What's New at Baraboo? A Field Trip For Educators

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ABSTRACT

The Baraboo Hills are an exceptional geologic resource. Dozens of universities send students on field trips to this region every year, taking advantage of world-class outcrops illustrating fundamental concepts, such as the use of bedding/cleavage relationships to define large-scale folds. Despite, or perhaps because of, such consistent attention, the geologic community has continued to learn from this locale.

This field trip showcases outcrops that illustrate well established, foundational relationships, but also highlights data illustrating more recent and less well known contributions to understanding the Proterozoic history of the area. The latter include both published and unpublished research on easily accessed sites. The trip is a time machine, elucidating tectonic and climatic controls on sedimentary structures as well as tectonic and rheologic controls on secondary structures. Evidence regarding deposition of the demonstrably supermature protolith to the Baraboo Quartzite; the history and kinematics of formation of the Baraboo Syncline; quartz shape preferred orientation and cleavage as a record of strain and rheology; the development of crenulation cleavage and associated F_2 folds; microearthquakes in both steeply dipping strike-slip and bed-parallel faults; hydrothermal fluids and brecciation; and details of the Cambrian unconformity that truncates the Proterozoic record – Baraboo has it all.

INTRODUCTION

This guidebook is designed to provide college educators and other interested individuals with updated descriptions and interpretations of the geologic record of the Baraboo Hills, Wisconsin. The majority of college classes that visit this area are focused on structural geology; hence, secondary structural features, which are Proterozoic in age, are emphasized in this guide. Secondary structures, however, cannot be fully understood without consideration of the larger geologic context. We therefore address not only the field record of the distribution, character, and relative ages of relevant lithologic units and both sedimentary and tectonic structures, but also the results of laboratory studies that provide numerical ages and geochemical data. Collectively, these data constrain variations in temperature, pressure, and fluid-rock interactions over time, and attest to the critical influence of both climate and tectonics in writing the record we now seek to read.

Complete consideration of this record would take longer than our one-day field trip. We therefore refer the reader to Medaris et al. (2011) for a wider array of possible field trip stops and a more comprehensive synthesis of petrologic, geochemical, and isotopic studies of the Proterozoic rocks as well as details of the Paleozoic sedimentary record and its significance. Their useful appendix also summarizes the history of geological investigations in the Baraboo District. Both this earlier guidebook and Dalziel and Dott's (1970) monograph and summary maps are critical resources, to which we will refer repeatedly in our guidebook.

The following sections provide the type of brief introductory framework a student could use to scaffold information acquired in outcrop. We build on this framework with summaries of each stop, in which laboratory data and interpretations follow descriptions of features visible in the field. We begin at the geologic beginning for Baraboo, with the Proterozoic growth of the North American continent.

REGIONAL TECTONIC SETTING

The North American continent is a geologic collage. The history of its growth and stabilization is written in the rocks and detailed in reviews by Hoffman (1988) and Whitmeyer and Karlstrom (2007), which together reference numerous studies that have allowed us to better understand the geologic history of our home turf. These studies show that the core of our continent consists of >2.5 Ga Archean provinces that were brought together by plate collisions between 2.0 and 1.8 Ga. The boundaries between these provinces are marked by zones that include reworked Archean crust, Proterozoic juvenile arcs and/or orogenic belts, or different combinations of these elements. Subsequent accretion of tectonostratigraphic terranes built the continent outward from this core. Timing of amalgamation is recorded in part by intrusion of stitching plutons and deposition of overlap sequences of sediments across tectonic boundaries.

Wisconsin includes a record of rocks and structures that is remarkably representative of these larger scale processes, with fragments of Archean crust; juvenile arc material accreted during the ca. 1.875-1.835 Ga Penokean orogeny (e.g., Van Schmus, 1976); intrusion of the 1.8 and 1.775 Ga 'stitching plutons' (Holm et al., 2005); and both intrusion of the 1.75 Ga Montello batholith, believed to be part of the Yavapai event and accretion of juvenile crust during the 1.71-1.68 Ga Yavapai orogeny (Holm et al., 2005, 2007; **Fig. 1**). A significant nonconformity separates these terranes from overlying quartz arenites, with present-day thicknesses that collectively record the regional character of this overlap sequence: a once southward-thickening wedge deposited over an area of at least 300,000 km^2 in the southern Lake Superior region (Medaris et al., 2011). Several observations suggest deposition of these sands on a stable craton of subdued topographic relief in a warm, humid climate with an oxidizing atmosphere. First, mature, feldspar-free paleosols have been documented beneath the Sioux quartzite (on Archean gneiss), the Barron quartzite (on Penokean metatonalite), and the Baraboo Quartzite (on Yavapai-age granite and rhyolite) (**Fig. 1**; Medaris et al., 2003; Driese and Medaris, 2008). Second, these are red, supermature quartzites that lack feldspar but include a suite of durable accessory phases, notably ca. 1.7 Ga and older detrital zircons (Holm et al., 1998; Medaris et al., 2003; Van Wyck, 1995; Van Wyck and Norman, 2004; Wartman et al., 2007) and hematite. These quartzites belong to the *Baraboo Interval*, a distinctive episode of weathering and sedimentation between ca. 1750 and 1630 Ma (Dott, 1983; Medaris et al., 2003), and thus are among the oldest of the world's red beds, which first appeared in the geological record ca. 2.4 Ga (Schröder et al.,

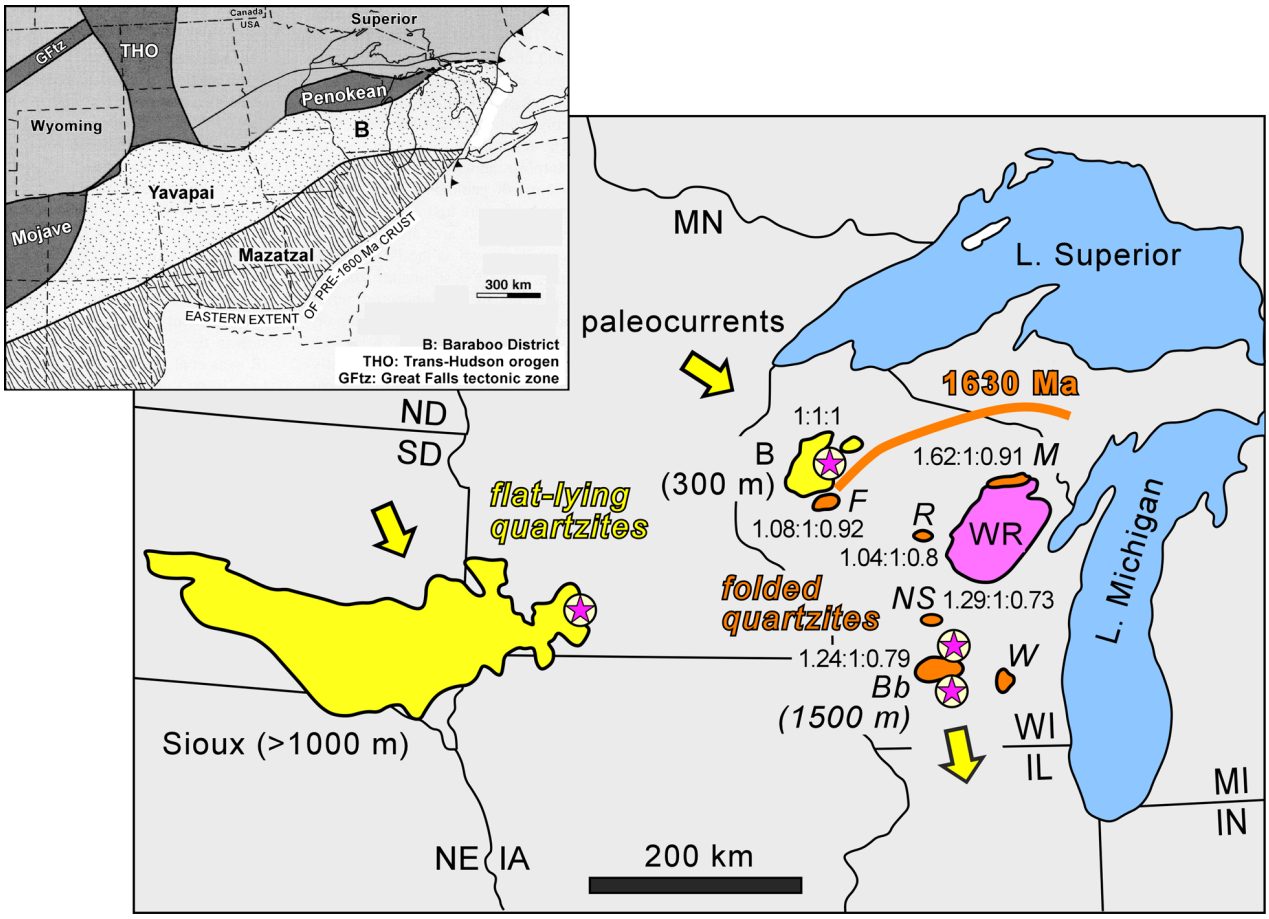

Figure 1. Upper left inset, modified from Holm et al. (2007), shows location of the Baraboo District with respect to Archean and Paleoproterozoic provinces of the North American midcontinent. Larger map, from Medaris et al. (2011), shows locations of paleosols (pink stars) on which Baraboo Interval quartz arenites were deposited. The latter include the Barron (B), Baraboo (Bb), Flambeau (F), McCaslin (M), Necedah and Seven Sisters (NS), Rib Mountain (R), and Waterloo (W) quartzites. Yellow arrows showing paleocurrent directions and present day thicknesses of quartzite sequences record thickening to the south-southwest. Orange line marks the position of the 1630 Ma (Mazatzal) thermal front, which appears to coincide with the boundaries between both flat-lying (yellow) and folded (orange) quartzites and equant and elliptical quartz grains. The latter, shown on this map by axial ratios, record strain broadly increasing to the south (Craddock and McKiernan, 2007). Outcrop of the Wolf River batholith, a ca. 1.4 Ga 'stitching pluton', is marked WR and highlighted in pink.

2011). Finally, metapelite in the Baraboo Quartzite, like fine-grained sedimentary rocks in the Barron and Sioux Quartzites, was derived from kaolinite-rich mud tone and exhibits extreme compositional maturity, with exceptionally low concentrations of K_2O, Na_2O, CaO, MgO, and MnO (Medaris et al., 2011).

Wisconsin bedrock also records the inboard affects of the 1.65-1.60 Ga Mazatzal orogeny, which added 1.70-1.6 Ga juvenile crust to the continent southeast of the Yavapai Province (**Fig. 1**; Karlstrom and Bowring, 1993; Holm et al., 1998; Shaw and Karlstrom, 1999). Specifically, a well defined thermal front in pre-Baraboo Interval basement separates post-Penokean $^{40}Ar/^{39}Ar$ and Rb/Sr cooling ages of 1750-1700 Ma to the north from younger, reset ages of ≤ 1630 Ma to the south (Holm et al., 1998; Romano et al., 2000). Two independent observations suggest that deformation of Baraboo Interval quartzites accompanied this heating: (1) Quartzites north and west of the thermal front are flat-lying, whereas those to the south are folded. (2) Quartz grain shapes north of the thermal front are equant; those to the south record strain, which generally increases with proximity to the Mazatzal Province (Craddock and McKiernan, 2007). Thus, although there is no direct isotopic evidence for the Mazatzal orogeny in Baraboo Interval quartzites, most

workers interpret deformation and metamorphism in these rocks as a record of foreland crustal shortening.

There is direct evidence for the impact of peraluminous to metaluminous, A-type granitoid magmatism, now known to have extended across the continent, that was first recognized in Wisconsin (Anderson et al., 1980; Anderson, 1983). The 1.47 Ga Hager granite of the Wolf River batholith intruded the McCaslin Quartzite; Wolf-River-type granitic dikes cut the Waterloo Quartzite; and the 1.522 Ga syenite of the Wausau Complex cut the Rib Mountain Quartzite (Dewane and Van Schmus, 2007). Hydrothermal alteration and K-metasomatism of broadly Wolf River age were more pervasive, as shown by $^{40}Ar/^{39}Ar$ dates on muscovite (Medaris et al., 2003; **Fig. 2**). In the Baraboo District, muscovite records the localization of fluids along veins and breccia zones in the Baraboo Quartzite, as well as along the apparently relatively permeable paleosol on which the quartzite was deposited. Similar ages are given by muscovite in the Seeley Slate, which overlies the quartzite. Although slates typically exhibit low permeability, this record suggests that the Seeley Slate might have been more permeable than either overlying Freedom Formation banded chert and dolomite or underlying quartzite. Evidence of K-metasomatism has also been documented in the Waterloo and Sioux quartzites as well as in Canada's Athabasca basin, suggesting that Late Paleoproterozoic sedimentary sequences provided pathways for hydrothermal fluids that were propelled far beyond their magmatic sources.

THE BARABOO DISTRICT

Stratigraphy

The Baraboo Quartzite, the stratigraphic unit on which we will focus much of our attention, was deposited nonconformably on diorite near the town of Denzer, granite in Baxter Hollow, and rhyolitic lavas and pyroclastic rocks beneath both limbs of the syncline in the eastern part of the district (Dalziel and Dott, 1970; Medaris et al., 2011). U-Pb zircon ages of the Baxter Hollow granite and rhyolite are the same, within error, at 1749 ± 12 Ma (Van Wyck, 1995). These igneous rocks exhibit evidence for pervasive chemical re-equilibration at greenschist facies conditions, although igneous textures are well preserved.

The 1500 m thick Baraboo Quartzite, like the igneous rocks on which it was deposited, exhibits evidence of relatively low-grade metamorphism. Unlike the igneous basement, it is not pervasively recrystallized, and primary structures are well preserved. The protolith to the quartzite was 85-90% quartz arenite and subordinate quartz wacke, characterized by prominent cross bedding and ripple marks (**Fig.2**). Thin layers and lenses of metaconglomerate constitute 5-10% of the formation. Pebbles are mostly white quartz and rare jasper, all less than 3 cm in diameter. The remainder of the formation is siltstone, now phyllite, and mudstone, now metapelite.

Original clastic textures, like mesoscopically visible primary structures, are remarkably well preserved in all but the metapelitic rocks, which are completely recrystallized. Microscopically visible textures record a moderately high degree of textural maturity in well rounded sand grains and a fair degree of sorting in most of the sandstone. These ancient red beds retain shades of red, pink, or maroon in most exposures, reflecting fine, interstitial hematite grains that indicate an oxidizing environment of deposition. Color variations generally reflect lithologic stratification, but secondary modification, producing elliptical color banding that crosses stratification (Liesegang bands), is locally visible.

Cross bedding, primarily trough-shaped, is ubiquitous. Both asymmetric and symmetric ripples are common; asymmetric, current-formed ripples dominate the lower half of the formation, whereas both types occur about equally in the upper half. Reactivation structures are locally visible in the upper half in unusually favorable exposures (see Stop 1, below). Roughly the lower half of the formation is interpreted as braided fluvial deposits with unidirectional paleoflow from north to south (in present-day coordinates). The upper half, which records both oscillatory and south-directed flow, is interpreted as very shallow marine with waves and currents (Medaris et al., 2011). Reactivation surfaces are interpreted as reflecting a tidal influence with the dominant hemicycle flowing south and the secondary hemicycle flowing toward the north. The overlying, fine-grained Seeley Slate represents a deepening of the sea, thus reflecting a marine transgression.

Both the Seeley Slate and overlying Freedom Formation (**Fig. 2**) occur in the subsurface along the hinge of the Baraboo syncline, and are known only from former underground iron mines and drill cores

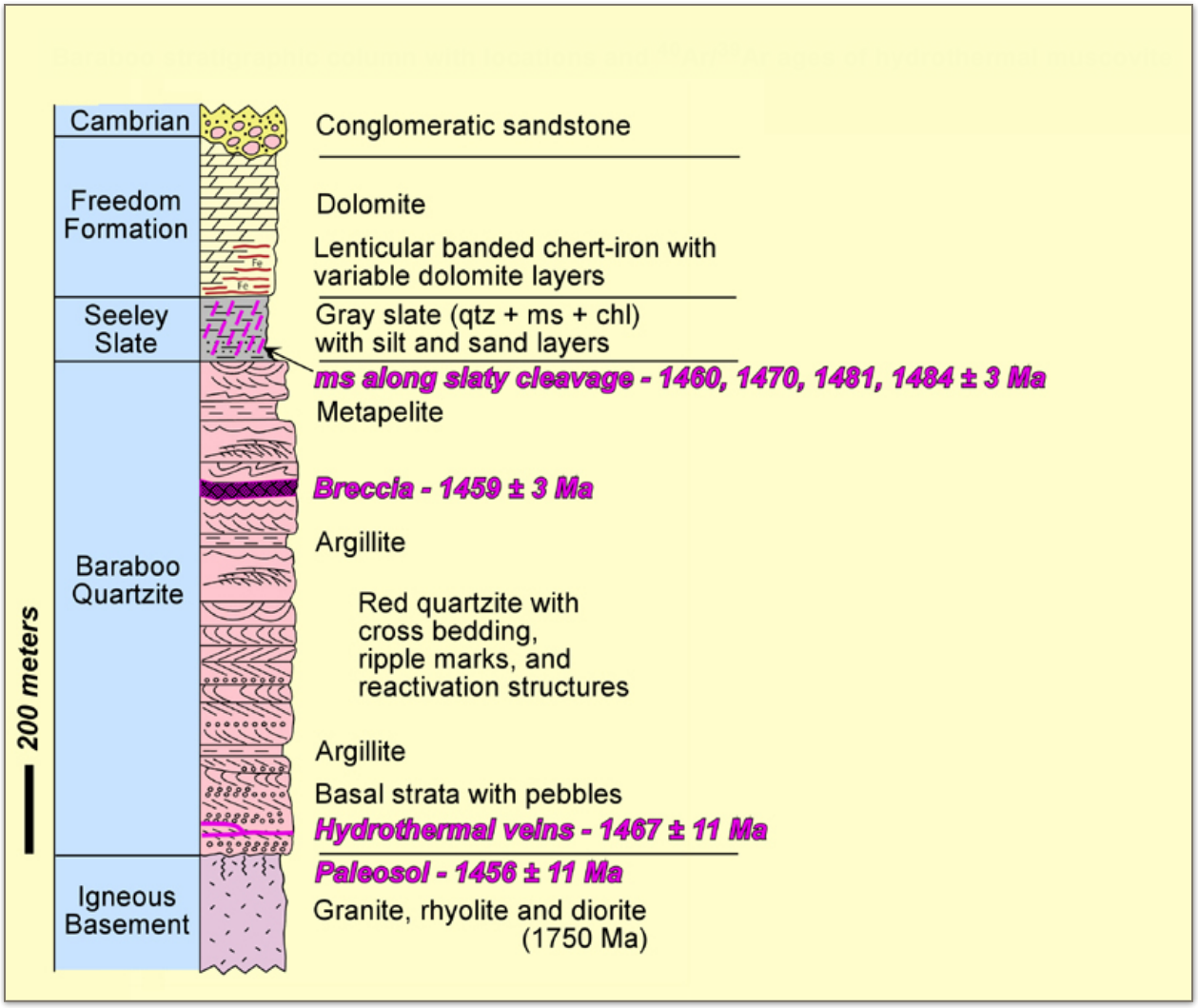

Figure 2. Stratigraphic column of Proterozoic rocks in the Baraboo District, with locations and $^{40}Ar/^{39}Ar$ ages of hydrothermal muscovite (ms).

(Weidman, 1904; Leith, 1935; Schmidt, 1951; Geiger, 1986; Medaris and Dott, 2008; Medaris et al., 2009). Because we are unable to visit these units in outcrop, we provide limited information on the Seeley Slate, which provides important radioisotopic information, and refer the reader to Medaris et al. (2011) for a more complete summary of information on both the slate and Freedom Formation.

Though predominantly a gray to green slate, the ~120m thick Seeley Slate contains variable amounts of thin layers and lenses of quartz silt, rare lenses of fine quartz sand, and dispersed laminae of siderite. Some silty laminae preserve graded bedding, and evidence for soft sediment deformation is locally present. Not surprisingly, the degree of slaty cleavage development is dependent on the proportion of clay-sized and silt-sized material in a given core sample. The Seeley Slate is high in Al_2O_3, reflecting its pelitic nature, but is unusual for shale in being virtually devoid of CaO and Na_2O and relatively rich in K_2O (Medaris et al., 2011). This unusual composition may reflect either a relatively mature illite-rich shale protolith or an extremely mature kaolinite-rich shale to which potassium was subsequently added.

Proterozoic Deformation and Metamorphism Associated with Folding

The asymmetric, doubly plunging Baraboo syncline is the geologic centerpiece of this field trip. As shown in the schematic cross section in **Fig. 3**, both

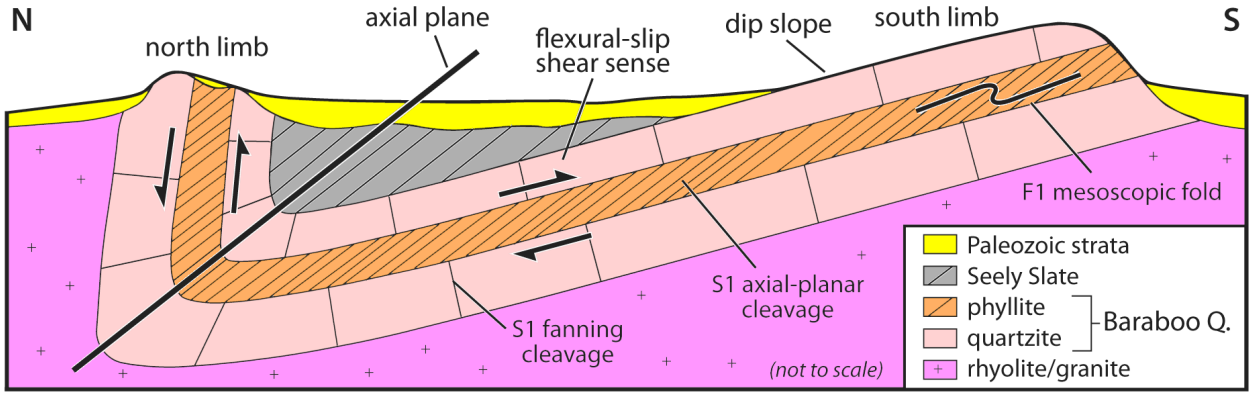

Figure 3. Schematic cross section of the Baraboo syncline, viewed looking east. The stratigraphy has been simplified, with phyllite shown as a single thick layer rather than the many thinner layers present in the field area, and the Freedom Formation omitted. Variations in orientation of cleavage around the fold and between different lithologic units, and asymmetry of minor folds, are represented by fine and bold lines, respectively. Figure from Defrates and Marshak (2007), reproduced in Medaris et al. (2011).

basement igneous rocks and Proterozoic strata define the fold form. In this section, we describe the syncline and smaller scale structures associated with the syncline, together with metamorphism recorded by the folded rocks. The purpose of integrating discussion of metamorphism and deformation is to explicitly consider the relationship between metamorphic minerals, deformation conditions, and the axial planar foliation to the fold.

Asymmetric ripples and trough cross beds preserved in the Baraboo Quartzite demonstrate that the Baraboo syncline is defined by vertical to overturned beds to the north and shallowly north-dipping beds to the south. The fold hinge plunges west in eastern exposures, and east in western exposures, where strata exhibit multiple smaller amplitude folds, and the structure is more appropriately termed a synclinorium. In three dimensions, then, the fold has a shape similar to the hull of a ship that is listing to one side.

All of the rocks involved in the fold have been metamorphosed. Although igneous textures are well preserved in basement rocks, igneous minerals are largely replaced by greenschist facies metamorphic phases. Biotite is generally replaced by chlorite; hornblende by chlorite + actinolite + cummingtonite; and calcic plagioclase by albite + epidote (sausserite) ± calcite. Alkali feldspar of intermediate composition exsolved into near end-member albite + microcline. A Rb-Sr whole-rock isochron for 1750 Ma granites and rhyolites from the Baraboo basement and Montello batholith yields an apparent age of 1635 ± 33 Ma (Van Schmus et al., 1975; Medaris et al., 2003), which suggests metamorphism occurred during the Mazatzal orogeny.

Although a clastic texture is still preserved locally in quartzite and phyllitic quartzite, the metamorphic mineral assemblage in the supermature Baraboo Quartzite records greenschist facies metamorphism. The clay mineral kaolinite, present in the quartz arenite protolith to the quartzite, was replaced by pyrophyllite through the reaction kaolinite + quartz = pyrophyllite + H_2O. The typical mineral assemblage of the quartzite is quartz + pyrophyllite + hematite + rutile. Although these minerals are not amenable to radioisotopic dating, they do constrain metamorphic grade. If we assume an H_2O activity of 1 and pressure of 400 MPa, quartz and pyrophyllite coexist between 300°C and 400°C (representing a geothermal gradient between 25°C and 33°C/km, reasonable for a contractional tectonic regime, at the assumed pressure). At this range of temperatures, quartz can deform by a variety of deformation mechanisms, including brittle fracture, solution mass transfer, and dislocation glide and creep, depending on water content and pore fluid pressure (see review by Passchier and Truow, 2005, and references therein).

The mesoscale structures visible in the Baraboo syncline reflect both deformation at these relatively low temperature, low pressure conditions and the thickness and abundance of quartzite versus phyllite layers. Foliation character and orientation illustrate this point. In phyllite and phyllitic quartzite, the foliation is defined by aligned plates of pyrophyllite,

elongate quartz grains, and solution seams (Czeck and Ormand, 2007), with the spacing between cleavage planes decreasing with quartz content. Phyllite layers sandwiched between thicker quartzite layers record evidence of shearing through flexural slip, causing the cleavage to rotate toward parallelism with the axial surface of the fold (**Fig. 3**). Quartzite records microstructural evidence of recrystallization and a weak shape preferred orientation of quartz grains as well as solution seams (Jank and Cambray, 1986). Only a widely spaced solution cleavage, formed at a high angle to relict bedding, is visible in outcrop (Dalziel and Dott, 1970). The degree of development of the solution cleavage in quartzite varies dramatically around the fold, from undetectable (Point of Rocks) to well developed and relatively closely spaced (Van Hise rock). This variation likely reflects pyrophyllite content, as the presence of a layer silicate appears to be necessary to development of solution cleavage. Traut (2011) found dissolution seams were only present with high pyrophyllite content (<77% quartz) regardless of position on the fold. Lithologic controls on cleavage development such as these have been described by Marshak and Engelder (1985), among others. Microstructures indicating dynamic recrystallization during dislocation creep are only found within layers with high quartz content (>65% quartz; Traut, 2011), marking a potential transition from dominant solution mass transfer to dislocation creep deformation processes at approximately 70% modal quartz content.

Where phyllite dominates and quartzite layers in phyllite are relatively thin, local mechanical instabilities allowed both boudins and mesoscale folds to develop. These structures record a remarkably rich history of fold development (Czeck and Ormand, 2007; Defrates and Marshak, 2007; Marshak and Defrates, 2015). To discuss this history, we will follow Defrates and Marshak (2007) in referring to the regional-scale deformation that produced the syncline as D_1. In this context, both the Baraboo syncline and mesoscopic, second-order folds referred to above, which have hinges parallel to the larger fold, are F_1 folds. Cleavage that is either parallel to the axial surfaces of F_1 folds (in phyllite layers) or fans around the F_1 folds (in thick quartzite beds) is S_1 cleavage. Bedding, by convention, is S_0.

Mesoscopic F_1 folds are restricted to the shallowly dipping south limb of the Baraboo syncline, which exposes portions of the upper quartzite in which phyllite is more common (e.g., Dalziel and Dott, 1970). These asymmetric folds record top-to-the-south shear, consistent with the asymmetry of the regional scale syncline. Associated boudins, however, also record both hinge-parallel and hinge-perpendicular extension (Czeck and Ormand, 2007). Relationships around the small fold in a thin quartzite layer at the entrance to Devils Lake State Park (Stop 2) demonstrate that the quartzite bed was first boudinaged, then folded.

Quartz-filled extension fractures, which locally exhibit quartz grains at a high angle to vein walls, record extension in quartzite on both the north and south limbs of the Baraboo syncline (Dalziel and Dott, 1970; Czeck and Ormand, 2007). The veins locally fill space between boudins. They occur both in isolation and in clusters, and locally form en échelon arrays of 'tension gashes'. The majority of these veins are perpendicular to the hinge of the syncline. Other common orientations include steeply dipping, roughly hinge-parallel veins, and veins that dip shallowly to the north or south. These structures, maximum finite strain axes determined from quartz grain shapes that parallel the fold axis (Craddock & McKiernan, 2007), chocolate-tablet style boudins described earlier, and the curvilinear hinge of the larger syncline collectively indicate that the fold is not a simple plane-strain structure, but records three-dimensional deformation with significant extension parallel to the fold axis (Czeck and Ormand, 2007). Czeck and Ormand (2007) suggest that all of these features and overprinting relationships record a two-stage fold development, in which initial layer-parallel buckling is followed by further buckling associated with bulk top-to-the-south shear ($\gamma \approx$ 0.3-0.55, shortening of ≈ 25%), a model compatible with the fold's asymmetry.

Other mesoscopic structures are restricted to phyllite layers. In phyllite layers that are more than 20 cm thick, S_1 cleavage is locally deformed. The resulting features are classified as D_2 structures because they are relatively younger than the cleavage, rather than because they are associated with a separate orogenic event (Defrates, 2007; Marshak et al., 2009). The D_2 structures developed in phyllite include S_2 crenulation cleavage and associated F_2 folds. F_2 folds range in shape from sharp kink bands to rootless, isoclinal hinges, with hinge orientations roughly parallel to both mesoscale F_1 folds and the regional

syncline axis. The F_2 folds, however, verge down-dip, opposite to the vergence of the F_1 folds (Defrates, 2007, and references therein; Marshak et al., 2009). The S_2 asymmetric crenulation cleavage is axial planar with respect to F_2 folds, and parallels kink band boundaries. Between crenulation cleavage domains, the earlier S_1 cleavage is sigmoidal.

The youngest structures present in the Baraboo quartzite are discrete slip surfaces decorated with quartz slickenfibers and quartz-cemented breccia. The former small faults (none is a map-scale structure) can be divided into two categories: shallowly dipping, bed-parallel dip-slip and near-vertical strike-slip faults. Bed-parallel faults have been documented on the shallowly north-dipping, south limb of the syncline only (Dalziel and Dott, 1970; Link et al., 2014, 2015). Slickenfibers on these surfaces plunge down-dip (perpendicular to F_1 fold hinges) and record top-to-the-south shear. Their association with bedding surfaces, particularly those separating phyllite from quartzite, and the consistent sense of shear recorded by phyllite cleavage, mesoscale F_1 folds, and slickenfibers, has led to the designation of slickenfibers as L_1 structures (Defrates and Marshak, 2007). However, in contrast with the bulk of the Baraboo Quartzite, in which the common phyllosilicate phase is pyrophyllite, phyllosilicates associated with quartz slickenfibers from LaRue Quarry are stably coexisting pyrophyllite and kaolinite (Link et al., 2013, 2015). Assuming a confining pressure no higher than 400 MPa and an H_2O activity of 1, this mineral assemblage constrains the temperature at the time of precipitation of quartz fibers to 280-330°C (Chatterjee et al., 1984). Slickenfibers therefore record slightly lower temperatures than host rock.

Slickenfibers on bed-parallel slip surfaces exhibit fluid inclusion trails perpendicular to quartz fibers, interpreted as healed microcracks formed by numerous crack-seal events (Link et al., 2013, 2015; **Fig. 4**). Analysis of fluid inclusions preserved between these microcracks in quartz fibers from two samples collected from LaRue Quarry show that fluids precipitating quartz were aqueous with ~12 wt% NaCl. Using the temperature constraints mentioned above, fluid inclusion analysis restricts entrapment pressures to 280-360 MPa, recording precipitation of slickenfibers at ~11-13 km depth. Because the fluid is not pure water, these PT conditions are considered

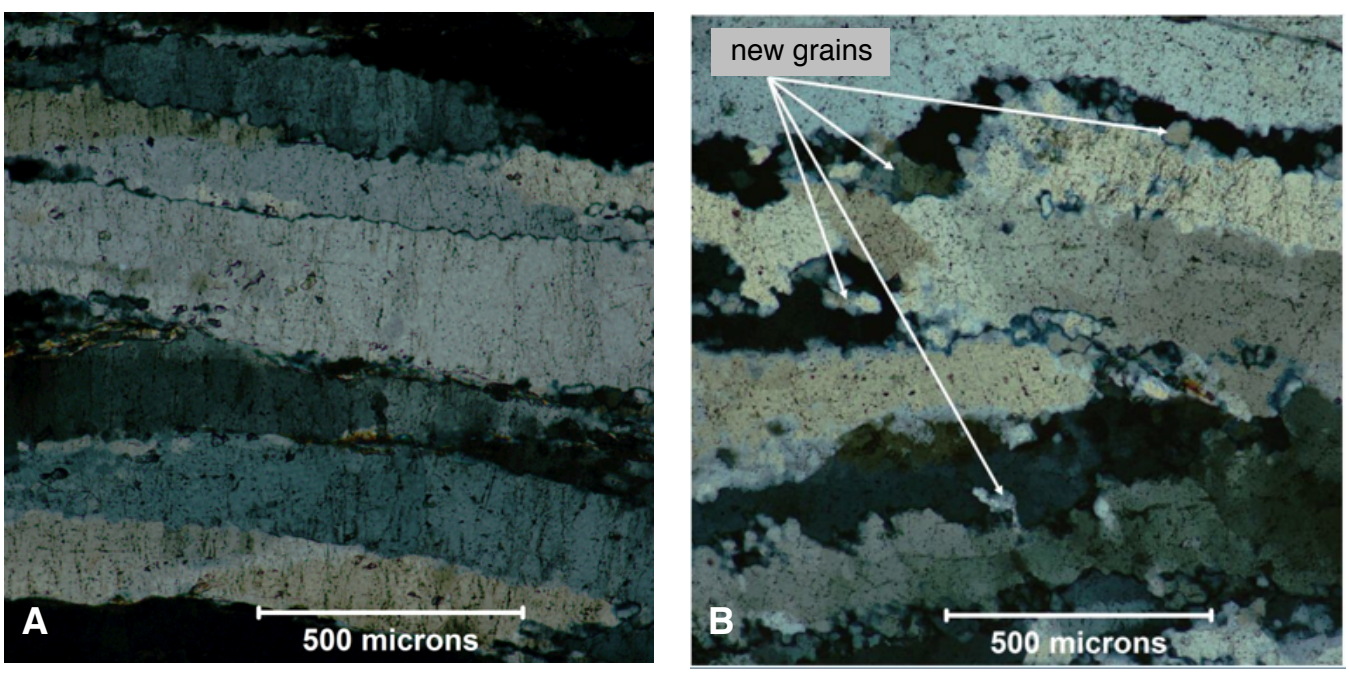

Figure 4. Photomicrographs of slickenfibers on bed-parallel slip surfaces from LaRue Quarry, taken with crossed polars. (A) Elongate quartz fibers showing fluid inclusion trails (nearly vertical in this image) Limited recrystallization is recorded by bulging and, locally, subgrains along fiber margins. (B) Sample from a different bed-parallel slip surface showing more extensive recrystallization, including the development of new grains. These different slickenfibers are indistinguishable in outcrop. Photographs courtesy of Benjamin Link.

maxima. They suggest deformation at the frictional-viscous transition in quartz, an interpretation consistent with microstructures, which record variable amounts of bulging and subgrain rotation recrystallization at the margins of slickenfibers (**Fig. 4**). Significantly, fluid inclusion planes both crosscut and locally terminate at recrystallized grain boundaries, suggesting that deformation in slickenfibers alternated between microfracture and dislocation creep.

Fagereng and Sibson (2011) documented remarkably similar slickenfibers on small faults in an exhumed accretionary prism in New Zealand. They interpreted each fluid inclusion trail as representing an extensional fracture event that was driven by elevated pore fluid pressure. Assuming the distance between healed microcracks represents the magnitude of slip for a given event, each microcrack would record a low stress drop, microearthquake. The authors proposed that hundreds of microearthquakes distributed among the densely spaced small faults they documented throughout the exposed prism would have produced the low frequency, slow slip events that have been documented in subduction zones, which have been attributed to elevated pore fluid pressure (Shelley et al., 2006; Ito and Obara, 2006). Link et al. (2013, 2015) propose that bed-parallel slickenfibers in the Baraboo syncline record episodic slow slip, inferred to be driven by elevated pore fluid pressure, separated by periods of aseismic creep when pore fluid pressure was low.

In contrast to bed-parallel slip surfaces, high-angle strike-slip faults are found on both the north and the south limbs of the syncline. The majority of these near-vertical faults strike northeast (oblique to the hinge of the syncline) and record evidence of dextral shear (Christian MacLeod, UW Senior Thesis, 2012). A few of these record sinistral slip, and a small number of both dextral and sinistral faults strike NNW, roughly perpendicular to the syncline hinge. The phyllosilicate associated with examples of these faults sampled at LaRue Quarry is pyrophylllite. At Point of Rocks, a single strike-slip fault is cut and displaced by multiple bed-parallel thrust faults. This is the only example of interaction between the two fault sets identified to date; however, it is consistent with the slickenfiber mineral assemblages documented from samples at LaRue Quarry, which suggests that the strike-slip faults, like the Baraboo Quartzite, record slightly higher temperature. Slickenfiber microstructures associated with these strike-slip faults, however, are generally similar to those shown in **Fig. 4**. Notably, the ductile deformation producing subgrains and new grains is most extensive in slickenfiber steps with the smallest amount of pyrophyllite.

Total displacement on either fault type is small, with slickenfibers typically measuring two centimeters or less in length. The strain they collectively accommodated is therefore also small. We hypothesize that these faults represent the last stages of shortening and shearing associated with folding, as the system cooled. In this scenario, bed-parallel flexural shear would have been accomplished by distributed deformation within phyllite layers at higher temperatures during the development of the Baraboo syncline. At lower temperatures, slip was localized on discrete phyllite/quartzite contacts. This localization would have been facilitated by the episodic increases in pore fluid pressure that microstructures are interpreted as recording (Link et al., 2013, 2015). Further research is clearly necessary to evaluate this hypothesis and more thoroughly investigate the source of fluids involved in this deformation.

Most researchers interpret the deformation and metamorphism discussed in this section as a record of foreland crustal shortening during the Mazatzal orogeny (Dott, 1983; Van Schmus et al., 1993; Holm et al., 1998; Craddock and McKiernan, 2007; Czeck and Ormand, 2007). However, the discrepancy between the progressive northward decrease in fold intensity, intragranular strain, and thermal resetting of basement (Fig. 1) and the southward vergence of the Baraboo syncline (Fig. 3) remains puzzling, particularly given the involvement of basement in the fold, which precludes detachment between basement and the cover sequence. Marshak and Defrates (2015) suggest a model that accommodates all of these observations, proposing that the Baraboo syncline formed in a tectonic setting comparable to that of the Pyrenees. They propose that the syncline formed in the footwall of a basement-penetrating thrust that initiated after closure and shortening of the basin containing the Baraboo Quartzite yielded a layer-parallel shortening fabric. As the fold evolved, flexural slip rotated the phyllitic cleavage into its current axial planar position, after which shear triggered the development of crenulation cleavage. This two-stage fold development

is consistent with the structures and deformation history (buckling followed by simple shear with buckling) documented by Czeck and Ormand (2007). Marshak and Defrates (2015) propose that loading produced by subsequent emplacement of overlying thrust sheets then rotated the crenulation cleavage into its current subhorizontal orientation.

Proterozoic Metasomatism

As mentioned earlier, potassium introduced by hydrothermal fluids locally stabilized muscovite. Where muscovites have been dated, they yield $^{40}Ar/^{39}Ar$ ages that are broadly coeval with 1465 Ma Wolf River magmatism (**Fig. 2**). Most zones thus shown to have transmitted fluids of Wolf River age exhibit evidence of enhanced permeability relative to surrounding rocks. These include the paleosol developed in basement beneath the Baraboo Quartzite (mineral assemblage of muscovite + quartz, 1456±11 Ma); hydrothermal veins near the base of the quartzite (muscovite + pyrophyllite + diaspore; 1467±11 Ma), and quartzite breccia (muscovite + kaolinite + quartz; 1459±3 Ma) (Medaris et al., 2002; Medaris et al., 2003). A significant exception is the Seeley Slate, which lies above the Baraboo Quartzite (**Fig. 2**). Permeability generally declines with decreasing grain size; however, the difference between permeability parallel versus perpendicular to slaty cleavage have not been investigated. In the Seeley Slate, sedimentary structures and textures on the sand-sized scale are commonly preserved, but both the S_1 axial plane and S_2 crenulation cleavage are locally well developed in finer grain sizes. The common mineral assemblage is quartz + muscovite + chlorite + tourmaline + rutile ± siderite ± various opaque phases (Medaris and Dott, 2008), with muscovite aligned parallel to S_1 slaty cleavage. The absence of biotite indicates lower greenschist facies metamorphism at temperatures no higher than ~350°C. Four samples of Seeley Slate yielded $^{40}Ar/^{39}Ar$ plateau ages ranging from 1460 to 1484±3 Ma (Medaris et al., 2009). These ages could record the timing of metasomatic replacement of a pre-existing, cleavage-parallel layer silicate by muscovite; a prolonged period during which the rock stayed above the closure temperature of fine-grained muscovite following Mazatzal deformation and metamorphism; or folding and axial cleavage development at Wolf River time.

The Cambrian Record

Proterozoic structures are truncated by the Great Unconformity, which records the profound physical and chemical changes that marked the beginning of the Paleozoic Era (Peters and Gaines, 2012), and overlain by Cambrian sediments (**Fig. 2**). We will view the quartz-rich, Cambrian, Parfrey's Glen Formation where both the underlying unconformity and basal conglomerate are visible.

STOP 1 POINT OF ROCKS NATIONAL AND STATE HISTORIC SITE, south limb of Baraboo Syncline [*UTM 16T 275360E 4812605N*]. *The descriptions that follow are focused on the large outcrop and roadcut on the east side of US Hwy 12. This section of US 12 is currently under construction which, when completed, will include a turnout to facilitate safe visits by geologists.*

We describe the features visible at four different locations along and above the outcrop that is visible from the road from south to north in the following sections. ***You may not hammer the naturally weathered quartzite outcrop (location 1) or the phyllite above the outcrop (location 4); modest collection from the blasted outcrop is permitted, though not encouraged.***

Location 1: Katabatic winds blowing from the Quaternary ice sheet that lay only 3 km to the east sandblasted the south-facing surface of the Point of Rocks outcrop, providing what is arguably the most instructive record of the depositional environment of the upper Baraboo Quartzite available in the Baraboo Hills. This beautifully polished surface illustrates cross-bedding that records Proterozoic currents as well as reactivation structures that attest to the influence of the tide (**Fig. 5**). Soft sediment deformation of several cross sets is interpreted as indicating incipient liquefaction due to sudden increases of current shear disturbing pore fluid and grain packing in dune bedforms. Collectively, these observations, reactivation structures noted at other localities, and an abundance of symmetric oscillation ripples indicate that the upper quartzite was deposited in a shallow marine environment. In contrast, the lower half of the quartzite is inferred to record braided fluvial deposition. Similar depositional environments have been inferred for the upper and lower parts of the

Sioux Quartzite (Ojakangas and Weber, 1984) and the Barron Quartzite (Rozacky, 1987) (**Figure 1**).

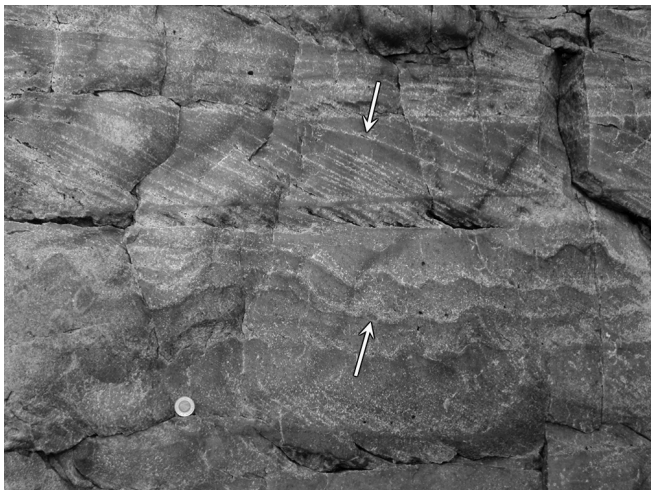

Figure 5. Quartzite outcrop at Point of Rocks (Stop 1, location 1) displays reactivation surfaces (upper arrow) and convolute bedding (lower arrow). Photo from Medaris et al. (2011).

Location 2: The central part of the road cut displays exceptional evidence for both mesoscopically ductile and brittle deformation. Interestingly, both are associated with the finer grained layers in the quartzite, and both fundamentally record competence contrast. In the following paragraphs, we discuss the older structures first.

The orientation of the S_1 cleavage varies across the fold and across layer boundaries, even where slip surfaces have not developed. Generally, quartzite cleavage is perpendicular to bedding and phyllite cleavage is subparallel to the axial plane, but subtle lithologic differences create more interesting patterns. Locally at layer contacts, the refraction of cleavage between these orientations is seen and gradational lithologies are evident by the gentle curving of cleavage at bed boundaries (**Fig. 6**). Initial inspection, suggests that cleavage refraction was largely controlled by mineralogy, but further analysis shows other factors were important as well.

Cleavage refraction can be used to estimate the relative effective viscosities of adjacent layers and evaluate rheologic models (e.g. Newtonian vs. non-Newtonian; Treagus, 1999). The largely bimodal mineralogy (dominantly quartz and pyrophyllite) of layers within the Baraboo Quartzite unit makes this an ideal location to study broader questions of rock rheology. Traut (2011) conducted a cleavage refraction study in combination with a mineralogical and microstructural analysis. She found that the effective viscosity ratios between layers are not constant throughout the fold, but nearly all calculated ratios are less than 5. This small range of effective viscosity ratios indicates that the rocks here can likely be approximated as Newtonian (Treagus and Treagus, 2002), a result seen in many field studies using a variety of techniques that find effective viscosity ratios less than approximately 10 (e. g. Treagus and Treagus, 2002; Czeck et al., 2009, Yonkee et al., 2013). Interestingly, the effective viscosity ratios for adjacent layer pairs with similar relative quartz abundances are not consistent between locations, indicating that, most likely, the rheology had a strain-dependent component. The controlling factors on the relative strength of different layers were most importantly intertwined modal mineralogy and grain size, but also fabric formation and possible fluid interaction.

Fluid-rock interaction was definitely involved in production of quartz slickenfibers, which are particularly well developed on steps provided by ripple marks on an exposed bedding surface at this site (**Fig. 7a**). Much more delicate slickenfibers extend from sheared folia at phyllite-quartzite contacts (**Fig. 7b**). Both the asymmetry of mineralized fault steps and displacement of an older sinistral strike-slip fault record top-to-the south slip on numerous bed-parallel slip surfaces, which are generally localized at either the bottom or top contacts between phyllite layers and

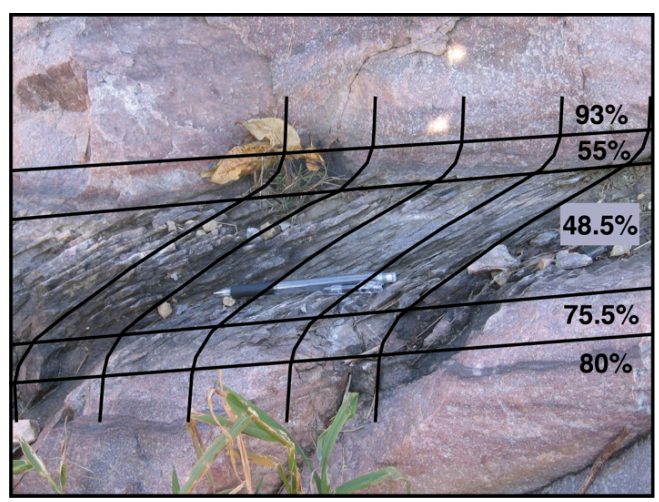

Figure 6. Cleavage refraction between layers with modal percent of quartz determined by point counting (modified from Traut, 2011). Stop 1, location 2.

Figure 7. Small faults exposed near southern end of Point of Rocks outcrop (Stop 1, Location 2). **A.** Slickenfibers extend from ripples on bed-parallel fault, recording top-to-south shear. **B.** Bed-parallel slip surfaces located at tops of thin, fine-grained layers. Note that the lower fine-grained layer is a true metapelite; the pastel green color indicates that is is composed virtually entirely of pyrophyllite. Top-to-south slip (arrows) recorded by curvature of cleavage in metapelite and slickenfibers visible only in outcrop results in small displacements of older, sinistral, strike-slip fault (**S-S**).

quartzite. Where they form at the top contacts, as in the bottom surface shown in **Fig. 7b**, cleavage in phyllite is steeper at the base than at the top, where it curves over into parallelism with the base of the overlying quartzite bed (note the trough cross beds recording way up in quartzite that allow us to distinguish bed tops from bottoms).

Finer slickenfibers are locally present on the strike-slip fault shown in **Fig. 7b.** The fault itself is unusually planar - perhaps an indication of the uniformity of grain size and composition of the quartzite at this locality.

Location 3: The absence of feldspar, rarity of muscovite, occurrence of the aluminosilicate pyrophyllite, and a durable heavy mineral suite dominated by zircon, tourmaline, rutile, magnetite, and hematite collectively attest to the extreme compositional maturity of the Baraboo Quartzite. Minor interbedded phyllite and metapelite, which are more informative geochemically than quartzite, consist almost entirely of SiO_2, Al_2O_3, Fe_2O_3, TiO_2, and H_2O (Medaris et al., 2011). Fine-grained sedimentary layers in the Barron and Sioux quartzites are similarly mature chemically, with Chemical Indices of Alteration (CIA) for samples from all three Baraboo Interval formations lying between 96.8 and 98.8, as compared to those of common mud rocks, which typically range from 50 to 65. These high CIA values require a source region that experienced extreme chemical leaching, resulting in the production of detritus consisting largely of quartz, kaolinite, hematite, and rutile.

The Point of Rocks outcrop is one of the rare sites that includes true metapelite. Because metapelitic

layers are composed almost entirely of pyrophyllite, they are a pretty pastel green (**Fig. 7b**). The northern quarter of the blasted outcrop is a particularly nice place to examine variations in the character the fine-grained layers.

Location 4: A distinct crenulation cleavage (S_2) occurs locally in phyllite exposures of the Baraboo Syncline. Examples are present in the northernmost section of the road cut at Point of Rocks, in the woods above the latter location (a small trail leads from the northern edge of the road cut to the area above) in outcrops on private land both to the east and west, and in exposures at the entrance to Devils Lake State Park (Stop 2). Dott and Dalziel (1970), and others have described the crenulation. DeFrates (2007), Defrates and Marshak (2007), and Marshak et al. (2009) note that the crenulation tends to have a moderate to subhorizontal dip, is axial planar to a set of F_2 mesoscopic folds, and is parallel to local kink bands (**Fig. 8**). Within crenulation domains, relict S_1 phyllitic cleavage typically has a sigmoidal form.

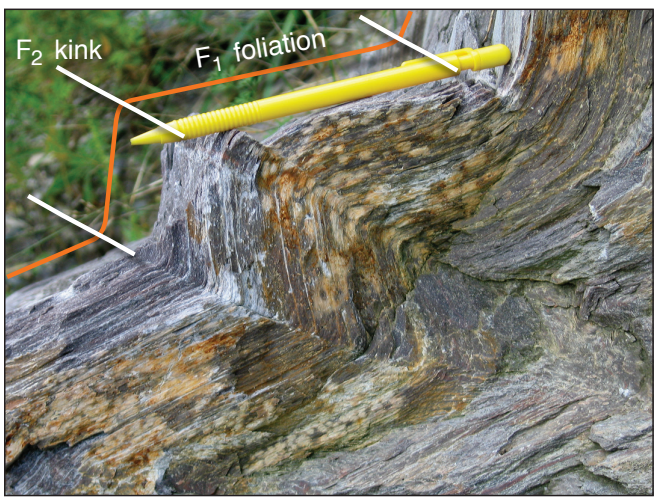

Figure 8. Beautiful kink bands in fine-grained, evenly laminated Baraboo phyllite, exposed at the top of the Point of Rocks road cut (Stop 1, locality 4). Associated crenulation cleavage (**S_2**) is axial planar to kink folds. North to left.

STOPS 2 and 3 Devils Lake State Park, south limb of Baraboo Syncline. We will divide the group in two at the north entrance to the park. Each group will visit one of the two following sites, then switch, after which we will have lunch overlooking the lake. ***Rock collecting is prohibited in the park.***

STOP 2 North Entrance to Devils Lake State Park. *[UTM 16T 278411E 4812532N]Follow Highway 123 to the north entrance to Devils Lake State Park. Park just outside the entrance, which is marked by a stonewall. Walk south along the road until you reach the Speed Limit sign. Turn west. From south to north, tare first a cliff and then a smaller ledge. Walk on a narrow path on the dip slope at the top of the cliff, below the smaller ledge. Walk up the hill ~60m.*

Competence contrast is recorded by splendid structures at this outcrop of thin, strong quartzite layers surrounded by relatively thick, weaker phyllite layers.

Along the path, thin quartzite layers, surrounded by phyllite, exhibit evidence of "chocolate-tablet" style boudinage, indicating layer-parallel extension (**Fig. 9**). This is a classic example of what happens when a relatively rigid layer of rock (quartzite) sandwiched between layers of weak rock (phyllite) undergoes extension. Quartzite rheology during deformation is recorded by the pinch-and-swell style of many of the boudins (recording strain localization and nonlinear plastic deformation) and white quartz veins roughly perpendicular to layering (recording elastic failure, possibly driven by episodic increases in pore fluid pressure). Thus, quartzite records evidence of both plastic and elastic failure. In contrast, phyllite displays evidence of ductile deformation and flow into the boudin necks of the quartzite.

Figure 9. Boudins in thin (~10-40 cm thick) quartzite layers (Q) exposed in roughly E-W transect visible along the path described above record hinge-parallel extension.

Chocolate-tablet boudins, which record extension both parallel and perpendicular to the local and regional fold axes (**Figs. 9 and 10**), provide one of the most compelling pieces of evidence for non-plane-strain deformation in the Baraboo syncline (Czeck and Ormand, 2007). Regionally, maximum finite strain axes determined from quartz grain shapes are oriented parallel (Craddock and McKiernan, 2007), and quartz-filled fractures are oriented normal, to the hinge of the syncline (Dalziel and Stirewalt, 1975; Czeck and Ormand, 2007). These structures therefore also record hinge-parallel extension, and suggest that extension parallel to the fold hinge continued throughout the fold history.

Further up the path at Site 2, the outcrop exposes a rare, spectacular, mesoscopic asymmetric fold. Viewed looking toward the face of the outcrop, it appears to be an "S-fold". It is technically a Z-fold, as viewed looking down the slight plunge of the fold hinge. The fold itself has a geometry similar to that of the larger syncline (approximately parallel hinge and axial surface) leading to the most simple conclusion that folding at the outcrop scale was contemporaneous with folding at the regional scale (i.e., this is an F_1 parasitic fold). We see two layers of quartzite involved in the fold, both of which experienced boudinage prior to folding. As a result, the hinge zone of the lower layer is beautifully exposed in 3-D, but its overturned limb is missing. The S_1 cleavage in the phyllite is roughly parallel to the axial surface of the fold. This cleavage was locally affected by the S_2 crenulation.

Why extension on the limb of the syncline? There are likely multiple reasons: 1) The phyllite/quartzite interval was caught up in a shear couple produced by the bed-parallel flexural slip that accompanied regional-scale development of the Baraboo syncline. The massive quartzite bed that once lay above the phyllite/quartzite interval sheared up-dip, relative to the massive quartzite bed that is exposed below the phyllite/quartzite interval. Layers roughly aligned with the long axis of the strain ellipsoid representing this deformation underwent stretching. The strain ellipsoid is roughly parallel to the axial plane of the Baraboo syncline, and therefore is inclined to bedding at this outcrop. 2) Boudins are only found in outcrops with thick phyllite layers surrounding thin quartzite layers on the south limb of the syncline, indicating that both the right location and necessary rheologic package were required for boudinage to occur. Note that there is evidence for flexural slip on the north limb, also, but boudins are not observed there; however, their apparent absence might be attributed to the lack of thick phyllite layers in the exposed section. The kinematic history responsible for boudinage likely included vertical shortening due to the load imposed by overlying rocks in addition to extension on the south limb during bulk top-to-the-south shearing (Czeck and Ormand, 2007). Rotated isolated boudins (evident at Stop 1) could record either of these shearing components.

Figure 10. Asymmetric fold in thin quartzite layers at Stop 2 records top-to-the-south shearing. Note fanning of fractures bounding boudins around fold, showing that extension in a plane perpendicular to that shown in **Fig. 9** preceded folding. Quartzite is thicker in fold hinge and thinner in limbs, indicating that further thinning of limbs accompanied folding. Foliation in phyllite is axial planar with respect to fold, and oblique to bedding.

The asymmetry of this small fold is consistent with regional-scale flexural slip. The asymmetry of both the overall Baraboo syncline and this mesoscopic fold is also consistent with a top-to-the-south shearing component of deformation. If the extension recorded by boudins is in part related to extension of the south limb during bulk shearing, then the presence of boudins within this small folded layer indicates that buckling continued during bulk shearing as well.

STOP 3 Base of East Bluff, Devils Lake State Park. [*UTM 16T 279264E 4811873N*] *This is the hardest site to find. Look for a building with restrooms, flanked by a small lawn, across from a small parking lot. Walk to*

the southernmost end of the yard (SE corner), then toward the rock, just slightly S of E. Follow a crude path to the outcrop that is just N of the large boulder field, about 7 m N of the boulders.

A hidden path through the woods leads through a talus of boulders to the base of a quartzite cliff. Near the base of the cliff, three dimensional outcrops provide good exposures of incipient solution cleavage in quartzite. Climbing up to the first ledge, a cliff face exhibits structures produced where a 15 cm-thick phyllitic layer with variable quartz content is sandwiched between thick quartzite layers, producing an exceptional record of strain and kinematics of deformation in a rheologic package quite different from that at Stop 2.

Structures in the phyllitic layer vary in character with quartz content (**Fig. 11**). To the north, where quartz content is relatively high and increases upward in the layer, it exhibits parting surfaces which intersect the base of the overlying quartzite at a high angle and curve smoothly to intersect the surface of the bed below roughly asymptotically. At first glance, these partings look like cross beds. They are not. Close examination shows that the surfaces cut across fine bed-parallel laminae within the phyllitic quartzite, so they cannot be cross beds. Rather, they are curving spaced cleavage domains. Their orientation reflects the top-up shear of the quartzite layer above, relative to the quartzite layer below, that developed during regional flexural slip, as well as bed-parallel extension recorded by normal motion on cleavage domains (**Fig. 11a**). The curved trajectory of cleavage domains indicates that the shear strain across the phyllitic quartzite layer increases progressively from top to bottom in the layer. The shear in this interval also opened quartz-filled extension fractures ("tension gashes"). The orientation of these veins is compatible with top-up shear and shortening across cleavage planes.

Further south in this outcrop, the quartz content of the phyllitic layer decreases. In this exposure, cleavage planes are more closely spaced, and exhibit a sigmoidal shape with respect to the bracketing quartzite (**Fig. 11b**), providing a different record of top-to-the-south shear.

North of the section of outcrop pictured in **Fig. 11a** is a series of rough 'steps' up to another ledge, above which a sub-vertical, hinge-perpendicular, strike-slip fault, subparallel to the main cliff, can be seen. Sinistral shear on this small fault is recorded by quartz slickenfibers.

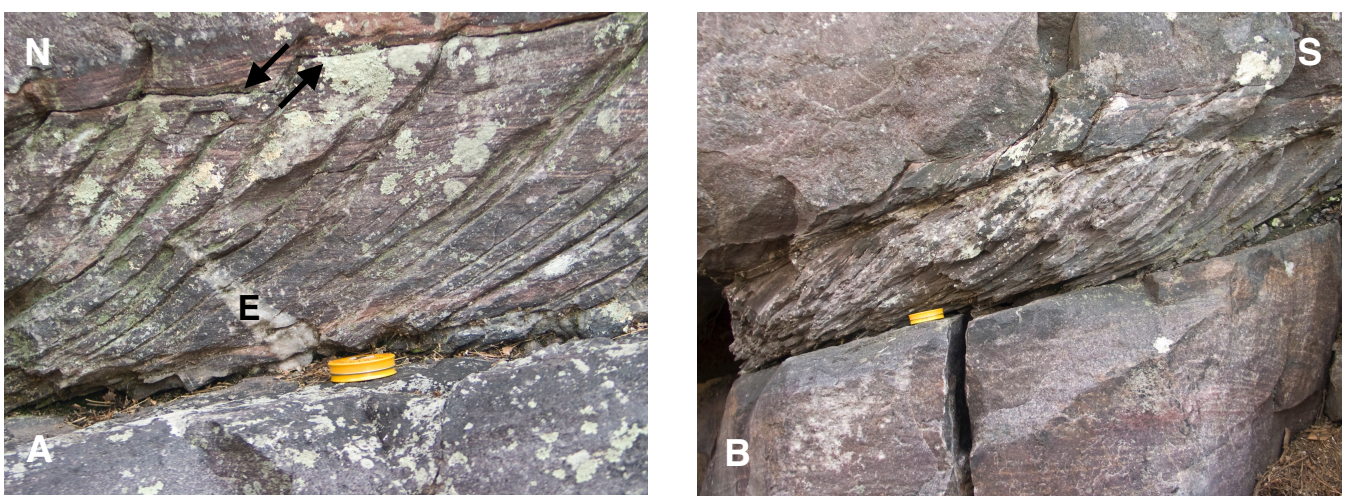

Figure 11. Meso-scale structures visible in a single, thin phyllitic layer sandwiched between thick quartzite layers. Yellow canister is ~2.5 cm in diameter. **(A)** Northern outcrop of phyllite exhibits increasing quartz content from bottom to top, accompanied by increases in both mesoscopically visible laminae and spacing between cleavage planes. Structures recording top-to-south shear and bed-parallel extension include asymptotic bending of cleavage at the base of the bed and normal slip along folia at the top of the bed (arrows). Note quartz-filled extension fracture (**E**) recording shortening roughly perpendicular to foliation. (**B**) Where the same layer is exposed to the south, it is thinner and exhibits a similar quartz content from base to top. In this location, cleavage is sigmoidal, recording flexural slip along both quartzite-phyllite boundaries.

The overlying massive quartzite contains many subtle, thin color bands. Some of these are cross beds, which demonstrate that the beds here are not overturned. Others are Liesegang bands, formed due to interaction of the quartzite with groundwater.

STOP 4 East Bluff Trail: **Short hike to Great Unconformity**. *From the eastern lot at the north shore of Devils Lake, follow the East Bluff trail south from the trailhead [UTM 16T 279305E 4812007N] to Elephant Rock and the overlook [UTM 16T 279406E 4811677N].*

Cambrian conglomerate of the Parfrey's Glen Formation is exposed above the Great Unconformity adjacent to the East Bluff trail at Devils Lake State Park (**Fig. 12**). Boulders in the conglomerate were derived from the underlying Baraboo Quartzite. The boulders are ≤2.5 m long, with most much smaller than this maximum size, and rounding generally increasing as boulder size decreases. The boulders indicate a nearby source. No relict sea cliff is evident today, but the quartzite does form a ramp rising to the south, which suggests a former cliff provided most of the clasts. Many may have originated as core stones produced by spheroidal weathering before the rise of the Cambrian sea allowed surf to attack and roll the boulders over the ramp.

This basal conglomerate passes abruptly upward into sandstone. Latest Cambrian fossils found in the overlying sandstone suggest correlation with the Jordan formation (Winchell, 1864). The elevation of the strata at this location also suggests correlation with the highest strata at Parfrey's Glen to the ESE. Quartzite conglomerate occurs in scattered patches along the sides of Devils Lake valley, demonstrating that this, like most valleys in the hills, was eroded by fluvial processes prior to Cambrian deposition. Conglomerate appears again on the quartzite ridge ~1.5 km southeast of Elephant Rock and extends discontinuously east from the lake toward Parfrey's Glen.

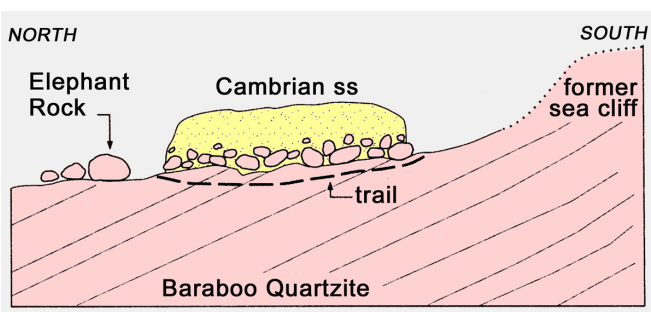

Figure 12. Sketch of Cambrian unconformity and basal conglomerate on the East Bluff Trail (from Dott and Attig, 2004).

STOP 5 Baraboo Quartzite, Cambrian Unconformity, Quartzite Breccia, and Van Hise Rock (National Historic Landmark), north limb of syncline. *Park in the small parking area [UTM 16T 264090E 4818610E] on the east side of State Highway 136, ~600 m north of Rock Springs; follow the nature trail [UTM 16T 264062E 4818652N] on the west side of the highway to Van Hise Rock [UTM 16T 264235E 4819260N]. The locations listed below follow this traverse, ending at Van Hise Rock.* ***YOU MAY NOT HAMMER Van Hise Rock.***

Ableman's Gorge provides a rich record of the depositional and deformational history of the Baraboo District. The following locations describe features from south to north across exposures of the north limb of the Baraboo syncline. These exposures, which are dominated by quartzite, formed a steep hill, which was gradually buried by sediments deposited in the Cambrian sea (**Fig. 13**). Above Dott's Detour, which is nestled in an old quartzite quarry, the nonconformity separating Cambrian conglomerate from underlying Baraboo Quartzite can be seen clearly. Staying on the nature trail from this point onward will lead to each of the following locations in succession.

Location 1: **Dott's Detour** provides a shady spot to view a remarkable exposure of ripples preserved on the base of a vertical quartzite bed. These symmetrical wave-formed ripples indicate that the exposure is part of the upper, marine section of the Baraboo Quartzite.

Location 2: **Post-folding brecciation**.

Many outcrops of Baraboo quartzite and similar quartzites elsewhere in Wisconsin (Necedah, Waterloo) contain breccias with angular quartzite fragments enclosed in a 3-D network of white quartz veins. A major brecciated zone is exposed in the abandoned quarry on the west side of Ableman's Gorge. The zone is vertical and east-west striking (broadly parallel with bedding), tens of meters thick, and traceable for several km westward along strike.

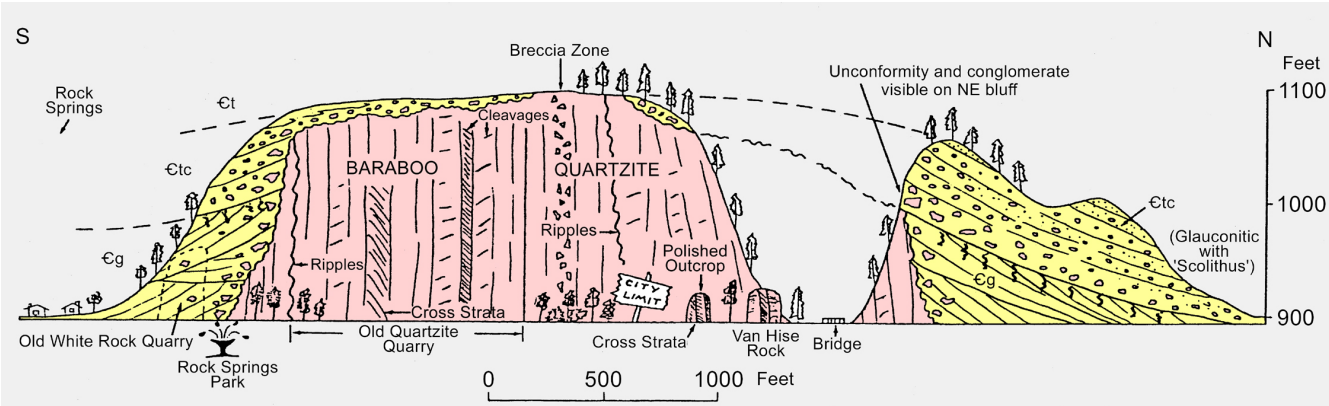

Figure 13. Schematic cross section of the Upper Narrows of the Baraboo River (Ableman's Gorge, Stop 5), showing key geologic features (modified from Dalziel and Dott, 1970). The symmetry exhibited by Cambrian strata abutting buried quartzite cliffs at either end of the gorge is particularly notable. Єg - Galesville Sandstone; Єtc - Tunnel City Group; Єt - Trempealeau Group.

The unstable upper wall is subject to rock fall. We will avoid it, and instead look at spectacular examples of the quartzite in blocks that have fallen on the trail.

Breccia fragments are angular and range in size from about 2 to 30 cm (**Fig. 14**). The bulk strain appears to be purely dilational; the rock fragments seem to have been separated almost isotropically through the growth of vein material, with no evidence for cataclasis or shear offset. The veins consist primarily of coarse-grained quartz with small amounts of kaolinite and hematite. Some crystals of vein quartz have pyramidal terminations, and void space is common in the breccias, although some of this space may once have been occupied by kaolinite.

Several types of textural evidence point to multiple stages of mineral growth in the veins. Pronounced differences in the size and shape of quartz grains within individual samples suggest distinct episodes of growth, and the coarsest, commonly euhedral, quartz crystals are optically zoned, with growth bands marked by planes of fluid and mineral inclusions. Surprisingly, much of the vein quartz exhibits undulose extinction and quasi-ductile microfractures, indicating that it was subjected to appreciable deviatoric stress some time after its formation, although the finite strains are very small.

The dilational nature of the breccias and textural evidence for multiple, distinct episodes of mineral growth point to elevated fluid pressures that transiently but repeatedly reached supralithostatic values. Fluid inclusion analyses of similar breccias from elsewhere

Figure 14. Quartzite breccia at Ableman's Gorge (Stop 5). Note angular fragments and dilation recorded by white quartz precipitated between once contiguous and now separated clasts. Hammer is ~13 cm long.

in the Baraboo district, combined with metamorphic phase equilibria, indicate that brecciation occurred at temperature between 200-260° C and pressures of 50 to 200 MPa (ca. 2 - 8 km) (Medaris et al., 2011).

The timing of formation of the breccias here is unclear. The fact that the brecciated zone parallels bedding could indicate that flexural slip during the main folding event played a role, but this is inconsistent with the lack of evidence for bedding-parallel shear. It is more likely that flexural slip

between stratigraphic layers during folding created layer-parallel zones of increased permeability that provided channels for later vein-forming fluids. Breccias from several other sites in the Baraboo district contain muscovite that yields well-defined $^{40}Ar/^{39}Ar$ ages of ~1450-1460 Ma, approximately coeval with the emplacement of the Wolf River Batholith (Medaris et al 2002). No muscovite has been found at Ableman's Gorge.

***Location 3*: Strike-slip fault and trough cross beds**.

Between breccia exposures and Van Hise rock is an outcrop, a few meters from the path to the west, that exhibits both exceptional trough cross beds and another example of a near-vertical strike-slip fault. The fault is decorated with subhorizontal slickenfibers that record sinistral motion.

***Location 4*: Van Hise Rock**, a National Historic Landmark, is named in honor of C.R. Van Hise, who first studied the relationship between cleavage and bedding in this outcrop (see Appendix, Medaris et al., 2011). This isolated erosional remanent of Baraboo Quartzite exhibits a dark-colored phyllitic quartzite layer sandwiched between two quartzite beds. Vertical bedding contrasts with S_1 cleavage, which dips ~50° north in the phyllitic quartzite (parallel to the axial surface of the Baraboo syncline), and is subhorizontal (roughly perpendicular to near-vertical bedding) in the quartzite bed that abuts it to the south - relationships typical of the syncline as a whole (**Fig. 3**). The outcrop provides an excellent example of mineralogical controls on the development of solution cleavage at greenschist facies conditions. Well developed cleavage in the phyllitic quartzite exhibits wavy, anastomosing cleavage domains with spacing that ranges from 2-6 cm. It is less well developed in quartzite; nevertheless, this is arguably the best exposure of solution cleavage in the Baraboo Quartzite. Quartzite cleavage surfaces are relatively closely spaced and exhibit characteristic mm-scale surface irregularities. The relationship between bedding planes and cleavage in the phyllitic quartzite indicates that the south side of the outcrop moved up relative to the north, providing evidence of flexural slip on this northern limb of the syncline.

Across the Road from Van Hise Rock is a phyllite bed in which a N-dipping, asymmetric S_2 crenulation cleavage is strongly developed. Sigmoids in the cleavage display a top-down shear sense on individual cleavage domains. *Be very careful crossing the road to access this exposure.*

Figure 15. Van Hise Rock, at the north end of Ableman's Gorge, provides three dimensional exposures of cleavage/bedding relationships in a roughly vertical phyllitic quartzite (dark gray) bed sandwiched between two quartzite layers. The smaller remanent of quartzite to the north also exhibits en échelon quartz veins that record roughly hinge-parallel extension and a small thrust fault with well developed quartz slickenfibers that record top-to-the-east slip.

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