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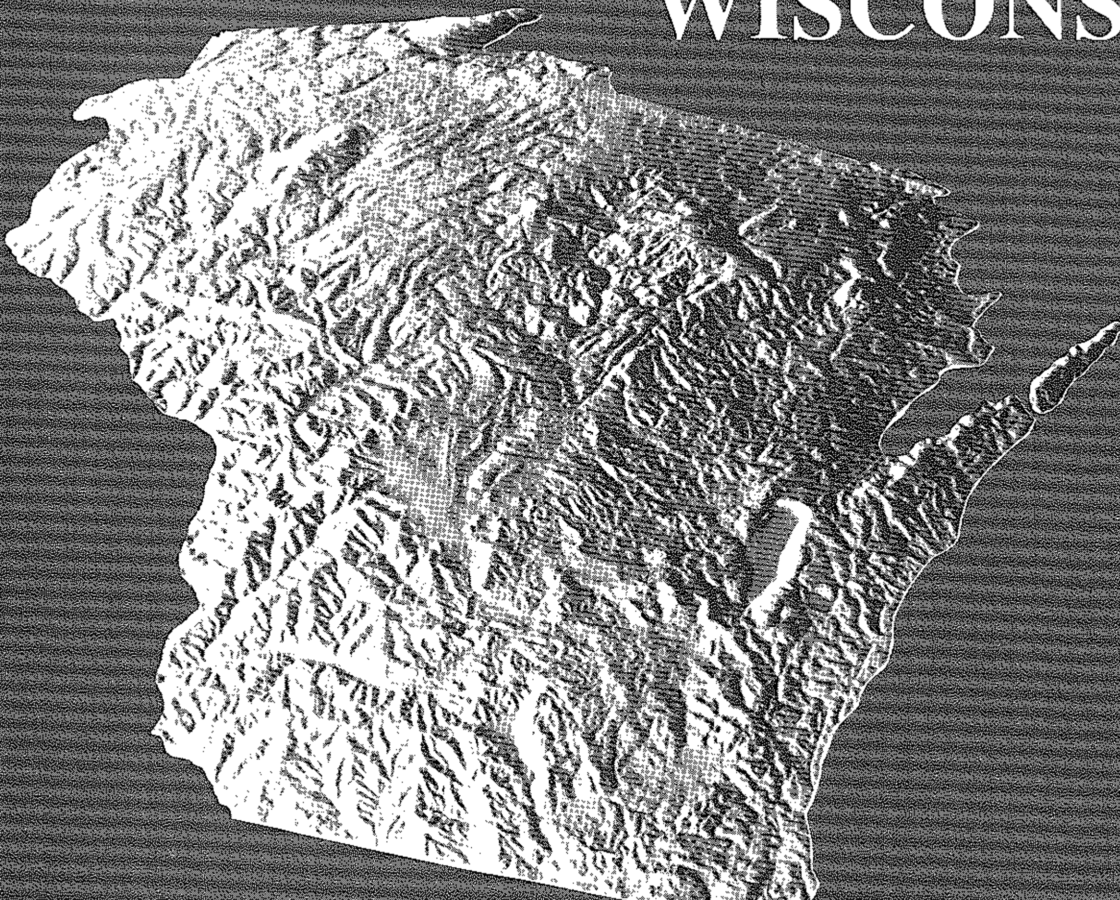
# GEOSCIENCE WISCONSIN



Quaternary Stratigraphy
Red Till Eastern Wisconsin
Cambrian Conglomerate Parfreys Glen
Gravity Modeling Vilas County
Density and Magnetic Susceptibility
Organic Geochemistry Decorah Formation

Wisconsin Geological and Natural History Survey
Geoscience Wisconsin • Volume 15 • 1995

# GEOSCIENCE WISCONSIN



edited by M.G. Mudrey, Jr.

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Geoscience Wisconsin ◆ Volume 15 ◆ 1995

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James M. Robertson, Director and State Geologist

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## PREFACE

*Geoscience Wisconsin* is a serial that addresses itself to the geology of Wisconsin — Geology in the broadest sense to include rock and rock as related to soil, water, climate, environment, and so forth. It is intended to present timely information from knowledgeable sources and make it accessible via scientific review and publication for the benefit of private citizens, government, scientists, and industry.

Manuscripts are invited from scientists in academic, government, and industrial fields. Once a manuscript has been reviewed and accepted, the authors will submit a revised copy of the paper, and the Wisconsin Geological and Natural History Survey will publish the paper as funds and time permit, distribute copies at nominal cost, and maintain the publication as a part of the Survey list of publications. This will help to insure that results of research are not lost in the archival system of large libraries or lost in the musty drawers of an open-file.

This collection of papers from academic colleagues represents a cross section of Wisconsin geology.

Rovey and Burucki present some new observations on well known exposures of Quaternary stratigraphy in southeast Wisconsin. Their work extends the southern margin of the oldest of the red tills in eastern Wisconsin, and suggest that the terminal ice front during deposition of the Keewaunee Formation was in standing water of glacial Lake Chicago.

McMillan, Vick and Shinn studied the well known Parfreys Glen, and determined from seismic methods that the present gorge is nearly coincident with a similar valley developed during deposition of the Parfreys Glen Formation. This observation may help to understand the evolution of the sea cliffs along the margins of the Baraboo hills.

Eckstein undertook a reevaluation of the gravity data in Vilas County. Her analysis helps to refine the position of the Niagara Fault of northern Wisconsin. This fault separates a dominantly young (1.8 Ga) terrane in the south (the Penokean fold belt) from a terrane defined by an older (2.6 Ga) Archean to the north. It is in this Archean terrane that much industrial exporation for diamonds has been understaken.

Dutch, Boyle, Jones-Hoffbeck, and Vandenbush systematically measured density and magnetic properties of Wisconsin rock. These data will permit refined geophysical modelling of gravity and magnetic surveys, and represents a significant increase in the amount of physical parameter data in the literature.

Blabaum evaluated the organic geochemistry of the Guttenberg Member of the Decorah Formation in the Upper Mississippi Valley zinc-lead district of southwestern Wisconsin. Her work finds that the Guttenberg has experienced a regional temperature no more than 100 °C, and that the depth of burial was about 700 m. This information, contrasted with the temperature of 200 °C in the ore horizons, will help to define regional fluid flow during mineralization.

We encourage submission of manuscripts relating to Wisconsin geology. Special consideration will be given to papers which deal with timely topics, present new ideas, and have regional or statewide implications.

M.G. Mudrey, Jr.
Editor, Geoscience Wisconsin
Wisconsin Geological and Natural History Survey

# RE-EXAMINATION OF QUATERNARY BLUFF STRATIGRAPHY, SOUTHEAST WISCONSIN

*Charles W. Rovey II[1] and Mark K. Borucki[2]*

## ABSTRACT

*Seven lithostratigraphic units from six ice advances are present within the Oak Creek Formation along Lake Michigan bluffs in southeastern Wisconsin, whereas only three till units were previously recognized. Two of the units are fine-grained glacial diamicton, apparently deposited into standing water, and five are till. To the north, each unit rises within the bluff and is successively truncated, exposing progressively older units at or near the bluff top.*

*Each till has a unique and consistent combination of color, texture and clay mineralogy. The characteristics of the oldest till are identical to the Shorewood till defined beneath Lake Michigan and is present in the bluff along the same reach delineated offshore for the Shorewood till. Based upon these similarities, the Shorewood till is provisionally correlated to the oldest Oak Creek till.*

## INTRODUCTION

Few geologists have studied the Pleistocene stratigraphic sequence in southeastern Wisconsin (fig. 1); Few, in particular, have studied the sequence exposed in the Lake Michigan bluffs. The first such study began in the mid-1970s when rising lake levels accelerated bluff erosion and sparked a comprehensive coastal inventory. As part of that inventory various geologists completed the first study on the lake-bluff stratigraphy. Results were compiled into a report with county by county appendices (Mickelson and others, 1977). In addition, Klauk (1978) and Acomb (1978) prepared theses as part of the study.

Results in the Mickelson and others (1977) study include identification of three formations. From oldest to youngest they were informally named till 1, 2 and 3; formally they are now the New Berlin, Oak Creek, and Kewaunee Formations, respectively (table 1; Mickelson and others, 1984). Each formation consists of multiple till units, however, only the Kewaunee Formation is formally subdivided. The oldest Kewaunee Formation Member (3a in Mickelson and others, 1977) is the only member present in this study area and is now known as the Ozaukee Member of the Kewaunee Formation.

One major aspect of the 1977 report can be revised; more than three Oak Creek till units are present in the bluff. This first became apparent during 1987, when record high lake levels coincided with a lack of ice armoring during an unusually mild winter. This resulted in erosion which exposed the most complete section probably ever seen along the shoreline. At numerous sites four distinct Oak Creek till units were unmistakably exposed in the bluff.

## PROCEDURES

### Field Sampling

Our major objective was to reinvestigate the sequence of Oak Creek till exposed in the Lake Michigan bluffs. Along one three-kilometer section (sections 25 and 36, fig. 2b) differentiation was simple; four tills were clearly superposed, separated from each other by stratified sediment or a boulder lag or both. To help trace and correlate the till units laterally, each was sampled along the three-kilometer section to establish characteristic parameters for unoxidized color, matrix texture and clay mineralogy. At each location samples were collected over a vertical and lateral distance of three to four meters or as thickness permitted. Mapping was then extended to both the north and south. In all cases care was exercised to collect only in-place material, and oxidized samples were avoided for color and clay mineral analysis.

### Textural Analysis

The percentage of sand, silt and clay in the sample matrix was determined using the total hydrometer method of Boyoucos (1962). Textural percentages determined using the Boyoucos method are not di-

[1] Geoscience Department, Southwest Missouri State University, 901 S. National Ave., Springfield, MO 65805

[2] Layne Geosciences, Inc, N. 4140 Duplainville Road, Pewaukee, WI 53072

*Figure 1.* Location map of the study area, southeastern Wisconsin.

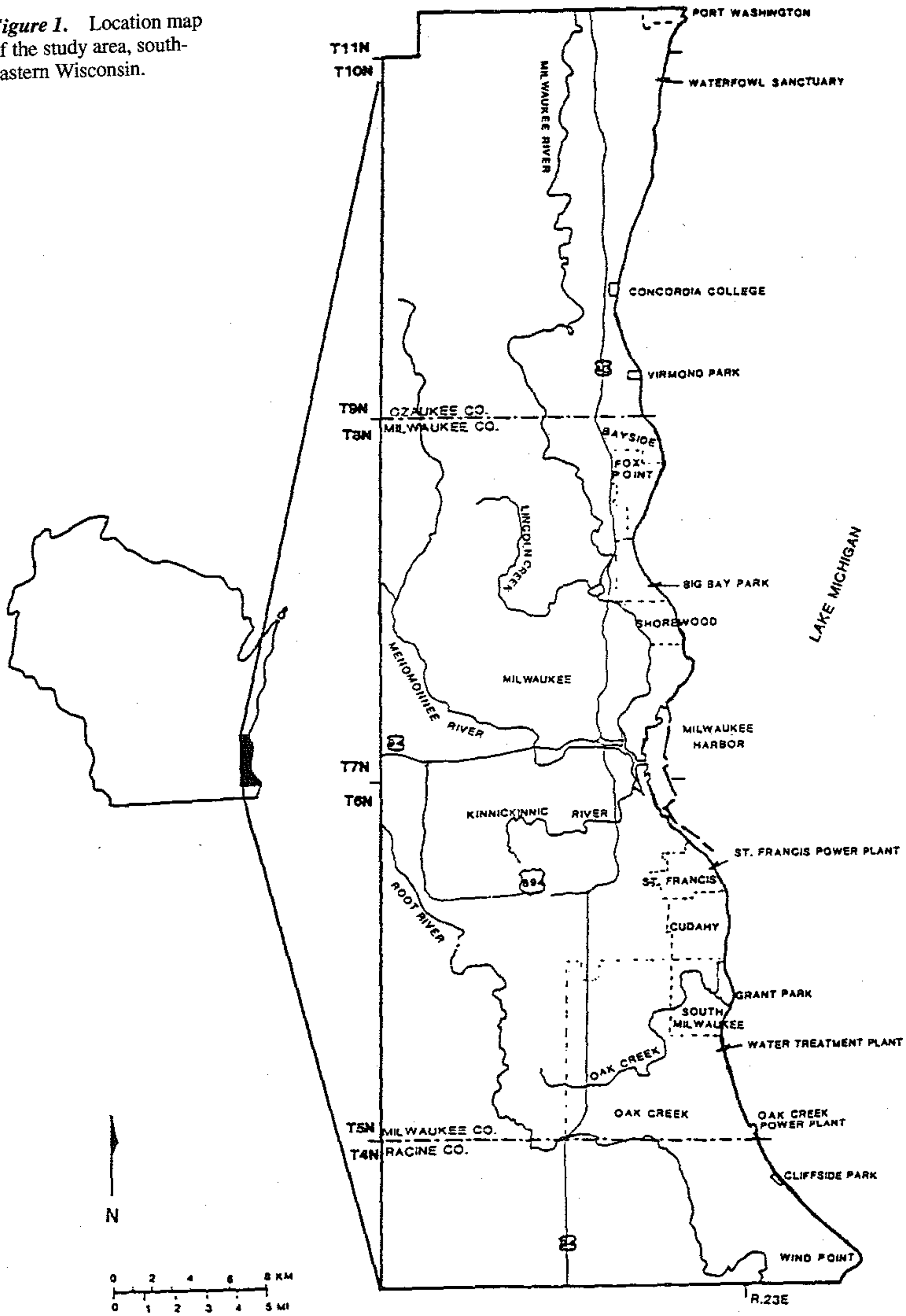

*Table 1.* Synopsis of glacial stratigraphy, southeastern Wisconsin.

| Mickelson and others, 1977 | Schneider, 1983 | Illinois | This study |
|---|---|---|---|
| Till 3A | Ozaukee Mbr. Kewaunee Fm. | Shorewood (Offshore) | Ozaukee Mbr. Kewaunee Fm. |
| Till 2 (a,b,c) | Oak Creek Fm. (1,2,3) | Wadsworth Mbr. Wedron Fm. | Oak Creek Fm. (1,2,3,4,5) |
| Till 1 (a,b) | New Berlin Fm. (1,2) | Haeger Till Mbr. Wedron Fm. | New Berlin Fm. (1,2) |

rectly comparable to those obtained using different procedures. This method is rapid and convenient for processing large numbers of samples, but differs from other hydrometer methods in that a larger sample (50 grams as opposed to 10-15 grams) is used. The sand fraction is also determined from a hydrometer reading, and not from sieving. The sand/silt boundary was 0.05 mm, because stable hydrometer readings could not be obtained for the usual 0.062 mm boundary.

Most previous studies in southeast Wisconsin used either the sieve and pipette technique, or determined clay percentage with the Boyoucos method, and determined sand percentage by sieving. Sieve and pipette analyses were conducted on 16 duplicate samples following procedures outlined in Walter and others (1978). The hydrometer sand values were, on average, 6% greater than the duplicate sieved samples, with no apparent dependence on clay or sand percentage. The deviation in clay percentage, however, is not as simple. A second order polynomial-fit between clay percents from the Boyoucos method and corresponding deviations from the pipette method gives the equation:

$$\text{Pipette\%} = \text{Hydrometer\%} + 1.0 - .16(\text{Hydrometer \%}) + 5.2 \times 10^{-3} (\text{Hydrometer\%})^2$$

for converting Boyoucos to pipette clay percentage. Details are available in Borucki (1988).

### Clay Mineral Analysis

The semi-quantitative method of clay mineral analysis by H.D. Glass was used in this study. The method is termed "semi-quantitative" in that the clay minerals are lumped into three broad categories, expandable clay, illite and kaolinite + chlorite. The percentages of each category are not deemed highly accurate, but they are very precise and diagnostic as demonstrated by Hallberg and others (1978) who documented the procedures and achieved standard deviations of approximately three percent in each fraction from duplicate samples. The technique is proven useful for stratigraphic differentiation of individual till units over wide areas (see for example Wikham and others, 1988; Hallberg, 1980). In southeastern Wisconsin, the method was previously used by Acomb (1978) and Acomb and others (1982) for bluff stratigraphy, and by Schneider (1983) and Hansel (1983) for inland localities. Duplicate samples (13) were sent to the Illinois State Geological Survey for analysis. The average deviation in each clay fraction was less than 1% (see Borucki, 1988).

## RESULTS

### General

Field mapping in combination with the laboratory analyses identified five Oak Creek till units and two stratified glacial diamicton units in the Lake Michigan bluffs. Their distribution and laboratory characteristics are found in figures 2 and 3 and table 2. All of the (sometimes) complex thickness variations are not shown on the figures, but rather general values and trends are shown, because the bluffs erode through time, and thicknesses invariably change in the new exposures. The primary purpose is to show the general stratigraphic relations among the different tills and diamicton.

As in the original survey numerous gaps in exposure were encountered. Differentiation between major and inconsequential gaps is subjective, but the gaps shown in figures 2 and 3 are those which we believe could be reasonably deemed to introduce uncertainty in lateral correlation. Each unit, oldest to youngest, is sequentially discussed below.

### Diamicton 1

The oldest mappable unit within the Oak Creek Formation is a deposit of uncertain origin. Schneider and Need (1985, p. 55) described this unit at the St. Francis Power Plant (fig. 2B) stating : "The basal part of the Oak Creek Formation is commonly a water-laid diamicton." This diamicton is found wherever the underlying New Berlin is itself exposed, at least as far north as Port Washington (fig. 2b, 3a-3c). The diamicton is

*Figure 2.* Comparison of bluff stratigraphy south of Milwaukee between this study and Mickelson and others (1977). Bluff elevations are taken from Mickelson and others (1977). Numbers along the bottom of the upper profile are one mile sections. Selected general sampling locations and average composition for the Oak Creek till units are shown as space permits. Abbreviations:
O.C. 1 - Oak Creek 1
O.C. 2 - Oak Creek 2
O.C. 3 - Oak Creek 3
O.C. 4 - Oak Creek 4
O.C. 5 - Oak Creek 5

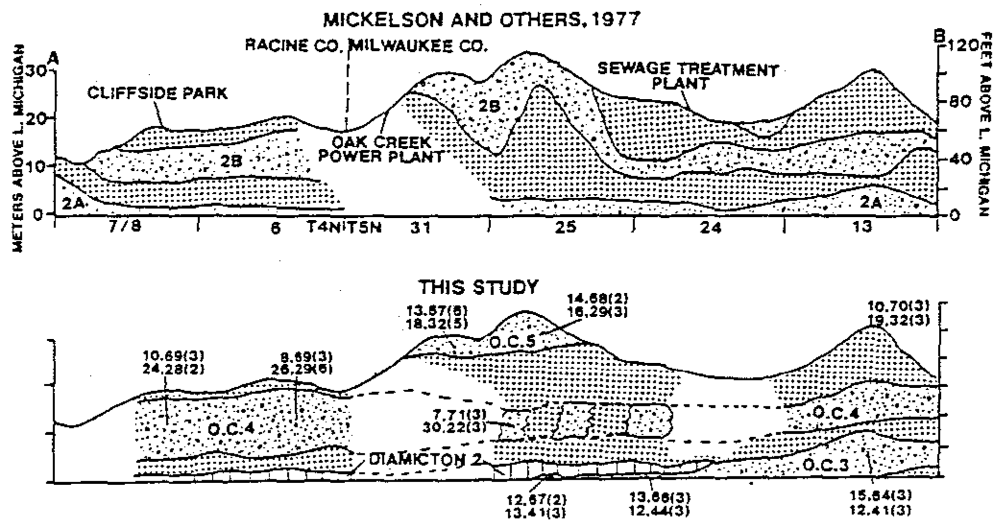

FIGURE 2A

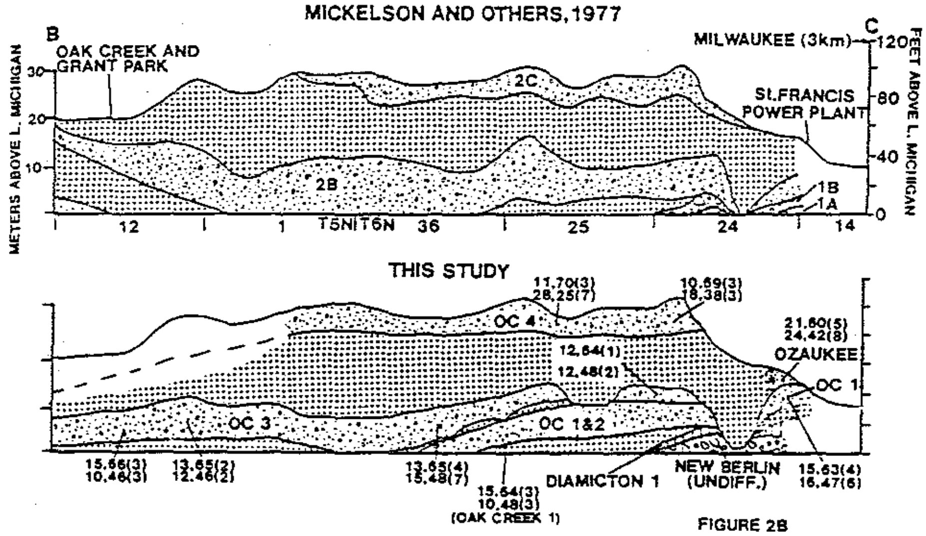

FIGURE 2B

**Explanation for figures 2 and 3**

15 . 64 (3) %expandables % illite (number of samples analyzed)
10 . 48 (4) %sand %clay (number of samples analyzed)
UNDIFFERENTIATED LAKE SEDIMENT
FINE GRAINED GLACIAL DIAMICTON
FINE GRAINED TILL
COARSE GRAINED TILL

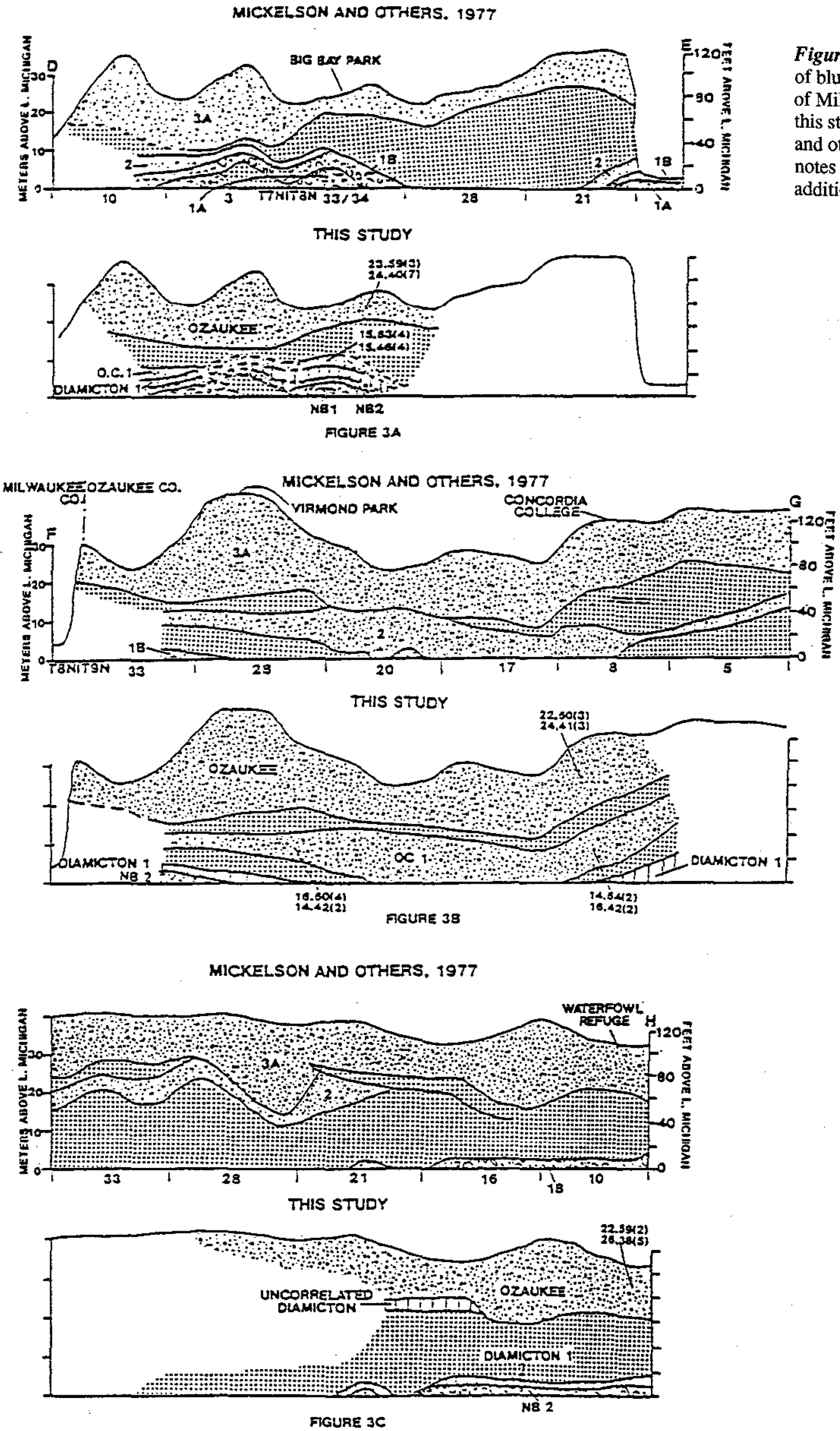


*Figure 3.* Comparison of bluff stratigraphy north of Milwaukee between this study and Mickelson and others (1977). See notes on figure 2 for additional descriptions.

*Table 2.* Mean composition of glacial units in Lake Michigan bluffs, southeastern Wisconsin. Textural percents determined by the Bouyoucos (1962) method using sand/silt and silt/clay boundaries of 0.05 mm and 0.002 mm, respectively. X = arithmetic mean (S) = standard deviation in percent, [N] = number of samples.

| BLUFF UNIT | TEXTURE | | | | | | | CLAY MINERALOGY | | | | | | | MUNSELL COLOR (DRY) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | Sand | | Silt | | Clay | | | Expandables | | Illite | | Kaolinite + Chlorite | | | |
| | X | (S) | X | (S) | X | (S) | [N] | X | (S) | X | (S) | X | (S) | [N] | |
| Ozaukee | 23 | (5.9) | 36 | (3.8) | 41 | (3.1) | [26] | 22 | (2.2) | 59 | (2.6) | 19 | (1.9) | [16] | Lt. Reddish Brown, 5YR |
| Oak Creek 5 | 19 | (3.8) | 52 | (2.1) | 29 | (3.7) | [11] | 14 | (.9) | 67 | (1.4) | 19 | (2.2) | [9] | Lt. - Dk. Gray, 10YR |
| Oak Creek 4 | 22 | (5.8) | 47 | (3.0) | 31 | (5.3) | [36] | 10 | (2.2) | 69 | (2.0) | 21 | 2.0) | [28] | Lt. - Dk. Gray, 10YR |
| Diamicton 2 | 13 | (8.7) | 45 | (3.9) | 42 | (10.5) | [8] | 13 | (2.4) | 66 | (2.7) | 21 | (0.9) | [8] | Lt. Gray, 10YR |
| Oak Creek 3 | 13 | (2.4) | 42 | (3.6) | 45 | (3.8) | [22] | 13 | (1.7) | 65 | (1.5) | 22 | (1.6) | [20] | Lt. Gray-Pinkish Gray,10YR- 5YR |
| Oak Creek 2 | 40 | (3.8) | 34 | (3.1) | 26 | (6.4) | [9] | 18 | (3.0) | 60 | (1.8) | 22 | (2.2) | [10] | Pink - Pinkish Gray, 5YR |
| Oak Creek 1 (Total) | 14 | (2.8) | 40 | (3.3) | 46 | (3.4) | [19] | 16 | (1.7) | 62 | (2.5) | 22 | (1.9) | [15] | Pinkish Gray, 5YR |
| Oak Creek 1 (N. of Milwaukee) | 15 | (1.5) | 41 | (2.5) | 44 | (2.2) | [8] | 15 | (1.6) | 62 | (2.3) | 22 | (2.0) | [10] | Pinkish Gray, 5YR |
| Oak Creek 1 (S. of Milwaukee) | 13 | (3.1) | 39 | (3.5) | 48 | (3.3) | [11] | 16 | (1.7) | 62 | (3.1) | 22 | (1.5) | [5] | Pinkish Gray, 5YR |
| Diamicton 1 | 19 | (12.9) | 42 | (7.0) | 39 | (11.4) | [20] | 14 | (4.5) | 65 | (4.8) | 21 | (2.0) | [19] | Dk. Gray, 10YR |
| New Berlin 2 | 30 | (5.0) | 45 | (6.0) | 25 | (7.1) | [12] | 18 | (3.2) | 63 | (3.0) | 19 | (1.3) | [8] | ———— |
| New Berlin 1 | 57 | (5.4) | 30 | (2.3) | 13 | (4.3) | [5] | — | — | — | — | — | — | — | ———— |

consistently 1.5 to 2.5 meters thick, but the nature of both contacts is variable. The lower ranges from a sharp contact directly above the New Berlin Formation to a gradational change from laminated to bedded sediment which overlies the New Berlin Formation. The upper contact is usually gradational to laminated silt and clay, but locally this same transition is abrupt. The texture and clay mineralogy are quite variable compared with the till (table 2). However, there is an apparent trend towards greater consistency in the upper parts where the deposit is characterized by the mean values in texture (table 2), but a higher illite and lower expandable clay content (approximately 70% and 10%, respectively) than average. The dark gray (10YR) color and high illite percentage are typical of the Oak Creek Formation, but the sedimentologic origin of this oldest Oak Creek deposit is uncertain. Although locally it appears massive and to be a till, close inspection invariably discloses a crude stratification (fig. 4). Therefore, the non-genetic term "diamicton" is applied to exposures of this unit in the bluff.

## Oak Creek 1 and 2 tills

The oldest definite till is here termed Oak Creek 1. It is distinguished by its pinkish gray (5YR) hue, finer texture and lower illite percentage than overlying Oak Creek till (table 2). The lower contact is abrupt, overlying stratified sediment above diamicton 1. The same sequence was previously reported at the St. Francis Power Plant by Christensen and Schneider (1984). Oak Creek till north of Milwaukee is identical in color, texture and clay mineralogy and vertical stratigraphic sequence to the Oak Creek 1 unit south of Milwaukee (table 2, figs. 2, 3). Therefore, all Oak Creek till north of Milwaukee is correlated to the Oak Creek 1 till.

Because of the color similarity between the pinkish Oak Creek 1 and 2 till units which are in strong contrast to the gray underlying and overlying units, and because the contact between the two is difficult to trace, the Oak Creek 1 and 2 tills units are undifferentiated south of Milwaukee. The two till units do, however, contain subtle differences in clay mineralogy and marked differences in texture (table 2).

The Oak Creek 2 till is only present south of Milwaukee, where the Oak Creek 1 till is occasionally absent or beneath lake level. Consideration must be given to whether the Oak Creek 1 and 2 till units are really distinct or whether they are different facies deposits of the same glacial advance. Might the Oak Creek 2 till be a supraglacial deposit overlying a basal Oak Creek 1 till, and might this origin account for the difference in texture and apparent patchy occurrence of the Oak Creek 1 till south of Milwaukee? Several observations are inconsistent with a supraglacial origin of the Oak Creek 2 till:

1. Both units have extremely uniform textures.

2. Neither unit contains interbedded stratified material, flow, or deformation structures, although locally a cobbly diamicton with flow structure interbedded with stratified sediments overlies the Oak Creek 2 till.

Both units appear to be basal till. Another question might be whether the Oak Creek 1 till is merely a basal contamination of the sandier Oak Creek 2 till? While we can not discount this possibility entirely, several observations are inconsistent with it also:

1. Where exposed, the contact is usually abrupt and planar, not gradational.

2. Discrete inclusions of the Oak Creek 1 till are present in the basal few feet of the Oak Creek 2 till, but no inclusions of the Oak Creek 2 till have been found within the Oak Creek 1 till.

3. Locally, the units are separated by laminated silt.

We conclude that the two units are deposits from two separate glacial advances.

## Oak Creek 3 till and diamicton 2

The Oak Creek 3 till is traceable from south of the St. Francis Power Plant to within 2 kilometers of the Milwaukee/Racine County line (figs. 2a, 2b). The till is separated from the undifferentiated Oak Creek 1 and 2 tills by coarse stratified sediment or boulder pavement. Matrix texture is similar to the Oak Creek 1 till (table 2), but the color and clay mineralogy are clearly different and distinguishable.

From section 24 (fig. 2a) southward the Oak Creek 3 till has a diffuse, undulating upper contact with a second, fine-grained diamicton, here termed diamicton 2. Diamicton 2 is variable in thickness, crudely bedded, and contains abundant flow and deformation structures such as folded silt laminae (fig. 5). Where it is best exposed 1.5 km north of the Racine-Milwaukee County line, its thickness ranges from one

***Figure 4.*** Diamicton 1 overlying New Berlin 2 Till. Contact is at hammer. Note crude stratification in diamicton. Photo taken at Virmond Park, figure 1, figure 2B, section 28.

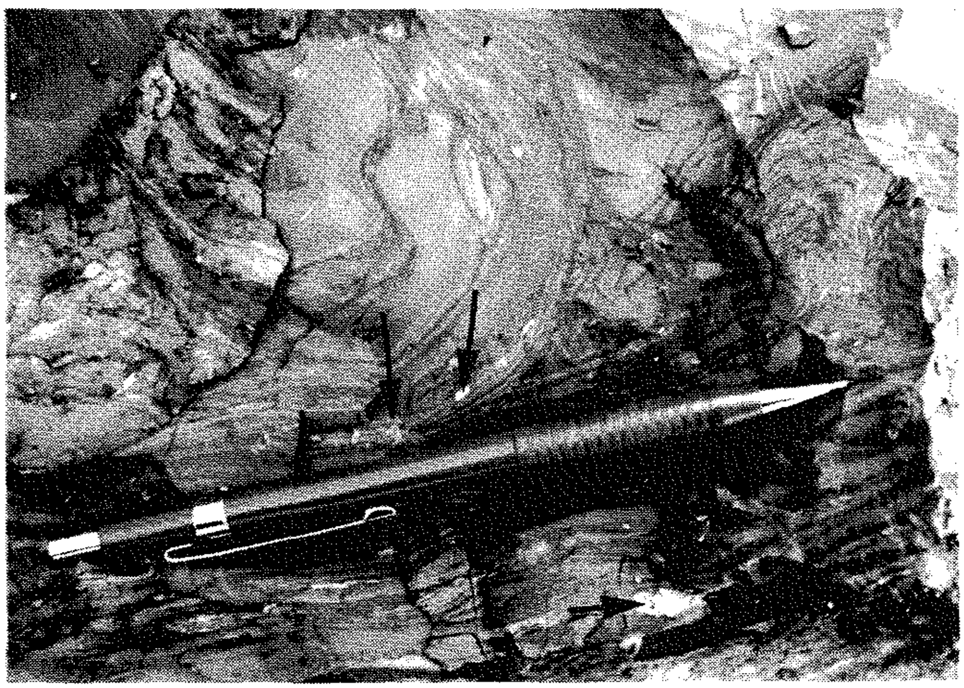

***Figure 5.*** Folded and contorted laminae in Diamicton 2. Arrows point to unconsolidated rip-up clasts. Photo taken 1.5 miles north of the Oak Creek Power Plant (figure 1, figure 2A, section 25).

to six meters. Its texture is similar, but more variable than the Oak Creek 3 till, and the clay mineralogy is identical (table 2). The upper contact is gradational to laminated silt and clay. The similarity in clay mineralogy and the diffuse lower contact are evidence that the origin of diamicton 2 is closely related to the Oak Creek 3 till. The genetic origin of diamicton 2 is discussed in more detail below.

### Oak Creek 4 till

The Oak Creek 4 till is easily traced in the upper six meters of the bluff from south of the St. Francis Power Plant (Fig. 2b) to north of Grant Park (fig. 2b). It has a distinctly coarser texture than the underlying Oak Creek 1 and 3 till units, and is also set apart by its higher illite/lower expandable clay percentages (69%, 10%) (Table 2). Another unique characteristic is the variability in matrix texture compared with other till. Samples taken within three meters of each other commonly vary by 8% or more in each size fraction. The continuous exposure at the bluff top ends in the vicinity of Grant Park (fig. 2b), however, south of the park an identical till is present in the mid-bluff (fig. 2a). Southward from section 13 (Fig 2a) the Oak Creek 4 till is even more difficult to trace. There, the occurrence is patchy, and the entire section above diamicton 2 and below Oak Creek 5 till is chaotic.

South of the 1.5 km gap in exposure at the Milwaukee/Racine County line (fig. 2a), the deposits are no longer chaotic and a continuous till from the mid-to-upper parts of the bluff is again present. Because of the identical vertical sequence above diamicton 2, and identical clay mineralogy and texture - including textural variability - this till is again correlated to the Oak Creek 4 till.

### Oak Creek 5 till

The Oak Creek 5 till caps the upper four to five meters of bluff from north of the Oak Creek Power Plant, northward to the central part of section 25 (fig. 2a). It is the same till described capping the bluff by Schneider (Mickelson and others, 1984) at the Oak Creek type section 1.5 km north of the Oak Creek Power Plant. Oak Creek 5 till has a similar average texture to the Oak Creek 4 Till, but averages 14% expandable clay and 67% illite; not a single sample contained less than 10% expandable clay and more than 70% illite, characteristics of the Oak Creek 4 till. The Oak Creek 5 till overlies the chaotic sediment described above, which in turn overlies a patchy occurrence of a 70% illite till, the Oak Creek 4 till. To the north, the Oak Creek 5 till is easily traced at the top of the bluff to a point in the northern half of section 25 beyond which it is not present.

## DISCUSSION AND IMPLICATIONS

### Glacial Diamicton

Three Oak Creek till units were described in the Mickelson and others (1977) study. All intervening sediment was undifferentiated. Not only are there five Oak Creek till units, but two mappable diamictons are included within the undifferentiated sediment. The diamicton units resemble till in many respects, but contain crude stratification. There are two explanations for the genesis of such diamicton. The first is that the stratification was produced by a series of mass flows emanating from the glacier margin (Hartshorn, 1958; Boulton, 1968). Dreimanis (1976) and Evenson and others (1977) extended this mechanism to flows into proglacial lakes and each modified the terrestrial term flow till to the analogous terms waterlaid till and subaqueous flow till. May (1977) subsequently proposed the term lacustrotill. Gibbard (1980) and Hicock and others (1981), however, retained versions of the term flow till. Regardless of nomenclature, the identifying characteristics include a crude, but deformed stratification, abundant evidence of flowage such as flow noses, roll-up structures and inclusions of unconsolidated silt rip-up clasts, and widely variable thickness. All of these attributes characterize diamicton 2, so we interpret it as the product of a series of mass flows. Because it has identical mean textural and clay mineral properties as the underlying Oak Creek 3 till, it was probably deposited during the same ice advance. But, because it commonly grades upward into laminated lacustrine silt, the flows were deposited into standing water during glacial retreat. A similar, but more detailed, description and interpretation of diamicton 2 was given by Jung and Powell (1985) for exposures on either side of the Oak Creek Power Plant (R.D. Powell, verbal communication, 1989).

The second explanation for stratified diamicton attributes deposition to basal melt-out beneath an ungrounded ice sheet (May,1977; Gibbard, 1980). Such deposits would have few, if any, flow structures, would be more constant in thickness, and would be vertically gradational into an unstratified till, deposited as grounded ice overrode deposits from the floating ice shelf. Most of these attributes characterize diamicton 1, but there is no conclusive evidence for a vertical transition into a true till. The upper part of diamicton 1 has a greater consistency in texture and clay mineralogy, possibly indicating less reworking, but it too appears crudely stratified. Therefore, although diamicton 1 was probably deposited into water, perhaps by melt-out beneath a floating ice margin, conclusive evidence is lacking and its genetic origin is still in doubt.

### Till Sequence

The pattern exposed in the southeast Wisconsin lake bluffs is counter to the regional trend. The general pattern of Lake Michigan Lobe Woodfordian deposits is described as shingled (Johnson and Hansel, 1986) with older till thinning and pinching out to the north beneath younger tills. However, the pre-Ozaukee pattern (fig. 2 and 3) is the reverse of this normal sequence. To the north, younger units are progressively truncated, successively uncovering older units. Two different mechanisms, non-deposition or erosion, may account for this pattern. Non-deposition would result if successively younger ice advances flowed farther south down the Lake Michigan basin - much as a valley glacier - before spilling out of the basin and depositing younger till in the vicinity of the present shoreline. However, this explanation stretches credulity. A more likely mechanism is that the truncation is due to pre-Ozaukee isostatic rebound, differential uplift and erosion to the north. As a result, younger units were eroded and older deposits exposed.

## CORRELATION OF THE SHOREWOOD TILL

The Shorewood was defined beneath Lake Michigan solely from color and clay mineralogy of three core samples (Lineback and others, 1974). The reported color (pinkish gray, 5 YR) and average clay mineral content (15, 60 and 25 percent expandable clay, illite and kaolinite + chlorite, respectively) raise speculation that the Shorewood unit could be the offshore equivalent of the Oak Creek 1 till. This seems plausible because no other onshore till (Mickel-son and others, 1984) has a composition approaching the Shorewood unit. Unfortunately previous workers were not aware of the existence of Oak Creek 1, and correlation attempts have been controversial.

Because of the pinkish hue, Lineback and others (1974) assumed that the Shorewood must be the equivalent of the oldest red till unit known, the Ozaukee. Because the Ozaukee was not formally defined at that time they gave it the name Shorewood for the prominent exposures at the bluff top along the Village of Shorewood, Wisconsin. They then averaged the clay mineral values from the offshore samples with an onshore sample of their presumed equivalent till exposed in the bluff (apparently an oxidized - and thus altered - sample of the Ozaukee), disregarding marked differences in both clay mineralogy and color. We believe there was insufficient evidence for correlation; the onshore sample was inappropriately averaged with the offshore material causing a false average clay mineral composition for the Shorewood. Hence, in later reports the incorrect averages in table 2 of Lineback and others were adopted. We use the correct averages in their table 1.

Mickelson and others (1977) also correlated the Shorewood to the oldest red till unit, 3A (Ozaukee). Acomb and others (1982) apparently recognized the discrepancies in color and clay mineralogy and provisionally correlated the Shorewood to till unit 2c (Oak Creek 4 till), which they believed was the youngest Oak Creek till. But, the Oak Creek 4 till color and clay mineralogy are even more distinct from the Shorewood than is the Ozaukee. Recognizing the problems in correlation, Mickelson and others (1984) correlated the Shorewood Till to a loosely defined interval spanning the upper Oak Creek and lower Kewaunee Formations.

All past correlation attempts assumed that the Shorewood Till, as defined beneath Lake Michigan, is stratigraphically above the Wadsworth/Oak Creek till. This assumption originated with Lineback and others (1974) from several offshore seismic profiles from which they concluded that the southern extent of the Shorewood Till was a terminal moraine north of points where Wadsworth Till was recovered. However, on none of the lines is the reflector beneath the Shorewood, and presumably marking the top of the Wadsworth Till traceable to a point where Wadsworth/Oak Creek till was recovered. Even if their interpretation is correct, and the Shorewood Till overlies some part of the Wadsworth/Oak Creek units, it does not follow that the entire Wadsworth/ Oak Creek Formation lies beneath the Shorewood unit. In the bluffs diamicton 1 with dark gray, high illite Wadsworth/Oak Creek composition is present below the younger Oak Creek 1 till. Therefore, the correlation of the Oak Creek 1 till to the Shorewood Till is compatible with the interpretation of the Shorewood till partly overlying the Wadsworth/Oak Creek Formation beneath Lake Michigan, but does not preclude the possibility that additional Wadsworth/Oak Creek till may overlie the Shorewood Till.

The interpretation that the Shorewood Till must be entirely younger than the Wadsworth/Oak Creek till was based on the implicit assumption that the normal or shingled pattern of tills continues northward in the Lake Michigan Basin. On this premise, a till with distinctly different composition found north of the Wadsworth/Oak Creek subcrop must be stratigraphically higher. At least along the current shoreline this premise is false. The same pattern can be present beneath Lake Michigan as well if differential uplift and erosion are responsible for the exposure of successively older tills northward along the current bluffs. The Shorewood till samples were all recovered from the mid-lake high which probably was exposed to subaerial erosion during low stands of Glacial Lakes Milwaukee and Chicago, and were all recovered at the approximate latitude as Milwaukee, the same latitude where the Oak Creek 1 till is finally uncovered from beneath younger Oak Creek tills (fig. 6).

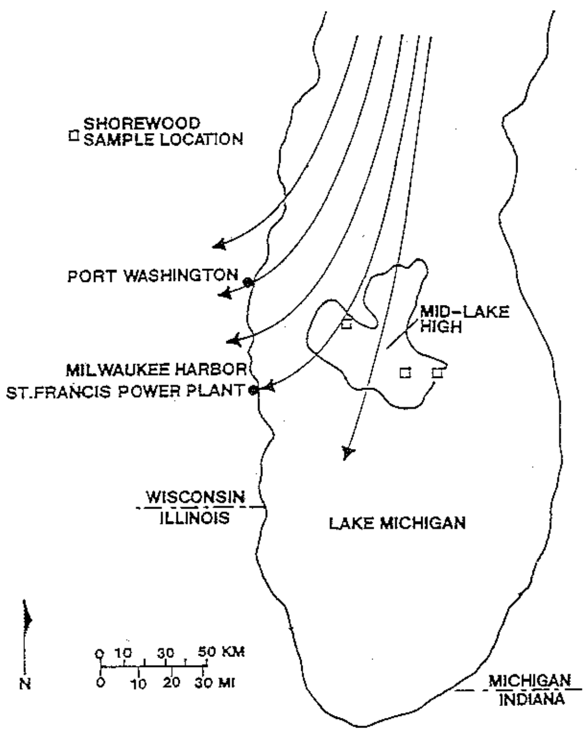


*Figure 6.* General flow paths (arrows) of the eastern half of a hypothetical Late Woodfordian ice-front.

The difference in composition between the onshore Ozaukee and offshore Shorewood tills might be explained by entrainment of debris and progressive alteration of clay mineral composition downglacier. Therefore, the Shorewood till actually could be the offshore Ozaukee till.

The questionable assumption that a glacier would entrain a significant amount of fresh, previously unincorporated, material near its terminus can be tested. The direction of flow for the Late Woodfordian Lake Michigan lobe is primarily north - south with an east - west component near the glacial terminus (fig. 6), implying a north - south compositional gradient, if compositions were indeed altered. A north - south compositional gradient does not exist within the two till units in question, the Ozaukee and Oak Creek 1 units. Younger tills also do not change in composition over approximately 70 km (Acomb, 1978). The composition of the Ozaukee till is constant from Port Washington, south to its terminus at the St. Francis Power Plant (Borucki and Rovey, 1989; fig. 2 and 3) This 50 kilometer traverse spans the distance over which Shorewood till samples were recovered. The Oak Creek 1 till also has constant composition over 50 km (fig. 2, 3). Therefore an east - west compositional gradient is unlikely, and these till units beneath Lake Michigan should have similar composition to their onshore equivalents.

If the term Shorewood till indeed has any stratigraphic significance, the most likely onshore correlation is to the Oak Creek 1 till. It follows that the Shorewood is not a younger overlying till, but part of the basal Wadsworth/Oak Creek sequence.

## SUMMARY

The glacial stratigraphy exposed within the Lake Michigan bluffs is more complex than previously mapped. The Oak Creek Formation includes seven mappable fine-grained units. Five of these are basal till, but the remaining two are crudely stratified diamicton apparently deposited into standing water. The northward succession of older units in the lake bluff result from the southerly dip of the seven units. The most likely explanation for this pattern is an episode of differential uplift and erosion due to isostatic rebound. The oldest definite Oak Creek till is uncovered at the same approximate latitude as offshore samples of the Shorewood Till beneath Lake Michigan, and is the only known onshore till with Shorewood composition. Therefore, the Shorewood Till beneath Lake Michigan is provisionally correlated to the basal portion of the Oak Creek Formation.

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# The Southern Limit of Red Till Deposition in Eastern Wisconsin

*Charles W. Rovey II*[1] *and Mark K. Borucki*[2]

## ABSTRACT

*Accelerated bluff erosion along Lake Michigan's Wisconsin shoreline exposed several features not previously noted at the St. Francis Power Plant exposure south of Milwaukee. Reddish rhythmites in the upper bluff grade northward into a massive red diamicton over a distance of 90 m. Based on stratigraphic sequence, lithologic similarity and re-mapping of the Ozaukee terminal moraine, the red diamicton is correlated with the Ozaukee Member of the Kewaunee Formation, the oldest of Wisconsin's red till units.*

*The absence of Ozaukee till south of this exposure proves that it is the southern limit of Wisconsin's red till. The gradation from till to rhythmites suggests that the terminal ice front was in standing water of glacial Lake Chicago.*

## INTRODUCTION

The record of Pleistocene glaciation exposed in the Lake Michigan bluffs at and around the St. Francis Power Plant exposure (named for the now largely-demolished Wisconsin Electric Power Plant and called the St. Francis exposure here, fig. 1) has intrigued geologists for years. The bluff profile south of the plant was described by Mickelson and others (1977), Lasca (1983), Christensen and Schneider (1983), Schneider and Need (1985) and Monaghan (1990) and was visited during the 1983 Northeast Section Geological Society of America meeting. The general stratigraphic sequence described in these reports is till of the New Berlin Formation at the base of the bluff overlain successively by lake sediment, water-laid diamicton and till of the Oak Creek Formation (fig. 2). A boulder lag progressively truncates the Oak Creek Formation and defines the base of a channel which was successively infilled by sandy sediment and then reddish rhythmites deposited in Glacial Lake Chicago. The bluff is planed by a wave-cut terrace slightly below the Glenwood (195 m) level of Lake Chicago.

These deposits represent significant events in the Late Woodfordian deglaciation history of the Lake Michigan basin (Hansel and others, 1985). After deposition of the New Berlin Formation during the later Cary ice advances, the basin was intermittently free of ice and occupied by a series of glacial lakes. The Oak Creek Formation was deposited during the latest Carry advances which successively left the prominent Valparaiso, Tinley and Lake Border morainic systems around the southern half of Lake Michigan. During retreat, phases of Glacial Lake Chicago formed at the Glenwood stage in the southern portion of the basin and expanded northward as the ice retreated

The site is notable for two reasons: it is the southernmost bluff exposure of the New Berlin Formation, and lake sediment overlying the New Berlin Formation as instrumental in recognizing and defining proglacial Lake Milwaukee as a separate predecessor to proglacial Lake Chicago (Schneider and Need, 1985). The stratigraphic record at the St. Francis exposure is more complete and important than previous authors thought. The importance was recognized during this investigation primarily because (1) the record-high lake levels of 1987 accelerated erosion and produced a fresh, uncovered bluff face, and (2) a stabilization project initiated by the City of St. Francis to combat that erosion progressively covered the entire bluff exposure from base to top with fill. Because the fill was emplaced in terraces, detailed inspection of the upper bluff was facilitated during the final stages of the project. The fill now completely covers the exposure, making further observations impossible without large-scale excavation.

[1] Geoscience Department, Southwest Missouri State University, 901 S. National Ave., Springfield, MO 65805

[2] Layne Geosciences, Inc, N. 4140 Duplainville Road, Pewaukee, WI 53072

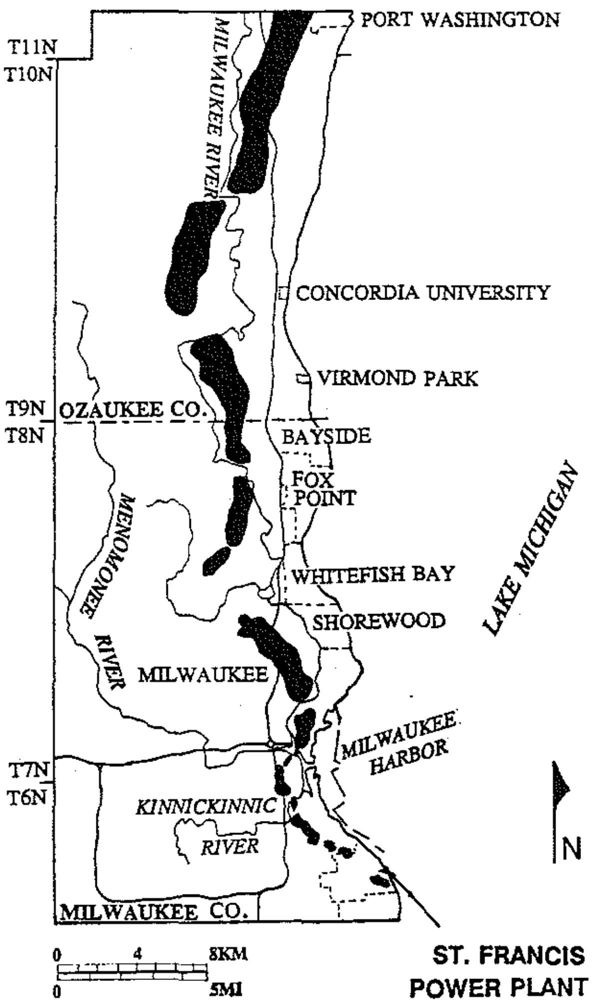


*Figure 1.* Location map, southeast Wisconsin. Darkened areas show the Ozaukee terminal moraine, as mapped in this study. Ridges in the Menomonee and Kinnickinnic River valleys are from Need (1983).

## NEW DESCRIPTION

### Boulder Lag

Several features not previously mentioned were noted during this study. There is not a single boulder lag, but two, defining two superposed erosion surfaces (fig. 3). The lower surface slopes gently to the south, and is present between the water-laid diamicton of the basal Oak Creek Formation and the overlying Oak Creek till. The middle part of the diamicton has a silty clay matrix and resembles till in many respects, but is crudely stratified, suggesting deposition of sedimentation in standing water instead of direct deposition from ice. The diamicton overlies the cobbly New Berlin Formation and grades into laminated silt and sand at both contacts. Thus, the diamicton unit as used here includes the lake sediment which Schneider and Need (1985) proposed as evidence for Glacial Lake Milwaukee.

The upper lag defines the previously described channel, originating between the Oak Creek till (above the lower lag) and a separate overlying massive, reddish diamicton, not previously reported. The upper lag drops steeply to the top of the New Berlin Formation where it merges with the lower lag. At the north end of the exposure the upper lag is not as pronounced, and up to 0.3 m of bedded sand (not depicted in fig. 3) are locally present between the Oak Creek Formation and the red diamicton.

### Channel Sand

The lower sandy sediment confined within the channel consists of alternating cross bedded and rippled sets. The occurrence is unusual and possibly diagnostic because it is the only bluff exposure of such an infilled channel known from south of Port Washington (Mickelson and others, 1977; Rovey and Borucki, 1995). It is similar geometrically and sedimentologically to channels and infilled sediment described by Rust (1976), who interpreted them as sub-aqueous channels cut by streams emanating from the margin of a nearby grounded ice mass into a proglacial lake. If such an interpretation is correct for the channel at St. Francis, it was cut and partially filled while a glacial ice front was within several hundred meters of the exposure.

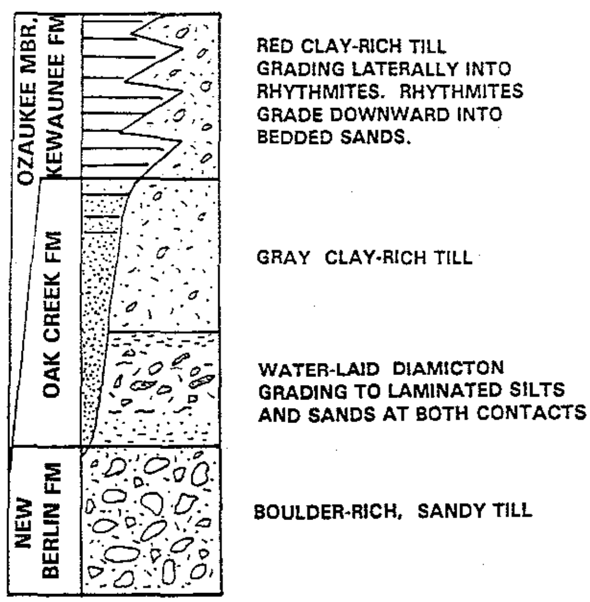


*Figure 2.* Late Woodfordian stratigraphy at the St. Francis Power Plant exposure.

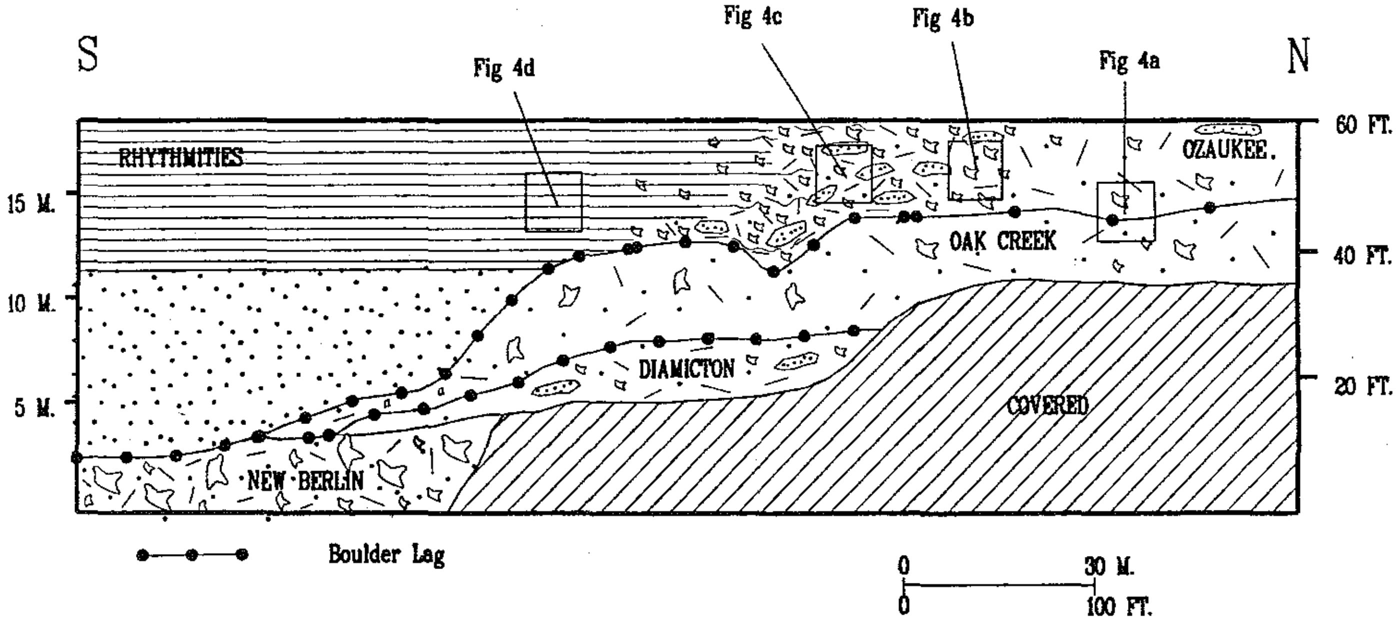


*Figure 3.* Schematic section of stratigraphic relationships formerly exposed at the St. Francis Power Plant Site. See Figure 2 for a description of the units and formal stratigraphy.

The channel sand grades upward into rhythmite. Within the upper meter of the sand are two thin (approximately 8 cm) layers of sandy red clay. These layers contain evidence of deposition from density currents, including scoured, erosional bases and normally graded bedding with a rippled sandy base grading upward to laminated clay. The clay is similar to that comprising the fine-grained part of each couplet in the overlying rhythmites. The color of the clay is distinct, and throughout the Lake Michigan basin it marks the initial appearance of a new sediment source generally associated with the Port Huron ice advances (Hansel and others, 1985). Apparently a connection was established between the Lake Michigan and Superior basins during the Cary-Port Huron retreat, and red clay from the Lake Superior region was transported and deposited in the Lake Michigan basin. Subsequent ice advances entrained the red clay, accounting for the distinct color of the Port Huron and Greatlakean red till.

Because density flows in glacial sediments are particularly diagnostic of a nearby ice margin in contact with a proglacial lake (Ashley and others, 1985; Gustavson, 1975 a,b), the interbedded flows in the channel are further evidence that the sands were deposited subaqueously near a glacial margin. We conclude that the ice front was the earliest Port Huron ice advance.

## Rhythmites

The cross bedded and rippled sands are overlain by distinct rhythmites consisting of red clay alternating with lighter-colored, coarser-grained sediment. The rhythmites overlap the channel sand at the channel margins and beyond rest directly on the upper boulder lag and Oak Creek till. The sand layers are generally rippled and, except near the base where they are normally graded and contain features consistent with density flow origins, contain internal laminae of silt and fine sand. The vertical change is evidence of an upward transition from slump-generated surge rhythmites to suspension-settling rhythmites (Ashley and others, 1985). The rhythmites are discussed in more detail in the following section.

## Red Till

The most important new observation is that the rhythmites grade northward almost imperceptibly into a massive red diamicton present at the top of the bluff. The continuous transition is proof that the red diamicton and the reddish rhythmites were deposited contemporaneously; they are facies equivalents.

Starting from the red diamicton (Fig. 4a), the transition to rhythmites occurs over a relatively short distance, scarcely more than 90 m. The transition begins where the diamicton's pebble and cobble con-

*Figure 4a.* Contact between the Ozaukee (upper) and Oak Creek (lower) till units. Contact is marked by approximately 8 cm of bedded sand (at shovel). See figure 3 for photo location.

*Figure 4b.* Cobbly Ozaukee diamicton. Note increase in pebble and cobble content from figure 4. See figure 3 for photo location.

tent increases abruptly (Fig. 4b). Traces of stratification are initially defined by clasts at the same elevation over horizontal distances of several meters, and then by concentrations of clasts and sand pods in elongate lens-shaped patterns. Over the following 30 m the deposit develops a definite, but discontinuous, stratification (Fig. 4c) with rudiments of rhythmites defined by a segregation into contorted layers of clay-rich and sand-rich sediment. The clay layers, in particular, contain evidence of flowage, including sharply erosional bases and rip up clasts of unconsolidated clay. Individual layers are folded, overturned, highly deformed, and are discontinuous. The clay units are interstratified with unsorted coarse, cobbly sand, variable in thickness, with distinct fining and sorting trends away from the diamicton and toward the rhythmites. Boulders and lens-shaped bodies of cobbles, interpreted as dropstone accumulations, are locally present within a finer-grained sand matrix. Pebbles and cobbles within the sand layers continue with decreasing frequency over another 30 m past the point where distinct stratification develops. Beyond that distance clasts and other evidence for ice-rafted debris are rare and rhythmites are fully developed (Fig. 4d).

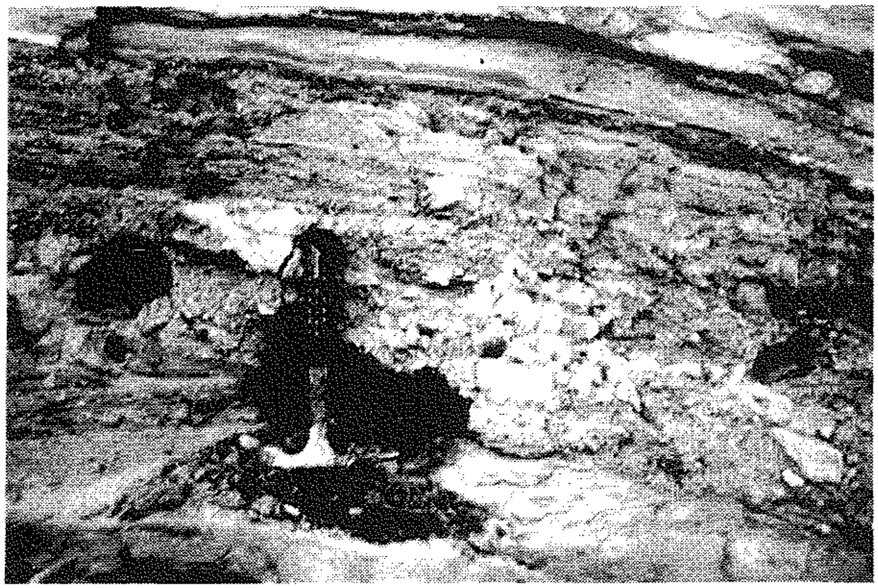

*Figure 4c.* Poorly stratified Ozaukee sediments. See figure 3 for photo location.

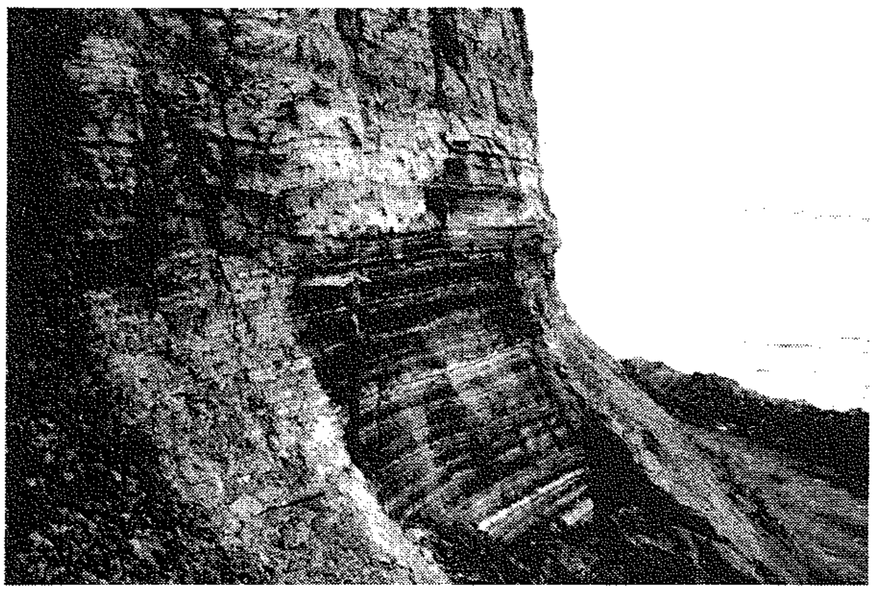

*Figure 4d.* Ozaukee rhythmites. See figure 3 for photo location.

The gradation between diamicton and rhythmites extends farther south in the lower part of the transition zone than above, where poorly bedded deposits in the lower meter are sharply overlain by well-bedded rhythmites. Because of the lateral fining trend in the rhythmite couplets, an extension of this contact can be traced southward over much of the exposure. The rhythmite thickness below this contact is variable, ranging from one to twelve centimeters. Above the contact the grain size in the sand couplets decreases, and the couplets thin to a consistent thickness of several centimeters, further evidence for an upward transition from surge to suspension rhythmites. The two fining trends, both upward, and laterally away from the transition zone, define a regressive sequence with rapid northward retreat of the glacial source.

The most obvious interpretation of the transition described above is of a glacial terminus in standing water. The transition apparently marks the southern limit of the initial Port Huron advance which deposited the first of the red till units in eastern Wisconsin. Further evidence supporting the terminal position is that there is no similar red diamicton in the bluffs south of this point (Mickelson and others, 1977; Rovey and Borucki, 1995), although an identical deposit is continuously present at the top of bluff exposures north of Milwaukee Harbor.

The transition between massive diamicton and rhythmites slopes generally to the south, implying that the transitional sediments were deposited as a series of sediment flows off the ice front, progressively building away from the ice and into Lake Chicago. However, the lack of discrete boundaries indicates a continuum beginning with subglacial melt-out and no resedimentation, progressing to slight remobilization, to extensive flowage, and ending with complete resedimentation and segregation into coarse and fine-grained rhythmite couplets.

## CORRELATION OF RED DIAMICTON

### Introduction

The upper red diamicton at the St. Francis site was first noticed by Lasca (verbal communication)in 1977, and he provisionally correlated it with the red till north of Milwaukee. However, subsequent slumping obscured the upper exposure; hence, Lasca (1983) later did not mention it. We propose that this uppermost diamicton indeed is the oldest of the red till units, now known as the Ozaukee Member of the Kewaunee Formation (Table 1).

### Red Till Controversy

The red tills of eastern Wisconsin, present along the margins of Lake Michigan and Green Bay, is of historical interest due to their long controversy. T.C. Chamberlin (1887, 1883) noted interbedded laminated sediment and the parallelism to shorelines and described them as entirely lacustrine deposits from a high-elevation precursor to Lake Michigan. He was also influenced by the relative lack of large erratics, and his descriptions show that, in his judgement, glacial sediment must meet certain textural criteria in order to be true till. His misdiagnosis was not corrected until Alden (1918) reported the presence of faceted and striated boulders, proving that most of the material is till.

The second major controversy involves stratigraphic subdivision. Thwaites (1943) established the name "Valders" for the red till at Valders Quarry, and later (Thwaites and Bertrand, 1957) applied the same name to all the red till in eastern Wisconsin, including that overlying the Two Creeks forest bed. Eventually, the term "Valderan" took on a time-stratigraphic meaning as well (Valderan Substage) for deposits stratigraphically above the Two Creeks (Twocreekan) forest bed (Frye and Willman, 1960; Frye and others, 1968). However, other authors (Evenson, 1973; Mickelson and Evenson, 1975) found evidence that the type-Valders is actually below the forest bed and, therefore, pre-Twocreekan (and pre-Valderan). The controversy is now largely resolved with the recognition of multiple red till units both older (Woodfordian) and younger (Greatlakean) than the forest bed (Mickelson and others, 1977; Acomb and others, 1982).

### Additional Evidence for Correlation

The diamicton composition is direct evidence that it is the Ozaukee. As shown in Table 1, lithologic properties are identical with undisputed Ozaukee till north of Milwaukee Harbor, and its properties are distinct from every other till unit exposed in the lake bluffs of southeast Wisconsin.

All previous workers, however, mapped the southern Ozaukee terminus at the mouth of the Milwaukee River at Milwaukee Harbor. This interpretation dates back to Alden (1918) who mapped the western margin of the red till (Ozaukee in this study area) roughly parallel to the Milwaukee River in Milwaukee and southern Ozaukee Counties (fig. 1). The implication is obvious; if the river occupies an ice-marginal position, then the Ozaukee could not extend southward beyond the river mouth. However, as shown in figure 1, the river does breach the moraine at several locations, and a series of morainal ridges are traceable from the St. Francis site northward to undisputed Ozaukee margins in northern Milwaukee and southern Ozaukee counties. The ridges south of Milwaukee Harbor are generally more subtle, but they are present at or just below elevations of proglacial Lake Chicago. Hence, they are likely to have been partially eroded and/or covered with lake or estuary sediments. Nevertheless, the ridges line up well with the Ozaukee morainal trend and are cored with a reddish-hued till (Need, 1983). Need also mapped these ridges as a moraine, although he interpreted them as the youngest recessional moraine of the Oak Creek Formation. He cautioned, however, that his assignment to the Oak Creek was based on visual inspection of soil boring samples without corroborating lab analysis. We unsuccessfully attempted to locate the samples in question for laboratory analysis. Without the samples we cannot eliminate the possibility that they are indeed Oak Creek till, but we think it unlikely for the following reason. Till with a slightly reddish hue is present within the Oak Creek Formation, but along the lake bluffs, at least, it always occurs at the base of the formation; it is the oldest or stratigraphically lowest Oak Creek till (Rovey and

*Table 1.* Clay mineralogy and textural parameters of glacial till exposed in the Lake Michigan bluffs, southeast Wisconsin. Subdivisions within the Oak Creek and New Berlin Formations are informal, based on Rovey and Borucki (1995). X denotes mean value, (S) standard deviation and [N] number of samples. The sand/silt boundary is .05 mm, the silt/clay boundary is 0.002 mm. See Rovey and Borucki (1994) and Borucki (1988) for methods of analysis.

| | Sand | | Silt | | Clay | | | Expandables | | Illite | | Kaolinite + Chlorite | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | X | (S) | X | (S) | X | (S) | [N] | X | (S) | X | (S) | X | (S) | [N] |
| **Ozaukee Member, Kewaunee Formation** | | | | | | | | | | | | | | |
| North of Milwaukee Harbor | 22 | (4.9) | 38 | (3.4) | 40 | (3.0) | [18] | 22 | (1.7) | 59 | (1.5) | 19 | (1.2) | [11] |
| St. Francis site | 24 | (6.2) | 34 | (3.5) | 42 | (3.2) | [8] | 21 | (2.9) | 60 | (4.1) | 19 | (2.9) | [5] |
| **Oak Creek Formation** | | | | | | | | | | | | | | |
| Oak Creek 5 | 19 | (3.8) | 51 | (2.1) | 29 | (3.7) | [11] | 14 | (.87) | 67 | (1.4) | 19 | (2.2) | [9] |
| Oak Creek 4 | 22 | (5.8) | 47 | (3.0) | 31 | (5.3) | [36] | 10 | (2.2) | 69 | (2.0) | 21 | (2.0) | [28] |
| Oak Creek 3 | 13 | (2.4) | 42 | (3.6) | 45 | (3.8) | [22] | 13 | (1.7) | 65 | (1.5) | 22 | (1.6) | [20] |
| Oak Creek 2 | 40 | (3.8) | 34 | (3.1) | 26 | (6.4) | [9] | 18 | (3.0) | 60 | (1.8) | 22 | (2.2) | [10] |
| Oak Creek 1 | 14 | (2.8) | 39 | (3.3) | 47 | (3.4) | [19] | 16 | (1.7) | 62 | (2.5) | 22 | (1.9) | [15] |
| **New Berlin Formation** | | | | | | | | | | | | | | |
| New Berlin 2 | 30 | (5.0) | 45 | (6.0) | 25 | (7.1) | [12] | 18 | (3.2) | 63 | (3.0) | 19 | (1.3) | [8] |
| New Berlin 1 | 57 | (5.4) | 30 | (2.3) | 13 | (4.3) | [5] | NOT ANALYZED | | | | | | |

Borucki, 1995). But, Need (1983) found two additional Oak Creek till units between the morainal ridges and the underlying New Berlin Formation. Therefore, the reddish till in the ridges cannot be part of the Oak Creek Formation unless two additional Oak Creek till units are present just a few kilometers inland that are not preserved anywhere along the bluffs.

## CONCLUSIONS

Based on the evidence so far available, we conclude that the Ozaukee moraine does not intersect the Lake Michigan shoreline at the mouth of the Milwaukee River as previous investigators maintained; rather, it continues south to the St. Francis Power Plant exposure. The Ozaukee till, however, is probably not present continuously behind (east of) the morainal front, because most of the area is slightly below 195 m in elevation, the Glenwood level of Lake Chicago, and would have been susceptible to removal by wave scour. The till is primarily preserved along the morainal ridges which themselves may be partially eroded or draped with later lake/estuary sediment.

Several unique features of the Ozaukee terminal ice front position are preserved at the St. Francis site. First, the Ozaukee till grades directly into rhythmite sediments. This proves that the ice front was advancing south into the standing water of Lake Chicago, and that the resulting transitional sediment is an ice-contact deposits. Such transitions and associated sedimentary features are frequently hypothesized and have been described conceptually (Powell, 1981; Powell and Molnia, 1989), but have seldom been conclusively recognized and described in the field. The most surprising aspect of the transition is its abruptness. The non-stratified till grades through a series of chaotic mass flow deposits into perfectly stratified rhythmites over a distance of approximately 90 m. Ice-rafted debris, likely to be an important sediment source near the glacial margin of most ice-contact lakes, is generally lacking beyond 30 m past the transition.

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# Variation in Cambrian Conglomerate Layers Exposed in Parfreys Glen, Wisconsin

*Mary E. McMillan[1,2], Timothy D. Vick[1], and Richard Shinn[1]*

## ABSTRACT

*Six measured stratigraphic sections and seismic refraction survey show the distribution and variation of Cambrian conglomerate units in Parfreys Glen and the configuration of the underlying Precambrian quartzite surface beneath the glen. The conglomerate units fall into four groups designated A, B, C, and D, in ascending order. Clast long axis measurements in units A and B range from 13 to 18 cm and units A and B have similar thicknesses and lateral continuity. Maximum clasts in units C and D have long axes ranging from 20 to 130 cm; these units show much greater variation in lateral continuity and thickness than do units A and B. The seismic refraction survey reveals that the Cambrian sediments in Parfreys Glen were deposited in a pre-existing valley which formed a cove in the Cambrian shoreline. The modern glen is off-center to the west in the ancient valley.*

## INTRODUCTION

Remarkable exposures of quartzite conglomerate interbedded with sandstone have been used to interpret the depositional environment and depositional mechanism for Cambrian sediment exposed in the Baraboo area. The sediment accumulated on the flank of the Baraboo Quartzite hills, the remains of a nearly east-west trending Precambrian syncline. Dott proposed that the Cambrian depositional environment was shallow marine, surrounding islands of resistant Precambrian quartzite (Dalziel and Dott, 1970; Dott, 1974; Dott and Byers, 1980). *Skolithos* are rare to absent in sandstones interbedded with the Parfreys Glen conglomerate, but their presence in comparable stratigraphic positions in other exposures in the region, as well as above the uppermost Parfreys Glen conglomerate, is strong evidence for shallow marine conditions. Further evidence for shallow marine conditions comes from the Cambrian Tunnel City Formation and Trempealeau Group, which contain marine invertebrate fossils, glauconite and algal stromatolites elsewhere in the Baraboo area (Dalziel and Dott, 1970).

Dott (1974, 1983); Dott and Dalziel, (1970); and Byers (1980) proposed a storm deposit origin for Parfreys Glen conglomerates and other similar deposits near Baraboo and the rounding of cobbles and boulders was attributed to storm waves and currents. However, the evidence supporting the storm deposit mechanism is based on the inferred shallow marine setting of the conglomerates, and the occurrence of coarse, rounded conglomerate in thin, sharply-bounded layers. The purpose of this paper is to describe the conglomerates in Parfreys Glen, including their lateral and vertical extent and variations, and to evaluate the storm deposit mechanism.

## METHODS

### Surface Methods—Conglomerates

Six sites for stratigraphic measurement and description were selected along the entire length of exposure (fig. 1). All sections were located on a pace-and-compass map of the trail and stream in Parfreys Glen. Layers were correlated from one section to another by walking contacts and by correlating several stratigraphic levels from one side of the stream to the other (fig. 2).

Within each conglomerate layer, several groups of cobbles were measured and a most-common size range and maximum clast size were recorded for each. Where exposure and accessibility permitted, guadrates for clast long-axis measurement were selected randomly. The observer looked away from the glen wall, stepped two paces along exposure (away from a measured section, parallel to the cliff

[1]Carleton College, Northfield, MN 55057

[2]Presently at BSCS, Pikes Peak Research Park, 5415 Mark Dabling, Colorado Springs, CO 80918

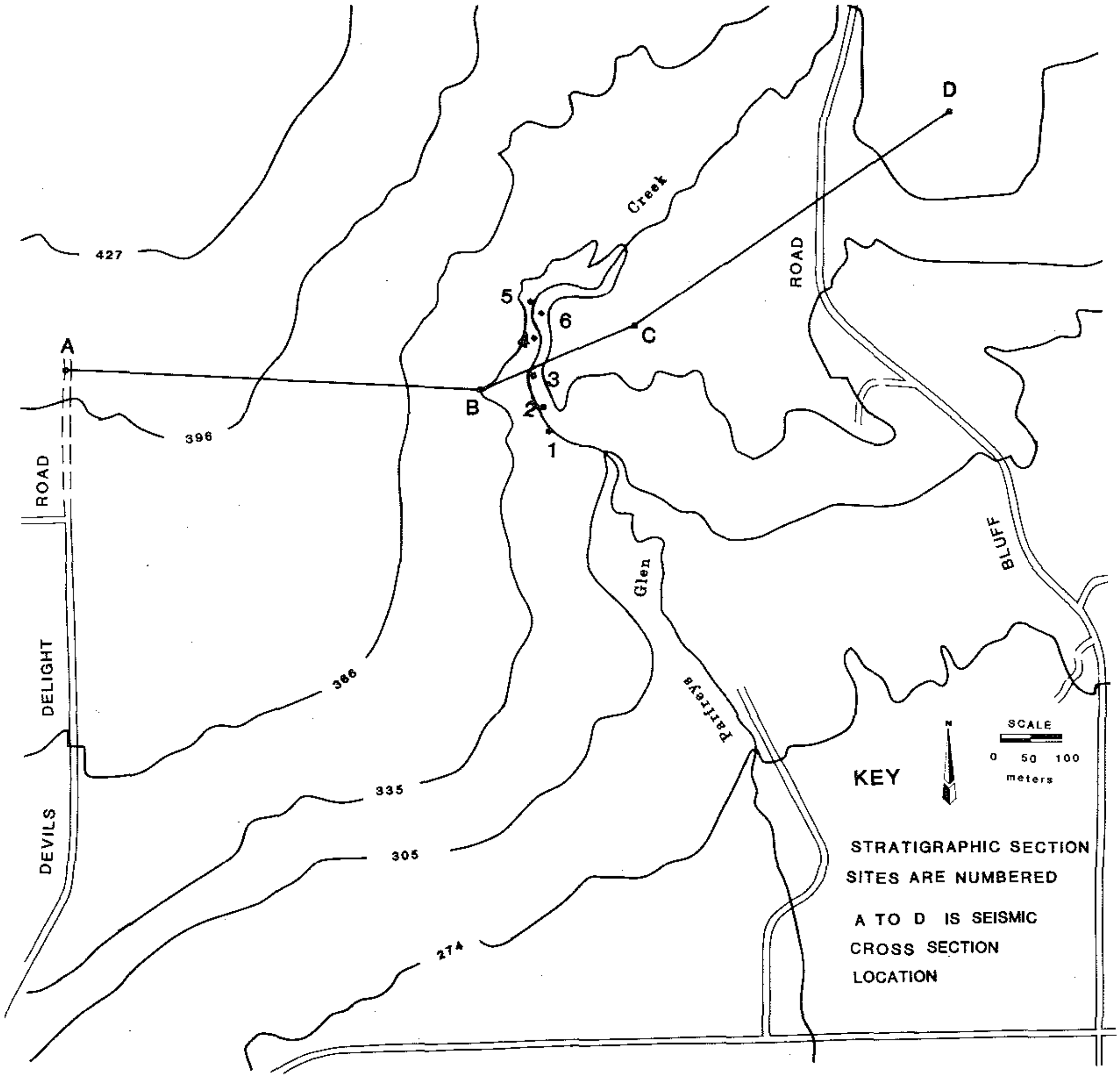


***Figure 1.*** Location map of stratigraphic sections and seismic cross section in Parfreys Glen, Wisconsin. Surface elevations in meters above sea level, contour interval is 30 meters.

face) and extended her arm to mark the right edge of a 40 x 42 centimeter rectangle. Mean clast size was calculated for the lower two units by measuring all clast long axes which lay within the rectangle. Exposures of upper units did not permit the random sampling technique, because the majority were on inaccessible cliff faces.

Other properties, such as angularity of clasts, orientation, type of matrix, and matrix or clast support were described as the sections were measured.

## Subsurface Methods — Seismic Refraction

The relief and position of the contact between Precambrian Baraboo Quartzite and Cambrian sedimentary rock was mapped using shallow seismic refraction. Contact elevation was determined by use of a Bison 1570C signal enhancement seismograph with one geophone. Seismic waves were generated with a sledge hammer. Seismic profiles were done in 21 locations in and near the glen (fig. 3). Seventeen profiles were reversed; difficult working conditions prevented reversal at four sites.

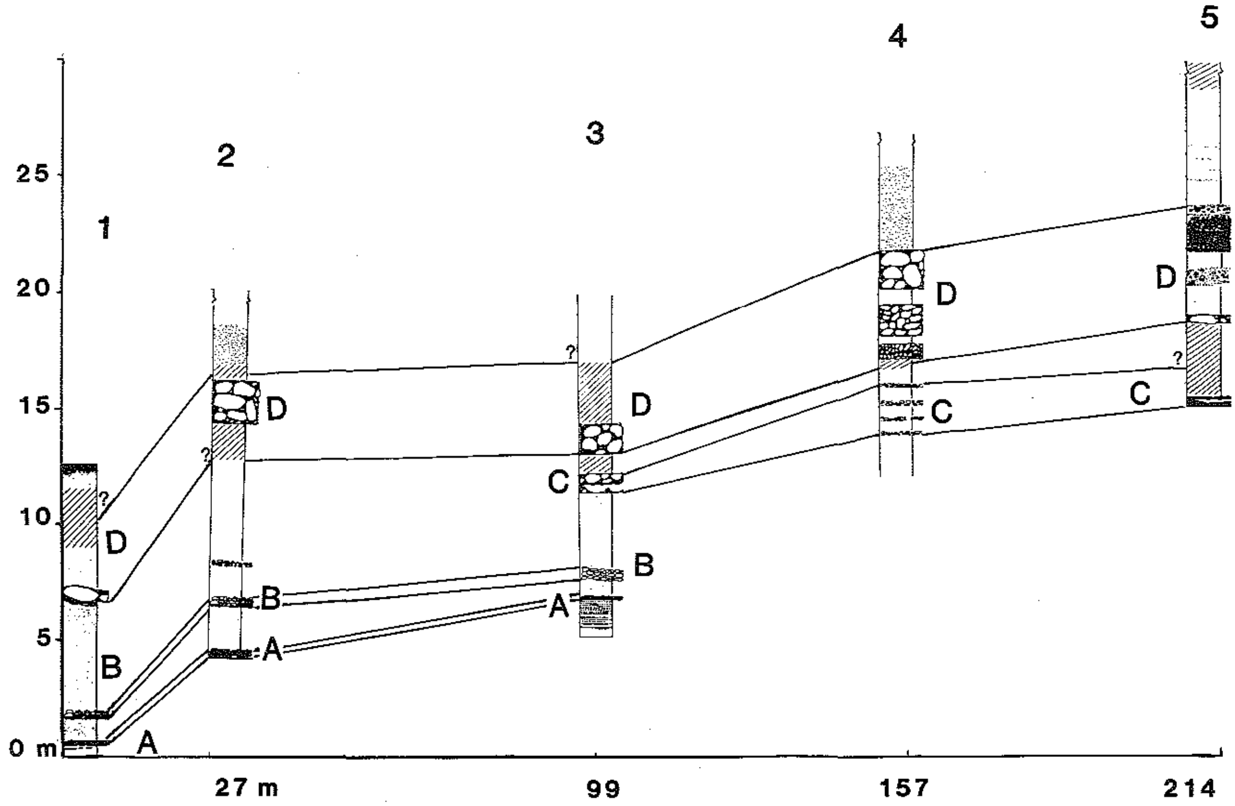


*Figure 2.* Stratigraphic sections 1 to 5 in Parfreys Glen, shown from south to north (left to right). (Also see fig. 1).

Two additional profiles, for the purpose of determining the seismic velocity of the quartzite and sandstone rather than to find depths, were located where there was no doubt as to the underlying material. A profile to determine the seismic velocity of the quartzite was done on an exposure near the north end of Devils Lake, and another to find the seismic velocity of the sandstone was located at the top of the bluff on the east side of Parfreys Glen. The profile lines were 50 to 100 meters long, with data points spaced at five or ten meter intervals.

In many cases the number of hammer hits required for a data point was fairly close to one tenth the number of meters between the impact point and the geophone. For example, four to six hammer hits might be required at a point 40 meters from the geohone. Since the seismograph is signal-enhancing, the seismic wave forms from the numerous impacts at each location are electronically accumulated and displayed as a single wave form. In most cases the data were not difficult to take. There were distinct differences in the seismic characteristics of the units and the stratigraphically lower units had successively higher seismic velocities so there were no blind zones. Seismic refraction, however, will reveal the depth only to layer boundaries which are distinct in terms of seismic qualities, but may or may not coincide with stratigraphic boundaries. For example, if the Baraboo Quartzite in a particular location has a weathered zone on its upper surface, the weathered zone may have seismic qualities similar to the overlying sandstone and thus be indistinguishable from it. The seismic discontinuity would be deeper than the actual depth of the quartzite-sandstone interface. We observed little chemical weathering of the quartzite in the study area, so any such error is minor.

The primary difficulties encountered in Parfreys Glen involved working conditions. In some places slopes were too steep to work on, and running water prevented reversing one profile in a streambed. A reasonable estimate of the accuracy of the soundings in Parfreys Glen would be plus or minus ten to fifteen percent. The seismic travel-time graphs resulting from the field work were interpreted using a FORTRAN computer program by Mooney (1980) and a short BASIC program by Vick (unpublished) based on formulae from the same volume.

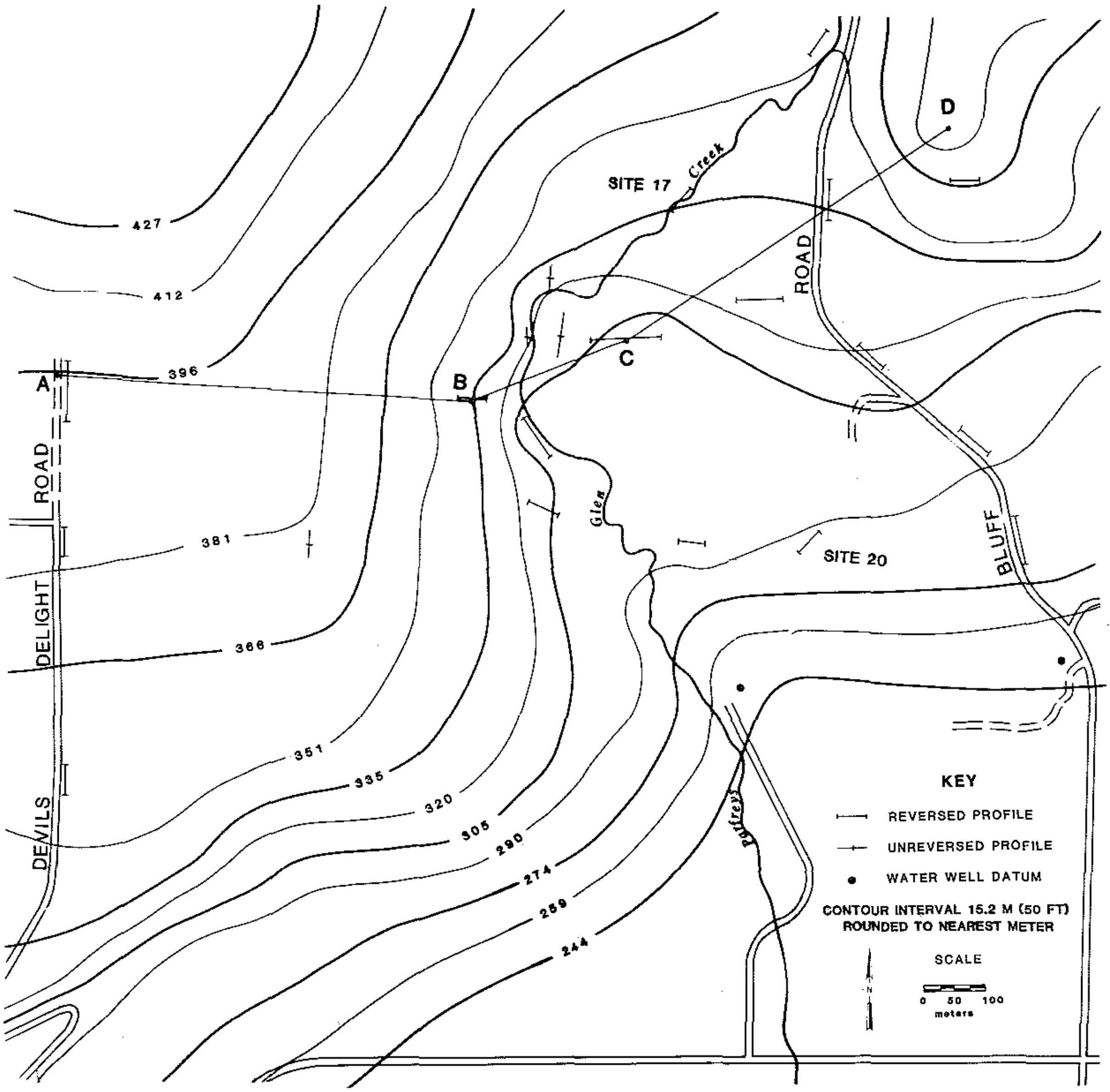


*Figure 3.* Precambrian Baraboo Quartzite surface map in Parfreys Glen area as determined by seismic refraction survey.

## SEDIMENTOLOGY

Four conglomerate units (labelled A through D) can be distinguished in Parfreys Glen. The lower units, A and B, consist of a single layer each, while the two upper units, C and D, consist of multiple and discontinuous conglomerate layers.

### Persistent Units A and B

Units A and B consist of pebbles and cobbles in a quartzarenite matrix. They share such similarities as sorting, layer thickness, matrix type and continuity. Most unit A clasts range from 4-7 cm (table 1). The clasts in Unit B are only slightly larger than those in unit A (table 1). It is possible to trace units A and B in the field over the entire range of exposure.

Clasts in units A and B range in shape from equant to disc-shaped and clasts in unit A are notably more angular than those in unit B. In general, the clasts support each other and are oriented with the long axes nearly horizontal (parallel to bedding planes). Dott and Byers (1980) found that most clasts in the glen dipped less than 20 degrees. Their data indicate a weak maximum of a-b planes dipping

*Table 1.* Clast size ranges, and conglomerate matrix descriptions

| UNIT | SECTION | MAXIMUM CLAST SIZE (in centimeters) | COMMON SIZE RANGE (in centimeters) | UNIT THICKNESS* (in meters) | MATRIX |
|---|---|---|---|---|---|
| A | 1 | 13 | 4-7 | 0.25 | sandy |
| A | 2 | 14 | 5-8 | 0.20 | sandy |
| A | 3 | 18 | 5-7 | 0.25 | sandy |
| B | 1 | 15 | 8 | 0.40 | sandy |
| B | 2 | 16 | 5-12 | 0.50 | sandy |
| B | 3 | 18 | 3-8 | 0.55 | sandy |
| C (?) | 3 | 20 | – | 0.90 | pebbly |
| C-1—C-4 | 4 | 58 | 8-18 | 2.30 | pebbly |
| C-1 | 5 | 25 | – | 0.45 | pebbly |
| C-1—C-5 | 6 | 30 | 3-8 | 3.20 | pebbly |
| D | 1 | 100 | 40 | 0.55 (exposed) | pebble & sand |
| D | 2 | 100 | 30 | 2.00 | pebbly |
| D | 3 | 30 | – | 2.20 | pebbly |
| D-1—D-3 | 4 | 100 | 30 | 4.80 | pebbly |
| D-1—D-3 | 5 | 130 | 5-7: 30 | 4.90 | pebble & sand |
| D-1—D-11 | 6 | 15 | 1-7 | 8.80 | pebble & sand |

**Unit thicknesses for C and D include sandstone lenses; A and B do not.*

| UNIT | MEAN CLAST SIZE (LONG AXIS)** | NUMBER OF MEASUREMENTS | MINIMUM; MAXIMUM | STANDARD DEVIATION | VARIANCE |
|---|---|---|---|---|---|
| A | 4.7 (cm) | 48 | 1.5; 13.0 | +2.43 | 5.92 |
| B | 5.8 | 49 | 1.5; 23.0 | +3.88 | 15.03 |

***Exposures and accessibility permitted random sampling technique for Units A & B only. Quadrate height was determined by thickness of conglomerate unit; quadrate width was the multiple needed to bring the area to 1680 square cm. e.g. (40x42 cm).*

toward the north.

The two lower conglomerate units show lateral continuity. For example, there is a layer of pebbles, one pebble thick, that appears below unit A at both sections 1 and 2. Thicknesses of units A and B remain uniform at approximately 0.20 m and 0.50 m, respectively. Unit B consists of two discrete cobble layers at section 2 and one cobble layer elsewhere. Upstream from section 3 (figs. 1,2), units A and B are not exposed. The two upperunits, C and D, show considerably more variation over short distances (a few meters) than do units A and B.

## Laterally Discontinuous Units C and D

With the exception of one layer in unit D, these layers cannot be traced laterally. Because of their discontinuity, they were correlated by stratigraphic position only. Other differences from units A and B exist. These upper units vary widely in sorting, layer thickness and matrix constituents.

Units C and D consist of layers of pebbles, cobbles and boulders. Pebble matrices packed between boulders are interbedded with discontinuous sandstone lenses and pebble layers. In both units C and D, the clasts and pebble matrix consist solely of Baraboo Quartzite, indicating a single source. Many of the clasts are oriented with the long axis parallel to bedding (nearly horizontal), but in a few places, long axes of clasts dip northeast. Previous work has also shown westward dipping clast orientations in some beds (Dott and Byers, 1980). As in unit B, most of the clasts in units C and D are rounded. Dott (1970) found rounded clasts up to 1.3 m in diameter, although the largest clasts in comparable conglomer-

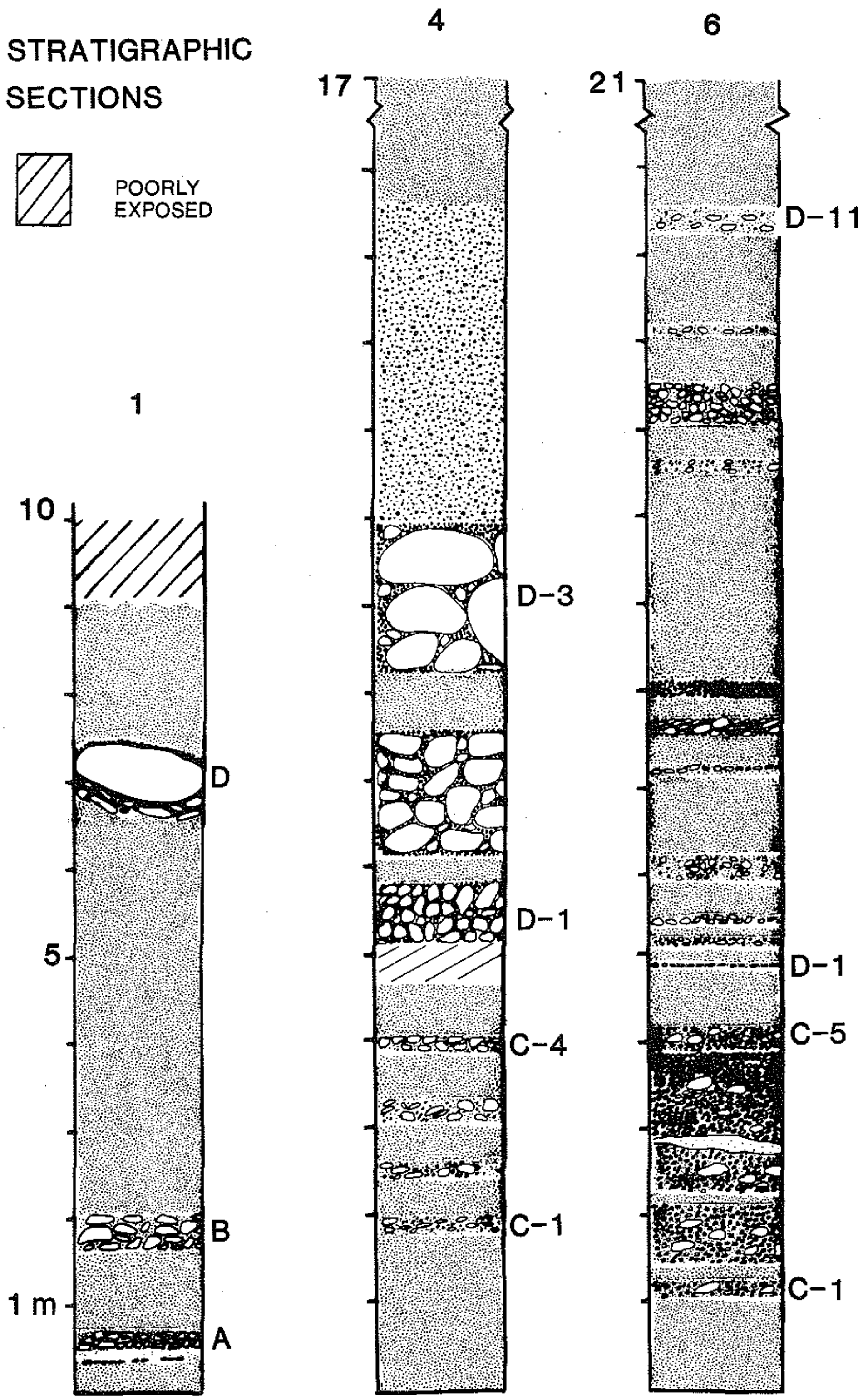


*Figure 4.* Stratigraphic sections 1, 4 and 6 (fig. 1) show some of the extreme size variations in clasts within Parfreys Glen.

ates on the flanks of the Baraboo hills are up to 3.0 m in diameter.

Unit C is the most enigmatic of the units exposed in Parfreys Glen, because it is not exposed in some areas where units A, B and D are present and it changes character over short distances. At the stratigraphic interval near section 3, unit C is only one cobble in thickness. Over a distance of 86 meters, unit C increases in thickness to 3.2 m at section 6, where it consists of as many as 5 layers of gravel and cobbles separated by sandstones, with maximum clast size of 15 cm. Twenty-four meters away at section 4, maximum clast size is 58 cm and the unit contains four poorly-sorted cobble layers with a pebbly matrix separated by sand lenses (figs. 2,4).

One layer in unit D is distinctive in all sections measured in Parfreys Glen. The bed contains an unsorted mixture of pebbles, cobbles, and boulders up to 1.5 m in diameter. The variability in clast size is most apparent at section 1, where the maximum clast is 1 m in diameter, whereas most clasts have long axis measurements of 40 cm or less.

Unit D is exposed at many points throughout the glen, but its strata change in thickness and character over the length of exposure. At section 1, the distinctive poorly-sorted layer is 55 cm thick, and at section 2, at least two m thick. In the lower part of the glen (sections 1, 2 and 3), this bed is the only exposure of unit D. Unit D contains numerous conglomerates at least four m thick separated by sandstones at sections 4, 5 and 6 (figs. 1, 2 and 4). These conglomerates extend only a short distance along the exposures in the glen, and the bed exposed in the downstream sections (1, 2 and 3) is the only continuous one. The largest and most numerous grouping of exposed boulders is at section 4 (fig. 2).

### Summary of Sedimentological Features

The four conglomerate units recognized in Parfreys Glen differ in their main features (table 1). The upper units C and D show wide variations in thickness, number of individual clast layers, and average and maximum size of clasts over the length of exposure. Units C and D are also characterized by pebbly matrices, larger average clast sizes, more variable sizes within layers and some non-horizontal clast orientations. The lower units (A and B) are much less variable. Units A and B have sandy matrices, smaller average clast sizes, and generally horizontal clast orientations.

### Paleocurrents

Information on paleocurrents indicated by trough and tabular cross-bedding orientations in Parfreys Glen was collected by Dott and Byers (1980). They discovered a strong unimodal dip toward the east and southeast with a smaller number of measurements dipping to the south-southwest. Judging from the diagram in Dott and Byers (1980), these measurements were taken from sandstones within units here designated as A, B, and C. In the sandstone above unit D, axes of trough cross-strata dip east-northeast and west-southwest, whereas tabular cross-strata in the same sandstone show modal dips to the east-southeast, south east and north-northeast (Dott and Byers, 1980). These data may mean that the transport directions changed radically between the deposition of the conglomerate layers and the overlying sandstone.

Limited observations of pebble orientations within the conglomerate layers tend to confirm the findings of Dott and Byers (1980) who concluded that most pebbles were oriented with the ab plane nearly horizontal. However, they found that clasts in unit D dipped gently west. Imbrication in the few exposures examined in this study consists of north-dipping clasts indicating flow to the south.

## SEISMIC REFRACTION RESULTS

A seismic refraction survey determined the three dimensional geometry of the deposits and provided enough information to refine paleogeographic interpretation at Parfreys Glen. The materials present in Parfreys Glen fell into three seismic units: 1) an unconsolidated surficial layer up to a few meters thick, including weathered sandstone, alluvium, colluvium and glacial materials. This layer conducted seismic waves at velocities from 309 to 744 meters per second, with an average of 416 meters per second with a standard deviation of 102 meters per second (table 2). In only one case was the velocity higher than 600 meters per second. 2) An intermediate layer of sandstone was present in about half of the traverses, mostly in the central part of the area between the creek and Bluff Road (fig. 3) where it was up to 122 feet (37.2 meters) thick. It conducted seismic waves at velocities form 634 to 1866 meters per second, with an average of 1156 meters per second with a standard deviation of 381 meters per second. 3) The Baraboo Quartzite conducted seismic waves at velocities of or greater than 2185 meters per

*Table 2.* Seismic refraction: velocity measurements and depths to sandstone and quartzite (Depths to contacts in meters, velocities in meters per second)

| SITE | SURFICIAL VELOCITY | DEPTH OF CONTACT | SANDSTONE VELOCITY | DEPTH OF CONTACT | QUARTZITE VELOCITY |
|---|---|---|---|---|---|
| 1 | 383 | 2.0 | 634 | 7.6 | 3005 |
| 2 | 443 | – | – | 10.2 | 2552 |
| 3 | – | – | 1111 | 1.8 | 3750 |
| 4* | 360 | 3.3 | 1192 | (Lower contact not found) | |
| 5 | 744 | 4.8 | 1866 | 21.8 | 4234 |
| 6 | 366 | – | – | 3.7 | 3664 |
| 7* | (See footnote) | | | | |
| 8 | 462 | – | – | 12.8 | 2636 |
| 9 | 377 | – | – | 6.1 | 2185 |
| 10 | 371 | 4.4 | 888 | 16.8 | 4872 |
| 11 | 395 | 3.0 | 794 | 17.7 | 2534 |
| 12 | 406 | – | – | 6.1 | 3000 |
| 13 | 333 | – | – | 3.4 | 4250 |
| 14 | 359 | 2.9 | 951 | 12.8 | 3574 |
| 15 | 577 | – | – | 3.7 | 4245 |
| 16 | – | – | 1381 | 4.9 | 3737 |
| 17 | 385 | 2.1 | 1857 | 7.1 | 4000 |
| 18 | 376 | 6.1 | 783 | 37.2 | 3966 |
| 19 | 335 | 3.5 | 1138 | 10.5 | 4805 |
| 20 | 370 | 3.8 | 1094 | 15.2 | 3548 |
| 21 | 433 | 6.1 | 1342 | 22.7 | 2987 |
| 22 | 509 | – | – | 5.3 | 5956 |
| 23 | 309 | – | – | 2.4 | 2334 |

**Seismic sounding done only to find a typical velocity for a rock unit (sounding #4: sandstone in Parfreys Glen; sounding #7: quartzite near Elephant Rocks in Devils Lake State Park).*

second with an average of 3641 meters per second with a standard deviation of 964 meters per second. There was no overlap in velocities between the sandstone and the quartzite and nowhere in the study area did the quartzite outcrop at the surface.

In about half of the seismic profile locations the interface between the quartzite and the overlying material was smooth and clearly visible on the travel-time graphs, as in figure 5. In other cases the boundary was topographically irregular or at a considerable angle to the modern ground surface, making the graphs more difficult to interpret, as in figure 6. In no case were the results impossible to use. The graphs which show a break in slope for the quartzite indicate irregularities in the floor of the ancient valley. Examples of irregularities include the ledges and knobs like those found on the hills in Devils Lake State Park and the sand-filled joint openings and joint slopes in La Rue Quarry.

The quartzite surface represents a wide, gentle valley or cove, (fig. 3). The valley measures at least 0.8 km in length from north to south, surrounded by steeper slopes on the southwest and northeast and ranges in width from 0.1 to 0.7 km. The valley in the quartzite surface trends southwest in the northern part of the area and southeast south of section A-D (fig. 3). Although this valley trends parallel to the trend of the modern stream in Parfreys Glen, the difference between the SSE direction of the modern stream and the SE trend of the older valley is important in explaining the SE trend of paleocurrents recorded in the sediments.

The ancient valley under Parfreys Glen is generally broad and flat-floored. Its average gradient is about nine percent, although there is a gentler-sloping area of about four percent near the lower part of the glen. The steepest-sloping part of the ancient valley is between the lower part of Parfreys Glen and the abandoned Parfreys Glen Park water well (NW 1/4, SW 1/4, Section 23, T 11 N, R 7 E) where the slope reached 15 percent.

The modern stream is displaced to the west relative to the axis of the ancient valley or cove in the quartzite surface (fig. 7). As a result, only sediments deposited in the western part of the old valley can be studied in detail in the walls of Parfreys Glen. Section A-D (fig. 7) shows that the stream has cut through most of the sediments in the western part of the sediment package. On the sides of the ancient valley, Phanerozoic sediments, including Cambrian sandstone and conglomerates and Pleistocene sediments, are less than 15 m thick. The axis part of the ancient valley, east of Parfreys Glen, is covered by a sediment package which is twice as thick as on the valley sides. Paleozoic sandstone and drift reaches a thickness of at least 37.2 m at point C (figs. 1, 3, 7). Evidence from wells in the vicinity of Bluff Road and at the mouth of Parfreys Glen indicates that conglomerate layers are found throughout the west-to-east extent of the Paleozoic sediment package and are not confined solely to the western flank.

South and east of the study area the Phanerozoic sedimentary blanket again thickens as the quartzite surface dips off into the uncharted depths of the subsurface. Near the Parfreys Glen Park water well (location given above), well log data shows that the quartzite is more than 30 meters below the surface. Driller's reports on water wells at Devils Head Lodge on the east side of the ancient valley indicate that near the barn at the base of the South Range on Bluff Road (NE 1/4, SW 1/4, Section 23) the quartzite is 58 m below the present topographic surface. It drops off to more than 80 m below the surface at an irrigation well in the SW 1/4 of Section 23 (east of Bluff Road, fig. 3). The ground surface south and east of the study area is nearly level.

## INTERPRETATION AND DISCUSSION

The results of the seismic survey indicate that some of the boulders exposed in the glen may actually be quite close to their source area. They may have become rounded within a short transport distance, since the glen runs along the base of a high quartzite slope. During the Cambrian this slope was the west wall of a nearly kilometer-wide valley. The height of the hillside west of the glen is more than 150 meters from the floor of the glen. The hill must have been at least as high and steep when the sandstone was deposited as it is now. It is conceivable

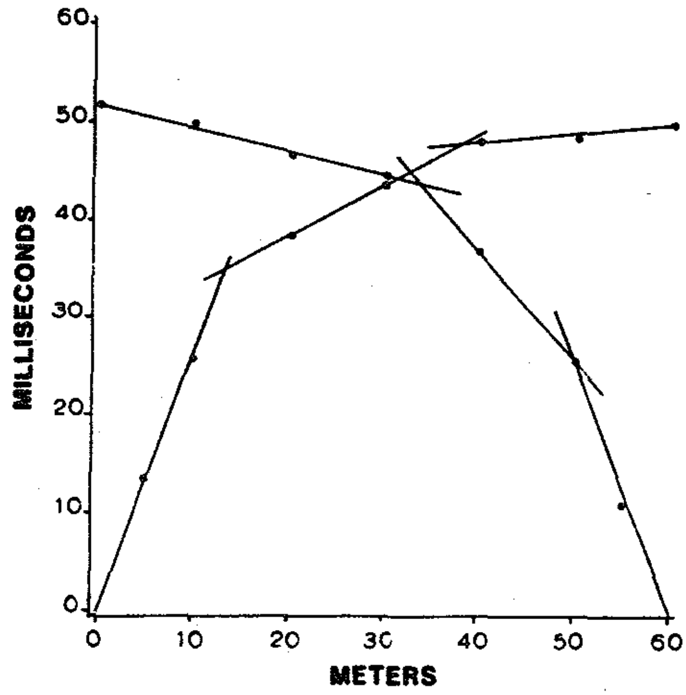


*Figure 5.* Site 20. Example of a travel-time graph displaying a close fit of points on the line segments. A change in slope indicates that wave speed changed, because the wave entered a different material (e.g. from sandstone to Quartzite).

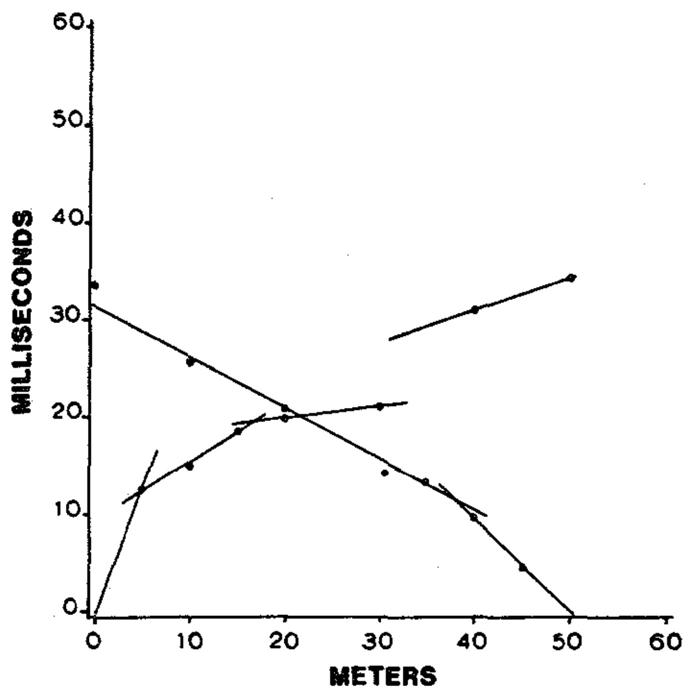


*Figure 6.* Site 17. Example of a seismic travel-time graph which was difficult to interpret. There is probably a ledge in the quartzite surface here; see Figure 3 for site location.

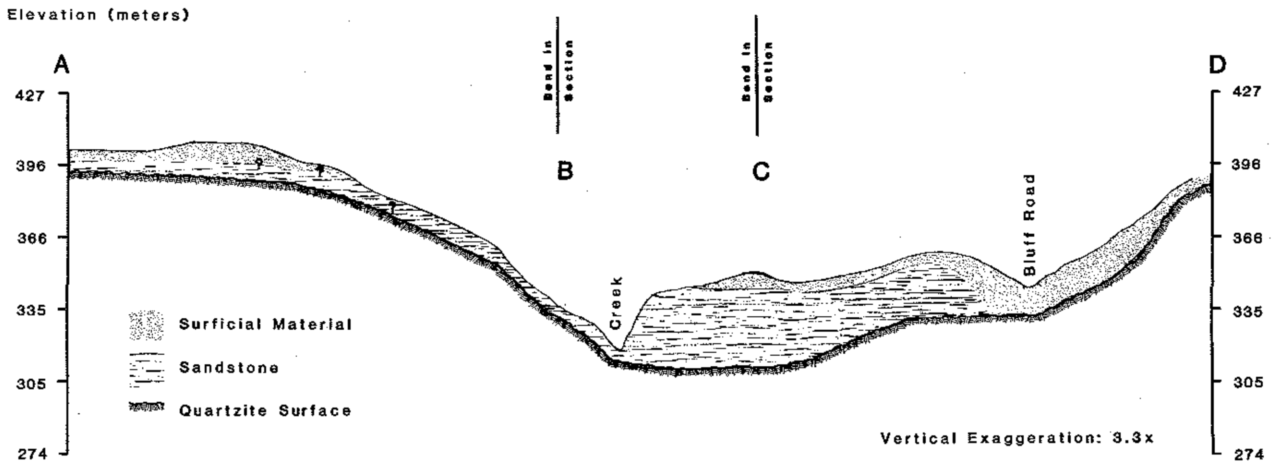


*Figure 7.* Seismic cross section A-B-C-D, showing Precambrian Baraboo Quartzite surface, upper Cambrian sandstone and conglomerate, and surficial material.

that the boulders in the conglomerates of Parfreys Glen could have weathered from topographic irregularities on the hills above the glen and washed down to be deposited with little lateral transport. The proximity of the glen to the hillside may also help explain the apparent eastward aspect to the dip of the conglomerate layers as sketched by Dott and Byers (1980). The Cambrian sediments may be draped into the Precambrian valley, dipping toward the valley center along the margins but dipping parallel to the valley axis in the center.

The vertical changes in the conglomerate layers exposed in Parfreys Glen indicate a coarsening upward sequence truncated and topped by a fine sandstone above the uppermost conglomerate. In individual sections, such as 1 and 2, where units A, B, and D are exposed, unit D is thicker, has larger clasts and a pebbly matrix. Unit B has a larger maximum and average clast size than unit A (fig. 2). The general upward increase in conglomerate layer thickness, proportion of pebbly layers to sand layers, and pebbly matrix indicates either increased proximity to the source of the quartzite clasts as might be caused by a prograding shoreline (Davis, 1983), or increased strength of the transporting and depositional mechanism (such as stronger storms or more extreme river floods). Furthermore, this coarsening upward developed despite the filling of the pre-existing valley by at least 25 m of sediment by the time unit D was deposited.

The lateral increase in clast-sorting from north to south (observed within individual units) suggests that most of the sediment moved down the axis of the pre-existing valley. The paleocurrent measurements recorded by Dott and Byers (1980) in the stratigraphic intervals of units A, B, and C also support the general pattern of sediment movement down the axis of the pre-existing valley with smaller amounts of sediment moving down the valley walls toward the axis.

Sedimentologic features of conglomerate units A and B support the hypothesis that these are storm deposits. The layers are uniformly thick over the extent of their exposure in the glen and contain uniformly-sized clasts. Both layers are laterally continuous. The horizontal orientations of the clasts in this unit is a feature that can be produced by storm surges moving away from cliffs and shorelines onto the nearshore (Kreisa, 1981). These layers seem to be widely distributed, uniform records of single storm events with a matrix that has filtered between the clasts after deposition.

The major difference between units A and B is the angularity of the clasts in unit A. The angular clasts probably were locally derived and did not undergo much transport before deposition. The more rounded clasts in unit B, however, indicate abrasion during transport. The disc-shapes common among these clasts suggest that they were worked on a beach prior to offshore deposition during a storm (Dobkins and Folk, 1970).

Sedimentological characteristics of units C and D are less consistent with a marine storm wave mechanism. Storm waves generally have the capacity to sort and winnow clasts of different sizes, yet the conglomerate layers in unit D are unsorted and many retain this unsorted character throughout the exposure (see unit D at section 1, fig. 2). Furthermore, the discontinuous layers within units C and D, many of which extend only a short distance down the glen, are inconsistent with a storm wave mechanism, which could be expected to spread cobbles over a wider area (Clifton, 1973).

Another problem with the storm deposit hypothesis is the paleogeographic environment. For much of the time represented by these deposits, there was a cove at this site which may have protected the area from storm waves generated further off-shore by refraction and attenuation of waves around the headlands (Leeder, 1982). Many (but not all) waves attenuate while passing over shallow shelves (Ritter, 1986).

An alternative is that of a stream moving into the shallow marine depositional basin from the north down the axis of the ancient basin. The conglomerates exposed in the Narrows section of Parfreys Glen (near sections 4 and 5, fig. 2) resemble deposits found in proximal braided stream environments (Miall, 1977). The unsorted nature of the conglomerate layers in units C and D, and the lateral variations in thickness and clast size over short distances are characteristics of proximal stream deposits. Dott and Byers (1980) have suggested that stream-deposited clasts would not be as rounded as those in Parfreys Glen. The rounding of the clasts could be explained by abrasion as well as transport in a stream (Schumm and Stevens, 1973).

Many of the cobble layers in units C and D have a pebbly matrix also composed of Baraboo Quartzite. Since clasts and pebble matrix have a uniform source, there is no problem explaining the discontinuous sandstone layers which may contain far-travelled sand grains. All quartzite clasts and pebbles may have come down the paleo-valley axis. As quartzite clasts and pebbles were transported down the valley axis, previously deposited sand layers may have been reworked as streams in flood carried cobbles onto the sands.

## CONCLUSIONS

This study of the Cambrian conglomerate layers in Parfreys Glen and the underlying Baraboo quartzite shows that Cambrian sedimentary rock is at least 37.2 m thick in the thickest area and no more than 15 m beneath Parfreys Glen. There are important vertical and lateral variations within and between conglomerate layers exposed in Parfreys Glen. The lower units, A and B, and specific conglomerate layers within unit C have sedimentological features which indicate a storm deposit origin. Parts of unit C and most of unit D, however, may have been deposited by streams in flood that rounded the clasts by abrasion and transport and deposited them in a shoreline embayment.

The presence of larger clasts in upper layers indicates that larger clasts were transported into the cove at the time represented by unit D. Either the shoreline had prograded into the cove (if these are storm deposits) or the water depth decreased, allowing streams to transport material further into the basin.

The deposits in Parfreys Glen have been cited as examples of episodic sedimentation by Dott (1983). This study indicates that two kinds of episodic mechanisms (marine storms and floods) acted in the Cambrian cove at Parfreys Glen.

## ACKNOWLEDGMENTS

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Wisconsin Geological and Natural History Survey, various years, water well logs.

# Gravity Modeling in Western Vilas County, Northern Wisconsin

*Barbara A. Eckstein*[1, 2]

## ABSTRACT

*The geology of Vilas county consists of Precambrian bedrock covered by thick glacial till. Outcrop and drill-hole data are rare; hence the geology must be inferred from surrounding areas and from geophysical surveys. Maps by the Wisconsin Geological and Natural History Survey and the Minnesota Geological Survey in 1982 have the same general features but the details do not match. Using the Bouguer gravity of western Vilas county, three subsurface geology profiles were derived based on the two maps, and differences in interpretations were discovered. The Wisconsin Survey map shows a much broader dome of Early or Middle Precambrian gneiss, north of the Niagara fault, than actually exists. The Minnesota Survey map places the dome in the metavolcanic region. Neither map contains a structure that would account for the high gravity to the southwest. Additional information about the bedrock is needed before the geology of Vilas county is completely understood.*

## INTRODUCTION

Gravity measurements from western Vilas county in north-central Wisconsin, near the Upper Michigan border, were used to model the subsurface structure in this region (fig. 1). Bedrock exposures are few, averaging about one small outcrop per township (LaBerge and Mudrey, 1979), so that geological field mapping methods cannot be applied. The geology must be inferred by using geophysical surveys to extrapolate from outcrops in surrounding areas and to interpolate between drill-holes. A gravity investigation can locate subsurface geological features and perhaps determine their depths and dimensions.

The field area was surveyed in the summer of 1982 with a LaCoste & Romberg Model G gravity meter. The gravity stations were chosen from 7.5- and 15-minute topographic maps at sites where spot and spirit level elevations were given. Additional data were provided by C.P. Ervin of Northern Illinois University and M.G. Mudrey, Jr. of the Wisconsin Geological and Natural History Survey. There are a total of 834 gravity stations, some of which were combined since their locations coincided, for a net of 730 observation sites. These measurements cover the western part of Vilas county. This area is bounded by latitude 46°15' N on the north, by latitude 45°48' N on the south, by longitude 89°18' W on the east, and by longitude 90°0' W on the west, and covers approximately 2,700 square km.

The Precambrian bedrock is overlain by glacial till (Attig, 1985). The topography is almost entirely the result of glaciation, although bedrock controls some of the larger features, such as kettle chains and belts of outwash and ground moraine. The average elevation is roughly 490 m above sea level. Local relief greater than 15 m is rare, but it can reach 45 m near terminal moraines and drumlins.

## PREVIOUS GEOPHYSICAL WORK

Numerous gravity and seismic studies have been undertaken in the Great Lakes region; however, most of these studies have been primarily concerned with distinguishing the Late Precambrian sequence from Early and Middle Precambrian rock, usually considered as one lithologic unit (Thiel, 1955; Bacon, 1966; Mooney and others, 1970; Cohen and Meyer, 1966; Ocola and Meyer, 1973; White, 1966). The regional seismic surveys indicate that the crustal thickness in northern Wisconsin is about 36 km, creating an average regional Bouguer anomaly of -40 mGal from an isostatically compensated crustal column (Cohen and Meyer, 1966).

Sternberg (1977) discovered an interesting feature of the region, the Flambeau anomaly, from resistivity data. This 20-km wide band of high conductivity within the earth's crust extends to the east from about latitude 46° N, longitude 91° 20' W for at least 170 km (Daneshvar, 1977; Sternberg and

[1] Department of Geology and Geophysics, University of Wisconsin-Madison, WI 53706

[2] Presently at Areté Associates, P.O. Box 6024, Sherman Oaks, CA 91413

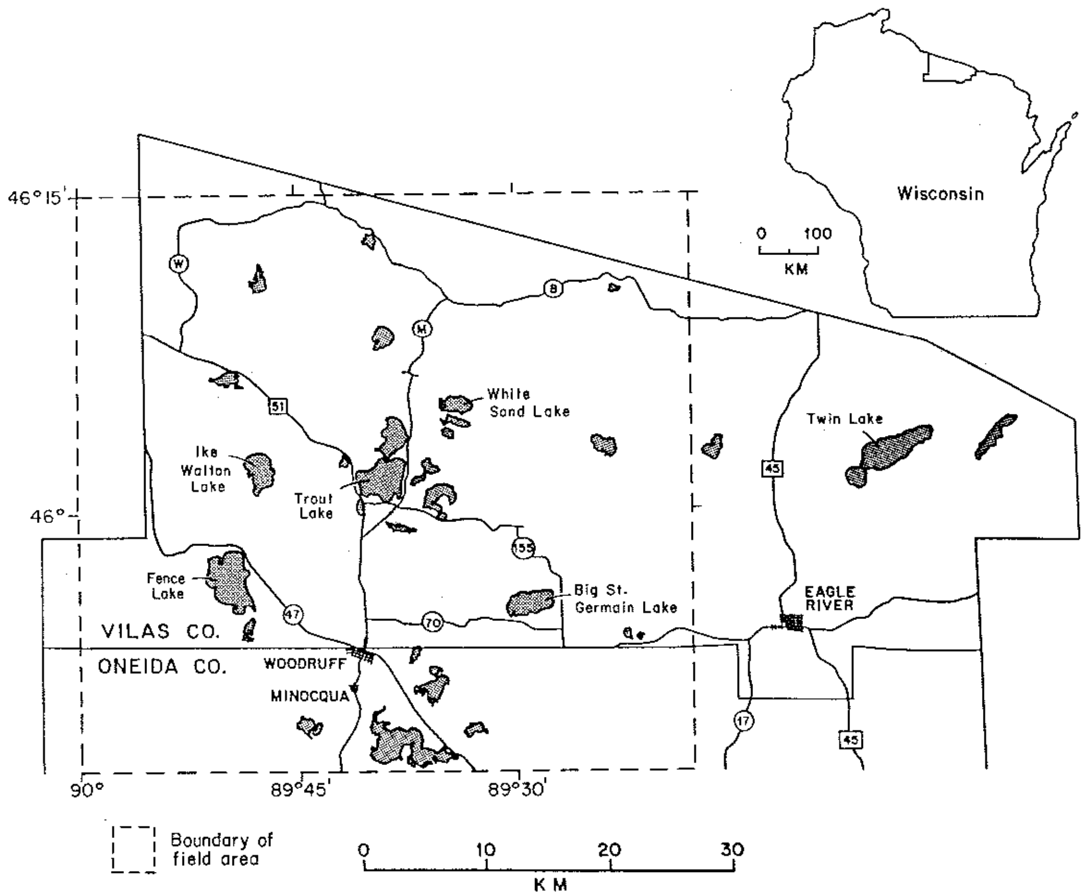


*Figure 1.* Location of the field area in northern Wisconsin. The dashed lines show the location of the field area with respect to Vilas county (solid lines).

Clay, 1977). The northern boundary of the anomaly lies roughly along an inferred east-west contact between an Early Precambrian gneiss terrane to the south and a younger Early Precambrian greenstone-granite terrane to the north (Sims, 1976). Daneshvar (1977) surveyed north-central Wisconsin from 90° W to 89°W and found that the Flambeau anomaly extends to at least 89° W through the Trout Lake region of Vilas county.

Several gravity surveys have been conducted in northern Wisconsin, but few have attempted to model the subsurface in any detail. Koo (1976) analyzed two north-south gravity traverses and concluded that a density contrast of 0.05-0.10 g $cm^{-3}$ accounts for the 20 mGal gravity high over the Flambeau anomaly. Fredricks (1975) studied the gravity of eastern Vilas county, found a single elliptical anomaly, and concluded that an extremely high density contrast (1.28 g $cm^{-3}$) is needed to model the feature. The Niagara fault separates two distinct crustal terranes: mafic intrusives are found north of the fault but are absent to the south. Davis (1977) obtained a gravity profile perpendicular to the strike of the Niagara fault and determined that two different basement terranes of contrasting density account for the measurements. This interpretation has been accepted by Greenberg and Brown (1983), but Mudrey and Ervin (1986, verbal communication) see no distinctive gravity pattern to justify this theory and instead consider mafic intrusives north of the fault to be the cause of the gravity high. Carlson (1974) studied the gravity field of northeastern Wisconsin and was able to distinguish several different units of the Middle Precambrian (granites, greenstones, and gneisses). His analysis of a profile across the Niagara fault led him to conclude that mafic intrusives, as well as the two terranes postulated by Davis, overly granitic gneiss.

## GEOLOGIC SETTING

Little is known about the bedrock geology of Vilas county because of the thick layer of glacial sediment overlying the Precambrian units. Most available information comes from geophysical data, exploration drill-hole logs, and extrapolation from surrounding areas where exposures are more common. Attig (1985) found only two bedrock exposures in the county, one of which is composed of white quartz and gray extrusive rocks. The other, also reported by Thwaites (1929), consists of coarse gray and pink granite intruded by pegmatite dikes. Drill holes penetrate granite, quartzite, slate, iron formation, schist, and gneiss (Dutton and Bradley, 1970). All of the bedrock is probably either Early or Middle Precambrian in age.

According to Attig (1985), the bedrock elevation in Vilas county ranges from 413 m above sea level in the south-central region to 512 m in the north-central and southeast areas. The two known Precambrian outcrops both occur at an elevation of 512 m, and nearby well logs show at least 50 m of local relief. Because of the scarcity of data, the bedrock elevation may vary considerably between contours. In general, elevation decreases from the north and northeast to the south and southwest. The Precambrian bedrock is covered by unconsolidated glacial drift ranging from 0 to 85 m thick. There is no simple relationship between the elevation of Precambrian bedrock and the thickness of the overlying unconsolidated sediment. The thickest overburden in eastern Vilas county (82 m) covers a relatively high bedrock surface, while the thick sediment in the south-central area (85 m) is underlain by the lowest Precambrian surface. The glacial features in western Vilas county consist chiefly of outwash, recessional moraines, and drumlins (Attig, 1985). Both the moraines and the outwash are heavily pitted by steep-sided kettles.

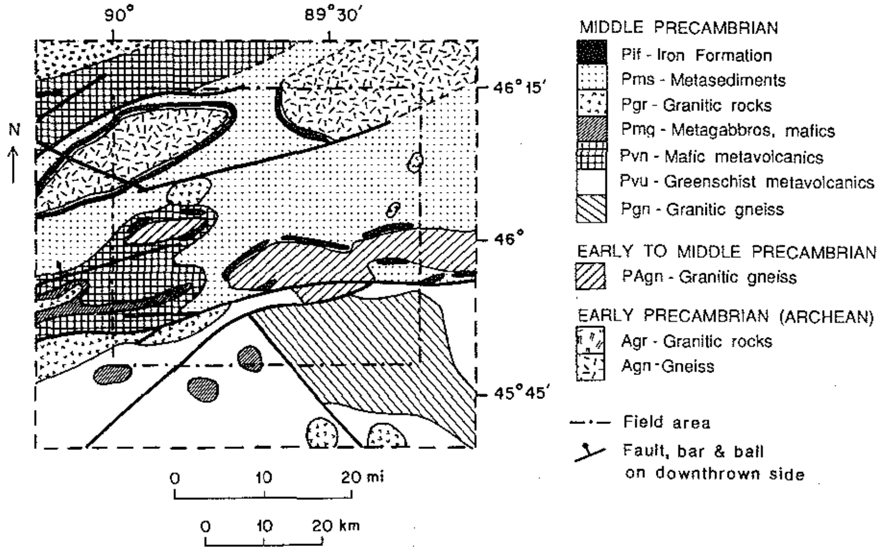


*Figure 2.* Geologic map of western Vilas county and adjacent areas, according to the Wisconsin Geological and Natural History Survey (Mudrey and others, 1982).

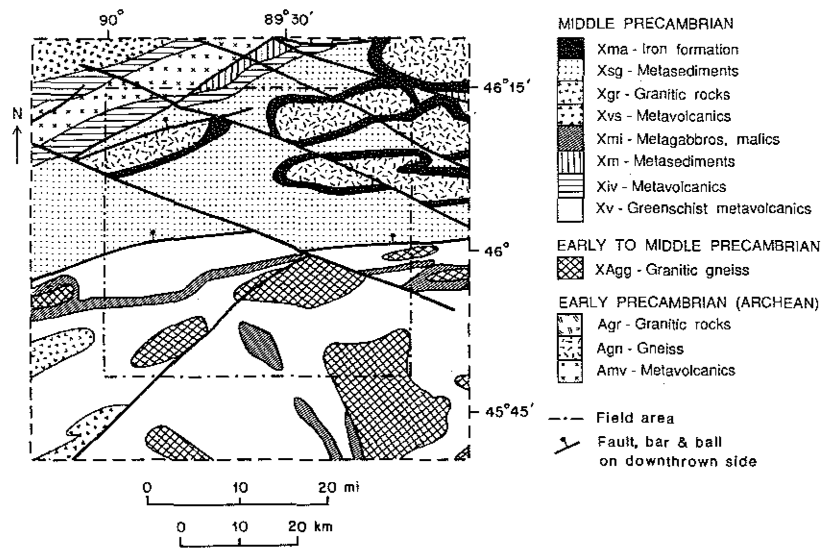


*Figure 3.* Geologic map of western Vilas county and adjacent areas, according to the Minnesota Geological Survey (Morey and others, 1982).

There are two very different 1982 maps of the bedrock geology of Vilas county, one published by the Wisconsin Geological and Natural History Survey and the other by the Minnesota Geological Survey (fig. 2 and 3; tables 1 and 2 ). Both maps have the same general characteristics, for example, Middle

*Table 1.* Descriptions of Early and Middle Precambrian rocks in northern Wisconsin. Symbols refer to the Wisconsin Geological and Natural History Survey (WGNHS) and the Minnesota Geological Survey (MGS) maps (from Mudrey and others, 1982; Morey and others, 1982).

| WGNHS SYMBOL | MGS SYMBOL | ROCK DESCRIPTION |
|---|---|---|
| MIDDLE PRECAMBRIAN INTRUSIVES: | | |
| Pgr | Xgr | Intermediate to granitic rocks |
| Pgn | XAgg | Banded and migmatitic gneiss with subordinate amphibolite and biotite schist |
| MIDDLE PRECAMBRIAN METAMORPHICS: | | |
| Pif | Xma | Magnetic iron formation, known or inferred from magnetic surveys |
| Pms | Xsg | Metasedimentary rocks—metagraywacke, slate, quartzite, meta-conglomerate, and marble—with some iron formation and metavolcanics |
| Pmg | Xmi | Metamorphosed mafic intrusives |
| Pvn | Xiv, Xm | Mafic metavolcanic rocks with subordinate felsic metavolcanics and metasedimentary rocks |
| Pvu | Xv, Xvs | Mafic metavolcanics with lesser felsic metavolcanic rocks |
| EARLY PRECAMBRIAN METAMORPHICS: | | |
| PAgn | XAgg | Metamorphic rocks of Early or Middle Precambrian age. Includes gneiss, biotite schist, and amphibolite |
| Agr | Agr, Amv | Metamorphosed mafic to intermediate rocks |
| Agn | Agn | Quartzofeldspathic gneiss, migmatite and, minor amphibolite |

Precambrian metasediments intruded by mantled gneiss domes to the north, and Middle Precambrian metavolcanics intruded by Early Precambrian mafic rocks to the south. The boundary between these two units is the Niagara fault (Dutton, 1971, and Greenberg and Brown, 1983). A 15-20 km wide band of Early Precambrian greenstone-granite terrane northwest of Vilas county is present on both maps, but the details of the two maps differ as to the location and extent of faults and geologic boundaries.

The Minnesota Survey map, compiled by Morey and others (1982), shows an extensive fault system throughout the region, which truncates many features such as the gneiss domes. South of the Niagara fault, the region is intruded by Early or Middle Precambrian mafic plutons. The Early or Middle Precambrian gneiss body in the center of the map is about 10 x 20 km at the surface of the bedrock.

According to the Wisconsin Survey map (Mudrey and others, 1982), there are fewer recognized faults in this area. The intrusives south of the Niagara fault are Middle Precambrian units. To the north, the geology is much more complicated. For example, the western region consists of mafics, gneiss, and granite (instead of simply metasediments as in the Morey diagram), and is substantiated by drill-hole data. The Early or Middle Precambrian gneiss structure is much longer in the east-west direction, based upon the presence of the mantling iron formation (Greenberg, 1986, verbal communication).

The differences between the two maps can be traced to the data from which they were created. The Minnesota Survey map was constructed predominantly from aeromagnetic measurements (Greenberg, 1986, verbal communication). The Wisconsin Survey map, however, used gravity and magnetic information, some drill-hole data not used by the Minnesota Survey, and, although outcrops are few, boulders in the glacial till were found to be indicative of the bedrock units to within a few kilometers. The gravity data will be used to determine the relative accuracy of the bedrock maps.

## COLLECTION OF DATA

The gravity data consist of measurements taken during the summer of 1982, using a LaCoste & Romberg Model G geodetic meter (number G-19), and data provided by C.P. Ervin, of Northern Illinois University, and M.G. Mudrey, Jr., of the Wisconsin Geological and Natural History Survey (fig. 4). The data pro-vided by the Wisconsin Survey are a composite from four different gravity studies that took place from 1972 to 1978 (Ervin and others, 1983). Most of these measurements were obtained with LaCoste & Romberg meters, but a small part of northwestern Vilas County was surveyed with a Worden meter. These readings have been corrected to the 1971 Inter-national Standardization Net (ISN-71).

The station spacing was kept to about 1.6 km along each traverse, but station sites were necessarily limited to places of known elevation accessible by reasonably passable roads. The entire gravity net was tied to the base station at Rhinelander, which has an absolute gravity of 980555.21 $\pm$ 0.05 mGal (Ervin, 1983).

The station elevations were obtained from 7.5- and 15-minute quadrangle topographic maps (contour interval = 10 ft) acquired from the Wisconsin Geological and Natural History Survey. The stations were located at U.S. Geological Survey and U.S. Coast and Geodetic Survey bench marks, and road corners where spot and spirit level elevations were given.

## DATA REDUCTION

By reoccupying the base station every two to three hours, the drift and tidal effects were eliminated with a simple linear interpolation between readings. The Geodetic Reference System (GRS-67) formula was used for the latitude reduction. The latitude, free-air, and Bouguer reductions were calculated using a density of 2.67 g $cm^{-3}$. The bedrock of western Vilas county is believed to be mafic metavolcanic and metasedimentary rock, with several large denser Early Precambrian gneiss bodies (Morey and others, 1982). However, this density value has been used for most gravity studies in northern Wisconsin (Fredricks, 1975; Koo, 1976) because core samples and density determinations are few and may not represent the entire region.

The 730 gravity station locations were converted to Cartesian coordinates by a Lambert projection, using a computer program supplied by C.P. Ervin of Northern Illinois University. The density of measurements ranges from about 7 points in the center of the region to 33 points per 50 $km^2$ in the southwest. The Bouguer gravity values were contoured manually, using a linear interpolation between the two nearest data points, and visually smoothed (fig. 5).

### Error Analysis

The accuracy of the measured gravity values is listed in table 3. The latitude reduction error ($\pm$ 0.08 mGal) was calculated using a $\pm$ 0.05 minute uncertainty in the latitude measurement. The Wisconsin gravity network base station values are reliable to within $\pm$ 0.05 mGal. The terrain correction for an area of similar topography was estimated by Carlson (1974) to be $\pm$ 0.25 mGal. The error of $\pm$ 0.02 mGal in the free-air correction comes from neglecting the higher order elevation terms. The net error is $\pm$ 0.62 mGal.

No error limits were provided with the data supplied by C.P. Ervin and the Wisconsin Geological and Natural History Survey. However, many sites coincided with those of the 1982 survey, and these were used to test the data's reliability. From 86 overlapping or nearby points, the absolute value of the Bouguer gravity discrepancies is 0.30 mGal, with a standard deviation of 0.24 mGal, which is well within the accuracy limits.

***Table 2.*** Correlation of rock units between the Wisconsin Geological and Natural History Survey (WGNHS) map and the Minnesota Geological Survey (MGS) map (from Mudrey and others, 1982; Morey and others, 1982). Ages of rock units are in billions of years (G.a.).

| WGNHS MAP | MGS MAP | AGE (G.a.) |
|---|---|---|
| Pif | Xma | 1.90–1.80 |
| Pms | Xsg | 1.90–1.80 |
| Pgr | Xgr | 1.90–1.80 |
| Pmg | Xmi | 1.90–1.80 |
| Pvn | Xiv, Xm | 1.90–1.80 |
| Pvu | Xv, Xvs | 1.90–1.80 |
| Pgn | XAgg | 1.90–1.80 |
| PAgn | XAgg | 3.00(?)–1.80(?) |
| Agr | Agr, Amv | 2.75–2.60 |
| Agn | Agn | 3.00(?)–2.75 |

## GRAVITY DATA ANALYSIS

### Density Values

The densities of the Precambrian rocks of northern Wisconsin are found in many sources, some of them unpublished. Most of these values, listed in table 4, are from Chandler (Open-file report, Minnesota Geological Survey). Additional data were obtained from Carlson (1974), Klasner and Cannon (1974), Klasner and others (1985), and Ocola and Meyer (1973). Densities range from 2.64 to 3.10 g $cm^{-3}$.

### Gravity Modeling Algorithm

The Bouguer gravity was modeled using both two- and two-and-one-half-dimensional algorithms developed by Talwani and others (1959) and by Cady (1980). The two-dimensional structure has infinite strike-length perpendicular to the profile, which defines the x-axis. The mass is truncated in the $\pm$ y-direction in the two-and-one-half-dimensional method. The gravitational attraction is then a function of the coordinates of the vertices of the polygon.

*Figure 4.* Locations of gravity stations in western Vilas county; the 730 measurements are scattered irregularly throughout the region. The southwest corner of the map is at latitude 45° 48.0' N, longitude 90° 00.0' W.

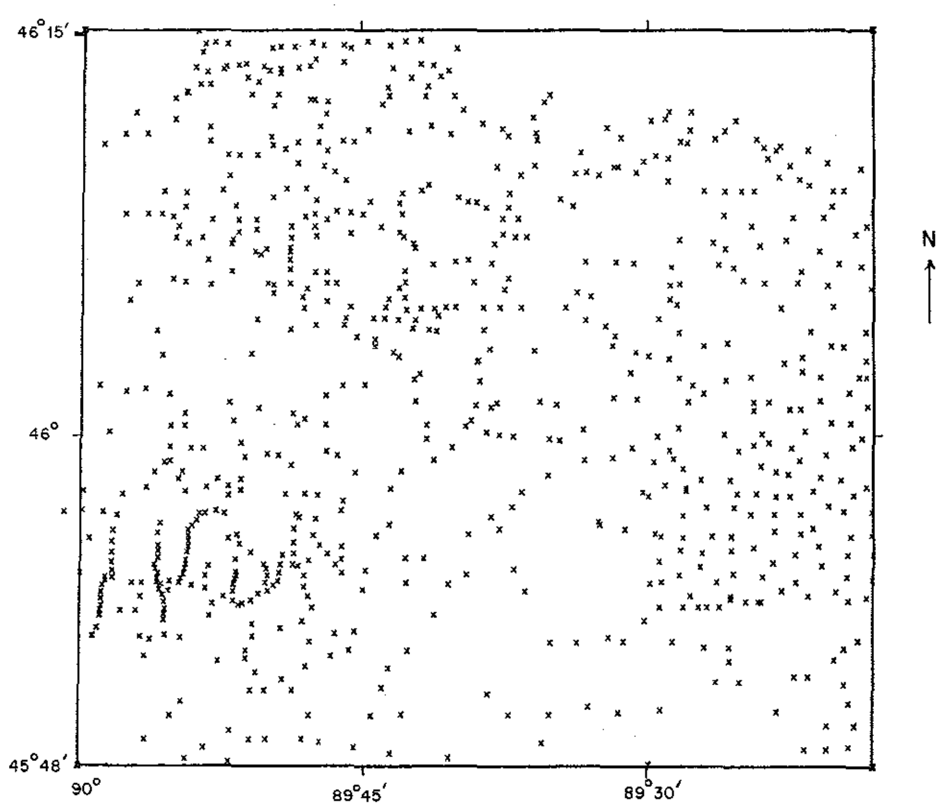


*Figure 5.* Bouguer gravity map of western Vilas county in northern Wisconsin, showing the three profiles AA', BB', and CC'. Gravity data are manually contoured and visually smoothed. The southwest corner of the map is at latitude 45° 48.0' N, longitude 90° 00.0' W. Contour Interval = 2.0 mGal.

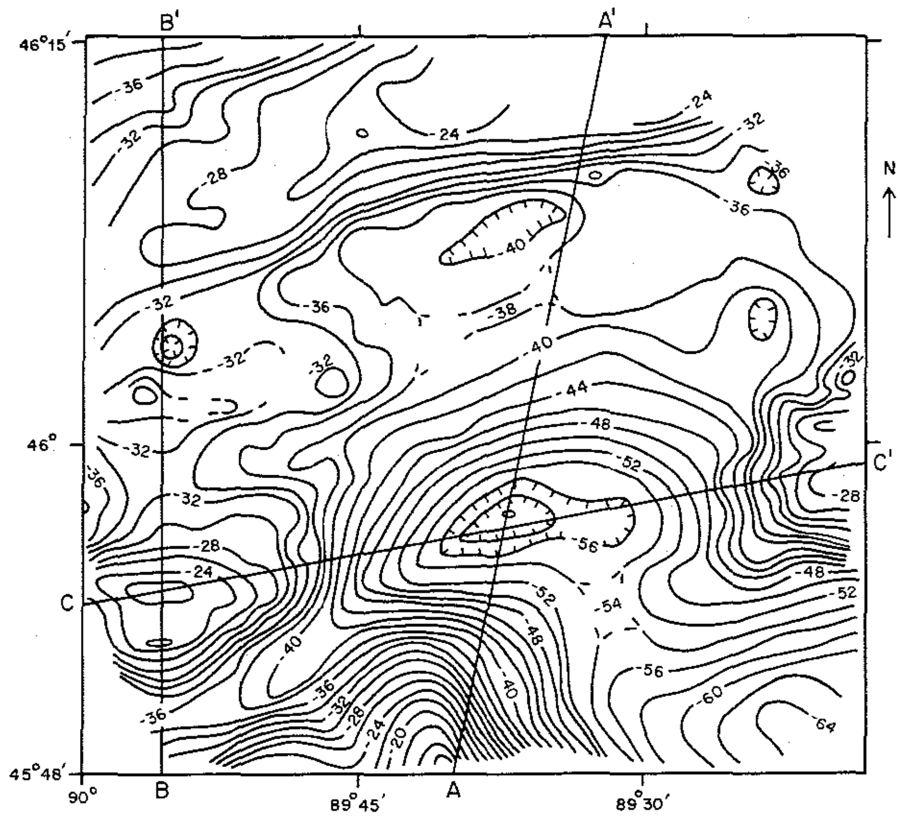

*Table 3.* Sources and magnitudes of error, in mGal, in Bouguer gravity values collected during 1982. The latitude error assumes an uncertainty of $\pm$ 0.05 minutes in latitude. Meter readings were accurate to within $\pm$ 0.05 mGal; hence errors are the same for the observed station value, the main base station reading at Rhinelander, and the local base station reading. The absolute gravity at the Rhinelander base station was 980555.21 $\pm$ 0.05 mGal with respect to the 1971 International Gravity Standardization Net (Ervin, 1983). The terrain correction of $\pm$ 0.25 mGal was estimated by Carlson (1974) for the area east of Vilas county, where the topography is similar. The free-air and Bouguer correction errors assume that altitudes are accurate to $\pm$ 0.30 m.

| SOURCE | ERROR (mGAL) |
|---|---|
| Latitude | ± 0.08 |
| Observed Station Value | ± 0.05 |
| Main Base Station Reading | ± 0.05 |
| Local Base Station Reading | ± 0.05 |
| Absolute Base Station Value | ± 0.05 |
| Terrain Correction | ± 0.25 |
| Free-air Correction | ± 0.09 |
| Bouguer Correction | ± 0.03 |
| Total Error | ± 0.62 mGal |

Three curves bisecting the major anomalies were selected for modeling (fig. 5). Profile AA' trends roughly north-south and transects both the high (-12 mGal) anomaly of the south-central region and the low (-60 mGal) feature in the central area. Profile BB' transects the positive (-22 mGal) anomaly to the southwest and the relatively low (-36 mGal) region due north. Profile CC' trends nearly east-west, crossing AA' at the -60 mGal area and BB' at the -22 mGal structure.

Each profile was first corrected for the glacial till layer and the greater bedrock density. Next, the regional gravity of -40 mGal was subtracted (Cohen and Meyer, 1966). Finally, the approximate boundaries of the geologic units were obtained from the bedrock maps, and polygons were selected to match the gravity curves. Since profile CC' crosses the other two, it was used as a constraint on both AA' and BB'. The fitting process was repeated until the polygons matched in both directions. Two models were derived for each profile: one based on the Wisconsin Geological and Natural History Survey (WGNHS) map, and the other on the Minnesota Geological Survey (MGS) map. Since outcrop and drill-hole data are scarce, the maps were used only as approximate guides to the bedrock surface. For the remainder of this work, all bedrock symbols correspond to those of the WGNHS map.

The average density of observations is about one reading for every 4 $km^2$ area, ranging from 0.5 readings to 2.6 readings per 4 $km^2$. Since the iron formation (Pif) typically occurs in long narrow bands about 0.1 km wide (Ervin, 1988, written communication), and is on the order of 0.2 km deep (Dutton, 1971), the gravity measurements cannot detect it except as high frequency noise. For example, a gravity profile bisecting a unit of iron formation 0.1 km wide, 0.2 km deep, and 10 km long will have an anomaly of +1.2 mGal directly over the iron formation, with a half-width of 0.07 km. Thus the iron formation cannot be accurately modeled with this data set.

## Till-Bedrock Correction

The Bouguer density of 2.67 g $cm^{-3}$ used to compute Bouguer anomalies is inaccurate since the till, which averages 40 m thick, has a density of 2.0 g $cm^{-3}$, and the average bedrock density is 2.74 g $cm^{-3}$ (Ocola and Meyer, 1973; Woollard, 1962). A till-bedrock correction was applied to each profile to compensate for these effects. Estimates of the till thickness and bedrock elevation were obtained beneath each profile from Attig (1985). The effects of the till and denser bedrock were then stripped from the Bouguer values. For example, the bedrock beneath profile AA' is very irregular, and the till thickness ranges from 85 m at the southern end to 5 m at the northern end. Figure 6 shows the cross-sectional view below AA', the till correction, bedrock correction, and the net correction along the profile. Since the till surface is relatively flat, the bedrock correction is essentially the negative of the till correction, but with a lower amplitude. Near the thickest till, the correction is as much as +1.0 mGal. As the till thins, the bedrock correction dominates, and an extremum of -1.4 mGal occurs. The net correction for profile BB' is always negative and ranges from -0.5 to -0.1 mGal. Profile CC' crosses a broad basin of till; its correction ranges from -0.5 to +0.5 mGal.

## Comparison of Gravity Models Based on the WGNHS and MGS Maps

The two-and-one-half-dimensional algorithm was used to derive structures that matched the observed

*Table 4.* Average densities (in g $cm^{-3}$) of Early and Middle Precambrian rock units in northern Wisconsin. The symbols refer to the Wisconsin Geological and Natural History Survey bedrock map. The references are: 1 – Chandler, 1986, written communication; 2 – Carlson (1974); 3 – Klasner and Cannon (1974); 4 – Klasner and others, (1985); 5 – Ocola and Meyer (1973).

| ROCK TYPE | SYMBOL | DENSITY | REF. |
|---|---|---|---|
| QUATERNARY: | | | |
| Glacial Till | – | 2.00 | 5 |
| MIDDLE PRECAMBRIAN: | | | |
| Iron Formation | Pif | 3.03 | 1 |
| Metasedimentary Rocks | Pms | 2.79 | 1 |
| Granitic Rocks | Pgr | 2.67 | 1 |
| Mafic Metaplutons | Pmg | 3.10 | 2 |
| Mafic Metavolcanics | Pvn | 2.92 | 4 |
| Greenschist Metavolcanics | Pvu | 2.91 | 3,4,5 |
| Granitic Gneiss | Pgn | 2.67 | 1,2 |
| EARLY PRECAMBRIAN: | | | |
| Granitic Rocks | Agr, PAgn | 2.64 | 1 |
| Metasedimentary Rocks | – | 2.74 | 1 |
| Granitic Gneiss | Agn | 2.64 | 1 |
| Felsic Gneiss | Agn | 2.89 | 1 |

gravity, using the WGNHS and MGS maps as guides to the subsurface geology. Figures 7-12 show the model cross sections and observed and calculated gravity fields. All gravity values were computed at the surface of the till, approx 500 m above sea level.

Using the WGNHS map, the gravity of profile AA' was calculated by placing the bodies of finite strike-length - mafic intrusives (Pmg), granitic gneiss (Pgn), Early or Middle Precambrian gneiss (PAgn), and Early Precambrian gneiss (Agn) - in infinitely broad units of either greenschist metavolcanics (Pvu) to the south or metasediments (Pms) to the north. The metavolcanic-metasedimentary boundary (the Niagara fault) was placed in the middle of the PAgn body. As shown by figure 7, a gravity maximum of -12 mGal at the southern end of profile AA' is modeled by a deep (approx 3.5 km) mafic intrusive body (Pmg), with a strike-length of $\pm$ 3 km (that is, 3 km both east and west of AA' ). The Pgn unit is only 1.3 km thick and extends from 4 km west of AA' to 30 km east of AA'. The Early or Middle Precambrian dome centered at 18 km lowers the field to -60 mGal and is about 8 km deep, becoming wider with depth. The dashed line on the cross section within the PAgn unit indicates that polygons of different strike-length (that is, different extent perpendicular to the profile) were used in the model. The felsic Early Precambrian gneiss (Agn) at the northern end of the profile is about 2.7 km deep, ranging from 4 km west to 10 km east of AA'. The greenschist metavolcanic layer is 1.1 km thick, and the metasediments range from 0.7 to 1.6 km thick.

Using the MGS map, the cross section beneath profile AA' is roughly the same as for the WGNHS profile, but the location of the greenschist metavolcanic (Pvu) and metasediment (Pms) boundary is further north (fig. 8). This makes a significant change in the Early or Middle Precambrian gneiss dome. Since the dome is now surrounded by higher density material, it must be much larger in order to decrease the gravity field to -60 mGal. Its uppermost portion, however, is very narrow because of the sharp drop in gravity directly over the dome's center.

Profile BB' has a considerably more complex sequence of bedrock units, based upon recent drill-hole data in the southwest region (Greenberg, 1986, verbal communication). Figure 9 illustrates the cross section as well as the observed and modeled gravity based on the WGNHS map. Although not shown on the diagram, the gravity was modeled by encircling the mafic plutons (Pmg) with mafic metavolcanics (Pvn) 1.4 km thick. Also, the Pvn unit was surrounded by metasediments (Pms). In this model, the mafic body (Pmg) had to be widened from 2 km in the north-south direction (as shown on the WGNHS map) to about 6 km to account for the broad gravity maximum of -23 mGal. The gravity over the Pmg unit ranges from -24.0 to -22.2 mGal, with a slight dip in the middle. A small unit of Middle Precambrian granite centered over this anomaly may account for the gravity drop but is entirely speculative. However, this interpretation corresponds with another structure just west of Vilas county mapped from drill-hole data (Greenberg, 1986, verbal communication). The 1.8 mGal fluctuation may easily be due to an undetected increase in the thickness of the glacial sediments. At 30 km, the lowered gravity (-33 mGal) is shown on the WGNHS map to be the result of another Early or Middle Precambrian dome, but this is not supported by the gravity: even a relatively shallow (approximately 3 km) and narrow dome decreases the field by almost 10 mGal below the observed values. Thin granitic masses, similar to those mapped throughout the region, may account for the lowered gravity. The lowest gravity, -36 mGal at 30 km, is above the metasediment (Pms) layer, 1.3 km thick and infinite in strike. The gravity increases 10 mGal at about 35 km over the denser Early Precambrian gneiss (Agn). At the northern extreme of BB', the rapid decrease in gravity is probably due to the large body of Early Precambrian granite (Agr) located just north of Vilas county.

The cross section for profile BB' based on the MGS map is almost exactly the same as that for the WGNHS profile (fig. 10). The MGS map does not distinguish between greenschist (Pvu) and mafic (Pvn) metavolcanics, and the density difference between these rocks is only 0.01 g cm$^{-3}$; hence the models for this area are the same. A step-like boundary between the mafic unit (Pmg) and the greenschist metavolcanic unit fits the gravity better than the sloping contact of the WGNHS profile. The major difference is the location of the metavolcanic-metasediment contact of the Niagara fault, which is about 13 km further north on this diagram. Since the metasediment has a relatively low density, it contributes little to the gravity at the southern end of BB', and the southern boundary cannot be determined.

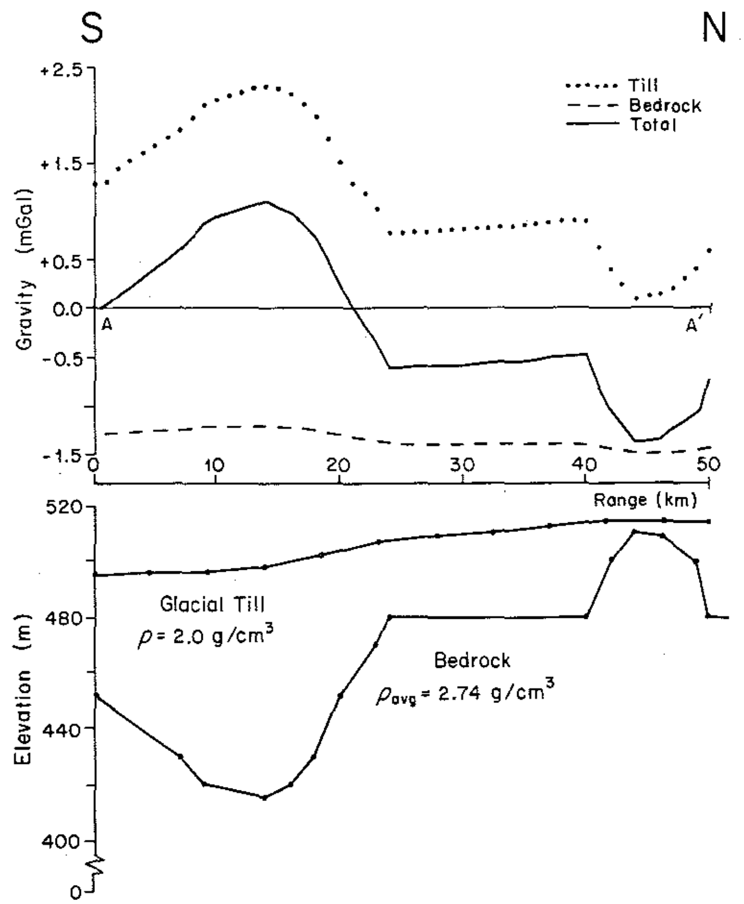


***Figure 6.*** Till-bedrock correction for profile AA'. The dotted line represents the correction term for the till layer, and the dashed line is the bedrock layer. The solid curve shows the values to be added to the Bouguer gravity to strip these layers from the observed field.

Profile CC' crosses both the -60 mGal dome of AA' and the -23 mGal feature of BB' (fig. 11). The bodies of Early or Middle Precambrian gneiss (PAgn) and mafic plutons (Pmg) have been translated into their corresponding two-and-one-half-dimensional polygons in the east-west direction. From the WGNHS map, the Pmg body is very narrow (approx 0.8 km) in the north-south direction at the extreme western edge of the profile. (The eastern boundary of the narrowest part is marked by dashed lines on fig. 11.) From the gravity model, the unit widens around the intersection of BB' and CC', then narrows once more. Not shown but included in the model is the surrounding mafic metavolcanics (Pvn). Further east, the sequence rapidly changes from metavolcanics to metasediments (Pms), followed by the dome of Early or Middle Precambrian gneiss (PAgn). This model differs from that shown on the WGNHS map in that the dome must end at about 43 km from the western end of CC' because of the increase in gravity from -60 to -30 mGal. On the map, the dome ends at about the eastern edge of Vilas county, but the gravity becomes 30 mGal less than the observed values using this model. Since the feature is mantled by iron formation, the increase in gravity may be due to another unit of mafic rocks (Pmg) within metavolcanics (Pvn), similar to those found at the western end of the profile. The gravity may have other causes, such as underlying material or a change in the dome's shape. In this example, the Pvn unit is 9 km wide and the Pmg rock is about 5 km wide.

Profile CC' based on the MGS data does not match the WGNHS diagram because the greenschist metavolcanic (Pvu) material is found along its entire length. Hence, as shown by figure 12, the Early or

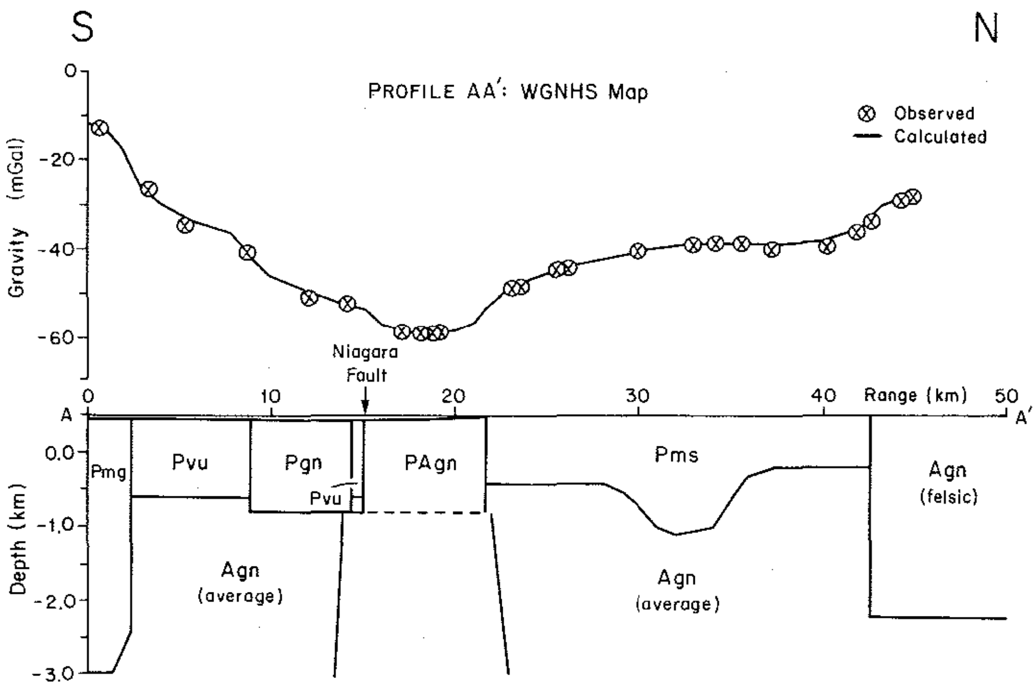


*Figure 7.* Observed (X···X) and calculated (—) gravity, in mGal, along profile AA', according to the WGNHS map. The axes of the cross section are in kilometers.

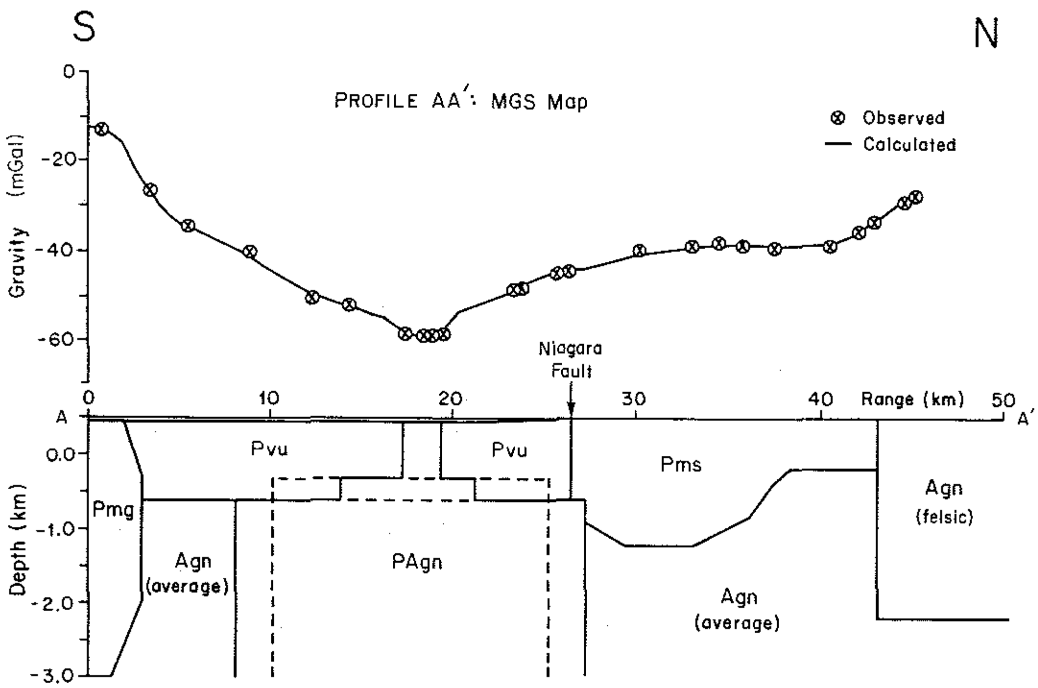


*Figure 8.* Observed (X···X) and calculated (—) gravity, in mGal, along profile AA', according to the MGS map. The axes of the cross section are in kilometers.

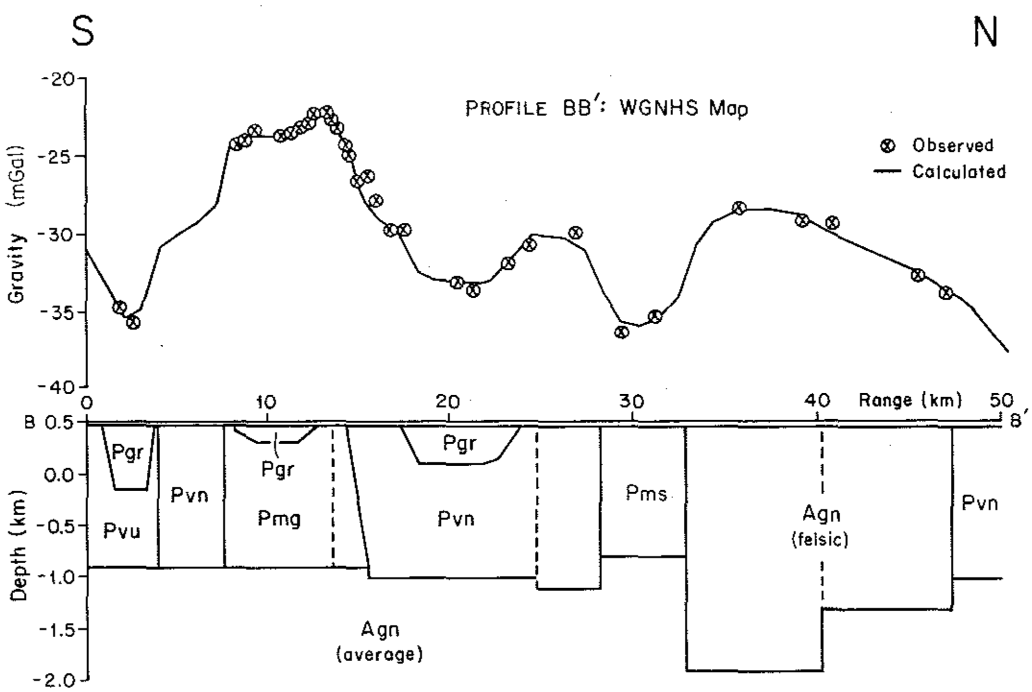


*Figure 9.* Observed (X···X) and calculated (—) gravity, in mGal, along profile BB', according to the WGNHS map. The axes of the cross section are in kilometers.

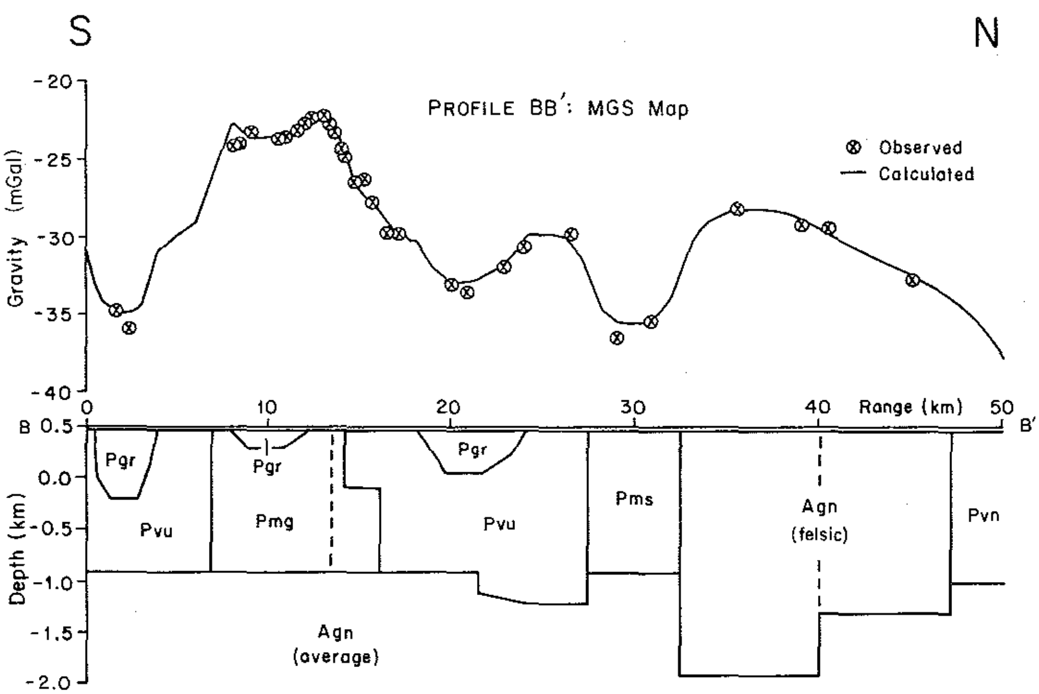


*Figure 10.* Observed (X···X) and calculated (—) gravity, in mGal, along profile BB', according to the MGS map. The axes of the cross section are in kilometers.

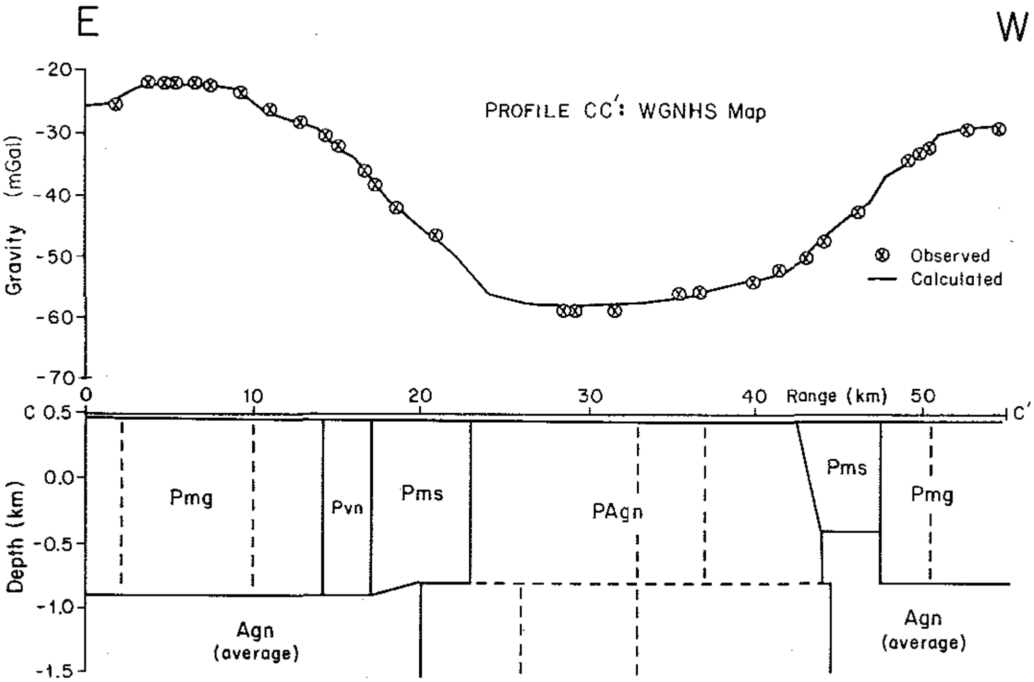


*Figure 11.* Observed (X···X) and calculated (—) gravity, in mGal, along profile CC', according to the WGNHS map. The axes of the cross section are in kilometers.

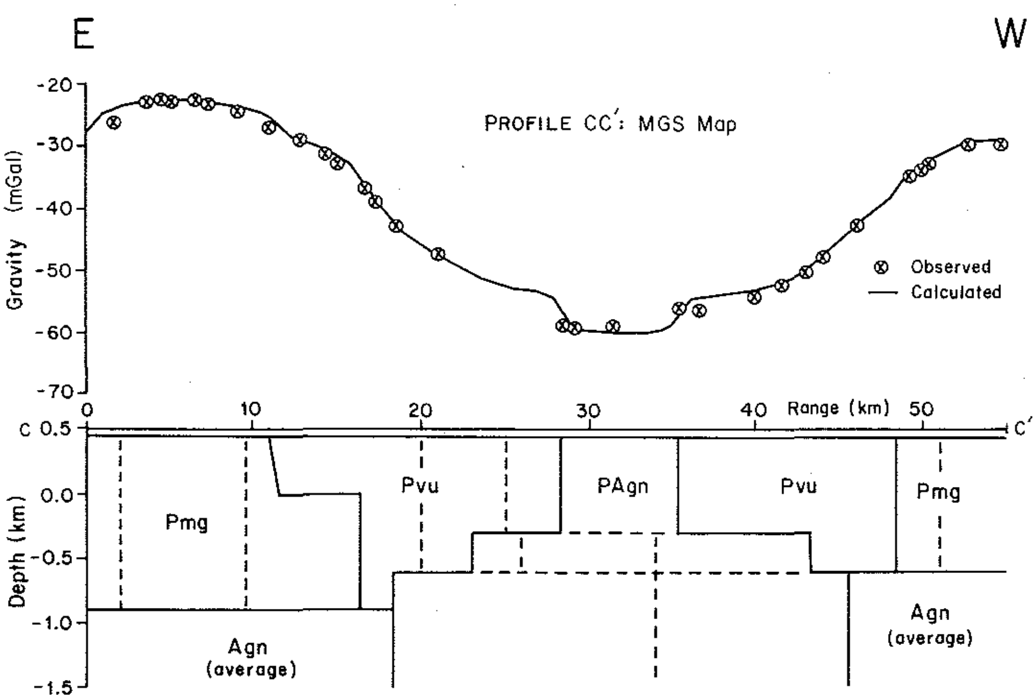


*Figure 12.* Observed (X···X) and calculated (—) gravity, in mGal, along profile CC', according to the MGS map. The axes of the cross section are in kilometers.

Middle Precambrian dome must be quite large yet narrow at the surface.

Figures 13 and 14 are three-dimensional views of the Early or Middle Precambrian dome models. From the WGNHS map (fig. 13), the structure is narrower on the eastern end to correspond with the gradually increasing gravity. The total depth is 8 km, the east-west length is a maximum of 26 km at the base, and the base in the north-south direction is 9 km long. A rough estimate of the total mass of the dome is approx $4 \times 10^{15}$ kg. In the MGS model (fig. 14), if the total depth is held at 8 km, the dome must be very wide and broad. Its maximum length along profile AA' (the north-south direction) is 18 km. Beneath profile CC', the base of the dome is 32 km long, and the net mass is approx $9 \times 10^{15}$ kg, double that of the dome from the WGNHS models.

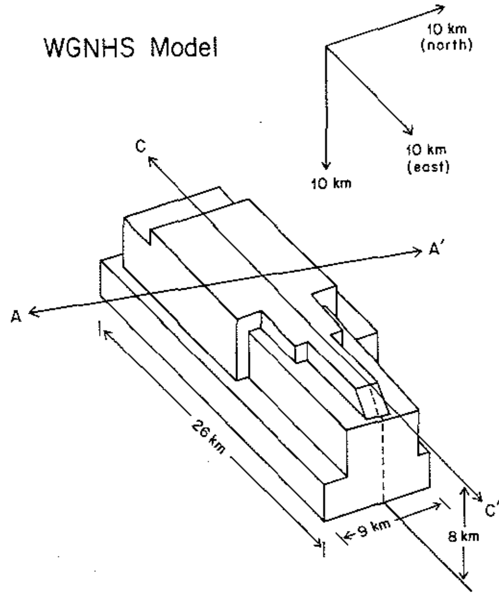


*Figure 13.* Early or Middle Precambrian gneiss dome in three dimensions, from the Wisconsin Geological and Natural History Survey gravity models.

## CONCLUSIONS

The variation in thickness of the overlying till and in the elevation of the bedrock surface may have a significant effect on the residual gravity. If the till-bedrock interface is fairly flat, the effect may be less than 0.5 mGal. However, when the interface is highly irregular, the effect is negative over low and positive over high bedrock and can easily range from -1.4 to +1.0 mGal, which will greatly alter a gravity model. By modeling the subsurface as two-and-one-half-dimensional polygons, one can effectively strip this layer from the gravity field.

Because of the greater amount of bore-hole data used in the WGNHS map, the bedrock surface should be generally more reliable than that of the MGS map. For example, the mafic metavolcanics (Pvn) in the southwestern region are not shown on the MGS diagram, and the Niagara fault, the contact between the greenschist metavolcanics (Pvu) and the metasedimentary units (Pms), is further north. Since Pvn only occurs in association with metasediments, this is also evidence for the more southern boundary hypothesized by the Wisconsin Survey, which in turn supports the smaller version of the Early or Middle Precambrian gneiss (PAgn) dome. However, the large eastward extent of the dome is unsupported by the gravity field. A more consistent interpretation is another unit of metavolcanics (Pvn) surrounding mafic intrusives, since this is also mantled by iron formation and thus consistent with the magnetics of the region.

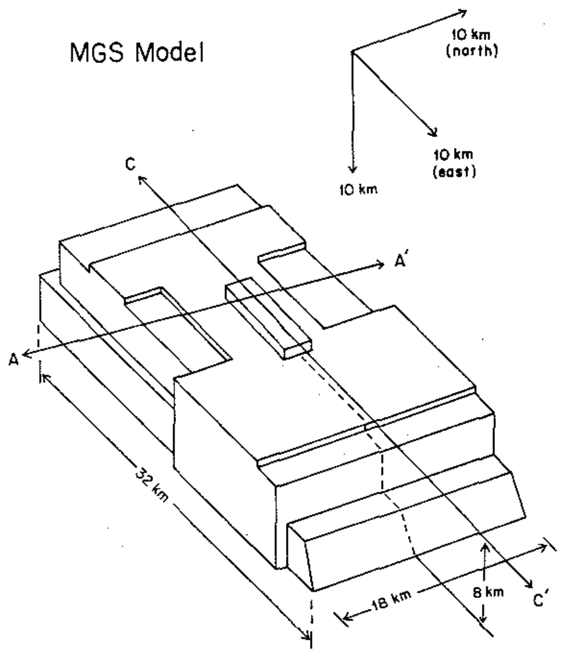


*Figure 14.* Early or Middle Precambrian gneiss dome in three dimensions, from the Minnesota Geological Survey gravity models.

The Early Precambrian gneiss (Agn) domes must be relatively dense in character since they increase the gravity field. The small PAgn body in western Vilas county, as shown on the WGNHS diagram, is not supported by drill-hole data (Greenberg, 1986, verbal communication), and the lowered gravity may more likely be the result of granitic (Pgr) bedrock. In the southwestern area, neither map shows a feature large enough to account for the high (-22 mGal) gravity. A mafic intrusive, about 7 km wide in the north-south direction and 15 km long in the east-west direction, matches the observed field. The exact location of the southern boundary of this body cannot be determined from the gravity because there are few measurements in this region.

The anomalous masses on either side of the Niagara fault prevent a determination of the fault's strike and dip from the gravity data. The changes in gravity from these bodies are much greater than the fluctuation due to the fault, and the fault's gravity field cannot be reliably separated from the total field. Also, the extent of the iron formation cannot be determined because the spatial wavelength of these long narrow features is much shorter than the measurement separation.

Northern Wisconsin, and particularly Vilas county, has few outcrops because of the thick glacial till, and drill-hole data are also scarce. The interpretation of the gravity field thus required several assumptions based upon the geology of the region surrounding the field area. Additional information, particularly concerning the density of the upper crust, may change the parameters determined from this analysis.

## ACKNOWLEDGMENTS

This study was undertaken as part of a Ph.D. thesis at the University of Wisconsin-Madison. I thank my advisor, C.S. Clay, and the members of my thesis committee, C.R. Bentley, H.F. Wang, J.S. Steinhart, and C.J. Bowser, for their guidance and support during the course of this study. This study could not have been completed without the data provided by J.K. Greenberg and M.G. Mudrey, Jr. of the Wisconsin Geological and Natural History Survey, by C.P. Ervin at Northern Illinois University, and by V. Chandler of the Minnesota Geological Survey. The Center for Limnology was also very generous with its facilities at the Trout Lake Biological Station in Vilas county.

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# Density and Magnetic Susceptibility of Wisconsin Rock

*Steven I. Dutch[1], Robert C. Boyle[1,2], Sandra K. Jones-Hoffbeck[1,3], and Sharon M. Vandenbush[1,4]*

## ABSTRACT

*Density and magnetic susceptibility values were determined for a suite of specimens from Wisconsin and northern Michigan that provide a sample of a wide variety of cover and basement rock types in the northern Midcontinent. In addition, we summarize a large number of previously published values. In general, the densities of Midcontinent rock agree with the normal range of densities for any given rock type, and therefore, the assumed density values used in most geopyhysical studies are valid. One significant exception is the Keeweenawan mafic rock suite, which shows a lower density contrast with granitic basement rock than has heretofore been assumed in many studies. Magnetic properties, which depend largely on magnetite content, show no simple correlation with rock type, but the overall pattern observed is consistent with the usual assumptions made in magnetic studies: cover rock is essentially nonmagnetic, mafic rock is generally magnetic, and granite is highly variable. The Wolf River Batholith has susceptibility of 100 to 1000 x $10^{-6}$ cgs units, whereas the rhyolite and epizonal granite of south-central Wisconsin are highly magnetic, with susceptibility over 1000 x $10^{-6}$ cgs units. Similar susceptibility has been previously reported for the younger rhyolite and epizonal granite of the St. Francois Mountains in southeastern Missouri. The possibility, therefore, exists that the widespread rhyolite and epizonal granite of the eastern Midcontinent is generally magnetic.*

## INTRODUCTION

Although many structural studies of the buried Precambrian basement of the central United States have been published in recent years (Dutch, 1983; Klasner and King, 1986; Sims and Peterman, 1986), there is as yet little published data on physical properties of Midcontinent rock, such as density or magnetic susceptibility. This study is an effort to fill part of that gap.

Since 1976 the senior author has been assembling a reference collection of Wisconsin rock. Wisconsin is particularly well situated for purposes of this study because every major Precambrian rock suite in the eastern Midcontinent except for the Eastern Rhyolite-Granite suite (1420-1500 Ma; Bickford and others, 1986) is exposed in or near Wisconsin (See Anderson, 1983; Sims and Peterman, 1983; 1986 for additional overviews of regional geology). Wisconsin also has a well-exposed and nearly complete lower Paleozoic section. The reference collection, at the time this study was done, numbered 236 specimens, of which 204 were measured for density. Magnetic susceptibilities were determined for 114 specimens, including most of the Precambrian specimens and a representative sampling of cover rock. Results are presented in tables 1-3, along with published data from other sources. The sample collection has been assembled as opportunity permitted, and is not exhaustive; Archean and Keeweenawan rock is particularly under-represented. Nevertheless, the data published here should serve a useful purpose in reducing uncertainties in geophysical interpretation and perhaps in inspiring additional data gathering. Density values from previous studies (about 1000 values) aid greatly in filling gaps in our own sample collection.

[1] College of Environmental Sciences, University of Wisconsin/Green Bay, Green Bay, Wisconsin 54301-7001

[2] Present address: 1512 Walls Street, Green Bay, WI 54311

[3] Present address: 1440 Applecreek Road, Appleton, WI 54915

[4] Present address: 2021 Deckner Avenue, Apt. 332, Green Bay, Wisconsin 54301

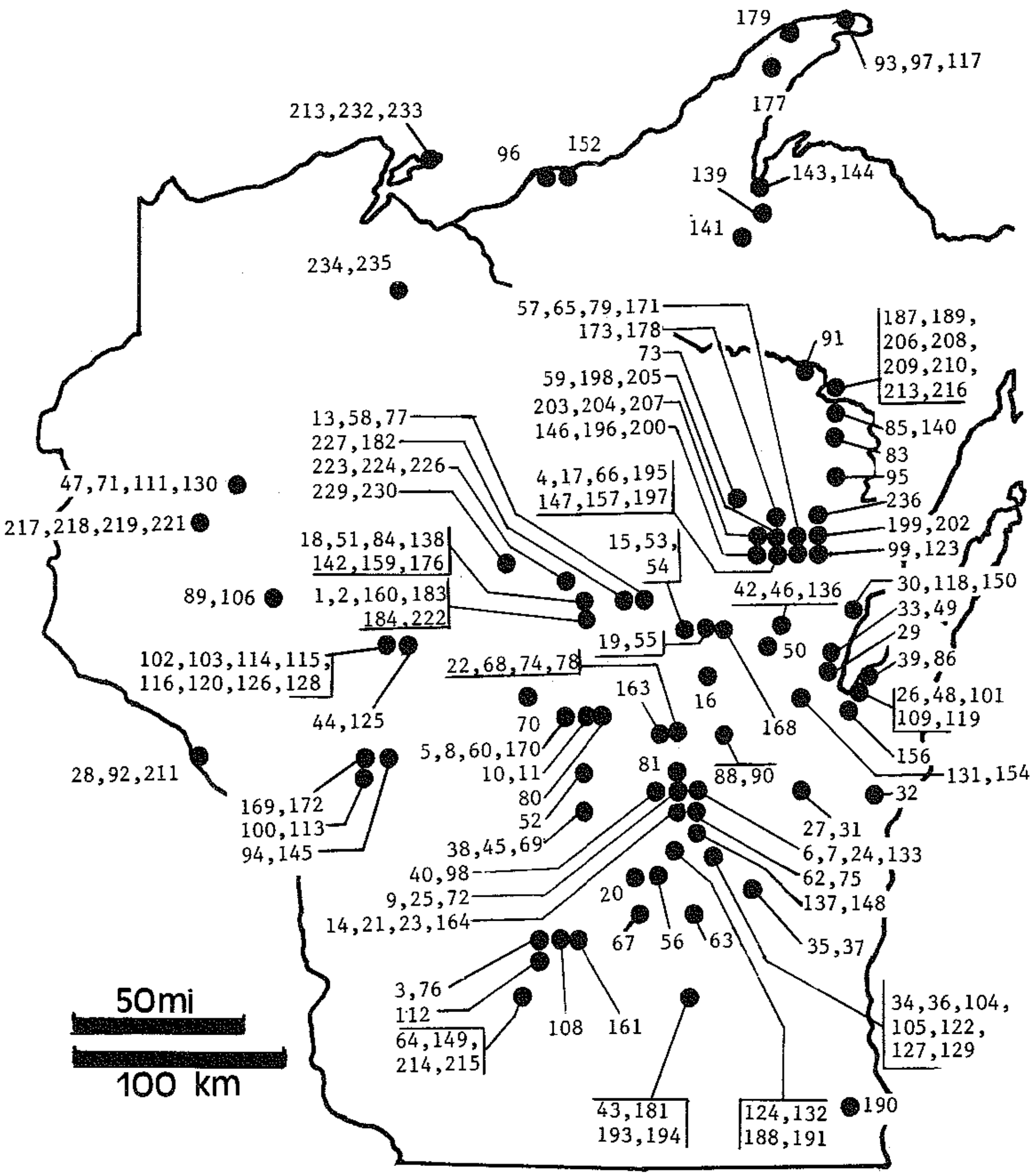


*Figure 1.* Map of Wisconsin and northern Michigan showing localities of samples measured by authors.

## METHODOLOGY

Density was determined by two methods. About 80 samples were measured by differential weighing in water and in air (Jolly balance method) with porous rock like sandstone being soaked for several hours before weighing in air. There was no significant difference in density between samples soaked for 24 hours and samples of the same rock soaked for one or two hours, so we conclude that the effect of trapped air is negligible. The coarseness of most Wisconsin sandstone accounts for the high permeability of the specimens. In addition, the full sample suite was measured by volumetric displacement. In general, densities determined by the two methods agreed to within 0.5 percent.

The apparatus for determining sample volume consisted of a pair of identical water-filled basins connected by a siphon. A sample was placed in one basin and displaced water ran through the siphon to the second basin, which was situated on a digital laboratory scale. The weight of water added to the second basin is proportional to the volume displaced by the specimen. The total volume of the sample is twice the volume of water displaced into the second basin. The apparatus was calibrated by adding known volumes of water from a volumetric flask to the sample basin; this calibration accounted for both instrumental errors and temperature corrections for the density of water.

The method described above offers several advantages: it is considerably faster and more convenient than the conventional method of using a balance with two pans. It seems to be much more suitable for large specimens (almost all samples weighed over 100 g, most were about 500 g), and large specimens are more likely to be representative of the rock as a whole. Finally, some sources of error, such as adhering water and sample loss in handling, are easier to control using our technique.

Magnetic susceptibility was determined using a Bison Model 3101 magnetic susceptibility meter. Samples were crushed to 1 cm or smaller and placed in glass or plastic vials (which measurements showed to have zero magnetic susceptibility). The instrument is calibrated for cylindrical (core) samples 2.54 cm in diameter by 7.62 cm long, and samples of other sizes must be corrected for length and cross-sectional area. In addition, corrections for void spaces in crushed samples are necessary. The full corrections are:

$$(1) \quad k = R\,(7.62/L)\,(2.54/D)^2\,(C/W)$$

where k is the true magnetic susceptibility, R is the reading from the instrument, L is the length of the sample in cm, D is the sample diameter in cm, C is the density of the rock and W the bulk density of the crushed sample. We can multiply numerator and denominator by $\pi/4$ and combine terms to obtain:

$$(2) \quad k = R\,C\,(\pi/4 * 2.54^2 * 7.62)/M$$

or

$$(3) \quad k = 38.611\,R\,C/M$$

where M is the weight of the sample. Even though all length terms have vanished in equation (3), it is still based on the assumption of a cylindrical specimen and should not be uncritically applied to irregular specimens.

The magnetic susceptibility meter is an inductance bridge. Susceptibility is read on a dial in terms of $10^{-6}$ cgs units. In this paper, magnetic susceptibilities will be reported as multiples of $10^{-6}$ cgs units. SI and cgs units differ only by a simple constant:

$$(4) \quad k(SI) = 4\ k(cgs) = 12.566k(cgs)$$

Repeated measurements on specimens suggest that the measurements are repeatable to about $10^{-5}$ cgs units or about 10 dial gradations. This level of error is insignificant for most crystalline rock but is significant for weakly magnetic cover rock. Because the Bison instrument drifts, the meter was re-zeroed for each measurement.

Statistical summaries of density and magnetic susceptibility of major rock suites are presented in tables 2 and 3. For density, table 2 tabulates minimum, mean, and maximum values. Magnetic susceptibility is much more variable than density; it can vary significantly within a single outcrop, be greatly reduced by weathering or alteration, or be increased greatly by small increases in magnetite content. Magnetic susceptibility readings showed such wide scatter that the arithmetic mean is not a suitable measure of central tendency because it can be greatly skewed by outliers. Accordingly, magnetic susceptibility data are tabulated in table 3 in terms of the median and logarithmic mean. For larger sample collections, quartile values are also tabulated.

## PREVIOUSLY PUBLISHED DATA

There is a large amount of published density data for Midcontinent rock in early reports of state geological surveys. These data were collected primarily in connection with reports on the dimension stone and aggregate industries. In some cases as well, it appears that density distinctions were also used to separate fine-grained igneous rock types before the advent of widespread thin-section analysis. The determination of density (usually by the Jolly balance or differential-weighing method) is straightforward and there is no reason to doubt that these values compare favorably in quality with contemporary measurements. The more recent data of Cain (1964), Leney (1966) and Carlson (1972) was collected for geophysical or petrographic studies.

Magnetic susceptibility values are not nearly as abundant in the literature, and tend to have been collected for two purposes: modelling a specific magnetic anomaly or determining the range of susceptibilities of various rock types. There seem to be few surveys of susceptibility aimed at tabulating reference values for regional geophysical interpretation. One such survey (Allingham, 1964) will be discussed later.

*Table 1.* Density and magnetic susceptibility of rocks from Wisconsin and northern Michigan. Density in gm/cm$^3$, susceptibility in $10^{-6}$ cgs units. All localities in Wisconsin unless otherwise noted.

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| **ARCHEAN** | | | | | | | |
| ***Black River Falls Area*** | | | | | | | |
| 94 | | IF | Black River Falls | 21N | 03W | 3.04 | 1968 |
| 145 | | GGN | Black River Falls | 21N | 03W | 2.60 | 120 |
| 169 | | GN | Black River Falls | 21N | 04W | 2.64 | 341 |
| 172 | | GR | Black River Falls | 21N | 04W | 2.61 | 67 |
| ***South-Central Wisconsin*** | | | | | | | |
| 10 | | MIG | W of Stevens Point | 23N | 07E | 2.63 | |
| 11 | | MIG | W of Stevens Point | 23N | 07E | 2.66 | 448 |
| 60 | | GR | Biron | 23N | 06E | 2.73 | 881 |
| 5 | | (1) | Biron | 23N | 06E | 2.61 | |
| 8 | | (1) | Biron | 23N | 06E | 2.60 | 273 |
| 80 | | (2) | Whiting | 23N | 08E | 2.63 | 61 |
| ***Northern Wisconsin and Michigan*** | | | | | | | |
| 234 | | GAB | S of Mellen | 44N | 02W | 2.98 | 766 |
| | High Bridge | GR | (B) | 45N | 03W | 2.67 | |
| | | GGN | MI (L12, Mean of 5) | 40N | 30W | 2.63 | |
| **PENOKEAN SUPRACRUSTAL ROCK** | | | | | | | |
| ***Northern Michigan Metavolcanic and Metasedimentary Suite*** | | | | | | | |
| 139 | | MB | Alberta MI | 49N | 33W | 2.77 | 188 |
| 141 | | MB | Covington, MI | 48N | 34W | 2.83 | 140 |
| | Michigamme | SL | MI (L2, Min. of 98) | 40N | 30W | 2.17 | |
| | Michigamme | SL | MI (L2, Mean of 98) | 40N | 30W | 3.00 | |
| | Michigamme | SL | MI (L2, Max. of 98) | 40N | 30W | 3.72 | |
| 187 | Vulcan | IF | Groveland Mine MI | 40N | 30W | 3.19 | |
| 189 | Vulcan | IF | Groveland Mine MI | 40N | 30W | 3.35 | |
| 208 | Vulcan | IF | Groveland Mine MI | 40N | 30W | 3.53 | |
| 208 | Vulcan | IF | Groveland Mine MI | 40N | 30W | 3.05 | |
| | Vulcan | IF | MI (L2, Min. of 103) | 40N | 30W | 2.87 | |
| | Vulcan | IF | MI (L2, Mean of 103) | 40N | 30W | 3.44 | |
| | Vulcan | IF | MI (L2, Max. of 103) | 40N | 30W | 3.87 | |
| 206 | Felch | (3) | Groveland Mine Mi | 40N | 30W | 2.94 | |
| | Felch | QZ,Phy | MI (L2, Min. of 48) | 40N | 30W | 2.22 | |
| | Felch | QZ,Phy | MI (L2, Mean of 48) | 40N | 30W | 2.72 | |
| | Felch | QZ,Phy | MI (L2, Max. of 48) | 40N | 30W | 3.22 | |
| 210 | Randville | (4) | Groveland Mine MI | 40N | 30W | 2.89 | |
| | Randville | (4) | MI (L2, Min. of 32) | 40N | 30W | 2.67 | |
| | Randville | (4) | MI (L2, Mean of 32) | 40N | 30W | 2.86 | |
| | Randville | (4) | MI (L2, Max. of 32) | 40N | 30W | 3.07 | |
| ***Quinnesec Volcanics and related rocks*** | | | | | | | |
| 83 | Quinnesec | MGR | Pembine | 37N | 20E | 2.86 | 357 |
| | | MR | (C2, Mean of 2) | 37N | 20E | 2.75 | |
| | | (20) | (C2) | 37N | 20E | 2.95 | |
| | Quinnesec | MB | (C2, Mean of 15) | 37N | 20E | 2.97 | |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| | Quinnesec | MB | (C2, Mean of 14) | 35N | 17E | 3.00 | |
| | Quinnesec | MB | (C2, Mean of 3) | 41N | 14E | 2.89 | |
| | Quinnesec | MB | (C2, Mean of 2) | 41N | 13E | 2.94 | |
| | | MB | (C2, Mean of 2) | 36N | 11E | 2.97 | |
| | | MR | (C2) | 36N | 11E | 2.65 | |
| | Quinnesec | MR | (C2) | 36N | 19E | 2.77 | |
| 85 | Quinnesec | (5) | S of Niagara | 38N | 20E | 2.80 | 152 |
| 91 | Quinnesec | MB | W of Niagara | 40N | 18E | 2.69 | 115 |
| 140 | Quinnesec | MB | S of Niagara | 38N | 20E | 2.94 | 124 |
| | Badwater | MB | (C2) | 40N | 17E | 2.97 | |
| ***Waupee Volcanics*** | | | | | | | |
| 12 | Waupee | MB | Butler rock | 31N | 18E | 2.84 | 80 |
| 197 | Waupee | MB | Mountain | 31N | 17E | 2.82 | |
| 147 | Waupee | MB | Mountain | 31N | 17E | 2.91 | |
| 157 | Waupee | MB | Mountain | 31N | 17E | 2.76 | |
| | Waupee | MB | (C2, Mean of 6) | 31N | 17E | 3.06 | |
| | Waupee | MB | (C2, Mean of 2) | 31N | 17E | 2.63 | |
| 174 | Waupee | MB | Butler Rock | 31N | 18E | 2.78 | |
| | Waupee | (20) | (C2, Mean of 4) | 32N | 18E | 3.06 | |
| ***Marathon County Mafic Suite*** | | | | | | | |
| 13 | | (6) | Eau Claire Dells | 29N | 10E | 3.00 | 2713 |
| 223 | | MB | Little Chicago | 30N | 06E | 2.76 | 225 |
| 226 | | (7) | Little Chicago | 30N | 06E | | 91 |
| 227 | | MB | SW Eau Claire Dells | 29N | 09E | 2.94 | 115 |
| 229 | | AM | Goodrich Dells | 31N | 03E | 2.94 | 168 |
| ***Marathon County Felsic Suite*** | | | | | | | |
| 58 | | (8) | Eau Claire Dells | 29N | 10E | 2.66 | 704 |
| 77 | | (8) | Eau Claire Dells | 29N | 10E | 2.54 | 2430 |
| 224 | | MR | Little Chicago | 30N | 06E | 2.65 | 65 |
| ***Post-Volcanic Sediments*** | | | | | | | |
| 51 | Marshall Hill | MCG | Brokaw | 29N | 07E | 2.84 | 3518 |
| 159 | | (9) | Brokaw | 29N | 07E | 3.01 | |
| 176 | Marshall Hill | MCG | Brokaw | 29N | 07E | 2.78 | 3633 |
| ***Penokean (?) Mafic Intrusive Rocks*** | | | | | | | |
| | | AM | (C2, Mean of 5) | | | 2.93 | |
| | | GAB | (C2) | 33N | 18E | 2.99 | |
| | | GAB | (C2) | 33N | 18E | 2.77 | |
| | | GAB | (C2) | 31N | 17E | 3.01 | |
| | | DIO | (C2, Mean of 2) | 33N | 18E | 2.91 | |
| | | DIO | (C2, Mean of 2) | 31N | 17E | 2.90 | |
| | | DIO | (C2, Mean of 3) | 35N | 20E | 2.77 | |
| | | MD | (C2) | 35N | 17E | 2.93 | |
| | | MD | (C2, Mean of 3) | 36N | 15E | 3.04 | |
| | | MD | (C2) | 36N | 11E | 2.83 | |
| | | AM | (C2, Mean of 3) | 33N | 12E | 2.99 | |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| ***Penokean Gneissic Rocks*** | | | | | | | |
| | Dunbar | GGN | (C2, Mean of 4) | | | 2.62 | |
| | Dunbar | GGN | (C2, Mean of 4) | | | 2.72 | |
| | Dunbar | GGN | (C2) | 37N | 16E | 2.70 | |
| | Dunbar | GGN | (C2, Mean of 4) | 38N | 16E | 2.64 | |
| | Dunbar | GN | (C2, Mean of 3) | 38N | 16E | 2.98 | |
| | Dunbar | GGN | (C2, Mean of 2) | 39N | 15E | 2.59 | |
| | Dunbar | GGN | (C2, Mean of 3) | 39N | 14E | 2.65 | |
| | Dunbar | GGN | (C2, Mean of 3) | 38N | 13E | 2.77 | |
| | Macauley | GGN | (C2) | 31N | 17E | 2.69 | |
| | | GN | (C2, Mean of 4) | 37N | 10E | 2.80 | |
| | | AM | (C2, Mean of 3) | 37N | 10E | 2.90 | |
| | | AM | (C2, Mean of 3) | 36N | 15E | 2.90 | |
| | | GN | (C2, Mean of 2) | 37N | 13E | 2.71 | |
| | | AM | (C2, Mean of 4) | 40N | 15E | 2.88 | |
| ***Penokean Granitic Rocks*** | | | | | | | |
| 95 | Amberg | QMZ | Amberg | 35N | 20E | 2.63 | |
| 216 | | | | | | | |
| | Amberg | QMZ | (C2, Mean of 6) | 35N | 20E | 2.67 | |
| 199 | | GR | Johnson Falls Dam | 32N | 19E | 2.68 | |
| 202 | | GR | Johnson Falls Dam | 32N | 19E | 2.67 | |
| 230 | | (1) | Goodrich Dells | 31N | 03E | 2.63 | |
| 366 | | | | | | | |
| 236 | Pomeroy | GR | Middle Inlet | 33N | 19E | 2.64 | |
| 1057 | | | | | | | |
| | Marinette | QD | (C2, Mean of 5) | 38N | 20E | 2.71 | |
| | Newingham | GRD | (C2, Mean of 4) | 38N | 20E | 2.71 | |
| | Newingham | GRD | (C1, Min. of 53) | 38N | 20E | 2.66 | |
| | Newingham | GRD | (C1, Mean of 53) | 38N | 20E | 2.70 | |
| | Newingham | GRD | (C1, Max. of 53) | 38N | 20E | 2.74 | |
| | Hoskin Lake | GR | (C2, Mean of 16) | 38N | 19E | 2.67 | |
| | Twelvefoot F. | QD | (C2, Mean of 2) | 36N | 19E | 2.90 | |
| | Athelstane | QMZ | (C2, Mean of 3) | 35N | 20E | 2.64 | |
| | Athelstane | QMZ | (C2, Mean of 9) | 33N | 20E | 2.74 | |
| | | GR | (C2, Mean of 2) | 34N | 18E | 2.64 | |
| | | GR | (C2, Mean of 3) | 34N | 16E | 2.69 | |
| | | GR | (C2, Mean of 2) | 34N | 17E | 2.69 | |
| | | GR | (C2) | 34N | 18E | 2.72 | |
| | Undivided | GR | (B, Mean of 6) | | | 2.68 | |
| | St Cloud, MN | GR | (B, Mean of 3) | | | 2.68 | |
| **POST-PENOKEAN RHYOLITE-EPIZONAL GRANITE TERRAIN** | | | | | | | |
| ***Metaluminous Granites*** | | | | | | | |
| 88 | "Cactus Rock" | GR | New London | 22N | 14E | 2.55 | |
| 490 | | | | | | | |
| 90 | "Cactus Rock" | GR | New London | 22N | 14E | 2.56 | |
| 22 | | GR | Waupaca | 22N | 12E | 2.64 | |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| 159 | | | | | | | |
| 24 | | GR | Poy Sippi | 19N | 13E | 2.65 | 573 |
| 6 | | CGR | Poy Sippi | 19N | 13E | 2.61 | |
| 7 | | CGR | Poy Sippi | 19N | 13E | 2.70 | 1179 |
| 72 | | GR | Pine River | 19N | 12E | 2.69 | 2258 |
| 9 | | CGR | Pine River | 19N | 12E | 2.63 | 2024 |
| 81 | | GR | Saxeville | 20N | 12E | 2.57 | 603 |
| 25 | | GR | Mount Morris | 19N | 12E | 2.66 | 220 |
| 21 | | GR | Redgranite | 18N | 12E | 2.56 | 300 |
| 164 | | GR | Redgranite | 18N | 12E | 2.64 | 2268 |
| 14 | | MD | Redgranite | 18N | 12E | 3.04 | 138 |
| 23 | | MD | Redgranite | 18N | 12E | 2.85 | 161 |
| | | GR | Waushara (B, 2 samples) | | | 2.64 | |
| 20 | | GR | Montello | 15N | 10E | 2.63 | 490 |
| | | GR | Montello (B, 2 samples) | | | 2.64 | |
| ***Metaluminous Rhyolites*** | | | | | | | |
| 75 | | (10) | Berlin | 17N | 13E | 2.45 | 149 |
| 62 | | RHY | Berlin | 17N | 13E | 2.71 | 2388 |
| | | RHY | Berlin (B, mean of 4) | | | 2.64 | |
| 63 | | RHY | Utley | 13N | 13E | 2.68 | 1681 |
| 161 | | RHY | NE of Baraboo | 12N | 07E | 2.62 | 1411 |
| | | RHY | NE of Baraboo (H) | 12N | 07E | 2.66 | |
| ***Peraluminous Rhyolites*** | | | | | | | |
| 56 | | RHY | Marquette | 15N | 11E | 2.59 | 1939 |
| | | RHY | Marquette | 15N | 11E | 2.65 | |
| 67 | | RHY | Marcellon | 13N | 10E | 2.57 | 445 |
| ***Unclassified*** | | | | | | | |
| 68 | | RHY | Waupaca | 22N | 12E | 2.72 | 2668 |
| 74 | | RHY | Waupaca | 22N | 12E | 2.78 | 3489 |
| 78 | | (11) | Waupaca | 22N | 12E | 2.53 | 142 |
| | Baxter Hollow | GR | (H, mean of 3) | 11N | 06E | 2.63 | |
| | Denzer | DIO | (H, 2 samples) | 10N | 05E | 2.83 | |
| **POST-PENOKEAN EPICRATONIC QUARTZITES (BARABOO INTERVAL)** | | | | | | | |
| ***McCaslin Syncline*** | | | | | | | |
| 4 | Baldwin | CGL | Mountain | 31N | 17E | 2.69 | |
| 66 | Baldwin | CGL | Mountain | 31N | 17E | 2.65 | 310 |
| 73 | McCaslin | QZ | Carter | 34N | 15E | 2.57 | 108 |
| 82 | | (12) | Mountain | 31N | 17E | 2.55 | |
| 79 | Thunder Mtn | MGR | Thunder Mtn | 32N | 18E | 2.70 | 150 |
| 173 | Thunder Mtn | QZ | Thunder Mtn | 33N | 17E | 2.64 | 462 |
| 178 | Thunder Mtn | QZ | Thunder Mtn | 33N | 17E | 2.68 | |
| | Thunder Mtn | QZ | (C2, Mean of 3) | 33N | 17E | 2.68 | |
| ***Rib Mtn*** | | | | | | | |
| 1 | Rib Mtn | QZ | Rib Mtn | 28N | 07E | 2.61 | 93 |
| 2 | Rib Mtn | QZ | Rib Mtn | 28N | 07E | 2.67 | 112 |
| 160 | Rib Mtn | QZ | Rib Mtn | 28N | 07E | 2.63 | |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| ***Barron*** | | | | | | | |
| 47 | Barron | QZ | Cameron | 34N | 10E | 2.59 | 111 |
| 71 | Barron | QZ | Cameron | 34N | 10E | 2.58 | |
| ***Central Wisconsin*** | | | | | | | |
| 52 | Hamilton Mound | QZ | Hamilton Mound | 20N | 07E | 2.65 | 44 |
| 70 | Powers Bluff | QZ | Bethel | 24N | 04E | 2.56 | 71 |
| ***Baraboo*** | | | | | | | |
| 3 | Baraboo | QZ | Rock Springs | 12N | 05E | 2.62 | |
| 76 | Baraboo | (13) | Rock Springs | 12N | 05E | 2.72 | 86 |
| | Baraboo | QZ | (H, mean of 5) | | | 2.66 | |
| | Dake | QZ | (H, mean of 3) | 12N | 06E | 2.68 | |
| | Freedom | SL,IF | (H, 2 samples) | 11N | 05E | 2.64 | |
| ***Southeastern Wisconsin*** | | | | | | | |
| 43 | Waterloo | QZ | Hubbleton | 09N | 13E | 2.64 | |
| 181 | Waterloo | MCG | Portland Quarry | 09N | 13E | 2.66 | 67 |
| 193 | Waterloo | QZ | Portland Quarry | 09N | 13E | 2.66 | |
| 194 | Waterloo | PHY | Portland Quarry | 09N | 13E | 2.81 | |
| **WOLF RIVER BATHOLITH** | | | | | | | |
| ***Northern Felsite Suite*** | | | | | | | |
| 17 | Hagar | (14) | Mountain | 31N | 17E | 2.56 | 481 |
| 57 | Hagar | (14) | High Falls Dam | 32N | 18E | 2.60 | 186 |
| 59 | Hagar | FSP | Crooked Lake | 32N | 17E | 2.65 | 293 |
| 198 | Hagar | FSP | Crooked Lake | 32N | 17E | 2.64 | |
| 205 | Hagar | FSP | Crooked Lake | 32N | 17E | 2.64 | |
| | Hagar | FSP | (C2) | 32N | 17E | 2.67 | |
| | Hagar | FSP | (C2) | 32N | 17E | 2.78 | |
| ***Northern Batholith*** | | | | | | | |
| 146 | | GR | Mountain | 31N | 16E | 2.64 | 213 |
| 171 | High Falls? | GR | High Falls Dam | 32N | 18E | 2.60 | 329 |
| 195 | | GR | Mountain | 31N | 17E | 2.61 | |
| 196 | | GR | Chute Pond | 31N | 16E | 2.61 | |
| 200 | | GR | Mountain | 31N | 16E | 2.61 | |
| | Belongia | GR | (C2, mean of 2) | 31N | 16E | 2.64 | |
| 203 | | GR | Lakewood | 32N | 16E | 2.67 | |
| 204 | | GR | Lakewood | 32N | 16E | 2.77 | |
| 207 | | GR | Lakewood | 32N | 16E | 2.68 | |
| ***Central Batholith*** | | | | | | | |
| 16 | Wolf River | QMZ | Marion | 25N | 13E | 2.60 | 53 |
| | Wolf River | QMZ | (C2, mean of 3) | 30N | 15E | 2.68 | |
| 19 | | GR | Bowler | 27N | 13E | 2.60 | 500 |
| 55 | | GR | Bowler | 27N | 13E | 2.72 | 162 |
| 163 | Waupaca | QMZ | Waupaca | 22N | 11E | 2.69 | 270 |
| | Waupaca | QMZ | (B, 2 samples) | 22N | 11E | 2.68 | |
| | Waupaca | QMZ | (C2) | | | 2.71 | |
| | Red River | QMZ | (C2, mean of 3) | | | 2.66 | |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| 168 | Wolf River | GR | W of Shawano | 27N | 14E | | 142 |
| 182 | | GR | W of Aniwa | 29N | 09E | | 266 |
| ***Wausau Syenite*** | | | | | | | |
| 18 | Wausau | SY | Wausau | 29N | 07E | 2.72 | |
| 84 | Wausau | SY | Wausau | 29N | 07E | 2.66 | 497 |
| 138 | Wausau | SY | Wausau | 29N | 07E | 2.69 | |
| 142 | Wausau | SY | Wausau | 29N | 07E | 2.71 | |
| ***Ninemile Pluton*** | | | | | | | |
| 183 | Ninemile | (15) | W of Wausau | | | | 24 |
| 184 | Ninemile | GR | W of Wausau | 28N | 07E | 2.63 | 119 |
| 222 | | MD | W of Wausau | 28N | 07E | 2.66 | 500 |
| | Wausau | GR | Wausau (B, 5 samples) | | | 2.63 | |
| ***Anorthosite and Gabbro*** | | | | | | | |
| 15 | Tigerton | ANO | E of Wittenberg | 27N | 12E | 2.64 | 68 |
| 53 | Tigerton | ANO | Bowler | 27N | 12E | 2.68 | 639 |
| 54 | Tigerton | ANO | Bowler | 27N | 12E | 2.72 | 967 |
| | Tigerton | ANO | (C2) | | | 2.74 | |
| 65 | | GAB | High Falls Dam | 32N | 18E | 2.71 | 761 |
| KEEWEENAWAN | | | | | | | |
| ***Igneous Rocks*** | | | | | | | |
| 152 | | (20) | Porcupine Mtns MI | 51N | 42W | 2.60 | 188 |
| 179 | | (16) | Eagle River MI | 58N | 31E | 2.70 | 92 |
| 235 | | BAS | Copper Falls St Pk | 44N | 02W | 2.81 | 2703 |
| | | BAS | Calumet, MI (L1, min. of 68) | | | 2.76 | |
| | | BAS | Calumet, MI (L1, mean of 68) | | | 2.89 | |
| | | BAS | Calumet, MI (L1, max. of 68) | | | 2.76 | |
| | | (16) | Calumet, MI (L1, min. of 50) | | | 2.70 | |
| | | (16) | Calumet, MI (L1, mean of 50) | | | 2.85 | |
| | | (16) | Calumet, MI (L1, max. of 50) | | | 3.09 | |
| ***Sedimentary Rocks*** | | | | | | | |
| 93 | | CGL | Copper HBR MI | 58N | 28W | 2.67 | 176 |
| 96 | | SS | Lake of Clouds MI | 51N | 43W | 2.55 | 83 |
| 97 | | SS | Copper Hbr MI | 58N | 28W | 2.36 | 142 |
| 117 | | CGL | Copper Hbr MI | 58N | 28W | 2.67 | |
| 177 | | (17) | Calumet MI | 56N | 32W | 2.97 | 178 |
| 231 | Bayfield | SS | Madeline Is | 50N | 02W | 2.21 | 31 |
| 232 | Bayfield | CGL | Madeline Is | 50N | 02W | 2.24 | 31 |
| 233 | Bayfield | SS | Madeline Is | 50N | 02W | 2.23 | |
| | | SS | WI (B, min. of 14) | | | 2.62 | |
| | | SS | WI (B, mean of 14) | | | 2.63 | |
| | | SS | WI (B, min. of 14) | | | 2.65 | |
| | | SS | WI (B, min. of 14) | | | 2.62 | |
| | | CGL | MI (B, min. of 10) | | | 2.62 | |
| | | CGL | MI (B, mean of 14) | | | 2.74 | |
| | | CGL | MI (B, max of 10) | | | 2.84 | |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| **LATE PRECAMBRIAN** | | | | | | | |
| 143 | Jacobsville | SS | L'Anse MI | 50N | 33W | 2.21 | 85 |
| 144 | Jacobsville | SS | L'Anse MI | 50N | 33W | 2.23 | |
| | Jacobsville | SS | MI (B) | | | 2.16 | |
| | Jacobsville | SS | MI (B) | | | 2.29 | |
| **CAMBRIAN** | | | | | | | |
| 40 | Undivided | SS | Mount Morris | 19N | 11E | 2.50 | 57 |
| 89 | Undivided | BCG | Chippewa Falls | 28N | 08W | 2.32 | |
| 98 | Undivided | SS | Mount Morris | 19N | 11E | 2.22 | |
| 99 | Undivided | SS | W of Pound | 31N | 19E | 2.29 | |
| 100 | Undivided | SS | Black River Falls | 20N | 04W | 2.14 | 16 |
| 102 | Undivided | SS | Fort Mc Coy | 17N | 03W | 2.33 | 125 |
| 103 | Undivided | SS | Fort Mc Coy | 17N | 03W | 2.33 | |
| 106 | Undivided | SS | Chippewa Falls | 28N | 08W | 2.09 | |
| 108 | Lodi | SH | Baraboo | 12N | 06E | 2.41 | 34 |
| 111 | Undivided | SS | Cameron | 34N | 10W | 2.20 | |
| 112 | Galesville | SS | Larue | 11N | 05E | 2.18 | |
| 113 | Undivided | SS | Black River Falls | 20N | 04E | 2.16 | |
| 114 | Undivided | SS | Fort Mc Coy | 17N | 03W | 2.37 | |
| 115 | Undivided | SS | Fort Mc Coy | 17N | 03W | 2.41 | |
| 116 | Undivided | SS | Fort Mc Coy | 17N | 03W | 2.08 | 20 |
| 120 | Undivided | SS | Fort Mc Coy | 17N | 03W | 1.95 | |
| 123 | Undivided | SS | W of Pound | 31N | 19E | 2.31 | |
| 124 | Jordan | SS | St Marie Quarry | 16N | 12E | 2.09 | |
| 126 | Undivided | SS | Fort Mc Coy | 17N | 03W | 2.31 | |
| 128 | Undivided | SS | Fort Mc Coy | 17N | 03W | 2.03 | |
| 130 | Undivided | BCG | Cameron | | | | 61 |
| 132 | Jordan | SS | St Marie Quarry | 16N | 12E | 2.05 | |
| 133 | Jordan | SS | Poy Sippi | 19N | 13E | 2.49 | |
| 136 | Undivided | SS | Cecil | 27N | 17E | 2.44 | |
| 137 | Undivided | BCG | Berlin | 17N | 13E | 2.10 | |
| 148 | Undivided | BCG | Berlin | 17N | 13E | 2.50 | |
| 170 | Undivided | SS | W of Stevens Point | 23N | 06E | 2.51 | |
| 211 | Undivided | SS | Alma | 21N | 13W | 2.06 | |
| 213 | Undivided | SS | Groveland Mine MI | 40N | 30W | 2.03 | |
| 216 | Undivided | SS | Groveland Mine MI | 40N | 30W | 2.47 | |
| | Undivided | SS | MI (L2, Min. of 37) | 40N | 30W | 2.17 | |
| | Undivided | SS | MI (L2, Mean of 37) | 40N | 30W | 2.42 | |
| | Undivided | SS | MI (L2, Max. of 37) | 40N | 30W | 2.82 | |
| 217 | Undivided | SS | Ridgeland | 32N | 12W | 2.52 | |
| 221 | Undivided | SS | Ridgeland | 32N | 12W | 2.30 | |
| | Undivided | SS | WI (B, min. of 12) | | | 2.50 | |
| | Undivided | SS | WI (B, mean of 12) | | | 2.59 | |
| | Undivided | SS | WI (B, max. of 12) | | | 2.63 | |
| | Undivided | SS | MN (B, min. of 3) | | | 2.34 | |
| | Undivided | SS | WI (B, max. of 3) | | | 2.38 | |
| **ORDOVICIAN** | | | | | | | |
| ***Prairie du Chien Group*** | | | | | | | |
| 28 | Pr. Du Chien | DOL | Alma | 21N | 13W | 2.45 | |
| 42 | Pr. Du Chien | DOL | Cecil | 27N | 17E | 2.68 | 52 |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| 46 | Pr. Du Chien | DOL | Cecil | 27N | 17E | 2.60 | |
| 0 | Pr. Du Chien | DOL | Shawano | 26N | 16E | 2.66 | |
| 92 | Pr. Du Chien | DOL | Alma | 21N | 13W | 2.35 | 59 |
| | Pr. Du Chien | DOL | WI (B, min. of 4) | | | 2.74 | |
| | Pr. Du Chien | DOL | WI (B, max. of 4) | | | 2.81 | |
| | Pr. Du Chien | DOL | MN (B, min. of 3) | | | 2.64 | |
| | Pr. Du Chien | DOL | MN (B, max. of 3) | | | 2.74 | |
| 188 | Oneonta | DOL | St Marie Quarry | 16N | 12E | 2.81 | |
| 191 | Oneonta | DOL | St Marie Quarry | 16N | 12E | 2.82 | |
| ***Saint Peter Ss*** | | | | | | | |
| 104 | Saint Peter | SS | Ripon | 16N | 14E | 2.32 | 126 |
| 105 | Saint Peter | SS | Ripon | 16N | 14E | 2.18 | |
| 122 | Saint Peter | SS | Ripon | 16N | 14E | 2.27 | |
| 127 | Saint Peter | SS | Ripon | 16N | 14E | 2.36 | 32 |
| 129 | Saint Peter | SS | Ripon | 16N | 14E | 2.32 | |
| 131 | Saint Peter | SS | Seymour | 24N | 18E | 2.28 | 39 |
| | Saint Peter | SS | WI (B, 2 samples) | | | 2.66 | |
| ***Sinnipee Group*** | | | | | | | |
| 30 | Platteville | DOL | Stiles Junction | 28N | 21E | 2.67 | |
| 34 | Platteville | DOL | Ripon | 16N | 14E | 2.72 | 85 |
| 36 | Shakopee | DOL | Ripon | 16N | 14E | 2.61 | 113 |
| 38 | Platteville | (21) | Glovers Bluff | 17N | 08E | 2.65 | |
| 45 | Platteville | (21) | Glovers Bluff | 17N | 08E | 2.50 | 83 |
| 69 | Platteville | (21) | Glovers Bluff | 17N | 08E | 2.75 | |
| 118 | Platteville | SHD. | SE of Lena | 28N | 21E | 2.47 | |
| 150 | Plattev/Galena | DOL | SE of Lena | 28N | 21E | 2.67 | |
| 154 | Platteville | DOL | Seymour | 24N | 18E | 2.75 | |
| 29 | Galena | DOL | Suamico | 25N | 20E | 2.78 | |
| 33 | Galena | DOL | Brookside | 26N | 20E | 2.74 | |
| 49 | Galena | DOL | Brookside | 26N | 20E | 2.67 | 80 |
| | Plattev/Galena | DOL | WI (B, min. of 4) | | | 2.78 | |
| | Plattev/Galena | DOL | WI (B, max. of 4) | | | 2.84 | |
| | Plattev/Galena | DOL | MN (B, 2 samples) | | | 2.77 | |
| ***Maquoketa*** | | | | | | | |
| 48 | Maquoketa | DOL | Edgewater Beach | 24N | 21E | 2.59 | 108 |
| 101 | Maquoketa | SH | Wequiock | 24N | 21E | 2.73 | |
| 109 | Maquoketa | DOL | Green Bay | 24N | 21E | 2.53 | 79 |
| 119 | Maquoketa | SH | Edgewater Beach | 24N | 21E | 2.67 | |
| **SILURIAN** | | | | | | | |
| 156 | Neda | (18) | Kolb Corners | 23N | 21E | 2.91 | |
| 86 | Neda | (19) | Bayshore Cty Pk | 25N | 22E | | 110 |
| 26 | Mayville | DOL | Wequiock | 24N | 21E | 2.67 | 49 |
| 27 | Mayville | DOL | High Cliff | 19N | 18E | 2.83 | 76 |
| 35 | Mayville | DOL | Oakfield | 14N | 16E | 2.74 | |
| 39 | Stalactitic | DOL | Bayshore Cty Pk | 25N | 22E | 2.66 | |
| 31 | | DOL | Stockbridge | 19N | 18E | 2.82 | |
| 32 | Byron | DOL | Grimms | 19N | 22E | 2.73 | |
| 37 | Byron | DOL | Oakfield | 14N | 16E | 2.70 | |

| SAMPLE NO. | | LITHOLOGY | LOCALITY | T | R | DENSITY | k |
|---|---|---|---|---|---|---|---|
| 190 | Racine | DOL | Racine | 03N | 21E | 2.30 | |
| | Undivided | DOL | WI (B, min. of 14) | | | 2.70 | |
| | Undivided | DOL | WI (B, mean of 14) | | | 2.82 | |
| | Undivided | DOL | WI (B, max. of 14) | | | 2.86 | |
| **POST-PALEOZOIC (CRETACEOUS-TERTIARY?)** | | | | | | | |
| 44 | "Windrow" | DUR | Fort Mc Coy | 17N | 02W | 2.33 | |
| 125 | "Windrow" | DUR | Fort Mc Coy | 17N | 02W | 2.75 | |
| 64 | | DUR | N of Spring Green | 09N | 04E | 2.72 | 132 |
| 149 | | DUR | N of Spring Green | 09N | 04E | 2.79 | |
| 214 | | DUR | N of Spring Green | 09N | 04E | 2.76 | |
| 215 | | DUR | N of Spring Green | 09N | 04E | 2.85 | |
| 218 | | DUR | Ridgeland | 32N | 12E | 2.52 | |
| 219 | | DUR | Ridgeland | 32N | 12E | 2.58 | |

*Table includes samples from previous studies, with source indicated under locality. These samples are not shown on Figure 1 but their location may be estimated by reference to plotted samples. Township locations determined from locality descriptions in original sources.*

**Explanation of Lithology Abbreviations used in table.**

| | |
|---|---|
| AM | Amphibolite |
| ANO | Anorthosite |
| BAS | Basalt |
| BCG | Basal Conglomerate |
| CGL | Conglomerate |
| CGR | Cataclastic Granite |
| DIO | Diorite |
| DOL | Dolomite |
| DUR | Duricrust |
| FSP | Feldspar Porphyry Felsite |
| GAB | Gabbro |
| GGN | Granitic gneiss |
| GN | Gneiss |
| IF | Iron formation |
| MB | Metabasalt |
| MCG | Metaconglomerate |
| MD | Mafic Dike |
| MGR | Metagreywacke |
| MIG | Migmatite |
| MR | Metarhyolite |
| PHY | Phyllite |
| QD | Quartz Diorite |
| QMZ | Quartz Monzonite |
| QZ | Quartzite |
| SH | Shale |
| SHD | Shaly Dolomite |
| SL | Slate |
| SS | Sandstone |
| SY | Syenite |

| | |
|---|---|
| (1) | Foliated Tonalite |
| (2) | Lineated Gneiss |
| (3) | Biotite Schist |
| (4) | Dolomite Marble |
| (5) | Cataclastic Metasediment |
| (6) | Sheared Amphibolite |
| (7) | Metabasalt Fault Breccia |
| (8) | Cataclastic Metarhyolite |
| (9) | Epidote |
| (10) | Weathered Metarhyolite |
| (11) | Fragmental Rhyolite |
| (12) | Vein Quartz |
| (13) | Quartzite Breccia |
| (14) | Quartz Porphyry Felsite |
| (15) | Granite Grus |
| (16) | Amygdaloidal Basalt |
| (17) | Copper Conglomerate |
| (18) | Oolitic Iron Formation |
| (19) | Iron-rich Clay |
| (20) | Mafic Tuff |
| (21) | Dolomite Breccia, possibly impact crater fallback |

**References:**


B. Buckley, 1896, Table V, p. 400-403 and XI, p. 413-414

L1. Lane, 1911, p. 98-99

L2. Leney, 1966, p.402

H. Hinze, 1959, p. 420

C1. Cain, 1964, p. 535

C2. Carlson, 1974, p. 20-22


*About 25 samples of various lithologies from this study were also measured by David Crimley at the University of Illinois-Champaign/Urbana using a Barrington surface probe. His measurements substantially agreed with this study for samples with large values of k (over 100), but were considerably lower for those samples with low k. Values of k on the order of a few tens are at the lower limit of sensitivity of the Bison instrument used in this study, and should be viewed as accurate to an order of magnitude only. The comparison measurements suggest that k values in the range of 0 to 50 reported in this study probably are systematically high. Some high values of k in Paleozoic cover rock might be due to detrital magnetite, especially in lower Cambrian units, or to pyrrhotite in shale and carbonate units.*

*Table 2.* Density Ranges of Principal Midcontinent Rock Types

| AGE OR FORMATION | STATE | N | MIN. | MEAN | MAX. | STD | REF. |
|---|---|---|---|---|---|---|---|
| **ARCHEAN** | | | | | | | |
| gneiss, granite | MI,WI | 15 | 2.61 | 2.64 | 2.73 | .035 | D,B1,L2 |
| **PENOKEAN (1800 MY)** | | | | | | | |
| Michigamme slate | MI | 98 | 2.18 | 3.00 | 3.72 | .281 | L2 |
| Vulcan iron fm | MI | 103 | 2.88 | 3.44 | 3.88 | .247 | L2 |
| Felch Formation | MI | 49 | 2.22 | 2.72 | 3.57 | .266 | L2 |
| Randville dolomite | MI | 32 | 2.67 | 2.86 | 3.07 | .082 | L2 |
| undivided granite | WI | 114 | 2.63 | 2.69 | 2.74 | .028 | D,B1,C1,C2 |
| St. Cloud Granite | MN | 3 | 2.63 | 2.68 | 2.71 | .044 | B1 |
| mafic metavolcanic | WI | 58 | 2.70 | 2.95 | 3.06 | 100 | D,C2 |
| felsic metavolc. | WI | 7 | 2.55 | 2.67 | 2.77 | .074 | D,C2 |
| post-volc. metased. | WI | 2 | 2.79 | | 2.85 | | D |
| **RHYOLITE/EPIZONAL** | | | | | | | |
| granite terrain | WI | 34 | 2.54 | 2.64 | 2.78 | .052 | D,B1,H1 |
| **BARABOO INTERVAL** | | | | | | | |
| quartzites | WI | 28 | 2.55 | 2.65 | 2.72 | .042 | D,H1,C2 |
| **WOLF RIVER (1500 MY)** | | | | | | | |
| granitic rocks | WI | 36 | 2.57 | 2.65 | 2.72 | .025 | D,B1,C2 |
| Wausau Syenite | WI | 2 | 2.67 | 2.70 | 2.72 | .015 | D |
| anorthosite/gabbro | WI | 5 | 2.65 | 2.69 | 2.72 | .030 | D,C2 |
| **KEEWEENAWAN (1100 MY)** | | | | | | | |
| ss and cgl | MI,MN,WI | 29 | 2.36 | 2.65 | 2.89 | NA | D,B1,L1 |
| massive basalt | MI | 70 | 2.61 | 2.88 | 3.09 | | D,L1 |
| amygdaloidal basalt | MI | 51 | 2.70 | 2.85 | 3.09 | | D,L1 |
| Bayfield Ss | WI | 3 | 2.21 | 2.23 | 2.24 | .002 | D |
| **PRECAMBRIAN Z** | | | | | | | |
| Jacobsville Ss | MI | 4 | 2.16 | 2.22 | 2.29 | .035 | D,B1 |
| Jacobsville Ss | MI | | | 2.77 | | | H2 |
| **CAMBRIAN** | | | | | | | |
| undivided ss | MI,MN,WI | 82 | 1.96 | 2.39 | 2.83 | .183 | D,B1,L2 |
| Mount Simon Ss | MI | | | 2.58 | | | H2 |
| Eau Claire Fm | MI | | | 2.67 | | | H2 |
| Dresback Ss | MI | | | 2.69 | | | H2 |
| Franconia Ss | MI | | | 2.72 | | | H2 |
| Trempeleau Fm | MI | | | 2.82 | | | H2 |
| **CAMBRO-ORDOVICIAN** | | | | | | | |
| undivided ls | MO | 7 | 2.77 | 2.79 | 2.80 | .013 | B2 |
| **ORDOVICIAN** | | | | | | | |
| Prairie cu Chien | MI | | | 2.70 | | | H2 |
| Prairie du Chien | MN,WI | 14 | 2.36 | 2.69 | 2.83 | .137 | D,B1 |
| Saint Peter | MI | | | 2.63 | | | H2 |
| Saint Peter | WI | 8 | 2.19 | 2.38 | 2.66 | .054 | B1 |
| Glenwood Fm | MI | | | 2.54 | | | H2 |
| Black River Fm | MI | | | 2.71 | | | H2 |

| AGE OR FORMATION | STATE | N | MIN. | MEAN | MAX. | STD | REF. |
|---|---|---|---|---|---|---|---|
| Trenton Fm | MI | | | 2.70 | | | H2 |
| Utica Shale | MI | | | 2.71 | | | H2 |
| Sinnipee Group | MN,WI | 18 | 2.51 | 2.72 | 2.84 | .083 | D,B1 |
| Galena | IL | 17 | 2.45 | 2.66 | 2.75 | .071 | K |
| Platteville | IL | 21 | 2.52 | 2.70 | 2.79 | .068 | K |
| Maquoketa/Richmond | IL | 5 | 2.66 | 2.72 | 2.79 | .053 | K |
| Maquoketa Fm | WI | 4 | 2.54 | 2.63 | 2.74 | .0088 | D |
| **SILURIAN** | | | | | | | |
| Cataract Fm | MI | | | 2.59 | | | H2 |
| Niagaran, Edgewood | IL | 71 | 2.39 | 2.65 | 2.71 | .082 | K |
| Niagaran | MI | | | 2.71 | | | H2 |
| Evaporites | MI | | 2.16 | 2.25 | 2.34 | | H2 |
| Salina Fm | MI | | | 2.79 | | | H2 |
| Bass Island Dol | MI | | | 2.89 | | | H2 |
| undivided dolomite | WI | 22 | 2.29 | 2.77 | 2.86 | .124 | D,B1 |
| undivided ls | MO | 2 | 2.71 | 2.74 | 2.76 | .022 | B2 |
| **DEVONIAN** | | | | | | | |
| Bois Blanc Fm | MI | | | 2.64 | | | H2 |
| Dundee Ls | MI | | | 2.81 | | | H2 |
| Bell Shale | MI | | | 2.59 | | | H2 |
| Traverse Fm | MI | | | 2.71 | | | H2 |
| Antrim Shale | MI | | | 2.48 | | | H2 |
| Berea Ss | MI | | | 2.62 | | | H2 |
| **MISSISSIPPIAN** | | | | | | | |
| Sunbury Shale | MI | | | 2.45 | | | H2 |
| Coldwater Shale | MI | | | 2.63 | | | H2 |
| Marshall Ss | MI | | | 2.48 | | | H2 |
| undivided ls | MO | 10 | 2.62 | 2.69 | 2.76 | .039 | B2 |
| Burlington Fm | IL | 8 | 2.54 | 2.62 | 2.66 | .040 | K |
| St Louis/Salem Fm | IL | 17 | 2.56 | 2.68 | 2.72 | .037 | K |
| St Geniveve Fm | IL | 5 | 2.69 | 2.69 | 2.70 | .005 | K |
| Menard Fm | IL | 4 | 2.63 | 2.67 | 2.70 | .031 | K |
| **PENNSYLVANIAN** | | | | | | | |
| undivided ls | MO | 4 | 2.45 | 2.61 | 2.69 | .094 | B2 |
| undivided ss | MO | 5 | 2.457 | 2.65 | 2.70 | .097 | B2 |
| McLeansboro Fm | IL | 6 | 2.60 | 2.66 | 2.71 | .039 | K |
| undivided ss | IL | 2 | 2.70 | | 2.79 | | K |
| **JURASSIC** | | | | | | | |
| undivided | MI | | | 2.47 | | | H2 |

*All Michigan data are for the northern peninsula except for those of Hinze and others, 1978.*

**References:**

B1. Buckley, 1896, Table V, pp. 400-403 and XI, pp. 413-414
B2. Buckley and Buehler, 1904, Table VII, p. 317
C1. Cain, 1964
C2. Carlson, 1974
D. This paper
H1. Hinze, 1959
H2. Hinze and others, 1978
K. Krey and Lamar, 1925, Table 5, pp. 47-62
L1. Lane, 1911, pp. 98-99
L2. Leney, 1966

*Table 3.* Magnetic Susceptibilities of Principal Rock Types

| SUITE | MIN | 1STQ | MED | LOG MEAN | 3RDQ | MAX | N |
|---|---|---|---|---|---|---|---|
| **Archean Granites** | | | | | | | |
| and Gneisses | 61 | 120 | 341 | 278 | 881 | 1968 | 8 |
| **Penokean Rocks** | | | | | | | |
| mafic metavolc. | 91 | 115 | 168 | 188 | 188 | 2713 | 13 |
| felsic metavolc. | 65 | | 704 | 481 | | 2430 | 3 |
| post-volc. metased. | 3518 | | | 3575 | | 3633 | 2 |
| granites | 216 | | 366 | 437 | | 1057 | 3 |
| **Rhyolite/epizonal** | | | | | | | |
| granite terrain | 142 | 445 | 1179 | 1914 | 2268 | 3489 | 19 |
| **Baraboo Interval** | | | | | | | |
| quartzites | 44 | 71 | 108 | 115 | 150 | 462 | 11 |
| **Wolf River Batholith** | | | | | | | |
| granites | 53 | 162 | 266 | 229 | 329 | 500 | 13 |
| anorthosite/gabbro | 68 | 639 | | 423 | 761 | 967 | 4 |
| **Wausau Syenite** | | | 497 | | | | 1 |
| **Keeweenawan Rocks** | | | | | | | |
| volcanic | 92 | | 188 | 360 | | 2703 | 3 |
| sediments | 31 | | | 72 | | 176 | 5 |
| **Jacobsville Ss** | | | 85 | | | | 1 |
| **Cambrian sandstones** | 16 | 20 | 34 | 41 | 61 | 125 | 6 |
| **Prairie du Chien** | 52 | | | 55 | | 59 | 2 |
| **St. Peter Sandstone** | 32 | | 39 | 54 | | 126 | 3 |
| **Sinnipee Group** | 80 | 83 | | 89 | 85 | 113 | 4 |
| **Maquoketa Formation** | 79 | | | 92 | | 108 | 2 |
| **Silurian Dolomite** | 49 | | | 61 | | 76 | 2 |

## DENSITY VALUES OF COVER ROCK

The dominant rock types in the cover of the Midcontinent are sandstone, shale, and carbonate. Unfortunately, no true argillaceous shale was represented in the collection analyzed in this study (the samples labelled "shale" are actually dolomitic). The friability of shale and the fine porosity make accurate density determinations difficult. The gravimeter study of Hinze and others (1978) includes several shale units, but those units are deeply buried and probably unusually dense because of compaction.

Midcontinent sandstone tends to be mature, and the upper bound of sandstone density would normally be that of pure quartz (2.67 gm/cm$^3$). Densities are generally lower because of pore space. Only in the case of local iron cementing does the density exceed that of quartz, and such rock is volumetrically insignificant. The effects of compaction at depth are obvious from comparing the densities of surface samples of Cambrian and Ordovician sandstones with the subsurface gravimeter results of Hinze and others (1978).

Pure carbonate rock can ideally approach the density of calcite (2.71 gm/cm$^3$) or dolomite (2.85 gm/cm$^3$) but actual values are generally significantly lower because of chert, gypsum or void space. There are a few reports of carbonate rock denser than dolomite. The extra density is likely due to sulfide minerals (pyrite, marcasite, sphpalerite or galena), dense carbonate (magnesite, ankerite, or siderite) or possibly other minerals like barite. Abnormally dense carbonate is minor in volume.

On the whole, carbonate densities are near or slightly above the average density of granite. The bulk density of the Paleozoic cover, a weighted average of the high density of carbonate and the low density of clastic rock, is probably close to the average density of the basement. The low density contrast between carbonate cover and basement rock probably accounts for the weak gravity expression of many Midcontinent basement highs and lows.

## DENSITY VALUES OF BASEMENT ROCK

Densities of crystalline rock determined in this study and published in older sources contain no surprises. They agree well with values typically assumed in most gravity modelling studies. The significant exception is the Keeweenawan basalt suite. The few samples measured in this study are not definitive, but the far greater number of densities reported by Lane (1911) are, especially since they were gathered during a survey of a mine and represent a considerable section of Keeweenawan rock. These data suggest that the densities of 2.95 gm/cm to 3.0 gm/cm assumed in most models of the Keeweenawan Rift System (King and Zeitz, 1971; Hinze and others, 1982; Chandler and others, 1982) are too high, and that densities of 2.85 to 2.9 gm/cm are more nearly correct. This revision is more significant than it appears, because the gravity anomalies associated with the Keeweenawan rift are due to the density contrast between the basalt and granitic basement, and the suggested density revision reduces the density contrast on the order of 20 percent. The basalts probably become denser at depth because of compression and the closure of void spaces, but the magnitude of the change is unknown.

Keeweenawan clastic rock is reasonably close to the density of 2.3 gm/cm commonly assumed in models of the Keeweenawan rift. The clastic rock of the Keeweenaw Peninsula is systematically denser than the stratigraphically higher Bayfield sandstone.

Granitic rock generally falls in the expected range 2.65-2.70 gm/cm$^3$. Some is slightly lighter, perhaps due to weathering, wheras granitic rock close to the quartz dioirite composition range tends to be denser than 2.70 gm/cm$^3$.

Cain's (1964) study is a unique contribution to Wisconsin geophysical data and one of few detailed studies of density variations in a single rock body. Cain found that the Newingham Granodiorite varied significantly in density (from 2.66 to 2.74 gm/cm$^3$) in a complex pattern. This result suggests that, though it may be useful to use a single assumed density for granitic rock as a first approximation, detailed gravity modelling of granitic plutons may require careful density control for accurate results.

## CAUSES OF MAGNETIC SUSCEPTIBILITY IN ROCK

Grant (1985) has summarized the factors that affect magnetic susceptibility of rock. Magnetic susceptibility is largely proportional to magnetite content, and magnetite is only marginally correlatable with lithology. Factors that tend to favor high magnetic susceptibility include:

a. High total iron content;
b. Intermediate oxidation level;
c. High grade metamorphism (as Fe-silicates decompose, they often form magnetite);
d. Silica undersaturation;
e. Pelitic protolith for metamorphic rock;
f. High aluminum content in metamorphic rock;

g. Low magnesium or titanium content; and
h. High-temperature hydrothermal alteration.

Magnetite, like all minerals, competes for cations with other mineral species, and the principal competitors of magnetite are ferromagnesian silicates and non-magnetic oxides like ilmenite or hematite. Low oxidation levels favor divalent iron minerals like ferromagnesian silicates and ilmenite, whereas high oxidation levels favor hematite. Scarcity of magnesium and titanium reduces the competition for iron and thus obviously favors magnetite, and abundant aluminum (likely in pelitic metasediments) favors the formation of muscovite + magnetite as opposed to biotite.

## MAGNETIC PROPERTIES OF COVER ROCK

Quartz, calcite, and dolomite are all weakly diamagnetic, with magnetic susceptibilities on the order of -2 x $10^{-6}$ cgs units but even miniscule amounts of magnetic minerals will produce positive susceptibilities.

Highly oxidizing conditions, including low-temperature alteration and weathering, destroy or prevent the formation of magnetite. Magnetic susceptibility in weathered specimens (75, 183) is dramatically lower than in unweathered equivalent rock. Because of the effects of oxidation and weathering, most interpretations of magnetic maps treat sedimentary cover rock as non-magnetic. The samples measured in this study support that assumption.

Keeweenawan sedimentary rock and quartzite of the Baraboo interval also shows low magnetic susceptibilities, rarely much over 100 x $10^{-6}$ cgs units. This low susceptibility is to be expected given the lack of Keeweenawan metamorphism and the purity of the Baraboo interval quartzite, although a few give high readings, apparently due to metamorphic magnetite. The post-Penokean Marshall Hill Conglomerate is remarkable for its high magnetic susceptibility.

## MAGNETIC PROPERTIES OF BASEMENT CRYSTALLINE ROCK

Although basalt is commonly assumed to be magnetic, it can actually have quite low magnetic susceptibility. Part of the Keeweenawan rift system has weak magnetic signatures, even though strong gravity signatures indicate the presence of large amounts of mafic rock. The magnetic susceptibilities found for the few Keeweenawan basalts measured are extremely variable. On textural grounds, specimens 152 and 179, with low susceptibilities, solidified in an oxygen-rich environment and were perhaps subaerially weathered after eruption, whereas 235, with a very high susceptibility, is part of a sequence of very thick, massive flows where conditions might be more favorable for magnetite formation and preservation.

The metamorphic rock of central Wisconsin consists of Archean gneissic basement and Proterozoic mafic and silicic metavolcanic rock. Both rock suites have susceptibilities between 100 and 500 x $10^{-6}$ cgs units, with some indication of higher susceptibility in Archean rock. Cataclastic rock, abundant in Wisconsin, often has high susceptibility (7, 9, 13, 58, 77).

Several suites of granitic rock are represented: Penokean synorogenic rock about 1800 Ma, post-orogenic rhyolite and epizonal granite about 1700 Ma, and the 1500 Ma Wolf River Bathlith. The granitic rock is rather more magnetic than the metamorphic basement rock, with the rhyolite-epizonal granite suite showing very high susceptibilities of over 2000 x $10^{-6}$ cgs units.

This pattern of low susceptibility in coarse batholithic rock, increasing in fine-grained epizonal and volcanic rock, is similar to that reported by Allinghan (1964) for rhyolite and granite from the St. Francois Mountains of southeastern Missouri, the only large exposure of a widespread 1420-1500 Ma rhyolite-granite terrane that covers much of the southeastern Midcontinent. He reported susceptibilities of 0-1000 x $10^{-6}$ cgs units for coarse-grained granite, over 2000 x $10^{-6}$ cgs units for fine-grained granite near the roof of the batholith, and over 3000 x $10^{-6}$ cgs units for rhyolite. A rough inverse correlation between magnetic susceptibility and grain size or depth of emplacement thus appears to exist in both Wisconsin and Missouri, in rock suites of two different ages. The petrologic explanation for this pattern is not known, but it may be due to a lower degree of oxidation in the coarser (and presumably deeper) granite.

## CONCLUSIONS

Most major geophysical studies of Midcontinent basement have assumed density and magnetic susceptibility values for Midcontinent rock that are in good accord with the values found in this study. Keeweenawan basalt, however, appears to be less dense than assumed in most studies. Rhyolite and epizonal granite are very magnetic and probably account for much of the magnetic fabric of the southeastern Midcontinent, as well as the high magnetic anomalies across southern Wisconsin.

It is obvious that many more values are needed, especially from basement drill holes. Density and magnetic susceptibility are neither difficult nor expensive to determine, and not even very time-consuming once a systematic procedure is established. Undoubtedly there are a large amount of data scattered throughout the literature and in theses that would be very useful if assembled.

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# Origin and Maturation of the Organic Matter in the Middle Ordovician Guttenberg Member of the Decorah Formation of Southwestern Wisconsin

*J.M. Blabaum*[1]

## ABSTRACT

*Samples from an unmineralized exposure of the Guttenberg Member of the Decorah Formation were characterized by hydrocarbon source rock parameters; organic carbon (22.16 percent), sulfur (0.13 percent), hydrogen index (989.2), oxygen index (17.0), asphaltenes (47.28 percent), resins (22.45 percent), and hydrocarbons (30.27 percent), a carbon preference index of 1.113, a pristane to phytane ratio of 0.93, and a pristane to C17 ratio of 0.08. The bitumen extract contained 20.75 percent paraffins and napthenes and 9.52 percent aromatics. Alkanes ranged from C15 to C36, with C15 to C19 predominating. The carbon preference ratio and the presence of C30 favor marine algae as the parent material for the organic matter. The transformation ratio (0.02), derived from the hydrogen and oxygen indices, reveals thermal immaturity and a history of shallow burial (less than 700 meters) within the basin. Temperatures of 50 °C to 90 °C are indicated by conodont color and kerogen color from the same sample. A geothermal gradient, in the neighborhood of 3.2 °C to 9.0 °C per 100 m, was calculated. All of these indicators of maturation are irreversible and thus represent average basinal temperatures experienced since Middle Ordovician deposition.*

## INTRODUCTION

Maturation of organic matter provides an indicator of the thermal history of sedimentary basins (Dow, 1977). Organic geochemistry is being used increasingly to provide insight into the origin of organic matter, its depositional environment and the diagenetic history of the enclosing formation. In this paper the characterization of organic matter in the Middle Ordovician Guttenberg Member of the Decorah Formation is used to estimate average basinal temperatures in southwestern Wisconsin. No direct estimates of basinal temperatures have been made previously. The results of the organic characterization are supplemented by kerogen and conodont thermal maturation color analysis (Peters and others, 1977, Epstein and others, 1977 and Rejebian and others, 1987) which serves as an independent indicator of maximum temperatures experienced by the samples.

Descriptions of the general geology and stratigraphy of the study area in addition to study of another organic-rich member (Quimbys Mill; fig. 1) below the Guttenberg can be found in work by Heyl and others (1959), and Hatch, Heyl and King (1985).

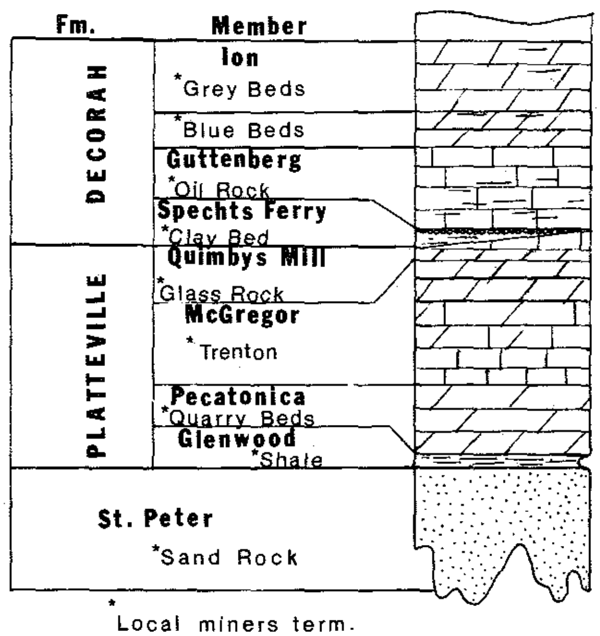


***Figure 1.*** Stratigraphic section of southwestern Wisconsin showing both the organic-rich Quimbys Mill Member and the Guttenberg Member.

[1]Geosciences Department, University of Wisconsin-Platteville, Platteville, WI 53818; now at Wisconsin Department of Natural Resources, Dodgeville, WI 53533

Two previous studies have been conducted on the geochemical nature of the Guttenberg organic matter and its relationship to the zinc-lead deposits in southwest Wisconsin. Gize and Hoering (1980) and Fowler and Douglas (1984) both used samples taken from mineralized areas. The sample taken for Rock-Eval analysis (see Tissot and Welte, 1984) in this study was taken from an unmineralized zone so that the basinal status of maturation might be determined.

## PROCEDURES

This study utilized organic geochemistry, kerogen and conodont thermal maturation color to evaluate the degree of organic maturation in the study area. Samples for this study were taken from three locations shown in figure 2. The unmineralized road cut from Highway 61 southwest of Platteville, Wisconsin was sampled for the Rock-Eval analysis. (The Rock-Eval process was developed by Espitalie' and others, 1977 and is amply described by Tissot and Welte, 1984.) Two other samples came from known mineralized zones (fig. 2) and were used for microscopic comparison with the sample from the unmineralized site. All samples were examined by transmitted and reflected light microscopy.

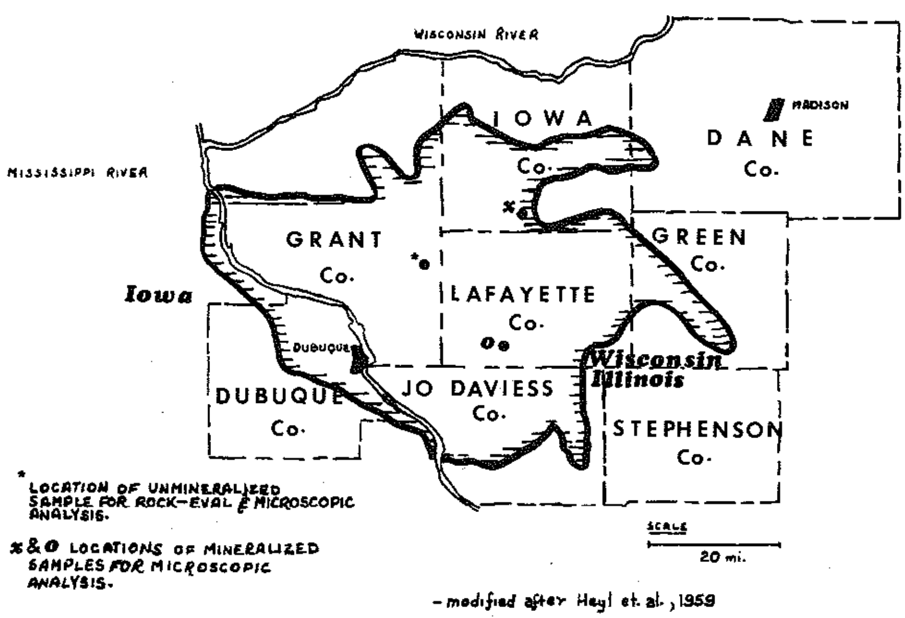


*Figure 2.* Location map of the study area (modified from Heyl and others, 1959). The sample for the Rock-Eval analysis was taken near Platteville, Wisconsin. The mineralized samples for microscopic comparison were from Mineral Point and the Shullsburg area.

## RESULTS

The kerogen from the unmineralized site is high in organic carbon (22.16 percent) and low in total sulfur (0.13 percent). Of the bitumen extracted, there was a larger fraction of the paraffin-napthene groups (20.75 percent) compared to the aromatics (9.52 percent). The resins made up 22.45 percent and the asphaltenes 47.28 percent of the bitumen.

The C15 Saturate Hydrocarbon Analysis (table 1) gave a range of C15 to C36 with the range weighted towards the lighter alkane compounds between C15 and C20. The major components are C15 at 26.10 percent and C17 at 28.74 percent, along with C16 at 17.82 percent, C19 at 11.67 percent, C18 at 4.34 percent, phytane at 2.57 percent and pristane at 2.39 percent. The pristane-to-phytane ratio of 0.93 shows phytane as slightly dominant. The carbon preference index (CPI) of 1.113 shows little odd-carbon chain bias.

The hydrogen index (HI) and the oxygen index (OI) reflect the nature of organisms that were the source of the organic matter and the degree of maturity that organic matter has reached. The HI and OI for the Guttenberg were found to be 989.2 and 17.0 respectively. When compared with other oils (fig. 3), the Guttenberg indeed appears to be a Type-I algaly derived kerogen that is very immature and has high genetic potential due to a very high hydrogen to carbon ratio. The Guttenberg transformation ratio (table 1), when plotted on the Depth versus Transformation Ratio Plot (Espitalie' and others, 1977, fig. 4) shows the Guttenberg basin has experienced a depth of no more than 700 m.

Deposition of the region's zinc-lead deposits was a hydrothermal event with temperatures of the ore-forming solutions reaching temperatures near 200 °C (Giordano and Barnes, 1981).

Differences between mineralized and unmineralized samples occur both in kerogen color and degree of carbonate dissolution. The southern-most sample, from a mineralized zone, displays only moderate carbonate dissolution with much of the carbonate intact and a red-brown kerogen color which corresponds to maturation temperatures between 100 °C and 200 °C (see Peters and others, 1977). The two northern samples, one mineralized and the other unmineralized (used for the Rock-Eval tests) show nearly complete carbonate dissolution with the kerogen occurring as grain coatings around insoluble quartz-silt and clay particles. Both of these samples also display the same light yellow kerogen colors (Peters and others, 1977) that indicate maturation temperatures below 100 °C. Conodonts were discovered in the unmineralized sample and their translucent light red-brown color (Epstein and others, 1977) indicates temperature (50 °C to 90 °C) in accordance to that shown by the surrounding kerogen.

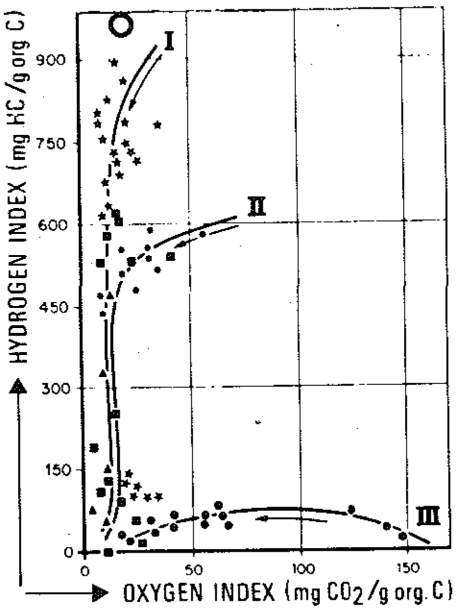


***Figure 3.*** Plot of HI and OI of Guttenberg compares the Guttenberg to other oil. It shows the Guttenberg is a Type-I immature alginite. Guttenberg is shown as an open circle (modified from Espitalie' and others, 1977).

These permanent, cumulative and irreversible thermo-color maturation temperatures can be used in conjunction with the maximum depth of burial (which was derived from the transformation ratio, see fig. 4) to estimate the maximum range of thermal gradients for the Guttenberg basin since the Middle Ordovician. This can be done by using the standard linear gradient model of $y = mx + c$ (Asquith and others, 1982) where y is the temperature indicated by the unmineralized Guttenberg sample (which ranges between 50 °C and 90 °C; x is the maximum depth of burial of the Guttenberg samples (700 m); and c is the mean surface temperature which is assumed to be near 27 °C (80 °F) for a shallow platformal carbonate environment from the Middle Ordovician to the present. The variable m then corresponds to the geothermal gradient for the Guttenberg which ranges between 3.2 °C per 100 m and 9.0 °C per 100 m. Schlumberger (1986) suggests a modern global range between 1.09 °C per 100 m and 2.92 °C per 100 m.

## DISCUSSION

The geochemical fossil indicators present in the unmineralized sample are characteristic of a relatively pure marine oil with no terrestrial plant contamination. The proof lies in the alkane range of C15 to C36 with a distinct weighting of occurrence between C15 and C20. In particular, the dominance of C15 and C17 definitely points to marine algae as the pre-cursor of the present kerogen (Tissot and Welte, 1984).

Algal origins are also indicated by the pristane and phytane proportions and ratio. Pristane tends to occur in low concentrations in marine benthonic algae (Laminarioles) and generally occur in planktonic algae and in assemblage with phytane in zooplankton. Therefore, when phytane predominates over pristane, a reducing type of environment is indicated, such as might be expected for marine benthonic algae (Tissot and Welte, 1984). Moldowan (1985) also has shown that presence of C30 steranes (found in this study) indicate organic matter from a marine environment. Likely contributing organisms during the Ordovician were Acritarchs, Chlorophyceae (green algae) and Cyanophyceae (blue-green algae). The environment at this time is typically ascribed to a shallow platformal carbonate-type of marine environment.

The hydrogen (HI) and oxygen (OI) indices are both related to the contributing organisms and do indicate that the Guttenberg kerogen is algally derived organic matter with excellent hydrogen saturation. This is evidenced by the hydrogen index of 989.2 and oxygen index of 17.0.

*Table 1.* Summary of the Rock-Eval analysis performed by Getty Research on the Guttenberg sample.

**C-15+ Saturate Hydrocarbon Analysis**

| n-Alkane | % |
|---|---|
| C15 | 26.10 |
| C16 | 17.82 |
| C17 | 28.74 |
| Pristane | 2.39 |
| C18 | 4.34 |
| Phytane | 2.57 |
| C19 | 11.67 |
| C20 | 1.39 |
| C21 | 0.80 |
| C22 | 0.76 |
| C23 | 0.83 |
| C24 | 0.37 |
| C25 | 0.33 |
| C26 | 0.24 |
| C27 | 0.27 |
| C28 | 0.20 |
| C29 | 0.16 |
| C30 | 0.12 |
| C31 | 0.11 |
| C32 | 0.12 |
| C33 | 0.08 |
| C34 | 0.15 |
| C35 | 0.34 |
| C36 | 0.11 |

*CPI* for C25 to C32 is 1.113
Pristane/Phytane is 0.93
Pristane/C17 is 0.08

**Group Type Analysis of Bitumen Extract**

| | |
|---|---|
| rock weight | 10.4009 grams |
| Hydrocarbons (%) | 30.27 |
| Paraffin-Naphthene (%) | 20.75 |
| Aromatics (%) | 9.52 |
| Resins (%) | 22.45 |
| Asphaltenes (%) | 47.28 |

**Rock-Eval pyrolysis**

| | |
|---|---|
| % organic carbon | 22.16 |
| HI (hydrogen index) | 989.20 |
| % sulfur | 0.13 |
| OI (oxygen index) | 17.00 |
| TR (transformation ratio) | 0.02 |
| TMAX (maximum temperature) | 440°C |

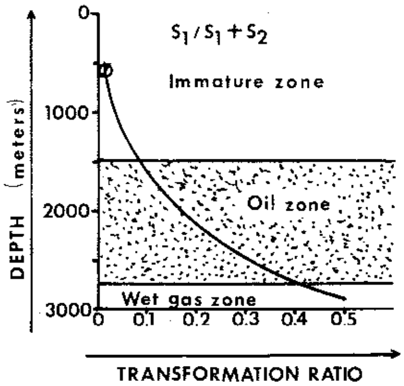


*Figure 4.* Plot of the transformation ratio vs. depth shows the Guttenberg to be at the top of the curve and at the corresponding depth of 700 m. Guttenberg is shown as an open circle (modified from Espitalie' and others, 1977).

The transformation ratio, when plotted on Espitalie's (1977) plot (fig. 4), seems to indicate a maximum burial depth for the Guttenberg of 700 m. When this information is linked to the indicated maturation temperatures of the organic matter, the resulting maximum thermal gradient range for the Guttenberg basin (3.2 °C and 9.0 °C per 100 m, fig. 5) is obviously high in comparison to the accepted modern range (1.09 °c to 2.92 °C per 100 m) for geothermal gradients.

The present geothermal gradient for an adjacent basin (Illinois Basin) ranges between 2.19 °C and 3.65 °C per 100 m (Barrows and Cluff, 1984). The Michigan Basin (Cercone, 1984) has a present gradient of 2.5 °C per 100 m. Barrows and others observed that Damberger (1971, 1974) found Pennsylvanian coal ranks in Illinois could not be accounted for by present burial depths and temperatures and suggests either deeper burial or that the geothermal gradient increased at some time in the past. Cercone also was troubled with unusually high paleogeothermal gradients (3.5 °C to 4.5 °C per 100 m) in the Michigan Basin.

It is premature to assert that these three basins experienced hydrothermal solutions from the same source responsible for the southwest Wisconsin zinc-lead district. A new theory by George deV. Klein and Albert T. Hsui concerning the origin of cratonic basins (1987) does indeed link the thermal subsidence histories of the Illinois, Michigan and Williston basins with other intercratonic basins throughout the world. They date the formation of these basins as around 550 to 500 Ma and suggest they formed concurrently with the rifting and breakup of a late Precambrian supercontinent (Klein and Hsui, 1987). Evidence from the Guttenberg samples, however, does indicate that hydrothermal solutions affected organic matter in and adjacent to mineralized zones as well as areas without even microscopic evidence

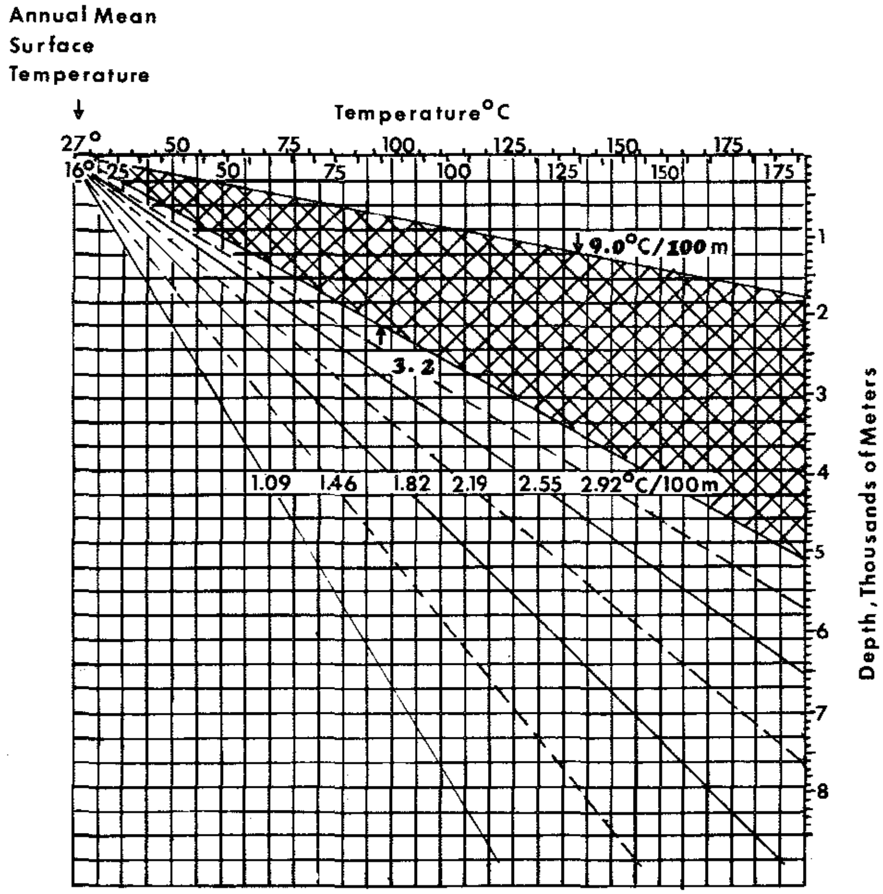


*Figure 5.* Standard Schlumberger, Inc. geothermal gradient plot showing the normal range of geothermal gradients currently found worldwide. The Guttenberg paleo-geothermal gradient in the cross-hatchered area shows a higher than normal range. Workers in both the Illinois and Michigan basins report paleo-geothermal gradients within this range for some time since the Middle Ordovician (modified from Schlumberger, Inc., 1986).

of mineralization. This global event may be the cause for the curious Guttenberg paleo-geothermal gradient and perhaps the actual mineralization of the region as well.

## CONCLUSIONS

1. The Guttenberg oil rock is a Type-I immature kerogen.

2. The Guttenberg Member experienced average regional temperatures of no more than 100 °C as confirmed by conodont and kerogen thermo-color development.

3. Probable burial depths are estimated at about 700 m from the transformation ratio.

4. Maturation of the hydrocarbon is locally affected by varying temperatures of the mineralizing solutions as evidenced by the 200 ° C maturation temperature in the southernmost mineralized sample land the 50 °C to 90 °C maturation temperatures in the samples to the north.

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Geological and Natural History Survey
University of Wisconsin–Extension
3817 Mineral Point Road
Madison, WI 53705
Phone: (608) 262-1705

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