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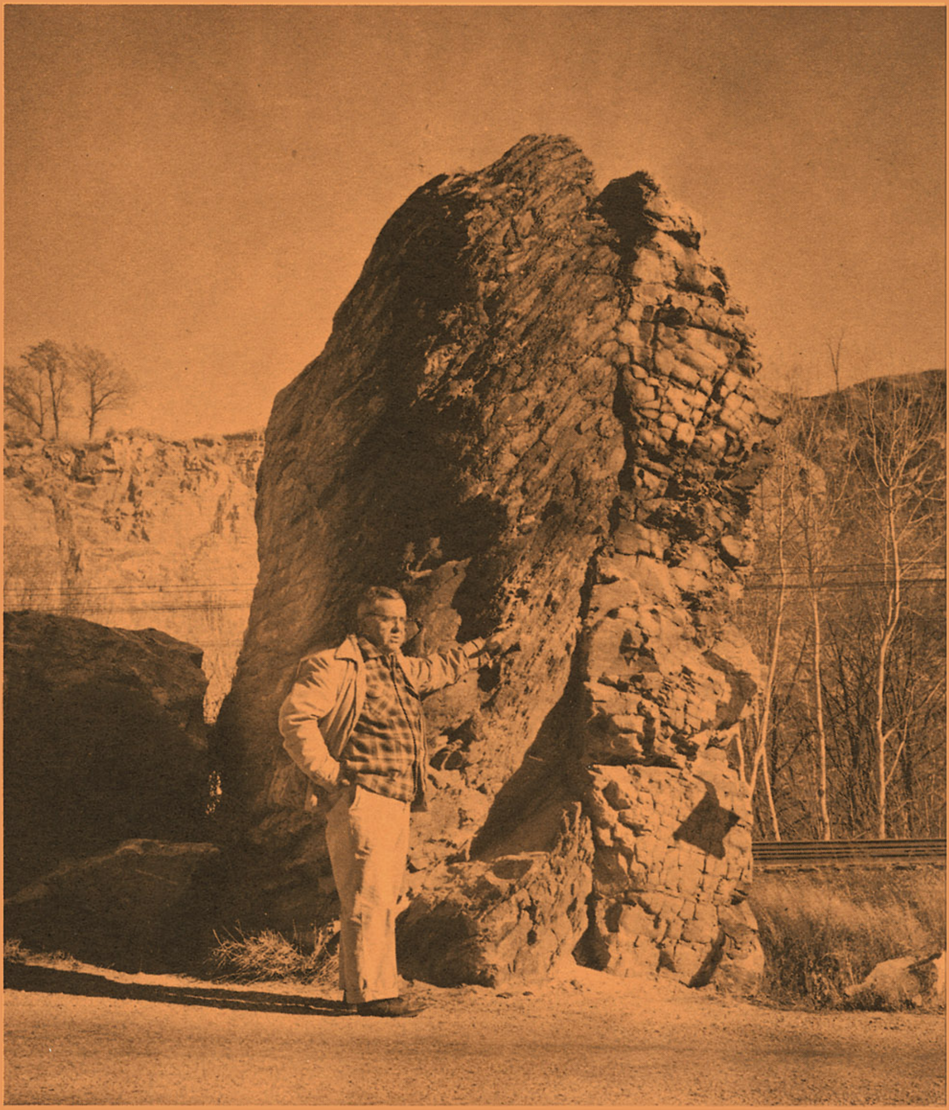
Geology of the Baraboo District Wisconsin



Geological and Natural History Survey

George F. Hanson,
State Geologist and Director

Madison, Wisconsin
November, 1970

Information Circular
Number 14

GEOLOGY OF THE BARABOO DISTRICT, WISCONSIN

A Description and Field Guide Incorporating
Structural Analysis of the Precambrian Rocks
and
Sedimentologic Studies of the Paleozoic Strata

by

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and

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With Summaries of the

GLACIAL GEOLOGY

by

R. F. Black
University of Connecticut

and

PLANT ECOLOGY OF THE BARABOO HILLS

by

J. H. Zimmerman
University of Wisconsin Arboretum, Madison

Available from the Geological and Natural History Survey, 1815 University Avenue, Madison, Wisconsin, 53706. Price $8.00

GEOLOGY OF THE BARABOO DISTRICT, WISCONSIN

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FOREWORD

For more than half a century, the Baraboo district has been an unusually important geological laboratory that has contributed to the education of countless thousands of students representing at least 100 different colleges. The list of teachers and research workers who have been drawn to the Baraboo hills reads like a "Who's Who" of American geology. Yet, in spite of all of the long-standing interest in the region, a satisfactory modern geologic map and detailed synthesis of Baraboo geology has not been available. It was to rectify this omission that this study was undertaken under the direction of the Wisconsin Geological and Natural History Survey, the Univ. of Wisconsin-Extension. We hope to have presented new data and ideas that will stimulate the further interest of both students and professional geologists.

At the same time that this guide to the geology of the Baraboo district is presented, we wish to voice a strong plea to preserve the priceless key outcrops from the onslaught of geology hammers in the hands of eager but unthinking visitors. Rock outcrops, too, sometimes require conservation-mindedness lest the hills be peneplaned by none other than geologists! Ironically, nothing is to be gained by hammering at such localities, for the structural relationships are best seen on weathered surfaces.

We are most grateful to George F. Hanson, State Geologist, and to Meredith E. Ostrom, Associate State Geologist, for their continual interest throughout the study and for their critical readings of the manuscript. Some financial support in addition to that of the Survey was provided by the Wisconsin Alumni Research Foundation (research grants to R.H. Dott, Jr.), the National Science Foundation (grant GA 12926 to I.W.D. Dalziel at Lamont-Doherty Geological Observatory of Columbia University), and Columbia University (field expenses to I.W.D. Dalziel).

Many of our colleagues at the University of Wisconsin and at Columbia University helped us in one way or another. In particular we thank Sturges Bailey (for unpublished X-ray data), Carl E. Dutton of the U. S. Geological Survey in Madison, Campbell Craddock, Marshall Kay, Charles V. Guidotti, Gerry L. Stirewalt (for permission to use unpublished microscopic data), and Richard E. Bischke. Kenneth O. Stanley assisted with petrographic work and Robert Blodgett and Paul Wells performed some of the size analyses. Unpublished geologic maps by Enis Usbug, Dennis Howe, Douglas Hackbarth, and Sharon Kreutzman were very helpful, as was some unpublished sedimentologic data obtained by Mary Roshardt.

Drafting was done in the University of Wisconsin Cartographic Laboratory by M. Czechanski, A.L. LeBlanc, and S.J. Thompson under the direction of Randall D. Sale, in-charge of Cartography for the Geological and Natural History Survey. We also wish to thank Kenneth I. Lange, Naturalist at Devils Lake State Park, Peter Monkmeyer, University of Wisconsin College of Engineering, Robert L. Miller of the University of Chicago, and Grant Cottam of the University of Wisconsin Department of Botany for assistance in criticizing the manuscript.

ISOTOPIC DATE PROVINCES OF THE GREAT LAKES REGION

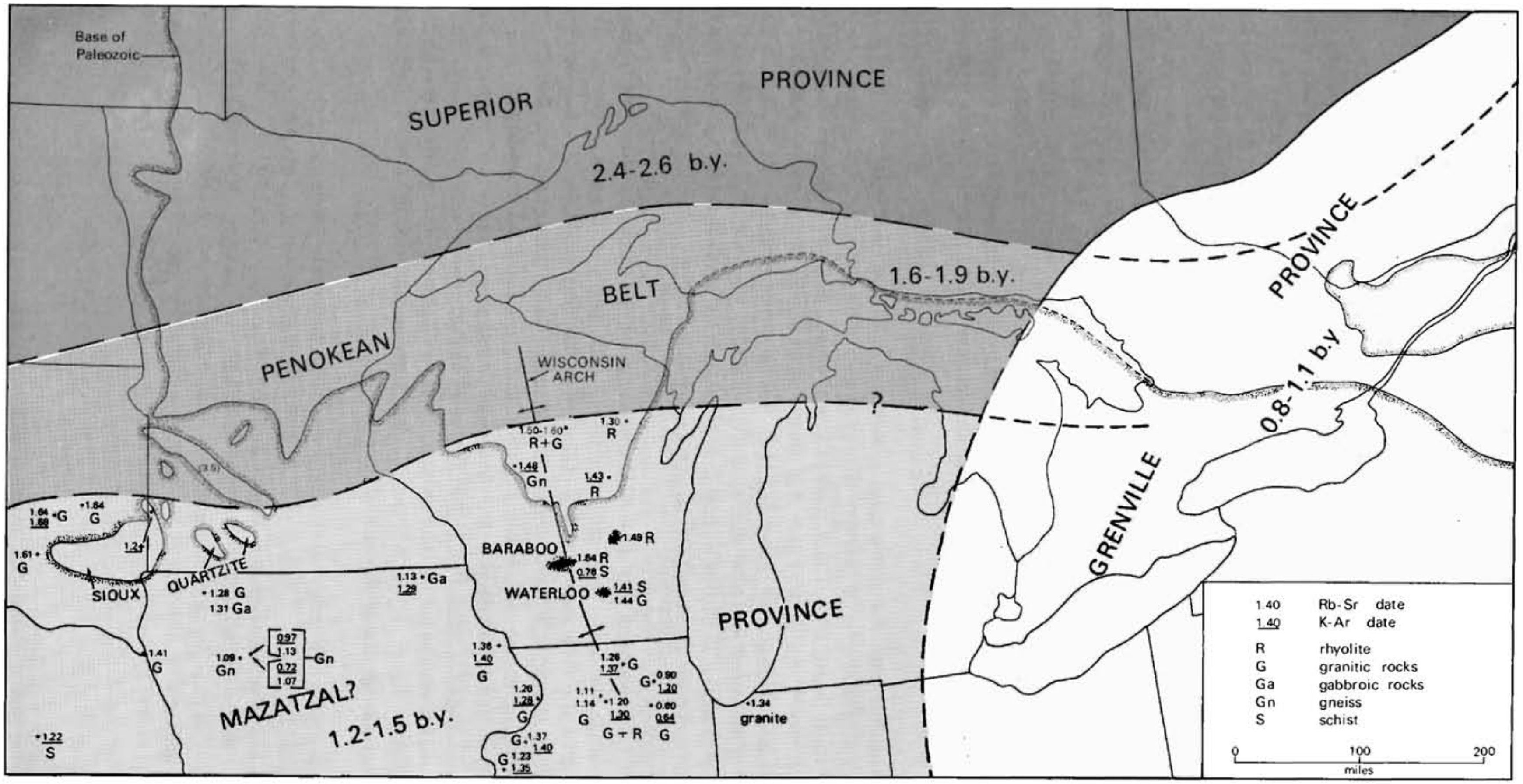

Figure 1. Regional tectonic and index map of the Great Lakes Regions, showing the Baraboo and Waterloo Quartzites with respect to major Precambrian isotopic-date and tectonic provinces. Note also limits of Paleozoic strata. Individual isotopic dates are included in the Mazatzal Province, which encompasses the Baraboo region. The north-south-trending Wisconsin arch is shown in southern Wisconsin. (Adapted from Goldich, et al, 1966).

INTRODUCTION

(Dalziel and Dott)

Regional Geologic Setting

Precambrian basement rocks are locally exposed along the axis of the Wisconsin Arch as inliers in the flat-lying lower Paleozoic sediments. The largest and best known of these inliers occurs in Columbia and Sauk Counties, south-central Wisconsin, where Precambrian rocks -- mainly quartzite -- form an elongate ring of hills known as the Baraboo Ranges (Fig. 1). The hills rise to a maximum elevation of 700-800 feet above the level of the Wisconsin River Valley (pl. I). Sedimentary rocks of Cambrian and Ordovician ages unconformably overlie the Precambrian basement and contain spectacular basal conglomerates in many places.

The Baraboo Ranges are exhumed monadnocks of massive, pink, maroon or purple-colored Baraboo Quartzite. This conspicuous rock unit is certainly more than 4,000 feet thick, and appears to rest stratigraphically upon poorly exposed acidic igneous rocks. Precambrian metasedimentary rock units overlying the Baraboo Quartzite have been recorded in drill and iron mining records (Weidman, 1904; A. Leith, 1935; Schmidt, 1951), but have not been positively identified in outcrop. The Precambrian succession in the Baraboo district inferred from subsurface as well as surface data is shown in Table 1.

TABLE 1

PRECAMBRIAN STRATIGRAPHY OF THE BARABOO DISTRICT

Rowley Creek Slate (maximum known thickness 149 feet)

Dake Quartzite (maximum known thickness 214 feet)

(Unconformity?)

Freedom Formation (dolomite and ferruginous slate; minimum thickness 1000 feet)

Seeley Slate (maximum known thickness 370 feet)

Baraboo Quartzite (thickness over 4000 feet)

(Unconformity?)

Rhyolitic "basement" (thickness unknown)

Subsurface information suggests that in places the Dake Quartzite rests on the lower (ferruginous slate) member of the Freedom Formation. This led A. Leith (1935) to propose an unconformity beneath the Freedom Formation. However, according to Leith, the records have been lost and the cores destroyed. He reported that 42 diamond drill cores in the eastern part of the basin penetrated the Dake Quartzite, and suggested that an outcrop on a low ridge northeast of Baraboo, and another in a railroad cut southwest of Baraboo, might be of Dake Quartzite. However, only a local pebbly unit in the former is unlike typical Baraboo Quartzite, and there is no strong reason for believing that the Dake Quartzite, if it exists, does indeed crop out.

The entire Precambrian succession has been folded into a complex doubly-plunging asymmetric syncline with an axial surface striking approximately east-northeast-west-southwest and dipping steeply north-northwest. At the present level of erosion, the syncline is about 25 miles long and has a maximum width of 10 miles. The north limb is nearly vertical and the south limb dips gently northwards.

The Precambrian rocks form a structural and topographic basin infilled with Paleozoic and Pleistocene sediments.

History of Research

The very existence of an inlier of metamorphic and igneous rocks in the region probably would have attracted much attention from geologists. However, the work of C. R. Van Hise, C. K. Leith and W. J. Mead (best known of the "Wisconsin school" of structural geologists), in the latter part of the nineteenth and early part of the twentieth centuries, has made the beautifully-displayed structures in the deformed metasediments at Baraboo famous to structural geologists throughout the world.

The earliest recorded geological observations at Baraboo were those of Shumard (1852) reported by Owen in his account of a survey of Minnesota, Iowa and Wisconsin. Percival (1856), in his report of geological survey commissioned by the State of Wisconsin, remarked on the similarity of the quartzite forming the North and South ranges at Baraboo, and suggested that they might in fact be part of the same formation. He regarded the quartzite as the metamorphic equivalent of the lower Paleozoic sediments. Hall (1862) first correctly assigned the quartzite to the Precambrian and correlated it with the Huronian of Ontario. Winchell (1864) reported Cambrian fossils near Devils Lake, but it was left to Irving (1872) during the first extensive study of the area to prove the unconformable relationship between the quartzite and the Cambrian sediments. Irving also made the first detailed structural observations, recording the general attitude of the cleavage in phyllitic lenses within the quartzite in the South Range (Irving, 1877). Like Chamberlin (1873) he published a tentative cross section in which he inferred the quartzite of <u>both</u> the North and South Ranges to lie on the north limb of a broad upright anticline. Irving and Chamberlin apparently were misled by the overturned north-dipping beds on the north limb of the syncline in the vicinity of the Upper and Lower Narrows of the Baraboo River (Pl. I). Salisbury and Atwood (1903) were similarly misled. Although they showed the synclinal relations of the quartzite beds on the North and South Ranges, they incorrectly showed a tight anticline within the North Range.

The most comprehensive account of the Baraboo Quartzite and the structural geometry of the Baraboo syncline is that of Weidman (1904). A. Leith (1935) later described the lithology of the overlying Precambrian units after extensive drill records became available. Weidman also seems to have been the first worker to record correctly the structural geometry of the quartzite; he based his interpretation on extensive observations of dip and strike. However, it is generally accepted that Van Hise was first to appreciate fully the structural configuration of the quartzite on the basis of bedding/cleavage relations. Van Hise, C. K. Leith and Mead largely concerned themselves with studies of "rock cleavage" and correlation of the succession at Baraboo with Precambrian stratigraphy in the Lake Superior region to the north. As a result of this work they came to be recognized as the "Wisconsin school" of structural geologists. The igneous rocks underlying the quartzite at Baraboo were assigned to the Archean, while the metasediments were correlated with the "Huronian", now "Animikean", of the Lake Superior region (Van Hise and C. K. Leith, 1911).

Numerous University of Wisconsin theses have been written on the structure of the Baraboo Quartzite, particularly the western end. These have proved particularly valuable in the present study for locating critical outcrops, but most contain no information on mesoscopic[1] structures other than bedding. An exception is the exhaustive study by Adair (1956) of the so-called "anomalous" structures of the syncline (secondary phase structures of this paper). Schmidt (1951) compiled a valuable synthesis of the subsurface data. Other theses include those of Damm and Mees (1943) on the northwestern end of the syncline, Gates (1942) on the Baxter Hollow granite, Griesell (1937) on the west end of the South Range, Mayer (1934), Wenberg (1936) and Kemmer and Kovac (1937) on the West Range, and C. K. Leith (1941) on the South Range. Ostenso (1953) and Hinze (1957) carried out magnetic and gravity surveys respectively.

An extremely valuable structural analysis of the Baraboo Quartzite, mainly on the microscopic scale, was carried out by Riley (1947). The extensive data obtained during his study of the deformed quartz grains has been valuable in the interpretation of subsequent experimental work (e.g. Raleigh and Christie, 1963; Friedman, 1964; Carter and Friedman, 1965).

The most recent published work on the Precambrian rocks of the area prior to the present study is that of Hendrix and Schaiowitz (1964), who studied the geometry and field relations of mesoscopic structures in phyllitic layers near the top of the quartzite exposed on the south limb of the syncline.

Paleozoic rocks also were observed, of course, by the earliest workers, but they received less attention than the Precambrian. Irving and Weidman recognized the profound unconformity between the two sequences and noted the spectacular coarse basal Paleozoic conglomerates. Weidman had a clear appreciation of profound mountain building and deep erosion prior to Cambrian deposition. He noted that the most resistant Precambrian rocks in Wisconsin remained as islands or monadnocks, and estimated that the Baraboo monadnocks must have been from 1000 to 1600 feet above the surrounding plain. He also recognized the unusually close control of the late Precambrian-Cambrian topography upon that of the present day. The general stratigraphy and several fossil localities became widely known among Upper Mississippi Valley geologists

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1. Mesoscopic structures are those on hand specimen or single outcrop scales.

during the first quarter of the century. Stratigraphic names borrowed from eastern states, such as Potsdam Sandstone and Trenton Limestone, gradually were displaced by newer Mississippi Valley terminology after 1920 (See Table 5). At the same time, the district was being used for field trips by many schools, and it figured in the great stratigraphic controversy promulgated by E. O. Ulrich. Based upon his interpretation of some stratigraphic relationships around Baraboo and Madison, as well as in Missouri, Ulrich proposed an entire new Ozarkian System between the Cambrian and Ordovician. A bitter controversy ensued between Ulrich and other stratigraphers such as Twenhofel. By 1935, the Ozarkian was discredited by the demonstration of several mis-correlations of lower Paleozoic formations, which had resulted from erroneous age assignments to certain critical faunal assemblages, including some from the Baraboo district.

By 1930, geologists who visited the area frequently, such as W. H. Twenhofel, G. O. Raasch, F. T. Thwaites, A. C. Trowbridge, and J. H. Bretz, were well versed in the stratigraphy around Baraboo, but this knowledge was not generally available until J. M. Wanenmacher mapped and described the Paleozoic formations in detail for a Ph.D. dissertation in 1932. Published stratigraphic syntheses for the area appeared soon thereafter (Wanenmacher, et al., 1934; Raasch, 1935). Two Bachelor's theses done at the University of Wisconsin (H. F. Nelson, 1940; Oetking, 1943) dealt with the Paleozoic rocks a few miles south of the syncline, and a recent Master's thesis (Usbug, 1968) treated the northwestern portion of the syncline. With extensive checking and additional mapping by Dott, maps included in these reports, that of Wanenmacher, three unpublished maps prepared as student independent study projects at the University in Madison by Douglas Hackbarth (southeastern Baraboo hills), Dennis Howe (southwestern Baraboo hills), and Sharon S. Kreutzman (eastern half of syncline), and old unpublished field maps in the files of the Wisconsin Survey were used in preparing the accompanying geologic map (Pl. I). Recent revisions of stratigraphic nomenclature proposed by the Wisconsin Geological and Natural History Survey (Ostrom, 1967) are adopted herein.

The origin of the Paleozoic sediments of the Baraboo area has received far less attention than has the general stratigraphy, a circumstance typical of the entire Mississippi Valley region. Comments in Wanenmacher, et al. (1934), Raasch (1958), and Farkas (1960) are practically the only specific references, and they are very general. On a regional scale, however, Ostrom (1964) and Raasch and Unfer (1964) interpret Cambro-Ordovician cyclic patterns of sedimentation that should have affected sedimentation in the Baraboo area; Hamblin (1961) and Emrich (1966) have provided regional paleocurrent data. Because the stratigraphy is now well known, the sedimentology is emphasized in this monograph, for the Baraboo area offers unusual opportunities for detailed analysis of ancient sedimentary environments and paleogeography. Results of recent sedimentologic investigations provide new data for interpretations presented here.

The Pleistocene geology of the Baraboo area has received almost as much attention as have the Precambrian rocks. The Devils Lake area, especially, provides unusually clear features easily comprehended by the neophyte, thus it has been for years a favorite for student trips; the origin of the lake is so readily demonstrable as to delight even the most casual tourist. Besides the obvious terminal moraine looping around the Devils Lake valley and thence northwestward through the west edge of Baraboo, striated rock surfaces, outwash deposits, and glacial lake sediments are well displayed in the district. The topographic contrast between the recently-glaciated eastern half and the western portion of the syncline region is indeed striking, and was noted by early workers (e.g., Weidman). This contrast, of which the Baraboo area provides a sample, is of regional extent, and long ago led to the concept of the famous Driftless Area of southwestern Wisconsin.

Comprehensive reports on geomorphology and Pleistocene deposits include early ones by Salisbury and Atwood (1897; 1900) and Alden (1918), while later ones were by Martin (1932), Smith (1937), Thwaites (1935; 1958), and Powers (1960). More specific local reports have dealt with Devils Lake (Trowbridge, 1917; Twenhofel and McKelvey, 1939; and Black, 1967, 1968) and with glacial Lake Merrimac, which surrounded the eastern end of the district (Bretz, 1950). Black's paper is especially designed to lead either the layman or professional to specific features that bear on the late Pleistocene history of the Devils Lake area. It provides a valuable supplement to the brief discussion of Pleistocene geology in this guidebook.

Economic Geology

The Baraboo quartzite was used originally for macadam aggregate and paving blocks. It was later quarried extensively for metallurgical (refractory) purposes. Now the main use is for railroad ballast, with some also being used for grinding pebbles.

Extensive accounts of the history of iron mining in the Baraboo district are available in Weidman (1904), Van Hise and C. K. Leith (1911), and Schmidt (1951). T. C. Chamberlin is generally credited with first recognizing the possibility of there being iron ore in the Baraboo district in 1882. He based his suggestion on the presence of iron minerals in quartz veins. Iron formation was discovered in Freedom Township (T.11N., R.5E.) in 1887 and during the period 1889-1899 it was mined just west of La Rue for paint pigment by the Chicago and Northwestern Railway Company. It was in April 1900 that W. G. La Rue struck ore-grade iron on the same property.

Mining was confined to the area south of Baraboo and North Freedom. The locations of the Illinois, Sauk and Cahoon mines, from which all the ore was shipped, are shown on the geologic map (Pl. I). The ore was mined from the lower member of the Freedom Formation, which is at least 400-500 feet thick in places (Pl. II). It was mostly red hematite with a small amount of limonite. The average iron content in 1,517 analyses of samples from the Illinois mine was 36.4% (Van Hise and C. K. Leith, 1911). At the end of 1909 only 0.06% of all the iron ore obtained from Precambrian sediments in the Lake Superior region had come from the Baraboo district, but it was estimated that there were reserves of 910 billion tons of ore with 35% or more of iron (Van Hise and C. K. Leith, 1911, p. 461 and 492). The Illinois mine operated from 1904-1908, then briefly from stockpiles in 1916, finally closing down the same year after producing 315,350 tons of ore. The Cahoon mine operated from 1916-1925 (except 1919 and 1921) and produced 327,683 tons. Records from the Sauk mine are not available. Flooding apparently was a major factor in causing the closure of all three mines (Schmidt, 1951). Unfortunately, Weidman's belief (1904, p. 152) that the district would become as important as some of the areas around Lake Superior was not fulfilled.

PRECAMBRIAN GEOLOGY

Relationships and Ages of the Precambrian Rocks

(Dalziel and Dott)

The determination of the age of the Baraboo Quartzite and associated rocks is conjectural because of isolation from the Great Lakes Precambrian rocks, wherein the standard time scale has been established for North America. Hall (1862) suggested a correlation with the Huronian of Ontario. Because of the presence of banded iron ores closely associated with dolomite, and all underlain by quartzite, correlation with Middle Precambrian (Animikean) rocks on the north and south sides of Lake Superior has been generally assumed ever since. But this type of correlation is tenuous at best, hence we judged it desirable to attempt to bracket the Baraboo sequence by establishing isotopically the ages both of slightly older and slightly younger rocks.

Relative age relations of the plutonic rocks around Denzer (southwest side of the syncline, Pl. I) to the sediments are ambiguous, so these rocks are of little value for establishing an older age limit as is borne out by the large scatter in Rb-Sr analyses of the Baxter Hollow granite (Table 2). The rhyolite complex offers more hope, for it underlies the Baraboo Quartzite concordantly on both limbs of the syncline, and nowhere shows offshoots penetrating the quartzite. Local rhyolite masses alleged to overlie the quartzite are, in our opinion, glacial erratics. The rhyolite has flow-banding and fragmental textures exactly like those characteristic of extrusive tuff breccias and rhyolitic glasses as Weidman noted many years ago (1895), and Stark more recently (1930; 1932). The rhyolites at Baraboo are strikingly similar to widely scattered dark rhyolites in central Wisconsin. Asquith (1964) studied the latter and found that several had microscopic shard and axiolitic structures characteristic of welded tuffs (ignimbrites). He did not detect such structures in the more deformed Baraboo rhyolites, but we have subsequently found clear axiolitic and faint shard-like relicts in a number of thin sections. Together with megascopic textures, these features prove not only a volcanic origin of the rhyolite around Baraboo, but probably a welded tuff origin for much of it.

Several workers (following Weidman, 1904) have claimed that small rhyolite pebbles occur in the basal Baraboo Quartzite, but we have been unable to confirm this. What Weidman referred to in 1904 as a basal rhyolite-bearing conglomerate is in reality the breccia that he earlier had interpreted correctly as of volcanic origin and belonging to the older rhyolite complex (see Stark, 1932). Most of the small, dark pebbles in the Baraboo appear to be jasper, although in fine particles without obvious phenocrysts these can be impossible to distinguish from red, devitrified rhyolitic glass fragments.

In spite of the lack of proven rhyolite pebbles in the quartzite, there was no doubt in our minds that the rhyolite is older. Therefore, we had Rb-Sr analyses performed for five specimens of rhyolite from three localities around the Baraboo syncline. The results (Table 2) allowed an isochron to be drawn whose slope indicated an age of 1.54 ($\pm$ 0.04) b.y. The similar appearing rhyolites of central Wisconsin (Fig. 1) have yielded dates of 1.30 to 1.60 (Goldich, et al., 1966), a range too large either to prove or disprove a direct age relation.

TABLE 2: Rubidium-strontium and potassium-argon data used in age calculations.

RUBIDIUM-STRONTIUM ANALYSES (by M. Halpern, The University of Texas, Dallas; September, 1968):

Whole Rock Sample No.	Sr^{87}/Sr^{86}*	Rb^{87} (µM/gm)	Sr^{86} (µM/gm)	Rb^{87}/Sr^{86}	Age** (10^9 yr)	Sr^{87}/Sr^{86}
Volcanic Rocks:						
U-4	0.7580	0.214	0.0891	2.40		
BA-1	0.7596	0.421	0.161	2.61		
U-1(A)	0.7605	0.228	0.0941	2.42	1.54 ± 0.04	0.705 ± -.005
U-1(B)	0.7607	0.227	0.0942	2.41		
U-6	0.7831	0.252	0.0760	3.32	(isochron)	(isochron)
U-2(A)	0.9609	0.367	0.0321	11.4		
U-2(B)	0.9616	0.368	0.0322	11.4		
Baxter Hollow Granite:						
68-6A(A)	0.7412	0.388	0.230	1.69	1.48 ± 0.12	0.705**
(B)	0.7414	0.381	0.227	1.68		
68-6B(A)	0.7376	0.310	0.208	1.49	1.53 ± 0.14	0.705**
(B)	0.7381	0.306	0.209	1.46		
68-6C(A)	0.7405	0.266	0.165	1.62	1.50 ± 0.13	0.705**
(B)	0.7403	0.267	0.165	1.62		

POTASSIUM-ARGON ANALYSES (by Geochron Laboratories, March, 1968):

Whole Rock Sample No.	Ar^{40+},ppm	Ar^{40+}/Total Ar^{40}	%K	K^{40},ppm	Apparent Age***
Phyllite in Baraboo Quartzite:					
US-12(A)	0.00603	0.276	0.091	0.113	760 (±50) x 10^6yrs.
(B)	0.00638	0.290	0.095		

*Normalized to Sr^{86}/Sr^{88} ratio of 0.1194. At the time of these analyses, the normalized Sr^{87}/Sr^{86} ratio of the Massachusetts Institute of Technology standard $SrCO_3$ (Lot# 492327) was measured as 0.7080 ± 0.00026 (mean of 7 analyses).

**Ages calculated using $\lambda_\beta = 1.47 \times 10^{-11}$ yr. $^{-1}$. Using $\lambda_\beta = 1.39 \times 10^{-11}$ yr. $^{-1}$, the calculated ages would increase by about 6%.
Ages for Baxter Hollow Granite have been calculated assuming a Sr^{87}/Sr^{86} initial ratio of 0.705. The plus and minus value assigned to each calculated age is the sum of the differences between duplicate dissolutions of each whole rock sample and the age differences assuming Sr^{87}/Sr^{86} initial ratios of 0.702 and 0.708. (See Goldich, et al., 1962) for Sr^{87}/Sr^{86} initial ratios used for rocks of similar age and geologic province.)

Ar^{40+} is radiogenic Ar^{40}

***Constants used: $\lambda_\beta = 4.72 \times 10^{-10}$/year; $\lambda_e = 0.585 \times 10^{-10}$/year
$K^{40}/K = 1.22 \times 10^{-4}$ g./g.

$$AGE = \frac{1}{\lambda_e+\lambda_\beta}\left[\ln \frac{\lambda_e+\lambda_\beta}{\lambda_e} \times \frac{Ar^{40+}}{K^{40}} + 1\right]$$

A younger age limit for Baraboo Precambrian rocks first was approached by attempting to date micaceous minerals in phyllite zones within the quartzite by the K-Ar method, which might pinpoint the date of later metamorphism. Results (Table 2) indicated a date seemingly too young (760 ± 50 m.y.) to relate to the known tectonic history of the Great Lakes region, thus it is a minimal age at best. We attach no significance to it because a great excess of K-free pyrophyllite over K-bearing muscovite in the phyllites made the K percentage so small that the analyses were done to the limits of sensitivity of the methods employed.

As an alternative, we turned to the identical appearing Waterloo quartzite, which is exposed in another but smaller inlier 25 miles east of Madison (Fig. 1). Here previous dating of a pegmatite dike cutting the quartzite (1.44 b.y.; Aldrich, et al., 1959) and of a muscovite-rich phyllite zone within the quartzite (1.41 b.y.; Goldich, et al., 1966) seems to limit quartzite deposition to some time prior to 1.4 b.y. ago. By extrapolation between Waterloo and Baraboo it appears that deposition occurred between about 1.45 and 1.5 b.y. ago, followed by deformation that produced the fold structures and metamorphism about 1.4-1.45 b.y. ago.

If our admittedly circuitous interpretations are correct, and if there has been no wholesale re-setting of whole-rock isotopic clocks, then the Baraboo-Waterloo Quartzite masses are younger than both the Middle Precambrian Animikean and Huronian rocks with which they have long been correlated. They even appear to postdate the well-known Penokean orogeny (1.6-1.9 b.y. ago), which affected at least the northern half of Wisconsin. Thus one could read some suggestion from our results of a post-Penokean orogenic event (circa 1.4-1.5; roughly Elsonian in the Canadian Survey or Mazatzal in western U. S. terminology) that has been suggested by several workers, but which is not yet fully accepted as valid (see Goldich, et al., 1966).

Rb-Sr analyses of the long-enigmatic Baxter Hollow granite northeast of Denzer suggest that it will remain enigmatic somewhat longer. It has been argued that the granite is younger than the quartzite (Gates, 1942), but this is not certain, and Stark (1932) believed it to be older. The two are separated by a narrow zone of sheared rock, rendering their relations rather ambiguous. Rb-Sr ratios were so similar in all samples of the granite analyzed that an isochron could not be constructed. Calculation of dates for individual samples indicates a total possible spread of values so large (1.36-1.67 b.y.) that the rock could as well be related to the post-Baraboo orogenic event or to the pre-Baraboo rhyolitic volcanism. Obviously "further study is indicated."

Petrology

(Dalziel)

Only a brief description of the various Precambrian rock types will be presented here. More extensive accounts of the igneous rocks and the Baraboo Quartzite are given by Weidman (1904), Stark (1932) and Gates (1942), and of the overlying Precambrian metasediments by Weidman (1904) and A. Leith (1935). The following descriptions are partly condensed from their articles.

Rhyolite. Much of the rhyolite is fine-grained, dark gray or red in color, crudely banded, and contains small quartz and feldspar phenocrysts. At some localities, however, it has the appearance of a volcanic breccia, and local tuffaceous sandstones also occur (Stark, 1930). In others compositional layering (presumably flow banding) is well developed. It shows relict microscopic textures typical of welded tuffs. A number of structural surfaces, some marked by the alignment of deformed phenocrysts, can be observed.

Diorite. Two small outcrops of diorite occur near Denzer (Pl. I). No field relations can be seen. The rock is massive, medium-grained, and reddish in color. It consists largely of plagioclase (70-85%), hornblende (10-20%) and quartz; mica, apatite and iron oxide are minor constituents, and some of the hornblende is altered to chlorite.

Granite. A few outcrops of granite are located on the south side of the south range in Baxter Hollow on the main branch of Otter Creek (Pl. I). The rock is medium to fine-grained and consists of plagioclase feldspar (50-75%), quartz 20-45%), mica, hornblende and apatite. The granite is sheared adjacent to the Baraboo quartzite, and jointing is common in all outcrops.

Baraboo Quartzite. The quartzite is a massive, vitreous rock, typically pink, maroon or purple, but locally white or gray. It is comprised of more than 80 percent quartz, which occurs as medium to coarse sand-size grains and sporadic rounded granules and fine pebbles. There is no pronounced basal conglomerate; rather, fine pebble bands and lenses are common throughout. The pebbles are predominantly white quartz, and rarely exceed one inch in diameter. In most outcrops, bedding parting parallel to pebble layers or argillaceous zones is pronounced, but in low, glaciated outcrops it may be obscure. Where such surfaces are well exposed, excellent ripple marks may be seen. Current bedding is almost universal. An intricate color banding is confusing. In some cases it follows bedding or current bedding surfaces, but commonly it is independent of both (See Weidman, 1904, p. 24) and forms circular, elliptical, and irregular patterns on planar joint surfaces (Fig. 27c; STOP 5). Joints are ubiquitous and mostly steeply dipping.

Numerous layers or lenses of more argillaceous material occur within the quartzite. Most are only a few inches thick, but some reach several feet. They have been referred to in the literature variously as argillite, slate, schist and quartz-schist. X-ray diffraction studies of specimens from a number of localities have been carried out by S. W. Bailey of the University of Wisconsin-Madison. He reports that they chiefly consist of pyrophyllite and quartz with minor quantities of muscovite and hematite (personal communication). They will be referred to here as phyllitic-quartzite, quartz-phyllite or phyllite because they exhibit more pronounced recrystallization and less regular foliation than a slate, but are finer grained than a schist.

The quartz-pyrophyllite-(muscovite)-(hematite) assemblage of the phyllitic layers indicates that the Baraboo Quartzite reached the lower greenschist facies of regional metamorphism. Some idea of the temperature involved can be obtained from experimental data. According to Kerrick (1968) the upper stability limit of pyrophyllite is 410-430°C at 2-4 lb. water pressure. It is interesting to note that Weidman (1904, p. 49) mentions that andalusite is "probably" present in the Seeley Slate. If this could be confirmed it would mean that the metamorphism reached the upper greenschist facies, and the temperature must have been higher than that of the upper stability field of pyrophyllite, at least locally. Weidman also mentions the presence of "kaolinite" in the Precambrian metasediments. Although this could, if confirmed, indicate lower temperatures, it should be borne in mind that the term was used very generally in Weidman's time; kaolinite could be hydrothermal in origin (see Bailey and Tyler, 1960 and discussion of breccia zone in Upper Narrows at Baraboo, STOP 2.) An unusual nickeliferous phosphate mineral, lazulite, occurs in veinlets southeast of Devils Lake (Olsen, 1962).

At a number of localities on the south limb of the syncline (e.g., northeast entrance to Devils Lake State Park, Skillet Creek, Highway 12 and Happy Hill School; STOPS 5 and 6, Supplementary Stops G and H respectively) there are outcrops of phyllite up to 10 feet thick containing thin quartzite beds. As these localities all occur at a high stratigraphic level in the Baraboo Quartzite and are roughly on strike with one another (Pl. III), they probably represent the same phyllitic zone.

A breccia consisting of relatively coarse, angular quartzite fragments in a quartz matrix containing dickite forms a prominent zone 150 yards wide on the west side of the Upper Narrows (Ableman's Gorge; STOP 2) in the North Range, and has figured extensively in the literature (e.g., Weidman, 1904, p. 25). A few narrower brecciated zones occur elsewhere, some containing kaolinite in the matrix (S.W. Bailey, personal communication).

Seeley Slate. The slate is described (Weidman, 1904; A. Leith, 1935) as gray or green in color, consisting of alternating bands varying slightly in texture and color, and being uniform in its appearance. A fine stratification and a well developed cleavage are both commonly present.

Freedom Formation. Frequently referred to as the Freedom Dolomite, this formation consists of a number of different rock types, slate, chert and iron ore in addition to dolomite (Weidman, 1904). Minerals present in varying proportions are quartz, dolomite, hematite and clay minerals. Dolomite forms the upper member and various types of ferruginous rocks form the lower member. A ferruginous slate is present at the base representing a transition from the Seeley Slate. Compositional banding is prominent throughout the formation; secondary cleavage is poorly developed, but filled fractures are common.

Dake Quartzite. This is reported from drill cores to be a coarse-grained quartzite with a high proportion of chlorite and sericite as a matrix (A. Leith, 1935). Pebbles, sometimes angular, are common, and consist only of quartz and quartzite. The matrix of the rock is ferruginous near the base.

Rowley Creek Slate is a gray quartz-chlorite-sericite slate. Bedding and cleavage are present and the rock oxidizes to a reddish color on these surfaces. (A. Leith, 1935).

Sedimentology of the Baraboo Quartzite

(Dott)

Composition and Texture. Besides quartz grains (0.25-2.0 mm) the Baraboo contains zircon, rutile, iron oxides, chlorite, pyrophyllite and sericite as accessories (see Tables 6, 7 for compositional data). Most quartz sand grains in the formation have very thin films of iron oxide (chiefly hematite), which gives rise to the characteristic reddish color. Conglomerates consist of more than 90 percent clear to milky, well-rounded and well-sorted quartz granules and pebbles, about 5 percent dark red jasper, and a few percent of other dark lithic fragments, which tend to be more angular. The maximum pebble size known is about 3 centimeters, but most are less than 1.5 cm. Approximate proportions of the original lithic types within the formation are: 85% well-sorted sand; 5% argillaceous fine sand and silt; and 10% conglomerate. Finest units occur chiefly in the upper Baraboo, whereas conglomerate is rather uniformly distributed in lenticular units a few inches to a foot thick.

<u>Sedimentary Structures</u>. Stratification is discernible in most Baraboo outcrops, although either secondary color banding or shearing obscures it in some cases. Over most of the syncline "master" or true bedding is revealed by essentially parallel parting planes assumed to have been originally horizontal; the majority of beds are from 6 to 12 inches thick. Asymmetrical ripple marks occur widely on the bedding surfaces, but are especially well displayed in quarries along the north (vertical) limb of the syncline.

Cross stratification and associated small scour features, reflected by slight variations of color and texture, are prominent internal sedimentary structures. Individual inclined laminae average about one-fourth inch in thickness. Roughly 60 percent of the exposed cross-stratified units are bounded by parallel, planar truncation surfaces, here termed <u>master stratification</u> or <u>master bedding</u> (generally abbreviated simply to bedding). The amplitude of most inclined sets of cross laminae is from 6 to 12 inches, while small-scale cross laminae with amplitudes on the order of one or two inches occur in some fine-grained units. The remainder of the strata show inclined truncation surfaces that produce wedge-shaped bodies with amplitudes of up to three or four feet. In the latter cases, the master bedding may be impossible to detect, making interpretation of structural measurements somewhat hazardous. Conversely, structurally-induced cleavages may be easily mistaken for cross bedding in certain cases. Typical cross stratification is most easily studied at the U. S. Highway 12 locality (Supplementary Stop G) the Upper Narrows (Stop 2), and in the bluffs at Devils Lake (STOP 5).

Brett (1955) described the cross bedding as mostly straight in cross section, thus with only slightly tangential bases (formerly known as the "torrential type"). Common lack of a clearly tangential base makes top-bottom determinations difficult in many outcrops, but classic, curved, tangential bases also are well displayed. Brett noted that very rare foresets are as much as 7 to 10 feet long, these being low-angle types. In general, the foresets are much shorter and show angles of inclination from about 12 to as much as 54 degrees. Repose angles greater than 35^{o} in well-sorted and rounded sandstones have been steepened structurally after deposition. Contortions clearly formed during or just after deposition are characterized by disharmonic folding of laminae in patterns geometrically unrelated to the syncline. In other cases, however, the origin of overturning of the tops of cross sets is less obvious. Brett (1955) and Pettijohn (1957) have noted that cross sets should be flattened on the north limb of the syncline and steepened on the south by differential slip between strata during folding; Brett found a 5-degree difference between average inclinations on the two limbs. This is oversimplifying the case and, needless to say, careful structural observations are required to separate cases of synsedimentary deformation from later structural deformation (see p. 16).

Because of the induration of the quartzite, cross stratification is rarely seen in three-dimensions except on talus blocks. This fact, coupled with a limited understanding of cross stratification prior to 1960, naturally led to an assumption that Baraboo cross bedding is practically all of the "textbook type", namely, simple uniformly inclined planes (see Fig. 10). But close examination in favorable outcrops (e.g., U. S. 12 roadcut, Supplementary Stop G) reveals that trough or scoop-shaped cross sets are not only common, but may be the dominant form, as is typical of many sandstones. Trough shapes are especially prominent where truncation surfaces are non-parallel, producing the well-known wedge-trough or "festoon" stratification (Fig. 10).

Paleocurrent analysis by Brett showed a pronounced average preferred dip of cross strata toward the south-southeast (Fig. 2). Apparently Brett did not recognize the possible importance of trough-type sets, which are difficult to measure, and his number of readings per locality was small (at only 5 of 45 localities were 10 or more readings reported). For simplicity of restoration to horizontality, he also assumed that folding had not rotated the foresets in the plan-view sense. In view of the fact that the strike of the cross-sets and the fold axis are near parallel, this seems reasonable. In spite of these possible shortcomings, Brett was able to verify the apparent mean foreset inclination direction with 21 asymmetrical ripple orientations; there was only a 22^{o} difference between their mean orientations, and locally the agreement was within 5 degrees. Qualitative observation by ourselves at many localities leaves no doubt of a general north-to-south average current flow during Baraboo deposition as reported by Brett.

<u>Sedimentary and Tectonic Environments</u>. We have no reason to question the long-assumed aqueous deposition of the Precambrian sediments, although criteria are almost totally lacking. Brett (1955) states that the <u>ripple-mark index</u> (amplitude divided by wave length) is characteristic for water deposition, but structural modifications preclude the use of cross-lamination and ripple inclination for environmental determination (see Fig. 33b, STOP 7). As usual, one must argue by indirect analogy with other strata whose environments are well established, which leads to the conclusion that the entire Precambrian sequence was deposited in shallow marine water. During Baraboo deposition vigorous current agitation prevailed, as attested by texture and stratification. Mud, carbonate, iron and silica deposition followed but presumably still in relatively shallow marine water.

The great purity and volume of preserved Baraboo sandstones is indeed impressive. For example, the quartzite is 40 or 50 times thicker than the mineralogically similar Ordovician St. Peter Sandstone. The contrast is even more staggering, however, when we consider that both the Baraboo and Waterloo Quartzite are but local remnants of what was presumably a much larger original body of sandstone! Such pure sandstones generally have been considered most characteristic of stable cratonic regions, but the Baraboo apparently formed within a mobile tectonic belt, or at least was caught up in such a belt soon after deposition. Together with several other pure quartzites of Middle and Late Precambrian ages in the Great Lakes region, which also occur within mobile belts, it constitutes inescapable evidence of a long and complex history of repeated weathering and depositional episodes. Furthermore, such voluminous ancient pure quartz sandstone attests to existence of still-larger volumes of quartz-bearing continental crustal parent rocks early in earth history (i.e., by 2.0 to 2.5 b.y. ago). To what extent the quartzites reflect peculiarities of Precambrian weathering due to the nature of the early atmosphere and a lack of land vegetation is not clear. The presence of very aluminous pyrophyllite suggests extremely thorough weathering of soils in source areas.

Presence of the but-slightly-older rhyolite complex, plutonic rocks, and deformation and metamorphism of the sedimentary sequence itself, all attest to the development of the Precambrian sequence within a mobile tectonic belt that extended east-west across Wisconsin. The great thickness of the strata, which apparently all formed in shallow-water environments, indicates profound subsidence within that belt. Based upon lithologic and thickness similarities, as well as recent isotopic dating, it appears that the Baraboo is part of a very large quartzite mass of late Middle or early Late Precambrian age (circa. 1.5 b.y. old) extending at least from South Dakota to Lake Michigan (see Fig. 1).

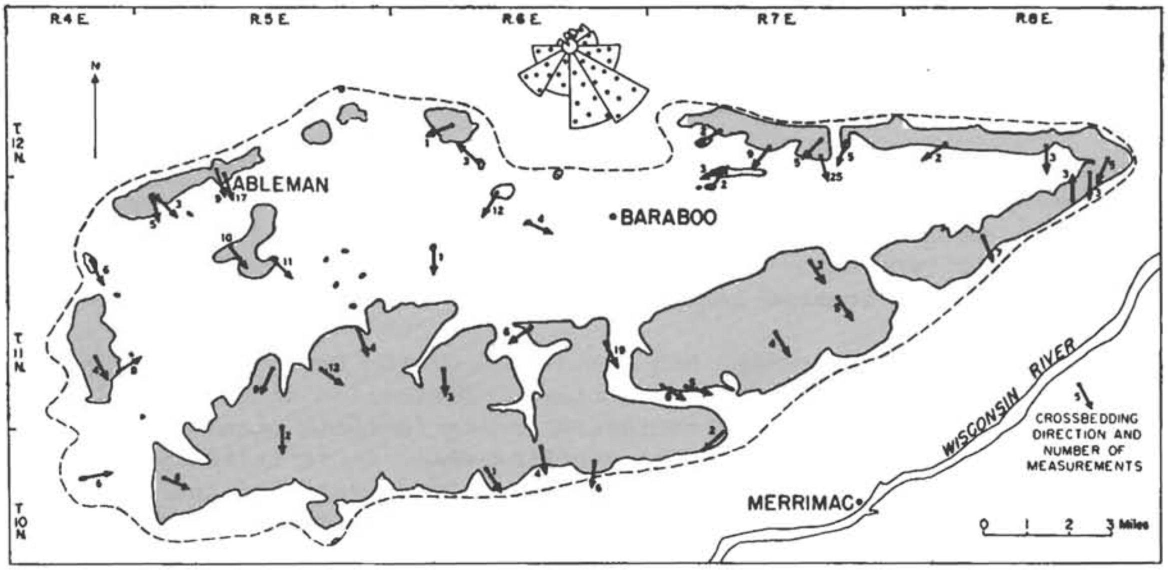

Figure 2. Restored dip orientations of cross stratification in the Baraboo Quartzite showing a north-to-south paleocurrent pattern. G. W. Brett, JOURNAL OF GEOLOGY, 1955, Fig. 1, from p. 145. Copyright, The University of Chicago Press. All rights reserved (published with permission).

Structural History

(Dalziel)

The Precambrian rocks of the Baraboo district were tectonically deformed in a Precambrian mobile belt. In common with the metasediments of all mobile belts investigated in detail to date, they show the effects of polyphase ductile deformation. Most of the mesoscopic structures observed can be ascribed to the effects of one main deformation episode during which the Baraboo syncline was formed. Other structures seen only in the phyllitic layers can be shown to have resulted from one, or possibly two, later but not necessarily unrelated phases. The structures are summarized in Table 3; in Table 4 they are correlated with those recorded by Riley (1947), Hendrix and Schaiowitz (1964), and earlier workers.

There is certainly no reason to suppose that the different sets of structures described here are the result of more than one orogeny. For example, in the Appalachian-Caledonian orogen, the Taconian and the Acadian orogenies in the Appalachians, and the first and second phases (early Ordovician and late Silurian) of the Caledonian orogeny in the British Isles, each resulted in similar sequences of structures (Dalziel, 1969b; Dewey, 1969a and 1969b). In fact, while the sequence of structural development outlined in Table 3 can be demonstrated clearly on the basis of criteria such as the deformation of a slaty cleavage by a later crenulation cleavage, there is good reason to believe that all the events distinguished were part of the progressive deformation of the Precambrian succession at Baraboo in response to one regional stress system. The various structures merely record certain stages within a continuous strain history.

Pre-tectonic structures. Master bedding (S_o) and current bedding in the quartzite were, to some extent at least, active elements during tectonic deformation for both types of surfaces are commonly slickensided. The importance of this type of deformation in the over-all strain of the Baraboo Quartzite is discussed later.

Oversteepening, overturning, and distortion of the cross-stratification is considered to be dominantly synsedimentary. Distortion is confined to discrete beds, and overturning of southward facing foresets occurs even in vertical beds on the north limb of the syncline where it is unlikely to have resulted from rotational strain during folding. Overturning of the foresets at the top of the beds on this limb is down dip to the south. The rotational strain must have been in the opposite direction, the higher beds moving up dip to the north out of the core of the syncline (see Fig. 3).

Preferred orientation of some pebbles in the quartzite may be primary. They lie with their longest and intermediate axes in the bedding or current bedding surfaces, and in some cases also have a crude, linear preferred orientation. Dimensional orientation of smaller quartz grains is probably largely tectonic.

Migration of solutions through the sediments to produce the complex color banding was post-depositional, but presumably predated metamorphism and deformation, as it is not structurally controlled (see Fig. 27c).

Main Phase Structures. Mesoscopic structures assigned to the main phase of deformation are the most common minor structures in the metasediments and are clearly related to the macroscopic[2] Baraboo syncline (Fig. 3). Cleavage and cleavage/bedding intersections are best developed and most widely recognizable.

There is a great deal of confusion over cleavage terminology generally, but the Baraboo area is associated in the minds of all structural geologists with classical studies of rock cleavage. The new names applied in this account have been selected in order to avoid genetic implications and do not conform to the classical terminology.

Early workers (Irving, 1877; Steidtman, 1910; C.K. Leith, 1913, 1923) recognized "strike joints" in the quartzite (near vertical on the south limb, near horizontal on the north limb), which were refracted into a north dipping "fracture cleavage" in the more phyllitic beds. In fact their "joints" are parallel to a more subtle, closely-spaced parting (S_1' on Tables 3 and 4; Figs. 3 and 4), which pervades the quartzite of the Baraboo Ranges. This is apparently controlled by tectonically-flattened quartz grains and aligned phyllosillicates at least in impure quartzite beds (Fig. 5B). Otherwise it is merely a closely spaced parting. The "fracture cleavage" referred to consists of very closely spaced, but nonetheless discrete, surfaces formed by concentrations of aligned phyllosilicates (S_1 on Tables 3 and 4; Figs. 3, 4, and 5a). Often slight displacements can be observed along these surfaces (see Fig. 27, and C. K. Leith, 1923, Fig. 54) which appear to deform (slightly crenulate) phyllosilicates forming a much more penetrative surface in the phyllitic beds, which is very faint in the field but readily visible in thin section (S_{1E} on Table 3, see Figs. 3, 4 and 5a).

Here the closely spaced parting in the quartzite (S_1') is termed simply the quartzite cleavage, and the "fracture cleavage" of C. K. Leith and others is called phyllitic cleavage (S_1). As Ramsay (1967, p. 180) points out, there can only be one true slaty cleavage in a tectonite (i.e. a completely penetrative cleavage, affecting all the platy minerals in the rock). In the Baraboo Quartzite this is the somewhat obscure S_{1E} surface formed by the scattered phyllosilicates in rather quartzose phyllites and deformed by slip on the more discrete surfaces of the (S_1) phyllitic cleavage.

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2. Macroscopic structures are those that occur on the scale of more than one outcrop. Hence their elucidation involves extrapolation from one outcrop to the next.
3. A "penetrative" structure is one which is closely spaced throughout the rock. The definition is therefore dependent on scale.

Table 3

MESOSCOPIC STRUCTURES IN THE BARABOO QUARTZITE

AGE	STRUCTURES			
Primary	S_o Bedding (master bedding) (also current bedding, and perhaps some preferred grain and pebble orientation)		Ripple marks ?Some preferred grain and pebble orientation	
Post-sedimentation, pretectonic	Color banding			
Early Main Phase	S_{1E} *Slaty cleavage (phyllite)	S_1' Quartzite cleavage (quartzite)	S_o/S_{1E} intersection (indistinct)	S_o/S_1' intersection
Main Phase	S_1 Phyllitic cleavage (phyllite), deforming S_{1E}		S_o/S_1 intersection	
	Axial surfaces of tight asymmetric minor folds (folding S_o)		Axes of tight asymmetric minor folds Longrain (mineral alignment) in S_1 (phyllite) Slickensides on S_o (quartzite) Boudin axes	transitional
Late Main Phase	S_{1L} Faint strain-slip or crenulation cleavage (deforming S_1 in phyllite)		Axes of fine crenulations (deforming S_1 in phyllite)	
Secondary Phase	S_2 Strain-slip or crenulation cleavage (conjugate; deforming S_o and, mainly, S_1 in phyllite) Axial surfaces of minor chevron folds (deforming S_o, S_1, and S_1')		Axes of crenulations (deforming S_o and, minly, S_1 in phyllite) Axes of minor chevron folds deforming S_o, S_1 and S_1')	
Late Phase	Axial surfaces of open minor folds (deforming S_o. S_1 and S_2)		Axes of open minor folds	

TABLE 3 (contd.)

Various	Joints, some quartz-filled; quartz "veins", including en echelon tension gashes in bands, some deformed gashes (probably post-second phase)	Slickensides on bedding, current bedding, fracture cleavage and joint surfaces
	Breccia zones and faults (?)	Boudin Axes
Doubtful	Local highly penetrative cleavage in quartzite	Intersection of highly penetrative cleavage in quartzite with bedding
	Second cleavage (closely spaced jointing) in quartzite	Intersection of second cleavage in quartzite with bedding
	Other cleavages (closely spaced jointing) in quartzite	Intersection of other cleavages in quartzite with bedding

* For definitions of cleavage terminology used here see text.

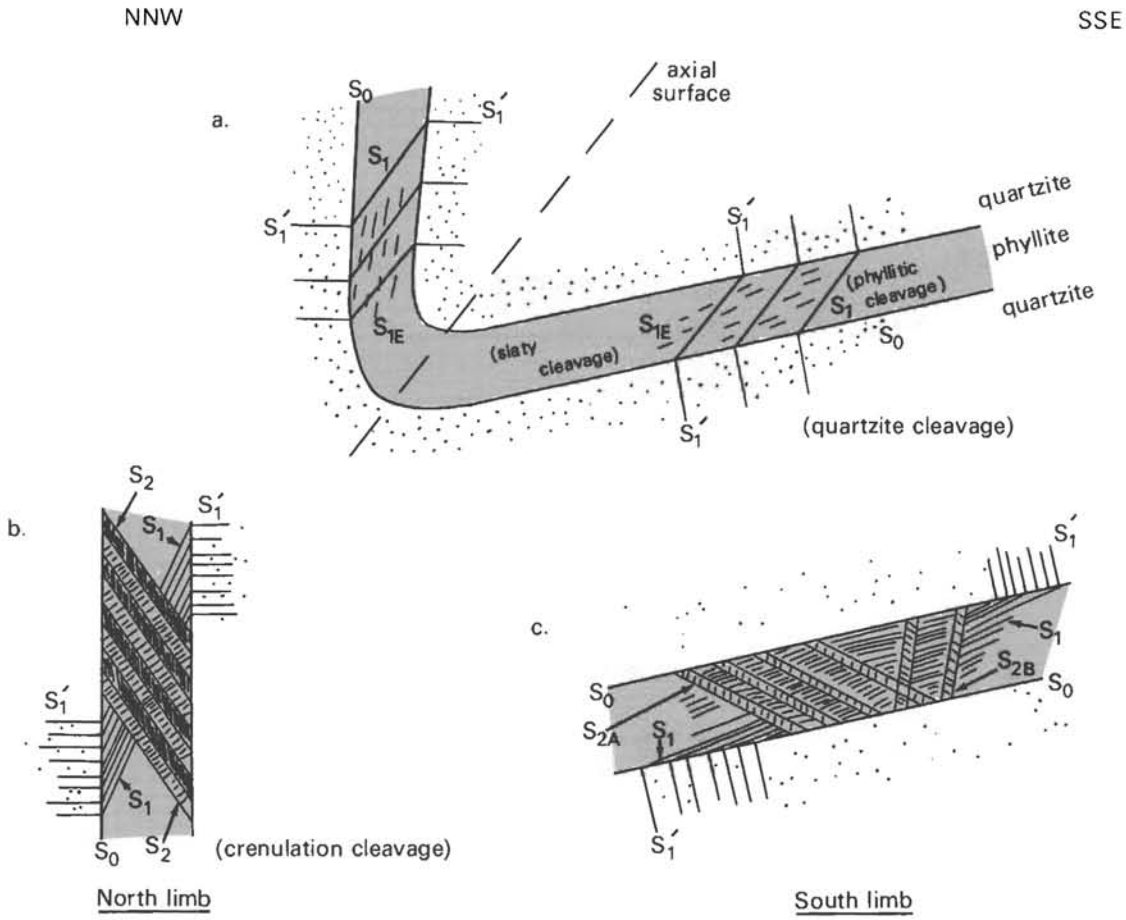

Figure 3. Diagrammatic cross-section of the Baraboo syncline showing the attitudes and relationships of cleavages.

Table 4

COMPARISON OF TERMINOLOGY USED FOR MESOSCOPIC STRUCTURES IN THE BARABOO QUARTZITE

Dalziel (this paper)	Riley (1947)	Hendrix & Schaiowitz (1964)
PRIMARY STRUCTURES		
(S_o) Bedding (master bedding)	Bedding and bedding plane foliation (principal foliation of south limb)	Bedding
EARLY MAIN PHASE STRUCTURES		
(S_{1E}) Slaty cleavage* (S_o/S_{1E}) Bedding/slaty cleavage intersection (indistinct)	---	---
MAIN PHASE STRUCTURES		
(S_1) Phyllitic cleavage**	Axial plane foliation (north limb only)	Normal rock or fracture cleavage
(S_1') Quartzite cleavage ***	Shear fracture	---
(S_o/S_1) Bedding phyllitic cleavage intersection	Bedding/axial plan foliation/ intersection	---
(S_o/S_1') Bedding/quartzite cleavage intersection	Bedding/shear fracture intersection	---
Axial surfaces of tight minor folds	---	Axial planes of normal drag folds
Longrain (mineral elongation lineation)	Grain elongation (?)	---
Slickensides on bedding	---	a lineation on bedding
Boudin axes	---	Boudin axes
LATE MAIN PHASE STRUCTURES		
(S_{1L} Faint strain-slip or crenulation cleavage	---	---
Axes of fine crenulations	---	---

TABLE 4 (contd.)

SECONDARY PHASE STRUCTURES (S_2) Strain-slip or crenulation cleavage	---	Reverse slip or rock cleavage; reverse shear zones
Axial surfaces of minor chevron folds	---	Axial planes of reverse drag folds
Axes of crenulations and minor chevron folds	Crests and troughs of chevron or buckling folds	Axes of reverse drag folds
LATE PHASE STRUCTURES Axial surfaces of open minor folds	---	---
Axes of open minor folds	---	---
STRUCTURES OF VARIOUS AGES Joints; quartz-filled joints	Joints; quartz-filled joints	Joints
Slickensides	Slickensides	Slickensides
Quartz "veins"; quartz filled tension gashes in bands, some deformed	Quartz veins; quartz filled ladder tension fractures; deformed and S-shaped quartz filled gash veins	Quartz filled joints
Breccia zones	Breccia zones	---
STRUCTURES OF DOUBTFUL AGE Cleavages in quartzite other than S_1'	Shear fractures (?)	---

* For definitions of cleavage terminology used here see text.

** "Fracture cleavage" of Van Hise, C. K. Leith and other geologists of the "Wisconsin school."

*** "Strike jointing" of geologists of the "Wisconsin school".

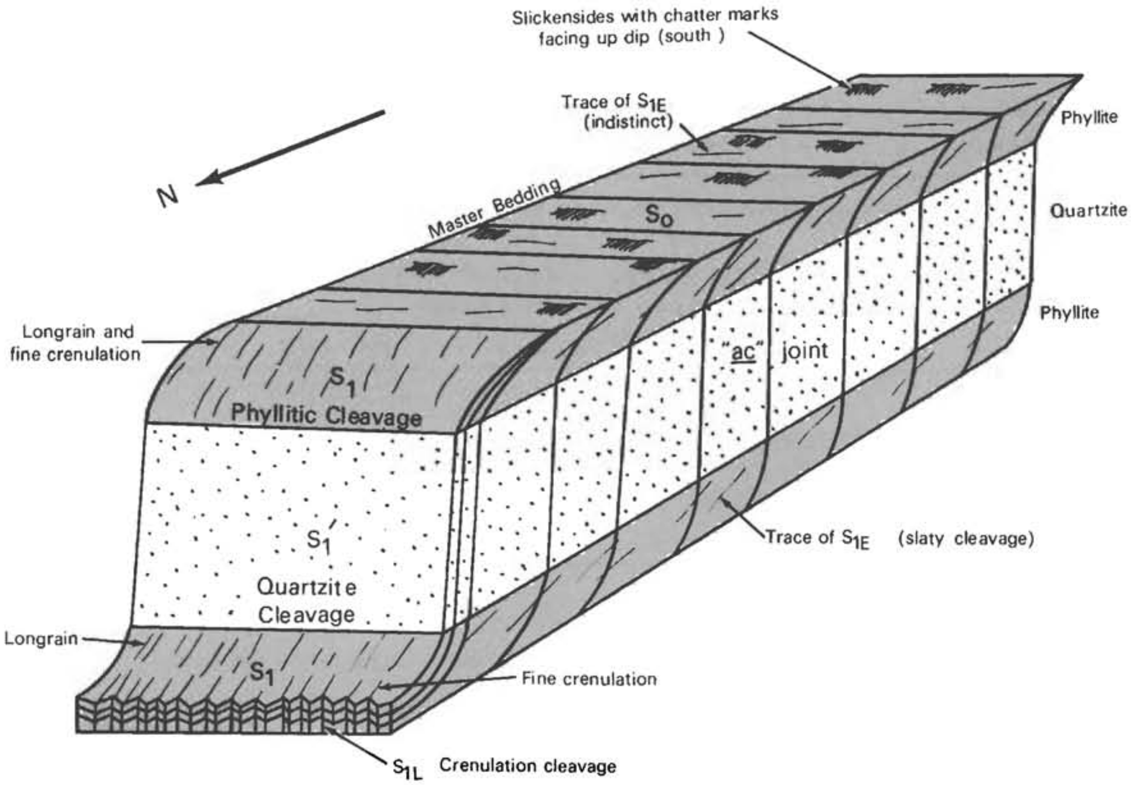

Figure 4. Block diagram showing the relations of main phase mesoscopic structures on the south limb of the Baraboo syncline.

It will be noted that the term "axial plane cleavage" has been avoided. This is because all three cleavages discussed up to this point are related to the formation of the Baraboo syncline but none are strictly axial planar (Fig. 3). The slaty cleavage (S_{1E}) converges on the axial surface, the phyllitic cleavage (S_1), though sometimes axial planar, is variable in attitude within a single bed, and quartzite cleavage (S_1') is variable over the syncline. It is everywhere nearly at right angles to the master bedding and hence axial planar only in the hinge zones.

On the phyllitic cleavage surfaces, a fine mineral alignment lineation can sometimes be discerned almost at right angles to the horizontal (east-northeast-west-southwest) S_o/S_1 intersection (Fig. 4). This is the structure known as "longrain" in slates. It is intimately related to the phyllitic cleavage as part of what Flinn (1965) has called an L-S system, the lineation being formed by the linear preferred orientation of the crystals that form the S_1 surface. It can be traced from thin phyllite beds, in which S_1 lies close to S_o, into slickensides on the bedding surfaces of the quartzite (Fig. 4). The genetic implications of these two types of lineation will be discussed later, but they are clearly related directly to the S_1 and S_1' cleavages. In some places a fine crenulation of the phyllosilicates lying in the main phyllitic cleavage (S_1) parallels the longrain and the slickensides on the bedding surfaces (Fig. 4). This can be observed, for example, at the northeast end of Devils Lake, the northwest entrance to the Devil's Lake State Park, and the east bank of Skillet Creek (STOP 5, Supplementary Stop F, and Stop 6, respectively).

The crenulation results in a faint, discontinuous strain-slip or crenulation[4] cleavage in phyllitic layers. It clearly postdates the formation of the (S_1) phyllitic cleavage, but nonetheless is intimately related geometrically to the longrain in S_1 and the slickensides on S_o, and can be genetically explained on this basis (see Fig. 7). No case was observed where the fine crenulation is at an angle to either of those linear structures. The associated strain-slip or crenulation cleavage is termed S_{1L} (Fig. 4).

All of the structural elements mentioned so far can be related directly to the formation of the Baraboo syncline and record particular stages in the progressive strain of the Baraboo Quartzite during its main phase of deformation. Three sets of structures resulting from this deformation phase can be recognized in the phyllitic layers (S_{1E}, S_1 and S_{1L}; see Table 3), but only one in the quartzite (S_1'). The quartzite cleavage (S_1') is refracted into the phyllitic cleavage (S_1), but it may have originated earlier, perhaps when S_{1E} was being formed in the phyllitic layers.

Mesoscopic folds resulting from the main phase of deformation are found at only two localities, both on the south limb of the syncline in the thick phyllite zone containing thin quartzite beds. At Skillet Creek (STOP 6, See Fig. 29b) and the northwest entrance to Devils Lake State Park (Supplementary Stop F, See Fig. 48), the phyllitic cleavage (S_1) is axial planar to folds with near-horizontal axes and a southerly vergence (direction of overturning). These are the "normal drag folds"[5] of Hendrix and

4. Strain-slip or crenulation cleavage is used here to describe a closely spaced parting marking an alignment of phyllosilicates resulting from the deformation of a pre-existing more penetrative structural surface (i.e., slaty or phyllitic cleavage.
5. Minor folds such as these most often originate from processes other than frictional "drag".

Schaiowitz (1964) in that they are congruent with the macroscopic geometry of the Baraboo syncline. The thin quartzite beds within the phyllitic layers have reacted in a much less ductile fashion than the phyllite. They form boudins with east-west axes crossed by tension gashes and fractures. Some of the boudins are deformed by secondary phase structures.

Secondary Phase Structures. Structures that postdate main phase structural elements, and are less obviously related to the formation of the Baraboo syncline, have been assigned to a secondary phase of deformation (see Table 3). At a number of localities on the south limb of the main syncline, and at one locality on the north limb, coarse east-northeast - west-southwest striking crenulation cleavages associated with chevron folds can be seen to deform the main cleavage of the phyllitic horizons (S_1) as well as bedding (see Fig. 3). Many of the chevron folds, which have near-horizontal axes, are overturned to the north on the south limb, hence they have long been called "reverse drag folds" (see Hendrix and Schaiowitz, 1964) as they are not congruent with the geometry of the syncline. It can be shown that there are conjugate crenulation cleavages on the south limb, dipping both north-northwest and south-southeast (Fig. 3).

While the main phase phyllitic cleavage can readily be traced into a cleavage in the quartzite, this is not true of the second phase crenulation cleavages. There are, however, a number of other structural surfaces in the quartzite at some localities, one or more of which may be equivalent to the crenulation cleavages in the phyllite.

Late Phase Structures. At two localities on the south limb (Skillet Creek, Stop 6, and the U. S. Highway 12 roadcut, supplementary Stop G) minor structures can be observed that appear to deform the second crenulation cleavages. These structures are rather open minor folds (Fig. 29e and 49d). They are rare, and their origin is rather problematical because of the complex geometry of the conjugate secondary crenulation cleavages (Fig. 49b, c and d).

Other Structures. Structures that result from brittle deformation of rocks are more difficult to relate in time than ductile structures. For instance, two joint surfaces merely intersecting one another most often yield no evidence of their age relations, but a crenulation cleavage can be deduced unequivocally to postdate a slaty cleavage because it deforms it.

Joints are common throughout the quartzite. Their attitude and tectonic significance will be discussed later. None are seen to be deformed so they are all likely to have developed late in (or after) the main phase of synclinal development.

Some of the joints are filled with recrystallized white or pale pink quartz. Numerous *en echelon* bands of quartz-filled tension gashes occur, and are locally conjugate. The gashes commonly have a sigmoid shape due to continued displacement along the bands. The (S_1') quartzite cleavage is deformed in the same sense (Fig. 54b). The gashes postdate the development of this cleavage, which cannot be traced through them (Fig. 54a).

Certain slickensides and boudins can clearly be related to the main phase of deformation (see above), but others cannot. Nearly all structural

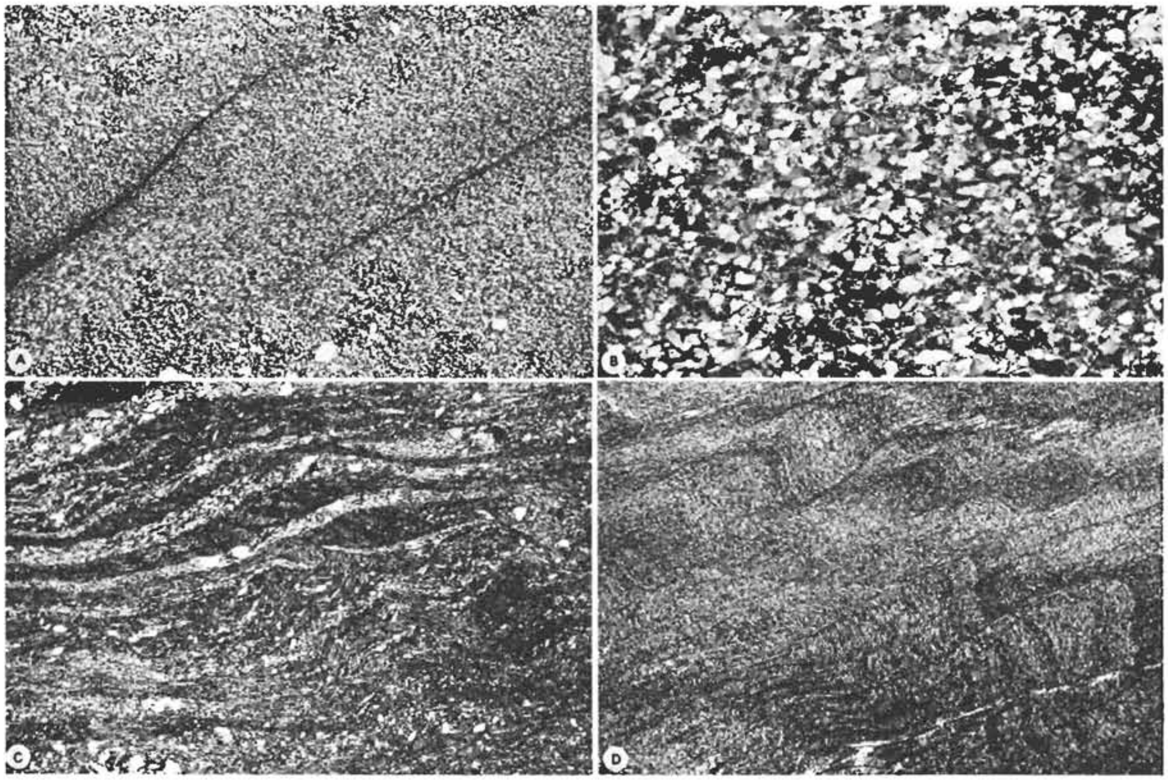

Figure 5. Photomicrographs showing the nature of cleavages in the Baraboo Quartzite.

A. Quartz-phyllite layer in quartzite at northeast corner of Devils Lake (STOP 5, see Figs. 3, 4 and 27a). Two discrete phyllitic cleavage (S_1) surfaces dip north (left on photograph) at a higher angle than the faint but much more penetrative slaty cleavage (S_{1E}) which they crenulate slightly. Plane polarized light.

B. Quartzite cleavage (S_1') in specimen from the quarry on the east side of the Upper Narrows (STOP 3). Alignment of phyllosilicate minerals and quartz grains in the cleavage is well developed, the quartzite being relatively impure. Crossed-nicols.

C. Bedding and phyllitic cleavage (almost parallel) deformed by secondary crenulation cleavage (S_2) which dips south (right on photograph). Phyllitic zone of the Baraboo Quartzite, U.S. Highway 12 roadcut (Supplementary Stop G, see Figs. 3 and 49a). Plane polarized light.

D. Phyllitic cleavage deformed by secondary strain-slip cleavage. Note complexity of strain-slip cleavage in detail and very faint conjugate surfaces. East side of Upper Narrows (Supplementary Stop B, see Fig. 3). Plane polarized light.

surfaces in the quartzite (master bedding, current bedding, jointing, S_1' cleavage) show slickensides. These may have formed at different times.

Finally, there are two types of structures whose relationship has not been resolved. At a number of localities near the east end of the syncline, a faint but persistent cleavage can be discerned in the quartzite that seems to be more penetrative than the S_1' cleavage, and may represent another stage in the progressive main phase deformation of the quartzite. Also, a number of partings less penetrative than S_1' can be observed at different localities. Some of these may be merely localized, closely-spaced jointing. Others may be related to the conjugate second-phase crenulation cleavages. Only in one area, the Upper Narrows (Figs. 24, 41, 42), does such a cleavage have a consistent attitude. It has not proved possible to determine the age relations of the surfaces.

Structural Geometry
(Dalziel)

General Structural Configuration. Comparison of Weidman's structural map (1904, Pl VI) with those accompanying the present account (Pls. III and IV) shows that the essentials of the structural geometry of the Baraboo syncline can be established from observations of dip and strike alone. Weidman was able to demonstrate the synclinal configuration of the quartzite by making observations all the way around the Ranges rather than just on the north and south synclinal limbs as Irving (1877), Chamberlin (1873), and Salisbury and Atwood (1903) had done.

The geometry of the syncline is further refined here not only by a still larger number of measurements of the attitude of the bedding in the quartzite, but also by the facing direction of the metasediments, mesoscopic structures, and subsurface data (Schmidt, 1951). This has resulted in the following changes in the picture previously presented:

1. The intermediate-scale folds northeast of Baraboo shown by Weidman (1904, Pl. VI), Wanenmacher (1932), and A. Leith (1935, Fig. 216), and later (following them) by Riley (1947, Fig. 14) and Hendrix and Schaiowitz (1964, Fig. 1), do not exist (see Pls. II and III).

2. On the south limb of the syncline west of Devils Lake, two additional open folds are recognized with axial surface traces parallel to those of the Pine Creek anticline and the adjacent syncline to the northwest mapped by Weidman (1904, p. 33) (see Pls. II and III).

3. The structure in the west is more complex than Weidman's data suggest. At the west end of the South Range, the bedding steepens to near vertical and swings around to strike nearly north-south (as shown by Riley, 1947, Fig. 14) while facing west. If there is a syncline between here and the West Range, it must be a very tight fold. Alternatively, a northeast-southwest-trending fault may be present between the South and West Ranges as suggested in a number of University of Wisconsin theses.

4. A fault just north of North Freedom recognized in the subsurface (Schmidt, 1951) has a downthrow to the south of at least 1200 feet, and confines the upper units of the Precambrian sequence to the immediate trough of the Baraboo syncline.

The general configuration of the Baraboo syncline and the geometry of the important mesoscopic structures can readily be understood from the geologic map and sections (Pls. I and II), the structural map (Pl. III), the map of main phase structures (Pl. IV), and the stereoplots of structural data. The area has been divided into four structural domains (Pl. III), comprising the north and south limbs of the syncline, the eastern closure, and the generally east-dipping strata of the West Range. While these domains are statistically homogeneous with respect to the bedding and all later structures, a number of smaller domains could be established in the west where the intermediate-scale folds develop. However, the geometry of these structures is quite clear from the maps of bedding attitudes and the cross sections, and establishing more domains seems unnecessary.

The bedding on the north limb of the fold is locally overturned and dips northward at angles as low as 55°, but statistically it is essentially vertical with a strike of 80°. There are two 6% maxima on Plate V, representing southeasterly dips of 20° and 60° and the smaller maximum representing a dip of 30° to the northwest. These maxima reflect attitudes in the extreme west of the domain in the Narrows Creek area where the bedding starts to swing into the West Range and is affected by intermediate-scale folding.

The stereoplot of bedding from the south limb shows a pronounced maximum representing a dip of 15°/340° (i.e., dip amount 15°, dip direction 340°). The partial girdle about a horizontal axis trending about 50° reflects the presence of the intermediate-scale folds in the west. In the hinge zone of the fold in the east, the bedding dips 40°/235°, and in the West Range it dips approximately 30°/98°.

Main Phase Structures. Most of the cleavage in the quartzite of the north limb (S_1') is flat lying, and therefore rather variable in strike. The spread of poles to S_1' (Pl. V) on a vertical girdle trending 350° results from refraction on a mesoscopic scale into the phyllitic cleavage (S_1), which dips steeply north (Fig. 3), and also from steeply-dipping S_1' cleavage in quartzite on the flatter limbs of the intermediate-scale folds in the Diamond Hill area. Intersection of cleavage (both S_1 and S_1') with bedding plunges at low angles to 80° or 260°. The spread of S_0/S_1' intersections to a more northeasterly trend once again reflects the situation near the west end of the north range.

On the south limb, the S_1' cleavage is very consistent, a strong statistical maximum indicating a dip of 80°/170°. The phyllitic cleavage dips at 35°/335°, the more north-easterly strike here (65°) reflecting the fact that phyllitic layers are exposed mainly in the western part of the South Range. Bedding/cleavage intersections mainly plunge east-northeast or west-southwest at angles of less than 20°. More northeasterly trends occur in the west.

Phyllitic layers are not common in the vicinity of the eastern closure or in the West Range. Hence cleavage in these domains is largely confined to quartzite cleavage, which dips at 65° or 70°/350°, approximately parallel to the axial surface of the Baraboo syncline. Bedding/cleavage intersections plunge 25°/275° in the east and 25°/75° in the west.

The early penetrative slaty cleavage (S_{1E}) has the same strike as the S_1 surfaces and has a variable dip. On the north limb it dips more steeply than S_1 (e.g., at Van Hise Rock, Stop 2); on the south limb it is either flatter than S_1 (as at the northeast end of Devils Lake, Stop 5) or steeper (as at La Rue quarry, Stop 7). At the latter locality it even dips steeply south-southeast.

The attitude of the longrain in S_1 and the slickensides on bedding throughout the area is shown on Plates IV and V. With the exception of a few slickensides on near-vertical beds on the north limb of the syncline, they mostly plunge at angles between 15° and 45° to the northwest of north-northwest. As would be expected, the plunge is steeper on the north limb.

At the few localities where it could be measured, the fine crenulation cleavage (S_{1L}) is near vertical, and strikes north-northwest-east-southeast. The axes of the crenulations are, as previously mentioned, parallel to the longrain in the slaty cleavage.

Secondary Structures. On the north limb of the syncline only one secondary phase crenulation cleavage occurs and can be seen in the field. This dips at 60° to 80° to the south. The axes of the crenulations have a plunge varying from 65°/262° to 78°/250°, being controlled by the S_0/S_2 intersection. A faint conjugate cleavage can be seen in this section (Fig. 5D). On the south limb there are two conjugate crenulation cleavages. These dip at 0°-40° to the southeast and 45°-80° to the northwest (Pl. V). The crenulation axes of both sets have a low plunge (generally less than 20°).

Late Phase Structures. Very few late-phase structures have been found. They are folds with axial surfaces dipping steeply just west of north and hinge lines that are rather more variable in attitude, but plunge at low angles east and west and low to moderate angles northeast (Pl. V).

Joints. The Baraboo quartzite is intensely fractured. Over 700 joints were measured on a purely random basis, a few of the most prominent surfaces at each locality being recorded. The results are surprisingly consistent (Pl. V). Throughout the area, joints approximately normal to the hinge line of the fold (so-called "ac" joints) are by far the most common. Flat-lying joints are sparsely developed, and are most common on the north limb of the syncline where the quartzite cleavage (S_1') is nearly horizontal. There are steeply-dipping joints striking east-northeast-west-southwest and northeast-southwest only on the south limb, where the quartzite cleavage has this orientation. Hence some of the joints, but not a large proportion, are a megascopic manifestation of the S_1' cleavage in the quartzite.

Faults. Owing to the lack of marker horizons, it is difficult to prove the presence of major faults in the Baraboo Quartzite. The important North Freedom fault recognized on the basis of subsurface data has already been mentioned. A. Leith (1935) interpreted a north-south fault, downthrown to the east, to lie in the Lower Narrows, apparently on the basis of subsurface information now missing. A fault through the Narrows Creek gorge in the west end of the syncline is indicated by a swing in the strike of the quartzite (Pls. I and III), but other faults, which have been postulated in the Upper Narrows area on the basis of the occurrence of breccia zones in the quartzite, are doubtful. The breccia seems more likely to be the result of explosive hydrothermal activity than cataclasis (see Stop 2). An important fault may well be present in the extreme southwestern corner of the Baraboo basin cutting out the Baraboo syncline and Diamond Hill anticline, but this cannot be verified on the basis of available data.

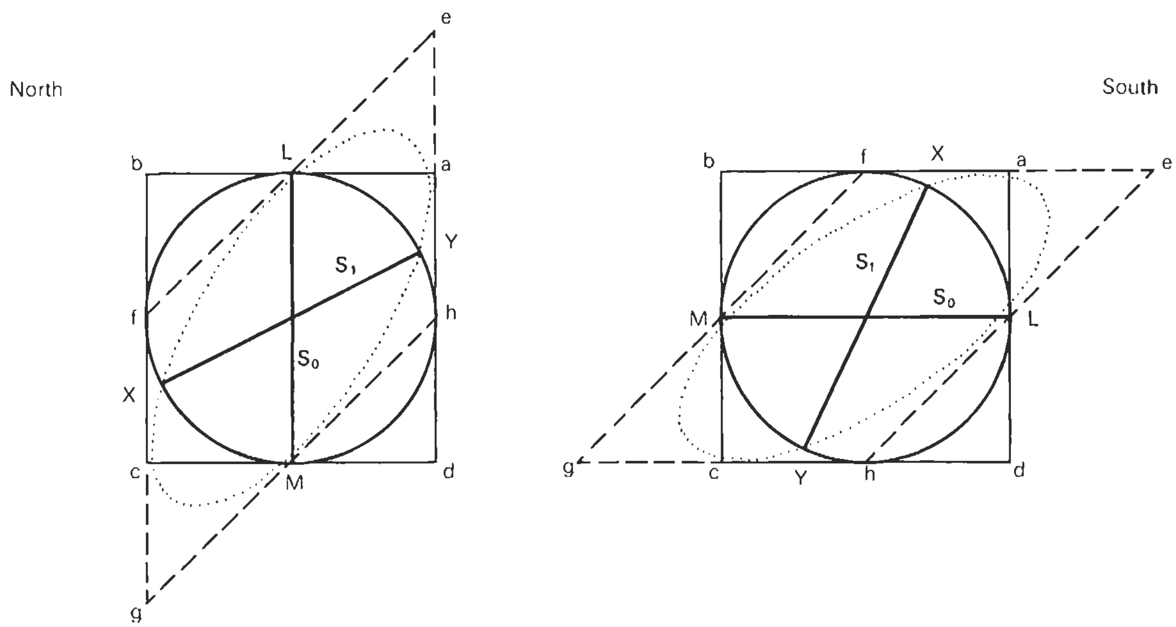

Figure 6. Mechanical significance of main-phase cleavage in the Baraboo Quartzite according to the "Wisconsin school" of structural geologists (after Steidtman, 1910). The diagram shows the approximate attitude of bedding (S_0) and main phase cleavage (S_1) on the north (left) and south (right) limbs of the Baraboo syncline.

It was assumed that the square abcd was deformed in each limb by interbed slip (upper beds moving out of the synclinal trough relative to those below) to become the parallelogram efgh. The diagram shows the shape of the ellipse (dotted lines) created by distortion of the inscribed circle of abcd. The bedding and cleavage were considered to be conjugate shear failure surfaces and to coincide with the "planes of no distortion" (LM and XY) in the ellipse just prior to failure. Refraction of the S_1' quartzite cleavage (strike joints of Steidtman and of C.K. Leith, 1913 and 1923) into the S_1 phyllitic cleavage (strike cleavage of Steidtman and fracture cleavage of Leith) was held to be the result of greater relative rotation of the "planes of no distortion" in the more ductile phyllitic layers prior to failure (see discussion in text of genesis of the Baraboo syncline).

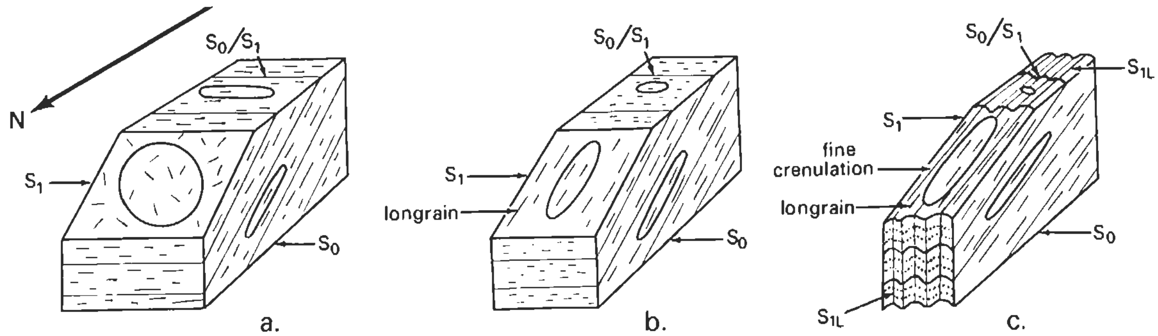

Figure 7. Block diagrams illustrating the suggested mode of origin of the longrain, phyllitic cleavage (S_1), and the late main-phase crenulations and cleavage (S_{1L}) in phyllitic layers of the Baraboo Quartzite. The diagrams show the situation on the south limb of the syncline.

The three principal sections of the finite strain ellipsoid resulting from deformation of an original reference sphere are shown in each diagram. Where flattening alone was involved (a) a phyllitic cleavage without any longrain would be developed. Where the ellipsoid was triaxial, however (i.e., where there was some shortening parallel to the hinge line of the Baraboo syncline), the phyllosilicates would tend to have been aligned parallel to the longest axis of the finite strain ellipsoid (b). More extreme shortening parallel to the hinge line would have led to crenulation of the S_1 surface about axes parallel to the longrain (c), resulting in the formation of the S_{1L} crenulation cleavage.

Note: (i) The use of the strain ellipsoid here as an indicator of the deformation does not assume any mechanical significance for the cleavage surfaces as in Figure 6.

(ii) It is not possible at present to be certain whether or not significant rotational strain was involved in the formation of S_1. If S_1 did not form in the principal section of the finite strain ellipsoid normal to the maximum shortening axis, it would have rotated towards this section (see text). The longrain and crenulations would have developed in the same way as that outlined above even if S_1 did not come to lie normal to the maximum shortening axis. They would be parallel to the longest axis of the elliptical section of the finite strain ellipsoid in which S_1 lay.

Genesis of the Baraboo Syncline
(Dalziel)

The Baraboo syncline developed when a thick pile of Precambrian sediments was subjected to regional tectonic stress in a mobile belt under conditions that favored ductile behavior. The sedimentary pile now exposed was anisotropic and inhomogeneous, consisting of arenaceous material with pronounced stratification (and cross stratification), containing numerous thin argillaceous partings, and overlain by other lithic units. Hence no matter how homogeneous the regional tectonic stress field may have been, the pattern of stress and of resultant strain within the deforming sediments must have been complex. Moreover, it must have been constantly changing as the fold developed.

Reinterpretation of Riley's data (1967) on the orientation of quartz deformation lamellae throughout the Baraboo Quartzite in the light of experimental studies (Carter and Friedman, 1965; see Fig. 8) shows that the lamellae were formed by a stress system in which σ_1 and σ_3 (respectively the greatest and least compressive principal stresses) lay in a vertical plane striking north-northwest--east-southeast. At a number of different localities where bands of quartz-filled tension gashes form distinct conjugate bands, the attitude and shear sense indicate that the gashes resulted from a stress system in which σ_1 and σ_2 (intermediate compressive principal stress) lay in the same plane (Fig. 8, Dalziel, 1969a). These interpretations are not at variance. Recent studies indicate that deformation lamellae characteristically reflect the stress field early in the fold history (Dieterich and Carter, 1969). Shear along the tension gash bands at Baraboo has deformed the (S_1') cleavage in the quartzite (Fig. 54; Dalziel, 1969a), which is locally marked by deformed quartz grains. Hence the tension gashes probably reflect the stress system at a later stage in the progressive development of the syncline than do the deformation lamellae.

The available evidence, therefore, indicates a regional stress system consistent with the over-all geometry of the Baraboo syncline. The plane normal to the hinge-line of the syncline, the hinge lines of the minor folds, and the strike of the quartzite, slaty, phyllitic, and secondary phase cleavages was the σ_1/σ_3 plane early in the fold history and later the σ_1/σ_2 plane. The orientation of σ_1 within this plane was variable (Fig. 8), probably in time as well as in space. The dominant joints in the quartzite (striking north-northwest--south-southeast and vertical, Pl. V) may have formed as extension fractures normal to σ_3 at approximately the same stage in the fold history as the tension gashes. Some are quartz-filled. No other joint sets consistent enough to warrant dynamic interpretation were distinguished.

In attempting to explain the process of folding in rocks in response to tectonic stress, geologists have tended to invoke relatively simple mechanisms that can account for recognizable fold shapes and associated small-scale structures such as slickensides and axial surface cleavage. However, it should be borne in mind that just as the various mesoscopic structures in the Baraboo Quartzite represent only recognizable stages in the progressive strain of the lithic units in which they occur, so the various mechanisms that can be called upon to explain them represent only the readily distinguishable components of a much more complex deformation. These components must be integrated in time and space before the strain is completely portrayed. Some possible components can be expressed in terms of slip or flow parallel to and across layer boundaries, giving rise to the terms flexural slip, flexural flow, passive slip and passive flow (Donath and Parker, 1964), which are commonly used to approximate the movement of material in layered rocks during folding.

The remainder of this section is devoted to a brief discussion of the possible mechanical significance of the most important mesoscopic structures in the Baraboo Quartzite during the genesis of the syncline. It should be emphasized that this discussion is largely based on field work. Detailed study of the structures under the microscope was started only recently.

The geometry of folded quartzite layers within the phyllitic zone on the south limb and of intermediate-scale folds at the west end of the syncline is essentially concentric. Slickensides on the bedding surfaces indicate active bedding slip during folding, the upper beds having moved up-dip towards anticlinal crests relative to those below. Thus a "flexural slip" component operated, and must have resulted in re-orientation of the principal stresses at and near layer boundaries.

Movement on cleavage surfaces across bedding (i.e., a "passive slip" component) is revealed by very slight displacements along S_1 and S_1' (Fig. 27a and b). The effect of this component was probably insignificant in the formation of the syncline, but it may have been significant in opening up partings and even joints along the cleavage surfaces. Joints parallel to S_1' in the quartzite are common.

Flow within individual layers is much harder to define. Theoretical studies reveal a complex pattern and history of stress and strain even within folding layers that behave as viscous fluids (Chapple, 1968; Dieterich and Carter, 1969). In the Baraboo Quartzite the preferred orientation of phyllosilicate minerals and locally of quartz grains, slickensides on cross stratification surfaces, and the presence of quartz deformation lamellae, all reveal ductile flow within individual layers. However, in the absence of reliable indicators of finite strain, and without information on the variation in orientation of σ_1 within the plane normal to the hinge line of the fold, the mechanical significance of the cleavages within the individual beds must remain in doubt.

In the field and in some thin sections, the quartzite cleavage (S_1') appears to be a classic "fracture" cleavage[6]. Towards the more phyllitic margins of the quartzite beds and in less pure quartzite, however, it is controlled by the alignment of phyllosilicates and inequidimensional quartz grains (Fig. 5b) and is "refracted" into the phyllitic cleavage. Geologists of the "Wisconsin School" interpreted both these cleavages as conjugate surfaces of shear failure and equated them with the planes of "no distortion" in a plane strain ellipsoid immediately prior to failure (see Fig. 6). There are many problems with this type of interpretation, as was pointed out by Griggs as early as 1935 (see also Leith, 1937). Principally, however, there is no sound basis for assuming that either of the cleavages is a surface of shear failure, or that either originated in its present orientation with

[6]Fracture cleavage is used here as a closely-spaced parting not controlled by mineral alignment.

Figure 8. Dynamic (stress) analysis of the Baraboo syncline.

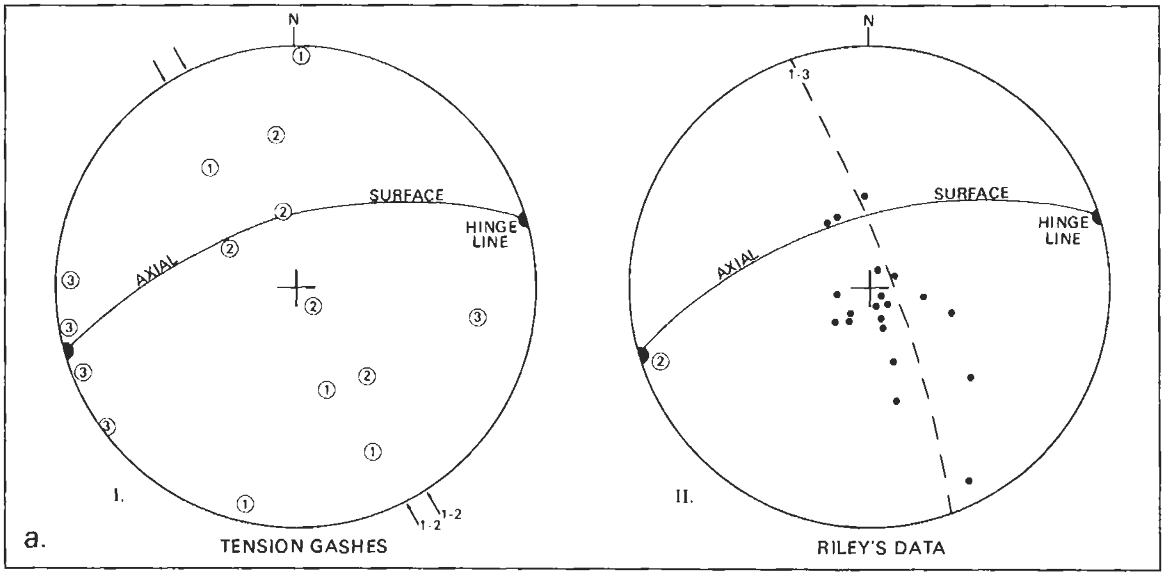

Figure 8a. Stereoplots showing the orientation of principal stress axes as deduced from tension gash bands (Dalziel, 1969a), and Riley's data (1947) on the orientation of quartz deformation lamellae (for interpretation of Riley's data, see Carter and Friedman, 1965).

The greatest compressive principal stress axes (σ_1) are denoted by the figure 1, the intermediate compressive principal stress axes (σ_2) by 2 and the least compressive principal stress axes (σ_3) by 3. The tension gash data was obtained from a number of bands at various localities around the syncline. For each the orientation of σ_1, σ_2, and σ_3, is plotted. Where the exact attitude of σ_1, and σ_2, within the plane normal to σ_3 could not be deduced (e.g., on a smooth glaciated surface), the strike of the σ_1/σ_2 plane is denoted by 1-2 on the stereoplot. Each black dot on the stereoplot of Riley's data denotes the σ_3 axis determined from the microscopic analysis of one specimen. 1-3 is the σ_1/σ_3 plane, and 2 the orientation of σ_2 determined from this data.

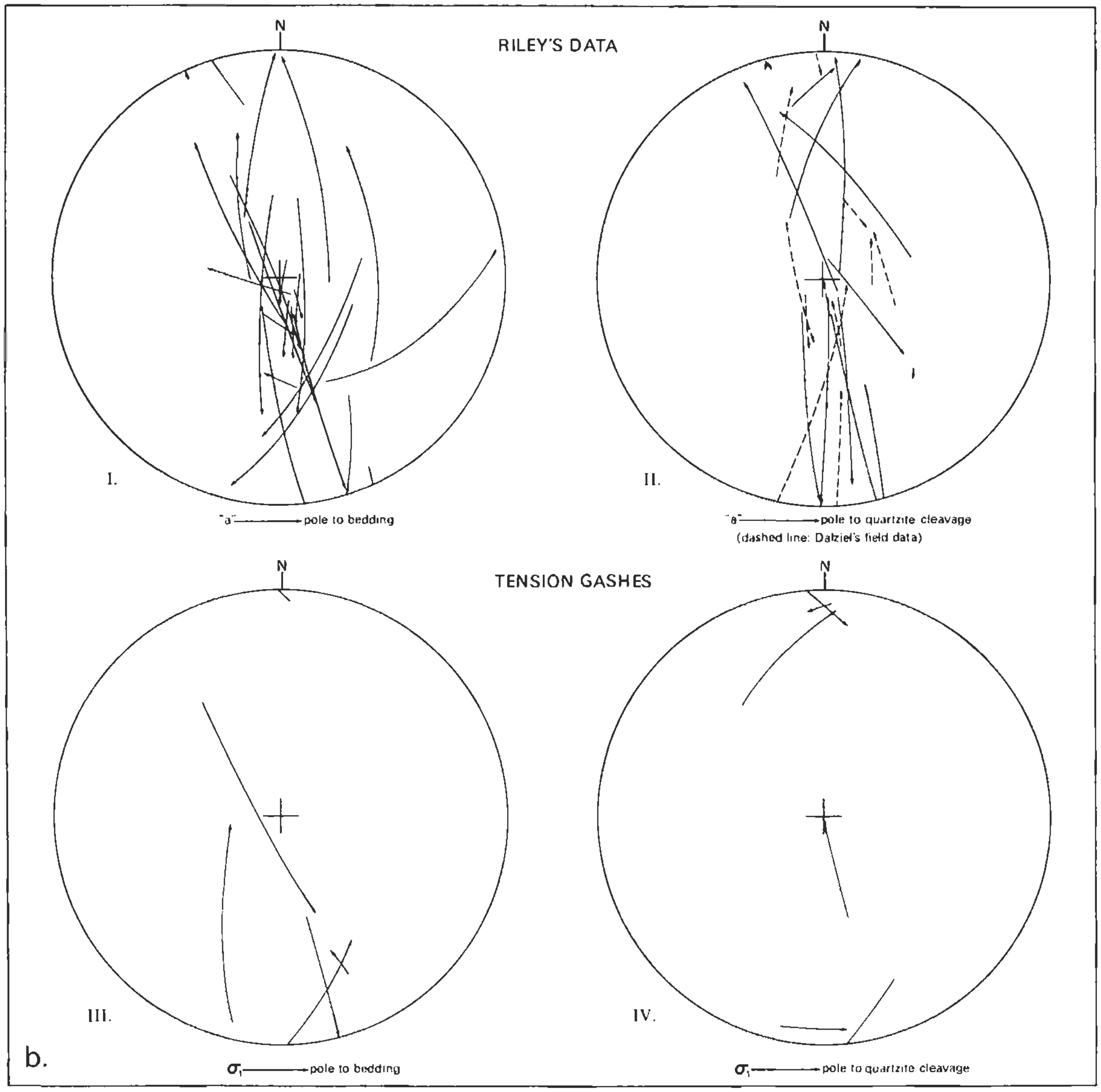

Figure 8b. Stereoplots showing the relationship of the stress axes (deduced from Riley's data and the tension gash data) to the bedding and quartzite cleavage. The arrows are drawn from Riley's "a" axes (σ_3 of Carter and Friedman, 1965) and the σ_1 axes determined from the tension gashes, to the pole to bedding or quartzite cleavage. They show no consistent angular relationship between the principal stress axes and these structural elements.

respect to bedding. Rotation of S_1 in the ductile phyllitic layers after its formation would certainly be expected.

Most studies of the relationship of principal stress axes to the geometry of simple folds indicate that σ_1 lay in or close to the bedding, and at right angles to the hinge line of the fold, at the time when the lamellae formed, probably early in the fold history (Carter and Friedman, 1965; Dieterich and Carter, 1969). Therefore, as the S_1' cleavage is almost invariably at right angles to bedding, it is possible that it may have formed essentially normal to σ_1, and normal to the direction of maximum shortening in the quartzite beds at an early stage in the folding. No consistent geometric relationship exists between the principal stresses that can be inferred from Riley's deformation lamellae data (1947) and the quartzite cleavage and bedding (Fig. 8). The exact location of the specimens studied by Riley is however unknown, and the problem is being studied at the present time using material collected on the basis of the field studies. The mechanical significance of S_1' is still uncertain.

The slaty (S_{1E}) and phyllitic cleavages in the phyllitic layers are both marked by aligned phyllosilicates. The former, however, is very penetrative and the latter formed by more discrete surfaces in which the deformed phyllosilicates aligned in S_{1E} appear to have been concentrated (Fig. 5a). The continuation of S_1' surfaces in the quartzite into S_1 in the phyllite (Figs. 4 and 29a) point to a close relationship between the development of the two surfaces.

Some structures may develop homogeneously throughout a multilayered sequence early in the strain history and later may undergo local heterogeneous strain, being deformed, for example, near layer boundaries. Others may originate at various times as a result of only localized stress fields such as those near layer boundaries. Hence the presence, nature, and attitude of the slaty (S_{1E}) and phyllitic cleavages in the phyllitic layers may result in part from the rotation and modification of pre-existing structural surfaces that formed at a high angle to bedding (like the S_1' cleavage in the quartzite), and in part from the stress field that must have been set up in the thin layers of more ductile phyllitic material between quartzite beds during folding. Once formed both surfaces would tend to rotate towards the principal section of the finite strain ellipsoid in the phyllitic layers normal to the direction of maximum shortening even if they did not originate in this position (Flinn, 1962; Ramsay, 1967). The phyllitic cleavage (S_1) is shown in this principal section in Figure 7, but may not have reached it in all phyllitic layers. It is important to note that the strain ellipsoid is used here purely as a measure of the deformation undergone by a reference sphere in the rocks and does not assume the mechanical significance of the cleavage surfaces.

Alignment of layer silicates to form a <u>longrain</u> lineation in S_1 nearly normal to the bedding/cleavage intersection, and the presence of the fine crenulation cleavage (S_{1L}) indicate that the finite strain ellipsoid within the phyllitic layers was triaxial rather than oblate, the lineation probably representing the direction of maximum finite elongation within the S_1 surface. This would be possible even if S_1 had not come to lie in a principal section of the ellipsoid (see Fig. 7). The fine crenulation on the phyllitic cleavage surfaces, which also plunges down-dip exactly parallel to the <u>longrain</u>, may be the result of a minor amount of shortening parallel to the bedding/cleavage intersection (Fig. 7). This shortening, being parallel to the hinge line of the Baraboo syncline, could account for the fact that the structure is doubly plunging.

Numerous mechanical explanations of the secondary phase structures have been put forward (see Adair, 1956, for a summary). Hendrix and Schaiowitz (1964) suggested on the basis of the northerly vergence of the associated folds on the south limbs that they were the result of gravitational sliding of the upper beds back into the trough of syncline after relaxation of the regional stress system. Structures of this type are, however, very common as late stage compressional features in mobile belts (Dewey, 1969b). Moreover, their conjugate geometry, which was not recognized by Hendrix and Schaiowitz, removes the appeal of an explanation based on "reverse" simple shear resulting from gravitational slumping. The fact that the axes of most of the secondary crenulations and associated folds are parallel to the hinge line of the Baraboo syncline suggests that they may be related to the same regional stress system.

There is no reason to assume that the secondary structures on the north and south limbs need be synchronous. They merely both postdate the formation of the phyllitic cleavage.

The Baraboo syncline has as much to offer structural geologists today as it did nearly 100 years ago when Irving (1877) first described the cleavage in the phyllitic layer at the northeast corner of Devils Lake. One of the reasons for this is the fact that the response of the Baraboo Quartzite to the application of a regional stress field was in part ductile and in part brittle. Hence the area affords an opportunity to study the relationship between structures resulting from a whole range of rheologic behavior. Moreover, the metamorphism did not result in complete recrystallization of the quartz grains, making it possible to analyze the structures in terms of the causative stress system by utilizing data on intracrystalline deformation.

Hopefully this description of the structures in the Baraboo Quartzite, and the numerous problems raised, will re-awaken the interest on the part of structural geologists that this classic area used to enjoy, and which it so fully deserves.

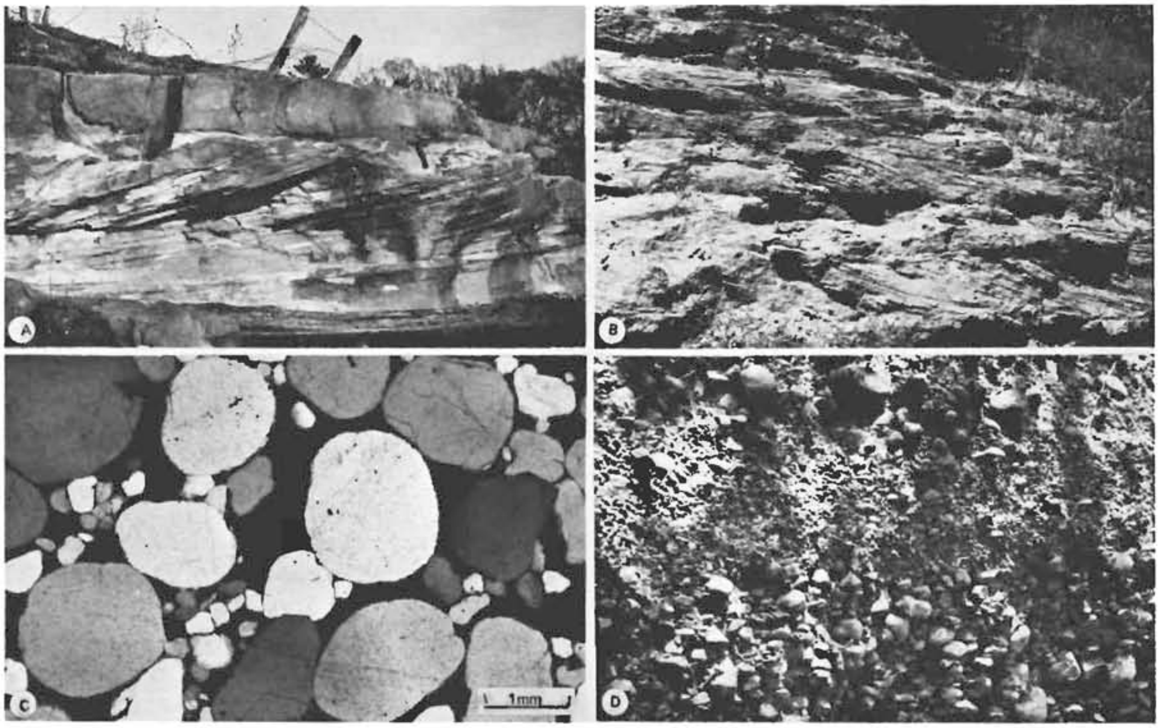

Figure 9. Photographs of some typical Cambrian sediments:

A. Large-amplitude trough cross bedding with parallel, planar truncation surfaces typical of the Galesville Sandstone. (NW$\frac{1}{4}$, Sec. 31, T.12N., R.4E.) (Photo by Enis Usbug).

B. Complex medium-amplitude wedge trough cross bedding typical of the Tunnel City Formation. Note the great variability of apparent dips of cross sets and a few reversals at trough axes. Pen near center provides scale. (One-eighth mile north of County Road V, SW$\frac{1}{4}$, Sec. 7, T.10N., R.8E.).

C. Photomicrograph of typical Galesville Sandstone from the Martin-Marietta Co. Sand quarry just south of Portage, Wisconsin, 5 miles east of the Baraboo syncline. Note the perfection of rounding of large grains and a conspicuous bi-modality of size distribution. (Crossed nicols; largest grains are 2 millimeters across).

D. Typical crudely-stratified conglomerate of rounded quartzite pebbles and cobbles with a glauconitic sand matrix deposited very near the outer side of a quartzite island. Largest cobbles are about 5 or 6 inches in diameter. (In deep valley in NE$\frac{1}{4}$, Sec. 36, T.12N., R.4E.) (Photo by Enis Usbug).

PALEOZOIC GEOLOGY
(Dott)

Stratigraphic Summary

General. Table 5 summarizes the Paleozoic stratigraphic succession of the Baraboo region. Wanenmacher, et al., (1934) discussed the local physical and bio-stratigraphy extensively and presented stratigraphic sections. Wanenmacher (1932) mapped the Paleozoic rocks, but his results were not portrayed on a topographic base map. Berg (1954), Berg, et al., (1956), C.A. Nelson (1956), and Ostrom (1964) presented the most up-to-date regional stratigraphic syntheses. Because the chief objective here is to elucidate paleogeography and sediment genesis, only brief stratigraphic comments will be made as necessary to update terminology and to indicate criteria used in mapping the formations. Biostratigraphy is not treated. In general, the Paleozoic units are very uniform even within the Baraboo basin (that is the Cambrian lagoon that was surrounded by quartzite islands). Only very near the quartzite is their identity completely masked by conglomerate; in such areas the formations were not mapped separately. Earlier workers (e.g. Wanenmacher, 1932) recognized that several formations became conglomeratic near the islands and also showed initial dips, but they did not map such features.

The oldest Late Cambrian formations generally recognized in western Wisconsin, the Mt. Simon and Eau Claire, are not exposed in the Baraboo region, but they have been tentatively identified by F. T. Thwaites from drill cuttings (unpublished file reports of the Wis. Geol. and Nat. Hist. Survey). Regional studies have suggested that the Eau Claire does not extend into this area, and it is entirely possible that the samples assigned by Thwaites to the Eau Claire actually represent a shaly part of the Mt. Simon Formation.

The Galesville Sandstone is the oldest exposed Paleozoic unit (see Table 5, footnote 3). It is characterized in most outcrops by its white to gray color and very friable nature in contrast to darker-weathering and more resistant overlying strata. The Galesville is unfossiliferous in the area, and is comprised of well-rounded and well-sorted medium to coarse pure quartz sandstone (Tables 6, 8; Fig. 9). The Ironton unit (see Table 5, footnote 2), which normally overlies the Galesville and contains fossil shell debris, is not clearly discernible in the immediate syncline area, but occurs a few miles to the northwest (e.g. along County Road H, Sec. 25,T.13N., R.4E., north of Reedsburg).

Cross stratification is universal, being chiefly of the trough type with parallel truncation surfaces (Figs. 9, 10), although in certain localities (e.g. around the Upper Narrows -- Stop 2 and Supplementary Stop C), wedge truncations are prominent. Amplitudes up to 8 or 10 feet are known, but the average is about 6-8 inches; trough widths are from 1-10 feet. Accessible localities where Galesville Sandstone is well exposed include the Upper Narrows, Skillet Creek Falls (Pewits Nest), Fox Glen (NE Sec.22,T.12N., R.8E.), around Leland, northeast of Reedsburg (especially in a road cut in north-central Sec.11, T.12N., R.4E.), and along the Wisconsin Dells 6-10 miles north of the syncline.

The Tunnel City was formerly called the Franconia Formation, but the

TABLE 5. PHANEROZOIC STRATIGRAPHY OF THE BARABOO REGION[1]

ERA	PERIOD	EPOCH	AGE	FORMATIONAL UNITS	CHARACTERISTICS
CENOZOIC	NEOGENE ("Quaternary")	Holocene	Late Woodfordian	Alluvium	Undifferentiated gravels, and other fluvial and eolian deposits
		Pleistocene		Cary till and glacial lake deposits	Unsorted moraine, outwash, and fine lake sediments
				UNCONFORMITY	
PALEOZOIC	ORDOVICIAN	Champlainian		Platteville Limestone	Doubtfully present on Gilbralter Rock
		Chazyan		St. Peter Sandstone	Tan to brown, well sorted and rounded, pure quartz sandstone; variable cross stratification; only locally preserved on tops of a few high hills (T = 100 - 120 ft)
				UNCONFORMITY	
		Canadian		Prairie du Chien Group: Shakopee Formation	(Not present in this area)
				Prairie du Chien Group: Oneota Dolomite	Resistant tan dolomite with white chert nodules; oolite in lower part; some quartz sandstone; rare fossil molds, small stromatolites. Caprock on many hilltops. (T = 80 - 100 ft)
	CAMBRIAN	Late Cambrian or Croixan	Trempealeauan	Trempealeau Group: Jordan Sandstone	White or brown, well rounded and sorted, medium, pure quartz sandstone; variable cross stratification; burrows locally abundant. (T = 30 - 50 ft)
				Trempealeau Group: St. Lawrence Formation: Lodi Member	Micaceous dolomitic siltstone and fine sandstone and thin dolomite (T = 10 - 15 ft) (Trilobites; Meristomes, etc.)
				Trempealeau Group: St. Lawrence Formation: Black Earth Member	Gray to buff, wavy-bedded dolomite and siltstone; locally fossiliferous (stromatolites, burrows, shells). (T = 15 - 20 ft)
			Franconian	Tunnel City Formation[2]	Buff, dolomitic and glauconitic fine sandstone and thin dolomite; glauconite variable, but increases upward; rare skeletal debris, burrows common; trough - wedge cross stratification universal ("festoon"). (T = 100 - 150 ft)
			Dresbachian	Galesville Formation[3]	White, unfossiliferous, well rounded and sorted, medium, pure quartz sandstone; trough cross stratification with dominant parallel truncation surfaces. (T = 100 ft. +)
				Eau Claire Formation (Subsurface only if present)	Glauconitic siltstone, fine sandstone and shale; fossiliferous. (T = 200 - 250 ft)
				Mt. Simon Formation (Subsurface only)	Sandstone, some conglomerate, and miner shale. (T = 200 ft ?)
				UNCONFORMITY	
				PREPALEOZOIC	

1. After Wisconsin Geological and Natural History Survey Information Circular No. 8 (Ostrom, 1967). Thicknesses (T) are maxima for the Baraboo region; all units thin to zero against quartzite hills. Several additional unconformities not shown have been postulated, but they would appear to be of relatively short duration -- if valid at all.

2. On a regional stratigraphic basis, Ostrom (1967) suggested Tunnel City as a group rank with two subdivisions, the Mazomanie and the glauconite-rich Lone Rock Formations, which have complex mutual facies relationships. In the Baraboo region the Lone Rock is not clearly represented, therefore, the Tunnel City has been treated here as one formational unit.

3. Ostrom (1967) suggested a new name -- the Wonewoc Formation -- to include Galesville and Ironton Members. In the Baraboo region, however, the Ironton can not be mapped, therefore the Galesville has been treated here as a formational unit.

latter name has been restricted by the Wisconsin Survey as a time-stratigraphic (Stage) name (Ostrom, 1967); stratigraphic nomenclature for this interval is in a state of transition (see Table 5, footnote 2). At many localities, the Tunnel City is disconformable on the Galesville (see Stop 9), but there is no proof that a single, major, widespread unconformity exists. The Tunnel City consists largely of subangular to subrounded, fine to medium quartz sandstone (Table 8) with appreciable dolomitic cement and rare, thin dolomite layers. Presence of the carbonate makes the strata more resistant than adjacent units; the Tunnel City also is darker-weathering. Glauconite is a conspicuous accessory mineral, but at only a few localities in this area is it abundant enough to form greensands (e.g. more than 50% of grains); it is least common inside the Baraboo basin. In general, glauconite increases in abundance upward in the formation.

Cross stratification is universal, and is invariably of the trough-wedge ("festoon") type (Figs. 9, 10) as Farkas showed (1960). Amplitudes average from 4 to 6 inches and, where fully exposed, troughs are generally 1-2 feet wide and have very gently-plunging axes; oppositely-plunging troughs appear to be rather common (Fig. 11). A spectacular zone of convolute lamination 6-8 inches thick occurs along Wall Street in the NW, SW, Sec.25, T.12N., R.7E. at about 1,020 feet elevation. Apparently it formed by temporary liquefaction of the strata with convolutions produced while in a quick state; it is the only such feature noted anywhere around Baraboo.

Intra-formational ("rip-up") conglomerates consisting of flat dolomite or dolomitic siltstone clasts are characteristic of the Tunnel City. Quartzite conglomerates are most common and comprise a greater total volume in the Tunnel City than in any other strata, although most were mapped within a separate lithologic unit of variable age. Most of the Tunnel City strata of the Baraboo area were assigned to the relatively non-glauconitic and prominently cross-stratified Mazomanie subdivision by Berg (1954) (see Table 5, footnote 2). Farther west, contemporaneous strata are more glauconitic and less cross stratified, apparently reflecting offshore deposition in less agitated water. Tunnel City strata are extensively burrowed, and they contain widely scattered fragmental trilobite, brachiopod, graptolite, and echinoderm (cystoid?) material, proving a marine origin. Extensive biostratigraphic zoning of the formation was attempted in the 1930's, but it has been shown that the zones are discordant regionally with lithic subdivisions (Berg, 1954). Inside the Baraboo basin, a significant unconformity within the formation was postulated by Wanenmacher, <u>et al</u> (1934), but seemingly on biostratigraphic criteria alone; the formation is thinner within the basin.

The Tunnel City unit is widely exposed both inside and outside the western half of the syncline area, and is well exposed one-half mile east of the Lower Narrows inside the syncline as well as on many hills southeast of the syncline.

<u>The Trempealeau Group</u>, consisting of the <u>St. Lawrence Formation</u> below and the <u>Jordan Sandstone</u> above, overlies Tunnel City strata with apparent conformity. Only a rather abrupt upward increase of dolomite distinguishes the Black Earth Dolomite Member of the St. Lawrence from the underlying unit. The Black Earth is a useful datum for mapping in the Baraboo area, although it crops out only sporadically. Wanenmacher, <u>et al</u> (1934), postulated a disconformity at the base of the dolomite on the basis of a "basal" greensand conglomerate. But, as there are dozens of similar conglomerates well down in the Tunnel City, there seems no reason to attach greater significance to that at

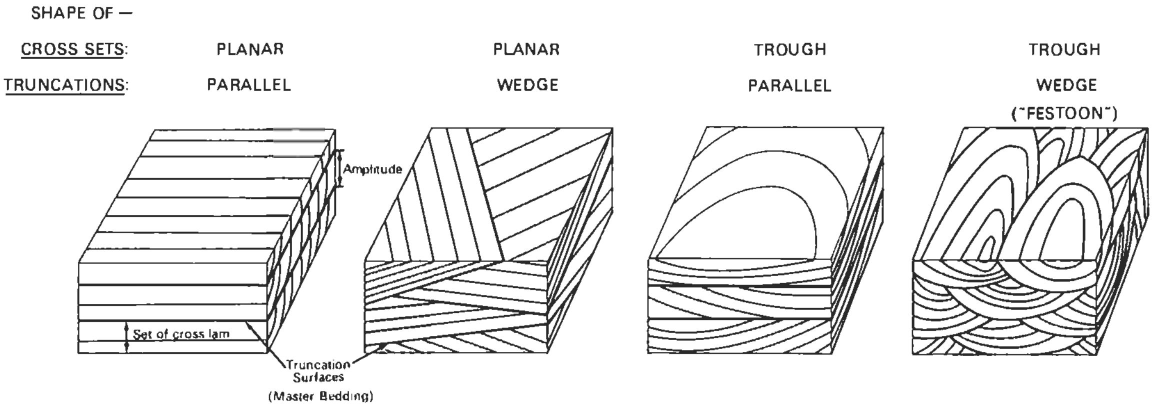

Figure 10. Types of cross stratification and geometric parameters recognized in this study. Trough cross sets with parallel planar truncations (or master bedding) characterize the Galesville Sandstone, trough sets with inclined or wedge truncations ("festoon" cross bedding) typify Tunnel City strata, and mixtures of the two occur in the Jordan and St. Peter sandstones as well as in the Baraboo Quartzite.

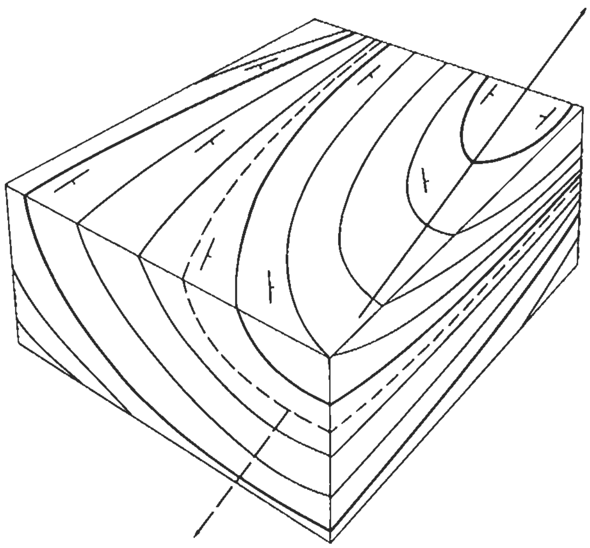

Figure 11. Diagrammatic portrayal of thickness changes vertically within one set of trough cross beds that produces opposite plunge directions of troughs in closely associated laminae. This phenomenon has been observed most clearly in Tunnel City strata, as have oppositely-plunging separate, but adjacent troughs. A few individual canoe-shaped troughs plunging from both ends toward their center have been observed in the Galesville Sandstone. (Compare data of Fig. 12 and Pl. VII).

TABLE 6. MAJOR SILICATE MINERALS OF SANDSTONES OF THE BARABOO REGION

(Collectively accounting for 95-99 percent of the rocks except for the Tunnel City Formation, which has about 30 percent carbonate. N = 35 thin sections, 300 points per section)

FORMATION	QUARTZ TYPES			GLAUCONITE	FELDSPAR (in part authigenic)
	Unstrained	Strained	Polycrystalline		
Jordan Sandstone	76%	22%	0.5%	0.1%	1.5%
Tunnel City Formation	62%	25%	6%*	6%**	1%
Galesville Sandstone	69%	30%	1%	----	0.5%
Baraboo Quartzite***	trace	85%	10%	----	----
Waterloo Quartzite***	----	87%	4%	----	trace

* Largely quartzite granules; all other data for sand-sizes only.
** Varies among individual strata from 0 to 90 percent.
*** Also includes about 1%-8% muscovite, about 2% opaque metallic grains, and traces of zircon and chert (jasper) grains.

TABLE 7: PRINCIPAL HEAVY ACCESSORY MINERALS OF SANDSTONES OF THE BARABOO REGION*

(D - Dominant, C - Common, R - Rare, VR - Very Rare)

ST. PETER SANDSTONE (Wanenmacher, 1932; Tyler, 1936):

Zircon (D)
Tourmaline (C)
Ilmenite-Leucoxene (C)
Garnet (R-C)

JORDAN SANDSTONE (Ockerman, 1930; Wanenmacher, 1932):

Garnet (D)
Zircon (C)
Tourmaline (C)
Ilmenite-Leucoxene (R)

TUNNEL CITY FORMATION (Pentland, 1931; Wanenmacher, 1932):

Garnet (D)
Zircon (C)
Tourmaline (C)
Ilmenite-Leucoxene (R)
Calcium phosphate shell debris (variable)
Rutile (VR)
Magnetite (VR)

GALESVILLE SANDSTONE (Wilgus, 1933; Wanenmacher, 1932; Emrich, 1966):

Zircon (D)
Ilmenite-Leucoxene (C)
Tourmaline (C)
Garnet (R)
Rutile (VR)

BARABOO QUARTZITE (Becker, 1931; Wanenmacher, 1932):

Zircon (D)
Magnetite (C)
Pyrite (R)
Rutile (VR)
Barite (VR)
(Only 1 grain of garnet, and no tourmaline)

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*Differences of proportions among heavy mineral suites of the different Cambrian sandstones suggested by investigations under the direction of W. H. Twenhofel in the 1930's are of doubtful statistical significance. First, the range of variation of abundance of many species was large within each formation. Secondly, garnet was the principal species that showed major variations between formations, but clear evidence of differential leaching of it casts doubt upon absolute abundance being of primary stratigraphic significance. (Andrew, 1965).

TABLE 8: SIZE DISTRIBUTION DATA FOR PALEOZOIC SANDSTONES OF THE BARABOO REGION

FORMATION	COARSEST 1 PERCENTILE	MEAN DIAMETER	MEDIAN DIAMETER	SORTING	SKEWNESS
ST. PETER SANDSTONE:					
Gibralter Rock (N = 2)	0.75ø (0.58mm)	2.35ø (0.20mm)	2.52ø (0.17mm)	0.65ø	-0.35
JORDAN SANDSTONE:					
Outside West End of Syncline (N = 9)	0.43ø (0.74mm) (0.18 - 1.40ø)	1.53ø (0.34mm) (0.93 - 2.08ø)	1.56ø (0.33mm) (1.00 - 2.10ø)	0.58ø (0.43 - 0.78)	-0.07 (+.01 to -0.39)
Inside West End of Syncline (N = 4)	0.30ø (0.8mm) (-.45 to +1.4ø)	2.07ø (0.24mm) (1.59 - 2.76Ø)	2.09ø (0.23mm) (1.70 - 2.72ø)	0.59 (0.41 - 0.72)	-0.02 (-.15 to +.10)
TUNNEL CITY FORMATION:					
Outside West End of Syncline (N =25)	1.36ø (0.37mm) (0.60 - 2.20ø)	2.16ø (0.23mm) (1.57 - 2.68ø)	2.51ø (0.18mm) (1.70 - 2.70ø)	0.43 (0.35 - 0.63)	-0.18 (-.53 to +.13)
Inside West End of Syncline (N = 16)	0.80ø (0.58mm) (0.10 - 1.80ø)	2.21ø (0.22mm) (1.30 - 2.77ø)	2.30ø (0.20mm) (0.25 - 0.75)	0.47 (0.25 - 0.75)	+0.01 (-.71 to +.45)
GALESVILLE SANDSTONE:					
Portage Quarry, 5 miles east of syncline (N = 11)	0.40ø (0.75mm) (-.60 to +.06ø)	1.96ø (0.26mm) (1.25 - 2.40ø)	1.93ø (0.26mm) (1.20 - 2.40ø)	0.66 (0.44 - 0.85)	+0.58 (-.75 to +.18)
Outside West End of Syncline (N = 26)	1.41ø (0.37mm) (0.65 - 2.10ø)	2.25ø (0.21mm) (1.65 - 2.75ø)	2.27ø (0.20mm) (1.65 - 2.75ø)	0.42 (0.20 - 0.68)	+0.05 (-.15 to +.42)
Inside West End of Syncline (N = 25)	0.62ø (0.68mm) (-.50 to +1.60ø)	2.00ø (0.25mm) (1.55 - 2.53ø)	1.95ø (0.27mm) (1.48 - 2.50ø)	0.49 (0.32 - 0.69)	+0.08 (-.03 to +2.70)

ø -- means and ranges of Phi Scale values the base of the Black Earth. It is referred here to the underlying formation. Like the Tunnel City, lithic subdivisions of the Trempealeau Group have been shown to cross faunal zones regionally (C.A. Nelson, 1956).

The two members of the St. Lawrence Formation were not differentiated in mapping because of their thinness, but they are very distinctive where exposed. The Black Earth Member is an impure, tan-colored dolomite with thin, wavy bedding. Glauconite is common in the basal portion. At the Wood's Quarry (SW Sec.10, T.11N., R.6E.), spectacular cylindrical algal stromatolites 1 to 3 feet in diameter and up to 4 feet high occur. Smaller stromatolites are present elsewhere, as are abundant burrow structures. The Black Earth was earlier known as the Mendota Dolomite (named for exposures on Lake Mendota in Madison). On the basis of fossil molds at the old Wood's and Eikie's quarries within the Baraboo basin (see geologic map, Pl. I), E. O. Ulrich considered the dolomite to be post-Trempealeau and a part of his Ozarkian System. Wanenmacher, *et al* (1934) showed that both localities belong to the basal Trempealeau, and that Ulrich's "Devils Lake Sandstone", alleged to be basal Ozarkian, actually represents both Franconian and Trempealeauan strata.

The Black Earth grades insensibly upward, as well as northeastward, into the Lodi Member, which is comprised of tan to gray, calcareous and micaceous siltstone, fine sandstone, and very thin impure dolomite. The Lodi is fossiliferous, being especially famous for *Dikelocephalus* and meristome fragments. It is seen only in a few artificial exposures.

The Jordan Sandstone is much like the Galesville (Tables 6, 8), but is coarser grained, contains more conglomerate and burrow structures near old quartzite hills, and is generally darker in the Baraboo area. Also it is more cemented, especially in its upper part where silicification has occurred due to apparent release of silica by weathering of the overlying cherty Oneota Dolomite. Thin red and green shale laminae occur within the Jordan locally. Regionally the lower Jordan (Norwalk Member) is finer-grained than the upper (Van Oser Member). This also seems to be true around Baraboo, as is well displayed in new cuts on a gravel road just north of the syncline (southeast corner of Sec.15, T.12N., R.5E). These subdivisions could not be mapped, but the upper Jordan -- the only portion widely exposed around Baraboo -- is everywhere characterized by coarse-grained sandstone with quartz overgrowths and quartzitic cement. Scattered oolitic spherules, which commonly are hollow, occur locally in the top few inches. Trough cross stratification is ubiquitous; both parallel and wedge truncations are about equally common.

The Sunset Point Member (formerly Madison Sandstone) at the top of the Jordan was not differentiated by us in the Baraboo area. Wanenmacher, (1932) mapped a "Madison Sandstone" unit, but apparently it included all of the Jordan. The Sunset Point is erratic in south-central Wisconsin, and is too thin to map at the 1:62,500 scale. We find none of the characteristic, thinly-stratified, very fine, dolomitic sandstone and fine dolomite with abundant burrow structures that typify the Sunset Point at its type locality in Madison. Instead, the upper contact of relatively coarse quartz sandstone containing scattered oolites, which probably represents Sunset Point-equivalent strata, is overlain directly by the Oneota Dolomite. The exact position of the Cambrian-Ordovician boundary is still debated, but apparently it falls within the interval represented by the Sunset Point.

The Oneota Dolomite, basal unit of the Early Ordovician Prairie du Chien Group, occurs at widely scattered localities, but is most extensive around the southwestern part of the syncline as a resistant caprock on many ridges. Numerous quarries have been opened in it to obtain road material. Only one exposure is known inside the east end of the syncline (SW, SE Sec. 30, T.12N., R. 8E.), but the Oneota occurs on several ridge-tops south of the eastern half of the syncline (see geologic map). Although a disconformity has been postulated at its base (e.g. Wanenmacher, et al, 1934; Raasch and Unfer, 1964), we find no compelling physical evidence in the Baraboo area itself other than an abrupt lithologic change. The Oneota is comprised of hard gray dolomite and considerable dolomitic quartz sandstone. Small pebbles of red quartzite occur rarely. Silicified oolite is common in the lower 20-30 feet as are white-weathering chert nodules. Dessication cracks and ripple marks are present, but algal stromatolites characteristic of the formation elsewhere are not prominent here. Unusual disturbed stratification occurs in the Denzer Quarry (Stop 10; south-central Sec. 14, T. 10N., R. 5 E.). The overlying Shakopee Formation is not present in the Baraboo area.

The St. Peter Sandstone is the youngest Paleozoic formation preserved. It is known to cap Pine Bluff and Gibralter Rock in the east part of the region, and may form unexposed outliers in the southwest (see geologic map). In addition, the topographically highest quartzite-conglomerate facies probably contains St. Peter age-equivalents, but they are impossible to differentiate from Jordan equivalents. The St. Peter here, as elsewhere, consists of brown-weathering, fine to coarse, pure quartz sandstone. It has trough cross stratification in some exposures, but essentially parallel flat lamination in others. It is moderately well cemented.

Younger Paleozoic Strata. It has been claimed that Middle Ordovician Platteville Limestone occurs above the St. Peter on Gibralter Rock (Wanenmacher, et al, 1934), but we find no such material in place today. A few pieces of limestone near the summit could be Platteville remnants, but the bluff has been glaciated, so all loose blocks are suspect. Silurian fossiliferous chert pebbles have been found in potholes on the East Bluffs at Devils Lake. The nearest in-place Silurian today is on Blue Mounds 25 miles south at an elevation of about 1700 feet, while the highest of the Baraboo Hills is 1600 feet. It is virtually certain from regional stratigraphic relationships, however, that Late Ordovician and younger strata formerly covered the Baraboo region.

Cambrian Sedimentology and Paleogeography

General Setting. Late Cambrian marine transgression of the North American craton began about 500 million years ago, and general submergence of most of the craton continued for about 50 million years until an equally extensive, but relatively brief, regression occurred at the end of Early Ordovician (i.e. pre-St. Peter) time when the entire craton again became emergent. The quartz-rich sandstones and dolomites that formed in southern Wisconsin in the Cambro-Ordovician epeiric sea comprise the "type" lithostratigraphic Sauk Sequence of Sloss (1963). Within the Sauk Sequence, several lesser transgressive-regressive "cycles" have been recognized (Ostrom, 1964).

Cambrian sandstones are noteworthy for their extreme purity and textural maturity (Tables 6, 7, 8). Shales, too, though not abundant, are characterized by a simple mineralogy dominated by illite with minor kaolinite and K feldspar (Emrich, 1966; S. S. Bailey, personal communication). The mineralogic maturity long has been interpreted as evidence of derivation from older, already pure sands, which in turn apparently had been formed by countless re-cyclings of still older sands whose ultimate sources were largely granitic terranes (Tyler, 1936). This long-standing concept has been verified recently by extensive studies by Blatt (1967). Purification presumably was favored by long emergence of most of the continent in Late Precambrian time (for 500 to 800 million years), and absence of a land plant cover to impede sediment transport and abrasion. Unusually high rounding values in the Cambrian sandstones (Fig. 9) indicate an important wind history, for recent experiments show that wind is 100 times more effective than water as a rounding agent (Kuenen, 1960). After abundant land plants appeared in Late Devonian time, such well-rounded sandstones did not form on a large scale.

Mature Middle Precambrian quartzites similar to the Baraboo are widely scattered around the Great Lakes region, and probably provided much of the Cambrian sand. Islands are known to have existed in the epeiric sea around Waterloo, Wisconsin (Fig. 1), near Taylor's Falls, Minnesota, and in northern Michigan, as well as at Baraboo. Many others, now buried, must have existed. The Baraboo islands contributed obvious local gravel, but comparison of heavy mineral suites between Cambrian sandstones and the Baraboo Quartzite (Table 7) suggests that very little of the sand was locally derived. Petrographic analysis of quartz types (Table 6) and preliminary studies of sand grain-size distributions in Cambrian sands near the old quartzite hills both confirm this conclusion. Therefore, we concur with most earlier authors that the bulk of the sand came from farther away (see also discussion for STOP 7).

Regional Sediment Dispersal. Early Paleozoic marine current and wind indicators in Wisconsin have been studied by Raasch (1958), Farkas (1960), Hamblin (1961), and Emrich (1966), and results have been reviewed by Ostrom (1964). All authors have concluded that the general average sand transport direction was southerly, which is probably correct, but most of their data show so much scatter that any interpretation of an average direction requires a large measure of faith, statistical significance tests not-withstanding. Emrich's data for the Ironton Sandstone showed an opposite (northeast) mean transport direction for western Wisconsin, and were random for Wisconsin and northern Illinois combined. Roshardt (1965) also found a random pattern of cross set orientations for the St. Peter Sandstone in southwestern Wisconsin. Figure 12 shows representative orientation data for cross strata north of the Baraboo hills, which should be relatively free of any influence of the old islands. It suggests a dominant southerly transport, but the scatter is large.

Popular secondary inferences of paleoslopes from current-structure orientations in epeiric sea deposits tax one's faith even more than do average flow directions, for currents on shallow shelves today are controlled not nearly so much by regional bottom slopes as by wind directions and topographic obstructions (see for example Klein, 1967). Tides, monsoonal climates, tsunamis, and storms -- especially in low latitudes -- also influence sediment dispersal more than paleoslope. The very shallow North Sea provides one of the most sobering modern examples, for there currents make a 360° rotary circulation parallel to shorelines on all sides.

Cross stratification orientation studies of epeiric sea and some eolian deposits are more difficult to perform and to interpret than for any other

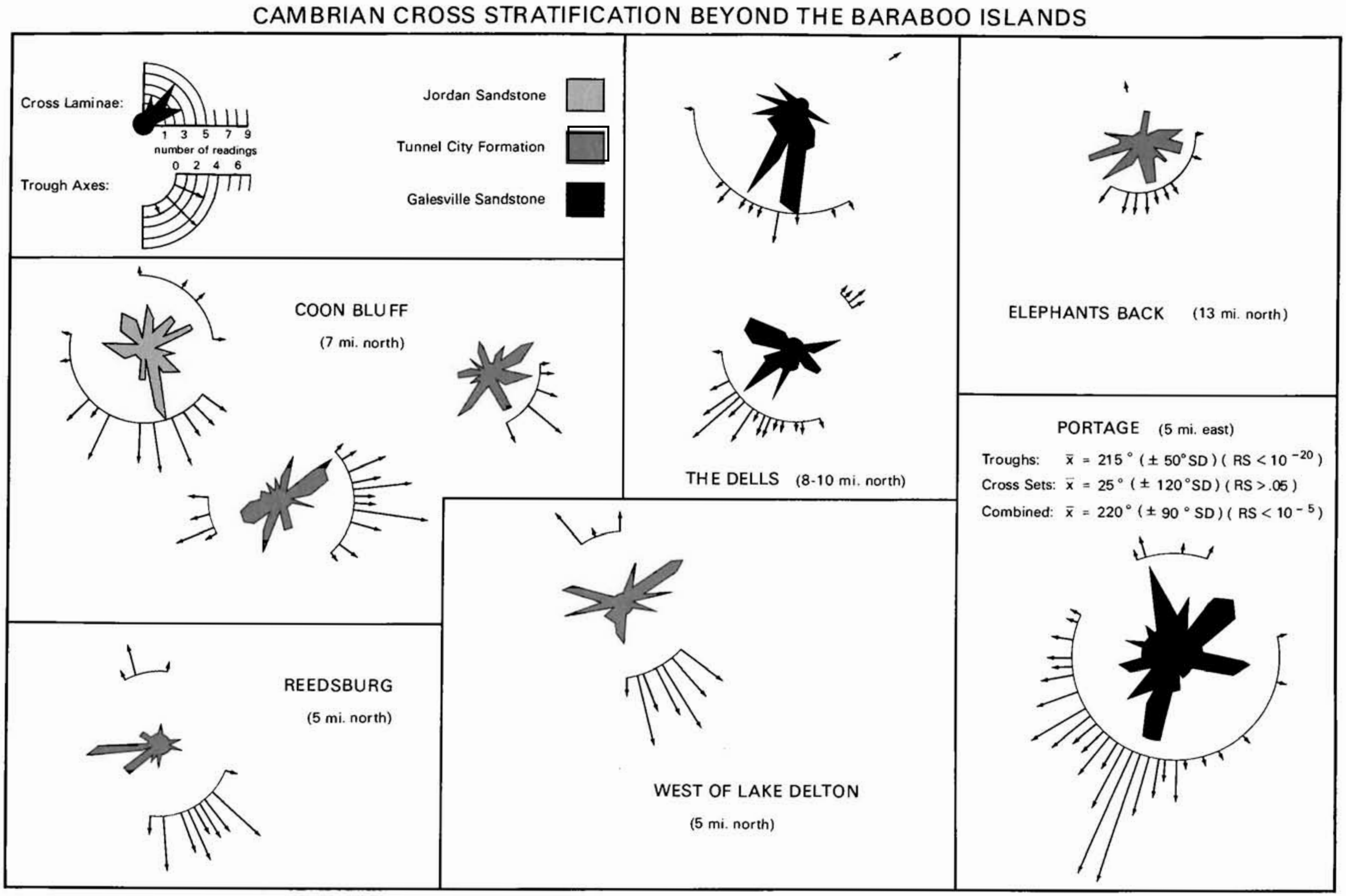

Figure 12. Representative cross bedding orientations north of the Baraboo Ranges. Although dispersion is fairly large, a north-to-south preferred orientation is evident. (Hamblin, 1961; Emrich, 1966). This data provides a standard for comparison of that from among and south of the quartzite islands, which shows much greater scatter (see Pl. VII).

types of deposits. Besides the above influences, the Coriolis Effect at certain latitude and depth combinations produces amarked deflection of wind-driven currents from the surface down to the bottom where current structures form; at some critical depth there is a complete reversal of direction (Von Arx, 1962, p. 113). Because of these complications, grain-size gradients and gross regional facies patterns are the best indicators of regional sediment dispersal in many deposits. As noted above, it appears from Roshardt's work (1965) that no preferred dispersal can be proven for the St. Peter Sandstone in southern Wisconsin from current structures, but Dapples (1955) demonstrated a southwestwardly St. Peter Sandstone dispersal from regional facies and grain-size trends. Hoffman (1953) reported from well samples finer Cambrian sediments concentrated on the southwestern sides of old Precambrian islands in Wisconsin, and he interpreted this to indicate the leeward direction. I have not been able to confirm this data from surface outcrops south of the Baraboo hills, however.

Sediments supplied from local islands in the Cambrian epeiric sea would provide an indicator of overall average sediment transport independent of, and less ambiguous than, current structures. Easily recognized red quartzite gravel is as good a "tag" in the Cambrian strata as the painted or radioactive particles used today by oceanographers. Raasch (1958) first noted that fine Baraboo Quartzite pebbles occur over a large region immediately south of the Baraboo syncline, but are very rare more than a half mile to the north.[7] I consider the pebble train, as he called it, a better indicator of average regional sediment dispersal toward the south than any paleocurrent feature.

As Raasch noted, paleomagnetic studies suggest that Wisconsin lay in the tropics during early Paleozoic time (see Fig. 13). Baraboo would have been in a Trade Winds belt, the paleoequator having extended then from northern Greenland to central Mexico. For the indicated paleomagnetic restoration, the Trade Winds here would have blown from north to south in terms of present coordinates (east to west in inferred Cambrian coordinates). Raasch interprets the quartzite pebble train to be the result of dispersal by shallow currents driven by trade winds. While the sedimentologic data do not uniquely prove the validity of the paleomagnetic interpretation, they are consistent with it.

<u>Sedimentary Environments</u>. The Late Cambrian transgression of the central craton was not a simple, continuous event as evidenced by several vertical and lateral alternations of pure quartz sandstones, glauconitic sandstones, shale, and dolomite. Repetitive lithofacies patterns and missing faunal zones, especially along the Wisconsin Arch where Baraboo is situated (Fig. 1), have suggested the presence of several widespread disconformities supposed to reflect transgressive-regressive oscillations. Such fluctuations seem mandatory to account for the widespread but thin, tabular nature of the sandstones. Dietz (1963) argues convincingly that most sand on the outer parts of modern shelves is relict from former low stands of sea level. He concludes that widespread thin sheets of cross stratified sand must have been deposited in the littoral and inner neritic zones during transgression and regression.

[7]Such pebbles are present at Lake Kegonsa (50 miles south-southeast), at Madison (35 miles southeast), and south of Spring Green (30 miles south-southwest). Pebbles so far from the Baraboo syncline, however, could have come from other islands such as that at Waterloo (Fig. 1).

At least three transgressive "cycles" are recognized at Baraboo, but more are postulated elsewhere (Raasch and Unfer, 1964; Ostrom, 1964). The Eau Claire, Black Earth and Oneota dolomitic units seem to represent the maximum transgressions evidenced around Baraboo. Workers have differed in their designation of the number of "cycles", and of the disconformities assumed to separate them, for example, the Galesville Sandstone is interpreted by many workers (e.g. Wanenmacher, et al, 1934; Berg, 1954; Raasch and Unfer, 1964) as a non-marine, regressive deposit that terminated a Dresbachian transgressive phase. Ostrom (1964), on the other hand, argues by analogy with northern Gulf of Mexico shelf sediments that it is largely a marine sand, which inaugurated a transgressive episode that culminated with deposition of Tunnel City and Black Earth strata.

With the possible exception of the Galesville Sandstone, all of the exposed Cambro-Ordovician sediments around Baraboo are considered to represent either littoral or neritic marine deposits. Fossils and glauconite demonstrate a marine origin, and structures formed by benthonic algae indicate maximum possible water depths on the order of 300 feet, but probably much less. The association of dessication cracks, flat-pebble conglomerates formed by the ripping up of shrinkage polygons, and evidence of the presence of primitive filamentous algae suggest, by analogy, with modern sediments, that, vast tidal and supratidal flats may have existed for long periods. Although burrow structures are common everywhere in Tunnel City and Black Earth strata, their greater abundance near the quartzite islands, suggests the littoral zone as an optimum habitat for the creatures that formed them. Vertical tubes are most common, but some curved and a few horizontal types also are present. Many recent studies elsewhere confirm that such Scolithus-type trace fossils are most characteristic of littoral and shallow neritic zones.

Tunnel City, St. Lawrence, and Oneota strata are the most clearly marine. Regionally the first two possess offshore facies to the southwest of Baraboo, where sands are finer textured, cross stratification is somewhat less prominent, and there is more dolomite and glauconite. The Oneota is the most homogeneous unit, apparently reflecting the regional "high water mark" for the Sauk Sequence of Sloss (1963). The processes and environments of deposition of the Galesville and some of the Jordan and St. Peter sandstones in south-central Wisconsin are not fully established.

As noted above, most earlier workers tended to regard the Galesville as non-marine. Hamblin (1961), however, interpreted it all as very shallow marine (littoral and neritic), while Seeland (1969) argued that much is eolian and some marine. Through a series of student sedimentology class projects, I have attempted to resolve these questions by modern grain-size analysis techniques. Various empirical plots of size statistics (e.g. Passega, 1957; Friedman, 1961; Sahu, 1964) do not provide unique solutions; they consistently suggest either wind or surf as the depositional process for all of the sandstones investigated (i.e. Galesville, Tunnel City, Jordan and St. Peter), which include some obviously marine ones! Wind, which was noted above to have been important during the early history of the sands, is such an efficient sand sorting agent that apparently any lower-energy process such as surf and shallow marine currents, could not significantly change the size distributions. We assume that most of the sands were re-worked and last deposited by marine agents, but apparently all of them were so texturally mature already, and new sediment was added so slowly, that their original size distributions were not modified significantly.

Alternatively, perhaps standard techniques of size analysis are not sensitive enough to discriminate unambiguously among the products of different Cambro-Ordovician processes. The dip angle of cross strata, cited in older literature as diagnostic of aqueous versus wind deposition, is likewise unreliable because of the great geometric complexity of the cross sets. The average inclination of 15^{o}-20^{o} (see Table 10) would tend, if anything, to support aqueous deposition for all of the sandstones.

Other criteria suggest that, at least locally adjacent to the old quartzite hills, the Galesville sands may have been wind deposited. There is the negative evidence of a complete lack of fossils, but more significant is the fact that the Galesville contains negligible quartzite debris even directly adjacent to the buried cliffs (see STOP 2 discussion). Wanenmacher, et al., (1934) argued that this should not be so if the sands were surf deposits, for overlying proven marine strata contain much conglomerate produced by surf attack. At the Upper Narrows (STOP 2), one might argue that the lack of quartzite gravel is not due to wind but rather to the presence there of nearly vertical cliffs and adjacent moderately deep water during Galesville deposition such that waves would not break with force against the cliffs (i.e., a sea wall effect). At other localities, however, the last argument does not seem to hold (e.g. STOP 7), for it appears that Galesville sand simply infiltrated and passively buried angular quartzite blocks as though transported by wind that was not competent to move the large debris. Much of the Galesville sand also is characterized by a distinctive bimodal texture (Fig. 9) that Folk (1968) has argued is indicative of eolian processes, specifically a mixture of dune and inter-dune sands.

Paleogeographic Conclusions. From the preceding critique, we conclude that during Galesville deposition, and perhaps earlier as well, the local strand lay slightly offshore from the quartzite hills so that surf was unable to attack the cliffs. A narrow zone of sand beaches and eolian dunes presumably lay between. Occasional angular blocks of quartzite that tumbled from the cliffs simply were buried in sand without being moved and rounded. Beginning with Tunnel City deposition, transgression brought the strand to the cliffs, where surf eroded, rounded, and dispersed large quantities of quartzite gravel. Erosion prior to Cambrian transgression had produced a complex topography so that there were many valleys among the quartzite hills, which became straits and coves upon inundation (Fig. 13). As transgression proceeded, sedimentation gradually buried the islands. Presence of the "offshore" dolomite lithofacies (Black Earth and Oneota) even within the Baraboo basin suggests that much of the time the main regional shoreline lay far to the north, and that a trade-wind fetch well in excess of 100 miles probably existed during much of latest Cambrian time (see Hamblin, 1961, for regional paleogeographic maps). A maximum depth of water in the vicinity of the islands on the order of 150 to 200 feet can be estimated from initial dips of up to 5^{o}-10^{o} in marine sandstones adjacent to the quartzite islands, but flattening to 1^{o} or less within one-quarter to one-half mile away from the quartzite. The large stromatolites in the Black Earth Dolomite at the old Woods Quarry (2 miles south of the west edge of Baraboo) must have formed in very shallow, clear water less than one-quarter mile from shore.

The presence of clearly-delineated ancient islands in the Baraboo area provides a unique opportunity to make detailed environmental studies of ancient sediments. Generally, paleogeography is the major "unknown" in the sedimentologist's equations. At Baraboo, however, where it is a "known", we can investigate detailed variations of other parameters, such as the distribution of size and shape of gravels and of current structures with respect to proximity to the islands, position inside or outside the central lagoon, and position to windward or leeward. The probable effects of storms also can be investigated, and the vigor of surf can be estimated.

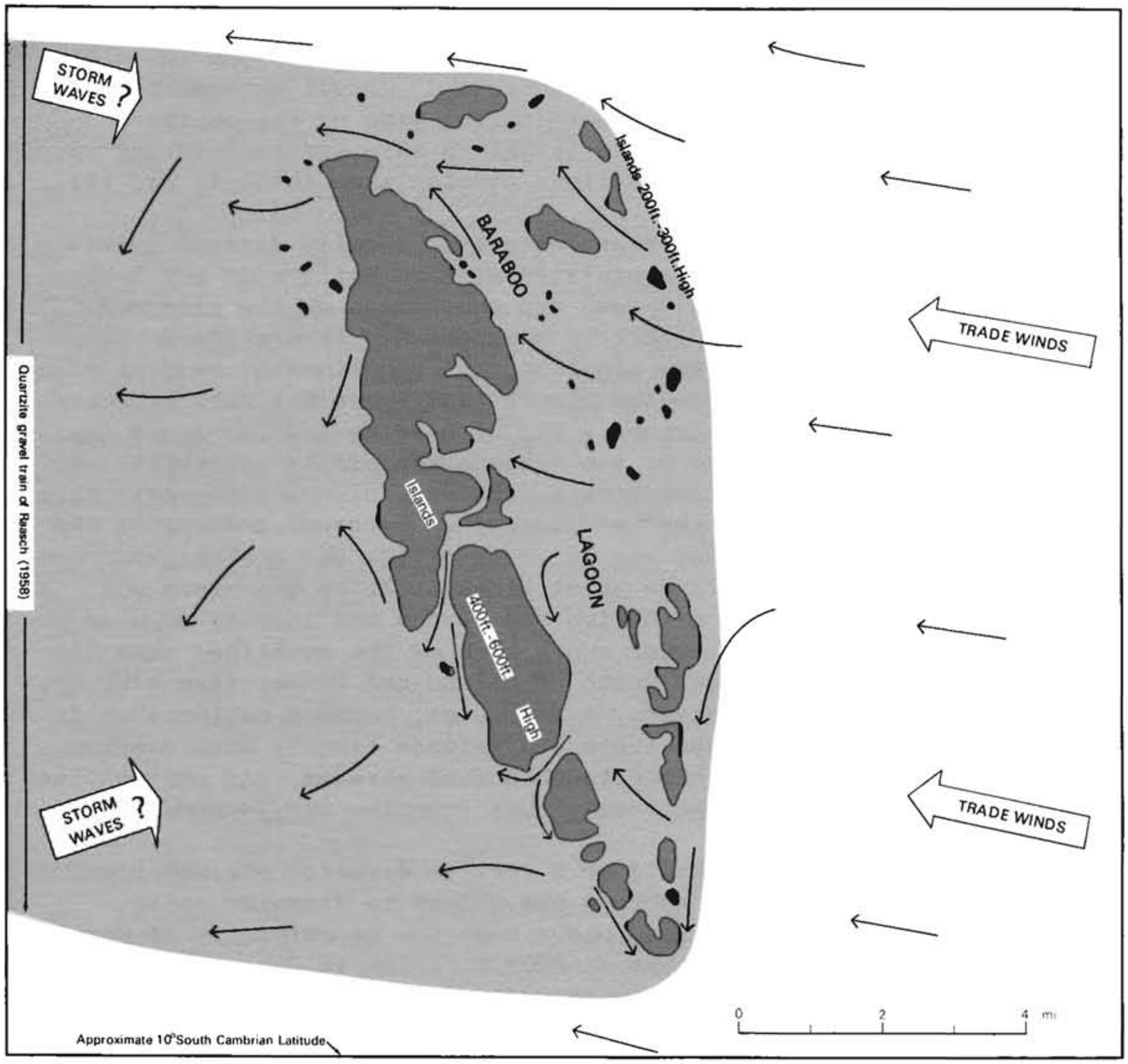

Figure 13. Hypothetical paleogeographic map of the Baraboo islands near the middle of Late Cambrian time (i.e. Franconian), showing inferred normal current patterns and some possible storm tracks. Map is drawn according to paleomagnetic restorations for North America, which suggest that the Baraboo region lay at about 10° South latitude and was rotated 90° relative to the paleoequator. The gravel train represents fine quartzite pebbles dispersed toward the left by trade-wind-driven "normal" currents after these were intermittently swept from the shores of the islands by "rare" storm waves. (Adapted in part from Raasch, 1958).

<u>Conglomerate Distribution</u>. Plate VI shows the distribution of clasts of conglomerate in Cambrian strata with respect to the present Precambrian outcrop, which also is a fair approximation of island configuration during Franconian and Trempealeauan time; most of the data shown represent conglomerate of those ages. Note that both relative size and rounding of clasts are indicated. Besides showing a general concentration of largest boulders nearest buried quartzite contacts and a crudely exponential diminution of size with distance from those contacts, the largest boulders and the greatest lateral extent of conglomerate occur along the outer southern side of the old island complex. Inside the basin and on the outer north side, nearly all cobbles and boulders (regardless of size) are confined to the immediate proximity of quartzite contacts. On the other hand, gravel extends farther to the south of the islands. On the inner (north) side of the southern islands, cobble and boulder conglomerate is just a thin veneer covering sloping quartzite surfaces where Cambrian formations overlap them (STOP 7; Pl. II).

As noted above, the greater southward extent of locally-derived quartzite debris is consistent with the postulated average pattern of north- to-south paleocurrents, but the great volume and coarseness on the presumed leeward side of the old quartzite atoll is unexpected. Several factors seem involved. On both limbs of the syncline, intersections of bedding with joints, controlled by cleavage, created blocks ready-made for surf attack. But the combination of a gentle northerly dip of bedding and the steep south-dipping joints made the south face of the islands especially susceptible to wastage by gravity collapse and wave attack. Bedding planes apparently were the most susceptible surfaces to surf erosion, thus vertical bedding in the north limb of the syncline produced nearly sheer cliffs and a straighter coast less susceptible to collapse of large joint blocks. It is therefore not surprising to find more evidence of buried sea stacks and immense angular quartzite talus blocks along the outer south side of the syncline; such features are especially well displayed north of Leland and Denzer (see STOP 8). On the summits of the northern islands, a few large, rounded boulders up to 8 feet in diameter are found. When these old islands finally were overrun by surf, large joint blocks formerly untouched were abraded, and the smaller of these were spread as far as a half-mile away from the old islands.

In general, boulders larger than 4 or 5 feet in diameter are not rounded, although locally a few exceptions up to 8 and 9 feet in diameter exist. Angular blocks much larger than 5 feet occur near the unconformity at many localities. The largest found is that at STOP 8, which is more than 25 feet long, but several from 10 to 20 feet in longest dimension are scattered elsewhere along the south side of the South Range (Pl. VI). Clearly most of these very large angular blocks could not be moved by Cambrian surf even during storms. Probably most originated by tumbling and sliding from sea cliffs and stacks into the sea or onto beaches where they were simply buried with little or no further movement. Wave erosion was accelerated wherever cavitation or sudden compression of air in cracks in the cliffs occurred, especially along low-dipping bedding planes. Undercutting of cliffs doubtless caused collapse of large masses of quartzite, while scour of sand or fine gravel from beneath blocks already resting on beaches would cause tipping and splitting by impacts. Rounding by such processes, however, would be minor.

Excellent rounding and presence of percussion marks on many boulders up to 4 feet in diameter attest to vigorous collisions. I infer that abrasion of the boulders was accomplished chiefly in the following ways:

1. Grinding and chipping between the boulders themselves through movement induced by impact of large breaking waves and by backwash.
2. Barraging of stationary larger boulders by small quartzite pebbles and cobbles thrown against them by waves -- anvil effect.
3. Abrasion by sand carried in suspension in waves -- sand blast.
4. Some abrasion in streams draining the islands (but valleys could have been only 1-2 miles long with gradients of about 200 feet per mile).

Presence of a broad spectrum of clast sizes and the clear discontinuity in rounding versus boulder size provides a rare opportunity to estimate the competence of ancient ocean waves that pounded the old islands. If rounding of the boulders was accomplished largely by tumbling of the blocks themselves (1. above), then it would appear that Cambrian breaking waves rather commonly could move blocks up to one ton, and rarely those weighing several tons (Table 9). But blocks weighing more than 10 tons apparently were never tumbled. It is important to note that the nature of the bed upon which boulders rested would greatly affect their ease of movement. A pile of interlocked boulders or scattered boulders on sand would be harder to rotate than scattered boulders resting on a more or less smooth, hard surface. Significantly, the largest rounded boulders (4-10 feet in diameter) almost exclusively are found directly upon quartzite bedrock at the sub-Cambrian unconformity.

While little quantitative empirical data exists on the competence of waves in nature to move large boulders, some observations of catastrophic river floods and empirical equations developed from flume experiments provide a basis for order-of-magnitude estimates of water velocities necessary to pivot and round the quartzite boulders. It is well known that as oscillatory waves enter shoalwaters, they are converted to solitary waves. When they break, they become translation waves in which much of the water mass is carried landward in a manner closely analogous to an ordinary unidirectional current. Similarly, the return flow of water after surf has struck a beach (the backwash) also can be treated as a unidirectional current. To rotate a quartzite cube 3 feet on a side, a threshold velocity of about 10 miles per hour is required (Table 9). As boulders become more rounded, of course, this velocity would decrease somewhat, and the velocity on a smooth, hard surface would be less than a rough surface. Maximum velocities that have been measured systematically for large rivers are on the order of 20 mph (Leopold, et al, 1964), but velocities at least twice as great are known for catastrophic floods (Malde, 1968). From Table 9, we conclude that the maximum water velocity in the largest Cambrian breaking (translation) waves at Baraboo would have to have been between 15 and 20 mph in order to move the largest well-rounded boulders.

Next we must ask if breaking-wave velocities of the magnitude suggested in Table 9 would be possible in the shallow Cambrian epeiric sea. Today the largest waves known in the deep oceans are generated in the circum-Antarctic seas where unusually persistent and strong winds -- the Roaring Forties -- produce deep-water waves at least 50 feet high with wave lengths of the order of 1,800 feet. When such waves move across the Pacific, they produce translational surf waves up to 40 feet high under the unusually favorable circumstances of very long fetch and narrow shelves along certain coastlines (e.g. Hawaii, southern Australia and South Africa). In the shallow Cambrian epeiric sea, however, with assumed continental shelf depths less than 600 feet extending for at least 500 miles in all directions from the Baraboo region, such large deep-water waves would be damped by conversion to shallow-water ones as they felt bottom at depths less than one-half their

TABLE 9. ESTIMATES OF CAMBRIAN DYNAMIC WAVE PARAMETERS FROM QUARTZITE BOULDER PROPERTIES.

QUARTZITE BOULDER SIZE (Diameter)	APPROX. MASS (Tons)	THRESHOLD VELOCITY (V_t)*		APPROX. DEPTH AND HEIGHT FOR BREAKING TRANSLATION WAVES OF SPECIFIED MAXIMUM WATER VELOCITY**	
		fps	mph	D_b	H_b
3.3 ft. (1 meter)	0.35	16	11	16 ft.	13 ft.
4 ft. (1.2 m) (Maximum commonly rounded)	0.7	17 - 18	12	20	16
5 ft. (1.5m)	1.2	19 - 20	13 - 14	25	20
10 ft. (3.3m) (Absolute maximum rounded)	10	24 - 28	16 - 19	49	38
12 ft. (3.6 m)	16	26 - 31	18 - 21	60	47
20 ft. (6 m)	25	27 - 42	18 - 29	110	86

* Water velocity required to initiate rolling or sliding from two empirical formulae (ranges represent values given by each):

<u>USBR</u> (Peterka, et al, 1956): $V_t = 9 \times d^{\frac{1}{2}}$ <u>Nevin-Hjulstrom</u> (Fahnestock, 1963): $V_t = 10 \times d^{\frac{1}{2.6}}$

** Maximum water velocity on bottom for breaking solitary wave (see Wiegel, 1964 for fuller treatment):

$$U_b^2 = \frac{gD_b}{2} \quad \text{and} \quad H_b = .78D_b$$

wave lengths. Thus it is improbable that surf of the maximum magnitude cited above could have broken frequently upon the Baraboo islands unless our estimates of depth of the Cambrian Sea are too small. Yet, large, very destructive waves generated by storms in the North Atlantic are legendary, producing stories of water engulfing lighthouses more than 100 feet high and of large cobbles being hurled through lighthouse windows; breakwater blocks weighing hundreds of tons also have been moved (Bascom (1964). But most of the sites where these dramatic phenomena have been observed are nearer to the edge of the continental shelf than was the Baraboo region. Moreover, such phenomena probably involve unusual local concentrations of energy, thus represent events of very low probability. A more realistic indication of large waves around the Cambrian islands may be provided by the New England coast, where breakers on the order of 20 feet high have been observed at Cape Cod. They are formed by winter "Nor'easters" blowing steadily for several days over the Atlantic shelf (R. L. Miller, personal communication). Such waves seemingly would be competent to move the rounded quartzite boulders up to 4 feet in diameter (Table 9).

From the above comparisons, it appears that very unusual conditions would be required to roll the largest rounded quartzite boulders at Baraboo (e.g. those greater than 5 feet in diameter and weighing more than one ton). It suggests either that much of the rounding of these blocks was accomplished through bombardment by cobbles, pebbles and sand, or that local topographic circumstances may have produced unusual concentrations of wave energy sufficient to tumble them on rare occasions. Still, we can not rule out rare, severe storm waves and tsunamis. If the tropical latitude suggested by paleomagnetic evidence is correct (Fig. 13), major tropical storms such as hurricanes with wind velocities on the order of 100 miles per hour could be expected. And storms like the New England "Nor'easters", which sustain high winds for longer periods, even if of lesser velocity, actually could produce larger waves than do hurricanes. Violent storms also raise the sea surface in local surges or "tidal waves". Under certain conditions these may advance as bores as in the Bay of Bengal where initial waves 30-40 feet high have wrecked havoc on the coast (Bascom, 1964). Finally, catastrophic tsunami waves must be considered. The tectonic stability of the craton itself indicates that such waves would have been generated at least 500-600 miles to the south or east in the Appalachian mobile belt; travel of such waves across the shallow epeiric sea would have damped them somewhat.

Waves produced by prevailing trade winds would have been refracted around the island atoll, causing much dissipation of energy before breaking on the present south faces of the islands. Yet, that is where there seems to have been the greatest absolute wave energy. To explain this seeming anomaly, I postulate that many major storm waves or tsunamis approached from directions in the southerly half of the compass in terms of present map coordinates (Fig. 13). Such a multiplicity of wave approach directions is common today on many coasts. Hawaii provides a useful example because it lies in the trade winds belt where steady waves of moderate power approach from the east. Less frequent but more powerful storm waves approach from the south and northwest. Hawaiian beach sand textures on different sides of the islands reflect these differences (Moberley, et al., 1965).

Episodic behavior of powerful Cambrian wave and current action is reflected at several localities around Baraboo. Most notable is Parfreys Glen (STOP 11), where discrete layers of conglomerate with clasts up to 4 feet in diameter alternate with nearly pebble-free cross stratified sandstone at a position at least one-quarter mile seaward from the ancient sea cliff. Pebble and cobble layers extending still farther seaward at other localities (Pl. VI) also attest to intermittent, unusually competent processes (particularly during Franconian time). Such occurrences are interpreted as the results of episodic strong storms, which moved coarse material unusually far during brief intervals of time. The 4 feet diameter boulders at Parfreys Glen and at Hemlock Draw -- both overlain and underlain by sandstone -- prove that, at least on rare occasions, waves or currents occurred that could roll boulders weighing nearly one ton some distance. Breaking waves tend to push sand and gravel shoreward, therefore these large boulders probably were swept offshore by the backwash of exceptionally large waves such as the bores noted above in the Bay of Bengal. Whenever normal conditions resumed, dominated by the inferred less vigorous north-to-south current regime, only sand and small pebbles up to 1 inch in diameter could be transported farther than about 1 mile from shore (although exceptional 4 inch cobbles occur as much as two miles from present quartzite outcrops near Leland). Apparently it was under the latter regime that the fine quartzite fragments -- probably originally made available by storms -- were dispersed for many miles to the south (see discussion for STOP 11).

There is additional evidence of vigorous (storm?) agitation induced from the south. Unusual stirring of the deeper sea bottom southwest of the Baraboo islands, where glauconite formed in abundance, seems required to have transported that mineral northeastward because it is found around Baraboo in the Tunnel City Formation; in rare cases, glauconite became concentrated in placer-like deposits. Flat pebble rip-up conglomerates widespread in Tunnel City strata also represent vigorous episodic scour of newly-deposited, cohesive sediments, most likely by storm-induced agitation. Finally, many of the most prominent and laterally extensive flat truncation planes in some of the cross stratified sandstones may represent episodic planing off by storm waves of the submarine sand dunes within which the cross sets had formed.

<u>Analysis of Paleocurrents</u>. The types of features formed by currents and the problems of paleocurrent interpretation from them in southern Wisconsin were summarized above. Previous authors (Farkas, 1960; Hamblin, 1961; Emrich, 1966) have noted the predominance of trough type cross stratification in the Cambrian sandstones, but the variability of this form has received little attention. It is somewhat surprising in view of the fact that this type may be the most prevalent in nature. Only the very small-scale cross lamination associated with simple ripplemarks typical of fine sands is commonly of the "textbook" planar type (Fig. 10), although some important exceptions are known on a larger scale (e.g. Meckel, 1967).

Several classifications of cross stratification have been proposed, some of which are so complex as to be of little practical value, and the principle that the "least possible classification is the best classification" seems to apply. Because the geometric variations represent a continuous spectrum, I prefer the simple, flexible scheme shown in Figure 10, otherwise, a separate formal pigeon-hole must be erected for practically every cross set encountered. The most significant parameters are <u>shape and orientation of the cross laminae</u> themselves, <u>shape and orientation of the</u>

truncation surfaces separating cross sets, the amplitude of sets, the maximum angles of inclination of the sets, and the plunge of trough axes (if present). Quantitative dimensions can be specified for each of these parameters (Table 10) supplemented as necessary by descriptive notations (e.g. tangency of sets, asymmetry of laminae within troughs, grain-size variations within laminae, etc.).

As noted above, Galesville cross stratification in the Baraboo-Dells region is of greater amplitude than that of the other formations (Fig. 9). Also, parallel or near-parallel truncations, while not universal, tend to be most common in the Galesville. They are practically unknown in the Tunnel City, whereas the Jordan has mixed types. Hamblin (1961) notes that the scale of trough sets in "Dresbach" sandstones (including the Galesville) are largest in northern Michigan, where troughs up to 600 feet wide occur, and troughs are present at the Wisconsin Dells, around Rock Springs (STOP 2; Supplementary STOP C), and at Reedsburg as noted above. Trough cross sets in the Tunnel City and Jordan Formations are uniformly of smaller scale. Many are asymmetrical in cross section, which, together with prevalence of wedge truncations, produces a chaotic looking mosaic of inclined surfaces in outcrops of these two formations (Fig. 9).

No exhaustive evaluation of the large dispersion of cross stratification orientation data has been attempted as yet for the lower Paleozoic sandstones of the Mississippi Valley region. Emrich (1966) made one study of local variations through time in the type section of the Galesville, where he found little variation through 150 feet of strata. Farkas (1960) believed that there was greater dispersion of cross set orientations in the Tunnel City Formation south of the Baraboo islands than elsewhere, but the contrast is not especially striking on his map. The standard deviation for a uniform (i.e. non-oriented) distribution is 104°, thus any value greater than this cannot be considered to show a statistically significant orientation. The published range of standard deviations from $\pm$ 81° to $\pm$ 113° (Table 10) suggests a very low degree of regional preferred orientation. Such large standard deviation values are by no means exceptional for data from cross strata. In general, linear features (e. g. flutes, oriented pebbles or graptolites, etc.) yield data with smaller deviations than do cross sets. This is due both to inherently greater geometric complexity of the latter and also to greater difficulties in determining true orientations from three-dimensional surfaces than from linear features. Meckel (1967) showed that tabular cross sets may have only half as much dispersion as associated trough sets, therefore the problem in Wisconsin is accentuated by the fact that practically all of the data are from trough type cross stratification. Obviously if trough-axis orientations were observed instead of many cross sets disposed around the limbs of the troughs, a smaller dispersion would result (Fig. 12). A more faithful current indication also should be obtained, for elongate trough axes tend to parallel the flow direction. In the literature, only Hamblin (1961) seems to have made this distinction for Cambrian sandstones heretofore; the effect is indicated by a 50 percent smaller standard deviation for his "Franconia" data than for that of other workers (Table 10).

Attempts have been made to characterize empirically depositional

TABLE 10: REGIONAL CROSS STRATIFICATION DATA FOR WESTERN WISCONSIN AND NORTHERN ILLINOIS

FORMATION	AMPLITUDE OF SETS	AVERAGE DIP ANGLE OF SETS	MEAN DIP DIRECTIONS* (±1 standard dev.)	VARIANCES (S^2) OF DIP DIRECTIONS
St. Peter Sandstone				
(Roshardt, 1965) Southwestern Wisconsin	(R = 1" - 6')	15° (R = 1 - 40°)	185° (±132) (vector magnitude = 8%)	17,424
(Potter & Pryor, 1961)	------	------	220° (Illinois)	--------
Jordan Sandstone				
(Farkas, 1960)	------	------	199° (±108)	11,664
(This study)	4" (R = 1 - 10")	19° (R = 8 - 30°)	--------	--------
Tunnel City Formation ("Franconia")				
(Farkas, 1960)	10" (R = 1 - 32")	18° (R = 2 - 36°)	151° (±81)	6,560
(Hamblin, 1961)	(R = 3 - 8")	(R = 2 - 36°)	Southwest (±44)	2,400
(This Study)	5" (R = 2 - 10")	19° (R = 5 - 38°)	------	--------
Ironton and Galesville Sandstones (combined)				
(Emrich, 1966)	4" (R = 1 - 144")	8 - 16° (R = 0 - 36°)	226° (±113)	12,819
Galesville Sandstone				
(Farkas, 1960)	------	------	253° (±81)	6,500
(This Study)	19" (R = 2 - 120")	16° (R = 4 - 38°)	------	--------
"Dresbach" of Hamblin (1961) (includes Galesville)	"Large scale"	------	South	--------

R -- Range of values where reported.

*Chiefly attitudes of trough limbs, although Hamblin (1961) included trough-axis plunge directions wherever observable. Note that standard deviation for a uniform or non-oriented circular distribution is ±104° (variance is 10,800); any larger value indicates no significant preferred orientation.

processes or "environments" by the relative dispersion of cross strata as measured by standard deviations (variance, which is the square of standard deviation, can be used if one wishes to magnify differences). A summary by Potter and Pettijohn (1963, Table 4-2) shows values between 42° and 88° for many fluvial and deltaic sands, between $53^{\circ}$and 77° for a few eolian sands, and between 71° and 86° for a few shallow marine ones (excluding the Cambro-Ordovician sandstones of the Great Lakes region). Wermund (1965) determined a standard deviation of 101° for an Eocene neritic bar sand. Standard deviation ranges for Wisconsin sandstones exceed all of these, verifying my earlier claim that their cross stratification is indeed exceptionally dispersed.

Paleocurrent analyses have proliferated in the past 15 years, and a variety of techniques have been developed (see Potter and Pettijohn, 1963). The object of most such studies heretofore has been to define regional average sediment dispersal patterns, but certain of the now rather standard practices used for regional reconnaissance studies can be misleading if subtle, local patterns are of interest. Calculated mean directions, especially if not accompanied by some indication of magnitude of dispersion of the data, can be especially deceptive. In some cases the mean direction is purely a mathematical artifact.[8] Lumping of data either stratigraphically or geographically also can obscure geologically significant information.

Some years ago I undertook an analysis of trough cross stratification that would investigate its complexity of orientations and, hopefully, might reveal meaningful geologic information that I suspected was being lost. Several Cambro-Ordovician sandstones of southern Wisconsin were investigated as class projects. The Baraboo area offered an unusual opportunity to document the great complexity of currents to be expected around an old island complex. I wished to see if by careful measurement and avoidance of the "averaging syndrome" (which, of course, is unavoidable in regional studies) one actually could map some of the local complexities in ancient sandstones. From Farkas' work and that of Roshardt (1965) in the St. Peter Sandstone, it seemed entirely possible that I would find a hopelessly random pattern at most stations around Baraboo, but that too could be a useful demonstration. This fear seemed fully borne out by the rude experience at the outset of the study of detecting what appeared to be portions of oppositely-plunging troughs in the same outcrop (Fig. 11). Later fully exposed and complete doubly-plunging troughs were found at the Dells.

Results of the Baraboo cross bedding study are summarized in Plate VII. The data are plotted in number-frequency histograms to avoid the possible aberrations of percentage or close-number data, and for more realistic comparisons among samples of different sizes. At most localities at least 30 cross set readings were taken, this being commonly designated by statisticians as a large sample. Sampling was done only where three dimensional orienta-

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8 A mean azimuth may be both geologically and statistically meaningless, especially if the data are polymodal as is expected with trough cross sets (Fig. 12). For widely dispersed data, the vector mean (Curray, 1956) is superior to a simple arithmetic mean, but even it may be nullified by polymodality. Certain measures of dispersion also can be misleading. The commonly used confidence interval is so sensitive to sample size that the interval is narrowed drastically simply by greatly increasing the number of observations even while true dispersion remains unchanged! Either the standard deviation or a polar histogram of the total sample is, therefore, preferable.

tions and trough axes could be seen through an average vertical interval of about 10-20 feet. Care was taken to measure cross sets representing all portions of the troughs wherever possible. Replications at several localities yielded similar results. Cross set orientations are shown in the histograms and, where observable, trough axes are shown for comparison by arrows proportional in length to the number of readings with a given azimuth. A very great complexity of orientations of both features is immediately apparent. Close examination reveals that a stronger orientation exists for the trough-axis plunges, which are taken to be more useful for paleocurrent interpretation than cross sets regardless of how large a sampling is made of the latter. At a number of localities, the expected bi-modal cross set orientations perpendicular to the trough axes are apparent, but in many other cases, no such idealized relation exists. Asymmetry of sets is revealed in some cases, which also can be observed visually in the field, and a larger number of trough-head than trough-flank cross set readings are revealed in other cases. Data from north of the Baraboo "atoll", where the islands should have had little or no influence on currents (Fig. 12), shows a general southerly preferred orientation, although dispersion is large there too. Near the islands, there is some evidence of deflection of currents parallel to old shores, as expected, but it is by no means universal. Inside the Baraboo basin there is great variation, but at a number of localities currents apparently flowed south across the lagoon directly toward the southern islands. South of the "atoll", great dispersion is the rule near the islands, but a weak southeasterly orientation seems to exist more than 2 miles away (especially southeast of the Wisconsin River). Figure 13 summarizes these inferences, but it is apparent that a still more detailed bed-by-bed analysis of very small areas would be necessary to realize fully the original objectives.

Fluid Mechanics Interpretations. For many years, most large-scale trough and wedge cross stratification was interpreted uncritically as of eolian origin, and eolian dunes were the only features cited as modern analogues. Yet, large sand waves or dunes were known in some rivers and estuaries 100 years ago, and studies of marine sediments have revealed recently that large-amplitude submarine dunes are common on shallow banks (e.g. Bahamas and Georges Banks) and in several straits (e.g. the English Channel). Flume studies also verify that relatively large and complex dune forms can be produced by flowing water (see Simons, *et al*, 1965, Fig. 6). Thus, cross stratification alone is a reflection of dune forms, and is not a diagnostic process or environmental indicator. Determination of the final environment of deposition of many lower Paleozoic cross-bedded sandstones of the Upper Mississippi Valley is particularly difficult because of the probability of mixing of eolian, surf, and shallow-marine products. The trough cross bedding in sandstones of the Tunnel City Formation is duplicated both in scale and form in near-shore Pleistocene marine sands that I have seen on elevated marine terraces in Oregon. Both appear to be the internal relict structures of submarine dunes or sand waves on a smaller scale than those on the modern Grand Banks (Jordan, 1962) or around Great Britain (Stride, 1963). The Galesville, Jordan, and St. Peter sandstones may include also some coastal eolian dunes more or less reworked during transgression much as has occurred on the Grand Banks during the late Pleistocene rise of sea level.

Understanding of ripple and dune forms, of which cross stratification is the preserved internal remnant, has been greatly augmented by experimental studies in recent years. Ever since G. K. Gilbert's milestone experiments in 1941, a sequence of bed forms from smooth to rippled back to smooth has been known to accompany a velocity increase. At certain combinations of discharge, depth, and granular sediment, the fluid-sediment interface becomes unstable, and rippling occurs much like the Helmholtz Effect between two immiscible flowing fluids. When rippling commences, the flow has become turbulent, and as velocity and turbulence increase, the scale of rippling tends to increase as well until finally <u>all</u> bedforms disappear at high velocity. With long-continued flow, straight ripples perpendicular to flow change to complex linguoid and lunate ridge forms roughly parallel to flow (Simons, <u>et al</u>. 1965; Allen, 1963). Cross stratification, representing as it does fossil slip faces of ripples and dunes, changes correspondingly from the planar type to increasingly complex trough forms. Trough cross sets apparently represent the fill from the head or from one or both sides in elongate depressions between dune ridges migrating more or less parallel to flow. Flow near the bed of the troughs may involve very complex helical turbulence roughly parallel to flow. Apparently the ridges are too ephemeral to be preserved in ancient sediments so that only the truncated trough fillings survive.

Doubly-plunging troughs around Baraboo apparently represent the fill of elongate depressions between coalescing elongate dune ridges and mounds (i.e. inter-dune depressions closed at both ends). Such troughs may not be so unusual as they at first seem, and should be sought elsewhere. The extreme complexity of the trough cross sets around Baraboo probably reflects oscillatory agitation of the bottom by waves as well as rippling by current flow near the old islands. Recent experiments with oscillatory flow produced very complex three-dimensional equilibrium dunes with ill-defined (i.e. not strongly linear) crests and troughs (Carstens, <u>et al</u>, 1969). The troughs or depressions so developed are geometrically very similar to many of the Cambrian trough cross sets. Oscillatory wave and tide-produced cross stratification may be common in many very shallow-marine sandstones, and where this is so, conventional paleocurrent determination is obscured. Tidal currents would have influenced the bedforms around the Baraboo islands. Cross bedding orientation diagrams for some modern tidal sediments (e.g. Klein, 1967) resemble closely the polymodal patterns for many of the Cambrian localities. At the time of writing, however, criteria in the sandstones of tidal influences had not been fully evaluated.

From the fact that the maximum size clast transported more than a mile from the ancient islands was only about 1 inch in diameter (and generally less rounded than associated sand), even though a wide range of gravel sizes was available, we can estimate the <u>maximum</u> velocity of bottom currents offshore as of the order of 5-10 feet per second (3-7 mph). Projecting the initial dips of Cambrian littoral sands, most of which are on the order of 5 degrees, we estimate that the maximum depth of water a mile or more offshore was about 100 to 150 feet. Intuition alone tells us that, whenever cross stratified sands were deposited, the flow was turbulent, but from the velocity and depth estimates, we can estimate the order of magnitude of turbulence. The Reynolds Number (N_r), a dimensionless quantity that relates inertial to viscous forces in fluid flow, is useful for defining relative turbulence (turbulent flow develops above $N_r = 600$). For the conditions estimated here, N_r would be on the order of 300,000 to 500,000, indicating a high degree of turbulence. It is dangerous to extrapolate, except in the most tentative way, the results from recent flume studies of flow (e.g. Simons, <u>et al</u>, 1965) to large-scale marine current regimes; such extrapolation is all the more difficult if oscillatory flow also were important, as surely was the case around Baraboo. Nonetheless, from such studies, it seems clear that ripples and dunes would be the predicted bed forms under the conditions specified, and current flow would be within the "lower flow regime" as defined by such studies.

Widespread smooth bed surfaces reflected by persistent parallel truncations of cross sets may have developed during storms when fluid dynamic conditions temporarily favored extensive scour, and dune forms could not persist. Commonly there are thin, coarse sand or very fine pebble lag concentrations along these truncations. Probably the flow at such times was in the "upper flow regime" of Simons, et al, (1965). After conditions returned to normal, dunes again formed and produced new sets of cross strata. Stokes (1968) suggested that such widespread truncations also could develop by deflation of eolian dunes down to a very shallow water table where moisture would give the sand sufficient cohesion to resist scour. Perhaps this mechanism was important in producing at least some of the planar truncations around Baraboo, notably those in the Galesville Sandstone.

Importance of Rare Events

The Baraboo district is an unusually good laboratory for demonstrating paleogeography and for investigating ancient sedimentary processes in an epeiric sea. The complexity of cross stratification orientations is almost without precedent in the current literature, but surely it is not unique in nature. Detailed studies in other very shallow-marine sandstones can be expected to show similar patterns.

There is much evidence of the importance of storms on sedimentation around the Baraboo islands, which raises the fundamental question of whether the sedimentary record is one largely of average, day-to-day processes -- a kind of strict uniformitarian viewpoint -- or of less frequent, more violent events (Gretener, 1967). Judging from many modern coasts and continental shelves, so-called rare events probably are responsible for most of the geologic work evidenced in the sedimentary record (e.g. Hayes, 1967). Following this more "catastrophic" point of view, one might regard every major stratification plane in the Cambrian clastic sediments as reflecting such a rare event. In interpreting the paleocurrent and paleowind data, one also must consider the rare event. Under certain conditions, as in the Mojave Desert (Sharp, 1966), storm winds and other factors leave more imprint on dunes than do prevailing winds. In other cases, however, as on the Oregon coast, prevailing summer northwest winds are so strong and steady that they impose a greater influence on coastal dunes than do strong but brief southern winter storm winds (Cooper, 1958). The great amount of winter moisture in the latter region may stabilize the dunes seasonally so that they are little-affected by the storms.

Although it is challenging to the imagination, there is much circumstantial evidence of independent kinds to suggest that 500 million years ago, the Baraboo islands lay in a Trade Winds belt about 10 degrees south of the equator, and that they were pounded periodically by vigorous waves, which very likely were produced by tropical storms (Fig. 13). If the implications of a shorter length of day in Cambrian time suggested by paleontologic criteria (Wells, 1963) and a suggested closer proximity of the moon to the earth prove to be correct, then seemingly the Baraboo islands may also have experienced very large tidal current effects. Perhaps we can look forward to further insights into ancient sedimentologic phenomena from this remarkable natural laboratory.

GLACIAL GEOLOGY
(Black)

Introduction

Wisconsin was among the first of the states to have an over-all map of its glacial deposits, including drifts of different ages (Chamberlin, 1882, Pl. II). That map clearly portrayed a boundary between the "Second", or younger, Drift and the Driftless Area, west of Devils Lake in the Baraboo area. Later the boundary was moved eastward on the relatively detailed maps and in the reports by Salisbury and Atwood (1900), Trowbridge (1917), and Alden (1918). Those reports provided us with lucid accounts, including geomorphology, of an area now famous for its intricate pattern of moraines and for its variety of other features. Chamberlin (1878 and 1883 a and b), Weidman (1904), Smith (1931), Martin (1932), Thwaites (1935, 1958, and 1960), Bretz (1950), Powers (1960), Black (1964 and 1967-68), Black, *et al*. (1965, p. 76-78), and Bachhuber (1966), among others, added new data on the geomorphology and glacial history of the area.

No field study of the geomorphology or glacial geology of the area of Plate 1 has been attempted for this report. Only the position of the prominent terminal moraine (Pl. I) of Late Woodfordian (Cary) age (perhaps 13,000--16,000 radiocarbon years B.P.) was refined by aerial photographic interpretation and limited field checks. Thus, this note largely is a synthesis of the literature. A brief description of the terminal moraine is followed in turn by descriptions of features associated with that moraine, by evidence of greater deployment of the ice, and by a synopsis of the glacial history, including the geomorphic evolution of some major land forms.

Terminal Moraine of Late Woodfordian (Cary) Age

The terminal moraine (Fig. 14a) of Late Woodfordian (Cary) age extends generally southward in a series of arcs from the vicinity of the Wisconsin Dells to the vicinity of Sauk City. The locally-crenulated border in the Baraboo Range is the most striking and clearest example to be found along the entire former margin of the Late Woodfordian ice sheet. That border has been traced from the Great Plains to the Atlantic Ocean. The deployment of the ice, as marked by the front, and the direction indicators and deposits behind it, show clearly the struggle of the ice to surmount the topographic irregularities in its path. Although the moraine can be traced essentially without break in the area of Plate I and has been considered to be the limit achieved by the ice as it flanked the Driftless Area, other evidence now available indicates that earlier ice went further westward into the Driftless Area as Chamberlin (1882, Pl. II) showed. That evidence, including deposits of distinct glacial origin and distinctive land forms, is cited later.

In the map area, the moraine is a narrow ridge generally less than 300 feet wide and commonly only 20 to 60 feet high (Fig. 14a). The morainal ridge is wider and higher locally, as in the lowlands between Wisconsin Dells and Baraboo, and is smaller on the Baraboo uplands and in the Wisconsin River valley to the south. The moraine is so symmetrically developed, on the upland surfaces particularly, that it resembles a railroad right of way. A youthful appearing, sharply linear ridge is the rule, but multiple parallel ridges are present locally. The outer moraine is characterized by a paucity

Figure 14. Photographs of prominent features:

A. Top of Late-Woodfordian (Cary) end moraine on the East Bluff of Devils Lake about one-quarter mile southwest of Wisconsin Highway 113, looking northeastward.
(Photo taken March, 1966)

B. Air view looking southward of Devils Lake, the morainal plug in the foreground of Late Woodfordian age, and the distant broad flat of the Wisconsin River at the Badger Ordnance works.
(Photo taken Sept., 1966)

of pits and hollows. Knob and swale topography characterizes the area behind the moraine and the interlobate areas.

Although stone counts in many gravel pits in the area have been made over the years (results are largely in the files of the Wisconsin State Geological and Natural History Survey), no detailed study of the composition of the till has been attempted. The till is generally stony and sandy, but highly variable in texture, composition and color. Some scattered stone counts in the outer moraine show a high proportion of crystalline rocks from the Precambrian shield to the north. However, in recessional moraines, some removed as much as several miles from the terminus, local dolomite and sand of Cambro-Ordovician age predominate.

Features Associated with the Terminal Moraine

Devils Lake. Devils Lake (Fig. 14b) is the only large natural lake today associated with the terminal moraine in the Baraboo area. It is a classic example of morainal damming on both ends of a former stream valley or gorge through the south flank of the Baraboo Range (Salisbury and Atwood, 1900). The lake is about 1.3 miles long, 0.4 to 0.6 miles wide, and generally 30 to 40 feet deep. Its surface elevation averages 963 feet. A shallow shelf extends south from the north shore a distance of about 500 feet. A narrow shelf surrounds the south end. The east and west sides drop abruptly into deep water. The water is soft and clear--on the border between eutrophic and oligotrophic (Twenhofel and McKelvey, 1939).

The lake has only two small streams entering it--Messenger Creek on the southwest and the smaller creek from Hells Canyon on the northeast (see Fig. 26). The total drainage basin is only about 5.5 square miles. No streams flow out of Devils Lake. Evaporation and seepage control the losses. The water table is perched at present lake level presumably by the fine sediments and organic matter in the lake basin (Thwaites, 1958, p. 157).

The sediments around the north and south shores are mostly clean, light-yellow, medium-grained sand with some pebbles of glacial origin. The sediments become finer and darker as water depth increases. The bottom of the lake, below about 25 feet of water, is covered with fine black organic muds (Twenhofel and McKelvey, 1939).

Deltaic deposits and glacial outwash sand and gravel near the south end of the lake were penetrated in a well to a depth of 383 feet without reaching bedrock (Thwaites, 1958). A complex sequence of events is only partially recorded by the glacial drift "plugs" at the north and southeast passes in the quartzite (Thwaites, 1958).

Drained Lakes. Glacial Lake Baraboo occupied the lowland west of the town of Baraboo when the ice front stood at the terminal moraine of Late Woodfordian age (Alden, 1918). Its water level stood at least to an elevation of 980 feet above sea level, making a lake well over 100 feet deep in places, but with numerous islands. The lake enlarged eastward as the ice retreated and merged more fully with Glacial Lake Wisconsin to the north with which it apparently had been connected by narrow passages. The much larger Glacial Lake Wisconsin reached an elevation of about 1,000 feet and drained westward into the Black River, about 78 miles northwest of Baraboo. Few shoreline features of these lakes are seen, but thick sediments with buried organic matter at 145 feet and a weathered zone at 215 feet are reported from wells near Baraboo (Thwaites, 1958) and from various other places (Alden, 1918). An organic-rich clayey zone is buried at 154 feet in sediments of Glacial Lake Wisconsin and is radiocarbon dated at more than 34,000 years B.P. (W--2052). Thus, it seems clear that both those glacial lakes had a long and complex history of which only the latest part can be related to the Late Woodfordian ice. They were drained completely after that as ice retreated to clear the east end of the Baraboo Range (Bretz, 1950).

During ice retreat, Glacial Lake Merrimac came into existence south of the ranges (Bretz, 1950). That lake was short lived and is evidenced by a torrential delta that formed when the waning Glacial Lake Wisconsin remnant north of the range suddenly burst through the drift at the east end of the ranges (Bretz, 1950). Glacial Lake Merrimac had an elevation of about 860 feet and was held in by the morainal dam in the Wisconsin River valley near Sauk City. One major abandoned channel through the moraine may have been produced during the discharge that built the delta.

Proglacial lakes were formed immediately in front of the moraine in places by Devils Lake Park. All of these former lakes have been drained, but their sediments remain behind. One unnamed lake formerly existed 1.3 miles southeast of Devils Lake on the north side of Devils Nose (Fig. 26). Similar but larger lakes were present in Sections 16, 17, and 18, northeast of Devils Lake. Peck and Steinke Glacial Lakes were named early, and Black (1967-68) gave the name Ott Lake to the easternmost and smallest basin in the Sauk Point loop. At one time those basins probably were merged into one lake that must have drained into Glacial Devils Lake via Hells Canyon. As the ice border retreated somewhat from the end moraine position shown on Plate I, the water level dropped and the lakes became isolated.

Trowbridge (1917) calculated that six miles of ice front drained into Steinke Lake, depositing over 2.5 billion cubic feet of debris. About 0.5 miles of ice front contributed at least 142 million cubic feet of debris to Peck Basin. The Devils Lake Gap between the two morainal dams on the north and southeast contains over 2 billion cubic feet of debris. These deposits were derived from 11 miles of the Cary (Late Woodfordian) ice front, which may have contributed 1.5 billion cubic feet of water per year to the area of Devils Lake. The Devils Lake basin alone has a capacity to the discharge level at Skillet Creek on the west side of about 7.5 billion cubic feet. Thus, only five years roughly was needed to fill Devils Lake Gorge to overflowing. That the ice stood in the vicinity for some time is demonstrated by the work of Bachhuber (1966) on Hansen Marsh. Rhythmically-bedded lacustrine sediments at least 25 feet thick were laid down along the ice margin and represent at least 600 years of time, if these are truly varved. Presumably, water drained through Skillet Creek for some time. Erratics up to the level of the Skillet Creek divide at 1155 feet have been found by Trowbridge (1917). However, Thwaites (1958) claimed that no proof exists that Devils Lake ever overflowed into Skillet Creek and suggested that water drained westward between the ice and West Bluff. I see no other mechanism for removing large angular quartzite blocks from the Skillet Creek gorge, than torrential discharge from overflow.

Outwash. In the lowland areas between Wisconsin Dells and Baraboo and in the Wisconsin River Valley, the terminal moraine of the Late Woodfordian ice fronts on prominent outwash aprons (in part deltaic) composed mainly of sand. These extend one to several miles beyond the terminus as distinct topographic forms only slightly dissected. Wells in the town of Prairie du Sac disclose up to 182 feet of clean material without reaching bedrock. Locally wood and other organic matter are reported at depths there, as in deposits of similar thickness at Baraboo.

Stagnant Ice Features. The retreat of the ice from Sauk Point at the crest of the south Baraboo Range was by melting *in situ*, for it left behind typical ice stagnation features with knob and swale topography. Many knobs are small kames of poorly-sorted but water-deposited sand and gravel. The depressions are almost invariably kettles produced by the melting out of buried ice blocks in the debris. Numerous ice stagnation features are found behind the terminal moraine. The lowland areas contain the largest and best developed features.

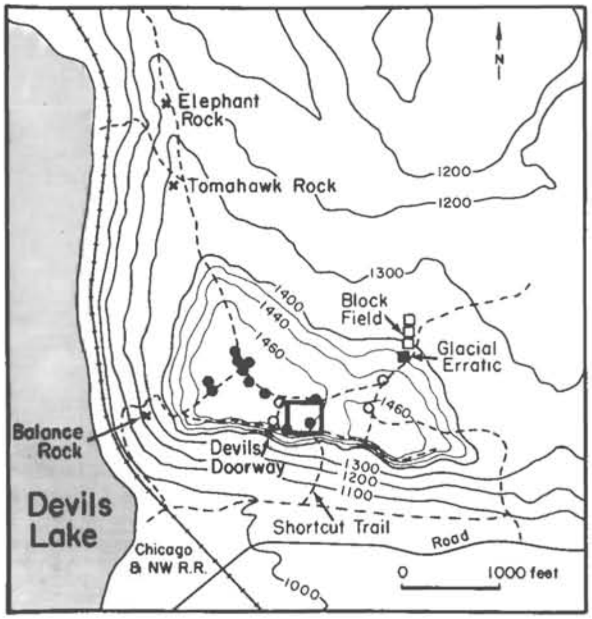

Figure 15. Generalized topographic map of East Bluff, Devils Lake Park, modified from the U.S. Geological Survey Topographic Quadrangle--Baraboo. Trails are partly diagrammatic (from Black, 1964, fig. 1).

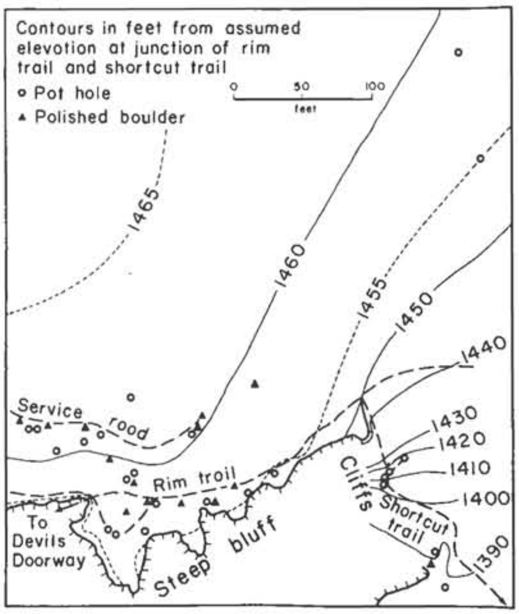

Figure 16. Sketch of pothole area, East Bluff, Devils Lake Park. Solid black circles represent locations of known Windrow Formation below quartzite rubble; open circles where the Windrow was not found. (from Black, 1964, fig. 2).

Potholes and Associated Windrow Formation. Black (1964) described and mapped the famous potholes on the East Bluff of Devils Lake (Fig. 15 & 16). They lie directly above the morainal plug east of the south end of Devils Lake. Some potholes are carved in bedding plane surfaces of the Baraboo Quartzite in situ on the upland. Others are in loose blocks of quartzite that are scattered irregularly on the beveled upland surface. Still others mark a former cascade down the south face of East Bluff. Polished chert and siliceous metamorphic rocks of gravel size from the East Bluff Member of the Windrow Formation of Cretaceous age are associated with those potholes in situ on the upland and have been found in them (Salisbury, 1895; Andrews, 1958). The Windrow Formation on East Bluff apparently is confined to the area above the 1460-foot contour and does not occupy all of that area. The Formation is sporadic but widespread in the Upper Mississippi Valley region (Andrews, 1958).

I dug two pits and used a power auger to drill 13 holes through the quartzite rubble scattered over the highest part of East Bluff and into clay with Windrow pebbles (Fig. 15). The holes were only 2.2 to 7.5 feet deep. Numerous other attempts failed to penetrate the rubble, or encountered no Windrow Formation to depths of 4 feet. Six of the general areas without Windrow pebbles are shown also on Figure 15. Maximum thickness known is 16 feet in a dug well at the trail junction northwest of the pothole area (Salisbury, 1895) (Fig. 15).

The East Bluff Member of the Windrow Formation encountered in the auger holes is silty clay with minor sand and about 5 to 10 percent well rounded and polished siliceous pebbles (but no Baraboo Quartzite) scattered uniformly throughout. The clay samples expanded many fold and gelled markedly when a sodium-rich solvent was used to aid disaggregation. Fred Madison (written communication, 1966) by X-ray study of two of my samples confirmed a marked concentration of montmorillonite and a variety of clay minerals, some kaolin and clay-sized quartz. Kaolin dominates in all other clayey Windrow sites (Andrews, 1958).

Black (1964) concluded that the pebbly clay on East Bluff might be an ancient glacial deposit. The high polish in the potholes attests to their recency of formation or of exhumation. The dating of the time of the formation of the potholes and of the pebbly clay deposit is still open from Precambrian time to the Pleistocene. The gravel on the edge of the bluff seems to be a lag concentrated from the clay during the Pleistocene. The initial clay deposit is not considered fluvial in origin. However, an electron microscopic study of surface textures of the sand grains from two samples of the Windrow Formation from the vicinity of the cistern, revealed "...no evidence of glacial action. What we do find is a good deal of mechanical action, probably violent fluvial action in a highly energetic environment; in other words, the sand grains were probably rolled around and battered in water with a great deal of turbulence." (David Krinsley, written communication, May 22, 1970). The sand grains, thus, confirm the multigenetic history of the well rounded pebbles. Both together make up less than 10 percent of the clay deposit in which they are found, but both are obviously not now in an environment of deposition (the clay) in which they were rounded. This leaves the origin of the deposit still open.

Windrow-type pebbles similar to those at East Bluff are found in many places west of U.S. Highway 12 in the Baraboo area on the Precambrian and Paleozoic rocks. Such pebbles are intimately intermixed with the thin soil. Their presence on easily eroded sandstones with relatively steep slopes poses problems if they are simply down-wasted products from an ancient Mesozoic landscape. Some soil profiles in the Baraboo silt loam (Geib, et al., 1925) locally west of U.S. Highway 12 are strongly oxidized and red, suggesting a Sangamon age in contrast to the yellow-brown and gray profiles of the Baraboo silt loam soil on the upland east of Devils Lake.

Potholes in Baraboo Quartzite also are found near the east nose of the Baraboo Range (Black, 1967-68; see description of supplementary stop M). One contains sandstone like that of the lower Paleozoic adhering firmly to its walls. Some of these potholes show high polish; some were striated inside their lips by the Late Woodfordian ice. It is not known how these relate to the potholes at Devils Lake.

Periglacial Features. Periglacial processes are those particularly involving frost action. Frost riving and gravity movements were especially important in the Baraboo area. Stabilized talus, block concentrations and block strewn slopes, choked valleys and block cascades are found in the Baraboo region (Smith, 1949; Black, 1967-68). The talus accumulations in the Devils Lake gorge are among the most striking features of the Baraboo region. The talus slopes are surmounted by pinnacles and monuments, also considered derived by frost action. Talus (excluding quarry detritus) and rocky cliffs are found in the gorge of the Upper Narrows north of Rock Springs and along the bluffs of the Lower Narrows, northeast of Baraboo. Talus is more abundant outside the area covered by the Late Woodfordian ice than inside. However, it does not necessarily follow that all pinnacles and monuments are much older, if any, than that ice.

Evidence of Greater Deployment of the Ice

Evidence of deployment of ice of presumed Woodfordian age beyond or west of the prominent terminal moraine shown on Plate I consists of features or material interpreted as kettles, an esker, till, a kame, and erratics. Three depressions interpreted as kettles are found well beyond the limits of the terminal moraine. One adjoining Wisconsin Highway 33, in the SE$\frac{1}{4}$, NW$\frac{1}{4}$, Section 18, T.12N., R.6E., is several hundred feet across and approximately 15 feet deep. A permanent small pond occupies its center. It lies west-southwest of a prominent ridge on the northeast side of Wisconsin Highway 33 (SW$\frac{1}{4}$, NE$\frac{1}{4}$, Section 18, T.12N., R.6E.). The ridge, 1,000 feet long, 150 feet wide, and 30 feet high, resembles the curving form of an esker; much material has been removed for road aggregate. The stratification and lithology of the material of which it is comprised is similar to the general outwash adjoining it; it contains more gravel than the outwash.

Another large gentle depression in the outwash west of U. S. Highway 12, in the west central part of Section 16, T.12N., R.6E., has relief of only a few feet. Its maximum enclosure shown by the contour line is 1,500 feet. These three features are respectively 4.3, 4.0, and 1.5 miles west of the prominent terminal moraine. They lie just north of the North Range of quartzite and in the headward reaches of the drainage flowing northwestward into the Mirror Lake valley. The outwash plain in the vicinity of these topographic forms has been only slightly dissected by runoff.

A third prominent depression interpreted as a kettle lies south of the Baraboo Ranges in the vicinity of the Badger Ordnance Works, on the grounds of the Mt. Ida Convent (SW¼, NE¼, Section 15, T.10N., R.6E.). This sharp kettle is about 500 feet in diameter and 30 feet deep. It lies in outwash sands, but is underlain with till (Thwaites, 1958, p. 157). It is one mile west of the terminal moraine.

Glacial deposits west of the terminal moraine and not directly related to the outwash or to lacustrine action are found locally. For example, south of Wisconsin Dells in the northwest portion of Section 27, T. 13N., R.6E., three power-auger holes up to ½ mile west of the prominent terminal moraine disclosed till-like debris below outwash and lacustrine sediments in each of the three holes. The till-like material rests directly on bedrock. It has not been studied in the laboratory.

Two miles northeast of North Freedom (SE¼, Section 25, T.12N., R.5E.) are distinctive gravel deposits; they are 3.4 miles west of the terminal moraine. They have been recognized for many years (Sailsbury and Atwood, 1900, p. 130; Weidman, 1904; Alden, 1918; Thwaites, 1958, p. 154), and have been interpreted both as ice-rafted debris and as deposits laid down directly by glacial ice. They are interpreted here as kames deposited directly from ice on the basis of their volume, stratification, and texture. Some 40,000 cubic yards of aggregate have been removed from one pit alone for road construction purposes. The material ranges in size from boulders of various Precambrian rocks several feet in diameter to clay. It is poorly sorted, but discrete foreset beds at the angle of repose are made up of poorly sorted, fine to very coarse units. The foreset beds indicate that water flowed westerly. These deposits lie just below the level attained by Glacial Lake Baraboo. Occasional ice-rafted or wave-rolled stones and boulders at the former shore may also be seen in the upper walls of nearby pits in bedrock. The deposit, however, is too large to ascribe to ice rafting or to wave or current modification of ice-rafted debris.

Another deposit of buried erratics is distinct above the level of glacial Lake Baraboo, on the northwest side of the West Bluff of Devils Lake. This site also is outside the area of impoundment in Devils Lake gorge. A bulldozer in 1967 exposed numerous large boulders of dolomite and mafic and acidic igneous rocks one-tenth mile east of Wisconsin Highway 123 at an elevation of 1,140 feet (SE¼, NE¼, Section 14, T.11N., R.6E.). These were below 1 foot to 4 feet of soil with apparently undisturbed soil profiles. They are 150 feet above the level attained by Glacial Lake Baraboo and 200 feet above the terminal moraine one-quarter mile to the north and northeast. They are too numerous and too large to have been transported by man without heavy equipment, and their position under a normal soil profile belies recent burial.

Identification of local rock types as erratic material now stratigraphically higher than source areas, present problems in interpretation. Black (1964 and 1967-68) has concluded that many quartzite blocks on the highest surfaces of the Baraboo Ranges are true glacial erratics deposited directly from the ice. These include boulders up to 85 tons that seemingly have been moved up slope. Hundreds of quartzite boulders also lie on the

Windrow Formation of supposed Cretaceous age (Andrews, 1958), which caps the highest part of East Bluff (Black, 1964) (Fig. 15). No higher quartzite surface is nearby. Quartzite blocks are common in the Cambrian sandstone, but they cannot be let down on Cretaceous material. Glaciation just beyond the prominent terminal moraine seems the only reasonable interpretation.

Some of the residual chert west of Devils Lake is associated with Windrow-type pebbles in rubbly zones on the Baraboo Quartzite or on Cambro-Ordovician strata. Many of these deposits are considered to be glacial erratic material, too.

Synopsis of Glacial History and Evolution of Some Major Landforms

Many interpretations by the earlier workers of the glacial history of the Baraboo area are plausible and may be correct. However, we are constrained by relatively few facts in our interpretations. A simple sequence of events does no justice either to the credibility of the reader or to the facts as we know them. Several alternative sequences are just as viable as any one we might select. It becomes necessary, then, to list some representative interpretations that have been proposed without attempting to negate or defend any one because of lack of space.

In 1882, T. C. Chamberlin's map of the glacial deposits of Wisconsin showed drift of the Second Glacial Epoch fronting on a line west of Devils Lake, from Sauk City northward to Wisconsin Dells (formerly Kilbourn City). The edge of the drift extended, in some instances, several miles west of prominent moraines (the outermost of which corresponds closely to that of Plate I. Alden (1918, p. 194) stated, "The relations of the fronts of either the Illinoian or the early Wisconsin glaciers to the Baraboo quartzite range are not known, and positive statement as to the age of the deposits in the Baraboo basin and in the Wisconsin Valley north of the quartzite range is not possible..."

We have not changed materially our status of knowledge of the Baraboo area during the last half century. A major question which Black (1967-68 and this note) has attempted to answer in part is: "Has glacial ice gone farther westward into the Driftless Area in the Baraboo area than indicated by the prominent terminal moraine"? Numerous features in the Baraboo area and elsewhere in southwest Wisconsin (Frye, Willman and Black, 1965) would indicate that it has. Weidman (1904) listed evidence for an earlier ice advance beyond the Late Woodfordian moraine. (Thwaites, 1958) rejected that evidence, but accepted the concept. Black (1967-68 and in this note) suggested that ice only slightly older than that which built the moraine went several miles westward. One or more older glaciations probably of early Pleistocene age are indicated also by weathering of some of the unusual deposits in southwest Wisconsin.

In the Baraboo area the Late Woodfordian ice left numerous prominent surface features like moraines, outwash plains, and lake deposits. Buried lake deposits and outwash should be attributed in large part to the pre-Late Woodfordian ice, if our interpretations of buried organic matter in the local deposits and of partly dated stratigraphy associated with the Late Woodfordian front elsewhere in Wisconsin are correct (Black and Rubin, 1967-68).

It seems logical to correlate the time of the most frost action yielding talus, monuments, pinnacles, and the natural bridge near Denzer (Black, 1959) with the permafrost climate when ice wedges formed in Woodfordian time (Black, 1965). Bachhuber (1966) by pollen studies documents the rapid warming of the climate subsequently. How many times cold-climate processes and glaciers directly affected the Baraboo area during the Pleistocene thus seems debatable--from once to many times. Certainly ice passed by or over the area many times during the Pleistocene on its way into Illinois (Frye, Willman, and Black, 1965). This situation then leaves open the time of origin of many major landforms in the Baraboo area (Thwaites, 1958).

The story of the Baraboo area must begin a billion years or so ago, in Precambrian time, with deposition in shallow seas of many hundreds of feet of very clean quartz sand, which became the Baraboo Quartzite. That chapter has been covered thoroughly by Dott. So too have those chapters on subsequent burial, metamorphism, tectonism, erosion and reburial under the Cambro-Ordovician sea.

Some unknown time after the folding and uplift of the Baraboo Range, subaerial erosion (Trowbridge, 1917), and probably marine erosion also (Thwaites, 1958), developed relief of a thousand feet between the top of the Range and the surrounding beveled Precambrian igneous and metamorphic rocks. Such relief was considered by them as due almost entirely to the great resistance of weathering and erosion of the quartzite. That relief obviously had a pronounced influence on the arcuate deployment of the ice front and the formation of opposed lobes as the ice was diverted around the north and south sides of the Ranges during the Pleistocene.

Beveling of the upland quartzite obliquely across the bedding produced surfaces that look smooth to the eye and have long been called peneplains (Thwaites, 1960). However, the interpretation that the region was in the end-stage of one or more cycles of erosion is now discredited (Thwaites, 1958 and 1960). Nonetheless, the mode of beveling of the resistant quartzite at such marked elevations above the surrounding plains is not truly understood. A possible wave-cut terrace lies on the northeast part of Happy Hill, six miles west of Devils Lake (Thwaites, 1958).

Several gaps are cut to the level of the surrounding plain through the narrow, steep North Range; only one is known, that with Devils Lake, in the broader south Range (e.g., Salisbury and Atwood, 1900; Trowbridge, 1917; and Alden, 1918, p. 105-107). Some, such as part of Devils Lake gap, are definitely pre-Late Cambrian in age, at least in part, for they contain Cambrian sandstone; Thwaites (1958, p. 145) attributes the north part of the gorge and several others (e.g., Upper Narrows, Narrows Creek, Lower Narrows, and entrance to Baxter Hollow) to the Paleozoic cycle of erosion; others likely are post-Paleozoic, and still others were modified by streams as young as Pleistocene to Recent age (Black, 1964 and 1967-68). Hence diagrams like Figure 17 are essentially hypothetical. We do not really know by what stream or when many gaps were cut--from the Precambrian to the Pleistocene.

Eaton (1872) credited Increase A. Lapham, (famous naturalist and at that time Secretary of the Wisconsin Academy of Sciences, Arts and Letters) with the idea that the Baraboo River once ran through the Devils Lake gorge but

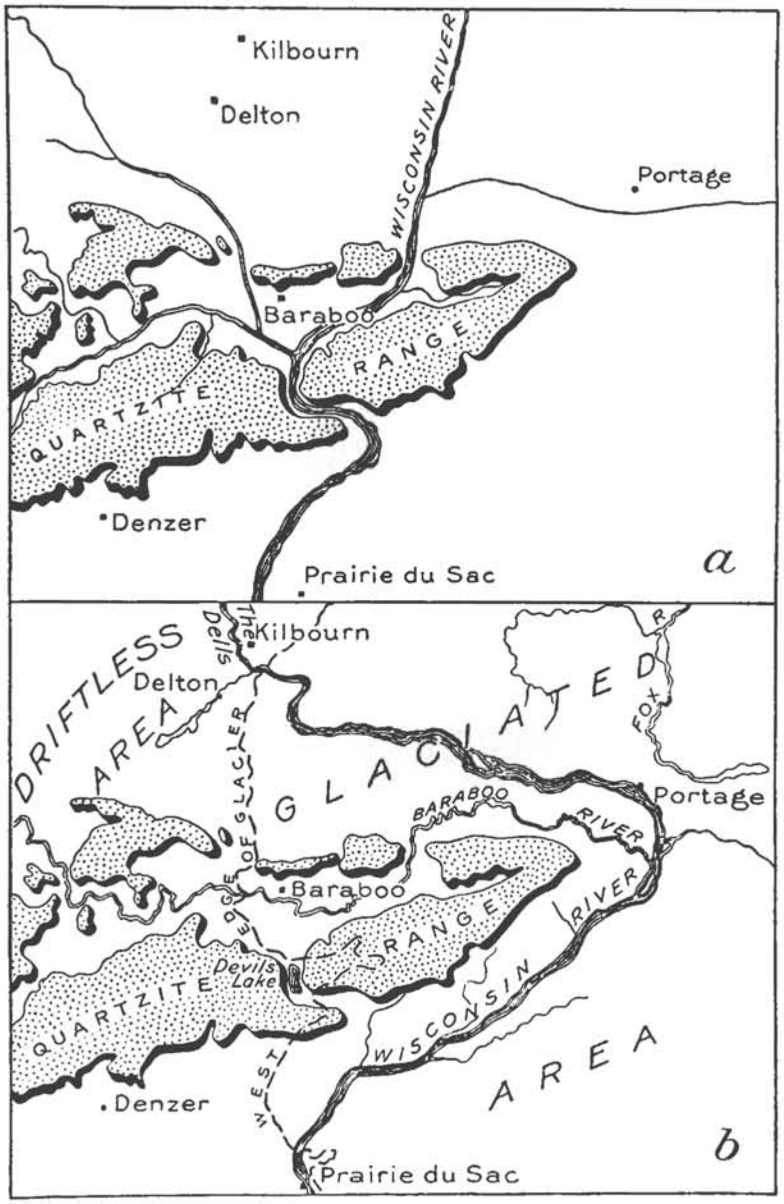

Figure 17. Hypothetical preglacial (a) and present drainage (b) in the Baraboo Ranges (from Alden, 1932, fig. 3). (Kilbourn is now Wisconsin Dells).

was diverted by glacial drift. Irving (1877, p. 508) believed the Devils Lake gap was more recent than the Potsdam (Cambrian) period and suggested it was cut by the Wisconsin River that also flowed through the Lower Narrows and was diverted by the ice of the "Glacial Period" to its present position. T. C. Chamberlin (1883b, p. 284-285) and Salisbury and Atwood (1897, p. 141) accepted Irving's suggestion. Later, Salisbury and Atwood (1900, p. 68) showed that the gap existed in Precambrian time, but they made clear they did not wish to assert that the gorge was as deep then as now. Trowbridge (1917) discusses at length various possibilities in the timing and the position of former streams that cut the gorge without strongly favoring any one. He says that neither the Lower nor the Upper Narrows seems large enough to be correlative with Devils Lake gap and they have not been proven to be of pre-Late Cambrian age (See discussions of STOPS 2 and 3). On the other hand, he cites the freshness of the gaps as an indication that they probably were cut at the same time during some post-Silurian period. He points to a broad, continuous stream gap filled with Cambrian sediments through the North Range northwest of Baraboo as a possible correlative with the Devils Lake gorge. Alden (1918, p. 105-107) also showed that at least part of the gorge was cut before the Cambrian sands were deposited in it and suggested that the Wisconsin River much later found that zone of weakness to reoccupy the gorge. He assigned the cutting of potholes and the deposition of the Windrow gravel on East Bluff to the time when the ancient river flowed at the summit level. Deep bedrock gorges lead from both the Upper and Lower Narrows to Devils Lake, but Martin (1932, p. 90) believed that the Wisconsin River was superposed on the quartzite at the Lower Narrows and Devils Lake gorge after cutting down through the Niagaran limestone and other Paleozoic rocks. Exhumation of the quartzite and restoration of the Baraboo Range to its former topographic prominence came simultaneously with the cutting of the Devils Lake gap. Thwaites (1958) and others have generally accepted Alden's (1918) interpretation. However, the writer suggests that other interpretations still seem equally plausible.

Hanging valleys in the quartzite of the south flank are anomalous also. They are broad and gently dipping in their upper reaches and plunge precipitously to the buried Precambrian surface hundreds of feet below. Some are filled partly in their upper reaches with Cambrian sandstone so date from the pre-Late Cambrian erosion cycle; some also have narrow notches cut into them that must postdate the Paleozoic. Pine Hollow, Parfreys Glen, Durwards Glen, and Fox Glen are typical.

A variety of drainage changes from glaciation characterize the Baraboo area from the supposed major shift of the Wisconsin River out of the Devils Lake gorge to the minor shifting of portions of streams like Skillet Creek (Salisbury and Atwood, 1900, p. 138-139). The Wisconsin Dells, north of the area of Plate 1, is perhaps the most famous example of relocated drainage in the region (Salisbury and Atwood, 1900; Martin, 1932, p. 345-353). (See Fig. 17). The main bedrock channel lies east of the present drainage, and the present route is considered to be entirely a postglacial channel eroded in Cambrian sandstone after the draining of Glacial Lake Wisconsin. Only the one major channel around Blackhawk Island is mentioned generally as an abandoned route occupied before the present narrow gorge was established by stream capture (Martin, 1932, p. 349). Air photos and aerial inspection disclose a veritable labyrinth of shallow abandoned water courses at the Dells. Some, like Rocky Arbor, seem to have been occupied by water more than once and also have had their direction of flow reversed. No detailed studies of the area have been attempted in the last half century, and the full story is not yet available.

During deglaciation of the area about 12,500 radiocarbon years ago (Black and Rubin, 1967-68), man seemingly appeared on the scene. Concentrations of charcoal in apparent hearths at Raddatz Rockshelter, Sk5, beneath the natural bridge near Denzer suggest man's presence along the front of the Late Woodfordian ice (Black, 1959; Black and Wittry, 1959; and Wittry, 1964); artifacts appeared later in the depositional sequence (Wittry, 1959; Parmalee, 1959). The lower concentrations of charcoal in the shelter are among the first indications of forests in Wisconsin since about 30,000 radiocarbon years ago (Black and Rubin, 1967-68). Periglacial climates seem to have precluded their development during the long interval from 30,000 to 12,500 radiocarbon years ago, even though some ice-free areas are thought to have been available in southwestern Wisconsin (Black, 1969). An extensive lake in Honey Creek Valley was present in Late Woodfordian time, but probably was drained shortly thereafter (Alden, 1918). An arm or bay lay immediately southwest of the shelter. Man apparently followed the forest into the area while looking for deer; at least his later diet indicates that late-winter and early spring hunting of white-tailed deer dominated his seasonal migrations to the shelter (Parmalee, 1959).

Since glaciation, gravity and frost have moved many large blocks of quartzite down slope. The present rate is very slow. Man's unsightly activities are now more important. Railroad and other construction, large pits for aggregates, and farming have left their marks. At present the most serious threats to the landscape are the rapidly increasing numbers of tourists and students. Inclusion of part of the area in the National Ice Age Scientific Reserve will help. However, immediate far-sighted planning and perhaps re-zoning of the whole Baraboo district is needed if many features and the esthetic value of the area are to be preserved in a form suitable both for scientists and tourists.

PLANT ECOLOGY OF THE BARABOO HILLS

(Zimmerman)

Influences on the Biota

The diversity of rock types, of topography, and of local microclimatic gradients in the Baraboo region have had a marked effect upon the present biota. The soils of the region are very diverse, reflecting the variety of parent materials such as glacial till, lake deposits, eolian sand, dolomite, sandstone, quartzite, and talus. Topographic influences are equally diverse. Larger valleys with talus slopes include the Upper and Lower Narrows, Narrows Creek, and Devils Lake valley. Narrow, shaded gorges include Pewits Nest, Parfreys Glen, Durwards Glen, Fox Glen, Pine Hollow, and a host of smaller unnamed glens along both Ranges (Pl. I). Broader valleys with bouldery swamp basins include Pine Creek and the valleys southwest and northeast of Devils Lake. Intermediate between these, and including some features of each, are both forks of Honey Creek, Hemlock Draw, and Baxters Hollow in the southwest; Seeley Creek in the western end and Rowley Creek in the eastern end of the central basin between the Ranges also are examples. Because of the richness of geology, plant and animal life, and archaeological history, many of these areas are in various stages of preservation as natural and scientific areas for specialized, limited teaching and research use. They require protection from over-use as well as abuse! Trampling away from trails by flower photographers and geologists is just as harmful as the air pollutants from Badger Ordinance Works, which have killed pines and rare lichens in the South Range.

Ecologic Diversity

The unusual biotic diversity of the Baraboo hills -- indicated by a floral count alone of over 600 species in almost any square mile area -- is due to the abundance of different microhabitats combined with a varied climatic history, all of which left many relic populations in favorable sites (see Table 11). The famous Driftless area, of which the western half of the Baraboo district is a part, acted as a biotic refugium at least during the last (Wisconsinan) glacial stage. It provided a topographic island within which many organisms survived the last major glacial advance. And because of the great topographic diversity of the region, a much wider range of microhabitats existed than if it had been a flat terrain. In the Baraboo hills, we find an especially rich collection of relic plant communities. The latitude is such that northern and southern species overlap to an unusual degree. More than half of the species of vascular plants of Wisconsin are found here, for example. A few species even represent outposts of Rocky Mountain forms. At least 100 species of birds nest in the region, and there are thirty species of mammals, six turtles, about 15 snakes, some lizards, amphibians and fish as well as many invertebrate animals.

Glens contain the most exotic floras characterized by many northern forms. The most common trees on north- and east-facing slopes include sugar maple, yellow birch, basswood, and, in the most deeply-shaded areas, hemlock. Pine and oak are more common on southern and western slopes (Table II). Smaller plants include ferns, mosses, liverworts, and a variety of flowers

TABLE 11. SUMMARY OF NATURAL PLANT COMMUNITIES OF THE BARABOO HILLS

TOPOGRAPHY	SOIL TYPES: Loess-Silt Clay	Sand	Rocky
Ridge Tops	-------Prairie flora ------	Pine flora-----------------	
Cliffs:			
North-facing	-------Yellow birch-hemlock-moss-lichens----------------		
South-facing	-------------Pine and small lichens---------------------		
Slopes:			
North-facing	Sugar maple flora	Yellow birch-basswood	Ferns, lichens
South-facing	Oak flora	Juniper and cactus	Oak flora
Valley Bottoms:			
Stream banks	Wet meadows	Lake shore flora------------------	
Flood plains	Silver maple-River birch	River island flora-----------------	
Terraces	Prairie or Dune flora		
Swamps	Tamarack-Sphagnum flora		

Note: Additional modifying factors include extra moisture (due to local seepage, poor drainage, and shading of some valley bottoms) and also fires, logging and farming.

For a comprehensive discussion of the Wisconsin floras and post glacial changes therein, as well as a summary of man's effects, see J. T. Curtis (1959) The Vegetation of Wisconsin, Univ. of Wis. Press, 657 p.

that are quite rare for the general region. The conifers and yellow birch commonly grow in seemingly inhospitable sand, talus and even in cracks in cliffs along the glen walls. Yellow birch roots can be seen grasping cliff faces like the arms of an octopus. Besides the local hemlock groves, dripping-cliff floras occupy the most unique microhabitats in the glens. The glen floras are favored by combinations of deep shade, moisture seepage, and the drainage of cool air into the valley bottoms. Temperature data for three April days at Pine Hollow showed valley bottom minima of 19^{o} - 28^{o}F and maxima of 54^{o} - 62^{o}F, while at the valley top, minima were 24^{o} - 32^{o}F and maxima were 72^{o} - 77^{o}F (Kopitzke, 1966). Besides being always warmer, the valley top not surprisingly also has more than a 10^{o} greater diurnal variation.

Above the gorges and on tops of bluffs, especially on south-facing slopes, many of the plants are postglacial invaders from the south. Oaks, shag-bark hickory, sumac, dogwood, and bracken fern are conspicuous. On very dry, sandy or rocky southern hillsides, juniper and prickly pear cactus occur, and in several areas in the Baraboo hills, small relic prairie foras of blue-stem grass, shooting-star, lead-plant, Baptisia, and others can be found. As if this were not enough, the several glacial outwash terraces along the Wisconsin River Valley to the south provide an additional array of biotas, including extensive sand prairies that for three millenia contained stabilized and enriched sand dune areas. When agriculture allowed the moisture-holding prairie organic matter to oxidize without replenishment, farming declined despite the introduction of pine windbreaks to hold snow and reduce evaporation. Then the abandoned lands, being too dry for European weeds, went right back to prairie vegetation.

Human Occupation

Man came to the Baraboo region more than 12,000 years ago. Archaeologic and geologic studies of the Raddatz and Durst Archaic Indian rockshelters near Leland (See Fig. 35) indicate the presence of man at least by the time the Late Woodfordian (Cary) glacier began to melt in the eastern part of the region (see preceding section on Glacial Geology). The Effigy Mound Builders, a group of Woodland Indians, lived in the Baraboo region from approximately 700 to 1200 A.D. They built many mounds in southern Wisconsin, most of which have the shapes of animals or birds, including several on the moraine north of Devils Lake. Man Mound east of Baraboo is unique for its human shape. In historic times, Sioux, Winnebago, Chippewa and Sauk Indians all lived within the district.

The first white men to view the Baraboo Ranges were French voyageurs and missionaries, who first entered the region in the 1630's from the Great Lakes and Fox River Valley. Undoubtedly they explored at least the southeastern Baraboo hills during canoe trips down the Wisconsin River to the Mississippi. Their passage through the region is memorialized in such local names as Portage and Prairie du Sac. Following the last Black Hawk Indian War, New Englanders began to settle here about 1835, soon to be followed by European immigrants. A Scottish community is remembered by the names Caledonia, Alloa, and Durward, just southeast of the hills.

Needless to say, white man's agricultural and industrial activities have altered the local ecology markedly, though less than might be expected; many local landowners have resisted attempts at "development." Not all of the effects, however, would be obvious to the layman. Whereas prairie fires had tended formerly to suppress oak and other trees -- thus preserving the prairies -- when agricultural practices began, burning virtually ceased. Then oak saplings quickly sprang up on unfarmable hills so that in southern Wisconsin, woods have expanded greatly over the past century due to farming. On the other hand, cessation of marginal farming on the rockier slopes of the Baraboo hills also has allowed large areas to return to woodlands. Pines and junipers, especially, have expanded on many cliffs and bluffs. Heretofore, such changes have been unplanned, but as stress on the environment increases, enlightened management will be required if the priceless natural ecology of the region is to be preserved.

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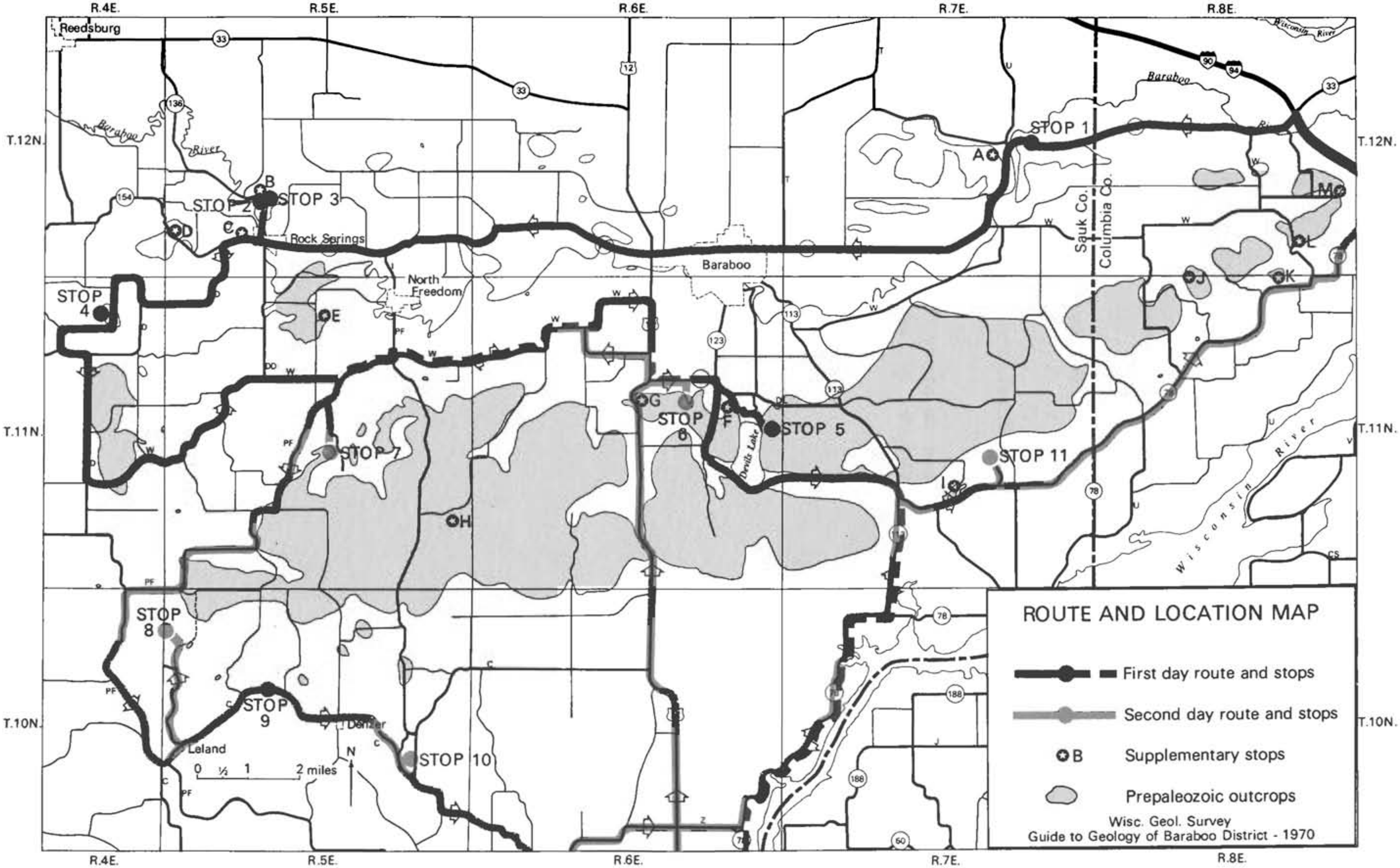
Figure 18. Route and Location Map



Road Log
(Dalziel and Dott)

FIRST DAY

Madison--Baraboo--Madison

0.0 Road log starts at corner of East Washington Avenue and Capitol Square. Drive east from the Square on East Washington Avenue (U.S. 151).

Madison is situated on an isthmus between two lakes. The lakes represent former drainage ways modified by glaciation. The last ice began retreating from the Madison area between 12,000 and 13,000 years ago. Glacial deposits overlie approximately 600-800 feet of Lower Ordovician dolomite and Upper Cambrian sandstones, which unconformably overlie Precambrian basement. Madison's water supply is derived from wells approximately 800 feet deep in Upper Cambrian sandstone, which yield 1,000-3,000 gallons per minute. From several hills in the city where glacial drift is thin, early settlers quarried dolomite and sandstone for some of the older buildings on the University campus and downtown. The north-south-trending Wisconsin Arch passes just west of Madison; the Baraboo Precambrian inlier lies on its axis (see Fig. 1). Paleozoic rocks tend to thin slightly across the arch, and they dip very gently away from it to the east toward Lake Michigan and to the southwest toward the Mississippi River.

2.1 Madison East High School on left.

3.1 Junction Wisconsin Highway 30.

4.2 Junction U.S. 51.

5.8 Howard Johnson's on left ahead; note outcrop of Cambrian rocks (Jordan Formation) behind gas station just west of motel.

6.2 Junction Interstate 90/94.
Turn on to Interstate westbound.

From here to the east for many miles there are fine examples of drumlins. They are oriented northeast-southwest, clearly reflecting the southwesterly movement of the Green Bay lobe of glacial ice. From here to the Baraboo area we travel over glacial deposits of Late Wisconsinan age.

15.6 Gravel pits on left in kames associated with stagnating ice rather than recessional moraines as mapped by Alden (1918).

19.8 On left and right America's Dairyland!

24.8 Hummocky morainal topography.

30.4 Crossing Wisconsin River, which is dammed up at the town of Prairie du Sac about fifteen miles downstream. Former Glacial Lake Merrimac flooded all of this region to an elevation of approximately 900 feet.

East end of Baraboo Ranges at eleven o'clock. Wisconsin River now flows around the east end of the Baraboo Ranges, but prior to glaciation apparently crossed through the eastern part of the Baraboo basin (see Fig. 17).

33.1 Junction with Wisconsin 78. Continue on Interstate.

34.5 The wooded ridge on the left is formed by the extreme eastern end of the Baraboo Syncline.

35.7 Leave Interstate on Wisconsin 33.
Proceed west on 33 across the Interstate toward the Baraboo Ranges. Figure 18 shows our route through the Baraboo district.

36.7 Road to left leads to Fox Glen. One of the better stratigraphic sections of the Cambrian rocks is exposed in Fox Glen, where a steep-gradient stream flows across Cambrian strata in a series of picturesque waterfalls.

37.3 Pigtail Alley Road

For the next few miles we travel at the foot of the North Range, which is held up by near-vertical, south-facing Baraboo Quartzite on the north limb of the Baraboo syncline. Cambrian sandstone and conglomerate abut against the quartzite in the lower wooded hills immediately to the left; each of the more prominent valleys exposes them.

38.8 Small road to left with a cliff of Galesville Sandstone (Dresbachian) visible just back from highway. To the right, the low flat plain of Glacial Lake Wisconsin is visible, with the modern course of the Wisconsin River in the distance.

41.6 'Narrow Bridge' sign at right: disembark.

STOP 1 -- LOWER NARROWS -- East Side.

From the roadside at this point one can see through the Lower Narrows of the Baraboo River into the basin to the south formed by the Baraboo syncline. The Lower Narrows is one of several "water gaps" in the area. There is much evidence that many present valleys in the area are exhumed early Paleozoic ones. It is probable that the Lower Narrows is such an old valley, although clear evidence is lacking here.

This gap is cut in the south-facing and near-vertical or steeply north-dipping quartzite beds of the north limb of the Baraboo syncline. On the basis of subsurface data, now missing, A. Leith (1935) drew a north-south fault through the gap. On the north slopes of the North Range on either side of the Narrows are the most extensive outcrops of rhyolite in the area The rhyolite structurally overlies the inverted quartzite, but the contact is not visible, and no rhyolite pebbles have been positively identified in the quartzite (see general discussion of ages and relationships of Precambrian rocks).

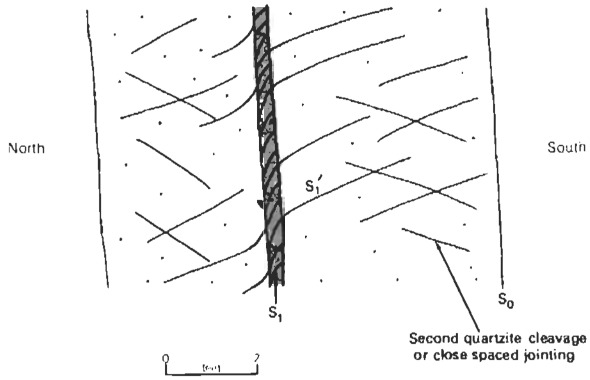


Figure 19. Bedding/cleavage relations high on the west side of the Lower Narrows (STOP 1 and Supplementary Stop A). Compare with those at Van Hise Rock (STOP 2, see Frontispiece and Fig. 23).

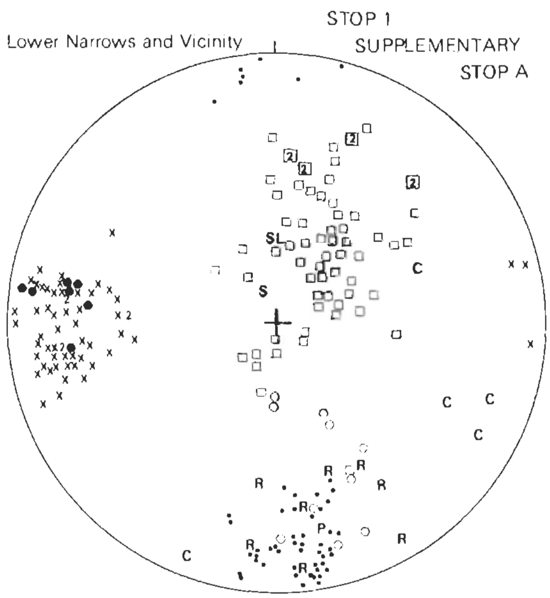


Figure 20. Stereoplot of structural data from the Baraboo Quartzite and underlying rhyolite at the Lower Narrows (STOP 1 and Supplementary Stop A). For key, see Table 13.

<u>Cross the road</u> - but watch out for fast-moving traffic - and proceed to rhyolite outcrops in low cliff to the east of the house.

The red, quartz-veined rhyolite at this locality (elsewhere it is black; see Supplementary Stop K) can be taken for quartzite at first glance! It is cut by a number of sets of structural surfaces, some flat-lying like the cleavage in the quartzite at the Lower Narrows (see Figs. 19 and 20), but there is no sound basis for correlation.

A north-dipping compositional layering, probably flow banding, can best be seen in weathered outcrops in the field above the roadcut. Angular clasts of various sizes can be seen to be aligned in this surface. Apparently clasts like these led Weidman (1904) to re-interpret erroneously his earlier-postulated (1895) volcanic breccia as a basal conglomerate of the Baraboo Quartzite containing rhyolite pebbles (see Stark, 1932, and discussion in text). Thin section study shows definite volcanic textures typical of welded tuffs, and the fragmental rock is, indeed, a rhyolite breccia as Weidman originally thought.

Continue on Wisconsin Highway 33 across bridge over Baraboo River. Junction County U at end of bridge; County U and an unnamed road provide access to main rhyolite exposures west of the Lower Narrows (see Supplementary Stop A).
Proceed through Lower Narrows. Note to the left outcrops of overturned beds of Baraboo Quartzite dipping steeply to the north. Some flat-lying joint surfaces also are visible. Overturning, which produced steep north dips, is common along the north side of the northern limb of the syncline.

42.9 Looking back and to the right, the upper surfaces of beds in the Baraboo Quartzite can be seen in a large quarry where the quartzite is being mined to produce grinding balls. In the quarry excellent ripple marks and many of the structural elements of the Baraboo Quartzite can be observed. Here we enter the synclinal basin. Flat-lying Cambrian and Ordovician strata filled this basin, which was surrounded by an atoll-like ring of quartzite islands (Fig. 13). In this part of the basin, Paleozoic strata are largely concealed by glacial deposits, but in the western portion we shall see many exposures of them. Directly ahead on the skyline is the South Range of the Baraboo hills in which the quartzite dips northward (toward us) at 25°. Hence, the syncline here at the eastern end is a simple asymmetric structure with a north-dipping axial surface.

43.5 Junction County X to left.
This road provides access into the east end of the basin (see Supplementary Stop M). High wooded ridge on skyline to left (east) is Pine Bluff. Its steep south face has exposed at the top flat-lying St. Peter Sandstone, which is the youngest Paleozoic formation preserved in the area.
Continue on Highway 33 toward Baraboo.

43.6 Glacial erratic boulders on left. Glacial deposits cover bedrock over most of the eastern basin; numerous striated pavements demonstrate east-to-west ice movement.

43.9 Very flat valley bottom at left (south) may have been the bed of a temporary glacial lake formed during early stages of ice retreat. It could have connected with Glacial Lake Wisconsin through the Lower Narrows. Travelling straight west toward Baraboo, one can see the North and South Ranges on either side; in other words, the two limbs of the syncline.

45.0 The low ridge half a mile north of the road is called Man Mound. It is one of the larger outcrop areas of Baraboo Quartzite within the basin; beds are vertical, and face south. The odd name of the ridge derives from a man-shaped Indian mound on the north side built by Effigy Mound Indians, who lived in the region from about 700 to 1200 A.D.

45.7 At two o'clock there is another low, wooded, quartzite ridge. The outcrops on this ridge have been said to be formed by one of the higher formations of the Precambrian, the Dake Quartzite (A. Leith, 1935). It has been positively identified only in drill cores now destroyed. Apart from a somewhat anomalous cleavage attitude, and at one place a rather unusual conglomerate,the rock looks like Baraboo Quartzite.

46.4 Rocky Point Road to right.

46.8 Notch in South Range at about ten o'clock on left is the valley in which Devils Lake is located. At STOP 5, the last today, we shall see the lake. It has long been felt that the pre-glacial route of the Wisconsin River was from the Lower Narrows across the basin to the present Devils Lake valley and out to the south. After glaciation, however, this route was blocked by morainal material so that the river sought its present route around the east end of the Baraboo hills, and the Baraboo River was diverted through the Lower Narrows to join the Wisconsin northeast of the syncline. Re-routing of the post-glacial river also caused the cutting of the Dells of the Wisconsin 12 miles to the north (Fig. 17).

47.6 Junction County T to north provides access to the North Range.

47.7 Fairgrounds on left. Hill beyond on the south skyline (with houses) is underlain by Cambrian sandstones with a veneer of glacial drift.
Proceed west through the north side of the city of Baraboo. Baraboo was the original winter home of the famous Ringling Brothers Circus, and it has today a fine museum of circus history.

48.6 Yellow and white Ringling Mansion on main road at corner of Ash Street is an example of elegant late-nineteenth-century mid-western architecture.

48.8 Junction with Wisconsin Highways 113 and 123. Proceed straight ahead.

49.3 Ochsner Park on left. Note wall built of glacial erratic boulders, many of which are dark red Baraboo Quartzite

50.1 Junction U.S. Highway 12.

Proceed straight ahead; leave Wisconsin Highway 33 and follow Wisconsin Highway 136.

Baraboo River visible to left. Ridge we climb straight ahead is the Late Wisconsinan (Late Woodfordian, or Cary, terminal moraine). Note erratic boulders and gravel pits. As we proceed west from the moraine, note the contrast in topography with that farther east. This is the boundary between the glaciated region to the east and the famous driftless area of southwestern Wisconsin. However, small patches of driftlike material and large, erratic boulders occur as much as five miles west of the moraine. These and other features suggest to R. F. Black that much of the driftless area <u>was</u> glaciated during an earlier stage of the Pleistocene by thin ice that contained very little rock debris and did not modify the topography appreciably. Glacial Lake Baraboo filled the valley ahead of us to at least 980 or 1,000 feet elevation above sea level; the valley floor is about 860 feet.

50.6 The wooded knoll on the right is another major quartzite inlier within the basin. There is a small quarry at the west end. The beds are vertical, and face south.

50.9 Baraboo River on left; note the flat valley bottom beyond. Much of the western part of the basin through which the river now flows was flooded by Glacial Lake Baraboo.

51.8 The North Range of the basin is rather subdued in this area. There are only a few small quartzite inliers. This may have been an area with few, if any, islands during early Paleozoic time; thus, it may have provided the main access of the sea to the interior of the basin (see Fig. 13). It also may have been a pre-glacial route of the Wisconsin River.

53.7 The low, flat Lake Baraboo plain is visible to left.

54.2 Road to right leads to small quarry in the Upper Cambrian Lodi Siltstone Member of the St. Lawrence Formation (see Table 5). Just north of the quarry there is a large deposit of sand and gravel considered to be a kame by Black. F.T. Thwaites believed that these were lake-beach gravels with some large ice-rafted erratics.

54.4 Junction County PF on left, which leads to the town of North Freedom. Proceed on Highway 136.

55.2 Roadcut at right on summit of wooded ridge exposes the Jordan Sandstone, which is the uppermost unit in the Trempealeau Group; the Oneota Dolomite directly overlies it, and is the youngest Paleozoic formation within this part of the basin (Table 5). As we go down the hill, we get a fine view of the southwestern portion of the basin. The skyline approximately delimits the quartzite outcrop along the south and west margins of the Baraboo syncline. Cambrian strata form most of the hills in the middle distance.

56.9 Small outcrop of Galesville Sandstone in roadcut on left.

57.5 Entering Rock Springs (formerly known as Ableman). To right, up valley, is a prominent outcrop of Galesville Sandstone.

58.0 Approaching railroad tracks. Outcrop on right of Galesville Sandstone.

58.1 Crossing Baraboo River.
Center of Rock Springs. Turn right on Highway 136 to Upper Narrows of Baraboo River (also known as Ablemans Gorge). We shall drive through the Narrows and then walk back (see Fig. 21). As we start to drive through, note at left old quarries in Galesville Sandstone (formerly known as the "White Rock" and Gall Stone Quarries).

58.4 Rock Springs Park, behind which is a large old quarry in the Baraboo Quartzite (Figs. 21 and 22). Note the vertical bedding of the quartzite throughout the gorge and local thin Cambrian sediments on top (Fig. 22). Across river to east is a large active quarry in the quartzite, which we shall visit later (STOP 3).

58.7 Straight ahead at north end of gorge is an excellent exposure of Cambrian conglomerates on top of the vertical quartzite. Those conglomerates are equivalent in age either to the Tunnel City or Jordan strata.

58.9 The prominent rock on the right hand side of the road is the famous Van Hise Rock in which the cleavages in the quartzite and a more argillaceous layer are beautifully displayed (see Cover Photo and Fig. 23). We shall look at the rock in more detail on our walk back through the gorge. It was named by friends of C. R. Van Hise at the University of Wisconsin, where he was a professor of geology and later president. It commemorates the contributions made by that great geologist to our understanding to the Precambrian geology of the Lake Superior region and of structures such as those seen in the Baraboo Quartzite. On the far side of the gorge, the cliffs described in Supplementary Stop B can be seen.

59.0 STOP 2 -- Just short of the bridge over Baraboo River. Leave buses and walk back through gorge.

The Upper Narrows of the Baraboo River provides the best possible introduction to the bedrock geology of the area, for it illustrates most of the critical structural features of the Baraboo Quartzite and displays admirably the general nature of Cambrian sedimentation near an old quartzite hill. Based upon evidence to be seen at STOP 3, it is probably that the Narrows is an exhumed Cambrian valley, although it may not have been so deep as it is now.

Disembark at the north end of the Narrows by a highway bridge. Visible

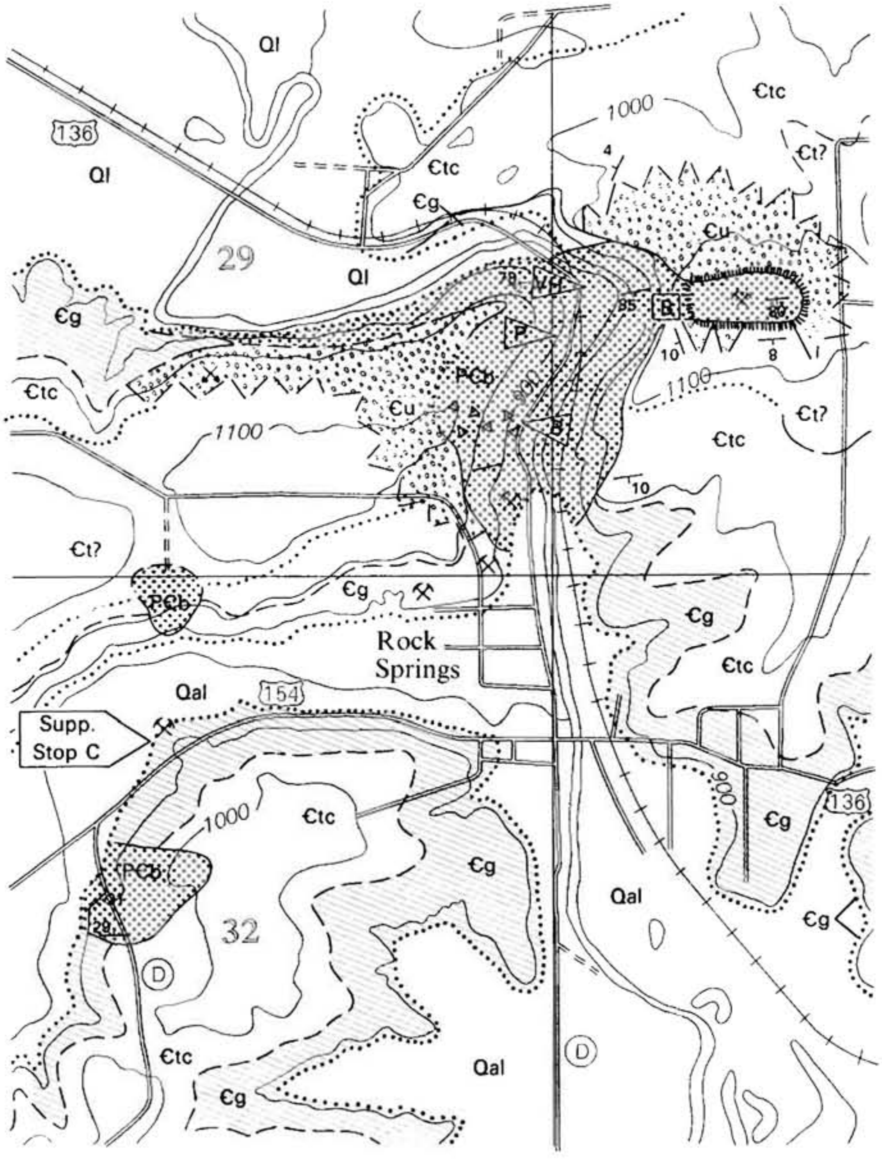


Figure 21. Geologic map of Upper Narrows (Rock Springs) area, which includes STOPS 2 and 3 and Supplementary Stops B and C. On west side of gorge: VH - Van Hise Rock; P - polished quartzite surface; B - breccia zone. Note initial dips and distribution of conglomeratic facies in Cambrian rocks. (Formational symbols same as Pl. I: PЄb - Baraboo Quartzite; Єg - Galesville Sandstone; Єtc - Tunnel City Formation; Єt - Trempealeau Group; Єu - undifferentiated conglomeratic sandstones; Qal - river alluvium; Ql - glacial lake beds). (Modified from Usbug, 1968).

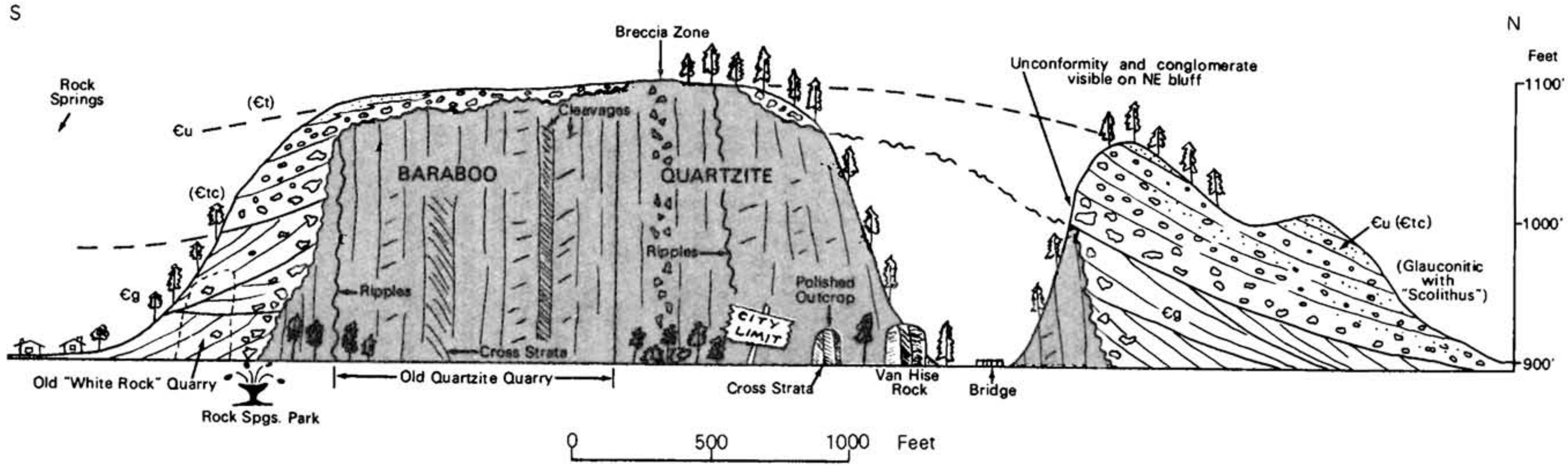


Figure 22. Diagrammatic cross section of Upper Narrows of the Baraboo River showing key geologic features referred to in discussion for STOP 2.

on the north bank of the river across the railroad tracks is a cliff of Galesville Sandstone with large-scale cross stratification overlain by conglomerate laterally equivalent either to Tunnel City ("Franconian") or Jordan strata. Note the initial dip to the northwest (Figs. 21, 22). The Galesville contains only a very few angular cobbles and boulders even though it abuts at right angles against a buried cliff of Baraboo Quartzite in the trees just above the railroad trestle, a relation sometimes described as a buttress unconformity. The overlying conglomerate, however, laps across the truncated quartzite as can be seen high on the river bluff to the east (Fig. 22); it contains boulders up to 5 feet in diameter on the unconformity, but most clasts are of cobble size. Wanenmacher, et al., (1934) interpreted the Galesville here as wind deposited because of the near-absence of conglomerate next to the quartzite as well as the perfection of rounding and sorting of the sand (see general discussion of Cambrian sedimentation). In an old quarry north of the railroad one-eight of a mile west and extending west for several miles, the conglomerate contains a glauconitic sand matrix, and has sandstone lenses rich in fossils (burrows). Those strata are equivalent to the Tunnel City Formation; similar ones can be traced for 5 miles to the west. The Tunnel City rocks clearly are marine, and that conglomerate was rounded and spread laterally by surf.

From here we shall walk back to the south through the Narrows. Be careful along the highway!

Van Hise Rock. The fame of this monolith is justified. Please do not hammer this rock! It is best for a large group to examine it from the eastern side away from the traffic! The rock consists of two current-bedded massive vertical quartzite beds, which face south and sandwich a darker, fine-grained argillaceous layer (Cover Photo and Fig. 23). We will call the latter the phyllitic layer here by way of contrast and in view of the cleavage developed in it, but it is seen to be surprisingly arenaceous in thin section, and is phyllitic quartzite.

Is Van Hise Rock in place? Examining the outcrops west of the road along strike reveals no equivalent phyllitic bed. However, two massive quartzite beds wall a vertical cleft and pinch together westwards. Presumably the phyllitic layer formed a lens of sedimentary or tectonic origin, which has been weathered away. The structural elements in Van Hise Rock have exactly the attitudes of their counterparts elsewhere in the North Range. In fact there is a "smaller edition" of Van Hise Rock, clearly in place, high on the west side of the Lower Narrows (see Fig. 19 and description of Supplementary Stop A).

The structures visible in Van Hise Rock are shown diagrammatically in Figure 23. It is the beautifully displayed cleavage refraction that has led this locality to figure so prominently in the literature. Well developed cleavage dipping gently south in the quartzite is refracted into a phyllitic cleavage dipping north at about 40° in the darker bed (for definitions of cleavage terminology used here, see general discussion of structural geology). X-ray patterns of powdered samples from the phyllitic layer here and others in the Baraboo Quartzite reveal an assemblage quartz-pyrophyllite- (muscovite)-(hematite). This is indicative of the lower greenschist facies of regional metamorphism and temperatures not greater than 410-430°C (see general discussion of petrology).

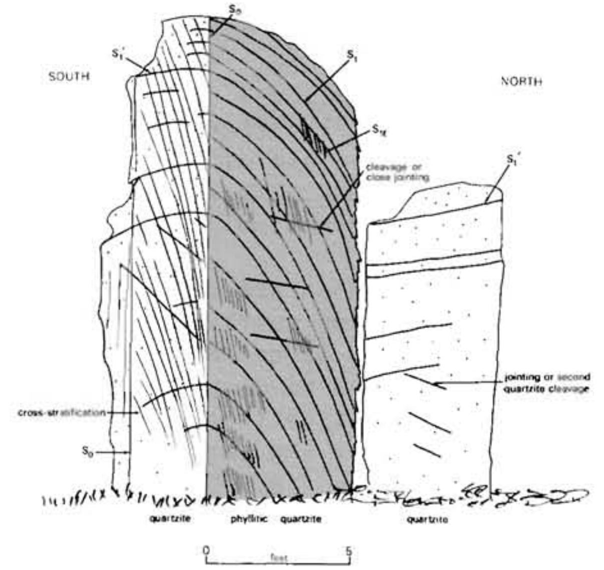


Figure 23. Sketch of structures in Van Hise Rock as seen from the east (STOP 2).

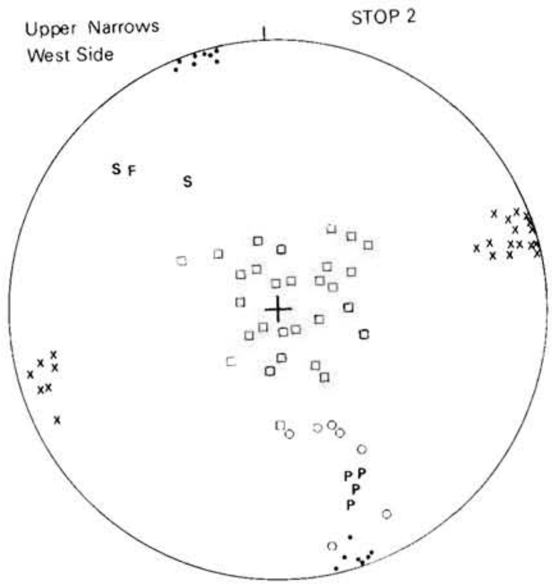


Figure 24. Stereoplot of structural data from the Baraboo Quartzite on the west side of the Upper Narrows (STOP 2). (See Table 13 for key).

Van Hise (1896), Steidtman (1910), C.K. Leith (1913, 1923), Swanson (1927), A. Leith (1935), Mead (1940) and others regarded the partings along the bedding and the cleavages as conjugate surfaces of shear failure. They equated them with the planes of no distortion (i.e., circular sections) of a triaxial strain ellipsoid portraying the state of strain immediately prior to failure (Fig. 6). There are, however, serious objections to this use of the strain ellipsoid, and the cleavages cannot be so easily explained. Their origin is discussed in the section on the genesis of the Baraboo Syncline.

The phyllitic and quartzite cleavages in Van Hise Rock are correlated with the prominent cleavage surfaces seen in phyllitic and quartzitic strata respectively throughout the syncline and here designated S_1 and S_1'. Their attitudes are typical of the North Range and there can be no doubt that they are closely related to the formation of the Baraboo Syncline (see general discussion of structural geology). Closer examination of the phyllitic layer in Van Hise Rock reveals a slaty type of cleavage (S_{1E} on Fig. 23), more penetrative than S_1', dipping north at a higher angle (50°). A similar structural surface on the south limb of the syncline can be shown to pre-date S_1 (Fig. 5). The geometric relationship of this cleavage to the syncline (Fig. 3) indicates that it developed during an early phase in the folding process rather than being an unrelated earlier tectonic structure.

On the more prominent S_1 phyllitic cleavage surface a lineation plunging down-dip can be observed (most easily on top of the rock and on a very thin dark phyllitic layer on the north side of the northerly quartzite bed). This lineation is seen to be both a mineral alignment lineation and (in the most phyllitic parts) a fine crenulation. It is suggested that they mark the direction of maximum finite extension within the phyllitic layers (see Fig. 7 and general discussion of structural geology).

Note that the bedding/cleavage intersections are near horizontal and oriented east-west, normal to a set of joints (sometimes quartz-filled), which are very prominent in the quartzite. These "_ac_" joints normal to the hinge of the syncline are statistically the most common, by far, in the area (see Plate V). They are interpreted as extension fractures normal to the regional least compressive principal stress.

A closely spaced jointing dips at a lower angle than the S_1 cleavage in the phyllitic layer, and there are structural surfaces in many other orientations. A second cleavage or closely spaced jointing is commonly developed in the quartzite of the Upper Narrows (see Figs. 24, 41, and 42), for instance here and on the west side of the road at the (dangerous!) bend 45 yards north of Van Hise Rock.

Finally, note the tension gash band on the north side of the northerly quartzite band. Most of these bands in the vicinity fall into two steeply dipping conjugate sets striking northwest-southeast and north-northwest-south-southeast and indicating right and left lateral motion respectively. (see Fig. 8 and general discussion of structural geology).

Through the rest of the gorge, cleavage in the quartzite (S_1') and in scarcer phyllitic layers (S_1) can be found with attitudes similar to those observed in Van Hise Rock (Fig. 24).

Just south of Van Hise Rock on the west side of the road is a unique, highly polished face on quartzite in which cross stratification shows clearly. The polish probably was produced by Pleistocene sand abrasion either by water or wind, but we cannot rule out the possibility of Cambrian sand-abrasion if the gorge was as deep then as now.

At the "Rock Springs City Limit" sign, walk up lane to right (west) into an old quartzite quarry. Above and to the right is a ripple-marked vertical face in the quartzite. To the far right is a curious breccia zone in the quartzite. This is best developed on the spur separating this quarry from the larger one to the south (0.25 miles south of Van Hise Rock).

Breccia Zone. The occurrence of brecciated quartzite at this locality has been known for many years. Blocks of red quartzite from $\frac{1}{4}$ inch up to two feet across and mostly very angular, are set in a stockwork of white quartz crystals. The latter are encrusted with dickite, a two layered, monoclinic mineral of the kaolinite group, especially in vugs. Some of these vugs now form cavities a few feet deep in the quarry walls.

The breccia zone is over 100 yards wide and runs vertically to the top of the cliff. Similar breccia occurs 100 yards further north on the east side of the Upper Narrows leading to suggestions of a north-south fault through the Narrows. There are also occurrences one mile to the west in an inlier on the south side of the North Range, in the Chicago and Northwestern Railroad Quarry (STOP 3), and at the southeast end of the Narrows Creek gorge (Supplementary Stop D). In the past faults have been drawn along the North Range in the Upper Narrows-Narrows Creek area, and through the Upper Narrows, on the basis of the breccia occurrences. There are certainly highly slickensided surfaces at the present locality. For instance, there are horizontal slickensides on the north wall of the spur between the two quarries, on a steep northwest-southeast trending face. Slickensides also are common on all types of surface within the quartzite throughout the district, however, and we know of no data that indicate a significant displacement along the breccia zone. Moreover there is no sign of cataclastic deformation (mechanical breakdown) of the blocks. Without the quartz matrix they would fit back together like a jig-saw puzzle.

S.W. Bailey of the University of Wisconsin in Madison, who identified the dickite by X-ray techniques, reports that fluid inclusions in the quartz crystals indicate a temperature of 105-107°C, which is consistent with that required for the formation of dickite (personal communication). The matrix of the breccia in the quarry on the east side of the gorge contains kaolinite, not dickite; so too does a similar breccia observed by Associate State Geologist M.E. Ostrom in an inlier of Precambrian quartzite similar to the Baraboo Quartzite in northern Adams County, Wisconsin (Bailey, personal communication).

It therefore seems possible that these breccias are associated with the hydrothermal activity that Bailey and Tyler (1960) suggested was responsible for the occurrence of kaolinite, dickite and other clay minerals in the Lake Superior iron ores. While a fault would provide a logical pathway for hydrothermal solutions, there is no strong evidence of faulting in the breccia zones.

Large, Old Quartzite Quarry behind Rock Springs Park. This quarry (Fig. 22) exposes the vertical bedding in the quartzite; cross stratification can be seen, and excellent large ripple surfaces occur on the south face. In the west wall, close examination will reveal the quartzite cleavage and a little of the phyllitic cleavage seen at Van Hise Rock (the very early penetrative slaty cleavage is not visible here). Note the Cambrian conglomerate veneer on top of the south and west walls. It is only a few feet thick for the most part, but fissures as much as 5-10 feet deep are filled with boulders (see south face). Clasts range up to three feet in diameter, and are rounded.

South End of Narrows. Cambrian Galesville Sandstone (locally known as "White Rock") is exposed in the old Gall Stone quarry just southwest of Rock Springs Park. Identical outcrops occur on the opposite side of the river as well. The sandstone here was deposited inside the Baraboo basin and next to a steep quartzite cliff; it shows initial dips up to 10°. Only two or three angular blocks of quartzite about 1 foot in diameter are visible in it here; apparently they simply fell from the cliff onto a sandy bottom and could not be abraded enough to become well rounded. We infer that either the sand was deposited here as wind dunes, or as marine sand in moderate depths of water next to a nearly vertical wall-like cliff against which waves could not break with great force. In either case, the location is in the lee of the quartzite hill.

Walk 200 yards north across a small stream to the base of the cliff forming the south end of the large quartzite quarry. This is the ancient buried cliff, and spectacular blocks of quartzite up to six feet long occur in a matrix of tan Galesville Sandstone. The jumbled, unstratified nature of the boulders suggests little lateral transport, rather only infiltration and burial in situ by sand. Step carefully up this slope as it is steep and can be slippery.

Return to Rock Springs Park.

59.6 Lunch at Rock Springs Park; potable artesian spring water available. Proceed south on Highway 136 back to Rock Springs.

60.0 Turn left with Highway 136 in center of Rock Springs back across Baraboo River, and cross two of the three railroad tracks. Turn sharp left on gravel road between tracks. Proceed north. Galesville Sandstone at right.

60.3 Road up hill to right leads to Chicago & Northwestern Railroad Quarry (see Fig. 21).

STOP 3 -- UPPER NARROWS -- East side (Chicago & Northwestern Railroad Quarry).

The large, active quarry on the top of the east side of the gorge is operated to provide ballast for railroad tracks. Quartzite is admirable for that purpose because of its physical and chemical durability. Sandstone along the road grade leading up the hill to the quarry is of the Upper Galesville and Tunnel City Formations, and was deposited inside the Baraboo basin (Fig. 21). It contains only scattered quartzite fragments much like the old "White Rock" flagstone quarry seen across the river. The railroad quarry itself contains excellent exposures of the stratification and ripple marks in vertical quartzite. The quartz-filled breccia zone has been observed, and phyllitic and crenulation cleavages occur locally in 4 to 6 inch phyllitic zones on the eastern wall (see Supplementary Stop B). Please do not climb the quarry walls!

Most spectacular is a Cambrian boulder conglomerate 15-25 feet thick (probably of Trempealeauan age) resting unconformably upon the summit of the quartzite hill; probably the surface was wave-cut at least in part. Identical deposits also occur on the west side of the gorge (Fig. 22). Intersection of joints and bedding produced rectangular blocks of quartzite that were readily moved and rounded as gradual flooding of the area brought this hill into the surf zone. The stratified conglomerate seen previously at the north end of the gorge resulted from lateral spreading of gravel by surf and currents as the top of the island was flooded. Note that many boulder surfaces show percussion marks. These, together with the large boulder sizes and well rounded shapes, attest to vigorous movement and powerful impacts during deposition. Note also that the basal conglomerate grades abruptly upward into cross-stratified sandstone with abundant "Scolithus" tubes, which can be seen at the entrance to the quarry opposite the main crusher (this is presumed to be the Jordan Sandstone). Initial (?) dip of the sandstone toward the gorge suggests that a valley of some sort already existed here in Late Cambrian time, although it may not have been very deep.

Return to buses.

Proceed south back to center of Rock Springs.

60.7 Proceed west on Highway 154 from junction with Highway 136.

60.9 Across river to right there is a small active flagstone quarry in Galesville Sandstone with excellent large-scale festoon cross bedding.

61.2 Large-scale cross bedding visible across valley to right (north) in Galesville Sandstone; vertical quartzite occurs in the trees to the left of that outcrop.

61.3 House on right has a small flagstone quarry in Galesville on its western side with spectacular large-scale festoon cross bedding (see Supplementary Stop C).

61.5 Turn left (south) on County D.

Highway 154 straight ahead passes through Narrows Creek, the third major gorge in the North Range (see Supplementary Stop D). The quartzite is again essentially vertical, and has swung round to a more northeasterly strike at the western extremity of the range. Local diversion in the strike of the quartzite near Narrows Creek may indicate the presence of a fault along the gorge. Coarse conglomerate occurs in Cambrian strata at both ends of the gorge.

61.7 Small outcrop of Baraboo Quartzite in bank to right of road. This is part of a small inlier where the Ablemans Syncline (see Pl. III) can be traced in the woods. It is one of a number of smaller-scale folds that together compose the Baraboo Syncline in the west.

61.9 Sandstones of the Tunnel City Formation ("Franconia") occur on both sides of the road, and are almost totally lacking in Baraboo pebbles in contrast with equivalent-aged strata both at Upper Narrows and Narrows Creek less than a mile away.

62.9 This flat upland surface is underlain by Trempealeau Group sandstones (Jordan Formation).

64.1 County D swings left; go straight ahead and, after fifty yards, turn right on unnamed road.

64.3 Jordan Sandstone in roadcut.

64.5 Jordan Sandstone in quarry on right.

64.7 Turn left on unnamed road (hardtop changes to gravel within a few hundred yards).

65.1 Gravel road turns 90° and heads south.

65.4 Road junction - proceed straight ahead.

65.8 STOP 4 -- West Range.

Cross the field to right. Two hundred yards west of the road there is a series of quartzite outcrops strung out in a north-south line. The quartzite dips gently (15-25°) east. It is cut by a strong cleavage (S_1') dipping 60-70° to the north, approximately parallel to the axial surface of the Baraboo Syncline (Fig. 25). In fact we are here close to the axial surface trace of Ablemans Syncline, which we crossed on County D, just southwest of Rock Springs (Mileage 61.7; see Pl. III).

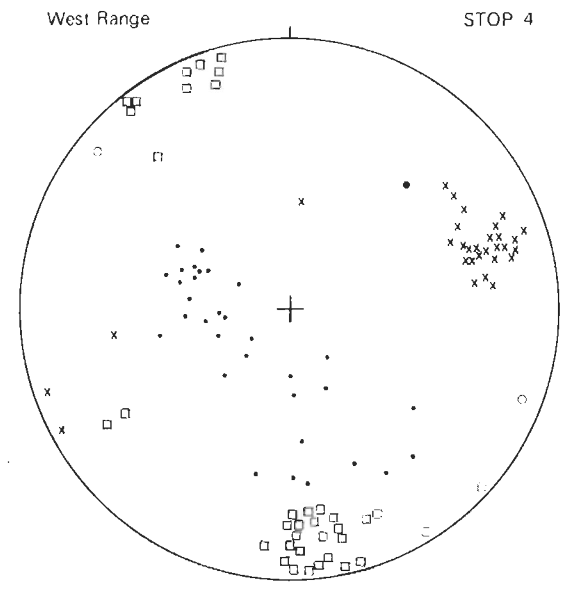


Figure 25. Stereoplot of structural data from the Baraboo Quartzite in the West Range (STOP 4). (See Table 13 for key).

In the outcrop immediately west of the gate into the field where the buses are parked, there are a few small tension gash bands that tend to form conjugate sets/striking northwest and northeast. The orientation of the gashes indicates a clockwise (right-lateral) sense of shear along the northwest trending bands and an anticlockwise (left lateral) sense along those striking northeast. Hence if we assume that they are contemporaneous, they resulted from a stress system with the greatest and intermediate compressive principal stresses oriented in a north-south vertical plane and the least compressive principal stress parallel to the hinge line of the syncline (see Fig. 8, and discussion of general structural geology).

Proceed south on gravel road.

66.2 Turn right (west) at T junction.
In field to the left about a hundred yards east of the road, there is a quartzite outcrop with a northwest-southeast strike. Between this locality and STOP 4 there is an open syncline that may correlate with Ablemans Syncline at Rock Springs (Plate III).

67.2 Turn left (south) on gravel road.
Blocks of Jordan Sandstone visible at head of valley just north of junction; Jordan underlies this entire upland area.

67.6 Turn left (east) at T junction on winding gravel road.

68.2 Junction with County D.

Turn right (south).

In roadcut on east side of junction, reddish-brown soil with chert and quartz geode blocks is typical residuum of the Oneota Formation. Similar material is widespread around the southwestern end of the syncline.

69.3 The low, wooded ridge behind the field to the left is the northern part of the largest quartzite inlier in the West Range. Several folds occur there.

69.6 Junction with road to Zion Lutheran Church. This road goes through a gap at the narrowest part of the quartzite inlier where there is an east-plunging syncline. In the stream bed farther east, Cambrian conglomerate lies as a thin veneer over the quartzite. This is another example of the close conformance between modern and pre-Late Cambrian topography.

70.7 Junction County W. At right, Elder Ridge School.

Turn left on W.

70.9 We are here in the gap between the West and South Ranges. Exposure of the quartzite is very poor. Although it reaches the road on the left, there are practically no more outcrops until the South Range is reached on the right.

A fault may be present in the gap, or else a tight syncline, part of the synclinorium that comprises the west end of the Baraboo "syncline" (see general discussion of structural geology). In the field below the road (right) conglomeratic Cambrian strata are exposed. Their elevation suggests that they are of Trempealeauan age, and they grade abruptly eastward into normal, non-conglomeratic strata.

72.3 Roadcut on left going downhill exposes the Tunnel City Formation with only minor interbedded conglomerate.

72.7 Junction - bear left down winding valley of Upper Seeley Creek.

73.6 Cliff of Tunnel City Sandstone just above road to left (north).

74.5 Junction DD on left.

76.5 The gravel road on left leads to the Diamond Hill quartzite inlier where there are a number of fairly tight macroscopic folds (Pls. II and III; see Supplementary Stop E). Good outcrop of Galesville Sandstone with large-amplitude cross bedding by farm at left (north). Continue east on W.

76.9 Junction County PF.

Turn left (east) on PF and W.

77.2 Ridge on right has Galesville Sandstone. Note large sandstone block called "Devils Chair" (the work of the Devil is everywhere in evidence).

78.1 Junction; turn left 90°.

78.2 Junction with PF.

Turn right 90° on W.

78.5 Bridge over Seeley Creek; Galesville Sandstone at left dipping south 4°.

79.3 We are here driving along the north side of the South Range, with the Baraboo Quartzite dipping northward toward us at a low angle. The structure of the South Range is here complicated by a number of relatively open folds (Pls. II and III). Farther east it is a simple homocline. Along the lower northern slopes of the South Range there is Cambrian conglomerate resting on quartzite. It appears to be of both Trempealeauan and Franconian ages. The conglomeratic facies ends in outcrops just south (right) of Highway W.

81.5 Junction with Rock Hill Road to left (north).

Rock Hill is capped by the Trempealeau Group and may have a trace of the Oneota Formation. Black Earth Dolomite has been seen in the ditch below Rock Hill Cemetery, while Jordan Sandstone crops out just above the cemetery on the south face of the hill.

81.9 Large cliff of Galesville Sandstone on right; note dip.

82.3 Junction Lehman Road to right (south).

82.4 Bridge over Skillet Creek. Cliff of Galesville Sandstone. Pewits Nest, a picturesque waterfall, is about 1/8 mile upstream. At that point Skillet Creek runs over excellent outcrops of the Galesville Sandstone. Another mile upstream, the contact of the Galesville with the Tunnel City Formation can be seen.

82.8 Rolling hills on right in glacial outwash gravels, which are exposed along road for next mile. This area is the face of a delta built into Glacial Lake Baraboo from the Late Woodfordian terminal moraine.

82.9 Highway W turns to right (east).

83.7 Terminal moraine.

83.9 Turn <u>right</u> on U.S. 12.

84.3 Margin of terminal moraine again. From here we drive onto the outwash plain. Moraine is clearly visible to the left (southeast).

85.3 Skillet Creek again.

85.5 Turn <u>left</u> (east) on 159.

86.0 Skillet Creek Farm on right. (We shall stop here tomorrow). Immediately past the farm, we climb up the terminal moraine.

86.7 Junction with Highway W and 123.

Turn <u>right</u>.
Bear left on 123 to Devils Lake State Park.

87.1 Park entrance; see Figure 26.

87.2 On sharp right bend, trail to Supplementary Stop F goes up hill on right. Here there is an exposure of the prominent phyllitic zone near the top of the Baraboo Quartzite on the south limb of the syncline. Within this zone, phyllitic cleavage of the main phase of deformation and conjugate crenulation cleavages of the secondary phase are well developed. There also is a large scale so-called "normal drag fold" to which the phyllitic cleavage is axial planar (See Fig. 48). Driving on from this point the north dip of the thick quartzite beds is readily seen on the right.

87.3 Small exposure on the right of Cambrian conglomerate and sandstone. This is one of several isolated patches of such rocks around the Devils Lake area. Its presence at this relatively low elevation strongly suggests here, as elsewhere, that there was an old valley in this area filled with Cambrian sediments. The sandstone here is quartzitic, which is commonly the case near the boundary with the Precambrian. This suggests that somehow silica has moved in solution from the quartzite into Cambrian strata. Besides being quartzitic, the sandstone is pink-colored, making it easily confused with the Baraboo.

87.5 West Bluff Trail. At the top of the bluff just north of the trail is a large patch of Cambrian conglomerate with boulders up to six feet in diameter.

88.0 Park headquarters; the Nature Center in the park contains excellent dioramas and other geological and biological displays.

Continue straight ahead across railroad.

Turn right through parking areas to the most extreme southeastern lot.

88.4 STOP 5 -- just beyond restroom on left and on top of East Bluff (Fig. 26).

STOP 5 -- NORTHEASTERN CORNER OF DEVILS LAKE AND ELEPHANT ROCK

This stop provides an introduction to South Range geology and rounds out the highlights of geomorphology. Devils Lake occupies a portion of a preglacial valley now dammed at both ends by the Late Woodfordian (Cary) terminal moraine. The lake is fed by two small streams; the only outlet is by seepage.

The buses will park in the easternmost parking lot, and we shall proceed to the foot of the East Bluff nearby to see structures in the Baraboo Quartzite (just beyond restrooms). Then we shall climb the East Bluff Trail by a stairway for a panoramic view of the geomorphology of the lake basin and an examination of well-exposed lower Paleozoic conglomerate and sandstone resting upon quartzite in Elephant Rock. Those who do not wish to climb the East Bluff may visit the Nature Center instead.

Quartzite structure

At the base of the cliff to the east of the road (some 35 yards north of the entrance to the final parking lot) is a locality which, like Van Hise Rock, has figured widely in the literature concerning the origin of cleavage. Sandwiched between massive quartzite beds up to nine feet thick, which dip gently northwards, are two thin (eight inches) and finely-bedded layers of quartz phyllite (Fig. 27). DO NOT HAMMER THIS ROCK!

The quartzite is cross stratified, but care has to be taken in identifying the cross stratification because a fine color banding, which is common in the Baraboo Quartzite, does not everywhere follow either the bedding, the cross stratification, or the tectonic structures. It frequently forms elliptical patterns on smooth joint surfaces (Fig. 27), and is generally taken to be the result of the migration of iron-rich ground water through the partly-lithified sandstone after deposition but prior to deformation and metamorphism.

Oversteepening of the cross stratification is common. It could be either syn-sedimentary or tectonic in origin here as the vast majority of the foresets dip south and the displacement of overlying beds during formation of the syncline was also in this direction (but see general discussion of structural geology).

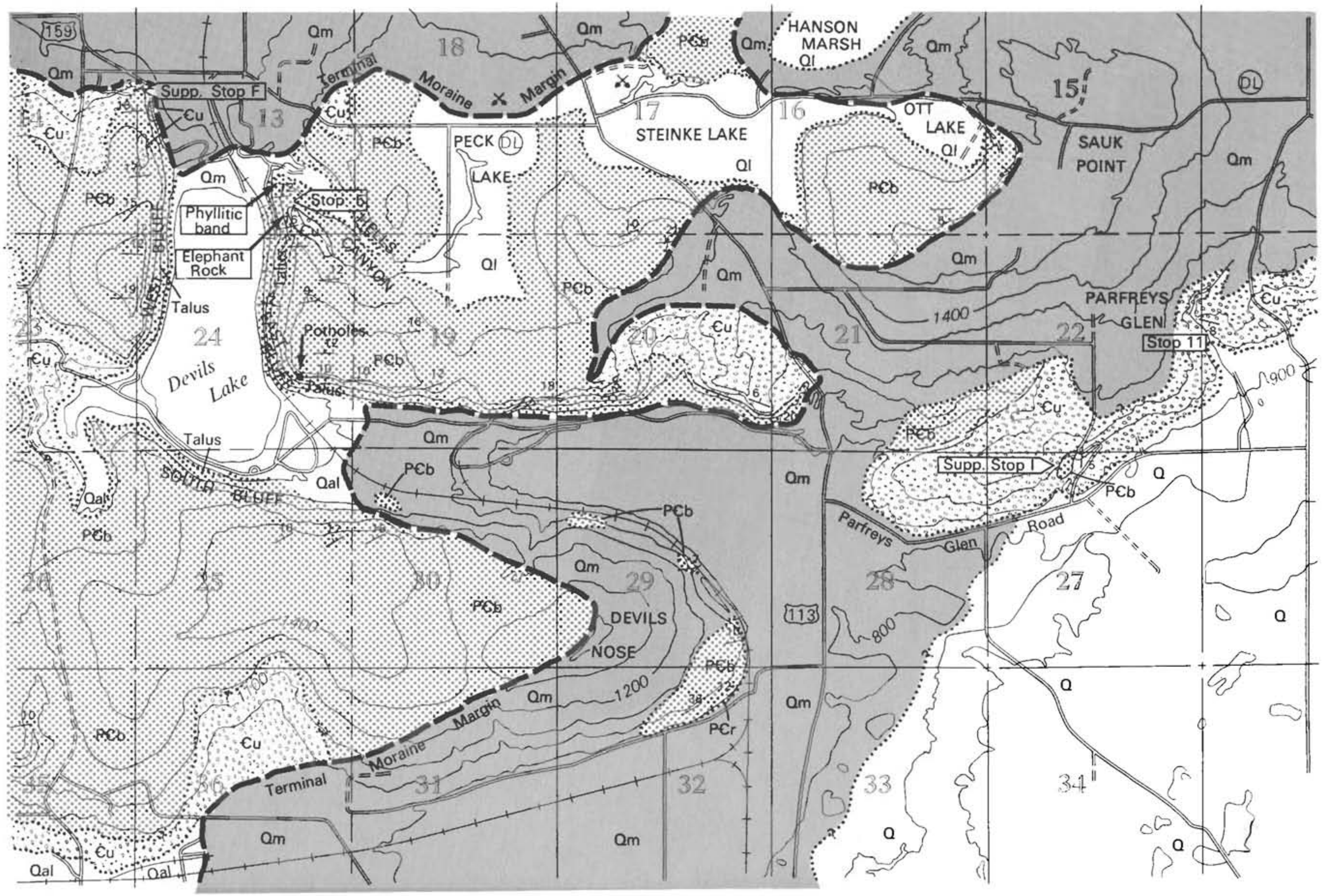


Figure 26. Geologic map of the Devils Lake-Parfreys Glen area, which includes STOPS 5 and 11, and Supplementary Stops F and I. (See Fig. 21 or Pl. I for key to symbols; refer also to Glacial Geology discussion by R.F. Black).

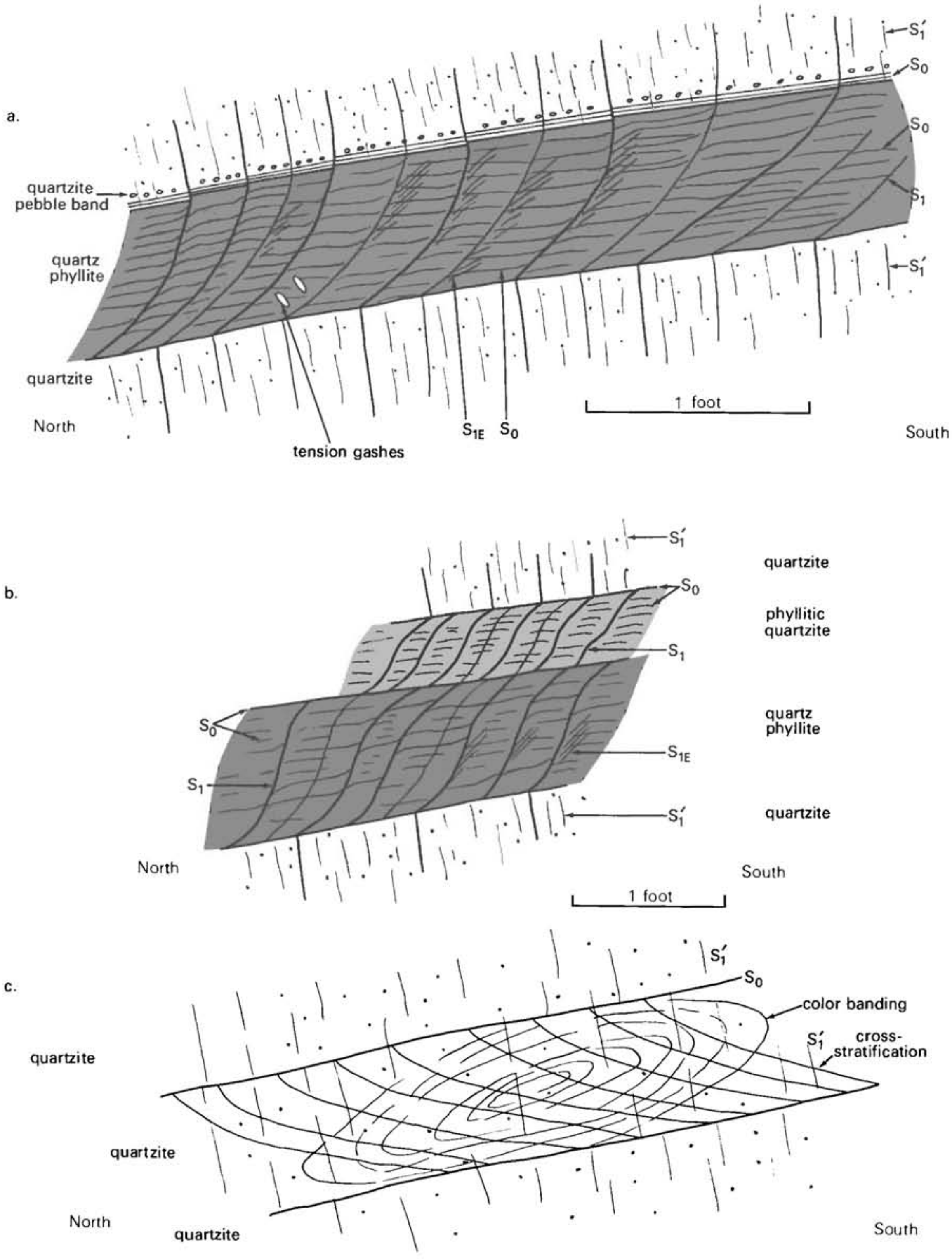


Figure 27. Sketches of field relations at the northeast corner of Devils Lake (STOP 5).

a. and b. Bedding/cleavage relations in the phyllitic layers and adjacent quartzite (for block diagram see Fig. 4; Photomicrograph, Fig. 5A).

c. Color banding in quartzite independent of bedding and cross stratification . This can be seen high on the cliff above the phyllitic layers shown in a and b.

Remarkably penetrative cleavage (S_1') in the quartzite dips steeply south and is refracted into a phyllitic cleavage (S_1), which dips north at various angles as it curves through the quartz-phyllitic layers (Fig. 27a and b). The cleavage at this exposure, as at Van Hise Rock, was formerly interpreted as a surface of shear failure (see Fig. 6, discussion of Stop 2, and text section on the genesis of the Baraboo syncline.

Closer examination of the phyllitic layers reveals the presence of a slaty cleavage (S_{1E}), fainter but much more penetrative than S_1, dipping north at a lower angle than the prominent S_1 surfaces (Fig. 27a and b). Examination of this rock in thin section (Fig. 5A) reveals that the S_1 surfaces are actually discrete zones in which layer silicates aligned in S_{1E} throughout the rock are deformed and concentrated. Thus S_{1E} is assigned to an early phase in the development of the Baraboo syncline. Intersections of all three cleavages (S_{1E}, S_1, S_1') with bedding are horizontal and trend east-west (Fig. 28).

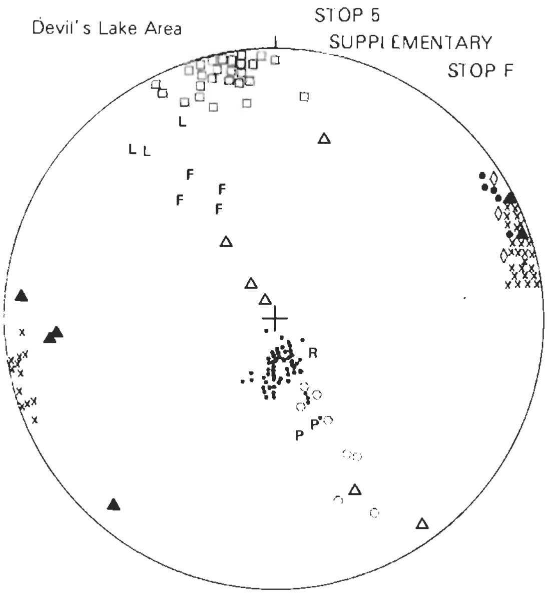


Figure 28. Stereoplot of structural data for the Baraboo Quartzite of the Devils Lake area (STOP 5). (See Table 13 for key).

There is a faint down-dip lineation on the discrete S_1 surfaces, caused in part by a longrain (or mineral alignment) lineation, but more by a fine crenulation of the layer silicates forming the S_1 surface. Slickensides on bedding planes of the quartzite also plunge down-dip. It is suggested (see general discussion of structural geology) that these structures result from the following mechanisms:

Slickensides on bedding planes - slip along bedding during folding.

Longrain - linear alignment of layer silicates in direction of maximum finite extension within phyllitic cleavage surfaces.

Fine crenulation - slight shortening parallel to the hinge line of the syncline (Fig. 7)

The fine crenulation actually forms a very faint steeply dipping and north-northwest - south-southeast trending crenulation cleavage in the thin zones of concentrated layer silicates forming the discrete S_1 surfaces (Fig. 4).

Finally, the prominent vertical north-south striking joint surfaces forming the cliff are again the "ac" joints most common throughout the syncline.

From the top of the East Bluff, the dip of the quartzite is apparent across the lake. Intersection of bedding and jointing has favored the formation of large talus accumulations not commonly seen in this part of the continent. It is felt that most of the talus formed as a periglacial phenomenon, and the abundance of large trees growing on the slope indicates little new accumulation at least during the past century. The terminal moraine dam at the north is apparent in the wooded ridge directly north of the lakeshore pavilion. On a clear day one can discern the moraine extending north-northwestward across the synclinal basin. The morainal dam at the southeast end of the lake is not visible from here, but will be seen enroute back to Madison.

Several potholes of different types occur in the quartzite near the south end of the East Bluff Trail (Fig. 16). Their age and geomorphic significance have been controversial for years. Similar potholes at the east end of the syncline contain indurated lower Paleozoic-type sandstone. (See general discussion of glacial geology by Black, and description of Supplementary Stop M).

Color-banding easily confused with stratification is displayed in the quartzite in this area. Along the trail in the cliff called Elephant Rock, we see a well-exposed example of Paleozoic basal conglomerate resting upon the quartzite in a low-angle unconformity. Quartzite boulders up to 5 or 6 feet in diameter occur at the base, but size decreases upward, and above 15 feet, there is only tan quartz sandstone. This locality is typical of the unconformity relationships along most of the north side of the South Range, that is the basal conglomerate is not very thick at any one point, and boulders more than about 2 feet in diameter occur only on the unconformity itself. Moreover, because of the low angle of the unconformity surface, the breadth of the conglomeratic facies in plan view may be more than a mile (unlike the "buttress" effect in the Upper Narrows seen at STOP 2).

No fossils are known here, but the elevation of 1200 feet is such that Elephant Rock may be laterally equivalent either to the Jordan or St. Peter Sandstones, which are virtually identical lithologically. As we have seen, near the old quartzite hills, initial dips up to 10 degrees are common in the Paleozoic strata. Therefore, elevation alone must be used with caution in correlating the atypical conglomeratic rocks with the more typical Cambro-Ordovician succession deposited farther from the quartzite hills. It is largely for this reason that the conglomeratic facies is shown as a separate map unit of variable age (Pls. I and II).

Return down the trail to the buses.

Turn around in parking lot

89.3 Return north to main road through park; proceed north at railroad crossing.

89.7 Turn left on County DL.

90.1 Cross Wisconsin 123.

90.3 Junction with South Shore Road

Turn left

90.6 Large blocks of conglomeratic sandstone in field at right. This is part of a large area of such sediments that lap onto the north side of the South Range quartzite. Here they are probably equivalent to the Jordan Formation.

91.1 Cross topographic divide, which is considered by some to have been an overflow outlet of Glacial Devils Lake into Skillet Creek toward the northwest.

91.6 Road swings left down to south shore of Devils Lake. From this turn all the way down to the lake, we are travelling over Cambrian conglomerate, which apparently filled an old valley within the South Range. Quartzite occurs both to the north and south. (See Fig. 26).

92.1 Large slumped blocks of conglomeratic sandstone at the south end of the West Bluff Trail; none are in place.

92.3 Southwest shore of Devils Lake with an excellent panoramic view of the East Bluff across the lake (See Fig. 15). The northern dip of the quartzite is evident as is the prominent jointing. The jointing dips steeply south, and is probably a megascopic manifestation of the S_1' cleavage, which it parallels. Note the large talus blocks controlled by the intersection of bedding and joints. The terminal moraine can be seen blocking the valley behind the pavilion at the north end of the lake, and similarly east of the south end of the lake. The course of the moraine between these two points is a complex double arc over the East Bluff, extending out of sight approximately three miles to the east. (See Fig. 26).

92.7 South end of Devils Lake. Extremely large talus blocks on right of road. On the South Bluffs quartzite *in situ* is now confined to the very top of the ridge. Note dense growth of large trees demonstrating that this talus slope has been inactive for a long period. Note also, the contrast in density of vegetation on this north-facing talus slope with the bare south-facing and west-facing ones across the lake (see summary of Ecology of the Baraboo Hills).

93.1 South Shore campground

93.6 Cross the summit of the terminal moraine east of the lake. The bulk of the material exposed in pits is outwash and deltaic sand and gravel.

94.3 Large quarry of quartzite on cliff face to the left. The bluffs east of here were glaciated by the ice that deposited the debris in the valley. The terminal moraine is on the upland to the left (north) (see Fig. 26).

95.0 Cliffs at eleven o'clock are of Cambrian sandstone overlying Baraboo Quartzite, which is exposed in the woods below.

95.8 Junction with Wisconsin Highway 113

Turn right (south)

96.0 Junction with road to Parfreys Glen on left (possible detour for high buses; see mile 98.3). At two o'clock, north-dipping quartzite on Devils Nose, the east end of the ridge south of Devils Lake, can be seen in railroad cuts (see Fig. 26). The terminal moraine arcs over the top of the Nose.

98.0 Water at left is a bay of Lake Wisconsin formed by a dam on the Wisconsin River near Prairie du Sac.

98.3 Junction with Wisconsin Highway 78 (Just before junction is an underpass with only 10.5 feet clearance; larger buses must detour east on Parfreys Glen Road and south to Merrimac, which is 2.5 miles east of this junction).

Turn right (west) on 78

For the next six miles we shall be crossing irregular topography produced by stagnant ice behind the terminal moraine; erratic boulders are fairly common.

99.7 Lake Wisconsin (Wisconsin River) on left.

101.1 Grounds of Badger Ordnance Works behind fence on right.

104.3 Junction with Business Route U.S. 12; bear left on 12 and 78. Abrupt change in topography marks the boundary of the terminal moraine (to the north and east) with an extensive outwash plain to the west. Here the present course of the Wisconsin River passes along the east side of this old outwash plain. At this point the terminal moraine crossed the river and dammed it up in immediate post-glacial time to form Glacial Lake Merrimac. Eventually the dam was breached and the lake drained, only to be partially "restored" later by a man-made dam. (The elevation of the present dam is 774 feet whereas Lake Merrimac reached 1000 feet.)

105.3 Entering Prairie du Sac. Lower bluffs on opposite side of river are in glacial sand and gravel; upper bluffs are Cambrian sandstones capped by the Oneota Formation. The northernmost, Blackhawks Bluff, was named for a famous chief who led the last Indian uprisings in the region (1832). Many Wisconsin names, such as Prairie du Sac, reflect the early entry of French fur traders and missionaries into Wisconsin (1634). The French travelled by canoe from the Great Lakes and Green Bay to the Fox River; they portaged from the Fox River across to the Wisconsin River just east of the Baraboo Ranges, and thence boated down the Wisconsin to the Mississippi River. The name "Wisconsin" derives, via the French "Ouisconsin", from an Indian word interpreted to mean "gathering of the waters."

106.8 Entering Sauk City, named for the Sauk Indians, who had an important village here.

107.6 Junction U.S. 12
Turn left, crossing Wisconsin River, and proceed south on 12.

107.8 Junction; large sand pit at left in glacial outwash material.

109.3 Travelling over outwash plain. Here we parallel approximately the terminal moraine, which is not so obvious in this area, but it can be traced south for many tens of miles; it is obvious again west of Madison. On the way back to Madison we shall be travelling over glacial material in which erratic boulders can be seen. Only locally do Paleozoic rocks appear as outcrops within this region.

118.8 Roadcut in Oneota Dolomite at left.

119.4 Roadcut in Oneota Dolomite with a pronounced eastward dip.

122.4 Outcrop of brown St. Peter Sandstone in hill just west of red barn on right.

124.1 Middleton city limits.

We descend a small hill onto the flat bed of Glacial Lake Middleton. The lake extended approximately two miles west, and drained to the west unlike present Lake Mendota (located one mile east of here), which drains southeast.

125.2 Major junction: To go into Madison - exit from four lane road to right and turn left on University Avenue; to go south to Illinois, stay straight ahead on U.S. 12/14 bypass; to go west toward Mississippi River, turn right and follow U.S. 14.

End of First Day Road Log.

SECOND DAY

Madison--Baraboo--Milwaukee

0.0 Road log begins in Middleton at junction of U.S. Highways 12 and 14 with University Avenue.

Drive north on U.S. 12 (route to Sauk City is same as that for first day's return to Madison).

17.6 Center of Sauk City

Continue west on U.S. 12.

19.4 Highway 12 swings to the north. Consult Figure 18 for today's route. Travelling across the outwash plain again. Small rise ahead is a river terrace covered with pine trees about 25 feet higher than the level at the curve. Old dunes, now stabilized, cover the terrace. Coming up onto the terrace, the South Range appears straight ahead.

22.4 Divided highway begins.

To the left (west) on the skyline are wooded, mesa-like hills underlain by Cambrian sandstone capped by Oneota Dolomite (see Pl. I).

24.6 Badger Ordnance Works straight ahead.

At eleven o'clock is the mouth of Baxters Hollow, wherein a granite and diorite complex lies structurally beneath the Baraboo Quartzite. Relative age relations are somewhat uncertain, as are Rb-Sr dating results (see Table 2).

27.8 Enter roadcut in the quartzite escarpment. Outcrops by roadside are highly fractured, and are veined with quartz.

28.3 Small outcrop of Cambrian conglomeratic sandstone on left side. This is another example of fill in an ancient valley that notched the entire southern quartzite ridge.

29.2 Divided highway ends.

30.8 Supplementary Stop G.

Large outcrop of quartzite to right of road, with exceptionally clear current bedding, some of which is the trough type. In the south end of the exposure, there are several zones of contorted cross stratification formed by synsedimentary deformation (probably resulting from loose packing and shear by a sudden current pulse). This outcrop exemplifies the north-to-south paleocurrent pattern of the Baraboo Quartzite as a whole (see Fig. 2). Above the small quartzite cliff (and also on the west side of the road) the phyllitic zone in the upper Baraboo Quartzite has well developed structures of all phases of deformation clearly exposed (see description of Supplementary Stop G).

31.4 Junction with Wisconsin 159.

Turn right on 159.

Reminder that terminal moraine again is just ahead of us.

32.0 Entrance to Skillet Creek Farm.

STOP 6 -- Skillet Creek - East Side.

After obtaining permission from the owners of the property, we shall drive as far south as possible down the west side of Skillet Creek to the turning circle in the campground. Cross the creek (at present there is a convenient fallen tree at this point) and proceed 100 yards upstream (south) and sixty-five yards east. This valley may have been the outlet of Devils Lake during its maximum-fill stage. We come to a large outcrop consisting of a massive quartzite overlain by a zone 6-8 feet thick of phyllite with interbedded quartzite layers up to a foot thick, and finally a zone of almost pure phyllite also 6-8 feet thick (Fig. 29a). DO NOT HAMMER THIS LOCALITY!

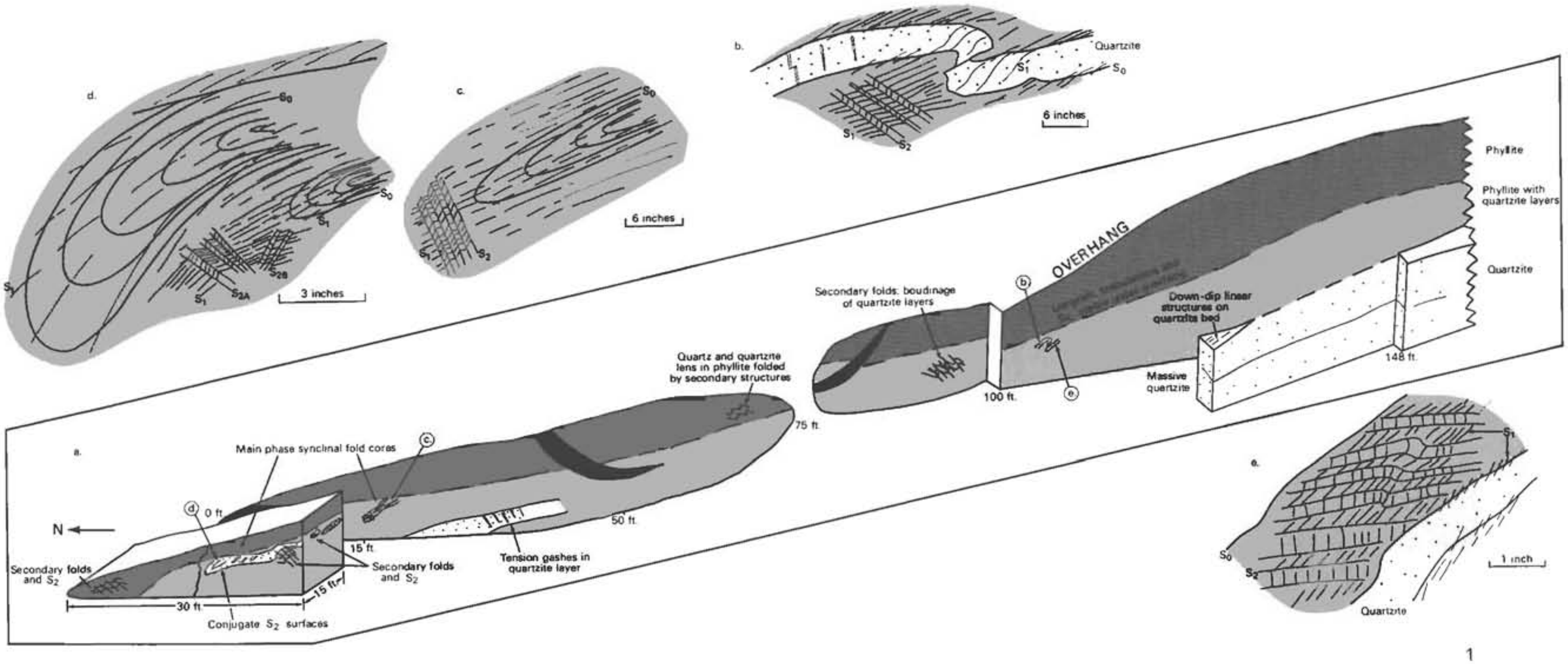


Figure 29. Structures in the phyllitic zone of the Baraboo Quartzite at Skillet Creek (STOP 6).

a. Diagram (modified after Adair, 1956) showing the outcrop on the east side of Skillet Creek.

b. Main phase minor fold with axial planar phyllitic (S_1) and quartzite (S_1') cleavages, the former deformed by a crenulation cleavage (S_2).

c. Synclinal main phase fold with axial planar S_1 cleavage deformed by S_2.

d. Synclinal main phase minor fold with fanning axial planar S_1 cleavage deformed by conjugate S_2 crenulation cleavages.

e. S_1 cleavage in phyllitic layer deformed by S_2 crenulation cleavage. The S_2 cleavage also appears to be deformed by what is here called a late phase minor fold (Table 3).

It should be noted that the phyllite here is part of a zone at the top of the exposed Baraboo Quartzite on the south limb of the syncline which can be traced southeastward from here to Happy Hill (Supplementary Stop H) (see Pl. III). This is a true phyllite, unlike the layers previously seen at Van Hise Rock (STOP 2) and the northeast corner of Devils Lake (STOP 5), which are merely significantly phyllitic by contrast with the massive quartzite beds.

The bedding here dips north at 20-35°. The S_1' cleavage is best seen in the lower massive quartzite zone, where it is nearly vertical and strikes east-west. Apparently strain in the thinner quartzite layers within the phyllite was taken up largely by buckling, for the cleavage is not obvious in them (see discussion of general structural geology).

In the interbedded phyllite/quartzite zone near the southern end of the outcrop under an overhand, there is a small asymmetric fold with a southerly vergence (sense of overturning) and a sheared middle limb (Fig. 29b). This has been called a "normal drag fold" because its vergence is congruent with its situation on the south limb of the Baraboo Syncline. Cleavage in the phyllite (S_1 is axial planar to this fold, and it undoubtedly was formed during the main phase of deformation of the Baraboo Quartzite. The term "drag fold", however, is to be avoided because structures such as this may form in other ways (for instance by buckling) and in fact probably do in most cases.

In the core of the anticlinal portion of this fold the phyllitic cleavage apparently was protected from further deformation by the quartzite layer. Tracing S_1 down-dip, however, it is seen to be deformed by a south-dipping crenulation or strain-slip cleavage (Fig. 29b). This cleavage is prominent in the phyllitic zone here, and is axial planar to a set of asymmetric "chevron" style folds with a northerly vergence. The term "reverse drag folds" has been applied to these folds for many years because they are not congruent with the geometry of the syncline. They clearly represent a later phase of deformation than the so-called "normal drag folds," the phyllitic cleavage, and the quartzite cleavage. Once again, use of the term "drag folds" is inadvisable for these secondary folds.

Although the secondary structures mostly have south-dipping axial surfaces, at many places in the phyllite outcrop there is a less obvious conjugate set of north-dipping crenulations (see Fig. 29, and Fig. 3 for diagrammatic representation of crenulation cleavage geometry). Numerous explanations of the "reverse" or "anomalous" structures of the phyllitic zone in the upper Baraboo Quartzite have been put forward (see Adair, 1956). Most recently gravitational gliding of the upper beds back towards the core of the syncline on relaxation of regional stress has been proposed (Hendrix and Schaiowitz, 1964). This type of structure is a common compressional feature in the later stages of the polyphase deformational history of mobile belts, however, and the conjugate geometry makes an explanation such as that proposed by Hendrix and Schaiowitz less appealing (see general discussion of structural geology).

While the chevron folds and associated crenulation cleavages are here assigned to a secondary deformation phase because they deform S_1, the parallelism of their axes with that of the syncline suggest that they merely may represent a later stage in a continuous deformation process.

Beneath the main phase fold shown in Figure 29a, there is some slight warping of the crenulation cleavage indicative of a still later phase of deformation (see Fig. 29e and Table 3).

Down-dip slickensides on the bedding, longrain, and fine crenulations on S_1 are visible as at Van Hise Rock (STOP 2) and the northeast corner of Devils Lake (STOP 5).

Lower down the outcrop (north) there are many isolated synclinal cores of main-phase folds to which S_1 is axial planar (Fig. 29). Apparently the anticlinal cores have been sheared out. One of the synclines at the extreme northern end is deformed by an open secondary phase fold.

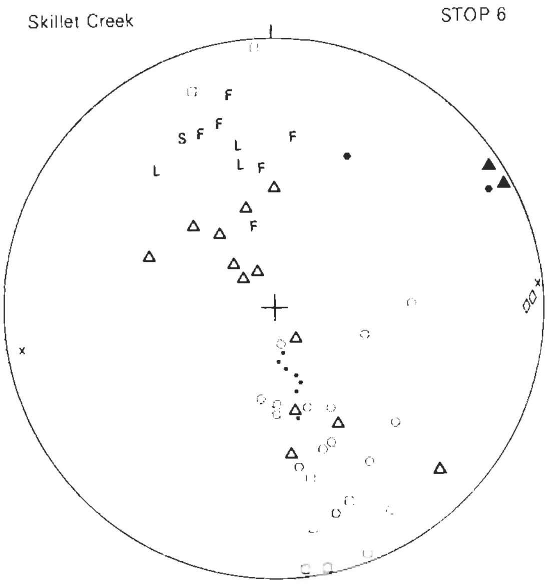


Figure 30. Stereoplot of structural data from the Baraboo Quartzite of the Skillet Creek area (STOP 6). (See Table 13 for key).

Boudinage of the thin quartzitic layers in the phyllite is common. Some of the boudins were deformed by second-phase folds, and at least partially recrystallized to give rise to white quartz lenses in the crenulated phyllite. The boudins have near horizontal east-west axes.

Return west on Wisconsin 159.

32.6 Junction U.S. 12.

Turn right (north) on U.S. 12.

33.1 Turn left on Gasser Road.

33.2 Crossing Skillet Creek again.

33.3 Ridge to left is capped by Black Earth Dolomite and has the old Woods Quarry at the south end exposing spectacular cylindrical algal stromatolites four feet tall. The dolomite there figures in the famous E.O. Ulrich stratigraphic controversy over the Cambrian-Ordovician boundary. It was considered "Lower Magnesian Limestone," i.e., Oneota, by early workers (see discussion of Paleozoic stratigraphy).

34.3 T Junction.

Turn right on Lehman Road.

34.8 T. Junction.

Turn left on County W.

For the next 5.4 miles we are retracing yesterday's route between mileages 76.9 and 82.3.

38.9 T Junction - continue on County W to left.

39.0 Continue on County W to right (west).

The road to the south at this junction leads to Supplementary Stop H southwest of Happy Hill School where excellent secondary phase structure can be seen in the phyllitic zone of the upper Baraboo Quartzite (Fig. 51). The same road also crosses the South Range and provides a short cut to the town of Denzer (gravel most of way).

40.2 Junction of County PF and W.

Continue straight ahead on PF.

40.4 Galesville Sandstone on the left as we enter La Rue.

40.6 Cross Seeley Creek.

40.7 Turn left on un-named road. Large quartzite quarry in distance on slope of South Range is our next STOP.

41.4 Entrance to La Rue Quarry. Galesville Sandstone on left.

STOP 7 -- LA RUE QUARRY.

The railroad spur, which runs into the quarry from North Freedom, was formerly used for the quarry and the iron mining industry, but is now used for steam train excursions. It is known from subsurface investigations that two Precambrian lithic units overlie the Baraboo Quartzite in the La Rue area. These are the Seeley Slate and the Freedom Formation (Pl. II). The Freedom Formation is unconformably overlain by Cambrian sediments, and its lower part is iron bearing. The Illinois and Sauk mines were operated near here (and the Cahoon mine 7 miles east) early in the century.

This old quartzite quarry (see Fig. 31) provides spectacular exposures of the Cambrian-Precambrian unconformity, of structures in the upper Baraboo phyllitic zone, joints, and an unusual weathered rind on the quartzite. The western part of the quarry exposes the unconformity high on its south face; angular blocks up to six feet long and rounded boulders up to 4 feet in diameter occur in a conglomerate lens within sandstone that is equivalent laterally to the Tunnel City Formation (Fig. 32a). On the opposite, low north face of the quarry, a jumble of angular blocks up to four feet long occurs along an undulating contact.

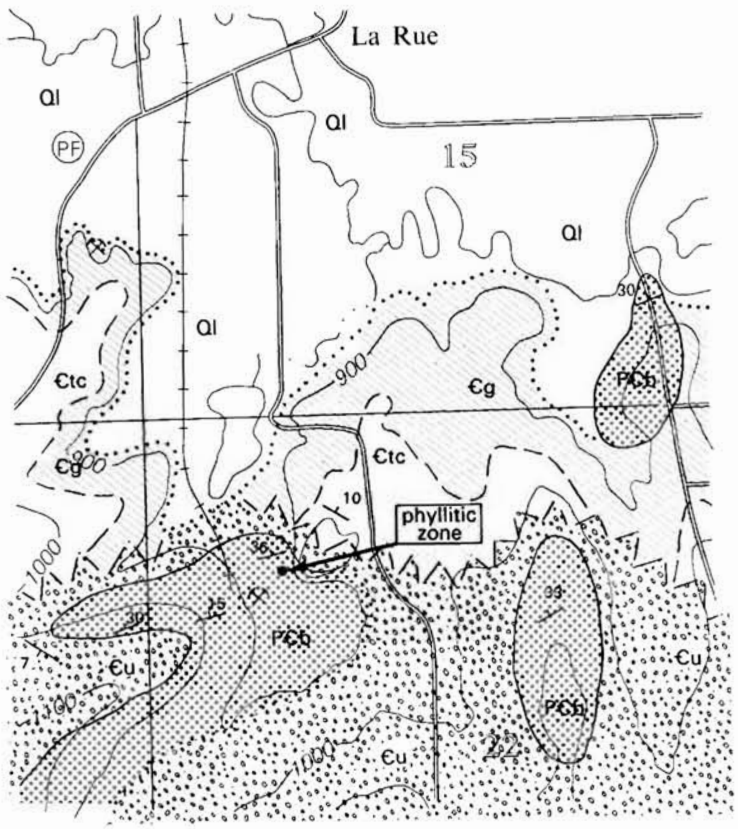
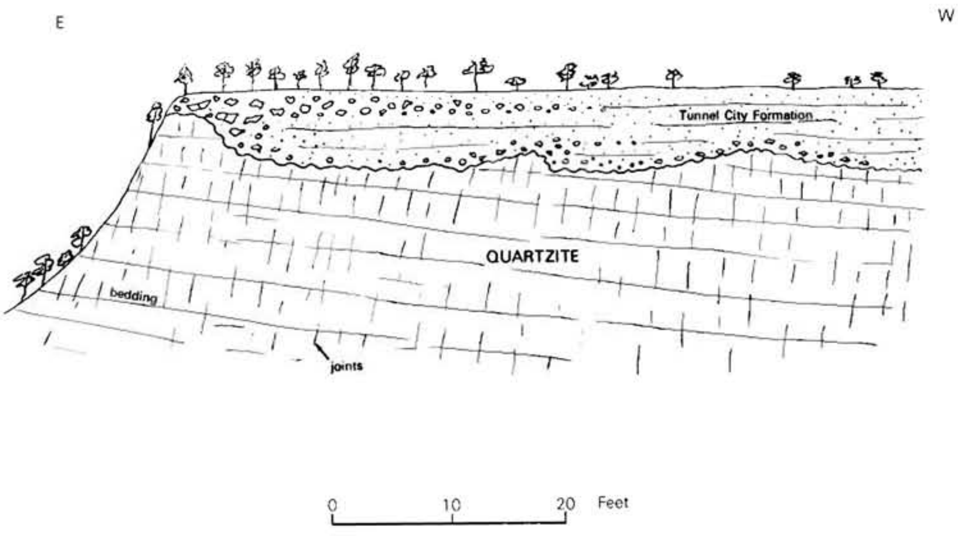
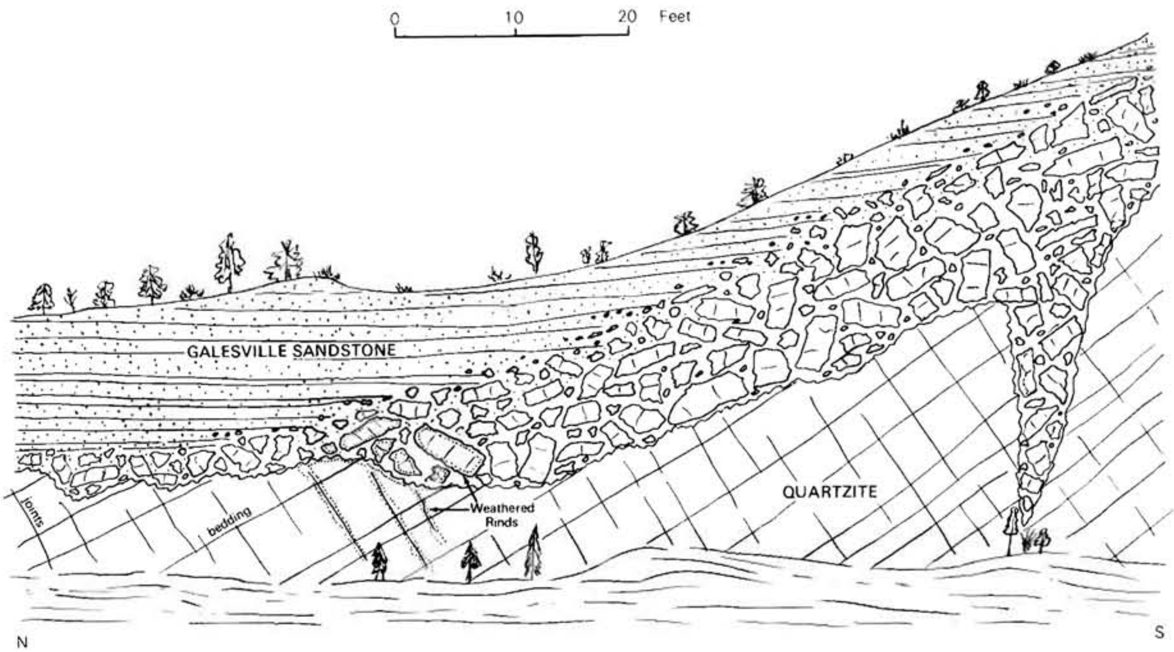


Figure 31. Geologic map of the La Rue quarry area (STOP 7), showing three quartzite hills surrounded by Cambrian strata. Two of the three iron mines of the Baraboo District were located just north of this map area. (See Fig. 21 or Pl. I for key to symbols). (Modified from Usbug, 1968).

Figure 32. Sketches of Cambrian-Precambrian unconformity in La Rue Quarry:

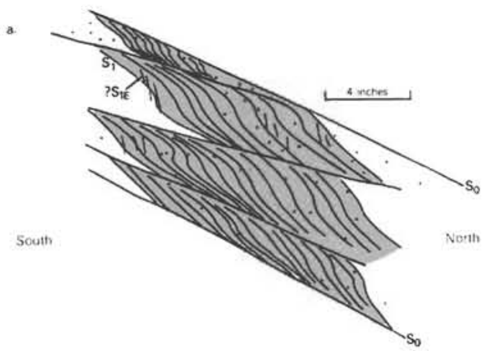


A. High southwest face of old quarry showing conglomerate and sandstone probably equivalent to Tunnel City Formation) resting upon Baraboo Quartzite. Note large, angular quartzite clasts near buried knob at left and conglomerate tongue extending to west (right). Smaller quartzite clasts are all well rounded and clearly were transported.

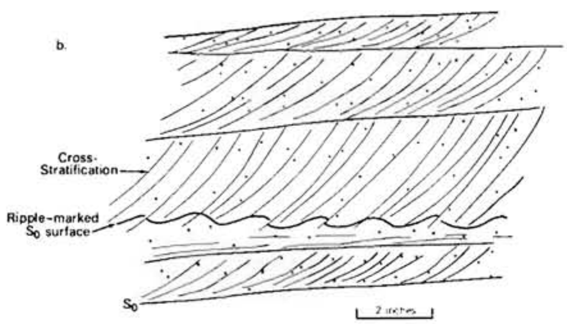


B. East face of quarry showing non-conglomeratic Galesville Sandstone onlapped over angular quartzite blocks that rest upon the unconformity (note deep fissure filled with quartzite blocks at right). Note bleached, weathered rinds on both quartzite bedrock and blocks discussed in text.

Figure 33. Sketches of structural relations at La Rue Quarry (STOP 7).

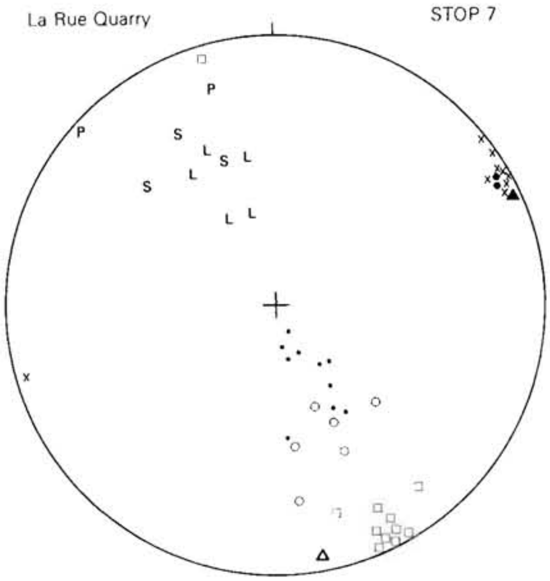


a. Bedding/cleavage relations in phyllitic quartzite, southwest corner of eastern part of quarry.



b. Relations of bedding, cross stratification and ripple marks, illustrating modification of ripple marks by slight slip along cross stratification surfaces during tectonic deformation. This can be seen on some of the large blocks high on the slope behind the south end of the eastern part of the quarry.

Figure 34. Stereoplot of structural data from the Baraboo Quartzite at the La Rue Quarry. (See Table 13 for key).



In the eastern part of the quarry, the unconformity is most spectacular. Here the quartzite dips 30°-40° to the north, and is prominently jointed; cross stratification also is apparent. White Galesville Sandstone overlies the quartzite in the northeast corner, and contains very few Baraboo fragments there. In the east face, however, Galesville strata lap higher and higher onto the sloping quartzite surface, and a jumbled zone 10-20 feet thick of angular quartzite rubble with blocks up to 8 or 10 feet long rests along the unconformity (Fig. 32B). Near the middle of the sheer east face of the quarry, a V-shaped fissure extending about 20 feet down into the quartzite is filled with angular blocks as well. At the top of the cliff to the southeast, brown sandstone represents the Tunnel City Formation, which finally buried this quartzite hill.

The abruptness with which the basal conglomerate passes northward into clean, white sandstone is remarkable. This feature, together with the angularity even of the smaller boulders--as contrasted with most of the conglomerate at previous stops and on the high, south face of the quarry here--reflects either infiltration of talus by wind-blown Galesville sand, or by water-borne sand in a situation where physical energy of waves was not competent to move the boulders (see general discussion of Cambrian sedimentology).

Bleached white, friable weathered rind from one to two inches thick on quartzite joint surfaces and on conglomerate fragments is a curious phenomenon at this locality. Weathering somehow has broken down interstitial silica so that individual sand grains stand out in relief. The age of the weathering is as uncertain as is the mechanism. Likely it was post-Cambrian, for the conglomerate boulders show it as well as underlying quartzite joint surfaces. If the weathering were all before Late Cambrian, one might expect some of the rind to have been removed from the boulders by abrasion. M.E. Ostrom (1966) feels that this locality indicates a probability that the Baraboo Quartzite was broken down sufficiently to provide much of the local Cambrian sands. He has observed similar weathering of quartzites at Rib Hill and Barron in Wisconsin, and at New Ulm, Minnesota. But this is an unusual locality for the Baraboo district, and petrologic evidence suggests that the Quartzite provided very little sand (see Tables 6-7).

<u>Structures in the Quartzite</u> (PLEASE DO NOT HAMMER HERE)

The general attitude of structures in the north-dipping quartzite of the quarry is shown in Figure 34. In the southwest corner of the eastern portion of the quarry, a well developed cleavage is refracted through successive impure quartzite layers, and subsidiary shear zones and crenulations developed (Fig. 33a). The amount of layer silicates (mostly pyrophyllite) crystallized along the cleavage planes is surprising, for the rock between the discrete cleavage surfaces looks like a pure quartzite (see general discussion of quartzite lithology). There is no real distinction between the quartzite and phyllitic cleavages here. Down-dip <u>longrain</u> and slickensides are present, and close examination reveals S_{1E} in the rock between the discrete white $S_1{'}/S_1$ surfaces.

On the talus slope above this part of the quarry, there is a bewildering variety of deformed ripple marks. Both cross bedding and ripple-marked bedding surfaces are invariably slickensided, and slip along the ripple-marked surfaces, and in particular along cross bedding surfaces, tended to distort the ripples (Fig. 33b). Bedding/cleavage intersections also modify the appearance of ripple marks, but they do not appear to have resulted in any "pseudo-ripples."

Return north to County PF.

42.0 Turn left on County PF.

42.1 Railroad crossing -- look out for steam trains!

42.5 Quarry in Galesville Sandstone, which lacks Baraboo pebbles at this locality even though it is very close to the South Range. On the floor of the quarry, festoon cross-stratification is exposed in plan view.

43.0 Fleeting view of La Rue Quarry to left (east).

At this elevation we are crossing the Tunnel City Formation, which is not well exposed there.

44.5 Highway makes right angle bend to west.

Upland surface here is underlain by cherty residuum of the Oneota Formation and by the Jordan Sandstone. Quartzite lies a short distance to the left (southeast).

47.7 In the woods to our left (southeast) as the road takes a right angle bend to the west (at a junction) are the most westerly outcrops of the large quartzite inlier of the South Range. We drive due west into more open country again underlain by the Oneota Formation. Note on the map the strike of the quartzite in this vicinity, which swings rather sharply to a northeast-southwest trend.

48.4 Scattered blocks of quartzitic sandstone on the highest part of this large upland may represent St. Peter Sandstone; however, no actual outcrops are known.

48.4 Junction with County D.

Continue left on County PF.

In the valley of Honey Creek, half a mile west of this junction, the Baraboo Quartzite can be seen in the stream bed forming what is known in the geologic literature as Weidman Falls (see Thwaites, 1958). This is the southwesternmost outcrop of the quartzite; it strikes north-south, is nearly vertical, and faces west. Clearly, the structural trend of the quartzite is substantially changed from that seen throughout most of the Baraboo district. The Weidman Falls inlier is surrounded and overlain by the upper Tunnel City Formation with local conglomerate; Black Earth Dolomite directly overlies the latter.

49.8 At a sharp bend in the road, basal contact of the Oneota Dolomite and the Jordan Sandstone at right; oolite is present at the contact. Steep grade for the next mile (taking us out of the Baraboo basin) passes down over Jordan Sandstone, St. Lawrence Formation (poorly exposed), and the Tunnel City Formation, which contains some conglomerate.

50.5 At the bottom of the grade, we see excellent exposures of the Galesville Sandstone, which is almost completely free of conglomerate. Typical Galesville trough cross bedding sets can be seen (right) separated by parallel, planar truncation surfaces.

50.9 Junction with Bluff Road to left (Wisconsin Society of Ornithologists' swamp preserve to right). Keep ahead on PF, but note cliffs along Bluff Road: the lower white rock is the Galesville; the upper buff-colored rocks are Tunnel City strata. The contact is conspicuous all along the Honey Creek valley. The Tunnel City is more resistant, so forms a cap rock above sheer cliffs. Honey Creek Valley was flooded temporarily by a lake as Late Woodfordian ice melted farther east. The lake was dammed by glacial outwash debris in the Wisconsin River Valley about 10 miles southeast of here.

52.3 Ridge to the right (west) again shows Galesville-Tunnel City contact very clearly.

52.8 Junction with County C. This valley is part of the Late Wisconsinan lake bed, which contains many tens of feet of sediments with buried organic layers. Mammoth remains were discovered a few years ago three miles downstream.

Turn left on C into town of Leland where we shall lunch in the town park. Water is available from an old fashioned pump; other fluids at "city hall" across street.

53.1 After lunch, proceed northeast on Highway C.

53.3 Junction; turn left on Bluff Road (see Fig. 35).

54.4 Roadcut in upper Galesville Sandstone; Tunnel City Formation just up the hill in the woods.

54.6 Junction -- keep going straight ahead. Across the valley at ten o'clock in Galesville Sandstone (in woods) is the Durst Archaic Indian rock shelter. It is a small, amphitheatre-shaped cave much like the larger cliff shelters of the Southwest.

55.1 Lane to right leads up the east side of the valley to Hemlock Draw; we leave the buses here and walk approximately one mile to STOP 8.

STOP 8 -- LOWER HEMLOCK DRAW

This is our first stop outside the Baraboo basin to the south.

At the mouth of Hemlock Draw there is an example of a buried quartzite sea stack and spectacular coarse conglomerate. We must walk about one-half mile up an abandoned road and through woods to see this locality, but we think that you will agree that it is worth it. Our botanist colleagues, who have had the hollow designated a State Scientific Area, hope that we will not trample the undergrowth unduly. This and other similar deep valleys in the area contain examples of an exotic hemlock-yellow birch flora (see discussion of Ecology of the Baraboo Hills).

On our right as we start up the old road is a cliff of Galesville Sandstone in which Baraboo pebbles are rare and of small size. About one-third of a mile up, we cross the creek (Fig. 35); 100 yards up the creek from the old bridge, at least 15 feet of conglomerate is exposed on the north bank, with rounded boulders up to 2.5 feet in diameter. It is a lateral facies of the non-conglomeratic sandstone we have just passed; the gradation between must be remarkably abrupt.

Continue on the old road, which bears right and uphill to a large quartzite outcrop completely surrounded by Cambrian strata. We interpret this as a buried sea stack; it is one-half mile from the nearest larger exposures of quartzite. A few other probable stacks are exposed nearby to the southeast (Fig. 35). The quartzite dips northwest at between 15 and 30°, and the essentially vertical S_1' cleavage has a strike of about 30°, which indicates that this mass is in place. Note that the regional strike of the Precambrian rocks swings to northeast-southwest in this southwesternmost exposed portion of the Baraboo syncline (Fig. 35 and Pl. III). A few thin phyllitic partings occur with cleavage (S_1) dipping northwest at 50-60°. Bedding/cleavage intersections are nearly horizontal and trend northwest-southeast. A few tension gash bands are visible.

We leave the road at the quartzite stack to follow the topographic contour westward through woods; walking will be a bit rough on this steep, south-facing slope. The next outcrops are of Cambrian Sandstone with conglomerate zones containing rounded boulders up to four feet in diameter; percussion marks can be seen on some boulders. These rocks are roughly equivalent to both the Tunnel City and Trempealeau strata, and are overlain on the upland surface 100 feet above and north, by residuum of the Oneota Formation. Moving on west we find similar strata, some of which show excellent cross stratification. About 200 or 250 yards west of the "stack", we see the largest quartzite boulder known in the region -- it is more than 20 feet long and is angular. Sandstone with finer rounded conglomerate fragments surrounds it completely. Several other blocks from 6 to 10 feet long occur nearby. Apparently the quartzite blocks larger than about six feet fell from sea cliffs and became buried with little or no movement or abrasion by wave action. The largest blocks probably fell from a now-buried cliff very nearby to the north rather than from the "stack" seen to the east. Occasional breaking waves with high velocities must have pounded the cliffs in this area, and were able to move and round boulders up to 4 feet in diameter, but were unable to move larger ones (see discussion of Cambrian sedimentology, especially Table 9).

Farther west and south, the conglomerate becomes finer and less abundant; festoon cross bedding and Scolithus tubes are well displayed (Fig. 36). These are interpreted as littoral zone deposits.

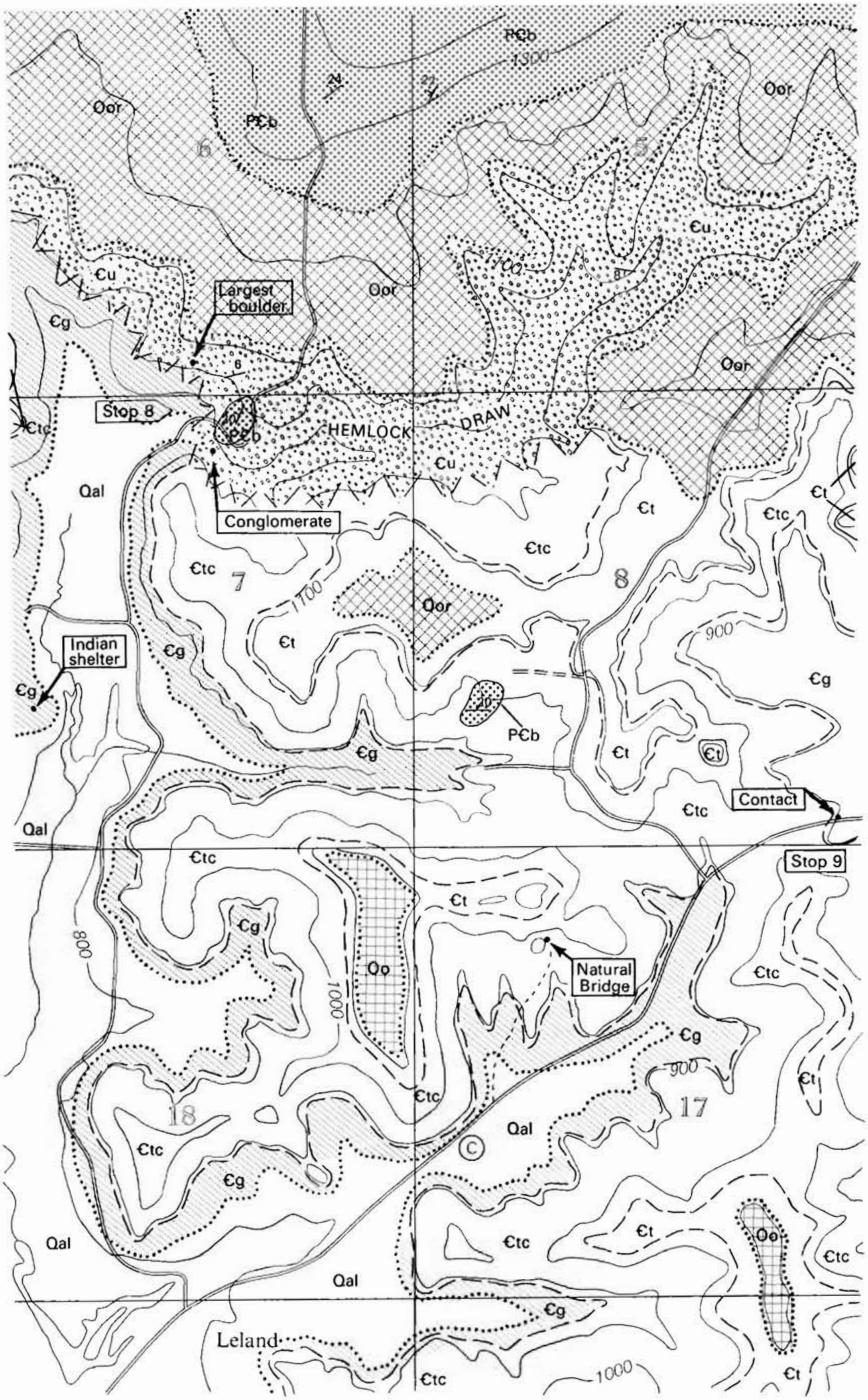


Figure 35. Geologic map of Hemlock Draw-Leland area (STOPS 8 and 9). (See Fig. 21 or Pl. I for key to symbols).

Return directly downhill (south) through the woods back to the old road, and thence to the buses.

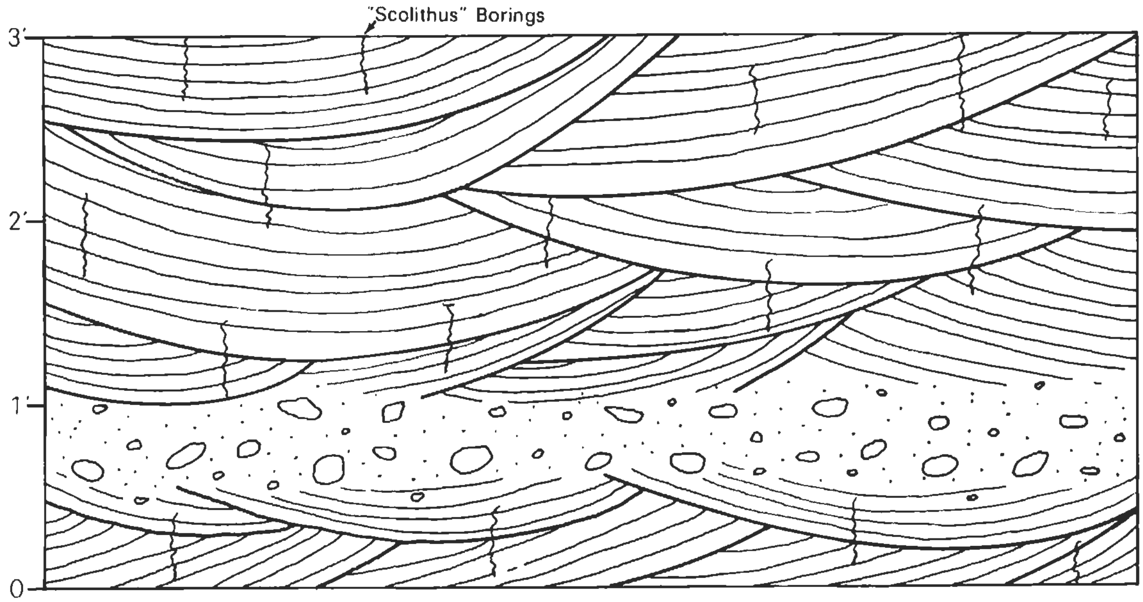


Figure 36. Cross section sketch of festoon cross bedding cut by Scolithus tubes west of the largest quartzite boulder at STOP 3, mouth of Hemlock Draw. These conglomeratic strata probably are equivalent in age to the Jordan Sandstone (Trempealeau Group).

Continue on to end of road.

55.4 Turn vehicles at farm and return south to Highway C at Leland.

57.5 Turn left on Highway C.

Note butte at eleven o'clock capped by Tunnel City Formation.

58.3 Entrance to Natural Bridge (privately owned). The bridge is cut from slightly conglomerate Tunnel City Formation sandstone. An Archaic Indian rock shelter (Raddatz, Sk-5), which was occupied first nearly 13,000 years ago, also is present under the bridge (see Black, 1959).

59.3 STOP 9 - Contact of Tunnel City Formation on Galesville Sandstone.

STOP 9 -- HIGHWAY C ROADCUT (Between Denzer and Leland)

The roadcut on the north side exposes the Galesville-Tunnel City contact and illustrates several aspects of Cambrian sedimentation south of the syncline; keep in mind that this locality is nearly a mile from any exposed quartzite (see Fig. 35), and note the contrast with STOP 8 in Hemlock Draw 1.5 miles northwest.

As seen here, this contact is typical for the Leland-Denzer region. The lower, white, very friable sandstone is Galesville with medium-amplitude cross strata faintly visible. Though it is not very clear here cross sets in this formation are generally truncated by parallel, planar surfaces. The sand is medium to coarse, well-rounded, pure quartz (maximum 0.8 mm) with very rare, angular granules and pebbles up to 1 cm of Baraboo material concentrated along a single two-inch zone. Angularity of the coarser (quartzite) clasts represents an inversion of the normal or expected rounding versus size relationship, and is further evidence of a marked difference of history (thus origin) of the quartz sand from that of the coarser, local Baraboo detritus (in general, rounding should increase with grain size). The top few inches of the white sandstone just below the wavy contact contains many tan-colored "Scolithus" borings, which are practically unknown in the typical Galesville. We presume that they were formed by organisms during a time of scour or non-deposition just before Tunnel City deposition. Immediately above the Scolithus zone is a sharp contact with 6-18 inches of brown rock containing white cobbles of Galesville sandstone up to 4 inches long. This interval represents scour and a marked change in sedimentation; because of the contrast in lithology across it, it is taken as the base of the Tunnel City Formation. Note that the rounded sandstone pebbles indicate that the Galesville somehow had gained at least slight cohesion prior to scour; the amount of time represented by the unconformity is, of course, impossible to estimate with any accuracy, and many similar scour surfaces and rip-up conglomerates characterize the Tunnel City Formation. The remainder of the cliff above is typical buff-brown weathering, very dolomitic sandstone with sporadic glauconitic zones. Cross bedding is of the trough type with amplitudes between 4 and 6 inches (rarely up to 12 inches). Truncations between most sets are neither planar nor parallel; the medium-scale festoon type cross bedding here is very typical of the Tunnel City (see Figs. 9 and 10). Orientations both of individual cross sets and of trough axes are extremely variable, with adjacent troughs plunging at right angles to one another on the hill just north of this cut. Opposite plunge directions are rather common (see Fig. 11 and Pl. VII).

The Tunnel City contains little conglomerate here, but "grit" zones can be seen both at the east end of the crop above the fence post and westward to the top of the hill. On the wooded hill south of the road, however, Baraboo cobbles up to 4 inches are present, as they are also in the Natural Bridge three-quarters of a mile west, and on a ridge 1.5 miles west of Leland. Only one mile north of here, near the quartzite outcrops in Hemlock Draw and Pine Hollow, strata equivalent to the Tunnel City have much more conglomerate, which contains clasts up to 3.5 feet in diameter. Thus this general area illustrates how abruptly quartzite gravels disappear away from their source (see Pl. VI).

Continue on Highway C

59.9 Pine Hollow Road. On the wooded hill one-quarter of a mile northeast of this junction, there is one of two inliers of Precambrian diorite. Relative age with respect to the quartzite and rhyolite is unknown because both inliers are surrounded by Cambrian sediments and the contact is not exposed. Pine Hollow Road extends for two miles up to the south face of the South Range (not recommended for buses). Pine Hollow exposes excellent outcrops of Cambrian conglomerate deposited next to an ancient quartzite sea cliff similar to that at Hemlock Draw (STOP 8). Like Hemlock Draw the Hollow itself is very scenic because of groves of exotic hemlock white pine, yellow birch and small ground plants growing on picturesque cliffs. The Nature Conservancy has been active in acquiring ecologically-choice sites in this region, including Pine Hollow, Hemlock Draw and Baxter's Hollow. It is hoped that more of the Baraboo region can be put under the supervision of the Wisconsin Department of Natural Resources, the University of Wisconsin Arboretum, or other conservation organizations.

61.2 Crossroads in Denzer - continue straight ahead; see Fig. 37. Denzer like Leland, lies on glacial lakebeds of Late Wisconsinan age.

Road to left (north) is another shortcut up over the South Range to La Rue.

61.9 Junction at bend in road.

Small quarry northeast of junction exposes contact of Galesville and Tunnel City Formations; considerable glauconite and some conglomerate occurs in the Tunnel City there (Fig. 37).

62.2 Turn left on gravel road to Denzer Quarry.

This is one of many quarries in this region from which Oneota dolomite is extracted for crushed rock for road materials.

STOP 10 -- DENZER QUARRY

This quarry exposes the uppermost Jordan Sandstone (Trempealeau Group) and the lower Oneota Formation, which caps many of the butte-like hills in this area (Fig. 37). The Cambro-Ordovician boundary occurs somewhere within this interval. The lithologies visible here are typical of these formations for the Baraboo region (the Sunset Point Member of the Jordan is not clearly developed in the Baraboo area, if indeed it is present at all). Unique structural features make this quarry of special interest--and puzzlement.

In the south part of the quarry, you will note a north dip of 10° in well-bedded, buff dolomite with thin, white chert and dolomitic sandstone layers. The Oneota here contains very sparse red quartzite pebbles up to 2 centimeters long, suggesting either that some quartzite islands still existed in Early Ordovician time or that Cambrian conglomerates were being re-worked. A small anticline near the middle of the quarry and another along the road just below the quarry entrance expose white Jordan sandstone and green siltstone. In the northeast part of the quarry, spectacular disharmonic folds occur in fine dolomite that overlies a thick dolomite breccia, the latter of which is a rather common lithology in the lower Oneota-upper Jordan interval. The crudely-stratified breccia also shows anomalous dips in the north end of the quarry. Some lenses of laminated and porous dolomite (fenestral texture) interbedded with the breccias suggest a supratidal environment by analogy with other, better-known carbonate rocks; the laminae may have been formed by algae (L.C. Pray, personal communication). Definite algal stromatolites, dessication polygons, and intraformational flat-pebble conglomerates common elsewhere in the

Oneota suggest extensive intertidal environments as well during Early Ordovician time. That the quartzite islands were very subdued by then is indicated by extensive dolomite very close to quartzite outcrops both in the west and east parts of the syncline (Pl. I).

The origin of the folding here (and extending for a least one-half mile to the east; Fig. 37) has been a puzzle to local geologists--some favoring a tectonic origin and others (probably the majority) favoring some sort of primary or syn-sedimentary processes. At least the structures in the northeast corner of the quarry must be syn-sedimentary because of the clear evidence of ductility and of intraformational truncations (Fig. 38). Proximity to known ancient hills of Precambrian rocks (Fig. 37) has suggested the possibility of initial dips and thickness variations followed by differential compaction; however, the lithologies visible here could not have been compacted significantly. Also appealing is the postulate of sea floor irregularities related to small algal reefs, which are not actually seen here but which are common in the Oneota elsewhere. The stratified breccias suggest material torn by waves from a supratidal or shoal-water carbonate prominence, while the disharmonically-folded and truncated fine-grained dolomite above points to syn-sedimentary slipping of cohesive sediments down the flank of such a prominence. The larger-scale folding of the Jordan and Oneota strata is more puzzling, especially near the top of the road grade where sandstone clearly is faulted against dolomite (Fig. 39). Such brittle behavior could be interpreted as a post-sedimentation feature, thus tectonic. However, lithification of carbonate rocks may occur very early, making brittle failure possible even during early post-depositional disturbances. We favor the interpretation that all of the structures here are essentially syn-sedimentary, and are the result of differential diapiric upward movement of unlithified, water-laden sand sealed beneath semi-lithified carbonates. Gravity-induced failures around prominences of the cohesive carbonate strata occurred at the depositional interface. While either an algal reef or a basement hill could have been the ultimate cause, through initial draping of sediments, it seems more probable to us that excess pore-fluid pressure in sealed, water-saturated Jordan sands was the initial cause of instability.

Proceed south on Highway C.

63.2 Straight ahead on Prairie Road. Junction with Prairie Road; one mile northeast of the last stop there is an outcrop of rhyolite and associated tuffaceous sandstone known informally as the "Denzer tuff". It was first described and identified by Stark (1930; 1932), and is a part of the pre-quartzite rhyolite volcanic sequence.

64.0 Junction to left (contact of Galesville and glauconitic Tunnel City Formation just above junction in ditch).

<u>Keep right</u>.

64.4 Junction -- bear <u>left</u> on Prairie Road.

65.4 Junction -- keep straight ahead.

66.5 Junction -- keep straight ahead.

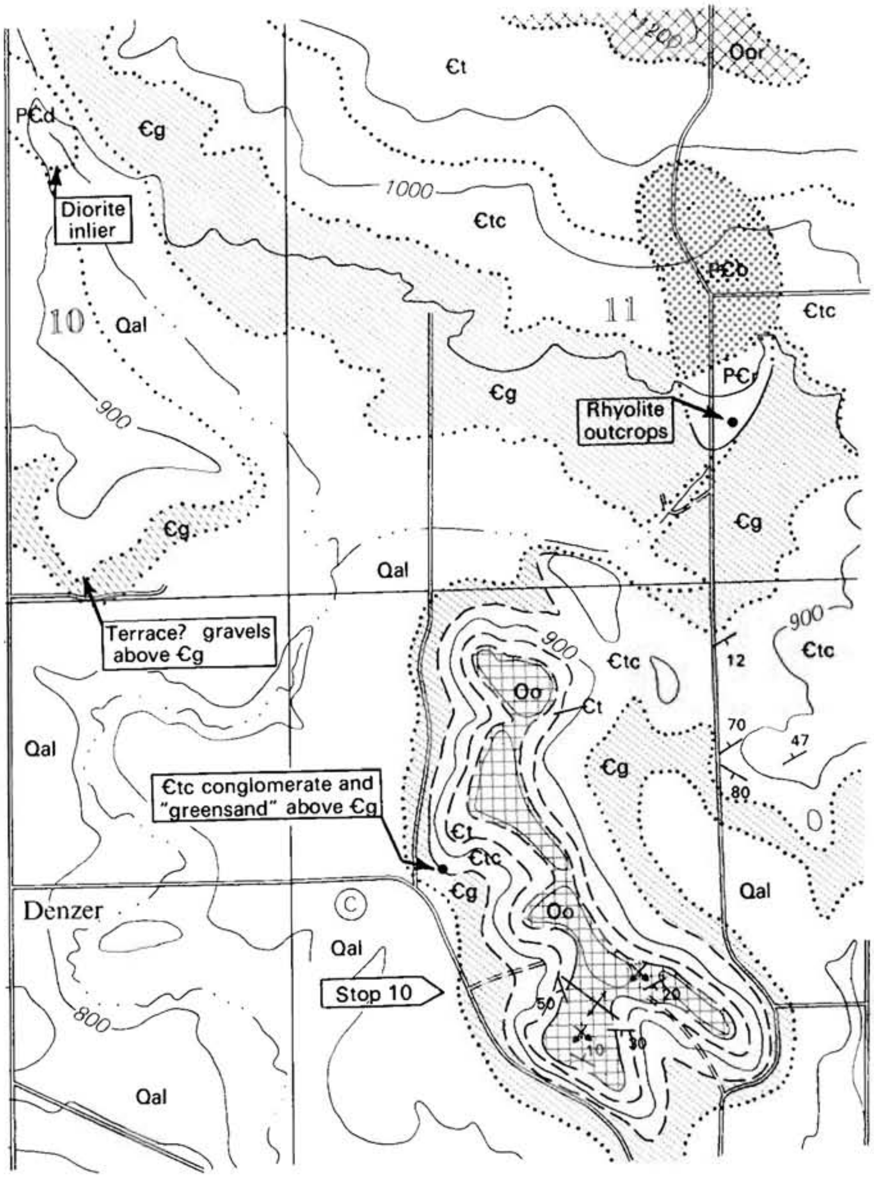


Figure 37. Geologic map of the Denzer area, containing STOP 10 (Denzer Quarry or "gravel pits"). Note the anomalously steep dips and folds at the quarry and to the northeast. Also note Precambrian diorite and rhyolite north of Denzer. The latter includes the "Denzer tuff" of Stark (1930, 1932). (See Fig. 21 or Pl. I for key to symbols).

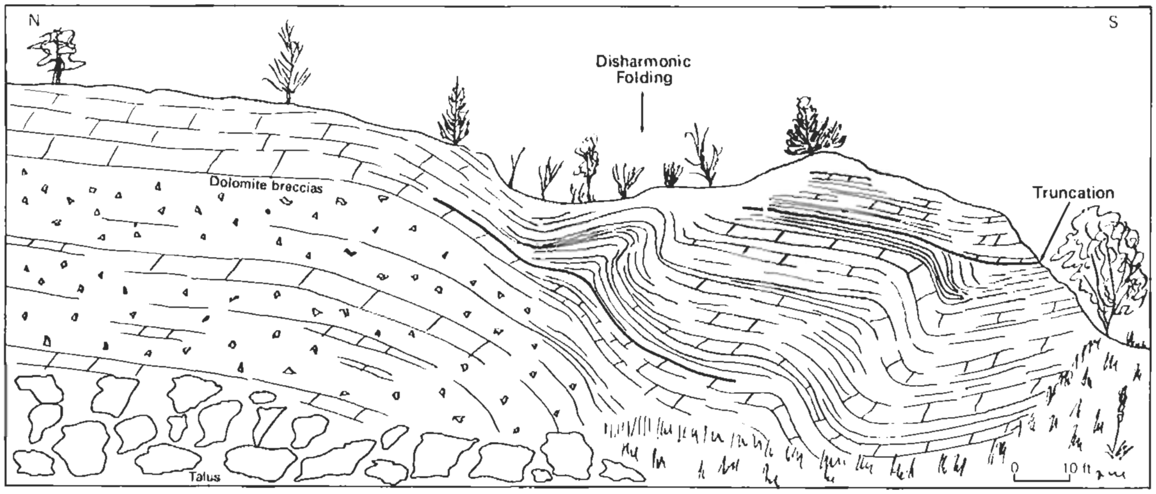


Figure 38. Sketch of disharmonic folds in fine dolomite overlying faintly-stratified dolomite breccias in the Oneota Formation at STOP 10. Ductile behavior evidenced by the folds and the intra-formational truncation attest to syn-sedimentary gravity failure--apparently a slump toward the right. (Northeast corner of Denzer Quarry).

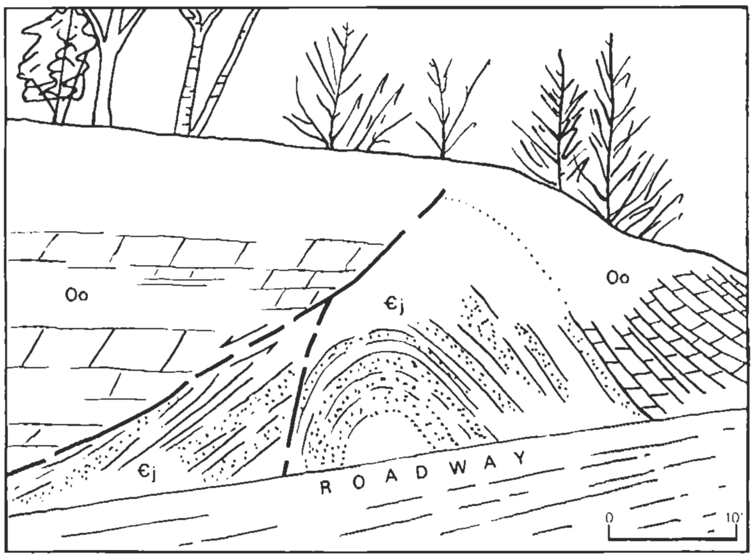


Figure 39. Faulted fold in uppermost Jordan and lowest Oneota strata in roadcut at entrance to Denzer Quarry, STOP 10. Faulting at left attests to brittle behavior in contrast to Fig. 38. (See accompanying discussion).

67.5 Tunnel City in cliff at left with excellent exposure of trough-type cross bedding. The farmer is friendly. Exposures of Black Earth Dolomite, Lodi Siltstone, Jordan Sandstone, and Oneota dolomites occur up the hill (north) in old quarries. This area provides a useful local stratigraphic "standard" for the region south of the syncline.

68.4 T Junction -- bear left (north).

69.6 U.S. Highway 12.

Stay straight ahead on County Z and City 12. We are recrossing the outwash plain; terminal moraine is evident ahead at the break in topography.

70.7 Junction Wisconsin 78 and Business 12.

Bear left on 78.

Terminal moraine crosses Highway at junction. We now are going to retrace our route of the first (late) afternoon in reverse order (between mileages 96.0 and 104.3).

76.7 Turn left (north) on 113 (if buses cannot clear 10.5 feet high underpass at left, a detour can be made east 2.5 miles to Merrimac and north 2.5 miles to Parfreys Glen Road, thence to STOP 11; see Fig. 18). Now approaching the rather poorly exposed eastern part of the South Range; we shall turn east along the bottom of the escarpment. Refer to Figure 26.

78.6 The large quarry at ten o'clock shows a strike section of the Baraboo Quartzite dipping away from us to the north.

79.1 Junction Parfreys Glen Road (mileage 96.0 of first day's log).

Turn right (east), leaving route of first day.

79.3 Kame deposits of sand and gravel on right.

80.3 Gravel road to left leads uphill a quarter of a mile to Supplementary Stop I, where spectacular Cambrian conglomerate with boulders 10 feet long can be seen resting upon in-place Baraboo Quartzite (see Fig. 26).

81.1 Turn off to Parfreys Glen State Scientific Area.

STOP 11 -- Buses proceed to parking area. We shall walk up the Glen about half a mile.

STOP 11 -- PARFREYS GLEN

Parfreys Glen Scientific Area is typical of many picturesque narrow gorges that notch the outer steep faces of the Baraboo Ranges, and contains the exotic yellow birch flora. Because the present streams are very small in proportion to their valleys and are choked with boulders that they cannot move, it is assumed that most of the gorges were cut in the past by glacial meltwater. Most of the cutting may be as young as the last 10,000 years. Remnants of large potholes in the upper gorge walls are relicts of the down-cutting history.

Gorges such as Parfreys Glen provide excellent exposures of Cambrian strata (see discussion for STOP 5). In several other examples, such as Supplementary Stop I only 1.5 miles west of here (Fig. 26), the Cambrian-Precambrian unconformity is perfectly exposed with very coarse angular or rounded basal conglomerate boulders above it. Farther from the Precambrian contact, as at Parfreys Glen, largest conglomerate fragments rarely exceed four feet in diameter, but tend to be well rounded (Pl. VI). Associated finer pebbles, however, are less rounded, a contrast characteristic of normal gravels.

The Glen is so located as to expose conglomerate facies with a nearly one-to-one conglomerate-sandstone ratio (Fig. 40). Based upon elevation alone, the strata should represent the Trempealeau Group, but the southward dip of 6° to 10° here makes it probable that Tunnel City equivalents also are present; there is no basis for separating the two here at the lower end of the gorge. Note pocked weathering of sandstone due to irregular silica cementation.

Orientation of the prominent cross bedding shows that Cambrian currents flowed toward the southeast here (See Pl. VII). Movement was nearly parallel to the general shoreline, which lay less than a mile to the north. The distribution of conglomerate in discrete, relatively thin layers separated by sandstone records episodic sedimentation processes. For purposes of argument, if we assume that the entire sequence exposed here (a maximum of nearly 100 feet) represents say 10 million years, then the average apparent rate of accumulation was only about 10 feet per million years; even a rate twice as fast for a total interval of 5 million years would be geologically slow! Regardless of the total time interval represented, "average" rates are clearly misleading. Geologists probably have tended to assume that most sequences of strata, especially in cratons, record the <u>normal</u> or continuous conditions of the past. Parfreys Glen, however, records chiefly the results of "unusual" or <u>episodic</u> conditions. Probably the "average" condition involved very minor transport of sand, but was interrupted by violent events, such as storms. The latter produced enough wave and current energy to sweep quartzite boulders up to about 1500 pounds from the foot of the nearby sea cliffs offshore for at least one-quarter of a mile. The gravel was spread out as thin layers, winnowed to produce lag gravels, and then buried by migrating submarine sand dunes (or sand waves) now reflected by cross bedding (Fig. 40).

Parfreys Glen, together with the regional pattern of conglomerate-size distribution (Pl. VI), provides considerable insight into the Cambrian paleogeography and history of sedimentation. The local source of gravels provided by the Baraboo islands makes it possible to see clearly the episodic nature of Paleozoic deposition here, which almost certainly also prevailed over much of the rest of the craton where its effects may be less obvious. Comparison with the modern oceans, especially in areas of violent tropical storms or of large tsunamis, provides further insight that leads one to wonder if most of the stratigraphic record may not record episodic "unusual" violent events rather than "average" tranquil conditions (see Gretener, 1967, and Hayes, 1967). Yet, one is forced to wonder if 20 or 30 violent hurricanes per century, such as characterizes much of our Gulf Coast, are really geologically very rare (see discussion of Cambrian sedimentology). How rare is "rare"?

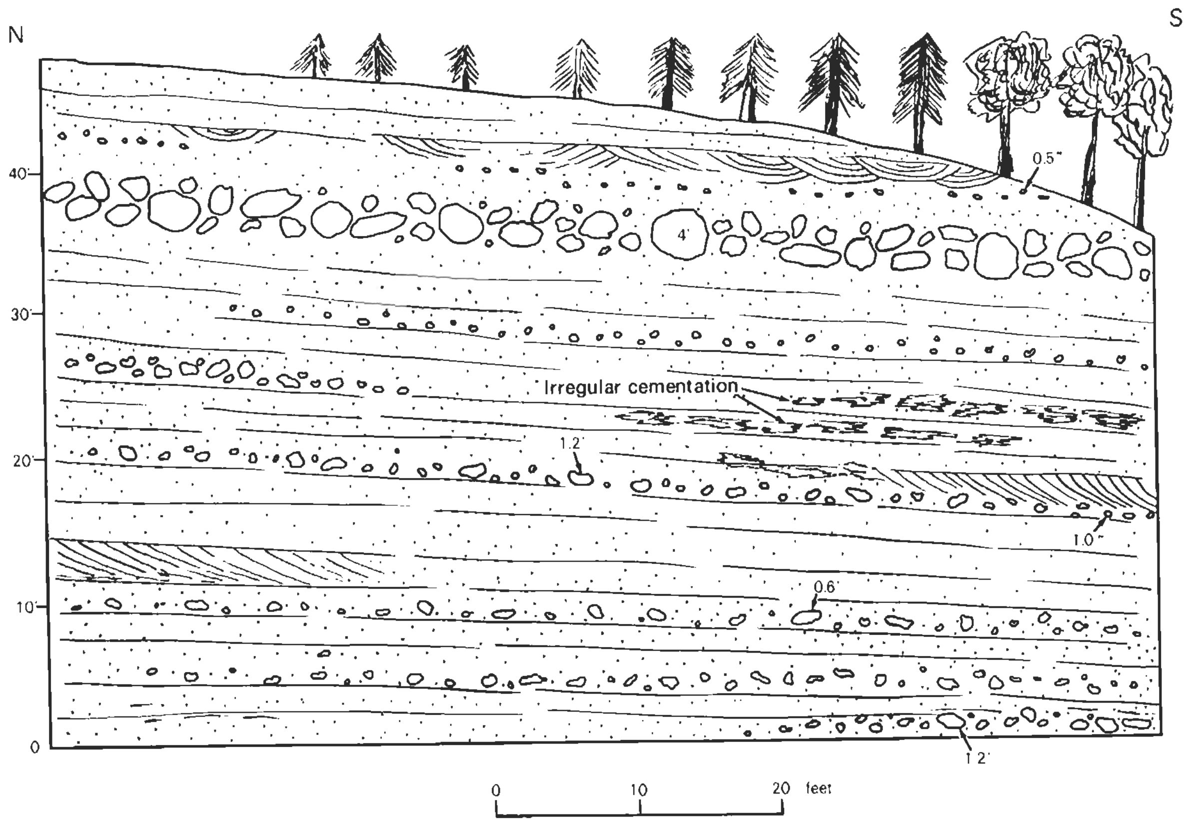


Figure 40. Sketch of east face of Parfreys Glen (STOP 11) showing thin zones of rounded quartzite cobbles and boulders (up to 4 feet diameter) interstratified with cross-bedded quartz sandstones (upper Tunnel City or Trempealeau Group equivalents). This locality is especially notable for evidence of episodic dispersal of coarse gravel for considerable distances offshore from old islands.

The suggested return route is by a steep trail from the waterfall up the east bank, and thence south along the top of the gorge back to the parking lot. The initial steep climb will be more than compensated by the view back down into the Glen. The rest of the group will be anxious to reach Milwaukee promptly, so do not step too close to the edge on the slippery pine needles. Moreover, defacing of the glen is prohibited, this being a Scientific Area.

Return to Parfreys Glen Road and proceed east (left).

81.5 Junction -- proceed straight ahead.

82.2 Climbing hill with Tunnel City and St. Lawrence Formations very poorly exposed.

82.8 Yum Yum Hill campground road to left!

83.3 Junction with Highway 78.

Proceed straight ahead on 78.

83.9 Rolling hills to right (south) of road are underlain entirely by glacial deposits, but the wooded ridge on the south skyline has good exposures of Cambrian strata.

84.4 Junction Durwards Glen Road to left. Durwards Glen (three-quarters of a mile northwest) is almost as picturesque as Parfreys Glen, and it also exposes the conglomeratic facies of Cambrian strata. The Glen was settled and named about 1860 by an early Scottish immigrant family; it is now a Catholic Novitiate, but the grounds are open to the public.

86.3 Junction with Messer Road on left. Messer Road leads to Supplementary Stop J, where excellent examples of conjugate tension gash bands occur in the Baraboo Quartzite.

88.3 Supplementary Stop K; Caledonia United Presbyterian Church (formerly Alloa)--more Scottish influence! There is an excellent exposure of rhyolite here with folded and cleaved steeply north-dipping compositional layering. Much of the layering appears to be flow banding; coarse fragmental material also is present, and in thin section, textures typical of welded tuffs can be seen. The contact with the quartzite to the north is not exposed, although the two are only about 30 yards apart.

89.2 Junction with Beich Road.

Continue, turning <u>right</u> on 78.

Beich Road leads to Supplementary Stops L and M. At Supplementary Stop K, on the south limb of the Baraboo syncline, an example of sigmoidally deformed tension gashes occurs where the S_1' cleavage in the quartzite is also seen to be deformed (see Fig. 54). Just southeast of L is a long series of excellent glaciated exposures of

Cambrian conglomerate and sandstone, and a half mile north of L along the roadside, glaciated basal conglomerate also is well exposed. At Supplementary Stop M, the strike of the quartzite can be followed around the east-closing and west-plunging hinge of the syncline. The surface of the quartzite has potholes, some of which contain indurated quartz sandstone of lower Paleozoic type.

91.2 Junction with Interstate 90/94.

Proceed onto the Interstate east bound for 27 miles. Where I-90 and I-94 split just east of Madison, follow I-94 east-bound to Milwaukee (75 miles).

For the first two-thirds of the Madison-Milwaukee route we cross typical rolling, glaciated terrain consisting chiefly of ground moraine, drumlins, and swamps. This entire region was traversed by the Late Wisconsinan (Cary) Green Bay ice lobe, which moved south-westward across eastern Wisconsin. The terminal moraine at Baraboo and west of Madison formed along its western margin.

About 25 miles west of Milwaukee, we cross the Kettle Interlobate Moraine. The Kettle Moraine State Forest to the north is proposed to be included in the Ice Age National Scientific Reserve. This moraine formed between the Green Bay ice lobe on the west and the Lake Michigan lobe on the east. Milwaukee itself lies on the slightly younger Lake Border and Valderan moraines and glacial lake deposits. Silurian and Devonian dolomites, which dip gently eastward toward the Michigan Basin, underlie those deposits.

End of Second Day Road Log. We hope that you have a pleasant and stimulating stay in Milwaukee.

APPENDIX

GUIDE TO SUPPLEMENTARY STOPS
(Dalziel and Dott)

SUPPLEMENTARY STOP A -- Lower Narrows - West Side

Location: $SW\frac{1}{4}$, Sec. 23, T.12N, R.7E; North Range immediately west of Wisconsin Highway 33.

Access: From Wisconsin Highway 33 turn north along County U at the bridge over the Baraboo River (see Road Log, First Day, Mileage 41.6). Drive 0.1 mile on U and turn left along an unnamed gravel road. Park at the first fence line on the left (0.3 mile west of County U). The farmer who owns the field, and from whom permission to enter should be obtained, lives in the house on the southwest corner of the junction of County U with the gravel road. Go south along the fence line and continue south up the shoulder of the North Range.

Features to be Seen: A short distance below the summit of the range you will reach a slight depression where sheared rhyolite and quartzite are seen in situ within 5 or 6 yards of each other. Both dip steeply north here. There is a pebbly zone at the base of the quartzite, but no pronounced basal conglomerate, and no obvious rhyolite pebbles. Continue over the crest of the spur and skirt south round the embayment in the North Range overlooking the Narrows. The bedding in the quartzite is near vertical and flat lying S_1' cleavage is prominent. Locally it can be seen to steepen into north-dipping cleavage (S_1) in thin phyllitic bands. Large quartz veins are common. At the top of the range there is a thin veneer of Cambrian conglomerate with rounded boulders up to six feet in diameter. Again the present topography closely approximates that of the Early Cambrian.

At a point on top of the next eastward spur on the range there is a long, low, outcrop where a structural situation essentially similar to that seen at Van Hise Rock (Cover photo and Fig. 23, see STOP 2) can be observed (see Fig. 19), thus helping to prove that the latter is in place. (The best way to find this locality is perhaps to climb to the top of the spur from the floor of the Lower Narrows twenty yards south of the fence line at the point where Wisconsin Highway 33 converges with the western wall of the Narrows). Please do not hammer this!

SUPPLEMENTARY STOP B -- Upper Narrows - East Side
(see cautionary note below)

Location: NW¼, SW¼, Sec. 28, T.12N, R.5E; on east side of Baraboo River opposite Van Hise Rock (see Fig. 21).

Access: Drive north from Highway 136 at Rock Springs along the gravel road between the railroad tracks just east of the Baraboo River bridge (same as for access to STOP 3, Road Log, First Day, Mileage 60.0). Continue on foot beyond Chicago and Northwestern Railroad quarry along east side of railroad tracks and down the hill beside the Baraboo River at the railroad bridge 135 yards south of Van Hise Rock. 150 yards beyond the bridge, there is (at time of writing, anyway!) a north-facing light-colored cliff with ripple marked quartzite surfaces above it (geographically and stratigraphically). Proceed to the base of this cliff 25 yards east of the dirt road.

CAUTION:
At present time the top of the cliff is being quarried from behind and, even apart from blasting, is highly dangerous. The only reason for mentioning this locality is that it is the only one on the North Range where secondary-phase structures occur as far as we are aware. Certainly the quarry manager should be approached concerning possible imminent blasting.

Features to be Seen: At the base of the light colored face a thin phyllitic layer with well developed cleavage (S_1) is discernible. This cleavage is deformed by a strain-slip cleavage assigned to the secondary phase. This is the only locality on the north limb where such a structure has been observed (apart from the quarry itself on the other side of the rock wall, Stop 3). Its geometry is shown diagrammatically on Figure 3. In thin section it is seen to be faintly conjugate (Fig. 5D).

About 25 yards north of this locality and halfway up the cliff (climb the pile of talus from old quarrying operations), the S_1' cleavage in quartzite can be seen refracting into cleavage (S_1) in a thin phyllitic band (Fig. 41). There is a second cleavage (or closely spaced jointing) in the quartzite whose relative age is unknown although it is surmised to be a later structure and could be related to the secondary cleavage in the phyllitic layers.

Note also that brecciated quartzite similar to that seen at STOP 2 is located near the railroad bridge.

<u>Remarks</u>: It is important to emphasize the nature of the structural time scale set up here (Table 3). Crenulation or strain-slip cleavages can be recognized on the north and south limbs of the Baraboo Syncline (and in the south associated mesoscopic folds). In both cases the (S_1) cleavage in the more phyllitic layers is deformed and the (S_2) crenulation or strain-slip cleavages can therefore be assigned to a "secondary" phase of deformation. This need not mean that the "secondary" structures are necessarily synchronous on both limbs nor that they are unrelated to the development of the syncline (see general discussion of structural geology).

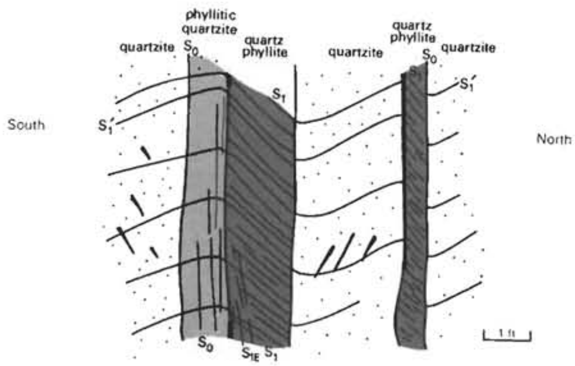


Figure 41. Sketch of bedding/cleavage relations on the east side of the Upper Narrows (Supplementary Stop B).

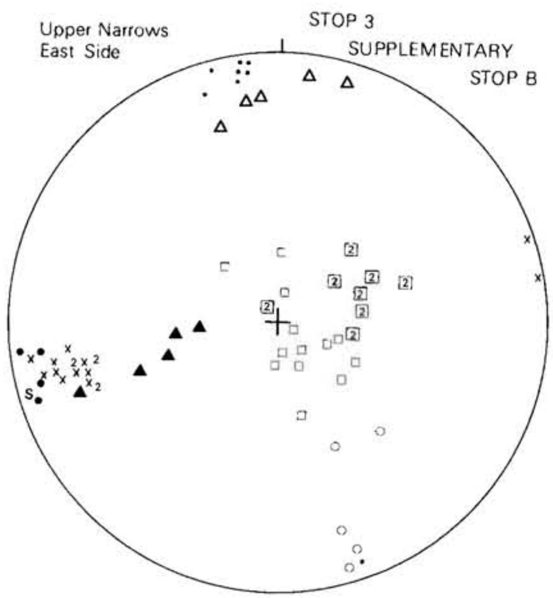


Figure 42. Stereoplot of structural data from the Baraboo Quartzite on the east side of the Upper Narrows (Supplementary Stop B). (See Table 13 for key).

SUPPLEMENTARY STOP C -- Flagstone Quarry at West Edge of Rock Springs

Location: NE¼, NW¼, Sec. 32, T.12N, R.5E.: 0.6 mile west of junction Wisconsin Highways 154 and 136 in the center of Rock Springs. (See Figure 21).

Access: Via Wisconsin Highway 154 (Road Log, First Day, Mileage 61.3); ask for permission at house next to quarry, and climb steps from driveway into quarry (spare the fence along highway); see accompanying detailed map (Fig. 43).

Features to be Seen: This quarry, like several others in Rock Springs, is in unusually well-indurated Galesville Sandstone that has large festoon cross strata clearly displayed both in section and plan (Figs. 43, 44). They suggest southerly-flowing currents (Table 12). The larger troughs are from 20 to 60 feet wide and a few are exposed for as much as 40 feet in length. Typical amplitudes of trough cross sets are 2 or 3 feet (maximum 8 feet). Many individual cross laminae thin toward the trough axes, suggesting that filling was largely from the sides rather than the ends. There is no consistent relationship between sand-grain size and thickness of laminae. Rare ripple marks near the west edge of the quarry have an average amplitude of 1 centimeter and wave length of 3 centimenters. The sandstone is medium-grained, very well sorted, and very well rounded (Table 12). Thin zones of fine quartzite and vein-quartz pebbles are present (mostly 3 to 10 mm, but one 3 cm in diameter). One prominent granule zone occurs in the lower strata (Fig. 44), and other concentrations occur in the upper edges of trough structures; they represent lag concentrations only one granule thick.

Remarks: The festoon cross stratification represents truncated fillings of troughs formed between cusp-like features, most likely migrating elongate dunes or sand waves. Their geometric relationships may be exceedingly complex (Fig. 43). Large festoons with low-dipping laminae of the sort seen here have been taken both as evidence of wind dunes and of beach cusps. But in light of recent oceanographic studies of very shallow marine sand banks and rivers, it is clear that subaqueous sand waves of identical type also exist (see general discussion of Cambrian sedimentology). Graphical plots of size-distribution parameters by M. Roshardt in 1964 yielded conflicting results for the sands at this quarry. The empirical plots of Friedman (1961) tended to show "eolian dune" characteristics, whereas those of Passega (1957) suggested "beach" deposition, thus diagnostic environmental criteria seem to be lacking at this locality. Nearby in the Upper Narrows (STOP 2), there is independent evidence suggesting eolian deposition of the Galesville sandstones against quartzite cliffs, but the pebbles, the ripple mark index, and the low angles of cross set inclination in the sandstones here at the quarry all suggest aqueous processes. Perhaps this was a sand shoal in a strait less than a mile wide between two quartzite islands (see Figs. 13 and 21). In such a constriction,

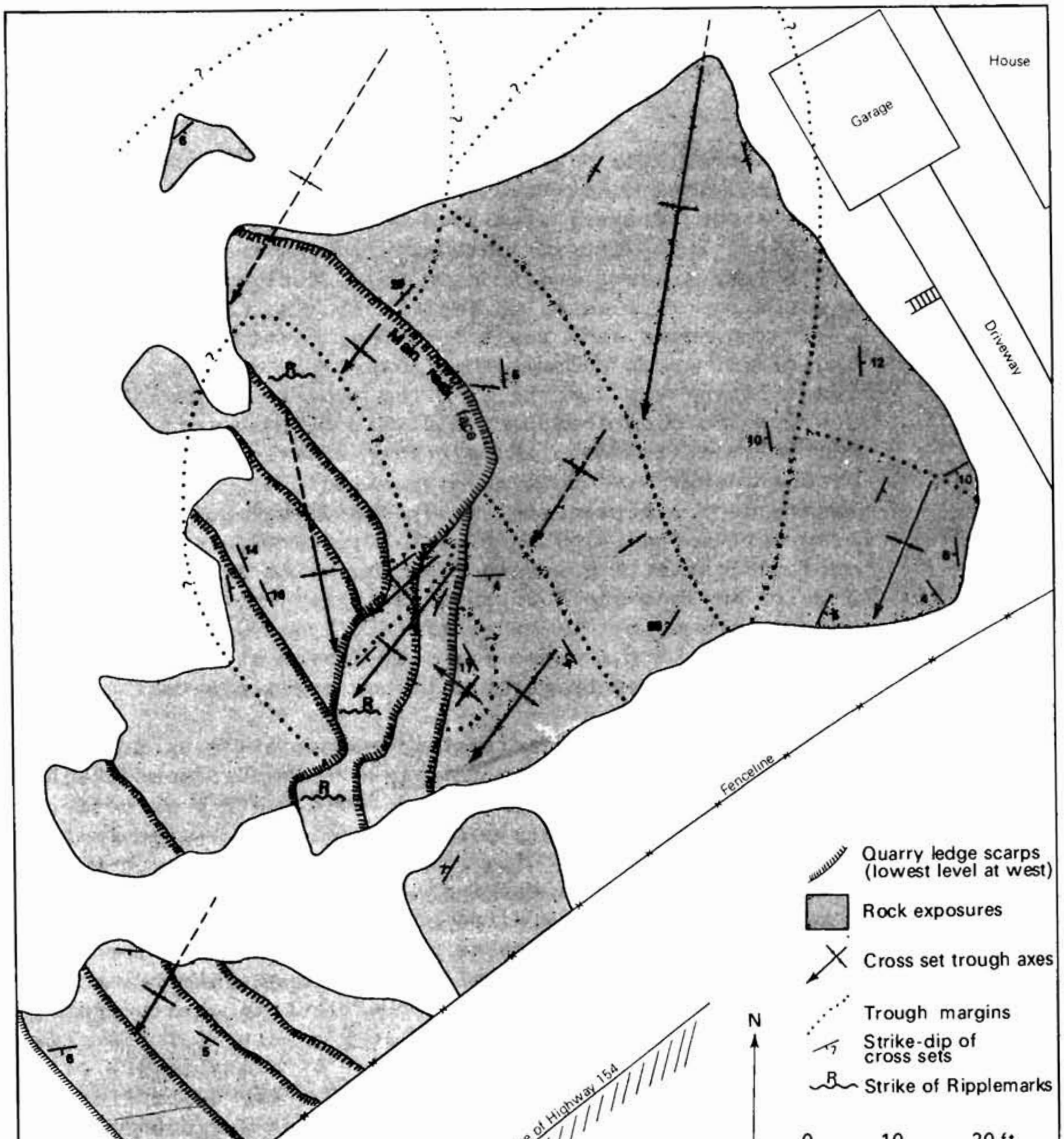


Figure 43. Pace and compass map of flagstone quarry at the west edge of Rock Springs (Supplementary Stop C) showing relationships in plan of well exposed large cross-bed troughs in Galesville Sandstone. (After M. Roshardt, 1964, unpublished student project report).

TABLE 12. SEDIMENTARY DATA FOR SUPPLEMENTARY STOP C AT ROCK SPRINGS*

Orientation Data for Cross Strata

Mean Dip Direction of 32 Cross Laminae:	218°	(Standard deviation: ±32°)
Mean Dip Angle of Cross Sets:	10.5°	(Standard deviation: ±6.3°)
Mean Plunge Direction of 9 Trough Axes:	198°	(Range: 145° - 245°)
Mean Strike of 4 Ripple Mark Sets:	90°	

Size Distribution Data

	Coarsest 1%	Mean	Median	Sorting
Means:	0.55ϕ (0.68mm)	1.82ϕ (0.28mm)	1.79ϕ (0.31mm)	0.48
Range:	-0.20-+0.95ϕ (1.12-0.52mm)	1.55-2.53ϕ (0.39-0.17mm)	1.48-2.50ϕ (0.36-0.18mm)	0.32-0.69

	Skewness	Kurtosis	Rounding
Means:	+0.11	1.22	0.9
Range:	-0.05-+0.37	1.00-1.79	---

(*After M. Roshardt, 1964, unpublished student project)

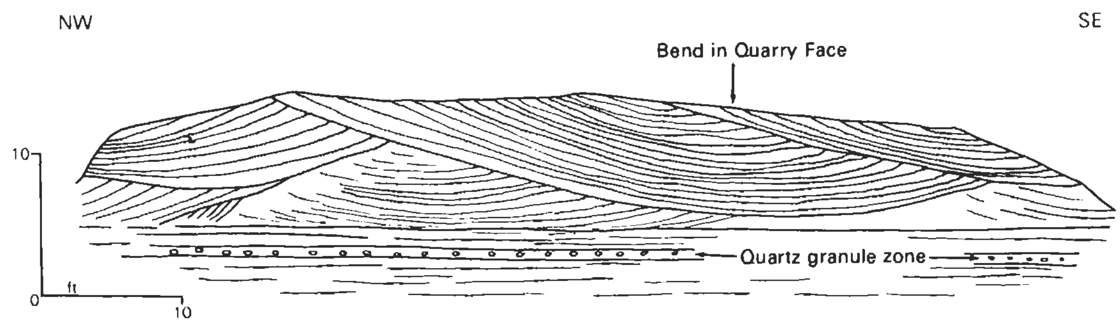


Figure 44. Sketch of main face of flagstone quarry at Supplementary Stop C, showing large trough (or festoon) cross bedding in section view. (See Fig. 43).

current velocities and turbulence might have been considerable, thus explaining the presence of unusually large sand waves. (See general discussion of Cambrian sedimentology for further details).

SUPPLEMENTARY STOP D -- Narrows Creek

Location: NW¼, Sec. 31, T.12N, R.5E. Alongside Wisconsin Highway 154 in Narrows Creek gorge, another gap cut through the North Range.

Access: Drive west on Wisconsin Highway 154 for 1.2 miles from junction with County D (see Road Log, First Day, Mileage 61.5); park at Wayside on right.

Features to be Seen: The strike of the quartzite has a more northeasterly trend here than in the Upper Narrows as it swings around into the West Range (Pl. III, Figs. 21, 45). Quartzite cleavage (S_1') is well developed as are other sets of closely spaced structural surfaces (Fig. 45).

Remarks: A number of former University of Wisconsin students have proposed in theses that a northwest-southeast-trending fault passes through Narrows Creek gorge. Deflection of the highly inclined bedding tends to support this idea. Horizontal shift, if any, would be left-lateral.

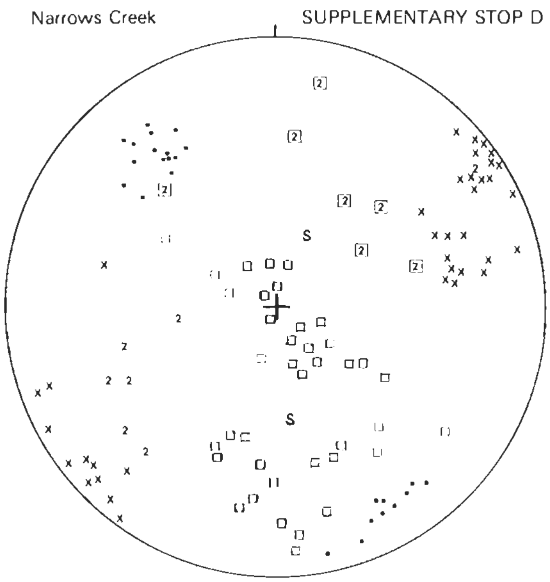


Figure 45. Stereoplot of structural data for Baraboo Quartzite at Narrows Creek, Supplementary Stop D. (For key see Table 13).

SUPPLEMENTARY STOP E -- Diamond Hill School Ridge

Location: NW$\frac{1}{4}$, SW$\frac{1}{4}$, Sec. 3, T.11N, R.5E. Helmer Erickson farm on ridge running northeast from Diamond Hill School at intersection of two unnamed gravel roads 1.1 miles north of County Highway W.

Access: Drive north from County D along gravel road (Road Log, First Day, Mileage 76.5) 1.1 miles to intersection at Diamond Hill School and a farther 0.2 mile to Helmer Erickson farm road on right. Drive up to farm and ask permission to look at the rocks on the hillock northeast of the barn.

Features to be Seen: There is a small syncline in the quartzite at this locality, on the crest of the dominantly anticlinal structure of the southeastern Diamond Hill area (see Pl. II, section B - B'). On top of the ridge there are excellent examples of deformed current bedding similar to those visible at the U.S. Highway 12 roadcut on the South Range (Supplementary Stop G). The deformation is probably synsedimentary, being confined to discrete sets. On the southeast slope of the ridge northwest-dipping phyllitic cleavage with a down-dip longrain type of lineation is deformed by the best developed secondary crenulation and strain-slip cleavage in the area (Fig. 46). This is essentially flat lying.

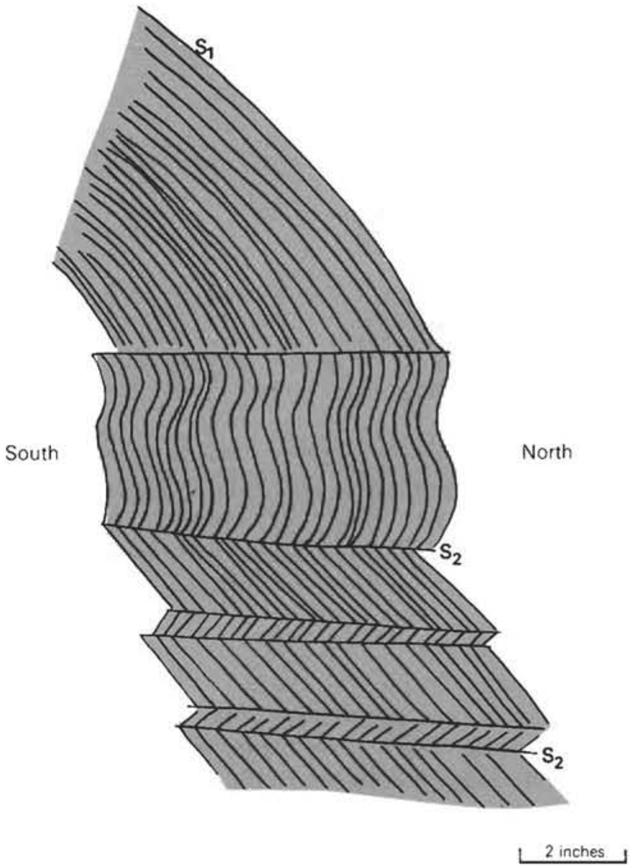


Figure 46. Secondary folds and associated cleavage in the Diamond Hill area (Supplementary Stop E).

Remarks: The Diamond Hill area comprises the most complex zone of folds exposed in the Baraboo district. Unfortunately, the exposure is rather poor. Older literature tends to overemphasize the "competent" behavior of the Baraboo Quartzite. The quartzite beds are certainly competent with respect to the more phyllitic layers, but the closure of a number of relatively tight folds can be seen in the Diamond Hill area, and thin-section studies such as that of Riley (1947) have revealed considerable deformation of quartz crystals in the massive quartzite beds by intracrystalline flow (see general discussion of structural geology).

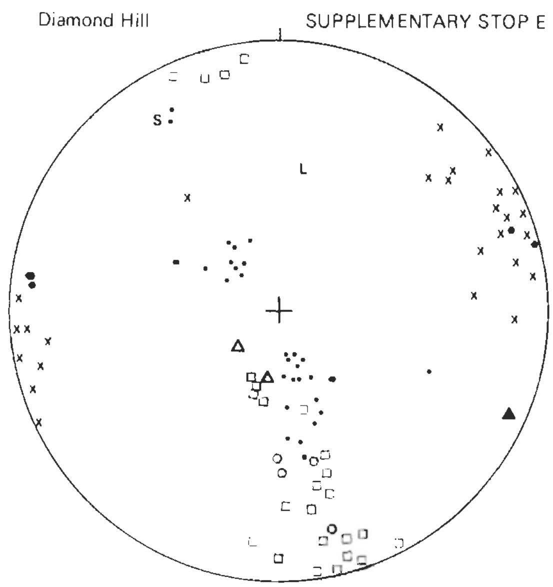


Figure 47. Stereoplot of structural data for the Baraboo Quartzite from Diamond Hill, Supplementary Stop E. (For key, see Table 13).

SUPPLEMENTARY STOP F -- Northwest Entrance to Devils Lake State Park

Location: SW$\frac{1}{4}$, NW$\frac{1}{4}$, Sec.13,T11N, R.6E. On west side of main road through Devils Lake State Park (Wisconsin Highway 123), 200 yards from northwest entrance (Fig. 26).

Access: Drive into park on Wisconsin 123 and stop on first right bend 200 yards inside (see Road Log, First Day, Mileage 77.2). There is space to park safely on the left (east) side of the one-way road. Walk up path on right (west) side of road.

Features to be Seen: A structure unique to this locality in the phyllitic zone of the upper Baraboo Quartzite is a large asymmetric fold in a quartzite layer approximately one foot thick enclosed by phyllite (Fig. 48). The fold (approximately 75 yards up the trail) has been called a "normal drag fold" (see discussion of STOP 6) because its southerly vergence (sense of overturning) is congruent with the geometry of the Baraboo Syncline. Phyllitic cleavage (S_1), which dips north at a higher angle than the bedding, is axial planar to this main-phase fold and is deformed by dominantly southward dipping secondary crenulation cleavage axial planar to a few asymmetric "reverse drag folds" (i.e., secondary minor folds whose vergence is not congruent with the syncline). Elsewhere in the outcrop, north-dipping crenulation cleavage also is present, thus forming a conjugate secondary structure. The hinge lines of the main-phase ("normal drag") folds, all bedding/cleavage intersections, and the axes of secondary crenulations, are all nearly horizontal and trend east-west (Fig. 28). In the core of the anticlinal portion of the so-called "drag fold" is another one-foot-thick quartzite layer isoclinally folded. Lineations that curve across this layer in the hinge zone appear to be secondary structures crenulating the main phase phyllitic cleavage (S_1) developed at the quartzite/phyllite interface (Fig. 48). Boudinage (on axes again nearly horizontal and east-west trending) and quartz veins (nearly at right angles to bedding) are well developed in the quartzite layer. Some boudins are deformed by the secondary phase asymmetric folds. **DO NOT HAMMER THIS ROCK!**

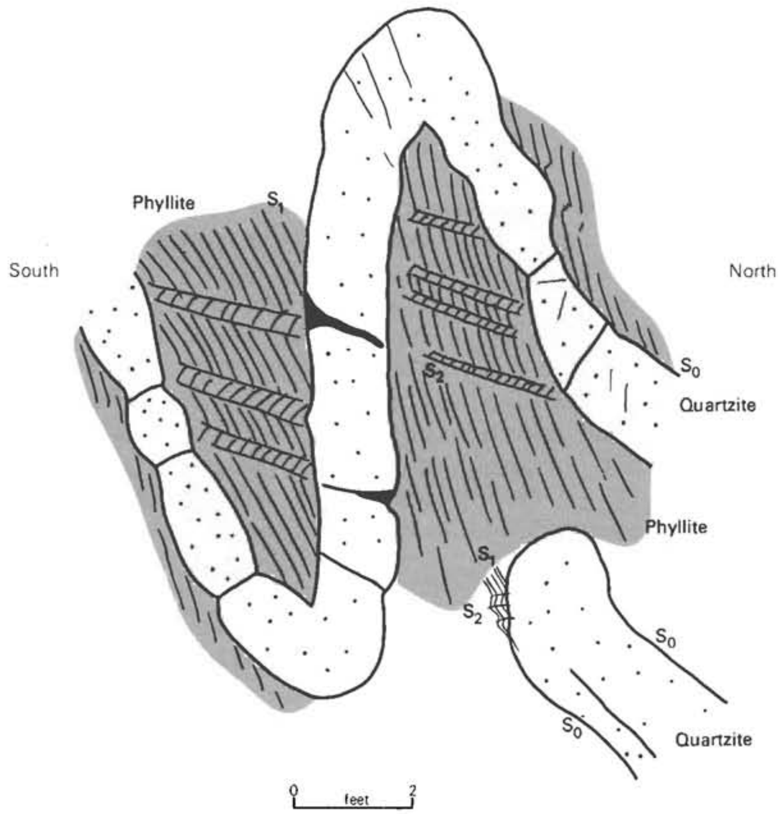


Figure 48. Structural relations in the phyllitic zone of the Baraboo Quartzite at the northwest entrance to Devils Lake State Park (Supplementary Stop F).

SUPPLEMENTARY STOP G -- U.S. Highway 12 Roadcut

Location: SW¼, NW¼, Sec. 15, T.11N, R.6E. Low cliff on east side of U.S. Highway 12 0.6 mile south of junction with Wisconsin Highway 159 (see Road Log, Second Day, Mileage 30.8).

Access: There is space to park safely on either side of the Highway, but the traffic is fast and heavy, and there is a blind corner to the north. Take care!

Features to be Seen: As mentioned in the road log, this locality exemplifies the dominantly north-to-south current direction revealed by cross-stratification of the Baraboo Quartzite (see Fig. 2). Oversteepening and contortion of the crossbedding is common and probably synsedimentary in origin, contortion being confined to discrete sets of cross laminae.

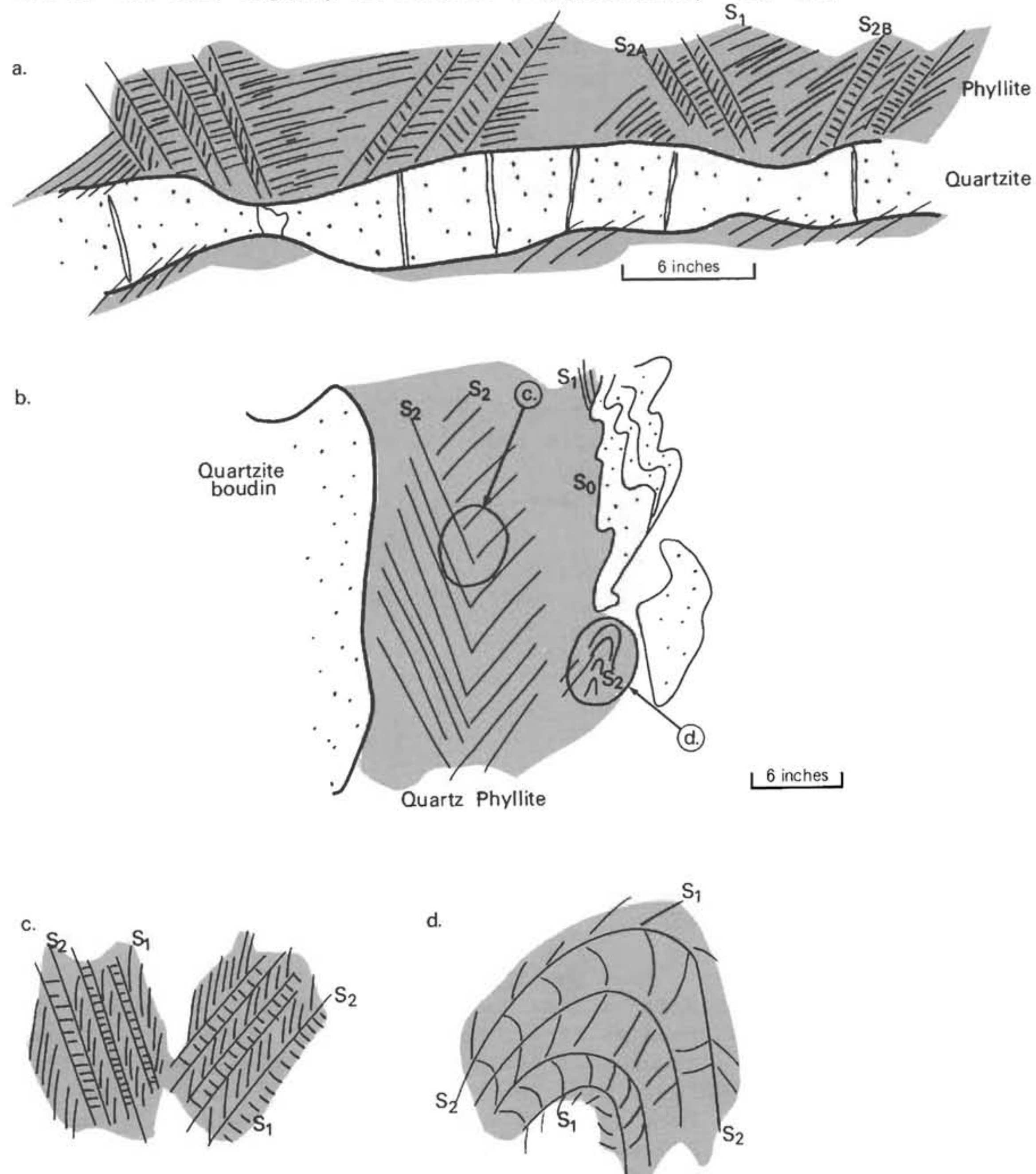
Very thin phyllitic partings occur in the quartzite and S_1 cleavage is near parallel to the gently north-dipping bedding and at right angles to the steeply south-dipping S_1' cleavage in the quartzite (as in Fig. 6). This is an excellent locality at which to observe the essentially parallel down-dip lineations on the surfaces of the phyllitic cleavage (longrain and fine crenulation) and bedding (slickensides). The chatter marks on the upper bedding surfaces face up-dip and seem to be the result of bedding/cleavage intersection in this case as elsewhere (e.g., La Rue Quarry, STOP 7). DO NOT HAMMER THIS ROCK!

Above the prominent quartzite cliff is the phyllitic zone of the upper Baraboo Quartzite. This can also be seen on top of the hill west of the highway 100 yards northeast of the farm. Secondary chevron folds with a south-dipping axial planar crenulation cleavage (S_2) deform the phyllitic cleavage (S_1). Locally - as at road level to the north of the quartzite outcrop - one can see the conjugate north-dipping crenulation cleavage. The north and south-dipping S_2 surfaces have an interesting relationship to boudins in thin quartzitic layers within the phyllite (Fig. 49a). Note that some of these boudins are deformed by the chevron folds and therefore related to an earlier phase of the deformation.

At the extreme upper (and southeastern) corner of the outcrop there is a fascinating and perplexing structure in the phyllite (Fig. 49b, c, d).

Bedding can be recognized in the adjacent quartzite layers, and phyllitic cleavage (S_1) deforming the margin of the quartzite beds. This cleavage is deformed by secondary crenulation cleavage in the phyllite and in places the secondary cleavage is apparently itself deformed by rather open minor folds. Due to the nature of the S_2 cleavage it looks as if those open folds have an axial planar foliation, but in fact this is the S_1 cleavage, which occupies an axial planar orientation.

Figure 49. Structural relations in the phyllitic zone of the Baraboo Quartzite at the U.S. Highway 12 roadcut (Supplementary Stop G).



a. Secondary crenulation cleavages (S_2) in phyllite showing commonly observed relationship to boudins in adjacent quartzite layer (at north end of outcrop, road level).

b. Highly complex structure in phyllite adjacent to quartzite layers (south end of outcrop above road level).

c. In part the structure shown in b consists of conjugate S_2 crenulation cleavages. (See b for location of this diagrammed area).

d. Elsewhere the crenulation cleavages appear to be deformed by what are here called late-phase folds (see Table 3). (Area of d is identified in b).

However, some question remains as to whether these "late phase folds" really exist. Could this complex structure be the result of interfering sets of conjugate S_2 cleavages deforming the S_1 cleavage in a zone of heterogeneous strain between a quartzite layer and a quartzite boudin?

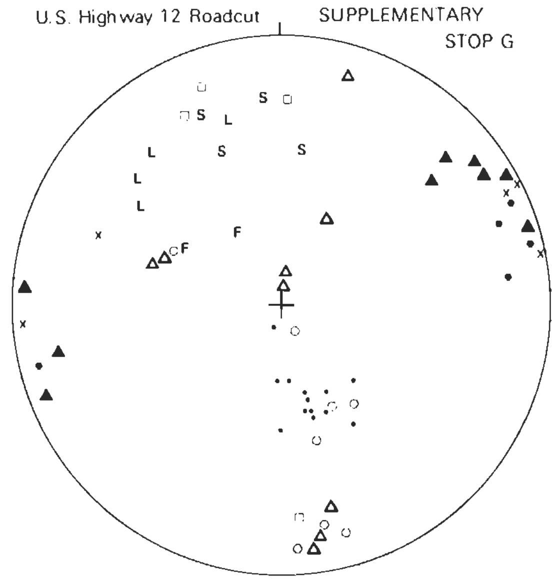


Figure 50. Stereoplot of structural data for the Baraboo Quartzite from the U.S. Highway 12 locality, Supplementary Stop G. (For key see Table 13).

SUPPLEMENTARY STOP H -- Happy Hill

Location: NW¼, SE¼, Sec. 25, T.11N, R.5E. 0.4 mile southwest of Happy Hill School.

Access: Drive south from junction of County Highways PF and W south of North Freedom (see Road Log, Second Day, Mileage 39.0). Road is paved at first, becomes gravel farther on. Take first gravel road on left (at 3.3 miles) towards Happy Hill School and drive east to point where power line crosses road (0.5 miles). Park and walk south along overgrown trail and fence line for 300 yards. The outcrops lie along a low cliff on the north slope of a heavily wooded valley.

Features to be Seen: This stop also is located in the phyllitic layer of the upper Baraboo Quartzite. The phyllitic cleavage is deformed by numerous asymmetric secondary phase folds (Figs. 51 and 52) with south dipping axial planes and near-horizontal axes (the "reverse drag folds" of Hendrix and Schaiowitz (1964) who described them extensively). Please do not use rock hammers!

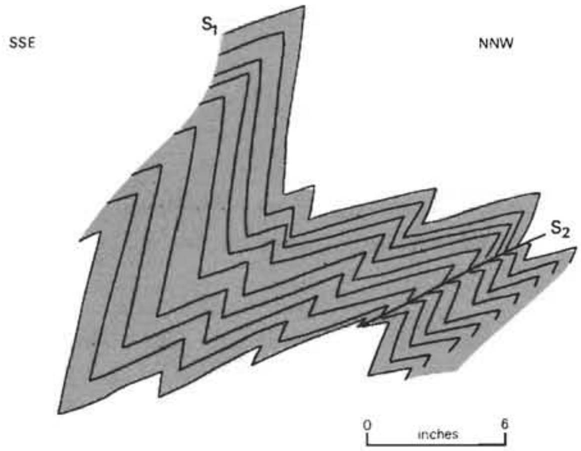


Figure 51. Secondary phase structures in the phyllitic zone of the Baraboo Quartzite near Happy Hill school (Supplementary Stop H).

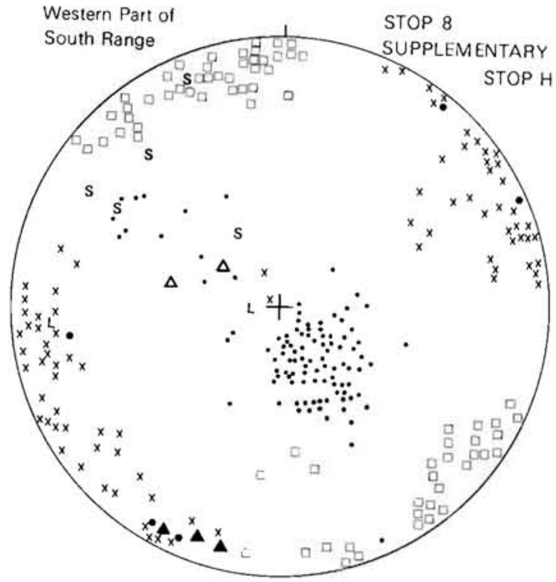


Figure 52. Stereoplot of structural data from the Baraboo Quartzite, in the western part of the South Range near STOP 8 and Supplementary Stop H. (For key see Table 13).

SUPPLEMENTARY STOP I -- Halweg Farm Glen

Location: SE$\frac{1}{4}$, SW$\frac{1}{4}$, Sec. 22, T.11N, R.7E. 4 miles east of the SE corner of Devils Lake and 1 mile SW of Parfreys Glen.

Access: Via steep gravel side road to left (Road Log, Second Day, Mileage 80.3) that intersects Parfreys Glen Road 1.2 miles east of Wisconsin Highway 113 (see Fig. 26). Park at second (upper) hairpin curve, and walk 200 feet southwest into the small, wooded valley over a low cliff. With a large group, one should inquire at the farm to the north.

Features to be Seen: Just below the road is a 20-feet-high cliff of sandstone with few pebbles, but conglomerate increases abruptly downward. In the valley bottom and on the northwest band, in-situ quartzite crops out. Up the east fork of the valley about 100 yards and just below the road, the Cambrian-Precambrian unconformity is exposed with angular boulders up to 12 feet long resting upon it. One rounded slab 9 feet long and many rounded boulders 4 to 6 feet in diameter are present.

Remarks: This locality is similar to Elephant Rock on the East Bluff of Devils Lake (STOP 5). It provides a fine, accessible complement to nearby Parfreys Glen where the unconformity and quartzite are not exposed. It is useful to let students discover both the quartzite and the unconformity here, and to interpret the relationship without coaching. Finally, the contrast of shape between the larger and smaller boulders is noteworthy with respect to interpreting ancient surf conditions (see general discussion of Cambrian sedimentology).

SUPPLEMENTARY STOP J -- Zamzow Farm Hill

Location: SE¼, SE¼, Sec.32, T.12N, R.8E. Spur high on eastern side of valley northeast of Zamzow farm valley drained by a left-bank tributary of Rowley Creek.

Access: Drive north from Wisconsin 78 along Messer Road (see Road Log, Second Day, Mileage 86.3). After three right angle turns (1.15 miles total) there is a farm set back from the road to the north. Walk to the northeast corner of the field behind the farm (400 yards) and then north along the east wall of the valley. There are three prominent quartzite spurs. Go on to the northernmost one immediately beyond an east-west barbed wire fence.

Features to be Seen: Sets of quartz-filled tension gashes occur on a glaciated surface of quartzite beds dipping gently north. The prominent sigmoid tension gashes in the quartzite form conjugate sets, S-shaped trending northeast-southwest and Z-shaped trending northwest-southeast. Less prominent sets occur in other orientations.

Remarks: Assuming synchroneity, the prominent gashes can be interpreted as having resulted from a least compressive principal stress parallel to the hinge line of the syncline and the greatest and intermediate compressive principal stresses in a north-south vertical plane (the smooth outcrop makes it difficult to measure attitude). This is the picture which emerges whenever any geometric pattern is discernible in the tension gashes of the quartzite (see STOPS 2 and 4, and general discussion of structural geology).

SUPPLEMENTARY STOP K -- Caledonia Church (formerly Alloa)

Location: NW¼, NE¼, Sec. 3, T.11N,R.8E. Just off Wisconsin Highway 78, 250 yards northeast of Caledonia United Church.

Access: Leave Wisconsin Highway 78 (Road Log, Second Day, Mileage 88.3) and park at the church. The best way to the locality is through the yard of the house to the east of the church and across the small creek in the valley behind the house. Ask permission first!

Features to be Seen: Compositional layering in dark-colored rhyolite dips north at a moderate angle concordant with that of the quartzite to the north. The rock has the appearance of a brecciated extrusive (see Stark, 1932). Microscopic features typical of welded tuffs (e.g., axiolitic structures) have been observed here. The attitude of the compositional layering is variable due to the presence of a few small folds to which there is an east-west trending axial surface foliation. Quartzite overlain by very coarse Cambrian conglomerate and conglomeratic sandstone can be seen in the woods to the north. One quartzite block just above the unconformity is nearly 20 feet long. Do not hammer excessively here.

Remarks: The axial plane foliation is roughly parallel to the S_1' cleavage in the quartzite to the north on the south limb of the syncline (Fig. 53), and there is every reason to believe that the rhyolite has been deformed with the Baraboo Quartzite.

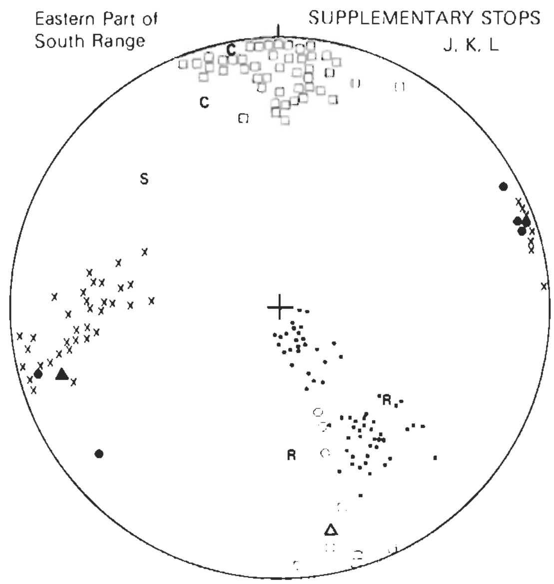


Figure 53. Stereoplot of structural data for the Baraboo Quartzite and underlying rhyolite from the eastern part of the South Range (Supplementary Stops J, K, and L). (For key see Table 13).

SUPPLEMENTARY STOP L-1 -- Beich Road

Location: SE$\frac{1}{4}$, NE$\frac{1}{4}$, Sec. 34, T.12N, R.8E. Field on east side of Beich Road at crest of hill as it climbs the South Range.

Access: Drive north along Beich Road from Wisconsin Highway 78 (Road Log, Second Day, Mileage 89.2). Continue round the hairpin bend and stop just beyond the following left bend at the crest of the hill (1.0 mile). Parking is not easy here, but traffic is fairly light. Enter the field to the east at the bend in the road and stop at the outcrops immediately across the fence.

Features to be Seen: In the quartzite, which dips north at 40-50°, there are tension gashes. Some are undeformed, others are sigmoid (Fig. 54). Quartz crystals in the gashes tend to be oriented normal to the boundaries of the gashes. The trace of the steeply-dipping and east-west-trending quartzite cleavage stops at the boundaries of all gashes. Where the gashes are deformed into a sigmoid shape, the cleavage is distorted in the same sense. Please do not hammer this rock!

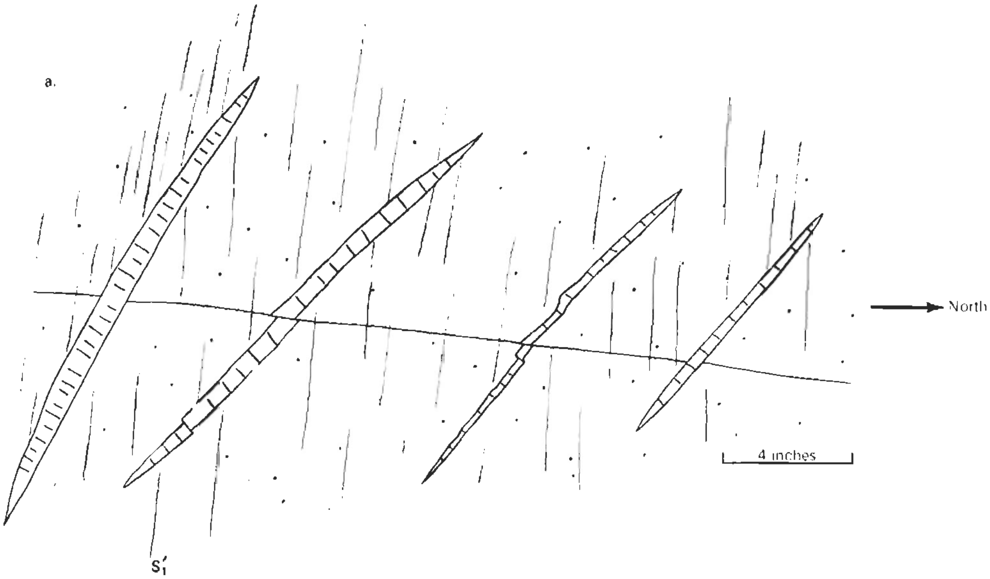
Remarks: Presumably the sigmoid tension gashes result from continued rotational strain along the discrete shear zones marked by the gash bands. Thus we are able to deduce that formation of the S_1' quartzite cleavage at least predated continued movement along the gash bands, and likely initial formation of the bands, also, as it does not cross the gashes. Hence the horizontal east-west orientation of least compressive principal stress deduced from the attitude of, and sense of displacement along, conjugate tension gash bands at a number of localities may reflect the stress system at a late stage in the tectonic history of the syncline (see general discussion of structural geology).

SUPPLEMENTARY STOP L-2

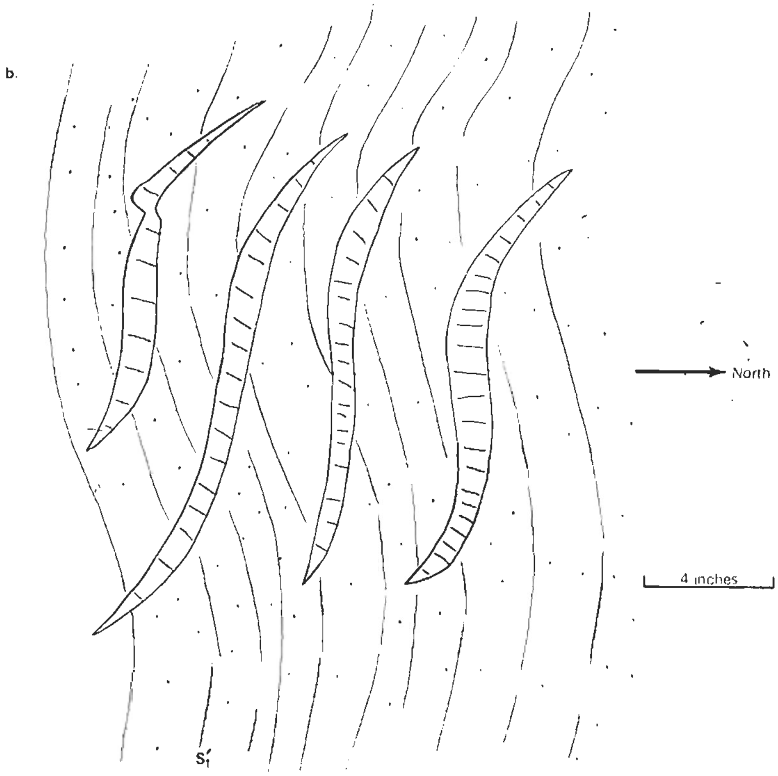
Cambrian conglomerate and sandstone in open field about 300 yards southeast of the quartzite outcrop above; at crest of steep, wooded slope (elevation about 1100 feet). Probably equivalent to Jordan sandstone.

Features to be Seen: Festoon cross stratification of medium-scale is very well exposed in plan by glaciation (widths of troughs 2-3 feet; amplitudes about 6-10 inches). Color variations emphasize the different laminae. The plunge orientation of the troughs is dominantly toward the east here, which is parallel to the old quartzite islands. Conglomerate also is conspicuous, especially just below the slope break at the upper edge of the woods near the eastern end of the crops. A few angular boulders are up to 12 and 15 feet long, but most are 1-4 feet in diameter, and rounded.

Figure 54. Tension gashes and quartzite cleavage (S_1') near the east end of the South Range (Supplementary Stop L1) seen in plan view.



a. Planar tension gashes and undeformed quartzite cleavage. Note that the cleavage does not pass through the gashes, which are formed of quartz crystals aligned perpendicular to the gash margins.



b. Sigmoid tension gashes and curviplanar quartzite cleavage. The curvature of both structures is interpreted to be the result of continued shear along the tension gash band after initial gash formation.

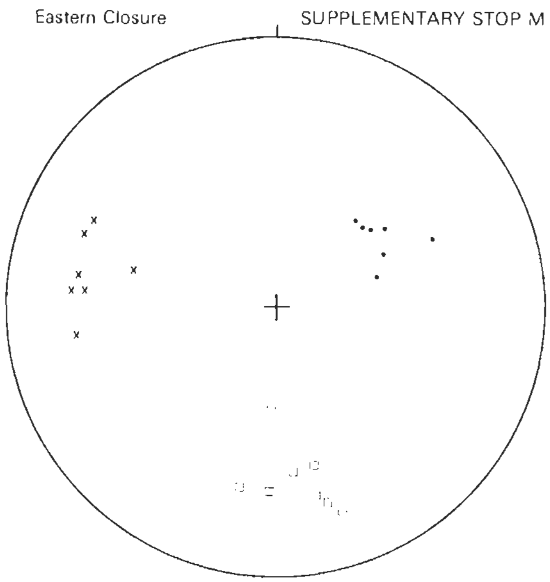


Figure 55. Stereoplot of structural data for the Baraboo Quartzite from the Eastern Closure of the syncline (Supplementary Stop M). (For key see Table 13).

Remarks: This is an excellent place for students to study festoon cross stratification three-dimensionally. Also it is a fine locality for them to make a paleogeographic interpretation before they have seen the nearby quartzite *in situ*. As at several other localities, some of the largest quartzite boulders can stimulate a debate over whether they are, in fact, boulders or the tops of buried hills. The stratigraphic facing direction criteria within the boulders (e.g., truncated tops of cross sets) in comparison with the main synclinal structure provide a clue.

SUPPLEMENTARY STOP M -- Eastern Closure

Location: NW¼, Sec. 26, T.12N, R.8E. The top of the eastern extremity of the Baraboo Ranges above Interstate 90/94.

Access: Continue north on Beich Road from Supplementary Stop L. Turn right (north) at junction with County W (1.0 mile) and right again (east) on Rowley Road (after .3 mile). Drive to the end of Rowley Road (another 1.5 miles). The outcrops at this locality can be seen ¼ mile north of the farmhouse across the field. At the time of writing, there was an access problem at this locality. The property was part of an estate being settled and no one lived there, but the local sheriff was apt to greet you on your return to your car! This will hopefully clear up soon, but it is best to check at the farm 1/3 mile west at the turn to Sky High Campground.

Features to be Seen: The quartzite can be traced around the east-closing and west-plunging (20-30°) hinge of the Baraboo Syncline by following the strike for a few hundred yards to the north and east. Quartzite cleavage (S_1') is prominent, dipping north at 35-55° roughly parallel to the axial surface of the syncline. Displacements of the bedding along the cleavage are clearly visible near the southern edge of the group of outcrops. There seems to be more than one penetrative quartzite cleavage present, but only one (identified here as S_1') can be consistently recognized. At the southern edge of the outcrop, there are potholes on the surface of the quartzite. One is lined with brown-weathering arenaceous sediment of lower Paleozoic type, and these are therefore likely to be Cambrian or Ordovician potholes. Once again, the present topography is coincident with that of the Cambrian. Please use restraint with rock hammers.

Remarks: The question of there being more than one cleavage in the quartzite arises in a number of localities around the syncline (see STOP 2, Supplementary Stops B and D). In some places a "second quartzite cleavage" probably is only late-stage local jointing unusually closely spaced. Elsewhere one may be observing a surface developed at a particular stage in the fold history due to local heterogeneities.

TABLE 13

TABLE 13

KEY TO STEREOPLOTS FOR FIELD TRIP LOCALITIES

	•	Pole to bedding (S_O)
	R	Pole to compositional layering in rhyolite
Main Phase	P	Pole to slaty cleavage (S_{1E})
	O	Pole to phyllitic cleavage (S_1) and / or axial surface of main phase minor fold
	●	Bedding / phyllitic cleavage (S_O/S_1) intersection and hinge lines of a few main phase minor folds.
	□	Pole to quartzite cleavage (S_1')
	×	Bedding / quartzite cleavage (S_O/S_1') intersection
	L	Longrain (mineral elongation lineation) in phyllitic cleavage
	S	Slickensides on bedding
	◊	Pole to faint strain – slip or crenulation cleavage (S_{1L})
	F	Axes of fine crenulations in phyllitic cleavage
Secondary Phase	△	Pole to strain – slip or crenulation cleavage (S_2) and / or axial surface of secondary phase minor fold
	▲	Axis of crenulation and / or secondary phase minor fold
Late Phase	3	Pole to axial surface of late phase minor fold
	3	Hinge line of late phase minor fold
Misc.	C	Pole to cleavage in rhyolite
	$\boxed{2}$	Pole to second quartzite cleavage (closely spaced jointing)
	2	Bedding / second quartzite cleavage intersection