

69th ANNUAL MEETING

Eau Claire, Wisconsin — April 24-25, 2023

INSTITUTE ON LAKE SUPERIOR GEOLOGY

Part 2 — Field Trip Guidebooks



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69th ANNUAL MEETING

INSTITUTE ON LAKE SUPERIOR GEOLOGY

**April 24-25th
Eau Claire, Wisconsin**

HOSTED BY

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**University of Wisconsin- Eau Claire and Wisconsin Geological
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Part 2 – Field Trip Guidebooks

Compiled and edited by Rob Lodge

Cover Photos. Upper — Photograph of E.O. Ulrich taking notes in the field describing the Cambrian Mount Simon Formation in the Chippewa Falls region in 1913. Lower — Photograph of geologists E.F. Bean and E.C. Edwards fording the Eau Claire River at Morrison's Ford in 1919.

69th INSTITUTE ON LAKE SUPERIOR GEOLOGY

VOLUME 69 CONSISTS OF:

PART 1: PROGRAM AND ABSTRACTS

PART 2: FIELD TRIP GUIDEBOOK

Trip 1: PRECAMBRIAN GEOLOGY OF THE CHIPPEWA RIVER VALLEY

Trip 2: WISCONSIN'S PALEOZOIC STRATIGRAPHY AND TOUR OF CRYSTAL
CAVE

Trip 3: PRECAMBRIAN GEOLOGY OF THE EAU CLAIRE RIVER VALLEY

Trip 4: QUATERNARY GEOLOGY AND GEOMORPHOLOGY OF THE EAU
CLAIRE REGION

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Part 2: Field Trip Guidebooks

Table of Contents

	Page
Field Trip 1: <i>Precambrian geology of the Chippewa River valley: A transect through the western Marshfield Terrane</i>	1
Field Trip 2: <i>Wisconsin's Paleozoic stratigraphy and tour of Crystal Cave</i>	27
Field Trip 3: <i>Precambrian Geology of the Eau Claire River Valley: Re-discovering the Eau Claire Volcanic Complex</i>	48
Field Trip 4: <i>Quaternary Geology and Geomorphology of the Eau Claire Region</i>	71

Field Trip 1 – Precambrian geology of the Chippewa River Valley: A transect through the western Marshfield Terrane

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Introduction

The erosional outliers of Precambrian bedrock in the Chippewa River Valley represent the southernmost extent of the Canadian Shield before it is completely covered by Paleozoic sedimentary strata. The rocks exposed here are part of the Paleoproterozoic Penokean Orogeny, a collisional orogen that resulted from the accretion of the Pembine-Wausau and Marshfield terranes onto the (present-day) southern margin of the Superior Province. This region was last visited by members of the Institute of Lake Superior Geology in 1980 when a field trip through the region was conducted by Paul Myers (Myers et al., 1980). Since this time, there has been ‘new’ U/Pb data collected by the USGS (Sims et al. 1989) and others (Van Wyke et al, 1997; Klier, 2019), regional syntheses of the Penokean volcanogenic massive sulfide (VMS) mineralization (DeMatties 1989; 1994; 2018; 2022), maps published by government surveys (Mudrey et al, 1987; Brown 1988), and orogen-wide tectonic model (Shultz and Cannon, 2007) that is being revisited based on new U/Pb data (Zi et al., 2021). After forty years of advancing our knowledge of the Penokean Orogen, it is worth touring again.

The Penokean Orogen is perhaps best known for hosting numerous VMS deposits. The passing of the “Prove-it-first” law, or sulfide mining moratorium, in 1997 effectively shut down mineral exploration and mining in Wisconsin. One of the most complete descriptions of several deposits was published by the Institute of Lake Superior Geology (LeBarge, 1996). More recently, the mineral exploration industry has been reinvigorated because of the 2002 discovery

of the 18.2 Mt Back Forty VMS deposit in Michigan, easing of the Wisconsin sulfide mining moratorium in 2017, and a recent national push for securing critical mineral resources. However, this has also highlighted the lack of modern datasets, notably lithogeochemistry, on these deposits that could be used to further our knowledge of the mineral-forming systems in the VMS belt. The Pembine-Wausau terrane has received most of the historic and recent attention since it hosts approximately 150 million tonnes of known VMS mineralization. However, little attention has been given to the Penokean volcanic deposits that overprinted the Marshfield Terrane that are presented in this guidebook. DeMatties (2022) recognized the gap in knowledge for these Penokean volcanic deposits within the Marshfield Terrane, also called the Eau Claire Volcanic Complex, and highlighted their exploration potential.

The portion of the Marshfield terrane that is visited in this guidebook is well known, but grossly understudied and much of its regional context is unknown. Students from the University of Wisconsin-Eau Claire have been visiting many of the locations in this guidebook for decades to learn how to map and describe rocks in the field, measure structures and interpret geologic histories, and learn the basic mechanics of field work. Faculty, students, and alumni from Eau Claire consider these outcrops classic. This guidebook will (re-)introduce these rocks and provide an updated view on their context to the Marshfield terrane and Penokean Orogen. Ongoing research in this region hopes to expand the lithogeochemistry and zircon petrochronology database to better delineate the

geodynamic evolution and crustal architecture of this region. Determining the presence or absence of Archean basement throughout the Marshfield terrane will help refine terrane boundaries and improve our understanding of the metallogeny of the region to assist in future mineral exploration efforts.

Regional Geology

The Paleoproterozoic Penokean Orogen (ca. 1.8 Ga) in the Lake Superior region (**Figure 1**) is a classic Precambrian orogenic belt comprised of dominantly sub-marine volcanic rocks and associated plutons. The Penokean rocks formed

in a suprasubduction zone setting and are now structurally juxtaposed along the southern edge of the Archean Superior Province during the earliest phases of forming the Columbia, or Nuna, supercontinent (LaBerge and Myers, 1984; Sims et al., 1989; Schulz and Cannon, 2007). The orogen is host to at least 150 million metric tonnes (Mt) of VMS and associated mineralization (DeMatties, 1994, 2018) but remains one of the more poorly understood and underexplored mineral districts in North America.

The Penokean Orogen has been divided into the Interior and Exterior domains. The Exterior

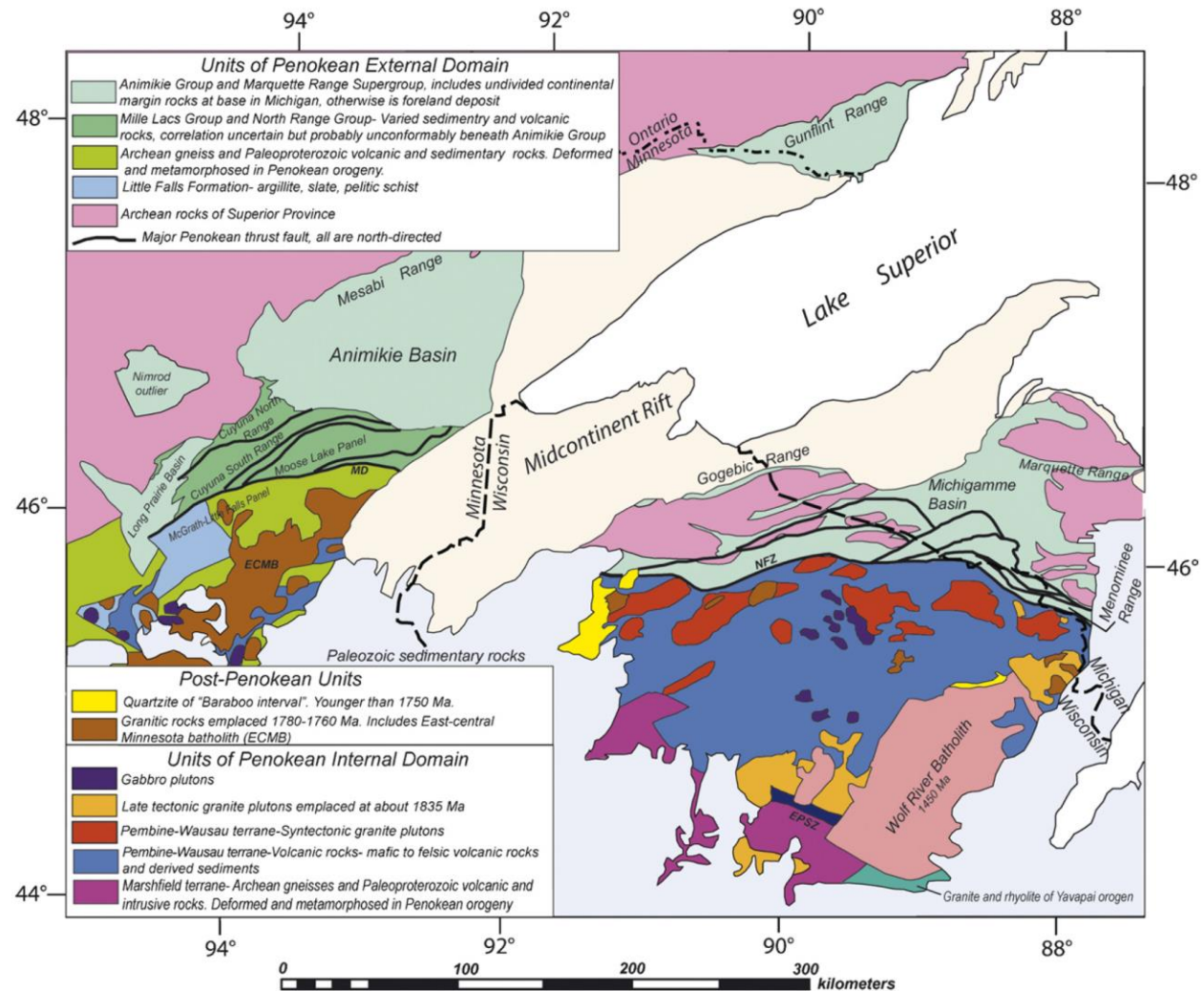


Figure 1 – Geologic map of the major tectonic assemblages and major structures of the Penokean Orogen. Notable and important abbreviations that are important for this guidebook are EPSZ, Eau Claire shear zone; NFZ, Niagara fault zone. Figure from Shultz & Cannon (2007).

domains are sutured to the Superior Craton by the Niagara fault zone (**Figure 1**). The Exterior domain consists of passive margin, rift, and forearc basin sediments and Archean crustal blocks from the Superior Province that were folded and faulted in the foreland part of the orogen. The Interior Domain consists of two accreted terranes, the Pembine-Wausau and Marshfield terranes, that are sutured by the Eau Pleine Shear zone (**Figure 1**). The Pembine-Wausau terrane is a composite accreted oceanic arc overprinted by continental-margin magmatism and hosts numerous VMS deposits and occurrences (DeMatties, 1994; Shultz & Cannon, 2007) (**Figure 2**). The Marshfield terrane is composed of Archean crustal fragments of unknown origin that were overprinted by Penokean magmas during the orogen (**Figure 2**) and is described in more detail in the section to follow.

Shultz and Cannon (2007) synthesized tectonic events during the Penokean Orogeny based on a detailed compilation of lithologic, structural, sedimentological, isotopic, and geochronological datasets. This classic model proposed that an oceanic arc, now referred to as the Pembine-Wausau terrane, collided with the southern Superior Province around 1880 Ma. Following a subduction flip from south-directed to north-directed subduction, continental arc magmatism and back arc extension followed until about 1850 Ma when convergence with an Archean crustal block, known as the Marshfield Terrane accreted to the southern edge of the Wausau- Pembine Terrane along the Eau Pleine Shear Zone (ESPZ). Sedimentation related to this convergence in a foreland basin setting continued until about 1835 Ma. The end of the Penokean orogen was constrained by a series of undeformed post-tectonic plutons dated at 1830 Ma which stitched the terranes.

However, this classic tectonic model for the evolution of the Penokean orogen has recently been re-evaluated in light of new U/Pb data obtained throughout the orogen. The first

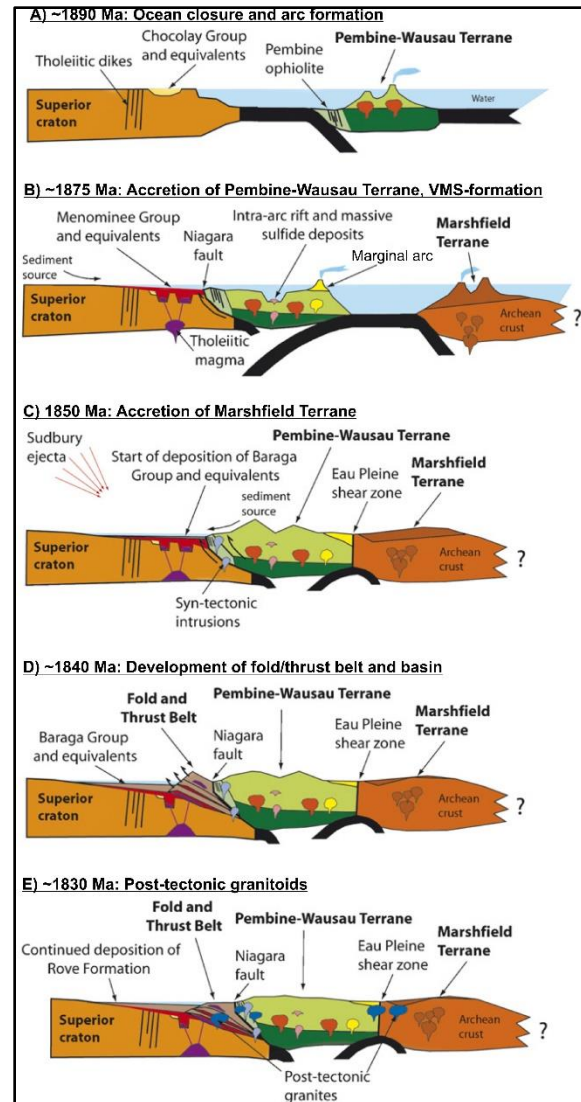


Figure 2 – Schematic tectonic evolution of the Penokean Orogen provided by Shultz and Cannon (2007) based on geophysical, sedimentological, and geochronological complications.

contradictory data came when Quigley (2016) obtained a high-precision U/Pb zircon age of 1832.98 ± 0.52 Ma from a rhyolite at the Back Forty deposit via CA-ID-TIMS. The other analyzed VMS deposits across the Pembine-Wausau terrane by Quigley (2016) provided consistent U/Pb zircon ages ca. 1875 Ma and supported the Shultz and Cannon (2007) tectonic model. Additional U/Pb zircon ages from volcanic units (Beecher Formation) and plutonic rocks (Dunbar Gneiss, Newingham Tonalite) in the eastern part of the orogen by Zi et al. (2021)

supported the younger extensional tectonic event proposed by Quigley (2016). These new ages resulted in a revised Penokean tectonic model where long-lived northward subduction along a continental margin with repeated extensional and contractional regimes in response to retreat and advance of the subducting oceanic plate (**Figure 3**). Weber and Lodge (2022) obtained a U/Pb age of 1831.4 ± 2.0 Ma on the dacite unit hosting the Eisenbrey deposit in the western part of the orogen, suggesting that this second VMS forming event was widespread. A summary of the geochronology is presented in **Figure 4**.

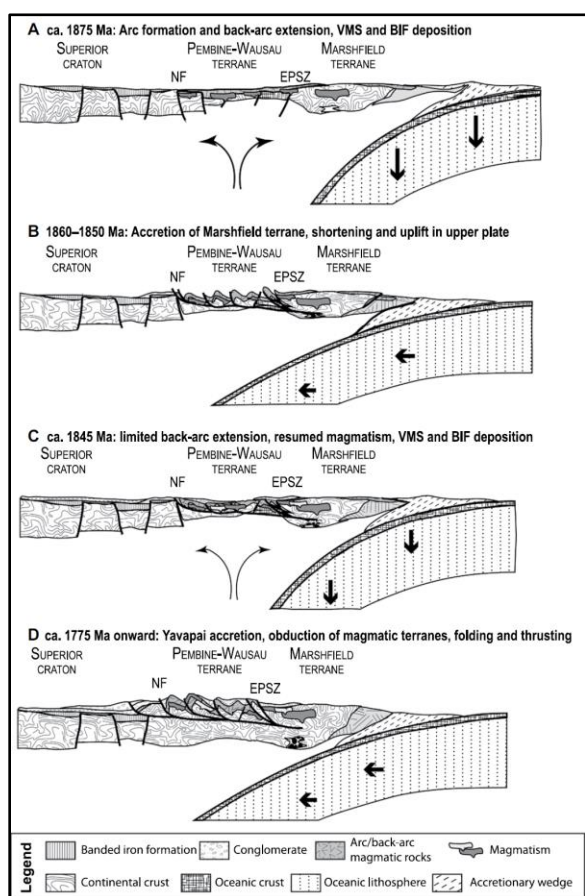


Figure 3 - Schematic illustration of the revised tectonic model of the Penokean Orogen. Figure is from Zi et al. (2021). Abbreviations: NF—Niagara fault zone; EPSZ—Eau Pleine shear zone; Marshfield Terrane

Marshfield Terrane

This guidebook visits field sites from the northwestern exposures of rocks interpreted to be

part of the Marshfield terrane and lie immediately south of the Eau Pleine Shear Zone. Current tectonic models suggest that the Marshfield Terrane represents an Archean microcontinent of uncertain origins (Sims et al., 1989; Schulz and Cannon, 2007; Zi et al., 2021). Some of the earliest work on the terrane by Sims et al. (1989) noted only eight Archean U/Pb ages from isolated outcrops along the Wisconsin, Black, and Chippewa Rivers, many of which were compiled from unpublished sources. One of those was the gneiss exposed at Jim Falls (Stop 3 in this guidebook) which was dated at 2522 ± 22 Ma. Current tectonic reconstructions usually have Paleoproterozoic volcanic rocks in the Marshfield terrane being deposited on Archean basement at about 1870–1860 Ma. The Paleoproterozoic volcanic sequence is referred to as the Eau Claire Volcanic Complex by DeMatties (2018; 2022) and are preserved only as erosional remnants. The Eau Claire Volcanic Complex consists principally of an interlayered sequence of felsic to mafic volcanic rocks, dacite porphyry, and a variety of clastic and chemical sedimentary rocks (Sims et al., 1989). Some conglomerates contain granitic gneissic clasts that were interpreted to be Archean (Myers et al. 1980), but no definitive ages were determined on the clasts. Throughout the Marshfield terrane there are various Paleoproterozoic intrusions of gabbro, diorite, and tonalite. These have U/Pb ages of 1835–1865 Ma (Sims et al., 1989; Van Wyck and Johnson, 1997; Weber and Lodge, 2022). Otherwise, our knowledge of the Archean basement of the Marshfield terrane and associated Paleoproterozoic volcanic rocks remains as sparse as the outcrop exposures.

The study of the Marshfield terrane remained stagnant until new U/Pb and Lu-Hf isotopic data from zircons was published as a masters thesis (Kleir, 2019). The new isotopic data in the Marshfield Terrane collected from the Chippewa and Yellow River valleys will be presented at various stops on this field trip. In our opinion, one of the most significant results was that the “Archean” rocks from the Jim Falls region of

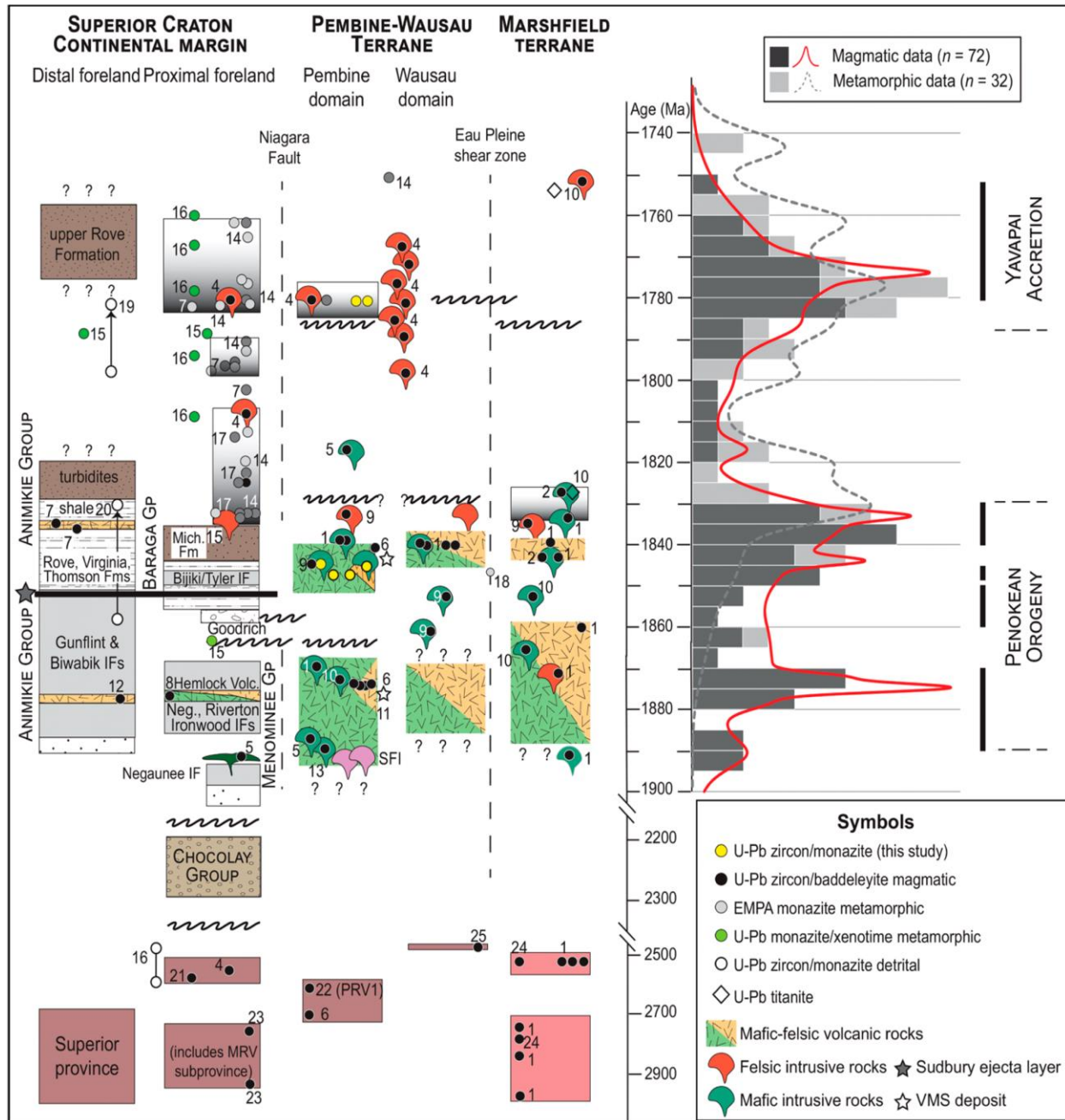


Figure 4 - Time-space plot for the tectonic components of the Penokean Orogen. Plot is from Zi et al. (2021). See citation for references on data sources.

Sims et al. (1989) is a metasedimentary rock that has a significant proportion of Paleoproterozoic zircons (Kleir, 2019). While the data clearly indicates the presence of Archean rocks in the sedimentary source region, the sedimentary provenance does not require that Archean rocks represent the basement architecture in the northern part of the Marshfield Terrane. U/Pb

ages from the northern part of the Marshfield Terrane collected in the Chippewa and Yellow River areas are interpreted as Paleoproterozoic in age (~1.83-1.88 Ga) and Hf isotopies indicate a juvenile source (without Archean contributions). This finding raises questions about the extent of the Archean basement in the Marshfield Terrane and consequently, the basement architecture in

the region. Preliminary geochemistry from Klier (2019) and our ongoing studies in the region show interesting trends that will help distinguish petrogenetic processes. **Figure 5** highlights the REE trace element characteristics of these deposits and their application to each stop in subsequent sections below.

Regional metamorphism in this region is at lower to upper amphibolite facies. Hornblende-plagioclase thermo-barometry from gneisses in the Chippewa River valley indicate metamorphism occurred at temperatures between 606-646°C and pressures between 5.74-6.64 Kbar (Hafften and Radwany, 2018). A sample of amphibolitic gneiss from the Eau Claire Volcanic Complex using the edenite-richertite

thermometry determined temperatures between 719-769°C (Hannack and Radwany, 2018). A rutile U/Pb age of 1835 Ma from Sims et al. (1989) in the Eau Claire Volcanic Complex (Big Falls – Fieldtrip 3 in this volume) may indicate the timing of metamorphism since new zircon U/Pb age from the same region provided a crystallization age of ~ 1875 Ma (Weber and Lodge, 2022).

Field Trip Stops

The overall objective to this guidebook is to tour the Precambrian exposures of the Marshfield terrane along a southwest-northeast transect as exposed in the Chippewa River Valley. Starting within the city of Chippewa Falls, the trip will work its way to the northwest along the river and presumably get closer to the terrane boundary at the Eau Pleine Shear Zone. Stops 1-4 and 6 are all within the Marshfield Terrane, whereas Stop 5 is considered the southernmost exposure of the Pembine-Wausau Terrane. Fieldtrip stops are summarized in **Figure 6**. New data have us questioning what we know about the Marshfield Terrane. For example: Where exactly is the northern boundary of the Marshfield terrane in the Eau Claire region, and how much of the Marshfield Terrane, as currently defined, has an Archean basement architecture?

Most of the locations in this guidebook are at the downstream side of hydro-electric dams. These areas are prone to sudden flooding and the upmost caution and careful planning should be used prior to visiting these locations. In addition, rocks here are uneven and slippery especially when wet. To access larger sections of outcrops, low water conditions or ladders (temporary bridges) may be required. In addition, all locations in this region may contain poisonous plants (e.g. nettle, poison ivy) and black-legged ticks that can transmit diseases. While this is unlikely to be a concern in early spring during the 2023 ILSG conference, future users of this manual should plan appropriately.

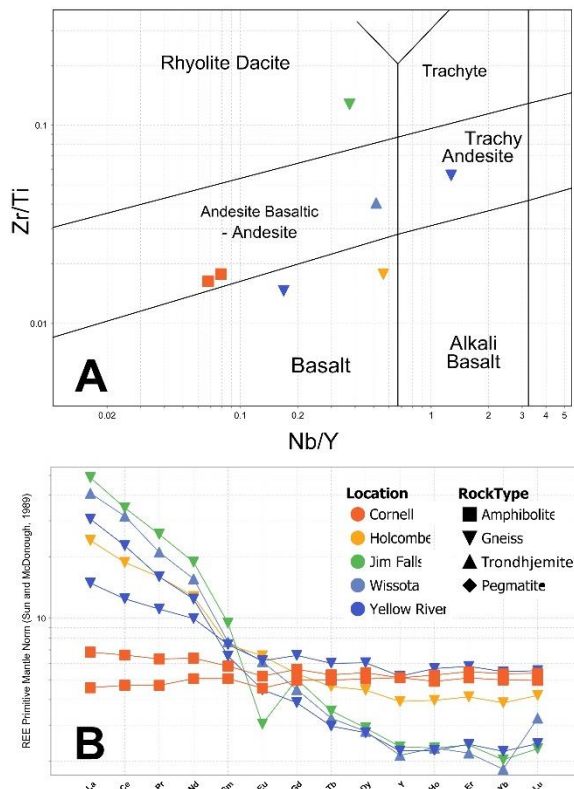


Figure 5 - Trace element diagrams from the rocks in the Chippewa River valley region. Data from Cornell is preliminary data from ongoing projects whereas the remainder is from Klier (2019). (A) Classification diagram from Pearce (1996) showing protolith compositions. (B) Primitive mantle-normalized rare earth element diagram using normalizing values from Sun and McDonough (1989).

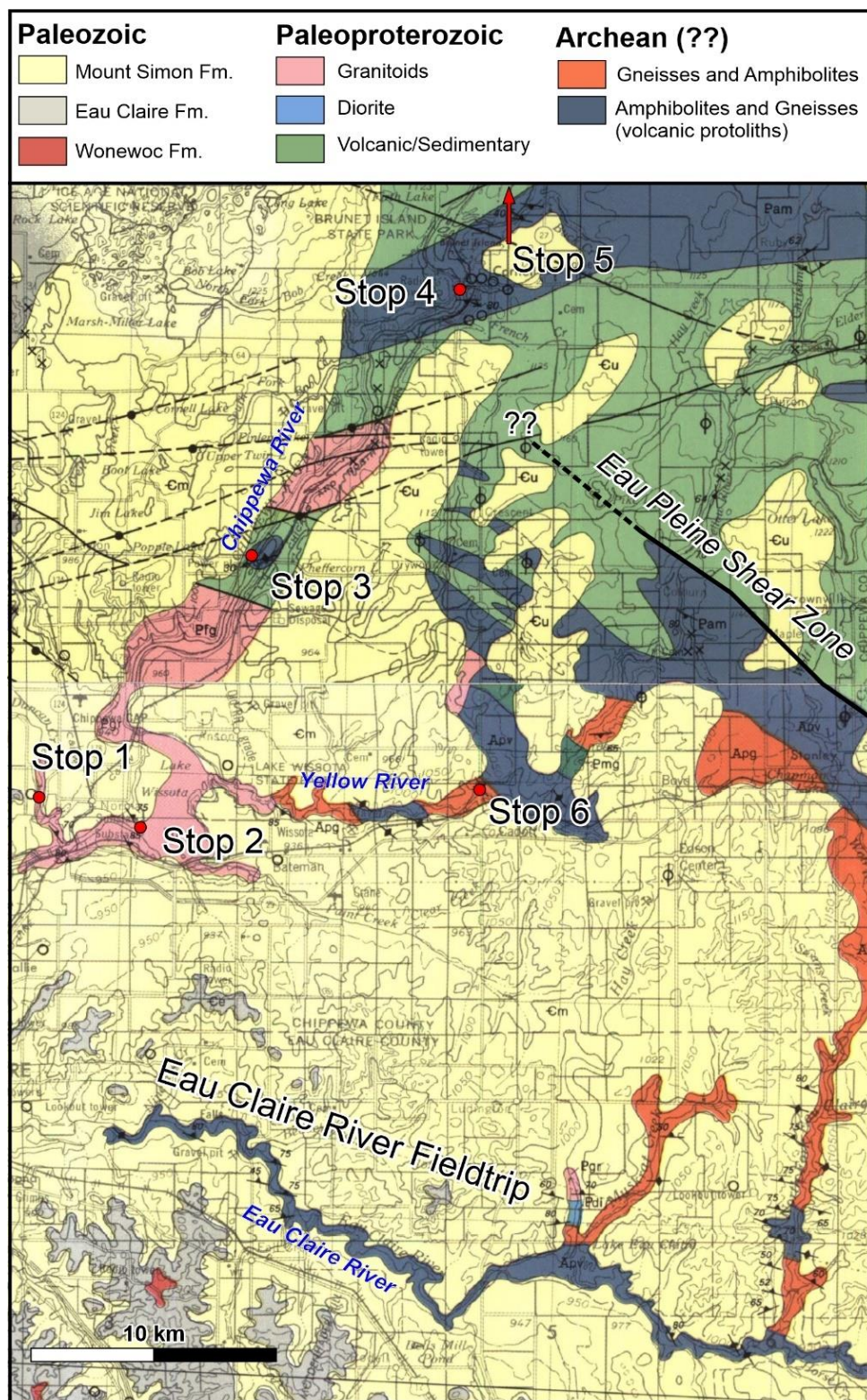


Figure 6 - Regional geology of the Chippewa Falls and Eau Claire region showing fieldtrip stops and approximate location of the Eau Pleine Shear zone. Rocks to the south of the Shear Zone are interpreted to be part of the Marshfield Terrane, whereas rocks to the north are part of the Pembine-Wausau terrane. Figure compiled from Mudrey et al. (1987) and Brown (1988).

Stop #1 – Nonconformity at Irvine Park

Lat: 44.9542° Long: -91.3972°



The outcrop described at this stop is located along the east bank of Duncan Creek within Irvine Park in Chippewa Falls. Upon entering the park, drive north on Irvine Park Drive past the zoo and bison enclosure until the intersection with Bear Den Road. There is ample parking in this area near the intersection that crosses Duncan Creek to the east and find the small foot trail that leads northward to the outcrop. Potential hazards include poisonous plants and ticks, but they are unlikely to be a problem in early spring. There is also uneven and potentially wet ground. The purpose of this location is to highlight the conditions that are impeding the study of the Precambrian bedrock in the region. The Eau

Claire and Chippewa Falls region are some of the southernmost exposures of the crystalline basement in the Lake Superior region before it disappears beneath the undeformed Paleozoic sedimentary strata. This is one of the many exposures of the “Great Unconformity” that is present throughout this region. Details of this unconformity in this region is described in detail in the most recent ILSG guidebook presented in Eau Claire (Chan et al. 1991) and is summarized below.

Precambrian Unconformity

The Precambrian- Cambrian boundary is represented by a highly variable surface in the mid-continent area. In west-central Wisconsin, the Precambrian surface forms an extensive planation surface with a regional SW dip of less than 1 degree. Archean iron formations in the Black River Falls region and Proterozoic quartzites throughout the state, most famously the Baraboo Syncline, form isolated monadnocks on the peneplain. The peneplain was mantled by a layer of paleosols as much as several hundred feet thick. In some areas, Cambrian rocks directly rest upon barren, moderately weathered Precambrian rocks. Considering the low paleolatitude of the continent during the Cambrian, deep weathering of the Precambrian surface must have occurred before the Upper Cambrian deposition. The Precambrian basement, however, shows variable degrees of weathering and the weathering is complicated by potassium metasomatism overprinting associated with Silurian and Devonian K-rich basinal brines (Lui, 1997; Lui et al., 2003). Potassium metasomatism along the unconformity is responsible for the development of illite, interlayered I/S and authigenic K-feldspar in both saprolites and in the Cambrian rocks in the Chippewa River Valley. Where the Precambrian is mafic (gabbros, amphibolites and gneisses) the potassium metasomatism commonly produces a distinctive bright blue-green clay (celadonite) seen at Stop 2 on this field trip and at Big Falls (Fieldtrip 3 in this guidebook). These potassic brines have been

implicated in the formation of MVT deposits in the Tri-state region (Aleinkoff et al., 1993). At this location (and numerous others) where the Cambrian formations are in contact with Precambrian plutonic rocks, the basement is heavily spheroidal weathered (**Photo 1, Figure 7**) and is generally deeply altered to kaolinite saprolite and then metasomatically altered to illite I/S and authigenic Kspar. Mt. Simon Formation sandstone and conglomerate fill the wedges among the spheroids. In some areas such as Little Falls and Big Falls in Eau Claire County (Fieldtrip 3, this volume) or Rock Dam in Jackson County, Cambrian sandstones rest upon Proterozoic amphibolite and meta-rhyolites that are only partially altered.



Photo 1 – Irvine Park outcrop photograph showing nonconformity between Cambrian Mt. Simon Formation (above) and Paleoproterozoic trondhjemite (below). Photo courtesy of Scott Clark (UW-Eau Claire)

Cambrian Mount Simon Formation

The sediment above the unconformity consists largely of Upper Cambrian Mt. Simon Formation deposited on the mid-continent region of North America during the Dreisbachian transgression. The Mount Simon Formation is a fine to coarse-grained, moderately to well sorted, quartz arenite with a local basal conglomerate. The Mount Simon Formation varies in thickness 40 to 180 meters in the Upper Mississippi Valley. Locally, in the Chippewa Valley area, the lower part of the

Mount Simon Formation is a coarse-grained, medium to thick-bedded quartz pebble conglomerate. The topographic relief on the Precambrian surface was probably only a few meters as sedimentary channels are typically less than 1 meter deep. The presence of trace fossils (rusophycus and Climactichnites, or trilobite burrows/tracks; **Photo 2**) and planar and bipolar cross-bedding (**Photo 3**) suggest a littoral or shallow marine tidal flat environment of deposition for the lower part of the formation. The upper part of the Mt. Simon Formation contains feldspathic quartz arenite with small-scale ripple bedding, brachiopod fragments (lingula sp.), and worm trails (planolites) The



Photo 2 - Climactichnites fossil from the lower Mt. Simon Formation collected along the Chippewa River near downtown Eau Claire. Climactichnites trace fossils are typical of tidal flats in the Cambrian. Field of view is ~1m across.



Photo 3 - Cross-bedded conglomerate and sandstone of the Cambrian Mount Simon Formation in the Irvine Park area, Chippewa Falls. Photo courtesy of Scott Clark (UW-Eau Claire).

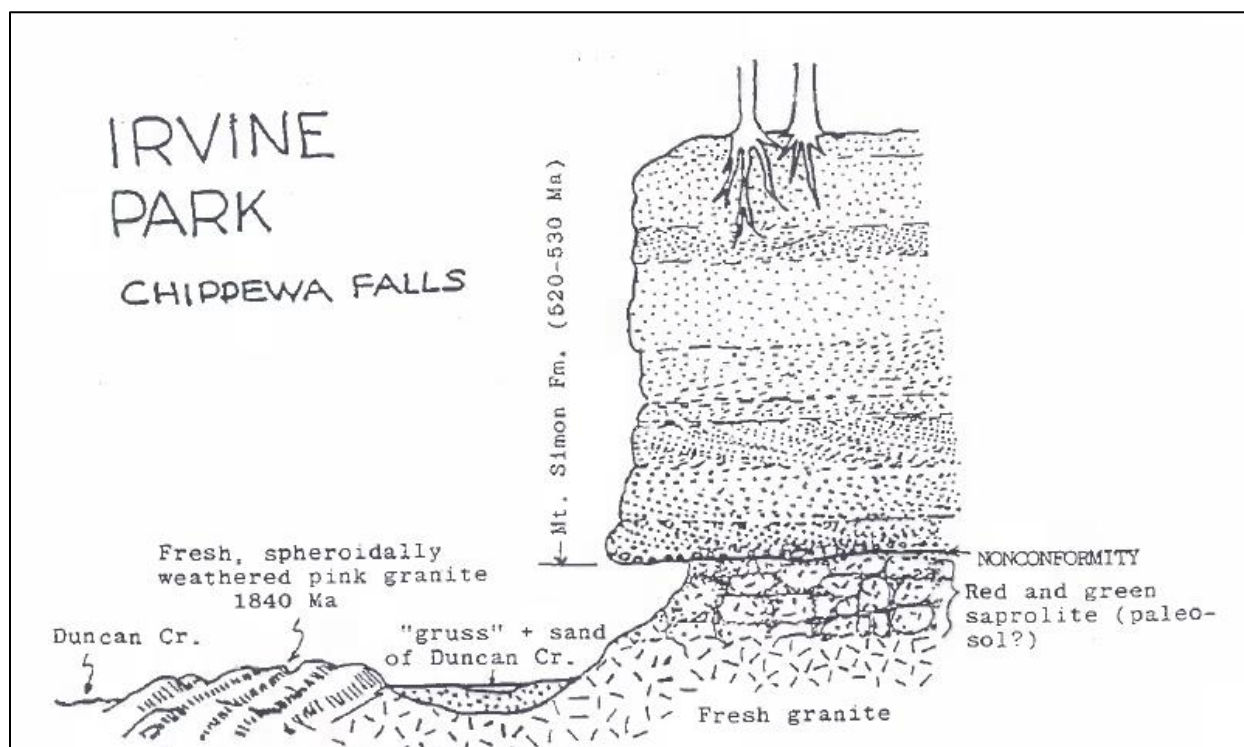


Figure 7 – Conceptual field sketch of the unconformity at Irvine Park. Figure from Chan et al. (1991).

grain size distribution shows a generally fining upward sequence.

Precambrian Intrusion

The Paleoproterozoic biotite tonalite (trondhjemite) showing spheroidal and saprolitic weathering at this location is interpreted to be similar to the larger trondhjemite intrusion that underlies much of the Chippewa River valley. The trondhjemite can be more easily observed below the Chippewa Falls hydroelectric dam in downtown Chippewa Falls and below the Wissota hydroelectric dam (Stop 2 in this guidebook). The biotite trondhjemite at Chippewa Falls Hydro was dated by Van Schmus (1980) at $1,840 \pm 15$ Ma. Saprolites like the one exposed here are characteristic of areas of prolonged tropical to subtropical weathering on a granitoid bedrock surface of low relief. The saprolite at this outcrop contains high clay content and angular quartz and feldspar. The alteration intensity increases approaching the Cambrian Mt Simon Formation.

Stop 2 – Penocean and Mid-Continent Rift Intrusions at Lake Wissota Dam

Lat: 44.9429° Long: -91.3425°

This outcrop is located on the downstream side the dam on Lake Wissota. Drive to the end of 74th Avenue in Chippewa Falls and park in the Chippewa Rod and Gun Club & Marina. From here, you can walk southward along the access road to the dam (about 1 km). There are several places to cross the small stream to access the largest part of the outcrop. The largest potential hazard at this location is the stream crossing and uneven, wet walking area. A ladder or other temporary structure might be required to assist in crossing the stream if water levels are high.

This outcrop highlights some of the magmatic history in this region. The majority of the exposure here is a Paleoproterozoic biotite tonalite (trondhjemite) that has local pods and dikes or medium gray biotite tonalite and alkali feldspar granite pegmatite. The outcrop is intruded by at least three gabbroic dykes associated with the mid-continent rift. The largest



of which is clearly visible in arial view (**Figure 8**). The entire area is covered by thin outwash gravels and silts that varies with seasonal flooding events. Some of the tonalite near the Chippewa River displays the same spheroidal weathering seen at Irvine Park so this location is just below the Great Unconformity and displays some of the same associated potassic alteration along faults and joints seen in Irvine Park (Stop 1). The potassic alteration is responsible for most of the pink color seen in outcrop.

Granitoid Intrusive Suite

Most of the Paleoproterozoic intrusive igneous rocks at Wissota are quartz diorites or tonalites with various proportions of hornblende and biotite. For clarity, and to be consistent with the terminology used by previous geologists working in the Chippewa River Valley, on this field trip we refer to the lightest colored tonalites (color index 15 or less) as trondhjemite and reserve

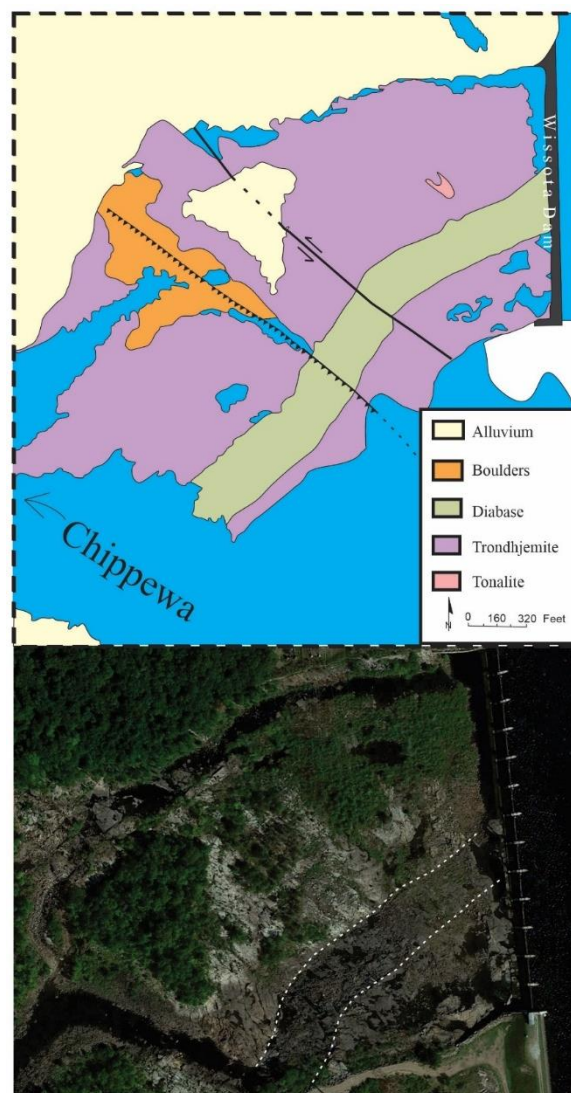


Figure 8 - (top) Generalized geology of the Wissota Dam region. Figure modified from Myers et al. (1980). (bottom) Aerial view of the outcrops with the mid-continent rift highlighted. Image obtained from Google Earth.

tonalite for rocks with higher mafic concentrations. The oldest, abundant rock at Wissota Dam is a weakly foliated hornblende, biotite trondhjemite composed of oligoclase (50%), quartz (30%), microcline (5%), biotite (10%), and 5% hornblende with common accessory euhedral to subhedral titanite (**Photo 4**). Weak foliation strikes N15°W and dips steeply east. This is intruded by small dykes and masses of medium-grained, medium-grey hornblende-biotite tonalite ($\pm$ epidote) that locally contains

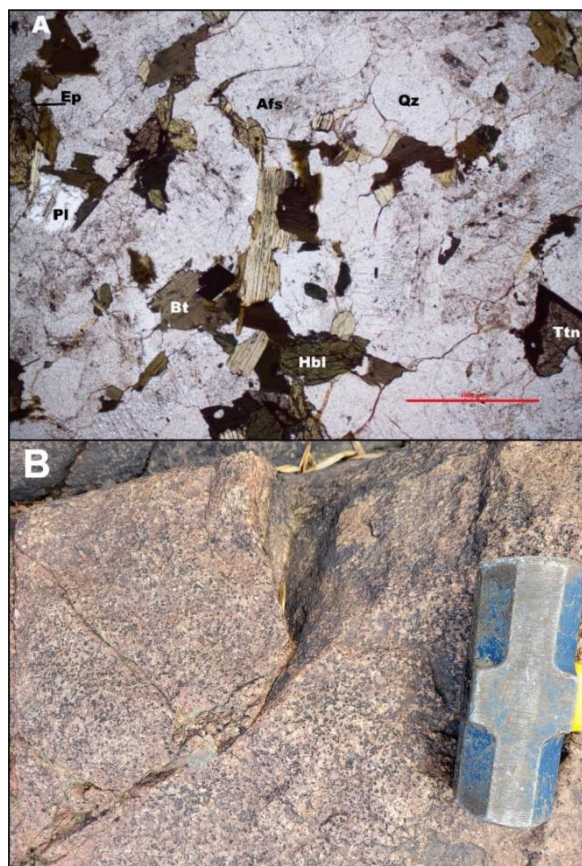


Photo 4 – Photographs of main lighter colored tonalite phase at Wissota Dam. (A)

Photomicrographs in plane-polarized light showing feldspar grains are lightly weathered with opaque rims around titanite. In cross-polarized light, quartz grains show moderate undulatory extinction. Photo from Klier (2019). (B) Field photograph of biotite tonalite (trondhjemite) showing medium-grained, equigranular texture. Feldspars weather pink in color and mafic phases tend to be recessively weathered.

lenticular xenoliths of banded amphibolite. The tonalite pods show no grain size diminution and sometimes have irregular shapes suggesting that some tonalites may be enclaves of earlier phases of the trondhjemite. In other cases, the tonalites are clearly dykes crosscutting the trondhjemite. The tonalite dikes tend to be unaltered with vitreous dark-brown biotite (~25%) and lack foliation. All minerals in the foliated trondhjemite show internal fracturing and dislocation, and contain quartz grains with undulatory extinction, display grain boundary migration and dynamic quartz recrystallization (Klier, 2019). The

trondhjemite is an FI-type felsic rock with strongly depleted HREE (**Figure 5**) representing deep crustal melting (Hart et al. 2004). The trondhjemite is cut by east-northeast-trending pegmatite veins and pods and quartz $\pm$ pyrite) and epidote veinlets.

Potassic alteration especially along any fractured surfaces the trondhjemites produces a pink color in outcrop (**Photo 5**). Minor cataclastic fault zones cut the granitoid intrusions with left-lateral displacement. A thin, branching discordant sheet of foliated biotite trondhjemite approximately 1-3 meters wide and trends N55°W. Drag folded foliation in the enclosing rocks indicates left-lateral displacement.

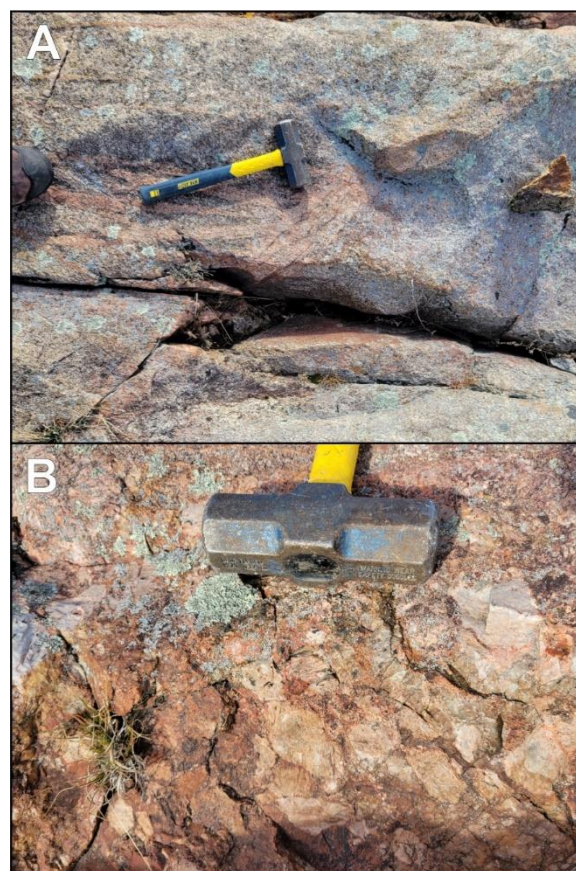


Photo 5 – Photographs of pegmatite and associated alteration at Wissota Dam. (A) Thin quartz-epidote veining and potassic alteration surrounding conjugate fracture sets adjacent to pegmatite. (B) Coarse grained texture of the pegmatite. Feldspar crystals can be as large as 5-7 cm in size.

Slickenside fault surfaces elsewhere in this outcrop have similar strike and dip with the slickensides plunging 5°NW.

Mid-Continent Rift Dykes

Three diabase dykes related to mid-continent rift extension intrude the granitoids (**Photo 6**) exposed below the Wissota Dam spillway and the largest dike is an ENE trending (~N65E) olivine tholeiite that averages 47m in width. The large dyke has a notable sharp and chilled margin. Unpublished data from the dyke at this location and others along the Chippewa River indicate a tholeiitic composition that shows slightly more Mg-enrichment trends on AFM diagrams in comparison to other parts of the dyke swarm in the region.

Ongoing student-faculty research at the University of Wisconsin-Eau Claire is examining the composition of the large dyke at Wissota Dam. Samples were collected as a cross section

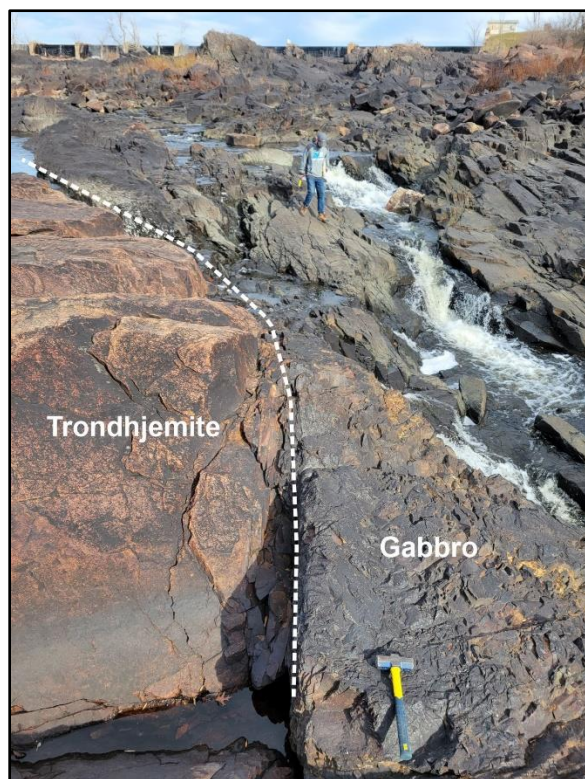


Photo 6 – Photographs showing sharp, chilled margin of mid-continent rift gabbro with Paleoproterozoic trondhjemite intrusion.

of the largest dyke and the mineral chemistry was examined using SEM-EDS and optical petrography. The main dyke has an aphanitic chilled margin a few cm wide along both the north and south sides and the grain size consistently coarsens toward the middle of the dike into a medium grained olivine gabbro (**Photo 7A**). A prominent set of joints perpendicular to the cooling surface along the dike walls are interpreted as columnar joints and these are especially prominent on the south side below the power lines. More pronounced columnar joints are also seen in one of the smaller (2m wide) dikes along the northwest side of the area next to the Chippewa River. No internal contacts are apparent at the outcrop scale suggesting that this large dike represents a single cooling unit of magma intruded into the upper crust. West of the power lines the dike is cut by two faults, one left lateral strike slip fault with a few meters of displacement and a low angle reverse fault with well-developed chlorite slickensides and extensive alteration including chlorite and hematite, and calcite filled tension fractures.

The chilled margins consist of very fine-grained plagioclase with variable compositions ranging from An₃₅ to An₆₃, in a devitrified-glass matrix crowded with submicron Fe-Ti oxides and sparse sub-calcic augite (Average cpx = [Mg₆₈Fe₆₀Ca₅₅Al₀₉Ti₀₂Mn₀₁] [Si_{1.91}Al_{0.09}O₆]). Within two meters of the north side of the diabase the dike contains single crystals of labradorite up to 10 cm across apparently sourced from a deeper magma chamber and transported (floated?) upwards during intrusion of the dike. Locally the chilled margin is altered to chlorite and very fine grained bright blue-green celadonite (**Photo 7D**), alkali-feldspar and dark red biotite.

Five samples collected from the central 35 m of the dike all consists of an olivine gabbro with a well-developed ophitic texture (**Photo 7B**). The mineralogy from the center includes both titaniferous augite (1-2wt% TiO₂) and titanaugite (>2wt% TiO₂) oikocrysts with pink and lavender

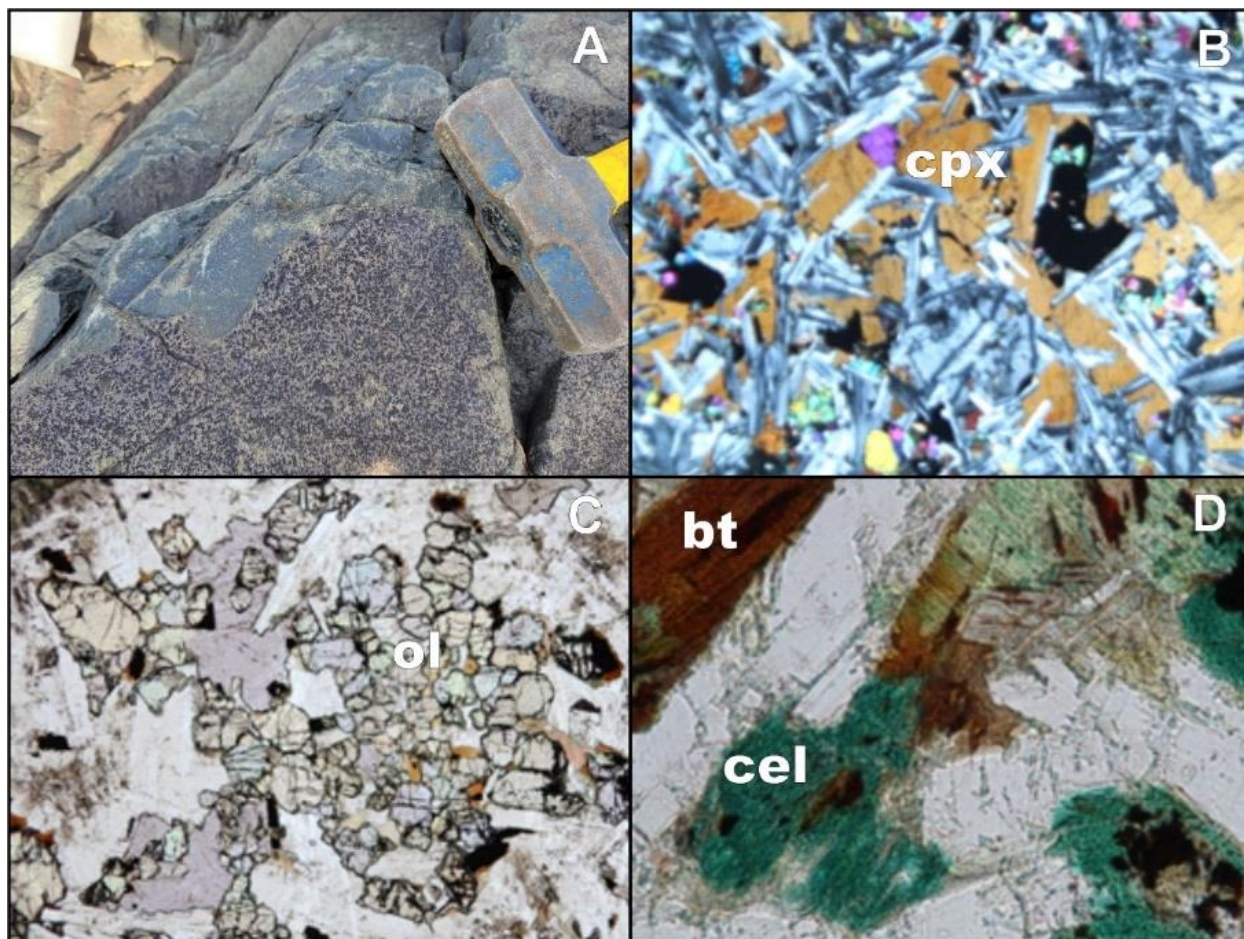


Photo 7 – Photographs of mid-continent rift gabbroic dyke at Wissota Dam. (A) Outcrop photo showing fractured and weathered surface of dyke. Weathered surface shows medium-grained texture. (B) photomicrograph in cross-polarized light (40x) showing ophitic texture. (C) Photomicrograph in plane-polarized light (40x) showing aggregate of euhedral to subhedral olivine (ol) crystals ($Fe_{0.40-45}$) in plagioclase and cpx matrix where cpx as pinkish purple pleochroism typical of the titanaugite composition. (D) Photomicrograph in plane-polarized light (100x) showing greenish blue celadonite (cel) replacing biotite (bt) in the transition zone between the chilled margin and dikes central olivine gabbro.

pleochroism, laths of normally zoned plagioclase with labradorite cores (An_{55-65}) and thin rims of andesine (An_{30-35}) and unusually large aggregates of euhedral to subhedral olivine containing over 50 individual olivine crystals (**Photo 7C**). The augite and biotite show little variation across the dyke but the olivine becomes progressively more Fe-rich towards the south with an average of $Fe_{0.43}$ in the north to $Fe_{0.35}$ near the southern contact. The opaque minerals are primarily ilmenite with titaniferous magnetite lamellae often rimmed with a highly titaniferous reddish orange biotite. In the transition zone between the chilled margin and the center 30 m of the dyke much of the

biotite is replaced (altered) with celadonite $K(Mg,Fe^{2+})(Al,Fe^{3+})[Si_4O_{10}](OH)_2$ with a brilliant blue-green color in plane polarized light (**Photo 7D**). Unusual olivine aggregates (glomerocrysts?) occur throughout the central 35m of the dyke and often consist of more than 50 crystals (**Photo 8**). In some of the olivine aggregates the minerals show at least some crystallographic alignment. Olivine within individual aggregates have a very narrow range of chemistry. In one aggregate 16 grains were analyzed and the average composition was $Fe_{0.47.5 \pm 0.2(2\sigma)}$; this standard deviation is about the

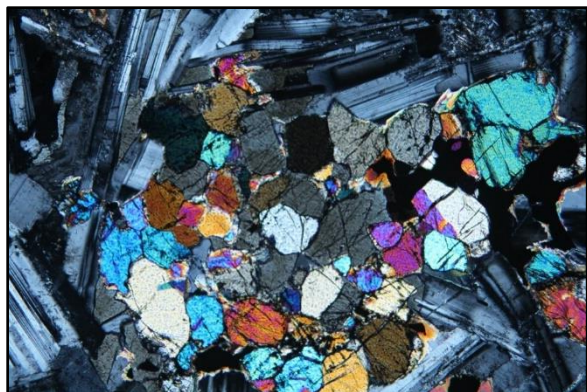


Photo 8 – Photomicrograph in cross-polarized light of olivine aggregate (glomerocrysts) showing consistent orientation of olivine crystals in the cluster. Gray crystals all have an optic axis almost perpendicular to the section. Magnification 100X.

same size as the analytical error for EDS analysis on olivine.

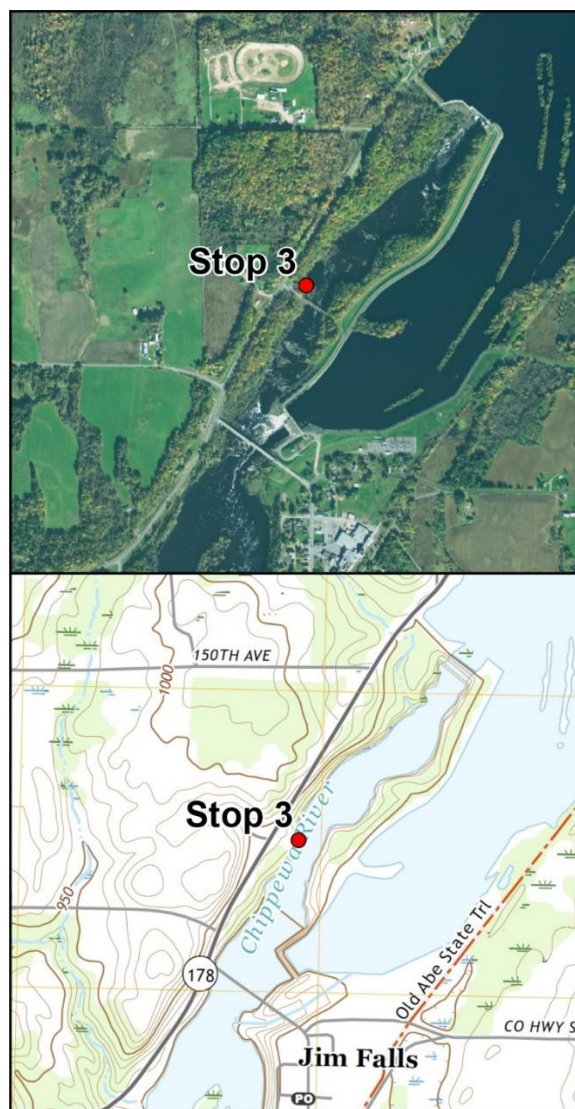
Olivine aggregates have been described from several basaltic conduits where they have been attributed to differential crystal movement during turbulent flow in an active conduit such as at Kilauea (Helz, 1987) or as xenocrysts extracted from a deforming cumulate. However, there is no reference to aggregates with such a large number of crystals. The texture and chemistry of the aggregates at Wissota come closest to matching olivine aggregates collected from lava flows at La Reunion (Welsch et al., 2013) which they ascribe to rapid dendritic crystal growth and ripening under a high degree of undercooling ($-\Delta T > 60^\circ\text{C}$) from low viscosity basalts.

Petrographic Interpretation: The dyke is sourced from a lower-level fractionated magma chamber of enriched basalt (E-MORB or alkali olivine parent) as evidenced by the olivine composition ($\sim\text{Fo}_{40}$), plagioclase (An_{60}) and titanite/ilmenite modal mineralogy. Plagioclase zoning from An_{60} cores to An_{30} rims suggests at least limited reaction with wall rocks. As a fractionated magma it seems likely that the ascending magma contained phenocrysts of both olivine and plagioclase that were kinetically fractionated by turbulent flow resulting in a chilled margin largely devoid of phenocrysts. The olivine aggregates form in equilibrium with

magma with limited chemical variation and could be produced by turbulent flow (synneusis) agglomeration or by a high degree of undercooling and ripening of dendritic olivine. It seems very likely that this dike was an active conduit for magmas reaching the surface to produce MCR lava flows even though Chippewa Falls is almost 200 km south of the main MCR rift axis. Geochemical results which are pending should help further constrain the system.

Stop 3 – Gneisses and Pegmatites at Jim Falls

Lat: 45.0549° Long: -91.2734°



This outcrop is located within the spillway of the hydroelectric dam near the community of Jim Falls. About 500 m north from the intersection of

Highway 178 and County Highway Y (the main road into the community of Jim Falls), there will be an old, abandoned bridge that that used to be the access bridge to the community. There is ample parking in front of this bridge. On the south side of the old bridge is a small foot path that leads down to the outcrops along the river. These outcrops are smoothly polished from the flooding at the dam. They are uneven and quite slippery when wet. If water levels are high, there are also outcrops immediately downstream of the spillway to the north.

This location has intensely folded amphibolite and biotite quartzofeldspathic gneiss that is intruded by granitoid intrusions and associated pegmatites (**Figure 9**). Intense shearing and metamorphism results in little preserved primary texture within the gneisses and amphibolites. On the east bank of the river is a gabbroic dyke associated with the mid-continent rift.

The outcrops at this stop are one of the original “Archean” exposures of the Marshfield terrane that was described in in Sims et al. (1989), but new data in the region casts doubt on that original interpretation. The gneisses at this location were assigned a U/Pb age of 2522 ± 22 by Sims et al. (1989). Little description of the data was provided in that original reference and the date itself was referenced as unpublished data from personal communication. Klier (2019) resampled the gneiss from the region and analyzed zircons using LA-ICPMS. The resulting data clearly shows a large spread of ages and a significant portion of those are Paleoproterozoic in age. There are clearly older sources of detritus for these meta-sedimentary rocks, some as old as 2841 Ma. However, the dominant source of detritus was Paleoproterozoic (**Figure 10**). Based on this new data, the Jim Falls region is not obviously an Archean crustal fragment.

Amphibolites and Gneisses

Myers et al. (1980) interpreted the amphibolites and gneisses in this region to be derived from mafic volcanic rocks and associated

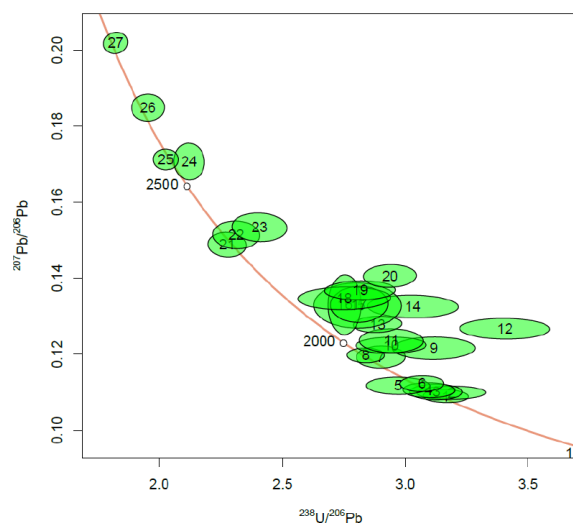


Figure 10 - Tera-Wasserburg concordia diagram of biotite quartzofeldspathic gneiss from Jim Falls. A wide spread of ages is suggestive of a detrital origin. Figure from Klier (2019).

sedimentary rocks. Trace element geochemistry from Klier (2019) was inconclusive in determining volcanic protolith because of low Ti abundance. Preliminary petrography from student projects at the University of Wisconsin-Eau Claire and Kleir (2019) seem to suggest that amphibolites are rather rare. Other geochemical results from ongoing research at the University of Wisconsin-Eau Claire are pending.

A biotite quartzofeldspathic gneiss was sampled by Klier (2019) for U/Pb geochronology. Based on recent field work, this rock appears to be the dominate lithology that exists in the immediate region around and under the bridge. Kleir (2019) describes the rock containing classic mylonitic textures and is comprised of quartz (60%), alkali feldspar (25%), biotite (15%), and trace zircon (**Photo 9A**). There are prominent bands of porphyroblastic quartz and cryptocrystalline biotite. Biotite is also present rarely as larger “destroyed” grains. Quartz has undulatory extinction and has undergone grain boundary migration recrystallization. Some feldspar grains display domino-type fragmented porphyroclastic textures. Portions of feldspar grains have diminished to sericite. Weakly

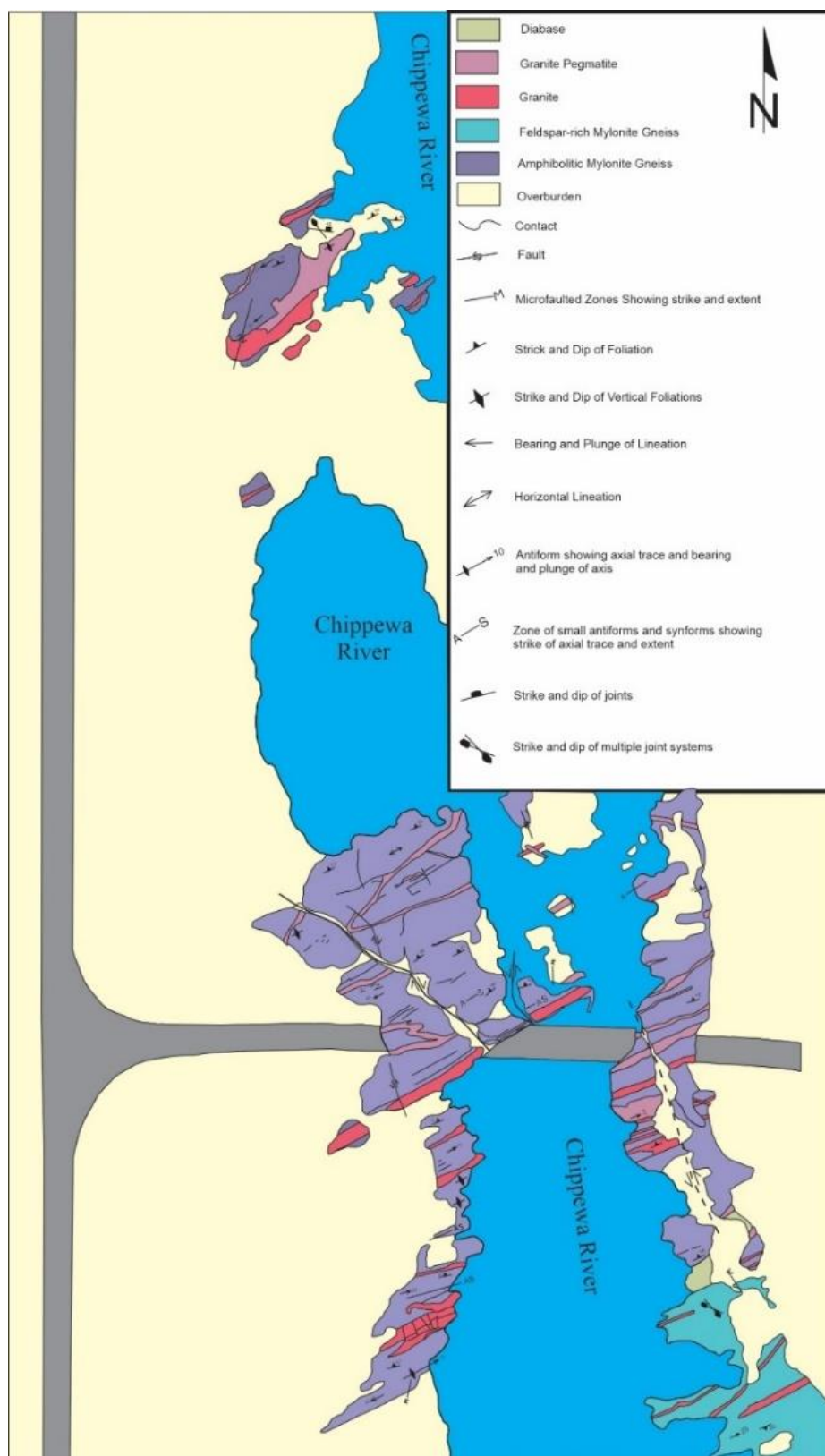


Figure 9 – Precambrian geologic map of Chippewa River near Jim Falls. Figure digitized from Myers et al. (1980).

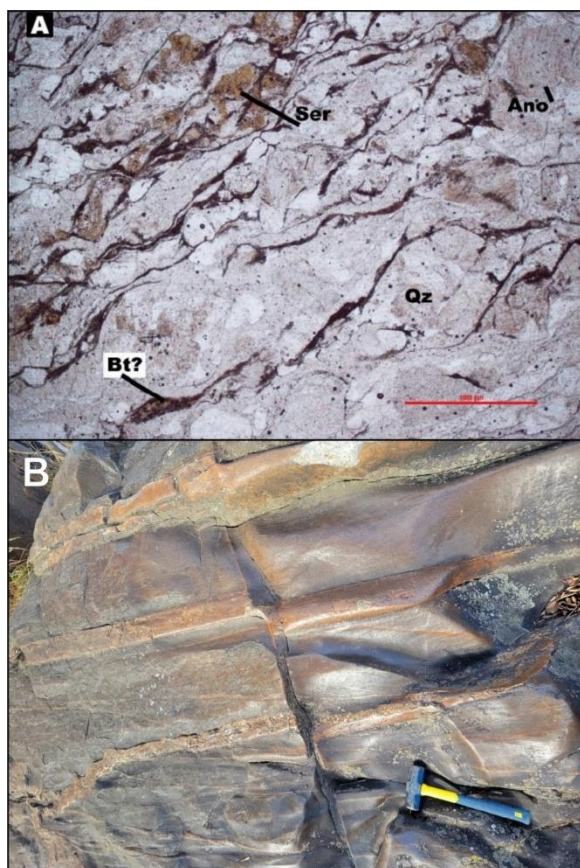


Photo 9 – (A) Photomicrograph in plane-polarized light of biotite quartzofeldspathic gneiss showing sericite-altered feldspar crystals and pronounced dynamic recrystallization of matrix. Foliation defined by elongation of grains and alignment of biotite. Photo from Klier (2019). (B) Outcrop photo showing gneiss intruded by boudinaged granitic dykes. Gneissic layering is very fine and difficult to see in this photo.

chloritized biotite bands define foliation (**Photo 9A**).

Garnetiferous hornblende gneiss and schist are folded with high-amplitude isoclinal folds with a persistent ENE strike. Small (F_2) folds plunge gently east-northeast. These are folded F_1 isoclinal folds, and a few hinges can be found in the outcrop. Some of the granitic pegmatites appear to be folded as well or are slightly boudinaged (**Photo 9B**), suggesting that pegmatites intruded prior to F_2 or were exploiting layering within the folded gneisses and amphibolites during emplacement. Xenoliths of

isoclinally folded amphibolite occur in the granitic rocks.

Granitoids and Pegmatites

Granitic rocks range in composition from trondhjemite to alkali feldspar granite. Pegmatite dike intrusion occurred at several stages of "granite" intrusion. Mineralogy includes alkali feldspar (65%) quartz (32%), plagioclase (3%), and trace zircon. The grains are subhedral to anhedral with intergrowths and granophyric textures occasionally present. Quartz grains appear stretched and strained and have undergone either subgrain rotation recrystallization or grain boundary migration recrystallization (Klier, 2019).

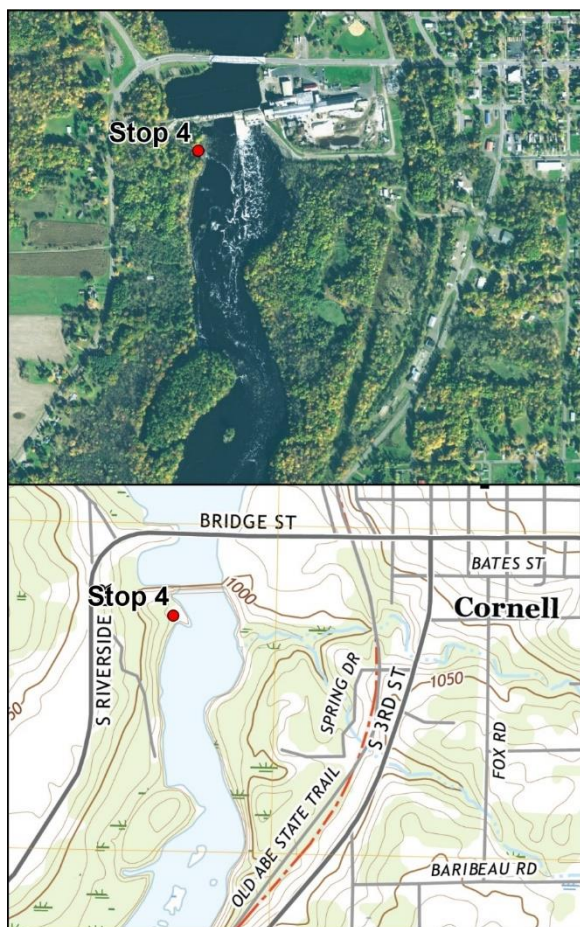
The older granitic rocks are foliated and locally mylonitized. Shearing and boudinage of pegmatite stringers transposed them into oblique concordance with lamination in the enclosing rocks. A rough correlation can be made between relative age and concordance of veinlets. Thinly laminated amphibolite was intruded by granite so that lenticular slices of the amphibolite were dragged en echelon away from the wall (**Photo 10**). The coarse granite pegmatite intruded under stress contains en echelon fractures filled with very coarse quartz.



Photo 10 – Typical intrusive relationship between pegmatite surrounding amphibolites and gneisses. Small pegmatite veinlets and en echelon fracturing along margins is common resulting in lens-shaped gneissic fragments.

Stop #4 – Amphibolites and Gneisses at Cornell Dam

Lat: 45.1625° Long: -91.1596°



This outcrop is located within the spillway of the hydroelectric dam near the community of Cornell. About 500 m southwest from the bridge into Cornell on Highway 178 is the Wisconsin Department of Natural Resources Ranger Station where there is ample parking. Just south of the Ranger Station is a small road (called Pine Point Road) that leads toward the water. There are foot trails and gated roads (accessible by foot) that lead toward the outcrops at the dam and by the river. These outcrops are uneven but are generally dry and easily traversed under normal river conditions. If water levels are high, outcrops can also be visited on the shoreline above the dam near the Municipal Works buildings in Cornell.

Myers et al (1980) described this location as a laminated (foliated) garnet amphibolite that

outcrops almost continuously for 4 km down the river. The amphibolite could also be classed as a gneissic, mafic hornblende tonalite or hornblende gneiss. This is also one of the few areas that this trip visits that you can potentially see primary depositional features! Immediately below the dam, the rock is a fine-grained amphibolite with elongate bulbous inclusions that appear to stretched pillows (**Photo 11A**) that contain local irregular to lens-shaped quartz-epidote nodules (**Photo 11B**).

The amphibolite is composed of subhedral to anhedral, lensoidal hornblende clusters (54%) with coarse, lensoidal porphyroclasts of twinned plagioclase (28%) and fine-grained quartz. Banding in the amphibolite is cut by lenticular segments of granite and quartz veinlets. Garnets

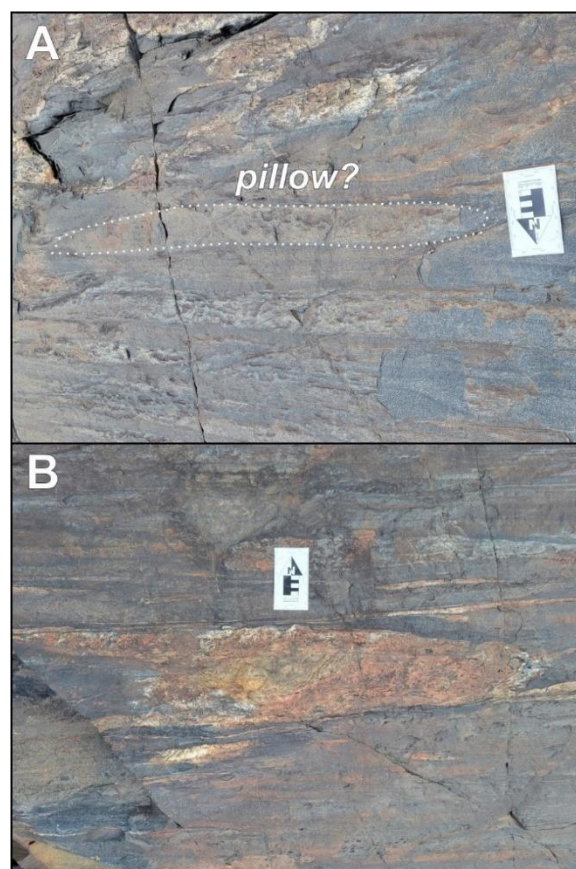


Photo 11 – Flow-like features in the amphibolites at Cornell Dam. (A). Stretched pillow-like structures with cm-scale darkened pillow margins. (B) Irregular quartz-epidote nodules that are common in submarine or hydrothermally-altered submarine flows.

in the amphibolite tend to be moderately poikiloblastic with minor rotational features. The distribution of garnet clusters shows little relation to banding. Trace element chemistry of these rocks show clear tholeiitic trends and flat REE patterns on normalized diagrams (**Figure 5**). Further downstream from the dam, the rock becomes notably lighter in color and there appears to be a lower percentage of amphiboles. These rocks share similar trace element patterns (**Figure 5**) and are interpreted to be genetically related. The reason for the change in texture may be due to increase structural modification and gneissic banding development.

The outcrops are also intruded by mafic dykes that clearly cross-cut the dominant foliation (**Photo 12**). These dykes trend N40°W and are approximately 30-50 cm in apparent thickness with no obvious chill margin. Since these dykes cross-cut the structural fabric, they are assumed to be related to the mid-continent rift. However, no petrography or chemistry has been done to confirm this hypothesis.



Photo 12 – Gabbroic dyke intruding through amphibolites at Cornell Dam.

Stop 5 – Amphibolites and Deformed Diorite at Holcombe Dam

Lat: 45.2251° Long: -91.1289°

As time permits, each side of the river at Holcombe Dam has different rock types to examine, but are vastly different approaches to see them. To visit the outcrops on the north bank



of the river, there is a small vehicle parking area and footpath that leads to the dam on 260th Avenue about 100 m west of the intersection with County Highway M. This trail will take you to the dam and carefully navigate to the north bank of the river downstream of the dam. To access the south bank of the river, drive to the end of Irvine Avenue before it turns into a private driveway. There are numerous small foot paths that will lead to the south bank of the river.

The outcrops at this location are considered part of the Pembine-Wausau terrane, or is it? While the location of the Eau Pleine Shear Zone becomes problematic in this region, Sims et al. (1989) consider the Jump River Shear Zone the northern boundary of the Marshfield Terrane. Magnetic lineaments mark this shear zone and extend it close to these outcrops. Depending on

the map, the shear zone lies just north or just south of the outcrops at this location (e.g. Mudrey et al., 1987). The rocks at this stop are a gneissic quartz diorite intrusion and an amphibolite schist (**Figure 11**). So, Marshfield or Pembine-Wausau terrane? The newest geochronology from the region is inconclusive.

The rock exposed on the southern bank of the river below the dam is a foliated amphibole-biotite schist (**Photo 13**). Klier (2019) describes this rock as banded at the microscopic scale. Quartz grains are well banded, fairly subhedral to anhedral and feature undulatory extinction. Their boundaries are somewhat irregular and indicative of bulging recrystallization. Amphibole grains are hornblende to tremolite. Biotite appears as brown to light green grains and typically feature

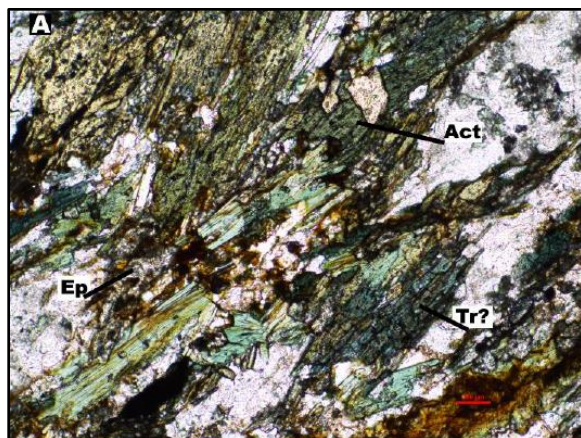


Photo 13 – Photomicrograph in plane-polarized light of amphibole-biotite schist with trace amounts of epidote in a quartzofeldspathic matrix. Figure from Klier (2019).

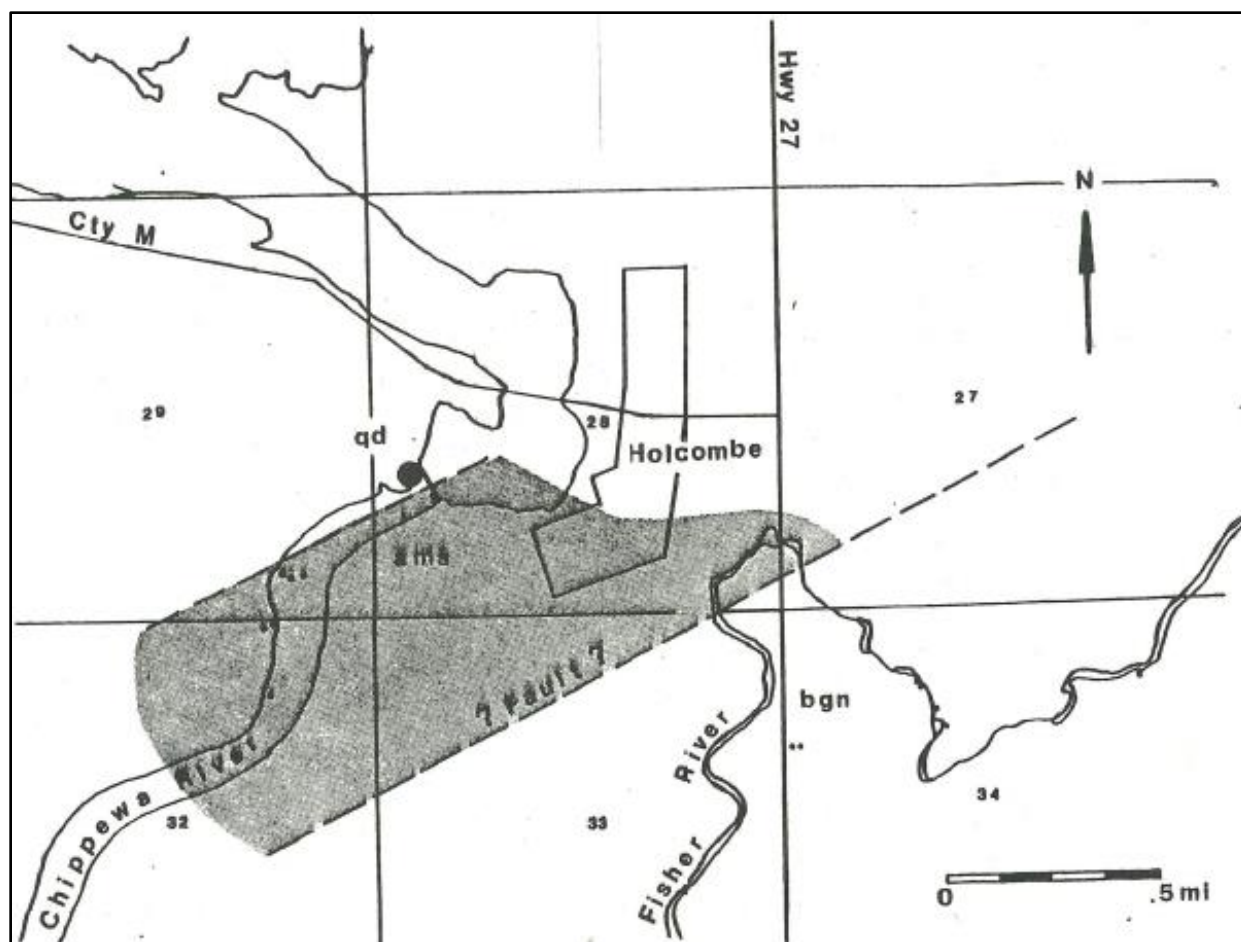


Figure 11 – Precambrian geologic map of the Holcombe Dam region. Unit Abbreviations: qd: quartz-diorite, bgn: banded gneiss, ams: amphibolite schist. Figure from Myers et al. (1980).

shear bands with cryptocrystalline biotite cross-cutting crystals. Klier (2019) obtained a U/Pb zircon age via LA-ICPMS of 1858 ± 1.0 Ma (**Figure 12**). This age does not definitively put the rocks in this region in Marshfield or Pembine-Wausau terrane. There does not appear to be any Archean inherited zircons, as one might expect if Penokean magmas were overprinting an Archean crustal block.

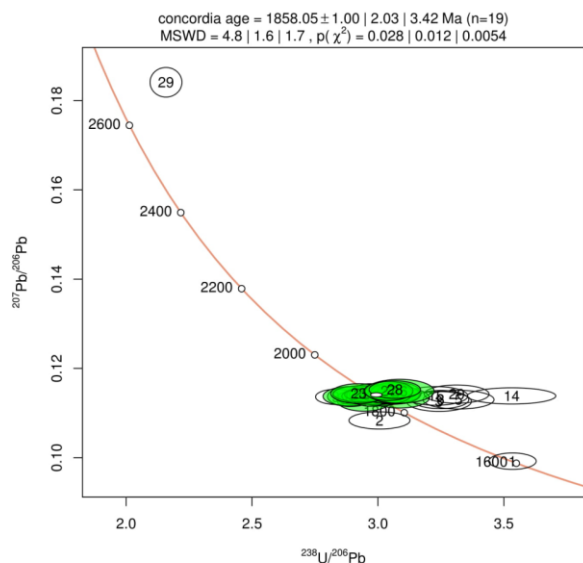


Figure 12 – Tera-Wasserburg concordia diagram of amphibolitic schist on the south bank of the Chippewa River near Holcombe dam. Figure from Klier (2019).

On the north bank of the river, the outcrop is predominately a synkinematic quartz diorite (**Photo 14; Figure 13**). The quartz diorite is a medium-grained, dark to medium grey rock with rusty weathering surfaces. It is faintly foliated and has white discontinuous bands and lenticles which are more quartz rich than the rest of the rock. Quartz diorite is composed of plagioclase (32-51%), quartz (11-31%) and mafic minerals (12-33%). Mafic minerals range from entirely hornblende to entirely biotite. The quartz diorite is cut by medium-grained granite pods with migmatitic contacts and by finer-grained dykes with sharp contacts. The granite is a pink, faintly foliated rock which locally contains porphyritic microcline grains reaching 1 cm in size. Granitic rocks consist of plagioclase, microcline and



Photo 14 – Outcrop of quartz diorite on north bank of Chippewa River at Holcombe Dam.

quartz, with minor amounts of biotite, muscovite and epidote.

The quartz diorite contains two types of inclusions: hornblende rich ultramafic inclusions and spotted mafic inclusions. Ultramafic inclusions occur along the northwest portion of the exposed quartz diorite, generally less than 0.5 meters in length, although one is at least 2 meters long. Ultramafic inclusions are composed of 75-85% hornblende and 11–13% biotite with a small amount of plagioclase (meta pyroxenites?). Chlorite occurs as an alteration product of biotite and less commonly of hornblende and can compose more than 20% of the rock.

Stop 6 – Tonalites and quartz diorites at Cadott Bridge

Lat: 45.9535° Long: -91.1508°

This outcrop shows is easily accessible under the Main Street bridge in Cadott. Just north of the bridge near the Main Street-Yellow Street intersection there is a parking area on the north bank of the river. From this parking area, there are footpaths that lead to the waters edge.

The predominant rock type here is foliated biotite quartz diorite to biotite tonalite (**Figure 14**) composed of plagioclase (An₂₅₋₃₅, 30-55%), quartz (10-40%), hornblende (0-25%) and biotite (0-15%) (Myers et al. 1980). Mafic minerals are partly replaced by chlorite (of several varieties), epidote, and sericite. Magnetite (1-5%) is a by-

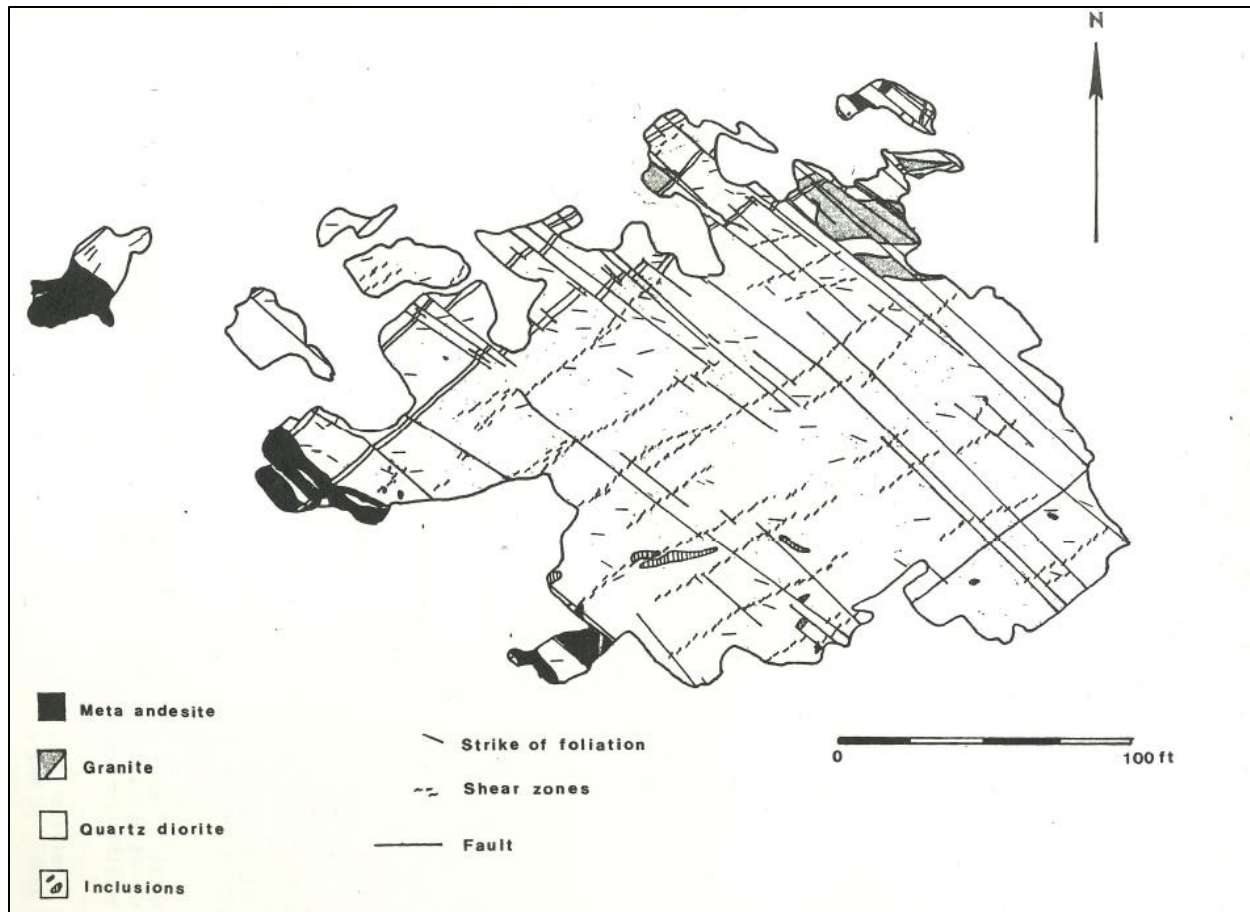


Figure 13 – Detailed outcrop map of the quartz diorite gneiss on the north bank of the Chippewa River at Holcombe Dam. Figure from Myers et al. (1980).

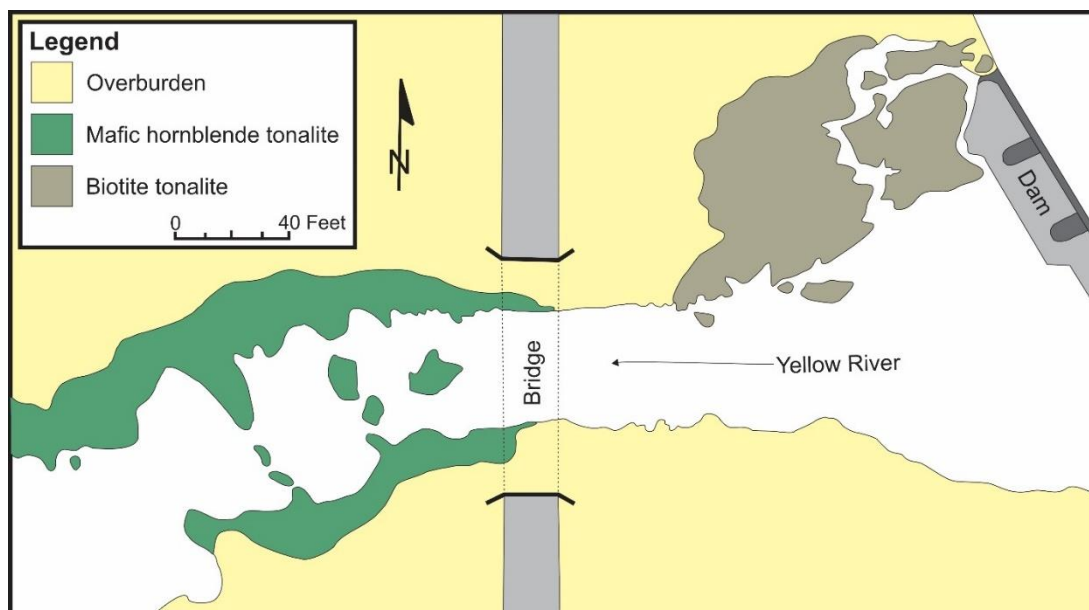
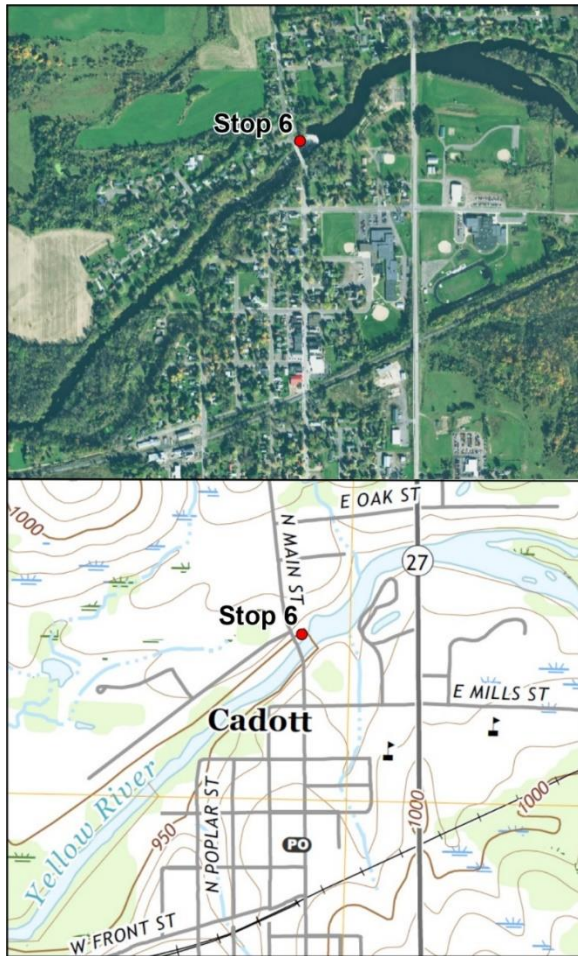


Figure 14 – Precambrian geologic map of the region downstream of Cornell Dam. Figure modified from Myers et al. (1980).



product in the chloritization of hornblende. The tonalites have been mylonitized (**Photo 15**) and locally recrystallized and contain lenticular xenoliths of chlorite and epidote-rich metavolcanic(?) rock (**Photo 16**). The older cataclastic foliation is axial-planar to isoclinally folded pegmatite, aplite, and quartz layers. These rocks are cut by a pervasive N65-75°W trending foliation and mylonitic shear zones.

West of the Yellow River bridge, foliated biotite tonalite encloses angular xenoliths of hornblende tonalite or amphibolite containing strongly deformed aplite and pegmatite stringers. Isoclinally folded quartz, aplite, and pegmatite veinlets exist as angular xenoliths in a lighter biotite tonalite.



Photo 15 – Mylonitized biotite tonalite at the Cadott Bridge on the Yellow River.



Photo 16 – Photograph of elongated xenoliths (?) of chlorite-rich metavolcanic rocks in biotite tonalite at Cadott Bridge on the Yellow River.

Acknowledgements

Despite decades of regular visits from groups from the University of Wisconsin-Eau Claire, the most extensive detailed maps and rock descriptions were provided by Paul Myers and collaborators in the 1980 ILSG guidebook (Myers et al, 1980). There is some recent research activity in the region, but between new but pending analyses and future ambitions, the descriptions and maps provided in that ILSG guidebook are the most detailed and accurate for the region. A lot of the geologic descriptions have been updated and figures have been digitized while adding new data and insights where available.

In addition, the authors of this guidebook would like to thank the countless undergraduate and graduate students that have worked on these outcrops and have continued to inspire new work in the region. Specific acknowledgement is deserving to Matt Leahy and his efforts in digitizing figures and compiling geochemistry for the guidebook.

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Field Trip 2 – Wisconsin’s Paleozoic stratigraphy and tour of Crystal Cave

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Introduction

The Cambrian-Ordovician strata exposed in western Wisconsin were deposited during the major Sauk and Tippecanoe marine transgressions onto the interior of the Laurentian continent (Sloss, 1963). These rocks compose the regional aquifer system, host disseminated sulfide mineralization that contribute to groundwater contamination, and are locally mined as proppant for fracking in the oil and gas industry. Additionally, variable hardness of these units in part controls the formation of ledges and hillslopes in the fluvially-dissected Driftless Area of southwestern Wisconsin. During this field trip, we will focus on Cambrian and lower Ordovician strata of the Sauk sequence (**Figures 1 and 2**). We start our day touring Crystal Cave, a cave system developed along joints within the Ordovician Prairie du Chien Group dolostone. For the rest of the day, we will visit outcrop exposures of the Cambrian Jordan Formation, Tunnel City Group, and Wonewoc Formation sandstones, and if time permits- the Eau Claire and Mount Simon Formations. We hope this field trip will provide an opportunity to discuss similarities and differences between units deposited on the western side of the Wisconsin Arch and those deposited on the eastern side, where field trip authors have focused much of their work. Additionally, we welcome and encourage discussion between participants that have knowledge of or experience working with these stratigraphic units.

A very brief geologic history of the Cambrian-Ordovician strata in Western Wisconsin

Regional depositional model and setting

CORRELATION OF MAP UNITS

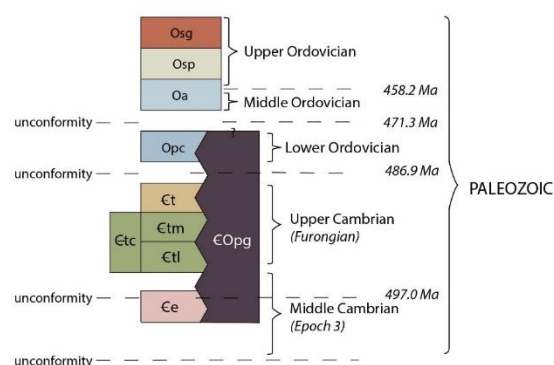


Figure 1. Correlation of map units showing relative ages of Cambrian-Ordovician units. COpg: Parfrey's Glen Formation, Ce: Elk Mound Group, Ctl: Lone Rock Formation of the Tunnel City Group, Ctm: Mazomanie Formation of the Tunnel City Group, Ctc: Tunnel City Group, Ct: Trempealeau Group, Opc: Prairie du Chien Group including the Oneota and Shakopee Formations, Oa: Ancell Group, including the St. Peter and Glenwood Formations, Osp: Sinnipee Group, including the Platteville, Decorah, and Galena Formations. From Stewart (in revision). Ages from Gradstein et al. (2020).

Cambrian-Ordovician strata in the southern Lake Superior Region were deposited on an essentially flat continental shelf in a shallow epeiric sea well within the Laurentian continent (**Figure 3**, Runkel et al., 2012, 2020). These strata overlie Precambrian bedrock of variable ages across the Great Unconformity, a surface characterized by locally significant topographic relief and weathering and exposed in outcrops around the Eau Claire area. The regional paleogeography that controlled sediment source to sink was defined by several structural highs, including the Transcontinental Arch, Wisconsin Dome, and Wisconsin Arch, and several basins, including the Hollandale Embayment, Illinois Basin, and Michigan Basin (**Figures 3 and 4**,

Bedrock stratigraphic units in Wisconsin

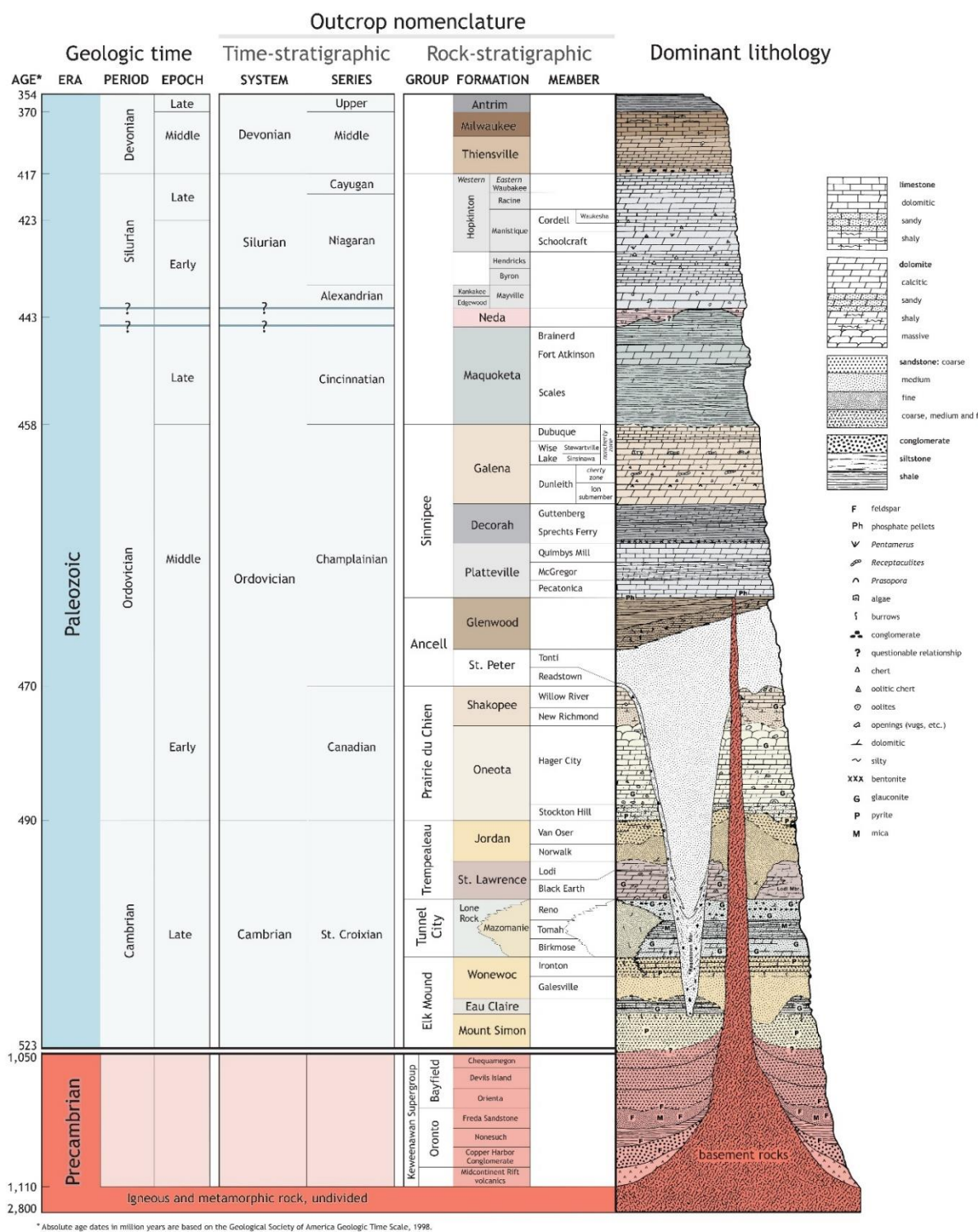


Figure 2. Generalized stratigraphic column of Wisconsin. From: [Bedrock Stratigraphic Units in Wisconsin - Bedrock Stratigraphic Units in Wisconsin \[small\] - WGNHS.](#)

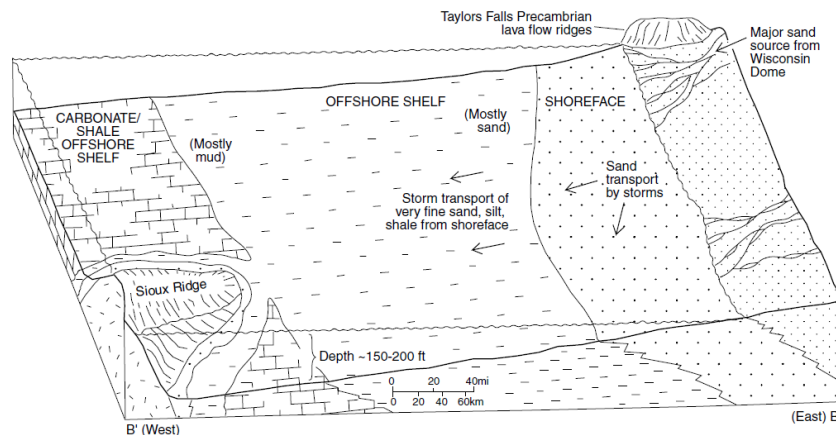


Figure 3. From Runkel 2020, Figure 4. This figure illustrates a depositional model developed for southeastern Minnesota. The Cambrian- Ordovician strata in Wisconsin are thought to have been deposited in a similar fashion.

Runkel et al., 1998). The field stops we will visit in western Wisconsin lie west of the Wisconsin Arch and straddle the eastern edge of the Transcontinental Arch and the southwest flank of the Wisconsin Dome. These structural highs were periodically subaerially exposed and eroded during deposition of Paleozoic units.

Cambrian-Ordovician siliciclastic and carbonate units were deposited in a nearshore, sandstone-dominated inner detrital belt that passed offshore into a relatively deeper water moat, which in turn transitioned into a carbonate-

dominated Great American Carbonate Bank (**Figures 3 and 5**; Runkel et al., 2012). Sandy Cambrian sediments of the Mt. Simon, Wonewoc, and Jordan Formations were deposited in shoreface, aeolian, wave-, and tide-influenced settings within the inner detrital belt. Mixed, fine-grained sandstone, siltstone, shale, and carbonate of the Eau Claire Formation, Trempealeau and Tunnel City Groups were winnowed and trapped within a transitional, relatively deeper water moat that separated the inner detrital belt from the Great American Carbonate Bank (Runkel et al., 2012). Dolomite of the Prairie du Chien Group was deposited in relatively shallow water, subtidal to peritidal settings on this carbonate bank. Interfingering sandstone, shale, and carbonate record marine transgressions and regressions that caused reciprocal expansion and contraction of the facies belts. During sea level rise, the carbonate bank advanced landward as siliciclastic-dominated nearshore environments were drowned.

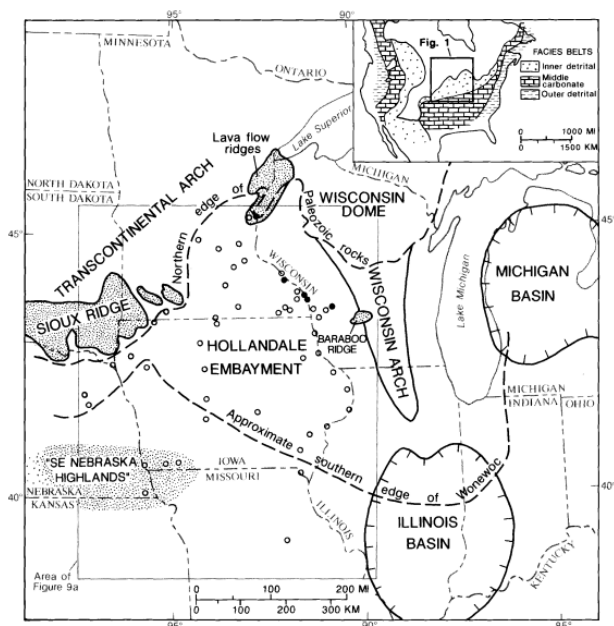


Figure 4. from Runkel and others 1998, regional map showing locations of the Wisconsin Dome, Wisconsin Arch, Transcontinental Arch and Hollandale Embayment. Other depositional basins are shown on map (Michigan and Illinois Basins), as well as regional extent of Paleozoic units.

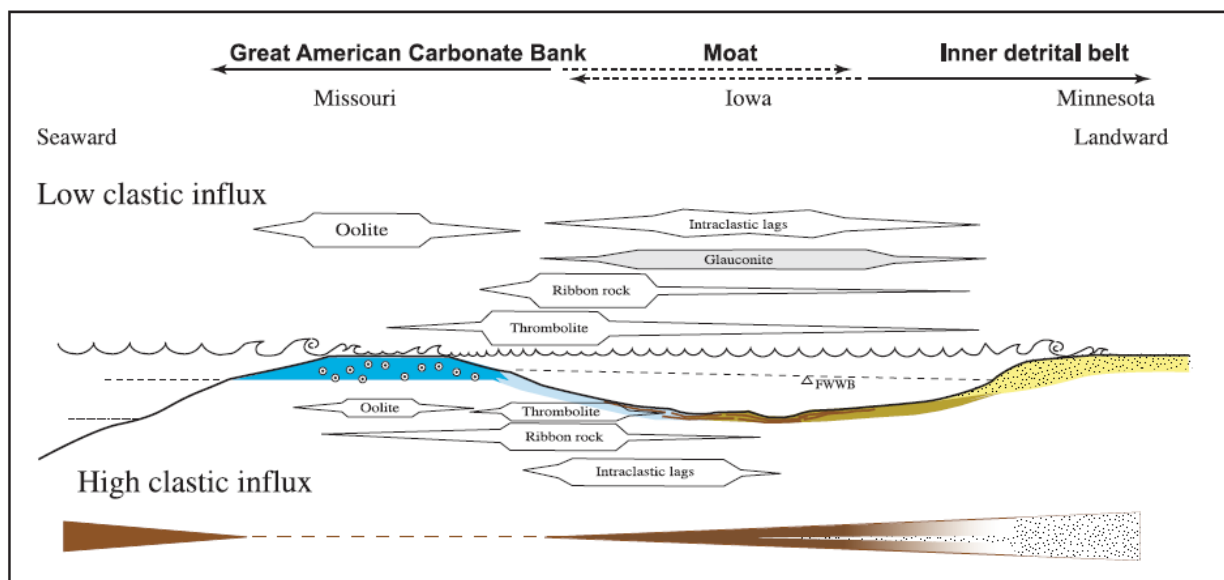


Figure 5. from Runkel et. al., 2012 showing the depositional environments that produce interfingering of different Cambrian-Ordovician siliciclastic and carbonate units across the Midwest.

Conversely, during sea level fall, sandy nearshore facies of the inner detrital belt expanded seaward, limiting carbonate deposition.

The Wisconsin Arch and its influence on Cambrian-Ordovician strata

Strata deposited in areas east (for example, Dodge, Fond du Lac, and Jefferson Counties, Wisconsin) and west (for example, the outcrops we will visit today) of the Wisconsin Arch (**Figure 4**) were deposited in different sub-basins and tapped different local sediment source areas. In addition, the Dodge, Fond du Lac, and Jefferson County map areas were situated in more proximal locations on the Wisconsin Arch relative to today's field stop locations. Therefore, the eastern sections include more pronounced exposure surfaces, condensed, or eroded sections, and typically include thinner and less abundant fine-grained intervals. **Figure 6** shows a generalized stratigraphic column for Jefferson County (east of Wisconsin Arch), with accompanying pXRF elemental data. **Figure 7** shows a stratigraphic column from Trempealeau County in western Wisconsin, south of this trip's field stops, and west of the Wisconsin Arch.

Structural observations on the Cambrian-Ordovician strata in Wisconsin

Paleozoic sedimentary rocks were gently folded and faulted in the Paleozoic, probably related to far-field effects of continental margin orogenic events. Structures in Wisconsin rarely exceed 200 feet in structural relief. Recent mapping in Wisconsin and Minnesota suggests folds and faults are probably related to reactivation of much older Precambrian structures (**Figure 8**). Deformation probably occurred in at least two pulses: once during the Ordovician (Mossler, 2006; Steenberg and Retzler, 2016; Stewart E.K., 2021) and at least once later in the Paleozoic (Heyl and others, 1959; Carlson, 1961). The importance of these folds and faults for groundwater studies is a topic of active interest. In northern Illinois, sandstones in the core of the Sandwich Fault zone have an order of magnitude reduction in horizontal hydraulic conductivity compared to the surrounding rocks (Hadley and others, 2020). In eastern Wisconsin, the Beaver Dam anticline is associated with a statistically significant increase in detection of dissolved arsenic in groundwater wells (Stewart E.D. and others, 2021).

Regional and county scale mapping

The most recent regional map for West-Central Wisconsin was published in 1988 by Bruce

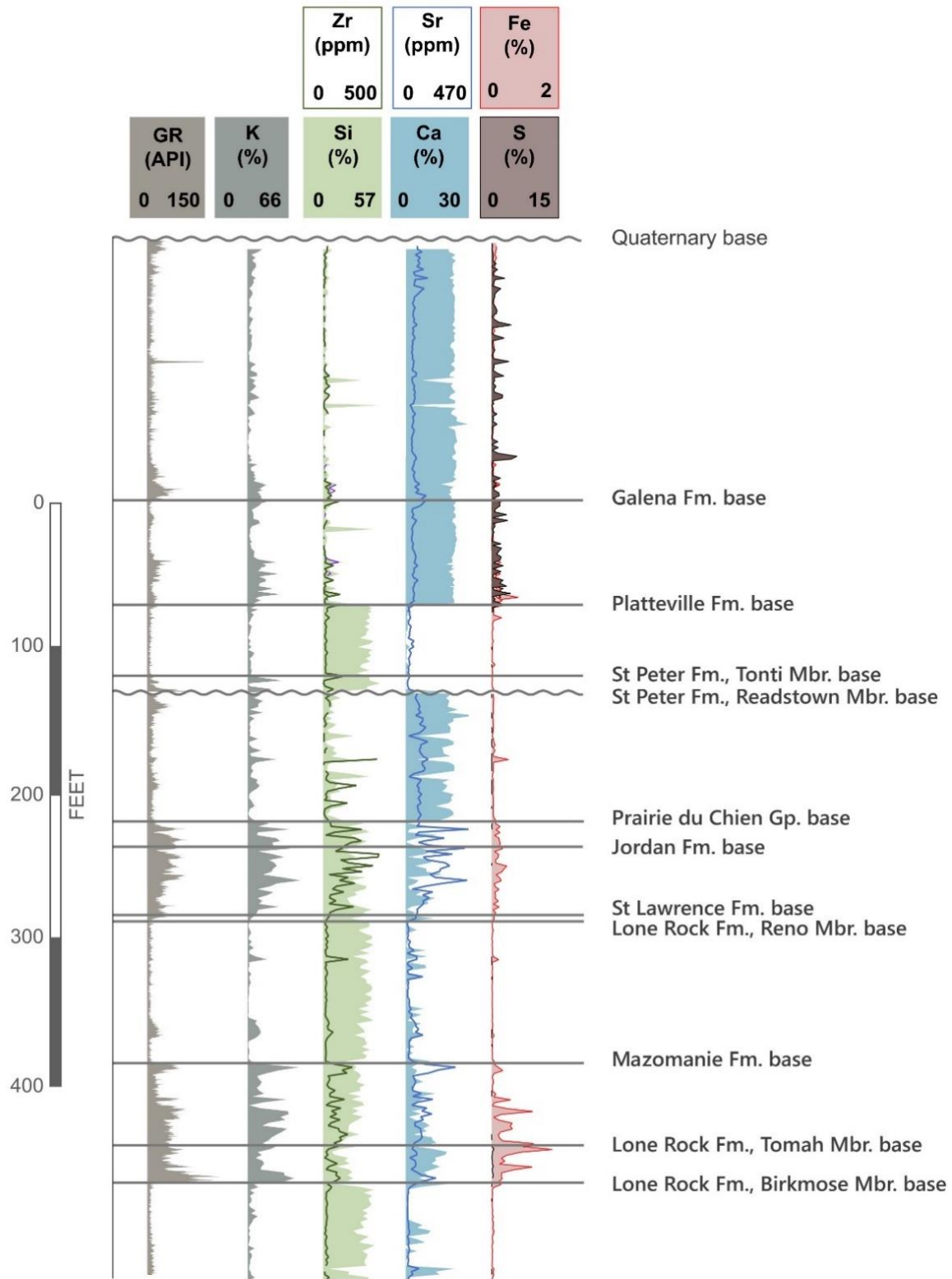


Figure 6. From Stewart (in revision), Bedrock geology of Jefferson County. Jefferson County is east of the Wisconsin Arch.

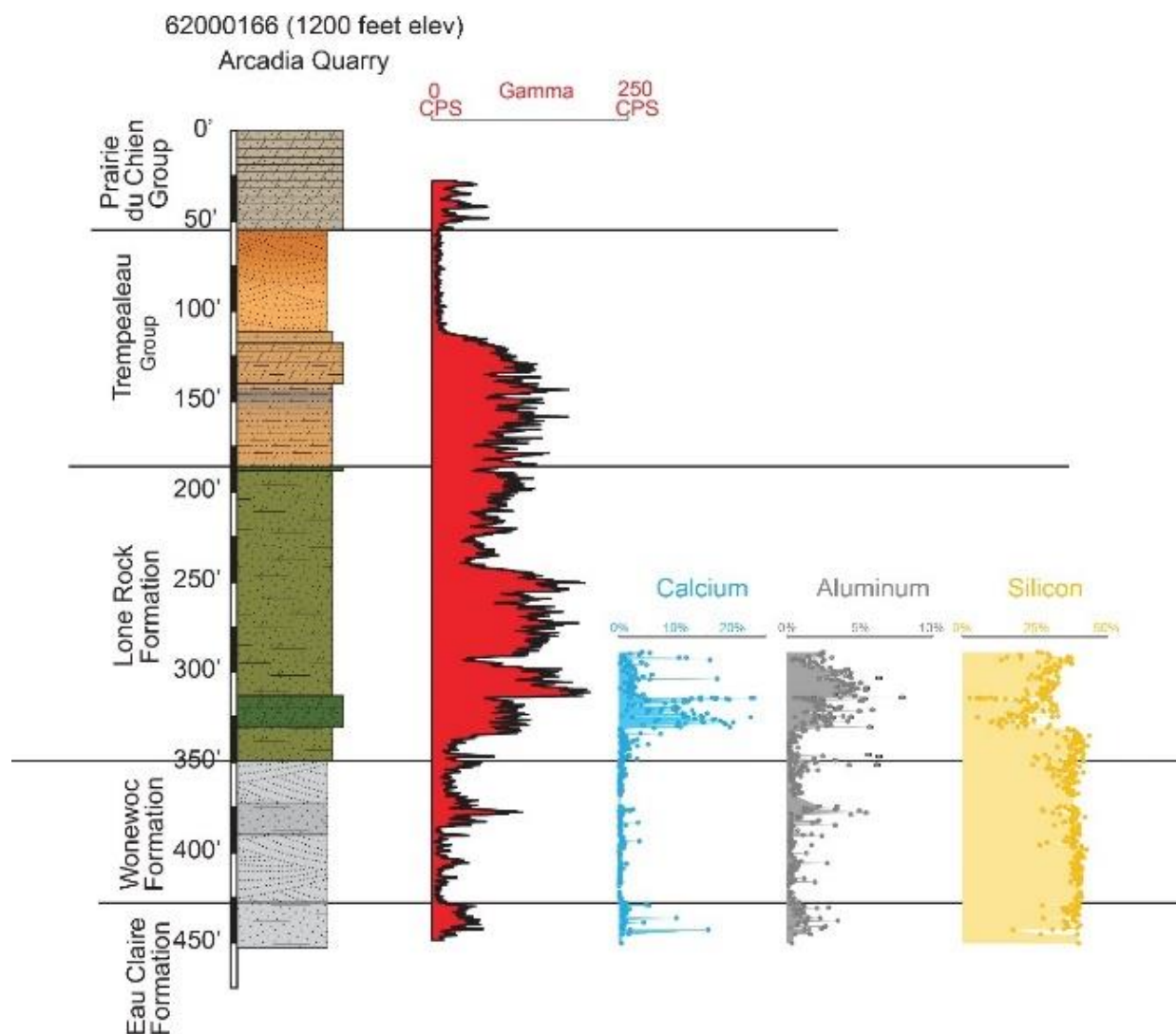


Figure 7. Core log, Gamma Ray log, and pXRF logs from the Arcadia core, Trempealeau County. Modified slightly from Zambito et al. (2018). Trempealeau County is west of the Wisconsin Arch.

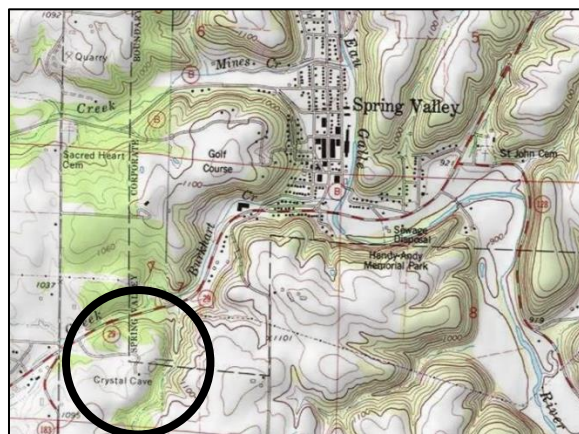
Brown (WGNHS). This map (**Figure 9**) includes all of the Paleozoic units we will see today, as well as the older Proterozoic and Archean rocks that make up the bedrock to the east of Eau Claire. Many of the stops for this field trip were found using the Cambrian contacts from this map.

Field Trip Stops

Stop 1: Crystal Cave, Spring Valley, WI (Contributed by Ian Orland, WGNHS)

UTM location for stop (559180.94E, 4964692.71N)

We will be touring the cave with staff from Crystal Cave. The tour is moderately strenuous



and includes walking, ducking, and climbing 7 stories. Please exercise caution while inside the cave as surfaces may be uneven. Below is a brief

synopsis of a recent collaboration between WGNHS and UW-Madison's Geoscience Department on speleothems from southern Wisconsin:

Caves are fascinating natural features, and can preserve geologic records of past environments. Relatively recent advances in the methods and precision of geochemical analyses have established cave formations (speleothems) as important scientific tools for understanding climate changes of the last 500,000 years.

A number of groups have studied the geochemistry of speleothems in the Lake Superior region. In Wisconsin, much of this work has happened at Cave of the Mounds in Blue Mounds, WI, just outside of the terminal moraine of the Laurentide Ice Sheet and some 20 miles southwest of Madison. While that cave is not the destination for this field trip, this section is intended to highlight the types of information we can learn from caves like Crystal Cave. Both caves are privately-owned show caves that were opened for tours in the late 1930s/early 1940s.

Crystal Cave is situated in Prairie du Chien Group dolomites of the Early Ordovician (~475 Ma), while Cave of the Mounds is in Sinnipee Group dolomites of the Middle Ordovician (~465 Ma). The formation ages of passages in each cave are poorly constrained. Stalagmites and stalactites from Cave of the Mounds, however, have recorded environmental signals for >250,000 years.

Cave of the Mounds: permafrost record

Researchers from UW-Madison collected the first seven stalagmite samples in 2015 for modern U-Th geochronological analysis at UM-Twin Cities. Initial results prompted further sampling and analyses; Batchelor et al. (2019) reports 141 U-Th dates from 19 cave carbonate (speleothem) samples ranging from 250–2 ka. The temporal distribution of these ages revealed hiatuses of stalagmite growth in the cave during both of the last glacial maxima, demonstrating the presence and duration of permafrost (**Figure 10**). Notably,

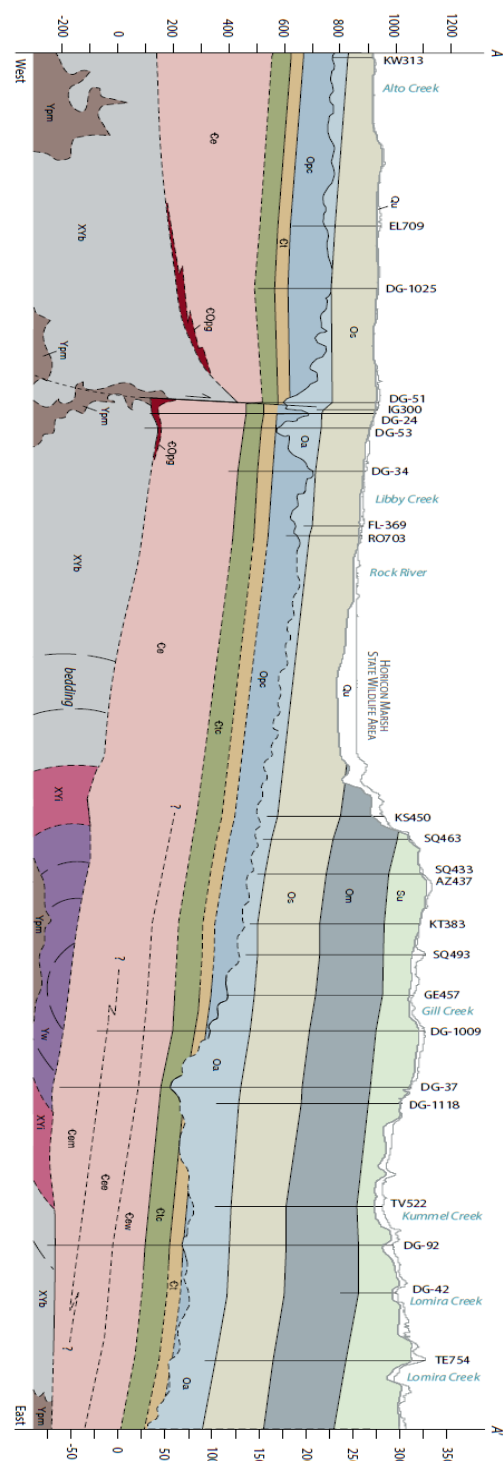


Figure 8. Cross-section from Dodge County, eastern flank of the Wisconsin Arch, south-central Wisconsin. Note offset of Precambrian basement and Cambrian Elk Mound Group (Ce) through Ordovician Prairie du Chien Group (Opc) and subtle folding of younger units. From Stewart E.K. (2021).

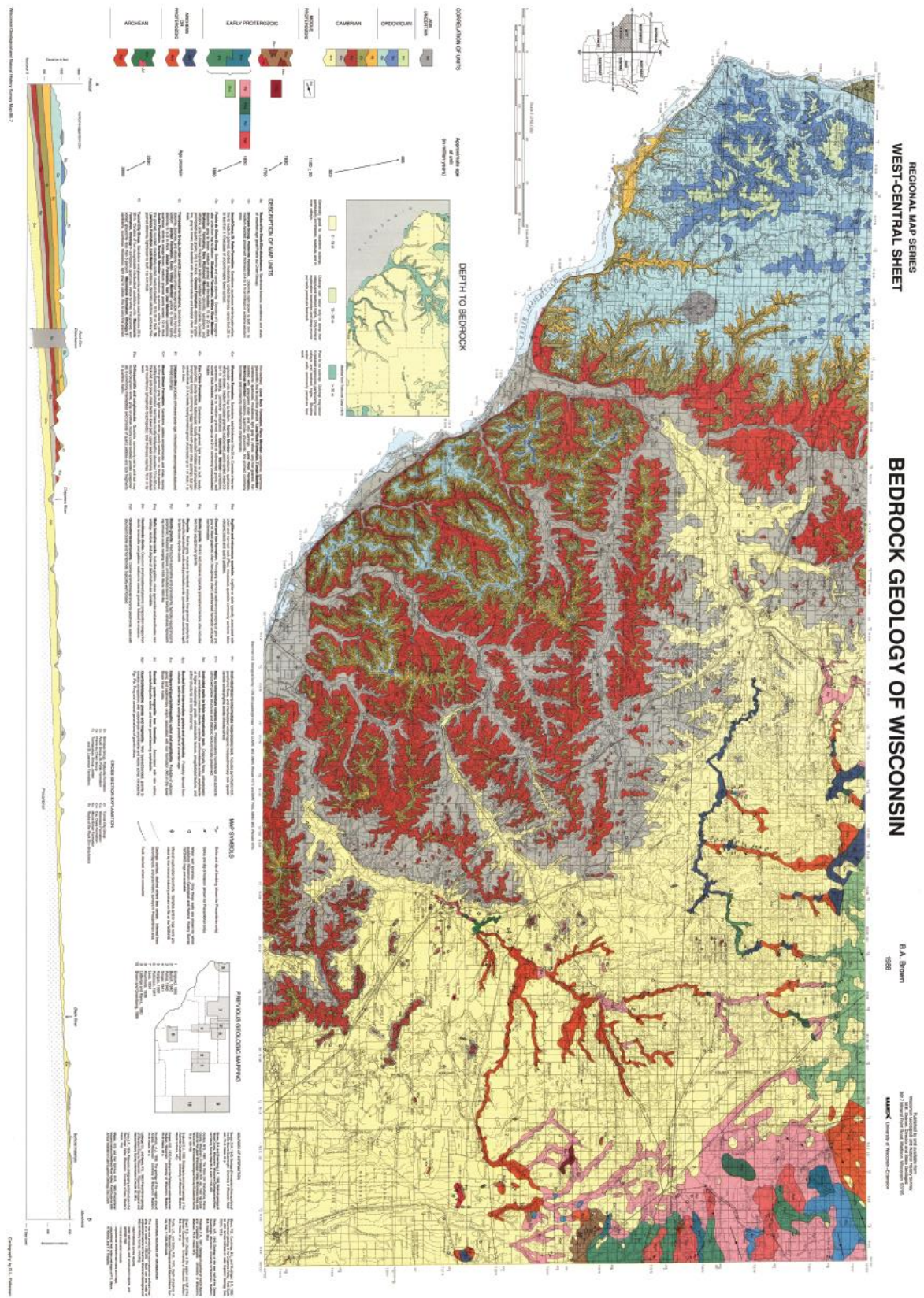


Figure 9. Bedrock geologic of west-central Wisconsin from Brown, 1988.

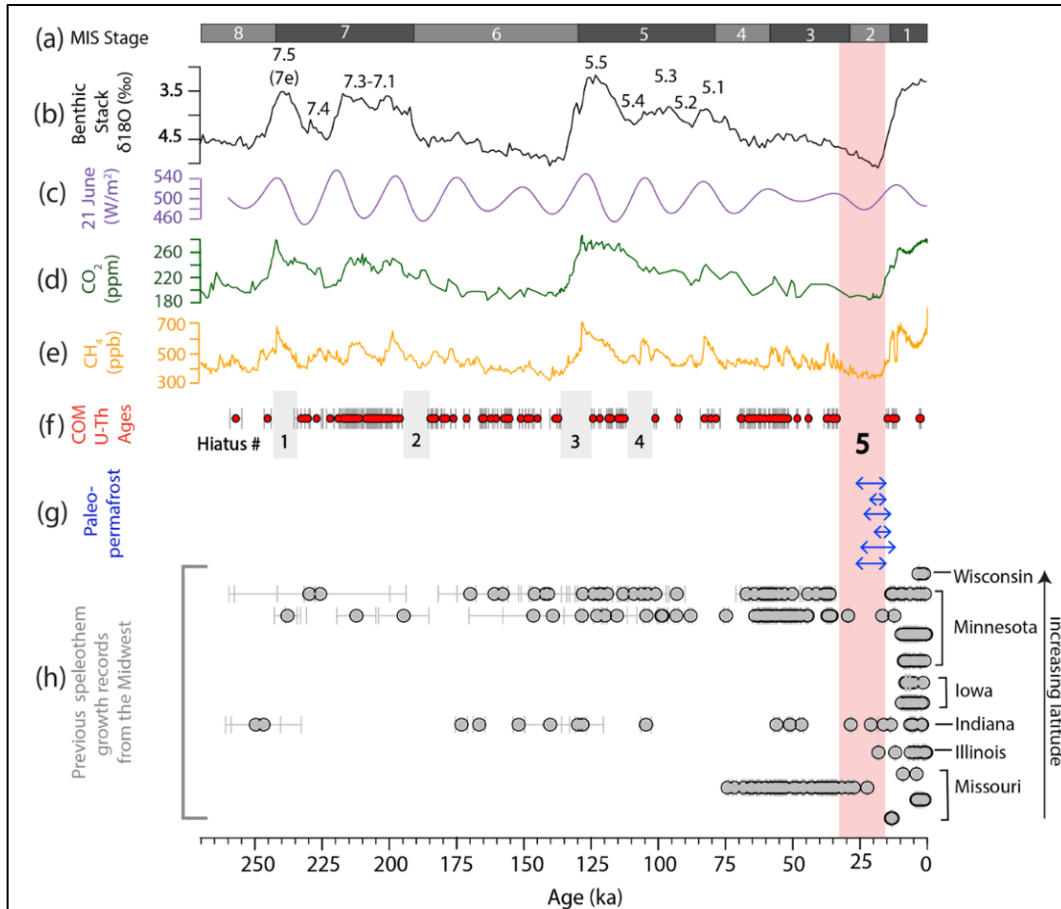


Figure 10 (*from Batchelor et al., 2019). Speleothem U-Th ages at Cave of the Mounds (COM) in context with regional and global paleoclimate records. (a) MIS boundaries (odd numbers=interglacial periods, even numbers=glacial periods). (b) Stacked records of $\delta^{18}\text{O}$ (‰) from benthic marine foraminifera annotated with MIS substage names (Lisiecki & Raymo, 2005). (c) Summer insolation (21 June at 43°N). (d) Atmospheric CO_2 (ppm) and (e) CH_4 (ppb) concentrations. (f) U-Th ages from COM speleothems with associated 2σ uncertainties (this study) and statistically significant growth hiatuses (gray and red vertical bars). (g) Paleo-permafrost reconstructions based on geomorphic features in Wisconsin, including ice wedge casts and polygons (Clayton et al., 2001). (h) U-Th dates of speleothems from caves in the Midwestern United States in order of decreasing latitude. References provided in the main text. MIS = Marine Isotope Stage

the 18 ky duration of the growth hiatus at MIS 2 was much longer than the hiatus that overlaps MIS 6 (5 ky), consistent with more extensive continuous permafrost in the region during the last glacial period.

Cave of the Mounds: Decadal warming events during the last glacial period

Earlier this year, Batchelor et al. (2023) published a record of the oxygen isotope ratios ($\delta^{18}\text{O}$) of calcite from a Cave of the Mounds stalagmite that grew during the last glacial period. Their interpretation focused on a number of rapid

changes in the $\delta^{18}\text{O}$ signal during a time period when warm periods are recorded in polar ice cores and stronger monsoons are recorded in tropical stalagmites (**Figure 11**).

A combination of microscopic imaging and analysis showed that the $\delta^{18}\text{O}$ changes each happened in ~ 10 years, and comparison to a climate model demonstrated that the $\delta^{18}\text{O}$ changes likely happened as a result of $>10^\circ\text{C}$ warming above the cave. These results speak to how quickly and dramatically those polar

warming events were propagated across the Laurentide Ice Sheet, which is important for better understanding the dynamics of rapid climate change.

As you enjoy the tour of Crystal Cave, consider what geologic stories might be captured in its

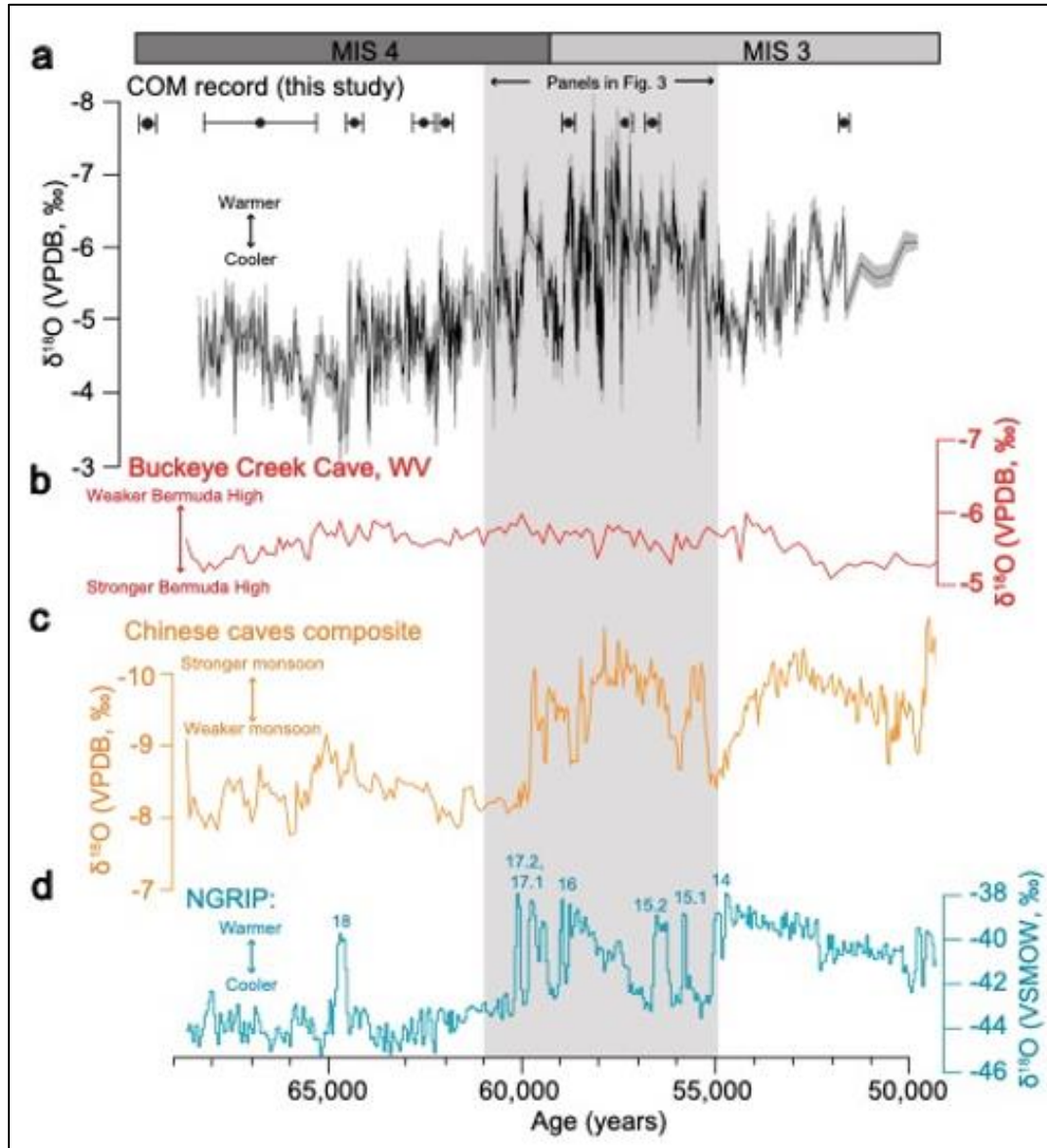


Figure 11 (*from Batchelor et al., 2023). Stalagmite CM-5 $\delta^{18}\text{O}$ record in comparison to other regional $\delta^{18}\text{O}$ records of the last-glacial period. a, Cave of the Mounds (COM; this study) $\delta^{18}\text{O}$ record (black line), with associated U-Th ages (black dots/2SD error). Note the error of our age model ranged from 520 to 2800 years and was on average 730 years. b, A stalagmite $\delta^{18}\text{O}$ record from Buckeye Creek Cave, WV (red line) showing relatively low-magnitude $\delta^{18}\text{O}$ changes during the last glacial period. c, A compilation of Chinese speleothem $\delta^{18}\text{O}$ records (orange line), showing high-magnitude $\delta^{18}\text{O}$ changes, which reflects the sensitivity of the East Asian monsoon system to high-latitude warmings (DO events) during the last glacial period. *Note the scale of the y-axis in panels A-C are the same to allow for one-to-one comparison. d, The North Greenland Ice Sheet Project (NGRIP) $\delta^{18}\text{O}$ record (blue line), showing the timing of abrupt warming DO Events (labeled #s).

speleothems. If you have ideas or questions, feel free to reach out to Ian Orland (orland@wisc.edu)!

Stop 2: Prairie Du Chien Group- Kraemer Quarry Entrance Outcrop- 850th Ave between Lincoln Rd and 870th Ave intersections.

UTM location for stop (564719.58E, 4966004.74N)



We do not have permission to enter the quarry- Do not enter the quarry. There is an outcrop of the Prairie Du Chien Group just outside of the quarry gate that continues down the hill from the quarry entrance. This outcrop appears to be a very sandy portion of the Prairie Du Chien Group, possibly representing the lower most Stockton Hill Member of the Oneota Formation, or an interfingering of the Jordan Sandstone within the basal Prairie Du Chien Group.

Just to the right of the quarry entrance are massive, 1-2m thick beds (**Figure 12**). To the left, and down the hill, the massive beds continue and just below them thinly bedded, lighter color units begin to appear (**Figure 13**). The portion of the outcrop that continues down the hill also contains what may be the Prairie Du Chien Gp./Jordan Fm. contact in the ditch just below the road grade between the outcrop and the road (**Figure 14**). While not recognized in the formal bedrock stratigraphic column for Wisconsin, the thinly bedded, lighter color units may also represent the Coon Valley Member of the Oneota Formation, often recognized and mapped in Minnesota (Steenberg, J.R., and Retzler, A.J., 2016). We will depart on 850th Ave. by continuing down the slope. To the left near the toe of the slope, there is a valley floor with a barn and small pasture.



Figure 12. Massive beds, of sandy, carbonate cemented Prairie Du Chien Group.



Figure 13. Massive beds of Prairie Du Chien Group- Stockton Hill Member? Possibly atop thinly bedded interfingerings of Jordan sandstone.

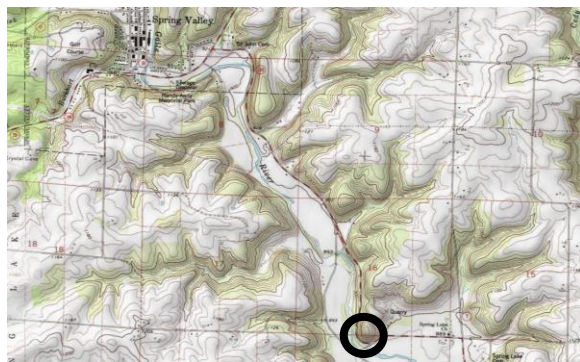
Looking across the valley floor, there is an outcrop of Jordan sandstone just across the creek (**Figure 15**).



Figure 14. Arrow pointing to possible Prairie Du Chien Gp/Jordan Fm. contact in ditch just below road grade.

Stop 3: Jordan Formation- Cth B and 770th (Spring Lake, Wisconsin)

UTM location for stop (562605.40E, 4962594.39N)



Lithofacies of the Jordan Formation are described in Runkel, 1994 and are as follows: 1) very fine-grained hummocky cross-stratified and burrowed sandstone, 2) fine-grained, trough cross-stratified and burrowed sandstone, 3) medium- to coarse-grained, large-scale cross-stratified sandstone and 4) thinly interbedded sandstone, mudstone and shale. They note that lithofacies 4 may only be relevant to certain areas in Minnesota (**Figure 18**). Authors are open to discussion as to where this particular outcrop falls in Runkel's 1994 classification schema (**Figure**

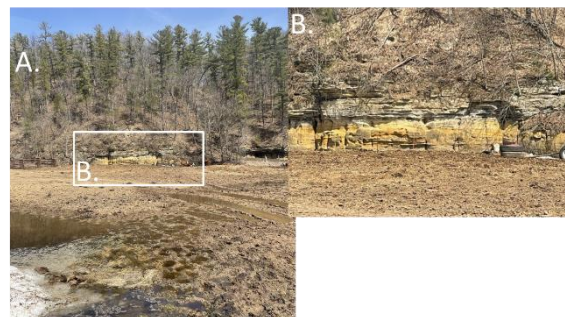


Figure 15. View of valley floor, at toe of slope driving down 850th Ave., looking across pasture towards Jordan outcrop just across creek.

18). This outcrop (**Figure 17**) is likely close to the base of the

Jordan/St. Lawrence contact (labeled in **Figure 16**), and the floodplain that the Eau Galle River runs through most likely represents the top of the St. Lawrence Formation.

The Jordan Formation of the Trempealeau Group has been highly studied in both Wisconsin and Minnesota (Mudrey, M.G. Jr. ed, 1997 and references therein). The distinction between and regional application of the quartzose and feldspathic sandstones in this formation have also been debated (Runkel 1994 and Byers and Dott, 1995). Overall, the Jordan Formation represents a coarsening upward sequence that is conformable



Figure 16. View of the outcrop across Cth B with approximate contacts labeled.



Figure 17. A. View of outcrop. Note the cross bedding and iron staining above hammer. B. Possible iron concretions?

at its basal contact with the St. Lawrence, and unconformable with the Prairie Du Chien Group contact at its top. The Coon Valley member of the Jordan is not formally recognized in the stratigraphic column of Wisconsin (WGNHS

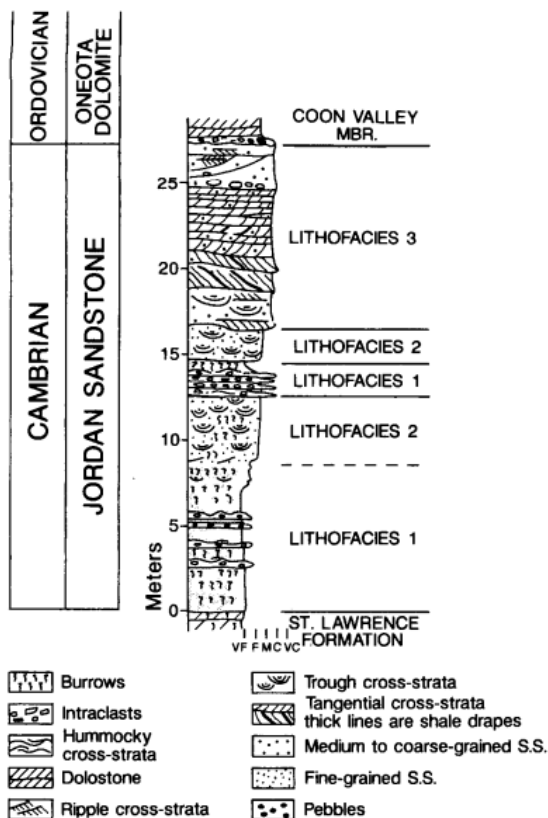


Figure 18. Figure 2 from Runkel, 1994 illustrating the different lithofacies of the Jordan Sandstone in Minnesota.

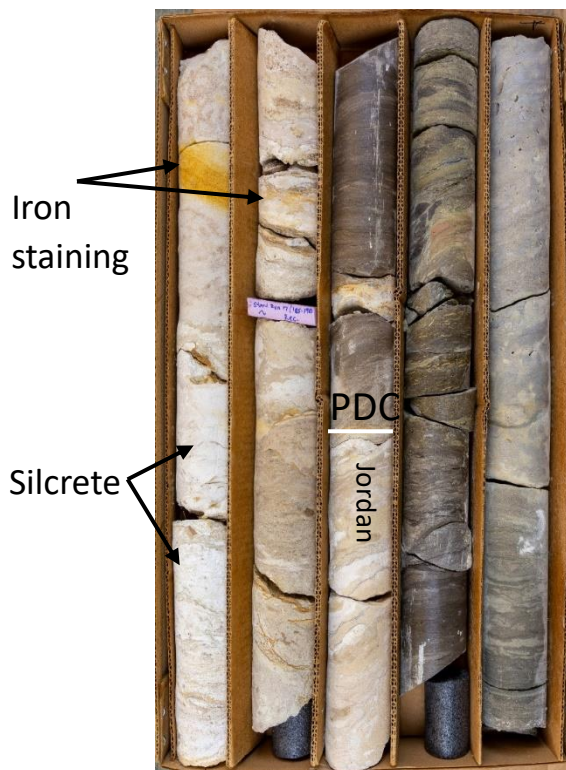


Figure 19. Photo of core from Jefferson County, Wisconsin (east of Wisconsin Arch) modified from Kusick, 2022 M.S. Thesis. This interval of core shows the contact between the Jordan Formation and the Prairie Du Chien Group.

2011), though it has been noted above the Jordan Fm. in southern parts of the state.

A recent M.S. dissertation (Kusick, 2022) discussed, in detail, both the stratigraphy and depositional environments of the Cambrian-Ordovician units east of the Wisconsin Arch. Kusick (2022) describes the Jordan sandstone as being comprised of only 2 facies of cross stratified sandstone and shaly sandstone, and as being deposited in an upper to lower shoreface environment. These authors would also like to note that locally, the Jordan Formation east of the Wisconsin Arch includes silcrete and clay, and hosts disseminated sulfides (**Figure 19**).

Stop 3a: Rock Elm Impact Structure- Rock Elm, WI- lunch at Nugget Lake County Park

UTM location for stop (561573.58E, 4948450.76N)



Figure 20. Core photos showing jumbled and deformed Cambrian strata from the southern edge of the Rock Elm central uplift - from WGNHS archives.

No set stop, informational only as we'll be eating lunch at a park within the crater.

The field trip will go through the Rock Elm impact structure (**Figure 20**), located in Pierce County around 35 miles WSW of Eau Claire (**Figure 21**). The Rock Elm impact structure is the largest deformation event recorded in the Paleozoic section of western Wisconsin. The structure contains a 6.5 km diameter ring boundary fault and a central uplift 1 km across (Cordua, 1985). Where control exists, the ring boundary fault is thought to have accommodated 45 meters of down-in-the-center displacement (French and others, 2004). Much of the interior of the ring boundary fault is filled with the relatively flat-lying Rock Elm shale and the overlying Washington Road sandstone, which have a combined thickness of approximately 48 meters. These units are unique to the area, and do not exist outside of the ring fault. These units are described based on numerous outcrops, many given in Cordua (1987) and Cunningham and others (2011). The central uplift contains outcrops of tilted Mt. Simon Formation (**Figure 20**), which suggests 250 to 300 meters of uplift within the core zone relative to rocks outside the impact structure (French et al., 2004). Reidite, a high pressure polymorph of zircon, has been

detected in detrital zircon grains and is interpreted to be caused by the impact (Cavosie et al., 2015).

Stop 4: *Skolithos* burrows in Tunnel City Group-330th Ave. between HWY 25 and Cth Y (Private Property!!!)

UTM location for stop (586184.90E, 4961335.76N)



We will park on a private drive and walk east along the road to this outcrop.

This stop in the Tunnel City Group is an excellent example of *Skolithos* burrows (**Figure 22**) which are common in the Tunnel City Group. This outcrop is likely the Tomah Member of the Lone Rock Formation. Excellent examples of cross-stratification can be seen at this outcrop as well.

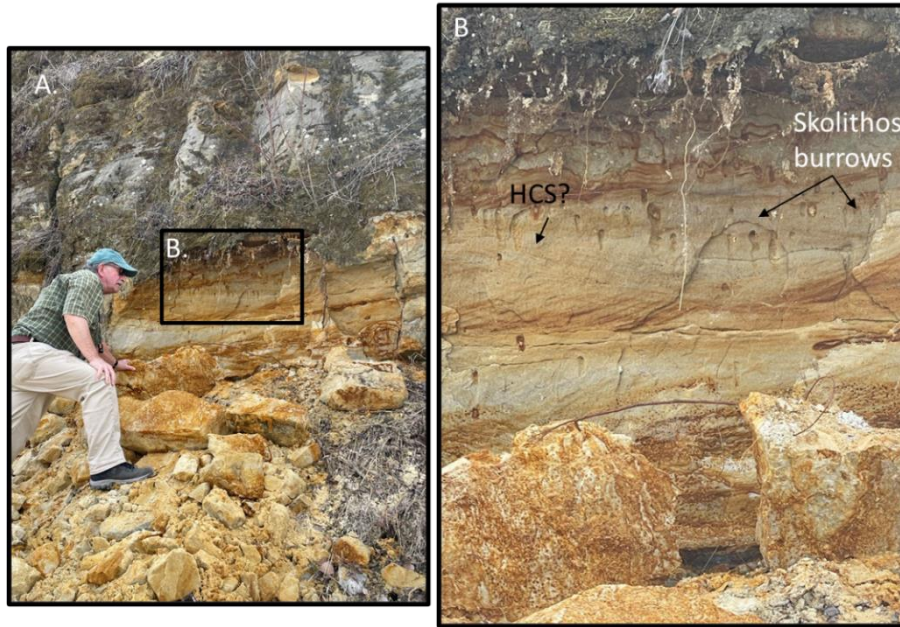


Figure 22. Tunnel City Group outcrop with excellent examples of *Skolithos* burrows and possibly multiple types of cross-stratification.

Stop 5: St. Lawrence Formation/Tunnel City Group road cut- Cth C and Cth Y

UTM location for stop (587368.83E, 4958918.89N)



The road cut is just west of the intersection of Cth C and Cth Y. We will park and walk to this outcrop. Cth C is a fairly busy road, please exercise caution when you decide to cross.

Recent mapping in Trempealeau County, southeast of stops 5 and 6, has produced interesting work on both the geological relationships of the Cambrian- Ordovician rocks, and the quality of groundwater in the west-central part of the state. Zambito and others, 2018 published the following unit description for the Tunnel City Group:

“Tunnel City Group (Cambrian)

The Tunnel City Group is comprised of the Lone Rock and Mazomanie Formations. Similar to neighboring La Crosse County (Evans, 2003), the Mazomanie Formation was not recognized in Trempealeau County.

Lone Rock Formation. The Lone Rock Formation (**Figure 23**) and its members are identifiable in the map area. The members, from oldest to youngest, are Birkmose, Tomah, and Reno; these are not differentiated at the map scale. The Birkmose Member is a dolomite-cemented, coarse-grained, glauconitic sandstone to sandy dolostone with flat-pebble conglomerates; the Tomah Member is a tan to white-colored, medium-grained, glauconitic quartz sandstone; and the Reno Member is a glauconitic medium- to coarse-grained quartz sandstone with flat-pebble conglomerates. *Palaeophycus* and *Skolithos* are common, as is hummocky cross-stratification and cross-stratification bounded by horizontal bedding surfaces. The contact with the overlying St. Lawrence Formation is sharp and unconformable.

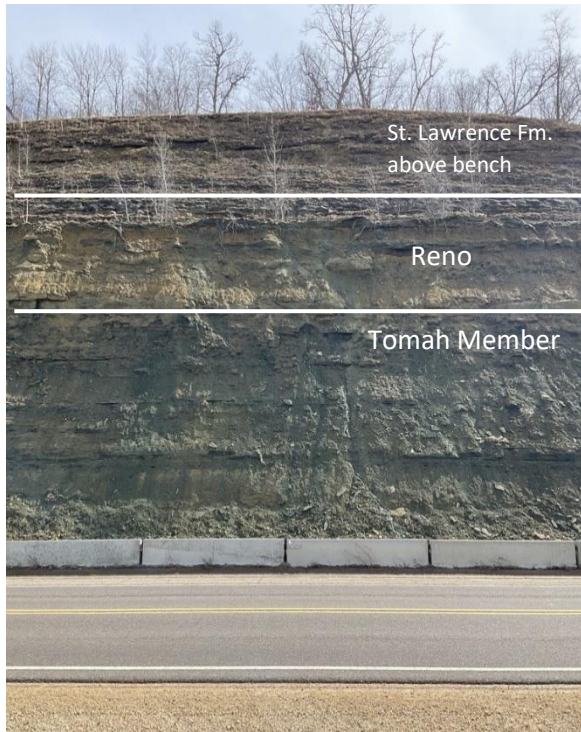


Figure 23. Road cut showing contacts between the St. Lawrence Fm., and Reno and Tomah Mbrs. of the Lone Rock Fm. This is the view from the north side of Cth C.

The Lone Rock Formation is commonly exposed in shale pits, along roads leading to ridgetops, and at the top of sand mine high walls where the Wonewoc Formation is extracted and the Birkmose Member forms the caprock. The formation is approximately 150 feet thick in the map area. Elemental data for part of the Lone Rock Formation is shown in plate 2 [figure 7]. These data show the formation's lithologic variability, in particular the distinct upper carbonate-cemented and lower sandstone dominated intervals in the Birkmose; the lower interval consists of reworked quartz grains from the underlying Wonewoc with interspersed, rare glauconite grains and phosphatic brachiopods. These authors find this to be an excellent, and representative description of the unit for the west-central region. – Zambito and others (2018)

Figure 24 shows a small part of the westernmost portion of the outcrop. The very dark, greenish-black bed just below the more resistant dolomitic bed is rich in the mineral glauconite, which is an iron potassium

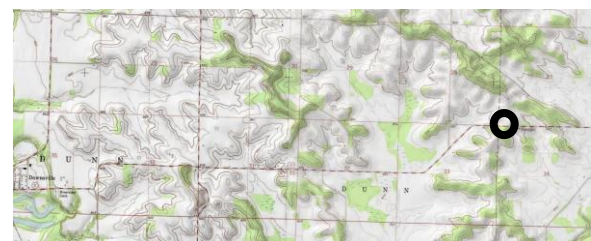
phyllosilicate mineral. Another interesting part of this road cut is the bench approximately 35ft up. This bench feature likely represents the unconformable contact Zambito and others, 2018 alluded to with the overlying St. Lawrence Formation. The Mazomanie Formation is generally not observed in this part of the state and is more prevalent in the southern parts of the state where it interfingers the Lone Rock Formation (Mudrey, M.G. Jr. ed, 1997 and references therein).

A 2019 study by Zambito and others investigated the relationship between groundwater quality and the geochemistry of the Tunnel City-Wonewoc units in western Wisconsin. This study notes that sulfide bearing minerals are disseminated between the two units in west-central Wisconsin, and they call for more work to better understand the geochemical effects of oxidation of sulfide minerals during groundwater pumping in this part of the state. Our next stop will be at an outcrop of Wonewoc sandstone, and we will pass other outcrops of this unit on our drive.

Overall, the Tunnel City Group both east and west of the Wisconsin Arch are quite similar. As examined at this stop, west of the Arch, the Tunnel City Gp. East of the Arch is also a quartz sandstone with glauconite and trace amounts of shale.

Stop 6: Wonewoc road cut- Cth Y

UTM location for stop (593272.27E, 4959793.83N)



The Wonewoc Formation is a fine to coarse sandstone unit with medium to thick beds, high-angle trough cross-stratification and some

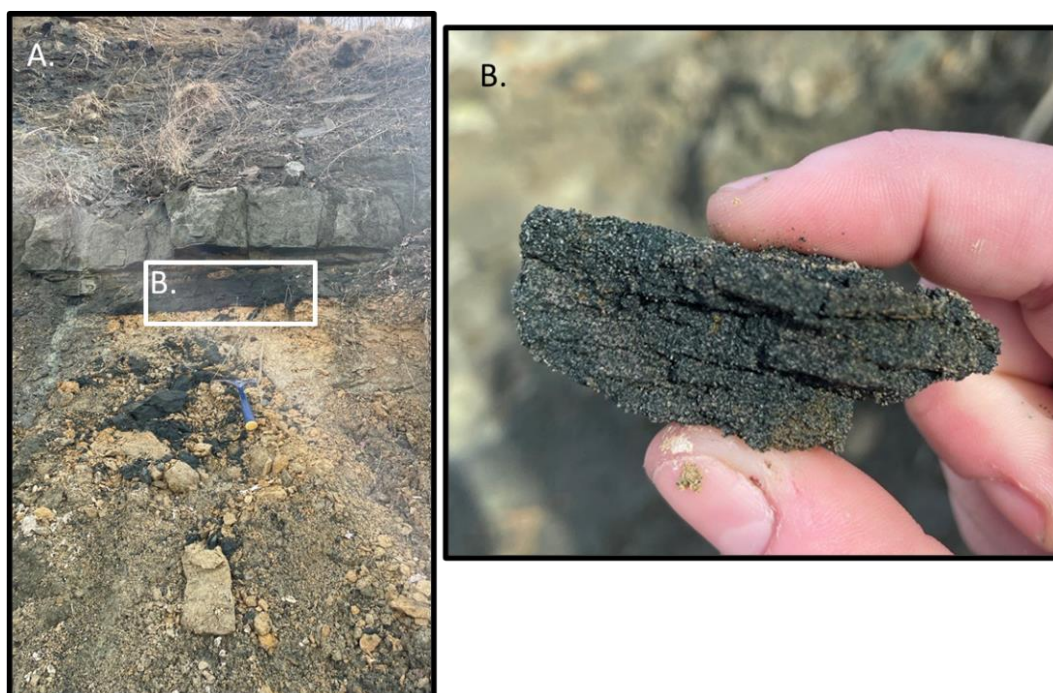


Figure 24. A. View of westernmost point of the outcrop. B. close up of the phosphatic rich, friable sandstone of the Tomah Member.

feldspar (Mudrey, M.G. Jr. ed, 1997). Brachiopod fragments, *Skolithos* burrows (which we observed at stop 4), and *Climactichnites* are somewhat common in the fossiliferous Ironton Member of this formation. The upward contact (**Figure 25**) with the Tunnel City Group is gradational and fines upward; the basal contact with the Eau Claire Formation has been debated as to whether it is gradational or not (Mudrey, M.G. Jr. ed, 1997 and Ostrom 1978). Note the

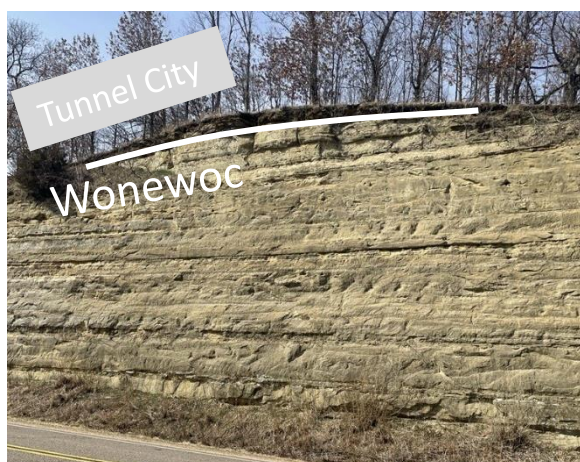


Figure 25. View of the Wonewoc outcrop on Cth Y.

orange, iron rich bed on the south side of Cth Y that doesn't seem to appear in the north face of the outcrop.

In the 1990s, to better characterize aquifer and confining units, the Minnesota Geological Survey began focusing on the hydrostratigraphic characteristics of geologic units (1998 ILSG field guide). Hydrostratigraphic subdivisions include: 1) fine clastic; 2) coarse clastic; 3) carbonate; 4) clastic/carbonate mix. While this approach has not been implemented as part of bedrock mapping in Wisconsin, its importance has been recognized by Wisconsin hydrogeologists in lithologically complex units such as the Eau Claire Formation (Bradbury and Runkel, 2011). The authors are open to questions, and discussion of this method as it may pertain to future groundwater study needs across the Midwest.

Additional stops if time permits:

Devil's Punchbowl – Eau Claire Formation:

UTM location for stop (582783.00N, 4966909.29E)

Park in the parking lot, walk east towards the stairs, and take them down the trail into the Punchbowl.

This is a classic stop for field trips in this part of the state. Devil's Punchbowl is managed by the Landmark Conservancy- **please do not use rock hammers on the outcrops and be good stewards of the landscape**. This outcrop shows the relationship between the Eau Claire Formation and the Wonewoc Formation. Expect to see a fine-grained sandstone with swaley cross beds in the Eau Claire Formation, and medium- to coarse-grained, cross-stratified sandstone in the Wonewoc Formation at this location (Mudrey, M.G. Jr. ed, 1997).

Hwy 37/Hendricks Ave and Silver Springs Dr.- Mt. Simon Formation:

UTM location for stop (615774.61N, 4958712.01E)

This is a typical Mt. Simon Formation exposure (**Figure 26**), coarse- to medium-grained, cross-bedded, iron stained, sandstone, interbedded with shale and fine grained sandstone (Mudrey, M.G. Jr. ed, 1997).



Figure 26. View of the Mt. Simon Formation outcrop. This location is heavily iron stained and exhibits excellent examples of sedimentary structures like trough cross stratification and channel forms.

Acknowledgements

A special thanks to Eric Stewart and Ian Orland for contributing content to this guide. A very special thanks to Bill Batten for helping scout field locations and always knowing where to find the best contacts. Additionally, this guide would not have been possible with consulting Dave LePain's mapping notes and field guides from Pierce and St. Croix counties and the work of others who have previously published work on these Paleozoic units.

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Field Trip 3 – Precambrian Geology of the Eau Claire River Valley: Re-discovering the Eau Claire Volcanic Complex

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Introduction

The erosional outliers of Precambrian bedrock in the Eau Claire River valley represent the southernmost extent of the Canadian Shield before it is completely covered by Paleozoic sedimentary strata. The rocks exposed here are part of the Paleoproterozoic Penokean Orogeny, a collisional orogen that resulted from the accretion of the Pembine-Wausau and Marshfield terranes onto the southern margin of the Superior Province. This region was last visited by members of the Institute of Lake Superior Geology in 1980 when a field trip through the region was conducted by Paul Myers and colleagues (Myers et al., 1980) when it was called the “Chippewa Amphibolite Complex”. Since then, the “Eau Claire River Complex” was defined and described in detail by Cummings (1984). There has been ‘new’ U/Pb data collected by the USGS (Sims et al. 1989) and others (Van Wyck et al, 1997; Klier, 2019; Weber and Lodge, 2022), regional syntheses of the Penokean volcanogenic massive sulfide (VMS) mineralization (DeMatties 1989; 1994; 2018; 2022), maps published by government surveys (Brown, 1988), and orogen-wide tectonic model (Shultz and Cannon, 2007) that is being revisited based on new U/Pb data (Zi et al., 2021). The rocks that will be visited on this trip are a critical part of evaluating the tectonic models for the Penokean Orogen and have not been examined using modern analytical techniques.

The Penokean Orogen is perhaps best known for hosting numerous VMS deposits. In fact, one of the most complete descriptions of several deposits was published by the Institute of Lake Superior Geology (LeBarge, 1996). The passing

of the “prove-it-first” law, or sulfide mining moratorium, in 1997 effectively shut down mineral exploration and mining activities in the region. More recently, the mineral exploration industry has been reinvigorated because of the 2002 discovery of the 18.2 Mt Back Forty deposit in Michigan, easing of the sulfide mining moratorium in 2017, and a recent national push for securing domestic critical mineral resources. However, this has also highlighted the lack of modern datasets on Wisconsin’s mineral deposits that could be used to further our knowledge of the mineral-forming systems in the belt. The Pembine-Wausau Terrane has received most of the historic and recent attention since it hosts approximately 150 million tonnes of known VMS mineralization. However, little attention has been given to the Penokean volcanic deposits that overprinted the Marshfield Terrane that are presented in this guidebook. These volcanic deposits host a VMS prospect (Butler Prospect) and therefore the geodynamic setting of these volcanic rocks clearly are favorable for submarine hydrothermal activity. DeMatties (2022) recognized the gap in knowledge for these Penokean volcanic deposits, known as the Eau Claire Volcanic Complex, within the Marshfield Terrane and their exploration potential. It is a little embarrassing how little we know about the Eau Claire region considering the mineral wealth of the rest of the orogen. Current research at the University of Wisconsin-Eau Claire is aimed at the addressing this issue.

The portion of the Eau Claire Volcanic Complex that is visited in this guidebook is not well exposed and its regional context is poorly constrained. Students from the University of

Wisconsin-Eau Claire have been visiting Big Falls and Little Falls locations in this guidebook for decades to learn how to map and describe rocks in the field, measure structures and interpret geologic histories, and learn the basic mechanics of field work. Faculty, students, and alumni from Eau Claire consider these outcrops classic. This guidebook will (re-)introduce these rocks and present some of the ongoing research with the Eau Claire Volcanic Complex. The outcrops visited in this guidebook are accessible by foot, but many others were accessed by kayaking in the Eau Claire River. Ongoing research in this region hopes to expand the lithogeochemistry and zircon

petrochronology database to better delineate the geodynamic evolution and crustal architecture of this region. Determining the presence or absence of Archean basement throughout the Marshfield terrane will help refine terrane boundaries and improve our understanding of the metallogeny of the region to assist in future mineral exploration efforts.

Regional Geology

The Paleoproterozoic Penokean Orogen (ca. 1.8 Ga) in the Lake Superior region (**Figure 1**) is a classic Precambrian orogenic belt comprised of

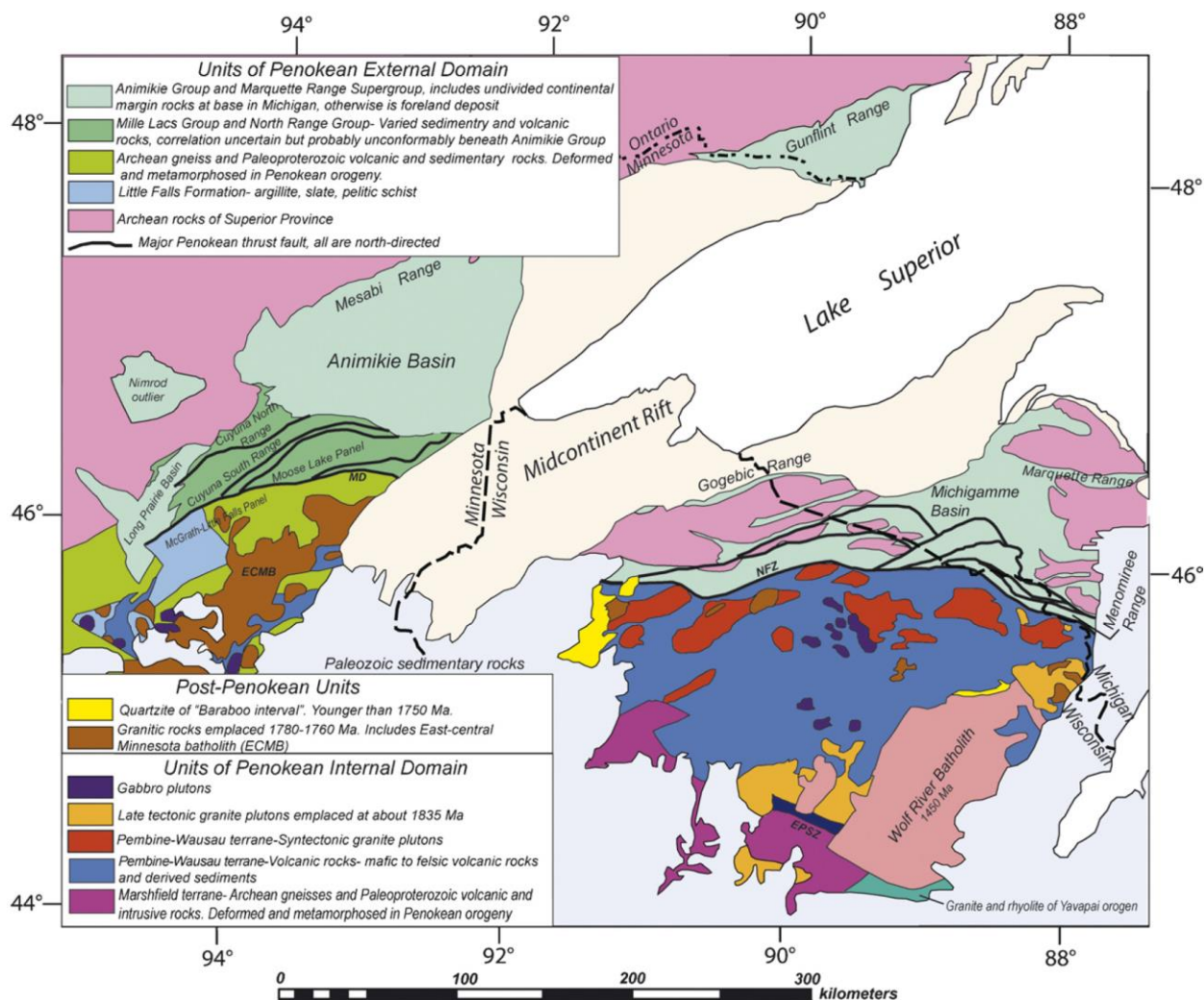


Figure 1: Geologic map of the major tectonic assemblages and major structures of the Penokean Orogen. Notable abbreviations that are important for this guidebook are EPSZ, Eau Pleine shear zone; NFZ, Niagara fault zone. Figure from Shultz & Cannon (2007).

dominantly submarine volcanic rocks formed in a suprasubduction zone setting that are now structurally juxtaposed along the southern edge of the Archean Superior Province during the earliest phases of forming the Columbia, or Nuna, supercontinent (LaBerge and Myers, 1984; Sims et al., 1989; Schulz and Cannon, 2007). The orogen is host to at least 150 million metric tonnes (Mt) of VMS and associated mineralization (DeMatties, 1994, 2018) but remains one of the more poorly understood and underexplored mineral districts in North America.

The Penokean Orogen has been divided into the Interior and Exterior domains. These domains are sutured by the Niagara Fault Zone (**Figure 1**). The Exterior domain consists of passive margin, rift, and forearc basin sediments and Archean crustal blocks from the Superior Province that were deformed in the folded and faulted foreland part of the orogen.

The Interior Domain consists of two accreted terranes, the Pembine-Wausau and Marshfield terranes. These terranes are sutured by the Eau Pleine Shear Zone (**Figure 1**). The Pembine-Wausau Terrane is a composite accreted oceanic arc overprinted by continental-margin magmatism and hosts numerous VMS deposits and occurrences (DeMatties, 1994; Shultz & Cannon, 2007) (**Figure 2**). The Marshfield Terrane is composed of Archean crustal fragments of unknown origin that was overprinted by Penokean-aged magmas during the Penokean orogen (**Figure 2**) and is described in more detail in the sections to follow.

Shultz and Cannon (2007) synthesized the tectonic events that formed the Penokean Orogen (summarized in **Figure 2**) based on a detailed compilation of lithologic, structural, sedimentological, and geochronological datasets. This classic model proposed that an oceanic arc, now the Pembine-Wausau Terrane, collided with the southern margin of the Superior Province around 1880 Ma. Following a subduction flip from south-directed to north-directed subduction,

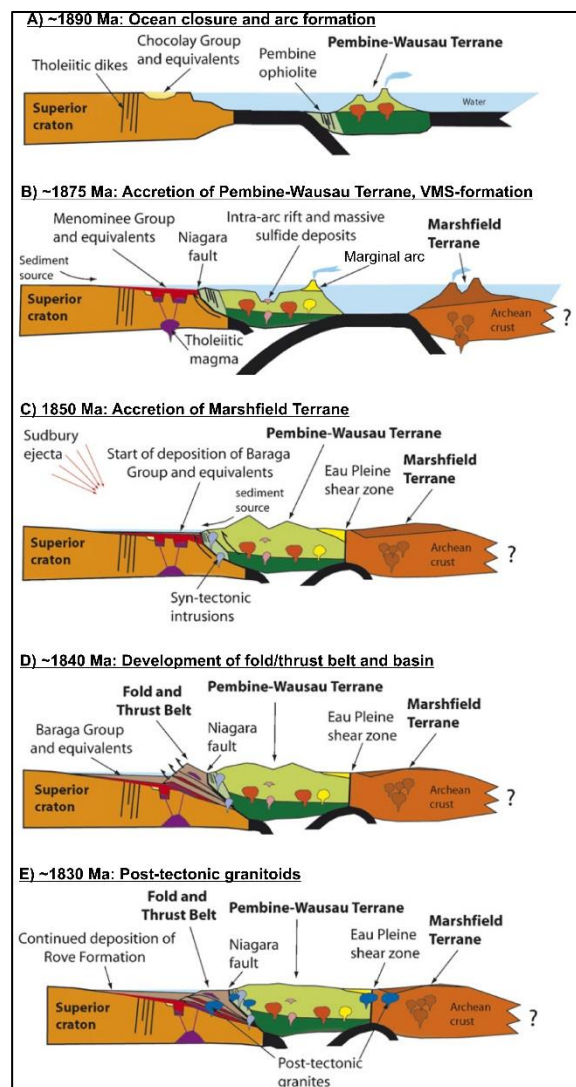


Figure 2 - Schematic tectonic evolution of the Penokean Orogen provided by Shultz and Cannon (2007) based on geophysical, sedimentological, and geochronological compilations.

continental arc volcanism and back arc extension developed until about 1850 Ma until the collision with the Marshfield terrane began. During this ocean closure, a double subduction zone with concurrent northward and southward subduction resulted in arc magmatism on both the Pembine-Wausau and Marshfield terranes. Sedimentation related to this convergence in a foreland basin setting continued until about 1835 Ma. The end of the orogen was constrained by undeformed post-tectonic plutons dated at 1830 Ma that stitch shear zones.

However, this classic tectonic model for the evolution of the Penokean Orogen has recently been re-evaluated considering new U/Pb data. The first contradictory data came when Quigley (2016) obtained a high-precision U/Pb zircon age of 1832.98 ± 0.52 Ma from a rhyolite at the Back Forty deposit via CA-ID-TIMS. This younger age was in stark contrast to the other VMS deposits that yielded U/Pb zircon ages of ca. 1875 Ma. Additional U/Pb zircon ages reported by Zi et al. (2021) from volcanic units (Beecher Formation) and plutonic rocks (Dunbar Gneiss, Newingham Tonalite) in the eastern part of the orogen supported the younger extensional tectonic event proposed by Quigley (2016). These new ages resulted in a revised Penokean tectonic model where long-lived northward subduction along a continental margin with repeated extensional and contractional regimes in response to retreat and advance of the subducting oceanic plate (**Figure 3**). Weber and Lodge (2022) obtained a U/Pb age of 1831.4 ± 2.0 Ma on the dacite unit hosting the Eisenbrey deposit in the western part of the orogen, suggesting that this second VMS forming event was widespread. A summary of the geochronology is presented in **Figure 4**.

Marshfield Terrane

This guidebook visits the only Penokean volcanic complex south of the Eau Pleine Shear Zone and is interpreted to part of the Marshfield Terrane. The Marshfield Terrane represents an Archean microcontinent of uncertain origins (Sims et al., 1989; Schulz and Cannon, 2007; Zi et al., 2021). Some of the earliest works on the terrane by Sims et al. (1989) noted eight Archean U/Pb ages from isolated outcrops along the Wisconsin, Black, and Chippewa Rivers; many of which were compiled from unpublished sources. Paleoproterozoic volcanic rocks in the Marshfield terrane were deposited about 1835-1865 Ma (Sims et al., 1989; Van Wyck, 1995; Klier, 2019; Weber and Lodge, 2022). These supracrustal rocks were referred to as the Eau Claire River Complex by Cummings (1984) or the Eau Claire Volcanic Complex by DeMatties (2018; 2022).

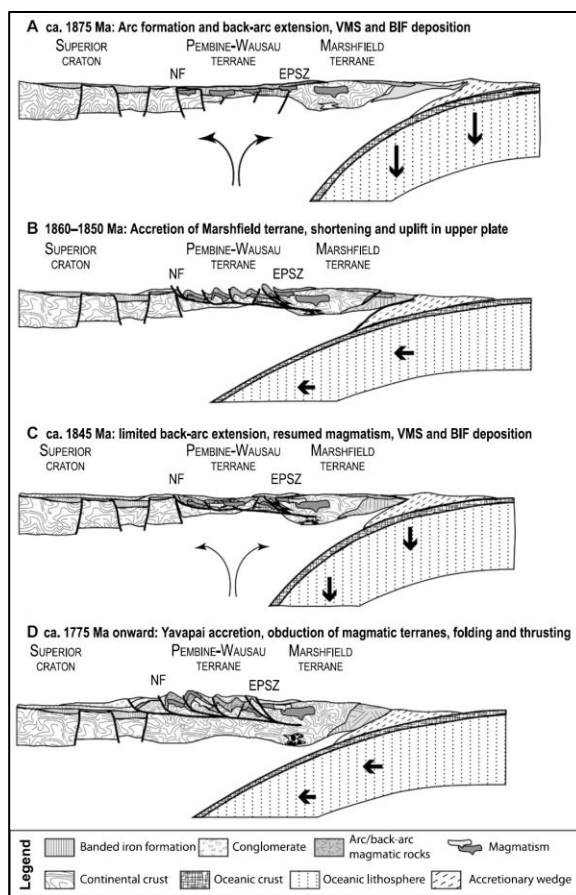


Figure 3 - Schematic illustration of the revised tectonic model of the Penokean Orogen. Figure is from Zi et al. (2021). Abbreviations: NF—Niagara fault zone; EPSZ—Eau Pleine shear zone.

They consist of an interlayered sequence of felsic to mafic volcanic rocks, dacite porphyry, and a variety of clastic and chemical sedimentary rocks (Sims et al., 1989). Some conglomerates contain granitic gneissic clasts that were interpreted to be Archean (Myers et al. 1980), but no definitive ages were determined on the clasts. Otherwise, our knowledge of the Archean Marshfield terrane and associated Paleoproterozoic volcanic rocks remains as sparse as the outcrop exposures.

Eau Claire Volcanic Complex

The Eau Claire Volcanic Complex is poorly documented and understudied mainly due to its inaccessible outcrops in remote parts of the Eau Claire River valley. Myers et al. (1980) described

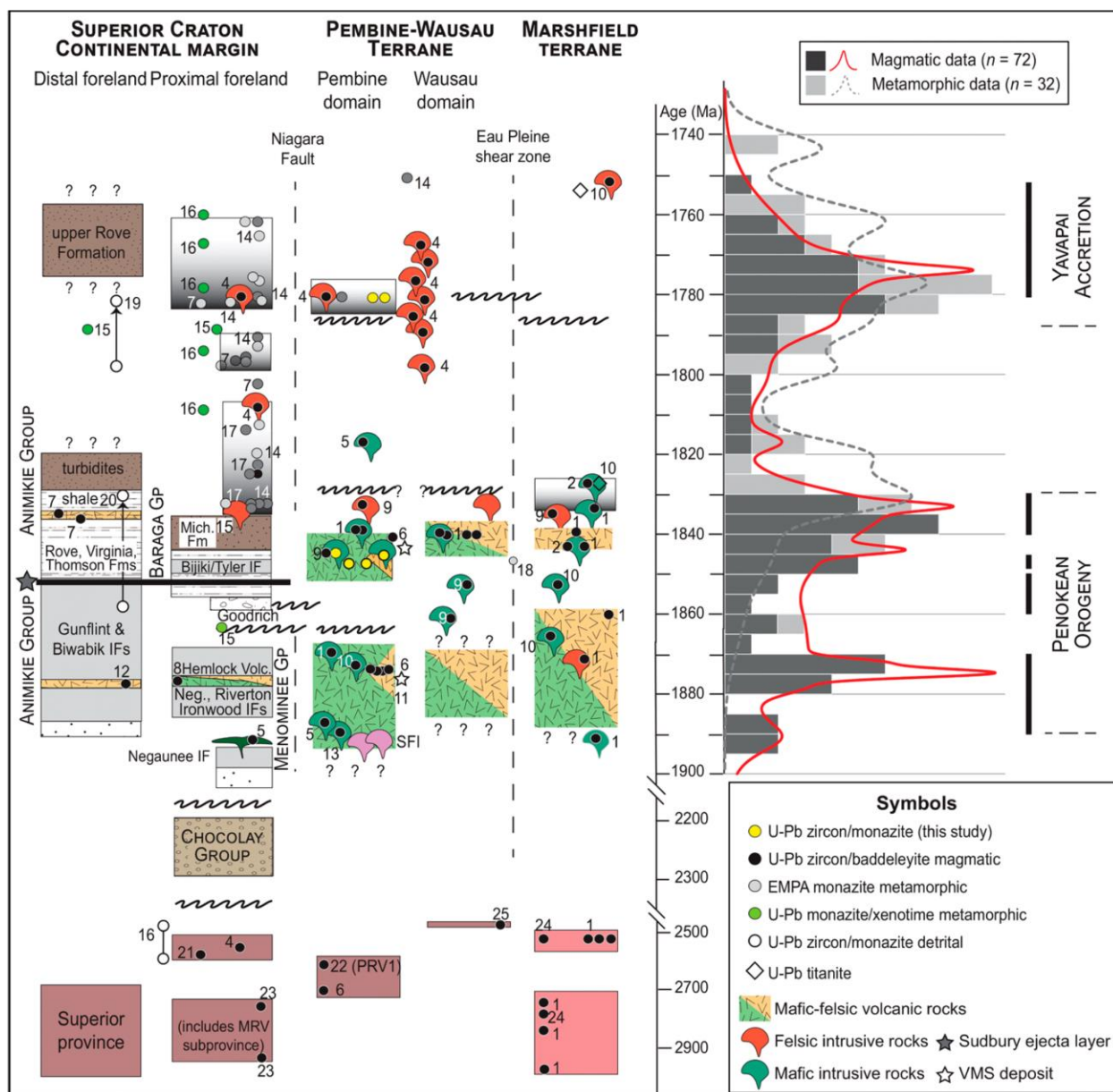


Figure 4 - Time-space plot for the tectonic components of the Penokean Orogen. Plot is from Zi et al. (2021). See citation for references on data sources.

supracrustal amphibolites, metarhyolites and metasediments in the Eau Claire River valley and classified them as part of the Chippewa Amphibolite Complex. This informal classification of the high metamorphic grade rocks in the Eau Claire-Chippewa River area was eventually grouped with the Marshfield Terrane by Sims et al. (1989) and Shultz and Cannon (2007). The first time that the Eau Claire “Complex” was officially referred to was by Cummings (1984) when discussing the petrology

and geochemistry of the gneisses in the Big Falls-Little Falls area (Stops 1 and 2 in this guidebook). After that, research in the Eau Claire Volcanic Complex essentially ceased. Sims et al. (1989) reported a U/Pb rutile age from Big Falls of ca. 1835 Ma. In fact, the words “Eau Claire” are not used in the Shultz and Cannon (2007) regional synthesis. DeMatties (2018; 2022) refers to the Eau Claire Complex when discussing the volcanic complexes in the Penokean, but largely cites the work of Myers et al. (1980).

Preliminary Hf-isotope data and zircon trace elements reveal that the rocks analyzed in the Eau Claire Volcanic Complex are juvenile, mantle-derived melts with no inheritance from older sources (See Weber et al. 2023, Part 1 of this volume). This suggests that these volcanic rocks are not forming on Archean basement, as one would expect if the Eau Claire Volcanic Complex was emplaced onto the Marshfield Terrane. Additionally, magnetic lineaments on aeromagnetic maps for the region appear to crosscut the interpreted position of the Eau Pleine Shear Zone and the overall fabric as outlined by magnetics appears constant (**Figure 5**). Ongoing research in the region seeks to better define the relationship of the Eau Claire Volcanic Complex to the Marshfield Terrane and the architecture of the basement in this area.

Field Trip Stops

The overall objective of this guidebook is to tour the accessible parts of the Eau Claire Volcanic Complex as exposed in the Eau Claire River valley and surrounding tributaries. The guidebook can be divided into two main regions: The Big Falls-Little Falls and North Fork regions. The Big Falls-Little Falls region represents the classic “Eau Claire Complex” originally described by Cummings (1984). The North Fork region is much more remote and rarely visited by geologists. In fact, it is not obvious that anyone has studied these rocks since they were first reported by Myers et al. (1980). These more remote parts of the complex are currently being studied (see Leahy and Lodge, Part 1 of this volume) to determine their regional context. Some of that data will be presented herein. The goal of this work is to determine if the Eau Claire Volcanic Complex is a volcanic center built upon Archean crust (continental arc) or is juvenile

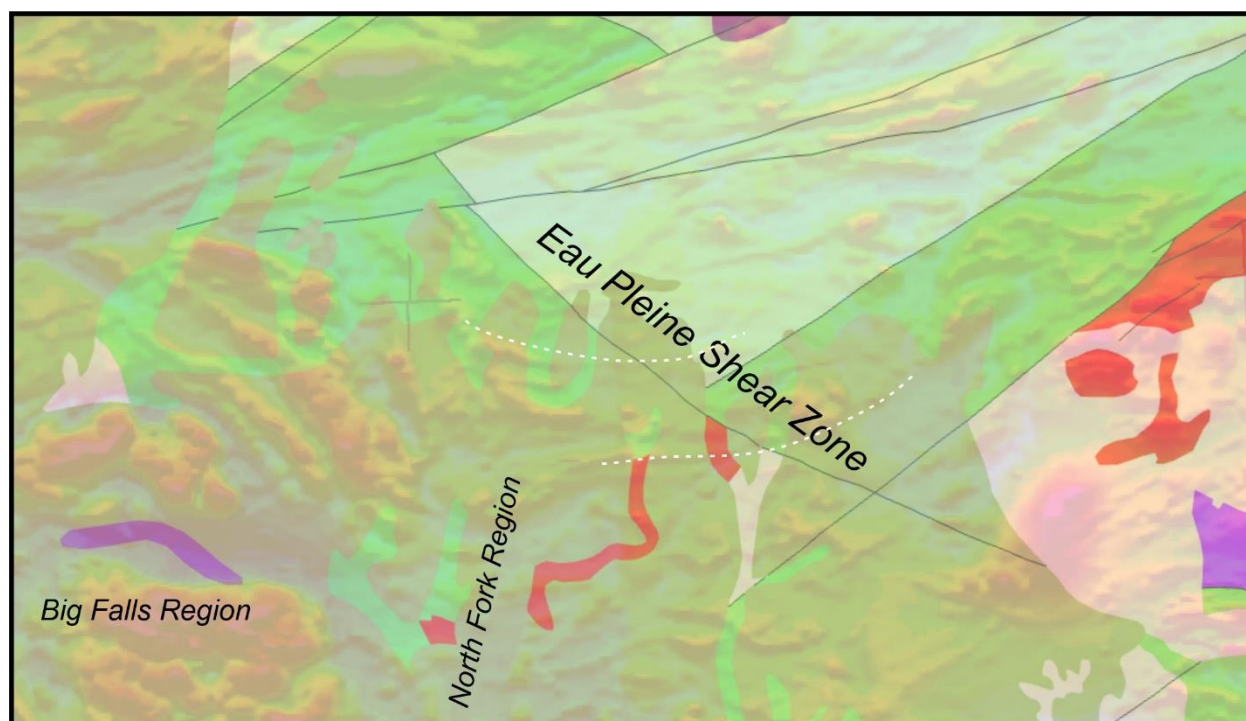


Figure 5 - Total field aeromagnetic map of the Eau Claire River region showing the location of Eau Pleine Shear Zone and field trip regions (Big Falls, North Fork). White dashed lines highlight a couple of magnetic lineaments that extend through the suturing shear zone. Magnetic maps from Daniels and Snyder (2002).

(oceanic arc) as this is important implications for the regional metallogeny and mineral systems.

Most of the locations in this guidebook are on riverside outcrops. These areas are prone to sudden flooding and the upmost caution and careful planning should be used prior to visiting these locations. In addition, rocks here are uneven and potentially slippery. To access larger sections of outcrops, low water conditions may be required. In addition, all locations in this region may contain poisonous plants (e.g. nettle, poison ivy) and black-legged ticks that can transmit diseases. While this is unlikely to be a concern in early spring during the 2023 ILSG conference, future users of this manual should plan appropriately.

Big Falls Region

The banded amphibolite, gneisses, and intrusions in the Big Falls region of the Eau Claire River are some of most studied Precambrian exposures in this region and are visited multiple times a year by introductory and upper division geology classes at the University of Wisconsin-Eau Claire. It is in this region that the term “Eau Claire River Complex” was first introduced by Cummings (1984) and this terminology has since been adopted by others (e.g. DeMatties, 2018) to describe the volcanic rocks present in the Marshfield Terrane.

The region consists of mostly amphibolitic and felspathic gneisses that are intruded and brecciated by tonalite (Figure 6). Regional metamorphism in this region is at lower to upper amphibolite facies. A sample of amphibolitic gneiss from the Eau Claire Volcanic Complex using the edenite-richterite thermometry determined temperatures between 719-769 °C (Hannack and Radwany, 2018). Unpublished data from University of Wisconsin-Eau Claire class projects using garnet-biotite thermobarometry estimate peak metamorphic conditions at 765 °C and 11.5 kbars (Figure 7). A rutile U/Pb age of 1835 Ma from Sims et al. (1989) in the Eau Claire

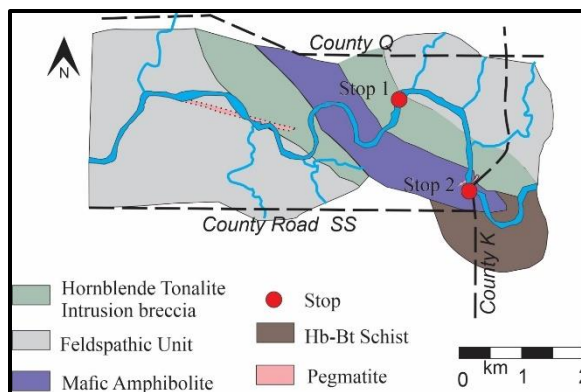


Figure 6 - Generalized Precambrian geologic map of the Big Falls-Little Falls area of the Eau Claire River. Figure modified from Cummings (1984).

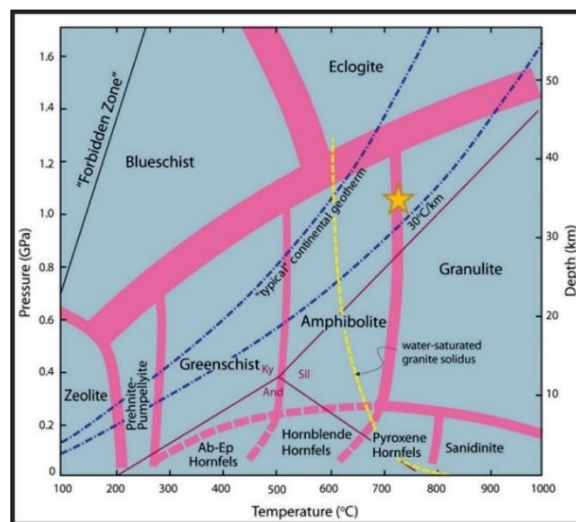


Figure 7 - Metamorphic conditions from geothermobarometric studies at Big Falls indicated by the yellow star. Data is from unpublished student project at the University of Wisconsin-Eau Claire.

Volcanic Complex may indicate the timing of metamorphism.

New research in this region provides our first glimpse into the trace element characteristics of these rocks (**Figure 8**). Rocks from both Big Falls and Little Falls have mafic protoliths with E-MORB to oceanic arc like abundances of Th, Nb, and Yb (Pearce, 2008). On normalized trace element diagrams, samples have elevated LREE, low Th/La ratios, negative Nb and Ti anomalies. These trace element characteristics features are common in back-arc environments. Additionally, felspathic units sampled have extremely

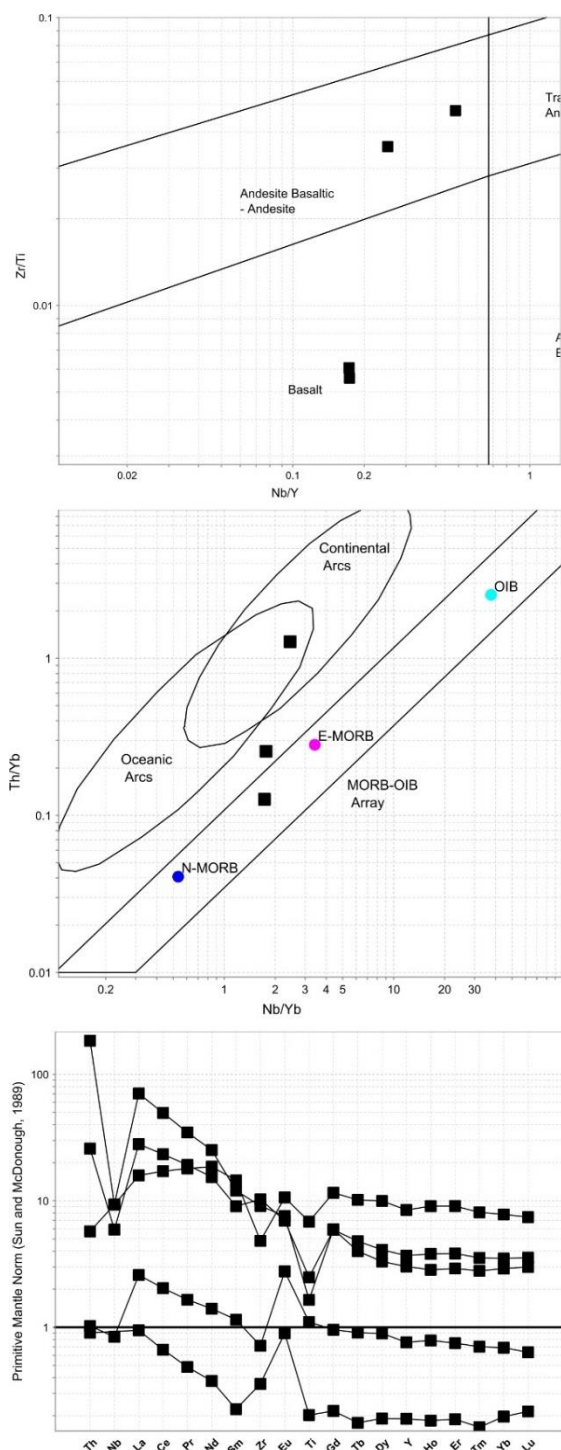


Figure 8 – Preliminary trace element geochemistry from the Big Falls-Little Falls area of the Eau Claire Volcanic Complex. Top: Trace element classification diagram from Pearce (1996) modified from Winchester and Floyd (1977). Middle: Mantle source discrimination diagram from Pearce (2008). Bottom: Primitive mantle-normalized trace element diagram using values from Sun and McDonough (1989).

depleted trace element signatures and positive Eu anomalies, suggesting that they may be fractionated crystal cumulates. This broadly supports the interpretation of Cummings (1984) that the protolith of the Big Falls gneisses are a layered mafic intrusion.

Stop 1: Amphibolite Gneisses and the “Great Unconformity” at Big Falls County Park

Lat: 44.8215° Long: -91.2953°



This location is accessible from the north entrance to Big Falls County Park off Eau Claire County Highway Q. There is a parking lot at this entrance with plenty of parking for park visitors. Follow the paved foot path eastward toward the river. Once on the riverbank, walk northward for

about 50 m to reach the outcrops at the falls. Note that outcrops on the south bank of the falls will have to be accessed via the south entrance to the park on County Highway K. During very low water conditions, it is possible to hop across the outcrops to access the south bank. This region has some steep-edged rock cliffs adjacent to the river and there are springs that keep some areas wet and slippery. Please watch your step.

This stop highlights the geology along the north side of Big Falls County Park where the rocks are exposed along the Eau Claire River. In addition to the Precambrian rocks, this location also has a great exposure of the “Great Unconformity” with overlying Cambrian Mount Simon Formation. The Eau Claire River flows along the nonconformity between the Cambrian and Precambrian rocks, where the river has eroded the overlying Cambrian units away exposing the Precambrian basement rocks. At this stop, we highlight four locations that highlight different units seen here at Big Falls County Park (**Figure 9**).

Location 1: The Banded Amphibole Gneiss

The banded amphibole gneiss (**Photo 1A**) is best described as a fine-grained banded gneiss with alternating hornblende-rich and plagioclase-rich layers. The hornblende-rich layers range anywhere from less than 1 cm to ~15 cm and contain about 85% hornblende and 15% granulated plagioclase with sparse idioblastic garnet. The plagioclase-rich layers are consistently thicker and contain approximately 15% hornblende. The garnets though scarce occur as coarse grained porphyroblasts in both layers (**Photo 1B**) although these garnets often show retrograde alteration back to hornblende. The garnets are typically poikiloblastic (**Photo 2**) with quartz, plagioclase and occasionally biotite inclusions. The hornblende occurs as euhedral to subhedral grains and the plagioclase as very-fine grained, dynamically recrystallized, matrix. The granulated plagioclase is typically labradorite to bytownite but anorthite (An_{92}) occurs as cores in some of the idioblastic hornblende to create an

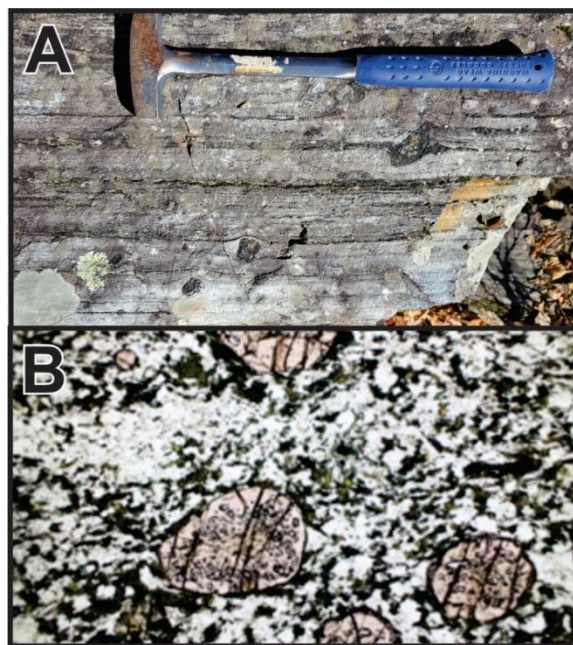


Photo 1: Photographs of the banded amphibolitic gneiss at Big Falls County Park. (A) Banded amphibole gneiss at Big Falls. B: Photomicrograph in plane-polarized light (25x) of large garnet porphyroblasts in quartzofeldspathic and hornblende matrix.

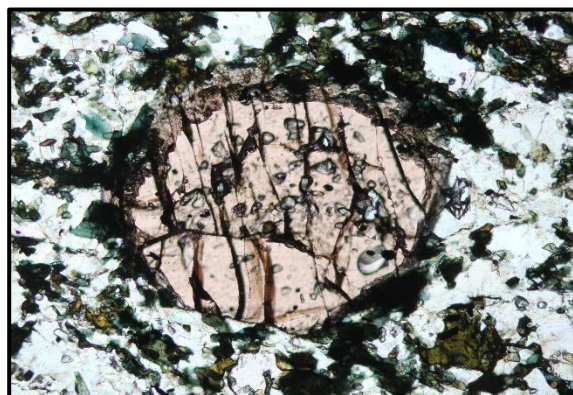


Photo 2: Photomicrograph in plane polarized light of poikiloblastic garnet in a hornblende and granulated plagioclase matrix (50X magnification).

unusual bi-modal plagioclase population (**Photo 3**).

Multiple shear zones and isoclinal folds are present throughout the outcrop, providing evidence for multiple deformation periods. Partially annealed shear zones can be traced across almost the entire unit that truncate and offset banding. There are also asymmetric

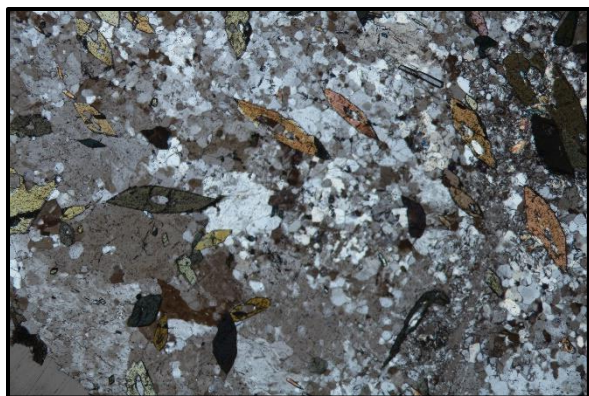


Photo 3: Photomicrograph in partially crossed polars of idioblastic hornblende with anorthite cores (An_{92}) in a matrix of granulated plagioclase (An_{72}) in the banded amphibole gneiss at Big Falls (Magnification is 50x).

amphibolite inclusions that show variable kinematics.

This unit was sampled for a recent zircon petrochronology study and produced significant results that question what is currently understood about the southern portion of the Penokean Orogen. This unit yields a U/Pb age of 1874.7 ± 2.1 Ma which temporally correlates with other VMS-forming events across the Pembine-Wausau terrane. Zircon trace element geochemistry of the sample indicate the sample formed in a hydrated but reduced melt in a back-arc setting where decompression was occurring in a metasomatized mantle (Weber and Lodge, 2022). Hf-Lu data from the banded amphibolite gneiss indicates a lack of older basement inheritance. The zircon petrochronology from these rocks contradicts the interpretation that Eau Claire Volcanic Complex was emplaced into the Archean Marshfield Terrane. These results have motivated additional research in the Eau Claire Volcanic Complex.

Location 2 Amphibole Schist

The further west along the Eau Claire River, the amphibole schist is exposed. This unit is best described as a dark green to black, fine grained thinly banded amphibole schist. Hornblende is the primary amphibole with lesser amounts of plagioclase. Based on whole rock XRF data, the

amphibole schist is enriched in MgO and FeO and depleted in CaO and Al_2O_3 in comparison to the banded amphibole gneiss (**Table 1**). Both major element (**Table 1**) and trace elements (**Figure 8**) characterize this as a MORB-like composition. Isoclinal folds of the foliation and the presence of ductile shear zones indicate deformation throughout this unit. The contact between the banded amphibole gneiss and the amphibole schist is buried by slumping blocks of the Cambrian Mount Simon Formation and vegetation.

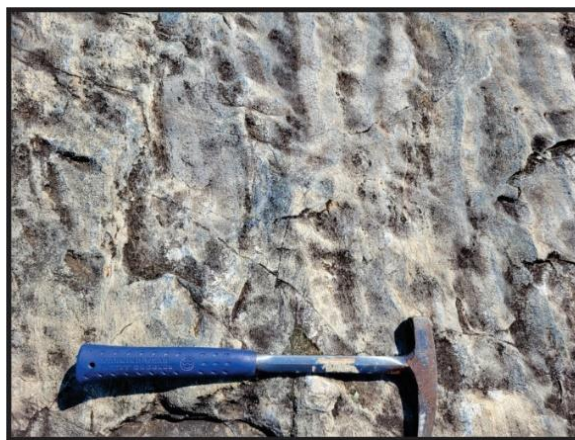


Photo 4: Outcrop photo of the feldspathic gneiss at Big Falls.

Location 3: Transition Gneiss and Feldspathic Gneiss

The amphibole schist gradually grades into the transition gneiss for a few meters as the unit contains fewer amphibole-rich layers and the plagioclase rich layers become more prominent (**Photo 4**). The feldspathic gneiss is primarily made of plagioclase, with 10-20% hornblende and lesser amounts of chlorite, epidote, and localized sulfidation with some pyrite mineralization.

Despite strong metamorphic recrystallization and structural fabric overprinting, there are some primary igneous textures that are preserved (Cummings, 1984). The compositional banding and layering throughout all units appear to be

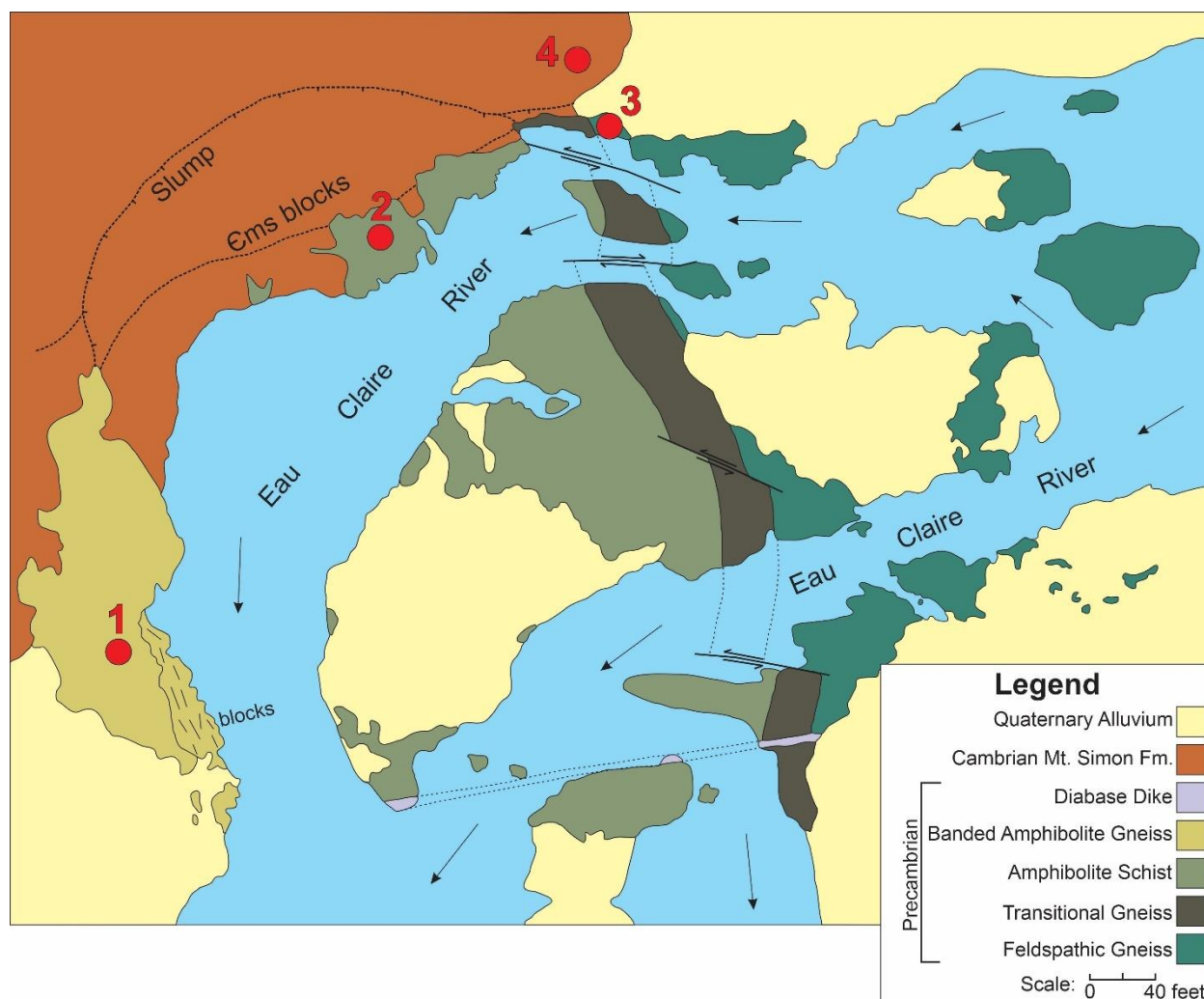


Figure 9: Geologic map showing outcrop locations at the Big Falls stop. Modified from Cummings (1984).

Table 1. Whole rock major element geochemistry (via XRF) from Big Falls for banded amphibole gneiss (location 1), amphibolite (location 2), and altered rocks at the Precambrian-Cambrian contact (location 3). Average MORB composition (Winters, 2010) is included for comparison.

Unit	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ T	CaO	MgO	MnO	Na ₂ O	K ₂ O	P ₂ O ₅	Totals
Unaltered											
Banded Gneiss 1	47.10	0.17	30.43	2.14	15.18	0.72	0.03	2.21	0.20	0.05	98.23
Banded Gneiss 2	48.43	0.21	30.56	3.35	14.65	1.38	0.04	2.97	0.25	0.04	101.88
Banded Gneiss 3	46.63	0.23	30.24	2.47	14.51	0.88	0.03	2.85	0.18	0.03	98.05
Altered (Depth)											
surface	52.06	0.44	16.66	6.72	0.88	5.21	0.04	0.11	9.28	0.03	91.43
.5m	50.26	0.74	15.28	11.56	0.63	6.18	0.05	0.05	9.14	0.28	94.17
1.0m	52.55	0.34	17.36	7.28	1.06	5.17	0.03	0.05	9.04	0.03	92.91
1.5m	52.92	0.32	18.23	5.79	0.80	4.91	0.02	0.09	9.37	0.03	92.48
Unaltered											
Amphibolite 1	51.10	1.34	15.50	13.73	9.07	5.79	0.19	3.11	0.29	0.23	100.35
Amphibolite 2	52.66	1.71	13.25	12.12	8.05	6.57	0.20	3.56	1.46	0.17	99.75
Average MORB (Winter 2010)	50.50	1.56	15.30	11.50	11.50	7.47	n/a	2.62	0.16	0.13	100.74
Altered (Depth)											
surface	61.93	0.94	17.36	4.98	0.63	2.70	0.02	0.32	7.10	0.19	96.17
.25m	62.79	0.99	17.18	6.10	0.65	3.05	0.04	0.20	7.22	0.19	98.41
.5m	61.48	0.84	17.04	5.04	0.64	3.02	0.02	0.21	6.97	0.18	95.44
.65m	62.62	0.88	16.26	5.42	0.64	2.68	0.04	0.23	7.15	0.18	96.10
.75m	60.07	0.95	16.84	5.60	0.64	2.89	0.04	0.07	7.45	0.21	94.76
1.0m	58.69	0.87	16.31	7.30	0.66	3.24	0.07	0.08	7.18	0.19	94.59

primary. Anorthositic autoliths are incorporated in a fine-banded, more mafic matrix near the transitional gneiss. The banding in the autolith is discordant to banding in the matrix and appear to be concentrated in bands but are not associated with boudinage fabrics. These observations were critical in interpreting the protolith of this region.

Location 4: The Great Unconformity and the basal portion of the Mt. Simon Formation

The large hillside on the north bank of the river and above the Precambrian outcrops is the Cambrian Mount Simon Formation. The base of the Mt. Simon Formation is a mix of coarse-grained quartz arenites and quartz pebble conglomerates.

The Great Unconformity creates a non-conformable contact between the Precambrian units at the previous three locations and the base of the Mt. Simon Formation. At Big Falls, a thin blue-green celadonite clay layer (**Figure 10**) has formed along the nonconformity as a result of K-metasomatism from basinal brines. The K-metasomatism at the unconformity is recognized throughout the midcontinent region and is related to MVT lead zinc deposits in the Tri-state region (See field trip 1 in this volume). Locally the celadonite acts as a fluid barrier for springs that flow along the unconformity. In other places, the contact appears to be relatively sharp with little alteration (**Photo 5**).

Stop 2 –Tonalite Breccia and Gneisses at Little Falls.

Lat: 44.8103° Long: -91.2825°

This location is just on the north side of the County Highway K bridge that crosses the Eau Claire River and is 300 m north of the exit to the south entrance of Big Falls County Park. Google Maps calls this place the East Eau Claire Canoe Landing, but the USGS refers to this location as Little Falls (so does the faculty and students at the University of Wisconsin-Eau Claire) and uses the



Photo 5: Nonconformity between the Cambrian Mount Simon Formation and the amphibolites at Big Falls County Park.

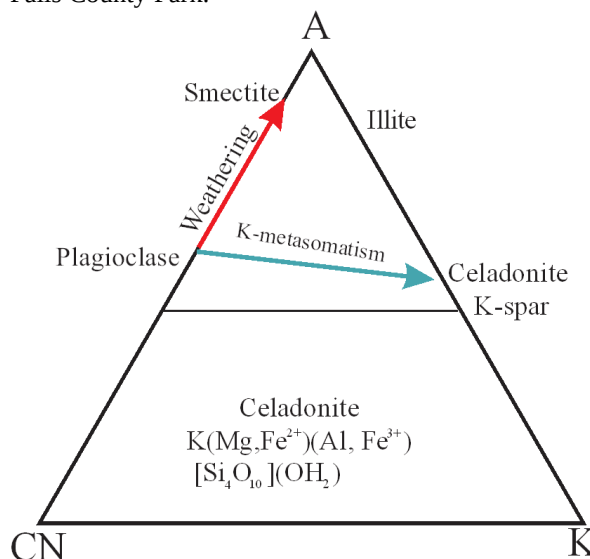


Figure 10: A-CN-K diagram showing chemical change due to weathering and the alternative path of K-metasomatism. The celadonite at Big Falls is not a weathering profile and requires adding substantial potassium and to produce the celadonite and authigenic K-spar seen along the unconformity (see Table 1 for chemistry).

bridge at this location for the Eau Claire River flood stage measurement. Just north of the bridge there is a small parking area on the west side of the road. There are several small foot paths that lead down to the river's edge. The outcrops are mostly exposed immediately around and under the bridge. The quality of exposure here changes all the time as flooding conditions sometimes

buries parts of the outcrop with sand and downed vegetation. To access the south bank of the river, a short bush traverse (100-300 m) will be required from the southside of the bridge along the riverbank.

If water level and time permit, this stop has four locations of interest that highlight the intrusive history in the region. The outcrops in this area expose an inclusion-rich intrusive contact between a foliated tonalite and lensoidal amphibolite and are cut by younger pegmatitic and mid-continent rift diabase dykes (**Figure 11**). This location is used to teach students at the University of Wisconsin-Eau Claire about interpreting relative geologic time and observing contact relationships. The absolute ages of these rocks are unknown as recent attempts to isolate zircons from the tonalite were unsuccessful.

Location 1: Gneissic Tonalite Breccia

The gneissic tonalite breccia is the most prominent unit at Little Falls (**Photo 6**). Roughly 90% of xenoliths in the breccia are characterized as banded gneiss to banded amphibolite containing 50-80% hornblende and 20-40% plagioclase with lesser amounts of biotite. These xenoliths range in size anywhere from less than 1cm to greater than 20 cm, and they are hosted in a biotite tonalite intrusion which destroys the older banded gneiss. The banded gneiss xenoliths are also elongated and contain folds. Ultramafic xenoliths are scarce but also present. These xenoliths contain over 90% hornblende with lesser amounts of epidote-clinozoisite and plagioclase. Occurring in localized clusters, the fragmented ultramafic xenoliths indicate these were most likely part of a larger block but separated during the tonalite intrusion event (Myers et al., 1980). The tonalite is composed of 35-40% plagioclase (An₅₀₋₅₅), about 30% hornblende, 25-30% quartz, 5-10% biotite, and accessory epidote. Myers et al. (1980) interpreted the fabric in the rock as flow-lamination, however it is parallel to regional magnetic lineaments suggesting it may be a structural fabric. Large variation in mafic mineral abundance indicates



Photo 6: Gneissic tonalite breccia highlighting some of the banded gneiss xenoliths. Some of the smaller xenoliths here are elongated.

significant assimilation of amphibolite. Biotite commonly produces a crude foliation that may have formed from hornblende during a later deformation.

It is clear the tonalite is metamorphosed and is an important part of determining the nature of the Eau Claire Volcanic Complex. However, efforts to constrain the timing of this intrusive event have yielded conflicting results. Van Schmus (1980) yielded a U/Pb age of 1842 ± 10 Ma utilizing zircon fractions (i.e. not modern single crystal methods). Sims et al. (1989) reported a U/Pb age of 1856 ± 5 Ma from a xenolith at Little Falls. Assuming that the amphibolite xenoliths at Little Falls are from the same amphibolite unit at Big Falls that was dated at 1875 Ma, then there is a clear conflict. The tonalite was sampled for that recent petrochronologic study (Weber and Lodge, 2022) but yielded very few zircons. Resolving the timing of emplacement and tectonic setting of this intrusion will help better understand the geodynamic setting of the Eau Claire Volcanic Complex.

Location 2: Diabase

Along the north side of the river and east of the bridge, lies one of many mid-Proterozoic diabase dykes associated with the mid-continent rift in the

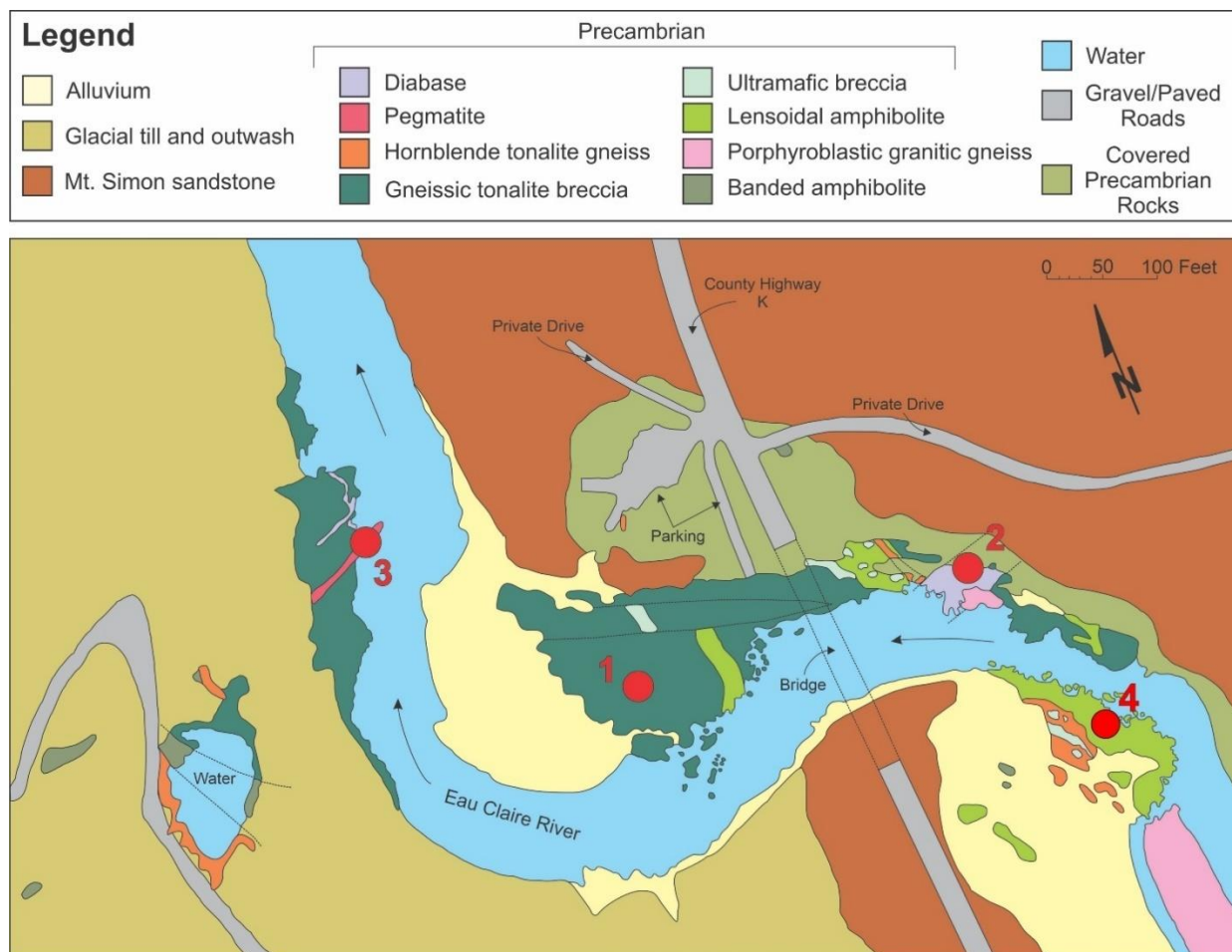


Figure 11: Geologic map of the Little Falls area showing the locations of interest at this stop. Figure modified from Myers et al (1980).

Eau Claire region. Like many other diabase dike outcrops in the area, this diabase exhibits both clean columnar jointing and well-defined chilled margins. This diabase is only a few meters in size and disappears beneath the surrounding overburden (**Figure 11**). The diabase has a medium-grained, equigranular texture and does not have any foliation or recrystallization textures and is clearly post-metamorphism.

Location 3: Pegmatite Dike

On the west bank of the river lies a 2 m wide pegmatite dyke (**Photo 7**). Outlier boulders of this pegmatite can be found on the eastern bank of the river. The alkali-feldspar crystals in this outcrop reach sizes greater than 30 cm. Various sizes of quartz veins also crosscut this unit. This granite pegmatite dike is one of a handful seen

throughout the Eau Claire volcanic complex along the Eau Claire River where the Precambrian rocks are exposed. Their macro- and microscopic characteristics indicate they are



Photo 7: 18m-wide pegmatite in the Eau Claire River downstream from Little Falls.

clearly younger than the Penokean deformation and metamorphism and could be related to Mazatzal or Yavapai orogenic events to the south like the Wolf River Batholith in northcentral Wisconsin.

The pegmatites, despite their pink color, are primarily composed of plagioclase with an overprint of potassic alteration. Most of the alkali-feldspar occurs along cleavages, crystal boundaries and fractures indicating it is a late phase in pegmatite formation. The plagioclase in the pegmatites is primarily albite but ranges from An₀ to An₃₀. The euhedral and inclusion free garnets (**Photo 8**) have a limited range of chemistry close to 50% almandine and 50% spessartine which is similar to magmatic garnet compositions in other garnet-quartz-albite pegmatites (Muller et al., 2018).

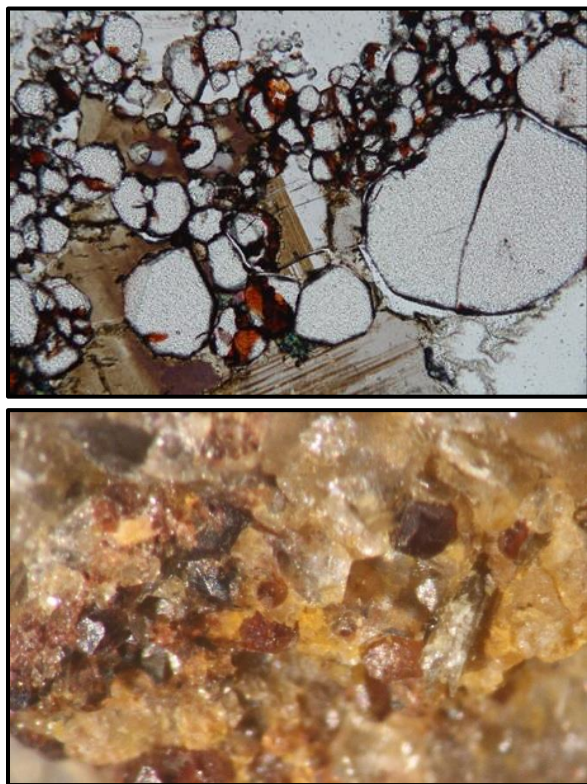


Photo 8. Top: Photomicrograph (25X in plane polarized light) of garnet cluster in quartz and albite from Little Falls pegmatite dike on the west side of the river. Bottom: Almandine/spessartine garnet clusters at the same location at Little Falls are magmatic in origin.

The mineralogy of the pegmatites is very complicated with many accessory carbonate, phosphate and oxide phases enriched in Nb, Y, F, U, Th and REE. Zircon in the pegmatites indicate extreme fractionation (**Figure 12**). The zircons also show considerable xenotime (Y,P) substitution and considerable solid solutions with both coffinite (USiO₄) and thorite (ThSiO₄).

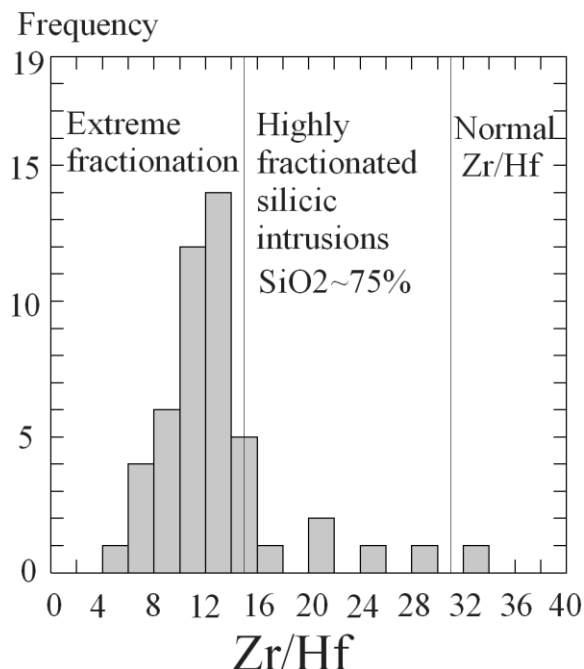


Figure 12. Zr/Hf in pegmatites from the Eau Claire River Complex.

The Eau Claire River pegmatites have many characteristics of pegmatites in the Nb/Y/F (NYF) family of rare element pegmatites and NYF pegmatites are always associated with metaluminous to alkaline (or peralkaline) granites (Cerny and Ercit, 2005)

Location 4: Amphibolitic Gneiss

The xenoliths within the tonalite are assumed to be derived from the nearby outcrops of the amphibolitic gneisses. Unlike the planar banding at Big Falls, the amphibolitic gneisses here is more lensoidal with cm- to dm-scale lens-shaped, hornblende-rich pods surrounded by more plagioclase-rich “matrix” and quartz-veining.

Detailed work has not been completed on these outcrops but are assumed to be petrogenetically related to the amphibolites at Big Falls. Future research will examine these exposures more closely.

North and South Fork Region

This is the part of the trip where information on these rocks is sparse and new data is only just becoming available. To our knowledge, the rocks in the North and South Fork areas of the Eau Claire River have not been studied in any detail since Myers et al. (1980). Much of the area is remote and sparsely developed and very few outcrops are easily accessible. Field work in the 2022 summer relied on one-way, day-long kayak trips along different segments of the North Fork. Aside from the occasional powerline, field work on these stretches of the Eau Claire River felt wild and remote. This field work aimed to characterize

the trace element geochemistry and zircon petrochronology of the rocks in this region to make better links with the rest of the Eau Claire Volcanic Complex. Much of that data is still pending or preliminary, so results will be forthcoming soon. The goal for the field trip in this region is to show as many of the rocks as possible, regardless of how much we know about them.

The region consists of amphibolites, feldspathic gneisses, and foliated granitoids and are cross-cut by younger, undeformed granitic pegmatites (**Figure 13**). A metarhyolite from this region yielded an age of 1858 ± 5 Ma (Sims et al, 1989) and was one of the key samples that linked Penokean volcanic processes to the Marshfield terrane. Myers et al. (1980) interpreted mappable contacts between intrusions and foliated

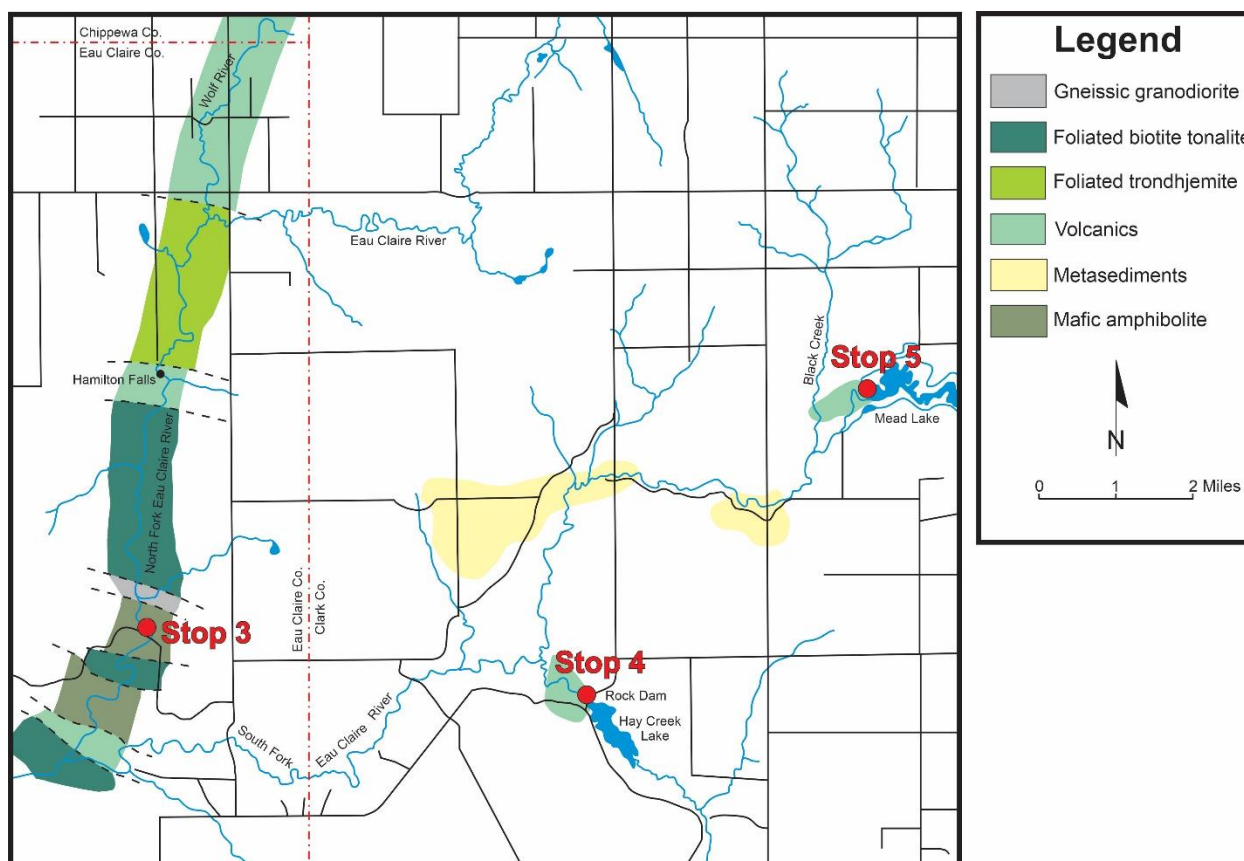


Figure 13: Geologic map of the North and South Fork of the Eau Claire River in eastern Eau Claire County and western Clark County. Figure modified from Myers et al. (1980).

supracrustal rocks in this area to be sheared and nearly vertical and that they enclose lensoidal fault slices which have been juxtaposed mainly by strike-slip displacement. Outcrops of amphibolite in the southern part of the map (**Figure 13**) near the confluence of the North and South Forks of the Eau Claire River were interpreted to be part of the “Chippewa Amphibolite Complex” (Myers et al., 1980) which is broadly supported by regional aeromagnetic maps (**Figure 5**). Myers et al. (1980) interprets the volcanic rocks in this region to unconformably on amphibolites, but geochronologic data is lacking to make any absolute local or regional correlations. Regional metamorphic grade is estimated to be upper greenschist to lower amphibolite based on the presence of garnet, epidote, muscovite, and hornblende.

Preliminary geochemical results from the North Fork region begin to reveal the setting of these volcanic and intrusive rocks. Volcanic protoliths are bimodal (**Figure 14**) with tholeiitic mafic rocks with oceanic affinities (**Figure 15**) and FI- to FII-type felsic rocks arc-like affinities (**Figure 16**). More work needs to be done before we can concretely interpret the setting of this region of the Eau Claire Volcanic Complex.

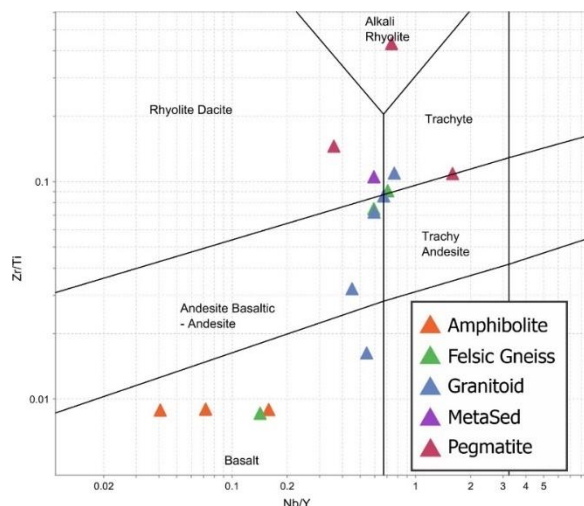


Figure 14: Trace element classification diagram from Pearce (1996), modified from Winchester & Floyd (1977).

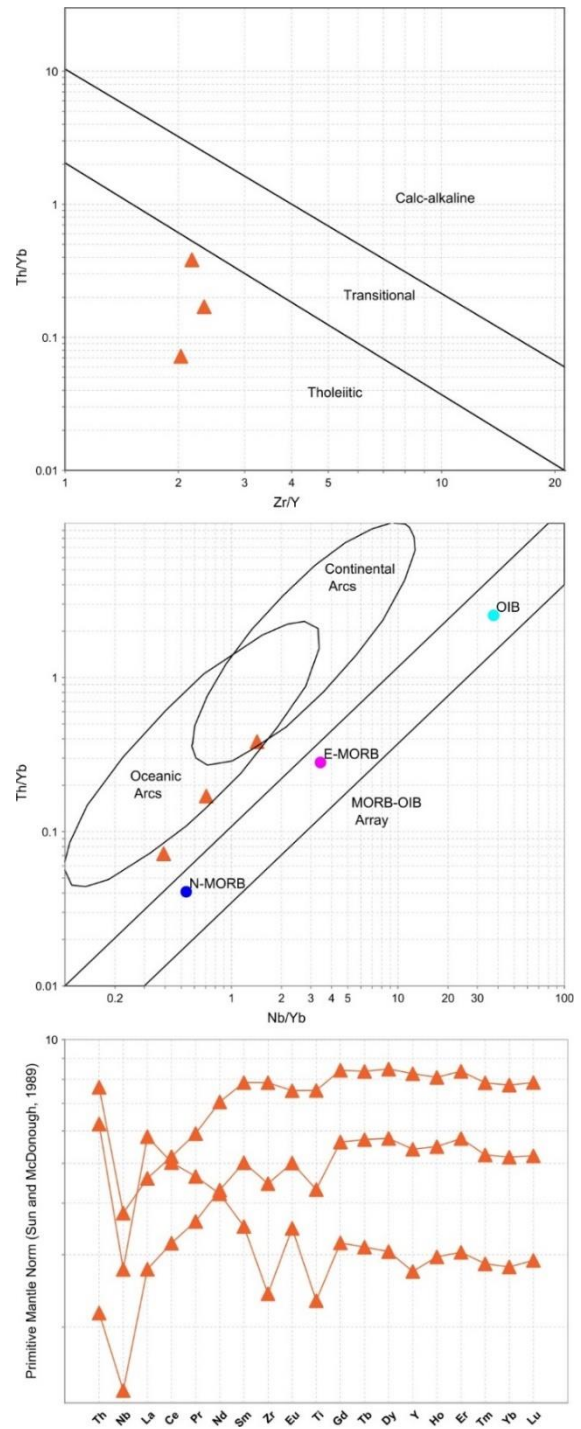


Figure 15: Trace element geochemical characteristics of the mafic rocks from the North Fork region of the Eau Claire River. Top: Magmatic affinity diagram for sub-alkaline basalts from Ross & Bedard (2009). Middle: Mantle source discrimination diagram from Pearce (2008). Bottom, Primitive mantle-normalized trace element diagram using values from Sun and McDonough (1989).

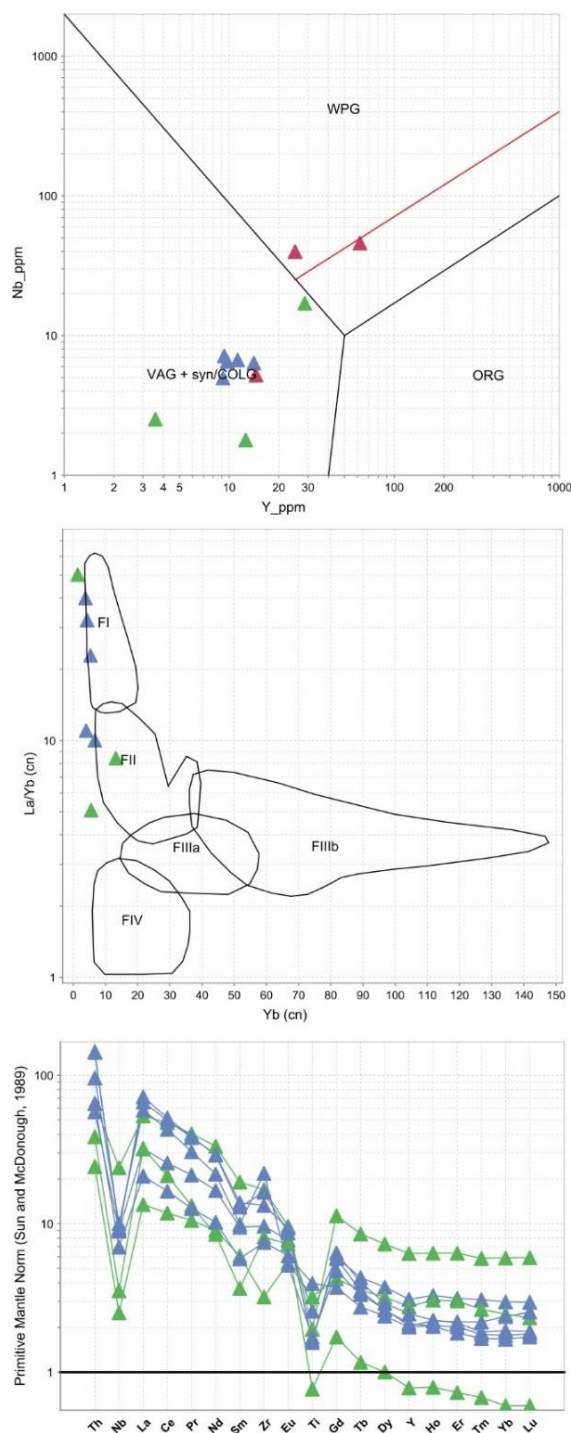


Figure 16 – Trace element geochemical characteristics of felsic rocks from the North Fork region of the Eau Claire River. Top: Nb/Y discrimination diagram for granites from Pearce (1984). Middle: F-type felsic discrimination diagram from Hart et al. (2004). Bottom: Primitive mantle-normalized diagram using values from Sun & McDonough (1989).

These settings are not typical of continental settings, which continues to question the relationship between the Eau Claire Volcanic Complex and Marshfield Terrane.

Stop 3 – Amphibolite and Intrusions at Knights Pool

Lat: 44.7482° Long: -90.9669°



The directions to get to this stop are a little more elaborate since it is in a more remote location. From the community of Augusta, take State Highway 27 north for 4.4 miles to County Road GG. Turn east on to County Road GG and drive 4.7 miles to the intersection of Channey

Road just after the bridge over the Eau Claire River. Turn east on Channey Road (note that this is an unpaved road) and drive for 4.6 miles until the road crosses the North Fork of the Eau Claire River. This location is called Knights Pool and is labelled by signage. The accessible outcrops in this region are immediately beneath the bridge and are accessible by foot trails. There are larger outcrops north of the bridge that are accessible by a small, 150 m bush traverse along a sparsely used trail. At both locations, rocks are immediately adjacent to a shallow but fast-moving river and caution should be used.

Knights Pool is located at the bridge on Channey Road as it crosses the North Fork of the Eau Claire River along the southern edge of the North Fork Eau Claire River State Natural Area. There is more of the Eau Claire Volcanic

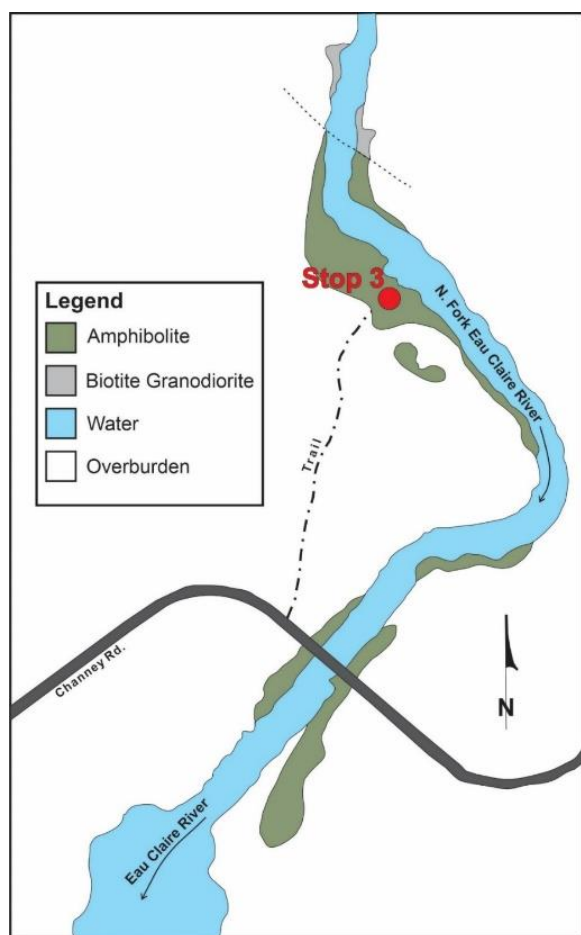


Figure 17: Geologic map of the Knights Pool area. Map is modified from Myers et al. (1980).

Complex exposed in this section of the river up- and downstream of this location. However, this is the only easily accessible section by foot. At Knights Pool, there are mostly strongly foliated and deformed amphibolites that are intruded by a biotite granodiorite (**Figure 17**).

Amphibolite: The amphibolite at Knights Pool is characterized by a fine to very fine grained mafic lined amphibolite with stretched quartz-filled amygdulites, relict pillow structures, and wispy textures suggesting a mafic flow (**Photo 9**). Thin sections of the amphibolite clearly show the lineations present in the amphibolite here (**Photo 10**). Several stages of deformation occurred starting with the amphibolite being isoclinally folded, then intruded by aplite veins, and intruded by large granodiorite body (Myers et al., 1980). The shearing in of the granodiorite body created a mylonite gneiss along the contact with the amphibolite.

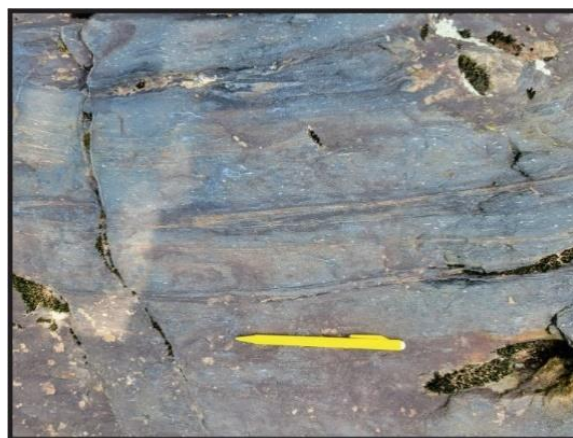


Photo 9: Outcrop of the amphibolite showing both the isoclinal folds and strained amygdulites present in the unit.

Trace element geochemistry of the amphibolites at Knights Pool are notably LREE-depleted with strong negative Nb anomalies (**Figure 15**). This suggests it was derived from strongly depleted but metasomatized mantle. This type of environment, presumably a mature back-arc, rarely exists in a continental setting. The granitoids in the map area have classic enriched LREE and Th with depleted Nb and HREE signatures suggesting they are related to a

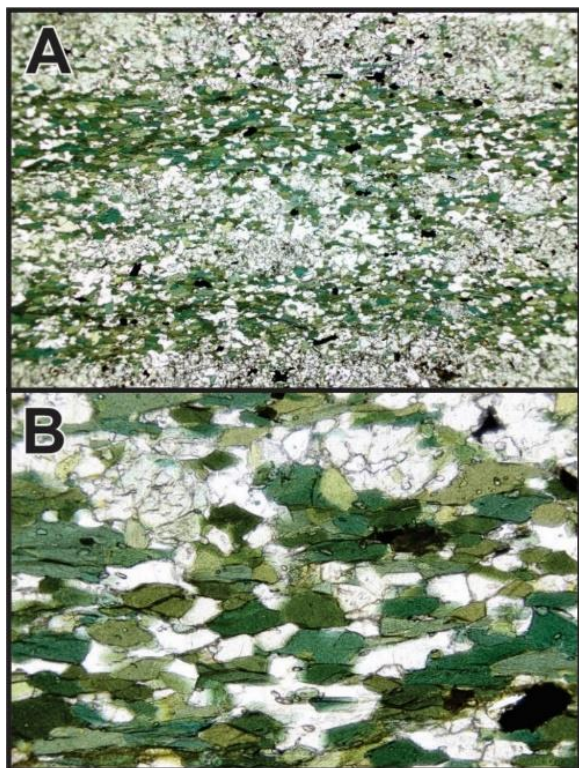


Photo 10: Photomicrographs in plane-polarized light of the amphibolite at Knights Pool amphibolite under thin section at 25x magnification (A) and 100x magnification (B). Hornblende forms a clear lineation.

different tectonic event when the crust was thicker, and garnet was stable to deplete HREE.

Granodiorite Intrusion: The medium to coarse grained biotite granodiorite intruded the amphibolite creating a mylonitic fabric along the contact. The intrusion, shearing along the contact, caused the folding of the aplite veins seen in the amphibolite. Quartz is strongly recrystallized and biotite concentrations define a weak foliation. Thin section photos highlight how the quartz is being recrystallized, as well as show the alignment of biotite aggregates (**Photo 11**).

Stop 4 – Metarhyolite and Phyllite at Rock Dam

Lat: 44.7338° Long: -90.8469°

From the previous stop, continue eastward on Channey Road until it ends at County Highway H. Turn southward and drive 1.5 miles to Rock Dam Road. Turn eastward and take this road into

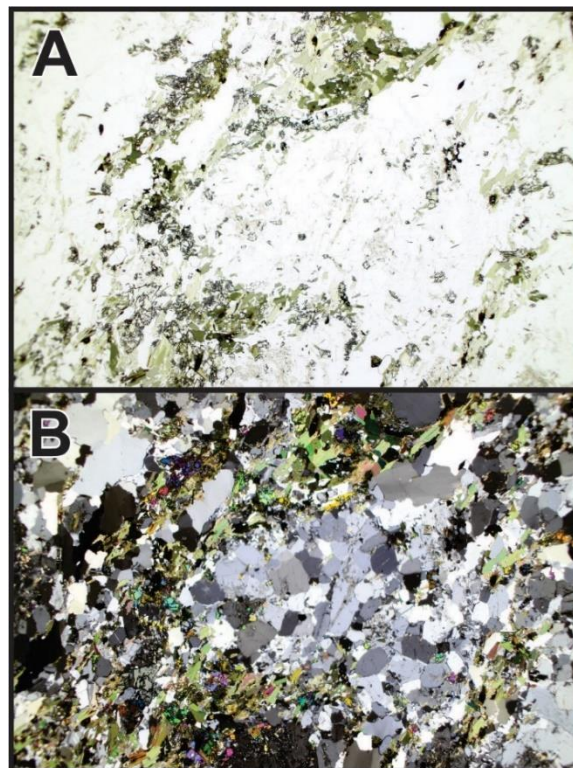
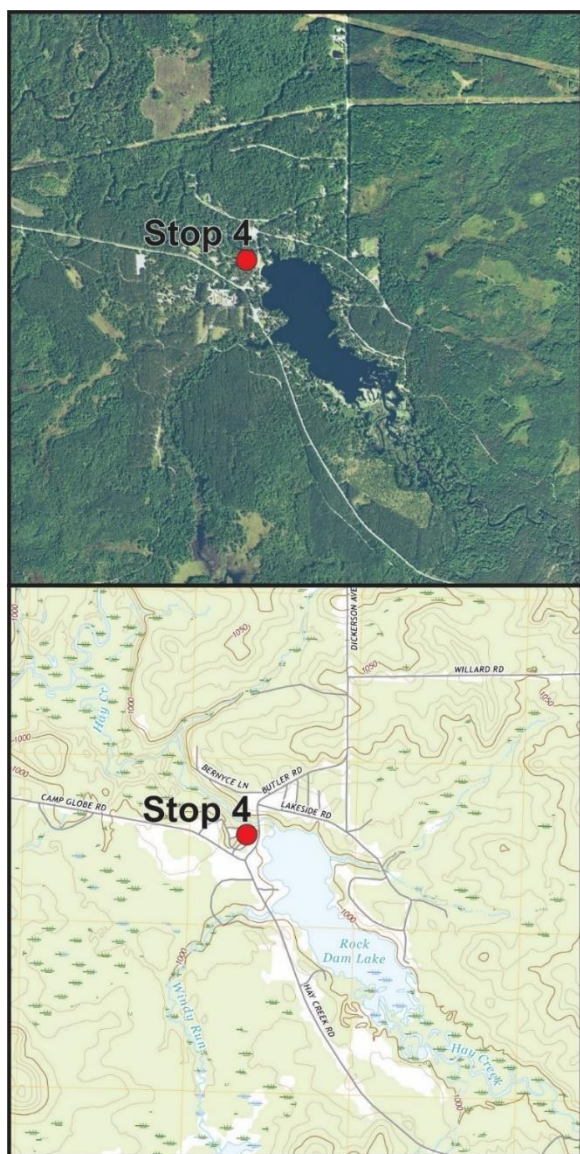


Photo 11: Photomicrographs of the biotite granodiorite at Knights Pool in (A) plane-polarized light, and (B) cross-polarized light. Both images are 25x magnification.

the village of Rock Dam. Turn northward on Butler Road and use the parking lot on Hay Creek Lake just south of the bridge over Hay Creek. The outcrop of phyllite are under the bridge and near the Rock Dam spillway. The nonconformity and metarhyolite can be better accessed from within the Rock Dam Campground near campsite 90.

This location reveals another nonconformity between the Precambrian and Cambrian Mount Simon Formation. The metarhyolites and phyllites in this area are strongly mylonitized and primary structures are difficult to interpret. The metarhyolite at this location described by Myers et al. (1980) is the only reference to a rhyolitic unit in the Eau Claire Volcanic Complex. Since Sims et al. (1989) dated a metarhyolite at 1858 ± 5 Ma in the Eau Claire River and cited Myers et al. (1980), it presumably came from this location. If that is the case, then the rocks at Rock Dam are of regional significance because it is one of



the very few dated Penokean supracrustal rocks within the Marshfield terrane.

Mylonitized Metarhyolite: Myers et al. (1980) admittedly conceded that determining the protolith of this outcrop can be challenging considering the similarities between sheared porphyritic rhyolites and leucogranites. The mylonite here is pale pink metamorphosed porphyritic rhyolite containing quartz eyes that can be described as phenocrysts or clasts. These quartz eyes are roughly 1-2.5 mm in size and under thin section show a subrectangular to lenticular shape (Myers et al. (1980). The

estimated mineral percentages and matrix of this rock is composed of a very fine-grained alkali-feldspar (57%), quartz (35%), muscovite (3%), magnetite (2%), and biotite (1%). The quartz eyes along with the absence of feldspar porphyroclasts suggests that the quartz either originated as phenocrysts or clasts in a tuff (Myers et al. (1980). Foliation trends east-west and is near vertical.

Phyllite: Closer to the base of Rock Dam lies a muscovite-rich phyllite composed of alkali-feldspar, quartz, and muscovite (**Photo 12**). This outcrop lacks both the quartz eyes and biotite possibly indicating a separate protolith than the metarhyolite (Myers et al. 1980).



Photo 12: Outcrop photo of strongly foliated phyllite at Rock Dam near spillway. Cambrian strata can be seen in background on opposite bank of river.

Mt. Simon Formation: The basal part of the Mt. Simon Formation, a conglomerate layer containing pebbles of vein quartz and rhyolite, is exposed at this location (**Photo 13**). The contact between the Mt. Simon and the Precambrian metarhyolite shows about 5 m of relief. Many of the locations exposing the Great Unconformity in the Eau Claire region show deep weathering of the underlying Precambrian rocks, however there is not much weathering of the Precambrian metarhyolite here.



Photo 13: Basal pebble conglomerate in Cambrian strata overlying metarhyolite at Rock Dam.

Stop 5 – Metavolcanic Rocks at Mead Lake

Lat: 44.7885° Long: -90.7742°

From the previous stop, continue northward on Butler Road for 0.8 miles to the intersection with Willard Road. Drive eastward on Willard Road for 2.0 miles and turn north onto County Road M. Drive northward on County Road M for 1.6 miles and turn eastward onto Rocky Run Road. Drive 1.2 miles on Rocky Run Road and turn northward on Bruce Mountain Road that will turn into South Lake Road. South Mead Lake Park will be 1.4 miles down this road. Park there, and the outcrops are on the riverbank west of the Mead Lake Dam spillway.

The bedrock exposed at this location is primarily a foliated, fine-grained chloritic metavolcanic rock (**Photo 14**). There has been no known study of this rock, and our data is still pending. Nonetheless, it is apparent that the metamorphic grade seems to be decreasing in this part of the Eau Claire Volcanic Complex. This is in stark contrast to the rocks in the Chippewa River valley (Fieldtrip 1, this volume) and Big Falls region (Stops 1-2). Future work in the region will utilize every outcrop, even small ones like this location, to better describe and define the tectonics and metallogeny of the Eau Claire Volcanic Complex.

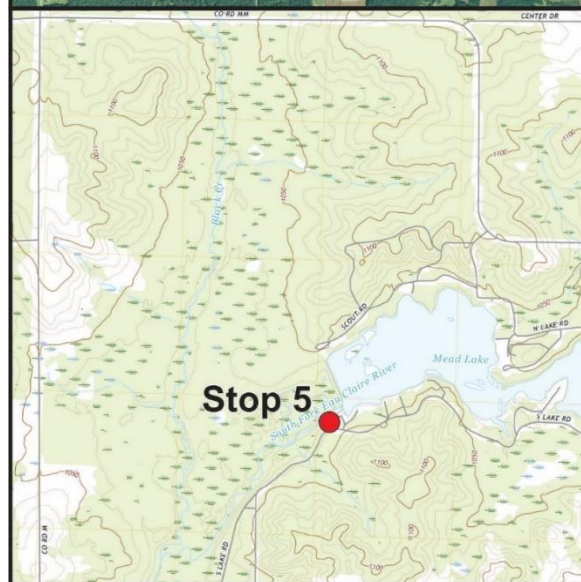


Photo 14: Outcrop photo of metavolcanic rocks at Mead Lake Dam.

Acknowledgements

Despite decades of regular visits from groups from the University of Wisconsin-Eau Claire, the most extensive detailed maps and rock descriptions were provided by Paul Myers and collaborators in the 1980 ILSG guidebook. Outside of Big Falls County Park, many of those locations have not been visited since then. A lot of the geologic descriptions from those lesser-known areas have been updated from Myers et al. (1980) and figures have been digitized while adding new data and insights where available.

In addition, the authors of this guidebook would like to thank the countless undergraduate and graduate students that have worked on these outcrops and have continued to inspire new work in the region. Specific acknowledgement is deserving to Matt Leahy and his undergraduate research project in the North Fork region in providing some insight into that part of the complex.

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Field Trip 4 – Quaternary Geology and Geomorphology of the Eau Claire Region

Douglas J. Faulkner

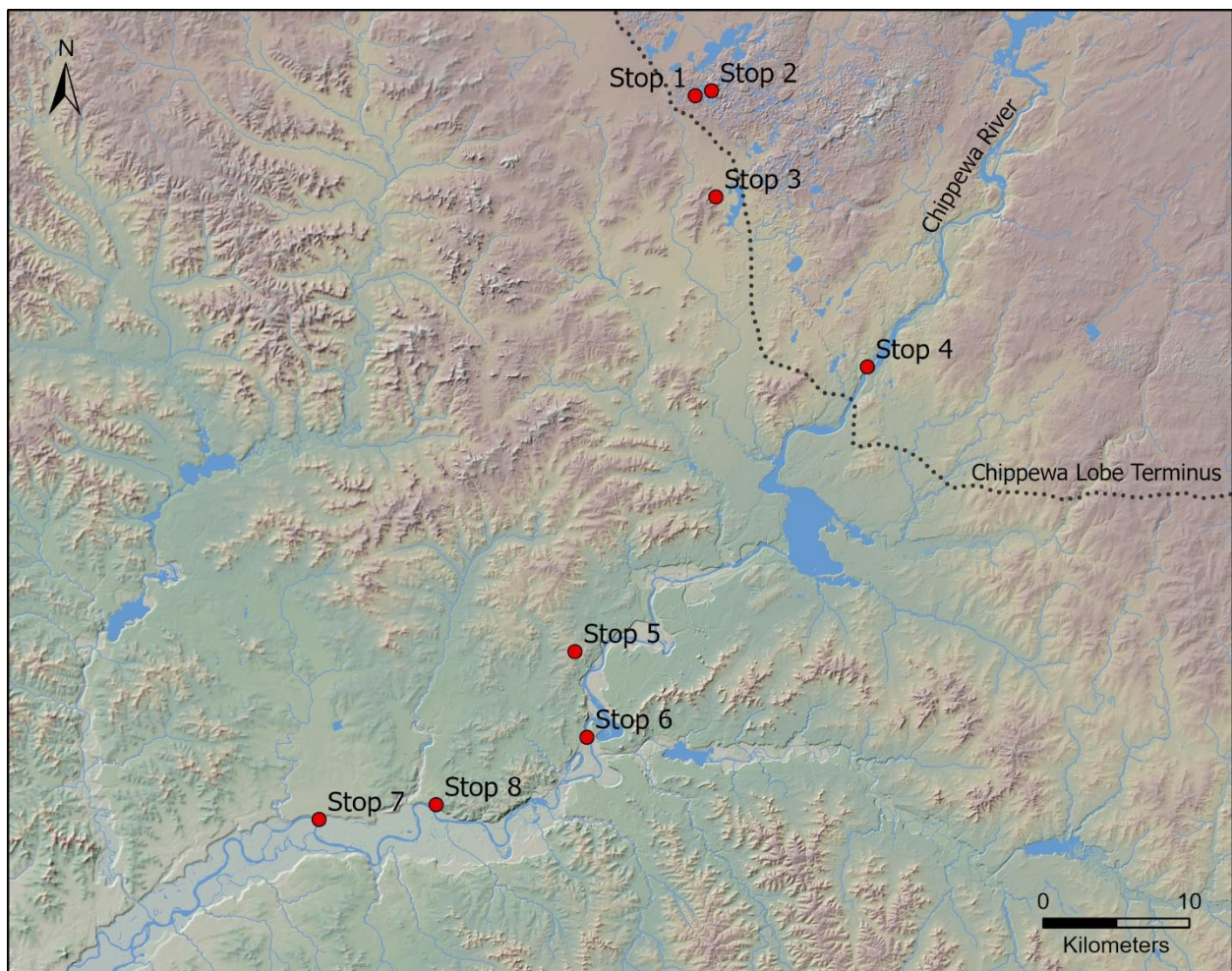
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Introduction

Eau Claire lies close to the outermost edge of the former Chippewa Lobe of the Laurentide Ice Sheet as it existed during late Wisconsinan time (MIS-2) (Fig. 1). To the south are older glacial deposits and then the Driftless Area, which apparently was never glaciated. This all-day field trip will concentrate on three aspects of the region's landscape development from the late Wisconsinan to the late Holocene: glacial, fluvial and aeolian.

Glacial Landscapes

Northern Wisconsin was glaciated multiple times in the Quaternary. The oldest glacial deposits were derived from the northwest and were likely deposited prior to 780,000 ka. These include the Pierce and Marathon Formations (Rawling et al., in review; Syverson et al., 2011). These deposits are poorly preserved where they occur at the surface (Rawling et al., in review) and although their occurrence is documented in the subsurface (Attig 1985 and 1993; Woodruff et al., 2004), their regional distribution is poorly documented. During the most recent glaciations, ice flowed from the northeast through the Superior Basin until it was thick enough to spill over the regional bedrock divide (Attig and Rawling, 2018). Ice formed during an earlier glaciation deposited glacial and meltwater sediment of the River Falls Formation (Syverson, 2007). River Falls tills and outwash are preserved on uplands in the Eau Claire area, and landforms associated with this advance have been eroded and are not preserved. The best preservation of landforms is associated with the late Wisconsinan ice (ca. 25–11.5 yr B.P.), which formed the Chippewa Lobe that reached its maximum extent at the Chippewa Moraine. This ice was subject to stagnation whenever the ice profile in the Superior Basin lowered, resulting in an ice margin landscape consisting of broad (10s of kilometers) zones of stagnant ice features such as disintegration ridges, ice-walled lakes, and kettles.

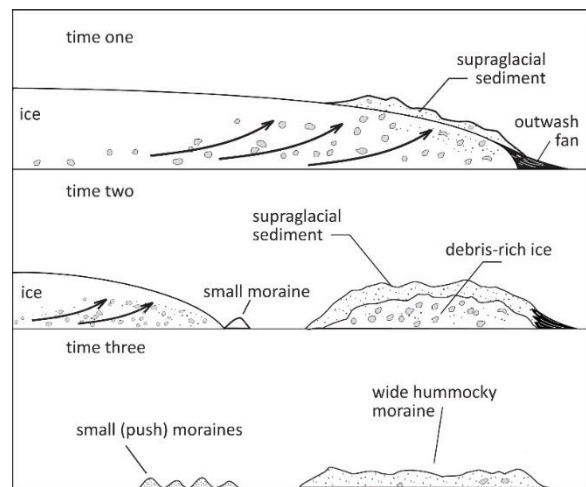
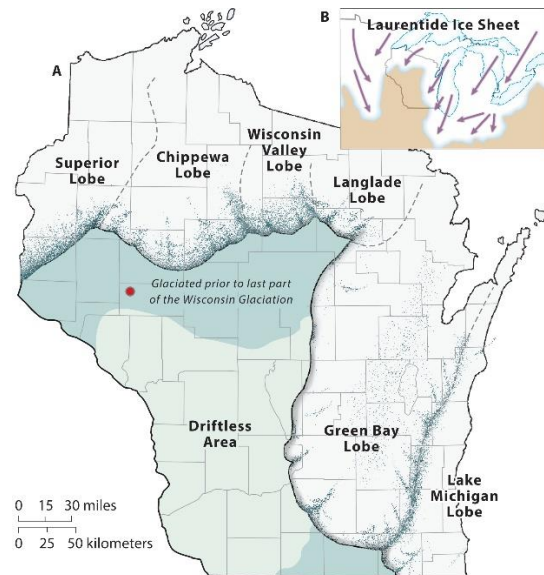


Figure 1. Top: Map of Wisconsin showing areas covered by lobes of the southern Laurentide Ice Sheet during the late Wisconsinan (MIS-2) Glaciation; inset map shows distribution of ice in the Great Lakes region. The red circle shows the location of Eau Claire near the southern edge of Chippewa Lobe. Bottom: Schematic illustration showing how moraines form over time. Supraglacial sediment accumulates at a stable ice margin (time one). As glacier retreats, buried ice is preserved under supraglacial sediment and minor moraines form if margin temporarily stabilizes (time two). The distribution of moraines after ice has melted (time three; modified from Attig and Rawling, 2018).

Fluvial Landscapes

The Chippewa River is the second largest stream in Wisconsin, draining a watershed of approximately 25,000 km² to the upper Mississippi River (UMR). During the Late Wisconsinan, it was the primary stream draining meltwater from the Chippewa Lobe. Overloaded with glacialigenic sediment, the lower Chippewa River (or LCR, which refers to the river beyond the Chippewa Moraine) filled its bedrock valley with sandy outwash to depths exceeding 50 meters. Then, sometime between 18-16 ka, the UMR incised 15 m, and at ~13.4 ka, it incised an additional 40 m (Knox 2007; Loope 2012; Gran et al. 2013). Each of these incision episodes abruptly lowered the base level of the LCR, creating knickzones that migrated up the LCR and its tributaries. The incision resulting from knickzone migration created the Wissota terrace, a prominent landform in the LCR valley that marks the maximum height of LCR aggradation during the Late Wisconsinan (Andrews 1965).

The process of knickzone migration and incision up the LCR was episodic and unexpectedly prolonged. The episodic history of knickzone migration and incision is clearly indicated by the number and spatial distribution of terraces found in the LCR valley below the Wissota. Instead of two terrace levels resulting from the two episodes of abrupt base-level fall, there are as many as seven (Fig. 2). Each of these levels represents a period when the river was migrating laterally and forming a floodplain, followed by an episode of renewed incision that left the floodplain as a terrace. The prolonged history of knickzone migration and incision along the LCR is revealed by the optically stimulated luminescence (OSL) ages of terrace alluvium from several sites in the LCR valley (Fig. 3). These OSL ages indicate that knickzone migration took thousands of years longer than studies of modern alluvial streams affected by minor base-level falls suggest it should've taken (Begin 1986, 1988; Begin et al., 1981)

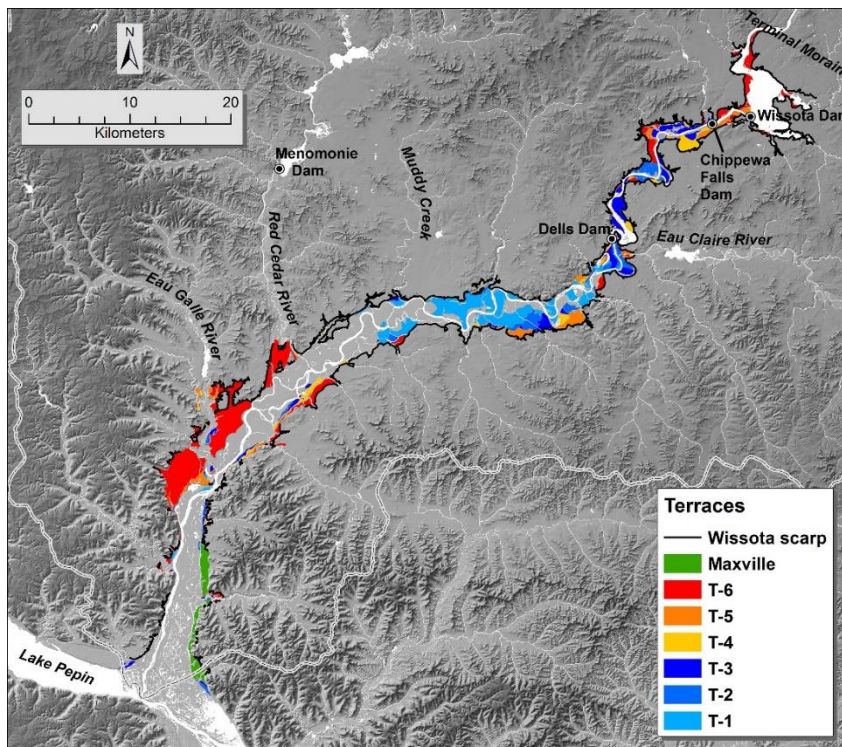


Figure 2. Terraces of the LCR valley. The names of terraces below the Wissota terrace are based on their height above the modern floodplain and distance below the Wissota. From lowest to highest these are T-1, T-2, T-3, T-4, T-5, and T-6. One terrace that does not fit into the T1-T6 schema is found in the relatively narrow bedrock valley downstream from the Eau Galle-Chippewa River confluence. Named the Maxville Terrace, this terrace slopes from the level of the Wissota at its upvalley end to a level equivalent to T-4 in the UMR valley. (Figure from Faulkner et al. (2016).)

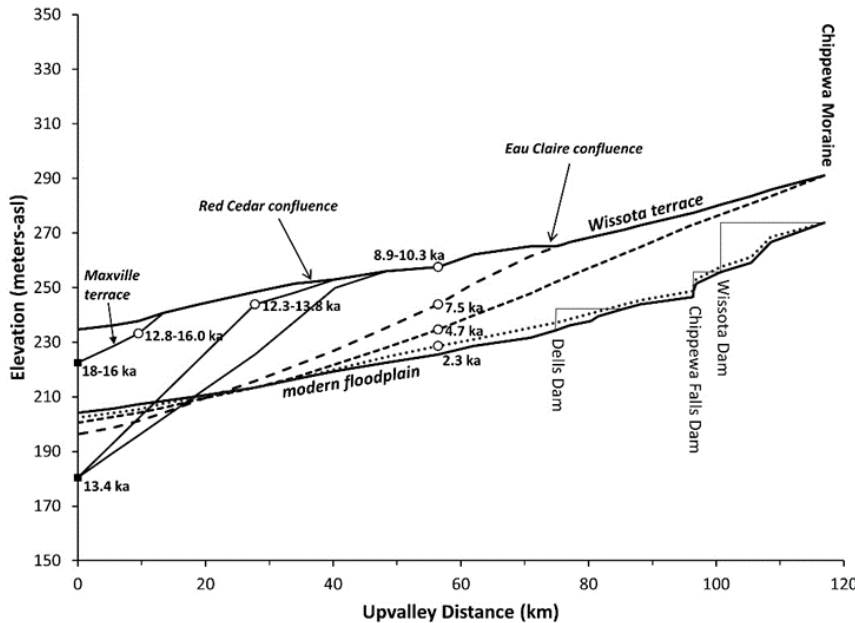


Figure 3. A proposed model of the evolution of the longitudinal profile of the LCR in response to UMR incision labeled with OSL ages of terrace alluvium. The profiles were constructed by connecting scattered terrace remnants, except for the Wissota and Maxville terraces, which are relatively continuous features. (Modified from Faulkner et al. (2016).)

Autogenic variations in the amount of sediment supplied to the river likely explain why the migration of knickzones and incision up the LCR was episodic and prolonged. Incision resulting from knickzone migration would've created a relatively deep narrow channel with steep, unstable banks. Bank collapse, promoted by lateral stream erosion, would have greatly increased the supply of sediment to the stream. With more sediment to transport, the stream would no longer have had excess power, causing knickpoint migration and incision to slow or cease altogether. This, in turn, would've allowed lateral stream migration and floodplain formation to occur. Over time, lateral erosion and bank failure would've caused the banks to move away from the stream and become less steep, leading to a reduction in the amount of sediment supplied from them to the LCR. With a declining sediment supply, the stream would've once again had excess power, resulting in renewed knickzone migration and incision, at least until the process repeated itself farther upstream.

The supply of sediment to the LCR from its tributaries was also subject to autogenic variations. LCR incision resulting would have lowered base level for its tributary streams, creating knickzones that then migrated up them.

The subsequent tributary incision would have caused a dramatic increase in the supply of sediment to the LCR. Sediment supplied from tributaries would have remained high until their incised channel banks began to stabilize. But until that happened, high amounts of tributary sediment would have affected knickpoint migration and incision on the LCR, slowing it down and possibly causing it to stop. It is likely that autogenic variations in sediment from the largest tributaries, the Red Cedar River and the Eau Claire River, had the biggest impact on the LCR.

Aeolian Landscapes

There is abundant evidence throughout the Eau Claire region that wind has been a significant geomorphic agent during the late Quaternary. The most widespread evidence of aeolian activity is provided by deposits of loess, which were mainly sourced from the outwash plains of meltwater streams, including that of the Chippewa River (Schaetzl et al., 2014; Schaetzl et al., 2018; Fig. 4). Loess deposition during the Late Wisconsin in the Eau Claire region began no later than 24 ka and continued until as recently as 10 ka (Schaetzl et al., 2014). Over this interval, the dominant processes of loess transport and deposition

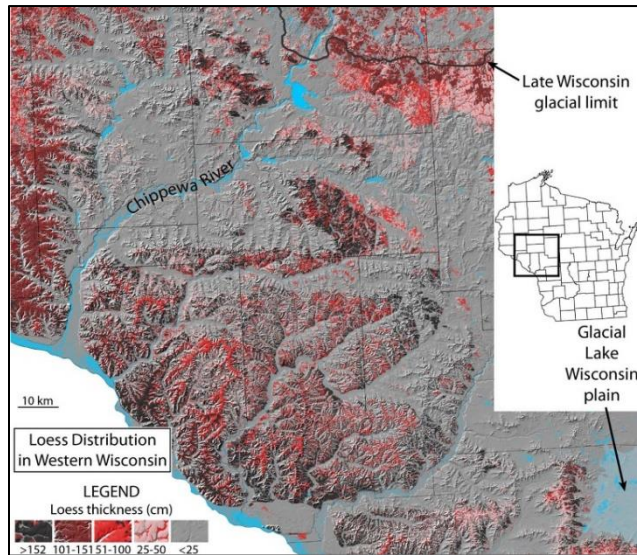


Figure 4: Extent and thickness of loess within in western Wisconsin, as derived from Natural Resources Conservation Service county soil surveys (from Schaetzl et al., 2018.)

apparently changed (Schaetzl et al., 2018). Existing evidence indicates that, early on, loess was primarily deflated from the outwash plains of the Chippewa River and its meltwater tributaries and deposited downwind of them (Schaetzl et al.,

2014). Later, strong northwest winds entrained sands from outwash and weathered sandstone. As they traveled over existing loess, saltating sands remobilized it and kept it in suspension until topographic barriers blocked further sand movement, which allowed loess to accumulate in their lee. Today, large swaths of the region are loess-free, with the thickest loess found on the southeast sides of prominent sandstone inselbergs and ridges (Schaetzl et al., 2018).

A variety of sandy aeolian landforms, such as parabolic dunes, sand sheets, sand ramps, and sand stringers, also attest to the geomorphic significance of wind in the Eau Claire region. While these landforms are generally subtle and apparent only on LiDAR-derived DEMs, they are widespread (Fig. 5). They also have a generally consistent orientation, which indicates that they were primarily formed by west-northwesterly winds. In addition, a dozen OSL ages from different landforms reveal that sandy aeolian landforms in the region were being deposited between 13 and 9 ka (Schaetzl et al., 2018; Millett, 2019; Mataitis, 2020; Shandonay et al., 2022).

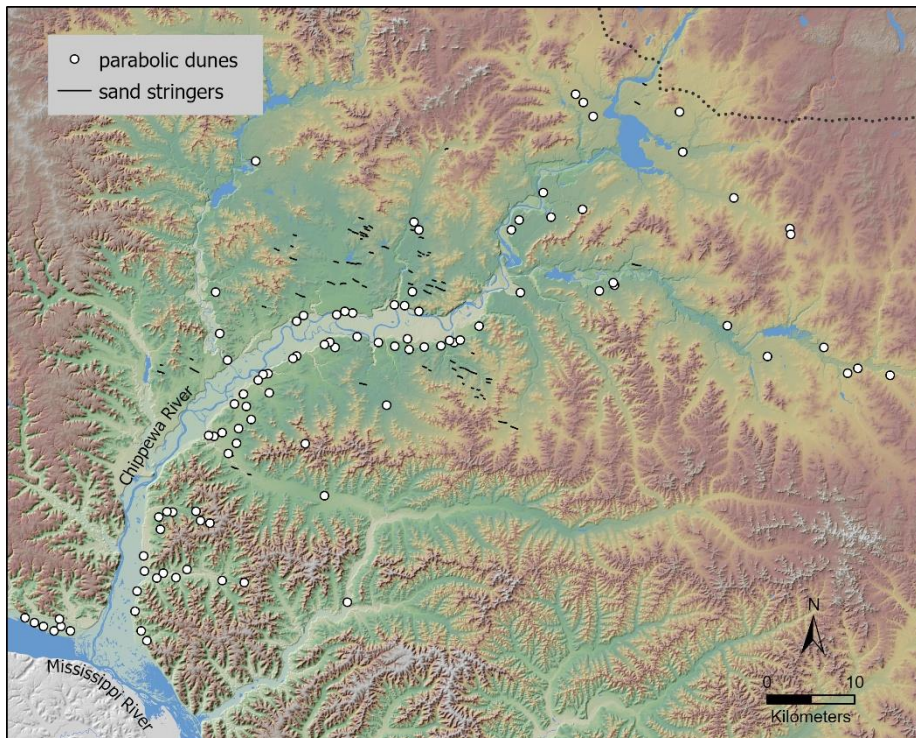


Figure 5. Parabolic dunes (Millett, 2019) and sand stringers (Schaetzl et al., 2018; Mataitis, 2020) identified from lidar-derived DEMs, aerial photographs, and soil survey data in and near the LCR valley.

Field Trip Stops

UTM coordinates are in zone 15, WGS84 datum

Stop 1: Copper Falls Glacial Deposits

UTM coordinates 623464E, 5008658N

Till of the Copper Falls Formation is reddish brown, sandy (~30 – 80% sand; Syverson, 2007; Syverson et al, 2011), and sourced from the northeast. Copper Falls till is distinguished from older River Falls till primarily by the landscape they underlie. Copper Falls sediment is found in relatively unmodified landscapes formed during the late Wisconsin. Glacial landforms (moraines, eskers, drumlins, ice-walled lake plains...) are well persevered. River Falls sediment is associated with a highly eroded landscapes and likely formed before the late Wisconsin Glaciation. This roadcut exposes typical unsorted glacial deposits of the Copper Falls Formation deposited as stagnant ice melted.

Stop 2. David R. Obey Ice Age Interpretive Center

UTM coordinates 624514E, 5009009N

The interpretive center is located in the Chippewa Moraine State Recreation Area and within the Chippewa Moraine. The moraine formed when late Wisconsin ice was at its maximum position (Syverson, 2007). The landscape here is generally described as hummocky, and formed as stagnant ice was

buried and melted. The dominant landforms in the area are ice-walled lake plains (Figs. 6 and 7; Clayton et al., 2001), such as the one the Obey Center is built upon. These form when lakes develop in the stagnant ice landscape preserved in permafrost conditions. It is likely that permafrost conditions were in northern Wisconsin until ~13.5 ka (Attig and Rawling, 2018; Batchelor, 2019). They are composed of laminated fine-grained sediment that is typically fine sand and silt. These landscapes contain organic material further south that have aided in interpreting the timing of the Lake Michigan Lobe (Curry et al., 2018); however, organic material is typically not preserved in northern Wisconsin.

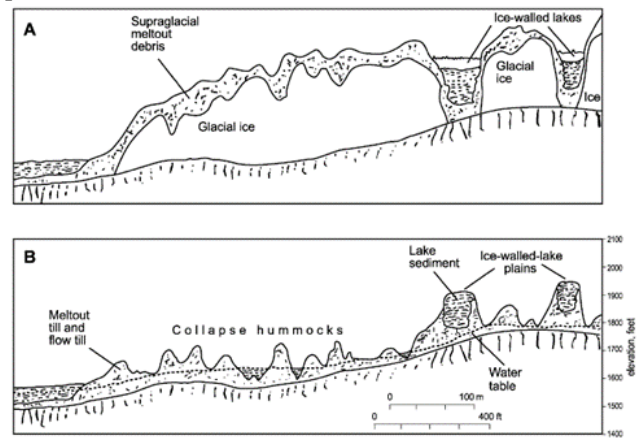
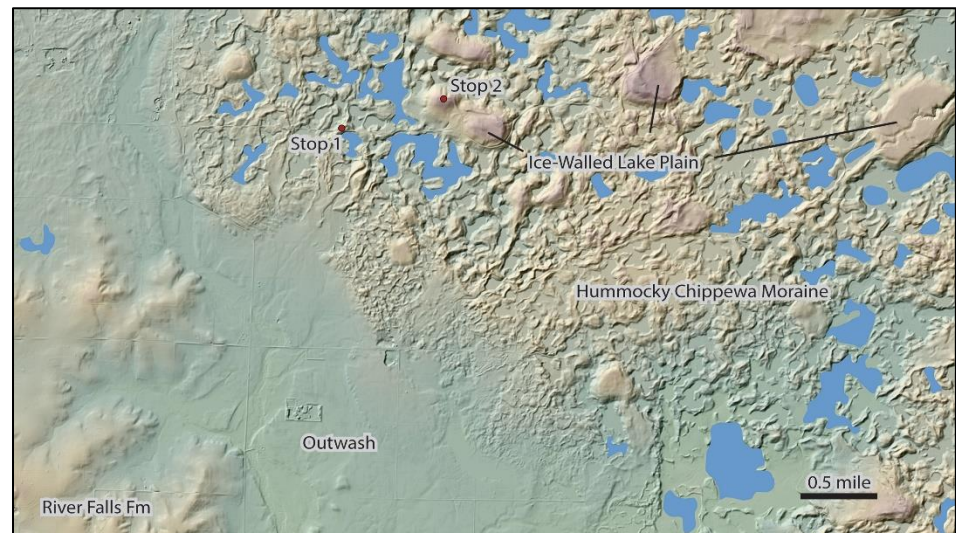


Figure 6. Schematic illustration showing the formation of ice walled lake plains (from Clayton et al., 2001). (A) Supraglacial sediment forms at the surface of stagnant ice and lakes occupies low areas where sorted sediment accumulates. (B) Hummocky topography with ice-walled lake plains remain after the ice has melted.

Figure 7. Lidar-derived DEM showing the hummocky topography of the Chippewa Moraine near Stops 1 and 2. Ice walled lake plains are abundant within the moraine, which contrasts with the flat landscape formed by meltwater streams (outwash).



Stop 3. Colluvium Exposure

UTM coordinates 625123E, 5001761 N

The landscape beyond the LIS margin was greatly affected by periglacial processes. One of the most profound effects was the stripping of hillslopes by the mass-wasting process of solifluction (Clayton et al., 2001). Evidence of solifluction is provided by relict deposits of colluvium (colluvial aprons) that mantle bedrock slopes in areas of former permafrost. Here we see a prime example of such a colluvial deposit, which consists of an unsorted mixture of sandstone clasts (pebble to boulder in size) supported in a matrix of silty sand. The sand and the sandstone clasts were likely derived from the underlying bedrock (sandstone of the Cambrian Eau Claire Formation) by intense freeze-thaw weathering during the Late Wisconsinan. The silty material probably is loess that winds deflated and transported to the sight from nearby outwash plains.

Stop 4. Jim Falls

UTM coordinates 635875E, 4990538 N

The Chippewa at Jim Falls is an example of a superimposed river (Fig. 8). Here, the Chippewa River incised through a cover of outwash and till and encountered a topographic high in the buried a bedrock landscape. It did not, however, incise to its present level in one episode of downcutting, as evidenced by two terraces that are apparent at and near this site. The highest terrace grades to the Wissota terrace, the maximum level of aggradation of the lower Chippewa River. This indicates that the river continued to flow at this level after the glacial margin had retreated to the north of this site. Remnants of a terrace approximately 3 meters below the Wissota (best seen downstream from the east end of the pedestrian bridge), which is cut into till, indicate a period of channel stability before the river incised further to the buried bedrock. When deeper incision occurred is unknown, although OSL ages of terrace alluvium farther down valley

suggest that it happened during the late Holocene (Fig. 3).

The falls consist of a series of small knickpoints formed in early Proterozoic bedrock consisting of banded amphibolites with granitic intrusions (Myers and Maercklein, 1978). The angular form of the knickpoints, along with angular boulders of the same lithology scattered along the channel bed, suggest that the river is incising here primarily by mean of hydraulic plucking. Hydraulic plucking occurs when flows are deep and fast, leading to a zone of flow separation and low pressure on the downstream face of knickpoints. If the bedrock of a knickpoint is sufficiently jointed and weathered, the resulting drag force will pull blocks away from the knickpoint face. Polished rock surfaces with rare grooves and potholes indicate that abrasion by bedload sediment is also playing a role here in channel incision, although its effects appear secondary to that of plucking.

While the Chippewa has clearly incised into the Jim Falls bedrock, incision overall has been minimal. The lack of incision is likely due, in part, to the weathering and erosion-resistant nature of the bedrock. An additional factor is the relatively short amount of time that the river has been incising at this location. Given the model of long-profile evolution in Figure 3, incision didn't start here until sometime after 4.7 ka.

The site today clearly is highly modified by a dam, appropriately named the Jim Falls Dam. The original Jim Falls Dam was built in 1923 to utilize the hydraulic head provided by the falls to generate electricity. It was designed so that the falls were bypassed and left dry except during high flows, when excess water was released through a spillway that was located at the falls upstream end. The dam was redeveloped in 1988 and now has the highest generation capacity of any hydropower dam in Wisconsin (~60 MW). This redevelopment included moving the main spillway from the head of the falls to a location adjacent to a new main powerhouse. It also included constructing a smaller spillway and

auxiliary powerhouse where the main spillway had been. This was done so that a minimum flow of 240 cfs could be released down the bypassed reach year-round, except for the period April 1-

May 31, when flow through the reach is increased to 850 cfs to enhance fish spawning habitat.

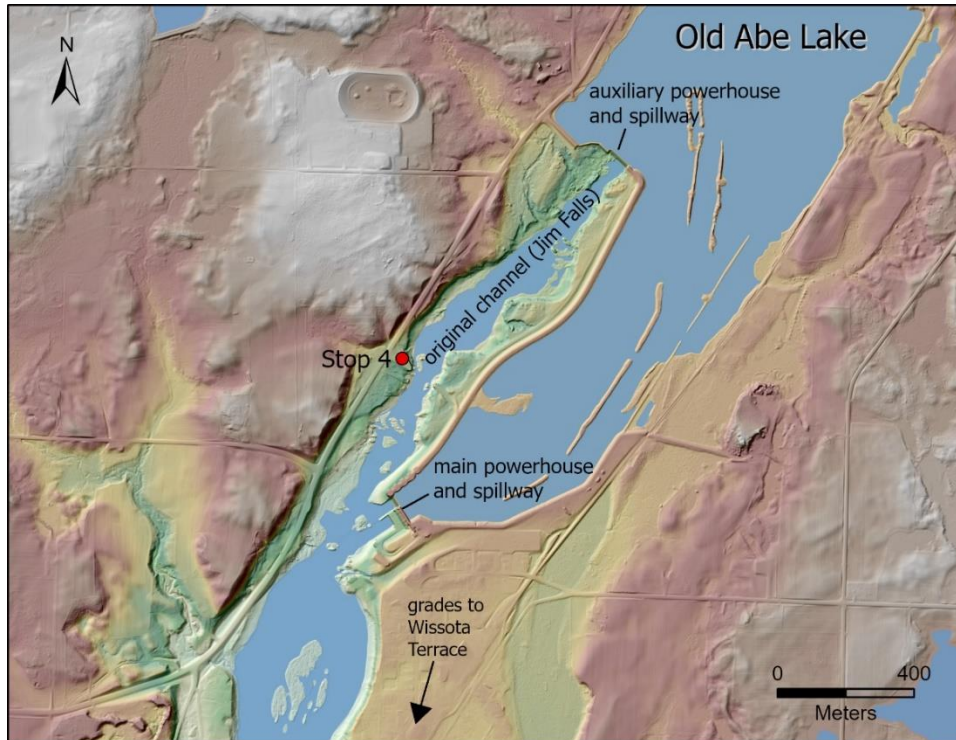


Figure 8. Lidar-derived DEM of Jim Falls and surrounding area.

Stop 5. Wildenberg Quarry

UTM coordinates 616648 E, 4970372 N

Note: This is private property. No access is allowed without owner's permission.

Glacial sediment of the River Falls Formation includes till that is lithologically like the Copper Falls Formation and melt-water stream sediment (Syverson, 2007; Severson et al., 2011). These can be distinguished from the Copper Falls formation because they occur at the top of highly eroded landscapes and the soils in them are more developed. There are no primary glacial landforms associated with the River Falls Formation, likely due to intense modification by periglacial processes in permafrost conditions during the late Wisconsinan.

Evidence for permafrost during the Late Wisconsinan is widespread in Wisconsin, with a hypothesized permafrost interval in central Wisconsin from ca. 33 to 15 ka (Batchelor et al., 2019) and as late as 13.5 ka in northern Wisconsin (Attig and Rawling, 2018). This interval is, however, poorly constrained in the Eau Claire region due to a lack of ¹⁴C datable materials in features diagnostic of permafrost. OSL dating now makes it possible to date proxy geomorphic features, like ice-wedge pseudomorphs, to help constrain this (Schaetzl et al., 2021).

The ice-wedge pseudomorphs in this quarry are sand wedges (Fig. 9). Sand wedges such as these form in periglacial settings when thermal contraction of frozen ground in winter forms

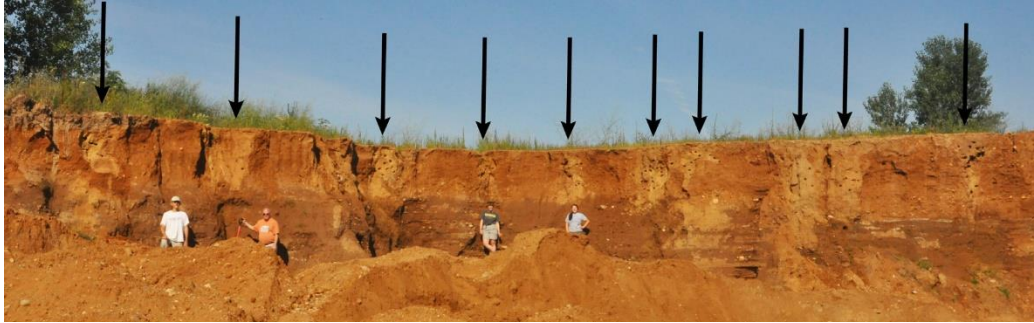


Figure 9. Sand wedges exposed in the headwall of the Wildenberg quarry in July 2018. (Photograph by Randy Schaeztl.)

cracks in the soil. If this happens in a cold, dry, wind-swept environment with sand available for transport, sand will blow into and fill the cracks. Over time, repeated cracking and filling will form vertical structures that generally taper with depth. OSL dating of sand wedges in this quarry and in another quarry located 60 km to the south indicate that thermal contraction cracks existed and were filling with sand from no later than 19.3 ka until 14.7 ka. Schaeztl et al. (2021) interpret these ages as documenting when permafrost in the region most likely ended. Interestingly, the OSL ages from this quarry (15.1 and 14.7 ka) are younger than those from the quarry 60 km to the south (19.3, 19.1, and 18.3 ka). These may represent a time-transgressive spatial relationship in that permafrost possibly degraded earlier at the more southerly location and remained longer at the more northerly one, although this is purely speculative given the large errors on the OSL ages (1.4 to 2.2 ka at 1σ). That said, the larger and more complex morphologies of the sand-wedges found at this quarry do suggest the possibility of more intense sand-wedge development due to more prolonged permafrost conditions.

Stop 6. High Bridge

UTM coordinates 617832E, 4964561N

This stop is on the so-called High Bridge in the city of Eau Claire. Standing 26 m above the Chippewa River, the High Bridge was built in 1880 as a railroad bridge and was an innovative bridge for its time. It was abandoned in 1992, and the City acquired ownership in 2007. In 2015, the

City incorporated it into its bicycle-pedestrian trail system.

In addition to being historically significant, the High Bridge (Fig. 10) affords an excellent view of many of the terrace levels found in the LCR valley. The High Bridge itself is at the level of T-6. To the east, trees and houses can be seen at the top of the Wissota terrace scarp, which is 6-7 meters above T-6. The Wissota can also be seen to the west where the pedestrian-bicycle trail cuts into its scarp. Looking downstream, lower terraces are difficult to discern from this vantage point, although the residential and business districts located near the river provide clues. These built-up areas are all above the 100-year flood level. That is, they all are on terraces. In Eau Claire, there is little active floodplain. This suggests that incision below the lowest terrace level occurred here recently (within the last 2.3 ka according to the model of long-profile evolution in Fig. 3).

Upstream from the High Bridge is the Dells Dam. This dam, which was built in 1924 for the purpose of generating electricity, is situated at an unusually good site for a dam on the LCR – a bedrock gorge. This gorge was formed when the river incised into a cover of glacial outwash that buried a low ridge of Cambrian sandstone (Mt. Simon Formation) connecting Mt. Simon (the conical bedrock hill located approximately 500 m northeast of the dam) to the bedrock uplands located west-northwest of it. In other words, the river here did not incise into its pre-glacial bedrock valley, which is located east of Mt.

Simon and the High Bridge. This is an example of an epigenetic gorge (Ouimet et al., 2008), and the river likely carved it sometime after 7.4 ka

(see Fig. 3). Incision is actually still occurring here, as evidenced by a 1.5-m bedrock knickpoint located 180 m downstream of the bridge.

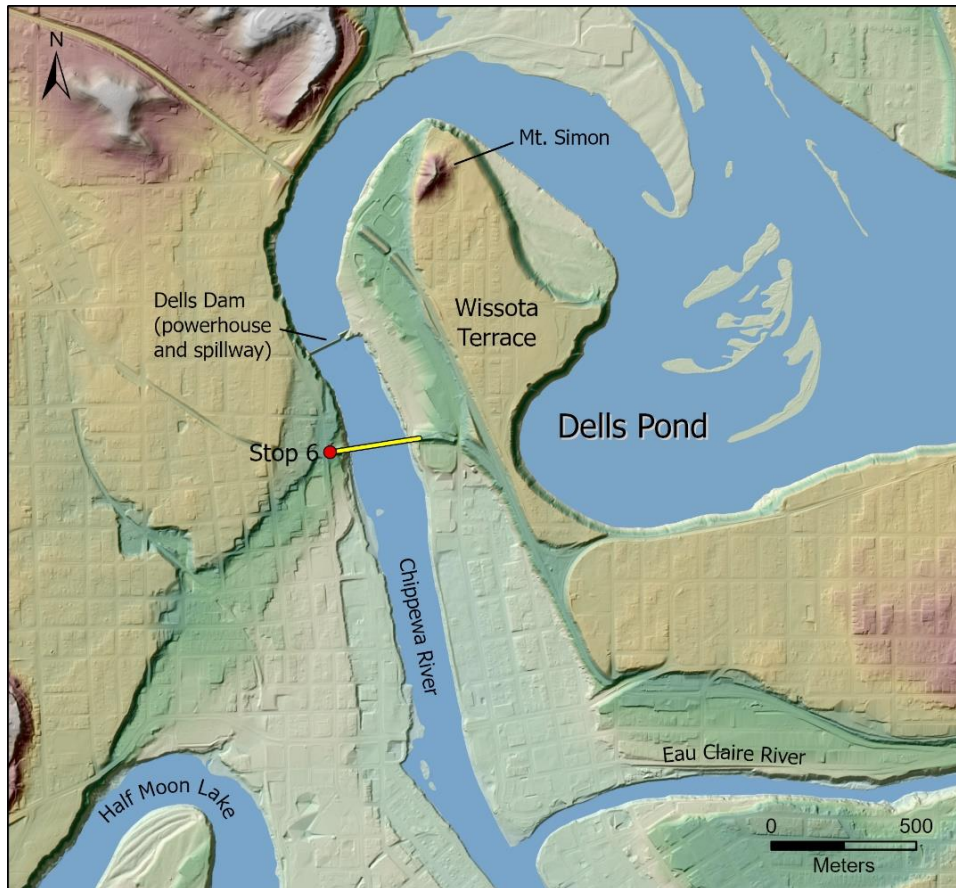


Figure 10. Lidar-derived DEM showing the stream terraces found in Eau Claire in the vicinity of the High Bridge.

Stop 7. Sand Hill Cemetery

UTM coordinates 599549E, 4958254N

This stop is at the edge of the Wissota terrace tread and top of the Wissota terrace scarp. To the north, a braided channel is apparent in the subtle rolling topography of the Wissota tread, providing evidence of the Chippewa's glacial past (Fig. 11). To the south, the Wissota scarp descends nearly 30 m to the lowest terrace in the valley (T-1). Looking upvalley, this terrace can be identified by the farmland situated on it. Low-lying land that is wooded is either floodplain or paleochannels cutting across the T-1 surface.

Incision here below the Wissota level occurred between 10 and 9 ka, based on four OSL dates of Wissota fill obtained from sites located 6.5 km

upvalley and on the valley's other side. The last episode of incision, below T-1, occurred within the last 2.3 ka based on an OSL date of T-1 fill from a site also located 6.5 km upvalley (Fig. 3). In the downstream direction, the valley below the Wissota is predominantly floodplain with only rare lower terrace remnants. In addition, the river's planform switches from a single-thread meandering shape to a multi-thread anabranching one that extends downvalley for a distance of 8.5 km. At that point, it returns to a single-channel meandering form. It is uncertain why this anabranching reach exists, although its similarity to the sedimentation zones of wandering gravel bed streams in British Columbia described by Desloges and Church (1987) suggests a cause. Given its location immediately downstream of

the reach that was recently incised below T-1, it could be the result of sedimentation resulting from that incision event (Fig. 12). A pronounced

convexity in the long profile of the modern river along the anabranching reach supports this hypothesis (Faulkner et al., 2016).

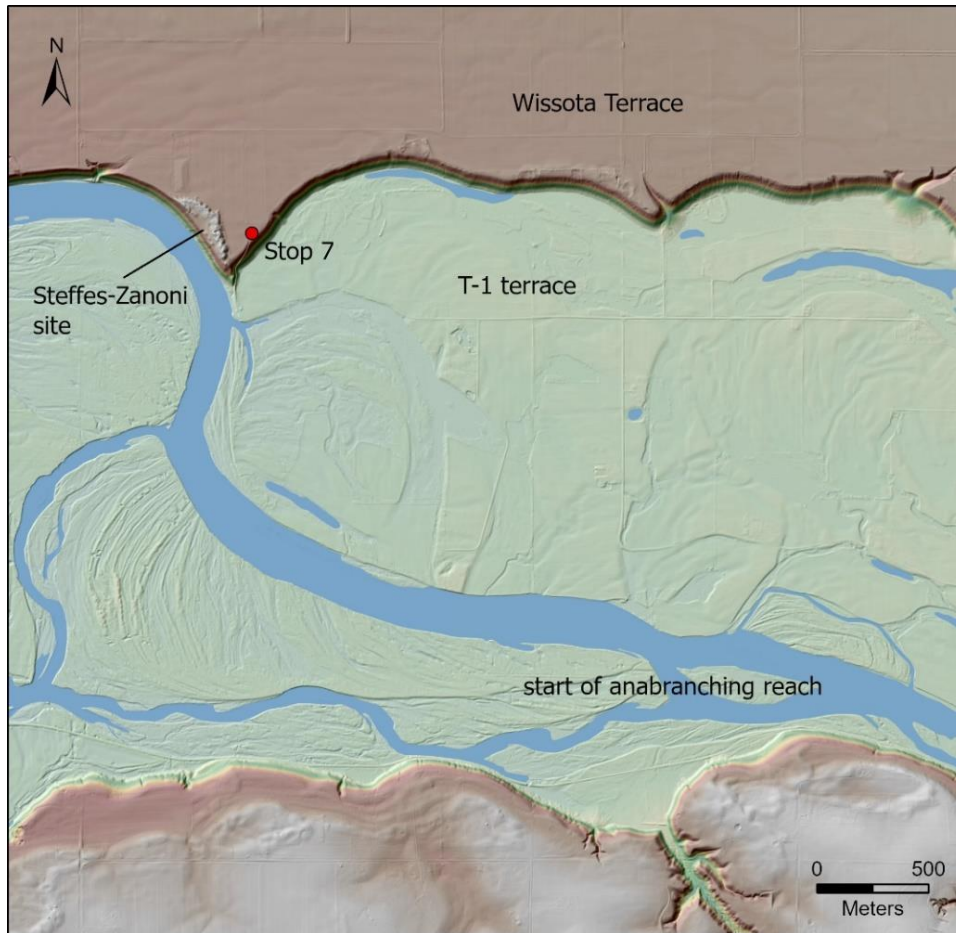
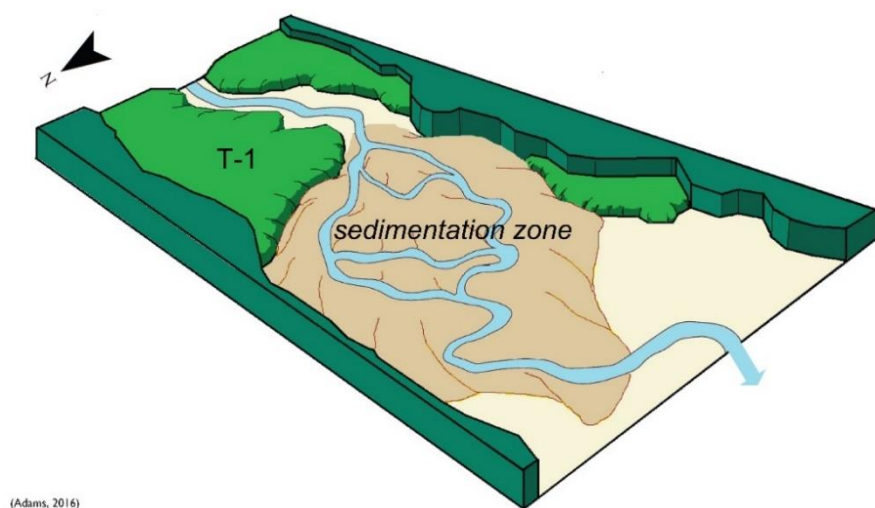


Figure 11. Lidar-derived DEM showing the fluvial and aeolian landforms in the vicinity of the Sand Hill Cemetery (Stop 7), which are discussed in the text.

Figure 12. Cartoon showing the setting of the anabranching reach (and hypothesized sedimentation zone) downstream of the reach that is incised below the T-1 terrace (adapted from Adams et al., 2016).



(Adams, 2016)

A short distance (~200 m) west of the Sand Hill Cemetery, an elongate wooded hill—informally named the Steffes-Zanoni site—can be seen rising above the low-relief landscape of the Wissota terrace. (This hill can also be seen in Fig. 11.) Its summit is approximately 8 m higher than the surrounding Wissota terrace surface, and its elongate form runs parallel to the Wissota terrace scarp. The landform is composed of sand that is mineralogically indistinguishable from the terrace sediments beneath it, although it is relatively finer-grained and more well-sorted, indicating that it is aeolian sand (Millett, 2019). It is also morphologically complex, consisting of multiple parabolic forms coalesced together with smaller parabolic forms on top of them. These forms—smaller parabolic dunes superimposed on larger older ones—indicate repeated aeolian activity at this site. The OSL ages of samples obtained from depths of 1.7 m and 2.5 m near the hill's northwest end suggest a depositional age of the landform's upper part to be ~0.5 ka. Larson et al. (2008) identified a dune similar to this one in the city of Eau Claire and called it a cliff-top parabolic dune (because of its form and its proximal relationship to the Wissota terrace scarp). Since then, many cliff-top parabolic dunes have been noted in the Eau Claire region (note the large number of parabolic dunes situated along the Wissota scarp in Fig. 5). Why these dunes exist will be discussed at our final stop.

Stop 8. Town of Union Conservancy

UTM coordinates 607504E, 4959523N

Note: This site involves walking on unpaved trails for an approximate distance of one mile. There are some short sections of the trail that are moderately steep.

Parabolic dunes in cliff-top positions are especially prominent at this location, which is informally called the Kiwanis Site (Fig. 13). Like those seen at the previous stop, these dunes are situated atop the Wissota terrace scarp. These dunes and others that are similarly situated in the LCR valley—above high Wissota terrace scarps

with parabolic forms oriented perpendicular to them—are hypothesized to have had a similar genetic origin (Larson et al., 2008; Millett, 2019). In their proposed model, a river cuts into the base of a high fill-terrace scarp. This creates an unstable cutbank and promotes mass wasting that removes vegetation from the scarp face. Wind that then blows against such a scarp gets compressed, which causes its velocity to go up. The increase in velocity enhances the wind's ability to entrain exposed sandy sediment and transport it up the scarp face. When this happens, the sandy sediment ultimately settles out at the top of the scarp as wind velocity is reduced there. This leaves behind “cliff-top dunes.” (Fig. 14).

Given this model of dune genesis, one should expect cliff-top dunes to have different orientations and depositional ages compared to other parabolic dunes not in cliff-top positions. This is indeed the case. Non-cliff-top dunes generally have a northwest-southeast orientation and depositional ages older than 10 ka. Cliff-top dunes display a variety of orientations (perpendicular to their scarps) and are generally much younger. OSL ages from a cliff-top dune in the city of Eau Claire indicate a period of aeolian deposition at ca. 6.0 ka, while two from the Steffes-Zanoni site (Stop 7) indicate that deposition in the uppermost dune sediments occurred at ca. 0.5 ka. At the Kiwanis site, eight OSL ages point to two depositional episodes: at ca. 0.9 ka and 0.5 ka.

The model of Larson et al. (2008) of cliff-top dune genesis suggests that these should be forming wherever the Chippewa River is eroding laterally into Wissota terrace fill. This, however, is not the case; today, all cliff-top dunes in the LCR valley are stabilized by vegetation and no longer moving. Thus, the genesis of these landforms is doubtless more complex than the model of Larson et al. (2008) suggests, with climatic variability likely playing a key role in the process of aeolian sedimentation and dune formation at cliff-top locations (Millett, 2019). For example, during humid climate intervals

(such as at the present), high cutbanks carved by the river into Wissota fill are colonized readily by vegetation, which inhibits the entrainment and transport of sand up terrace scarps. In contrast, during arid intervals, vegetation cover is greatly reduced, especially on steep well-drained terrace scarps, allowing wind to entrain and transport sand up them. OSL ages from the Kiwanis Site and the Steffes-Zanoni Site support the significance of climate variability in the formation of cliff-top dune. At Kiwanis, these ages indicate two pulses of aeolian deposition – the first at ca. 0.9 ka and a second at ca. 0.5 ka, with the latter coinciding with ages from the Steffes-Zanoni Site. If correct, these pulses happened during the Medieval Climatic Anomaly when well documented dry periods affected the mid-continent of North America (reviewed in Millett, 2019).

Lastly, closer examination of the Kiwanis Site (Fig. 13) reveals landforms in close proximity to the dunes that do not look to be of natural origin. Several hemispherical (or conical) mound-like

forms enclosed by a subtle linear ridge appear to be anthropogenic features. It is likely, based on the morphology and distribution of these features, that their creation was tied to the genesis of the prominent aeolian dunes found nearby and, potentially, culturally linked to a well-established late Woodland period of mound building in the upper Mississippi River valley – a period which would have coincided with the formation of the Kiwanis and Steffes-Zanoni dunes. Based on the configuration of these features, it is hypothesized that this site represents the “Thunderer,” an effigy of a bird-like deity, or sky being. The spotted Thunderer, with “spots” represented by the location of the mounds within the linear structure, is associated with the West and the bringer of storms. If true, Native peoples may have watched the dunes form during a period of more aridity during episodes of higher winds/storms, resulting in this site being spiritually significant at that time and now an important site of cultural heritage (R. Schirmer, personal communication).

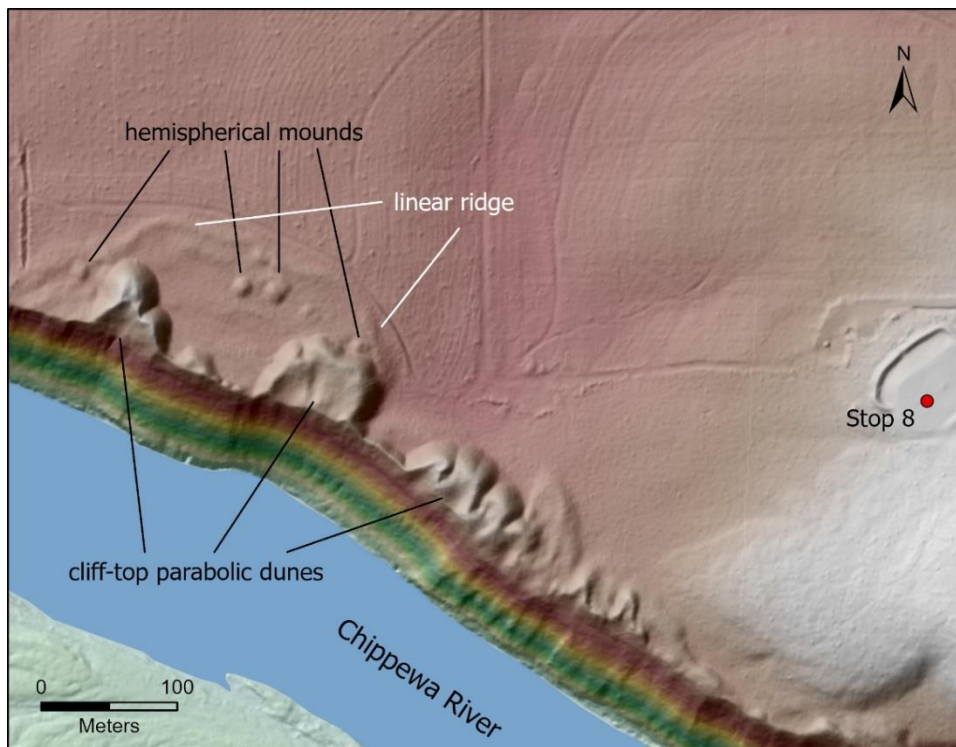


Figure 13. The Kiwanis Site. Cliff-top parabolic dunes are situated directly above the Chippewa River on top of the Wissota terrace scarp. Also note the hemispherical mounds and linear ridge of hypothesized anthropogenic origin.

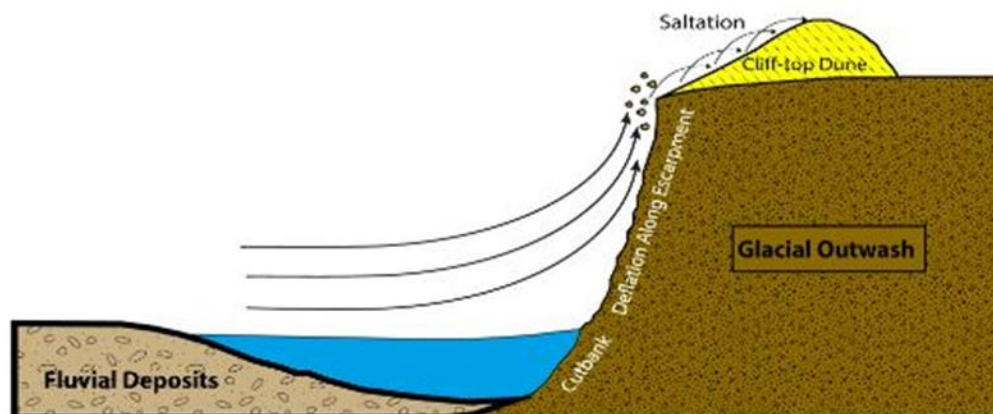


Figure 14. Model of cliff-top dune genesis induced by lateral river erosion (adapted from Larson et al., 2008).

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