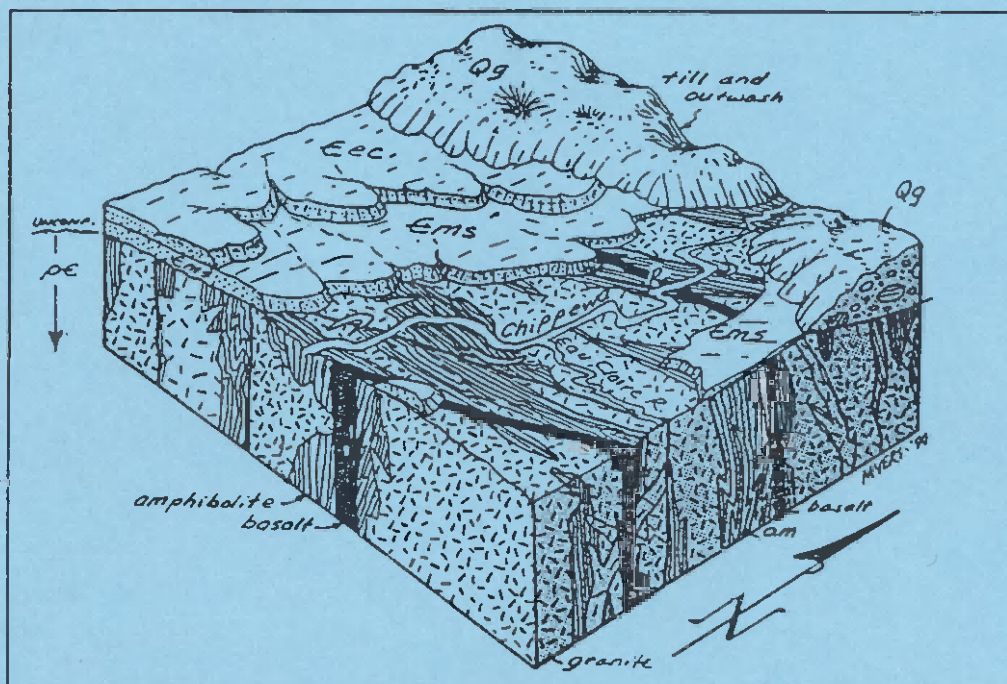




THIRTY-SEVENTH ANNUAL
INSTITUTE ON LAKE SUPERIOR GEOLOGY



PROCEEDINGS: PART 2 - FIELD TRIPS

MAY 1-4, 1991
EAU CLAIRE, WISCONSIN

Organization Committee, 37th Annual Meeting ILSG, 1991

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PROCEEDINGS VOLUME 37

PART I: ABSTRACTS (Includes Meeting Program)

PART 2: FIELD TRIPS

Reference to material in this volume should follow the example below.

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Copies of the 37th ILSG Proceedings (Part I, Abstracts and Part II, Field Trips) may be purchased during the meeting for \$5.00 (US) each.

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Issues of Proceedings and Abstracts, Part I, and Field Guidebook, Part 2, from this and previous meetings may be ordered from:

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37th Annual

Institute on Lake Superior Geology

Proceedings

Ray Wachs Civic Center

Eau Claire, Wisconsin

May 1-4, 1991

Organized by

P.E. Myers, University of Wisconsin-Eau Claire

L.S. Chan, University of Wisconsin-Eau Claire

Volume 37

Part 2 Field Trips

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CALENDAR OF EVENTS

TUESDAY, APRIL 30

- 3:00-5:00 p.m. Early Registration for Field Trips #1 and #2 Participants - Civic Center Inn (CCI) Lobby
- FIELD TRIP #1 P.K. Sims, and J.S. Klasner (Leaders) "The Mountain shear zone - a post-Penokean discrete ductile deformation zone"
5:00 p.m. Two vans depart for Antigo (Park Inn)
8:30 p.m., Park Inn, Pre-Field Trip Conference

WEDNESDAY, MAY 1

- FIELD TRIP #1 8:00 a.m. Vans depart from Park Inn, Antigo, for the mountain area. Vans return to CCI by about 9:00 p.m.
- FIELD TRIP #2 L.S. Chan, P.E. Myers, and R.L. Hay, (Leaders): "Features and significance of the Precambrian-Cambrian contact in western Wisconsin"
8:00 a.m. Pre-trip Seminar CCI, Farwell Rm.
8:20 a.m. Vans depart from CCI Lobby
- 2:00-5:00 p.m. Pre-Conference Planning Session - Staff, CCI Gibson Room
- 4:00-8:00 p.m. Registration and check-in for pre-registrants, Civic Center Inn Lobby
- 8:00-11:00 p.m. Informal Get-together and Cash Bar
Music by: Bill Jordan Jazz Quintet

THURSDAY, MAY 2

- 7:15 a.m. Registration - Lobby, Civic Center Inn (CCI)

TECHNICAL SESSION

All Technical Sessions are in Room C, Ray Wachs Civic Center (RWCC); Poster Sessions in RWCC Rooms D and E. Schedules of oral papers are shown on pages xiii and xiv. Contributors should have posters up by 9:00 a.m.

- 8:00 a.m. Introduction and Welcome: Paul Myers, Conference Chairman, and Larry Schnack, Chancellor, University of Wisconsin - Eau Claire
- 8:20-10:00 a.m. Morning Session #1. Chairs: R.W. Ojakangas and A.B. Dickas
- 10:00 a.m. COFFEE BREAK
- 10:20-11:50 a.m. Morning Session #2. Chairs: G.L. LaBerge and K. Schultz

- NOON-2:00 p.m. BOARD LUNCHEON AND ANNUAL MEETING: *Farwell Room, CCI*
- 12:30-1:30 p.m. POSTER SESSION #1 - Authors present (Poster papers, p. xv)
- 1:40-3:00 p.m. Afternoon Session #1. *Chairs: V.W. Chandler and G.B. Morey*
- 3:00 p.m. COFFEE BREAK -- *Remove Poster Papers*
- 3:20-4:50 p.m. Afternoon Session #2. *Chairs: M.M. Kehlenbeck and C. Craddock*
- 6:00-7:30 p.m. CASH BAR: RWCC ROOM A
- 7:30-10:00 p.m. ANNUAL ILSG BANQUET -- RWCC Rooms A and B
- Presentation of the Goldich Medal to W.J. Hinze, Purdue University
by V.W. Chandler
- Speaker: Paul K. Sims, "Archean and Early Proterozoic Geology of the
Lake Superior Region -- An Overview"; introduced by G.B. Morey.

FRIDAY, MAY 3, 1991

- 8:00 a.m.-3:30 p.m. Registration and Sale of Extra Proceedings Volumes
- 8:20 a.m. Morning Session #1. *Chairs: M. G. Mudrey, Jr. and T. A. DeMatties*
- 10:00 a.m. COFFEE BREAK
- 10:20-11:30 a.m. Morning Session #2. *Chairs: G.W. Adams and B.A. Brown*
- 12:30-1:30 p.m. POSTER SESSION #2 -- Authors present (poster papers, p. xvi)
- 1:40-2:40 p.m. Afternoon Session #1. *Chairs: W.S. Cordua and M. Jirsa*
- 2:40 p.m. Student Award Presentation, John Klasner
- 3:00 p.m. COFFEE BREAK -- *Remove Poster Papers*
- 3:20-4:30 p.m. Afternoon Session #2. *Chairs: J. Kaliokoski and J. C. Green*
- 7:00-8:30 p.m. Pre-Trip #3 Seminar, CCI, Farwell Room, M.G. Mudrey, Jr. and B.A. Brown

SATURDAY, MAY 4, 1991

- FIELD TRIP #3 M.G. Mudrey, Jr., and B.A. Brown (Leaders), "Proterozoic
volcanogenic massive sulfide deposits of northwestern Wisconsin"
8:00 a.m. Bus departs from front entrance of CCI

WEDNESDAY, MAY 1, 1991

FIELD TRIP #1

P.K. Sims, and J.S. Klasner (Leaders)

**"The Mountain shear zone - a post-Penokean
discrete ductile deformation zone"**

MOUNTAIN SHEAR ZONE, OCONTO COUNTY, WISCONSIN--
A POST-PENOKEAN DISCRETE DUCTILE DEFORMATION ZONE

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Field trip, 37th Annual Meeting of Institute on
Lake Superior Geology, Eau Claire, Wisconsin

May 1, 1991

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INTRODUCTION

Reconnaissance mapping in 1980 by Sims and M.G. Mudrey, Jr. revealed the existence of a belt of remarkably well exposed mylonitic rocks near Mountain in Oconto County, Wisconsin. Subsequently, the mylonitic rocks were systematically mapped to determine their relation to the regional deformation (Sims, 1989), and Klasner made detailed structural studies of critical outcrops to determine the sequence of deformational events and the sense and direction of movement within the shear zone (Sims and others, 1990).

The Mountain shear zone is a discrete ductile deformation zone as much as 1.5 km wide that post-dates Early Proterozoic regional deformation of the Penokean orogeny. A body of quartz diorite, formally named the Hines Quartz Diorite, that was intruded into the shear zone at about the close of the shearing event has a U-Pb zircon concordia intercept age of $1,812.7 \pm 3.6$ Ma (Sims and others, 1990), which places a firm minimum age on the time of shearing. The Mountain shear zone and other N. 55° - 60° E. shear zones in the Wisconsin magmatic terranes record a significant episode of ductile deformation that probably resulted from a collision of unknown rocks to the south of the area after the Penokean orogeny.

PENOKEAN OROGEN

The 1,990-1,835 Ma Penokean orogen developed as an embayment in the southern margin of the Archean Superior craton. Rocks of the orogen are exposed in Wisconsin, northern Michigan, and Minnesota (Cannon, 1973; Morey, 1973; Sims and Peterman, 1983).

The orogen consists of two distinct lithologic domains (fig. 1): a northern deformed continental margin assemblage overlying an Archean basement; and a southern assemblage of island arcs, the Wisconsin magmatic terranes (Sims and others, 1989). The northern rocks include the Marquette Range Supergroup in Michigan and Wisconsin (Cannon and Gair, 1970), which correlates with the Animikie and Mille Lacs Groups in Minnesota (Morey, 1983), and comprise a south-facing continental-margin assemblage. They consist of a lower rifted passive-margin sequence overstepped northward by a synorogenic foredeep sequence (Barovich and others, 1989; Southwick and others, 1988). Early Proterozoic regional deformation involved northward-directed thrusting and related folding of the supracrustal rocks and produced metamorphism and basement gneiss domes caused by crustal thickening (Sims and Peterman, 1983; Holm and others, 1988; Attoh and Klasner, 1989). The southern magmatic terranes consist of Early Proterozoic calc-alkaline and tholeiitic volcanic and calc-alkaline plutonic rocks (Sims and others, 1989).

The Niagara fault zone separates the south-facing continental margin assemblage and the oceanic arc terranes in Wisconsin and Michigan (fig. 1). It is a broadly arcuate, convex-northward system of ductile shears as much as 10 km wide that contains flattened steeply dipping rocks having dominantly down-dip stretch lineations (Larue and Ueng, 1985; Sedlock and Larue, 1985). It contains a dismembered ophiolite and has been interpreted as a paleosuture zone (Schulz, 1987) between the two terranes, which were juxtaposed at about 1,860 Ma (Sims and others, 1985).

The oceanic arc rocks and related intrusive rocks south of the Niagara fault are part of the Pembine-Wausau terrane of the Wisconsin magmatic terranes (Sims and others, 1989). The metavolcanic rocks are 1,860 to 1,889 Ma. The granitoid rocks are mainly granodiorite and tonalite; they range in age from 1,760 to 1,870 Ma, with the younger group being post-tectonic. Sims and others (1989) have proposed that the Penokean orogeny terminated between 1,835 Ma and 1,840

Ma, when the Marshfield terrane (see fig. 1) was sutured to the Pembine-Wausau terrane along the Eau Claire shear zone.

ROCK UNITS, MOUNTAIN AREA

The Mountain shear zone is in the eastern part of the Pembine-Wausau terrane. The major rock units exposed within and adjacent to the Mountain shear zone are metavolcanic rocks and minor related volcanogenic sedimentary rocks, assigned to the Waupee Volcanics, granodiorite and its mylonitized equivalent in the shear zone--granite gneiss (unit Xgn, fig. 2)--and the Hines Quartz Diorite. The sheared metavolcanic rocks (units Xwi and Xwa, fig. 2) and the granite gneiss are overlain along the northern margin of the shear zone by a metaconglomerate, named the Baldwin Conglomerate. These rocks are Early Proterozoic in age. They are intruded by and occur as a roof pendant in anorogenic rocks of the Middle Proterozoic (~1.47 Ga) Wolf River batholith.

The stratigraphic nomenclature used in this report was adopted earlier (Sims, 1989), and is based largely on previously proposed geologic names. Mancuso (1957) assigned the name Waupee Volcanics to the metavolcanic rocks exposed in this and adjacent areas to the east because of typical exposures near Waupee Creek, and this formal designation is retained herein. Lahr (1972) renamed this unit the Waupee Formation, but the earlier nomenclature appropriately emphasizes the volcanic affinity of the rocks. Mancuso (1957) also proposed the name Baldwin Conglomerate for the metaconglomerate exposed in the vicinity of Baldwin Creek, which is accepted herein. Lahr (1972) assigned the undeformed quartz diorite exposed in section 12, T. 31 N., R. 16 E., east of the village of Mountain (see Sims, 1989), to the Hines Quartz Diorite because of its exposures near Hines Creek. Previously, Mancuso (1957) had included these rocks in his "Macauley Granite." Although Lahr (1972) correctly renamed the granitoid rocks "Macauley Gneiss," neither of these terms are used herein because they each include substantial amounts of metavolcanic rocks as well as granitoid rocks.

Because the rocks within the Mountain shear zone differ profoundly in structure and to some extent in mineralogy and degree of metamorphism from those outside of it, the rocks within and outside of the shear zone are described separately below. Slightly to moderately deformed rocks (D_1) of the Waupee Volcanics and the intrusive granodiorite were converted to mylonitic schists and gneisses within the shear zone (D_2). The undeformed Hines Quartz Diorite was emplaced within the shear zone at or near the end of the ductile deformation.

Outside of Shear Zone

Waupee Volcanics

Within the map area (fig. 2), the Waupee Volcanics is divided into four mappable units. Outside of the shear zone, it consists of a unit of mafic lava flows (unit Xwm, fig. 2) and a unit of predominantly felsic volcanic rocks (unit Xwf). Sheared correlative rocks within the shear zone are: (1) amphibole schists (unit Xwa) and (2) interlayered quartz-biotite and amphibole schist (unit Xwi), respectively. A body of quartz diorite schist (unit Xwq), presumably a subvolcanic intrusion, occurs within the shear zone. Chemical compositions of two representative rocks are listed in table 1.

Mafic lava flows

This rock unit (Xwm, fig. 2) is dominantly exposed southeast of the shear zone, and is composed mainly of dark-gray, fine-grained porphyritic lava flows that are highly altered. The rocks are generally massive and unfoliated to weakly foliated but locally have a distinct compositional layering visible on weathered surfaces. Possible relicts of pillow rinds can be seen at places. The lavas are metamorphosed to greenschist facies.

In the least altered rocks, plagioclase megacrysts occur as stubby laths with broad albite twinning and as aggregates of randomly oriented stubby laths. The plagioclase is altered variably to epidote/clinozoisite, sericite, and calcite. The mafic minerals are completely altered to pale-green actinolite and, locally, to sphene, which shows incipient crystallization, and skeletal opaque oxides. Quartz generally is absent, but in two sections occurs as anhedral, strained, sutured grains. Biotite is dark brown.

Felsic volcanic rocks and minor interbedded mafic volcanic rocks

This rock unit (Xwf, fig. 2) is mainly exposed in sections 4 and 5, T. 31 N., R. 17 E. (see Sims, 1989). It consists of massive to layered metamorphosed felsic tuff, intercalated metasedimentary rocks, and generally thin interbedded layers of mafic tuff. The rocks have a penetrative cleavage (S_1) and, at least locally, are folded (F_1).

The felsic tuff is a light-gray, fine-grained, inequigranular, xenomorphic granular rock of general dacitic composition. It is cut by widespread narrow shears but lacks mylonitic textures.

Plagioclase occurs mainly as anhedral grains generally less than 0.3 mm in diameter, but locally forms phenocrysts as much as 1.5 mm in diameter. It is poorly twinned. Some grains contain abundant rounded quartz inclusions. Quartz grains are anhedral, clear, and approximately the same size as the plagioclase grains. Stubby brown to pale-red biotite is preferentially orientated in the (S_1) foliation and is intergrown with about equal amounts of muscovite. The biotite is locally altered to chlorite. Sparse, poorly crystallized sphene, opaque oxides, allanite, and zircon are accessory minerals.

The sedimentary rocks are medium- to dark-gray, fine-grained, thinly layered, foliated rocks composed of plagioclase, quartz, and biotite and, locally, either garnet or cummingtonite. Good exposures can be seen in the center of section 5, T. 31 N., R. 17 E., on the ridges just north of Bear Paw road, west of Waupee Creek (stop 5, fig. 8).

Granodiorite

Porphyritic granitoid rocks (unit Xg, fig. 2) that have an average composition of granodiorite (table 2; fig. 3) intrude the metavolcanic rocks on the southeast side of the Mountain shear zone. They are most extensively exposed in section 7, T. 31 N., R. 17 E., but extend into sec. 5 (see Sims, 1989). The granodiorite intrudes the surrounding metavolcanic rocks lit-par-lit, to form migmatite, and contains moderate to abundant elliptical inclusions of them (mainly amphibolite) that are consistently oriented N. 10°-30° E. The granodiorite extends northward into the shear zone, where it has been deformed into a gneiss, termed granite gneiss (unit Xgn, fig. 2). The chemical composition of one sample of granodiorite (number 35-85) is given in table 1.

The granodiorite is a pinkish-gray, generally massive, medium-grained rock characterized by light-gray or pinkish-gray subhedral microcline porphyrocrysts (~1 cm long) and pink plagioclase phenocrysts (as much as 1.5 cm long) in a finer grained, essentially equigranular groundmass of feldspar, quartz, hornblende, and biotite. The mafic minerals typically form clusters between feldspar and quartz grains. Adjacent to the southeast margin of the shear zone (fig. 2), the rock is partly deformed into protomylonite, but it lacks a penetrative structure.

Plagioclase (oligoclase) occurs as blocky to lath-shaped, subhedral crystals that have Carlsbad-albite and, less commonly, pericline twins. A weak but distinct concentric normal zoning can be distinguished in a few sections. Myrmekite commonly occurs adjacent to microcline. Phenocrysts locally have a mortar structure (type IP core-mantle aggregates and type IIP structures; Hanmer, 1982) that comprise intragranular, fine-grained recrystallized plagioclase. Adjacent to the shear zone, the plagioclase contains abundant tiny rounded inclusions of quartz.

Quartz generally occurs as sutured grains that are somewhat crackled and show undulose extinction. Most quartz is interstitial but some forms anastomosing aggregates as much as 5 mm long.

Microcline commonly occurs as phenocrysts that have undulose extinction and contain inclusions of myrmekitic plagioclase and quartz. At places, elliptical quartz grains and aggregates of grains are clustered near grain margins. Type IM mantle aggregates and type IIM structures (Hanmer, 1982) are present locally. Judged from its morphology, microcline behaved more brittly than associated plagioclase to the strain related to the shearing. In a thin section, from the center of section 7 (fig. 8), for example, plagioclase has been partly reduced to fine-grained intragranular and mantle aggregates, whereas nearby microcline is only strained and crackled. Not uncommonly, shears transect microcline megacrysts.

The principal mafic mineral is dark-green to bluish-green, poikilitic hornblende commonly 0.5 to 1 mm in diameter. It is partly intergrown with biotite of approximately the same grain size, which probably is a primary mineral, but also is partly altered to biotite. The secondary biotite and associated reaction products--magnetite, sphene, and epidote--tend to form ill-defined pseudomorphs after hornblende. The secondary biotite is much finer grained than the original biotite. The accessory minerals are mainly zircon, apatite, and altered allanite.

Shear Zone

Waupee Volcanics

Amphibole schist

Mafic rocks in unit Xwa (inside shear zone) superficially resemble those in unit Xwm (outside shear zone) in being dark-gray porphyritic lavas, but they are variably mylonitized and foliated, less altered, and have a slightly higher metamorphic grade (lower amphibolite facies). An exception is the mafic rocks in the extreme northeastern part of the shear zone, west of Nelligan Lake (see Sims, 1989), which are highly altered and bleached and definitely greenschist facies. The chemical composition of a representative sample of amphibole-facies lava is listed in table 1.

Plagioclase occurs as sparse anhedral porphyroclasts as much as 5 mm in diameter, but mainly has been recrystallized to finer grained aggregates. Core-mantle aggregates (Hanmer, 1982) are common. Quartz grains are sparse.

The amphibole is either a green or a bluish-green hornblende; the latter appears to replace the former. The hornblende commonly contains oriented blebs of magnetite. Olive-green to brown biotite and associated sphene is locally associated with the hornblende and appears to be mainly an alteration product of it. Skeletal opaque oxides occur in a few sections. Zircon is a sparse accessory mineral.

Quartz diorite schist

Quartz diorite schist is a dark-gray, speckled, medium-grained metamorphosed rock that intrudes the amphibole schist in section 33, T. 32 N., R. 17., and presumably is a subvolcanic body. It has a conspicuous foliation and lineation expressed by aligned hornblende aggregates. This structure is modified by local mylonitic structures, characterized by grain-size reduction through recrystallization. Narrow mylonitic shears are common. Compositionally (table 3), the rock overlaps that of the Hines Quartz Diorite but it differs from it in megascopic appearance and by being deformed.

Plagioclase (An_{35-40}) occurs as concentrically zoned, elongate subhedral crystals having combined Carlsbad-albite twins and pericline twins. It is partly recrystallized to IP core-mantle aggregates (Hanmer, 1982) and fine-grained intragranular aggregates (mortar structure). In shears, recrystallized plagioclase is intergrown with recrystallized hornblende.

Hornblende is altered to varying degrees. The least altered, possibly original hornblende, is medium to dark green. It is intergrown with a bluish-green hornblende that is strongly poikilitic and is probably an alteration product of the green hornblende. Locally, either of the hornblendes have a partial rim of a secondary fibrous pale-green actinolitic hornblende or actinolite. Recrystallized hornblende in mylonitic shears is fine grained and dark green. Much of the original(?) hornblende is replaced by pseudomorphs of fine-grained secondary grayish-olive to brown biotite.

Quartz grains are interstitial and paragenetically late. The grains have undulose extinction and are recrystallized to sutured grains ~0.5-0.8 mm in diameter that retain the approximate outline of the original 3-4-mm-diameter grains. The accessory minerals are zircon, as much as 0.15 mm in diameter, apatite, and altered allanite.

The quartz diorite schist (unit Xwq, fig. 2) is interpreted as a coeval, subvolcanic intrusive rock associated with the metamorphosed porphyritic lava flows (units Xwm and Xwa).

Interlayered biotite and amphibole schist

Interlayered biotite and amphibole schist (fig. 4; table 4) is the most abundant rock unit in the Mountain shear zone (unit Xwi, fig. 2). The rocks are dark-gray to pinkish-gray, generally fine grained schists, which can be designated primarily as biotite, biotite-amphibole, and amphibole schist. More restricted rock types include cummingtonite-bearing schist, gedrite-bearing schist, and magnetite-rich schist; none of these units were mapped separately (Sims, 1989), and only the magnetite-rich schist is discussed following. The modes in table 4 are representative of the major rock types; the cummingtonite-, gedrite-, and magnetite-rich schists are generally too fine grained

for accurate modal analysis. The metamorphic grade of the rocks is mainly lower amphibolite facies. The rocks are grossly equivalent to the interbedded felsic and mafic volcanic rock unit (Xwf) outside the shear zone, and represent strongly deformed variants of the volcanic rocks within the shear zone. The schists are intruded by granite gneiss and, adjacent to the main body of granite gneiss, by dikes of pink leucogranite gneiss that are part of the granite gneiss body.

Biotite schist

Biotite schist is the dominant rock type in the interlayered schist unit (Xwi). It consists principally of plagioclase and quartz and contains variable amounts of K-feldspar and biotite; it is gradational into biotite-amphibole schist and cummingtonite-, gedrite-, and magnetite-bearing schists.

Plagioclase, the most abundant mineral (table 4), occurs as anhedral to subhedral porphyroclasts as much as 2 mm in diameter, and as smaller, recrystallized grains (oligoclase) as mantle aggregates and intragranular aggregates (Hanmer, 1982). Combined Carlsbad-twinning in the porphyroclasts is partly obliterated by strain. Locally, the porphyroclasts are antiperthite. The recrystallized, intragranular plagioclase generally is untwinned or weakly twinned. The porphyroclasts commonly contain small rounded blebs of quartz; the recrystallized grains lack quartz blebs. The porphyroclasts are generally weakly to moderately altered to sericite, and (or) epidote/clinozoisite and calcite.

Quartz commonly is recrystallized as lensoid aggregates of sutured grains, which in part express a foliation; the recrystallized grains are generally somewhat larger than the associated recrystallized plagioclase. Many quartz aggregates contain inclusions of recrystallized plagioclase and biotite, indicating that recrystallization of the quartz occurred after grain size-reduction of plagioclase and biotite.

Biotite is various shades of olive brown, and for the most part occurs in shears in association with recrystallized plagioclase, K-feldspar, and minor quartz. This biotite commonly is obliquely oriented to the strike of the shears. In relatively unsheared rocks, the biotite retains its original grain size and orientation; commonly, this biotite is associated closely with sphene.

K-feldspar is sparse or absent in most of the biotite schist. It occurs as interstitial twinned microcline grains, rare microcline porphyroclasts, and as recrystallized, very fine grained material in shears and core-mantle aggregates. Accessory minerals are zircon, apatite, opaque oxides, and allanite (altered).

Biotite-amphibole schist

This rock type differs mainly from biotite schist in containing a few percent of amphibole. The amphiboles are mainly hornblende or related actinolitic hornblende or actinolite, but include colorless clinoamphiboles, mainly cummingtonite(?). The hornblende is a green or bluish-green variety, which locally is variably altered to actinolitic hornblende or actinolite. It is commonly intergrown with biotite, and occurs as both original crystals and as recrystallized hornblende with associated biotite, plagioclase, and locally, K-feldspar and quartz in shears. The clinoamphibole is relatively widespread as a minor component of the rocks. It is typically colorless, acicular or bladed, and has a high relief and an extinction angle of 10°-15°. Tentatively, the mineral is identified as cummingtonite.

Amphibole schist

Amphibole schist is interlayered with the biotite and biotite-amphibole schists, mainly as subordinate discrete compositional layers a meter or more thick. Contacts with the biotite-bearing schists are sharp. The schist is dark gray, fine to medium grained, generally equigranular, and strongly foliated. It ranges from massive and homogeneous to a thinly layered rock. It is dominantly of lower amphibolite facies metamorphic grade.

The rocks are mainly composed of amphibole and plagioclase (table 4) but locally contain as much as 10 percent quartz and (or) magnetite and sparse biotite. Epidote and chlorite are the principal alteration minerals. The amphibole varies widely from green hornblende to bluish-green hornblende, or actinolitic hornblende, to pale-green actinolite, reflecting a decrease in metamorphic grade from lower amphibolite to actinolite facies. Plagioclase is partly recrystallized to ca. 0.1-0.3 mm-diameter grains, which are finer grained than the associated amphibole. Quartz is an interstitial mineral. Magnetite is disseminated through the rocks as small subhedral crystals.

Magnetite-rich schist

Magnetite-bearing schist containing as much as 20 modal percent of magnetite is exposed on the southwest side of the broad hill in the SE1/4 of section 6, T. 31 N., R. 17 E. and on the lower and smaller hills immediately to the west (fig. 8). The magnetite units trend northeastward but their lateral extent and length have not been delineated. The magnetic rocks are interlayered with biotite-amphibole schist and amphibole schist.

Two principal lithotypes contain uncommon amounts of magnetite: (1) actinolite-magnetite schist and (2) cummingtonite- or grunerite(?) -magnetite schist. Although both of these rock types are compositionally layered, the magnetite is disseminated through them, apparently indiscriminantly, indicating that the magnetite is a metamorphic mineral formed by the breakdown of iron-rich silicate minerals that were redistributed during the ductile-brittle deformation accompanying the shearing. Possibly the protolith had a higher iron content than most of the other volcanic rocks in the interlayered schist unit.

Granite gneiss

Granitic gneiss ranging in composition from tonalite to granodiorite (table 5; fig. 3) is a major constituent of the Mountain shear zone (unit Xgn, fig. 2). It crops out in the western part of section 5 and the eastern part of section 6, T. 31 N., R. 17 E. (fig. 8), within an area of about 1 km², and forms spectacular, nearly continuous exposures on several hills in this area. The gneiss contains local rafts of amphibolite. It is distinguished from other felsic rocks in the shear zone mainly in being coarser grained and lighter in color.

The gneiss is megascopically a light-gray, generally medium-grained rock characterized by narrow, discontinuous, commonly intersecting dark-gray mafic-rich zones and by coarse, lensoid quartz aggregates. In thin section, the dark-gray zones are seen to be ductile shear zones (S-C structures; Lister and Snoke, 1984) composed mainly of fine-grained (commonly less than 0.2 mm in diameter) biotite, rare hornblende, plagioclase, and microcline that have formed by grain-size reduction of the original minerals. Overall, the rocks are mainly protomylonite, but in part they are orthomylonite, as defined by Wise and others (1984).

Plagioclase (oligoclase) is typically bimodal. It occurs as (1) sub-rectangular to lath-like porphyroclasts 1-5 mm across associated with somewhat finer grained, anhedral, nearly equigranular plagioclase (~0.5-1 mm diameter) and (2) as fine-grained, recrystallized, intragranular plagioclase commonly less than 0.2 mm in diameter. Both the porphyroclasts and the associated intragranular plagioclase are twinned according to the combined Carlsbad-albite laws, and locally, pericline laws, and at places have a small amount of patch antiperthite; in a few sections, quartz blebs are visible. Rarely, K-feldspar blebs are present in the plagioclase. The recrystallized plagioclase occurs as type IP mantle aggregates (Hanmer, 1982), intragranular plagioclase that at places completely obliterates the original grains, and as fine-grained material in narrow shear zones that transects all the original minerals except biotite. The recrystallized plagioclase generally is untwinned. It is typically less altered by weathering than the original grains.

Quartz is recrystallized into aggregates of coalesced, sutured and strained grains that are distinctly coarser grained than the associated recrystallized plagioclase, microcline, and biotite. Typically, the aggregates are lensoid, as much as 4 or 5 mm in length, and consist of several individual grains 0.3 to 1 mm in diameter. Some quartz is crackled by late brittle fractures. The aggregates locally contain considerable plagioclase and are transected by late shears filled with recrystallized biotite, plagioclase, and microcline. The quartz lenses impart a foliation to the rock, which is conspicuous on weathered surfaces because the quartz stands out in relief.

Microcline generally is sparse. It occurs mainly as interstitial grains between plagioclase and quartz and as recrystallized fine-grained (~0.2-0.3 mm in diameter) material intergrown with plagioclase and biotite in narrow shears.

Biotite is grayish olive to greenish brown and occurs entirely as clots of very fine (0.1-0.2 mm) grain size that are nearly confined to narrow shears--the C- and S-surfaces of Berthé and others (1979) and Lister and Snoke (1984). The biotite generally is intergrown with recrystallized plagioclase and microcline, and commonly is oriented obliquely to the S-surfaces (foliation), thus indicating a unidirectional movement along the S-surfaces (Simpson, 1986). With few exceptions, the biotite is closely associated with epidote, sphene, and magnetite, suggesting that it probably is secondary after hornblende. Sparse garnet is associated with biotite in one section.

Amphibole is uncommon in the granite gneiss. Where present, it is a dark-green or dark-bluish-green hornblende that is intergrown with biotite. In two sections, it is in late shears and appears to have formed from biotite by prograde metamorphism. Possible pseudomorphs of an original hornblende consist of epidote, biotite, sphene, magnetite, quartz and, locally, hornblende. Accessory minerals include zircon, apatite, and allanite (altered).

The chemical composition of 3 samples of granite gneiss is given in table 1.

The protolith of the granitic gneiss is the granodiorite (fig. 2). The granodiorite borders the southeast margin of the shear zone, strikes into it, and was incorporated into the shear zone, where it was deformed and metamorphosed under conditions of the lower amphibolite facies during the ductile shearing. The deformation was accompanied by pronounced mineralogic changes in the amounts of modal quartz and amphibole. Comparison of modal analyses of the granitic gneiss (table 5) and the granodiorite (table 2) indicate that the gneiss contains about 15 percent more quartz than the protolith and has much less amphibole. The apparent decrease in K-feldspar in the gneiss, as indicated by modal data, probably results from original variations in the amount of K-feldspar in the granodiorite, and thus almost certainly is a sampling problem.

The increase in modal quartz in the gneiss (fig. 3) over the protolith appears to be related to metamorphic reactions accompanying the deformation. Specifically, the increase in quartz from an average of about 21 percent in the granodiorite to an average of more than 36 percent in the gneiss is attributed to the separation of silica from plagioclase during recrystallization. Relatively undeformed plagioclase in the granodiorite appears homogeneous in thin section, whereas in the partly deformed rocks, plagioclase contains abundant tiny inclusions of quartz. Within the shear zone itself, the plagioclase appears to lack tiny quartz blebs, but instead contains larger, generally rounded inclusions of quartz or coalesced groups of inclusions sufficiently large to be counted modally. Quantitative estimates of the amount of quartz clearly associated with particular plagioclase grains have not been determined, but we suggest that the aggregate amount of silica derived from recrystallization of the plagioclase is on the order of 10 percent, to account for the increase of modal quartz in the granite gneiss. Chemical analyses (table 1) also show higher SiO_2 values in the shear zone. The processes involved in the release of silica from plagioclase during progressive deformation are presumed to be, first, exsolution of silica as a subsolidus reaction during initial recrystallization of plagioclase and, second, the migration and coalescence of individual quartz blebs into larger grains during ductile deformation of the rock. Inasmuch as the quartz was relatively more mobile than the associated plagioclase (and microcline) in the stress field, it differentially flowed into lensoid masses oriented perpendicular to the major axis of stress.

Baldwin Conglomerate

The Baldwin Conglomerate (unit Xb, fig. 2) is a lens-shaped metaconglomerate unit about 2.3 km long and a maximum of 150 m wide that is exposed locally along the northwest margin of the shear zone. It unconformably overlies metavolcanic rocks and granite gneiss within the shear zone and is truncated and intruded on the north by the Hager Rhyolite (Lahr, 1972).

The metaconglomerate has relict sedimentary layering and possible scour features and contains subrounded to subangular clasts of metavolcanic rocks, gneiss, and quartzite. All but the quartzite are of local derivation; the quartzite clasts probably were derived from the more distal quartzite bodies exposed at McCaslin Mountain and Thunder Mountain, 20 km and 10 km, respectively, north of the Mountain shear zone (Sims, 1990; Sims and Peterman, 1988). Clasts of the Hines Quartz Diorite (unit Xh, fig. 2) have not been reported in the metaconglomerate (Lahr, 1972). The matrix is a gray, fine- to medium-grained rock containing quartz, feldspar, biotite, muscovite, and hornblende. A weak foliation exists that is approximately parallel to the bedding.

Hines Quartz Diorite

The Hines Quartz Diorite (unit Xh, fig. 2) occurs mainly as a lenticular intrusive body about 3 km long and 0.4-0.5 km wide in the southwestern part of the Mountain shear zone. Except for exposures on hills in the SW1/4 section 12, T. 31 N., R. 16 E. (see Sims, 1989), outcrops are small and dispersed. The quartz diorite intrudes adjacent metavolcanic rocks and contains local, small inclusions of them. It is intruded to the southwest by the Belongia Granite (unit Ywb, fig. 2) of the Wolf River batholith (Lahr, 1972), which is not discussed in this report.

The quartz diorite is virtually undeformed, indicating that it was intruded into and along the shear zone as the shearing waned. A meter-thick dike in the SE1/4NW1/4 section 7, T. 31 N., R. 17 E. (fig. 8) is intruded along the S_2 foliation. It has a U-Pb zircon age of $1,812.7 \pm 3.6$ Ma (Sims and others, 1990).

The rock is dark gray, locally slightly mottled, massive, and medium grained. Primary subophitic textures are retained by plagioclase laths, but the primary mafic mineral--apparently clinopyroxene--is so highly altered to a pale-green amphibole (actinolitic hornblende), actinolite, biotite, and opaque oxides and locally chlorite that its primary texture is largely obliterated. Although little deformed, the quartz diorite is cut by local mylonitic shears.

Plagioclase (An_{35-48}), the major mineral (table 3), is weakly altered, has a moderate concentric zoning, is tabular to lath-shaped, and is twinned according to the combined Carlsbad-albite and the pericline laws. Crystals are typically 2-4 mm long. Quartz is interstitial, somewhat strained, and highly variable modally (table 3).

The biotite associated with clinopyroxene as an alteration mineral is dark reddish brown. Elsewhere in the rock, the biotite is dark yellowish brown, and is always associated with opaque oxides. Cumingtonite is a local alteration product of actinolitic hornblende, occurring as colorless needles extending outward from the actinolitic hornblende. Apatite, zircon, and allanite are common accessory minerals.

REGIONAL STRUCTURE

The rocks southeast of the shear zone have a single penetrative structure (cleavage or schistosity, S_1) that is interpreted as the regional foliation. It is generally subparallel to layering and pillow structures (S_0) but locally is crosscutting. The structure is expressed by sub-planar fractures, aligned biotite sheets, and flattened clasts. S_1 occurs both in the metamorphic and granitoid rocks. Flattened mafic volcanic inclusions in the granodiorite are oriented parallel to S_1 . S_1 is oriented northward and dips steeply; within the map area, the strike ranges from about N. 15° W. to N. 25° E. (fig. 5).

SHEAR ZONE

The Mountain shear zone sharply truncates and is distinctly younger than the regional foliation (S_1) and resulted from a ductile shear deformation (D_2) that was superimposed on the regional foliation (fig. 2; table 6). D_2 structures are modified locally by two younger ductile deformation events (D_3 and D_4), which do not appreciably affect the distribution of the rocks, and still younger kink bands formed by brittle deformation (Sims and others, 1990; F_k , table 6).

Mesoscopic Structure

The shear zone strikes about N. 55° E. and is 1 to 2.5 km wide. S_2 foliation within the shear zone has an average strike of N. 55° E. and a moderately steep southeast dip. Northwest-oriented S-surfaces within the shear zone are local relicts of S_0/S_1 . The southeastern margin of the shear zone--the transition between mylonitic and non-mylonitic rocks--is abrupt. It is composed of narrow zones of mylonitic rocks that clearly truncate S_1 structures (as can be seen at stop 2). The transition zone is 100 m or less wide. The boundary is a zone of deformation and differential displacement of the walls accomplished by ductile flow, without apparent separation.

The shear zone is truncated on the north and northeast ends by the younger Hagen Rhyolite, and on the southwest end by granitic rocks (unit Ywb, fig. 2) of the 1.47 Ga Wolf River batholith, which had little or no tectonic effect on the rocks in the shear zone. Accordingly, neither the full width nor the termination of the shear zone is known.

The dominant structures in the shear zone are related to D_2 and consist mainly of an S_2 surface (S_2) resulting from tectonic layering and a mylonitic foliation; a stretching lineation (L_2), expressed by elongate clasts, rodding, and biotite streaks; and rare folds of S_2 surfaces (L_2). The resulting structure is an L-S tectonite (Ramsay, 1980). The younger ductile structures (D_3 , D_4) that are superposed on the older structures (table 6) do not appreciably modify the geometry of the rocks.

In the granitoid rocks, which were relatively isotropic initially, S_2 consists dominantly of thin anastomosing, commonly sigmoidal biotite-rich zones and oriented, flattened aggregates of quartz (S-C structure); a related lineation is given by streaked biotite aggregates and quartz rods, composed of fine-grained recrystallized aggregates of quartz. In strongly anisotropic rocks, such as bedded tuff, S_2 is locally expressed by brecciated and boudinaged, hard (siliceous) layers that have been tightly folded so that S_0/S_1 has been tectonically deformed into parallelism with S_2 . Elsewhere, segments of S_0/S_1 surfaces have not been completely transposed. In layered felsic-mafic rocks (unit Xwa, fig. 2) in the northeastern part of the shear zone, the felsic layers commonly have a conspicuous S_2 foliation expressed by recrystallized biotite that obliterates S_1 , whereas the mafic layers have only a weak S_2 foliation generally expressed by aligned hornblende or chlorite. In these rocks, S_0 commonly is not appreciably reoriented or transposed. The felsic-mafic layer contacts may show a slightly zig-zag, sawtooth (chevron fold) pattern similar to that formed by a slip cleavage.

F_2 folds are not common, but have been observed at a few localities, especially in interlayered felsic-mafic volcanic rocks. The folds generally plunge gently to moderately northeast or southwest.

Microscopic Structure

The mylonitic rocks in the shear zone are characterized by recrystallization as indicated by conspicuous grain-size reduction. Biotite and quartz are entirely, or nearly entirely, recrystallized, whereas plagioclase and microcline, which were harder than the matrix, were only partly recrystallized. According to the classification of Wise and others (1984), the rocks are mainly protomylonite or orthomylonite. Biotite occurs in irregular, anastomosing trains, commonly associated with recrystallized plagioclase and, locally, microcline. Both the biotite and feldspar are very fine grained (~ 0.2 mm diameter). Quartz is recrystallized into oriented lenses or rods, which are coarse-grained aggregates of individual sutured grains < 0.5 mm in diameter; commonly, it contains inclusions of plagioclase and biotite and, locally, sphene and amphibole. Plagioclase commonly is only partly recrystallized, it occurs as porphyroclasts as much as 5 mm in diameter and commonly has mantles (type IP structures; Hanmer, 1982) of fine-grained (~ 0.1 - 0.2 mm) plagioclase. Also, fine-grained recrystallized plagioclase occurs as veinlets (shears) that traverse the rocks.

Kinematic Analysis

The sense of movement during the F_2 shear deformation has been determined from S-C fabrics (Berthé and others, 1979; Lister and Snoke, 1984) in the mylonitic granite gneiss (unit Xgn, fig. 2), orientation of stretch fabrics (long axes of strain ellipsoids), and other field data such as described by Simpson (1986). S-C mylonite fabrics (fig. 6) in granite gneiss from several localities show that the southeast block of the Mountain shear zone is upthrown relative to the northwest block. Stretch fabrics, as indicated by the orientation of the long (X) axes of strain ellipses, also suggest the same sense of movement. Stereoplots show that stretch axes (L_{2b}) plunge steeply south to east. This steeply plunging stretch fabric is possibly caused by a simple shear couple having a

shallower dip than the dip of the stretch axis, as shown on figure 7. Stereoplots at some localities indicate a steep easterly L_{2b} stretch fabric, whereas a stereoplot at one locality indicates a steep southerly stretch fabric. This difference in orientation of stretch fabrics suggests a varying sense of movement of the shear couple along the length of the shear zone.

Field data suggest that the sense of movement in the horizontal plane was generally dextral (right lateral). At the map scale (fig. 2), S_1 foliation appears to have been rotated clockwise toward parallelism with the shear zone, which is consistent with dextral movement. Also, the granodiorite body (unit Xg, fig. 2) is displaced dextrally in the shear zone as granitic gneiss (unit Xgn). In addition, S-C fabrics suggest dextral movement. Such movement is supported also by measured axes of extension (L_{2b}) as well as by computed ellipsoid axes; these axes plunge toward the east-southeast quadrant of stereoplots, indicating that the upthrust southeast block had a dextral component of movement in the horizontal plane, as shown on figure 7 (Sims and others, 1990). On the other hand, the variation in orientation of L_{2b} suggests a different horizontal shear sense from place to place along the shear zone.

GEOBAROMETRY AND GEOTHERMOMETRY

The regional D_1 deformation textures and associated M_1 metamorphic minerals were overprinted in the shear zone by shear-induced D_2 fabrics and M_2 metamorphic mineral assemblages (Day and others, in press). Hornblende geobarometry of granitoid rocks and metabasalt indicates that the regional M_1 pressure of about 4.5 ± 1 Kbar was elevated during M_2 to approximately 6.0 ± 1 Kbar both within and adjacent to the shear zone. Garnet-biotite geothermometry indicates that M_1 ranged from 585° to 615° C, but that M_2 equilibrium temperatures may have increased to as much as 654° C. Inasmuch as garnet is rare both outside and inside the shear zone, the estimated peak M_2 thermal conditions are only approximate. Two-feldspar geothermometry indicates that inside the shear zone the feldspars underwent continuous subsolidus reequilibration from peak M_2 conditions (574° C at 6.0 Kbar) down to very low temperatures (280° C at 6.0 Kbar). These data indicate that both pressure and temperature increased during shear-induced M_2 recrystallization within the Mountain shear zone.

CONCLUSIONS

Rocks within the Mountain shear zone are L-S tectonites that fit with the classic definition of a shear zone (Ramsay, 1980). The shear is a local (discrete) linear zone of high ductile deformation. Neither the width nor the length are known, but we presume that the shear zone had parallel sides prior to emplacement of the Wolf River batholith. In this area, the batholith was emplaced passively.

Ramsay (1980) classified shear zones according to their displacement fields. The Mountain shear zone appears to fit best Ramsay's (1980) type (iii) classification; that is, it is a heterogeneous simple shear with accompanying heterogeneous volume change. Ramsay and Graham (1970) noted that most deformation from natural orogenic processes lead to states of heterogeneous strain, and Ramsay (1980, p. 89) noted that "the basic component of practically all shear zones is that of heterogeneous simple shear . . . made up of a number of elements showing homogeneous simple shear." The Mountain shear zone has the following features that fit these criteria.

(1) Deformation within the shear zone was not uniform; domains of relatively low strain, where pre- F_2 features are preserved, as well as domains of high strain, where the country rocks have been mylonitized, are present.

(2) The nature, magnitude, and orientation of displacement varies both across and along the shear zone, as suggested by the variation in (a) shape of the strain ellipsoids at different localities, (b) percentage of flattening and stretching among the outcrops where strain analyses were carried out (Sims and others, 1990), and (c) movement pattern along the shear zone; the orientation of the maximum principal shortening direction apparently rotated from nearly west in the northern part of the exposed shear zone to north-northwest at its southwestern exposed segment, as indicated by the orientation of deformed clasts (L_{2b}).

(3) L-S tectonites of the shear zone appear to have formed primarily by plane strain, as indicated by the Flinn K values (Flinn, 1962) at two localities, which are very close to 1, the value for plane strain (Sims and others, 1990). Rocks at another locality, however, have a K value of 2.4, indicating that they lie in the field of apparent constriction.

(4) The magnitude of both flattening and stretching increases from the southeast margin of the shear zone toward the interior.

(5) Finally, Roy (1977) has pointed out that mylonites within shear zones develop in stages and that this development is accompanied by chemical mobility. Chemical mobility is suggested in the Mountain shear zone by an apparent substantial gain in modal quartz and some increase in SiO_2 (table 1) of the granodiorite gneiss (unit Xgn) within it, as compared to undeformed granodiorite (unit Xg), but this apparent gain has not been confirmed by quantitative chemical studies.

REGIONAL SIGNIFICANCE

Prior to the study of the Mountain shear zone (Sims and others, 1990), major northeast-trending shears in the Wisconsin magmatic terranes were known to have formed subsequent to the main regional deformation in the Penokean orogen (Sims and Peterman, 1983, p. 8), but data on the precise timing were lacking. The fortunate existence of a datable intrusive rock (Hines Quartz Diorite, $1,812.7 \pm 3.6$ Ma) within the Mountain shear zone, which was emplaced during the waning stages of the ductile deformation and thus establishes a firm younger limit for the shearing, provides a benchmark for the time of shearing on a regional basis. Thus, it can be presumed now that the steep Jump River and Athens shear zones (fig. 1) as well as others of N. 50° - 60° E. trend (Sims and others, 1989) have comparable ages of about 1,815 Ma. A deformed intrusive rock within the Athens shear zone, which has a crystallization age of $1,832 \pm 9$ Ma (Sims and others, 1989), is consistent with this interpretation that the shearing post-dates 1,840 Ma. The recognition of a major ductile deformation event 25 to 35 m.y. after the main continent-arc collisional event ($\sim 1,860$ Ma; Sims and others, 1985) along the Niagara fault zone indicates the complexity and long duration of tectonic events in the Penokean orogen and the need to identify specific events, perhaps by assigning names to them, such as has been done in the Wopmay orogen in the Northwest Territories (Hoffman and Bowring, 1984). It seems improbable that the $\sim 1,815$ Ma ductile deformation could be a late tectonic episode related to the collision that sutured the Wisconsin magmatic (arc) terranes to the continental margin at about 1,860 Ma. Instead, we suggest that the $\sim 1,815$ Ma ductile deformation probably resulted from a continent-continent or continent-arc collision far removed from northern Wisconsin, probably to the south or southeast. A possible candidate was accretion of the Central Plains orogen (Sims and Peterman, 1986) to the North American continent.

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FIELD STOPS AND ROAD LOG

FIELD STOPS ARE SHOWN ON FIGURE 8

- STOP 1 Riverview road, 0.7 mi east of County Highway (Co. hy.) W. Park in open space at road junction. Walk east 0.1 mi through woods to prominent knob.

Massive to pillowed porphyritic basalt. Pillows are about 1 ft thick and 2 ft long; they trend N. 50° W., dip steep, and top to southwest. Outcrops are typical of Xwm unit of Waupee Volcanics (Sims, 1989); they are outside the shear zone.

Drive west on Riverview road to Co. hy. W; turn left (southwest) on hy. W and proceed for 1.3 mi; turn left (east) on Bear Paw road and proceed for 1.1 mi; park on north side of road.

- STOP 2 This outcrop area extends westward for about 0.2 mi. It contains dominantly felsic volcanic rocks (Xwf, fig. 2; Sims, 1989) that are intruded on the west by porphyritic granodiorite (Xg, fig. 2). The outcrops are outside the shear zone.

The most easterly outcrops are a layered biotite-garnet schist of Xwf unit; a crude layering is expressed by garnet-rich zones. Numerous flattened clasts are present. S_1 foliation is subparallel to layering (S_0); it is oriented N. 28° E., 78° NW. (fig. 9). A 3 ft-wide zone of folding deforms S_1 . Axial planes to these folds trend N. 70° E. and dip 82° SE. This orientation roughly corresponds to S_2 inside the shear zone. The folded zone is interpreted to be a small domain of F_2 deformation that developed outside the shear zone.

Westward, narrow dikes of granodiorite-tonalite and pink aplite of Xg unit cut the schist. The westernmost outcrops are dominantly porphyritic granitoid rocks (Xg unit, fig. 2) that contain inclusions of amphibolite and biotite schist. The granitoid rocks are nearly massive.

Return to Bear Paw road. Proceed west for a distance of 0.4 mi. Walk north about 0.05 mi to outcrop of a pink granite gneiss.

- STOP 3 The granite gneiss (Xgn, fig. 2) is cut by pink aplitic granite; these rocks are within the Mountain shear zone. The granite gneiss has a prominent S_2 foliation expressed by aligned biotite sheets that is oriented N. 53° E., 86° SE (fig. 10). The southeastern margin of the shear zone lies in the northeast-trending valley to the east of Stop 3.

Return to Bear Paw road. Proceed west on this road to junction with County Highway W. Turn right (northeast) on Highway W and proceed to Riverview road. Park on Riverview road and walk southwest on Highway W for a distance of 0.35 mi. Walk southeastward onto outcrops of mylonitic granite gneiss.

- STOP 4 These outcrops of granite gneiss (mylonite) (Xgn, fig. 2) have well-defined S-C structures (Berthé and others, 1979; Lister and Snoke 1984). C is a foliation oriented subparallel to the southeastern boundary of the shear zone. The relation of S (an oblique foliation) to C indicates that the southeast block of the shear zone is upthrown

relative to the northwest block. The sense of movement in the horizontal plane was generally, but not everywhere, dextral. The stretching lineation (X finite strain axis), as indicated by deformed clasts, plunges steeply south to east; it is subparallel to the intersection of S and C.

Return to Riverview road. Proceed southwestward on County Highway W to junction with Bear Paw road, and continue southwest on Highway W for a distance of 1.3 mi. Park in low area, near house on northwest side of road, and walk southwestward on County Highway W for 0.1 mi. Proceed northwestward to outcrops in the woods.

- STOP 5 These outcrops are within the shear zone and are dominantly felsic biotite schist of unit Xwi (fig. 2). The schist has a prominent S_2 foliation that has been folded into a series of east-southeast plunging folds (F_3) (fig. 11). The axes of these folds have, in turn, been folded by younger north-trending folds (F_4). The axes and axial planes associated with F_4 cannot be measured on these outcrops.

In these outcrops, the S_2 foliation is expressed by brecciated and boudinaged siliceous layers that are generally parallel to S_2 . At places, however, S_0/S_1 surfaces (siliceous layers) are not completely transposed parallel to S_2 . The occurrence of isolated domains in which the older S_0/S_1 foliation is preserved is characteristic of the mylonite (shear) zone as a whole.

Return to parking area. Proceed west on Highway W for a distance of 1.4 mi to road junction on left. Proceed on this road into the village of Mountain and to Wisconsin Highway 32-64. Turn left (southeast) on Highway 32-64 and proceed southward for a distance of 0.9 mi to low outcrop on north side of highway in borrow pit.

- STOP 6 This outcrop is the Hines Quartz Diorite (unit Xh, fig. 2) (named by Lahr, 1972), a late-to post-tectonic rock intruded into the Mountain shear zone. (It has not been noted outside the shear zone). The Hines Quartz Diorite has a U-Pb zircon concordia intercept age of $1,812.7 \pm 3.6$ Ma (Sims and others, 1990).

The Hines Quartz Diorite is a lensoid body about 3 km long and 0.4-0.5 km wide that was intruded subparallel to the S_2 foliation; it is virtually undeformed, although locally cut by mylonitic shears, and therefore it establishes a minimum age for the Mountain shear zone. The quartz diorite is dark gray, slightly mottled, massive, and medium grained. Primary subophitic textures are retained by plagioclase (An_{35-48}) laths, but the primary mafic mineral--probably clinopyroxene--is highly altered to amphibole, biotite, and locally chlorite. The amphibole is mainly actinolitic hornblende and actinolite, but includes cummingtonite.

Note: A facies of the anorogenic Belongia Granite (unit Ywb, fig. 2) of the 1.47 Ga Wolf River batholith is exposed on the low knob on the southwest side of Highway 32 and 64. This intrusive rock truncates the southwest margin of the Mountain shear zone.

This concludes the field trip, but, an additional stop is described below because of its significance to the geology of a broader region. To reach Stop 7, return to Mountain,

then proceed to County Highway W; turn right (east) and proceed on Highway W for a distance of 2 mi. Outcrop on left (north) side of Highway W is stop 7.

STOP 7 The Baldwin Conglomerate (unit Xb, fig. 2) is exposed on the south-facing slope of the hill.

The Baldwin Conglomerate is a lensoid unit about 2.3 km long and a maximum of 150 m wide that is exposed locally (sec. 6, T. 31 N., R. 17 E.) along the northwest margin of the Mountain shear zone. It overlies metavolcanic rocks within the shear zone and is truncated and intruded (by dikes) on the north by the Hager Rhyolite (Ywh, fig. 2) of the Wolf River batholith (Lahr, 1972). The conglomerate has relict sedimentary layering and possible scour features and contains subrounded to subangular clasts of metavolcanic rocks, granite gneiss, and quartzite (as much as 25 cm in diameter). All but the quartzite clasts are derived locally from the shear zone itself. Clasts of the Hines Quartz Diorite have not been reported in the conglomerate (Lahr, 1972). The matrix of the conglomerate is a gray, fine- to medium-grained rock containing quartz, feldspar, biotite, muscovite, and hornblende of a proximal source.

The conglomerate has a foliation that is oriented N. 66° E., 78° NW. It is interpreted as S_2 foliation inasmuch as it is parallel to S_2 in the mylonitic rocks of the shear zone.

The source of the quartzite clasts almost certainly was the quartzite now exposed at McCaslin and Thunder Mountains, less than 20 km to the north (Sims and Peterman, 1988; Sims, 1990). Thus, inasmuch as the Baldwin Conglomerate was deformed during the ductile deformation along the Mountain shear zone, and prior to emplacement of the Hines Quartz Diorite, the quartzite clasts and their parent rocks are older than 1,812 Ma, the age of the Hines Quartz Diorite.

Table 1.--Chemical analyses of selected rocks in Mountain area, Wisconsin

[Major oxide analyses by A. Bartel, E. Brandt, S. Roof, J. Taggart, and K. Stewart. FeO, H₂O, and CO₂ by E. Brandt and S. Roof.
Minor element analyses by K. Barovich, K. Futa, and Z. Peterman]

Sample No.	Major oxides (weight percent)								
	20-85	23-85B	27-85	35-85	39-85	129-85A	130-85A	150-85	153-85
SiO ₂	51.80	49.00	72.40	69.30	72.70	75.60	73.20	50.10	63.10
Al ₂ O ₃	17.70	18.90	14.50	14.20	13.20	12.40	13.00	19.00	17.40
Fe ₂ O ₃	2.69	2.00	0.31	0.98	0.89	0.61	0.83	0.90	0.21
FeO	7.30	9.54	2.19	3.14	3.18	1.52	3.10	10.80	2.89
MgO	4.77	3.54	1.69	1.00	0.50	0.22	0.55	4.57	1.64
CaO	7.35	10.90	2.19	1.47	2.22	0.41	2.21	9.68	7.52
Na ₂ O	3.29	3.40	3.44	2.79	3.60	2.39	3.62	2.64	2.68
K ₂ O	2.28	1.09	1.33	5.52	2.50	6.48	2.34	0.64	2.97
TiO ₂	0.73	0.66	0.29	0.23	0.19	0.02	0.18	0.70	0.65
P ₂ O ₅	0.25	0.10	0.05	0.09	0.05	0.05	0.05	0.17	0.23
MnO	0.16	0.16	0.02	0.04	0.05	0.02	0.04	0.19	0.05
H ₂ O+	1.48	1.28	1.15	0.68	0.58	0.29	0.45	0.99	0.40
H ₂ O-	0.04	0.01	0.02	0.01	0.04	0.06	0.03	0.02	0.02
CO ₂	0.03	0.19	0.01	0.04	0.15	0.02	0.09	0.05	0.01
Sum	99.87	100.77	99.59	99.49	99.85	100.09	99.69	100.45	99.77
Minor elements (parts per million) by X-ray fluorescence									
Rb	48	17	22	70	47	148	48	14	35
Sr	555	299	168	431	175	54	195	346	208
Y	21	12	22	8	27	22	18	9	15
Zr	92	35	165	115	93	42	97	27	91
Nb	9	2	8	2	6	7	4	1	7
K/Rb	396	541	511	658	440	363	402	377	711
Rb/Sr	0.086	0.056	0.129	0.161	0.270	2.735	0.248	0.041	0.167
Peacock	1.32	2.43	0.46	0.18	0.36	0.05	0.37	2.95	1.33

Table 1 (continued)

20-85	Amphibolite (unit Xwa), SW1/4 SW1/4 sec. 33, T. 32 N., R. 17 E.
23-85B	Porphyritic mafic metavolcanic rock (unit Xwm), NE1/4 NW1/4 sec. 4, T. 31 N., R. 17 E.
27-85	Felsic metavolcanic rock (unit Xwf), SW1/4 NW1/4, sec. 4, T. 31 N., R. 17 E.
35-85	Porphyritic granodiorite (unit Xg), NE1/4 SW1/4, sec. 5, T. 31 N., R. 17 E.
39-85	Granitic gneiss (unit Xgn), NE1/4 SW1/4, sec. 5, T. 31 N., R. 17 E.
129-85A	Granitic gneiss (unit Xgn), SW1/4 NW1/4, sec. 5, T. 31 N., R. 17 E.
130-85A	Granitic gneiss (unit Xgn), NW1/4 NW1/4, sec. 5, T. 31 N., R. 17 E.
150-85	Hines Quartz Diorite (unit Xh), NW1/4 NW1/4, sec. 13, T. 31 N., R. 16 E.
153-85	Intermediate metatuff (unit Xwi), SE1/4 SW1/4, sec. 1, T. 31 N., R. 16 E.

Table 2.--Approximate modes of granodiorite

Constituent	101	60B	35	112	165	34A	34B	Average
Plagioclase	42.8	49.5	40.7	44.8	51.8	55.2	23.3	47.5
Quartz	18.2	15	37.6	14.6	18.2	21.6	35.2	21
K-feldspar	24.6	17.5	15.4	20.6	11	5.7	40	16
Biotite	8.8	10	5.4	10.6	9	7	1.4	7.5
Hornblende	1.8	7.5	0	7.3	9.8	9	0	6
Alteration minerals	3.3	Tr.	0.1	Tr.	Tr.	Tr.	Tr.	) 2
Accessory minerals	0.5	0.5	0.8	2.1	0.2	1.5	0.1	

Sample localities

101	NE1/4 SW1/4 sec. 7, T. 31 N., R. 17 E.
60B	SE1/4 SW1/4 sec. 5, T. 31 N., R. 17 E.
35	NE1/4 SW1/4 sec. 5, T. 31 N., R. 17 E.
112	NW1/4 SE1/4 sec. 7, T. 31 N., R. 17 E.
165	SE1/4 NE1/4 sec. 7, T. 31 N., R. 17 E.
34A	NE1/4 SW1/4 sec. 5, T. 31 N., R. 17 E.
34B	Same as 34A. Pink leucogranite that cuts granodiorite.

Table 3.--Approximate modes of Hines Quartz Diorite
and quartz diorite of the amphibole schist unit

Constituent	Hines Quartz Diorite		Quartz diorite	
	147	150	134	172
Plagioclase	47.7	54.2	65.6	48.8
Quartz	32.6	2	9.7	22.6
K-feldspar	0	0	0	0
Biotite	12.4	0.9	5.6	12.2
Amphibole	6.3	¹ 41.7	18.4	15.8
Alteration minerals	0.1	Tr.	Tr.	0.4
Accessory minerals	0.9	1.2	0.7	0.2

¹Includes relicts of clinopyroxene.

Sample localities

147	Roadcut, NE1/4 NW1/4 sec. 14, T. 31 N., R. 16 E.
150	Outcrop, NW1/4 NW1/4 sec. 13, T. 31 N., R. 16 E.
134	Outcrop, SE1/4 NW1/4 sec. 33, T. 32 N., R. 17 E.
172	Outcrop, NE1/4 SE1/4 sec. 33, T. 32 N., R. 17 E.

Table 4.--Approximate modes of interlayered biotite and amphibole schist

	Biotite and biotite-amphibole schist											Amphibole schist			
	13	12	64	199	76	79	88	141-1	167	168	170	173	91	18	164
Plagioclase	50.8	46.2	50.5	47.7	53.3	46.7	41	42	31	50.5	40	51.5	27.5	25	32
Quartz	41.2	47.8	38.7	3.5	37.3	23.6	37	41.5	29	37.5	21.5	22.5	31	1	Tr.
K-feldspar	2	0.6	8.3	0.6	2.1	21.7	10	0	0	0	0	10	31.5	0	0
Biotite	4.9	4	1.8	0	5.8	7.2	8.5	13	28	11.5	37	15	3	0	0
Amphibole	0	0	0	2.4	0	0	2.5	0	5	Tr.	0	0	37	68.5	64
Muscovite	0.8	Tr.	0.3	0.1	Tr.	0.8	0	2	0	0	0	Tr.	Tr.	0	0
Epidote/clinozoisite	0	0.5	0.2	44.2	0.5	0	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.	3	2
Sphene	0	Tr.	0	0	Tr.	0	1	0	1	0.3	1	1	Tr.	2	Tr.
Chlorite	Tr.	0.2	0	0	0	0	0	3.5	0	Tr.	Tr.	Tr.	Tr.	Tr.	Tr.
Calcite	Tr.	0.2	Tr.	0.1	Tr.	0	0	Tr.	Tr.	Tr.	Tr.	Tr.	0	0.3	Tr.
Magnetite	Tr.	Tr.	Tr.	Tr.	1	Tr.	Tr.	0	6	Tr.	Tr.	Tr.	0	0.2	2
Accessory minerals	0.2	0.5	0.2	0.4	Tr.	Tr.	Tr.	Tr.	Tr.	0.2	0.5	Tr.	Tr.	Tr.	Tr.

¹ Plagioclase has abundant tiny quartz droplets.² Included with biotite in mode.³ Amphibole is a clinoamphibole, probably cummingtonite.

Sample localities

13	NW1/4 NE1/4 sec. 5, T. 31 N., R. 17 E.	167	NE1/4 NE1/4 sec. 7, T. 31 N., R. 17 E.
12	NW1/4 NE1/4 sec. 5, T. 31 N., R. 17 E.	168	NE1/4 NE1/4 sec. 7, T. 31 N., R. 17 E.
64	SW1/4 NE1/4 sec. 6, T. 31 N., R. 17 E.	170	SW1/4 NW1/4 sec. 27, T. 32 N., R. 17 E.
199	SE1/4 SW1/4 sec. 1, T. 31 N., R. 16 E.	173	NE1/4 SE1/4 sec. 33, T. 32 N., R. 17 E.
76	NE1/4 NW1/4 sec. 7, T. 31 N., R. 17 E.	91	SE1/4 SE1/4 sec. 12, T. 31 N., R. 16 E.
79	NW1/4 NW1/4 sec. 7, T. 31 N., R. 17 E.	18	SW1/4 SW1/4 sec. 33, T. 32 N., R. 17 E.
88	SW1/4 SW1/4 sec. 7, T. 31 N., R. 17 E.	164	NW1/4 SW1/4 sec. 7, T. 31 N., R. 17 E.
141-1	NE1/4 NE1/4 sec. 5, T. 31 N., R. 17 E.		

Table 5.--Approximate modes of granite gneiss

Constituent	39	46	50	129- 85B	130- 85A	130- 85B	55	40A	45B	44	18-1	129- 85A	Average
Plagioclase	44.2	49.4	48.2	54.0	51.0	47.0	46.7	46.1	44.4	48.4	57.7	29.5	48.8
Quartz	39.0	41.2	40.4	32.2	35.3	37.0	38.3	34.4	30.5	36.1	30.7	37.1	35.9
Microcline	10.6	0	1.5	1.0	5.0	8.0	9.4	1.3	19.0	8.7	0	30.9	5.9
Biotite	6.0	8.4	8.4	12.0	8.2	6.5	3.4	15.0	5.4	6.5	10.7	1.1	8.2
Hornblende	0	0	0	0	0	0	1.0	0.7	0	0	0	0	0.2
Alteration minerals	0.2	0.3	0.5	0.1	0.1	0.5	1.2	2.0	0.7	0.3	0.9	1.0	0.6
Accessory minerals	Tr.	0.7	1.0	0.7	0.4	1.0	Tr.	0.5	Tr.	Tr.	Tr.	0.4	0.4

Sample localities

39	NW1/4 SW1/4 sec. 5, T. 31N., R. 17E.	55	NE1/4 SE1/4 sec. 6, T. 31N., R. 17E.
46	SW1/4 NE1/4 sec. 5, T. 31N., R. 17E.	40A	NW1/4 NE1/4 sec. 5, T. 31N., R. 17E.
50	SE1/4 NW1/4 sec. 5, T. 31N., R. 17E.	45B	SE1/4 NW1/4 sec. 5, T. 31N., R. 17E.
129-85B	SW1/4 NW1/4 sec. 5, T. 31N., R. 17E.	44	NW1/4 NE1/4 sec. 5, T. 31N., R. 17E.
130-85A	SW1/4 NW1/4 sec. 5, T. 31N., R. 17E.	18-1	NW1/4 SW1/4 sec. 33, T., 32N., R. 17E.
130-85B	Same as 130-85A.	129-85A	Same as 129-85B, Leucogranite gneiss.

Table 6.--Sequence of events, structural elements, and nature of events in the Mountain shear zone

Event	Structural element	Symbol	Explanation	Nature of event
Deposition	Bedding, pillow structures intrusive contacts	S ₀	Primarily is layering between felsic and mafic rocks	Original depositional and intrusive feature
F ₁	Regional foliation Flattened clasts	S ₁	Best observed south of shear zone, but present in isolated outcrops within shear zone	Flattened clasts suggest regional compression
F ₂	Foliation	S ₂	Tectonic layering, mylonitic foliation in granitic gneiss (unit Xgn)	Foliation associated with simple shear due to relative upthrow from southeast. L _{2a} formed by fold of S ₁ and S ₀ . L _{2b} caused by simple shear associated with upthrow
	Fold axes	L _{2a}	Folds of S ₂	
	Long axes of strained clasts Stretched	L _{2b}	Stretched clasts that lie in S ₂ plane	
F ₃	Foliation	S ₃	Geometric element. Not generally observed	Late folding that created a west- northwest-trending fabric
	Fold axes	L ₃	Seen as fold axes on S ₂ surface	
F ₄	Foliation		Fabric is poorly developed Appears to fold F ₃	Rarely observed. Yielded a north- northwest-trending fabric
	Fold axes	L _{4a}		
	Lineations	L _{4b}		
	Shears			Shears are parallel to cross faults
F _k	Kink bands	S _k	Variously oriented kink bands and fold axes associated with kink bands	Late-state brittle deformation. Not systematically studied. Number of events not known.
		L _k		

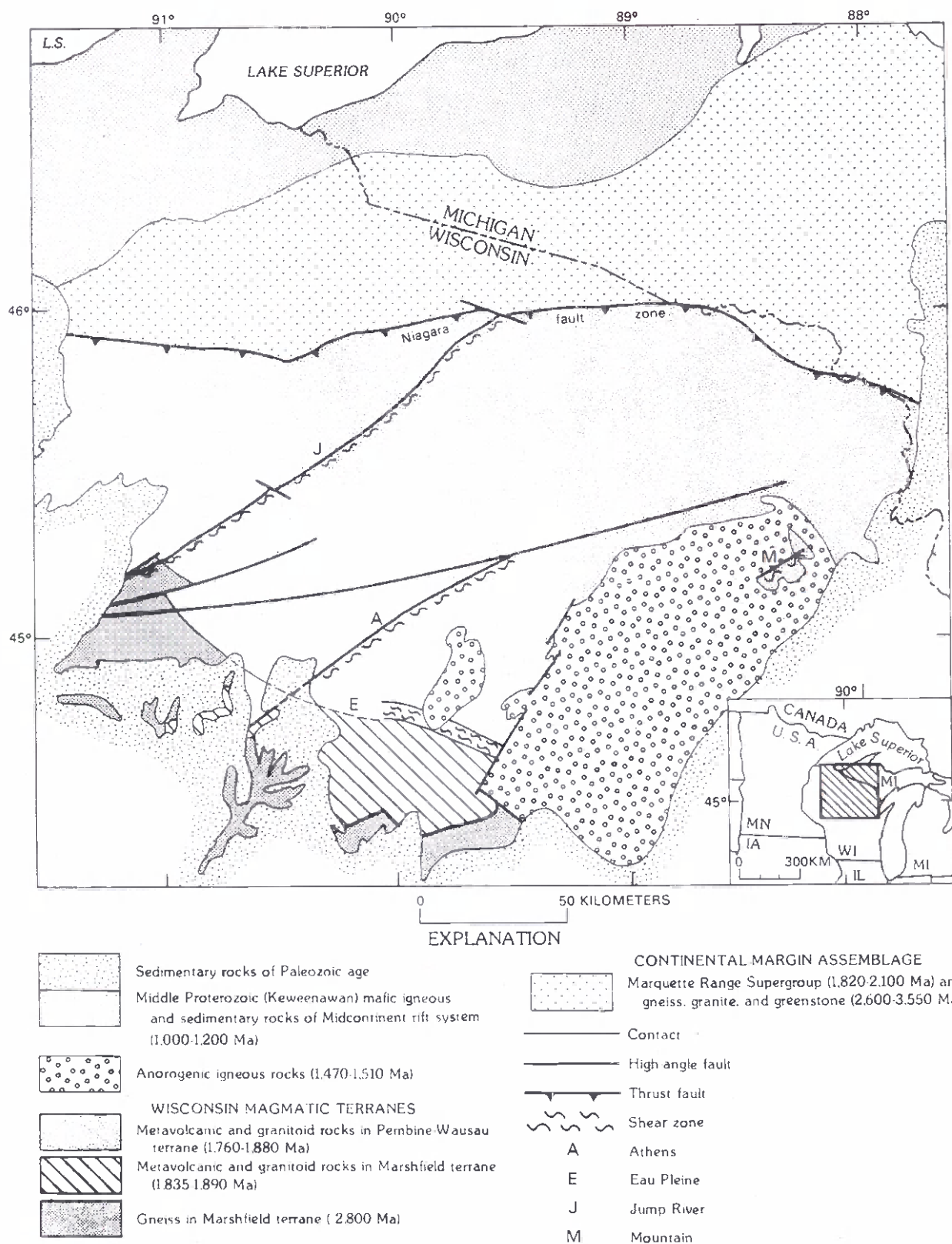


Figure 1. Generalized geologic map of the eastern part of the Lake Superior region, northern Wisconsin and Michigan. Note the location of the Mountain shear zone.

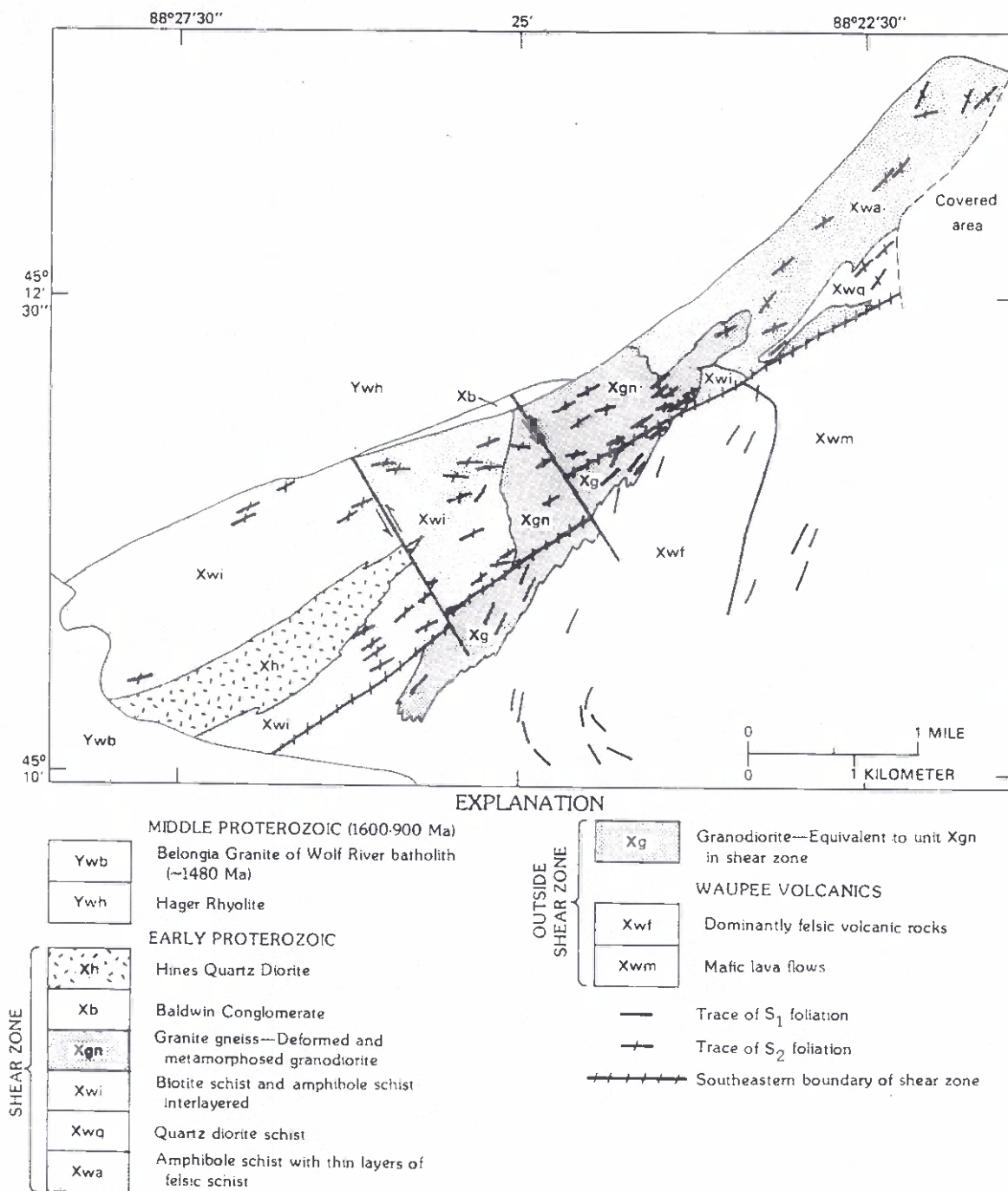


Figure 2. Geologic map of the Mountain shear zone and vicinity, Oconto County, Wisconsin

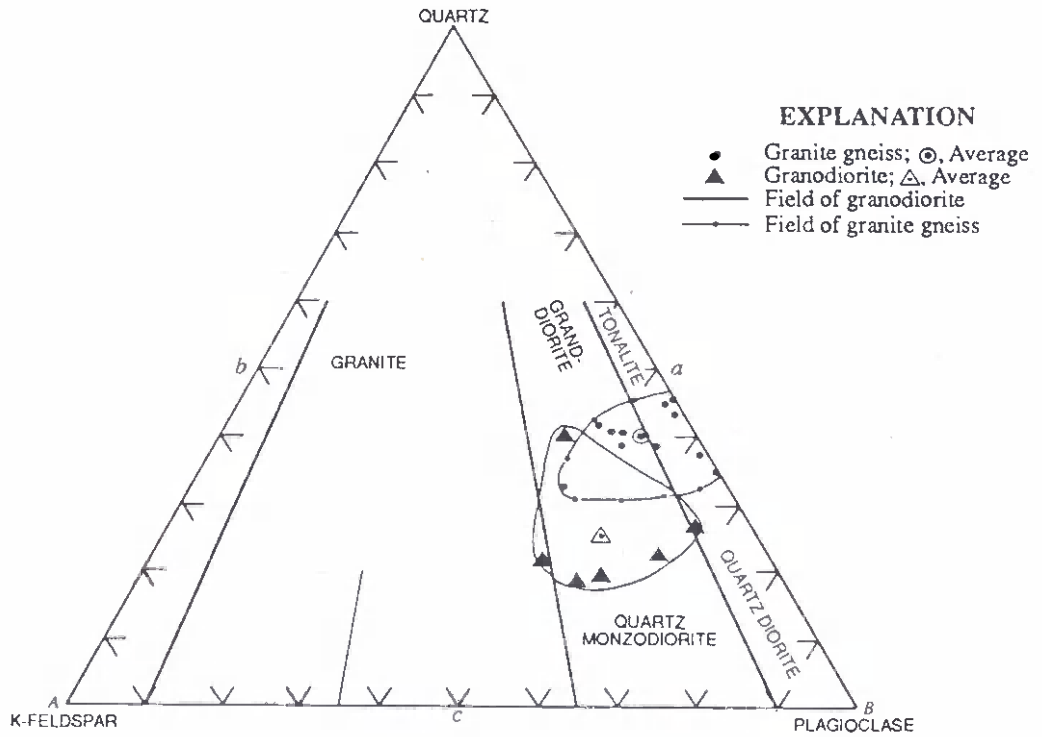


Figure 3. Quartz-K-feldspar-plagioclase diagram showing modal composition of samples from the granitic gneiss and granodiorite units

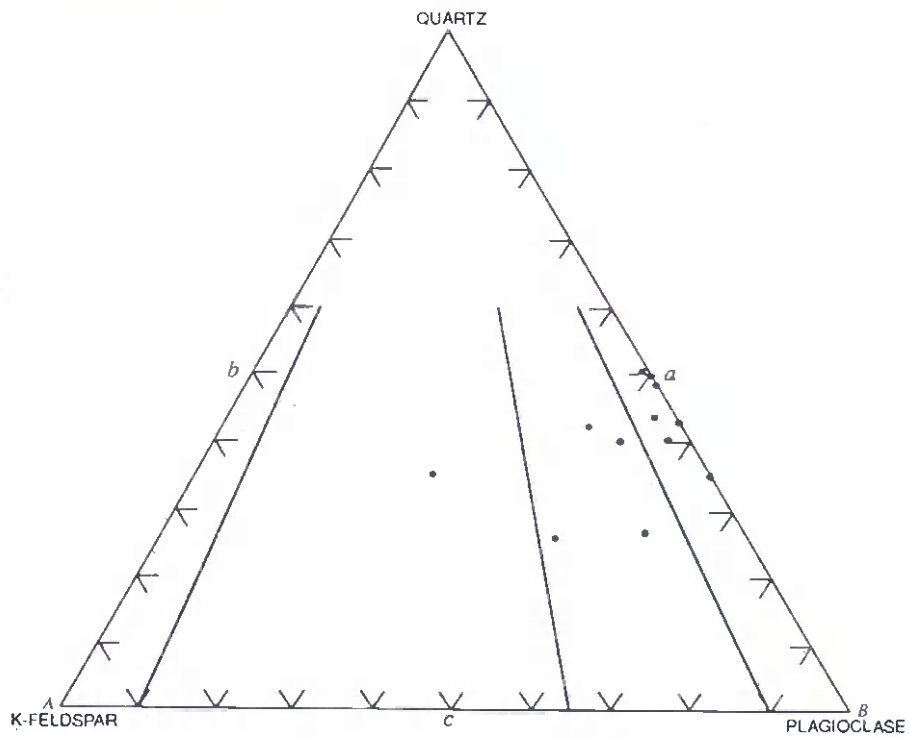


Figure 4. Quartz-K-feldspar-plagioclase diagram showing modal composition of samples from interlayered biotite and amphibole schist unit

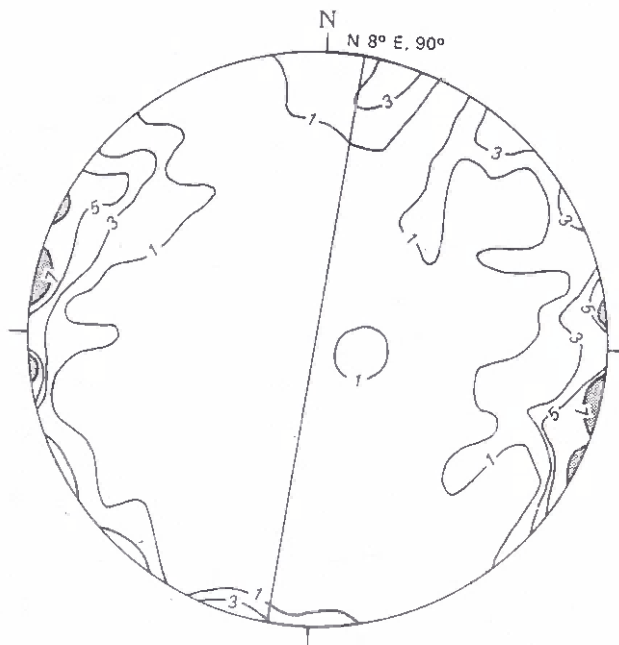


Figure 5. Contoured, equal-area projection of 68 poles to foliation outside of shear zone. Contours are 1, 3, 5, and 7 percent. The great circle (N. 8° E., 90°) is the calculated mean of S_1 foliation

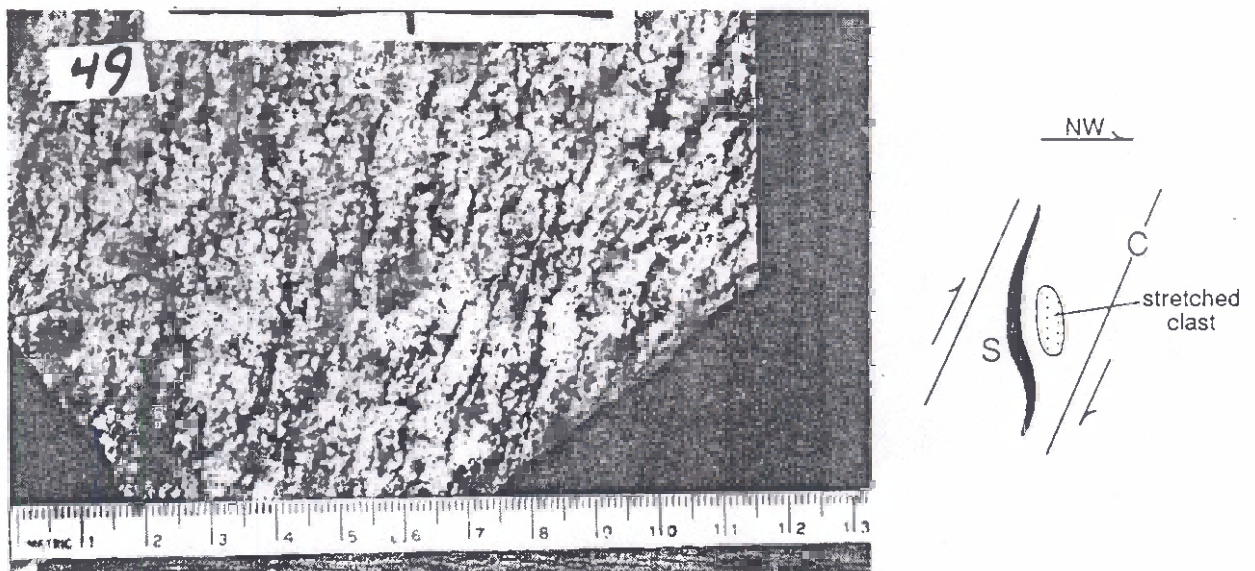


Figure 6. Photograph of granitic gneiss slab showing S-C fabric viewed parallel to the shear zone from the northeast. The direction of relative movement is shown diagrammatically adjacent to the photograph. The long axis of the stretched clast lies within the S plane

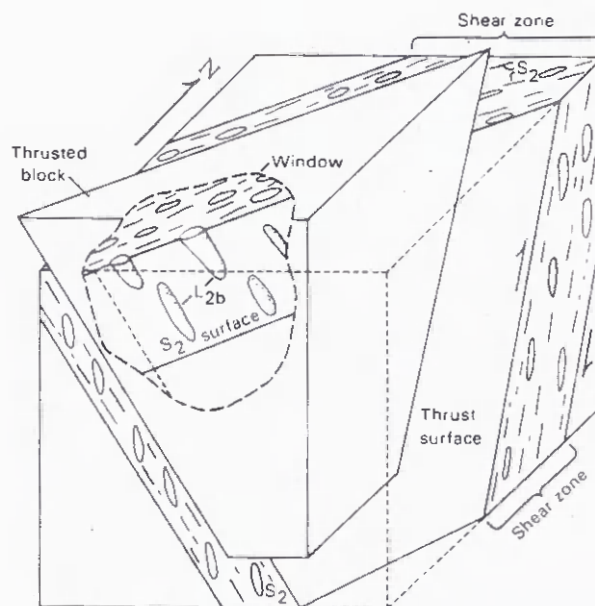


Figure 7. Block diagram showing relationship of strained clasts (strain ellipsoids) to simple shear envelope. Note that the stretch (long) axis of the clasts (L_{2b}) dip more steeply than the shear zone

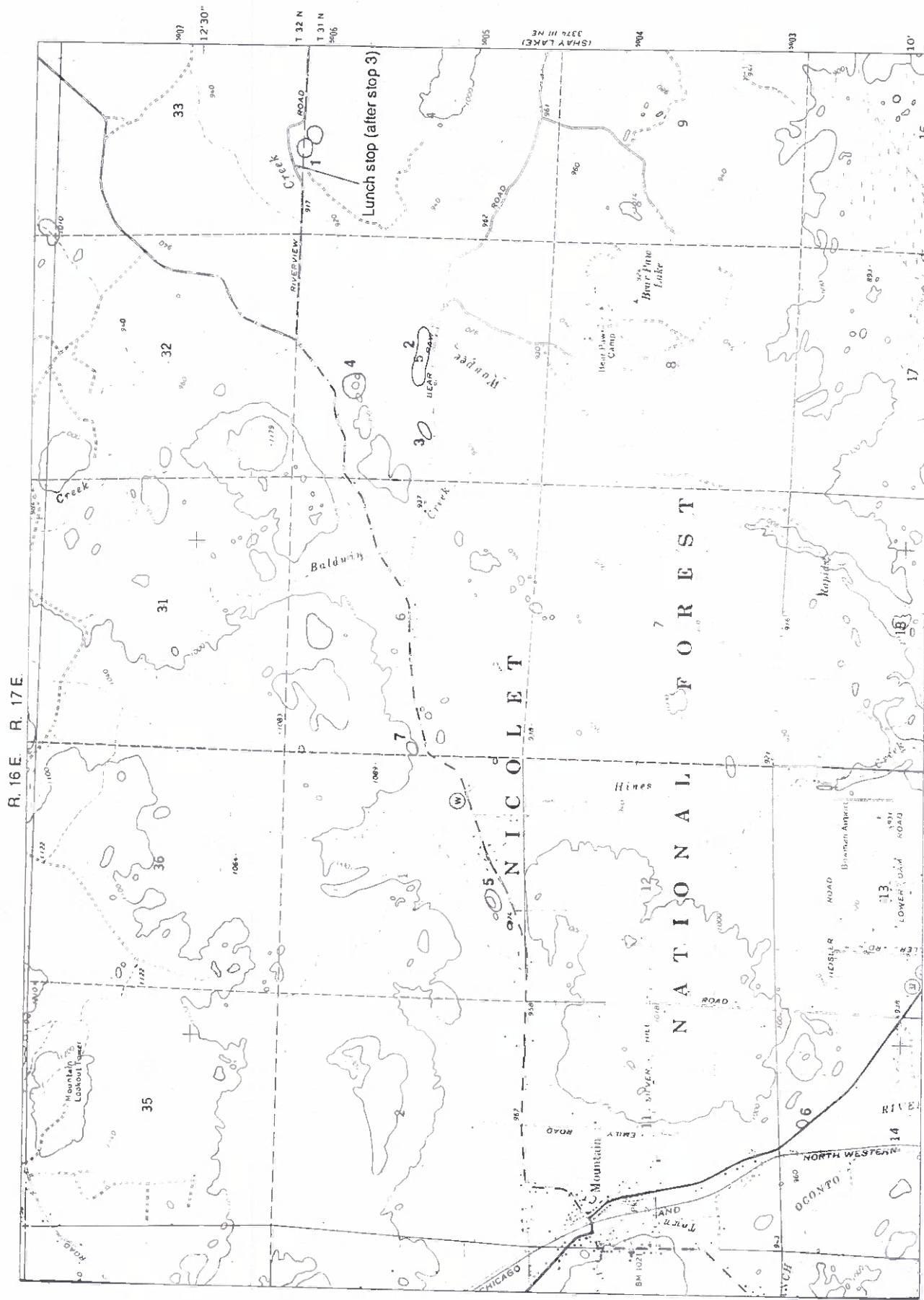


Figure 8. Map of part of Mountain 7 1/2-minute quadrangle showing field stops

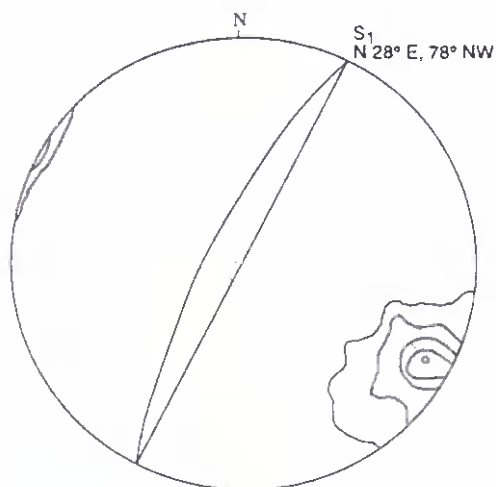


Figure 9. Contoured equal-area projection of 33 poles to S_1 foliation. Contours are 1, 16, 20 percent. The great circle is the calculated mean of S_1 foliation

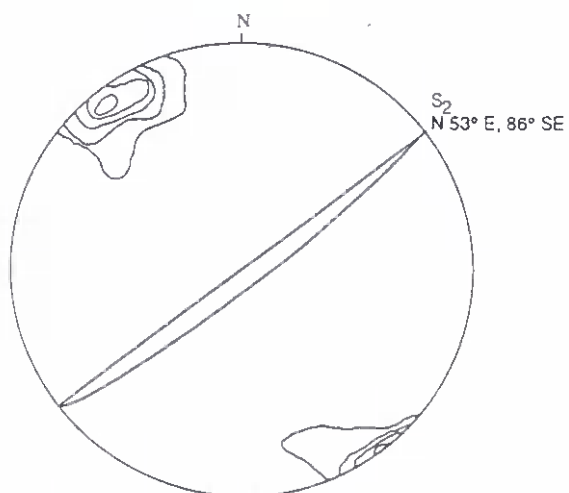


Figure 10. Contoured equal-area projection of 43 poles to S_2 foliation. Contours are 1, 10, 20, 30, 40 percent. The great circle is the calculated mean of S_2 foliation

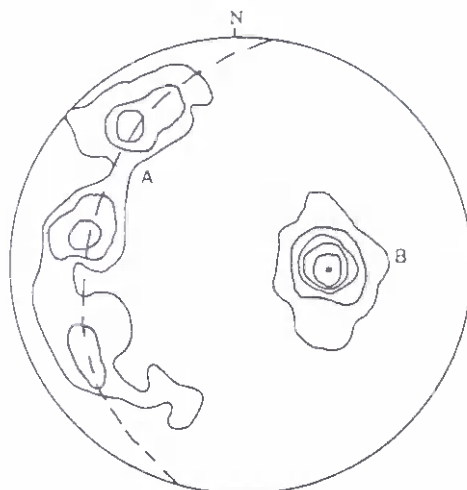


Figure 11. Contoured equal-area projection of 50 poles to S_2 foliation. Contours are 2, 6, and 12 percent; 50 points to long axes of deformed clasts (L_3) (contours are 2, 10, 20, 30, and 40 percent). A, girdle defined by folded S_2 foliation (F_3 folds); B, maximum concentration of lineations represented by dot

WEDNESDAY, MAY 1, 1991

FIELD TRIP #2

L.S. Chan, P.E. Myers, and R.L. Hay (Leaders)

**"Features and significance of the
Precambrian-Cambrian contact in western Wisconsin"**

Field Trip #2

Precambrian-Cambrian Boundary in West-Central Wisconsin

Lung. S. Chan, Paul E. Myers, and Richard L. Hay

INTRODUCTION

Geologists have realized long ago the predominance of gaps in the stratigraphic record. Recent developments in event stratigraphy and the introduction of the idea of 'catastrophic uniformitarianism' to geology have called for increasing effort on the study of geologic gaps and boundaries [Ager, 1981]. The Precambrian-Cambrian boundary is a very big topic. It represents a time when dolomite accumulation and stromatolite formation decreased, when red biogenic limestone and phosphorite deposits increased, and when living organisms abruptly became skeletalized [Rozanov, 1984]. Most studies of the Precambrian-Cambrian (PC-C) boundary are concerned with its stratigraphic aspects, as exemplified by the objectives of the Working Group on the PC-C Boundary (IUGS) [Crowe and Brasier, 1989]. Consequently, the major controversies about the PC-C boundary are focused upon its biostratigraphic definition and geochronometry. In the mid-continent, the PC-C boundary is generally represented by a gap almost 500 m.y. long. In most areas, the 0.9-1.1 Ga Keweenaw Event that affected most of the Mid-continent represents the last geological record in the Precambrian. A few intracratonic basins formed near Lake Superior while most of the mid-continent remained emergent above the sea-level during this vast interval of time. The absence of continuous stratigraphic sections across the PC-C boundary has perhaps limited the potential usefulness of the area for studying the major events and phenomena at the boundary. The non-stratigraphic aspects of the PC-C contact, however, still present many geological problems that may be important for the understanding of the geological processes and history of the boundary.

One of the problems is the age and manner of the peneplanation process. The PC-C unconformity in west-central Wisconsin is represented by a gentle surface with regional dip less than 1°. The smooth morphology of the PC-C contact presumably resulted from prolonged weathering and peneplanation. The geology of the surface, however, is highly variable. Precambrian crystalline rocks below the unconformity are weathered to a thick saprolite in some areas, but may be unweathered or partially weathered in others. The formation and condition of the Precambrian surface, the age of the saprolitization, the spatial variations in the degree of weathering of the Precambrian basement, and possible pre-transgression uplift of the mid-continent, are among the problems awaiting further examination.

The alteration profile developed on Precambrian rocks by weathering prior to the Mt. Simon Sandstone has been modified by potassic diagenesis over a wide area of the mid-continent [Duffin, 1989]. The authigenic potassic minerals are K-feldspar and illite-rich mixed-layer illite-smectite (I/S). The K-feldspar occurs principally as a replacement of primary igneous feldspar, and the I/S may be chiefly a reaction product of kaolinite with potassic fluids. Potassic alteration of granite is exhibited at Stop 1 (Chippewa Falls) and Stops 3 and 4 (Neillsville). Authigenic K-feldspar is also widespread in the Mt. Simon Sandstone, and authigenic K-feldspar from the Illinois Basin has given a K-Ar date of 394 ± 6 Ma [Duffin et al., 1989]. Dates of 390-400 Ma have been obtained from the authigenic K-feldspar of Ordovician tuffs of Minnesota and Iowa [Hay et al., 1988], suggesting a widespread K-feldspar episode in the early Devonian. Basinal brines have been postulated for the K-feldspar episode, and oxygen-isotopic data suggest that the K-feldspar formed at elevated temperatures. Illitization episodes at 350-375 Ma (late Devonian-early Mississippian) and 250-275 Ma (Permian) have been documented by K-Ar dating of I/S in Ordovician tuffs and the Mt. Simon Sandstone [Hay et al., 1988; Duffin et al., 1989].

The K-feldspar event may have important implications for the lead-zinc mineralization in the Upper Mississippi Valley. Lead isotope data for some Upper Mississippi Valley ores reveal Pb-leaching event at possibly early Devonian time [Doe et al., 1983]. The similar age of the potassic diagenesis suggests a possible link between the hydrothermal event and the Pb-Zn mineralization. Duffin et al. [1989] suggested that the fluids in the basal sandstone aquifers could have interacted with the Precambrian granitic basement and the paleosols. The nature of the Precambrian-Cambrian surface, thickness and distribution of the paleosols, and the fracture pattern of the PC-C surface could have played an important role in guiding the fluid flow and in the ore mineralization.

Late Proterozoic to Late Cambrian also represents a time of substantial latitudinal migration and climatic changes of the paleocontinent. Paleomagnetic evidence shows that the center of the paleocontinent must be located at high latitudes during the Late Proterozoic and at equator during the Late Cambrian [e.g. Irving, 1979; Van der Voo, 1981]. Late Keweenawan age glacial deposits in Lake Superior area are also evidence of continental glaciation during the late Proterozoic [Murray, 1955]. Despite the large horizontal motions, tectonic activity was relatively quiet in this part of the paleocontinent and was represented mainly by formations of gentle arches and highlands. Why was there such a long interval of apparent lack of tectonogenesis? How did paleozoic formations manage to maintain an almost perfect horizontality? What was the geometry of the epeirogenic blocks? How did the drastic climatic changes and continental glaciation affect the morphological development of the Precambrian surface and the paleosols? We may have to resolve to geophysical means to answer some of these questions. However, field observations can undoubtedly provide important constraints and tests on geophysical models and hypotheses.

We have outlined above several questions related to the Precambrian-Cambrian unconformity in the mid-continent. The traditional focus of the Institute on Lake Superior Geology has been Precambrian geology and ore mineralization. It is our hope that this field trip to the PC-C boundary outcrops in western Wisconsin could promote discussion and encourage research on Paleozoic geology of the Lake Superior region.

PRECAMBRIAN TERRANES

Numerous papers on the Precambrian geology of the Lake Superior region have been published in recent years. The correlation of the Precambrian rock sequences in the Lake Superior area is best summarized in three U.S.G.S. professional papers by Morey and Van Schmus [1988], Denison et al. [1984], and Harrison and Peterman [1984]. The most comprehensive descriptions of the Precambrian outcrops in west-central Wisconsin are given in two field trip guidebooks by Myers et al. [1980] and Maass and Van Schmus [1980]. LaBerge and Myers [1984], and Sims et al. [1989] presented two different syntheses on the Precambrian magmatic terranes in west-central Wisconsin.

Based on structure and metamorphic facies, LaBerge and Myers [1984] identified two major Proterozoic metamorphic sequences in west-central Wisconsin: a younger greenschist facies calc-alkaline sequence showing northeasterly trending fold-axes and an older amphibolite-facies quartz-feldspathic gneiss and amphibolites. The spatial extents of the two Proterozoic successions roughly coincide with the tectonostratigraphic terranes postulated by Sims et al. [1989]. The latter model, however, contends that the two early Proterozoic terranes are separate entities. The Eau Pleine Shear Zone between Ladysmith and Mosinee that separates the Marshfield tectonostratigraphic terrane in the south from the Pembine-Wausau Terrane in the north probably represents the paleosuture between the two terranes [Sims et al., 1989]. The Marshfield Terrane supposedly contains mafic to felsic volcanic sequences that formed on an Archean basement. All the field stops on this trip are located in the Marshfield Terrane.

South of the town Neillsville near the Black River, Precambrian bedrock consists primarily of Archean gneissic rocks and early Proterozoic metamorphic and plutonic rocks [Myers et al., 1974]. At Lake Arbutus near Hatfield, the sub-Cambrian consists of high grade metamorphic rocks or gneiss and migmatite (Stop 5). The principal rock types are an interlayered sequence of quartzo-feldspathic gneiss and amphibolite complex [Maass and Van Schmus, 1980]. Uranium-lead dating on two gneiss specimens has yielded ages of 2.5 and 2.8 Ga [Sims et al., 1989]. No Archean rocks are known north of Neillsville. In the Chippewa Valley area, layered amphibolite and feldspathic gneiss form the core of the eastward plunging Penokean orogen [Myers et al., 1980]. Recent zircon dating has yielded an age of 1835 ± 10 m.y. [Sims et al., 1989].

In both Archean and Proterozoic rocks, cataclasis played a major role in the metamorphism and deformation throughout the region. At least two generations of fold deformation and shearing can be observed in the Archean feldspathic gneiss exposed below Lake Arbutus Dam [Maass and Van Schmus, 1980]. In the Chippewa amphibolite complex, cataclasis is manifested by development of interlensing foliation, transposition of originally discordant structures such as dikes and xenoliths, and transposition of fold fragments into tectonic xenoliths [Myers et al., 1980]. The amphibolite terrane is flanked by volcanic-sedimentary provinces that are less deformed and metamorphosed. The nature of the contact between the amphibolite terrane and the volcanic-sedimentary terrane is uncertain. Both the Archean and Proterozoic terranes were synkinematically intruded by tonalitic and gabbro plutons between 1880 and 1820 m.y. [Sims et al., 1989]. In many locations, such as Lake Wissota, Little Falls, and Neillsville, rocks that belong to the amphibolite complex are present in generally foliated plutons as xenoliths.

Late Precambrian dikes of the Keweenawan age cut all the other rocks with strong discordance. Most Keweenawan-age dikes have an ENE strike. Prolonged pre-intrusion erosion is evident since the mafic dikes often contain chilled margins indicative of intrusion at shallow depth and high thermal gradient across the dike margins [Myers et al., 1980]. Two sets of dike orientation can be observed. Paleomagnetic results from central Wisconsin dike swarms reveal episodic dike intrusions during the Keweenawan event [Chan, Part 1 of Proceedings].

PRECAMBRIAN-CAMBRIAN UNCONFORMITY

The PC-C boundary is represented by a highly variable surface in the mid-continent area. The morphotectonic characteristics of basement surface in the Midwest were described by Rudman et al. [1965] and Thwaites [1980]. In west-central Wisconsin, the PC-C surface forms an extensive planation surface with a regional dip of less than 1 degree. Archean iron formations and Proterozoic quartzites form isolated monadnocks on the peneplain. Some of these monadnocks were several hundred feet in elevation. The peneplain was mantled by a layer of paleosols as much as several hundred feet thick [Cummings and Scrivner, 1980]. In some areas, Cambrian rocks directly rest upon barren, moderately weathered Precambrian formations.

Considering the low paleolatitude of the continent during the Cambrian, deep weathering of the Precambrian surface must have occurred before the Upper Cambrian deposition. The Precambrian basement, however, shows variable degree of weathering. Where Cambrian formations are in contact with Precambrian plutonic rocks, the Precambrian basement is generally deeply saprolitized to kaolinite-rich soils [Cummings and Scrivner, 1980; Stop 1]. In Neillsville, for example, spheroids over 6 m across developed in weathered granites; Mt. Simon sandstone and conglomerate fill the wedges among the spheroids [Stops 3 & 4]. In some areas such as Big Falls in Eau Claire County or Rock Dam in Jackson County [Stop 2], Cambrian sandstones rest upon Proterozoic amphibolite and metarhyolites that are only partially weathered. In the Jackson County Iron Mine, Upper Cambrian Mt. Simon and Eau Claire formations are represented by angular breccia and sandstones deposited

on the slopes of the iron formation monadnock [Stop 7].

The variable degree of weathering of the PC-Є contact has raised questions about the timing of the planation process. It is unquestionable that the PC-surface was already a gentle and smooth surface with isolated monadnocks by Late Cambrian time. The chilled margins of mafic dikes indicate prolonged erosion even during the Proterozoic time. The relatively unweathered Precambrian outcrops in some localities, however, suggest relatively immature weathering or intense erosion and removal of weathered material during the transgression event. Büdel [1957] presented a model on the formation of peneplains which consists of two concurrently active leveling surfaces: an exposed wash surface at the top of the deep saprolite and a buried basal surface of weathering at the bottom in contact with unweathered bedrock. The limited weathering in some sub-Cambrian profiles suggests a very late-stage development of the peneplain or a rapid denudation of the wash surface shortly before the deposition of the Mt. Simon Sandstone. The latter theory points to the possibility of Cambrian uplift or eustatic movement before the Dreisbachian transgression.

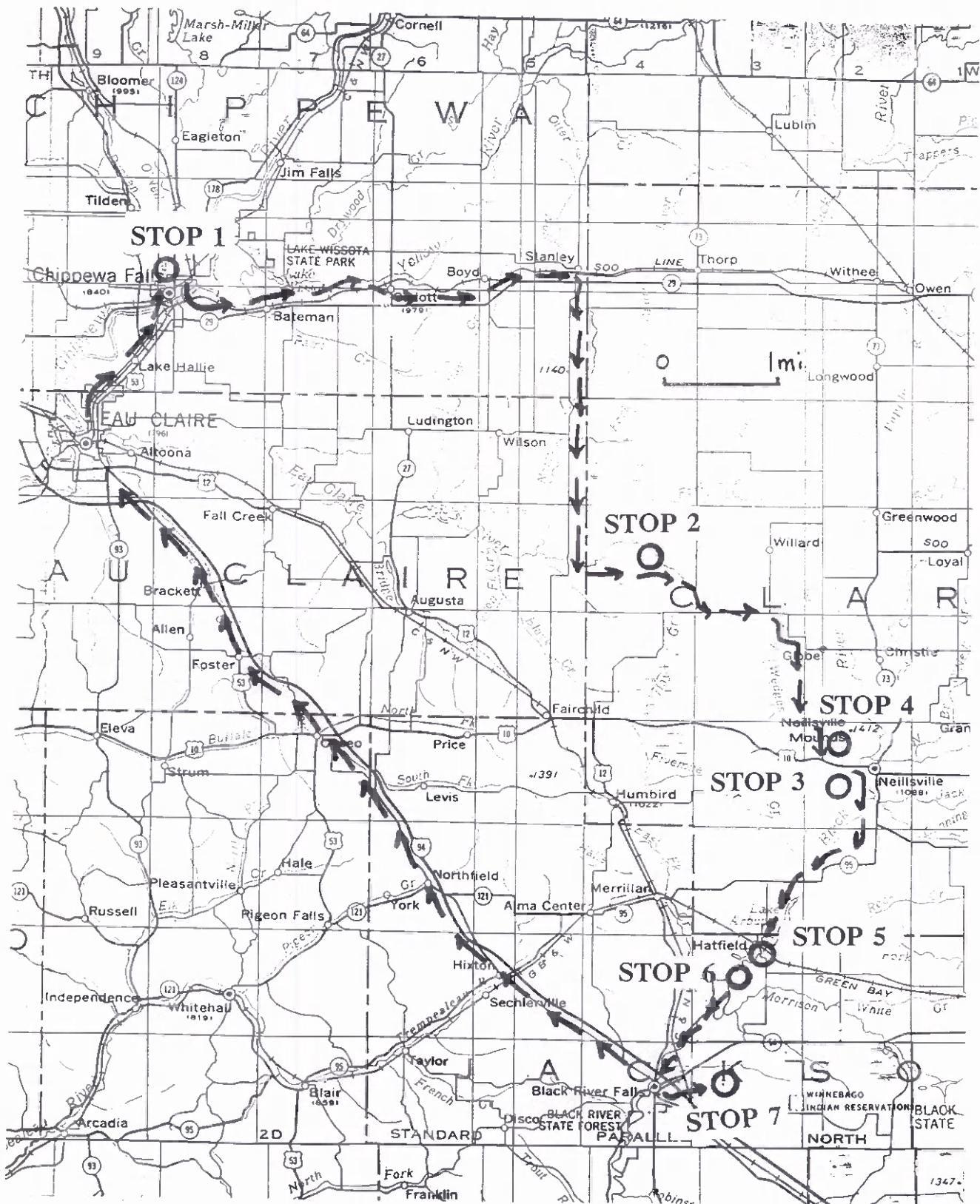
UPPER CAMBRIAN MOUNT SIMON FORMATION

The sediment above the unconformity consists largely of Upper Cambrian Mt. Simon formation deposited on the mid-continent region of North America during the Dreisbachian transgression. The Mount Simon formation is a fine-coarse-grained, poorly sorted, quartz arenite with a local basal conglomerate. In the Upper Mississippi Valley, the thickness of the formation varies from 180 to 40 meters. Lochman-Balk [1971] gave a summary on the Cambrian paleogeography and paleoecology of the North America craton. Previous studies by Thwaites [1923], Ostrom [1964, 1966, 1967], Asthana [1969], and Ostrom [1988] have outlined the general stratigraphy and paleodepositional environment of the local sequences in west-central Wisconsin.

In the Chippewa Valley area, the lower part of the Mount Simon formation is a coarse-grained, thick-bedded quartz sandstone and conglomeratic sandstones. The relief of the deposition was probably very shallow; sedimentary channels are less than 1 meter deep. The presence of trace fossils (*rusophycus*?), and planar and bipolar cross-bedding suggest a littoral or shallow marine tidal flat environment of deposition for the lower part of the formation. The upper part of the Mt. Simon formation contains feldspathic sandstone with small-scale ripple bedding, brachiopod fragments, and worm trails (planolites). Iron-oxide concretions are commonly present. Some concretions are as large as 1.5 cm in diameter. The grain size data obtained by Asthana [1969] indicate an unimodal grain size in the lower part of Mt. Simon, a bimodal distribution in the middle and the upper parts. The grain size distribution shows a generally fining upward sequence. The upper part of the formation represents a subtidal shallow marine deposition. The presence of 10-20 mm diameter iron-oxide concretions and horizontal worm trails suggest a much broader transgression toward the top of Mt. Simon Formation. Presence of brachiopods and ripple bedding in the upper part of Mt. Simon Formation is diagnostic of a shallow-marine subtidal deposition, probably above the wave base at a depth of 10-15 m. The part of the section containing iron-oxide concretions may represent a slightly deeper water than the sandstones containing brachiopod fragments.

There are a few questions concerning the origin and source of the detrital grains and the age of the diagenesis. Asthana [1968] reported a large range in K-feldspar content in the formation. Qi and Chan have also identified several lithotypes in the formation [Suppl. notes in this field guide]. The age of the potassic diagenesis in the Mt. Simon formation in Wisconsin is still uncertain. Further mineralogical and stable isotope studies will be conducive to our understanding of the sediment source, the diagenetic environment, and the regional tectonics at the time of deposition.

Field Trip 2: Route Map and Stop Descriptions



STOP 1: GEOLOGY OF IRVINE PARK, CHIPPEWA FALLS, WISCONSIN

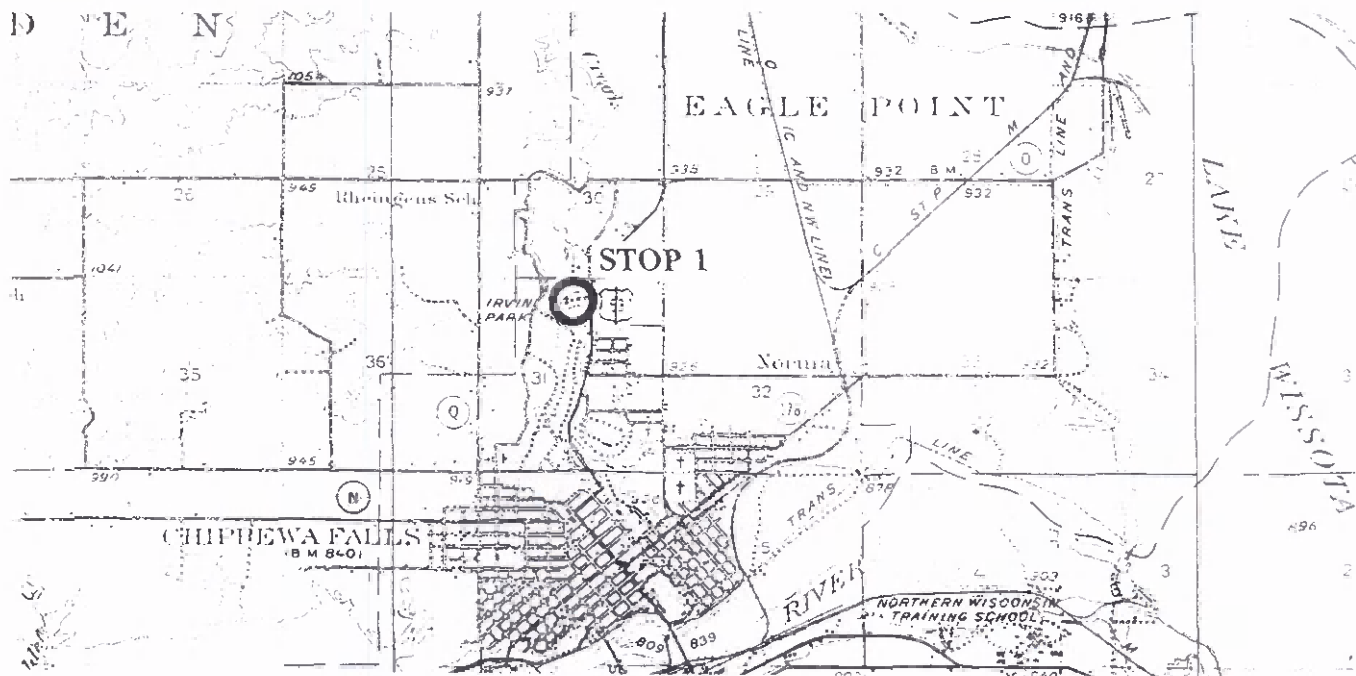
Location: NW1/4 SW1/4 Section 31 T29N R8W; Chippewa Falls 15' Quadrangle

Summary of Features: Mt. Simon basal conglomerate on Precambrian saprolitized granite

Middle Proterozoic biotite trondhjemite showing spheroidal and saprolitic weathering is unconformably overlain by feldspathic conglomerate and sandstone of the Mt. Simon Sandstone of Late Cambrian age. The configuration of the Precambrian surface was mapped by Thwaites [1957]. The Precambrian-Cambrian unconformity at Irvine Park is described in detail by Ostrom [1978, 1987].

The Precambrian at Irvine Park is composed of a deeply saprolitized trondhjemite. Myers et al. [1980] reported that the trondhjemite at Irvine Park is probably part of a larger granitic pluton which occupies the Chippewa Valley between Eau Claire and Jim Falls and that it is part of a complex WNW-trending granite-amphibolite complex of Middle Proterozoic age. The trondhjemite at Chippewa Falls was dated by Van Schmus [1980] at $1,840 \pm 15$ m.y.

Saprolite is a "soft, earthy, clay-rich, thoroughly decomposed rock formed in place by chemical weathering of igneous and metamorphic rocks" [AGI, 1972]. It is characteristic of areas of prolonged tropical to subtropical weathering on a granitoid bedrock surface of low relief. Saprolites in the Piedmont Province of United States attain a thickness of over 100 m. The saprolite at this outcrop contains high clay content. The feldspar and quartz grains tend to be quite equidimensional and possess sharp edges and corners. The saprolite becomes increasingly altered upward until at the



unconformity. It is composed mostly of quartz, clay minerals and chlorite. Was the saprolite formed before or after deposition of Mt. Simon Sandstone? Evidence suggests that most of the weathering and saprolitization preceded Mt. Simon Sandstone deposition, although there is also evidence of later modification.

A comparison of the clay mineralogy, heavy mineralogy, and clast texture classification in the saprolite and overlying sandstone reveals a strong provenance relationship between the two units. Cummings and Scrivner [1980] indicated that most of the clasts over 1.0 mm in the saprolite are composite grains composed of quartz and K-feldspar, whereas the coarser size fraction in the Mt. Simon Sandstone is composed predominantly of quartz grains. The clay minerals, according to Cummings and Scrivner [1980], are 80-90% kaolinite and 10-20% illite and vermiculite, and maroon color of the saprolite at this locality is a result of more recent weathering of chlorite. Reddish spheroids of altered trondhjemite are found in the lower part of the saprolite. Plagioclase and part of the microcline are replaced by K-feldspar and lesser phyllosilicate. This is an excellent example of the regional potassic alteration of Precambrian rocks below the Sub-Cambrian unconformity.

The Mt. Simon Sandstone thickens from about 35 m at Chippewa Falls to about 270 m in southern Wisconsin [Asthana, 1969]. The northernmost exposures of this formation are near Ladysmith and Conrath, Wisconsin, and probably represent the approximate original limit of Mt. Simon deposition in this region. Ostrom [1964], in a detailed regional synthesis of the Cambrian stratigraphy of Wisconsin [1964], concluded that the Mt. Simon Sandstone formed in a marine littoral and nearshore shelf environment by a slowly transgressing sea, which migrated SE-NW over a deeply eroded Precambrian surface. It has been suggested that the basal conglomeratic sandstones of the Cambrian sequence were best interpreted as subtidal shelf or tidal-channel deposits. The Mt. Simon-Eau Claire sequence was regarded as progradational. Driese [1979], based on sedimentary structures and paleontological evidence, argued that the Mt. Simon Sandstone is "largely progradational, shoaling- and fining-upward tidal sequence." According to Driese [1979] the sandstones at Irving Park belong to the lower and middle units of the Mt. Simon Sandstone.

Asthana [1969] showed that the Mt. Simon Sandstone averages 18% feldspar of which 81% is K-feldspar; the feldspar content ranges from 2.85 to 40.07% and that its abundance decreases generally upward in the formation. Authigenic orthoclase occurs commonly as rhombic overgrowths on detrital grains. A petrographic description of several specimens of the Mt. Simon formation is given in the supplementary section [Qi and Chan, Suppl. 1]. Cummings and Scrivner [1980] compared the mineralogy and texture of the Mt. Simon Sandstone at Irvine Park with the underlying saprolite, and concluded that the silt and clay fractions of both show considerable similarities. The saprolitized trondhjemite could have been a source of clasts for the Mt. Simon Sandstone. The coarser fractions of the Mt. Simon Formation are devoid of the composite quartz-feldspar grains characteristic of the underlying saprolite. The sandstone also contains abundant epidote and garnet not found in the underlying saprolite. These minerals, however, are abundant in the amphibolites nearby (e.g. Jim Falls) and could have been derived from them.

Based on the work of many geologists, it appears that the Precambrian basement was deeply weathered to form a saprolite in Late Precambrian time (between 900 and 500 m.y.). The saprolite was locally preserved during slow transgression of the Late Cambrian seas and deposition of the saprolite resulted in their incorporation as interstitial materials in the Mt. Simon Sandstone. With more recent exposure of the lower Paleozoic formations to ground water circulation, some alteration of chlorite to iron oxides occurred.

IRVINE PARK

CHIPPEWA FALLS

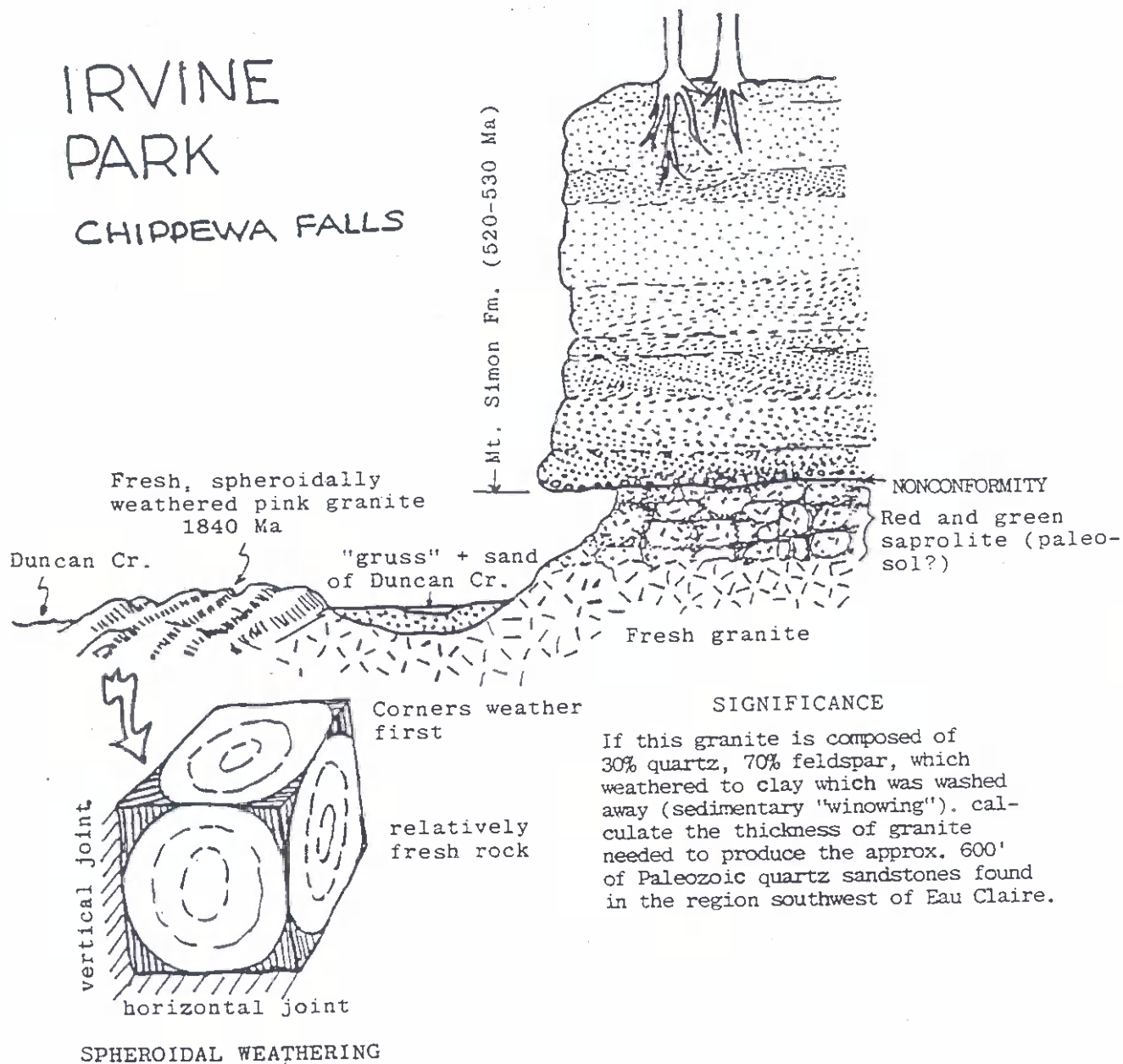


Fig. 1. Precambrian-Cambrian unconformity at Irvine Park.

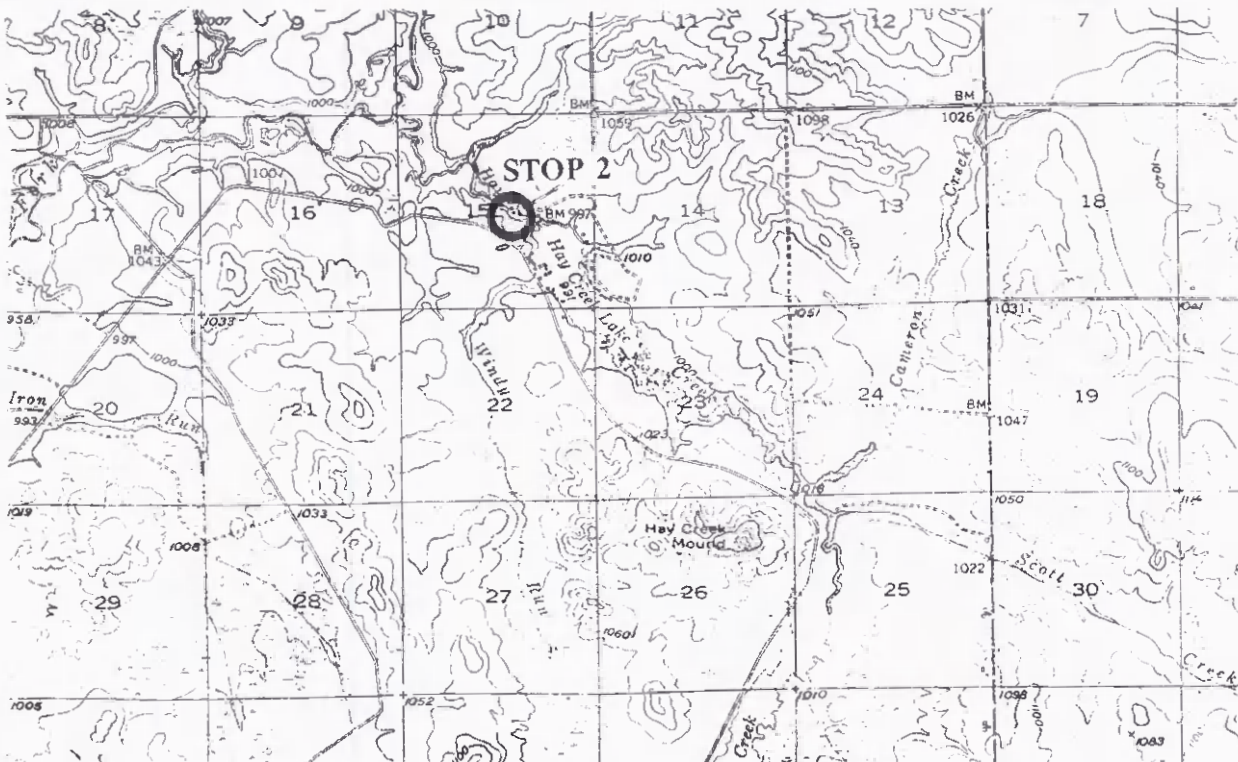
STOP 2: PRECAMBRIAN-CAMBRIAN CONTACT AT ROCK DAM, HAY CREEK

Location: Hay Creek, 0.15 mile downstream from Rock Dam; SE 1/4 NW 1/4 Sec 15
T26N R4W; Fairchild 15' Quadrangle

Summary of Features: Mt. Simon basal conglomerate on partly weathered
Precambrian metarhyolite

Conglomeratic Mt. Simon Sandstone overlies unconformably Penokean-age metamorphic rocks. Myers et al. [1980] contended that the Precambrian unit is a metarhyolite: a felsic mylonite containing lenticular clusters of strained quartz in a very fine matrix of K-feldspar, quartz and muscovite. In thin section, the clusters of strained quartz are 1.0-2.5 mm long, with a coarse-grained core that grades outward into a fine-grained matrix. Estimated composition is: K-feldspar 57%, quartz, 35%, muscovite, 3%, magnetite 2%, and biotite, 1% [Myers et al., 1980]. A muscovite-rich phyllite crops out near the base of Rock Dam. The rock is composed of K-feldspar, quartz, and muscovite. The foliation is N87W, 85N. Complementary joint sets at N18W, 73E and N77E, 8S account for the blocky appearance of the outcrops.

The basal Mt. Simon is a conglomeratic layer containing large pebbles of vein quartz and metarhyolite. Some of the pebbles are over a foot across. The contact surface with the underlying metarhyolite shows a local relief of about 5 meters. The Precambrian rocks do not appear to be deeply weathered in this particular outcrop.



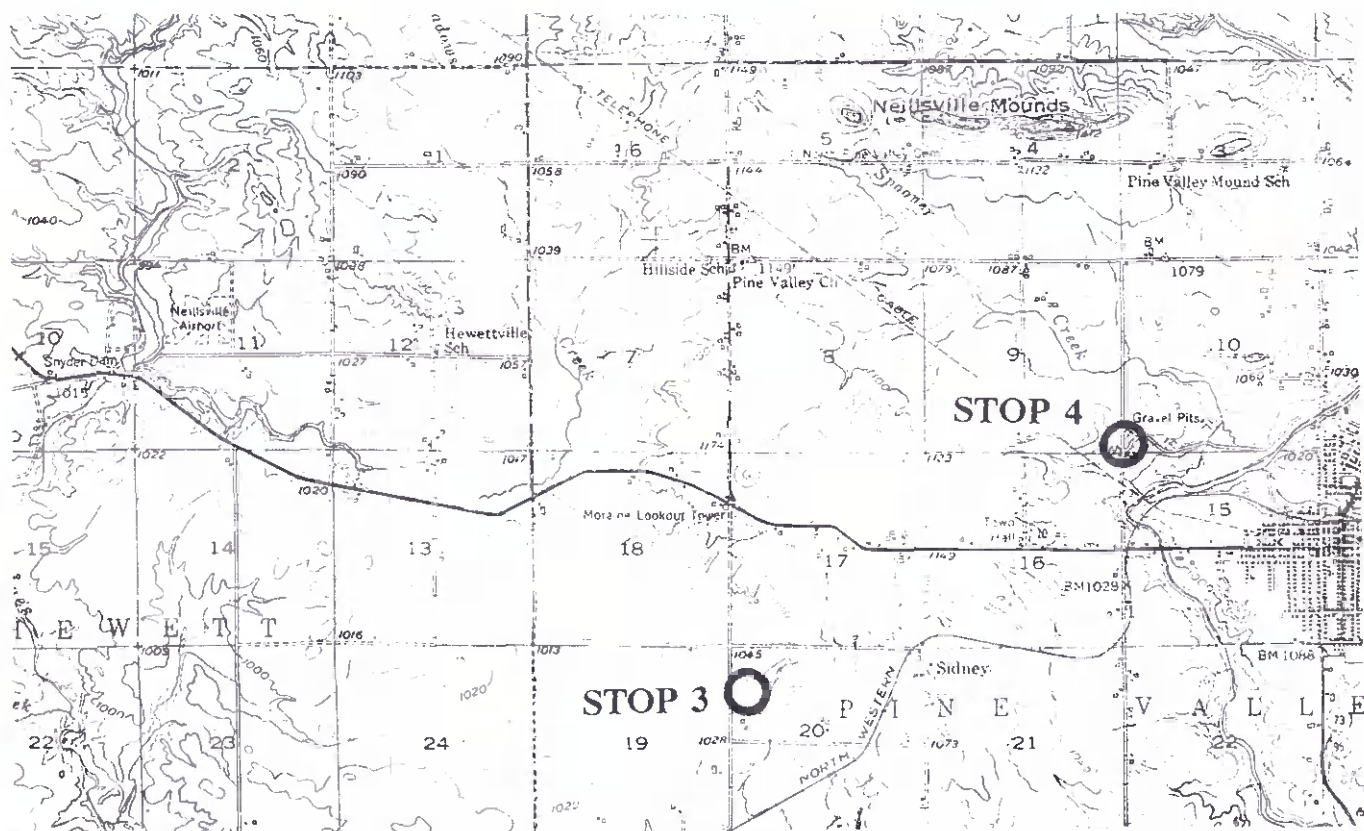
STOPS 3 & 4: SPHEROIDAL WEATHERING IN NEILLSVILLE GRANITE QUARRIES

Location: Stop 3 NW 1/4 NW 1/4 SEC 20 T24N R2W; Neillsville 15' Quadrangle

Stop 4 SE 1/4 SW 1/4 SEC 10 T24N R2W; Neillsville 15' Quadrangle

Summary of Features: The Precambrian-Cambrian boundary displays a spectacular spheroidal weathering features in the Neillsville granite in two quarries in the Neillsville area.

The Neillsville granite, located in the Black River Valley, is a Penokean pluton. A description of its petrography and composition is given by Maass and Van Schmus [1980]. Zircon dating yields an age 1875 ± 25 m.y. [Sims et al., 1989]; the age is about 40 Ma older than other similar Penokean plutons in the area. The granite is composed of subequal amounts of quartz, microcline, and plagioclase (An 26-28); and approximately 2% biotite [Maass and Van Schmus, 1980]. Fresh rock is gray and altered rock is pink, both because of pyrite weathering to ferric oxide and replacement of igneous feldspar by authigenic K-feldspar. A strong set of lineation trends N45E and plunging 82° pervades the granite. The granite contains xenoliths of mica schists which display at least three generations of deformation. Two sets of crenulation form an oblique angle of about 30°. A third set of lineation aligns in the same direction as the lineation in the granite. The northern end of the quarry is a mixture of granite and granodiorite which contain similar structures. The nature of the contact between the two units is unknown.



The Neillsville granite is exposed in several old quarries at Stop 4. In the largest quarry, the sandstone on top of the PC-Є boundary is about 10 meters thick and overlain by recent glacial outwash. The spheroidally weathered granite is spectacularly exposed in this quarry. The PC-Є boundary is at about 1070' elevation. The spheroids are sometimes over 6 meter across. The granite is identical to those found at Stop 3. Relatively unweathered corestones ranging from 1 m to over 6 m have been isolated by zones of intense weathering where grus formed. The granite shows a foliation that strikes N45W and dips 30°SW. A sub-planar fabric in the grus strikes N58W and dips 16E [Nelson, 1980].

In both quarries, the Precambrian surface weathers into a clay-rich spheroidally weathered pinkish granite. Sandstone and conglomerate can be found on top of and in the wedges between the weathering spheroids. The granite/Mt. Simon unconformity has a few feet of relief, which follows the tops of the granite spheroids, and coarse material has filled the wedges between the spheroids. The clasts in the basal conglomerate are entirely quartz pebbles. A dacite dike cut across the Neillsville granite in the northern end of the quarry. The dike is composed of quartz, oligoclase, biotite and minor microcline. Microscopic deformation and recrystallization textures are nearly identical to those in the granite and granodiorite [Maass and Van Schmus, 1980]. The weathering sequence for minerals in the Neillsville Granite determined by Nelson [1980, unpubl. work] suggests that plagioclase in the saprolite is weathered into a clay. Biotite mica is similarly weathered by hydrolysis to chlorite. Illite-rich I/S of diagenetic origin is the principal clay mineral in the saprolite at Stop 3. Here K-feldspar replaces primary microcline and plagioclase in pink bodies of granite within the saprolite. In the quarry at Stop 4, exfoliation sheets of the spheroids consist largely of authigenic K-feldspar, and primary igneous textures have been largely obliterated by diagenetic alteration superimposed on pre-Mt. Simon weathering. Potassic alteration extends to the quarry floor in fractured area, ~10 m below the sub-Cambrian unconformity.

The weathering and saprolitization at the two outcrops occurred before the deposition of the Mount Simon Sandstone. Uplift and erosion occurred after the plutonic emplacement. The time of the uplift is uncertain. The presence of corestones not entirely converted to grus suggests either an immature weathering, which implies a short duration between the exposure of the surface and the Dreisbachian transgression, or denudation of substantial portion of the saprolite by erosion, which may suggest a pre-transgression uplift that led to rapid erosion of the weathered profile.



Fig. 2 Basal conglomerate of Mount Simon Sandstone in Neillsville quarry.

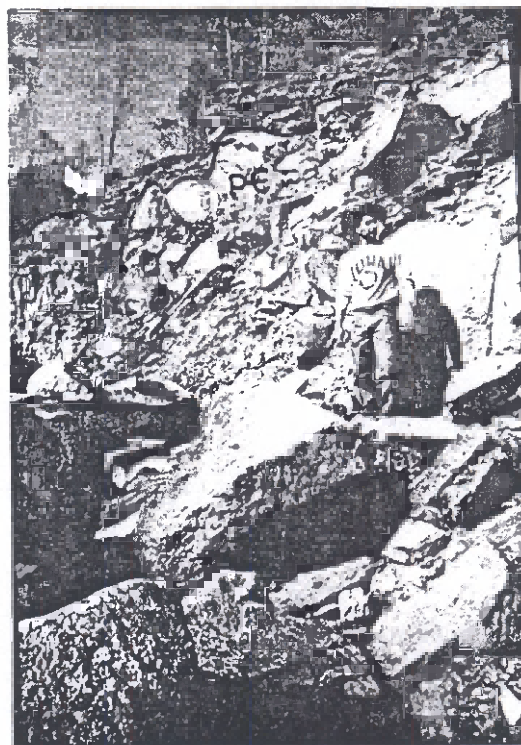
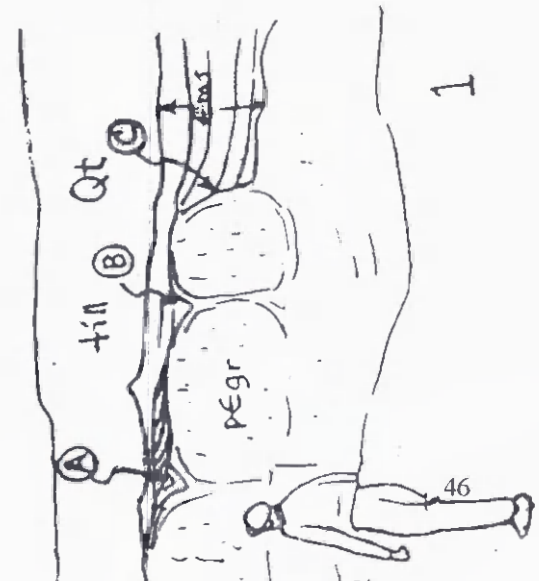
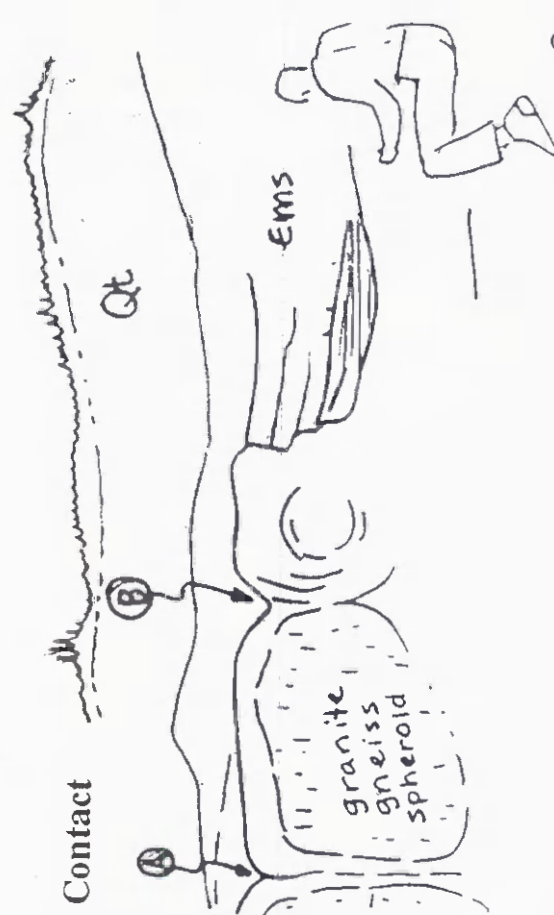


Fig. 3 Spheroidal weathering features in Neillsville quarry.

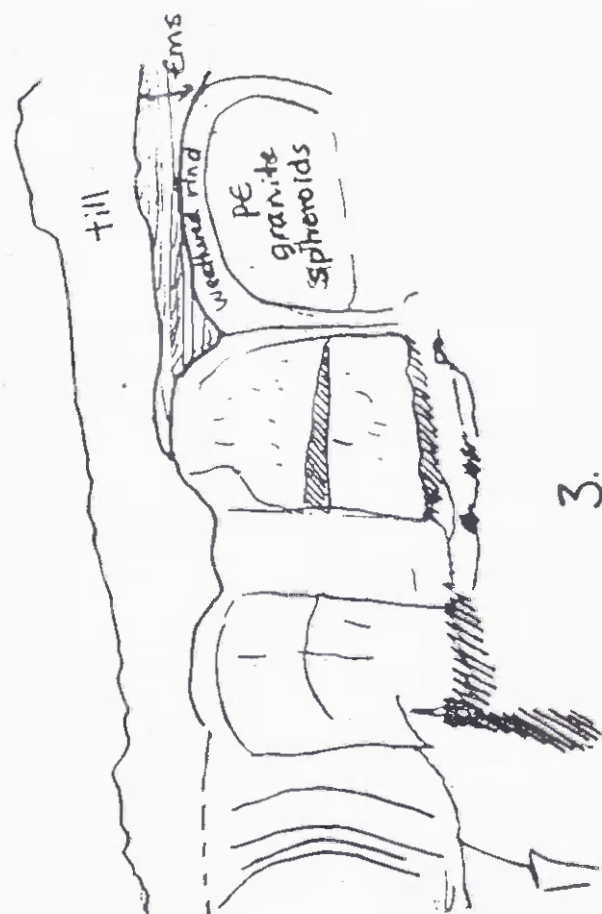
Neillsville Granite-Sandstone Contact



1



2



3

sketches from photographs.

PE granite gneiss with strong, steeply west-plunging lineation (75°) is spheroidally weathered in concentric layers 5-10 cm thick. The outermost rinds are bleached; inner ones are iron-stained. Cross-bedded quartz sandstone of the Mt. Simon Fm. was deposited in the depression between the spheroids. (A) and (B) a steep, SW-dipping surface of one spheroid is overlain by Mt. Simon Sandstone at (C). The sandstone is overlain by orange-brown cobble till on the south side of the exposure. The till directly overlies weathered granite to the north.

STOP 5: LAKE ARBUTUS DAM

Location: SE 1/4 SEC 3 T22N R3W, Hatfield 7½" Quadrangle

Summary of Features: Weathered Mt. Simon Sandstone overlying Archean gneisses.

In the Hatfield and Lake Arbutus area, glacial outwash generally overlies directly on top of Precambrian rocks. Archean rocks are well exposed below the Lake Arbutus Dam. The Archean rocks are interlayered sequence of quartzo-feldspathic gneiss, amphibole gneiss, and metarhyolite that formed about 2.8 Ga ago [Van Schmus and Maass, 1980]. The feldspathic gneiss shows a gneissic banding ranges from 0.1 to 3 cm thick and contains principally quartz, plagioclase and microcline [Van Schmus and Maass, 1980]. The rocks underwent at least two generations of fold deformation; the first one produced a foliation which was refolded about an easterly fold axis plunging at about 40°. The metavolcanics were sheared to form a phyllonite. Several mafic dikes intrude the gneissic and phyllonites. Some of the dikes show a strong foliation, indicative of a probable synkinematic origin.

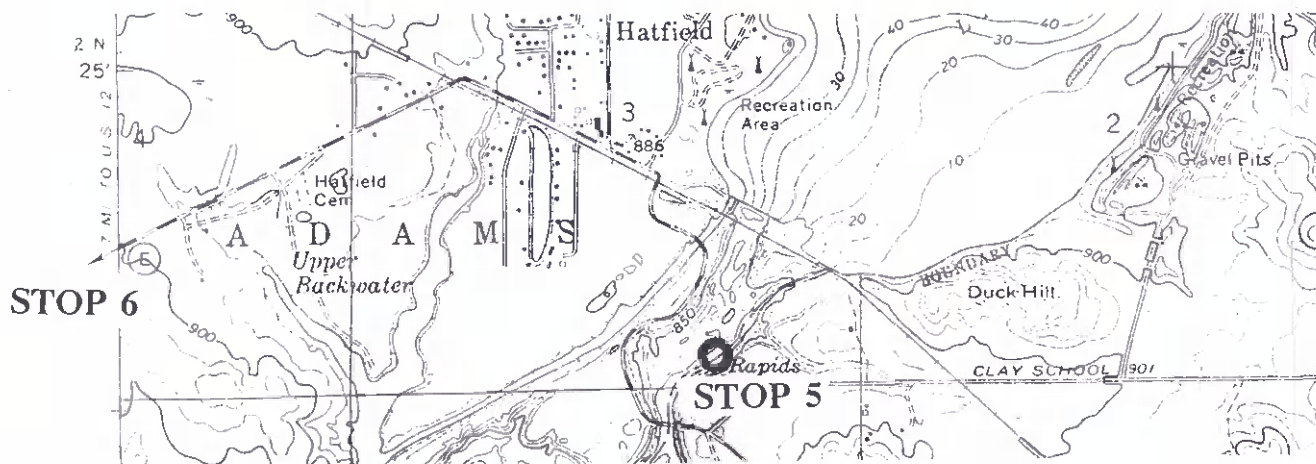
The boundary between the Mt. Simon Sandstone and the Archean rocks is present about 400 ft. south of the Dam. The boundary, however, is presently covered by talus, but could be located to within 15 feet. The upper 20 feet of the Archean rocks are partially weathered. A 6-10' section of Mt. Simon is exposed on a talus slope above the Archean metavolcanics. The sandstone is deeply weathered and crumbles easily in hand specimens, but relict sedimentary structures like lamination and cross bedding are still visible. On the lake side of the Lake Arbutus Dam, Mt. Simon sandstone is exposed at water level.

STOP 6: HALLS CREEK

Location: NW 1/4 SE 1/4 SEC 19 T22N R3W

Summary of Features: Mt. Simon conglomeritic sandstone over Precambrian metavolcanics

The PG-C boundary is exposed along a meander about 300' from Hwy E. The boundary is located at about 2' above the water level and may be covered by a thin layer of gravel and talus. The Mt. Simon basal layer is a conglomeratic sandstone. The Precambrian is a weathered metavolcanics of unknown age (Proterozoic?). The weathered rocks do not appear to contain iron oxide.



STOP 7: PRECAMBRIAN-CAMBRIAN SURFACE IN JACKSON COUNTY IRON MINE

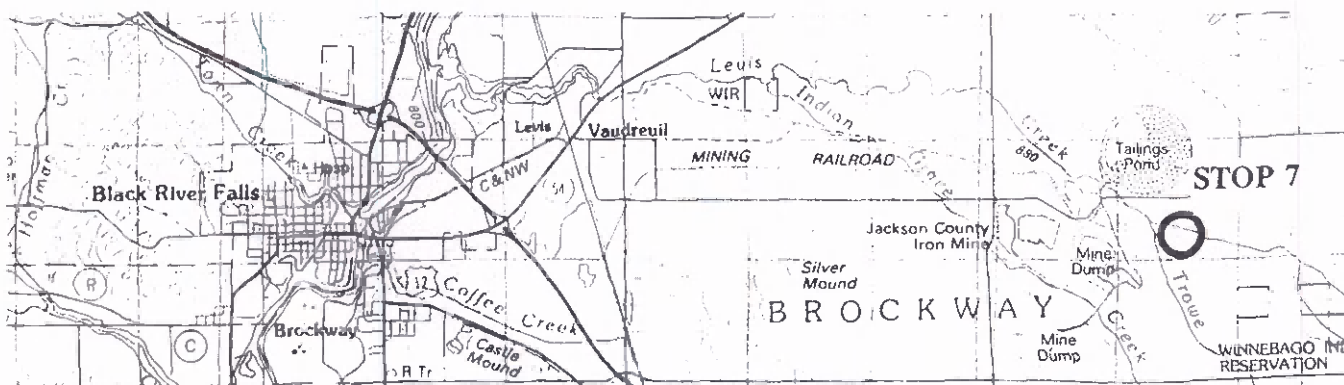
Location: NE 1/4 SE 1/4 SEC 15 T21N R3W; U.S.G.S. 1:50,000 Jackson County Map

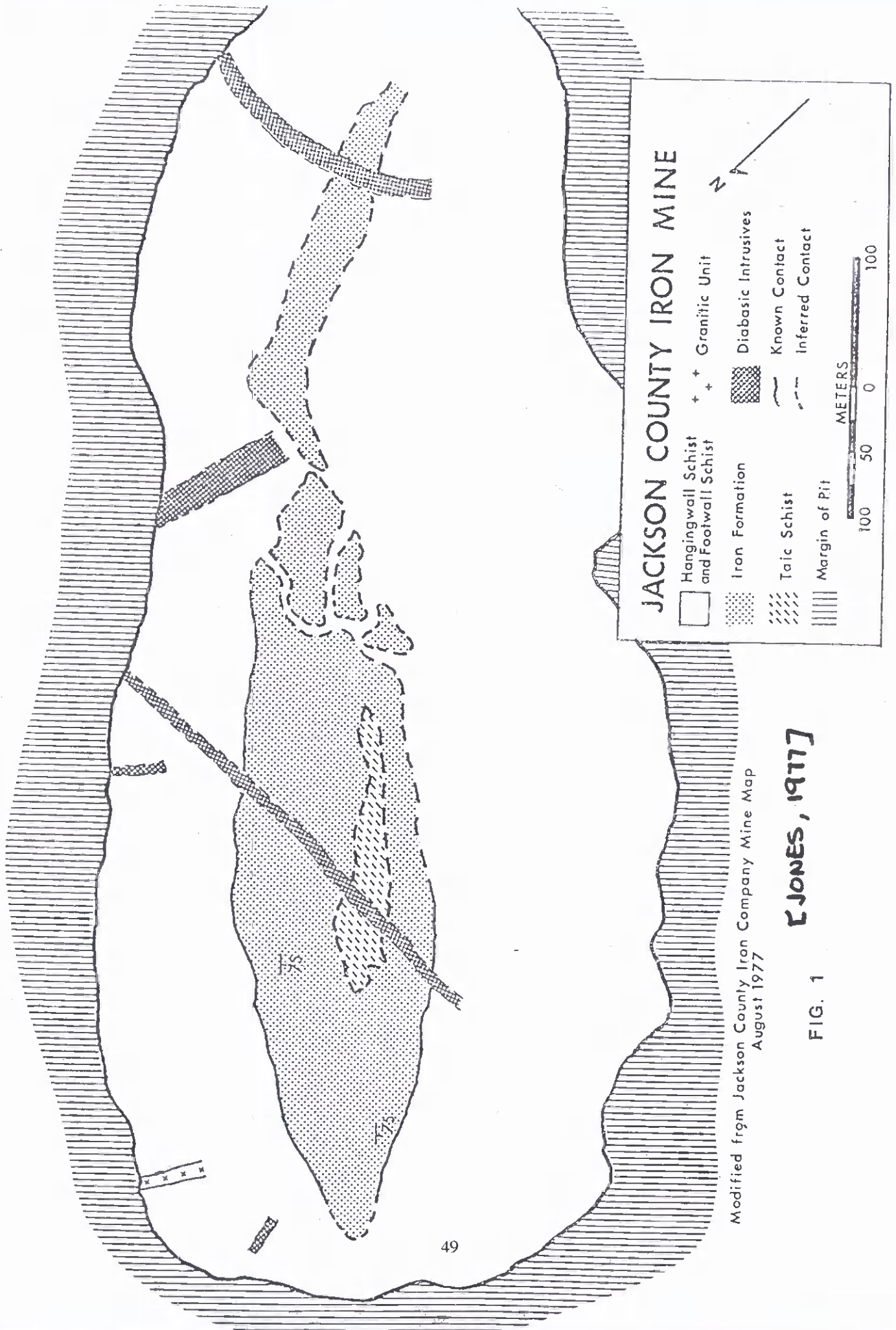
Summary of Features: Precambrian iron formation and quartz feldspathic gneiss overlain by Upper Cambrian Eau Claire formation containing a basal breccia layer.

The Jackson County Iron Mine located about 6 miles east of Black River Falls was an open pit taconite operation active during the early eighties. The ore body is a lens-shaped 915 m x 550 m body [Jones, 1977, Unpublished work]. The primary ore mineral is magnetite residing in Archean iron formation which is embodied in quartz-feldspathic schist and gneiss. The iron formation is composed almost entirely of quartz, magnetite, cummingtonite-grunerite, biotite, garnet, calcite, ferroactinolite, Ca-rich hornblende, sphene, apatite, pyrite, and K-feldspar. The iron-formation is highly deformed and exhibits prominent compositional bands ranging from a few mm to several meters. The compositional bands often formed isoclinal folds and were transposed along foliation. The iron mound has been completely excavated and the open pit is at present filled with over 200 ft. of water. Fragments of the iron formation could be found on the slopes of the pit. The rocks surrounding the iron mound are upper greenschist-amphibolite facies metamorphic rocks. The principal rock types are quartz-biotite-plagioclase schist, mica-chlorite-andalusite schist, mica-garnet-staurolite schist, and amphibolite gneiss. Quartz veins over 1 ft wide are present.

The contact between the Archean and Paleozoic formation is an undulatory surface. The Archean rocks are generally weathered. The depth of the weathered zone varies from 6 to 50 meters, probably controlled by fracture patterns [Jones, 1977]. The Paleozoic rock immediately above the unconformity is a framework-support breccia containing angular to subangular fragments of iron formation, schists, and quartz pebbles. Fragments from the iron formation are tabular in shape and occasionally imbricated. Quartz fragments, some of them over 1/2 m in diameter, tend to concentrate in certain levels. Upper Cambrian glauconitic siltstone and shale can be found above the breccia layer. Mudcracks, iron concretions, and worm trails are commonly present in the argillaceous layers. The rock is lithological similar to the Eau Claire formation which overlies stratigraphically the Mount Simon Sandstone.

The iron formation probably formed a monadnock during the late Proterozoic time. The basal breccia layer represents a slump deposit on the slopes and base of the monadnock. Since the Eau Claire formation is in direct contact with the iron formation, this particular outcrop was probably located at a relatively high part of the slope.





Modified from Jackson County Iron Company Mine Map
August 1977

[JONES, 1977]

FIG. 1

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SUPPLEMENT 1: NOTES ON PETROGRAPHIC CHARACTERISTICS OF THE MOUNT SIMON FORMATION, EAU CLAIRE, WISCONSIN

Authors: Qi Zhu (Visiting Scholar from Chongqing Petroleum College, China) and Lung S. Chan (University of Wisconsin at Eau Claire)

We have conducted a petrographic study on Mt. Simon Sandstone samples collected from the type locality of the formation in Eau Claire. Nineteen specimens were collected at different stratigraphic levels of the formation. Based on the petrographic observation, we have identified the following types of lithologies in the formation. The approximate levels of the specimens and the mineralogical abundance are shown in Fig. S1.1.

Type I. Coarse-Fine Conglomeratic quartz sandstone (MS-1, Plate 1).

The conglomerate consists of 20% quartz pebbles and 80% quartz cobbles. Magnetic grains are occasionally present. The quartz grains show high degree of rounding and maturity. About 10% of the grains display dissolved edges. Individual quartz shows wavy extinction indicating a source from metamorphic rocks. About 20% of the detrital grains are polycrystalline and quartz-vein fragments. All the quartz grains are clear and structureless, with few impurity inclusions. Cementing material consists mostly of matrix material with 80% silt (0.1 - 0.05 mm) and 20% euhedral potash feldspar or irregular feldspar or kaolinite. Both detrital grains and cementing matrix are very well sorted.

The petrographic characteristics of the sandstone suggest a shoaling deposition with multiple cycles of erosion. Original or primary cementing materials are replaced entirely by authigenic potash feldspars. The matrix material suggests a multiple cycle of deposition, indicating strong wave or current motion and active pore current flow during the deposition.

Type II. e.g. MS-2. Quartz sandstone.

This type of quartz sandstone contains about 60% medium-fine grained sand and 40% medium grained sand and minute amount of silt size particles. The sand is 95% quartz and about 5% feldspar. The quartz grains are clean and monocrystalline; polycrystalline quartz and impurity inclusions are absent. The quartz sands are subrounded to subangular and moderately sorted. The feldspars are mostly authigenic minerals that develop around a detrital core of microcline or plagioclase. Most of the authigenic feldspars show euhedral form. Some of the authigenic feldspars tend to form suture-type intergrowth pattern in interstitial spaces among the quartz grains. Many feldspars develop around detrital feldspar grains. Detrital feldspars are normally not stained, microcline in composition while authigenic feldspar are well stained. Plagioclase twinning is observed in some of the detrital feldspar grains. The orientation in the authigenic feldspar often makes a 10-12° angle with the twinning orientation of the detrital core. Fragments of quartz veins and granites are present in minute amount. The cementing material also contains small quantity of silica. The same lithological characteristics are present in MS-3, 4, 5, 7, 8, 11, 12, 13, 14, and 15 (Plate 2). Feldspar composition in these specimens ranges from 5 to 15%, and normally not exceeding 10%.

Type III. Volcaniclastic-like feldspathic quartz sandstone

This type of quartz sandstone differs from Types I and II in that morphology and composition of the feldspars. This type of quartz sandstone contains 40-60% feldspar and shows a strong dissolution of the crystal edges. In MS-10 collected at about 15 m from the base of the Mt. Simon Formation, mica grains present in matrix material seem to show a preferred orientation, which may be due to strong current flow through the interstitial pore spaces during or after the deposition. We suggest that the presence of a strong interstitial pore current may be an important factor in the formation of authigenic feldspar.

MS-9 (Plate 3)

This specimen contains 50% feldspar and 50% quartz. About 20% of the quartz grains are subrounded, 30-40% subangular and 40% show dissolution edges. The dissolution grains show bone-like, curve or sharp shards. Feldspar grains are mostly fine sand size, and in the forms of single crystals. Most of the feldspars are subhedral to euhedral and form intergrowths. Many form around a core grain of dissolved residue and enlarge around the core. Other feldspars do not indicate a core grain. Cementing material are inconspicuous. The feldspars basically form the interstitial materials.

MS-10

The lithological characteristics of this specimen are similar to that of MS-9. Feldspar grains constitute about 50% of the specimen. It contains about 10% of coarse-grains pebble-size detrital quartz fragments with good degree of rounding and uniform composition, and about 2% sericite mica fragments.

Type IV. Feldspathic Quartz Siltstone

MS-19

This specimen contains 80% quartz, about 15% irregular-shaped quartz grains showing strong dissolution edges and 20% feldspar with or without a detrital core. Quartz grains are subrounded to subangular and moderately sorted. Cementing materials are mostly iron-rich mineral or sericite micas in interstitial pores.

MS-18

This specimen contains 80% silt and 15% fine silt with 70% quartz and >25% feldspar. Sericite and ferric mineral are present at about 2-3%. Quartz grains are mostly subrounded to subangular and well sorted. The feldspars are mostly secondary pore filler, crystal forms are less well developed as MS-8 or MS-9 and are similar to kaolinite cement. Part of the feldspars are authigenic and authigenic overgrowth. Feldspar abundance is also lower than in MS-9 and MS-10.

Based on the petrographic observations, the Mt. Simon formation in Eau Claire can be divided into several zones. The basal layer of the Mt. Simon Sandstone is represented by a conglomerate and coarse sandstone layers. The dominance of rounded quartz sand indicates high degree of maturity. It is also evident that the sandstone underwent multiple cycles of erosion and deposition. The detrital quartz grains and quartz cement show corroded edges that resembles quartz that formed in a volcanic setting. It is questionable whether the cementing material represents the primary original cementing medium. The clay content in the conglomerate is very small, suggesting a second cycle cementation process.

The Mt. Simon Formation contains several beds that are enriched in feldspathic composition. Based

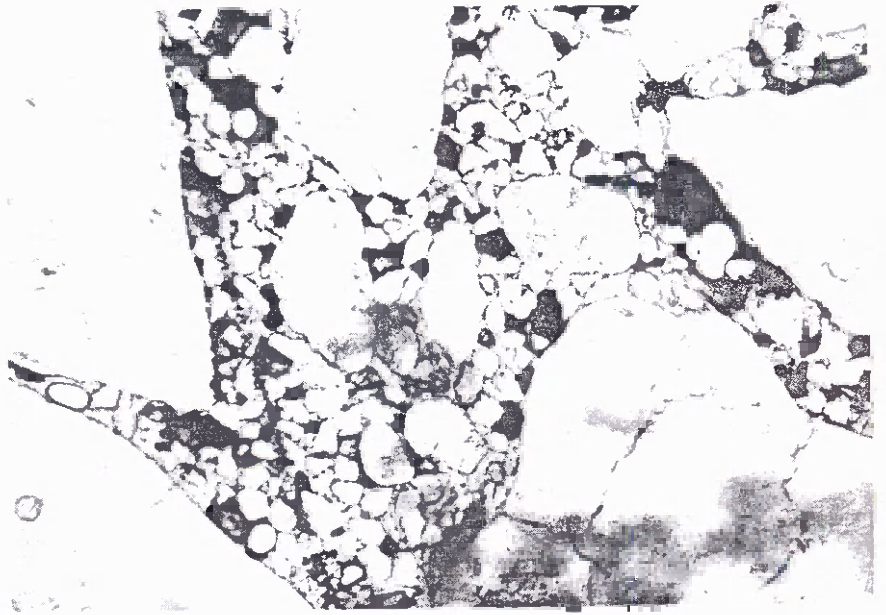
on the form and crystallinity, the potash feldspars are authigenic in nature. About 50% of the quartz grains, on the other hand, are terrigenous detrital grains derived there through normal transportation means. The other quartz grains, however, exhibit dissolved edges and form irregular shards. This form of quartz bears similarities to quartz that forms in a volcanoclastic environment.

The authigenic feldspar crystals and their formation conditions present a complicated problem. Some of the feldspars contain a core of dissolved feldspar grains. Others do not contain any core and are entirely authigenic in origin. Some of the core grains contain obvious twins, others do not. Some of the core grains are possibly dissolved quartz, with brainlike structures. Some authigenic feldspars simply follow the forms of existing detrital quartz and feldspar grains or fill-in cracks and veins in existing grains.

Most of the authigenic feldspars are euhedral to subhedral. When the growth condition is confined by surrounding quartz grains, the feldspar may form dog-teeth like crystal edges. The source of feldspars, morphology and authigenesis have presented some interesting questions. First, the source of the feldspar has been an enigma. The depositional medium and water conditions at the time of deposition also represent problems for understanding the origin of the Mt. Simon Formation. The high percentage of feldspar in particular beds of the formation may be indicative of the following conditions. (1) The transported debris may contain substantial amount of volcanic quartz fragments, as suggested by the forms of the quartz grains. (2) During the early deposition of the Mt. Simon sandstone, there might have been volcanic eruptions in a nearby area. The deposit surely contains quartz grains sufficiently washed and eroded by ocean waves. The primary cementing material may contain volcanoclastic material that were being eliminated during the redeposition. The eroded primary cement might be enriched in potassium feldspar and constituted the source of the authigenic feldspar in the Mt. Simon Formation. (3) The observed dissolution of quartz is probably related genetically to the formation of the authigenic feldspars and may imply an volcanoclastic event at the time of the deposition.

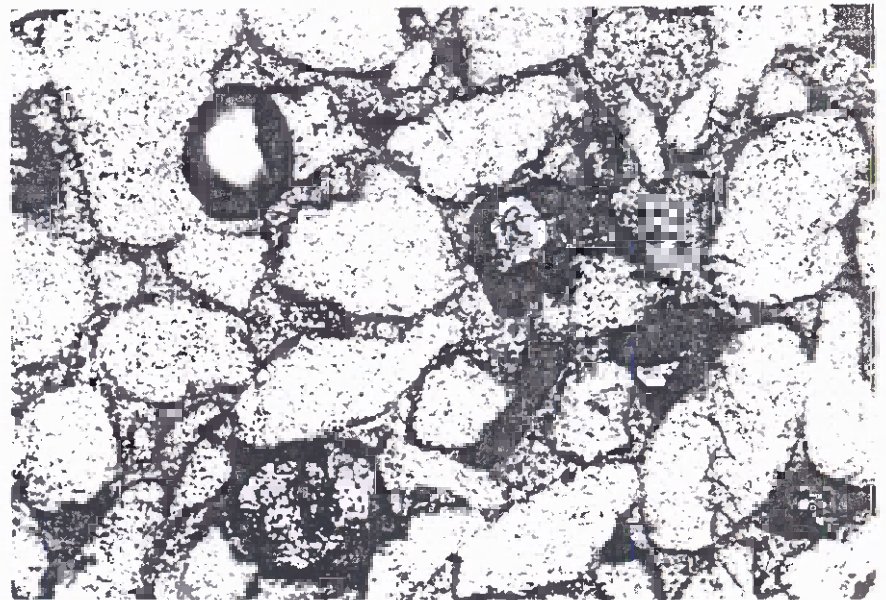
Type I. Conglomeratic
Sandstone

MS-1.
Sig. 4x10



Type II. Quartz Sandstone

MS-3
Sig. 10x10



Type III. Volcaniclastic-like
feldspathic sandstone

MS-9
Ri. 10x10

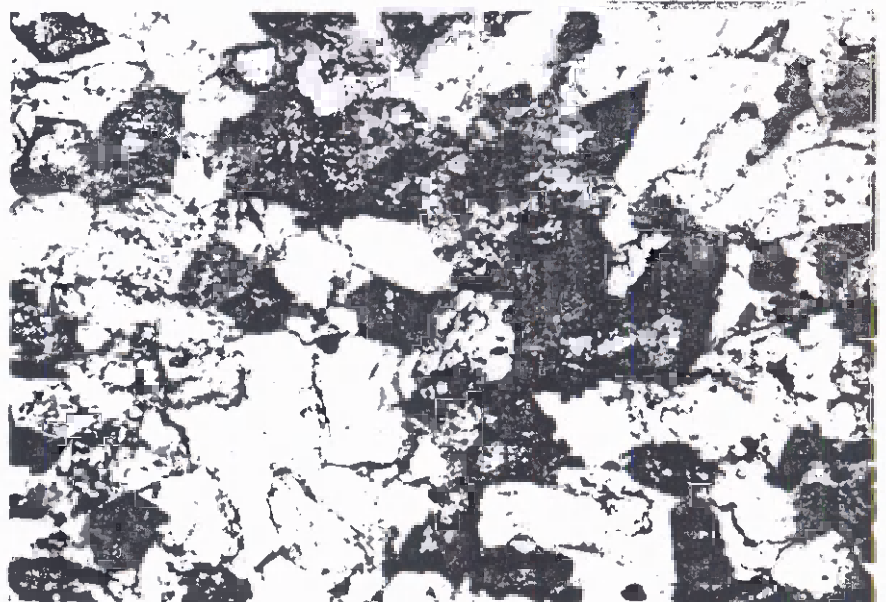
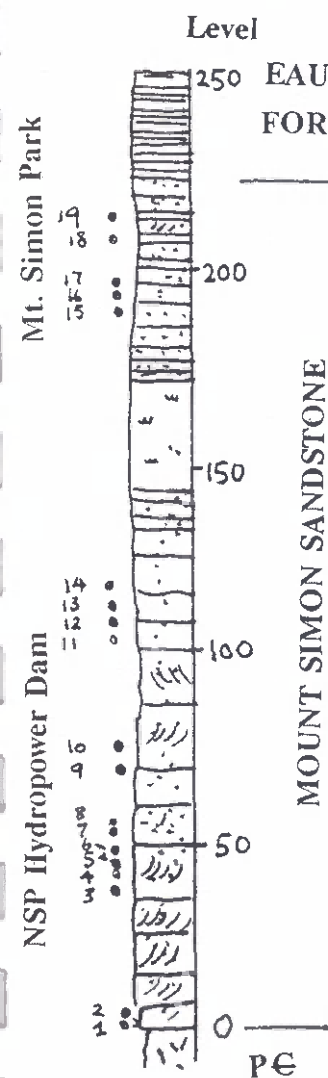


Fig. S1.1 Grain Composition of Mt. Simon Sandstones based on 500 points



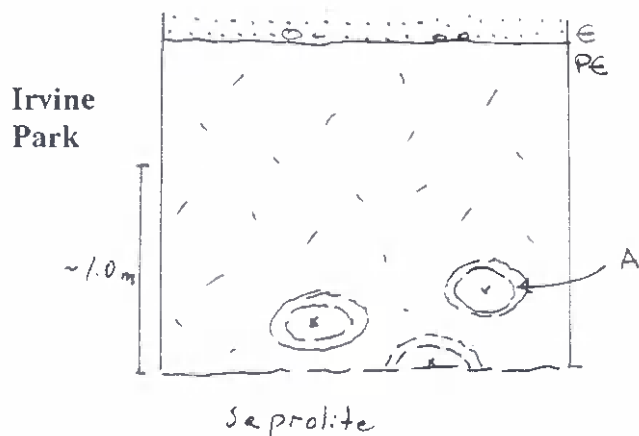
Specimen	DMQ	DPQ	DSQ	QO	DO	DM	AKF	PHY	O
MS-1	44.4	32.2	3.2	11.2	0.8	0.8	7.4	0	-
MS-2	68.6	3.8	1.2	20.2	0	0.2	5.6	0	-
MS-3	64.6	2.2	0.8	25	0	1.4	2.4	0.2	2.8
MS-4	55.8	3.0	0	29.4	0.2	2.4	6.8	0.4	2.0
MS-5	65.2	1.6	2.4	24.8	1.6	2.0	1.6	0.6	0.6
MS-7	65.2	2.8	2.6	24.0	1.0	0.8	2.4	1.0	-
MS-9	26.6	2.4	2.2	10.2	4.0	4.0	43.6	6.6	0.2
MS-10	33.6	8.0	1.8	11.6	2.2	1.8	36.6	4.0	-
MS-11	60.7	6.5	4.5	17.7	2.7	3.7	3.0	0.4	2
MS-12	69.6	2.8	0	9.0	0.4	2.2	11.4	4.0	0.4
MS-13	58.6	6.4	5.8	23.6	1.2	1.4	2.0	0.4	1.0
MS-14	61.4	8.6	6.0	13.0	5.0	1.4	3.0	0.6	1.6
MS-15	61.1	8.8	10.8	12.9	0.2	0.5	4.7	0	0
MS-16	72.2	2.5	3.7	7.4	1.7	2.5	9.1	0	0
MS-17	60.0	5.6	2.7	6.4	9.9	6.7	4.3	0.9	1.6
MS-18	50.3	6.0	0.3	13.6	10.6	4.3	13.6	0	1.0
MS-19	54.0	3.3	0.6	7.6	3.6	2.7	25.9	0	2.0

DMQ - Detrital monocrystalline quartz; DPQ - Detrital polycrystalline quartz;
 SQ - Strained quartz; QO - Quartz overgrowth; DO - Detrital orthoclase;
 DM - Detrital microcline; AKF - Authigenic K-feldspar;
 PHY - Phyllosilicate; O - Others.

Supplementary Notes 2.

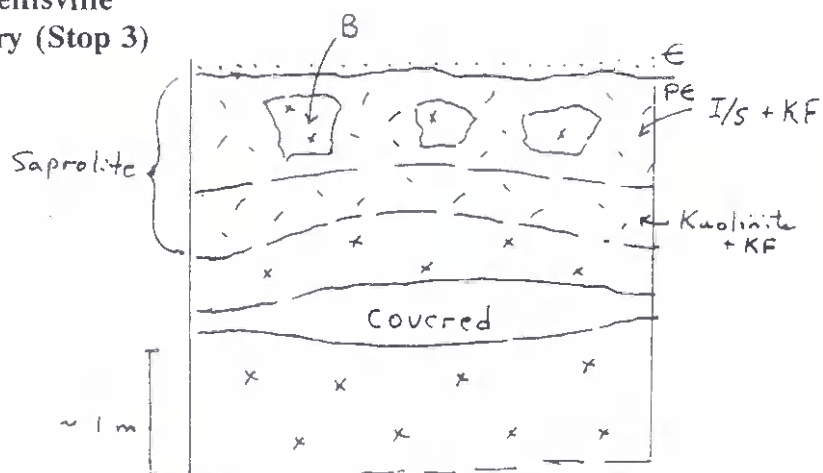
Mineralogical data for diagenetically altered paleosols at Stops 1, 3, & 4

Analysis done by Richard L. Hay



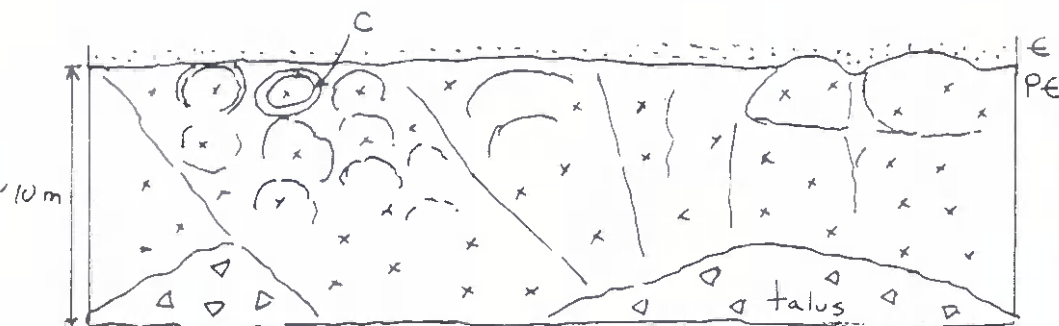
	Fresh Oligoclase Granite	A (Altered)
Q	35	24
Mr	14	8
Pc	47	0
Mc	3	5
KF	0	52
Ph	0	10

1st Neillsville Quarry (Stop 3)



	Microcline Granite	B (Altered)
Q	43	43
Mr	40	0
Pc	12	0
Mc	5	5
KF	0	49
Ph	0	3

2nd Neillsville Quarry (Stop 4)



Q - Quartz
 Mr - Microcline
 Pc - Plagioclase
 Mc - Mica
 KF - Authigenic K-feldspar
 Ph - Authigenic Phyllosilicate

 Saprolite & gruss

 Granite & granite gneiss

	Granite Gneiss	C (Altered)
Q	24	13
Mr	51	31
Pc	18	0
Mc	-1	-1
KF	0	53
Ph	0	2

SATURDAY, MAY 4, 1991

FIELD TRIP #3

M.G. Mudrey, Jr., and B.A. Brown (Leaders)

**"Proterozoic volcanogenic massive sulfide
deposits of northwestern Wisconsin"**

PROTEROZOIC VOLCANOGENIC MASSIVE SULFIDE

DEPOSITS OF NORTHWESTERN WISCONSIN

M.G. Mudrey, Jr., and B.A. Brown

Wisconsin Geological and Natural History Survey, Madison, WI

INTRODUCTION:

Field trip 3 represents a departure from the usual format followed by previous Institute field trips. The emphasis is on massive sulfide mineralization, and particularly on the Flambeau, Lynne, and Bend deposits. The regional geology has been covered in previous Institute field guides (1980, 1984, 1986) and in several papers about the regional geology (Sims and others 1990, Greenberg and Brown, 1983). Because time is limited, we will depart from tradition and not visit any outcrops. We will proceed directly to Hawkins, where we will examine core from the Bend and Lynne deposits, guided by personnel from E. K. Lehman and Associates and Noranda. Then we will go to Ladysmith, where we will examine core from the Flambeau Deposit, guided by personnel of the Flambeau Mining Co. The following notes, based largely on papers by DeMatties (1989) and Mudrey and others (1991), are intended to provide a brief outline of the geology, mineralization, and exploration history, and will be supplemented by reprints provided to trip participants.

GEOLOGIC FRAMEWORK:

The Penokean orogen in northern Wisconsin consists of two distinct terranes. The northern terrane contains platform sediment, including oxide-facies iron formation, turbidite, and minor tholeiitic volcanic units deposited on an Archean continental margin. The northern terrane extends east into Michigan and west across the Midcontinent Rift into Minnesota. This terrane has historically been a major producer of iron but is not known to contain major massive sulfide deposits.

The southern terrane, known only in Wisconsin, represents an island arc

volcanic terrane accreted to the Superior Craton during the Penokean orogeny (1850 MA). This terrane, the Penokean Volcanic Belt of Greenberg and Brown (1983) or Wisconsin Magmatic Terrane of Sims (1987), consists of calc-alkaline volcanic rock and volcanogenic sedimentary rock. The Penokean arc terrane, in contrast to the continental margin terrane, contains significant massive sulfide mineralization. To date, seven deposits totalling nearly 100 million tons have been discovered, including the 60 million ton Crandon deposit.

MASSIVE SULFIDE DEPOSITS

DeMatties (1989) recognized two mineralized districts in the western area of the volcanic belt, the Ladysmith district centered on the Flambeau deposit in Rusk County and the Somo district to the east in Price, Lincoln, and Taylor Counties. The massive sulfide deposits belong to the Cu-Zn group of Franklin and others (1981) and represent at least three district host environments.

The first environment is exemplified by the Flambeau and Ritchie Creek main zone deposits, and consists of syngenetic strata-bound and stratiform mineralization associated with felsic volcanic centers. This association is typical of many of the classic deposits in the Canadian Superior Province. The felsic-hosted deposits consist of tabular to lenticular layers of massive to semi-massive sulfide, enveloped by a disseminated sulfide halo with no distinctly recognizable alteration pipe. The mineralization is copper (chalcopyrite) rich and associated with quartz-sericite $\pm$ amphibole or chlorite-carbonate alteration.

Host rock in this environment is commonly felsic crystal tuff (quartz-sericite schist). Cherty tuff and metachert commonly occur interbedded with or stratigraphically above the host tuff. Disseminated sulfide of the footwall grade upward into massive sulfide, which is commonly in sharp contact with the hanging wall rock. Primary jointing may be present in these deposits

with zinc enrichment occurring stratigraphically above the copper rich ore. Precious metals may be uniformly distributed through the massive sulfides or occur as assay zones throughout the sulfide halo. DeMatties (1988) reported that gold appears to be locally enriched in the Ritchie Creek ore body near crosscutting faults that have been intruded by mafic-intermediate dikes.

The second environment described by DeMatties (1989) is that of syngenetic strata-bound and stratiform massive sulfides associated with cherty magnetic iron formation within the main volcanic arc sequence. The Thornapple deposit north of Ladysmith is an example of this type. The deposit consists of lenses of copper and zinc bearing pyrite-pyrrhotite in quartz-amphibole schist, which is interpreted to have been a mafic tuff. The mineralized lenses grade laterally into cherty magnetic iron formation. The Thornapple deposit is not associated with a felsic center but is located in predominantly mafic rocks of the volcanic arc sequence. Zoning is also evident within the Thornapple deposit, with sphalerite enriched in the upper ore lenses. Some stringer and disseminated sulfide occurs stratigraphically below the main chalcocite-bearing lens. Limited information suggests that there are precious metals (0.029 oz/ton Au) in this deposit as well.

DeMatties' third type comprises epigenetic stringer sulfide mineralization and syngenetic strata-bound and stratiform massive sulfide at or near the top of mafic volcanic piles within back arc basin sequences. The Kivela zone on the Ritchie Creek prospect occurs at the contact between massive mafic to intermediate flows and overlying cherty tuff and volcanogenic metasedimentary rock. A sulfide halo is developed in the flows and is associated with intense quartz-sericite alteration. The halo grades into an alteration pipe of stringers and veinlets of pyrite-pyrrhotite chalcocite $\pm$ quartz in brecciated and altered country rock.

The environments of mineralization described by DeMatties (1989) provide

the basis for a geological model to guide exploration in the western part of the volcanic belt. A similar stratigraphic model has yet to be developed for the eastern part of the belt, although known deposits such as Crandon, Pelican, and barren sulfides in Marinette County occur in similar host rock settings. The lithologic similarity, similar tectonic setting, and presence of mineralization throughout the belt suggests a potential for more discoveries (Mudrey and Kalliokoski, in press; Mudrey, 1979; and Mudrey and others, 1991).

EXPLORATION AND DEVELOPMENT HISTORY

The history of exploration in the Penokean volcanic belt up to 1976 is summarized in detail by Mudrey and others (1991). An important factor was the existence of a good database from early surveys of the Wisconsin Geological and Natural History Survey in which C.E. Dutton recognized the presence of "greenstone belt" assemblages and hints of mineralization similar to that of massive sulfide bearing Archean terranes of Canada. Although exploration of the terrane began in the 1950s, real progress was made only after the successful application of ground and airborne geophysics allowed conductive anomalies to be identified beneath the thick glacial overburden. The result has been a series of significant discoveries, from the announcement of the Flambeau deposit in 1968 and Crandon in 1976 down to the Lynne and Bend projects of the 1990s.

With the permitting of the Flambeau deposit in early 1991 and with startup scheduled for the summer of 1991, the Penokean terrane may finally become a producing district.

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