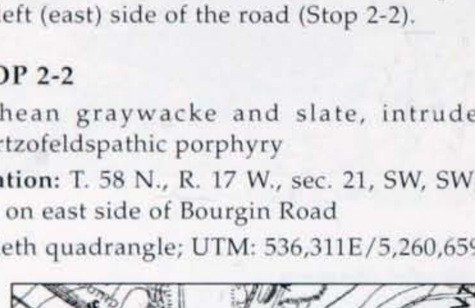
NEXT: Return to Highway 37, travel northeast to State Highway 135. Follow 135 west approximately 2.5 miles to Bourgin Road. Turn left (south) on Bourgin Road and continue about 0.4 mile to a large cut on the left (east) side of the road (Stop 2-2).

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 up-hill 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.

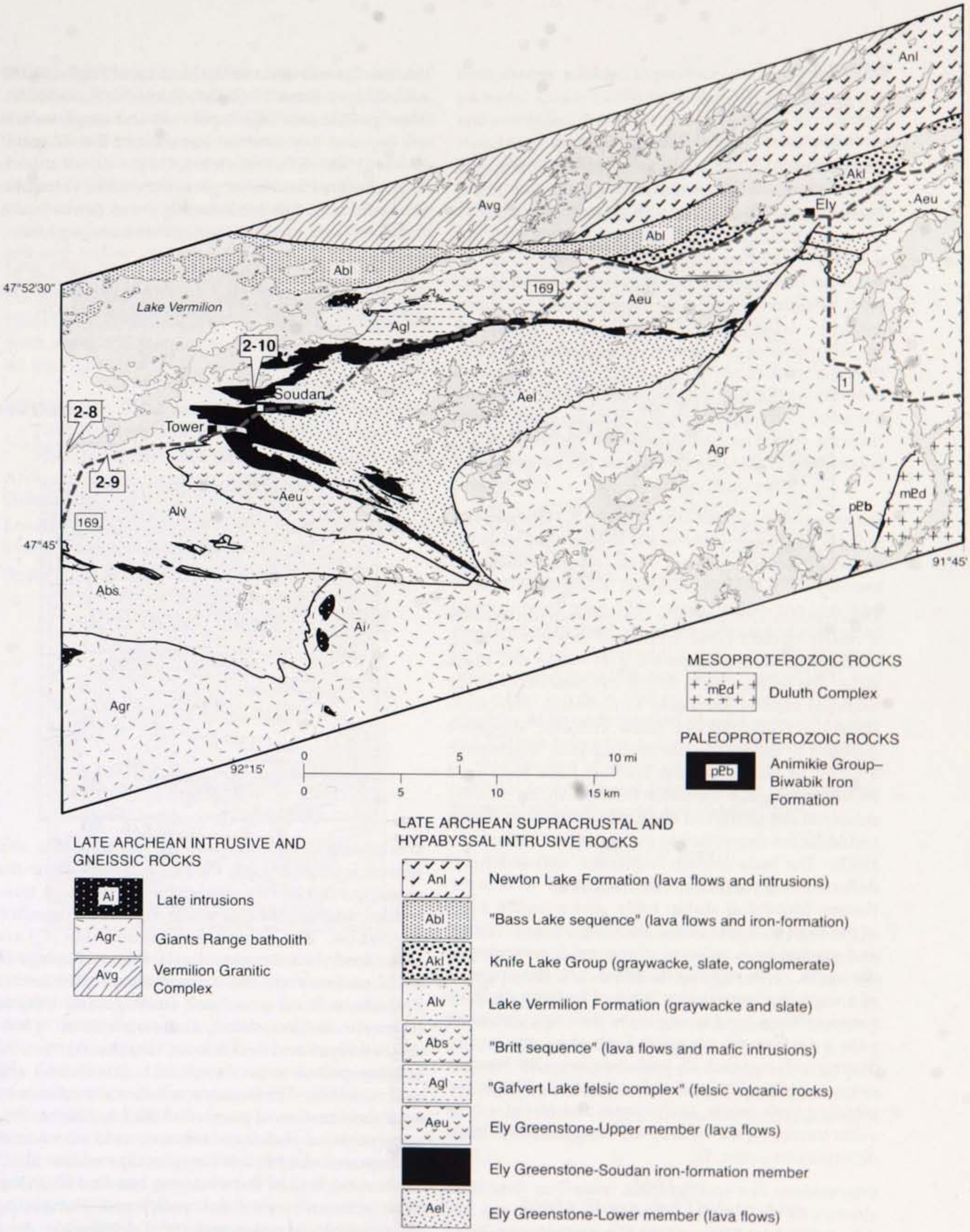


Figure 2.4. Geologic map of the western Vermilion district (modified from Peterson and Jirsa, 1999) including the Tower–Soudan and Ely areas, and showing details of field trip Stops 2-8 to 2-10.

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.

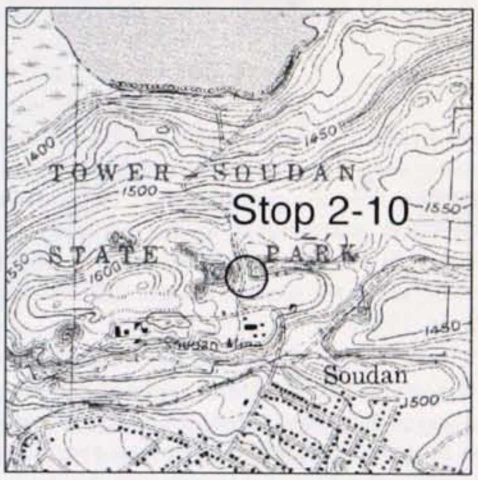
STOP 2-10

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)

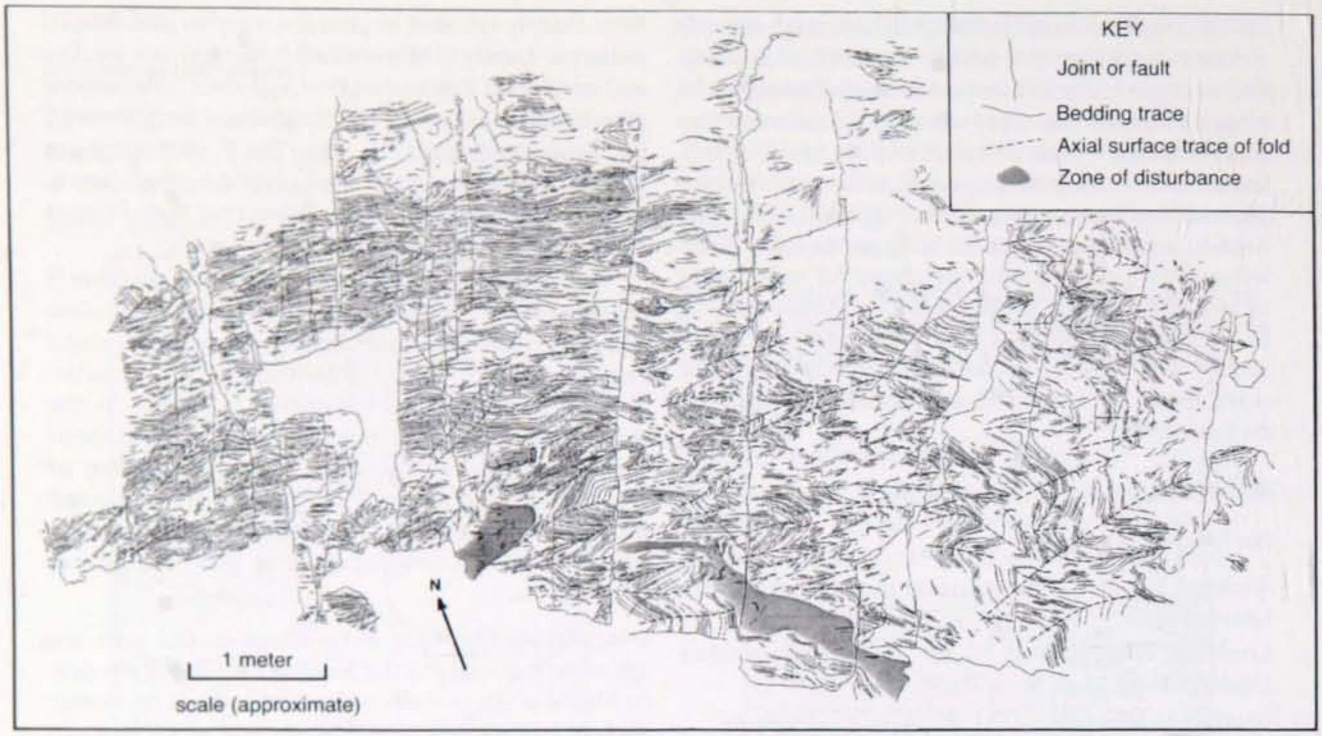


Figure 2.5. Detailed map of Soudan iron-formation outcrop showing bedding trajectories and several generations of folds and faults (from Lundy, 1985).

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.

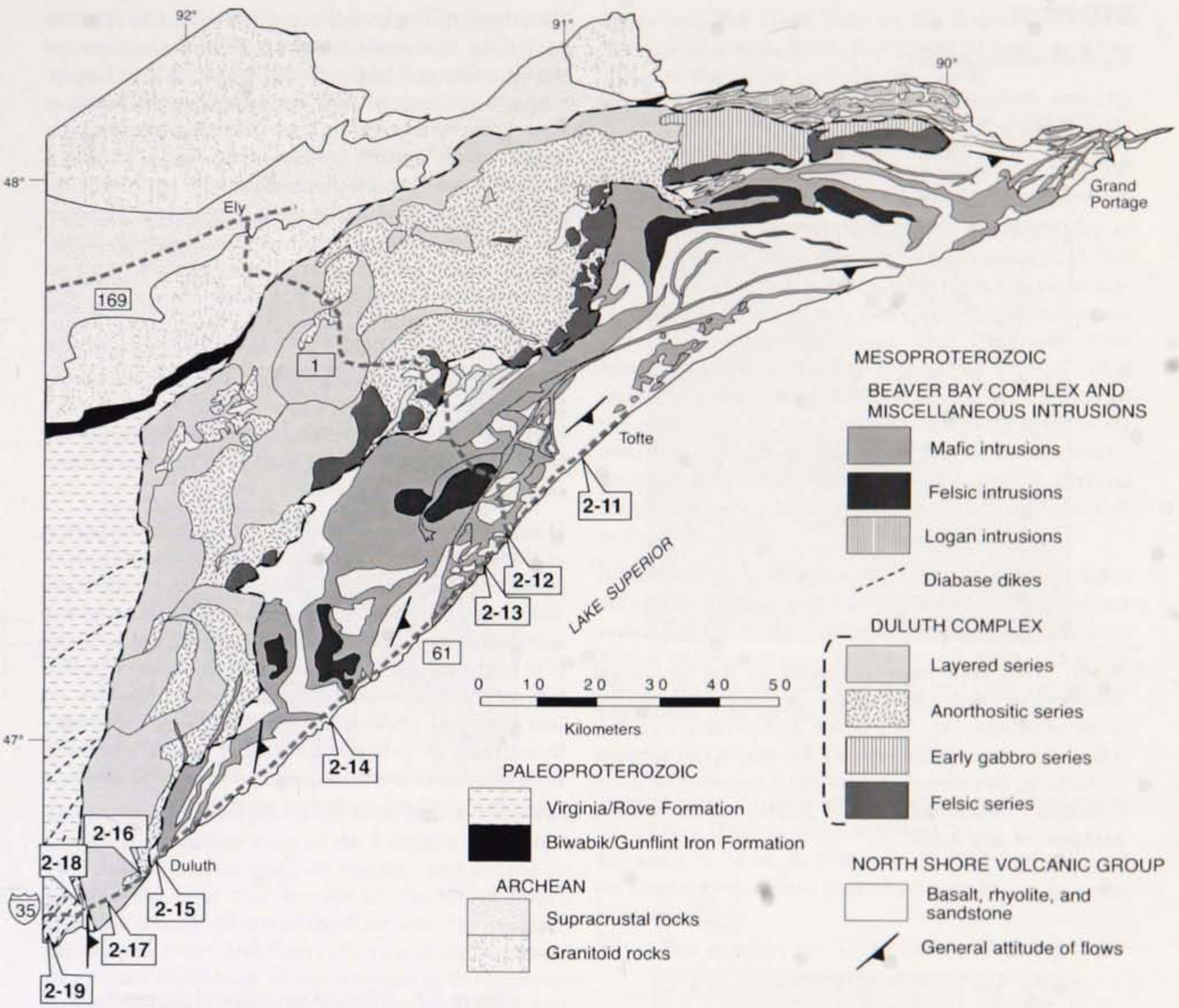


Figure 2.6. Geologic map of northeastern Minnesota (modified from Miller and others, 2002) showing field trip Stops 2-11 to 2-19 on the North Shore and in the Duluth area.

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.

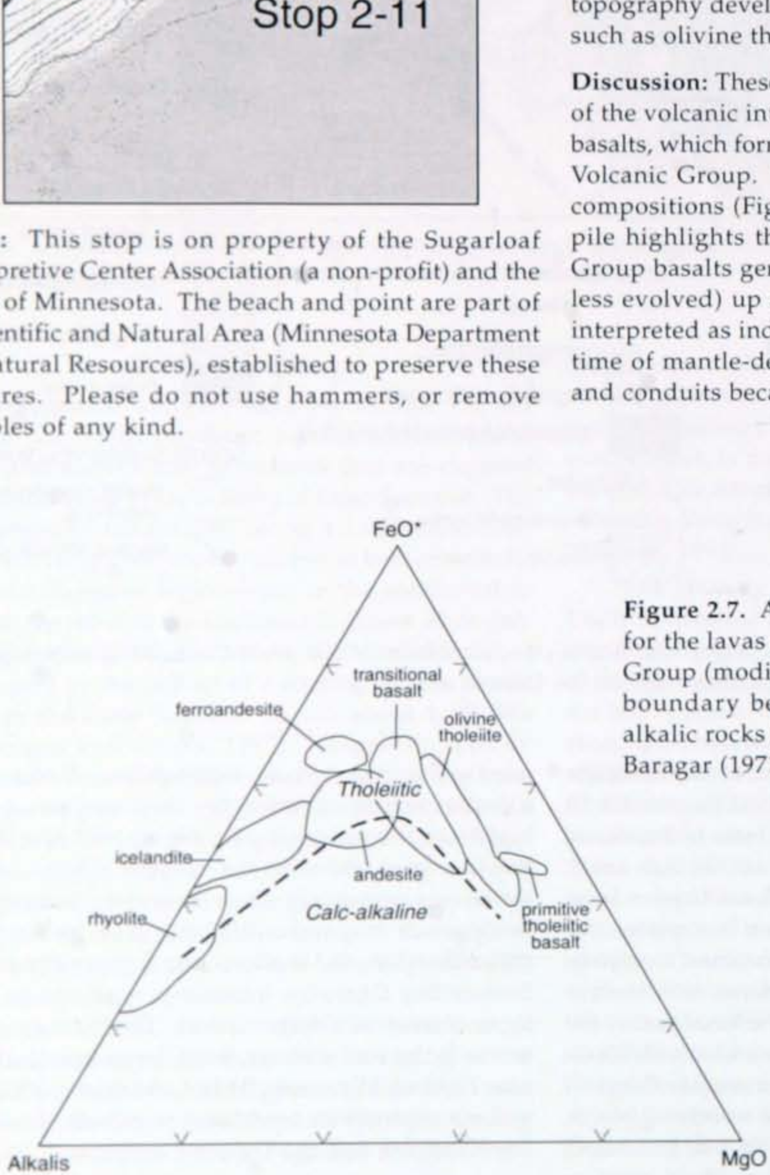


Figure 2.7. AFM compositional diagram for the lavas of the North Shore Volcanic Group (modified from Green, 1982). The boundary between tholeiitic and calc-alkalic rocks is modified from Irvine and Baragar (1971). $FeO^* = FeO + 0.9Fe_2O_3$.

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).

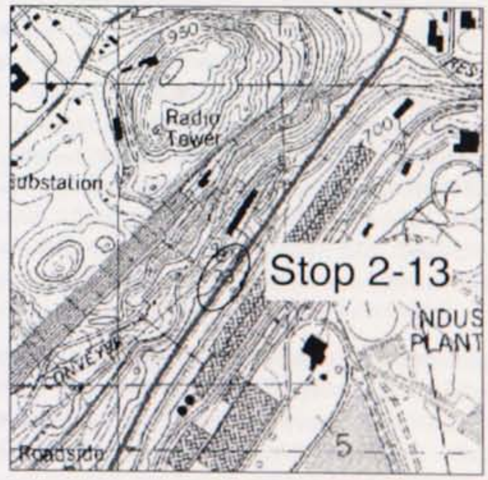
NEXT: Return to Highway 61. Turn southwest (right) on 61 and drive approximately 4.4 miles to a pullout area on the southeast side of the highway adjacent to the gated entrance to the North Shore Taconite Plant. Walk across the highway to a road cut on the northwest side of the road.

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).

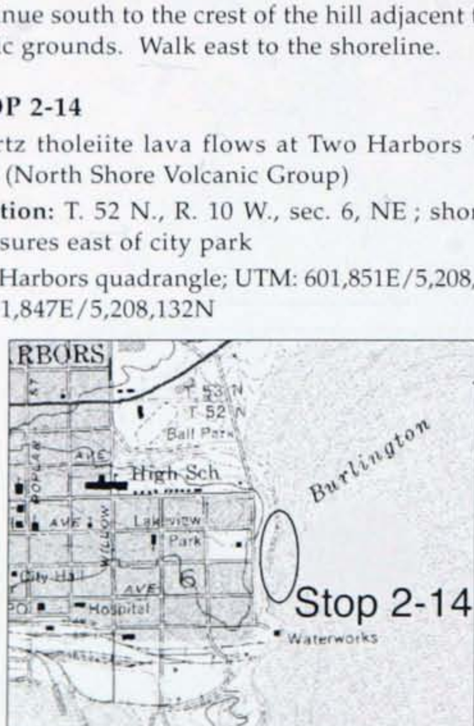
NEXT: Return to Highway 61, turn southwest (left), and continue approximately 29 miles to Two Harbors. At the north end of town, turn left on 1st Street and continue south to the crest of the hill adjacent to the picnic grounds. Walk east to the shoreline.

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

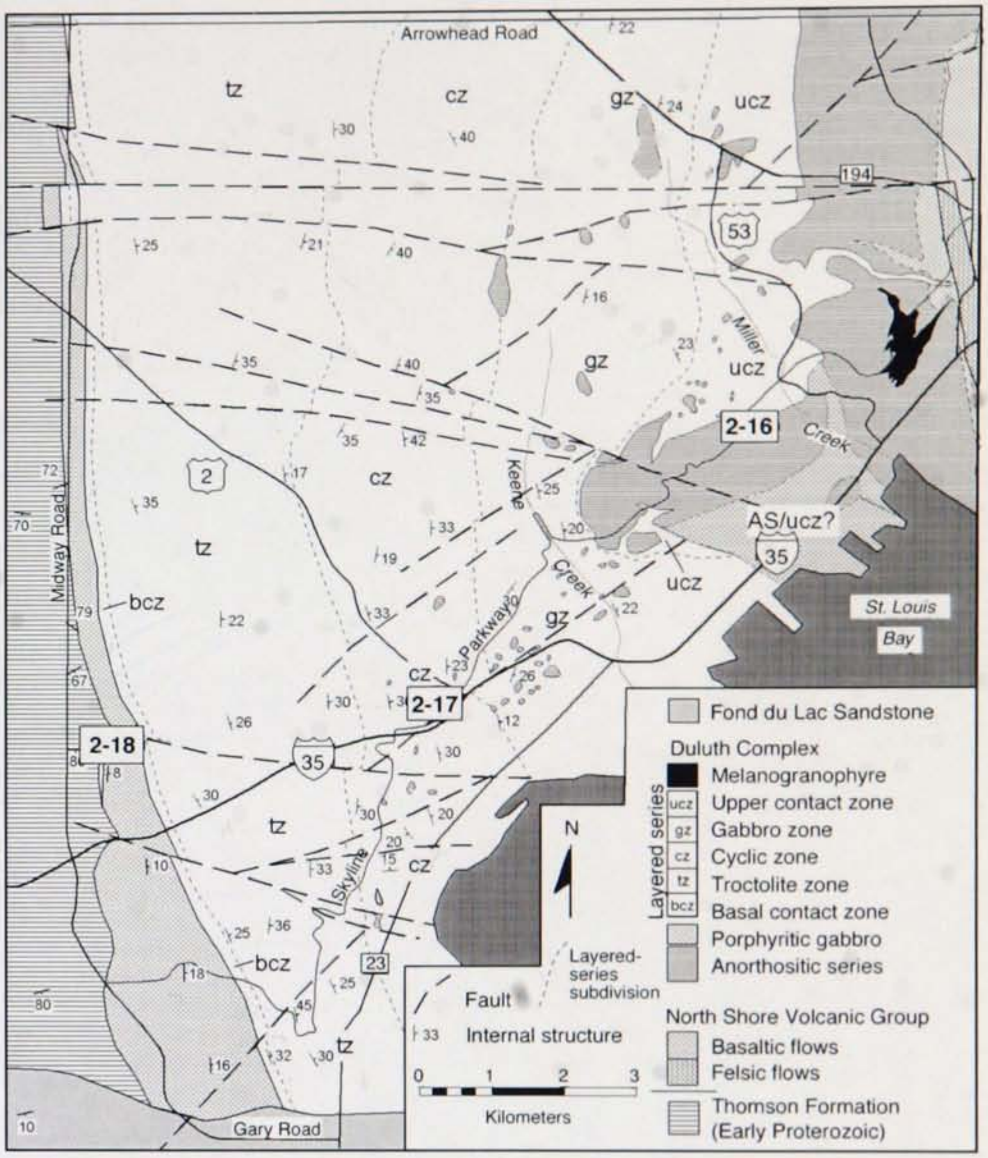
Description: Exposed at the northeast end of this road cut is a fine-grained mafic rock with intermingled granophyre that together cut coarse-grained olivine gabbroic anorthosite. The gabbroic anorthosite observed here and along much of Skyline Parkway above West Duluth is typical of what is termed the 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 ($MgO/(MgO + FeO) = \sim37$; 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.

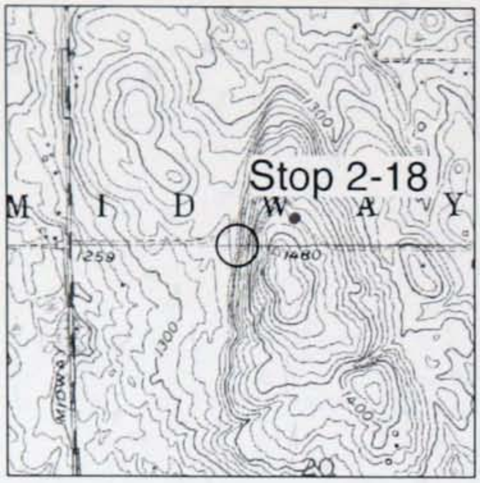
NEXT: Return to Interstate 35 and travel southwest approximately 3 miles to St. Louis County Road 13 (Midway Road). Turn north on Midway Road and continue 0.8 mile to a road to the east (right). Follow this road east approximately 0.2 mile, disembark and walk east to the base of the bluff.

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.

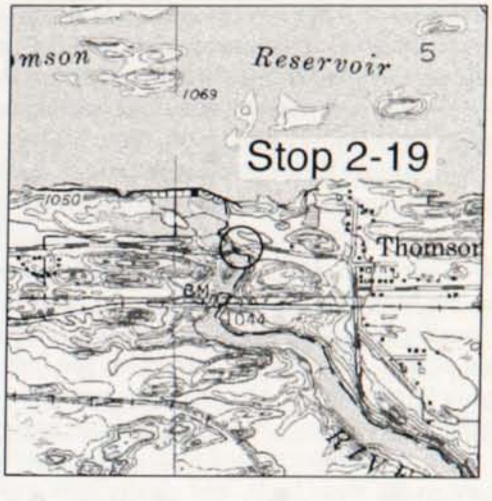
NEXT: Return to Interstate 35 and head south (actually west). Take the Esko/Thomson exit (St. Louis County Road 98/Carlton County Road 1) and head south (left) toward Thomson. At the junction with State Highway 210, turn right and follow it several blocks to a pull out next to the bridge over the St.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

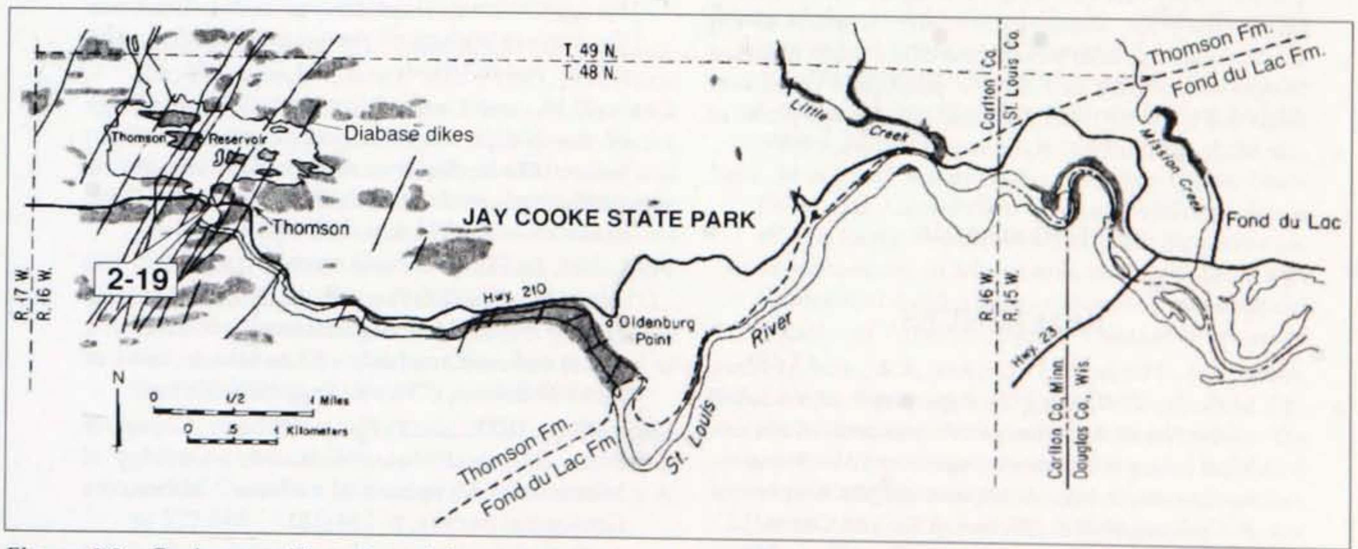


Figure 2.9. Geology and location of the stop in the Fond du Lac, Jay Cooke State Park, and Thomson Dam areas (modified from Jirsa and Morey, 1987).

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.

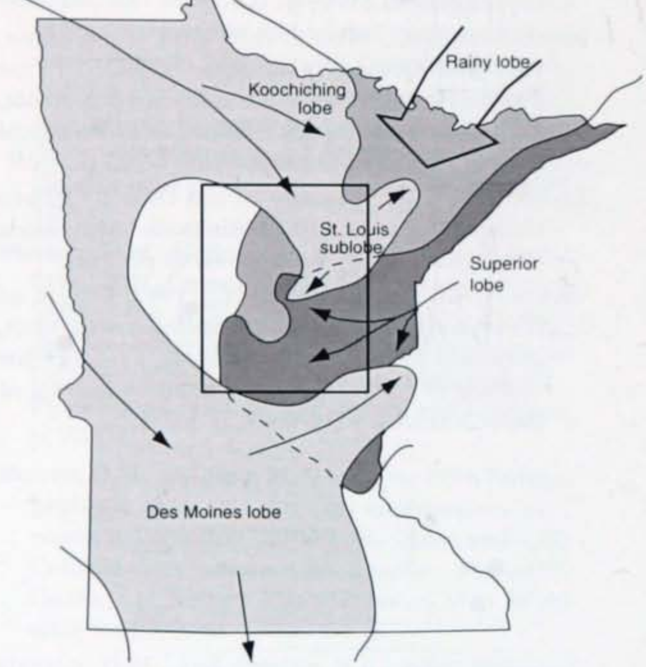


Figure 3.1. Ice-lobe phases with approximate flow direction and ice-marginal extent of each lobe (modified from Knaeble and Meyer, 2004). Box shows the day one field trip area. During the last glaciation, the Late Wisconsinan, the north-central region of Minnesota was covered by ice of the Rainy lobe, 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

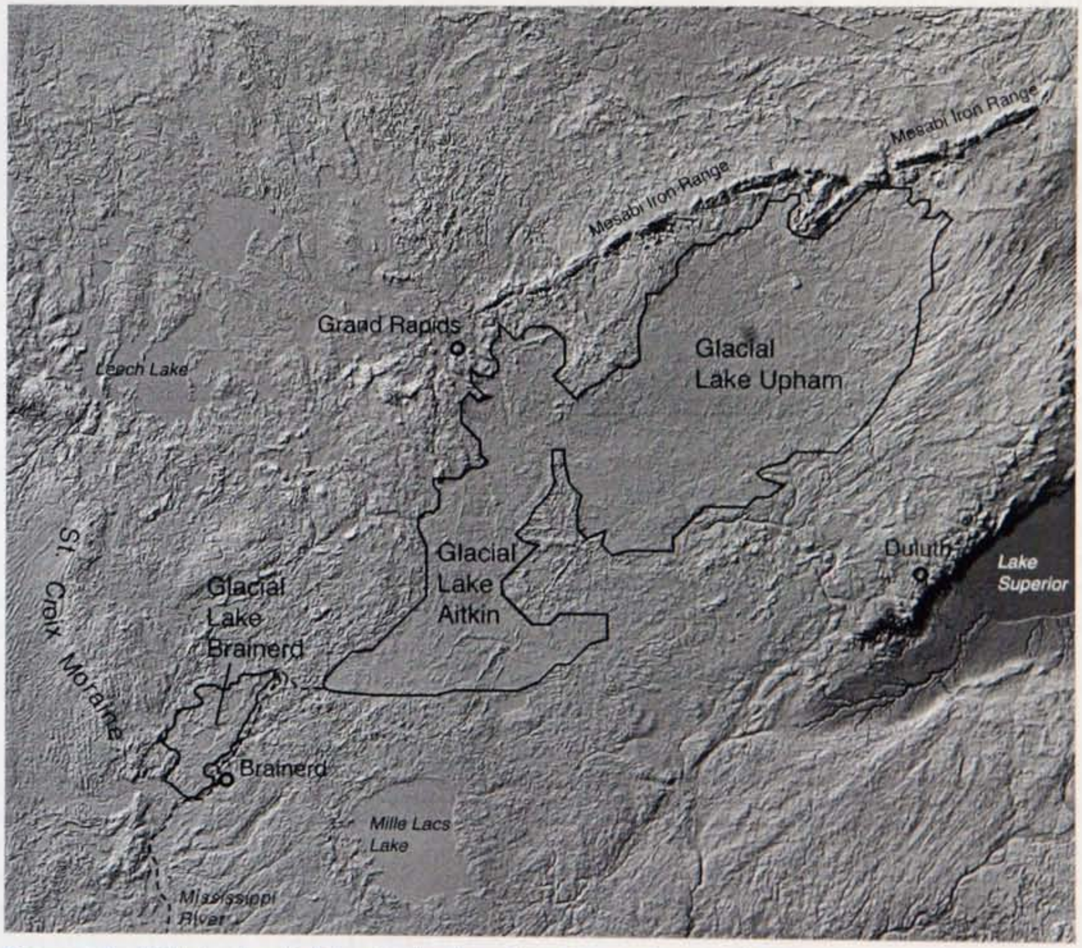


Figure 3.2. Hill-shade relief for the land surface for north-central Minnesota, showing glacial lake basins, geomorphic features, drainage paths (dashed lines), and a few modern-day lakes and cities.

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.

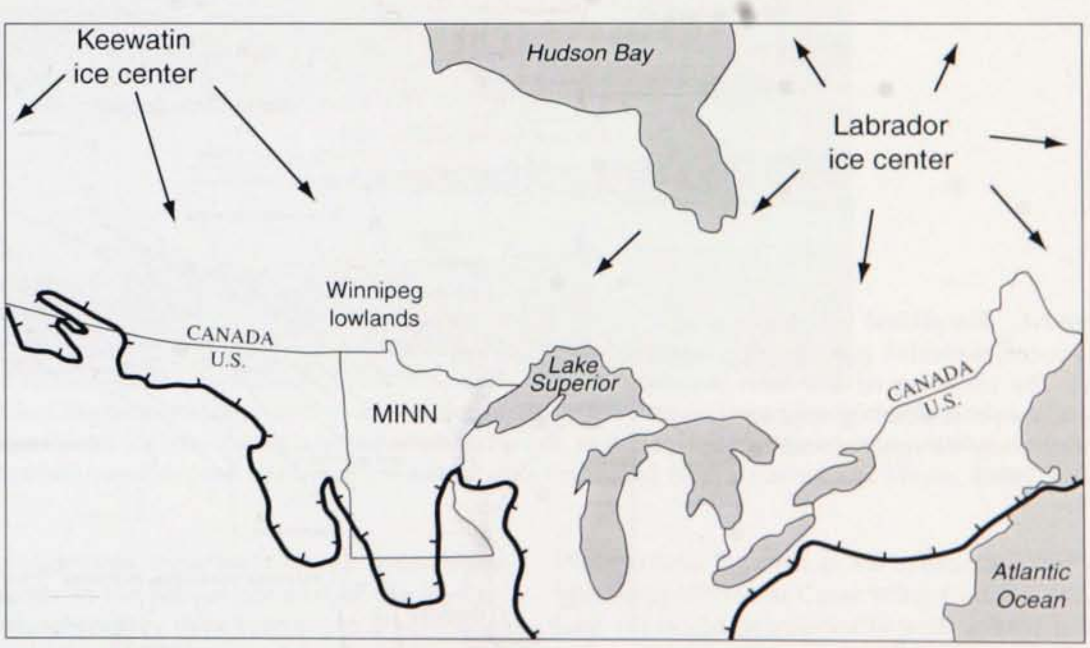


Figure 3.4. Location of Laurentide Ice Sheet accumulation centers. Includes ice lobe flow directions and Late Wisconsinan ice margins (hachured line) at approximately 14,000 years ago (from Knaeble and others, 2004; modified from Lusardi, 1998).

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

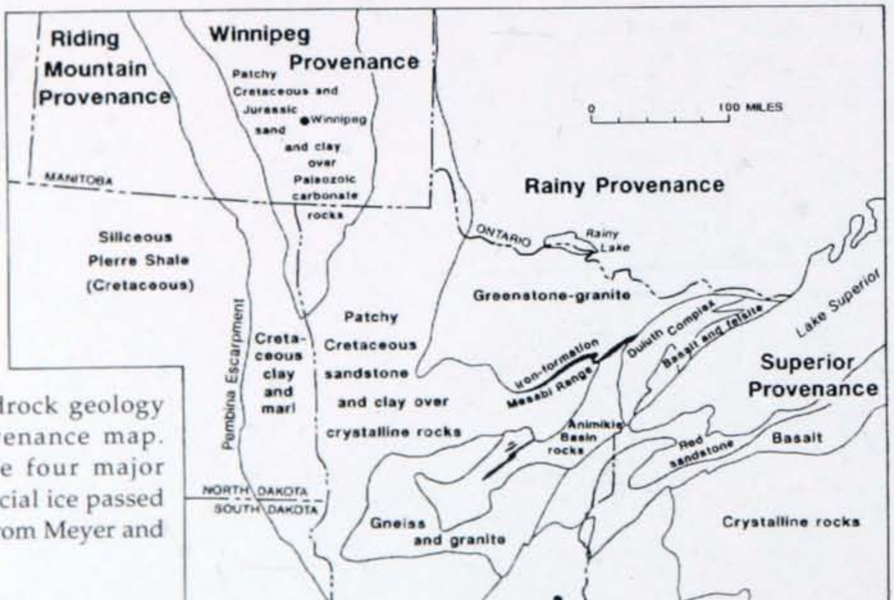


Figure 3.5. Simplified bedrock geology and source-material provenance map. Shows the location of the four major provenances over which glacial ice passed before entering Minnesota (from Meyer and Knaeble, 1996).

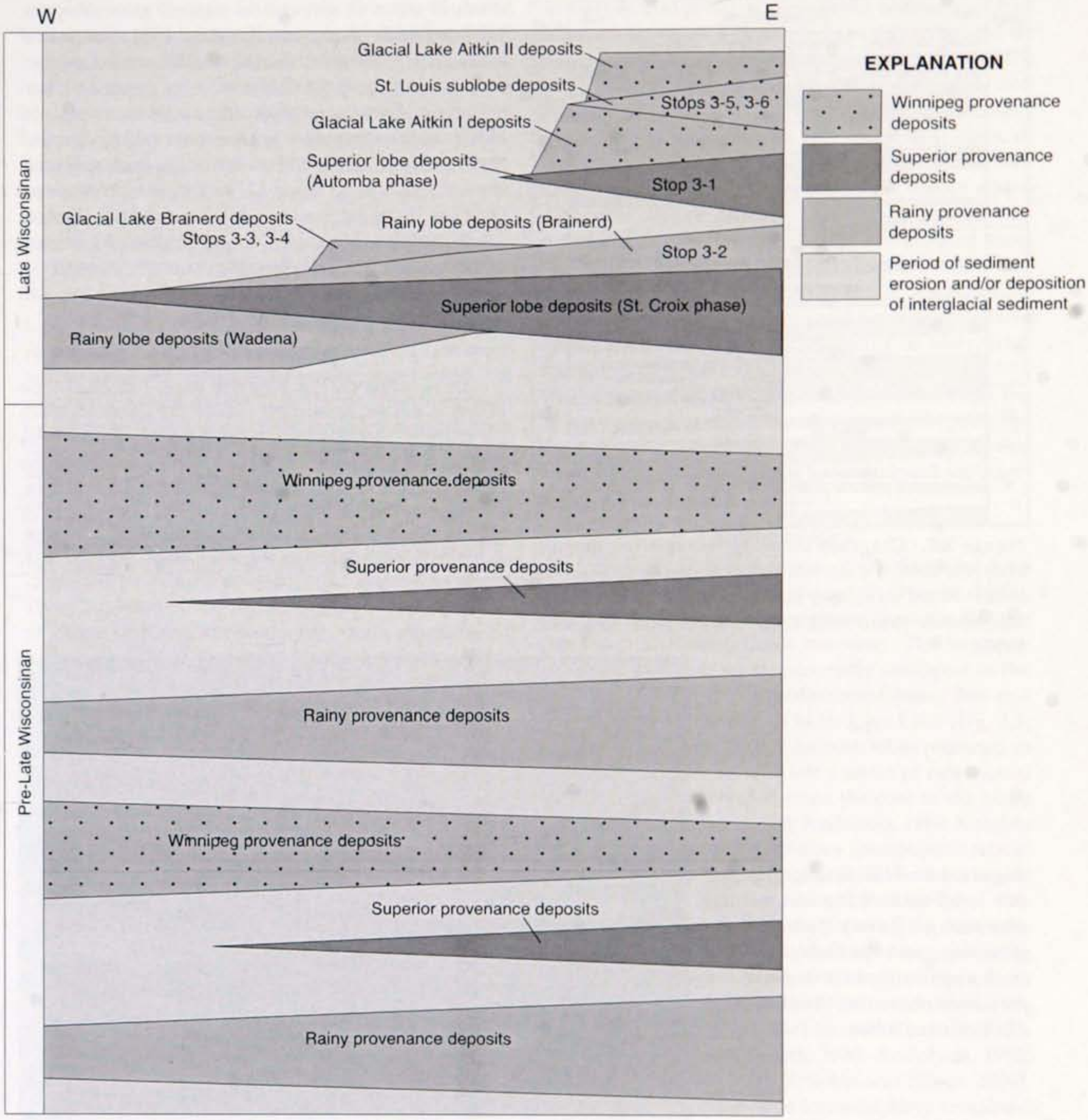


Figure 3.6. Diagram showing relative age and deposit locations (across the region of the field trip) related to the Late Wisconsinan ice lobes, their associated glacial lakes, and the pre-Late Wisconsinan ice lobe provenances. The age column and deposit drawings are not to scale (modified from Knaeble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

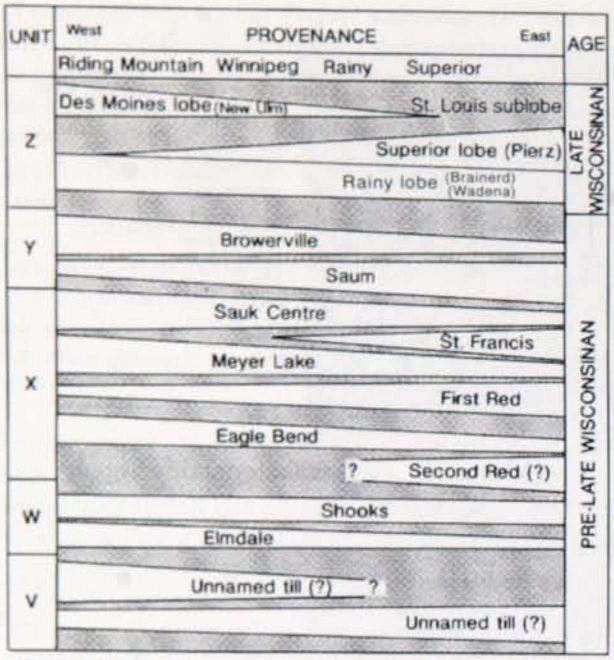


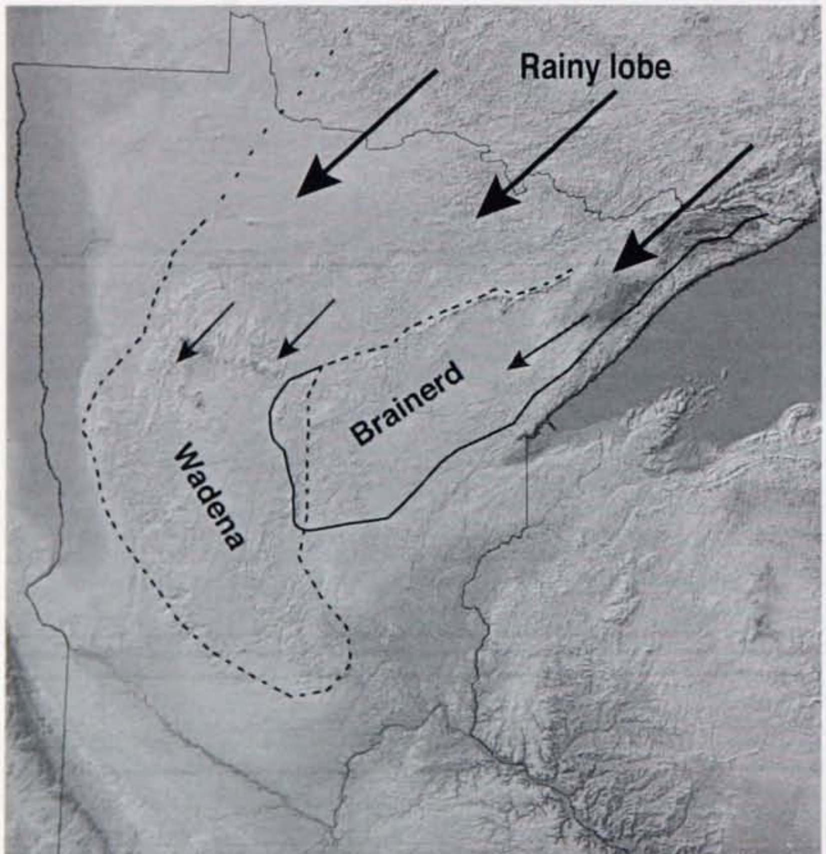
Figure 3.7. Diagram showing the relative timing, bedrock source-material provenance, and relative extent of ice advances that deposited till in central Minnesota (modified from Meyer and Knaeble, 1996).

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

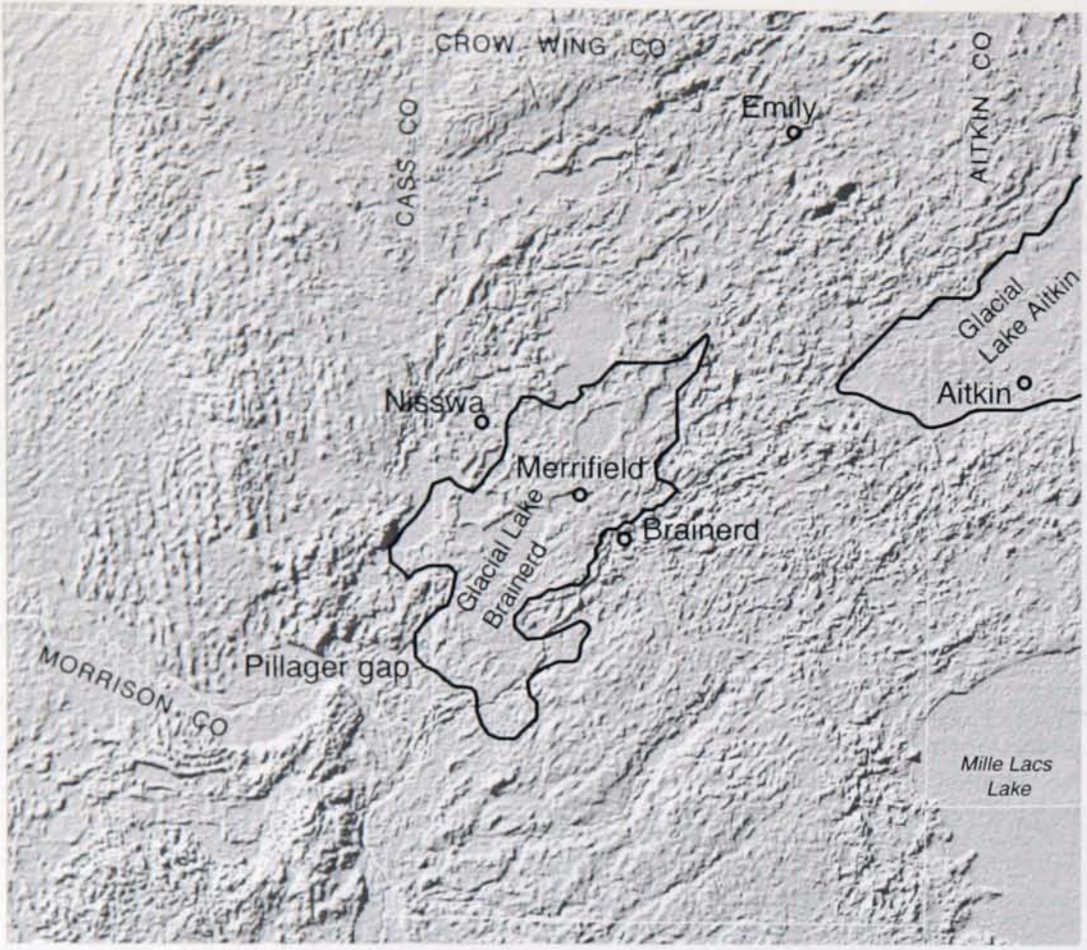


Figure 3.9. Hill-shade relief for the land surface for north-central Minnesota, showing glacial Lakes Aitkin and Brainerd, the Pillager gap, and modern-day counties, cities, and lakes.

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

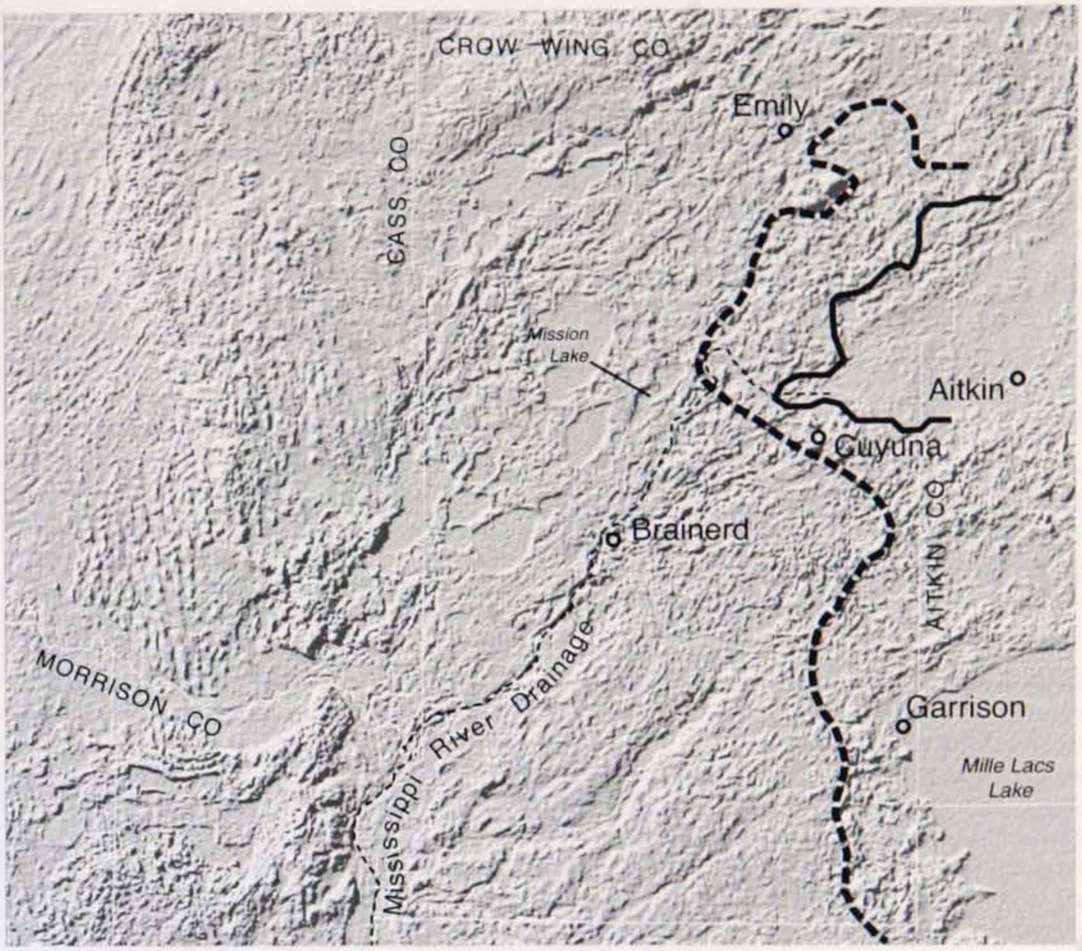


Figure 3.10. Hill-shade relief for the land surface for north-central Minnesota, showing the approximate ice margin of the Automba phase of the Superior lobe (heavy dashed line) and St. Louis sublobe (heavy black line), Mississippi River drainage (dashed line), and modern-day counties, cities, and lakes.

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.

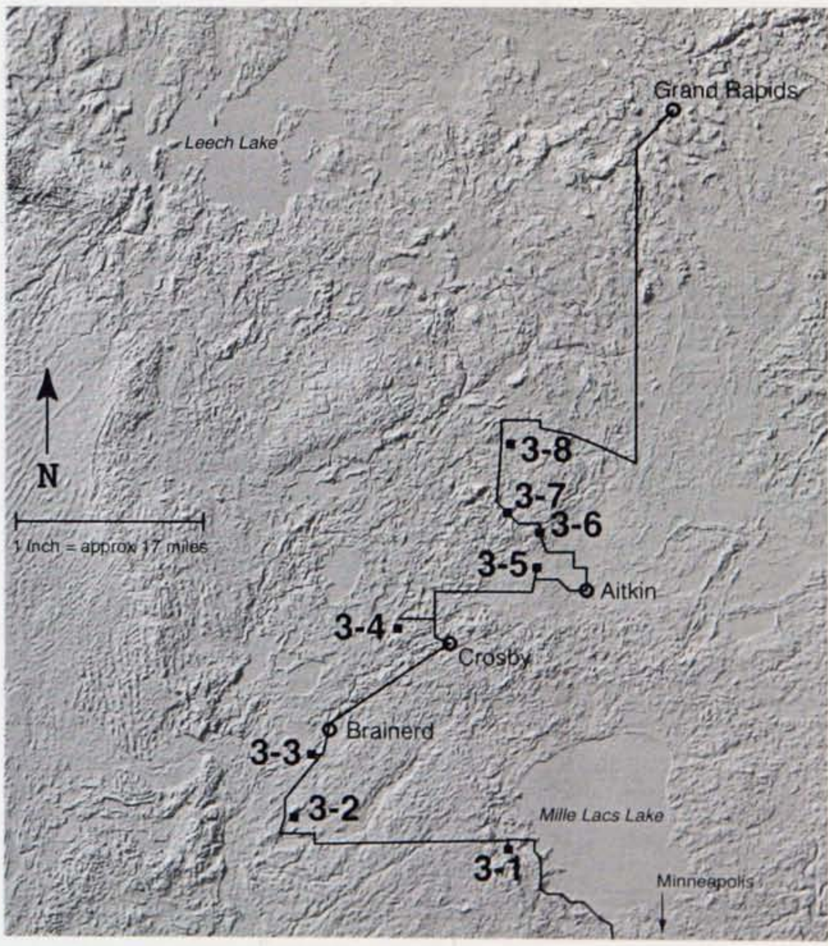


Figure 3.11. Map of the field trip route.

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.

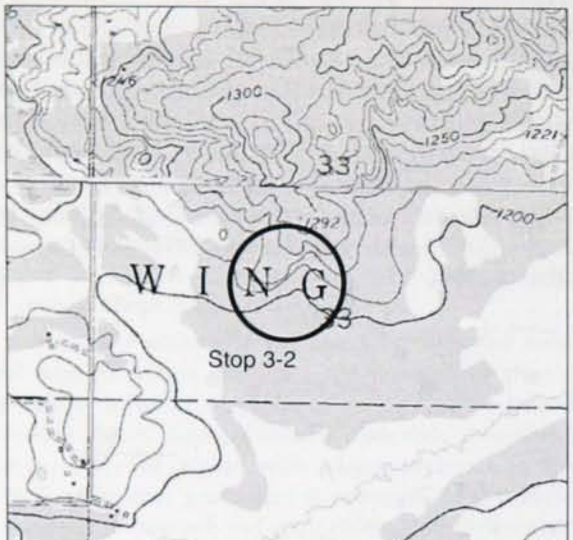
NEXT: North ~0.5 mile on township road to County Road 2. West ~17 miles to Crow Wing County Road 45. North 1.5 miles to Crow Wing County Road 131. West ~4 miles to township road. North ~1.5 miles to a pit road entrance on the east side of the road.

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.

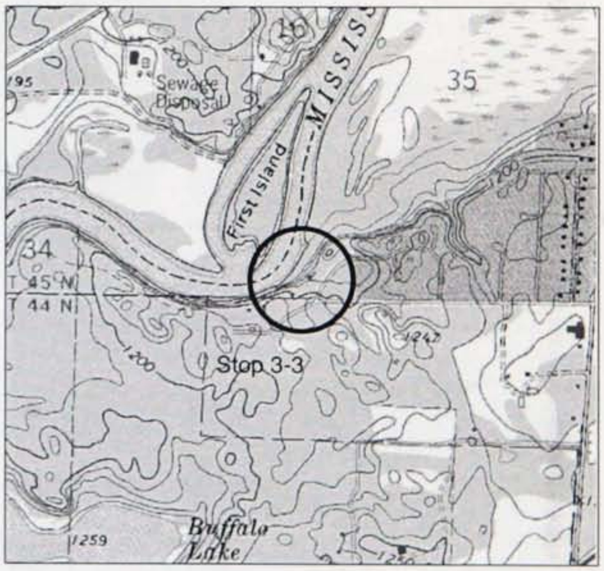
NEXT: North ~3 miles to State Highway 371. Northeast ~4 miles to Buffalo Hills Lane. West ~1 mile to dunes on the south side of the road.

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

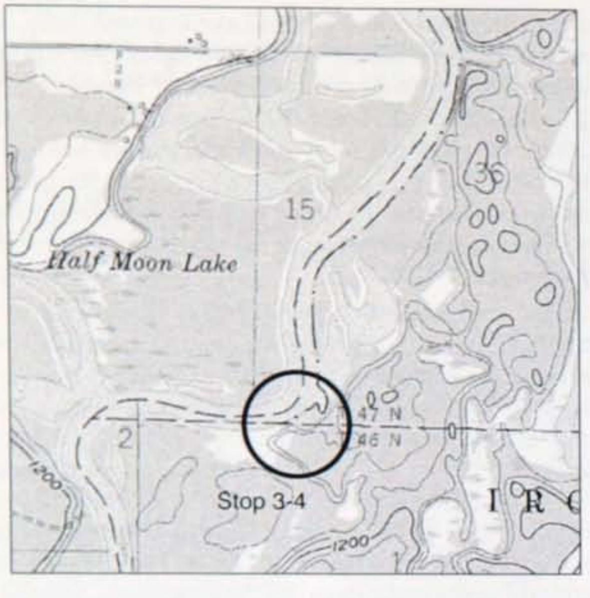
NEXT: West ~1 mile to Crow Wing County Road 30. North and west ~3 miles to Crow Wing County Road 34. South and west ~1.5 miles to Trommald. West on township road located at the northwest corner of Trommald for ~2.5 miles to an ATV access road on the south side of the township road. Hike to the river.

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?

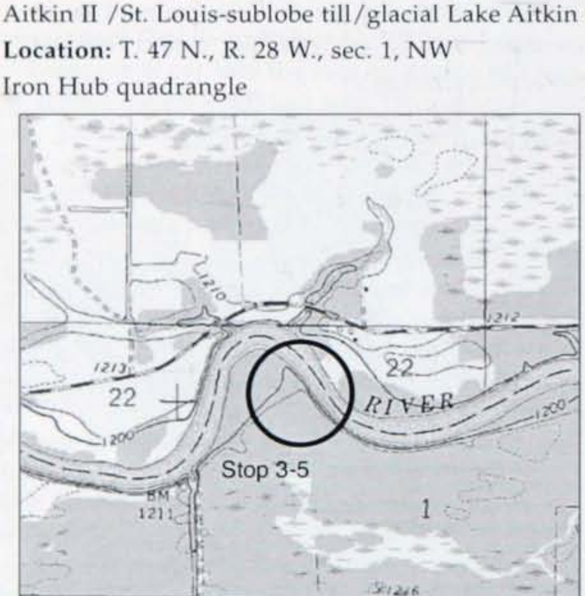
NEXT: East on township road for ~2.5 miles to County Road 34. East and north ~1.5 miles to County Road 30. North and east ~10.5 miles to Crow Wing County Road 32. North for 2 miles to township road. Continue north for ~1.5 miles on the township road and a path along the ditch. Walk to the river.

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.

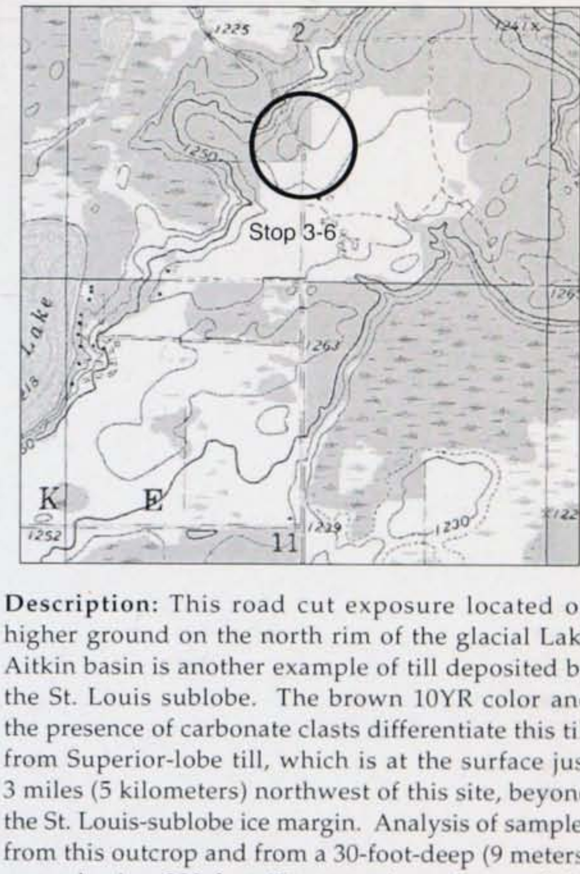
NEXT: South ~1.5 miles to County Road 32. East ~6.5 miles (County Road 32 changes to Aitkin County Road 15) to Aitkin County Road 22 in Aitkin. North and west ~9 miles (County Road 22 changes to Crow Wing County Road 11) to Crow Wing County Road 105. North ~2 miles to an exposure on the east side at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

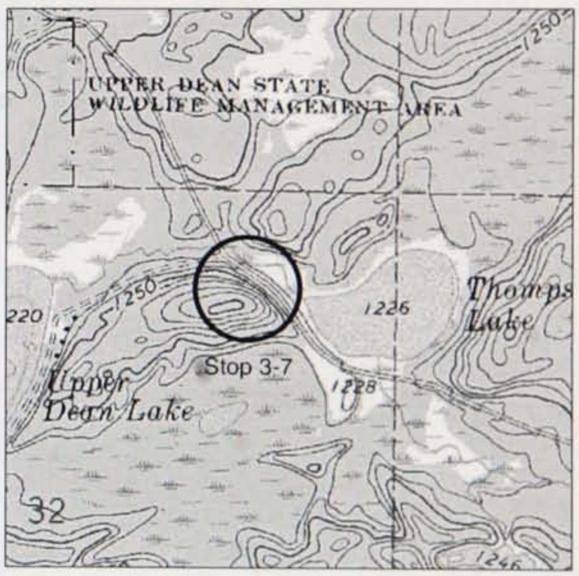
NEXT: North and west on County Road 105 for ~3.5 miles to an exposure on the south side of the road.

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.

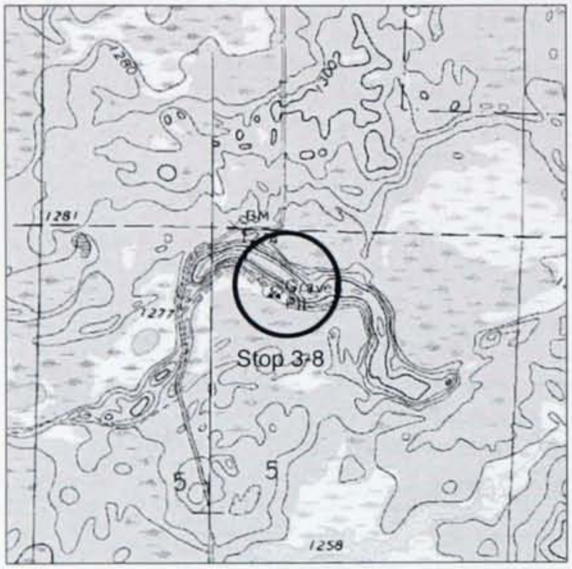
NEXT: Continue northwest ~1.5 miles on County Road 105 to Crow Wing County Road 36. North ~1 mile to Crow Wing County Road 106. North ~3.5 miles to the pit entrance on the east side of the road.

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

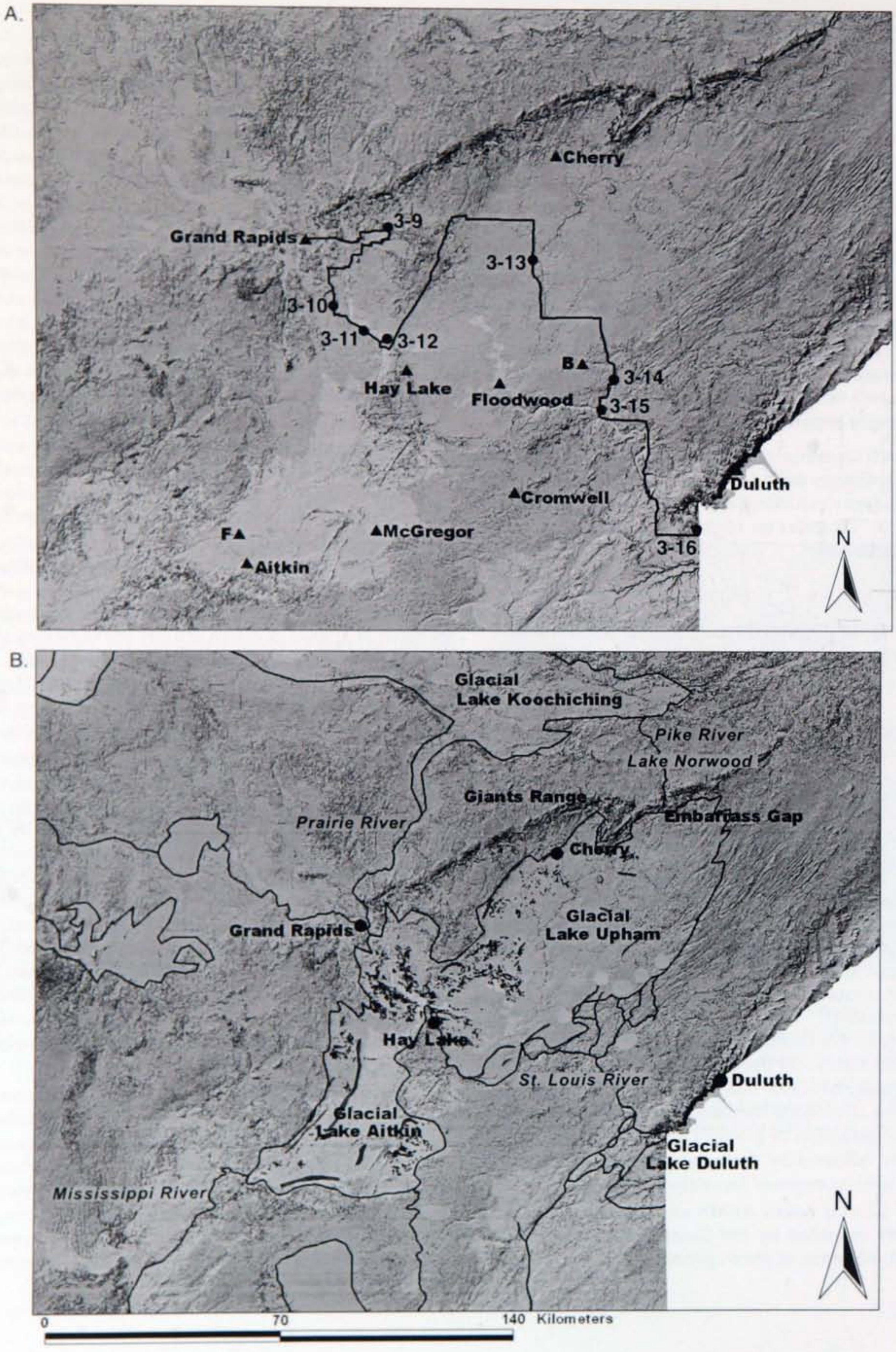
Many deposits and landforms related to the advance of the St. Louis sublobe are found in and around the glacial Lakes Aitkin and Upham basin. Recent investigations (Marlow, 2004; Marlow and others, 2004) into the glacial geology of northeastern Minnesota have considerably revised the glacial chronology. 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
Prairie River
Embarrass Gap
Cherry
Glacial
Lake Upham
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Hay Lake
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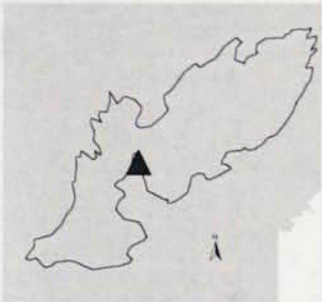

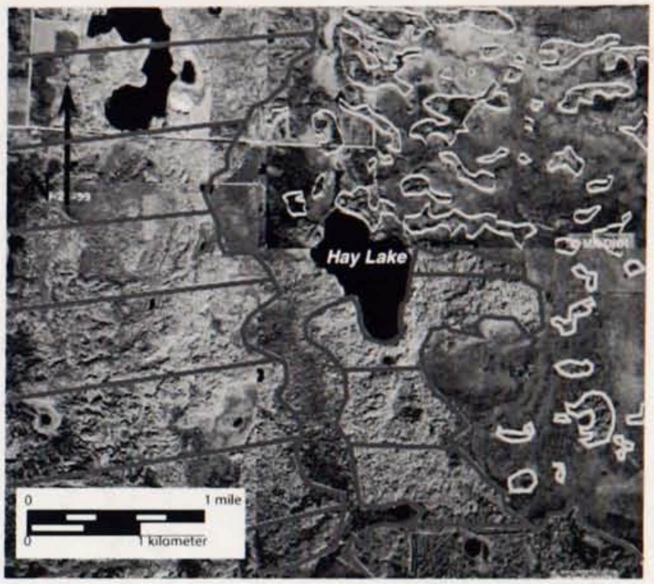

Figure 3.13. Location of Hay Lake. The lined region indicates the Lake Upham upland area; white polygons are dunes (Vanduse quadrangle).

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).

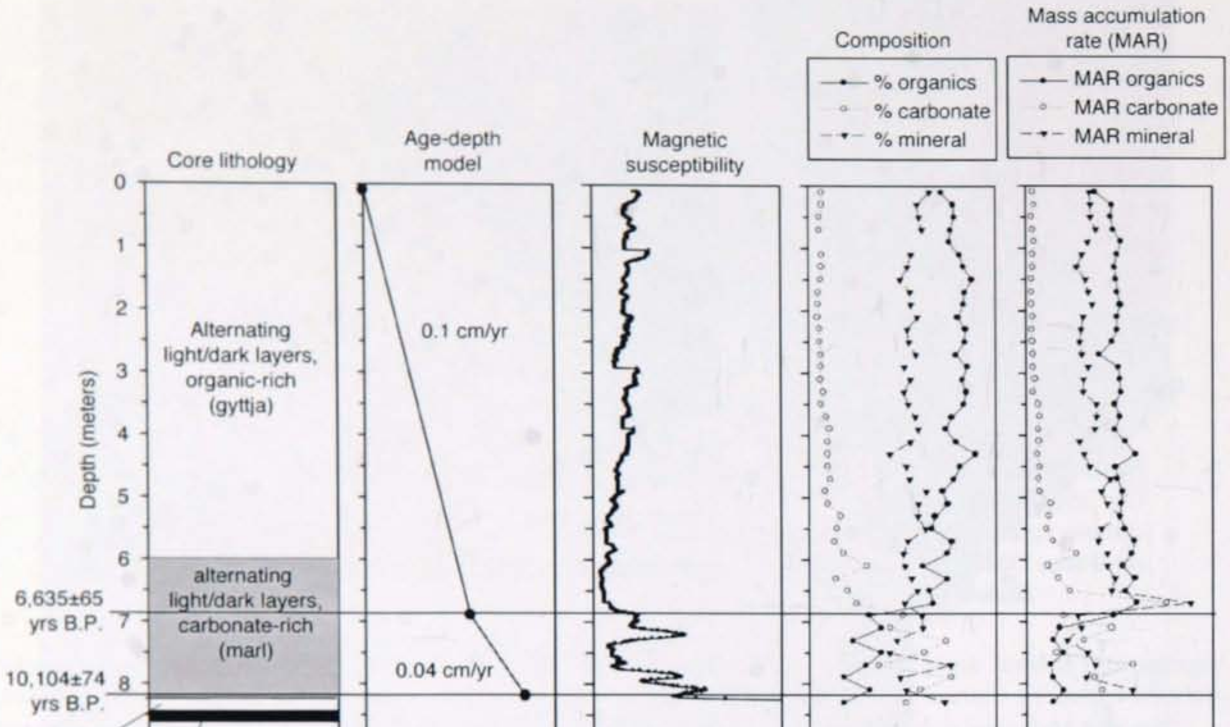

Figure 3.14. Hay Lake sediment core (8.57 meters).

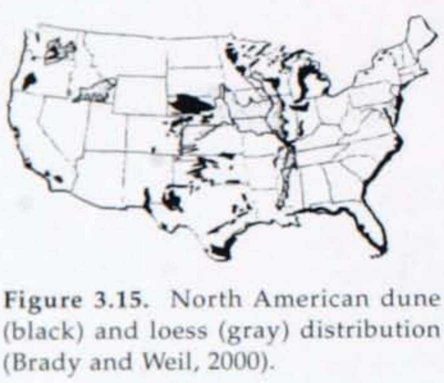

Figure 3.15. North American dune (black) and loess (gray) distribution (Brady and Weil, 2000).

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

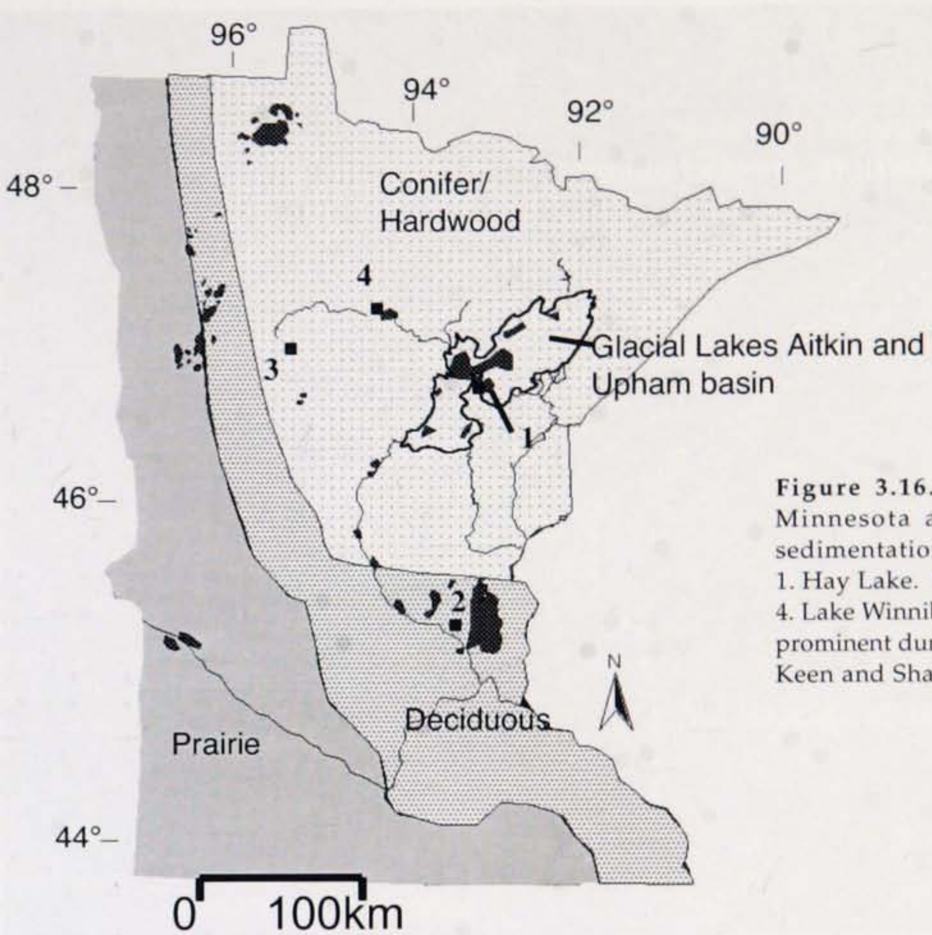

Figure 3.16. Location of sites in Minnesota associated with eolian sedimentation referenced in the text. 1. Hay Lake. 2. Lake Ann. 3. Elk Lake. 4. Lake Winnibigoshish. Filled areas are prominent dune colonies (modified from Keen and Shane, 1990).

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.

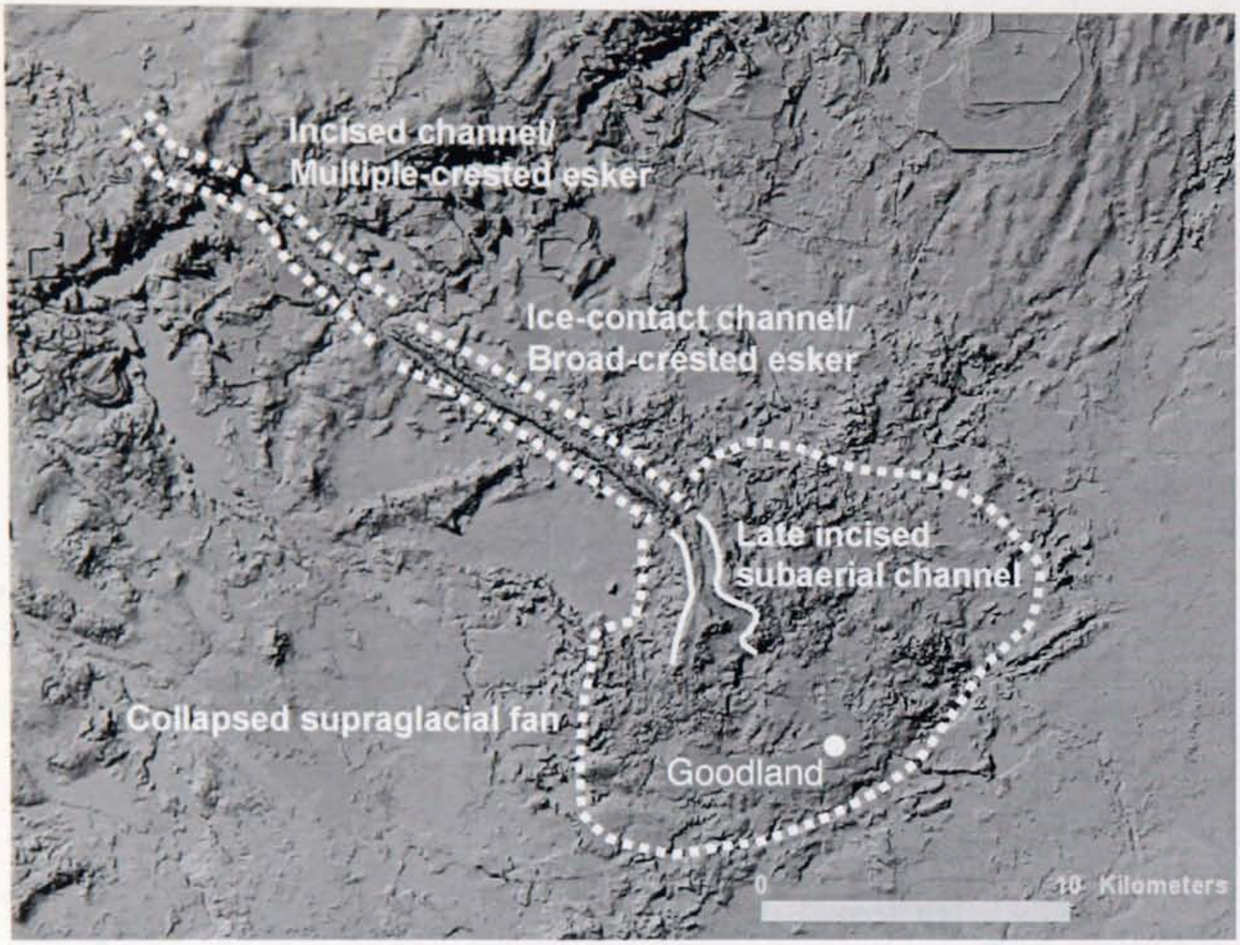

Figure 3.17. Main features of the Goodland esker system (image from 30-meter DEM of Minnesota).

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.

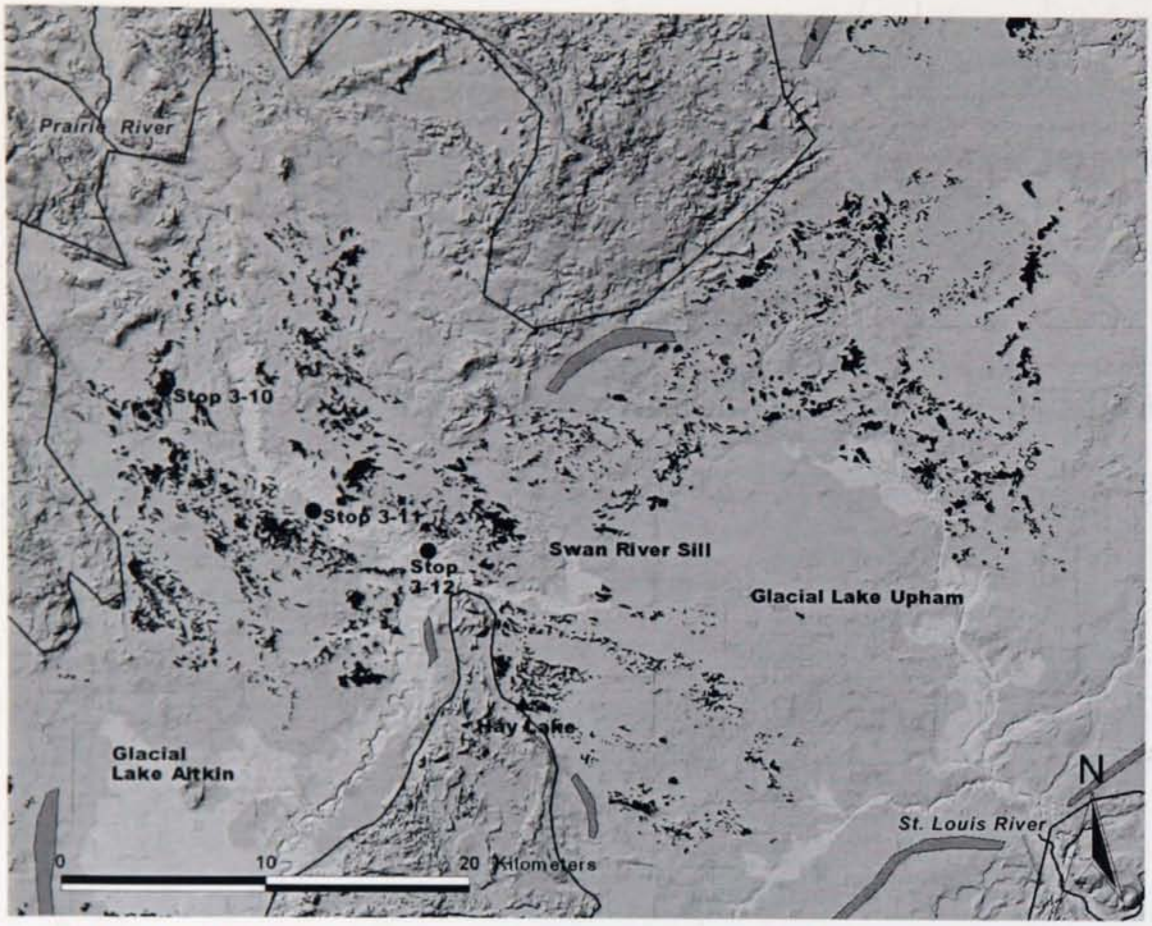

Figure 3.19. Trend in dune clusters (black) indicating a northwest–southeast wind direction. Elongate shaded polygons are beaches.

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

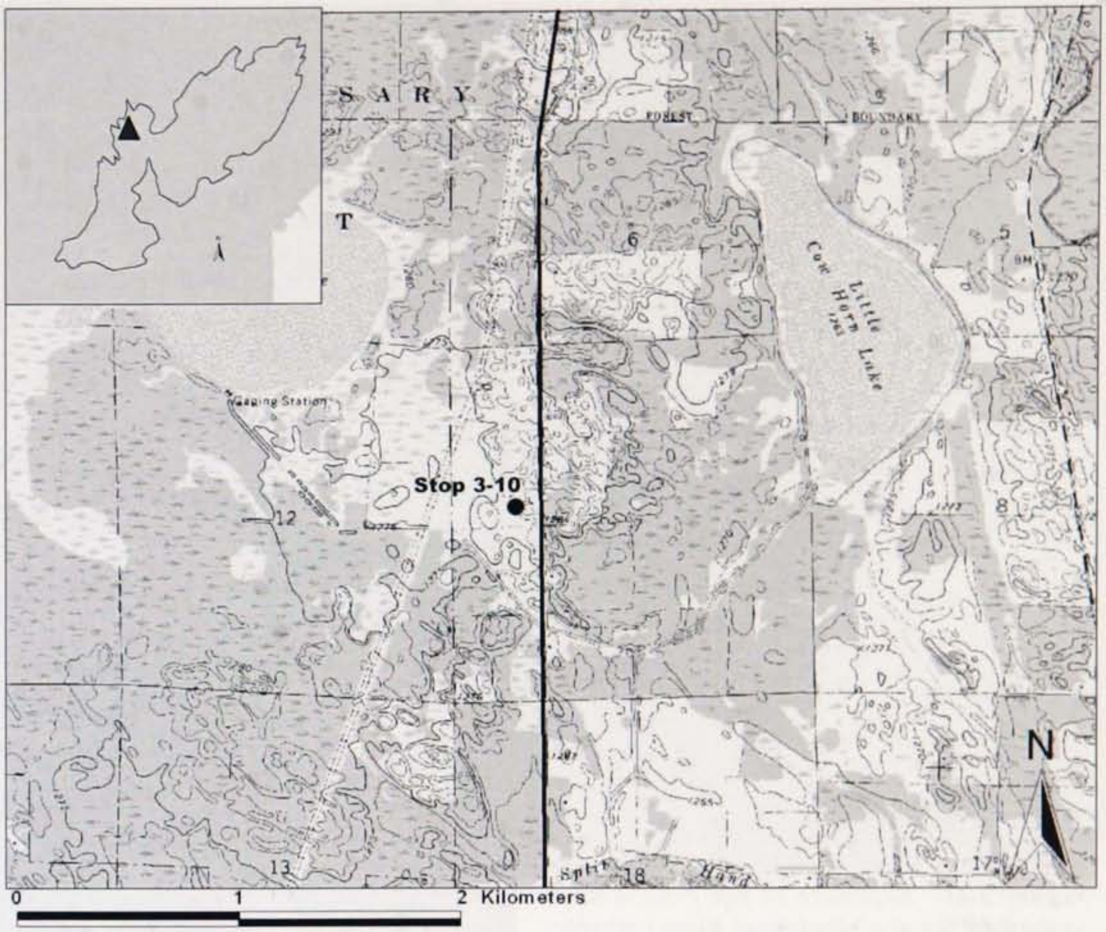

Figure 3.20. Topographic map of distorted dunes in the northern end of the glacial Lake Aitkin basin (see inset for location) showing a combination of crescentic and longitudinal dunes.

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

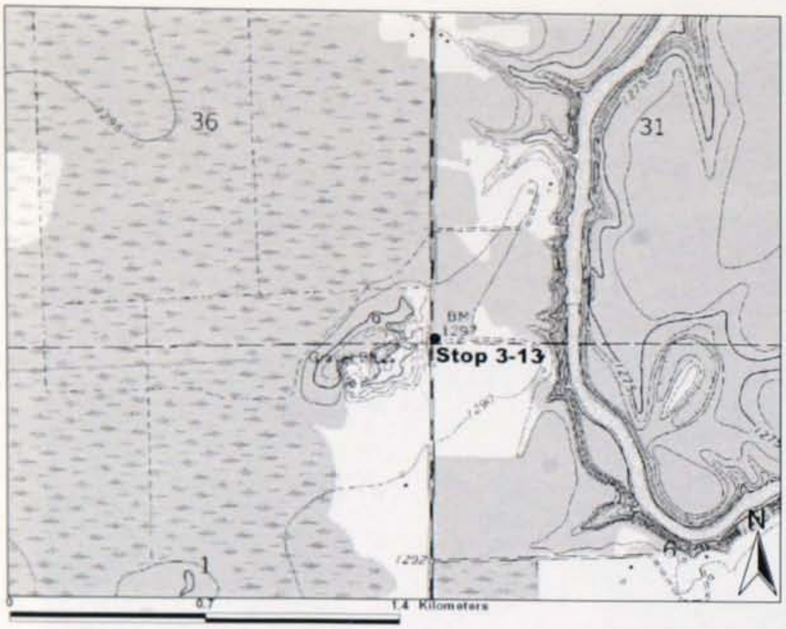

Figure 3.23. Topographic map of a wave-washed esker island in glacial Lake Upham II.

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).

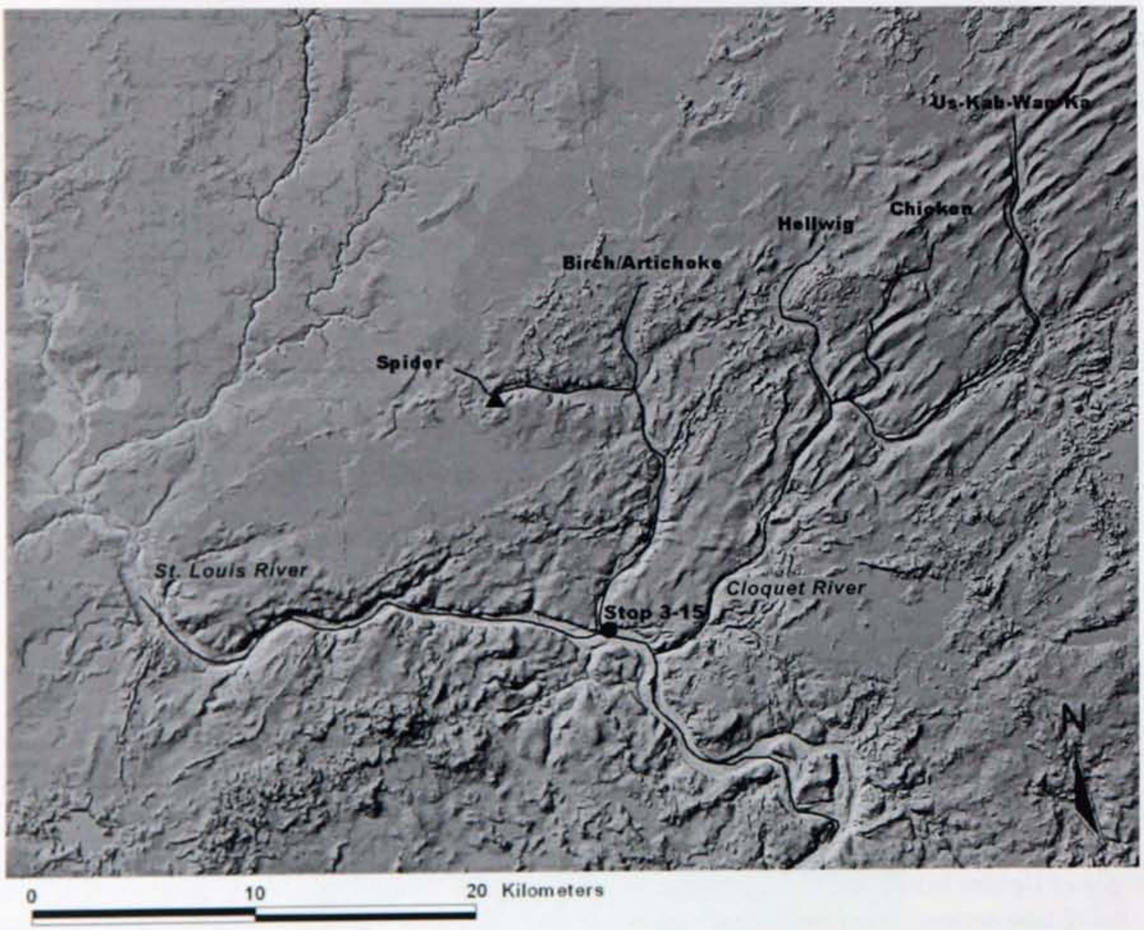

Figure 3.26. Location of early outlets to glacial Lake Upham; triangle marks the location of the core taken by Baker (1965) containing marl.

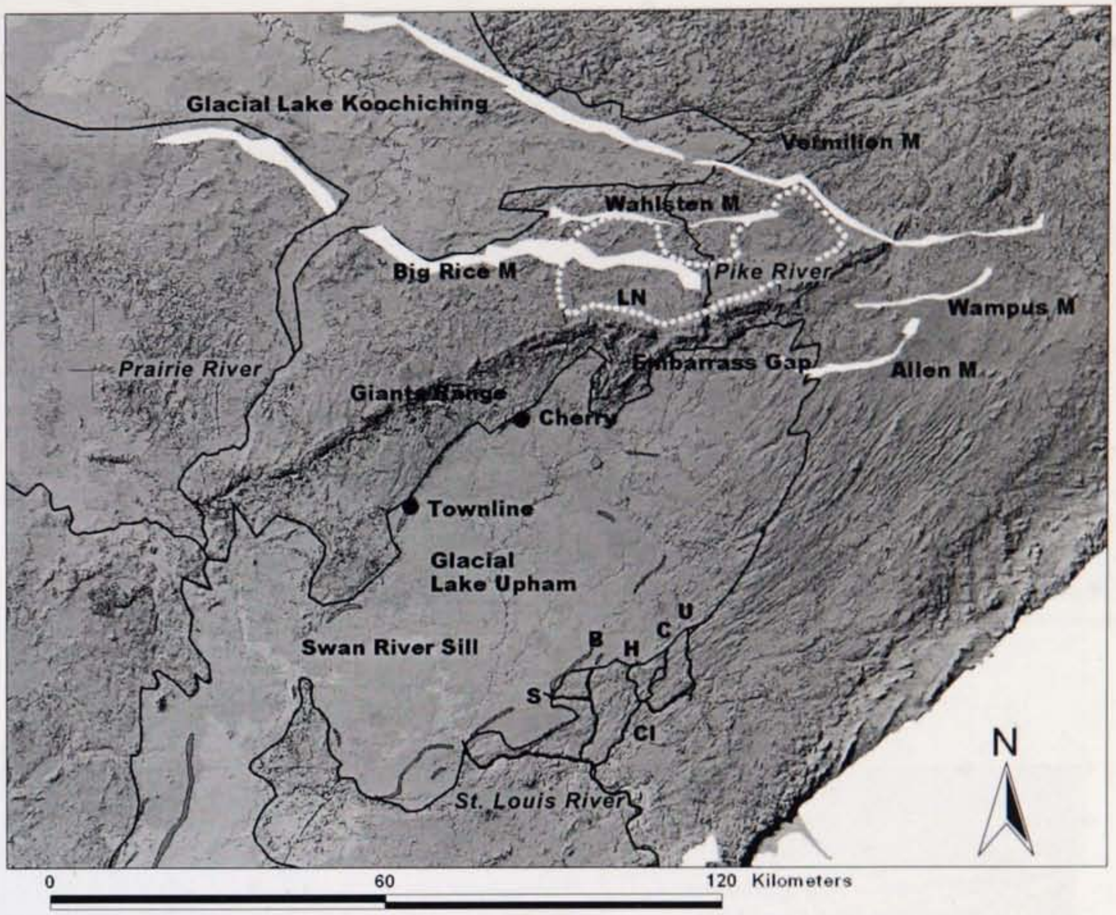

Figure 3.27. Significant features associated with the recession of the Rainy lobe after the St. Louis-sublobe advance. The extent of glacial Lake Norwood (LN) is indicated by a dotted line. White polygons are recessional moraines.

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

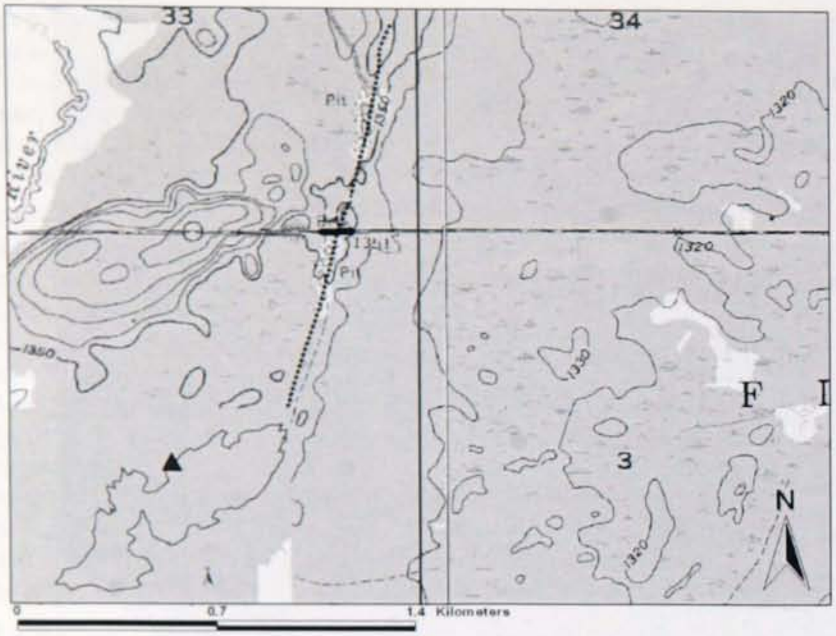

Figure 3.28. Topographic map of the shoreline of glacial Lake Upham II along Townline Road south of Hibbing. Shoreline is indicated by a black dotted line. The heavy solid line indicates where GPR data were collected (Silica and Riley quadrangles). See inset for location.

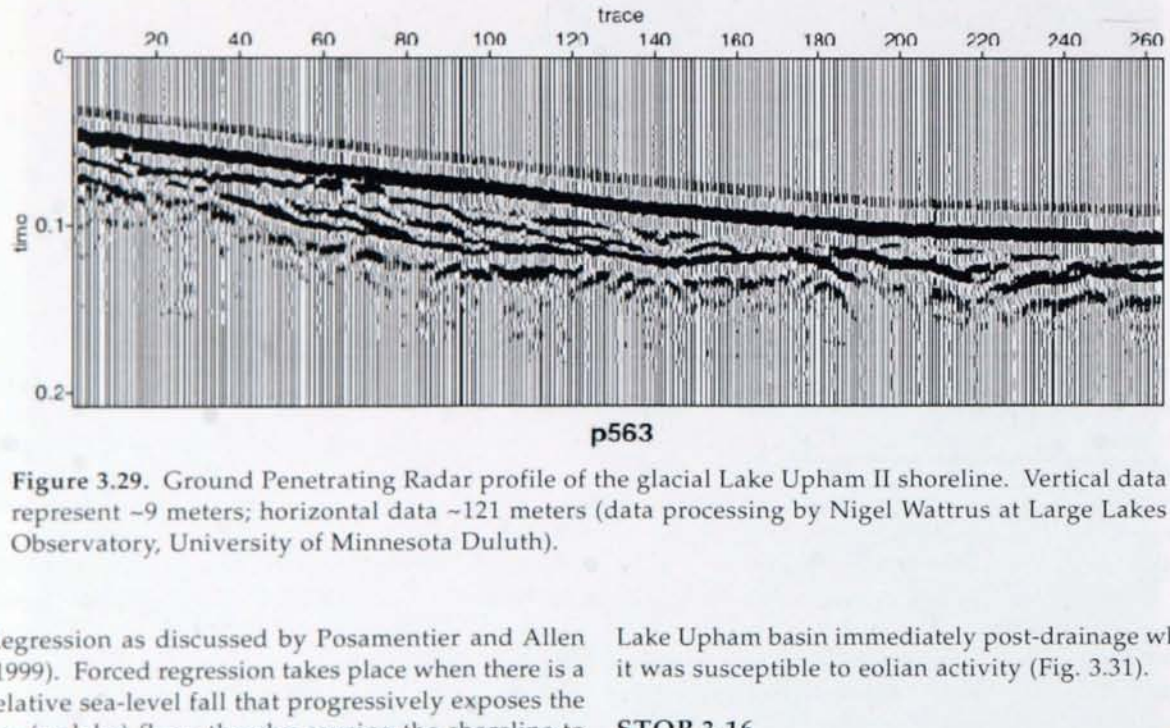

Figure 3.29. Ground Penetrating Radar profile of the glacial Lake Upham II shoreline. Vertical data represent ~9 meters; horizontal data ~121 meters (data processing by Nigel Wattrus at Large Lakes Observatory, University of Minnesota Duluth).

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

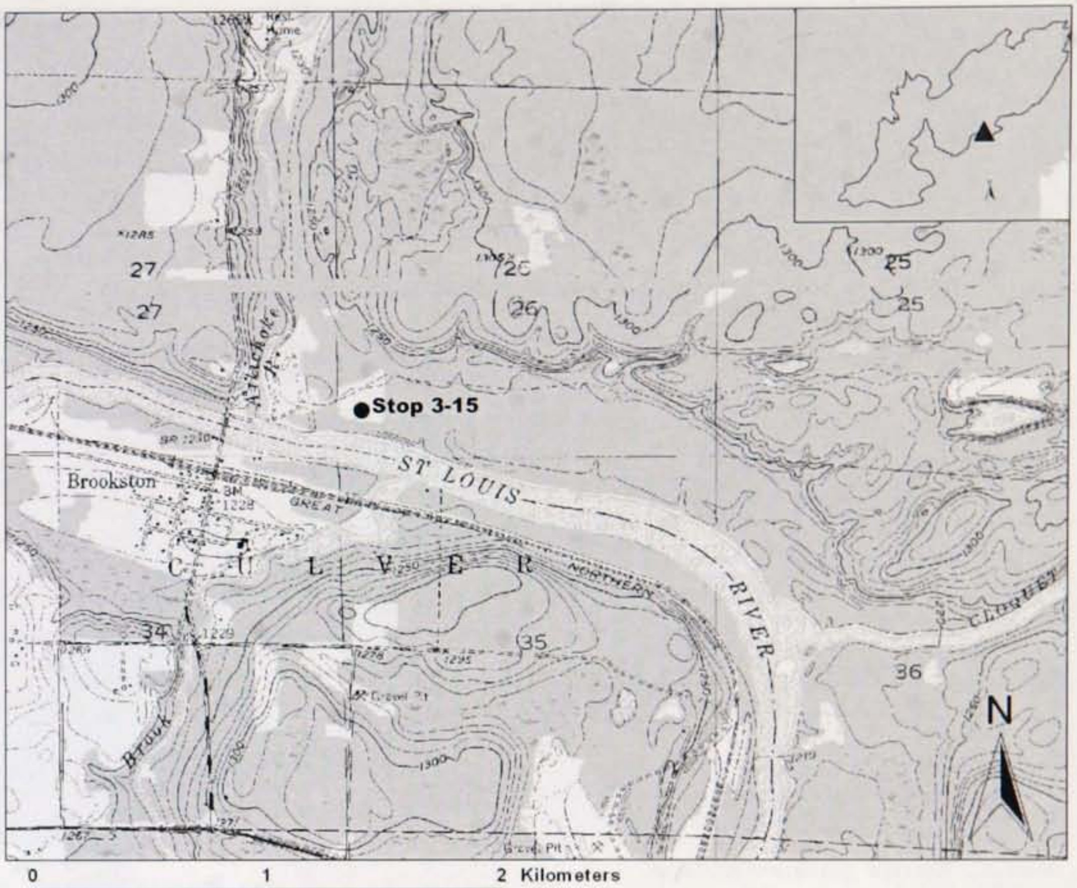

Figure 3.30. Topographic map of the Artichoke/St. Louis River terrace.

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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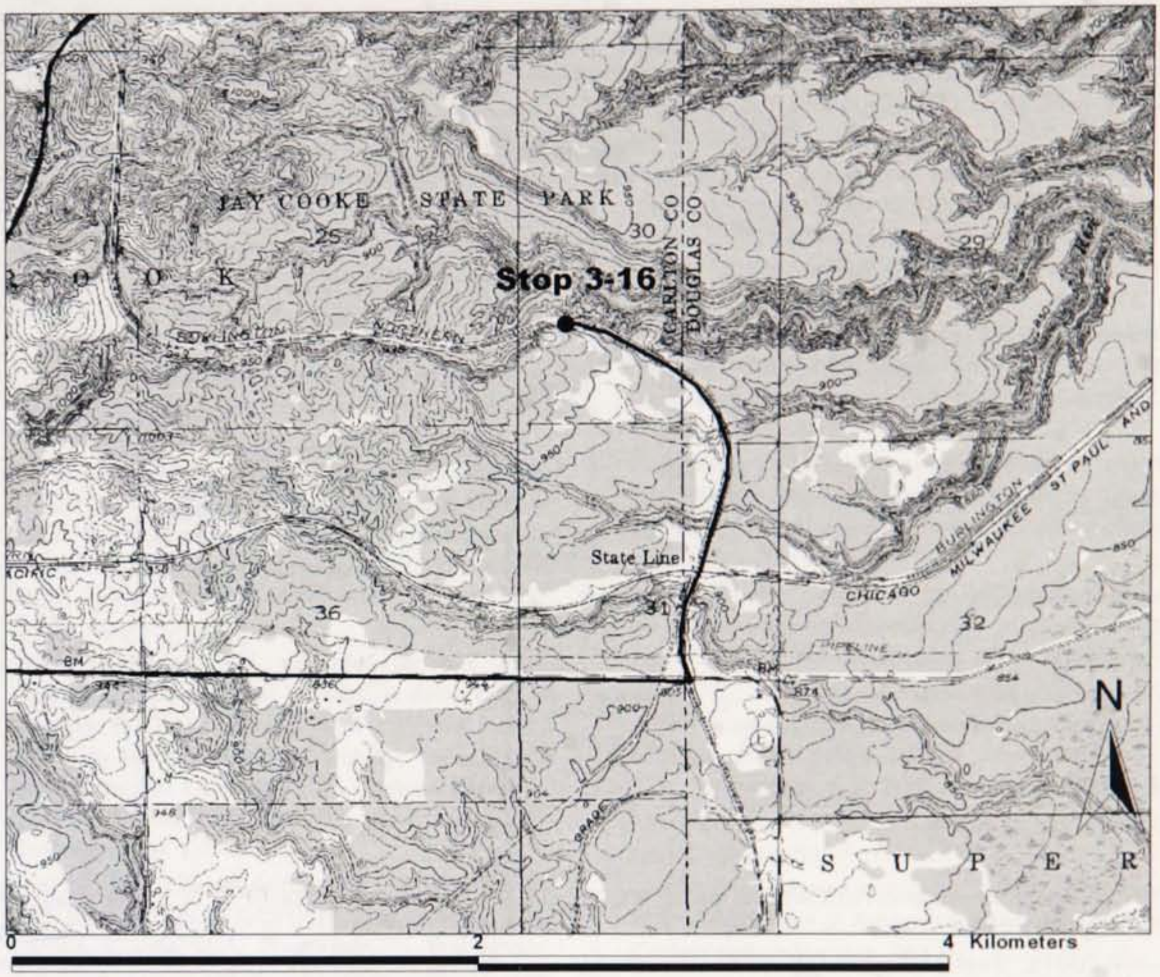

Figure 3.32. Topographic map of the underflow fan at the mouth of the St. Louis River drainage in the glacial Lake Duluth basin. The sediments were deposited by meltwater from glacial Lake Upham II.

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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.)
				Drainage of Aitkin II	
Marquette		Superior			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
			Aitkin II; Upham II; Duluth	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

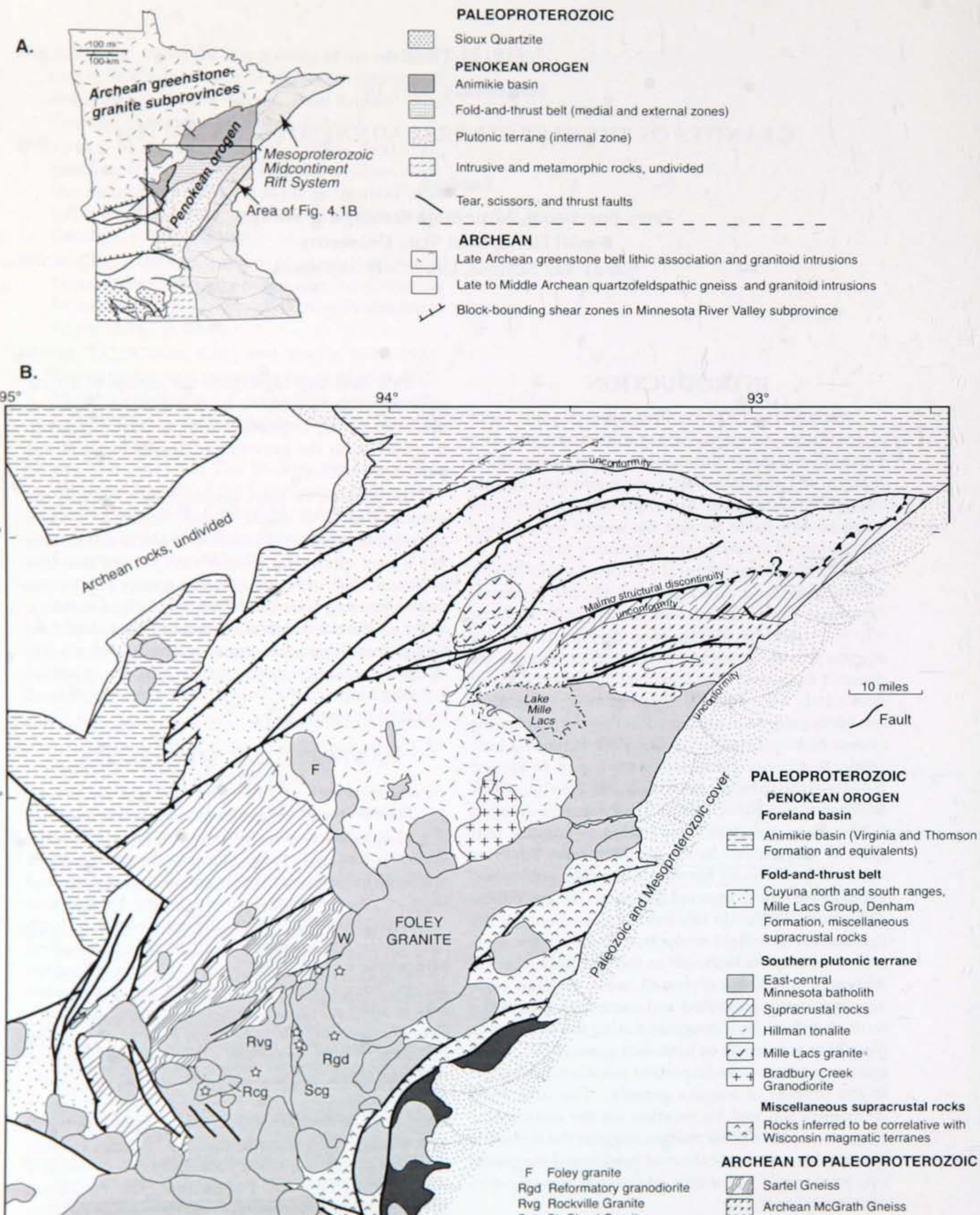
The 1,870 to 1,830 Ma Penokean orogenic rocks of the North American midcontinent are part of a vast belt of juvenile crust accreted onto the southern margin of Laurentia-Baltic continent during late Paleoproterozoic time. Subsequently, the Penokean orogenic belt twice formed the source region for a generation 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
Lake Mille Lacs
F
W
FOLEY GRANITE
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	——

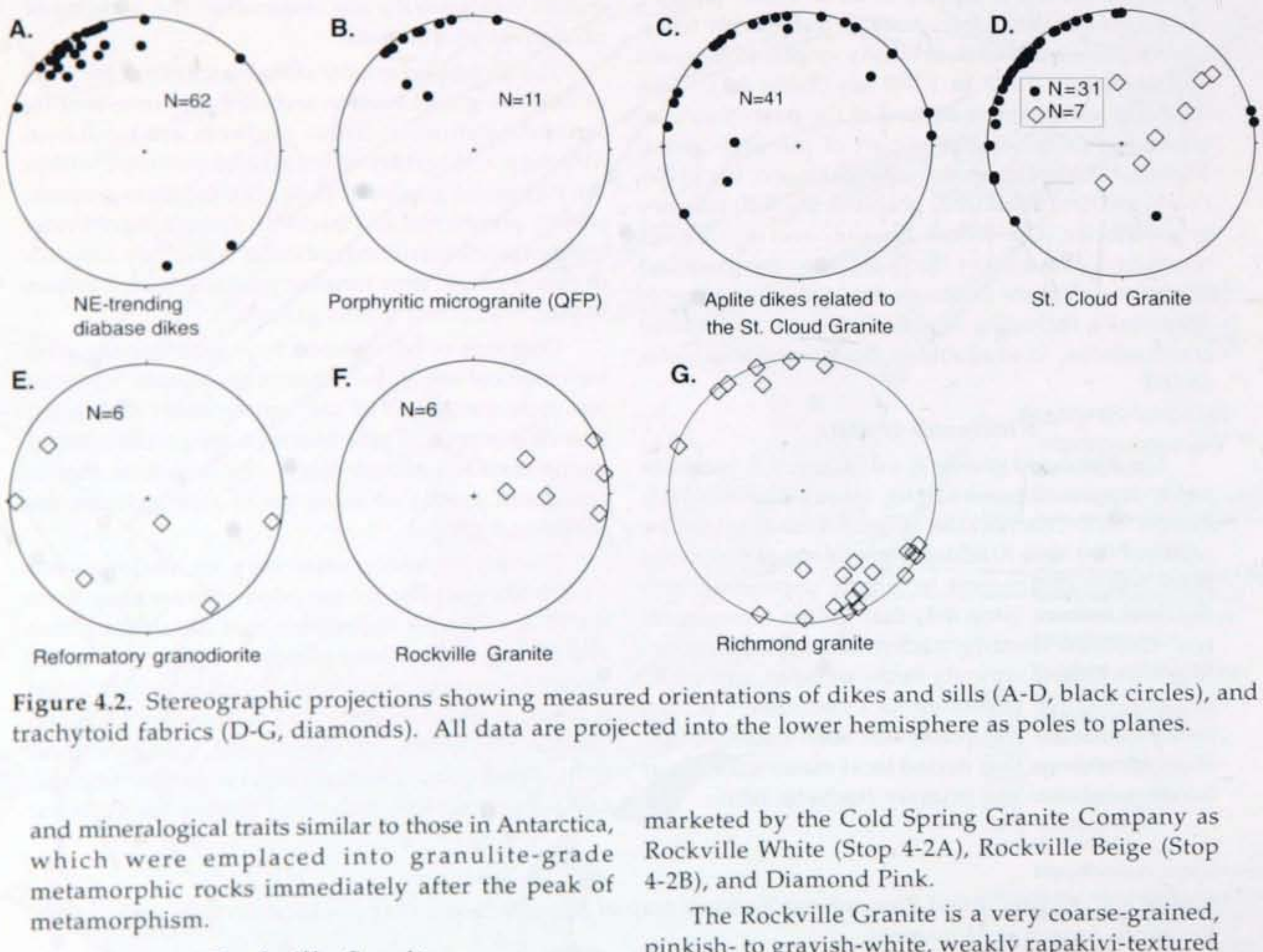


Figure 4.2. Stereographic projections showing measured orientations of dikes and sills (A-D, black circles), and trachytoid fabrics (D-G, diamonds). All data are projected into the lower hemisphere as poles to planes.

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,

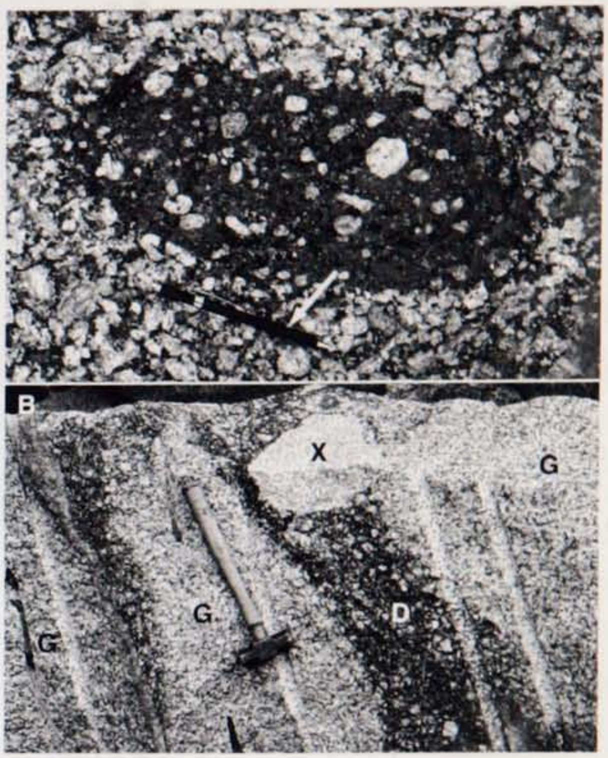


Figure 4.3. A. K-feldspar and plagioclase phenocrysts in a mafic dioritic enclave in the Rockville Granite, from the quarry at Stop 4-2A. Arrow points to a pencil for scale.

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.

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 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,

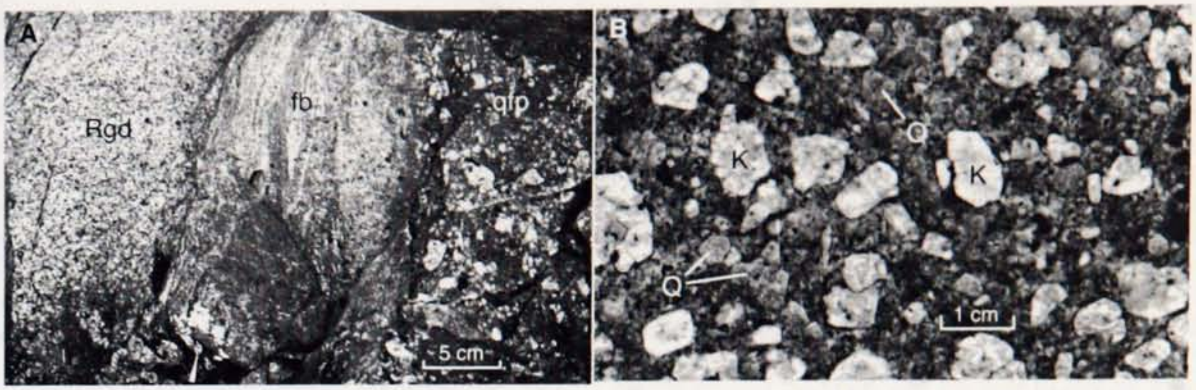


Figure 4.4. **A.** Porphyritic microgranite (quartz-feldspar porphyry) dike with flow-banded chilled margin (fb) that cuts the Reformatory granodiorite (Rgd).

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

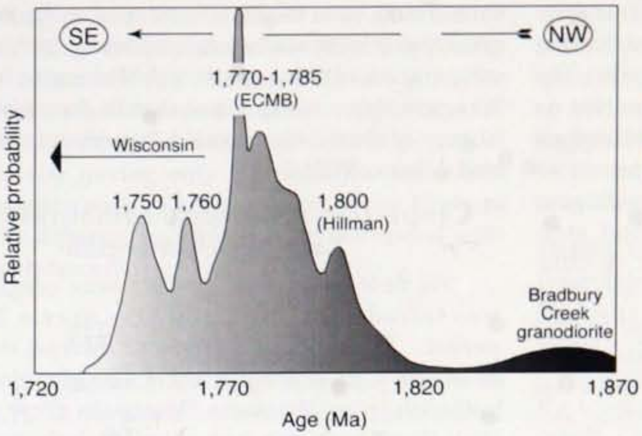


Figure 4.7. Plot of new U-Pb zircon ages for plutonic rocks across the Penokean orogen.

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

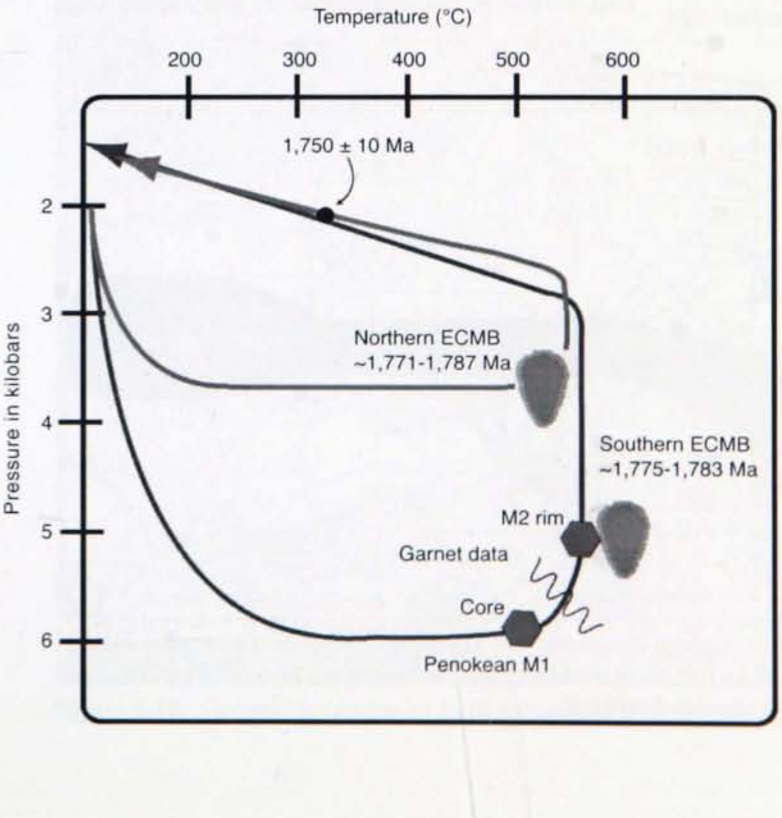


Figure 4.8. Comparison of P-T-t paths for Early Proterozoic rocks near St. Cloud (southern) and those several miles to the northeast (northern). The black circle indicates cooling through closure temperature of biotite. Pressure estimates were taken from Holm and others (1998).

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-

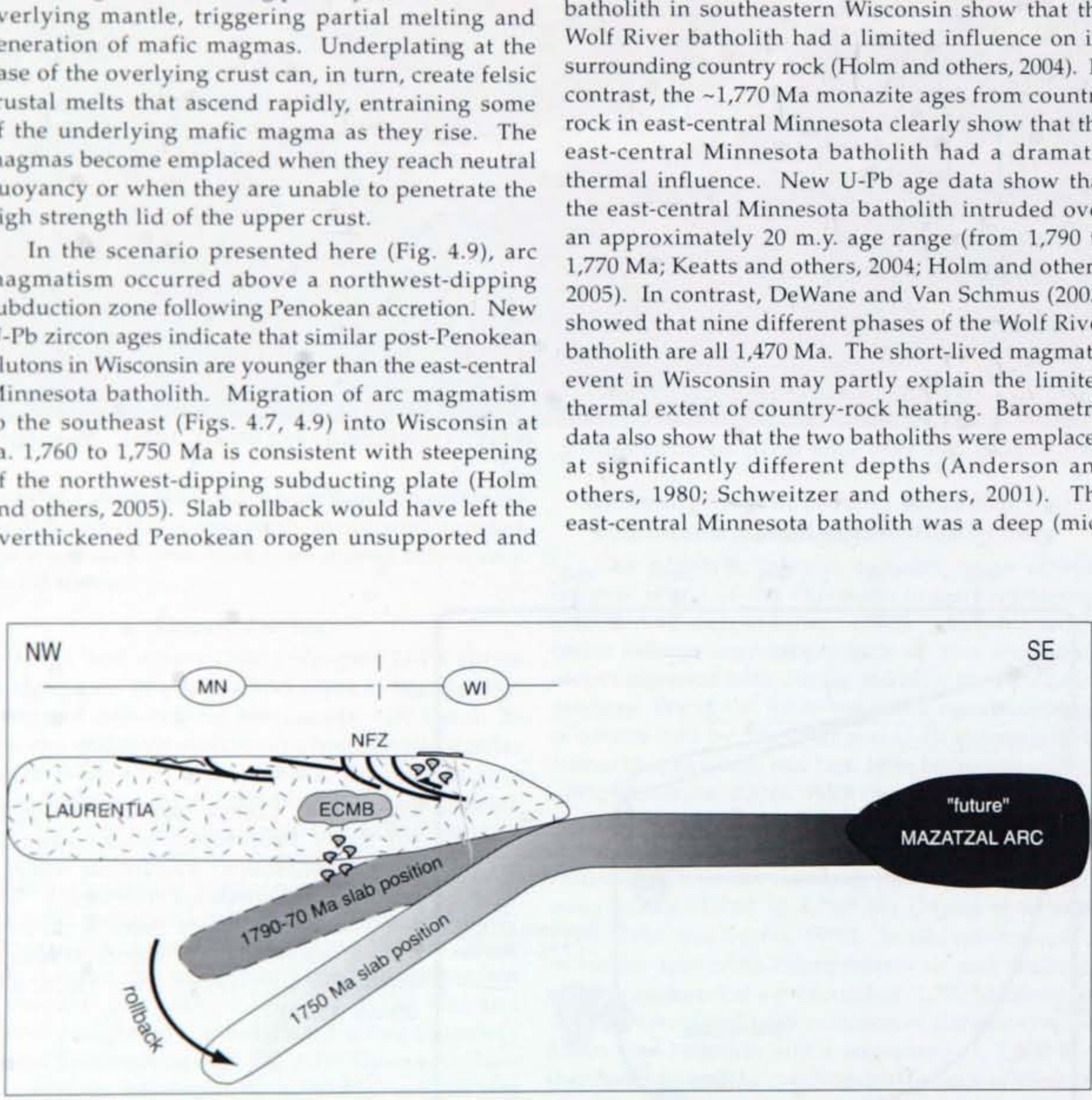


Figure 4.9. Subduction model proposed to explain magmatic arc age progression across the Penokean orogen. South-dipping thrust faults such as the Niagara fault zone in northern Wisconsin (NFZ) are Penokean in age. ECMB—east-central Minnesota batholith.

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

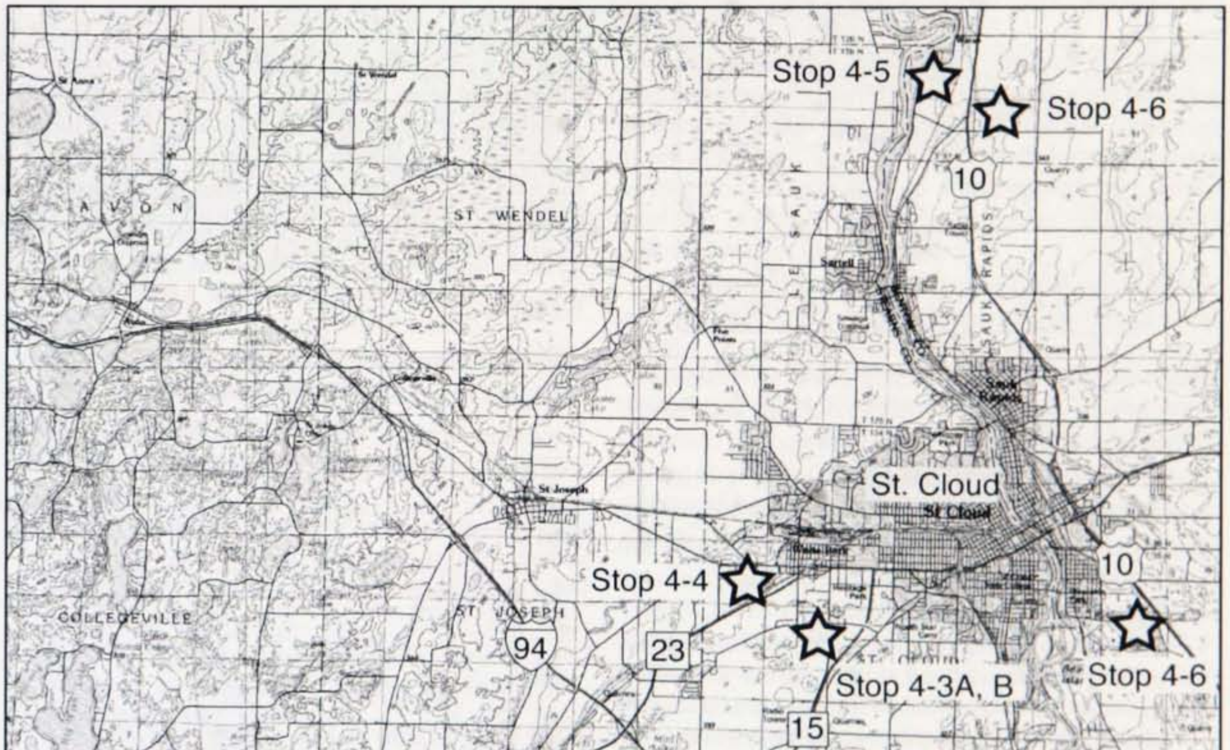


Figure 4.10. General locations of field stops for Field Trip 4.

(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.

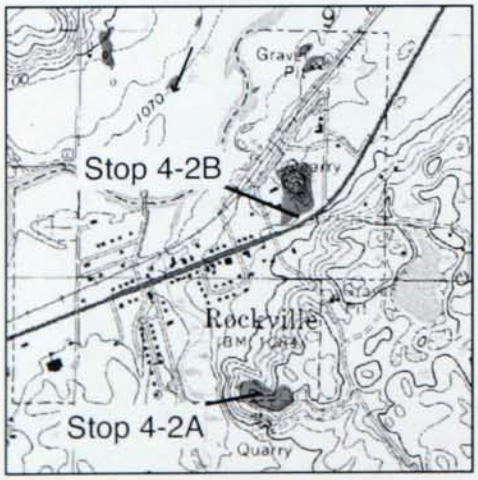
STOP 4-2A

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."

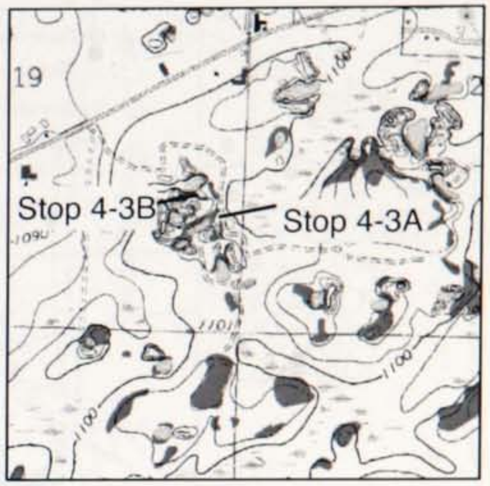
NEXT: Drive back to old Highway 23. Turn right (east), drive a few hundred feet, and stop adjacent to another large quarry just north of the highway. Look north into the quarry from the edge of the road.

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.

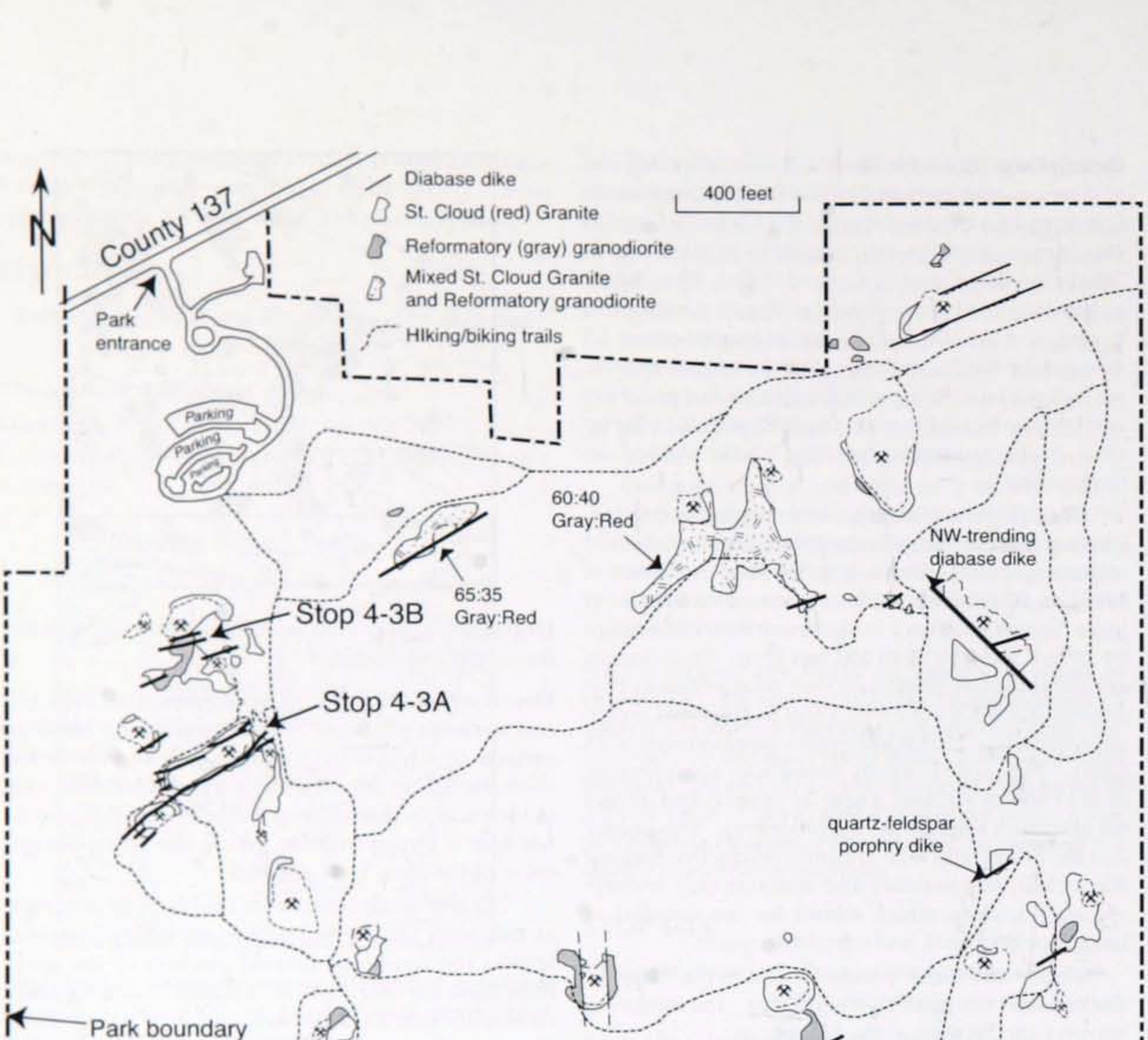


Figure 4.11. Sketch map of Stearns County Quarry Park showing general rock types, quarry locations, and trails.

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

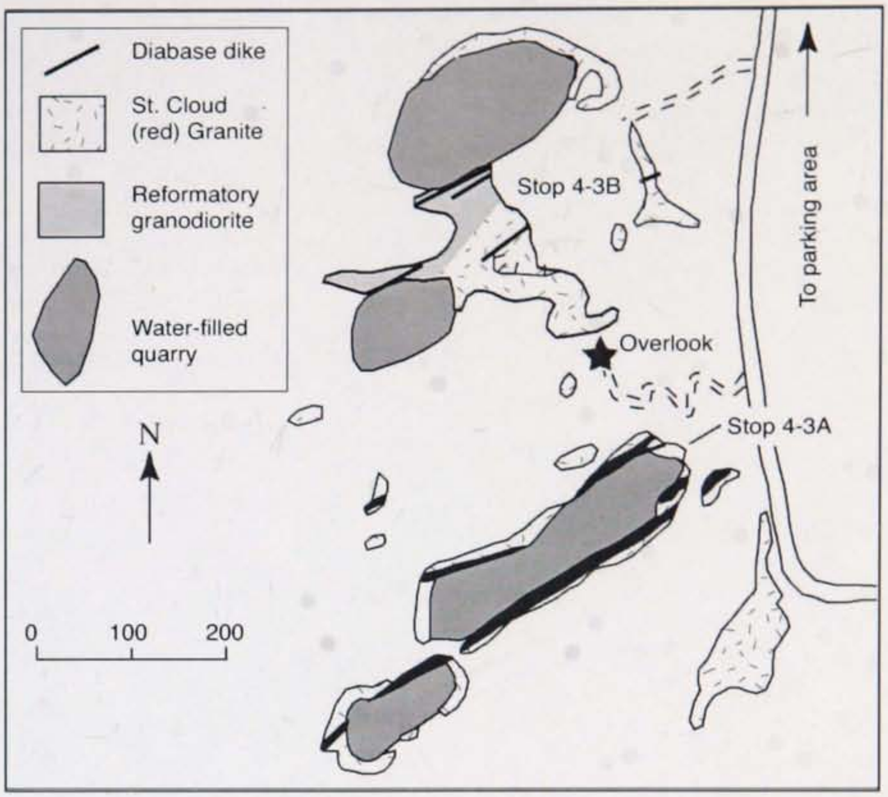
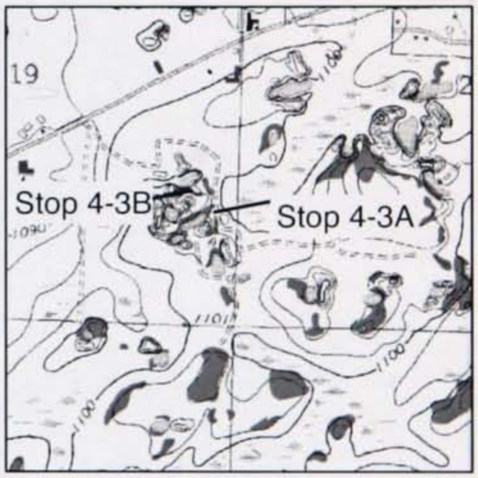


Figure 4.12. Detail of Stops 4-3A and 4-3B.



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.

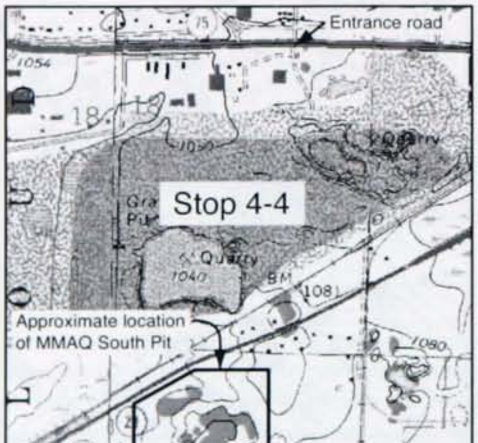
NEXT: Leave the park and turn right on County Road 137. Proceed approximately 0.7 mile to 10th Avenue. Turn left (north) on 10th Avenue and proceed approximately 0.9 mile to Stearns County Road 75/Division Street. Turn left (west) on 75, travel approximately 0.6 mile, and turn left into the Martin-Marietta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

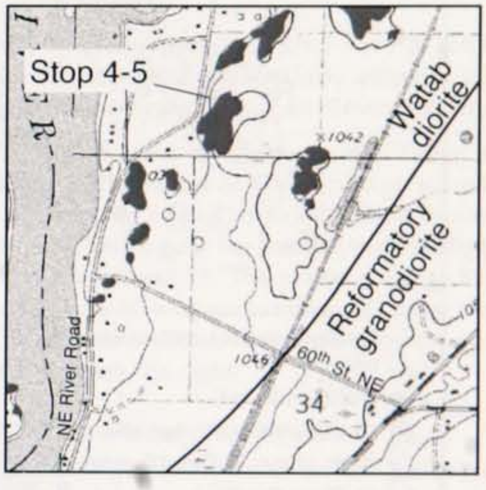
NEXT: Drive east on County Road 75 1.5 miles to the intersection with State Highway 15. Turn left (north) on 15 and stay on it to U.S. Highway 10, a distance of approximately 5 miles. Go north on Highway 10 for 2.5 miles to the 60th Street/Frost Road NW intersection and turn left (west) on Frost Road. 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

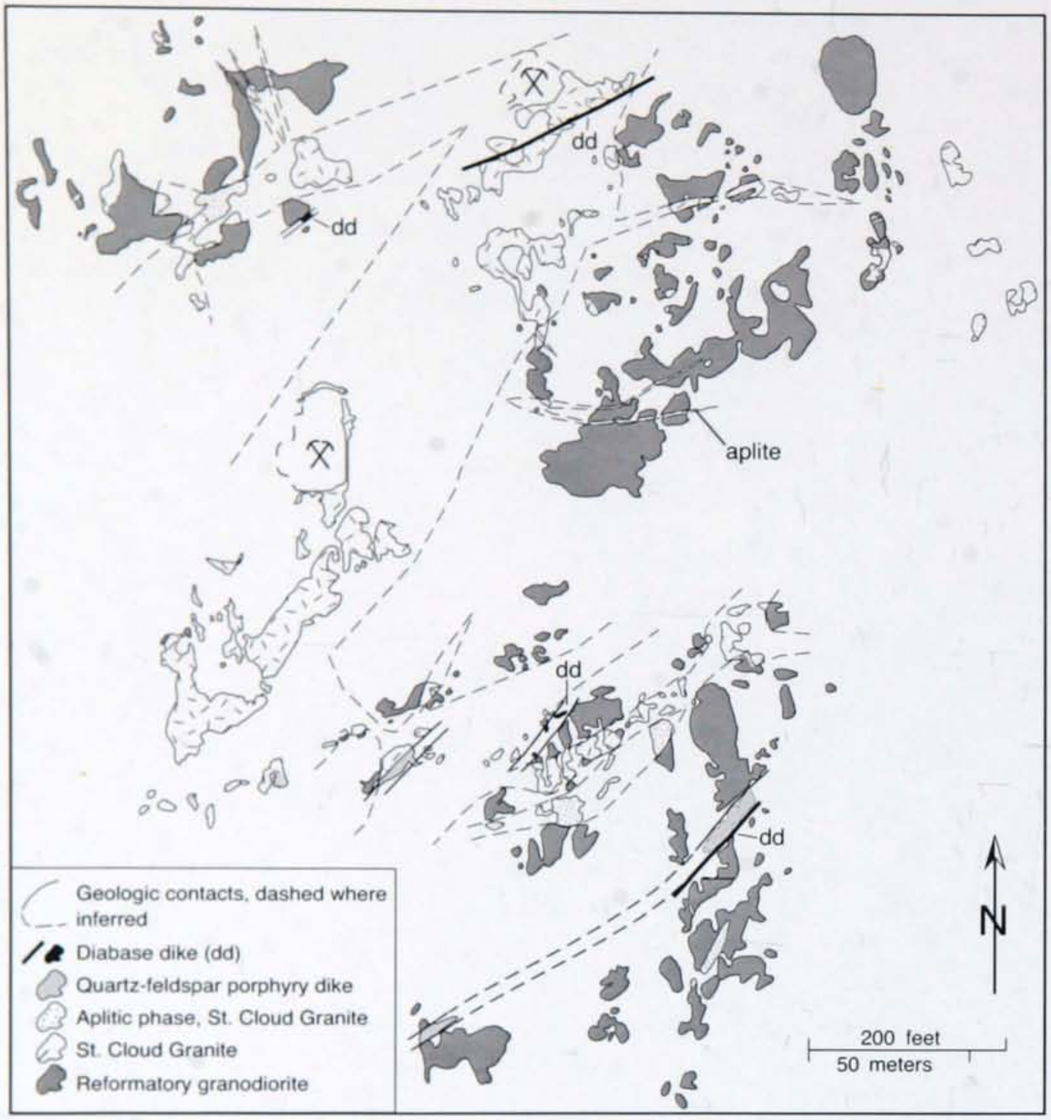


Figure 4.13. Pre-quarry geologic outcrop map of the area now occupied by the Martin-Marietta quarry south of Highway 23; modified from Boerboom and Holm (2000).

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

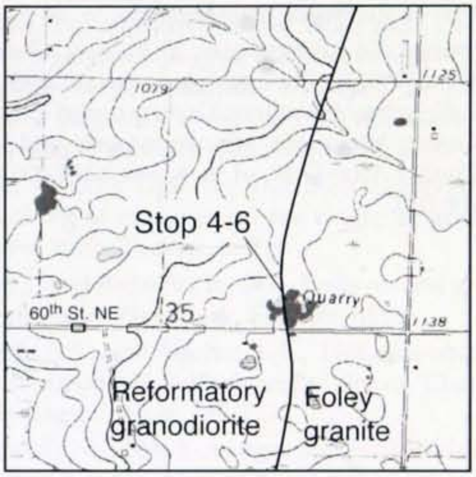
Private property! Permission must be obtained before entering!

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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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).

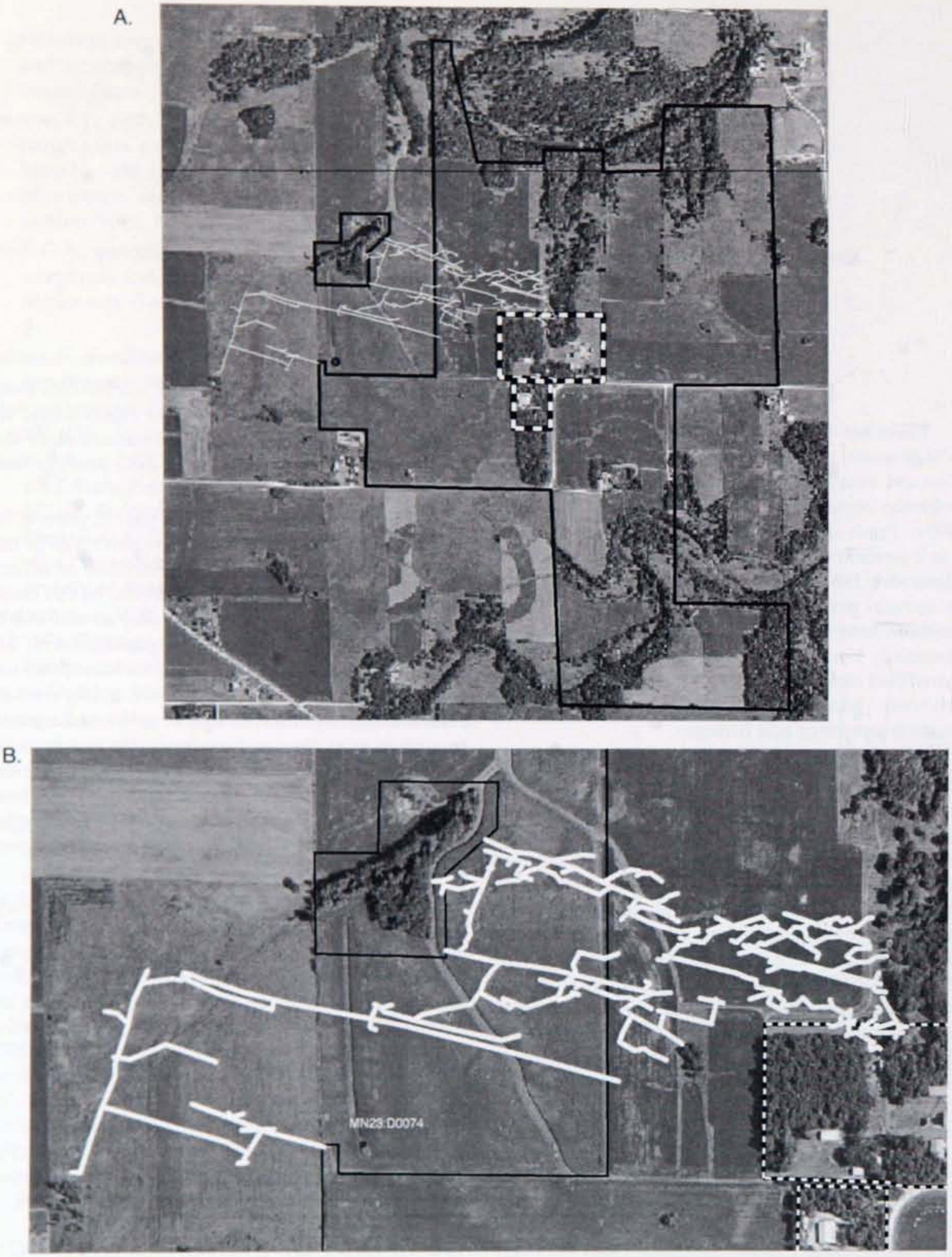
Based on his experience of the past 15 years at the site, Mr. Ackerman has found that sinkholes in this setting typically identify the intersection of several passageways; by excavating a sinkhole, he can create a new access point to a cave that may or may not be connected to previously mapped passageways. 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.

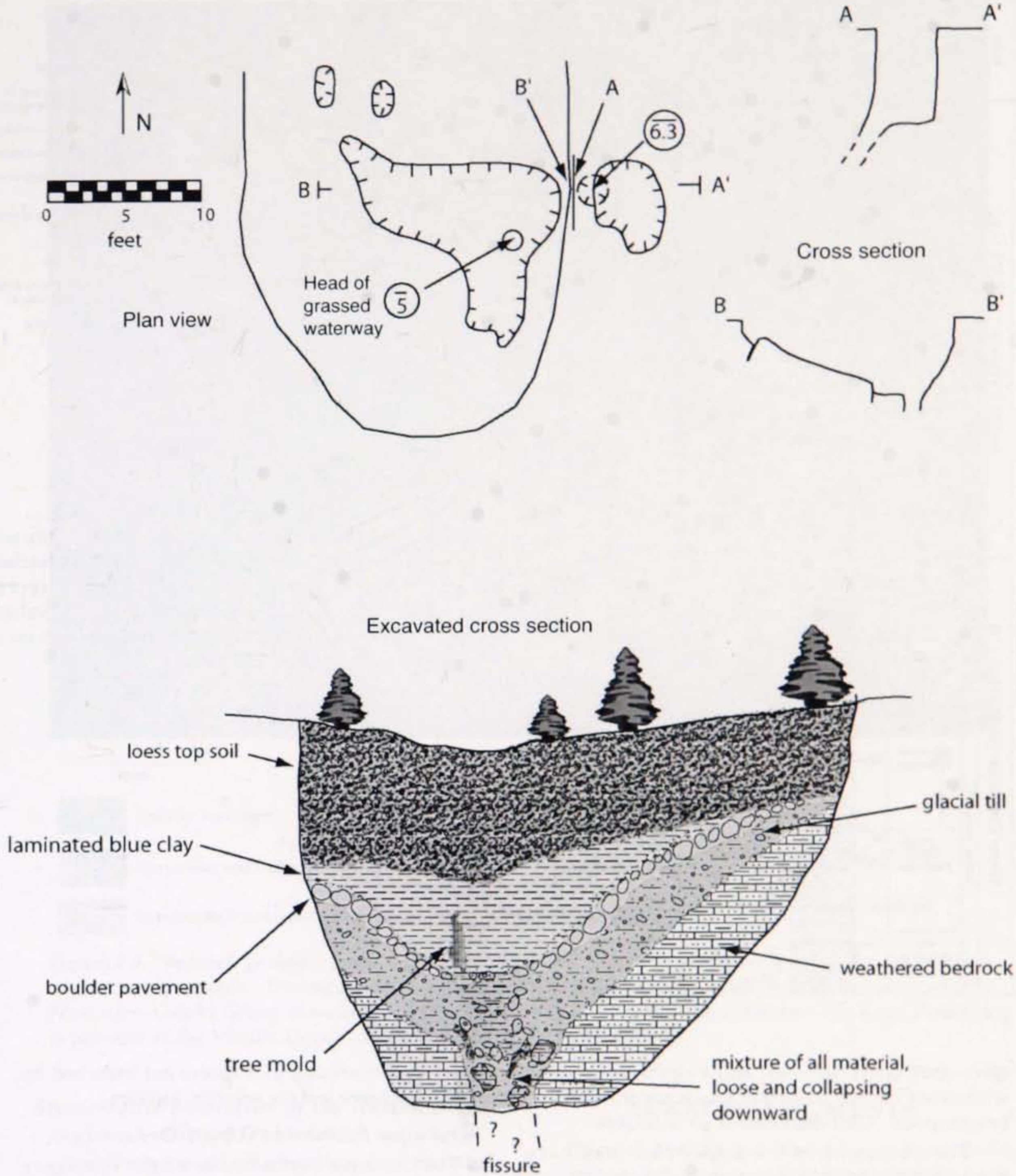


Figure 5.3. Cross section and plan view of sinkhole MN23:D0074. The sequence at the site includes "A" horizon top soil, thick loess, laminated blue clay, boulder pavement, till, and weathered bedrock. A black, funnel-shaped column of "A" horizon material from the land surface to the bedrock fissure is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

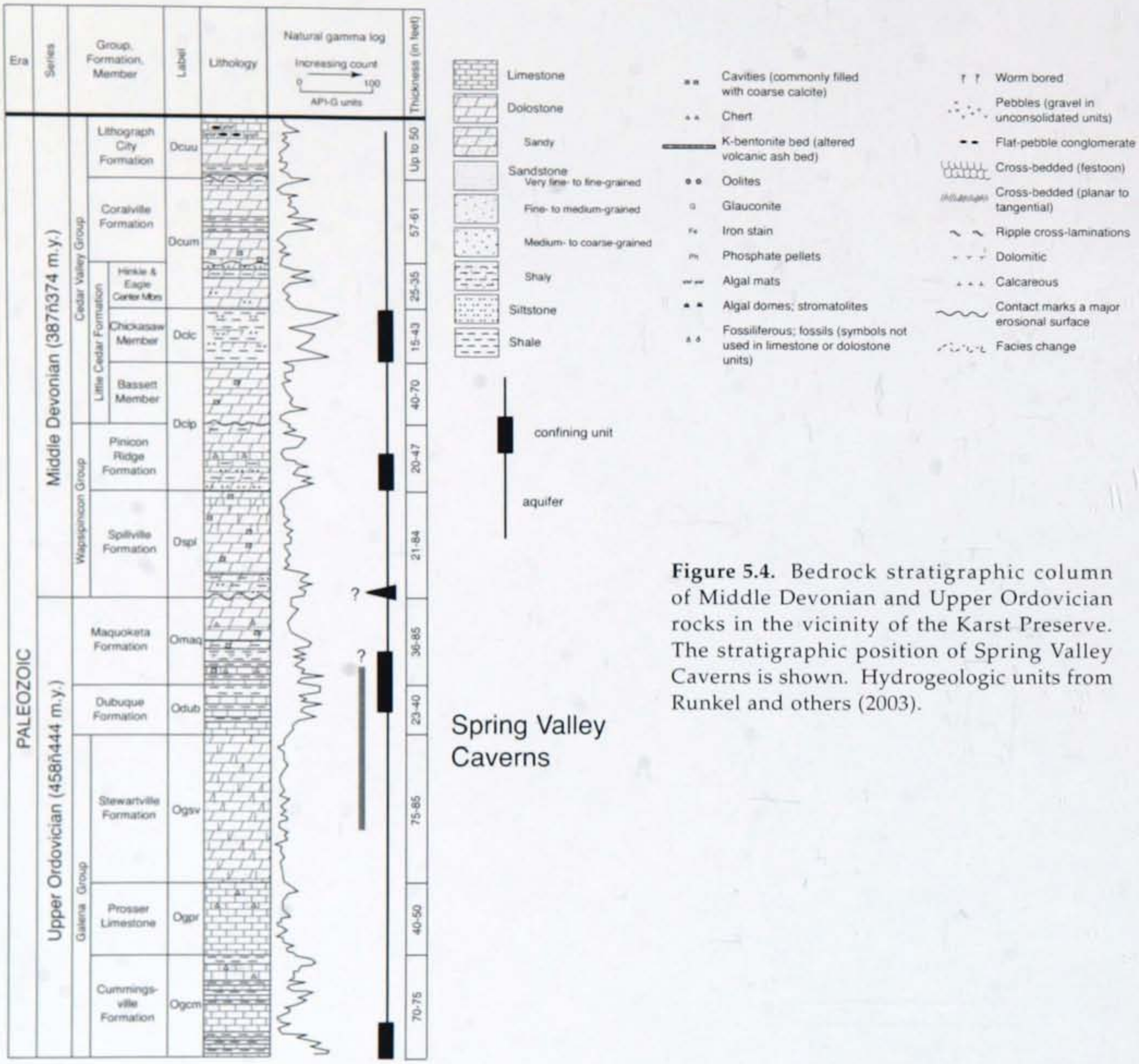


Figure 5.4. Bedrock stratigraphic column of Middle Devonian and Upper Ordovician rocks in the vicinity of the Karst Preserve. The stratigraphic position of Spring Valley Caverns is shown. Hydrogeologic units from Runkel and others (2003).

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).

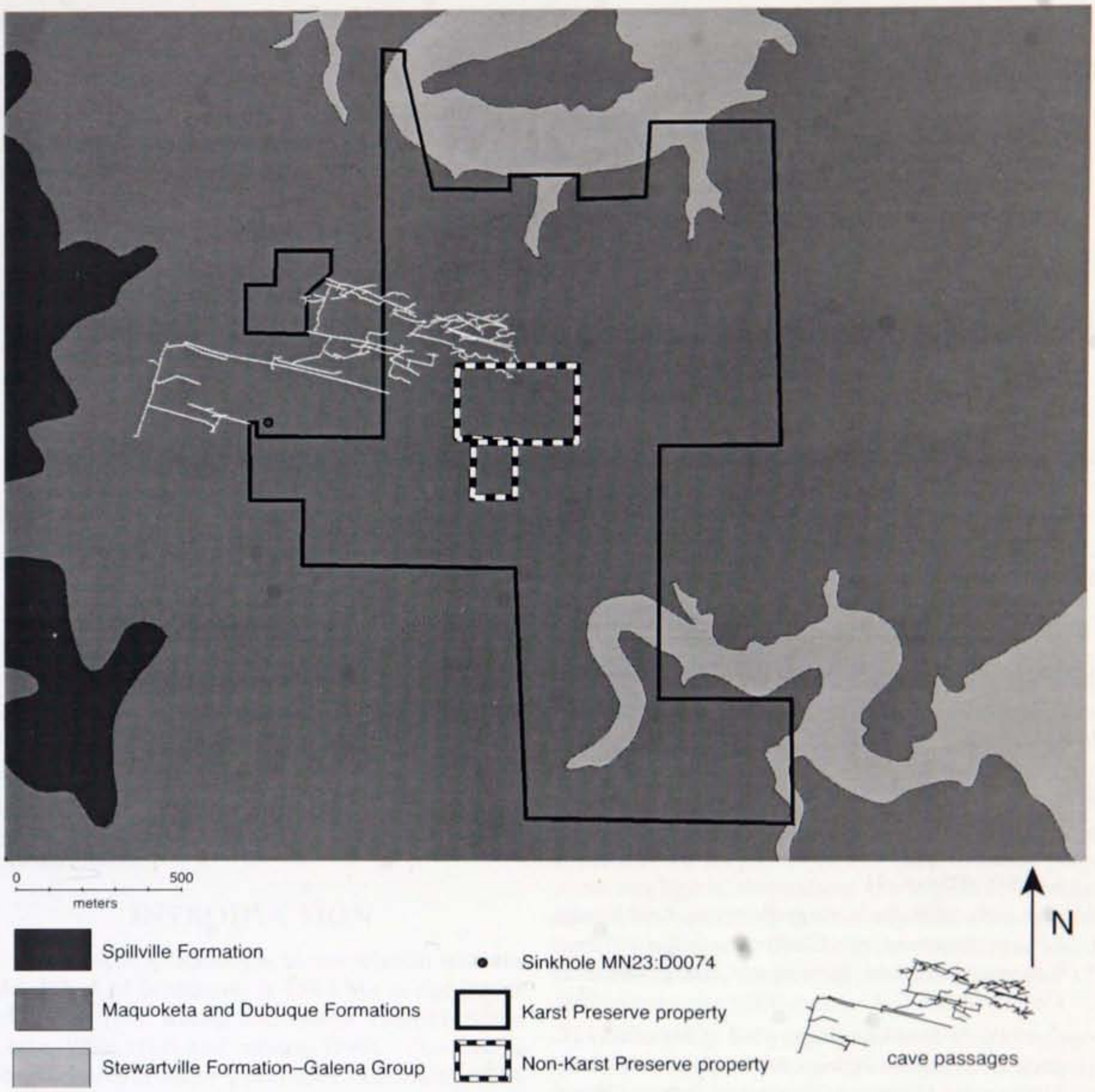


Figure 5.5. Bedrock geologic map of the Karst Preserve property, along with approximate location of the cave passages. Downcutting of the bedrock surface to the Upper Ordovician Stewartville Formation–Galena Group in surface drainageways to the north and south is shown. The Karst Preserve is just east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

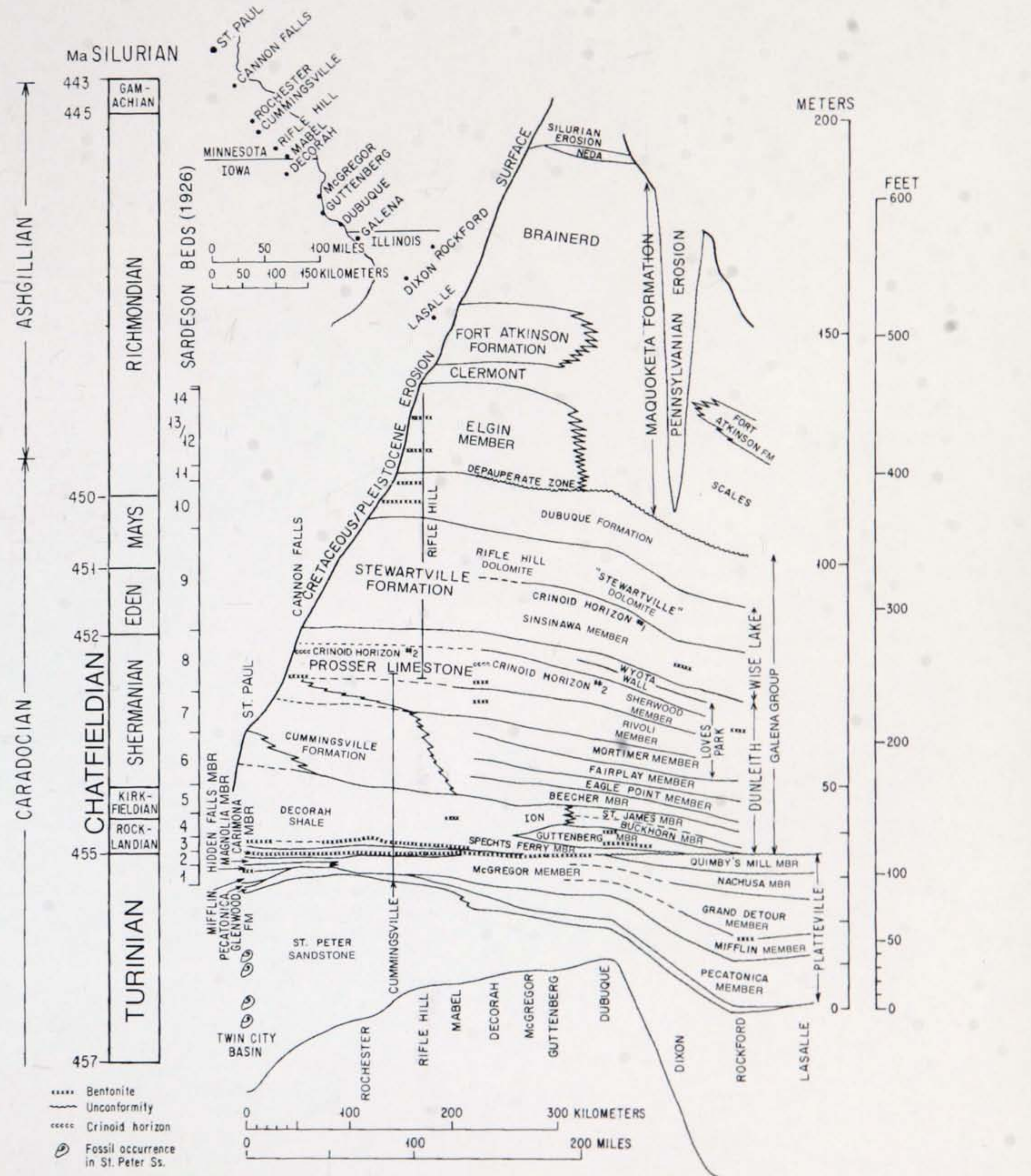


Figure 6.1. Stratigraphic cross section of Late Ordovician rocks between St. Paul, Minnesota and LaSalle, Illinois, with correlations, zones, and ages. Modified from Sloan (1987) with input from Goldman and Bergstrom (1997) and Webby and others (2004).

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.

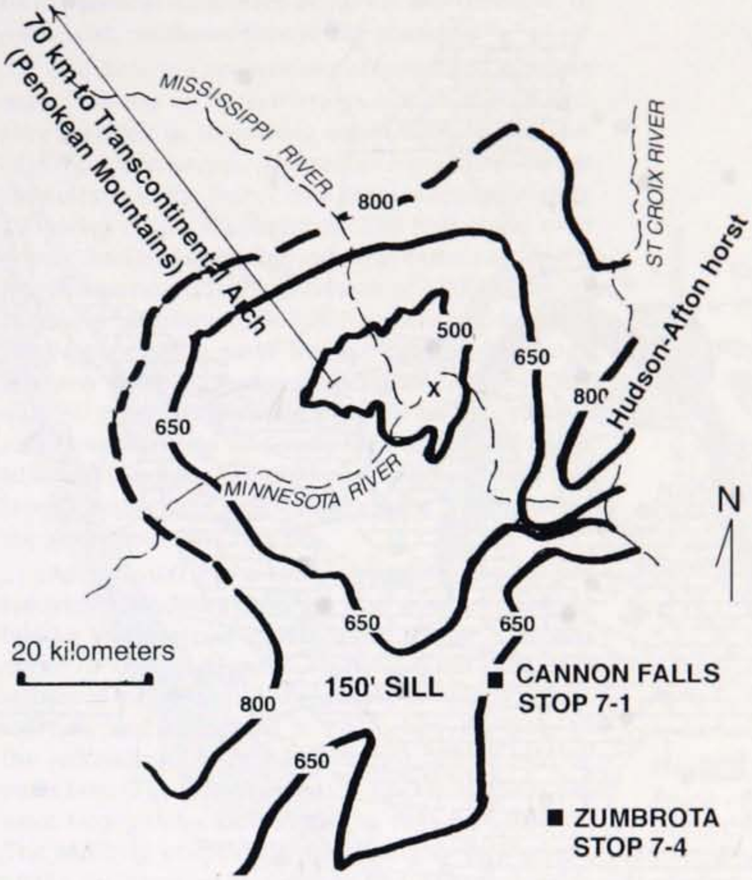


Figure 6.2. Structural contour map of the Twin Cities basin, based on the Jordan Sandstone–Oneota Dolomite contact (approximately the Cambrian–Ordovician contact). Contour interval is 150 feet, the locality of Lilydale Park is marked by an X, Cannon Falls and Zumbrota stops (Stops 7-1 and 7-4, respectively)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

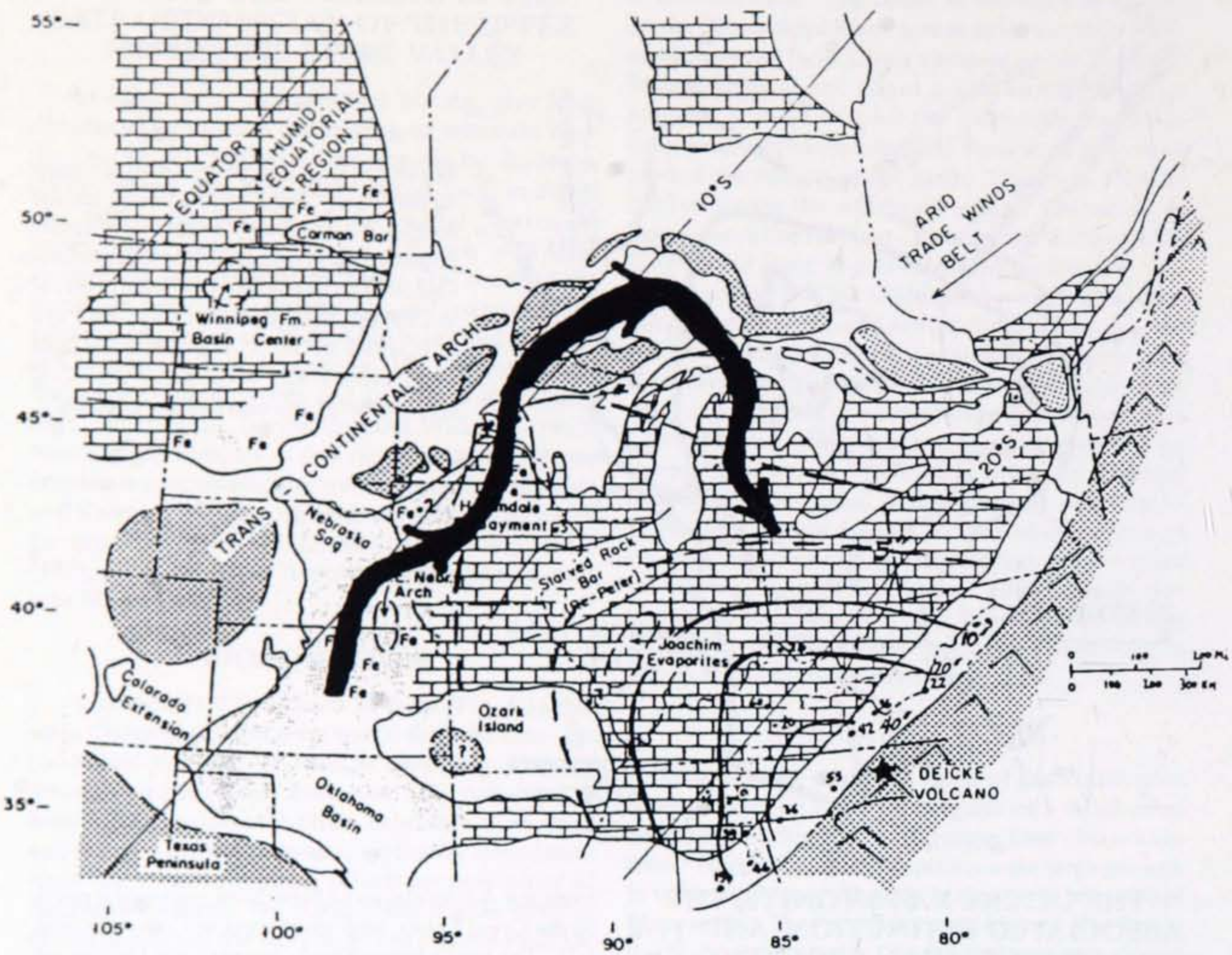


Figure 6.3. Late Ordovician paleogeography of the eastern half of the United States and Canada showing the Transcontinental Arch. Heavy stipple is persistent highlands, light stipple is siliciclastic dominated. Heavy arrow is the axis of the 1.1 Gy Keweenawan Rift, suspected route of post Deicke event 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

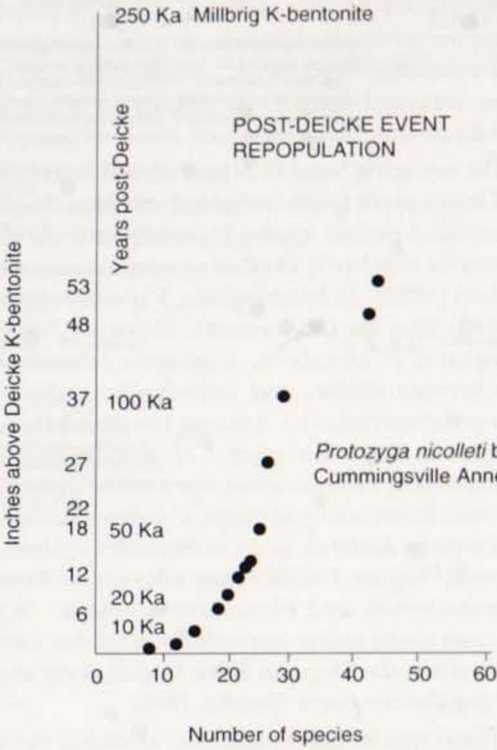


Figure 6.4. Plot showing the increasingly varied fauna of the Spechts Ferry Member part of the basal Chatfieldian stage, to show the rate of repopulation after the Deicke extinction. Data from Weiss (1953), Sloan (1972, unpub. data), and Rice (1987), all projected into the Cummingsville Annex section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

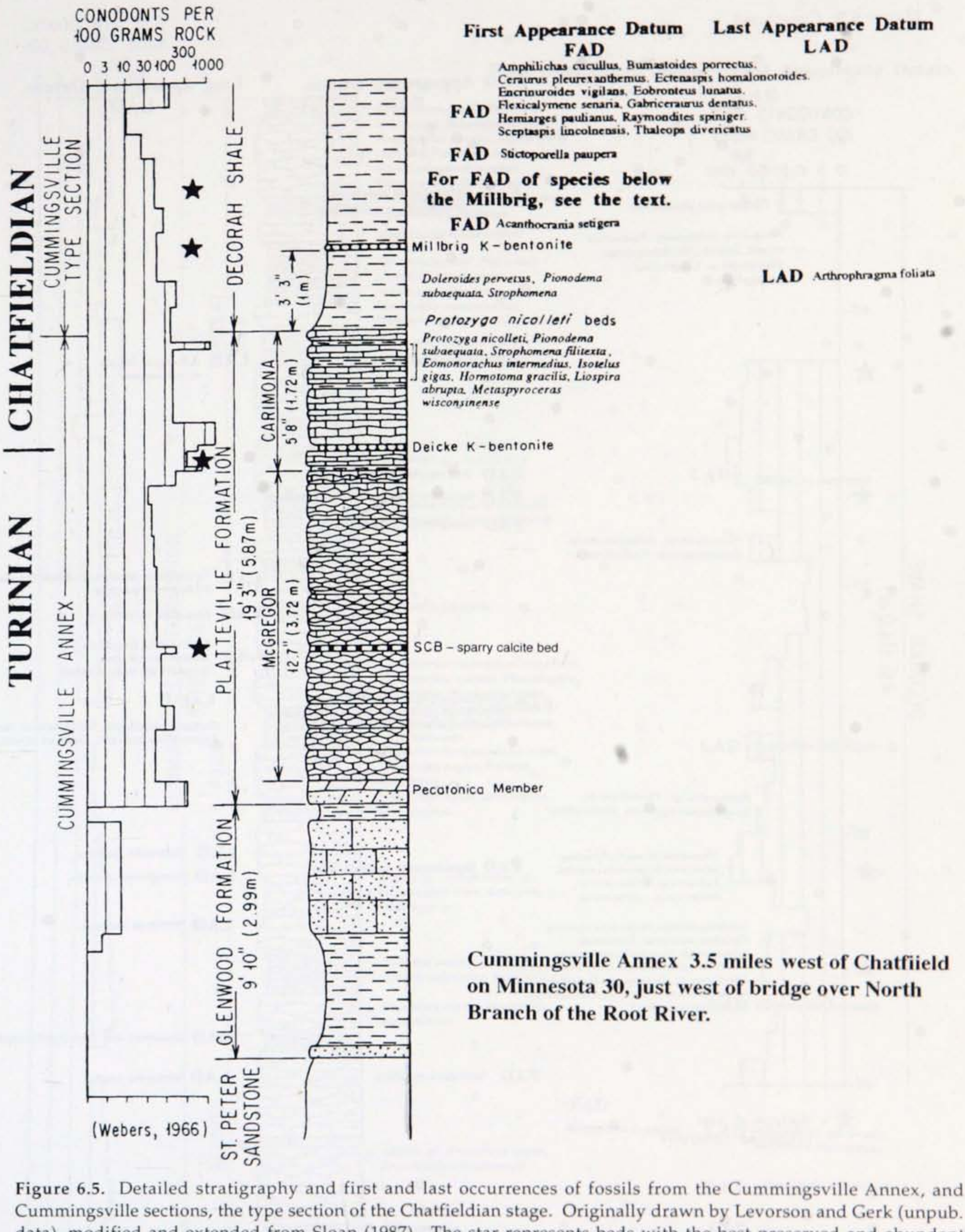
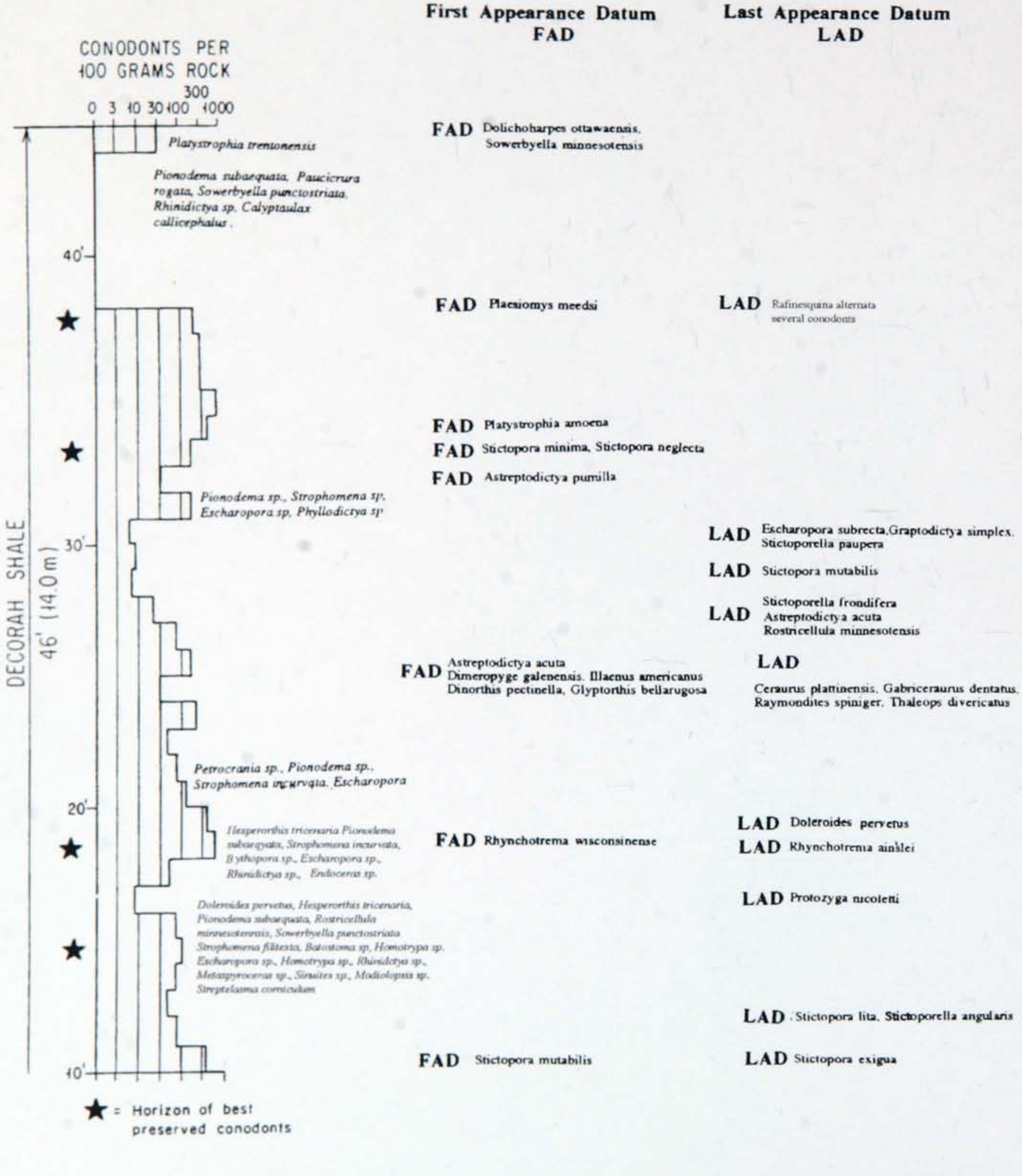


Figure 6.5. Detailed stratigraphy and first and last occurrences of fossils from the Cummingsville Annex, and Cummingsville sections, the type section of the Chatfieldian stage. Originally drawn by Levorson and Gerk (unpub. data), modified and extended from Sloan (1987). The star represents beds with 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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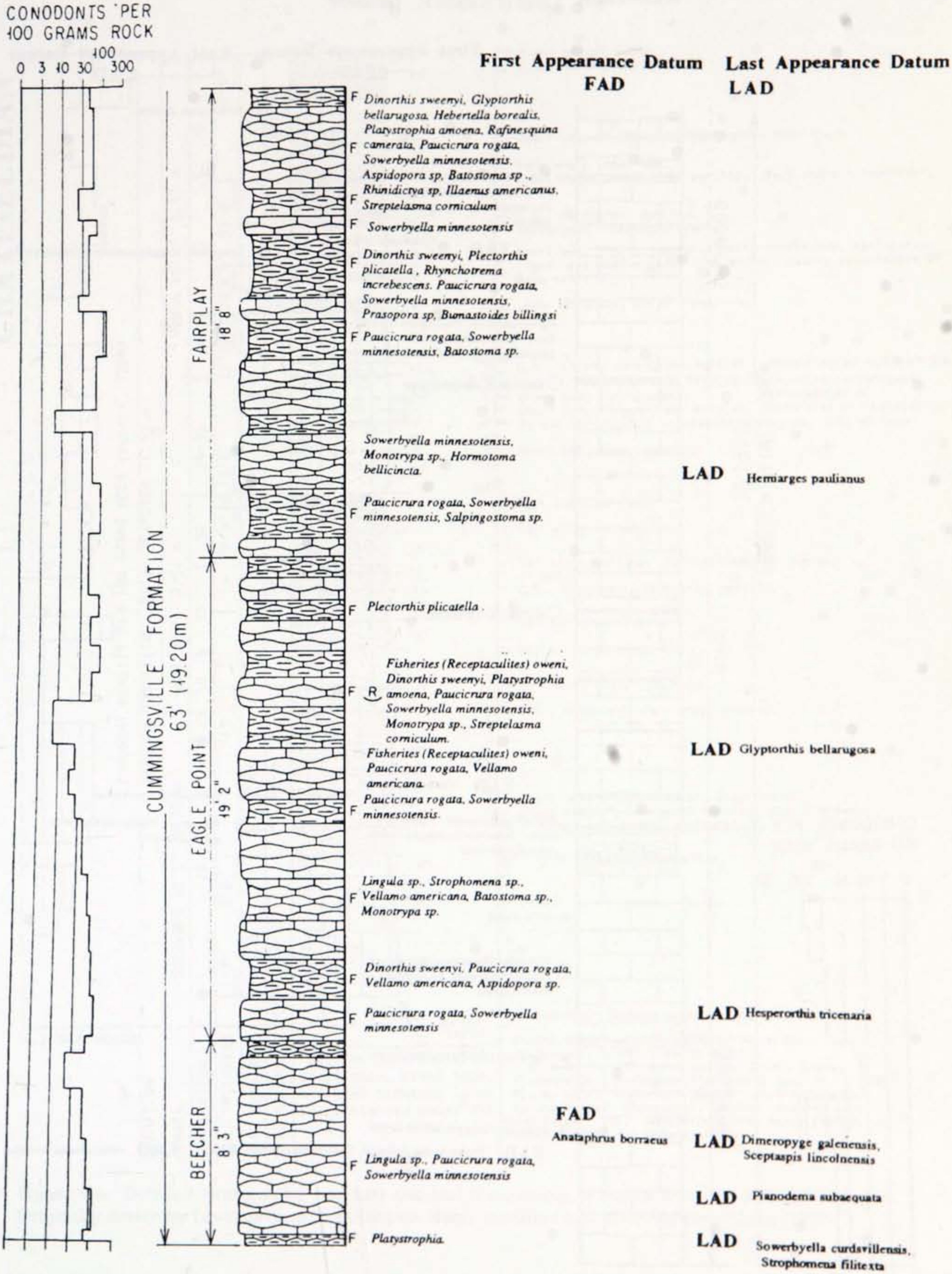

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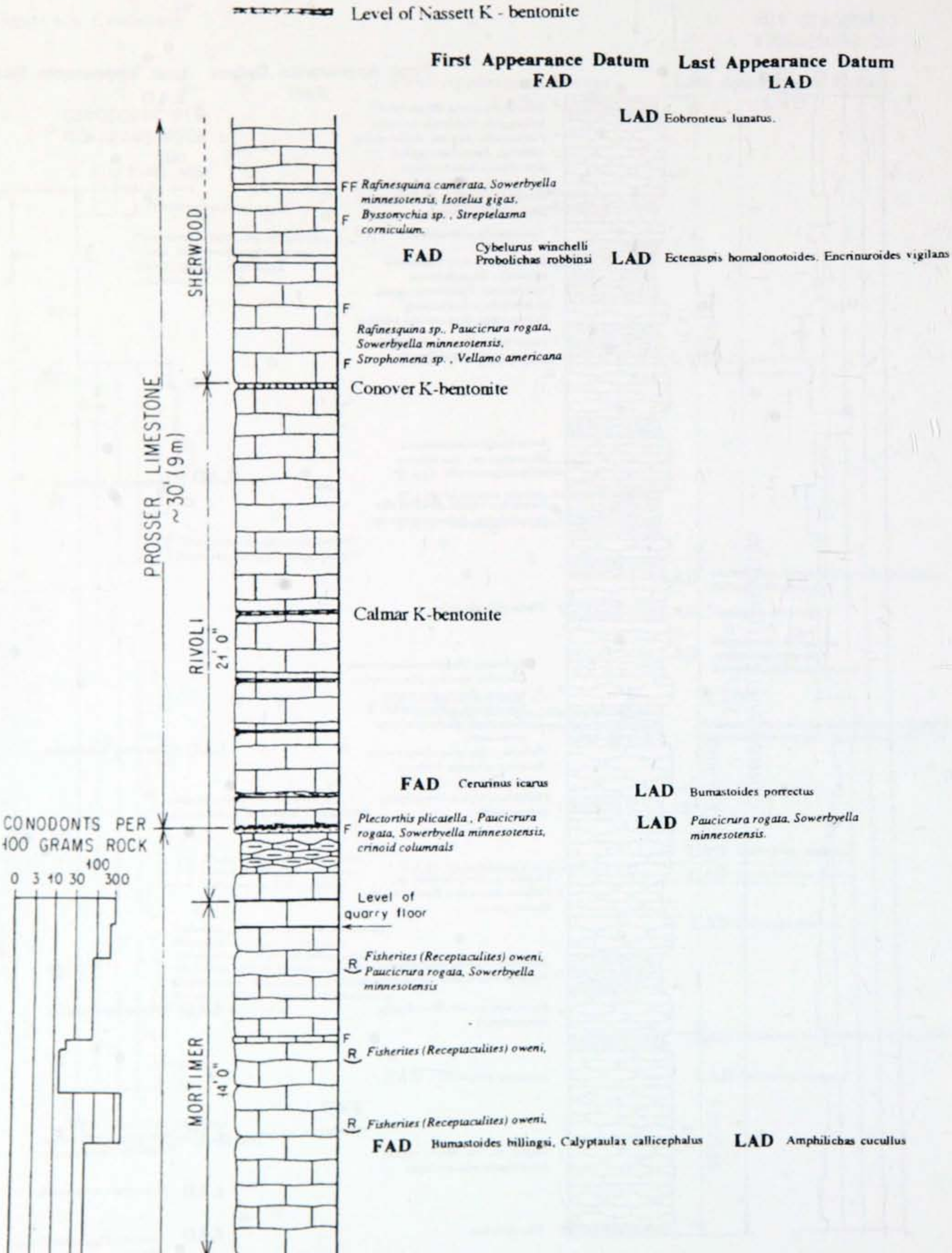


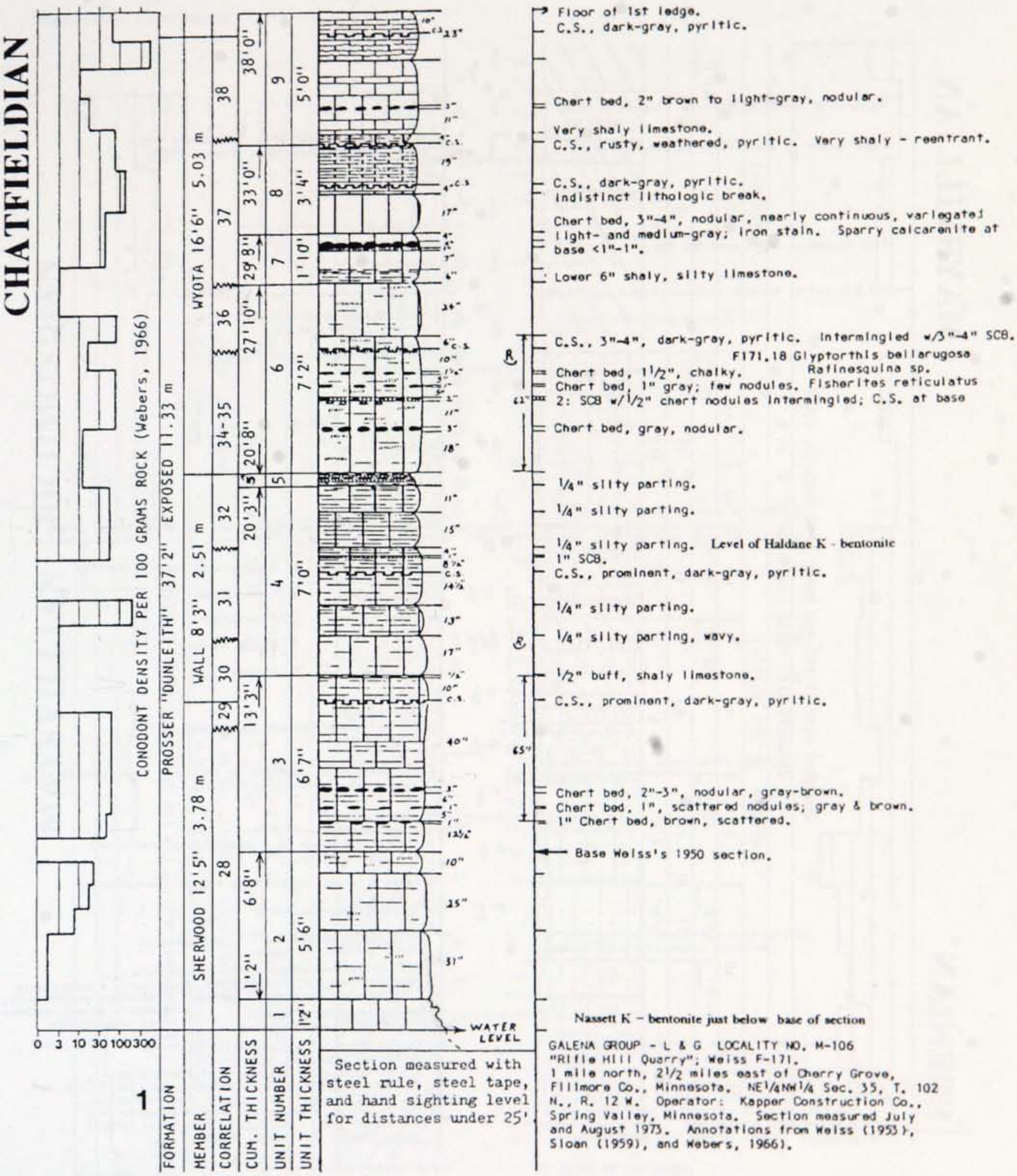
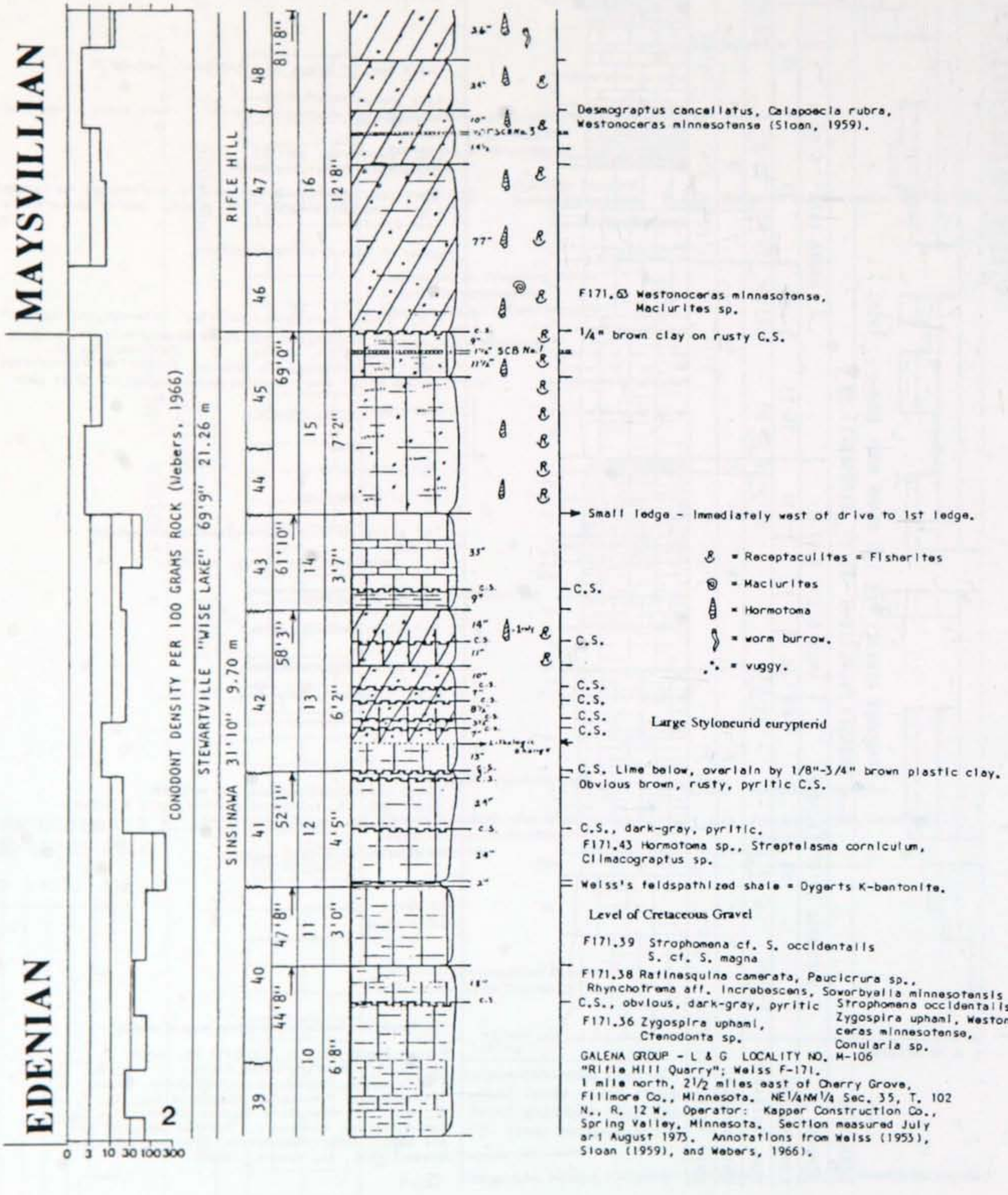


Figure 6.6. Detailed stratigraphy and first and last occurrences of fossils from the Rifle Hill section. Originally drawn by Levorson and Gerk (unpub. data), modified and extended from Sloan (1987).

Figure 6.6. Continued



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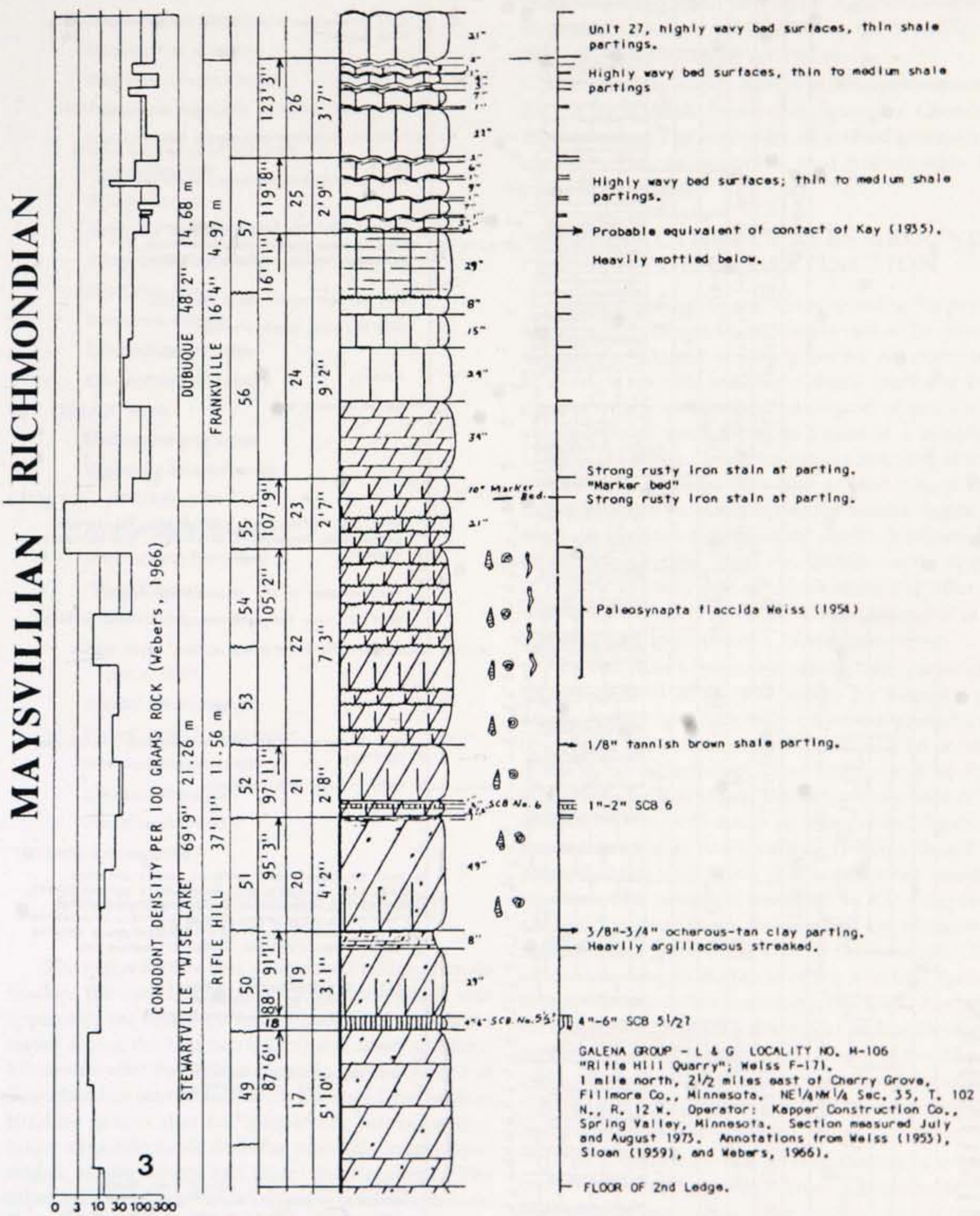

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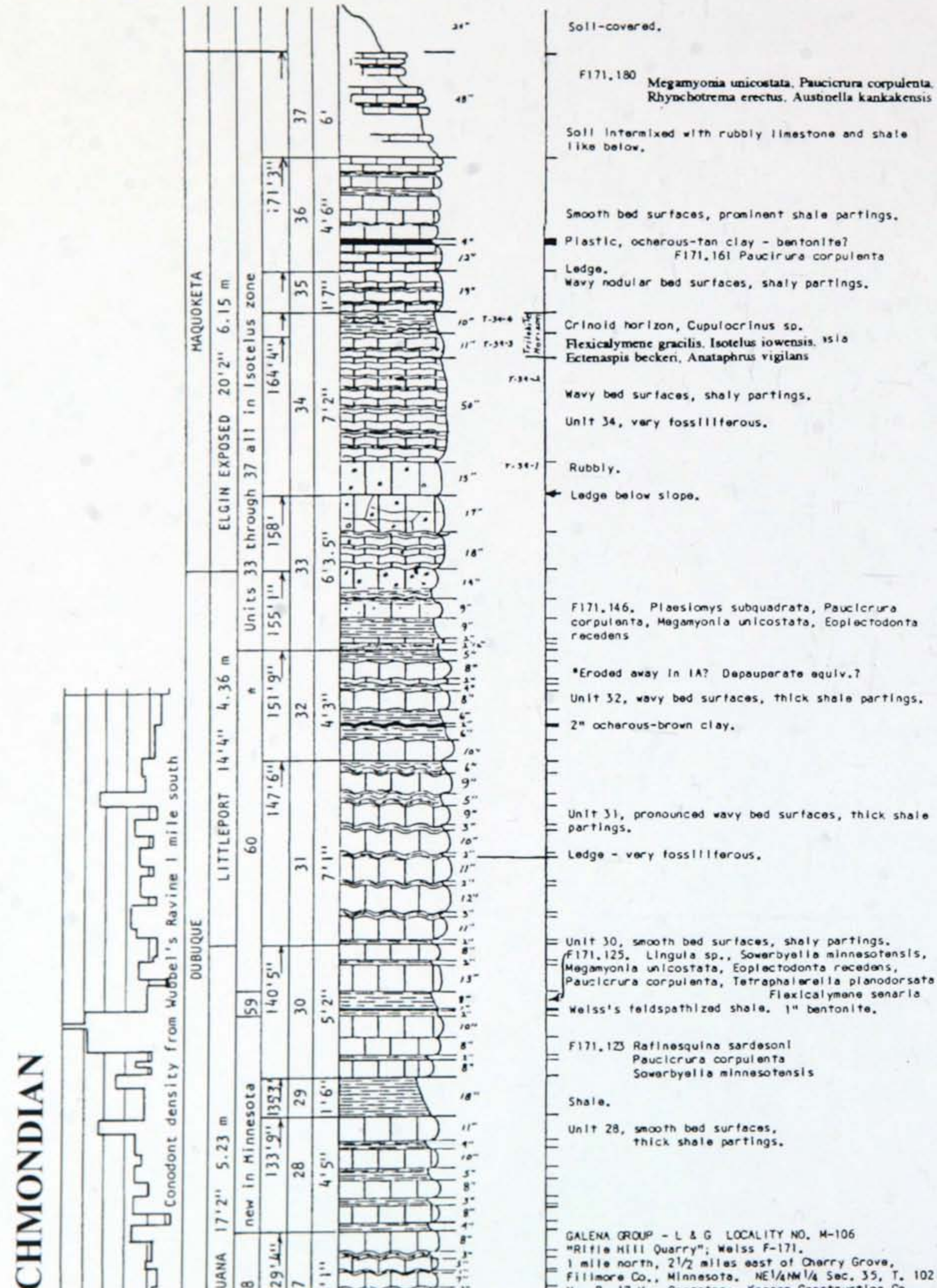


Rostricellula minnesotensis

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

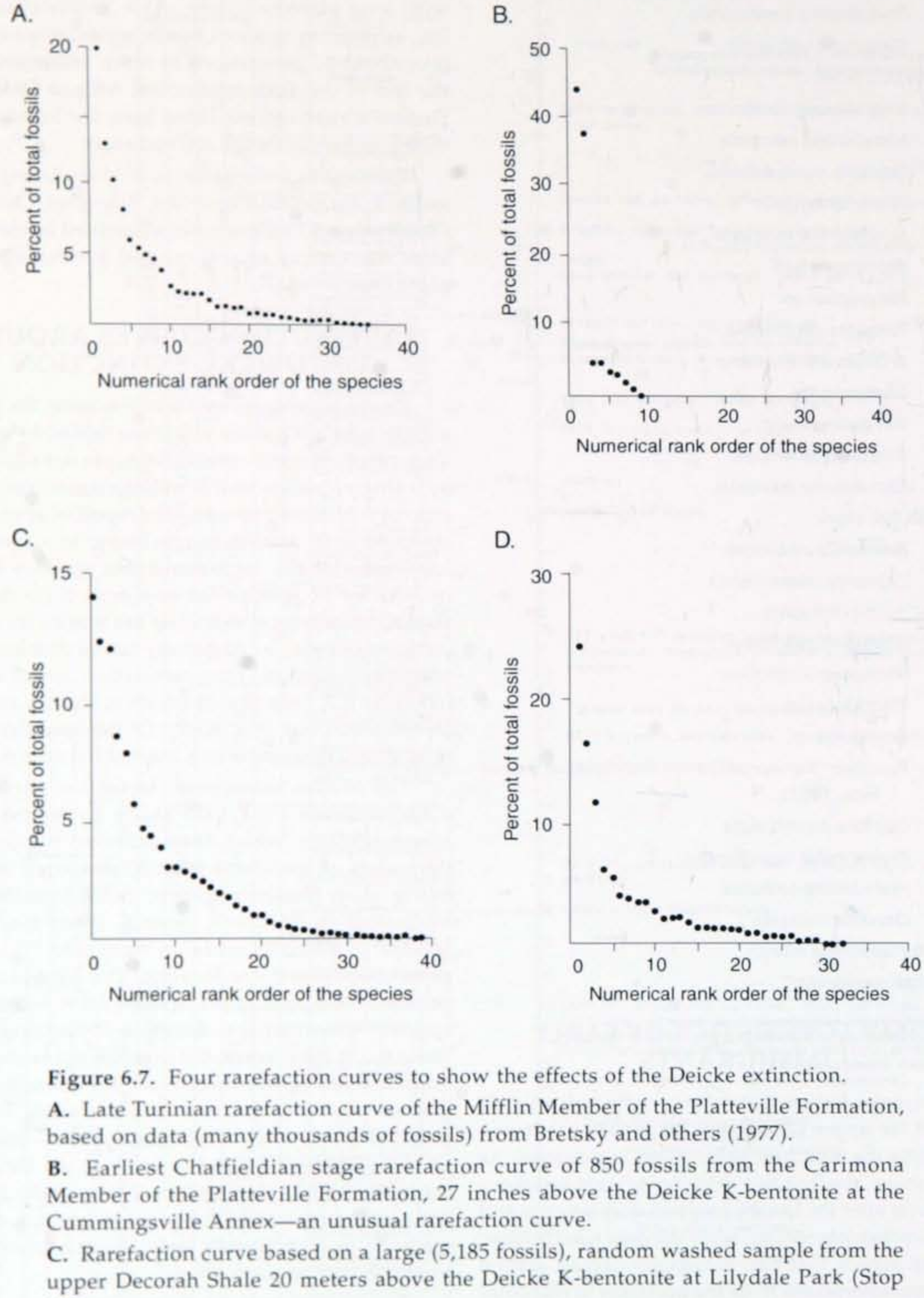


Figure 6.7. Four rarefaction curves to show the effects of the Deicke extinction.

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

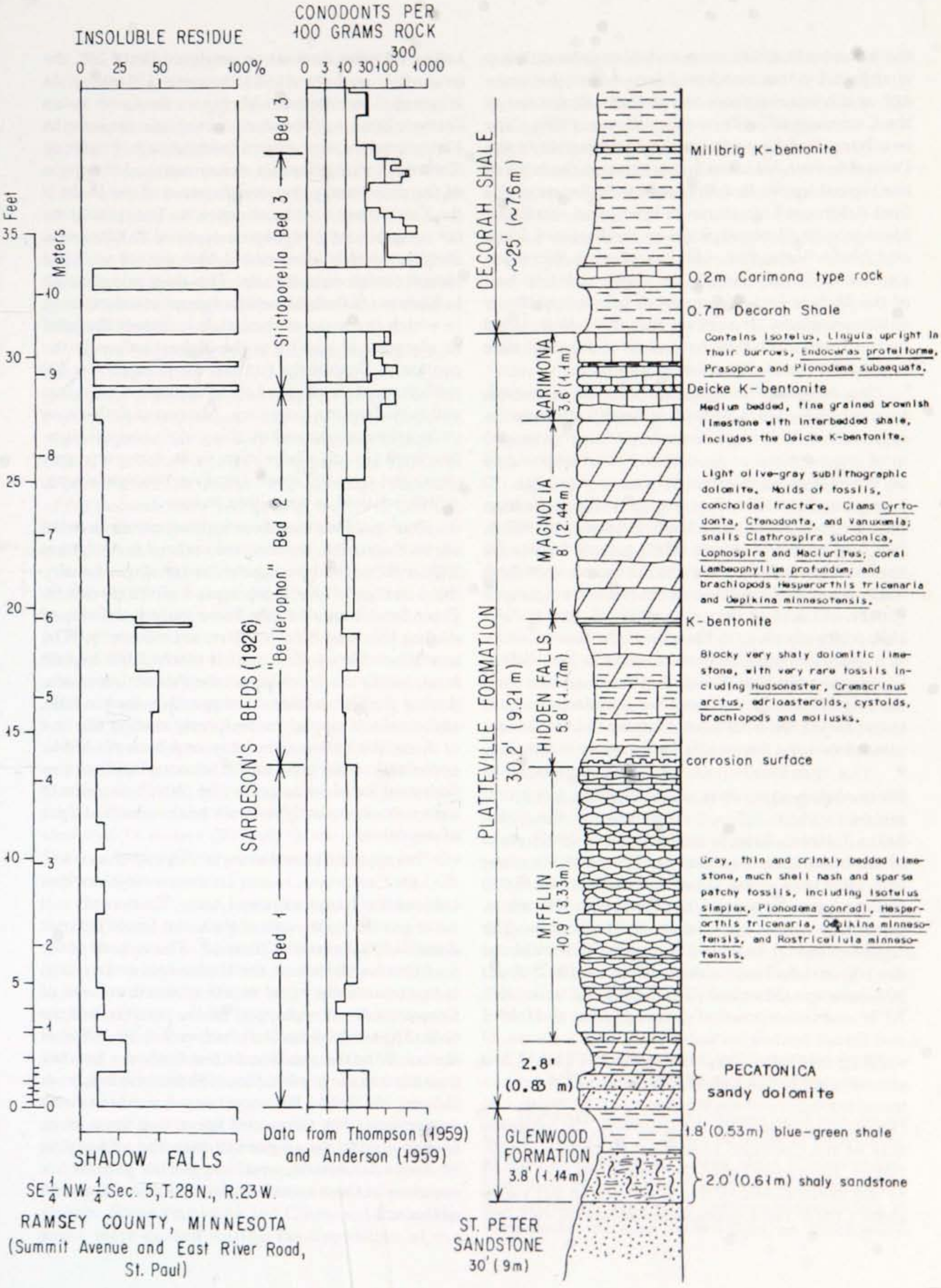
Primary productivity and standing crop of plankton in the sea is usually determined by concentrations of nitrates and phosphates, which are the limiting nutrients in Liebigs' "Law of the Minimum." It is usually measured in grams of wet plankton per square meter. The Diecke K-bentonite provides a spot 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
Bed 3
Stictoporella Bed 3
"Bellerophon" Bed 2
Bed 1
SARDESON'S BEDS (1926)
DECORAH SHALE
~25' (~7.6m)
PLATTEVILLE FORMATION
30.2' (9.21 m)
CARIMONA
3.6' (1.1 m)
MAGNOLIA
8' (2.44 m)
HIDDEN FALLS
5.8' (1.77m)
MIFFLIN
10.9' (3.33m)
2.8' (0.85 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 Plonodema 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, conchoidal 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
Data from Thompson (1959) and Anderson (1959)
SHADOW FALLS
SE 1/4 NW 1/4 Sec. 5, T.28N., R.23W.
RAMSEY COUNTY, MINNESOTA
(Summit Avenue and East River Road, St. Paul)

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 *Sphaerocoxochus* 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 *Sphaerocoxochus* 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.

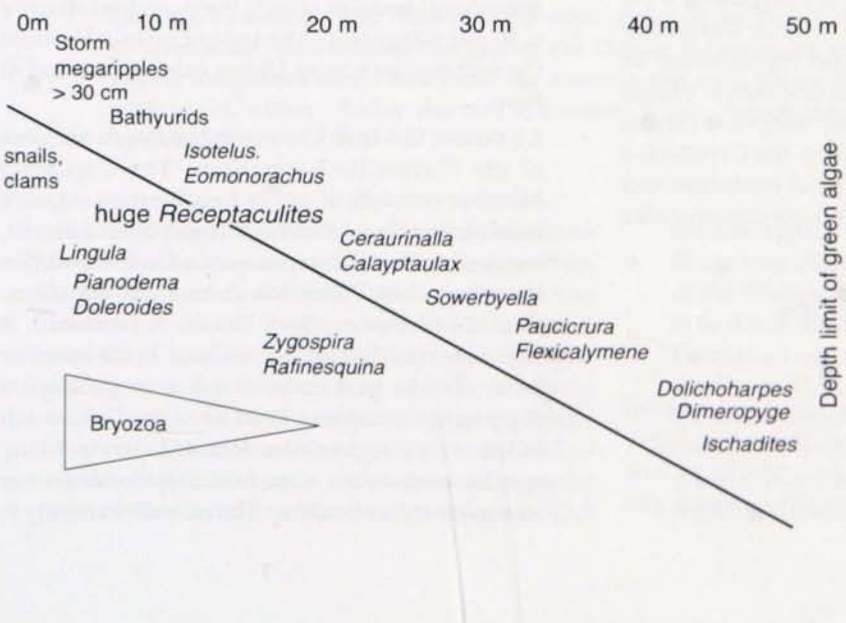


Figure 6.9. An estimate of Chatfieldian paleoecology, with genera plotted at what appears to be their optimum depth. Based on unpublished data collected by Sloan, Schneider, Rice, and DesAutels.

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

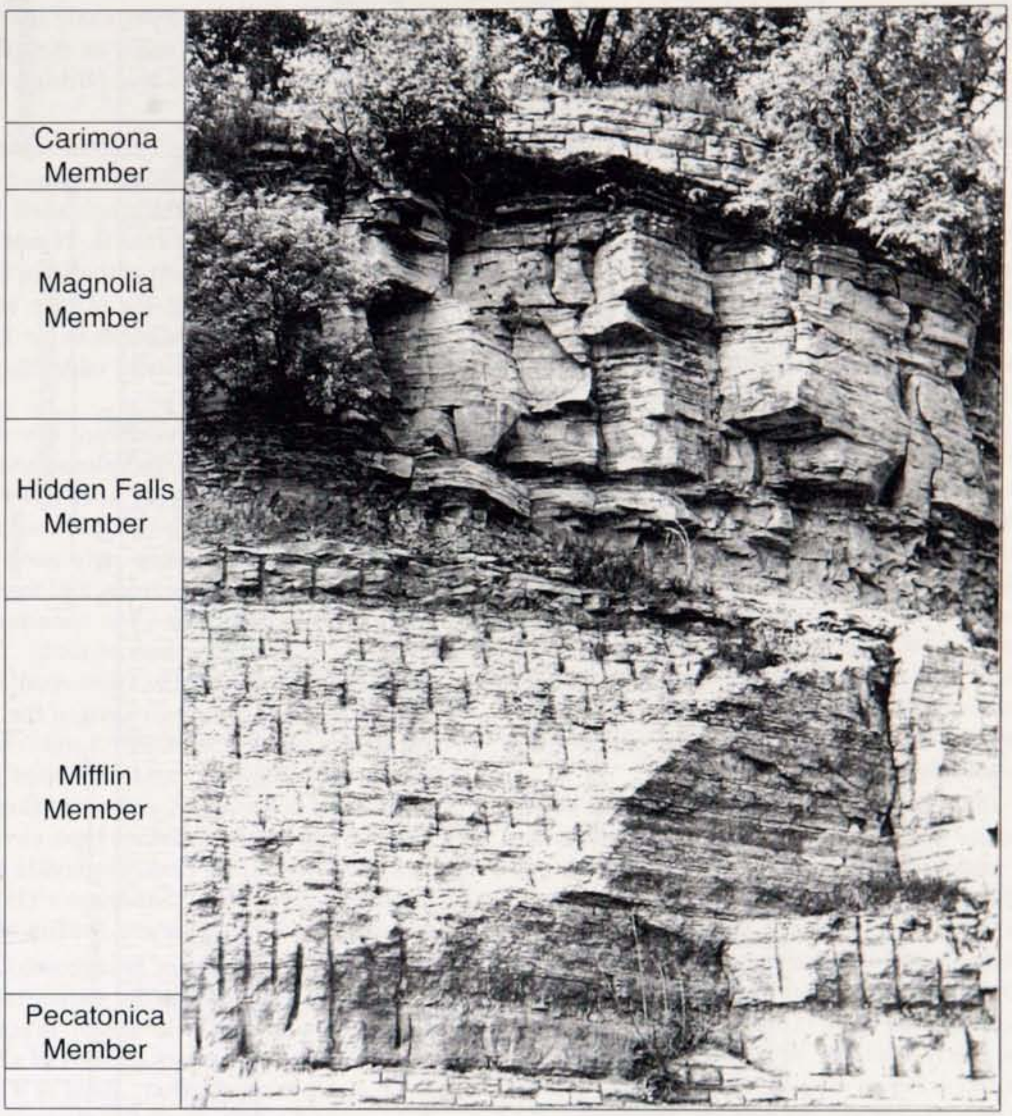


Figure 6.10. Former cliff exposure of the Platteville Formation below Fairview Riverside Hospital as seen in 1958. This is at the exact center of the Twin Cities basin. Minneapolis ground water flows along the top of the Deicke K-bentonite and forms a year round spring that flows down this face. As a 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.

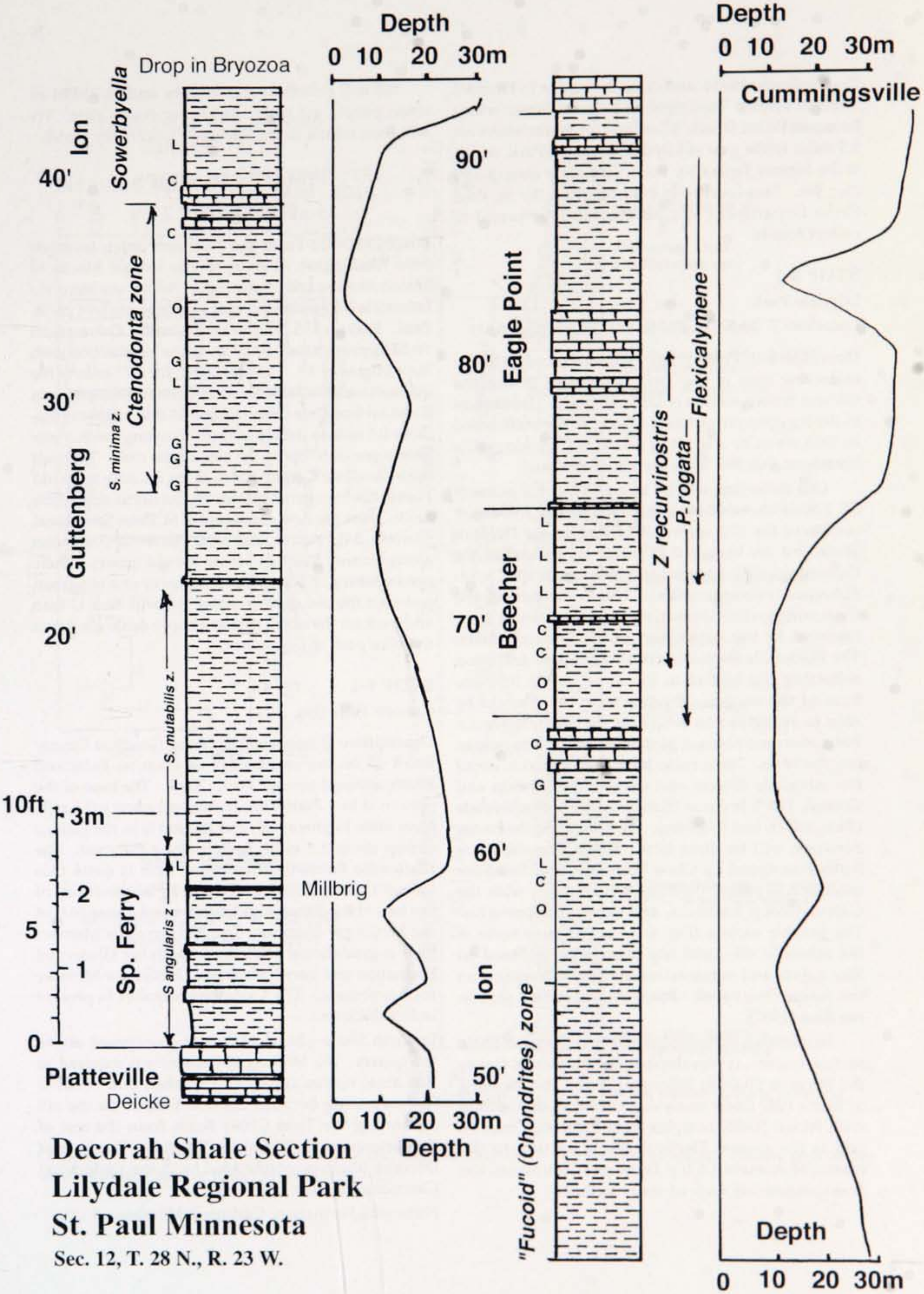
about 100 per 100 grams of rock in the limestone beds. As best as we can determine, based on trilobites and conodonts, this represents the boundary between the Turinian (old Blackriveran) and the Chatfieldian (old Trentonian) stages. Please note the interbedding of Decorah Shale and Carimona Member lithologies 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".

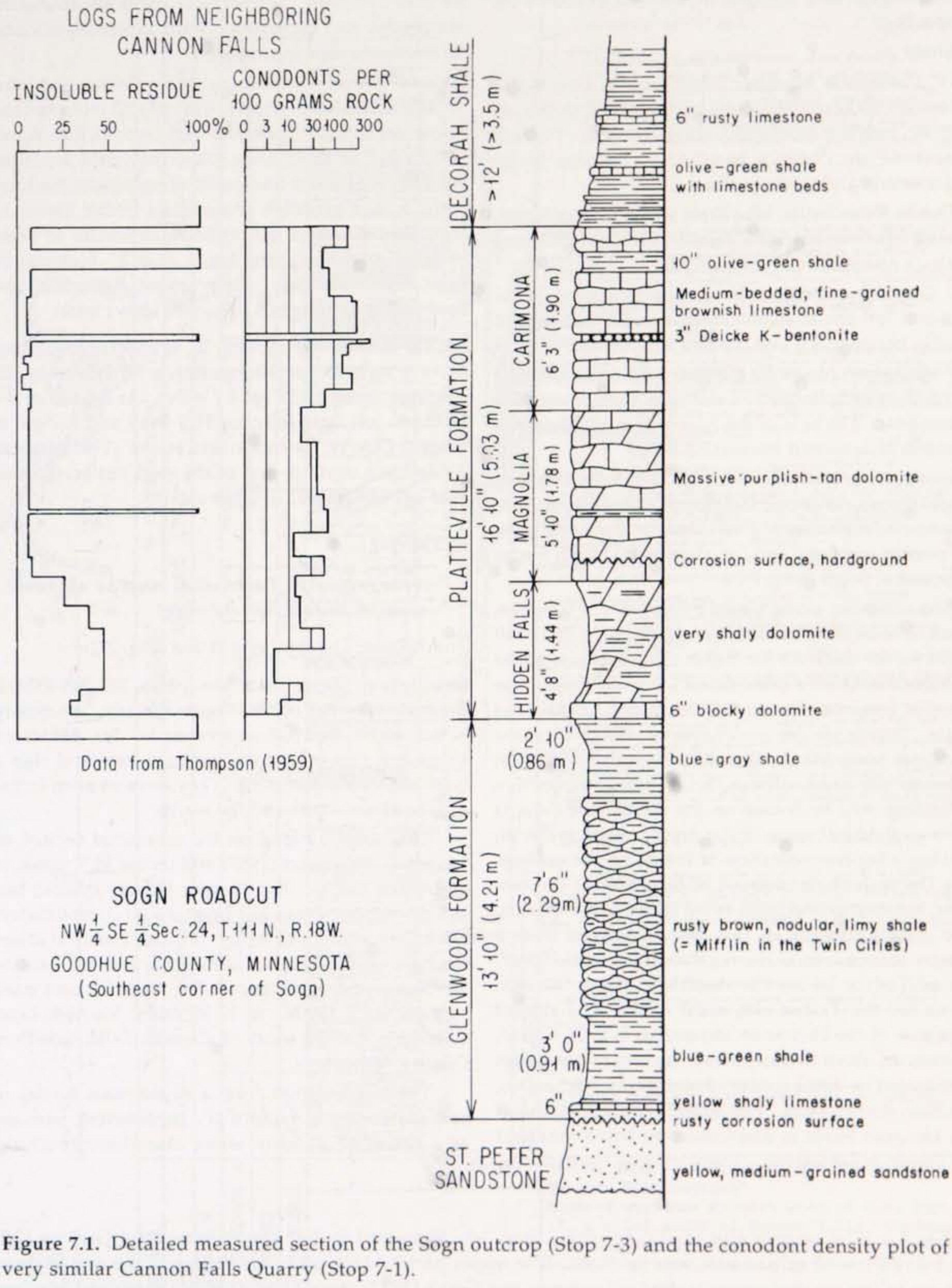


Figure 7.1. Detailed measured section of the Sogn outcrop (Stop 7-3) and the conodont density plot of the very similar Cannon Falls Quarry (Stop 7-1).

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!

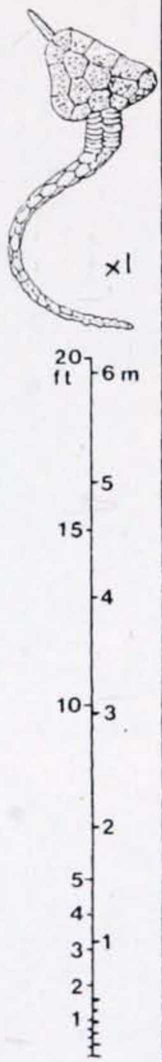
The Wagner Quarry cystoid bed (Fig. 7.2)

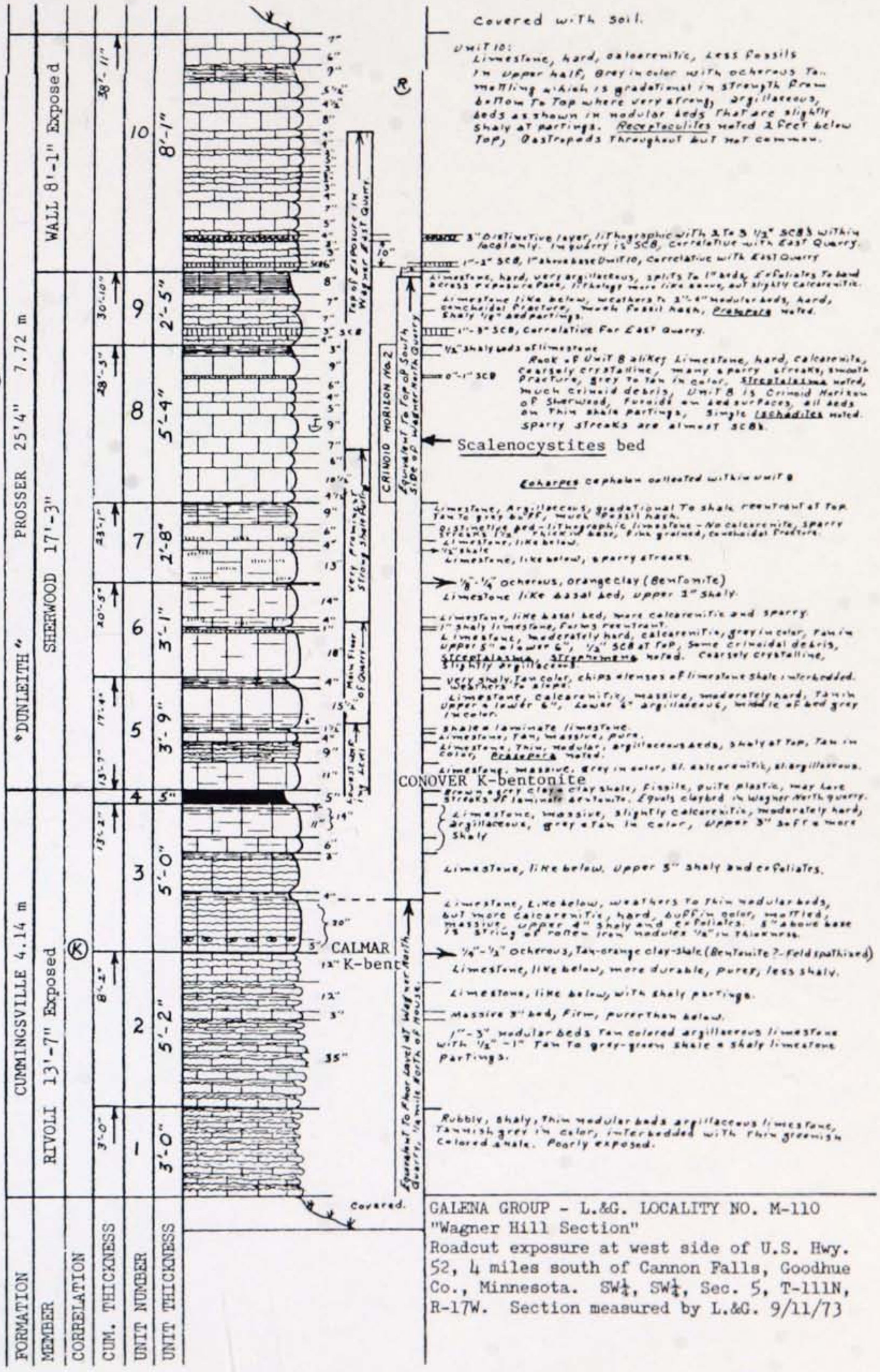
Description: Darren Anderson (phone 507-263-3526) is the current owner of the Wagner Quarry. The quarry is still active, but not on weekends. Mr. Anderson requested that collectors use hard hats and sign a hold harmless agreement. They have an even larger quarry about 4 miles to the south.

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. SW¼, SW¼, 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

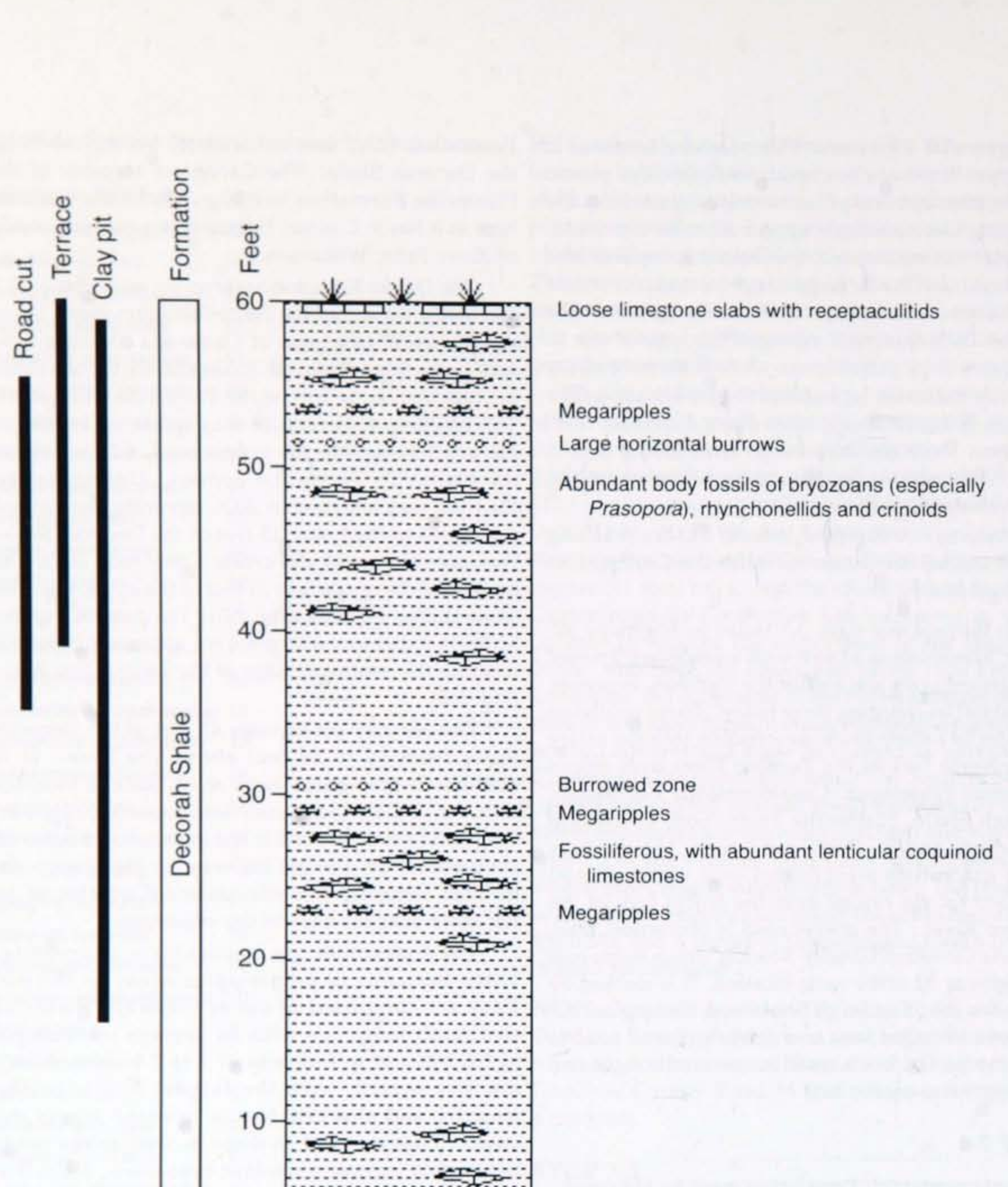


Figure 7.3. Composite stratigraphic section of the Barrclay pit in the Decorah Shale near Zumbrota, T. 110 N., R. 6 W., sec. 21. Exposed intervals are shown at left. Measured by Johnson (unpub. data, 1995) and Middleton (unpub. data, 2004).

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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