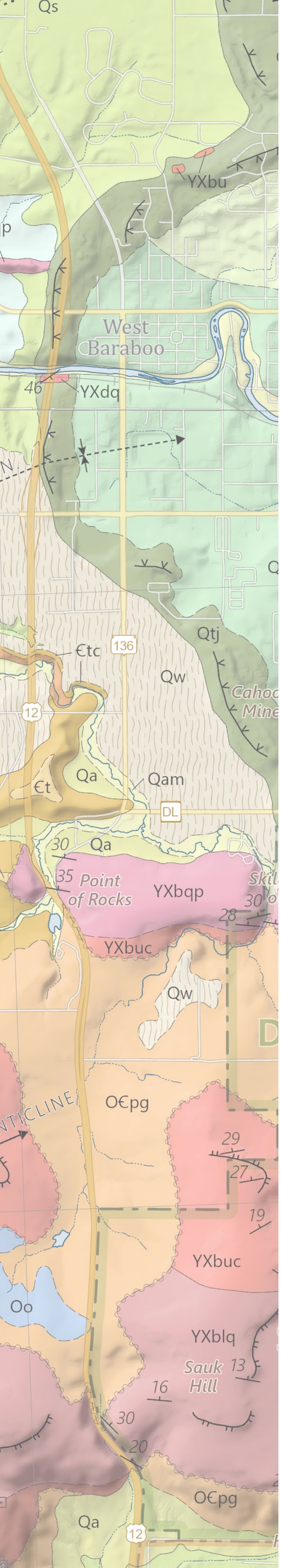
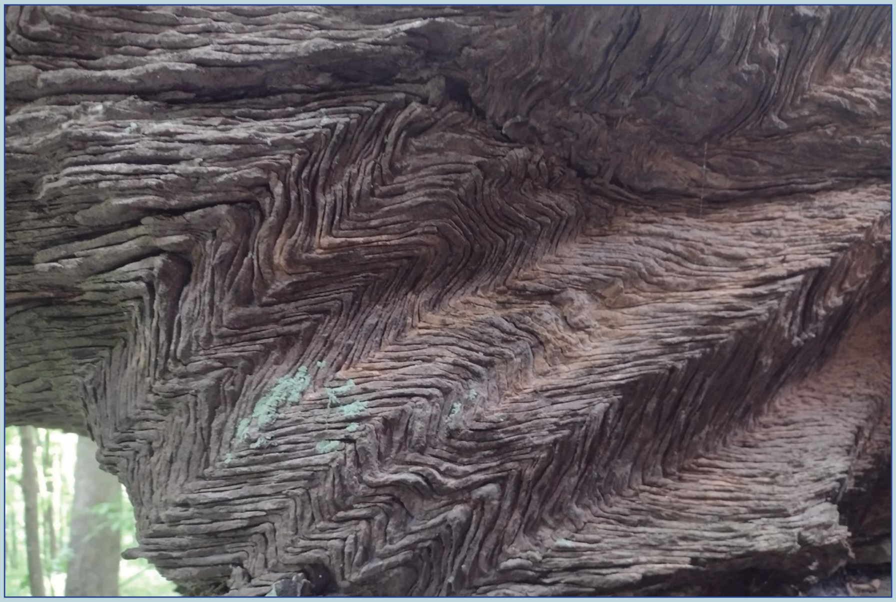


Geology of the Baraboo Hills, Sauk and Columbia Counties, Wisconsin



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Photos

Front cover: Kink folds in the Baraboo Quartzite

Back cover: Outcrop of the Baraboo Quartzite in Leopold Memorial Woods, North Freedom, WI

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Supplemental material

The following materials are available for download at: https://doi.org/10.54915/jkkg8792.

Plate 1: Geologic map
A map (.pdf format) of the geology of the Baraboo Hills.

Dataset 1: GIS data
GeMS Level 3 geodatabase (.gdb) and GeoPackpage (.gpkg file) that include map units and contacts, faults, folds, structural measurements, outcrop locations, core and drill log locations, Quaternary features, and cross section lines.

Abstract

New geologic mapping done at the 1:50,000-scale in the Baraboo Hills of Sauk and Columbia counties, Wisconsin, provides updated baseline information on the distribution of Precambrian, Paleozoic, and Quaternary geologic materials in the area. Paleoproterozoic rhyolite, granite, and diorite, and Paleoproterozoic to potentially Mesoproterozoic quartzite are the oldest rocks in the area. In the Cambrian and Ordovician periods of the Paleozoic, rising sea levels resulted in the deposition of a series of mostly siliciclastic sedimentary units unconformably over Precambrian units along the flanks and interior of the Baraboo Hills. The Paleozoic-Precambrian contact is part of the global Great Unconformity, a major gap in the geologic record prior to the rapid diversification of life during the Cambrian Period. During the Quaternary, the Green Bay Lobe of the Laurentide Ice Sheet reached its maximum extent in central Wisconsin approximately 24,000 years ago, covering the eastern Baraboo Hills. To the east of the hills, the ice sheet deposited till and outwash as well as constructed moraines and drumlins. West of the maximum extent of the ice sheet, glacial lake sediment and loess were deposited.

New and recent geologic mapping has focused on the Precambrian rocks of the Baraboo Hills, and has resulted in a revision of the Precambrian stratigraphic section. The Baraboo Quartzite is now divided into four informal members: a lower conglomerate and quartzite member, a lower quartzite member, an upper conglomerate and quartzite member, and an upper quartzite and phyllite member. Overlying the Baraboo Quartzite is the Seeley Slate, Freedom Formation, Dake Quartzite, and Rowley Creek Slate. The base of the Dake Quartzite forms an important angular unconformity that represents an unknown amount of missing time. The maximum depositional age for the Baraboo Quartzite is 1714±17 Ma and the maximum depositional age for the Dake Quartzite is 1630±9 Ma. Regional mapping constraints combined with detrital zircon populations indicate the Dake Quartzite/Rowley Creek Slate package is correlative with the base of the Waterloo Quartzite. The combined age and map constraints indicate deposition of the Baraboo Quartzite may have pre-dated the ca. 1600 Ma Mazatzal orogeny.

Stratigraphic revisions have provided marker horizons that have helped constrain the Precambrian structural history. The Baraboo syncline and associated smaller folds were formed in a protracted southeast-vergent fold-thrust belt. The Baraboo Quartzite, Seeley Slate, and Freedom Formation were initially deformed into a series of folds with axes that trended east to northeast. After deposition of the Dake Quartzite and Rowley Creek Slate, a second pulse of shortening tightened the early folds and deformed the younger Dake Quartzite and Rowley Creek Slate. Mapping has identified at least three thrust faults in the Baraboo Hills, and the construction of restorable cross sections suggests the folds are genetically related to the faults. Recrystallization and cooling ages from Precambrian rocks across the region include one set of ages approximately 1600 Ma (Mazatzal orogeny) and a second set of ages between 1493 and 1411 Ma (Baraboo/Picuris orogeny). Thus, the Precambrian history of the Baraboo Hills contains evidence for a complicated and prolonged history of sedimentation and deformation. Together with mapping constraints, these suggest deformation and sedimentation may have spanned both the Mazatzal and Baraboo/Picuris orogenies.

Introduction

The Baraboo Hills are located in Sauk and Columbia counties, south-central Wisconsin. The hills are cored by Precambrian quartzite, rhyolite, and granite, and are partially mantled by Paleozoic sedimentary rock and Quaternary sediment. They rise abruptly about 150 to 200 m above the surrounding plains. The unique Precambrian, Paleozoic, and Quaternary history of the area, and the picturesque setting, draw geology classes and tourists from across the United States.

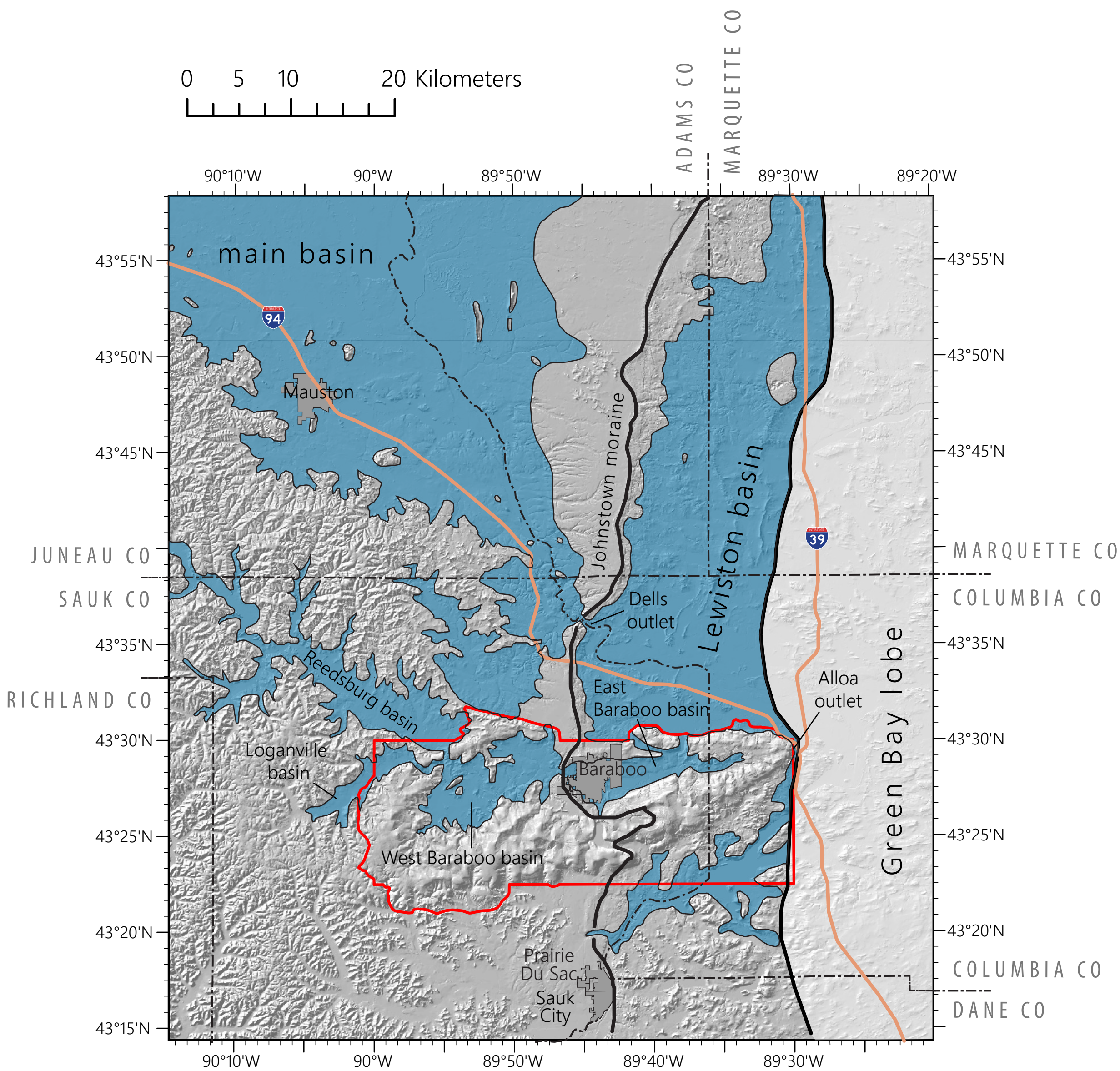
Precambrian quartzite, rhyolite, and granite are the oldest rocks in the Baraboo Hills area. Igneous rocks were intruded and extruded approximately 1750 Ma (Van Wyck, 1995). These rocks were subsequently eroded so that they serve as the basement for deposition of subsequent Precambrian sedimentary strata. Supermature sandstones (now quartzites) were deposited between 1750 Ma and 1450 Ma. Dott (1983) named this period of supermature deposition in the Baraboo area and across the upper midwest the Baraboo-interval. The Baraboo Quartzite, Waterloo Quartzite, Sioux Quartzite, and Barron Quartzite are four of the thicker preserved sections making up the Baraboo-interval. Later detrital zircon dating narrowed the depositional age of the Baraboo Quartzite to less than 1714±17 Ma (Stewart and others, 2021a). Precambrian slate, iron formation, meta-carbonate, and more quartzite and slate occur stratigraphically above the Baraboo Quartzite in the Baraboo Hills. The iron formation was mined for iron ore in the early 1900s at the Sauk, Illinois, and Cahoon mines in the interior of the Baraboo Hills. During the two proposed contractional events described in the Discussion section, deformation recrystallized the sedimentary units and rhyolite, resulting in the development of weathering resistant, positive relief units on the landscape beginning in the Precambrian.

Paleozoic sedimentary rocks were deposited unconformably over Precambrian rocks in the Baraboo Hills area. The contact between the base of the Paleozoic section and the eroded surface of the Precambrian is widely known as the Great Unconformity. In the Baraboo area, it represents a time gap of over 1 billion years (Marshak and others, 2016). The surface of the Great Unconformity has significant topographic relief and represents a time transgressive surface spanning many millions of years. Eroded remnants of the Paleozoic section are found throughout the Baraboo Hills. Most of the sedimentary section was deposited during the worldwide Cambrian to Ordovician Sauk transgression (Sloss and others, 1949), which was named after Sauk County, Wisconsin. During the Cambrian, seawater inundated lowlands surrounding the Baraboo Hills, leaving the Precambrian-cored uplands exposed as a chain of islands within a tropical sea (Dalziel and Dott, 1970). Impressive boulder conglomerates were deposited by storms that battered these islands (Dott and Byers, 2016). During Ordovician time, the oceans carved a terrace into the quartzite at a modern elevation of approximately 395 m, which is still visible today (Thwaites, 1935). Eventually, the Precambrian hills were entirely buried by Paleozoic sediment. After burial in the Paleozoic, Precambrian units were re-exhumed in the Mesozoic or Cenozoic (Clayton and Attig, 1990) and still stand as topographically high hills and ridges.

In the Quaternary, the Baraboo Hills straddled the western margin of the Green Bay Lobe of the Laurentide Ice Sheet. The effects of glaciation are impressive and can be seen today across the area. The Johnstown moraine is the terminal moraine of the Green Bay Lobe of the Laurentide Ice Sheet, and it crosses the eastern part of the Baraboo Hills. Cosmogenic exposure ages from glacially transported boulders within or near the Johnstown moraine varied from 22,100 to 19,400 years ago, providing an age for the onset of glacial retreat in the area (Ullman and others, 2015). Glacial Lake Wisconsin developed west of the Johnstown moraine, when the ice sheet dammed outlet drainages. The west Baraboo Basin, a subbasin of Glacial Lake Wisconsin (fig. 1), filled much of the interior of the Baraboo Hills with water (Clayton and Attig, 1990). Smaller ice-marginal lakes formed high in the hills (43°24'0"N, 89°42'12"W), and dating of the sediment in these lakes found the Green Bay Lobe did not begin to recede from the area until 18,500 years ago (Attig and others, 2011; Carson and others, 2020), similar to ^{10}Be ages from glacial boulders (Ullman and others, 2015). When the Green Bay Lobe receded past the eastern edge of the Baraboo Hills at the Alloa outlet (fig. 1), Glacial Lake Wisconsin catastrophically drained, carving the Wisconsin Dells and depositing a delta with foresets reaching 12 m in height (Bretz, 1950; Clayton and Attig, 1989; Clayton and Knox, 2008). Today, outwash deposits and glacial lake sediment fill low-lying areas in the western Baraboo Hills. Till deposits blanket much of the eastern Baraboo Hills. The remains of rock glaciers can be seen on south-facing slopes of the South Range (Stewart and Stewart, 2021). East of the Baraboo Hills, abundant kettle lakes mark locations where stagnant ice was left behind as the glacier receded.

This map (see plate 1) compiles new and recent 1:24,000-scale geologic mapping at 1:50,000-scale. It culminates a long-term Wisconsin Geological and Natural History Survey (WGNHS) mapping objective of refining and revising the Precambrian Baraboo-interval stratigraphy. This revised stratigraphy is used to reinterpret the structural history. Precambrian rocks in the Baraboo Hills were long thought to have been deformed in

Figure 1. Map showing the greatest extent of Glacial Lake Wisconsin in blue. Modified from Clayton and Attig (1990). As noted by Clayton and Attig (1990), the lake did not occupy the entire area at any given time. Area of plate 1 is highlighted in red. Interstate roads shown in orange. Select cities shown in gray. Base map: Hillshade derived from U.S. Geological Survey National Elevation Dataset, 2017.



Wisconsin Transverse Mercator projection, 1991 Adjustment to the North American Datum of 1983 (NAD 83/91); EPSG 3071.

a fold-thrust belt (e.g. Dalziel and Dott, 1970), however, no map-scale thrust faults had been mapped prior to the 21st century. As a result of the new mapping herein, we present three stratigraphic advances that have allowed a re-interpretation of the map-scale structural history of the Baraboo Hills and surrounding areas. The first advance is the re-establishment (following the original work of Leith, 1935) of the Dake Quartzite and Rowley Creek Slate at the top of the Baraboo-area Precambrian stratigraphic section (Stewart and others, 2018). The work of Leith (1935) had been progressively questioned over several decades. Initially, select Dake outcrops were remapped as Baraboo Quartzite (Schmidt, 1951; Dalziel and Dott, 1970). Later, workers believed there was not enough evidence to determine whether the Dake and Rowley Creek existed or not (Brown *in* Clayton and Attig, 1990). Recently, many authors have focused on other Precambrian problems and simply left these units off their stratigraphic columns (Medaris and others, 2011; Bjørnerud, 2016; Marshak and others, 2016). The second stratigraphic advance is the identification and mapping of marker horizons within the Baraboo Quartzite. At the onset of this project, we hypothesized that marker horizons could be identified and mapped, and faults could be mapped based on offset of these marker horizons. Initial work focused on defining and subdividing the stratigraphic section to provide marker horizons from which to interpret structure. Stewart and Stewart (2020) and Stewart and others (2021a) built on the ideas of Henry (1975) and subdivided the Baraboo Quartzite into four informal members based on grain size, lithology, and sedimentary lithofacies. Finally, in addition to refinement of the Precambrian stratigraphy within the Baraboo Hills, we present regional stratigraphic correlations between the Waterloo Quartzite of southeast Wisconsin, and the Dake Quartzite and Rowley Creek Slate of the Baraboo Hills.

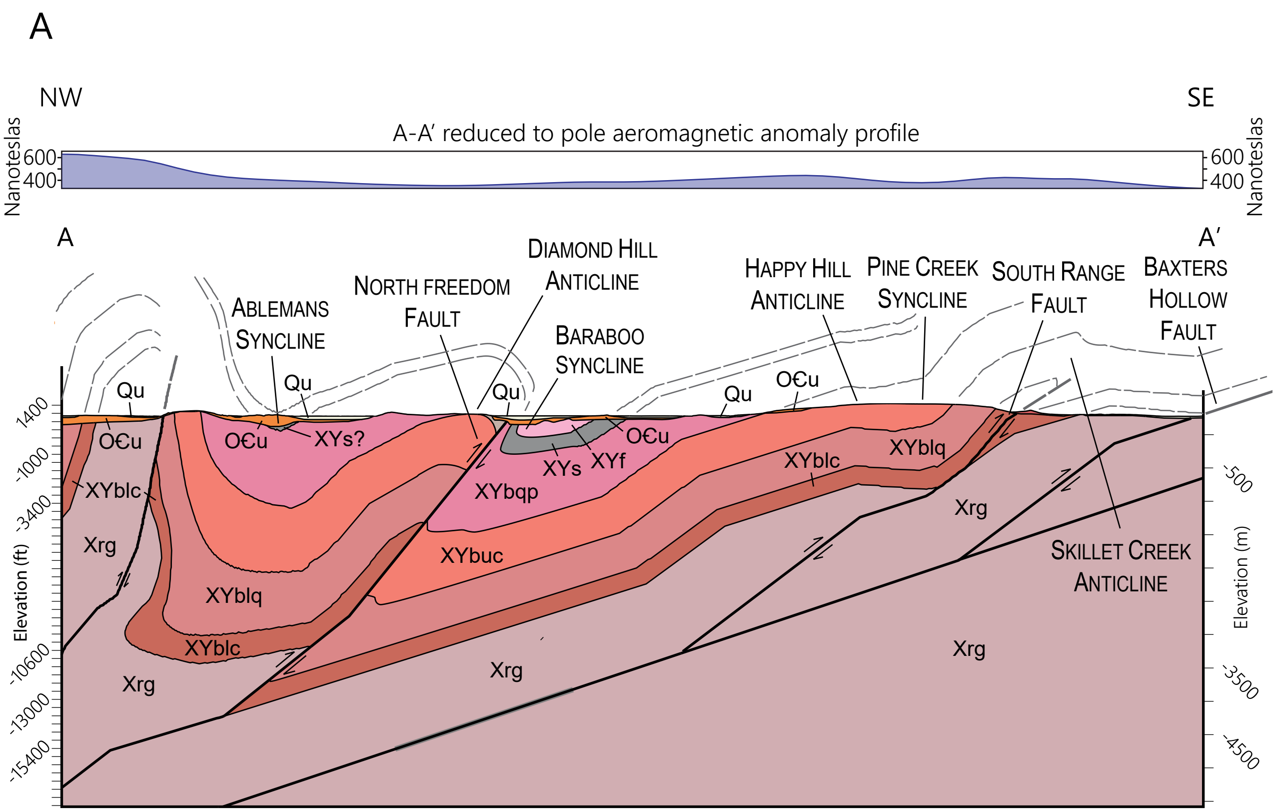
The following report focuses on Precambrian geology. It contains a Cross Section Methods section, a Description of Maps Units section, a Map Structures section, and a Discussion section. The Description of Map Units describes the revised Baraboo Quartzite stratigraphy, as well as all previously defined Quaternary, Paleozoic, and Precambrian units in the Baraboo Hills shown on plate 1. The Map Structures section utilizes the revised Precambrian stratigraphy to map new Precambrian faults, provides an overview of folding, and describes planar and linear deformational fabrics in the Baraboo Quartzite. The Discussion section focuses on revisions to the Precambrian history using the results of the new mapping. We first review depositional models for the Baraboo-interval and use the revised stratigraphy to constrain the timing of deposition. The revised depositional constraints provide a framework for reassessing the timing and tectonic setting of deformation, which are presented at the end of the Discussion section. Readers interested in the details of the Quaternary history of the Baraboo Hills are referred Clayton and Attig (1989), Clayton and Attig (1990), Syverson and Colgan (2011), Davis (2016) and Carson and others (2019). Those interested in the details of the Paleozoic history of the Baraboo Hills are referred to Clayton and Attig (1990) and Dott and Byers (2016).

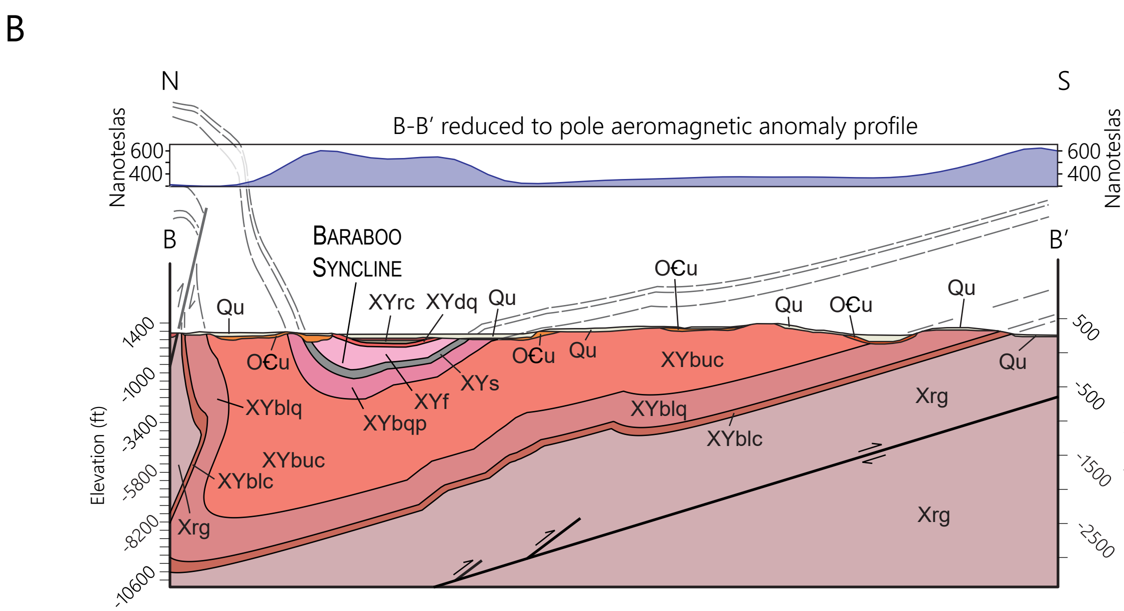
Cross section methods

Cross sections in figure 2 are close approximations of the surface geology, and are modified from plate 1. They were constructed as forward models from an originally undeformed section using FaultFold v.7.1.2, a program based on the work of Allmendinger (1998) and Zehnder and Allmendinger (2000). All Precambrian units were forward modeled except the Dake Quartzite and Rowley Creek Slate. FaultFold uses the trishear kinematic model for fault propagation (Erslev, 1991), which uses a triangular shaped shear zone in advance of the tip line of the fault to create a zone of distributed shear and deformation. Distributed shear is effective at modeling footwall synclines, which is the likely setting for the Baraboo syncline (e.g. Marshak and others, 2016). Trishear produces folds with thickened hinges and thinned limbs. It also assumes deformation of incompressible solids and assumes constant slip along faults. However, actual faults can stick, slip, or aseismically creep along different segments of the same fault (e.g. Byerlee and Brace, 1969; Beeler and others, 2001).

Kinematic models such as trishear are generally applied to well stratified sedimentary rocks that produce predictable flats and ramps. However, the cross sections shown on figure 2 show a basal thrust deep within igneous basement. The placement of the basal thrust within the basement may be justified because most of the Baraboo-interval is underlain by bedded volcanic rocks, reaching at least 1 km in thickness in the North Range (plate 1). We place the basal thrust just over 1 km beneath the Baraboo-interval based on the thickness of the volcanic package in the North Range, and assume the bedded volcanic units behaved similarly to the bedded sedimentary units during deformation. Two

Figure 2. Panel A, cross section across the western Baraboo Hills. Panel B, cross section across the eastern Baraboo Hills. See figure 3 for locations of cross section lines.





intrusive bodies, the Baxters Hollow granite and the Denzer diorite, outcrop in the western South Range, and these are also placed within the thrusted package. The extent of intrusive versus extrusive igneous rock in the southwestern Baraboo Hills is unclear. Though less common, the trishear kinematic model has been successfully applied to intrusive and metamorphic basement in a number of studies (Erslev and Rodgers, 1993; Giambiagi and others, 2009; Sánchez and others, 2018; others).

The sections in figure 2 are restorable but are unlikely to be balanced. Fold axis parallel stretching occurred in quartzites during early folding (Czeck and Ormand, 2007; also see Precambrian structure section below), which indicates deformation was not plane strain, a requirement for balanced cross sections. Hence the sections probably underestimate the original thickness of the Baraboo-interval section due to an unknown amount of stretching into and out of the plane of cross section. Nevertheless, these sections can still provide first-order controls on the architecture of the thrust belt in the Baraboo Hills. The cross section shown on figure 3 was not modeled using FaultFold, and as a result does not show thickening of the hinge and thinning of the fold limbs.

The figure 3 section is not restorable. It is included because exploration cores drilled during iron ore mining in the Baraboo area on the fold limbs suggest an angular relationship between the Dake Quartzite and the underlying Freedom Formation.

Aeromagnetic profiles along both lines of section are also shown on figure 2. Aeromagnetic data is from Daniels and Snyder (2002). The dataset was modified using a reduced-to-pole transform (E. Anderson, written comm., 2015).

Description of Map Units

The following unit descriptions are described from oldest to youngest.

Paleoproterozoic

Paleoproterozoic igneous rocks in the Baraboo Hills area include both intrusive and extrusive rocks. Isolated outcrops occur along the northern and southern margins of the Baraboo Hills. The WGNHS core repository includes 1 core of the Denzer diorite and 9 cores of either rhyolite or the Baxters Hollow granite. The latter were drilled less than a km east of Baxters Hollow (ASDC cores, fig. 4).

Rhyolite

Map unit **Xr**. Rhyolite crops out on both the northern and southern limbs of the Baraboo syncline. Volcanic rocks at Caledonia Church are mostly weakly flow foliated rhyolite, with subordinate welded crystal tuff and volcanic breccia (Medaris and others, 2011). Cleavages are not well developed (LaBerge and others, 1991), but thin cm-scale cataclastic bands crosscut parts of the rhyolite. Plagioclase phenocrysts have been altered to albite and calcite (Medaris and others, 2011). Near the Lower Narrows of the Baraboo River, gray to brown porphyritic rhyolite is interbedded with welded tuff containing possible fiamme. At the Lower Narrows, flow banding in the rhyolite parallels nearby bedding in the Baraboo Quartzite. Van Wyck (1995) using the U-Pb method on zircon found an age of 1754±44 Ma for rhyolite close to the contact with the Baraboo quartzite on the north limb of the Baraboo syncline.

Figure 3. **Cross section modified from Stewart and others (2024). Drill logs and core along each line of section are shown in gray. Section C-C′ shows the angular discordance between the Dake Quartzite/Rowley Creek Slate (XYdr) and underlying units. The thin, 65 m thickness of the Dake Quartzite combined with geologic logs provides control on the angular discordance. Unit abbreviations include undivided Quaternary and Paleozoic units (Qu, O€u), Freedom Formation (XYf), Seeley Slate (XYs), Baraboo Quartzite (XYbqp, XYbuc, and XYblq). This smaller cross section was not drawn using FaultFold.**

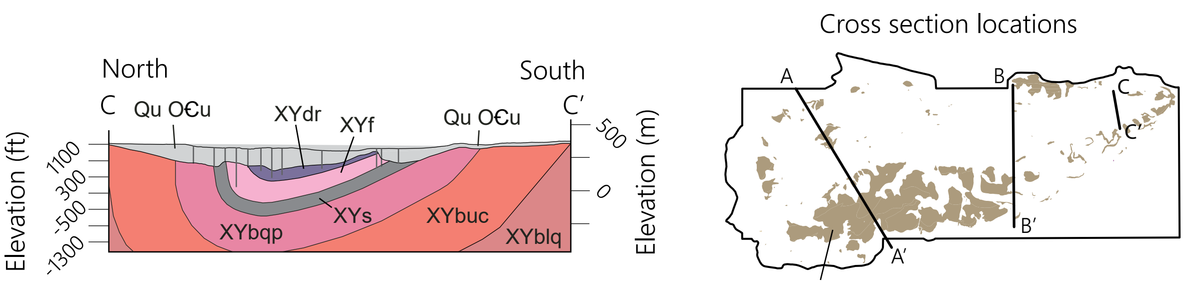
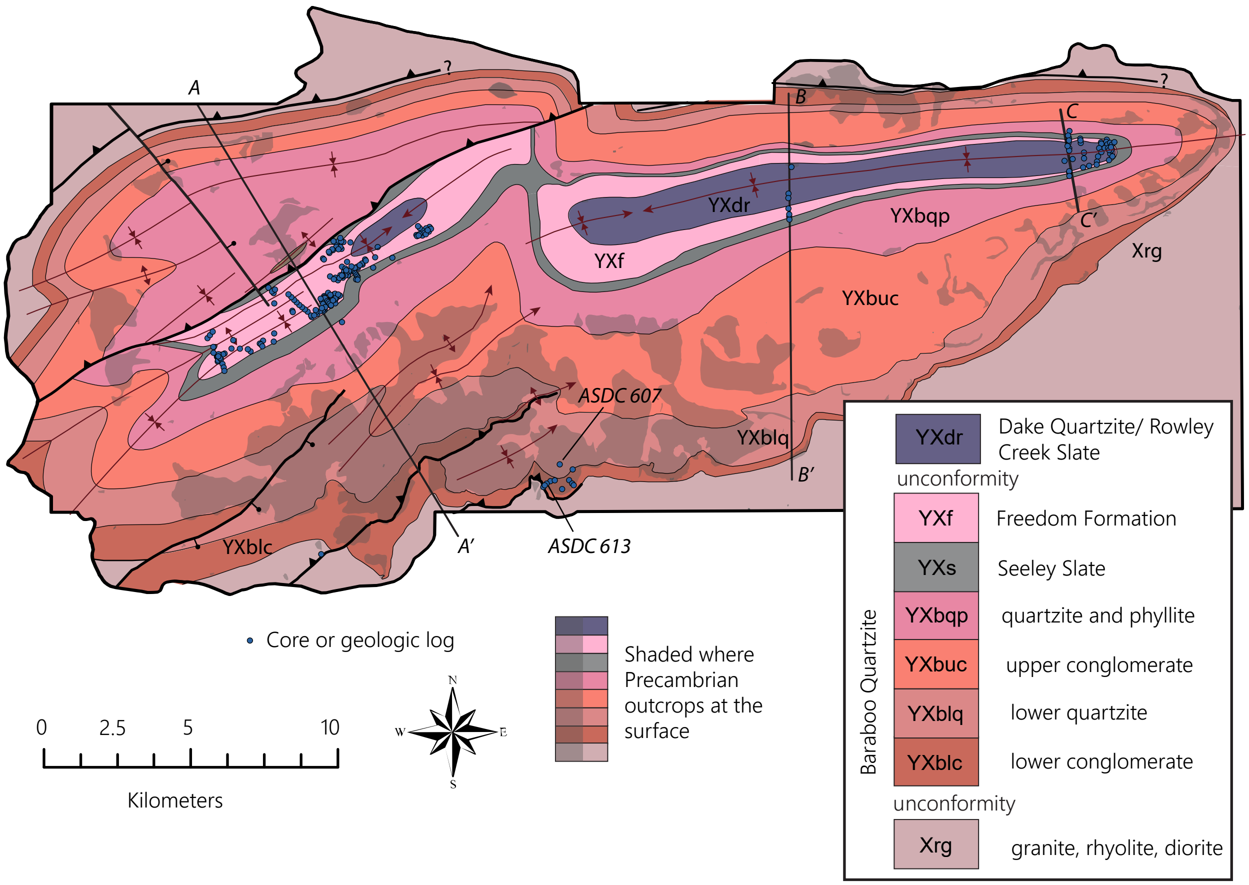


Figure 4. Precambrian geology map of the Baraboo Hills area. The map shows areas of outcrop exposure (shaded gray) and drill core and geologic logs. The location and ID of cores and wells referenced in this figure are available in accompanying dataset 1.



Views of the Lower Narrows of the Baraboo River

Baxters Hollow granite

Map unit **Xg.** The Baxters Hollow granite is a pink, fine-grained granite composed of 29% quartz, 66% feldspar, and 4.6% biotite and chlorite (Smith, 1978). The granite has been recrystallized to a greenschist facies mineral assemblage that includes quartz, epidote, chlorite, microcline, and albite (Medaris and others, 2011). It contains high CaO and relatively low Rb/Sr, but is related chemically to other calc-alkaline granites and rhyolites that make up the Montello batholith and extrusive equivalents (Smith, 1978). Smith (1978) report the granite intrudes into related rhyolite. The top of the granite immediately below the contact with the Baraboo Quartzite was weathered into a well-developed paleosol. The weathered profile includes a < 0.67 m thick regolith and a > 6.1 m thick saprolite (Medaris and others, 2003; Driese and Medaris, 2008). The granite has been dated using the U-Pb method on zircon to 1752±15 Ma (Van Wyck, 1995).

Denzer diorite

Map unit **Xdd.** The Denzer diorite is a black and white, medium grained diorite that originally contained approximately 70–85% plagioclase, 10–20% hornblende, and lesser quartz, biotite, apatite and iron oxides (Dalziel and Dott, 1970). However, the original mineralogy has been recrystallized to a greenschist facies metamorphic assemblage that includes chlorite, cummingtonite, actinolite, albite, epidote, and microcline (Medaris, 2001). Naymark and others (2001) report Ar^{40}/Ar^{39} ages of 1746±12 Ma Ar^{40}/Ar^{39} for biotite, and 1596±16 Ma and 1427±15 Ma for hornblende partially recrystallized to actinolite, cummingtonite, and chlorite.

Paleoproterozoic to Mesoproterozoic

Paleoproterozoic to Mesoproterozoic unit descriptions are based on various outcrop locations and drill core. The WGNHS has 45 cores archived at its core repository in Mount Horeb, Wisc., which include portions of the Baraboo-interval section. The drilling locations of historic drill core from the early 20th century that contain the Seeley Slate, Freedom Formation, Dake Quartzite, and Rowley Creek Slate are given in Stewart and others (2018). The WGNHS also has paper records of geologic logs and location maps for an additional 204 exploration holes from the interior of the Baraboo Hills, though no core is preserved from these holes. Core and core log locations are shown on figure 4. The Dake Quartzite description is based on field exposures at Dake Ridge (43°28'53"N, 89°41'15"W), North Freedom (43°27'23"N, 89°52'7"W), and West Baraboo (43°28'12"N, 89°46'31"W). The Freedom Formation and Seeley Slate descriptions are based entirely on drill core. Descriptions of the Baraboo Quartzite are largely derived from field exposures.

Baraboo Quartzite

The Baraboo Quartzite was broken into four informal members based on grain size and sedimentary lithofacies packages. Lithofacies were defined based on sedimentary structures, grain size, bed geometry, and grain sorting (Stewart and others, 2021a). Outcrops were visited across the Baraboo Hills (see dataset 1 for locations), and members were broken out and mapped based on map-scale patterns. These patterns were consistent across the North and South Ranges of the Baraboo Hills. The nearly full Baraboo Quartzite section can be viewed by walking north-northwest from Pine Hollow (43°23'16"N, 89°45'9"W) to Sauk Hill, and continuing north to the Skillet Creek Campground (see plate 1). The maximum depositional age for the Baraboo Quartzite is 1714±17 Ma (Stewart and others, 2021a).

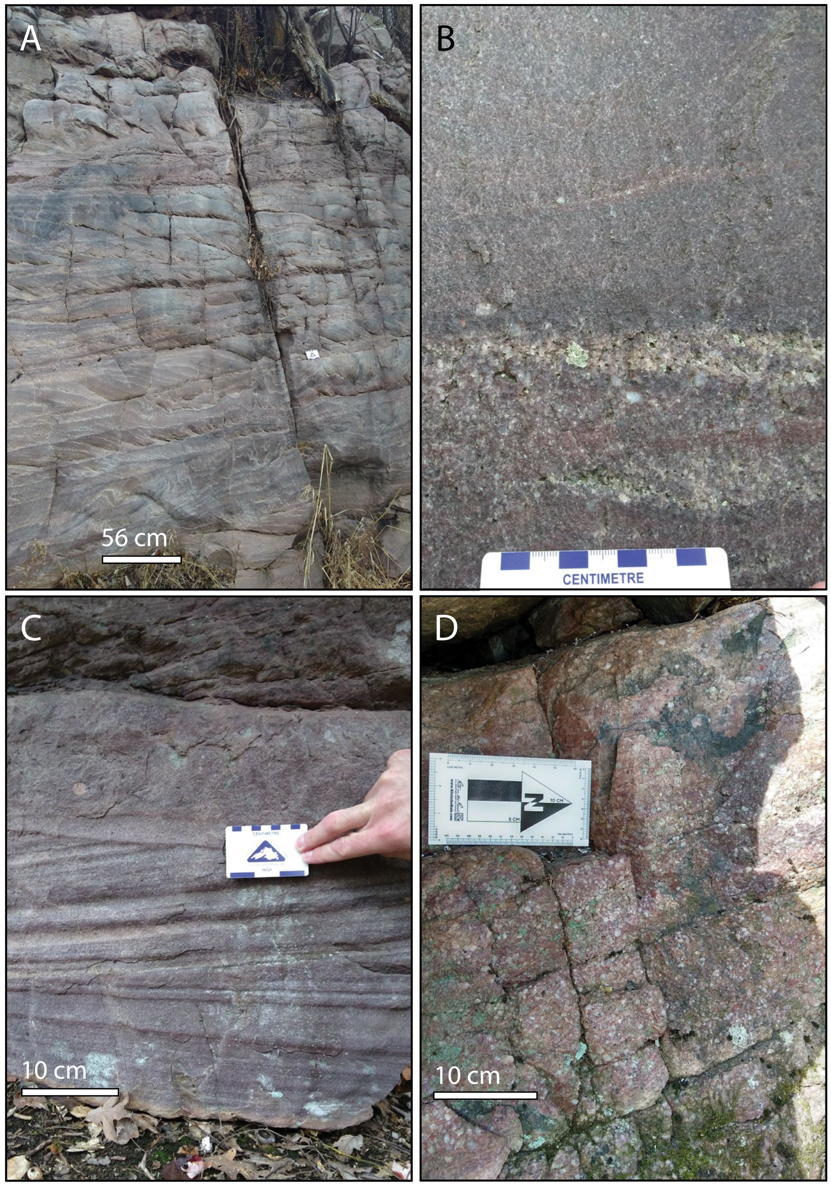
Lower conglomerate and quartzite member

Map unit **YXblc.** Light purple to pink pebble conglomerate (fig. 5) with fine- to coarse-grained clean quartzite and minor argillite. Pebbles are largely composed of monocrystalline and polycrystalline quartz, with subordinate populations of jasper, possible rhyolite, and slate (Stewart and Stewart, 2020). Most pebbles do not exceed 1 cm in diameter. Pebble conglomerate is most abundant at the base of the unit. Sedimentary structures in conglomerate-rich beds near the base of the unit include cut-and-fill structures, basal scour surfaces, planar upper bounding surfaces, and high-angle planar to tangential cross-bed sets that are normally graded (Stewart and others, 2021a). Sedimentary structures in quartzite-rich beds include low-angle scour surfaces with overlying thin pebble conglomerate, high-angle and low-angle cross-beds, reactivation surfaces sometimes draped by pebbles, climbing ripples, brinkpoints, planar upper bounding surfaces, and uncommon decimeter-scale, high-angle concave erosion surfaces. Paleocurrents have a dominant, nearly unimodal southeast directed peak (Stewart and others, 2021a). The lower conglomerate and quartzite member is thought to have been deposited in a range of settings, from a medial to distal fluvial braidplain to a proximal deltaic setting (Stewart and others, 2021a). Strata with reactivation surfaces combined with the pebbly composition suggest in places the proximal deltaic setting was mixed tidal and fluvial influenced. The thickness of the lower conglomerate and quartzite member varies from 75 m at Devils Nose (plate 1) to approximately 250 m at Sauk Hill.

Lower quartzite member

Map unit **YXblq.** Light purple to pink to red, fine- to medium-grained clean quartzite to argillaceous quartzite (fig. 5). Near the base of the unit, phyllite and sandy phyllite beds reaching approximately 1 m in thickness are discontinuously present. These beds are often spectacularly deformed into kink folds. Isolated pebbles up to 2 cm in diameter are found in the upper half of the unit. Thin lags of granules and pebbles, typically a single clast thick, are also uncommonly found in the unit. Overall, the unit is finer-grained and contains far fewer pebble-sized clasts than the overlying and underlying units. It also does not contain relatively thick, >3 cm thick pebble beds typical of the overlying and underlying units. Discontinuous phyllite and sandy phyllite beds contain planar laminated and planar thin bedded decimeter-scale cosets. These beds are thought to have been deposited in an interdune, ephemeral lake setting (Stewart and others, 2021a). Medium-grained to argillaceous quartzite beds contain a variety of sedimentary structures, including broad tangential cross-beds, low-angle basal scours, flame structures, reverse grading within cross beds, and planar-horizontal or pinstripe laminae. These deposits are thought to have been deposited in a subaerial dune setting or interdune setting (Stewart and others, 2021a). Cross-beds for these aeolian deposits are west directed. Distinctive pebble lag deposits are associated with wedge-shaped beds containing tangential cross-beds. Basal scours are present, and the upper bounding surfaces are often planar. These beds are thought to have been deposited by ephemeral streams within an overall aeolian setting (Stewart and others, 2021a). The thickness of the unit probably increases westward. It is approximately 285 m thick near Lost Lake close to the eastern nose of the Baraboo syncline, but increases to 520

Figure 5. Outcrop photos showing characteristic sedimentary structures in the Baraboo Quartzite. Panel A, upper quartzite and phyllite. Panel B, upper conglomerate and quartzite. Panel C, lower quartzite. Panel D, lower conglomerate and quartzite.



m at Sauk Hill in the central portion of the Baraboo Hills. It may be as much as 650 m thick at Ableman's Gorge.

Upper conglomerate and quartzite member

Map unit **YXbuc.** Light-purple to pink, fine- to medium-grained clean quartzite with pebbly beds reaching 20 cm (fig. 5). Pebbles are comprised of monocrystalline vein quartz and polycrystalline quartz, with subordinate numbers of red jasper and slate pebbles. Pebbles rarely exceed 1 cm in diameter. The base of the unit is marked by ≥3 cm thick pebble beds, and the overall abundance of pebbles increases relative to the lower quartzite member. However, pebble conglomerate is not as common in the upper conglomerate and quartzite member compared to the lower conglomerate and quartzite member. Grain size tends to decrease near the top of the section. Quartzite and conglomerate beds contain a number of sedimentary structures that help identify depositional environment, including mud chips, climbing ripples, and tabular beds of moderate-angle to high-angle planar and tangential cross-beds capped by reactivation surfaces. Reactivation surfaces are sometimes draped by pebbles. Quartzite-dominant lithologies in the unit contain reactivation surfaces, soft-sediment deformation, mud chips, tabular beds of moderate-angle to high-angle planar and tangential cross-beds overlain by laminations or thin beds of argillaceous quartzite. Stewart and others (2021a) interpreted these sedimentary structures to reflect deposition in a delta front that was tide and wave influenced. Pebbly beds reflect a more proximal setting, while sand-sized beds (now quartzite) reflect a medial to distal delta front environment. Paleocurrents for the upper conglomerate and quartzite member are either south-directed or southeast-directed (Stewart and others, 2021a). The unit has an estimated thickness of 1205 m east of Devils Lake, 975 m at Sauk Hill, and only 480 m at Ableman's Gorge where it may be tectonically thinned.

Upper quartzite and phyllite member

Map unit **YXbqp.** Light-purple to pink, fine- to medium-grained clean quartzite (fig. 5). Lesser gray phyllite and sandy phyllite with interbedded fine- to medium-grained quartzite is also present. Phyllite and sandy phyllite beds are typically 10–20 cm thick, but can reach several meters (for example, Skillet Creek outcrop in Marshak and others, 2016). Rare, isolated quartz pebbles occur in the unit. Quartzite beds contain sedimentary structures that include reactivation surfaces, soft-sediment deformation, and tangential and planar cross-beds in tabular sets capped by ripples or argillaceous quartzite. These beds are interpreted to record a medial to distal delta front environment that was tidal and wave influenced (Stewart and others, 2021a). Interbedded phyllite and quartzite contain low-angle or planar laminations. Most other sedimentary structures are not visible due to strain localization during deformation. Interbedded phyllite and quartzite is interpreted to record a distal delta front to prodelta environment that was wave influenced (Stewart and others, 2021a). Thickness estimates are difficult because the top of the unit is nowhere exposed. It is probably approximately 200 m thick east of Devils Lake, approximately 300 m thick near Sauk Hill (plate 1), and probably 400 to 500 m thick at Ableman's Gorge (43°28'58"N, 89°55'6"W) though folding has made this estimate very uncertain.

Seeley Slate

Map unit **YXs.** Cross section only. Portions of the Seeley Slate are preserved in 18 cores stored at the WGNHS core repository in Mount Horeb, Wisc. (see Stewart and others, 2018 for drilling locations). The Seeley Slate is mostly a black to gray slate. Near the top of the unit, siltstone and very-fine-grained sandstone laminations become interbedded with slate. Bedding changes from massive near the bottom of the unit, to thin to laminated at the top. Graded beds, scours, and ripples are found near the top of the unit. Granule-sized clasts (2–4 mm in diameter) can be found within the slate (Stewart and Stewart, 2020). Medaris and others (2021) report 1473±3 Ma, 1483±3 Ma, and 1493±3 Ma whole-rock ^{40}Ar/^{39}Ar plateau ages for three axial-planar muscovite samples in the Seeley Slate. Thickness estimates range from 110 m (Dalziel and Dott, 1970) to 150–300 m (Weidman, 1904).

Freedom Formation

Map unit **YXf**. Cross section only. Portions of the Freedom Formation are preserved in 31 cores stored at the WGNHS core repository in Mount Horeb, Wisc. (see Stewart and others, 2018, for drilling locations). The Freedom Formation includes an upper gray to white or yellow dolomite member and a lower red to white to brown to green iron formation, slate, and minor carbonate member. The upper dolomite is often massively-bedded. Bedding is locally visible where detrital quartz, clays, or iron oxides are more concentrated (Weidman, 1904). The lower iron-rich member locally includes a basal iron formation composed of silicate and carbonate facies banded iron formation. This is overlain by a hematite-rich chemical sedimentary facies. The basal banded silicate facies locally contain a thin layer of chamosite and stilpnomelane granules, cemented by large, euhedral carbonate crystals. Most typically, iron formation of the lower iron-rich member contains mm to cm-scale planar to wavy bands. Mineralogically, the banded iron formation transitions from a quartz, chamosite, magnetite assemblage with minor local stilpnomelane at the base, to a carbonate, hematite dominated assemblage at the top, with interbedded intervals of clastic minerals (detrital quartz and clays). Historic exploration drill logs from the early 1900s suggest facies changes occur within the iron-rich member. Weidman (1904) observed mudcracks in the lower iron-rich member at the historic Illinois mine. In places, the iron-rich member contains abundant interbedded cherty iron-rich slates and dolomitic slates. Iron ore was mined from the iron-rich lower member in the early 1900s at the Cahoon, Illinois, and Sauk mines.

Dake Quartzite

Map unit **YXdq**. The Dake Quartzite consists of purple to gray, fine- to coarse-grained quartzite with interbedded pebble conglomerate, that form tabular and lenticular bedsets that range in thickness from 5 to 50 cm. Pebble conglomerate is probably most common near the base of the unit, but a lack of continuous exposures and drill core make this uncertain. Quartzite ranges from poorly sorted to well sorted with uncommon phyllite beds. Phyllite or sandy phyllite interbeds range from 1 mm to 5 cm thick, and they often mark the boundaries of bedsets (Stewart and others, 2018). Pebble clasts reach 3 cm, and in outcrop are composed of quartz with lesser jasper and slate. Cores of the Dake Quartzite where it overlies the lower Freedom Formation also contain angular iron formation pebbles and quartzite pebbles. Sericite is locally present bounding quartz grains. Geiger (1986) examined drill core of the Dake Quartzite and observed pockets of clay reaching 4–5 mm in size. X-ray scans of separates showed a major kaolinite peak, and a lesser feldspar peak, suggesting the pockets could be weathered feldspar (Geiger, 1986). Detrital zircons from the Dake Quartzite at North Freedom (43°27'23"N, 89°52'7"W) contain a youngest zircon population that gives a maximum depositional age for the unit of 1630±9 Ma (Stewart and others, 2021a). Besides the clasts of iron formation and clay pockets, the Dake Quartzite is lithologically similar to the Baraboo Quartzite. The unit is recognized as younger than the Baraboo Quartzite because it occurs stratigraphically above the Freedom Formation. This interpretation is based on Dake Quartzite outcrop locations and bedding orientations relative to the magnetic anomaly signature of the Freedom Formation (Stewart and others, 2018). The Dake Quartzite is approximately 65 m thick (Leith, 1935).

Rowley Creek Slate

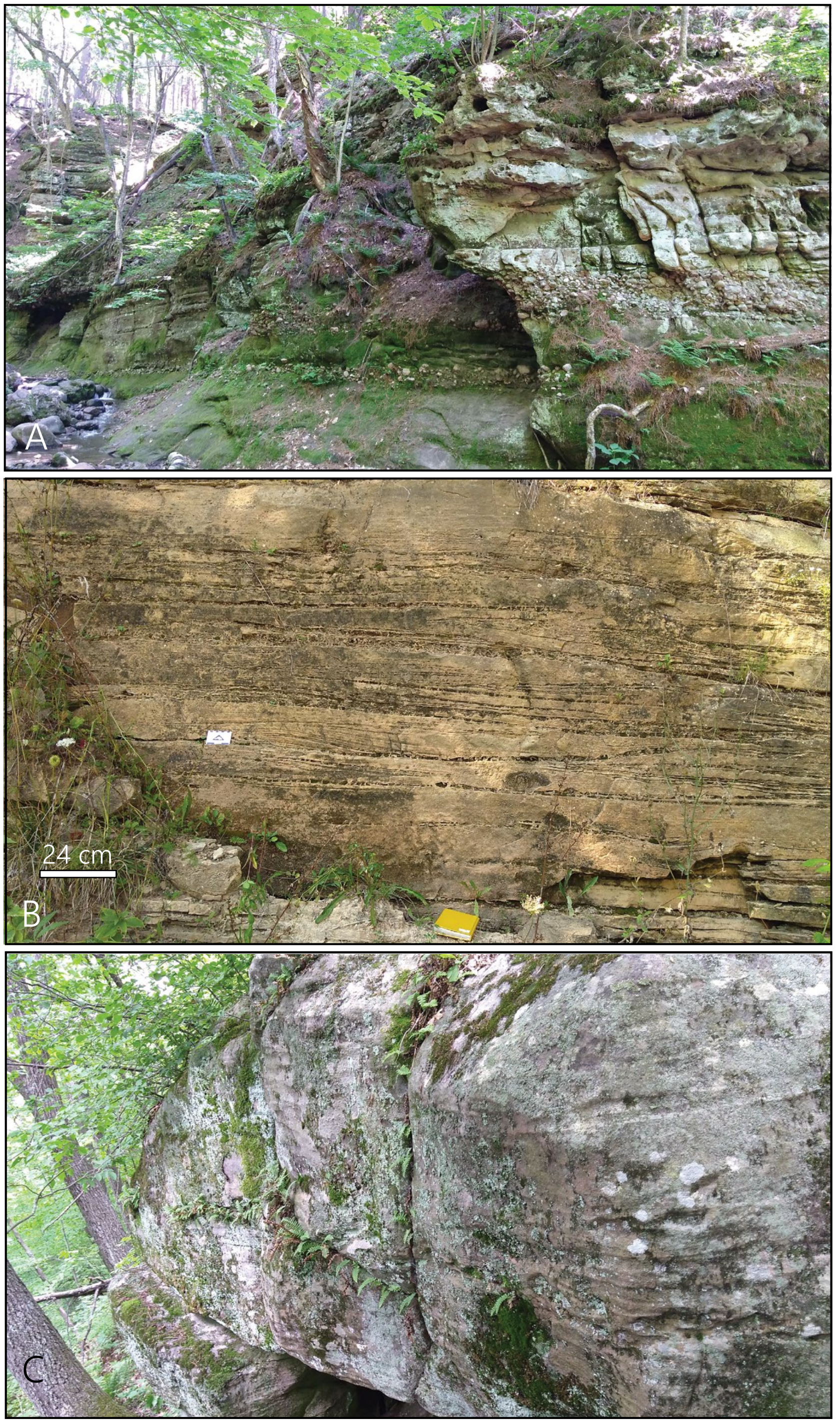
Map unit **YXrc**. Cross section only. The Rowley Creek Slate is a gray slate found only in the eastern interior of the Baraboo Hills. It is composed of sericite with lesser chlorite and quartz. It is at least 45 m thick (Leith, 1935).

Cambrian

Wonewoc Formation

Map unit **Ꞓw**. White, tan, and brown fine- to medium-grained sandstone (fig. 6). Generally, the unit is a clean quartz arenite with little feldspar. Cross-beds occur throughout much of the unit. The uppermost 1–3 m of the Wonewoc, known as the Ironton Member, contains *skolithos* burrows and has a brown or reddish-brown color due to iron oxides coating quartz grains (Clayton and Attig, 1990). The underlying Galesville Member generally lacks trace fossils. The unit is approximately 30 m thick (Dalziel and Dott, 1970).

Figure 6. Outcrop photos. Panel A, interbedded conglomerate and sandstone of the Parfreys Glen Formation. Panel B, hummocky cross stratification and burrowed horizons in the Jordan Formation of the Trempealeau Group. Panel C, planar and cross bedded sandstone of the Wonewoc Formation of the Elk Mound Group.



Tunnel City Group

Map unit **€tc**. The Tunnel City Group in the Baraboo Hills is a tan, white, or light-brown, fine- to medium-grained sandstone. Trace amounts of glauconite are locally present. Carbonate cemented sandstone is interbedded with clean quartz arenite. Hematite cement is also common. Clayton and Attig (1990) note minor detrital feldspar. The Tunnel City Group is moderately sorted but locally can contain Baraboo Quartzite pebbles. Cross-beds are common. Clayton and Attig (1990) found the unit typically weathers to undulating hills and slopes in Sauk County. Crinoids are locally present in the lower Tunnel City Group near Prentice Creek in the Durwards Glen quadrangle. The unit has an estimated thickness of 30 m.

Trempealeau Group

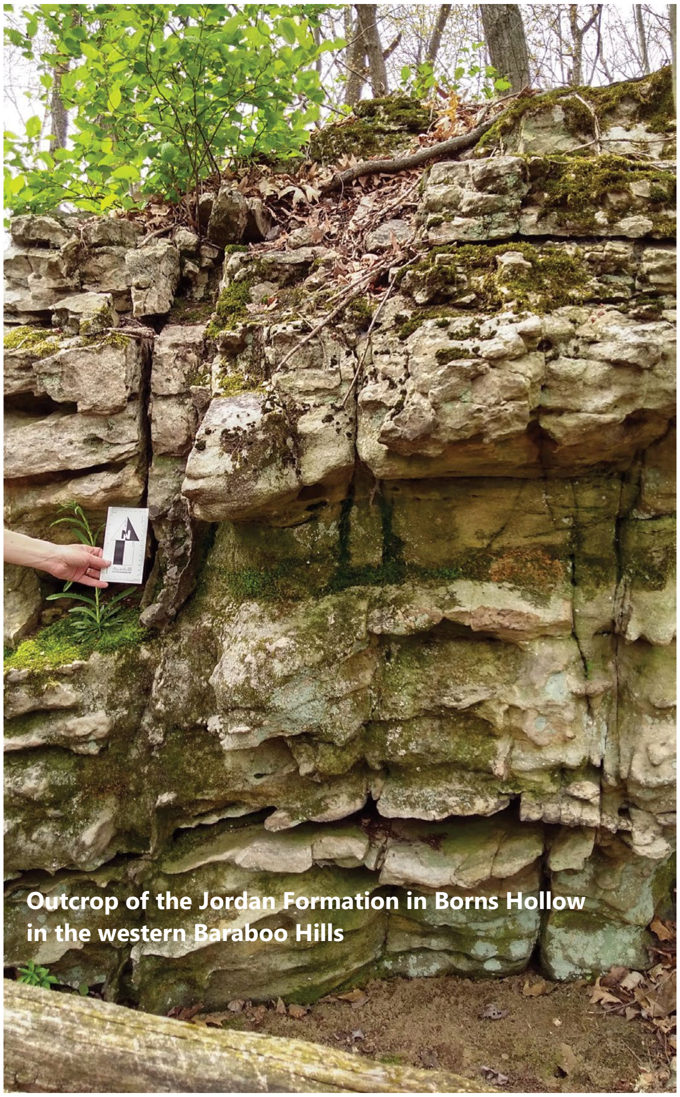
Map unit **€t**. The Trempealeau Group includes the upper Jordan Formation and the lower St. Lawrence Formation. The Jordan Formation is a tan, gray, or white, fine- to coarse-grained sandstone. The lower Jordan contains detrital feldspar, but feldspar is largely absent near the top (Clayton and Attig, 1990). The Jordan can include pebble sized clasts of Baraboo Quartzite. The Jordan Formation is typically moderate to well sorted, and bedding is packaged into thin to medium tabular sets with common internal trough and low-angle cross-beds. Hummocky cross stratification and *skolithos* burrows, evidence of a marine environment, are nicely exposed at the junction of Scenic Rd and Wisconsin State Highway 136 (fig. 6; 43°28'44"N, 89°51'31"W) in the North Freedom quadrangle. The Jordan Formation often forms cliffs (Clayton and Attig, 1990). The unit is approximately 20 m thick.

The St. Lawrence Formation in the Baraboo Hills area is a white to tan, fine- to very fine-grained sandstone to siltstone. Some beds contain carbonate cement. The unit often forms a bench below the cliff-forming Jordan Formation, and is generally poorly exposed (Clayton and Attig, 1990). Away from the Baraboo Hills, the St. Lawrence Formation is more easily recognizable. The base of the unit is the Black Earth dolomite, and the top is the Lodi siltstone. These members are not present in the Baraboo Hills area. The unit has an estimated thickness of approximately 12 m.

Ordovician to Cambrian

Parfreys Glen Formation

Map unit **O€pg**. The Parfreys Glen Formation (Clayton and Attig, 1990) includes pink, tan, and brown fine- to medium-grained quartz arenite and pebble, cobble and boulder conglomerate (fig. 6). Conglomerate clasts are composed of the Baraboo Quartzite, and are set in a matrix of clean quartz sand or glauconitic sand (Stewart and Stewart, 2021). Trough cross-beds and planar cross-beds are common in sandstone intervals. Pebbles occur both as drapes along cross beds and as floating pebbles within sandstone beds (Stewart and Stewart, 2021). Clast sizes diminish away from Precambrian topographic highs, but along the southern slopes of the Baraboo Hills, clast sizes tend to be larger and diminish less rapidly (Dalziel and Dott, 1970). Burrows are locally present in glauconitic sand beds (Stewart and others, 2024). The unit drapes Precambrian hills. Adjacent to Precambrian cliffs, the Parfreys Glen Formation contains angular talus (Clayton and Attig, 1990). The unit has a variable thickness.



Outcrop of the Jordan Formation in Borns Hollow in the western Baraboo Hills

Ordovician

Oneota Formation

Map unit **Oo**. The Prairie du Chien Group in the Baraboo Hills area is restricted to the Oneota Formation. The unit is mostly a tan crystalline dolomite with dolomite-cemented sandstone. The base of the unit includes abundant silicified dolomite breccia (Stewart and others, 2021b). Chert is common in the unit. Dalziel and Dott (1970) report silicified oolite near the base. Bedding varies from laminated to massive (Stewart and others, 2024). Dalziel and Dott (1970) note desiccation cracks and ripples within the unit. Stromatolites are locally present in exposures in the interior of the Baraboo Hills (Stewart and others, 2024). The unit is poorly exposed across the hills, but probably forms the caprock for a number of upland surfaces. The thickness varies because of the unconformity with the overlying St. Peter Formation.

St. Peter Formation

Map unit **Osp**. The St. Peter Formation is a white, tan, or orange, fine- to medium-grained sandstone. Minor coarse-grained sand occurs in some beds. Sorting varies from moderate to poor. The St. Peter Formation is only exposed in the eastern interior of the Baraboo Hills at Pine Bluff, where quartz overgrowths are well developed and the unit forms a resistant cap to the bluff. The base of the unit marks an important unconformity with significant relief. Approximately 50 m of section is exposed at Pine Bluff, but the unit can be thicker or thinner elsewhere in southern Wisconsin.

Pleistocene

Rock glacier

Map unit **Qr**. Talus piles of quartzite forming linear trains meters to tens of meters in thickness. They are generally found beneath quartzite ledges and flowed downslope. Locally, they contain preserved ridges which may reflect flow banding (Stewart and Stewart, 2021).

Glacial sediment

Yellow to brown till deposited by the Green Bay Lobe of the Laurentide Ice Sheet. Till is generally composed of poorly sorted clay (5–10%), silt (15–30%), sand (60–75%), and gravel (10%) (Clayton and Attig, 1990). Glacial sediment belongs to the Horicon member of the Holy Hill Formation. Map unit **Qt**. Till draped over bedrock. The morphology of the unit is controlled by the underlying bedrock topography. It contains abundant sediment-cored drumlins and bedrock-cored drumlins. Map unit **Qtc**. Till containing numerous ice collapse features, resulting in modern kettle lakes and ponds. Map unit **Qts**. Mixture of till and lake sediment. The unit is restricted to the interior of the Baraboo Hills east of the Johnstown moraine. It formed during retreat of the glacier as the West Baraboo basin portion of glacial Lake Wisconsin expanded eastward. Map unit **Qtj**. Glacial sediment of the Johnstown moraine, defining the maximum extent of the Green Bay Lobe of the Laurentide Ice Sheet. The moraine locally contains basal outwash deposits, formed during ice advance (Lundqvist and others, 1993). Above the discontinuous outwash deposits, it typically contains boulder gravels likely deposited by eskers, stratified till of subglacial origin, and thin boulder gravels of supraglacial origin (Lundqvist and others, 1993).

Lake sediment

Map unit **Qls**. Offshore lake sediment varying from sand to silt to clay. Deposited as part of Glacial Lake Wisconsin. Lake sediment varies by depositional basin and location. The east Baraboo basin consists of mostly sand with minor silt and clay. The west Baraboo basin near Diamond Hill and the Reedsburg basin north of Ableman's Gorge consist of laminated silt and clay. Lake sediment deposited in the Loganville basin west of the Narrows Creek gorge consists of laminated clay and silt with interbedded sand lenses (E. Carson, oral commun., 2020).

Windblown sediment

Map unit **Qw**. Sand deposited in both sheets and dunes. Well sorted. Windblown deposits in excess of 1.5 m accumulated on flat or low-angle slopes in the late Pleistocene towards the end of the last glaciation (Clayton and Attig, 1990).

Meltwater stream sediment

Gravel and sand deposited by streams draining glacial meltwater. Map unit **Qs**. Found adjacent to the Johnstown moraine, where it forms a prominent flat outwash plain. Gravel is common near the Johnstown moraine. Sand becomes common farther west from the moraine edge (Clayton and Attig, 1990). In places it is deeply incised by Holocene streams. Map unit **Qsc**. Found east of the Johnstown moraine. The unit occurs in areas of hummocky topography where meltwater stream sediment buried stagnant ice during glacial retreat. Till may be present in places and generally becomes more common to the east.

Holocene to Pleistocene

Alluvium

Map unit **Qam**. Dark brown to tan mud, silt, and fine-grained sand. Modern overbank stream deposits. Map unit **Qa**. Largely sand, silt and mud, but can contain cobbles and gravels. Mostly premodern alluvial overbank stream deposits, but also includes some premodern colluvial deposits on the toes of hillslopes. Map unit **Qat**. Sand, silt and mud terrace deposits. The terrace formed from incision into lake deposits, but the unit also contains a veneer of alluvial overbank stream deposits. It is only present on the south-side of the Baraboo River southeast of a constriction in the river near Rock Springs. It contains two terraces; the lower terrace is 1–2 m above Qam, and 2 m below the upper terrace. The upper terrace is 4–5 m below glacial lake deposits. Map unit **Qalo** Sand, pebble, cobble, and boulder stream deposits. The unit formed from incision into Qad. Map unit **Qad**. Tan, poorly sorted pebble, cobble, and boulder conglomerate. The unit is unlithified. Cuts in quarry faces indicate the conglomerate was deposited in continuous foresets 12 m in height (Bretz, 1950). Foresets are composed of pebbles and cobbles, with isolated boulders reaching up to 5 feet in diameter (Bretz, 1950). Cobbles and boulders are mostly composed of Baraboo Quartzite, with subordinate dolomite, sandstone, and even some clay-rich till (Bretz, 1950).

Holocene

Peat

Map unit **Qp**. Organic-rich sediment and muck. Present in low-lying, flat, wet areas. Probably 1 to 5 m thick in most places (Clayton and Attig, 1990; Hooyer and others, 2021).

Artificial fill

Map unit **Af**. Human-made deposits of quartzite fragments from aggregate mining.

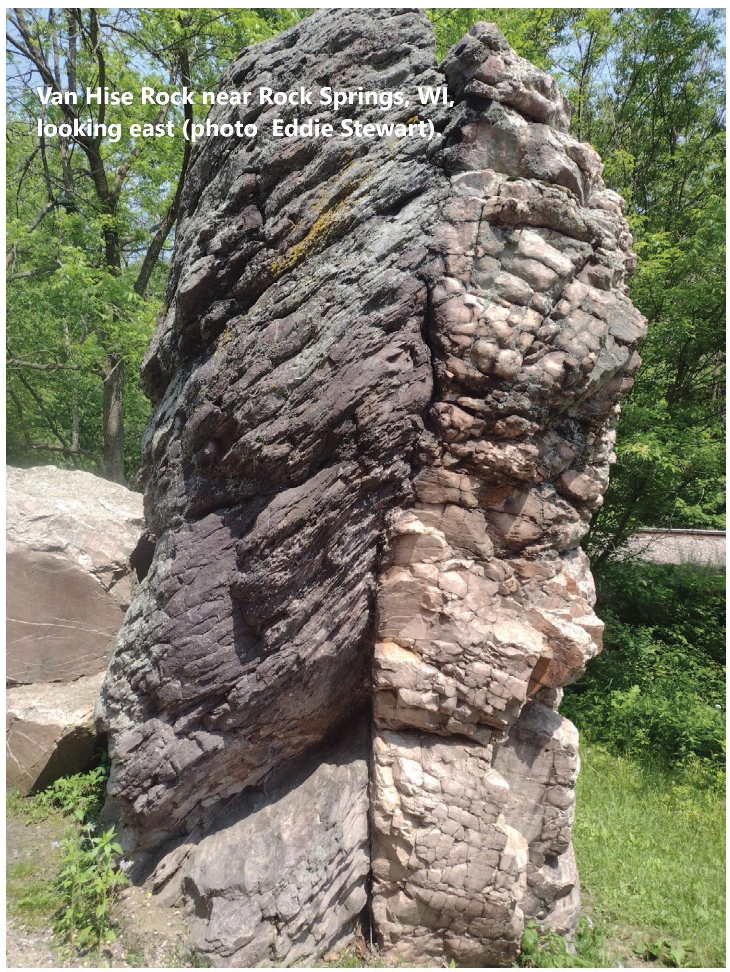
Map Structures

This section begins by describing planar and linear deformational fabrics in the Baraboo Quartzite. Next, field and microstructural evidence for contractional and extensional faults are described. Finally, major folds are described. Orientation data of structural features on plate 1 are simplified from Stewart and Stewart (2020; 2021), and Stewart and others (2021; 2024). The full set of orientation features used to define contacts, folds, and faults is available in dataset 1.

Foliation

Planar deformational fabrics in the Baraboo Quartzite record two episodes of deformation (Dalziel and Dott, 1970; Marshak and others, 2016; Marshak and others, 2024). Early deformation in quartzite beds (D1) includes a disjunctive, or spaced cleavage (S1) that typically is oriented perpendicular to bedding. The disjunctive cleavage is not only present in the Baraboo Quartzite but is visible in the Dake Quartzite at West Baraboo as well (43°28'12"N, 89°46'31"W). D1 fabrics in the quartzites are considered to have formed early from bedding parallel shortening during the initial stages of folding (Czeck and Ormand, 2007). Quartzite cleavage (S1) is nearly horizontal on the sub-vertical north limb of the Baraboo syncline, and dips steeply south across much of the South Range where bedding is shallowly north-dipping. Cleavage development in the quartzite began prior to folding. After the quartzite cleavage formed, it was rotated into its current two major orientations during folding. D1 in phyllitic beds of the Baraboo Quartzite resulted in a north dipping S1 penetrative cleavage that approaches an axial planar orientation. Phyllites with abundant pyrophyllite are weaker than quartzite layers (Czeck and others, 2020), and strain localization in these layers during progressive folding led to the rotation of cleavage planes to near axial planar orientations. Therefore D1 cleavages, overall, display the classic pattern of cleavage refraction. The Seeley Slate also contains an axial planar S1 cleavage consisting of aligned, very-fine grained muscovite (Medaris and others, 2021). All D1 fabrics are considered to have formed and be related to the development of the larger map-scale folds (Dalziel and Dott, 1970; Czeck and Ormand, 2007; Marshak and others, 2024). Van Hise Rock at Ableman's Gorge (43°29'22"N, 89°54'56"W) famously displays these cleavage-bedding relationships.

Later deformational fabrics (D2) are only observed in relatively thick phyllitic beds in the South Range. In the South Range of the Baraboo Hills, bedding typically dips gently to moderately north. Here, locally present are D2 fabrics including north vergent kink bands and a south-dipping crenulation cleavage (S2) (Marshak and others, 2024). The north vergence implied by these structures, opposite to the south-vergent map-scale folds in the Baraboo Hills, led some workers to assume an episode of extension post-dated contraction (Hendrix and Schaiowitz, 1964; Hempton and others, 1986). However, recent work by Marshak and others (2024) showed the S2 crenulation cleavage and north-vergent kink bands were not created in an extensional setting. Instead, they found D2 structures are likely the result of progressive folding and bed rotation, and are caused by the S1 cleavage plane accommodating a component of the regional NNW-SSE horizontal shortening.


Van Hise Rock near Rock Springs, WI, looking east (photo Eddie Stewart).

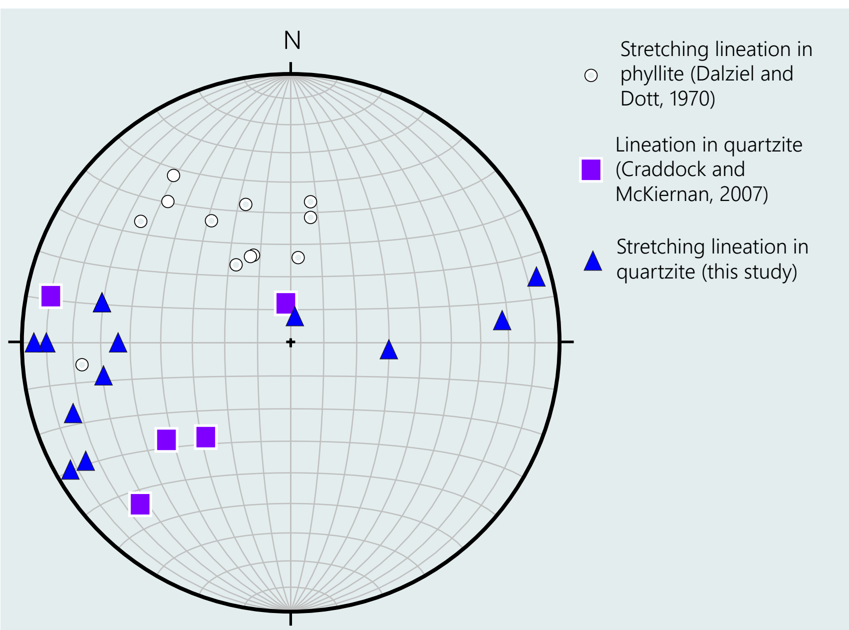
Lineation

Twelve new mineral stretching lineations were measured from oriented hand samples of the Baraboo Quartzite. Stretching lineations are subtle in the field, so saw cuts were made on the oriented samples parallel to the spaced cleavage, and where possible the orientation of aligned, elongated quartz grains was measured on the foliation plane. Stretching lineations were easiest to observe within the spaced cleavage, and very subtle in the rock matrix around the cleavage. Care was taken to avoid measuring intersection lineations between bedding and cleavage.

Most lineations measured in quartzite plunged shallowly NE or SW with some scatter, sub-parallel to the Baraboo syncline fold axis (fig. 7). These lineations are interpreted to reflect stretching during D1, and so could be referred to as l1. Some scatter in the data was probably introduced from folding along a sub-parallel fold axis. Maximum stretch directions approximated by quartz grain shapes measured in oriented thin sections of the Baraboo Quartzite (Craddock and McKiernan, 2007) show a similar pattern (fig. 7).

Quartzite lineations differ from most phyllite stretching lineations reported by Dalziel and Dott (1970). Phyllite lineations (longrain in the terminology of Dalziel and Dott, 1970) pitch steeply within the plane of phyllitic cleavage (fig. 7), which is approximately axial planar to the Baraboo syncline. This phyllitic cleavage is an S1 fabric described above. This down-dip lineation is roughly perpendicular to the Baraboo syncline fold axis.

Figure 7. **Equal area stereographic projection of stretching lineations in phyllite (open circles) and quartzite (solid squares and solid triangles). Most quartzite lineations trend close to the ENE-WSW trending Baraboo syncline. Phyllite lineations pitch steeply within the S1 foliation plane.**



Contractional Faults

North Freedom Fault

The North Freedom thrust fault strikes northeast, dips northwest. The dip angle is unconstrained. The history of the fault's recognition is complicated and is summarized here. A more complete history is provided in supplemental materials of Stewart and others (2018). Questions pertaining to the location of the North Freedom fault deal heavily with the stratigraphic assignment of two quartzite outcrops (43°27.23"N, 89°52'7"W and 43°27'36"N, 89°51'30"W) near the town of North Freedom along the axis of the Baraboo syncline. Leith (1935), using exploration drill logs from the early 1900s, assigned these outcrops to the stratigraphically higher Dake Quartzite, and used a NE-plunging Baraboo syncline to bring the stratigraphically higher quartzite to land surface. Schmidt (1951) was the first to map a fault in the North Freedom area. He mapped a fault south of the town of North Freedom using some of the same exploration logs from the early 1900s. Schmidt's mapping required a fault because he assigned the quartzite near North Freedom to the older Baraboo Quartzite, which in his interpretation was outcropping within the trough of the Baraboo syncline in an otherwise stratigraphically impossible location. The exploration logs, critical to understanding the geology of the area, were lost for a stretch of time after publication of Schmidt (1951), stalling progress. Dalziel and Dott (1970) relied heavily on Schmidt (1951). They also mapped the quartzite near North Freedom as Baraboo Quartzite and included a fault south of the town of North Freedom. In a tremendous stroke of good fortune, numerous iron ore exploration drill logs, maps, and a number of historic cores were acquired by Bruce Brown of the WGNHS from US Steel and local landowners in the early 1980s. The logs and maps were rediscovered in 2015 in the basement archives of the WGNHS, with the maps providing location information for cores stored by the WGNHS at the core repository. Stewart and others (2018) examined over 80 of the rediscovered logs, from drill cores located within 3 km of the town of North Freedom (fig. 4), and found the mapping of Leith (1935) was largely correct and the quartzite near North Freedom belonged to the Dake Quartzite. Schmidt (1951) did not include logs from Weidman (1904), which made his interpretation untenable. Later detrital zircon dating of this quartzite found a population of 1630 Ma detrital zircons, younger than any ages found in the Baraboo Quartzite (Stewart and others, 2021a). This result supported the assignment of these outcrops to the younger Dake Quartzite. Stewart and others (2018; 2021b) found a fault was still needed in the area (described below), but they shifted the trace of the fault north of the town of North Freedom, and extended the fault to the boundaries of the Baraboo Hills.

Map relations require the North Freedom fault, but the fault is not exposed at land surface. Near Zion Lutheran Church in the western Baraboo Hills (43°25'21"N, 89°58'34"W), the fault is interpreted to underlie tens of meters of alluvium and Parfreys Glen Formation. On the adjacent hills, the contact between the upper conglomerate and quartzite member (XYbuc) and the upper quartzite and phyllite member (Xbqp) is offset across the valley. At a second location to the northeast near Diamond Hill (43°27'2"N, 89°53'41"W), field control combined with nearby drill logs in the Seeley Slate indicate the thickness of the upper quartzite and phyllite member (XYbqp) is too small and a fault is needed. The fault is thought to thrust the lower portion of the upper quartzite and phyllite member (XYbqp) over either the upper portion of the upper quartzite and phyllite member or over the Seeley Slate near Diamond Hill. Differences in the hanging wall and footwall units along the strike of the fault are probably not caused by differences in slip, but are related to a NE-directed plunge in the entire section. This causes deeper rocks and a deeper portion of the fault to be exposed along the western periphery of the Baraboo Hills.

Quartzite on the hills adjacent to the North Freedom fault in the far western Baraboo Hills are deformed locally by the fault (Kemmer and Kovac, 1937; Stewart and others, 2018), but a fault core is not exposed. Quartzite near the fault has a strong deformational cleavage oriented N70E/65°NW. Brecciation and veining are also abundant near the mapped fault (Stewart and others, 2018).

South Range Fault

The South Range fault was first mapped and named by Stewart and Stewart (2020), but a fault in a similar location was mapped by Schmidt (1951). The South Range fault strikes roughly N60E, dips 30°NW. The fault is interpreted to be a thrust fault that places the lower conglomerate member of the Baraboo Quartzite over either the lower quartzite member of the Baraboo Quartzite or younger portions of the lower conglomerate member. Deeper structural levels are exposed along the southwest trace of the fault, and structurally higher levels are exposed along the northeast trace of the fault, accounting for the variation in footwall units (for additional discussion, see the discussion on the Skillet Creek anticline and Otter Creek syncline).

Outcrop observations

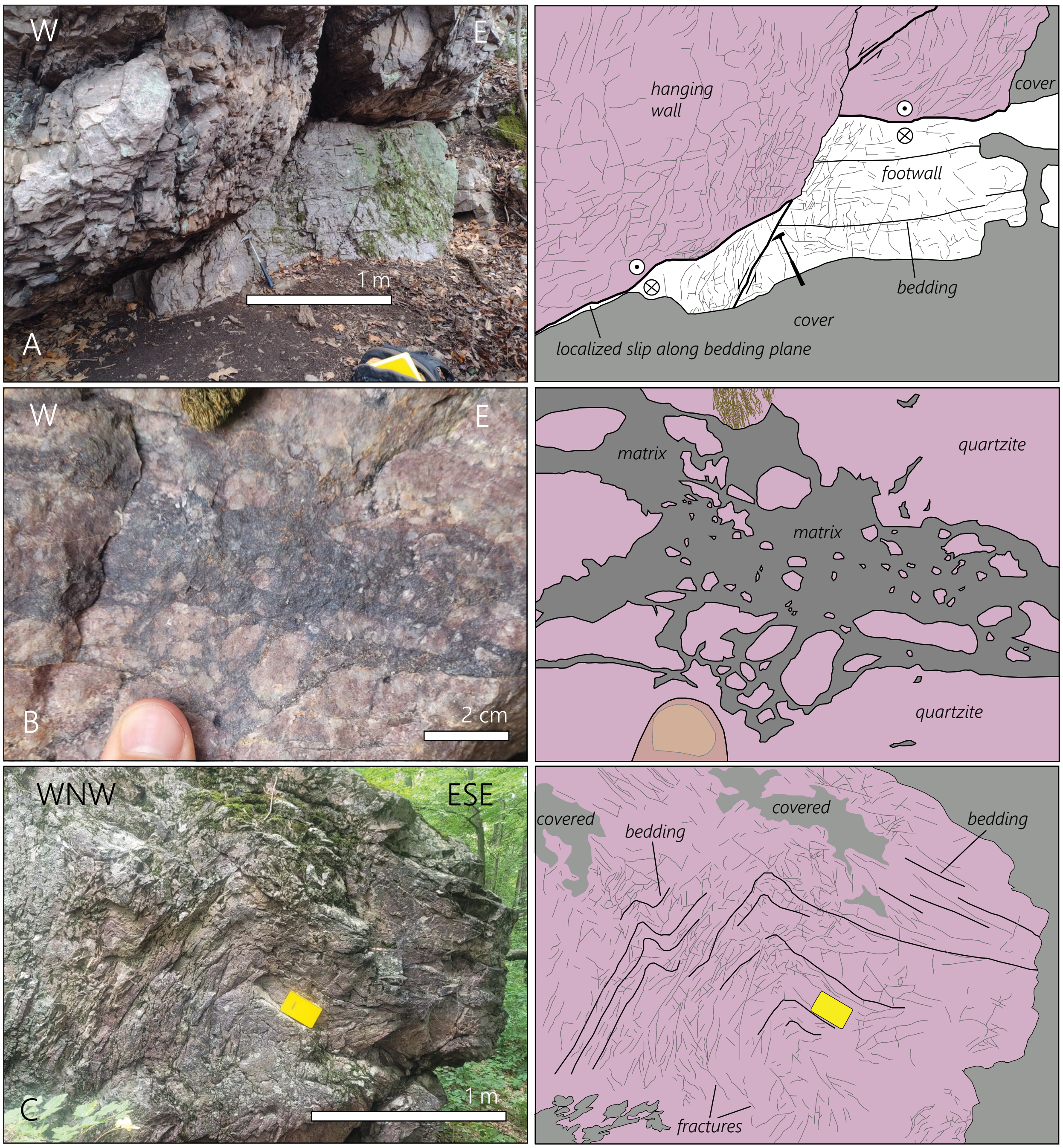
The South Range fault shows evidence for localized deformation near its fault trace within the damage zone of the fault. Both discrete slip surfaces and zones of cataclastic quartzite or quartzite breccia occur. Minor faults within the damage zone of the fault generally dip shallowly to the north and have weakly defined slickenlines that plunge NW to NNW. In the hanging wall, faults in the damage zone reactivate bedding surfaces where the fault forms a hanging wall flat (fig. 8a). Collectively, these small bedding parallel faults serve to distribute strain across a roughly 100-m-thick zone in the hanging wall. Small faults most commonly appear as narrow millimeter-scale slip surfaces composed of a fine-grained mixture of microcrystalline hematite and white micas, muscovite and pyrophyllite. Locally, shear bands and imbrication can be observed that show top to the SE to SSE transport direction. Broader zones of cataclasite and breccia also occur in the hanging wall in a narrower 10–20-m-thick zone adjacent to the footwall. The widest zone observed is about 15 cm thick at its widest extent and along strike narrows to a mm-scale zone of slip akin to the previously described fault surfaces. Deformation within this zone is expressed as variably sized angular clasts of Baraboo quartzite and brecciated quartz veins. The matrix is either composed of finer grained cataclastic quartz or a dark matrix of microcrystalline hematite and white micas (fig. 8b), similar to the finer slip surfaces.

Mesoscale open to tight folds are also present within the hanging wall with wavelengths up to 1 m (fig. 8c). These folds occur in quartz-rich lithologies without phyllitic beds. Such folds have only been observed adjacent to the South Range fault. At the map scale, bedding is variable adjacent to the fault, which is probably related to folding in the fault damage zone.

Microstructural observations

Cataclastic textures are common in quartzite from the damage zone of the fault. Relict quartz grains are commonly mantled by a rim of fine-grained (<10 μm) quartz. In the fault damage zone, the grain size variation is interpreted to be caused by brittle deformation mechanisms and cataclasis. Grains exhibit varying degrees of microfracturing, and fluid inclusion trails are common. Internal strain of quartz grains is visible, however subgrain growth is weak and most relict quartz grains exhibit only undulose extinction. Cataclasis of quartzite is highly localized, occurring primarily as a network of narrow, up to several mm thick bands of heavily grain-sized reduced quartz. Such bands occur at a variety of angles but are commonly steeply oblique to the orientation of local mesoscopic thrusts. These bands interlock around macroscopic blocks of quartzite, producing a block-supported network of cataclasite.

Figure 8. Field photos from the damage zone of the South Range Fault. Panel A, discrete slip surfaces localize strain along bedding planes in the hanging wall of the fault. The main thrust is cut by a later normal fault. Panel B, breccia zone within a small fault in the damage zone of the South Range fault. The dark matrix is probably composed of microcrystalline magnetite, quartz, and mica. Panel C, folded Baraboo Quartzite adjacent to the South Range fault. Numerous small cataclastic faults are believed to accommodate the folding in the pure quartzite.



Specular hematite slip surfaces are present locally. Individual hematite grains are very fine-grained (<5 μm) and are elongate, with their long axis aligned with the shear direction of the slip surface (NW-SE). Rare red polygonal grains of hematite are locally present around the hematite lenses.

Adjacent to the microcrystalline hematite layers are layers of fine and sheared phyllosilicates. Energy dispersive spectroscopy (EDS) results on the scanning electron microscope indicate the layer silicates contain potassium and suggests these zones are a mixture of muscovite and pyrophyllite. Locally, the layer silicates exhibit shear banding that reflects a top to the SE sense of shear. Other regions exhibit a more chaotic shear pattern or a turbulent vorticity pattern when viewed parallel to lineation and perpendicular to foliation.

While deformation mostly is accommodated by brittle mechanisms, crystal plastic deformation of quartz is locally visible. Crystal plastic deformation of quartz has been observed at other Baraboo Quartzite exposures (Dalziel and Stirewalt, 1975). Relict quartz grains within the damage zone variably possess a shape-preferred orientation, suggesting ductile deformation. Some quartz grains exhibit subgrain formation and have interfingering boundaries with neighboring quartz grains. Bulging recrystallization is common around interfingering boundaries. Bulging recrystallization of quartz is most developed along localized (<3 mm wide) cataclastic zones where heavily grain-size reduced quartz is in contact with larger relict grains with little or no composition of other phases. Such large grains exhibit abrupt transitions from interiors of distributed strain (e.g. undulose extinction) to zones of abundant subgrain formation to the cataclastic band matrix. As such, these grains exhibit a weakly developed but evident core-and-mantle structure characteristic of bulging recrystallization. Evidence of dissolution of quartz is visible as well, such as the formation of stylolites and truncation of detrital grains, indicating fluid-rock interaction.

Baxters Hollow Fault

The Baxters Hollow fault has not been previously mapped, though shearing and deformation in cores from the Baxters Hollow area were noted by LaBerge and others (1991) and Medaris and others (2011). The Baxters Hollow fault strikes roughly N90W, and dips approximately 9 degrees north. We interpret the fault at Baxters Hollow as a tilted detachment surface near the base of the Baraboo Quartzite. Detachment surface is used here and elsewhere in the report to describe a low-angle, bedding parallel surface that forms a footwall flat in a thrust system. We note that while detachment surfaces are commonly used to describe low-angle normal faults, they can also be used to describe footwall flats in thrust systems. The Baxters Hollow detachment horizon marks an upper footwall flat above an inferred basal thrust that occurs more than 1 km below the Baraboo Quartzite. The footwall flat at Baxters Hollow likely localized near the base of the Baraboo Quartzite due to the inherent weakness of the Baraboo Quartzite-Baxters Hollow granite contact. Driese and Medaris (2008) characterized the top of the Baxters Hollow granite as a highly altered paleosol, and the weakness of the paleosol may have helped localize the fault. Cores previously studied by LaBerge and others (1991) are stored at the WGNHS core repository and provide transects across the Baxters Hollow fault zone. The paragraphs below detail observations from cores ASDC 607 (43°23'21"N, 89°47'12"W) and ASDC 613 (43°23'2"N, 89°47'33"W). ASDC 607 was drilled approximately 1 km northeast of ASDC 613 (fig. 4). The location and ID of all cores are available in accompanying dataset 1. The fault is not exposed at land surface.

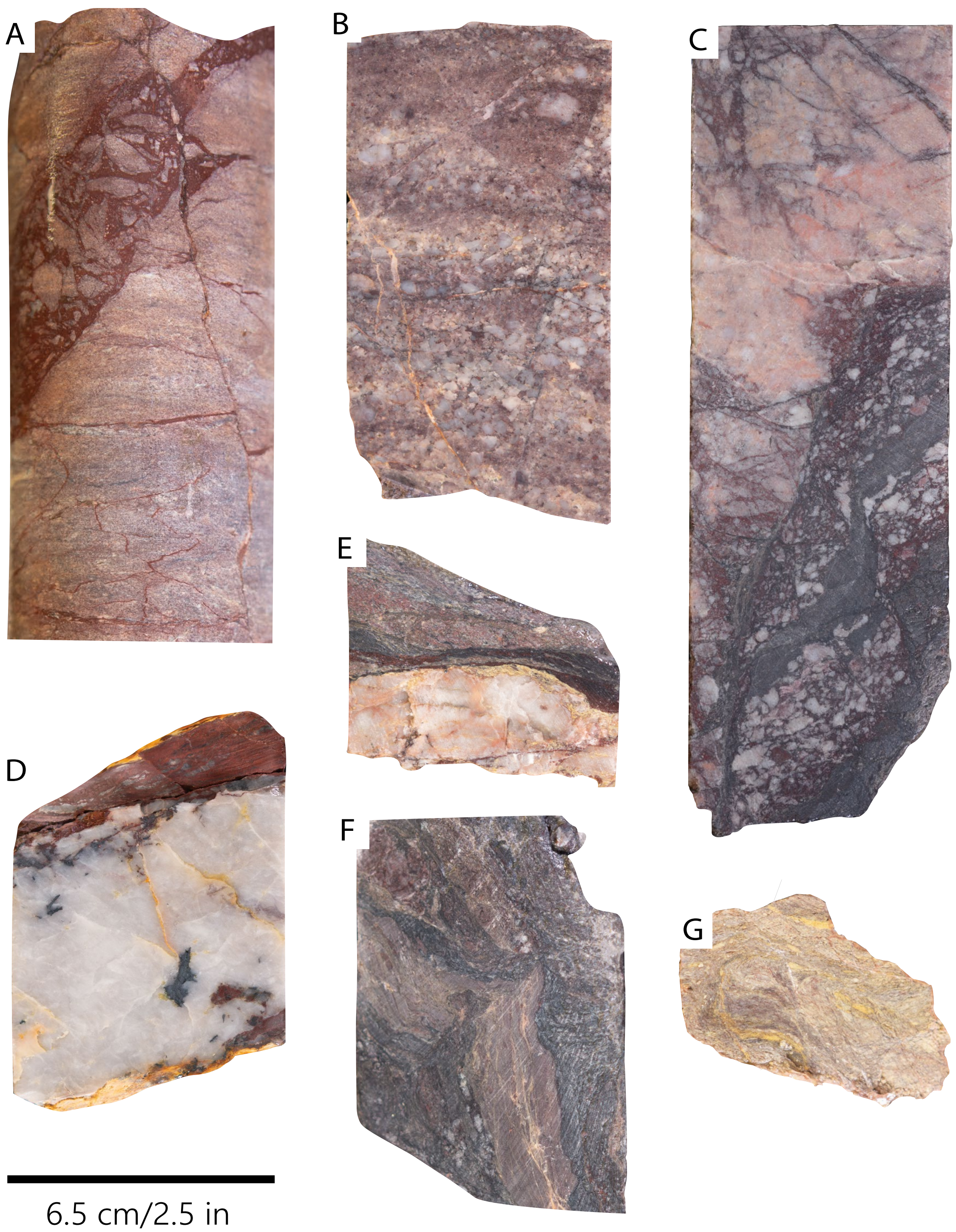
Core observations

Drill cores ASDC 607 and ASDC 613 sample the underlying Baxters Hollow igneous suite, the overlying Baraboo Quartzite, and the fault zone between the two units. The underlying igneous suite contains granitic rock and a fine-grained, gray to pink to purple volcanic rock varying in composition from chloritic andesite to chloritic basalt (fig. 9a; LaBerge and others, 1991). Some of the volcanics show flow features and have alternating light and dark layers. The rock directly overlying the fault zone consists of the lower conglomerate and quartzite member of the Baraboo Quartzite (fig. 9b). The quartzite in the drill cores manifests as 1) conglomeratic quartzite with pebble- to sand-sized clasts (fig. 9b–c), and 2) sections of pure white to pink quartzite with little to no visible sedimentary features (fig. 9d).

Rocks in the fault zone exhibit clay and clay gouge, sheared zones, cataclasite, thin slip surfaces, veins, and breccia (Figure 9a, c–g). Zones of clay gouge, cataclasis, and brecciation indicate the mechanical breakup of host rock due to slip along a fault. Clay-rich zones within the drill core occur largely as rubble, and are tan to pink, light gray, and dark gray in color, with some layers of bright yellow clay (fig. 9g). Foliation and fragments of reworked parent rock are visible within the clay gouge. We define these ~1 m wide clay-rich zones as the main fault core. Zones of brecciation show angular to subrounded clasts within an iron oxide or clay-rich matrix (fig. 9a, c). Cataclastic zones are often foliated (fig. 9c–f). Discrete slip surfaces exhibit sharp boundaries (fig. 9d–e) and can show up to 8 mm of offset.

The majority of slip occurred within the fault core zones, with some slip being accommodated within a broader damage zone. Drill core ASDC 607 shows two distinct clay-rich fault cores, which may indicate a splay or migration of the main fault trace here. ASDC 613, drilled approximately 1 km to the

Figure 9. Drill core through the Baxters Hollow fault from boreholes ASDC 607 and ASDC 613. The location and ID of cores referenced in this figure are available in accompanying dataset 1. Up on each image is up on the core. Panel A, sample of the Baxters Hollow protolith. The igneous rock here presents as a fine-grained, pink to purple volcanic rock. Note the dilation breccia with a red Fe-oxide matrix. Veins of Fe-oxide crosscut the sample. (ASDC 607-522.7). Panel B, sample from the lower conglomerate and quartzite member of the Baraboo Quartzite, showing pebble- to sand-sized clasts. (ASDC 613-90.4). Panel C, brecciated quartzite with localized zones of cataclasite and dark Fe-oxide slip surfaces. (ASDC 607-500). Panel D, sharp contact between quartzite and red Fe-rich slip surface. The Fe-oxide zone is foliated, with elongate clasts of quartz in an Fe-oxide clay matrix. (ASDC 613-92.8). Panel E, sharp slip surfaces of red and dark Fe-oxide. (ASDC 607-502). Panel F, clay-rich sheared sample with foliated fabric. Sample shows folding and discrete slip surfaces. (ASDC 607-500.9). Panel G, clay gouge with foliated fabric. Gouge shows folds and clasts from parent material. (ASDC 607-508).



southwest (fig. 4), shows only 1 clay-rich fault core. A broader fault damage zone <10 m wide is evidenced by brittle deformation features and slip surfaces observed in the rock surrounding the clay-rich fault cores. Cross-cutting relationships of iron oxide-rich and quartz-rich veins indicate complex fluid-rock interactions within the fault zone.

Microstructural observations

Microstructural analysis of the rocks in the drill cores reveal evidence for ductile and brittle deformation (fig. 10). Quartz grains throughout the drill core show deformation lamellae, undulose extinction, and bulging recrystallization, with evidence for some subgrain development (fig. 10a–b) and rotation. These features suggest that the rock experienced ductile strain at elevated temperatures, though all features but the bulging recrystallization could be inherited from detrital sources. Additionally, some of these features may be attributed to the burial and metamorphism of the original rock. Rocks within the shear zones are often strongly foliated, particularly where there is an abundance of phyllosilicate minerals (fig. 10c, f–g), and show kink bands and microfolds. Shear bands show penetrative fabrics and largely follow the same orientation within the same sample.

Brittle deformation features recorded in the fault zone include inter- and intragranular fractures, veins, slip surfaces, fault breccia, cataclasite, and clay gouge. Clasts are often highly fractured. Cross-cutting relationships, such as deformation features offset by discrete faults (fig. 10c, e), are present at the hand sample and thin section scale. The rocks show different stages of brittle deformation in shear zone evolution as described in Hausegger and others (2010). Joint-bounded slices formed at high angles to shear direction and stylolites formed subparallel to shear direction (fig. 10d). With continued shear, the rock was further fractured and brecciated to accommodate strain. Cataclastic flow developed as shear localized (fig. 10e). The presence of implosion breccia, which occurs by the sudden pressure release (opening) of material during rapid slip, indicates coseismic deformation (Sibson, 1986). Reworked clasts of cataclasite within cataclastic zones and cross-cutting relationships of deformation features signify multiple seismic events recorded in the rocks.

The Baxters Hollow Fault is a structurally complex fault zone that shows evidence for multiple cycles of brittle failure and fluid-assisted healing. Fluid inclusions and the cementation of brecciated and fractured rock suggest that the Baxters Hollow Fault experienced fluid-assisted healing between seismic ruptures (fig. 10c-f). The seismic ruptures may have been driven by intermittent high fluid pressure. Veins and the breccia and cataclasite cement consists largely of quartz, Fe-oxide, and phyllosilicates. Stylolites often form in contractional regimes as a result of compression and dissolution of minerals due to fluids in the fault zone. Fluid-induced mineralization and dissolution of mobile elements in the sheared zone resulted in the concentration of Fe-oxide in slip surfaces and sheared material and the alteration of clay minerals.

Extensional Faults

Hemlock Draw Fault

The Hemlock Draw fault strikes N40E and dips approximately 60 degrees SE. It is interpreted as a down-on-the-southeast normal fault. The fault crosses much of the southwestern portion of the South Range. It was first mapped by Stewart and others (2021b) based on an observed shift in map contacts. Stewart and others (2021b) estimated approximately 300 m of displacement across the fault.

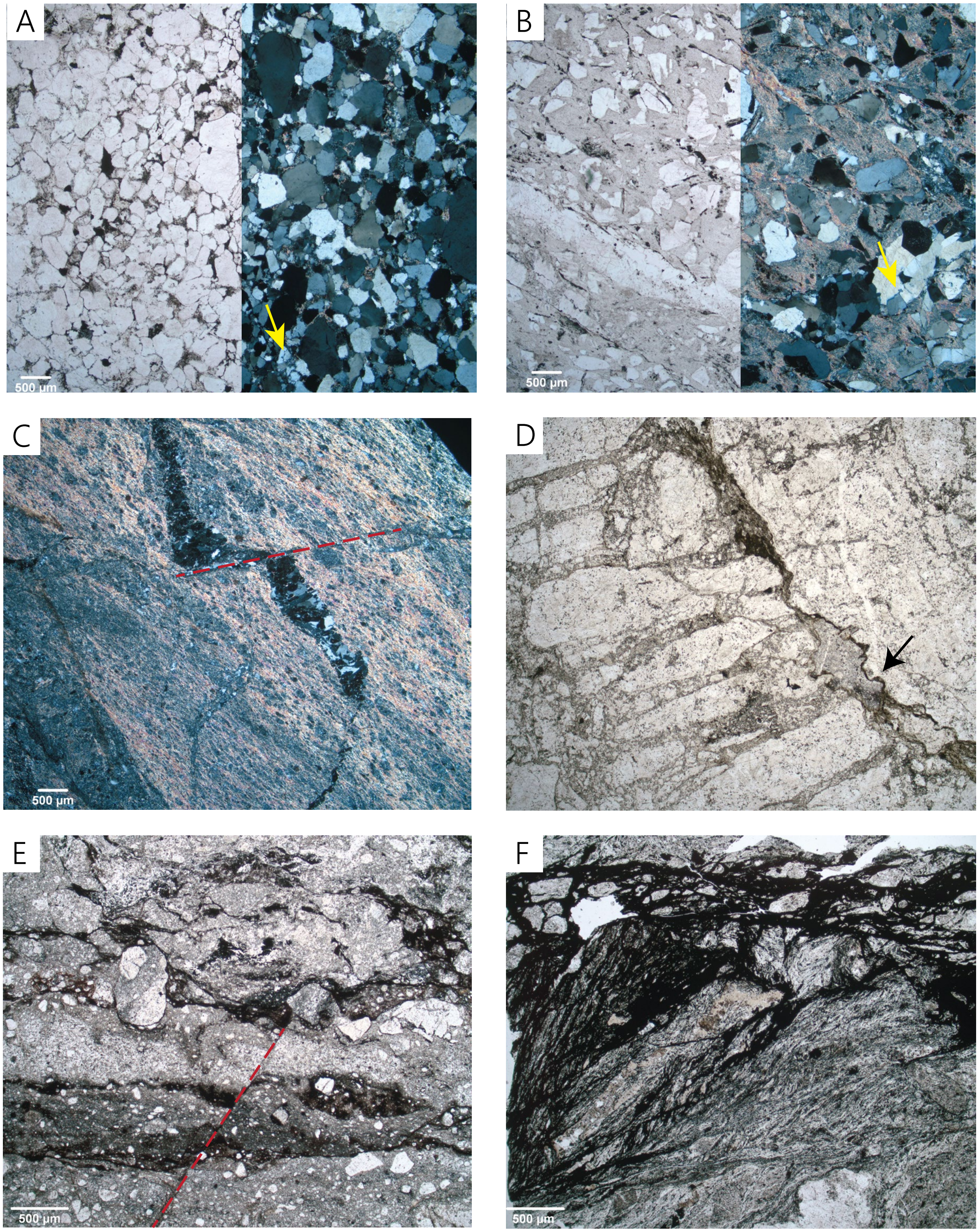
The damage zone of the Hemlock Draw fault is tens of meters wide, and part of it is well exposed in the South Range (43°22'47"N, 89°54'51"W). Small listric normal faults (fig. 11a) and en echelon quartz veins (fig. 11b) indicate down-to-the-southeast kinematics. Slickenlines on fracture surfaces suggest down-dip motion. Quartz veining and fracturing increase in intensity towards the inferred fault core, which does not outcrop. A small dilational breccia is present near the fault trace close to Honey Creek in the far southwestern Baraboo Hills (43°21'50"N, 89°56'38"W). The breccia shows no evidence of cataclasis or slip.

The timing of the fault is unclear. Because it cuts dipping Baraboo Quartzite beds, it probably post-dates the main episode of folding in the area. If fluids that formed the dilational breccia adjacent to the fault were focused through the fault zone, then faulting must have pre-dated breccia formation. Elsewhere in the Baraboo Hills, the dilational breccia has been dated to 1472±3 Ma (Medaris and others, 2021). The occurrence of en echelon quartz veins also suggests elevated temperatures and pressures during faulting.

Narrows Creek Fault

The Narrows Creek fault is a northwest striking fault running along Narrows Creek in the northwestern Baraboo Hills. It was first named by Stewart and others (2021b), but was originally identified by Kemmer and Kovac (1937), who identified and mapped this fault on the basis of a lithology change across Narrows Creek. They interpreted the fault as a left-lateral strike-slip fault on the basis of a counter-clockwise rotation of bedding towards the fault. Later work by Usbug (1968) did not support a fault. Dalziel and Dott (1970) supported the presence of a fault with a left-lateral component of slip based largely on deflected bedding orientations along Narrows Creek. Stewart and others (2021b) applied their revised stratigraphy and found an apparent left-lateral shift in the top of the upper conglomerate member across Narrows Creek, but did not observe a significant rotation in bedding strike.

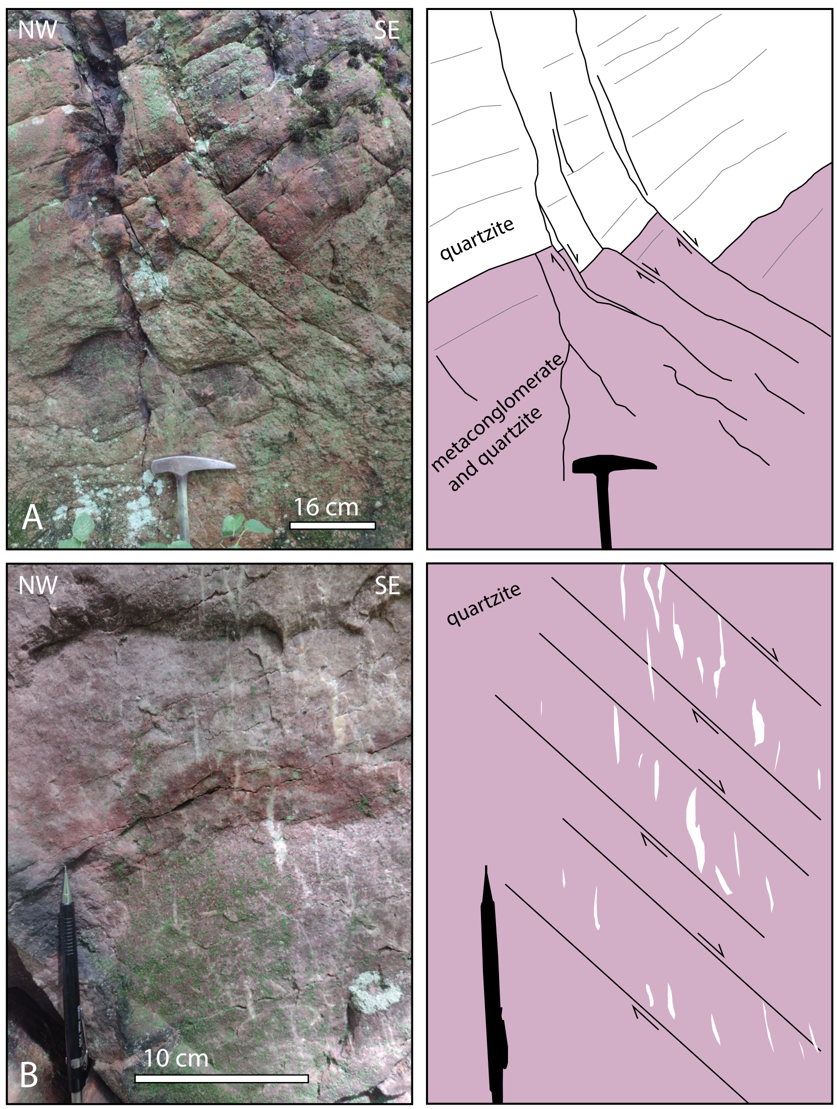
Figure 10.



◀ **Figure 10.** Photomicrographs of thin sections showing evidence for the Baxters Hollow fault. Sections are not oriented. The location and ID of cores referenced in this figure are available in accompanying dataset 1. Panel A, plane (left) and cross (right) polarized photomicrograph of the lower conglomerate and quartzite member of the Baraboo Quartzite. Shows bulging recrystallization of quartz grains (yellow arrow). (Core ASDC 613, 89 ft depth). Panel B, plane (left) and cross (right) polarized photomicrograph of the Baxters Hollow igneous unit. Shows bulging recrystallization of quartz grains (yellow arrow). (Core ASDC 607, 524 ft depth). Panel C, cross polarized photomicrograph showing foliated phyllosilicates from the Baxters Hollow igneous unit. The red dashed line indicates a microfault offsetting a ductile deformed quartz vein. (Core ASDC 607, 509 ft depth). Panel D, joint-bounded slices with secondary fractures through the internal slices. Hausegger et al. (2010) describes this feature as the mid stages of shear zone development. The slices show some rotation and brecciation is visible in the lower left corner of the photomicrograph. Black arrow points to stylolite. Photomicrograph is from the lower conglomerate and quartzite member of the Baraboo Quartzite, and is in plane polarized light. (Core ASDC 607, 490.6 ft depth). Panel E, foliated cataclasite with dark opaque bands of Fe-oxide, also from the lower conglomerate and quartzite member of the Baraboo Quartzite. Cataclasite is cut by discrete fault (red line). Photomicrograph is in plane polarized light. (Core ASDC 607, 490.6 ft depth). Panel F, distinct Fe-rich slip surfaces with fractured clasts composed of quartz, phyllosilicates, and reworked cataclasite. The lighter portion in the lower half largely consists of foliated phyllosilicates with microfolds and some quartz grains. Photomicrograph is in plane polarized light and is located near the contact between the Baraboo Quartzite and the Baxters Hollow igneous unit. (Core ASDC 607, 502 ft depth).

We tentatively reinterpret the Narrows Creek fault as a northwest striking, down-on-the-northeast normal fault rather than a strike-slip fault. This revised interpretation is based on problems with the strike-slip model, and the ability of a normal fault to explain otherwise inconsistent map patterns described below. Specifically, a strike-slip fault interpretation poorly explains the change in dip angle and dip direction across Narrows Creek. On the east side of the fault, bedding flips from steeply overturned to sub-vertical around the contact between the upper conglomerate member and quartzite and phyllite member. Nowhere on the west side of the fault is bedding

Figure 11. Field photos from the damage zone of the Hemlock Draw fault showing normal sense displacement. Panel A, listric normal faults displacing conglomerate beds. Panel B, en echelon veins indicating normal sense kinematics. The field photos show evidence for kinematics on the Hemlock Draw fault; they show why a normal fault was interpreted rather than a thrust or strike-slip fault. These outcrops plus heavy veining and fracturing nearby, combined with a shift in map unit contacts indicate a normal fault.



overturned, despite approximately 475 m of the same upper conglomerate member exposed (identical to the full thickness of the upper conglomerate member at Ablemans Gorge 3 km to the east). A pure left-lateral strike slip fault is unable to explain the lack of overturned bedding on the west side of the Narrows Creek fault because most, if not all, of the upper conglomerate member is present on both sides of the fault. A down-on-the-northeast normal fault rectifies this problem, and can explain an apparent left-lateral shift in the location of the upper conglomerate-quartzite and phyllite contact. Cross section A–A′ (fig. 2), drawn east of Narrows Creek, shows how along the north limb of Ablemans syncline, bedding dip is predicted to change from subvertical at the surface to steeply and eventually moderately south-dipping at depth. Deeper footwall rocks on the southwest side of a normal fault could both shift member contacts in a relative left-lateral sense and change bedding dip direction and magnitude to those observed. Finally, we also note that the distance between Ablemans syncline and Diamond Hill anticline is also predicted to decrease with depth (see cross section A–A′), and a down-on-the-northeast normal fault helps explain the abrupt reduction in the distance between the axial trace of these two folds south of Narrows Creek (see plate 1).

Outcrops around Narrows Creek are heavily fractured (fig. 12a, b) and contain an impressive dilational breccia (fig. 12c), but good kinematic indicators were not observed. Fracture orientations are variable, but populations of northwest striking, northeast-dipping and northwest striking, southwest dipping fractures are common (fig. 12a, b). These populations are probably part of a conjugate fracture set, which is consistent with a vertical maximum principal stress that is typical of extensional settings. Tight quartzite folds and well developed cataclastic zones typical of contractional structures in the Baraboo Hills were not observed. Quartz veins occur in the fault zone outside of the breccia, many of which contain vugs partially filled with well-developed hexagonal quartz crystals (fig. 12d). The dilational breccia is not thought to record any fault movement, but to be related to overpressured hydrothermal fluids (Dalziel and Dott, 1970) that probably rose through the pre-existing fault zone.

The age of the fault is unclear, but the dilational breccia may provide some control. Much like the Hemlock Draw fault, if fluids that formed the dilational breccia adjacent to the fault were focused through the fault zone, then faulting must have pre-dated breccia formation. The dilational breccia elsewhere in the Baraboo Hills has been dated to 1472±3 Ma (Medaris and others, 2021).

Folds

Baraboo syncline

The Baraboo syncline is an east-northeast trending, south verging asymmetric syncline with a vertical to overturned north limb and a shallowly north-dipping south limb. Phyllitic beds within the fold contain a cleavage that is approximately axial planar, while quartzite beds contain a spaced cleavage that is perpendicular to bedding. The axial trace of the fold is concealed by Quaternary and Paleozoic units along most of its length. Where concealed, the trace was mapped using core logs and regional aeromagnetic data following the approach of Stewart and others (2018). In the far east of the Baraboo Hills near Cascade Mountain (43°29'22"N, 89°30'31"W), the Baraboo syncline has a west-southwest plunge that varies from approximately 35 degrees to horizontal (Stewart and others, 2024). In the far western Baraboo Hills near Elder Ridge (43°25'34"N, 89°57'22"W), the fold plunges east-northeast at approximately 25 degrees (Marshak and others, 2016). However, aeromagnetic and drill logs indicate the Baraboo syncline is not a simple doubly plunging fold. Instead, it is composed of two doubly plunging segments that are probably not connected at depth (fig. 13; Stewart and others, 2018). The western segment is interpreted to be a footwall syncline formed in advance of the tipline of the North Freedom fault (fig. 2, section A-A′, fig. 13). The shift between the axial trace of the eastern and western segments suggests the eastern segment formed as a footwall syncline on a separate, unrelated fault. The region where the two folds terminate may be a relay or transfer zone, where displacement is transferred from one fault to another. Alternatively, it may also be possible that the two segments of the Baraboo syncline formed as detachment folds rather than fault propagation folds (Czeck and Ormand, 2007). Ultimately, a better understanding of the linkage zone between the eastern and western segments of the Baraboo syncline is needed before a compete kinematic model of the Baraboo Hills area can be established.

Ablemans syncline

Ablemans syncline was first recognized by Weidman (1904), but Dalziel and Dott (1970) were the first to name the fold. It trends northeast, and is an asymmetric, southeast vergent fold. Phyllitic beds within the fold contain a cleavage that is approximately axial planar, while quartzite beds contain a spaced cleavage that is perpendicular to bedding. Like the Baraboo syncline, the axial trace of the fold is concealed by Quaternary and Paleozoic units along most of its length. The northwestern limb contains bedding that varies from vertical to overturned. The southeastern limb is poorly exposed, but probably contains beds that dip shallowly to the northwest. The fold may be a footwall syncline similar to the western segment of the Baraboo syncline. If so, it would have formed in advance of the tipline of an unnamed, unexposed northeast-striking fault buried just northwest of Narrows Creek and Ablemans Gorge. This favored explanation is depicted in figure 2.

Figure 12. Field photos from the damage zone of the Narrows Creek fault. Good kinematic indicators were not observed. Panel A, heavily fractured quartzite adjacent to the inferred fault. White box shows location of panel B. Panel B, conjugate northwest striking fractures. Panel C, dilational breccia. Panel D, quartz vein with vugs and hexagonal quartz crystals. The Narrows Creek fault was largely inferred based on map patterns, but heavy fracturing as shown in panels A and B are consistent with the presence of a fault.

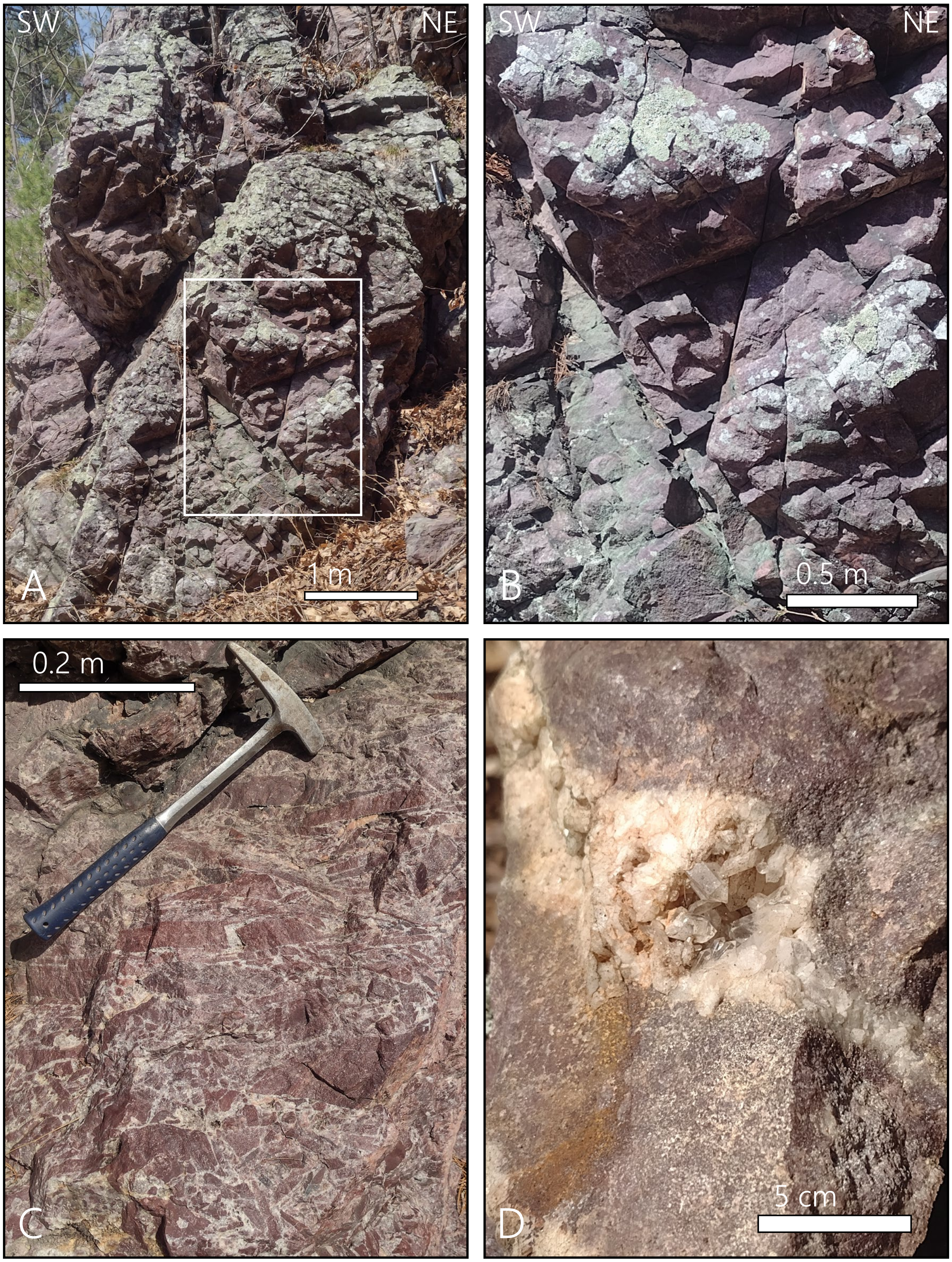
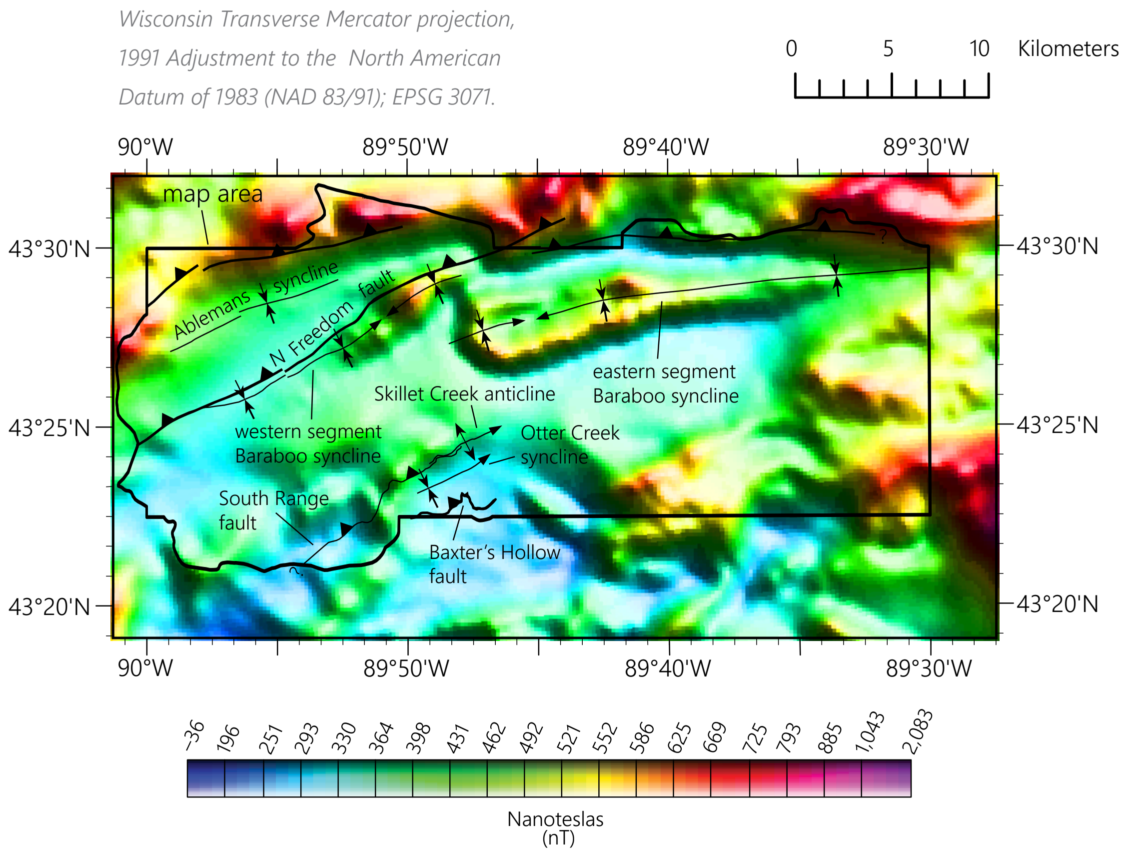


Figure 13. Simplified contractional structure map of the Baraboo Hills, underlain by the aeromagnetic anomaly map of Wisconsin (Daniels and Snyder, 2002). Normal faults and small folds were not included. The anomaly map was modified by a reduced to pole transform (E. Anderson, oral comm, 2015). The high aeromagnetic values along the Baraboo syncline show the subsurface location of the iron-rich Freedom Formation. The subsurface pattern of the Freedom Formation helps identify the location of the two segments of the Baraboo syncline. The origin of the eastern segment of the Baraboo syncline is unclear, and it does not appear to be connected to the western segment. The map also shows the close association of the North Freedom fault and the western segment of the Baraboo syncline, and the South Range fault and the Skillet Creek anticline and Otter Creek syncline. The folds are believed to be fault propagation folds related to the adjacent thrust faults.



Skillet Creek anticline and Otter Creek syncline

The Skillet Creek anticline and Otter Creek syncline were first named and mapped by Dalziel and Dott (1970). However, parts of the Skillet Creek anticline can be seen in the mapping of Leith (1941). Stewart and Stewart (2020) calculated a fold axis that plunges 6 degrees at 068 for the Skillet Creek anticline, and 3 degrees at 067 for the Otter Creek syncline. Both folds decay and end southwestward. The folds are interpreted to represent an anticline-syncline pair formed as a fault propagation fold in advance of the tipline of the South Range fault. Due to the regional plunge of all structural elements into the interior of the Baraboo Hills, deeper portions of this fault-fold pair are exposed to the southwest, and shallower portions to the northeast. This is the favored explanation for why the fold decays southwestward, and why the footwall unit in the South Range fault changes from the lower quartzite member in the northeast, to the lower conglomerate member in the southwest. A–A' (fig. 2) can be used to visualize how the hanging wall and footwall of the fault change with structural level, and how the Skillet Creek anticline appears at higher structural levels and disappears at deeper structural levels.

Discussion

The discussion section focuses on revisions to the Precambrian history of the Baraboo Hills. We first describe the history and recent advances in understanding the stratigraphy of the Baraboo-interval. We include all stratigraphic units from the Baraboo Quartzite through the Dake Quartzite and Rowley Creek Slate within the Baraboo-interval. Next, we review models for the depositional environment of Baraboo-interval strata, ongoing questions about the tectonic setting of Baraboo-interval deposition, and recent advances in the timing and regional correlations stemming from geologic mapping in southern Wisconsin. These stratigraphic advances provide important timing constraints that help feed a revised deformational history for the area, which is discussed in the Precambrian structure section at the end of the discussion.

Precambrian stratigraphy

Depositional environment of Baraboo-interval strata

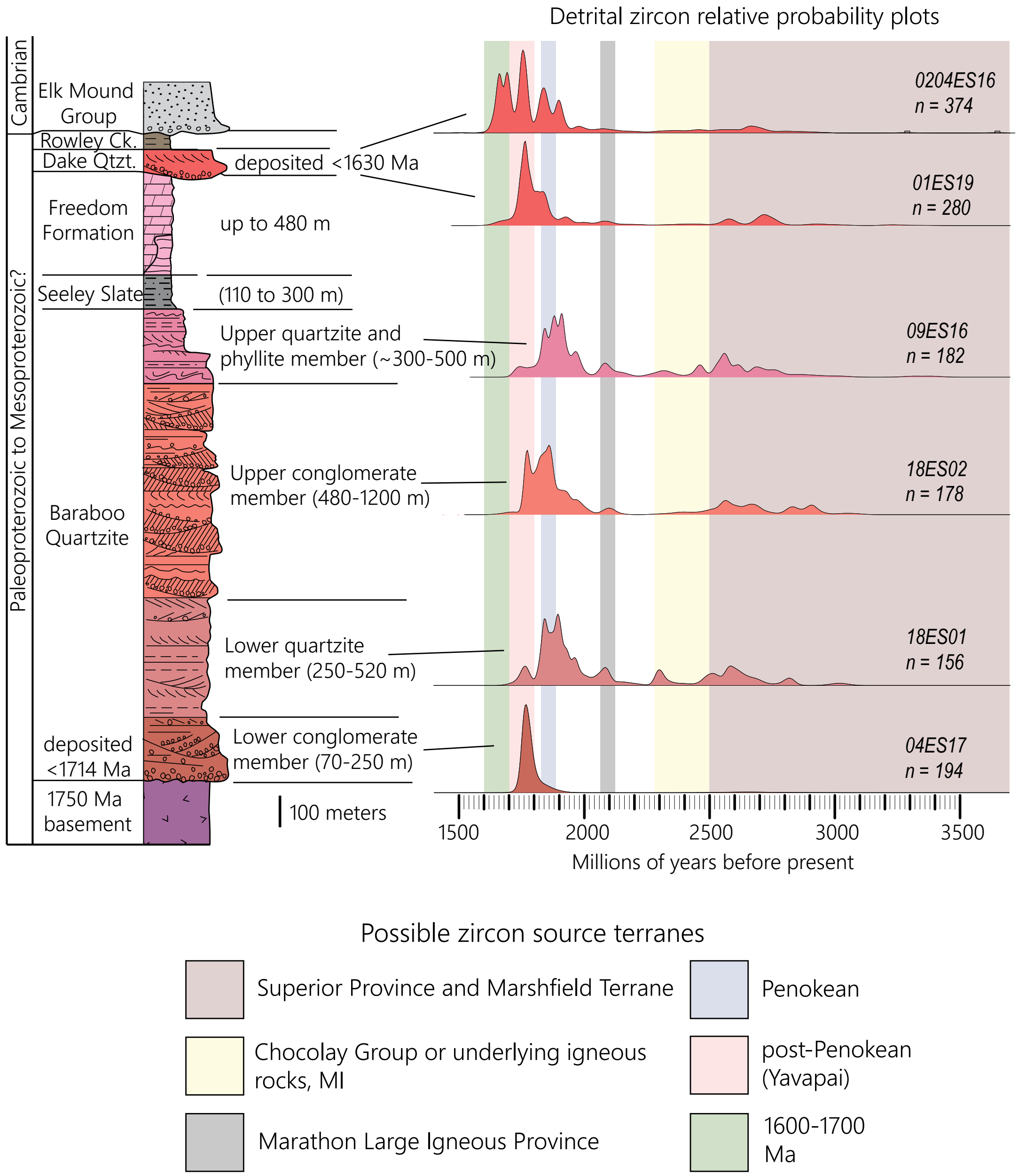
All modern workers agree the Baraboo Quartzite was deposited in both a braided fluvial and marine depositional environment, but successive studies have progressively refined the details of this model. Dott (1983) suspected the majority of the Baraboo Quartzite section was deposited in a braided fluvial setting, with the transition to a marine environment occurring in the uppermost several hundred meters. Davis (2006) emphasized the influence of tides in the upper Baraboo Quartzite based on observations of sedimentary structures. He broke the Baraboo Quartzite into a lower braided fluvial package and an upper tidally influenced marine package. Medaris and others (2011) interpreted the transition from fluvial to marine to occur around the middle of the Baraboo Quartzite section. Stewart and others (2021a) described the lithofacies assemblages that compose four informal members of the Baraboo Quartzite and examined lithofacies stacking patterns to interpret changes in depositional environment. A simplified stratigraphic column showing Baraboo Quartzite members from Stewart and others (2021a) and their associated detrital zircon signatures is given on figure 14. They found the lower conglomerate and quartzite member was deposited in a backstepping coastal setting, with fluvial-deltaic lithofacies overlying fluvial lithofacies. The overlying lower quartzite member was deposited in a coastal dune setting with ephemeral streams and lakes. Above, the upper conglomerate and quartzite member recorded the return of shallow marine conditions. It was deposited in a tidal- and fluvial-influenced delta front environment. At the top of the section, the upper quartzite and phyllite member was interpreted to record a tidal- and wave-influenced delta front to prodelta environment. The overlying Seeley Slate and Freedom Formation were deposited in a setting with minimal evidence for wave action.

Tectonic setting of Baraboo-interval deposition

Dott (1983) suggested the fluvial to marine package of sediments in the Baraboo-interval was deposited in a passive margin setting on the southern edge of Laurentia. He noted a reduction in the maximum clast size southward from the Barron Quartzite to the Baraboo Quartzite, south-directed paleocurrents, and an overall thickening of preserved section southward. All suggested the Baraboo-interval was deposited in a southward thickening wedge of passive margin sediment. Additionally, the lithology and extreme chemical maturity of the Baraboo Quartzite was used to support a passive margin setting for deposition. The complete absence of detrital feldspar in the 2-km-thick Baraboo Quartzite has been used as evidence of a warm, humid, stable continental setting without topographic relief (Dott, 1983; Medaris and others, 2003). Low relief, typical of a passive margin, coupled with a warm climate could allow slow weathering and maturation of the sediment source.

The top of the Baxters Hollow granite beneath the Baraboo Quartzite is a highly altered contact that not only shows evidence for faulting (see description of the Baxters Hollow fault above), but also has been interpreted as a paleosol (e.g. Medaris and others, 2003). Studies on this paleosol are the basis for interpreting a warm, humid, passive margin setting during

Figure 14. Stratigraphic column of the Baraboo-interval with representative detrital zircon relative probability plots for the four members of the Baraboo Quartzite and the Dake Quartzite. Sample numbers are from Stewart and others (2021a). Figure modified from Stewart and others (2021a).



deposition of the Baraboo Quartzite. The paleosol contains a 0.67 m thick upper heavily weathered regolith, and a lower saprolite of weathered granite that is at least 6.1 m thick (Driese and Medaris, 2008). Fe enrichment and depletion features, or redoximorphic features, observed in thin section from the regolith suggest alternating wetting and drying and a mean annual soil temperature between 5 degrees and 20 degrees Celsius (Driese and Medaris, 2008). These temperatures suggest a cool temperate to warm paleoclimate during weathering (Driese and Medaris, 2008). Medaris and others (2017) used weathering characteristics in the paleosol and modified modern empirical relations to estimate a mean annual temperature of 14.3 degrees Celsius during paleosol development, similar to the average annual temperature of Tennessee. Finally, Medaris and others (2022) compared the degree of weathering in the Baraboo paleosol to other Proterozoic, Cambrian, and Cretaceous paleosols as well as 5 modern soils from a variety of climate settings. They found the percentage of SiO_2, CaO, Na_2O, and K_2O removed in the Baraboo paleosol was 22.4%, similar to other Proterozoic paleosols but less than modern paleosols, which averaged 34.2%. The difference was attributed to the lack of land plants in the Precambrian, which produce organic acids that are effective at weathering basement sources. Taken together, these studies showed Proterozoic terrestrial environments were not as effective at weathering and removing feldspars as modern environments and also showed that the climate in the Baraboo area during the Baraboo paleosol formation was warm but not tropical.

Tectonic subsidence calculations represent a test of the passive margin depositional model for the Baraboo-interval. Thick basins require some form of subsidence to create the space needed to deposit sedimentary units. Subsidence can be generated passively by loading from sediments and water, but the amount of subsidence created through this mechanism is limited. Subsidence can also be generated tectonically, either through faulting, lithospheric flexure, or thermally due to cooling such as after rifting (Angevine and others, 1990). Thermal cooling following rifting, combined with passive loading, are the typical sources of subsidence in passive margin sequences (e.g. Watts, 1982). Stewart and others (2018) calculated the necessary tectonic subsidence needed to deposit the Baraboo Quartzite, Seeley Slate, and Freedom Formation using thicknesses of 1372 m, 112 m, and 480 m for each unit, respectively. They found 1280 m of tectonic subsidence was required, and that thermal subsidence and loading alone were insufficient. Their analysis likely underestimated tectonic subsidence for several reasons, including their use of an Airy isostasy model, additional Freedom Formation thickness that has since eroded away, fold-axis parallel stretching (see Precambrian structure section below), and more recent 1:24,000-scale mapping that has shown a better estimate of the actual thickness of the Baraboo Quartzite is between 1900 and 2000 m (Stewart and others, 2021a). These results strongly suggested Baraboo Quartzite deposition did not begin in a passive margin setting. Instead, a tectonically active setting was needed to deposit such a thick section of siliciclastic units. Other Baraboo-interval quartzites may also have been deposited during active tectonism. Southwick and Morey (1984) and Southwick and others (1986) interpreted that deposition of the Sioux Quartzite was controlled by local faults.

These various studies have created a challenging paradox. How could 2 km of supermature sediment in the Baraboo Quartzite deposit in a tectonically active setting? While the tectonic setting could be either a rift or a foreland basin, presumably both would occur with local topographic relief. Local relief is supported by detrital zircon populations near the base of the Baraboo Quartzite that contain largely local igneous sources of sediment (fig. 14; Stewart and others, 2018). Local sources would limit the transport and weathering time. Furthermore, climate was warm but not tropical, and the efficiency of weathering in the Precambrian was lower than modern settings due to the lack of land plants (Medaris and others, 2022). One possible explanation for this paradox was provided by Cox and others (2002). They suggested the higher percentages of quartz arenites and supermature quartz pebble conglomerates in the Precambrian were due to post-deposition alteration and removal of feldspar and other labile clasts. Ultimately, more work is needed to determine the cause of the supermature composition and its role in constraining the tectonic setting of deposition.

Basal Dake quartzite unconformity

Stratigraphically above the Freedom Formation is the Dake Quartzite, and the base of the Dake Quartzite is interpreted to have been deposited over an important angular unconformity (Leith, 1935; Stewart and others, 2018; 2024). Near the town of North Freedom, the Dake Quartzite is deposited over the upper dolomite member of the Freedom Formation, but east of the town of Baraboo the Dake Quartzite is deposited over the lower iron-rich member of the Freedom Formation (Stewart and others, 2018). Cross sections perpendicular to the Baraboo fold axis consistently suggest an angular discordance between the Dake Quartzite and underlying units (e.g. fig. 3; Stewart and others, 2018). The thin, 65 m thickness of the Dake Quartzite combined with historic geologic logs provide control and require some amount of angular discordance.

The amount of missing time across the unconformity is unknown. The maximum depositional age for the Baraboo Quartzite is 1714±17 Ma, and the maximum depositional age for the Dake Quartzite is 1630±9 Ma (Stewart and others, 2021a). While the Dake Quartzite was likely deposited after an initial pulse of folding, it is unclear if it was deposited synchronously during a progressive folding event (such as in a piggyback basin) or if it was deposited between periods of folding and contraction during tectonic quiescence. It is also unclear how deep the Baraboo Quartzite was buried prior to the development of the angular unconformity. It is possible it was not deeply buried and little section was eroded prior to deposition of the Dake Quartzite.

Regional stratigraphic correlations

Freedom Formation

Recent drilling in Dodge County, eastern Wisconsin, discovered an iron-rich metasedimentary Precambrian rock (fig. 15a; Stewart, 2021) that is probably correlative with the Freedom Formation. The mineral assemblage in the Dodge County iron formation includes iron- and magnesium-rich silicates, iron-oxides, and quartz. There are also common carbonate veins and other detrital minerals. The Dodge County mineral assemblage closely resembles the Freedom Formation from the Baraboo Hills (see Freedom Formation unit description, above), and is typical of iron-rich chemical sedimentary rocks. We correlate this new iron formation to the Freedom Formation, but also acknowledge that it is possible this iron formation is a completely new iron formation stratigraphically above the Freedom Formation and Dake Quartzite.

The location of the core directly overlies an elliptical aeromagnetic high (fig. 15b). This elliptical high is continuous with a second aeromagnetic high to the southeast. These elliptical features run roughly parallel to the strike of bedding in the Dodge County iron formation, and parallel to bedding in the nearby Waterloo Quartzite to the southeast (fig. 15b, c). For this reason, the aeromagnetic feature is believed to represent a stratigraphic interval rather than an intrusive body. This enables the aeromagnetic high to act as a marker horizon in the buried Precambrian basement in southeast Wisconsin, much like how member contacts in the Baraboo Quartzite serve as marker horizons in the Baraboo Hills. This helps identify where in the regional stratigraphic section isolated quartzites, such as the Waterloo Quartzite, are located.

Waterloo Quartzite

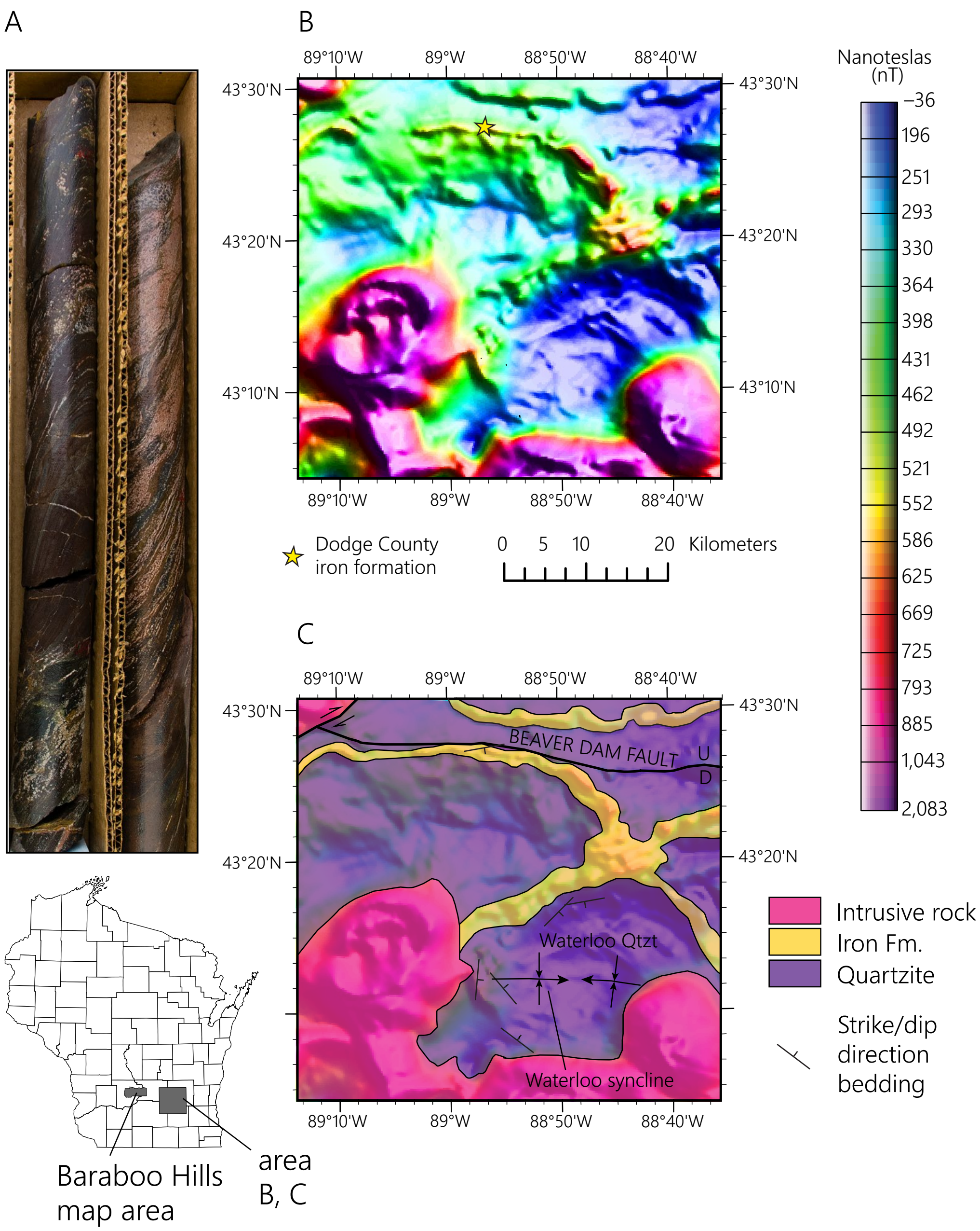
We correlate the lower Waterloo Quartzite to the Dake Quartzite and Rowley Creek Slate (fig. 16) based on aeromagnetic and mapping relations seen in figure 15, and detrital zircon populations. Bedding orientations from scattered outcrops of the Waterloo Quartzite in Dodge and Jefferson counties, southeast Wisconsin, define the east-plunging Waterloo syncline (fig. 15c). The continuation of the same aeromagnetic high observed at the site of the Dodge County iron formation lies stratigraphically below the Waterloo Quartzite. If the Dodge County iron formation is correlative with the Freedom Formation, then the Waterloo Quartzite is younger than the Freedom Formation, and exposed outcrops in the Waterloo syncline are correlative with unexposed and now eroded quartzite stratigraphically above the very thin Dake Quartzite and Rowley Creek Slate. Detrital zircon populations support these map relations. The youngest population of zircons from the Waterloo Quartzite define an age population of 1643±11 Ma (Medaris and others, 2021) and the youngest population of zircons from the Dake Quartzite define an age population of 1630±9 Ma (Stewart and others, 2021a). Malone and others (2022) did a multidimensional scaling analysis on 4082 zircon ages from 23 Baraboo-interval quartzites from across the upper midwest, and they found the two datasets from the Dake and Waterloo quartzites above defined their own grouping independent of all other samples, strongly supporting a correlation between the two units.

Timing of Baraboo-interval deposition

The depositional age of Baraboo-interval metasediments is important because it has been used to constrain tectonic models for south-central Laurentia in the Paleo and Mesoproterozoic. Detrital zircon ages in Medaris and others (2021) found a 1643±11 Ma maximum depositional age for the Waterloo Quartzite in southeast Wisconsin. From this dataset, it was assumed this age reflected the maximum depositional age for not just the Waterloo Quartzite, but the entire Baraboo-interval section. Medaris and others (2021) argued deposition of the entire Baraboo-interval must have occurred either near the end or after the Mazatzal orogeny. In southern Wisconsin, a regional Rb-Sr isochron age of 1635±33 Ma constrains the timing of Mazatzal deformation (Medaris and others, 2003). Based on these depositional constraints, they argued folding and deformation in the quartzites must have taken place after the Mazatzal orogeny for all metasedimentary Baraboo-interval units (Medaris and others, 2021). This provided the grounds for reinterpreting ca. 1.47 Ga anorogenic hydrothermal ages (e.g. Medaris and others, 2003) as 1.47 Ga deformational ages as part of the Baraboo orogeny (Medaris and others, 2021).

Detrital zircon maximum depositional ages are useful tools for constraining the age of deposition, but they only provide partial control. Maximum depositional ages for the Waterloo Quartzite (1643±11 Ma) and Dake Quartzite (1630±9 Ma) indicate deposition of these units could have occurred either during or after the Mazatzal

Figure 15. Panel A, core of iron formation discovered in Dodge County. Panel B, reduced to pole aeromagnetic anomaly map of part of eastern Wisconsin (see inset map on bottom left for location). The star marks the location of the iron formation, which directly overlies an aeromagnetic high. Panel C, geologic interpretation of the aeromagnetic data. Based on bedding orientations in the Waterloo Quartzite, the iron formation must be stratigraphically below the Waterloo Quartzite. The Beaver Dam fault is speculative and based on ideas in Stewart (2021).

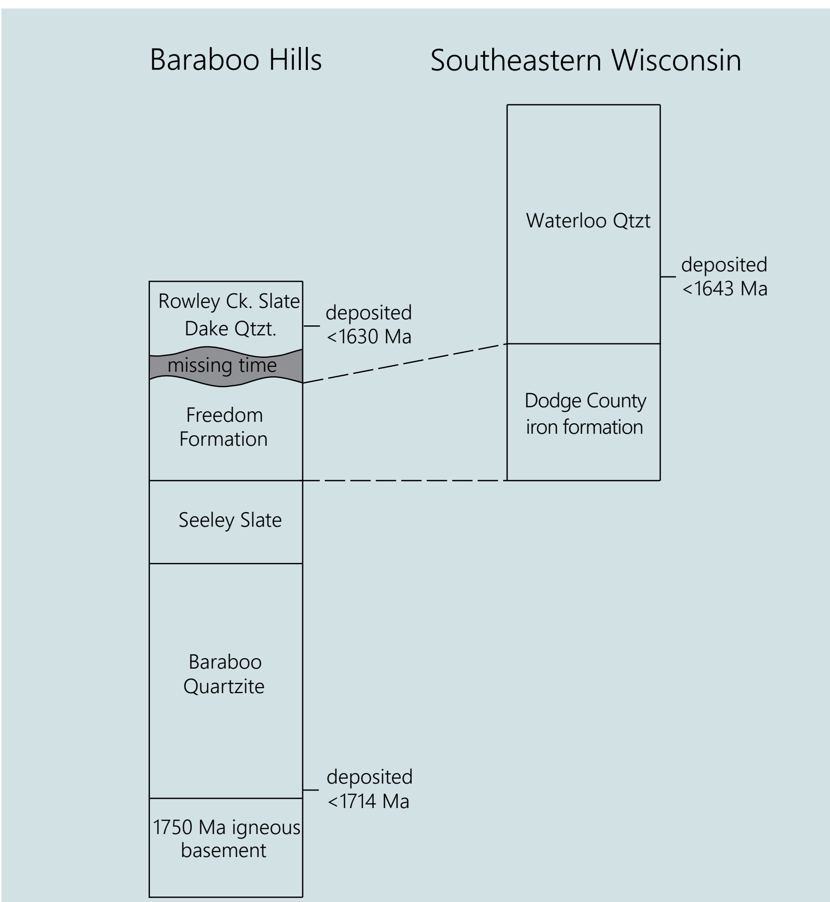


orogeny. The lithostratigraphic placement of the Waterloo Quartzite above the Baraboo Quartzite in the Baraboo-interval section (fig. 16), and the presence of an unconformity below the correlative Dake Quartzite indicate a 1643±11 Ma maximum depositional age should not be used for older strata like the Baraboo Quartzite, which has a maximum depositional age of 1714±17 Ma (Stewart and others, 2021a). This older age does not provide a unique solution to the question of whether the Baraboo Quartzite was deposited before, during, or after the Mazatzal orogeny. Instead, the dataset is consistent with all models. The timing of Baraboo Quartzite deposition is important to deformation models of the area, and is discussed more below in the Precambrian structure portion of the Discussion section.

Precambrian structure

This section begins by restoring the deformed-state cross sections (fig. 17) to their undeformed state and interpreting the significance of the results. Next, the development and importance of lineations in the Baraboo Quartzite is reviewed. Finally, cross sections, map-scale relations and improvements in the stratigraphic framework are used to constrain the timing of deformation and the geometry of the newly mapped fold-thrust belt is used to test existing tectonic models.

Figure 16. **Correlation chart showing proposed correlations between the Freedom Formation in the Baraboo Hills and iron formation in Dodge County, and The Dake Quartzite and Rowley Creek Slate with the base of the Waterloo Quartzite. It is unclear if the angular unconformity beneath the Dake Quartzite continues eastward beneath the Waterloo Quartzite, or if the section becomes conformable in Dodge County.**



Interpretation of deformed and restored cross sections

The km-scale asymmetric Baraboo syncline and the strong cleavage development in the Baraboo Quartzite are often used as evidence for a fold-thrust belt setting for deformation (e.g. Dalziel and Dott, 1970; Craddock and McKiernan, 2007; Czeck and Ormand, 2007; Marshak and others, 2016). Figures 2 and 17 present a subsurface interpretation consistent with this setting, depicting a series of south vergent thrust faults deformed within a tilted “thin-skinned” style thrust belt. Thin-skinned thrust belts are characterized by a subhorizontal basal thrust separating deformed rocks above from undeformed rocks below. The basal detachment or thrust within this thin-skinned system is placed within the igneous basement beneath the Baraboo Quartzite (figs. 2, 14). Detachment surfaces within crystalline basement have been recognized or inferred in many orogenic belts (Lacombe and Mouthereau, 2002; Pfiffner, 2016) despite the traditional view that basal thrusts in thin-skinned systems form at the base of a sedimentary cover sequence. In the case of the Baraboo area, over 1 km of bedded volcanic rocks underlie the Baraboo Quartzite in the northern Baraboo Hills. These igneous rocks may have behaved similar to bedded sedimentary rocks during deformation allowing a deeper basal thrust to form.

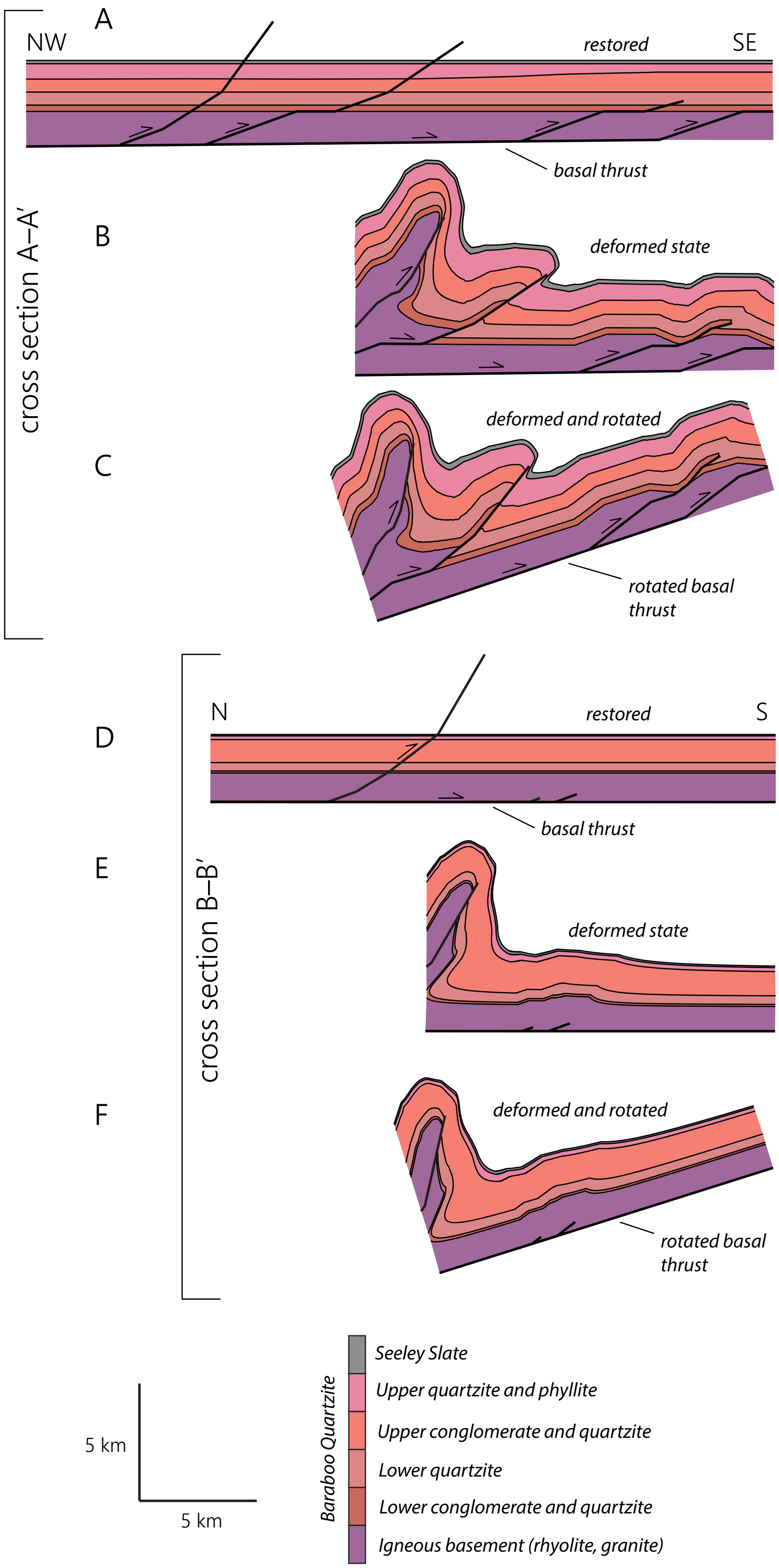


Figure 17. Restored state, deformed state, and rotated deformed state cross sections, respectively for cross sections A–A′ and B–B′ (fig. 2). The restored state cross section removes all folding and faulting, which returns the stratigraphy to flat-lying as it would have been immediately after it was originally deposited.

Restoration of cross section A-A' yields approximately 14 km of horizontal shortening in the deformed-state cross section and B-B' yields approximately 9 km of horizontal shortening (fig. 17a, b, d, e). Following horizontal contraction, both sections include approximately 18 degrees of counter-clockwise rotation (looking northeast and east, respectively; fig. 17c, f). The rotation was required to bring the deformed section into alignment with the surface geology. The rotation causes the southern limb of the Baraboo syncline to dip gently north, which manifests today as the gently north sloping South Range of the Baraboo Hills when viewed from the town of Baraboo, Wisc.. The additional rotation may suggest a deeper ramp within the crystalline basement, or could result from injection of ca. 1470 Ma Wolf River batholith-age intrusions beneath the southern Baraboo Hills (Allen and Hinze, 1992). Shortening and rotation combine to produce 4 to 6 km of vertical exhumation in the igneous basement. However, the cross sections and shortening numbers should be used with caution. Lineations in quartzite are often perpendicular to the line of cross section (see section below), indicating non-plane strain conditions were locally present within the Baraboo Quartzite. Plane strain is a requirement of balanced cross sections, so the shortening numbers and deformed state cross sections may give only a rough estimate of deformation.

Impact of lineation data on kinematic models

Stretching lineation data fit a general history of early distributed strain across all rock lithologies changing to localized strain in weaker phyllites during progressive deformation. Lineations in quartzite generally pitch shallowly within the quartzite spaced cleavage, sub-parallel to the axis of the Baraboo syncline, while lineations in phyllites generally pitch steeply within the phyllitic S_1 cleavage at a high angle to the Baraboo syncline axis (fig. 7). Czeck and Ormand (2007) provide a three-dimensional kinematic model that may explain this change in lineation orientation with lithology. Citing Craddock and McKiernan (2007) and the presence of chocolate-tablet boudinage (stretching and boudinage in two perpendicular directions), Czeck and Ormand (2007) suggested a component of axis parallel extension occurred during the early stages of folding. Axis parallel stretching creates room problems at depth, but they note that this can be overcome by the development of doubly plunging folds like the Baraboo syncline. Once bedding was rotated into a favorable orientation, strain became localized within phyllite beds (flexural flow folding) and along bedding planes (flexural slip folding), S_1 and S_2 cleavages formed during progressive deformation in phyllites, and little if any axis parallel stretching occurred. Due to the localization of strain later in the deformation history, earlier distributed axis parallel stretching was preserved in the spaced cleavage of the quartzite. This model helps explain a rotation in the pitch of stretching lineations from shallow (quartzite) to steep (phyllite). However, there are other possible explanations for the geometry of the fold and the orientation of the lineations. The doubly plunging character of the Baraboo syncline could be related to a change in ramp displacement along strike (Laberge, 1994), and the axis parallel stretching could be caused by the differential propagation of the fold-thrust belt into the foreland region.

The timing of axis parallel stretching relative to depositional age of the Dake Quartzite is partially controlled. Outcrops of Dake Quartzite at Dake Ridge (43°28'52"N, 89°41'17"W) have a strong Baraboo fold axis parallel stretching direction, and outcrops of Dake Quartzite at North Freedom (43°27'23"N, 89°52'7"W) have a subtle fold axis parallel stretching direction (see plate 1). Thus some axis parallel stretching must have occurred after 1630±9 Ma, the depositional age of the Dake Quartzite. However, this does not preclude earlier axis parallel stretching in the Baraboo Quartzite. Cross section construction (fig. 3) shows angular discordance between the Dake Quartzite and underlying units, indicating prior folding before deposition of the Dake Quartzite. It is unclear what the pressure-temperature conditions were for this earliest episode of folding. Microstructural studies comparing deformation conditions in the Dake Quartzite and Baraboo Quartzite may be needed to determine if both units experienced identical peak metamorphic deformational histories.

Timing of deformation

The Baraboo-interval was long thought to have been folded and deformed in the ca. 1.63 Ga Mazatzal orogeny (Dott, 1983; Holm and others, 1998; Craddock and McKiernan, 2007; Czeck and Ormand, 2007). This event reset basement $^{40}Ar/^{39}Ar$ mica geochronometers in northern Wisconsin in areas south of a 1630 Ma cooling front (Holm and others, 1998). Areas south of the cooling front correspond to areas with folded and deformed Baraboo-interval meta-sediments, while areas north of the cooling front contain undeformed Baraboo-interval sediments. Taken together, this was interpreted as evidence that folding of the Baraboo-interval occurred at the same time as the 1630 Ma isotopic resetting event, and was connected to the Mazatzal orogeny of the southwestern United States (Holm and others, 1998; Romano and others, 2000; Holm and others, 2007).

Evidence for this isotopic resetting event also exists in the Baraboo area. Dott and Dalziel (1972) calculated a Rb-Sr isochron age of 1640±40 Ma from rhyolite samples taken around the periphery of the eastern Baraboo Hills. This age was calculated using an older Rb decay constant of $\lambda=1.39 \times 10^{-11}\ yr^{-1}$. A younger age of 1605 Ma is calculated when combining Dott and Dalziel's best fit of the Rb-Sr isochron line with the more commonly used decay constant of $\lambda=1.42 \times 10^{-11}$ yr^{-1}. These same rhyolites were later dated with U-Pb methods by Van Wyck (1995) to approximately 1750 Ma. Medaris and others (2003) combined Rb-Sr data from Van Schmus and others (1975) and Dott and Dalziel (1972), and using the updated $\lambda=1.42 \times 10^{-11}$ yr^{-1}, calculated a regional Rb-Sr isochron age of 1635±33 Ma. Importantly, igneous basement in the Baraboo Hills area show evidence of widespread recrystallization of primary igneous minerals into a greenschist facies assemblage, including biotite to chlorite, plagioclase to albite and epidote, and hornblende to actinolite, chlorite, and cummingtonite (Medaris and others, 2003). Thus, Medaris and others (2003) interpreted the 1635 Ma Rb-Sr age to reflect the age of greenschist facies basement recrystallization, and to correlate to 1630 Ma basement mica cooling ages from northern Wisconsin (Holm and others, 1998; Romano and others, 2000). Finally, Naymark and others (2001) report a 1596±16 Ma $^{40}Ar/^{39}Ar$ age for partially recrystallized hornblende, which may reflect Mazatzal-age deformation. Folding of the Baraboo Quartzite was assumed to have occurred at the same time as basement recrystallization.

Recently, Bjørnerud (2016), Medaris and others (2021), and Marshak and others (2024) proposed folding and deformation of Baraboo-interval rocks occurred approximately 1470 Ma around the time of the Picuris orogeny instead of during the Mazatzal orogeny. Medaris and others (2021) named this event the Baraboo orogeny. Evidence includes axial planar muscovite from the Seeley Slate with $^{40}Ar/^{39}Ar$ ages of 1473±3, 1483±3, and 1493±3 Ma (Medaris and others, 2021). Metamorphic overgrowths on monazite in the Seeley Slate were dated on an electron microprobe using the U-Th total Pb method, with a weighted mean age 1488±20 Ma (Medaris and others, 2021). Additionally, neoblastic hematite within pyrophyllite from a phyllitic bed in the Baraboo Quartzite yielded a 1411±39 Ma (U-Th)/He age. Sheared potassium bearing micas in the South Range fault also suggest the fault was active during the Picuris event (see description of South Range fault above). Medaris and others (2021) argued the maximum depositional age of 1643±11 Ma for the Waterloo Quartzite also represents the maximum depositional age for all Baraboo-interval metasediments, and due to this age, all deformation must post-date the Mazatzal orogeny. However, based on observations presented above, the Waterloo Quartzite is stratigraphically above and younger than the Baraboo Quartzite, and the argument that all deformation must post-date the Mazatzal orogeny is not required.

P-T-t problems with a single ca. 1490-1410 Ma event

Map relations are critical for assessing pressure-temperature-time paths (P-T-t) during folding and deformation of both the Baraboo Quartzite and underlying volcanic rocks, but many of these relations have not been considered in existing tectonic models. Several field relations are particularly important and are highlighted here. First, bedding in the Baraboo Quartzite is parallel to primary flow and depositional banding in the underlying rhyolite (Dalziel and Dott, 1970), so the Baraboo Quartzite was originally deposited nonconformably over horizontally flow-bedded rhyolite. Additionally, near the Lower Narrows of the Baraboo River, the rhyolite contains a N72E/46°NW cleavage close to the axial plane of the Baraboo syncline, and the intersection lineation between rhyolite depositional layering and cleavage is parallel to the Baraboo syncline fold axis (LaBerge and others, 1991). All these data suggest that the rhyolite was not deformed prior to the Baraboo Quartzite, and that both units were deformed only by post-Baraboo Quartzite deformation.

Considering the map relations described above, the P-T-t path for rhyolite underlying the Baraboo Quartzite is difficult to explain if the Baraboo Quartzite was deposited after the Mazatzal orogeny and only deformed during the Picuris orogeny (fig. 18a; Medaris and others, 2021). Assuming an age of 1605 Ma represents the age of greenschist facies recrystallization in the rhyolite (recalculated from Dalziel and Dott, 1972, using the updated Rb decay constant of $\lambda=1.42 \times 10^{-11}$ yr^{-1}), the underlying rhyolite must have been buried to upper mid-crustal levels and recrystallized under lower greenschist facies conditions (resetting the Rb-Sr clock, fig. 18a), then exhumed back to the surface all without being rotated or folded (fig. 18a). Mazatzal-age burial, recrystallization, and exhumation could not have resulted in any rotation of the rhyolites, because the Baraboo Quartzite was deposited nonconformably over flat-lying rhyolites (fig. 18a). Such a Mazatzal-age P-T-t path seems highly fortuitous and unlikely. It is more likely the rhyolite stayed near the surface after eruption and emplacement and was buried by the same basin-forming event that resulted in deposition of the Baraboo Quartzite and Dake Quartzite (fig. 18b). In this case both the Baraboo Quartzite and the underlying igneous suite were metamorphosed at greenschist facies conditions approximately 1605 Ma (fig. 18b). Younger 1490-1470 Ma $^{40}Ar/^{39}Ar$ ages associated with the Picuris orogeny may reflect the lower closure temperatures of the $^{40}Ar/^{39}Ar$ system.

It is also unlikely all deformation of the Baraboo Quartzite occurred during the ca. 1490-1410 Ma Picuris event in the upper Midwest, though we note the ages ranging from 1493–1473 Ma for the axial planar cleavage in the Seeley Slate represents strong evidence for some deformation. If all recrystallization and greenschist facies deformation in the Baraboo Quartzite occurred during the Picuris orogeny, then this event must have selectively recrys-

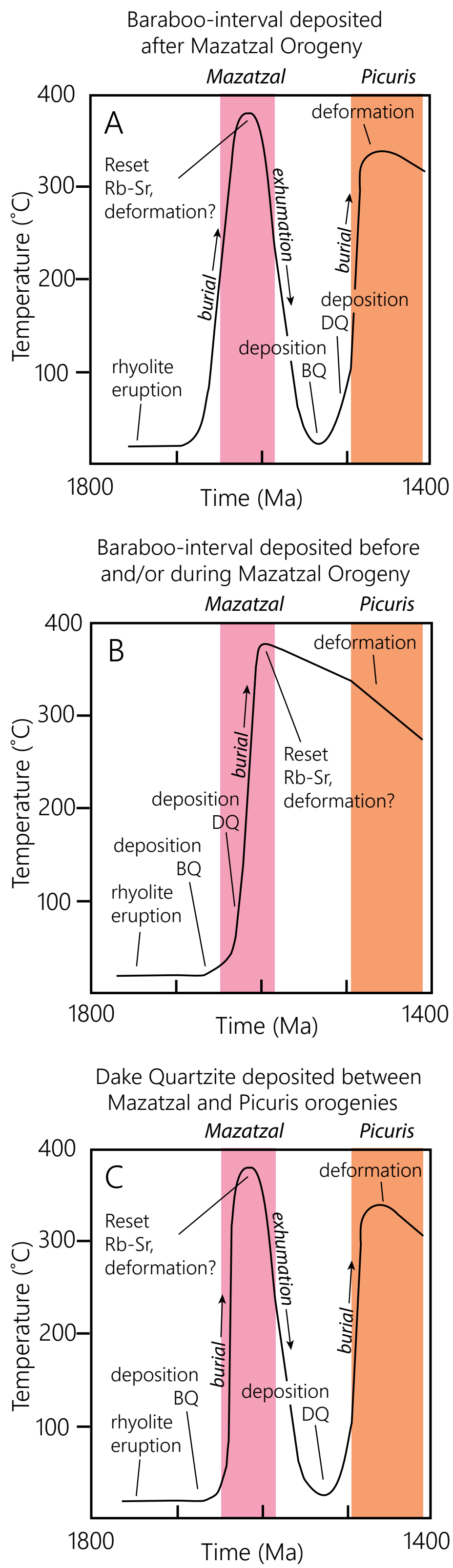


Figure 18. Conceptual temperature-time plots for igneous basement in the Baraboo Hills area. Panel A, Baraboo-interval deposited after Mazatzal orogeny (Medaris and others, 2021). This model requires exhumation of basement rhyolite to the surface following resetting of the Rb-Sr system. No rotation could have occurred in the rhyolite during Mazatzal burial and subsequent exhumation because the younger Baraboo Quartzite was deposited over horizontally bedded rhyolite. Panel B, preferred model where the Baraboo-interval is deposited before and/or during the Mazatzal orogeny. Panel C, Baraboo Quartzite deposited before or during Mazatzal orogeny, but Dake Quartzite deposited after Mazatzal orogeny. BQ represents Baraboo Quartzite. DQ represents Dake Quartzite plus Rowley Creek Slate.

tallized only the Baraboo Quartzite and not the underlying rhyolite, which contains a 1605 Ma whole-rock Rb-Sr age (recalculated from Dott and Dalziel, 1972). Yet the rhyolite was caught up in the same fold-fault event as the quartzite (fig. 2), and contains a cleavage parallel to the axial plane of the Baraboo syncline (LaBerge and others, 1991) suggesting recrystallization associated with folding.

Evidence for two orogenic events

We suggest the Baraboo-interval in the Baraboo Hills was impacted by both the Mazatzal and Picuris orogenic events (fig. 18b) rather than only the Picuris (fig. 18a). The Baraboo-interval contains evidence for a complicated and prolonged history of sedimentation and deformation, and rocks in the area contain two distinct geochronologic ages. Additionally, the hypothesis that the Baraboo-interval was deposited after the Mazatzal orogeny (Medaris and others, 2021; fig. 18a) requires relatively rapid exhumation of igneous basement from upper mid-crustal levels without rotation. Such a scenario is unlikely. Based on these clues, we suggest the Baraboo-interval was involved in two orogenic events rather than one. A brief, preferred possible history is described below, and graphically illustrated in figure 18b.

First, the Baraboo Quartzite, Seeley Slate, and Freedom Formation were deposited nonconformably over undeformed 1750 Ma rhyolite, granite and diorite after 1714±17 Ma (fig. 18b). Minor folding during the early stages of the Mazatzal orogeny was followed by deposition of the Dake Quartzite and Rowley Creek Slate over an angular unconformity with the Freedom Formation. This section was subsequently buried during the later stages of the Mazatzal orogeny to greenschist facies conditions around 1605 Ma, and the Rb-Sr clock was reset in the igneous basement (fig. 18b).

During the Picuris orogeny, contractional deformation continued (fig. 18b). Faulting during the Picuris may have reactivated earlier Mazatzal faults or initiated new faults. Sheared potassium-bearing micas in the South Range fault support active thrusting during the Picuris orogeny. An axial planar muscovite cleavage formed in the Seeley Slate (Medaris and others, 2021). Deformation became increasingly localized, as established faults with thick cataclastic cores accommodated the bulk of deformation. This may explain why basement rhyolites retained a whole-rock Rb-Sr isochron age of approximately 1605 Ma despite later deformation during the Picuris orogeny. Alternatively, a higher closure temperature in the whole-rock Rb-Sr system compared to the $^{40}Ar/^{39}Ar$ system in muscovite may have allowed the earlier Rb-Sr ages to be preserved in the igneous basement.

An alternative but less preferred history is depicted in figure 18c. It is possible the basement, Baraboo Quartzite, Seeley Slate, and Freedom Formation were gently folded and buried to greenschist facies conditions during the Mazatzal orogeny, and later were exhumed back to the surface prior to the Picuris orogeny (fig. 18c). Following deposition of the Dake Quartzite and Rowley Creek Slate over an angular unconformity, the entire section was again buried and deformed during the Picuris orogeny (fig. 18c). This is less favored because it requires a more complicated double sequence of burial and exhumation. Ultimately, better age control is needed to identify a unique P-T-t history.

Vergence of the Baraboo thrust belt, and its impact on tectonic models

The vergence, or transport direction, of thrust faults in a fold-thrust belt provides constraints on the architecture of an orogenic belt. Most thrusts verge away from the hinterland, or central core of an orogen. However, isolated backthrusts can also develop with vergence opposite the main belt. Thrusts exhume deeper, older rocks in their hanging walls, with the effect that older, deeper rocks become exposed in the hinterland, and younger rocks are progressively exposed towards the foreland in the direction of vergence. Thrust belts develop on both sides of an orogenic core, often with characteristic geometries that taper away from the hinterland in both directions, resembling wedges in cross section.

The architecture of an initial Mazatzal orogenic belt that included the folded and faulted Baraboo Quartzite is unclear. Numerous models have been proposed. Craddock and McKiernan (2007) found finite strain in Baraboo-interval quartzites decreases northward from the Baraboo Quartzite, similar to the observations of Dott (1983). They suggested this is most consistent with a convergent margin located south of the Baraboo Hills, and the northward reduction in finite strain was probably related to an overall north-directed fold-thrust belt. They assumed deformation in the quartzites occurred during the Mazatzal orogeny, and was most likely driven by a north-dipping subduction zone occurring well south of the Baraboo Hills. The Baraboo syncline would have formed beneath a backthrust in such a scenario.

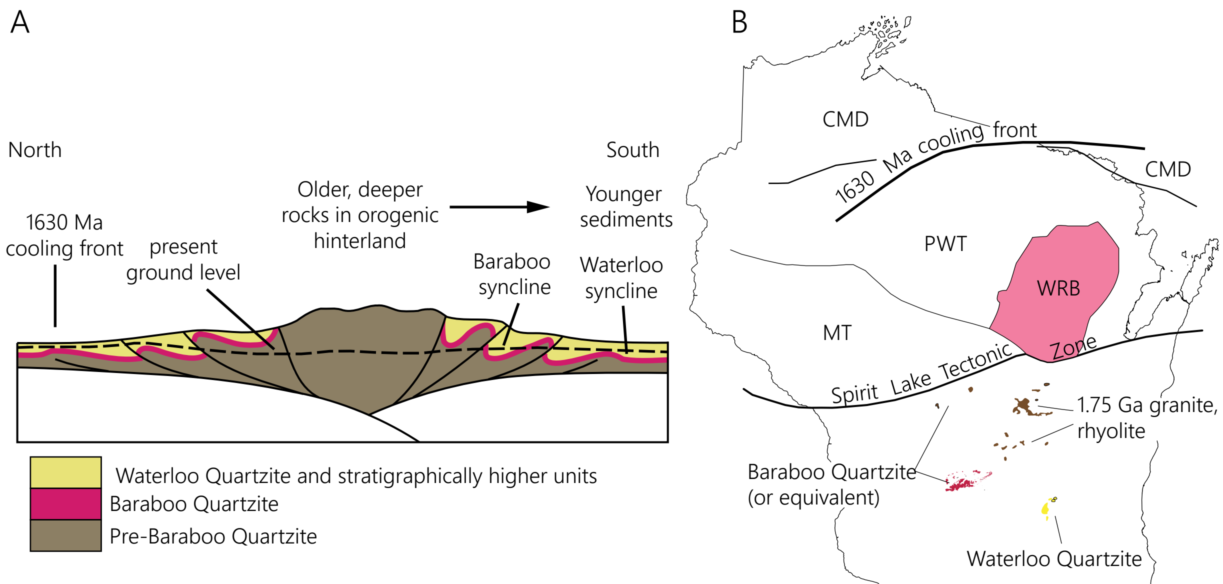
Holm and others (2005), Czeck and Ormand (2007), and Van Schmus and others (2007) also suggested north-directed subduction occurred during the Mazatzal orogeny, but instead subduction and arc accretion drove the development of a southeast-directed fold-thrust belt that included the Baraboo syncline. In this case, the vergence of the folds in the Baraboo Quartzite would be in the same direction as the larger thrust belt. However, any Mazatzal-age volcanic arc must be located south of the Baraboo Hills (NICE working group, 2007). With a north-directed subduction zone and volcanic arc to the south, this means the folded Baraboo rocks occur on the retroarc side of the Mazatzal arc. The southeast vergence of the fold-thrust belt would be in the direction of the hypothesized sutured volcanic arc and orogenic core, opposite of the typical transport direction in orogenic belts.

Marshak and others (2016; 2024) suggested a bivergent thrust wedge centered around the Spirit Lake Tectonic Zone in central Wisconsin (fig. 19) could explain both the southeastward vergence of the Baraboo syncline and the reduction in finite strain in Baraboo-interval rocks in the far north. They recognized that the core of the orogenic belt did not need to be connected to a subduction zone boundary, but instead could be caused by contraction centered on older rift deposits (Marshak and others, 2016). Stewart and others (2018) independently suggested an active rift or extensional setting for Baraboo-interval deposition. Marshak and others (2016) placed the 1630 Ma Mazatzal-age cooling front at the northern edge of the bivergent wedge, but the model could also apply to a Picuris event. (e.g. Marshak and others, 2024).

The architecture of a later Picuris-only event is also unclear. Medaris and others (2021) suggested evidence for contractional deformation occurred between 1493 and 1411 Ma. This time interval overlaps with the undeformed Wolf River batholith. Daniel and others (2023) suggested the Wolf River batholith formed in a back arc setting, and rapid switching from extension to contraction led to the development of folds and thrusts more or less coeval with regional magmatism and undeformed plutons. The relationships between thrust-belt vergence, back-arc rifting, and the undeformed Wolf River batholith were not developed in these papers, and how this scenario is possible is unclear.

We conclude that cross sections across the Baraboo Hills and regional map patterns across central and southern Wisconsin are most consistent with the bivergent wedge model centered on the Spirit Lake Tectonic Zone (fig. 19a) during the Mazatzal orogeny, as originally suggested by Marshak and others (2016). Burial of the Baraboo-interval section in the Baraboo Hills area is interpreted to have occurred during the Mazatzal orogeny. Early folding and thrusting may also have occurred. Early folding could be responsible for the angular unconformity between the Dake Quartzite and Freedom Formation. During the Picuris orogeny, continued contractional deformation occurred. Cross sections across the Baraboo Hills (fig. 2) suggest the North Freedom fault, South Range fault, and Baxters Hollow fault all verge southeastward, making an isolated backthrust unlikely. Instead, it is likely the folds and faults in the Baraboo Hills are part of a larger southeast vergent fold-thrust belt (fig. 19a). The southeastward vergence implies an orogenic hinterland to the north, potentially at the Spirit Lake Tectonic Zone. Additionally, regional stratigraphic correlations show a general younging-to-the-southeast pattern between the Spirit Lake Tectonic Zone and outcrops of Waterloo Quartzite, consistent with this model (fig. 19b). South of the Spirit Lake Tectonic Zone, granitic rocks of the Montello batholith are exposed along with small, isolated occurrences of Baraboo-interval quartzites. In the Baraboo area, overall younger rocks are exposed, such as a thick section of Baraboo quartzite and extrusive rhyolite. Farther southeast, the youngest rocks are present, where the Waterloo Quartzite is exposed, and older 1750 Ma rhyolites and granites are deeply buried and not near the Precambrian-Cambrian unconformity surface.

Figure 19. **Panel A, schematic cross section of the bivergent wedge model modified from Marshak and others (2016). Older, deeper rocks are exposed in the central core of an orogenic belt, and younger rocks are found in foreland regions. Panel B, normal (southeast) to the Spirit Lake Tectonic Zone, older and deeper 1.75 Ga intrusive rocks broadly give way to stratigraphically higher sedimentary units like the Baraboo Quartzite and Waterloo Quartzite. This map-scale pattern of younging-to-the-southeast resembles the predicated pattern in a bivergent wedge centered near the Spirit Lake Tectonic Zone, and agrees with the southeastward vergence of thrusts in the Baraboo Hills. Abbreviations are MT=Marshfield terrane, PWT=Pembine-Wausau terrane, CMD=continental margin domain, WRB=Wolf River batholith. Constructed from Holm (1998), NICE working group (2007), Mudrey and Brown, (1982), and this map.**



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