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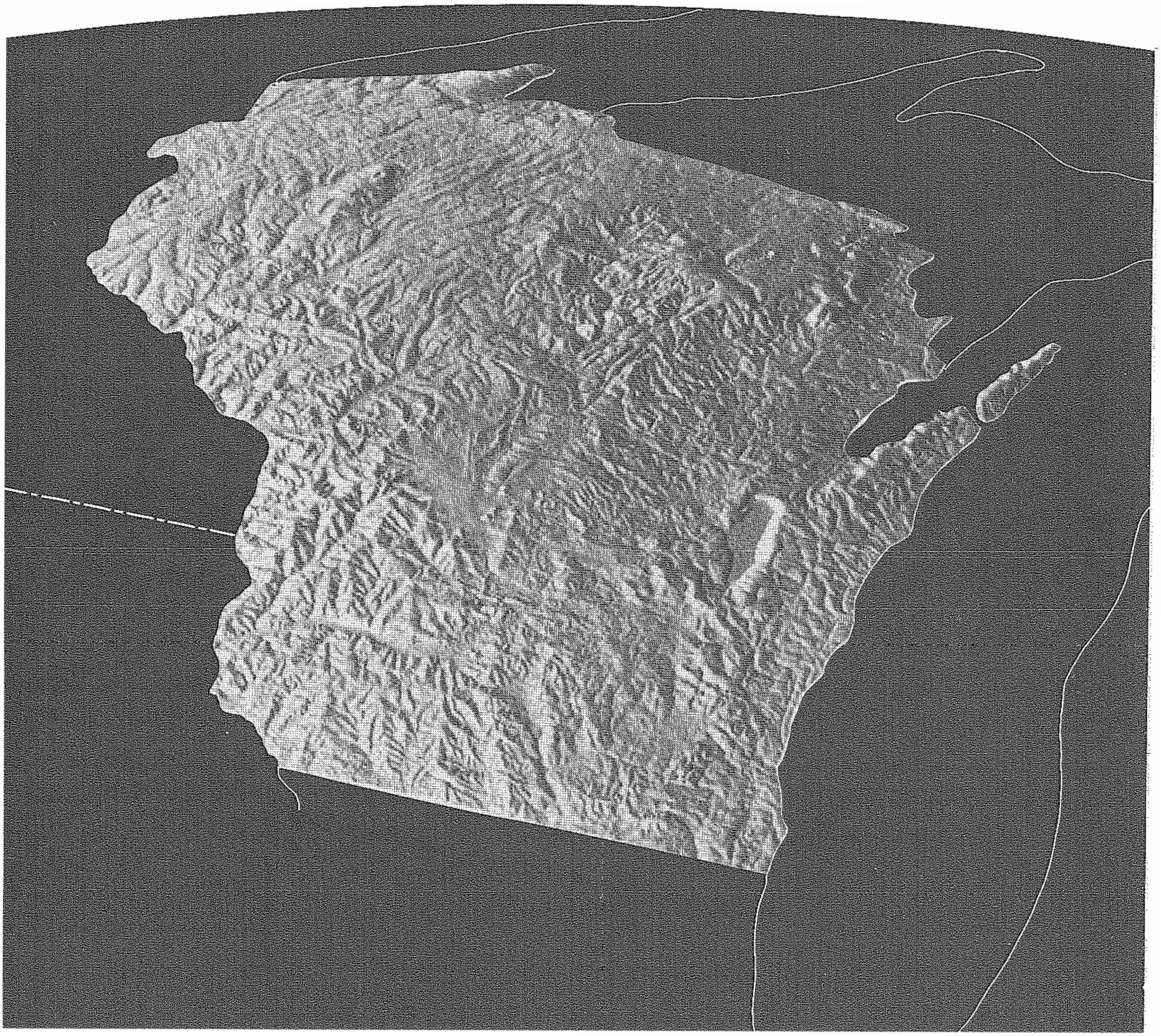
PROTEROZOIC BARABOO INTERVAL IN WISCONSIN



Volume 10

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PROTEROZOIC BARABOO INTERVAL IN WISCONSIN

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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 with minimal time in review and production to 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 Geological and Natural History Survey will publish the paper as funds 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 systems of large libaries, or lost in the musty drawers of an open-file.

In conjunction with the Seventeenth Annual Meeting of the North-Central Section of the Geological Society of America in Madison in late April and early May of 1983 Jeff Greenberg and Bruce Brown coordinated a symposium on Baraboo-interval quartzite. Following the meeting they led a field excursion to central Wisconsin. The proceedings of this symposium include Wisconsin Geological and Natural History Field Trip Guide Book Number 8 on the Middle Proterozoic to Cambrian geology of central Wisconsin and Geoscience Wisconsin Volume 10, Proterozoic Baraboo interval in Wisconsin. Jeffrey K. Greenberg and Bruce A. Brown of the Wisconsin Geological and Natural History Survey coordinated the symposium and field excursion and Greenberg coordinated the papers presented in this volume.

Submission of manuscripts relating to Wisconsin geology is encouraged. 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
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FOREWORD

The studies in this volume and other recent articles on the Baraboo interval (rock deposited between 1,760 Ma and 1,630 Ma in the Lake Superior region) indicate an accelerated interest in this package of metasedimentary and associated igneous rock. For many decades the Baraboo syncline has been an area studied for its classical structural geology. Modern investigations by R.H. Dott, Jr. and I.W.D. Dalziel have served to continue this emphasis and develop an appreciation of Baraboo sedimentology. Baraboo-type metasedimentary rock is also considered an economic resource providing large reserves of road and building material, railroad ballast, abrasive-grinding pellets, and potential supplies of iron ore and catlinite (pipestone). More speculative potential resources include gold and other metallic ores. Studies of the distribution and composition of the rock have practical and not simply academic implications.

This interest resulted in a special session of the North-Central Section meeting of the Geological Society of America in 1983. This dedicated volume, Proterozoic Baraboo interval in Wisconsin, is a follow-up to that meeting and offers a more formal outlet for new ideas, even where controversial and not yet firmly established. In certain cases the studies are ongoing and will later result in more detailed reports.

Included in this Baraboo-interval collection are papers on the stratigraphy and correlation of units (Brown), the description and interpretation of quartzite in the subsurface of Iowa (Anderson and Ludvigson) and from the Baraboo syncline (Geiger), the geology of Flambeau Quartzite from northwestern Wisconsin (Campbell), paleomagnetic data from Baraboo and Waterloo (Kean and Mercer), an analysis of cleavage from Baraboo (Jank and Cambray) and tectonic interpretation of Baraboo deformation (Hempton, Gordon, and Kirschner), interpretation of jointing problems in the Rib Mountain quartzite of north-central Wisconsin (Dockstader), the chemistry of granite which intrudes quartzite at Hamilton Mound in central Wisconsin (Taylor and Montgomery), and the association of quartzite breccia with intrusions and metasomatism of Baraboo interval units throughout (Greenberg).

In the last few years our own work has stressed the significance of the Baraboo interval in terms of tectonic environment prevailing about 1,760 to 1,500 Ma. The Baraboo-type metasedimentary and associated granitic rock are analogous to anorogenic suites of similar age recognized world wide. In general the anorogenic activity was superimposed on the region of transition between Archean craton and early Proterozoic terranes.

There are some diverse perspectives on tectonism and deposition expressed in the present volume. In some cases the lack of agreement may be philosophical, in that different models are proposed for essentially the same data. However, limitations such as sparse sample material from drill cores and/or incomplete awareness of available data also appear to have led to differing interpretations. For example, metamorphism and deformation of quartzite and rhyolite in southern Wisconsin have been attributed to a collisional orogeny about 1,630 Ma. The location of the collisional zone (suture?) is supposedly in northern Illinois or southernmost Wisconsin. This is based on just a few bits of evidence, including: (1) the occurrence of amphibolite-facies metamorphism at Waterloo Wisconsin, lower metamorphic-grade rock at Baraboo to the northwest, and extremely low-grade Barron and Sioux Quartzite futher to the northwest; (2) deformation is more intense at Waterloo and Baraboo than in the Barron and Sioux; and (3) 1,630 Ma-old Rb/Sr isochron ages are regionally persuasive even though they are not the original age of any known unit in Wisconsin. In Greenberg (this volume) and more specifically in Brown (this volume) there are summaries of published and some previously unpublished data that do not support the idea of metamorphism and deformation of decreasing intensity away from the orogenic front. In addition the paper by Hempton, Gordon, and Kirschner introduces the totally novel hypothesis of 1,000 Ma-Keweenawan deformation of Baraboo Quartzite. It is hoped that the innovation contained in this volume will stimulate more investigations. The reference lists at the end of each paper provide almost all of the additional information available on the Baraboo interval. Readers should also be aware of recent studies and publications by the Minnesota Geological Survey on the Sioux Quartzite (see Southwick, 1984 in Brown's paper).

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THE BARABOO INTERVAL IN WISCONSIN

by

Bruce A. Brown[1]

ABSTRACT

Recent geologic mapping in Wisconsin has uncovered significant new data on the nature, extent, and tectonic setting of rock deposited during the Baraboo interval (1,760 to 1,500 Ma). It is now known that argillite, chert, conglomerate, micaceous quartzite, and low-grade iron-formation are important constituents of the sequence in addition to the well-known red quartzite. These other lithologies are widely distributed in the upper part of the sequence in southern Wisconsin. They commonly occur directly over older basement rock in central Wisconsin, suggesting that the thick, red quartzite may not have been deposited in this area.

Dating of intrusive rock and local stratigraphic relationships suggest that Baraboo interval sedimentation may have begun as early as 1,760 Ma, contemporary with the episode of felsic volcanism which followed the Penokean orogeny. Current evidence suggests that Baraboo interval sediment may have been deposited in a complex environment with facies distribution in part controlled by local uplift and basin development. The Baraboo interval sedimentary rock was deformed and metamorphosed probably around 1,630 Ma. The interval ended with emplacement of the Wolf River and related granitic rock (1,500 Ma).

INTRODUCTION

The term Baraboo interval was first used by Dott (1983) to designate the period of time (1,450 to 1,750 Ma) during which the well-known red quartzite, (Baraboo, Sioux, Barron) of the southern Lake Superior district was deposited. Greenberg and Brown (1983a, 1984) redefined Baraboo interval as a period of anorogenic igneous and sedimentary activity which followed the end of the Penokean orogeny (1,760 to 1,860 Ma) and ended with the emplacement of the Wolf River batholith (1,500 Ma). The time span of the Baraboo interval is based on dating of basement rock and igneous rock intrusive into Baraboo interval metasedimentary rock. Preliminary paleomagnetic studies (W. F. Kean, unpublished data) have shown promise as a means of distinguishing Baraboo interval rock from older and younger (Keweenawan) sedimentary rock.

Dott originally defined the Baraboo interval on the basis of the apparent age and lithologic similarity of the quartzite units. The Baraboo type quartzite, named for the best exposed and studied section at Baraboo, Wisconsin, is a unique rock type in the Precambrian of the Lake Superior region. Recent field studies in Wisconsin (Greenberg and Brown 1983b, 1984) have identified many new exposures of a diverse suite of metasedimentary rock. This suite includes argillite, bedded chert, micaceous and conglomeratic quartzite, and low-grade iron-formation. The rock overliesbasement ranging from 2,800 Ma to 1,760 Ma and are intruded by anorogenic igneous rock ranging from 1,760 Ma to 1,500 Ma which were apparently deposited along with the quartzite during this interval. This paper will briefly describe the distribution and character of the Baraboo interval rock in Wisconsin, and discuss their depositional history, correlation, and tectonic setting.

DISTRIBUTION OF BARABOO INTERVAL ROCK

With the exception of the Sioux Quartzite of southwestern Minnesota and eastern South Dakota, the best examples of Baraboo interval rock are exposed in Wisconsin. Other quartzite which is known from deep wells in Iowa, Nebraska, and Kansas may also be Baraboo interval rock (Anderson and Ludvigson, this volume). The informal type area for this rock is the Baraboo region of Sauk and Columbia Counties, Wisconsin (Dalziel and Dott, 1970). Similar red quartzite has long been known to occur in other areas of the state, most notably the Barron Quartzite in northwestern Wisconsin, the Waterloo Quartzite of Dodge and Jefferson counties, the McCaslin and Thunder Mountain Quartzites in northeastern Wisconsin, and an extensive area of red quartzite, argillite and iron formation in the subsurface of southeastern Wisconsin. Recent mapping in central Wisconsin (Greenberg and Brown, 1983b, 1985) has led to the recognition of numerous outcrops of Baraboo interval rock that were previously unknown or poorly known. The locations of known exposures and subsurface occurrences of these rocks in Wisconsin are shown in figure 1.

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ROCK OF THE BARABOO AREA

The Precambrian rock of the Baraboo area has been known and studied for many years (Irving 1872, 1877; Weidman 1904; Dalziel and Dott 1970; Dott 1983). The Baraboo Quartzite crops out in an elliptical pattern which outlines a syncline overturned to the south (fig. 2). Rhyolite and fragmental felsic volcanic rock which have been correlated with similar rock in the Fox River Valley and the subsurface of southeastern Wisconsin (Smith, 1978) have generally been interpreted as basement for the quartzite, although the actual contact is not exposed. The volcanic rock is part of an extensive anorogenic granite-rhyolite terrane (Smith, 1978) and has been dated elsewhere by U/Pb zircon method at 1,760 Ma (Van Schmus and others, 1975; Van Schmus, 1980). The stratigraphic succession at Baraboo consists of an undetermined thickness of rhyolite overlain by at least 1500 m of Baraboo Quartzite, 100 m of Seeley Slate, a dominantly argillaceous unit, and 300 m of the Freedom Formation, an iron-rich dolomite (Dalziel and Dott, 1970). Leith (1935) described two younger formations, the Dake Quartzite (65 m) and the Rowley Creek Slate. The Rowley Creek Slate is known only from early iron exploration records.

The lower 1500 meters of section at Baraboo consists of the typical red quartzite described by Dott (1983). Weidman (1907) described an extensive basal conglomerate developed on top of the underlying rhyolite. Most of the coarse fragmental rock such as those exposed at the Lower Narrows and Caledonia Church (fig. 2) are now known to be volcanic breccia belonging to the rhyolite sequence (Smith, 1978). The lowermost beds of the quartzite are locally pebbly, but do not contain much recognizable fresh volcanic debris. However, several of Weidman's thin sections taken from several hundred meters above the base of the section in the north range, (Greenberg and Brown, 1983) contain what are probably volcanic rock fragments. The exact nature of the basal contact is not known, although outcrops of both units exposed less than 15 meters apart suggest an abrupt change from rhyolitic fragmental rock to mature quartz sandstone.

The sedimentology of the quartzite at Baraboo has been discussed by Henry (1975) and Dott (1983) who concluded that sedimentary characteristics are most consistent with "braided fluvial and/or littoral and inner shelf deposition." This environment graded upward into a marine environment at the time of deposition of the argillite and carbonate of the Seeley and Freedom Formations. Dott (1983) presents convincing sedimentological arguments for the similarity and probable correlation in the broad sense of the Barron and Sioux Quartzites, with the quartzite at Baraboo.

Argillaceous rock is most abundant in the upper part of the Baraboo section. However thin argillite beds are known to occur throughout the lower 1500 meters of quartzite. Beds of argillite up to several centimetres in thickness occur near the base of the quartzite at Baxter Hollow on the south limb of the syncline, and increase in abundance and thickness upwards to the Highway 12-Skillet Creek road cut (Greenberg and Brown 1983; Hempton, this volume) where they are abundant and are commonly up to 2 m thick. No other outcrops of quartzite occur on the south limb to the north of Skillet Creek (fig. 2). The core of the syncline is composed of the Seeley Slate and younger formations.

The Seeley Slate was described by Weidman (1904) from core drilled during iron exploration and from mine workings. The Seeley is commonly gray with bedding marked by slight variation in color and texture. Weidman reported that a well developed slaty cleavage at high angle to bedding is commonly present. Wiedman described the contact with the underlying Baraboo quartzite as conformable and the contact with the overlying Freedom Formation as gradational.

The Freedom Formation was extensively explored for commercial iron ore in the early 1900s. The only complete description of this unit is that of Weidman (1904). Schmidt (1951) described the subsurface geology of the Freedom Formation based on Weidman's notes and iron company records. A petrographic description of core samples from the Cahoon Mine site south of Baraboo is discussed by Geiger (this volume). The Freedom Formation consists of a variety of lithologies, ranging from relatively pure slate and dolomite to ferruginous slate and dolomite and to chert. The iron ore was largely hematite. Much of the waste rock consisted of jasper (ferruginous chert), slate and ferruginous dolomite. In describing the Freedom formation in a letter Van Hise stated that the more siliceous types closely resembled any other Lake Superior type iron-formation. The dolomitic members of the Freedom Formation are unique among Baraboo interval rock in Wisconsin, but the ferruginous slate, bedded ferruginous chert, and lean cherty iron formation are similar to rock which occurs in association with Baraboo-type quartzite elsewhere, particularly in central Wisconsin and in the subsurface of southeastern Wisconsin

WATERLOO--SOUTHEAST WISCONSIN SUBSURFACE

Quartzite crops out along Stony Branch Creek and the Crayfish River east of the town of Waterloo (fig. 3). The southernmost outcrops, north of Lake Mills, are purple to red quartzite with well developed cross bedding. This quartzite is identical in appearance, texture, and composition to quartzite in the lower 1500 m of the Baraboo section. At Portland quarry and in the area east of Waterloo the quartzite is interbedded with argillite and pebble conglomerate. At Portland (fig. 3) all lithologies show evidence of recrystallization and metamorphism. The quartzite becomes gray in color and takes on a coarse mosaic texture. The scattered 10- to 20-cm thick argillite beds contain andalusite and iron-rich chloritoid. Geiger and others (1983) identified two stages of metamorphic mineral growth, an early mineral assemblage consisting of muscovite, quartz, chlorite, andalusite, chloritoid with or without plagioclase and hematite which they

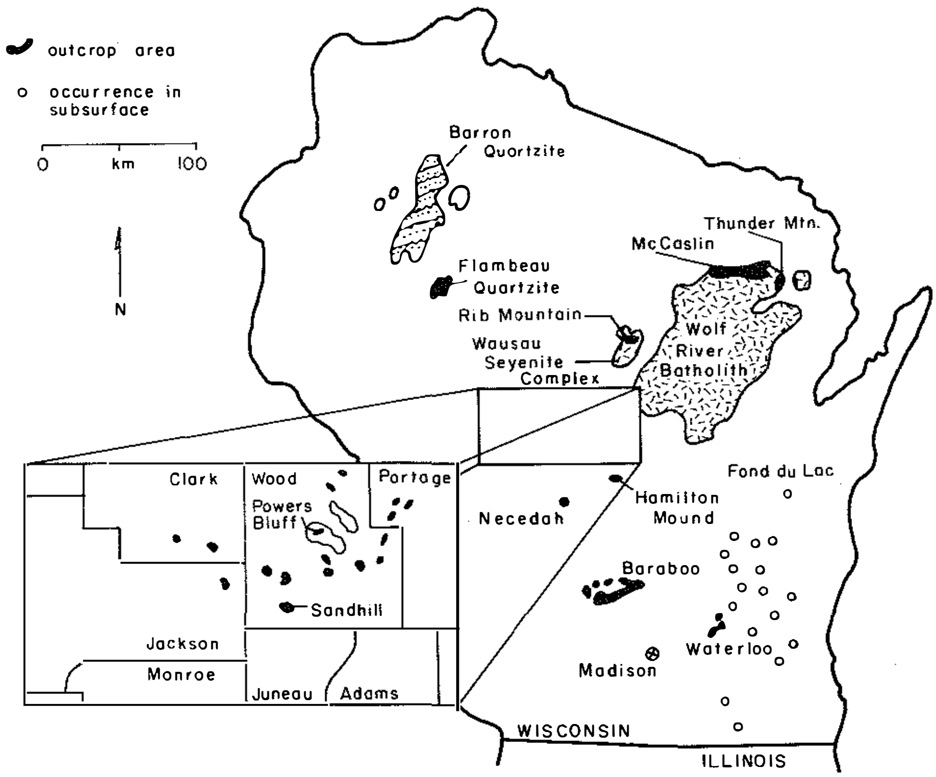

Figure 1.--Map showing locations of exposure and subsurface occurrences of Baraboo interval rock in Wisconsin.

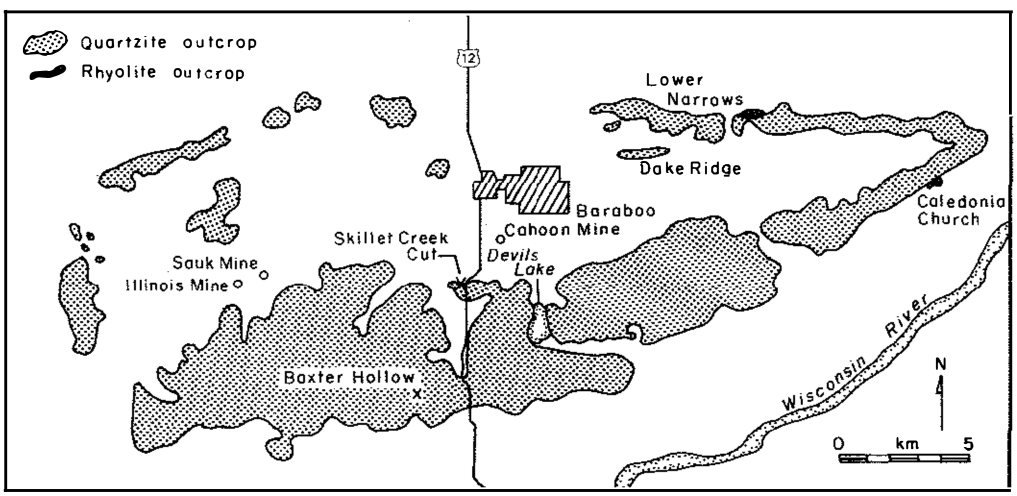

Figure 2.--Map of the Baraboo area showing location of outcrop and mines.

attribute to static growth at greater than 3.8 Kb and 550 °C. They also identified a later overprinting assemblage of coarse chlorite and muscovite. Both metamorphic events were static recrystallization which occurred after deformation.

A structural study of the Waterloo area by Brandon (1983) identified two sets of folds. Early northeast plunging mesoscopic folds were reoriented by later broad, easterly plunging, macroscopic folds. The second folds of Brandon are major structures several kilometers wide, which are defined in the subsurface by basement topography and aeromagnetic trends. These major folds are of similar magnitude and orientation and may be equivalent to the Baraboo Syncline. Brandon's study agreed with Geiger and others (1983) that folding was followed by static metamorphism.

Pink-granitic pegmatite dikes up to 1 m across cut quartzite exposed on Rocky Island in the Crayfish River northeast of Waterloo (fig. 3). The pegmatite has been dated at 1,440 Ma by the Rb/Sr method (Aldrich and others, 1959). This age for the pegmatite and an Ar/Ar release age of 1,450 to 1,500 Ma on mafic dikes in drill core taken at Portland Quarry (Guidotti, unpublished data) establishes a minimum age for the quartzites that coincides with the age of emplacement of the Wolf River batholith in central Wisconsin. Thermal metamorphism associated with intrusion of dikes at around 1,500 Ma may account for the static metamorphic events found by Brandon (1983) and Geiger and others (1983). Further evidence of intrusive activity is found in the northernmost outcrop of this area near Reeseville where hydrothermal quartz breccia zones similar to those described at Necedah and Hamilton Mounds (Greenberg this volume) cut the quartzite.

Quartzite is encountered in deep wells throughout southeastern Wisconsin, (Brown and Greenberg 1983; Thwaites 1931, 1940). The dominant material is pink to purple quartzite with minor amounts of argillite and iron formation. Petrographically, these rocks are similar in all aspects to the rocks of the Baraboo section. The quartzite generally defines buried topographic highs similar to the Baraboo Range, which commonly influenced distribution of facies within the overlying Cambrian sandstone (Thwaites 1940). A prominent bedrock high surrounds the Waterloo outcrops (Smith, 1978). Quartzite is the dominant basement rock over a broad area as far north as Fond du Lac (fig. 1).

HAMILTON MOUND

The quartzite at Hamilton Mound in Adams County (fig. 1) is a typical Baraboo-type quartzite which has been intruded by granite (Greenberg and Brown, 1983b; Taylor and Montgomery, this volume). Away from the granite the quartzite has the color, texture, and sedimentary structures typical of the Baraboo Quartzite. Near the granite contact an extensive zone of alteration is developed in which the two rock types grade into each other (fig. 4) over an interval of 10 to 12 m (Greenberg and Brown, 1983b). A structural map by Ostrander (1931) identified two synclines and an intervening anticline in the Hamilton Mound outcrop. The macroscopic folds trend N. 80° W., roughly parallel to the axis of the Baraboo syncline.

The granitic intrusion is discussed by Taylor and Montgomery (this volume). Two notable features seen at Hamilton Mound and other locations, where Baraboo-type quartzite is intruded by granite, are the breccia zones consisting of angular quartzite fragments in a white vein quartz matrix and the alteration of the red color to a greenish-yellow by reduction of iron (Greenberg, this volume).

Preliminary U/Pb zircon studies (W.R. Van Schmus, unpublished data) suggest an age of 1,760 Ma for the granite. If this age is confirmed, it is significant in that this provides a minimum age for the quartzite. An age of 1,760 Ma would imply a time of deposition closely following the end of the Penokean orogeny and contemporary with rhyolite volcanism in south-central Wisconsin.

NECEDAH

Precambrian quartzite is exposed in a large bluff and as several smaller outcrops within the town of Necedah in Juneau County (fig. 1). The quartzite in the smaller exposures is gray to pink in color and is nearly identical in composition and texture to other Baraboo-type quartzite. Recrystallization has destroyed most primary textures and structures with the exception of poorly preserved cross bedding. Recent excavations in a large quarry near the west end of the bluff have exposed some ferruginous argillite, which appears to strike east west and dip steeply (80°) to the north. Most structures in the quarry exposure have been destroyed by brecciation and quartz veining associated with a granitic intrusion of unknown age.

Large pieces of intrusion breccia (fig. 5) which consist of quartzite fragments in a granitic matrix have been observed in the quarry. The quartzite fragments are green in color indicating iron reduction similar to that which occurred near the intrusive contact at Hamilton Mound. The actual intrusive contact is not exposed due to back filling and stockpiled materials, but the in-

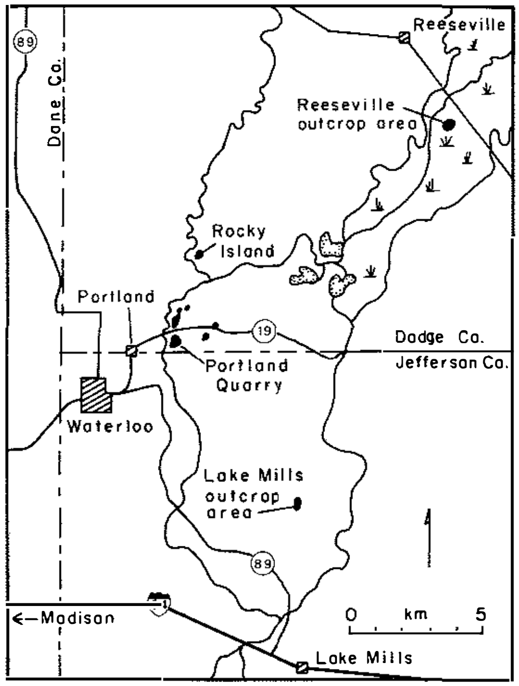

Figure 3.--Map of the Waterloo area showing location of quartzite outcrop.

trusion breccia along with extensive quartz breccia zones, (Greenberg, this volume) and hydrothermal minerals such as muscovite and feldspar confirm the presence of an intrusion. Early drilling records (S. Wiedman, field notes) reported diorite and granite bedrock in wells located within 500 m of the Necedah outcrops.

The lithologic similarity of the quartzite to that of Baraboo, and the similarity of the ferruginous argillite to that associated with other Baraboo interval quartzite argues for their inclusion in this group.

CENTRAL WISCONSIN

Quartzite, chert, argillite, and ferruginous metasediment which closely resemble rock in the Baraboo section are abundantly exposed in Wood and Portage counties (Greenberg and Brown, 1985). Isolated exposures of quartzite, chert, and brecciated rhyolite with extensive quartz veining also occur in Clark and Jackson counties. The best exposures occur in the area of Powers Bluff County Park and in the Sandhill Wildlife Refuge at Babcock (fig. 1).

Powers Bluff is an elongate ridge consisting of fine-grained quartzite (fig. 6) best described as a bedded chert, (Greenberg and Brown, 1983b). Bedding is commonly 5 to 10 cm thick, and the rock varies from nearly white to dark bluish-gray to red in color. In thin section (fig. 7) the chert is aphanitic, consisting of fine-grained quartz with hematite as the dominant opaque mineral. The chert is similar in texture to some of the chert described by Weidman (1904) from the Freedom Formation. The common occurrence of this distinctive chert in glacial deposits and the abundance of small outcrops suggests that this unit was once widespread in central Wisconsin.

A small quarry to the southeast of Powers Bluff (fig. 9) exposes bedded chert and up to 20 m of argillite. The argillite is finely laminated (5 to 10 mm) and varies in color from brown to greenish gray to black. The argillite is similar in appearance, texture, and composition to the ferruginous argillite at Necedah and in the Freedom Formation at Baraboo. Large-scale bedding, marked by color variation, is on the scale of 40 to 50 cm. Preliminary X-ray analysis (S.W. Bayley, unpublished data) indicates that the black zones are high in amorphous carbon, and that and that the red zones are highly ferruginous. Quartz and kaolinite are dominant materials in the argillite, suggesting a low grade of metamorphism. The contact between the chert and argillite is gradational with thin 5 to 10 cm beds of chert and isolated chert nodules occurring in the argillite.

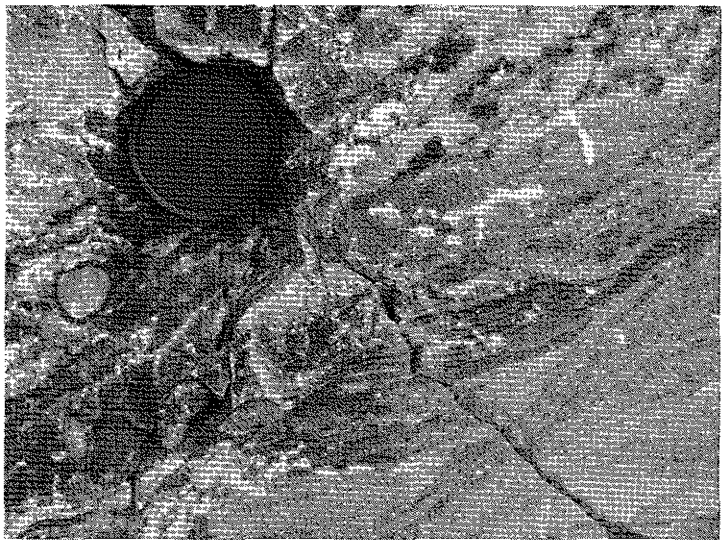

Figure 4.--Distorted bedding and inclusion with reaction rim in quartzite from the intrusive contact at Hamilton Mound. Lens cap is 5 cm in diameter (from Greenberg and Brown, 1983b).

In a road cut southeast of the quarry (fig. 9) chert and argillite occur in association with more typical, red, Baraboo-type quartzite. Quartzite ranging from quartz arenite to micaceous pebbly conglomerate occurs in numerous small outcrops within 2 to 3 km of Powers Bluff. This association and the similarity of the argillite to the Seeley and Freedom Formations suggests that these rocks are Baraboo interval metasediment.

No detailed structural work has been done in the Powers Bluff area, and no reliable top indicators have been found to establish relative ages of the chert, argillite and quartzite. Bedding strikes N. 60° to 80° W., parallel to the trend of the bluff. Dips vary from near vertical to 80° north or south. Local variation in strike suggests gentle folding about steeply plunging axes. A distinct cleavage trending N. 20° E. is evident throughout the argillite units. These rocks have been folded but show evidence of only very mild metamorphism. Preservation of fine lamination in the argillite (fig. 8) and the lack of significant recrystallization in the chert suggests low temperature and pressure. The low-grade rock of the quartzite-chert-argillite sequence stand in marked contrast to the high-grade (amphibolite faces) Penokean and Archean basement rock exposed nearby (Greenberg and Brown, 1985). The basement/cover relationship is apparent in that Baraboo-type rocks occur on knobs and hills with the high-grade rock in low areas, often only tens of meters apart.

Baraboo-type, red quartzite occurs in intimate association with felsic metavolcanic rock at the Sandhill Wildlife Refuge (Greenberg and Brown 1983). The volcanic rock consists of flow-banded rhyolite and felsic fragmental rock containing rhyolite clasts in a matrix of quartz-sericite phyllite. The large outcrop on North Bluff consists of massive and fragmental rhyolite (fig. 10). A small exposure in a quarry 0.5 km to the east of the bluff consists of rhyolite in direct contact with quartzite. The quartzite varies from a micaceous red quartzite containing up to 25 percent sericite to a pebble conglomerate. These micaceous quartzite, which are common in the region, are similar in texture and composition to the Dake Quartzite in the upper part of the Baraboo sequence. The quartzite and rhyolite appear to be in conformable contact with no evidence of a major unconformity. A common foliation suggests that both underwent deformation and low-grade metamorphism at the same time.

Micaceous quartzite grading to quartz/sericite phyllites occurs at several other localities in Wood and Portage Counties, (Greenberg and Brown, 1983b, 1985). Notable occurrences are in a quarry northwest of Vesper and in two quarries near Veedum, southwest of Pittsville (Greenberg and Brown, 1983b), where the micas are high in chromium (S.W. Bayley, unpublished data). The high chromium content suggests local derivation from chromite-bearing ultramafic rock which occurs in the basement rock of the area. At Vesper the micaceous pebbly quartzite overlies granite dated at 1,900 Ma (Van Schmus, 1984) and grades upward into a clean buff to red quartzite similar in appearance to other Baraboo interval quartzite. The Rb/Sr isotopic system in the granite shows evidence of disturbance at 1,500 Ma (C. Montgomery, unpublished data).

Figure 5.--Quartzite breccia cemented by white vein quartz in quarry at Necedah Bluff, Juneau County (from Greenberg and Brown, 1983b).

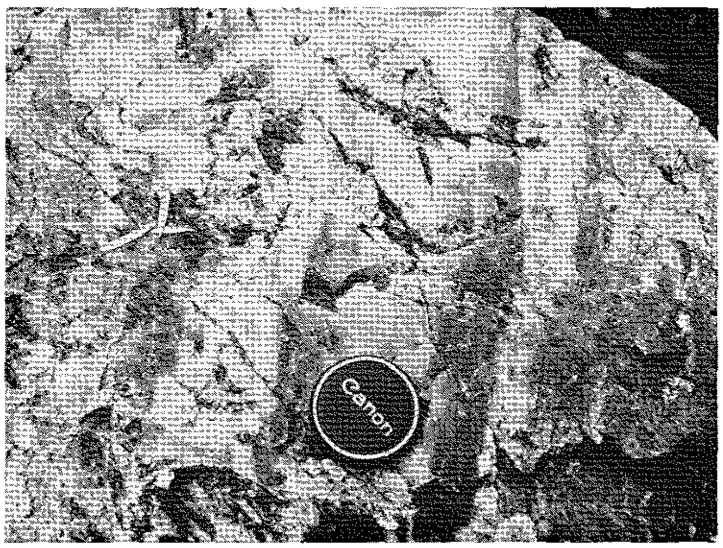

Figure 6.--Bedded chert at Powers Bluff County Park, Wood County. Lens cap is 5 cm in diameter (from Greenberg and Brown, 1983b).

Little systematic work has been done on the structural geology of the Baraboo interval rock of central Wisconsin. Reconnaissance work has shown that the rock is folded, generally about shallow plunging, roughly east-west trending axes, a similar trend to the Baraboo syncline and the macroscopic folds at Waterloo and Hamilton Mound. The relatively simple style of deformation and low metamorphic grade suggest that the rock is younger than the highly deformed and metamorphosed Archean and Penokean basement rock. Retrograde metamorphism apparent in the older rock of the area (R. Maass, unpublished data) may be related to the event which deformed the Baraboo interval rock, but no definite connection has yet been established.

RIB MOUNTAIN

The quartzite at Rib Mountain and smaller outcrops on nearby Mosinee and Hardwood Hills (fig. 1) are actually large xenoliths in the 1,500 Ma Wausau complex (Dockstader, this volume). Ansfield (1967) described ripple marks and cross bedding, but original structures and textures are largely obliterated by recrystallization. Laberge and Myers (1984) argued that these quartzite xenoliths represented basement quartzite caught up in the syenite of the Wausau complex. They based their reasoning on metamorphic textures and the presence of sillimanite which they attributed to regional metarmorphism in the early Proterozoic. In the absence of other criteria to establish age this interpretation cannot be accepted. Sillimanite is known to occur in other Baraboo-equivalent quartzite associated with 1,500 Ma intrusive rock (Olson, 1984; Greenberg and Brown, 1984; Maass and others, 1985). Further evidence is presented by Dockstader (this volume) which suggests that the quartzite blocks in fact rotated downward into the intrusive granite and syenite. An additional suggestion of Baraboo affinity is the occurrence of meta-catlinite layers in the Rib Mountain quartzite.

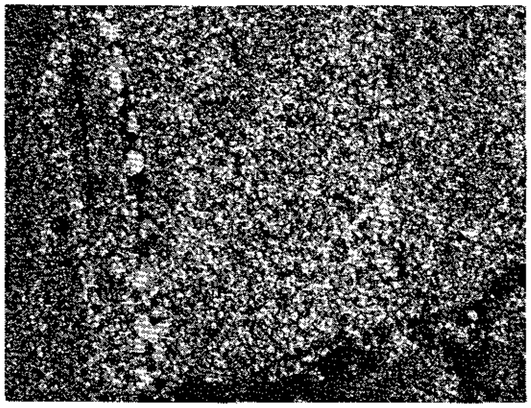

Figure 7.--Photomicrograph of bedded chert, Powers Bluff County Park. Rock consists almost entirely of fine quartz grains with minor iron oxides. Field of view is 8 mm in long dimension (from Greenberg and Brown, 1983b).

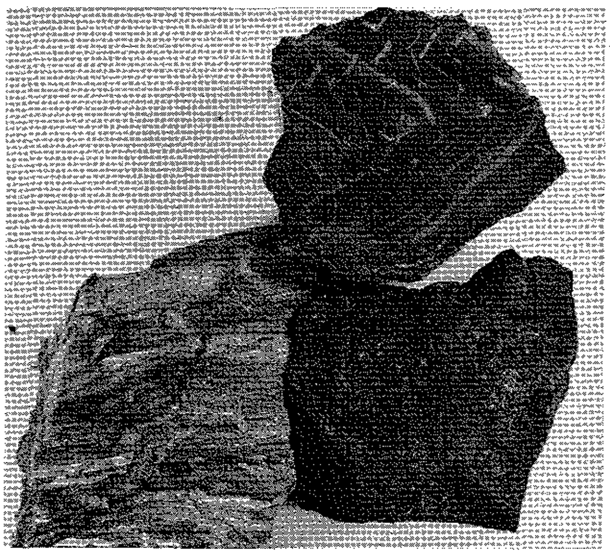

Figure 8.--Hand specimens of argillite interbedded with chert at Powers Bluff. Clockwise from top: folded ferruginous argillite; carbonaceous argillite; and banded, kaolin-rich argillite. Specimens are approximately 10 cm in diameter (from Greenberg and Brown, 1983b).

McCASLIN QUARTZITE

The McCaslin Quartzite of northeastern Wisconsin is similar in lithology and texture to other quartzite included in the Baraboo interval group. The McCaslin Quartzite overlies metavolcanic rock of presumed Penokean age, and is intruded by rock of the 1,500 Ma-old Wolf River batholith. Olson (1984) described the petrology of this unit at McCaslin Mountain, Thunder Mountain, and Deer Lookout Tower in Forest and Oconto Counties (fig. 1). Quartzite is the dominant lithology, although a thin basal pebble conglomerate is present. Total thickness is at least 1220 m. Olson concluded that the McCaslin Quartzite was deposited in a braided fluvial environment, similar to that proposed for the Baraboo Quartzite by Dott (1983). True argillite is present only as thin, phyllitic partings in the McCaslin Quartzite.

As at Rib Mountain, original texture in the McCaslin has commonly been obliterated by contact metamorphism from the Wolf River batholith which produced metamorphic sillimanite, garnet, chloritoid, and andalusite in the more argillaceous beds.

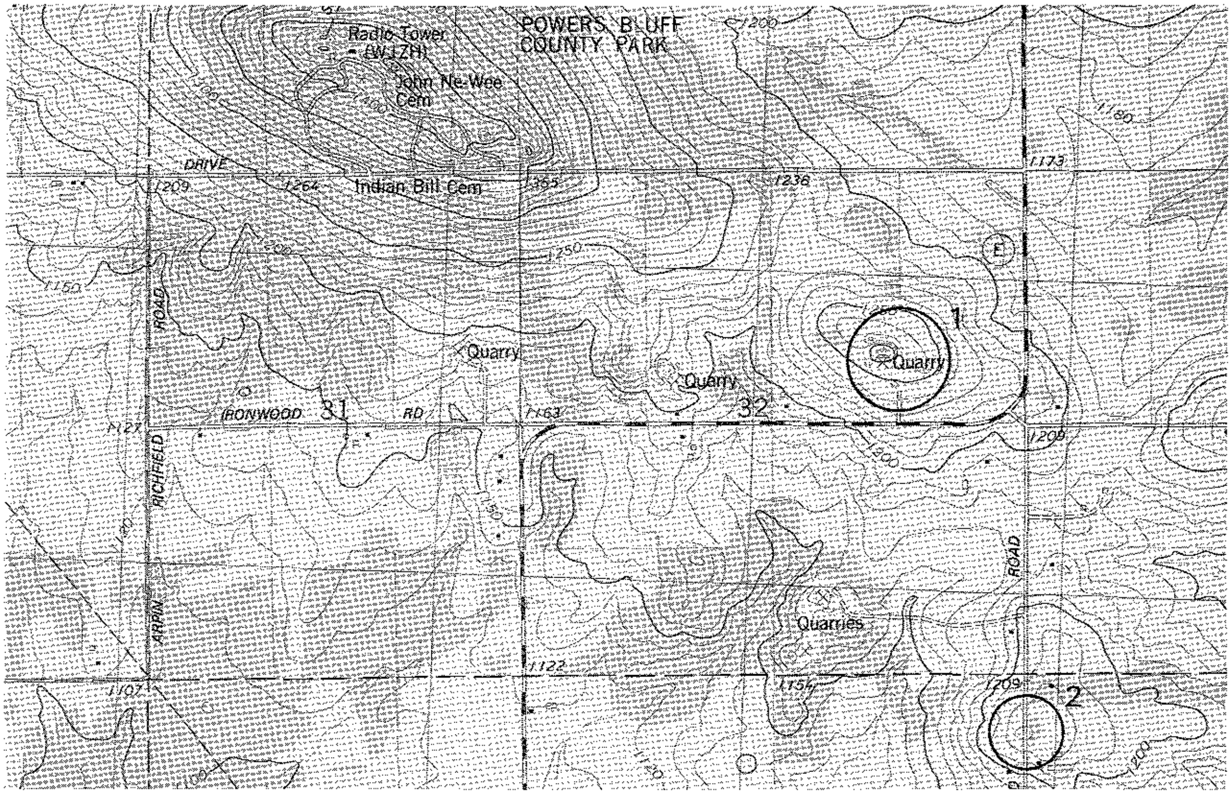

Figure 9.--Map showing location of outcrop in the Powers Bluff area, Wood County: 1, location of quarry in which argillite/chert sequence is exposed; and 2, location of road cut which exposes chert, argillite, and quartzite (from Greenberg and Brown, 1983b).

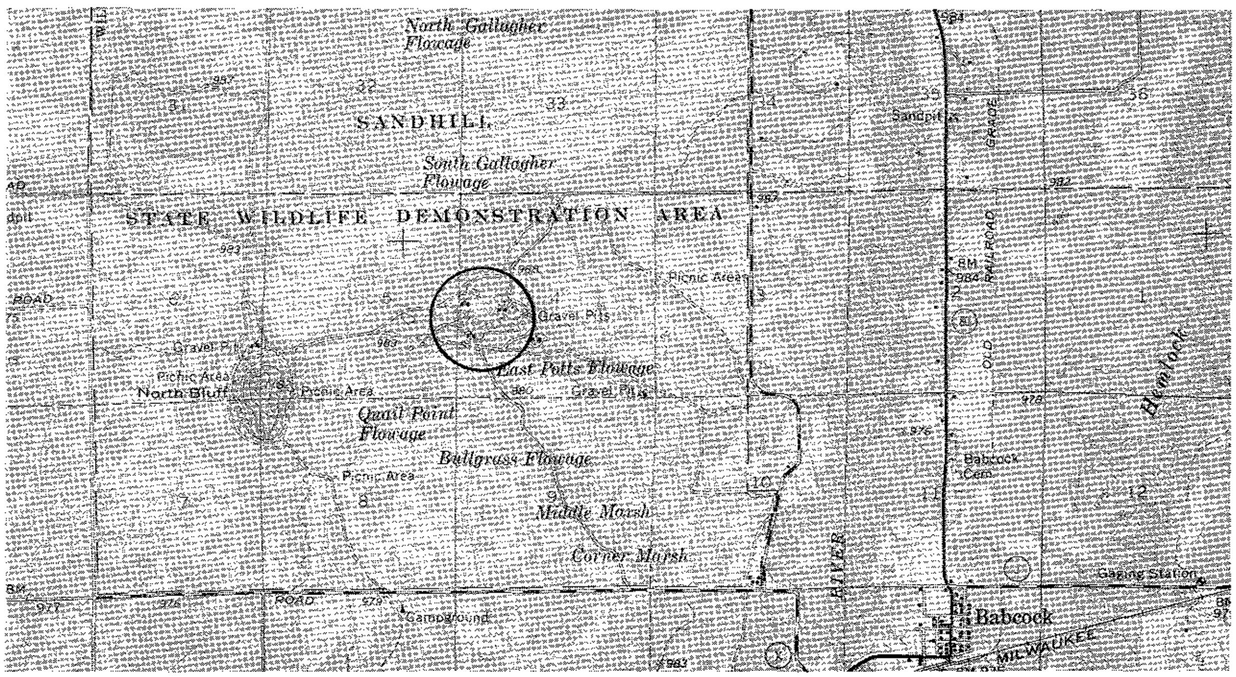

Figure 10.--Map of the Sandhill Wildlife Refuge area, Wood County. Quartzite and rhyolite are exposed in the gravel pits. North Bluff is a large outcrop of rhyolite (from Greenberg and Brown, 1983b).

BARRON AND FLAMBEAU QUARTZITES

The Barron and Flambeau Quartzites are the westernmost exposures of Baraboo-type quartzite in Wisconsin. The Barron covers a large area of Barron, Rusk, Sawyer, and Washburn Counties. The Flambeau crops out in Chippewa and Rusk Counties (fig. 1).

Johnson (1985) described the Barron Quartzite as a fine-grained, red to buff, quartz arenite with little variation except for scattered zones of quartz pebbles, a thin basal conglomerate, and 10- to 20-cm thick beds of catlinite in the upper part. Johnson concluded on the basis of a sedimentologic study that the Barron Quartzite was deposited in a braided fluvial environment superseded by a marine shelf environment, similar to that proposed for other Baraboo-type quartzite in Wisconsin (Dott, 1983; Greenberg and Brown, 1984). A shallow marine environment for the upper part of the Barron is suggested by bimodal, bipolar paleocurrent directions. Johnson and Dott both found little or no evidence for folding or metamorphism in the Barron, although northeast-trending faults are common. The Barron overlies granitic and volcanic rock which may be in part as young as 1,760 Ma (M.G. Mudrey, Jr., unpublished data) and is intruded by diabase dikes of presumed Keweenawan age.

The Flambeau Quartzite, which forms a prominent ridge, is located only 30 km to the south of the outcrop edge of the Barron (fig. 1) but is different in several significant ways (Campbell, 1981; this volume). The Flambeau is predominantly a quartz arenite, similar to the Barron and other Baraboo types, but is more conglomeratic, containing pebble conglomerate throughout. Campbell concluded that the Barron and Flambeau Quartzites were lithologically similar and probably shared a common source terrane. He also notes that the Flambeau is folded, whereas, the Barron is flat lying and nearly undeformed. The Flambeau is nearly 800 m thick. Only 200 m of Barron section has been documented; the original thickness of this unit may have been greater.

The Barron and Flambeau present a problem in direct correlation, which is typical of other Baraboo interval sediments. Lithologic and sedimentological evidence argue for correlation in the broad sense. What is known of the basement rock suggests that both are post-Penokean, implying that they belong in the Baraboo interval. It is agreed by most workers that the Barron and Flambeau are probably both members of the Baraboo-interval group of metasedimentary rock, but whether they are different sedimentary sequences deposited in that interval or whether they are equivalent lateral facies affected differentially by the tectonic event that deformed the Baraboo-interval rock, is a question yet to be resolved (Greenberg and Brown, 1984).

STRATIGRAPHY AND SEDIMENTATION

Discontinuous exposure and the lack of fossils or other stratigraphic markers makes correlation among the units in the strict sense impossible. However, the lithologic similarity within the group, and the relationship of the rock to dated units strongly supports the conclusion that they are related and that they were deposited during the roughly 250 m.y. period now called the Baraboo interval.

Available sedimentological data indicate that the characteristic red-quartzite sequences were largely fluvial deposits. Detailed studies of the Baraboo and Barron Quartzites (Dott, 1983; Johnson, 1985) suggest that the fluvial environment later changed to shallow marine conditions. At Baraboo argillite and chemical sediment dominate the upper part of the section. In the Barron, argillite is much less common, but bimodal and bipolar paleocurrent directions suggest a tidal influence in the upper beds (Johnson, 1985). The Baraboo and Barron areas are the only places where a sufficiently complete section is preserved for the fluvial/marine transition to be observed. All other exposures in Wisconsin are relatively small, and probably represent only a partial stratigraphic section. The majority of quartzite occurrences described consist of the fluvial facies, which evidently was present over much of Wisconsin at one time.

The marine facies (argillite, carbonate, chert) are known only from Baraboo, central Wisconsin, and the subsurface of the southeast. The marine facies may have been present at the top of the section throughout the entire area of quartzite deposition, but they have not been preserved. The similarity of chert and argillite in central Wisconsin to rock of the Freedom and Seeley Formations at Baraboo suggests that this facies was deposited as far north as Wood and Portage counties. Their absence in the northern areas of quartzite (Rib Mountain, McCaslin, and Barron) suggests that they were either not deposited or were removed by erosion. The transition from fluvial to marine conditions in the Barron Quartzite without significant argillite deposition suggests that these lithologies might never have been deposited in the north and west.

Greenberg and Brown (1984) also suggested a model which involved multiple sequences separated by major unconformities. In this model (fig. 11c) the chert, volcanogenic sediment and argillite of central Wisconsin could be part of an early sequence, eroded and later covered by marine transgression as in model 1. Later uplift and erosion would have removed the overlying upper sequence in this area. However, no major regional unconformities have been identified in the sequence. There is no evidence for more than one major episode of deformation and metamorphism in the rock. The only evidence in support of a multiple sequence is the suggested unconformity between the Dake and Freedom Formations at Baraboo and the difference in deformational intensity between the Barron and Flambeau Quartzites. Neither of these is, however, sufficient to establish two depositional, tectonic sequences in the absence of other data. The unconformity below the Dake Quartzite is not well documented and may represent only a local change in source or fluctuation in sea level. The Flambeau-Barron problem may reflect only local variation in deformational intensity, a question which cannot be resolved at present without more detailed structural work and dating. Considering the limited preservation of the rock and the length of the Baraboo interval, the possibility of multiple sedimentary cycles certainly cannot be eliminated at this time.

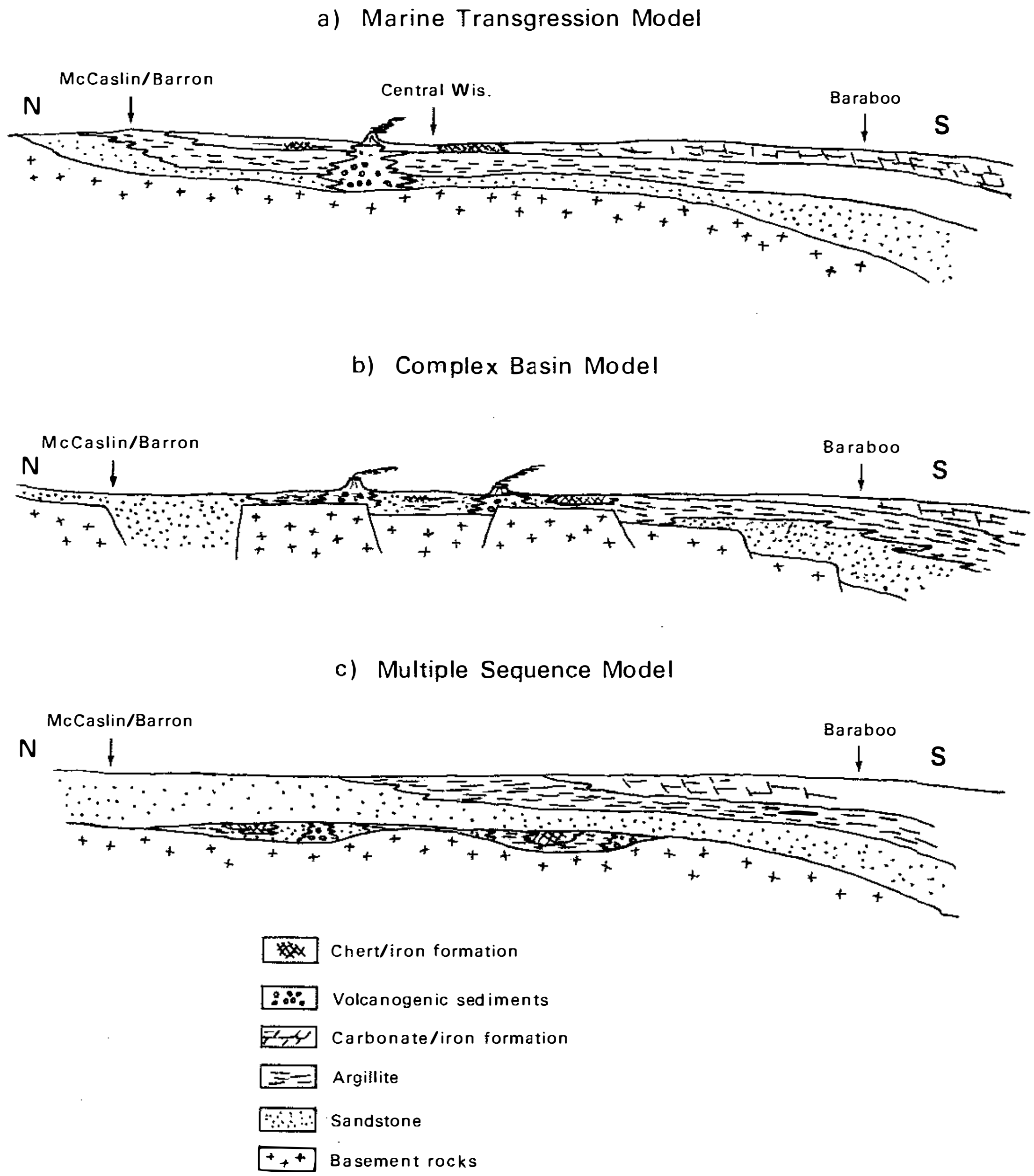

Figure 11.--Possible models for deposition of the Baraboo interval rock of Wisconsin (from Greenberg and Brown 1983b).

TECTONIC SETTING OF THE BARABOO INTERVAL

Greenberg and Brown (1984) presented three possible depositional models for Baraboo interval sediment in Wisconsin. The first model (fig. 11a) consisted of a simple south to north marine transgression, which resulted in a fining-upward sequence. This is essentially the model suggested by Dott (1983) based on his analysis of stratigraphic and sedmentological data from the Baraboo, Barron, and Sioux Quartzites. This model implies increased subsidence to the south with resulting accumulation of a thicker section in the Baraboo area. It is also consistent with the absence of the marine facies in the far north. However, the presence of argillite and chert deposited directly over older basement in much of central Wisconsin, particularly in the absence of a significant thickness of quartzite, suggests that it may be too simplistic.

A simple model that is consistent with existing data is the complex basin model (fig. 11b) of Greenberg and Brown (1983, 1984). Marine transgression in combination with local topographic highs within the basin of deposition, possibly the result of block faulting, controlled facies distribution. This model would provide for the accumulation of thick fluvial quartzite in the north and south and the relative absence of fluvial quartzite in central Wisconsin, where marine facies were deposited directly on top of basement highs at the time of maximum transgression. A similar model involving fault controlled basins has been suggested for the Sioux Quartzite by Morey (1984). This model allows for restricted basins of accumulation for the McCaslin, Barron, Rib Mountain, and other fluvial quartz-arenite sequences in the north, and the similar Baraboo sequence in the south. The positive area in central Wisconsin did not receive these deposits, but was covered by locally derived micaceous and arkosic quartzite with minor local volcanogenic contributions as in the case of the interbedded rhyolite at Sandhill. After the fluvial quartzite sequence was deposited in local basins the positive area in central Wisconsin was covered by the transgressing marine environment. Argillite and chemical sediments were deposited in this region and at the top of the Baraboo sequence at this time. Depending upon how far north marine conditions reached, similar rock may have at one time covered the McCaslin and Barron Quartzites and later been removed by erosion. Alternatively, the fluvial to marine transition in the north may have been marked only by the change from fluvial to tidal sandstone as occurred in the Barron Quartzite.

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BARABOO INTERVAL QUARTZITE IN WASHINGTON COUNTY, IOWA

by

Raymond R. Anderson[1] and Gregory A. Ludvigson[1]

ABSTRACT

A previously unreported red Proterozoic quartzite was encountered in two drill holes in Washington County in southeastern Iowa. At least 46.7 m of the unit, the Washington County quartzite, was penetrated with the uppermost 5.5 m cored and cuttings recovered from the remainder of the well. Petrographic examination and modal analyses of thin sections prepared from the cored interval disclose a sequence of poorly-sorted, fineto coarse-grained quartzite interbedded with thin phyllitic units. The quartzite exhibits extreme compositional maturity. Larger constituent quartz grains are well-rounded. Phyllites and phyllitic quartzite units include a matrix dominated by muscovite with minor iron oxides, and are characterized by a steeply dipping foliation which is cut by a later cleavage, indicating polyphase deformation. These observations indicate strong compositional, sedimentologic, and structural similarities between the Washington County and Baraboo-type quartzite, and may indicate that the Washington County Quartzite is genetically related to the Baraboo, Sioux, and other generally correlative quartzite units of the mid-continent. As such it represents the southernmost reported occurrence of a probable Baraboo interval quartzite.

INTRODUCTION

The name Washington County Quartzite is here assigned to a sequence of Precambrian, probably Baraboo interval (Dott, 1983a), quartzite and phyllite encountered in two drill holes in west-central Washington County, Iowa. The formation was described from a diamond-drill core and rock-bit cuttings from two wells drilled in 1972 by the Natural Gas Pipeline Company of America at the SE¼SW¼SW¼ sec. 19, T. 76 N., R. 9 W., M. Flynn M-1 (IGS W-23141), and at the NW¼SE¼NW¼ sec. 20, T. 76 N., R. 9 W., W. F. Flynn M-1 (IGS W-23148) (fig. 1). Based on the structural relief on Paleozoic sedimentary units which drape over the Washington County Quartzite forming the Keota Dome, we infer that the unit is limited to the northwest and central Washington County and adjacent areas of northeast Keokuk County.

The Washington County Quartzite is similar in composition and general appearance to the Baraboo Quartzite, located 300 km to the northeast, and the Sioux Quartzite, 350 km to the northwest (fig. 1). It has only previously been reported by Yaghubpur (1979) and Anderson and Ludvigson (1983). Rock of the formation probably is the product of braided fluvial deposition, and is described in detail later in this paper.

The discovery well, completed early in 1972, was drilled by the Natural Gas Pipeline Company of America as a part of exploration of the Keota Dome for underground natural gas storage. This well, M. Flynn M-1 (IGS W-23141), was drilled on the flank of a domal structure that was mapped on an Ordovician datum (Hase and Koch, 1968). It reached quartzite at the Precambrian surface at a depth of 870 m (640 m below sea level) and penetrated 13.5 m into the unit. Rock bit cuttings were collected at 1.5 m intervals, but are of generally poor quality.

Upon completion of the discovery well a second test hole was drilled on the crest of the Ordovician structure. Named the W. F. Flynn M-1 (IGS W-23148), this well is located about 2 km northeast of the discovery well. It encountered the Washington County Quartzite at a depth of 706 m (461 m below sea level), 179 m higher than the discovery well (fig. 2). In this well the basal 107.6 m of the Cambrian System and the upper 5.5 m of the underlying Washington County Quartzite were cored. Recovery of the 113.1 m of 7.6 cm diameter core was virtually complete. The well was then deepened an additional 40.5 m to a final depth of 752.2 m using a rock bit. The cuttings collected at 1.5 m intervals during this deepening are of excellent quality. The core and cuttings from both wells are reposited at the Iowa Geological Survey Rock Library.

UPPER CONTACT

The contact between the Washington County Quartzite and the overlying Mt. Simon Formation displays a marked unconformity. The basal Mt. Simon is dominated by clasts of quartzite, apparently eroded from the underlying Washington County Quartzite.

1. Iowa Geological Survey, 123 North Capitol Street, Iowa City, Iowa 52242

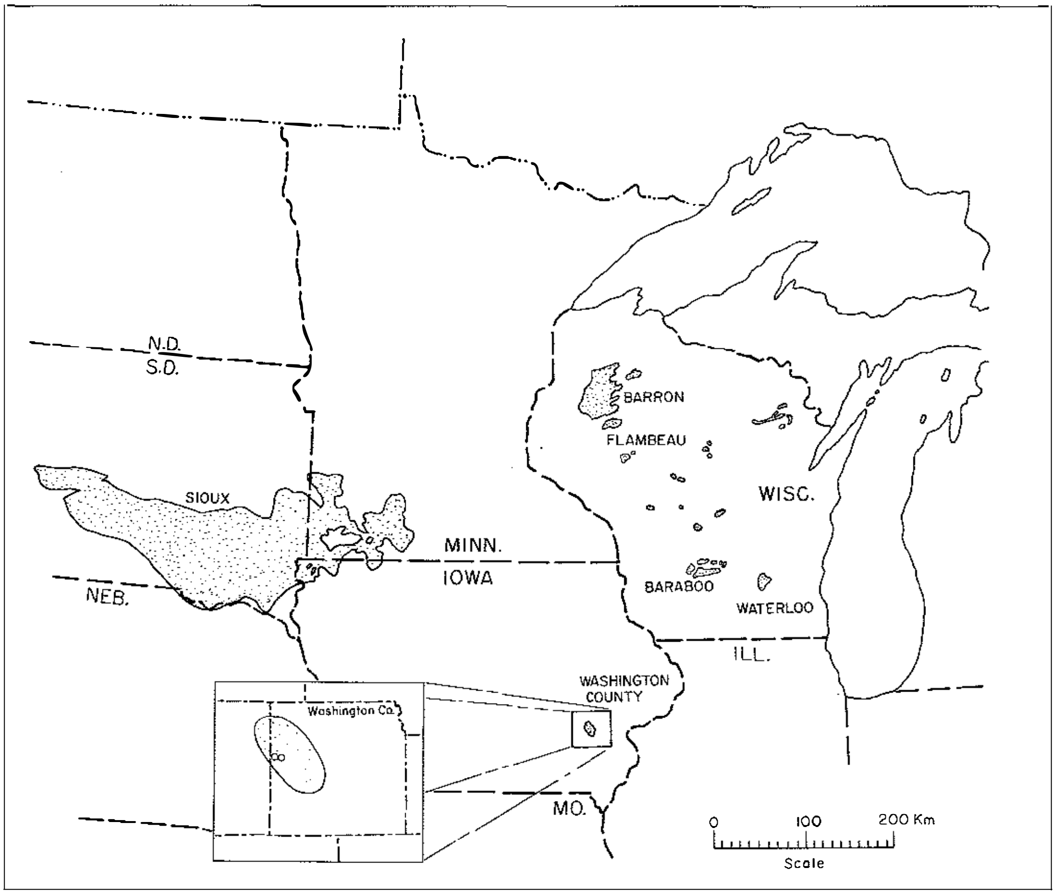

Figure 1.--Location of the Washington County Quartzite and known Baraboo interval quartzite.

The upper one metre of the Washington County quartzite encountered in the W. F. Flynn well is a gray-green to maroon phyllite with a pervasive foliation dipping at 52°. All dip angles were measured from the horizontal plane, assuming that the core axis is vertical. Modal analysis of this unit identified a 72 percent matrix content dominated by sericite with minor iron oxide (fig. 4, table 1). The possibility of other phyllosilicate minerals being present in this rock should be investigated. Muscovite has been noted from the Waterloo Quartzite, but pyrophyllite is the dominant phyllosilicate mineral from most of the other Baraboo interval quartzites (Dott, 1983b). The presence of muscovite is generally consistent with greenschist facies metamorphism of mudstone within the Washington County Quartzite. Silt-size grains identified were all quartz with 25 percent of the counts polycrystalline quartz and 3 percent monocrystalline.

PHYLLITIC QUARTZITE

The upper phyllite unit grades downward into a 2.5-m thick unit of maroon and gray mottled phyllitic quartzite with a pronounced foliation dipping 42°. The foliation is manifested as an alignment of mica grains which are evenly dispersed throughout the rock (fig. 5), similar to the quartzite cleavage illustrated from the Baraboo quartzite by Dalziel and Dott (1970, p. 25). Two modal analyses in this unit averaged 25 percent matrix, again dominated by phyllosilicates with minor iron oxides. The iron oxide is dominantly hematite with both rhombohedral and hexagonal crystal habits petrographically observed, as was a spinel group mineral, possibly magnetite. The presence of minor sulfide minerals in these rocks was reported by Yaghubpur (1979), but none were observed in this investigation.

The coarse component of the phyllitic quartzite was entirely quartz, with 65 percent of the counts polycrystalline and 10 percent monocrystalline grains (fig. 7). Phyllitic quartzite intervals were also encountered in rock bit drilling beneath the cored interval at depths from 718.8 m to 719.4 m and 723.7 m to 728.6 m.

Table 1. Modal analyses (in percent) of lithologic samples from the Washington County Quartzite and Baraboo interval rocks from other midwestern localities, from thin sections reposited at the Iowa Geological Survey. These reconnaissance analyses were performed by making 100 point counts for each thin section for the purpose establishing the mineralogical composition of the major framework grains.

Rock Suite	Sample	Lithology	Mono. Qtz.	Poly. Qtz.	Phyllo-silicate	Fe-oxide	K-feldspar	Muscovite schist lithoclast	Zircon	Plucked grain
Washington County Quartzite W.F. Flynn M-1 core	WQ-1 (2318.5')	Phyllite	3	25	71	1	--	--	--	--
	WQ-2 (2320')	Phyllitic Quartzite	1	71	27	1	--	--	--	--
	WQ-4 (2324.5')	Phyllitic Quartzite	16	61	14	9	--	--	--	--
	WQ-5 (2328')	Phyllitic Quartzite	7	25	55	13	--	--	--	--
	WQ-6A (2330.5')	Quartzite	19	61	13	15	--	--	--	--
	WQ-6B (2330.5')	Quartzite	36	54	--	3	--	--	--	7
	WQ-7 (2331.4')	Quartzite	12	71	2	7	--	--	1	7
	WQ-8 (2332')	Quartzite	13	84	1	1	--	--	--	1
Flambeau Quartzite	#1	Quartzite	42	48	9	1	--	--	--	--
	#2	Quartzite	24	51	14	9	2	1	--	--
Baraboo Quartzite	Wis-1	Quartzite	17	78	5	--	--	--	--	--
Sioux Quartzite	SD-1	Quartzite	54	44	--	2	--	--	--	--

QUARTZITE

Below the phyllitic quartzite the lithology of the Washington County Quartzite changes to a poorly-sorted, fine-grained to pebbly quartzite with an overall appearance similar to the Sioux, Baraboo, and some other Baraboo interval quartzites. Quartz grain sizes observed in this section ranged from a minimum of 0.004 mm to a maximum of 5 mm. The color of this section of the core is dark maroon. This interval displays bedding dipping 70° and numerous healed fractures dipping about 35°. Some of the fractures display local color bleaching. Bedding can be recognized as alternating bands of finer and coarser grains. One 5 cm thick band of pebbly quartzite at 711.3 m exhibits apparent tabular cross bedding, disclosed by a possible clay drape over foreset laminae.

Modal analyses of thin sections from three quartzite intervals showed a matrix content ranging from 2 percent to 9 percent and averaging 2.7 percent, dominated by iron oxide with subordinate amounts of sericite (fig. 6). Quartz grains comprise an average of 90 percent of this unit, with 70 percent polycrystalline quartz and 20 percent monocrystalline.

STRUCTURES

The most obvious structure evident in the W. F. Flynn M-1 core is the extensive foliation (S_1), especially pronounced in the finer-grained rock of the Washington County Quartzite. This folia- tion is expressed as a planar alignment of phyllosilicate minerals. In the phyllitic quartzite a penetrative muscovite foliation (S_1) was observed dipping at about 60°.

A phyllitic cleavage (S_2) which cross-cuts the penetrative muscovite foliation (fig. 7) was observed in the phyllitic quartzite, indicating a complex deformational history. A series of en echelon, muscovite-filled cleavage surfaces (S_2 in fig. 7) were identified dipping at about 30°. These features closely resemble deformational fabrics illustrated from the Baraboo Syncline (Dalziel and Dott, 1970, p. 25).

DISCUSSION

The compositional maturity, stratigraphic succession, and sedimentologic characteristics of the rock of the Washington County Quartzite closely resemble those noted in the better-studied Baraboo interval quartzites in the midcontinent. Terrigenous sandstone in the quartz arenite compositional field (fig. 4) are believed to be derived from stable cratonic source terranes of low relief, although deposition may have ultimately taken place in other tectonic settings (Dickinson and others, 1983, p. 223; Dickinson and Suczek, 1979, p. 2175).

The interbedded silty mudstone and poorly-sorted sandstone, which comprised the terrigenous sediments of the Washington County Quartzite prior to metamorphism, are generally consistent with braided fluvial deposition, as hypothesized by Dott (1983a,b) for much of the exposed Baraboo interval sequence in Wisconsin. The limited sample available for study precludes detailed interpretation of the depositional environments of the Washington County Quartzite. Nevertheless, the dominance of coarse-grained over fine-grained clastic sedimentary units and the poorly-sorted nature of the known stratigraphic succession are most easily reconciled with deposition by low-sinuosity braided stream systems (Brown, 1973, p. 13; Ore, 1964, p. 12). The possible presence of tabular cross bedding with clay drapes over foreset laminae, previously noted from the core at 711.3 m, would be consistent with this depositional setting.

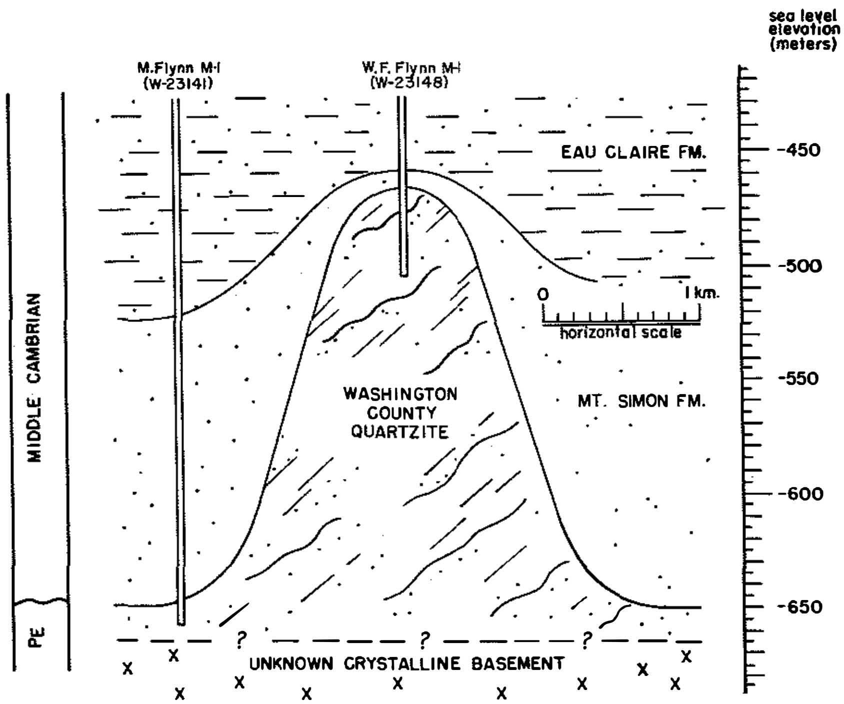

Figure 2.--Schematic geologic cross section of the Washington County Quartzite.

Estimates of the extent and geometry of this deposit are presently controlled only by the two drill holes, neither of which fully penetrate the unit. Therefore, all that can be stated with assurance is that the Washington County Quartzite has a linear extent of at least 2 km, that these rocks form a buried Precambrian hill with a paleotopographic relief of no less than 180 m, and that the thickness of the unit exceeds 46.7 m. Yaghubpur (1979) proposed that the difference in elevation of the quartzite surface between the two drillholes could be explained by vertical faulting of the basement surface. Overlying Phanerozoic deposits, however, do not display similar vertical displacement, and comparison with well-known Baraboo interval quartzite from other localities suggests that the relief of the Washington County Quartzite may be most easily explained as paleotopographic relief on the pre-Phanerozoic surface. The resistant hills of the Baraboo Range in Wisconsin have a documented pre-Phanerozoic topographic relief of greater than 300 m (Dalziel and Dott, 1970, Plate II). Shurr (1981, p. 28) has documented similar paleotopographic relief on the pre-Phanerozoic surface of the Sioux Quartzite, which forms a large buried resistant range called the Sioux Ridge.

The style of deformation and general metamorphic grade of the Washington County Quartzite is very similar to that observed in the Baraboo Syncline. While no radiometric age data are presently

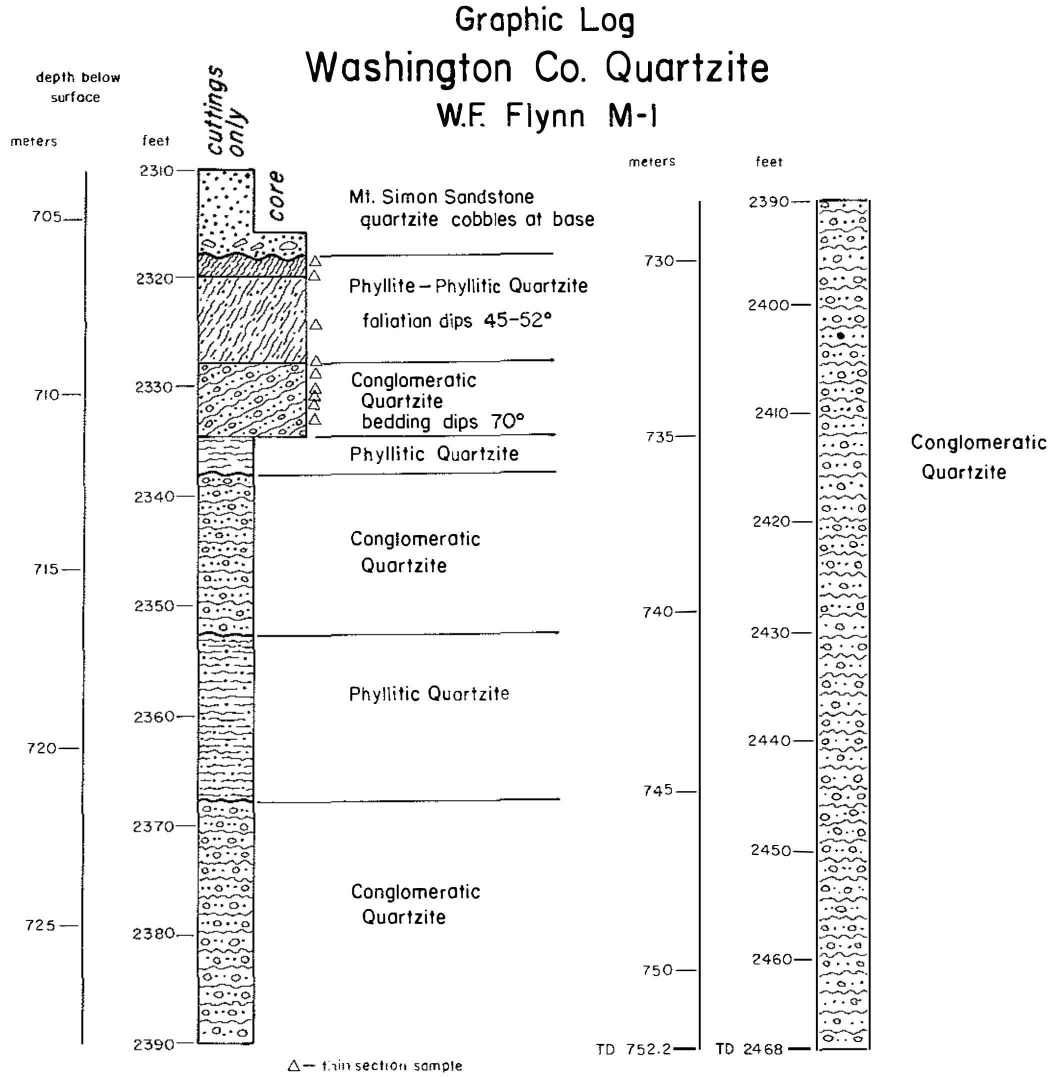

Figure 3.--Graphic lithologic log of the Washington County Quartzite from W.F. Flynn M-1.

available for this rock, their macroscopic appearance, petrology, and location far to the south of known Archean and Penokean terranes (Anderson and Black, 1983) argue that they were probably deposited and deformed contemporaneously with other Baraboo interval quartzites between 1,760 and 1,500 Ma (Van Schmus and others, 1975; Smith, 1978; Van Schmus, 1979; Van Schmus and Bickford, 1981; Dott, 1983a,b; Greenberg and Brown, 1984). This argument is strengthened by a recent review (Ojakangas and Morey, 1982) of the known Proterozoic sedimentary sequences of Lake Superior region that postdate the Penokean orogeny (1,860 to 1,800 Ma, Anderson and Black, 1983) but preceded Keweenawan volcanism about 1,100 Ma. Ojakangas and Morey (1982, p. 85) grouped these rocks into two major suites: (1) the Baraboo interval quartzites to the south of Lake Superior; and (2) a northern group of subarkosic, sublitharenitic, and arkosic quartzite whose outcrop belts fringe the Lake Superior Basin, including the Sibley Group (1,339 $\pm$ 33 Ma) and younger units. The Washington County Quartzite resembles their Suite 1 (Baraboo interval quartzites) much more closely than their Suite 2 quartzites.

REGIONAL TECTONIC IMPLICATIONS

Two differing tectonic interpretations have recently been proposed for the Baraboo interval. Dott (1983a,b) proposed that the Baraboo interval metasedimentary rock is a remnant of a miogeoclinal sequence that was deformed and regionally metamorphosed by plate collisional orogeny about 1,600 Ma. An alternative anorogenic setting for the deposition and deformation of the Baraboo interval rock has been proposed by Greenberg and Brown (1983a,b,c, 1984). According to their hypothesis, epicratonic Baraboo interval sedimentation was preceded by and partly contemporaneous with anorogenic felsic magmatism, and was followed by gravity-induced folding related to epiorogenic uplift in central Wisconsin. Greenberg and Brown (1983b, p. 11; 1984, p. 168) further argue that all instances of high-grade metamorphism in Baraboo interval rocks are related to contact metamorphic effects of the 1,500 Ma Wolf River batholith and related intrusive rock. We are using (and presently favor) the miogeoclinal hypothesis of Dott (1983) because: (1) it specifically makes testable predictions about the crustal evolution of buried Proterozoic terrane to the south of Wisconsin, whereas Greenberg and Brown's hypothesis does not; (2) it can be broadly integrated into the North American chronology of Proterozoic crustal accretion proposed by Van Schmus and Bickford (1981); and (3) many of the claims which Greenberg and Brown (1984) cite as evidence against Dott's (1983b) hypothesis have not been documented well enough for us to evaluate their validity.

Dott (1983b, p. 138) described the Baraboo interval quartzites as the "remnants of a vast sedimentary wedge that blanketed at least the southern margin of the post-Penokean craton of Proto-North America." He suggested that a "combination of braided rivers...and shallow marine processes dispersed clastic sediments across that continental margin." These sediments were subsequently "subjected to strong compressive tectonics, which folded and metamorphosed the... sequences" (Dott, 1983b). The "most plausible hypothesis to explain the seemingly anomalous deformation and metamorphism of the mature red quartzites...is either an arc-continent or continent-continent collision about 1600 m.y. ago" (Dott, 1983b p. 139). The suture proposed by Dott (1983b, p. 130, fig. 1) extends generally east to west across southern Wisconsin and northern Iowa, passing to the south of the outcrop belts of the Waterloo, Baraboo, and Sioux Quartzites (fig. 8).

Data obtained in studies of samples of the Washington County Quartzite are compatible with a braided fluvial mode of deposition. Metamorphism and deformation observed in the W. F. Flynn M-1 core might be the product of convergent tectonics. The Washington County quartzite, however, is located about 200 km south of the Proterozoic continental margin and suture zone proposed by Dott (1983b) If they are Baraboo interval rock, the quartz-rich sediment of the Washington County Quartzite could not have been located on the plate to the south of the suture, as their deposition would have been influenced by a nearby volcanic arc highland required by the southward subduction of oceanic crust in Dott's (1983b, p. 139) model. Terrigenous sediments derived from such a provenance would not compositionally resemble the Washington County Quartzite (Dickinson and Suczek, 1979). A modification of Dott's (1983b) plate tectonic model, including Washington County rock as authentic Baraboo interval quartzites, follows.

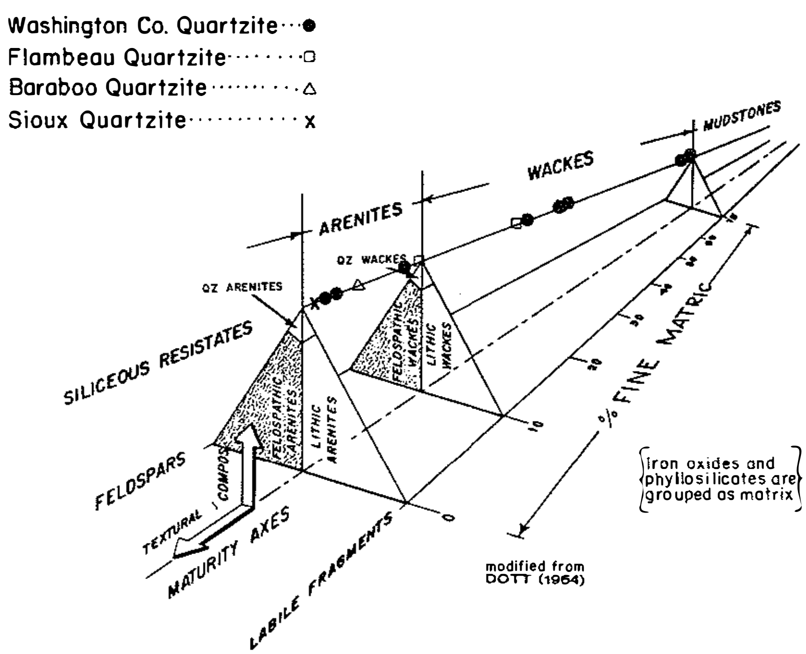

Figure 4.--Modal analyses of samples from the Washington County Quartzite and known Baraboo interval rocks from table 1. Triangular composition diagram in the prism corresponds to the QFL diagram of Dickinson, and others (1983).

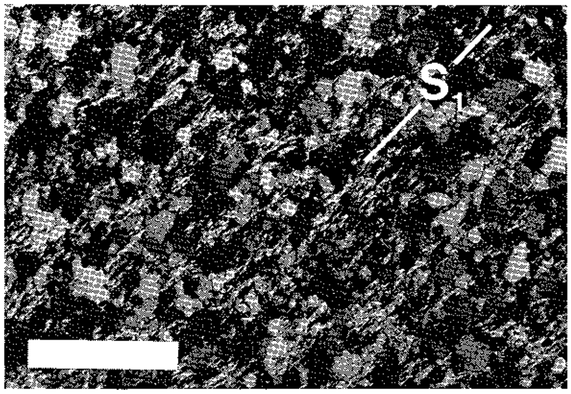

Figure 5.--Photomicrograph of phyllitic quartzite from W.F. Flynn M-1 (IGS W-23148), 707.8 m, showing foliation (S_1) developed by alignment of phyllosilicate grains. Scale bar is 0.5 mm. (polarized light).

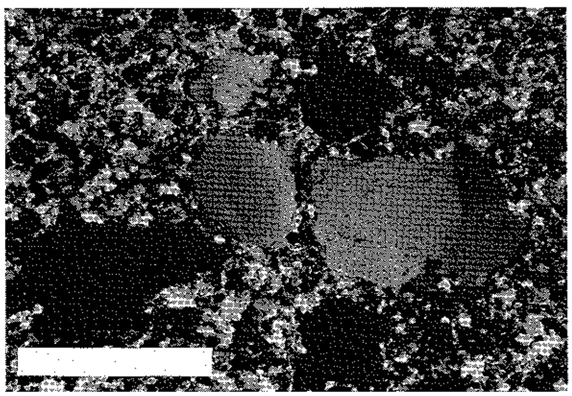

Figure 6.--Photomicrograph of pebbly quartzite from W.F. Flynn M-1 (IGS W-23148), 711.5 m. Note difference in original detrital quartz grains. Scale bar is 2 mm. (polarized light).

We propose that the Washington County Quartzite was deposited as part of the same sedimentary wedge and on the same passive trailing continental margin as the other Baraboo interval quartzites. To accommodate this interpretation, the continental margin and subsequent suture zone would have to be located to the south of Washington County, Iowa. A possible location for this suture is along a northeast trending zone of apparent crustal thickening first interpreted by Black (1981) and trending from panhandle Texas, across southeastern-most Iowa, and into Lake Michigan in the area of the Wisconsin-Illinois border (fig. 8). We feel this location for the suture has some merit, for the reasons outlined below.

The 1,630-Ma suture proposed by Dott (1983b, fig. 8) was discussed earlier by Van Schmus and Bickford (1981, p. 272, p. 280), who located it further south and referred to the feature as the Mazatzal Belt, a probable 1,680 to 1,610 Ma convergent plate boundary which they tentatively extended from southern Arizona to northern Illinois (fig. 8). Black (1981) identified a trend of high values on a map of the earth's magnetic field as recorded by NASA's MAGSAT satellite (fig. 9). The zone of high magnetic values corresponds directly to a pronounced gravity low, a combination of which is "indicative of crustal thickening" (Black, 1981, p. 78). Black (1981, p. 82) further noted the geographic coincidence of these paired geophysical anomalies with the Mazatzal

Belt of Van Schmus and Bickford (1981), and discussed possible relationships between them. This relationship suggests that the geophysical anomalies of Black (1981) may be a deep crustal signature of the Mazatzal Belt, the collisional orogen which is believed to have caused the foreland deformation of the Baraboo interval rocks north of the Baraboo suture (Van Schmus and Bickford, 1981, p. 272). Determination of the exact location of the Baraboo suture along Black's zone of crustal thickening is complicated by the subsequent emplacement of rocks of the "granite-rhyolite terrane" (Van Schmus and Bickford, 1981), the Wolf River batholith, and other felsic volcanic and plutonic rocks between about 1,500 and 1,380 Ma. This later terrane partially overlaps our proposed location for the Baraboo suture.

Our proposal is to use three admittedly speculative criteria to constrain our interpretation of the location and trend of the Baraboo suture. These are: (1) the MAGSAT contours of Black (1981) may be used as form lines to show the general trend of the Baraboo suture (fig. 9), which is the boundary between the Mazatzal Belt orogen and the passive continental margin upon which the Baraboo interval miogeoclinal rocks were deposited; (2) the Washington County Quartzite, which we propose to be part of this miogeoclinal sequence, would thus have been deposited on a basement of continental crust, and should be north of the Baraboo suture; and (3) the intensity of foreland deformation and regional metamorphism in Baraboo interval rock should decrease to the north away from the Mazatzal Belt orogen, and thus the metamorphic grade of pelitic rocks interstratified with the Baraboo interval quartzite can be used to further constrain the interpreted position of the Baraboo suture.

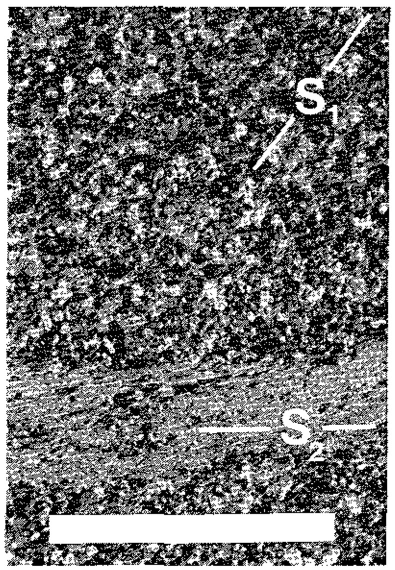

Figure 7.--Photomicrograph of phyllitic quartzite from W.F. Flynn M-1 (IGS W-32148), 707.1 m, showing a muscovite-filled cleavage surface. S_1 is the penetrative muscovite foliation, and S_2 is the crosscutting phyllitic cleavage. Scale bar is 3 mm (polarized light).

Metapelites from the Waterloo Quartzite apparently record a complex deformational and metamorphic history that was culminated by low-grade amphibolite-facies metamorphism, all of which Geiger and others (1981, p. 33) interpreted to be caused by a regional thermo-tectonic event related to the deformation of the Baraboo interval quartzite. Phyllite from the Baraboo and Washington County Quartzite, however, has apparently been metamorphosed to the greenschist facies. Finally, mudrock from the Barron (Campbell, 1981, p. 138) and Sioux (Austin, 1972, p. 455) Quartzites display only minor metamorphism, indicated by their alteration to argillite or pipestone. These observations can be used to construct a crude regional metamorphic zonation in the quartzites, as is shown in figure 8. High-grade metamorphism observed in the Rib Mountain and McCaslin Quartzites (putative Baraboo interval rock from northern Wisconsin) do not relate to this proposed regional metamorphic pattern, since they apparently are roof pendants to the 1,500 Ma Wolf River batholith, and have been contact-metamorphosed (Greenberg and Brown, 1983b, p. 11). The origin of the high-grade metamorphism in the Waterloo Quartzite (regional versus contact) however, is pivotal to the zonation shown in figure 8 and further attention should be focused on this problem.

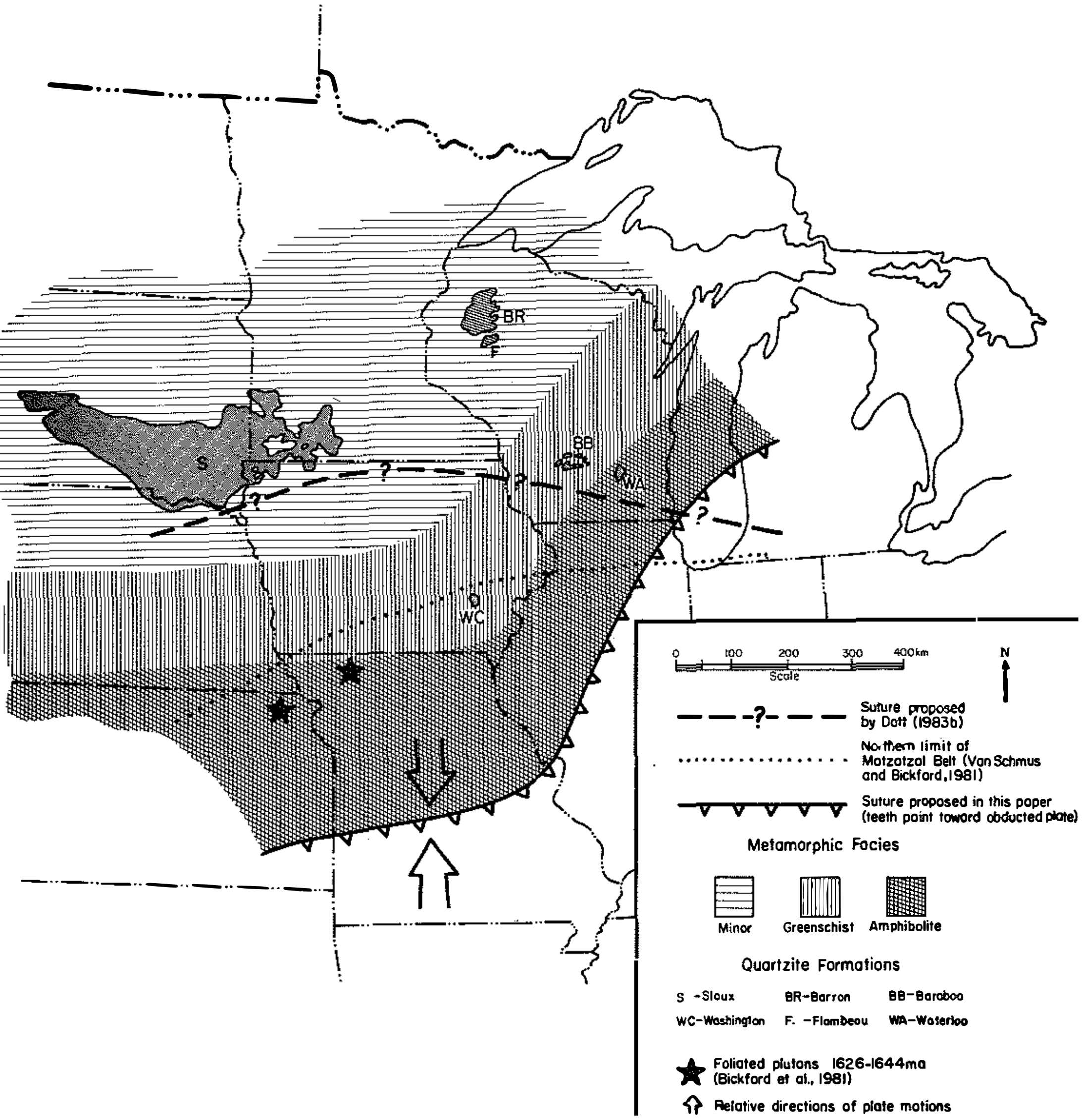

Figure 8.--Proposed locations for the Baraboo suture (1630 Ma) by Dott (1983b)and this paper with possible associated metamorphic zonation.

A possible location and trend for the Baraboo suture, using the constraints outlined above, is shown in figure 8. Note the positions of the 1,626- to 1,644-Ma, foliated plutons in northeast Kansas and northwest Missouri with respect to our proposed suture (fig. 8). These units were cited by Van Schmus and Bickford (1981, p. 275) and Bickford and others (1981, p. 339) as critical evidence for extending the Mazatzal Belt into the midcontinent region. The petrology of these foliated granitic plutonic rock indicates that they have been deformed by cataclasis, and that the granite from Gentry County, Missouri appears to be "a sheared and recrystallized igneous rock of granitic to quartz monzonitic composition" (Bickford and others, 1981, p. 330). The chemical affinities of this rock to continental crust is consistent with its location to the north of the proposed Baraboo suture, as shown in figure 8. Since the collisional orogeny leading to the development of the Baraboo suture is postulated to have been preceded by southward subduction of oceanic crust (Dott, 1983b, p. 139), volcanic-arc or Andean-type volcanism would be expected to have occurred on the plate to the south of the suture. We tentatively suggest that the foliated granitic plutons in northeast Kansas and northwest Missouri may have originated as the continental basement rock upon which the Baraboo interval miogeoclinal sediment was deposited (the 1,690 to

1,780 Ma Belt of Van Schmus and Bickford, 1981, p. 282), and that the rock possibly may be anatexite, deformed because of their proximity to the collisional Mazatzal Belt orogen (as in Schermer and others, 1984, p. 116).

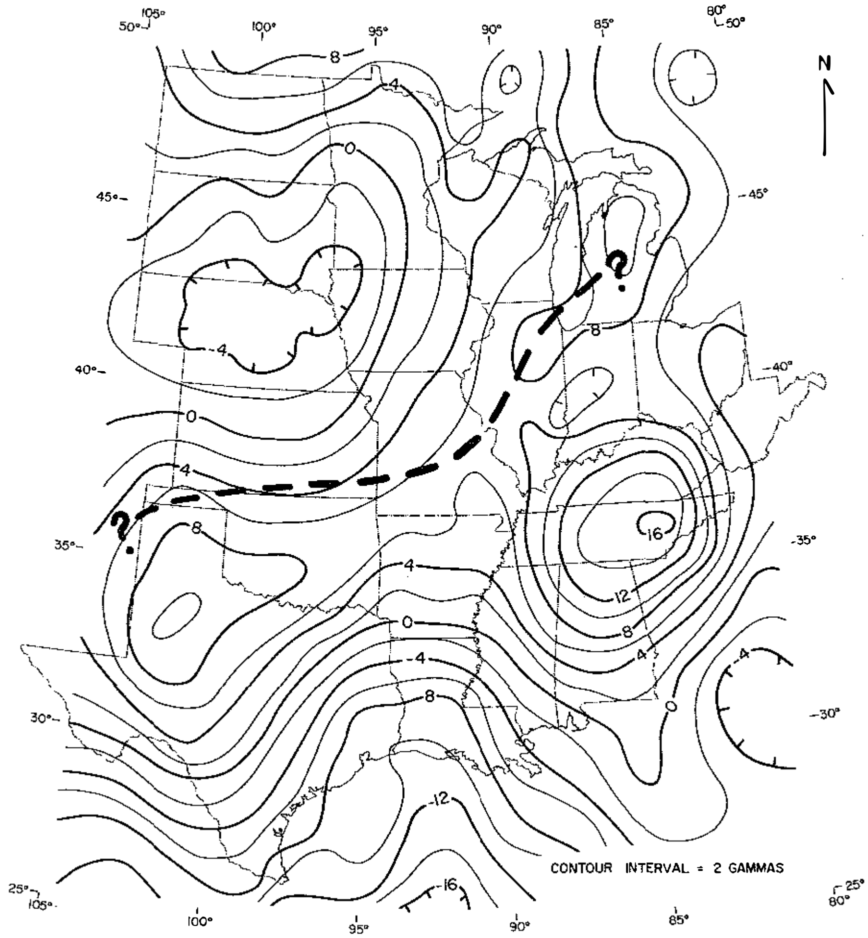

Figure 9.--MAGSAT vector magnitude map, reduced to the magnetic pole, filtered, one degree weighted-average (from Black, 1981, p. 74). Heavy dashed line shows the location for the Baraboo suture proposed in this paper.

One apparent weakness of our hypothesis is the troubling lack of parallelism between the east-west trending fold axes known from the exposed Baraboo interval quartzite (Dalziel and Dott, 1970; Brandon, 1983) and the generally northeast-trending collisional suture proposed here. This disparity might be explained by oblique continental convergence. It is hypothesized that the east-west trending folds record compressive foreland deformation along a northeast-southwest trending continental margin by a north-south directed plate collision (fig. 8).

Two alternative plate tectonic interpretations might be applied to the Washington County Quartzite: (1) the Washington County Quartzite was carried to its present position as part of an exotic terrane that was accreted onto the proto-North American continent during the 1,630 Ma collision or later. Thus, the Washington County Quartzite records the tectonic history of a different continental shield; or (2) the Washington County Quartzite was deposited and deformed on a passive trailing continental margin that developed to the south of Dott's (1983b) proposed suture during a post-Baraboo interval Wilson Cycle, and is younger than the Sioux, Baraboo, or related quartzite.

It seems unlikely that Washington County Quartzite was accreted onto the proto-North American continent as suggested in the first alternative interpretation. Integration of available drill data, outcrop studies, and geophysical signatures by Anderson and Black (1983) has led to the interpretation of the position of the Penokean (1,860 to 1,800 Ma old) suture, extending west and south from Green Bay (south of Escanaba, Michigan), across northern Wisconsin along the trend of the Niagara tectonic zone (Brown, 1983), then across southeastern Minnesota, northwestern Iowa, and into Nebraska following the Storm Lake geo physical trend (Anderson and Black, 1983). The nature of the gravity and magnetic signatures between the Washington County Quartzite and Penokean suture to the north (broken only by the Midcontinent Geophysical Anomaly, about 1,000 Ma) argues against a post-Penokean suture zone and any post-Penokean microcontinental accretion in this area.

The second proposed interpretation, that the Washington County Quartzite records post-Baraboo interval deposition and deformation, is also considered unlikely. The only presently known post-Baraboo interval, pre-Keweenawan Proterozoic clastic rock sequences in the region are located far to the north (Ojakangas and Morey, 1982), and are petrologically distinctive from Baraboo interval quartzite. If the rock of the Washington County Quartzite do indeed record a post-Baraboo interval Wilson Cycle, it would be the first known sedimentary rock from that interval.

In our opinion the relationships between the exposed Baraboo interval rock and the Washington County Quartzite will ultimately be resolved by investigation of the Proterozoic geology in areas presently covered by Phanerozoic sedimentary rock. It is apparent from Van Schmus and Bickford (1981) that a roughly east-northeast trending convergent plate boundary (the Mazatzal Belt) existed in the region during the 1,680 to 1,610 Ma interval. Since all interested workers seem to agree that the Baraboo interval rock was folded approximately during this time (Van Schmus and Bickford, 1981, p. 272; Greenberg and Brown, 1983b, p. 11; Dott, 1983b, p. 138), the precise location, trend, and tectonic history of the Mazatzal Belt in the midcontinent is obviously of great interest in resolving the regional tectonic setting for the Baraboo interval. This paper is intended to help further that end.

CONCLUSIONS

Petrologic studies of the Precambrian quartzite from Washington County in southeastern Iowa indicate that the sedimentology and style of deformation of this unit are closely similar to those known from the Baraboo Quartzite of Wisconsin. We suggest that the Washington County Quartzite may be reasonably correlated with the general depositional and tectonic episodes of the mid-Proterozoic Baraboo interval (Dott, 1981), during which the Baraboo, Barron, Flambeau, Sioux, Waterloo, and other unnamed quartzites were deposited and deformed. The compositional maturity of the Washington County Quartzite (table 1, fig. 4) is identical with that observed in samples from the Baraboo interval quartzite. The interbedding of terrigenous silty mudstones and poorly-sorted sandstones in the Washington County Quartzite is compatible with the braided fluvial origins interpreted by Dott (1983b) for most of the Baraboo interval sequence in Wisconsin. Alignment of phyllosilicate minerals in the finer-grained units imparts a steeply-dipping foliation (S_1) in the rock of the Washington County Quartzite. This feature, and later phyllitic cleavage surfaces (S_2 fig. 6), closely resemble structures illustrated from the Baraboo Quartzite (Dalziel and Dott, 1970, p. 25). Finally, the texture, color, and overall megascopic appearance of the unit strongly resembles other quartzite of the Baraboo interval.

The Washington County Quartzite represents the southernmost reported occurrence of probable Baraboo interval quartzite in the midcontinent. Its identification may provide an important constraint on interpretation of the former areal extent of the quartzite, and the location and orientation of a Proterozoic passive trailing continental margin and later collisional suture zone, which may have controlled the deposition and deformation of the quartzite (Dott, 1981, 1983a,b).

Both core and cutting samples of the Washington County Quartzite are reposited at the Iowa Geological Survey and are available for additional study. Future work may provide information important to the development of an understanding of these areally extensive but extremely enigmatic deposits.

ACKNOWLEDGMENTS

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MINERALOGY AND SEDIMENTOLOGY OF ROCK
OVERLYING THE BARABOO QUARTZITE

by

C.A. Geiger[1]

ABSTRACT

The thick Proterozoic Baraboo Quartzite of south-central Wisconsin is overlain by two sequences of metasedimentary rock. The first sequence consists of the Seely Slate and the Freedom Formation. The Seely Slate conformably overlies the Baraboo Quartzite and displays a low-grade mineral assemblage indicative of greenschist-facies metamorphism. The Freedom Formation consists of a lower, banded iron-formation with interbedded and overlying dolomite. Unconformably overlying the Freedom is the Dake Quartzite and Rowley Creek Slate. The Dake Quartzite appears to be a more immature lithology than the Baraboo Quartzite. It contains some detrital feldspar, although most has been replaced by kaolinite.

An environment of deposition on a continental margin or platform for the stratigraphically lowest three lithologies at Baraboo is consistent with the present data and the sedimentary model described by Dott (1983). The Dake may have micaceous quartzite analogues in central Wisconsin, but correlation of the Baraboo Quartzite and those in central Wisconsin is difficult to demonstrate and is open to considerable uncertainty at this time.

INTRODUCTION

The Proterozoic Baraboo Quartzite has been studied for over one hundred years, and thus one would expect that the tectonic setting and the related environment of deposition would be well understood. Unfortunately this is not the case, because diametrically opposite tectonic settings have been proposed recently to explain the metamorphic, structural, and sedimentologic features at Baraboo and possibly related central Wisconsin quartzite (Dott, 1983; Greenberg and Brown, 1984). As Dott has stressed, the absence of fossils in this rock makes the environment of deposition difficult to define. Therefore, sedimentologic criteria must be used. Much sedimentologic data are available for the Baraboo Quartzite. However, information is limited on the buried metasedimentary units overlying the quartzite. Much of the information that is available is old (Weidman, 1904) and not easily confirmed.

The purpose of this note is to further describe and document mineralogic and petrologic data for several lithologies that overlie the Baraboo Quartzite. Some of the sedimentologic features and stratigraphic relationships around the Baraboo area will be analyzed with this data. The information presented herein may help in the potential correlation of the Baraboo area and with central Wisconsin Proterozoic quartzite. It will also aid in the understanding of the sedimentary environment of deposition and thus the evolution of the Baraboo region as a whole.

GEOLOGIC HISTORY

Based upon structural studies, Irving (1872) was the first to demonstrate that the Baraboo Quartzite was older than the surrounding Paleozoic strata. He proposed, as Hall (1862) did earlier, that the Baraboo Quartzite was Huronian in age. Later, geologists working in the Baraboo area found and described several buried formations overlying the quartzite (Weidman, 1904; Leith, 1935). They formed the geological column of the Baraboo district as shown in table 1 and figure 1. Their work was initiated because of the discovery of iron ore in the western part of the syncline around 1900. Since most of these formations are not exposed, the descriptions of lithologies overlying the Baraboo Quartzite are known primarily from subsurface drill core and iron mining records. Leith (1935) proposed, however, that parts of the Dake Quartzite crop out. The metasedimentary rock around Baraboo overlies basement rhyolite and granite and have been described as an upper and lower series, with an unconformity separating the two (Leith, 1935). The Baxter Hollow granite on the southern part of the Baraboo Range is intrusive into the quartzite according to Gates (1941) and Schmidt (1951).

Early geologists correlated rock of the Baraboo district with the Menominee Group metasedimentary rocks and iron-formation in the Marquette area of Upper Michigan, because of the general stratigraphic similarities and the presence of iron-formation in the Freedom Formation. Both were considered Huronian in age. Today, the term Huronian is restricted only to metasedimentary and

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volcanic rock found on the north shore of Lake Huron (Card and others, 1972; Young, 1981). The early Proterozoic rock in Canada is now considered to be older than Proterozoic metasedimentary rock directly to the west, such as those found near Marquette and at Baraboo. Huronian sedimentary rock is considered to be about 2,500 to 2,100 Ma (Young, 1981). The metasedimentary rock and iron-formation in Michigan, which constitute the Marquette Range Supergroup (of which the Menominee Group is part, Cannon and Gair, 1970), are now thought to be Penokean, about 1,950 to 1,850 Ma in age (Van Schmus, 1982). Furthermore, Baraboo metasedimentary rock was interpreted by Dalziel and Dott (1970) and Dott and Dalziel (1972) as being distinctly younger than the somewhat lithologically similar Menominee Group Strata. The Baraboo Quartzite rests unconformably upon 1760 Ma rhyolites (Van Schmus, 1976). Therefore, Baraboo metasedimentary rock is also post-Penokean. Dott (1983) introduced the term Baraboo interval (1,760 to 1,500 Ma ago) for the time period in which sedimentation, deformation, and metamorphism formed the Baraboo and similar southern Lake Superior quartzite, such as at the Waterloo, Barron, and Sioux localities (fig. 2). Greenberg and Brown (1984) have recently described a number of quartzite, chert argillite conglomerate, and iron-formation localities in central Wisconsin (fig. 2), and have proposed that units were also deposited during the Baraboo interval.

PREVIOUS DESCRIPTIONS

Cambrian sandstone and conglomerate uncomformably overlie the Proterozoic metasedimentary rock around Baraboo. The oldest Cambrian sedimentary rock of the Mt. Simon and Eau Claire Formations is found in western Wisconsin, but are not exposed in the Baraboo region (Dalziel and Dott, 1970). They may, however, be present in the subsurface. The Galesville sandstone is the oldest Paleozoic unit exposed in the area. It is generally a white to gray rock, friable, unfossiliferous, and consists of well-rounded and well-sorted mediumto coarse-grained quartz sand, except around Baraboo where very coarse storm deposits have been noted (Dalziel and Dott, 1970).

The Precambrian sedimentary rock overlying the Baraboo Quartzite (table 1) has been described by Weidman (1904), Leith (1935), and Schmidt (1951). The Seely Slate is a soft gray and/or green chloritic slate. Weidman (1904) originally termed it a slate, although some parts might better be termed a phyllite or schist. Lenses of pure quartzite up to 8 m thick have been found in the Seely (Leith, 1935). Weidman described it to be uniform throughout the Baraboo region. The Seely displays a fine stratification and has a well-developed tectonic cleavage that cuts diagonally across bedding. Weidman listed chlorite, quartz, andalusite, and probably kaolinite as constituting the bulk of the rock. Brown mica, tourmaline, and apatite were also mentioned.

Table 1.--Precambrian Stratigraphy of the Baraboo District (modified slightly from Dalziel and Dott, 1970)

≥ 1,500 Ma	Rowley Creek Slate (maximum known thickness 45 m)
upper Baraboo series	Dake Quartzite (maxium known thickness 65 m)
	--- (Unconformity ?) ---
	Freedom Formation (dolomite, ferruginous slate, and iron-formation, minimum thickness 305 m)
lower Baraboo series	Seely Slate (maximum known thickness 110 m)
	Baraboo Quartzite (thickness over 1220 m)
	--- (Unconformity ?) ---
1,760 Ma	Rhyolitic basement (thickness unknown)

The Freedom Formation conformably overlies the Seely Slate. It is a thick (roughly 300 m) and lithologically diverse unit. Weidman proposed that the Freedom could be divided into two separate formations, although he did not do so. The upper member consists of a dolomite with or without chert. The lower member contains a thick banded-ferruginous chert, with thinner interbedded sideritic, calcitic, and dolomitic slate (Leith, 1935). Located in the lowermost Freedom is a ferruginous kaolinitic slate which is a transitional lithology between the underlying Seely and overlying ferruginous chert. The iron-formation of the lower member has a variable thickness ranging from roughly 60 to 160 m over relatively short lateral distances (Schmidt, 1951).

The Dake Quartzite was described by Leith (1935), and apparently was not recognized by Weidman (1904) in his study of the Baraboo region. The Dake unconformably overlies the Freedom, so that in places it rests on the lowermost part of the Formation. It also displays a flatter dip than the underlying Baraboo Quartzite (Leith 1935). These data were the basis for placing an unconformity between the upper and lower series (Leith 1935). The Dake is generally a coarse-grained quartzite containing a large amount of matrix sericite and chlorite, thus giving it a dirty appearance. Lower parts of the Dake are loosely cemented with iron oxide. Much of this formation is coarsely conglomeratic and has large angular pebbles. Most of the pebbles consist of quartz or quartzite. At Dake Ridge (fig. 3) where this formation is thought to crop out, red jasper pebbles, dark purplish quartzite and a few black or green pebbles also occur in conglomeratic beds (Leith, 1935). Greenberg and Brown (oral communication, 1983) have recently reexamined this outcrop and their observations agree with that given above. However, at least part of this outcrop consists of a cleaner well-sorted quartzite that is similar to the Baraboo Quartzite (Schmidt, 1951). The presence of jasper and dark purplish pebbles suggests local derivation from the lower Freedom and Baraboo Formations, respectively (Leith, 1935).

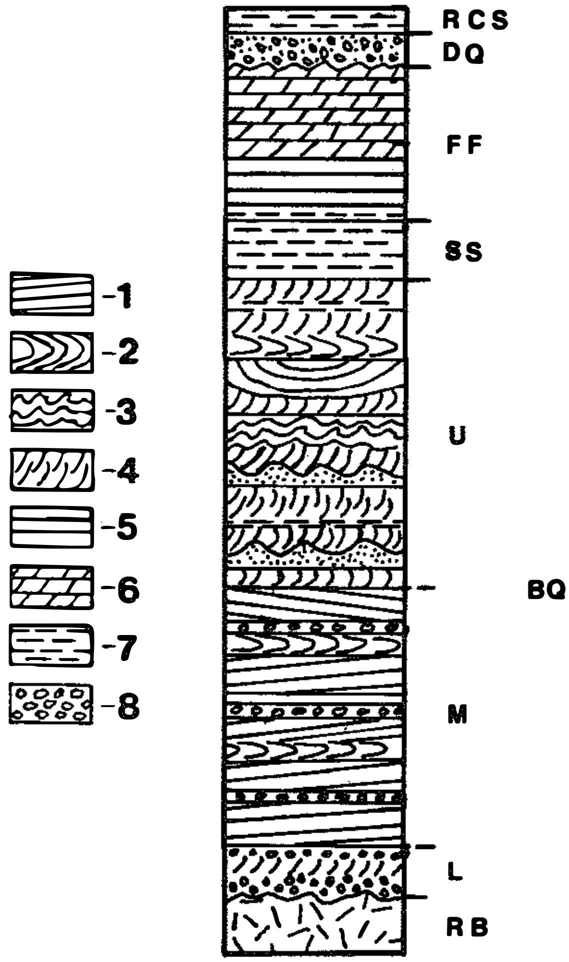

Figure 1.--Schematic geologic column of rock around Baraboo, Wisconsin, not to scale. Symbols and abbreviations: 8, conglomerate; 7, slate or argillite bed; 6, dolomite; 5, banded iron-formation; 4, high angle cross bedding; 3, ripple marks; 2, deformed cross bedding; 1, low-angle cross bedding; RB, rhyolitic basement; BQ, Baraboo Quartzite (L, lower; M, middle; U, upper); SS, Seely Slate; FF, Freedom Formation; DQ, Dake Quartzite; and RCS, Rowley Creek Slate (modified slightly from Dott, 1983).

The Rowley Creek Slate is the youngest Precambrian metasedimentary unit in the Baraboo area. It is only known in the eastern part of the district, as is the Dake (Leith, 1935). This may explain why it and the Dake were not described by Weidman (1904). Most of his work appears to have concentrated on the western part of the district in the vicinity of the Illinois mine (fig. 3). The Rowley Creek Slate is composed dominantly of sericite, with lesser amounts of chlorite and quartz. It is gray in color, but oxidizes red along bedding and cleavage planes. It has been folded and deformed and has not been found in outcrop.

METHODS

Several pieces of drill core taken from strata overlying the Baraboo Quartzite were made available by M.G. Mudrey, Jr. This core came from undescribed holes drilled many years ago (60-70?) at the Cahoon Mine. The Cahoon Mine is located about 1.5 km south of the town of Baraboo in the central part of the Baraboo Syncline (fig. 3). No information is available concerning the exact orientation or location of the drill core, except that it came from holes numbered 134 and

135 and depths between 410-509 ft. Most of the drill core from the turn-of-the-century iron mining period has since been destroyed or lost, and much of the information concerning the formations overlying the Baraboo Quartzite is based on an unpublished report (Dake, cited in Leith, 1935).

The core was studied using thin section and X-ray diffraction techniques. Most of the samples examined are relatively fine-grained and identification of some minerals was only possible in X-ray scans. X-ray scans often indicated the presence of 3 or 4 minerals, with quartz predominant. Due to diffraction-peak overlapping, a few minerals were tentatively identified using one or two peaks. Whenever possible, mineral separates were made by crushing and hand picking. All major phases were identified, but minerals in small modal amounts may not have been detected. Because of the fine-grain size, visual modal estimation of the various minerals was not attempted.

DISCUSSION

Four distinct lithologies are evident in the five pieces of core studied. The first lithology examined was a fine-grained green chloritic phyllite. X-ray scans showed chlorite, muscovite, quartz, and possibly albitic feldspar. Tourmaline and apatite were observed in thin section. Red hematite-rich bands or segregations are also evident in certain samples. Alternating arenaceous and argillaceous layers are visible in some thin section samples, thus producing a fine stratification (fig. 4). These bands probably represent original depositional layering. The most prominent tectonic foliation, S_1, strikes at a low-angle to this bedding. A second, weaker crenulation cleavage, S_2, is observable in some sections (fig. 5). The S_1 cleavage either partly accompanied or occurred after the development of small porphyroblasts or cross micas of chlorite which display cleavages at a high angle to S_1 (fig. 6). These observations fit the description of the Seely Slate as given by Weidman (1904). However, Weidman reported the presence of kaolinite and andalusite. The latter is very likely a misidentification (for albite?), as no andalusite was observed in several different samples. Furthermore, based upon a simple petrographic examination, this rock appears to have a metamorphic grade too low for the development of andalusite. Kaolinite may be present in small amounts, but it was not positively identified.

The second lithology can be best described as a red, iron-rich, cherty marble or carbonate. White or gray chert layers or semi-irregular patches constitute a major fraction of the core. A typical macroscopic chert layer is on the order of 1 to 4 mm thick. Some samples in thin section display alternating chert or carbonate or both with hematite layers about 0.1 to 1 mm thick. This

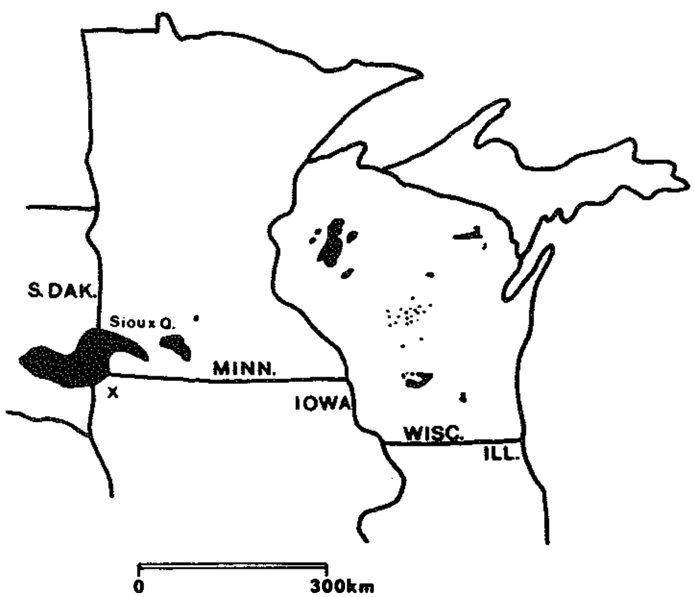

Figure 2.--Distribution of Baraboo interval exposures in the upper midwestern US shown in black. Veedum and Vesper are located in one of the small central Wisconsin outcrops Figure 2, p. 160 reproduced with permission of The Journal of Geology, by permission of the University Press, from J.K. Greenberg and B.A. Brown, 1984, Cratonic sedimentation during the Proterozoic: An anorogenic connection in Wisconsin and the upper midwest: v. 92, p. 159-171.

banding is distinct and usually additional to the large macroscopic banding. X-ray scans show the presence of calcite, quartz, hematitie, and goethite in the core sample. Other core samples that were only examined in thin section contain small amounts of white mica. One of these thin sections contains a small band of quartz and white mica, which appears to be detrital in origin. The presence of goethite is typical of many soft iron ores, and suggests in this case that significant

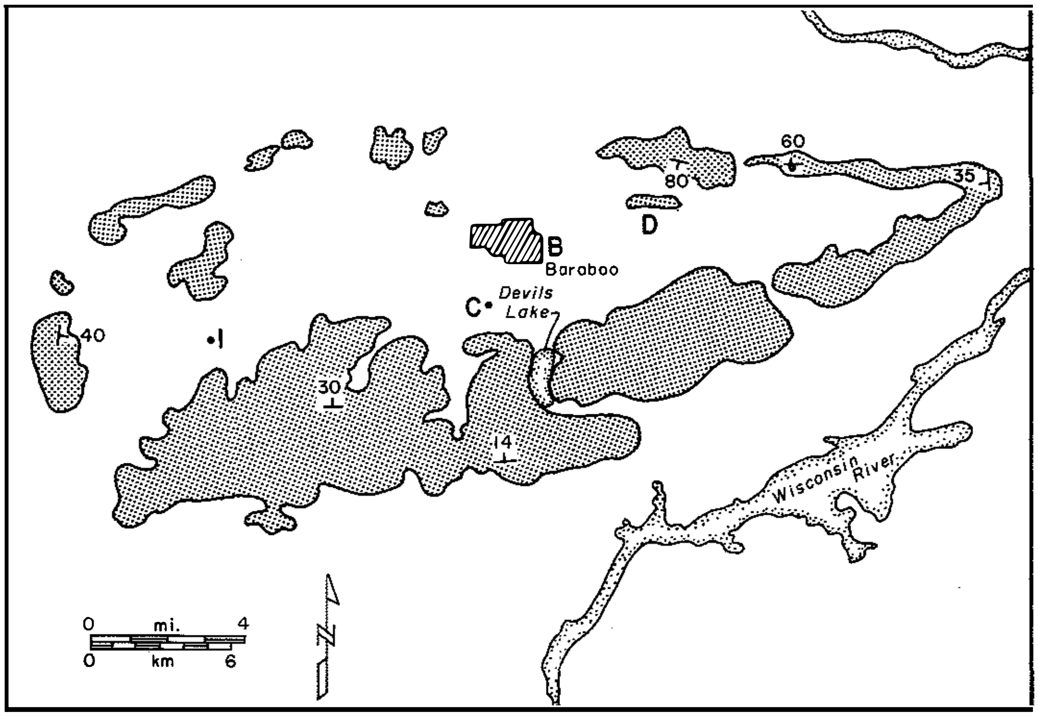

Figure 3.--Generalized geologic map of the area around Baraboo, Wisconsin, showing outcrop locations of the Baraboo Quartzite. Abbreviations: B, Baraboo; C, Cahoon Mine; D, Dake Ridge; and I, Illinois Mine.

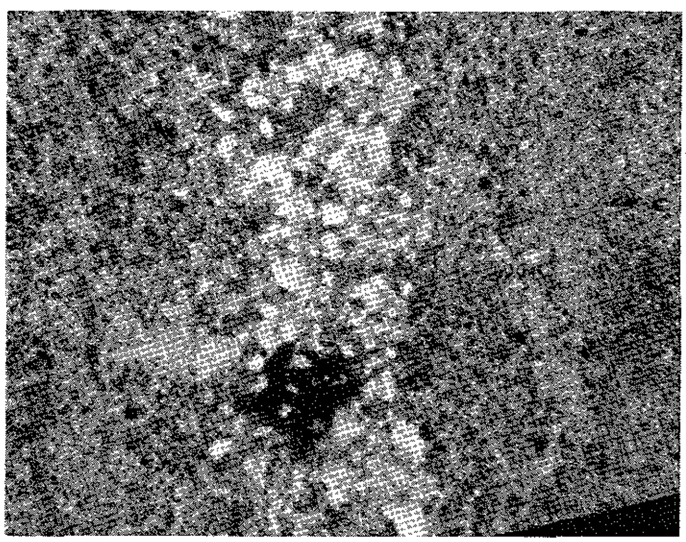

Figure 4.--Photomicrograph of the Seely Slate showing a thin arenaceous layer composed dominately of quartz. The large opaque grain is hematitie. Long direction is about 2.5 mm; crossed nicols.

alternation of a primary chert-carbonate-magnetite or hematite lithology occurred. The core and the thin sections match the description of the Freedom Formation, as given by Weidman (1904) and Schmidt (1951).

The third rock type is a medium-grained quartzite or subarkose. This rock is composed of quartz (95-90%), kaolinite, white mica, and trace feldspar. On a weathered surface this rock has a red iron-stained color, but freshly broken surfaces disclose a relatively clear quartzite. Some of the quartz grains are cloudy or milky colored. In thin section the quartz grains show a slight dimensional preferred elongation. These grains range up to 3 to 4 mm in diameter, but average about 0.5 mm. The kaolinite is concentrated in pockets that are up to 4 to 5 mm in size (fig. 7). X-ray scans of separates taken from these pockets give major kaolinite and small feldspar peaks. Matrix sericite or clay is sparse and is broadly disseminated between the larger quartz grains. The kaolinitic pockets probably formed from the breakdown of detrital feldspar. Thus, this rock is petrographically similar to some Baraboo interval rocks at Veedum and Vesper in central Wisconsin (fig. 2 in Greenberg and Brown, 1984).

Lithology type four can be best described as a fine-grained arkose. X-ray scans show mainly quartz, with substantial albitic feldspar, and probably microcline. Thin sections show also about 5 percent of muscovite, carbonate, chlorite, and heavy accessory minerals such as zircon. Feldspar is often sericitized and sometimes displays albite twinning. In addition the rock is colored red due to the presence of interspersed hematite. Grain sizes are fairly uniform averaging around 0.05 mm. Sorting is good and the majority of grains are subangular in shape. The core displays a distinctive intersecting cross-stratification in hand sample. This rock shows no evidence of pervasive metamorphism or recrystallization as in the other rocks mentioned above.

CONCLUSIONS

Dott (1983) proposed that the lower beds of the Baraboo Quartzite could be interpreted as fluvial clastic sands that were deposited on a continental margin. These sands grade upwards into a series of transgressive marine sediments called the Seely Slate and Freedom Formation (Dott, 1983). The iron-formation in the lower part of the Freedom contains a significant amount of ferruginous chert (Weidman, 1904) or what might better be called banded iron-formation. This lithology probably constituted a major fraction of the lower Freedom prior to the alteration which produced the soft iron ore. Weidman described the ferruginous chert as "layers of red and grayish chert alternating with layers of nearly pure hematite, and with layers of chert mixed with hematite. These alternating layers generally vary from one-fourth of an inch in thickness." This general description would also apply to much of the approximately 1,900 Ma-old oxide facies banded iron-formation located throughout the Lake Superior region (James, 1954). But it is now known, however, that the sedimentary rock at Baraboo was deposited 150-250 m.y. after the Lake Superior-type iron-formation.

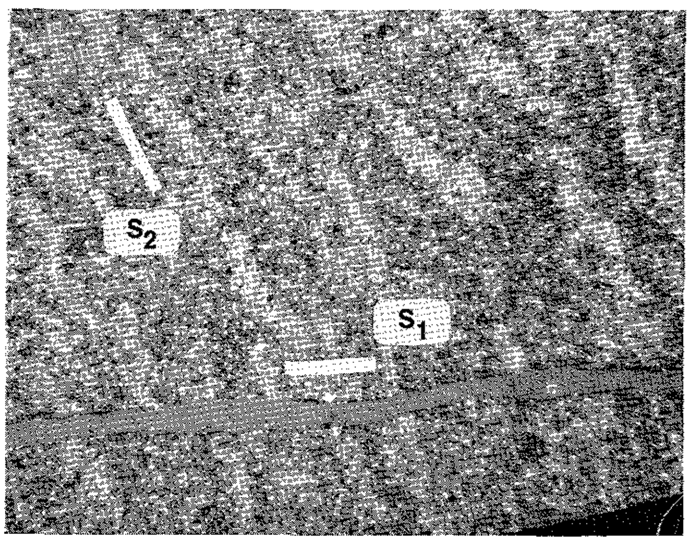

Figure 5.--Photomicrograph of the Seely Slate showing S_2 crenulation (light streaks) superimposed at a high-angle to S_1 which is parallel to the parting or crack in this slide. The major minerals are quartz, white mica (muscovite), and chlorite. Long direction is about 2.5 mm; crossed nicols.

Age differences aside, the environment of deposition could well be similar for both groups of iron-formation. Many workers believe that the banded iron-formation in the Lake Superior region was deposited in shallow water on a continental shelf or platform environment (James, 1954; Gross, 1983; Lougheed, 1983; however see Larue, 1981). The banded chert-hematite iron-formation and associated carbonate in the Freedom may have formed under similar conditions. This sedimentary environment for the Baraboo region agrees with the sedimentary model proposed by Dott (1983), which was based upon studies conducted primarily on the lower Baraboo Quartzite.

James (1954) described four distinct iron-formation facies that could be recognized in the Lake Superior region. The oxide facies which is characterized mainly by chert-magnetite-hematite, is conjectured to have formed in shallow water adjacent to the slightly deeper water carbonate facies iron-formation. At Baraboo not only does a thick unit of dolomite overlie banded chert-hematite iron-hematite iron-formation, but lesser amounts are also interbedded with it. These two facies, carbonate and oxide, also grade laterally into one another over relatively short distances (Schmidt, 1951). These observations suggest that deposition in the lower part of the Freedom occurred in a transitional facies, and that the upper Freedom represents a transgression to deeper water levels. Moreover, the presence of dessication cracks and carbonaceous matter in the lower Freedom (Weidman, 1904) are not definitive criteria for, but are consistent with, an intertidal environment of deposition. More petrographic work on the iron-formation in the Freedom could disclose the type of textures observed by Lougheed (1983) in early Proterzoic Lake Superior type iron-formation, and would enable correction or refinement to be made regarding the enviroment of deposition of the Freedom at Baraboo.

The Dake Quartzite unconformably overlies the Freedom (Leith, 1935). The kaolinitic quartzite or former subarkose sample described above could be from the Dake. This rock has pockets of kaolinite which still contain a little detrital feldspar. Therefore, it is unlike anything described from the lower Baraboo Quartzite, because most of the Baraboo is devoid of feldspar (Dott, 1983). Thin, widely spaced argillite beds are found in the Baraboo, but these are composed of the highly aluminous mineral pyrophylite. Such aluminous-rich bulk compositions usually occur in areas of intense chemical weathering (for example, tropical areas). For these reasons Dott (1983) argued that intense chemical weathering occurred at the source region for the Baraboo. The Dake as described by Leith (1935) is a less mature lithology. If the above hypothe- sis is correct, the presence of detrital feldspar indicates that the sedimentary environment or provenence changed between deposition of the Baraboo and the Dake. Hence, the unconformity be- tween the upper and lower series is a real possibility.

The composition of the Seely Slate, which comformably overlies the Baraboo Quartzite, may also indicate less maturity than the highly aluminous interbeds in the Baraboo. The Seely is richer in iron, magnesium, and the alakali elements because of the presence of abundant muscovite and chlorite. Its source region may not have been as deeply weathered, as that which provided the material for the pyrophyllite beds in the Baraboo. According to Dott (1983) the Seely Slate marks the transition between a fluvial (Baraboo) and a truly marine environment (Seely).

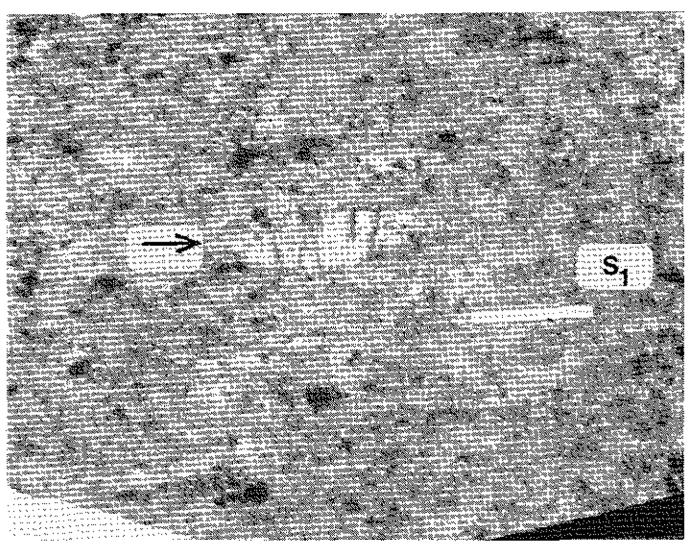

Figure 6. Photomicrograph of the Seely Slate displaying porphyroblast or "cross mica" of chlorite with growth cleavages roughly perpendicular to S_1 Long direction is about 1.0 mm; crossed nicols.

Greenberg and Brown (1984) have recently documented Proterozoic quartzite in central Wisconsin that is intimately associated with micaceous conglomerate, argillite, bedded chert, and iron-formation (fig 2.). The Dake could be correlative with these micaceous quartzites in central Wisconsin or they could be transgressive equivalents of the Baraboo Quartzite (Greenberg and Brown, 1984). The central Wisconsin quartzite is not always similar to the Baraboo interval quartzite described by Dott (for example, the Baraboo, Waterloo, Barron, and Sioux), although a few are. Furthermore, they generally do not appear to show the kind of thicknesses between 300 and 1000 m as indicated by the isopach map of Baraboo interval sandstone presented by Dott (1983). The Proterozoic metasedimentary rock and stratigraphic relationships in central Wisconsin are not well exposed or well understood. Hence, obvious or strict correlations between the Baraboo area and central Wisconsin are difficult, and at this time are open to considerable interpretation. The Dake may have analogues in central Wisconsin, but further work is needed to test this hyothesis.

It is not possible from this study alone to choose between the orogenic model of Dott (1983) and the anorogenic model of Greenberg and Brown (1984) for the tectonic setting of the Baraboo interval. More metamorphic and geochronologic studies (Geiger and Guidotti, in preparation), in combination with structural studies of all the quartzite are needed to resolve this problem. The data herein are consistent with the sedimentologic scenario for the lower series as presented by Dott (1983). Dott did not consider in detail the upper series, and no sedimentologic model has been advanced for these younger and poorest known lithologies at Baraboo. Unfortunately, at this time it is not wise to formulate any model based upon the limited data in this paper and in Leith (1935). However, in the future the presence of the upper series should be considered when constructing geologic models of the Baraboo region.

The fine-grained arkose described above may be a basal Cambrian lithology. This was suggested by J.K. Greenberg (oral communication, 1983), as he has observed similar lithologies from other drill core taken in the Baraboo region. The lack of any evidence for significant metamorphism in this rock is consistent with this proposal, with which I agree. Yet, this arkose is unlike many of the mature quartz arenites found among the Wisconsin Paleozoic sandstone. This rock may reflect local changes in Cambrian facies in the Baraboo region (for instance, a storm related deposit--Dalziel and Dott, 1970) as compared to elsewhere in Wisconsin.

Based upon the reconnaissance work presented in this paper, it appears that the sedimentological history of the Baraboo district remains a complex problem. Obiously much more work needs to be done, primarily sedimentologic in nature, on the buried and unexposed rock overlying the Baraboo Quartzite to gain a better understanding of the depositional history there.

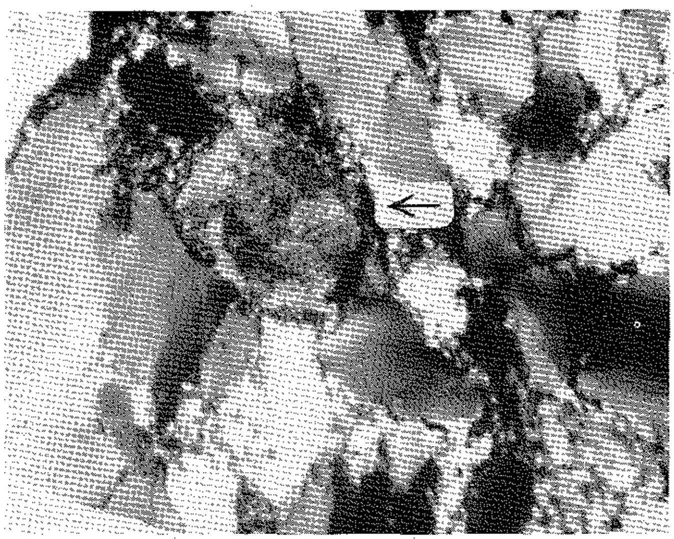

Figure 7. Photomicrograph of the Dake Quartzite showing a kaolinite pocket in left-center of this slide. Long direction is about 1.25 mm; crossed nicols.

ACKNOWLEDGMENTS

I thank M.G. Mudrey, Jr., for the core and many helpful discussions regarding the geology of Wisconsin. M.G. Mudrey Jr., B.A. Brown, and J.K. Greenberg read early drafts of this manuscript and made many useful comments and suggestions which improved this paper.

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PETROLOGY AND SEDIMENTATION OF THE FLAMBEAU QUARTZITE

by

Frederick K. Campbell[1]

ABSTRACT

The Early Proterozoic Flambeau Quartzite of northwestern Wisconsin is an outlier in the Penokean Volcanic Belt underlain by felsic volcanic and plutonic rocks. It has a minimum thickness of 800 metres and is folded into a syncline. The Flambeau Quartzite consists of quartz arenite to lithic graywacke. Quartz arenite is by far the most abundant rock type with conglomerate common. The most common framework grains are rounded common quartz, polycrystalline quartz, vein quartz, chert, argillite, magnetite, hematite, and zircon. These components suggest a dominant quartzrich sedimentary source. Paleocurrent data indicate a source terrane to the west and south of the outcrop area.

The poorly sorted pebbly texture, local abundance of clayey fragments and matrix, and unimodal paleocurrent pattern with low variance (2500) strongly support this deposition in a braided stream . Important controls on the sedimentation were the lack of land vegetation, aeolian transport, intense weathering, and a relatively stable tectonic environment.

INTRODUCTION

The Early Proterozoic Flambeau Quartzite is exposed at the surface only on Flambeau Ridge near the confluence of the Flambeau and Chippewa Rivers in northwestern Wisconsin (fig. 1). Flambeau Ridge is an east to west trending erosional remnant composed of well-indurated, cross-bedded sandstone and conglomerate. Contact between the Quartzite and adjacent rock is covered by glacial deposits. Diamond drill holes, water wells and outcrops provide strong evidence that the Flambeau is an outlier in the Penokean Volcanic Belt. Rock units in this belt include intermediate to felsic volcanic and metasedimentary rock (Myers and others, 1974). Model lead ages from volcanogenic massive sulfides near Ladysmith (fig. 1) suggest an age of 1,850 Ma for this belt (Sims, 1976).

The precise age of the Flambeau Quartzite is difficult to determine due to the lack of observable contacts with adjacent rock units. In order to clarify the age of the Flambeau, comparisons with other quartzite exposures in the region (for example, Baraboo, Waterloo, and some others) must be made. The other areas of quartzite have been studied extensively. In some cases there is direct evidence for their relative and absolute ages. It is not within the scope of this paper to make detailed comparisons between the Flambeau and other quartzite units. However, on the basis of lithology, structural and geological relationships, and geographic location, the Flambeau Quartzite is probably correlative with the Baraboo, Sioux, Waterloo, Barron and Rib Mountain Quartzites (Dott and Dalziel, 1972). Radiometric ages and other data indicate deposition of these quartzites after a 1,760 Ma igneous event and prior to a 1,630 Ma metamorphic-deformational event (Smith, 1978).

STRUCTURE AND THICKNESS

The Flambeau Quartzite is exposed in patches along Flambeau Ridge (fig. 2). Outcrop data indicate that the structure of the Quartzite is a steep syncline (fig. 2), although the axis of the syncline is not exposed in outcrop. A stereonet plot (fig. 3) of 80 poles-to-bedding (see Billings, 1972, p. 100-104) indicates that the axis of the syncline trends approximately N. 50° W. and plunges approximately 45° to the northwest. Stratigraphic tops were determined with the aid of cross-bedding, which is visible in nearly every outcrop.

No minor folds or related small-scale structures were observed in the Flambeau, but fractures, joints, slickensides and veins are fairly common in most outcrop. Veins are typically filled with milky quartz. Some veins resemble tension gashes but are not sigmoidal in form.

Due to lack of continuous outcrop the total thickness of the Flambeau Quartzite is unknown. I estimated the exposed thickness of the formation by measuring the stratigraphic distance between outcrop on the southern limb of the syncline (fig. 2). This method yielded approximately 800 metres of section, which will serve as an estimate of the minimum thickness of the Flambeau Quartzite.

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STRATIGRAPHY AND GROSS LITHOLOGY

The northern limb of the syncline is exposed only on the eastern end of Flambeau Ridge, whereas the southern limb is exposed on the central and western parts of the Ridge (fig. 2). The geometry and the structure suggest that the lowermost exposed beds are on the northern limb of the syncline, and the uppermost exposed beds are on the southern limb (fig. 2).

Three main lithologies occur in the Flambeau Quartzite: quartzite, mudchip conglomerate, and quartz-pebble conglomerate. All three lithologies are found in the north limb of the syncline, but mud-chip conglomerate is absent in the south limb. It appears that the mud-chip conglomerate ise the lower-most exposed units in the Flambeau, whereas the uppermost exposed units are mostly quartzite. The three main lithologies all exhibit a poorly sorted fabric, a well-indurated texture

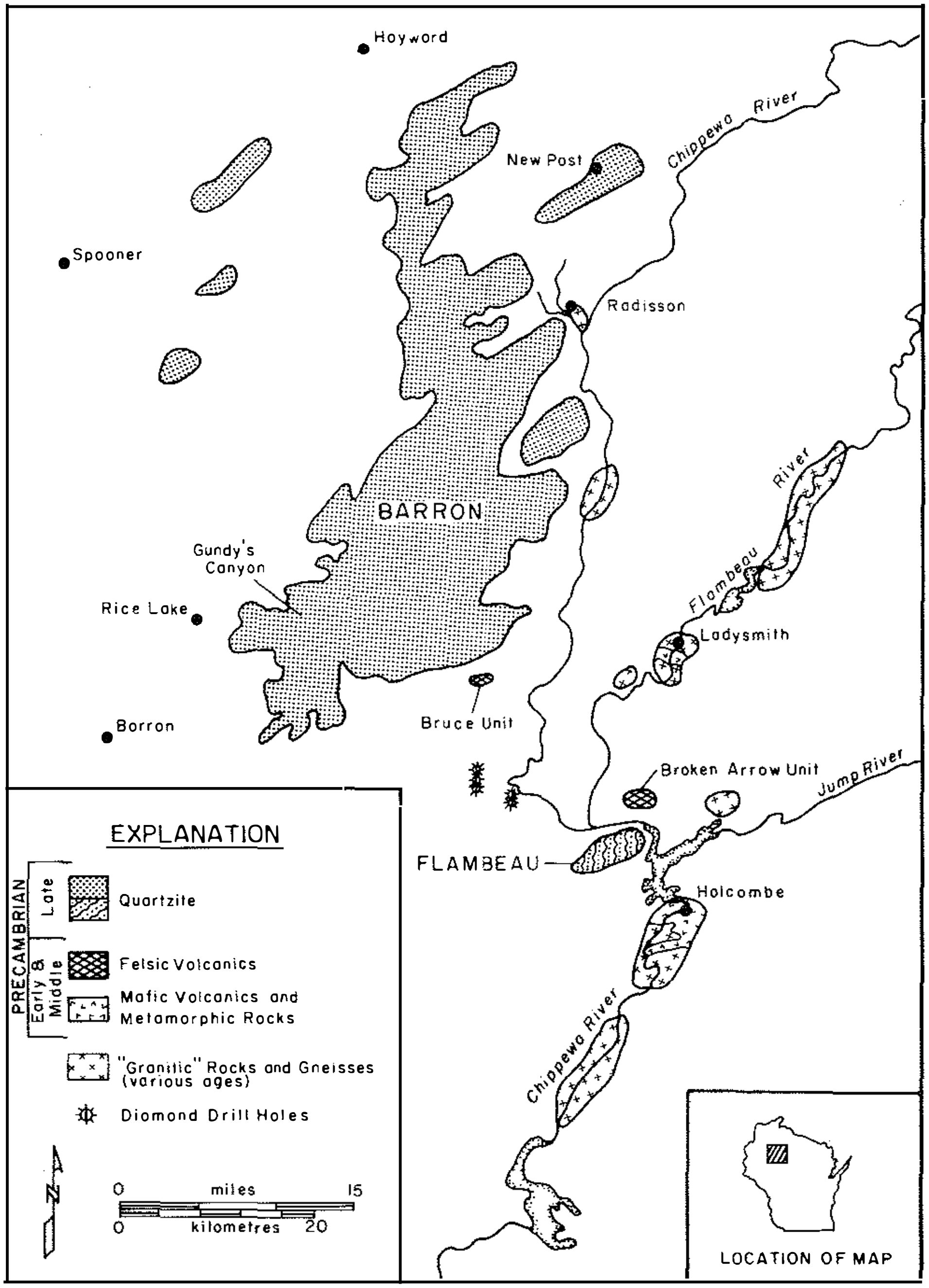

Figure 1. Map of Flambeau and Barron Quartzite Formations, modified from Hotchkiss and others (1915), Dutton and Bradley (1970).

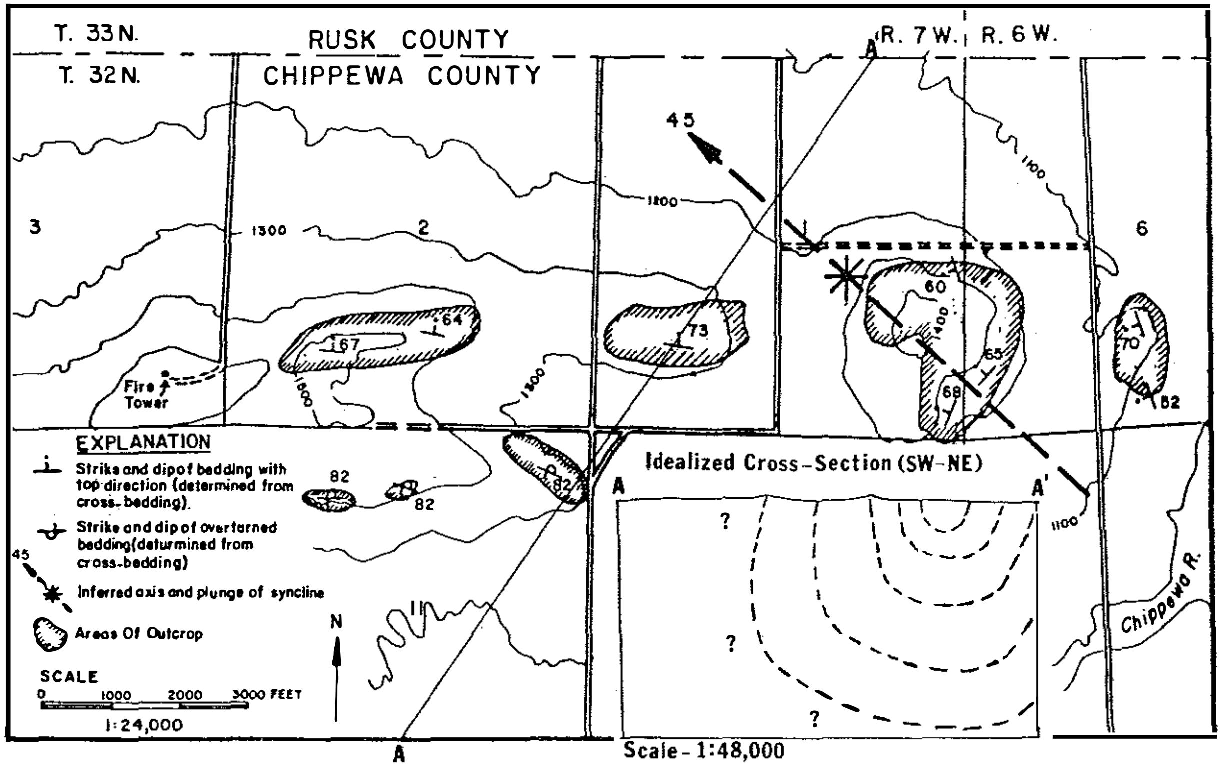

Figure 2.—Map of Flambeau Quartzite.

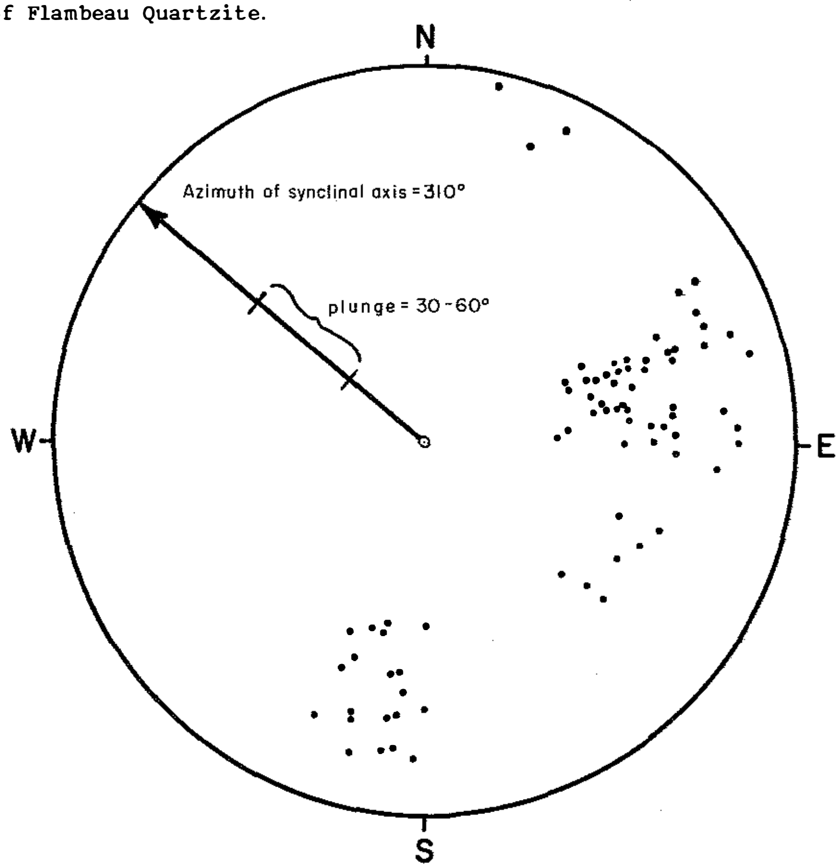

Figure 3.—Poles to bedding plotted with inferred axis and plunge of syncline (80 measurements).

texture, and stratification or cross-bedding or both. Pebbles occur in varying proportions in all three lithologies. Quartzite typically contains less than five percent pebbles, whereas the mud-chip conglomerate contains up to thirty percent pebbles, and the quartz-pebble conglomerate contains more than twenty-five percent pebbles. Pebble counts indicate that vein quartz is the most common pebble lithology (75 percent), followed by argillite (19 percent), chert and iron-formation (4 percent), rhyolite (1.5 percent) and quartzite (0.5 percent). These clasts are generally well-rounded except for the argillite pebbles, and are concentrated in stringers at the tops and bottoms of beds. Pebbles are also scattered randomly in the fine to coarse sand-sized matrix. The clasts range in size from a few millimetres to ten centimetres.

Bedding in the Flambeau Quartzite is visible in nearly every outcrop, although joints, fractures, and liesegang bands may be mistaken for bedding or may obscure it. The average thickness of bedding is 60 cm and ranges from 2 to 180 cm. Most beds are of consistent thickness in outcrop, however some beds are lensoidal or wedge-like. No bed could be traced to other outcrop areas.

In addition to bedding, cross-bedding is another common primary sedimentary structure. It is present in all lithologies. Tabular (planar) cross-bedding appears to be much more common than trough cross-bedding. The average thickness of foreset beds is approximately 30 cm, whereas the average dip of foreset beds is about 20°.

Ripple marks or possible pseudo-ripple marks are visible in a few outcrops. The form of the ripple-like features is irregular and the lateral extent is limited. The ripples are symmetrical with rounded crests and range from 7 to 15 mm in amplitude and from 5 to 15 cm in wavelength. These ripple-like features may be of sedimentary origin or they may be pseudo-ripple marks, the result of post-depositional deformation.

MINERALOGY AND PETROLOGY

The Flambeau Quartzite is a mature quartz sandstone composed mostly of sand-sized quartz with lesser but varying amounts of rock fragments and matrix material and minor accessory minerals (table 1). Four varieties of quartz were distinguished by the author: common, polycrystalline, stretched polycrystalline, and recrystallized quartz. Rock fragments in the formation include argillite, dacite to rhyolite, iron-formation and quartzite (see above). Matrix material is mostly sericite with minor kaolinite. Both quartz and hematite cement are present. The accessory minerals hematite, magnetite, leucoxene, zircon and tourmaline occur in trace amounts throughout the formation.

Rock types in the Flambeau range from quartz arenite to lithic graywacke (fig. 4). Most of the exposed sequence is quartz arenite and many units are conglomeratic. Most of the variation in composition can be seen in a few outcrops on the north limb of the syncline (fig. 2). The lithologic change from lithic arenite to quartz arenite (fig. 4) occurs over a 90 to 120 metre interval. This localized change is probably due to a local change in the environment of deposition. Muddy layers which accumulated during low energy conditions were probably the local source of the argillite fragments comprising the mud chips which formed during high energy conditions. Argillite fragments in the Flambeau are up to ten centimetres across. A fairly high energy, perhaps fluvial, environment is necessary to transport clasts of this size. Experimental studies by Smith (1972) showed that mud chips are broken down in fluvial environments within a few tens or hundreds of metres of their source. Thus, the argillite clasts were probably transported a short distance and were locally derived.

The sand-sized quartz grains which form most of the framework in the Flambeau appear to have been transported from distant sources. These grains show evidence of much abrasion since they are subrounded to well rounded in shape. Substantial abrasion of sand-sized quartz implies that the sediments were well worked and possibly wind transported to some extent. Double (abraded?) overgrowths on quartz grains, although extremely rare in the Flambeau indicate a multicycle origin for at least some of the grains. Multicyclicity is also indicated by the presence of quartzite clasts. The sand-sized rock fragments and heavy-mineral grains are mostly subrounded to well rounded in shape, suggesting substantial transport.

It is apparent from petrographic and petrologic considerations that the Flambeau Quartzite is comprised of both locally-derived (intrabasinal) and transported (extrabasinal) material. The framework grains in most lithologies exhibit fair to poor sorting and subrounded to rounded shapes. The abundance of matrix material, mostly sericite, in some Flambeau lithologies (table 1, fig. 4) is probably due to diagenesis of clays such as illite, montmorillonite, or both.

TABLE 1.--Modal Composition of the Flambeau Quartzite (figures in percent, x = less than 1%)

Sample Number	Common Quartz	Polycrystalline Quartz	Plag.	Rock Fragments	Opaques	Total Matrix	Total Cement
FR-1a	76	7	x	1	2	4	10
FR-1b	76	9	x	1	1	7	6
FR-2a*	48	34	-	5	2	10	3
FR-2b*	59	28	-	2	2	5	4
FR-2c*	55	26	-	11	2	4	5
FR-2e	52	6	2	7	2	25	8
FR-2s	52	34	x	1	1	6	5
FR-3a	69	14	1	2	x	7	6
FR-4a	73	10	1	3	3	3	8
FR-4d	71	11	-	3	2	5	7
FR-5c-1*	42	47	-	2	1	4	3
FR-5c-4*	53	32	-	4	x	7	5
FR-5c-6*	54	26	-	4	5	7	3
FR-5c-8*	29	50	1	4	6	8	3
FR-6b	62	24	-	2	x	8	4
FR-6d	73	16	-	1	1	4	5
FR-6e	71	17	-	1	x	4	7
FR-6f	70	22	-	1	x	3	4
FR-6g	72	11	x	3	1	7	6
FR-7a-1	72	14	x	3	1	7	4
FR-8a-1	60	17	-	11	1	2	7
FR-9c	63	15	-	6	x	12	4
FR-10a-3*	43	10	-	36	x	6	5
FR-11a-1	58	9	-	3	2	24	2
FR-12b*	52	13	-	13	4	11	7
FR-15b*	45	10	-	27	x	11	2
FR-17b*	49	8	x	27	1	9	2
FR-19b	71	14	x	2	1	3	8
FR-20a-1	75	13	-	1	2	4	7
FR-21a	75	12	1	2	1	3	5
FR-22a-2	72	14	x	2	x	8	3
FR-26a	71	12	x	5	1	6	5
FR-26b*	55	27	1	1	1	4	2
FR-26c*	38	50	-	8	x	2	x
FR-27	78	10	x	2	x	2	4

(*indicates conglomeratic sample)

PROVENANCE AND SEDIMENTATION

Thin section and heavy mineral data from the Flambeau strongly suggest a dominant quartz-rich sedimentary source for the Quartzite. Subrounded to well rounded chert, iron-formation and quartzite fragments are present in most of the formation and are direct evidence of siliceous sedimentary rock in the source area. The preponderance of rounded, sand-sized, common quartz grains, some with double overgrowths, is also indicative of sedimentary derivation. The dominant heavy-mineral assemblage consists of rounded zircon, tourmaline, magnetite, hematite, ilmenite and leucoxene. This mature assemblage implies a dominantly reworked sedimentary source (Pettijohn and others, 1972, p. 304).

Petrographic data indicate important igneous and metamorphic sources for the Flambeau sediments. Volcanic rock fragments and volcanic quartz grains are present and are direct evidence of igneous rock in the source area of the Flambeau. The heavy-mineral assemblage of augite, apatite, biotite and sphene is not common, but does imply an igneous source terrane. The presence of abundant vein-quartz, mostly in the form of pebbles, suggests an important igneous or metamorphic source or both. Polycrystalline quartz comprises up to 50 percent of some Flambeau samples (table 1). The ratio of polycrystalline quartz to total quartz is an indicator of provenance; a high ratio suggests a metamorphic source (Pettijohn and others, 1972, p. 300). This ratio may be as high as 5:8 in some conglomeratic samples of the Flambeau Quartzite (table 1); a metamorphic source for the formation seems likely. The presence of rare schistose rock fragments supports this conclusion.

An interpretation of cross-bedding orientation suggests that the source terrane for the Flambeau Quartzite was probably located to the west and southwest of the present outcrop area. A total of 101 measurements of cross-bedding were taken and were plotted on a stereonet. The azimuths of cross-beds were corrected for the plunge of the syncline (usually 30° to 60°) and for the dip of bedding (50° to 90°) (Ramsay, 1961). A unimodal pattern was obtained for all 101 measurements and also for 68 measurements which were less biased towards the better exposed outcrops (fig. 5).

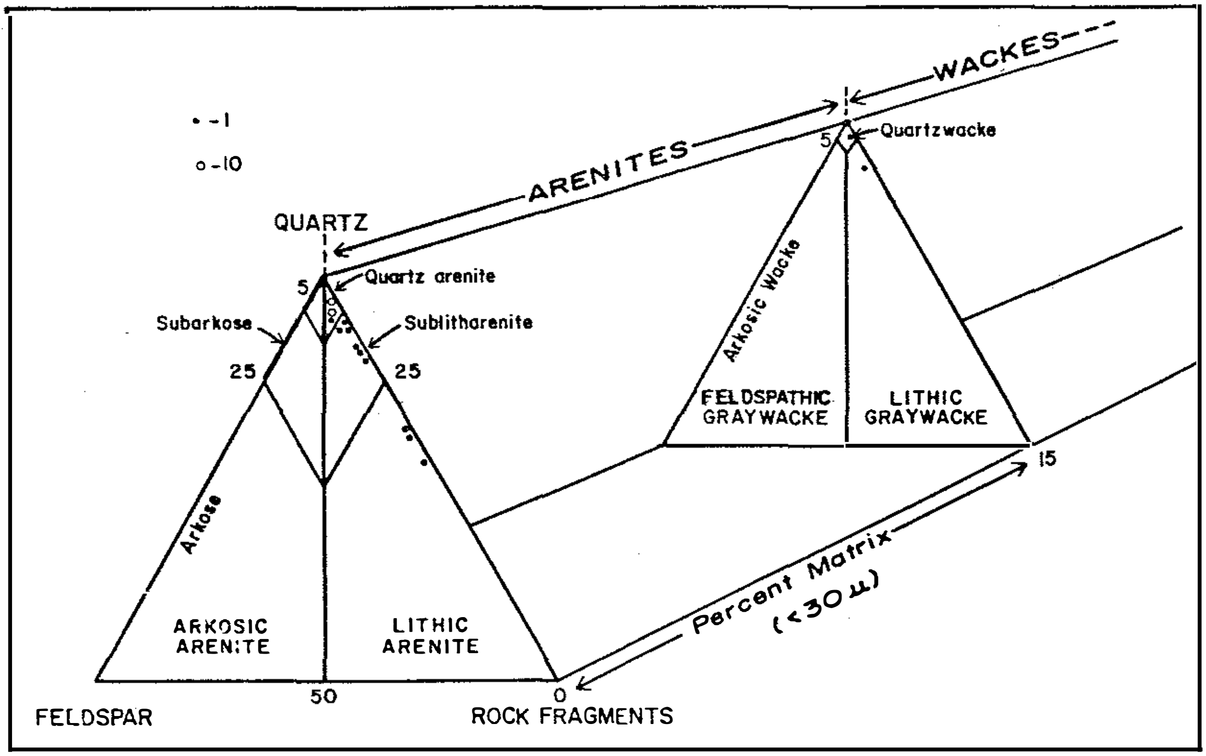

Figure 4.--Thirty-five samples of Flambeau Quartzite plotted diagrammatically to show compositional range.

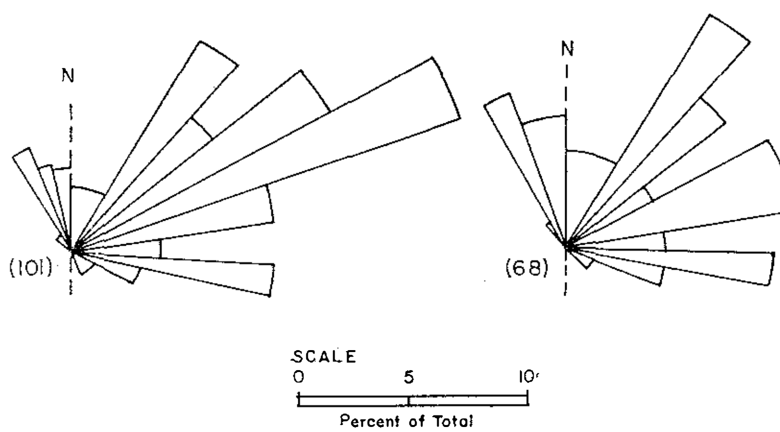

Figure 5.--Paleocurrent directions in Flambeau Quartzite based on doubly rotated cross-beds. Number in parentheses indicates number of measurements.

The variance of cross-bed azimuths in the Flambeau is fairly low (2500). A low variance is typical of a fluvial environment (Long and Young, 1978). The unimodal paleocurrent pattern for the Flambeau (fig. 5) also suggests fluvial deposition. The generally poorly sorted framework, the pebbly and conglomeratic texture, the local abundance of clayey fragments and matrix and unimodal paleocurrent pattern with low variance of cross-bed azimuths is strong evidence that the Flambeau was deposited in a braided fluvial environment (Rust, 1978).

Modern braided alluvial deposits exhibit various facies types which are based upon grain size and sedimentary structures (Rust, 1978). The Flambeau Quartzite exhibits many of the facies types which are recognized in modern environments. The facies types that are well-represented include Sp (planar cross-stratified sand), Sh (horizontally stratified sand), Gp (planar cross-stratified gravel), Gm (massive or horizontally bedded gravel) and Se erosional scours with muddy intra-clasts). These facies types are typically the result of longitudinal and transverse bars and their related alluvial features (Miall, 1978). A representative, composite vertical profile of the Flambeau Quartzite (fig. 6) shows the relationship between stratigraphy and facies types.

The braided stream system which probably deposited the Flambeau sediments may have resembled an alluvial plain due to the lack of vegetation-stabilized river banks. This type of environment permitted the rapid erosion and removal of the finer clastic material. Wind transport of sand, silt and clay was probably characteristic of the environment and is strongly suggested by the well-rounded, sand-sized material. Clay and silt probably accumulated during low-energy stages of sedimentation. Later current action destroyed the muddy layers and produced the observed mud-chip conglomerate (facies type Se) in the lower Flambeau.

Important controls on the sedimentation of the Flambeau Quartzite include the nature of source materials, the depositional environment, tectonics and climate. Tectonics is probably the primary factor and has a large influence on the other variables (Pettijohn and others, 1972, p. 243). However, sedimentation in the Precambrian may have been quite different from that in post-Devonian time (Long, 1978). Intense weathering, the lack of land vegetation and a quartz-rich source terrane (see above) may have been major factors in the sedimentation of the Flambeau. The possible large effect of these three factors on the compositional uniformity of the Proterozoic Athabasca Formation of northwestern Canada is postulated by Ramaekers and Dunn (1976).

The tectonic environment in which the Flambeau was deposited is implied by the lithology thickness and structure of the Quartzite. It seems likely that the formation was deposited at or near the edge of a craton. The sediment was well-worked and probably became mature in a tectonically stable area before reaching a more rapidly subsiding basin. The mature quartzite could have accumulated to a significant thickness in a rapidly subsiding basin. After deposition and fairly deep burial the Flambeau was probably deformed in a zone of mild tectonic activity.

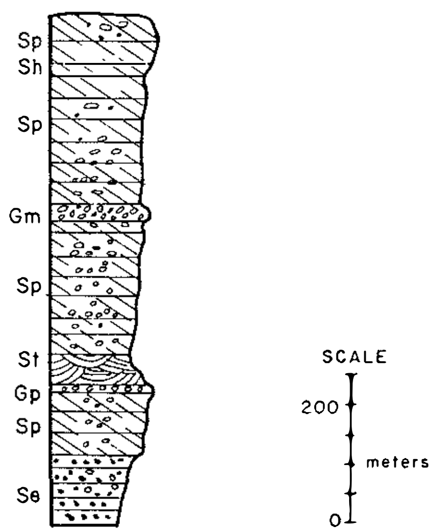

Figure 6.--Vertical profile of facies types in Flambeau Quartzite.

GEOLOGIC HISTORY

Based on lithology, structure, and stratigraphic relationships, the Flambeau Quartzite appears to be correlative with other early Proterozoic quartzites in the region including the Baraboo, Waterloo, Sioux, and Barron Quartzites. If this correlation is valid, then these quartz- ites probably had a similar geologic history. The available data suggest that the following se- quence of events led to the present condition of the Flambeau Quartzite: (1) syn or post-Penokean (1,760 Ma) felsic volcanic and intrusive activity in the source area; (2) erosion of an Archean or early Proterozoic source terrane; (3) intense chemical weathering of the source terrane, producing large volumes of quartz-rich sediment; (4) braided fluvial and aeolian transport of sediment; (5) slow but continuous subsidence of the depositional basin; (6) sheet-like sedimentation of sand- sized material with local lag deposits of gravel and pebbles; (7) burial and diagenesis of the sediment; (8) folding and metamorphism during a 1,630 Ma event; (9) fracturing; (10) uplift and erosion; (11) marine transgression and regression; deposition of sandstone during late Cambrian, later transgression and regression; (12) uplift and erosion; and (13) erosion and deposition of glacial material by the Chippewa Lobe during late Pleistocene time.

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PRELIMINARY PALEOMAGNETIC STUDY OF THE BARABOO QUARTZITE, WISCONSIN

by

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and
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ABSTRACT

Preliminary paleomagnetic data for the Baraboo Quartzite of southern Wisconsin are presented in this report. Nineteen hand samples from areas on both sides of the Baraboo Syncline were analyzed using both alternating field (AF) demagnetization and thermal demagnetization techniques. AF demagnetization up to 100 millitesla had little effect on the magnetic directions or intensity. Thermal demagnetization indicated that the stable remanence is probably caused by hematite (curie temperature 650 to 700 °C). The statistics of the magnetic directions improve for the samples from the western end of the Baraboo Syncline when structural corrections are made. This indicates that the magnetization was set prior to folding. The virtual geomagnetic pole (VGP) is located at 19°S., 89°W., which is consistent with currently published pole positions for the time range of 1,800 Ma to 1,700 Ma for North America. The statistics for samples from the eastern end of the syncline do not improve when structural corrections are made. This suggests that the magnetism for these samples may be post folding. The VGP for these samples before structure corrections is 57°W., 44°N., which is similar to the VGP obtained for several of the associated rhyolite.

INTRODUCTION

The Precambrian geology of Wisconsin represents a complex succession of tectonic events. Of particular interest to this study is a suite of rocks in southern Wisconsin comprised of rhyolitic volcanic rock, associated granite, and overlying quartzite, which were originally deposited as sands sometime between 1,760 Ma to 1,450 Ma. The geologic history of this sequence and the tectonic causes are not fully understood, although there are several recent interpretations of it.

It is known that the rhyolite and granite intruded Penokean age rock (about 1,850 Ma old) about 1,760 Ma (Smith, 1978; 1983; Van Schmus and Bickford, 1981). This was followed by the extensive deposition of sand (now quartzite) in fluvial to shallow marine environments (Dalziel and Dott, 1970; Dott, 1983). Folding of the quartzite probably took place sometime between 1,630 and 1,500 Ma, the time of intrusion of the Wolf River batholith anorogenic granite complex.

Dott (1983) has suggested that the red sands which are now the Baraboo Quartzite formed on a passive continental margin between 1,700 to 1,600 Ma; deformation and metamorphism of the quartzite took place about 1,600 Ma by either an arc-continent or continent-continent collision from the south. Evidence for a tectonic event at about 1,630 Ma comes from Rb/Sr systems of the 1,760 Ma-old rhyolite which has been reset at 1,630 Ma (Van Schmus and others, 1975).

Greenberg and Brown (1984) present a different tectonic history for the area, one similar to the Basin and Range province of the western United States (Atwater, 1970). The deposition of the Baraboo Quartzite began on a subsiding continental crust during the late stages of anorogenic magmatism 1,760 Ma. The 1,630 Ma event is considered a time of epeirogenic uplift. This uplift brought about relatively mild deformation of the Baraboo rocks, with some metamorphism. Finally, the widespread intrusion of the 1,500 Ma-old Wolf River batholith further deformed the region and caused thermal metamorphism. No collision tectonics is considered in this model.

It is apparent from these two models that the time of the folding and metamorphism of the Baraboo Quartzite is still uncertain. There are certain specific questions which may be answerable from paleomagnetic studies. Two of great interest are the actual deposition age of the Baraboo Quartzite and the time of folding. Presently, the deposition and folding are bracketed between about 1,760 Ma and 1,500 Ma from the ages of igneous activity in the area. There is also uncertainty with the final age of intrusion and metamorphism in the Baraboo area. Dott (1983) suggests that the last thermal event was reheating at about 1,600 Ma. Greenberg and Brown (1984) place the final reheating activity closer to 1,500 Ma. The magnetic characteristics and paleomagnetic directions from the granite and rhyolite may help to determine the igneous history of the area.

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In this paper we present preliminary paleomagnetic results from the Baraboo Quartzite. Other studies are in progress on the associated rhyolite and granite near Baraboo, such as the Denzer rhyolite, Baxter Hollow granite, and the granitic and rhyolitic inliers of the Fox River Valley.

PREVIOUS PALEOMAGNETIC STUDIES

Runcorn (1964) undertook preliminary paleomagnetic studies of the Barron and Sioux Quartzites which are quite likely contemporaneous with the Baraboo. However, no cleaning methods were used, and the results show great scatter. No paleomagnetic studies have been reported for the Baraboo Quartzite. Irving (1979) and Roy (1983) have included in their reviews the paleomagnetic data for the geologic time of interest for the Baraboo, that is 1,800 Ma to 1,500 Ma, however, there is no data from locations in Wisconsin.

GEOLOGY

The Baraboo Quartzite crops out near the town of Baraboo in southern Wisconsin (fig. 1). It is a sequence of massive, vitreous rock, typically pink, maroon, or purple, and up to 1500 metres thick (Dalziel and Dott, 1970). It is composed of more than 80 percent quartz, which occurs as medium to coarse sand grains. Numerous layers or lenses of more argillaceous material occurs within the quartzite. Dott (1983) suggests that this unit is the result of braided-stream deposition. The quartzite is formed into a doubly plunging syncline with the major axis oriented east-northeast. The north limb of the major syncline is almost vertical. In some places it is overturned. The south limb has shallower dips on the order of 15° to 30° (fig. 2). The location of the associated rhyolite and granite is shown in figure 3 (Smith, 1978).

PALEOMAGNETISM: FIELD AND LABORATORY PROCEDURES

Three to eight, oriented, hand samples were collected from the 12 locations noted on figure 1. The sites were selected such that both limbs of the major syncline were sampled. The local strike and dip of the beds were also noted for each site. Each sample was cored in the laboratory to obtain three to eight core specimens of size 2.54 cm diameter by 2.54 cm long. However, all of these samples have not been measured as yet. The numbers of samples and specimens actually used in this report are listed in table 1.

An initial suite of 10 cores were chosen for demagnetization studies by both alternating field (AF) demagnetization and thermal demagnetization. This type of analysis is used to determine the magnetic stability of the rock samples. The natural remant magnetism (NRM) of all samples were initially determined using a two-axis cryogenic magnetometer. The samples were then subjected to either AF demagnetization or thermal demagnetization. The AF demagnetization was performed in a Schonstead single-axis AF demagnetization unit, stepwise from 0 to 100 millitesla. The remanent magnetic direction was measured after each demagnetization step. In general the Baraboo Quartzite does not respond to AF demagnetization; neither the directions nor the intensity of magnetization changes significantly (less than 5° and 5 percent) with peak fields to 100 millitesla. The remaining samples from this preliminary study were thermally demagnetized by stepwise heating and cooling cycles. The heating was performed in a noninductive, wound furnace set inside a three-axis set of helmholtz soils; the samples were then cooled in a three-stage, mu-metal shield. The helmholtz coils and mu-metal shield isolate the sample being measured from the ambient magnetic field of the earth. Thermal demagnetization (fig. 4) shows very little decrease in the magnetic intensity or changes in magnetic direction until the demagnetizing temperature reaches 600 °C or higher. These results indicate that the magnetic mineralogy is single phase and probably hematite (curie temperature 650 to 700 °C).

From this initial suite of samples the remaining samples were measured for NRM and also thermally demagnetized at values of 200, 400, 650, and 750 °C.

PALEOPOLE ANALYSIS

The mean inclination and declination were determined for each site from the averaged directions for each hand sample. This calculation was made before and after the removal of the known struc- ture for each site. This involves unfolding the beds and in some cases rotating them as well. We have chosen N. 90° E. as the direction of the main synclinal axis. Fisher statistics (Fisher, 1953) were determined after each set of corrections to determine what structural corrections, if any, should be made on the data. This represents the classic fold-test of Graham (1949). If the data has a tighter cluster or better statistics before the structure is removed, it generally in- dicates that the measured magnetism is post folding. However, if the data clusters better after the structural corrections are made, then the magnetic directions in the rocks are prefolding directions.

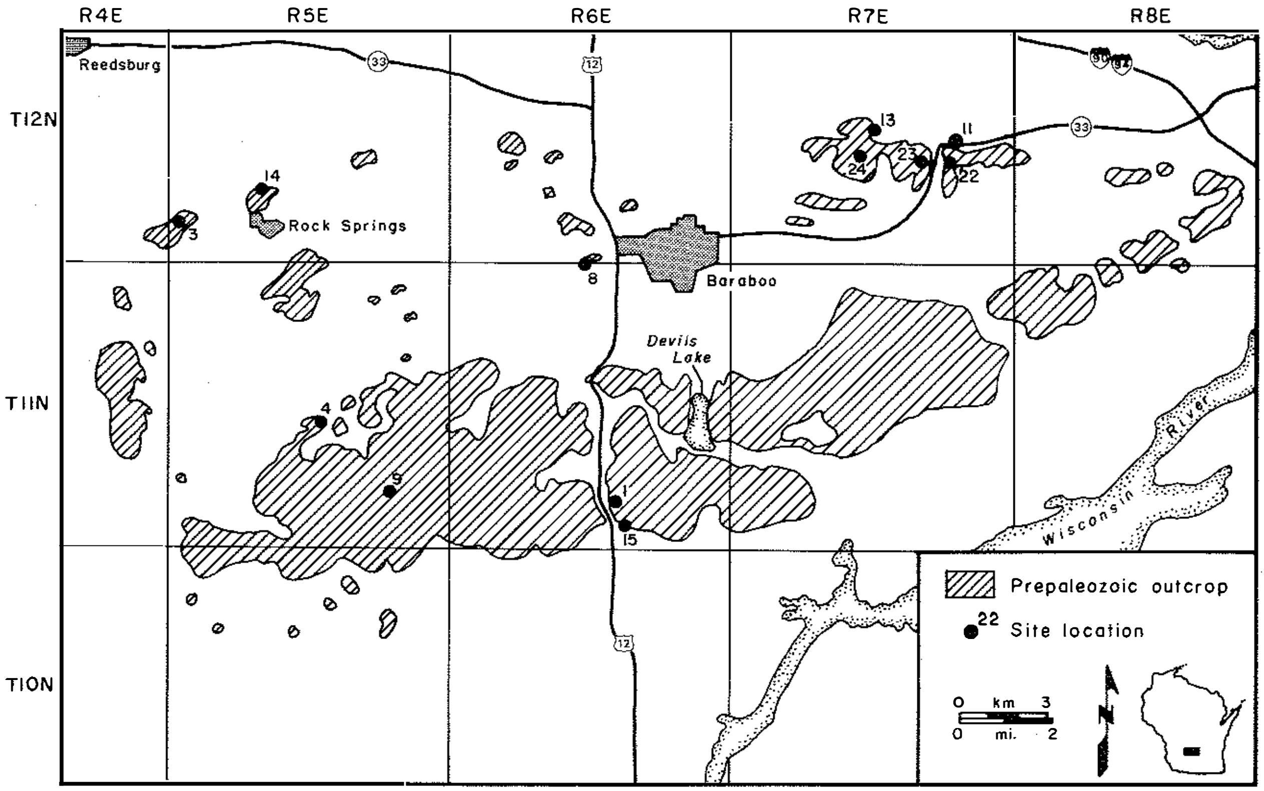

Figure 1.--Outcrop map of the Baraboo Quartzite modified from Dalziel and Dott (1970). Sample location numbers correspond to samples listed in table 1.

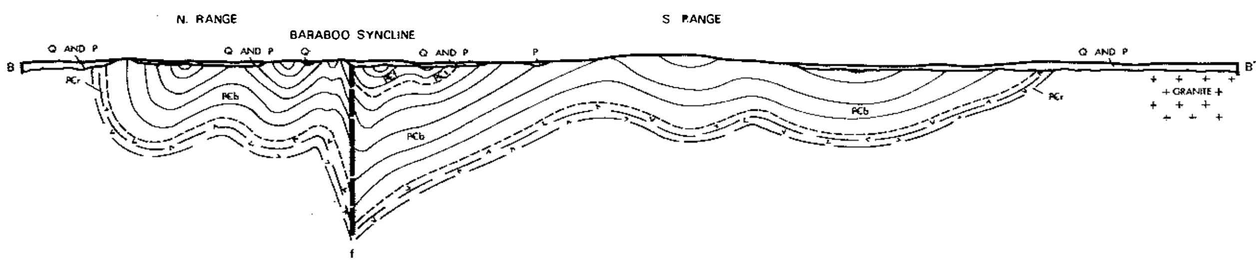

Figure 2.--Geologic cross section of the Baraboo Syncline (from Dalziel and Dott, 1970). The cross section is from the northwest to the southeast on the western end of the map, figure 1. PCr = rhyolite, PCb = Baraboo Quartzite, PCs = Seeley Slate, PCf = Freedom Formation, P = Paleozoic sedimentary rocks, Q = Quarternary deposits.

The results of the Fisher averages for all the sites are presented in table 1 and figure 5 and 6. The data is listed as original orientation, rotated, and rotated and unfolded. The original strike and dip of beds at the sampling locations are also given. This shows the changes in the mean declination, inclination, and data quality with progressive removal of the structure. The data is divided into two groups. The grouping was originally based on the similarity of the pole positions before structural corrections were made. The major difference between these two data sets is in the improved quality of the data in group I with the application of the structural corrections (table 2). For group I the precision parameter κ increases from 4.1 to 18.9 when the structure is removed. In group II the precision parameter κ improves only slightly (3.4-5.8). In addition, when the mean pole positions for group I and II are compared after the structure has been removed, they represent two distinct pole positions at the 95 percent confidence level.

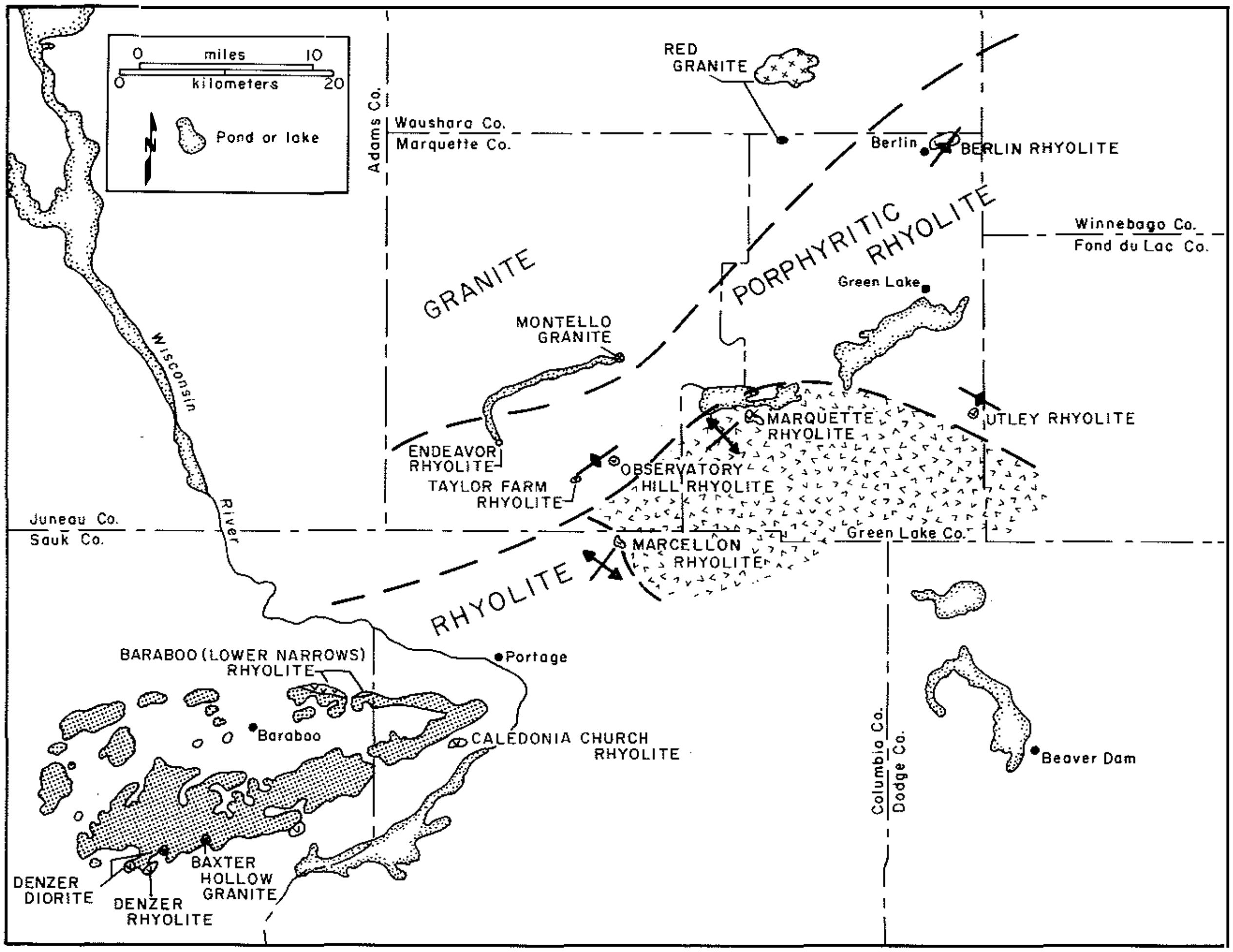

Figure 3.--Associated granite and rhyolite outcrop in the vicinity of Baraboo, Wisconsin (from Smith, 1978).

Data for group I come from the western end of the syncline and consist only of quartzite samples. Data for group II are from the eastern end of the syncline, where the quartzites are more deformed, and adjacent to outcrops of rhyolite. Group II also includes some phyllites, which are magnetically softer (Kean and others, 1983).

INTERPRETATION

The overall data quality improves as the structural corrections are made on the group I data. This provides a positive fold test (Graham 1949), which implies that the magnetization was set in previous to the major folding of the quartzite. This in itself is a significant finding. The thermal demagnetization results indicate that the only magnetic carrier of the magnetism in the quartzite is hematite. The only likely time of acquisition is during the original formation of the hematite, by oxidation processes during or soon after the deposition of the sand. The possibility of a major heating event after the deposition and before folding cannot be ruled out. However, it would require temperatures on the order of 600 °C to cause a post-deposition remagnetization of the hematite. There is no indication from any geologic studies that this area was ever heated above 250 °C after 1,760 Ma (Smith, 1978).

The data from group II are more problematic. The removal of the structure in this area does not statistically improve the quality of the data, and it does not move the magnetic directions closer to that of group I data.

Table 1.--Statistical data for the samples from all the sites indicated on Figure 1. **Asterisk** indicates overturned beds.

Sample Location	Strike & Dip	No. of Samples	No. of Specimens	Before Structural Correction Dec.	Inc.	Kappa	After Structural Correction Dec.	Inc.	Kappa
1	N.90°E., 14°N.	2	4	194	45	82.9	199.9	58.7	27.5
3	N.67°E., 83°NW.a	3	8	276	61	7.0	197	17.1	8.0
4	N.80°E., 15°NW.	2	7	145	55	48	139.9	67	49.6
8	N.66°E., 45°SE.	1	3	168	75	1500	183	30	8552
9	N.15°E., 21°NW.	1	3	88	39	178	157	59	134
14	N.68°E., 86°NW.a	2	6	346	57	15	176	30	15
15	N.8°W., 3°SW.	1	3	258	53.4	60.2	176.9	50.6	64.7
11	N.62°E., 84°NW.a	2	4	63	68	13	165	-2.9	12
13	N.73°E., 83°NW.a	2	3	30	63	16	160	11	16
22	N.83°E., 65°NW.a	1	3	352	70	345	182	-5	282
23	N.91°E., 75°NE.a	1	3	166	65	44	181	-39	46
24	N.98°E., 78°NE.a	1	3	127	63	14	171	-34	14

a - overturned bed

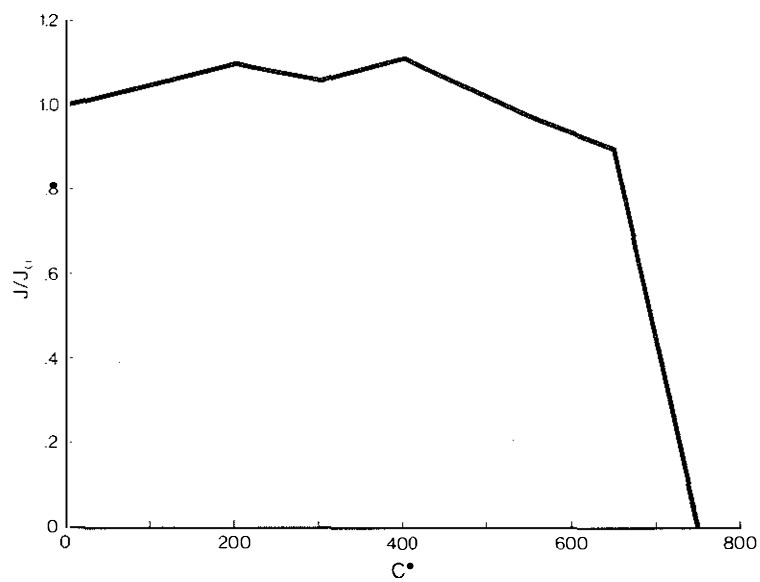

Figure 4.--Thermal demagnetization of Baraboo Quartzite, J/Jo is magnetic intensity normalized to the initial intensity.

When the pole positions from group I and group II samples are compared to published paleomagnetic poles for the time interval 1,800 Ma to 1,400 Ma, certain conclusions can be made. The pole position for group I data after the structural corrections are made (that is, 19° S, 89° W) is located in the general vicinity of most North American paleopole positions for the time range of 1,800 Ma to 1,700 Ma as seen in figure 7. The polar-wander path on figure 7 (from Irving, 1979) was chosen for comparisons. However, as was pointed out by Roy (1983), the limited amount of data for this time interval warrants considerable caution in assigning a definitive date to a new pole position based on currently drawn polarwander paths. The pole position for Group II data is much more enigmatic. At present, it does not fit well with any presently proposed polarwander curve.

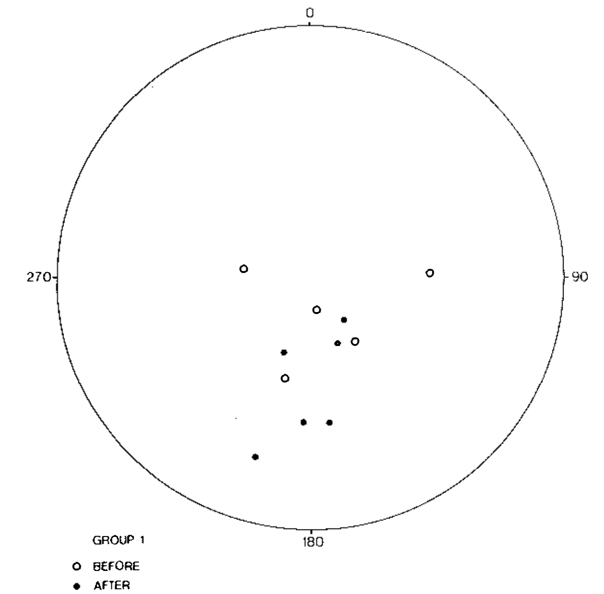

Figure 5.--Stereographic plot of declination and inclination for group I data, before and after structural corrections.

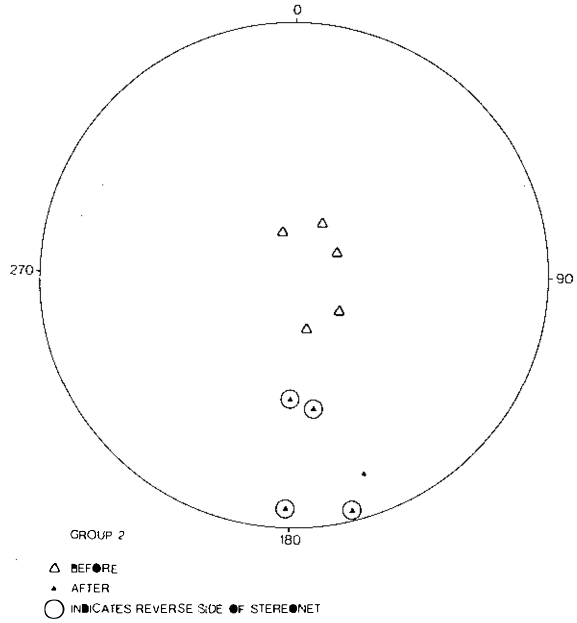

Figure 6.--Stereographic plot of declination and inclination for group II data, before and after structural corrections.

Table 2.--Statistical data for group I and group II data, before and after structural corrections were made.

	Mean Dec.	Mean Inc.	Kappa	α	VGP	dp	dm
Group I							
Before	193.5	80.3	4.7	31	25°N., 94.3°W.	57	10.3
After rotation	182	70	5.8	27	8°N., 90.7°W.	41.0	15.9
After unfolding	179	46	12.9	17.4	19.1°S., 89°W.	14.3	15.5
12 samples 34 specimens							
Group II							
Before	77	78	11	23	44°N., 57°W.	41.5	41.4
After rotation	89.3	75	13.9	21.2	37°N., 58°W.	35	38
After unfolding	165	-12.6	13.5	22	50°S., 66°W.	11	22
7 samples 16 specimens							

It may be a problem of improper removal of the structure or the results of a heating event. We expect a more definitive answer to this when additional samples are measured and when our study of the associated rhyolite and granite is completed. Initial results from some of the rhyolite give pole positions near to those of Group II (Kean and others, 1983).

CONCLUSIONS

A preliminary paleomagnetic study of the Baraboo Quartzite has produced the following results: (1) the magnetism of the Baraboo Quartzite is due primarily to hematite which formed prior to the major folding in the area; (2) two dominant pole positions are found, one definitely predates the folding based on a positive fold test. The pole position is consistent with published Precambrian poles for North America in the time range of 1,850 to 1,700 Ma (Irving, 1979). A second pole position is also noted but not understood. This second pole position is similar to some pole positions obtained for a few of the associated granite and rhyolite in the area. However, additional work is needed to confirm this; and (3) this initial study suggests that paleomagnetic studies will be useful in deciphering the geologic history of the Baraboo interval rock of southern Wisconsin.

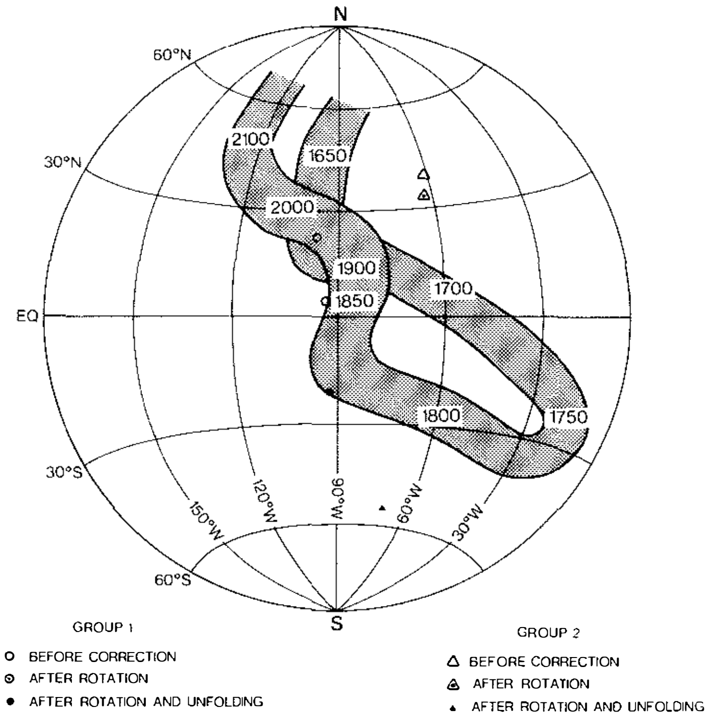

Figure 7.--Paleopole positions for group I and group II data in relation to Irving's (1979) polar-wander path.

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PRESSURE SOLUTION AND CLEAVAGE DEVELOPMENT IN THE BARABOO QUARTZITE

by

Mary E. Jank[1] and F. William Cambray[2]

ABSTRACT

This study demonstrates two main points. The first is that the shape of the quartz grains makes a significant contribution to the development of penetrative cleavage in both quartzite and phyllite. The second is that the penetrative cleavage becomes more defined and the spaced cleavage becomes more common with increasing phyllosilicate content. Samples from two localities on the south limb of the Baraboo Syncline were sectioned at right angles to the bedding cleavage intersection. The length to width ratio of the quartz grains and the orientation of the long axes were recorded as a function of the phyllosilicate content. The spacing of the spaced cleavage was also measured as a function of the phyllosilicate content. In quartz-rich horizons the grains have low axial ratios and exhibit a long-axis orientation close to bedding. With increasing phyllosilicate content the axial ratio increases and a long-axis orientation distinct from bedding is observed. The latter is parallel to the spaced cleavage where this is present. The long axes of the grains and the spaced cleavage both curve toward the bedding (refraction) as the phyllosilicate content increases, and the distinction between bedding parallel orientation and cleavage parallel orientation of the quartz grains becomes difficult to measure.

In an analogy with the development of pressure solution in sedimentary rock it is suggested that the increase in the length to width ratio of quartz grains and the spacing of spaced cleavage with the increase in phyllosilicate content supports the origin of the cleavage by pressure solution. The presence of truncated grains, pressure shadows, and interpenetrative grains provides additional support. The concentration of sulphide and zircon in the spaced cleavage planes suggests that they too are related to a process of selective solution transfer.

INTRODUCTION

The Baraboo Quartzite of Wisconsin is a relatively small outcrop of Proterozoic quartzite deposited sometime between 1,760 Ma and 1,630 Ma (Van Schmus and Bickford, 1981). It is isolated from other Proterozoic rocks by Paleozoic cover (fig. 1). Correlation, age relationships, and tectonic setting of the quartzite are obscure, but the structures have been of interest for geologists since the last century (Van Hise, 1893). The descriptive classification of cleavages into slaty, fracture and strain slip (penetrative, spaced and crenulation cleavage in modern usage) was well demonstrated here, and continued investigations have attempted to determine the mechanism of cleavage formation using rock from this locality (Dalziel and Dott, 1970; Dalziel and Stirewalt, 1975). The cleavage investigated in this study consists of a penetrative structure controlled by tectonically flattened quartz grains and aligned phyllosilicates, S_1' (Dalziel and Dott, 1970) and a spaced cleavage, often referred to as a fracture cleavage (Irving, 1877; Steidtman, 1910; Leith, 1913, 1923) which consists of closely spaced, but nonetheless discrete, surfaces formed by concentrations of aligned phyllosilicates (S_1' of Dalziel and Dott, 1970). The phyllosilicates consist of pyrophyllite and muscovite (Dalziel and Dott, 1970).

The concept of pressure solution is an old one (Thomson, 1862a,b) and in some of the older geologic literature it was referred to as the Riecke Principle (Riecke, 1912). Sorby applied the concept of pressure solution to several of his geologic observations. His studies still form the basis of many of the more modern concepts (Sorby, 1879; Kerrich, 1977). In general, it is thought that a body of rock in a state of non-hydrostatic stress will preferentially undergo solution on surfaces normal to the maximum principle stress direction (σ_1) even without precipitation of the dissolved material on the surfaces of the grains which are normal to the minimum principle stress direction (σ_3); the grains would tend to become more elongate in the σ_3 direction (fig. 2). If precipitation does occur, then the effect is enhanced and structures which are commonly called pressure shadows are formed (fig. 2). Such a mechanism could be responsible for the development of penetrative cleavage in a variety of lithologies. While it is a theoretically sound mechanism it is difficult to test its role in a natural situation. An alternative mechanism might be termed strain solution. As stress builds up at the grain contacts normal to the σ_1 direction, the points of contact will become regions of potentially high strain. If the strain is in the form of increased dislocation density, then the potential for an increased rate of dissolution exists at these points. The two mechanisms could be distinguished by using the transmission electron microscope to look for the distribution of dislocations. However, the dislocations have

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a tendency to anneal and recrystallize under the conditions thought to exist during metamorphism. There is also the possibility that the grain shape variation may be due to homogeneous distortion of the grains. It would be hard to explain the increase in the axial ratio of the grains in areas of high phyllosilicate content if this were the case. The correlation between grain shape and phyllosilicate content and spacing of spaced cleavage and phyllosilicate content suggests to us that pressure solution has played an important role in the production of spaced and penetrative cleavage in the Baraboo Quartzite.

It has been proposed that the presence of clay minerals increases the potential for pressure solution (Heald, 1956; Weyl, 1959; Thompson, 1959; Greseno, 1966; Durney, 1972b; Sibley and Blatt, 1975; Kerrich and others, 1975; Bosworth, 1981). These authors based their ideas on petrographic observations of the concentration of clay minerals in rock containing structures which they thought might be formed by pressure solution. The theoretical basis for the observations is difficult to test but several suggestions have been made. It is generally thought that the migration of the dissolved components is the rate controlling factor in the process. With this in mind it has been proposed that the presence of clay films might provide pathways for transport (Weyl, 1959; Sibley and Blatt, 1975). Clay could also be responsible for local chemical variations such as a high pH (Thompson, 1959) or act as catalysts (Heald, 1956). Hydrolytic weakening of quartz as a result of the water associated with the clay has also been suggested (Weyl, 1959; Jones, 1975; Hobbs, 1981). A good review of the role of pressure solution in geologic processes can be found in Kerrich (1977).

The association of pressure solution with phyllosilicate and the development of cleavage in greenschist-facies metamorphic rock is the focus of this paper. The length to width ratio and the orientation of the long axes of the phyllosilicate were recorded together with the phyllosilicate content. In addition the spacing of the spaced cleavage was recorded. The hypothesis to be tested is that if pressure solution is a function of phyllosilicate content then the length to width ratios should increase, the long axis orientation should become better defined, and the spaced cleaving should become more dense with an increase in phyllosilicate content. The success of this test does not mean that pressure solution is the operative mechanism, only that it could be under the conditions described here.

There are petrographic features which also suggest selective removal and precipitation of quartz. Quartz grains which are adjacent to the spaced cleavage bands commonly have flat margins facing the band and a more rounded shape on the side facing away from the band (fig. 3a). In the area between the spaced cleavage the quartz grains have delicate finger-like overgrowths of quartz extending from the grains in the direction of the longest dimension (fig. 3b). These pressure shadows do not extend from surfaces which are parallel to the cleavage; such surfaces tend to be smooth. Finally, there is a higher concentration of opaque minerals in the spaced-cleavage zones compared with the rest of the rock.

The data were plotted using SURFACE II GRAPHICS SYSTEM of the Kansas Geological Survey (Sampson, 1978), a licensed package of software.

PROCEDURE

Large, oriented specimens were collected from two localities--one set from the LaRue quarry (SW¼NW¼ sec. 22, T. 11 N., R. 5 E.) and the other set from an outcrop near Lake Jerdean (SW¼SW¼SE¼ sec. 15, T. 11 N., R. 5 E.) (fig. 1). The samples contain a complete cross section of a bed which grades from a phyllite at the base to a quartzite in the center and back to a phyllite at the top. The phyllitic horizons exhibit a well developed penetrative foliation lying close to the bedding. The quartzite shows a well developed spaced cleavage consisting of narrow seams of phyllosilicate which are at a high angle to the bedding in the quartz-rich parts and progressively approach parallelism with the bedding as the quartz content decreases toward the top and bottom of the layer.

Several sections were cut from each bed form in order to obtain a complete cross section of the bed. The sections were cut at right angles to the bedding-cleavage intersection. This was done on the assumption that the same cross section of the finite strain ellipsoid was being examined in each case. The Baraboo Syncline has an essentially horizontal hinge line at the localities chosen, and the bedding cleavage intersection is parallel to it. It is assumed for the purpose of this study that the sections are close to the X-Z plane of the finite-strain ellipsoid and therefore exhibit the maximum strain ratio. It is realized that this is not always the case, but the simple nature of the Baraboo Syncline justifies the assumption for the purpose of this study. A more complete three-dimensional study to determine strain variation is now underway.

Point counts were made on each thin section to determine the ratio of quartz to phyllosilicate (1000 points per section). In addition the length to width ratio of the quartz grains was measured together with the orientation of the long axis (150 grains per section). The spacing of the spaced cleavages was measured for the same specimens. These results are plotted on figures 4 through 13. Sixteen samples were measured; for consideration of space only eight are shown in detail on figures 6 through 13. The interchange of stress and strain terms does not mean that we have proved any geometric similarities; on the other hand, we were not able to define any differences.

RESULTS AND CONCLUSIONS

The simplest measurement to make is the spacing of the spaced cleavage as a function of lithology. It is obvious in the field that the spaced cleavage is less common in the quartz-rich centers of the beds than in the phyllitic margins. Although there is a wide spread in the data, the spaced cleavage in the LaRue quarry varies from zero to five per centimetre over the range of phyllosilicate content recorded (fig. 4) with a correlation coefficient of 0.709 ($p<0.01$, df = 74). Fewer measurements were made at Lake Jerdean but a similar trend was found with a correlation coefficient of 0.894 ($p<0.01$, dfn= 7) (fig. 4b). The spaced cleavage consists of phyllosilicate seams varying in width from 0.1 to 1 millimetre. Concentrations of opaque and heavy minerals are found in these seams. This last mentioned feature is independent evidence in support of the idea of pressure solution. It is proposed that the quartz has been selectively removed during deformation leaving behind the phyllosilicate, the opaque and heavy minerals. This process was more effective in the phyllosilicate-rich horizons for reasons outlined in the introduction.

The shape of the quartz grains shows a progressive increase in length to width ratio with increasing phyllosilicate content (fig. 5). The long axes of the quartz grains change from a bedding-parallel orientation to one which becomes increasingly parallel to that of the spaced cleavage. It is interesting to note that it is still common in introductory books to state that penetrative (slaty) cleavage is only common in rocks rich in phyllosilicates and is due to the alignment of the phyllosilicate cleavage (Press and Siever, 1982). This study clearly shows that the shape of the quartz grains makes a major contribution to the penetrative cleavage in the Baraboo Quartzite. The easiest way to represent these results is by plotting the axial ratio against the orientation of the long axis with respect to bedding (figs. 6b through 13b) and the

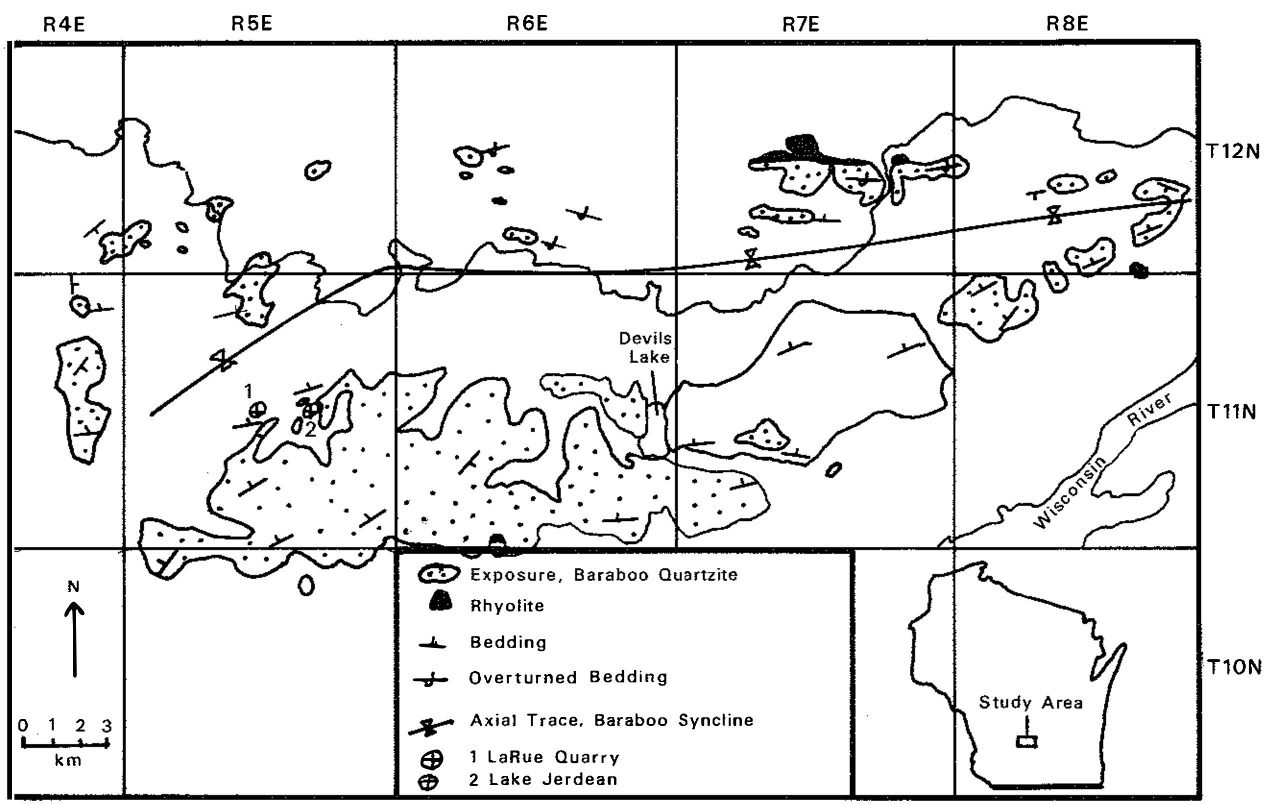

Figure 1.--Map of Baraboo Quartzite (after Dalziel and Dott, 1970).

frequency of a given angular relationship (fig. 5a through 12b). The series of block diagrams (figs. 6c through 13c) show the same data in pictorial form. On each diagram bedding is the 0° orientation for reference. The series of thin sections representing the LaRue quarry show that for a low phyllosilicate content, 5.8 percent, there is a strong concentration of long axes parallel to the bedding (fig. 6a). The plot of the distribution of axial ratios as a function of orientation shows that there is a broad distribution of grains with axial ratios between 1 (circular) and 2, and that there is a small maximum of grains with larger axial ratios close to the bedding orientation (fig. 6b). The main features of this sample are grains with low axial ratios, a preferred orientation of long axes in the bedding plane, and a tendency for the more elongate grains to lie in the bedding plane. The specimen containing 38.6 percent phyllosilicate shows a concentration of long axes distinct from the bedding orientation (fig. 7a). The ratioplot (fig. 7b) illustrates the change in orientation of the long axes and the increase in axial ratio of grains with this orientation. In addition, the mean orientation of the spaced cleavage and its range are plotted. It can be seen that the spaced cleavage and the orientation of the long axes of the quartz grains have similar trends. These observations are further illustrated in the block diagram (fig. 7c). The specimen with 47.7 percent phyllosilicate (fig. 8) shows a more distinct concentration of quartz-grain long axes separate from the bedding and again subparallel to the spaced cleavage. The increase in axial ratio of grains with this orientation is also more clearly defined. The specimen with the highest phyllosilicate content at the LaRue Quarry, 70.6 percent (fig. 9) shows the largest axial ratios. The orientation of quartz-grain long axes still shows a distinct maximum parallel to the spaced cleavage. However, because of the refraction of the cleavage, the bedding and cleavage are almost parallel at this point. The samples analyzed from the Lake Jerdean locality were taken from one large specimen. They show the same pattern of quartz-grain long-axis orientations and axial ratios in relation to the bedding and spaced cleavage. The sample with the lowest phyllosilicate content, 27.5 percent, shows a broad distribution of long axes and low ratios (fig. 10). There is a weak maximum close to, but not parallel to, the bedding. There is also a well developed spaced cleavage in the specimen almost at right angles to bedding. The section containing 38.1 percent phyllosilicate (fig. 11) shows a marked change; there

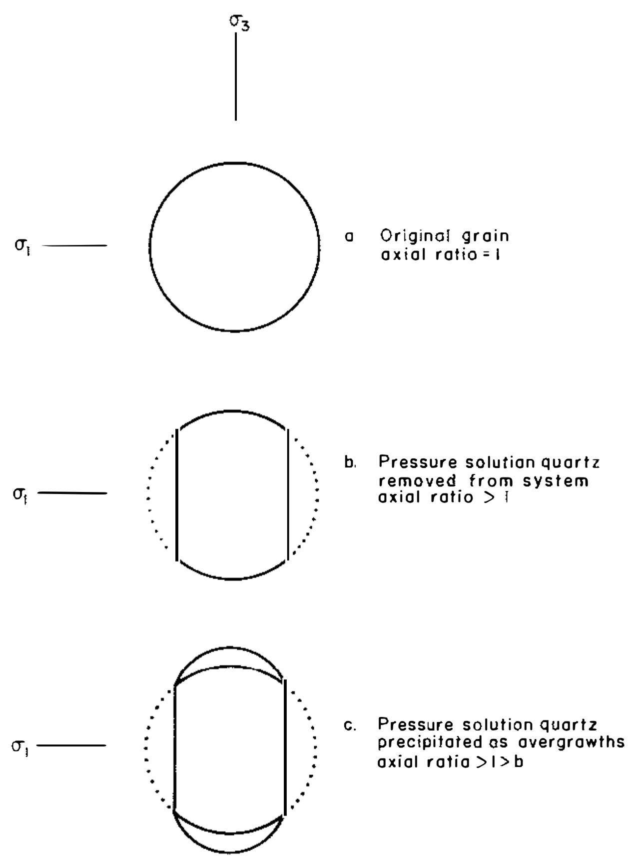

Figure 2.--Effect of pressure solution when quartz is removed from the system (b) or stays within the system (c).

is a concentration of long axes parallel to the spaced cleavage and an increase in axial ratio of grains with this orientation. In the sample containing 56.1 percent phyllosilicate (fig. 12) the maximum orientation of the grains with the largest axial ratio is still parallel to the spaced cleavage though the pattern is more diffuse. At the margins of the bed where the phyllosilicate content reaches 84.0 percent (fig. 13), the axial ratios show a further increase with a strong concentration of the long axes parallel to the spaced cleavage which at this point has been refracted so that it lies close to the bedding.

The main conclusions are that both the penetrative and the spaced cleavages in the Baraboo Quartzite become stronger with increasing phyllosilicate content (fig. 4 and fig. 5). This supports the concept of pressure solution as a mechanism for the origin of both cleavages by analogy with the studies of pressure solution in sedimentary rock. The penetrative cleavage is formed by a mixture of quartz grain-shape and phyllosilicate cleavage orientation. The spaced cleavage is

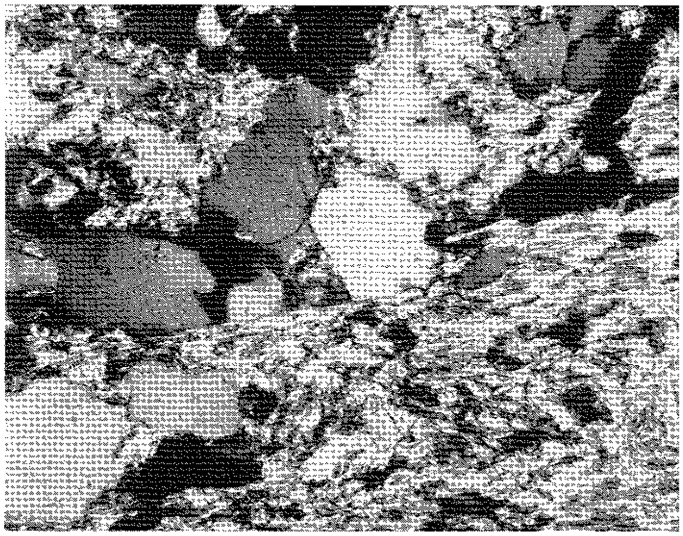

Figure 3a.--Straight-sided quartz grains abutting against spaced cleavage surface (lower left of photo). Note the irregular surfaces on parts of quartz grains not abutting against the spaced cleavage. (magnification x 160).

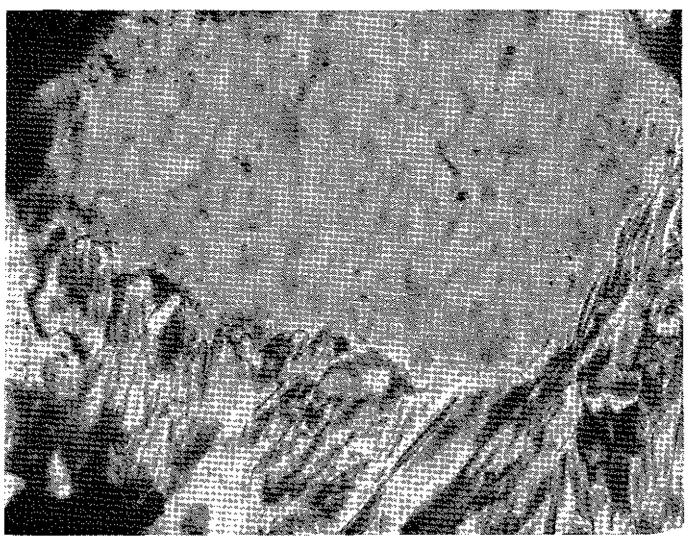

Figure 3b.--Fine overgrowths on quartz grain running parallel to the penetrative cleavage (bottom left to top right), (magnification x 620).

formed by selective removal of quartz along bands which form at regular intervals. It also demonstrates large strain variations within a single layer. There seems to be a critical level of about 20 percent present phyllosilicate content below which no spaced cleavage forms. The penetrative cleavage formed by the elongation of the quartz grains is also difficult to detect below this level.

There remains the possibility that the phyllosilicate content we have measured is entirely due to the strain. If we are to argue that the rock has deformed by removal of quartz as a result of pressure solution, then it is equally true to state that the high strain regions will inevitably become proportionally richer in phyllosilicate. We have not attempted to resolve this dilemma in the present study. If pressure solution was an important mechanism in inducing strain in the rock, then it is possible that the phyllosilicate-rich regions developed higher strain for the reasons suggested above and that in the course of this deformation became even more phyllosilicate rich.

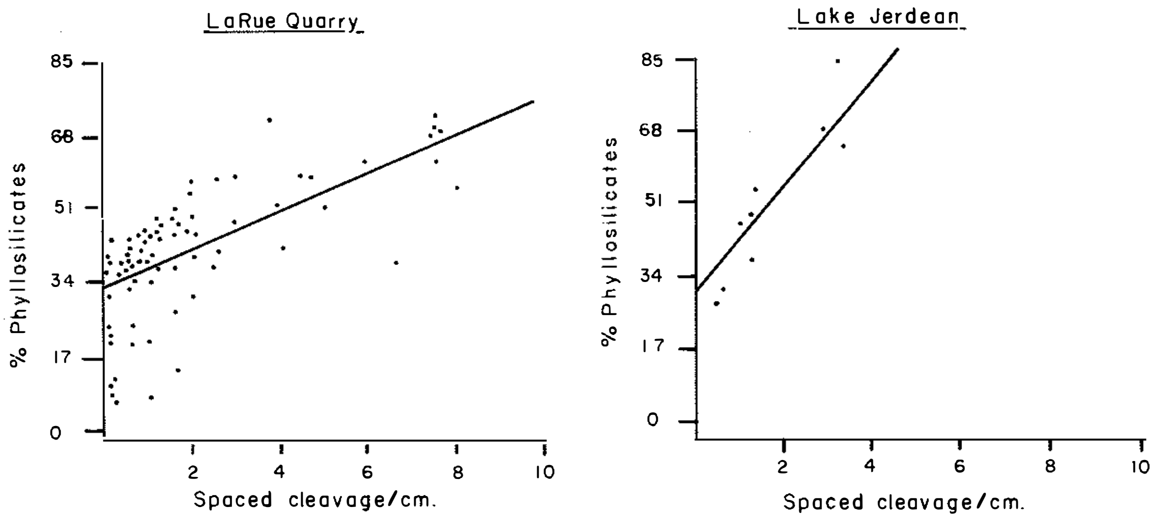

Figure 4.--Relationship between phyllosilicate content and spacing of spaced cleavage for the LaRue quarry (a) and the Lake Jerdean outcrop (b).

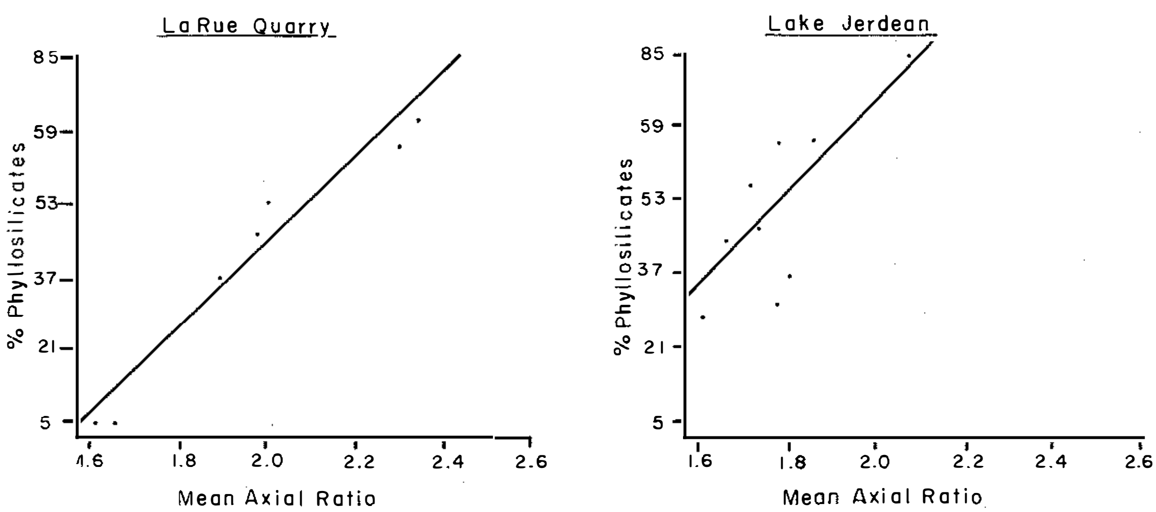

Figure 5.--The relationship between phyllosilicate content and mean axial ratio of quartz grains for the LaRue quarry (a) and the Lake Jerdean outcrop (b).

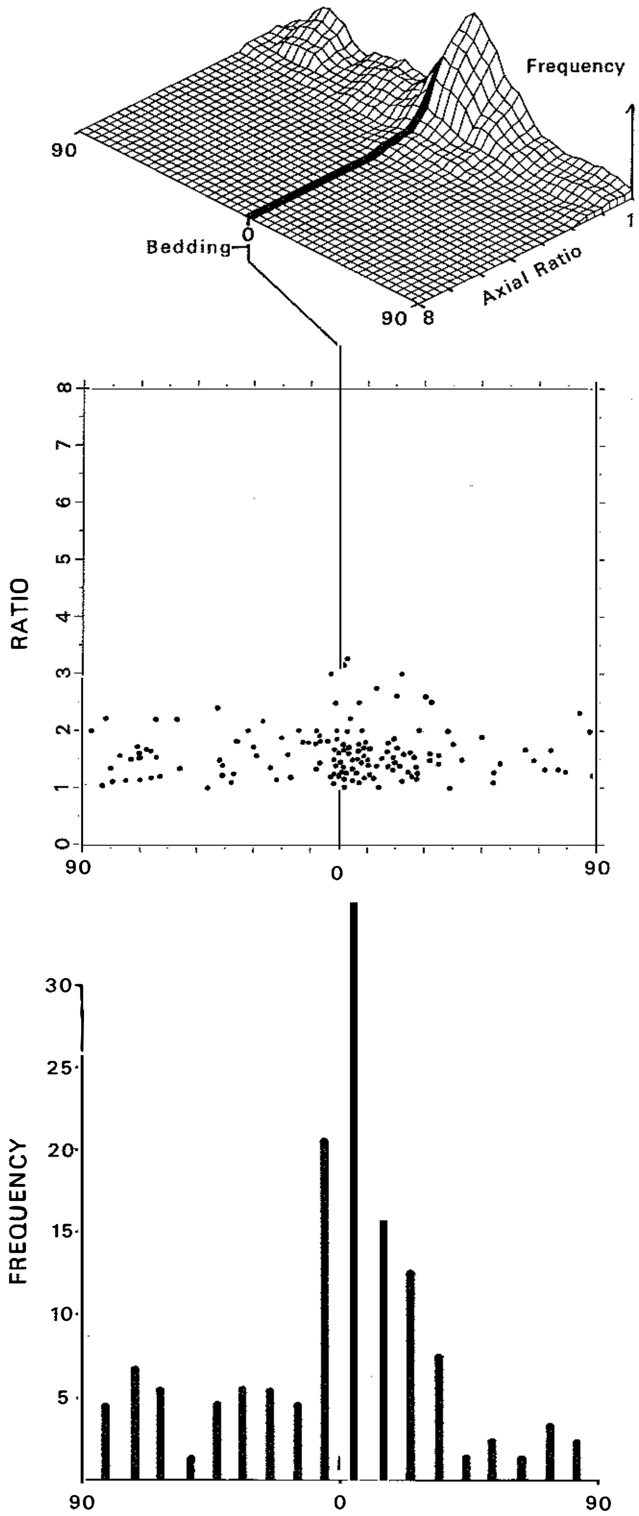

Figure 6.--Plot of axial ratio and orientation of long axis to bedding for LaRue Quarry (5.8 percent phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

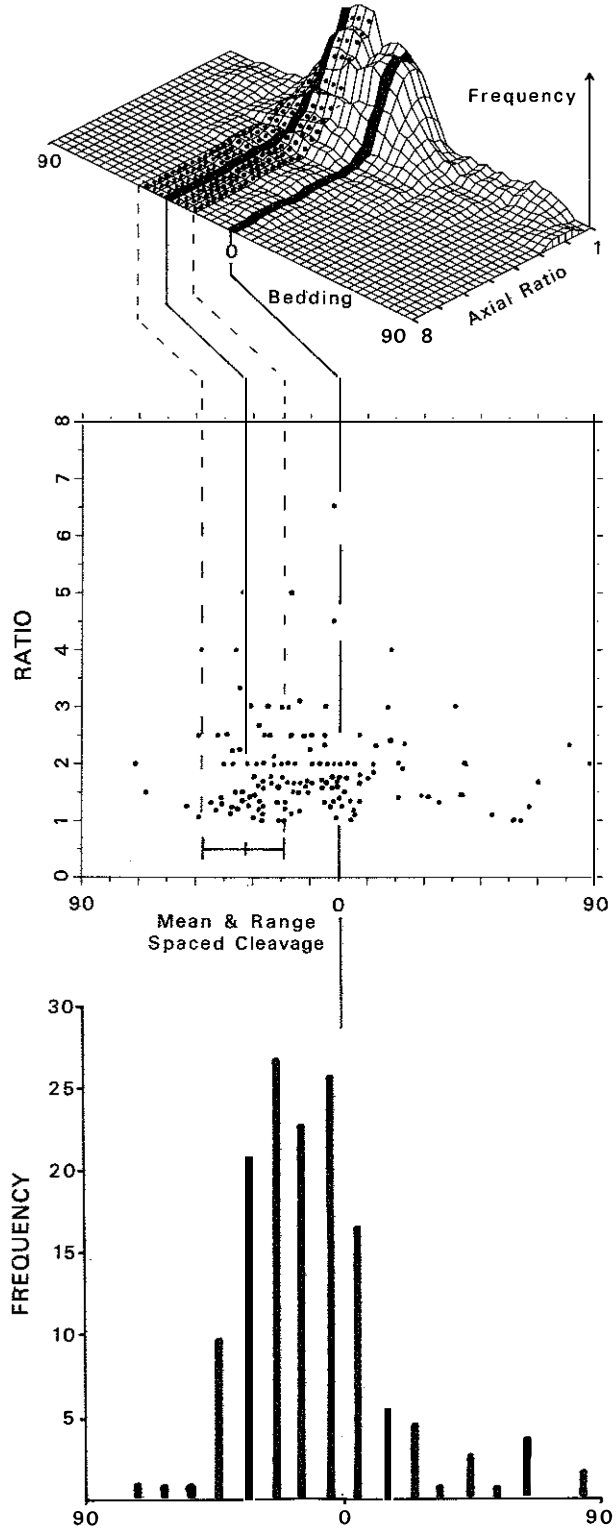

Figure 7.--Plot of axial ratio and orientation of long axis to bedding for LaRue Quarry (38.6 percent phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

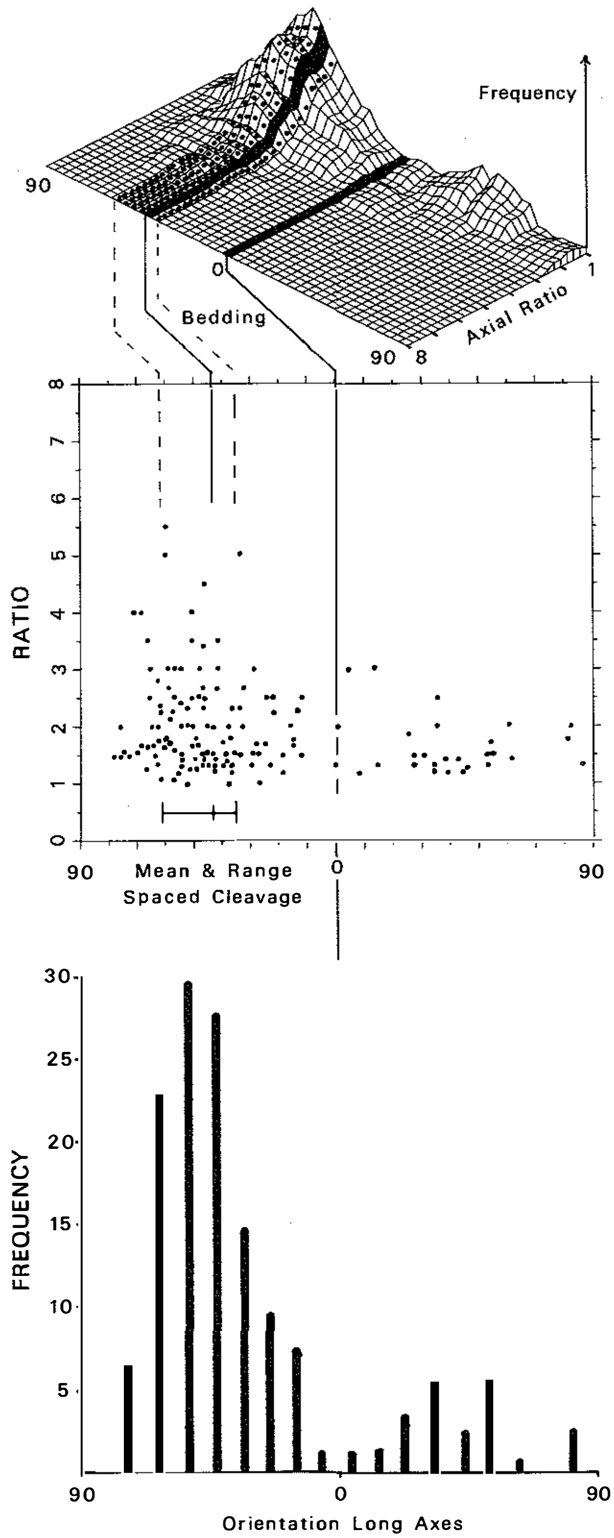

Figure 8.--Plot of axial ratio and orientation of long axis to bedding for LaRue Quarry (47.7% phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

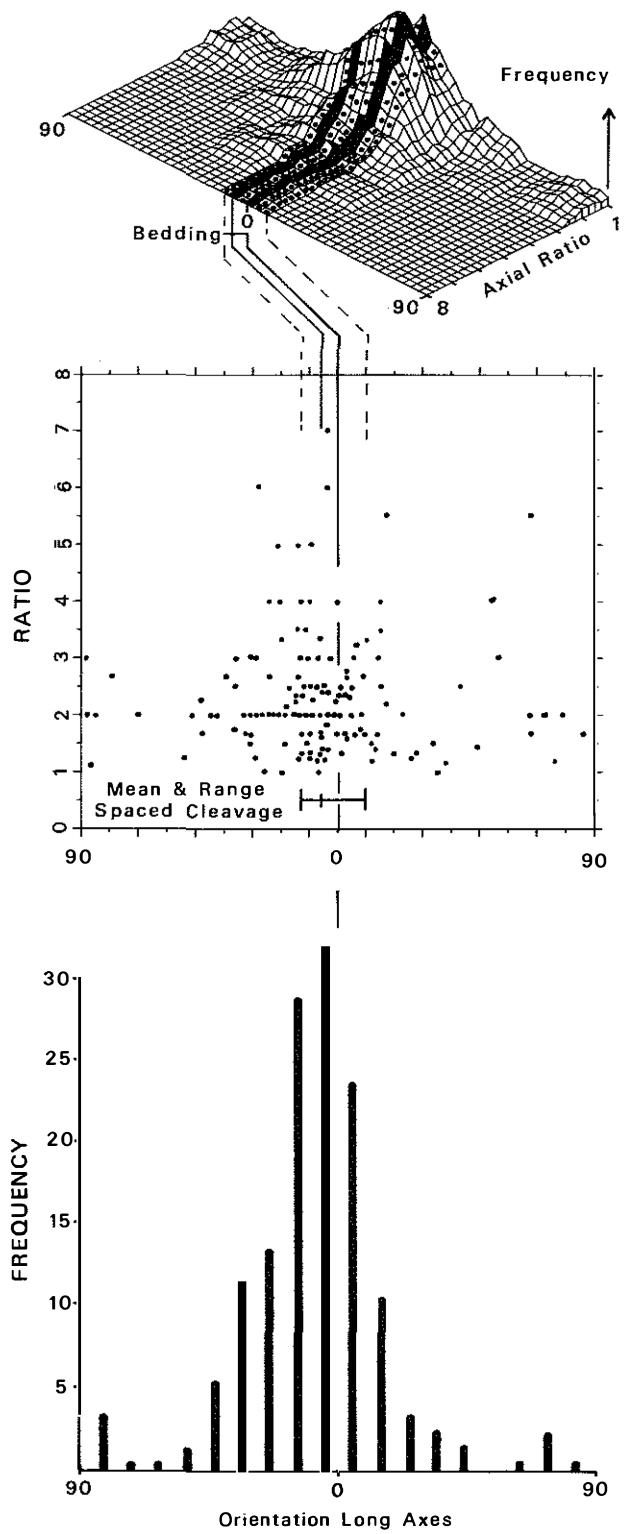

Figure 9.--Plot of axial ratio and orientation of long axis to bedding for LaRue Quarry (70.6 percent phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

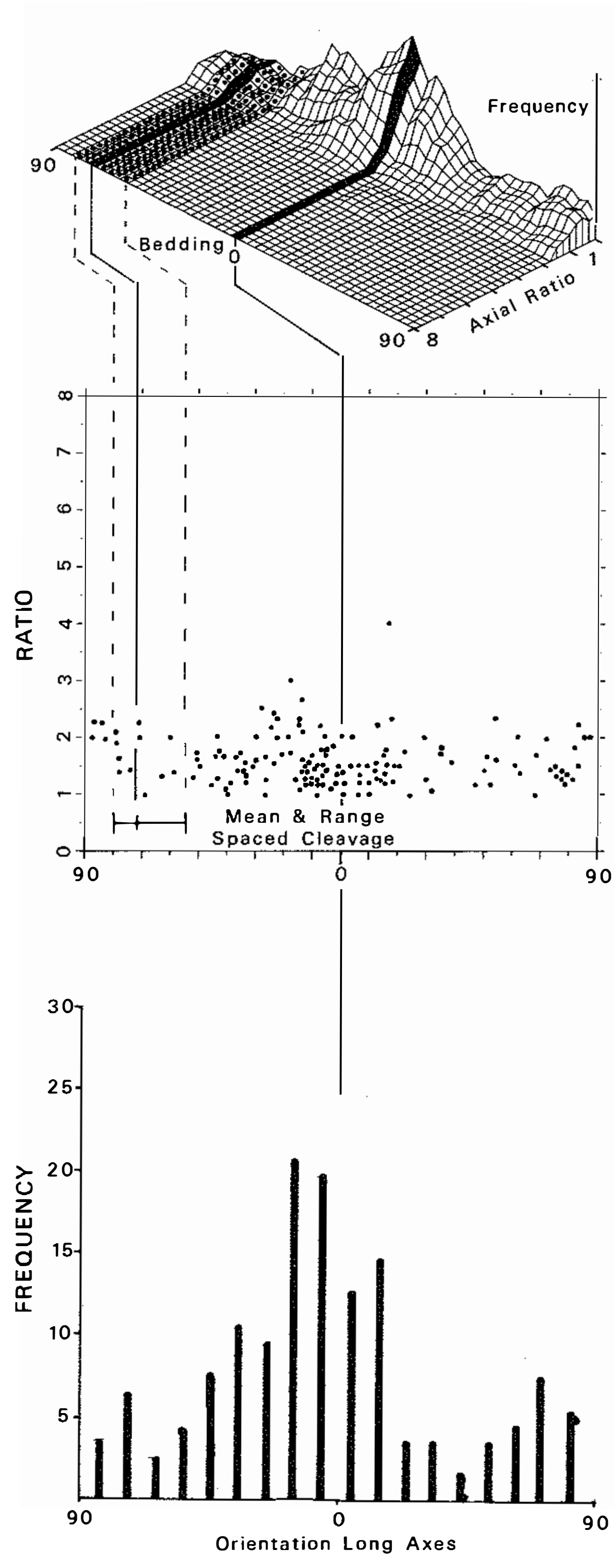

Figure 10.--Plot of axial ratio and orientation of long axis to bedding for LaRue Quarry (27.5 percent phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

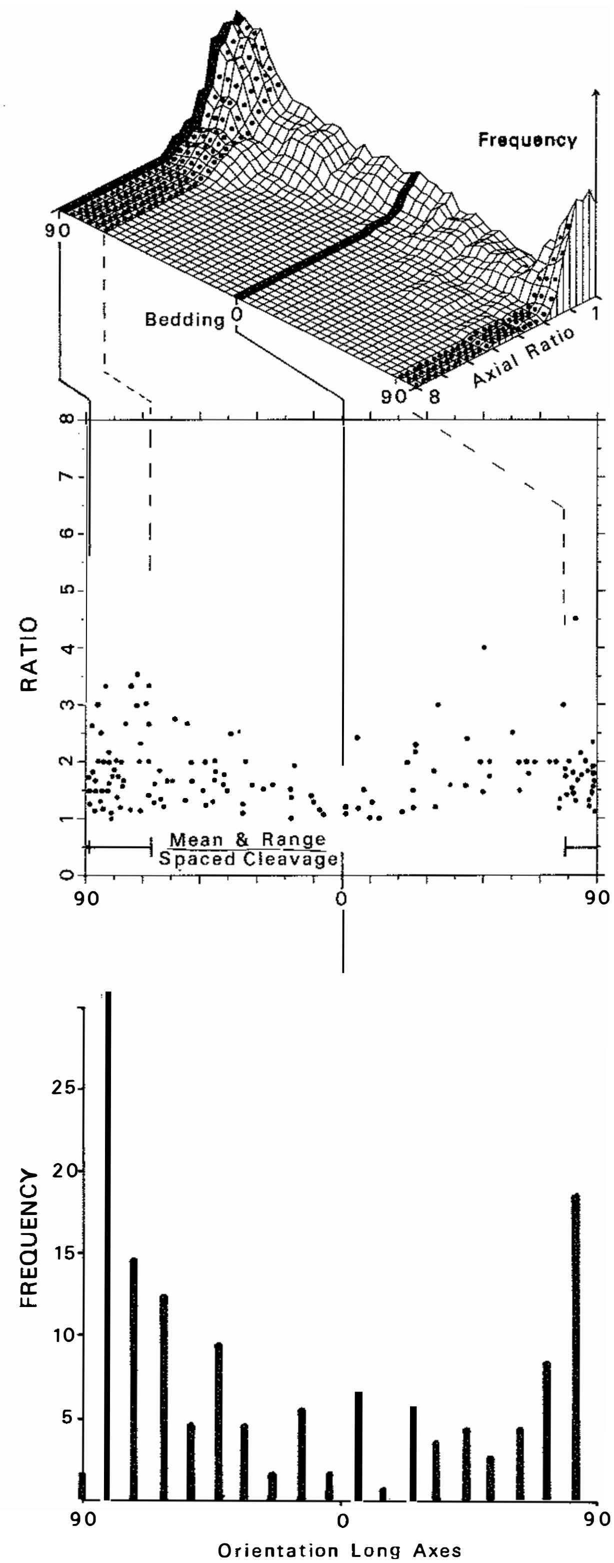

Figure 11.--Plot of axial ratio and orientation of long axis to bedding of LaRue Quarry (38.1 percent phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

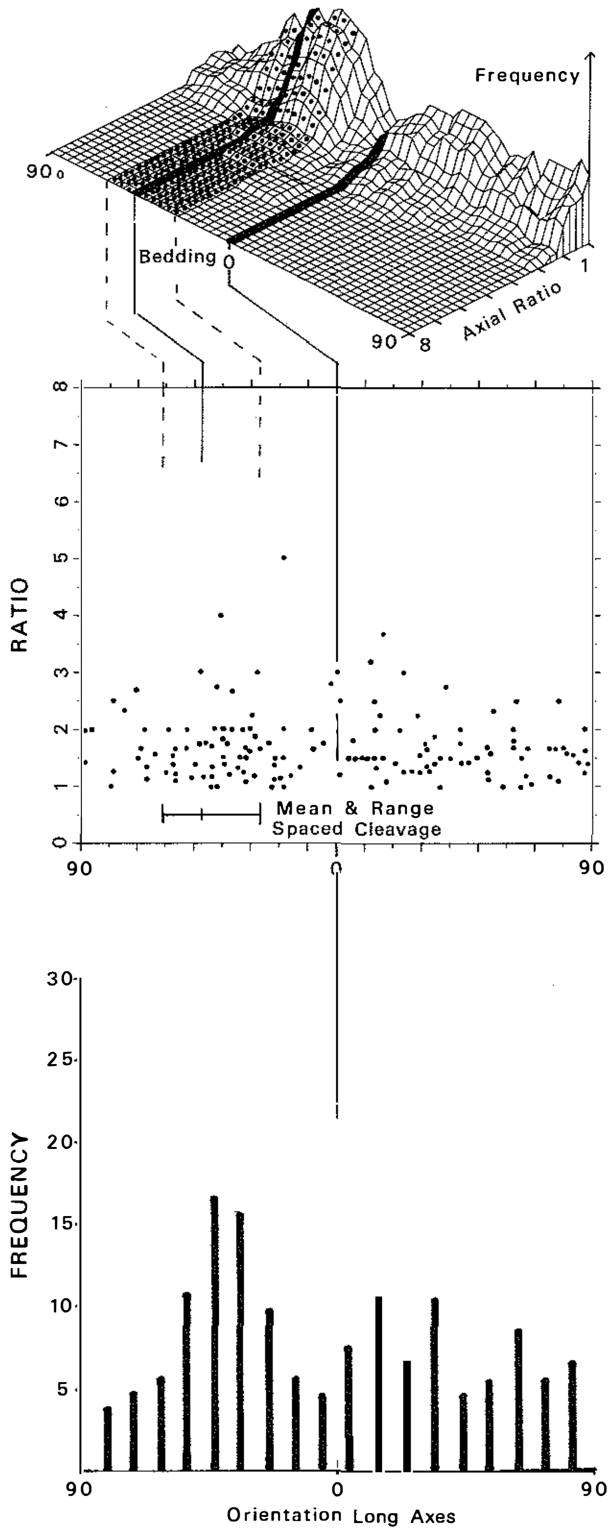

Figure 12.--Plot of axial ratio and orientation of long axis to bedding for LaRue Quarry (56.1 percent phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

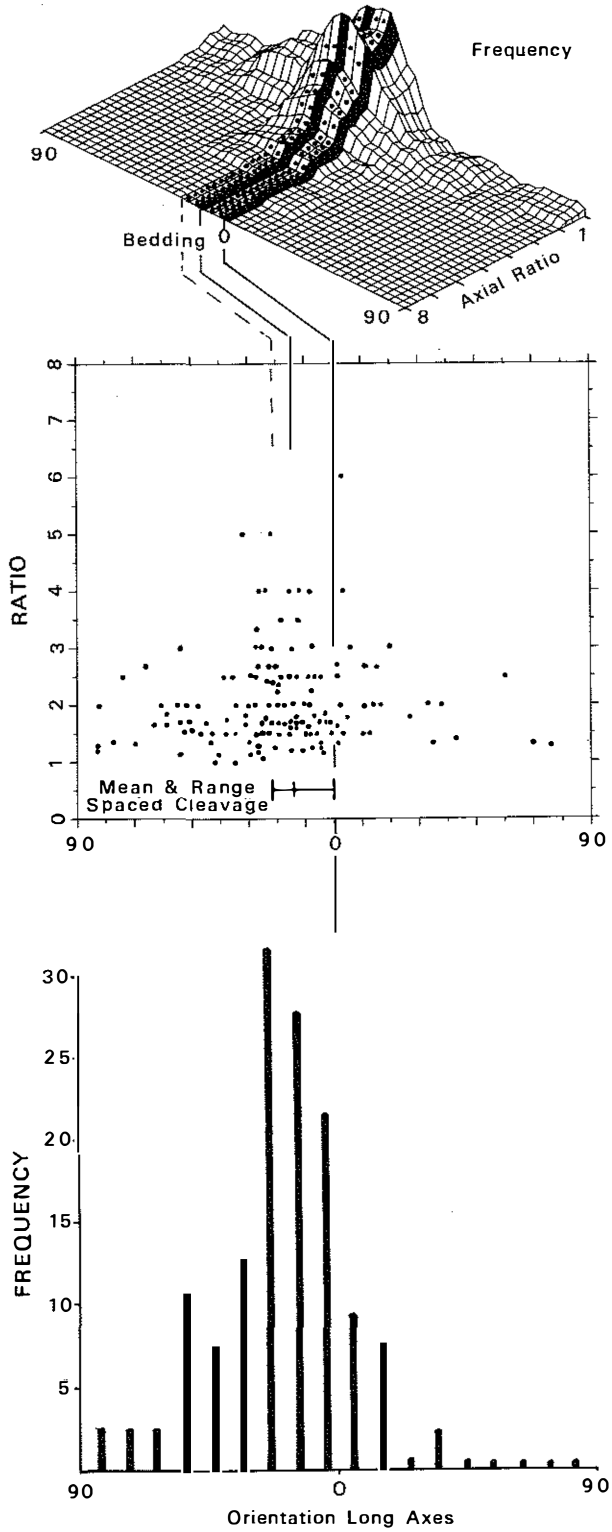

Figure 13.--Plot of axial ratio and orientation of long axis to bedding for LaRue Quarry (84.0 percent phyllosilicate): a) histogram of orientation; b) orientation to axial ratio; c) block diagram of a and b.

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The authors acknowledge invaluable discussions with students and faculty of the Department of Geological Sciences at MSU. In particular, one of us (FWC) would like to express thanks to Bill Monaghan for writing a computer program which made it possible to produce the block diagrams, and to John Wilband for general help in data reduction.

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ANALYSIS OF STRUCTURES WITHIN PHYLLITIC LAYERS OF THE BARABOO SYNCLINE, WISCONSIN: A NEW INTERPRETATION OF DEFORMATION HISTORY

by

Mark Hempton[1, 2], Mark Gordon[1], and David Kirschner[1]

ABSTRACT

Analysis of structures within phyllitic layers of the Baraboo Syncline suggests a deformation history involving two independent generations of shear strain. First-generation structures include (1) foliation parallel to layer boundaries, (2) foliation oblique to layer boundaries, (3) quartzite boudins, (4) asymmetric folds, and (5) rotated tension gashes. The orientation of these structures suggests that they formed by updip shear strain. Secondgeneration structures are superimposed on first-generation structures and include (1) crenulation cleavage, (2) kink bands, (3) chevron folds, (4) folded late-stage quartz veins, and (5) tension gashes. The orientation of these structures suggests an opposing downdip shear strain. We propose that (1) the first-generation structures were formed by updip shear strain related to the flexural genesis of the Baraboo Syncline during a 1,630-Ma deformation event, (2) the second-generation structures were generated by downdip shear strain related to opening of the Syncline in a northwest-southeast direction during Keweenawan rifting at 1100 to 1200 Ma.

INTRODUCTION

Early studies of the Baraboo Quartzite in south-central Wisconsin showed that it crops out as a large doubly-plunging syncline whose hinge line trends east-northeast (Weidman, 1904). Bedding/cleavage relations exhibited within minor phyllitic layers confirmed the synclinal geometry (Van Hise and Leith, 1911). It was proposed that the cleavage was axial planar and had formed during the formation of the syncline (Leith, 1923). On the basis of a microstructural study of the quartzite, Riley (1947) suggested that the Baraboo Syncline may have experienced a second phase of deformation. This idea was supported by Adair (1956) who reported anomalous structures in phyllitic layers that were apparently related to a second deformation event. Hendrix and Schaiowitz (1964) conducted a structural analysis of mesoscopic structures within the phyllite layers and concluded that there are two generations: (1) normal structures related to updip shear during the formation of the syncline; and (2) reverse structures caused by "downslope flow and thinning of the argillite under the load of the overlying quartzite." The latest interpretation of the deformation history of the Baraboo Syncline is that of Dalziel and Dott (1970) and Dalziel and Stirewalt (1975). They conducted a detailed field and microscopic study which led them to propose that all the mesoscopic and microscopic structures in the Baraboo Syncline could be explained as the products of a single progressive deformation event. This hypothesis was inspired by recent findings from orogenic belts which showed that overprinting structural relationships could form as progressive deformation phases within a single orogenic event (Dewey, 1969).

The purpose of this paper is to test the proposal of Dalziel and Dott (1970) and Dalziel and Stirewalt (1975) by following the methods of Hendrix and Schaiowitz (1964) and examining the mesoscopic structures within the phyllite layers in greater detail. The phyllitic layers have accommodated the most strain within the syncline and possess the best record of deformation history. Since the time of Hendrix and Schaiowitz (1964), Dalziel and Dott (1970), and Dalziel and Stirewalt (1975), many advances have been made in understanding the geometry of mesoscopic structures (Berthe and others, 1979; Logan and others, 1979; Simpson and Schmid, 1983). It is within this context that this study has been undertaken. We hope to shed new light on the kinematic significance of mesoscopic structures within the phyllite layers and to explore their implications for the deformation and tectonic history of the Baraboo Syncline.

MESOSCOPIC STRUCTURES IN PHYLLITIC LAYERS

Phyllitic layers composed of phyllite, phyllitic quartzite, and quartzite as much as 8m thick are found near the top of the 1200 m-thick Baraboo Quartzite (Dalziel and Dott, 1970; Dott, 1983). Layers thinner than 1 m usually occur as discontinuous lenses within the quartzite. This

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study concentrated on zones of phyllite layers from six sites (fig. 1) on the north and south limbs of the Baraboo Syncline at (1) Abelman's Gorge (Van Hise Rock), (2) Happy Hill, (3) Skillet Creek, (4) northwest park entrance, (5) Highway 12, and (6) northeast Devil's Lake. Two generations of structures were identified within the phyllitic layers.

First-generation Structures

First-generation structures are the most prominent structural features within the phyllite layers. They include in decreasing order of abundance foliation oblique to layer boundaries, foliation parallel to layer boundaries, quartzite boudins, asymmetric folds composed of quartzite, and rotated tension gashes in quartzite. The nature and origin of these structures is interpreted by assuming that the phyllitic layers sustained much simple shear strain and behaved as shear zones between the most competent quartzite layers. The geometric and kinematic characteristics of the structures suggest two different types of models of structural evolution within the phyllitic shear zones.

Model A

This model accounts for structures in the most phyllitic layers where the ratio of mica to quartz is highest (such as the Skillet Creek and the northwest park entrance outcrops on the south limb of the syncline, fig. 1). These phyllitic shear zones have behaved in a very ductile manner.

Foliation oblique to layer boundaries in these zones typically makes angles of 20° to 35° with layer boundaries. Some of these foliation planes exhibit the same sense of displacement as the larger shear zone. Foliation parallel to layer boundaries exhibits the same sense of displacement. These two types of foliation are interpreted according to a modified version of the model of Berthe and others (1979) for the development of foliation in ductile shear zones (fig. 2a). The oblique foliation was produced as the result of flattening perpendicular to the X-Y plane of the finite strain ellipsoid during the initial stages of shear. During later stages of shear foliation parallel to the layer boundaries developed and accommodated shearing parallel to the boundaries which began to truncate the oblique foliation. At first these shear surfaces were widely spaced, but as shearing progressed they became more closely spaced. In many locations the two foliations intersect to form shear surfaces around retort-shaped pods of less-deformed phyllite similar to those described by Simpson and Schmid (1983) for other areas.

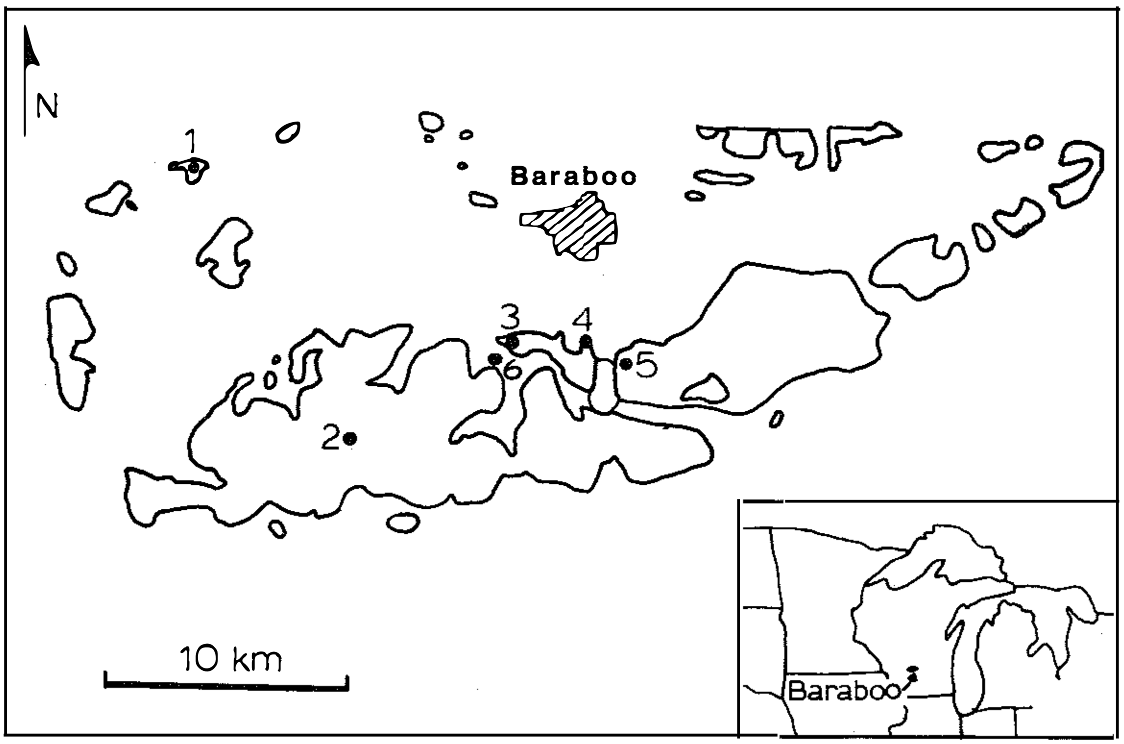

Figure 1.--Map showing distribution of Baraboo Quartzite in south-central Wisconsin. Numbers refer to phyllite zone locations: (1), Abelman's Gorge; (2), Happy Hill; (3), Skillet Creek; (4), northwest park entrance; (5), northeast park entrance; and (6), Highway 12.

The orientations of the two foliations (fig. 3b, 3d, 4) indicate that updip simple shear strain occurred in the phyllite between quartzite layers on both the north and south limbs of the syncline as would be expected during a flexural genesis (fig. 4). The updip shear affected the thin interbedded quartzite beds in a different fashion. It extended parts of these beds into boudins (fig. 3a, 4c) and at the same time compressed other parts to form asymmetric folds. The asymmetric folds (fig. 3a, 4c) previously termed drag folds by Riley (1947) and Hendrix and Schaiowitz (1964) all verge in a direction in support of updip shear. According to the model of Berthe and others (1979), such folds form in the most advanced stages of shear. Rotated tension gashes within extended quartzite layers also indicate strain associated with updip shear. They correspond to the S-Y plane of the updip finite strain ellipsoid.

Our petrographic observations confirm that the phyllite consists predominantly of pryophyllite and quartz with minor muscovite (Dalziel and Dott, 1970). These minerals indicate greenschist-facies conditions. Much of the quartz is recrystallized, which supports a ductile response to strain.

Model B

This model accounts for structures in less mica-rich phyllite layers in which phyllitic quartzite predominates (Van Hise Rock, Happy Hill, and northeast Devil's Lake). Phyllitic shear zones in these layers have behaved in a more brittle fashion (fig. 2b).

The only first-generation structure present within the zones is an oblique foliation consisting of fractures that typically make angles of 70° to 20° with layer boundaries. These fractures exhibit a sense of displacement opposite to that of the overall shear zone (fig. 2b). This is the fracture cleavage of Leith (1923). It could correspond to the X fractures of Logan and others (1979). Near the edges of the phyllite zones in contact with the adjacent quartzite this foliation typically decreases its angle of obliquity, forming a sigmoid shape.

This foliation is interpreted to have formed in a phyllitic shear zone characterized by a falling book style of strain (fig. 2b). It represents more brittle behavior than Model A. The sense of shear indicated by Model B shear zones supports the concept of updip shear generated during flexural genesis of the Baraboo Syncline (fig. 4).

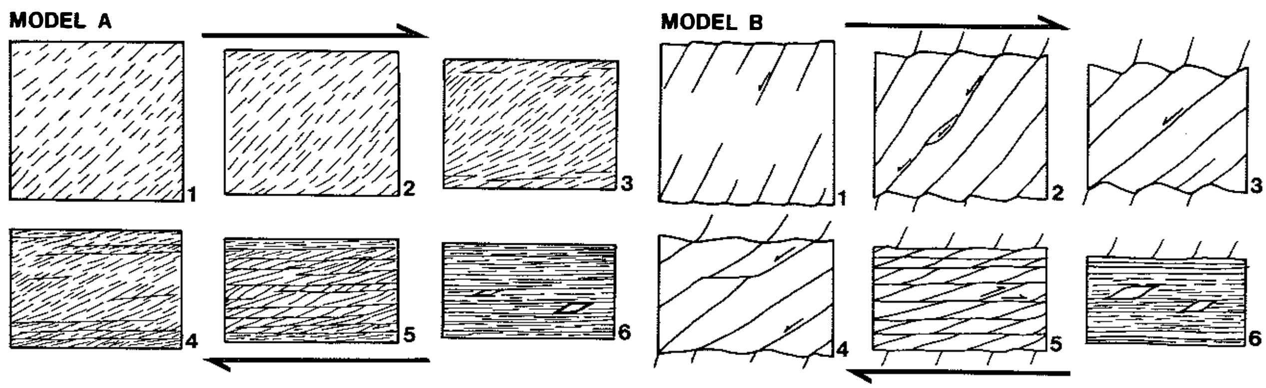

Figure 2.--(a) Six stages in the evolution of foliation in Model A based on a modified version of the model of Berthe and others (1979). In this example the shear zone boundaries are parallel to the top and bottom borders of each frame and the shear sense is right-lateral. (1) An early stage of shear when platy minerals are alligned parallel to the XY plane of the finite strain ellipse to form S surfaces or schistosity. (2) More S surfaces develop. Those near the shear zone boundaries experience some drag. (3) and (4) S surfaces rotate toward parallelism with the shear zone boundaries especially near the shear zone margins. C surfaces develop parallel to the shear zone boundaries as small shear zones with the same sense of displacement. (5) C surfaces become more numerous especially near the shear zone boundaries. S surfaces have rotated into parallelism with C surfaces. Only a few remain and these intersect with C surfaces to form retort-shaped pods. C surfaces are closely spaced and accommodate most of the shear strain. (b) six stages in the evolution of foliation in Model B. (1) An early stage of shear where fractures propogate from the shear zone boundaries. (2) These fractures link to form left-lateral shears separating blocks of shear zone rock. (3) The fractures and blocks rotate clockwise. (4) Shears parallel to shear zone boundaries develop. (5) These shears grow more numerous and accommodate shear parallel to that of the overall shear zone. The fractures oblique to the shear zone boundaries rotate clockwise so far that their sense of displacement changes from left lateral to right lateral. (6) Shears parallel to shear zone boundaries predominate. Some oblique shears remain and intersect with dominant shears to form retort-shaped pods.

<u>Second-generation Structures</u>

Second-generation structures are less common than the first generation and are developed to different degrees in different exposures of phyllite layers (fig. 5). They are especially well displayed at the mica-rich Skillet Creek, Happy Hill, and northwest entrance exposures (fig. 1). These structures include, in decreasing order of abundance, crenulation cleavage, kink bands, chevron folds, folded late-stage quartz veins, and tension gashes in quartzite. They are found only within Model A phyllite zones on the south limb of the syncline where the ratio of mica to quartz is high. The composition of these zones suggests that they were most easily deformed by a later deformation event.

All the second-generation structures deform or overprint the first-generation. The axial surfaces of crenulation cleavage (fig. 5), kink bands, chevron folds, and folded late-stage quartz veins are parallel and oriented at N. 55° E., 30° SE. These structures verge to the north, opposite to the vergence of the asymmetrical first-generation structures. Second-generation rotated tension gashes offset first-generation tension gashes within interbedded quartzite layers (fig. 5). The orientation of the second-generation tension gashes and northward vergence of the other second-generation structures indicates that the phyllite zones experienced a simple shear strain directed in the opposite direction to the first-generation simple shear strain. In other words the second-generation structures were formed as a result of a downdip shear (fig. 6). The crenulation cleavage, kink bands, and chevron folds manifest the same folded geometry but at different scales. In each case, the folds form where first-generation C foliation parallel to layering predominates. The C foliation has been folded sharply at the hinges. The axial surface of these folds corresponds to the X-Y plane of the finite strain ellipse associated with the downdip shear.

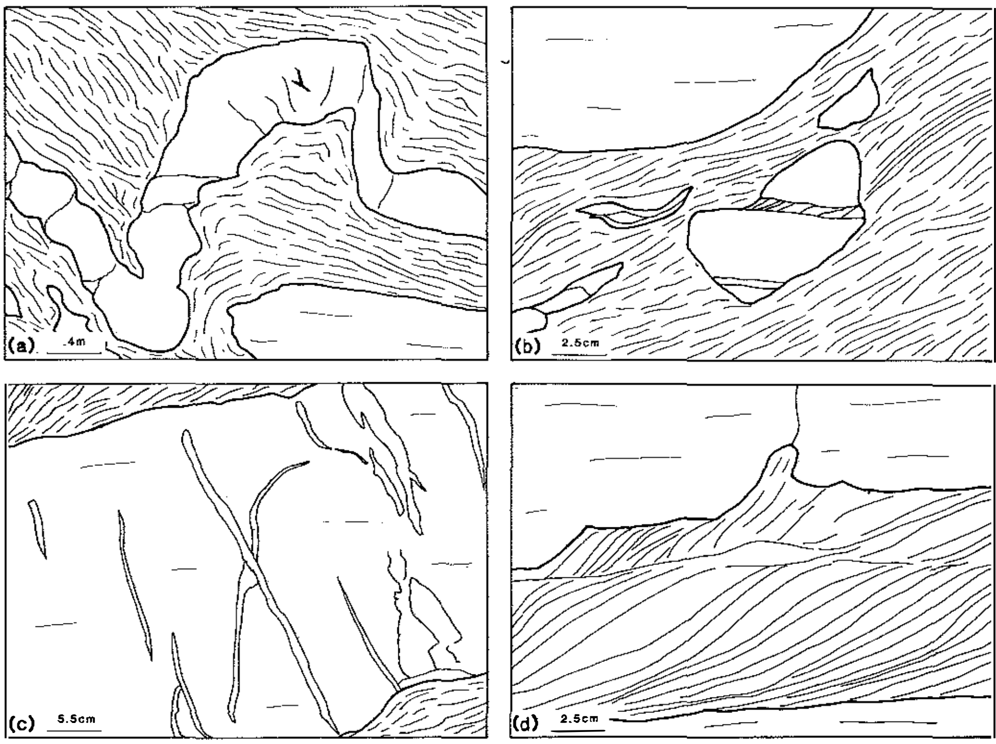

Figure 3.--Structures in Baraboo phyllite zones indicative of the first-generation, updip, simple shear. In all frames phyllite is represented by lined pattern (the orientation and spacing of the lines corresponds to the orientation and spacing of the foliation), and quartzite is represented by the clear pattern. (a) Asymmetrically folded boudinaged quartzite layer as observed at the northwest park entrance. In this exposure the updip simple shear strain is represented by a left-lateral shear couple relative to the borders of the frame. (b) Offset quartzite nodule surrounded by phyllite at Skillet Creek. Note the recrystallized sigmoidal quartz grains indicative of a dextral simple shear. In this Skillet Creek exposure the updip, simple shear strain is represented by a right-lateral shear couple relative to the borders of the frame. (c) Quartz-filled tension gashes in a 0.6 m thick quartzite layer within phyllite at Skillet Creek. Note the rotated tension gash offset right laterally by another first-generation tension gash. (d) S and C surfaces at the northeast park entrance indicative of dextral, simple shear strain. At this exposure updip simple shear is represented by a right-lateral shear couple relative to the borders of the frame.

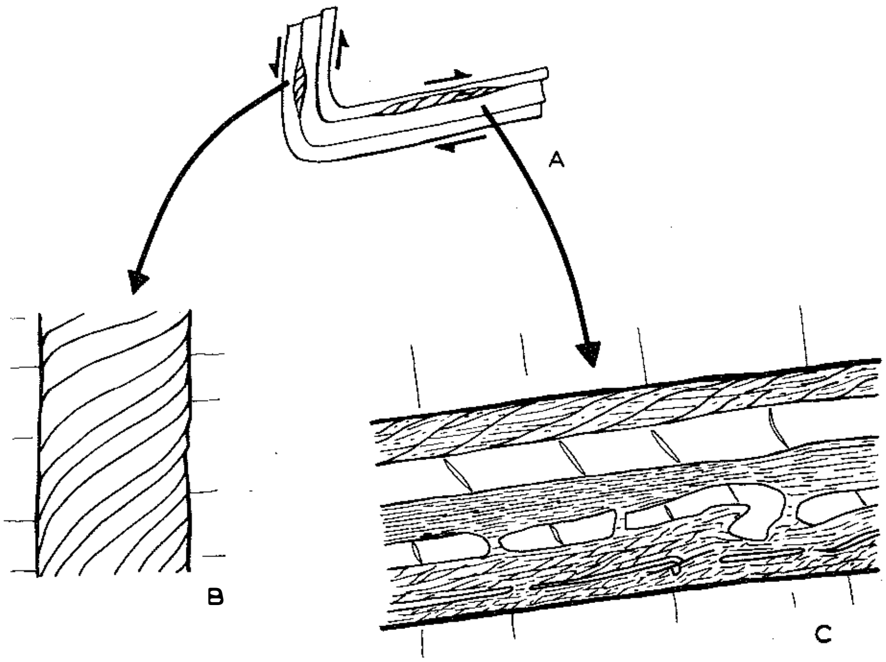

Figure 4.--Schematic diagram to illustrate how first-generation structures were formed by updip, flexural shear within the Baraboo Syncline. (b) Nature and orientation of structures on north limb (for example at Van Hise Rock). (c) Nature and orientation of structures on south limb (for example at Skillet Creek).

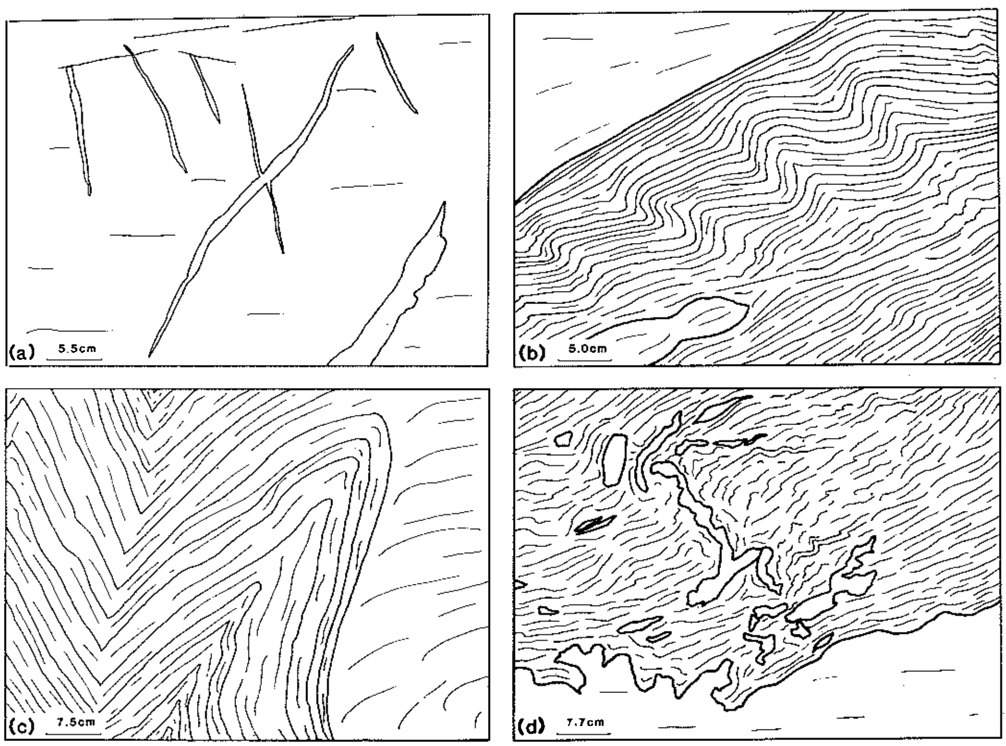

Figure 5.--Structures in Baraboo phyllite zones indicative of the second-generation downdip simple shear. (a) A second-generation tension gash that has offset a first-generation tension gash at Skillet Creek. The down dip second-generation shear is represented by a sinistrel shear couple in this exposure. (b) Kink bands and crenulation cleavage planes at Skillet Creek. In this exposure the downdip shear is represented by a sinistral shear couple. (c) Chevron fold at Happy Hill. In this exposure, downdip shear is represented by right-lateral shear couple. (d) A folded and dismembered late-stage quartz vein at Skillet Creek. Downdip shear is represented by a sinistral shear couple.

The tension gashes correspond to the X-Z plane of the finite strain ellipse (fig. 5b). the less ductile style and orientation of the second-generation structures and lack of recrystallization suggests that the downdip shear strain was of lesser magnitude and possibly represents a lower temperature and pressure regime than the first-generation updip shear strain. Orientation data plotted on stereonets indicate that the principal stress (σ_1) associated with the second generation of shear strain was probably different than that associated with the first generation of shear (fig. 7). The hinge line of the Baraboo Syncline is approximately N. 75° E. whereas the hinge line of the second-generation structures trends more north of east.

DEFORMATION HISTORY

Summary

The deformation history proposed by our study involving two independent and opposite generations of simple shear strain supports the original proposals of Riley (1947) and Hendrix and Schaiowitz (1964) for two phases of deformation. However, our study does not support the proposal of Dalziel and Dott (1970) and Dalziel and Stirewalt (1975) for a single progressive phase of deformation; there are several problems with their model. It is difficult to reconcile the opposing shear sense indicators of the first- and second-generation structures delineated in this study with the single progressive sense of shear they propose. This is especially true with the intersecting and offsetting first- and second-generation, rotated tension gashes which clearly indicate opposing senses of shear. The model of Dalziel and others may be overly complex. We did not find their S. 1° E. slaty cleavage within the phyllite. Their crenulation cleavage bands conjugate to the ones described in this study are minor and are not incompatible with the second-generation shear. They also neglect the microstructural data of Riley (1947) that suggests two independent generations of strain. Even their own microstructural data (Dalziel and Stirewalt, 1975, p. 1681) show a heterogeneous quartz fabric that could be interpreted to suggest more than one period of deformation. Dalziel and Stirewalt (1975) based part of their argument for one progressive phase of deformation on similar features observed in orogenic belts where there has clearly been one progressive phase of deformation, for example the British Caledonides (Dewey, 1969) and the south-

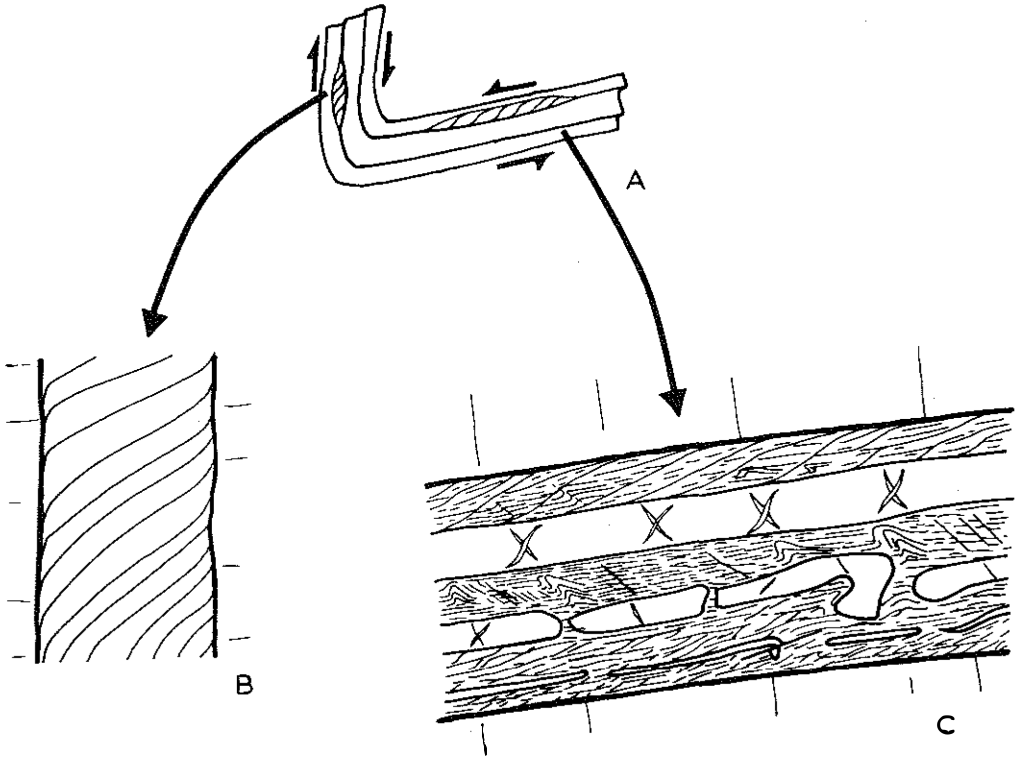

Figure 6.--(a) Schematic diagram to illustrate how second-generation structures were formed by down-dip shear within the Baraboo Syncline. (b) This shear zone consists of phyllitic quartzite. It possesses sufficient strength to resist deformation during the second phase of deformation. (c) This shear zone contains a greater proportion of mica and so is strained during the second episode of strain. Crenulation cleavage, kink bands and asymmetric folds deform the first-generation, C surfaces. Second- generation, tension gashes cut first-generation, tension gashes.

ern extremity of the Andes (Dalziel and others, 1974; Dalziel and Palmer, 1979). However, in these orogenic belts the sequence of mesoscopic structures formed in pelitic sequences 2 to 3 km thick and the structures associated with progressive stages are all coaxial. In the Baraboo Syncline the phyllitic layers are now no thicker than 8 m and the first- and second-generation structures are not coaxial. Hence, we propose that the Baraboo Syncline experienced two distinct episodes of strain each associated with stress patterns of different orientation.

Tectonic Implications

Van Schmus and Bickford (1981) have proposed that the Baraboo Syncline was produced 1,630 Ma in the foreland of the Mazatzal Orogeny to the south. Zircon from rhyolite below the Baraboo Quartzite have been dated at 1760 $\pm$ 10 Ma which marks the age of extrusion (Van Schmus and others, 1975b; Van Schmus and Bickford, 1981). Mineral and whole-rock Rb-Sr systems in the region were reset at 1630 Ma (Van Schmus and others, 1975c; Sims and Peterman, 1980; Van Schmus and Bickford, 1981). Both the rhyolite and overlying quartzite have been folded. These data led Van Schmus and Bickford (1981) to propose that the folding and resetting of the Rb-Sr systems occurred during the 1630 Ma event (after the suggestion of Smith, 1978). Alternatively, Brown and Greenberg (1981) and Greenberg and Brown (1983) acknowledge folding at 1630 Ma but suggest a different tectonic mechanism. They discount evidence for orogenic activity south of Baraboo and propose that deformation be envisioned as a result of mostly vertical, gravity-type tectonics within an epicontinental, anorogenic environment. Eugene I. Smith (oral communication, 1984) suggests that the Baraboo interval quartzite synclines, such as Baraboo, could have formed in localized strike-slip zones. He bases his argument on the fact that there is little evidence suggesting that these synclines were formed in a broad, region-wide fold-belt, or orogenic front. We agree that 1,630 Ma is the most likely time for forming the Baraboo Syncline and the first-generation structures within phyllite layers. However, we decline to support any tectonic mechanism until more data become available.

The nature and age of the deformation event responsible for producing the second-generation structures within the phyllite layers of the Baraboo Syncline are less clear. The downdip shear indicated by the second-generation structures suggests an opening or extension of the syncline during some kind of extensional tectonic event. There are at least two possible periods of extensional activity in the Upper Midwest after 1,630 Ma that may have affected the Baraboo Syncline: (1) a volcanic-plutonic event at 1,380 to 1,500 Ma possibly associated with minor rifting; and (2) Keweenawan rifting at 1,100 to 1,200 Ma.

Emslie (1978) and Van Schmus and Bickford (1981) report numerous 1,380 to 1,500 Ma granite, rapakivi granite, anorthosite, and granodiorite plutons and rhyolitic volcanics in the midcontinent region. Emslie (1978) and Klasner and others (1982) propose that these igneous rocks were produced by a rifting or incipient rifting event. However, there is no structural evidence for this event. Brown and Greenberg (1981) and Greenberg and Brown (1983, 1986) suggest that central Wisconsin experienced additional tectonic strain during intrusion of these 1,500 Ma igneous bodies.

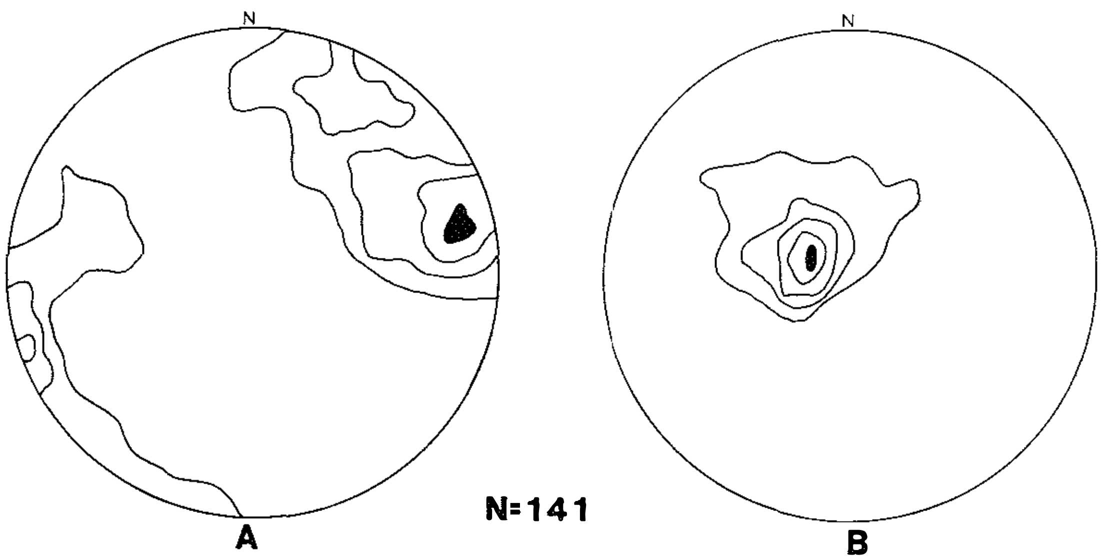

Figure 7.--Stereograms summarizing second-generation, mesostructural data from phyllite zones. Contour intervals are 5%, 3%, 2%, 1%, and ½%. (a) Poles to planar mesostructures, for example crenulation cleavage, kink bands and axial surfaces of asymmetric folds. (b) Hinge lines of crenulation cleavage, kink bands and asymmetric folds.

Greenberg and Brown (1984) show that the Wolf River batholith deformed host rock up to 10 km from its margins. However, the nearest 1,500 Ma igneous rocks to the Baraboo Syncline are at Waterloo, Wisconsin, and northern Illinois, 30 km and 100 km away, respectively. It is unlikely that the intrusion of these bodies could extend to the distant Baraboo Syncline to form the second-generation structures within the phyllite zones. The Baxter Hollow Granite lies adjacent to the Baraboo Syncline. Although it has not been adequately dated, it is texturally and chemically similar to other granites in south central Wisconsin dated at 1,760 Ma (Smith, 1978; W.R. Van Schmus, written communication, 1983). We conclude that the 1,380 to 1,500 Ma plutonic event in south central Wisconsin probably did not impose a second period of deformation on the Baraboo Syncline.

A more likely tectonic event to have deformed the Baraboo Syncline a second time was the late Proterozoic Grenville/Keweenawan event. The Keweenawan rift system (Craddock, 1972) opened to the west of Baraboo at 1,200 to 1,100 Ma according to the best U-Pb and Rb-Sr dates available (Van Schmus and others, 1982; Anderson and Burke, 1983). The Grenville Orogeny existed to the east of Baraboo from 1,250 to 1,100 Ma according to the best U-Pb and Rb-Sr age dates (Baer, 1981; Anderson and Burke, 1983). Hence, Keweenawan rifting and Grenville collisional orogenesis was roughly synchronous. Based on Stockwell's (1964) synthesis of K-Ar dates, many previous investigators have asserted that Grenville orogenesis postdated Keweenawan rifting (Craddock, 1972). In light of the new data, this notion must be revised.

Recognizing the sychroneity of Keweenawan rifting and Grenville orogenesis, Gordon and Hempton (1983) have suggested that rifting was genetically related to orogenesis. They propose that the Keweenawan Rift (including the Michigan Basin) formed as a partially coalesced series of large en-echelon pull-apart basins along intracontinental strike-slip faults generated in the foreland of the Grenville Orogeny. According to Gordon and Hempton (1983), the area including the Baraboo Syncline, between the compressional Grenville Province and the extensional Keweenawan Rift experienced much strike-slip and extensional strain in associated pull-apart basins. This strain pattern is similar to that present in the forelands of modern convergent systems such as the Himalayas (Molnar and Tapponier, 1975), the Bitlis/Zagros (Hempton, 1982), and the Alps (Illies, 1975, 1981). Strike-slip faulting associated with the Keweenawan had already been proposed by Chase and Gilmer (1973) and Weiblen (1982). In the Keweenawan/Grenville system the orientation of σ_1 may have been variable. Near the collision front, σ_1 was probably horizontal and trended northwest-southeast. In the foreland between the rift system and the orogenic front where strike-slip faults predominated, σ_1 was probably horizontal except near pull-apart basins where it would have been vertical. The trend of σ_1 would have depended on the orientation and sense of displacement of the faults. Near the en-echelon pull-apart basins that coalesced into the rift system, σ_1 was probably vertical with σ_3 horizontal and oriented northwest-southeast. This particular stress pattern is reflected in the numerous northeast-southwest dikes associated with rifting. An important point to emphasize is that within the whole convergent/rift system σ_1 was regionally and locally variable. The strain that affected the Baraboo Syncline for the second time involved extension, perhaps local, that opened the syncline and caused down-dip simple shear on the limbs. That shear produced second-generation structures in phyllite whose axes are oriented northeast-southwest. The extension was probably associated with a strike-slip or pull-apart basin along a large northwest-southeast strike-slip fault between the collisional front to the east and the rifting to the west.

Gordon and Hempton (1983, 1984) point to the relationship between the Bitlis/Zagros orogenic belt and the Red Sea/Gulf of Aden rift system (Cochran, 1981, 1983) as a modern actualistic analog for Grenville/Keweenawan tectonic development (fig. 8). In both cases the geometry and kinematic evolution is similar and involves synchronous orogenesis and rifting by the coalescense of pull-apart basins along large strike-slip faults in the foreland. The Bitlis/Zagros orogenic system resulted from the collision of the Arabian Peninsula with Eurasia in the Miocene-Pliocene (Hempton, 1982). The Red Sea/Gulf of Aden system formed a rift in the Miocene-Pliocene that parallels the orogenic front of the Bitlis/Zagros (Cochran, 1981, 1983). Many large strike-slip faults cut the Arabian Peninsula between the rift and the orogenic front (Hempton, 1982), including the well-known Dead Sea Fault. Although the geometry of the Keweenawan and Red Sea/Gulf of Aden rift systems is not exactly identical in terms of size and location of parts of the rift, the important point to emphasize is that the processes responsible for the formation of the two rift systems and their relationship to the synchronous orogenic events is similar.

We argue that because there exists today a synchronous orogenic/rift system with strike-slip related deformation occurring between the rift and orogenic front a similar strain pattern may have existed for the Late Proterozoic Keweenawan/Grenville system. In that type of setting, rocks in Wisconsin (for example, the Baraboo Syncline) could have been redeformed by strike-slip related tectonism.

CONCLUSION

Analysis of structures within phyllitic layers of the Baraboo Syncline suggests a two-stage deformation history involving two independent and opposing generations of shear strain. First-generation structures were formed by updip shear strain related to the flexural genesis of the Baraboo Syncline probably during a 1,630 Ma deformation event of uncertain tectonic origin. Second-generation structures were possibly made by downdip shear strain related to opening during the time of Keweenawan rifting and Grenville continent/continent collision at 1,100 to 1,200 Ma.

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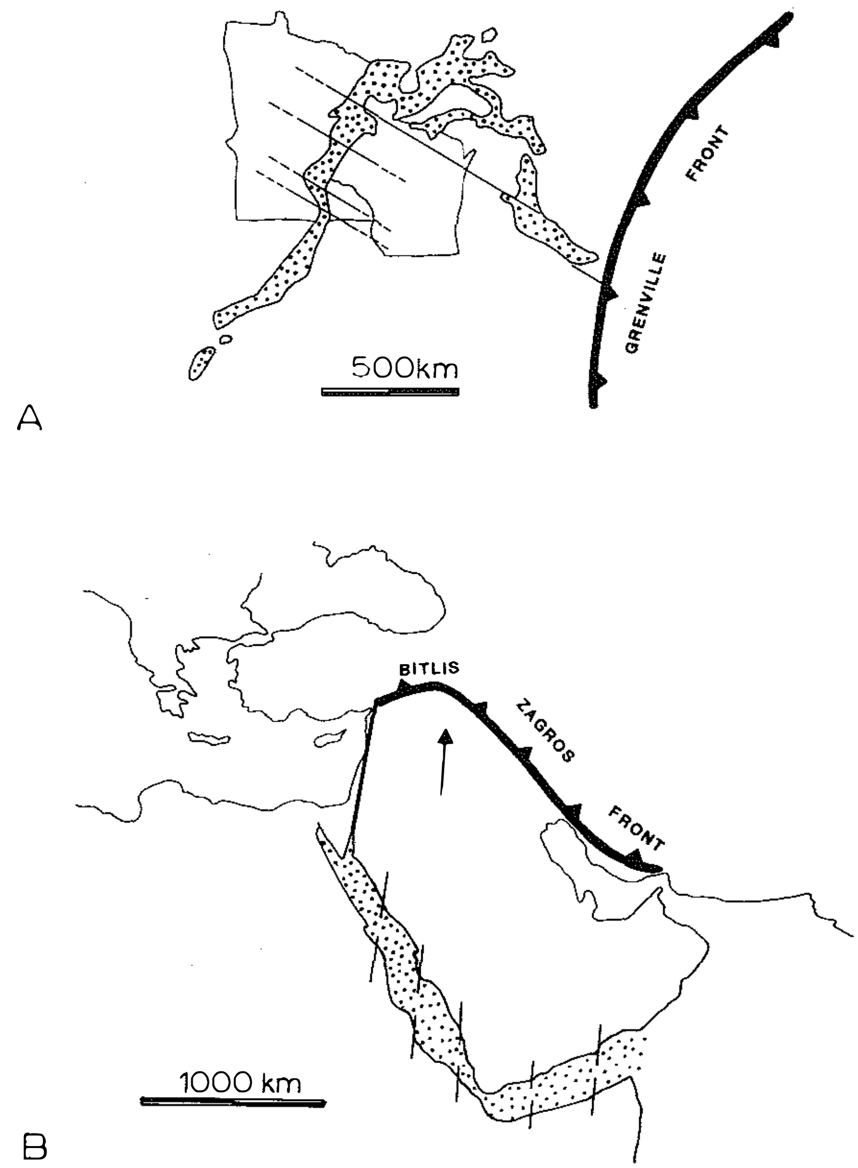

Figure 8.--Schematic, tectonic maps illustrating large rifts developed in the forelands of convergent zones. A. An interpretation of the relationship between the Keweenawan Rift (dot pattern) and synchronous, Grenville continent-continent collision, frontal thrust (thick barbed line). The area between the Keweenawan Rift and Grenville Front underwent strike-slip related shear and extension. B. An actualistic example from the Middle East. The Red Sea and Gulf of Aden Rift System (dot pattern) is presently extensional, whereas the Bitlis-Zagros Orogenic Belt is convergent (after Hempton, 1982; Cochran, 1983).

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JOINTS IN THE RIB MOUNTAIN AREA, WAUSAU, WISCONSIN

by

David R. Dockstader[1]

ABSTRACT

Four distinct joint sets have been identified in the large quartzite xenoliths of the Wausau pluton, southwest of Wausau, Wisconsin. Planes parallel to three of these joint sets have a common line of intersection, while the fourth set is perpendicular to this line of intersection. These four joint sets with the same relative angular relationships have been found in the quartzite of Rib Mountain, Hardwood Hill and Mosinee Hill. However, the attitude of the joints is different in each of these locations. This difference in attitude may be accounted for by simple rotations of the xenoliths. It is proposed that these xenoliths formed during magma emplacement as a result of monoclinal bending and faulting of an overlying quartzite unit. In this model rotation of the xenoliths occurs as they break away from the roof of the magma chamber and settle into the melt.

INTRODUCTION

The Wausau pluton is located in Marathon County, Wisconsin, southwest of the city of Wausau (fig. 1). This pluton has a reported age of 1,520 Ma (Van Schmus, 1980) and consists of two parts separated by the Rib River. The rock of this pluton exhibits many interesting and often subtle variations which are reported by Sood, and others (1980) and by LaBerge and Myers (1983). The present study has concentrated on the southern section of the Wausau pluton which is made up of two intrusions. The first intrusion is syenitic in composition and forms a crescent shaped outcrop. The second, referred to as the Ninemile pluton, is granitic in composition and has intruded the syenite forming an elliptical outcrop measuring roughly 15 x 19 km, elongated northeasterly (LaBerge and Myers, 1983). Rib Mountain, one of the highest points in Wisconsin, as well as Hardwood Hill and Mosinee Hill, are formed by erosionally resistant quartzite xenoliths within the syenite. These hills form a ring rising as much as 200 m above the Ninemile Swamp, which has formed in the center of the pluton over the more easily weathered Ninemile granite (fig. 2).

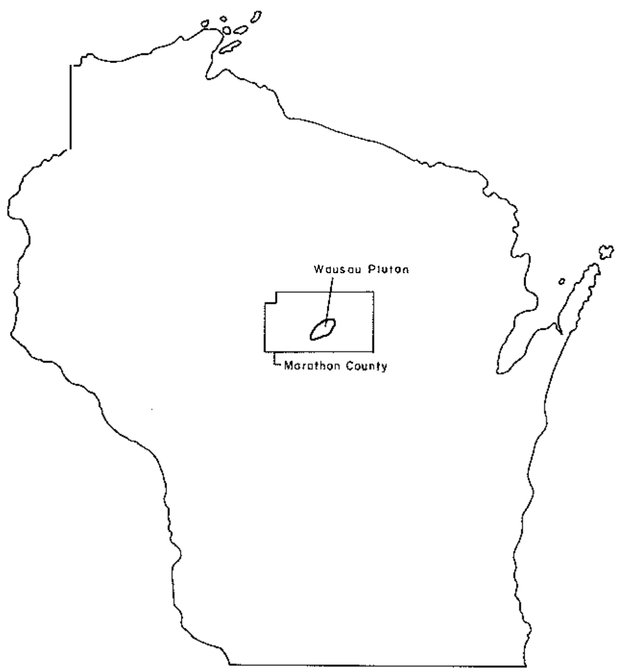

Figure 1.--Map of Wisconsin showing location of Marathon County and the Wausau pluton.

Outcrop surrounding the Wausau pluton is early Proterozoic metavolcanic country rock. Xenoliths of this country rock, as well as quartzite xenoliths, are abundant in the syenitic intrusion, although no outcrop of quartzite is found in place (LaBerge and Myers, 1983). Xenoliths range in size from small (cm) up to the large (km) hill-forming blocks. The large quartzite xenoliths are quite uniform in both texture and composition, suggesting that they formed during a regional event rather than from contact metamorphism in the pluton. All xenoliths are angular in shape and display sharp contacts with the syenite. This evidence clearly indicates brittle behavior of the country rock during the intrusion process and suggests shallow emplacement with rapid cooling. A precise determination of overburden thickness from this mechanical evidence is impossible, but a thickness from 1 to 3 km is indicated.

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QUARTZITE XENOLITHS

The quartzite xenoliths consist of a hard, white quartzite. Hand specimens from Rib Mountain consist of well-rounded, well-sorted, coarse sand-sized quartz grains which account for 90 to 95 percent of the rock volume. Mica and iron oxide appear to make up most of the remainder. LaBerge and Myers (1983) report the presence of K-feldspar, which they attribute to granitization, near contacts with the syenite. Dark bands (concentrations of interstitial mica and iron oxide) suggestive of bedding are found in the quartzite of Rib Mountain, but were not observed in other quartzite xenoliths. Ripple marks are found in the Queen's Chair area of Rib Mountain (fig. 2) but were not found in other outcrops on Rib Mountain, nor in outcrops of other xenoliths. In the Queen's Chair area several quartzite blocks contain clear exposures of ripple marks. The surfaces exposed are quite smooth showing showing ripple marks with rounded crests and troughs and a wavelength of 4 to 5 cm. However, no blocks with these clear exposures were found in place. A park information sign calls attention to ripple marks in one large block which does seem to be in place. The indicated markings are on a vertical face which strikes east to west. This face parallels the dark bands in the block and the markings have the same periodicity, 4 cm, as the smooth ripple marks observed. However, unlike the smooth, rounded markings mentioned above, these markings are sharp, jagged fracture surfaces. Thus, while the exposed surface apparently parallels bedding in the quartzite and the periodic markings seem to correspond to ripples, the actual outcrop surface is the result of brittle fracture and does not correspond to the original ripple surface. This modification of the ripple surface prevents reliable determination of younging in the quartzite from asymmetry in the surface markings.

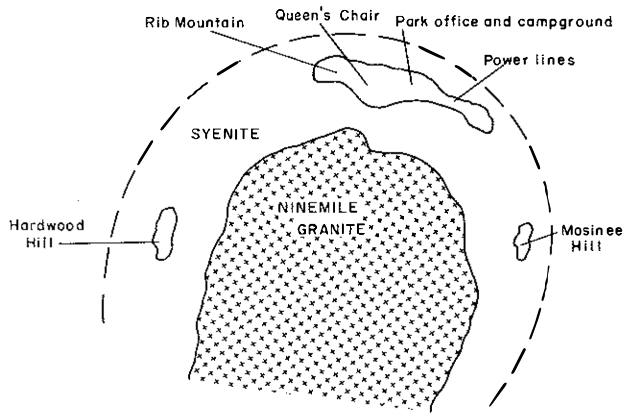
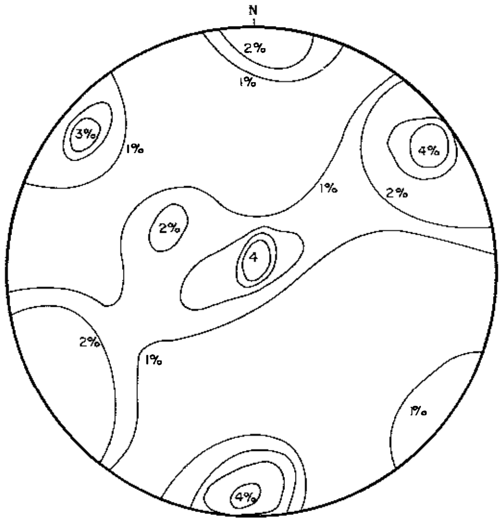


Figure 2.--Generalized geologic map of the Wausau pluton (southern part).

Several joint sets are clearly evident in the quartzite xenoliths. Measurement of joint attitudes in the three largest xenoliths was undertaken with the idea of correlating patterns in the different xenoliths and thus determining relative rotations. Preliminary results of this effort have been reported by the author (Dockstader, 1980).

Four joint sets were readily apparent in the state park area of Rib Mountain. Orientation of these joint sets was initially determined by simply measuring the attitude of a few prominent joints from each set. As a check on the reliability of this procedure and to give the results greater statistical validity, fifteen students in the University of Louisville summer field camp were assigned to measure 20 joint attitudes each, both in the Queens Chair area and just north of the state park office. These results are summarized in the contoured pole diagrams of figure 3. Although the relative frequency distributions for these two locations differ somewhat, the same four concen-trations do show up in each diagram. These concentrations correspond to the same attitudes identified initially for the four readily apparent joint sets. Centers of concentrations are shown in figure 4. To facilitate discussion, these pole concentrations are labeled a, b, c, and d. Pole a is a horizontal joint set. Pole b corresponds to joints in the vertical plane identified above as the bedding plane. Joint surfaces corresponding to poles c and d are also nearly vertical in these outcrops. To return this Rib Mountain quartzite to a position with horizontal bedding requires a rotation of 180 degrees about an east-west horizontal axis. Following this rotation, planes corresponding to poles b, c, and d will intersect along a horizontal north-south line, and pole a now corresponds to a vertical plane with an east-west strike. Since the direction of younging in the quartzite is not clearly discernable in outcrop, the direction of rotation required to return the quartzite to an upright position is also uncertain. For illustrative purposes figure 5 shows the results of a rotation which brings south-facing surfaces into an upward facing position. *In situ* positions are shown with lower case letters, whereas upper case is used for positions after rotation.

Four joint sets are readily apparent in outcrops along the length of the Rib Mountain Ridge. Attitude measurements of prominent joints from the west end of Rib Mountain to the power line crossing (fig. 2) show the same orientation found in the Queen's Chair and park office areas. This evidence suggests that this block of quartzite, roughly 3 km in length, rotated as a single unit. Along the 1.2 km section of Rib Mountain from the power line to the east end of the ridge, four joint sets are again apparent; however, they are found in a new orientation. Poles to these



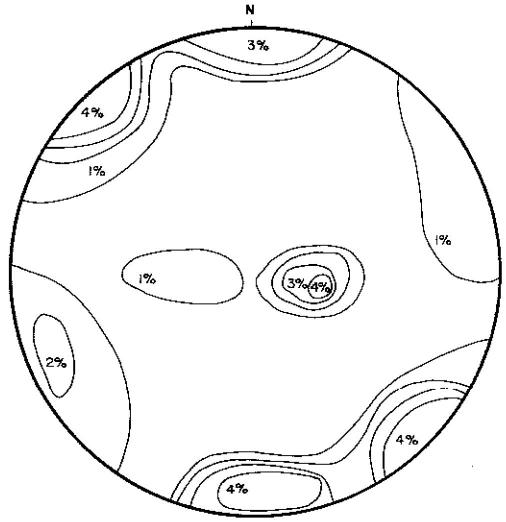


Figure 3.--A. Poles to joints planes near the Queens Chair, Rib Mountain. Countours show 1, 2, 3, and 4% concentration per 1% area. B. Poles to joint planes north park office, Rib Mountain.

joint planes are indicated by lower case letters in figure 6. Although the orientation of these poles differ from those of figure 4, the joints show the same pattern relative to each other as those represented in figure 4. The new joint orientation seems to be simply the result of a rotation of the whole outcrop. Using relative positions to identify corresponding poles, figure 6 has been labeled so that poles are identified with the same letter as corresponding poles in figures 4 and 5. Note that the near symmetry of joint sets c and d about the bedding (b) makes it impossible to distinguish these sets without an indicator of younging. The procedure followed in marking corresponding joints was to assume that although different blocks of quartzite may have rotated independently, they probably rotated in the same sense relative to the center of the pluton. This assumption is discussed further below. The identification of bedding (pole b) agrees with orientation of dark bands occasionally observed on this part of the ridge. However, on this part of the ridge the dark bands were much more difficult to find than on the western section, and no ripple marks were apparent. Bedding on the east section has an attitude of N. 60° W., 65° SW so that a rotation of 65 degrees about a horizontal axis trending N. 60° W. will return the beds to a horizontal position. The results of this rotation are illustrated with upper case letters in figure 6. A 105-degree rotation would also return the beds to a horizontal position. Because sym-

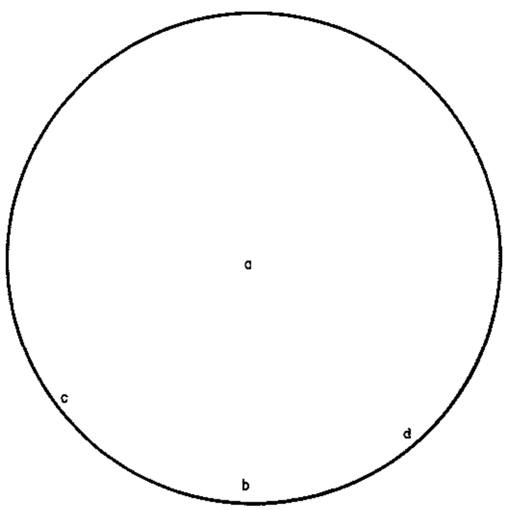


Figure 4.--Poles to prominent joints in Rib Mountain determined from centers of concentration in figure 3A, and figure 3B.

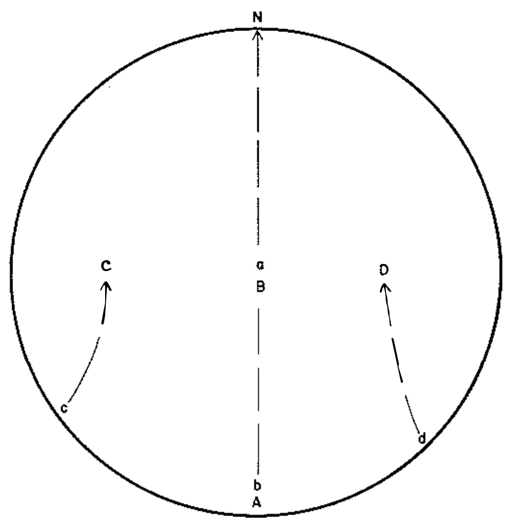


Figure 5.--Poles to prominent joints in Rib Mountain. Lower case represents joint <u>in situ</u> and upper case represents joints after a rotation which brings bedding represented by pole b into a horizontal position represented by pole B.

metry in the joint set geometry either rotation would yield very similar joint attitudes. Note also that once the beds on the eastern part of the ridge are returned to a horizontal position an additional rotation in the counterclockwise direction about a vertical axis is required to return the joints to the same attitudes illustrated by the upperase letters in figure 5.

The difference between joint attitudes in the two parts of Rib Mountain indicates the presence of two separate xenoliths, which rotated independently in the magma. In both cases the major component of rotation is about an axis tangential to the pluton margin. This observation suggests that the xenoliths were not tumbled about in the magma, but were simply turned up or down in a single rotation and provides the basis for the assumption mentioned above in identifying relative orientations of joint sets c and d in different outcrops.

Outcrops on Hardwood Hill and Mosinee Hill show neither color bands nor ripple marks to indicate bedding. However, a comparison of joint orientation with Rib Mountain makes a determination of bedding possible. In determining joint attitudes on Hardwood Hill and Mosinee Hill the procedure was simply to take a few measurements of the visually prominent joint sets. On Hardwood Hill four joint sets were found in the same relative position as on Rib Mountain. Corresponding joints are labeled with the same letters as in earlier diagrams (fig. 7). As before, lower-case letters indicate _in situ_ orientations, whereas upper case indicates positions after rotation of bedding into a horizontal position. Note that a fifth joint set was also observed on Hardwood Hill and is labeled x. A few joints in this relative position show up in the measurements taken on Rib Mountain (fig. 3). Hardwood Hill is thickly vegetated, limiting good outcrops. Not enough information was gathered to prepare a frequency diagram, so no significance is attached to the fifth joint set. What is judged significant is that the bedding is nearly vertical with a strike parallel to the local margin of the Wausau pluton, suggesting the simple rotation of the xenoliths mentioned above. Results from Mosinee Hill (fig. 8) also show four joint sets in the same position relative to each other as in the xenoliths discussed above. Bedding is vertical with a strike parallel to the local margin of the pluton. In addition to the rotation observed for the other xenoliths, an additional rotation of 90 degrees about an axis perpendicular to the bedding is also apparent. This rotation is analogous to the 17 degree rotation observed for the eastern section of Rib Mountain.

In summary the large quartzite xenoliths ring the southern section of the Wausau pluton. Relative joint orientation in conjunction with primary structures indicate that bedding in the xenoliths is roughly vertical and faces either into or away from the center of the pluton.

IGNEOUS INTRUSION

Two, vertical, orthogonal joint sets, striking north-south and east-west, pervade the igneous rock in the southern part of the Wausau pluton. It is my observation that such orthogonal joint sets occur in tabular intrusions (thickness substantially less than diameter) whereas more complex jointing occurs when the thickness is roughly the same or greater than the intrusion diameter. Figure 9, an aerial photograph taken over the Shonkin Sag laccolith, illustrates orthogonal

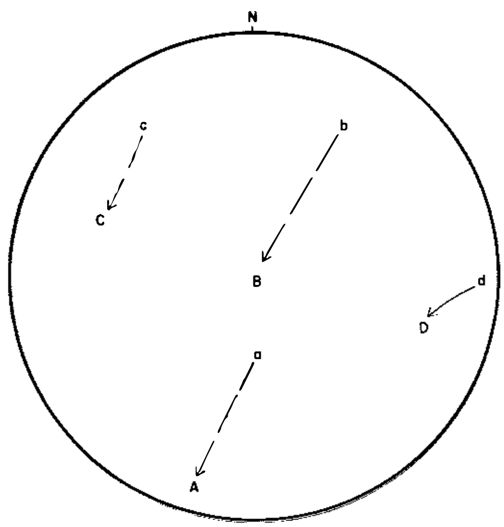


Figure 6.--Poles to prominent joints on the east end of Rib Mountain.

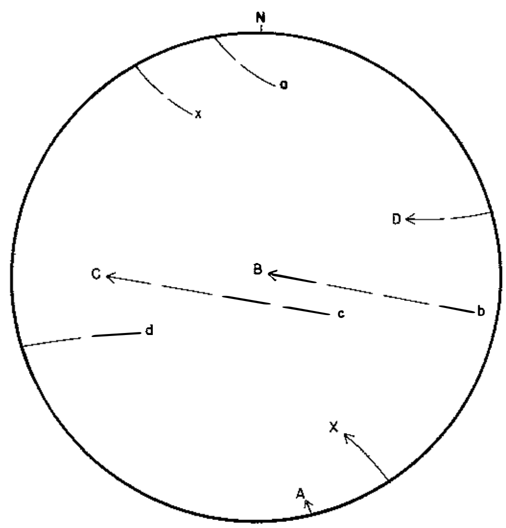


Figure 7.--Poles to prominent joints on Hardwood Hill.

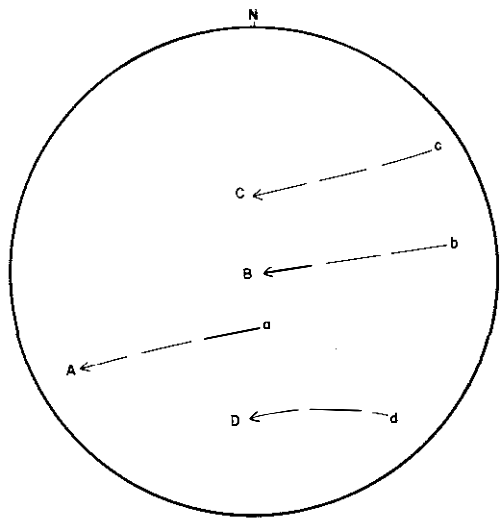


Figure 8.--Poles to prominent joints on Mosinee Hill.

jointing over a tabular intrusion. Devils Tower, Wyoming, is a well known intrusion with more complex jointing. The joint pattern, as well as the roughly circular plan view (LaBerge and Myers, 1983) are evidence to the author of a tabular or laccolithic form for the Wausau pluton. The Grizzly Peak cauldron, Sawatch Range, Colorado (Fridrich and Mahood, 1984), is an example of multiple intrusions in a laccolithic form, which seems in many ways to be analogous to the Wausau pluton. This laccolithic model predicts a thickness of roughly 1 to 2 km for the Wausau pluton.

FORMATION OF XENOLITHS

Host-rock deformation during emplacement of a laccolith creates a mechanism for the formation of both large and small xenoliths. This is the mechanism proposed for the formation of the xenoliths found in the Wausau pluton. A laccolith begins as a sill which spreads until enough upward force is produced from the magma pressure to lift the overburden. Once this critical area is reached the intrusion will thicken. The roof of the intrusion will be nearly flat, and monoclinal bending of the overlying strata will accommodate the additional thickness of the magma (Koch and others, 1981). As the intrusion continues to thicken, bending of the overburden progresses into fracturing and faulting. At this point the overburden is pushed up by magma pressure, in a process which operates much like a hydraulic piston. Piston faults develop around the intrusion periphery allowing the piston of overburden to be pushed up relative to the surrounding country rock. During this entire process, the floor of the laccolith remains virtually undeformed (Johnson, 1970; Pollard and Johnson, 1973; Dockstader, 1982).

Large xenoliths may be easily formed from the roof of a laccolith during the transition from monoclinal bending to faulting. This process is illustrated in figure 10. As uplift progresses, fractures will develop in those areas where the overburden is most sharply bent. In the case of a competent overburden like quartzite, a large intact block of roof rock may be detached from the adjacent host by these fractures. As the block falls into the magma, the angle of the fractures allows the end of the block closest to the intrusion center to fall freely, while the opposite end will catch. The end which is caught will then act as a pivot as the xenolith swings down into the magma. After the block has swung part way into the magma, the hinge end will come free and the xenolith will rotate about its center of mass, causing this end to kick in toward the center of the laccolith. The final result of this process is to place the xenolith some distance in from the intrusion margin with formerly horizontal surfaces now vertical, and the former top of the xenolith facing the intrusion center. Small xenoliths of country rock will be produced during the fracture process and will surround the large xenoliths. If the hinge end of the large xenolith breaks free unevenly, an additional component of rotation may be introduced. This is the mechanism proposed for formation of the xenoliths in the Wausau pluton.

Just as folding roof rock down into the magma provides a plausible mechanism for the formation and orientation of a xenolith, folding floor rock up might possibly provide similar results. Two observations argue against this mechanism. In areas where floor rock has been observed--near the Highwood Mountains, Montana, and Cascade, Montana (personal observations by the author) and in the Henry Mountains, Utah (Johnson and Pollard, 1973)--no deformation of the floor rock has been observed; whereas folding and fracturing of roof rock, as described above, has been observed. Second, where inclusions of floor rock have been identified in an intrusion, the inclusions were concentrated toward the intrusion core (Fridrich and Mahood, 1984).

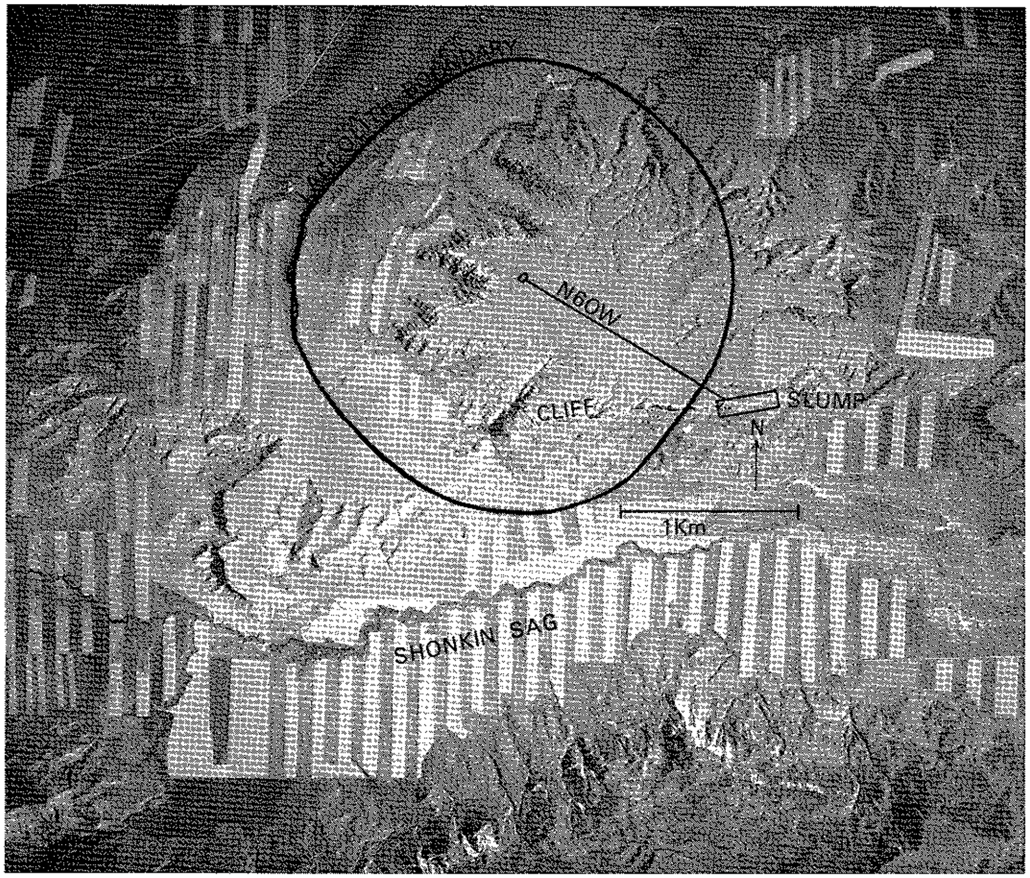


Figure 9.--Aerial photograph of Shonkin Sag laccolith, Montana. The orthogonal joints in the igneous rock are visible where gulleying has exposed the rock.

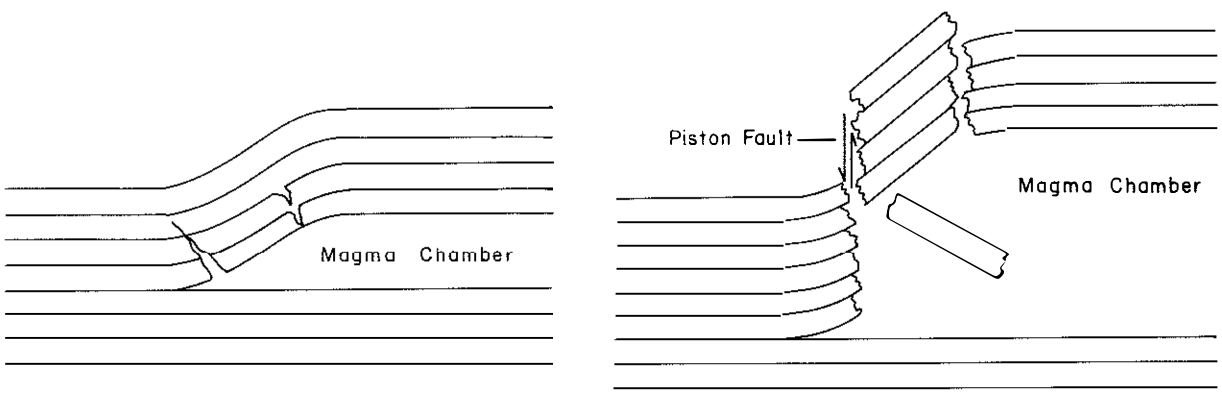


Figure 10.--The formation of xenoliths during bending and fracturing of the roof rock. A. Early stage of fracture formation. B. Advanced stage of fracture with xenolith settling into the magma chamber.

CONCLUSIONS

Emplacement of the Wausau pluton took place through metavolcanic rock and overlying metasedimentary rock. A quartzite layer at least 0.8 km thick (the north-south dimension of Rib Mountain) was near the bottom of the metasedientary sequence. Magma rising through the metavolcanic sequence spread beneath the quartzite and pushed up the overburden bending and then faulting the roof margins. During this process, blocks of roof rock from the fractured roof margins fell into the magma. The largest blocks (km in size) fell into a position where surfaces which had faced upward now faced into the center of the intrusion. This initial intrusion, syenitic in composition, was followed by a second granitic intrusion.

ACKNOWLEDGMENTS

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PETROLOGY, GEOCHEMISTRY AND RB-SR SYSTEMATICS OF THE PORPHYRITIC GRANITE AT HAMILTON MOUND, WISCONSIN

by

Sheryl M. Taylor[1] and Carla W. Montgomery[2]

ABSTRACT

The Hamilton Mound porphyritic granite is petrologically and chemically typical of what would be expected of a eutectic, two-feldspar granite. Field relationships suggest that the granite may have been emplaced along fractures in the quartzite that overlies it. Rb-Sr isotopic data for the granite yield a whole-rock errorchron age of 1,940 $\pm$ 400 Ma with initial $^{87}Sr/^{86}Sr$ = 0.702 $\pm$ 0.008 (2σ). A Rb-Sr mineral isochron records a post-emplacement overprint producing strontium isotopic homogenization on the mineral scale at approximately 1,580 Ma. This event is probably the major cause of scatter in the whole-rock errorchron. A U-Pb zircon age obtained by Van Schmus (oral communication, 1983) indicates a primary crystallization age of 1,764 $\pm$ 4 Ma, so the Hamilton Mound granite is in fact post-Penokean. However, chemical comparison with other Penokean and post-Penokean felsic rock in the region indicates that the Hamilton Mound granite is chemically more similar to the Penokean rock. This raises some doubt as to the validity of previous attempts to distinguish between products of the two periods of igneous activity on geochemical grounds.

INTRODUCTION

The focus of the present study was the red, porphyritic granite at Hamilton Mound, Wisconsin. The isotopic and chemical properties were investigated together with the possible relationship of the granite to other associated lithologies within the study area. Features or rock types which appeared in the field to be unrelated to the granite were not considered. Other lithologies which were studied include the quartzite that the granite intruded, a quartz breccia, and discontinuous veins found primarily within the quartzite. The complex nature of the local rock association was a major reason for undertaking the study. An attempt was made to relate the Hamilton Mound granite to other lower to middle Proterozoic granite within the region.

REGIONAL SETTING

The early to middle Proterozoic of Wisconsin was dominated by three periods of magmatic activity (Van Schmus, 1976). The first was associated with the Penokean orogeny, 1,850 to 1,900 Ma. The volcanic and plutonic rock associated with this orogeny form a complex which underlies most of northern Wisconsin. Scattered exposures of rock associated with the Penokean orogeny are also found in central and western Wisconsin.

The second period of magmatic activity post-dates the Penokean orogeny by approximately 100 m.y. The post-Penokean rock types are primarily epizonal granite and rhyolite. The rock is exposed mainly in central and southeastern Wisconsin (Smith, 1978; Van Schmus and Bickford, 1981). Some isolated exposures are found in northern Wisconsin. The post-Penokean rock has been dated by Van Schmus (1976) using U-Pb zircon methods at approximately 1,760 Ma. The rock yields younger ages by Rb-Sr rock and mineral dating methods. These younger ages are attributable to a pervasive low-grade, regional geochemical alteration and deformational event approximately 1,630 Ma (Van Schmus, 1976).

The third episode of magmatic activity is associated with the emplacement of the Wolf River batholith, 1,400 to 1,500 Ma. The ages of the various plutons were determined by Rb-Sr whole-rock and U-Pb zircon methods (Anderson and Cullers, 1978), which yielded generally concordant ages.

GEOLOGY OF HAMILTON MOUND

The Hamilton Mound area is located in Adams County in central Wisconsin (fig. 1). Precambrian rock is exposed extensively only in a quarry there. The major lithologies at the Hamilton Mound quarry are a red, porphyritic granite, a quartzite, and a quartz breccia. Minor features include a series of discontinuous pegmatitic and aplitic veins and schist inclusions within the granite. The only previous study of the area, by Ostrander (1931), identified just two lithologies present--the quartzite and the quartz breccia. Only recently has the quarry become deep enough, approximately 5 metres, to expose the granite.

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The quartzite is an extensive unit which varies in color from pinkish-gray to green-gray. The upper part of the unit contains obvious relict cross-bedding. Cross-bedding is much less obvious in the lower part of the quartzite. This lack of apparent cross-bedding in the lower part of the unit may be due in part to the influence of the granite intrusion. The quartzite is arkosic, with minor biotite or hornblende or both as the mafic component. Sericitization of the feldspar is common; this is the only mineralogic alteration observed in the quartzite. Some quartz grains have also been recrystallized. The discontinuous veins are found primarily within this unit. These do not appear to have caused significant mineralization in adjacent host rocks. The quartzite is also heavily fractured with conjugate joint sets prominent throughout it.

The breccia is exposed in a triangular-shaped mass in the topographic center of the Mound. The clasts within the breccia are mostly quartzite; however, there are a few granite clasts. The granite clasts are somewhat different in character from the granite in the main intrusion. Generally, these fragments are finer grained than the intrusion proper, and more equigranular than porphyritic.

The main body of porphyritic granite consists of a dark-colored matrix with phenocrysts of red feldspar. The matrix is primarily quartz, plagioclase, and microcline. In a few places the granite changes character and becomes more equigranular. In such places its color also becomes lighter with pink feldspars giving the rock an overall pink appearance rather than the dark, spotted appearance of the porphyritic parts.

The majority of the phenocrysts are microcline. Plagioclase is nearly as abundant in the rock overall, but does not usually form phenocrysts. Composition of plagioclase was determined optically from extinction angles, and ranges within the study area from albite to andesine, but individual grains are generally unzoned. Epidote, chlorite, and sericite occur as secondary minerals. Perthitic and myrmekitic textures are common throughout the granite. Myrmekite is found between plagioclase and quartz and occurs throughout entire grains rather than just along the grain boundaries. This type of texture is generally considered to be a product of eutectic crystallization (Cox and others, 1979). Zircon is conspicuous in most thin sections, although not present in all. There seem to be two populations of zircons. One appears to be of primary igneous origin with euhedral crystals. The other population has rounded cores with euhedral overgrowths and may indicate a minor inherited detrital component.

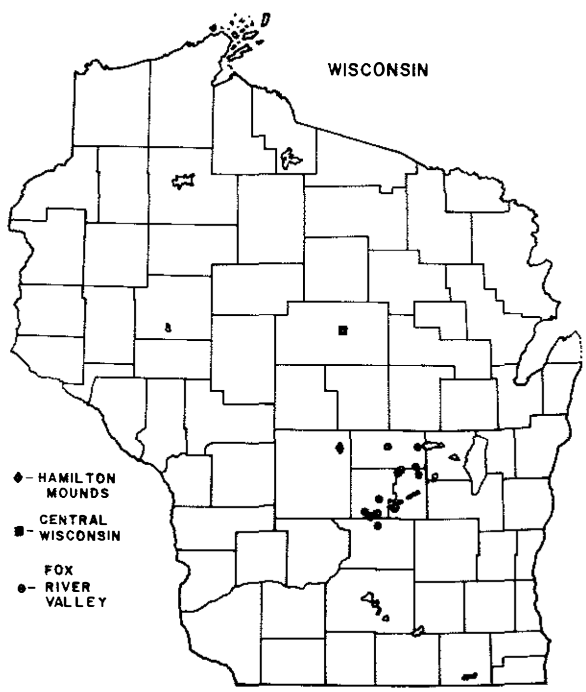

Figure 1.--Distribution of post-Penokean granite and rhyolite. Note location of Hamilton Mound (diamond).

The samples are altered to varying degrees. Sericite is the principal alteration product of the primary minerals in the granite, with microcline less strongly affected than other feldspar. There is a rough correlation between the depth in the vertical section at which the sample was collected and the degree of sericitization, the sericitization decreasing with depth. Small veins cutting the granite consist principally of calcite and quartz, with minor albite and epidote. Epidote is also found along contacts between the veins and host granite. It does not appear to be the product of alteration of minerals within the granite.

At several places within the granite, schist inclusions are found. These inclusions, which consist of quartz, biotite, hornblende, and trace amounts of albite, are finer grained than the granite host. The minerals within the inclusions do not show any secondary alteration. The mineralogy would suggest that temperatures within the schist never exceeded 500 °C. The outer part of the inclusions, although finer grained than the interior, has the same mineralogy. This would suggest that little reaction has occurred between the granite host and the inclusion. The origin of the inclusions is problematical, as the inclusions do not resemble any of the exposed lithologies in the area. It seems likely that these inclusions have been brought up from depth by the granite during emplacement.

The field relationship between the porphyritic granite and the quartzite suggests that the granite has been emplaced along fractures in the quartzite. Several such fractured areas contain quartz veins and granitic material. The fracture filling is not associated with any significant metamorphic effect. Epidote is found along some vein contacts, but otherwise there is little evidence of interaction between granitic material and quartzite. These fractured areas occur throughout the study area.

ANALYTICAL METHODS AND DATA REDUCTION

Details of all analytical methods may be found in Taylor (1983). For both chemical and isotopic analyses samples were chosen to be free of vein material and as little altered as possible. Granite samples were up to 10 kg in mass; smaller samples were taken of finer grained lithologies.

Most major and some minor elements were analyzed by atomic absorption (AA) spectrophotometry; silica and alumina were determined by inductively-coupled plasma (ICP) spectrometry. USGS rock standards G-2, GSP-1, AGV-1, and W-1 were the reference standards. Volatile contents reported are based on total weight loss on ignition. Accuracy of AA and ICP analyses can be estimated by treating some standards as unknowns and noting deviations from accepted values. Average deviations so measured were less than 0.5 percent of standard value for CaO and Na_2O, 1-1.6 percent for SiO_2, Al_2O_3, MgO, and MnO, and 2.2 percent for K_2O.

Rubidium and strontium for isotopic analyses were extracted from approximately 0.2 g splits of rock powder by standard ion-exchange methods. All samples were analyzed on a semi-automated, 6-inch, solid-source, NBS-type mass spectrometer. Rb and Sr contents were determined by isotope dilution. Mineral separates were taken from the -60+150 mesh fraction of rock powder using acetylene tetrabromide separation followed by hand-picking. The feldspar separate is estimated to be about 98 percent pure. Because of the low total mafic content, a mixed biotite-hornblende separate was used for the other mineral fraction. This will not affect the determination of a mineral isochron, as a mixture of two minerals will still fall along the same isochron. Mineral separates were then analyzed by the same procedures as whole-rock samples. Precision of $^{87}Sr/^{86}Sr$ ratios averaged $\pm$0.1 percent (2σ) and is estimated at $\pm$1 percent for Rb/Sr ratios.

Rb-Sr isochrons were fitted using a York (1969) Model II regression technique. This allows for correlation of errors, weighting of points on the basis of analytical precision, and consideration of geologic scatter in determining slope and intercept errors. Uncertainties are reported at the 95 percent confidence level ($\pm 2\sigma$). The ^{87}Rb decay constant used was $1.42 \times 10^{-11} y^{-1}$.

ISOTOPIC DATA

Rb-Sr analyses for 11 samples of Hamilton Mound granite and several associated lithologies are reported in table 1. A Rb-Sr whole-rock isochron plot for the granite samples is shown in figure 2. The data show sufficient scatter that the regression line is really an errorchron with a poorly constrained age of 1,940 $\pm$ 400 Ma, and initial $^{87}Sr/^{86}Sr$ of 0.702 $\pm$ 0.008. The scat- ter is possibly due to a low-grade, post-emplacement metamorphic event, as recognized in thin sec- tion (see petrographic description above). A mineral isochron, defined by a mafic fraction, a K-feldspar fraction, and the corresponding whole-rock point, indicates the approximate time of this event (fig. 2) of 1,580 $\pm$ 160 Ma with an initial ratio of 0.708 $\pm$ 0.004.

It is also apparent from figure 2 that the whole-rock data point furthest away from the intercept (sample 56) is strongly controlling the slope of the whole-rock errorchron. The remaining points scatter along the trend of the mineral isochron. If the regression program is performed for the whole-rock data, excluding sample 56, the result is a date of 1514 $\pm$ 314 Ma, the same error as for the mineral isochron. The whole-rock Rb-Sr system thus appears to have been incompletely reset by the metamorphism with the rock data rotated toward the mineral isochron. There is no evidence that the controlling whole-rock point represents questionable data. A replicate analysis produced the same data point within error. In thin section the sample appears to be among the least altered, and thus may simply have been least reset by the thermal event.

Table 1.--Isotopic data for Hamilton Mound granite and related rock

Sample	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	Rb, ppm	Sr, ppm
Granites				
WIHM-1	1.679	0.7499	70.93	122.7
WIHM-4	1.710	0.7499	78.42	124.7
WIHM-7	1.084	0.7311	72.97	195.3
THM-40	1.335	0.7381	74.91	162.8
THM-42	1.298	0.7401	70.80	158.4
THM-43	1.274	0.7361	103.9	236.7
THM-50	0.7665	0.7275	73.55	278.2
THM-52	1.839	0.7519	93.51	147.7
THM-54	0.9610	0.7307	64.43	194.4
THM-55	1.391	0.7374	87.53	182.6
THM-56	2.308	0.7737	150.9	190.4
schist	2.481	0.7585	159.1	186.4
quartzite	2.422	0.7662	71.26	85.60
arkose	2.486	0.7592	116.7	136.5
mineral separates				
K-feldspar	1.940	0.7524	168.0	251.6
mafic	2.627	0.7668	208.8	231.3

A linear trend may in principle result from mixing of material from two different sources having different chemical and isotopic compositions. In the study area mixing between an incoming magma and an older lithology could have occurred by assimilation. The result would be a pseudo-isochron which yields a meaningless date. The possibility of mixing may be tested by plotting the ratio of $^{87}Sr/^{86}Sr$ against total strontium content. If mixing has occurred, the data should fall along a hyperbolic curve between the two end member components which have been involved in the mixing (Faure, 1977). Such a diagram for the granite samples fails to define a good binary mixing hyperbola; therefore the granite errorchron, although imprecise, probably has age signifi- cance.

MODAL AND CHEMICAL DATA

The results of AA and ICP analyses are given in table 2; modal analyses are reported in table 3 and are based on 750 counts per thin section. When normalized and plotted on a quartz-alkali feldspar-plagioclase (QAP) ternary diagram, the majority of porphyritic granite samples indeed fall in the granite field of Streckeisen's (1976) classification (fig. 3).

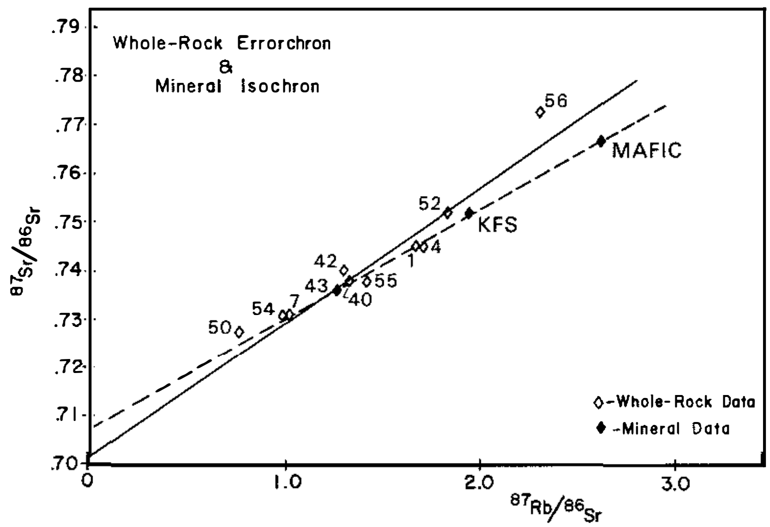

Figure 2.--Composite of whole-rock and mineral Rb-Sr data, Hamilton Mound granite.

The modal data were also converted to CIPW norms and plotted in the system quartz-albite-orthoclase to determine the relationship to a plausible ternary eutectic (fig. 4). The eutectic boundary line for P_{H_2O} = 1 kb is plotted for comparison. The granite analyses do scatter about the thermal minimum. Increasing P_{H_2O} moves the thermal minimum toward the albite vertex, and increase in Ca content would cause a shift in the thermal minimum toward the quartz-orthoclase join. There is a broad linear trend toward the albite vertex which may indicate varying P_{H_2O} during emplacement. The scatter may in part also be a function of the later thermal event having caused some chemical alteration.

DISCUSSION

The attempt was made to relate the chemical and isotopic information obtained for Hamilton Mound to the three earlier periods of Proterozoic magmatic activity in Wisconsin. The whole-rock Rb-Sr errorchron obtained for Hamilton Mound has an uncertainty which is sufficiently large to encompass two of these periods of igneous activity; only the igneous activity at 1,500 Ma may be ruled out because it falls outside of the indicated age range of the granite. U-Pb dating of zircons has, however, pinpointed the time of crystallization of the granite at 1,764 ± 4 Ma (Van Schmus, 1983, oral communication). The porphyritic granite thus formed during the 1,760 Ma magmatic event.

Taking into account the error, the date indicated by the mineral isochron could also correspond to either of two of the recognized events in the region. The resetting could be correlated with either the regional thermal event near 1,630 and 1,700 Ma, or with the emplacement of the plutons associated with the Wolf River batholith between 1,400 and 1,500 Ma. Because the thermal effects of the latter intrusive rock are more localized, it is more likely that the overprint corresponds to the earlier, regional event.

According to Van Schmus and Bickford (1981), Anderson and others (1980), and Smith (1978), the granite and rhyolite dated at approximately 1,760 Ma are considered to be distinct from the earlier Penokean suite. In addition to the age difference the separation had been supported by the hypothesis that the Penokean rock has a different chemical signature from rock regarded as post-Penokean. As the granite at Hamilton Mound appears on the basis of age determinations to belong to the 1,760-Ma group, one would expect its chemistry to be more similar to that of the post-Penokean felsic rock, rather than to Penokean units.

To test this expected relationship, the Hamilton Mound granite and related rock analyses were plotted on various chemical diagrams with the Penokean and post-Penokean fields as defined by Anderson and others (1980). The first of the diagrams is the quartz-alkali feldspar-plagioclase ternary plot (fig. 3). Anderson and others (1980) used this diagram to demonstrate that the post-Penokean granite is compositionally distinct from the earlier, synorogenic Penokean rocks, and also that the post-Penokean rock represents true granite (syenogranite of Streckeisen, 1976). According to Anderson and others (1980), this true granite represents a major change from the

Table 2.--Chemical analyses of Hamilton Mound granite and related rock

	granite													calcite
	WIHM-1	WIHM-4	WIHM-7	WIHM-8	THM-40	THM-50	THM-51	THM-52	THM-55	THM-56	schist	quartzite	arkose	vein
SiO_2	69.66	67.44	69.68	69.13	67.42	70.00	64.57	69.48	68.68	66.34	68.45	79.85	63.11	42.04
Al_2O_3	14.10	14.66	14.05	14.83	15.51	14.54	16.14	14.97	15.29	14.91	6.93	9.03	19.51	10.52
FeO	3.80	4.66	2.66	3.90	3.89	3.20	6.11	4.03	4.30	5.94	11.46	1.86	3.47	1.15
MgO	1.19	1.07	1.44	1.39	1.04	1.14	1.71	1.03	1.27	1.91	3.90	0.99	1.90	4.02
CaO	1.25	1.45	1.49	1.78	1.69	2.38	1.41	1.62	1.77	1.96	2.82	1.47	2.56	20.26
Na_2O	3.56	2.87	3.34	3.34	3.94	2.24	1.47	2.81	3.34	2.38	0.73	0.38	1.12	0.87
K_2O	4.21	4.08	3.96	3.26	3.32	4.43	4.84	4.76	3.95	4.43	3.23	3.34	7.74	6.34
MnO	0.11	0.11	0.10	0.17	0.06	0.05	0.05	0.03	0.03	0.10	0.19	0.02	0.04	0.11
Volatiles	1.55	1.94	1.63	2.25	1.55	1.22	3.12	1.38	1.55	2.15	3.09	1.15	0.98	15.23
Total	99.43	98.28	98.35	100.14	98.42	99.00	99.42	100.19	100.18	100.12	100.80	98.09	100.43	100.54
Cu	7.8	12.0	14.9	25.0	13.0	14.9	3.9	5.0	2.0	10.1	4.3	2.8	11.9	25.7
Zn	85.7	73.2	76.2	97.0	67.7	57.6	89.3	64.4	55.7	69.5	84.4	51.7	127.4	61.1
Co	34.0	65.8	22.6	20.5	36.0	57.0	33.5	36.8	45.0	45.1	37.4	43.2	19.2	27.8
Ni	3.9	9.3	5.5	13.4	16.1	3.9	5.9	3.1	13.2	13.2	7.0	1.6	21.2	6.0
Rb	70.93	78.42	72.97		74.91	73.55		93.51	64.43	150.9	159.1	71.26	116.7	
Sr	122.7	124.7	195.3		162.8	278.2		147.7	182.6	190.4	186.4	85.60	136.5	

Major components in weight percent; trace elements in ppm.

Volatile contents determined by total weight loss on ignition.

Table 3.--Modal analyses for Hamilton Mound granite.

sample	quartz	K-feldspar	plagioclase	biotite	hornblende
WIHM-1	29.70	32.08	35.96	1.50	1.31
WIHM-4	30.66	33.53	31.61	2.00	2.20
WIHM-7	27.72	35.37	32.50	2.80	1.61
THM-40	29.04	37.75	30.01	3.20	----
THM-42	31.42	39.98	23.80	2.01	2.80
THM-50	32.78	36.63	26.99	2.60	1.00
THM-52	30.72	38.40	26.88	2.00	2.00
THM-54	28.92	40.49	26.99	3.00	<1
THM-55	30.80	34.40	30.59	2.71	1.50
THM-56	26.54	39.82	28.44	2.50	2.70

Modes based on 750 counts per sample; all numbers in volume percent.

calc-alkaline activity of the Penokean orogeny to anorogenic, potentially rift-related igneous activity. Figure 3 shows the distribution of each group. The modal analyses for Hamilton Mound samples plotted with respect to those two fields fall between the Penokean and post-Penokean fields, and are clearly separate from the post-Penokean group.

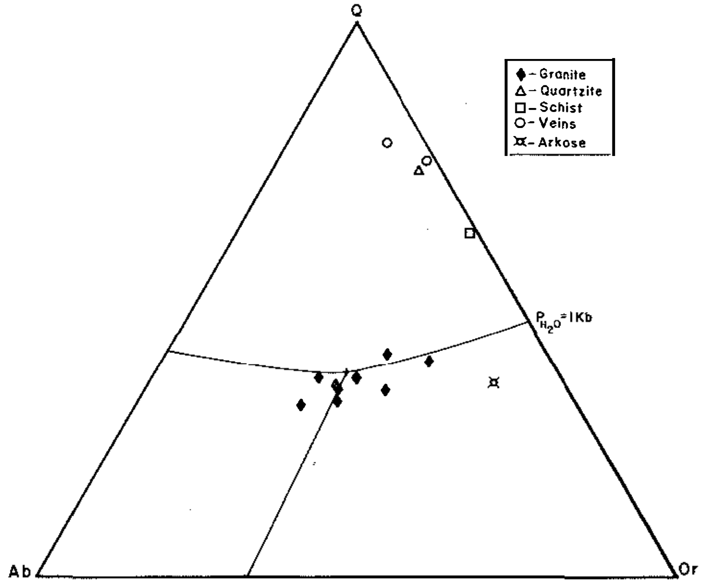

Figure 3.--Quartz-alkali feldspar-plagioclase ternary plot with Hamilton Mound samples plotted together with Penokean and post-Penokean groups of Anderson and others (1980).

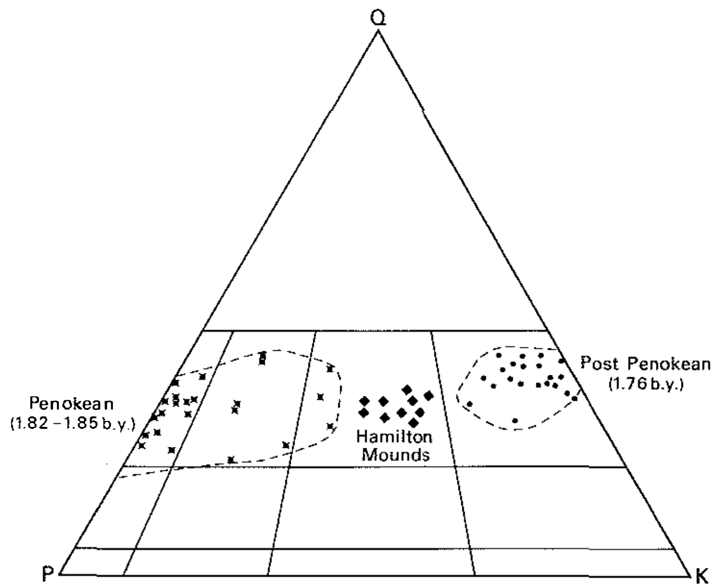

Figure 4.--Ternary eutectic diagram based on normative compositions for granite. Compositions of lithologies at Hamilton Mound plotted; granite is represented by diamonds.

Chemical distinction between the Penokean and post-Penokean has also been made on the basis of total alkali content and relative iron enrichment in comparison to the amount of SiO_2. A plot of total alkalies versus SiO_2 was initially used by Anderson and others (1980) to demonstrate that both Penokean and post-Penokean rock fall within the subalkaline region for igneous rock. However, Penokean rock generally has lower total alkali contents than granite and rhyolite associated with the post-Penokean episode (fig. 5). On this diagram the data for Hamilton Mound fall entirely within the field for the Penokean. It is true that loss of alkalis could shift samples from the post-Penokean to Penokean fields. In the case of the Hamilton Mound samples a corresponding loss of SiO_2 would also have been required. There is, however, no evidence for alkali migration at Hamilton Mound, and the granite contains abundant alkali acceptor phases. The presence of adjacent quartzite makes silica gain far more likely than silica loss.

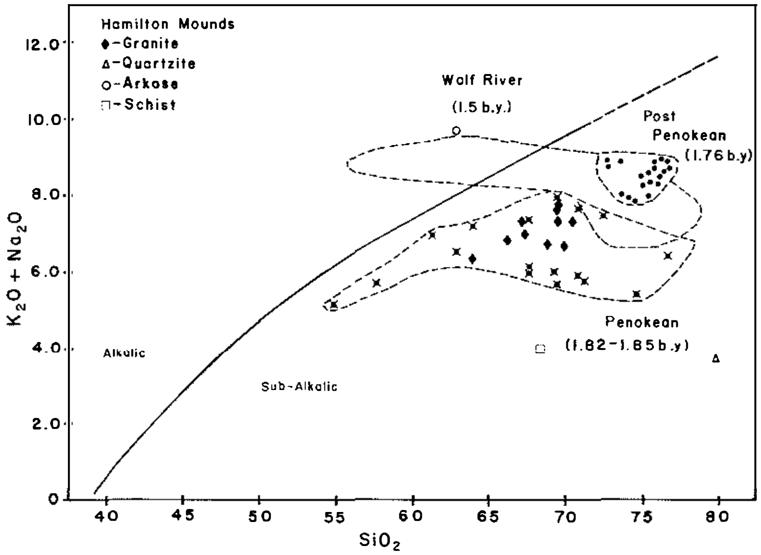

Figure 5.--Plot of total alkalies vs. SiO_2. Data for Penokean and post-Penokean samples from Anderson and others (1980).

Anderson and others (1980) argued further that whereas the Penokean and post-Penokean groups are both subalkalic in nature, the rock related to the post-Penokean is too iron-rich relative to magnesium to have the calc-alkaline character associated with the orogenic plutonism of the Penokean. The post-Penokean suite is then believed to be anorogenic. To demonstrate the compositional difference between the two periods of igneous activity, Anderson and others (1980) plotted the FeO/(FeO + MgO) ratios against SiO_2 for each rock suite. Hamilton Mound samples were plot- ted on the same diagram for comparison (fig. 6). The Hamilton Mound granite samples fall within the Penokean field; they seem compositionally to be more similar to the calc-alkalic, orogenic suite of the Penokean than to the anorogenic granite.

Granitic magmas which are derived from a single source should have similar initial strontium isotopic compositions as well as other chemical similarities. Other post-Penokean granite occurs near the Fox River and in central Wisconsin (fig. 1). A comparison of Rb-Sr data among Hamilton Mound and these two areas (data from Van Schmus and others, 1975) shows major differences with few similarities. The central Wisconsin samples are generally lower in total rubidium than the Fox River Valley suite, usually less than 100 ppm versus 100-200 ppm. Both regions have low total strontium contents (most samples below 50 ppm). As would be expected, the $^{87}Rb/^{86}Sr$ and pre- sent $^{87}Sr/^{86}Sr$ ratios are correspondingly high. The Hamilton Mound samples present a striking contrast to these data. The Hamilton Mound granite has a similar total rubidium content to the central Wisconsin granite. However, it contains much higher concentrations of strontium than either of the two other regions, mostly 100-200 ppm, some samples over 200 ppm, and therefore far lower average $^{87}Rb/^{86}Sr$ ratios and less radiogenic modern Sr. Some of these differences are apparent in figure 7.

The chemical variation among the three suites of rock may be explained by two processes. The first possibility is that the source for the Hamilton Mound granite is different from that of either of the other two groups of post-Penokean felsic rock. The second is that the melt and source were initially similar for all three regions, but that contamination of one or more units occurred during emplacement. The later explanation is more easily evaluated, at least with respect to lithologies now exposed in association with the Hamilton Mound granite.

There are three identifiable candidates for the assimilated material exposed near the granite. Two of these are an arkosic unit and the quartzite which the granite appears to intrude. The third is represented by the schist pieces included in the granite. The fact that fragments of this last rock type, which is not otherwise exposed in the study area, are found in the granite would indicate that some assimilation has taken place, and makes the schist a particularly obvious candidate for a melt-contaminating component.

Composition of the granite and the three possible contaminant lithologies at Hamilton Mound were plotted on a diagram of $^{87}Sr/^{86}Sr$ vs. Sr content (fig. 7), together with Fox River Valley/central Wisconsin samples. There is a broad trend of higher $^{87}Sr/^{86}Sr$ with lower Sr

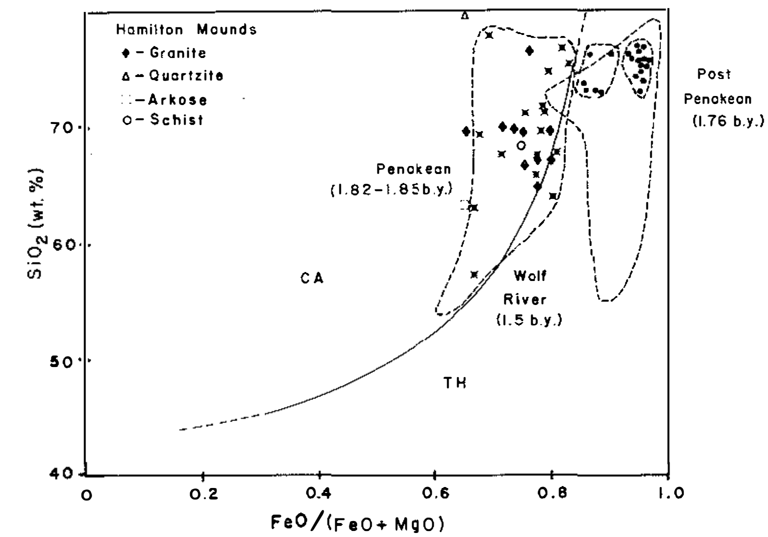

Figure 6.--Plot of SiO_2 vs. FeO/(FeO + MgO). Symbols same as in figure 5. Data for Penokean and post-Penokean samples from Anderson and others (1980).

content which might be interpreted as evidence of mixing. Several points should, however, be made. One is that such a trend will to some extent always be expected, especially when similar rock types are involved, simply as a consequence of the geochemistry of Rb and Sr. Second, simple binary mixing would produce a much tighter hyperbolic curve, although the observed considerable scatter does not necessarily rule out multi-component mixing. It can be seen by inspection of figure 7 that it is not possible to produce the composition of the Hamilton Mound granite by contaminating a Fox Valley/central Wisconsin-type magma with any of the potential contaminants observed at Hamilton Mound. The same conclusion may be reached by inspection of figures 5 and 6, on which the schist, arkose, and quartzite compositions are also included for reference. The diagrams indicate that the distribution of Hamilton Mound granite data points is not consistently attributable to the assimilation of one or more of these components by a melt similar to typical post-Penokean composition. This is true even if the present analyzed composition of the schist or other lithologies is partially altered by reaction with the granite, for the reacted compositions should fall along the same mixing trend. At least one additional high-Sr, low-$^{87}Sr/^{86}Sr$ component is required, and no such material was observed in the area. Figure 7 might alternatively suggest the possibility of the reverse process, assimilation of an unknown contaminant by a Hamilton Mound-type initial magma to produce the Fox Valley/central Wisconsin array. However, in either case the heat budget of a granitic magma is unlikely to be sufficient to allow assimilation and complete digestion of enough contaminating material to cause the large shifts in bulk chemistry noted in figures 5 and 6, or the decrease of 50 percent or more in total Sr which would be required. In the case of the Hamilton Mound granite a modest heat budget is suggested both by the lack of high temperature mineralogy in the schist inclusions, and by the absence of flow structures, which supports batch emplacement of the magma. Also, boundaries of schist inclusions are generally sharply defined.

Mixing among the tested end members thus cannot account for the composition of the Hamilton Mound granite, and the apparent discrepancy between its age and the correspondingly expected composition. This is not to say that there has been no assimilation, but rather that a melt similar to that of the Fox River Valley and central Wisconsin felsic rocks could not have assimilated any of the medasedimentary rock types visible at Hamilton Mound to produce the composition of the porphyritic granite. The chemical and isotopic data require at least one additional component, so far unseen, if the granite's composition is to be accounted for by the mixing hypothesis tested. Taken together with the heat-budget argument above, the data strongly suggest that the original melt for the granite at Hamilton Mound was distinctly different from the source of the granite and rhyolite of the Fox River Valley and the central Wisconsin granite. This raises serious doubts about the possibility of distinguishing Penokean from post-Penokean felsic rock solely on the basis of geochemical signature.

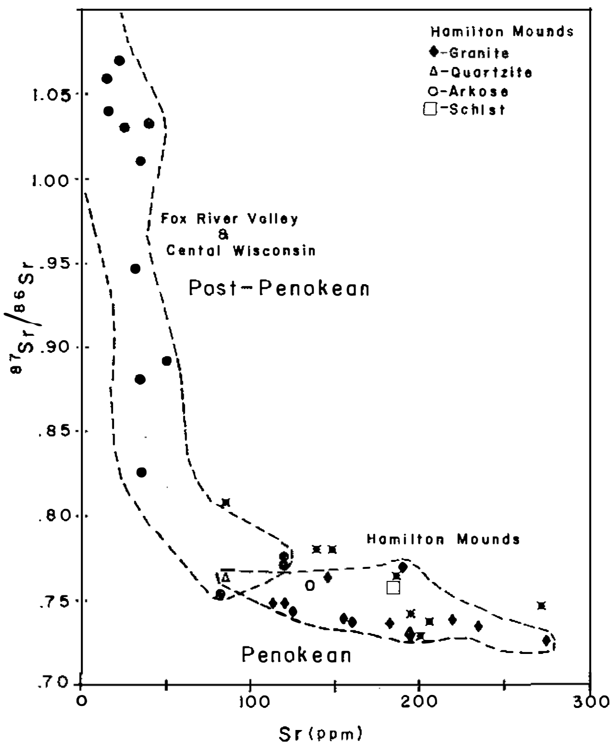

Figure 7.--Test of mixing hypothesis relating Hamilton Mound to post-Penokean rock of the Fox River Valley and elsewhere in central Wisconsin, and to Penokean felsic rock. Data on non-Hamilton Mound units from Van Schmus and others (1975).

SUMMARY AND CONCLUSIONS

Petrologically, the porphyritic granite at Hamilton Mound is a typical granite. The most abundant minerals are quartz, microcline, and plagioclase. The phenocrysts are usually microcline. Perthitic, myrmekitic, and graphic intergrowths are fairly common throughout the granite. Ternary plots of the normative and modal compositions indicate little compositional variation, as would be expected for a small, fairly homogeneous granite body. The overall composition is close to a plausible thermal minimum on a QAP diagram. The granite appears to intrude the overlying quartzite; its emplacement may have been fault controlled.

Rb-Sr whole-rock analysis of the granite yielded an errorchron with slope age of 1,940 $\pm$ 400 Ma, and initial $^{87}Sr/^{86}Sr$ = 0.702 $\pm$ .008. The age is far better defined by a U-Pb zircon age obtained by Van Schmus (1983, oral communication) of 1,764 $\pm$ 4, Ma which presumably indicates the time of primary crystallization of the porphyritic granite. A Rb-Sr mineral isochron records a post-emplacement metamorphic event which accounts for the scatter in the whole-rock data. The age of this event as defined by the mineral isochron is 1,580 $\pm$ 160 Ma; it probably corresponds to the pervasive regional overprint at 1,630 to 1,700 Ma observed throughout Wisconsin.

A comparison of chemical and isotopic compositions of the Hamilton Mound granite with other granite and rhyolite of similar age in central Wisconsin disclosed chemical differences. Other felsic rock has previously been divided into Penokean and post-Penokean (1,760 Ma-old) suites on the basis of apparently distinct chemical signatures. However, the Hamilton Mound granite of post-Penokean age is chemically far more similar to the Penokean-age rock. Simple assimilation cannot readily account for the data. It may, therefore, not be possible to distinguish the Penokean from post-Penokean felsic rock in central Wisconsin on a chemical basis.

ACKNOWLEDGMENTS

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MAGMATISM AND THE BARABOO INTERVAL: BRECCIA, METASOMATISM, AND INTRUSION

by

J.K. Greenberg[1]

ABSTRACT

Breccia consisting of quartzite fragments surrounded by white, vein quartz is known to occur in Wisconsin at several exposures of Baraboo-type metasedimentary rock. These quartzite breccias include those at Rock Springs on the north limb of the Baraboo Syncline, Hamilton Mound, Necedah, Battle Point, Vesper, Waterloo, and McCaslin Mountain. With the exception of McCaslin Mountain in the northeast, all the breccias occur in the central or south-central part of the state. Most of the brecciated quartzite has intrusive contact with plutonic rock. Various types of hydrothermal alteration (metasomatism) are apparent in the brecciated outcrop and other exposures of quartzite intruded by granitic or dioritic magma. The most common metasomatic features are quartz crystal-lined pockets and clay-mica segregations, feldspar porphyroblasts in altered quartzite, hematite segregations, and quartz-tourmaline veinlets.

A present interpretation of the breccias is that they are analogous to the stockwork of quartz veins produced around the upper levels of porphyry-copper mineralized plutons. During magma intrusion, the roof rock of quartzite was fractured and soaked in hydrous granitic fluids. The fluids and their particular effects vary with distance from source plutons. Thus, as in some Wisconsin examples quartz veins and breccia grade into pegmatite dikes as an intrusion is approached. Another possible analogue for the Wisconsin examples are explosive breccias developed in quartzite above volatile-rich appinite intrusions.

INTRODUCTION

Several exposures of quartzite deposited during the Proterozoic Baraboo tectonic interval (Dott, 1983; Greenberg and Brown, 1983, 1984) contain breccia with a white vein-quartz matrix. These exposures are widely distributed across central and northern Wisconsin (fig. 1). Intrusions and metamorphic effects of various types have also been observed in brecciated and some unbrecciated outcrops of Baraboo-type quartzite. Intrusive contact and the character of the affected sedimentary rock suggest that these features may be related to anorogenic magmatism which occurred 1,760 and 1,500 Ma. Intrusive bodies are viewed as sources of the combined heat, fluid mobility, and stress necessary to cause much of the observed metamorphism, commonly hydrothermal metasomatism and deformation. Evidence linking the various phenomena is at present mostly empirical. However, the volume of evidence is substantial and sufficient to warrant more comprehensive investigation.

Baraboo-type quartzite and associated metasedimentary rock in Wisconsin, Minnesota, Iowa, and South Dakota are considered to be a product of anorogenic, epicratonic deposition (Greenberg and Brown, 1984). The period of deposition, probably between 1,760 and 1,630 Ma, was preceded by orogenesis from 1,900 to 1,800 Ma, the Penokean orogeny. After 1,800 Ma, magmatism, sedimentation, and deformation in Wisconsin were apparently responses to epeirogenic subsidence and elevation of maturing continental crust (Greenberg and Brown, 1984; Rogers and others, 1984).

Intrusion, metamorphism, and brecciation appear to post-date most folding and associated cleavage in Baraboo interval rock. If, as several investigators have proposed, the quartzite and associated metasedimentary rock were folded about 1,630 Ma (Smith, 1978; Geiger and others, 1982; Dott, 1983; Greenberg and Brown, 1984), then this could be considered a maximum age of intrusion. At Rib Mountain, McCaslin Mountain, and Waterloo (fig. 1) quartzite was cut by granite about 1,500 Ma. However, a recently obtained U-Pb zircon age is anomalous; the granitic rocks intruding deformed quartzite at Hamilton Mound (fig. 1) are near 1,760 Ma (R. Van Schmus, unpublished data; Taylor and Montgomery, this volume). A tectonic sequence similar to Hamilton Mound is apparent at Baxter Hollow (Baraboo, fig. 1) where granitic magma intruded quartzite, possibly sometime between 1,760 and 1,630 Ma (Dott and Dalziel, 1972; Smith, 1983). A potential explanation for the apparent age discrepancies is that the broad, open folds at Hamilton Mound (Ostrander, 1931) and Baraboo (Dalziel and Dott, 1970) actually followed intrusion and imposed only minor, brittle strain on plutonic rock. Another possible explanation is that some folding occurred about 1,760 Ma as well as later. Because some foliated granitic rock in northern Wisconsin were recently

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determined to be about 1,760 Ma (Van Schmus, unpublished data), the folding at Hamilton Mound, Baraboo and elsewhere may also have taken place about the same time. There is, however, no direct evidence for this.

With the exception of Mancuso (1960) and Dalziel and Dott (1970, p. 29, 105), there has been little previous discussion of the quartzite breccia. Dalziel and Dott attributed the brecciation to some unspecified "explosive hydrothermal activity." Mancuso reported breccias associated with intrusive igneous activity at McCaslin Mountain. Other intrusive relationships have been recognized at Rib Mountain (mafic dike, LaBerge and Myers, 1983) which is itself a quartzite xenolith in syenite, Waterloo (granite pegmatite, Dott and Dalziel, 1972), Hamilton Mound (Greenberg and Brown, 1983; Taylor and Montgomery, this volume), Baxter Hollow (Gates, 1942; Petro, 1983), and most recently, Necedah (fig. 1, see below). Metasomatic alteration in the form of microcline intergrown with quartz in quartzite had previously been described by Ostrander (1931) at Hamilton Mound. Before quarrying exposed the intrusive contact, Ostrander speculated correctly that the Hamilton Mound quartzite may have been intruded. Gates (1942) also suggested that granitic veins in the Baraboo quartzite and the petrography of Baxter Hollow granite indicated intrusion-related hydrothermal activity. His interpretation has not been widely accepted, but reexamination of Baxter Hollow in this study provides additional evidence supporting many of the observations of Gates (1942).

DESCRIPTION AND INTERPRETATION OF LOCALITIES

Baraboo

There are three areas of interest in the Baraboo Quartzite, two on the south and one on the north limb of the Baraboo Syncline (fig. 1).

Rock Springs

In the Rock Springs area (sec. 28, 29, T. 12 N., R. 5 E.) well developed breccia zones were produced without any overt evidence of intrusion. White, vein quartz commonly first appears as bedding-parallel fracture filling near distinct contacts between conglomerate and finer-grained beds (fig. 2). The veins become more abundant and grade into breccia zones where there is much less planar preference of vein orientations. Dalziel and Dott (1970) concluded that the breccia was not fault phenomena in that neither quartzite within or on either side of the zone had been displaced. Typical breccia (fig. 3) consists of angular fragments of purplish quartzite surrounded by white to colorless, vein quartz. The veins possess pockets up to about a metre in diameter that are lined with quartz crystals. Other pocket constituents include interstitial dickite or kaolinite and well crystallized hematite (to 2 cm in diameter). S.W. Bailey of the University of Wisconsin reported that fluid inclusions in quartz crystals indicate temperature of formation near 106 °C (Dalziel and Dott, 1970, p. 105). Geiger (this volume) has determined by x-ray diffraction that some feldspar is also preserved in clay-rich quartzite pockets. Although intrusive rock is not exposed at Rock Springs or elsewhere on the syncline's north limb, gravity and magnetic anomaly maps (Hinze, 1957) indicate the possibility of plutons beneath the brecciated zones.

Highway 12 - South Bluff

On the east side of Highway 12 where it crosses the south limb of the syncline (sec. 34, T. 11 N., R. 6 E.) the presumably basal Baraboo quartzite is unusually fractured in a boxwork of joints and quartz veins (fig. 4). Most of the joints are coated with clay and secondary quartz. No displacement along any fracture surface is apparent. Quartz veining and the characteristic pockets are less well developed here than at Rock Springs, but the fracturing and quartz-clay concentrations are similar. Drill core from the area between Highway 12 and Baxter Hollow contains quartzite intruded by granitic rocks.

Baxter Hollow

Baxter Hollow designates an area with over 400 m of continuous outcrop along the south limb of the syncline (sec. 32, 33, T. 11 N., R. 6 E.). This area is of particular interest because of the controversial nature of the contact between granite and quartzite. Previous workers have ascribed the contact either to intrusion (Gates, 1942) or were uncertain as to the nature of the contact (Dott and Dalziel, 1972). Recent investigations by the Wisconsin Geological and Natural History Survey and Petro (1983) have concluded that the Baxter Hollow granite was a complex intrusive into quartzite.

An unusual zone of interaction varying from less than one to several metres in width separates relatively unaltered Baxter Hollow granite from unaltered quartzite and phyllite. Within this

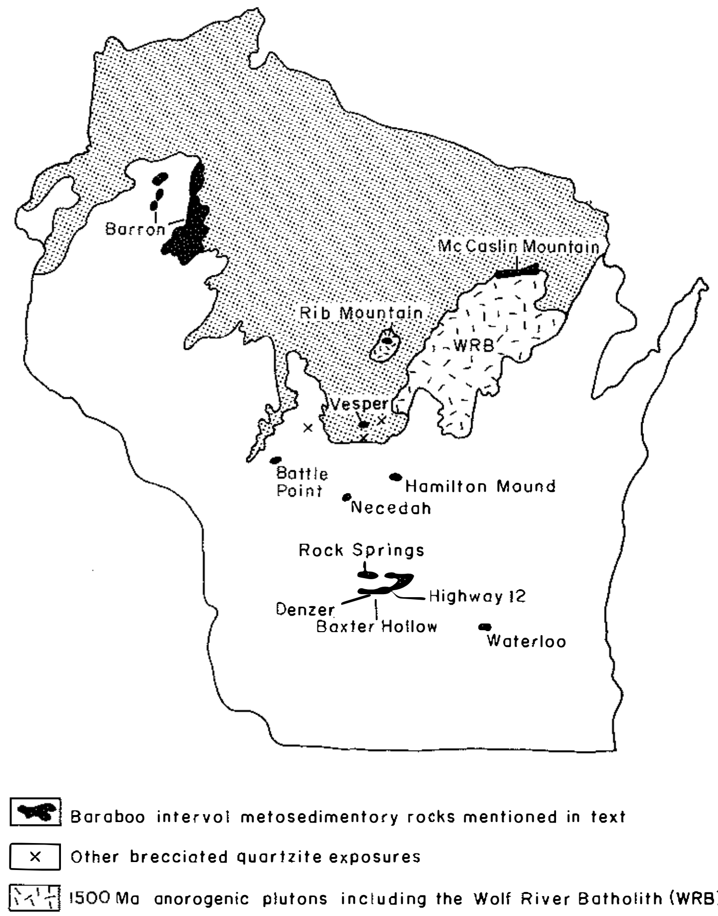

Figure 1.--Distribution of brecciated quartzite of the Baraboo interval. The unbrecciated Barron Quartzite is also shown. The northern region of Precambrian exposures is shaded.

zone the intrusion becomes albite rich, is granophyric in texture, and plagioclase becomes progressively sericitized toward the quartzite. Nearest the contact intense foliation developed in the altered granite led Gates (1942) to propose the existence of a shear zone. The foliation can also be explained mechanically as a result of forceful intrusion. The sericite-rich rock probably received ductile strain along the intrusive contact. Where observed, the foliation is parallel or subparallel to the contact. Above the contact within quartzite, laths of albite have been introduced into quartz-rich rocks (greater than 65 percent quartz). These laths are referred to as porphyroblasts by Gates (1942). The albitized and sericitized rocks are commonly green, perhaps due to the reduction of iron.

Hematite concentrations occur sporadically on both sides of and close to the contact. The granitic rock, especially where altered, contains as much as 60 percent hematite with quartz and sericitized plagioclase. Near the intrusion a unique quartzite breccia with highly strained angular rock fragments in a matrix of over 40 percent hematite (fig. 5) appears to be a brittlely deformed version of hematitic conglomerate. Some samples of undeformed conglomerate from the same area also contain hematite in high proportions as matrix, but surrounding more coherent quartz and phyllite pebbles. In most cases hematite is well crystallized, perhaps indicative of hydrothermal development. The iron oxide could have originally been sedimentary and was later mobilized by intrusion. Another possibility is that the iron was a metasomatic addition from a magmatic source (Hauck and Kimball, 1984). The sericite, hematite, and quartz mineralogy of altered rock at Baxter Hollow represent the same chemical mobility present in the more conventional Baraboo breccias.

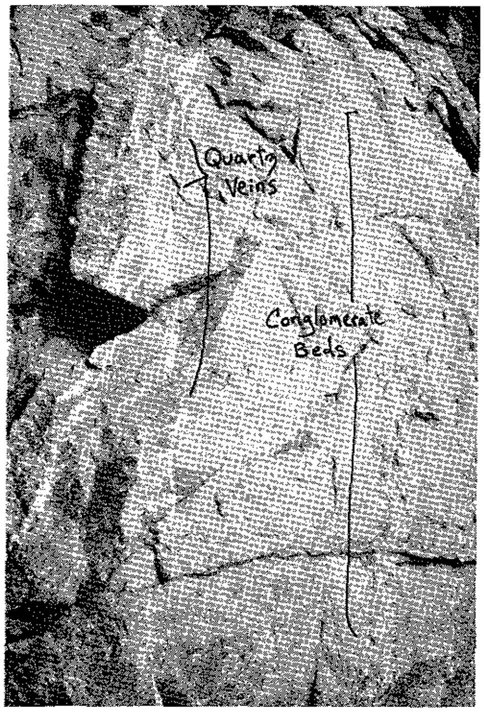

Figure 2.--Outcrop of Baraboo Quartzite at Rock Springs: White quartz veins are oriented parallel to bedding as shown by pebble conglomerate on right.

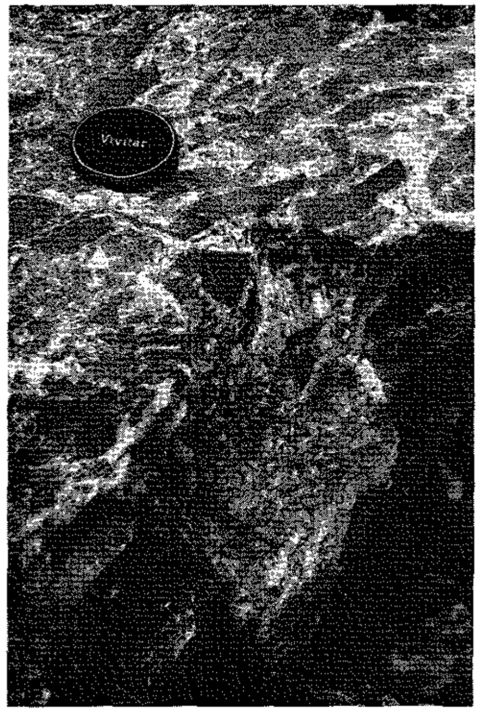

Figure 3.--Close up of typical Baraboo quartz breccia. Dark fragments are purplish quartzite.

Figure 4.--Outcrop of Baraboo Quartzite on Highway 12 (south range). Note the network of thin fractures lined with clay minerals and some quartz.

Gates (1942) and Petro (1983) interpreted quartz-tourmaline veinlets as additional late-state hydrothermal phenomena in or near Baxter Hollow. The thin veins cut both altered intrusive rocks and quartzite.

Necedah

Quartzite exposures at Necedah (fig. 1) include a large quarry cut into a prominent ridge and two other smaller outcrop ridges. All of these are overlain by Cambrian sandstone which has been variably eroded. The quarry (sec. 24, T. 18 N., R. 3 E.) exposes ferruginous quartzite which is locally well brecciated (fig. 6). Many of the fragments and blocks in the breccia are metachert, red-gray to black or green in color (Greenberg and Brown, 1983) and are surrounded by networks of white vein quartz. Color is a function of the oxidation state of iron, grain size of recrystallized quartz, and fracturing. Quartz crystal pockets with partial to complete clay filling occur between fragments in the veins. Unlike Baraboo the pockets and fracture fillings at Necedah also contain well crystallized books of muscovite as large as 1 cm in diameter. The presence of coarsely crystalline muscovite in quartzite breccia at the quarry suggested the possibility of an unexposed intrusion (Greenberg and Brown, 1983). Recent quarry expansion at Necedah has confirmed this suspicion with the uncovering of breccia with granitic matrix and fragments. Water-well records also indicate that granite and diorite occur below the quartzite in the immediate area.

Deformation and alteration effects in the quartzite fragments developed before and perhaps during brecciation. Previously-foliated quartzite fragments are randomly oriented in the breccia. The typically dark green to black fragments are limited to the area where igneous material is exposed in the quarry. Reddish or lighter-colored quartzite occurs elsewhere at Necedah. At present there is no evidence to suggest any deformation of Necedah quartzite after brecciation.

The igneous breccia contains relatively few quartz veins compared with the quartzite breccia. Constituents of the igneous breccia include fragments of variably altered aplitic granite, dark colored quartzite and metachert, ferrugenous metaargillite, and quartz, perthite and plagioclase megacrysts. All these components range in size from several mm to blocks near 1 m in the longest dimension. The breccia matrix surrounds the large fragments with highly strained quartz, relatively unstrained microcline and sodic plagioclase, chlorite, muscovite-sericite, and hematite. Zircon occurs as a relatively abundant minor phase.

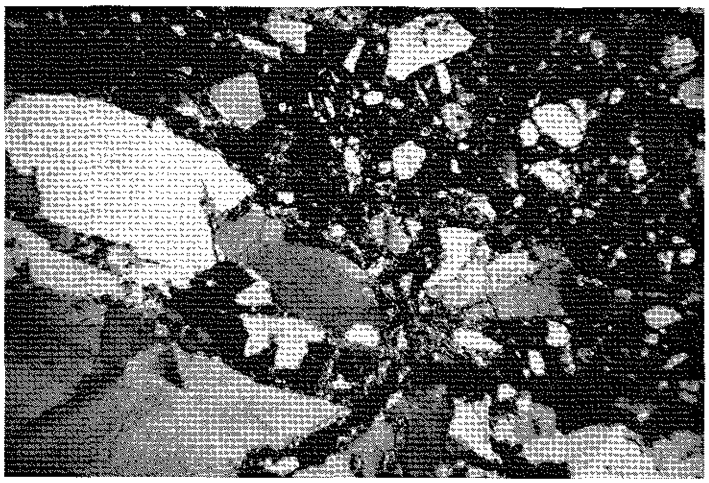

Figure 5.--Photomicrograph of brecciated quartzite from Baxter Hollow. Dark matrix is predominantly hematite crossed nicols. Long dimension about 8 mm.

Although mineral grains and rock fragments in this igneous breccia are variably strained, no uniform recrystallization is apparent in the rock as a whole. There is much evidence for the near solid-state reaction of various breccia components. Some of the large feldspar grains have been partially replaced by quartz and sericite. Radial clusters of chlorite occur in some samples as a boundary phase between argillaceous and granitic fragments. Well-crystallized muscovite has grown within irregular veinlets that transect feldspar megacrysts and continue into the matrix.

Another major Necedah outcrop is located southeast of the quarry (sec. 19, T. 18 N., R. 4 E.) and consists of a different, unbrecciated rock. The exposed ridge is glacially polished and is composed of a white to gray quartzite with recognizable, but contorted, cross beds (fig. 7). At Baraboo similar contorted structures have been interpreted as slumped cross beds oversteepened by synsedimentary deformation (Dalziel and Dott, 1970, p. 16, 154). This may be the correct interpretation at the exposure described by Dalziel and Dott, but the degree of strain and evidence of thermal metamorphism from intrusion at Necedah, Hamilton Mound (see below), and possibly Baraboo suggests the possibility of a tectonic origin for some of the contorted structures.

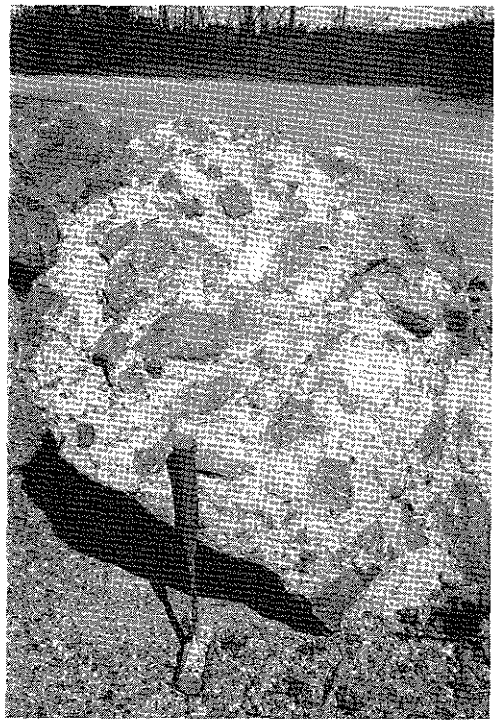

Figure 6.--Boulder of intrusion breccia at Necedah. Dark fragments are green to black quartzite and argillite. Lighter-colored materials are altered granitic rock and clay/quartz-rich matrix.

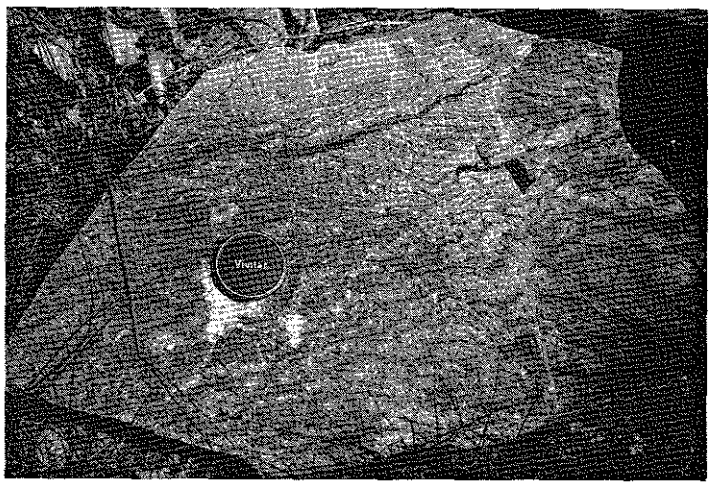

Figure 7.--Broken outcrop at Necedah showing contorted (distorted?) crossbedding.

In the unbrecciated Necedah outcrop quartz veins are present both parallel to and transecting bedding. Dark, chlorite and muscovite-coated fractures also transect bedding at high angles. Coarse-grained perthite with or without muscovite has been identified during detailed examination of some of the larger quartz veins. These veins are appropriately termed granite pegmatites.

Hamilton Mound

A cluster of quartzite ridges in northeast Adams County, Wisconsin (sec. 36, T. 20 N., R. 6 E. and sec. 31, T. 20 N., R. 7 E.) are collectively known as Hamilton Mound (fig. 1). Studies of the quartzite and intruding granitic rock (Greenberg and Brown, 1983; Taylor and Montgomery, this volume) have only recently taken place after initial description by Ostrander (1931). Granitic intrusion, quartz veins, breccia, and evidence of metasomatism, are exposed in one large quarry and are similar to corresponding features described for Baraboo.

The intrusion in the quarry is present as two discernable types, both apparently contaminated. One type is porphyritic with red microperthite and plagioclase phenocrysts in a matrix of chlorite and strained quartz. Alkali feldspar, plagioclase, and biotite are less abundant matrix phases. Taylor and Montgomery (this volume) determined little evidence for contamination of the porphyritic intrusion, but their chemical data alone is inconclusive. The nature of the other intrusive rock type points more conclusively to contamination. This second type occurs in an area within 40 m of the porphyritic variety, but is finer grained and otherwise different in appearance. The intrusive contact in this area is nearly horizontal with quartzite at the top of the quarry and magmatic rock exposed in a vertical surface for about 15 m below. Where exposed furthest below the contact, the intrusion appears least modified. Here the rock is an unusually quartz-rich medium- to fine-grained foliated granodiorite. Quartz content varies from about 40 percent to 65 percent at the contact. Even the most quartzose rocks contain sodic plagioclase, microcline, biotite, and hornblende in various proportions. Nearest the intrusive contact, rocks are greenish, some exhibiting an oddly distorted banding and inclusions composed of concentric zones rich in biotite, muscovite, chlorite, and quartz (fig. 8). The rock in the contact zone is certainly neither of unmodified sedimentary or magmatic origin. The green color is interpreted as a result of hydrothermal iron reduction. Undeformed, tourmaline-bearing pegmatite which cuts the hybrid contact-rock (fig. 9) indicate a later stage of magmatic activity.

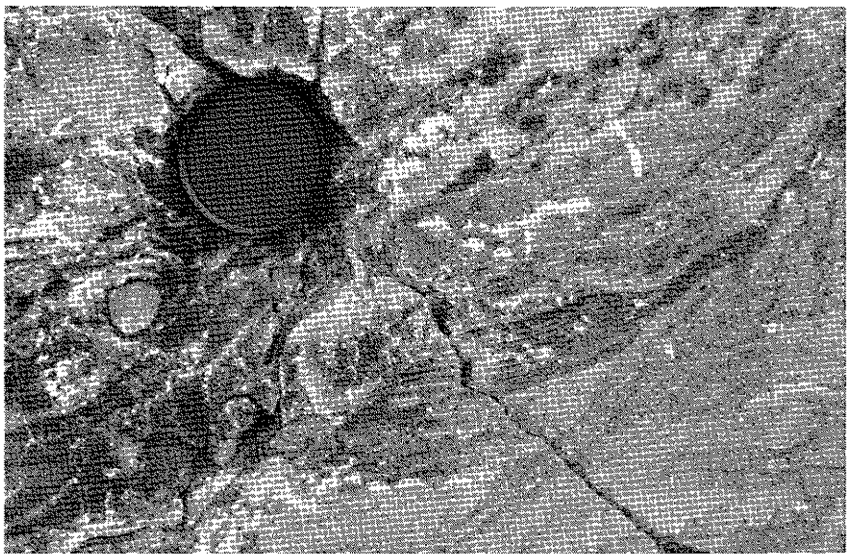

Figure 8.--Close up of outcrop at Hamilton Mound showing flow lamination (distorted cross beds?) and xenolith to upper right of lens cap. The rock shown is over 65 percent quartz.

Banding in the contact zone is not unlike the appearance of contorted cross-bedding preserved in the green quartzite above the contact (Greenberg and Brown, 1983). In another, smaller quarry away from the intrusion quartzite is purplish and contains only undeformed cross beds. This may be a specific analogue of the nature of bedding structures at Baxter Hollow and Necedah.

Silica mobility is evident at Hamilton Mound, especially within and above the porphyritic granite. The intrusion contains abundant quartz veins rich in chlorite and calcite. These veins increase in number and tend to merge with subhorizontal silicified faults and fractures which occur near the contact. Ductile and brittle strain is most intense near the contact. Small quartz veins that penetrate both altered granite and metasomatized quartzite have stylolites developed along quartz grain boundaries (fig. 10). Secondary silica, presumably mobilized by pressure solution, is observed as dark, fine-grained to cryptocrystalline matrix material in fractures, faults, and some hybrid rocks. The timing of the causal deformation is interpreted as late to post intrusion.

Brittle fracturing and brecciation occur respectively over 75 m horizontally and 25 m vertically from the porphyritic intrusion as it is now exposed. The fractures seen in one face of the quarry cut near vertical bedding at high angles (fig. 11). The fractures may represent either fold-related cleavage or jointing. Fillings of kaolin and doubly terminated quartz crystals also suggest that these fractures may be the result of magmatic hydrothermal activity.

Breccias at Hamilton Mound are present both in the larger quarry and in the smaller quarry, where there is no overt indication of intrusion. Above the intrusion the quartzite breccia appears virtually identical to most of the others described in this paper. As Taylor and Montgomery (this volume) described, some of the blocks surrounded by white vein-quartz are granite or are quartzite cut by pink granite aplite. All that this means for sure is that at least some distinctly magmatic activity preceded some hydrothermal activity. Hamilton Mound is one place where complex magmatic-hydrothermal relationships are amenable to further analysis. A temporal sequence of cross-cutting features could be derived from the well exposed and diverse types of evidence.

Waterloo

Quartzite and associated metapelite exposed near Waterloo, Wisconsin (fig. 1) are similar in composition and general appearance to rock 50 km to the northwest at Baraboo. Some of the Waterloo outcrops contain granite pegmatite dikes (fig. 12), sec. 27, T. 9 N., R. 13 E. and others (sec. 28, T. 9 N., R. 13 E.) contain andalusite-bearing, phyllitic beds (Geiger and others, 1982). Peak metamorphism and dikes are evidently the same age, about 1,500 Ma (Dott and Dalziel, 1972; C. Guidotti, unpublished data) and are interpreted as products of a major episode of anorogenic thermal activity (Greenberg and Brown, 1984; Greenberg and Brown, 1986) A penetrative foliation and cross-cutting crenulations in the phyllites predate peak metamorphism and may reflect 1,630 Ma tectonism.

The pegmatite dikes at Waterloo consist mostly of red microcline, perthitic orthoclase, muscovite, and quartz. Dike margins are sharp with only minor growth of microcline grains in the adjacent quartzite wall rock. No other metasomatic features were observed in these exposures.

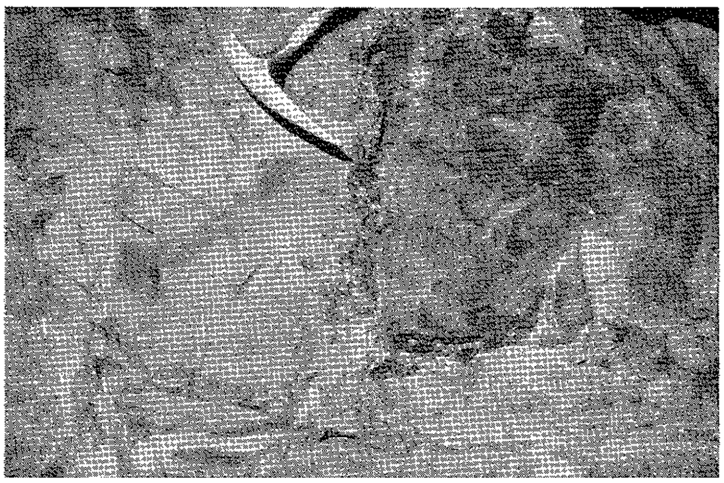

Figure 9.--The hammer points to a nearly vertical pegmatitic segregation in the quartz-rich Hamilton Mound exposure. Tourmaline crystals occur as black rods up to 2 cm in diameter.

A large outcrop area of brecciated quartzite occurs several kilometres north of the pegmatite (sec. 3, T. 9 N., R. 14 E.). Joints at high angles to bedding become vein-quartz filled as the zone of brecciation is approached. This zone is about 30 m wide along strike of bedding and perhaps twice as wide perpendicular to bedding. The breccia is well developed with dark gray quartzite fragments in white quartz matrix (fig. 13). Some quartz-crystal lined pockets are present, but no clay filling was in evidence. There is no direct evidence associating the breccia at Waterloo with intrusion.

McCaslin Mountain

McCaslin Mountain is a 40 km-long ridge of quartzite exposed along the northern margin of the intruding 1,500 Ma-old anorogenic Wolf River batholith in northeast Wisconsin (fig. 1). Studies have documented the intrusive and thermal metamorphic effects at different locations along the ridge. Chloritoid, andalusite, garnet, and sillimanite occur in various combinations in quartzose-pelitic rocks of appropriate composition at the main ridge and at other nearby exposures

Figure 10.--Photomicrograph of "shear zone" microstylolite in quartzite above granite at Hamilton Mound. Crossed nicols, long dimension about 3 mm.

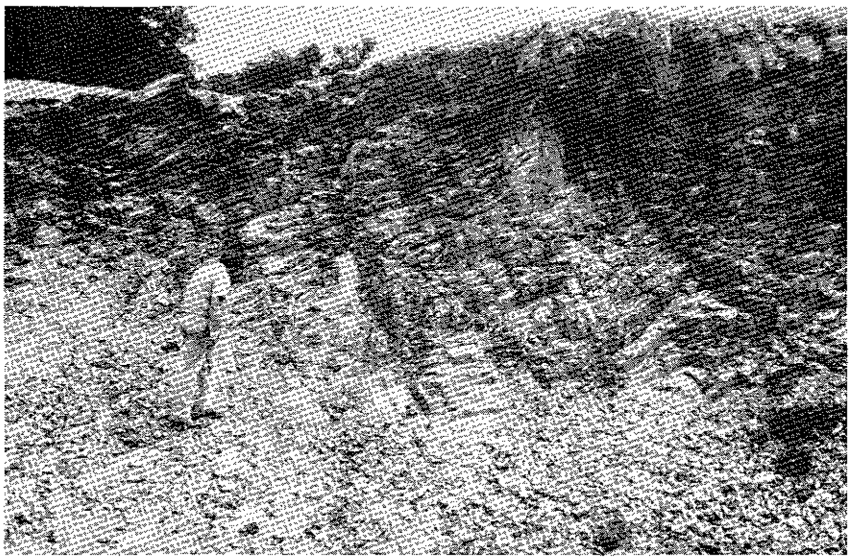

Figure 11.--Well-developed fracture cleavage at high angles to bedding in Hamilton Mound quartzite. Long dimension about 20 cm.

to the north and east (Mancuso, 1960; Olson, 1984; Greenberg and Brown, 1986). The development of these metamorphic phases is attributed to contact with rock of the Wolf River batholith.

Mancuso (1960, p. 35) described outcrop including steep rock walls along the eastern end of the ridge which shows well developed brecciation. The breccia occurs above an intrusive contact where massive granite and granite dikes have been injected into fractures. Mancuso (1960) wrote that "a continuous transition exists from granite to quartzite invaded by large granite veins and dikes, to quartzite breccia healed by later secondary quartz" and "the white secondary quartz can be traced downward to an origin in a granite vein or stringer."

There has been some metasomatism of the quartzite near contact with the Hager porphyry and other granitic members of the Wolf River batholith. Pink microperthite and albite grains occur in McCaslin quartzite at various locations (Mancuso, 1960) but are particularly conspicuous in conglomeratic rocks cut by granitic dikes. The extent of metasomatic replacement at McCaslin Mountain, Waterloo, and Rib Mountain, where quartzite was intruded about 1,500 Ma, is not as great as Baxter Hollow and Hamilton Mound, where the intrusions are thought to be older. It is possible that the earlier magma was more fluid and volatile rich. It is also likely that the sedimentary rock would be more permeable and retained more of the original water of deposition 1,760 Ma than about 250 m.y. later.

Rib Mountain

Rib Mountain near Wausau, Wisconsin (secs. 7, 8, 9, 10, 15, 16, 17, T. 28 N., R. 7 E., fig. 1) is a 3.5 x 1 km quartzite xenolith in syenite of the 1,500 Ma-old Wausau granite-syenite complex

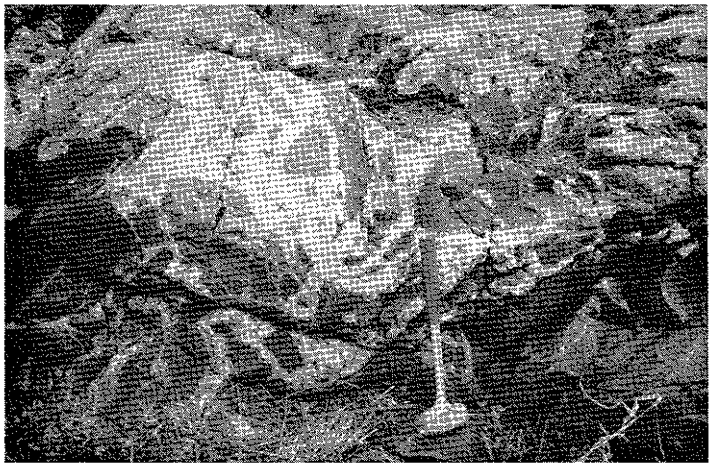

Figure 12.--Outcrop of 1,450 Ma pegmatite cutting Waterloo Quartzite. Quartzite is the darker, gray lithology along the bottom.

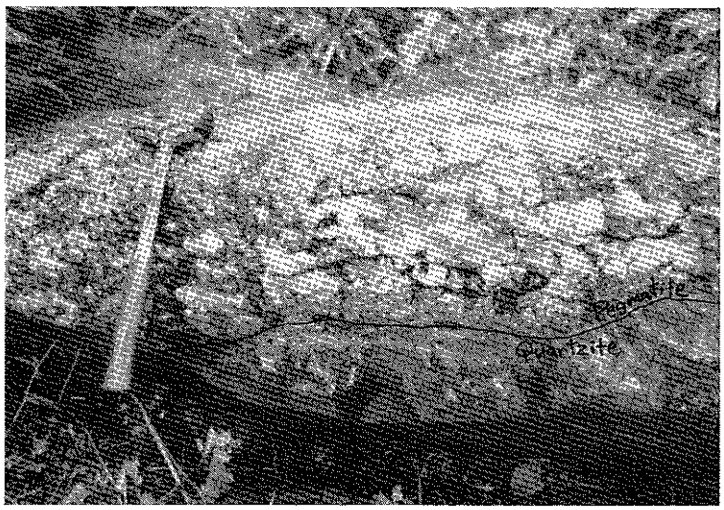

Figure 13.--Quartzite breccia outcrop at the far northern end of the Waterloo quartzite exposure area.

(LaBerge and Myers, 1983). Due to thermal metamorphism the quartzite is generally recrystallized to coarse grain sizes, and contains sillimanite in the more aluminous layers. The typical white to gray color of the quartzite may be a result of a metamorphic bleaching which mobilized iron, precipitating the iron along discrete fractures.

There is no direct evidence of granitic magma penetrating the quartzite at this location although a large, undeformed (Keweenewan-age?), basaltic dike occurs in the quarry atop Rib Mountain (sec. 8, T. 28 N., R. 7 E.). No breccias have been observed here, but large quartz veins and unusual zones of alteration exist in several areas of the quarry.

The quartz veins contain crystal pockets, some with large crystals of muscovite (to 2 cm) grown over quartz. The veins and alteration zones also contain 0.5 mm to 2 mm crystals of anatase and brookite (A. Falster, unpublished data). The exact nature of the alteration zones has not been determined. Detailed petrographic and chemical analysis will be necessary to determine the origin of the zones. The tabular geometry of the alteration which parallels some joint surfaces may imply dike or fracture control. As seen in thin section alteration consists of sericite and secondary quartz which have overgrown the preexisting rock fabric. In two samples the protolith appears to have contained pyroxene phenocrysts, possibly suggesting earlier dikes. As with other described exposures, the source of the distinctly aluminous and siliceous metasomatizing fluids is interpreted to be an intrusion acting on variably aluminous quartzite.

Battle Point

At Battle Point (sec. 9, T. 21 N., R. 2 W., fig. 1) a single outcrop of highly brecciated and fractured pink quartzite is exposed at the eastern end of a ridge oriented parallel to faint east-west bedding (Brown, 1983, p. 31). Major white quartz veins are subparallel to bedding. The general macro- and microscopic appearance of the angular rock fragments suggests a metachert. However, although all of the fragments appear to be similar, some are silicified felsic volcanic rock containing feldspar phenocryst laths replaced by sericite, clay, and quartz. These pseudomorphs were also observed in a clay-rich, quartz vein. This atypical breccia is another example of granitic magma-quartzite interaction.

Other Brecciated Quartzite

Several other quartzite breccias similar to those described here have been recognized during regional geologic mapping in northcentral Wisconsin. These other exposures include a small quarry near Vesper (Greenberg and Brown, 1983, p. 17-20), excavated farm ponds and road-side, ditch outcrops (fig. 1). None of these examples show any overt evidence of associated intrusive activity. Their one common attribute is geographic distribution in a region of extensive anorogenic magmatism.

Non-brecciated Quartzite

Among quartzite correlated with Baraboo interval rock, only the westernmost exposures (the Barron Quartzite in northwestern Wisconsin and the Sioux Quartzite in Minnesota, Iowa, and South Dakota) have experienced neither brecciation nor metasomatic alteration. One reason for the apparent absence may be the lack of detailed studies on the Barron exposures in particular. It is more likely that these only mildly deformed and metamorphosed western quartzites (Greenberg and Brown, 1984) were not in the main region of anorogenic intrusion. Keweenawan-age, diabase dikes which have been found in the Barron and Sioux contributed little more than minor contact thermal effects. Altered felsite known from subsurface data to occur in the Sioux Quartzite are possibly about 1,500 Ma (Lidiak, 1971), but these have not been well studied.

DISCUSSION

The above evidence strongly implies that brecciation and metasomatism of quartzite in central and northern Wisconsin originated with intrusion. The relationship between mostly thermal metamorphism of Baraboo interval rock and anorogenic magmatism has been previously discussed (Greenberg and Brown, 1984). The interaction between anorogenic plutons and quartzose cover rock is not well documented in the literature. Two pluton-cover rock relationships that may be pertinent to the association in Wisconsin are the nature of porphyry-copper mineralization and breccia zones developed around appinitic intrusions.

Burnham (1979) has provided a general physical-chemical framework for hydrothermal activity attending high-level granitic intrusions. His model shows in great detail how H_2O saturated melt ---> crystals + vapor produces brittle deformation, dike intrusion, and chemical alteration of wall rock. This scheme is well demonstrated around porphyry-copper host plutons. Complexities of magmatic crystallization and second boiling vapor release often develop multiple stages of hydrothermal activity and metallogenesis (Wallace and others, 1978). There can be several comagmatic generations of cross-cutting dikes and fracture-breccia systems (Burnham, 1979, fig. 3.5b). In Burnham's illustrations (p. 112, 113) and the Wisconsin examples breccia zones attain a vertical orientation from hydrothermal expansion and hydrofracting with a horizontal σ_3 (minimum stress axis).

During crystallization of volatile-rich plutons vapor build up and separation may cause fracturing of the wall rock and saturation of the fracture system in a silica-rich fluid. In porphyry-copper deposits the stockwork of quartz veins and fractures are sites of extensive sulfide mineralization. Distinctive zones of metasomatic alteration are also characteristic of porphyry systems (Lowell and Guilbert, 1970).

The quartzite-intruding magmas in Wisconsin produced physical effects similar to porphyry-copper systems, but they were chemically distinct. Neither porphyry-type mineralization nor metasomatic zonation have yet been identified in Wisconsin. Evidence of any chlorine or sulfur enrichment in the magmatic hydrothermal systems affecting the quartzites is also lacking. Other than silica, only feldspar (both K and Na varieties), phyllosilicates, iron oxide, tourmaline, calcite, and traces of fluorite were hydrothermally precipitated from mobile components within definable country rock.

There may be reasonable doubt as to the feasibility of granitic magma assimilating quartzite at Baxter Hollow or Hamilton Mound. Silica-saturated melt should not be capable of dissolving siliceous wall rocks, especially in any appreciable volume. However, the corrosive action of a volatile-rich intrusion on potentially hydrous country-rock involves vapor-rich fluid separated from the host magma (as described by Burnham, 1979) and not simply the magma itself. Chemical and petrographic data from Baxter Hollow and Hamilton Mound also suggest that the original uncontaminated magma may have been more dioritic than after intrusion. Some plagioclase laths at Baxter Hollow are zoned having cores as calcic as An_{32} with albitic rims. Textures and the crystalli- zation sequence of feldspar and mafic minerals apparent at Baxter Hollow are consistent with those in the Denzer diorite exposed 4 km to the west (fig. 1). Both diorite and granite display grano- phyric-diabasic textures with euhedral plagioclase. Smith (1983) has shown chemically that the diorite is a reasonable source of the Baxter Hollow granite. Large blocks of Denzer-type diorite are included in the granite near quartzite contacts (fig. 14). Burnham (1979, p. 104) specified that a moderately hydrous dioritic magma could crystallize plagioclase and pyroxene (also some hornblende) and enable the remaining melt fraction to dissolve relatively large amounts of quartz and potassium feldspar. These two phases are progressively incorporated into the residual melt and vapor involved in dike, vein, and hybrid-rock formation. Metasomatic quartz and microcline hybrids occur above the more mafic intrusive rock at Hamilton Mound.

Although porphyry-type mineralization may be absent from Wisconsin, the mobility of iron, fluorine, and CO_2 in the hydrothermal zone at Hamilton Mound, Baxter Hollow, and Necedah may be consistent with certain other ore-forming environments (Hauck and Kimball, 1984). Specific examples are associated with anorogenic iron-rich dioritic to granitic intrusion.

Explosion breccia (Wright and Bowes, 1968) or breccia pipe (Norton and Cathles, 1973) are phenomena related to the violent action of a water-rich vapor phase exsolved from a cooling magmatic body. The mechanism and force of vapor separation were discussed above with reference to Burnham (1979). Wright and Bowes (1968) mentioned several types of magmatic systems, including high-level granite, which may involve the development of explosion breccia. The common factor in all cases is the presence of ample volatiles. Caledonian appinite stocks from Scotland are described by Wright and Bowes (Bowes and McArthur, 1976) as classic, breccia-forming systems. The appinites are moderately alkaline dioritic rock with both quartzand amphibole-rich varieties. These compositions are similar in many respects to the more mafic varients proposed as members of the Wisconsin anorogenic suite.

The occurrence of Scottish appinite strongly implies particular structural controls (Wright and Bowes, 1968; Bowes and McArthur, 1976). The appinitic plutons and accompanying breccia are commonly exposed in antiformal structures of thick quartzite roof rock. The quartzite served as a nearly impermeable cap over the cooling magmas. In Scotland the quartzite was folded before and after intrusion (Bowes and McArthur, 1976). Folding may have allowed volatiles to collect in the hinges of antiforms and also provided fractured zones for escape of trapped fluids. In Wisconsin, the structural role of some Baraboo interval quartzite appears analogous to the appinite-suite environment. The timing of Wisconsin quartzite deformation relative to intrusion is not yet well understood, but brecciated rocks do occur in the eroded axial regions of large anticlines at Baraboo, Hamilton Mound, and perhaps elsewhere. The limited vertical exposure of the plutons associated with Wisconsin breccia typically does not allow an appreciation of the explosiveness envisioned by Bowes and McArthur (1976). At the present level of exposure, only Necedah, McCaslin Mountain, Hamilton Mound, and Battle Point breccia exhibit actual igneous fragments. Only the first two of these could be considered true magmatic breccia with igneous matrix material.

An idealized cross section through brecciated, metasomatized quartzite is depicted in figure 15. Two or more of the features shown have been observed at each of the Wisconsin exposures described above. An oversimplified sequence would place these features in the following progression from pluton outward: intrusion breccia, metasomatism, dikes, quartz veins, and quartzite breccia. There is some spatial overlap, especially with metasomatism and igneous dikes.

One additional implication can be drawn from the exposures of quartzite discussed above. It may seem strange that most areas of Baraboo-type metasedimentary rock contain quartz veins and vein quartz breccia. It has been shown that these veined exposures also commonly cap intruding granitic or dioritic plutons. The apparent coincidence suggests that intrusion-related hydrothermal quartz contributed to the preservation of some quartzite whereas other deformed but unintruded parts of the sedimentary pile were more easily eroded.

CONCLUSIONS

There is a previously unrecognized association of anorogenic intrusion with brecciated and metasomatized quartzite deposited during the Baraboo interval in central and northern Wisconsin. Specific exposures and characteristic features are summarized in table 1. Apparently the Barron Quartzite in western Wisconsin was the only major area of outcrop unaffected by overt hydrothermal activity.

The breccia is typically composed of recrystallized orthoquartzite or metachert surrounded by white, vein-quartz. The quartz-vein networks often contain open pockets of quartz crystals with or without clay and hematite. Fluid inclusion analysis from one quartzite exposure indicates a vein-quartz origin near 106 °C. With an average geothermal gradient (stable craton) this temperature would exist at pressures that may be too high to allow open crystal-pockets of sedimentary or origin groundwater. This suggests a hydrothermal origin for the quartz veins and breccia.

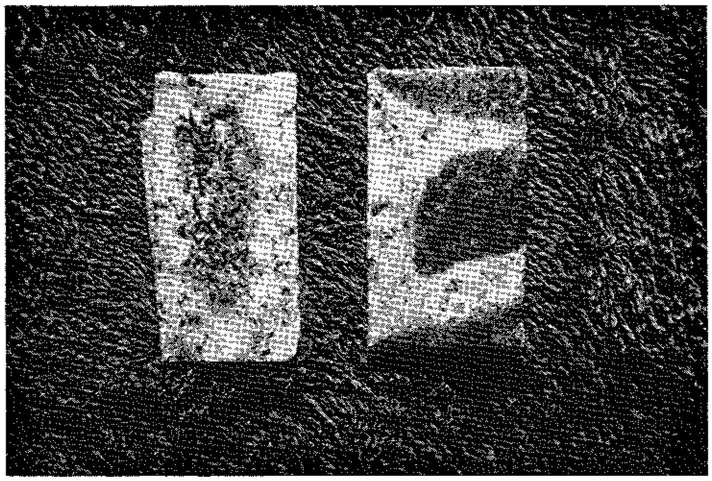

Figure 14.--Polished core samples of Denzer-type diorite inclusions in Baxter Hollow granite. Long dimensions about 12 cm.

Most of the brecciated exposures display complex metasomatic alteration of quartzite and intrusive rock. Hybrid-rock zones several metres wide span the contacts at Baxter Hollow and Hamilton Mound. Si, Al, Fe, K, and Na were the primary components mobilized in the hydrothermal systems. F, CO_2, and Ca were also present in metasomatizing fluids. The more extensive metasomatic changes described at Baxter Hollow, Hamilton Mound and Necedah were developed in conjunction with the anorogenic plutons of the older, 1,760 Ma, type. This may imply that the earlier magma was effectively wetter than the 1,500 Ma-old intrusions which generally produced less alteration. It is also possible that original water of deposition trapped in the Baraboo-type units contributed to alteration 1,760 Ma. Less water was probably available 1,500 Ma after deformation and thermal metamorphism.

Contorted bedding structures, including cross beds, appear in certain cases to be related to ductile deformation and recrystallization of metamorphosed quartzite. Similar structures had previously been attributed to synsedimentary processes, which may also be a correct interpretation in some exposures.

Two types of intrusion-country rock interaction have been evaluated for their similarity to the described features in Wisconsin. Each proves to be a reasonable model analogous to certain features. Both porphyry-copper host plutons and appinite intrusions produce well developed vein-breccia zones. Porphyry-copper deposits are also characterized by hydrothermal metasomatism. Folded quartzite roof rock appears to be especially important to the development of appinite brec- cias. Similarities to porphyry-copper magmatism and hydrothermal systems imply the possibility of economic mineralization in Wisconsin. Although sulfide enrichment has not been observed, the mo- bility of other volatiles and iron in magma intruding the quartzite is consistent with certain ore-forming environments.

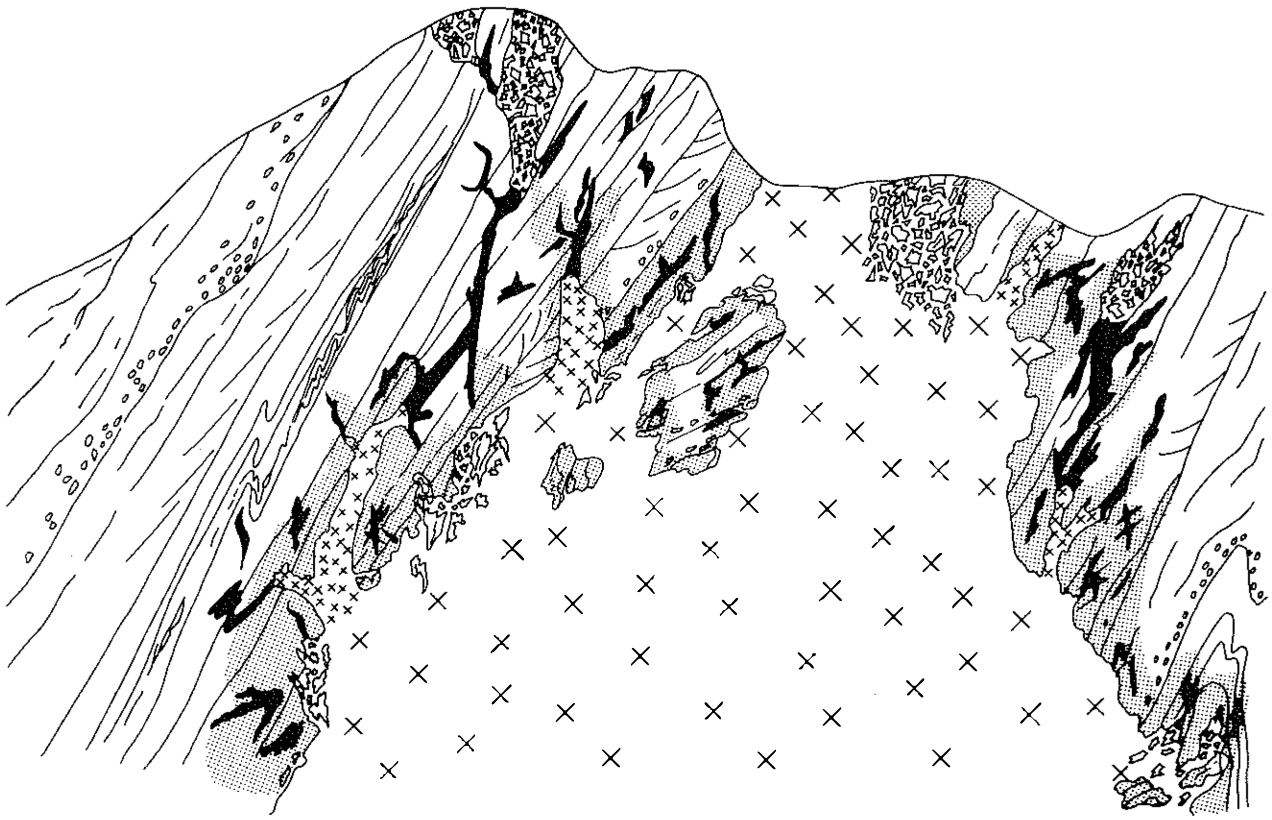

Figure 15.--Idealized cross section sketch representing the major features associated with intrusion of Baraboo-interval quartzite.

The coincidence of quartz-vein breccia with most Baraboo interval quartzite exposures is significant. Intrusions and silicification are considered to be at least partly responsible for outcrop preservation.

A final conclusion is that detailed investigations are needed to further study problems brought out in the present reconnaissance study. Metasomatism can be characterized by oxygen isotope and other geochemical analyses as well as by careful petrographic work. Trace element data would be particularly helpful to test models of hydrothermal processes and ore formation. Whenever possible, rocks intruding quartzite should be isotopically dated to provide adequate control for tectonic models.

Table 1. Features observed at various locations.

Wisconsin quartzite	Intrusion type (age)	Intrusion breccia	Quartz veins	Quartzite breccia	Hydrothermal metasomatic features	Other significant metamorphic phases
Baraboo	-	-	-	-	-	pyrophyllite
Baxter Hollow	granite-diorite (1,760 Ma?)	not observed	yes	hematite rich	sericitization, albite porphyroblasts-iron mobility, quartz-tourmaline	-
Rock Springs Area	subsurface?	-	abundant, bedding parallel near breccia	well developed	quartz-clay-hematite pockets	-
Highway 12 - South Bluff	subsurface granite	-	yes	-	clay-lined fractures	-
Necedah	granophyric granite (similar to 1,760 Ma) pegmatites, diorite(?) in subsurface	yes	bedding parallel and across bedding	yes	clay-muscovite pockets chlorite-muscovite lined fractures	-
Hamilton Mound	contaminated porphyritic granite,(1,760 Ma) foliated granodiorite, pegmatites containing tourmaline	not observed	abundant and varied, some include chlorite and calcite	contains granitic dikelets in quartzite	quartzite-intrusion hybrid rocks, Si mobility, quartz crystal clay lined fractures clasts	-
Waterloo	granite pegmatites (about 1,500 Ma) metabasalt(?) dike (1,430 Ma)	-	increase near breccias	yes	quartz crystal pockets feldspar grains in quartzite	andalusite chloritoid hornblende
McCaslin Mountain	granite and feldspar porphyry (about 1,500 Ma)	yes	grading into breccia	yes	feldspar grains in quartzite	sillimanite, andalusite, chloritoid, garnet
Rib Mountain	granite-syenite (1,500 Ma) altered mafic dike (?about 1,000 Ma?)	not observed	large and complex	(?) not observed	sericitic alteration, quartz-coarse grained muscovite pockets, anatase, brookite, tourmaline	sillimanite
Battle Point	?felsic porphyry inferred	?probably	yes	contains altered feldspar-rich clasts	sericite-clay-quartz pockets	-
Barron	basaltic dike (about 1000 Ma)	-	-	-	-	-
Others including Vesper	-	-	yes	yes, in various locations	quartz-clay pockets	-

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After the manuscript has been reviewed by an associate editor and reviewers (specialists in the field addressed by the paper) a final decision on acceptance or rejection will be made by the editor. The manuscript, if accepted, will be returned to the author for modification, based on the recommendations of the associate editor and reviewers, and for preparation of the final copy. The final manuscript will then be submitted to the editor who will see the manuscript through its publication stages.

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