

57TH ANNUAL MEETING



INSTITUTE ON LAKE SUPERIOR GEOLOGY

ASHLAND, WISCONSIN MAY 18-21, 2011



PROCEEDINGS VOLUME 57

PART 2 – FIELD TRIP GUIDEBOOK



INSTITUTE ON LAKE SUPERIOR GEOLOGY

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ASHLAND, WISCONSIN

HOSTED BY:

NORTHLAND COLLEGE

TOM FITZ

Chair

Proceedings Volume 57

Part 2 – Field Trip Guidebook

Edited by: Tom Fitz, Allison Mills, Kristi Wilson, Cassandra Bodette, and Drew Cramer

57TH INSTITUTE ON LAKE SUPERIOR GEOLOGY

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Reference to material in Part 2 should follow the example below:

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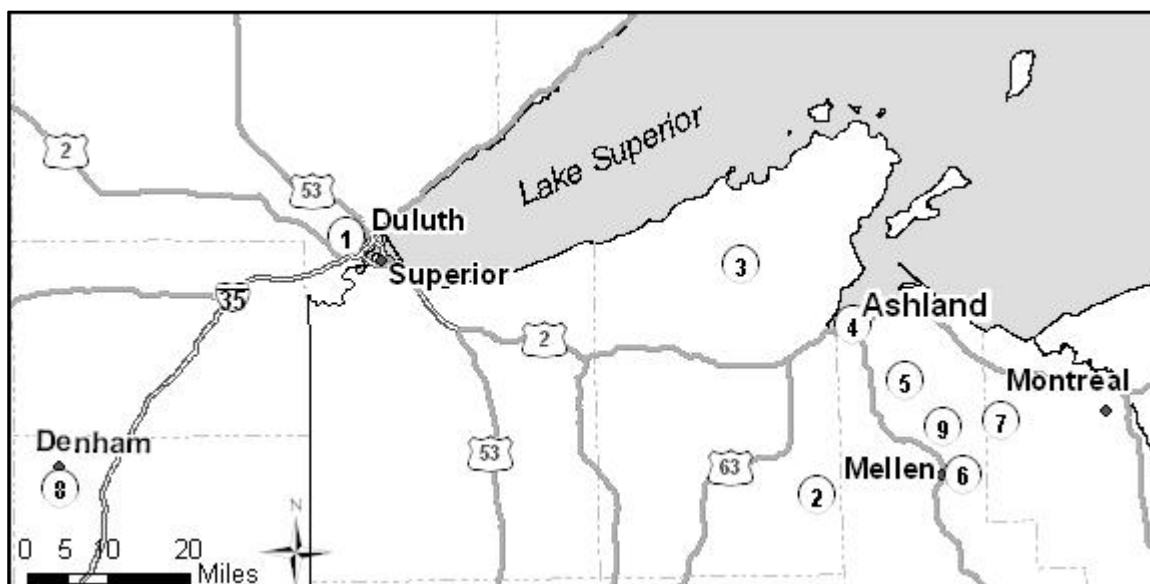
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Locations of ILSG 2011 field trips. Numbers on the map correspond to field trip numbers.

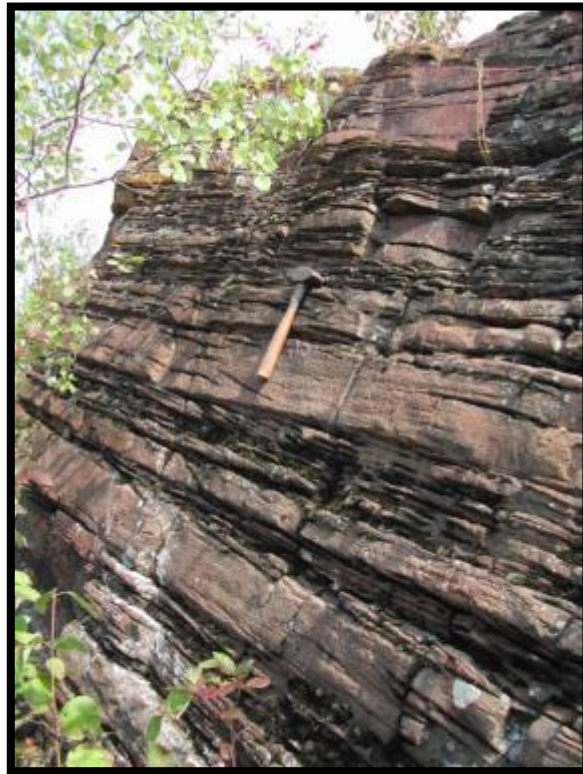
The lead editor, Tom Fitz, would like to thank the field trip leaders, the reviewers, and the dedicated team of editors who have contributed their time and effort to creating this field trip guidebook. Their efforts are greatly appreciated by everyone who will use this book. We hope the guidebook will help foster understanding and appreciation of the remarkable geology the Lake Superior region.



57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 1

IGNEOUS STRATIGRAPHY OF THE LAYERED SERIES AT DULUTH – TYPE INTRUSION OF THE DULUTH COMPLEX



Rocks of the Duluth Layered Series
-J. Miller

Field Trip 1

Igneous Stratigraphy of the Layered Series at Duluth – Type Intrusion of the Duluth Complex

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This trip guide is modified from a guidebook entitled Field Guide to the Geology and Mineralization of Mafic Layered Intrusions of the Duluth and Beaver Bay Complexes, Northeastern Minnesota (Miller and Severson, 2009). The guidebook was developed for a field trip associated with the Precambrian Research Center Professional Workshop on Field, Petrographic and Mineralization Characteristics of Mafic Layered Intrusions held October 4-10, 2009

GEOLOGY AND PGE MINERALIZATION OF THE LAYERED SERIES AT DULUTH

The well-exposed gabbroic rocks forming the escarpment above the city of Duluth have long been recognized as the type section of the Duluth Complex. While early surveys recognized the presence of two distinct rock types in the Duluth area (Winchell, 1899), Grout was the first to interpret the layered gabbros as a product of convection and magma differentiation (Grout, 1918a-e). Taylor (1964) produced the first detailed-scale (1:24,000) geologic map of the complex in the Duluth area and defined the main distinctions between the layered and anorthositic series. Moreover, based on field, petrographic, and very limited geochemical data, Taylor recognized basic similarities between the layered series at Duluth and the Skaergaard intrusion, which Wager and Deer (1939) had established as the classic example of a mafic intrusion formed by closed-system fractional crystallization of a tholeiitic magma. More recently, detailed mapping in the Duluth area has delineated much more about the structure and cumulate stratigraphy of the layered series (Miller and Green, 2008a & b; Green and Miller, 2008). The petrology of the layered series was summarized by Miller and Ripley (1996) and Miller and Severson (2002).

The Layered Series at Duluth (DLS) is a well-differentiated, 3- to 4.5-km-thick, east-dipping sheet-like mafic layered intrusion that is the southernmost intrusion of the Duluth Complex. Exposure is very good along the 600'-high escarpment above Lake Superior and the St. Louis River estuary, but is spotty inland. Aeromagnetic data show that the DLS has a north-south strike length of about 60 kilometers long and eventually pinches out into the Boulder Lake intrusion. The hanging wall of the DLS consists of olivine gabbroic and troctolitic anorthosite of the anorthositic series, which according to U-Pb dating was emplaced just prior to DLS intrusion at 1099 Ma (Paces and Miller, 1993). The footwall of

the DLS at its southern end is reversed polarity lavas of the Ely's Peak basalts (Fig. 1-1). Layering and foliation in the DLS, which dip 20°-40° to the east, and modeling of aeromagnetic and gravity data indicate that the basal contact is not conformable with shallow-dipping (~15°) Keweenaw basalt flows in the footwall, but instead dips at a greater than 35° angle (Miller and Green, 2008b). Aeromagnetic data further imply that the basalt is cut out by the DLS to the north, whereupon the footwall becomes graywacke and slates of the Thomson Formation.

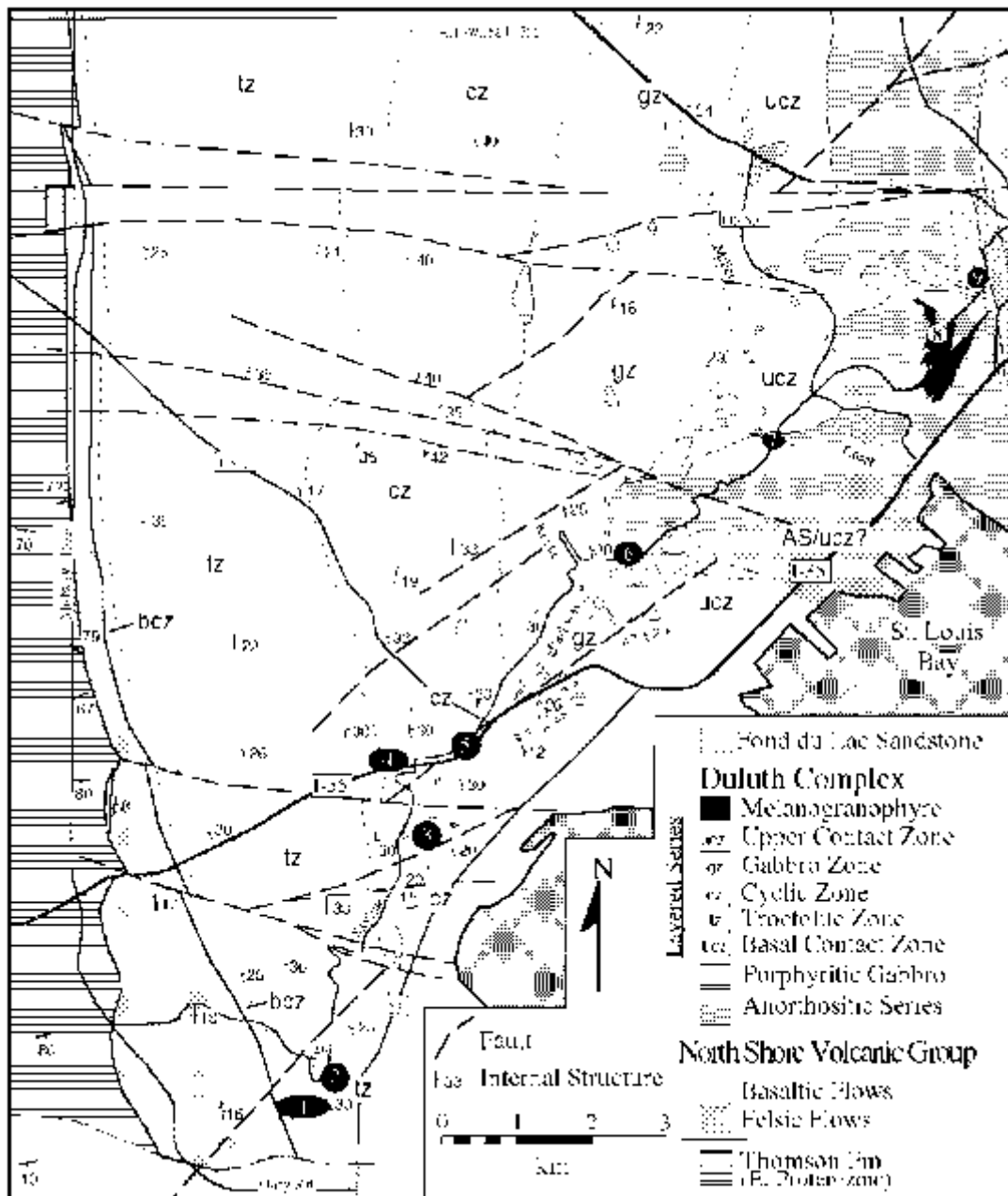


Figure 1-1. Geology of the Duluth Complex at Duluth showing the field trip stops.

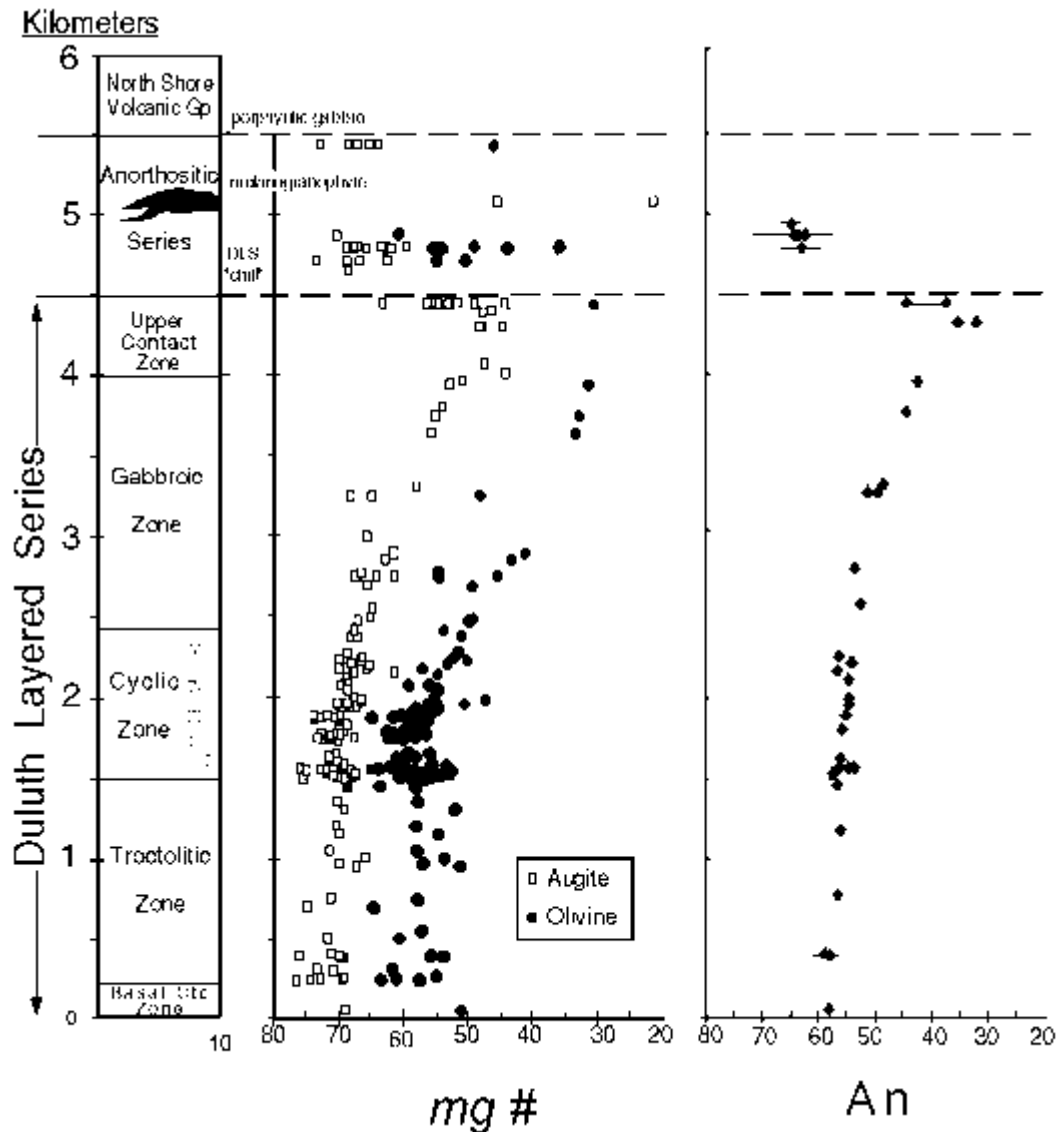
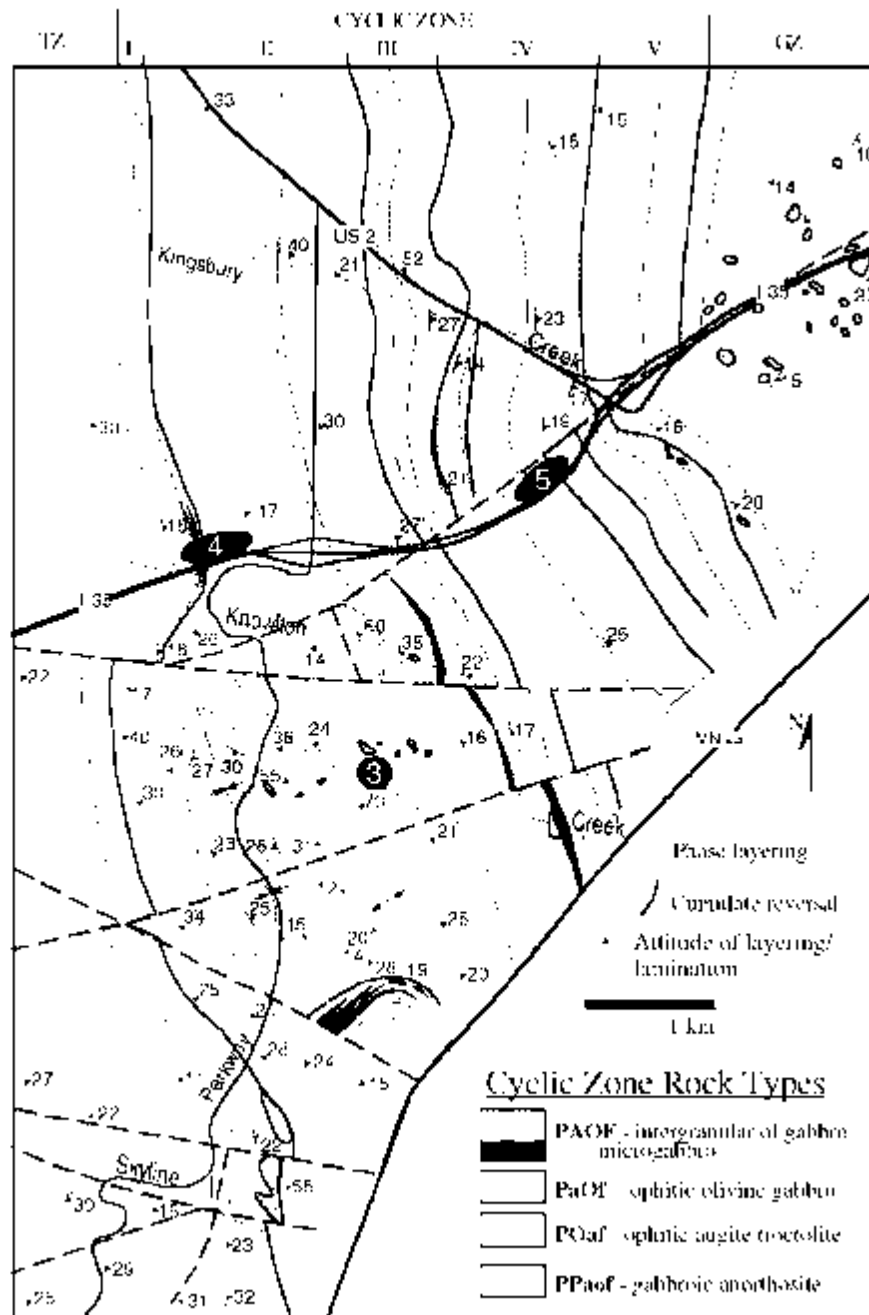


Figure 1-2. Igneous stratigraphy and cryptic variation of olivine, pyroxene and plagioclase composition through the Layered Series at Duluth. mg# values of olivine and pyroxene and An content of plagioclase are based on multiple microprobe analyses (error bars indicate one standard deviation). From Miller and Sevenson, 2002, Figure 6.8. Note that the cumulate codes for augite are A/a rather than the C/c (clinopyroxene) used in the text.

The DLS is divided into five major zones based on dominant rock type (Fig. 1-1). The basal contact zone is composed of coarse-grained, taxitic olivine gabbro and augite troctolite (Stop 1). The lowermost cumulate sequence is the troctolite zone (Stops 1, 2A & 2B). It is 1 to 2 kilometers thick, consists mostly of homogeneous foliated troctolitic (PO) cumulates, and locally displays modal and textural layering, especially in its lower third. The cyclic zone (Stops 3, 4, & 5) forms the medial section of the DLS and is characterized by cyclical variations in cumulus mineralogy between troctolitic and gabbroic (PCFO) cumulates (Fig. 1-4). The persistent occurrence of gabbroic cumulates defines the gabbro zone. Gabbroic cumulates, in turn, grade upward into unlayered (noncumulate) apatitic

Figure 1-3. Geology of the Cyclic Zone of the DLS showing field stops 1-3 to 1-5. Note that the cumulate codes for augite are A/a rather than the C/c (clinopyroxene) used in the text.



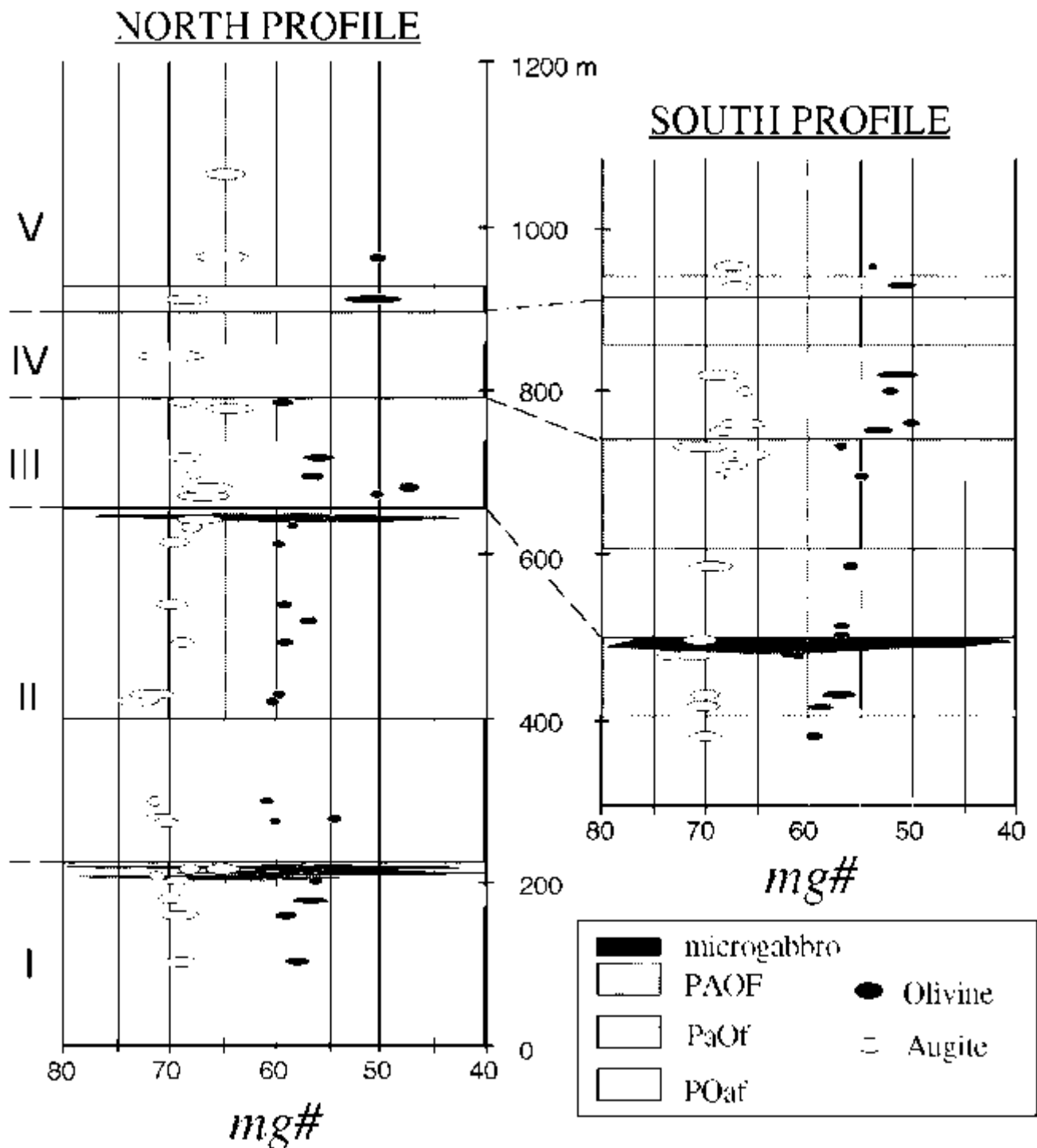


Figure 1-4. Cryptic variation of Fo and En in outcrop samples from the cyclic zone taken along two profiles north and south of Interstate 35 (Fig. 1-3). Note that the cumulate codes for augite are A/a rather than the C/c (clinopyroxene) used in the text.

quartz DLS ferromonzodiorite, which composes most of the upper contact zone (Stop 6). This quartz ferromonzodiorite complexly mixes with a fine-grained biotitic ilmenite ferrodiorite, which ultimately forms a "chilled" contact with anorthositic series rocks (Stop 7). A body of melanogranophyre that irregularly cuts through the anorthositic series probably represents the uppermost differentiate of the DLS (Stop 8). This igneous stratigraphy is complimented by cryptic layering of cumulus mineral compositions (Fig. 1-2) and together these imply the generally formed by bottom-up fractional crystallization of a moderately evolved, olivine tholeiitic parent magma.

Although the DLS is overall, a well-differentiated intrusion, the repeated progression from troctolitic to gabbroic cumulates in the cyclic zone indicates that it did not fractionally crystallize as a closed system. The cyclic zone consists of at least five macrocycles (each 50-200 m thick) within which troctolitic cumulates grade upward to gabbroic cumulates (Figs. 1-3, 1-4). Macrocycle boundaries are marked by the abrupt regression in the cumulus mineralogy from gabbroic back to troctolitic cumulates. The gabbroic parts of the macrocycles commonly contain inclusions of anorthositic series rocks and the very uppermost parts of the macrocycles locally have discontinuous layers of fine-grained gabbroic adcumulate (microgabbro). This cyclicity in phase layering does not correspond to a complimentary cryptic variation in mineral chemistry (Fig. 1-4). Based on these characteristics, Miller and Ripley (1996) suggested that the macrocyclic phase layering is predominantly related to devolatilization and decompression that attended magma venting events from a shallow (< 5 km) chamber, with magma recharge possibly having played a secondary role.

Geochemical data were acquired from 83 hand samples that profile the general stratigraphy of the DLS. The sampling focused on specific horizons, particularly in the cyclic zone, that contain visible sulfide mineralization or appear to represent perturbations to the magmatic system (Fig. 1-5, 1-6). Most of the samples have Pt+Pd concentrations below 15 ppb; 14 samples have concentrations between 15 and 150 ppb and five samples have concentrations >150 ppb. Whole rock S and Cu concentrations (Fig. 1-5) imply that the DLS magma 1) was consistently sulfide-undersaturated during troctolite-zone crystallization; 2) achieved intermittent saturation during crystallization of the cyclic zone; and 3) was fairly consistently saturated as the gabbro zone accumulated. In contrast to the singular large increase in Cu abundance and Cu/Pd ratio observed in the Sonju Lake intrusion (See Day 3 description), the DLS shows erratic variability in these parameters (Fig. 1-5). Such variability probably reflects the openness of the DLS system to magmatic recharge. Nevertheless, a two to three order-of-magnitude range in the Cu/Pd ratio through the cyclic zone indicates some level of efficiency in the ability of sulfide melt to extract PGE from silicate magma (e.g., Maier and others, 1996).

Six hand samples that contain the most enriched Pt+Pd and sulfide concentrations come from the upper portions of macrocycles I and II (Figs. 1-5 & 1-6) and the two highest values are associated with gabbro-microgabbro interfaces. The model of magma venting to explain the phase layering of the cyclic zone (Miller and Ripley, 1996) also can explain the elevated concentrations of sulfide and PGE at cyclic zone boundaries of cumulus regression. Experimental data at low pressures (1-2 kb) suggest a positive correlation between pressure and sulfide solubility in hydrous tholeiitic magmas (Carroll and Rutherford, 1985). If valid, this implies that magma decompression due to venting from shallow differentiated magma chambers may trigger sulfide saturation (or oversaturation in a saturated magma).

Devolatilization resulting from venting of a volatile-rich magma would also cause an abrupt increase in fO_2 , thereby having a compounding effect on reducing sulfur solubility (Poulson and Ohmoto, 1990). The attractiveness of a magma venting as a trigger for sulfide saturation in terms of PGE enrichment is that it would promote chamber-wide sulfide segregation if the system was well mixed, or a rain of sulfide out of the roof zone if the magma chamber was compositionally zoned. Both situations would promote high R factors, though the second scenario would require enough overproduction of sulfide so as to survive the descent to the cumulate floor.

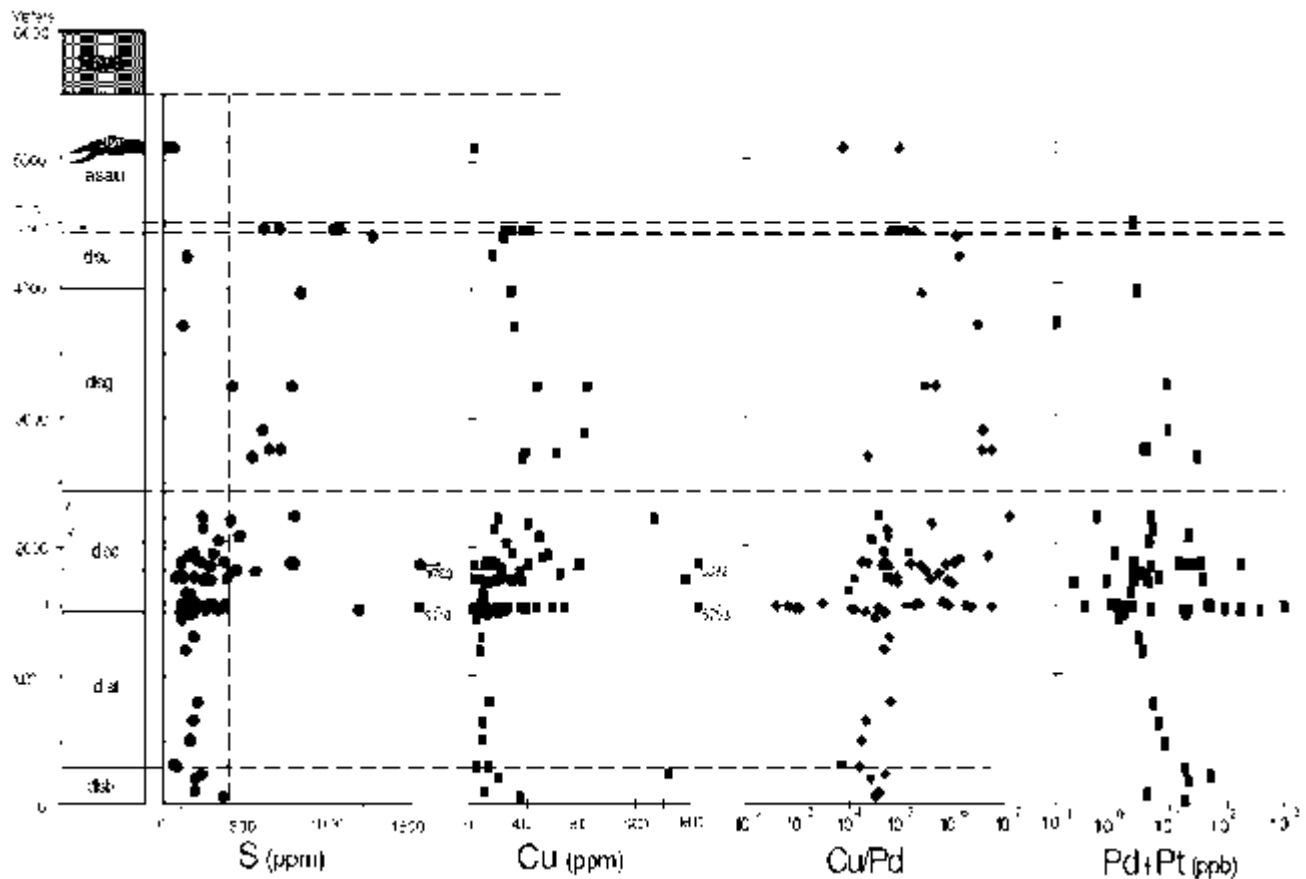


Figure 1-5. Stratigraphic variation in sulfur, copper, copper/palladium, and palladium + platinum through the Layered Series at Duluth (Fig. 8.11, RI-58).

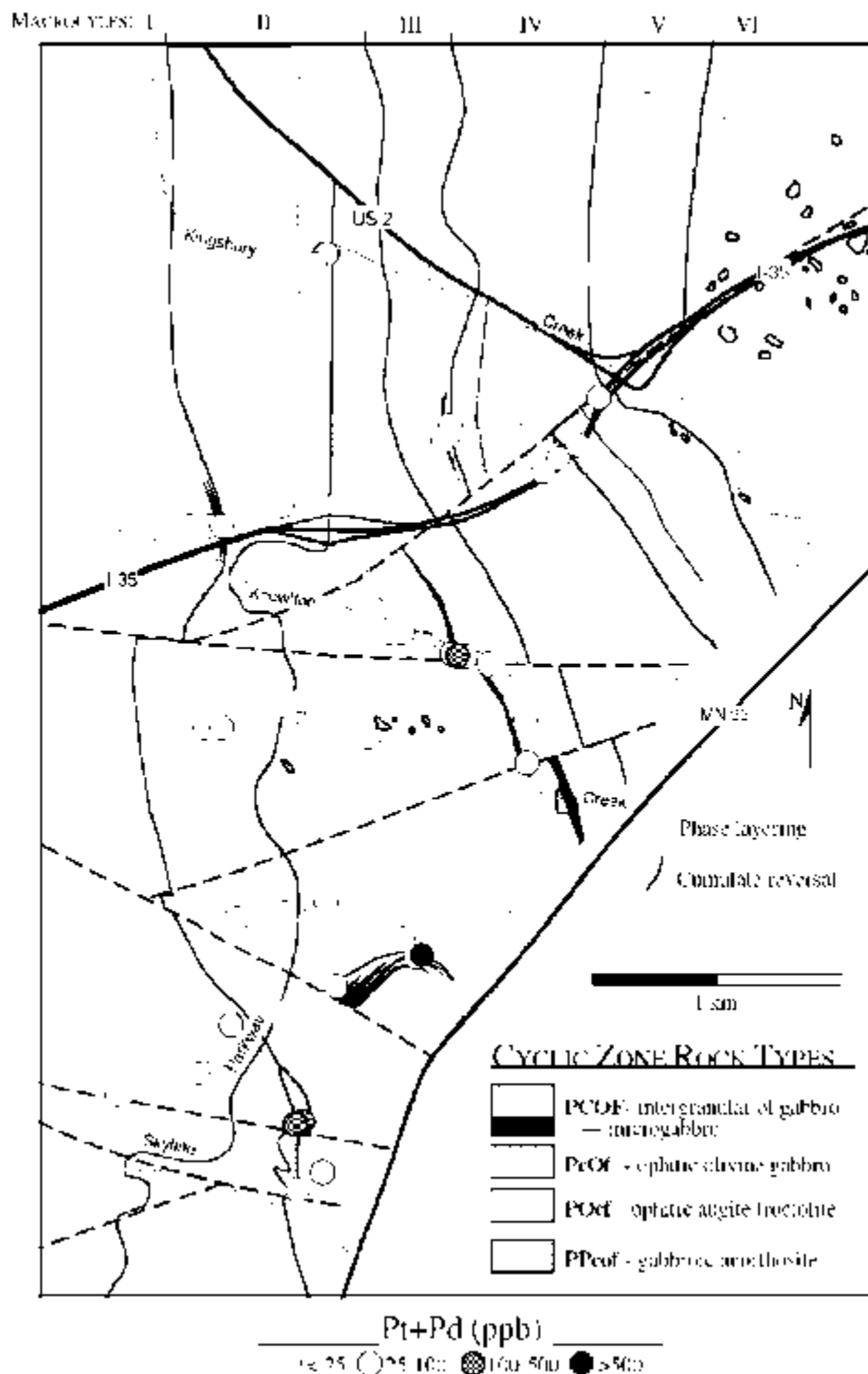


Figure 1-6. Geology of the cyclic zone of the DLS showing locations and concentration ranges of Pt+Pd from outcrop samples. (Modified from Fig. 8-12, Severson et al., 2002).

FIELD STOP DESCRIPTIONS

STOP 1-1: Basal Contact Zone and Lower Troctolite Zone, Duluth Layered Series and Ely's Peak Basalts

Location: Railroad Grade of the former Duluth, Winnipeg, and Northern Railway, Bardon Peak, Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Secs. 33 SE & 34 SW; approx. NAD 83 UTM – Start: 557740E, 5170300N; End: 558460E, 5170040N).

Duration: 90 min.

Description: This stop traverses about 400 meters of stratigraphic section of the lower part of the Layered Series at Duluth (DLS). The exposures will be accessed by walking the railroad grade from the western side of Ely's Peak. Along the way, several extensive exposures of Ely's Peak basalts (including a tunnel) will be encountered. The Ely's Peak basalts form the footwall of the Duluth Complex and show the effects of intense thermal metamorphism by the gabbro. The Ely's Peak basalts are comprised of about 1.5 kilometers of gently (10-15°) dipping, tholeiitic basaltic lavas and represent the earliest volcanic eruptions into the Midcontinent Rift.

Starting at the basal contact, which trends ~N-S in a valley between Ely's and Bardon's peaks, a variety of rock types belonging to the basal contact zone and the lower part of the troctolitic zone of the DLS are exposed in roadcuts and outcrops along the abandoned railroad grade as it arcs around Bardon Peak. Several macrocyclic layers grading in mode and texture are apparent in the lower part of the troctolitic zone and are thought to represent major recharge and differentiation/cooling episodes associated with the early integration stage of DLS emplacement. The exposures on the north side of the tracks are schematically shown in Figure 1-7 and areas of particular interest are flagged along the route and described below.

A. Basal Contact

The traverse starts at the irregular contact between strongly hornfelsed mafic volcanics of the Ely's Peak basalts and taxitic olivine gabbro and augite troctolite of the basal contact zone (Flag A). Granophyre dikes cut both rock types, but display lobate contacts with the coarse gabbro suggesting two-liquid mixing. Grout (1918e) cited this mixing of granophyre and gabbro as evidence of silicate immiscibility. An alternative explanation is that the granophyre represents anatectic melts from the footwall that were not completely assimilated into the mafic magma. Contacts between the hornfels basalt and the gabbro are commonly characterized by coarse augite prisms oriented perpendicular to the contact. Roadcuts on the downslope side of the railroad grade (Flag A') show a small, irregular intrusion of coarse gabbro into hornfels basalt.

B. Basal Contact Zone

This exposure of taxitic olivine gabbro to augite troctolite with its variable textures (medium-grained to pegmatitic, subophitic to ophitic, nonfoliated to poorly foliated) and variable modal compositions are characteristic of the basal contact zone. In contrast to other layered series intrusions along the northwestern margin of the Duluth Complex that locally contain abundant Cu-Ni(-PGE) sulfide (see Day 2 field stops), the basal contact zone of the DLS is devoid of significant sulfide mineralization.

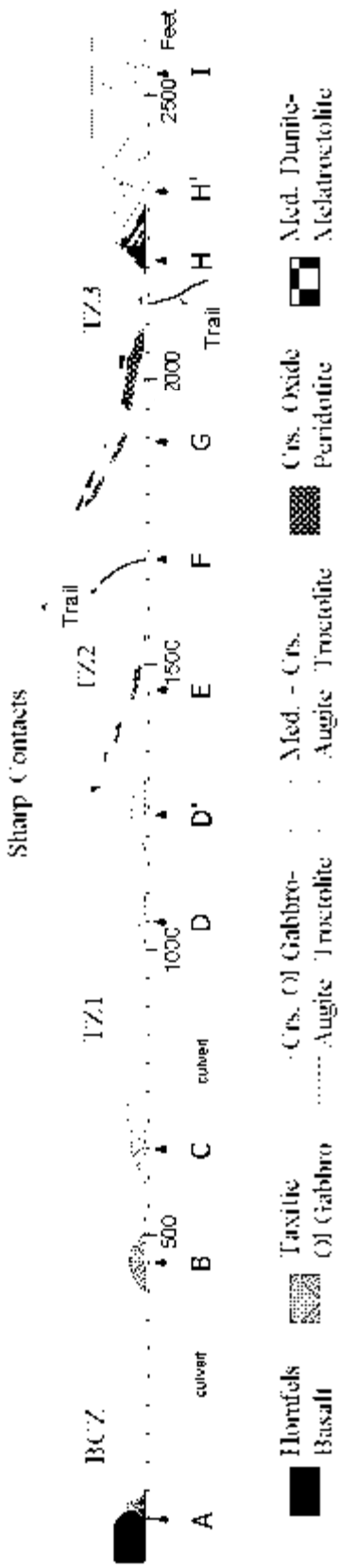


Figure 1-7. Outcrop exposed along the north side of the Duluth, Winnipeg and Northern railroad grade showing variability in rock type across the basal contact zone (BCZ) and several macrocycles in the lower troctolitic zone (TZ1-3).

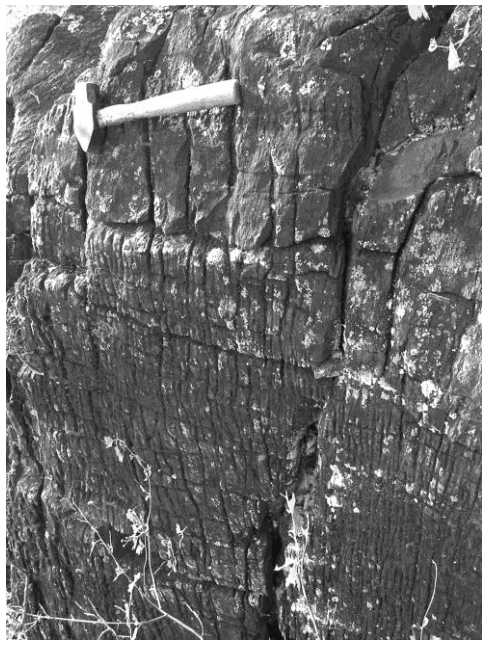


Photo 1-1.
Strong sheet jointing in feldspathic dunite formed from intense serpentine veining. Area 1G'

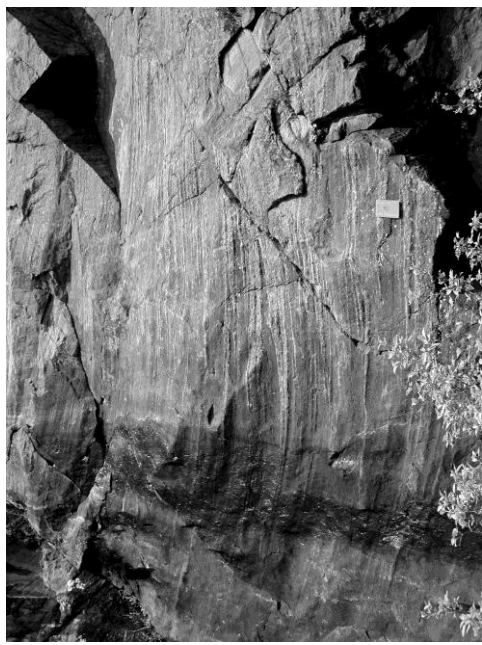


Photo 1-2.
Trough layering at Area 1-I.
Note thickening of leucocratic layers toward the axis of the trough.

C. Lower Macrocylic Unit of the Troctolite Zone

Exposed here is a modally layered, medium-grained, moderately foliated, ophitic augite troctolite. This foliated POcf cumulate marks the base of the troctolitic zone and the first of three macrocylic layers exposed along this traverse. At the east end of this exposure, the rock grades to a medium coarse-grained, augite troctolite to olivine gabbro.

D. Upper Part of Lower Macrocycle

This exposure of medium coarse- to very coarse-grained, non-foliated to moderately foliated, ophitic to subophitic olivine oxide gabbro characterizes the upper part of the lower macrocycle (TZ1, Fig. 1-7). The exposure at D' is coarser grained and generally non-foliated.

E. Contact between Lower and Medial Macrocycles

Down from the grade is a sharp contact between coarse-grained non-foliated olivine oxide gabbro in a lower exposure (similar to outcrop D') and a coarse-grained feldspathic, biotitic oxide peridotite above. Following this contact upslope to the west, the peridotite gives way to a medium-grained melatroctolite overlying coarse olivine oxide gabbro. This abrupt transition from coarse olivine gabbro to medium melatroctolite is interpreted to mark a major recharge event into the DLS magma chamber. The origin of the coarse-grained oxide peridotite is unclear (see Stop 1-2A for a discussion). The oxide peridotite commonly contains trace amounts of sulfide (note local Cu staining) and has higher PGE concentrations (10-30 ppb) than surrounding rock types.

F. Middle Section of Medial Macrocycle

Medium- to medium coarse-grained, moderately foliated, locally layered augite troctolite is exposed in small outcrops on either side of trail leading north of tracks. These exposures occur in the midsection of the second macrocycle (TZ2, Fig. 1-7).

G. Contact between Medial and Upper Macrocycles

Exposures nearest the north side of the railroad grade are of massive, biotitic coarse-grained oxide peridotite which form the upper part of the medial macrocycle (as at E). Following the flagging to an outcrop ledge, an abrupt contact is observed between the oxide peridotite and medium fine-grained, well foliated, feldspathic dunite to melatroctolite. The orientation of the contact between the feldspathic dunite and the oxide peridotite is locally irregular. It is unclear if this irregularity is due to intrusion of the peridotite into the dunite or due to scouring of the dunite into the peridotite. Evidence for both processes are found elsewhere in this area. Here and in exposures along this ledge, the feldspathic dunite/melatroctolite display a very pronounced sheet jointing formed by intense serpentine veining (Photo 1-1). Following the flagging over the top of this ledge leads to another 5-meter-high ledge outcrop that displays lenticular interlayering of troctolite and melatroctolite. Lenses of troctolite consistently thicken to the west suggesting that they may denote channel or trough layering. A similar type of lenticular (channel?) layering is more completely displayed at I, which is approximately at the same stratigraphic level. Also evident at the eastern end of this upper exposure is steeply inclined serpentine veining that refracts through the different layers.

H. Lower Part of Upper Macrocycle.

At the beginning of a very deep cut is well layered melatroctolite grading up from the feldspathic dunite observed at G and forming the lower part of the uppermost macrocycle (TZ3, Fig. 1-7). The melatroctolite grades upward into a medium-grained, well foliated and intermittently layered, ophitic augite troctolite. At two locations (flagged as H'), the troctolite is cut by irregular steeply oriented bodies of oxide olivine gabbro pegmatite. Some have suggested that this pegmatite and the oxide peridotite are related late magmatic features.

I. Layered Troctolite of the Upper Macrocycle

At the highest part of the cut is an exceptional exposure of lenticular layered melatroctolite and leucotroctolite. Similar to the upper exposure at G, trough/channel structures are seemingly implied by the bowing of the layering and the consistent thickening of leucocratic layers in the axis of the trough. Several trough structures have been recognized in the Troctolitic Zone in the Bardon Peak area. The axis of the troughs tend to trend easterly (i.e., in the down dip direction of regional layering).

Stop 1-2A: lower troctolite zone, Duluth Layered Series, Duluth Complex (OPTIONAL)

Location: Skyline Parkway near Bardon Peak, Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Sec. 34, SW of NW; NAD83 UTM (Area E) 558540E, 5170280N).

Duration: (60 min.)

Description: This stop examines rocks comprising the lower part of the Duluth Layered Series in roadcuts and barren knob exposures at the south end of Skyline Parkway. This outcrop area is approximately 500 m above the moderately dipping (~45°) basal contact of the Duluth Complex against shallow-dipping (<15°) basalt flows, which form Ely's Peak (rocky hill to the west). This area is situated in the lower part of the troctolitic zone, but upsection of the upper exposures observed in the traverse at Stop 1-1. As seen in that section, this part of the Troctolite Zone continues macrocyclic layering of melatroctolite grading upward to coarser augite troctolite. Because the layering is similar to that seen below, it is unlikely that we will be able to visit this exposure. It is included here for those who wish to return at a later date. Figure 1-8 shows the geology of the area with eight areas of interest (A-H) noted and generally described below.

The PO to OP cumulate rocks exposed at this stop include ophitic olivine gabbro, augite troctolite, troctolite, melatroctolite, and feldspathic dunite. In general, the more olivine-rich melatroctolites and dunites tend to be finer grained than the more augite-rich troctolites and olivine gabbros. These rocks display a variety of types and scales of layering. Macrocyclic layering of rock types is evident on a meter to decimeter scale—typically finer grained troctolite, melatroctolite, and feldspathic dunite upward to from coarser grained augite troctolite to olivine gabbro. Cryptic differences in olivine composition are evident with melatroctolite to feldspathic dunite being more Mg-rich (Fo₆₂₋₆₅) compared to augite troctolite-olivine gabbro (Fo₄₈₋₆₀). Contacts between macrocycles are abruptly (<10 cm) to narrowly gradational (<1 m) (areas C, D, and G, Fig. 1-8). Within macrocycles, various type of finer-scale layering is common. Subtle grain-size layering locally occurs in the medium- to coarse-grained augite troctolite to olivine gabbro intervals (area E). Centimeter-scale isomodal layering and decimeter-scale graded modal layering is very common in the more melatroctolitic rocks and locally has very rhythmic alternations (area F). Trough layering is locally implied by variable orientations of modal layering (areas A and C) and the tendency

for the more melatroctolitic layers to pinch out along strike. However, some of what appears to be pinch out may be due to faulting along ENE-trending structures through the area (Fig. 1-8). Very abrupt reorientation and steepening of lamination in Area G suggests that some faulting may be contemporaneous with crystallization.

Deeply weathered, coarse-grained, biotitic oxide dunite to peridotite, similar to that seen at Stop 1-1 (areas E & G), occurs in two locations here (areas B and G, Fig. 1-8). Petrographic studies show the rock to be composed of 50-90% granular olivine, 3-20% Fe-Ti oxide (mostly ilmenite), 2-25% ophitic augite, 1-4% biotite, and 0-20% interstitial plagioclase. Primary minerals are commonly altered to chlorite, serpentine, talc, and actinolite. Olivine compositions in these bodies are Fo₅₀₋₅₉ and therefore are similar to the augite troctolite and olivine gabbro cumulates in the area. The limited aerial extent of these oxide ultramafic bodies suggest that many may occur as small pipes or dike-like bodies. However, as seen at Stop 1-1, they also occur as stratabound, sub-conformable lenses at olivine gabbro-melatroctolite contacts.

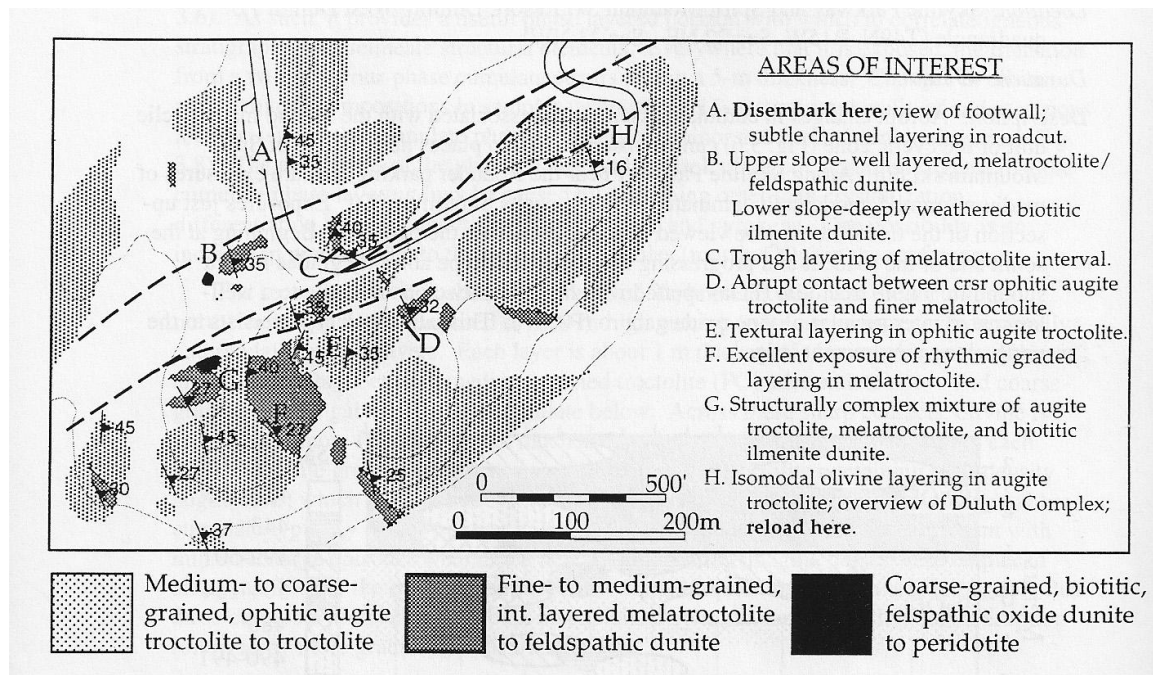


Figure 1-8: Outcrop geology of the Bardon Peak area denoting areas of interest at Stop 1-2A. Dashed lines are inferred faults. Narrow lines are topographic contours portraying 50 foot intervals.

Ross (1985) suggested that these small oxide plugs and lenses are metasomatic replacement bodies formed by volatile fluxing out of the footwall basalts. Severson (1995; Severson and Hauck, 1990) found similar oxide ultramafic (OUI) bodies along the western and northwestern margin of the Duluth Complex to be spatially related to iron-formation inclusion and suggested that partial melting and assimilation of these inclusions may have generated the bodies. Another possible explanation is that they formed by mobilization of

interstitial volatile-rich magma extracted from lower cumulates in the basal contact zone which became ponded beneath fine-grained melatroctolite/dunite layers at the base of macrocycles. More study, especially isotopic data, are clearly needed to resolve the origin of these bodies.

STOP 1-2B: troctolite zone, Duluth Layered Series and Overview of Duluth Complex.

Location: Skyline Parkway, Bardon Peak, Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Sec.34 SW; approx. NAD 83 UTM - 5170450N, 558740E).

Duration: 15 min.

Description: This brief stop, which corresponds to Area H in Figure 1-8, provides a panoramic view of the breadth of the Duluth Complex at Duluth. A 5 kilometer-thick stratigraphic section of the layered series and the overlying anorthositic series are exposed along the 200 meter high, and 15 kilometer long escarpment above the St. Louis River estuary. The top of the Duluth Complex on the rise just above downtown Duluth which is marked by a cluster of radio towers and the Enger Tower landmark. This is where our trip will end today (Stop 9). As exemplified by the layering of the troctolitic rocks exposed at this overlook, the general internal structure of the DLS dips 20-25° to the east, thus the northeast trending escarpment cuts an acute angle across the intrusion. The anorthosite series overlying the DLS can be recognized by its higher topographic expression.

STOP 1-3: Cyclic Zone, Duluth Layered Series

Location: Spirit Mountain Ski Resort, Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Sec 23 NW; approx. NAD 83 UTM - 5174250N, 559952E).

Duration: 30 min.

Description: About 100 m east of the north side of the ski lodge, near the top of the northernmost chairlift is series of outcrop ledges that display over a 10-meter-thick section about 16 successive meter-scale mesocycles (Fig. 1-9) that mimic the decameter-scale macrocycles that characterize the medial Cyclic Zone of the DLS (Fig. 1-3). Each mesocycle is composed of a medium fine grained troctolite at its sharp base with the cycle below. About 1/3 of the way up, the troctolite develops small clots (~5mm) of subpoikilitic augite and lesser oxide, which then grades on a cm-scale to a medium-grained, intergranular oxide olivine gabbro (Photos 1-3 and 1-4). All rocks display well developed foliation of plagioclase (in troctolite) and plagioclase and augite (in oxide olivine gabbro). Although the thickness of each mesocycle gradually increases from 30 cm to 100 cm upsection, the proportion of gabbro to troctolite comprising each cycle remains generally constant at about 2:1 (Fig. 1-9). The mesocycles fundamentally show a rapid progression from a two-phase (PO) to a four-phase (PCOF) cumulate. The mesocyclic boundaries represent even more abrupt regressions from four-phase back to two-phase cumulates.

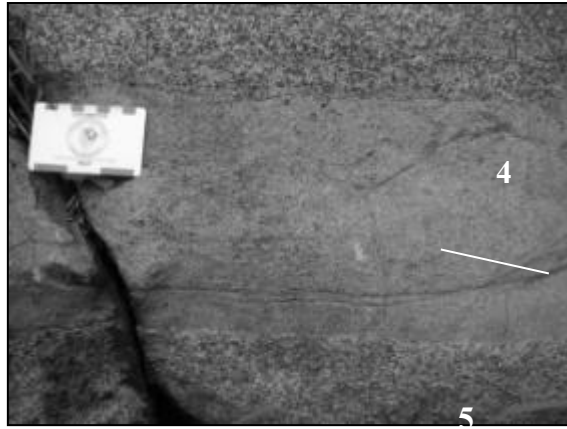


Photo 1-4. Close-up of contact relationships between oxide olivine gabbro (OG) and troctolite (Tr).

Photo 1-3. Mesocyclic phase layering at Stop 1-3. Cycles 4 and 5 and the upper part of Cycle 6 are visible in this photo.

Microprobe analyses of olivine and augite in samples collected across mesocycles 1 through 4 and 10 through 13 (Fig. 1-9) show subtle but systematic variations in mg# ($\text{MgO}/(\text{MgO}+\text{FeO})$, mole%). Olivine (and less consistently augite) show a 2-3% increase in mg# in the troctolite intervals compared to the gabbro intervals. The mg# increase can be recognized on the scale of a single thin section at the sharp mesocyclic boundaries. The maximum mg# typically is found in the central part of the troctolite intervals then gradually decreases as subophitic augite clots appear. The mg# stays remarkably constant through the gabbro intervals. The average mg# of each mesocyclic zone is also remarkably constant across the multiple cycles investigated (Fig. 1-9).

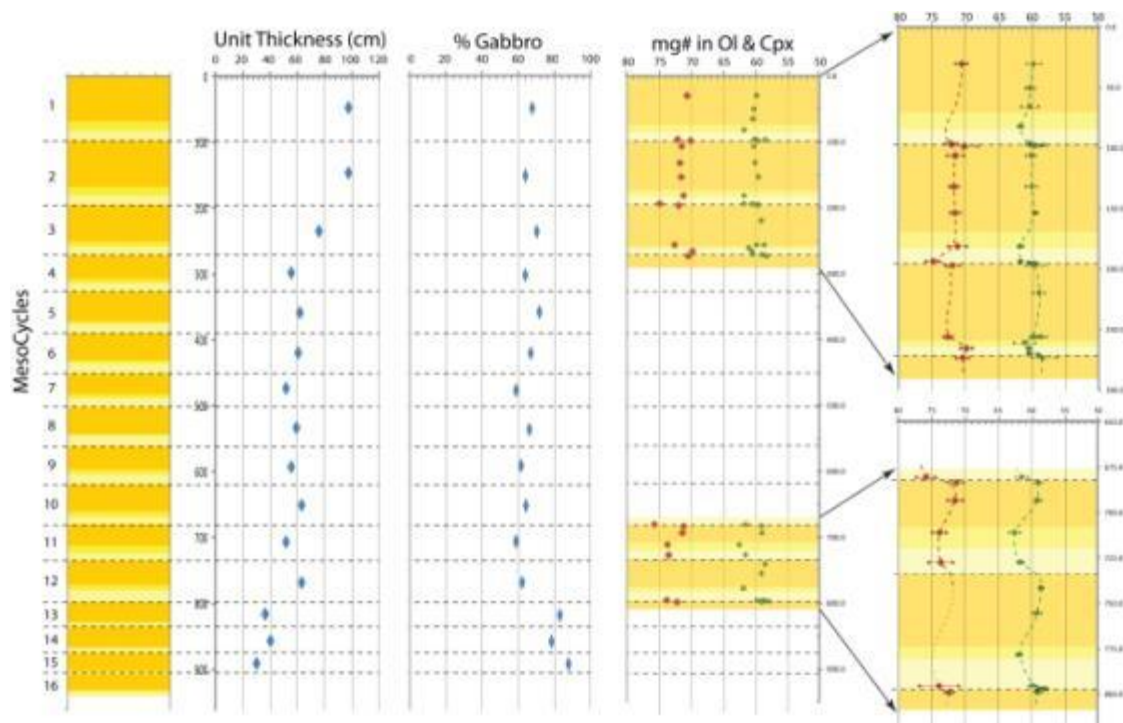


Figure 1-9. Stratigraphic variations in mesocycle thickness, volume % of gabbro composing each mesocycle, and mg# ($\text{MgO}/(\text{MgO}+\text{FeO})$, mole%) of olivine and augite through the mesocycles exposed at Stop 1-3. White parts of the mesocycles indicate troctolite intervals (PO cumulate), yellow indicates subophitic augite troctolite (POcf cumulate), and orange indicate olivine oxide gabbro intervals (PCFO cumulate). Figure after Miller and Stifter (2010).

This may at first glance seem to argue for recharge of more primitive magma as the cause for the phase layering. However, the shift in mg# in olivine may also reflect a shift in the Fe-Mg equilibrium between melt and olivine when augite becomes a cumulus phase. Trace element data, particularly REE, also show a significant difference between the troctolite and gabbro layers, but this can be attributed to differing partition coefficients for the main cumulate phases (Miller and Stifter, 2010). Collectively, the field, petrographic, mineral chemistry, and lithochemical attributes of the mesocyclic layering observed here does not unambiguously favor either a magma recharge or decompression venting model for its origin. More data and evaluation are needed to definitely determine the origin of the mesocyclic and larger macrocyclic layering of the Cyclic Zone of the Duluth Layered Series.

Stop 1-4: Cyclic Zone, Duluth Layered Series

Location: North of I-35, west of Boundary Ave., slope north of Northern Engine & Supply Co. and Spirit Mtn. Red Roof Motel, Duluth. West Duluth 7.5' quadrangle (T49N, R15W, Sec 15 SE, NAD 83 UTM - 5175100N, 559350E (area E))

Duration: 60 min.

Description: This stop examines the first macro-scale transition from troctolite cumulates to olivine gabbro cumulates that defines the base of the Cyclic Zone (Fig. 1-3).

From behind the Northern Engine and Supply Co., head north and uphill to outcrops of medium-grained ophitic augite troctolite (A, Fig. 1-10). This POcf cumulate contains about 7-10% interstitial Fe-Ti oxide and augite, with the latter commonly forming 2-4 cm oikocrysts. Progressing about 70 m to the east over intermittent outcrop ledge, the augite troctolite gradually becomes coarser grained and more enriched in augite and oxide (B, Fig. 1-10).

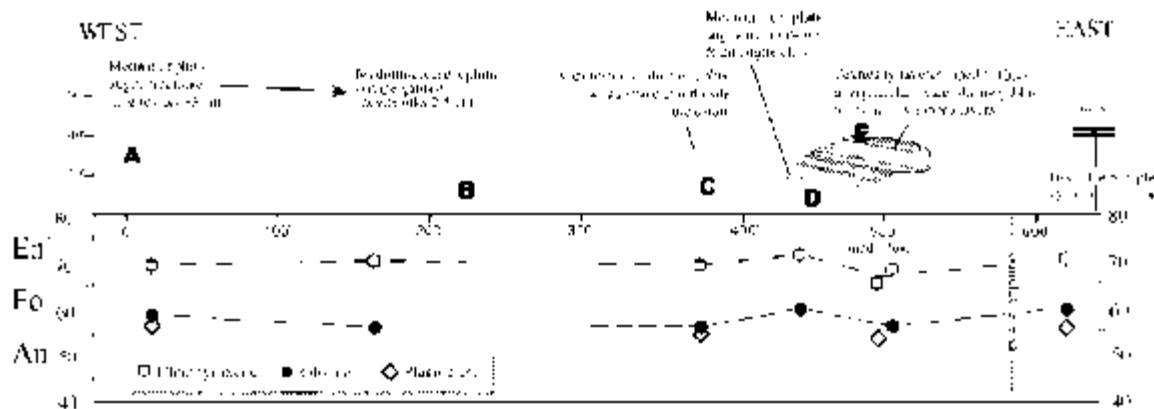


Figure 1-10: Outcrop geology and cryptic variation along south facing slope at Stop 1-4. Igneous foliation and layering dip at 15-25° to the east. Exposure corresponds to the top of macrocyclic unit I marking the base of the cyclic zone (Fig. 1-3). Troctolitic rocks 120 m east of the powerline behind the motel indicate the presence of an unexposed cumulus reversal.

Crossing a 35-m gap in exposure, the next outcrop is of a medium- to very coarse grained, subophitic to ophitic olivine gabbro (PcOf; C, Fig. 1-10). In one area of the exposure, this rock type is in sharp, discordant, but unchilled contact with a coarse-grained, moderately laminated gabbroic anorthosite that is probably an inclusion of the anorthositic series. Beyond an 8-m gap is a 10-m-high outcrop knob that grades from a medium-grained, subophitic augite troctolite with 1/2-cm high density augite oikocrysts near the base (D, Fig. 1-10) to a texturally layered, medium- to fine-grained, intergranular oxide olivine gabbro at the top (E, Fig. 1-10).

Textural layering on top of knob is defined by 2- to 20-cm-thick, laterally discontinuous layers of fine grained, moderately laminated, intergranular oxide olivine gabbro (microgabbro) that intermittently occurs within medium-grained, well-laminated, intergranular oxide olivine gabbro. Contacts between the medium gabbro and the microgabbro range from sharp to gradational on a scale of centimeters. In some areas, the textural layering drapes over football-sized blocks of a coarser grained, subophitic olivine gabbro, similar to the rock type seen just to the west (down section). Gossan of sulfide-bearing areas are locally present.

If time allows, we will follow flagging tape about 200 meters from the powerline (Fig. 1-10) to the parking lot behind the Red Roof motel. Exposed in an 80m-long roadcut and pavement exposure is medium-grained, well foliated, locally layered augite troctolite, thus implying that a cumulus reversal was crossed between site E and here. This rock is interpreted to form the lower part of the second macrocycle unit (Fig. 1-3). Exposure of this macrocycle boundary to the south shows that the texturally layered oxide olivine cumulate

interval (E, Fig. 1-10) is only locally present and appears to be lenticular in form (Fig. 1-3). In some exposures, the boundary between macrocycles I and II is marked by a coarse-grained ophitic olivine gabbro with anorthositic series inclusions (as seen at C, Fig. 1-10) in sharp contact with troctolite.

With the exception of the medium-grained, subophitic augite troctolite at D, cryptic variations through this sequence show a general decrease in An, Fo, and En (Fig. 1-10). Plagioclase, olivine, and augite compositions in the troctolite above the layered gabbro define a regression to more primitive values. Mineral compositions are also more primitive in the microgabbro layers compared to the enclosing medium-grained gabbro. Across the cumulus reversal from oxide olivine gabbro to troctolite, Fo, En and An increase slightly.

Again a straightforward interpretation of this sequence is that it represents progressive crystallization differentiation leading to multiple saturation in plagioclase, olivine, augite, and ilmenite, followed by recharge of a more primitive magma causing a regression to saturation in just plagioclase and olivine. The occurrence of anorthosite inclusions and the microgabbro layers suggest however that any such recharge event may have been preceded by an eruption event from the chamber. Miller and Ripley (1996) interpreted the microgabbro lenses to have formed by decompression quenching of magma due its reaching water saturation in the roof zone (see Stops 1-5 and 1-7 for further discussion).

Decompression may also have triggered sulfide saturation which caused local stratiform enrichment of PGE at this and other cumulus regression. Although some enrichment in PGE above background is noted from samples taken from site E, more pronounced sulfide and PGE enrichment are evident at this macrocycle interface to the south and at the interface between the second and third macrocycles (Figs. 1-5, 1-6).

Stop 1-5: Cyclic Zone, Duluth Layered Series

Location: Roadcut on Skyline Parkway at Thompson Hill Rest Area. West Duluth 7.5' quadrangle (T49N, R15W, Sec 14, center, NAD 83 UTM - 5175350N, 560700E)

Duration: 90 min (including lunch).

Description: In the layered sequence of gabbroic rocks exposed along this 200-m-long roadcut (Fig. 1-11), a reversal in cumulus paragenesis can be seen that is characteristic of higher levels in the cyclic zone. The west end of the roadcut begins with a coarse-grained, moderately laminated, subophitic olivine gabbro, which locally is intergranular and elsewhere is leucocratic (sample A, Fig. 1-11). Augite in this rock is marginally cumulus but becomes definitely so quickly upsection where it consistently has an anhedral granular to subprismatic habit. This coarse-grained, intermittently layered (locally graded), olivine-poor (<5%) oxide gabbro (sample B) classifies as a PCFO cumulate. Beyond a poorly exposed interval, this gabbro includes a 2-m-thick interval wherein minor olivine becomes subpoikilitic and concentrated in layers (sample C). Another 3 m above this, the coarse gabbro passes into a medium-grained, well-laminated, subpoikilitic olivine-bearing oxide gabbro (samples D and D') which displays layering of olivine oikocryst concentration and elsewhere isomodal layers rich in Fe-Ti oxide and pyroxene. The very strong foliation and subhedral to euhedral habit of cumulus phases (plagioclase, pyroxene, and ilmenite) impart an adcumulate texture to this rock. Over a poorly exposed interval about 15 m long is an altered, coarse-grained, ophitic gabbroic anorthosite (sample E) that is texturally and mineralogically identical to rocks in the anorthositic series. At the beginning of the next well-exposed section of roadcut, several similar gabbroic anorthosite inclusions are found in

a coarse-grained, subophitic to intergranular olivine gabbro (sample F), which gradually grades upward into a more consistently subophitic texture over the remainder of the roadcut (sample G). This rock type closely resembles that at the west end of the roadcut and indicates a downgrading in the cumulus status of pyroxene (and oxide?) and a reemergence of cumulus olivine.

Figure 1-11: Geology and cryptic variation along Skyline Parkway roadcut near Thompson Hill rest area; Stop 1-5. View is to the north. Dip of lamination and layering is exaggerated; averages about 20° to the east. Contacts between units are gradational over thicknesses of 10 cm to 1 m.

Decompression of a volatile-enriched magma would cause supercooling and multiple saturation of the magma and thereby explain the abrupt decreases in grain size and cumulus phase changes without much compositional variation. Magma expulsion through the roof of the layered series would also explain the occurrence of a gabbroic anorthosite inclusion and elsewhere throughout the cyclic zone. The hydrothermally altered nature of the gabbroic anorthosite is consistent with a volatile-rich environment in the anorthositic series cupola of

the DLS magma chamber. The cumulus reversal to ophitic olivine gabbro without an increase in mg# could be explained by repressurization of a devolatilized magma.

STOP 1-6: Gabbro Zone, Duluth Layered Series and Anorthositic Series

Location: Roadcuts on Skyline Parkway above Oneota Cemetery. Duluth Heights 7.5' quadrangle (T49N, R15W, Sec 1, SE, approx. NAD 83 UTM - 5177970N, 562995E)

Duration: 30 min.

Description: In roadcuts along a 0.5-km section of Skyline Parkway, rocks forming the contact between the layered series and the anorthositic series may be examined. Exposed in low, deeply weathered outcrops at the west end of the section is a poorly to well foliated, intergranular, apatitic olivine ferrodiorite cumulate (PCFOAp). It is composed mostly of plagioclase (zoned An_{38-50} , avg. An_{44}), augite (avg $En_{33}Fs_{27}Wo_{40}$), and bladed ilmenite. In addition, as much as 10% olivine (Fo_{30}), 5% orthopyroxene (En_{45}), 5% apatite, 3% biotite, and 5% granophyric mesostasis is present. Generally medium-grained, it locally varies in grain size from fine to coarse. Medium- to fine-grained variants have well-developed foliation, whereas coarser grained ferrodiorite tends to be poorly foliated to nonfoliated. Coarser ferrodiorite tends to be more granophyric, as well. By being locally well-laminated, olivine-bearing, and lacking ferromonzodioritic component, the ferrogabbro here is not typical of the upper contact zone of the layered series; relationships more characteristic of the upper contact zone will be seen at the next stop. Rather, this rock is more typical of gabbroic cumulates of the gabbro zone.

About 35 meters east of the ferrogabbro, a prominent roadcut displays several common variants of anorthositic rocks that compose the anorthositic series. The western end of the roadcut is composed of a poorly to moderately laminated, poikilitic olivine gabbroic anorthosite (OGA) that contains olivine (Fo_{42-46}) oikocrysts up to 6 cm across, as well as interstitial augite and oxide. Progressively eastward, this rock type grades into a subpoikilitic to granular olivine gabbroic anorthosite. At the high point in the roadcut, this granular OGA contains an inclusion of olivine (Fo_{60})-bearing anorthosite (Photo 1-5). Whereas the enclosing OGA contain about 80-85% plagioclase (avg. An_{62}), this inclusion contains more than 95% plagioclase (unzoned An_{65}). The contact between OGA and the anorthosite inclusion is sharp with some concordant alignment of plagioclase in OGA.

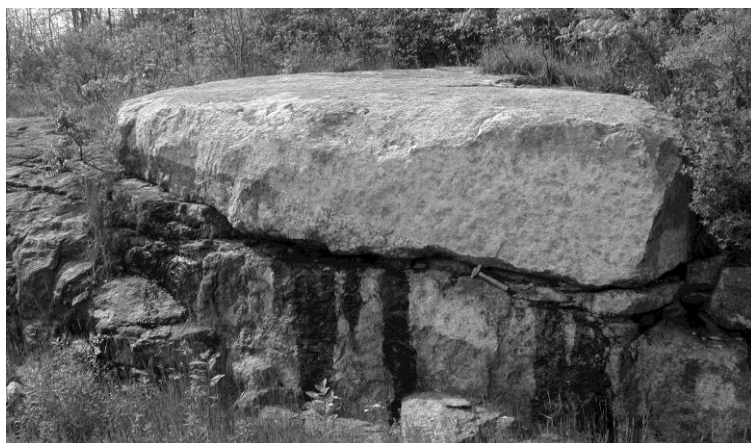


Photo 1-5. Olivine-bearing anorthosite inclusion in subpoikilitic olivine gabbroic anorthosite as Stop 1-6. Hammer at contact.

The inclusion relationships and the outcrop-scale variability in internal structure and lithology (mode, texture, olivine habit, and lamination development) of the anorthositic rocks here are ubiquitous features of the anorthositic series. These complexities led Taylor (1964) to characterize the anorthositic series as an igneous breccia. He and later Miller and Weiblen (1990) concluded that these rocks formed by repeated intrusions of plagioclase crystal mush. This stop is at the western end of a westward projection of the layered series-anorthositic series contact (Fig. 1-2). A possible explanation for the irregular trace of the contact is that it reflects the original shape of the DLS roof. This would seem to be a difficult shape to maintain if the anorthositic series rocks were hot and lighter than the DLS magma. Rather, it seems more likely that the westward projection of the anorthositic series represents a large inclusion that detached from the roof zone and settled down to the cumulate floor of the DLS at the time of gabbro zone crystallization. Evidence for this is suggested by a truncated cryptic variation in the DLS leading up to this contact compared to the thicker DLS section to the north and the atypical character of the ferrodiorite in contact with the anorthositic series here (compare with next stop).

Stop 1-7: DLS “Chill”, Granophyre, and Anorthositic Series

Location : Skyline Parkway, Duluth Heights 7.5' quadrangle (T50N, R14W, Sec 32, SE, NAD 83 UTM - 5179675N, 564880E).

Duration: 30 min.

Description: Exposed at the northeast end of this roadcut is a fine-grained mafic rock with intermingled granophyre that together cut coarse-grained olivine gabbroic anorthosite of the anorthositic series. Since early mapping by Grout in the Duluth area in 1912, this rather innocuous exposure has been a keystone outcrop in interpreting the intrusive history of the Duluth Complex. This fine-grained mafic rock can be traced up over the ridge to the west where it merges into the upper contact zone of Layered Series. Grout (1918a) and later Taylor (1964) saw this exposure as evidence that the anorthositic series was considerable older and cold when the layered series was intruded. This paradigm was accepted by all subsequent workers on the Duluth Complex up through the 1980's. It came therefore as a shock when high resolution U-Pb age dating (Paces and Miller, 1993) showed that the anorthositic series and the layered series were virtually identical in age (within 0.5 Ma relative to the 22 Ma span of Midcontinent Rift magmatic activity). This precipitated a major paradigm shift in the perception of the intrusive relationships between these two series here and throughout the Duluth Complex.

A closer look at this DLS “chill” reveals that it is not a thermal quench of the Layered Series at all. This rock type is found at the contact with the anorthositic series throughout most this area and has a remarkably homogeneous composition with an mg# of about 37 (Table 1-1). In thin section, it is a subprismatic biotitic oxide ferrodiorite. Applying its composition to the MELTS routine indicates that it should be in equilibrium with augite, ilmenite, and plagioclase with compositions comparable to gabbroic cumulates found in the cyclic zone and gabbro zones of the DLS. In sum, this rock is much too evolved to have produced the troctolitic rocks of the layered series. Rather than this being a thermal quench, Miller and Ripley (1996) have interpreted this rock to represent decompression quenching of an evolved, water-saturated DLS magma during venting at a time when the cyclic zone was crystallizing. Whereas decompression of a water-undersaturated magma should result in superheating and a suspension of crystallization (or at least a significant change in phase

equilibrium), decompression under water-saturated conditions should cause supercooling and quenching. This model fits nicely with the explanation for the cyclic zone with which this composition is apparently comagmatic.



Photo 1-6. Lobate contacts between fine-grained oxide ferrodiorite (fd) and medium-grained melanogranophyre (mg), which is in sharp contact with coarse-grained gabbroic anorthosite (ga) at Stop 1-7.

The lobate contacts between the irregular masses of medium-grained granophyre and the fine-grained ferrodiorite host (Photo 1-6) gives the appearance of two magma mixing. The ferrodiorite is not of a composition that would indicate that these two liquids formed by immiscibility. An alternative explanation is that these felsic magmas were derived from anatectic melting of various inclusions carried into the DLS chamber. Because of their high silica content and low density, these felsic melts did not readily mix or assimilate with mafic melt, but rather rose to the roof zone where they ponded beneath the anorthositic series cupola. During magma venting from the chamber, the felsic melts became entrained and irregularly mixed with the mafic magmas. While decompression under water-saturated conditions caused rapid crystallization of the mafic magma, the felsic melt became irregularly entrapped in the quenched mafic host and cooled more slowly to a medium-grained texture.

Table 1-1. Compositions of six fine-grained ferrodiorite samples collected at the DLS-AS contact.

	Analyses of DLS "Chill" samples						Avg. "Chill"
	MD268	MD279A	MD293A	MD247A	DG 103	DG 241	
SiO ₂	48.3	48.5	49.1	47.2	~51	~51	49.32
TiO ₂	3.35	2.97	3.10	3.49	3.92	3.74	3.43
Al ₂ O ₃	13.3	13.1	13.2	12.5	10.6	10.5	12.2
Fe ₂ O ₃	2.3	2.3	3.0	2.8			2.61
FeO	12.0	10.6	10.8	12.3	17.0	16.7	13.2
MnO	0.22	0.18	0.22	0.23	0.20	0.19	0.21
MgO	5.0	4.8	4.8	4.9	4.2	5.5	4.9
CaO	9.2	9.3	8.8	9.3	7.6	8.6	8.8
Na ₂ O	3.10	2.89	2.70	2.68	2.78	2.36	2.75
K ₂ O	0.42	1.05	0.93	0.71	1.19	0.47	0.80
P ₂ O ₅	0.58	0.53	0.55	0.65	0.75	0.71	0.63
TOTAL	98.9	98.2	98.9	98.1			98.9
mg#	38.8	40.1	38.9	37.0	30.8	37.0	37.1
Trace Elements (ppm)							
Cr	110	110	100	110	59	120	102
Ni	76	77	73	63	55	83	71
Rb	15	46	34	19	<15	50	30
Sr	156	134	266	288	190	209	207
Ba	180	186	180	181	400	280	235
Zr	306	341	343	278			317
Y	56	55	56	61	88	56	62
La	35.6	35.6	38	33.9	61	28	38.7
Sm	10.5	10.0	11.0	10.7	14	9.2	10.9
Yb	5.0	4.9	5.2	5.2	9.4	5.5	5.9

(Analyses by ActiLab, Ancaster, Ontario; Miller unpublished data)

Stop 1-8: Ferrodiorite/Melanogranophyre Composite Intrusion in Anorthositic Series

Location : Roadcuts on Skyline Parkway on the south and north sides of Twin Ponds swimming area. Duluth 7.5' quadrangle (T50N, R14W, Sec 21, E central; NAD 83 UTM - 5180870N, 566980E).

Duration: 30 min.

Description: Outcrops and roadcuts north and south of Twin Ponds expose contact relationships between very altered anorthositic series rocks and an irregular composite body of intermediate to felsic rocks. Exposed in the roadcut south of the T-junction is an altered, coarse-grained, granophyric gabbroic anorthosite. All ophitic pyroxene has been replaced by fibrous amphibole, and a granophyric mesostasis locally composes 5-15% of the rock. The altered and granophyre-enriched character of anorthositic rocks is common over most of the upper part of the anorthositic series in the Duluth area (Fig. 1-1).

About 15 m east of the T-junction, an abrupt but unchilled contact is exposed between the gabbroic anorthosite and a medium-grained, apatitic olivine ferromonzodiorite. As seen in thin section, all the olivine and most of the clinopyroxene has been replaced by amphibole, chlorite, and iron oxide, and plagioclase is commonly mantled by dusty K-feldspar which becomes intergrown with quartz to form a granophyric mesostasis. Locally the ferromonzodiorite is moderately laminated, and in one area it displays a streaky modal layering that is moderately inclined to the southeast. About 75 m east of the junction, a very complex contact between the ferromonzodiorite and a plagioclase-porphyritic quartz ferromonzodiorite is encountered which appears to be a hybrid mixture of gabbroic anorthosite and ferromonzodiorite. Over a 25 m interval, this hybrid rock grades into variably altered and granophyric gabbroic anorthosite. With the exception of small irregular bodies of quartz monzodiorite about 140 m from the junction, the remainder of the roadcut is altered and granophyric gabbroic anorthosite.

On the north side of the Twin Ponds, a roadcut exposes a pink, mafic quartz monzonite (or melanogranophyre) cut by a 2-m-wide tholeiitic diabase dike. The melanogranophyre is composed of about 50% subhedral plagioclase, 5-10% skeletal prisms of Fe-pyroxene and amphibole, and 5% granular Fe-oxide in a felsic matrix of micrographically intergrown dusty K-feldspar and quartz. This rock type makes up 80% of the irregular composite body labeled as melanogranophyre in Figure 1-1. Uphill (north) from the roadcut in intermittent pavement outcrops, the melanogranophyre can be observed to grade into more mafic compositions (quartz ferromonzodiorite to ferrodiorite) as sharp but unchilled contacts with gabbroic anorthosite are approached.

The alteration, contact relationships, and variety of rock types observed here indicate that evolved magma and hydrothermal fluids emanating from the underlying DLS passed through the anorthositic series over a protracted period of time and at a variety of scales. The very altered and granophyre-rich character of anorthositic rocks observed here and over much of the upper part of the anorthositic series suggests that late-stage melts and hydrothermal fluids fluxed through the intercumulus pore spaces of the (partially molten?) anorthositic series over a large area. In addition to this widespread percolation, the composite melanogranophyre body appears to represent a discrete conduit through which DLS-originating magmas passed and partially crystallized on their way to higher level intrusions or surface eruptions. The crude zonal distribution of composition more ferrodioritic at the margins and more quartz monzonitic in the interior—is consistent with this conduit being open during crystallization of at least the upper third of the DLS. Moreover, the hybridization between the composite body and the gabbroic anorthosite host suggest that the anorthositic series was partially molten at least in the vicinity of this conduit.

Stop 1-9: Upper Contact of Duluth Complex and North Shore Volcanic Group.

Location : Radio Tower Hill. East of corner on gravel road corresponding to 5th Ave and 9th Street. Duluth 7.5' quadrangle (T50N, R14W, Sec 28, NE; NAD 83 UTM - 5182260N, 567650E).

Duration: 45 min.

Description: The east side of Radio Tower Hill overlooking downtown Duluth approximates the slope of the upper contact of the Duluth Complex with the overlying lava flows of the North Shore Volcanic Group, which now underlie all the lower ground to the northeast along the shore. As observed at the southeastern crest of the hill, the rock exposed

over most of this slope is a medium- to fine-grained, plagioclase-porphyritic (5-15%), subophitic to ophitic olivine gabbro that contains many decimeter- to meter-sized basaltic hornfels inclusions. In thin section, the gabbro typically displays intense hydrothermal alteration resulting in sericitic breakdown of plagioclase and uraltitic, chloritic, and oxide replacement of pyroxene and olivine, though its primary subophitic to ophitic texture is typically preserved. The basaltic inclusions are locally intensely recrystallized and the gabbro shows little to only a weak chill around the inclusions. Less commonly, an inclusion of gabbroic anorthosite is present in the gabbro. To the west (downsection), basaltic inclusions become less common and plagioclase phenocrysts become more abundant, but in a nonsystematic way. The contact with the main suite of gabbroic anorthosite is not well exposed but seems to be abrupt and may actually finger into the anorthositic series as dike-like apophyses (Fig. 1-12).

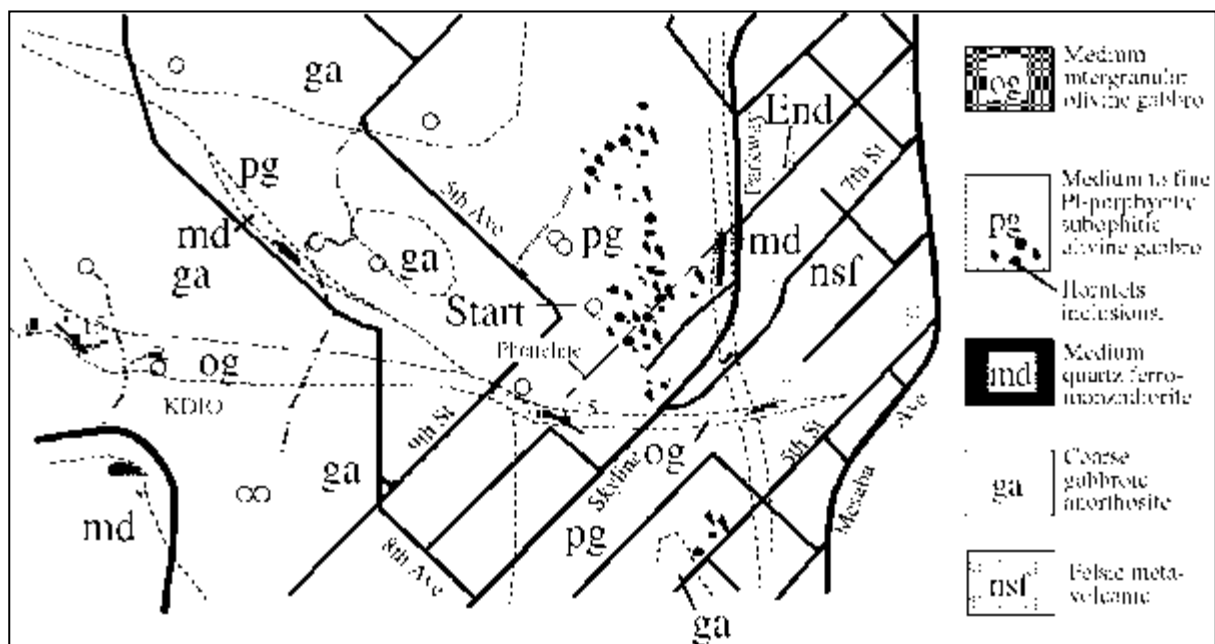


Figure 1-12: Outcrop geology of Radio Tower hill area, showing the starting and ending locations of Stop 1-9. Open circles indicate the location of radio towers.

Along a telephone line down the slope to the northeast, the gabbro gradually becomes finer grained and less porphyritic. At the base of the slope, the gabbro abruptly grades into a pink, medium-grained, intergranular to micrographic hornblende quartz monzonite (Fig. 1-12). The quartz monzonite is composed of ~15% amphibole and ferroaugite, 5% granular Fe-oxide, 40% plagioclase, 30% dusty K-feldspar, and 10% graphitic to granular quartz. Abundant miarolitic cavities indicate a shallow depth of crystallization. The contact relationship between the gabbro and the quartz monzonite is best seen in a roadcut along the driveway to 415 W. Skyline Parkway. Here, the two phases, along with some basaltic inclusions, occur in a complex mixture over a 15-m-wide zone. The melanogranophyre

probably formed by partial melting of intermediate to felsic lavas of the overlying North Shore Volcanic Group which form the hanging wall of the Duluth Complex. A strongly recrystallized porphyritic felsic (icelandite) volcanic rock can be observed in low roadcuts about 40 m northeast of Skyline Parkway on W. 8th St.

A possible interpretation of the porphyritic gabbro is that it represents the upper border phase of the anorthositic series that formed by flow differentiation of plagioclase crystals away from the contact with overlying volcanic rocks. An alternative explanation is that this body represents a discrete intrusion, perhaps comagmatic with the layered series, which was emplaced at the anorthositic series-volcanic contact. Except for its highly altered state, its primary mineral compositions and texture are similar to the basal contact zone rocks of the layered series.

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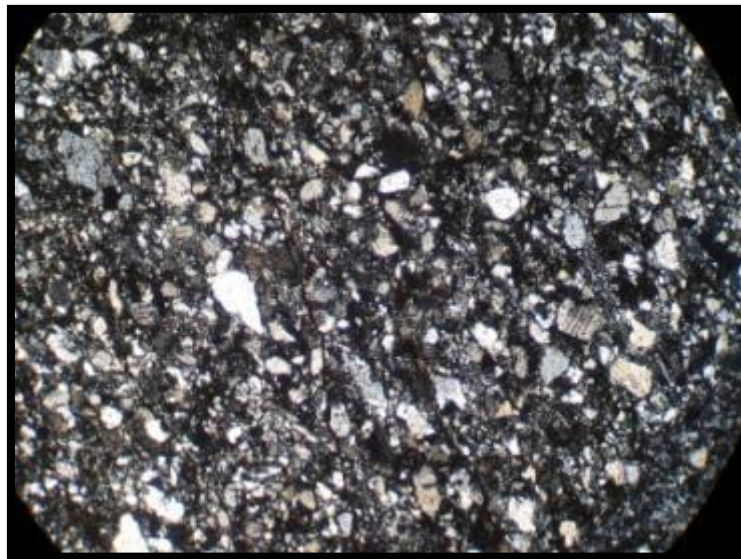
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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 2

MIDCONTINENT MICROCOSM: GEOLOGY OF THE ATKINS LAKE –MARENGO FALLS AREA



Photomicrograph of ultracataclasite from the north side of the Marengo River below Marengo Falls. The protolith is probably basalt of the Siemens Creek Volcanic series. The field of view is about 2 mm. – M. Bjørnerud

Field Trip 2

Midcontinent Microcosm: Geology of the Atkins Lake –Marengo Falls area

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OVERVIEW OF FIELD TRIP STOPS

1. Paleoproterozoic Ironwood Iron-formation, metadiabase sills, and breccias near Atkins Lake
2. Mesoproterozoic Mellen-type granite near Atkins Lake
3. Mesoproterozoic Siemens Creek Volcanics contact metamorphosed to pyroxene hornfels and granophyre
4. Penokean shear zone in Paleoproterozoic Palms Formation, Marengo River
5. Paleoproterozoic Bad River Dolomite: Stromatolites, Penokean deformation, contact metamorphism, Grandview Quarry, Marengo River
6. Marengo Falls, South side: Neoarchean Puritan Batholith; fault breccias, cataclasites and pseudotachylyte along the Atkins Lake-Marengo Fault
7. Marengo Falls, North side: Ultracataclasite along the Atkins Lake-Marengo Fault

INTRODUCTION

Archean and Proterozoic rocks exposed over about 16km² between Atkins Lake and Coffee Lake in southeastern Bayfield County (Fig. 1) chronicle almost all of the major Precambrian geologic events in the history of the southern Superior Craton. The oldest rocks are part of a locally gneissic quartz monzonite complex, the Puritan Batholith, with an igneous Rb-Sr age of 2710±140 Ma (Sims et al., 1977). At the regional scale, this complex is part of one of the youngest Archean granite-greenstone belts in the Superior Province, and it intrudes greenstones of the Neoarchean Ramsay Formation. In the Atkins Lake – Marengo River area, the Puritan Batholith is nonconformably overlain by the Paleoproterozoic (ca. 2200 Ma) Bad River Dolomite. The Bad River Dolomite is in turn separated by an unconformity from rocks of the ca. 1875 Ma Menominee Group (Palms Formation and Ironwood Iron-formation), which locally contain mafic volcanic rocks and diabase sills (Cannon et al., 2008). These Paleoproterozoic rocks provide insight into climate and biogeochemical cycles during the transition to an oxidizing atmosphere (Bekker et al., 2006) and have deformational fabrics (folds, strong cleavage, local mylonite zones) that record the ca. 1850 Ma Penokean Orogeny. The youngest rocks in the area are Mesoproterozoic

basaltic lava flows (Siemens Creek Volcanics, ca. 1110 Ma) and a layered mafic complex (the Mineral Lake Intrusion, also ca. 1100 Ma), both related to the Mid-continent Rift. All of the stratified units show static contact metamorphic textures near their contacts with the Mineral Lake Intrusion. Thus the area constitutes a microcosm of the regional bedrock geology, and the cross-cutting relationships among the units provide clear constraints on the relative timing of different phases of deformation and magmatism (Cannon et al., 2008, Bjørnerud, 2010a).

The rocks in the Atkins Lake-Marengo Falls area are part of a regional structure called the Montreal River Monocline (Fig. 2), a northward-dipping sequence of Archean through Mesoproterozoic rocks that exposes a 35-km thick section of the crust as it existed at the time of the closure of the Midcontinent Rift (see also chapter 7 of this guidebook). The monocline extends almost 100 km along strike, from Gogebic County in northern Michigan to Bayfield County, Wisconsin (Schmidt, 1976). The regional northerly tilt of the strata is thought to have been caused by south-directed post-rifting reverse displacement on a system of lithosphere-scale, north-dipping listric faults, including the Atkins Lake-Marenisco and Pelton Creek Faults, which may have formed initially as normal faults during the development of the rift (Cannon et al., 1993). Because the rift-filling lavas and sediments formed a much thicker stratigraphic section to the north of these normal faults, the subsequent reactivation of the normal faults as reverse faults caused the Mesoproterozoic rift-related rocks to be thrust over older, Archean and Paleoproterozoic rocks (in contrast to the usual older-over-younger stratigraphic pattern caused by reverse faulting in a simple 'layer cake' sedimentary sequence). The Atkins Lake-Marenisco fault cuts higher in the crustal section on its western terminus, where Paleoproterozoic rocks form the footwall, than in the east, where it transects Archean crust (Cannon et al., 2008). Apart from the major faults bounding the Montreal River Monocline on the south, there is no evidence for significant stratigraphic disruption or repetition within the exposed section in north-central Wisconsin. Therefore, the steeply dipping strata of the monocline provide a complete transect of the crust at ca. 1000 Ma.

The listric shape of the Atkins Lake-Marenisco fault seems well constrained by the outcrop patterns in the hanging wall (north side): the northward-shallowing, northerly dips of Proterozoic strata, and the fanning pattern of southward dips seen in (originally vertical) Midcontinent rift-related dikes intruded into Archean rocks (Fig. 2) (Cannon et al., 1993, 2008). Also, in the eastern half of the monocline, where the fault cuts through Archean basement rocks, there is a well-defined biotite Rb-Sr age front as one approaches the fault from the north. The Archean rocks north of this line preserve Archean and

Paleoproterozoic (Penocean) ages, showing that they have remained below the 270° C Rb-Sr blocking temperature for biotite since at least early Proterozoic time. South of the front, the Archean rocks yield anomalous Mesoproterozoic Rb-Sr ages (ca. 1040-1060 Ma), indicating that they cooled through the blocking temperature at the time of closure of the Midcontinent rift and were brought up from mid-crustal depths by rotational displacement along the curved surface of the Atkins Lake-Marenisco fault (Cannon et al., 1993; Holm et al., 2007). The Rb-Sr age front thus represents an exhumed Mesoproterozoic isotherm. These regional relationships indicate that the Atkins Lake-Marenisco fault was a major structure that accumulated many kilometers of reverse displacement in Mesoproterozoic time (Aldrich, 1929; Cannon et al., 1993, 2008).

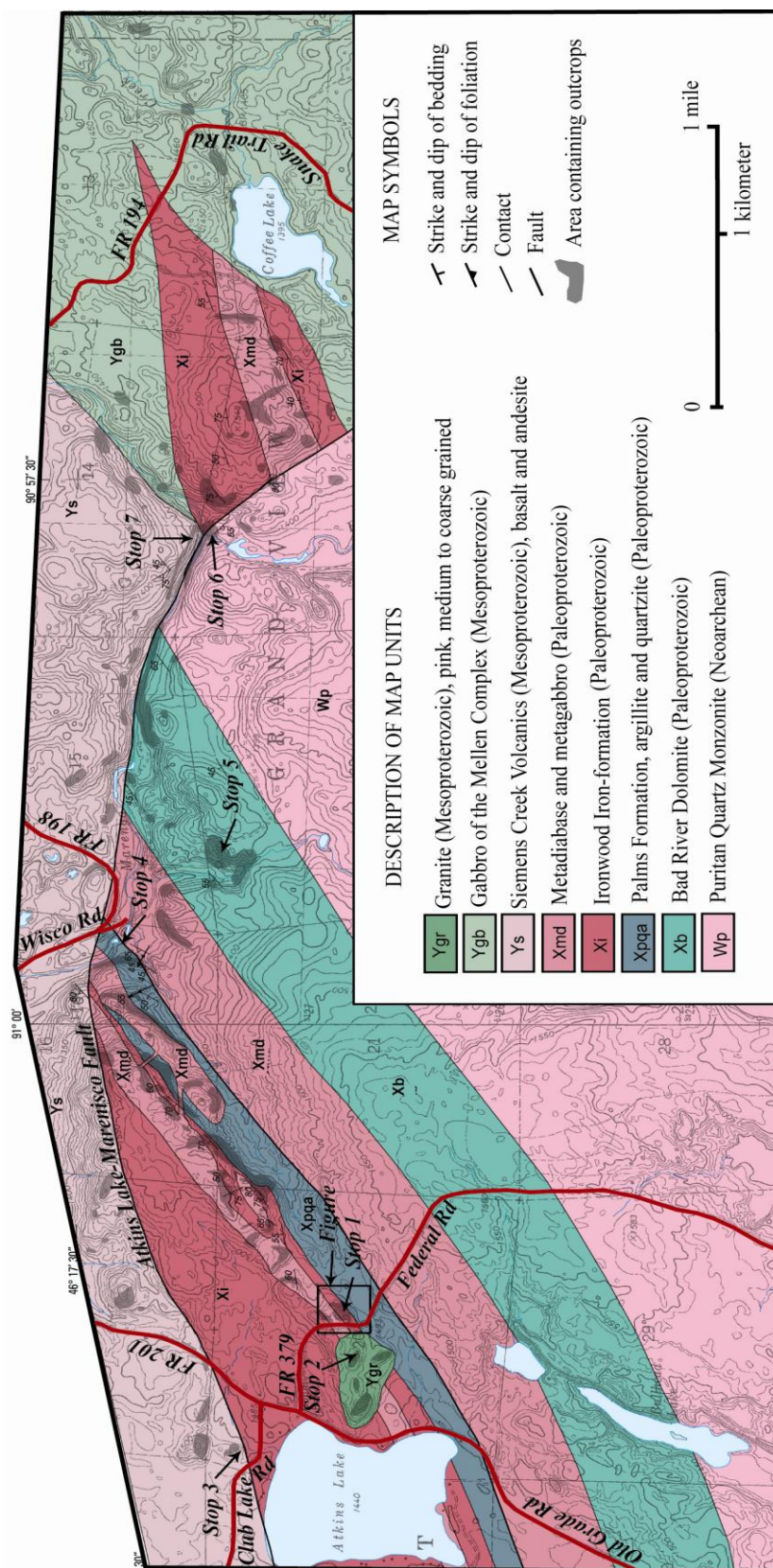


Figure 1. Geologic map of the Atkins Lake-Marengo Falls area showing the location of field trip stops. Geology from Cannon et al. (1996)

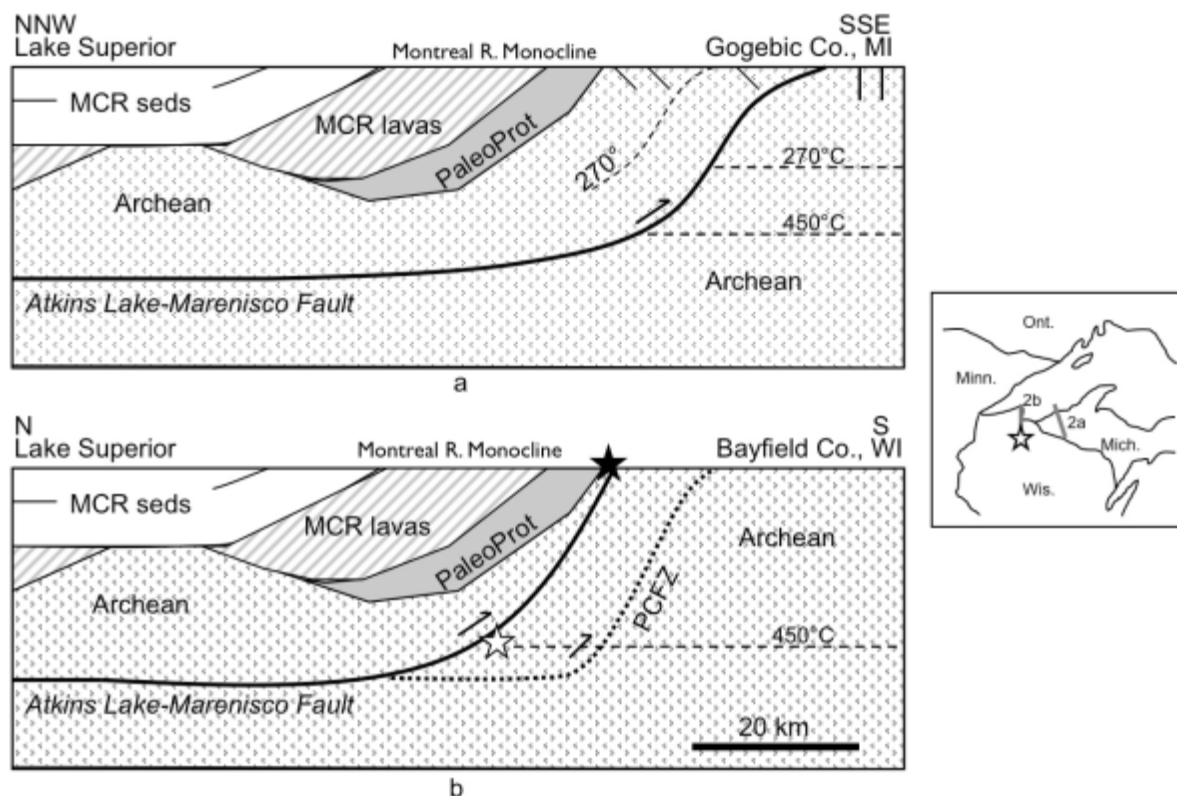


Figure 2. Schematic cross sections through the Montreal River Monocline in northern Michigan and Wisconsin (Modified from Cannon et al., 1993). Sections (a) and (b) depict eastern and western parts of the monocline, respectively, as indicated in the inset. The major reverse faults originated as rift-bounding normal faults, creating a thicker stratigraphic section to the north. MCR = Midcontinent rift; Paleoprot = Paleoproterozoic metasedimentary rocks; PCFZ = Pelton Creek Fault Zone (inferred). Dashed lines indicate the positions of isotherms prior to closure of the rift. Thin lines in Archean rocks in upper section represent orientations of Midcontinent rift-age mafic dikes. Filled star shows location of Marengo Falls (Stops 6, 7); open star shows inferred depth of formation of pseudotachylytes at Stop 6.

Stop 1. Paleoproterozoic Ironwood Iron-formation, metadiabase sills, and breccias

(Text modified from Klasner and LaBerge, ILSG Field Guide 1996)

Numerous exposures of the Ironwood Iron-formation and the Palms Formation, both part of the Menominee Group, occur in a 4,000 m long northeast-trending zone from Atkins Lake to the Marengo River (Fig. 1), where the Early Proterozoic units are truncated by the Atkins Lake-Marenisco fault. The Palms Formation consists of an upper massive quartzite member and a lower argillaceous quartzite member, similar to exposures elsewhere on the Gogebic Range. (Due to time limitations, we will not visit exposures of the Palms Formation at this stop.) The Ironwood Iron-formation consists of magnetic wavy-bedded and laminated cherty iron-formation and argillite, some of which is also magnetic. In places, mafic igneous bodies—either sills or flows—occur in contact with iron-formation.

A group of closely spaced outcrops along a ridge east of Atkins Lake (N46.27904, W 91.02187) shows typical metamorphosed Ironwood Iron-formation, unusual intraformational breccias, and gabbro sills (Fig. 3). These outcrops are a good representation of the Ironwood in the western extremity of the Gogebic Range, and of the igneous and tectonic overprints that are more pronounced here than in the central part of the range.

At this stop, enigmatic breccia units are spatially associated with the upper and lower contacts of a concordant mafic body in the Ironwood Iron-formation (Fehrer and Flood, 1995). These breccias contain angular to sub-rounded clasts of banded siliceous sedimentary rock, ranging from cm to several m in length, within a matrix of a fine-grained, massive mafic igneous rock. There is a marked gradation in grain size in the mafic igneous material, from basalt-like at the contacts with the iron-formation, where the breccia occurs, to diabasic and gabbroic away from the contacts (Fig. 3). Clasts within the breccia are mainly siliceous argillite; a minority is magnetic. Puzzlingly, the population of clasts appears to be more siliceous than the rock in the immediately adjacent iron-formation. Most clasts, regardless of size and composition, are elongate and tabular. Bent or folded clasts are not uncommon. The argillite clasts consist of alternating layers of recrystallized quartz, fine-grained chlorite, biotite or actinolite, garnet, and variable amounts of magnetite. The fine-grained mafic matrix contains conspicuous vesicles, especially around clasts. The matrix is composed of medium-grained amphiboles and plagioclase laths, probably a relict ophitic texture. Plagioclase phenocrysts are present locally. Limited geochemical analysis indicates a tholeiitic composition for the matrix. It should be noted that the mineralogy and, to some extent, the textures of these rocks have likely been modified by Mesoproterozoic contact and regional metamorphism.

The origin of the breccias is problematic. Formation by debris flows, faulting or impact ejecta can be ruled out because the matrix is igneous and not fragmental. Outcrop-scale relationships are arguably consistent with either intrusive or extrusive magmatic activity. The presence of breccia units at both the upper and lower margins of the igneous body seems most simply interpreted as evidence that they formed during intrusion of a sill. The presence of angular, and locally deformed, blocks of country rock within mafic material could be evidence for intrusion into semi-consolidated sediments at shallow depth -- or could indicate that the clasts were hot enough to deform ductilely in a mobile magma that was shearing against the walls of more rigid host rock. Alternatively, the breccia and mafic rocks could represent an extrusive basaltic flow that incorporated 'rip-up' clasts from the sediments over which it flowed. Sometime later, this stratigraphic horizon may have become the site for the concordant intrusion of a mafic sill that separated the clast-bearing basalt flow into two parts. Features consistent with this origin include: 1) perfectly concordant contacts with thinly laminated sediments at several outcrops, and the lack of any observed cross-cutting contacts; and 2) the marked textural contrast between the fine-grained breccia matrix and presumably intrusive coarse-grained, inclusion-free metadiabase and gabbro. Whether the breccias are the result of extrusive or shallow intrusive igneous activity, they do indicate that mafic magmatism occurred during or soon after iron-formation deposition in the western part of the Gogebic Range.

Field relationships at this site are similar to those in the eastern part of the Gogebic Range, where the Emperor Volcanic Complex is interlayered with the Ironwood Iron-formation (Klasner et al., 1998). In both areas, platformal sediments, including the Bad River Dolomite, the Palms Formation, and the Sunday Lake Quartzite in the east, are similar to

equivalent rocks in the central part of the Gogebic Range. However, evidence for igneous activity during the time of deposition of the Ironwood Iron-formation is much scarcer in the central part of the range, limited to a few minor ash beds (R. G. Schmidt, personal communication, 1995). The larger volume of igneous rock on either end of the Gogebic Range indicates a far more volcanically active environment in these regions.

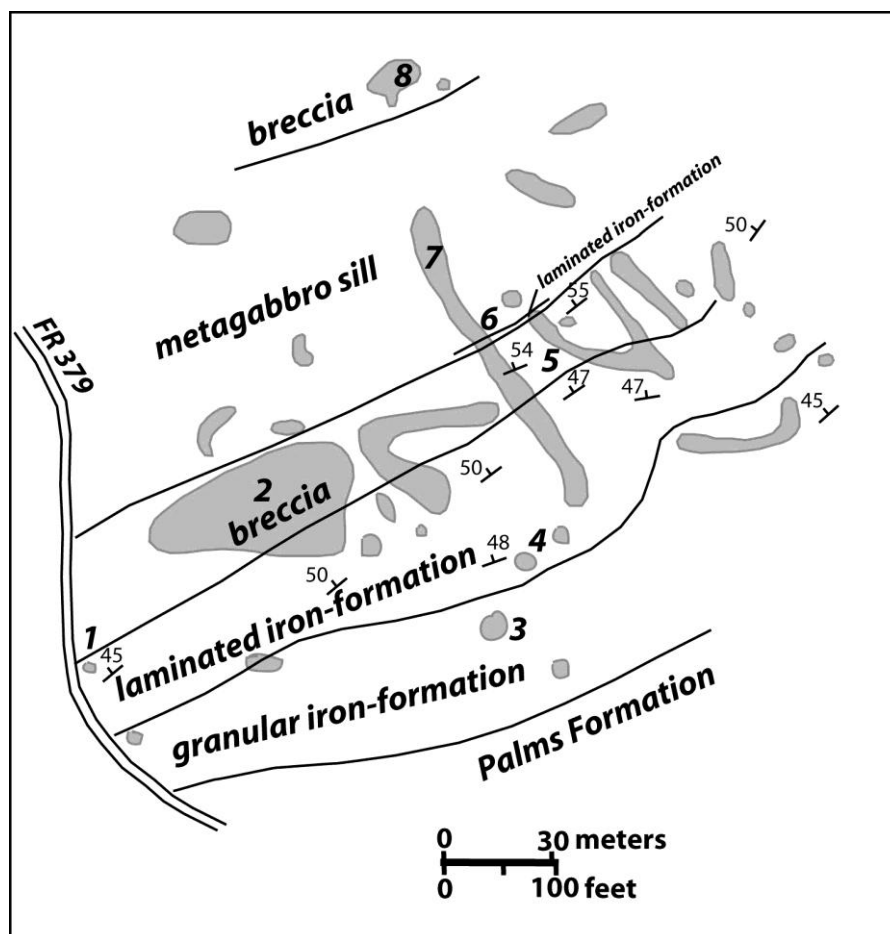


Figure 3. Detailed map (Klasner and LaBerge, unpublished) of the area east of Atkins Lake showing outcrops and geologic relationships among iron-formation, breccia units, diabase and gabbro. Descriptions of the numbered localities are provided below.

Locality 1. This outcrop consists of layered magnetic, greenish, actinolitic iron-formation with metadiabase, and a breccia zone with clasts of iron-formation in a fine-grained metadiabase matrix. Bedding orientations at this outcrop are consistent with orientations of bedding elsewhere in the area, indicating that the outcrop is in place and not a detached raft of iron-formation within the igneous rock. *Walk northeast about 60 m to the top of a hill.*

Locality 2. At this location, large thin slabs of argillite lie in various orientations in a matrix of locally vesicular metadiabase. Some of the slabs are contorted or folded, and one is rolled into a ball with a central layered portion surrounded by a concentric zone of smaller

clasts in the metadiabase matrix. Slabs range in length from less than 1 cm to more than 1 m. *Walk east along the crest of the hill about 90 m to a small NNW-trending valley. When in the valley, follow the edge of the hill toward the SSW for about 60 m.*

Locality 3. Strongly magnetic, wavy-bedded, granular cherty iron-formation. Here lenticular chert beds are up to 15 cm thick. This bedding style is typical of wavy-bedded iron-formation farther east along the Gogebic Range. *Walk north-northeast along edge of hill about 25 m.*

Locality 4. Thin-bedded, non-magnetic chert-argillite or argillaceous iron-formation. Some cherty layers appear boudinaged. The argillite unit is about 50 m thick and grades into a more argillaceous unit with less cherty layers to the north. *Walk northeast about 55 m.*

Locality 5. The top of the argillaceous iron-formation layer is in sharp contact here with the overlying breccia unit, in which argillite clasts are surrounded by metabasalt matrix. Clasts in the breccia tend to become larger higher in the section (northward from the contact). Remnants of what appears to be a thin slab of iron-formation lie along the top of the breccia unit both here and at locality 6. *Walk east-southeast about 10 m.*

Locality 6. Exposed here is the southeast continuation of the thin slab of iron-formation that lies along the contact between the breccia on the southwest and massive metagabbro on the northeast. *Walk northeast about 50 m.*

Locality 7. Massive metadiabase/metagabbro. *Walk northwest about 35 m.*

Locality 8. Contorted clasts, mainly argillitic, in a vesicular metabasalt matrix.

Stop 2. Mesoproterozoic (?) Mellen-type granite

A small undated granite body is exposed in several outcrops east of Atkins Lake, in the open woods south of Forest Service Road 379 (N46.27739, W91.02542). The granite is pink and white, with a medium- to coarse-grained equigranular to porphyritic texture. Principal minerals are potassium feldspar, plagioclase, biotite, and quartz. The granite intrudes Paleoproterozoic metasediments of the Ironwood Iron-formation, and the lack of metamorphic effects and or deformational fabric within the granite suggests that it is Mesoproterozoic in age.

This granite is very similar in appearance to the well-known Mellen granite, which is a phase of the Mellen Intrusive Complex, a major magmatic unit of the Midcontinent Rift exposed over a large area more than 20 km to the northeast of this site. In addition to the Mellen granite, the Mellen Intrusive Complex is made up of two large, layered gabbroic bodies -- the Potato River and the Mineral Lake intrusions -- and several related igneous bodies, including the peridotitic Rearing Pond intrusion, and a group of gabbro and granophyre sills. A granophyre within the Mineral Lake intrusion has yielded a U-Pb zircon age of 1102.0 ± 2.8 Ma (Zartman et al., 1997). An indistinguishable U-Pb zircon age (within error) of 1100.9 ± 2.8 Ma has been determined for the Mellen granite, which intrudes and brecciates gabbro of the Mineral Lake intrusion (Zartman et al., 1997).

Stop 3. Mesoproterozoic Siemens Creek Volcanics contact metamorphosed to pyroxene hornfels and granophyre(text from Cannon et al., *ILSG Field Guide* 1996)

The Siemens Creek Volcanics, the lowermost formation within the Powder Mill Group, is the oldest volcanic unit in the Midcontinent Rift in northern Wisconsin. It consists of a sequence of flood basalts erupted as relatively thin flows. These Keweenaw rocks are thrust southward over Paleoproterozoic rocks (such as those seen at Stops 1, 4, and 5) along the Atkins Lake-Marenisco fault (Figs. 1 and 2). The northward tilt caused by the faulting has exposed the basal contact of the Mineral Lake intrusion together with a structurally lower sequence of intruded and metamorphosed volcanic rocks. At this stop (N46.2805, W-91.0268) an unusually highly metamorphosed section of the Siemens Creek Volcanics is exposed on an isolated hill about 100 m north of Forest Service Road 210 (Club Lake Road), just north of Atkins Lake. The Atkins Lake-Marenisco fault lies under the low ground between the road and the hill. No indications of faulting have been found in the outcrops we will examine, but other low outcrops just south of the road (on private land) contain zones of markedly sheared basalt.

The Siemens Creek Volcanics at this locality are intensely recrystallized to pyroxene hornfels. Most original textures are obliterated. Locally, textures possibly representing relict amygdular flow tops can be seen, but these occur discontinuously, making it difficult to recognize distinct lava flows. Small bodies of pink, granophyric granitic rock are common and appear to represent segregations of magma generated by partial melting of the basalt during high-grade metamorphism. These intensely metamorphosed and partly melted rocks can be traced for about 10 km along strike, beyond which they grade both east and west into less metamorphosed equivalents. This unit is up to about 1 km thick and is bounded on the north by the western extension of the Mineral Lake intrusion and related intrusive rocks. Although it is tempting to ascribe the intense metamorphism at this stop to contact effects of the Mineral Lake intrusion, the grade and geographic extent of the metamorphism seen here are much greater than those in the border zone of the gabbro in other areas. Additional conductive heating during the time that these rocks resided deep in the crust, at the base of the thick sequence of lavas and sediments that filled the Midcontinent Rift, may have contributed to the unusually high metamorphic grade.

Stop 4. Penokean shear zone in Paleoproterozoic Palms Formation, Marengo River

On the south bank of the Marengo River immediately downstream from the private bridge off Forest Road 198 (N46.28965, W90.99405), laminated argillite of the lower part of the Palms Formation is exposed in several subdued outcrops. The rockshere have been intensely sheared within a NE-trending zone that is at least 60 m wide. This shear zone, presumed to be of Penokean age, is essentially bedding parallel and now strikes 50° NE and dips 50° NW. It separates a thick metadiabase sill in the footwall from the upper Palms Formation and thinner metadiabase sills in the hanging wall. At this site, bedding in the Palms is tightly to isoclinally folded and transposed into a mylonitic foliation. The folds are commonly dismembered, with limbs so strongly attenuated that only relict fold hinges can be seen. These small folds typically plunge about 50 degrees to the northwest and their axes lie within the plane of the mylonitic foliation. Some of the relict fold hinges apparently acted like rigid porphyroclasts during shearing and display sigma-type tails that indicate dextral shear (in the present orientation).

We interpret this fault to have formed as a bedding-parallel thrust during the Penokean Orogeny. The current northwestward dip of the Paleoproterozoic strata and the fault are a result of regional tilting that occurred during inversion of the Midcontinent Rift at about 1060 Ma. If the strata are restored to their attitude prior to this rotation, they are nearly flat lying and the fault likewise is a nearly flat surface across which the upper plate was thrust northward. Thus, the shear zone seen here is, to our knowledge, the best exposed example of a Penokean thrust fault in this part of the Lake Superior region.

Stop 5. Paleoproterozoic Bad River Dolomite: Stromatolites, Penokean deformation, contact metamorphism, Grandview Quarry, Marengo River

Please note that Stops 5-7 are on private land owned by the S.C. Johnson family, who have been generous in making these outcrops accessible to geologists for educational purposes. Steep slopes and loose rocks at these sites pose hazards, and field trip participants are responsible for their own safety.

The oldest Paleoproterozoic unit in the western Gogebic Range, the Bad River Dolomite of the Chocolay Group, is exposed here in an abandoned quarry that was developed on a natural glaciated bedrock outcrop (N46.28509, W90.98608). In this area, the Bad River Dolomite lies nonconformably and discontinuously on top of the Neoproterozoic Puritan Batholith (Stop 6), but in the eastern Gogebic range it overlies the Sunday Quartzite, interpreted as a shallow marine tidally influenced unit. At this site, the Bad River Dolomite strikes approximately N50°E and dips 50°-60° NW, concordant here with the overlying Palms Quartzite (Stop 4), but regional relationships indicate that the contact with the Palms is disconformable (Cannon et al., 2008).

The Bad River Dolomite is a fine-grained chert-bearing dolostone in which massive, meter-scale strata give way up-section to laminated dolostone with horizons of silicified stromatolites. The stromatolites are well exposed in cross section on the glacially scoured surface at the rim of the quarry and are typically 20-30 cm in diameter. Along the Marengo River, the Bad River Dolomite experienced metamorphism as a result of the emplacement of the nearby Mineral Lake intrusion, and radiating bundles of pale green tremolitic amphibole can be seen in and around chert nodules and within stromatolitic layers, where silica and dolomite are interbedded at the centimeter scale. These relationships record the classic calc-silicate reaction: $\text{Dolomite} + \text{SiO}_2 + \text{H}_2\text{O} \rightarrow \text{Tremolite} + \text{Calcite} + \text{CO}_2$, which occurs either 1) in the presence of very CO_2 -rich fluids under low pressure conditions (ca. 1-2 kbar) or 2) over a wider range of fluid compositions at higher pressures (>3 kbar) (Spear, 1993). The isotropically radiating clusters of tremolite indicate nucleation-limited growth and static metamorphism (i.e., under low deviatoric stress). In places the tremolite is very fibrous, even asbestiform. Fine mineralogical specimens can be found in the waste rock piles in the old quarry. The quarrying also exposed several meter-scale Keweenaw dikes that cut discordantly across the tilted beds.

In a detailed geochemical and isotopic study of carbonate rocks from the Chocolay Group across the Lake Superior region, Bekker et al., (2006) found that the Bad River Dolomite (where it is less metamorphosed) has a $\delta_{13}\text{C}$ signature distinct from that of other Chocolay Group carbonates (e.g., the Kona Dolomite in the Marquette Range and the Gordon Lake Formation in Ontario). They concluded that the Bad River is slightly younger than these units, deposited after the early Proterozoic (Gowganda) glaciations but just before the

global biogeochemical upheaval recorded by the large positive carbon isotope excursion known as the Lomagundi event.

Stop 6. Marengo Falls, S side: Neoarchean Puritan Batholith; fault breccias, cataclasites and pseudotachylyte along the Atkins Lake-Marenisco Fault

The deeply incised valley of the Marengo River in southeastern Bayfield County, Wisconsin (sections 14 and 15, T44N, R5W) provides a 0.5 km transect through the Atkins Lake-Marenisco fault zone and is one of the few places within the Midcontinent Rift where rocks along a major rift-closing fault are well exposed (Fig. 4). At the falls of the Marengo River (N46.286, W90.964), the Puritan Batholith is juxtaposed across the Atkins Lake-Marenisco Fault with the Ironwood Iron-formation and Paleoproterozoic diabase sills, and also with both basaltic lava flows (Siemens Creek Volcanics) and a layered mafic complex (the Mineral Lake Intrusion) of the Midcontinent Rift. Based on regional outcrop relationships, discussed in the Introduction, the vertical throw on the Atkins Lake-Marenisco Fault is probably on the order of 15 km (Fig.4).

The river flows northward over Archean rocks until it intersects the fault zone, marked by a 15 m high waterfall near the southern boundary of section 14 (Fig. 4). At that point, the river bends sharply to the west-northwest, subparallel to the fault zone, and cuts through rocks in the core of the zone. Along this transect, the Puritan Quartz Monzonite can be traced from an undeformed condition above the falls to an ultracataclastic state a few hundred meters downstream. Rocks exposed between these two locations preserve intermediate textures that record the temporal evolution of the fault zone (Bjørnerud 2010a).

The nearly undeformed Puritan Quartz Monzonite is a medium- to coarse-grained (0.5-2.0 cm), locally pegmatitic granitoid with consertal igneous texture in thin section. It is comparatively poor in mafic minerals and those present (mainly biotite, some hornblende) are largely altered to chlorite. The feldspars, principally albitic plagioclase and microcline, are commonly sericitized, and this becomes more pronounced near the fault zone. Away from the fault, feldspars show little evidence of internal deformation, although in places where a weak gneissic fabric is developed, a fine-grained (0.1 mm) polygonalized texture indicates partial annealing following ductile deformation, presumably during Neoarchean time (Schmidt, 1976).

The earliest fault-related microstructures, found in rocks immediately downstream from Marengo Falls, are quasi-ductilely deformed albite and microcline grains with kinked or bent twin planes and patchy extinction. In some samples, very small (ca. 0.05 mm) subgrains have developed along the most deformed parts of these crystals. These textures are overprinted by more brittle features. In the steep cliffs on the south side of the river, the quartz monzonite has been broken into heterogeneous breccias and cataclasites. These rocks lack any prominent planar fabric, slickenlines or other indications of slip sense; they are simply pervasively fragmented.

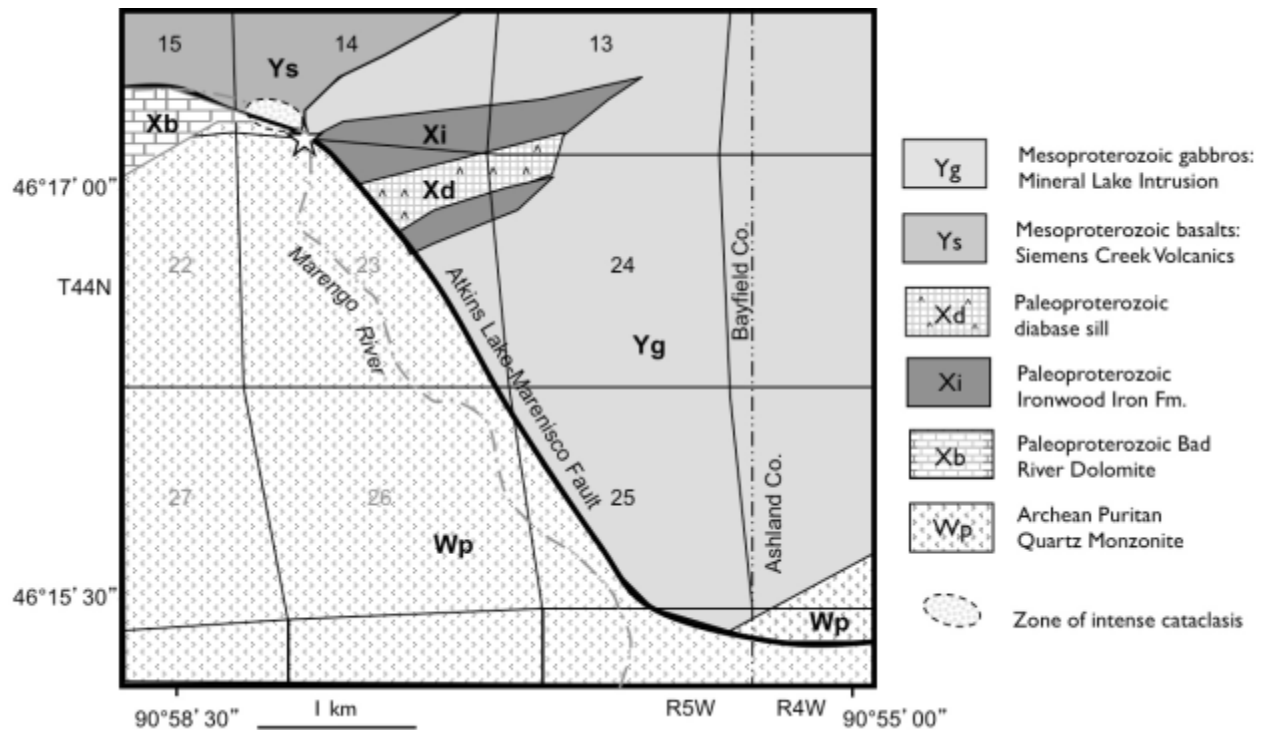


Figure 4. Geologic map of the Atkins Lake-Marenisco Fault zone in the Marengo Falls area, Bayfield County, Wisconsin (modified from Cannon et al., 2008). Star marks location of Marengo Falls (Stops 6 & 7). The location of Stop 5 is just west of the area marked Xb. The non-parallelism of section lines reflects the difficulty early surveyors had in mapping areas with extensive outcrops of magnetite-bearing iron-formation.

In a series of outcrops of the Puritan quartz monzonite on the south side of the river, extending about 200 m immediately west from an abandoned hydroelectric station, breccias and cataclasites are complexly interlaced with very dark veins, 0.2 to 6 cm thick, of what appears to be devitrified (mainly chloritized) pseudotachylyte (Bjørnerud, 2007, 2010a). Several types of textural evidence support this interpretation. First, the chlorite in this material is texturally isotropic, occurring as uniformly spaced radial bundles 0.1-0.2 mm in diameter. This suggests that it nucleated in a homogeneous and isotropic medium – i.e., glass or on primary microlitic spherules in glass sometime after the deformation had ceased. Second, this chloritic material typically includes relatively large (0.5-2 mm), angular clasts from the protolith (including quasi-ductilely deformed feldspar grains), while finer clasts are conspicuously absent. The absence of fines is consistent with preferential melting of smaller rock fragments, a phenomenon documented in natural and experimental pseudotachylytes (e.g., Tsutsumi, 1999). Third, the contacts between some of the chloritic material and the host rock are distinctly scalloped, a partial melting texture observed in pseudotachylyte injection veins (Swanson, 2006). Fourth, in some of the thicker veins, a disproportionate number of the surviving clasts are embayed grains of titanite, a refractory accessory mineral in the protolith that could have withstood temperatures sufficient to melt the major minerals (quartz and alkali feldspar). The irregular, cusped edges of these crystals are again suggestive of partial assimilation by a melt phase. Finally, some of the chloritic material

preserves ‘pseudomorphs’ of apparent flow textures similar to those observed in younger, still-glassy pseudotachylytes. Collectively, these observations support the interpretation that much of the black chloritic material in the fault zone originated as pseudotachylyte.

The spatial relationships among the various types of fragmented rock and the pseudotachylytes shed light on the dynamic processes in the fault zone over time. In thin sections of some pseudotachylyte-bearing samples, a continuous gradient in grain size can be observed from cataclasite to ultracataclasite to pseudotachylyte, suggesting that the textures formed in a single slip event. More commonly, though, fine-grained cataclasites or microbreccias transect and/or include scattered pieces of the pseudotachylytic material, indicating that these granulated rocks are generally younger than the pseudotachylytes. One group of samples shows two distinct generations of pseudotachylyte, which in turn are both incorporated as clasts in a cross-cutting breccia.

Some of these microbreccias appear to have dilational and intrusive (fracture Mode I), rather than shearing (Mode II), cross-cutting relationships with the pseudotachylyte and the surrounding rock. As such, they are not cataclasites in the strict sense of having been produced only by grinding and frictional wear. They occur as dike-like fingers that ‘explosively’ disaggregate the pseudotachylyte and make inroads into fractures in the host rock. These microbreccia veins are typically matrix- rather than clast-supported and contain parallel bands of uniform grain size, commonly fining inward from wall to interior, possibly a result of dynamic sorting of fluidized granular material. Such textures may record thermal pressurization of fluids during a seismic event (Otsuki et al., 2003; Bjørnerud, 2010b).

Stop 7. Marengo Falls, N side: Ultracataclasite along the Atkins Lake-Marenisco Fault

The most extreme comminution of grains has occurred in rocks exposed in prominent bluffs on the north side of the river. In outcrop, these ultracataclasites are dark gray, massive, and transected by anastomosing veins of chlorite, quartz and zeolites. In thin section, one can discern very fine (<0.1 mm) angular feldspar grains in a still finer matrix of chlorite with a weak preferred orientation. These rocks are so modified by deformation and mineralization that their protolith is unclear; they are likely derived from the basaltic lava flows of the Siemens Creek Formation, which crops out to the north of the river below the falls (Fig. 4). In the very first geologic description of these rocks, C.E. Wright (1879), who mapped the Penokee Gap under the direction of T.C. Chamberlin in 1876, noted their unusual character, giving them the non-genetic designation “chloro-silicious rocks” (Fig. 5).

These fault rocks are overprinted by multiple generations of vein networks, with a wide array of orientations, indicating a protracted period of fracture-hosted fluid flow that continued after the main period of faulting and cataclasis had ceased. Many of these veins have a fibrous texture, indicating a ‘crack-seal’ mechanism of formation, and some appear to record almost isotropic dilation of the fault rock. Vein minerals include chlorite, zeolites, quartz, and calcite. These same minerals are common as vein fillings and slickenfibers in Midcontinent Rift basalts in northern Wisconsin (Garbowicz and Bjørnerud, 2003) and in the Upper Peninsula of Michigan, where they occur together with hydrothermal native copper deposits known to have formed between about 1060 and 1050 Ma (Bornhorst et al., 1988).

Summary: The Atkins Lake-Marenisco fault zone of north-central Wisconsin provides a rare window into faulting processes along the flanks of the Midcontinent Rift during its inversion at ca. 1060 Ma. The presence of quasi-ductilely deformed feldspars and cm-thick

pseudotachylite fault veins indicate that faulting began at mid-crustal levels and that the fault slipped in one or more large earthquakes. Cross-cutting cataclasites and matrix-supported microbreccias record progressive comminution and intermittent fluidization of rocks in the fault zone. Elevated fluid pressures almost certainly played a role throughout the life of the fault. Reverse slip on the steeply dipping Atkins Lake-Marenisco fault and other rift bounding faults coincided in time with widespread hydrothermal mineralization, including native copper deposition, in rocks of the Midcontinent Rift. The Atkins Lake- Marenisco fault is a thick-skinned lithosphere-scale structure, and its characteristics support the hypothesis (Cannon, 1994) that large deviatoric stresses were transmitted far inland during the Grenville Orogeny.

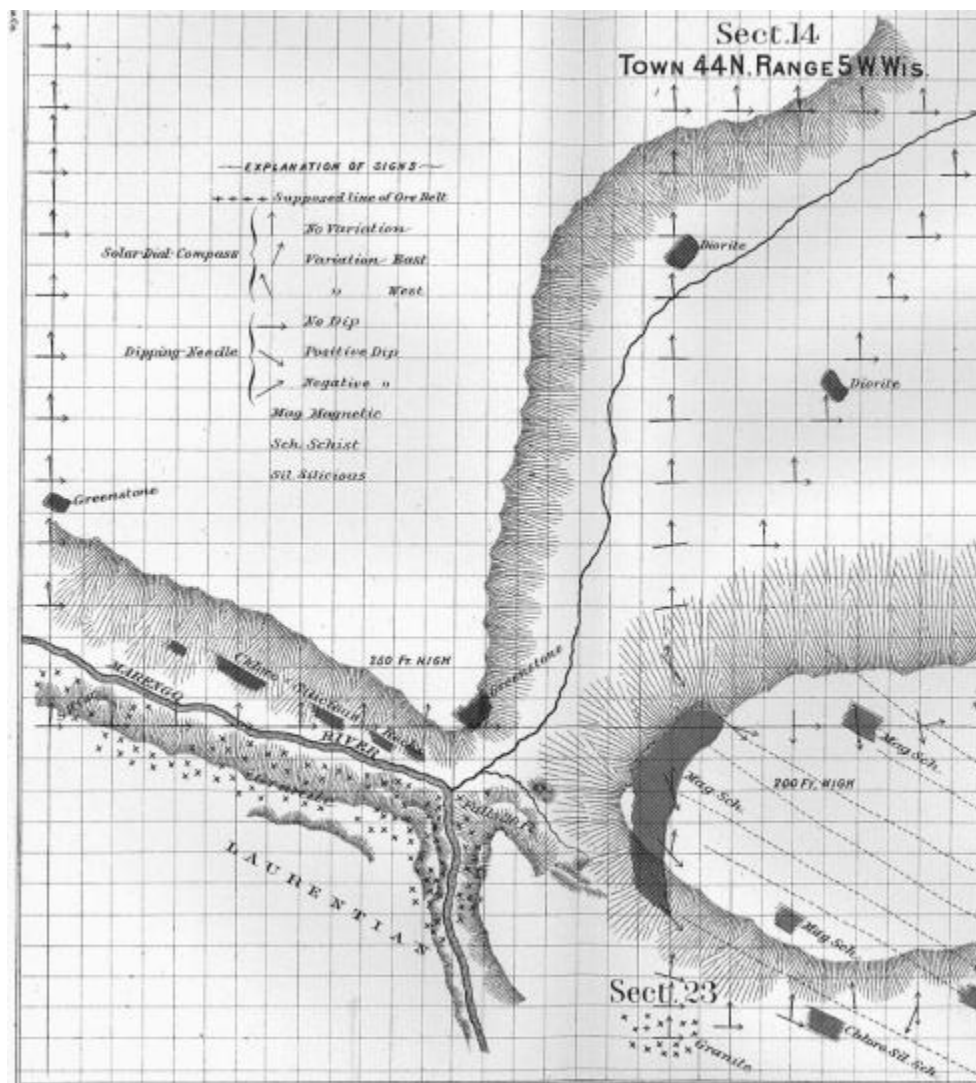


Figure 5. The first geologic map of the Marengo Falls area, by C.E. Wright (1879), based on field work done in the summer of 1876. The ultracataclasites along the Atkins Lake-Marenisco fault zone are mapped as “Chloro-silicious rocks”; the Puritan Quartz Monzonite as “Granite” and Siemens Creek Volcanics as “Greenstone”. Only the strike of the Ironwood Iron-formation (“Magnetic schist”) is inaccurate (shown as NW rather than NE).

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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 3

GEOLOGY OF THE BAYFIELD PENINSULA: KEWEENAWAN BAYFIELD GROUP AND PLEISTOCENE DEPOSITS



Bayfield Peninsula – K. Wilson

Field Trip 3

Geology of the Bayfield Peninsula: Keweenaw Bayfield Group and Pleistocene deposits

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THE KEWEENAWAN BAYFIELD GROUP

Introduction

In the Lake Superior region of the Midcontinent Rift System (MRS), as much as 20,000 m of dominantly basaltic sub-aerial flood lavas are overlain by 10,000 m of sedimentary rocks (Figs. 1, 2) (Behrendt et al, 1988; Cannon et al, 1989). The exposed volcanic rocks range in age from 1109-1086 Ma; the greatest bulk was extruded during a magnetically reversed interval around 1108-1105 Ma, and during a magnetically normal interval around 1097-1094 Ma (Green, 1995).

The overlying, unmetamorphosed, sedimentary rocks, comprise the Oronto Group and the younger Bayfield Group (Fig. 3). Although we will not see the Oronto Group rocks on this field trip, a short description is provided here to set a sedimentological and petrographic background for the Bayfield Group rocks that we will see.

Cessation of volcanism was followed by the deposition of the Oronto Group, which includes the Copper Harbor Conglomerate, the Nonesuch Formation, and the Freda Sandstone (Daniels, 1982). On the Keweenaw Peninsula of Michigan, the Copper Harbor Conglomerate, which is as thick as 1340 m, contains basaltic flows that represent some of the last volcanism within the basin. The Lakeshore Traps in the Copper Harbor have been dated at 1087.2 $\pm$ 1.6 Ma (Davis and Paces, 1990), and an intrusive on Michipicoten Island in eastern Lake Superior has been dated at 1086 Ma (Palmer and Davis, 1987). The Copper Harbor is largely a coarse, clast-supported conglomerate; the clasts are dominantly volcanic and were derived from a variety of volcanic rocks. Sandstones, which are also largely composed of volcanic detritus, are common, especially in the upper half of the formation; generally, they are lithic arenites. Siltstones and shales are less abundant, at least in the exposed portion of the formation, but may be more abundant at depth nearer the axis of the basin. Deposition was on alluvial fans and associated braided alluvial plains; rare stromatolite horizons formed in standing bodies of water during pauses in sedimentation. A probable correlative unit is the subsurface Solor Church Formation of southeastern

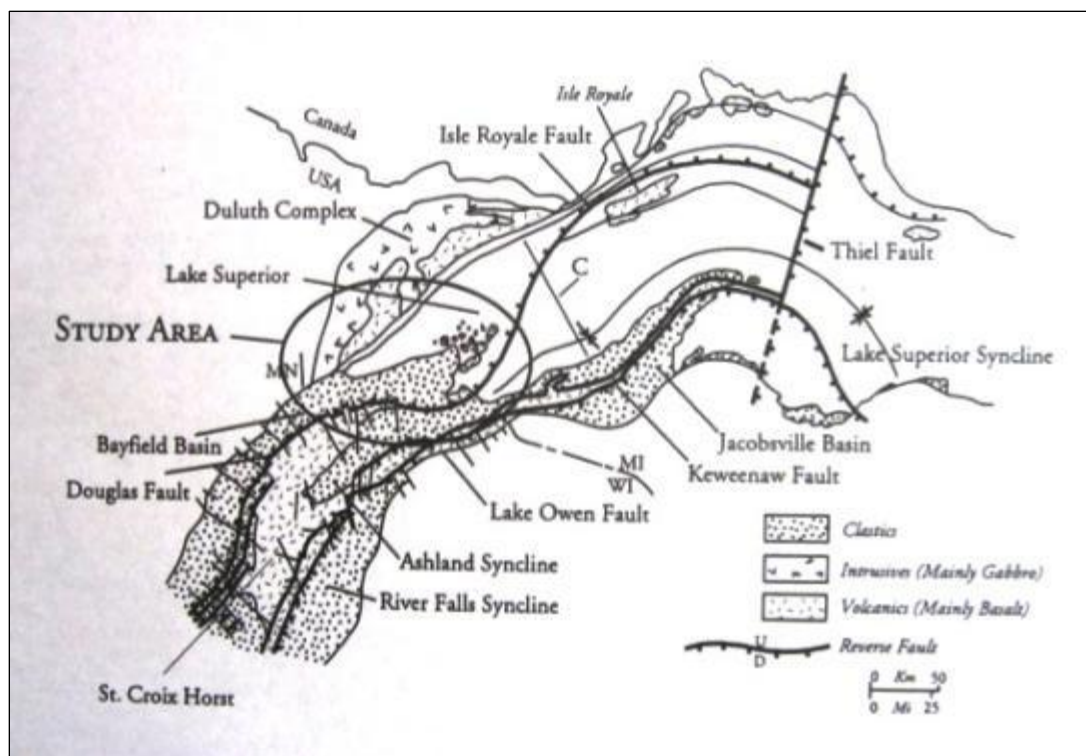


Figure 1. Generalized geologic map of the Midcontinent Rift System in the Lake Superior region showing the study area. From Adamson, 1977, modified from Dickas, 1986.

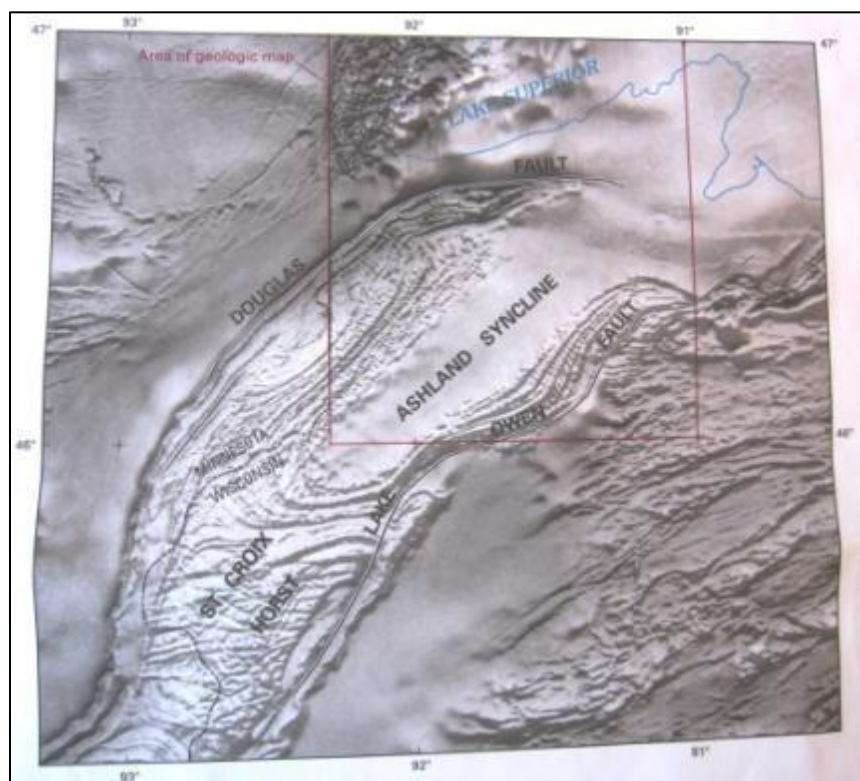


Figure 2. Aeromagnetic map of Northern Wisconsin. From Nicholson et al., 2006.

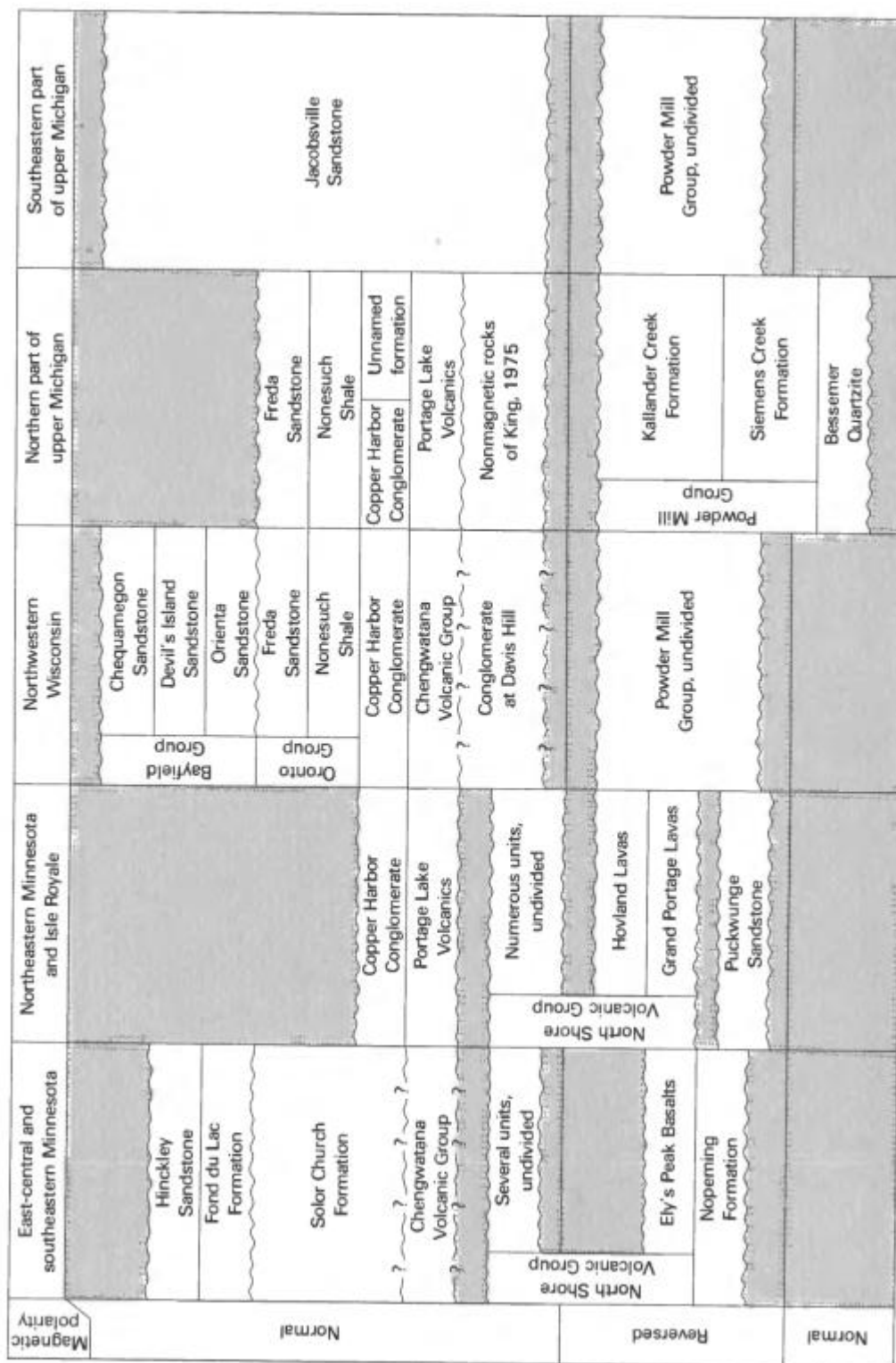


Figure 3. Stratigraphic correlation chart for Keweenaw Supergroup rocks in Minnesota, Wisconsin, and Michigan. From Morey and Van Schmus, 1988.

Minnesota. The Nonesuch Formation, which is as thick as 200 m, is noted for its copper content (mainly chalcocite) and the presence of petroleum seeps in the now-closed White Pine Copper Mine. The Nonesuch is largely a fine-grained sequence of gray to black siltstone with minor thin shales and carbonates as well as local, graded sandstone beds and conglomerates. The fine sediment was deposited in a restricted lacustrine basin, where organic material could accumulate as a result of a reducing chemical environment (Elmore et al, 1988; Suszek, 1995). The Freda Sandstone, as thick as 1525 m, is comprised dominantly of red to buff sandstone with intercalated red siltstone and shale; the sandstones are more mature than those lower in the group. The environment of deposition fits a braided stream model (Fig. 4).

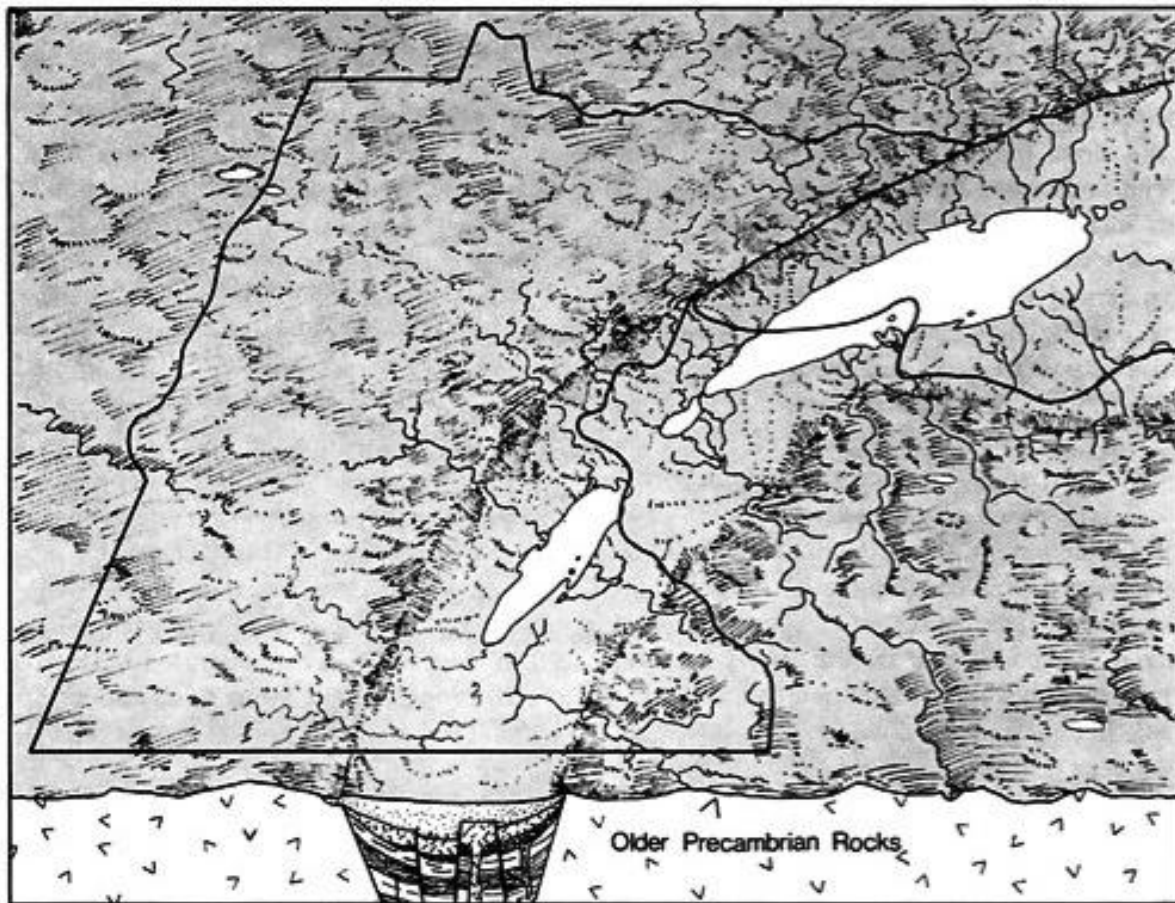


Figure 4. Generalized reconstruction of what the Midcontinent Rift System may have looked like after cessation of volcanism about 1086 Ma. Note the streams entering the rift from the adjacent highlands and the presence of lakes (white). From Ojakangas and Matsch, 1982.

The overlying Bayfield Group, 2100 m thick, is comprised from the base upwards of the Orienta Sandstone, the Devils Island Sandstone, and the Chequamegon Sandstone. All three units are best exposed on the Bayfield Peninsula and the Apostle Islands. Correlative units are found in Minnesota, Michigan, and Ontario (Ojakangas and Morey, 1982). The

contact between the subhorizontal Bayfield Group and the steeply dipping underlying Oronto Group is not exposed, but field and geophysical data can be interpreted as indicating an unconformable relationship (Morey and Ojakangas, 1982). Figure 5 shows the distribution of the three formations in the Bayfield Group.

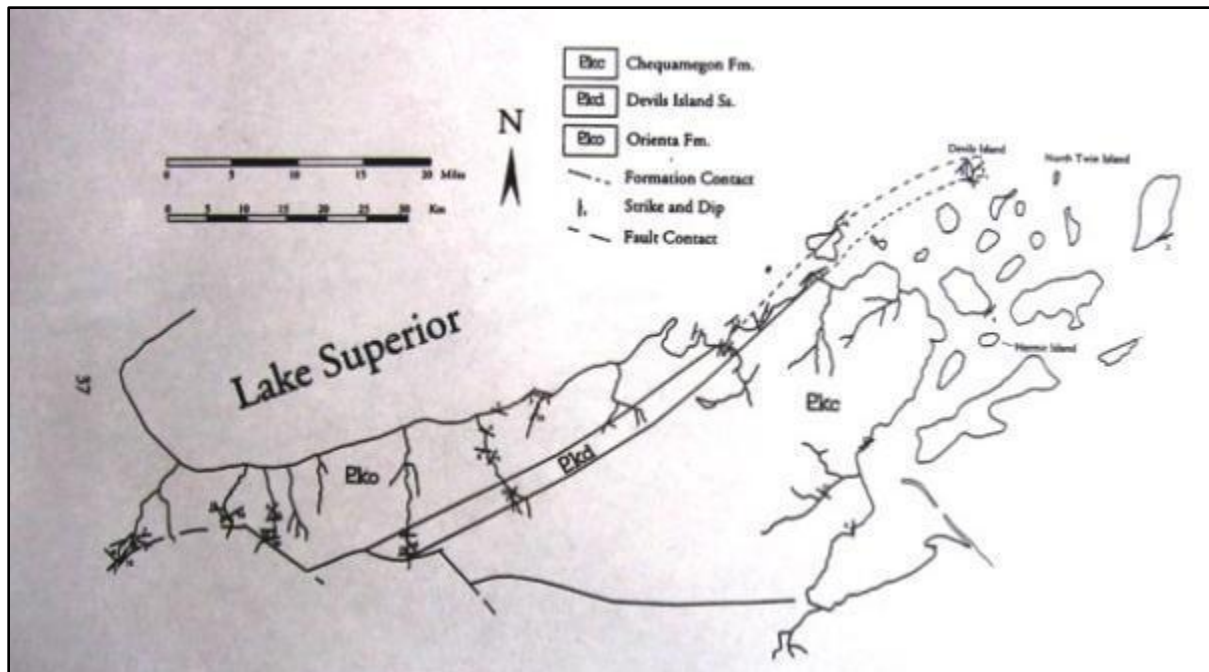


Figure 5. Distribution and attitudes of the three Bayfield Group formations.
From Adamson, 1997.

Orienta Sandstone

The Orienta, and its equivalent several miles to the west in Minnesota (the Fond du Lac Formation), are comprised of red feldspathic-lithic arenite, red siltstone, and shale units with fining-upward sequences that are typical of deposition in meandering river systems; large-scale trough cross-beds are common and parting lineation and current ripple marks are also present.

Devils Island Sandstone

The white, buff, and orange Devils Island Sandstone is about 100 m thick. It is a quartz arenite, as is its equivalent unit, the Hinckley Sandstone, to the west in Minnesota, (Tryhorn and Ojakangas, 1972). Symmetrical ripples suggest a lacustrine environment as the final site of deposition, but eolian processes must have been an important factor in both the textural and the mineralogical maturation of the sand; creating a quartz arenite from feldspathic-lithic sands (Fig. 6). The discovery of adhesion ripples and other eolian features by Johnson et al (2001) in the Hinckley support this suggestion, although the latter works also suggest the presence of a braided stream environment. More recently, Galston and Havholm (2008) have identified sedimentary features that they interpret as eolian dunes, adhesion ripples and wind ripples that were deposited on a floodplain with braided streams. An

equivalent to the east in Michigan may be an upper portion of the Jacobsville Sandstone (Kalliokoski, 1982) as exposed on the Middle Branch of the Ontonogan River on Michigan Highway 28. Quite likely, the maturation of the sand on the vegetation-less surface of that time was due to eolian processes, which are the most effective agents of such maturation.

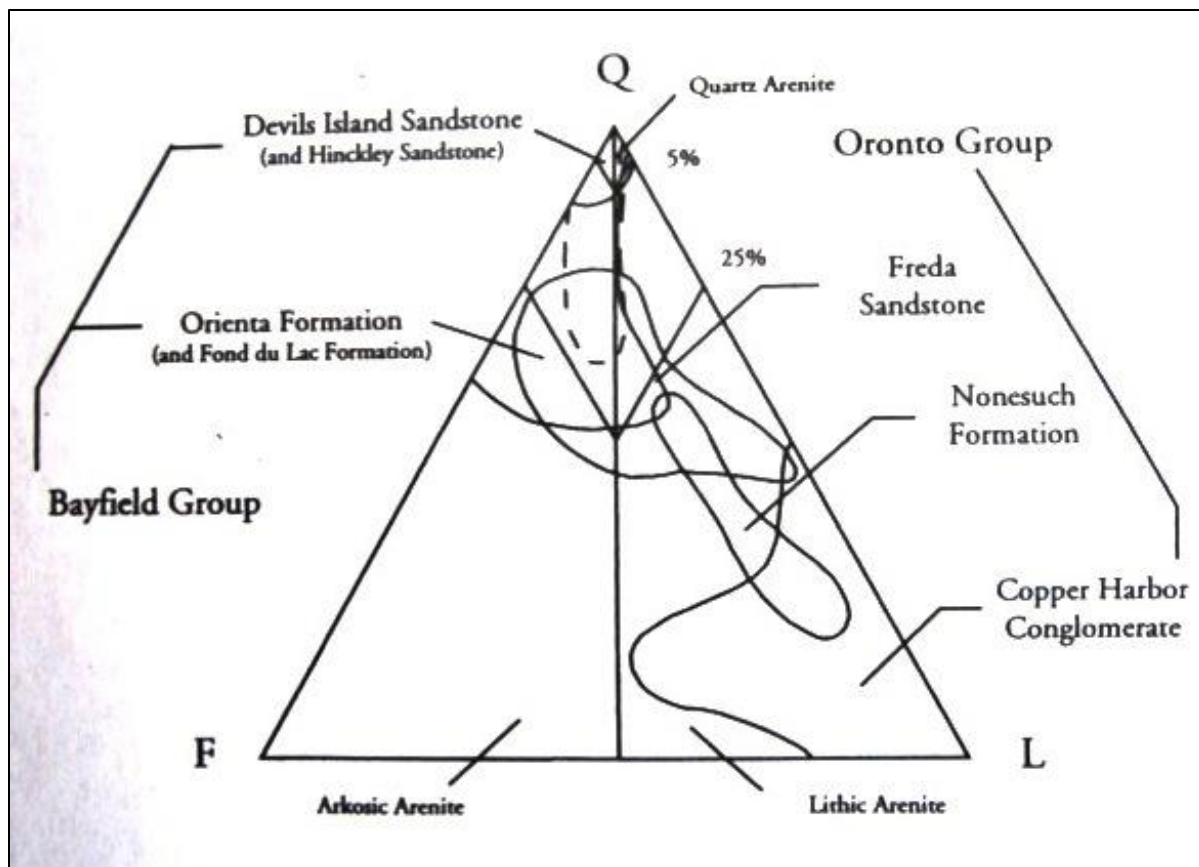


Figure 6. Compositional triangle showing the relative abundances of quartz, feldspar, and lithics (rock fragments) in the formations of the Oronto and Bayfield Groups. The Chequamegon Formation is outlined with the dashed line. Note the increasing compositional maturity upwards from the base of the Oronto Group through the Bayfield Group. From Adamson, 1977, after Ojakangas, 1986.

Chequamegon Sandstone

This feldspatho-lithic unit resembles the Orienta Sandstone, but overlies the Devils Island. The similarity and stratigraphic relationship has long been questioned, largely because the formations are subhorizontal and exposures are scattered. However, Adamson (1997), on the basis of both field and well data, concluded that the Chequamegon does indeed overlie the Devils Island. Another question involving the Bayfield Group is whether the unfossiliferous units could be nonmarine Cambrian deposits (e.g., Hamblin, 1958; Ostrom, 1967; personal communication, A.C. Runkel). Paleocurrent data from several sources show that the dominant currents moved toward the north or northeast, toward the basin axis

in both sedimentary groups (Fig. 7). Whereas intrabasinal volcanogenic detritus is dominant in the Oronto Group, extrabasinal granitic detritus dominates in the younger Bayfield Group.

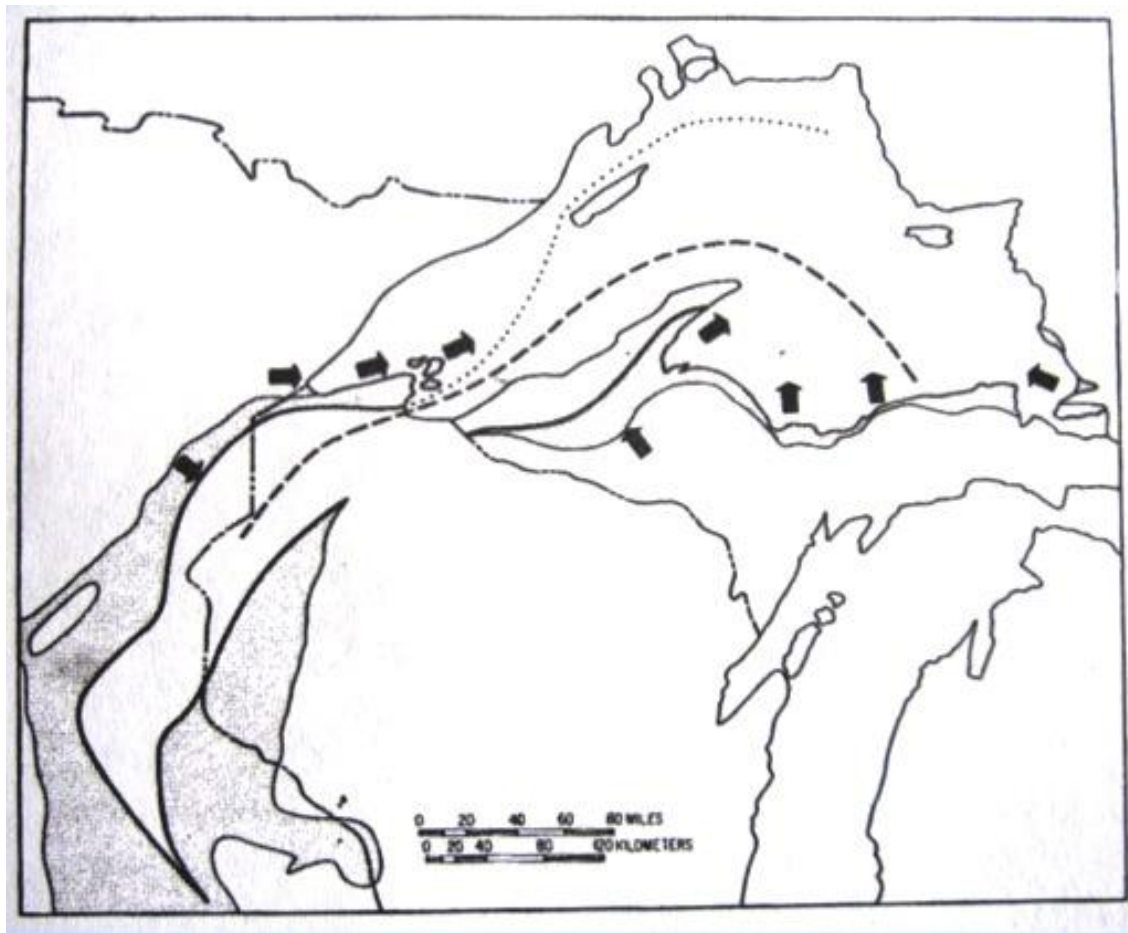


Figure 7. Generalized paleocurrent trends in the Bayfield Group of Wisconsin, the Fond du Lac Formation and the Hinckley Sandstone of Minnesota, and the Jacobsville Sandstone of Michigan and Ontario. The stippled areas in southeastern Minnesota and southwestern Wisconsin are subsurface occurrences. The dashed line represents the axis of the Lake Superior Syncline. From Ojakangas and Morey, 1982.

Beginning in 1869, the Chequamegon was extensively quarried for “brownstone” on the Bayfield Peninsula as well as on several of the Apostle Islands. There were also quarries in the Orienta and in the equivalent Fond du Lac Formation near Duluth; in total, there were 12 quarries. The stone was shipped to Chicago, Milwaukee, Detroit, Toledo, Cincinnati, St. Paul and Kansas City, as well as used locally in lighthouse construction and in buildings such as Washburn’s main street bank building (now a museum) and Northland College’s Wheeler Hall.

Apostle Islands

Because of both time and uncertain weather, it was decided that a boat trip to the Apostle Islands was not possible as an ILSG field trip. However, we have included a short summary of the islands. (Take an excursion boat trip on your own some day from Bayfield.)

Why the name? Presumably, the early French explorers concluded there were 12 islands, and named them after the 12 apostles of the Bible. Actually, there are 22 islands, 21 of which are in the Apostle Islands National Lakeshore, established in 1970, that also includes part of the Bayfield Peninsula north of the Red Cliff Indian Reservation (Fig. 8).

The bedrock on 19 of the islands is the Chequamegon Sandstone; not so surprisingly, the bedrock on Devils Island, the northernmost island, is the Devils Island Sandstone, and Sand Island, the north-westernmost island, consists of the Orienta Sandstone. Well-exposed sedimentary structures are common.

Glacial deposits of the Miller Creek Formation blanket most of the islands, but red to buff bedrock is visible in cliffs, sea caves, sea stacks, arches and elsewhere along the shorelines where erosion has exposed the bedrock. Several excellent beaches and a few tombolos are present. Locally, there are sand dunes.

The islands look quite pristine, but they were extensively logged for white pine, hemlock, yellow birch and maple between 1850 and 1970. Brownstone quarries were established on Basswood Island in 1869, and later on Hermit and Stockton Islands. There are six lighthouses on the islands, built between 1857 and 1891.

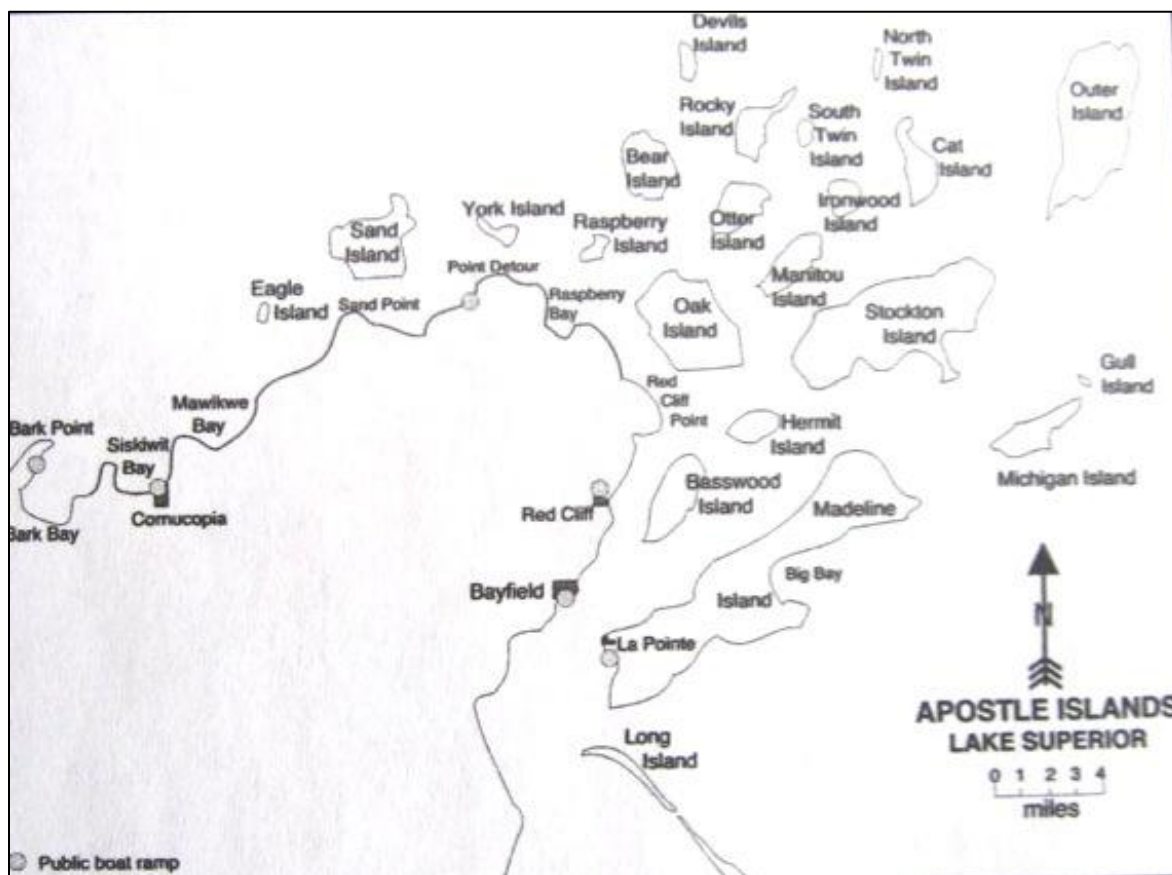


Figure 8. The Apostle Islands. From Nuhfer, 2004.

PLEISTOCENE GEOLOGY

Introduction

The Bayfield Peninsula has been repeatedly glaciated over the last 2.4 million years, but the most recent glaciations have wiped out evidence of previous glaciations, leaving a record of only the last 30,000 years. Glaciers scoured down to bedrock, then deposited many tens of meters of till and outwash as they melted away. Glacial lakes intermittently occupied the margins of the ice sheets where deposits of lacustrine sediment accumulated, much of which was reworked by minor glacial advances. These sediments have been divided into two units: the sandy Copper Falls Formation deposited as till and outwash before 11,500 BP, and the younger, finer-grained Miller Creek Formation consisting of lacustrine and glacially reworked lacustrine sediments deposited between 11,500 and 9,500 years BP (Fig. 9) (Clayton, 1984).

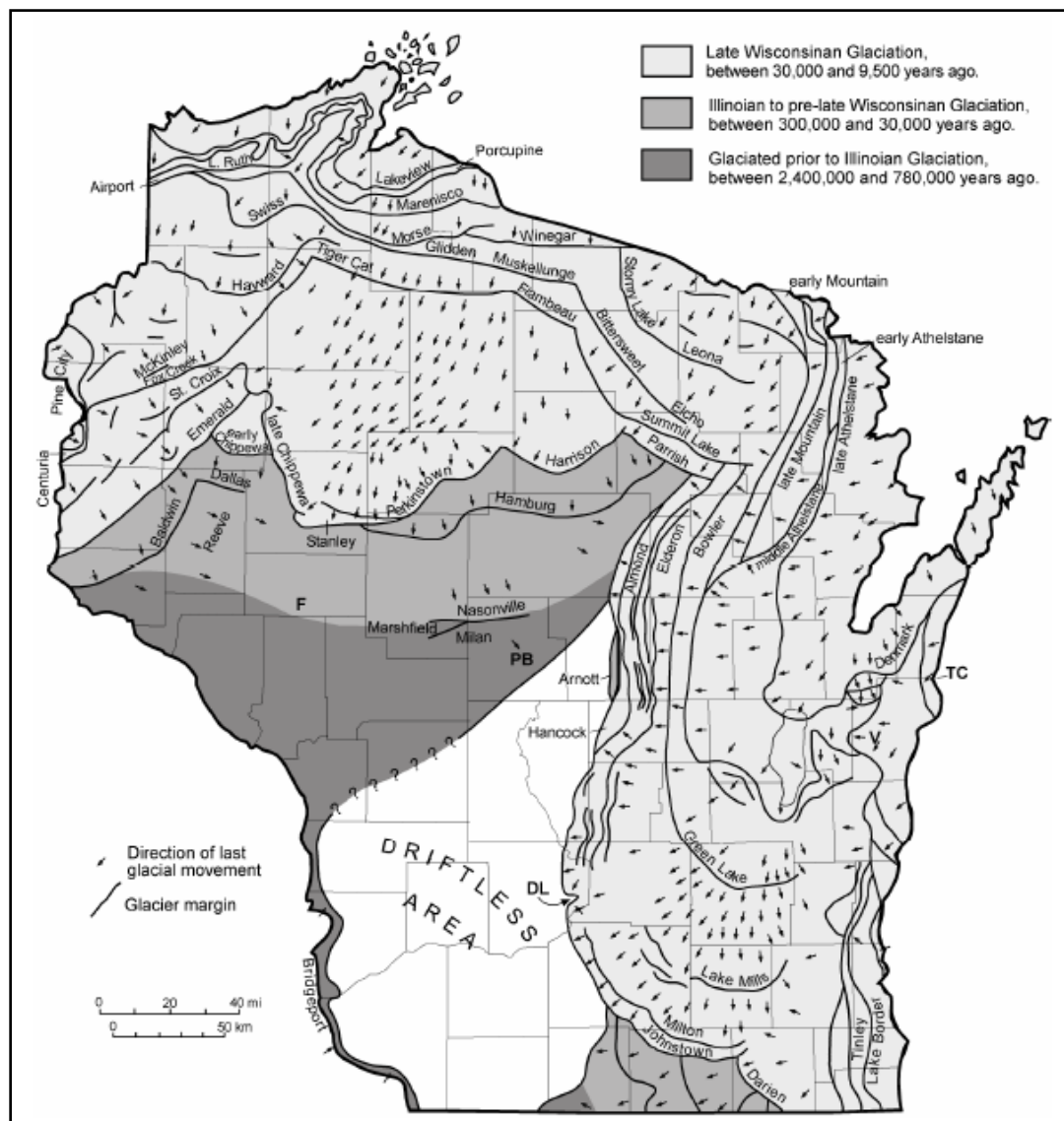


Figure 9. Glaciations in Wisconsin. From Syverson and Colgan, 2004.

Glaciation

The Bayfield Peninsula was most recently glaciated by the Superior Lobe and the Chippewa Sub-lobe of the Laurentide Ice Sheet, during the Wisconsin Glaciation from 25,000 BP to 9,500 BP. Deposits from several glacial advances have been identified in northern Wisconsin, the youngest of which is referred to as the Lake View Advance (Fig. 10). The ice-margin positions have been determined by a combination of lithostratigraphic evidence, ice-margin landforms, and inferred connections in areas where no evidence remains (Clayton, 1984).

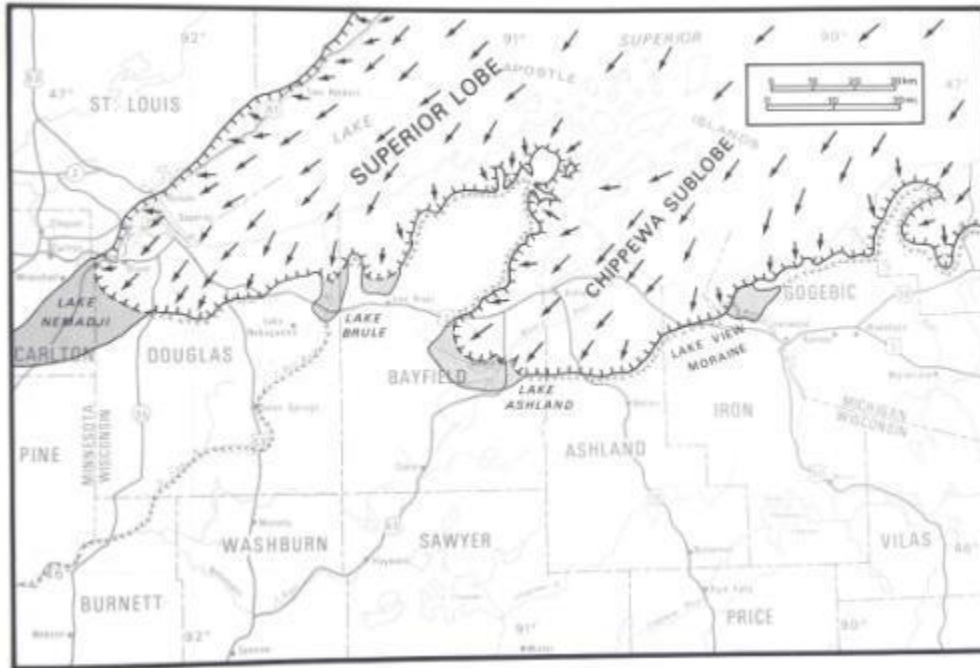


Figure 10. Furthest extent and flow direction of the Lake View Advance, between 10,000 BP and 9500 BP. From Clayton, 1984.

Surficial Units

Copper Falls Formation

The Copper Falls Formation lies above pre-Pleistocene rock types and is generally composed of sandy till derived from Keweenawan sandstones in the Superior basin (Fig. 11). The Copper Falls Formation is at the surface in large areas of northern Wisconsin, and is prevalent at elevations higher than 1000 feet where it has not been covered up by younger deposits of lacustrine and reworked lacustrine sediments. The type section for the Copper Falls Formation is a river cutback near Copper Falls. Although the Copper Falls Formation is primarily composed of sandy till, it contains other sediment types including common proglacial stream deposits. These proglacial stream deposits are often interbedded with lithologically similar till. The formation on average is 40-80% sand, 15-50% silt, and 2-20% clay. The Copper Falls is slightly calcareous and the clay component is mostly composed of illite and smectite (Clayton, 1984).

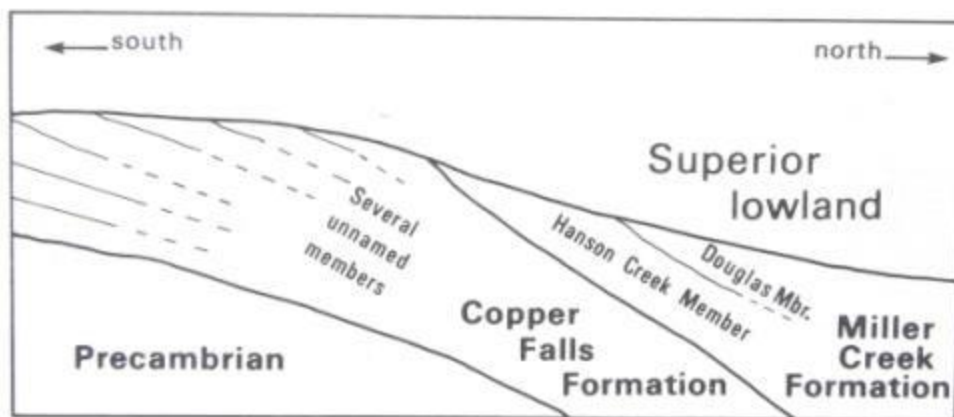


Figure 11. The stratigraphic relationship between the Keweenaw rocks, the Copper Falls Formation and the Miller Creek formation. From Clayton, 1984.

Miller Creek Formation

The Miller Creek Formation, overlying the Copper Falls Formation, was deposited between 11,500 and 9,500 BP, and is composed of offshore clayey sediment and clayey till derived from fine-grained offshore sediment. The Miller Creek Formation has two subunits on the western half of the Bayfield Peninsula: the Hanson Creek Member and the overlying the Douglas Member. It is the majority of the subsurface material between 1000 feet and lake level (602 feet). The Hanson Creek Member is generally unlaminated and is 45-70% clay, 20-45% silt, 3-20% sand, trace pebbles, cobbles, and boulders. The silt and clay particles are typically 10% carbonates. The Douglas Member generally is 45-85% clay, 10-40% silt, 3-20% sand, trace pebbles, cobbles, and boulders. Where it overlies sand it may contain up to 60% sand. It is usually dull reddish brown and calcareous. East of the Bayfield Peninsula the Miller Creek Formation is not divided into named members. It is 30-65% silt, 10-60% clay, 5-30% sand, trace pebbles, cobbles, and boulders. It is usually reddish brown and where it overlies sand may be more sandy (Clayton, 1984). The presence of well-preserved laminations in some parts of the Miller Creek Formation, and their absence in other places, is indicative of the variable depositional environments and post-depositional histories of the sediments. Some parts are undisturbed lake sediments, while other parts were reworked, mixed with coarser sediment, and deformed by glacial action by later ice advances (Clayton, 1985; Shumway and Iverson, 2008).

FIELD TRIP STOPS

Head west out of Ashland on U.S. Highway 2. Just outside of town, we are driving to a young beach ridge built up at the mouth of Fish Creek (Fig. 12). This formed as the result of very rapid sedimentation from Fish Creek at the shallow southwest end of Chequamegon Bay. Three miles west of Ashland, past the junction of U.S. Highway 2 and Wisconsin Highway 13, is the Northern Great Lakes Visitor Center, which has excellent displays of natural and cultural history, including some that are geologic. Between the junction with Wisconsin Highway 13 and County Road E in Ino, a distance of 12 miles, we are driving on

the Pleistocene Miller Creek Formation. The glacial topography has been greatly modified by Glacial Lake Duluth. From just west of Ino, traveling 20 miles to Brule, we are driving on the older Pleistocene Copper Falls Formation.

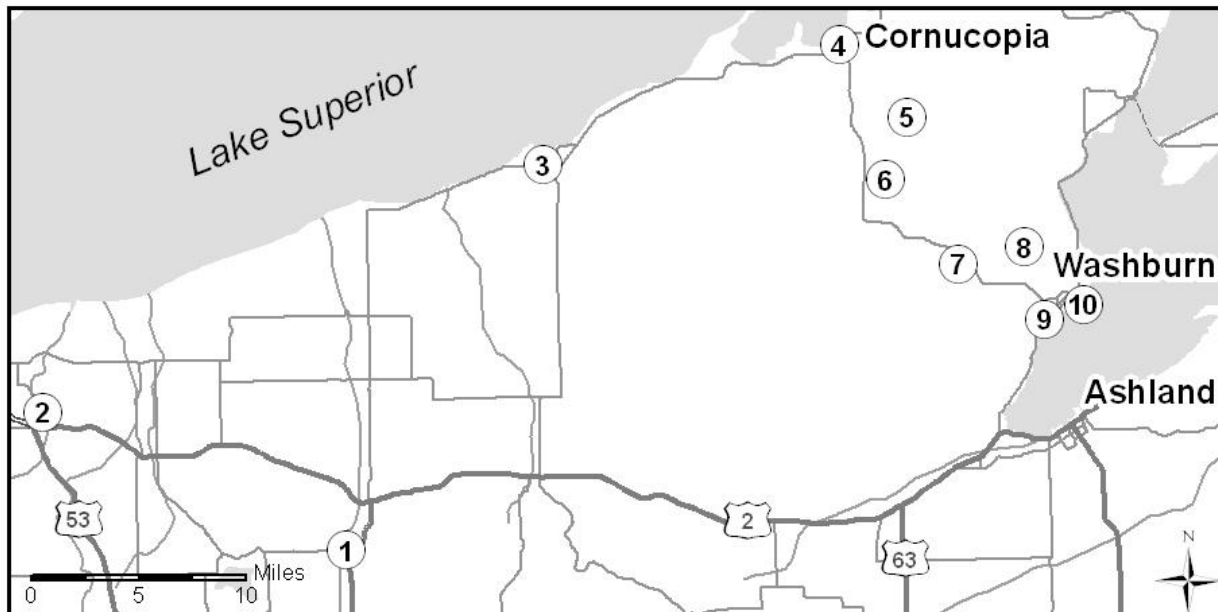


Figure 12. Map of field trip stops.

Stop 1: Brule River Outlet of Glacial Lake Duluth

In Brule, turn south on Wisconsin Highway 27 and proceed for 3 miles to Winneboujou. Winneboujou is located at the bottom of a steep-walled outlet channel, presently 100 feet deep, about one-mile-wide, that was cut when Glacial Lake Duluth drained via the Bois-Brule River outlet into the St. Croix River, which enters the Mississippi River (Figs. 13, 14).

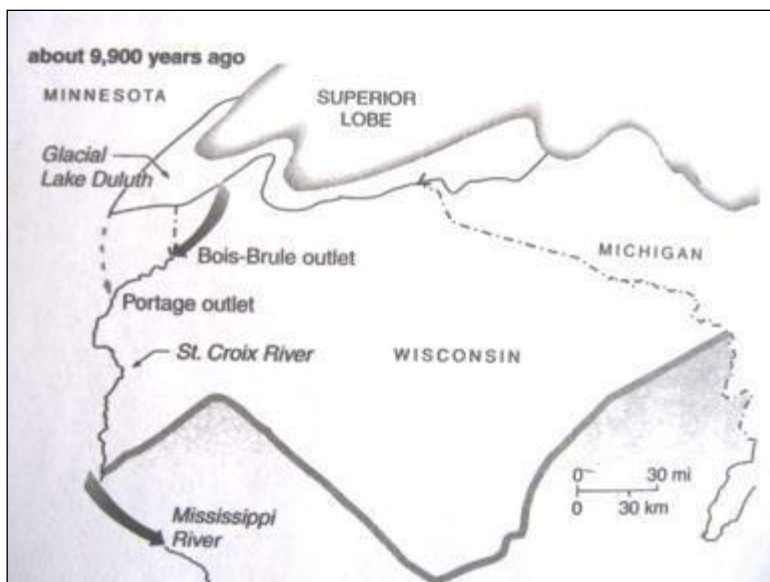


Figure 13. Generalized map showing the Bois-Brule outlet that carried waters southward from Glacial lake Duluth to the Sta. Croix River and hence to the Mississippi River. The heavy line near the bottom represents the southernmost advance of the Late Wisconsin glaciers. From Dott and Attig, 2004.

When Glacial Lake Duluth breached this outlet, at an elevation of about 1100 feet, the Bois-Brule flowed southward, cutting this deep and wide channel. The present underfit Brule River now flows northward into Lake Superior, with its headwater 30 miles to the south. Native Americans, early explorers, and fur traders came up the Mississippi and St. Croix Rivers and then traversed a two-mile portage from the St. Croix River into the present Brule River. This divide is located a few miles north of Solon Springs.



Figure 14. Landscape image of the Bois-Brule outlet of Glacial Lake Duluth just south of the village of Brule. Note the Glacial Lake Duluth plain. From Dott and Attig, 2004.

Between Brule and Maple, a distance of 8 miles, we are again driving on the younger Miller Creek Formation. At Maple, we descend over several low, sandy beach ridges (elevations about 1080 feet) onto lake-modified glacial topography, the bed of Glacial Lake Duluth. Three miles west of Maple is the village of Poplar, located on this lake bed. Proceed about 5 miles and turn right on Country Road U. Follow County Road U a short distance to the Amnicon State Falls Park entrance on the left.

Stop 2A: Amnicon Falls State Park: Orienta Sandstone, Keweenaw Volcanics, and the Douglas Fault.

The Douglas Fault has thrust Middle Keweenaw basalt (Chengwatana Volcanics) northward over the Late Keweenaw Orienta Sandstone (Fig. 15). A waterfall exists as the Amnicon River flows over the basalt into the gorge cut into the much softer, subhorizontal red Orienta Sandstone. The basalt flows dip at 30-40 ° to the southeast whereas the Orienta dips at about 5°. These relationships are easily seen after crossing the bridge and observing the east bank from the west side of the river. Note that the Orienta has been dragged into a near-vertical attitude adjacent to the few-meter-thick gouge and breccia zone between the fault and the Orienta. Prior to a collapse of the side of the gorge several years ago, at the base of the short stairway on the east bank, steep slickensides were visible on the sandstone beneath the breccia.

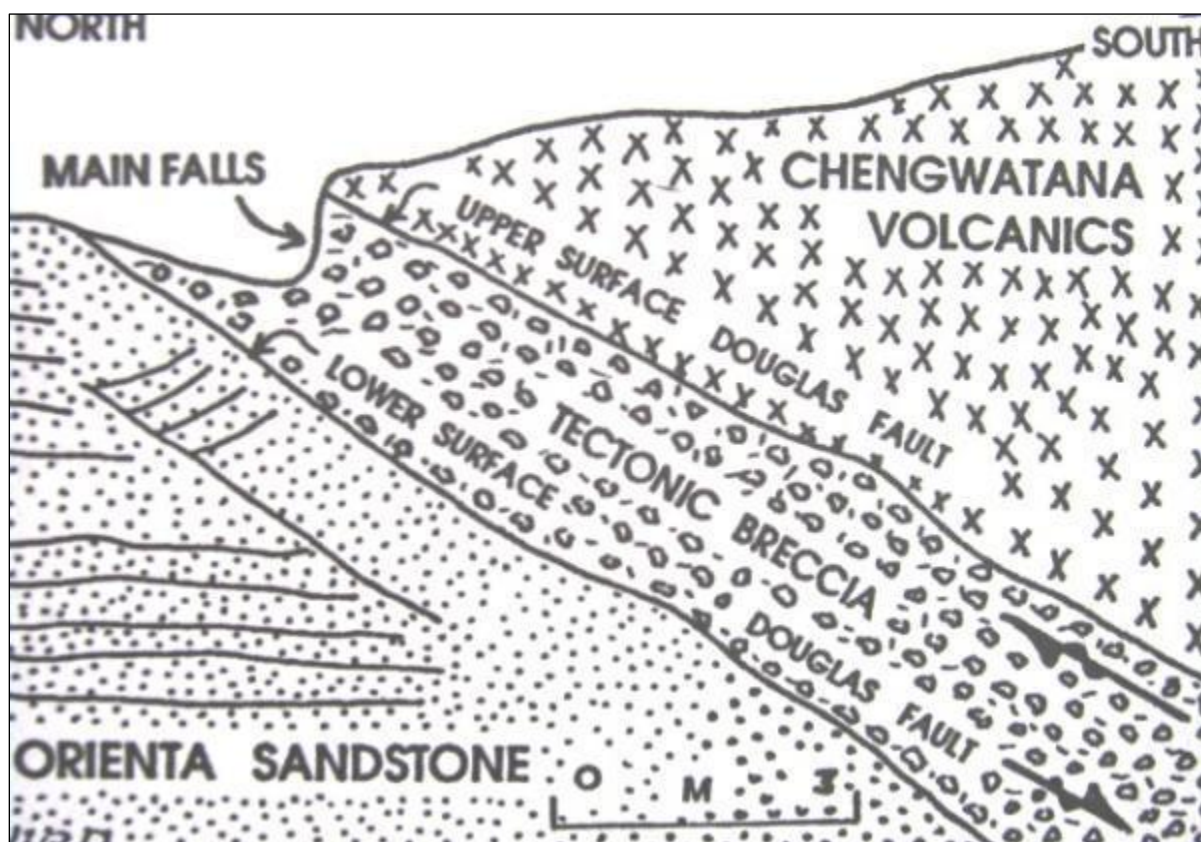


Figure 15. Sketch of the Douglas Fault as exposed at the base of the main falls in Amnicon Falls State Park. The arrows show the relative motion along this fault. From Dickas, 1989, adapted after Thwaites, 1912.

You can walk on the Orienta Sandstone a short distance downstream from the bridge on the northeastern bank. Note trough cross-bedding and parting (current) lineation. The Douglas Fault defines the northwestern edge of a major feature of the Midcontinent Rift, the St. Croix Horst. (see Fig. 2). It has been traced by geophysical methods from the Twin Cities through northwestern Wisconsin and into Lake Superior, and may have 3000 m of displacement (Craddock, 1972). There is now some question about whether it connects with the Isle Royale Fault beneath Lake Superior. The southeastern side of the horst is marked by the Lake Owen-Keweenaw Fault system, also reverse faults. It is generally assumed that these faults were originally normal faults during the formation of the rift, but became reverse faults later due to compressional forces perhaps caused by the collision of the Grenville micro-continent with eastern North America. A paleomagnetic study by Watts (1981) estimated the age of the compressive event at 950-1040 Ma.

Stop 2B: (Optional) Quarry in Keweenaw Volcanics

Exit the park and return to U.S. Highway 2. Turn right (west) and drive a short distance to an entrance road into a quarry operated by Northwoods Paving of Ashland. Drive a short distance behind the crushed rock piles to basalt exposures on the quarry wall. Nearby is gabbro of the Amnicon Intrusion (Nicholson et al, 2006) that has altered and deformed these flows.

Return to County Road U on which the entrance to Amnicon Falls State Park is located, and proceed north for 5 miles to Wisconsin Highway 13. Proceed eastward for about 35 miles to Port Wing. Note that we are driving this entire distance on the flat bed of Glacial Lake Duluth (the Lake Superior Lowland), and that this flat area (the Douglas Member of the Miller Creek Formation) is deeply dissected by numerous rivers and smaller streams, all flowing northward into Lake Superior (Fig. 16). Locally some of the rivers have cut down to the underlying Pleistocene Copper Falls Formation and to the Bayfield Group sandstones.

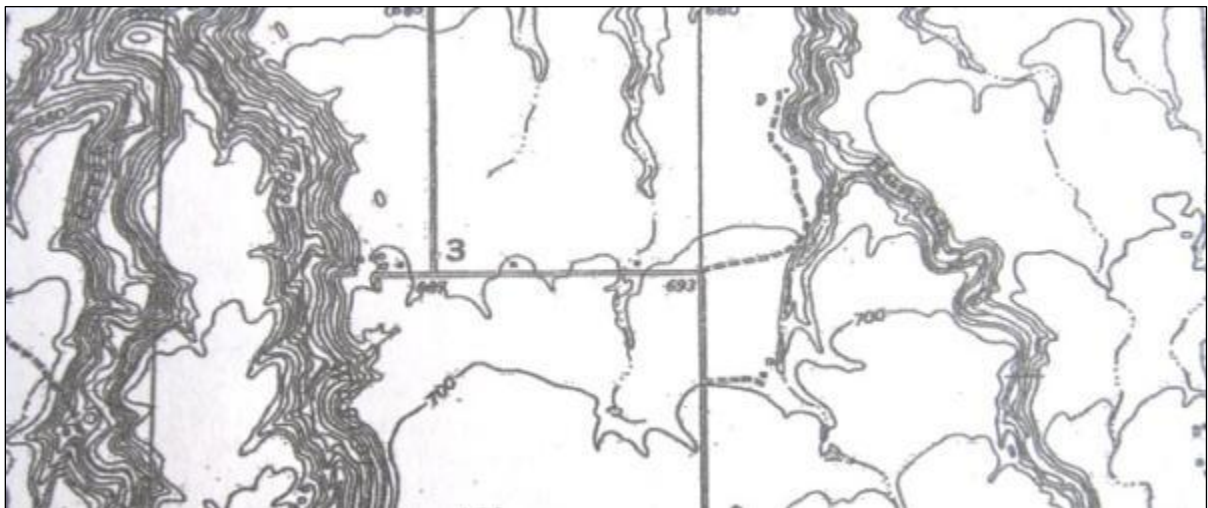


Figure 16. Lake-modified glacial topography intersected by incised stream valleys. U.S. Geological Survey Poplar NE Quadrangle (7.5-minute, topographic, 10-ft contour interval. T. 48 N.. R. 12 W. From Clayton, 1984.

Stop 3: Orienta Sandstone at Twin Falls Park in Port Wing

Upon entering Port Wing, pull off to the right into Twin Falls Park. Follow the path a few hundred feet to an observation deck above the falls. Note that the Flag River has cut a deep gorge through the glacial beds and into the Orienta Sandstone.

Stop 4: Devils Island Sandstone at Siskiwit Falls, Cornucopia

In Cornucopia, turn right (south) on Superior Avenue and continue for one block. Turn left (east) on County Road C and continue 0.4 miles to Siskiwit Falls Road. Turn left and continue for one block. Park on shoulder just before the bridge over the Siskiwit River. The falls, really a rapids, formed by water passing over a series of sandstone beds, is just upstream from the bridge. This, unfortunately, is the only easily accessible exposure of the quartz arenite of the Devils Island Sandstone. At low water, a total thickness of about 6 meters can be seen. Look for current ripple marks and cross-bedding. If the water is high, the only observation possible may be a close look at the composition of loose pieces of the formation under a hand lens—99% fine-grained quartz. Good rounding of grains may be visible in coarser-grained pieces.

Transect Across the Bayfield Peninsula

Stops 5 through 10 constitute a transect across the peninsula, from the highest elevations in the middle of the peninsula down to the lakeshore on the east side (Fig. 17). All of the stops are on public land, with the exception of Stop 7, which is on private land. Locations are given in public land survey grid, and UTM coordinates (Zone 15T), datum NAD 1927. The transect crosses the transition zone between sandy Copper Falls sediments and clay-rich Miller Creek sediments on both sides of the Peninsula (Fig. 18).

From Siskiwit Falls, go east on Siskiwit Falls Road, then turn right (south) on Stage Road, which becomes Star Route Road. Take Star Route road southeast for approximately three miles uphill to where the road levels off and turns east. After passing over the top of the hill, go left on Settlement Dump Road. Stop 5 is at the end of this road. The road has a steep grade and loose sand, so it may not be passable at all times of the year.

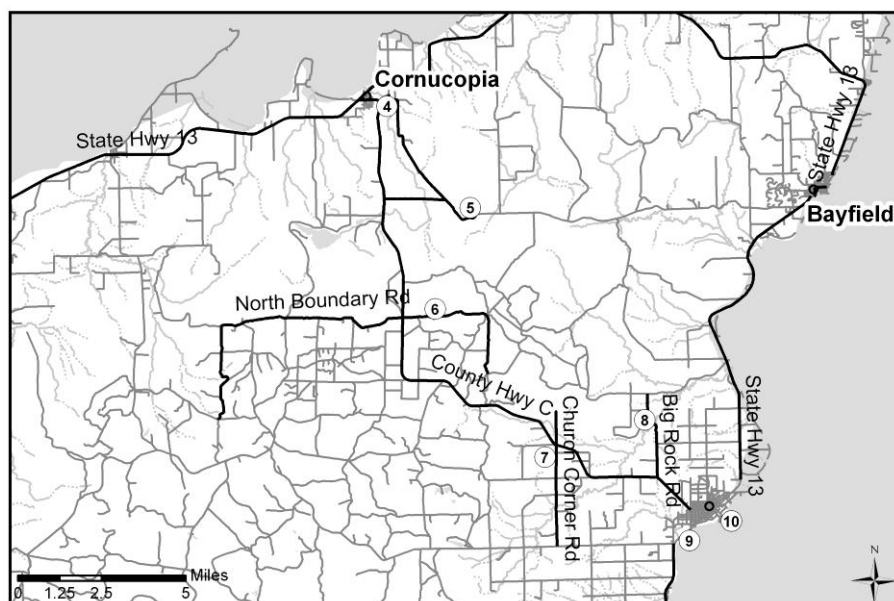


Figure 17. Map showing locations of field trip stops on the Bayfield Peninsula.

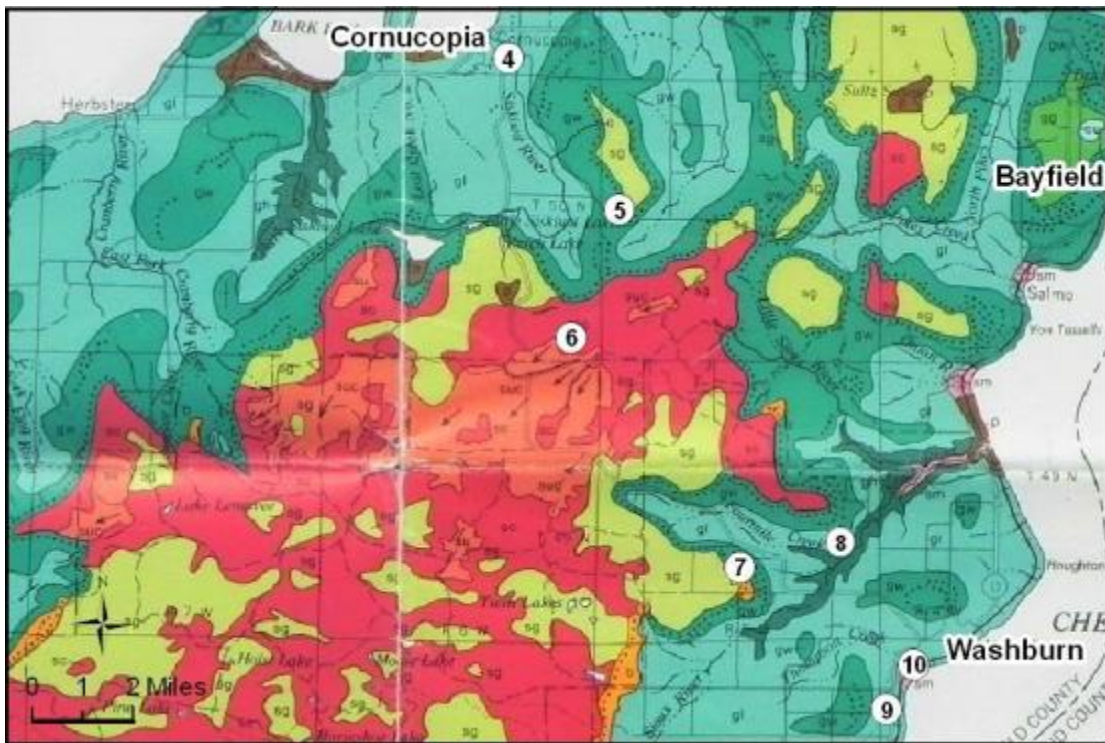


Figure 18. Surficial geologic map showing field trip stops 4-10. Miller Creek Formation: gl—lake-modified glacial topography (teal on map), gw—wave-plained topography (green), gh—valley sides (darkest green), b—shoreline sediment (orange); Copper Falls Formation: suc—uncollapsed proglacial stream sediment (Valhalla surface) (reddish orange), sc—collapsed proglacial stream sediment (red), sg—hummocky stream sediment overlain by silty material (light green). Modified from Clayton, 1985.

Stop 5: Glacial Outwash and the Glacial Lake Duluth Beach

Location: Gravel pit at the north end of Settlement Dump Road; UTM coordinates: 0649710 mE, 5185261 mN; T50N, R5W, Sec 18, south-central. (Fig. 19)

Stop 5A: Glacial Outwash Above the Lake Duluth Beach

Travel up Settlement Dump Road watching for exposures of sediment; the best exposures are in a gravel pit at the top of the hill (UTM coordinates listed above). This is glacial outwash of the Copper Falls Formation deposited by meltwater rivers flowing out of the Superior Glacial Lobe to the west and the Chippewa Lobe to the east. Here in the central part of the Bayfield Peninsula, at elevations above 1100 feet, outwash of the Copper Falls Formation is gravelly sand with few cobbles and boulders. There is a wide variety of clast types, which is typical of the Copper Falls Formation. The hilltop to the north of the pit is the highest point on the Bayfield Peninsula, with an elevation of 1426 feet – which is 824 feet above the present surface of Lake Superior and well above all the glacial lake beaches. The sandy soils and relatively high topographic relief in the area make for excessively well drained soils. These soils support a sparse forest consisting of only drought-tolerant trees such as jack pine, red pine, and scrubby red oak. Be careful of glass and debris in this gravel pit.

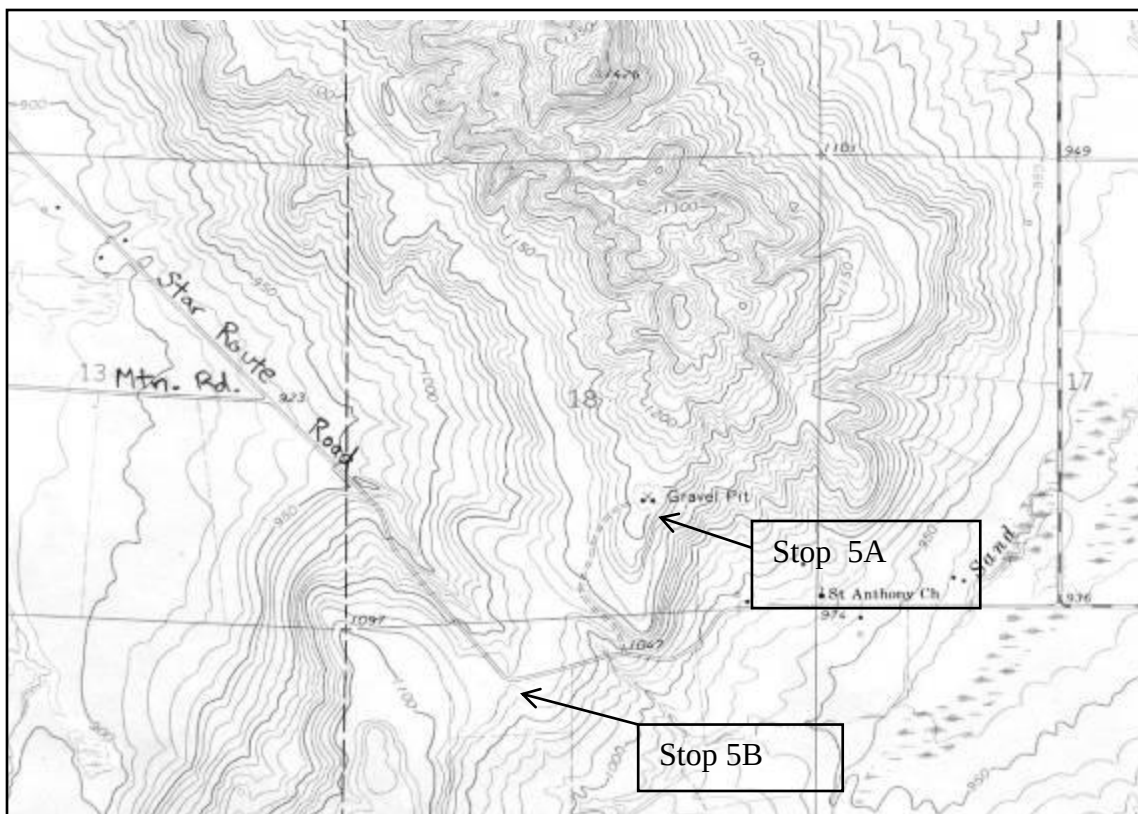


Figure 19. Topographic map showing location of Stop 5 area (U.S.G.S., 1964a).

Go back down Settlement Dump Road and go right (west) on Star Route Road. In 0.25 miles stop on the flat stretch where the road turns from west to northwest. You are now on a beach of Glacial Lake Duluth – stop 5B.

Stop 5B: Glacial Lake Duluth Beach Terrace

Location: On Star Route Road; UTM coordinates: 0647828 mE, 5180050 mN; T50N, R5W, Sec. 19, north-central (Fig. 19).

There is a distinct terrace at an elevation of 1100 feet throughout this part of the southern Lake Superior region, and is especially well defined here on the Bayfield Peninsula. This beach terrace was created 9500 BP when Lake Duluth was at its highest level (Fig. 20).

The terrace at 1100 feet is a distinct step in the landscape and an important transition zone from very sandy sediments on and above the beach level to finer sediments on the slopes below. The sandy outwash seen in stop 5A underlies all areas above the beach – deep deposits of gravelly sand with no silt or clay. The beach terrace goes all the way around this hill, indicating that at one time it was an island within Glacial Lake Duluth.

Continue northwest on Star Route Road back toward Siskiwit Falls. In 0.7 miles, turn left (west) on Mountain Road and travel two miles to County Road C. Turn south on County Road C and go 3.5 miles to North Boundary Road. Turn left (east) on North Boundary Road East, which is also labeled FR 439, and head toward the Bayfield County Sportsman's Club. Stop 7 is one mile east on this road.

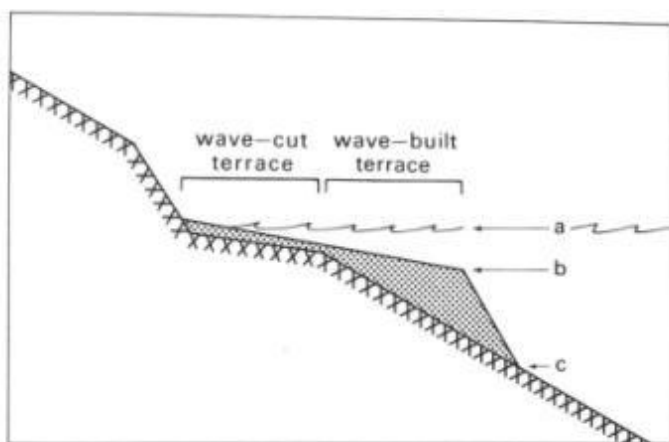


Figure 20. Cross section of a beach terrace. Letter a represents the lake level, b is the outer edge of the wave-built terrace, and c is the base of the wave-built terrace. From Clayton, 1984.

Stop 6: Pitted Outwash Plain

Location: North Boundary Road East; UTM coordinates: 0647828 mE, 5180050 mN; T50N, R6W, Sec 36, south-central (Fig. 21).

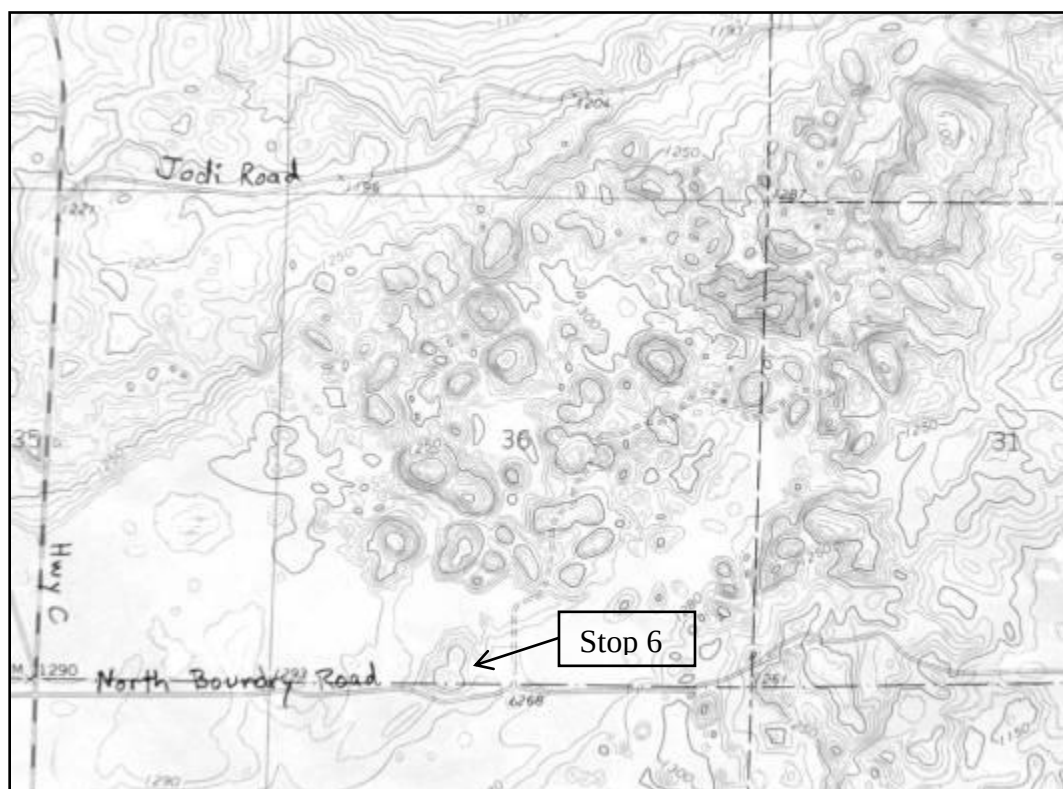


Figure 21. Topographic map of Stop 6 area (U.S.G.S., 1964a).

Areas above the 1100-foot beaches have a variety of landforms including outwash plains, eskers, outflow channels, and kettles of all sizes. These features were created during melting of the ice of the Airport Advance, the last glacial advance to cover the high interior of the Bayfield Peninsula, about 12,300 BP (Clayton, 1984). The large flat outwash plain, at an elevation of 1250-1280 feet, in the middle of the Peninsula is the Valhalla Surface.

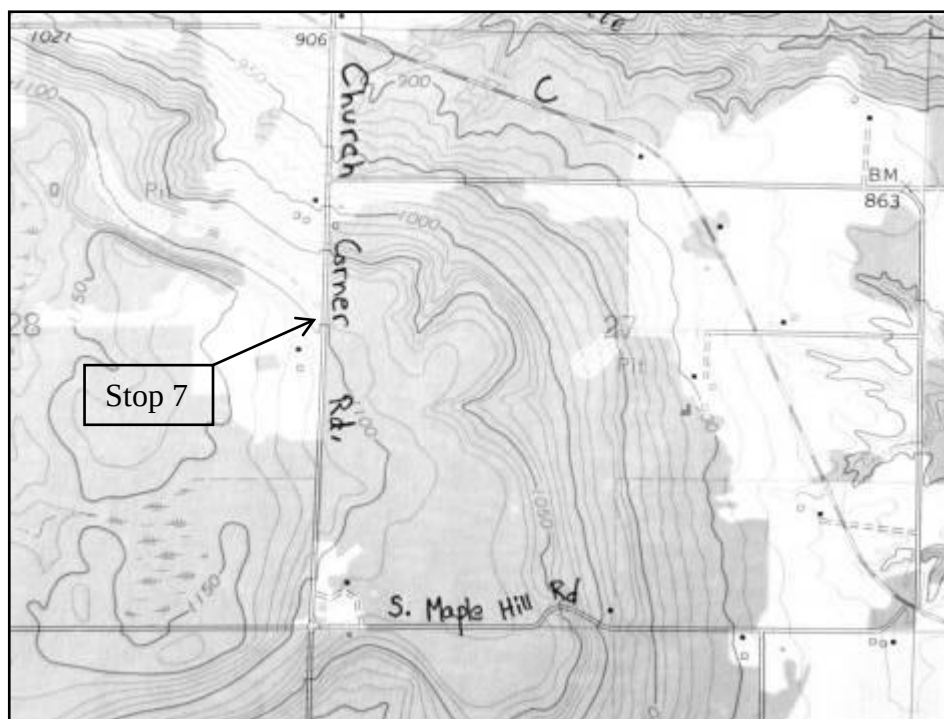
Numerous large blocks of ice were left behind in some areas as the ice margin retreated, and subsequently the blocks were buried in thick deposits of sandy outwash. As the ice blocks melted, kettles formed in the otherwise flat outwash plain. These pits are especially abundant in Sections 30, 31, and 36 (T50N, R6W), some up to 180 feet deep and 3000 feet across with very steep slopes (Figure 22).

Here at stop 6 is a relatively small pit, the southern-most kettle in a large kettle field. Just to the east is a road that goes north into the heart of the kettle field if time permits, we will walk the road north into the kettle field. The absence of surface water here is very revealing about the high permeability of the sands of the Copper Falls Formation. There are no rivers in this high interior part of the Bayfield Peninsula because all water infiltrates and is drained by groundwater. The presence of kettles nearly 200 feet deep that are dry on the bottom is evidence of the great depth of the water table.

Return to County Road C and turn south. Travel 7.5 miles to Church Corner Road and turn right (south). In 0.5 miles stop at the crest of the hill. This is stop 8.

Stop 7: Beach Terrace and Former Bay

Location: Church Corner Road; UTM coordinates: 0653786 mE, 5173208 mN; T49N, R5W, Sec. 27, west-central (Fig. 22). *This is private land so do not explore far off the road without seeking permission from land owners.*



This stop is on the terrace of the highest Lake Duluth beach, the 1100-foot beach level visited on stop 5B. This terrace marks the highest and longest lived lake level of the numerous lake levels that existed in the Superior Basin during deglaciation, thus it is the topographically most distinct terrace and the only one which has significant sandy beach deposits (Farrand, 1969). Below this terrace the topography was planed by waves of at least

five short-lived lower lake levels, so lower beaches are less well defined. On the east side of the road, there is an exposure of pebbly sand typically present on this terrace. The sand of this beach is now used as a resource for road construction and other uses, and several sand pits exist along contours of this hill. The view to the north is of the Fourmile Creek valley and the Bayview ridge beyond. At the time that the Glacial Lake Duluth shoreline was here, about 9500 bp, this was a large bay closing to the west.

The hydrology changes dramatically below 1100 feet, where the sand of the Copper Falls Formation is overlain by silt and clay of the Miller Creek Formation. There is typically a transition zone between these two areas, below which fine sediments and soils prevent much infiltration of precipitation. As a result, nearly all the rivers on the Bayfield Peninsula start at the edge of the fine sediments just below an elevation of 1000 feet.

This stop provides an excellent view of the transition zone where the sediment changes from sand on the hill slopes to mud on the bottom of the valley. The transition is rapid on steep slopes such as this one, with clay-rich sediments of the Miller Creek Formation dominating below elevations of about 950 feet. At that elevation the hydrology and soils change, so there is a corresponding change in forest types and ecosystems, from the dry jack pine and red oak forests above to wet-tolerant species at lower elevations. Although the differences in the forest types are much less pronounced today than they were prior to the forests being cut over in the late 19th century, the differences are still apparent. At the time of first settlement, the “Lake Superior clay plain” was occupied by spruce, fir, and cedar, whereas today it is primarily red maple, aspen, and birch.

Travel 0.5 mile and turn left (east) on Maple Hill Road. Travel 1 mile to County Road C and turn right (east) toward Washburn. In 2 miles, turn left (north) on Big Rock Road. Travel 1.5 miles to Big Rock Park and stop at the parking lot just north of the bridge over Sioux River -- this is stop 9.

Stop 8: Chequamegon Sandstone and Miller Creek Formation

Location: The parking lot at Big Rock Park; UTM coordinates: 0658541 mE, 5174862 mN; T49N, R5W, Sec. 24, east-central (Fig. 23)

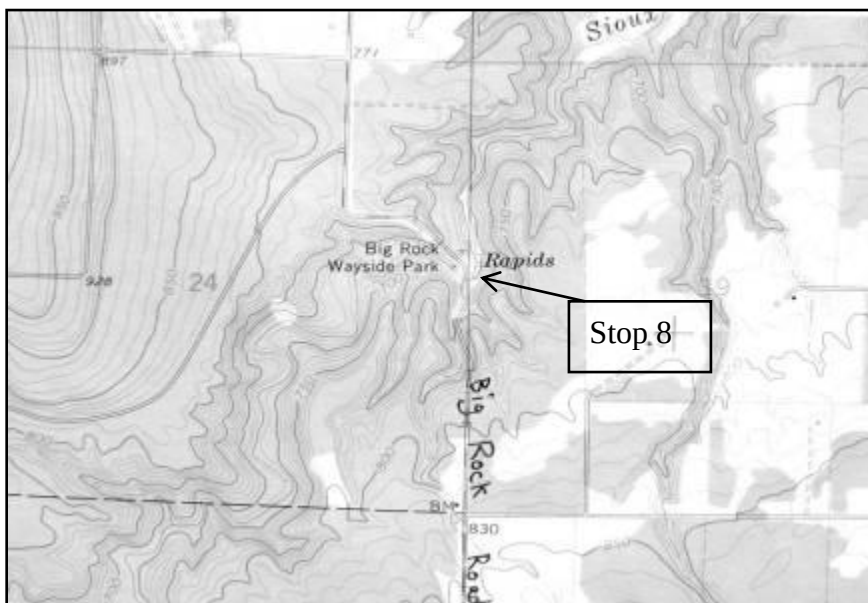


Figure 23. Topographic map of Big Rock Park area at Stop 8 (U.S.G.S., 1964b).

Stop 8A: Chequamegon Sandstone

Location: The Sioux River valley downstream of Big Rock Park; UTM coordinates: 0658541 mE, 5175068 mN (Fig. 23).

From the parking lot hike north on the trail along the Sioux River for about ¼ mile to two long outcrops of Chequamegon Sandstone. The northern outcrop is located at 0658541 mE, 5175068 mN.

Pleistocene deposits have been eroded away in the Sioux River valley to expose the Chequamegon Sandstone, the stratigraphically highest formation in the Bayfield Group. The Chequamegon Sandstone here is a feldspathic sandstone that has enough clay on weathered surfaces to give it a chalky white appearance. It is poorly cemented and friable, so it is not very resistant to erosion. Although it is composed primarily of sand-sized clasts, there are sparse quartz pebbles and some mud chips; the chips are soft red mud, which easily erodes out of the sandstone, leaving deep holes in the surface of the outcrop. Some of the mud chips in this outcrop are up to 25cm in length.

Bedding here is horizontal and has large cross beds 0.5 to 1m in height, with dip of cross strata indicating the flow was in a northerly direction, similar to paleocurrent indicators throughout the Bayfield Group. Some of the bedding is disturbed by various soft-sediment deformation features, including at least one excellent ball-and-pillow structure exposed in the southern outcrop.

The sediment along the base of the cliff is sand derived from weathering of the friable sandstone immediately above. However, about 2m away from the cliff face the sediment is hard sandy mud. Most of the Sioux River valley is filled with this red mud – sediment typical of the Miller Creek Formation. The presence of Miller Creek mud deep within a bedrock gorge indicates that the gorge was carved into the sandstone prior to the deposition of the mud. Thus, it is likely that the Sioux River gorge was carved during a Pleistocene interglacial time, then filled with mud during deposition of the late Pleistocene Miller Creek Formation, then re-excavated to create the river valley present today. This situation is similar to that of Fish Creek west of Ashland, and Sand River on the northern side of the Bayfield Peninsula (Nuhfer, 2004).

Hike back out the trail, past the parking lot, south along the road to the banks on the east side of Big Rock Road – this is stop 9B.

Stop 8B: Fine mud of the Miller Creek Formation

Location: East side of Big Rock Road, just south of the bridge over Sioux River.

UTM coordinates: 0658570 mE, 5174755 mN (Fig. 23).

The bank is composed of red mud of the Miller Creek Formation. The Bayfield County Highway Department has attempted to stabilize the bank, with moderate success; the clay of the Miller Creek Formation is notorious for its instability, especially where oversteepened by erosion or excavation. The mud here in the Sioux River valley is especially fine grained for the region, consisting of approximately 60% clay, 35% silt, and 5% sand, with few, if any, pebbles. Although no laminations have been seen in this exposure, there are excellent examples of horizontal, undeformed, thin laminations in the mud upstream along west side of the Fourmile Creek valley (just downstream of the Hwy C bridge).

Drive south on Big Rock Road, back to Hwy C. Go left (east) on C into Washburn. At the intersection of C and Hwy 13, go straight across the intersection (south) onto South 8th Avenue West. In about 0.5 mile enter Thompson's West End Park. The artesian well is on the right (west) as you first enter the park – stop 10.

Stop 9: Flowing Artesian Well

Location: Thompson's West End Park in Washburn; UTM coordinates: 0660194 mE, 5170054 mN; T48N, R4W, Sec. 6, SE1/4 (Fig. 24).

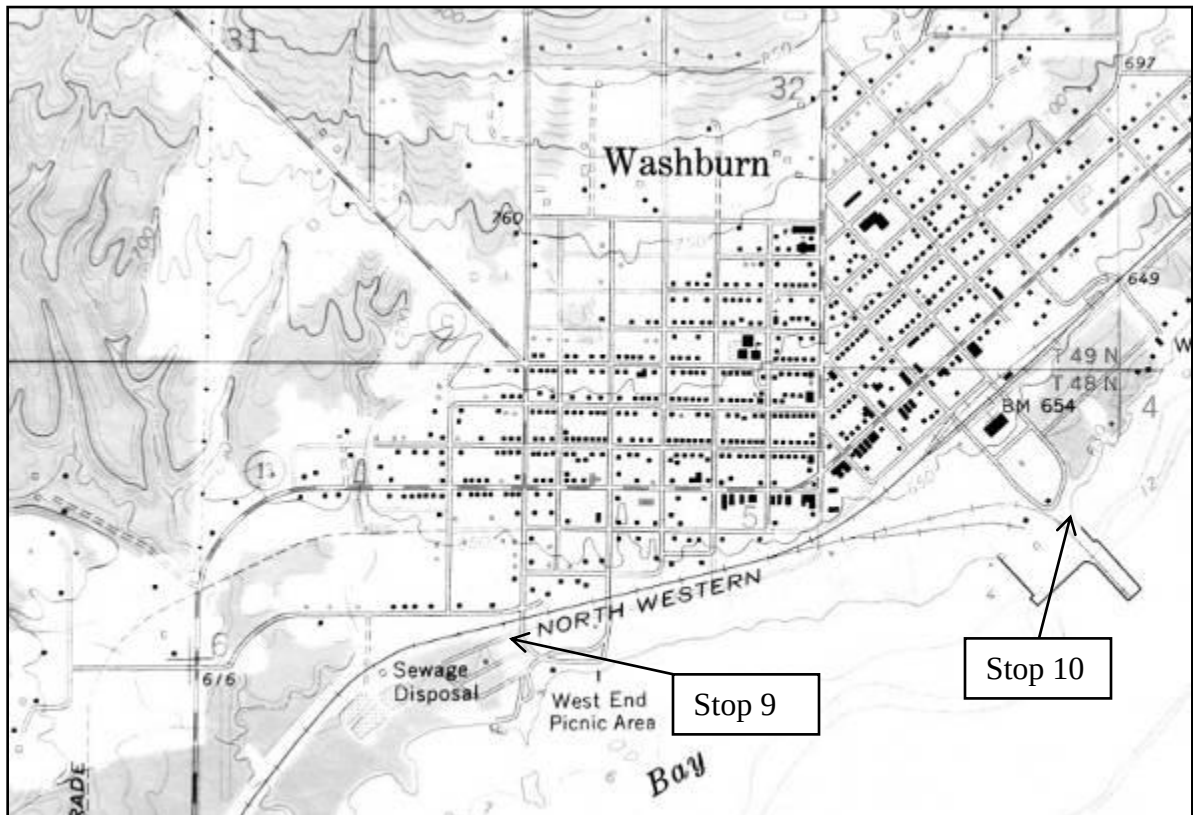


Figure 24. Topographic map of Washburn and stops 9 and 10 (U.S.G.S., 1964b).

The fine-grained sediments of the Miller Creek Formation create a confining layer over more-permeable sediments of the Copper Falls Formation. The very permeable sands of the Copper Falls Formation exposed at the surface in the middle of the Bayfield Peninsula – areas above the 1100-foot-elevation beaches – serve as the recharge area for this confined aquifer. This situation creates high enough artesian pressure that the piezometric surface is above the land surface and flowing wells can exist. This hydrogeologic condition exists all around Chequamegon Bay, and there are numerous flowing artesian wells in the area. There are also natural springs, although none near here that have high flow rates. The artesian wells are popular sources of drinking water, and often there is a line of people waiting to fill their bottles.

This is the historic Sprague Well: drilled in 1903, it was the first artesian well in Bayfield County; it is 119 feet deep and currently produces 43 gallons per minute (gpm).

The water has a constant temperature of 45° F and total dissolved solids concentration of 104 ppm. The interpretive sign at the well says that it originally produced 224 gpm from a four-inch pipe and was “so free of minerals that it was piped directly into the sawmill boilers.”

Return to Hwy 13 and turn left (east). Go east on Hwy 13 to downtown Washburn, then go right (southeast) on Central Avenue toward Lake Superior. Just before the town dock there is a parking lot on the left (north) and a small beach – stop 10.

Stop 10: View of Chequamegon Bay

Location: Washburn beach, near the town dock; UTM coordinates: 0661718 mE, 5170422 mN; T49N, R5W, Sec. 24, NE ¼ (Figs.24, 25)



Figure 25. Satellite image of Chequamegon Bay (image from UW SSEC and Wisconsin View).

From the beach you can look out over Chequamegon Bay, which has served as a safe harbor to vessels on Lake Superior for centuries because it is so well protected from the wind

and waves of the open lake. This protection is provided by the Apostle Islands and Bayfield Peninsula to the north, and Long Island to the northeast (Figure 25).

Long Island is a large barrier spit that has been created by deposition of sediment carried by longshore transport from the southeast. The shore of Lake Superior southeast of the bay has high bluffs of Miller Creek Formation which supply large volumes of sediment to longshore currents. The spit has grown since the lake level stabilized at its current level about 2000 years ago. Smaller spits first grew in the area where Oak Point is located today, and when the White River captured the Bad River 800 bp, the sediment supply increased, eventually leading to the development of the spit in its current position (Meeker, unpublished data). The power of the waves continues to deliver sediment and reshape the spit today. In historic times, it had been a series of unconnected barrier islands, thus its name Long Island (Bona, 1990; Collie, 1901). But deposition and erosion has caused it to alternate between being a barrier spit connect to the mainland, to being a barrier island. The last time it changed dramatically was in November 1975 when a large storm, the same storm that sank the Edmond Fitzgerald, caused deposition of enough sand to connect the sand bars into the large barrier spit present today.

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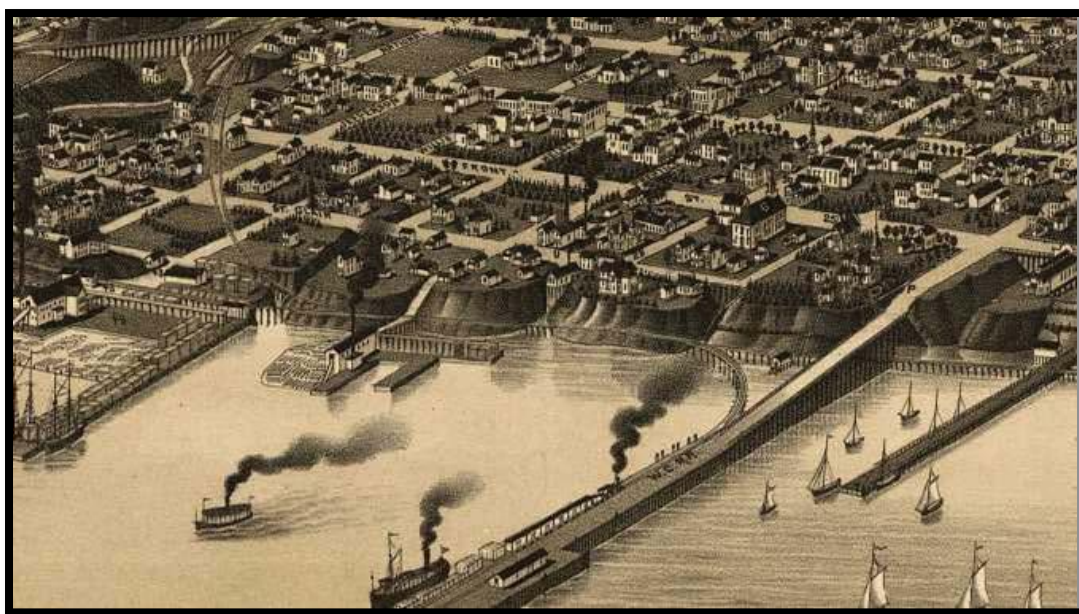
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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 4

GEOLOGY AND REMEDIATION AT THE ASHLAND/NORTHERN STATES POWER SITE



Ashland Lakefront circa 1900

Field Trip 4

Ashland/Northern States Power Company Lakefront Superfund Site

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As with many cities of the late 1800's, Ashland Wisconsin's lights and heat were supplied by manufactured gas. From 1887 into the 1940's a manufactured gas plant (MGP) operated under various owners, making gas and producing wastes. The Ashland MGP produced mainly carbureted water gas from feed stocks of coal and oil. MGP wastes included oil tars (coal tar) and wastewater as well as boiler ash, retort brick, and purifier box wastes.

The City of Ashland sits on the shores of Lake Superior's Chequamegon Bay. The MGP was located less than a block from the historic shoreline transected by a ravine that discharged to the bay. Over time the ravine was filled with wastes from the MGP and the lakebed area was filled mainly with wood waste from a series of saw mills that operated up to 1930.

Regional geology in the Ashland area consists of unconsolidated glacial deposits overlying Precambrian aged sedimentary bedrock. Unconsolidated deposits consist of the Miller Creek Formation, which overlies the Copper Falls Formation. Precambrian-aged sandstone of the Oronto Group, is the uppermost bedrock unit in the Ashland area. Thickness of the sandstone unit has not been determined. The Oronto sandstone is most likely underlain by Precambrian basalt.

Surficial soils in the Ashland area are underlain by red clay and silt deposits of the Miller Creek Formation, the predominant surficial lithologic unit in the Ashland area. The Miller Creek Formation is a fine-grained clayey silt to silty clay. It consists of lacustrine deposits and glacial till deposited during the last major advance of glacial ice. The thickness of the Miller Creek Formation in the Ashland area ranges from approximately 15 to 50 feet based on local well logs. The Miller Creek Formation encountered at the Site consists of clays and silts and range in thickness from 5.5 to 40 feet. The thinnest portion of the Miller Creek Formation is found at the base of the bluff where the former lake shoreline was located. This thin zone likely resulted from wave erosion of the Miller Creek Formation, forming the "bench" shaped bluff and former lake shoreline. Data from soil borings and downhole geophysics confirm that the Miller Creek Formation thickens to the north of the bluff. Sand and gravel layers interbedded with silty clay lenses have been encountered near the contact of the Miller Creek Formation and the underlying Copper Falls Formation. The Copper Falls Formation consists of interbedded glacial clays, sands, and gravels.

In addition to unconsolidated glacial deposits, fill soil units are present beneath the upper bluff area and Kreher Park. Fill soil encountered in the upper bluff area was placed in a former ravine that dissected the Miller Creek Formation in the vicinity of the former MGP

facility. Historic documents indicate that the ravine was filled by 1909. Kreher Park consists of fill material that was used to fill the former lakebed. Historic documents indicate that the former lakebed was filled beginning in the late 1800's, and continued to be altered by filling activities until the 1950's.

The ravine fill unit consists of silty clay fill material mixed with solid and liquid MGP wastes, ash, cinders, slag, and fragments of concrete and brick, glass bottles, scrap steel, and wire.

Fill material at Kreher Park consists predominantly of silty clay soil mixed with rocks, bricks, concrete, wood, and miscellaneous debris. A buried clay berm is also located along the shoreline on the northeast side of the Site near the former wastewater treatment plant WWTP. This fill unit ranges in thickness from 3 to 7 feet. A wood waste layer consisting of wood planks was encountered beneath the fill soil. Native soil underlying the fill unit includes a beach sand layer (predating filling of the lakebed) that ranges in thickness from 0 to 5.5 feet at Kreher Park. The Miller Creek Formation was encountered below the fill and beach sand and ranges in thickness from 7 feet at the MW-7 well nest to 40 feet at the MW-26 well nest. Offshore geology of the Chequamegon Bay inlet area consists of a discontinuous layer of submerged wood chips and debris of varying size on the lake bottom underlain by variably fine to medium grained sediments. The sediments are underlain by silts and clays of the Miller Creek Formation. The Copper Falls Formation was not encountered during investigation of the offshore sediments. Consequently, the thickness of the Miller Creek Formation below the bay is unknown.

Regional hydrogeologic units correspond to regional geologic units. The three aquifers that occur in the Lake Superior Basin in the vicinity of Ashland include the following:

- The Pleistocene sand and gravel aquifer (Copper Falls Formation);
- The Precambrian sandstone aquifer (Oronto Formation); and
- The Precambrian basalt aquifer.

The Copper Falls aquifer occurs between 25 to 55 feet below ground surface in the Ashland area. Based on deep piezometer boring information, the Copper Falls aquifer is over 150 feet thick. Sandy till units within the aquifer yield low volumes of water (5 to 10 gallons per minute gpm), while sand and gravel lenses can yield up to 100 gpm. The Copper Falls aquifer is confined by the overlying Miller Creek Formation near the Chequamegon Bay shoreline where the Miller Creek Formation has a higher plasticity and clay content. The low permeability Miller Creek Formation functions as an aquitard for the underlying Copper Falls aquifer in this area.

Based on a series of investigations of the site, MGP wastes including free product oils and tars are present within the filled ravine, Miller Creek Formation, Copper Falls Formation, Kreher Park (filled lakebed) fills, and approximately 17 acres of Chequamegon Bay. The contaminants within the sediments manifest themselves as slicks during windy conditions.

A pilot test, pump and treat system has operated in the Copper Falls Formation since 2001. In 10 years of operation over 11,000 gallons of oil/tar free product has been collected and over 2 million gallons of contaminated groundwater treated.

In consultation with the Wisconsin Department of Natural Resources, the USEPA has approved the Record of Decision for the remediation of contaminated soil, sediment and groundwater at the Ashland/NSP Superfund site. EPA's final cleanup plan includes:

- Removing soil from the most contaminated areas of Kreher Park and the Upper Bluff/Filled Ravine that overlooks the park area, treating the soil on-site and reusing it after treatment.
- Using barriers to contain and stop the movement of contaminants in ground water, treating the ground water in-place, and adding wells to extract and treat ground water.
- Digging up wood waste and contaminated sediment near and along the shore of Chequamegon Bay and dredging contaminated sediment in the bay, covering excavated and dredged sections nearshore and offshore with six inches of clean material, and treating contaminated sediment after removal.

The estimated cost for this cleanup plan is between \$84 million and \$98 million.



Photo 1: Product slick on Chequamegon Bay, Ashland, Wisconsin. Photo by Ed Monroe



Photo 2: Product slick on Chequamegon Bay, Ashland, Wisconsin. Photo by Ed Monroe

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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 5

BAD RIVER WATERSHED CULVERT RESTORATION PROGRAM



Hagger Road west culvert – before replacement.
-T. Fitz

Field Trip 5

Bad River Watershed Culvert Restoration Program

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INTRODUCTION

The Bad River watershed (BRW), a sparsely populated, largely forested area of northern Wisconsin, covers 700,000 acres in the Lake Superior basin. It is one of the largest watersheds in the Great Lakes basin. The BRW encompasses parts of Ashland, Bayfield and Iron Counties, draining approximately 1,094 square miles. With 1,140 miles of streams, the BRW supports one of the most diverse fish assemblages in the Lake Superior basin.

Watershed Geology

The BRW is divided into three primary drainage and topographic zones: the upper basin, transition zone and clay plain (Figure 1).

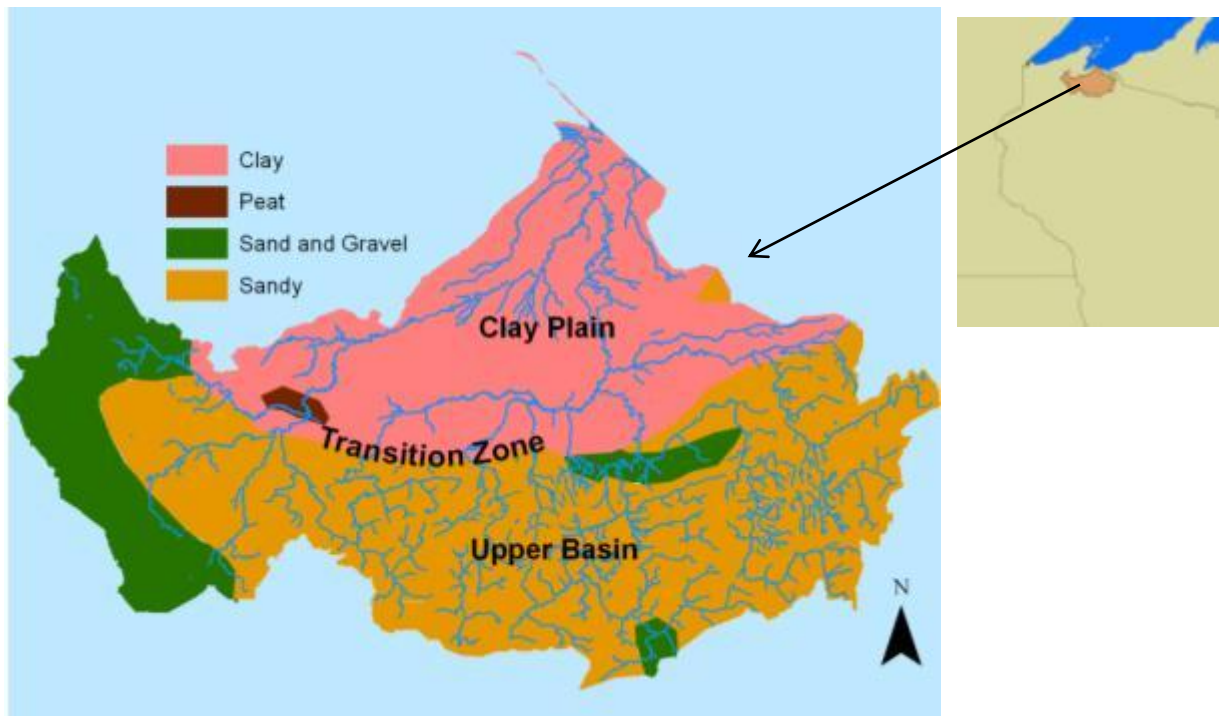


Figure 1. General watershed zones in the Bad River system

The upper basin is characterized by sandy glacial till with frequent rock outcroppings and a poorly developed drainage network with no valleys (Fitzpatrick, 2005). The lower portion of the watershed is known as the Lake Superior clay plain. This area is dominated by

clay till soils deposited across the landscape by glaciers and glacial Lake Duluth. Streams are often found in entrenched, alluvial valleys in this zone. The middle portion of the watershed that separates these two zones is known as the transition zone. This area is characterized by shallow sandy soils, as well as mixed sand and clay soils. The topography is characterized by steep unstable stream banks, entrenched valleys with steep ravines, with many springs and seeps (Fitzpatrick, 2005).

Culvert Program Description

Background

When culverts are employed to cross streams with roads, fish habitat may be significantly compromised in at least two ways. First, culverts can indirectly reduce habitat quality by accelerating sedimentation. Sediment can come from roadside ditches that empty directly into stream channels, in stream erosion associated with high water velocities, and bank erosion caused by misaligned crossings. In addition to these regular, smaller scale inputs, major crossing failures deliver large volumes of sediment into waterways. As a result of sediment loading from “blow outs,” stream channels can gradually fill in, further affecting fish habitat. As the stream substrate becomes covered with sand and fine sediment, both food and water quality can be diminished, and spawning gravels can be covered. The result can affect the growth and survival of fish. In addition, the Bad River watershed is the largest producer of sediment into Lake Superior from Wisconsin waters (Figure 2), largely as the result of non-point sources like failed crossings.



Figure 2. Mouth of the Bad River, and a destroyed culvert after a blowout event.

Second, culverts may block the ability of fish to move up and downstream. In this case, juvenile fish may not be able to reach food or find cover from predators in critical rearing habitats. Adult fish may not be able to reach spawning habitats. Fish may not be able

to retreat to tributary streams during floods or find relief in cold water during the heat of summer.

Consistent with the mission of the United States Fish and Wildlife Service (USFWS) and the Bad River Watershed Association (BRWA) to maintain the integrity of the Bad River Watershed, the culvert project was initiated to build community support and capacity to address problems at road stream crossings. The Culvert Program includes inventory, prioritization, coordination, restoration and monitoring elements in the overall goals to restore naturally functioning stream channels in the Bad River watershed.

Inventory

Inventory protocols were developed to identify sites that are likely fish barriers and/or significant contributors of sediment passage problems. Protocols included recording dimensions of the culvert, measurements from the stream channel, and conditions at the crossing that reflect how the stream and culvert interact. Partners throughout the region combined efforts to conduct surveys with local citizen volunteers, Ashland and Iron counties, US Forest Service, Northland College and the Bad River tribe contributing to survey efforts. To date, 996 of the 1,154 crossings in the watershed have been inventoried.

Prioritization

Site specific criteria have been developed for both fish passage and sedimentation concerns. Existing fisheries criteria include a drop greater than 6 inches and in-pipe velocity greater than 1 foot/second. Existing sedimentation criteria include sites with steep, un-vegetated and eroding embankments. Erosion of stream banks above and below crossings is used to estimate alignment issues. Town road crew knowledge of eroding or regularly washing out sites was also considered as a sedimentation criteria. Results are summarized into this Needs Assessment that outlines problem sites by sub basin. The Needs Assessment shows that blockages to fish migration and sedimentation problems at road stream crossing are widespread and identifies sites with the most significant problems. There are 164 crossings on the priority list that were identified using these criteria, 69 of which are on trout streams.

Strategic Plan

The primary purpose of this document is to provide local road and fish managers with information that can be used as a planning tool for crossing replacements.

Restoration

In the past 5 years, about 10 culvert replacement projects in the watershed have been designed and funded by fishery interests to ensure proper fish passage and to reduce excess sediment inputs into the system.

Monitoring

Assumptions have been made about the benefits of these projects, based on past experience, design standards and scientific literature:

- More fish pass through the culverts after replacement than before and this is a positive thing for fishery populations and fish communities.
- Erosion from sites has decreased immediately and in the long-term.

- Sediment from upstream head-cuts (created when new culverts are placed into channel beds rather than perched above them) move effectively through the culvert and are carried downstream, improving upstream and within-culvert habitat.
- Mobilized sediment from culvert replacement is not negatively impacting downstream habitat in gentle-sloped channels or coastal wetlands, such as the Bad River Slough, even though the influence of excess sediment load to downstream habitat quality is not known.

The BRWA and USFWS led an effort to test these assumptions by initiating a monitoring program of culvert replacements in the Bad River watershed. Companion fish, habitat, and channel stability protocols were being developed as a result of a coordinated monitoring workshop held in December 2008. The purpose of the workshop was to identify monitoring protocols that can be used to evaluate how well restoration work in the BRW is achieving program goals. At the workshop, participants (listed in Appendix A) developed specific recommendations for Culvert Program monitoring in the Bad River watershed. These recommendations included objectives, critical questions to be answered, parameters, and metrics for both physical and biological aspects of a monitoring program for the Culvert Program. Specific protocols followed are provided in Appendix B.

Monitoring Objectives and Questions

Objective 1: Reconnect artificially fragmented stream channels.

Question: Have we restored fish passage?

Objective 2: Determine changes in species assemblages associated with culvert replacements.

Question: Are there differences in species richness between upstream and downstream reaches of culverts prior to restoration?

Question: Are there differences between treatment and reference reaches prior to restoration?

Question: Are treatment reaches more similar to reference reaches post restoration?

Objective 3: Restore or improve instream habitat & prevent accelerated erosion and sedimentation

Questions: Has the channel morphology, slope, and sediment characteristics improved or restored (relative to reference reach), upstream, downstream and within the culvert and does this result in quality habitat? This question relates to the following sub-questions:

Is there deposition of fine sediment and bar formation?

Is the composition of streambed sediment appropriate for biotic health?

Is there habitat complexity in the channel geomorphic features? (riffles/pools)

Can bedload move appropriately through the culvert?

Is the stream aggrading/degrading or widening/narrowing?

Is the channel morphology similar to a preferred reference reach?

Are the banks eroding?

Education

Although many culverts that pose fisheries and sedimentation problems have been replaced in the Bad River watershed by natural resource professionals, road managers from towns, counties and state agencies are involved in a far greater number of replacements each year. Investment by road managers in fish friendly culvert installation techniques is essential to adequately restore connectivity and reduce sedimentation in the watershed. Therefore, the BRWA has invested in education and outreach activities to transportation managers to encourage consideration of stream channel integrity as much as road maintenance at road stream crossings. Thanks to BRWA and partner sponsored events, elected officials and/or road crew members from all 15 townships in the watershed have attended fish friendly culvert installation trainings. BRWA staff have prepared township specific summaries of culvert inventory data, and have worked with townships to strategically plan culvert replacements. Outreach sheets summarizing problems at existing culverts, restoration solutions and monitoring results have been prepared and distributed to all streamside landowners and town officials. Regular press releases and articles in local newsletters inform the general public of restoration efforts to conserve fish and their habitats, building public support for future restoration projects.

FIELD TRIP STOPS

We will be visiting three to four sites within the transition zone of the Bad River Watershed. At each stop, we will describe conditions at the crossing prior to restoration and compare the 3 different treatment techniques.

Stop 1: Baffled culvert on a Marengo Tributary

Location: Hager Road West - Latitude 46.398877 N Long. 90.931833 W (Fig. 3)

Description: The shallow, fast-moving water and drop at the culvert exit prevented fish from moving upstream. If a new pipe was set sufficiently low enough to eliminate the drop, the stream would have headcut substantially. Therefore the new culvert was installed at a 3.75% slope (based on natural channel conditions) with baffles throughout the pipe to prevent a velocity barrier (Figs. 4 and 5).

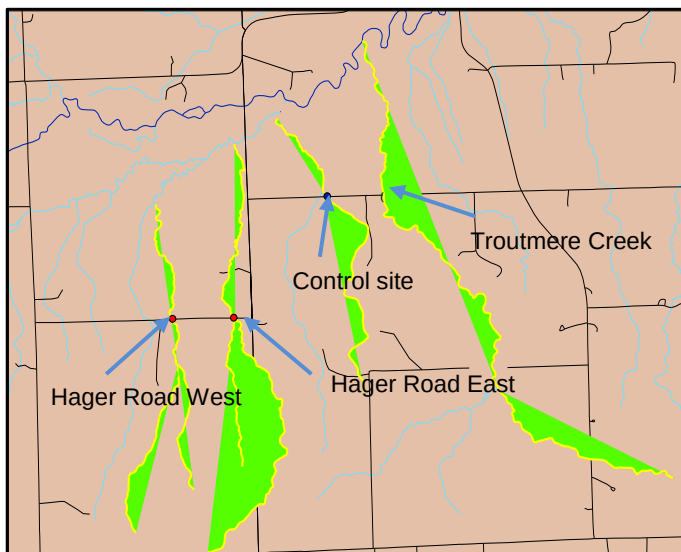


Figure 3. Map showing locations of Hager Road and Troutmere Creek sites.



Figure 4. Hagar Road west culvert; the photo on the left shows the old culvert before replacement, the photo on the right shows the new culvert.

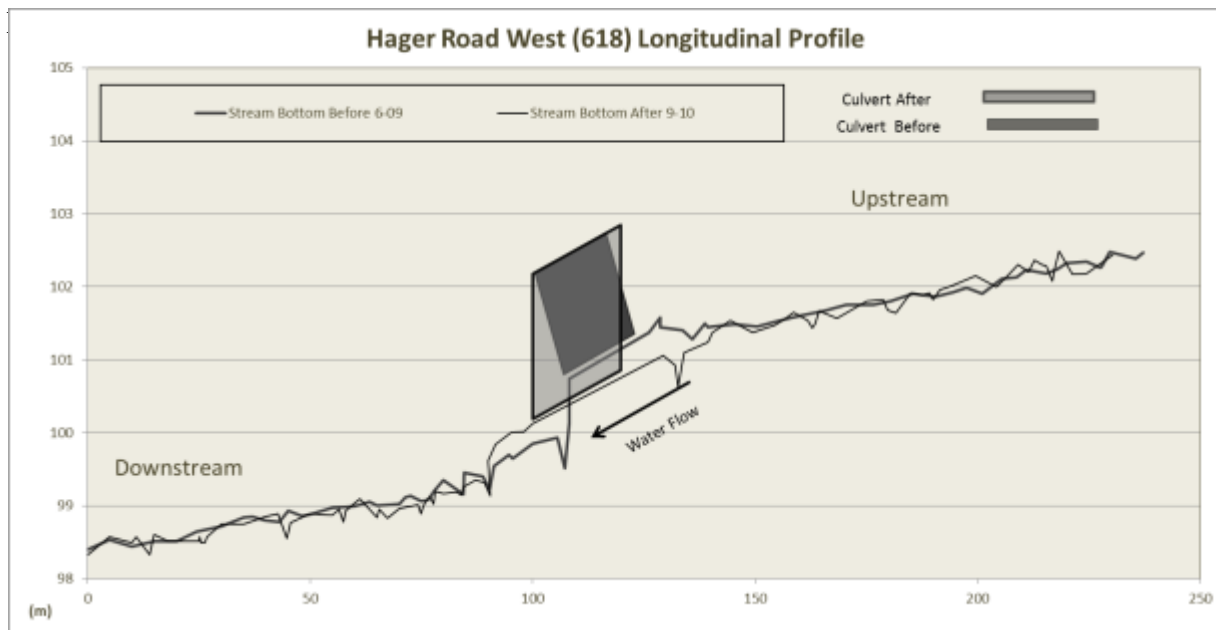


Figure 5. Longitudinal profile of the unnamed tributary of the Marengo River showing the stream profile before and after replacement of the Hagar Road west culvert.

Stop 2: Culvert Replacement on a Marengo Tributary

Location: Hager Road East -Latitude 46.399013 N Longitude 90.925979 W

Description: The culvert was old and needed to be replaced (Figure 6). There was no sand, gravel, or rock within the culvert to mimic a natural stream bottom. The shallow, fast moving water and plunge at the culvert exit prevented smaller fish from moving upstream. The old culvert was replaced with a new, larger culvert set into the stream bottom to simulate a natural channel. Larger culverts prevent erosion by allowing water to pass through during high flow events. Larger culverts also slow the water down, making it easier for fish to swim up through the pipe.

Figure 6. The Hager Road east culvert before replacement (top photo), and after replacement (bottom photo).



Stop 3: Raising the grade at Troutmere Creek

Location: County C - Latitude 46.406291 N Longitude 90.911559 W

Description: The pipe was in pretty good shape at this crossing (Figure 7). The 10-inch drop at the end of the culvert prevented many fish from accessing the 2 miles of stream above. Instead of replacing the culvert, the stream channel will be built up by placing large boulders in the stream below the culvert. This will cause sediment to fill in around the boulders, raising the stream bed and eliminating the 10-inch drop at the outlet of the culvert. It is anticipated that the boulders and sediment will form a terrace-like stream bottom similar to a staircase.



Figure 7. Culvert at Troutmere Creek.



Stop 4 (Time allowing): Control for Hager Road sites

Location: Midway (County Line) Road; Lat. 46.406302 N Long. 90.916723 W

Description: This site is a control site for transition zone restoration sites.

Discussion

Evaluation of the monitoring results calls for clarification and refinement of what we intend to achieve with our Culvert Program. Overall, monitoring set out to characterize site scale and broad scale population response to culvert replacement restoration. Movement protocols served to establish if passage was restored and if habitat was affected (site scale), and fish community metrics served to provide insights to population response (catchment scale). For the later to be effective, a better understanding of current fisheries habitat, distribution, status and limiting factors is required. In addition, a desired outcome, or fish population target, that is agreed upon among agencies is paramount to success.

At the outset of the program, natural resource managers recognized that these building blocks of effective management are currently lacking for the watershed. Monitoring protocols intended to improve understanding of population response to restoration given these constraints. Recommendations for the monitoring program are provided for watershed

scale context in evaluating restoration projects (catchment scale) and specifically for future culvert restoration projects (site scale).

Recommendation for monitoring at catchment level

- Increase examination of habitat-fisheries linkages
- Improve catchment level assessments as context for restoration benefits
- Continue fisheries monitoring to examine efficacy of various treatment sites

Monitoring protocols were developed with the intention of considering subwatershed scale effects of culvert replacements. Evaluation of monitoring program data has shown a good assessment of reach scale effects, however only a small proportion of the watershed has been sampled. At present, the status of fisheries habitat, fish distribution and status is not well known. In order to consider culvert restoration projects in a landscape scale context, additional data is needed. Existing modeling efforts may help provide this context.

Some work is underway through U.S. Fish and Wildlife Service (USFWS) initiatives to provide landscape level analysis on these topics. The USFWS Great Lakes Basin Fish Habitat Partnership is in the process of conducting habitat condition assessments for several species. Fish habitat condition can be estimated and changes in habitat conditions from restoration and protection actions could be tracked.

The USFWS is also currently involved in classifying catchments in the Lake Superior basin that maintain self-sustaining brook trout populations. Researchers have identified subwatershed metrics found useful in identifying changes in brook trout status. By compiling data from state and federal partners, this effort will describe the distribution, status, and stressors of brook trout throughout the region, including in the Bad River watershed. This landscape scale modeling work will include data inputs to ground truth predicted brook trout status in areas where status is unknown.

Upon review of monitoring program results, local managers recommend focusing local monitoring on efficacy of different structure types, in different zones of the watershed. In particular, the steep erosive channels in the transition zone area of the watershed raise question about longer term channel adjustments following culvert restoration. In particular, continued monitoring of sites with different culvert techniques (in pipe baffles, grade control below, or a larger pipe embedded into the stream bottom) will provide insights into the longer term efficacy of these techniques. This is a research oriented approach to evaluate culvert replacements in the different zones and different types of culvert designs.

This program has demonstrated that collaboration among groups is essential to implementing responsible watershed restoration programs over the long term. This collaboration assists all partners in achieving their respective objectives, in a way that maximizes watershed integrity.

ACKNOWLEDGEMENTS

The BRWA's Culvert Project has been part of an integrated effort in concept and structure from its inception. We sincerely thank all partners who have contributed to this project, including:

Bad River Band of Lake Superior Chippewa Indians, Natural Resource division, the Nature Conservancy, Sigurd Olson Environmental Institute, BRWA volunteer culvert inspectors, Ashland County Land and Water Conservation, Iron County Land and Water Conservation, Bayfield County Land and Water Conservation, U.S. Forest Service, Wisconsin Department of Natural Resources, U.S. Fish and Wildlife Service, U.S. Geologic Survey, Trout Unlimited, local townships, Natural Resources Conservation Service, and landowners adjacent to culvert sites.

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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 6

GEOLOGY HIKE AT COPPER FALLS STATE PARK



Devils Gate at Copper Falls State Park
-T. Fitz

Field Trip 6

Geology hike at Copper Falls State Park

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COPPER FALLS GEOLOGIC SETTING

Introduction

Copper Falls State Park, near Mellen, Wisconsin, is 2,676-acre park with hiking trails, scenic waterfalls, excellent bedrock exposures, and beautiful surficial geologic features (Figure 1). The Bad River flows from south to north through the park, and has cut through glacial sediments, eroding deeply into bedrock to form the Copper Falls gorge. After a 12-foot drop at Copper Falls, the Bad River travels about two miles through the gorge. Downstream, the Tyler Forks River plunges 30 feet over volcanic ledges to join the Bad River before flowing through the narrow passage of Devil's Gate. The gorge has excellent exposures of Keweenaw volcanic and sedimentary rocks overlain by Pleistocene glacial deposits. This guidebook explains the geology along the 1.7-mile Doughboys' Nature Trail that takes hikers around the top of the gorge (WDNR, 2009).

Copper Falls State Park was established in 1929, aided by the Civilian Conservation Corps and the Works Progress Administration. Several buildings at the park are from this era and contain some nice gabbro and slate building stones. While Copper Falls State Park was preserved for its natural beauty, people once sought native copper at the site. Though no significant deposits were discovered, mine shafts can still be found within the park. In the early 1900s, the Bad River was diverted to prevent flooding in the mine shafts; the original channel is visible along the trail at stop 8 (WDNR, 2009).

Bedrock Geology

The bedrock exposed in the park is part of the 1100 Ma Keweenaw Midcontinent Rift (MCR) sequence. The MCR was created when the Laurentian craton began to pull apart, resulting in huge outpourings of flood basalts within major rift valleys. Many thousands of feet of volcanic rock accumulated in the rift's central graben, but the tectonic activity stopped before an ocean basin could form. Volcanic activity was most voluminous between 1109-1094 Ma, and, with waning extension and volcanism, the rift filled with sediments. The MCR then underwent tectonic compression, and the central graben was partially inverted. As a result rocks along the southern arm of the rift now dip steeply to the north. One volcanic and three sedimentary bedrock units related to the MCR can be seen in Copper Falls State Park: the upper part of the Kallander Creek Volcanics, and the Copper Harbor Conglomerate, the Nonesuch Formation, and the conglomeritic facies of the Freda Sandstone, all part of the Oronto Group (Figure 2).

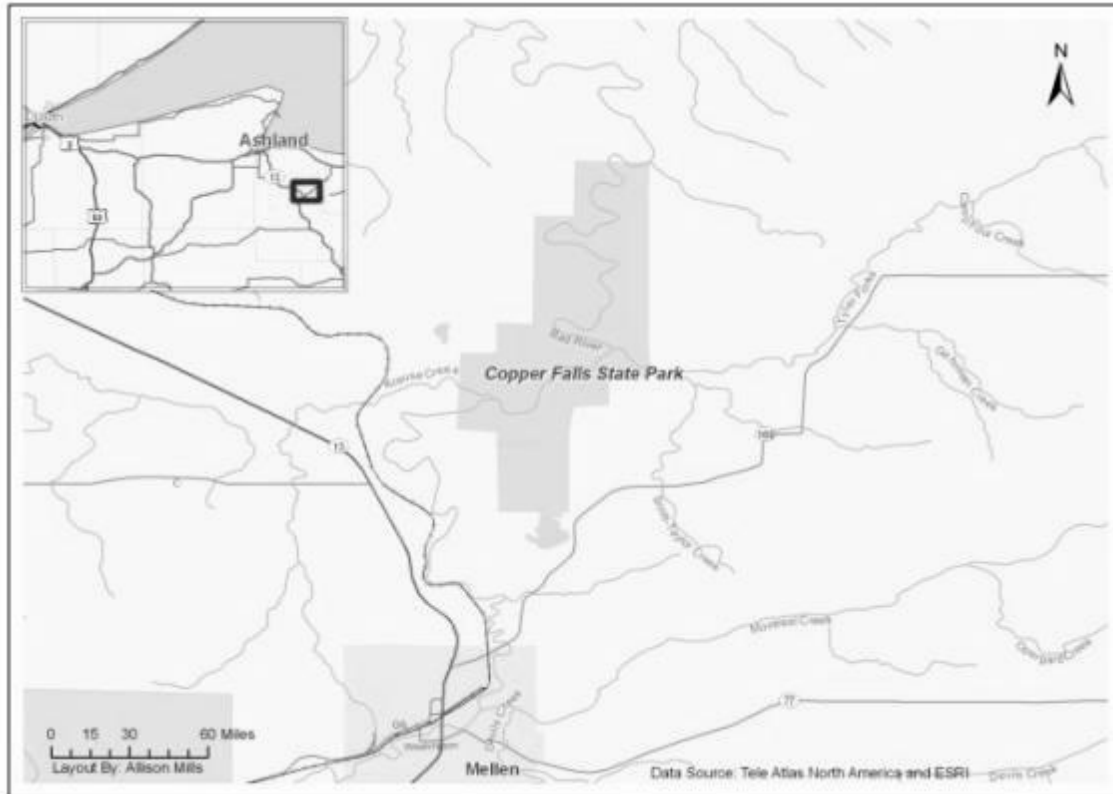


Figure 1. Location of Copper Falls State Park.

These rock units are situated on the southern flank of the Montreal monocline—a large structure created when the rocks of the central graben of the MCR were tipped north during compression of the rift about 1060 Ma ago (Figure 3). Erosion of the monocline has exposed more than 35 km of crustal rocks. Layering in the monocline strikes northeast-southwest and dips steeply north (Cannon et al., 1996; Cannon et al., 1993b; and Chapter 7 of this guidebook volume).

The Kallander Creek Volcanics consist of mostly basalt, with some andesite and rhyolite—the lowermost flows are flood basalts, while the upper sections are dominated by andesites and rhyolites—erupted from a large central volcano that was probably located in the region of Copper Falls Park (Cannon et al., 1993a). This central volcano acted as a topographic high during the early phase of volcanism of the MCR at the time, and somewhat after, the Kallander Creek eruptions. Overlying bedrock units pinch-out in the Mellen area and thicken to the east away from this area. The intrusive Mellen Complex is interpreted to have been emplaced within the volcano late in its magmatic activity. The time of igneous activity is known from U-Pb zircon ages—rhyolite in the middle of the Kallander Creek Volcanics has an age of 1107 $\pm$ 2 Ma, and granite in the Mellen Complex has an age of 1102 $\pm$ 2 Ma (Cannon et al., 1993a). Thus, the central volcano was probably active from about 1108 Ma to 1102 Ma ago (Cannon et al., 1993a).

<u>Formation</u>		Magnetic <u>Polarity</u> (Keweenawan Rocks)	<u>Age</u>
Copper Falls / Miller Creek			Pleistocene
Freda	Oronto Group	Normal	Mesoproterozoic
Nonesuch			
Copper Harbor			
Portage Lake	Bergland Group	Normal	
MellenIntrusions (1102 Ma)			
Kallander Creek Volcanics (~1108 - 1099Ma)	Powder Mill Group	Reversed	
Siemens Creek Volcanics		Normal	
Bessemer Formation		Reversed	
Tyler	Marquette Range Supergroup		Paleoproterozoic
Ironwood			
Palms			
Bad River Dolomite			
Puritan Batholith (2750 Ma)			Neoproterozoic
Ramsey Formation			

Figure 2. Stratigraphic table of formations in the Copper Falls area. The Kallander Creek through Copper Falls Formations are visible in Copper Falls Park. Note that the table shows rocks by age, not position – the Mellen Intrusions were mostly emplaced between the Siemens Creek and Kallander Creek Volcanics (after Green, 1982; Nicholson et al., 1997).

The eastward thickening of the units above the Kallander Creek is referred to as “The Wedge” on interpretive signs within the park. In fact, the Portage Lake Volcanics, just above the Kallander Creek Volcanics and below the Oronto Group, pinches out east of Copper Falls State Park and is not visible on this field trip. Therefore, going up-section, the sedimentary units above the Kallander Creek are the Copper Harbor Conglomerate, the Nonesuch Formation, and the Freda Sandstone, all of which are much thicker to the east in Michigan (Figure 2). The shale and siltstone of the Nonesuch is overlain by interbedded red and brown conglomerate and sandstone of the Freda. These units record the waning of volcanic activity in the MCR and the subsequent filling of the rift with sediments.

Rocks of the Mellen Complex are located nearby, but are not exposed in the area of the Copper Falls gorge. The Mellen Complex consists primarily of gabbro and granite intrusions that were emplaced along the base of the MCR volcanic pile. The southern part of Copper Falls park is underlain by granite of the Mellen Complex, and a small gabbro quarry

can be seen on the west side of State Highway 169 south of the park, and along State Highway 13 north of Mellen (see Chapter 9 of this guidebook volume).

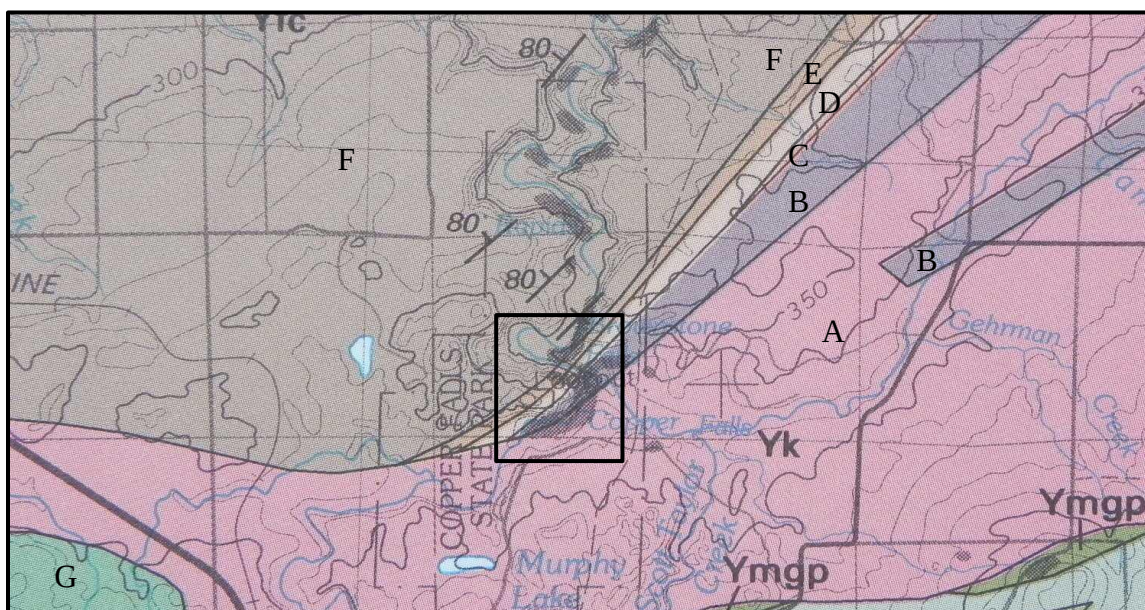


Figure 3. Bedrock geologic map of the Copper Falls area (from Cannon et al., 1996). Formations are: A, Yk) Kallander Creek andesite and basalt; B) Kallander Creek rhyolite; C) Portage Lake Volcanics, D) Copper Harbor Conglomerate; E) Nonesuch Formation; F) Freda Sandstone; G) Mellen Granite; Ymgrp) Granophyre of Potato River Intrusion in Mellen Complex. The inset box shows the area of this field trip.

Surficial Geology

Bedrock exposed in the park is overlain by Pleistocene glacial deposits of the Copper Falls and Miller Creek Formations that record the retreat of the last of the Wisconsin glacial lobes prior to 9,500 BP (Clayton, 1984). The Copper Falls Formation contains sandy till and outwash deposited by the Chippewa Lobe prior to 11,500 BP, whereas the younger Miller Creek Formation is comprised of clay-rich deposits associated with Glacial Lake Duluth that were deposited between 11,500 and 9500 BP (Figure 3). Areas occupied by these two different deposits are often separated by a ribbon of sand – part of the Miller Creek Formation – at an elevation of 1100 feet that marks the highest beach of Glacial Lake Duluth. Deposits of sandy till of the Copper Falls Formation cover the bedrock in the southern part of the park and can be seen in the banks of the Bad River along the field trip route. Here the Copper Falls Formation is a mix of sand and silt with some gravel and boulders but has considerable local variability. The topography in the vicinity of the park is primarily controlled by the form of bedrock hills that have little or no glacial sediment on top, separated by intervening valleys where the Copper Falls Formation is tens of feet thick (Figure 4). Fine-grained sediments of Miller Creek Formation form large banks just downriver of the field trip route.

Soils that developed atop these sediments on topographic highs, such as in the landscape of the park, are moderately well-drained. These soils support a mixed forest with abundant white pine, sugar maple, and hemlock. In spite of the permeable sediments in the region, topographic lows are poorly drained and have a high water table because of the deranged drainage pattern. In these wetter areas, soils support spruce, fir, and cedar forests. There are also numerous kettle lakes and bogs that have thick accumulations of peat.

Downstream (north) of the gorge the bedrock is covered by clay-rich sediment of the Miller Creek Formation. It is the distinctive red clay of the Lake Superior “clay plain” in this region. The red clay is hard and stable when dry, yet soft and unstable when wet, which results in the erosion of very steep banks in the meanders of the Bad River immediately north of the gorge and Doughboys’ Nature Trail.

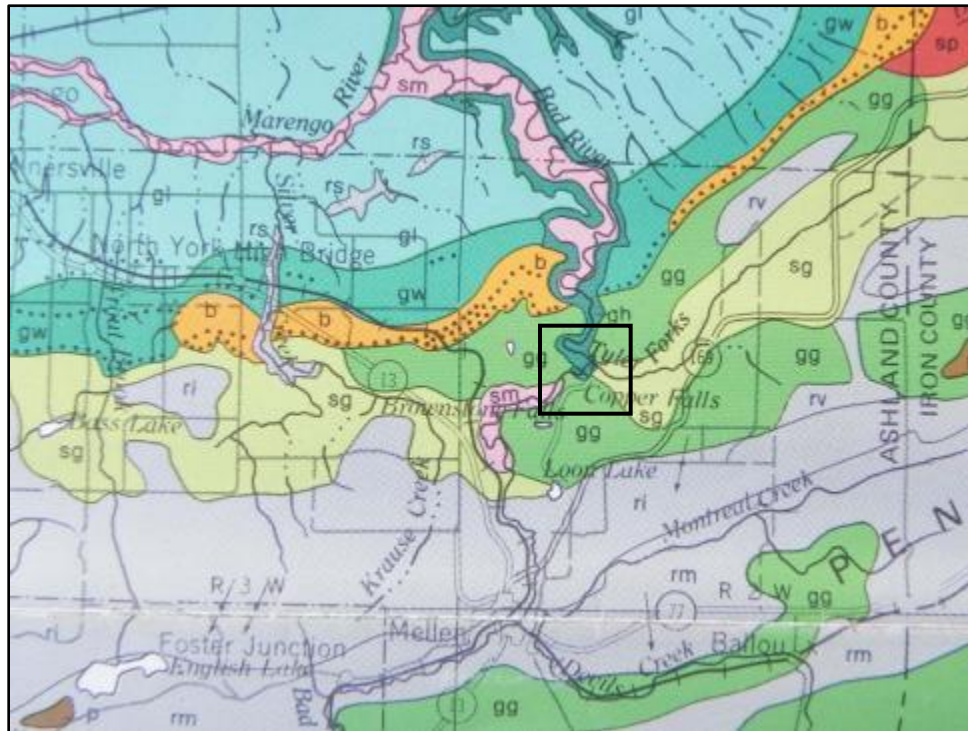


Figure 3. Surficial geologic map of the Copper Falls area (from Clayton, 1984). The area of this field trip is shown in the inset box. Areas represented in green are the Copper Falls Formation; those in blue-green are Miller Creek Formation. The strip of orange represents a beach deposit from Glacial Lake Duluth.

Hydrology

The Bad River starts at Caroline Lake, southeast of the park and the town of Mellen. The land in the headwaters of the Bad River is poorly drained and has numerous bogs. The water flowing from this headwater region is rich in tannins and organic acids derived from organic material in the peaty bogs. This gives the Bad River its characteristic brown color and is the source of the name “Copper Falls”.

From Caroline Lake the river flows southwest through a large area of poor drainage, then turns north and goes through the bedrock gorge of the Penokee Gap. The Bad River then flows through the City of Mellen, continuing north into the park. Within the park, the river meanders across a floodplain in a relatively flat area probably held in place by the nick point defined by the hard basalt at the head of the Copper Falls Gorge. The river then plunges over the ledge at Copper Falls and flows through the deep gorge. The presence of the gorge is likely due to nonresistant rock, which may be due to a combination of poor cementing in sedimentary rocks and closely spaced fractures in the felsite. Also, the rock may be highly fractured due to faulting.



Figure 4. Topographic map of the Copper Falls area. From U.S.G.S. maps of the Mellen quadrangle (1967) and the High Bridge quadrangle (1984).

FIELD TRIP STOPS

This field trip guide is for a walking tour of the Copper Falls State Park Doughboys' Nature Trail starting and ending at the parking lot of the concessions building near the river. The stops are listed in order going around the Copper Falls trail loop in a clockwise direction, although the hike can be done in either direction (Figure 5). Helpful interpretive signs mark many of the significant geologic features along the trail and will be mentioned here. Also, bring along binoculars; some of the outcrops are inaccessible by foot.

This area is represented on the Mellen and Highbridge topographic quadrangle maps (U.S.G.S.), on the Pleistocene map of Clayton (1984), and the bedrock geologic maps of Cannon et al. (1996) and Sims (1992). All locations are given in UTM coordinates (Zone 15 T), NAD 1927. Maps and brochures are also available at the park; some of which are accessible on-line through the Wisconsin DNR website.



Figure 5. Field trip stops 1-9 along the Doughboys' Nature Trail. Map courtesy of the Wisconsin Department of Natural Resources.

Stop 1. Bridge over the Bad River

Location: UTM coordinates: 681444 mE, 5137895 mN

The bridge is located past the concessions building and marks the trailhead for the Doughboys' Nature Trail.

Upstream of the bridge, there is a large bank of silty sand and some gravel of the Copper Falls Formation. Boulders eroded out of the slope make up the channel and banks of the river. There is a wide variety of lithologies in these boulders, reflecting the diverse bedrock geology of the region.

Rock exposed under the bridge is reddish gray basalt of the Kallander Creek Volcanics. The rock is slightly amygdaloidal and extensively fractured. The amygdules and fractures are filled with calcite stained red with hematite. During the Keweenaw eruptions, these flows were deeply buried by younger lava flows and sedimentary rocks, deep enough to undergo zeolite facies metamorphism. Oxidation of ferromagnesian minerals created abundant hematite that gives the rocks their dark reddish tint. No lava flow contacts are

visible in this part of the section, but the lava flows seen from the bridge are tipped to near vertical, dipping steeply to the north.

The bridge, built in the 1930s by the CCC, sits on a foundation constructed with Mellen gabbro. During mining activities in the early 1900s, the Ruggles Mining Company blasted the basalt exposed below the bridge in order to remove a small bedrock ridge to prevent flooding in the mine shafts. This allowed the river to flow directly east before turning north into the gorge; prior to blasting, the river flowed south around a big meander bend, visible at Stop 8.

Stop 2. Copper Falls Overlook

Location: UTM coordinates: 681510 mE, 5138011 mN

Along bend in trail with good view of the falls; history-focused interpretive sign nearby.

In 1902, Copper Falls was reportedly nearly 30-feet high, but several large bedrock pieces have broken off, and today the falls is about 12-feet high. During spring floods, the water can rise high enough to flow over the basalt island separating the channels.

The rock of the falls is basalt of the Kallander Creek Volcanics, similar to that seen at Stop 1. Look carefully to see if you can see any lava flow contacts or interflow sedimentary units. The interpretive sign here says that the gorge formed because of a fault—what are your thoughts?

Stop 3. Brownstone Falls Overlook

Location: UTM coordinates: 681682 mE, 5138203 mN

Along trail in bend that offers view of Brownstone Falls; interpretive sign nearby.

Brownstone Falls marks the confluence of Tyler Forks River and Bad River and reveals a contact between the mafic and felsic rocks of the Kallander Creek Volcanics. Felsic rocks are visible along the west side of the gorge, just left of the falls, apparent by their reddish color and closely spaced fractures. This quartz- and plagioclase-phyric unit is informally named the Sheep Farm rhyolite and is the uppermost unit of the Kallander Creek Volcanics. It is at least 300 m thick and is likely the product of a single eruption. It has been dated at 1099 Ma, and lies just above the horizon of the principle magnetic reversal of the Keweenaw from reversed to normal polarity (Zartman et al., 1996; Cannon, chapter 7 of this guidebook volume). Don't climb over the fence, but examine the rocks alongside the path, and see what you can find.

Along the trail, watch for some rare Northern Wisconsin vegetation: cutover-surviving white pines, eastern hemlock, and ravine-hugging cedars. The gorge, with its steep sides, has offered protection for the hemlock and cedar from over-browsing by deer. The cedars also show evidence of soil creep by the way their trunks curve down slope.

Stop 4. Devils Gate Overlook

Location: UTM coordinates: 681506 mE, 5138285 mN

Overlook along the trail above Devils Gate

Despite the name, this part of the field trip is really quite tame and the river's gradient has decreased. From this vantage, several formations of the Oronto Group can be seen across the river. The beds strike about 24 degrees northeast-southwest and dip northwest at 86 degrees. The Copper Harbor Conglomerate, characterized by mainly red and brown conglomerate, sandstone, and siltstone, is straight across the gorge; there is some crude

bedding apparent. These rocks represent a prograding alluvial fan complex in the MCR (Daniels, 1982). The Nonesuch Formation—up-section and to the left—is characterized by mostly gray to black shale, siltstone, and mudstone. These fine-grained sediments were derived from local sources and deposited in alternating lacustrine and fluvial environments (Cannon et al., 1993a). The conglomeritic facies of the Freda Sandstone, the farthest outcrop visible to the left, is characterized by red and brown interbedded sandstone and conglomerate with thick bedding. The clasts of this facies are approximately equal proportions of immature, subangular to well-rounded, and poorly sorted pebbles of felsic and mafic rocks (Cannon et al., 1996). These clasts indicate that volcanoes were still serving as the main source of sediment in the rift.

Stop 5. View of Devil's Gate from Bridge

Location: UTM coordinates: 681389 mE, 5138371mN

Looking upstream into Devil's Gate from the bridge is another view of the Oronto Group members. The name 'Devil's Gate' may come from the wicked turbulence and power of the Bad River in this constriction, especially notable during floods.

This stop is unique as three units of the Oronto Group are visible. The different units here are relatively thin as they pinch-out towards Mellen; the thinning is especially apparent when these same formations are compared at different locations, particularly as they thicken northeastwards into the Michigan Upper Peninsula.

The varying slopes indicate different rock resistance: the Nonesuch Formation has clearly eroded back, while the conglomerates and sandstones of the Copper Harbor and Freda have formed more resistant outcrops protruding into the river. The Copper Harbor Conglomerate is the farthest upstream, while the Freda Sandstone is closest to the bridge. From the next overlook along the trail, there is a good close-up view of the siltstones and sandstones of the Freda Sandstone.

Stop 6. Overlook of Tyler Forks River

Location: UTM coordinates: 681742 mE, 5138243mN

Follow the wooden bridge path down to the farthest overlook, which is almost directly over Brownstone Falls.

From this vantage, north of the confluence, we are looking upstream into the Bad River gorge with Tyler Forks River on the left, just to the right, Brownstone Falls plunges over a ledge of the Kallander Creek Volcanics. Felsic rock is visible on the right; a close-up of the outcrops seen at Stop 3. To the left is Tyler Forks Cascades—a series of short turbulent falls tumbling over felsite. A better view of the cascades can be seen at the overlook a short distance back along the wooden walkway.

Stop 7. Brownstone Falls: a Different Perspective

Location: UTM coordinates: 681723 mE, 5138192 mN

Past the bridge, along the trail, is another good view of Brownstone Falls from the east side of the confluence.

At this stop, there is another view of Brownstone Falls: to the right, the Tyler Forks River flows in from the northeast, joins the Bad River below the falls, and continues northwest through Devil's Gate. Across the gorge, looking towards Brownstone Falls, an overflow channel can be seen carved into the rock; this abandoned channel funnels water into the Bad

River during floods. More mafic and felsic rocks of the Kallander Creek Volcanics are visible—do you see any evidence of a contact?

Stop 8. Old Bad River meander

Location: UTM coordinates: 681477 mE, 5137861 mN

Along trail, just before the first bridge; old channel is to the left.

In the early 1900s, the Ruggles Mining Company put in several shafts near the current concessions building, prospecting for copper. To prevent flooding, workers blasted a new channel so the Bad River could be redirected. Today, the old meander bend is still visible, though trees and other vegetation have successively reclaimed the channel. As for the mining, although some of these same formations are rich in copper farther east, this endeavor proved unsuccessful.

Stop 9. Flagstones in the path at the concession building

Locally quarried slate flagstones in the sidewalk near the concessions buildings are from the Paleoproterozoic Tyler Formation. They show good examples of thin bedding, graded beds, concretions, and other sedimentary structures. Additionally, more slate and gabbro were used in the bridges, buildings, and along the trails, particularly on the steps. Around the corner from the first waypoint, on the west side of the concessions building, is a large chimney constructed of anorthositic gabbro of the Mellen Complex.

Stop 9a.

Location: UTM coordinates: 681383 mE, 5137847 mN

Along the paved path, near the southwest corner of the concessions building kitchen.

This particular location, along the south wall of the concessions building, has several flagstones that show thinly interbedded mudstone and sandstone. Look for interesting sedimentary structures in other flagstones. The angle of slaty cleavage varies from being parallel to bedding to being perpendicular to bedding, with many angles in between.

Stop 9b.

Location UTM coordinates: 681330 mE, 5137846 mN

Along the paved path, closer to the parking lot.

This location, closer to the parking lot, has some good examples of graded bedding. All of this slate comes from the Paleoproterozoic Tyler Formation. The Tyler Formation crops out sporadically south and east of the park, and is considerably down-section of the Keweenaw rocks found in Copper Falls State Park.

ACKNOWLEDGEMENTS

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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 7

GEOLOGY OF THE MONTREAL RIVER MONOCLINE



Steeply dipping Freda Sandstone at the mouth of the Montreal River
-W.F. Cannon

Field Trip 7

Geology of the Montreal River Monocline

William F. Cannon

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Note: This fieldtrip was originally run for the 42nd Annual Institute on Lake Superior Geology in 1996. At that time, poor road conditions including deep snow on some back roads, “vehicular traction issues”, and high water in streams prevented a complete examination of all intended localities. New developments in the past 15 years and hopes for better field conditions in 2011 lead us to rerun the trip for the 57th Annual Meeting. The Field Guide is based largely on the original publication in 1996 (Cannon, 1996) with a few modifications based on new exposures and recent studies.

INTRODUCTION

South of Lake Superior, near the Wisconsin-Michigan border, a remarkable monoclinical sequence of rocks exposes, at the present land surface, what had been the upper 35 km of the crust (Figure 1) prior to uplift and monoclonal tilting during inversion of the Midcontinent rift. This section is seen in a sequence of steeply to vertically dipping and north-facing volcanic and sedimentary units of Paleoproterozoic and Mesoproterozoic age and underlying Neoproterozoic volcanic rocks.

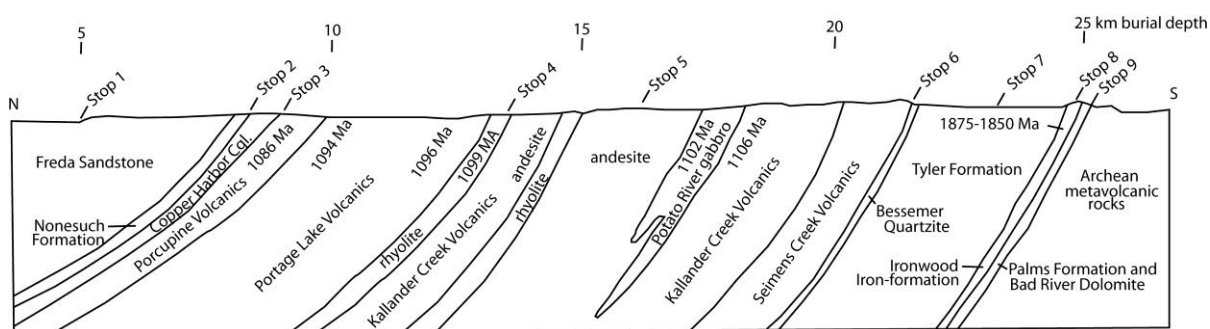


Figure 1. Cross section of the Montreal River monocline showing the stratigraphic position of field trip stops and approximate burial depths.

This structure, named the Montreal River monocline by Cannon and others (1993), resulted from southward-directed thrusting on a northward-dipping listric thrust of crustal scale, the Atkins Lake-Marenisco fault (Figure 2). The age of thrusting is known to be about 1040-1060 Ma, the date of reset biotite Rb-Sr ages in Archean rocks that were upthrust and cooled on the upper plate above the fault (Cannon and others, 1993).

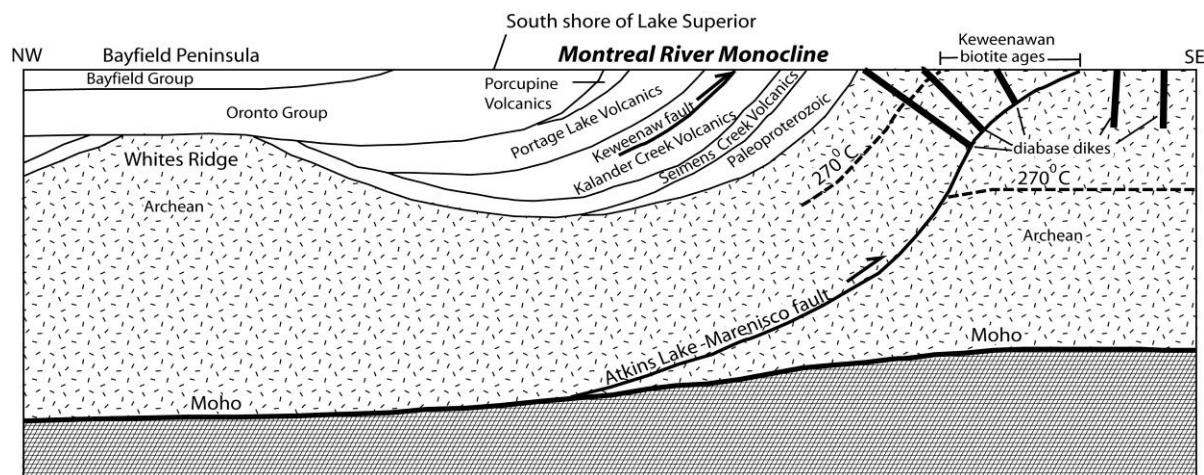


Figure 2. Schematic cross section of the Montreal River monocline showing the inferred crustal-scale structures responsible for its formation.

Figure 2 shows the regional structural context of the Montreal River monocline. The section extends approximately from the Bayfield Peninsula on the northwest to Mercer, Wisconsin on the southeast, a distance of roughly 100 km. To the northwest, a prominent basement high, White's Ridge, limited the depositional extent of much of the Keweenaw volcanic section and remained a positive topographic feature until the close of the extensional phase of the rift. As more widespread thermal subsidence supplanted extension, the ridge was buried beneath the Oronto and Bayfield Groups, which remain nearly flat-lying. Southeastward from White's Ridge to the present outcrop belt of the Keweenaw volcanics, the volcanic section becomes progressively thicker. This geometry suggests that the deepest parts of the rift in this area may actually have been to the south of the preserved volcanic rocks, in an area since uplifted and deeply eroded. The present steeply north-dipping attitude of the rocks of the Montreal River monocline is believed to be the result of crustal-scale ramping on the Atkins Lake-Marenisco fault. The Atkins Lake-Marenisco fault is interpreted to penetrate the entire crust and to have thrust rift units and their Paleoproterozoic and Archean basement southward. The listric geometry of the fault surface caused the upper plate to tilt northward, after which erosion has exposed the present crustal-scale cross section. The involvement of Archean rocks in the monoclonal structure is shown by the southerly dip of originally vertical diabase dikes and by Mesoproterozoic Rb-Sr ages of biotite, which mark the time of uplift and cooling. The 270° isotherm, the Rb-Sr blocking temperature of biotite, was located by determining the Rb-Sr age of biotite in Archean rocks. Rocks north of the isotherm (originally above the isotherm) remained cooler than 270° throughout Keweenaw burial and retained ages reflecting Paleoproterozoic and Archean conditions. Rocks between the isotherm and the Atkins Lake-Marenisco fault were heated above 270° by Keweenaw burial and the Rb-Sr ages were reset. In this area biotite ages

range from 1040-1060 Ma and date the time of uplift, erosion, and cooling caused by thrusting along the Atkins Lake-Marenisco fault. South of the fault Mesoproterozoic burial was not deep and biotite ages continue to reflect Paleoproterozoic thermal events.

The trip begins at the mouth of the Montreal River (Figure 3) where extensive exposures of vertically dipping Freda Sandstone, the youngest unit exposed in the monocline, form bluffs along the Lake Superior shore. Based on projections of seismic data from Lake Superior these strata may have been buried by as much as 5 km of overlying clastic rocks now preserved beneath the lake. The trip progresses southward for about 25 km through successively older and steeply dipping to vertical strata of the Mesoproterozoic Keweenaw Supergroup, the Paleoproterozoic Marquette Range Supergroup, and into Neoproterozoic metavolcanic rocks. The Paleoproterozoic and Neoproterozoic rocks seen at stops 8 and 9 have been exhumed from lower crustal depths, but lack the expected high metamorphic grade and deformation typical of rocks from such depths. The generally low-grade metamorphism and lack of Mesoproterozoic penetrative deformation attest to a short residence time at these depths. Rapid subsidence during rifting and rapid uplift during rift inversion brought these rocks to great depth and returned them to shallower depths at a rate that did not allow conductive heating to achieve typical deep crustal temperatures.

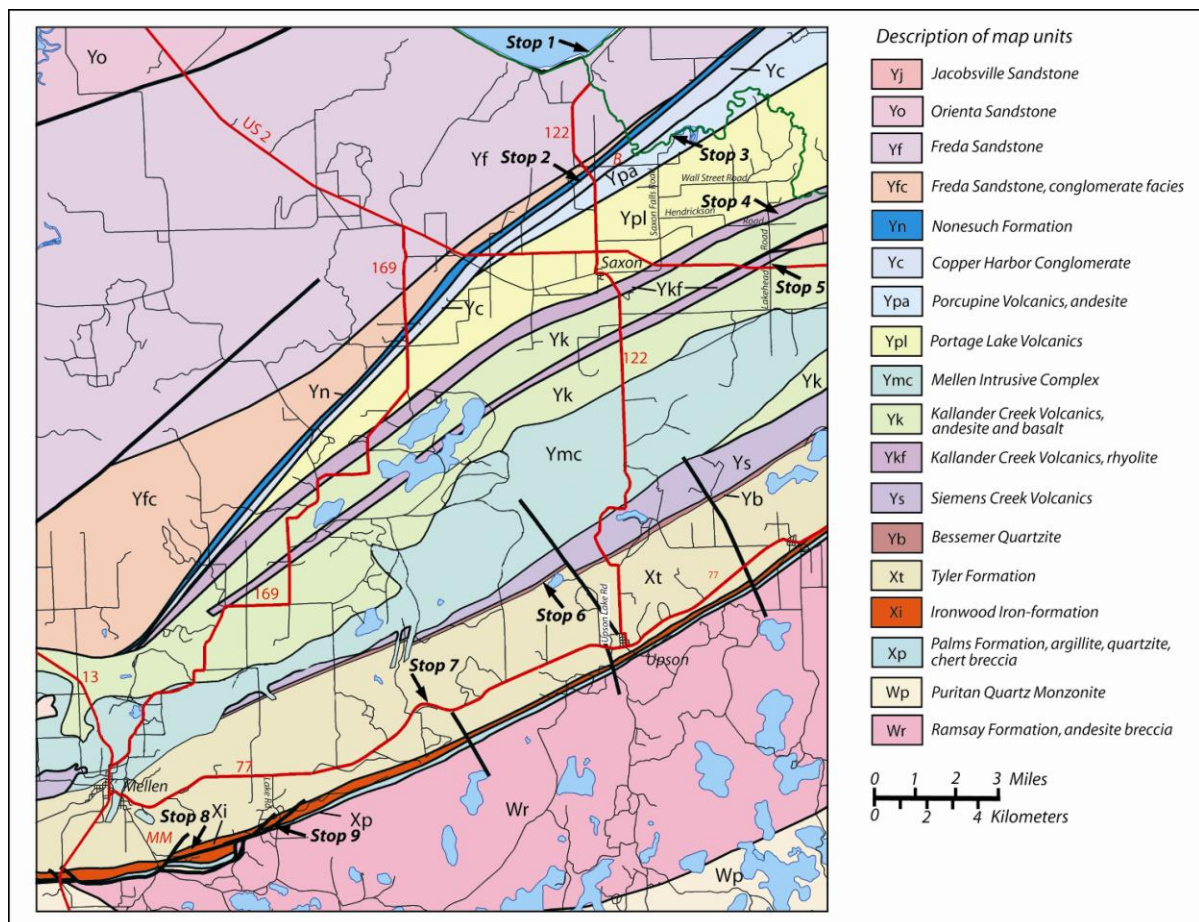


Figure 3. Geologic map of the field trip area showing location of stops. Geology generalized from Cannon and others (1996).

Stop 1: Freda Sandstone at the mouth of the Montreal River

(46.566, -90.4169) From the parking area at the top of bluff overlooking Lake Superior walk down the trail to exposures along the lakeshore.

The Freda Sandstone is the youngest unit exposed in the Montreal River monocline. The rocks here, representing the middle to upper parts of the Freda, may have been buried by as much as 3-5 km under younger parts of the Freda and younger sediments of the Bayfield Group. The sandstone is well exposed in high continuous bluffs along the Lake Superior shoreline. It is mostly a well bedded and commonly crossbedded feldspathic sandstone. Dips are essentially vertical. Volcanic rock fragments are a major constituent, indicating that rift-derived volcanic rocks composed much of the source area. Muscovite is also a common detrital mineral in many layers indicating that an Archean basement terrane was also exposed that contributed a substantial part of the detritus.

Stop 2: Middle part of Oronto Group along Parker Creek

(46.5241, -90.4199) These exposures are best reached by walking west from Highway 122 to Parker Creek. The coordinates are approximately at the contact between the Copper Harbor Conglomerate and Nonesuch Formation. These exposures are on private land.

The top of the Copper Harbor Conglomerate, the Nonesuch Formation, and the lower part of the Freda Sandstone can be seen along the incised bed of Parker Creek. The Parker Creek section is the most complete exposure of the Nonesuch and adjacent units in the Lake Superior region. Beds dip from 75°N to vertical. Detailed stratigraphy of the Parker Creek section was described by Suszek (1991) on whose work the following generalized description is based. The upper part of the Copper Harbor Conglomerate consists of massive to crudely bedded clast-supported conglomerate with cross-bedded, pebbly sandstone interbeds. Clasts consist mostly of basalt and rhyolite with lesser granite and gabbro, all probably derived from slightly older Keweenawan rocks. Banded iron-formation, jasper, quartzite, and chert clasts are a minor component but are widespread, indicating that pre-Keweenawan basement was also exposed to the south within the drainage basin.

The Nonesuch Formation is about 130 m thick and comprised of mostly gray to pale brown siltstone, sandstone, and minor shale. It is generally thinly bedded to laminated, but thicker massive beds also occur. Small-scale trough crossbedding is common. There is an overall coarsening-upward trend to the formation, although many individual units display fining-upward textures. The fine-grained rocks of the Nonesuch Formation lie on coarse conglomerate of the Copper Harbor. A transitional bed, one meter thick, contains reddish-brown, trough-crossbedded sandstone at the base and grades upward into gray-brown calcareous siltstone. The upper contact of the Nonesuch with the Freda Sandstone is gradational over about 20 m, in which gray siltstone and shale are interbedded with reddish-brown, medium to coarse sandstone.

Stop 3: Top of the Porcupine Volcanics at Saxon Falls

(46.5380, -90.3751) Drive to the end of the road to Saxon Falls dam. The exposures are along the river bed below the dam.

Basalt flows and interflow conglomerate and sandstone of the uppermost part of the Porcupine Volcanics are exposed along the Montreal River below the dam at Saxon Falls flowage. The uppermost exposed unit is thin-bedded red sandstone and siltstone, which is underlain by a massive basalt flow about 100 m thick. This flow is underlain by about 30 m of interbedded conglomerate and sandstone. The lowest exposed unit is another basalt flow with an amygdular, partly brecciated top. About the upper half meter of the flow shows typical rubbly flow-top breccia, and the upper three meters are coarsely amygdular. The most abundant amygdule-filling mineral is thomsonite, which occurs in masses of coarse radiating blades in the larger amygdules. The predominance of this low temperature zeolite indicates that rocks at this level of burial were only weakly heated before uplift.

Stop 4: Rhyolite at top of Kallander Creek Volcanics

(46.5108, -90.3265) This exposure is on private land on an active farm. Permission for access should be requested from the land owner.

The uppermost unit of the Kallander Creek Volcanics is a thick, mostly massive, quartz- and plagioclase-phyric rhyolite. We informally designate this unit the Sheep Farm rhyolite and this stop is our informal type locality. The unit is at least 300 m thick and appears to be the product of a single eruptive event. It can be traced from this locality westward for about 30 km to Copper Falls State Park near Mellen, where it is the resistant rock that holds up Brownstone Falls on the Tyler Forks River. At that locality, we have located by field measurement the horizon of the principal magnetic reversal recorded in the Keweenaw section. The reversal lies a few flows beneath the rhyolite. The rhyolite has been dated at 1099 Ma (Zartman and others, 1996), so the age of the magnetic reversal must also be approximately 1099 Ma. At this stop, the exposures reveal the basal part of the flow, which contains numerous blocks of rhyolite up to about 20 cm across. These blocks have a slightly different texture and phenocryst content than the host rhyolite.

Stop 5: Icelandite in Kallander Creek Volcanics

(46.4936, -90.3292) The exposures to be examined are in a gravel pit to the northeast of the intersection of Lakehead Road and US Highway 2

The upper member of the Kallander Creek Volcanics contains abundant rocks of intermediate composition in addition to basalt and rhyolite. We have proposed that the upper part of the Kallander Creek Volcanics represents the partly eroded remains of a broad central volcano, and that the Mellen Complex is the crystallized remnants of a magma chamber that was intruded into the volcanic edifice late in its eruptive history (Cannon and others, 1993).

At this locality our data show that the rock knob in the gravel pit is a fine-grained reddish-brown icelandite ($\text{SiO}_2=64\text{--}66$ wt%; $\text{TiO}_2=1.26$ wt%; $\text{Ce}_N/\text{Yb}_N=4.5$) that contains small plagioclase phenocrysts. The rock has a strong primary foliation, perhaps indicating that it is the basal zone of a flow, and weathers into platy fragments parallel to the nearly vertical foliation. Small outcrops nearby show a coarser, more massive icelandite. The icelandite is overlain by a basalt ($\text{SiO}_2=50.24\text{wt}\%$; $\text{TiO}_2=3.78$ wt%; $\text{Ce}_N/\text{Yb}_N=3.7$) that is exposed in low outcrops along the power-line north of the gravel pit.

Stop 6: Upson Lake area—Siemens Creek Volcanics, Bessemer Quartzite, and Tyler Formation

(46.3867, -90.4442) Drive road along the south shore of Upson Lake to the small parking area at the west end of the lake. Walk north and east along logging road about 300 meters and then north to exposures on higher part of the south-facing slope.

At this locality the base of the Keweenaw section is exposed in contact with the underlying Tyler Formation. At the top of the prominent bluff north of Upson Lake the lowermost basalt flows of the Siemens Creek Volcanics are exposed. They are pillowed and brecciated as a result of eruption into a shallow lake in which the Bessemer Quartzite had been deposited. The flows have a distinctive chemical composition marked by low Al_2O_3 (8–12 wt%), and high Cr, Ni, and MgO. In addition, rare earth element (REE) patterns are quite steep (Ce_N/Yb_N = about 15) and heavy REEs have low abundances. These chemical characteristics indicate that the magmas formed by a small degree of partial melting of a deep, garnet-bearing enriched mantle source (postulated to be a mantle plume), and that the magma erupted directly to the surface without undergoing significant crystal fractionation. Flows of this chemical character can be recognized in the field by the presence of small phenocrysts of clinopyroxene. These clinopyroxene-phyric flows have been traced from near Bessemer, Michigan, to just west of this locality. They apparently represent a unique magma composition that was erupted only at the very outset of volcanism. Identical flows have been recognized at the base of the North Shore Volcanics in Minnesota, both at Ely's Peak and at Pigeon Point and Grand Portage (Green, 1977). An interflow sedimentary bed is exposed between the first and second flows. The sediments are, in part, an algal laminite and are now substantially metamorphosed so that original calcareous layers are a fine-grained mat of wollastonite. The metamorphism was caused, at least in part, by emplacement of the Potato River Gabbro about 1 km up-section.

The Siemens Creek Volcanics are underlain by the Bessemer Quartzite, the oldest unit in the Keweenaw Supergroup. Scattered exposures of white to gray laminated quartzite are on the slope below the basalt exposures. Passing down-section, the Bessemer contains basal beds of coarse conglomerate composed largely of Paleoproterozoic metasedimentary clasts, including iron-formation. Small garnet porphyroblasts are common in the matrix. A small exposure near the base of the slope shows an angular unconformity between basal conglomerate of the Bessemer and thin-bedded argillite of the Tyler Formation. This is the only exposure known to us where the unconformable relationship between the Keweenaw Supergroup and the Marquette Range Supergroup can be seen.

Stop 7: Tyler Formation

(46.3500, -90.5051) Exposure is a roadcut on the north side of High 77.

This roadcut exposes typical thin-bedded graywacke and argillite of the Paleoproterozoic Tyler Formation. Although dips are nearly vertical, the lack of penetrative fabric indicates little structural effect of the Penokean orogeny on the Tyler in this area. The highest grade metamorphic mineral is biotite, which is common throughout the matrix.

Stops 8 and 9: Mt. Whittlesey area

The prominent ridge of Mt. Whittlesey (Figure 4) is held up by the Ironwood Iron-formation. An unusually great thickness of iron-formation here resulted from structural repetition by thrust faults and folds produced during the Penokean orogeny. Mt. Whittlesey lies at the easternmost extent of the Penokean (Paleoproterozoic) foreland fold and thrust belt. Here, a series of thrusts repeats the Paleoproterozoic section, having detached it from Archean rocks along a basal decollement. The thrusts all rise in the section to the east along a series of lateral ramps. Displacement along the basal decollement diminishes eastward as splays from the decollement successively rise in the section. East of Stop 9 the Paleoproterozoic section lies unconformably on Archean rocks.

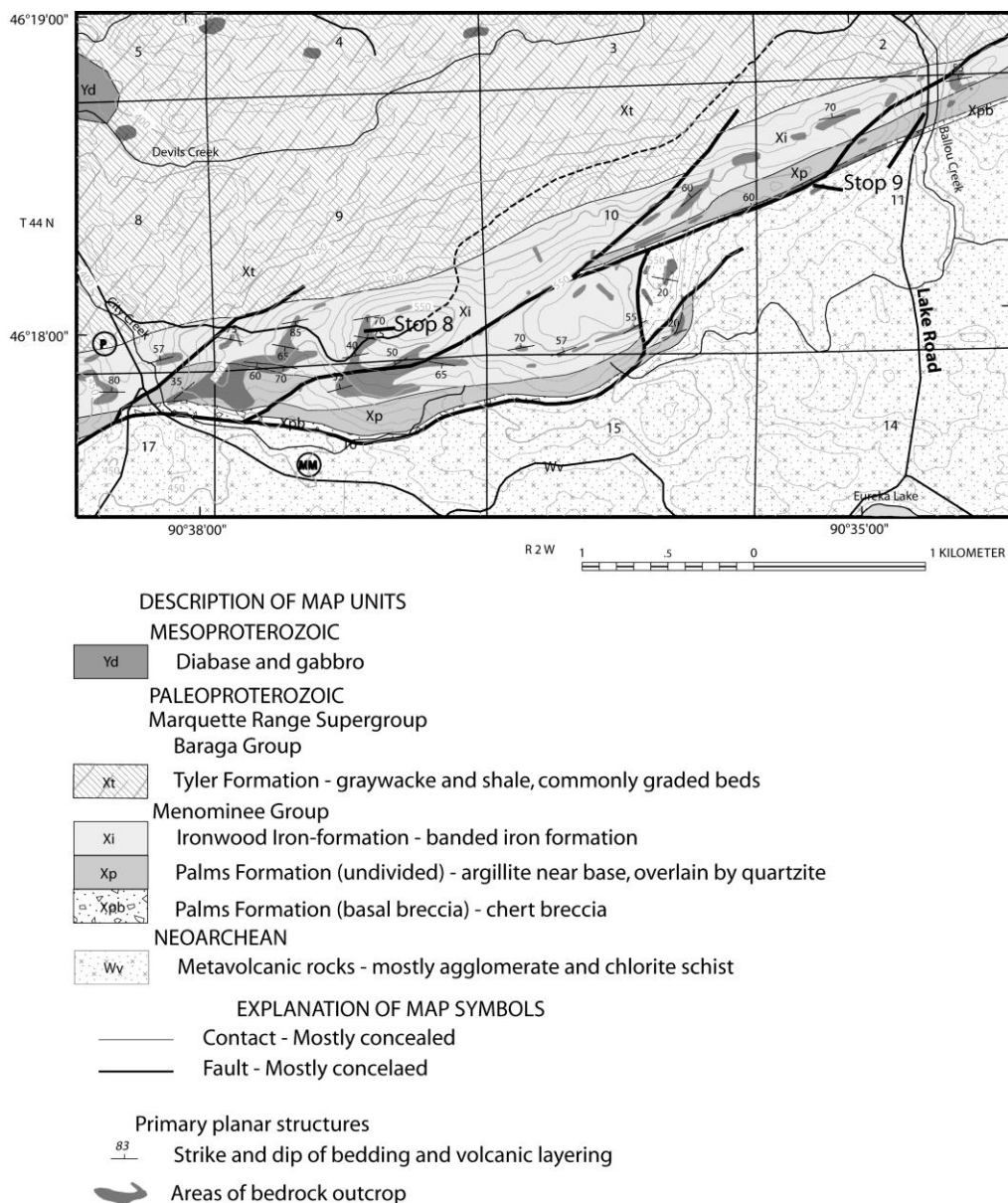


Figure 4. Geologic map of the Mt. Whittlesey area showing the locations of Stops 8, and 9.

Stop 8: Ironwood Iron-formation in Mt. Whittlesey

(46.30226, -90.61633) Vehicle access is via a poorly maintained road that extends eastward from County Road MM. Some steep sections of road are susceptible to washouts and may be impassable at times. From the Berkshire mine ruins walk eastward on an ATV trail up the slope to large bedrock exposures.

Excellent exposures of the Ironwood Iron-formation occur on the north face of the ridge west of the summit, and along a railroad grade near the Berkshire mine ruins. At the former locality, a large area was cleaned of overburden in the 1920s in anticipation of open pit mining. Only a small pit was developed near the base of cleared area; the remainder provides an excellent, glacially polished exposure in which many details of sedimentary

features are extraordinarily well displayed. At the Berkshire mine ruins an abandoned railroad grade provides a continuous exposure of several hundred meters of section and a tight fold is exposed in the cut.

The iron-formation contains several lithologic variations. The classic five member internal stratigraphy defined to the east (Hotchkiss, 1919; Huber, 1959) consists of alternating members of even-bedded, carbonate iron-formation (slaty iron-formation) and irregularly-bedded hematitic jasper. This stratigraphy becomes more complex and irregular westward along the iron range; we have not been able to apply it west of Mt. Whittlesey. The iron-formation exposed here characteristically has many lithologic variations on a small scale so that individual exposures may show a complete range of lithologies interbedded on a scale of meters or less.

The following detailed description of the Ironwood exposed on the northwest flank of Mt. Whittlesey was prepared by Philip Fralick as part of an informal guidebook for the Workshop on Geology, Mineralogy, and Genesis of Precambrian Iron-Formations held at the University of Minnesota-Duluth Precambrian Research Center in October 2010. It is based largely on a detailed study of these exposures reported in Pufahl and Fralick (2004).

Scattered outcrops in the vicinity of Mt. Whittlesey define three upward-coarsening, asymmetric cycles composed of a basal unit of slaty iron-formation that is gradational into an upper unit of trough cross-stratified cherty iron-formation. The middle cycle provides an unparalleled opportunity to investigate cycle sedimentology as it is completely stripped of overburden to appraise the Fe resources of this area. Like the cycles above and below, it consists of ~14 m of slaty iron-formation gradationally overlain by ~16 m of cherty iron-formation, which is capped by a ~1 m thick unit of convolute-bedded, slaty iron-formation.

Pufahl and Fralick (2004) conducted a detailed description of a 40x44 m glacially polished exposure through 1.5 cycles combined with a reconnaissance study of poorly exposed strata above and below the logged section. This area provides an excellent opportunity to investigate the depositional controls governing the accumulation of iron-formation facies. Their research was augmented with geochemical analysis of iron-formation facies in order to elucidate the chemical controls on iron-formation precipitation. This approach has yielded important insights into the relative roles tectonics, fluctuating hydraulic regime, and changing water chemistry played in governing the distribution of iron-formation facies.

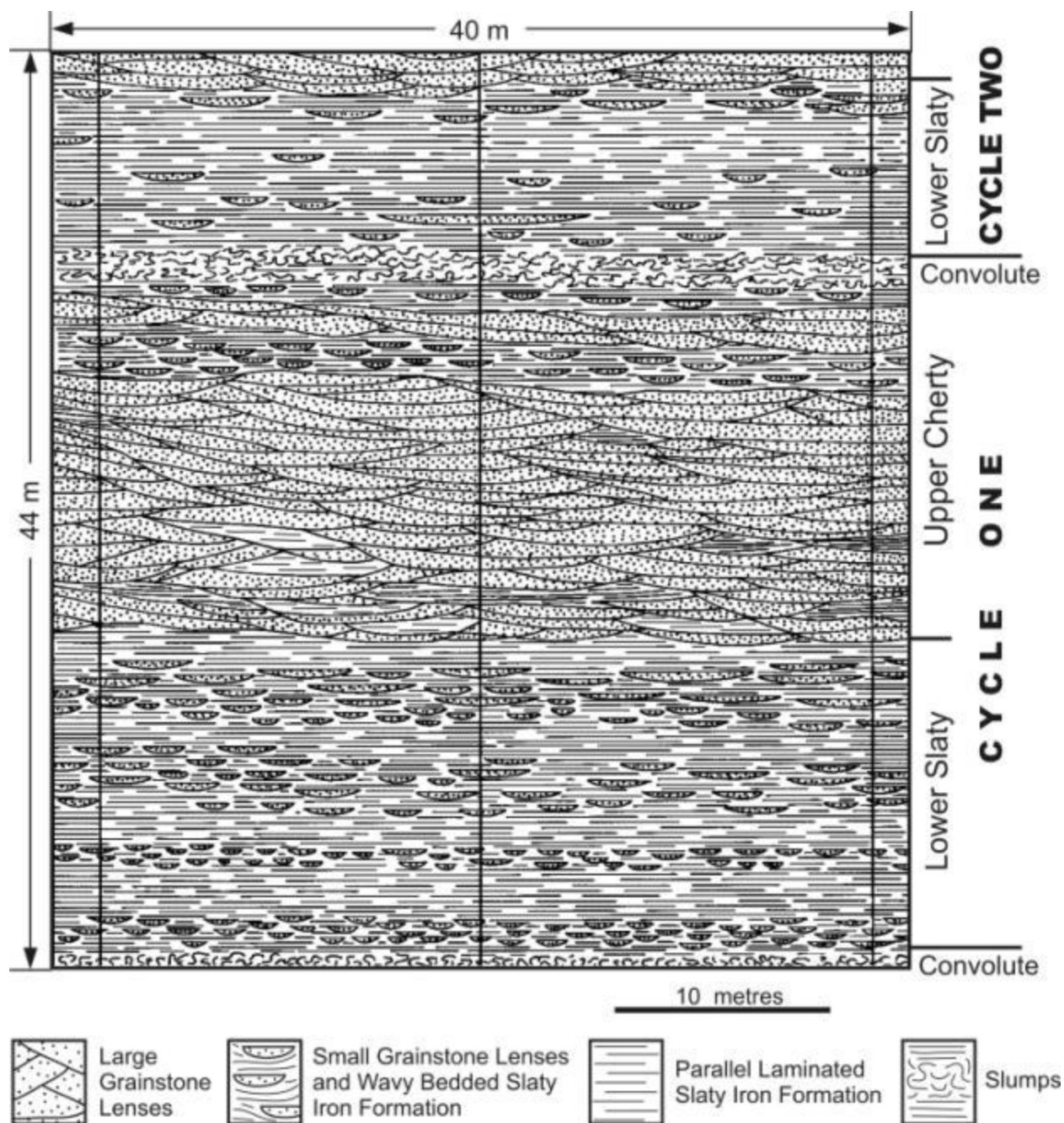


Figure 5. Schematic map of vertically dipping asymmetric cycles developed in Superior-type iron-formation of the study area. The slaty unit forming the lower portion of cycle 1 overlies a slumped unit of convolutedly-bedded chemical mudstones and is gradational into a succession of cherty grainstone lenses. These coarse-grained beds are sharply overlain by another slumped chemical mudstone unit. The base of a second coarsening upward cycle (cycle 2) forms the upper portion of the outcrop.

Grainstone lenses become progressively thicker upwards in cycles with the largest at cycle tops, where they are sharply overlain by a unit of slumped chemical mudstone. Cycles formed through progradation when offshore-directed storm currents transported progressively greater quantities of chert sand intraclasts that were formed in nearshore settings, outboard into middle and distal shelf environments (Fig. 5). Abrupt subsidence events, probably resulting from normal faulting associated with extensional tectonism, repeatedly terminated shallow-water chert grainstone accumulation and may have also generated the slumped units at cycle boundaries. The episodic storm currents are also interpreted to have transported biologically oxygenated waters over the shallow-water, inner shelf outboard along the seafloor into otherwise anoxic bottom waters of the strongly stratified distal shelf (Fig. 6). The consequence of such transport and mixing was rapid deposition of chemical mud, mainly as precipitated Fe-oxide. In many cases, the resultant decrease of Fe^{2+} in the water column, together with pelagic inorganic precipitation of SiO_2 and rainout of terrigenous clays, resulted in submillimeter- to millimetre-thick, chemically graded laminae. The concomitant decreasing $\text{Fe}^{2+}/\text{Mn}^{2+}$ ratio also led to increasing Mn-compound precipitation and enrichment in the upper portions of some chemically graded laminae (Fig. 7).

The outcrop is mostly two-dimensional and glacially polished so that sedimentary features are well displayed. It is sub-parallel to the trend of the paleoshoreline but oblique enough so that paleocurrent trends can be ascertained.

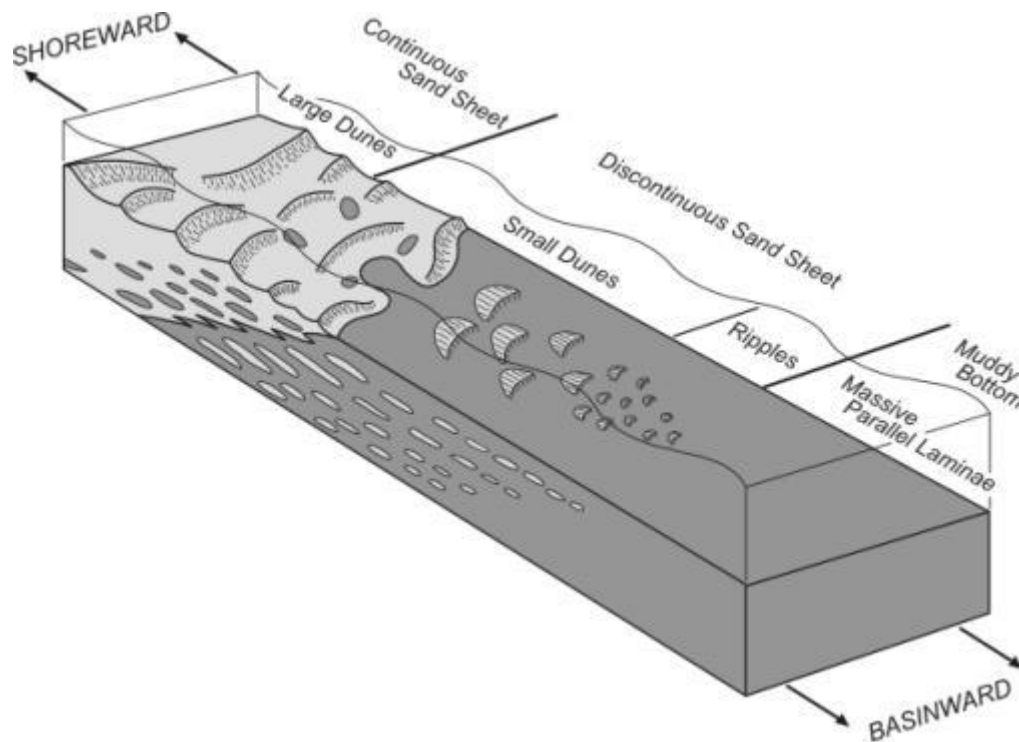


Figure 6. Physical depositional model for an asymmetric iron-formation cycle. Rapid transgression was followed by formation of a shoaling upwards, prograding cycle. Offshore flow during or immediately after major storm events built shelf sand sheets from chert and Fe-oxide intraclasts which progressively migrated into more distal environments as water depth decreased.

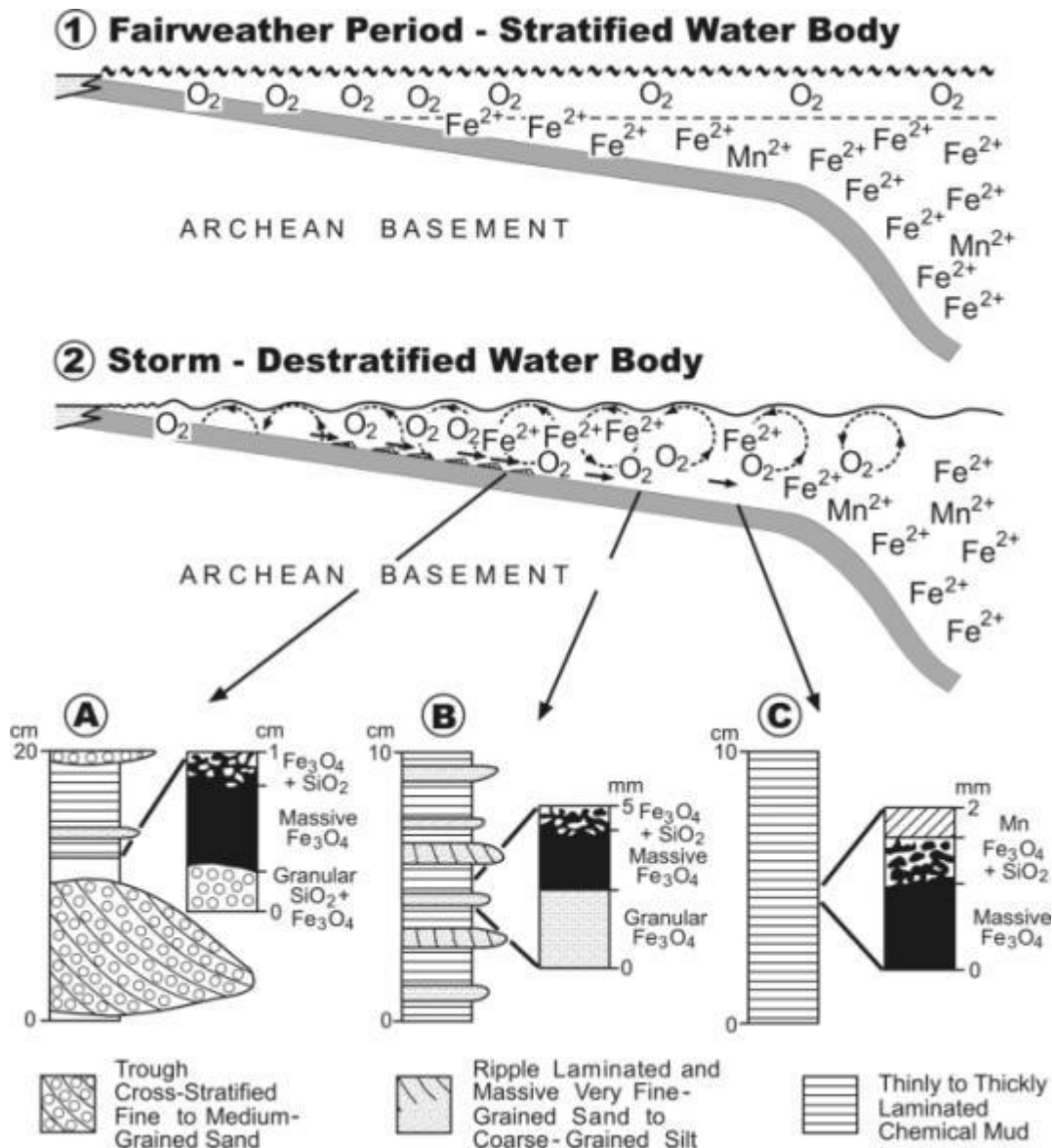


Figure 7. (1) Chemical depositional model for an asymmetric iron-formation cycle. Low free oxygen levels allowed Fe and Mn to remain dissolved in a water mass with near-neutral pH. The main O_2 production took place near-shore where light penetration to the seabed allowed cyanobacteria to flourish. (2) Storm-generated offshore flows, which were responsible for generating hummocky cross-stratification in more shore-proximal locations, transported the oxygenated coastal waters to the mid and outer shelf where they mixed with Fe- and Mn-bearing offshore waters, with resultant chemical precipitation. Storm-wave mixing of the stratified water body may have also played a role in this process, although an O_2 -bearing surface layer is not necessary. The succession represented by Column A reflects the formation of thick Fe-oxide laminae overlying grainstone lenses. The succession illustrated in Column B would have developed in a more distal shelf position resulting in thinner and finer grained sand lenses and thinner chemical layers than in Column A. The assemblage which developed furthest offshore is shown in Column C.

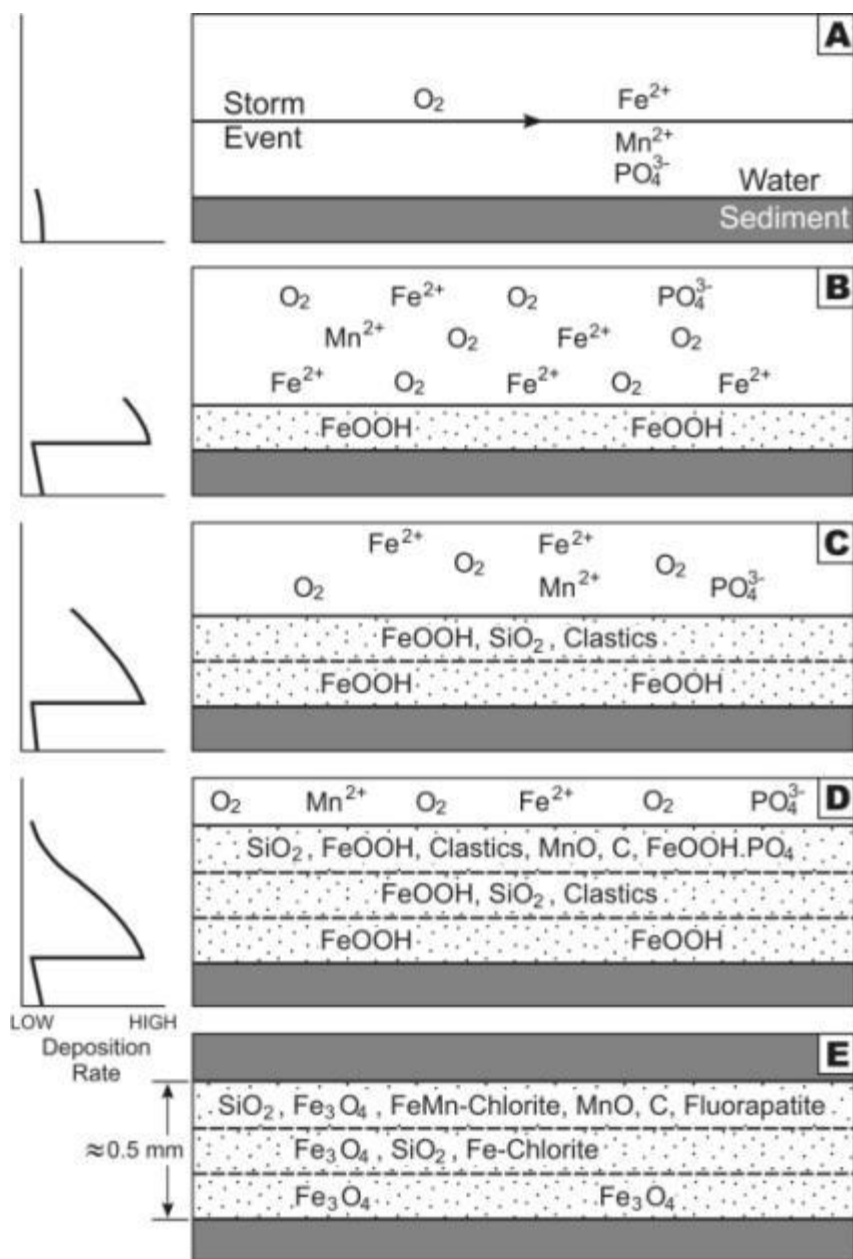


Figure 8. Progressive formation of a millimetre-scale chemical mud laminae. Chemical grading is interpreted to be the result of changes in water chemistry driven by the precipitation process. (A) Offshore currents generated during storm events delivered near-shore oxygenated water to deeper portions of the shelf, where it mixed with normal (Fe²⁺-, Mn²⁺- and PO₄³⁻-bearing) ocean water. (B) Precipitation proceeded relatively rapidly as the abundant Fe²⁺ and O₂ combined to probably form FeOOH's. (C) As Fe content decreased in seawater precipitation rate slowed and ambient clastic and SiO₂ deposition gained in importance. (D) As the Fe/Mn ratio in the aqueous phase decreases, Mn²⁺ more effectively combined with oxygen. A progressively decreasing Fe²⁺/PO₄³⁻ ratio results in more scavenging of PO₄³⁻ by the FeOOH's. Carbon is delivered to the sediment by the rainout of phytoplankton from the surface ocean. (E) Diagenesis and low-grade metamorphism resulted in the mineral assemblage present today.

Stop 9A: Palms Formation and Ironwood Iron-formation along Ballou Creek, east end of Mt. Whittlesey

(46.31083, -90.57756) *The exposures are in a long road cut along the west side of Lake Road.*

Recent road improvement along Lake Road has produced a set of roadcuts along the west side the valley of Ballou Creek that provide a cross section of most of the Palms Formation and part of the lower Ironwood Iron-formation. Starting at the south, at the lower part of the section, a nearly continuous exposure, about 75 m long, of laminated argillite and sandstone illustrates the lower part of the Palms Formation. Proceeding north, up-section, along the outcrop many typical sedimentary structures of the Palms are well exposed. These features have been interpreted to indicate sedimentation on a tidal flat or tidally influenced lagoon. Farther north, a covered interval about 75 m wide is probably underlain by the upper part of the laminated lower Palms Formation and quartzite of the upper Palms Formation. Next, to the north, is a 30 m exposure of the upper part of the Palms Formation quartzite and the Ironwood Iron-formation. The uppermost quartzite is coarse-grained and cross-bedded. The transition to the Ironwood Iron-formation takes place in a 0.5 m interval containing reworked fragments of the Palms surrounded by magnetite. The overlying Ironwood Iron-formation varies from even-bedded magnetite-rich material to irregularly bedded jasper at a scale of a meter or less.

Stop 9B. Contact of Paleoproterozoic strata and Archean metavolcanic rocks on the southeast flank of Mt. Whittlesey

(46.30834, -90.58632) *Walk west from Lake Road about 600 meters along the base of the ridge.*

Low outcrops, near the base of a prominent ridge supported by the Palms Formation, reveal the basal contact of Paleoproterozoic strata with underlying icelandite breccia of Neoarchean age. The breccia has a penetrative structural fabric, best expressed by the prominent elongation of clasts. Long axes plunge moderately southward, but must have plunged moderately to steeply northward prior to tilting of the rocks into the monoclinial structure. The basal Paleoproterozoic unit is a chert breccia that locally is the basal unit of the Palms Formation. We have interpreted this breccia to be residual concentration of chert produced by dissolution of the Bad River Dolomite, a cherty carbonate rock that is preserved sporadically beneath the breccias. The chert fragments were reworked during the transgression of the Palms Formation over a long-lived erosion surface that separates the Bad River and Palms. The upper meter or two of the icelandite breccia is strongly sheared parallel to the contact with the chert breccia, and the Archean fabric typically is completely overprinted by this secondary fabric. Shearing is weak to absent in the chert breccia. To our knowledge, these outcrops are the only ones in the western part of the Gogebic Range where a basal detachment between the Paleoproterozoic strata and Archean basement rocks can be directly observed.

We believe the Paleoproterozoic strata were thrust northward along a nearly flatlying basal contact during the Penokean orogeny. Because of later tilting of the Montreal River monocline, the present apparent displacement reflects a down-to-the-north normal fault. This fault can be traced westward, and about 1 km to the west of these exposures it passes up-

section so that the Bad River-Palms-Ironwood sequence of the upper plate is juxtaposed over Ironwood Iron-formation in the lower plate, producing the unusually wide outcrop belt of Ironwood along higher parts of Mt. Whittlesey. Just east of this stop, the basal thrust appears to pass upward into the Paleoproterozoic strata. Exposures just east of Ballou Creek show that the base of the Paleoproterozoic rocks lies unconformably on Archean volcanic rocks without an intervening shear zone. The Archean rocks seen at Stop 9B are the structurally lowest parts of the monocline that are seen on the field trip. But the trace of the Atkins Lake-Marerinsco fault, the master thrust that formed the Montreal River Monocline, lies about 8 km farther to the south. Thus, all of the stratigraphic section to the north that is seen on the trip, plus an additional 8 km of Archean rocks to the south, form the upper plate of the Atkins Lake-Marenisco fault, a coherent structural unit that was uplifted and tilted late in the Mesoproterozoic history of the region.

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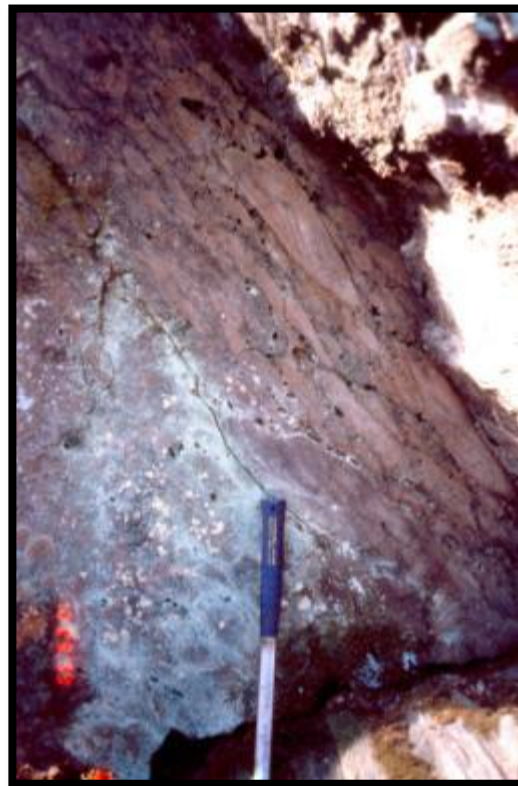
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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 8

THE ARCHEAN/PALEOPROTEROZOIC UNCONFORMITY NEAR DENHAM, MINNESOTA



Pillow basalt flow near Denham, Minnesota
- T. Boerboom

Field Trip 8

Transect from Archean basement to the Animikie basin, East-central Minnesota

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INTRODUCTION

This field trip runs parallel to the western margin of the Mesoproterozoic Midcontinent Rift System, where the continental crust is dominated by the geon 18 (1,800 – 1,900 Ma) Penokean Orogen and is intruded by rocks of Yavapai age (1,800 – 1,772 Ma; geon 17). Together, these orogenic components form a major curvilinear orogenic belt that spans an area in Minnesota from the Mesabi Iron Range to central Minnesota, with outlying intrusions as far south as the Iowa border (Figure 1).

The first part of the trip will start on Archean basement rocks (the McGrath Gneiss dome) which are non-conformably overlain by the Paleoproterozoic Denham Formation, which in turn is overlain by the eastern extension of the overlying Little Falls Formation. The Denham Formation consists of basal coarse clastics interlayered with pillow basalts that were erupted into a shallow-water environment, overlain by a thick unit of dolomitic marble, in turn overlain by a thick sequence of turbidites correlative with the Little Falls Formation. The Denham Formation and the lower part of the Little Falls Formation have been metamorphosed to staurolite-zone amphibolite facies.

The remainder of the trip will examine outcrops of turbiditic sediments of progressively lower metamorphic grade, ending in the low-grade Thomson Formation near Carlton, Minnesota. The stops will examine the structural attributes of the rocks and the gradual decrease in metamorphic grade from south to north, highlighting the difficulty of defining a boundary between the Animikie basin to the north and the older rocks of the Penokean orogen to the south. One of the stops will be in an area traditionally mapped as part of the older rocks south of/underneath the Thomson Formation, but which are now considered to be part of the Thomson Formation. We will also discuss the different generations of deformation evident in the rocks, and the likelihood that the later deformation may be related to the geon 17 (1,700 – 1,800 Ma) Yavapai Orogeny.

The reader is greatly encouraged to make use of the numerous references listed, which provide detailed studies that tie together the various components of the Penokean Orogen as well as the effects of the Yavapai and Mazatzal orogens in Minnesota, Wisconsin, Michigan, and Ontario. A special volume of Precambrian Research (Volume 157, August 2007) contains the latest interpretations of the Paleoproterozoic history of the Lake Superior region.

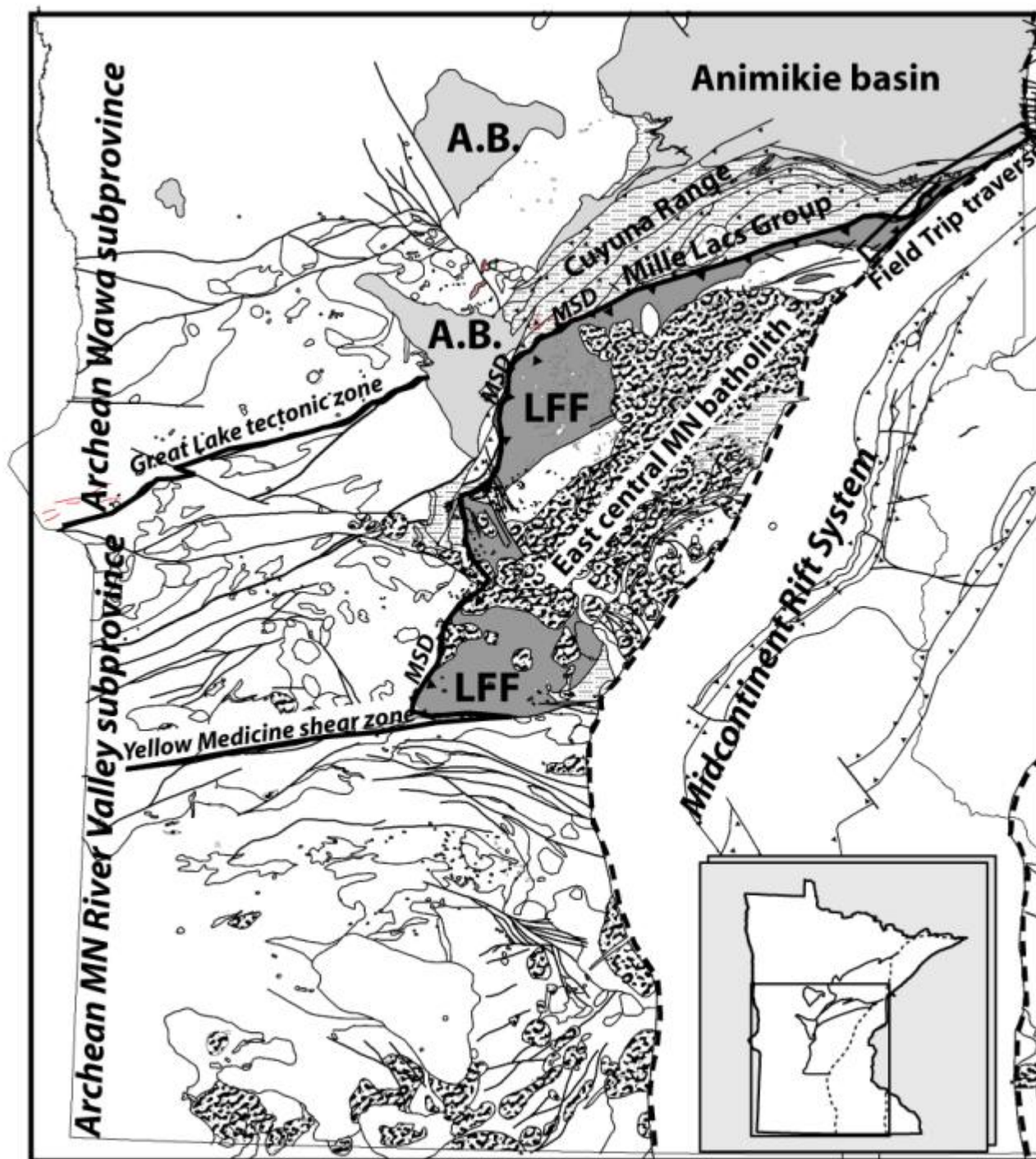


Figure 1. Simplified geologic map of southern two-thirds of Minnesota showing extent of east-central Minnesota batholith and other intrusions of known and inferred Yavapai age in the Minnesota

River Valley subprovince; also extent of Yavapai-interval intrusive rocks in central Minnesota (stippled pattern).

TECTONIC ELEMENTS OF EAST-CENTRAL MINNESOTA

Early workers (e.g. Marsden, 1972, Morey and others, 1981) correlated components of the Mesabi Iron Range (Pokegama, Biwabik, and Virginia Formations) with those of the Cuyuna North and South Ranges (Mahnomen, Trommald, Rabbit Lake Formations), and considered the Thomson Formation (equivalent to the Virginia Formation) to include the higher metamorphic grades south of Thomson Dam, e.g. stops 8-4 and 8-5 of this trip. They considered the Denham Formation to be part of an older sequence termed the Mille Lacs Group, which also included such units as the Trout Lake dolomite north of the Cuyuna Range, the Little Falls Formation, the Glen Township Formation, and other mafic volcanic and high-grade metasedimentary units.

Subsequently Southwick and others (1988) reinterpreted the entire Penokean Orogen in terms of a plate-tectonic model, with the aid of high-resolution aeromagnetic data that was unavailable to earlier workers. They considered the Cuyuna Range and everything to the south (Figure 1) to be part of an older series of folded and thrust-faulted panels that increased in metamorphic grade (hence greater degrees of exhumation), from north to south, culminating in the southern-most plutonic terrane near St. Cloud. In this model, northward-directed thrust-stacking created uplift along the southern margin of the Animikie basin, and the basin developed to the north of the uplifted crust in response to flexural loading. All of the components were considered to be of Penokean age.

The next step in refining our understanding of the Penokean Orogen in Minnesota was and is the acquisition of age dates. The first significant studies began in the 1990's, through multiple geochronological studies initiated by Daniel Holm at Kent State University, and later by others. Ar-Ar ages define widespread ca. 1,760 Ma metamorphism throughout the Penokean Orogen (e.g. Holm and others, 1993; Holm and Lux, 1996; Holm and others, 1998). Extensive dating of various components of the east-central Minnesota batholith (Holm and others, 2005) show that nearly all of the intrusions in the heart of the 'Penokean Orogen' internal zone are actually between 1,772 – 1,800 Ma (Yavapai), and only the Bradbury Creek granodiorite (1,858 – 1,877 Ma) is Penokean in age. More recently, *in-situ* U-Pb geochronological studies of hydrothermal xenotime and metamorphic monazite have further documented metamorphic and hydrothermal events related to both the Penokean and Yavapai orogens (e.g. Schneider and others, 2004; Vallini and others, 2007; Holm and others, 2007). Detrital zircon studies (e.g. Wirth and others, 2006) have provided ages for source materials for many of the units. Ongoing U-Pb zircon dating of thin ash layers and interbedded volcanic rocks within the lower parts of the Animikie basin and the Baraga basin in Michigan (e.g. Schneider and others, 2002; Fralick and others, 2002; Addison and others, 2005) and the important recognition of the Sudbury ejecta layer (e.g. Addison and others, 2005; Jirsa and others, 2008; Cannon and others, 2010) has allowed for confident correlation of the major depositional basins in the region and allowed for a more thorough interpretation of the Paleoproterozoic evolution of the Lake Superior region.

For example, Schulz and Cannon (2007), using newly available dates for igneous and metamorphic events in combination with the Penokean-wide Sudbury impact marker layer, have compiled a robust interpretation of the geologic history of the Penokean and Yavapai

events in the Lake Superior region. In their synopsis, the Penokean orogeny began ca. 1,880 Ma via docking of the Pembine Wausau terrane to the southern margin of the Superior craton, resulting in subduction jumping to the south and the development of back-arc basins in both the Superior craton edge and the arc. Extension and subsidence of the back arc basins produced volcanic grabens within a broad, shallow sea into which the major iron formations (i.e. Mesabi-Gunflint, Ironwood, and Negaunee) were formed. The iron formations at the south edge of the basin (e.g. Marquette and Gogebic ranges) are coeval with volcanic rocks whereas deposition at the north edge of the basin (e.g. Mesabi and Gunflint ranges) was distal to the volcanic activity. At around 1,850 Ma (a time for which the Sudbury impact layer provides a basin-wide marker horizon; Figure 2), a fragment of Archean crust (Marshfield terrane in Wisconsin) arrived at the subduction zone and converged with the Superior craton, causing subsidence in front of the collision zone, and forming a broad foreland basin. Some parts of the basin saw uplift and possible subaerial exposure of the iron formation for a time, prior to inundation and burial by a transgressive sediment sequence (Animikie and Baraga basins). Sedimentation into the southern part of the Animikie foreland basin began at 1,850 Ma and ended by 1,833 Ma, whereas in the northern part of the basin sedimentation did not start until ca. 1,835 Ma and continued until at least 1,770 Ma. Tectonic thickening produced medium-grade metamorphism of the sediments, which continued at the south edge of the basin until ca. 1,830 Ma when the deformed sediments and accreted arc terranes were intruded by post-tectonic granitic rocks (Van Schmus, 1976), marking the end of Penokean orogenesis.

Younger tectonic and thermal events that overprint the Penokean orogen, previously ascribed to the Penokean orogeny, are now known from metamorphic and thermochronologic studies to be Yavapai-interval events that post-date the Penokean by tens of millions of years (e.g. Holm and others, 2007; Tohver and others, 2007). Renewed Yavapai-interval activity that peaked in mid-geon 17 produced doming of the Archean crust (e.g. McGrath Gneiss dome in Minnesota and other gneiss domes in Michigan – the ‘gneiss dome corridor’), accompanied by mid-crustal level emplacement of the east-central Minnesota batholith, and widespread resetting of metamorphic ages in the rocks of east-central Minnesota. Detrital zircons as young as 1,770 Ma (Heaman and Easton, 2005) in the northern reaches of the Animikie basin (Rove Formation in Ontario) show that deposition continued into the Yavapai-interval accretion event, at least in the upper portions of the basin. Deformation at the southern margin of the Animikie basin may have begun in the late stages of Penokean orogenesis, but likely continued during geon 17 (Scheiner, Boerboom, and Holm, 2011).

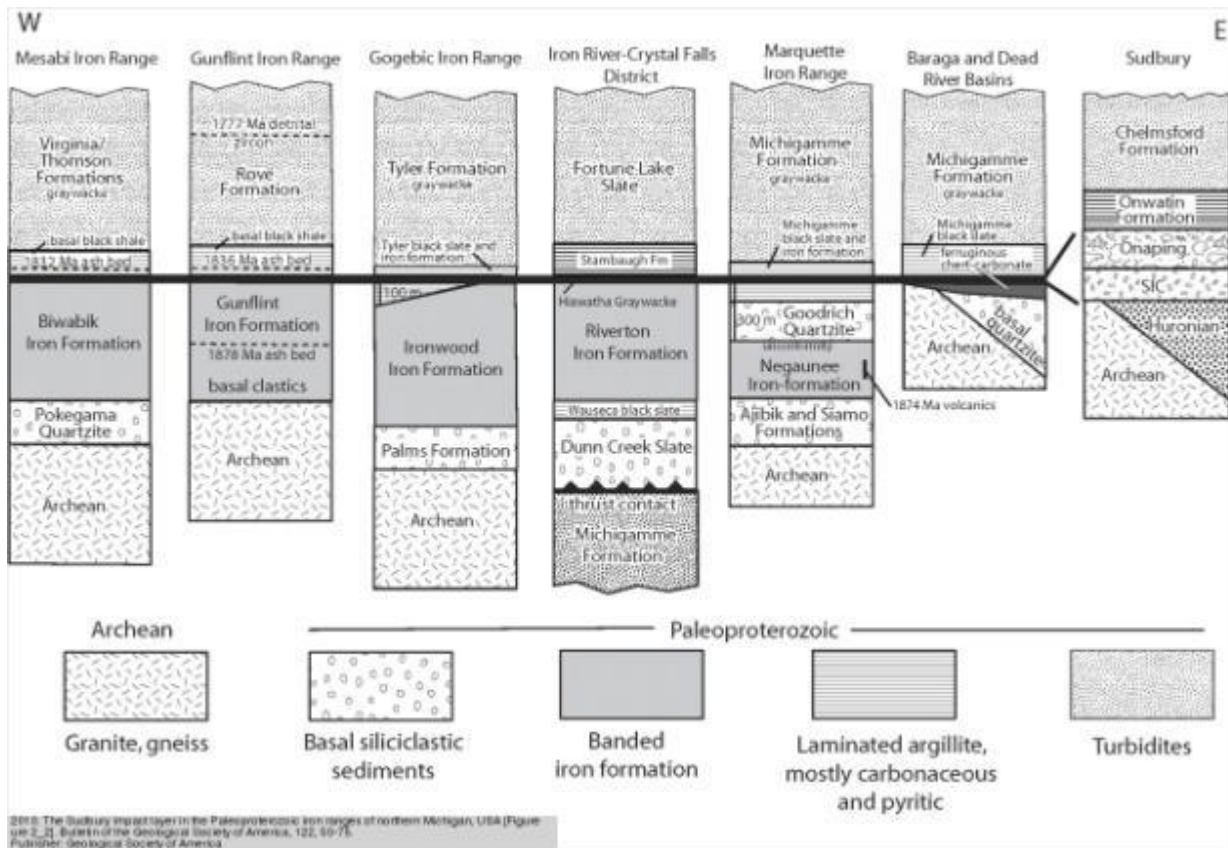


Figure 2. Correlation diagram of Paleoproterozoic strata across Minnesota, Wisconsin, and Michigan, showing the 1,850 Ma Sudbury impact layer timeline. From Cannon and others, 2010.

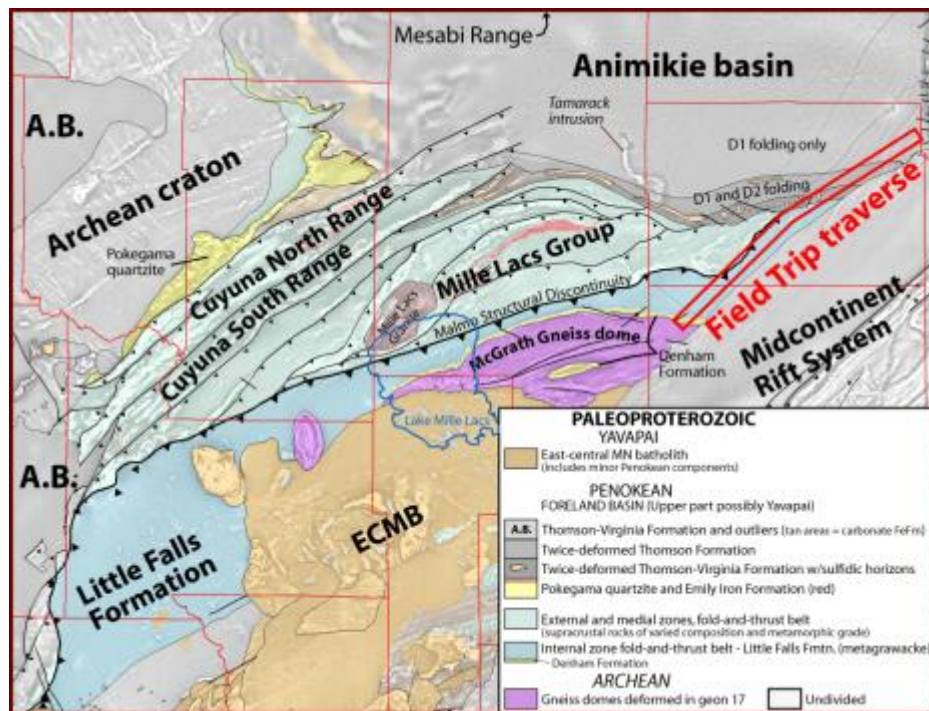


Figure 3. General geologic map of east-central Minnesota showing the major subdivisions of the Penokean Orogen. Red box outlines field trip traverse.

LITHOLOGIC UNITS

Animikie Basin Foredeep

The Animikie basin is conceptualized as a foreland basin that developed in front (north) of the fold and thrust belt (everything south of the Animikie basin in Figures 1 and 3). It includes the main depositional basin as well as two outliers that probably represent erosional remnants of a larger, formerly continuous area. Strata of the Animikie basin unconformably overlie both Archean crust and deformed rocks of the Penokean Orogen fold and thrust belt (e.g. the Cuyuna North Range). The strata at the north edge of the Animikie basin are essentially undeformed, with a gentle south dip, whereas at the southern margin they are folded into a series of generally upright and open, east-west trending folds (stop 8-9). Slaty cleavage associated with folding first appears about 15 km south of the Mesabi Iron Range, and increases in strength southward in conjunction with progressively tighter folds (Southwick and others, 1998). A narrow belt along the southern edge of the basin contains an early bedding-parallel metamorphic foliation that is folded into upright folds which are apparently coaxial with folds related to the single deformation to the north.

The Animikie Basin is composed of a thin lower unit of quartzite and siltstone (e.g. the Pokegama quartzite on the Mesabi Iron Range; Figures 1, 2, 3, 5) overlain by the Biwabik and Gunflint Iron Formations, and a thick upper sequence of turbidites and shale (the Virginia-Thomson Formations and related outliers, and the Rove Formation). The Pokegama quartzite onlaps ca. 2,700 Ma Archean granite-greenstone terrane to the north, and contain only locally-derived 2.9 – 2.6 Ga detrital zircons eroded from adjacent Neoproterozoic crust (Wirth and others, 2006). In Ontario, the 1,850 Ma Sudbury ejecta layer (Addison and others, 2005) lies above an $1,878 \pm 1.3$ Ma volcanic ash layer near the top of the Gunflint Formation (Fralick and others, 2002) and below an $1,836 \pm 5$ Ma ash layer near the base of the Rove Formation (Addison and others, 2005). In addition an ash layer near the base of the Virginia Formation on the Mesabi Iron Range in Minnesota yields a U-Pb zircon age of 1832 ± 3 Ma (Addison and others, 2005). The nearly 40 m.y. hiatus at the north side of the basin is ca. 15 m.y. greater in duration than the same hiatus associated with the same unconformity at the south side of the basin in Michigan, which was more proximal to the Penokean tectonic front. (Shulz and Cannon, 2009). The position of the Mille Lacs Group and Cuyuna Range strata in Minnesota relative to deposition of the Animikie basin fill is still unresolved, but geophysical evidence implies that the Cuyuna North Range Group continues beneath, and is unconformably overlain by, strata of the Animikie Group (Figures 3 and 5).

Detrital zircon studies to date from turbiditic sediments at the southern edge of the Animikie basin in the Thomson Formation (e.g. stop 8-9 of this trip) yield ages mostly in the 2.05 – 1.80 Ga range (Wirth and others, 2006). They report similar ages for detrital zircons from the Rove Formation in northeastern Minnesota and from the Tyler Formation in northern Wisconsin. Heaman and Easton (2005), obtained detrital zircon ages as young as 1,777 Ma from near the top of the Rove Formation in Ontario, approximately 400 m above the base (Maric and Fralick, 2005). This age overlaps with the age of the east-central Minnesota batholith in Minnesota (Holm and others, 2005), and indicates that sedimentation, at least in the uppermost strata of the Rove portion of the Animikie basin, was synchronous with Yavapai magmatic activity to the south. Using the interpretation that the Penokean orogeny ended at ca. 1,830 (Schulz and Cannon, 2009), these data imply that much of the deposition into the basin may have occurred during Yavapai time.

Thomson Formation

The Thomson Formation is correlative with the Virginia Formation, but is situated at the southeastern margin of the Animikie Basin (Figures 1, 3, and 5). As reported by Morey (1978), the type locality of the Thomson Formation (stop 8-9) is composed of 34-46 percent graywacke, 27-39% siltstone, and 27% slate. The Thomson Formation has been metamorphosed to the greenschist facies, and is cut by a series of northeast-trending diabase dikes that are presumed to be associated with the Midcontinent rift system.

The bulk of the Thomson Formation is deformed into a series of open, east-trending folds with axes that dip vertical to steeply south and plunge shallowly to the east and west. However, Holst (1984) recognized a 'southern structural terrane' which contains an early, near bedding-parallel, S_1 foliation that is inferred to be parallel to poorly exposed recumbent isoclinal folds, and the entire area is proposed to be on the upper limb of a large recumbent fold. This early fabric is refolded along open and upright folds having axial traces that plunge gently both east and west, which has crenulated the S_1 cleavage and produced an S_2 cleavage. The second generation of folds has the same style and orientation as those to the north, where only one generation of folding (regional D_2) has produced a single slaty cleavage. The boundary between the once- and twice-deformed rocks has been interpreted by Holst (1984) as the edge of a thrust nappe, south of which the rocks contain two deformations and north of which the rocks were only affected by the second deformation. However, recent mapping (e.g. Boerboom, 2009) across this transition shows that there seems to be a progressive decrease in metamorphic grade from south to north, and the transition between twice-deformed and once-deformed rocks (e.g. Figure 3) is not easily demarcated. The twice-deformed 'southern structural terrane' includes rocks interpreted to be part of the Penokean fold and thrust belt as well as the southern edge of the Thomson Formation/Animikie basin, without a clear-cut structural front to separate the two entities.

An alternate explanation, based on magnetic and petrofabric studies, proposes that the early subhorizontal S_1 cleavage in the southern zone formed at the same time as the single subvertical cleavage in the northern zone (Sun and others, 1995 and references therein).

Their paleomagnetic studies showed that the natural remnant magnetism of the Thomson Formation record the field during deposition and compaction, and was not reset during deformation. This field is recorded by ferromagnetic minerals (e.g. hematite). Anisotropy of magnetic susceptibility, recorded primarily by the preferred orientation of metamorphic chlorite, records minima perpendicular to the single upright cleavage in the northern zone (i.e. subhorizontal), whereas it is normal to the early S_1 cleavage in the southern zone (i.e. subvertical: Figure 4).

They invoke a model, based on studies of other foreland orogens, involving simple shear on subhorizontal planes (south) transitioning into pure shear with subhorizontal shortening (north). Late-stage regional shortening produced the second deformation, which shows up preferentially in the southern zone due to favorable orientation there of the S_1 cleavage perpendicular to compression, whereas in the northern zone the later deformation is not apparent because the preexisting fabric was already oriented subparallel to the later compressional event.

This model may be a means to explain the transitional nature observed between the northern and southern (once- and twice-deformed) panels of the Thomson Formation (e.g. between stops 8-8 and 8-9 of this trip), where the strength of S_2 cleavage in the southern zone seems to progressively weaken to the north. However, the samples for the paleomagnetic

study were confined to the immediate area of the transition, and do not include the higher-grade metamorphic facies to the south. The promising use of future paleomagnetic studies should incorporate samples from the higher-grade terrane to the south.

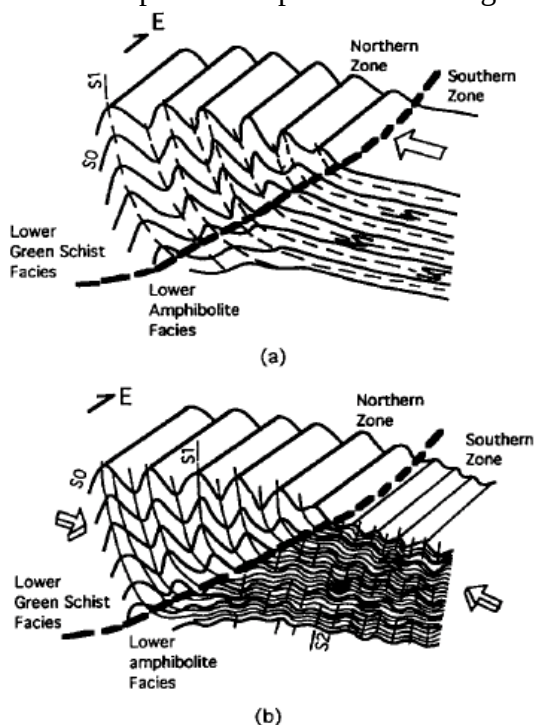


Fig. 17. A possible explanation of the deformational history of the Thomson formation. (a) N-S shortening and N-directed (upper plate) shear of the Thomson Formation during the Penokean orogeny. Cleavage subparallel to bedding was formed in the southern zone where deformation and translation was greatest, while upright open folds with axial plane cleavage were formed in the northern zone. (b) Late-stage regional shortening continued to affect the Thomson Formation. Crenulation cleavage was formed only in the southern zone due to the high angle between S_1 and the direction of shortening.

Figure 4. Alternative model for explaining the contrasting deformational history of the northern vs. southern zones of the Thomson Formation. From Sun and others, 1995.

On prior geologic interpretations, the southern margin of the Thomson Formation was considered to be the point at which the large area of featureless aeromagnetic signature associated with the Animikie basin ends, and where characteristically ‘busy’ signatures associated with the Mille Lacs Group begins. However, based on remapping and relogging of drill cores and cuttings, the southern margin of the Thomson Formation is now considered to include an east-west belt marked by moderately high gravity in combination with an east-west belt of discontinuous linear aeromagnetic anomalies (Figure 5). This linear zone wraps southwest into apparently tightly-folded strata between rocks of the Cuyuna South Range and the Mille Lacs Group, and the argument could be made that the entire Cuyuna South Range is the continuation of this southern-most belt now mapped as Thomson Formation. The discontinuous lenses shown as ‘sulfidic horizons’ in the base of the Thomson Formation on Figure 3 are characterized by large cubic pyrite porphyroblasts. Most of these zones are confined to the very base of the Thomson, but some lenses occur as far as 4 km north of the

base, well within classic Thomson Formation. In addition to the sulfidic horizons, this lower unit includes generally thin units of chert, and minor amounts of mafic hypabyssal intrusions (Boerboom, 2009). These sulfidic horizons may be similar to those in the base of the Baraga basin in Michigan, e.g. the Bijiki Iron Formation.

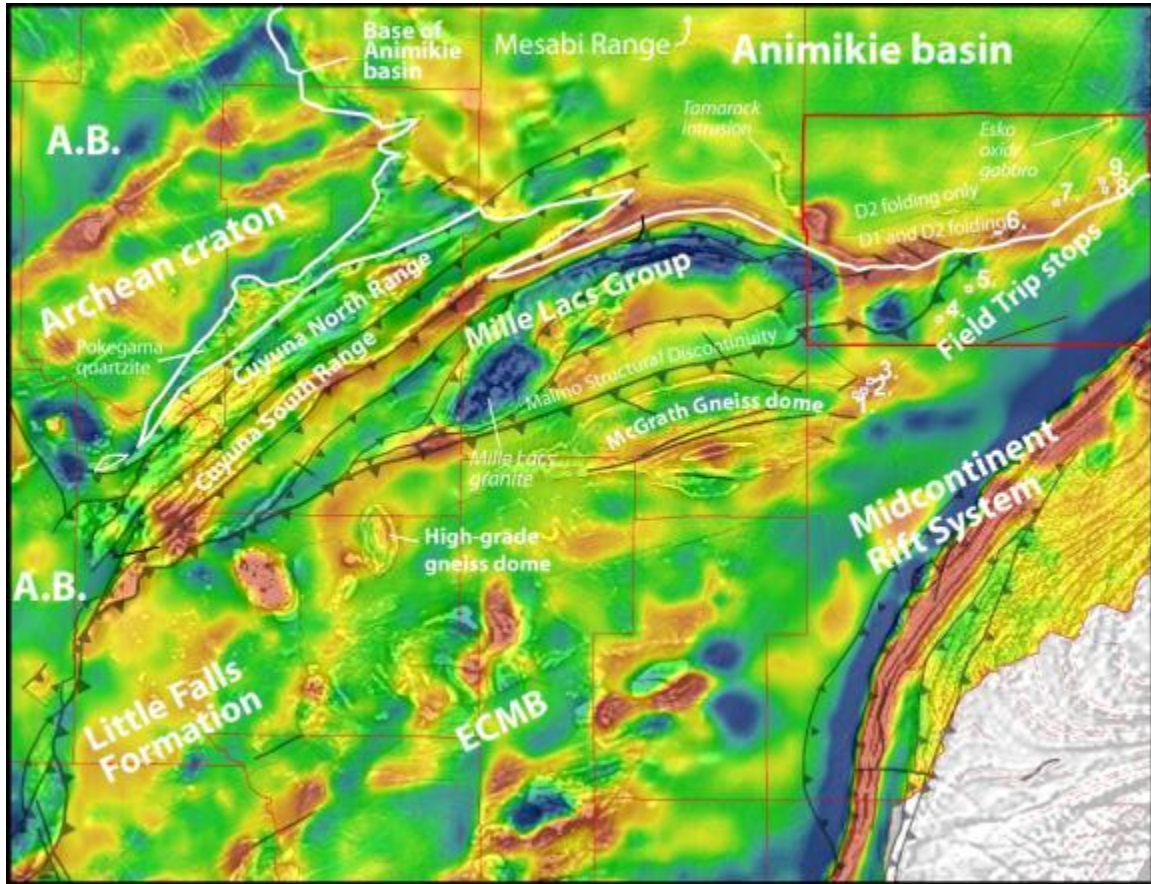


Figure 8-5. First-vertical derivative aeromagnetic anomaly map superimposed on second-vertical-derivative gravity. Note the moderate gravity high that along the southern margin of the Animikie basin that extends southwest into rocks currently mapped as Cuyuna South Range. Compare to Figure 3.

Fold and Thrust Belt

The fold and thrust belt (Figure 3) shows decreasing metamorphic grade and structural complexity (i.e. a decreasing depth of tectonic burial) from south to north. It contains a complex assemblage of thrust-stacked, folded and faulted metasedimentary and metavolcanic rocks and associated hypabyssal mafic sills, which overall are of low to moderate metamorphic grade except for the southern margin which have been metamorphosed to the amphibolite facies. The medial zone includes the South range of the Cuyuna district and the Mille Lacs Group and associated rocks (Figure 1, 3 and 5), which encompass a broad area that extends from the Moose Lake area south of the Animikie basin to as far southwest as Stearns County. Iron-formations in the medial zone are closely associated with mafic volcanic rocks and euxinic shale. The external zone, which forms the

northwest-most panel and includes the Cuyuna district North range, is composed of folded and weakly metamorphosed strata including the Mahnomen, Trommald, and Rabbit Lake Formations, but also includes a substantial proportion of mafic volcanic rocks. Iron-formations (mainly the Trommald) are associated with fine-grained argillaceous rocks as well as mafic volcanic rocks. Aeromagnetic data (Figure 5) and geophysical models clearly indicate that tightly folded strata of the Cuyuna North range district continue to the east beneath unconformably overlying strata of the Animikie group, a relationship that has been verified by geophysical models (Carlson, 1985). Also clearly visible on Figure 5 is the truncation of magnetic anomalies associated with the Mille Lacs Group and Cuyuna South Range by the overthrust Little Falls Formation, which is magnetically quiet.

Denham and Little Falls Formations

The Denham Formation (Morey, 1978) is composed of a heterogeneous mixture of interbedded pelitic to carbonate-cemented, arenitic arkosic sedimentary rocks, calc-alkaline pillow basalt, dolostone, and fragmental volcanic rocks. This sequence nonconformably overlies the Archean McGrath Gneiss, and has been subjected to amphibolite-facies metamorphism. Drill holes (cores and cuttings) from north and east of the Denham locality show that the Denham is transitional upward into graywacke now inferred to be the eastern extension of the Little Falls Formation (Figure 1, 3, and 5), with the transition marked by a 1m thick zone of graphitic argillite. Aeromagnetic and widely scattered outcrop data indicate that the Denham Formation extends westward along the north edge of the McGrath Gneiss to Mille Lacs Lake, and the Little Falls Formation continues west and south a considerable distance past Mille Lacs Lake. Old drilling records and corroborating aeromagnetic anomalies also indicate that infolded remnants of the Denham Formation are present within the McGrath Gneiss (Figures 3 and 5). Stops 8-1 through 8-3 of this field trip will cover the transition from the McGrath Gneiss to the base of the Little Falls Formation.

The age of the Denham Formation is constrained by the underlying McGrath Gneiss ($2,557 \pm 15$ Ma; Figure 6; Holm and others, 2005). The Denham Formation and overlying Little Falls Formation are interpreted as a panel thrust northwest over the Mille Lacs Group, which is cut by a granite stock (Mille Lacs granite; Figure 3) that yields a U-Pb zircon age of 2,009 Ma (Holm and others, 2005). Volcanic rocks in the Denham Formation have provided a Sm-Nd isochron age of 2197 ± 39 Ma (Beck, 1988). Thus, the age of the Denham Formation is poorly constrained to somewhere between 2,550 to possibly as young as 2,009 Ma, depending upon the relationship to the Mille Lacs Group. Monazite U-Th-Pb single-spot electron microprobe analyses and single-spot ion microprobe $^{207}\text{Pb}/^{206}\text{Pb}$ analyses of samples from the Little Falls Formation yield mainly 1,741 – 1,776 Ma Yavapai-interval ages, and a sample from stop 8-3 of this trip also yielded an age of $1,844 \pm 7$ Ma in addition to a $1,788 \pm 4$ age, thus this location records both Penokean and Yavapai metamorphic events (Schneider and others, 2004; Holm and others, 2007).

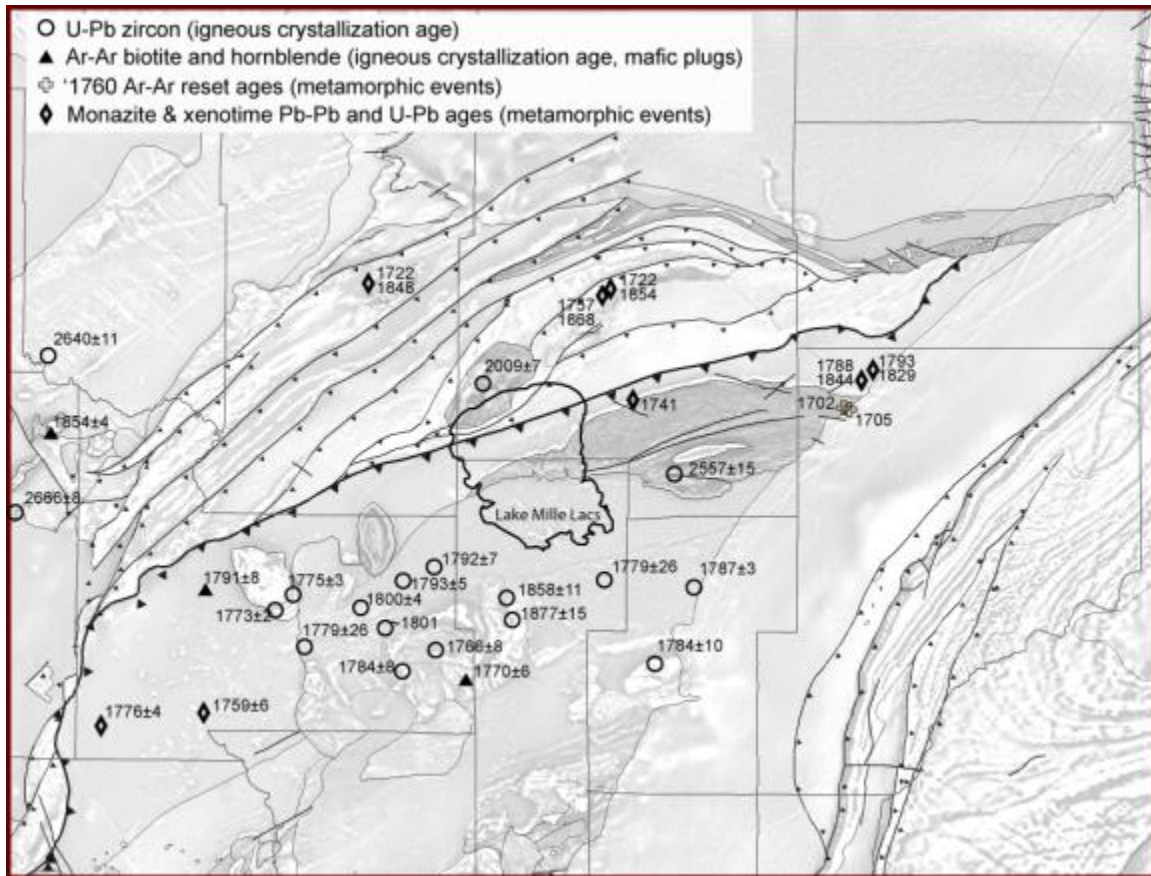


Figure 6. Locations of various types of age date determinations in east-central Minnesota. Base map is same as in figure 8-3.

Rocks of the Denham and Little Falls Formations have undergone regional amphibolite-grade metamorphism and at least two periods of deformation attributed to Penokean and Yavapai Orogenic events. The first deformation event was synchronous with metamorphism to the garnet zone of the amphibolite facies (Holm, 1986). It produced an early S_1 foliation that is typically bedding-parallel, and a locally strong, shallowly plunging, stretching lineation (Figure 7a). S_1 foliation and S_0 bedding were subsequently folded along steeply dipping axes (Figure 7b), concurrent with or followed by ca. 1,760 Ma peak metamorphism that produced staurolite. In the Denham valley (stop 8-2), the stratigraphic sequence dips variably north, having local F_2 folds with overturned limbs. North of the valley in the overlying Little Falls Formation (stop 8-3) bedding and S_1 in metagraywacke are nearly horizontal, and are deformed into open F_2 folds with local crenulation features. Farther north the bedding dips mostly south, defining a broad, regional-scale, upright F_2 syncline.

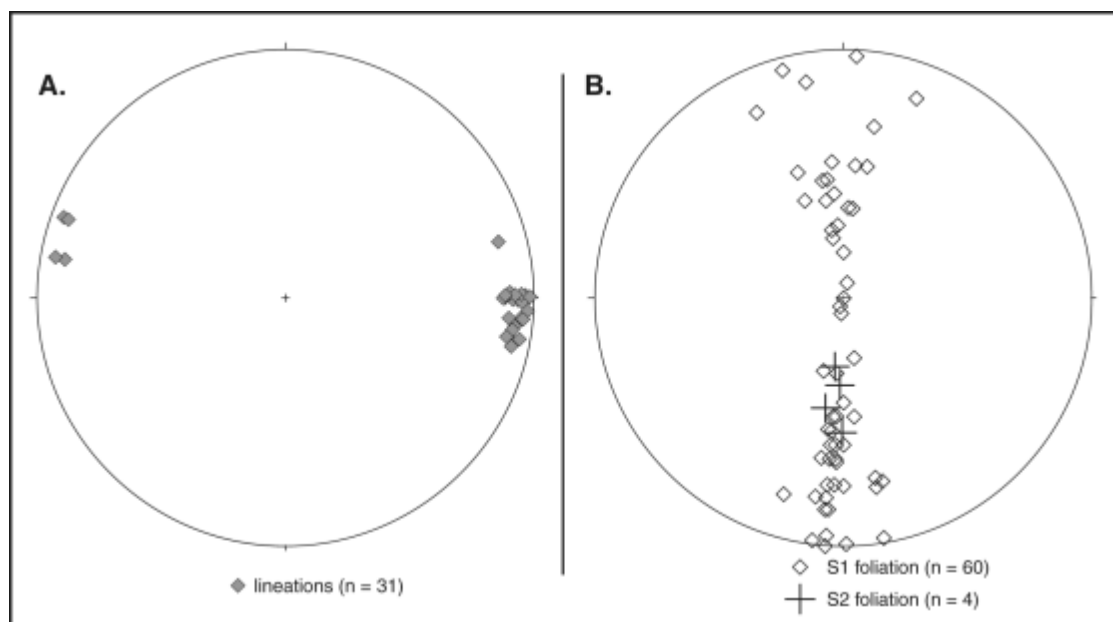


Figure 7. Equal area, lower hemisphere projections of structural elements of the Denham Formation. A. Measured lineations; B. Poles to measured foliations. Open diamonds represent the main, early foliation, which has been refolded by a second deformation that produced only a weak north-dipping foliation (crosses).

Despite deformation and metamorphism, the stratigraphy of the Denham Formation forms a coherent package that is shown schematically in Figures 8. Omitting the prefix 'meta' for clarity, the base of the Denham Formation consists of interbedded siltstone and cross-stratified pebble conglomerate. This is overlain by coarse-grained and locally conglomeratic arkose that apparently pinches out laterally. The primary clastic grains in the dolomitic arkose are well preserved owing to the abundance of dolomite in the matrix, which accommodated most of the deformation. The arkose is interbedded with amygdaloidal basalt flows that contain well-preserved primary macroscopic flow features. The flows grade stratigraphically upward (northward) from massive bases to pillowed interiors to fragmental upper portions. The volcanic flows are thickest at the eastern outcrop limit, where at least four flows of nearly 1000 feet total thickness were recognized, and thin westward to two flows of 300 feet total thickness. An upper sequence of mafic fragmental volcanic rocks nearly merges to the east with the pillow basalts (see Figure 10 under stop 8-2), possibly implying that later stages of volcanic activity were more explosive in nature and that both volcanic horizons emanated from a common center to the east, which may presently be buried beneath the Mesoproterozoic Fond du Lac Formation. Arkosic and pelitic strata (now staurolite-garnet-mica schist) between the flows and fragmental volcanic sequence pinch out to the east where the flow package thickens, and are not present in drill holes to the north and east of the Denham valley. The northern-most outcrops in the valley consist of very pure dolomite, now marble, which contains ptgmatitically folded and strongly lineated quartz veins. Drill cores show that the dolomitic marble is at least 500 feet thick, and is overlain by graywacke (e.g. stop 8-3) that is exposed discontinuously to the north for some distance. A thin layer of graphitic argillite marks the contact between the dolomite and overlying graywacke.

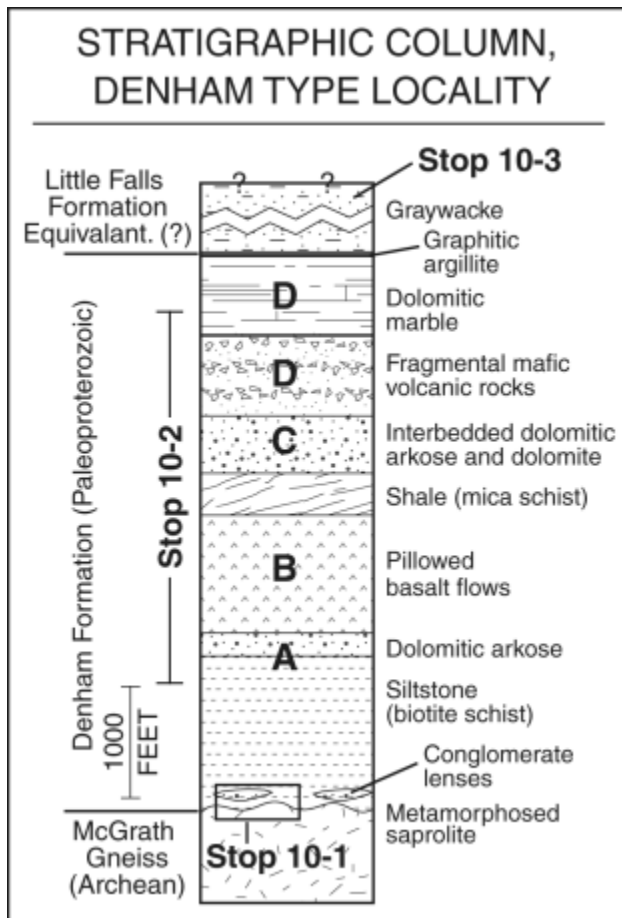


Figure 8. Schematic stratigraphic section of the Denham Formation, showing relative location of field trip stops.

Field and petrographic observations imply that clastic detritus in the Denham Formation was derived in large part from a weathering residuum on the subjacent McGrath Gneiss. Near the contact with the Denham Formation, the McGrath grades abruptly from granite gneiss containing quartz, orthoclase, plagioclase, and biotite to strongly foliated, quartz- and sericite-rich schist that contains orthoclase, but no plagioclase. The arkosic parts of the Denham Formation similarly lack plagioclase, and are composed of quartz and orthoclase grains, together with small clasts of granitic gneiss. Studies of saprolite developed beneath Cretaceous sedimentary rocks on Precambrian crystalline rocks in southwestern Minnesota may provide an analog (Setterholm and others, 1989). These studies demonstrate that plagioclase and mafic minerals such as biotite and hornblende are the first minerals to alter to clay during the weathering process, and that orthoclase and quartz are the most resistant to weathering. The basal Cretaceous strata locally consist of reworked saprolite, including beds of cross-stratified sandstone and nearly pure kaolinitic shale. Exposures of basal Cretaceous sedimentary rocks locally contain detrital orthoclase and quartz derived by slight reworking of weathered granite grus, set in a carbonate matrix. Similar processes apparently occurred in the Paleoproterozoic by erosion and reworking of weathered McGrath Gneiss into beds of calcareous arkose and kaolinitic shale. These were subsequently metamorphosed to produce recrystallized dolomitic arkose and staurolite-garnet-sericite

schist. Slight weathering of orthoclase may have liberated potassium for the inferred conversion of kaolinite to sericite during metamorphism.

The Denham Formation is interpreted to represent a rift-margin assemblage deposited during the Paleoproterozoic, similar to and perhaps temporally equivalent to the Chocoma Group in Michigan. In this setting, the McGrath Gneiss was part of the continental margin that was weathered and eroded to provide detritus to an evolving rift basin undergoing active, shallow water volcanism. Interbedded arkose and dolomite higher in the stratigraphic section represent foundering of the shelf and deepening water, possibly by subsidence of localized grabens. The paucity of clastic material in the upper, dolomite-dominated part of the sequence indicates that deposition of coarse detritus was restricted to the shallow, nearshore environment adjacent to the McGrath Gneiss. The stratigraphically upward sedimentological gradation of dolomite to graywacke (aka Little Falls Formation) indicates further deepening water and associated turbidite deposition. The deformation of the Denham Formation is inferred to be the product of basin closure during the Penokean Orogeny, later influenced by deformation related to Yavapai orogenic activity which produced doming of the underlying McGrath Gneiss.

McGrath Gneiss Dome (gneiss dome/plutonic terrane)

The Neoarchean McGrath Gneiss occurs within the internal zone of the Penokean orogen as discussed prior (Southwick and others, 1988). The McGrath Gneiss is exposed in several localities between Denham and Mille Lacs Lake to the west. Although this unit locally contains layered gneissic structure, it is best characterized as a metamorphically foliated porphyritic granite. The 2557 Ma McGrath Gneiss is only slightly younger than the undeformed 2600 Ma Sacred Heart Granite of the Minnesota River Valley in southwestern Minnesota (Doe and Delevaux, 1991). Holm and Lux (1996) interpreted the basement-cover contact of the McGrath gneiss dome to be a detachment fault because of an apparent 50 m.y. cooling age difference between basement and cover rocks. However, more recent detailed Ar ion laser data show that the basement and cover rocks have essentially the same lower temperature thermal history (Figure 9; McKenzie, 2004). Also, as discussed under stop 8-1, field and petrographic evidence (Boerboom and others, 2001) indicates that the McGrath Gneiss had a saprolite developed on it at the time of deposition of the overlying Denham Formation. Together, these data indicate that the boundary may simply be a domed nonconformable contact. Further evidence of this lies in the inlier of Denham Formation that is infolded into the top of the McGrath Gneiss dome (e.g. Figure 3). The McGrath Gneiss dome and adjacent Denham Formation represent the culmination of the north to south increase in metamorphic grade, from greenschist facies at the base of the Animikie basin to upper amphibolite facies. This southward increase in metamorphic grade is considered to reflect an increase in depth of exposure from the foreland basin at the north to the deeply eroded crust of the Penokean orogen at the south, related to geon 18 accretion. Plutonic activity in east-central Minnesota between 1,800 – 1,772 Ma, likely due to Yavapai-interval convergence (Holm and others, 2005) was more or less concurrent with formation of the McGrath Gneiss dome, and with widespread regional ca. 1,750 Ma regional metamorphism.

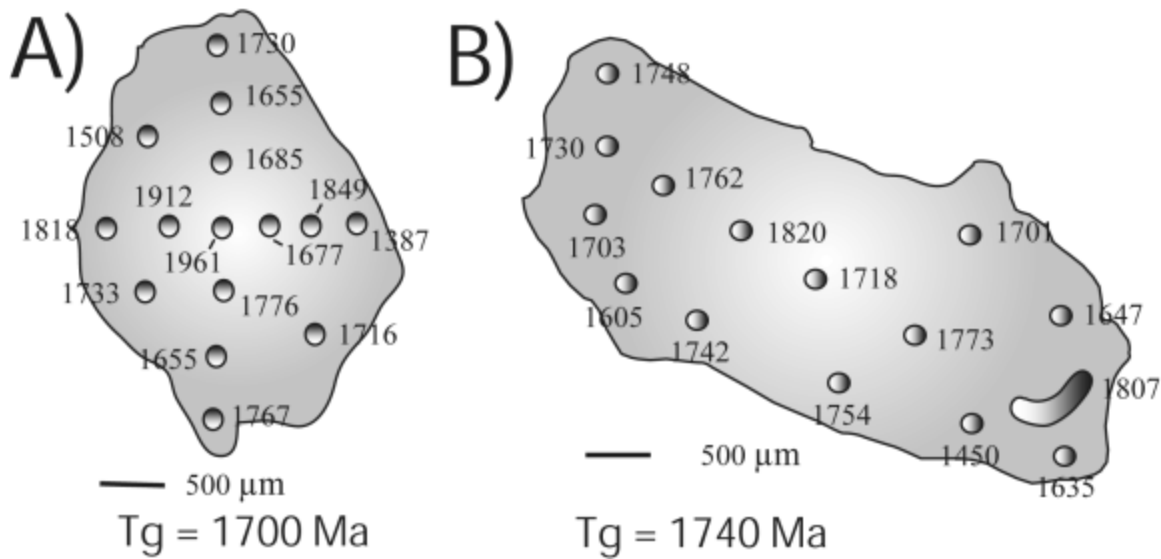


Figure 9. Spot age data (in Ma) of muscovite grains from Ar ion laser data (McKenzie, 2004). A. McGrath gneiss basement near stop 8-1; B. metapelite cover rocks at stop 8-3.

Malmo Structural Discontinuity (MSD)

The internal zone of the Penokean Orogen (the gneiss dome – plutonic terrane and the Denham-Little Falls Formations) are separated from the Penokean fold and thrust medial zone (e.g. Mille Lacs Group) by the Malmo structural discontinuity (originally termed the Malmo thrust), inferred to be a major, south-dipping fault (Wunderman and Young, 1987). The western end of the Malmo structural discontinuity (MSD: figure 1, 3, and 5) separates relatively low metamorphic grade sedimentary and volcanic rocks to the north from relatively high-grade metamorphosed granitic and sedimentary rocks and Yavapai intrusive igneous rocks to the south. The western portion of the MSD is clearly marked by the sharp truncation of linear aeromagnetic anomalies associated with thin iron-formations in the Cuyuna south range and Mille Lacs Group, indicating that the ‘internal’ zone containing the Little Falls panel was thrust west-northwest over lower-grade rocks of the medial zone. Also, recent metamorphic geochronology indicates that the ages of metamorphism differ across the MSD (geons 17 and 18 to the north and dominantly geon 17 to the south). These data all imply that the MSD is a post-Penokean geon 17 structure along which the gneiss dome – plutonic terrane (i.e. the Penokean orogen internal zone and the McGrath Gneiss dome) was uplifted and subsequently exhumed, most likely during or after emplacement of the Yavapai-interval east-central Minnesota batholith.

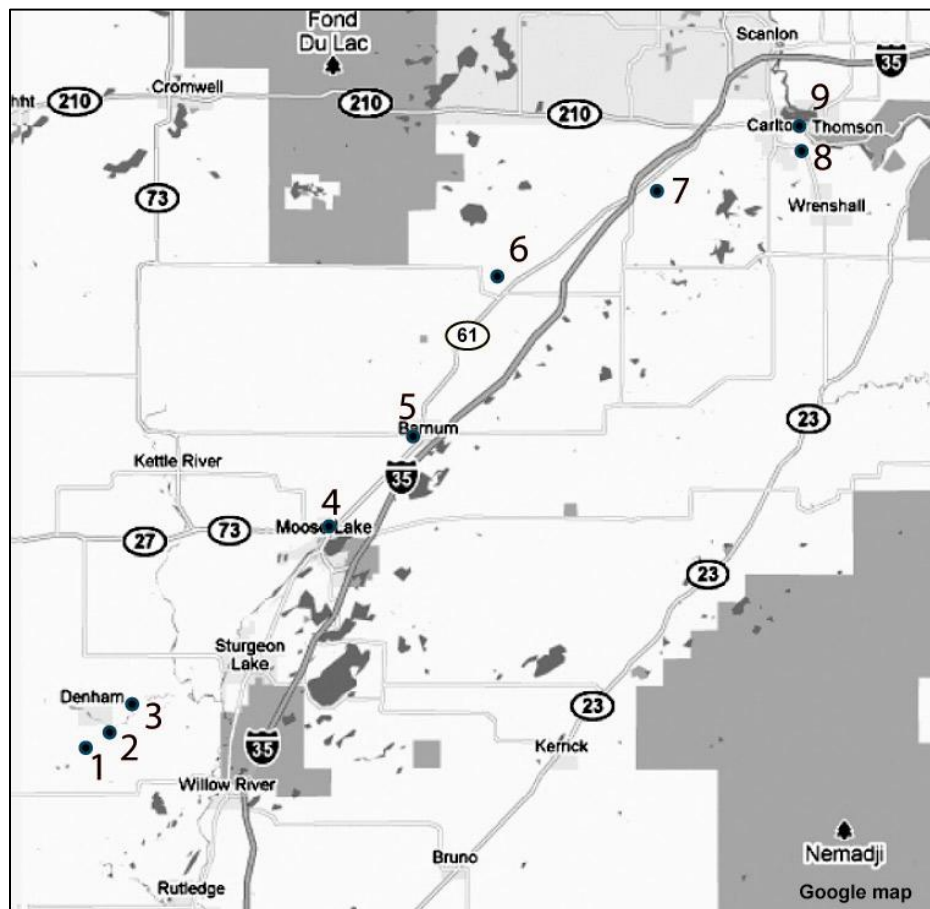
East of Mille Lacs Lake, the location of the MSD is more obscure due to the lack of contrast in geophysical data. Historically the eastern end of the MSD has been placed at the northern margin of the McGrath Gneiss. However, more recent mapping (e.g. Boerboom, 2009) indicates that the MSD may lie a short distance north of the McGrath Gneiss (Figure 3). The graywacke unit (stop 8-3), which overlies the Denham Formation (stop 8-2), contains the same metamorphic and lithological attributes as the Little Falls Formation, and if this correlation is correct the MSD must lie north of the Denham and Little Falls Formations, an observation supported by outcrop and drill core lithologies. South of where the MSD is currently mapped the rocks are composed of only metagraywacke and related rocks with no interbedded volcanic rocks. North of where the MSD is mapped, the bedrock types are a

diverse assemblage of carbonate, iron formation, graphitic schist and phyllite, and mafic intrusive and volcanic rocks. West-northwest-directed thrusting may have also caused more displacement and hence a greater degree of exhumation to the west, whereas to the east the MSD fades into more of a strike-slip regime, with less vertical displacement and hence less contrast in metamorphic grade.

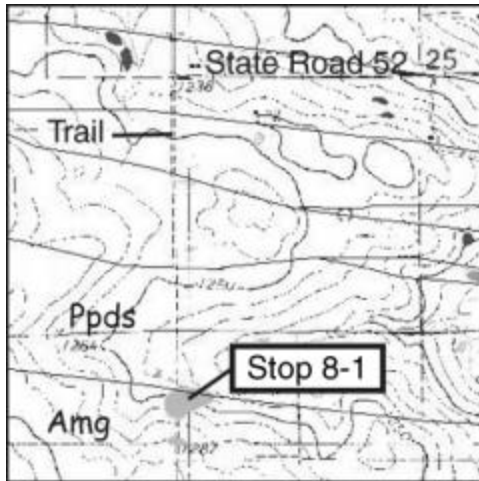
FIELD TRIP STOPS

The general geologic setting and locations of stops are shown on Figure 1. A detailed location map is included with each stop description, with a base made at scale from a portion of the appropriate 7.5' quadrangle map. The UTM coordinates are reported in NAD83, zone 15. The regional setting of the various rock units is given in the introduction, and only facts pertinent to each specific stop are given below.

--From the parking area near Carlton go west on Highway 210 through Carlton then south on I-35 to the Willow River exit (no. 205) – a distance of approximately 34 miles. Go west to Willow River, turn north on Highway 61 for 3 miles to intersection with State Road 52, go west on 52 for approximately 6.2 miles (road changes from pavement to gravel). The road makes a right angle turn to the north towards the town of Denham, but instead turn south and walk or drive on small dirt path to gate. Proceed past gate into open pasture until road crosses a gentle rise and west end of outcrop (stop 8-2a). Outcrop is 0.65 miles south of 52.



Road map showing stop locations



Stop 8-1: McGrath Gneiss and contact with overlying Denham Formation

Location: T.45N., R.21W., Sec. 36
Denham 7.5' quadrangle
UTM 503979 easting; 5132367 northing

**HIGHLIGHTS: ARCHEAN MCGRATH GNEISS
AND METAMORPHOSED SAPROLITE (AMG),
CONGLOMERATE AT BASE OF
PALEOPROTEROZOIC DENHAM FORMATION
(PPDS).**

Note: Stops 8-1, 8-2, and 8-3 are shown in apparent stratigraphic context on Figure 8. These stops will examine in order the north-facing transition from the Neoarchean McGrath Gneiss (stop 8-1), to the overlying Paleoproterozoic Denham Formation (stop 8-2), and then to graywacke of the Little Falls Formation that stratigraphically overlies the Denham Formation (stop 8-3).

****Stops 8-1 and 8-2 are on **PRIVATE PROPERTY**.

Please procure permission from landowner before entering****

At this locality, the northernmost outcrops of the McGrath Gneiss (metamorphosed porphyritic granite) grade abruptly from the typical mineral assemblage of quartz, orthoclase, plagioclase, and biotite to strongly foliated, quartz- and sericite-rich schist that contains coarse relict orthoclase, but no plagioclase. The latter assemblage, present just below the base of the Denham Formation, may be the product of metamorphism of partially weathered granite, in which the plagioclase feldspar had weathered to clay, but quartz and potassium feldspar remained fresh, analogous to the weathering patterns noted in younger saprolites that underlie Cretaceous rocks in Minnesota.

A sample of McGrath Gneiss from 20 miles southeast of this locality yielded a U-Pb zircon age of 2557 ± 15 Ma. (Holm and others, 2005). Detrital zircons from the metasedimentary rocks just above the McGrath Gneiss contact (see next paragraph) yield ages as young as $2,072 \pm 17$ Ma and as old as $3,447 \pm 17$ Ma, with spikes around 2,625, 2,900, and 3,300 Ma. Detrital zircons from stratigraphically higher dolomitic arkose (stop 8-2) yield ages as young as $2,173 \pm 11$ Ma and as old as $3,505 \pm 8$ Ma, with groups around 2,600, 2,800, and 3,360 Ma (Wirth and others, 2006).

Just north of the McGrath gneiss is a thin layer of very poorly exposed biotite schist of sedimentary protolith which contains irregular lenses of coarse conglomerate with abundant pebbles of quartz and pink orthoclase, but no plagioclase (unit Ppds on Figure 10). Similarly,

the clastic component of the arkosic beds stratigraphically higher in the Denham Formation (stop 8-2b) is comprised almost entirely of quartz and orthoclase feldspar with essentially no plagioclase feldspar. Together, these observations are consistent with the interpretation that the McGrath gneiss was weathered to the point that plagioclase and mafic minerals had altered to clay minerals, and this weathered residuum acted as a source of detritus for the overlying Denham Formation.

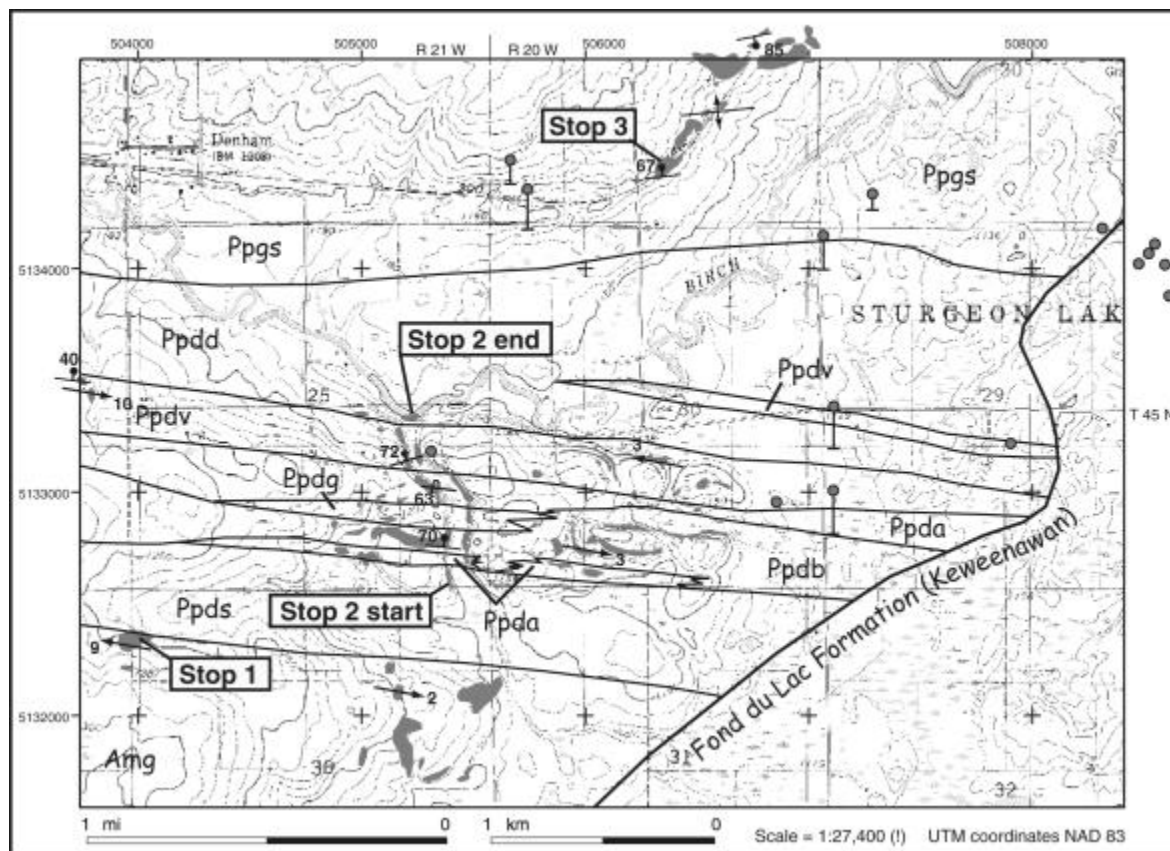
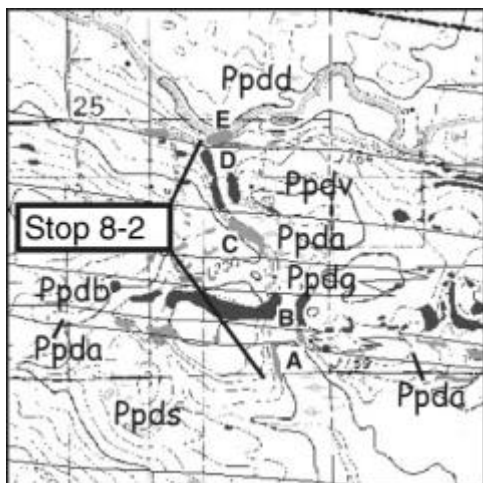


Figure 10. Bedrock geologic map of the Denham Formation type locality, showing stop locations. Black circles are locations of exploratory drill holes and gray areas are mapped outcrops. **McGrath Gneiss**-Amg; **Denham Formation**-Ppds (metasiltstone), Ppda (meta-dolomitic arkose), Ppdb (metabasalt), Ppdg (metapelite), Ppdv (meta-fragmental mafic volcanic rocks), Ppdm (dolomitic marble); **Little Falls Formation** equivalent metagraywacke – Ppgs.

--Return to vehicles. Drive back east 0.75 miles on State road 52 to a small gated trail to the south. Walk south down trail for approximately 0.5 miles following edge of ravine to south-most outcrop on west side of ravine (see inset map). This stop will traverse from south to north over outcrops located along west edge of ravine. There are also outcrops on the east side of the ravine that show parts of the stratigraphy not exposed on the western valley traverse. ****Private Property****



Stop 8-2: Denham Formation

Location: T.45N., R.21W., Sec. 25, SE 1/4

Denham 7.5' quadrangle

UTM Start: 505406 easting; 5132575 northing

UTM end: 505208 easting, 5133330 northing

HIGHLIGHTS: INTERBEDDED BIOTITE SCHIST(SILTSTONE), PILLOW BASALT, DOLOMITIC ARKOSE, AND DOLOMITIC MARBLE.

From bottom to top, the Denham Formation consists of fine-grained metamorphosed siltstone (Ppds), dolomite-cemented arkosic arenite with rare cobbles of McGrath gneiss (Ppda), pillowed basalt flows (amphibolite; Ppdb), shale (staurolite-garnet-mica schist; Ppdg), interbedded dolomitic arkose and dolostone (Ppda), fragmental basaltic volcanic rocks (amphibolite; Ppdv), and at the top of the section, pure dolostone (marble; Ppdm). See Figure 8 (repeated below) for stratigraphic section of the Denham Formation.

A. (South most). Gray biotite schist interpreted as metamorphosed siltstone (unit Ppds), overlain by brownish-tan weathered dolomitic arkose (unit Ppda).

B. Pillow basalt flows (unit Ppdb). Classic basalt flow sequences, with massive bases that grade upward (north) into amygdaloidal pillow basalt, which in turn grades to coarsely fragmental or blocky flow tops. The fragmental and amygdaloidal character of the flows indicates eruption into a shallow-water environment. Two distinct flows are present in the valley, but in hills to the east where the unit thickens, as many as four flow sequences were identified. Strong stretching lineation evident in fragmental portions of flow is accentuated by differential weathering between clasts and matrix. Basalt geochemistry shows slight calc-alkaline tendencies (Southwick and others, 2001).

Outcrop gap is inferred to represent garnet-staurolite-mica schist (unit Ppdg) that is exposed both east and west of this gap.

C. Dolomitic arkose (unit Ppda). Weathered reddish-tan color due to abundant dolomite in matrix. Coarse sand-sized detrital grains of orthoclase and quartz, and rare cobbles identical to the McGrath Gneiss. The original clastic grains in the arkosic rocks are well preserved despite metamorphism, largely because strain has been partitioned into the comparatively more ductile dolomitic matrix. As discussed at the McGrath gneiss stop (8-1), the arkosic rocks contain almost no plagioclase, presumably due to sediment derivation from a weathered McGrath Gneiss source.

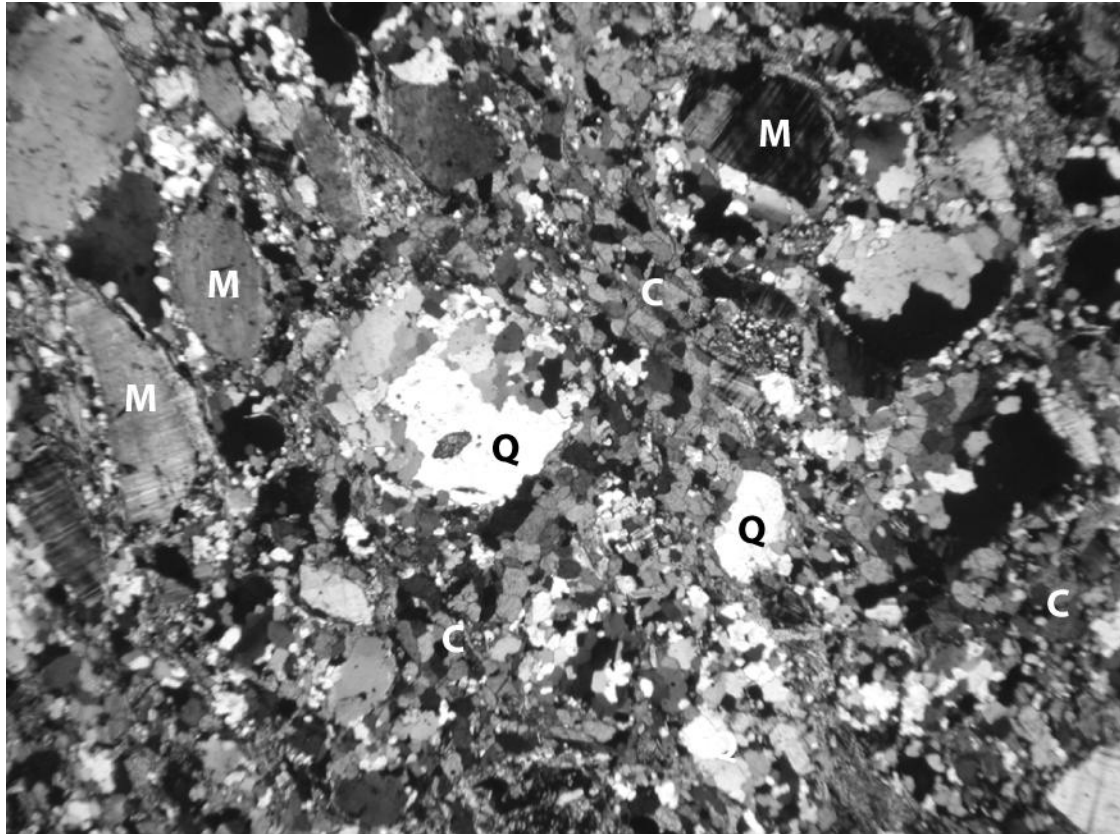


Figure 11. Photomicrograph of dolomitic arkose from Stop 8-2C.

D. Fragmental volcanic rocks (unit Ppdv). Dark green amphibolitic clasts and scattered clasts of white cherty material are strongly flattened and lineated. In map view, this fragmental unit nearly merges to the east with the underlying pillow basalt, possibly indicating a common volcanic source located to the east beneath the Keweenaw Fond du Lac Formation. Holm and others (1993) obtained Ar/Ar plateau ages of $1,756 \pm 16$ Ma on a hornblende separate, and 1754 ± 13 Ma on a biotite separate from this amphibolitic unit.

E. Dolomitic marble (unit Ppdm). Tan weathered, dark gray fresh. In drill holes (cores and cuttings) to the east marble changes from gray to white with depth, with intercepts of massive dolomite as much as 500 feet.

STRATIGRAPHIC COLUMN, DENHAM TYPE LOCALITY

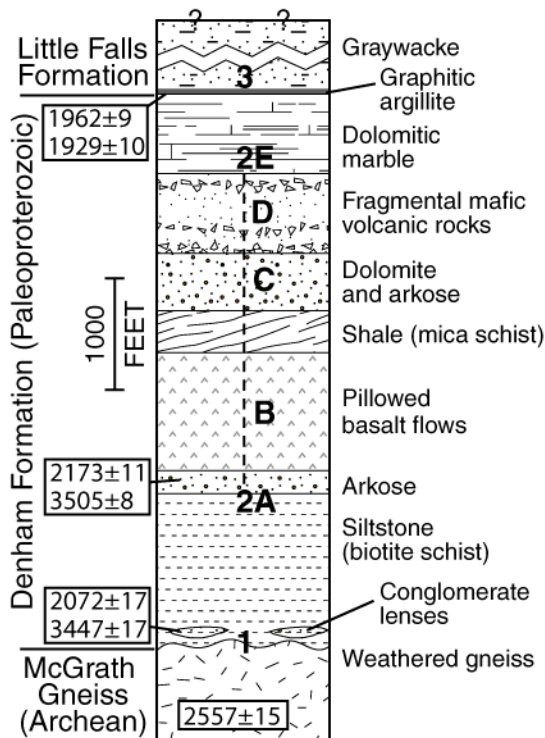
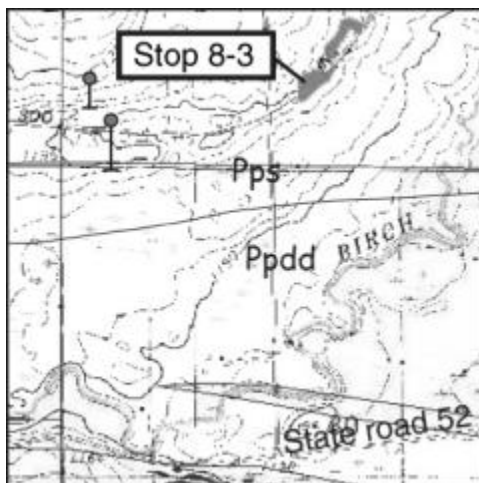


Figure 7 repeated. Stratigraphic section of the Denham Formation, showing relative locations of field trip stops. Age of McGrath gneiss from nearby outcrop, and age ranges of detrital zircons shown in boxes.

---Return to vehicles. Drive east approximately 0.75 miles on State road 52 to a "T" intersection with gravel road to the north. Go north on gravel road 0.65 miles to junction of the Soo Line Trail – a former railroad grade now utilized as an ATV trail. Park at trail junction and walk northeast on trail about 500 feet to outcrops and low road cuts.



Stop 8-3: Metagraywacke and argillite

Location: T.45N., R.20W., Sec. 19, SE 1/4
Denham 7.5' quadrangle
UTM: 506343 easting; 5134441 northing

**HIGHLIGHTS: GRADED BEDS,
METAMORPHIC ASSEMBLAGE**

Railroad cuts and flat outcrops of metagraywacke with pelitic beds. Pelitic units are micaceous and contain small garnet and staurolite crystals, although the latter are more abundant further to the northeast.

Abundant 1 mm diameter garnets are synkinematic with respect to S_1 , and have a well-developed internal schistosity defined by quartz and ilmenite inclusions (Figure 12). At this locality, inclusion-rich cores are surrounded by haloes of inclusion-free garnet, and staurolite has overgrown both the schistosity and the crenulation cleavage. Garnet rim analyses give final equilibration temperatures of 520-590 degrees C and a pressure of ca. 6 kbar (Holm and Selverstone, 1990). Chemically homogeneous monazite grains at this locality give a prominent metamorphic age domain of ca. 1788 ± 4 Ma and a less prominent age domain of 1840 Ma (McKenzie, 2004).

In the series of outcrops to the northeast of this stop, bedding and S_1 cleavage are gently folded along a series of shallow east-plunging, northeast-striking F_2 fold axes which are oriented similar to those at the last stop but are more open and upright. A series of 10-30 cm-thick graded beds and scoured crossbedding visible in the flat outcrop north of the trail give a sense of younging to the north.

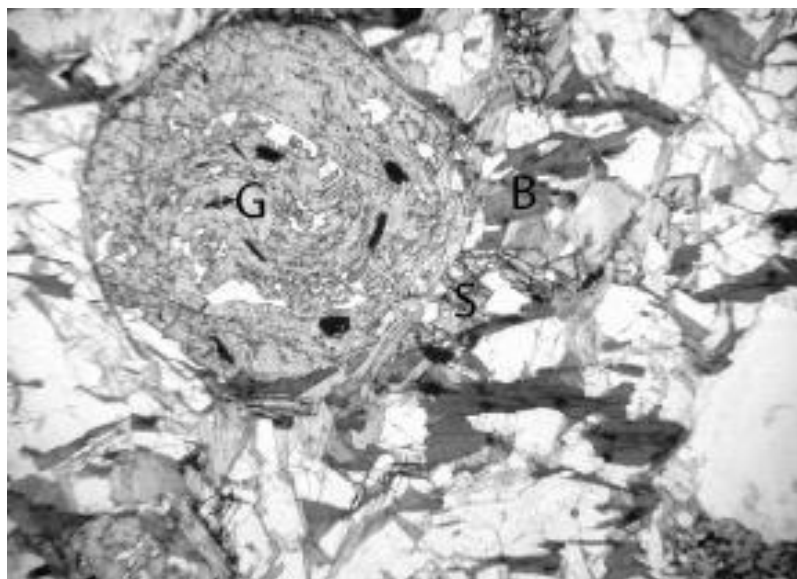
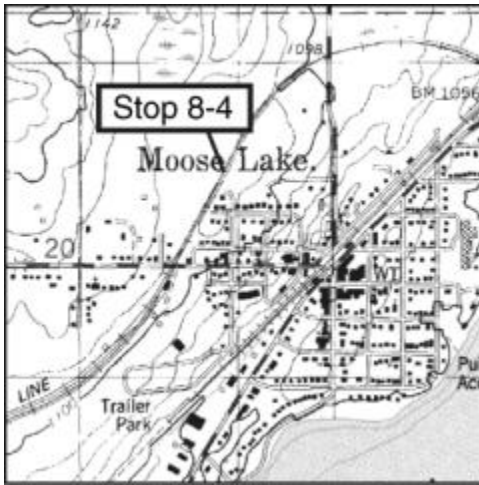


Figure 12. Photomicrograph of rolled garnet from stop 8-3. G – garnet; S – staurolite; B- biotite, matrix is quartz and plagioclase.

Drill cores from approximately 1 mile southeast of this stop (locations shown on Figure 10) show that this metagraywacke conformably overlies the marble to the south (Stop 8-2). The transition between the graywacke and marble is marked by a thin (1 meter) interval of carbonaceous argillite.

---Return to vehicles. Take State road 52 back east to Highway 61. Turn left (north) on Highway 61 and go approximately 7.5 miles to junction with Highway 27 west at stop light in Moose Lake. Turn left (west) on Highway 27 a short distance to the Soo Line ATV trail; just before trail turn left into parking lot. Walk north along Soo Line trail for approximately 0.2 mile to outcrops along trail.



Stop 8-4: Schist at Moose Lake

Location: T.46N., R.19W., Sec. 20, NE 1/4

Moose Lake 7.5' quadrangle

UTM start: 517815 easting; 5144635 northing

UTM end: 518205 easting, 5145203 northing

**HIGHLIGHTS: GARNET-GRADE
METAGRAYWACKE AND PELITIC SCHIST;
FOLDED BEDDING AND S1 FOLIATION**

This pelitic schist is part of the same belt of rocks as the last stop, but here is north of the staurolite isograd (McSwiggen, 1987). A well-developed, bedding-parallel, moderately south-dipping S1 foliation is folded by steeply to moderately south-dipping F2 folds which are accompanied by an axial-planar crenulation cleavage (Figure 13).

Internal inclusion trails in small (0.5 mm diameter) garnets show as much as 160 degrees of rotation. These garnets give final equilibration temperatures of 440-500 degrees C. The age of metamorphism here is ca. 1830 Ma based on monazite ages obtained from a similar grade rock south of this locality.

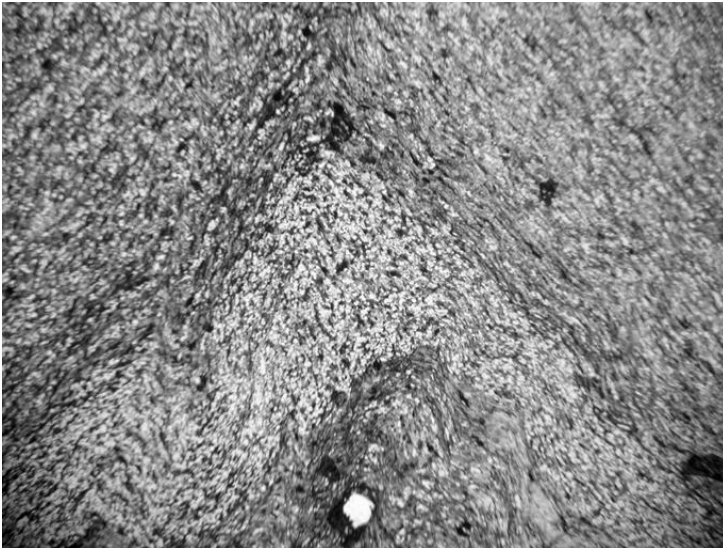
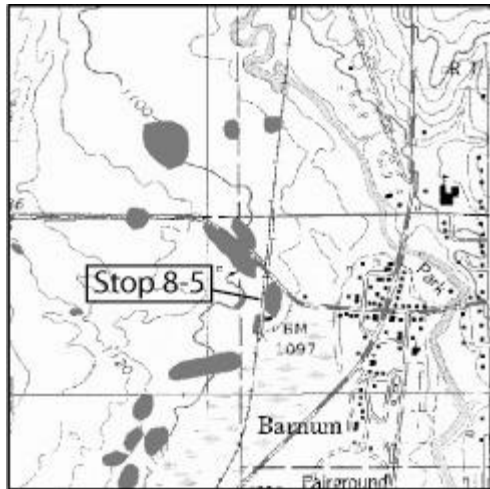


Figure 13. Photomicrograph of crenulated micaceous phyllitic schist from this stop. Fine-grained micas define S1 fabric which is parallel to bedding; both are folded by D2 deformation.

An east-northeast trending diabase dike of presumed Mesoproterozoic age may be visible towards the north end of this set of outcrops, but the exposure has become quite overgrown. The dike is well exposed farther north along this trail but this trip will not go there.

---Return to vehicles and drive back east to Highway 61. Turn left (north) on Highway 61 for approximately 4.5 miles to Barnum, turn right (west) on County Highway 6 for a short distance and park at the Willard Munger trail parking lot.



Stop 8-5: Phyllitic schist at Barnum

Location: T.46N., R.19W., Sec. 1,
NW ¼ and Sec 2, NE ¼.

Barnum 7.5' quadrangle

UTM: 523180 easting; 5149900 northing

**HIGHLIGHTS: BIOTITE-GRADE
METAGRAYWACKE AND PELITIC SCHIST;
FOLDED BEDDING AND S1 FOLIATION.**

Fine-grained, phyllitic, silvery micaceous schist metamorphosed to biotite isograd; this location is north of the garnet isograd (McSwiggen, 1987). Contains a strong subhorizontal to south-dipping, approximately bedding parallel foliation that is weakly folded along small east-plunging (5-10 degrees) fold axes. A weak S2 cleavage parallels the fold axes; in thin section a strong S2 crenulation cleavage is obvious (Figure 14). Micaceous minerals are dominantly chlorite and sericite with minor biotite.

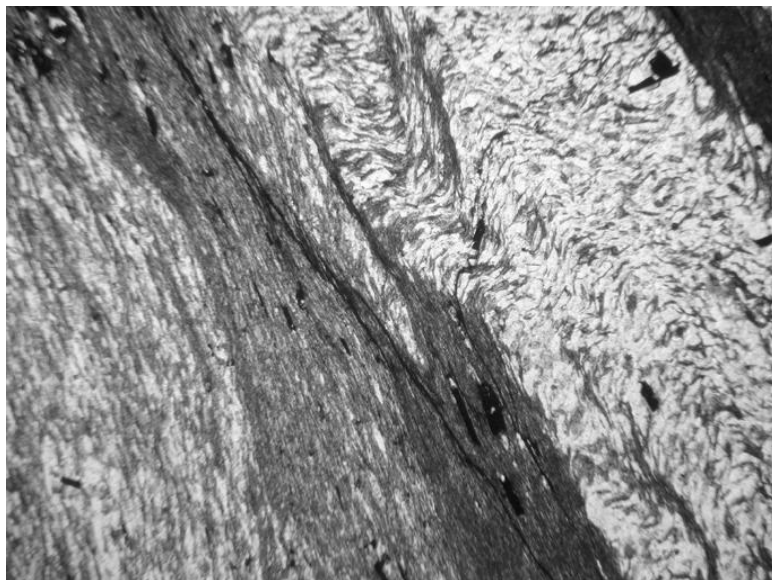
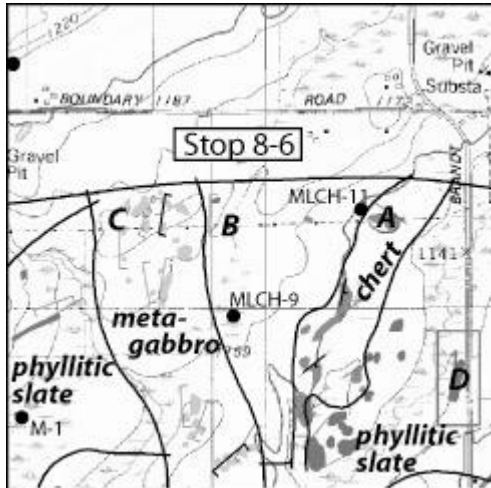


Figure 14. Photomicrograph of crenulated micaceous phyllitic schist from this stop. Earlier S1 foliation is strongly overprinted and transposed by S2 fabric; S1 is refracted through a sandy bed, in upper right part of photograph.

--Return to Highway 61. Turn left (north) on Highway 61 for 6 miles to Mahtowa, continue northeast for another 0.5 miles. Veer left (north) on Brandt Road for 0.8 miles and turn left (west) onto Boundary Road/Township Road 85 after the first curve and proceed to second driveway on the

left (3260 Boundary Road). ****Private property permission must be obtained before entering****
 From farmhouse walk south on trail across pasture to powerline, then east across low ground to knob of white quartz.



Stop 8-6: Base of Thomson Formation, mafic intrusion

Location: T.47N., R.18W., Sec. 4, NW ¼ and Sec 5, NW ¼.

Barnum 7.5' quadrangle

UTM start: 528370 easting; 515950 northing

UTM end: 527540 easting, 5159462 northing

Highlights: Chert, pyrite porphyroblasts, metagabbro.

This stop is just south of what was ‘traditionally’ considered to be the south edge of the Animikie basin, but is now considered to be the lowermost part of it, based on remapping and analogues to Michigan. There are a variety of rock types here including recrystallized chert, metagabbro/diorite, and pyritic slate. We will first walk east along the powerline to point ‘A’ on the above figure, then will go back west to examine outcrops of phyllitic slate (B), and then to outcrops of sheared, porphyritic metagabbro (C). On the index map above the three black dots are the locations of core holes drilled in the 1980’s by Rocky Mountain Energy. Summaries of those drill core logs are listed at the end of this stop.

Substop A: This is a small knob composed mainly of recrystallized black chert that is flooded by white to pinkish-colored, pre-cleavage, quartz veins. A smaller outcrop to the west contains abundant small cubes of oxidized pyrite and is stained yellow and red. Petrographically this chert is composed of fine-grained saccharoidal quartz, minor muscovite, and (where black) abundant submicroscopic dusty opaque inclusions that are not identifiable by optical methods. Scattered rhombic shapes imply that there were dolomite porphyroblasts, which are now replaced by fine-grained quartz. McSwiggen (1987) identified thin intervals of dark gray chert in the drill cores drilled adjacent to this outcrop as being phosphatic. Other drill cores obtained along this lower, basal-most unit of the Animikie basin further west also intersect thin to thick beds of chert interbedded with pyritic slate.

Substop B. Between substops A and C are some small outcrops of dark gray, fine-grained phyllitic slate/schist in which bedding and S1 cleavage are kinked by a later S2 fabric. Scattered 2-3 cm square pits represent weathered-out cubes of pyrite. Also visible are small dark green clots that may be retrograde-altered porphyroblasts formed during an earlier prograde metamorphic event. The dark retrograded porphyroblasts and the pyrite cubes are both found elsewhere at this same stratigraphic horizon, for example near Kalevala township and near the Tamarack intrusion to the west (Figure 3). The retrograded

porphyroblasts (possibly after garnet or andalusite) contain an earlier, rotated S1 metamorphic foliation; pressure shadows are also evident (see Figure 15).

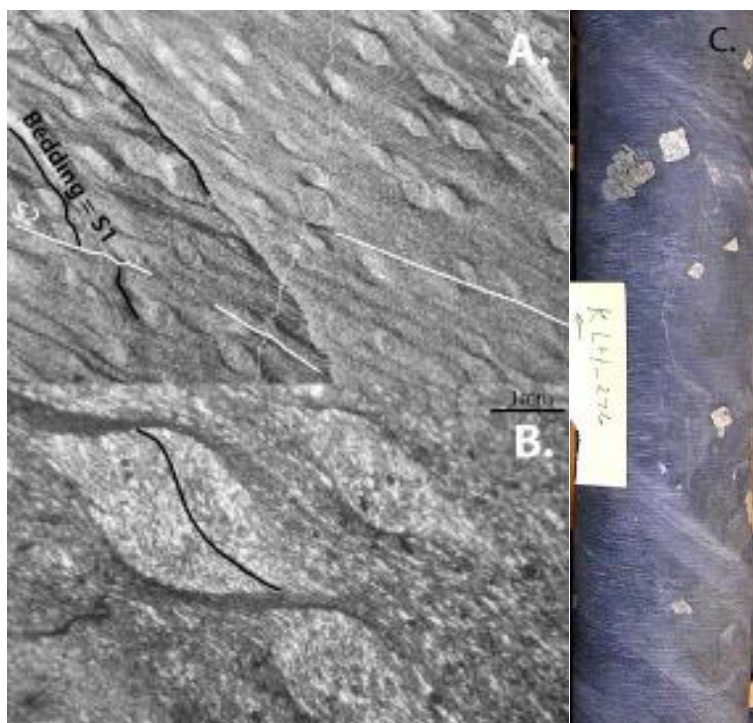


Figure 15. A.) Photomicrograph of drill core samples showing rotated, retrograde-altered porphyroblasts similar to those present at this stop. Bedding-parallel S1 foliation is highlighted by black lines and S2 fabric by white lines).

B.) Close-up of rotated porphyroblasts.

C.) Photograph of drill core from approximately 15 km west of this stop from the same stratigraphic horizon, showing characteristic pyrite porphyroblasts.

Substop C: Green, variably shear-foliated metagabbro metamorphosed under greenschist-facies conditions. In less deformed lozenges 5-10% white plagioclase phenocrysts up to 3cm in size are preserved, but most of the phenocrysts are broken and strung out in the foliation plane and appear as flaser-like grains. Schist is composed of sericite, quartz, carbonate, chlorite, sphene/leucosene, and minor sulfides, and is locally overprinted by rhombic carbonate and rarely, coarse-grained muscovite porphyroblasts. The foliation dip varies from 25 degrees east at the north end of the unit to 60 degree east at the south end; this is at a high angle to the regional fabric and is likely the result of cleavage refraction through this more competent unit. In thin section the main (presumably S1) fabric is weakly folded by a later D2 crenulation cleavage.

Summary of drill core logs; locations of core holes shown on location map above:

Drill Core MLCH-9 (vertical hole)

39-53 – Thinly laminated graphitic argillite.

53-430 – Strongly sheared/mylonitic, porphyritic metagabbroic rock. Flattened plagioclase phenocrysts give the rock a flaser-gneiss like appearance. Below 395 is generally less deformed.

430-550 – Thinly laminated tightly folded graphite-rich interbedded slate and metagraywacke with pyrite cubes, ankerite clots, and some bedding-parallel pyritic layers. Last 5 feet is strongly graphitic.

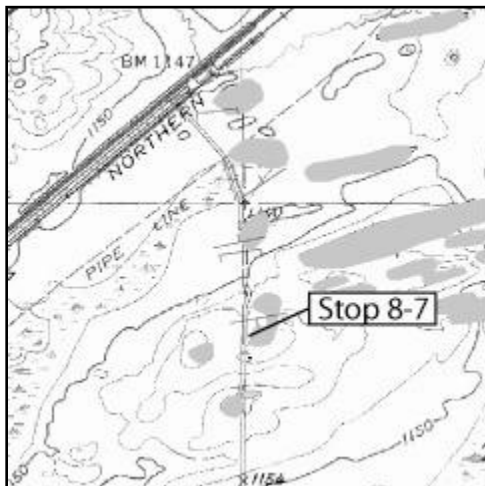
Drill Core MLCH-11 (0, -60)

- 9-36** – Dark gray sooty graphitic phyllitic argillite, variably pyritic, with pyrite cubes and local thin layers semi-massive sulfides.
- 36-42** – Fine-grained siliceous graywacke with white siliceous lapilli-like fragments.
- 42-45** – Dark grayish-black chert, similar to nearby outcrop.
- 45-83** – Meta-argillite and graywacke, variably sulfidic and graphitic, minor chert.
- 83-137** – Variably shear-foliated metagabbroic rock.
- 137-242** – Fine-grained thinly laminated argillite with carbonate spots, locally graphitic, local layers rich in disseminated sulfides.
- 242-354** – Variably shear-foliated, porphyritic metagabbroic rock.

Drill Core ML-1 (azimuth and inclination unknown; core incomplete)

- 9-105** – Weakly crenulated phyllitic slate, argillite, and graywacke, locally graphitic with minor stratiform sulfides.

---Return to Highway 61. Turn left (north) on Highway 61 for 6.6 miles to Gillogly Road. Turn right (south) on Gillogly Road and proceed 0.64 miles around curve to small outcrop just north of the crest of the hill, on the east edge of the road.---



Stop 8-7:

Weakly D2 – crenulated slate and graywacke

Location: T.48N., R.17W., Sec. 21, NW 1/4

Iverson 7.5' quadrangle

UTM Start: 537374 easting; 5164270 northing

HIGHLIGHTS: S1 FOLIATION AT A HIGH ANGLE TO BEDDING, BOTH AFFECTED BY D2 FOLDING.

This stop is near the northern edge of the twice-deformed portion of southern edge of the Animikie basin. Graded beds strike roughly N60W, 60°N, with younging directions upright to the northeast. Bedding can easily be traced by following layers of carbonate concretions. This is one of the rare places where S1 foliation (developed mainly in the upper pelitic portions of the graded beds) is observed at a high angle to bedding. F2 fold axes plunge approximately 15° to N85E. This is one of the key outcrops used by Holst (1984; Figure 16) to conclude that the entire area of twice-deformed metasedimentary rocks is on the upper limb of a large, north-directed thrust nappe. The following sketch is from Holst (1984), showing the relationships between bedding (S0), early S1 cleavage, and later S2 cleavage.

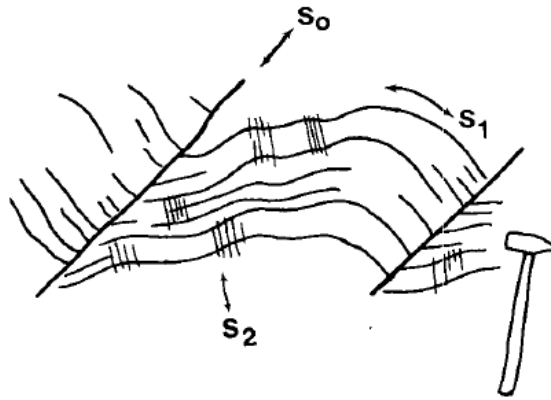
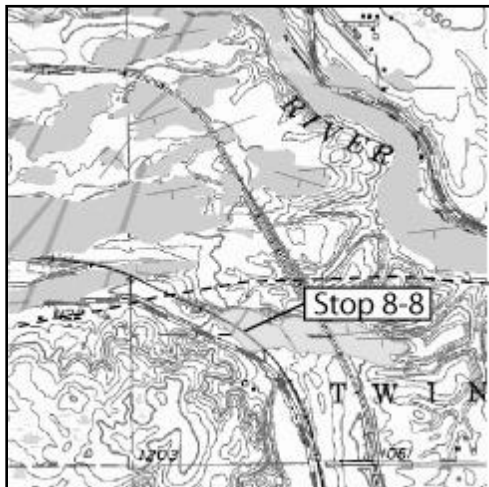


Figure 16. Sketch showing relationships between bedding (S_0), early S_1 cleavage, and later S_2 cleavage for stop 8-7. From Holst, 1984).

--Go back to Highway 61 and continue north approximately 3 miles to the junction with Highway 210. Turn right (east) on Highway 210 and go approximately 2 miles into town, then take a right on County Highway 3/1 for 0.6 miles then stay left (east) on County Highway 1. Follow Highway 1 for 1.3 miles then turn left (east) on small road and park at Munger Trail adjacent to Highway 1. Walk north approximately 400 feet on the Munger trail to rock cuts on both sides of trail. ---



Stop 8-8:
Slate and graywacke with minimal S_1 foliation

Location: T.48N., R.17W., Sec. 21, NW 1/4
Cloquet 7.5' quadrangle
UTM Start: 545770 easting; 5166742 northing

**HIGHLIGHTS: VERY WEAK S_2 FOLIATION,
KEWEENAWAN DIABASE DIKE**

This stop is the northern-most location within the southern, twice-deformed terrane, and at this point the S_1 fabric which is obvious to the south is nearly absent. On the index map above, the dashed line is the break between one deformation (north of the line) and two deformations (south of the line), as mapped by Clark (1985). Here the dominant cleavage is subvertical and parallel to gentle east-west trending F_2 fold hinges; the earlier S_1 cleavage is very weak and difficult to see. Thin sections show a very weak bedding-parallel foliation cross-cut by the stronger S_2 cleavage (Figure 17). Bedding varies from thinly bedded slate-graywacke sequences to thicker sandy graywacke beds up to 3 meters thick. Due to the nature of the outcrop the bedding can be hard to observe and a good way to see it is by

observing patterns of cleavage refraction through pelitic and sandy beds, and in a few places graded beds can be seen which show upright younging directions. Carbonate concretions are weakly flattened in the slaty cleavage plane.

This stop and the last are both within classic Animikie basin-type sediments, i.e. low-grade graywacke-slate sequences, but unlike the bulk of the Animikie basin to the north, these have been twice-deformed, in the same manner as the phyllites and schists to the south which show progressively higher grades of metamorphism from north to south, as seen on earlier trip stops. The metamorphic grade and style of deformation here is virtually identical to that of the Thomson Formation at the type locality at Thomson Dam, which will be the last stop of this trip.

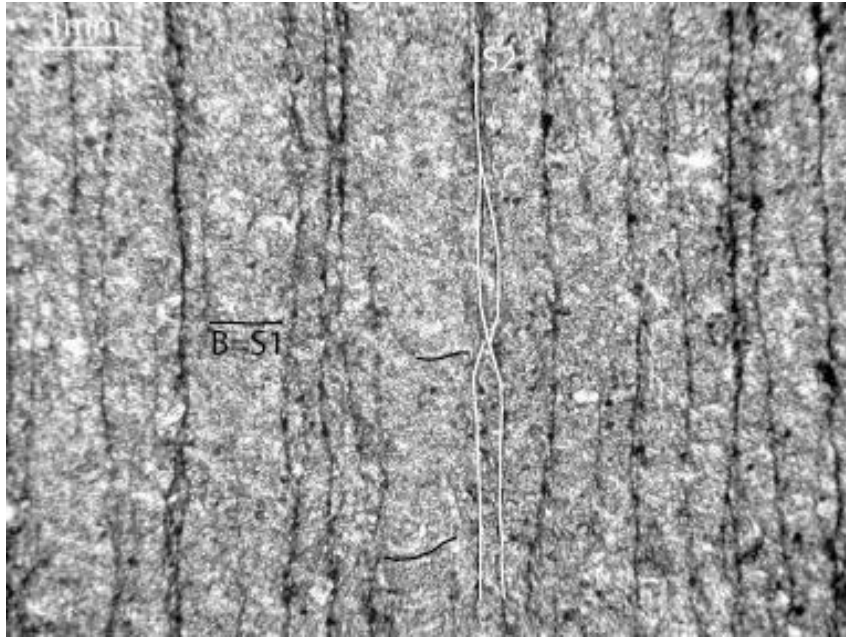


Figure 17. Photomicrograph of slate from stop 8-8 showing very weak early S1 foliation (highlighted by black lines; parallel to bedding which is not visible in this photo), cut by S2 cleavage that is parallel to D2 fold axes (highlighted by white lines).

Diabase Dike: Towards the north end of this outcrop the graywacke is cut by a near vertical, 6m-wide diabase dike that is oriented approximately N45E. The dike has strongly chilled margins and in the center is fine-medium grained. Petrographic examination shows the chilled margin to be devitrified glass with small spherulitic or variolitic structures; the dike interior is composed of plagioclase, sub-prismatic augite, opaque Fe-Ti oxides, and minor olivine. Approximately 30% of the rock in the dike interior is turbid, opaque-dusted mesostasis that is likely devitrified glass which contains abundant small prismatic pyroxene grains. This dike is part of the Carlton dike swarm; see stop 8-10 for further description.



**Stop 8-9:
Slate and graywacke of Thomson Formation;
Keweenaw diabase dikes**

Location: T.48N., R.16W., Sec. 5, SW 1/4
Cloquet 7.5' quadrangle
UTM Start: 546612 easting; 5168119 northing

**HIGHLIGHTS: VARIETY OF SEDIMENTARY
STRUCTURES, FOLDS, DIABASE DIKES**

This is the type locality for the Thomson Formation, which is southern equivalent of the Virginia Formation along the Mesabi Iron Range. The Virginia Formation overlies the Biwabik Iron Formation, and collectively the Virginia/Thomson Formations cover a broad area (the Animikie basin) that extends south from the Mesabi Iron Range to a few miles south of this locality (Figure 1). Deformation of the Animikie strata increases to the south, in closer proximity to the Penokeyan fold and thrust belt. Here, near the south edge of the Animikie basin, the strata are folded into a series of gently east- and west-plunging, open symmetric to locally asymmetric, slightly overturned folds with near-vertical axial-planar cleavage.

Several northeast-trending diabase dikes of Mesoproterozoic age cut the Thomson Formation. These dikes are part of the Carlton County dike swarm, which is one of a number of dike swarms that flank the Midcontinent rift system. The dikes here were emplaced along northeast-trending joints in the graywacke, and they exhibit well-developed subhorizontal columnar cooling joints and have chilled margins. The approximately 2 meter wide dike just below the parking area is reversely polarized, and like most of the dikes in the Carlton County swarm has a composition of high Fe-Ti continental tholeiite (Green and others, 1987). Overall, dikes of the Carlton swarm range from centimeters to greater than 60 meters in width. Narrow contact metamorphic albite-epidote hornfels are found adjacent to the contacts of only the thicker dikes.

Green and others (1987) have summarized the geologic, geochemical, and paleomagnetic data on the dikes, and has demonstrated that these swarms are most likely Mesoproterozoic in age.

---Return to your vehicles in the parking lot to end the trip.---

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57TH ANNUAL INSTITUTE ON LAKE SUPERIOR GEOLOGY

FIELD TRIP 9

GRANITIC, GABBROIC, AND ULTRAMAFIC ROCKS OF THE KEWEENAWAN MELLEN INTRUSIVE COMPLEX



Intrusive breccia near Mellen, Wisconsin – T. Fitz

Field Trip 9

Granitic, gabbroic, and ultramafic rocks of the Mellen Intrusive Complex in northern Wisconsin

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INTRODUCTION

The Mellen Intrusive Complex (MIC) is a group of Keweenawan Midcontinent Rift (MCR) intrusions underlying 380 km² in northern Wisconsin (Fig. 1). The Complex was intruded as large sill-like bodies into the lower part of the Keweenawan volcanic sequence 1102 Ma ago (Zartman et al., 1997). The rocks of the MIC, along with rest of the rocks in the region, were tilted steeply north during the compression and folding that created the Lake Superior Syncline (Cannon et al., 1993a). Subsequent erosion has exposed a complete cross section of this magmatic system – from the feeder dikes that brought magma from depth, to the subvolcanic magma chambers, and the overlying volcanic sequence (Woodruff, 2005). The MIC is bi-modal in composition, meaning mafic and felsic rocks dominate and there is a relatively small volume of intermediate rocks. This is similar to the rest of the igneous rocks of the MCR, and is typical of continental rifts. Anorthositic gabbro and granite are the two most common lithologies in the MIC, but there is a wide variety of rock types. This variety of rocks, as well as the fact that they are mostly pristine and well exposed, makes the MIC an excellent showcase of igneous petrology. The goal of this trip is to see this variety of rocks.

The Mellen Complex is one of four major intrusive complexes of the MCR – these are: the Duluth Complex in northeastern Minnesota; the Nippigon sills near Lake Nippigon in Ontario; the Coldwell Complex on the north shore of Lake Superior in Ontario; and the Mellen Complex in Wisconsin (Miller, 2007). Another MCR layered mafic intrusion, the relatively small Clam Lake Intrusion, occurs at depth in the Clam Lake area 25 km southwest of the MIC. It is not exposed at the surface but is known from its magnetic and gravity signatures, and from two drill cores (Mudrey et al., 2003).

Midcontinent rift magmatism produced an estimated 1,000,000 km³ of igneous rock over a period spanning 29 Ma. (Klewin and Shirey, 1991; Miller, 2007). It occurred in three main pulses: an early episode from about 1115 to 1107 Ma; the main phase that produced a huge volume of volcanic and plutonic rocks from about 1102 to 1092 Ma; and a third, less prodigious phase, from 1092 to 1086 Ma (Paces and Miller, 1993; Miller, 2007). Magnetic

polarity reversals occurred during rifting, so the remnant magnetism of the rocks has been important in deciphering the history of the MCR and making correlations between rocks in different parts of the rift: The early pulse of magmatism occurred during a time of reverse magnetic polarity, followed by a switch to normal polarity before the main phase of activity (Halls and Pesonen, 1982).

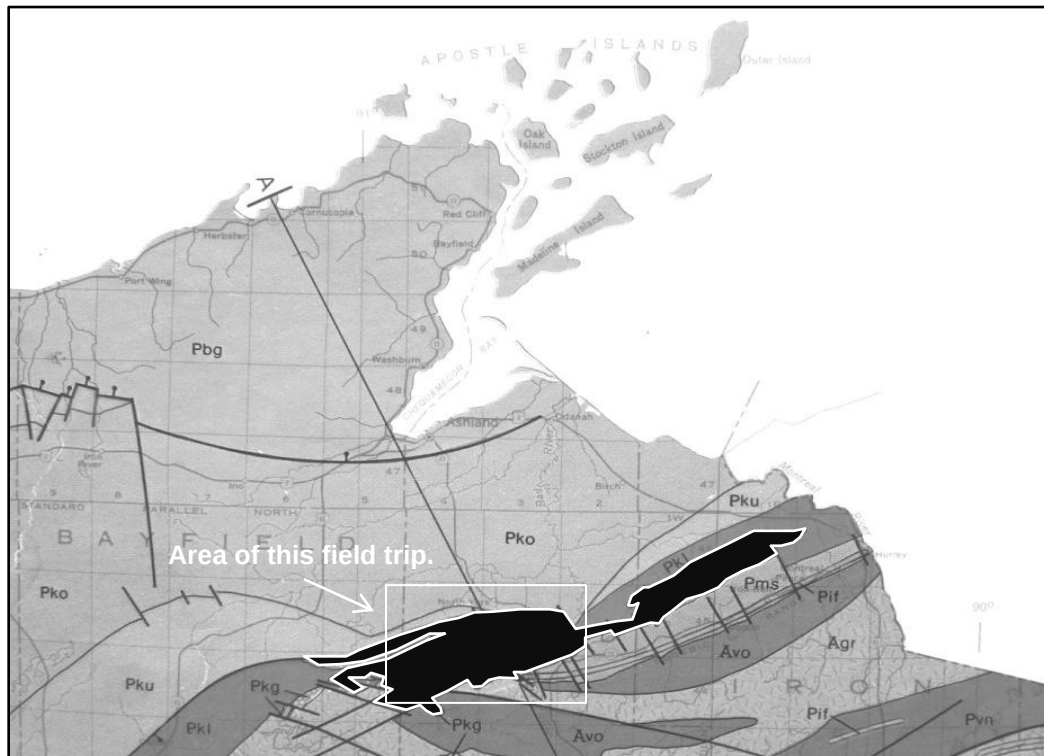


Figure 1. Location map showing the Mellen Intrusive Complex in black, and the area of this field trip outlined in the white box. From the Bedrock Geologic Map of Wisconsin (Mudrey et al., 1982)

The age of the MIC is known from U/Pb dating of zircon crystals from granite which yield an age of 1100.9 ± 1.4 Ma, and from granophyre that give an age of 1102 ± 2.8 Ma (Zartman et al, 1997). These dates, along with the contact relationships within the MIC, indicate the magma was intruded over a relatively short period of time very early in the main pulse of MCR magmatism. Its age is nearly the same as the Kallander Creek Volcanic rocks which it intruded, so the intrusions of the MIC probably represent crystallized magma chambers in the subvolcanic plumbing system of a large volcanic center located near what was is now Mellen, Wisconsin (Cannon, 1993b).

INTRUSIONS OF THE MELLEN COMPLEX

The MIC consists of four intrusions— from west to east these are: the Mineral Lake Intrusion, the Rearing Pond Pluton, the Mellen Granite in the central region, and the Potato River Intrusions in the east (Fig. 2). The Mineral Lake and the Potato River Intrusions are each made up of two or more intrusions, whereas the Rearing Pond and Mellen Granite appear to be single intrusive bodies (Olmsted, 1979; Klewin, 1990; Seifert et al., 1992).

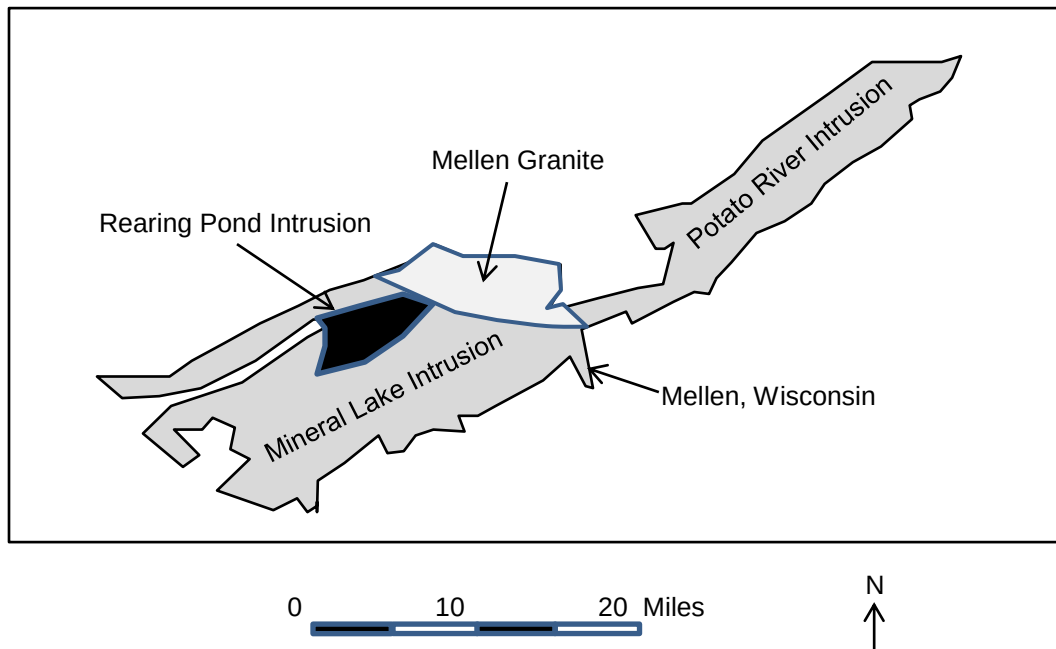
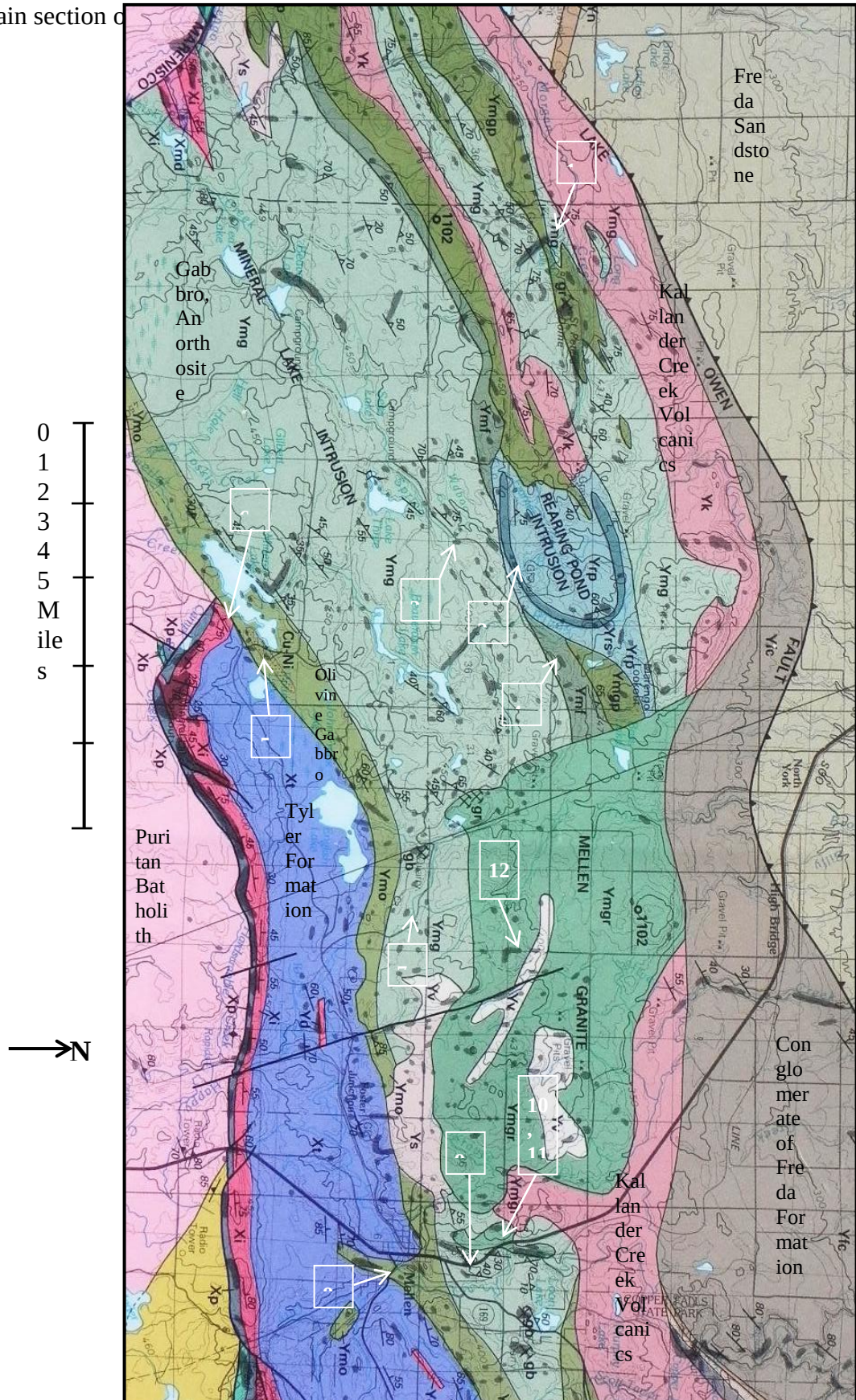


Figure 2. Map of the four main intrusions of the Mellen Intrusive Complex.

The intrusions were emplaced into the lower parts of the volcanic sequence and likely fed magma upward to the surface where it was extruded as lava flows. In the eastern part of the Potato River intrusion, the sill-like gabbro bodies are nearly concordant and are within the lower part of the 2- to 4-km-thick Kallander Creek Volcanics (Klewin, 1990; Seifert et al., 1992). The intrusions are not exactly concordant however, and the level of emplacement gets progressively deeper toward the west (Fig. 3). The bottom of the MIC cuts through the Kallander Creek Volcanics and the underlying Siemens Creek Volcanics in the area of Mellen, Wisconsin. West of Mellen, the intrusions are in contact with the Paleoproterozoic Tyler Formation, and completely separate overlying volcanic rocks from the basement rocks on which they were deposited. In the area of Mineral Lake, the MIC crosses the Tyler and into the underlying Ironwood Formation. Then, finally, at the western end of the MIC the intrusions cut across Archean rocks, so the entire Paleoproterozoic and lower Keweenaw volcanics are missing. In some areas, the intrusions occur at multiple levels and are separated by screens of volcanic rock preserved between parallel sill-like intrusive bodies. The upper intrusions must have been emplaced at relatively shallow levels since miarolitic cavities are quite common. The total thickness of the MIC varies from about 1 km to just over 5 km.

Strong igneous lamination defined by parallel alignment of plagioclase crystals is very apparent in many of the gabbroic rocks. Additionally, the large mafic bodies have large-
ing, with more mafic, olivine-rich gabbroic rocks near the base, which transition
a main section of

Figure 3.
Geologic map
of the western
part of the
Mellen
Intrusive
Complex.
Field trip
stops are
shown in white
boxes.
Formations
are labeled on
the map and
rocks of the
Mellen
Complex have
these
symbols:
Mineral Lake
Intrusion
(shown light
green and
dark green):
Ymo: olivine-
bearing
gabbroic
rocks; Ymg:
gabbro,
anorthositic
gabbro,
gabbroic
anorthosite;
Ymf:
ferrodiorite;
Ympg:
granophyre.
Rearing Pond
Intrusion
(shown in
blue): Yrp:
olivine
gabbro; Yrs:
serpentinized
peridotite.
Mellen
granite
(shown in
medium
green): Ymgr.



felsic rocks near the top. The cumulate texture and broad compositional layering indicates the magmas were emplaced as crystal mushes that underwent crystal settling and fractional crystallization, and possibly some contamination, to create the variety of rocks present today (Seifert et al., 1992). However, well-defined small-scale compositional layering is not common. Some of the granitic intrusions have sharp chilled contacts against the gabbroic rocks, which means that not all of the felsic magma was created by fractionation within the preserved intrusions.

Rocks of the Mellen Complex were extensively quarried for building stone starting in the early 20th century and some quarrying activity continues today. The black, coarse-grained gabbro, which has the trade name “black granite”, was widely used for decorative stone and monuments. It serves as structural stone in many historic buildings in the area, including numerous buildings in the town of Mellen and at Copper Falls State Park. In 1963, a monument made of Mellen “black granite” was shipped to Arlington National Cemetery for the tombstone of the late President John F. Kennedy. Although no cutting and polishing operations remain in the area, several quarries still extract gabbro and granite for ballast rock and rip rap. Small abandoned quarries are common in the area and provide excellent bedrock exposures. The importance of quarrying in the local history is evident in the name of the Mellen school athletic teams – the “Granite Diggers”.

Mineral Lake Intrusion

The Mineral Lake Intrusion (MLI) consists of two intrusive bodies separated by a screen of Kallander Creek Volcanics (Seifert et al., 1992). The lower intrusion is 4.5 km thick and has a basal olivine gabbro, a thick central section of anorthositic gabbro, and an upper ferrodiorite zone. The composition of the plagioclase changes gradually upward in the intrusion from a modal value of An₆₄ near the base to An₂₃ near the top (Seifert et al., 1992). Cumulate texture is common and Seifert et al (1992) concluded that trapped intercumulus liquid constituted 25–35% of the volume of the accumulated crystal mush. The central portion of the intrusion consists of a large volume of coarse-grained anorthositic gabbro composed of plagioclase feldspar, augite, and minor amounts of olivine and magnetite. The igneous lamination defined by the alignment of large plagioclase crystals is distinct in most of this zone. The lamination dips about 70° northwest and is generally parallel to the regional structure, which indicates the crystals are aligned parallel to the floor of the intrusion, probably due to crystal settling.

The central zone grades upward into gabbroic rocks with hornblende instead of olivine and augite, and eventually into diorite with small amounts of quartz and alkali feldspar. The diorite is iron rich and has a relatively high proportion of magnetite, a characteristic special enough to warrant the name “ferrodiorite”. It is a dark-colored rock with an appearance similar to gabbro, but close inspection reveals that its mineralogy is distinctly different than that of gabbroic rocks – it is composed of plagioclase feldspar, hornblende, magnetite, in some places biotite, and minor amounts of quartz and alkali feldspar. It does not contain olivine or augite as do most of the gabbroic rocks lower in the MLI. Based on REE data, Olmsted et al. (1986) concluded that the ferrodiorite is the product of fractional crystallization and that final crystallization took place at a temperature of 850°–750°C.

Granophyre is abundant in the upper parts of the MLI, especially in the upper part of the higher of the two intrusions. The granophyre is brick red and usually fine to medium

grained for a granitic rock. It has spectacular micrographic intergrowths of quartz and microperthitic orthoclase feldspar that often penetrates the entire rock. Although it is usually visible only at the microscopic scale, it is this textural characteristic that distinguishes it from granite. Leighton (1954) found that the micrographic texture is well developed only in rocks in which the An content of the plagioclase is within the feldspar exsolution lamellae is within a restricted range. Klewin (1990) concluded that the roof-zone granophyres were residual liquids intruded along the upper margin of the MLI.

Rearing Pond Intrusion

The Rearing Pond Intrusion (RPI) is a relatively small intrusion mostly contained within the larger MLI. It is interpreted to have intruded as a single magma body into the MLI, but the nature of the contact between the two bodies is unknown because of poor exposure. The RPI is the most mafic intrusion of the MIC, being dominated by Mg-rich olivine gabbro and troctolite. Cumulate layers change upward from an olivine-rich peridotite layer near the base of the intrusion, through olivine gabbro in the middle, to gabbro near the top. The composition of plagioclase changes vertically as well, from An₈₀ near the base to An₆₀ near the top. This sequence is interpreted to have been created by fractional crystallization of olivine, followed by olivine, plagioclase, and clinopyroxene (Olmsted, 1979). Although the basal peridotite layer is partially serpentinized, it locally has well-preserved olivine crystals and igneous textures.

Potato River Intrusion

The Potato River Intrusion (PRI) is similar to the MLI in that it has a lower zone of olivine gabbro overlain by a thick section of anorthositic gabbro that has conspicuous igneous lamination, and granophyre near the top (Klewin, 1990; Tabet and Mangham, 1978). It differs from the MLI in that it has a larger volume of anorthosite.

Mellen Granite

The Mellen Granite (MG) occupies the central portion of the MIC and is in contact with the MLI on the west and the PRI on the east (Fig. 3). It is a medium-grained, grey to pink biotite granite. It includes xenoliths of gabbro and ferrodiorite that closely resemble those rock types elsewhere in the MIC and are probably locally derived. The lack of significant chilled margins in the contact zone indicates the granite was intruded while the mafic rocks were still relatively hot, and the presence of miarolitic cavities indicates the granite was emplaced at a depth of less than about 6 km. Together those facts attest to the rapid succession of emplacement of the MIC high in the Keweenawan volcanic sequence. The granite is interpreted to be a stock intruded near the end of the igneous activity that created the MIC and Kallander Creek Volcanics (Cannon et al., 1993b).

In some areas on the east and west sides of the MG, the contact between the granite and the gabbroic country rocks contains a zone of extremely brecciated gabbro xenoliths (Sikkila, 2002). The gabbro country rocks were complexly intruded by the granite, creating intrusive breccia comprised of angular blocks that appear to have been shattered by the intrusion. Additionally, there are often xenoliths of several different types of gabbroic rocks and ferrodiorite mixed together in the breccia zone, suggesting that the brecciation was followed by mixing of the inclusions within the granite (Cervin, 2009; Goscinak, 2010). The granite in some breccia zones has highly variable texture showing a complex arrangement of

pegmatite, graphic granite, and hypidiomorphic granular granite. The combination of a diverse range mafic rock xenoliths, complex brecciation, and highly variable granitic textures makes this a very distinctive and intriguing intrusive breccia.

FIELD TRIP STOPS

The field trip starts near Sanborn, Wisconsin and goes through the Chequamegon National Forest east to Mellen, Wisconsin on gravel roads maintained by the U.S. Forest Service (Fig. 4). The roads are generally well maintained but may be impassible during mud season. The U.S.G.S. topographic quadrangle maps covering the area of this field trip are, from west to east: Marengo Lake, Mineral Lake, and Mellen, Wisconsin. All locations in this chapter are given in UTM coordinates (Zone 15 T) with NAD 1927, which is the datum for all the local topographic maps. This guide focuses on the western part of the MIC—specifically the MLI, the RPI, and the MG. It has one stop in the very western part of the PRI and a dike that penetrated up through the underlying Tyler Formation into the PRI. A detailed description of the eastern part of the Mellen Complex, with numerous stops in the PRI, can be found in the 1989 ILSG field guide chapter by Klewin, et al.

In this chapter, rock names for the gabbroic rocks are based on the classification system of Miller et al., 2002, and the felsic and intermediate rocks are classified according to the IUGS classification system (Middlemost, 1991; Streckeisen, 1976).

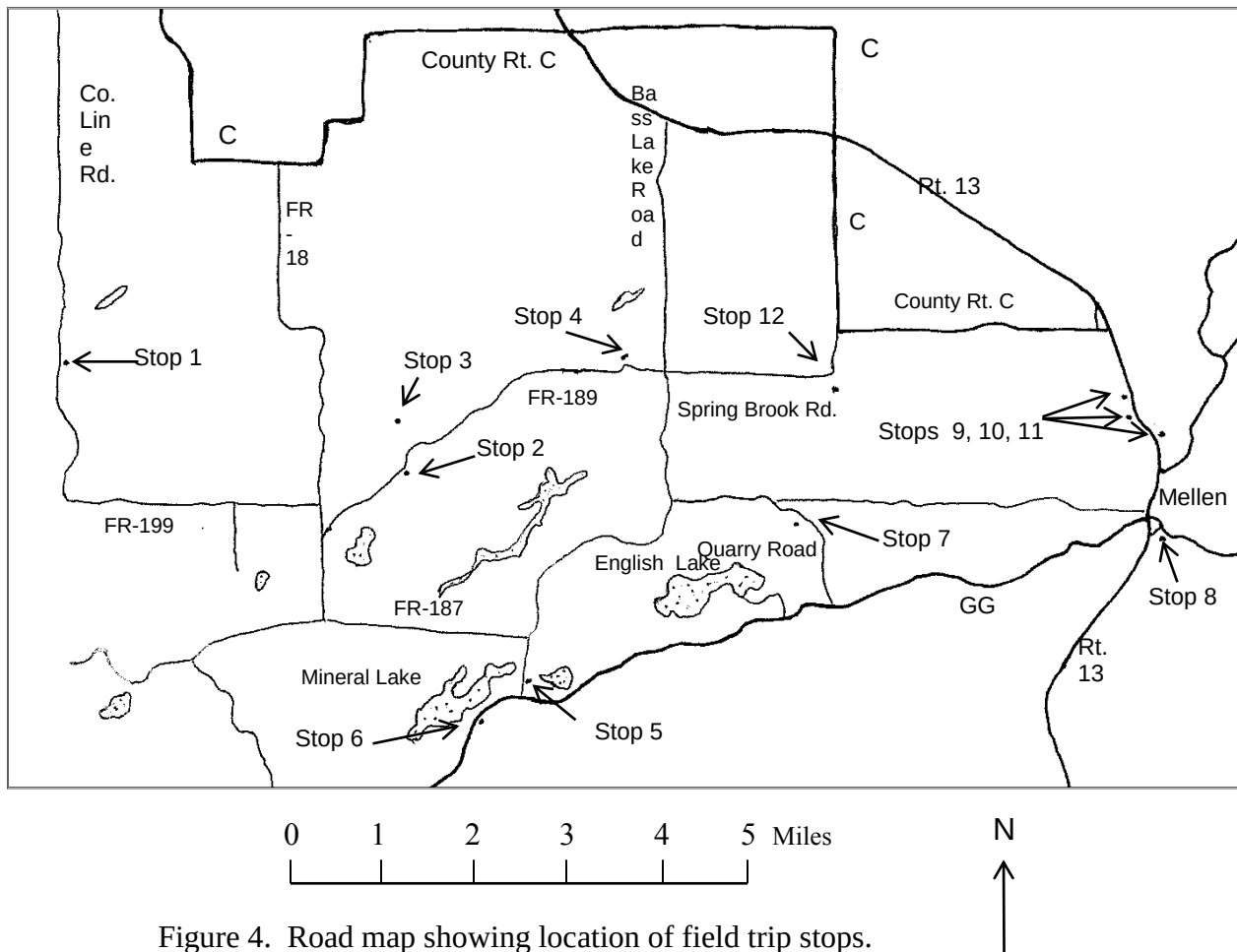


Figure 4. Road map showing location of field trip stops.

Stop 1: Granophyre at Morgan Falls

Location: 0659762 mE, 5134883 mN; T45N, R4W, Sec. 30, west-central.

Park at the Forest Service parking lot for Morgan Falls and St. Peter's Dome located on the east side of Forest Service road FS-199 (Fig. 5). The UTM coordinates listed above are for the parking lot. From the parking lot, walk east on the trail to St. Peter's Dome for approximately 0.25 miles, then bear right (south) at the fork and take the trail to Morgan Falls.

This area has good exposure of gabbro and granophyre in the upper part of the MLI. These intrusions were emplaced as large sills concordant with lava flows of the Kallander Creek Volcanic rocks, which are exposed on the hills across the valley to the north. Screens of Kallander Creek Volcanics are preserved between the sills but are not well exposed because they are relatively non-resistant in comparison to the hard intrusive rocks. In the area of this stop, volcanic rocks of the Kallander Creek served as the roof rocks, below which is a relatively thin roof zone of gabbro, below which is a large granophyre sill that is exposed at Morgan Falls and on St. Peter's Dome (Cannon et al., 1996).

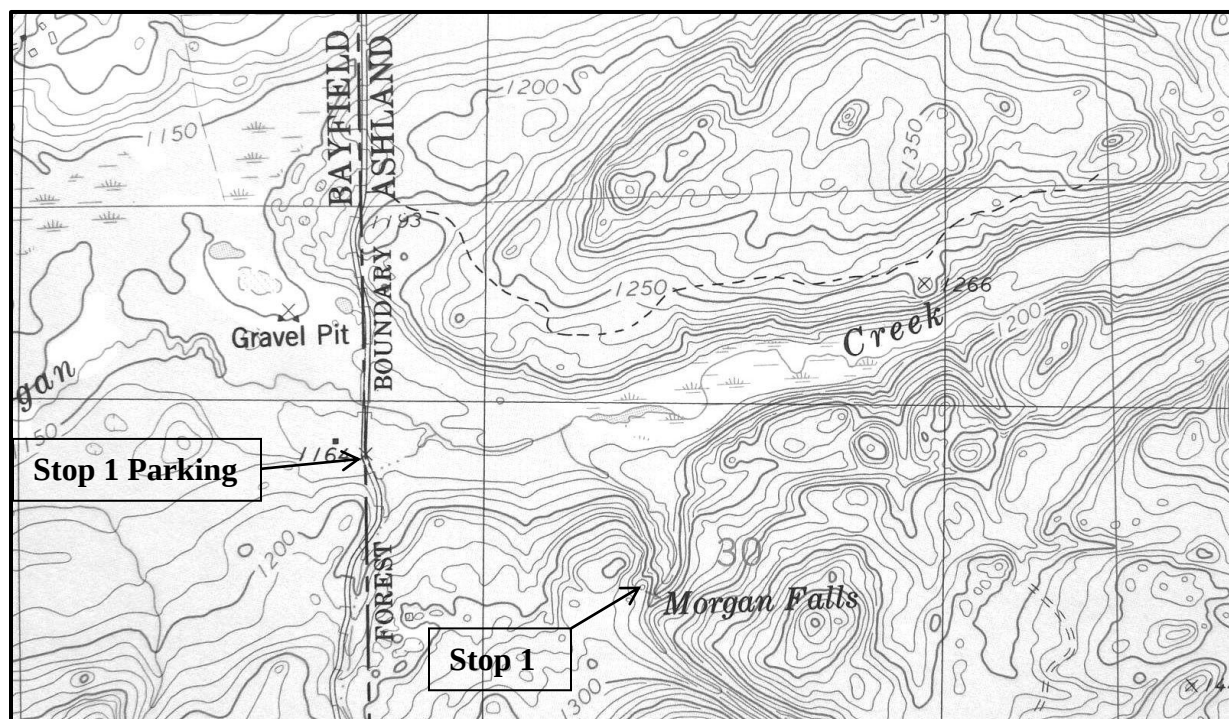


Figure 9. Topographic map showing the area of Stop 1.

Stop 1-A: Gabbro of the Mineral Lake Intrusions

Location: Near Morgan Falls, on the west side of trail; 0660363 mE, 5134684 mN.

The first outcrops encountered along the trail are just south of the bridge over Morgan Creek, on the west side of the trail. These outcrops are all in gabbro of the MLI. This is the uppermost intrusion of the MLI, and this outcrop is approximately 200 – 300 m below the roof of the intrusion. The rock is gabbro, or oxide gabbro, containing approximately 80% plagioclase feldspar, 15% oxide mineral, at least some of which is magnetite, 5%

intercumulus augite and hornblende, and a trace of quartz. This lithology is typical of this part of the MLI.

Stop 1-B: Granophyre at Morgan Falls

Location: At the end of the Morgan Falls trail; 0659762 mE, 5134884 mN.

Scenic Morgan Falls is a 70-foot cascade in a narrow rock canyon where Morgan Creek falls over dark red granophyre. The joint pattern in the rock is distinct and exerts a strong control on the form of the rock face and the flow of the water over the falls.

The rock at the falls is a dark red, medium-grained granophyre. It is composed almost entirely of quartz and microperthitic orthoclase feldspar crystals intergrown in a micrographic texture. The micrographic texture is not readily apparent in hand sample, especially in finer-grained samples, but is very distinctive in thin section (Fig. 6).

The contact between the granophyre and the overlying gabbro is exposed in the sides of the valley to the west, but the slopes are extremely steep and treacherous so we will not be visiting the outcrop. The contact is sharp and the granophyre has a distinct chilled margin against the medium-grained gabbro, indicating the granophyre intruded into the gabbro after the gabbro had cooled.

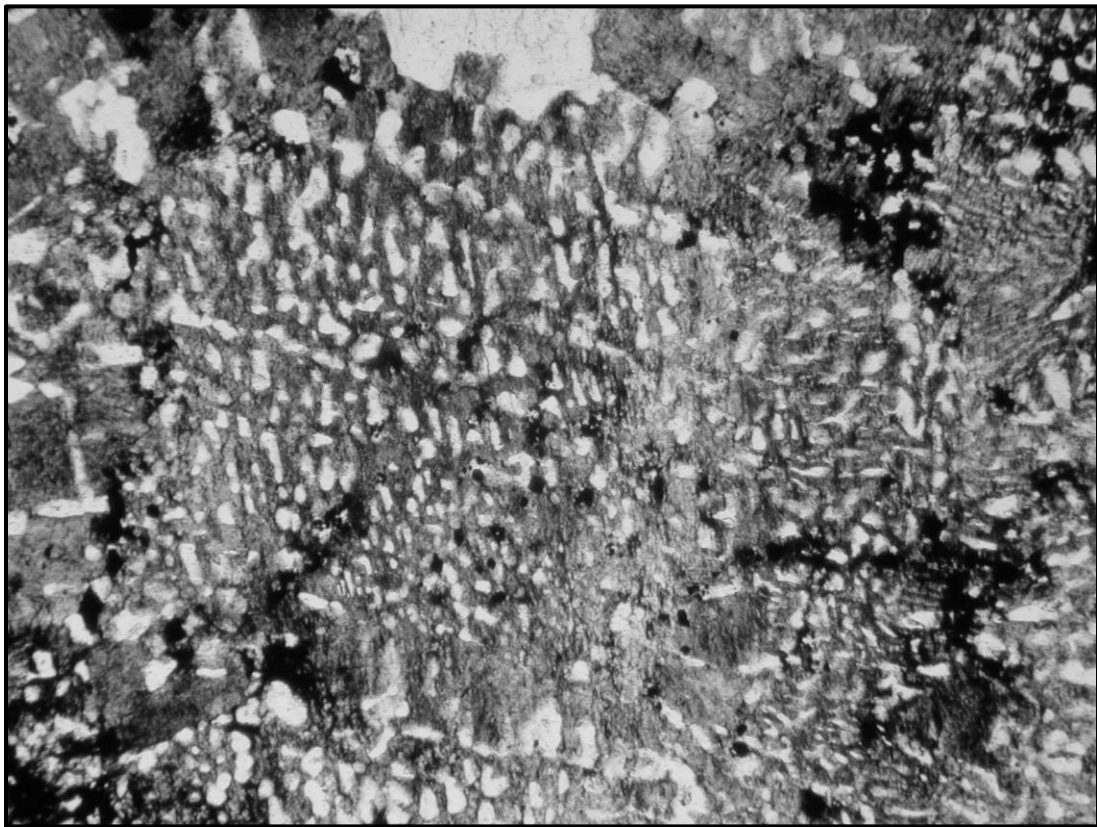


Figure 6. Photomicrograph of granophyre from Morgan Falls at Stop 1. The image was taken with plane transmitted light, and is shown here in black-and-white. The rock is composed primarily of alkali feldspar (dark grey) and quartz (light grey) intergrown in micrographic texture. The field of view is 2 mm.

Stop 2: Anorthositic Gabbro of the Mineral Lake Intrusion

Location: South side of road FS-189; T45N, R4W, Sec. 30, west-central; 0666203 mE, 5133192 mN. There are several large outcrops in the woods immediately south of the road.

The rock at this location is coarse-grained anorthositic gabbro with a strong parallel alignment of plagioclase feldspar defining a lamination that dips 70° northwest. This rock is very typical of the central zone of the MLI, and is the most common rock type in the intrusion overall. The rock is composed of 80–85% plagioclase, the rest being intercumulusferromagnesian silicate minerals, now biotite and hornblende. The southern part of the outcrop has some small-scale modal layering.

Stop 3: Olivine Melagabbro and Peridotite of the Rearing Pond Intrusion

Location: T45N, R4W, Sec. 26, SW1/4; start of hike: 066638 mE, 5133660 mN; hilltop destination: 666713 mE, 5134266 mN. This stop involves a 0.4-mile bushwhack north from FS-189 to a hilltop with exposures of olivine melagabbro and peridotite. From the starting location hike $N10^{\circ}W$ to the ridge, then $N65^{\circ}E$ to the end of the ridge where there are several large outcrops of olivine melagabbro. There is a concrete foundation and detritus from a work camp on the highest point.

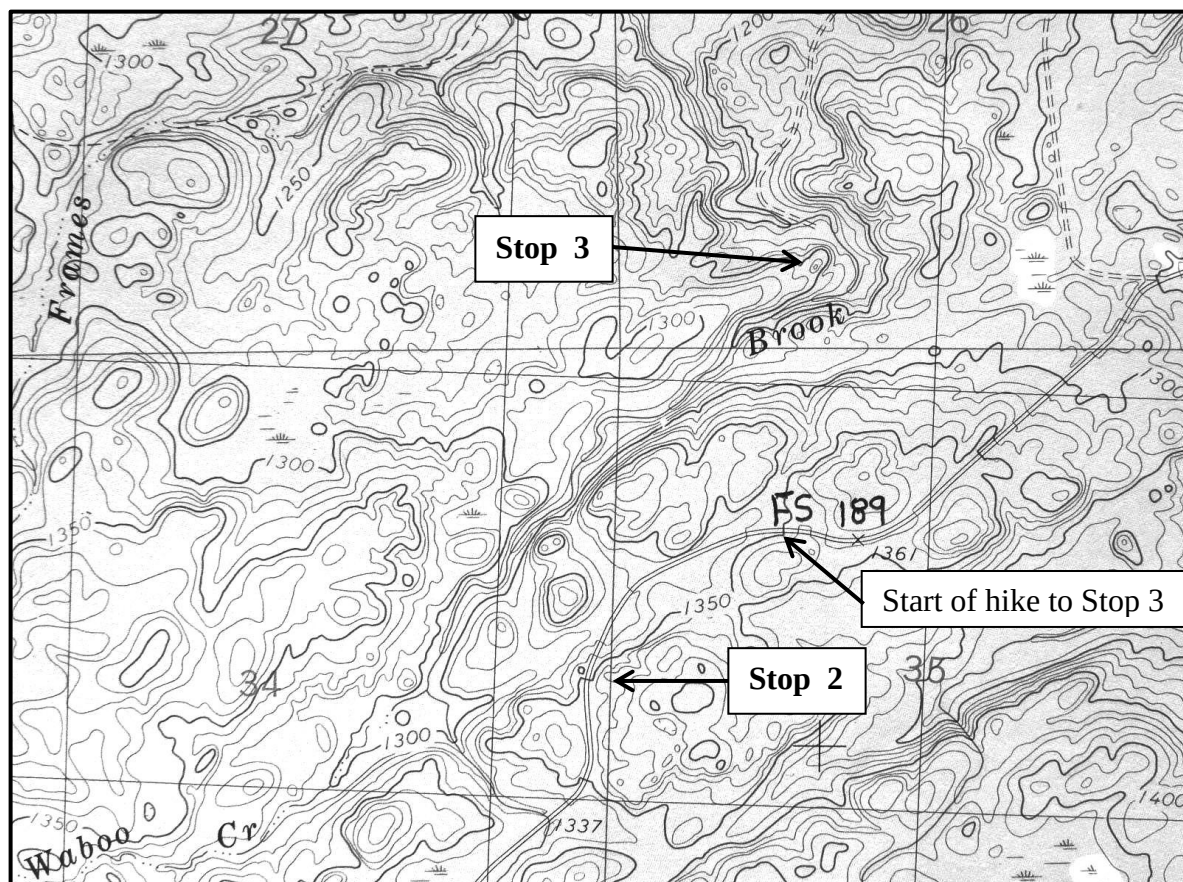


Figure 7. Topographic map of the area of Stops 2 and 3.

The rock exposed on the hilltop is olivine melagabbro composed of approximately 50% plagioclase feldspar, 25% olivine, 20% augite, and 5% oxide mineral. There is no lamination, which is typical of the RPI.

Hike northwest, downhill a short distance to outcrops of peridotite. The rock is black, medium-grained, olivine-rich, partially serpentinized peridotite. The outcrop extends down the steep hill to the valley floor – hike carefully down this slope and then return to FR-189.

There are few outcrops of the peridotite, but Klewin et al., 1987 describe another outcrop of this unit about 1 km northeast of this stop. Perhaps the best exposures are located in the ledges above the Brunswailer River in T45N, R4W, Sec. 23 SW1/4; 0667514 mE, 5135895 mN. We will not be visiting these outcrops on this trip because of their remote location. At that location the rock is slightly coarser grained and less serpentinized. It is composed of approximately 50% olivine, 15% orthopyroxene, 10% opaque oxide mineral, and 25% serpentine (Fig 8). The large orthopyroxene crystals enclose euhedral olivine crystals. Some of the serpentine contains thin veins of parallel-aligned acicular crystals oriented perpendicular to the margins of the vein, which is typical of serpentinites.

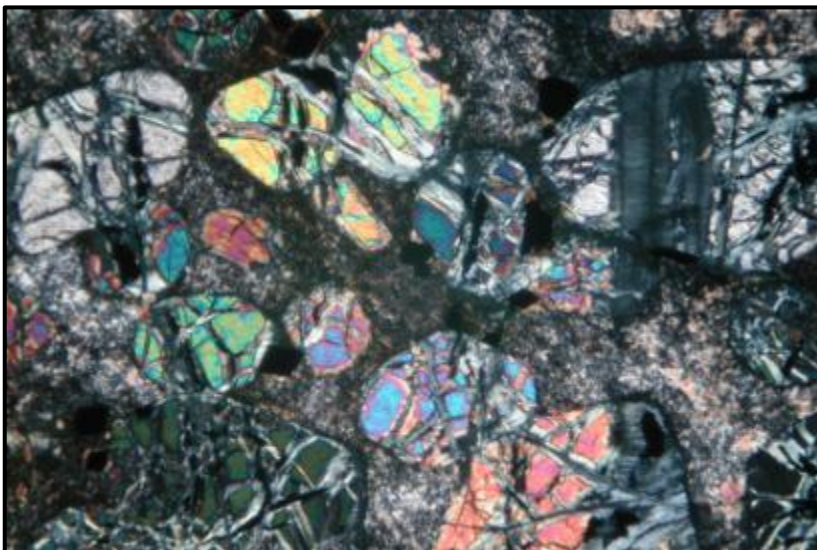
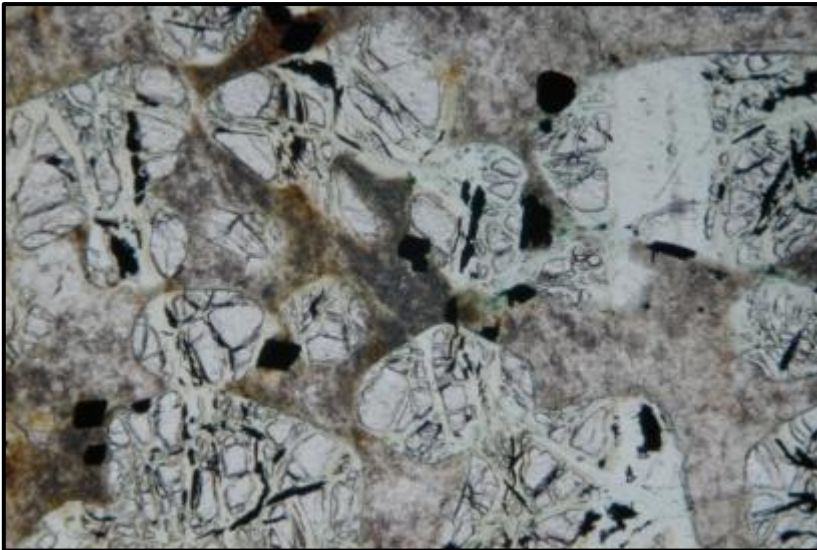


Figure 8.
Photomicrographs
of serpentinized
peridotite of the
Rearing Pond
Intrusion. The
upper image was
taken with plane
light, the lower
image is the same
view, but with
crossed polars. The
rock is composed
primarily of olivine
(highly
birefringent),
magnetite (opaque),
and serpentine
(fine-grained, low
birefringence).
Field of view is 2
mm. This rock is
not from the
outcrop at Stop 3,
but is from an
outcrop of the same
unit at 0667514
mE, 5135895 mN.

Stop 4: Ferrodiorite in the Upper Part of the Mineral Lake Intrusion

Location: T45N, R4W, Sec 25, north-central, 0668365 mE; 5134955 mN

The rock at this stop is within the thin ferrodiorite unit at the top of the MLI. The dark-colored ferrodiorite somewhat resembles gabbro, but its mineral composition is distinctly different than that of the underlying gabbroic rocks. The rock is composed of plagioclase, hornblende, magnetite, and minor amounts of alkali feldspar and quartz. The proportion of alkali feldspar and quartz increases to the north, toward the top of the unit.

Stop 5: Olivine Gabbro of the Mineral Lake Intrusion

Location: T44N, R4W, Sec. 13, NW1/4; Park along the road near the entrance to the Mineral Lake boat ramp (0668420 mE, 5129159 mN). This stop includes one outcrop a short distance into the woods on the west side of Mineral Lake Road (Stop 4A), and several outcrops on along the ridge on the east side of the road (Stop 4B) (Fig. 9).

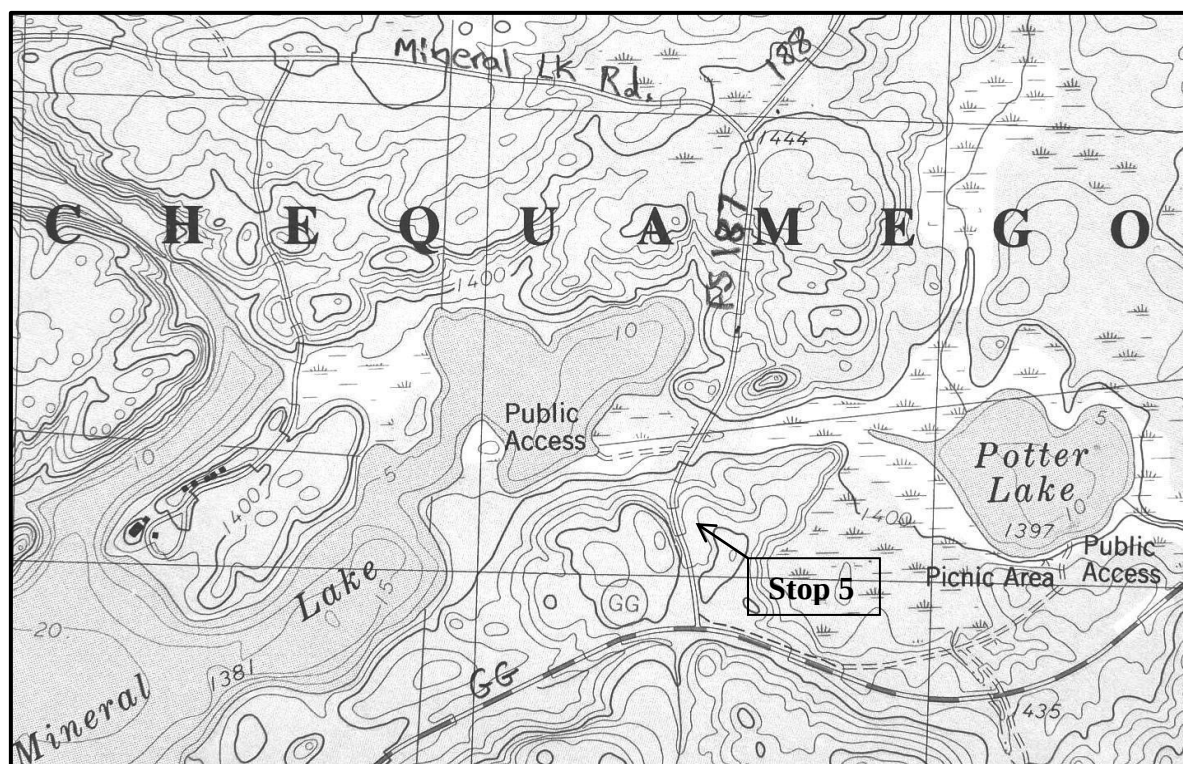


Figure 9. Topographic map showing the area of Stop 5.

Stop 5A: Coarse-Grained Gabbro

Location: T44N, R4W, Sec. 13, NW 1/4; 0668392 mE, 5129146 mN. This outcrop is located in the woods on west side of Mineral Lake Road.

The rock is coarse-grained gabbro composed of approximately 80% plagioclase feldspar, 20% intercumulus pyroxene that is now mostly actinolite, 5% oxide minerals, some of which is magnetic, some of which is not. It does not show laminations or compositional layering.

Stop 5B: Troctolite and olivine melagabbro

Location: 0668468mE, 5129193 mN.

The outcrops of this stop are located in the woods a short distance east of Mineral Lake Road. The low ridge trending northeast toward Potter Lake has numerous exposures of the olivine gabbro, troctolite, and melatroctolite of the basal zone of the MLI. The rock varies in grain size and proportion of olivine, with some of the finer-grained rocks containing up to 50% olivine. There is a trace of sulfide mineral present. The map of Cannon et al, 1996 shows a Cu-Ni zone here, and Bakheit (1981) studied the petrography of Cu-Ni mineralization in this area.

It is likely the rockshere are compositionally layered, although it is not apparent in individual exposures because of the relatively small size of most outcrops. These rocks are within 200 m of the base of the MLI.

The basal contact zone is not exposed in the area of Mineral Lake, but it is exposed in one place approximately 2.5 km east of here near English Lake. This field trip will not be visiting that basal contact exposure because of difficulties with access, but the site is described in the Klewin et al., 1989ILSG guidebook chapter (their Stop 7). The exposed contact has a chill zone where fine-grained gabbro of the MLI is in contact with contact-metamorphosed rocks of the Tyler Formation.

Stop 6: Metamorphosed Iron Formation Below the Mineral Lake Intrusion

Location: T44N, R4W, Sec 14, NE 1/4; 0667764 mE, 5128518 mN. This outcrop is located in the woods just south of County Road GG. Use caution while parking along the road because the shoulder is narrow; there a small pull-off on the shoulder on the north side of the road not far west of the outcrop that has adequate room for one car.

This outcrop is in the upper part of the Ironwood Iron Formation where iron formation has been contact metamorphosed by the MLI. Although the contact is not exposed, olivine gabbro of the MLI is not far to the north. The rock here is black, dense, metamorphosed siliceous iron formation. It is composed of approximately 50% dark green to black olivine, 30% black amphibole, and 20% quartz. This is a relatively rare lithology – rocks containing both quartz and olivine are not common and are restricted to just a few geologic settings, one of which is where siliceous iron formation has undergone high-temperature metamorphism. The situation here, where the Ironwood Iron Formation was contact metamorphosed by the MLI, hadone of these special circumstances in which olivine crystals could grow in a silicic metamorphic rock. The fact that the olivine is in an iron-rich rock and occurs with quartz – which is not stable with Mg-rich olivine – suggests it is iron-rich (fayalitic) olivine.

Stop 7: Quarry in Coarse-Grained Anorthositic Gabbro of the Mineral Lake Intrusion

Location: T44N, R3W, Sec. 4, NW 1/4; 0672101 mE, 5132430 mN.

This abandoned quarry is in very coarse-grained anorthositic gabbro. The rock is composed of plagioclase feldspar, hornblende, some augite, and non-magnetic oxide minerals. The quarry dump contains large blocks of freshly broken gabbro displaying a wide range of textures, including some pegmatitic gabbro. Hornblende is more common than augite as the ferromagnesian mineral in these rocks, which is common in upper parts of the MLI.

Stop 8: Feeder Dike at Mellen

Location: T44N, R2W, Sec. 6, NE 1/4; 0680368 mE, 5132423 mN; near the intersection of Fayette Avenue and Highway 77 in Mellen, Wisconsin.

This outcrop is fine-grained olivine gabbro within a dike that cuts across the Tyler Formation. It lies immediately below rocks of the PRI and is likely a feeder dike approximately 100 – 200 m wide that carried magma upward into the base of the PRI. This rock is considerably finer grained than most of the gabbroic rocks of the MIC because the relatively small size of this intrusion caused it to cool quite rapidly. The dike creates a resistant ridge that trends south from here, down-section, into the area of less-resistant slate of the Tyler Formation.

Stop 9: Anorthositic Gabbro of the Potato River Intrusion

Location: East side of Wisconsin Highway 13, 0.5 miles north of Mellen, Wisconsin; T45N, R2W, Sec. 31, NE ¼; 0680249 mE, 5133847 mN.

This is a large highway road cut outcrop of anorthositic gabbro in the western part of the PRI. It is composed of plagioclase feldspar with a strong parallel alignment that dips north, and minor amounts of augite and magnetite. At the south end of the outcrop there is a 5-cm-wide basalt dike that dips south, cutting across the lamination. Further north along the outcrop, the rock surface is not freshly broken gabbro, but instead is joint and minor fault surfaces covered with epidote and calcite. In a few areas where freshly broken gabbro is exposed, there are tabular zones approximately 2 cm wide of very fine-grained light-colored rock interpreted to be gabbroic mylonite created along minor faults. The next outcrop to the north along the highway is similar in that its surface is mostly fine-grained epidote. The area apparently has numerous small-scale faults, along which the rock broke when it was blasted during highway construction, so what is exposed on the surface is mostly epidote.

Stop 10: Intrusive Breccia of the Mellen Granite – South Outcrop

Location: T45N, R2W, Sec. 30, SW ¼; 0679689 mE, 5134810 mN. This is a large highway road cut on the west side of Wisconsin State Highway 13, 1.3 miles north of Mellen. The rocks on the upper slopes of this outcrop are unstable so climbing would be hazardous, and if knocked out of place, the blocks would pose a significant hazard to anyone below. Thus, do not climb the outcrop, examine it from ground level.

This outcrop is in the contact zone of the MG; the main body of the MG lies to the west, and the PRI lies to the east. Grey granite of the MG has intruded and brecciated mafic and intermediate rocks of the MIC, creating an intrusive breccia that is volumetrically about half inclusions and half granite. The granite is grey, medium-grained hypidiomorphic granite typical of the MG. The xenoliths are composed of anorthositic gabbro, biotite-rich gabbro, ferrodiorite, minor troctolite, and probably others as well. The ferrodiorite inclusions tend to be finer grained than other inclusions. The biotite-rich gabbro, informally called the “biotite cluster gabbro”, tends to be the coarsest grained of the inclusions. Cervin’s (2009) study of this outcrop and that of Stop 11 included whole-rock chemical analyses of the MG and five different xenolithic rocks from the breccia zone (Table 1).

The dark-colored mafic- and intermediate-rock xenoliths have been broken into a complex array of angular fragments, some of which are not completely fragmented but have thin intrusions of granite that nearly penetrate the xenolith (see photograph on the cover page

of this chapter). In some xenoliths, the granite is finer grained within the fractures than in the surrounding granite. This is interpreted to be the result of filter pressing of the magma – liquid was forced into the cracks but larger feldspar crystals in the crystal mush could not travel into the small nascent fractures.

No rocks other than MIC lithologies are present as inclusions, so they are probably all locally derived. However, the inclusions constitute a wide variety of rock types. Some of this heterogeneity may be due to intrusion of the granite into a zone of the PRI that had a complex mixture of rock types in one area, but at least some of the compositional heterogeneity is likely the result of entraining and mixing of inclusions from different sources within the MIC.

Table 1. Whole-rock chemical analyses of Mellen Granite and selected xenoliths in the intrusive breccias.

Sample	8-10-13	8-10-11	8-10-16	8-10-1	8-10-14	8-10-15	Units	Detection
Rock	Mellen Granite	Ferro- Diorite	Anortho Gabbro	Gabbro	Hbl-rich Fe-Dior	Fn-grain Troctolite		Limit
Location	SB	NB	SB	NB	SB	SB		
Analyte								
SiO ₂	72.22	55.82	51.24	48.07	46.95	43.44	%	0.01
Al ₂ O ₃	13.45	12.74	22.81	20.72	11.14	12.11	%	0.01
Fe ₂ O ₃ (T)	1.9	13.09	6.71	9.48	20.35	20.08	%	0.01
MnO	0.02	0.174	0.076	0.11	0.236	0.229	%	0.001
MgO	0.25	3.01	1.97	6.61	4.83	18.08	%	0.01
CaO	1.13	5.67	10.29	9.53	9.1	5.57	%	0.01
Na ₂ O	3.47	3.29	4.3	2.76	3	1.64	%	0.01
K ₂ O	4.9	2.02	0.62	0.62	0.75	0.21	%	0.01
TiO ₂	0.161	2.461	1.28	0.682	3.439	0.16	%	0.001
P ₂ O ₅	0.03	0.47	0.25	0.06	0.54	0.03	%	0.01
LOI	0.52	0.7	0.39	0.94	-0.01	-0.99	%	
Total	98.04	99.44	99.94	99.58	100.3	100.6	%	0.01
Ba	2449	594	197	123	211	62	ppm	2
Sr	102	171	395	376	176	199	ppm	2
Y	77	72	20	6	61	2	ppm	1
Sc	1	25	10	9	35	6	ppm	1
Zr	626	518	117	36	321	35	ppm	2
Be	4	3	1	< 1	3	< 1	ppm	1
V	11	266	213	93	441	31	ppm	5

Locations: SB=South Breccia outcrop; NB=North Breccia outcrop
Method: FUS-ICP
Actlabs Report
Date: 14/04/2009
Also reported in: Cervin, 2009

Stop 11: Intrusive Breccia of the Mellen Granite – North Outcrop

Location: T45N, R2W, Sec. 31, SW ¼; 0679517 mE, 5135384 mN. This is a low road-cut outcrop on the west side of Highway 13, 1.8 miles north of the town of Mellen.

The rock at this outcrop is similar to that of Stop 9 – intrusive breccia with granite enclosing angular inclusions of a variety of mafic and intermediate lithologies of the MIC. The granite is different here though – it is pink granite with a wide variety of textures. Some is granite with hypidiomorphic granular texture typical of the MG, but this grades into zones of graphic granite that in turn grade inward to pegmatitic zones. Some of the pegmatitic zones have concentric zoning around central miarolitic cavities that contain euhedral quartz crystals, others are filled with calcite. There is also zoning of the granite outward from inclusions (Fig. 10), suggesting the inclusions influenced the crystallization of the granite. The textures of this granite indicate complex crystallization processes in the granite of the breccia zone. Sikkila (2002) concluded there was significant potassium metasomatism here, which altered plagioclase to potassium feldspar and caused alteration of ferromagnesian minerals in both the granite and the xenoliths. The complex variation in textures and the potassium metasomatism indicate that water was important in petrologic processes during and after crystallization of this rock.



Figure 10. Intrusive breccia at “the north breccia outcrop” of Stop 11. The rock has dark-colored, angular inclusions of ferrodiorite enclosed in pegmatitic granite. Quarter for scale.

Goscina's (2010) detailed mapping of this outcrop revealed that the lithology of the inclusions defines three zones within the outcrop (Fig. 11). Ferrodiorite is the dominant rock type of inclusions in the central section, whereas “biotite-cluster gabbro” occurs at the north and south ends of the outcrop.

If, as the evidence suggests, the intrusive breccia was created by forceful injection of granite at a shallow level, then perhaps the magma exploded to the surface as a pyroclastic eruption. This hypothesis is consistent with the presence of large rhyolite units in the Kallander Creek Volcanics, so perhaps there is a connection between the intrusive breccias and the older rhyolites.

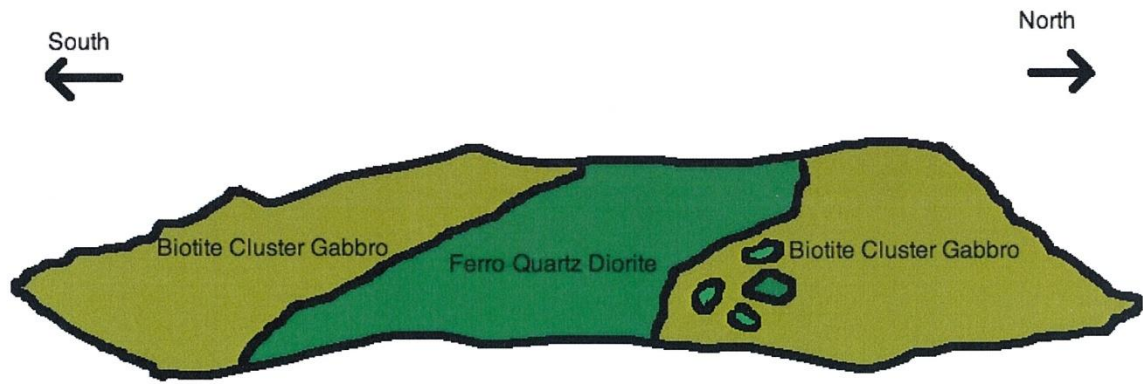


Figure 11. Zones of xenolith composition in the “north breccia outcrop” of Stop 10. From Goscinak, 2010.

Stop 12: Fault in Mellen Granite

Location: T45N, R3W, Sec. 27, west-central; 0674116 mE, 5134740 mN.

This quarry is owned and operated by Milestone Materials Corporation. Permission must be obtained from Milestone Materials before venturing into the quarry. This is an active quarry with significant hazards such as loose rock on quarry walls and in rock piles. Use great caution while in the quarry; do not climb on any rock faces, and do not even approach the tall faces on the south side of the quarry.

Piles of freshly broken rock in the quarry allow detailed examination of the granite, which is typical pink granite of the MG. Inclusions of ferrodiorite are common, some of which have gradational contacts with the granite.

The outstanding feature at this stop is the large fault surface on the east side of the quarry. The fault surface is green due to a neomineral coating of epidote on and within the fault zone. Slickenlines and fault mullions on the surface clearly indicate strike-slip motion. The fault is not a single surface but instead is a zone of nearly parallel surfaces that enclose lens-shaped bodies of largely undeformed granite. The fault zones are 1–2 cm wide and contain angular fragments of granite enclosed within the fine-grained epidote-rich cataclasite. The potassium feldspar in the granite along the fault is slightly darker red, evidently due to metasomatism. The timing of faulting and the amount of displacement on the fault are unknown, but since the rock is identical on both sides of the fault, the offset is not on the order of kilometers.

Ferrodiorite is exposed along both sides of the access road into the quarry. There is an especially good outcrop of the ferrodiorite on the west side of the road: 0674139 mE, 5134942 mN. The contact between the ferrodiorite and the granite is not exposed here, but based on their relationships elsewhere, the granite is probably intrusive into the ferrodiorite.

ACKNOWLEDGEMENTS

Dan Cervin and Chris Goscinak studied the “north breccia outcrop” of Stop 11 for their senior capstone research projects at Northland College. Their companionship during our exploration of that outcrop and many others within the MIC is greatly appreciated. Their

work has both answered and raised many questions about the origin of the intrusive breccias. I am grateful to Bill Cannon for his suggestions for field trip stops and for his time in the field showing me some especially interesting rocks in and around the MIC.

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Northland College Van #1 on the verge of meeting its deserved demise at a “black granite” quarry near Mellen, Wisconsin.

