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By KARIN M. BAROVICH, P. JONATHAN PATCHETT, ZELL E. PETERMAN, and P.K. SIMS

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By WARREN C. DAY, P.K. SIMS, and ZELL E. PETERMAN

Chapters G and H are issued as a single volume and are not available separately

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P.K. SIMS and L.M.H. CARTER, Editors

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Neodymium Isotopic Evidence for Early Proterozoic Units in the Watersmeet Gneiss Dome, Northern Michigan

By Karin M. Barovich,[1] P. Jonathan Patchett,[1] Zell E. Peterman, *and* P.K. Sims

Abstract

In the Watersmeet gneiss dome, Upper Peninsula of Michigan, neodymium isotopic data for three gneiss units previously thought to be entirely Archean in age demonstrate that one unit, biotite gneiss and schist, must be primarily Early Proterozoic in origin. T_{CHUR} (chondrite uniform reservoir) ages for six of the gneiss samples range from 3.72 to 3.54 Ga, confirming the Early Archean origin of the majority of the gneisses in the dome. T_{DM} (depleted-mantle) ages for four samples from a biotite gneiss and schist unit from the dome, however, range from 2.37 to 2.25 Ga, and $\varepsilon_{Nd}(1.8)$ values range from −3.1 to −2.3, indicating a mixture of an Early Proterozoic source with some recycled Archean material. The biotite gneiss and schist unit has uncommonly high samarium and neodymium concentrations relative to the other gneisses in the dome, between 13 and 26 parts per million and 71 and 142 parts per million, respectively. These and other high incompatible element concentrations suggest that the protolith of the Early Proterozoic unit possibly was volcanic, representing quite a strongly chemically differentiated magma. The presence of disseminated fluorite in the Early Proterozoic rocks suggests that they formed in a volatile-rich environment.

INTRODUCTION

Two Archean terranes are recognized in the Lake Superior region: (1) a Late Archean greenstone-granite terrane and (2) an Early to Late Archean gneiss terrane. The gneiss terrane lies south of the greenstone-granite terrane and is separated from it by the Great Lakes tectonic zone (GLTZ), a presumed paleosuture of subcontinental length (Sims and others, 1980). The greenstone-granite terrane is the southernmost part of the Superior province, and is termed the Wawa subprovince (Card and Ciesielski, 1986).

The Watersmeet gneiss dome, in the Upper Peninsula of Michigan, is one of several gneiss domes south of the GLTZ (Morey and others, 1982) that are cored by gneiss, migmatite, and amphibolite and are partly overlain by Early Proterozoic metasedimentary and metavolcanic rocks of the Marquette Range Supergroup (Sims, 1990). They owe their origin to Early Proterozoic (Penokean) deformation (Sims and others, 1984). Of the domes that have been studied, the Watersmeet dome formed at a higher pressure (7 kbar) than the others and at temperatures in the range of 600–650 °C (Attoh and Klasner, 1989). As a consequence, rocks in the core of the Watersmeet dome were metamorphosed to amphibolite facies and were complexly folded and tectonically interleaved and mylonitized during the doming. Sims and others (1984) concluded from the geochronological data available that all the gneisses in the core probably are Archean in age. A subsequent Sm-Nd whole-rock isotopic study of representative samples of the gneisses, reported herein, however, has shown that some of the gneisses are Early Proterozoic.

The purpose of this report is to (1) describe the principal types of gneisses in the dome, (2) present the Sm-Nd data, and (3) point out the utility of Nd isotopic work in distinguishing crustal formation ages in highly complex, metamorphosed terranes. Subsequent to recognition of the Early Proterozoic gneisses, two of us (Sims and Peterman) reexamined the field relations, collected additional samples, and studied additional thin sections. Figure 1, revised from the earlier published geologic map (Sims and others, 1984, fig. 2), is based on these new studies combined with the Nd isotopic data.

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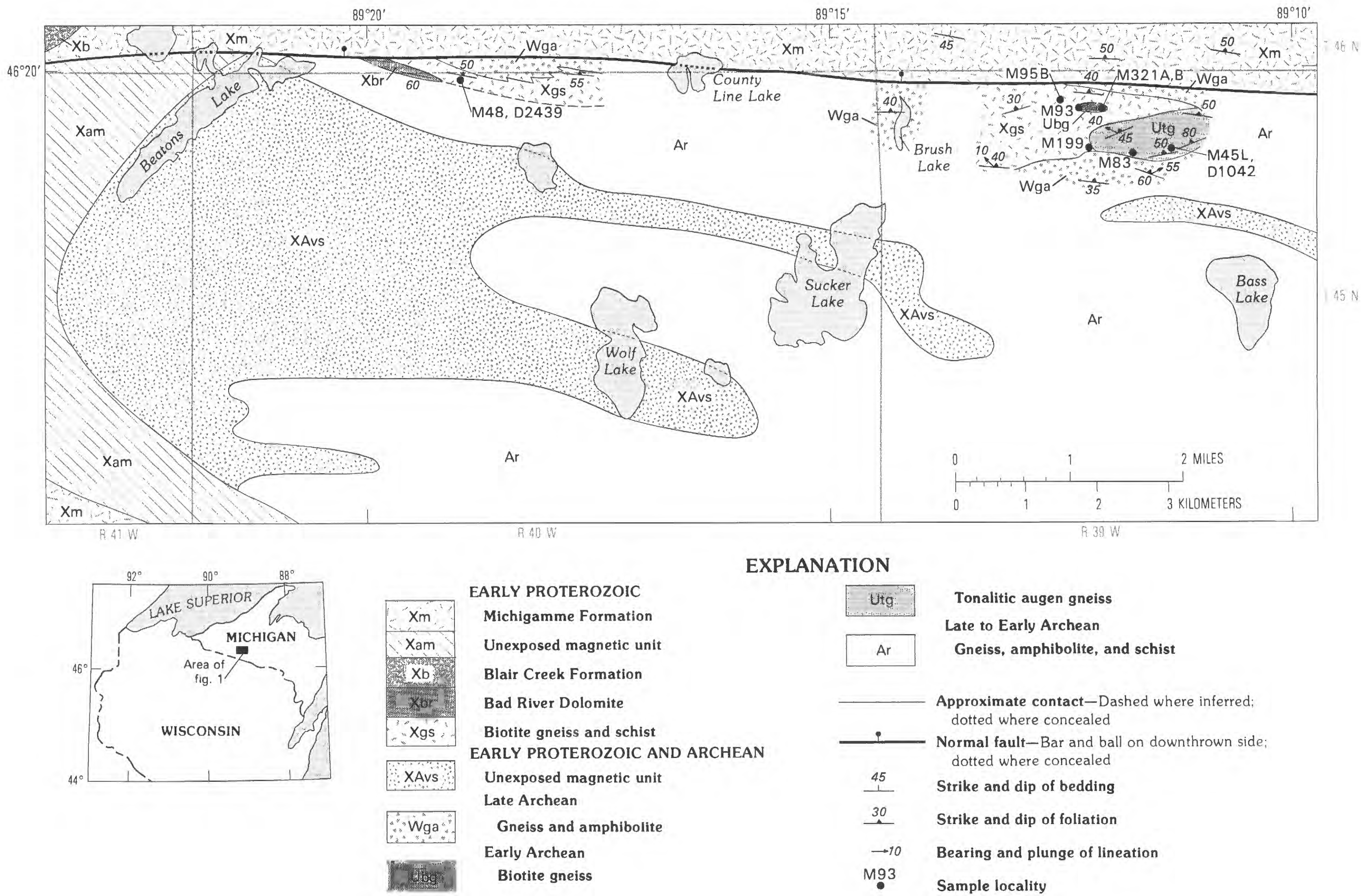

Figure 1. Geologic map of the northern part of Watersmeet gneiss dome, northern Michigan (modified from Sims, 1990). No contact line between units indicates position of contact not known.

GEOLOGY

Three major gneiss units were delineated during mapping in the area of the Watersmeet dome (Sims and others, 1984). These units are (1) tonalitic augen gneiss, exposed mainly in the northern part of the dome (fig. 1); (2) a complexly interlayered biotite gneiss succession, which overlies the tonalitic augen gneiss, and (3) an interlayered bimodal amphibolite and biotite gneiss succession, which crops out along the northern part of the dome. Our Sm-Nd study shows that part of the biotite gneiss succession (biotite gneiss and schist) is Early Proterozoic in age.

The tonalitic augen gneiss, the oldest unit (Utg on fig. 1), is a medium-gray, medium- to coarse-grained irregularly layered rock that is invariably mylonitized and recrystallized to finer grain sizes.

The biotite gneiss succession consists of two major rock types, which are intercalated: (1) a medium-grained biotite gneiss (Ubg, fig 1) that varies from massive to layered, on a scale of meters, and (2) a fine- to medium-grained biotite gneiss and schist (Xgs, fig. 1). Both rock types show multiple deformation and recrystallization, indicated by wide variations in texture and fabric. Some facies of the finer grained gneiss and schist resemble garnetiferous schist of the Early Proterozoic Michigamme Formation (Fritts, 1969). The latter (Xgs, fig. 1) is apparently the dominant unit within the biotite gneiss succession. It differs from the older biotite gneisses in having as much as 6 percent muscovite and as much as 1 percent fluorite.

The third unit (Wga, fig. 1), interlayered amphibolite and biotite gneiss, is a distinctive bimodal volcanic succession that occurs both inside and outside the dome (Sims, 1990).

GEOCHRONOLOGY AND GEOCHEMISTRY

Peterman and others (1986) determined a U-Pb zircon age of 3,562±39 Ma for the tonalitic augen gneiss of the Watersmeet dome. Zircons from one unit of the medium-grained biotite gneiss within the biotite gneiss succession also plotted on a chord with an upper intercept of 3.56 Ga. Rubidium-strontium whole-rock and mineral analyses on these units showed that these systems were reset, and indicated ages of about 1,750 Ma. The interlayered amphibolite and biotite gneiss unit has an age of about 2,640 Ma (Sims and others, 1984).

One sample (M48, fig. 1) of biotite gneiss and schist of the biotite gneiss succession gave a concordant U-Pb zircon age of 1,755 Ma, which is virtually indistinguishable from the whole-rock and mineral Rb-Sr age of 1,750±90 Ma on the rocks (Sims and others, 1984). Because of the intense deformation and metamorphism this unit underwent during the Penokean orogeny, Sims and others (1984) concluded that the Rb-Sr isochron was a secondary isochron, and interpreted the 1,755 Ma zircon age as resulting from recrystallized zircon formed during the 1.9–1.7 Ga Penokean orogeny. As discussed herein, however, the analyzed sample (M48) represents a definite Early Proterozoic rock.

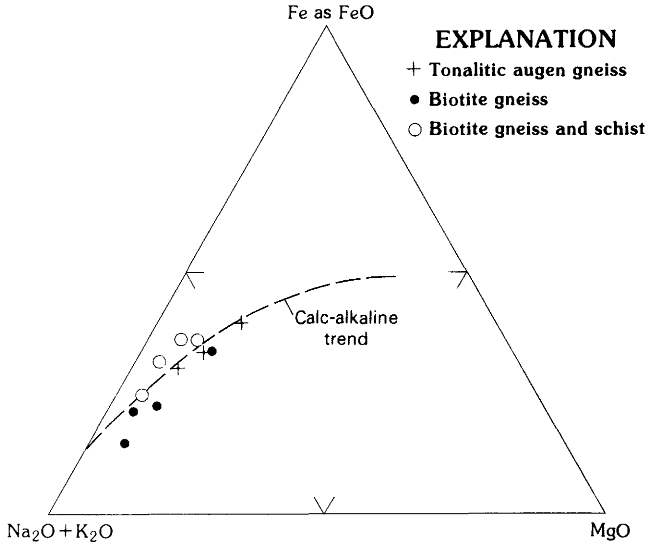

Figure 2. AFM diagram for rocks from Watersmeet gneiss dome. Calc-alkaline trend from Ringwood (1974). Data from Sims and others (1984).

Sims and others (1984) presented some major element geochemistry for all the three major rock bodies delineated in the Watersmeet gneiss dome. The range of compositions for samples from the tonalitic augen gneiss and the biotite gneiss succession is narrow, as shown in figure 2; the samples plot as calc-alkaline on this diagram. Accordingly, on the basis of major element geochemistry alone, there is no indication of differences in the age of any of the analyzed rocks of the tonalitic augen gneiss and the biotite gneiss succession.

ANALYTICAL METHODS

The analytical techniques used for Sm-Nd for crustal rocks are summarized here. Details are given by Patchett and Ruiz (1987).

1. Whole-rock powders were placed in open Teflon bombs with a HF–HNO_3 solution on hot plates to dissolve the major minerals, then sealed in the Teflon bombs with fresh HF–HNO_3 for 7 days.
2. Concentrations and $^{147}Sm:^{144}Nd$ ratios were determined with a $^{149}Sm–^{150}Nd$ combined spike.
3. Two duplicate analyses are reported (table 1, M45L and D1042). They show that for a rock powder, $^{147}Sm:^{144}Nd$ can be reproduced to within 0.5 percent.

Table 1. Sm-Nd data for rocks from Watersmeet gneiss dome

[Leaders (--), not determined]

Sample No.	Sm (ppm)	Nd (ppm)	^{147}Sm: ^{144}Nd	^{143}Nd: ^{144}Nd*	T_{CHUR} age (Ga)	T_{DM} age (Ga)
Tonalitic augen gneiss						
M45L	4.58	28.43	0.09737	0.510216±9	3.69	--
M45L	4.63	28.61	0.09768	0.510223±9	3.69	--
M83	10.92	67.17	0.09821	0.510262±9	3.65	--
D1042	5.62	35.46	0.09847	0.510251±13	3.64	--
D1042	5.88	36.16	0.09833	0.510268±7	3.64	--
Biotite gneiss succession						
Biotite gneiss:						
M93	4.10	23.90	0.10423	0.510387±13	3.68	--
M321A	2.56	15.11	0.10220	0.510315±7	3.72	--
M321B	1.82	10.96	0.10012	0.510376±10	3.54	--
Biotite gneiss and schist:						
M95B	26.33	142.27	0.11189	0.511502±7	2.04	2.30
M199	23.77	139.63	0.10287	0.511405±9	2.00	2.25
M48	19.70	102.91	0.11570	0.511519±8	2.10	2.37
D2439	13.01	70.94	0.11088	0.511508±3	2.00	2.27

* Errors are ±2 standard errors of the mean.

SAMPLE DESCRIPTIONS

Tonalitic augen gneiss (3.56 Ga):

M45L Medium-grained irregularly layered gneiss. NW¼SE¼ sec. 4, T. 45 N., R. 39 W.

M83 Similar to M45L. Coarser layering. NW¼SW¼ sec. 4, T. 45 N., R. 39 W.

D1042 Similar to M45L. NW¼SE¼ sec. 4, T. 45 N., R. 39 W. (Same locality as M45L.)

Biotite gneiss (3.56 Ga):

M93 Light-gray fine- to medium-grained layered gneiss. SW¼NE¼ sec. 5, T. 45 N., R. 39 W.

M321A Medium-grained distinctly layered biotite gneiss. NE¼ sec. 5, T. 45 N., R. 39 W.

M321B Similar to M321A, but finer layering. NE¼ sec. 5, T. 45 N., R. 39 W. (Same locality as M321A.)

Biotite gneiss and schist (Early Proterozoic):

M95B Gray fine-grained biotite gneiss. NW¼NE¼ sec. 5, T. 45 N., R. 39 W.

M199 Gray fine-grained biotite schist. NE¼SE¼ sec. 5, T. 45 N., R. 39 W.

M48 Strongly cataclasized biotite schist with porphyritic plagioclase. NE¼NW¼ sec. 4, T. 45 N., R. 40 W.

D2439 Gray fine-grained biotite schist. NE¼NW¼ sec. 4, T. 45 N., R. 40 W. (Same locality as M48.)

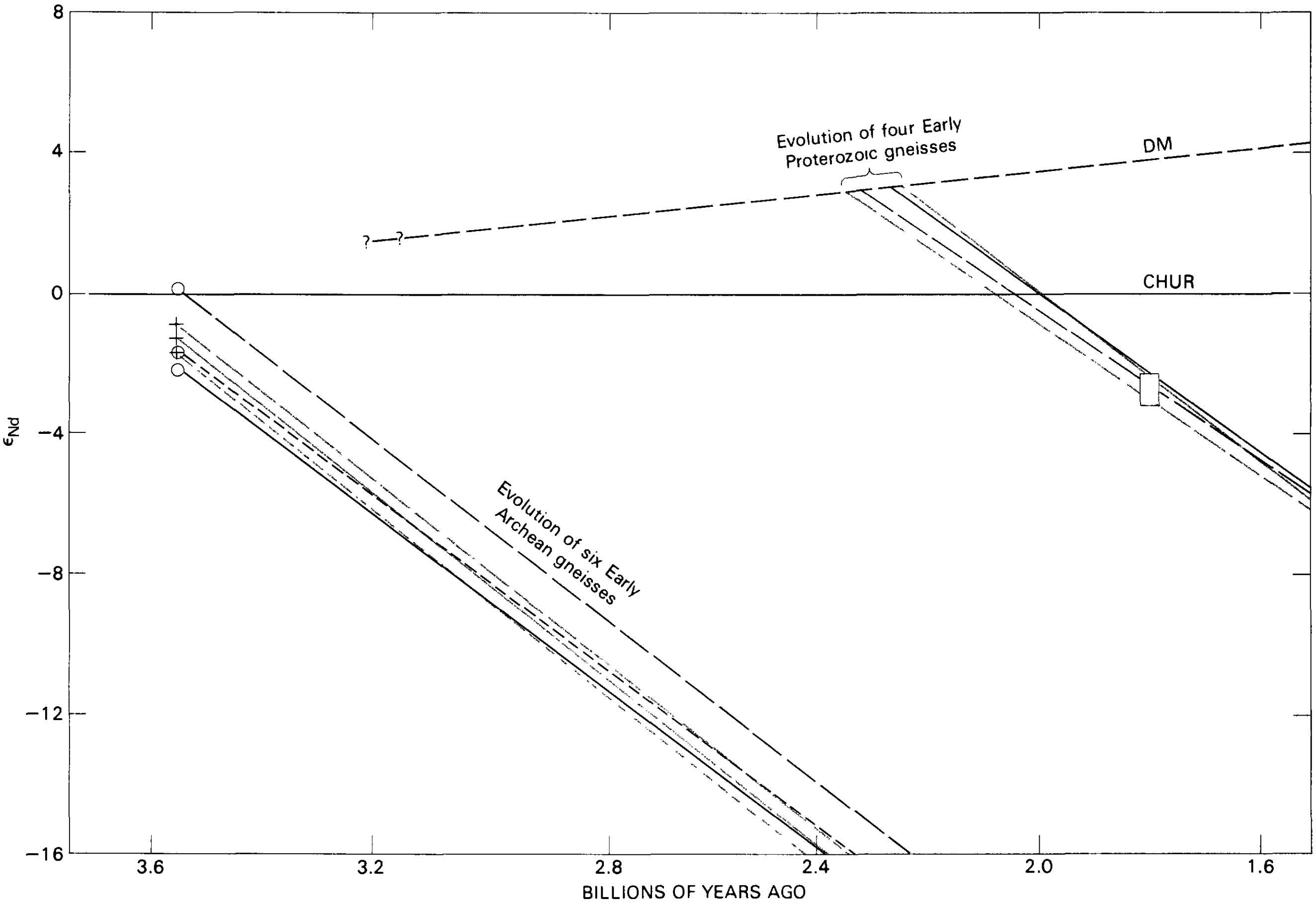

Figure 3. Nd isotopic data for rocks from Watersmeet gneiss dome, plotted as ε_{Nd} versus time. CHUR, the chondrite uniform reservoir or bulk Earth isotopic evolution; DM, depleted mantle (DePaolo, 1981). Epsilon, the deviation in parts per ten thousand of the ^{143}Nd: ^{144}Nd ratio of the sample from the chondrite value of the same age. Cross, Early Archean tonalitic augen gneiss; open circle, Early Archean biotite gneiss; open rectangle, range of $\varepsilon_{Nd}(1.8)$ values for the Early Proterozoic samples.

4. Throughout the period of this study, the La Jolla Nd standard gave $^{143}Nd:^{144}Nd=0.511862\pm7$ (2σ of the mean); 2-sigma errors reported correspond to the last two significant figures.

RESULTS

We analyzed 10 whole rock samples, 3 from the tonalitic augen gneiss, 3 from the biotite gneiss unit, and 4 from the biotite gneiss and schist unit for Nd isotopic composition and Sm and Nd concentrations. Descriptions and locations of the samples are given in table 1. Approximate sample localities are shown in figure 1. Neodymium isotopic data are presented in table 1, and isotopic evolution, ε values, and T_{CHUR} and T_{DM} model ages[2] are shown in figure 3. T_{DM} ages were calculated only for those samples that were determined to have an Early Proterozoic crustal origin. The soundness of the depleted-mantle model used (DePaolo, 1981) has been well documented for the Early Proterozoic in this region (Nelson and DePaolo, 1984; Patchett and Arndt, 1986).

The data fall into two groups, according to their model ages. The three tonalitic augen gneiss samples (M45L, M83, and D1042) and the three biotite gneiss samples (M321A, M321B, and M93) form the first group. This group yields Early Archean T_{CHUR} ages between 3.72 and 3.54 Ga. These model ages are in agreement with the U-Pb upper intercept age of 3.56 Ga determined for the tonalitic augen gneiss and one layer of the biotite gneiss unit (Peterman and others, 1980). The evolution through time of these Early Archean gneisses is plotted for reference. At the U-Pb crystallization age of 3.56 Ga, ε_{Nd} values for these samples range from −2.0 to +0.2. Sm and Nd concentrations and $^{147}Sm:^{144}Nd$ ratios generally are typical for Early Archean felsic gneisses. (See Taylor and McLennan, 1985, for summary.) The second group includes the biotite gneiss and schist samples (M48, D2439, M199, and M95B). These

[2]Ages derived using chondrite uniform reservoir or bulk Earth isotopic evolution, and depleted-mantle method.

samples have a narrow range of Early Proterozoic T_{DM} ages between 2.37 and 2.25 Ga (fig. 3). Because the crystallization ages of these gneisses and schists are unknown, we have plotted their evolution back to their intersection with the depleted mantle curve of DePaolo (1981). Whereas the $^{147}Sm:^{144}Nd$ ratios of these samples are representative of typical felsic upper crust, the Sm and Nd concentrations are anomalously high: Sm values range between 13 and 26 ppm and Nd values between 71 and 142 ppm. These values are as much as 4.5 times higher than average upper crustal felsic material.

DISCUSSION

The T_{CHUR} model ages and ε_{Nd} values for the first group of samples (the tonalitic augen gneiss and the biotite gneiss samples) show that these rocks represent new crustal additions approximately 3.60 Ga. Their model ages are also in agreement with Nd model ages for samples of the tonalitic augen gneiss of the Watersmeet dome analyzed by McCulloch and Wasserburg (1980) and Futa (1981).

The second group of samples has a narrow range of Early Proterozoic T_{DM} ages, varying by only 0.12 b.y. (from 2.37 to 2.25 Ga). It is theoretically possible that this unit is also Early Archean in age, and that its whole-rock system was reset during a later metamorphic episode. That these rocks contain fluorine (in the form of disseminated fluorite) suggests that they may have been exposed to a volatile-rich environment at some point in their history. But if they are Early Archean in age, it would be necessary to invoke large-scale REE (rare earth element) mobility to explain their Early Proterozoic model ages. An implausible scenario such as massive flooding of the rocks by extremely REE enriched hydrothermal fluids with an Early Proterozoic isotopic signature during a major metasomatic event would be needed. To achieve the consistent Early Proterozoic model ages, it would be necessary to replace all the original Sm and Nd in the rock, and completely delete any Archean isotopic signature. We find this unlikely for two reasons: (1) These rocks show no evidence of the severe and pervasive hydrothermal alteration that would have to result from the circulation of such large volumes of REE-enriched fluids. Megascopically they do not show any evidence of hydrothermal alteration, and microscopically they are unusual only in the presence of fluorite as a common accessory mineral. If the fluorite was introduced by metasomatic fluids, then its occurrence as disseminated grains throughout the unit (not in confined veinlets) would strongly suggest pervasive as opposed to selective hydrothermal alteration. Yet no other minerals record a pervasive alteration event. (2) Given the complicated structural relationships whereby the Early Archean tonalitic augen gneiss and biotite gneiss are interleaved with the biotite gneiss and schist, it is impossible to see how this REE replacement during massive fluid circulation could have been so selective—as long as we assume that the rock types were in their current structural positions relative to each other before the proposed alteration event. We suggest instead that the biotite gneiss and schist unit, represented by the samples with Early Proterozoic model ages, was added to the crust around 1.80 Ga, probably during an early phase of the 1.9–1.7 Ga Penokean orogeny, and that it was then structurally interleaved with the Archean gneiss units of the Watersmeet dome during deformation associated with the Penokean event. Barovich and others (1989) have already demonstrated that the Penokean orogeny was a time of major growth of crust from the mantle. The ≈2.30 Ga T_{DM} ages of the biotite gneiss and schist samples do not correlate with any known igneous or volcanic activity. Instead, we suggest that mixing of two components, depleted mantle and a small percentage of preexisting Archean crustal material, produced the precursors of these rocks, around 1.80 Ga. Thus the T_{DM} model ages of ≈2.30 Ga are older than the proposed crystallization age of 1.80 Ga, and represent a weighted average of the two components, primitive mantle and recycled Archean crustal material. Existence of a depleted mantle in this area and others in North America has been well documented (Nelson and DePaolo, 1984; Barovich and others, 1989). That some Archean material was incorporated during formation of the gneiss protoliths is reasonable to assume, because the terrane from which these samples were taken contains considerable amounts of Archean basement.

As a result of subsequent field and laboratory work, we have distinguished the Early Proterozoic biotite gneiss and schist from the Early Archean units within the Watersmeet dome. In thin section, the biotite gneiss and schist unit is characterized by 3–6 percent modal muscovite and accessory fluorite. In addition, trace element analyses by K.J. Schulz (unpub. data, 1990) have shown that the rocks are uncommonly enriched in incompatible elements, including U, Th, Ta, Hf, and Zr.

The elevated REE and incompatible element concentrations of these Early Proterozoic rocks are uncommon. The minerals that host the REE budget of the whole rock have not been identified. The presence of fluorite as an accessory mineral disseminated throughout the rocks suggests a volatile-rich environment for the precursor of this unit. Although little experimental evidence exists to gauge the mobility of rare earth and other incompatible elements as fluorine complexes and explain the elevation of incompatible element concentrations in fluorine-rich rocks, much evidence of an empirical nature is available. Alderton and others (1980) attributed REE loss in an altered granite in southwest England to REE complexing by fluorine in the altering fluids. Christiansen and others (1986) cited characteristic enrichments in incompatible lithophile elements in Cenozoic fluorine-rich topaz rhyolites from the Western United States.

CONCLUSIONS

Although most of the rock units in the Watersmeet gneiss dome are Early Archean in age, as confirmed by our Nd isotopic data for the tonalitic augen gneiss and the biotite gneiss, our data have shown that the biotite gneiss and schist unit within the dome is Early Proterozoic. Trace element concentrations, both from this study and unpublished data by K.J. Schulz, show that, geochemically, the Early Proterozoic gneiss and schist unit is quite distinct from the Early Archean gneisses of the Watersmeet dome. Further work is needed to determine the origin of this Early Proterozoic unit and the manner and timing in which these rocks were so complexly interleaved with the Early Archean units of the gneiss dome.

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Chapter H

Temperature-Pressure Estimates of Dynamically Recrystallized Rocks in the Early Proterozoic Mountain Shear Zone, Northeastern Wisconsin

By WARREN C. DAY, P.K. SIMS, and ZELL E. PETERMAN

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CONTRIBUTIONS TO PRECAMBRIAN GEOLOGY OF LAKE SUPERIOR REGION

P.K. SIMS and L.M.H. CARTER, Editors

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Temperature-Pressure Estimates of Dynamically Recrystallized Rocks in the Early Proterozoic Mountain Shear Zone, Northeastern Wisconsin

By Warren C. Day, P.K. Sims, *and* Zell E. Peterman

Abstract

The Mountain shear zone, in northeastern Wisconsin, is a discrete structural belt 2 kilometers wide and 12 kilometers long that transects older (1,889 to 1,860 Ma) supracrustal and granitoid rocks of the Pembine-Wausau terrane of the Early Proterozoic Wisconsin magmatic terranes. The shear zone is intruded by the 1,812 Ma Hines Quartz Diorite, which provides a minimum age of deformation. Dextral transpressive second-phase (D_2) ductile deformation within the zone reoriented original regional (D_1) deformational fabric and caused dynamic recrystallization, which is recorded in the second-phase (M_2) mineral assemblages. The weak, nearly bedding-parallel regional (S_1) foliation preserved outside the shear zone is overprinted within it by a strong second-phase (S_2) schistosity. Protomylonite and orthomylonite, as well as S-C tectonite, are developed within the shear zone.

Hornblende geobarometry of granitoid rocks and metabasalt indicates that the regional M_1 pressure of about 4.5±1 kilobars was elevated during the second metamorphic episode (M_2) both within and adjacent to the shear zone to approximately 6.0±1 kbar. Garnet-biotite geothermometry indicates that M_1 ranged from 585 °C to 615 °C, but that M_2 equilibrium temperatures may have increased to as much as 654 °C. However, garnet is a rare mineral both outside and inside the shear zone, and hence, the peak M_2 thermal conditions are only an estimate. 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 kilobars) down to very low temperatures (280 °C at 6.0 kilobars). These data indicate that shear-induced dynamic M_2 recrystallization effectively increased pressure and temperature and produced significant alteration in rock and structures during its operation in the Mountain shear zone. Maximum temperature-pressure conditions in the shear zone were followed by cooling to very low temperatures, equivalent to that outside the shear zone.

INTRODUCTION

This study consisting of field and geochemical analysis documents the mineralogical changes and associated temperatures and pressures attendant to deformation within the Mountain shear zone, northeastern Wisconsin. Early regional D_1 deformation textures and associated M_1 metamorphic minerals were overprinted in and adjacent to the shear zone by shear-induced D_2 fabrics and M_2 metamorphic mineral assemblages. Geobarometry and geothermometry applied to rocks of similar bulk composition from within and outside the shear zone provide estimates of the temperature and pressure changes produced by the ductile deformation.

Regional Setting

The Mountain shear zone is a discrete zone of ductile deformation in metavolcanic and granitoid rocks of the Pembine-Wausau terrane, within the Early Proterozoic Penokean orogen (Sims, 1987; Sims and others, 1989). The Penokean orogen lies along the south margin of the Superior province of the Canadian Shield, and in northern Wisconsin and Michigan consists of two lithostratigraphic assemblages (fig. 1). To the north is a continental-margin assemblage of metasedimentary rocks of the Early Proterozoic Marquette Range Supergroup, which overlies an Archean basement. To the south is an assemblage of volcanic and granitoid rocks of the Wisconsin magmatic terranes. The two assemblages are separated by the east-west-trending Niagara fault zone, which is thought to be an Early Proterozoic paleosuture (Sims and others, 1989). The supracrustal rocks of the Mountain shear zone were generated in an island arc and (or) back-arc basin environment in the interval 1,889–1,860 Ma, and were intruded by granitoid rocks in the interval between 1,870 and 1,760 Ma. The Hines Quartz Diorite intruded the Mountain shear zone at about 1,812 Ma, thus providing a minimum age for the shear zone. For a more comprehensive regional geologic review the reader is referred to reports by Sims and others (1989; 1990) and Sims (1987).

Local Geologic Setting

The Mountain shear zone is a northeast-trending zone of ductile deformation that is 2 km wide and exposed for a length of 12 km (fig. 2). Early Proterozoic metavolcanic rocks of the Waupee Volcanics as well as granodiorite occur along the southeast margin of the shear zone. The eastern part of the shear zone is covered by Quaternary glacial deposits. The shear zone is intruded by the 1.47 Ga Middle Proterozoic anorogenic rocks of the Wolf River batholith along the northern and western periphery (Sims, 1989; fig. 2).

Outside the shear zone, the Waupee Volcanics are massive to thinly bedded, moderately foliated dacitic to rhyolitic tuff and tuff breccia and minor volcanogenic metasedimentary rocks (unit Xwf, fig. 2; tables 1 and 2) containing M_1 mineral assemblages. Interbedded with these felsic rocks are minor layers of massive to thinly layered mafic metavolcanic rocks. Along the southeast margin the Waupee Volcanics are olivine-normative metabasaltic rocks with low Mg-numbers (cation ratio of 100×Mg:(Mg+0.85 Fe_{total})) that form massive and pillow lava flows and thinly layered mafic tuffs (table 2). Inside the shear zone, the Waupee volcanic rocks are schistose and comprise M_2 mineral assemblages. In the western part of the shear zone, they consist of interlayered biotite schist, interpreted to have had a felsic volcanic and volcanogenic protolith, and minor amphibole schist. The eastern part of the shear zone is composed dominantly of amphibole schist and quartz diorite schist. The protolith for the amphibole schist was mafic volcanic rocks of the Waupee Volcanics, similar to those outside the shear zone (unit Xwm, fig. 2).

An Early Proterozoic peraluminous granitoid rock intrudes the Waupee Volcanics both outside and inside the shear zone (fig. 2; tables 1 and 2). Outside, the rock is a medium-grained, moderately foliated hornblende granodiorite (fig. 3) with a porphyritic texture resulting from plagioclase and potassium-feldspar phenocrysts; the rock has regional M_1 mineral assemblages. Inside the shear zone, the granitoid is a mylonitic biotite-hornblende granodiorite to tonalite gneiss with M_2 mineral assemblages. The unit contains porphyroclasts of plagioclase and microcline and lensoid recrystallized quartz aggregates in a fine-grained recrystallized groundmass of plagioclase, alkali feldspar, quartz, and minor biotite and amphibole. Using the terminology of Wise and others (1984), the unit is dominantly a protomylonite with zones of orthomylonite. The protolith of the unit is the granodiorite exposed outside the shear zone (unit Xg, fig. 2).

Analytical Methods

Representative samples (5–10 kg) for geochemical analysis (table 2) were selected from outcrops that exhibited the least alteration. The samples were crushed, split, and powdered to less than 100 mesh. All analyses were done in U.S. Geological Survey laboratories in Denver, Colo. Major-oxide sample splits were fused and analyzed by X-ray fluorescence using the method outlined by Taggart and others (1987). Determinations for FeO followed procedures outlined by Peck (1964), for total H_2O by Jackson and others (1985), and for CO_2 by those reviewed by Jackson and others (1987). Analytical precision and accuracy for the major oxides are less than 3 percent of the reported value. The trace elements Rb, Sr, Y, Zr, and Nb were determined by X-ray fluorescence using an energy-dispersive spectrometer with a ^{109}Cd source. Estimated analytical error for this technique is less than 10 percent.

The mineral compositions reported in tables 3–5 were obtained using an analytical research laboratory electron microprobe with six movable spectrometers. Mineral standards chosen were those that are well characterized and had a mean atomic weight similar to that of the unknown. The samples were analyzed with an electron beam current of 10 nanoamps and an accelerator voltage of 15 kev. The data were reduced using the MAGIC4 matrix correction program.

TEXTURE AND MINERAL COMPOSITION OF ROCKS WITHIN THE MOUNTAIN SHEAR ZONE

The rocks of the Mountain shear zone have undergone a regional M_1 metamorphism, which is overprinted within the shear zone by the shear-induced M_2 metamorphism. The regional M_1 metamorphism was of lower amphibolite facies. Outside the shear zone, M_1 metamorphism of the felsic volcanic and volcaniclastic rocks produced weakly foliated equigranular, medium-grained biotite-garnet-muscovite schists. Regional M_1 metamorphism of the basaltic rocks produced hornblende-plagioclase amphibolite (table 1). Muscovite, epidote, and chlorite are M_1 alteration minerals of the granodiorite both inside and outside the shear zone.

M_2 Mineralogy and Texture

Ductile D_2 deformation caused dynamic recrystallization (M_2) both within and adjacent to the shear zone. Adjacent to the shear zone, the felsic volcanic rocks have

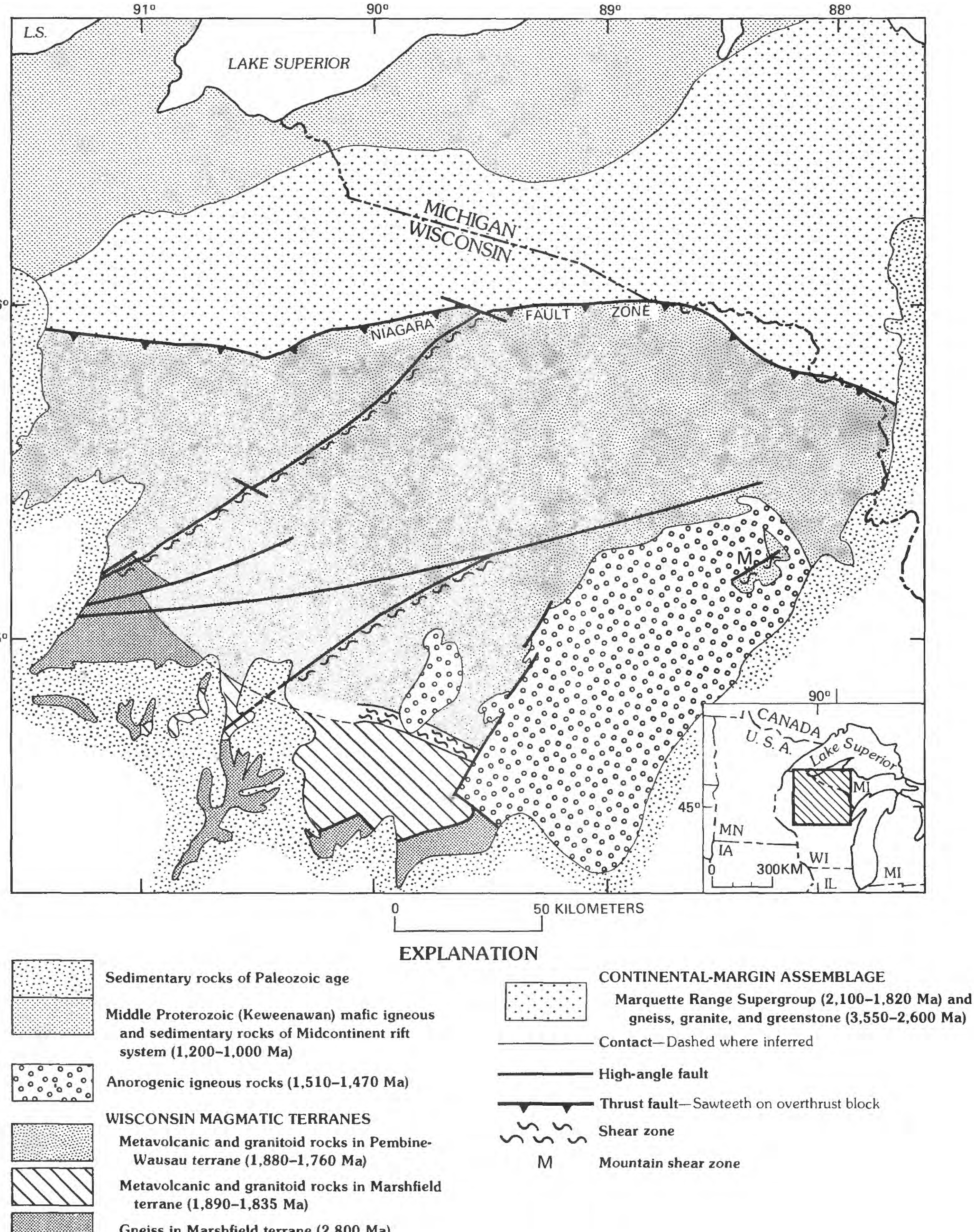


Figure 1. Regional geology of northern Wisconsin and adjacent Michigan showing relationship of continental-margin assemblage rocks to north and Wisconsin magmatic terranes to south of Niagara fault zone. Study area is near Mountain, Wis. (M). Modified from Sims and others (1990).

Table 1. Modal mineralogy for rocks of the Mountain area, northeastern Wisconsin

[Values in volume percent. Tr, trace; blank, not determined; amphibole in sample 91–85 is a clinoamphibole, probably cummingtonite]

	Outside the shear zone			Waupee Volcanics and Inside the						
Column No.	1	2	3	4	5	6	7	8	9	10
Sample No.	23–85	32–85	32B-85	18–85	164–85	20–85	134–85	172–85	144–85	91–85
Rock type	Basalt	Dacite	Dacite	Basalt	Basalt	Basalt	Quartz diorite	Quartz diorite[1]	Mafic schist	Mafic schist
Plagioclase	47.2	55.2	42.0	25.0	32.0	40.8	65.6	48.8	42.0	27.5
Quartz	1.0	8.0	10.0	1.0	Tr		9.7	22.6	41.5	31.0
Microcline		Tr								
Biotite		33.6	27.5			18.6	5.6	12.2	13.0	3.0
Hornblende	51.4	Tr		68.5	64.0	35.6	18.4	15.8		7.0
Muscovite		Tr	2.5						Tr	Tr
Garnet		2.0	15.5							
Epidote				3.0	2.0				Tr	Tr
Sphene	0.2			2.0	Tr	4.0				Tr
Chlorite				Tr	Tr				3.5	Tr
Calcite				0.3	Tr				Tr	
Magnetite				0.2	2.0					
Accessory	0.2	1.2	2.5	Tr	Tr	1.0	0.7	0.6	Tr	Tr

	Granodiorite Outside the shear zone								Hines Quartz Diorite Inside the shear zone		
Column No.	22	23	24	25	26	27			28	29	30
Sample No.	101–85	60B–85	35–85	112–85	165–85	34A–85	Average	Standard deviation	147–85	150–85	88–85
Plagioclase	42.8	49.5	40.7	44.8	51.8	55.2	47.5	5.1	47.7	54.2	41.0
Quartz	18.2	15.0	37.6	14.6	18.2	21.6	20.9	7.8	32.6	2.0	37.3
Microcline	24.6	17.5	15.4	20.6	11.0	5.7	15.8	6.2			10.0
Biotite	8.8	10.0	5.4	10.6	9.0	7.0	8.5	1.8	12.4	0.9	8.5
Hornblende	1.8	7.5		7.3	9.8	9.0	7.1	2.8	6.3	[2]41.7	2.5
Accessory	0.5	0.5	0.8	2.1	0.2	1.5	0.9	0.7	0.9	1.2	Tr
Alteration minerals	3.3	Tr	0.1	Tr	Tr	Tr	0.6	1.2	0.1	Tr	Tr

[1]Unit Xwg.
[2]Includes relicts of clinopyroxene.

minor fine-grained intercrystalline M_2 mineral phases. Phenocrysts of feldspar and quartz locally have interlocking sutured grain boundaries. The interstitial fine-grained minerals (quartz, feldspar, mica, and opaque minerals) are randomly distributed around the margins of the larger phenocrysts. The felsic volcanic rocks inside the shear zone are strongly foliated and medium grained with ubiquitous fine-grained granular intercrystalline M_2 minerals. The primary medium-grained quartz and feldspar crystals form a well-developed mosaic with distinct interlocking sutured crystal boundaries. The intercrystalline M_2 mineral matrix forms an interconnected groundmass of fine-grained equigranular "bleb-shaped" quartz, feldspar, mica, and opaque minerals that defines the strong foliation.

Similar textural variations are observed in the mafic volcanic rocks. Outside the shear zone the unit (Xwm, fig. 2) has a weak foliation (S_1), with medium- to coarse-grained plagioclase phenocrysts within fine-grained hornblende aggregates. Fine-grained, granular M_2 mineral phases (hornblende, feldspar, and minor quartz) form along the interstices of the M_1 mineral grain boundaries with no preferred orientation. Inside the shear zone, the foliation within the mafic volcanic rocks (units Xwa and Xwi) is intense, and is defined by a matrix of interlocking elongated fine-grained hornblende crystals and quartz-feldspar blebs and stringers (fig. 4). Both inside and outside the shear zone, the plagioclase phenocrysts are commonly altered to sericite, which is thought to be an M_1 phase. However, inside the shear zone the fine-grained M_2 plagioclase essentially lacks sericite alteration.

Crystal-plastic (D_2) deformation of the granitoid rocks (unit Xgn, fig. 2) within the shear zone has caused M_2 recrystallization of the coarse-grained quartz crystals, producing large (centimeter-scale) multicrystalline quartz stringers in which the individual crystals have sutured boundaries. The crystal boundaries of the coarse-grained phenocrysts of plagioclase and potassium feldspar differ in being more angular and less recrystallized than the quartz. Biotite is generally restricted to the fine-grained interstitial M_2 material. Small 50–150 micrometer-sized blebs of fresh M_2 plagioclase and quartz occur within the primary plagioclase phenocrysts (fig. 4).

associated quartz diorite shear zone										
11 13–85 Dacite	12 199–85 Dacite	13 76–85 Dacite	14 141–1–85 Dacite	15 167–85 Dacite	16 168–85 Dacite	17 170–85 Dacite	18 18–85 Dacite	19 64–85 Rhyolite	20 173–85 Rhyolite	21 79–85 Rhyolite
50.8	49.1	53.3	42.0	31.0	50.5	40.0	57.7	50.5	51.5	46.7
41.2	3.5	37.3	41.5	29.0	37.5	21.5	30.7	38.7	22.5	23.6
2.0	31.5	0.6	2.1					8.3	10.0	21.7
4.9		5.8	13.0	28.0	11.5	37.0	10.7	1.8	15.0	7.2
	2.4			5.0	Tr					
0.8	0.1	Tr	Tr					0.3	Tr	0.8
	44.2	0.5	Tr	Tr	Tr	Tr		0.2	Tr	
		Tr		1.0	0.3	1.0			Tr	
Tr			3.5		Tr	Tr			Tr	
Tr	0.1	Tr	Tr	Tr	Tr	Tr		Tr		
Tr	Tr	1.0		6.0	Tr	Tr		Tr		Tr
0.2	0.0	Tr	Tr		0.2	0.5	0.9	0.2	Tr	Tr

Granodiorite gneiss Inside the shear zone												
31 39–85	32 46–85	33 50–85	34 129A–85	35 129B–85	36 130A-85	37 130B–85	38 55–85	39 40A–85	40 45B–85	41 44–85	Average	Standard deviation
44.2	49.4	48.2	29.5	54.0	51.0	47.0	46.7	46.1	44.4	48.4	46.3	6.0
39.0	41.2	40.4	37.1	32.2	35.3	37.0	38.3	34.4	30.5	36.1	36.5	3.1
10.6		1.5	30.9	1.0	5.0	8.0	9.4	1.3	19.0	8.7	9.5	8.8
6.0	8.4	8.4	1.1	11.0	8.2	6.5	3.4	15.0	5.4	6.5	7.3	3.5
				1.0			1.0	0.7			0.9	0.1
Tr	0.7	1.0	0.4	0.7	0.4	1.0	Tr	0.5	Tr	Tr	0.4	0.4
0.2	0.3	0.5	1.0	0.1	0.1	0.5	1.2	2.0	0.7	0.3	0.6	0.6

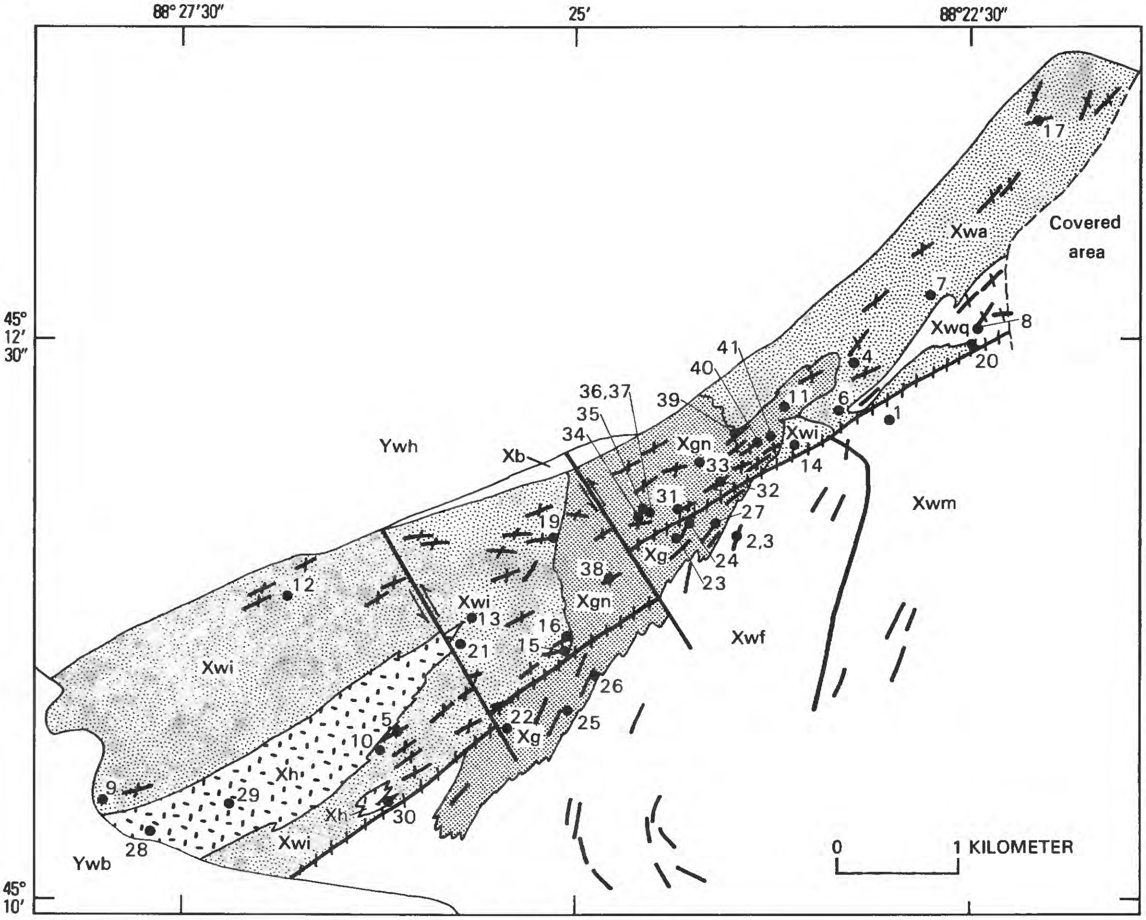
Minerals produced by M_2 are also developed (although in less abundance) outside the shear zone, indicating that the D_2 deformation was not restricted to the shear zone itself. Small M_2 blebs of quartz and feldspar are present locally in the granitoid rocks, and thin veinlets (0.1 mm to 0.5 cm) of fine-grained, comminuted quartz and feldspar occur along the margins of the primary minerals (fig. 5).

Mylonitic fabric, similar to the type II S-C mylonite fabric of Lister and Snoke (1984), is developed within the granitoid rocks inside the shear zone. The matrix material is oriented parallel to the dominant foliation and represents zones of high-strain (C-surface). Fine-grained mica within the matrix material is commonly oriented at an acute angle of 20°–40° to the dominant foliation, forming a secondary foliation (S-surface) that is compatible with a dextral sense of shearing. Locally, muscovite and (or) biotite "fish," common in S-C mylonites, are developed in the interstitial M_2 material and indicate a dextral sense of shear. Sims and others (1990) have described both the regional and local effects of the D_1 and D_2 deformation events.

Mineralogic Compositions

Recrystallization (M_2) resulting from the D_2 deformation within the Mountain shear zone produced distinct changes in mineral compositions. The hornblende composition changed slightly as a result of reequilibration. Figure 6 shows the microprobe data (table 3) for hornblende from basaltic and granodioritic rocks from the study area. In general, in the same sample the early M_1 hornblende is slightly lower in ^{IV}Al and Na+K abundances than the M_2 hornblende. The lower calculated ^{IV}Al content for the M_1 hornblende is a result of relatively higher silicon content; however, the M_1 hornblende also has lower total aluminum content.

The plagioclase composition of rocks within the shear zone changed dramatically as a result of M_2 dynamic recrystallization. In general, plagioclase became more



EXPLANATION

MIDDLE PROTEROZOIC (1,600–900 Ma)

Ywb	**Belongia Granite of Wolf River batholith (≈1,480 Ma)**
Ywh	**Hager Formation**

EARLY PROTEROZOIC (2,500–1,600 Ma)

SHEAR ZONE	Xh	**Hines Quartz Diorite**
	Xb	**Baldwin Conglomerate**
	Xgn	**Granite gneiss**—Deformed and metamorphosed granodiorite
	Xwi	**Biotite schist and amphibolite schist**—Interlayered
	Xwq	**Quartz diorite schist**
	Xwa	**Amphibole schist with thin layers of felsic schist**
OUTSIDE SHEAR ZONE	Xg	**Granodiorite**—Equivalent to unit Xgn in shear zone
		Waupee Volcanics
	Xwf	**Dominantly felsic volcanic rocks**
	Xwm	**Mafic lava flows**

Contact—Dashed where approximate

Transcurrent fault

Trace of S_1 foliation

Trace of S_2 foliation

Southeastern boundary of shear zone

•7 **Sample locality**

Table 2. Whole-rock geochemistry and normative minerals of rocks from the Mountain shear zone, northeastern Wisconsin

[Major oxide analyses by J. Taggart, A.J. Bartel, and K. Stewart; FeO, H_2O, and CO_2 by R. Brandt and S. Root; minor elements Rb and Sr by isotope dilution; chemistry by K. Futa; mass spectrometry by K. Barovich and T. Ball; Y, Zr, and Nb (EOXRF) by K. Barovich; Mg-number=Mg×100/(Mg+Fe^{+2}), Fe^{+2} calculated as 0.85×Fe total. A:KNC=Al_2O_3/(K_2O+Na_2O+CaO), molar ratio. Blank, not determined]

	Relation to shear zone								
	Outside	Inside		Outside		Inside			
Column No.	1	2	3	4	5	6	7	8	9
Sample No.	23–85 Basalt	20–85 Basalt	153–85 Dacite	27–85 Rhyolite	35–85 Grano-diorite	39–85 Granite gneiss	130–85 Granite gneiss	129–85 Granite gneiss	150–85 Hines Quartz Diorite
Major oxides, in weight percent									
SiO_2	49.0	51.8	63.1	72.4	69.3	72.7	73.2	75.6	50.1
Al_2O_3	18.9	17.70	17.40	14.50	14.20	13.20	13.00	12.40	19.00
Fe_2O_3	2.00	2.69	0.21	0.31	0.98	0.89	0.83	0.61	0.90
FeO	9.54	7.30	2.89	2.19	3.14	3.18	3.10	1.52	10.80
MgO	3.54	4.77	1.64	1.69	1.00	0.50	0.55	0.22	4.57
CaO	10.9	7.35	7.52	2.19	1.47	2.22	2.21	0.41	9.68
Na_2O	3.40	3.29	2.68	3.44	2.79	3.60	3.62	2.39	2.64
K_2O	1.09	2.28	2.97	1.33	5.52	2.50	2.34	6.48	0.64
TiO_2	0.66	0.73	0.65	0.29	0.23	0.19	0.18	0.02	0.70
P_2O_5	0.1	0.25	0.23	0.05	0.09	<0.05	<0.05	<0.05	0.17
MnO	0.16	0.16	0.05	<0.02	0.04	0.05	0.04	<0.02	0.19
H_2O+	1.28	1.48	0.40	1.15	0.68	0.58	0.45	0.29	0.99
H_2O–	<0.01	0.04	0.02	0.02	<0.01	0.04	0.03	0.06	0.02
CO_2	0.19	0.03	<0.01	<0.01	0.04	0.15	0.09	0.02	0.05
Total	100.8	99.9	99.8	99.6	99.5	99.9	99.7	100.1	100.5
Mg-number	22.10	30.86	32.67	38.43	18.41	10.25	11.49	8.80	26.35
A:KNC	0.71	0.83	0.82	1.31	1.07	1.04	1.04	1.06	0.84
Trace elements, in parts per million									
Rb	17	48	35	22	70	47	48	148	14
Sr	299	555	208	168	431	175	195	54	346
Y	12	21	15	22	8	27	18	22	9
Zr	35	92	91	165	115	93	97	42	27
Nb	2	9	7	8	2	6	4	7	1
K:Rb	541	396	711	511	658	440	402	363	377
Normative minerals of rocks									
Quartz			18.37	38.90	25.68	35.07	35.95	34.81	
Corundum				3.58	1.28	1.00	0.83	0.87	
Orthoclase	6.39	13.49	17.59	7.89	32.79	14.80	13.87	38.26	20.20
Anorthite	32.84	26.83	26.74	10.52	6.49	9.75	10.10	1.58	
Nepheline	1.77								
Wollastonite	7.93	3.28	3.79						
Enstatite	2.98	9.64	4.09	4.23	2.50	1.25	1.37	0.55	9.62
Ferrosilite	5.09	8.34	4.16	3.34	4.68	4.89	4.80	2.30	15.40
Fosterite	4.05	1.58							1.19
Fayalite	7.62	1.51							2.11
Magnetite	2.88	3.91	0.31	0.45	1.43	1.29	1.21	0.88	1.30
Ilmenite	1.24	1.39	1.23	0.55	0.44	0.36	0.34	0.04	1.32
Apatite	0.24	0.60	0.55	0.12	0.21	0.12	0.12	0.12	0.40
Calcite	0.43	0.07	0.02	0.02	0.09	0.34	0.21	0.05	0.11
Magnesite	2.14	2.88	0.99	1.02	0.60	0.30	0.33	0.13	2.76

Figure 2 (facing page). Geology of the Mountain shear zone, northeastern Wisconsin, showing sample localities. Numbers refer to column numbers in tables 1 and 2. Modified from Sims and others (1990).

albitic for rocks of equivalent composition (table 4). Plotted in figure 7 are the combined core and rim analyses for plagioclase for the various rock compositions. A significant shift to more albitic plagioclase compositions occurred in each rock type. The plagioclase composition in the felsic

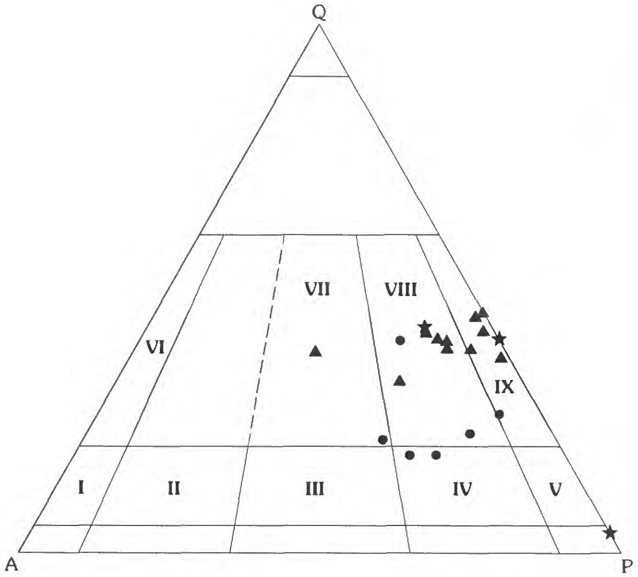


Figure 3. Quartz–alkali feldspar–plagioclase diagram for granodiorite outside the shear zone (dots), granodioritic gneiss inside the shear zone (triangles), and the Hines Quartz Diorite (stars), Mountain area, northeastern Wisconsin. Fields for diagram: I, quartz alkali syenite; II, quartz syenite; III, quartz monzonite; IV, quartz monzodiorite; V, quartz diorite; VI, alkali granite; VII, granite; VIII, granodiorite; IX, quartz diorite (tonalite).

volcanic rocks (fig. 8) changed from oligoclase (An_{24-31}) to albite (An_{10-13}). Inside the shear zone the medium-grained plagioclase phenocrysts have essentially equivalent compositions to the recrystallized M_2 matrix plagioclase, suggesting that the larger crystals equilibrated with the matrix during D_2 shearing.

A similar shift in plagioclase composition is recorded in the granodioritic rocks. Outside the shear zone (unit Xg, fig. 2) the cores of the plagioclase phenocrysts are dominantly andesine (An_{27-40}) with rims of oligoclase (An_{17}). Inside the shear zone (unit Xgn, fig. 2), however, the composition becomes distinctly more albitic in the phenocryst cores (An_{16-22}), rims (An_{23-14}), and within the recrystallized fine-grained matrix (An_{16-22}), again indicating that the plagioclase reequilibrated to more albitic compositions as a result of shear-induced M_2 recrystallization.

GEOBAROMETRY

The mineral assemblages of the rocks within the Mountain shear zone are simple, and, therefore, allow application of only one geobarometer (table 3). Recently, Hammerstrom and Zen (1986) presented an empirical geobarometer in which they calibrated the aluminum content of hornblende observed in calc-alkaline plutons as a function of depth of crystallization (pressure). They estimated pressure from the metamorphic mineral assemblages of the associated country rocks. Hammerstrom and Zen (1986) noted that the ratio of total Al (Al^t) to tetrahedral Al (^{IV}Al) in hornblende varies linearly, and that the Al^t content is principally a function of pressure. Subsequent work by Hollister and others (1987) and Rutter and others (1989) has verified the hornblende geobarometer for use in calc-alkaline plutonic rocks. Hollister and others (1989) independently confirmed the approach of Hammerstrom and Zen (1986) with additional data for hornblendes from calc-alkalic plutons from nine localities in the Central Gneiss Complex of British Columbia. They refined the equation for determining pressure, which resulted in decreasing the estimated error from ±3 kbar to ±1 kbar. Rutter and others (1989) experimentally calibrated the geobarometer at higher pressures (10 kbar) and confirmed its application to specific multimineral assemblages in which hornblende is in equilibrium with plagioclase, biotite, potassium feldspar, quartz, sphene, and magnetite or ilmenite±epidote. These phases are present in the granodiorite both inside and outside the Mountain shear zone.

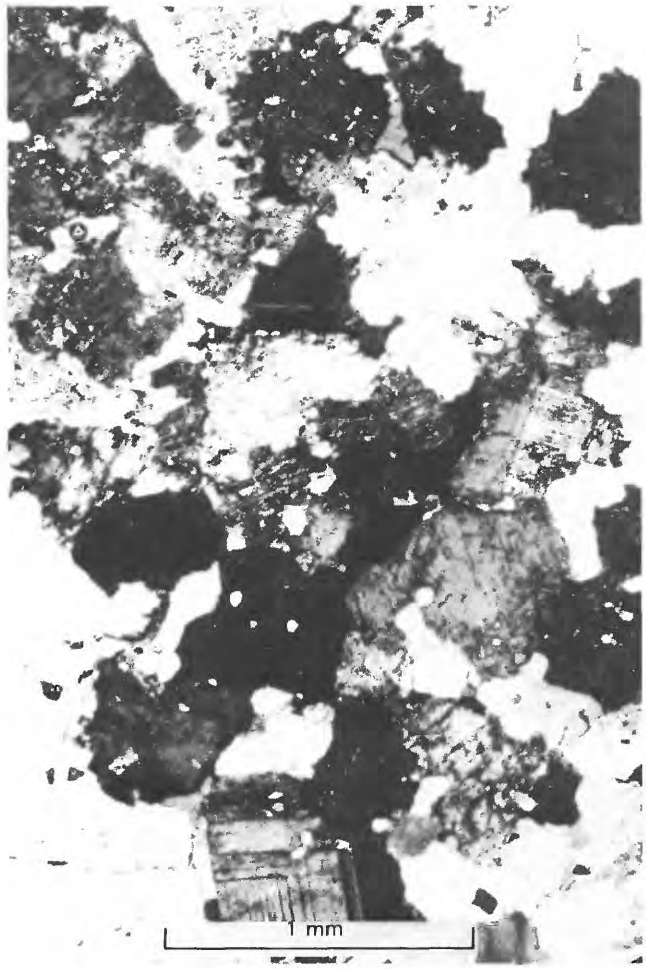


Figure 4. Photograph showing mosaic texture of quartz and feldspar in a granodiorite (129–85) inside the shear zone.

Microprobe data and calculated pressures for hornblende from the granodiorite and metamorphosed basalt inside and outside the Mountain shear zone are presented in

Figure 5. Photograph showing a small M_2 veinlet formed along margin of primary minerals in a granodiorite (165–85) outside the shear zone.

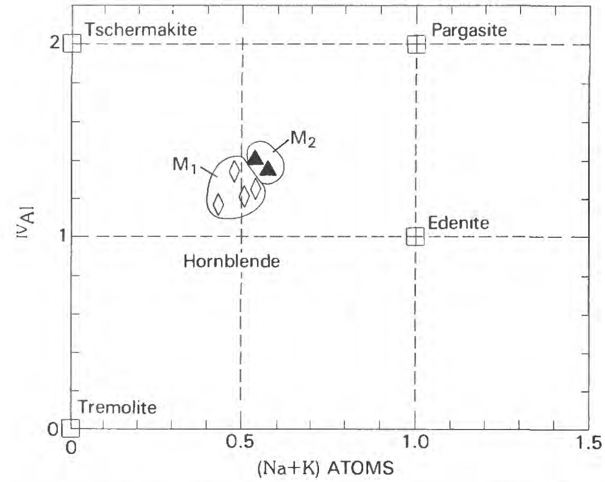


Figure 6. Hornblende composition data for rocks of the Mountain shear zone, northeastern Wisconsin. M_1 hornblende (diamond) contains less Na+K and ^{IV}Al compared to the younger M_2 hornblende (triangle). Data listed in table 3; analytical error approximately the symbol size.

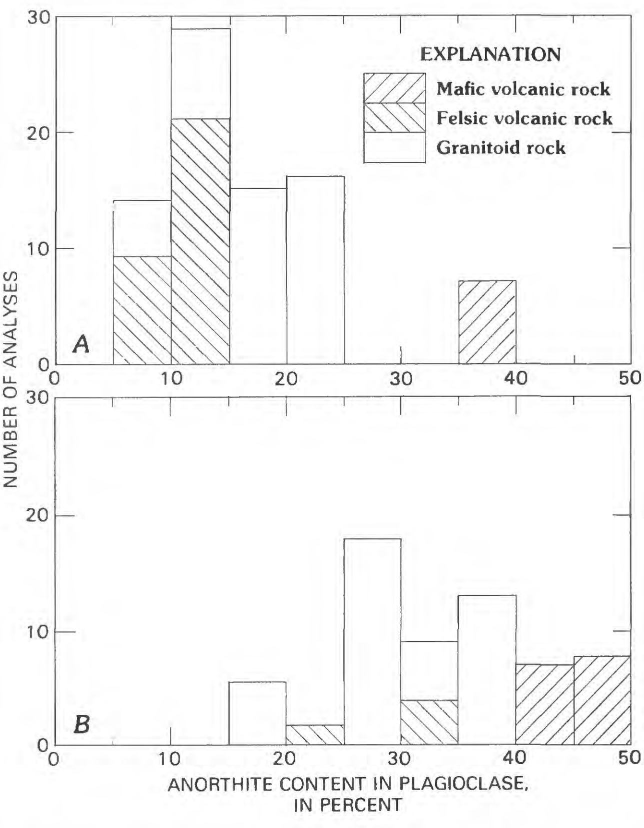


Figure 7. Histogram of plagioclase anorthite (An) content of rocks from *A*, inside, and *B*, outside the Mountain shear zone.

table 3. The basalt does not strictly meet the multimineral assemblage criteria in that it lacks potassium feldspar. However, the hornblende compositions in the basalt are very similar to those in the granodiorite. Outside the shear zone the estimated equilibrium pressures using the method of Hollister and others (1987) for the primary M_1 hornblende range from 4.0 kbar for the granodiorite to 4.7 kbar for the basalt, both of which are within the estimated error for the geobarometer of ±1 kbar, suggesting that the geobarometer yields a reasonable estimate for the metabasalt. The secondary M_2 hornblende in the basalt, which occurs along the rims of the M_1 hornblende, yields an estimated pressure of 5.8 kbar. This is significantly higher than the M_1 hornblende outside the zone.

Inside the shear zone the estimated pressure for the primary M_1 hornblende is 4.3 kbar for the basalt and 5.5 kbar for the granodiorite gneiss, similar to M_1 outside the shear zone. However, the calculated pressure of equilibration of the secondary M_2 hornblende from the granodiorite gneiss is elevated (6.1 kbar), as is the M_2 hornblende from outside the zone. These data suggest that the M_1 generation of hornblende equilibrated at similar pressures both inside and outside the shear zone at about 4.5±1 kbar. Subsequent deformation (D_2) associated with the development of the shear zone, as recorded in the M_2

Table 3. Summary of microprobe data for hornblende and pressure estimates for rocks of the Mountain shear zone

	Outside shear zone			Inside shear zone		
Sample No.	23–85		165–85	20–85	129–85	
Rock type	Basalt		Granodiorite	Basalt	Granodiorite gneiss	
	Primary M_1	Secondary M_2	Primary M_1	Primary M_1	Primary M_1	Secondary M_2
SiO_2	44.05	43.20	43.55	45.48	43.14	42.66
Al_2O_3	9.33	10.31	8.45	9.11	10.04	10.56
FeO	19.95	20.07	20.30	16.78	22.49	22.79
MgO	8.68	8.16	8.40	11.04	7.34	7.02
CaO	11.62	11.80	11.24	11.93	11.49	11.65
Na_2O	1.15	1.19	1.01	1.06	0.90	1.00
K_2O	0.98	1.10	1.01	0.64	1.06	1.18
TiO_2	0.86	0.81	1.15	0.59	0.52	0.54
MnO	0.40	0.38	0.68	0.48	0.49	0.52
Total (wt. pct.)	97.02	97.02	95.79	97.11	97.66	97.92
No. analyses	6	6	3	10	6	7
Number of cations on the basis of 23 oxygens						
Si	6.752	6.641	6.790	6.834	6.659	6.586
^{IV}Al	1.248	1.359	1.210	1.166	1.341	1.414
Al^{VI}	0.438	0.509	0.343	0.448	0.486	0.508
Fe^{+2}	2.558	2.580	2.647	2.109	2.903	2.943
Mg	1.983	1.869	1.952	2.472	1.689	1.615
Ca	1.909	1.944	1.878	1.921	1.901	1.927
Na	0.342	0.355	0.305	0.309	0.269	0.299
K	0.192	0.216	0.201	0.123	0.209	0.232
Ti	0.099	0.094	0.135	0.067	0.060	0.063
Mn	0.052	0.049	0.090	0.061	0.064	0.068
Al total	1.686	1.868	1.553	1.614	1.827	1.922
Pressure estimates (kbar)						
Hammerstrom and Zen (1986) (error ±3 kbar)	4.6	5.5	3.9	4.2	5.3	5.7
Hollister and others (1987) (error ±1 kbar)	4.7	5.8	4.0	4.3	5.5	6.1

hornblende, occurred at elevated pressures of approximately 6.0±1 kbar. Therefore, the dynamic recrystallization recorded in the shear zone was accompanied by a relative increase of approximately 1.5 kbar pressure.

GEOTHERMOMETRY

Garnet-biotite geothermometry was applied to measure the thermal peaks of the M_1 and M_2 mineral assemblages, whereas a two-feldspar geothermometer was used to unravel the effects on low-temperature subsolidus recrystallization (post-M_2) observed in the Mountain shear zone. The garnet-biotite geothermometers utilize the Mg:(Fe–Mg±Mn) ratio of coexisting garnet and biotite, and, therefore, the calculated temperatures should be internally consistent, allowing a comparison of the relative temperatures calculated for the M_1 and M_2 mineral assemblages. Although rare, garnet does coexist in textural equilibrium with biotite in the metamorphosed felsic volcanic rocks (unit Xwf, fig. 2) outside the shear zone (fig. 9) (samples 32–85 and 32B–85; tables 1 and 5). An extensive effort was made to find garnet within the shear zone; however, it is not present in the felsic volcanic rock (table 1). Garnet was noted in only one sample inside the shear zone, which was granodiorite gneiss (46–85). The garnet is an M_1 metamorphic mineral, crosscutting the original igneous minerals (fig. 10). The garnet in this sample is compositionally zoned, having a more Fe-rich overgrowth rim (M_2-stage).

The calculated temperatures using the biotite-garnet geothermometer of Ferry and Spear (1978), Hodges and Spear (1982), and Perchuck and Lavrent'eva (1983) are listed in table 5. These temperatures assume a pressure of 4.5 kbar for the M_1 metamorphism (garnet cores) and 6.0 kbar for the M_2 metamorphism (garnet rims) as calculated from the hornblende geobarometer (table 3). The effect of pressure variation on the temperature calculations is

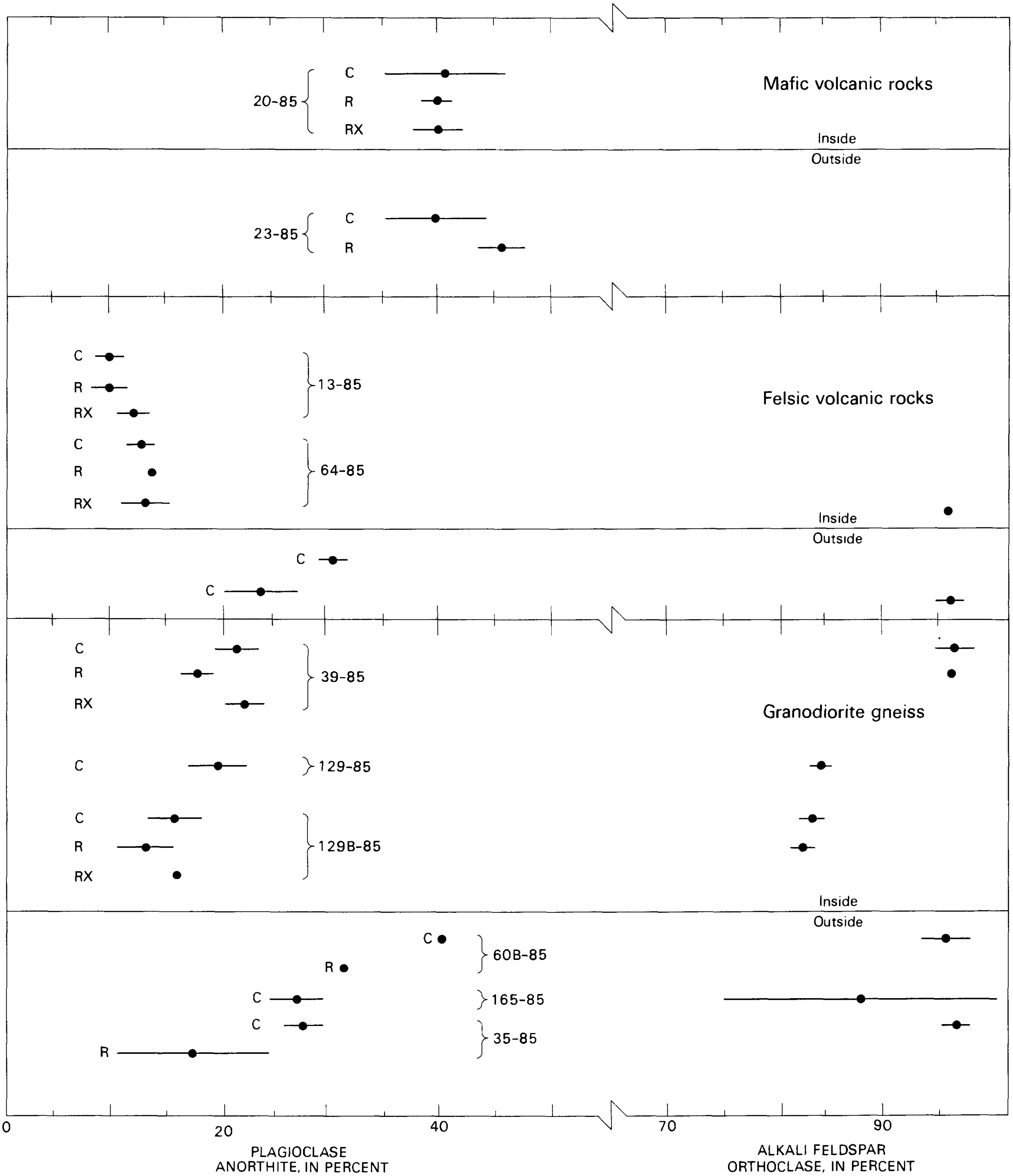


Figure 8. Plot of plagioclase anorthite (An) and alkali feldspar orthoclase (Or) content for rocks of various bulk compositions from inside and outside the Mountain shear zone. Error bars are one standard deviation from the mean (dot) for a sample. C, core; R, rim; RX, recrystallized M_2 feldspar.

minimal (3 °C/kbar using Hodges and Spear, 1982), which implies that any error in the calculated pressure is negligible when comparing the *relative* temperature variation.

The Ferry and Spear (1978) geothermometer yields an estimated temperature of 595 °C for M_1 metamorphism (4.5 kbar) and 591 °C for M_2 metamorphism (6.0 kbar) for

Table 4. Summary of mean and standard deviation for feldspar microprobe data and temperature estimates using the two-feldspar geothermometer of Haselton and others (1983) for rocks in the Mountain shear zone area

[(No.), number of analyses; blank, not determined]

		(No.)	An	Ab	Or	T (°C) 4.5 kbar	T (°C) 6.0 kbar
			Outside the shear zone				
23–85 Basalt:							
Plagioclase	Core	(7)	40.3±4.6	59.2±4.6	0.4±0.1		
	Rim	(10)	46.1±1.8	52.9±2.0	1.0±1.0		
32–85 Biotite schist (felsic metavolcanic):							
Plagioclase	Core	(2)	24.2±4.0	74.8±2.7	1.0±0.3	297	312
Alkali-feldspar	Core	(3)	0.0±0.0	3.5±1.0	96.5±1.1		
32B–85 Biotite schist (felsic metavolcanic):							
Plagioclase	Core	(4)	31.3±1.7	68.3±2.9	3.8±0.9		
35–85 Granodiorite porphyry:							
Plagioclase	Core	(9)	27.1±2.0	72.2±2.0	0.6±0.4	310	325
	Rim	(3)	17.5±7.6	69.5±4.6	13.0±9.2	278	293
Alkali-feldspar	Core	(5)	0.1±0.2	3.7±0.3	96.1±0.9		
	Rim	(1)	1.3	3.0	95.7		
60B–85 Granodiorite:							
Plagioclase	Core	(7)	39.9±1.2	58.2±1.1	1.9±0.3	420	439
Alkali-feldspar	Core	(6)	0.0	6.2±2.4	93.8±2.4		
Myrmekite							
Plagioclase	Host	(4)	26.1±2.0	73.5±0.8	0.4±0.1	340	356
Alkali-feldspar	Bleb	(1)	0.2	4.8	95.0		
Intergranular recrystallized feldspar (with sericite alteration):							
Plagioclase	Host	(5)	31.6±0.2	67.7±3.6	0.6±0.2		
	New	(6)	36.5±1.8	62.9±1.6	0.6±0.1		
165–85 Granodiorite:							
Plagioclase	Core	(5)	26.7±2.3	83.6±1.6	0.9±0.1	441	458
Alkali-feldspar	Core	(3)	1.2±1.0	11.0±5.7	87.8±13.8		

dacite outside the shear zone. These temperatures are identical within the estimated error of the technique (±50 °C). Inside the shear zone, the estimated M_1 temperature is relatively elevated to 645 °C, whereas the M_2 temperature of 612 °C (at 6.0 kbar) was essentially the same as that recorded outside the shear zone (table 5).

The Ferry and Spear (1978) geothermometer is based on ideal solid solution model for Fe-Mg cation exchange between garnet and biotite. Hodges and Spear (1982) recalibrated the geothermometer by incorporating manganese cation exchange between garnet and biotite and calcium exchange with plagioclase. This method yields temperature estimates outside the shear zone of about 610–615 °C for M_1 and a slightly lower 601 °C for M_2, all of which are within the estimated error of ±50 °C for the technique. However, inside the shear zone the calculated temperatures are elevated to 661 °C for M_1 and 654 °C for M_2.

Perchuck and Lavrent'eva (1983) presented another geothermometer that is based on the Fe-Mg-Mn cation exchange between garnet and biotite. In a recent review, Chipera and Perkins (1988) compared garnet-biotite geothermometers and suggested that the Perchuck and Lavrent'eva (1983) method yields the most precise and accurate results. However, as Indares and Martignole (1985) noted, the temperatures estimated from the Perchuck and Lavrent'eva (1983) geothermometer generally are lower for garnet with elevated manganese content. Outside the shear zone, the manganese content of the garnet is relatively low (about 3.5 wt. percent; table 5), and temperatures

Table 4. Summary of mean and standard deviation for feldspar microprobe data and temperature estimates using the two-feldspar geothermometer of Haselton and others (1983) for rocks in the Mountain shear zone area—Continued

[(No.), number of analyses; blank, not determined]

		(No.)	An	Ab	Or	T (°C) 4.5 kbar	T (°C) 6.0 kbar
Inside the shear zone							
20–85 Basalt:							
Plagioclase	Core	(7)	38.1±6.2	60.8±5.6	1.1±1.0		
	Rim		37.3±0.9	61.8±0.8	0.9±0.5		
	Groundmass		37.6±1.7	61.8±1.6	0.6±0.1		
13–85 Biotite schist (felsic volcanic):							
Plagioclase	Core	(5)	9.8±1.2	89.3±1.9	0.9±0.5		
	Rim	(4)	9.9±1.4	87.8±3.9	2.2±2.0		
	Groundmass	(5)	11.8±1.1	87.2±1.0	1.0±0.4		
64–85 Felsic metavolcanic:							
Plagioclase	Core	(8)	12.6±1.1	83.6±5.4	3.8±4.8	268	282
	Rim	(3)	13.8±0.5	85.1±2.1	1.1±1.1	266	280
	Groundmass	(5)	13.1±2.0	85.2±1.5	2.8±2.3	266	280
Alkali-feldspar	Groundmass	(4)	0.3±0.2	3.3±0.3	96.4±1.1		
39–85 Granodiorite gneiss:							
Plagioclase	Core	(6)	21.5±2.0	75.9±1.3	2.5±2.0	294	308
	Rim	(3)	17.8±1.4	76.1±5.5	6.1±5.0	293	307
	Groundmass	(2)	22.0±1.3	73.0±1.5	4.9±0.7	306	321
Alkali-feldspar	Core	(10)	0.2±0.1	3.5±1.0	96.4±1.5		
	Groundmss	(2)	0.1±0.01	3.7±0.4	96.2±0.5		
129–85 Granite gneiss:							
Plagioclase	Core	(5)	20.1±2.7	79.2±2.0	0.7±0.2	469	487
	Rim	(3)	22.8±2.7	76.5±2.2	0.6±0.1	554	574
Alkali-feldspar	Core	(4)	2.8±2.0	13.3±6.6	83.9±10.5		
	Groundmass	(2)	1.0±0.2	16.6±0.8	82.4±0.8		
129B–85 Granite gneiss:							
Plagioclase	Core	(7)	15.8±2.9	83.6±2.0	0.6±0.5	443	460
	Rim	(8)	13.5±2.7	83.5±3.7	3.0±3.1	513	532
	Groundmass	(5)	16.4±0.4	82.8±1.2	0.8±0.4	519	538
Alkali-feldspar	Core	(4)	3.3±1.0	13.3±6.0	83.4±0.8		
	Groundmass	(2)	1.0±0.2	16.6±1.3	82.4±0.8		
Crosscutting the shear zone							
147–85 Hines Quartz Diorite:							
Plagioclase	Core	(5)	42.3±0.7	57.1±0.4	0.3±0.1		
	Rim	(5)	41.5±2.0	58.0±1.1	0.5±0.5		

calculated are similar to those of the other two geothermometers. The estimated temperatures for M_1 of about 590 °C and of about 576 °C for M_2 metamorphism are essentially equivalent to those calculated by the other geothermometers (table 5). However, inside the shear zone the garnet has appreciable Mn content (about 11 wt. percent; table 5) and the temperatures calculated for both M_1 (565 °C) and for M_2 (543 °C) are lower. The calculated M_2 temperature is approximately 100 °C cooler than that estimated by either the Ferry and Spear (1978) or the Hodges and Spear (1983) method and results from the effect in the calculation of elevated Mn content in the garnet.

Two-feldspar geothermometry indicates that the feldspars underwent subsolidus reequilibration during the waning stage of D_2 deformation. Plotted in figure 11 are calculated temperatures of feldspar equilibration using the

Table 5. Summary microprobe data for garnet and biotite, and temperature estimates for rocks of the Mountain shear zone, northern Wisconsin

[Major oxides in weight percent; cation proportions in terms of number of oxygens; na, not analyzed]

	Outside shear zone					Inside shear zone		
Sample No.	32–85		32B–85			46–85		
Rock type	Dacite		Dacite			Granodiorite gneiss		
	Garnet core	Biotite	Garnet core	Garnet rim	Biotite	Garnet core	Garnet rim	Biotite
SiO_2	37.43	33.84	37.04	36.57	33.49	37.82	37.54	34.80
Al_2O_3	20.71	19.42	21.41	20.97	19.39	20.97	20.81	17.93
FeO	34.39	22.86	35.19	35.12	23.00	25.60	26.61	20.86
MgO	2.00	7.12	1.85	1.78	6.42	2.29	2.32	8.71
CaO	1.71	0.04	1.24	1.38	0.03	1.37	1.08	0.05
Na_2O	0.06	0.14	0.07	0.06	0.10	0.01	0.02	0.11
K_2O	na	9.68	na	na	9.32	na	na	9.90
TiO_2	0.06	1.54	0.05	0.05	1.59	0.09	0.05	1.90
MnO	3.60	0.08	3.22	3.15	0.18	11.36	10.78	0.35
F	na	0.31	na	na	0.24	na	na	0.33
Cl	na	0.07	na	na	0.11	na	na	0.08
		95.10			93.87			95.02
–O=F		0.13			0.10			0.14
–O=Cl		0.02			0.02			0.02
Total	99.96	94.95	100.07	99.08	93.75	99.51	99.21	94.86
No. analyses	3	5	4	4	6	6	8	8
No. oxygens	12	24	12	12	24	12	12	24
Si	2.983	5.759	2.995	2.994	5.776	3.050	3.045	5.884
^{IV}Al	0.017	2.241	0.005	0.006	2.224	0.000	0.000	2.116
Al^{VI}	1.929	1.655	2.025	2.019	1.718	1.994	1.989	1.456
Fe	2.293	3.253	2.380	2.404	3.317	1.727	1.805	2.950
Mg	0.238	1.806	0.222	0.218	1.649	0.276	0.260	2.195
Ca	0.146	0.007	0.107	0.121	0.005	0.118	0.094	0.009
Na	0.010	0.047	0.012	0.009	0.035	0.002	0.003	0.037
K	na	2.101	na	na	2.052	na	na	2.136
Ti	0.006	0.197	0.003	0.003	0.206	0.005	0.003	0.242
Mn	0.216	0.012	0.220	0.219	0.027	0.776	0.741	0.050
F	na	0.166	na	na	0.133	na	na	0.177
Cl	na	0.020	na	na	0.032	na	na	0.024
Temperature estimates (°C)								
Ferry and Spear (1978):								
4.5 kbar	594		595	585		645	607	
6.0 kbar	600		601	591		650	612	
Hodges and Spear (1982):								
4.5 kbar	615		610	601		661	648	
6.0 kbar	620		616	606		667	654	
Perchuck and Lavrent'eva (1983):								
4.5 kbar	589		591	586		565	552	
6.0 kbar	579		581	576		556	543	

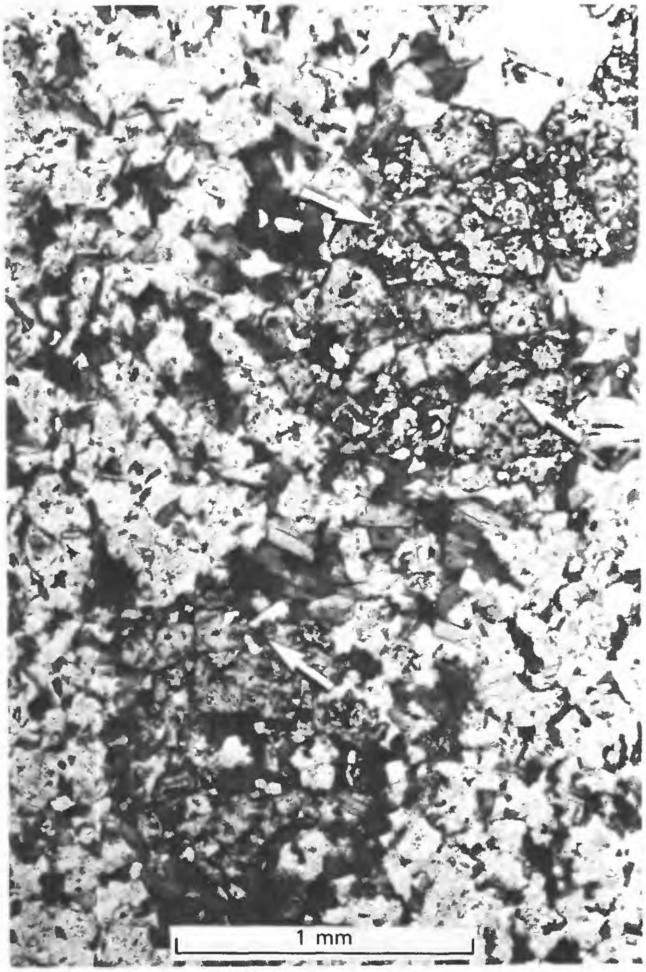


Figure 9. Photograph showing biotite and (M_1) garnet (arrows) in a fine-grained felsic volcanic rock (32–85) outside the shear zone.

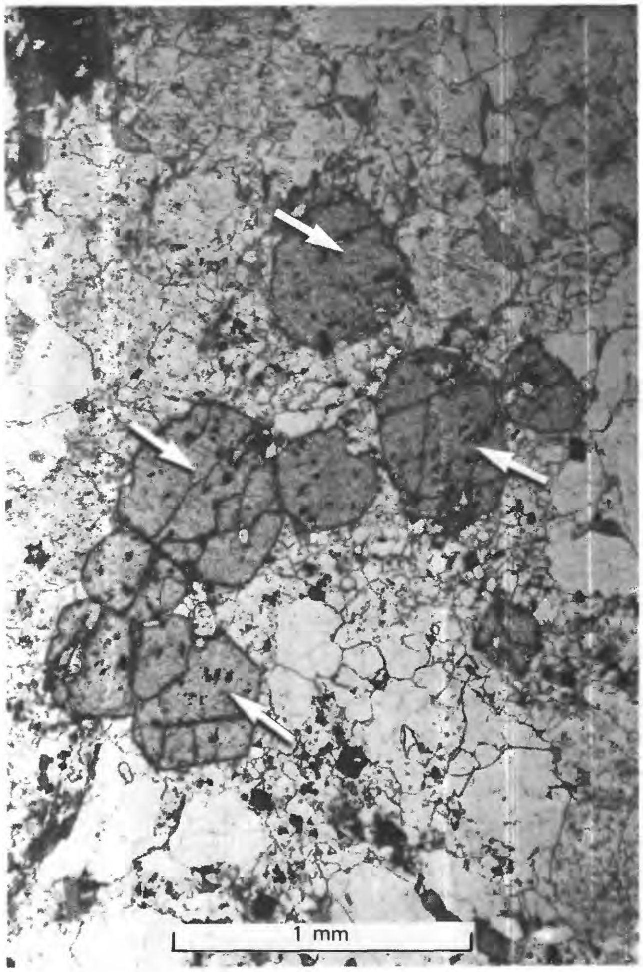


Figure 10. Photograph showing biotite and (M_2) garnet (arrows) in a granitoid rock (46–85) inside the shear zone.

method of Haselton and others (1983). Although deformation and recrystallization were concentrated within the Mountain shear zone, each rock type sampled in the study area has fine-grained recrystallized M_2 quartz-feldspar matrix material. Both inside and outside the shear zone, the temperatures of feldspar reequilibration range down to about 275 °C (table 3), which is well below the peak M_2 thermal conditions of about 650–670 °C recorded in the garnet-biotite data (table 5). The feldspar within the shear zone records a continuum in temperature from just slightly below the peak M_2 (574 °C at 6.0 kbar) to the low similar to that observed outside the shear zone (about 293 °C at 6.0 kbar). Outside the zone occurs a distinct temperature gap between the M_2 temperatures measured in garnet-biotite (575–600 °C) and those recorded in the feldspars (275–450 °C).

The retrograde metamorphism may have caused reequilibration of the Fe-Mg cation exchange within the biotite and garnet pairs analyzed to establish the peak M_2 metamorphic conditions. Therefore, the peak M_2 conditions calculated are minimum estimates. The effects of the retrograde metamorphism should not drastically alter the relative peak M_2 temperatures calculated, however, because the feldspar data indicate that the retrograde overprint is equivalent both inside and outside the shear zone.

CONCLUSIONS

Dynamic M_2 recrystallization within the Mountain shear zone overprinted the regional D_1 deformation fabric. Although the D_2 deformation was localized within the shear zone, the effects can be observed in rocks immediately adjacent to the south. Hornblende geobarometry indicates a relative increase in pressure of about 1.5 kbar as a result of D_2 deformation. Garnet-biotite geothermometry suggests a relative increase in temperature of as much as 60 °C. However, the geothermometer developed by Perchuck and Lavrent'eva (1983), points to a possible slight relative

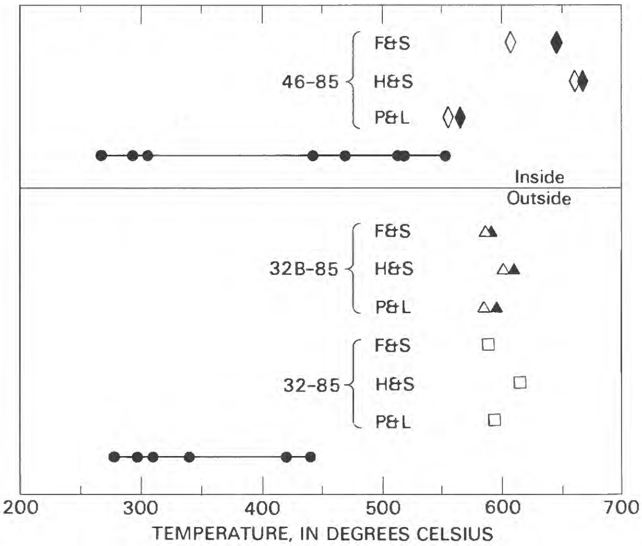


Figure 11. Comparison of the temperature estimates (at 4.5 kbar) from inside and outside the Mountain shear zone. Temperature estimates use the garnet-biotite geothermometers F&S (Ferry and Spear, 1978), H&S (Hodges and Spear, 1982), and P&L (Perchuck and Lavrent'eva, 1983). Solid symbols, core compositions; open symbols, rim. Sample 46–85 (diamond), 32B–85 (triangle), and 32–85 (square). Temperature estimates using the two-feldspar geothermometer indicate (solid dots) that both inside and outside the shear zone, feldspar postpeak M_2 reequilibrated down to about 275 °C.

decrease amounting to about 30 °C during D_2. The lower estimated temperature of the Perchuck and Lavrent'eva (1983) geothermometer results in part from elevated manganese content in the garnet inside the shear zone. However, garnet is found only rarely both inside and outside the shear zone; therefore, these temperatures are only an estimate for the peak M_2 thermal conditions. Reequilibration of the feldspars extended well below subsolidus temperatures. Only within the shear zone does a two-feldspar geothermometer yield temperatures similar to those of the garnet-biotite geothermometry.

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