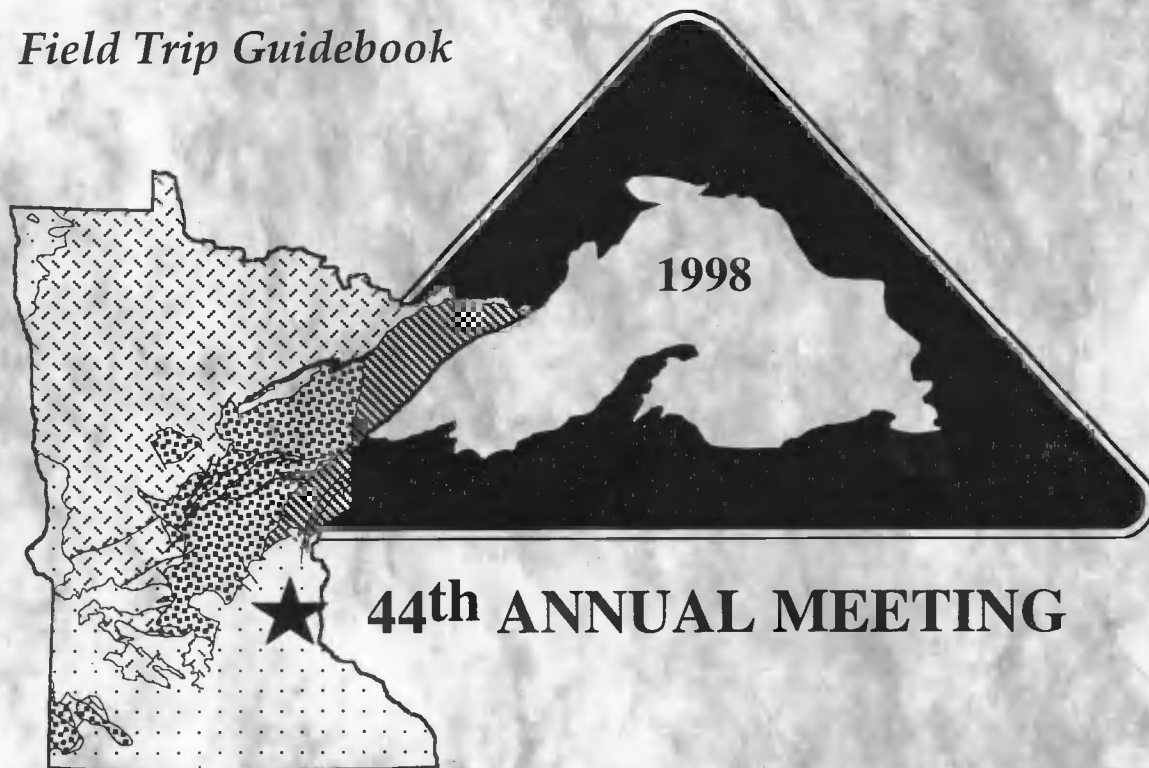


INSTITUTE ON LAKE SUPERIOR GEOLOGY

Field Trip Guidebook



**Minneapolis, Minnesota
May 6-10, 1998**

Organized by

MINNESOTA GEOLOGICAL SURVEY
UNIVERSITY OF MINNESOTA

PROCEEDINGS VOLUME 44
Part 2—Field Trip Guidebook

44th Annual Meeting INSTITUTE ON LAKE SUPERIOR GEOLOGY

Volume 44 contains the following parts:

Part 1: Program and Abstracts

Part 2: Field Trip Guidebook

- 1—Early Proterozoic intrusive rocks of east-central Minnesota
- 2—Geology of the southeastern portion of the Midcontinent Rift System, eastern Minnesota and western Wisconsin
- 3—Glacial exotica of the Twin Cities area
- 4—Stratigraphy and hydrogeology of Paleozoic rocks of southeastern Minnesota
- 5—Minnesota River Valley and vicinity, southwestern Minnesota

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44th Annual Meeting
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Sponsored by:
Minnesota Geological Survey
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Field Trip Guidebook

Volume 44

Part 2—Field Trip Guidebook

EDITORS:
Terrence J. Boerboom and Barbara A. Lusardi
Minnesota Geological Survey

PROCEEDINGS VOLUME 44, PART 2—FIELD TRIPS

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FIELD TRIP 1

EARLY PROTEROZOIC INTRUSIVE ROCKS OF THE ST. CLOUD DISTRICT, EAST-CENTRAL MINNESOTA

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FIELD TRIP #1

EARLY PROTEROZOIC INTRUSIVE ROCKS OF EAST-CENTRAL MINNESOTA

Leaders: Terry Boerboom, Mark Jirsa, and Daniel Holm

INTRODUCTION

The Early Proterozoic (Paleoproterozoic) granitoid rocks of the St. Cloud District in east-central Minnesota (Fig. 1) have long been a source of both building materials and geologic controversy. Quarrying for dimension stone began in the St. Cloud District in the late 1800's and peaked at about 1930-1940, when there were some 18 different quarry companies operating in the district (Thiel and Dutton, 1935). Currently, Cold Spring Granite is the only dimension stone company operating, with 5 active quarries here, and 4 others in different parts of the state. Meridian Aggregates Company also operates a large crushed-stone operation in the St. Cloud District. The St. Cloud area contains abundant active and inactive quarries, as well as natural outcrops. Despite these exposures, many of the critical relationships between various intrusive units are concealed—leaving details of the magmatic history unknown and speculative. Until recently, geologic maps of the district were reconnaissance-style (Morey and others, 1981), and little geochronologic work had been conducted. This field trip highlights the results of more current and thorough mapping endeavors that utilized quarry and natural exposures, drill core, and geophysical interpretation. These include the Stearns County Geologic Atlas (Boerboom and others, 1995; Boerboom, 1996a), and mapping north and east of St. Cloud (Jirsa and others, 1995a; 1995b; Jirsa and Chandler, 1997). In addition, we review both pre-existing and newly acquired geochronologic data (e.g., Holm 1997, 1998) in light of the field, petrographic, and geophysical observations.

Intrusions of the St. Cloud District were emplaced relatively late in development of the Penokean orogen—a belt of Early Proterozoic volcanic, sedimentary, and plutonic rocks affected by the Penokean orogeny, which culminated at approximately 1870 Ma (Van Schmus and Bickford, 1981). Rocks of the orogen are divided from north to south (Fig. 1) into a variably deformed foreland basin; a fold and thrust belt; and what has been inferred to be an accreted island-arc terrane, broadly equivalent to the Wisconsin magmatic terranes (Jirsa and others, 1995b, Boerboom, 1996b). All of these terranes were deposited upon or accreted against the southern margin of the Archean Superior Province. Using the terminology of Southwick and others (1988), the fold and thrust belt is further subdivided into distinct segments on the basis of lithologic and structural attributes: 1) an external zone comprised of the Animikie Group and related outliers; 2) a medial zone characterized by fault slices of interlayered volcanic and clastic strata including the Cuyuna north and south (iron) ranges; and 3) an internal zone, containing fault-bounded blocks of various metamorphic and plutonic rocks.

The Early Proterozoic granitoid rocks of the St. Cloud District occur within the internal zone of the western Penokean orogen, where they form a physically continuous mass of late- and post-orogenic intrusions. They represent some of the most deeply emplaced magmas of the

orogenic belt. Regionally, the Early Proterozoic supracrustal country rocks—including the Little Falls Formation to the north and west, and unnamed volcanic and clastic strata to the east and southeast—were metamorphosed at depths of 22-25 km during the Penokean orogeny (Holm and Selverstone, 1990). Reconstruction of P-T-t histories indicate that the rocks currently exposed remained at those depths until well after the orogeny, then were rapidly exhumed at about 1770 Ma during emplacement of post-tectonic magmas (Holm and others, 1998). The significant depth of emplacement and the great volume of igneous rock provide an excellent opportunity to investigate midcrustal magmatic processes that occurred immediately after the orogenic belt was constructed, and later as the orogen collapsed.

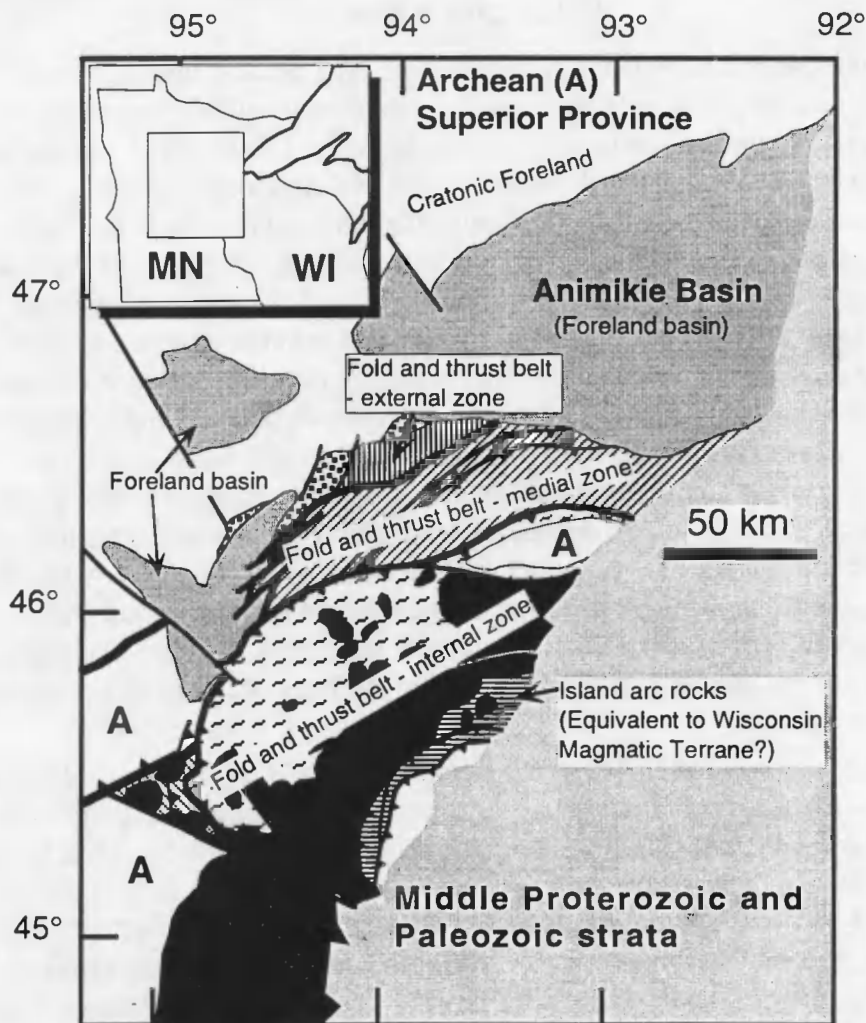


Figure 1. Simplified geologic map of the Penokean orogen, showing major tectonic elements. Late- to post-tectonic Penokean intrusions shown in black.

GEOLOGY OF INTRUSIVE ROCKS

Intrusions of the St. Cloud District (Fig. 2) are comprised of several discrete units that include the Reformatory granodiorite, the charnockitic Richmond granite, the Rockville Granite, the St. Cloud Granite, the Foley batholith, and several types of granitic and diabasic dikes. Only scattered xenoliths of metamorphic country rocks are recognized within the intrusions,

but numerous mafic enclaves interpreted to be comagmatic with the enclosing granitoid rocks are present in some units. The intrusions can be divided on the basis of composition and relative temporal setting into three categories: 1) gray to pinkish-gray to greenish gray granitic, granodioritic, and dioritic rocks—Reformatory granodiorite, Richmond granite, and Rockville Granite; 2) red and pink granitic intrusions—St. Cloud Granite, Foley batholith, and associated aplite; and 3) diabasic and porphyritic microgranite dikes. Color is not always an adequate distinguishing characteristic between various intrusions because components of the “gray” intrusions can be quite pink. Other attributes that distinguish “gray” from “red” varieties include the observation that “gray” intrusions contain mafic enclaves, and have a better developed and more pervasive magmatic foliation than do the “red” ones. These differing attributes have important implications about the magmatic history that will be discussed below. A description of individual plutonic units follows.

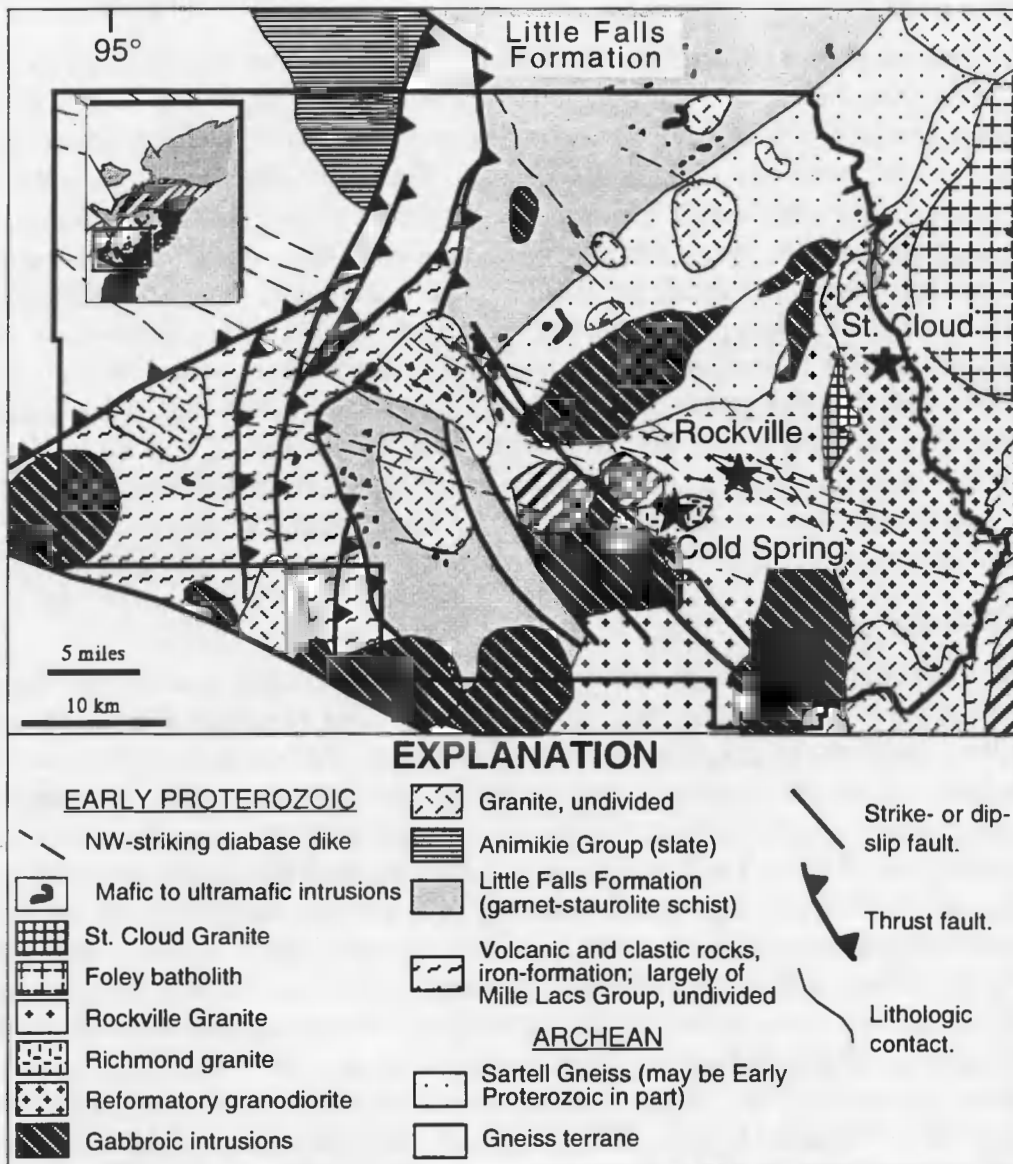


Figure 2. An extremely busy and largely illegible regional geologic map of the area around Stearns County (outlined). Inset map references location within Penokean orogen (Fig. 1).

Reformatory granodiorite

The Reformatory granodiorite (Reformatory Granite of Morey, 1978) is presently extracted from only one quarry (trade name Charcoal Granite), just south of the city of St. Cloud. However, there are dozens of inactive pits from which this rock has been quarried in the past, including those that supplied stone to the unit's namesake—the State Reformatory! The rock is medium- to dark-gray, medium- to coarse-grained, weakly porphyritic biotite-hornblende granodiorite characterized by small tabular plagioclase phenocrysts (oligoclase to andesine; Johnson, 1978) in a groundmass of plagioclase, microcline, hornblende, biotite, and accessory pyroxene, sphene, and apatite. A weak, generally subhorizontal trachytoid fabric in the granodiorite can be observed in many places. Small ovoid mafic enclaves are scattered throughout the intrusion.

Richmond granite

The charnockitic Richmond granite is well exposed between the towns of Richmond and Cold Spring, but has never been significantly exploited as a dimension stone. Although generally dark greenish-pink, it locally retains a unique deep green color where the rock has not been subjected to oxidation along tight brittle fractures. The granite displays a prominent rapakivi texture defined by plagioclase-mantled, 2-6 cm long K-feldspar megacrysts, as well as a consistent east-northeast magmatic trachytoid fabric defined by aligned phenocrysts. Hypersthene, red-brown biotite, and green hornblende are the dominant mafic minerals; apatite, ilmenite, and zircon form abundant accessory phases concentrated with the late-formed mafic minerals. Quartz typically occurs as more-or-less equant, but anhedral-interstitial, monocrystalline clots as large as 1 cm. Although elongate grains of quartz occur in localized, meter-scale shear bands, the bulk of the intrusion contains a well-developed magmatic foliation and no evidence of metamorphic recrystallization. Scattered outcrops and geophysical maps (Fig. 3) indicate that the Richmond granite may be one component of a large composite charnockitic intrusion composed largely of gabbroic and noritic intrusive rocks.

Rockville Granite

The Rockville Granite is presently taken from three separate quarries and marketed under the trade names of Diamond Pink, Rockville Beige, and Rockville White. It is a very coarse-grained, pinkish- to grayish-white rapakivi granite characterized by zoned 1-3 cm microcline phenocrysts and smaller plagioclase (oligoclase; Johnson, 1978) phenocrysts in a groundmass of coarse quartz, feldspar, biotite, and hornblende, with accessory zircon, apatite, sphene, epidote, and allanite. The Rockville locally shows a weak trachytoid fabric, defined by aligned feldspar phenocrysts, that strikes generally N-S and dips variably to the west. This granite contains scattered enclaves that vary from coarse-grained black diorite, to fine-grained, gray quartz monzonite; both having abundant hornblende and biotite. The dioritic enclaves commonly contain microcline phenocrysts or xenocrysts; whereas the monzonitic ones contain scattered plagioclase phenocrysts in a pristine igneous-textured, trachytoid matrix of microcline, quartz, biotite, and hornblende. Dikes 5-30 cm wide of rock like the coarse-grained dioritic phase locally cut the Rockville Granite. These two conflicting occurrences of the coarse-grained diorite (as both enclaves and dikes), the presence of microcline 'phenocrysts' in the coarse-grained enclaves, the pristine trachytoid igneous texture of the finer grained enclaves, and the rapakivi texture, all imply an origin by the commingling of mafic and felsic magmas during magma ascent (see DISCUSSION).

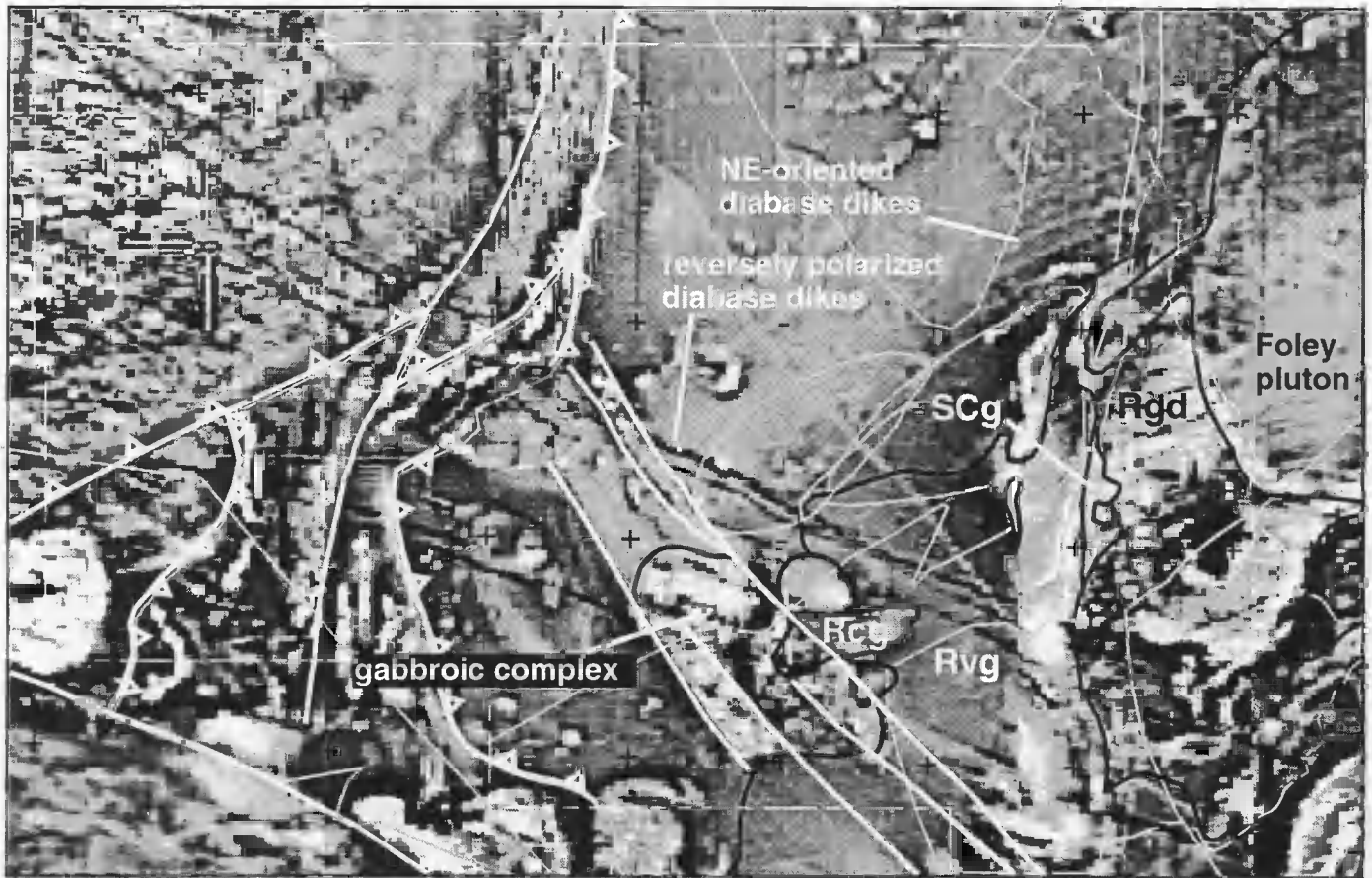


Figure 3. First vertical derivative, reduced to pole aeromagnetic anomaly map of Stearns County (white dashed line, also see Figure 2) and the fringing area. Outlines only shown for rock units discussed in text.

St. Cloud Granite

The St. Cloud Granite occurs as a zone of variably wide dikes and sills that permeated into the Reformatory granodiorite, resulting in a complex geometry not easily depicted at small map scales. At larger scale, dikes of St. Cloud Granite are irregular in shape, and contacts most commonly trend to the north and northeast. Small dikes and irregular pods of pink aplite that are associated with the St. Cloud Granite also cut Reformatory granodiorite.

The St. Cloud Granite has been quarried extensively in the past, but no longer. It is a coarse-grained, deep-colored, grayish-pink to red, hornblende granite having a weak porphyritic texture defined by pink microcline phenocrysts only slightly larger than the matrix. The main minerals are microcline, plagioclase (albite to oligoclase; Johnson, 1978), quartz, hornblende, and biotite, with accessory Fe-Ti oxides, apatite, zircon, sphene, and rare epidote. Using terminology of Streckeisen and Le Maitre (1979), the St. Cloud varies in modal mineralogy from alkali-feldspar granite to syenogranite (Fig. 4). A vague trachytoid fabric defined by aligned feldspar and hornblende crystals is common and is parallel to the margins of the dikes and sills in which the St. Cloud Granite occurs. Narrow dikes of pink aplitic granite locally cut the St. Cloud and Reformatory, and are inferred to be late differentiates of the St. Cloud on the basis of lithologic similarity. Coarse pegmatite is exceedingly (and conspicuously) rare. The St. Cloud commonly contains angular xenoliths of Reformatory granodiorite, along with rare xenoliths of metamorphic country rock.

Foley batholith

The Foley batholith is a red to salmon-colored, coarse-grained, hornblende-biotite granite that is cut by minor amounts of granitic porphyry and aplite—similar in nearly all respects to the St. Cloud Granite. It forms a large (60 x 20 km), ovoid gravity low that extends from the east side of St. Cloud, northeastward to near Mille Lacs Lake (Jirsa and Chandler, 1997). The batholith is moderately well exposed along its western terminus, where its intrusive contact against Reformatory granodiorite can be seen, and in several places east of the map area shown on Fig. 2. Small quarries in the western part of the unit operated during the period 1896-1909. The Foley is massive to vaguely foliated, having a poorly developed and localized trachytoid fabric. Although it will not be visited during this field trip, the Foley is described here because it is a large and distinctive unit that appears to be magmatically related (parental?) to the St. Cloud Granite.

Porphyritic microgranite (AKA QFP) dikes

Thin northeast-trending dikes of distinctively textured porphyritic microgranite cut the Reformatory granodiorite, Rockville Granite, St. Cloud Granite, and Foley batholith. None of these dikes were noted in the Richmond granite, but they would logically post-date that unit. Most of these dikes are 2-4 m wide, but one is at least 30 m wide, and another may be as much as 100 m wide, although correlation of the latter with this dike set is tenuous. The margins of these dikes are chilled and locally flow-banded, texturally similar to volcanic rhyolite. The porphyries are characterized by 3-10 percent quartz phenocrysts as large as 4 mm across that have a primary euhedral shape modified by resorption, and 3-10 percent microperthite or sanidine phenocrysts as large as 10 mm. The fine-grained groundmass varies in texture from anhedral-granular to spherulitic. In addition to the main quartz and feldspar, the rock contains minor amounts of fine-grained biotite, hornblende, actinolite, and chlorite, and accessory fluorite, apatite, Fe-Ti oxides, epidote, sphene, zircon, and allanite.

Diabase dikes

There are two main groups of diabase dikes recognized in the area: a northeast-trending set, and a northwest-trending set. Both sets are steeply dipping and are interpreted to be the youngest Precambrian rocks in the St. Cloud area, as they cut nearly all other rock types. Northeast-trending diabase dikes are typically non-magnetic and lack geophysical expression, but are exposed in several outcrops and old quarries throughout the area. They are most commonly 1-3 m thick, but some are as wide as 8 m, and dike margins are sharply chilled against the neighboring granitic rocks. These dikes contain fresh plagioclase, but low-grade metamorphism or deuteric alteration (?) has altered the primary mafic minerals to actinolite, chlorite, and dusty oxides.

Northwest-trending diabase dikes are rarely exposed, but are inferred from geophysical data to be as much as 100 m wide. They are geophysically traceable by their distinctive, generally reversely polarized, aeromagnetic signature (Fig. 3). One exposed northwest dike (Fig. F6 of stop descriptions) has sharply chilled margins, and compared to the northeast-trending set is strongly magnetic (pyrrhotitic) and is less altered/metamorphosed, having substantially more fresh clinopyroxene. This dike is normally polarized and only 6 meters thick—attributes that contrast somewhat with the geophysically-defined northwest swarm, making correlation

uncertain. Another small outcrop of apatitic ferrodiorite, in northwestern Stearns County, lies over one of the reverse-dike anomalies and is probably part of this swarm, but no magnetic polarity tests have been performed.

GEOCHEMISTRY

A standard alkali-silica chemical classification plot characterizes the granitoid rocks as subalkaline to alkaline composition. As shown in Figure 4, both the St. Cloud and Rockville Granites plot as subalkalic using the dividing line of Irvine and Baragar (1971), but as alkalic using the dividing line of Miyashiro (1978). This alkalic tendency reflects the perthite-dominated mineralogy of the St. Cloud Granite. In addition, all the granitoid rocks in Stearns County are peraluminous, calc-alkaline, and I-type, consistent with the abundance of biotite and hornblende and lack of muscovite.

DISCUSSION

Temporal Relationships—Field Evidence

Based on field and petrographic observations, geophysical maps and measurements, and limited geochronologic data, the following sequence—from oldest to youngest—summarizes the emplacement history of igneous rocks in the St. Cloud District:

- 1) Reformatory granodiorite, and Rockville and Richmond granites—emplaced at about 1812 Ma;
- 2) St. Cloud Granite and associated aplite, and Foley batholith—emplaced at approximately 1770 Ma;
- 3) northeast-trending dikes of porphyritic microgranite and diabase; and
- 4) northwest-trending diabase dikes.

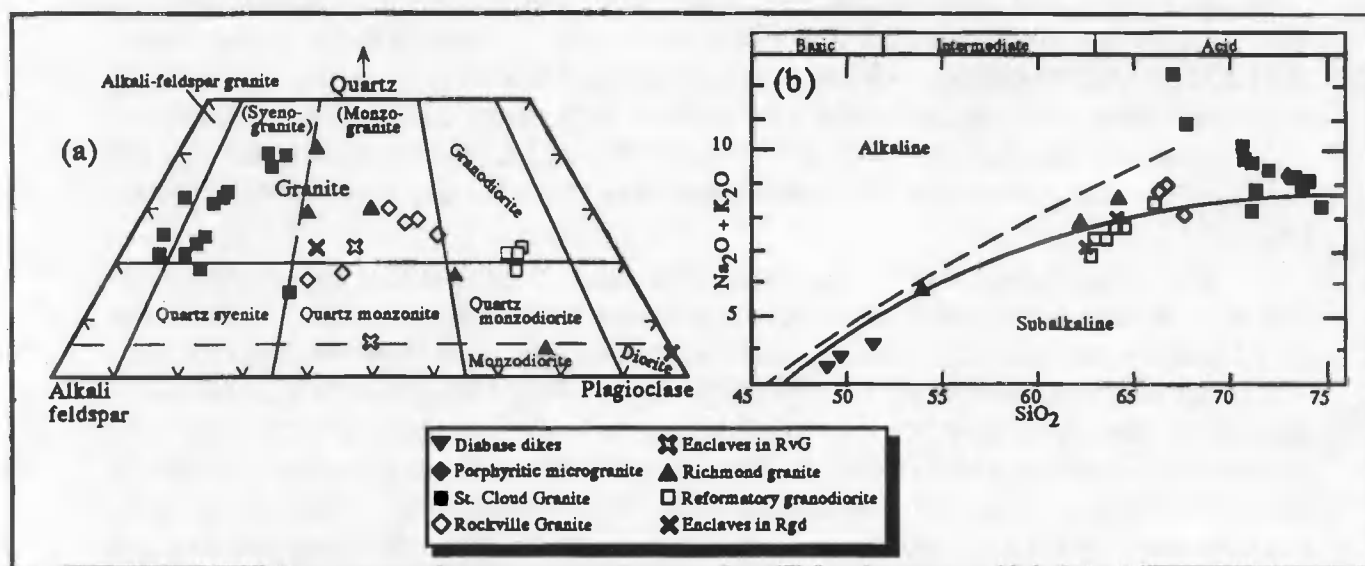


Figure 4. Modal and chemical variation diagrams of intrusive rocks from the St. Cloud area. (a) selected modal analyses compiled from Dittman (1971), Morey (1972), Johnson (1978), Skillman (1946), and Boerboom (in prep.); (b) alkali-silica classification, with dividing lines of Miyashiro (1978, solid line) and Irvine and Baragar (1971, dashed line).

The St. Cloud Granite intrudes the Reformatory granodiorite, as established above and by earlier workers (Morey, 1978 and references therein). The intrusive relationships between the Rockville Granite and other granitoid rocks cannot be seen in outcrop. Although geochronological data imply that the Rockville and Reformatory are generally synchronous (as discussed below), some outcrops of Rockville Granite contain inclusions of rock equivocally similar to the Reformatory granodiorite. Geophysical maps imply that the non-magnetic Rockville Granite forms an eastward-thinning wedge that overlies the more magnetic Reformatory granodiorite (Fig. 3); however, the observations contribute little to their relative emplacement history.

Porphyritic microgranite dikes are inferred to be either cut by or synchronous with northeast-trending diabase dikes. It is interesting to note that the microgranite dikes commonly have a similar northeast-trending diabase dike near them. At one locality a thin diabase dike is chilled against the Rockville Granite on one side, but retains felty-textured plagioclase against the contact with a porphyritic microgranite dike on the other, implying that the microgranite was still hot when diabase was emplaced. The diabase contains abundant xenocrysts of porphyry-derived quartz and feldspar grains adjacent to this contact. These textures indicate that the diabase may be comagmatic with the microgranite, or that diabase post-dates microgranite only slightly. A northwest-trending diabase dike exposed in a quarry in Waite Park (Fig. F6 of stop descriptions) unequivocally cuts and is chilled against a small northeast-trending diabase dike. If this northwest-trending dike is representative of the aeromagnetically defined dikes having the same orientation, they are the youngest intrusive rocks in the area.

Temporal Relationships—Geochronologic Data

All the intrusions described above are inferred to be Early Proterozoic, with the exception of diabasic dikes that cut the granitoid rocks and yield somewhat equivocal dates. Unfortunately, U-Pb data are rare, and field observations together with Ar-Ar and K-Ar dates make much of the emplacement history inferential. Reported U-Pb dates of 1,770 Ma for the St. Cloud Granite, and $1,812 \pm 9$ Ma for both the Rockville and Reformatory (Goldich, pers. comm., as reported in Horan and others, 1987) are consistent with the field observations that show Reformatory cut by St. Cloud Granite. Although not dated by the U-Pb method, the Richmond granite has Nd and Pb isotope and trace element systematics consistent with it being Early Proterozoic (Spencer, 1987).

Most controversial are the dates for diabasic dikes. Hanson (1968) gives K-Ar whole-rock ages of 1280, 1460, and 1570 Ma for the northeast-trending diabase dikes. Of these, the 1570 Ma date is considered a minimum age, and some of the dikes yield Pb isotope data consistent with an age of approximately 1800 Ma (Horan and others, 1987). Using these data, the northeast-trending diabasic dikes may be more or less comagmatic with enclosing granitic rocks. By contrast, Holm and Lux (1997) report Ar-Ar whole-rock ages of 1164 ± 12 Ma and 1190 ± 9 Ma from two of the northeast-trending diabase dikes. Field relationships do not permit us to unequivocally resolve these contrasting age dates. However, some field, geophysical, and petrographic observations raise questions about the younger (Keweenawan) emplacement date. For example, nearly all of the diabase dikes—including those dated by Holm and Lux (1997)—show alteration of mafic minerals to actinolite. The northeast-trending dikes in the St. Cloud District are only weakly to non-magnetic, in contrast with Keweenawan dikes exposed elsewhere that are strongly magnetic. Furthermore, paleomagnetic measurements on a limited suite of

samples failed to isolate any magnetization direction that is consistent with either the reversed or normal directions typical of Keweenaw dikes (Val Chandler, unpub. data, Minnesota Geological Survey). The granites surrounding the diabase dikes are not metamorphosed, implying that the alteration of mafic minerals to actinolite in the diabase may be of deuteric rather than metamorphic origin. In this case the Ar-Ar ages in the range of 1152-1199 Ma may indeed reflect emplacement, or alternatively a weak metamorphic overprint related to the onset of Keweenaw igneous activity to the east. A third scenario is that two different ages of northeast-trending dikes exist; however, no systematic differences between dikes of this trend have been observed in the field or during subsequent petrographic and geochemical study. The apparent conflict of dates and observations from diabasic and quartzofeldspathic dikes, and uncertainties about the age of other plutonic elements in east-central Minnesota, underscores the need for modern U-Pb zircon or baddeleyite dates.

Mafic microgranular enclaves-evidence for magma mingling

Mafic microgranular enclaves that occur in many intrusions of the St. Cloud area are interpreted as the product of commingled, coeval felsic and mafic magmas. The textural and mineralogical attributes of mafic microgranular enclaves in the Rockville Granite and Reformatory granodiorite, and the rapakivi texture of the Rockville and Richmond granites, are similar to features attributed elsewhere to the commingling of mafic and felsic magmas. Although some of the enclaves may have originated by the reincorporation of chilled intrusion margins by upwelling magma, textural evidence clearly shows that some of the enclaves coexisted as a melt with the host granite magma. Textures implying commingling of mafic and felsic magmas beg the question: Under what conditions are these coeval yet disparate magmas generated? The mingling of mafic and felsic magmas is most likely to occur in regions where crustal melts are generated by underplating of continental crust by mafic mantle-derived magma. Melt generated by this process and in a quantity sufficient to accumulate at the base of the crust can become fully mobilized and ascend rapidly, entraining the underlying mafic magma as it rises (Cruden and others, 1995, and references therein).

We suggest that the numerous mafic enclaves present in the ~1812 Ma intrusive suite give evidence that melting of the lower crust by mafic mantle underplating gave rise to the late-tectonic suite of granitic intrusions, during the waning compressive stages of Penokean orogeny. Furthermore, the numerous gabbroic to noritic intrusions such as that associated with the Richmond granite (Fig. 2) may represent mafic end-members of this same late-Penokean suite of intrusions.

In the case of the ~1770 Ma suite of post-tectonic granites, mantle underplating as the result of mantle lithospheric delamination (e.g., Holm and Dahl, 1996; Cruden and others, 1995; Turner and others, 1992) may have generated the melts. In this model, delamination and thinning of the orogenically overthickened crust allows hot asthenosphere to well up, raising the asthenosphere-lithosphere boundary and increasing the thermal input into the base of the crust. This process is attractive because granitic melt generation is delayed until some 40-50 m.y. after the peak of orogenesis, coincident with the time necessary for orogenic collapse and thinning of the mantle lithosphere (Turner and others, 1992).

There are several lines of evidence which indicate that the post-orogenic granites of the St. Cloud area may have formed from crustal underplating by high-temperature mantle material following crustal collapse and mantle lithospheric delamination.

1. The granite melts are inferred by Spencer (1987) to have formed at high temperatures (800-1000°C), and are nearly coeval with the isotopically similar mantle-derived diabasic rocks (Horan and others, 1987).
2. The close and probable cogenetic association of the porphyritic microgranite dikes with diabase dikes. This may reflect bimodal magma generation, a common feature of post-orogenic plutonic complexes elsewhere (Turner and others, 1992).
3. The emplacement of granite lags behind peak orogenesis by ~50 Ma.
4. Holm and others (1998) concluded from garnet and staurolite core-rim thermobarometric, geobarometric, and thermochronologic studies (see following section) that crust thickened by the Penokean orogeny remained so for 50-60 m.y. before collapse, which then may then have triggered sudden lower crustal melting.
5. Spencer (1987, p. 174, e.g.) interpreted from isotopic evidence that the granite melts were generated at a convergent continental margin that recycled a source material with a short-lived crustal history. However, he allows for an extensional environment superimposed on an earlier orogenic event, to produce basic mantle melts beneath the crust to melt the crustal rocks.

Thermobarometric data

The significant proportion of igneous rocks, together with the regional staurolite grade of metamorphism of the country rocks, implies that outcrops in this area represent one of the deepest exposed levels of the entire Penokean orogenic belt. Holm and others (1998) recognized that Early Proterozoic country rock within the internal zone (Little Falls Formation) were metamorphosed twice. Thermobarometric estimates indicate that early, Penokean-age, garnet-grade (M1) metamorphism occurred at pressures of 5-6 kbar and temperatures of 450-500°C. Thermobarometric estimates of the M2 metamorphism yield similar pressures, and only slightly higher temperatures (500-550°C). This indicates that Penokean plutons exposed in the St. Cloud District were emplaced during the interim between the two metamorphic events at 19-23 km midcrustal orogenic depths. The age of M2 metamorphism must be older than the 1750-1760 Ma mica Ar/Ar cooling ages obtained throughout the region because of the 300°C closure temperature of biotite (Fig. 5; Holm and Lux, 1997; Holm and others, 1998). These ages indicate that the area has not been reheated to temperatures above 300°C since that time. Five hornblende Ar/Ar dates obtained from central Minnesota are concordant with the 1760-1750 Ma mica dates. Hornblende has a typical closure temperature of 500°C, which matches the temperatures obtained for M2 metamorphism (500-550°C). This is strong evidence that the rocks of the internal zone of the orogen in central Minnesota were widely remetamorphosed to staurolite grade during the same event responsible for generating the post-tectonic magmas at 1770 Ma. The M2 pressure estimate of 5-6 kbars indicates that little or no unroofing of the Penokean midcrustal root occurred between about 1820 and 1770 Ma.

Emplacement pressure estimates using an igneous barometer indicate that some of the 1770 Ma post-Penokean plutons were intruded at 4.6 kbar pressures (17 km depth); whereas, others were emplaced at lower pressures of 3.5 kbar, or about 13 km depth (Darrach, 1996). These contrasting pressure estimates—derived from suites of rock tentatively inferred to be comagmatic—indicate that post-tectonic magma emplacement was synchronous with crustal uplift (Fig. 6).

Early Proterozoic tectonomagmatic model

Our mapping observations and analytical results of the last five years indicate that there are significant differences in the overall character of the late vs. post-Penokean plutons (Table 1) that must be explained in any tectonic model. We believe these differences reflect two very different tectonomagmatic settings; one that occurred at the end of orogenic belt construction, and one during its demise approximately 40-50 my later.

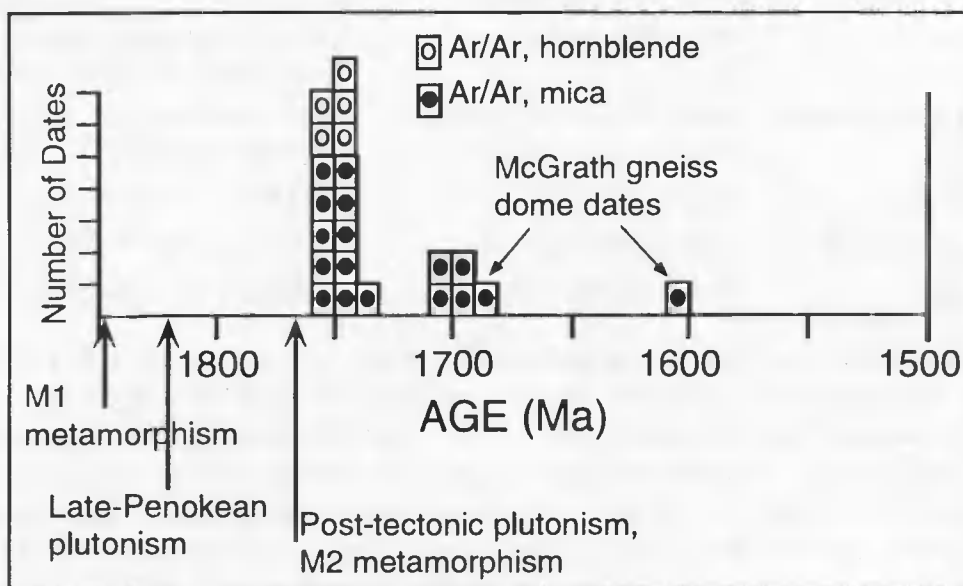


Figure 5. Histogram of Ar/Ar dates from the Early Proterozoic of central Minnesota (after Holm and Lux, 1996; Holm and others, 1998).

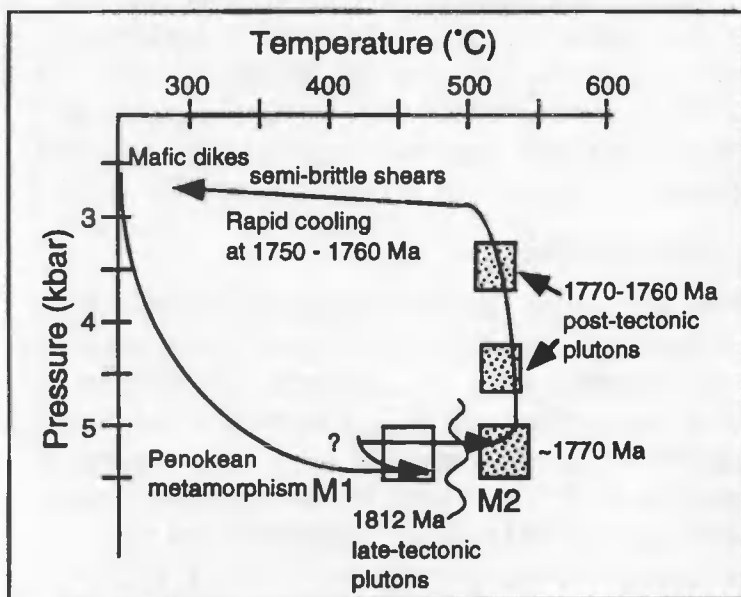


Figure 6. Summary of P-T-t diagram showing the relation between metamorphism (M1 and M2), plutonism, and formation of semi-brittle shears in the internal zone of the Penokean orogen (modified after Holm and others, 1998).

Figure 7 presents a schematic synopsis of our interpretation of pluton emplacement after the compressive phase (Fig. 7A) of Penokean deformation in central Minnesota. The Richmond, Reformatory, and Rockville—the “gray granites”—represent lower crustal melts intruded into the midcrust (20-22 km) of the Penokean orogen (Fig 7B). These magmas were generated during the waning stages of Penokean compression and hence, developed a moderate magmatic fabric.

The evidence for commingled mafic and felsic magmas in these rocks indicates

TABLE 1—Comparison of late and post-Penokean intrusions

TYPE/AGE	LATE PENOKEAN/~1812 Ma	POST-PENOKEAN/~1770 Ma
Composition/Color	Granodiorite-granite—typically gray—with mafic enclaves	Granite (alkali-feldspar granite to syenogranite)—typically red—with few mafic enclaves
Magmatic Fabric	Moderately developed, NE orientation, commonly flat	Very weak to absent
Pluton Shapes	Broad and irregular	Dikes, sills, and larger plutons with a tendency for NE orientation
Geophysical Characteristics	Variable - moderate and busy magnetic signature; dense	Typically non-magnetic, pronounced gravity lows
Depth Emplaced	20-22 km (5-6 kbar)	13-17 km (3.5-4.5 kbar)
Associated Metamorphism	Garnet grade M1 (5-6 kbar)	Staurolite grade M2 (5-6 kbar)
Solid-state Fabric	Cut by semi-brittle shears	Cut by semi-brittle shears

that crustal underplating may have played an important role in their generation. However, these crustal melts were not able to penetrate the stronger upper crust of the orogen, but instead formed irregular shaped plutons at depths of 20-22 km where they remained for approximately 40 m.y. The period of tectonic quiescence and amagmatism after the Penokean orogeny (during the period approximately 1812-1770 Ma) was terminated by a short-lived period of post-tectonic magma generation, metamorphism, and rapid midcrustal uplift and cooling at about 1770 Ma. This abrupt episode may have been triggered by mantle-lithospheric delamination that would have caused both melting of the lower crust (Fig. 7C) and heating of the midcrust to temperatures above 500°C during M2 metamorphism. This thermal heating softened the overthickened crust, causing it to collapse and thin (Fig. 7D). At this time significant unroofing occurred and shears formed as the rocks cooled and became more brittle. The St. Cloud Granite, Foley batholith, and associated plutons—the “red, massive granites”—were advected during this event into uplifting crust that was extending NW-SE, parallel to the previous orogenic shortening direction. Early Proterozoic diabasic and quartzofeldspathic dikes apparently lagged behind a bit and were emplaced just after brittle shears formed, using the shears as conduits (Fig. 7E).

ACKNOWLEDGMENTS

This field guide is an outgrowth of studies by the Minnesota Geological Survey (MGS) that were supported by several programs of the State Legislature. The Stearns County segment (Boerboom and others, 1995, 1996a) was funded through the Legislative Commission on Minnesota Resources; funding for drilling and mapping east of St. Cloud (Jirsa and others, 1995a) was made available on recommendation of the Minnesota Minerals Coordinating Committee. On-going mapping and analytical work in both areas has been supported by the MGS. The thermobarometric and geochronologic work of Daniel Holm and others was supported by a National Science Foundation grant (EAR-9304070). We are grateful to Mr. Frank Lyons and the Meridian Aggregates Company for access and guidance in the aggregate quarry, and for permission to visit their developing southern property (before the natural outcrops are gone!). A very special thanks to Charles Muehlbauer and other employees of the Cold Spring Granite Company for assistance, advice, samples, permission, and guidance (on their day off!).

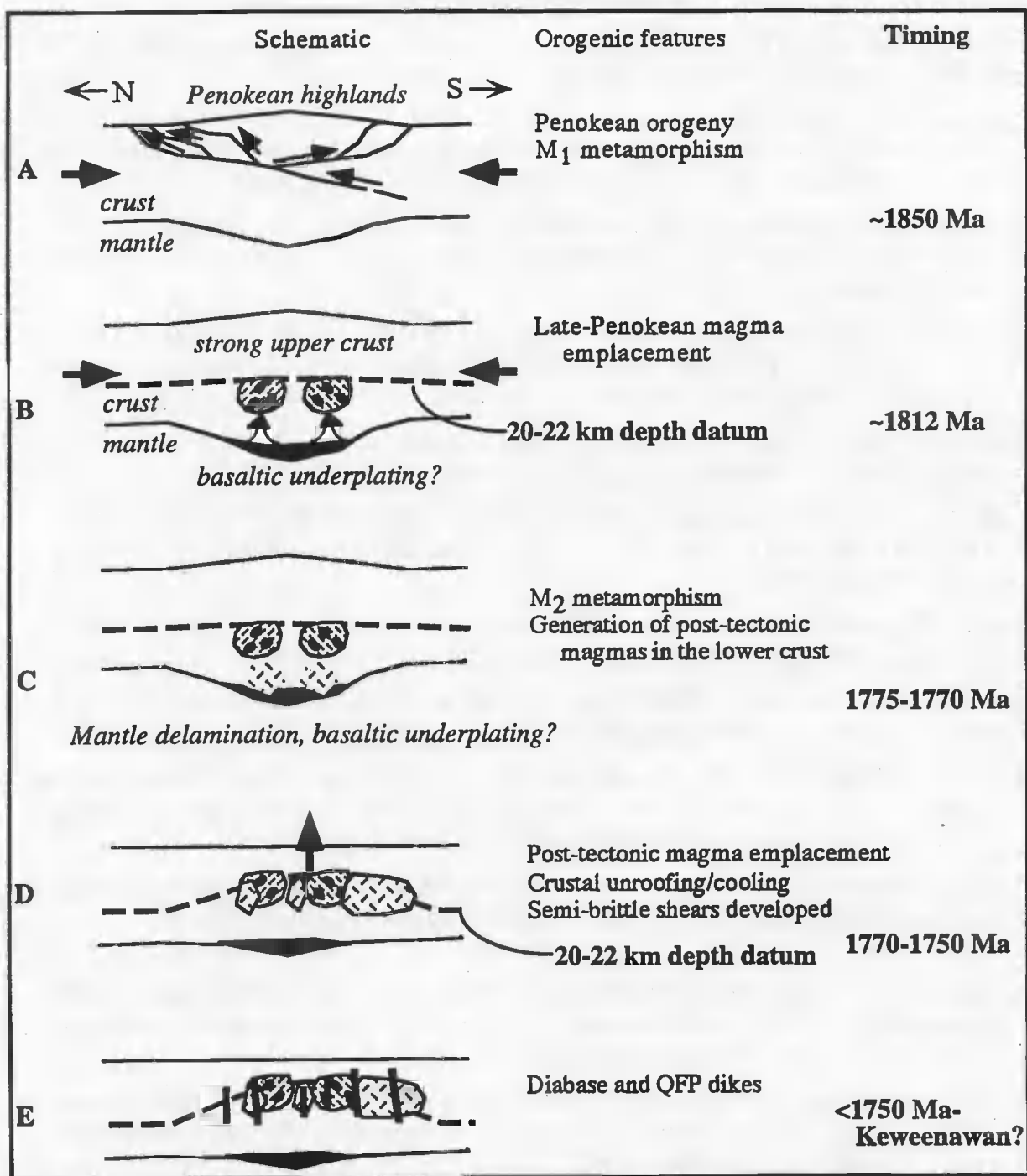


Figure 7. Proposed tectonomagmatic mode for generation of late- and post orogenic plutons, internal zone of the Penokean orogen, central Minnesota. Heavy dashed line represents 20-22 km depth datum prior to crustal unroofing (B and C). Crustal unroofing and exhumation causes upwarping of this datum level in (D and E).

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FIELD TRIP STOPS

The general locations of all stops are shown on figure F1. A more detailed location and geologic map is included with each stop description. The various rock units are fully described in the field trip introduction, and only the details pertinent to the individual sites are given below.

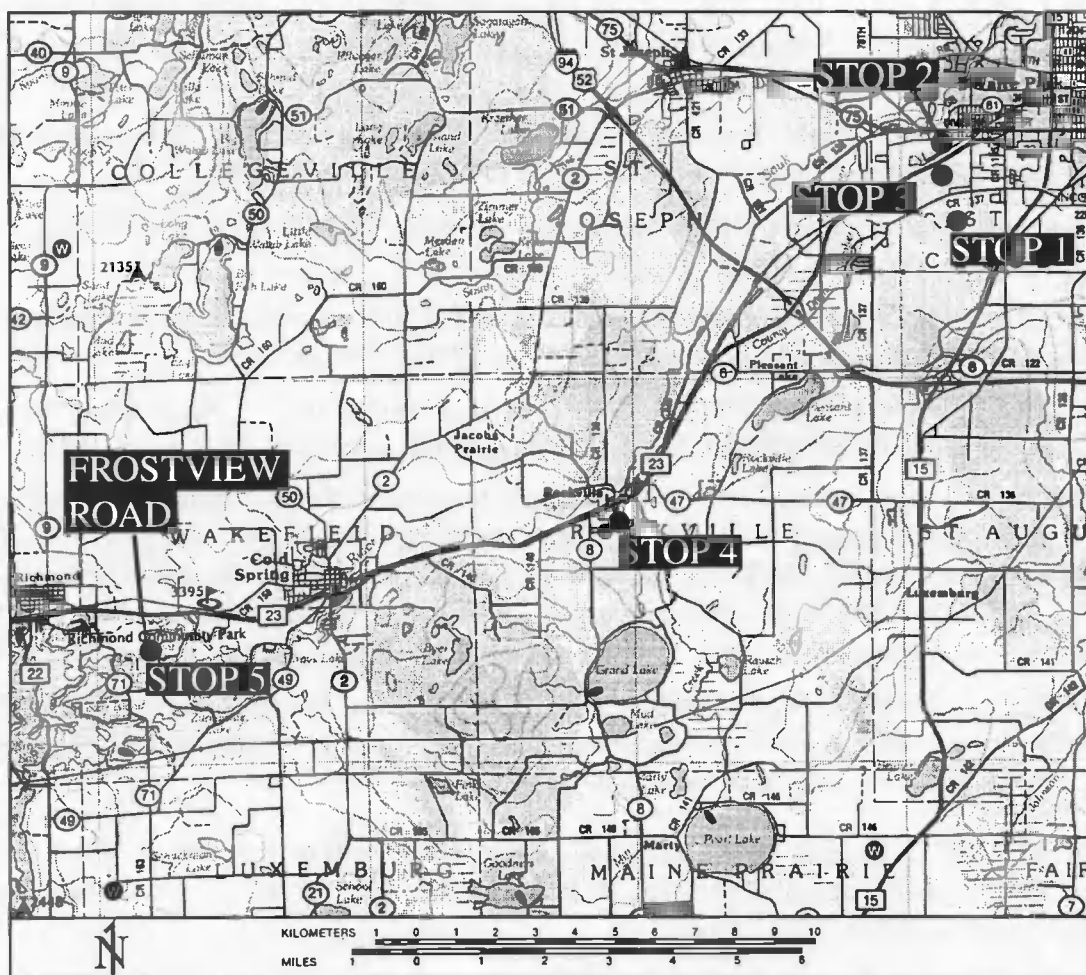


Figure F1. Locations of field trip stops.

STOP 1. STEARNS COUNTY QUARRY PARK

- **Location:** T. 124 N., R. 28 W., Section 19 DDA.
- **St. Cloud (red) Granite, diabase dikes, semi-brittle shears in granite, glacially polished outcrops.**

This area contains several inactive quarries (Figs. F2 and F3), and was recently made into a unique county park. The park provides excellent exposures in which one can view the relationship between the Reformatory granodiorite and the St. Cloud Granite, as well as several diabase dikes. Although the park has exposures of both northeast- and northwest-trending diabase dikes, we will only see northeast-trending ones and St. Cloud Granite, due to time constraints.

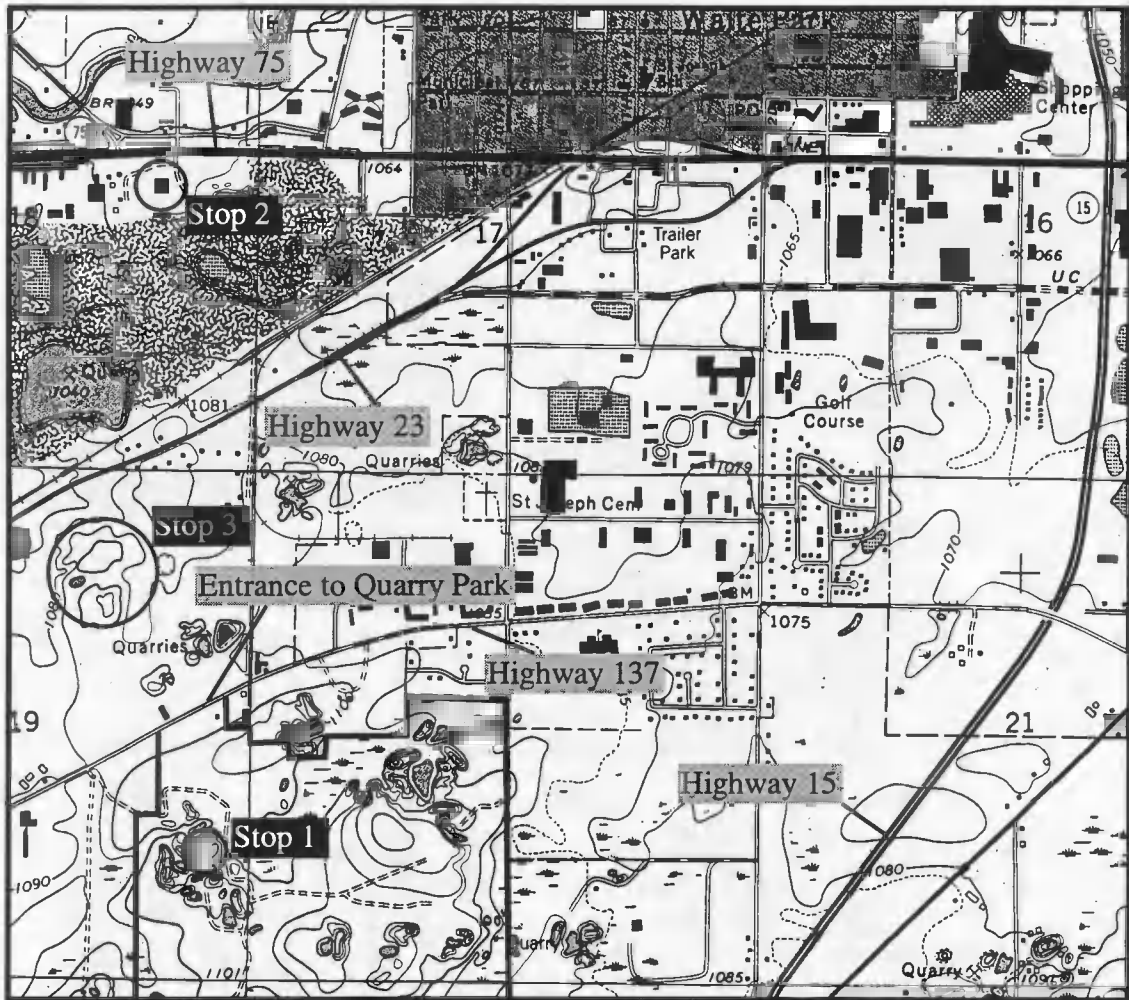


Figure F3. Location map for stops 1, 2, and 3. Scale 1:24,000; St. Cloud 7.5-minute quadrangle.

At least seven different quarry companies were active in what is now the park area—all utilizing the St. Cloud Granite. Production started in the early 1900's, peaked in about 1940, and ceased in the mid-1950's. The stone they produced was marketed under a variety of "red" names, including "North Star Red," "Mahogany Red," "Indian Red," and "Rose Red." Most of the pits were (and still are!) 60 to more than 100 feet deep (Thiel and Dutton, 1935).

PLEASE EXERCISE EXTREME CAUTION! THE QUARRY PILES ARE NOT STABLE, AND THE PITS ARE DEEP!

Follow the park road south from the park entrance to the parking area. From here walk 400 feet south on a dirt road to a trail on the right (west) that leads up to quarried area. Proceed up this trail to the edge of a small quarry and Stop 1A (Fig. F4).

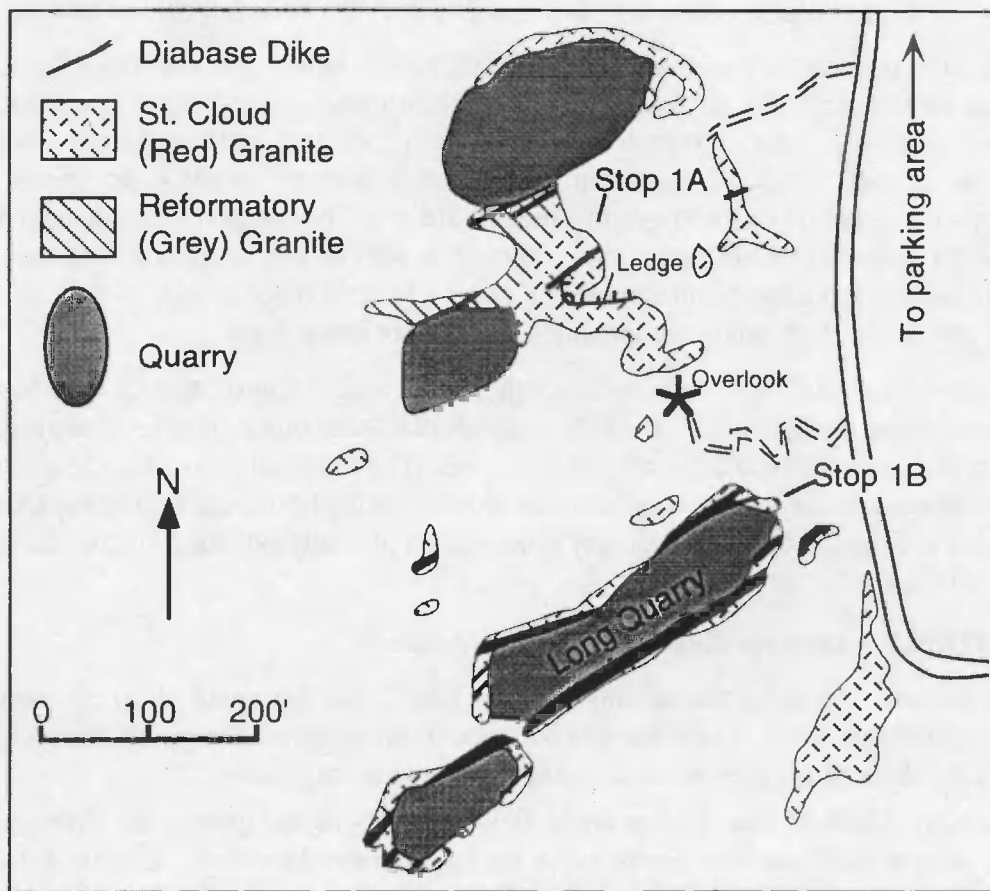


Figure F4. Outcrop map of Stop 1A and 1B.

STOP 1A—Three-dimensional view of diabase and quartz-epidote shears in St. Cloud Granite.

Directly west, across the rubble in the pit, is an excellent exposure of a northeast-striking, steeply south-dipping, meter-thick diabase dike that cuts St. Cloud Granite. The dike was sampled for both paleomagnetic study and Ar/Ar dating (see introduction). This is typical of the diabases in the area, having chilled margins and scattered cm-scale xenoliths of granite wallrock. The granite adjacent to the dike contains numerous thin, northeast-trending, quartz- and epidote-lined, semi-ductile shears that

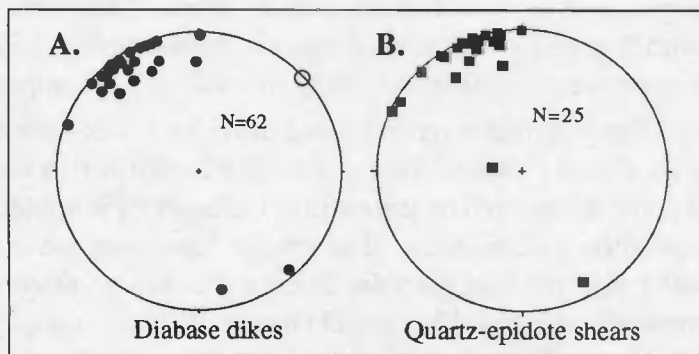


Figure F5. Stereographic projections of orientations of diabase dikes (A) and quartz-epidote-filled semi-brittle shears (B).

are clearly truncated by and are probably in part occupied by the diabase dike. This nearly coincident trend of dikes and shears is typical of the regional pattern shown on the stereoplots in figure F5. Two other diabase dikes, up to 2 meters thick, can be seen north of this dike.

Although the majority of blocks in the waste rubble before you are indigenous, several exotic rock types are present, including blocks of Reformatory granodiorite (gray) that we will see in place at Stop 3. The Reformatory granodiorite is presently extracted from a quarry just south of the city of St. Cloud, marketed as "Charcoal Granite" by the Cold Spring Granite Company. The maroon-colored granite blocks are probably imported from quarries near Ortonville, Minnesota and Milbank, South Dakota. Blocks of grayish-pink, complexly banded Morton gneiss are from the Minnesota River Valley. Several other blocks of pink granite and pale pink gneiss are from unknown localities, but are not indigenous.

Return to the dirt road, and walk south (to the right) approximately 250 feet to the switchback trail on the right. Note the spheroidally-weathered outcrops of red granite along the west (right) edge of the road between the two trails. The aggregate covering the switch-back trail illustrates one of the many uses of crushed stone from the Meridian Aggregates Quarry that we will visit at Stop 2. At the next quarry proceed to a glacially polished outcrop on the north (right) side of the pit.

STOP 1B—Diabase dike in St. Cloud Granite.

The diabase dike splits and rejoins along its length, and has small, 4-10 cm, north-south 'jogs' at its northeast trend. There are no obvious cross-fractures in the granite that might have controlled these small offsets, making them somewhat enigmatic.

This stop affords a better look at the chilled dike margins and granite xenoliths, as well as slivers of granite wallrock torn partly loose during dike emplacement. Granite xenoliths in various stages of assimilation can be seen, and xenoliths are concentrated in the center of the main dike, presumably by flow entrainment. Also visible here, with some searching, are ± 0.5 cm amygdules filled with fibrous actinolitic amphibole.

Walk around the quarry to its south side, where two subparallel, northeast-trending diabase dikes occur. These dikes are approximately 3 m thick—the maximum thickness of dikes in this set. They are cut by thin, white, silicic, and sugary-textured veinlets that emanate from the dike margins and pinch out in the dike interiors. The veinlets—interpreted as partial melts of the wallrock granite—are rarely present in dikes less than 2.5 m thick, presumably because narrower dikes were not capable of heating the wallrock to temperatures sufficient for melt generation.

The polished outcrop surface provides a good look at the mineralogy and coarse texture of the St. Cloud Granite, characterized by its distinctive black and white mottled, red color. The red color is imparted by microcline, the black by hornblende and minor biotite, and the white by saussuritized plagioclase. The granite has a very weak trachytoid foliation, via alignment of blocky feldspar and hornblende crystals. Where discernible, this fabric parallels the granite dike or sill margins. This granite has an amazing lack of pegmatite (very coarse-grained phase) and a preponderance of hornblende as the mafic phase.

Glacial features are well displayed here and include polished outcrops, striations, and differential scouring of the diabase/granite contacts. 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.

An interesting hydrologic observation 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 diabase dike's width, and the hydrologic gradient exists despite the abundance of cross-fractures in the diabase.

Addendum to Stop 1

The sketch map shown on figure F6 is included for those who wish to visit the park to view other exposed geologic features. Of particular interest is a pit at the east side of the park where a northwest-trending diabase dike cuts a northeast-trending one. The next pit to the northwest of that shows St. Cloud Granite having many angular, meter-scale xenoliths of Reformatory granodiorite—typical of regional intrusive relationships between the two units.

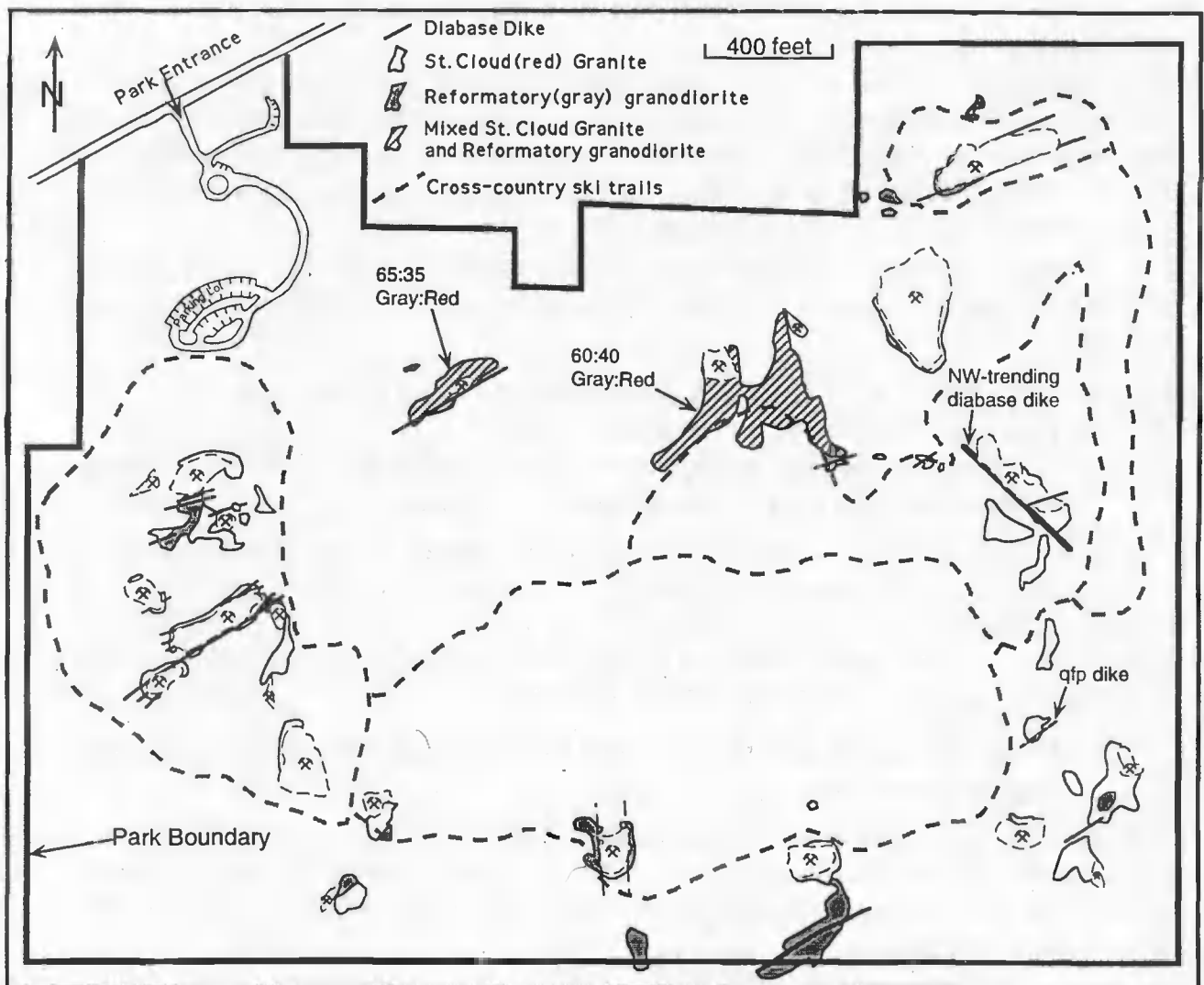


Figure F6. Sketch map of the Quarry Park showing outcrop distribution and simplified bedrock geology. Modified from Stearns County Park winter trail map.

STOP 2. MERIDIAN AGGREGATES QUARRY

- **Location:** T. 124 N., R. 28 W., south half of Section 18.
- **Quarry overlook, crushing facilities, restrooms.**

Due to constraints of time, safety, and group size, we will not go into the Meridian Aggregates quarry, but we will get an excellent view of the pit from an overlook point.

The Meridian Aggregates Company produces an average of 1.2 million tons of crushed rock aggregate per year. Mining and crushing operations run 7 months of the year, although stockpiled product is shipped year-around. This quarry utilizes the complex mixture of igneous rock types present in the pit, mainly the Reformatory granodiorite and St. Cloud Red Granite, and lesser proportions of diabase and quartz-feldspar porphyry dikes, such as will be seen at Stop 3. Unlike dimension stone quarries, joints and fractures in the rock are not detrimental—blast away! Quarrying methods involve drilling and blasting, primary crushing in a gyratory crusher, and secondary crushing to various sizes. The primary use of the aggregate is railroad ballast, followed by road construction and other uses, including breakwater walls, decorative stone, and driveway material.

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 in the Reformatory granodiorite. Although there are undoubtedly some small xenoliths, textural and geochemical evidence indicates that the bulk of this gray material is fine-grained border phase of the Reformatory granodiorite.

If the light conditions are proper, a cross-section is visible on the far wall of the pit showing fresh rock, transitionally overlain by weathered bedrock residuum (saprolite), overlain by glacial deposits.

STOP 3. MERIDIAN AGGREGATES COMPANY—SOUTH PROPERTY

- **Location:** T. 124 N., R. 28 W., Section 19 AB.
- **St. Cloud (red) Granite, aplite, Reformatory granodiorite, diabase dikes, quartz-feldspar porphyry (aka porphyritic microgranite) dikes.**

This stop is located in an area of planned large-scale mining like that in Meridian's main pit (Stop 2) just across Highway 23 to the north. There are a few small abandoned quarries on the property, but most of the outcrops are natural (Fig. F7). Permission must be obtained from Meridian Aggregate Company, which has kindly granted access for this trip, before entering their property, but these outcrops will soon be mined out.

STOP 3A—NE-trending dikes of St. Cloud (red) Granite intrusive into Reformatory (gray) granodiorite.

The St. Cloud Granite is not a distinct single body, but rather an intricate series of vertical and horizontal dikes emplaced into the Reformatory granodiorite. As represented in this series outcrops, the intrusions have complex geometries and varied thicknesses from few centimeters to more than 50 meters.

STOP 3B—A variety of intrusive rock units, including (oldest to youngest) Reformatory granodiorite, St. Cloud Granite, aplite related to the St. Cloud Granite, and diabase.

The aplite occurs as small, irregular pod-shaped masses that apparently cut both the red and gray intrusions. Although the timing of emplacement of the aplites relative to that of the quartz-feldspar porphyry and diabase dikes is uncertain, the irregular shapes of the former vs. the straight and continuous (fracture-controlled emplacement) of the latter two leads to the inference that the aplites are a late differentiate of the St. Cloud Granite. At this site and the next (stop 3C), note the contrast in jointing pattern of the aplite and porphyry compared to the surrounding plutonic rocks.

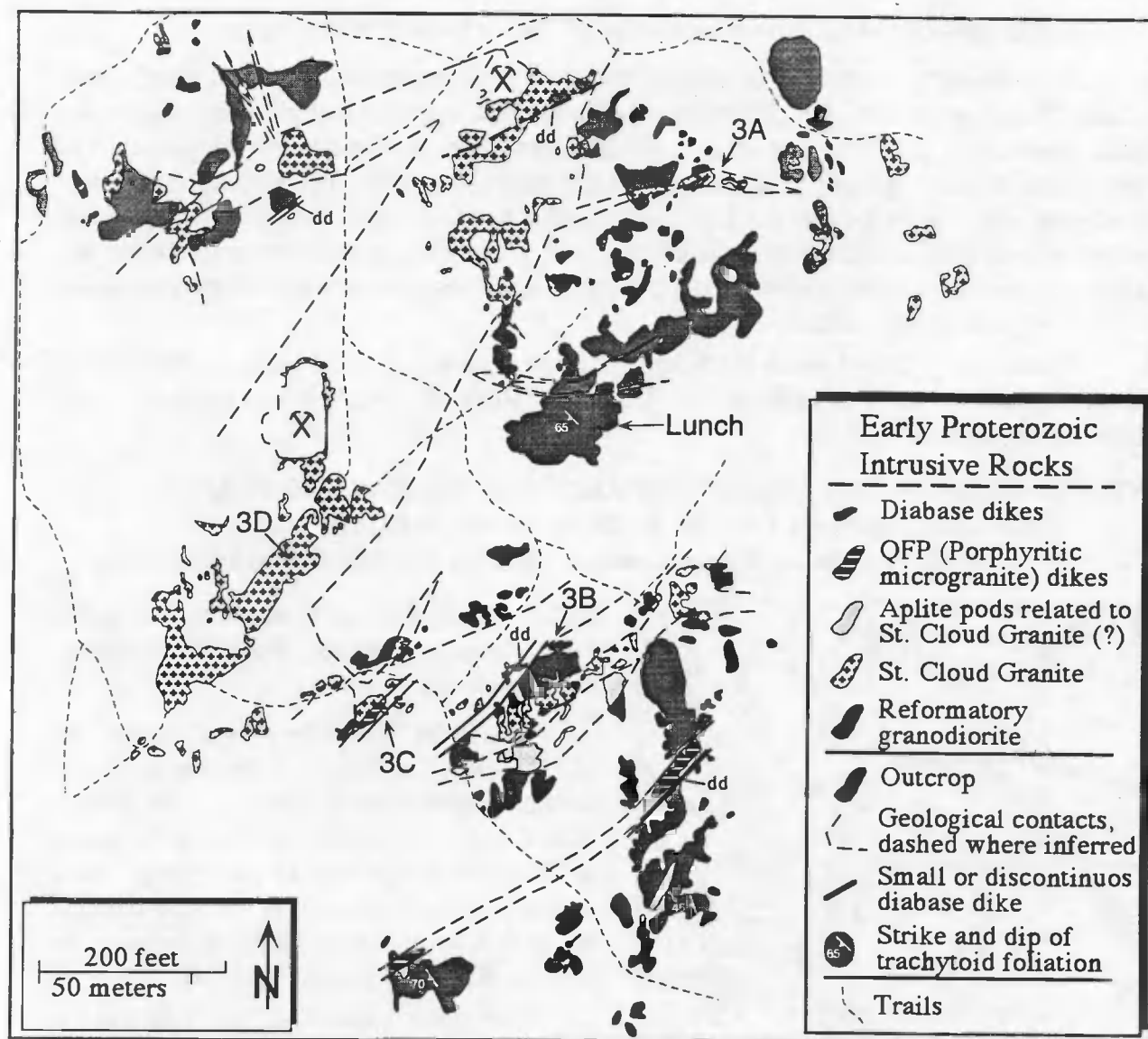


Figure F7. Outcrop and geologic map of Stop 3.

STOP 3C—Porphyritic microgranite (aka quartz-feldspar porphyry).

This dike strikes N45E, is 4-5 m thick, and has 4-6 percent glassy subhedral quartz phenocrysts and 3-4 percent subhedral K-feldspar phenocrysts in an aphanitic, siliceous, groundmass. Dikes such as this are scattered around the St. Cloud area, and like the diabase dikes, they cut the Reformatory granodiorite, Rockville Granite and the St. Cloud Granite. The phenocrysts vary from the 2-4 mm size here, to as large as 2 cm elsewhere. The margins of the quartz-feldspar dikes are chilled, and in some cases flow-banded. A similar dike in the aggregate quarry north of here is clearly comagmatic and commingled with a diabase dike that contains fresh ophitic clinopyroxene, olivine, and chromite. However, that diabase dike contrasts with all the other diabase dikes in this area in that the mafic minerals are not altered/metamorphosed to actinolitic amphibole.

STOP 3D—Small xenolith of garnet amphibolite in St. Cloud Granite.

True xenoliths of external origin (vs. cogenetic mafic enclaves) are rare in all of the St. Cloud District intrusions. Xenolith types noted in the St. Cloud area include garnet-magnetite-amphibole schist (probably metamorphosed iron-formation), biotite-quartz-feldspar schist of graywacke protolith, granulitic gneiss that contains garnet, sillimanite, magnetite, and abundant cordierite, and charnockitic garnet-biotite-orthopyroxene granofels. The general metamorphic mineral assemblages indicate that the country rock was metamorphosed to amphibolite and higher grade prior to incorporation into the granites, with minor retrograde metamorphism caused by contact with granite magma.

On the way to Stop 4 watch for the derrick in the deep quarry north (right) of the highway on the east edge of the town of Rockville. This is the “Rockville Beige” quarry operated by the Cold Spring Granite Company.

STOP 4. COLD SPRING GRANITE QUARRY AND PROCESSING PLANT

- **Location:** (quarry) T. 123 N., R. 29 W., Section 19 BDAB.
- **Rockville Granite, quarry methods, production of finished dimension stone.**

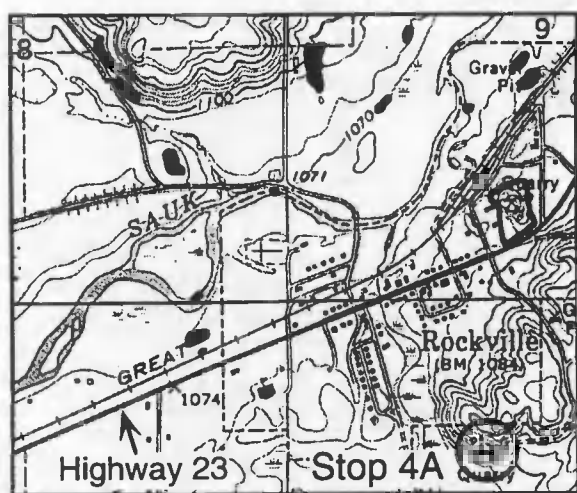


Figure F8. Location of Stop 4A. Other outcrops of Rockville Granite shown in black, except over the two quarries.

STOP 4A—Cold Spring Granite Rockville White quarry (Fig. F8), Rockville Granite, mafic enclaves.

This is the “Rockville White” quarry, one of the three Cold Spring Granite quarries developed in Rockville Granite. The granite from the other Rockville quarries is marketed as “Rockville Beige” and “Diamond Pink.” 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.

This quarry uses the “drive-in quarry method,” a system that replaces the traditional cable-operated derrick method of stone removal. Surprisingly, the method was developed only in

the mid-1970's (Muehlbauer and Fuchs, 1997). The drive-in method utilizes a series of benches, from which the stone is removed as a series in large "loaves" that vary in size—15-25 feet (5-8m) high x 20 feet (6m) wide x 30-150 feet (10-50m) long. These loaves are subsequently drilled off into smaller blocks that are optimally 5.5 x 5.5 x 10.5 feet in size, weighing in at approximately 25,000 kg, or nearly 28 tons. This size is small enough to transport legally by flatbed trailer without a special permit, and allows for minimum wastage during processing.

Charles Muehlbauer of Cold Spring Granite will give a more extensive overview of the quarrying process at the quarry, and will also answer questions about the quarry and stone products in general.

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

1. Porphyritic texture, rapakivi feldspars.
2. Mafic enclaves, some with feldspar phenocrysts in them. The enclaves are interpreted as the product of the mingling of mafic and felsic magmas (see INTRODUCTION). Most occur as ovoid inclusions, although discontinuous and irregular dike-like masses of the same composition have been noted in another quarry.
3. Weak trachytoid texture defined by aligned feldspar phenocrysts. This is typical of the "gray granites."

Stop 4B—Cold Spring Granite Company Processing and Fabrication Plant.

Cold Spring personnel will guide us in small groups through two different buildings. One is the facing plant, where rough quarry blocks are cut, finished (polished vs. the rougher "fire-finish"), and made into facing panels of varied size, depending on product specifications. The other is the monument plant, where the stone is cut, shaped, polished, and engraved. Notice surprisingly ornate works!

*****PLEASE OBSERVE SAFETY REGULATIONS—**

HARD HAT AND SAFETY GLASSES MANDATORY***

STOP 5. RICHMOND GRANITE

- **Location:** T. 123 N., R. 30 W., Section 29 BB center.
- **Charnockitic, rapakivi granite with prominent trachytoid K-feldspar phenocrysts.**

*****PRIVATE PROPERTY—MUST HAVE PERMISSION OF HOMEOWNER TO ENTER*****
—HAMMERING AND SAMPLING PROHIBITED—

The Richmond granite was interpreted by earlier workers as a high-grade metamorphic gneiss of Archean age. By contrast, this stop demonstrates that the granite has a primary igneous texture, which is typical of the unit overall. It contains undeformed unit quartz grains as large as 1 cm and euhedral feldspar phenocrysts as large as 5 cm. Because quartz is one of the first minerals to exhibit strain and recrystallization during deformation and metamorphism, the rock clearly has not experienced substantial amounts of either process, and only localized semi-brittle shearing exists. We interpret this and other lithologic evidence to indicate that the Richmond granite was emplaced after most of the regional deformation ceased, and infer that it probably had an emplacement history like that of the Reformatory granodiorite and the Rockville Granite.

Key features at stop 5:

1. Dark green charnockitic granite. This color phase grades into the more typical pink-tinted Richmond granite, due to oxidation introduced along 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.
2. 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.
3. Magmatic-trachytoid alignment of feldspar phenocrysts (see Fig. F9).
4. Small dark-colored mafic enclaves. From a petrographic standpoint, these enclaves are mineralogically identical to a fine-grained, trachytoid, apatitic, quartz-orthopyroxene-clinopyroxene ferrodiorite (or jotunite, in charnockite terminology) that occurs in several outcrops 1/4 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 to the west is inferred to be a border phase of the Richmond, and the small enclaves scattered throughout the Richmond are interpreted as cognate xenoliths derived from this dioritic border phase.

-----END OF FIELD TRIP. THANK YOU FOR ATTENDING!-----

You are encouraged to visit the poster titled
“DIMENSION STONE PRODUCTS OF MINNESOTA”
on display at the poster session.

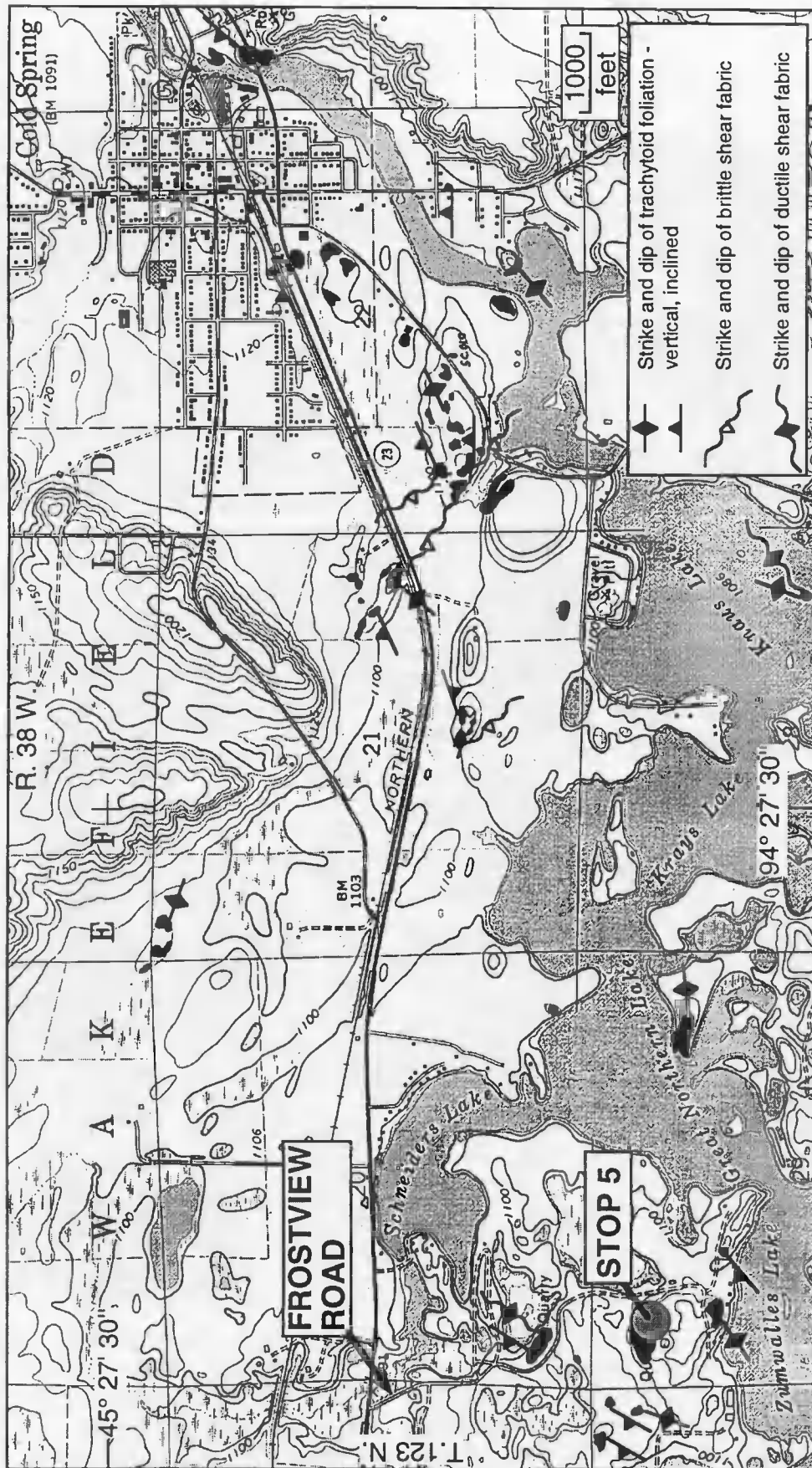
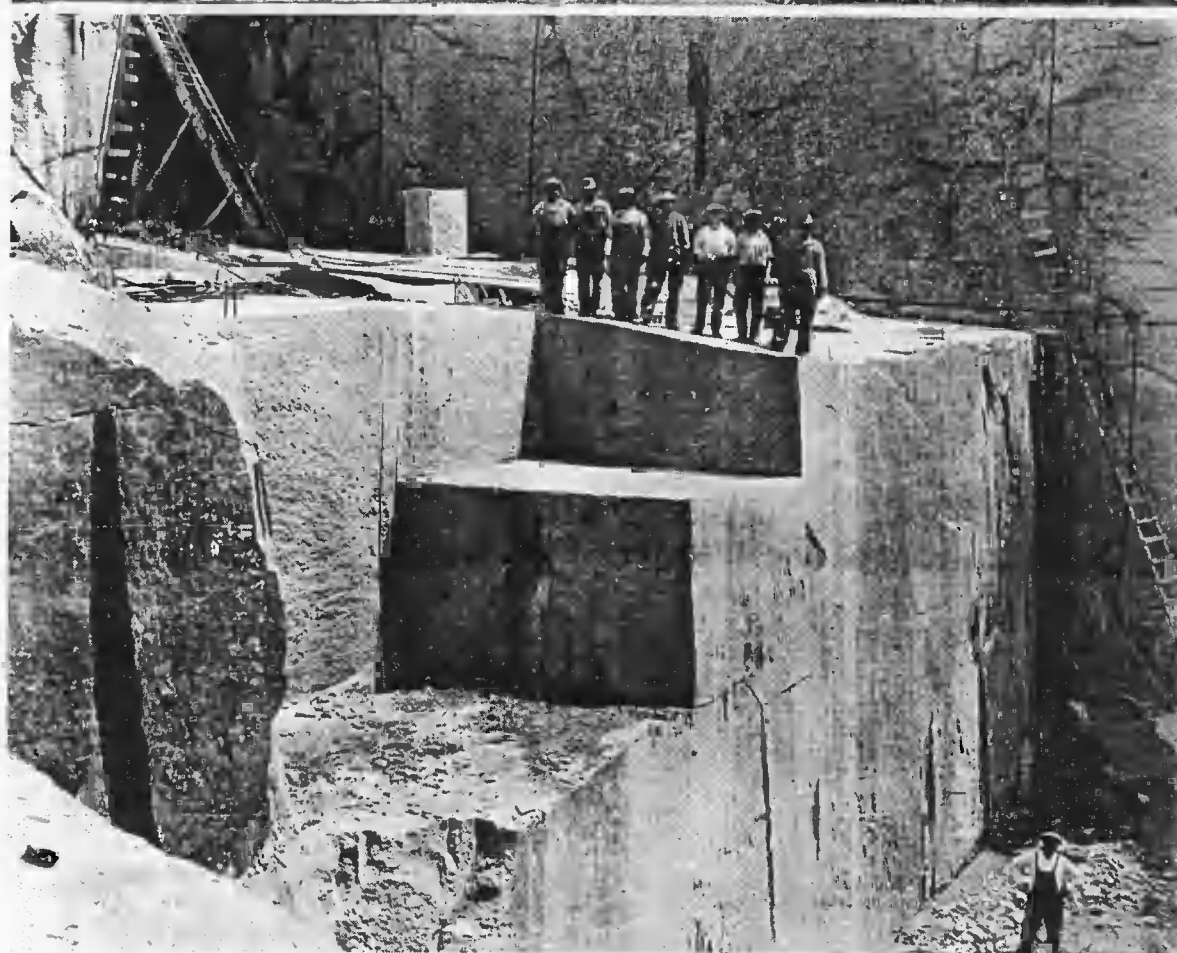
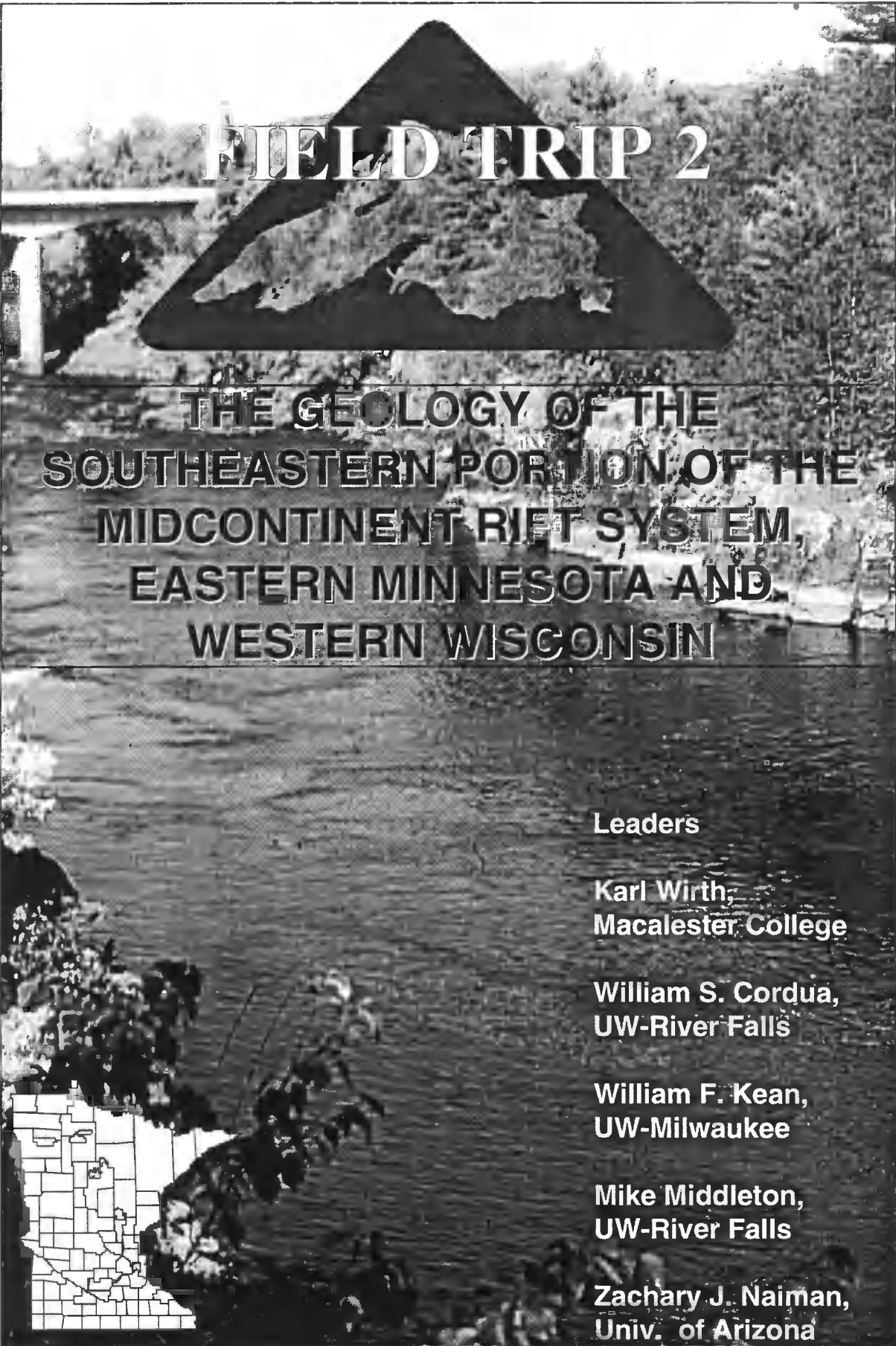


Figure F9. Map of central part of the Cold Spring 7.5-minute quadrangle, showing outcrops of Richmond granite. Structure symbols touch the outcrop from which they were taken.



PHOTOGRAPHS OF QUARRIES IN THE ST. CLOUD AREA FROM THE EARLY 1900's. TOP, QUARRIES OPERATED BY NORTH STAR GRANITE COMPANY, NOW OVERRUN BY THE PRESENT-DAY MERIDIAN AGGREGATES QUARRY?; BOTTOM, A QUARRY OPERATED BY THE MELROSE GRANITE COMPANY, LOCATED IN THE VICINITY OF STOP 1 OF FIELD TRIP 1. REPRODUCED FROM JOURNAL ARTICLE OF THE BRILLIANT COLORED GRANITES FOUND IN AMERICA'S GREAT NORTHWEST IN AMERICAN STONE TRADE, PUBLISHED MAY, 1931.



FIELD TRIP 2

THE GEOLOGY OF THE SOUTHEASTERN PORTION OF THE MIDCONTINENT RIFT SYSTEM, EASTERN MINNESOTA AND WESTERN WISCONSIN

Leaders

Karl Wirth,
Macalester College

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Zachary J. Naiman,
Univ. of Arizona

FIELD TRIP #2

FIELD GUIDE TO THE GEOLOGY OF THE SOUTHEASTERN PORTION OF THE MIDCONTINENT RIFT SYSTEM, EASTERN MINNESOTA AND WESTERN WISCONSIN

**Leaders: Karl R. Wirth, William S. Cordua, William F. Kean,
Mike Middleton, Zachary J. Naiman**

INTRODUCTION

The Middle Proterozoic (Keweenawan) Midcontinent rift system extends over 2,000 kilometers in an arc across central North America from Kansas to central Michigan. Magmatic and sedimentary rocks in this major tectonic feature record the evolution of continental lithosphere as it interacted with a mantle plume 1100 million years ago. Rifting was aborted before the production of oceanic lithosphere began. Field, geochemical, geochronologic, and geophysical studies during the past decade have made significant contributions to our understanding of the evolution of the Midcontinent rift system. The interested reader is referred to several recent reviews (Van Schmus and Hinze, 1985), special volumes (Tectonophysics, v. 213, 1992; Canadian Journal of Earth Science, v. 31, 1994; Canadian Journal of Earth Science, v. 34, 1997; Ojakangas et al., 1997), and guidebooks (Miller et al., 1995) on the Midcontinent rift for additional information.

Geophysical studies have provided much information about the nature of the rift, which lies buried beneath younger deposits along much of its length. Regional gravity and magnetic surveys were important in establishing the extent and character of the rift (e.g., Hinze et al., 1992). Our knowledge of the detailed structure and stratigraphy of the rift has been greatly enhanced by several seismic surveys, most notably those of the COCORP (Serpa et al., 1984; Zhu and Brown, 1986) and GLIMPCE (Behrendt et al., 1988; Cannon et al., 1989) programs. These surveys revealed the presence of fault-bounded basins filled with thick sequences of volcanic and sedimentary rock. Detailed seismic images of the eastern portion of Lake Superior indicate the extent of crustal thinning (~30 percent of original thickness) in some portions of the rift; the rift in this region is more than 30 kilometers deep and consists of roughly 20 kilometers of basalt overlain by 10 kilometers of Keweenawan sedimentary rock.

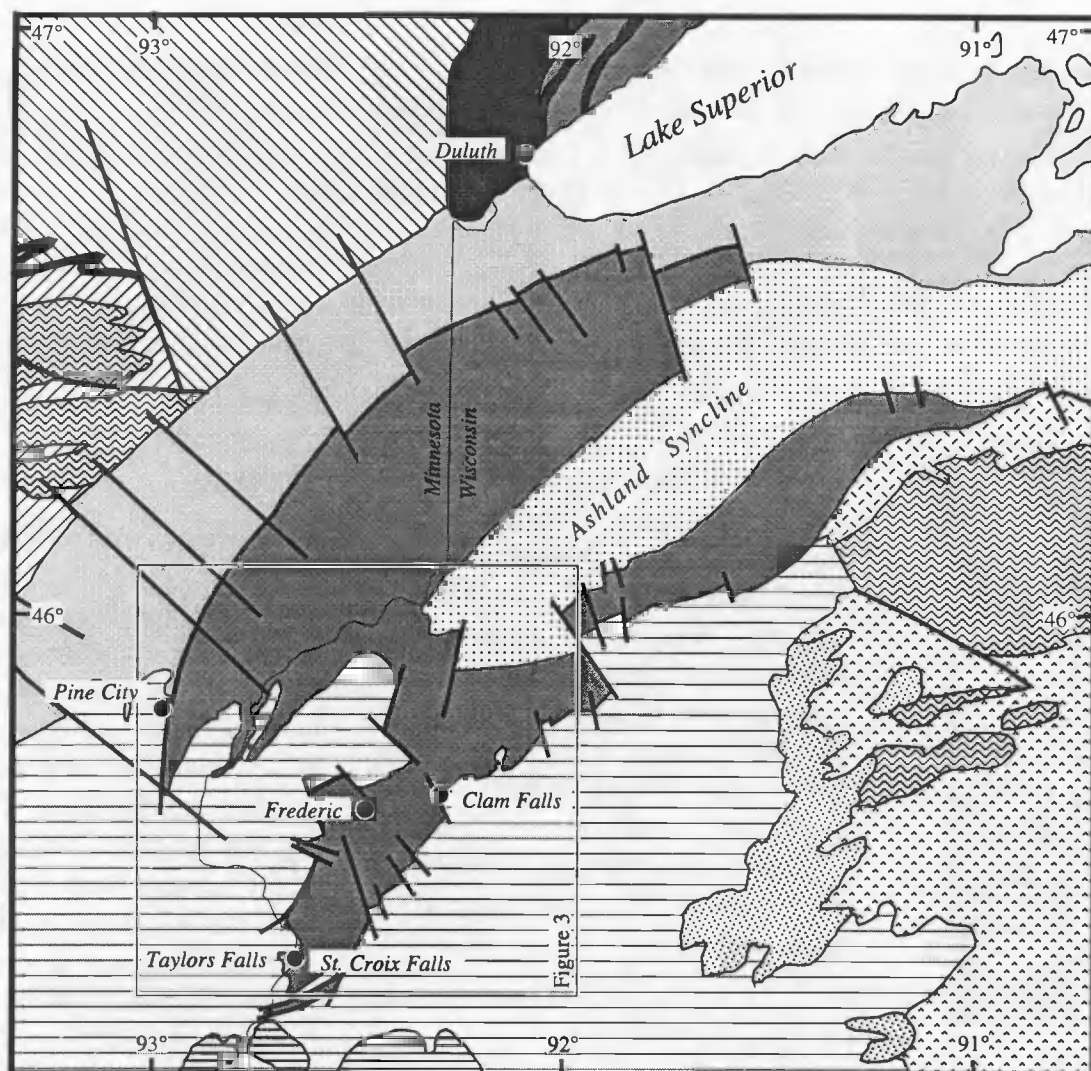
The voluminous volcanic rocks of the rift consist predominantly of tholeiitic basalt flows and minor rhyolite (Green, 1977; 1983) that were erupted during a relatively short time span (1110-1090 Ma; Davis and Green, 1997). Trace element and isotopic studies indicate that the basalts were derived from mantle melts that were variably contaminated by continental lithospheric mantle and crust (Nicholson and Shirey, 1990; Shirey et al., 1994; Nicholson et al., 1997). The combined petrologic, structural, and geophysical evidence are consistent with a "hot-spot" origin for the Midcontinent rift system (Hutchison et al., 1990).

SOUTHWESTERN ARM OF THE MIDCONTINENT RIFT SYSTEM

Most ground-based studies of the Midcontinent rift have been conducted in the Lake Superior region where Keweenaw volcanic and sedimentary rocks are well-exposed. Along much of the southeastern (Michigan) and southwestern (southern Minnesota to Kansas) extensions of the rift, these rocks are largely buried beneath thick Phanerozoic deposits. The objective of this field trip is to provide an overview of recent studies of the Midcontinent rift in east-central Minnesota and west-central Wisconsin. Despite the relatively poor exposure in this region, there are several compelling reasons for studying this portion of the rift. Most studies of the rift in the Lake Superior region have focused on the portion of the rift formed near the location of the inferred Keweenaw plume beneath northern Wisconsin (e.g., Peterman and Sims, 1988). New models of mantle plume processes (e.g., Campbell and Giffiths, 1990) have contributed significantly to our understanding of large igneous provinces and suggest that rift processes may vary with distance from the plume center. Unfortunately, the distal portions of the Midcontinent rift system are buried, making geologic study difficult and costly. The rocks of the Taylors Falls - Clam Falls region are the southernmost exposures of the Midcontinent rift system and provide an opportunity to examine variations in rift geology and processes along the length of the rift. A second reason for studying this segment of the rift relates to regional geologic provinces. The Midcontinent rift system crosses several major geologic provinces as it traverses from Minnesota to Kansas (Van Schmus, 1992). In northern Minnesota and Ontario the rift is superimposed on Late Archean granite-greenstone terranes (2.6-2.8 Ga) of the Superior Province. South of Lake Superior, the rift crosses a younger east-west belt of immature volcanic and sedimentary rocks (Southwick et al., 1989) that formed in a southward-dipping subduction zone during the Penokean orogeny (~1.85 Ga). Subduction processes during the Penokean may have modified the mantle beneath these terranes, and thus might have had an effect on the Keweenaw magmas erupted through them. Studies of Keweenaw volcanics south of the Malmo Structural Discontinuity (Minnesota) and Niagara Fault Zone (Wisconsin) provide an opportunity to study these effects.

PREVIOUS STUDIES

Studies of the Taylors Falls - Clam Falls region date back to the time before Minnesota became a state in 1858. A thorough review of the history of investigations is beyond the scope of this guide, and the interested reader is referred to the review by Craddock (1972). Since the early 1900's it has been recognized that the rocks of the Taylors Falls - Pine City region constitute a thick, fault-bounded volcanic and sedimentary sequence that has been broadly folded into a synform (Strong, 1880; Berkey, 1897, 1898; Hall, 1901). The term "Chengwatana" was first assigned by Hall in 1901 to refer to a sequence of flows and interflow sediments of "unusual extent" that are exposed along the Snake River east of Pine City (Fig. 1). As stated by Hall (1901) "The flood of 1898, which tore away the dam at the foot of Cross lake and poured down the river gorge a vast volume of water, cleaned out in an admirable manner for examination the channel of the stream for several miles." At the time, Hall (1901) was able to describe a nearly continuous sequence consisting of 65 lava flows and 5 interflow sedimentary units, as well as the relationships of the bounding Douglas fault. He also recognized the typical three-fold flow morphology and described the igneous and metamorphic minerals. Unfortunately, the rocks along the Snake River are not nearly so well-exposed today. Later, the term "Chengwatana volcanics" was used to refer to Keweenaw volcanic rocks exposed over a much broader region



(Modified from Morey et al, 1982)

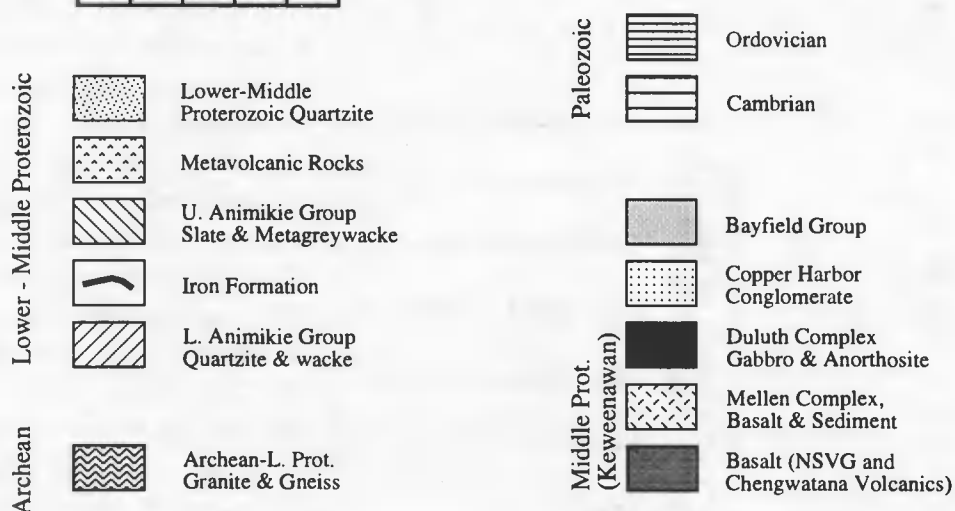


Figure 1. Generalized bedrock geology of the southern Lake Superior region showing the location of the field trip in the Taylors Falls - Clam Falls region. (Modified from Morey et al., 1982).

between the Douglas fault (west) and Lake Owen fault (east) in east-central Minnesota and west-central Wisconsin (Morey and Mudrey, 1972; Hassan and Mudrey, 1980).

The distribution and stratigraphy of the Chengwatana volcanics in the Taylors Falls - St. Croix Falls region was established by Cordua (1989a and b; unpublished mapping). He divided the nearly 3,000 meter thick volcanic section into seven flow units based on field and textural relationships (Fig. 2). The mapping by Cordua also delineated several east-west faults in the region. Nearly 3,000 meters of volcanic and sedimentary rock are also exposed in the Clam Falls - Frederic region, however corresponding mapping and stratigraphic studies have not been conducted there yet.

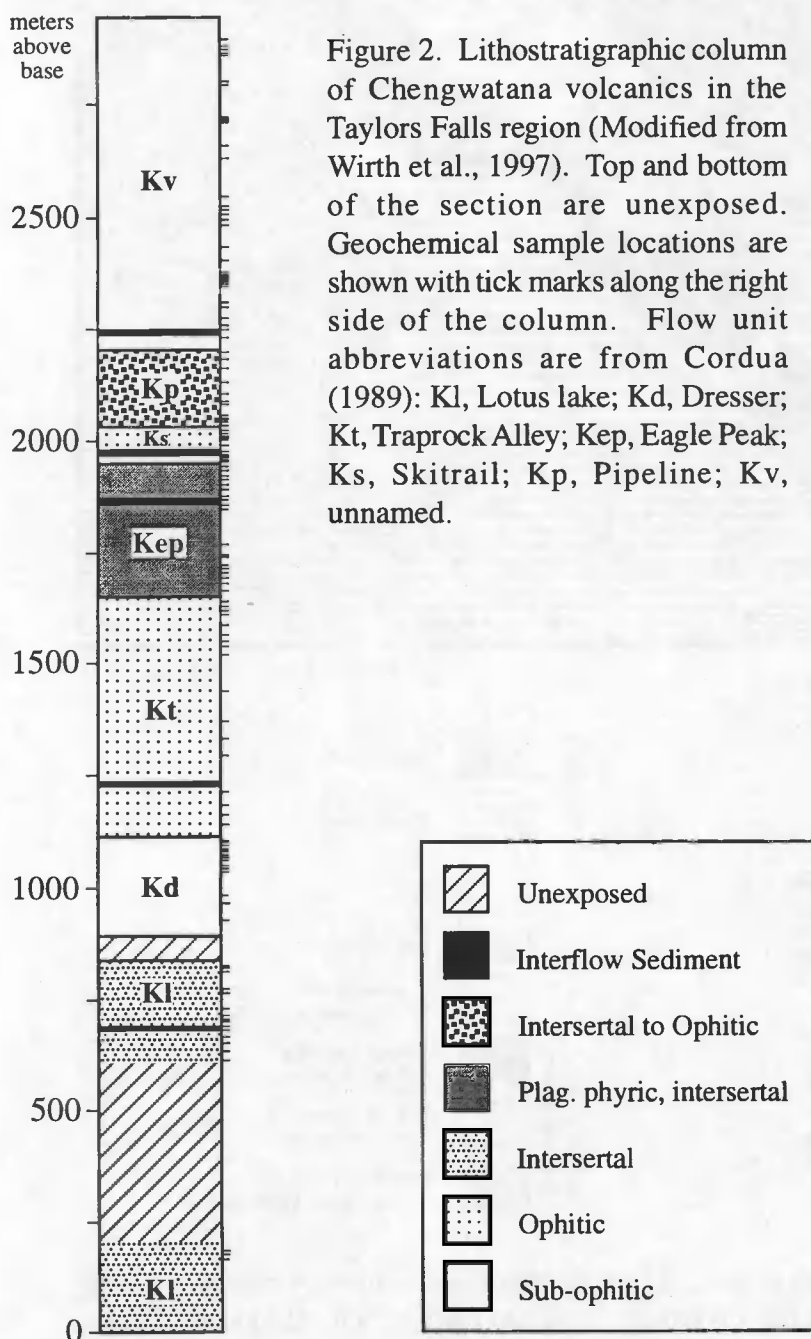


Figure 2. Lithostratigraphic column of Chengwatana volcanics in the Taylors Falls region (Modified from Wirth et al., 1997). Top and bottom of the section are unexposed. Geochemical sample locations are shown with tick marks along the right side of the column. Flow unit abbreviations are from Cordua (1989): Kl, Lotus lake; Kd, Dresser; Kt, Traprock Alley; Kep, Eagle Peak; Ks, Skitrail; Kp, Pipeline; Kv, unnamed.

GEOPHYSICAL STUDIES

Gravity, magnetic, and seismic studies were essential to developing detailed models of the buried parts of the Midcontinent rift system in Minnesota and Wisconsin (see review by Craddock, 1972). The regional magnetic character of the Keweenaw Rift in eastern Minnesota and western Wisconsin is given in Chandler et al. (1989) and Allen et al. (1997). The rocks of the Taylors Falls - Clam Falls region are part of the St. Croix horst, which is bounded by the Douglas (northwest) and Hastings (southeast) faults (Fig. 3). On the basis of three-dimensional gravity models, the thickness of Chengwatana volcanics in this region ranges from 4 kilometers at Clam Falls to 14 kilometers north of Frederic, and vertical uplift of the horst ranges from 3 kilometers in the northwest northwest to 8 kilometers in

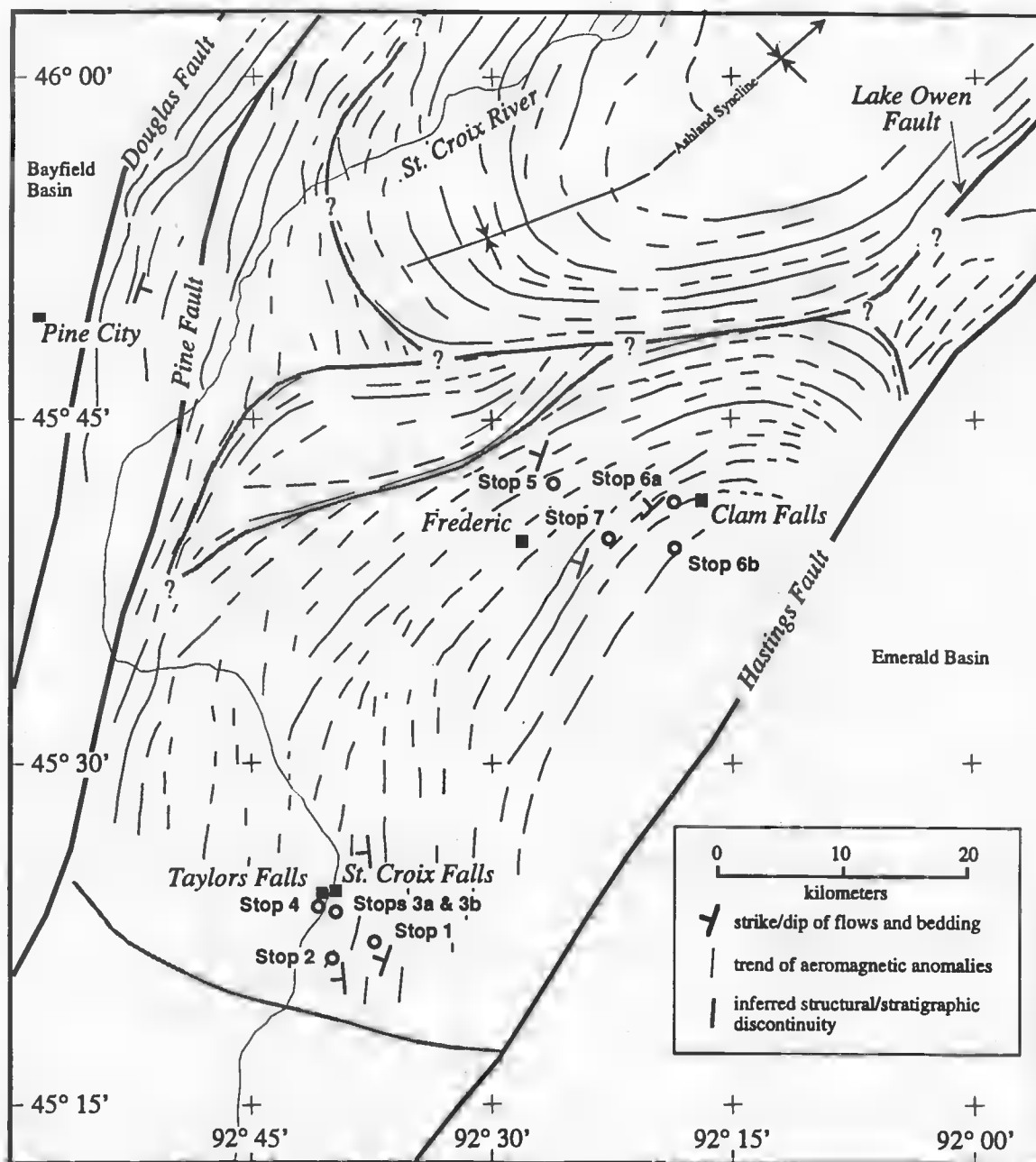


Figure 3. Sketch map of aeromagnetic anomalies in the Taylors Falls - Clam Falls region interpreted from new aeromagnetic maps of western Wisconsin (Cannon et al., 1997; in prep).

the southeast (Allen et al., 1997). Thick (~5 km), wedge-shaped accumulations of Keweenaw sediment (Oronto and Bayfield Groups) occur in basins on the margins of St. Croix Horst (Allen et al., 1997).

The magnetic character of Polk County is provided by Reich (1989), who conducted a ground magnetic survey at 0.1 mile intervals, transecting the county along seven primarily N-S and E-W lines (Fig. 4). Time-corrected raw data and 3-point smoothed data for line one are presented in Fig. 5. This line covers some of the outcrops seen on this field trip, and shows the strong magnetic signature of the Chengwatana flows that is about 1000 nanotesla (nT) above the regional value. The minimum in the magnetic line at approximately 15 kilometers from the

west end of the line is interpreted by Reich (1989) as the result of offset in the bedrock caused by the Cottage Grove fault. The high-frequency nature of the magnetic signal on the west end of the line is common to all the lines that cross outcrops of the flows. These high-frequency "spikes" may in part represent locations hit by lightning. Lightning strikes on highly magnetic rock can produce large localized anomalies. Many outcrops in this region are magnetically strong enough to significantly deflect a compass! Reich (1989) interpreted the variations in the filtered data as resulting from half-graben faults that parallel the Cottage Grove fault. Recent work by Cannon et al. (1997; in prep.) suggest that these features are the result of folding within the flows. Variations in the magnetic susceptibility of the different flows may also be a contributing factor. Recent studies by Kean (1992, 1993, 1995) and Kean et al. (1997) on the paleomagnetic character of Chengwatana flows in the Taylors Falls region are described in the section for Stop 2.

New aeromagnetic data from western Wisconsin (Cannon et al., 1997; in prep.) provide additional constraints on the stratigraphy and structure of the Chengwatana volcanics in the Taylors Falls - Clam Falls region. Previously, the relationship between the stratigraphic sections in the Taylors Falls and Clam Falls regions was not well understood. Preliminary interpretation of the data suggest that the stratigraphy continues north of the Taylors Falls region before trending

to the east around a northeast-southwest antiformal structure (Fig. 3). No major structural breaks are apparent. If this interpretation is correct, then the flows in the Frederic - Clam Falls region are stratigraphically below those of the Taylors Falls section. Also apparent on the aeromagnetic map are several structural or stratigraphic discontinuities surrounding the Ashland syncline, and northwest of Frederic. These relationships, combined with the results of recent geochemical and geochronologic studies, suggest that there may be significant differences between the flow sequences and that further stratigraphic subdivision of the Chengwatana volcanics may be warranted.

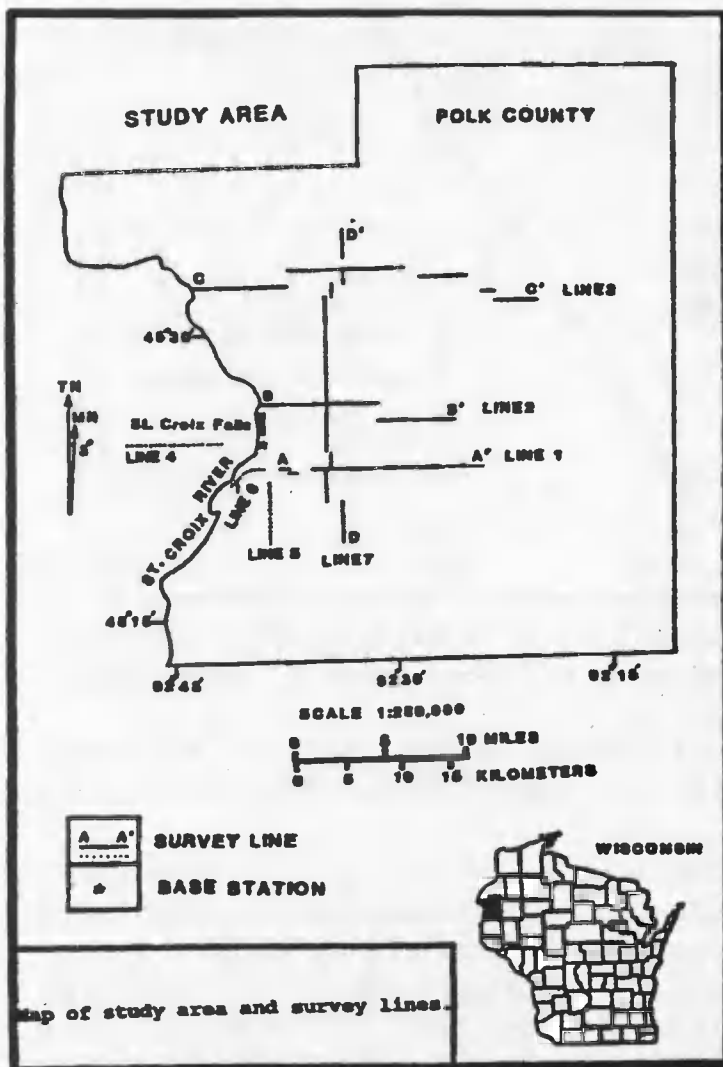


Figure 4. Location map for magnetic profiles of Polk Co., Wisconsin (after Reich, 1989).

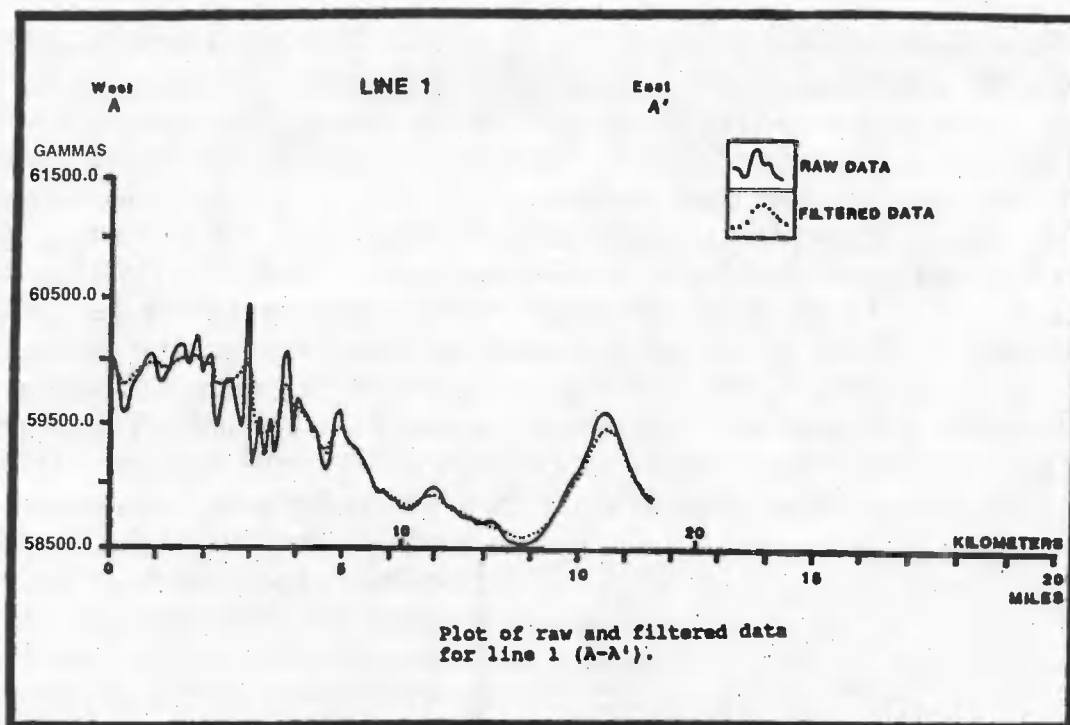


Figure 5. Magnetic profile of Line 1 in Figure 4 (after Reich, 1989).

METAMORPHISM AND MINERALIZATION

The secondary minerals in the Keweenawan volcanic rocks form by a combination of burial metamorphism, hydrothermal activity and contact metamorphism. They form 20-80 percent of a typical sample. They replace phenocrysts and groundmass, fill amygdules, form veins and line joint and fault planes. The minerals replacing ground mass and phenocrysts typically consist of the assemblage epidote+actinolite+chlorite+quartz+ albite +/- calcite +/- muscovite +/- titanite +/- Fe-Ti oxides. The common minerals filling amygdules are quartz, chlorite, epidote, K-feldspar and calcite. The quartz may be present as coarse crystalline material or, more rarely, as agate. Joint surfaces and fault planes may have veins or coatings of any of the above mentioned minerals, but may also show tremolite, and chrysotile. Minor copper mineralization in the form of native copper, chalcopyrite, bornite, chalcocite and malachite and azurite occurs in amygdules, flow-top breccias or later gash veins (Grant, 1901; Dutton, 1972; White, 1978; Cordua et al., 1979).

Epidote occurs as both granular and crystalline grains. Granular epidote is commonly observed in the groundmass or as alteration of the plagioclase, whereas crystalline epidote occurs within the amygdules. Actinolite is present mainly in the groundmass, but also replaces clinopyroxene and occurs in amygdules. Chlorite is ubiquitous in all of the flows and is present in two forms: ripidolite (brown birefringence) and pycnochlorite (anomalous blue birefringence). Although both types are present throughout the sequence of Chengwatana flows the Fe-poor variety (pycnchlorite) is most abundant in the uppermost flows, and the high-iron variety (ripidolite) is more common in amygdules (Wirth et al., 1996).

The pressure and temperature conditions during metamorphism in these rocks are constrained by mineral reactions (Liou et al., 1987) and by the Na content of the actinolite. The minimum temperature of metamorphism of the Chengwatana volcanics is constrained by the

presence of actinolite and epidote (Fig. 6) and by the absence of coexisting prehnite and pumpellyite. The upper temperature boundary of metamorphism is constrained by the lack of hornblende coexisting with epidote, chlorite, actinolite and quartz. The maximum pressure conditions of metamorphism (4-6 kbar) are constrained by the absence of glaucophane. Further constraints on the pressure conditions during metamorphism can be estimated from the Na and Al contents of the calcic amphiboles in the presence of quartz + albite + chlorite + magnetite (Brown, 1977). Analyses (SEM-EDS) of actinolite indicate that it contains Na (M-4 site) and tetrahedral Al (Al^{IV}). The results are similar to the composition of amphiboles found in the contact aureoles of the Sierra Nevada Batholith and suggest low pressure burial metamorphism of the Chengwatana Volcanic Group. This suggests that the presently exposed Chengwatana flows were overlain by an additional 6-7 kilometers of volcanic flows or sediment. This estimate is greater than the estimated burial depths made by Cordua (1980; 1989b) and Cavaleri (1987) by analogy with the Portage Lake Volcanic Group in the Keweenaw Peninsula, but are consistent

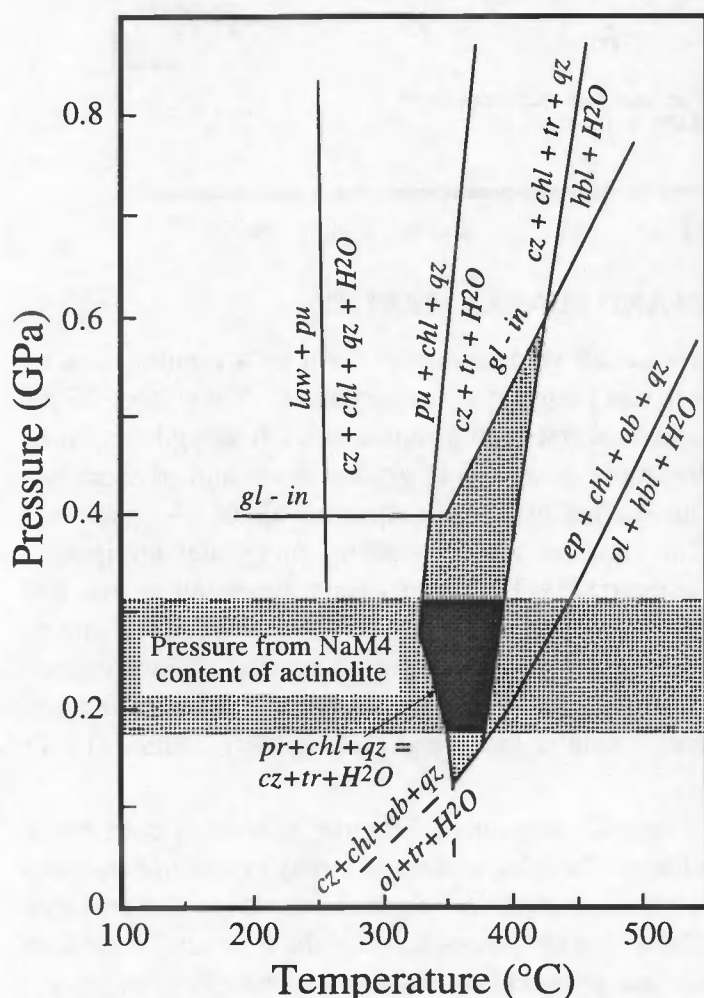


Figure 6. P-T diagram for univariant reactions relevant to low-grade metamorphism of mafic volcanic rocks (modified from Wirth et al., 1997). Estimated conditions of metamorphism of Chengwatana volcanics are shown in shaded regions.

with the amount of uplift (<8 km) estimated from geophysical models (Allen et al., 1997). Assuming a peak metamorphic temperature of approximately 350° C and a burial depth of approximately 7.5 km., the metamorphic data from the Chengwatana flows suggest a geothermal gradient of approximately 45-50° C within this portion of the midcontinental rift. Similar results were obtained by Livnat (1983) using oxygen isotopes in the Portage Lake volcanic sequence on the Keweenaw Peninsula.

The metamorphic minerals in Polk County are distinctly different from those seen further north in Wisconsin. In the volcanic rocks exposed in northern Douglas County (south shore of Lake Superior), prehnite, laumontite and other zeolites are common secondary minerals (Grant, 1901, Dutton, 1972, White, 1978; Cordua, 1990, 1991). In southern Douglas County, in the region between Dairyland to Solon Springs, prehnite is found, but the zeolites are rare to absent (Grant, 1901). To the east, in the Minong Copper Range, which stretches from Minong through Grandview in

Washburn, Bayfield and Douglas Counties, prehnite is common and pumpellyite has been reported. (Grant, 1901; Smith, 1947; Dutton, 1972; White, 1978; Ali, 1982). Zeolites such as natrolite occur along later joint surfaces near Chittamo and heulandite and natrolite are reported occurring in amygdules with quartz, epidote and chlorite near Grandview (LaPoint, 1976). South of Grandview, southeast of the Lake Owen fault, epidote, chlorite, actinolite, quartz, biotite and ferrostilpnomelane occur, with no prehnite, pumpellyite or zeolites (LePoint, 1976). A hornblende-epidote-biotite assemblage is likely due to the presence of nearby Mellen Intrusive Complex plutons (Cannon et al., 1996). Toward the Michigan border, at Saxon Falls near Hurley, prehnite and zeolites, such as thomsonite and laumontite, are again reported (Irving, 1883; Cannon, 1996).

The higher metamorphic grade of the Chengwatana basalts could reflect either deeper burial in the volcanic pile than rocks to the north, or the presence of a higher geothermal gradient in that part of the Keweenawan terrain. Allen et al. (1997) estimate that the St. Croix Horst has undergone 3-8 kilometers of uplift in the Taylors Falls-Clam Falls region. This contrasts with estimated uplift of 2-4 kilometers in the Ashland syncline and 1-2 kilometers in the Twin Cities Basin (SE Minnesota).

GEOCHEMISTRY

Petrographic and geochemical studies of the Chengwatana volcanics in the Taylors Falls - Clam Falls region have been conducted by Cordua (1980; 1989b), Cavaleri (1987), Leslie et al. (1994), Wirth et al. (1995; 1996; 1997), Naiman et al. (1996; 1997; in press), and Abbott et al. (in press). These studies established the tholeiitic, high-alumina and high-iron character of the basalts in the region (Table 1; Fig. 7); intermediate and felsic rocks, with the exception of rare rhyolite near Clam Falls, are not observed. Most flows are weakly olivine-normative (0-15 percent norm ol) and the remaining flows have trace amounts of quartz in their norms (0-5 percent norm qz).

Nickel abundances (36 - 185 ppm) and Mg#'s ($Mg\# = Mg/[Mg+Fe^{+2}] = 0.37 - 0.58$) are generally low, suggesting that the Chengwatana Volcanic Group flows have undergone significant fractionation. Chengwatana basalts can be subdivided on the basis of TiO_2 (less than and greater than 2.3 weight %), a distinction that has been recognized elsewhere in the Midcontinent rift. In general, the high- TiO_2 basalts are more abundant in the upper part of the St. Croix Falls section and are characterized by having generally low Mg# (<0.5) and high

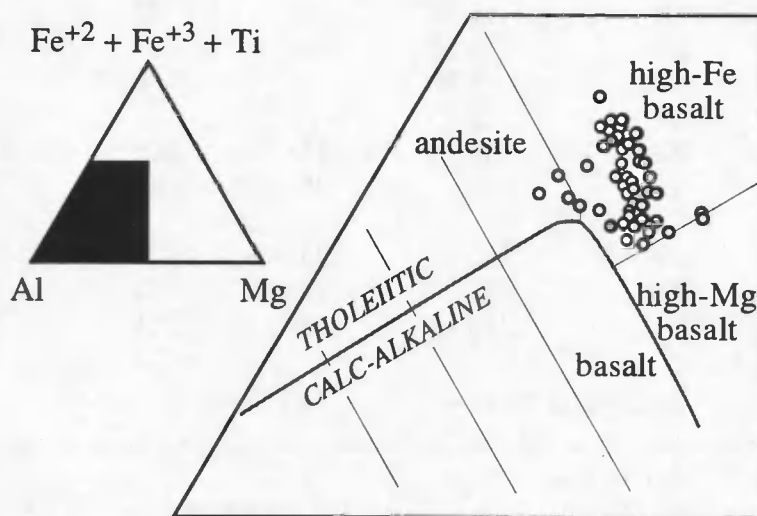


Figure 7. Plot of Chengwatana volcanics from the Taylors Falls region on a Jensen cation plot (Jensen, 1976) (after Wirth et al., 1997).

Table 1. Representative analyses of basalt and sediment from the Taylors Falls and Clam Falls region.

	basalt KC-1b	basalt KC-7	basalt KC-70	sediment KC-330g	sediment KC-331f
SiO ₂	47.60	48.30	48.13	64.75	59.64
TiO ₂	1.69	1.71	2.64	0.93	0.84
Al ₂ O ₃	15.57	15.65	14.07	11.11	13.55
Fe ₂ O ₃	13.34	13.61	16.40	7.37	8.75
MnO	0.19	0.19	0.21	0.08	0.10
MgO	7.48	6.69	5.17	0.97	2.66
CaO	8.14	9.56	8.41	11.92	11.55
Na ₂ O	2.38	2.60	2.45	0.21	1.28
K ₂ O	1.13	0.55	1.43	0.73	0.61
P ₂ O ₅	0.16	0.19	0.31	0.10	0.13
LOI	3.09	1.63	1.25	2.02	1.59
Total	100.77	100.68	100.47	100.19	100.70
Mg#	0.57	0.43	0.42	0.23	0.48
<i>qz</i>	—	—	—	39.54	0.57
<i>ne</i>	—	—	—	—	—
<i>hy</i>	17.73	17.01	20.38	—	26.68
<i>ol</i>	9.24	6.41	2.66	—	—
Rb	42	14	48	16	15
Ba	206	277	546	206	128
Th	1.03	—	—	—	—
Nb	11	9	13	9	11
La	9.58	—	—	—	—
Sm	3.50	—	6.94	—	—
Yb	2.15	—	—	—	—
Zr	108	121	209	122	23
Y	23	26	42	19	142
Cr	165	125	128	79	105
Ni	185	124	82	67	50
V	248	254	318	169	161
Initial ϵ_{Nd}	+0.1—	-1.7 —	-1.86		

Sample Locations:

KC-1b	-	NE1/4, NW1/4, section 5, T 33 N, R 18 W
KC-7	-	SE 1/4, NE 1/4, section 6, T 33 N, R 18 W (Stop 1)
KC-70	-	NW 1/4, NW 1/4, section 14, T 33 N, R 19 W
KC-330g	-	NE1/4, NE1/4, section 6, T 33 N, R 18 W
KC-310	-	NE1/4, SW1/4 of section 11, T 37 N, R 17 W (Stop 5)

incompatible trace-element abundances. REE patterns of the Taylors Falls basalts are sub-parallel and variably enriched in the light REE, with abundance levels 40-100 times chondritic values for La, and from 12-22 times chondrites for Yb. Most samples have prominent negative Eu anomalies, a feature that is consistent with the widespread petrographic evidence of plagioclase fractionation. Normalized trace element abundances (Fig. 8) of the flows are enriched in the highly incompatible elements (e.g., Th, Nb, La) relative to the less incompatible trace elements

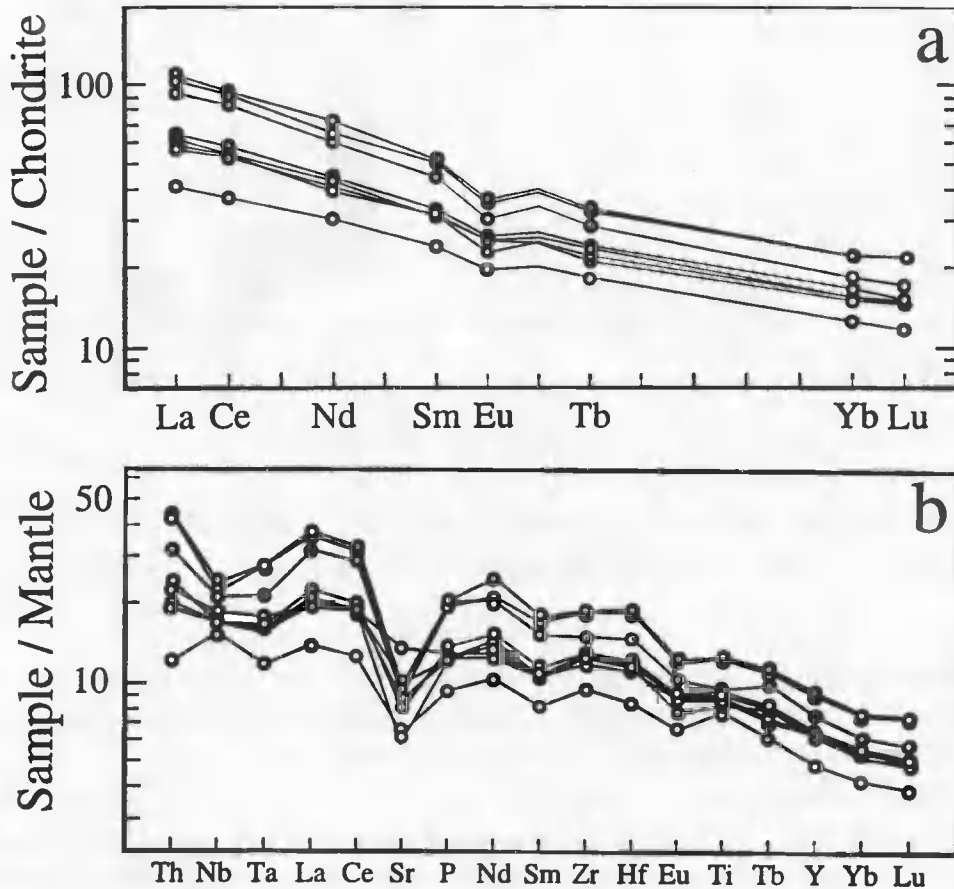


Figure 8. (a) Chondrite-normalized plot of REE abundances in representative Chengwatana volcanics from the Taylors Falls region (after Wirth et al., 1997). (b) Plot of trace-element abundances normalized to primitive mantle.

(e.g., Zr, Y, Yb); Ta, Nb, and Sr are relatively depleted. These relationships suggest that at least some of the major and trace element variations exhibited by the Chengwatana volcanics are a result of fractional crystallization and variable contamination. Basaltic flows of the Chengwatana volcanics exhibit a relatively narrow range of Nd isotopic compositions (Fig. 9). Although initial ϵ_{Nd} values of the flows range from +3.4 to -4.5, the majority of samples plot within a more limited range (-1.5 and -3.0). The combined geochemical data suggest that the basalts originated from plume-derived melts that were variably contaminated by continental lithospheric mantle or crust (Wirth et al., 1997). Rhyolites from the Clam Falls region has initial ϵ_{Nd} values of +1.0 and -0.1. These values are close to the composition of the inferred plume (initial ϵ_{Nd} value = 0) implying that the rhyolites formed by crystal fractionation (Abbott et al., in press).

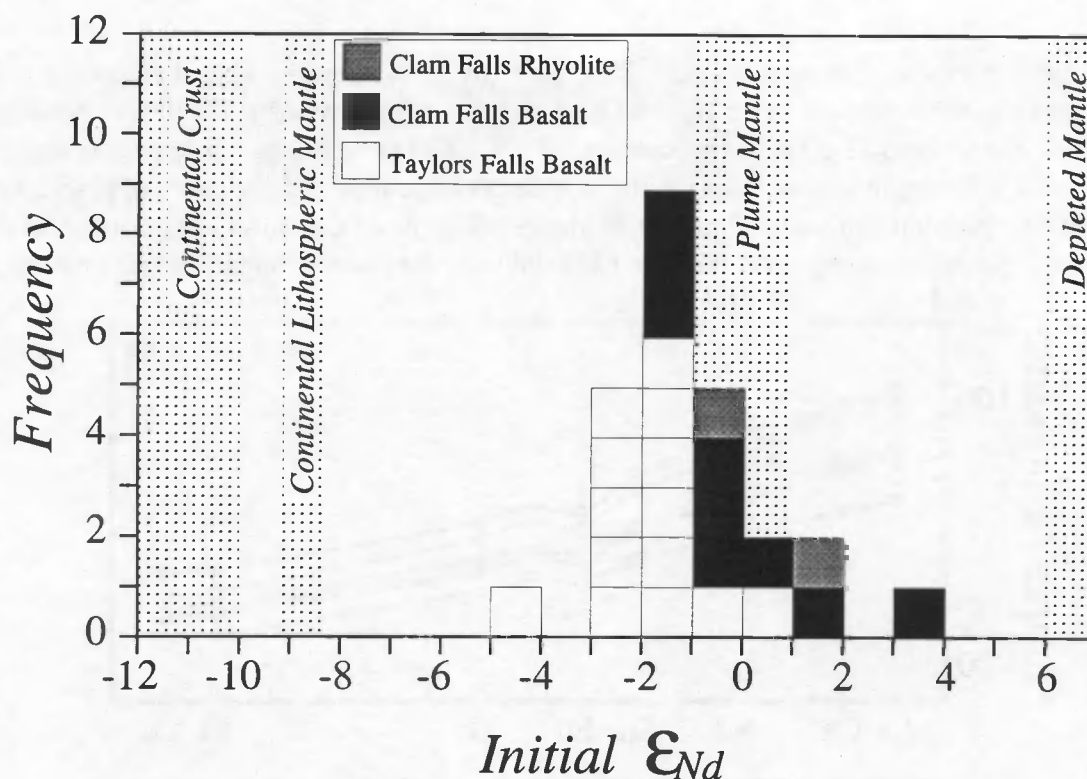


Figure 9. Histogram of initial ϵ_{Nd} values (1100 Ma) of Chengwatana volcanics from the Taylors Falls and Clam Falls regions (after Naiman et al., in press).

ITINERARY

This guide describes a one-day trip beginning and ending in Minneapolis, Minnesota. A regional map of the route is given in Figure 10, and more detailed maps of the stops are given in Figures 11 and 12. A brief description of the route is given below:

Proceed from Minneapolis about 45 miles to Osceola Wisconsin. One route to Osceola would be as follows: Take I-35 north to the exit for Minnesota 97 east bound. Follow Route 97 for 11 miles to its intersection with Minnesota 95. Turn left (north) on Minnesota 95 for 6 miles. Turn right on route 243, cross the St. Croix River and follow it to its intersection with Wisconsin Route 35.

Go north on Route 35. In about 4.5 miles (about 1 mile north of Dresser Wisconsin) turn right on Ravine Drive. Follow Ravine Drive a short distance to the entrance of the Dresser Trap Rock Company Quarry (Stop 1; Fig. 13).

Return to Route 35. Turn left (south). Go 1 mile to State Street. Turn right (west) on State Street. This soon becomes 100th Avenue. Follow this road about 3 miles to the intersection with County S. Turn north and go about 1/2 mile to the pipeline crossing of Count S (Stop 2; Fig. 15).

Return to Route 35. Turn left (north). Go three miles to the city limits of St. Croix Falls. Turn left into Interstate Park, Wisconsin side (fee area). Follow the park road to Eagle Peak and the group camp area (Stops 3a and 3b; Fig. 17).

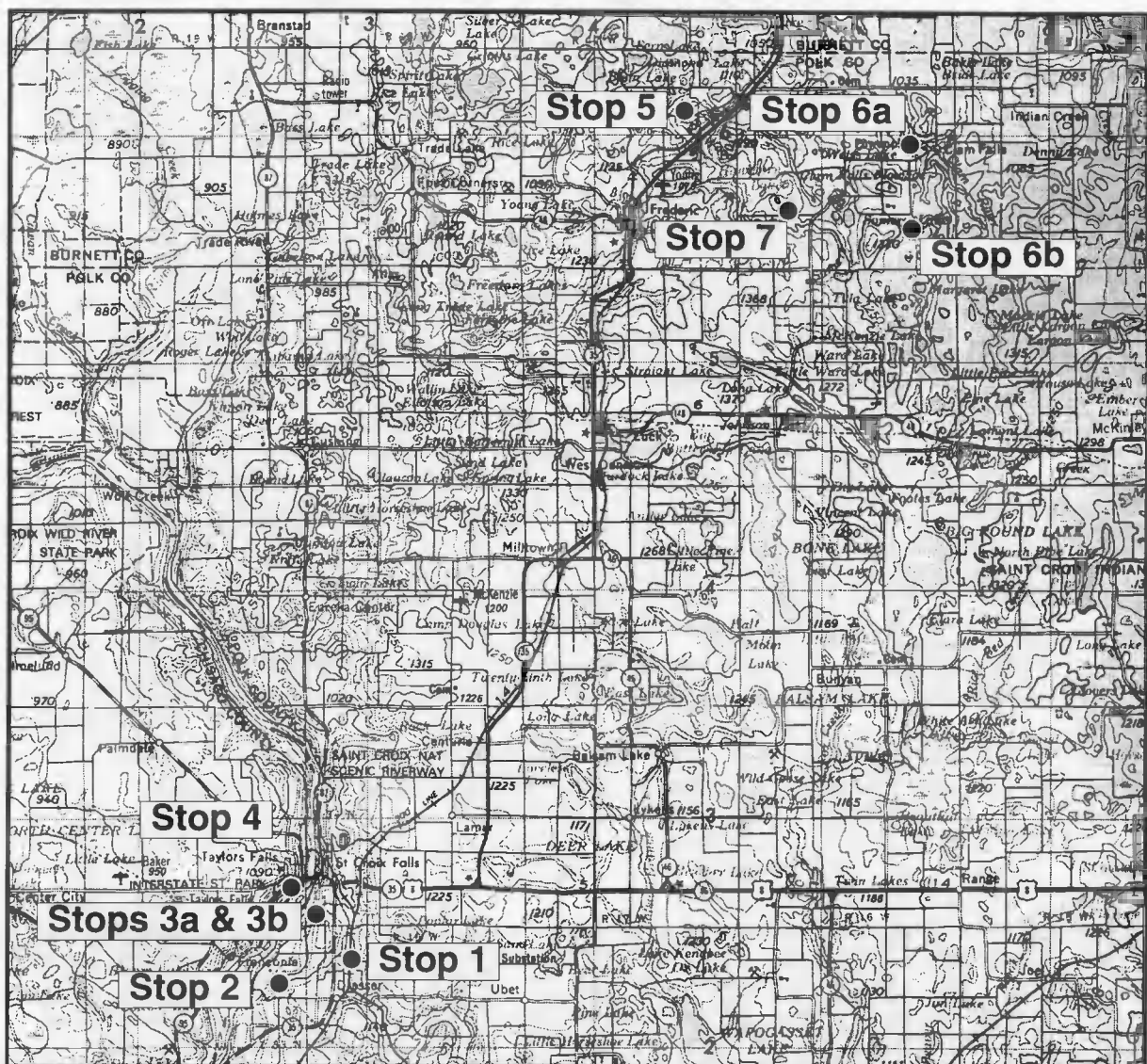


Figure 10. Scanned image (not shown to scale) of USGS Stillwater, Minnesota - Wisconsin topographic quadrangle (1 x 2 degree series; 1:250,000 scale) showing the locations of the field trip stops.

Return to Route 35. Turn left. Follow the signs to merge onto U.S. Highway 8 - WEST bound. Cross the St. Croix River into Minnesota. Immediately after the bridge turn left at the traffic light into Interstate Park - Minnesota side (a different fee area) (Stop 4; Fig. 17).

Return to Highway 8. Turn right (east). Recross the St. Croix River. Follow U.S. 8 for 5 miles to the exit for Route 35 north. Follow Route 35 north about 25 miles. About 2.5 miles north of Frederic, turn left onto 130th Street. In about 0.5 miles turn left on 345th Street. In 0.7 miles, look for an outcrop in the field to the north (Stop 5; Fig. 19).

Return east on 345th Street to Lewis. Cross Route 35 and continue east on County Road E for 5 miles; turn south on 65th Street to the town of Clam Falls. Turn west at the junction with County Road I. Stop 5 is NNE of the junction with 73rd Street. (Stop 6; Fig. 20).

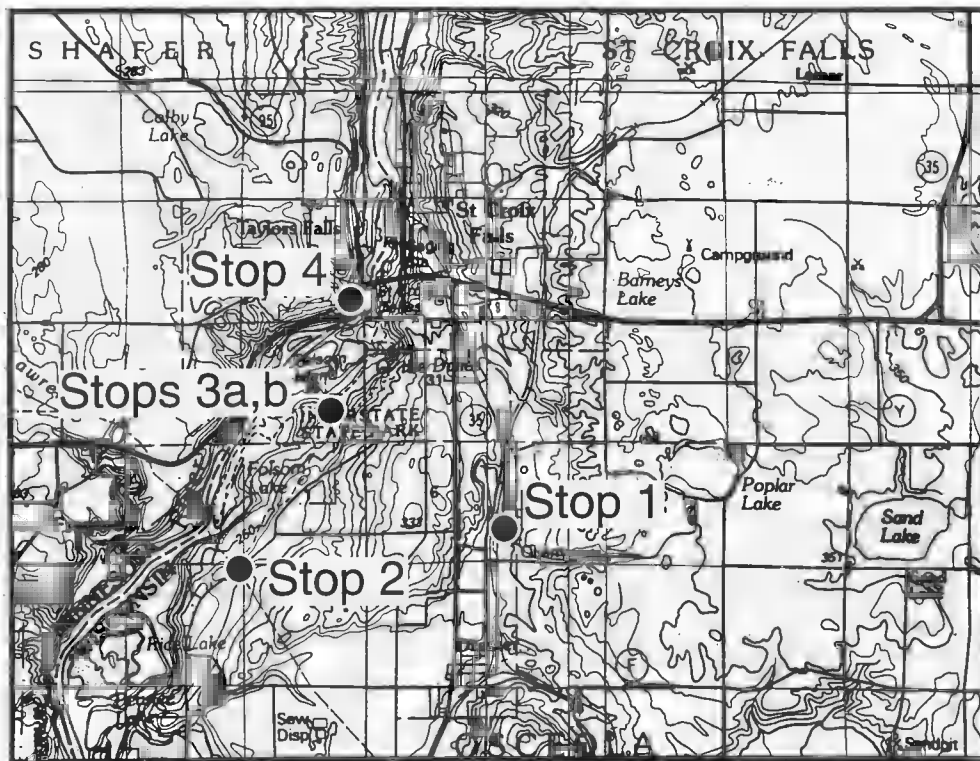


Figure 11. Scanned image of a portion of the USGS Stillwater, Minnesota-Wisconsin topographic quadrangle (30 x 60-minute series; 1:100,000 scale)

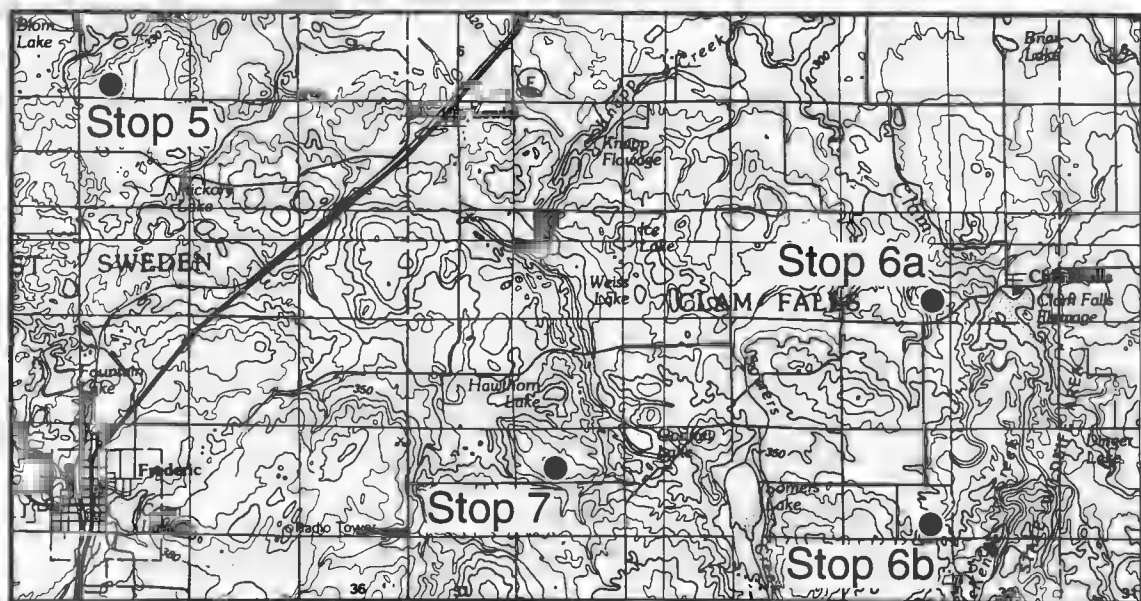


Figure 12. Scanned image (not shown to scale) of a portion of the USGS Grantsburg, Wisconsin - Minnesota topographic quadrangle (30 x 60-minute series; 1:100,000 scale) showing the locations of the field trip stops 5-7.

Alternate Stop 6 is reached from stop 6 by continuing south on 73rd to the junction with 315th Ave. Continue west on 315th Ave to McKenzie Trail. Follow McKenzie Trail south and around the bend to the east. The road veers south around an outcrop on the north side of the road (Stop 6b; Fig. 23).

Return to 315th Ave and continue west to 90th Street along the east side of Sommers Lake. Turn north and continue on 90th Street to the junction with 320th Ave. Continue west on 320th and on County Road I until the junction with 315th Ave. Continue west on 315th Ave. Outcrops on north side of road near top of hill (Stop 7; Fig. 24).

Return to Frederic and Route 35 by travelling west on 315th Ave., south on 120th Ave., and west on 310th Ave. A simple way to return to Minneapolis is to follow Route 35 south to U.S. Highway 8. Follow U.S. 8 east to its intersection with I-35. I-35 southbound leads directly to the Twin Cities of Minneapolis and St. Paul.

Stop 1—Chengwatana Lava Flows in the Dresser Trap Rock Quarries.

Location: Approximately one mile NNE of Dresser, Wisconsin. North Quarry - NW 1/4 Sec. 5 and NE 1/4 Sec. 6, T 33 N, R 18 W. East Quarries - SW 1/4 Sec. 5, T 33 N, R 18 W.

The lava flows at Dresser have been mined for “trap rock” since 1855. The large quarry complex, owned by Dresser Trap Rock Inc., currently produces 750,000 tons per year in crushed stone of all sizes. The rock is extremely tough and durable, due in large part to the interlocking of microscopic unoriented crystals of chlorite and actinolite. All visitors are urged to seek permission to enter, and to use fundamental rules of caution. The rock walls are shear, blasting is frequent, rock piles are unstable, and quarry truck drivers expect you to yield right-of-way. We will visit several spots in the quarry, the exact localities dependent on the areas accessible at the time of the field trip.

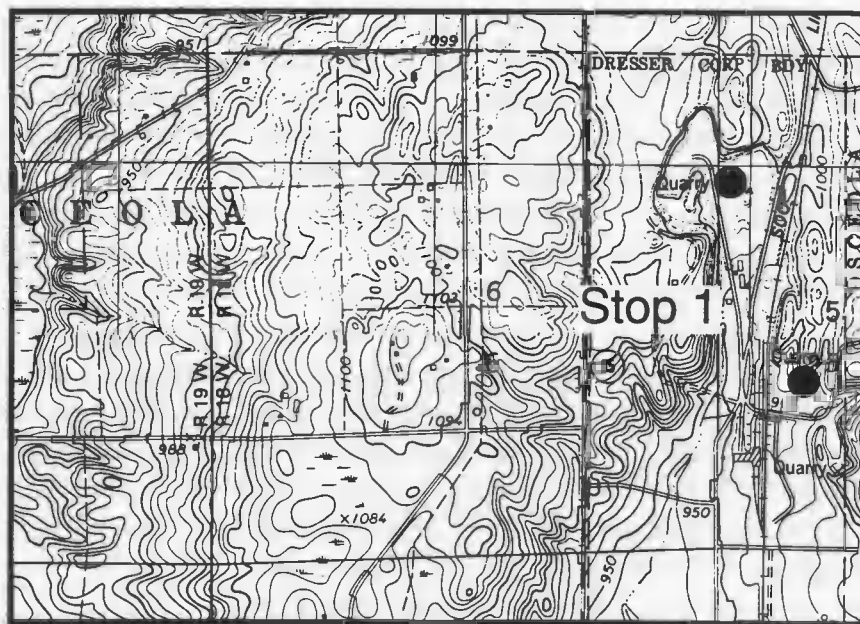


Figure 13. Scanned image of portions of the USGS Osceola, WI-MN and St. Croix Dalles, WI-MN topographic quadrangles (7.5-minute series; 1:24,000 scale) showing the location of Stop 1.

Both quarries expose basaltic lava flows that have been recrystallized incipiently to completely to greenschist facies minerals by static burial metamorphism, due to burial within the volcanic pile.

Geology of the north quarry

The north quarry is a large quarry cut into a single lava flow over 35 meters thick, which is informally named the Trap Rock Alley flow. The flow is a generally subophitic to diabasic porphyritic basalt. The majority of the phenocrysts are plagioclase. Augite shows up as lighter gray patches on weathered surfaces, giving the rock a distinctly mottled appearance (Fig. 14 a and b). Other primary igneous minerals are plagioclase and ilmenite. In places these minerals are relatively fresh, but usually they are incipiently to completely replaced by a mixture of chlorite, epidote, actinolite, albite, zoisite and titanite. The groundmass usually consists entirely of this mix of secondary minerals.

Locally, bodies of coarse gabbro are present as sill-like masses. Particularly common on the lowest level of the quarry, they average 0.5-1.0 meters in thickness, and consist of coarse augite, plagioclase and magnetite. Elongated augite crystals may be up to 4 cm long. Secondary minerals, such as those found in the surrounding basalt, are commonly developed at the expense of the primary minerals. Angular vesicles occur in these gabbros, and are filled with euhedral quartz, epidote and felted masses of chlorite. These gabbros are interpreted as lenses of slowly cooled magma trapped within this exceptionally thick flow, rather than as later intrusions. The lack of chilled margins in the gabbro against the basalt support this conclusion.

Flow-top features are common along the west wall of the quarry. Here an interflow zone is found, containing clasts of fine-grained and extremely amygdaloidal basalt. The clasts are angular and up to 20 cm in diameter. Interstitial to these clasts is a fine-grained matrix consisting of angular quartz and feldspar. This may be an interflow sediment consisting of wind-blown scoriaceous flow-top material or perhaps pyroclastic debris from a nearby cinder cone. Burial metamorphism has resulted in the recrystallization of the matrix and growth of secondary minerals, especially quartz, epidote, and chlorite. Pay close attention to the sedimentary features at this stop for later comparison with sediments exposed at Stop 5.

Amygdules are ubiquitous throughout the flow, but increase in size and number toward the top of the flow. The minerals commonly found filling the amygdules are quartz, chlorite, epidote, K-feldspar, calcite and actinolite. The amygdules are usually zoned, with chlorite forming the rinds. Growing on these rinds are inward projecting crystals of quartz, epidote and pink perthitic K-feldspar. Calcite, or a second generation of chlorite, occupies the centers of the amygdules. Actinolite occurs as microscopic fibers projecting from the groundmass into the amygdules and intergrown with the other amygdular minerals.

Another group of minerals form veins or occur along joint surfaces within the basalt. The most prominent veins are a series of quartz veins formed along reverse faults. These veins range from 10-40 cm in thickness and are surrounded by zones of particularly extensive chloritization and epidotization of the basalt. The quartz is coarsely crystalline and milky. At times the veins open up into vugs containing euhedral quartz crystals up to 15 cm long. Locally calcite occurs on the quartz as yellow-green flattened rhombohedral or globular clusters of radiating crystals. Pink to white K-feldspar druses, radial groups of dark-green chlorite, and copper minerals such as chalcocite, cuprite and malachite also occur from time to time in the veins.

Joint surfaces may be coated by a variety of minerals, including chlorite, hematite, tremolite (growing in fibers up to 3 cm long), white cleavable calcite, chrysotile, chalcopyrite, bornite, malachite, and native copper.

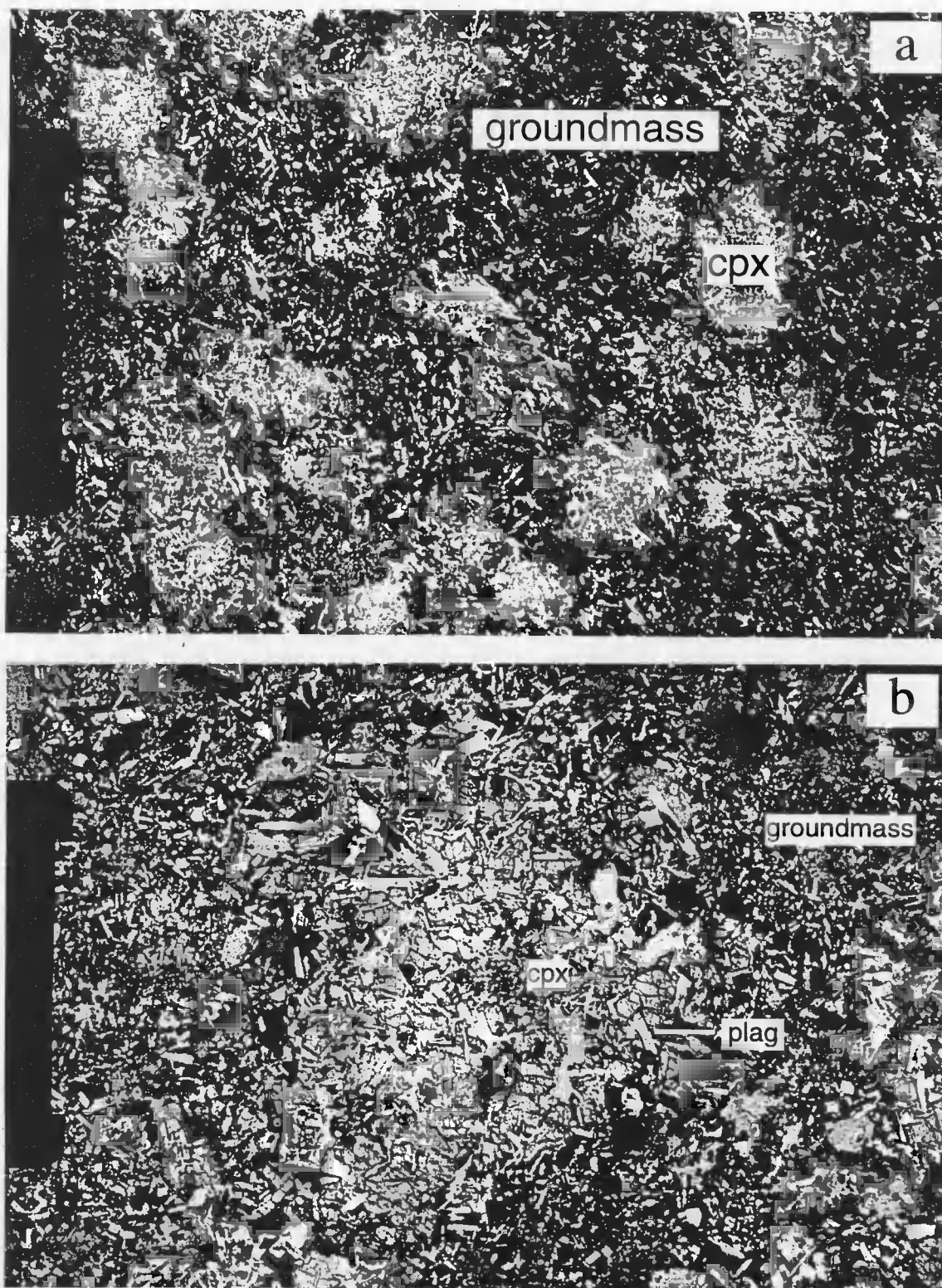


Figure 14. (a) Photomicrograph (crossed polars) of "ophimottled" basalt (scale bar along left side is 5 mm long). (b) Photomicrograph (uncrossed polars) of augite exhibiting ophitic texture (scale bar along left side is 2.5 mm long). Groundmass in (a) and (b) contains plagioclase microcrysts, Fe-Ti oxides, chlorite, quartz, epidote, actinolite, and calcite.

Nuggets of native copper massing up to several kilograms have been found in the quarry, but the author has not seen any in place. These reportedly are most common when quarry operations are active in the flow-top area in the northwestern corner of the quarry. These can have disastrous effects on the crusher, and are thus not viewed as a resource by the quarry owner.

The flows of the north quarry are cut by numerous faults. Numerous east-dipping (45°) thrust faults are cross-cut by a west-dipping (35°) thrust fault that trends north-south across the center of the quarry (Leslie et al., 1994). Along most of the west-dipping fault surface is massive and milky-white quartz. Also found within the gouge zone are epidote-filled en-echelon fractures, striated and stepped chlorite-quartz-epidote surfaces, and late-stage syntaxial calcite fillings. All of these kinematic indicators indicate that the hanging wall moved up and east relative to the footwall. Motion along this surface offsets other thrust faults in the quarry that dip to the east; the magnitude of the motion is not known but is likely minor. The age of faulting is not known, but the similar metamorphic minerals found in both the host rock and the gouge zone suggest that motion occurred while the rocks were still relatively deeply buried.

Geology of the east quarries

Approximately 0.5 km SSE of the north quarry, across the railroad tracks, is another quarry complex. These quarries expose two flows underlying the Trap Rock Alley flow seen in the north quarry. These flows are informally named the Dresser flows. Ten to twelve meters of each flow are exposed. The flows dip westerly at about 10 degrees. The contact between the two flows is well exposed.

Both flows are subophitic diabase, and like the Trap Rock Alley flow, have been converted to greenstones during burial metamorphism. The lower flow contains a primary igneous assemblage of augite, plagioclase and ilmenite. The rock is microporphyritic. The groundmass and phenocrysts are incipiently to completely replaced by a mixture of chlorite, epidote, titanite, actinolite, albite and zoisite. The same mineral assemblage, plus white quartz and pink K Feldspar occur in the amygdules. The top of this flow is well exposed, and has been virtually entirely recrystallized to a metamorphic greenstone. Large amygdules are abundant, and contain an attractive coarse-grained assemblage of white quartz, pink K-feldspar, red hematite, yellow-green epidote, and dark green chlorite. Only minor flow-top breccia can be seen.

The upper flow has a fine-grained chilled base grading upward into a diabasic subophitic interior. The primary igneous and secondary minerals are the same as that of the lower flow. The upper flow contains fewer amygdules, and these are dominated by nearly isotropic chlorite with minor intergrown epidote and quartz. The top of this flow is not exposed.

Stop 2—Magnetic Characteristics of the Chengwatana Lava Flows.

Location: Outcrops are located on both sides of the road near the intersection between the pipeline and County Road S. NE 1/4, NE 1/4, Sec. 11, T 33 N, R 19 W.

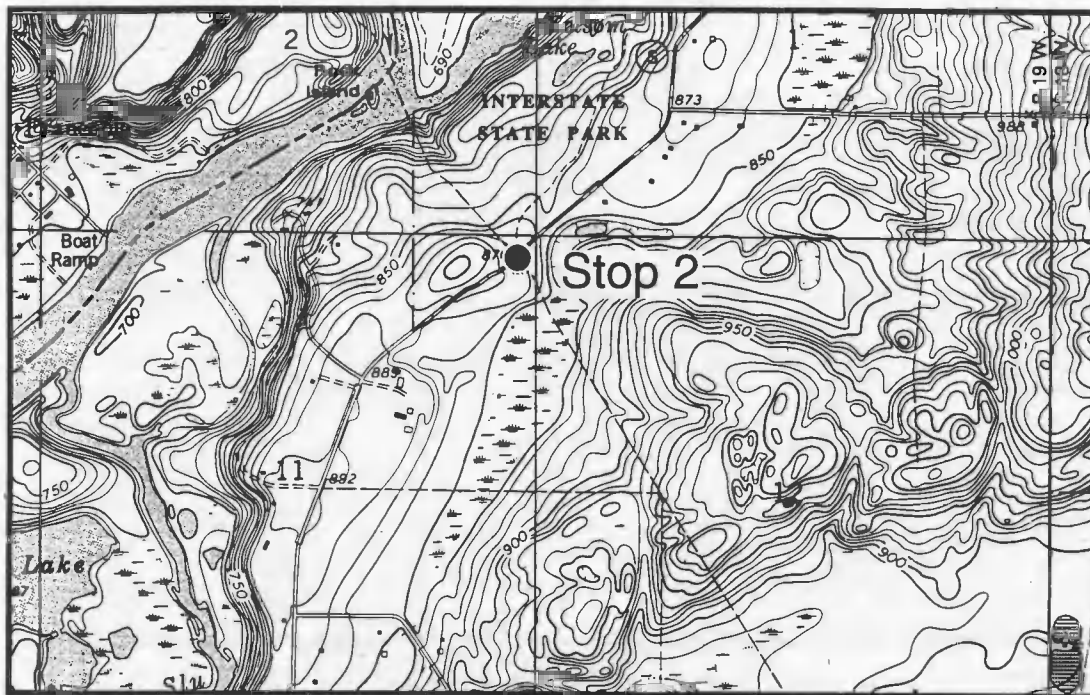


Figure 15. Scanned image of a portion of the USGS Osceola, WI-MN topographic quadrangle (7.5-minute series; 1:24,000 scale) showing the location of Stop 2.

At this stop we will examine and discuss the magnetic characteristics of the Chengwatana lava flows as exposed in Polk County, Wisconsin. Paleomagnetic studies have been conducted in this region over the past 10 years to better understand the stratigraphic relationship of these flows within the context of the regional Keweenawan magmatic sequence.

Detailed paleomagnetic studies for this region are reported by Kean et al. (1997). Here we concentrate only on the general interpretation of the paleomagnetic results that help to place the Chengwatana flows in the context of the other Keweenawan rocks in the Lake Superior region. Keweenawan paleopole positions for the craton are well known and a major polarity transition from reverse to normal occurred during the middle Keweenawan (1105-1102 Ma) has been documented by Palmer (1970) and Halls and Pesonen (1982). Table 2 shows this polarity change as determined in some of the major basalt flows of the southern Lake Superior region. Feeney (1990) determined that the lava flows throughout Polk Co. have predominantly normal polarities, with a mean direction of $D=275^\circ$, $I=41.5^\circ$, $\alpha-95=15.5^\circ$; which is consistent with other Keweenawan directions (Table 2) (Halls and Pesonen, 1982) such as the North Shore Volcanic Group (Books, 1972). He also noted several locations with reversed directions. Insufficient geologic control north of the city of St. Croix Falls rendered a more precise correlation impossible for those locations. However, the volcanics south of St. Croix Falls are better exposed and have been mapped in detail by Cordua (1989a and b) which allowed for a more detailed study. Sixty-five oriented cores were collected at the 6 sites noted in Figure 16. Sampling was

Table 2. Magnetic polarity and directions for many Keweenawan rocks in the Lake Superior region (from Halls and Pesonen, 1982).

Formation	Declination	Inclination	Apha 95	Pole Position	Polarity
Jacobsville Sandstone	263	-11	6	183°E, -0.9°S	N
Freda Sandstone	271.3	0.7	2.8	179.5°E, 1.2°N	N
Nonesuch Shale	279.8	9.8	6	176.5°E, 10.3°N	N
Portage Lake Volc (Avg)	288.6	34.9	2.5	181.2°E, 26.5°N	N
Upper North Shore Vol. (Avg)	290.3	44.5	4.7	183.8°E, 32.4°N	N
Lower North Shore Vol. (Avg)	120.9	-63	7.8	197.8°E, 9.6°N	N
Logan Sills	111	-72	2.5	219°E, 49.4°N	N
Upper Powder Mill Gr.	74	-71.5	4.8	232.4°E, 28.9°N	N

concentrated in the areas where Feeney (1990) found reversed directions (sites 1 and 2), and which are considered stratigraphically equivalent. Most of the samples (Table 3) have high intensities of natural remanent magnetism (NRM) (1.0-2.0 A/m) and volume magnetic susceptibility (4×10^{-2} SI). The magnetic properties differ somewhat between flows. The middle flows (Pothole trail, Trap Rock Alley and Dresser flows) have remanent intensities and magnetic susceptibilities which are at least an order of magnitude less than the other flows (Table 3). For samples with very strong remanent values, the intensities are typically reduced 100 fold with A.F. demagnetization to 20 millitesla (mT). Petrographic studies of the opaque minerals in samples from each site also show differences which are consistent with the bulk magnetic parameters of susceptibility and intensity. The upper lava flows (sites 1-4; Fig. 16) have a high percentage of coarse-grained magnetic minerals (25 percent), which are generally more altered than the older flows (sites 5 and 6), and which appear to have been produced during metamorphism. Samples from sites 5 and 6 have only about 10 percent magnetite grains and the grains are smaller than the upper flows. Principal component analysis of the magnetic directions (Kirschvink, 1980) for samples from the various sites reveals two directions that are approximately antiparallel to each other. The most prominent direction, after removal of a viscous component is a northwesterly declination, with inclinations of about 43° (normal magnetization). A second direction with a southeasterly declination and approximately -50° inclination (reverse magnetization) is present only in some samples at sites 1 and 2. There are no significant differences either in petrographic characteristics or in the rock magnetic properties between the normal magnetized samples and the reversely magnetized samples at both sites 1 and 2. Wirth et al. (1997) report only minor geochemical differences for the entire sequence of flows. The detailed magnetic information for these sites is given in Table 4. Combined data from both components yield a Virtual Geomagnetic Pole (VGP) at 30.9° N, 186° E, which

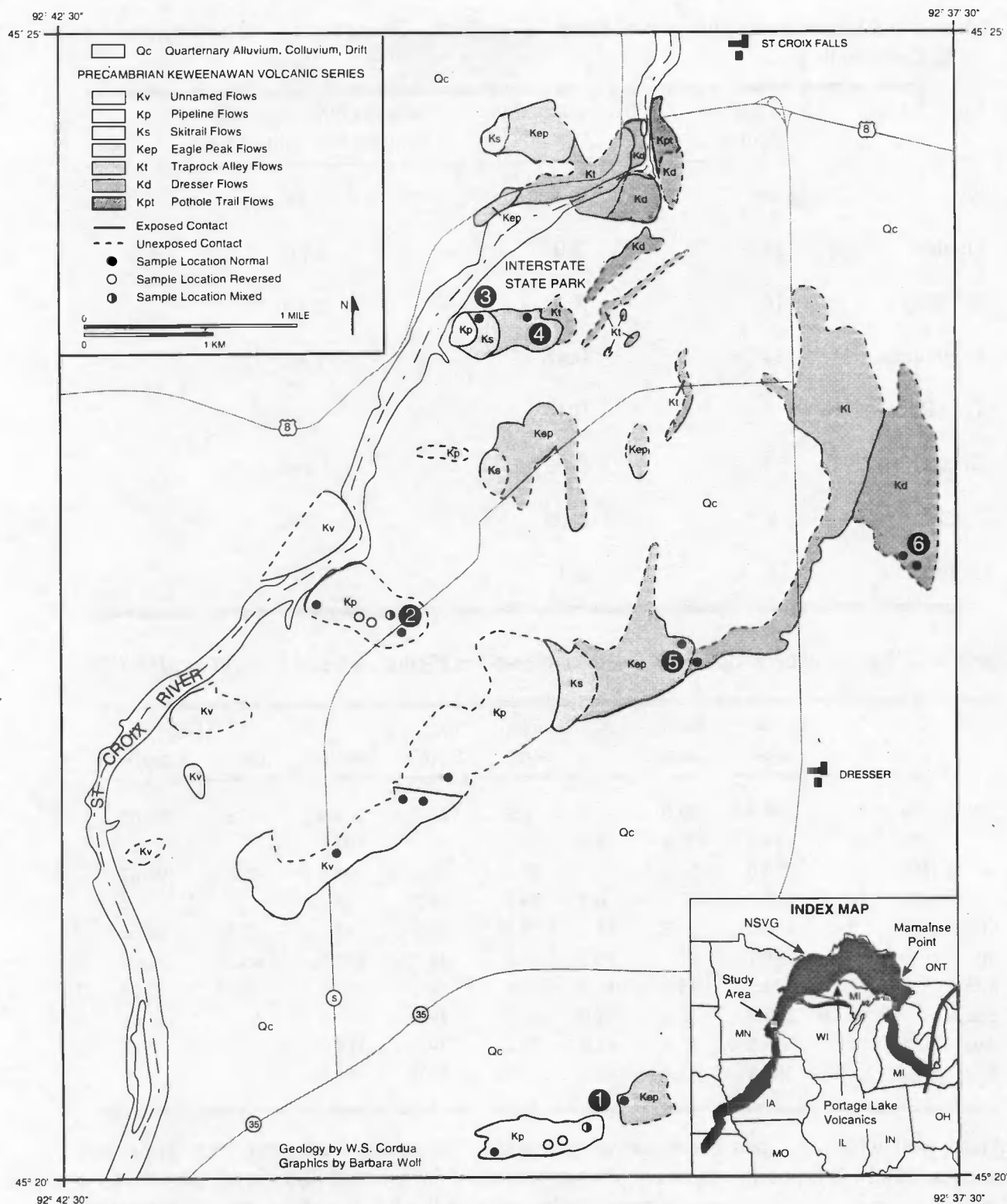


Figure 16. Map showing paleomagnetic sampling sites (after Kean et al., 1997).

Table 3. NRM intensities and susceptibility values for the Chengwatana volcanics in the St. Croix Falls area.

Flow Name	Number of samples	NRM Intenstity A/M	Magnetic Susceptibility Volume S.I. units x10-5
Kv	9	2.3	3591
Pipeline	14	2.0	5118
Ski Trail	16	2.1	5326
Eagle Peak	14	0.15	1061
Trap Rock Alley	6	0.14	23
Dresser	11	0.057	299
Pothole Trail	6	0.05	742
Lotus Lake	6	2.1	3647

Table 4. Paleomagnetic data for the sites indicated in Figure 16 (after Kean et al., 1997).

	N	Declination	Inclination	K	Apha -95	VGP		SITE	
						Lat(N)	Long(E)	Lat(N)	Long(W)
Site 1 (N)	10	288.9	55.0	49.5	6.9	36.7	193.4	45.33	92.65
(R)	7	114.8	-57.8	5.2	29.2	41	193.4		
Site 2 (N)	7	274.0	51.1	13	17	24.6	198.4	45.37	92.67
(R)	8	106.7	-47.5	6.7	23.1	31.2	187.9		
Site 3	5	280.0	33.2	14	20.8	19.9	183	45.39	92.66
Site 4	8	285.1	31	16.3	14.1	22	178.9	45.39	92.66
Site 5	7	306	45.5	11.2	18.8	43	173	45.36	92.64
Site 6	8	287.9	42.6	25.4	11.2	29.6	183.6	45.37	92.63
Avg (N)	6	286.8	43.4	43.9	10.2	29.7	184.9		
Avg (N&R)	8	287.5	45.8	48.5	8.04	30.9	186.1		

agrees well with published Keweenawan paleopoles for stable North American (Halls and Pesonen, 1982). It is possible there are other exposures of the reversed flows in the area, because detailed magnetic studies have not been conducted over the full length of the two prominent ridges in the area.

Paleomagnetic Correlation

The reversed flows in this area are important because they appear within a thick series of flows that are normally polarized and considered to be equivalent in age, or younger than, the

North Shore volcanics. Past interpretations suggested that all the Keweenawan-age flows equivalent or younger than the North Shore volcanics were normally polarized (Green, 1977; Halls and Pesonen, 1982), and those flows older than the North Shore volcanics were reversely polarized. Similar short reversals at apparent equivalent stratigraphic levels have been documented at only a few other locations in the Lake Superior region. The basalt flows at Mamainse Point in the eastern Lake Superior region show a similar short reversal (Palmer, 1970; Robertson, 1973). The dikes in the alkali syenite-carbonatite complexes in the Kapuskasing structural zone in Canada dated at 1097 Ma years show reversals in a normal polarity sequence (Symons et al. 1994). The geochemical data of Wirth et al. (1997) are consistent with this equivalence in age between the Chengwatana flows and the middle section at Mamainse Point. A short reversal at this time is also consistent with recent ages by Wirth and Gehrels (in press). They report precise U-Pb zircon ages of $1102 \text{ Ma} \pm 5 \text{ Ma}$ for slightly older Chengwatana rhyolites near Clam Falls, which is on the eastern extreme of the Chengwatana exposures.

Field Identification of Reversed Flows

Unfortunately, it is not possible to visually recognize normal- from reverse-polarity rocks in the field. Occasionally, the reversed flows in this area will cause a compass needle to deflect by more than 90° , but then so will areas struck by lightening. Only detailed demagnetization of samples in the laboratory can resolve the polarity question.

Stop 3a—Chengwatana Volcanic Flows of Interstate Park, WI.

Location: Access to Eagle Peak is from a hiking trail leading west from the group camping area, Interstate Park (WI). NW 1/4, SE 1/4, Sec. 36, T 34 N, R 19 W. (Fig. 17).

NOTE: Specimen collecting is not allowed in the park!

The crest of Eagle Peak is on top of the dipping Eagle Peak flow, a porphyritic basalt with large plagioclase phenocrysts. The panoramic view from the crest allows visitors to see the contrasting topography on east-facing flow fronts versus west-sloping flow tops, the general distribution of Cambrian sedimentary rocks deposited around basaltic islands, and the Quaternary features of the Dalles of the St. Croix River.

The Eagle Peak flow map unit consists of several flow units containing conspicuous pink plagioclase phenocrysts up to 6 cm long in a fine crystalline matrix. The matrix and phenocrysts are partly to completely replaced by secondary metamorphic minerals such as epidote, chlorite, actinolite and albite. The flow top contains many amygdules filled with quartz, epidote, K-feldspar and chlorite. The flows dip gently westward at an angle of about 15 degrees.

This flow is an important stratigraphic marker, traceable in the area for over 9 kilometers along strike. In places the unit consists of only one flow; in others it thickens to as much as 60 meters and includes at least 3 flow units, one on top of the other. In the Eagle Peak area, only one flow is evident.

The slope continuing down to the west is the dip slope marking the top of the Eagle Peak flow. This is in contrast to the steep east-facing escarpment one ascends to reach the summit from the group camping area. A small escarpment in the trees marks a quarry exposing the overlying flow. This emphasizes the stair-step topography of the terrain formed by the dipping flows.

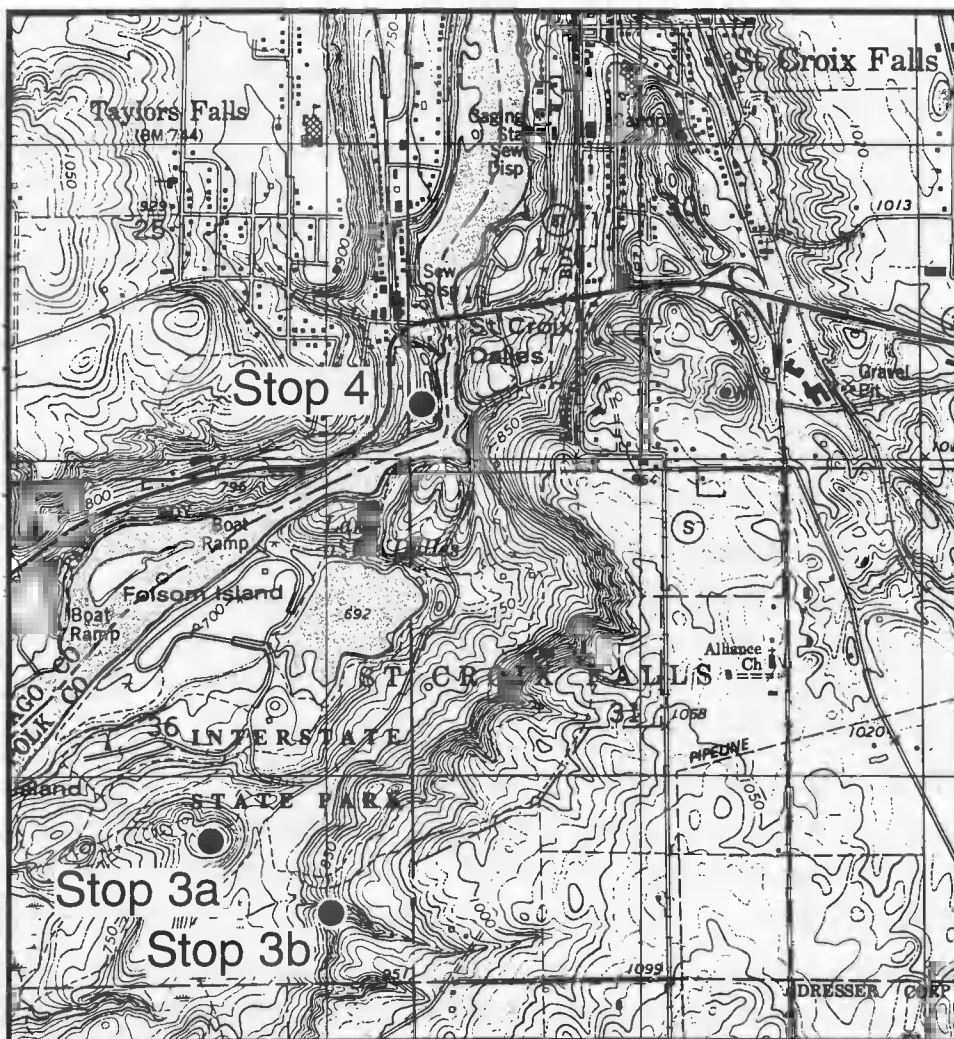


Figure 17. Scanned image of a portion of the USGS St. Croix Dalles, WI-MN topographic quadrangle (7.5-minute series; 1:24,000 scale) showing the locations of Stops 3a, 3b, and 4.

One can see outcrops of Cambrian sediments deposited unconformably on the complex landscape developed in the Chengwatana flows. Looking slightly south of west, across the river, one sees buff-colored Cambrian Mazomanie sandstones exposed in road cuts along U.S. Highway 8. Cambrian strata continue down to river level on the west side of the St. Croix River. The sandstone is traceable through the trees along the bluffs north nearly to the town of Taylors Falls. Here a ridge of basalts (predominantly the Eagle Peak and Trap Rock Alley flows) rise up from the river to form the hills on which the town rests. The scenic gorge called the St. Croix Dalles is cut through these flows. Locally along the contact of the Cambrian sediment, is a coarse basalt boulder conglomerate facies called the Mill Street conglomerate. This is not visible here, but will be visited at the next stop. Another ridge of basalt is found near the Northern States Power Plant and forms the highlands on which St. Croix Falls is built. East of the river are many outcrops of basalt. A few exposures of Cambrian sandstone, particularly along the bluffs to the east, can also be found.

Reconstruction of the paleogeography of the Cambrian-Precambrian unconformity was done using outcrop and water-well data. The unconformity is complex and shows a relief of over 100 meters (Scott and Anderson, 1983). In some areas, the relief occurs along buried shear cliffs, which were progressively covered as the Cambrian sea rose. The Mill Street conglomerate marks these environments. In other areas, calmer embayments and lagoons formed in areas of lower relief, sometimes isolated behind basaltic islands. Fine sandstones and shales of the Eau Claire and Mazomanie Formations were deposited in these areas. The shapes of these islands were controlled by the stair-step topography of the flows, joint sets, and small faults. The continuity of the Eagle Peak flow from island to island suggests that faults with major displacement did not control the shapes of the islands.

Knowledge of the distribution of basalt in the subsurface has considerable practical importance. A company building a pipeline through the area in the 1970's went bankrupt because they failed to take into account the expenses involved in blasting through these buried basalt ridges when submitting their bid. It is also not uncommon for one parcel of land to have basalt near the surface to have little or no ground water, while a neighboring parcel a few 100 meters away will have a plentiful supply of water from a thick section of Pleistocene and Cambrian sediments.

The scenic dalles of the St. Croix River are where the river, after excavating the soft Cambrian sediments in a gap between paleo-islands, has finally cut down to the hard basaltic shoal separating deeper Cambrian embayments.

Immediately to the north, one can see the Lake of the Dalles, a possible whirlpool or plunge pool formed when the St. Croix flowed along a more easterly course. Finally, numerous glacial striations and chatter marks can be found on the bedrock surface at Eagle Peak.

Stop 3b—Cambrian Rocky Shoreline Deposits of Interstate Park, WI.

See Figure 17 for location map.

Location: Ravine immediately south of Skyline Trail, approximately 400 m from trailhead at group campground in Interstate Park (WI). SE1/4, SE1/4, Sec. 36, T 34 N, R 19 W.

NOTE: Specimen collecting is not allowed in the park!

The Skyline Trail skirts a low, marshy area for its first 150 meters. It then begins its slow ascent onto the Trap Rock Alley flow of the Precambrian Chengwatana basalt, with outcrops of basalt prominent on both sides of the trail. After approximately 400 m, a steep-walled, narrow ravine can be seen by looking down to the right (south) of the trail. This ravine marks the approximate contact between Precambrian basalts and Cambrian sedimentary deposits. Basalt outcrops occur on the floor of the ravine and along the north wall. Along the south wall, 8 m of Upper Cambrian conglomerates and sandstones tentatively assigned to the Mazomanie Formation of the Tunnel City Group (Mazomanie Member of the Franconia Formation in Minnesota) lie directly above and lateral to the basalts. The Cambrian sediments can be traced for some 240 m to the east up the ravine, but the best exposures are in the first 70 m.

On the south wall of the ravine, a basalt boulder conglomerate facies ("the Mill Street conglomerate" of Berkey, 1897) lies against the lava flows and grades vertically and laterally into the sandstone facies more typical of the Cambrian section in this area (Fig. 18). In sharp contrast to the gray-green basalt flow outcrops on the north side of the ravine, the basalt boulders in the conglomerate are heavily weathered and stained orange-brown by iron oxides. The largest

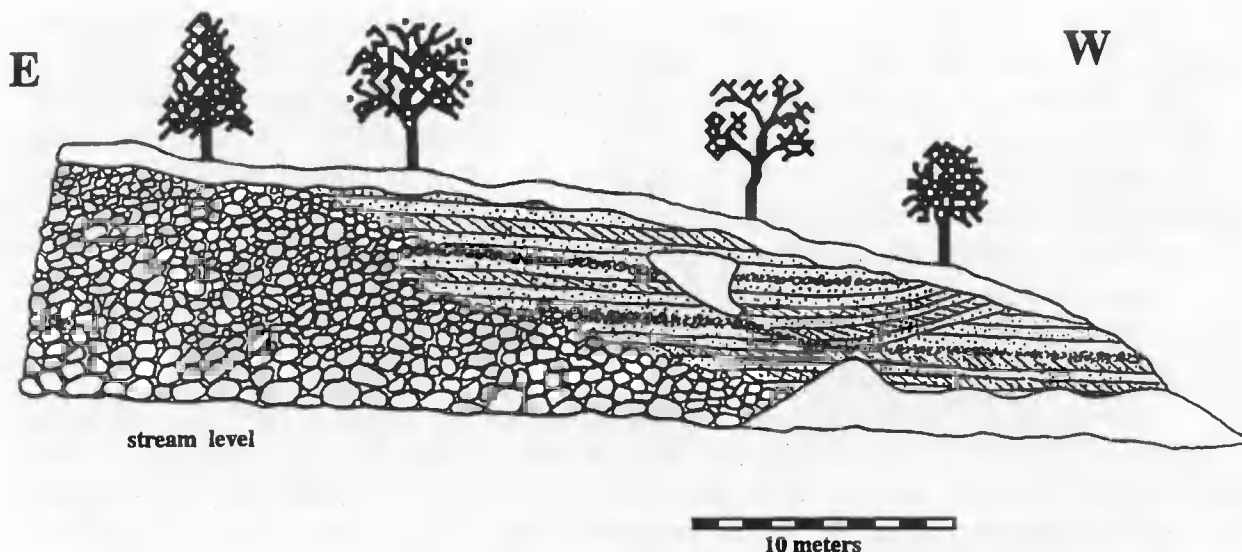


Figure 18. Sketch of south wall of ravine along Skyline Trail (Stop 3b) showing vertical and lateral relationships between basalt boulder conglomerate and sandstone facies. Covered areas indicated by lack of pattern. No vertical exaggeration.

boulders near the base of the exposure are over 1 m in diameter. The boulders are rounded but of low sphericity, perhaps due to original jointing in the basalt. A quartz sandstone matrix with abundant inarticulate brachiopods fills the spaces between the boulders. No trace of bedding is present within the main body of the conglomerate, although a few thin conglomerate beds extend out into the sandstones. These beds contain rip-up clasts of sandstone as well as basalt cobbles and pebbles, in contrast to the main body of the basalt conglomerate in which sandstone clasts have not been seen. They also show weak imbrication suggesting current flow to the west.

The contact between the conglomerate and sandstone facies is surprisingly sharp. The sandstones are orange-brown, thin- to medium-bedded quartz arenites. The sandstone beds vary in texture from fine- to coarse-grained, from sub-rounded to rounded, and from moderately to well-sorted. Some fine- to medium-grained sandstone units contain scattered coarse, frosted quartz grains. A few beds are graded, and some contain quartz pebbles and rip-up clasts of sandstone and siltstone. Many units show high-angle planar cross-bedding, with foresets dipping to the west. Iron-oxide staining and concretions are common, as are inarticulate brachiopod shell- hash layers. A large trough cross-bed set (or possibly a slump) near the center of the exposure is indicated by a curved erosional surface marked by basalt cobbles near its base, and sandstone beds which dip slightly to the east, in contrast to the slight westward dip of most of the sandstone beds in the outcrop.

Interpretation

The Taylors Falls-St. Croix Falls area of Minnesota and Wisconsin is one of the few localities in North America where Upper Cambrian conglomerates lie unconformably on Precambrian rocks (Webers and Austin, 1972; Cavaleri et. al., 1987).

Apparently, the Cambrian Sauk sea transgressed onto a land surface of very low relief as it flooded the North American craton, leaving a blanket of sand as it advanced. At a few localities such as Taylors Falls-St. Croix Falls and Baraboo, Wisconsin (Dalziel and Dott, 1970; Dott, 1974), resistant bedrock hills stood above the subdued land surface and acted as local sources for coarse conglomerate deposits as the sea advanced. In Baraboo, the conglomerate was formed of quartzite boulders from local outcrops; in the Taylors Falls-St. Croix Falls area, the conglomerate was formed of basalt boulders from the local lava flows.

The basalt boulder conglomerates, informally called the "Mill Street conglomerate" are exposed at only a handful of sites on both sides of the St. Croix River south of Taylors Falls-St. Croix Falls. Of these sites, the Skyline Trail locality, shows most clearly the vertical and lateral relationships between the Precambrian basalts and Upper Cambrian conglomerates and sandstones. As can be seen in the ravine, the basalt boulder conglomerate was deposited on top of and against a steep face cut into the basalt lava flows. Very high energy conditions must have been present to round the corners of the basalt boulders. Dott (1974) estimated the height of the waves required to round boulders of similar size in Baraboo. His conservative estimates of waves 5 m high suggest storms were important in producing these conglomerates. A paleogeographic reconstruction of the Taylors Falls-St. Croix Falls area as the Sauk sea began to submerge the basalt highlands shows that the lava flows along the Skyline Trail would have been fully exposed to waves from the west or southwest. It is tempting to envision basalt sea cliffs on the windward side of islands or headlands, with storm-generated waves crashing into and undercutting the cliffs, and the rubble in the surf zone at the base being pounded and jostled to produce rounded boulders. Such an environment evokes images of the Hawaiian Islands or perhaps the Pacific Northwest coast today. However, very little has been written in the geologic literature on the characteristics of rocky shoreline deposits (Johnson, 1988) and care should be taken in reconstructing the exact depositional environment.

The sharp contact between the conglomerate and sandstone and the difference in lithology between the two facies suggest that the sands came from a more distant source and were washed along the shoreline and deposited against and between the basalt boulders. The graded bedding, rip-up clasts and shell-hash layers show that periodic storms may have been more important in deposition of the sandstone beds than normal day-to-day events. Storms could also account for the two conglomerate beds extending out into the sandstone; each bed with a mix of basalt cobbles and sandstone rip-up clasts from the ocean floor.

The sandstones are tentatively assigned to the Mazomanie Formation of the Tunnel City Group by their similarity to outcrops of that formation across the St. Croix River in Minnesota. However, they may represent a near-shore facies of the Ironston Member of the Wonewoc Formation as well. The Mill Street conglomerate varies in age from Dresbachian to Franconian (Webers and Austin, 1972) and thus may grade laterally into sandstones of either formation. Detailed study of the fossil assemblages, particularly the trilobites, in the sandstone and conglomerate would be the best method to tell the age and formation, but as yet this has not been attempted. Such a study would be especially valuable since the Mill Street conglomerate is known to contain an unusual assemblage of trilobites, inarticulate brachiopods and monoplacophoran molluscs representing an intertidal, shoreline environment (Webers, 1972). According to Johnson (1988), the oldest known rocky-shore faunas are Ordovician in age. It may well be that the unusual Upper Cambrian conglomerate deposits around St. Croix Falls and Taylors Falls contain the oldest rocky shoreline faunas in the world.

Stop 4—Potholes of Interstate Park, MN.

See Figure 17 for location map.

Location: Paths leading from the parking lot by the interpretative center, Interstate Park (Minnesota). SW 1/4, SW 1/4, Sec. 30, T 34 N, R 18 W.

NOTE: Specimen collecting is not allowed in the park!

This stop provides an opportunity to examine outcrops of basaltic lava flows with spectacular potholes. From the parking lot, follow the service road (also called Trap Rock Alley) a short distance. Many short, well-marked trails to the east imaginatively and conveniently lead over, around and through world-class potholes cut in the top of the Dresser basaltic lava flow.

The base of the Trap Rock Alley flow, here at its type locality, is well exposed along the west side of Trap Rock Alley. The top of the Dresser flows is exposed to the east. This flow dips gently west under the Trap Rock Alley. The flow top is typically amygdaloidal and extensively epidotized. It is well exposed in abundant outcrops, as well as along the sides of the many enormous potholes.

The potholes formed on the Dresser flows are thought to be some of the deepest in the world (Crawford, undated). They range from fist-size depressions to “giant’s kettles” that can be over 6 meters across and more than 20 meters deep. The potholes are generally partly filled with a mixture of silt, mud, peat and coarse grind stones ranging from pebble to boulder size. The sediments, swirled by the turbulence of the river as it raced over the basalt, were the tools cutting these potholes.

These potholes are part of a larger region of river-scoured terrain. In addition to them, the bedrock surface is scoured, smoothed and polished in many areas. The potholes were produced when localized areas, facilitated by joint intersections, wore down slightly faster than the surrounding rock. Eddies formed over these low spots, trapping sediment. Large boulders occasionally were stranded in these depressions. The strong current swirled them about, but lacked the force to free them. This action cut the potholes. Notice in places the grooves cut in the sides of the potholes as large boulders ground their way down. The boulders themselves were shaped by the process. Grind stones recovered from the potholes are as smooth and rounded as cannon balls.

The potholes here are 7.5-18 meters above the current river level. Some potholes have been found as high as 34 meters above river level.

The potholes formed when water from the melting glaciers filled Glacial Lake Duluth to the north (Cahow, 1975; Ojakangas and Matsch, 1982). The water overtopping this lake flowed south down the ancestral St. Croix River. At Interstate Park, these waters were further funneled down the narrow channel now seen as the Dalles. Below the Dalles, the river encountered soft Cambrian sandstones and shales, which cut away faster than the basalt. This resulted in a major knickpoint in the St. Croix at the Dalles. This high gradient sped up the water, increasing its ability to gouge out the potholes as it excavated its valley into the resistant basaltic bedrock.

Many of the potholes here have been given names by the park service. Be sure to enjoy the Lily Pond, Devil’s Parlor, Cauldron, Bottomless Pit and Bake Oven as time permits. Also note the west-dipping thrust fault exposed in the cliff by the riverboat dock just west of the potholes area.

Stop 5—Volcanic Flows and Interflow Sediments west of Lewis, WI.

Location: Outcrops are located on 345th Ave one-half mile west of the intersection with 130th Street. N 1/2, SE 1/4 and S 1/2, NE 1/4, Sec. 11, T 37 N, R 17 W; Lat. 45° 42' 07.6" N, Long. 092° 26' 25.8" W.

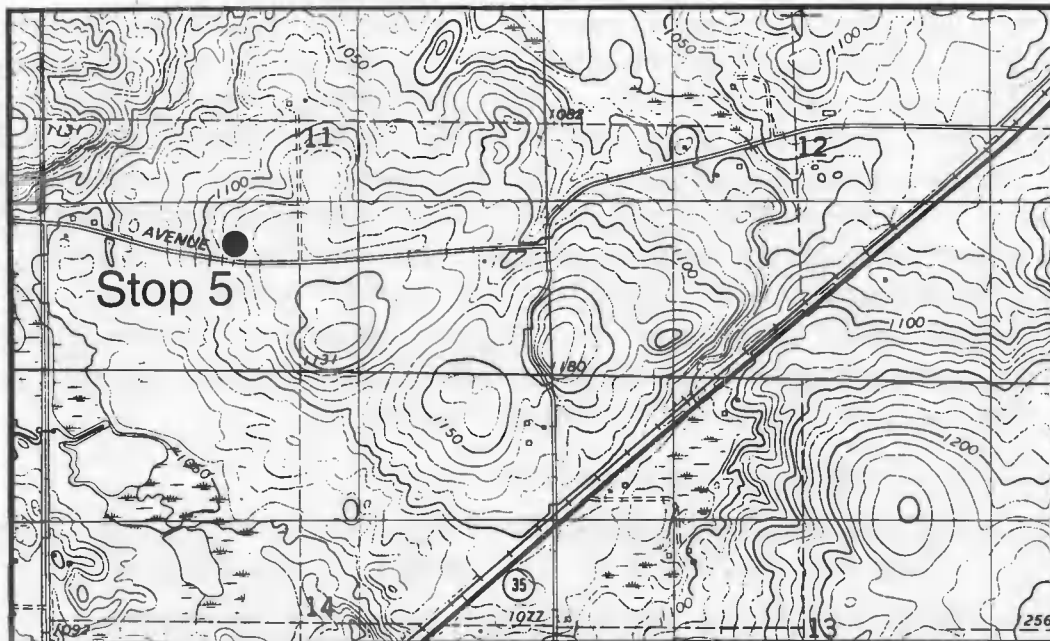


Figure 19. Scanned image of a portion of the USGS Frederic, Wisconsin topographic quadrangle (7.5-minute Series; 1:24,000 scale) showing the location of Stop 5.

These outcrops are on private land and you should seek permission before entering. Basalt and interflow sediments are exposed in a series of outcrops 20-100 meters north of the road.

Near the road, fine-grained basalt is exposed in a flow that is at least 4 meters thick. Near the top of the flow, the basalt contains rare large quartz-filled amygdules.

Interflow sediment is exposed further north and is particularly well-exposed in an east-west trending cliff face. Please do not hammer on this face. Excellent examples of the interflow sediment can be found in the many loose blocks along the base of the cliff. The basalt is overlain by a mixed basalt and sediment zone that is roughly 3 meters thick. The sediment weathers pale reddish-orange and consists of fine-grained, poorly sorted, and thinly laminated quartz, feldspar, and volcanic fragments. The fine-grained matrix material is completely recrystallized to epidote and other greenschist facies minerals. Beds typically show planar bedding structures, although some cross-beds (ripple structures), channel-fill structures, and convoluted beds are also visible. Bedding structures often trend uninterrupted right up to the margins of clasts, though in some places bedding surfaces appear to be deflected by clasts. There are also patches of sediment that are surrounded by basalt and appear “isolated” in the two-dimensional outcrop face. These, as well as many narrow sub-vertical “veins” of sediment, exhibit well-defined, sub-horizontal layering that is sharply truncated along its margins.

Fragments of basalt range from less than a millimeter to more than 15 centimeters. The clasts are typically sub-angular, consist of microcrystalline plagioclase, and are highly vesicular (1-5 mm diameter); the outer surfaces of many clasts are highly embayed and exhibit very delicate “sponge-like” textures. Some clasts are sub-rounded and exhibit decreasing vesicle sizes near their outer edges; the long axes of highly elliptical vesicles also parallel the clast surfaces. Many of these features are suggestive of chilled margins surrounding some of the clasts. Many larger clasts also have numerous small “floating” fragments around their margins. Viewed closely, many of these fragments appear to be disrupted “glassy” walls between adjacent vesicles. Vesicles in the interiors of fragments are typically filled with coarse epidote. Near the top of the outcrop, a nearly continuous layer of vesicular basalt has a highly irregular base that in some places cross-cuts layering in the underlying sediment. All of these features point to a highly immature sediment. Elsewhere in the rift, interflow sediments are interpreted to record deposition in predominantly sedimentary environments (e.g., Jirsa, 1984). However, some features at this stop are suggestive of hyaloclastites, in which basaltic melt interacts with water, or water-saturated sediment. Preliminary initial ϵ_{Nd} values of these sediments (~ 2.0 ; Naiman, unpublished data) are similar to those of the surrounding basalts (Table 5) suggesting that the source of the sediments is primarily immature rift volcanics.

Stop 6a—Rhyolite and Basalt near Clam Falls.

Location: 435 m NNE of the intersection between County Rd I and 73rd St. SE 1/4, SE 1/4, Sec. 14, T 37 N, R 16 W; Lat. 45° 41' 09.8" N, Long. 092° 18' 27.4" W.

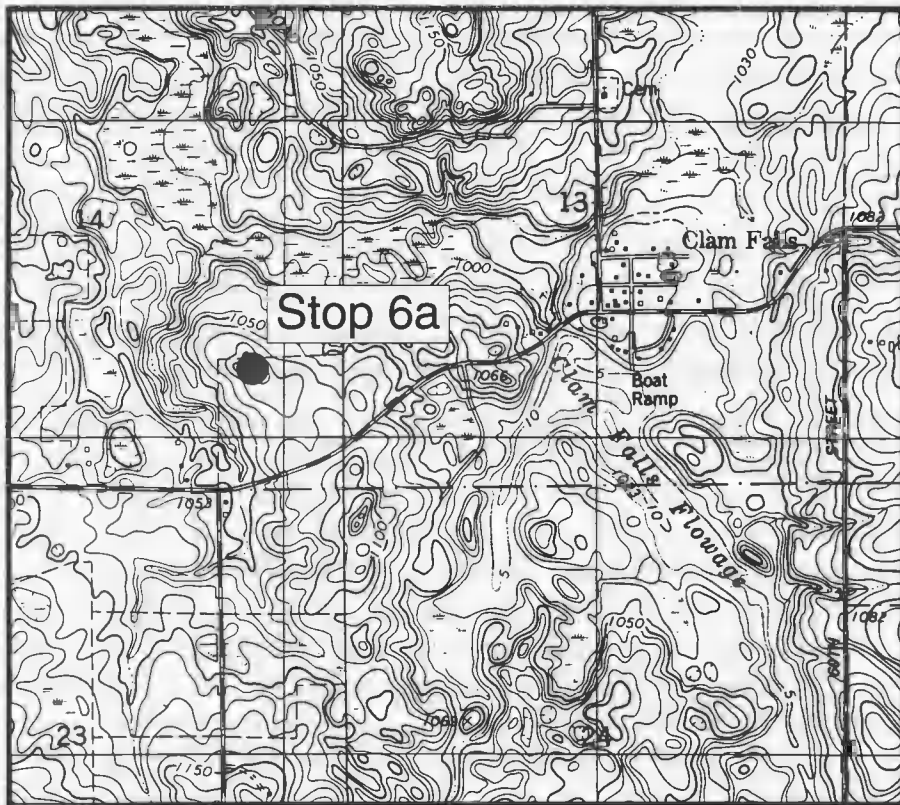


Figure 20. Scanned image of a portion of the USGS Clam Falls, Wisconsin topographic quadrangle (7.5 minute series; 1:24,000 scale) showing the location of Stop 6a.

Table 5. Representative analyses of basalt and rhyolite from the Clam Falls region.

	basalt KC-301b	basalt KC-303	rhyolite KC-302a	rhyolite KC-302d	rhyolite KC-310
SiO ₂	47.08	49.89	72.46	71.34	72.18
TiO ₂	1.60	1.84	0.49	0.49	0.48
Al ₂ O ₃	15.66	14.91	11.61	11.52	11.93
Fe ₂ O ₃	14.10	13.61	4.87	5.30	5.78
MnO	0.18	0.17	0.09	0.06	0.12
MgO	6.84	5.10	0.48	1.36	0.40
CaO	10.00	8.61	1.08	2.32	1.32
Na ₂ O	2.31	3.17	2.89	2.63	3.52
K ₂ O	0.56	0.55	5.10	4.11	4.21
P ₂ O ₅	0.15	0.20	0.08	0.08	0.09
LOI	1.72	2.31	0.54	1.01	0.40
Total	100.20	100.36	99.69	100.22	100.43
Mg#	0.53	0.47	0.19	0.37	0.14
<i>qz</i>		0.56	31.39	31.85	29.75
<i>or</i>	3.41	3.36	30.56	24.62	25.02
<i>ab</i>	20.08	27.66	24.75	22.51	29.90
<i>an</i>	31.54	25.60	3.65	7.55	4.30
<i>hy</i>	14.07	21.60	6.43	8.21	7.24
<i>ol</i>	9.23				
Rb	19	13	151	91	123
Ba	115	82	1036	889	831
Sr	164	216	80	108	90
Nb	8	12	68	62	67
Zr	111	178	794	713	965
Y	31	38	141	118	155
Cr	205	117	14	24	19
Ni	133	76	6	14	9
V	281	299	21	26	17
Initial ϵ_{Nd}	+3.4	—	-0.10	—	+0.95

Sample Locations: T 37 N, R 16 W.

KC-301b - SW1/4, SW1/4, section 13

KC-302a - SE 1/4, SE 1/4, section 14 (Stop 6)

KC-302d - SE 1/4, SE 1/4, section 14 (Stop 6)

KC-303 - SE1/4, NW1/4, section 13

KC-310 - SE1/4, SE1/4, section 26 (Stop 6 Alt.)

These outcrops are on private land and permission to access them should be sought before entering. The outcrops can be accessed from the property (north of the intersection of Country Rd I with 73rd St. approximately 1 mile west of Clam Falls. Direct access to the outcrops from County Rd I requires permission from another landowner. An alternate stop (6b) near the McKenzie Creek State Hunting Grounds is given below.

The purpose of this stop is to examine an example of rhyolite. Rhyolite is extremely rare in this part of the rift and only two localities are known in the area extending from Taylors Falls to Clam Falls. This is in stark contrast to the relatively abundant rhyolites (up to 25 percent of the volcanic section; Green and Fitz, 1993) that are exposed on the North Shore of Lake Superior. This locality was also sampled for U-Pb zircon studies (see below).

At this exposure, the rhyolite is exposed over an area that measures approximately 150 x 75 meters. The upper surface of the outcrop at the north end of the exposure trends N 50° E and dips 15° NW, sub-parallel to the trend of flow layering in the area. At the southern end of the outcrop, the basal contact with the underlying basalt flow can be seen. In a few isolated exposures approximately 3-4 meters below the top of the knoll one can observe isolated exposures in which basalt and rhyolite are complexly inter-fingered. Sub-angular to rounded clasts consist of fine-grained basalt, up to 15 cm in diameter, can be found in rhyolite. The upper contact of the rhyolite with the overlying flows is not exposed.

Weakly porphyritic rhyolite consists of phenocrysts of euhedral albite ($\text{An}_2\text{Ab}_{94}\text{Or}_4$) and rare quartz occur in a groundmass of intergrown quartz and potassium feldspar ($\text{An}_0\text{Ab}_5\text{Or}_{95}$); textures range from fine-grained intergranular to medium-grained spherulitic (Fig. 21). Spherulitic intergrowths are commonly nucleated on phenocrysts. Minor anhedral Fe-Ti oxides, rare calcic amphibole (ferro-edinite-hornblende), and accessory apatite and zircon are also commonly present. Groundmass quartz is typically euhedral to subhedral and tabular to elongate (length:width ratios = 10:1 to 50:1) indicating crystallization as tridymite, the orthorhombic, high temperature ($>867^\circ\text{C}$) form of quartz. The preservation of these delicate devitrification textures is remarkable given the metamorphic alteration and recrystallization that has occurred in this portion of the rift.

The rhyolites of the Clam Falls region are classified as rhyolite or rhyodacite on most major and trace element discriminant diagrams (Abbott et al., in press). The major and trace element compositions (Table 5) are similar to some icelandites and granophyres from the North Shore Volcanic Group (Vervoort and Green, 1997) and to type I rhyolites from the Portage Lake volcanics (Nicholson, 1992; Nicholson and Shirey, 1990). Nd isotopic studies of the rhyolites (Abbott et al., in press; Naiman and Wirth, in press) indicate that they have initial ϵ_{Nd} values ≈ 0 . This evidence, combined with their relatively minor abundance in the volcanic section, suggests that they originated as residual melts by magmatic fractionation. This is in contrast to many of the rhyolites on the North Shore that have negative initial ϵ_{Nd} values and are interpreted to be the result of crustal melting (Vervoort and Green, 1997). Basalt from this part of the volcanic section (e.g., KC-301b and KC-303; Table 5) is similar to low-Ti basalt of the Taylors Falls region and has initial ϵ_{Nd} values ranging from 0 to +3.4, consistent with derivation from multiple mantle sources, including enriched mantle plume and depleted asthenosphere (Naiman and Wirth, in press).

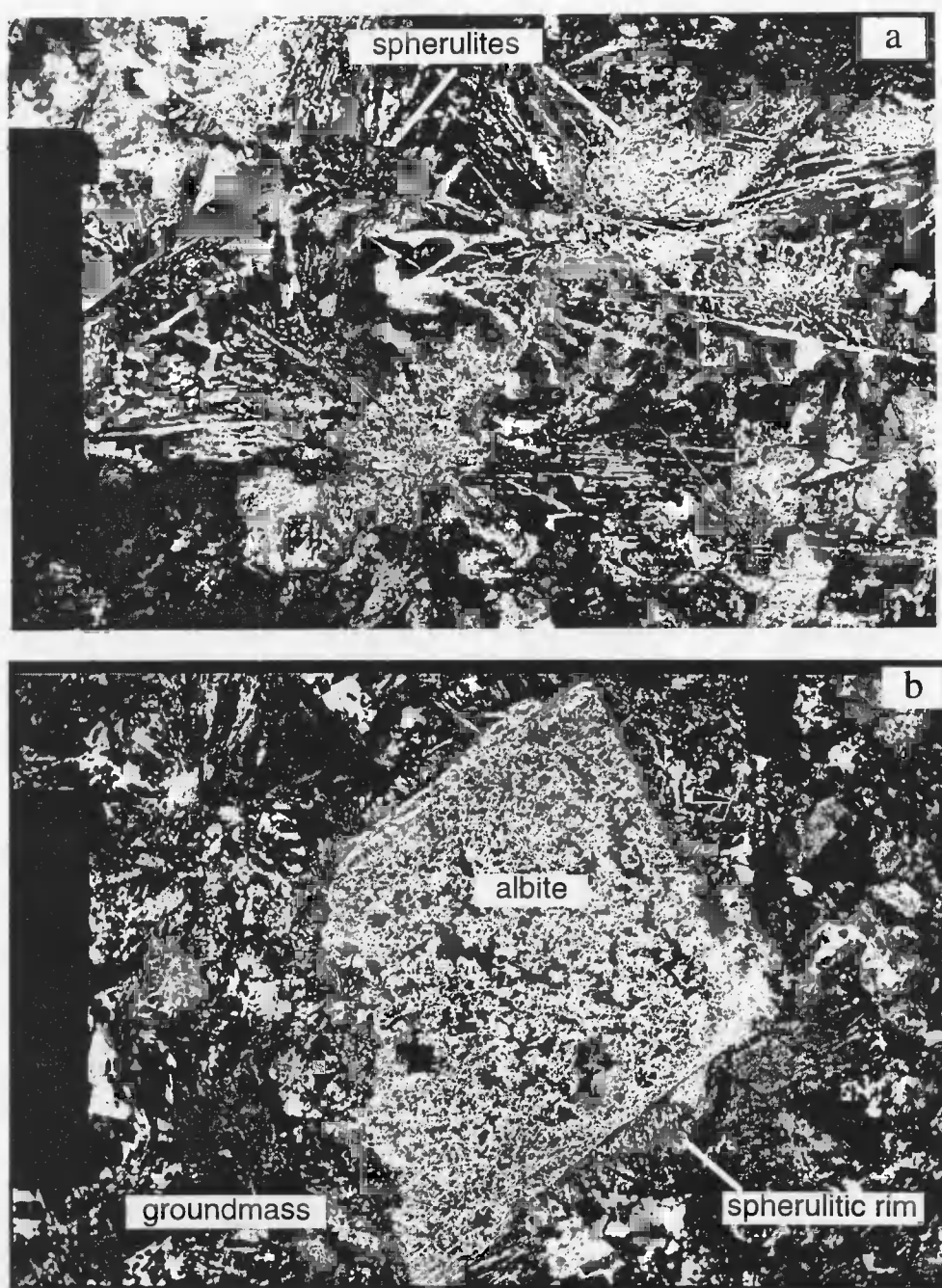


Figure 21. (a) Photomicrograph (plane-polarized light) of spherulitic intergrowths of tridymite paramorphs and feldspar in Clam Falls rhyolite (sample KC-302a). Scale bar along left side is 0.25 mm long. (b) Photomicrograph showing plagioclase (albite) phenocryst with spherulitic intergrowths around the margin. Scale bar along left side is 1.25 mm long.

One of the major difficulties in integrating geologic information from this portion of the rift with models of the rift has been the lack of geochronologic control in the Chengwatana volcanics. With this objective in mind, large samples of rhyolite were collected from both Stop 6a and Stop 6b for U-Pb zircon dating. The sample material from Stop 6b (KC-310) has not been processed at this time, but two, nearly identical ages were determined for different rhyolite phases (KC-302a and KC-302d) from Stop 6a. Zircons from both samples have relatively simple prismatic forms, are translucent, and vary in color from reddish-brown, to honey brown, and pale pink. Three zircon grains from KC-302a yielded concordant ages (Fig. 22) with mean $^{206}\text{Pb}^*/^{238}\text{U}$ and $^{207}\text{Pb}^*/^{206}\text{Pb}^*$ ages of 1102 ± 5 Ma (2σ). Three additional grains are slightly discordant, presumably due to small amounts of lead loss, and a discordia through these points, projected from 1102 Ma, yields a lower intercept of 196 ± 205 Ma (MSWD = 0.15). Analyses of three abraded grains from KC-302d are analytically concordant and yield mean $^{206}\text{Pb}^*/^{238}\text{U}$ and $^{207}\text{Pb}^*/^{206}\text{Pb}^*$ ages of 1101 ± 5 Ma (Fig. 22). Three additional grains and the two multigrain fractions are slightly discordant; a discordia through these points, projected from 1101 Ma, yields a lower intercept of 63 ± 217 Ma (MSWD = 0.45). Given the uncertainties of ± 5 Ma on each of the ages, the 1 Ma age difference between the two samples is probably not significant.

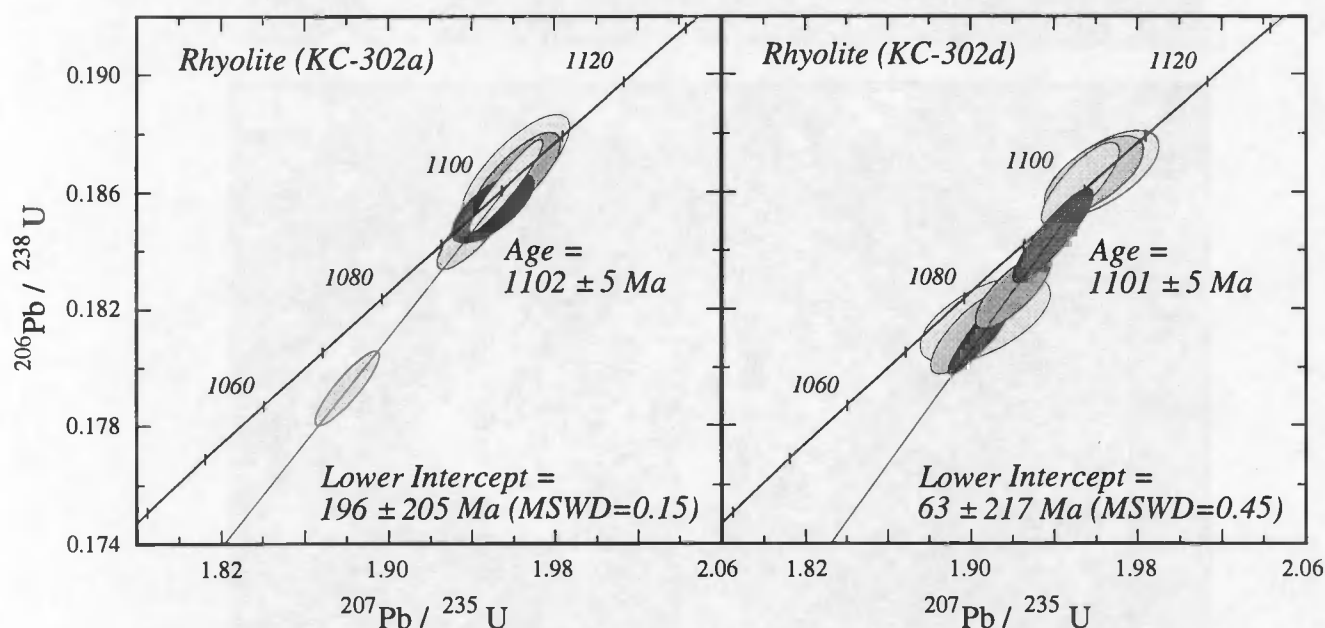
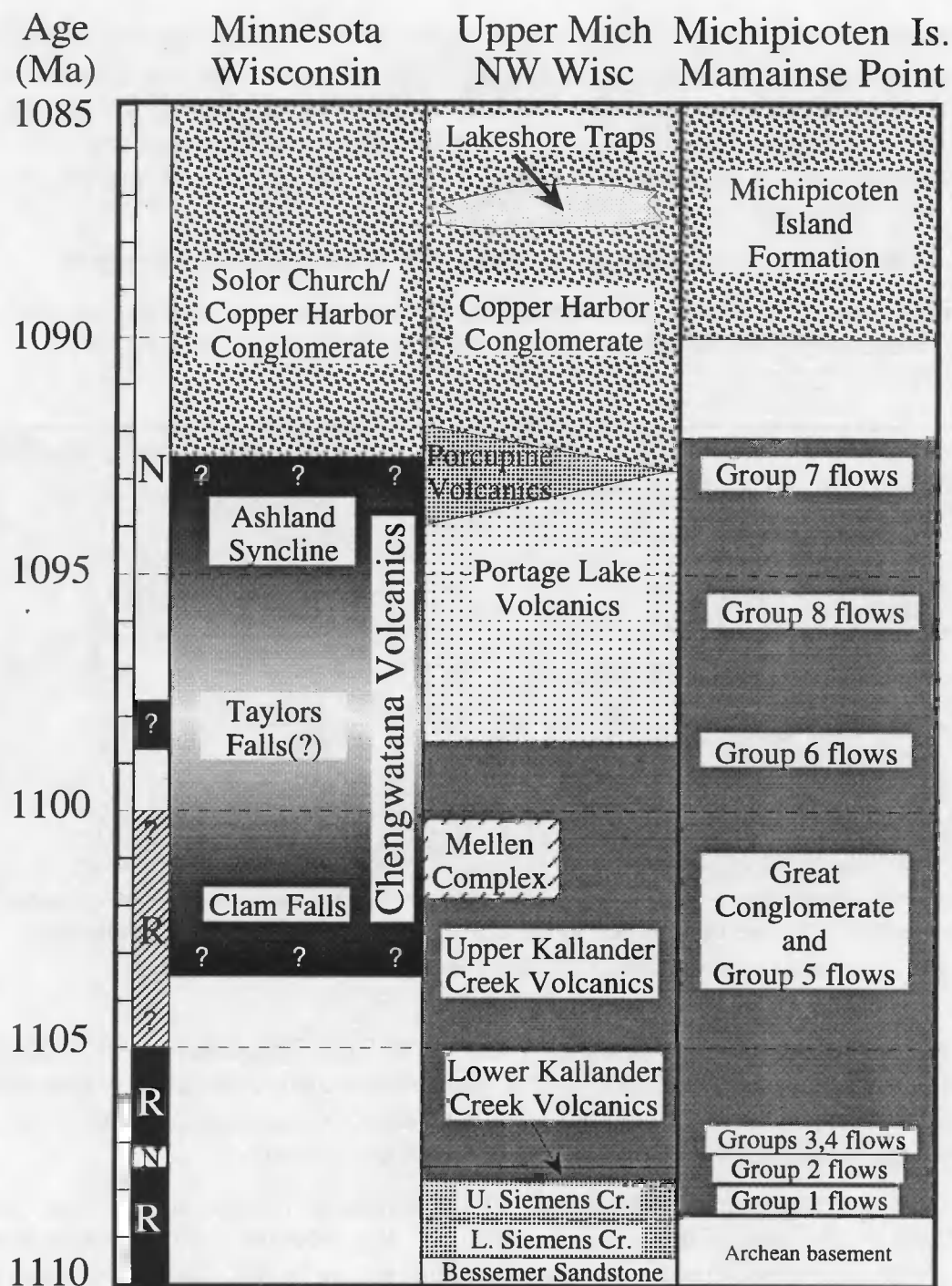


Figure 22. Concordia diagrams for Chengwatana volcanics from the Clam Falls area (Stop 6a). Error ellipses and regression age uncertainties represent 95 percent confidence limits (after Wirth and Gehrels, in press).

Analyses of the Chengwatana rhyolites near Clam Falls indicate that magmatism began by at least 1102 Ma in this portion of the rift, considerably earlier than the age reported by Zartman et al. (1997) for rhyolite (1094.6 ± 2.1 Ma) exposed near the base of the Chengwatana section in the southeastern limb of the Ashland syncline. This implies that lower flows exposed near Clam Falls likely formed contemporaneous with the Upper Kallander Creek volcanics and the Mellen Complex of Northern Wisconsin, and the Group 5 flows and Great Conglomerate of Mamainse Point (Fig. 23). The 1102 ± 5 Ma age of Chengwatana volcanics in the Clam Falls



Modified from Nicholson et al. (1997)

Figure 23. Diagram illustrating the correlation of Midcontinent rift rocks from the Taylors Falls-Clam Falls region with those of northern Wisconsin and eastern Lake Superior. The upper limit of the Chengwatana volcanics is estimated on the basis of U-Pb zircon studies by Zartman et al. (1997) in the Ashland syncline; the lower limit is based on the age of Clam Falls rhyolite. The age of Taylors Falls flows is based on their predominantly "normal" paleomagnetic signature (Kean et al., 1997) (after Wirth and Gehrels, in press).

region implies that the lower Clam Falls flows may have “reversed” paleomagnetic directions since they are approximately coeval with the reverse-to-normal magnetic polarity change (1105-1100 Ma) that has been widely observed throughout the Lake Superior region. The 1102 ± 5 Ma age of the Clam Falls rhyolites also implies that this section of the Chengwatana volcanics was erupted during a period regarded as the transition between the “latent” and “main-stage” of rift magmatism (Miller and Vervoort, 1996).

Stop 6b —(Alternate) Rhyolite near the McKenzie Creek State Hunting Grounds.

Outcrops are located on 310th Ave approximately 0.8 miles south and east of the intersection with 315th Ave. SE 1/4, SE 1/4, Sec. 26, T 37 N, R 16 W; Lat. $45^{\circ} 39' 15.1''$ N, Long. $092^{\circ} 18' 28.4''$ W.

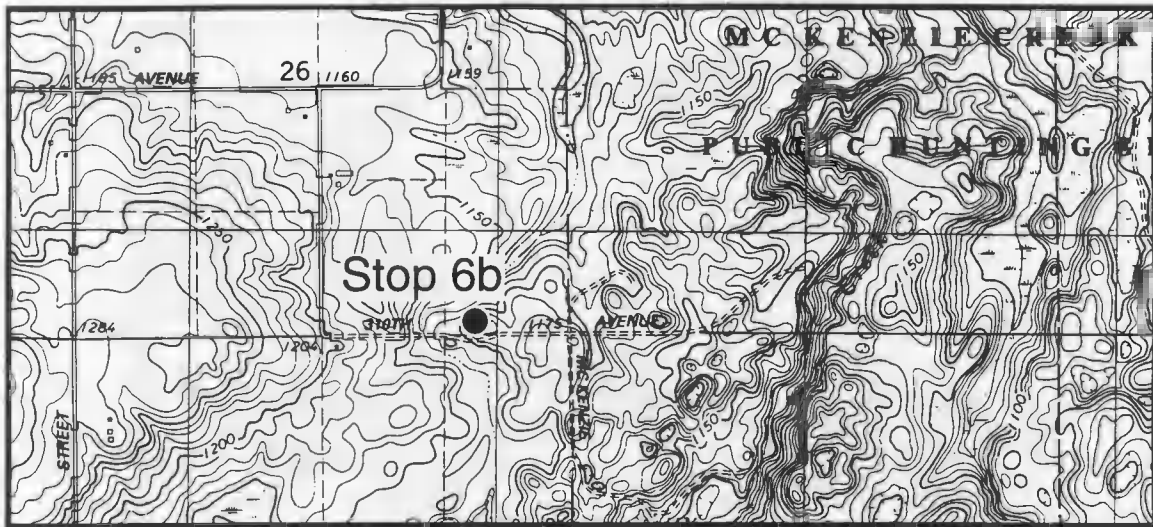


Figure 24. Scanned image of a portion of the USGS Clam Falls, Wisconsin topographic quadrangle (7.5 minute series; 1:24,000 scale) showing the location of Stop 6b.

This stop is an alternate stop in the rhyolites of the Clam Falls region. Although the contact with surrounding basalts is not exposed at this locality, the stop does permit examination of a large exposure of rhyolite. The outcrops are located along the north side of 310th Ave after the end of the pavement near where the road curves south around outcrop.

This outcrop is the easternmost exposure of Chengwatana volcanics in the Clam Falls region and defines the base of the exposed section. At this location, a several-meter-thick section of rhyolite is exposed over an area several hundred square meters. The rock consists of pale purple and aphanitic rhyolite. No compositional or textural layering is observed. The outcrop is cut by several sets of closely-spaced sub-vertical fractures that have coatings of iron oxides and other alteration minerals. Quartz-rich veins up to several millimeters thick are also commonly observed.

Stop 7—Porphyritic flows near Godfrey Lake, WI.

Outcrops are located 0.72 miles west of County Rd I on 315 St. SE 1/4, NW 1/4, Sec. 29, T 37 N, R 16 W; Lat. 45° 39' 43.0" N, Long. 092° 22' 43.4" W.

The purpose of this stop is to examine some of the strongly plagioclase-porphyritic flows of the Clam Falls region. Several interesting features can be observed in the exposures along the north side of the road at this stop. The eastern end of the exposure consists of coarse

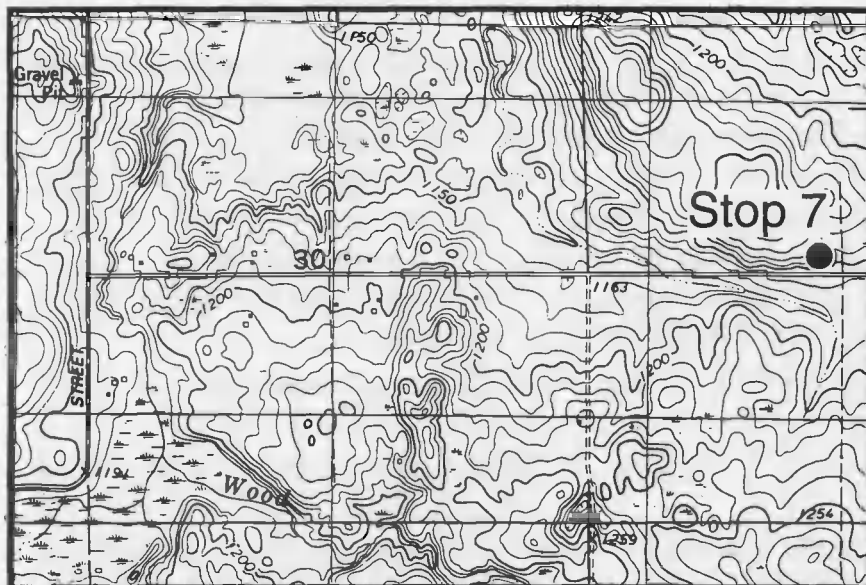


Figure 25. Scanned image of a portion of the USGS Clam Falls, Wisconsin topographic quadrangle (7.5-minute series; 1:24,000 scale) showing the location of Stop 7.

porphyritic basalt with large (<3 cm long) plagioclase phenocrysts (40-60 percent) in a groundmass consisting of microcrysts of plagioclase and secondary minerals; no primary mafic minerals are present. Possible zoning structures are still apparent in some of the feldspar phenocrysts. The rock has been extensively altered as evidenced by the strongly saussuritized and epidotized plagioclase, abundant chlorite,

disseminated iron-oxides, quartz, and secondary feldspar (pink). Near the top of the porphyritic basalt is a ten centimeter thick "epidosite" that is sub-parallel to flow layering. This epidosite is similar to many interflow sediments seen at other localities. Further west is a sparsely porphyritic basalt, which is overlain by repeating layers of fine-grained basalt and vesicular basalt that might represent several thin flow units. Some cross-cutting vesicular basalt is also observed.

Other coarse porphyritic basalts can be found several hundred meters south of this stop. There, plagioclase phenocrysts typically comprise 15-25 percent of the rock. Interflow sediment is also found. Other exposures of coarse porphyritic basalt are also found further south and west of this stop (one mile NW of Straight Lake). The orientations of flows in this area are poorly known, but they generally trend NNE and dip gently ($\sim 15^\circ$) to the NW. On the basis of these general trends, and the similarity of rock types, these units are tentatively considered to be the same stratigraphic unit. Considering that flows with coarse porphyritic textures are relatively rare in the Chengwatana volcanics (Eagle Peak flows near St. Croix Falls and stop in the Clam Falls region), it is also tempting to suggest that these geographically disparate flow units might be a roughly coeval eruptive event. A large sample from this outcrop was processed for baddeleyite and zircon, but did not yield either in sufficient quantity for U-Pb analysis. Preliminary interpretation of new USGS aeromagnetic data suggests that the porphyritic basalts at Stop 7 are stratigraphically below those of Stop 3a. Additional geochemical and stratigraphic studies are needed to confirm this hypothesis.

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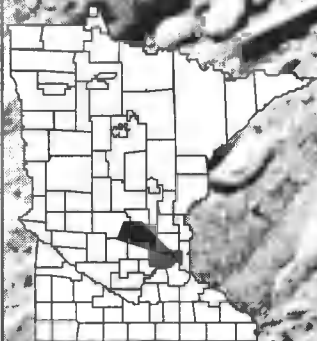
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FIELD TRIP 3

GLACIAL EXOTICA OF THE TWIN CITIES AREA



Leaders

Howard Hobbs

Alan Knaeble

Gary Meyer

Minnesota Geological Survey

FIELD TRIP #3

GLACIAL EXOTICA OF THE TWIN CITIES AREA

Leaders: Howard Hobbs, Alan Knaeble, Gary Meyer

OVERVIEW OF LATE QUATERNARY GEOLOGIC HISTORY

Howard Hobbs

The landforms and surface sediments of this area formed during the most recent glaciation—the late Wisconsinan. The area was covered by previous glaciations some 10 to 20 times, but sediments from these earlier glaciations have been partially eroded by subsequent glaciations and are now deeply buried by sediments deposited during the last glaciation. Although the volume of late Wisconsinan drift in Minnesota is less than half the volume of all the older drifts put together, the amount remaining per glaciation is so much more that it can be explained only by recycling large amounts of older glacial sediment into each new drift sheet. In other words, only a portion of the sediment that we will see was brought from its bedrock source to its present position in one glaciation.

The Twin Cities-St. Cloud area is dominated by the deposits of two ice lobes, each from a different direction (Fig. 1). The Superior lobe advanced into central Minnesota through the Lake Superior basin from northeast. Its till is reddish-brown, sandy to loamy, and rich in red and black rocks from the Superior basin and the north shore of Lake Superior. Typical rocks are red sandstone and siltstone, rhyolite and granophyre, basalt and gabbro. Within the Superior basin, the Superior lobe drift contains practically no carbonate rocks or matrix carbonate, but in the field trip area it contains a modest amount, presumably recycled from older, calcareous drifts.

The St. Croix moraine roughly marks the late Wisconsinan maximum of the Superior lobe. There is evidence from beyond the moraine of a somewhat more extensive lobe that may have preceded moraine-building, but this zone is not very wide. On the southeast side of the lobe, from south of Minneapolis to the St. Croix River, the moraine is wide and high. It ranges from 5-10 miles wide, and the thickness of morainic material ranges from 100-300 feet. Its surface morphology is highly collapsed, indicating that the final event of moraine development was the melting of stagnant ice from beneath thick debris.

At the west end of this same segment, the moraine turns sharply and runs northwest to the St. Cloud area. Most of this segment is buried by glacial sediment of the Des Moines lobe, so is not known in detail. Only the highest parts (generally kames) are exposed at the surface (for example, Stop 1). These kames may represent fans of sand and gravel built at the edge of the ice by subglacial tunnel valleys (Fig. 1.1). The top of the moraine in this segment is lower, due to a combination of the following factors: the base on which it was built seems to be lower; the average thickness of moraine material is less; and some of the material was eroded off by the overriding Des Moines lobe. Inasmuch as this segment represents the frontal part of the moraine, the debris may have been spread over a wider zone by minor fluctuations of the ice front.

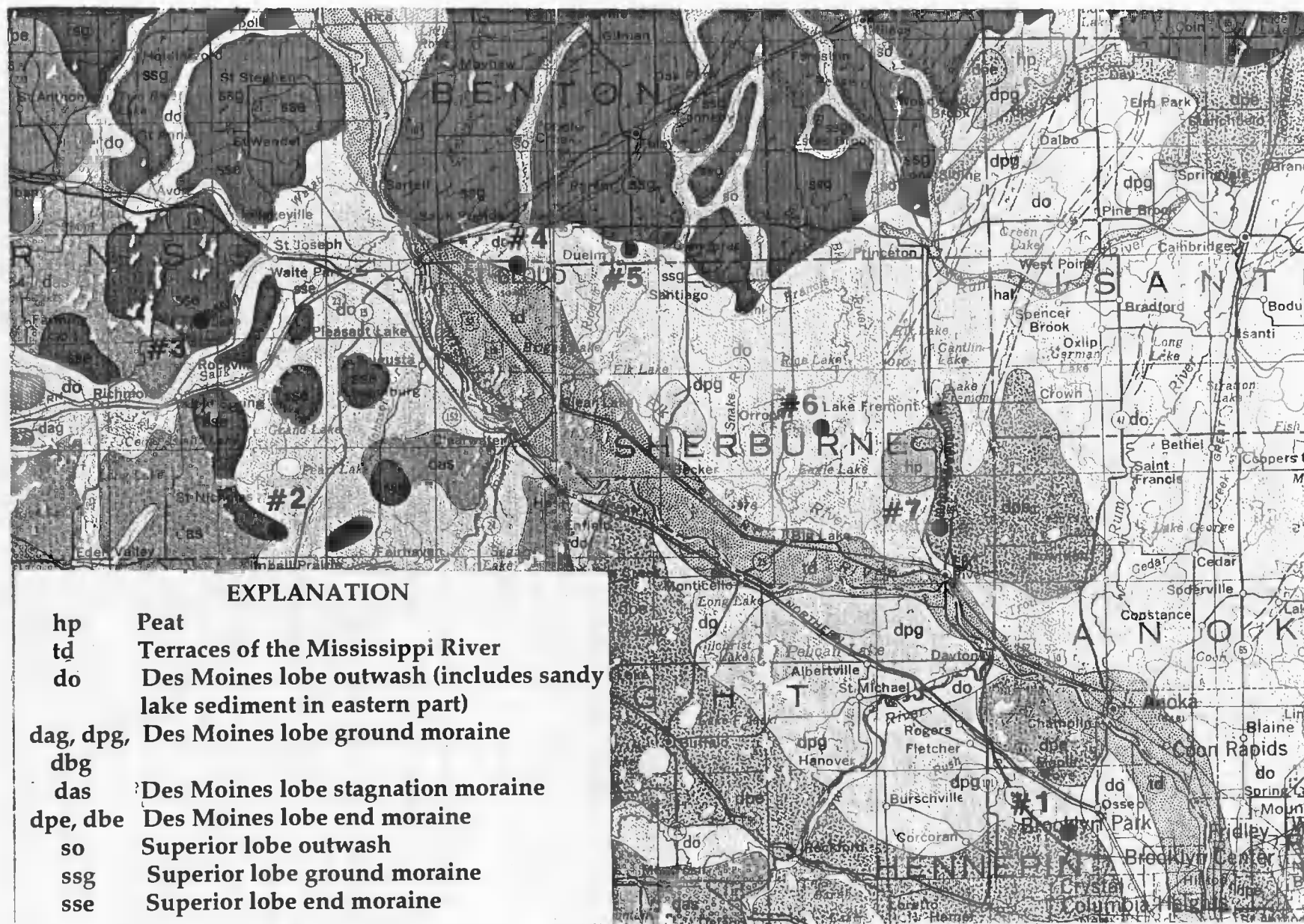


Figure 1. Simplified Quaternary geology of the field trip area, showing major roads and location of stops.

About 15 miles south of St. Cloud, the moraine begins to emerge from under its Des Moines lobe cover. The higher parts of the moraine form isolated bumpy knobs, while the lower parts are covered with outwash of the Des Moines lobe. Many of the higher parts appear to be ice-thrust features, such as the one at Stop 2.

The Des Moines lobe advanced from the northwest, following the broad lowland now occupied by the Minnesota River Valley. Its till is gray (oxidizing to olive brown), loamy to clayey, and rich in sedimentary rock fragments—mainly carbonate and gray shale. Granite and greenstone fragments are also fairly common. An offshoot of the Des Moines lobe, called the Grantsburg sublobe, extended northeast from the main lobe, burying the St. Croix moraine, and advancing over land that had recently been vacated by the Superior lobe.

It appears that the Grantsburg sublobe was a late advance, during a time when the main Des Moines lobe was retreating. The retreat was interrupted by many minor re-advances, each one less extensive than the last. Apparently the Bemis advance of the Des Moines lobe—14,000 years ago—was blocked and diverted by the St. Croix moraine. How was this moraine able to block the earliest, most extensive advance, and yet was overridden by a later, less powerful pulse? The most reasonable explanation is that early on stagnant ice buried in the moraine made it effectively higher. As the ice melted, the barrier became low enough to be breached by the advancing ice.

One could postulate that at the time of the Bemis advance, the St. Croix moraine was backed by active glacier ice, which later retreated enough to allow the Grantsburg sublobe to cross the moraine. However, glacial lake deposits have been discovered well back from the St. Croix moraine, underneath sediments of the Grantsburg sublobe. Johnson and Hemstad (in prep.) estimate at least 900 varve couplets from the thickest part of the lake sediments as extrapolated from thinner counted sections. This suggests that active ice had melted back from the moraine long before the Grantsburg advance.

The Des Moines lobe advance over the St. Croix moraine was in two phases: first to cross was the Duelm Bulge (Stops 4 and 5). The northern limit of this advance is defined by the Duelm channel, which carried the discharge of the upper Mississippi River around the bulge of ice. The eastern limit of the bulge is less certain, but it may have been pinned by the high ground north of Elk River (Stop 7). This advance appears to have been short-lived, and quickly succeeded by the advance of the Grantsburg sublobe.

The Grantsburg sublobe crossed the St. Croix moraine over a broad front, and advanced to the northeast, past the St. Croix river to the town of Grantsburg, Wisconsin. It was hemmed in by the back side of the St. Croix moraine on the southeast side, and by high ground to the north. The Grantsburg appears to have been one main pulse, followed by widespread stagnation and melting.

Till of the Grantsburg advance is quite variable in composition; in places, it is essentially the same as till of the Des Moines lobe. In others, it is banded red and gray, on the scale of feet to inches. In other places, it appears the same as till of the Superior lobe, except for a trace of shale. Apparently, the ice of the Grantsburg sublobe incorporated a large amount of Superior lobe debris as it was passing over the St. Croix moraine. This debris was partially, but not completely, mixed with sediment already in the ice. In places, especially in the center of the lobe, Grantsburg till is thin, or missing altogether.

The ice melted more rapidly in the center than along the margins, presumably because it had less of a debris cover. Glacial Lake Anoka formed in the center of the sublobe, fed by large amounts of meltwater from the upper Mississippi (a continuation of the same flow that eroded the Duelm Channel). The lake declined in stages as ice melted out of the margins, uncovering successively lower outlets into the St. Croix River, and finally drained altogether down the Mississippi River. This drainage event connected the upper and lower parts of the Mississippi for the first time in its modern course.

The Anoka Sand Plain is composed of two parts: the lakebottom sediment (mostly fine sand), and a sandy outwash plain-delta which extends from the Mississippi valley train into the lake. The lake was shallow, and apparently most of the silt and clay from the meltwater stayed in suspension until it was carried downstream. The boundary between the delta and the lakebottom sediment is difficult to pin down, because the texture of the sand is not much different, and because the boundary shifted as the lake filled in.

STOP 1—Superior Lobe Tunnel Valley Fan Sand and Gravel Buried by Superior Lobe and Grantsburg Sublobe Till. (Gary Meyer)

A radiocarbon date of $32,300 \pm 2000$ years B.P. (Beta-40098) obtained from wood fragments in a core sample taken below 130 feet of Superior lobe sand and gravel in Ramsey County (Meyer, 1992), gives a possible maximum age for the advance of the Superior lobe into east-central Minnesota. As the front of the Superior lobe became stabilized, it formed the St. Croix moraine—a complex linear series of high hills encircling the Twin Cities on the east, south, and west, and paralleling the Mississippi River to the St. Cloud area (Fig. 1.1). During the eventual retreat of the ice front, meltwater issuing from the Superior lobe became channelized through the moraine. The Superior-provenance (Table 1.1) sand and gravel deposit at Maple Grove in northeast Hennepin County (Fig. 1.1), extensively exposed in gravel pits (Fig. 1.2), was likely deposited during formation of the North Ramsey mounds (Patterson, 1992), which mark a retreatal position of the Superior lobe behind the St. Croix moraine.

Table 1.1. General characteristics of mapped glacial deposits of the Twin Cities area.

Feature	Riding Mountain Provenance	Superior Provenance
Till texture	loamy	sandy
Color		
oxidized	yellow brown to olive brown	red brown
unoxidized	gray	gray to red gray
Pebble type		
white to buff carbonate	common	rare to uncommon
dark gray to gray-green rocks	uncommon to common	common to abundant
red felsite and sandstone	rare to uncommon	common
gray shale	uncommon to abundant	absent

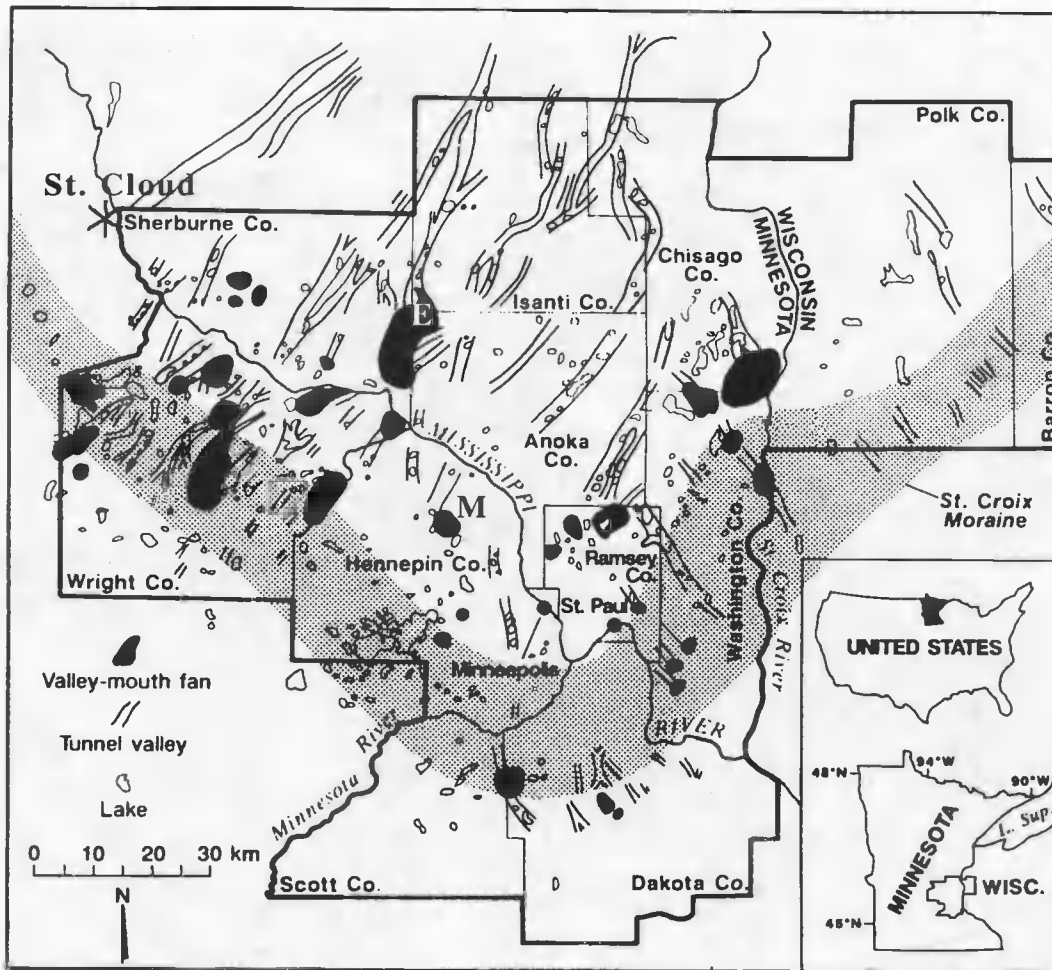


Figure 1.1. Tunnel valleys, valley-mouth fans, and the St. Croix moraine in east-central Minnesota and adjoining counties in Wisconsin. Lettered features are landforms discussed in the text: M, Maple Grove fan; E, Elk River fan complex. Modified from Patterson (1994).

The Maple Grove sand and gravel deposit was laid down as a fan at the mouth of a tunnel valley—a deeply incised, broad, subglacial drainageway (Patterson, 1994). The trend of the tunnel valley that fed the fan is marked to the northeast in Anoka County by a chain of lakes (Fig. 1.1) formed when thick ice filling the valley later melted. The sand and gravel laid down at the mouth of the valley evidently buried stagnant ice, as the fan deposit is deeply pitted (Fig. 1.2). The buried ice was well-insulated, as it apparently was still melting during a readvance of the Superior lobe, which buried the sand and gravel below reddish sandy till, and then during an advance of the Grantsburg sublobe, which further buried the sand and gravel with yellowish clayey till. The sand and gravel is thickest where the buried ice was thin, forming ridges in the postglacial landscape. Areas underlain by the thickest ice became depressions into which first Superior lobe till from the readvance (Automba?) phase flowed, and then, following final deglaciation, both Superior lobe and Grantsburg sublobe tills flowed as the ice finally completely melted, apparently thousands of years after it was first deposited. A log at the base of collapsed lake sediments overlying collapsed Grantsburg and Superior lobe tills, exposed in a Maple Grove fan pit, yielded a date of $11,930 \pm 60$ radiocarbon years B.P. (Beta-23746; Meyer and Hobbs, 1989), indicating that buried St. Croix phase ice was still melting less than 12,000 years ago.

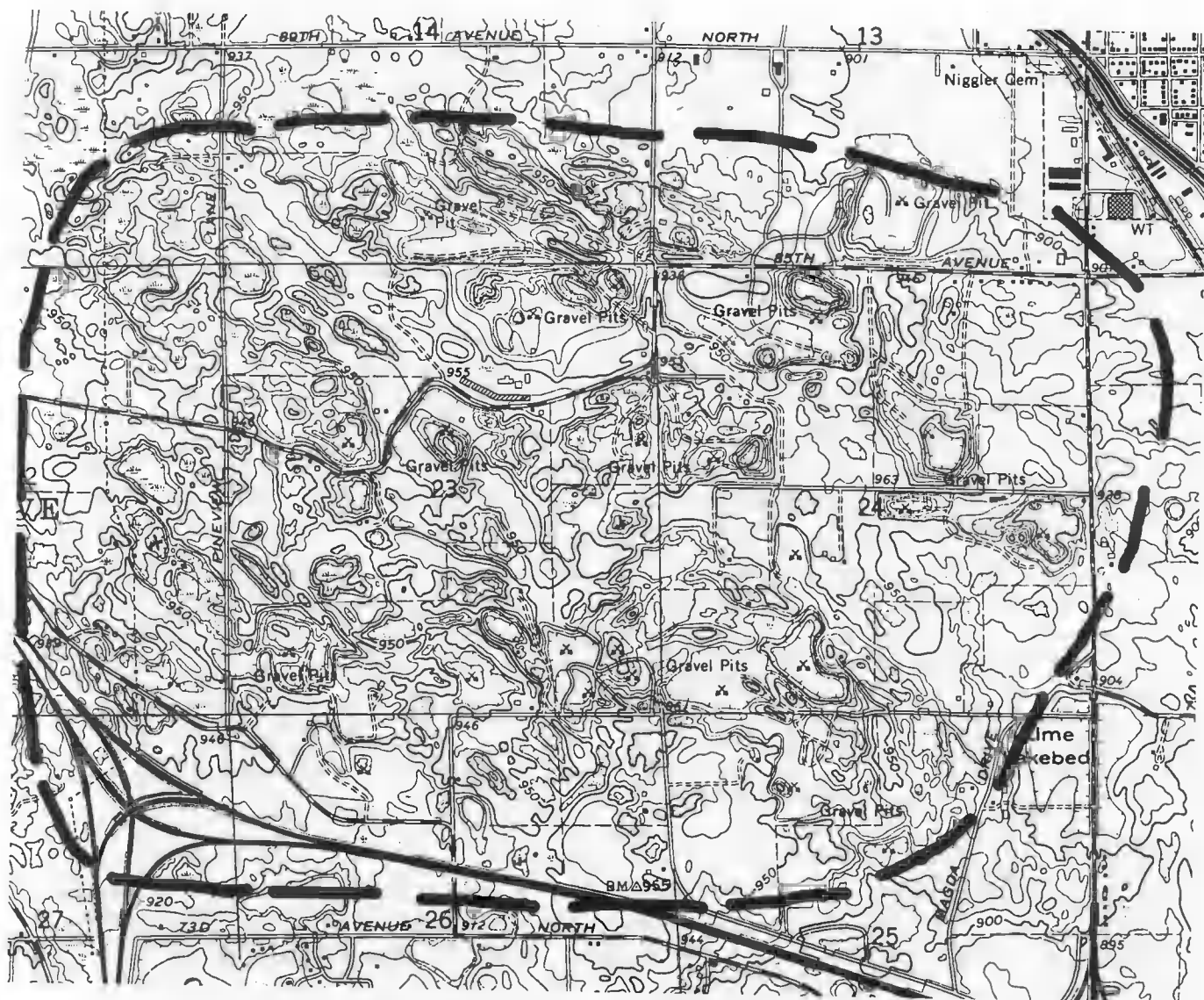


Figure 1.2. Topographic map with Maple Grove fan deposit outlined. The landform likely extended to the northeast prior to being planed off by Grantsburg sublobe meltwater. Area shown is about 2.8 miles wide; contour interval 10 feet. (U.S. Geological Survey Osseo quadrangle, 7.5-minute series, 1967.)

The movement of the bulk of overlying tills into depressions during differential ice-block melt-out reduced the “overburden” thickness above the sand and gravel, allowing it to be exploited for construction aggregate. Gravel mining has inverted the topography once again, turning the gravel ridges into deep pits and the till-filled depressions into hills. Gravel from the Maple Grove fan deposit is the most durable in the Twin Cities area as it consists mostly of Precambrian rock clasts (Meyer and Jirsa, 1984). Superior lobe deposits to the east and southeast contain greater amounts of locally-derived, relatively soft Paleozoic clasts, eroded as the lobe moved up the escarpment formed by the Oneota Dolomite (Jirsa and others, 1986). Des Moines lobe and Grantsburg sublobe deposits contain clasts of both Paleozoic carbonate and Cretaceous shale, as they are of Riding Mountain provenance (Table 1.1).

STOP 2—Powder Ridge Borrow Pit Site—Glacial Thrusting of Drift and Cretaceous Bedrock. (Alan Knaeble)

A large borrow pit is located in the St. Croix moraine (Fig. 2.1). The moraine was deposited prior to 16,150 years BP (Birks, 1976) when the Superior lobe was at its maximum. This site is perched about 225 ft above the surrounding Des Moines lobe outwash plain (Fig. 2.2). The borrow pit was excavated to supply material to build up the summit of the ski hill and regrade existing ski-run slopes. The pit is 1/4 mile long and oriented north-south. There are good exposures on the north, west, and east walls (Fig. 2.3). Six texturally and lithologically distinct units are recognized in the pit including: Cretaceous shale (K); pre Wisconsin sand (S); unoxidized, dark gray, clayey pre-Wisconsin till (Elmdale till) (wt-w); unoxidized, dark red-brown pre-Wisconsin till ("Old Red till") (st-x); oxidized, yellow-brown or unoxidized, gray pre-Wisconsin till (wt-x); and reddish-brown, sandy late Wisconsin till, lake sediments, and sand/gravel (Superior lobe deposits) (st). Fossil and mineral content have helped identification of the Cretaceous deposits, and texture and grain count analyses indicate that at least four glacial advances are represented (Table 2.1; Fig. 2.4) (Knaeble, 1996).

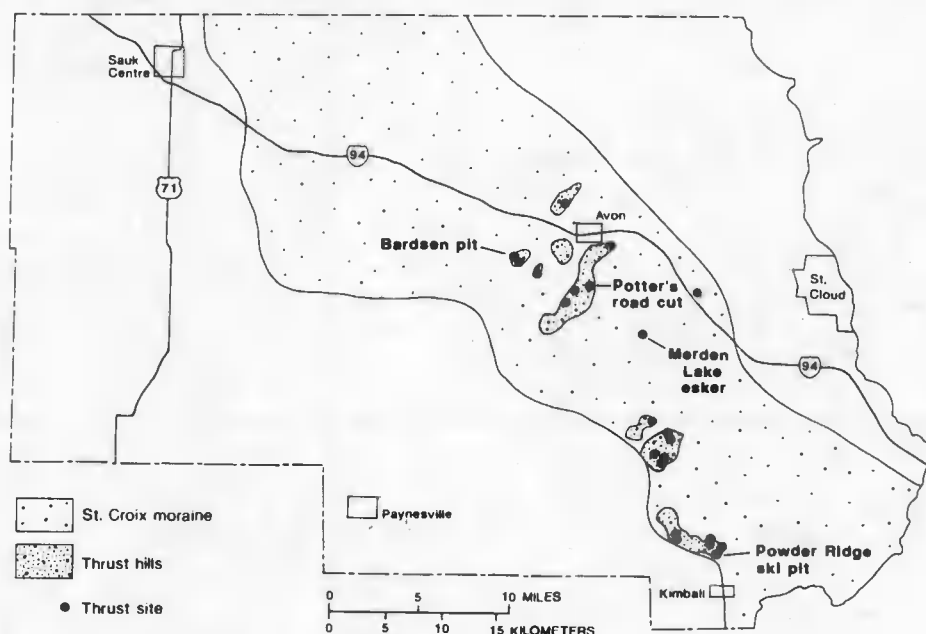
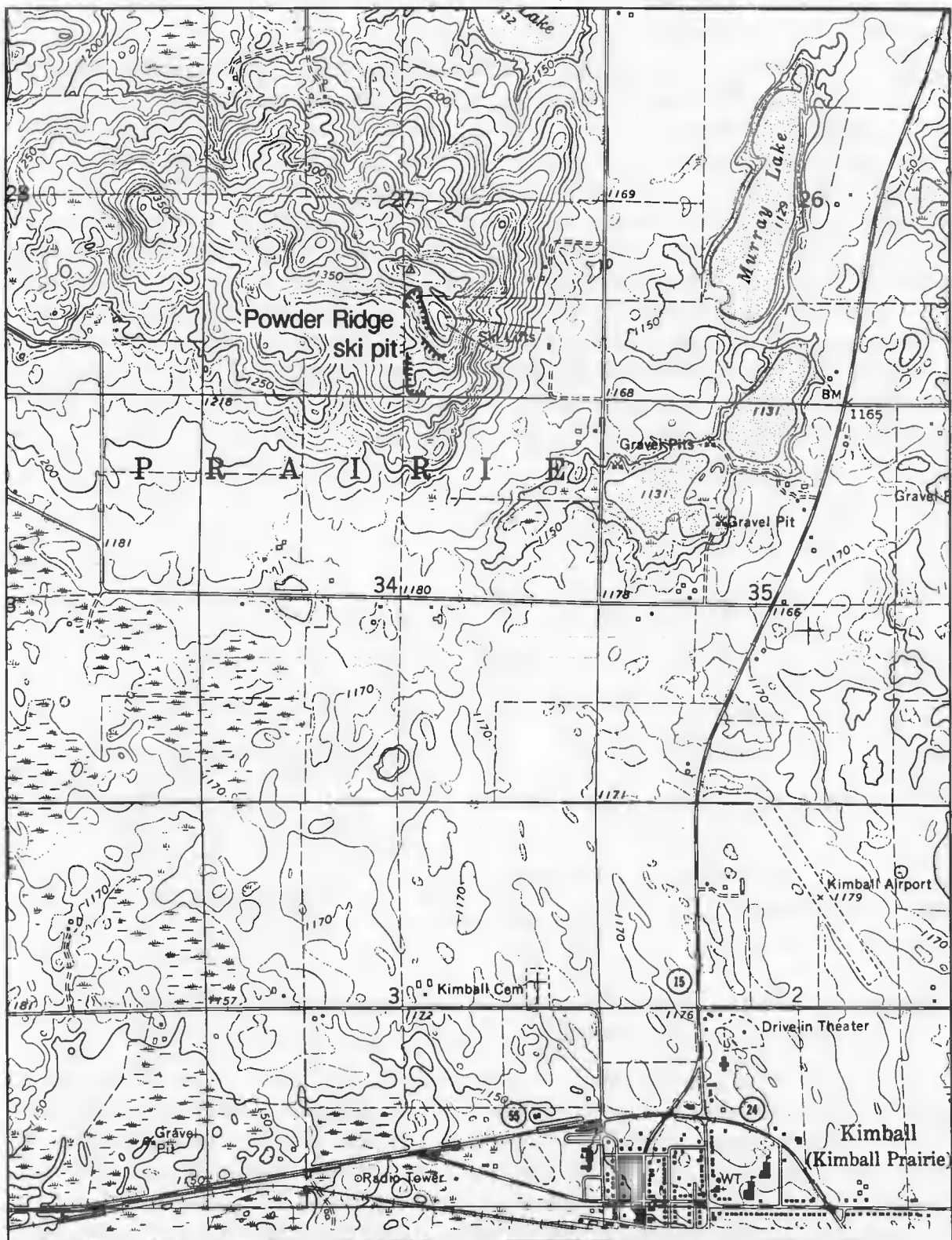


Figure 2.1. Simplified map of Stearns County showing the location of thrust masses (hills) within the St. Croix Moraine and 22 thrust sites. From Knaeble (1996).

Exposures in the pit walls show that these sediments form slabs 50-250 ft thick, which appear to extend laterally for more than 400 ft. The slab contacts dip 30°-60° north-northeast, in what would have been the up-glacier direction for the Superior lobe. The internal stratification within these slabs largely subparallels slab boundaries, although two zones, each 50-100 ft thick, are intensely deformed; the sediments are sheared, streaked, mixed and, in the intensely deformed zone, display distorted overturned folds. Based on comparisons with undeformed exposures in Stearns County, the sequence of sediments in this pit can be divided into at least 6 thrust blocks. Each thrust block contains two or more distinct units differentiated by color, texture, and lithology. The youngest till (unit st) caps each thrust block. Superior lobe ice, associated with deposition of this youngest till, is interpreted to have caused the thrusting (Knaeble, 1996).



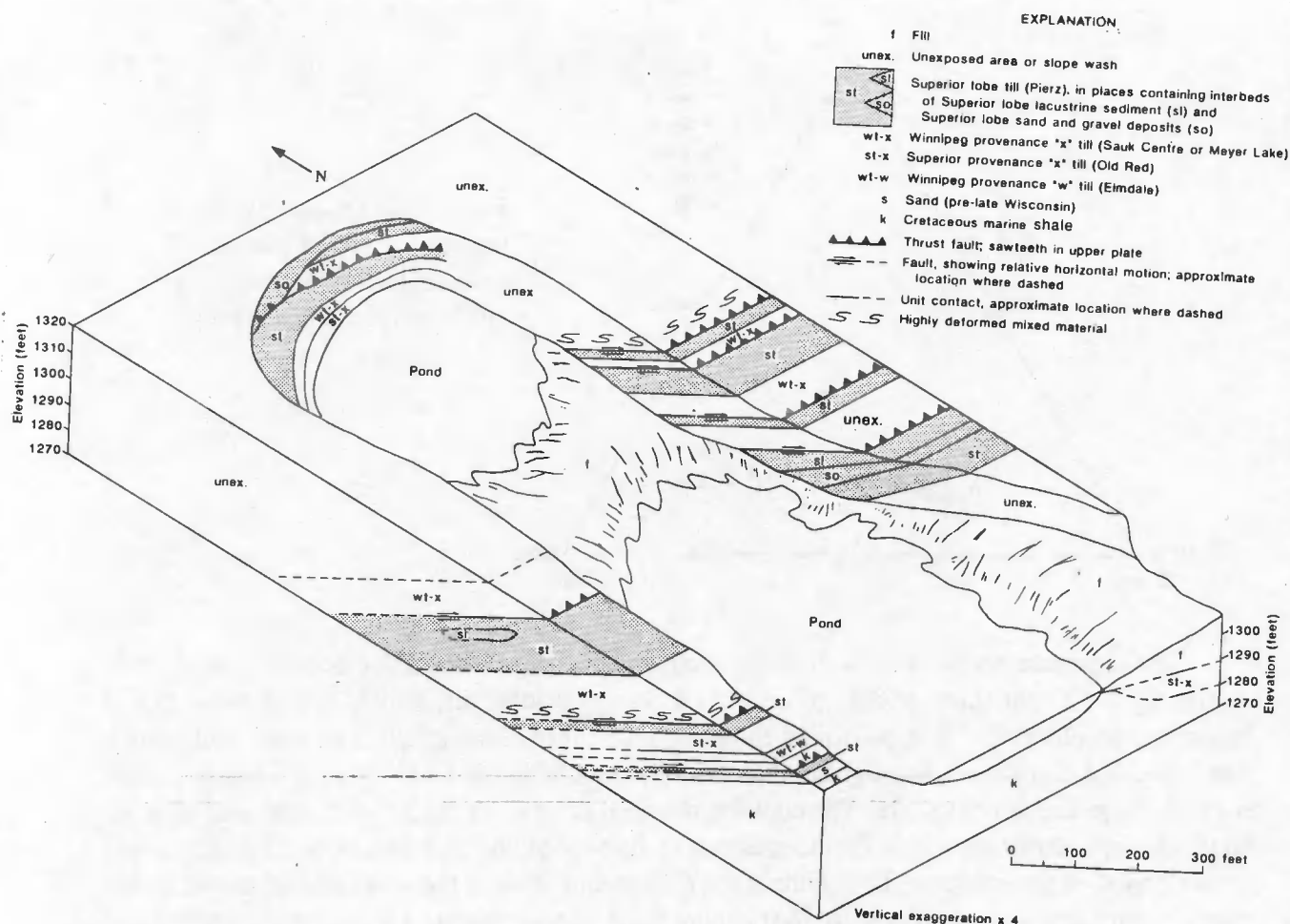


Figure 2.3. Powder Ridge ski pit showing interpreted thrust blocks. Quaternary sediments and Cretaceous clays are exposed on the north, east, and west walls of the pit; the exposure on the west wall has been extended through to the outside face of the block diagram. See figures 2.1 and 2.2 for pit location. From Knaeble (1996).

Table 2.1. Matrix texture of glacial tills in the Powder Ridge ski pit.

Unit	No. of Samples	Percent of		
		Sand	Silt	Clay
wt-w	4	29	40	31
st-x	6	41	34	25
wt-x				
Meyer Lake	5	28	57	15
Sauk Centre	8	41	41	18
st	20	58	29	13

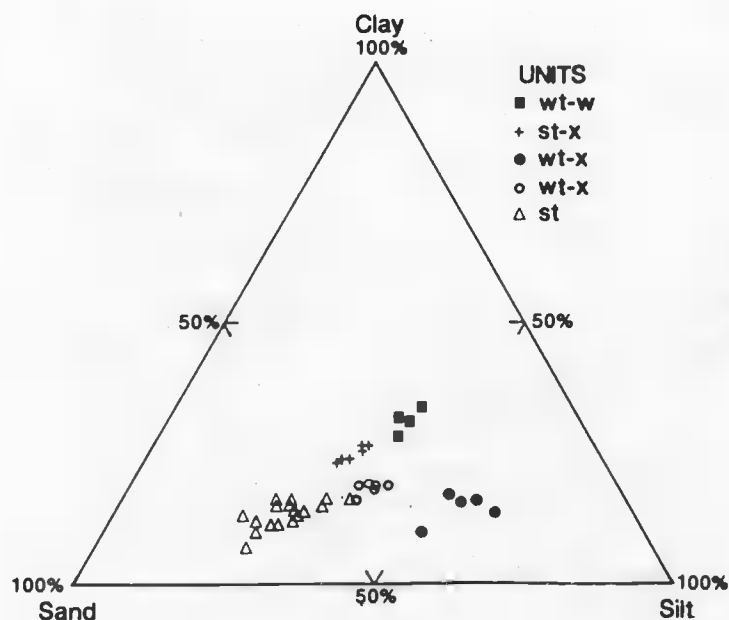


Figure 2.4. Triangular plot of matrix texture of glacial till units in the Powder Ridge ski pit. From Knaeble (in prep.).

The exposure on the west wall at the south end of the pit, shows Cretaceous marine shale overlying a 10-foot-thick packet of deformed Superior lobe till, which contains streaks of incorporated older till. The deformed Superior lobe till overlies 25-30 ft of very well sorted, finely bedded, fine- to medium-grained sand interpreted to be pre-late Wisconsin, which in turn overlies more Cretaceous shale. The contacts between these units dip 30°-60° north and indicate south-directed thrusting with a Cretaceous unit at the base of one thrust block overlying younger units in another thrust block. Drill core of the Cretaceous shale at the south end of the pit shows them to extend to a depth of at least 56 ft, where the drill core ended. The elevation at the top of the Cretaceous shale is about 1300 ft; yet well logs show the highest undisturbed Cretaceous bedrock in the region to be less than 1100 ft above sea level. Some of the thrust blocks contain Cretaceous shale, which is interpreted to have been exhumed and stacked on top of other thrust blocks to form the moraine during ice advance (Knaeble, 1996).

The St. Croix end moraine is about one mile wide at the Powder Ridge ski pit. Water-well records for three wells located on the flanks of the end moraine (two about a half mile northwest of the pit, one located about a quarter mile northeast of the pit) also describe materials that appear to be thrust (i.e., Cretaceous deposits over till, and repetitive sequential color changes in the till units).

There is evidence that the presence of permafrost (Mackay and Mathews, 1964; Clayton and Moran, 1974; Bluemle and Clayton, 1984; Aber, 1988), an increase in pore-water pressure in the underlying sediment (Clayton and Moran, 1974; Bluemle and Clayton, 1984; Aber, 1988), and sufficient ice-flow velocities (Clayton and Moran, 1974; Aber, 1988) are conditions that contribute to thrusting of substrata at ice margins. When the Superior lobe advanced over soft bedrock, such as Cretaceous shale, or older glacial deposits, these conditions may have been present and appear to have fostered thrusting (Fig. 2.5) during the formation of the St. Croix moraine at Powder Ridge.

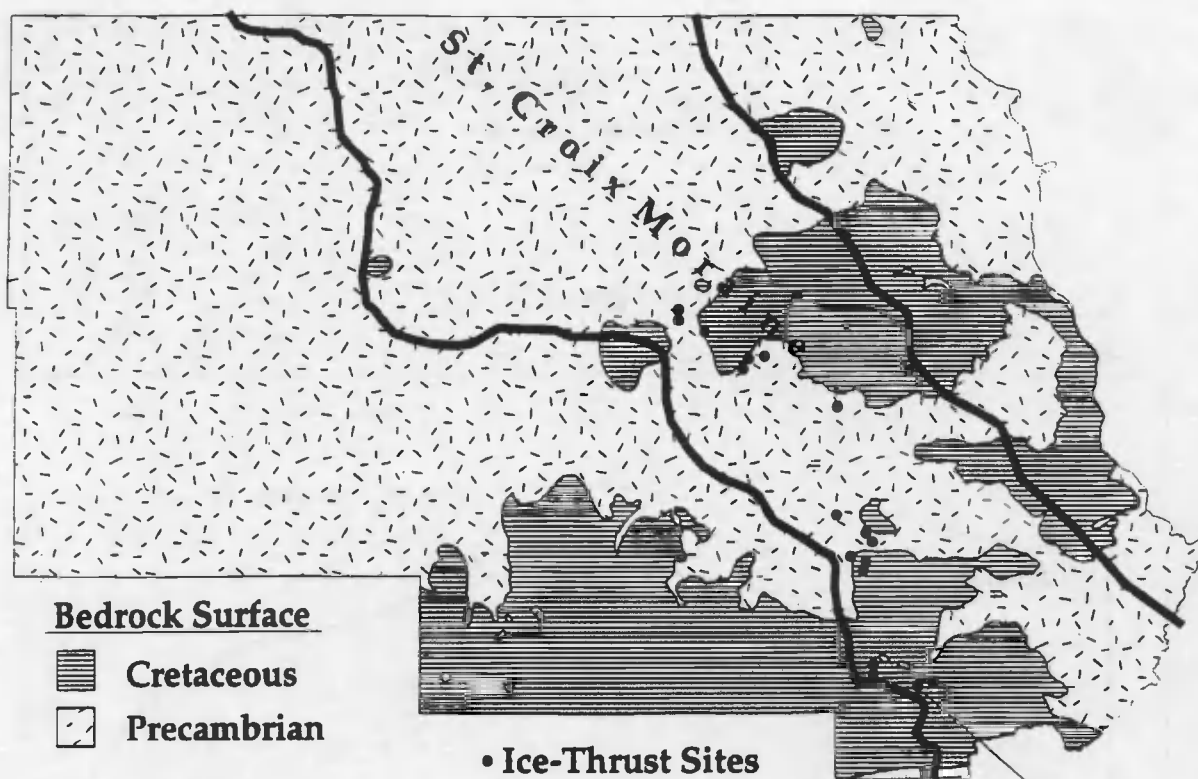


Figure 2.5. Geologic map of Stearns County showing the St. Croix moraine and the association of ice-thrusts with the underlying bedrock.

**Powder Ridge
Borrow Pit**

STOP 3—Merden Lake Esker Gravel Pit—Small-scale Thrusting of Cretaceous Bedrock and Glacial Drift. (Alan Knaeble)

The gravel pit cuts through a section of the Merden Lake esker and provides evidence for small-scale thrusting that is not associated with end-moraine formation (Fig. 2.1). The esker formed beneath the Superior lobe as the ice advanced towards the southwest.

Typical esker deposits are exposed along the east wall of the pit. Here, about 20 ft of Superior lobe sand and gravel are overlain by a thin 5-foot layer of sandy Superior lobe till (Fig. 3.1). In the northwest wall, a Superior-lobe-source sand and gravel is also interpreted to represent the core of an esker. Mantling this core is a 5-10-foot cap of Superior lobe till that has incorporated a 3-5-foot slab of old 'Elmdale till' of Meyer (1986). Thin layers of dark gray Cretaceous shale are present at the base of the Superior lobe till cap. The Cretaceous shale and the older Elmdale till are interpreted as substrata that were picked up by Superior-lobe ice, transported at the base of the glacier, and then deposited on top of the esker as the ice wasted (Knaeble, 1996).

Scattered in the sand and gravel on the pit floor you can find pebbles and cobbles of pisolitic lateritic clay. These clasts are interpreted to be derived from the uppermost weathering profile of the Precambrian igneous and metamorphic saprolite, and transported here by Superior lobe ice and meltwater. Setterholm (1990) states, "The pisolitic layer marks a soil of the landscape upon which the bulk of the Upper Cretaceous rocks in Stearns County were deposited."

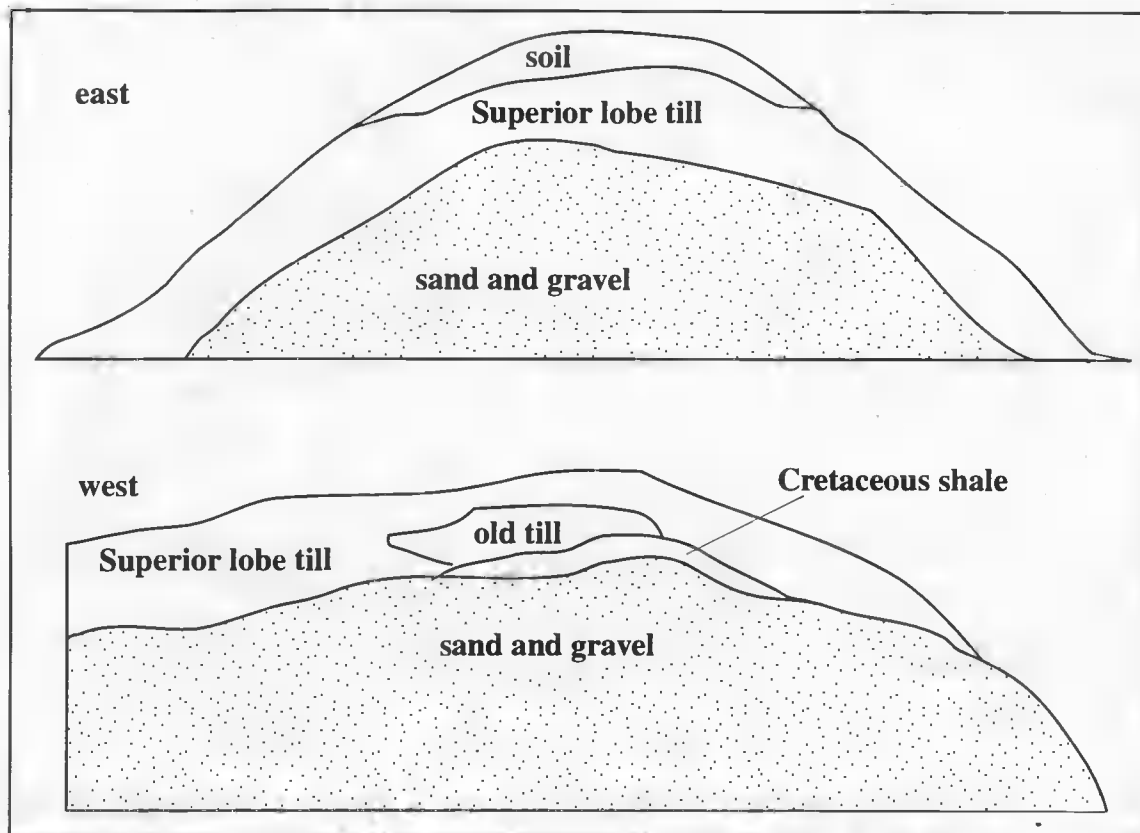


Figure 3.1. Merden Lake Esker gravel pit: profile of exposed east and west walls.

STOP 4— Bauerly Brothers Gravel Pit. (Howard Hobbs)

This pit is dug into the side of a Superior lobe drumlin (Fig. 4.1). The near side of the drumlin is truncated by the edge of the Mississippi valley train. There is about 25 feet of Superior outwash overlain by as much as 25 feet of reddish-brown till. The till cover gets thinner on the lower flanks of the drumlin. It is not unusual for drumlins of the Pierz drumlin field to contain meltwater sediment, presumably deposited in an earlier phase of glaciation, as well as till.

Both the till and the gravel are typical of Superior-provenance deposits. They are reddish brown, and contain many red and black pebbles; carbonate pebbles are sparse. The till is a rocky gravelly sandy loam. In sharp contrast, this sequence is overlain in places by a thin layer of buff-colored loamy till, which contains only a few rocks and pebbles. This material is interpreted as till deposited by the Duelm bulge of the Des Moines lobe. This site is only a few miles from the edge of the bulge, judged by the soil surveys of Benton and Sherburne Counties, and by the position of the Duelm channel. The thin and patchy nature of the till is interpreted as a sign that the ice did not persist long.

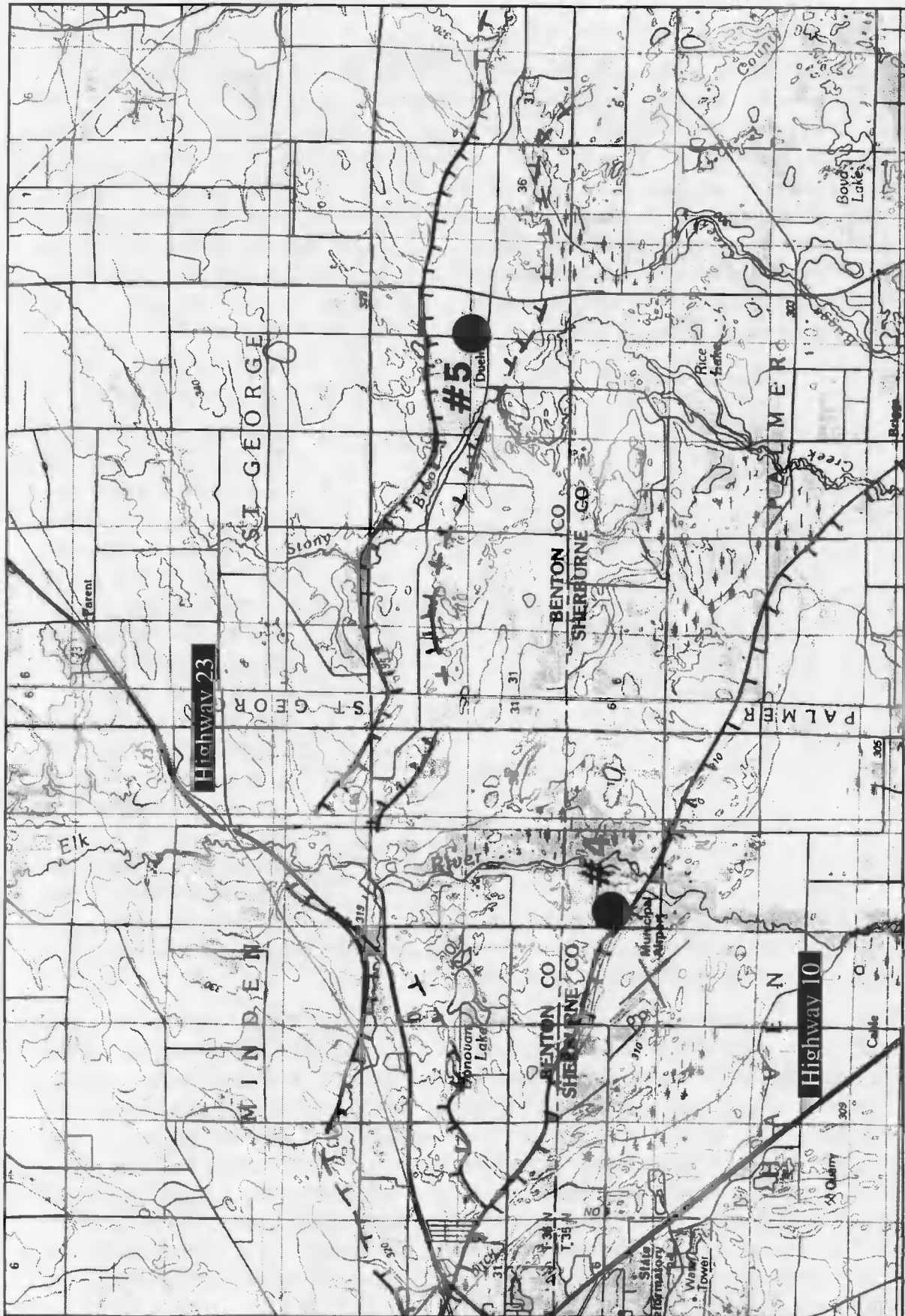


Figure 4.1. Duelm Channel and Mississippi valley train.

STOP 5— Duelm Channel (Howard Hobbs)

From this vantage point, on an erosional remnant of till, the essential elements of the Duelm channel can be seen (Fig. 4.1). The channel stretches as far as the eye can see to the east and west. The northern bank of the channel is about 1/2 mile away, rising from 30-60 feet above the flat channel floor. The land surface to the north is made up of drumlinized till, underlain in places by outwash, of the Superior lobe. To the south, the edge of the channel is not so well defined. In places, the channel cuts several tens of feet into a drumlinized topography similar to the north side, but in other places there is scarcely any elevation difference between the channel bottom and the land outside the channel. In some places, the land actually drops off to the south. It is difficult to see how the channel could have held in any water unless it was fronted by glacial ice on the south side.

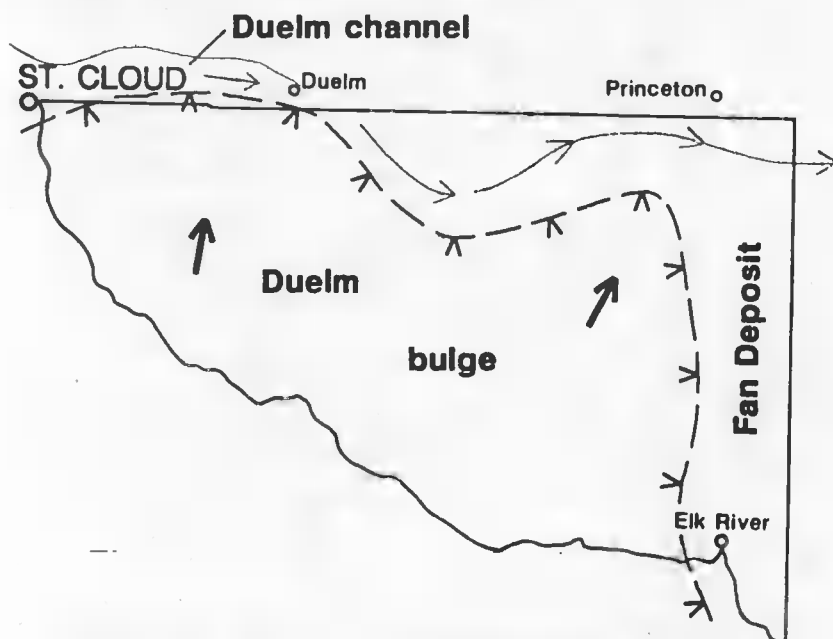


Figure 5.1. Duelm channel and Duelm bulge.

This channel defines the limit of what I have called the "Duelm bulge" (Fig. 5.1). It is lobate, but too small to be an ice lobe or even a sublobe, in the usual sense. All the meltwater flowing down the Mississippi valley train was diverted around this bulge in the ice front, on its way to the Mississippi south of St. Paul. The thin skim of gray, shale-bearing till at Stop 4 indicates that this was a bulge of the Des Moines lobe, which was otherwise held back by the

St. Croix moraine. Because the till is thin, and has not altered the underlying drumlinized topography inherited from the Superior lobe, I interpret the Duelm bulge as a short-lived event. It was probably a single push followed by stagnation.

The Duelm channel originates just east of St. Cloud. Its original junction with the Mississippi valley train cannot be observed, because of subsequent erosion along the valley train. At the upper end, the watercourse is broad and its edge is poorly defined, suggesting that it was not very erosive in this section. In the middle section, it is as we see it here—well-defined and erosive on the north side, variable on the south side. At its lower end, the channel gradually loses its identity as it is buried by younger deposits of the Anoka Sand Plain. South of the Duelm channel is a set of smaller, anastomosing channels at lower elevations. They range from well-defined to obscure, and contain many shallow depressions, apparently a result of ice collapse. Like the main channel, they trend east to southeast, and disappear beneath the Anoka Sand Plain. These channels represent short-lived diversions around the ice as it was advancing, and possibly also as it was melting back.

The northwestern part of the Duelm bulge advanced over relatively high drumlinized till of the Superior lobe, which was not subsequently covered after the ice melted off. By contrast, the southern and eastern parts of the bulge passed over lower ground, which was covered by sand of the Anoka Sand Plain after the ice melted off. Here, it is impossible to tell where the boundary of the bulge was, or where the Duelm channel went. Clearly, the water must have found its way eventually to the Mississippi River below the Twin Cities.

Stop 6—The Lake Ann Dunes, East-Central Sherburne County, Minnesota. (Gary Meyer)
[From Stop 2 by Kerry Keen in Goldstein (1987); see figure 7.1 for location].

Sand dunes are an uncommon feature of Minnesota's landscape. They typically occur as small distinct areas on much larger glacial outwash plains or on terraces along rivers (Cooper, 1935; Grigal and others, 1976). Dunes also occur adjacent to lakes and on beach ridges of Glacial Lake Agassiz in northwestern Minnesota. Some of the largest tracts of well developed dunes are scattered across the Anoka Sand Plain, just north of the Twin Cities Metropolitan Area. Of these, the largest areas of dunes are north of Elk River in eastern Sherburne County. These dune areas are best described as parabolic dune blankets (David, 1977) that contain narrow, elongate, parabolic dunes and en echelon dune ridges.

The Lake Ann dunes (Fig. 6.1) display the typical geometry of dune blankets on the Anoka Sand Plain (Fig. 6.2A). This dune blanket's distinct southeast lobe appears to have advanced into the southern area of Lake Ann. The dune blanket is highly asymmetric, with an indistinct southwest margin and a well-developed northeast wing. It is composed of several southeast-trending dune ridges, which also are very asymmetric, with northeast slopes steeper than southwest slopes. The dune blanket grades northwestward into wind-deflated outwash plain, containing a number of northwest-southeast elongate deflation depressions.

The eolian features of the Anoka Sand Plain were originally attributed to formation by southwest winds (Leverett and Sardeson, 1932; Cooper, 1935). However, study of air photos and other data indicates that the major eolian features of the sand plain — the dune blankets — were formed by northwest winds (Keen, 1985). Modification by west, southwest, and south winds created the asymmetry of the dune blankets and dune ridges. Sand was eroded and transported to the northeast over the dune ridges, steepening their northeast slopes. This modification was so severe that the original parabolic dune shapes were nearly obliterated (Fig. 6.2B). The present wind regime of the sand plain is dominated by northwest winds that are strongest in spring and fall. During the summer, winds from the west and south would result in a smaller net northeastward movement of sand. The present wind appears to be similar to that responsible for eolian activity on the Anoka Sand Plain, although components of this regime probably varied in importance during the Holocene.

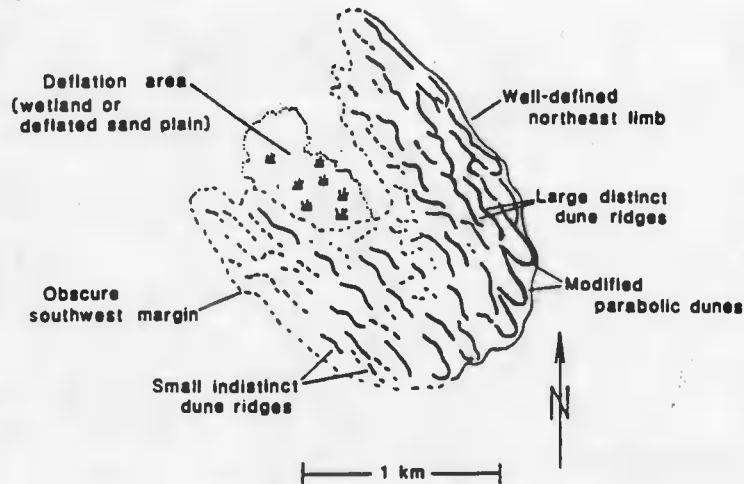
A number of vegetational communities occur on the dunes of the Anoka Sand Plain and are well displayed in the Lake Ann area. Prairie grows on the dry sunny southwest sides of dune ridges, whereas oak woods occupy the moister northeast slopes. An area of low dunes in the Lake Ann dunes supports an oak savanna community. Various wetland communities occupy dune depressions.

Sediments from Lake Ann were studied with the goal of interpreting the vegetation and eolian history of the adjacent Lake Ann dunes (Keen and Shane, 1990). As determined by



Figure 6.1. Topographic map showing part of the Lake Ann dune field. Area shown is about 2.8 miles wide; contour interval 10 feet. (U.S. Geological Survey Orrock quadrangle, 7.5-minute series, 1991).

A.



B.

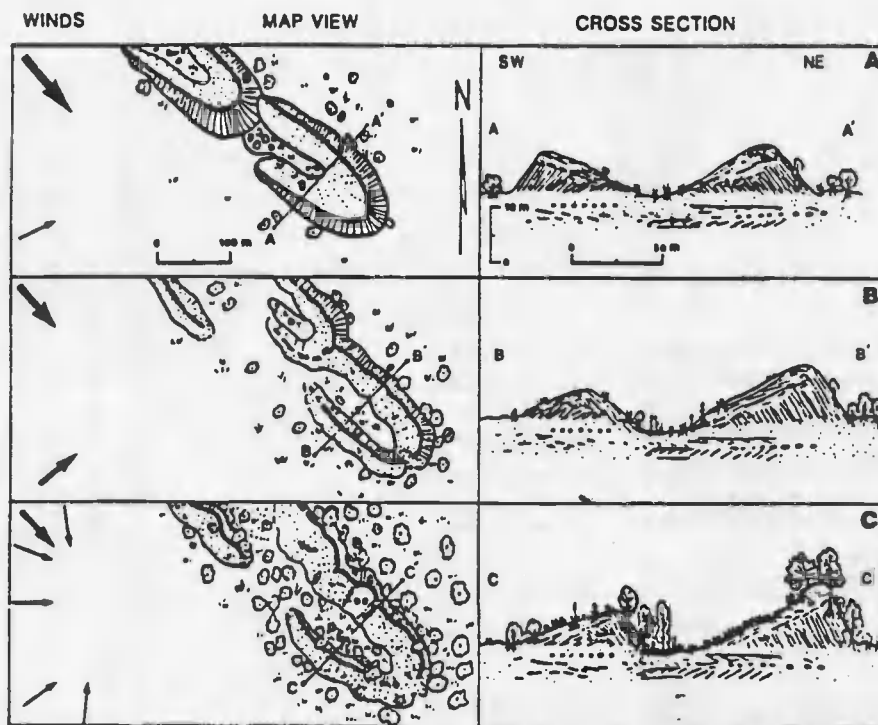
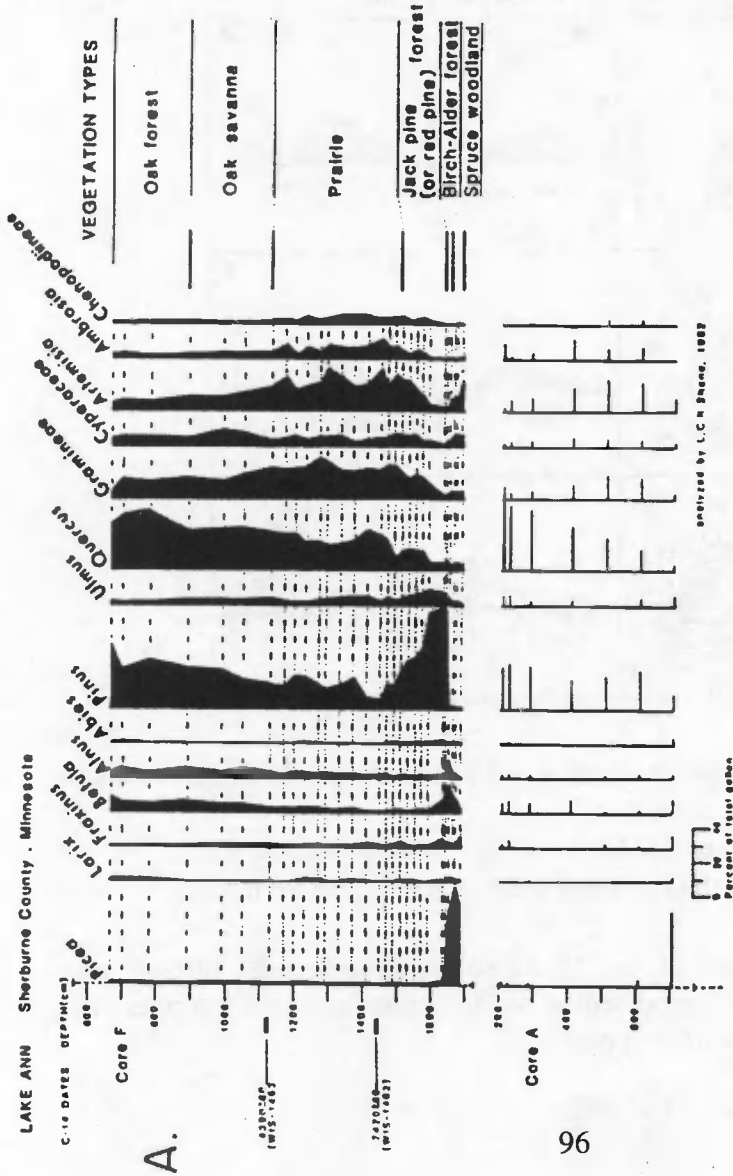


Figure 6.2. A. Typical asymmetric dune blanket on the Anoka Sand Plain. B. Proposed stages in the development of modified parabolic dunes:

- A. Initial development of symmetric parabolic dunes by strong northwest winds.
- B. Modification by cross-winds. Northwest dunes are still strong, but an important southwest wind component exists.
- C. Stabilization and revegetation of the dunes. Minor eolian erosion under variable wind directions. Steep slope (slipfaces) represented by lined pattern; less steep, eroding slopes (backslopes) represented by dotted pattern.

(Stop 2 by Keen, from Goldstein, 1987)

EOLIAN COMPONENT OF MAGNETIC SUSCEPTIBILITY
CORE F, LAKE ANN, SHERBURNE COUNTY, MINNESOTA



pollen analysis (Fig. 6.3A), the postglacial vegetation history of spruce woodland, birch-alder forest, pine forest, prairie, and finally oak woodland is similar to that inferred from other studies in east-central Minnesota. The degree of representation of prairie-indicator species implies very dry conditions on the Anoka Sand Plain during the mid-Holocene.

The chronology of eolian activity for the dunes near Lake Ann, derived from analyzing lake sediment for eolian coarse silt and very fine sand, correlates extremely well with the vegetation history determined from pollen analysis. Because no wind-derived sediment is found in early Holocene sediments, it appears unlikely that dunes were present on the sandplain in late glacial or early Holocene time, when forest covered the land. Eolian activity appears to have begun around 8000 years ago, when the percentage of tree pollen falls to its lowest postglacial value. Major eolian activity, as shown best by the magnetic susceptibility curve (Fig. 6.3B), occurs during prairie conditions of the mid-Holocene, until about 4000 years ago. Minor eolian activity follows throughout the late Holocene.

Stop 7. The "Feature" (Howard Hobbs)

North of Elk River, an extensive highland rises above the Anoka Sand Plain (Fig. 7.1). Its largest dimensions are a little over 15 miles in the east-west direction and a little less than 15 miles north-south. Its average height above the sandplain is about 100 feet; local relief ranges up to 100 feet, but many parts are flatter.

The origin of this highland has been interpreted as a tunnel-valley fan (Fig. 1). However, it has been mantled and modified since its formation, so its surficial sediments are quite variable. Much of it has been mantled by reddish-brown sandy till, similar to till of the Superior lobe. In places, banded red and gray till has been observed. The till-covered part of the feature is crossed by eskers, some of which appear to be Grantsburg, some of which are left over from the Superior lobe. On the west side of the highland is a set of north-south ridges of gravel. They are similar to eskers, but are straighter. Stop 7 is a large gravel pit in one of them.

The gravel is clean and rather coarse in most of the pit at first glance, it looks just like Superior gravel. It has lots of black and red pebbles from the Superior basin and the North Shore. But look closely, and you will find a few gray shale pebbles, which are never found in pure Superior gravels. It appears that most if not all the shale is coming out of the upper part of the exposure, but this is difficult to confirm, because the lower parts slump over so quickly.

Our best interpretation of these ridges is that they were formed as a sort of ice-marginal meltwater deposit where the Duelm bulge was hemmed in by the highland. Most of the gravel is reworked from the existing fan, but a little gravel was derived from the Des Moines lobe. This is where the shale came from.

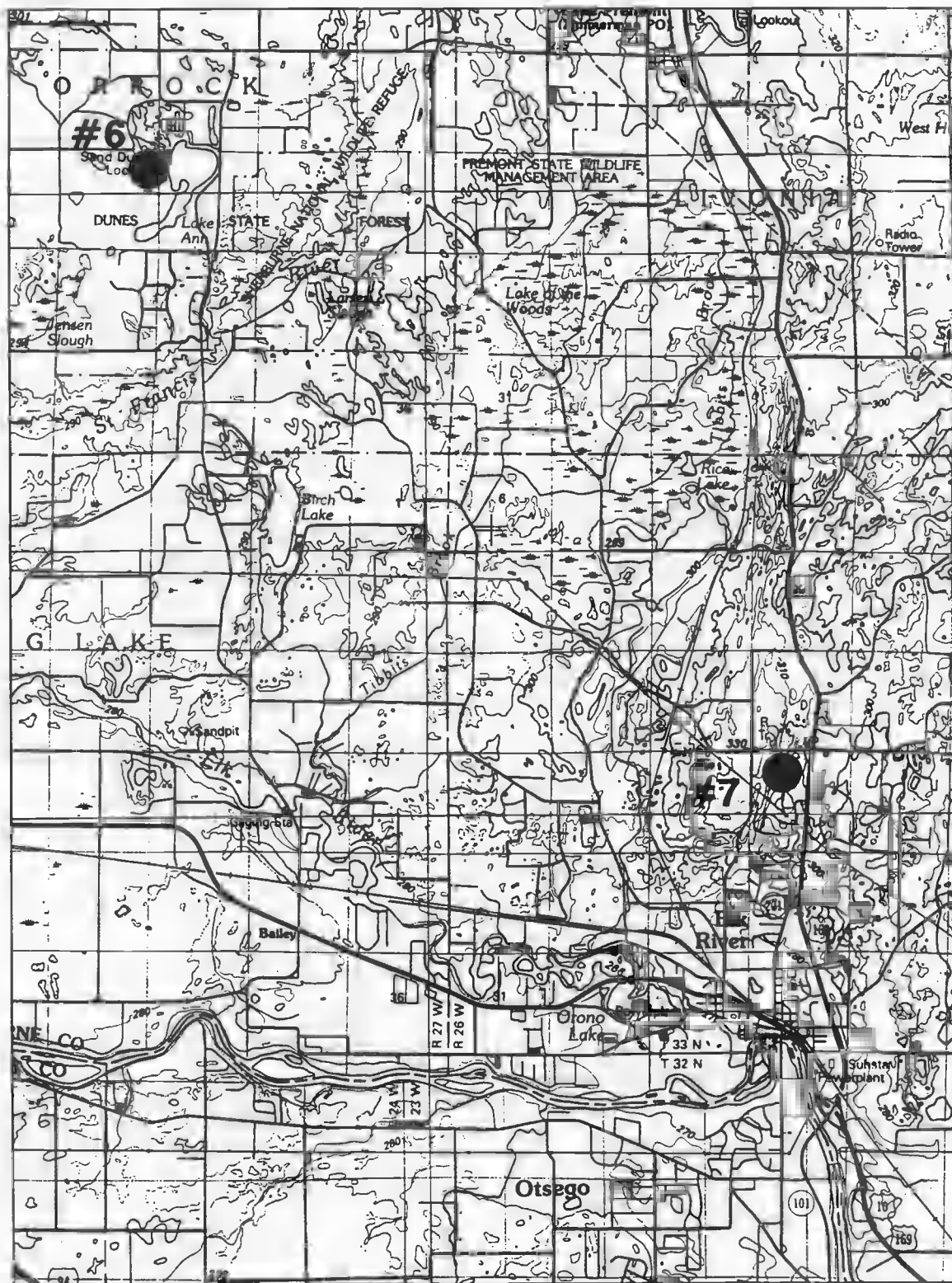


Figure 7.1. Location map for stops 6 and 7.

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FIELD TRIP 4

STRATIGRAPHY AND HYDROGEOLOGY OF PALEOZOIC ROCKS OF SOUTHEASTERN MINNESOTA



Leaders

Tony Runkel
Bob Tipping

Minnesota Geological Survey

FIELD TRIP # 4

STRATIGRAPHY AND HYDROGEOLOGY OF PALEOZOIC ROCKS OF SOUTHEASTERN MINNESOTA

Leaders: Anthony C. Runkel and Robert G. Tipping

Paleozoic strata of the central midcontinent region are among the longest studied and best known sedimentary rocks in North America. This field trip will provide an opportunity to examine the diverse suite of lower Paleozoic rocks exposed along the Mississippi River and its tributaries in southeastern Minnesota, focusing on two interrelated topics: 1) the depositional history of these strata, highlighting advances made over the past fifteen years in understanding the origin of their well documented enigmatic features, and 2) the hydrogeologic properties of these strata, with particular emphasis on characterization of hydrostratigraphic components.

INTRODUCTION

Paleozoic strata in the central midcontinent region of North America consist of thin units of carbonate, sandstone, and shale distributed across tens of thousands of square kilometers. This overview focuses on the northernmost extent of lower Paleozoic strata that were deposited on the stable cratonic shelf northwest of the Illinois and Michigan basins (Figs. 1 and 2). The rocks are exposed in a sinuous belt of outcrops along the southern flank of the Transcontinental arch and the Wisconsin dome and arch in southern Minnesota, Wisconsin and northern Michigan.

Deposition of Paleozoic sediments began in the Middle to Late Cambrian. Coarse siliciclastic sediments of what is now the Mt. Simon Sandstone covered the eroded Precambrian surface. Overlying strata were deposited in two broad, laterally equivalent facies belts across the central midcontinent region (Fig. 1 inset) (Palmer, 1960). An inner detrital belt was composed of shallow marine siliciclastics derived from subaerially exposed Precambrian Shield areas on and to the north of the Wisconsin Dome. Thin, laterally extensive units dominated by either fine- to coarse-grained quartzose sandstone, very fine grained sandstone, or shale were deposited in this inner detrital belt. The middle carbonate belt consisted of subtidal to intertidal carbonate and shale that accumulated to the south. Over time, the boundary between these depositional belts shifted: sedimentation in the inner detrital belt was dominant during the Cambrian and part of the Middle Ordovician; and sedimentation in the middle carbonate belt was dominant during the Early and Late Ordovician, as well as in the Silurian and Devonian (Fig. 2).

Early Paleozoic deposition occurred on a virtually horizontal, cratonic shelf that subsided very slowly. Late Cambrian subsidence averaged less than 10 m/m.y. (Sloss, 1988)—about one-fifth to one-tenth the rate in the contemporaneous Illinois basin (Sargent, 1991), and orders of magnitude slower than that of better known, younger basins in North America. Maximum paleobathymetry was typically less than 100 m (e.g., Byers and Dott, 1995; Ludvigson and others, 1996) and the shelf had a low gradient slope of about 0.1 m/km.

The Wisconsin dome and arch and Transcontinental arch were positive structural features that influenced patterns of early Paleozoic sedimentation, and eventually controlled the distribution and configuration of the gently folded and faulted Paleozoic strata we see today. In

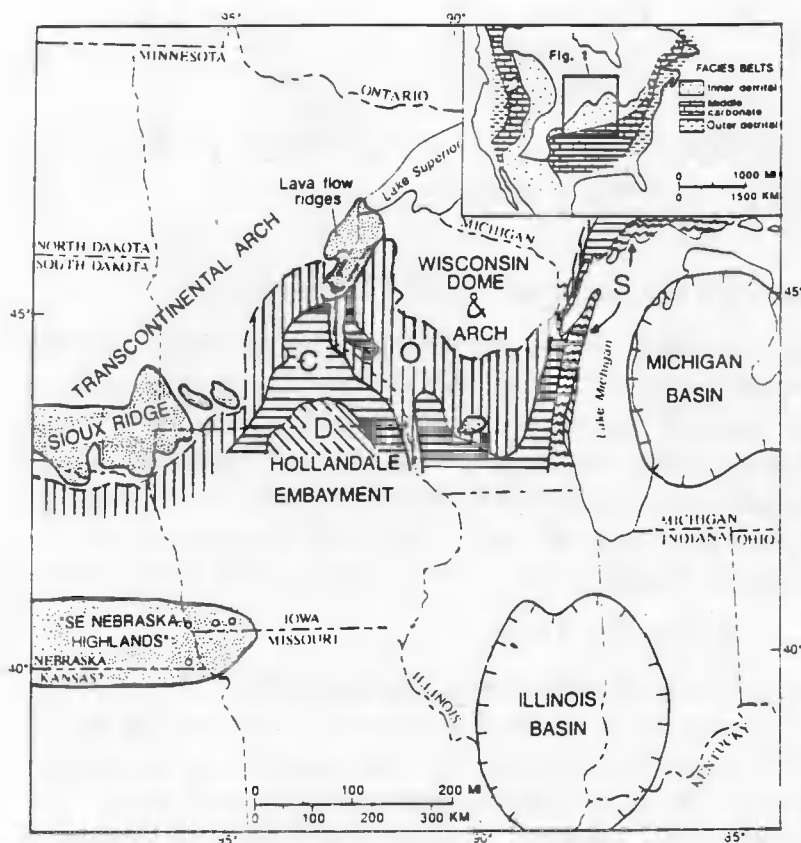


Figure 1. Location map showing early Paleozoic tectonic features, major paleotopographic highs of Precambrian rocks (stippled) and the approximate distribution of Cambrian (C), Ordovician (O), Silurian (S), and Devonian (D) strata in the northern part of the central midcontinent region. The inset shows the study area relative to the Late Cambrian inner (stipple) and outer (dashes) detrital belts and the middle carbonate belt (block pattern) of Palmer (1960). Modified from Runkel and others (1998).

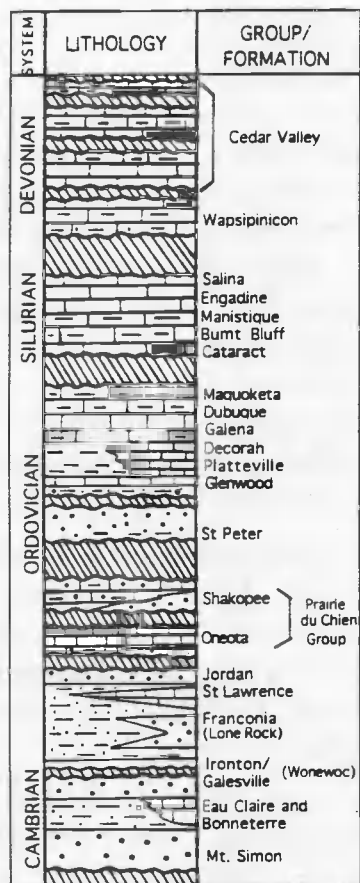
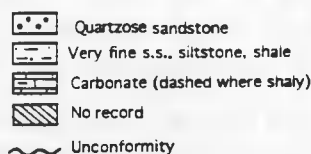


Figure 2. Generalized stratigraphic column (no scale) for lower Paleozoic rocks in the northern part of the central midcontinent region. The Cambrian and Ordovician nomenclature is that used in southeastern Minnesota and Wisconsin, the Silurian from eastern Wisconsin and northern Michigan, and the Devonian from southernmost Minnesota and northern Iowa.



northern Iowa, southern Minnesota, and southwestern Wisconsin, Paleozoic strata are preserved in what is known as the Hollandale embayment, which is a broad syncline that lies between these two positive features. South and east of the Wisconsin dome and arch, Paleozoic rocks dip less than 1 degree into the Illinois and Michigan basins.

DEPOSITIONAL MODELS

Despite over 100 years of study, the depositional history of lower Paleozoic siliciclastic strata of this region remains poorly understood. Many workers have lamented an apparent absence of modern or ancient depositional analogues. In addition, while dozens of local studies have been conducted (e.g., Nelson, 1956; Haddox and Dott, 1990), few are regional-scale investigations that incorporate several lithofacies into a temporally constrained stratigraphic framework. Present-day understanding of early Paleozoic siliciclastic deposition is based mostly on work conducted since 1950, beginning with the predominantly stratigraphic investigations of Berg (1954), Nelson (1956), Berg and others (1956), Bell and others (1956), and Ostrom (1964, 1970). Subsequent work has been chiefly sedimentologic—focusing on the interpretations of near-shore marine facies (Fraser, 1976; Driese and others, 1981; Dott and others, 1986; Haddox and Dott, 1990; Barnes and others, 1992; Mossler, 1992; Runkel, 1994).

In a very important and widely cited study, Dott and others (1986) examined the transition from nonmarine to marine environments recorded in quartzose sandstones, and provided the first set of criteria for identification of nonmarine facies. They described an early Paleozoic terrestrial setting with broad braided streams and eolian sheet and erg systems that delivered sand to a shallow marine environment dominated by small three-dimensional dunes. Dott and others (1986) stressed the *differences*, such as the lack of vegetation, between early Paleozoic depositional environments and those inferred for both younger rocks and modern settings.

A depositional model of a simple, storm-dominated, texturally graded shelf was proposed by Runkel and others (1998) (Fig. 3), and is noteworthy in its *similarity* to models based on both younger rocks and modern systems. Their study was essentially a regional-scale outgrowth of the excellent outcrop work of Charlie Bell and his students from the University of Minnesota in the 1950's. Runkel and others (1998) suggested that early Paleozoic depositional processes and facies distribution are analogous to those depicted in well-known models of the Cretaceous interior seaway and the modern Bering Shelf. The thin, widespread siliciclastic units characteristic of lower Paleozoic strata result from large lateral facies migrations that occurred in response to continental-scale changes in relative sea level.

Carbonate rocks within the lower Paleozoic sequence historically have been regarded as more amenable to comparison with both modern and ancient facies models. Deposition occurred in environments that range from evaporitic, supratidal conditions for parts of the Prairie du Chien Group (Smith and others, 1993) to subtidal settings in depths of 100 m or more for parts of the Decorah Shale (Ludvigson and others, 1996). Most investigations have been local in scope relative to the great lateral extent of these units, and there are too many to cite them individually. Ordovician carbonate strata of the Hollandale embayment are the focus of a monograph edited by Sloan (1986). Silurian carbonates on the western flank of the Michigan basin have recently been studied by Harris and Waldhuetter (1996). There are many excellent studies of Ordovician- through Devonian-age strata in northern Iowa, largely conducted by the Iowa Geological Survey; the results can often be extrapolated to Minnesota and Wisconsin (for

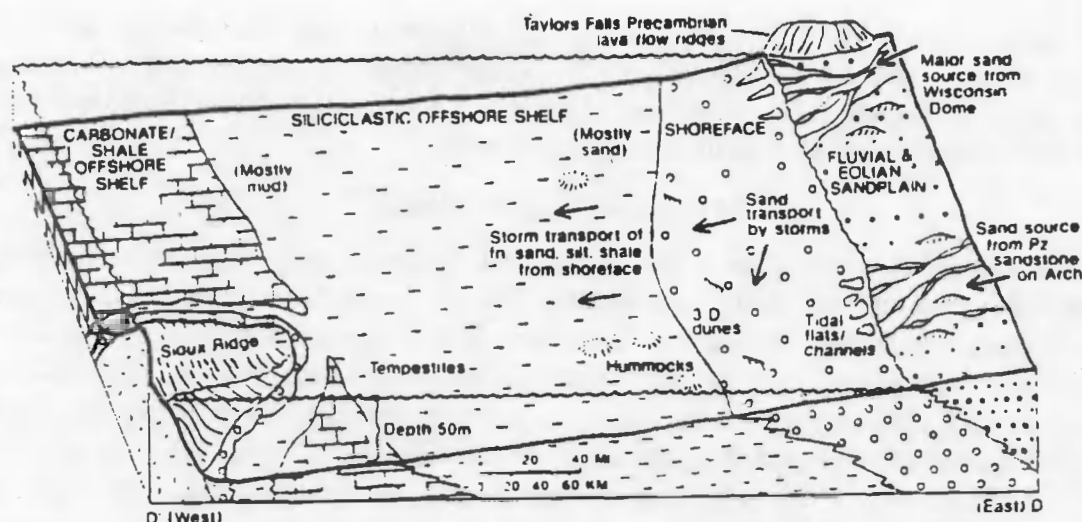


Figure 3. Conceptual depositional model of Late Cambrian storm-dominated, texturally graded shelf. Quartzose fine- to coarse-grained sand accumulated in nearshore terrestrial and shoreface environments. Finer grained, feldspathic siliciclastic sediments and subtidal carbonates accumulated in deeper water of the offshore shelf.

example, several papers in Witzke and others, 1996). Some regional-scale studies have recently been completed; a stratigraphic framework has been developed for the Prairie du Chien Group from the Michigan basin to the Hollandale embayment (Smith and others, 1993, 1996; Barnes and others, 1996). Simo and others (1997) are currently constructing a large scale stratigraphic framework for Ordovician rocks of the central midcontinent in order to address long-standing problems such as identification of the controls responsible for the apparently synchronous regional-scale changes from "tropical" to "temperate" styles of carbonate deposition.

NOTABLE FEATURES AND LONGSTANDING PROBLEMS

Lower Paleozoic strata in the central midcontinent region are well known to geologists outside of the area for several enigmatic features (Dott and Byers, 1980). Most notable is the extreme textural and mineralogical maturity of the fine- to coarse-grained sandstones, and the sheet-like geometry of these and other siliciclastic units. The overall dearth of shale has puzzled sedimentologists, and the fundamental controls on the episodic change from siliciclastic-dominated to carbonate-dominated sedimentation remain poorly understood. Lastly, the presence, position, and magnitude of unconformities has been debated for decades. The remainder of this discussion summarizes the progress made since Dott and Byers (1980) published a similar overview, and offers a guide to topics that can be discussed at the individual field trip stops.

Textural maturity of sandstones

The extreme textural and mineralogical maturity of the fine- to coarse- grained sandstone units, such as the St. Peter Sandstone, may be the best known and longest studied feature in the lower Paleozoic strata of the central midcontinent. The sandstones contain more than 98 percent quartz and most grains are moderately to well rounded. Such textural and compositional maturity could not have been achieved solely by fluvial and marine abrasion, even over transport distances

of hundreds of kilometers, if the grains were derived directly from crystalline source rocks. Rare grains with abraded overgrowths are reworked from older sedimentary rocks, but the volumetric significance of such recycling is uncertain and a source of suitable sedimentary rock has never been identified. Odom (1975, 1978) conducted the most comprehensive mineralogical study of lower Paleozoic sandstones. He noted that the very fine grained fraction is feldspathic, whereas the fine- to coarse-grained fraction contains more than 98 percent quartz (with the exception of some lower Mt. Simon Sandstone beds). He suggested that a long history of abrasion in a marine setting could account for both the maturity of the quartzose grains and the reduction in size of feldspar grains. Dott and others (1986) were the first to clearly identify the presence of substantial eolian deposits within some quartzose sandstones, and they suggested that eolian abrasion could contribute to the textural and mineralogic features described by Odom (1975, 1978). Lastly, chemical weathering in the source area must be taken into consideration (Morey, 1972). Odom (1975, 1978) noted that the very fine grained, feldspathic sandstones are rich in K-feldspar but have only a "trace" amount of plagioclase grains, even though plagioclase was presumably dominant in the source area of the Precambrian Shield. Selective diagenetic leaching of plagioclase grains from the sandstone was discounted by Odom (1978). Thus, chemical weathering probably preferentially dissolved plagioclase crystals and reduced the size and amount of other relatively unstable minerals prior to mechanical abrasion during transport to the shoreline.

While each of the processes described above could have been responsible in part for the mineralogic and textural attributes of lower Paleozoic sandstones, they do not entirely account for the compositional record that appears to indicate that virtually *all* fine- to coarse-grained sand was exceptionally mature when it arrived to the early Paleozoic shoreline. Fine- to coarse-grained immature sand would have almost certainly been deposited locally if the marine abrasion model of Odom (1975) were accurate. Similarly, it is difficult to reconstruct a terrestrial setting in which *all* sand is subjected to prolonged eolian abrasion prior to deposition in a marine environment as suggested by Dott and others (1986).

Sheet geometry

Another long-standing question is how sheet-like siliciclastic layers were initially deposited and then preserved across tens of thousands of square kilometers of the cratonic shelf. Most workers have attributed the formation of such sheets to depositional processes and environments that are markedly different from those described for most younger rocks and modern settings. Lochman-Balk (1970) inferred the presence of enormous tidal flats extending well over two hundred kilometers perpendicular to the shoreline. Dott and Byers (1980) suggested that quartzose sand aggraded more or less vertically into a sheet under high-energy conditions across a vast shallow sea. The most widely cited hypothesis is one that attributes the origin of quartzose sandstone sheets to fluvial and eolian processes that dispersed sand across a vegetation-free landscape prior to marine reworking during a transgression (Dott and others, 1986). Recent studies, however, have demonstrated that some quartzose sheets were deposited entirely within the marine realm under regressive conditions (e.g., Runkel, 1994; Hughes and Hesselbo, 1997; Runkel and others, 1998), and that the depositional processes responsible for the dispersal of siliciclastic sediments were similar to those operating during the deposition of ancient sandstones that are *not* sheet-like.

The lateral persistence and sheet-like geometry of lower Paleozoic siliciclastic units probably reflect the roles of basin physiography and tectonics in controlling the lithostratigraphic architecture, rather than the existence of atypical sedimentary environments. Deposition was characterized by a continuous and abundant supply of sediment to a relatively stable, nearly flat basin that had a slow, uniform rate of subsidence. Individual sheets of siliciclastic sediment were deposited when discrete facies (e.g., Fig. 3) migrated great distances during changes in sea level. Deep incision of the individual sheets during episodes of subaerial exposure did not occur, resulting in more or less uniform preservation.

Dearth of Shale

Lower Paleozoic rocks of the central midcontinent region are noted for a dearth of shale compared to other shallow marine deposits. Two very different hypotheses have been proposed to account for this dearth of shale. In a "marine bypassing" model, Pettijohn and others (1973) suggested that clay- and silt-sized particles were delivered to the shoreline, but subsequently carried in suspension by repeated storms across the shallow shelf before final burial in basinward areas. In contrast, Dalrymple and others (1985) suggested that strong winds blew clay and silt particles hundreds to thousands of kilometers offshore, to be ultimately deposited in the outer detrital belt (Fig. 1).

Recent investigations of Upper Cambrian siliciclastic strata support the marine bypassing model of Pettijohn and others (1973), demonstrating that the absence of shale is characteristic *only* of the outcrop belt on the flanks of the Wisconsin dome and arch (McKay, 1988; Runkel and others, 1998). Basinward of the Wisconsin arch, subsurface records from Minnesota and Iowa show that facies dominated by shale and siltstone are common. These shale and siltstone facies are the offshore equivalent of the shale-poor sandstone facies of the outcrop belt, and their geographic position changed through time in response to changes in the position of the storm-fairweather wave base. The outcrop belt has a dearth of shale simply because the depositional record is dominated by highstand shoreface facies deposited above fairweather wave base. This stratigraphic and sedimentologic attribute is not unique to the lower Paleozoic rocks of the central midcontinent region—it is common in many younger wave-dominated sequences along the landward margins of individual basins. The notion of a shale-free early Paleozoic epeiric sea is a deeply entrenched idea that simply reflects the bias of the outcrop belt. The dearth of shale in lower Paleozoic rocks should be removed from the list of enigmatic features unless the scarcity can be demonstrated to exist on a scale larger than that of the outcrop belt.

Siliciclastic-Carbonate cycles

The fundamental controls on the transition from nearshore siliciclastic- to carbonate-dominated deposition have never been satisfactorily understood. For example, in the Early Ordovician, carbonate deposition apparently dominated across the entire region, even in the shallowest water conditions (Smith and others, 1993). Both older (Jordan Sandstone) and younger (St. Peter Sandstone) shallow-marine environments were dominated by siliciclastics. Only speculative and vague hypotheses have been suggested to account for such fundamental changes. For example, Adams (1978) suggested that the Early Ordovician change in depositional styles may have been in response to drowning of the siliciclastic source area by a shallow sea or the result of siliciclastics being "deflected" to another region. Early Paleozoic carbonate-dominated

systems developed following extended periods of high sea level (Smith and others, 1993), which suggests that the siliciclastic source area may have been covered by a blanket of carbonate rocks. The spread of certain forms of terrestrial life such as bacterial encrustation, and later the development of vascular plants, may have also reduced siliciclastic input.

Cryptic unconformities

Previous investigations of lower Paleozoic strata of the central midcontinent region have also been hindered by an inability to recognize unconformities. The unconformities bounding the major sequences of Sloss (1963) are fairly well established within the strata of the outcrop belt. The presence and position of "lesser" unconformities, especially within siliciclastic units, have been a matter of considerable debate. Stratigraphic relations and interpreted depositional histories indicate that such unconformities "should" be present. However, little or no evidence for regionally extensive subaerial erosion has been documented. Such sequence-bounding unconformities are difficult to recognize where they are contained within coarse siliciclastics because they separate quartzose sandstones that are texturally and mineralogically similar, and because they are relatively flat—reflecting erosion that occurred on a loose, sandy substrate along a low, uniform gradient, and in a nonvegetated terrestrial environment. Furthermore, the ultra-mature mineral composition of the exposed substrate inhibits development of a distinctive weathering profile.

Recent work has shown some promise for identifying subtle subaerial erosion surfaces. Recognition of cryptic unconformities requires interpretation of regional stratigraphic relations in conjunction with physical evidence collected at individual outcrops and high resolution biostratigraphic data. Smith and others (1993) interpreted local deposits of silica cement in uppermost Cambrian and Lower Ordovician strata as subaerially formed silcrete. Runkel and others (1998) suggested that cryptic unconformities within the Ironston and Galesville Sandstones and at the top of the Jordan Sandstone and can be identified only by overlying lag deposits. These lag deposits are the coarsest bed within nearshore sandstone successions; they separate a regionally traceable, decameter-scale, coarsening-upward interval below from a decameter-scale, fining-upward sequence above. Preliminary results of high resolution biostratigraphic dating using conodonts has verified the presence of an erosion surface on top of the Jordan Sandstone, and demonstrates promise for measuring the magnitude of such cryptic unconformities with detailed paleontologic work.

HYDROGEOLOGY

The Paleozoic strata of southeastern Minnesota are the most widely used source of groundwater in the state. Groundwater management and scientific investigations, including predictions of contaminant transport, ground water models, estimates of aquifer yields and water well construction codes have, in the past, relied on a generally accepted hydrogeologic framework that we believe is inaccurate in many important respects. This commonly used classification of aquifers and confining beds (e.g., Kanivetsky and Walton, 1979; Delin and Woodward, 1985) is based on the assumption that Paleozoic lithostratigraphic units in Minnesota are more or less internally homogeneous in hydrogeologic character, and that they have boundaries that correspond to hydrostratigraphic boundaries. Furthermore, this paradigm is based chiefly on the hydrogeologic characteristics of the Paleozoic strata in the Twin Cities Metropolitan Area, which differ markedly from equivalent strata in southeastern Minnesota. A

more accurate hydrogeologic framework is necessary to advance our understanding of ground-water flow in this region.

We have recently made fundamental changes in the methods used to characterize hydrogeologic attributes of Paleozoic strata, focusing on *hydrostratigraphic* components rather than lithostratigraphic units. Fig. 4 shows our preliminary results in delineating hydrostratigraphic components, distinguished and characterized by their porosity and permeability. The Paleozoic bedrock of southeastern Minnesota consists of four distinct hydrostratigraphic components that have been described in studies at both the regional and local scale (e.g., Setterholm and others, 1991; Miller and Delin, 1993; Runkel, 1996a, 1996b). They are: 1) fine clastic; 2) coarse clastic, 3) carbonate rock, and 4) mixed carbonate and clastic. The values for porosity and permeability of these components, discussed below, have been determined through laboratory tests of plug samples and through hydraulic testing of boreholes in southeastern Minnesota.

The fine-clastic component consists of very fine grained, feldspathic sandstone, siltstone and shale in thin to medium beds that are strongly to moderately cemented. This component has low to very low relative permeability—several orders of magnitude less than that of the coarser-grained sandstone of the quartzose component described below. Vertical conductivity based on pumping tests commonly ranges from 10^{-3} to 10^{-5} ft/day for interbedded very fine sandstone and shale (Miller and Delin, 1993), to as low as 10^{-7} ft/day for units composed almost entirely of shale (Freeze and Cherry, 1979). In the former, horizontal permeability is typically more than 100 times greater than vertical permeability (Miller, 1984; Setterholm and others, 1991; Miller and Delin, 1993).

The coarse-clastic component is a mostly uncemented, moderately to well-sorted, fine- to coarse-grained sandstone composed of about 98 percent quartz. Plug-sample tests indicate it has a high to very high permeability and porosity due to relatively large, well-connected intergranular pore spaces. Pump tests of wells open mostly to this component typically range from about 2 -20 ft/day, indicating a moderate to high conductivity, corroborating the results of the plug tests (Miller, 1984; Setterholm and others, 1991; Miller and Delin, 1993; Runkel 1996a).

The carbonate-rock component consists of limestone or dolostone with minor shale and sandstone. Porosity and permeability is due to fractures and solution features, and karst features. The permeability varies substantially from place to place depending on the size, extent, degree and interconnectivity of fractures and solution features (Libra and Hallberg, 1985; Visocky and others, 1985). Pumping tests indicate that this component in most places is moderately to highly permeable where such features are well developed—typically ranging from 1-40 ft/day (Delin and Woodward, 1985). However, in places where fractures and solution features are minimally developed, such as where deeply buried by younger bedrock, the component has a very low conductivity, even at a large scale (Libra and Hallberg, 1985; Visocky and others, 1985).

The mixed carbonate and clastic component is composed of interbedded siltstone, very fine to very coarse grained sandstone, sandy dolostone, and shale (Setterholm and others, 1991; Runkel, 1996b). Cementation is variable; beds of medium- to coarse-grained sandstone are typically friable, but very fine to fine-grained sandstone and siltstone are commonly strongly cemented. As a heterolithic unit, it varies markedly in permeability from bed to bed. Coarse-grained sandstone beds that are too weakly cemented to provide coherent samples are likely to be highly permeable. Laboratory tests of the more strongly cemented, finer-grained, clastic

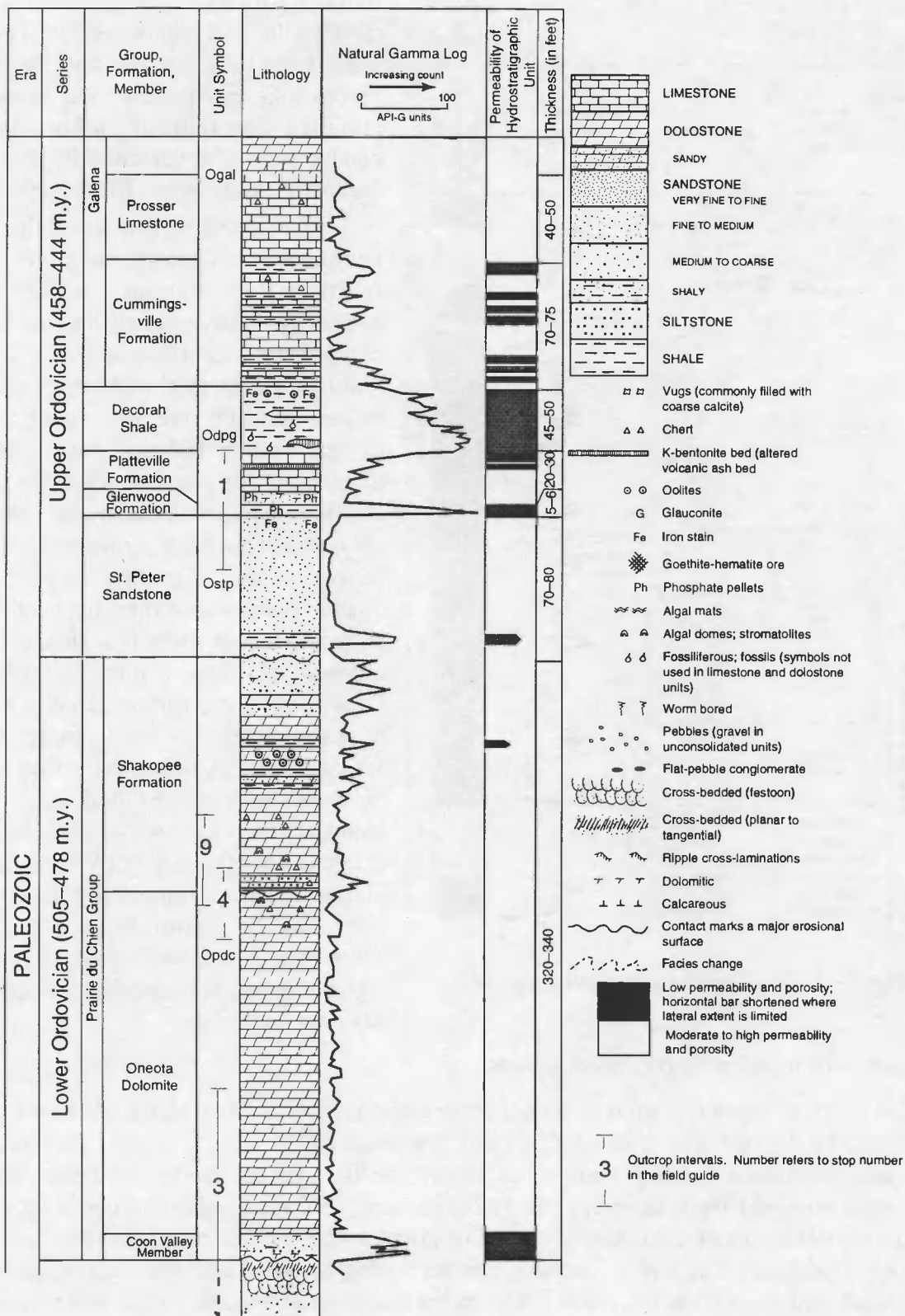


Figure 4. Bedrock stratigraphic column showing units from the Prosser Limestone of the Upper Ordovician Galena Group to the Upper Cambrian Eau Claire Formation. Column includes hydrostratigraphic units based on estimated permeability and porosity.

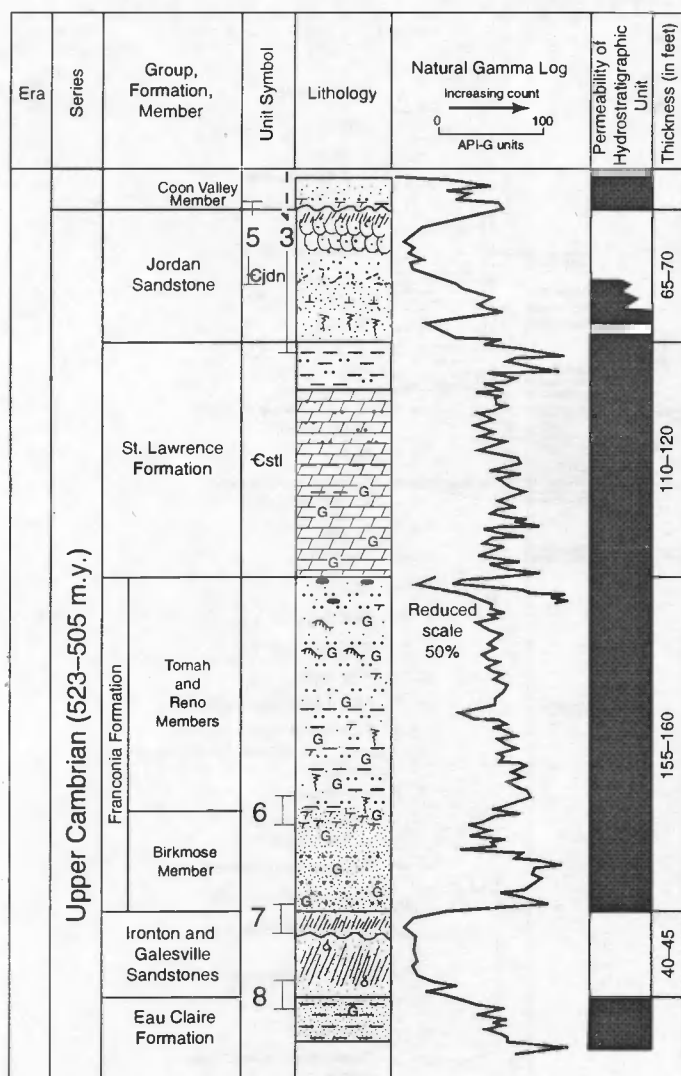


Figure 4. Continued from previous page.

Delineation of aquifers and confining beds

Given the hydrostratigraphic character of the components described above, the Paleozoic section can be divided into moderate/high and low permeability units (Fig. 4). The coarse clastic and carbonate rock component (where karstic) are likely the sources that contribute most of the yield to water wells developed in Paleozoic strata. The fine-clastic component can potentially yield moderate quantities of water, in particular where it is fractured near the surface, but more importantly it acts as a confining unit separating coarse-clastic and karstic aquifers (e.g., Wenck and Associates, Inc., 1997). The mixed carbonate and clastic component contains individual beds that can yield moderate quantities of water, but where greater than 10 feet thick, this unit apparently also acts as an aquitard (Setterholm and others, 1991). Unfractured carbonate rock can also serve as a confining bed (e.g., Visocky and others, 1985; Donahue and Associates, 1991; Barr Engineering, 1996).

beds, indicate a low permeability (Setterholm and others, 1991). These latter beds make up the bulk of this heterolithic component, and meager pumping data indicates a horizontal conductivity of about one-fifth that of the coarse clastic facies (Runkel, 1996a).

Lateral and vertical variability in abundance and interconnectivity of fractures can markedly affect the hydrologic behavior of each of the four components described above. Such features are most abundant and best interconnected where the bedrock is at or near (within 100 feet) the surface. Siliciclastic components such as the fine clastic component, which has low to very low intergranular permeability, may be orders of magnitude higher in conductivity where they lie near the surface because there is a substantial amount of flow along fractures. Conversely, some carbonate units may have a relatively low conductivity and act as confining beds where they are covered by younger bedrock and secondary porosity is not well developed (Libra and Hallberg, 1985; Visocky and others, 1985). Additionally, fracture flow may be dominant even in the coarse-clastic component, which has a high intergranular permeability where it lies near the surface.

Of particular importance to this field trip is that this more rigorous hydrogeologic approach indicates that two widely used aquifers, the Prairie du Chien-Jordan Aquifer and the Franconia-Ironton-Galesville Aquifer, are not single, hydraulically connected aquifer systems as commonly believed (e.g., Kanivetsky and Walton, 1979). Pumping tests (Miller, 1984; Delta Environmental Consultants, Inc., 1992; Miller and Delin, 1993) and carefully collected local static water level measurements (Delta Environmental Consultants, Inc., 1992; Wenck and Associates, Inc., 1997) clearly demonstrate that the fine clastics of the Franconia Formation hydraulically separate ground water in the fractured, upper part of the Franconia Formation from ground water in coarse clastics of the Ironton-Galesville Sandstones below. Some fractured, carbonate-cemented rock in the lower part of the Franconia, where it occurs at or near the surface, is the source of many springs in the area and is perhaps the only highly permeable conduit for groundwater within the Franconia Formation. The Prairie du Chien-Jordan Aquifer is likewise two distinct and separate aquifers—an upper carbonate aquifer and a lower, coarse clastic aquifer—separated by an intervening aquitard composed of the mixed carbonate and clastic component. Hydraulic separation of the carbonate and coarse-clastic aquifers is indicated by several lines of hydrologic evidence including: potentiometric data (Kanivetsky, 1988; Donahue and Associates, 1991; N.O. Janick, Winona State University, oral communication 1994), pumping tests (Barr Engineering, 1996; N.O. Janick, Winona State University, oral communication, 1994), and by ground-water chemistry (Alexander, 1990; Setterholm and others, 1991, Wall and Regan, 1994).

FIELD TRIP STOPS

Introduction

As we examine these rock exposures, think about the enigmatic features and long-standing sedimentologic problems associated with these units. For example, what created the extreme maturity of the fine to coarse-grained sandstones? Is there *really* a dearth of shale? Why is it difficult to place your hand on an unconformity in these rocks? In addition, think about the water-bearing characteristics that are determined by grain size, cementation, fractures, and dissolution features. With this in mind, judge for yourself the validity of some of the commonly accepted depictions of the hydrogeologic characteristics of the Paleozoic rocks. For example, does the Platteville Formation look like a confining unit that will impede the downward transport of contaminants, as it is commonly shown to be on maps? Compare the Franconia Formation to the Eau Claire Formation, and try to determine why the former is most often regarded as an aquifer and the latter as a confining bed. Examine the dramatic variability in the porosity and permeability within the Prairie du Chien Group and Jordan Sandstone and ask yourself whether the commonly accepted hydrogeologic depiction of these units as internally homogenous, hydraulically connected aquifers is accurate.

The location of individual stops is shown on Figures 5, 7, 8, 10, and 11. Directions from major road intersections, or street addresses, are also provided for each stop.

STOP 1—Rochester Public Utilities, 4000 E. River Road NE, Rochester, Minnesota.

An overview of the depositional history and hydrogeologic attributes of the Paleozoic strata of southeastern Minnesota will be presented at this stop. Paleozoic bedrock in the Rochester area are nearly flat lying, and have been eroded into an upper and lower plateau. The resistant limestone and dolomite formations of the Galena Group form an upper plateau that skirts the

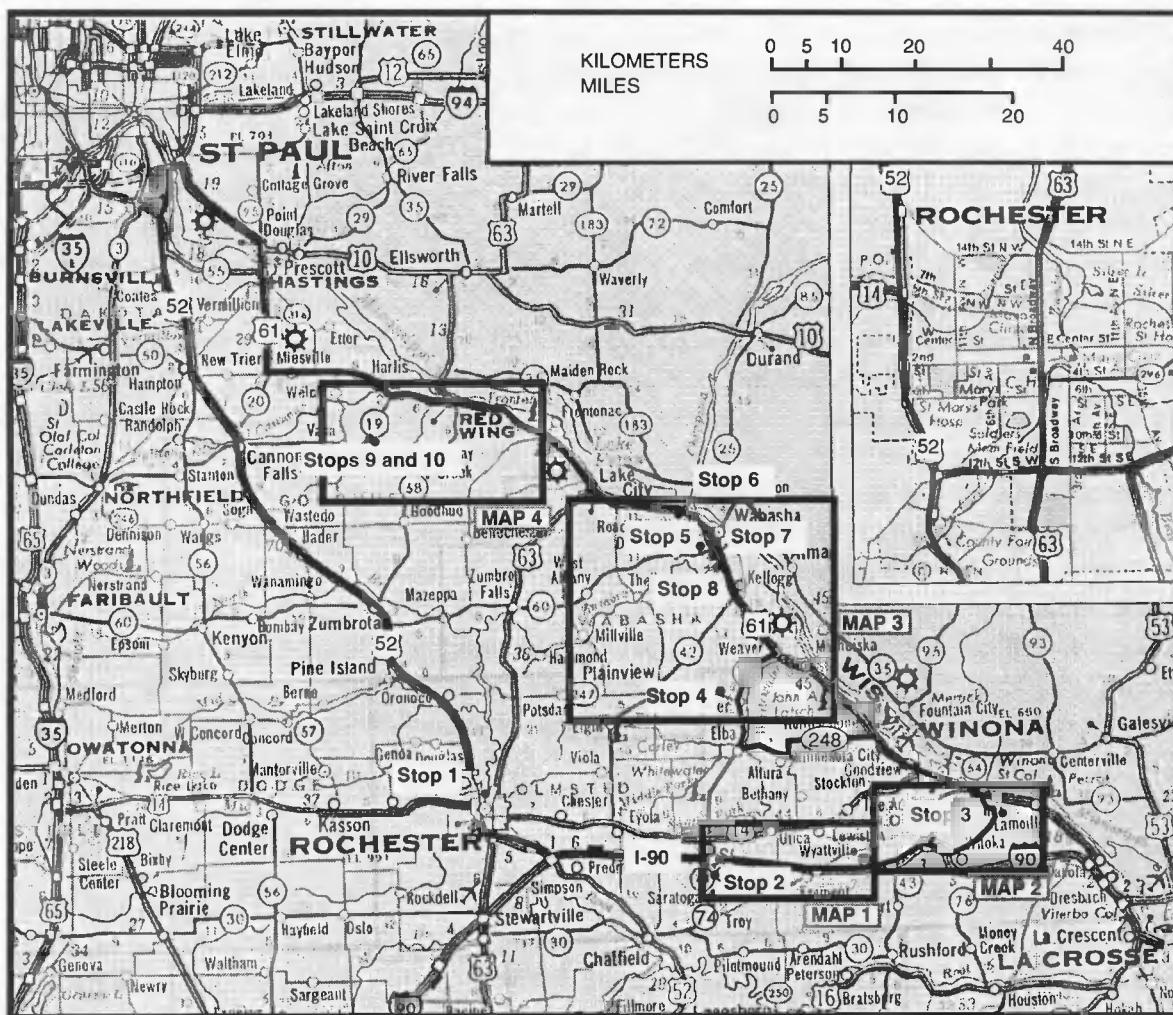


Figure 5. Index map of field stops. Black boxes refer to maps in Figures 7, 8, 10, and 11.

perimeter of Rochester. The mostly carbonate rocks of the Prairie du Chien Group form a lower plateau along the South Fork of the Zumbro River and the lower ends of its tributaries, in and around the city of Rochester. These two plateaus are separated by an escarpment, composed of Decorah Shale, Platteville Formation, Glenwood Shale, and St. Peter Sandstone, which is exposed on steep bluffs within and around the city. We will examine most of these units on the trip today, as well as several older units, such as the Jordan Sandstone.

The hydrostratigraphic attributes of many of the Paleozoic units that we will examine today have a direct or indirect impact on the available water supply for the city of Rochester. The city has devoted a considerable amount of effort and resources to understanding the geologic controls on ground-water quality and quantity in their region. An investigation of variable yield from city wells that draw water from the Jordan Sandstone and lower part of the Oneota Dolomite (Runkel, 1996a) provides an example of where lithostratigraphic and hydrostratigraphic boundaries do not coincide. The open-hole interval of these city wells cross three or more distinct hydrostratigraphic components—the fine-clastic, coarse-clastic, and mixed carbonate

and clastic components (Fig. 6). Permeability values of these components vary by as much as four orders of magnitude. Grain size, sorting and degree of cementation affect the permeability of these rocks, all within boundaries of what is currently designated as a single aquifer unit—the Prairie du Chien-Jordan Aquifer. Pumping records of municipal wells finished in these three components, all considered “Jordan” wells, show that the productivity of individual wells is directly proportional to the thickness of the coarse-clastic component exposed in the open-hole interval (Runkel, 1996a).

This stop shows the sandy dolomite of the Shakopee Formation in the large quarry west of the road. Notable features include the thin cover of the bedrock and the highly fractured nature of the carbonate rock hydrostratigraphic component. The high density of fractures is typical of all bedrock units near the land surface in this region. Also visible to the southeast is the quartzose facies of the St. Peter Sandstone and the overlying Glenwood Shale and Platteville Limestone. Delin (1991) showed that ground-water recharge to the Prairie du Chien Group and Jordan Sandstone is focused along the edge of the plateau above the St. Peter Sandstone. Ongoing investigations into the phenomena of focused recharge by the USGS based on earlier work (Delin, 1991; Delin and Almendinger, 1991) will provide better data on the hydrologic budget of the region by quantifying the amount of water recharging from upper plateau.

STOP 2—Bailey Quarry near St. Charles, Minnesota. West side of Minnesota Highway 74, about 1.2 miles south of Interstate 90. Winona County, Minnesota (Fig. 7).

This quarry has one of the most accessible, continuous exposures across the upper St. Peter Sandstone, Glenwood Shale, and Platteville Formation in Minnesota. Our work on this part of the Paleozoic section is of a reconnaissance nature at this time. Most workers who have studied these units have suggested that they are part of an overall transgressive sequence (e.g., Dott and others 1986; Mossler, 1985). A rise in sea level is recorded by the facies change upsection from fine- to coarse-grained quartzose sandstone deposited in a shoreface (St. Peter SS.), to shale deposited in deeper offshore shelf areas (Glenwood Shale), culminating in micritic carbonates that accumulated in the most distal offshore conditions (Platteville Formation). While this scenario is plausible, some features suggest that the depositional history may be more complex. For example the upper half to two-thirds of the St. Peter Sandstone coarsens upward in southeastern Minnesota, which in other sheet sandstones is indicative of shallowing rather than deepening. Workers in other states (e.g., Simo and others, 1997) have suggested that the St. Peter Sandstone is capped by an unconformity, which also suggests that its deposition was culminated by a regressive event. Additionally, the Glenwood Shale contains discrete beds of medium- to coarse-grained sandstone, some with a channel-shape, and the lower part of the Platteville Formation contains matrix-supported, coarse quartz grains. These attributes are incompatible with deposition in a relatively deep, distal offshore shelf setting.

There are three hydrostratigraphic components exposed here. The Platteville Formation consists chiefly of the fractured carbonate rock component, the Glenwood Shale consists mostly of the fine clastic component, and the St. Peter Sandstone consists of the coarse clastic component. Hence, across this part of the Paleozoic section lithostratigraphic boundaries nearly (within a meter) coincide with the principle hydrostratigraphic boundaries. We will see examples at subsequent stops where the lithostratigraphic boundaries do not correspond to the principal hydrostratigraphic boundaries.

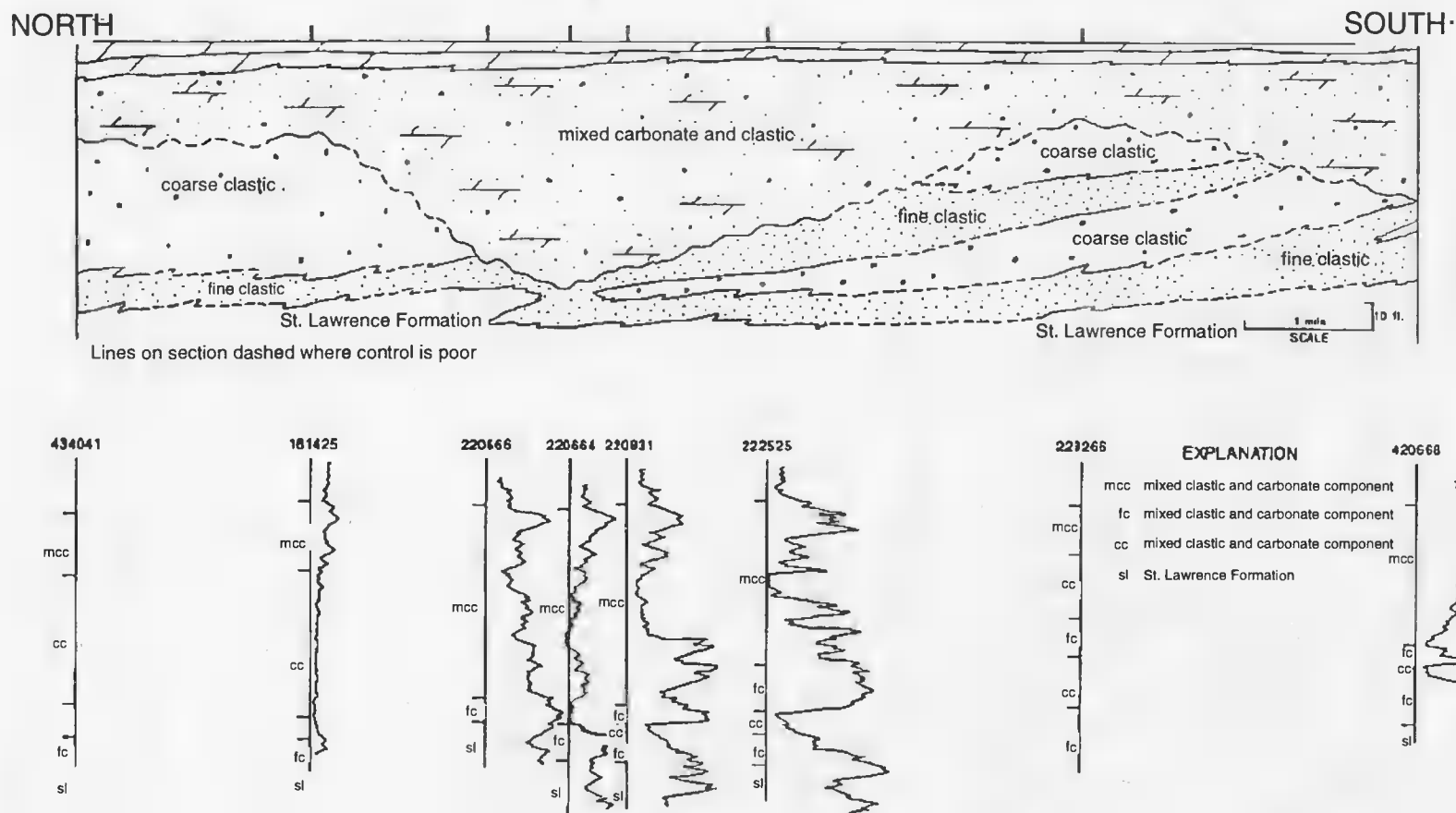


Figure 6. West-east stratigraphic cross-section and corresponding gamma logs for the Jordan Sandstone and the lower part of the Oneota Dolomite in the Rochester Metropolitan area, showing the variable thickness of the coarse-clastic and fine-clastic components of the Jordan beneath the mixed carbonate and clastic Coon Valley Member of the Oneota. (Modified from Runkel, 1996a.)

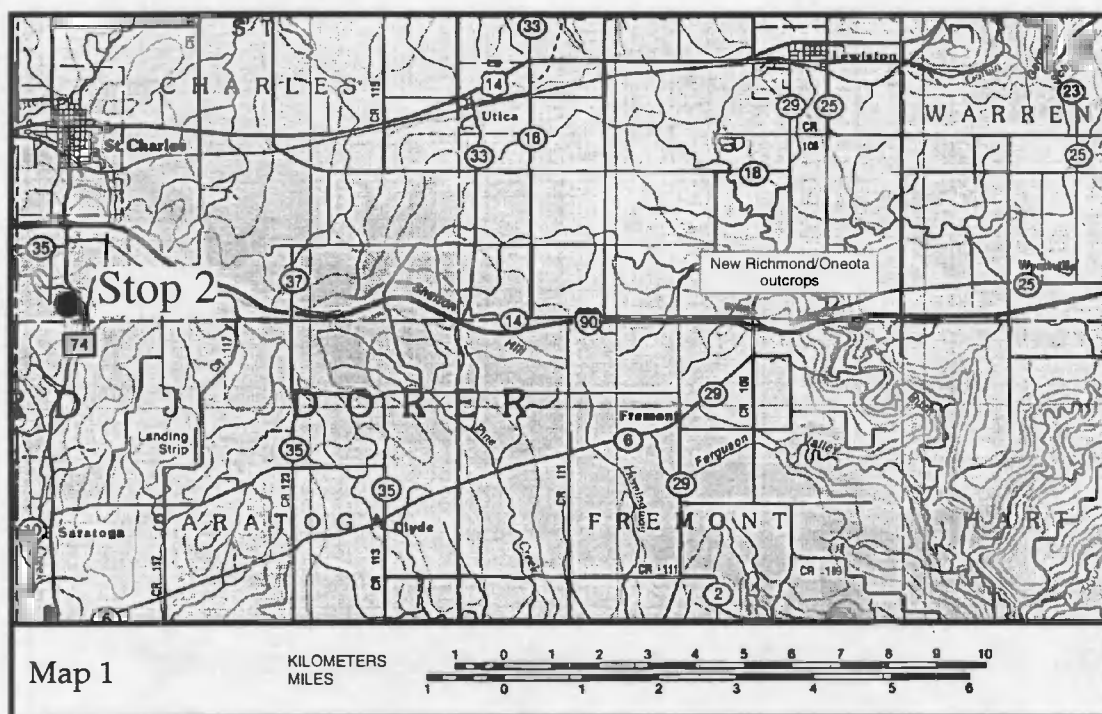


Figure 7. Map 1 on index map (Fig. 5) showing the location of stop 2.

This quarry is a good example of the geologic controls on ground-water recharge discussed at Stop 1. Ground-water flow through the carbonate Platteville Formation is through horizontal fractures along bedding planes and vertical joints. Matrix porosity and permeability does not contribute substantially to ground-water flow. Water pools on top of the lower Platteville bench because the underlying Glenwood Shale restricts the downward movement of ground water. Hydraulic conductivity of the Glenwood Shale is estimated to be similar to values for shale reported in the literature of 10^{-7} to 10^{-3} m/day (Freeze and Cherry, 1979). The surface water entering the quarry runs across the Platteville bench and onto the top of the Glenwood Shale before spilling into the St. Peter section of the quarry. This stream does not account for all the water recharging the St. Peter Sandstone. Joints are visible in the Platteville that extend into the Glenwood Shale. The Platteville Formation and Glenwood Shale form a flat plateau across much of the Twin Cities area. The Glenwood Shale is an important confining bed in that it restricts contaminant transport to lower bedrock aquifers where it is not crossed by fractures.

STOP 3—Road Cut near Homer Minnesota. From intersection of State Highway 43 and U.S. 61 in Winona, take State Highway 61 south for 3.5 miles. Turn right on County Road 15 (at Homer) and travel 2.2 miles to large road cut (Fig. 8).

This is one of the most complete, readily accessible exposures of the Jordan Sandstone in the entire central mid-continent region. The section is continuously exposed from the upper part of the St. Lawrence Formation, through the entire Jordan Sandstone, and most of the Oneota Dolomite. A complete regressive-transgressive cycle is recorded. The very fine grained sandstone, shale and dolostone of the St. Lawrence Formation was deposited in relatively deep water during a high stand of sea level (Hughes and Hesslbo, 1997), which was followed by

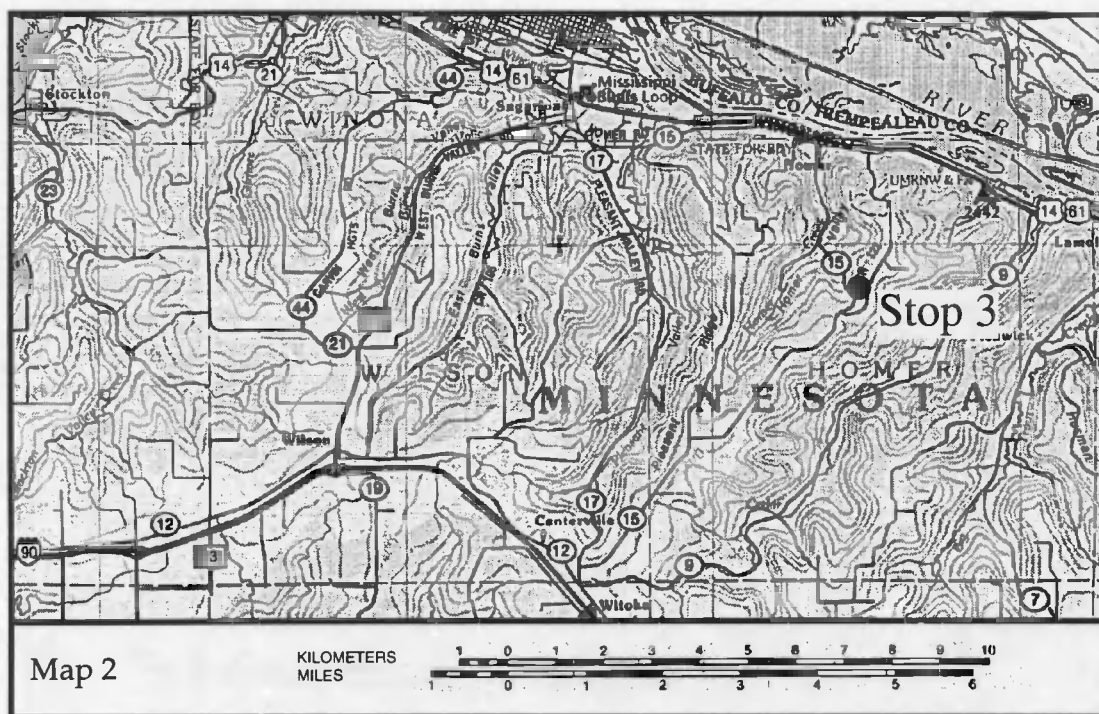


Figure 8. Map 2 on index map (Figure 5) showing the location of Stop 3.

progressive shallowing recorded in the coarsening upward Jordan Sandstone, culminated by subaerial erosion marked by a pebbly sandstone lag. The Oneota Dolomite above the unconformity records the return of the shoreline and gradual deepening in the Early Ordovician. Lowermost Oneota strata (Coon Valley Member) are mixed clastic carbonate deposits that accumulated in supratidal to shallow subtidal conditions. Upper Oneota strata (Hagar City Member) are mostly sand-free dolostone deposited in deeper, entirely subtidal conditions.

Particularly interesting sedimentologic features at this stop include large-scale, high-angle cross strata with reactivation surfaces and shale drapes in the Jordan Sandstone that may record a strong influence of tidal currents, and “cabbage-head” stromatolites in the lower Oneota. The depositional model depicted in Figure 3 can account for the sedimentologic and stratigraphic attributes of the St. Lawrence Formation and Jordan Sandstone, but does not readily explain the attributes of the overlying Oneota Dolomite

This exposure includes an example, where a major hydrostratigraphic boundary does not correspond to a lithostratigraphic boundary, and also shows the great hydrostratigraphic variability within lithostratigraphic units that are widely regarded as internally homogenous in hydrogeologic properties. The St. Lawrence Formation and lower part of the Jordan Sandstone are both composed of the fine-clastic component. The St. Lawrence is commonly regarded as a confining bed and the Jordan as an aquifer, yet there is no distinct hydrogeologic boundary between the two lithostratigraphic units as is commonly believed. As you walk up the outcrop, consider where you would define the bottom and the top of the “Jordan Aquifer.” The mixed carbonate and clastic component that composes the Coon Valley Member of the Oneota Dolomite (Prairie du Chien Group) is markedly lower in porosity and permeability than the coarse clastics of the uppermost Jordan Sandstone. The overlying carbonate rock component of the Hagar City Member is tightly cemented, with primarily horizontal fractures. Fracture tops show evidence

of dissolution. Larger scale vertical dissolution along a vertical joint is also visible here. If such fractures and dissolution features were not developed in the subsurface this unit would be a confining bed.

BIOSTRATIGRAPHY OF CONODONTS FROM JORDAN SANDSTONE AND LOWERMOST ONEOTA DOLOMITE NEAR HOMER, MINNESOTA

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The physical evidence for the presence of a large magnitude, regional unconformity associated with the Jordan Sandstone has been debated for decades. Smith and others (1993) described subaerially formed silcrete that occurred at or below the Jordan Sandstone-Oneota Dolomite contact at a few outcrops in Wisconsin. Runkel (1994) suggested that a pebbly sandstone that caps the Jordan Sandstone at many outcrops in Wisconsin and Minnesota is an erosional lag of regional extent. It everywhere separates a major, decameter-scale coarsening upward sequence below from a decameter-scale fining upward sequence above, and was formed when the regressive shoreline was in extreme basinward positions in latest Cambrian time. The Jordan Sandstone was shown to be more deeply truncated beneath this lag on the margins of the Hollandale Embayment (Runkel, 1994).

The work described above suggests that an unconformity is present near or at the top of the Jordan Sandstone but the physical evidence is subtle and cannot be recognized at all outcrops (Fig. 9). In an effort to substantiate the presence of an unconformity and determine its magnitude, conodont samples were collected at the Homer section from the Jordan Sandstone and from the overlying Coon Valley Member of the Oneota Dolomite by A. C. Runkel and were processed by J.F. Miller. Samples from the Jordan Sandstone contain only two conodont species, *Proconodontus muelleri* (Miller) and *Eoconodontus notchpeakensis* (Miller), which range throughout the Jordan. These species are widespread in North America and are diagnostic of the *Eoconodontus notchpeakensis* Subzone of the *Eoconodontus* Zone, approximately of middle Trempealeauan Age. This fauna has been recovered from samples 1 through 10. Sample 11 is from the Coon Valley Member of the Oneota Dolomite and yielded a completely different conodont fauna. This sample includes *Acanthodus uncinatus* (Furnish), *Aloxoconus iowensis* (Furnish), *Aloxoconus propinquus* (Furnish), *Aloxoconus* sp., *Cordylodus intermedius* (Furnish), *Cordylodus lindstromi* (Druce & Jones), a new species of *Cordylodus* known from Utah and Texas, and *Polycostatus oneotensis* (Furnish). Many of these species were named by Furnish (1938) from the Oneota in the Iowa-Minnesota area. The presence of these species in the Coon Valley Member indicates a close age relationship between sandstones of the Coon Valley and the overlying dolomite member of the Oneota, and supports the assignment of these sandstones to the Oneota rather than to the Jordan. This assemblage of species is at least as young as the *Cordylodus angulatus* Zone but could be from the overlying *Rossodus manitouensis* Zone. These two zones are from the middle to upper part of the Skullrockian Stage of the Ibexian Series (Lower Ordovician) in the Ibex area of western Utah (Ross and others, 1997).

These faunas indicate that the Jordan Sandstone and Coon Valley Member are separated by a large hiatus compared with the continuous succession in the Ibex area of Utah, which was

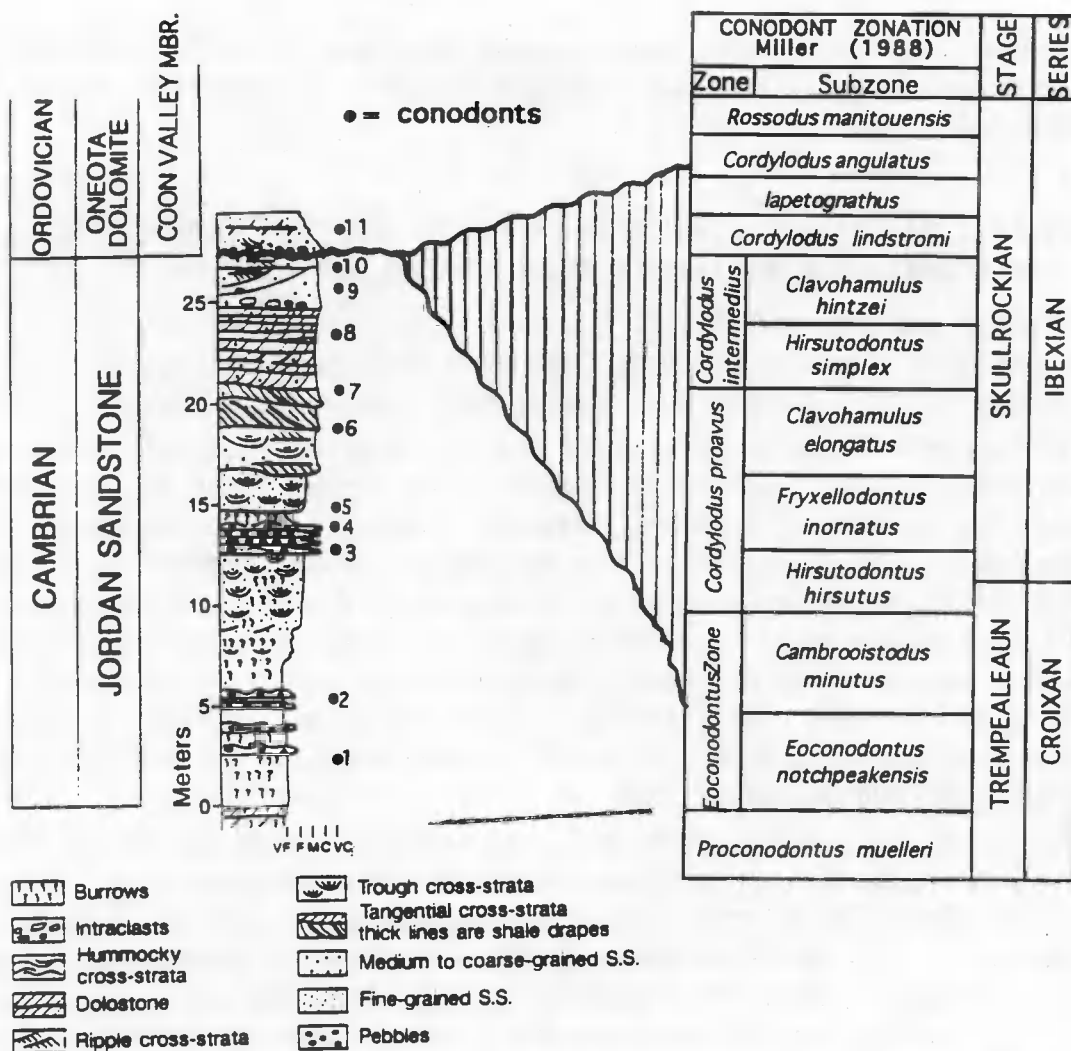


Figure 9. Measured section of the Jordan Sandstone and adjacent units at Homer, Minnesota outcrop, highlighting biostratigraphic interpretation based on conodonts as reported herein. The vertical lines show the relative magnitude of the hiatus represented by the unconformity on top of the Jordan Sandstone.

deposited on a rapidly subsiding miogeoclinal carbonate platform. The following conodont biozonal units known in Utah are missing between the Jordan and Coon Valley at Homer: *Cambroostodus minutus* Subzone of the *Eoconodontus* Zone (uppermost Trempealeuan, Upper Cambrian), three subzones of the *Cordylodus proavus* Zone (basal Skullrockian, Lower Ordovician), two subzones of the *Cordylodus intermedius* Zone, the *Cordylodus prolindstromi* Zone, the *Cordylodus lindstromi* Zone, and the *Iapetognathus* Zone (approximately middle Skullrockian). In the Ibex area of Utah, Upper Cambrian carbonate strata are approximately 2850 ft (870 m) thick and are overlain by approximately 5000 ft (1525 m) of Ordovician strata. The missing interval at Homer is equivalent to more than 500 ft (150 m) of limestone in Utah. The interval includes several stratigraphic sequences that include at least two major lowstands of sea level, as well as several transgressive and highstand sequences. A similar large hiatus is known on the crest of the Transcontinental arch in western Colorado (Myrow, Ethington, & Miller, 1995), which is a western extension of the Minnesota craton.

STOP 4—Road cut along north side of County Road 30 (time permitting). From County road 14, travel east on County road 8 for one mile (becomes County road 30 in Winona County). Continue east on County road 30 for about 0.7 miles to road cut (Fig. 10).

The upper part of Oneota Dolomite and lowermost Shakopee Formation of the Prairie du Chien Group are exposed at this stop. The Oneota Dolomite consists chiefly of tan to gray microcrystalline dolostone in thick to medium irregular beds. The Shakopee consists of sandy, oolitic and intraclastic dolostone, with intercalated sandstone. The New Richmond Sandstone of the Shakopee Formation forms the uppermost part of this exposure. Deposition of these strata occurred on a carbonate-dominated, laterally extensive platform in environments ranging from supratidal to shallow subtidal (Smith and others, 1993). The Oneota Dolomite and Shakopee Formation are separated by an unconformity less than 2 meters below the New Richmond Sandstone. Features diagnostic of depositional origin, as well as the unconformity, are largely inaccessible in this steep road cut. Stop 9 will provide a better opportunity to more closely examine the Prairie du Chien Group.

The primary purpose of this stop is to discuss the karst features exposed in this road cut. A well developed cave in the Oneota exists at the solution horizon near the Oneota-Shakopee contact. The unconformable contact shows subaerial exposure, as evidenced by considerable karst development below it, and is characteristic of this contact throughout southeastern Minnesota and western Wisconsin (Smith and others, 1993, 1996). Sinkholes occur near this contact (Dalgleish and Alexander, 1984; Alexander and Maki, 1988; Magdeline and Alexander, 1995) and have resulted in catastrophic collapse of several sewage treatment lagoons (Alexander and Book, 1984; Alexander and others, 1993; Jannik and others, 1991).

The cave at this outcrop apparently formed in a joint trending NE to SW. Evidence for the development of this solution feature is visible. Ground water originally flowed horizontally along a permeable bedding plane, widening itself through dissolution as it became the dominant hydraulic conduit at this horizon. At the intersection of a vertical joint, water moved downward, dissolving the bedrock and increasing the size of the cavity. This feature widened from top to bottom, as the conduit captured increasing amounts of ground-water flow. Characteristic features include a flat-topped ceiling—top of the original hydraulic fracture—and sinuous walls formed by the water dissolving beds of variable permeability. A large collapse adjacent to the cave, presumably into a void extending into the outcrop, and the alignment of this feature with a creek bed on south side of the highway, indicate the extensive nature of this type of dissolution within the Prairie du Chien.

The regional direction of ground-water flow with the Prairie du Chien Group is to the southeast—perpendicular to the trend of this large dissolution feature. Monitoring of ground-water quality within the Prairie du Chien is complicated by these types of features throughout southeastern Minnesota. Dye-trace studies have documented cases where local flow is in a direction perpendicular to the regional gradient (Donohue and Associates, 1991). The presence of these large-scale dissolution features leads to extreme variability in transmissivity, and preclude the treatment of the Prairie du Chien Group as isotropic porous media in ground-water monitoring and modeling.



Figure 10. Map 3 on index map (Figure 5) showing the locations of stops 4 through 8.

STOP 5—Scenic overlook of Mississippi River Valley near Wabasha. On north side of State Highway 60 about 1.3 miles east of U.S. 61 (Fig. 10).

This overlook provides a visual estimate of the thickness of the lower Jordan Sandstone, St. Lawrence, and Franconia Formations that lie below our feet. The upper 50 ft of Jordan Sandstone, and lower 20+ ft of Oneota Dolomite (Coon Valley Member) are exposed here. This outcrop provides an excellent opportunity to examine the sequence bounding unconformity that separates well-sorted, fine- to coarse-grained, quartzose Jordan Sandstone below from the overlying heterolithic Coon Valley Member. The Coon Valley Member consists of very fine- to coarse-grained sandstone, siltstone, shale, and sandy dolostone. The erosional lag marking the unconformity was poorly developed at Homer, Minnesota (Stop 3) but here consists of an intraclastic, poorly sorted, complexly cross-bedded sandstone—1 - 2 m thick—that contains abundant pebbles of Precambrian rocks as large as 2 cm in diameter.

Note the considerable difference in grain size and degree of cementation between the Jordan Sandstone and the Coon Valley Member. The former is dominated by the coarse clastic hydrostratigraphic component, the latter by the mixed carbonate and clastic component. Permeability of the mixed carbonate and clastic component is as much as three orders of magnitude less than that in the underlying coarse clastics. In unfractured subsurface settings, this unit and overlying unfractured carbonate rocks of the Hagar City Member of the Oneota Dolomite can act as a confining beds, hydraulically isolating ground water in the Jordan from ground water in the upper Prairie du Chien. This separation can be seen in chemical (Alexander, 1990; Setterholm and others, 1991; Smith and Nemitz, 1996; Wall and Regan, 1995) and hydraulic data (Barr Engineering, 1996).

STOP 6—Lower Franconia Formation near Wabasha, Minnesota. Outcrop is along west side of U.S. Highway 61, 1.2 miles north of State Highway 60 (Fig. 10).

The lower Franconia Formation at this stop is an excellent example of the heterolithic, very fine grained sandstone and shale facies characteristic of Cambrian storm and fairweather deposits that accumulated on the offshore shelf seaward of the quartzose shoreface sands. Thin to medium beds of feldspathic, variably glauconitic sandstone, siltstone, and shale commonly occur in normally graded sequences as much as 30 cm thick. Sandstone beds pinch and swell as form sets of small-scale hummocky cross-stratification, which in a complete depositional sequence, pass transitionally upward into ripple cross-stratified or structureless, very fine grained sandstone and siltstone, which in turn passes upward to shale. Bioturbation is common, and ranges from isolated burrows (*Planolites* and *Skolithos*) to nearly homogenized sediment.

The fine-clastic component of the lower Franconia Formation is hydrostratigraphically similar to the same component that makes up the bulk of the St. Lawrence Formation seen at Stop 3. The St. Lawrence is widely regarded as a confining bed, yet the Franconia is commonly designated as an aquifer along with the underlying coarse clastics of the Ironton and Galesville Sandstones. We believe that this characterization of the Franconia is inaccurate. This part of the Franconia has permeability values several orders of magnitude lower than underlying coarse sandstones (Miller and Delin, 1993). Core samples taken from outcrop had permeabilities of 0.01 - 0.006 md. (MGS unpublished data, 1997). In addition, wells finished above and below this part of the Franconia have over a 40-foot difference in hydraulic head (Wenck and Assoc., 1997).

STOP 7—Road cut on U.S. Highway 61 near Wabasha. From State Highway 60, take State Highway 61 south 1.8 miles to road cut on west side of road (Fig. 10).

The uppermost 5 meters of Ironton Sandstone and about 4 meters of the overlying Franconia Formation are exposed here in a steep road cut. The Ironton consists of poorly sorted sandstone that has a pervasive bioturbated texture, with burrows as much as a centimeter in diameter. Shale and siltstone occur as a matrix between sand grains and as lenticles. Moderately to well-sorted sandstone beds are internally structureless or contain ill-defined trough and tangential cross-strata with rare shale drapes and reactivation surfaces. These features are indicative of deposition in a shoreface and tidal flat environment. The overlying strata of the Franconia Formation consists mostly of feldspathic, glauconitic, fine- to very fine grained sandstone with hummocky cross-strata and tempestites that were deposited on the offshore shelf. These lower

Franconia Formation strata are better exposed at Stop 6. The contact between the Ironton Sandstone and Franconia Formation is transitional and records gradual flooding during an overall rise in sea level.

The quartzose sandstone of the Ironton exposed includes a pebbly sandstone bed that may mark a regional unconformity. The upper part of the underlying Eau Claire Formation and lower approximately two-thirds of the Ironton/Galesville Sandstones (below level of this exposure) form a coarsening-upward sequence that records deposition in progressively shallower water conditions. The upper one-third of the Ironton/Galesville Sandstones fines upward, recording deeper water conditions. The nearly structureless, resistant bed about 2 m above the base of this outcrop is a 1.5-2 m pebbly sandstone lag that separates these decameter-scale sequences. This pebbly lag can be physically traced from the Twin Cities area south into Iowa, and from southwestern Wisconsin westward at least 80 kilometers into the subsurface of south-central Minnesota. It may represent a transgressively re-worked lag that developed subaerially when the "Ironton-Galesville shoreline" was in extreme basinward positions.

Compare the hydrostratigraphic properties of the Franconia Formation and Ironton Sandstone. The Ironton is composed mostly of the coarse-clastic component, specifically of a poorly cemented mostly fine - very coarse grained, shaly sandstone with moderate porosity and permeability. The Franconia above is dominated by the fine clastic component, which is finer grained and better cemented than the Ironton, and contains discrete, continuous laminations of shale. It is markedly lower in porosity and permeability than the Ironton. The common practice of lumping these two units together as a single aquifer is not reasonable, and gives the erroneous impression that they share similar hydrogeologic attributes (see Stop 6 for additional discussion).

STOP 8—Road cut on U.S. Highway 61 near Wabasha. From State Highway 60, take U.S. 61 south 2.3 miles to road cut on west side of road (Fig. 10).

Interfingering Eau Claire Formation and Ironton-Galesville (Wonewoc) Sandstones are exposed in this long road cut. There are several interesting sedimentologic features, including hummocky cross-strata, tempestites, trough cross stratified sandstone, and coarse sand dune forms draped by shale. Trilobite molds are abundant at this stop in the hummocky stratified sandstone facies. This outcrop is one of only two known exposures in this area that record the boundary between the Marjumiid and Pterocephaliid biomes, a major isochronous extinction event correlatable across all of North America.

This exposure is a good example of the same fine-clastic hydrostratigraphic component that dominates the lower Franconia and St. Lawrence Formations seen at stops 3 and 6. Core samples from all three units show a considerably lower density of fractures along bedding planes.

STOP 9—Vern Keller Quarry. Along west side of County Road 6, 1.15 miles south of State Highway 19. Goodhue County (Fig. 11).

About 30 m of the middle part of the Prairie du Chien Group is exposed in this quarry. Relatively pure dolostone beds of the uppermost Oneota Dolomite are unconformably overlain by sandy dolostone and sandstone of the Shakopee Formation about 2 meters above the floor of the quarry. This unconformity is regional in extent, traced from this area into the Michigan basin by Smith and others (1993). Deposition of the Prairie du Chien Group occurred on a shallow carbonate platform during a relatively highstand of sea level that flooded most of North

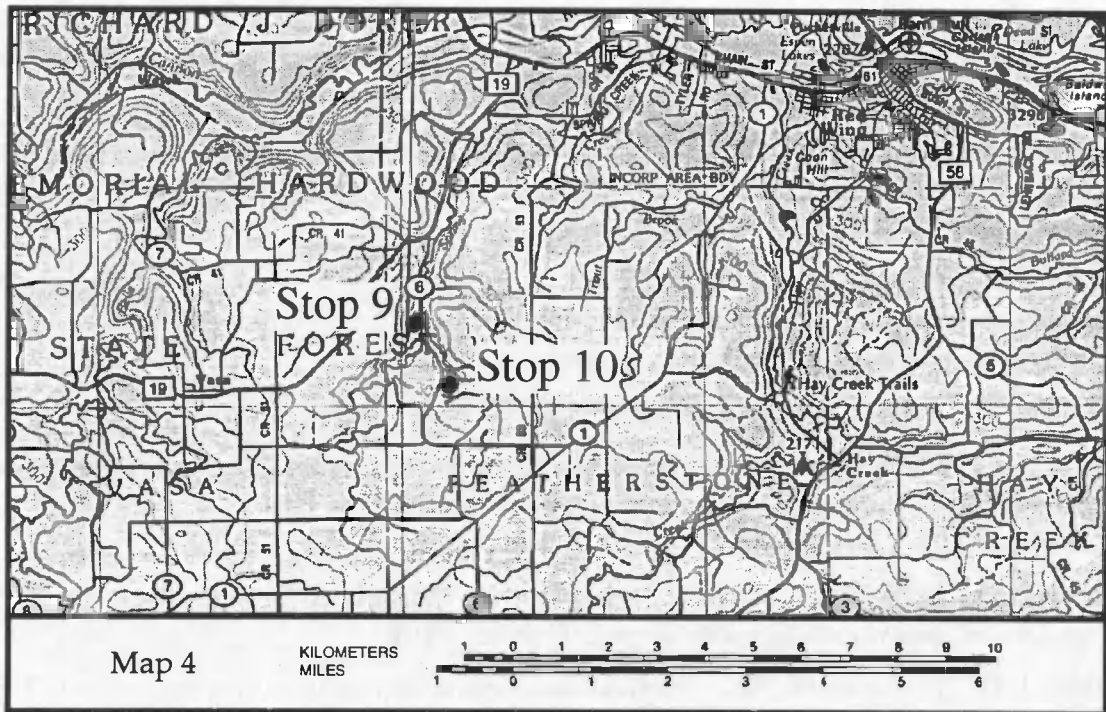


Figure 11. Map 4 on index map (Figure 5) showing the locations of stops 9 and 10.

America during the Early Ordovician (Smith and others, 1993; Smith and Clark, 1997). These strata are typical of “tropical” carbonate systems, dominated by oolitic and peloidal grainstones and packstones, and stromatolitic and wavy-laminated boundstones deposited in shallow subtidal to supratidal settings. Stromatolites are abundant in this quarry: digitate forms can be discerned in a few places in the uppermost Oneota, and cabbage-head stromatolites are common in the Shakopee. Evidence for subaerial exposure includes abundant mud cracks, intraclastic breccias, and silicified and moldic anhydrite.

There are some interesting questions that have yet to be answered about the deposition of the Prairie du Chien Group:

- 1) What is the fundamental control that blocked detrital input when most of the Oneota was deposited? Much of the Oneota consists of strata that were apparently deposited along a regional shoreline in which siliciclastics were nearly absent.
- 2) What are the fundamental controls that led to an arid, tropical carbonate style of deposition, when both younger and older carbonate deposition was reminiscent of modern cool-water, temperate conditions?

This part of the Prairie du Chien Group consists chiefly of the carbonate-rock component. It is an excellent example of the importance in delineating the presence and position of fractures that are the principal groundwater conduits. A prominent horizontal interval of dissolution features, some with stalactites, occurs along the Oneota-Shakopee unconformity in this quarry. A second dissolution layer is visible high on the quarry’s south wall, just below a 1-2 m bed of coarse sandstone. Work by Calvin Alexander from the University of Minnesota suggests that these dissolution horizons have an important control on the hydrogeologic attributes of the Prairie du Chien. They are likely major conduits for ground-water flow, and enhance the development of sinkholes where they are nearer the surface. Other notable features include the

tight, unfractured nature of the carbonate rock exposed on the south wall of the quarry—in contrast with the carbonate rock of the Shakopee seen at Stop 1 where it is closer to the surface, and secondary porosity is better developed.

Stop 10—Spring near Vern Keller Quarry. From quarry continue south 0.7 miles on County Road 6. Turn east on to gravel road and travel 0.1 miles (Fig. 11). Spring is on north side of road on PRIVATE PROPERTY.

Elevation of this resurgent spring approximately the same as the Oneota-Shakopee contact seen at the previous stop. This supports the hypothesis that dissolution features along the unconformity may be a major conduit for groundwater flow.

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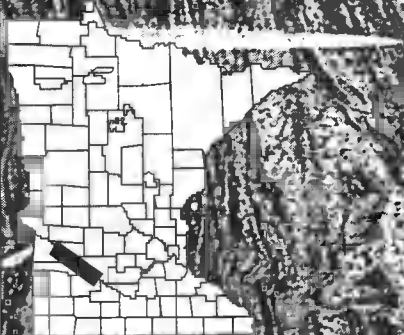
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FIELD TRIP 5

Archean and Quaternary Geology of the Minnesota River Valley



Loaders

L. Southwick
Camp Patterson

Minnesota Geological Survey

FIELD TRIP #5

MINNESOTA RIVER VALLEY AND VICINITY, SOUTHWESTERN MINNESOTA

Leaders: D.L. Southwick and Carrie J. Patterson

INTRODUCTION TO THE GEOLOGY OF PRECAMBRIAN ROCKS IN THE MINNESOTA RIVER VALLEY

D. L. Southwick

The Precambrian rocks exposed in the Minnesota River Valley are within an Archean gneiss terrane that constitutes the southernmost subprovince of the Superior Province of the Canadian Shield. The Minnesota River Valley (MRV) subprovince is further subdivided into four tectonic blocks by regional ENE-trending shear zones (Fig. 1; Southwick and Chandler, 1996). This trip will focus on exposed rocks of the Morton and Montevideo blocks, which are

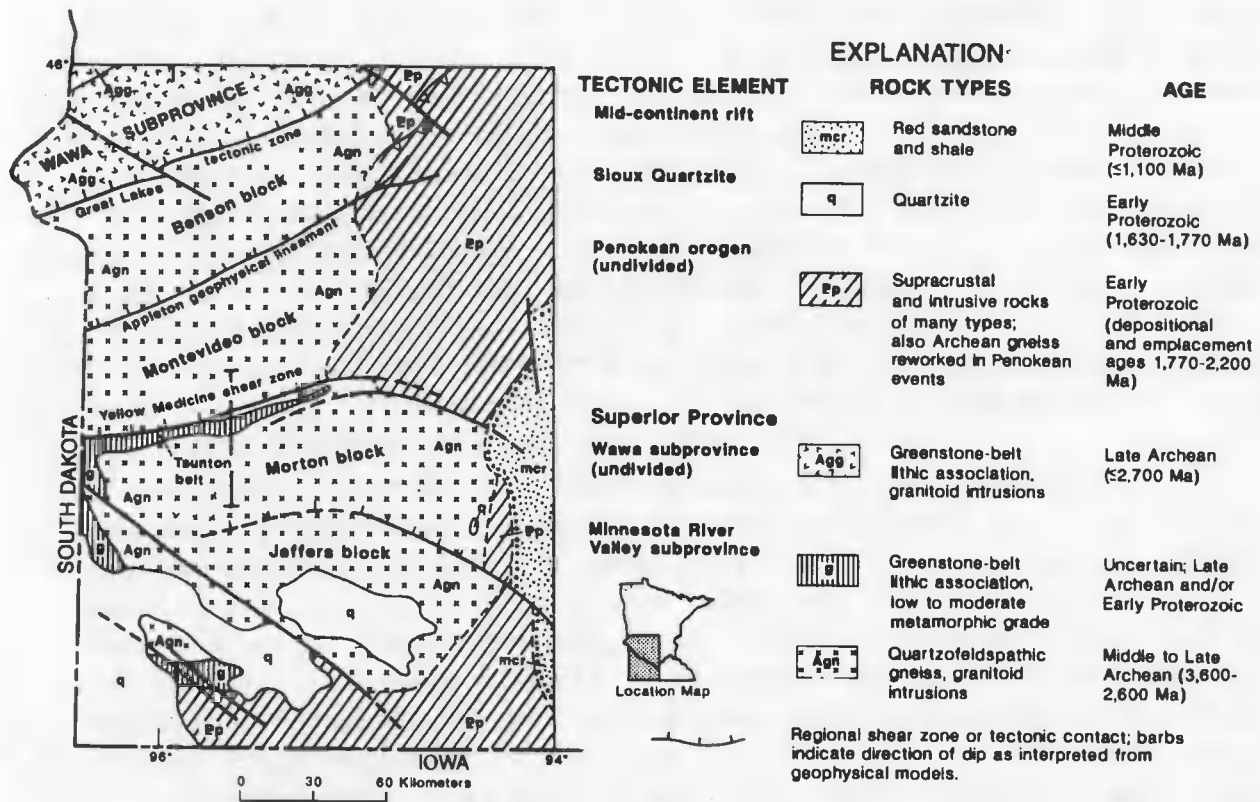


Figure 1. Tectonic sketch map of Precambrian terranes in southwestern Minnesota. The Minnesota River Valley subprovince consists of the Benson, Montevideo, Morton, and Jeffers blocks. The Taunton belt is a narrow septum of relatively low grade metavolcanic rocks along the Yellow Medicine shear zone, which marks the boundary between the Montevideo and Morton blocks. The line indicates the location of geophysical profiles. From Southwick and Chandler (1996).

the central two blocks of the subprovince. The boundary between these is the Yellow Medicine shear zone (YMSZ), a major crustal feature that is well delineated aeromagnetically but does not crop out. The YMSZ crosses the Minnesota River between trip stops 6 and 7 (Figs. 1 and 2). At stops 1 through 6, south of the YMSZ, we will see features of the Morton block and at Stops 6 through 11 we will see features of the Montevideo block.

The characteristic rock type of both blocks is a modally layered gneiss that falls generally within the tonalite to granodiorite range in bulk composition (Goldich and others, 1980a and b). This rock type, together with less abundant layers, lenses, and irregular blocks of mafic gneiss, amphibolite, and various types of aluminous gneiss, constitutes that Morton Gneiss south of the YMSZ and the Montevideo Gneiss north of it. Alert readers will note the correspondence between the block name and the name of the principal map unit within it. Mineral-assemblage data indicate that the Morton and Montevideo gneisses were metamorphosed under conditions of the upper amphibolite and granulitic facies (Grant, 1973; Perkins and Chipera, 1985). Retrograde processes have affected a considerable area of the Montevideo Gneiss but are evident only locally in the Morton.

The Morton and Montevideo gneisses were invaded by granite (adamellite, in the terminology favored by Goldich and others, 1980a and b) at two different times during the Archean. The first episode of granite emplacement (adamellite-1 of Goldich and others, 1980a) preceded the development of the predominant metamorphic fabric in the gneissic host rocks; that fabric passes through the granite bodies, most of which are aplitic or pegmatitic, without discernable change in orientation. In addition, much granite of this generation possesses metamorphic texture. The second episode of granite emplacement (adamellite-2 of Goldich and others, 1980a) occurred after the main fabric-forming events. Most intrusions of this generation lack internal fabric, although some show a weak foliation and therefore may be late- to post-tectonic. Good igneous textures are the norm. Adamellite-2 intrusions range from aplitic to medium-grained granitic in grain size and from thin dikes and sills to batholiths in form and scale. The Sacred Heart Granite (Stop 6) forms a regional-scale, sheet-like intrusion several kilometers in thickness that is interpreted to be of the adamellite-2 generation. The Ortonville and Fort Ridgeley granites are late- to post-tectonic masses that also are of the adamellite-2 type and are assumed to be approximately coeval with the Sacred Heart.

Between 1950 and 1980, S.S. Goldich and colleagues devoted huge efforts to pioneering geochronologic studies of the Morton and Montevideo gneisses and the various granitic rocks that are associated with them (Goldich and Wooden, 1980; Goldich and others, 1980a and references therein). The gneisses proved to be (1) very old, with components of both the Morton and Montevideo as old as about 3,600 Ma, and (2) geochronologically complex, with several discernable Archean events recorded isotopically. Moreover, the Goldich group found the "adamellite 1" rocks to be isotopically complex as well, and even the geologically straightforward-appearing "adamellite 2" rocks to yield less than definitive crystallization ages. Goldich's interpretation of the geologic history of the MRV subprovince is presented in Table 1.

Goldich emphatically believed that the granite of the adamellite-1 type represented genuine igneous importation of magma from a distant source; i.e., true intrusion. Others have contended that some, if not all, of the early granite was derived more or less locally through partial melting of the gneissic host. A major question persists as to whether the pink leucosomatic components of the Morton Gneiss in the vicinity of Morton crystallized from injected melt of adamellite-1

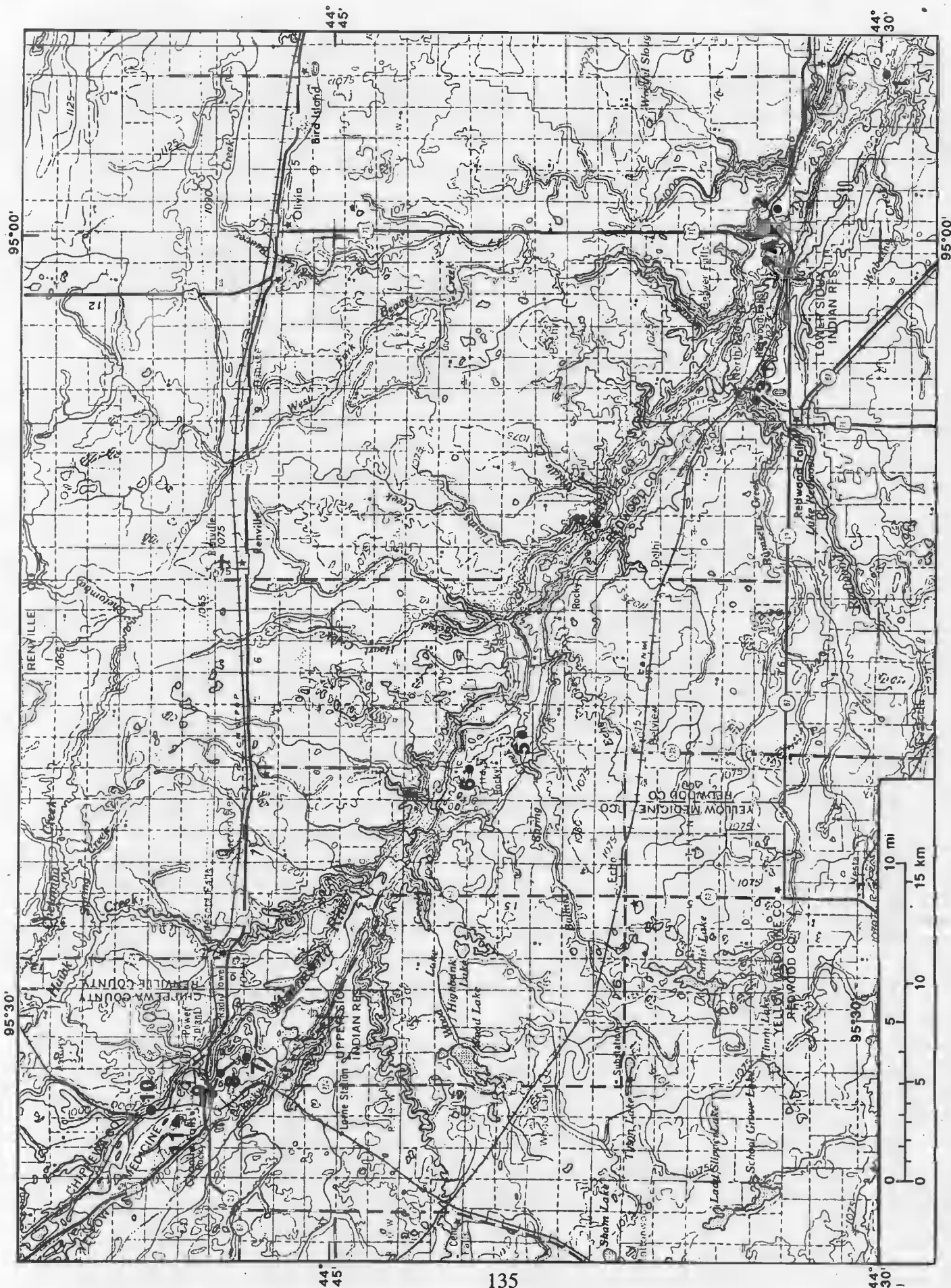


Figure 2. Locations of field-trip stops for Day 1. More detailed location maps accompany the individual stop descriptions.

Table 1. Summary of principal Precambrian events in the Minnesota River Valley subprovince. Modified and interpreted from Goldich and Wooden (1980); Goldich and others (1980a).

1800 Ma event (Proterozoic): Emplacement of small intrusions; local retrograde metamorphism, shearing
2600 Ma event Emplacement of post-kinematic aplite and pegmatite dikes Emplacement of Sacred Heart Granite, other granite batholiths, and associated minor intrusions ("adamellite-2"; late-kinematic) High-grade metamorphism, deformation of gneiss
3100 Ma event High-grade metamorphism, deformation of gneiss Emplacement of minor granitoid intrusions ("adamellite-1")
3600 Ma event Intrusion of tonalite, granodiorite into heterolithic precursor of layered gneiss

type or from melts developed through anatexis of pre-existing granodioritic to tonalitic gneiss. We will have an opportunity to form our own conclusions at stops 2 and 2A.

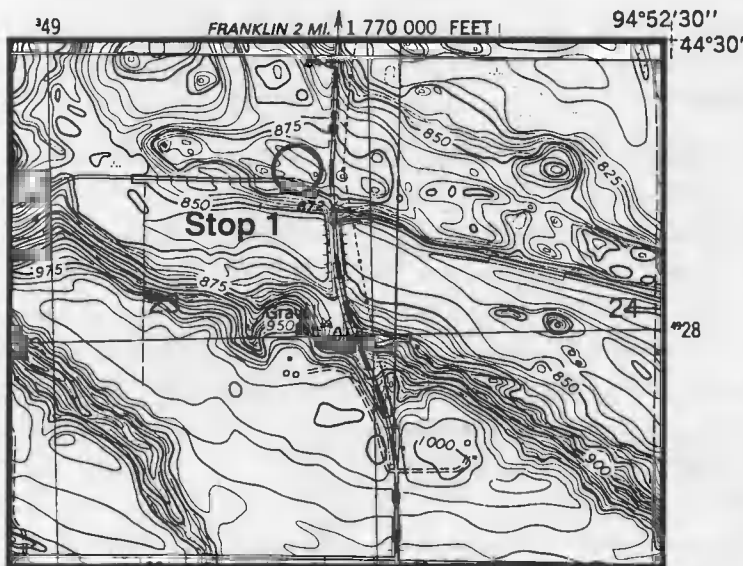
Although the principal rock-forming events responsible for the formation of the MRV subprovince occurred over a billion-year interval of Archean time, there are clear indications of further geologic activity in the Early Proterozoic. The subprovince lay in the cratonic foreland (Superior craton) with respect to Penokean tectonism in the interval between 2.1 and 1.7 Ga. We are just beginning to recognize and quantify events and responses recorded in the foreland rocks that were engendered by tectonic processes within the Penokean orogen (Southwick and Chandler, 1996; Holm and others, 1998), the axis of which lies only a few 10's of kilometers to the east of present MRV exposures. Proterozoic events include (1) the emplacement of at least two and possibly as many as four swarms of diabase dikes; (2) the emplacement of small intrusive plugs that range in composition from peridotite to granite; and (3) differential vertical movement on block-bounding shear zones, with concomitant development of brittle structures and retrograde metamorphic effects.

No discussion of the MRV subprovince, regardless of brevity, is complete without an acknowledgment of the excellent geologic mapping done by Ernest Lund (1950, 1956) over 45 years ago. Lund's work was the starting point for subsequent generations of field investigations (Himmelberg, 1968; Grant, 1972; Bauer, 1980) and was the principal framework for the geochronologic investigations of Goldich and colleagues. Those of us working today find Lund's maps to be proof of the mantra that careful geologic observations and descriptions are timeless; their value will outlast the interpretational framework within which they were made.

FIELD TRIP STOPS—DAY 1

Stop 1—MRV geomorphology; Proterozoic diabase dikes and wallrock of Archean Morton Gneiss. Morgan quadrangle, T. 112 N.-R. 34 W.-Section 23AAC.

At this locality south of Franklin, newcomers can readily comprehend several elements of post-Precambrian geological history that bear on the condition and distribution of the Precambrian rock outcrops on the valley floor. Point (1): The Precambrian rocks were deeply weathered prior to continental glaciation in the Quaternary. Actually, most of the weathering occurred prior to the Late Cretaceous. Many MRV outcrops, such as these, are incipiently to substantially weathered; they represent material that lay just at the base of the pre-Cretaceous weathering



profile. Upper parts of the weathering profile, remnants of Late Cretaceous marine strata, and the overlying glacial deposits were stripped away as the Glacial River Warren (a high-discharge, low-gradient meltwater river ancestral to the Minnesota) eroded downward. River downcutting slowed or perhaps ceased where hard rock was encountered at the base of the weathered section. Point (2): The present valley-floor outcrops were rapids and islands in the Glacial River Warren. Their morphology is due primarily to river erosion.

Scoured cuts (former shoots in rapids) and potholes are common. Point (3): Because of the history discussed in points (1) and (2), outcrops tend to be clusters of low knobs of approximately the same height. In general they do not occur on the valley walls, Grand-Canyon style, but instead are located toward the middle of the valley floor where the stream velocity and downcutting power of the River Warren were greatest. The sides of the valley floor are commonly occupied by constructional terraces (as here) and the valley walls are chiefly the Quaternary sediments through which the River Warren cut.

Diabase dikes of Proterozoic age form prominent valley-floor ridges at this locality. Compositionally, the dikes are good continental tholeiites. They are very similar in composition and petrographic characteristics to tholeiitic dikes near Granite Falls, some 40 miles up-river, that yield K-Ar dates of about 2,080 Ma (Hanson and Himmelberg, 1967). Several of the larger tholeiite dikes at Granite Falls strike about az. 095-100, approximately parallel to the dikes here at Franklin. Some better geochronology and a little paleomagnetic work would be very helpful in deciding whether the Franklin and Granite Falls tholeiitic diabase dikes are or are not genetically related.

About a mile northwest of Stop 1 is a prominent brushy knob that is held up by the Cedar Mountain Complex, a compositionally zoned igneous plug of Proterozoic age. The outer part of the plug is melanocratic hornblende diorite that has a chilled external rind of olivine trachybasalt porphyry. The core of the plug is granite and granodiorite. Biotite and hornblende from the core yield essentially identical K-Ar ages of 1,750 Ma (Hanson, 1968).

About 1.5 miles to the east of Stop 1 is a small plug of highly silicified and carbonatized serpentinite. This body is post-Archean but is unconstrained otherwise as to age.

The wallrock to the dikes and other intrusions is a granodioritic variant of the middle Archean Morton Gneiss. Because the gneiss here is in such poor condition (incipient weathering), it has not been studied in any great detail. We will see much better outcrops of the Morton at Stops 2, 4, and 5.

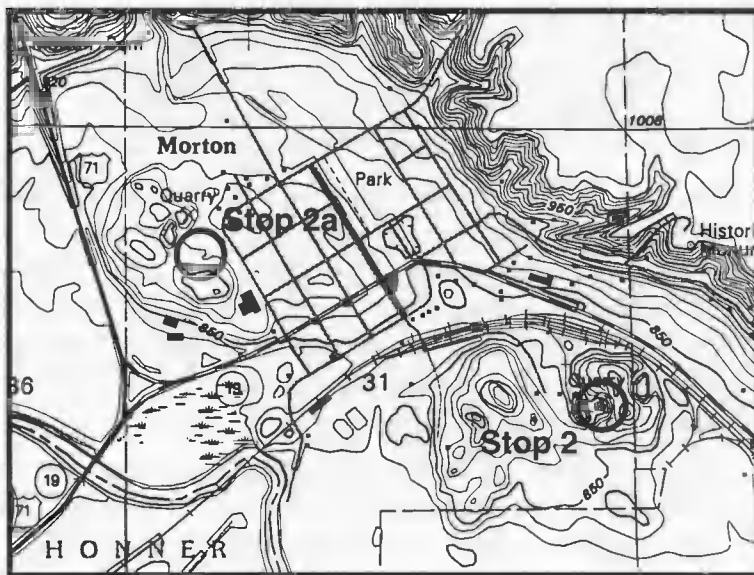
Features to see:

- Grain size and fabric variations within diabase dikes; chilled margins.
- Paleomagnetic drill holes: Where are the paleomagnetic data??

Stop 2—Morton Gneiss, Morton Quarry owned and operated by the Cold Spring Granite Company. Morton quadrangle, T. 113 N.-R. 34 W.-section 31DAA. Hard hats must be worn everywhere on company property.

The flamboyant dimension stone quarried here is a variant of the Morton Gneiss. It is a classic migmatite in which a pink, granitic to pegmatitic component of the neosome (the “adamellite-1” rock suite of Goldich and others, 1980b) is unusually abundant. More typical Morton Gneiss from the standpoint of areal abundance is a predominantly gray rock of granodioritic to tonalitic composition that is more uniform in texture than the gneiss in the Morton area. This is another case where the concept of “type locality” for a complex metamorphic rock turns out to be flawed. The Morton area supplied the name for the gneiss because the many quarries provided accessible, unweathered, clean exposures to study, not because the rock exposed is particularly typical of a mappable geologic unit.

The gneiss from this quarry is marketed under the trade name “rainbow granite.” It can be seen as a facing and ornamental stone on buildings throughout the Nation, and in recent years it has become a favorite material for up-scale kitchen counters and tabletops. The patterns, suggestive of plastic, inhomogeneous bulk deformation under conditions at which the rock was in a partially molten state, are as attractive and intriguing to architects and interior designers as they are to geologists.



Features to see:

- Paleosomatic “enclaves” of several types.
- Evidence of rheological contrasts among different rock types during partial melting and deformation.
- Gently dipping, broadly undulating gneissosity; small, tight folds reoriented into the undulating “regional” fabric.

- Zones of extensional ductile shear, many of which contain granitoid segregations interpreted to have formed as partial melts.

Stop 2A (Alternative to Stop 2)—Morton Gneiss, small quarry behind former school building in town of Morton. Morton quadrangle, T. 113 N.-R. 34 W.-section 31BCBA.

Because of insurance liability considerations, the Cold Spring Granite Company does not normally permit field-trip groups into the active Rainbow quarry. As of press time we were not cleared for quarry access. The small quarry behind the old Morton school shows many of the features of the gneiss described under Stop 2 (above), but the rock surfaces are smaller and less spectacular than those in the Rainbow quarry. Please read the discussion of Stop 2 and transfer it to the Stop 2A exposures.

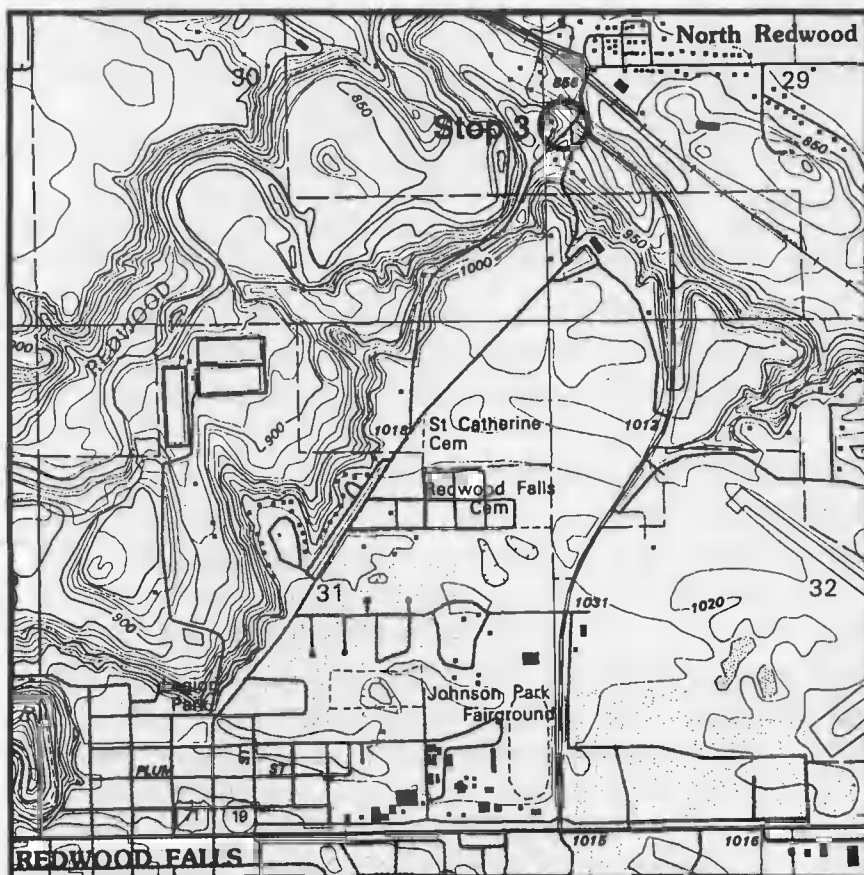
Be very careful as you move about Stop 2A. Rock surfaces that are shaded most of the time acquire a slimy coating that can be treacherously slick.

Stop 3—Weathering profile developed on Morton Gneiss prior to the Late Cretaceous. Redwood Falls quadrangle, T. 113 N.-R. 35 W.-section 30DAB.

Clay-rich saprolith of the general type exposed here occurs widely in the subsurface of southwestern Minnesota. Where Late Cretaceous marine strata have not been removed by post-Cretaceous erosion they overlie the saprolith directly, confirming that most of the weathering occurred before Late Cretaceous marine transgression and the deposition of marine sedimentary

rocks. In places where the Cretaceous rocks were eroded away prior to glaciation, remnants of the weathering profile directly underlie Quaternary glacial deposits. The clay, grus, and corestones from the weathered Precambrian rocks undoubtedly contributed substantially to the mass of subsequent Cretaceous shales and Quaternary tills.

Reworked and in-situ clay deposits of Late Cretaceous and post-Cretaceous age are mined at several sites along the Minnesota River, including one about a mile southeast of



these exposures. The clay is used mainly in the manufacture of portland cement and for various agricultural purposes.

The weathering profile at this stop was studied by Sam Goldich in the 1930's; his classic paper on the weathering stability of rock-forming minerals (Goldich, 1938) was based on materials collected here. That fundamental paper on weathering has probably achieved wider notoriety than all of Sam's geochronological contributions combined.

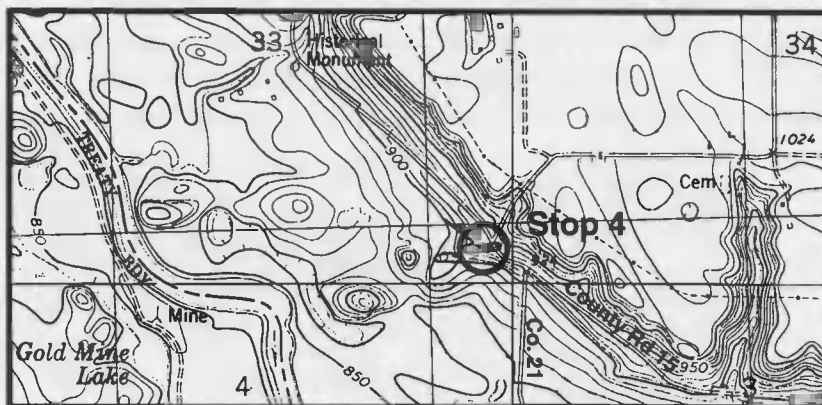
Features to see:

- Upward loss of source-rock minerals and fabric in the bottom two-thirds of the weathering profile.
- Pisolitic clay layer toward the top of the section.
- Bedded clay (reworked by streams?) above the pisolitic layer and below Quaternary till.

Stop 4—Morton Gneiss cut by adamellite dikes.

Vicksburg quadrangle, T. 114 N.-R. 36 W.-section 33DDD.

The gray tonalitic gneiss exposed in these outcrops is a common, widespread variant of the Morton. Much of the rock mapped as Morton Gneiss is basically this lithology with varying quantities of included mafic "enclaves" and varying degrees of intrusion by younger granitoid rock types. The gray gneiss here is about as inclusion-free as it gets. Furthermore, this is a locality where there is a clear-cut intrusive relationship between late granite of the Sacred Heart type (sharp-walled dikes, including an easily accessible one about two meters thick) and the tonalitic Morton Gneiss. The relative age of the two is not a matter of great debate, in that it can be reliably inferred from textural evidence in the rocks and from field relationships observed at a number of places in the Valley. Accessible outcrops that show the critical field relations are rather rare, however.



Samples of tonalitic Morton Gneiss from this site form part of the dataset for the "old" Rb-Sr isochron age of about 3,500 Ma (Goldich and Wooden, 1980) that is widely cited in the literature. Data from here anchor the lower or less radiogenic end of the Rb-Sr diagram (Goldich and Wooden, 1980; their fig. 2 and data from locality 5). In addition, zircons from the gneiss at this locality yield an interpreted U-Pb age of 3,662 $\pm$ 42 Ma (Goldich and Fischer, 1986).

Lengthy discussion of geochronologic details is beyond the scope and purpose of this field guide. It must be pointed out, however, that healthy disagreements exist as to the exact antiquity of the Morton Gneiss and the significance to be attached to the dates determined. A major source of controversy derives from the complexity of the neosomatic components of migmatitic rocks (as seen in the quarries at Stops 2 and 2A, for example) and conflicting interpretations of the processes and sequence of processes involved in neosome genesis. The

more straightforward sequence of cross-cutting relationships displayed at the present stop is informative because it eliminates some of the interpretational ambiguities associated with the more complex phases of the Morton.

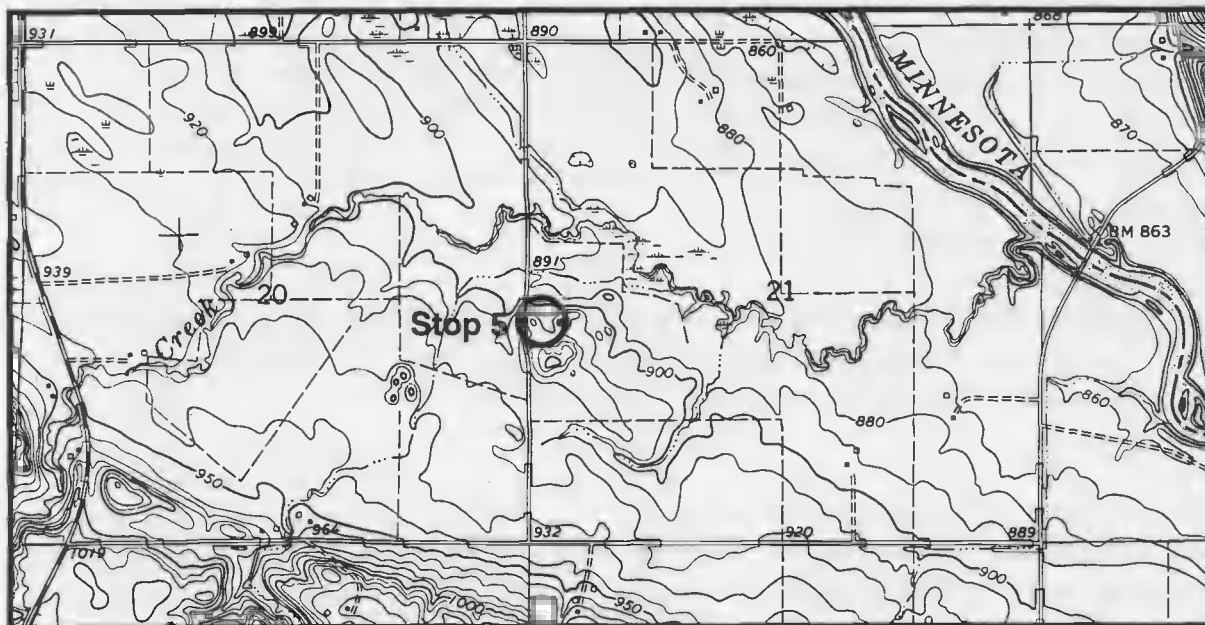
Features to see:

- Fabric and fabric variations in tonalite gneiss.
- Folded mafic layer interpreted as a deformed diabase dike. (Do you agree?)
- Textural and fabric variations within post- Morton adamellite intrusions.

Stop 5—Morton Gneiss that displays a variety of enclaves and small-scale structures.

Iverson Lake quadrangle, T. 114 N.-R. 37 W.- section 21CB.

These outcrops are located about 0.5 mile south of the the largest exposed mass of Sacred Heart Granite in the Minnesota River Valley. The gneiss here is characterized by abundant amphibolitic inclusions of tholeiitic and komatiitic composition, complex fold structures



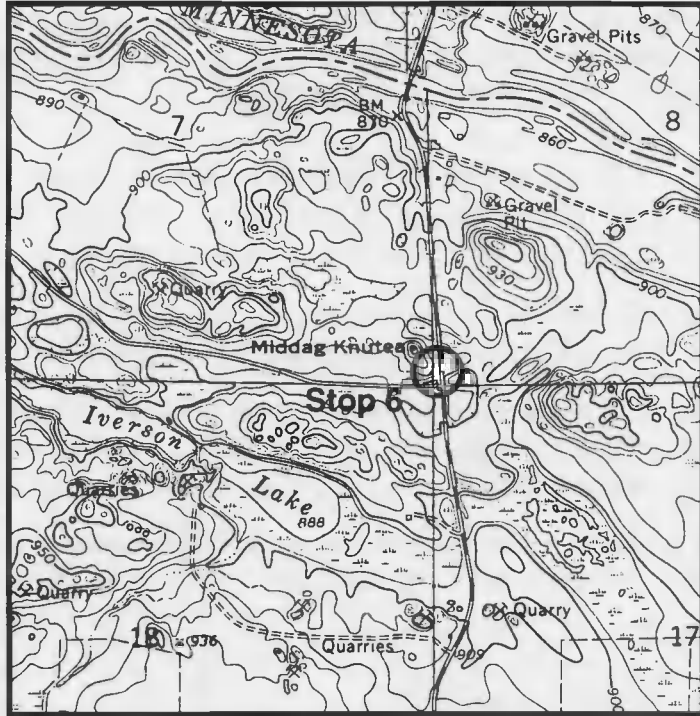
indicative of multiple folding events, deformed (boudinaged, folded) aplitic dikes presumably of the “adamellite 1” suite, and straight, undeformed granite dikes of Sacred Heart Granite. The quartzofeldspathic gneiss that encloses the various mafic rock types is variable in modal composition and texture. Its overall composition is approximately in the granodiorite range.

Wander around among the cedars and thornbushes to find examples of the features mentioned above. Look before you kneel or sit! Little cacti with big spikes grow all over these knobs.

Stop 6—Sacred Heart Granite and associated rock types. Iverson Lake quadrangle, T. 114 N.-R. 37 W.-section 7DD, near prominent knob named Middag Knutea.

The Sacred Heart Granite is one of several large granite masses that intruded MRV gneisses in the 2.55 - 2.7 Ga time interval. Available radiometric age determinations on these intrusions are rather imprecise, owing to the scarcity and poor quality of the contained zircons. Further efforts to obtain high-precision dates would be most welcome.

Most of the Sacred Heart is a rather ordinary medium-grained granite that consists of subequal amounts of quartz, oligoclase, and K-feldspar. It was quarried as dimension stone in an earlier era but is not being worked now. Although much of the rock mapped as Sacred Heart is quite uniform in texture and composition, there are parts of the mass that display strongly gneissose fabric and quartz-poor compositions trending toward monzonite. Some features indicative of a complex intrusive history are displayed in the road cuts at this stop.



Features to see:

- Massive, uniform Sacred Heart Granite (road cuts about 1,000 ft south of Middag Knutea).
- Gneissose fabric in granite; modal and textural layering (SE flank of Middag Knutea).
- Monzonite with pyroxene granulite inclusions.

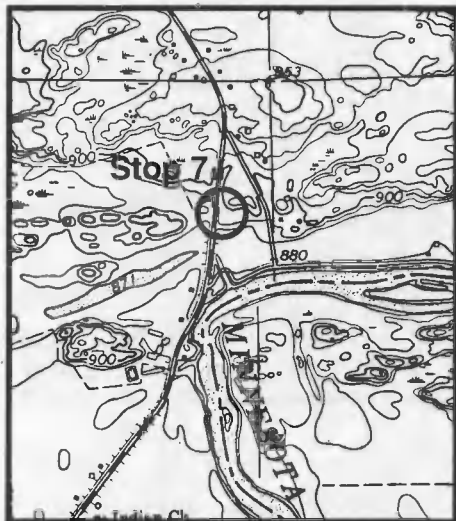
Drive to Stop 7—The bedrock surface exposed here was exhumed by Glacial River Warren, however, significant erosion of fresh rock was not required to form this smoothly undulating bedrock surface with rounded bedrock knobs and boulders. Such surfaces can be formed simply by the removal of deeply weathered bedrock. Large rounded boulders were not shaped during fluvial and glacial transport; they represent chemically rounded corestones resting at or near their source.

Pre-Cretaceous weathering penetrated to an average depth of 30 m, but the fresh rock/ weathered rock interface is irregular and undulates with as much as 45 m of relief. Joints and fractures are preferentially exploited and penetrated to depths of 60 m in places, locally isolating meter-sized volumes of fresh rock. The rounded, fresh rock remnants are known as corestones and are a common feature of the weathered bedrock residuum in Minnesota and when incorporated into glacial sediment, may be misinterpreted.

On the south side of the road is Swedes Forest, Homme-Kollin Unit, Scientific and Natural Area. Its main scientific significance is that it is the site of several old quarries that are hibernating areas for the rare, five-lined skink (also called blue-tailed lizards and “blue devils”). They prefer sunny bedrock exposures near oak woodlands. The main range of this species is the eastern United States and these are an isolated population.

Stop 7—Garnet-biotite-quartz-feldspar gneiss (Montevideo Gneiss). Granite Falls quadrangle, T. 115 N.-R. 39 W.-section 10AA.

The Yellow Medicine shear zone (YMSZ), a major crustal structure within the Minnesota River Valley subprovince (Southwick and Chandler, 1996), crosses the Minnesota River about two miles southeast of here. The shear zone rocks do not crop out; the zone was first recognized from geophysical mapping and the existence of sheared rock within it was later verified through core drilling. The YMSZ is now informally defined as the boundary between the Morton Gneiss (to the south) and the Montevideo Gneiss (to the north).



The rock in the rubbly road cuts at this locality is a garnet-biotite-quartz-plagioclase gneiss, a mappable sub-unit of the Montevideo Gneiss that is fairly abundant in the Granite Falls area (Himmelberg, 1968). Lund (1950) mapped it as the Granite Falls garnetiferous quartz diorite gneiss, a name that never caught on. The protolith of the unit as a whole is not immediately apparent. Parts of it are quite aluminous, suggestive of a sedimentary origin,

and other parts are very feldspathic, at least locally suggestive of igneous derivation.

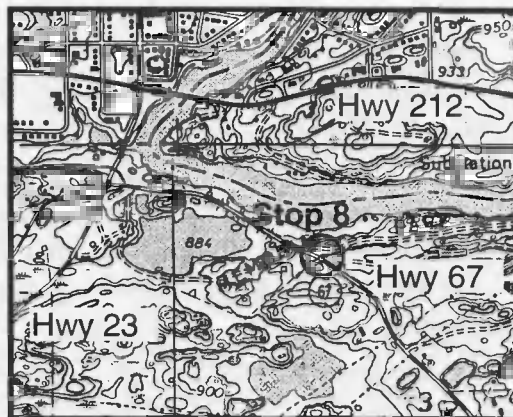
This and other rock units in the Granite Falls area are folded broadly about east-trending, gently east-plunging axes (Himmelberg, 1968) that Bauer (1980) interpreted as F2 structures. The complete deformational history involves four folding events, which Bauer (1980) discusses in detail. Faults and brittle-ductile shears at and near the south edge of the garnet gneiss outcrop belt are probably related to late deformation along the nearby YMSZ. There also is scattered evidence of earlier, more deep-seated shear deformation that may have been related to earlier YMSZ activity.

Features to see:

- Lithologic variations in the gneiss.
- Late shear zones.
- Rotational structures indicative of ductile shear.

Stop 8—Granulite-facies mafic and pelitic gneiss (Montevideo Gneiss). Granite Falls quadrangle, T. 115 N.-R. 39 W.-section 3BBD.

Garnet-bearing gneiss similar to that at Stop 7 is in contact with granulite-grade mafic gneiss (cpx-opx-plag) in these road cuts. Small amounts of biotite in the granulite may record retrograde re-equilibration. The garnet gneiss is quite micaceous and well layered, and has been regarded as a metamorphosed sedimentary rock. The mafic gneiss may have been a basalt flow originally, although other mafic protoliths are certainly possible.



Stop 9—Mafic gneiss (Montevideo Gneiss). Shear zones and pseudotachylite veins. Granite Falls quadrangle, T. 115 N.-R. 39 W.-section 4AAB.

Shear structures in these outcrops have been studied by John Craddock and several of his students at Macalester College. The following summary is taken from material provided by John.

Twenty-two discrete shear zones have been found here that contain pseudotachylite seams. The zones are subparallel, trend about az. 070 on average, and dip subvertically. This orientation is approximately that of the regional-scale Yellow Medicine shear zone and several of the Proterozoic diabase dikes in the area. Within a given zone the individual pseudotachylite seams (black) are of diverse orientations.

The pseudotachylite matrix is compositionally layered and preserves flow-foliation swirls and relics of glass. Dextral winged porphyroclasts in the pseudotachylite preserve dextral and sinistral kinematic displacement, primarily opposite the sinistral sense of S-C structures in the country rock and of sinistral fault drag indicators along the strike-slip zones that contain the pseudotachylite seams. Seam-margin forking directions are dominantly toward the west; porphyroclast imbrication within the pseudotachylite is dominantly up and easterly.

Calcite in veins that appear to be contemporaneous with the pseudotachylite seams yield twinning strains indicative of zone-parallel and zone-normal horizontal shortening. Magnetic fabric measurements on the pseudotachylite seams give K_{max} subvertical, K_{min} subhorizontal and zone-normal. If the magnetic fabric correlates with flow, these results indicate that the flow of pseudotachylite glass was upward within the zone of shear.

Features to see:

- Discrete shear zones (many are reddened).
- Pseudotachylite seams.
- Sense-of-shear indicators in shear zones.

Stop 10—Proterozoic "Section 28" intrusion and wallrocks of Montevideo Gneiss. Granite Falls quadrangle, T. 116 N.-R. 39 W.-section 28BCA.

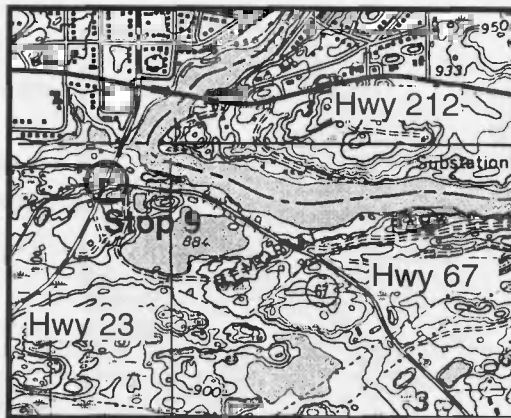
At this locality a small plug of Proterozoic granite is in complex contact with a body of hornblende andesite (also Proterozoic). Contacts with the Montevideo gneiss are not exposed, but they can be mapped rather closely. The granite yields a zircon Pb-Pb age of 1825 Ma (Catanzaro, 1963).

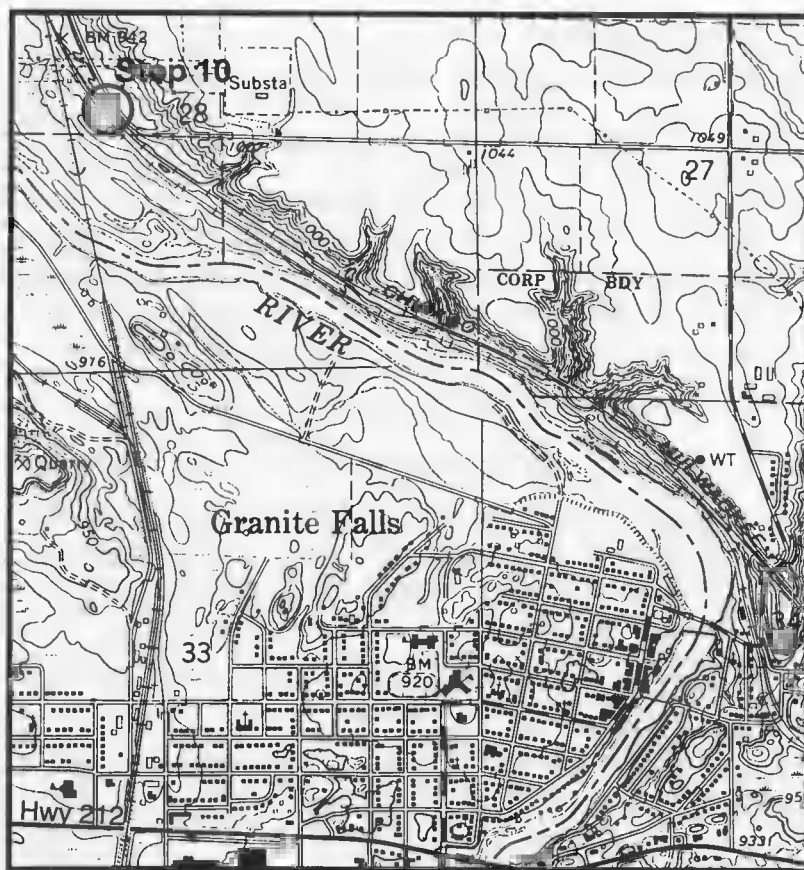
The standard explanation of the relationships seen here is as follows:

(1) The body of dark rock is a dike of hornblende andesite. Dikes of this composition occur elsewhere in the Granite Falls area (e.g., Stop 11).

(2) The granite intrudes the dike and thus is the younger rock, by classical geological reasoning.

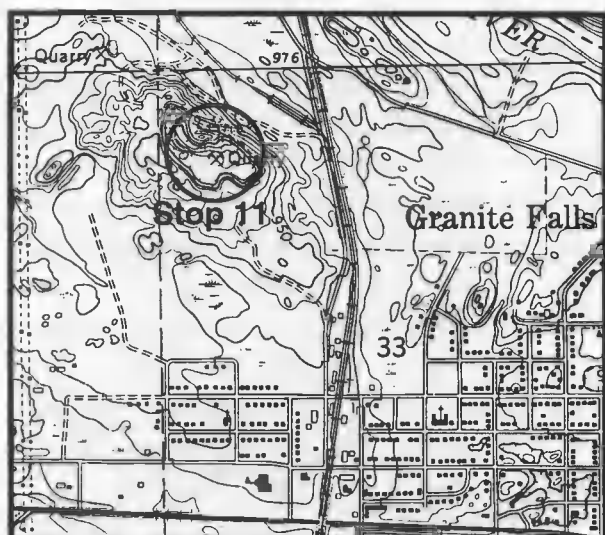
Do we agree that the observed contact relations necessarily imply the interpretation outlined above?





Stop 11—Montvideo Gneiss, Proterozoic dikes in the Yellow Medicine quarry owned and operated by the Meridian Aggregates Company. Granite Falls quadrangle, T. 115 N.-R. 39 W.-section 33BB. Hard hats must be worn everywhere on company property.

The Yellow Medicine quarry produces crushed stone, much of which is used as railroad ballast for the Burlington Northern Railroad. Aggregate from here is used on railroad tracks across the Dakotas and Montana to the Rocky Mountain front.



The principal rock in the quarry is a granitic gneiss composed of quartz, plagioclase, microcline, and biotite that is modally layered on various scales. Scattered within it are conformable lenses and layers of hornblende-pyroxene gneiss. This general rock type is the commonest variant of the Montevideo Gneiss in terms of map area and is essentially the rock exposed up-river at the “type locality” outcrops near the town of Montevideo. Samples of gneiss from the quarry were used by Goldich and others (1980a, their locality 4) to identify a sequence of Rb-Sr isotopic events that range from 3,530 to 2,265 Ma. Various interpretations of these

isotopic results are discussed at length by the original authors. U-Pb zircon data reported for an "older" adamellite intrusion into the layered gneiss yield an interpreted age of 3,230 Ma (Goldich and others, 1980a).

Dikes of Proterozoic hornblende andesite are prominently displayed in the quarry walls. These and all other rocks in the quarry are shot through with chloritized, slickensided, brittle shears on all orientations. This type brittle shearing and associated evidence of hydrothermal activity (chlorite films; calcite-quartz veins) are associated with a pronounced reddening of the rocks, presumably due to disseminated hematite. The reddening phenomenon affects many MRV exposures north of the YMSZ but is uncommon south of it.

END OF DAY ONE.

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INTRODUCTION TO THE GLACIAL HISTORY OF THE MINNESOTA RIVER VALLEY

Carrie J. Patterson

Contrary to the four-fold glacial chronology devised around the turn of the century for terrestrial glacial sediments (Wisconsin, Illinoian, Kansan and Nebraskan (Flint, 1957)), the current model for the Quaternary record of glaciations is much more elaborate (Ruddiman and Wright, 1987). The oxygen isotope record of ocean carbonate sediments records the effects of glaciations. During glaciations, ^{16}O , which is preferentially evaporated from the oceans because of its lesser mass, is stored on land in the ice sheets, enriching the glacial oceans in ^{18}O . According to this record of global ice volume, there were 40 glacial/interglacial oscillations involving moderate global ice volumes between 2.4 Ma and 0.9 Ma and there were 22 oscillations involving greater ice volumes from .9 Ma to the present (Fig. 3) (Shackleton et al., 1984). Counting back from our current oxygen isotope stage, stage 1, odd-numbered stages represent interglaciations and even-numbered stages, glaciations. Oxygen-isotope stages 2, 6, 12 and 16 show the greatest ice volumes, perhaps explaining the historical recognition of a four-fold record of terrestrial glaciation.

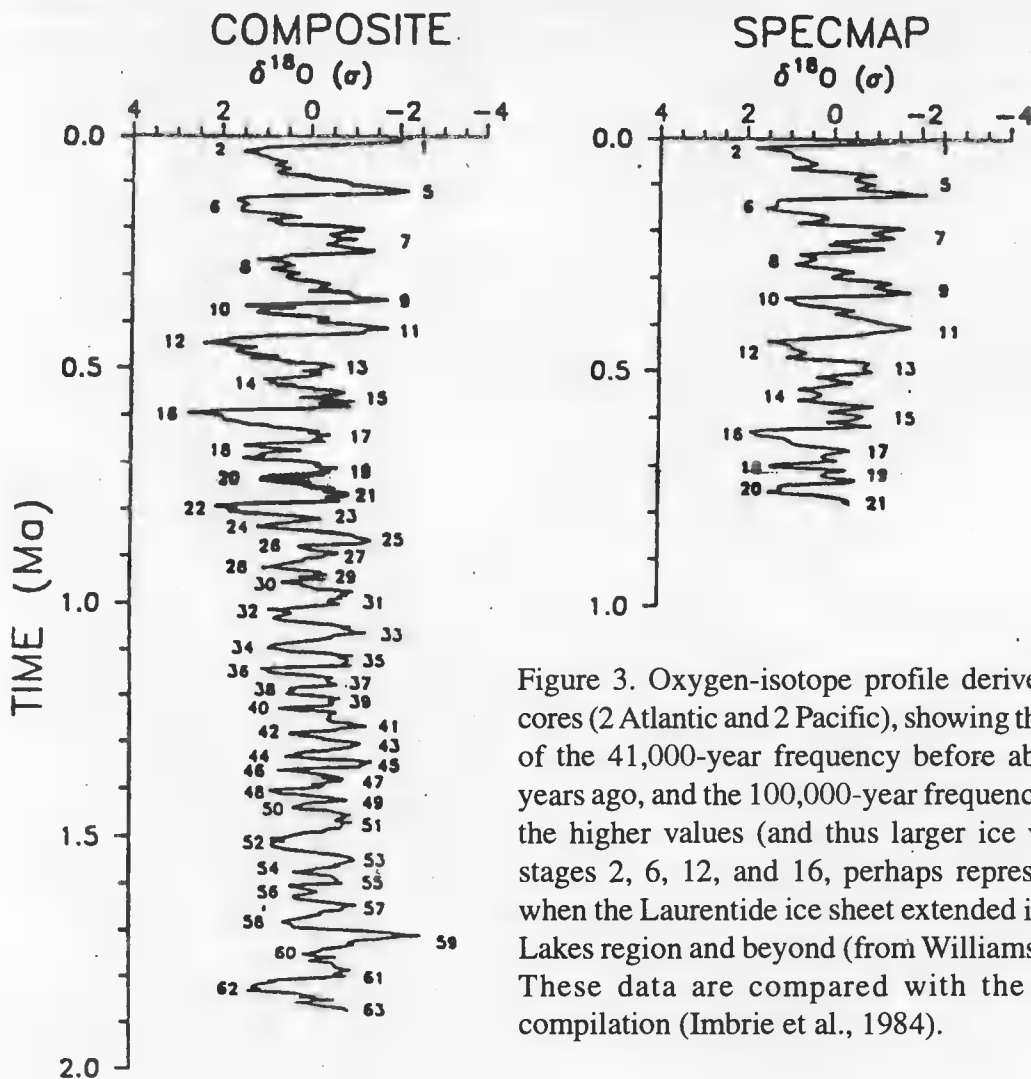


Figure 3. Oxygen-isotope profile derived from four cores (2 Atlantic and 2 Pacific), showing the dominance of the 41,000-year frequency before about 700,000 years ago, and the 100,000-year frequency after. Note the higher values (and thus larger ice volumes) for stages 2, 6, 12, and 16, perhaps representing times when the Laurentide ice sheet extended into the Great Lakes region and beyond (from Williams et al; 1988). These data are compared with the SPECMAP compilation (Imbrie et al., 1984).

During oxygen-isotope stage 2 (the late Wisconsin, retained from the former chronology) and possibly during stages 6, 12 and 16, Minnesota was at the edge of an ice sheet that covered most of Canada (Fig. 4).

Minnesota was affected mainly by thin, dynamic ice lobes that advanced, stagnated, and retreated independently of one another (Wright, 1972). Ice-lobe dynamics were controlled by the mass balance and bed conditions of their respective icesheds. Owing to the different substrates the ice flowed over, it is possible to distinguish the till of the different ice lobes using matrix color, texture and mineralogy, and clast lithology, and thus reconstruct the activity of the lobes. Landforms remain for the last few advances and aid in the reconstruction of the nature of ice advance and retreat.

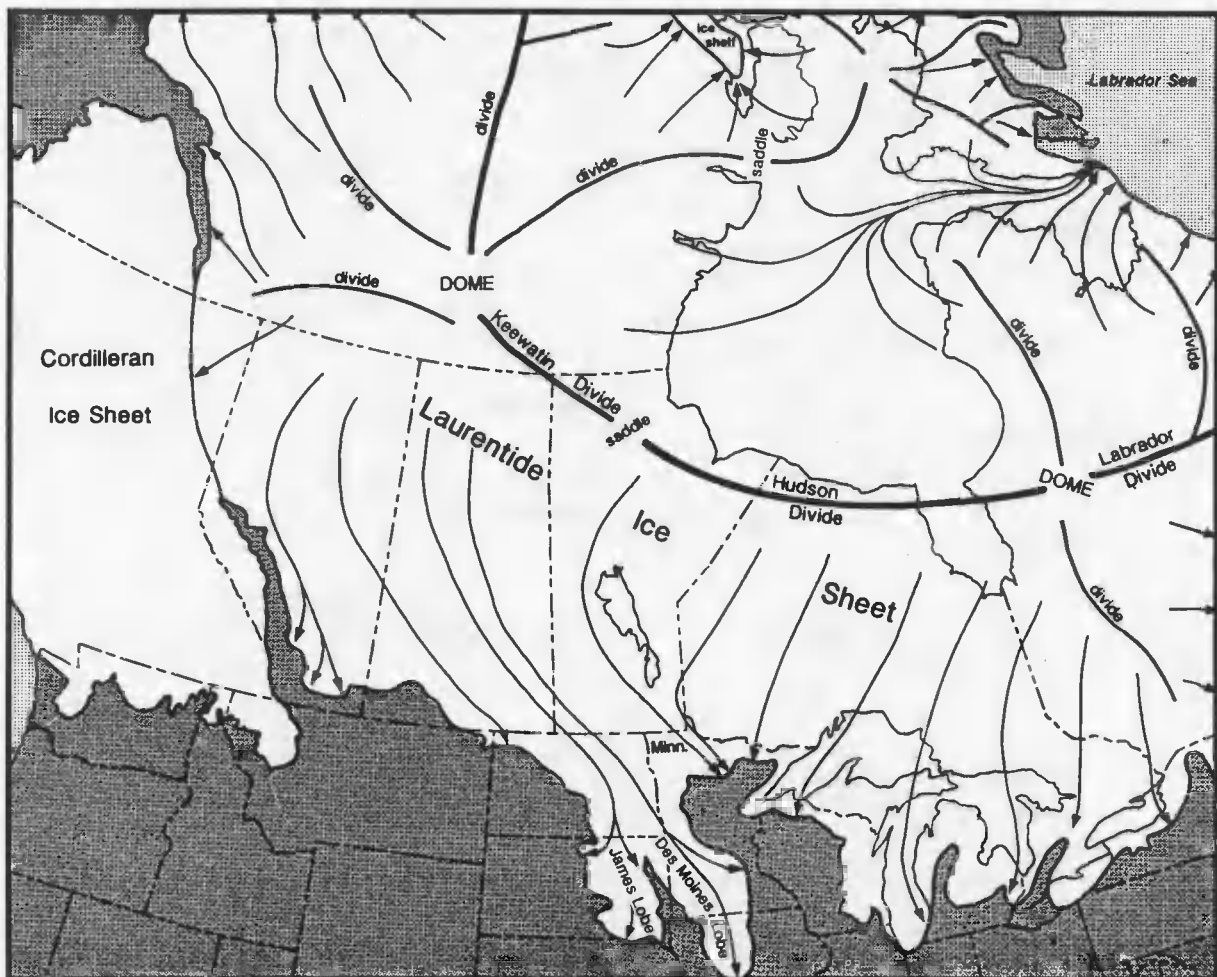


Figure 4. Laurentide ice sheet at 14 Ka. (Modified from Dyke and Prest, 1986).

Absolute dating of the late Wisconsin glacial sediments is possible using radiocarbon-dating methods if organic remains are found. Pre-late Wisconsin units are difficult to date absolutely (Ruddiman and Wright, 1987). A few scattered dates have been secured in the area using the presence of volcanic ash (Richmond and Fullerton, 1986), magnetic reversals preserved in till and lake sediment (Patterson, 1993), amino-acid dating of shells (Gilbertson, 1990) and cosmogenic isotope age estimates of striated rock surfaces (Bierman et al., in press).

The history of work in this area and the basic stratigraphy was described by Matsch (1972). The oldest glacial sediment in southwestern Minnesota is at least of mid-Pleistocene age (Fig. 5 A and B) (Patterson, 1997) and may even be of Late Pliocene/Early Pleistocene age (2.1-1.2 Ma) based on correlations with glacial sediment in Iowa and Nebraska (Hallberg, 1986). These deposits are attributed to northwest-sourced ice advances, which deposited gray, loam-to-clay-loam tills containing clasts of Paleozoic carbonate and Cretaceous shale. There are at least two tills of this age that are difficult to differentiate. Similar tills in South Dakota are thought to be .6 Ma to 2.1 Ma (Gilbertson, 1990).

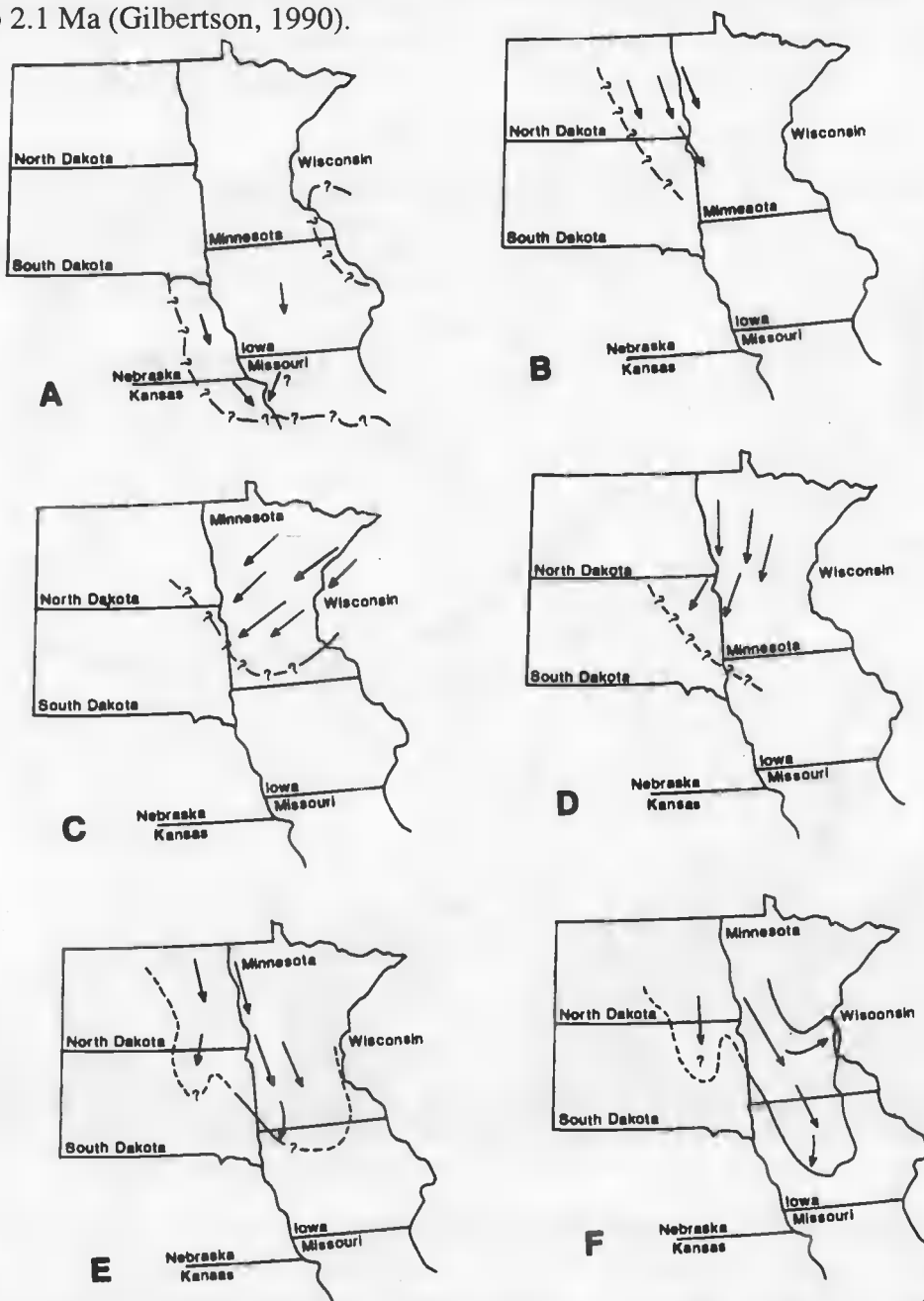


Figure 5. Ice advances into the region. A. Early Pleistocene advances, greater than 0.6 Ma; B. Ice advance from the northwest; C. Ice advance from the northeast; D. Ice advance from the north, deposition of Granite Falls till >40 ka; E. Late Wisconsin advance of early Des Moines lobe, 30-20 ka; F. Late Wisconsin advance of the Des Moines lobe, 14 ka. (From Patterson, 1993).

Stratigraphically above these gray tills is a northeast-sourced glacial sediment (Fig. 5C) that can be distinguished by its red-brown color, sandier texture, and clasts originating in the Lake Superior area including banded iron formation, agate, red clastics and basalt. This unit, the Hawk Creek till (Matsch, 1972) is only locally preserved and its southwestern extent is not known. It is thought to be >50 ka (Gilbertson, 1990).

The last pre-late Wisconsin glacial advance was from the north-northwest (Fig. 5D) and is similar to the older gray units but has less Cretaceous shale. Wood in this, the Granite Falls till, is >40,000 years old (Matsch, 1972).

The surficial glacial sediment in the Minnesota River basin was deposited by the late Wisconsin Des Moines lobe, which advanced, stagnated and retreated repeatedly. Its first advance may have been as early as 30-20 ka (Fig. 5E) (Gilbertson, 1990). The advance to Des Moines, Iowa is dated at 14 ka (Fig. 5F) (Clayton and Moran, 1982; Hallberg and Kemmis, 1986). The loam-textured till is gray when unoxidized and has Cretaceous shale and Paleozoic carbonate clasts. It generally contains Pierre Shale, whereas the older gray tills contain other Cretaceous formations such as the Greenhorn and Carlisle. The base of the Des Moines lobe till is locally marked by a planar concentration of faceted, striated boulders.

Glacial Lake Agassiz and Glacial River Warren

It is common for lakes to form in front of a retreating glacier, with meltwater ponding between a moraine and the ice front. A proglacial lake may not be very large or long-lived because of the instability of the moraine dam and ice margin. However, an unusually stable and large proglacial lake began to form in west-central Minnesota at approximately 12 ka as the ice retreated north of the Big Stone Moraine and the nonspectacular continental divide at Brown's Valley (Fig. 6).

For much of its early history, Glacial Lake Agassiz was drained by Glacial River Warren, which flowed southeast from the southern tip of the lake (Upham, 1883, 1895; Matsch and Wright, 1967). Initially Glacial River Warren cut a series of anastomosing channels with streamlined residual bars and boulder lags (Matsch, 1983). The style of channel formation is characteristic of sudden and large discharge from a lake that overwhelms the existing drainage system (Kehew and Lord, 1986). Flow was eventually focused in one main spillway which became more deeply incised, stranding earlier channels (Fig. 7).

Continued retreat of the ice front eventually opened lower outlets for Glacial Lake Agassiz to the northeast (Clayton, 1983) and possibly the northwest (Smith and Fisher, 1993) and Glacial River Warren was abandoned as an outlet. Tributaries to Glacial River Warren, which had steepened their gradients in response to the downcutting, were now contributing too much sediment to be carried away by the reduced flow of the river. Fans spread across the valley floor partially damming it, forming a series of long lakes (Wright, 1972; 1990) including Lake Traverse, Big Stone Lake, Marsh Lake and Lac qui Parle. The levels of these lakes are now controlled by dams for recreation and flood abatement.

The field stops will first address the most recent event related to glaciation, the formation of the Minnesota River Valley, and then move to progressively older events that are exposed because of river downcutting. See plate 1 (page 168) for stop locations.



Figure 6. Early stages of Glacial lake Agassiz. (Adapted from Teller, 1987).

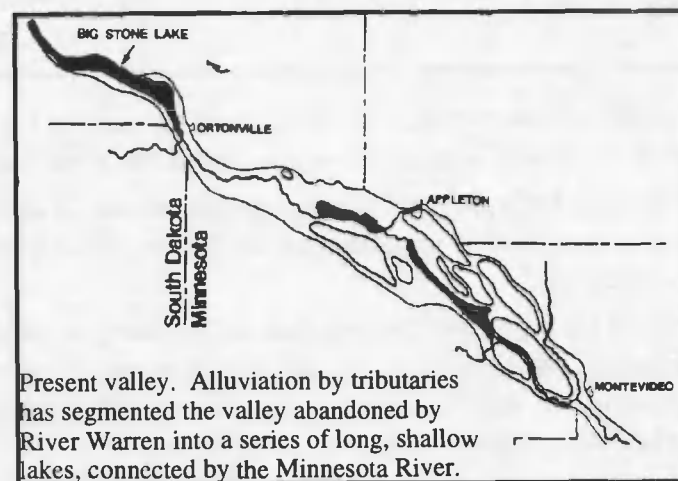
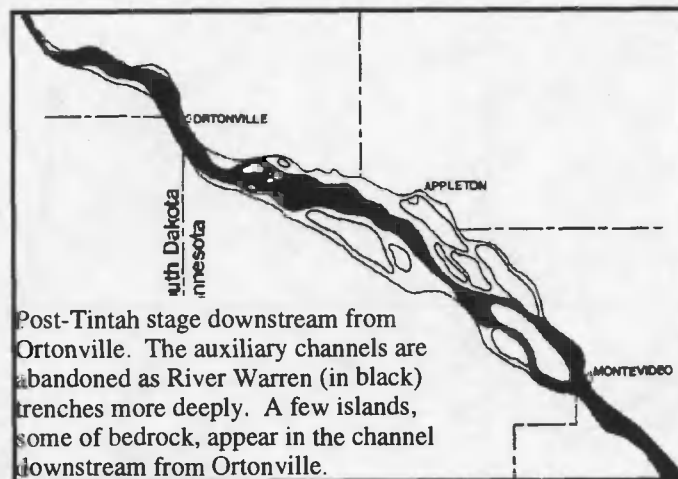
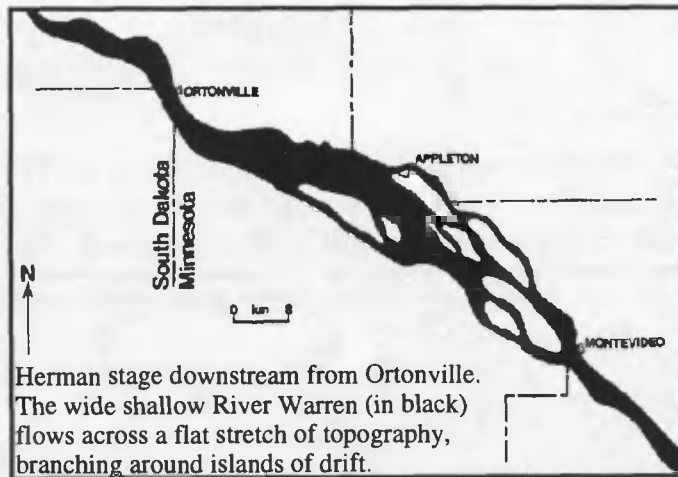
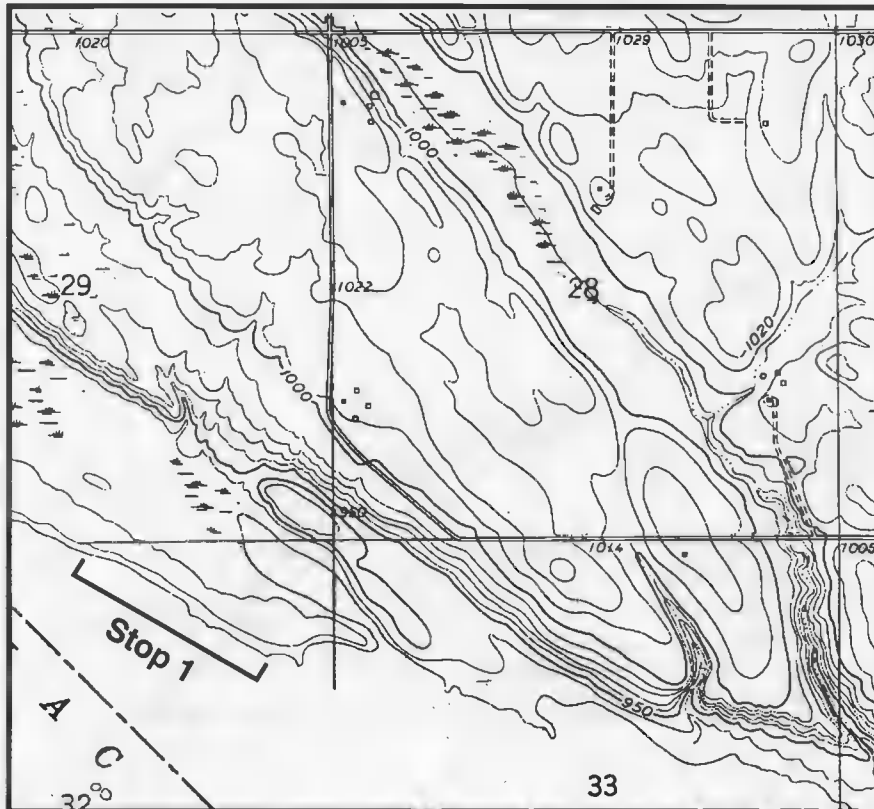


Figure 7. A. Initial stages of development of Glacial River Warren. B. Formation of main river trenches. Abandoned channels may be marked by river lags. C. Present-day Minnesota River Valley. Reproduced from Matsch, 1983.

FIELD TRIP STOPS—DAY 2

Stop 1—Ingebretsen's Landing, or "the clay banks" on Lac qui Parle. Milan quadrangle, T. 119 N.-R. 42 W.-section 32

The beach and low bank are formed in 2-3 m of reddish brown till. The till matrix has 43-48 percent sand and the clasts have a northeastern provenance; you may find Lake Superior agates, banded iron formation, basalt and red clastic rocks on the beach. This color, texture and



grain type are unusual for tills in this general area. This is not the surface till in the region but is normally found deeper in the glacial section and was exhumed by river downcutting.

In places there are remnants of another till overlying this red till. It can be distinguished by its siltier texture and clasts typical of an ice advance from the northwest, including Paleozoic carbonate and Cretaceous shale.

A discontinuous boulder lag represents the erosional discontinuity created as Glacial River Warren, the spillway for Glacial Lake Agassiz, cut into the glacial sediments, selectively removing the finer grains. The following sections from a paper by Kehew and Lord (1986) describe the general formation of glacial-lake spillways (Fig. 8).

The instability of the spillway rivers was exceptional; torrential discharges from glacial-lake outlets, following dam failure, had high energy and no sediment load. Consequently, flows which incised spillways were highly erosive, and channel sediment from these events is scarce. The flood deposits which do exist consist of homogeneous masses of very poorly sorted, pebbly cobble gravel, which indicates rapid aggradation in very turbulent water. p. 165

The morphological characteristics of spillways...indicate that the spillways were channels that contained flow at bankfull stage during one or more highly erosive floods. The size of the spillways qualitatively demonstrates that the flood discharges were enormous, and the estimated volumes of most proglacial lakes (Kehew and Clayton, 1983) require that bankfull discharges couldn't have been sustained for more than a few weeks. In short, spillways in the mid-continent region were incised by a few, geologically instantaneous events of colossal magnitude in comparison to modern river discharges. p. 165

At this location in Glacial River Warren, two subparallel channels were incised rather than one main channel: Lac qui Parle and Watson Sag. They are separated by the erosional residual that the town of Watson is located on. According to Kehew and Lord, (1986) erosional residuals are key to identifying floods that overwhelmed existing drainage systems.

In places there is an organic horizon associated with the boulder lag. A silty lake sediment caps the section. The organic horizon and lake sediment may represent Glacial River Warren

overbank deposits or more recent Holocene deposits. An examination of the organics for pollen to determine their approximate age before radiocarbon dating was considered yielded no identifiable grains.

The bank has retreated significantly in recent years and the DNR-fisheries is planning to riprap it to prevent further retreat. Their concern is that sediment loading of the lake is negatively affecting fish habitat.

Features to see:

- Lake Superior agates, banded iron formation, basalt and red clastic rocks on the beach indicative of a northeastern provenance for the till.
- Joints in till with narrow zones of reduced till indicating water movement along the joints.
- Boulder lag and associated organic horizon.

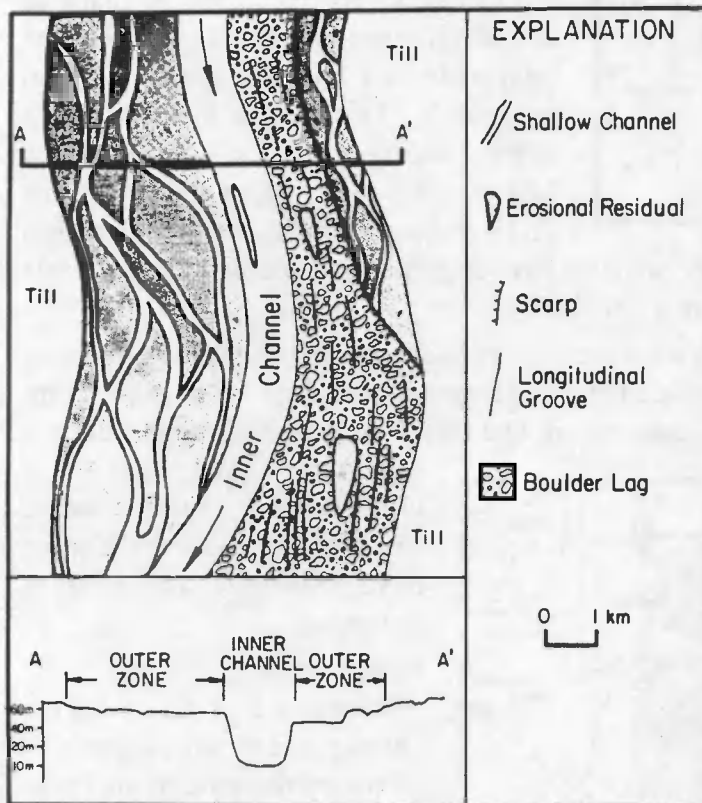
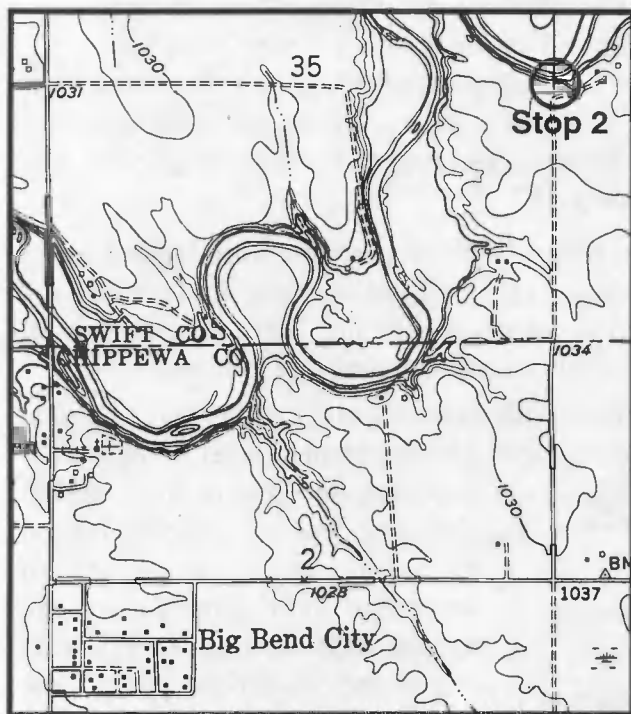


Figure 8. There are two basic spillway levels: (1) the trench-like, low sinuosity, inner channel; and (2) the broad outer zone with: (a) slightly scoured areas of anastomosing channels; and (b) the deeply scoured zone which displays longitudinal grooves, streamlined erosional residuals and boulder lags. From Kehew and Lord (1986).

Stop 2—Glacial Lake Benson sediment exposed along the Chippewa River. Big Bend City quadrangle, T. 120 N.-R. 41 W-section 36



Glacial Lake Benson, a moraine-dammed, 2800 sq km, proglacial lake in west-central Minnesota, formed during the late Wisconsin retreat of the Des Moines lobe (Fig. 9). It was the last in a series of lakes before the Des Moines lobe retreated north of the divide to form Glacial Lake Agassiz (Diedrick and Rust, 1975; Patterson, Cotter and Knaeble, 1998; Rittenour, Geiger and Cotter, in prep.)

Two-m-thick exposures of lake sediment along the Chippewa River in Swift and Chippewa counties show as many as 23 rhythmically bedded couplets interpreted as annual layers (varves) underlain by Des Moines lobe till. In a series of undergraduate and NSF-sponsored REU (Research Experience for Undergraduates) studies supervised by Jim

Cotter of University of Minnesota, Morris, students have described and counted varves at five sites; 44 varves were correlated (Rittenour, 1994; 1995).

In this exposure you will note that in some areas the beds are overturned. We interpret this as a result of the lake forming in an area where there was stagnant glacial ice which melted out, disrupting some of the beds. The upper, massive silt and clay may have been deposited in a

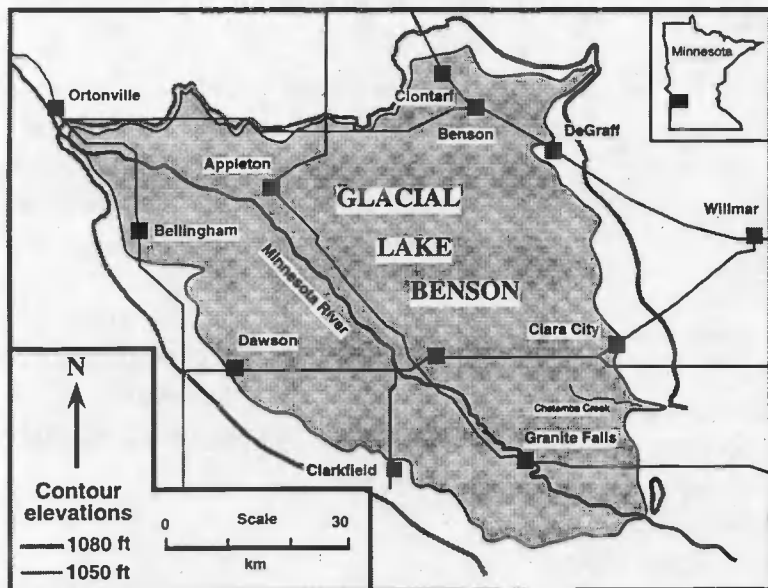


Figure 9. Approximate extent of Glacial Lake Benson (from Rittenour et al., in prep).

shallower lake that was unfavorable to varve development because the wave base disturbed the bottom sediment.

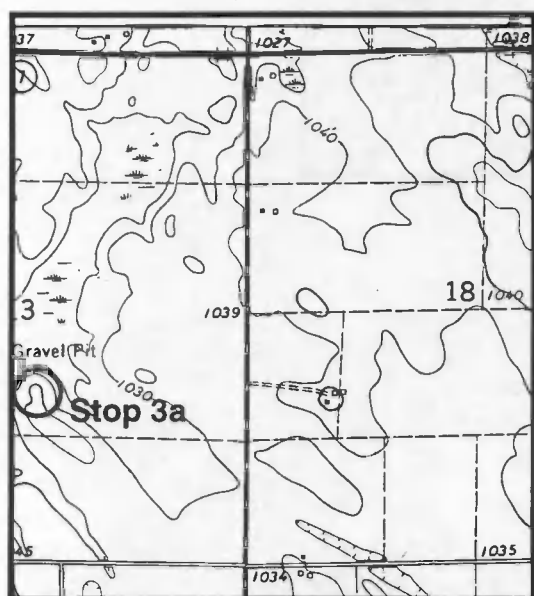
Features to see:

- Difference in grain size of the winter and summer layers.
- Upward decrease in thickness of summer layer indicating lessening sediment load and retreating glacier.
- Overturned beds.
- Dropstones.
- Nature of contact with underlying Des Moines lobe till.

Stop 3—Ice-confined streams. Montivideo landfill and nearby pits. Montevideo quadrangle, T. 117 N.-R. 40 W.-section 12; Asbury quadrangle, T. 117 N.-R. 40 W.- section 13 and T. 117 N.- R. 39 W.-section 33.

These three pits have been developed in southeast-oriented, discontinuous ridges that are generally less than 7 m high. We may only visit one or two of the pits depending on the quality of the exposures and time constraints. These ridges and others oriented transverse to them form a rectilinear pattern that is common in the interior of the Des Moines lobe.

Ridges parallel to ice flow, and therefore to former ice surface slope, are predominantly composed of sorted sediments. The transverse ridges are composed of sediment ranging from diamicton to sand and gravel. Most of the sampled transverse ridges were only slightly coarser and more sorted than the underlying and surrounding till.

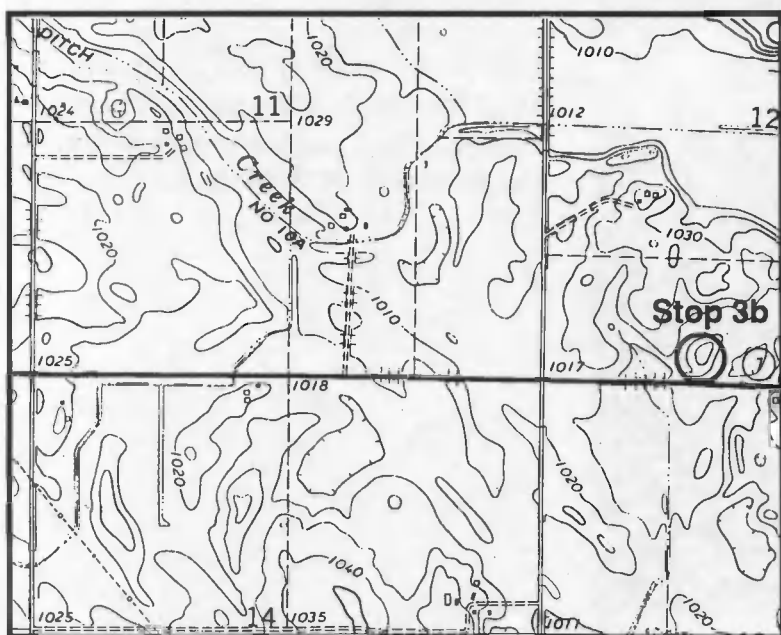


The ridges are interpreted to be a result of deposition localized in the pervasively crevassed stagnant ice of the Des Moines lobe (Fig. 10) (Matsch, 1972; Patterson, 1997).

Crevasses oriented parallel to the ice surface slope were preferentially used by supraglacial streams. Final melting of the ice walls of the streams inverted the topography and left sorted stream sediment as a positive element in the landscape.

Glacial Lake Benson outlets were ice-confined and the pits east and north of the landfill are interpreted as part of a late-stage outlet of Glacial Lake Benson. In the northern pit, well sorted sand is unconformably overlain by poorly sorted pebbly sand which has an indistinct,

conformable contact with the overlying diamicton. This is interpreted as steady discharge followed by a larger, short-lived discharge causing the collapse of stagnant ice resulting in flow of diamicton into the channel.



Features to see:

- Climbing ripples.
- Uncollapsed beds indicating no ice beneath the stream when the sediments were deposited.

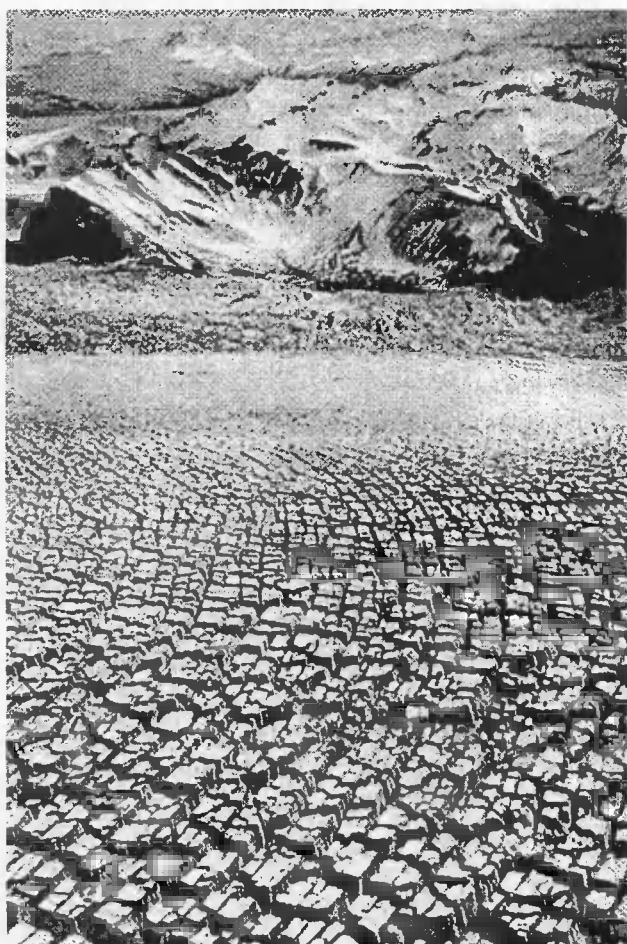
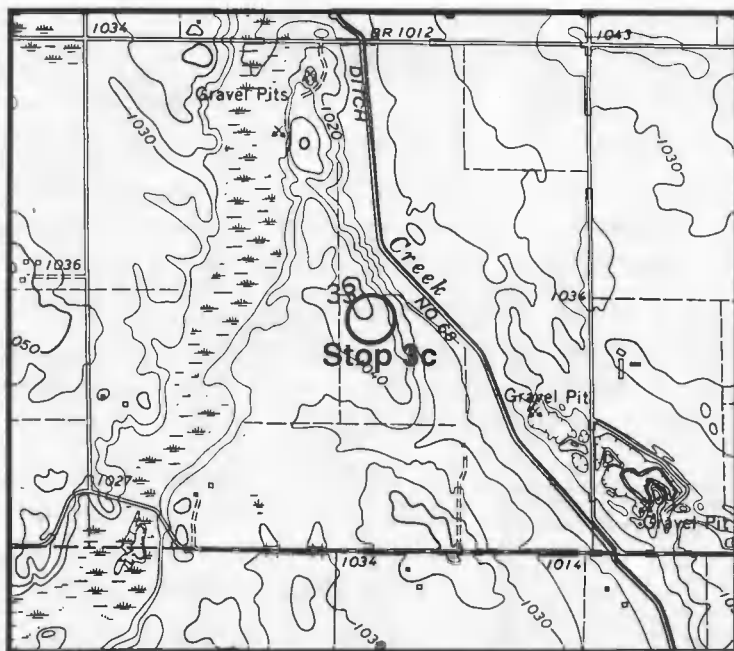


Figure 10. Heavily crevassed, post-surge ice surface of the Bering Glacier, Alaska. Photograph reproduced with the permission of B. Molnia.

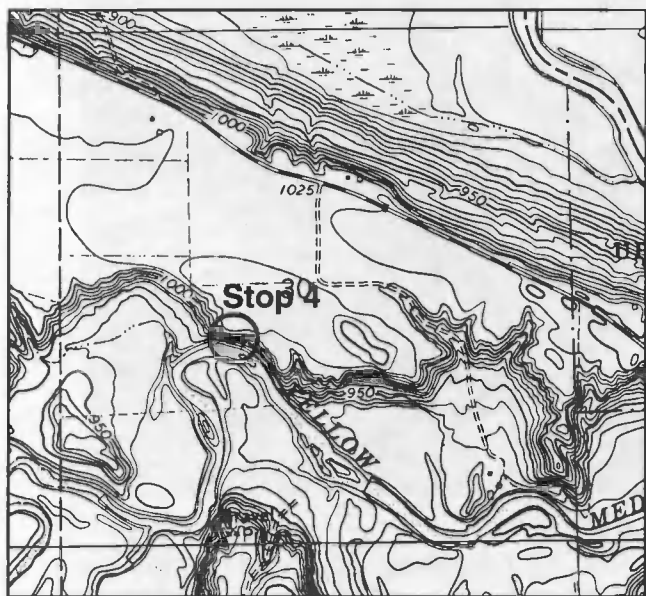
Stop 4—Glacial Stratigraphy, Upper Sioux Agency State Park, exposure on the Yellow Medicine River. Lone Tree Lake quadrangle, T. 115 N.-R. 38 W.-section 30

WARNING

This is an unstable outcrop. Be aware of the position of others on the slope. Do not follow another person up a gully. Look for people below you before digging or descending. Climbing up is easier than getting down. Hardhats are recommended.

If you chose not to climb around on this outcrop, samples of the till and washed pebbles from the different units will be provided for viewing on top of the bluff. Do not step too close to the edge. There are rotational slump blocks waiting to descend; avoid becoming colluvium.

Nearly 100 ft of glacial sediment is exposed along this meander bend of the Yellow Medicine River (Fig. 11). This is almost the entire thickness of glacial sediment at this location.



Cross sections constructed from water-well data show the regolith surface at approximately 930 ft in the immediate area; this is just a few feet below the river bed.

The uppermost unit is a 2-3 m layer of Des Moines lobe till. It is not accessible from this outcrop. Please do not try!

There is a discontinuous boulder pavement at the base of the Des Moines lobe till. The boulders are faceted and striated and form a planar surface. The boulder pavement, although not continuous, is found at this stratigraphic position over most of the interior of the Des Moines lobe in Minnesota (Wright et

al., 1973). The faceting and striating had to develop beneath a moving glacier that did not, however, have the shear stress to move the boulders very far because underlying beds are generally not disturbed. The boulders may have been concentrated at the surface before the glacier overrode them by water or periglacial processes. Alternatively, some have proposed that the boulder concentration developed subglacially, by large clasts settling through an incompetent till matrix (Clark, 1991). The issue is still unresolved.

The next unit is exposed is the Granite Falls till. This may be an atypical exposure of the Granite Falls till; it is siltier than normal and layered with sorted sediment. Toward the base of this unit, near the contact with the underlying fine sand, discrete lobes of diamict can be recognized. These features are interpreted as the result of flow of the till into standing water. This is a very plausible scenario given the unit immediately below which has been interpreted as a lake sediment.

The unit interpreted as lake sediment is mainly a fine sand with silt and some clay. In places it is finely bedded. The extent of the lake sediment beyond this outcrop is not well



Figure 11. Yellow Medicine River exposure in Upper Sioux Agency State Park looking downriver. Photograph by Star Mulder

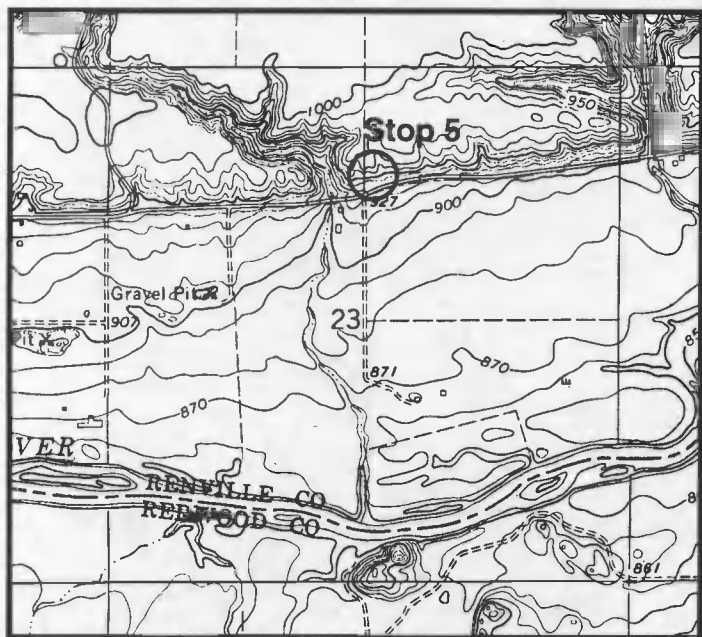
known. From this exposure it appears to be a relatively deep lake and therefore could have covered a large area.

Beneath the lake sediment is a thick, gray loamy till. Note the many well-striated Paleozoic carbonate clasts. The lower till has not been well exposed here but at other cuts along this river there are two gray tills separated by a glacial stream sediment at this stratigraphic position.

Features to see:

- Striated, bullet-shaped clasts of Paleozoic limestone from lower till.
- Path of groundwater flow through glacial units. Note areas of lateral flow.
- Contact of the Granite Falls till with underlying lake sediment.
- Evidence for depositional environment of Granite Falls till.
- Boulder pavement (from a distance, please).

Stop 5—Inge Kea's climbing ripples. Iverson Lake quadrangle, T. 114 N.-R. 37 W.-section 23 (this stop is out of stratigraphic order along the field trip route)

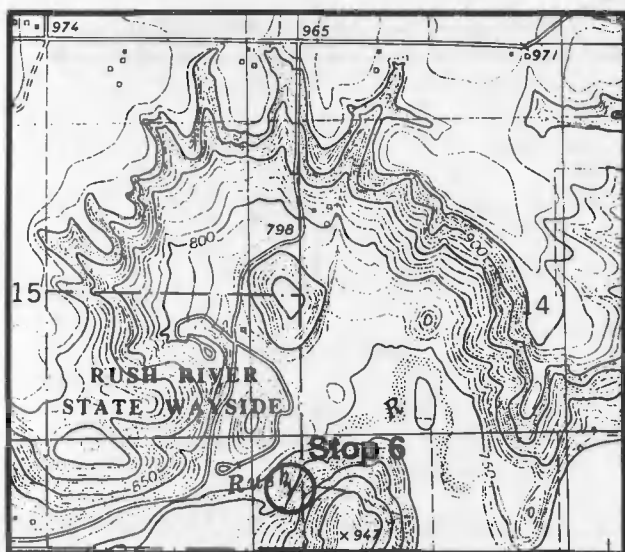


This sand pit is located in an upper terrace along the Minnesota River. The deep channel below us is part of the Glacial River Warren spillway. Glacial outwash streams flowed from the retreating Des Moines lobe in the general area of Glacial River Warren prior to its formation. This exposure is most likely a remnant of an earlier outwash stream. The depositional environment is distinctly different from the first stop of the day.

Features to see:

- Variation in migration and aggradation rates (angle of climb).

Stop 6—Rush River Wayside; Deformation of glacial sediment. Henderson quadrangle T. 112 N.-R. 26 W.-section 15



The downcutting of this east-flowing tributary to the Minnesota River has exposed glacial units that display deformation induced by the advance a glacier. Mark Johnson of Gustavus Adolphus College has directed undergraduates from his school and from an NSF-sponsored field program in work on the details of this outcrop (Fig. 12).

The lowest unit is a gray loamy till and it is conformably overlain by lake sediment, that, judging by the color, formed in front of the retreating, gray-till-bearing ice lobe. The gray lake sediment grades into a red sand with silt layers. This sandy unit records the advance of an ice lobe from the northeast.

The next unit is a mixture of the underlying sand with a red sandy loam till. There is a sharp contact between this and the final unit exposed, a homogenous red sandy till.

Folds: The lower two units are folded into three anticlines that span the outcrop.

Faults: The red sand displays a subparallel series of high-angle normal faults.

Fabric: The long axis of prolate pebbles in a till are commonly measured and plotted on a Rose diagram. Pebbles in a subglacial till may be oriented by the shear stress of the overriding glacier; pebbles are generally parallel to flow, with a slight up-glacier dip. The red till has a fabric suggesting ice flow from the north-northeast.

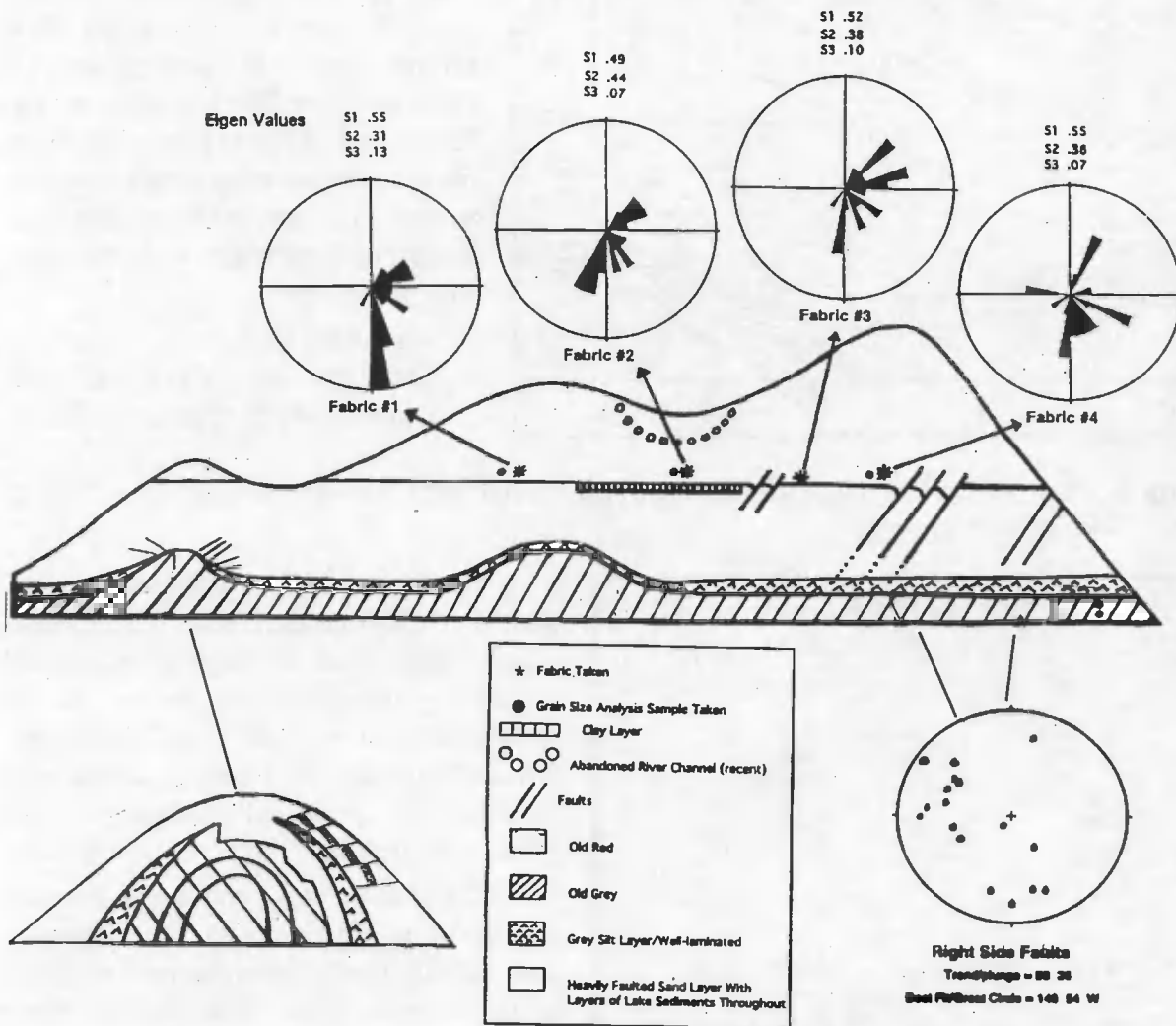


Figure 12. Structural data for Rush River outcrop. Unpublished work of Amy Moe, Gustavus Adolphus College student, 1996.

Question

Is the structural data consistent with the simple interpretation of proglacial deformation and eventual overriding by the advance of the red-till-bearing ice lobe from the north-northeast (Fig. 13)?

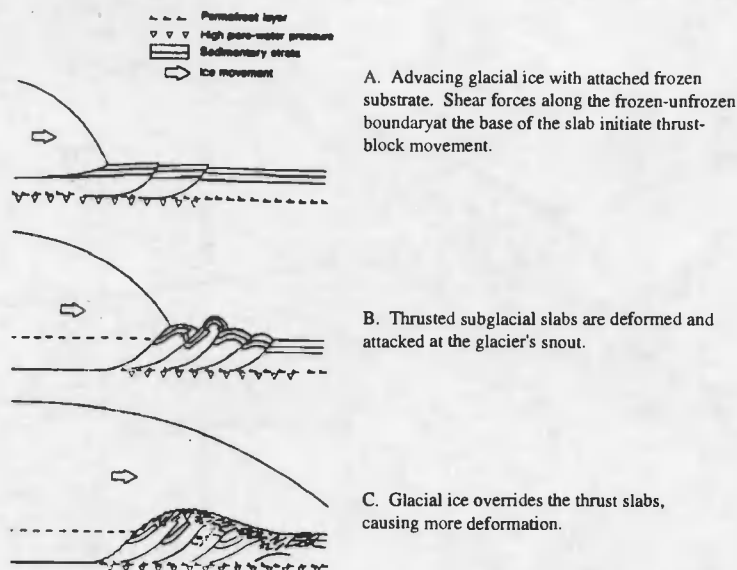


Figure 13. Sequence of proglacial deformation A and B with eventual overriding by the advancing glacier (from Aber, 1988).

TERRACES ALONG THE MINNESOTA RIVER NEAR SHAKOPEE, MINNESOTA

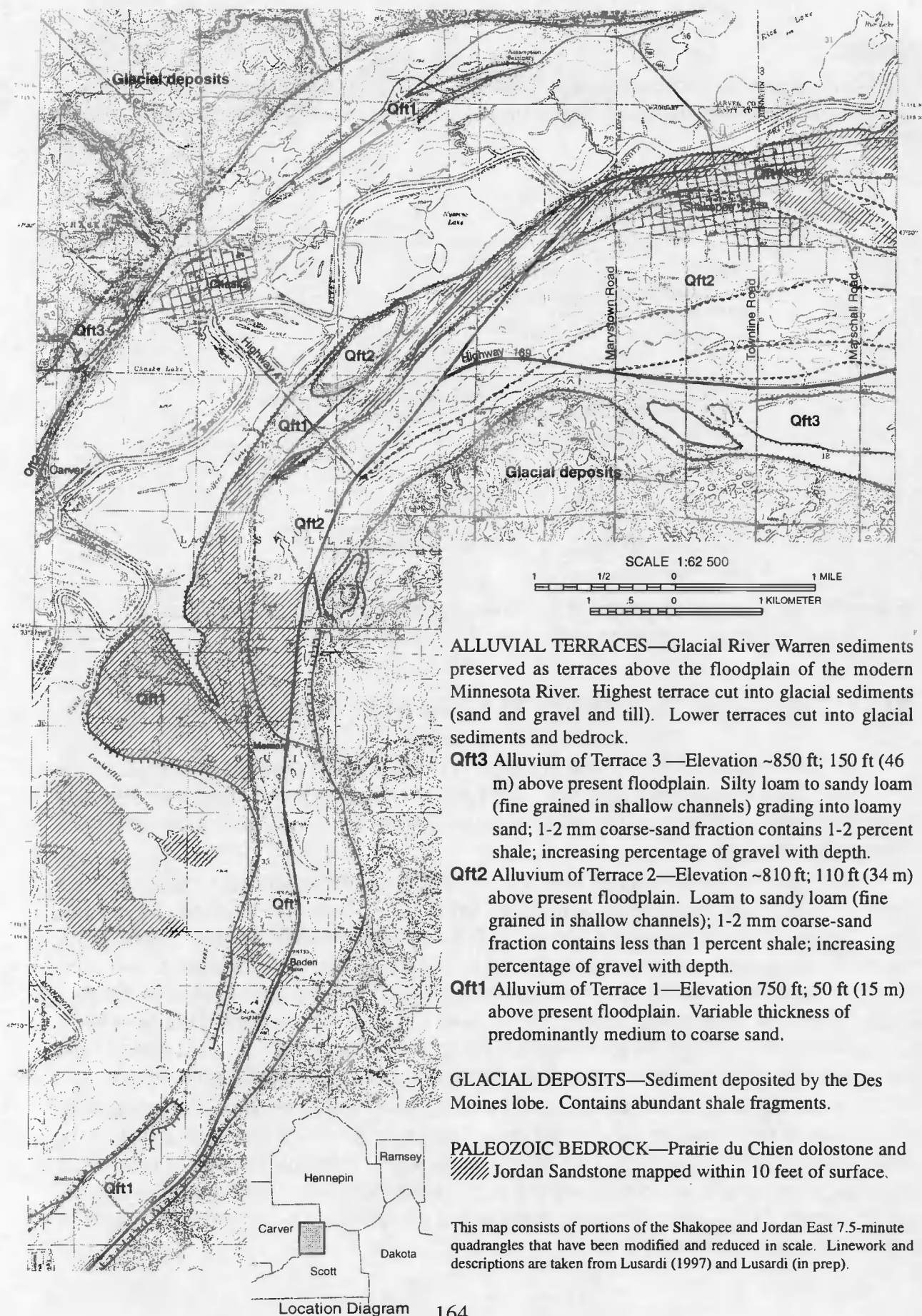
Barbara Lusardi

The Minnesota River Valley near Shakopee is almost 200 feet deep and over 3 miles wide in places. Till highlands on either side of the valley rise to an elevation of over 900 feet. Distinct terraces step down into the valley where the underfit Minnesota River meanders across a broad alluvial plain.

Three terraces are distinguished based on elevation. The uppermost terrace, terrace 3 (Qft3), is visible just to the south of highway 169 between Marschall Road and Marystown Road. Terrace 3 is at an elevation of about 850 feet—150 feet above the modern floodplain. A distinct scarp separates the terrace from the till upland to the south, and smaller channel scars crisscross the surface breaking the terrace up into three small “islands” above the level of terrace 2 (Qft2). Alluvial sediments of terrace 3 are up to 40 feet thick over bedrock. Data from water well logs indicate that sand and gravel extends south beneath the till upland. This suggests that the terrace sediments may be outwash deposits that have been eroded to form the terrace.

The new highway 169 follows the middle terrace, terrace 2, at an elevation of approximately 810 feet—110 feet above the modern floodplain. Numerous shallow channels crisscross the terrace surface. Alluvial sediments are 10-50 feet thick over Paleozoic bedrock. Just south of Shakopee, terrace 2 is about 1.5 miles wide. It narrows to the southwest and eventually disappears against terrace 1 (Qft1) near Merriam. Many gravel pit operations occur within sediments of terrace 2 and into the bedrock below.

Figure 14. Terraces of the Minnesota River Valley near Shakopee and Chaska.



The lowest terrace is at an elevation of approximately 750 feet—50 feet above the modern floodplain. In many places the underlying bedrock is at or near the surface. Ordovician Prairie du Chien dolostone is the principal source for crushed stone. Cambrian Jordan Sandstone is exposed in limited areas near Merriam and Baden. The only known sandstone quarry, just west of Baden, is now abandoned.

On the opposite side of the river, the terraces are narrow or obscured. Just northeast of Chaska, sediments of terrace 1 are buried by colluvial sediments eroded from the valley wall. Only small remnants of this lowest terrace are visible above the floodplain. Terrace 2 is present as a very narrow strip, less than 1/4 mile wide, between Chaska and Carver. Terrace 3 is present south of Chaska and widens dramatically south of Carver. Sand at the same elevation of terrace 3 continues to the south in a band almost 2.5 miles wide along the river, although no distinct scarp is obvious. Unlike the other terrace surfaces, this area is pitted and channeled, and may be related to an earlier flood event. Similarly, thick sand buried by 50-100 feet of glacial till on the north side of the valley, north of Shakopee, suggests that the river valley at one time extended further to the north. The sand was covered by glacial sediment during the last glacial advance and was never re-excavated by subsequent floods.

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Plate 1. Preliminary surficial geologic map of the upper Minnesota River basin showing stop locations.