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Significance of the chloritic breccia zone Bitterroot Dome western Montana

Sharon A. Myers

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SIGNIFICANCE OF THE CHLORITIC BRECCIA ZONE, BITTERROOT DOME, WESTERN MONTANA

by

Sharon A. Myers

B.S. University of Arkansas, 1984

Presented in partial fulfillment of the requirements for the degree of Master of Science

UNIVERSITY OF MONTANA

1986

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Dean, Graduate School

June 5, 1986
ABSTRACT

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Significance of the Chloritic Breccia Zone, Bitterroot Dome, Western Montana (72 pp.)

Director: Donald W. Hyndman

The Bitterroot dome of the Idaho batholith is a probable metamorphic core complex. A zone of frontal mylonites located on the flank of the dome dips approximately 25 degrees east. This zone of amphibolite-facies mylonites is thought to have formed in late Cretaceous time when the 15 km thick Sapphire block moved eastward relative to the Idaho batholith. Mylonite S and C surfaces and abruptly higher regional metamorphic grades on the west indicate top-to-the-east movement.

Affecting and overlying the amphibolite-facies mylonite zone is a zone of later chloritic breccias. This zone ranges from tens of meters to nearly 100 m thick and is concordant to or slightly steeper than the mylonitic foliation. In the chloritic breccia, second-stage greenschist-facies mylonitization overprints the original amphibolite-facies mylonite. Orientations of S and C surfaces, mica "fish", and pull-apart veins formed during retrograde mylonitization show top-to-the-east sense of rotation and eastward extension.

Brecciation and the development of brittle structures followed mylonitization. Slickenside lineations within the chloritic breccias generally plot in east-west girdles and probably formed during flexural slip folding of small folds with north-south hinges. Small folds within the chloritic breccias show top-to-the-east sense of rotation and down-dip movement.

The chlorite-muscovite-bearing breccias formed under conditions of lower temperatures and lower confining stresses than the older mylonites. The chloritic breccias show the transition from deep ductile deformation in late Cretaceous time to shallower brittle deformation which was probably related to reactivation of the amphibolite-facies mylonite zone during Eocene extension.
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CHAPTER I

INTRODUCTION

The chloritic breccia which borders and locally truncates the Bitterroot dome mylonitic shear zone (Fig. 1) records a transition from ductile to brittle deformation related to eastward movement of the Sapphire tectonic block. These shear zone rocks preserve features which formed at distinctly different crustal levels over an extended period of time. The oldest fabric preserved in these breccias is an amphibolite-facies mylonitic fabric which formed during intrusion of the late

Fig. 1. Map of the Bitterroot lobe of the Idaho Batholith (Hyndman, 1980).
Cretaceous Idaho batholith. The second mylonitic fabric is a greenschist-facies fabric which overprints the original amphibolite-facies mylonite. Orientations of S-C surfaces, mica "fish", and pull-apart veins which formed during retrograde mylonitization show a top-to-the-east sense of rotation and eastward extension. The third fabric is a chloritic breccia which formed by upward and outward expansion with late-stage brittle fracturing causing an increase in rock volume. Influx of meteoric waters caused sericitic alteration of feldspars, replacement of biotite by chlorite, and the growth of epidote in tensile fractures. Near the contact between amphibolite-facies mylonite and brecciated greenschist-facies mylonite, small, tensile en-echelon fault sets cut steeply across the mylonitic granite.

The majority of slickenside lineation measurements, taken from the chloritic breccia between Mill and Sheafman creeks and from the region surrounding Gash Creek fall in east-west girdles and are related to the development of folds with north-south hinges.

The purpose of this study is to determine if a structural continuum exists between the development of Cretaceous amphibolite-facies mylonite and the younger brecciated greenschist-facies mylonite. If this is the case, fabrics within the mylonites and the breccias should record similar directional stresses under changing deformational environments.
PREVIOUS INVESTIGATIONS

Several studies have been conducted in the area which includes the chloritic breccia. Lindgren (1904) did a geologic sketch of the complete Bitterroot Range. Langton (1935) studied the area from Sweathouse Creek north. Ross (1952) mapped the geology of the Hamilton 30' quadrangle, from just north of Big Creek to Rye Creek in the south. Chase (1961) mapped in detail the region surrounding Sweathouse Creek. Chase (1968) mapped a north-south-trending elongate rectangle which extended from Kootenai Creek to south of Blodgett Creek. Hyndman (1980, 1983a, b) studied the Idaho batholith and in particular the Bitterroot dome. Mineralogical changes which accompanied mylonitization were studied at the Sweathouse Creek quarry by La Tour and Barnett (1986). Kerrich and Hyndman (1986) have done 180/160 studies to determine thermal and fluid regimes in the Bitterroot Dome-Sapphire Block detachment zone. Waren (1985) mapped the mylonite-chlorite breccia contact between Sheafman Creek and Mill Creek.

PRESENT STUDY

I studied an area of Ravalli Co., Montana at Sweathouse and Gash creeks and between Sheafman and Mill creeks that includes the contact zone between Cretaceous amphibolite-facies mylonites and brecciated greenschist-facies mylonites of unknown age (Fig. 2). I closely examined the contact zone between the Cretaceous mylonite and the chloritic breccia zone, as it is often called.
Fig. 2. Map of the study area.
collected both oriented and nonoriented samples, and studied oriented thin-sections cut perpendicular to foliation and parallel to the mineral elongation lineation. I measured tectonically polished joint surfaces, slickenside lineations, and en-echelon fault set orientations, and tried to determine the senses of rotation for en-echelon fracture sets. I mapped the chloritic breccia zone between Sheafman Creek and Mill Creek and did stereonet analysis of slickenside lineations and en-echelon vein arrays. I especially focused on offsetting and crosscutting features, outcrop patterns, and small folds that provided data on the sense of shear in the fault zone.

U.S. highway 93 provides service to the Victor-Hamilton region. Forest Service roads and trails along with privately owned roads provide access to the study area.

METAMORPHIC CORE COMPLEXES WITH ASSOCIATED CHLORITIC BRECCIAS

Cordilleran core complexes form a discontinuous belt extending from southern Canada southward into Sonora, Mexico (Fig. 3). These complexes have characteristic rock types, fabrics, and structures, and commonly have assymetrical domal or anticlinal profiles. Coney (1980) divided the core complexes into two distinct domains, a metamorphic-plutonic basement terrain and an overlying or adjacent cover. Abrupt changes in rock types, metamorphic
Fig. 3. Location of metamorphic core complexes of the North American Cordillera (adapted from Davis and Coney, 1979).

grade, and deformational structures separate these domains (Coney, 1980).

In the metamorphic-plutonic basement, gently-dipping foliation planes contain mineral lineations with a nearly constant orientation within a given complex. Mineral foliation on domal flanks rarely exceeds 20 to 30 degrees (Coney, 1980). Maximum shortening is perpendicular to the foliation plane and maximum extension parallels the lineation (Compton, 1980). Small late-stage normal faults strike perpendicularly to the extension direction (Davis and others, 1975; Davis, 1975, 1977, 1980) and seem to record progression from ductile to brittle deformation. These fault surfaces contain slickensides which are
subparallel to the mineral lineation.

Overlying the mylonitic rocks of core complexes is a decollement that is also a zone of steep metamorphic gradient. The decollement marks a contrast between dominant ductile behavior below and brittle behavior above. Low-angle younger-on-older faults and extensional listric normal faults cut and brecciate rocks above the decollement (Coney, 1974; Hose and Danes, 1973). Retrograde chlorite alters the extremely brecciated rocks near the decollement (Coney, 1980). These chloritic breccias are found in the Spokane Dome, Washington (Rhodes and Hyndman, 1984), the Newport fault, Washington (Harms and Price, 1983), the Kettle Dome, Washington (Cheney, 1980), the Okanagan Dome, Washington (Hansen, 1983), the Bitterroot Dome, western Montana (Chase, 1973; Hyndman, 1980, 1983a), metamorphic core complexes of southern Arizona (Davis, 1980), and the South Mountains, central Arizona (Reynolds, 1983). The association of chloritic breccias with metamorphic core complexes is almost ubiquitous.

REGIONAL GEOLOGY OF THE BITTERROOT DOME

The 39,000 km Idaho batholith complex located in western Montana and northern Idaho consists of the southern Atlanta batholith and the northern Bitterroot batholith. Proterozoic pre-Belt Supergroup rocks of the Salmon River Arch separate these batholiths (Armstrong, 1975) (Fig. 4). The Idaho batholith intruded directly
east of a suture with the Triassic Seven Devils volcanic arc (Hyndman, 1983b). Dominant rock types within the Idaho batholith are medium-grained, massive to moderately foliated, muscovite-biotite granite and granodiorite (I.U.G.S. classification). The 12-16 km wide western border zone of the batholith is a more mafic gneissic tonalite. Granodiorite comprises a 10-20 km wide border zone surrounding much of the Idaho batholith, enclosing the granite in the batholith interior (Hyndman, 1983b). Country rocks surrounding the Idaho batholith comprise Proterozoic Belt metasediments and pre-Belt basement orthogneisses (Hyndman, 1983b). Jurassic-Cretaceous pre-batholith, sillimanite-zone metamorphism, surrounds the northern half of the Idaho batholith and extends for a few to several kilometers beyond the batholith contact (Hyndman, 1983b).

The Bitterroot dome is located in the northeastern portion of the Bitterroot batholith. Chase (1973, 1978), Chase and others (1983), Hyndman (1980, 1983a, b), and Kerrich and Hyndman (1986) identified the Bitterroot dome as a metamorphic core complex. A zone of frontal mylonites located on the east flank of the dome dips at approximately 25 degrees to the east. The amphibolite-facies mylonitic fabric presumably developed as the Sapphire tectonic block moved eastward relative to the Bitterroot batholith (Hyndman and others, 1975; Hyndman, 1980). S and C surface orientations and the abruptly
Fig. 4. Map of the Idaho batholith and related plutons (simplified from Hyndman, 1980).
higher metamorphic grades on the west confirm this sense of movement (Hyndman, 1980, 1983a, b). The domal character of the Bitterroot dome may result from isostatic adjustments in response to unloading of the Sapphire block. Brittlely fractured chlorite-bearing fault rocks lie above and sometimes crosscut the Bitterroot mylonites.

GEOCHRONOLOGY OF THE BITTERROOT DOME

Radiometric ages of granitoid rocks in the Bitterroot lobe of the Idaho batholith and surrounding metamorphosed country rock range from 66 to 81 m.y. (Hyndman, 1983b). Chase and others (1978) date main-stage batholith emplacement at 66+/-10 m.y. Unloading of the Sapphire block and development of frontal mylonites in the Bitterroot dome appears to coincide with or immediately follow emplacement of the Idaho batholith (Hyndman and others, 1975; Hyndman, 1980; Kerrich and Hyndman, 1986; La Tour and Barnett, 1986).

Within and adjacent to the Bitterroot dome are large areas of Tertiary volcanic rocks and shallow plutons. These plutons are quartz monzonitic to granitic in composition (I.U.G.S. classification), pink in color, and contain smoky quartz and miarolitic cavities. Eocene plutons and volcanic rocks within and surrounding the Bitterroot dome include: the Bungalow Stock (K-Ar, 43 m.y.), the Running Creek batholith (K-Ar, 49 m.y.), the Whistling Pig pluton; the Lolo batholith, and the Challis volcanics (Bennett, 1980). The emplacement of epizonal
Eocene intrusions within the Bitterroot dome indicates that 10-15 km of rock initially overlying the Bitterroot dome was removed previous to middle Eocene.

Unloading of the Sapphire block before the Eocene may explain the removal of 10-15 km of overburden. However, zircon lower-intercept ages ranging from 52+/-1 m.y. to 48+/-1 m.y. obtained from sheared plutons in the mylonite zone by Chase and others (1983) remain difficult to explain. Incremental 40 Ar/39 Ar spectra undertaken by Garmezy and Sutter (1983) on the Bitterroot mylonites suggest that mylonitization began 45.5 to 47.5 m.y. ago and continued for approximately 2 m.y. Thus they argue that mylonitization was not related to "initial intrusion and crystallization of the batholith." The above workers argue that the late Cretaceous Sapphire Block thrusts (Hyndman and others, 1975; Hyndman, 1980) are older than what they consider to be Tertiary Bitterroot dome mylonites.

In studies conducted to determine the conditions of deformation within the mylonite zone, La Tour and Barnett (1986) observed the following:

1. Better developed microstructures within the mylonitic pegmatites suggest they deformed in a more ductile manner than the granites, contrasting with the more competent behavior usually displayed by pegmatites within mylonite zones.

2. The occurrence of pegmatites as streaks and layers
entirely parallel to granitic foliation.

These findings suggest that the foliation existed within the granite prior to pegmatite crystallization, and that the pegmatite was deformed late in the crystallization of the batholith when the pegmatite was more easily deformed than granite due to high water content (La Tour and Barnett, 1986).

Two-feldspar geothermometry studies by La Tour and Barnett (1986) yielded a minimum deformation temperature of 500 °C, which is consistent with expected late-stage crystallization temperatures within the Bitterroot dome. These findings are substantial because they indicate that initial amphibolite-facies mylonitization occurred in the latest stages of crystallization of the Idaho batholith and probably during the late Cretaceous rather than during the lower-grade greenschist-facies conditions of the chloritic breccias.

The present study shows that a retrograde mylonite event occurred under greenschist-facies conditions. The timing of this event is bracketed between late Cretaceous mylonites and Tertiary extension which may have opened the Bitterroot Valley (Fig. 5).
### CHAPTER II

**CHLORITIC BRECCIAS OF THE BITTERROOT DOME**

The zone of later chloritic breccias overlies and locally cuts the amphibolite-facies mylonite zone (Fig. 6). Contained within the zone of brecciation is a thin greenschist-facies mylonite which overprints the earlier high-grade mylonite (Fig. 7). The greenschist-facies mylonite has in turn been brecciated under lower pressure conditions. Alteration minerals within the breccias include muscovite, quartz, chlorite, epidote, and sericite. This zone of chloritic breccias which includes the greenschist-facies mylonites, ranges from several tens

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Fig. 7. Schematic diagram of the relationship of the chloritic breccia with the underlying amphibolite-facies mylonite (adapted from Hyndman, 1983a).
of meters to nearly 100 m thick and is concordant to or slightly steeper than the mylonitic foliation. These green rocks were retrogressively metamorphosed, deformed under ductile-brittle transition conditions, and brecciated at near-surface conditions. Mylonitic S and C surfaces within these greenschist-facies mylonites indicate a top-to-the-east shear direction. Transgranular fractures and chloritic alteration pervasively modified the amphibolite-facies mylonites, hence the often used name chloritic breccia.

En-echelon fracture sets found in the narrow transition zone between the brecciated greenschist-facies mylonite and the underlying amphibolite facies-mylonite (Fig. 7) dip steeply west and are filled with epidote, quartz, and chlorite with a surrounding halo of white mica a few mm thick.

The chloritic breccias flanking the Bitterroot dome are the erosional remnants of a zone which probably extended over a larger zone of movement. Geomorphologic studies done by Pain (1985) on remnant chloritic breccias in Arizona indicate that once a fine-grained resistant upper layer has been eroded through, the breccias below are rapidly removed. In many cases, the Oligocene chloritic breccias of the southwest core complexes have been completely removed by erosion.
GENERAL CHARACTERISTICS OF MINERALOGY AND TEXTURES WITHIN THE CHLORITIC ROCKS

Mineralogy in the Bitterroot dome chloritic rocks places the conditions of alteration in the greenschist-facies of metamorphism (Turner, 1981, p. 342). The presence of chlorite, epidote, and muscovite and the textures present within these rocks identify them as greenschist-facies mylonites (c.f.: Simpson, 1985).

Tabulation of the mineralogy of the chloritic breccias (Appendix A) shows a change from the original mineral and textural assemblage of granitic to granodioritic mylonites and coeval pegmatites (Hyndman, 1980, 1983b; La Tour and Barnett, 1986) to an assemblage that includes muscovite, chlorite, epidote, and white mica with a texture that developed during the transition from ductile to brittle conditions.

The extent of alteration within the chloritic mylonites and breccias is variable (Appendix A) throughout the zone. Contacts between the amphibolite-facies mylonites and greenschist-facies rocks are seemingly sharp over a range of 2-5 m. This may vary though, because only a few areas along the contact are well exposed. Alteration within the secondary greenschist-facies mylonites is variable throughout their area of exposure. With this comes variability in degrees of exposure, mylonitization, and brecciation. The most obvious manifestation of these variations in the map area is prominent resistant outcrops with shattered appearances.
(Fig. 8). Silica contents within these resistant outcrops are elevated and a secondary mylonite fabric has developed (Fig. 9) which shows a grain-size reduction beyond that of the original mylonitic fabric. Good examples of these highly deformed and retrograded rocks are located on the south side of Sheafman Creek.

Fig. 8. Photo of the shattered appearance of the greenschist-facies mylonites.
Fig. 9. Photomicrograph of greenschist-facies mylonitic fabric. Field of view is 8.5 mm.

GREENSCHIST MYLONITE MINERALOGY AND TEXTURE

Well-developed micromylonites:

Rocks within the silica-rich greenschist-facies mylonites (Fig. 7), hereafter called "micromylonites" because of their fine grain size, are composed of approximately 55 percent quartz grains ranging in size from 0.1 by 0.1 mm up to 0.1 by 0.2 mm with undulose extinction. Quartz deformation occurred by dislocation creep, producing well developed ribbon structures (Fig. 10a). Recovery growth has led to the formation of subgrains and recrystallized new grains (Fig. 10b). Quartz and muscovite define the mylonitic foliation.
Fig. 10a. Photomicrograph of quartz ribbons. East is to the right. Field of view is 3.5 mm.

Fig. 10b. Photomicrograph of subgrains of quartz. Field of view is 3.5 mm.
Orthoclase grains comprise 20 percent of these samples of greenschist-facies micromylonites. They have been slightly rotated and are surrounded by fine-grained quartz and orthoclase. Orthoclase grains range in size from 0.2 by 0.3 mm to 2 by 4 mm. The orthoclase grains are often crushed and rounded suggesting that they were subject to brittle mechanisms of grain size reduction (Fig. 10c). Fractures and microfaults within the orthoclase grains are filled with quartz. The orthoclase grains are slightly to pervasively altered to sericite.

Fig. 10c. Photomicrograph of crushed and rounded orthoclase grains. Field of view is 8.5 mm.
Plagioclase ranges in size from 0.2 to approximately 2 mm in size and comprises nearly 12 percent of these rocks. Grains are crushed, rounded, and crossed by fractures and microfaults filled with quartz. Albite twins are bent and contorted (Fig. 10d). Plagioclase is slightly to pervasively altered to sericite.

Muscovite grains comprise 10 percent of most samples and range in size from tiny stringers defining a C-surface foliation up to grains of 0.4 by 1 mm across. Larger muscovite grains are strung out and form well developed mica "fish" (Fig. 10e).

Approximately 4 percent of these rocks are comprised of chlorite which is less than 0.1 mm by 0.2 mm in size. Fine-grained chlorite is oriented along the C-surface (Fig. 10f.) which is defined by muscovite.

Fine-grained epidote less than 0.1 by 0.1 mm in size fills a series of steeply dipping crosscutting fractures and comprises 2 percent of the sample (Figs. 7 and 10g). Grains in vein centers appear random in orientation, while those along the margins are oriented with (001) perpendicular to the margins.

These veins probably developed after greenschist-facies mylonitization and during periods of high water pressure based on a study of tensile fractures by Sibson (1981).

Opaque minerals make up less than 1 percent of these rocks and tend to be associated with chlorite.
Fig. 10d. Photomicrograph of bent plagioclase grain. Field of view is 3.5 mm.

Fig. 10e. Photomicrograph of mica "fish". East is to the left. Field of view is 8.5 mm.
Fig. 10f. Photomicrograph showing parallelism of chlorite. Field of view is 8.5 mm.

Fig. 10g. Photomicrograph of epidote filled veins. Field of view is 3.5 mm.
More-poorly developed greenschist-facies mylonites:

Less altered and more texturally poorly developed greenschist mylonites tend to vary somewhat in mineralogy and texture from well developed micromylonites. In these rocks, the original amphibolite-facies mylonite remains clearly visible but is affected by deformation under lower temperature and pressure conditions. Quartz comprises 35 to 40 percent of these rocks and ranges from 0.1 by 0.1 mm up to 0.4 by 0.4 mm in size. Well developed fine-grained quartz ribbons (Fig. 11a) indicate recrystallization of quartz. Undulose extinction indicates that the quartz has been strained after recrystallization. Fine-grained quartz fills fractures within both orthoclase and plagioclase.

Orthoclase forms 15-25 percent of the samples and ranges in size from 0.5 by 2 mm up to 2.5 by 3 mm. Orthoclase is intensely fractured and faulted (Fig. 11b), even more so than plagioclase. Orthoclase grains are slightly to pervasively altered to sericite.

Plagioclase varies in amount from 25 to 35 percent of the samples. Sizes range from 0.2 by 0.5 mm up to 2 by 2 mm. Albite twins are in some cases bent, indicating that plagioclase may have undergone a small amount of ductile deformation. Fractures within plagioclase grains are offset along twin planes (Fig. 11c). Plagioclase is slightly to pervasively altered to sericite.
Fig. 11a. Photomicrograph of quartz ribbons. East is to the right. Field of view is 8.5 mm.

Fig. 11b. Photomicrograph of a microfaulted orthoclase grain. East is to the right. Field of view is 3.5 mm.
Muscovite comprises approximately 4 percent of these samples and ranges in size from 0.5 by 0.7 mm up to 2 by 2 mm. Larger muscovite grains are kinked and torn.

Chlorite is fine-grained, ranging in size from less than 0.1 by 0.2 mm increasing to 0.5 by 1 mm and constituting 2 to 4 percent of most samples. Smaller chlorite grains are oriented with (001) cleavage lying in an east-west direction (Fig. 11d). Larger grains are found as non-oriented inclusions in feldspar grains and are thought to have replaced biotite.

Apatite and an opaque mineral are present as accessory minerals and tend to be associated with chlorite. Epidote is present in some but not all steeply-
dipping crosscutting veins.

INCRIMINATING TEXTURES

The greenschist-facies mylonites are foliated and the foliation is marked by secondary S-C surfaces formed by quartz, muscovite, and chlorite (Figs. 10e, 11d, and 12a). Relict orthoclase and plagioclase grains have undergone grain-size reduction by microfaulting and microfracturing (Figs. 11b, 11c, and 12b).

The behavior of plagioclase and orthoclase in these secondary greenschist-facies mylonites agrees with the deformational behavior of feldspar grains deformed under lower greenschist-facies conditions as observed by Simpson
Fig. 12. Microstructures useful in determining the sense of shear in mylonites. (a) orientation of S-C surfaces. (b) orientation of quartz ribbons. (c) orientation of mica "fish" (Lister and Snoke, 1984; Simpson and Schmid, 1983; White and others, 1980).
Feldspar grains are crosscut by veins containing quartz, and these veins show evidence of eastward pull-apart and top-to-the-east shear (Fig. 11c). Fine-grained chlorite is scattered throughout, giving the rocks a green appearance. Relict primary muscovite is kink banded and torn as is any remaining biotite. Mica "fish" were produced by boudinage and microfaulting of muscovite grains (Fig. 12c) (c.f.: Lister and Snoke, 1984). Biotite shows replacement by chlorite and is not in equilibrium with the final mineral assemblage. In most rocks, (001) cleavage of the chlorite has a preferred east-west orientation.

The presence of extensional pull-apart veins within these rocks (Figs. 10g and 11c) and the sense of shear producing mica "fish" (Fig. 12c) and quartz ribbons (Fig. 10a and 12b) indicate that the deformation controlling these secondary mylonites involved top-to-the-east rotation (c.f.: Simpson and Schmid, 1983; Lister and Snoke, 1984; White and others, 1980).

The (lower) greenschist-facies mylonites of the Bitterroot dome show a ductile-brittle microstructural assemblage which includes plastically deformed quartz, kink banded biotite and muscovite, formation of mica "fish", and brittlely fractured feldspars. This structural assemblage places deformation of these rocks in a ductile-brittle transition zone as in a study by Simpson (1985). White and others (1980) note that feldspar grains
undergo limited ductile behavior in low-grade greenschist-facies mylonites and that intracrystalline plastic behavior of plagioclase, excluding albite, and of alkali feldspar occurs in amphibolite-granulite facies mylonites.

Simpson (1985) notes that the transition to textures indicative of completely ductile rock behavior occurs within the middle to upper greenschist-facies range. Sibson (1977) places the ductile-brittle transition at the onset of the lower greenschist-facies. Three important factors affecting the onset of ductile behavior and discussed in White and others (1980) are:

1. Fluid content.
2. Strain rate.
3. Temperature.

Lower plate rocks of the Snake Mountains, Nevada were deformed and metamorphosed in the amphibolite-facies (Miller and others, 1983). The high-grade metamorphic fabric was cut or transposed into a penetrative subhorizontal lower metamorphic-facies fabric. This later deformation was accompanied by a lower greenschist-facies metamorphic event and occurred at depths of 6-7 km. At the lower temperatures and pressures which accompanied deformation, textures which developed included ductile behavior of quartz, brittle behavior of feldspars, and brittle behavior of micas in the form of kink-banding or mechanical rotation of mica into parallelism with the new fabric.
The following (Table 1) is a summary of the characteristic microstructures of typical granitic mylonites of lower and middle greenschist-facies observed by Simpson (1985). The secondary mylonites of the Bitterroot dome contain textures of deformation compatible with their lower greenschist-facies mineralogy.

The best developed greenschist-facies mylonite fabrics within the chloritic breccias also are the finest grained. Mylonites formed at lower grades tend to be finer grained than high-grade mylonites (White and others, 1980). The same study points out that the preservation of small grains indicates that little growth has occurred after recrystallization.
Table 1. Table summarizing microstructures common to lower and upper greenschist mylonites (after Simpson, 1985).

<table>
<thead>
<tr>
<th>Lower Greenschist</th>
<th>Upper Greenschist</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Multiple fracturing of orthoclase grains in a ductile quartz ribbon matrix</td>
<td>1) Orthoclase grains contain planar zones of finely recrystallized feldspar.</td>
</tr>
<tr>
<td>2) Grain size reduction of feldspars by microfracturing and microfaulting.</td>
<td>2) Plagioclase shows evidence of low temperature plasticity with the formation of kink bands and minor recrystallization along curved microcracks in the host mineral.</td>
</tr>
<tr>
<td>3) Progressive rotation of microfault blocks resulting in greater separation of feldspar fragments along the foliation with an increase in bulk strain.</td>
<td>3) Quartz retains original ribbon texture but has undergone recovery to produce polygonal elongate grains, with planes of flattening facing the direction of shortening.</td>
</tr>
<tr>
<td>4) Quartz grains deformed mainly by dislocation creep, producing ribbon structure with deformation bands at a high angle to the ribbon boundaries.</td>
<td>4) Deformation of mica by kink band formation. Ribbon structure develops in biotite with ilmenite and and finely recrystallized biotite along ribbon boundaries.</td>
</tr>
<tr>
<td>5) Mica showing open, simple kink bands.</td>
<td></td>
</tr>
</tbody>
</table>
MACROSCOPIC BRITTLE STRUCTURES

Structural features within the map area include numerous joints, slickenside lineations, small folds, small normal faults, and en-echelon vein arrays.

Although a systematic measurement of joint surfaces was originally planned for this study, it became apparent during mapping that most joint surfaces contained slickenside lineations and were surfaces where movement and readjustment had occurred. Slickenside lineations comprised both a mineral lineation and a limonite stain lineation that overprinted the mineral lineation (Fig. 13). Recorded mineral and limonite stain lineations taken in the Gash Creek region, plotted in slightly differing orientations on separate stereonet projections (Figs. 14a and 14b). Both of these stereonet projections show an east-west girdle related to folding during brittle deformation (Fig. 14c). Most of the lineations are perpendicular to a north-south fold hinge, but vary in trend and plunge. The limonite lineations record a small deviation from the direction of movement which created the earlier mineral lineations. The presence of limonite may indicate that temperature and fluid conditions changed somewhat between the time of formation of the earlier mineral lineations and the later limonite lineations.

Analysis of a stereonet projection from the Sheafman-Mill creeks region shows a similar east-west girdle of slickenside lineation measurements (Fig. 15).
Slickenside lineations in the study area may in part be remnants of the original mylonite lineation, but a large number appear to have developed during flexural slip folding of the chloritic breccia. The lineation may be referred to as "serpentinite" style. That is, the lineations often surround and bend around fractured blocks of rock ranging in size from a few to tens of centimeters across.

The development of slickensides by streaking of limonite stains also suggests that post-mylonitization generation of slickensides did occur within the chloritic breccia.
Fig. 14a. Lower hemisphere equal area stereonet projection of limonite-stain lineations from Gash Creek. Contours, 9%, 6%, 3% per 1% area.
Fig. 14b. Lower hemisphere equal area stereonet projection of mineral lineations from Gash Creek. Contours, 8%, 6%, 3% per 1% area.
Fig. 14c. Diagram showing relationship of slickensides to folds.

Structures within the map area which are most indicative of the direction of movement which deformed the breccias are a series of drag folds which are best exposed on the north side of Gash Creek along a Forest Service road cut and in the Sheafman-Mill creeks area. These folds (Figs. 16a and 16b) probably began as joint surfaces and were folded during top-to-the-east rotation and downdip movement that occurred during late-stage eastward unloading of the Sapphire block.

Several small high-angle normal faults are located in the study area. One such fault near the mouth of Sweathouse Creek occurs in a highly deformed and
Fig. 15. Lower hemisphere equal area stereonet projection of lineations from Sheafman-Mill creeks. Contours, 10%, 8%, 6%, 4%, 2% per 1% area.
Fig. 16. Photos of small folds within the study area. East is to the right. (a) small fold near Sheafman Creek. (b) small fold near Gash Creek.
slickensided region of the breccia, and the fault surface itself is heavily slickensided and gouged (Fig. 17a). The fault surface strikes northeast and dips 60 degrees southeast. A normal fault in the Sheafman-Mill creeks area dips 45 degrees to the east with a strike of 296 degrees. A one-meter-wide zone of intense brittle fractures oriented perpendicular to the fault is located above and below the fault plane (Fig. 17b).

In Sweathouse Creek Canyon, a series of en-echelon vein arrays have developed below and slightly cross-cutting the mylonite-chloritic breccia contact (Fig. 7). The majority of the vein sets dip steeply westward to
nearly vertical. Individual fractures within the sets intersect at small angles, the orientations of the overall vein sets which include them.

Veins within en-echelon vein arrays located in Sweathouse Creek Canyon, are essentially parallel to one another (Fig. 18b) suggesting that they originated as tensile fractures and not by shear which often causes significant vein rotation (Beach, 1975). Non-rotated to slightly rotated veins (Fig. 18a and 18b) located randomly throughout the same zone support the possibility of dilation as a mechanism for the development of the veins in Sweathouse Creek. Individual veins within larger vein sets (Fig. 18b) dip steeply east and west, although the majority dip steeply east. Pressure solution processes were not active in these arrays, indicating that the veins simply dilated perpendicular to their lengths. Elsewhere within the study area, pressure solution features are also absent.

The steronet plot (Fig. 19) of orientations of en-echelon fracture sets shows small variations in dip direction and slightly more in strike direction. Strike direction point scatter may have been amplified by error in measurement on a smooth vertical surface perpendicular to the orientation of the veins.

If these vein arrays developed as a result of extension, as this study suggests, then their orientations could be used in determining the orientation of the
Fig. 17. Normal faults within the study area. (a) small fault near Sweathouse Creek. East is to the right. (b) small fault near Sheafman Creek. East is to the left.
Fig. 18. Diagrams of tensile fractures found in the mylonite/chloritic breccia contact. (a) non-rotated fractures. (b) slightly rotated fractures. (c) pull-apart fractures.

principle stresses at the time of brittle failure (Hancock, 1985) (Fig. 20). The direction of extension is the direction of sigma 3 and the principle stresses are located perpendicular to the extension direction. Making use of the orientation of the vein arrays, sigma 3 would be nearly horizontal in a northwest-southeast direction.

Larger pull-apart veins (Fig. 18c) up to 4 mm across, located in Sweathouse Creek Canyon, kinked under shearing deformation then pulled apart to give veins with a parallelogram shape. This is in keeping with a model of eastward extension.
Fig. 19. Lower hemisphere equal area stereonnet projection of poles to en-echelon vein arrays. Contours, 10%, 8%, 5%, 3% per 1% area.
A study by Sibson (1981) suggests that arrays of parallel extension fractures and veins such as those at Sweathouse Creek are often associated with exhumed faults and may be the product of repeated hydro-fracturing. The inward growth of epidote prisms and other minerals from the margins of veins is characteristic of a progressively opening fissure (Sibson, 1981). He also suggests that fluid flow occurs intermittently through veins and each episode is ended as deposition of hydrothermal minerals destroys fracture permeability.

When fluid pressure builds and exceeds sigma 3, hydro-fracturing occurs. Once the veins fracture, fluid
pressures should fall off to hydrostatic, resulting in strengthening of the fault and the beginning of a new cycle of hydro-fracturing (Sibson, 1981).

CHAPTER III

P,T CONDITIONS DURING DEFORMATION

The transition from granitic amphibolite-facies mylonite to greenschist-facies brecciated mylonite occurs quite sharply over 1 to 2 meters where the contact is exposed. The transition zone is marked by a very narrow transition from ductile to brittle deformation. Geothermometry studies by Kerrich and Hyndman (1986) have shown that the late Cretaceous, high-grade mylonites deformed under ductile conditions at temperatures near 500 °C and that fluids associated with the mylonites were either of magmatic or high temperature metamorphic origin, whereas the chloritic breccia zone a small distance above deformed at the ductile-brittle transition under lower temperatures with meteoric fluids.

Fluid inclusion studies conducted by Kerrich and Hyndman (1986) on the chloritic breccias yield alteration temperatures of 250 °C to 370 °C. Calculated delta O of fluids in apparent equilibrium with albite and chlorite at the determined high and low alteration temperatures, range from -11.8 at 250 °C to -7 at 370 °C. This implies that alteration of the chlorite-bearing rocks occurred during an influx of meteoric waters, giving a negative range of
$^{18}$O values which is indicative of terrestrial meteoric waters. Bitterroot dome chloritic breccias were deformed first under ductile, possibly lower greenschist-facies conditions, followed later by brittle deformation. These deformational events occurred subparallel to the same zone of shear which deformed the Cretaceous amphibolite-facies mylonites.

Studies of the Newport fault by Harms and Price (1983) and of the South Mountains of Arizona by Reynolds (1983), found a continuum in stress orientation across the ductile-brittle transition. Reduction of confining pressure along with a reduction in temperature led to formation of lower-grade mylonites followed by brittle fracture, allowing the incursion of meteoric waters along these fractures (Kerrick and others, 1984). Paterson (1978, p. 120) suggests that transgranular fracturing increases volume and provides permeability for movement of meteoric waters. Flow of meteoric water in large convective cells may have played an important role in the cooling of the Idaho batholith (Taylor, 1978; Criss and Taylor, 1983; Kerrich and Hyndman, 1986).

Beach (1980) notes that metamorphism in shear zones is often retrograde and that the mineral assemblages produced are usually more hydrated than the original assemblage, suggesting that secondary assemblages formed at lower P,T conditions and higher water contents than preexisting rocks within the shear zone. Transitional ductile-brittle
conditions are exposed in the Santa Rosa mylonite zone, California (Simpson, 1985), the Alpine Fault zone (White and White, 1983), and the Lewisian Complex, Scotland (Beach, 1980).

Sibson (1977) notes that variations in styles of deformation and fault rock types vary with crustal levels (Fig. 21). Rocks at deep and mid-crustal levels deform plastically while rocks at shallow crustal levels deform brittlely in seismically active fault zones. Between these two end-members lies a poorly defined region known as the ductile-brittle transition. Changes in textural types between the two styles are gradational, and some
fault rocks contain mixed textures (White and White, 1983). An important starting point in determining the conditions under which the greenschist-facies mylonites formed is knowing that it is not until lower greenschist-facies conditions are reached that penetrative fabrics appear in rocks of granitic composition (Sibson, 1977). Sibson (1977) points out that the isotherm which defines the onset of greenschist metamorphic conditions was estimated by Turner (1968, p. 366, 1981, p. 420) to lie somewhere between 250-300 °C for P load=PH2O at 2 kb. This is probably the low-temperature limit for extensive crystallographic fabric development in quartz.

Also important in determining conditions during deformation is the amount of water present. The introduction of water into a shear zone may lead to retrograde metamorphism (Beach, 1980) and the quartz may take up small amounts of water (White and others., 1980). The transition to plasticity in quartz is dependent upon water to promote what is known as hydrolytic weakening (Sibson, 1977). Estimates by Sibson on the depth of the 300 °C isotherm which marks the lower temperature boundary for greenschist facies conditions, as well as the location of the ductile transition within quartzofeldspathic crust, are based upon a "normal" geothermal gradient of 20-30 °C/km. Using this gradient, the transition would lie between 10 and 15 km. Considering the thermal history of the Idaho batholith region, with both Cretaceous and
Eocene thermal events, the 300 °C isotherm was probably at considerably less than 10 km depth during the formation of the greenschist-facies mylonites. As noted above, temperature of retrograde mineralization of the chloritic breccias was between 250 and 370 °C and accompanying fluids were meteoric in origin. Hence the rocks were fairly shallow but were hot enough to deform in a ductile manner. Brecciation of the greenschist-grade rocks occurred as the rocks were brought to shallow levels in the crust where deformation became brittle.

CHAPTER IV
TIMING OF CHLORITIC BRECCIA FORMATION

Spencer (1984) conducted a study of the role of tectonic denudation in the warping and uplift of low-angle normal faults. In this study, he points out that based on geologic evidence, the antiformal uplifts characteristic of metamorphic core complexes are at shallow depths during and immediately after low-angle normal faulting. Spencer (1984) refers to a study by Pashley (1966) in the Catalina-Rincon Mountains, which found that clasts of mylonitic rock and chloritic breccia derived from lower plate rocks were present in the detached and tilted Rillito Conglomerate at the base of the mountain range. This indicates that lower plate rocks reached surficial levels during or immediately following tectonic denudation. The statement holds true only if the mylonite/breccia-bearing conglomerate is nearly the same
age as the lower plate rocks from which they were derived.

As for the timing of formation of the Bitterroot dome mylonites and subsequent erosion of mylonitic rocks and chloritic breccias, several studies suggest late Cretaceous formation of the amphibolite-facies mylonites and the thrust faults located east of the Sapphire tectonic block. These include:

1. Late Cretaceous dates for the crystallization of the Idaho batholith (Hyndman, 1983b).

2. Well developed mylonitic textures within pegmatites of the high-grade mylonite indicating that the Idaho batholith was in the late stages of cooling during mylonitization (La Tour and Barnett, 1986).

3. Intrusion of pegmatites along planes parallel to the mylonitic foliation suggesting that a directional fabric was present in the Bitterroot dome granites and granodiorites during late-stage crystallization of the batholith (La Tour and Barnett, 1986).

4. Oxygen isotope studies by Kerrich and Hyndman (1986) indicating a magmatic or high-grade metamorphic source for fluids active during amphibolite-facies mylonitization.

5. Dates constraining movement on thrust faults east of the Sapphire block between 72.0 and 78 m.y. of age (Hyndman, 1980).

6. The presence of shallowly emplaced Eocene plutons and volcanic rocks within deep-seated rocks of the Bitterroot dome indicates that 10 to 12 km of cover had been removed prior to intrusion of these plutons (Hyndman, 1980).

Bickford and others (1981), Chase and others (1983), and Garmezy and Sutter (1983) disagree. Based on U-Pb zircon studies and 40 Ar/39 Ar studies Bickford and others (1981), Chase and others (1983), and Garmezy and Sutter (1983), contend that the mylonites formed during the
Eocene and are totally unrelated to the Cretaceous thrusts to the east. All workers do agree on Eocene dates for the emplacement of shallow granitic intrusions and their associated volcanics within and upon the Idaho batholith. Formation of the greenschist mylonites and the brittle deformation that later affected them are not well dated, but probably occurred during the Tertiary because Eocene volcanic rocks capping the amphibolite-facies mylonites in the southern Bitterroot Valley have been subjected to similar pervasive brittle deformation (Hyndman, 1980).

Mylonites that developed within the rocks of the Bitterroot dome were possibly related to a low-angle normal shear zone that tectonically denuded the Bitterroot dome. This shear zone, along with a component of thermo-gravitational movement as proposed by Hyndman (1980), was probably responsible for late Cretaceous thrusts east of the Sapphire block. When the Sapphire block moved to the east it probably did not leave the present Bitterroot Valley in its wake. The possibility remains, however, that the valley first developed in the late Cretaceous during a wet period so that sediments were not deposited but were carried through by streams leaving no definitive evidence (Alt, personal communication, 1986). Extension in the Eocene may have served to open the valley farther. The presence of thrusts related to eastward movement of the Sapphire Block (Hyndman 1980) indicates that the allochthonous block was moving eastward rather rapidly to
have created these thrusts. Considering that the oldest sediments in the Bitterroot Valley are Tertiary and that post-middle Eocene extension is probably responsible for the numerous intermontane valleys which dot southwestern Montana and eastern Idaho (Hyndman and others, 1986), post-middle Eocene extension may also have been responsible for the renewed opening of the Bitterroot Valley. This is further supported by the presence of brecciated Eocene volcanics which lie upon the east-dipping mylonite of the Bitterroot dome (Hyndman, 1980) and by studies undertaken by Rehn and Lund (1981) on the Eocene Painted Rocks pluton, which indicate extensionally controlled emplacement.

The younger episode of extensional shear took advantage of a weak zone in the crust caused by the presence of the late Cretaceous mylonite zone. White and Brefan (1985) argue that a mylonite zone in the crust is weaker than the parent rock due to a number of factors, including localization of fluid flow leading to retrograde metamorphism and development of fabric which increases ease of glide and sliding of grains. Also, the narrow thickness of the lower-grade shear zone as compared to the thickness of the amphibolite-grade shear zone corresponds to a more localized zone of shear and to the lower P,T conditions under which these rocks were deforming. As the shear zone reached near-surface brittle conditions, movement probably became even more localized into a nearly
planar fault surface separating lower plate rocks below from upper plate rocks of the Sapphire block above.

TECTONIC MODELS

Two tectonic events appear to have affected the Bitterroot dome mylonite zone. The first event is evidenced by an amphibolite-facies mylonite and probably developed during the late Cretaceous due to eastward movement of the Sapphire tectonic block. Hyndman (1980) proposed that the Sapphire block moved off a metamorphic high to the west. Movement was quite rapid, geologically speaking, and was responsible for thrust faults which bound the east side of the Sapphire block (Hyndman, 1980). The Bitterroot dome had to have been at least partially unroofed during this movement in order for epizonal plutons to have intruded deep levels of the batholith during the Eocene. Results published by several workers further support a late Cretaceous shearing event for the development of amphibolite-facies Bitterroot dome mylonites (Table 2a).

It is not conclusively known whether or not this eastward movement of the Sapphire block was thermo-gravitationally motivated (Hyndman, 1980), occurred during the development of a ramp anticline over a major eastward moving thrust (Sears, personal communication, 1986), or was due, at least in part, to low-angle detachment faulting as proposed by Wernicke (1981, 1985) for the development of core complexes in Nevada and Arizona (Fig.
Miller and others (1983) suggest that extensional detachment faults can develop between brittlely extending rocks above and underlying plastically extending rocks and intrusions, as would have been the case with the Bitterroot lobe and the overlying Sapphire block during the late Cretaceous. These low-angle, "thin-skinned" normal faults are analogous to thin-skinned thrust faults.

A second tectonic event is recorded by the development of greenschist-facies mylonites and by their subsequent brecciation. During the Eocene, extensional faulting reactivated the late Cretaceous mylonite zone creating a lower-grade mylonite which was brecciated at higher crustal levels. Microscopic textures within the greenschist-facies mylonites and brittlely developed macroscopic structures record top-to-the-east rotation. Evidence recorded in Table 2b. lends support to an Eocene tectonic event.

In summary, movement along the Bitterroot dome mylonite zone during the late Cretaceous is only in part responsible for the opening of the Bitterroot Valley (Fig. 23a). Movement along the same zone further opened the Bitterroot Valley during the Eocene (Fig. 23b).
Table 2a. Data supporting late Cretaceous movement along the Bitterroot front.

1. Better developed mylonitic textures in pegmatites of the Bitterroot dome than in surrounding granitic rocks. This indicates that the pegmatites had high water contents and developed late in the crystallization of the Idaho batholith (La Tour and Barnett, 1986).

2. The occurrence of pegmatites as streaks and layers parallel to granitic foliation suggests that a fabric had already developed in the Bitterroot lobe rocks by the time pegmatites began to crystallize (La Tour and Barnett, 1986).

3. The amphibolite facies mylonites formed under the influence of magmatic or high-grade metamorphic waters (Kerrich and Hyndman, 1986). Geothermometry studies by Kerrich and Hyndman (1986) indicate that the amphibolite-facies mylonites formed under ductile conditions at temperatures near 500 degrees C. This approaches the expected temperature of granitic rocks during late-stage crystallization.

4. Age dating reviewed by Hyndman (1980, 1983b) places main stage batholith intrusion for the Bitterroot lobe of the Idaho batholith between 85-66 m.y. ago.

5. Thrusts east of the Sapphire tectonic block developed during the late Cretaceous (Hyndman, 1980).

6. The intrusion of shallow plutons into deep levels within the Idaho batholith during the Eocene demands that the batholith was at least partially unroofed before the intrusion of epizonal plutons (Hyndman, 1980).
Table 2b.

1. Brecciated Tertiary volcanic rocks lying on the mylonite zone were deformed in a manner similar to the chloritic breccias (Hyndman, 1980) and thus probably under similar conditions.

2. Oxygen isotopic studies by Kerrich and Hyndman (1986) have shown that the chloritic breccias formed under the influence of meteoric waters and at temperatures ranging from 270 to 350 degrees C. This range of temperature falls within that of greenschist-facies metamorphic conditions.

3. Age dating by Garmezy and Sutter (1983) and by Chase and others (1983) record a major Eocene thermal event superimposed on older rocks.

4. The presence of a narrow greenschist-grade mylonite zone overprinted by shallow level brecciation documents rocks deforming at progressively shallower levels in the crust.

5. The Bitterroot Valley may be another example of the numerous intermontane valleys of southwest Montana and Idaho which formed during a period of Eocene extension (Hyndman and others, 1986).

6. Rehn and Lund (1981), in a study of the Eocene Painted Rocks pluton in southwestern Montana and eastern Idaho, suggest that both geochemical and structural data indicate emplacement in an extending environment. Intrusion was controlled by normal faulting and chemistry indicates subalkalic affinity.

Table 2b. Data supporting Eocene movement along the Bitterroot front.
Fig. 22. Schematic diagram of the geometry of low-angle normal faults (adapted from Wernicke, 1985).
Fig. 23. Schematic diagrams of structural relationships of the study area (a) Post-late Cretaceous (b) Post-Eocene.
DISCUSSION AND CONCLUSIONS

The late Cretaceous amphibolite-facies mylonites of the Bitterroot dome have been crosscut and overprinted by greenschist-facies mylonites that formed during a period of shear which occurred at lower P,T conditions than the amphibolite-facies mylonites. The younger mylonites have taken advantage of a zone previously weakened by crustal shear. Studies and reason indicate that once a shear zone is activated, that zone is likely to be reactivated during later movement. Greenschist-facies mylonitization occupied a narrow zone within the total available shear zone because deformation was under lower pressures and temperatures. Orientations of S-C surfaces, mica "fish", and offset twin planes within the greenschist-facies mylonites, indicate that shearing involved top-to-the-east rotation and eastward extension.

As the greenschist-facies mylonite reached shallow depths the chloritic zone was deformed at near-surface conditions. Eastward extension is recorded by top-to-the-east, down-dip rotated folds, steeply eastward-dipping faults, and east-west extensional tensile fractures and pull-apart veins. Slickensides oriented perpendicular to north-south fold hinges formed during top-to-the-east rotation.

Shearing and extension initially took place during the late Cretaceous, deforming the amphibolite-facies
mylonites and possibly being related to the thrusts located east of the Sapphire block. Later shearing at higher levels in the crust deformed the greenschist-facies mylonites. Brecciation is younger than either mylonite zone and may be related to Tertiary extension that was responsible for renewed opening of the Bitterroot Valley.
## Appendix A. Modal Analysis of Chloritic Breccia Rocks (Volume % visual estimate)

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Appendix B. Sample location map
REFERENCES CITED


Snoke, A.W. 1980, The transition from infrastructure to

Spencer, J.E., 1984, The role of tectonic denudation in the warping and uplift of low-angle normal faults. Geology, 12, p. 95-98.


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