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

by

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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.

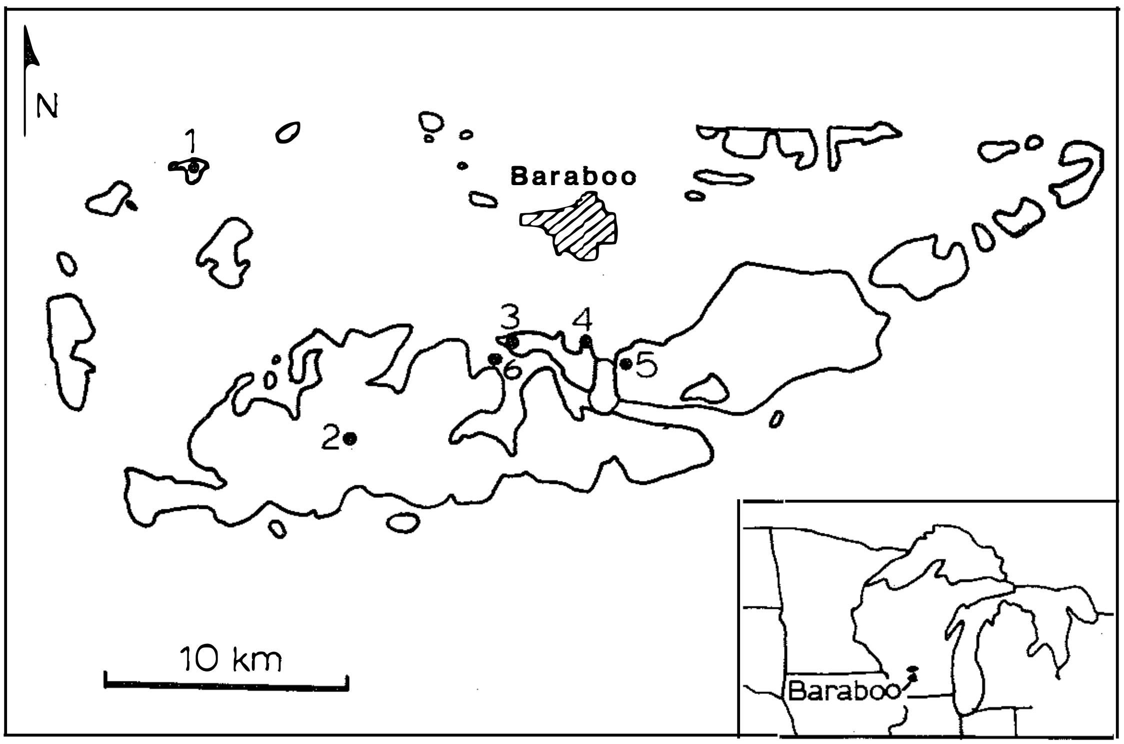

Figure 1.--Map showing distribution of Baraboo Quartzite in south-central Wisconsin. Numbers refer to phyllite zone locations: (1), Abelman's Gorge; (2), Happy Hill; (3), Skillet Creek; (4), northwest park entrance; (5), northeast park entrance; and (6), Highway 12.

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).

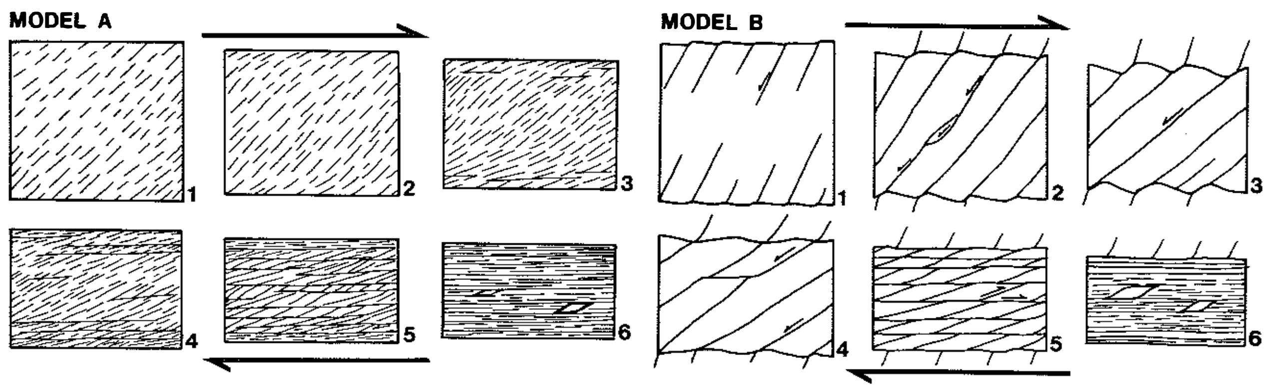

Figure 2.--(a) Six stages in the evolution of foliation in Model A based on a modified version of the model of Berthe and others (1979). In this example the shear zone boundaries are parallel to the top and bottom borders of each frame and the shear sense is right-lateral. (1) An early stage of shear 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.

Second-generation Structures

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.

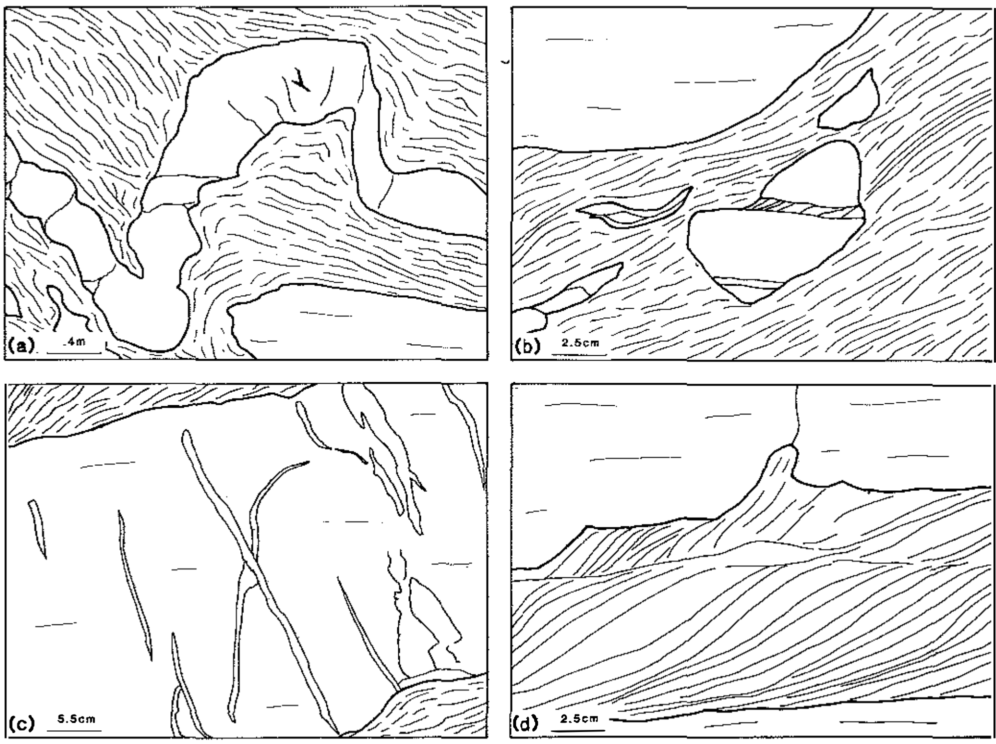

Figure 3.--Structures in Baraboo phyllite zones indicative of the first-generation, updip, simple shear. In all frames phyllite is represented by lined pattern (the orientation and spacing of the lines corresponds to the orientation and spacing of the foliation), and quartzite is represented by the clear 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.

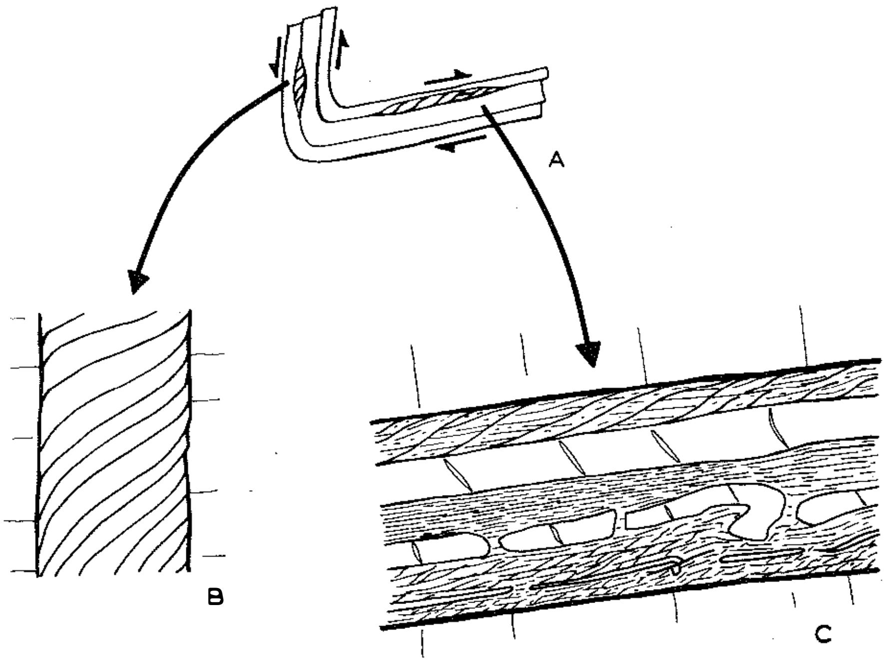

Figure 4.--Schematic diagram to illustrate how first-generation structures were formed by updip, flexural shear within the Baraboo Syncline. (b) Nature and orientation of structures on north limb (for example at Van Hise Rock). (c) Nature and orientation of structures on south limb (for example at Skillet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

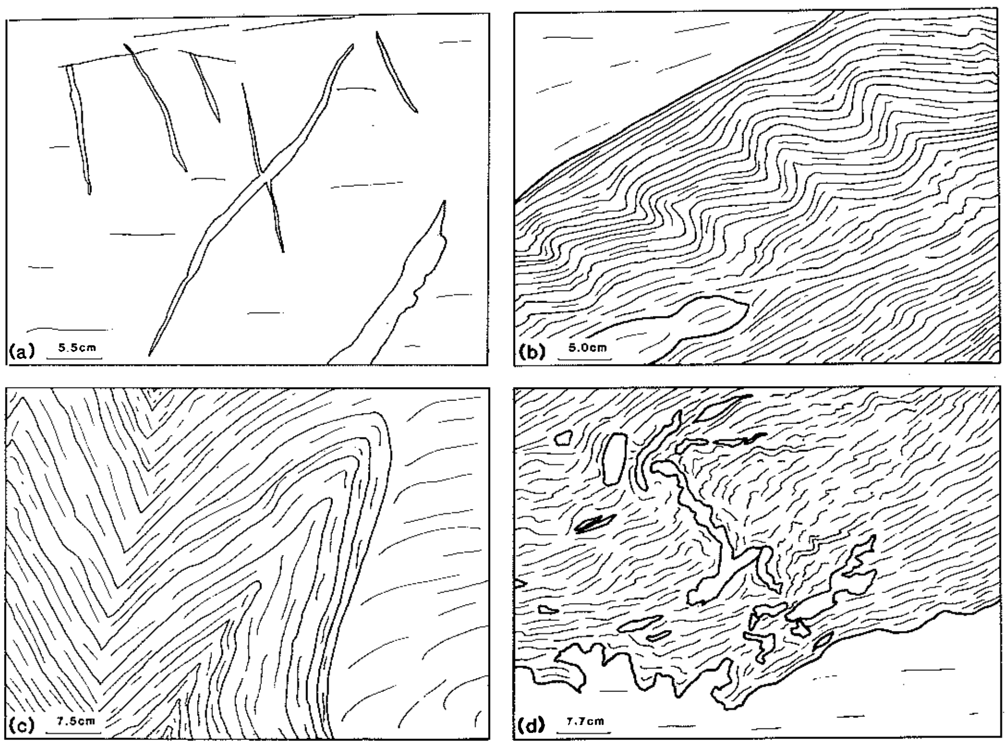

Figure 5.--Structures in Baraboo phyllite zones indicative of the second-generation downdip simple shear. (a) A second-generation tension gash that has offset a first-generation tension gash at Skillet Creek. The down dip second-generation shear is represented by a sinistrel shear couple in this exposure. (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-

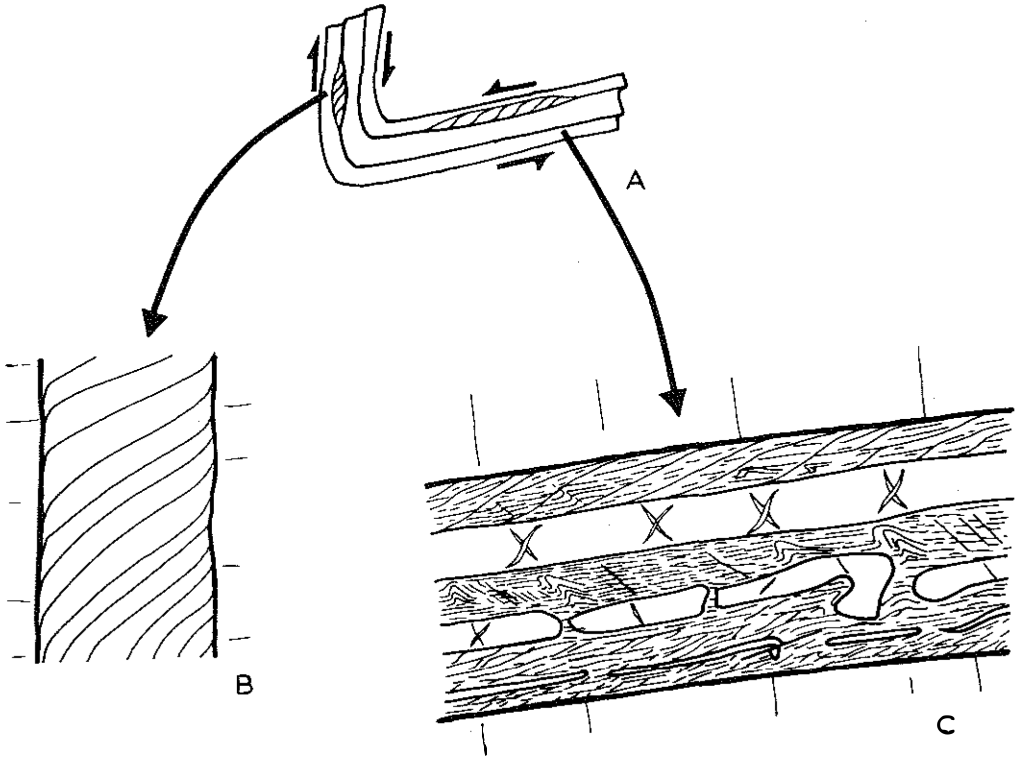

Figure 6.--(a) Schematic diagram to illustrate how second-generation structures were formed by down-dip shear within the Baraboo Syncline. (b) This shear zone consists of phyllitic quartzite. It possesses sufficient strength to resist deformation during the second phase of deformation. (c) This shear 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 ± 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.

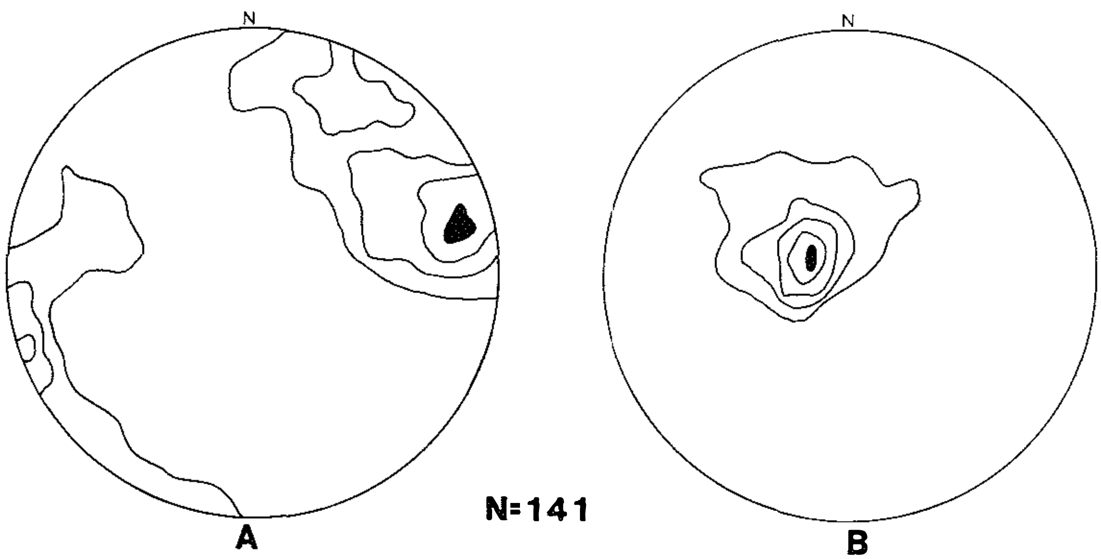

Figure 7.--Stereograms summarizing second-generation, mesostructural data from phyllite zones. Contour intervals are 5%, 3%, 2%, 1%, and ½%. (a) Poles to planar mesostructures, for example crenulation cleavage, kink bands and axial surfaces of asymmetric folds. (b) Hinge lines of crenulation cleavage,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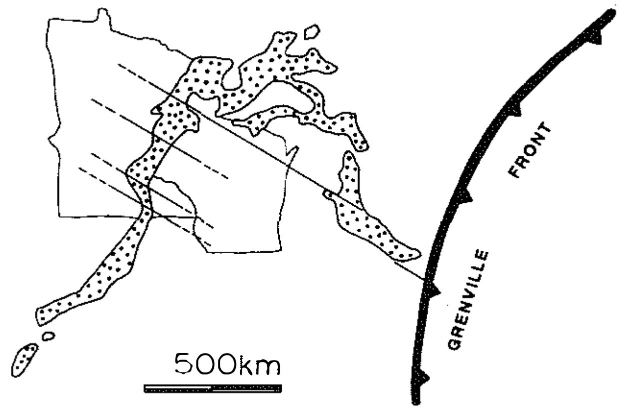

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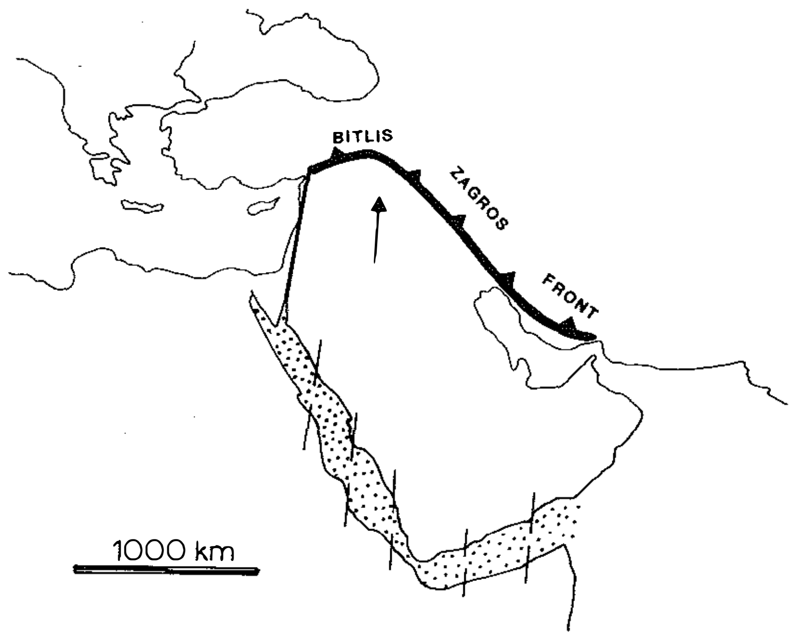

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Figure 8.--Schematic, tectonic maps illustrating large rifts developed in the forelands of convergent zones. A. An interpretation of the relationship between the Keweenawan Rift (dot pattern) and synchronous, Grenville continent-continent collision, 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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