Petrology and Origin of Precambrian Metamorphic Rocks in the Eastern Ruby Mountains Southwestern Montana

Kevin Smith
The University of Montana

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PETROLOGY AND ORIGIN OF PRECAMBRIAN METAMORPHIC ROCKS
IN THE EASTERN RUBY MOUNTAINS,
SOUTHWESTERN MONTANA

by

Kevin Smith

B.S., Western Washington State College, 1977

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Master of Science

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Director: David Alt

Precambrian schists and gneisses are exposed in much of the Ruby Mountains of southwestern Montana. The major rock types include abundant quartz-feldspar gneiss, hornblende gneiss, amphibolite, quartzitic gneiss, and marble. Minor rock types important in the study area include anthophyllite gneiss, hornblende-pyroxene-plagioclase gneiss, quartz-hypersthene-garnet granulite, iron-formation, and hornblende-hypersthene granulite. These rocks form a concordant series folded into a broad, open antiform-synform pair in the study area.

The marble, quartzitic gneiss, and iron-formation in the area clearly originated as sedimentary units. The hornblende gneiss assemblage was also found to be dominantly sedimentary in origin. Evidence for this includes its association with marble and quartzitic gneiss, the latter intimately mixed within it, the presence of meta-conglomerates, and the abundance of quartz making the composition distinctly different from normal igneous rocks. A petrographic estimate of composition for one amphibolite showed it to most resemble a dolomitic shale. It is possible that some amphibolites in the area are metamorphosed dikes and sills. The uniform granitic composition and lack of structures in the quartz-feldspar gneiss suggest an igneous origin for this unit.

Metamorphism in the area exceeded the sillimanite-orthoclase zone of the amphibolite facies as indicated by the presence of sillimanite, perthite, and calcic plagioclase An42. The presence of the assemblage quartz-hypersthene-garnet, and textures indicating the reaction anthophyllite = enstatite + quartz show that conditions ran locally into the orthopyroxene zone of the granulite facies. These probably represent areas locally undersaturated in water.

The area went through two phases of regional deformation. The first event, roughly contemporaneous with the highest metamorphic grades, produced similar-style, tight isoclinal folds and the regional foliation. Subsequent to that, the area was folded into a large, concentric, open antiform-synform pair trending northeasterly and plunging at moderate to steep angles. Small super-imposed folds of undetermined age relationship to the second event also exist.
ACKNOWLEDGMENTS

I greatly appreciate the assistance and encouragement given by Doctors David Alt, Don Hyndman, Gray Thompson and Tom Margrave. I would like to thank the various ranchers upon whose property I conducted my field work, especially for the generous hospitality given me by Mr. and Mrs. Jack Kephart, and Mr. and Mrs. Coy Brown. Lastly, thanks to the many graduate students for the hours of discussion over many aspects of this paper.


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Pre-beltian rocks underlie much of the Ruby Range of southwestern Montana. They include hornblende gneisses, amphibolites, mica schists, quartzites, marbles and quartz-feldspar gneisses. Marbles and quartzites clearly represent metasediments. Bielak (1978) has shown that hornblende gneisses and amphibolites in the region are separable into distinct metasedimentary and metaigneous units respectively. The Dillon Gneiss unit has so far resisted the clear designation as metaigneous or metasedimentary. Although it was originally mapped as a "granite gneiss" by Heinrich (1953), recent investigations have called that interpretation into question. Garihan (1973 and 1974), and Garihan and Williams (1976) have postulated a mudstone or shale rich in illite and quartz as a possible sedimentary parent. In each case though, lack of hard evidence prevents any final conclusion. The resolution of this problem is important before a resonable pre-metamorphic history can be postulated.

The purpose of this study was to map a portion of the contact between the Dillon Gneiss and overlying metasedimentary units. By determining through field relationships and petrographic analysis the original nature of the Dillon Gneiss, I could go on to suggest a
premetamorphic depositional environment that resulted in the present rock assemblage. Another purpose of this study was to determine the metamorphic and deformational history of the area. The study area was chosen for its lack of retrograde effects and interesting structure. Well developed retrograde effects would tend to mask prograde mineralogy and equilibrium textures essential to determining the highest temperature and pressure conditions reached.

Location

The Ruby Range lies east of Dillon along the Beaverhead-Madison County line in southwestern Montana (Fig. 1). The Hinch Creek map area covers about 18 square kilometers on the eastern edge of the range. The Ruby River and the range front bound the area on the east. Vertical faults bringing the basement against Paleozoic sedimentary rocks bound the area on the west. The area includes portions of the Alder, Laurin Canyon, Metzel Ranch, and Ruby Dam 7 1/2-minute quadrangles.

Physiography

The area has been uplifted to elevations approaching 2135 meters by differential movement on range-bounding vertical faults. Activity on the faults dates from the Eocene or Oligocene until the present time (Garihan, 1973). Gently rolling upland surfaces cut by deep drainages characterize the area. Sage and grass cover most of the surface but groups of juniper and pine are scattered about, especially
Figure 1. Schematic index map of the Ruby Mountains showing the Pre-Cherry Creek (PCC), Dillon Gneiss (DG), Cherry Creek (CC) units and overlaying paleozoic segments (PS) with the Northwest-trending faults. The heavy outline indicates the study area.
on north facing slopes. Greasewood chokes the upper portions of many gullies. Major drainages include Hinch Creek, Dry Hollow, and Beatch Canyon. They all flow northeastward normal to the range front. Hinch Creek has an elevation of 1615 meters at its exit from the range giving a total relief in the area of 520 meters.

Previous Studies

Earliest descriptions of the region were made by Hayden (1872) who noted excellent exposures of banded gneiss, hornblende gneiss, and veins of feldspar and quartz. Peale (1896) described the rocks along Cherry Creek in the Gravelly Mountains. Equivalents of the Cherry Creek Group form much of the basement along the northwest flank of the Ruby Range.

Winchell studied the Crystal Graphite deposits near Dillon in 1911. In 1914 he published a general survey of the important mining districts of the Dillon one degree quadrangle. Klepper (1951) described the southern Ruby Range in a general reconnaissance of southwestern Montana. The first detailed studies of the Ruby Mountains were made by Heinrich (1948, 1949a, 1949b, 1950, 1960, and 1963).

Recent studies in the range include the following: Tysdal (1970) mapped the Paleozoic section in the northern Ruby Range; Okuma (1971) studied the petrology and structural geology in relation to talc deposits in the southern Ruby Range; Garihan (1973, 1974, 1976a, and 1976b) studied the structure, petrology, and talc deposits of the central Ruby Range; Dahl (1978) determined metamorphic conditions using
electron microprobe geothermometry and geobarometry; Desmarais (1978) studied the origin of the ultramafic bodies; Bielak (1978) examined the origin of amphibolites. Wooden (1973), Giletti (1966), and James and Hedge (1980) have obtained radiometric dates of rocks from the Ruby Range.

Included in the area of this study is some mapping done by James and Wier (1960) of the Kelly iron deposit, and Berg (1976) of the portion between Hinch Creek and the Ruby Reservoir. All previous mapping in the present area was reviewed in the field for this study.
General Statement

The Ruby Range is a block of basement rock uplifted since Eocene or Oligocene time. The range core contains crystalline schists and gneisses unconformably overlain at the northern end by Paleozoic sediments (Fig. 1). The crystalline complex consists of broad belts of different rock types that parallel the northeast trend of the range. From northwest to southeast they are grouped into three units: 1) the "Cherry Creek" Group, 2) the "Dillon Granite Gneiss", and 3) the "Pre-Cherry Creek" rocks (Heinrich, 1960). The regional foliation strikes northeast parallel to the units and dips northwest placing the Cherry Creek Group highest structurally. The uppermost part of the Cherry Creek is not exposed in the Ruby Range (Garihan, 1973).

Cherry Creek. The Cherry Creek group occupies the northwest side of the range and contains several different rock types. Peale first described it near Ennis, Montana in 1896. It crops out regionally in the Gravelly, Madison, Tobacco Root, Greenhorn, and Ruby Ranges. The presence of marble distinguishes it from other similar units. The various rock types of the Cherry Creek Group include:
1) marble
2) calc-silicate schist
3) quartzite
4) hornblende gneiss and amphibolite
5) quartzofeldspathic gneiss
6) sillimanite schist
7) biotite schist and gneiss
8) chlorite schist
9) iron formation

**Dillon Gneiss.** The Dillon Gneiss commonly contains elongate stringers of quartz, perthitic microcline, and plagioclase imparting a conspicuous banding to the rock. It forms a sheet-like mass of quartz-feldspar and biotite-quartz-feldspar gneiss separating the Cherry Creek from the Pre-Cherry Creek units. Heinrich (1960) originally named it the "Dillon Granite Gneiss". Garihan and Williams (1976) have proposed renaming it the "Dillon Gneiss" because the igneous nature of the unit is not firmly established.

**Pre-Cherry Creek.** The Pre-Cherry Creek rocks crop out along the southeast side of the range. Heinrich (1960) used the name to distinguish these from the marble bearing Cherry Creek unit to the northwest. The various rocks are mostly gneissic, coarse grained, migmatitic, and discontinuous along strike. The main rock types include:
1) quartz-feldspar gneiss
2) biotite-garnet-quartz-feldspar gneiss
3) biotite-garnet gneiss
4) hornblende-quartz-feldspar gneiss
5) hornblende gneiss and amphibolite

Field Description and Petrography

The field work for this study took place during parts of the summers of 1978, and 1979. Petrographic study of thin-sections was made from representative samples. Rock slabs and chips were etched in hydrofluoric acid and stained with sodium cobaltanitrite as an aid in distinguishing and estimating feldspar percentages. Marbles were stained with Alizarine Red S to distinguish calcite from dolomite (Friedman, 1959). Anorthite content was determined with the bisectrix method in combination with extinction angles on cleavage fragments immersed in oils.

Marble. Marbles in the Hinch Creek area form resistant marker beds. They form rounded outcrops easily recognized by their characteristic tan- to grey-weathered surface commonly covered by orange lichen.

Mappable units range from 4 to over 500 meters thick. The width of beds varies along strike. Unrecognized folding or plastic flowage causes much of the thickness variations and makes it impossible to measure true thickness. In the southeast corner of section 32, beds of marble as thin as 0.5 meters interbed with hornblende gneiss and quartz-feldspar gneiss.
The layering in much of the marble ranges from 5 centimeters to over a meter in thickness. Differential weathering along the parting surfaces defines most of the layering, especially in pure marbles. In quartz bearing varieties, beds and stringers of quartz up to 20 meters thick also define the layering. Bedding everywhere parallels the foliation. Compositional differences, the presence of concordant quartzite beds, and interlayering with quartz-feldspar gneiss and hornblende gneiss on a small scale indicate that the compositional layering represents original sedimentary bedding. In some massive marbles the quartz lenses and beds 2-20 millimeters thick are isoclinally folded and broken (Fig. 2). The marble flowed plastically during deformation while the more competent quartz beds folded and broke. Bedding in the marble generally persists along strike, but on the noses of some folds it stretches and pinches out into lenses best observed in the Hinch Creek valley about 2 kilometers from the range front.

Coarse grained dolomitic marble predominates in the area with grain sizes ranging from .1 to 5 millimeters. Compositions include pure dolomite marble and quartz-bearing calcite marble containing up to 15% quartz. Quartz grains show strain shadows and some subgrain development. Some contain rounded inclusions of calcite and/or microcline. Accessory minerals include muscovite, microcline, and hematite. Calc-silicate marbles contain calcite, dolomite, diopside, serpentine, phlogopite, graphite, and garnet (Fig. 3).
Figure 2. Folded and broken lenses of quartz within fine-grained structurless marble. The head of the hammer is approximately 10 cm long.
Figure 3. Calc-silicate marble showing diopside, relict grossular garnet altering to chlorite, phlogopite mica, graphite and dolomite.

0.6 mm

Figure 4. Randomly oriented grains of tremolite.

1.0 mm

Figure 6. Large ameboidal grains of quartz embaying smaller grains of microcline.

1.0 mm

Figure 7. Ameboidal quartz grains embaying microcline and hornblende.

0.6 mm
The unit apparently originated as beds of limestone with local thin beds of quartz. In several locations undeformed layers of quartz were found within it but for the most part deformation and recrystal-
ization destroyed the primary sedimentary structures leaving the grey massive marble as it now exists.

Tremolite rock. The rock is a mass of ragged tremolite crystals 3 to 5mm long (Fig. 4). Crystals up to 15mm exist in places. Tremolite has difficulty in nucleating and commonly forms as coarse-grained, unoriented crystals. Calc-silicates normally lack preferred orien-
tations and may differentiate into monomineralic masses at high grades of regional metamorphism (Spry, 1969).

In outcrop the rock forms a low, massive, and very resistant bed or lens. It is discontinuous along strike and weathers tan to brown, brown on a fresh surface. In is most likely genetically related to the adjacent marble unit.

Quartz-feldspar gneiss. Quartz-feldspar gneiss occurs most com-
monly as a conspicuously foliated or banded rock with a distinctive orange-pink color on the weathered surface. Bands of elongate quartz and feldspar define the foliation. Where present, biotite is dis-
seminated or, more commonly, concentrated in bands a few millimeters to many centimeters thick. Biotite parallels the foliation and helps define it. Biotite-quartz-feldspar gneiss is less common and on the average more calcic than "normal" quartz-feldspar gneiss.
The gneiss in the center of the antiform (see map), contains a few beds of hornblende gneiss concordant to the foliation. In general, it has uniform composition and color except near the hornblende gneiss contact. The two units become interlayered at the contacts. The southeastern body of gneiss varies in color from gray to orange-pink and contains abundant biotite in some exposures and none in others. It contains numerous, scattered, and concordant beds of hornblende gneiss and amphibolite from 0.5 to 10 meters thick which cannot be followed along strike due to poor exposure.

Severely deformed migmatites crop out locally and augen gneiss is fairly common throughout. Augen range from 0.5 to 5 centimeters. Porphyroblasts of garnet from 0.2 to 2 centimeters form a spotted appearance in many outcrops. Garnets are disseminated or form concentrations in either the quartzofeldspathic or more commonly in the biotite-rich layers.

Structurally these gneisses form the lowermost unit in the area. They occupy the center of the antiform and the edge of the synform. The foliation everywhere parallels the contacts with marble or hornblende gneiss. The gneiss makes sharp contacts with the marble, but it grades into, and interbeds with the hornblende gneiss. Gradation consists of increasing amounts of mafic minerals, especially hornblende and biotite with conspicuous light and dark layers. Hornblende, where present, generally appears in quantities of 10% or more and is used as the basis for distinguishing hornblende gneiss from quartz-feldspar gneiss.
Compositions of the unit concentrate near the center of a quartz-plagioclase-alkali feldspar variation diagram (Fig. 5). They show some variation in composition across the diagram, ranging from quartz-plagioclase to quartz-plagioclase-microcline gneisses. Assuming isochemical metamorphism, this indicates a fairly homogenous unit with the average bulk composition of granite.

Quartz ranges from one fourth to one half of the rock (Table 1). Except for augen, it usually forms the largest grains but sizes are seriate from 5 millimeters down to very fine. Crystals normally are elongate and strained. They have irregular and ameoboid shapes with lobes growing into and enveloping adjacent minerals. In some cases this leaves relict inclusions within the quartz (Figs. 6 & 7).

Potassium feldspar normally forms about one third of the rock with microcline the dominant variety. Shapes vary from minor elongate to equigranular grains, some with 120° triple junctions. Some specimens show strong development of ribbon and patchy microperthite. The albite in many cases has exolved to the edge of the grain (Fig. 8). Microcline augen grow to as large as 3 centimeters. Plagioclase, An27-33, ranges from minor amounts up to two thirds of the rock. Minor to almost complete sericite alteration exists in the mineral, but is normally less than 10%. Many plagioclase grains have a zoned edge adjacent to microcline which the sericite nowhere invades. Twinning is present but not abundant.
Figure 5. Modal compositions of quartz-feldspar gneiss plotted on an I.U.G.S. quartz, alkali feldspar, plagioclase variation diagram (from Streckeisen, 1973).
Table 1. Modal analyses of quartz-feldspar gneiss
Volume % visually estimated from thin section

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Modal analyses of biotite-quartz-feldspar gneiss

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Figure 8. Quartz-feldspar gneiss. Large perthite grains with albite exolved to the edges.

0.6 mm

Figure 9. Texture in garnet-biotite-quartz feldspar gneiss showing suboriented grains of biotite and broken grains of garnet.

1.0 mm

Figure 10. High magnification view of sillimanite end-sections forming a lineation in the sample.

0.2 mm

Figure 13. Texture in amphibolite section 06 showing hornblende plus plagioclase.

0.6 mm
Biotite grains characteristically define the foliation, but in some cases show no preferred orientation. Pink almandine garnet ranges up to one fifth of the rock occurring as anhedral to euhedral, fractured poikioblasts (Fig. 9). It forms compositional layers usually associating with biotite layers.

Accessory sillimanite occurs in one thin-section (27g), as euhedral end-sections associated with the garnet-biotite layer (Fig. 10). In handspecimen sillimanite occurs as parallel blades up to 5 millimeters long associated with biotite and radiating blades growing on the foliation surface which apparently shows that they grew during or after formation of the foliation. Other accessory minerals include rounded zircon, apatite, hematite, chlorite altering from biotite, magnetite, and rutile.

The origin of the quartz-feldspar gneiss presents a special problem. Throughout most of the Ruby Range, no firm evidence has been found to indicate an igneous or sedimentary parent. The contacts between the gneiss and adjacent rocks, especially marble, lack cross-cutting relationships. Garihan and Williams (1976) suggested a sedimentary parent of a mudstone or a shale rich in illite and quartz to explain the granitic composition.

The quartz-feldspar gneiss/marble contact in the study area also shows no cross-cutting relationships. The two become interbedded in the southcentral part of the area. This could result from either interbedding of original sediments or rhyolite volcanics, intrusion of granite
sills into the marble, or isoclinal folding. Synkinimatic intrusion in the catazone takes place with minor disturbance of the wall rock (Buddington, 1959).

The granitic composition of the unit argues in favor of a plutonic origin. The Dillon Gneiss is an extensive unit in the Ruby Range. It would take a very thick pile of sediments to form the gneiss. It is structureless and compositionally uniform. No compositional layering was found, no lenses or beds of conglomerate could be located. Meta-conglomerates were found in some hornblende gneiss units showing their ability to survive the metamorphism (Fig. 11). No primary sedimentary structures have been found in the gneiss even after many regional and detailed studies in the Ruby Range. Because of this, I feel that a plutonic origin is the most permissible conclusion.

Hornblende-gneiss assemblage. Structurally, the hornblende-gneiss unit sits just above the marble and grades "upsection" into quartzitic gneiss and biotite-quartz-feldspar gneiss. In the northwest corner of the map area the unit becomes very thick and contains the Kelly Iron Formation mapped in detail by James and Wier (1960). Quartzitic gneiss beds occur both in the hornblende gneiss and marble units.

The assemblage forms abrupt contacts with the marble but grade concordantly into the quartz-feldspar gneiss. Near the "contact" the volume of microcline in the rock is much greater than in the rest of the assemblage. The presence of hornblende defines the unit although there exist several minor related rock types within it. Because of their
Figure 11. Metaconglomerate within hornblende gneiss. Exposure is too poor to determine the original attitude and position of sedimentary layer. Pebbles are quartz.
variety and distinctiveness, they are described separately in this section.

**Hornblende gneiss and amphibolite.** This dominant rock type of the assemblage normally crops out as a medium grained banded gneiss. Massive "salt and pepper" amphibolite beds are made up almost entirely of hornblende and plagioclase. These crop out within the lower part of the assemblage and locally within the quartz-feldspar gneiss unit. Grain sizes range from .1 to 5 millimeters with 1 or 2 millimeters most common. Layers consist of dark hornblende-rich and light quartz-feldspathic lithologies from a few millimeters to over a meter thick. Garnets in some cases concentrate in mafic layers. The unit forms poor outcrops that are best exposed in gullies or ridges. The layers are compositionally distinct parallel to the foliation. In places they are isoclinally folded with an amplitude ranging from 2 to 50 meters (Fig. 12). This is best observed on the north side of Hinch Creek in section 32. Foliation passes through the noses of the folds. Folding within the unit makes the original thicknesses of beds impossible to determine. Metaconglomerates are noted in several locations within hornblende gneiss units (Fig. 11).

Hornblende content ranges from minor amounts to nearly one half of the gneiss and 60% in amphibolite (Table 2 and Fig. 13). Anhedral to euhedral crystals define the foliation and in some cases a lineation. The crystals commonly contain inclusions of biotite or rounded quartz. Crystals show brown to dark green pleochroism except in amphibolite
Figure 12. Isoclinal folding in interlayered hornblende and quartz-feldspar gneiss. The folding is not apparent on the vertical portion of the outcrop (the top of the figure). The brim of the hat is approximately 35cm wide.
Table 2. Modal analyses of hornblende gneiss and amphibolite
Volume % visually estimated in thin-section

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*Thin-section count of 534 points
sample.34 where it is blue-green to green near garnet. This suggests decreased availability of ferrous iron near the iron-rich garnet. In hornblende this lightens the pleochroism from dark olive-green and brown to light shades of blue-green and green (Troger, 1979). Hornblende also appears to be altering to diopside in some rocks as evidenced by diopside crystals with hornblende cores.

Plagioclase is more calcic in the gneiss at An$_{42}$. The anhedral to subhedral grains show minor sericitization, but sericite may replace up to one third of the plagioclase. Near microcline, the plagioclase rims are zoned and the sericite does not invade these rims. In one amphibolite, recrystallized plagioclase grains have common 120° grain junctions.

Quartz is an important, although variable constituent in the gneiss, but minor in the amphibolite. Grains have strain shadows, with elongate to equigranular shapes. The grains become larger on the average and more ameboidal with increasing quartzofeldspathic content of the rock. Garnet is minor but generally present in the rock. Porphyroblasts often contain inclusions of quartz, plagioclase, and in some cases hornblende. Garnets are commonly fractured and broken. The fragments spread out along the foliation and long dimensions lie parallel to it. Foliation wraps around the porphyroblasts in some samples. As in the quartz-feldspar gneiss, garnets sometimes show disequilibrium textures with biotite (Fig. 14). Symplektite on garnet was noted in sample 19a (Fig. 15). In some samples equilibrium textures exist between the two.
Figure 14. Disequilibrium textures between garnet and biotite. They are separated by a plagioclase rim.

0.6mm

Figure 15. Symplektite on garnet in the presence of biotite. Garnet plus hornblende show equilibrium textures elsewhere in the section.

1.0mm

Figure 16. Randomly oriented biotite grains in hornblende gneiss. Note the perthite grain which is present in samples taken near the contacts with quartz-feldspar gneiss.

1.0mm

Figure 17. Skeletal enstatite growing on anthophyllite from the anthophyllite gneiss sample 3xa.

2.0mm
Biotite ranges up to 10% but rarely amounts to more than one percent. It generally grows parallel to foliation, but may show no preferred orientation (Fig. 16). In some rocks it shows disequilibrium textures such as those with garnet described above. It may also have a ragged, splintery appearance and wormy intergrowths of quartz or plagioclase.

Perthitic microcline is normally absent in the rock, but may become important near contacts with quartz-feldspar gneiss. Diopside is important in some samples as described above under hornblende. Other accessory minerals include apatite, rounded zircon, minor rutile, magnetite, hematite, and in one sample, secondary epidote.

The hornblende gneiss beds vary greatly in mineralogy and texture, and generally contain abundant quartz (Table 2). The beds are compositionally layered and grade into other units. Because of the textural and mineralogical variations and concordant beds grading into other units, I conclude that they are derived from a calcareous shale. A shaley limestone derivative would contain less quartz. The large amount of quartz eliminates the possibility of metamorphosed basalts because the composition departs too radically from a natural basalt.

Amphibolites in the study area are more basic in composition than the hornblende gneiss, but they are not necessarily metamorphosed basic volcanics. A petrographic estimate of the composition of sample 06 shows it to resemble most closely a dolomitic shale (Table 3). Bielak (1978) found that amphibolites in the Winnipeg Creek area could be separated
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1. Oxide approximation of modal amphibolite (this study)
2. Average shale (Clarke, 1924, page 552)
3. Average shale from 2 recalculated with one part in six pure dolomite, results after removal of volatile constituents and renormalization to 100%
4. Average andesite (Hyndman, 1972, page 166)
5. Average continental tholeiite (Hyndman, 1972, page 171)
into both meta-sedimentary and meta-basalt units. This is most likely the case in the Hinch Creek area as well. Many amphibolite units have massive, uniform textures whereas the hornblende gneiss is generally streaky, quartz bearing, and in places conglomeratic. The entire assemblage consisted of minor basic volcanics interbedded with calcareous shales. Volcanics could have formed at the same time as deposition or intruded later as sills. The assemblage was then subjected to upper amphibolite-facies metamorphism.

**Garnet-enstatite-quartz-anthophyllite gneiss.** This bed crops out as a dark purplish-brown rock with foliation defined by oriented anthophyllite grains. Hornblende gneiss concordantly surrounds the bed. It is very hard and resistant to weathering and therefore caps the ridge where it crops out. This rock type was found only once in the study area.

Parallel anthophyllite grains dominate the mineralogy. Growing in the anthophyllite are skeletal enstatite crystals with quartz inclusions (Fig. 17). This texture suggests the reaction:

\[
\text{Anthophyllite} \rightarrow 7 \text{Enstatite} + \text{Quartz} + H_2O
\]

(Winkler, 1974)

Anthophyllite reacts at nearly 800°C over most pressures under conditions of \( P_{H_2O} \) equals \( P_{\text{load}} \). Because the prograde reaction yields water, it would be favored where \( P_{H_2O} \) is less than \( P_{\text{load}} \) and would react at lower temperatures.
Table 4. Modal analyses of minor lithologies
Volume % visually estimated in thin section

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<tr>
<th>Sample #*</th>
<th>21b</th>
<th>3xa</th>
<th>15b</th>
<th>08</th>
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</table>

*21b. tremolite rock  
3xa. garnet-enstatite-quartz-anthophyllite gneiss  
15b & 08. hornblende-pyroxene-plagioclase gneiss  
30f & 38. quartz-hypersthene-garnet granulite  
37a. graphite gneiss  
42 & 10. hornblende-hypersthene granulite  
05b & 33. quartzitic gneiss  
28b & 36. pegmatite
Sieve textured garnets are an important part of the rock. Inclusions in them weakly to strongly parallel the foliation with no evidence of rotation. The porphyroblasts are irregular in outline and elongate parallel to the foliation. Cordierite is absent. Rabbit (1948, p. 314) lists an anthophyllite gneiss from the "Ruby Dam area" which also has no cordierite. I believe it is from the same outcrop as I know of no other in the vicinity. Accessory minerals include biotite and rutile.

**Hornblende-pyroxene-plagioclase granulite.** This rock type crops out in two places about one kilometer apart. Sample 15b was collected from a bed about a meter thick within the quartz-feldspar gneiss. Plagioclase-rich layers about 1 millimeter thick define a crude foliation concordant with that of the enclosing gneiss. Sample 08 was collected near the contact between the quartz-feldspar and hornblende gneisses with which it is also concordant. Both samples have average grain sizes in the range of 0.1 and 2 millimeters.

Essentially unaltered plagioclase dominates the rock type. It occurs both as a mosaic of recrystallized grains commonly intersecting at 120° angles and as a wormy intergrowth in garnet (Fig. 18). Quartz is minor to absent in the rock, small amounts occurring as mosaic grains intersecting each other and plagioclase grains at 120° angles. Diopside forms up to a quarter of the rock and also occurs as a mosaic of recrystallized grains intersecting at 120° angles.
Figure 18. Wormy intergrowths of plagioclase into garnet in hornblende-pyroxene-plagioclase granulite.

1.5mm

Figure 19. Relict grains of hornblende within a large grain of hypersthene. The gradational contact suggests replacement of hornblende by hypersthene.

1.0mm

Figure 20. Quartz-hypersthene-garnet granulite. Section showing the granulitic texture.

1.0mm

Figure 21. Quartzitic gneiss. Wide-angle view in reflected light showing elongate feldspar grains within quartz. Feldspar grains consist of plagioclase (light stipple) with minor orthoclase along the edges (coarse stipple).

5.0mm
The other minerals in this rock show common disequilibrium textures. Hornblende and hypersthene make up small to moderate amounts of the rock. In one grain of sample 15b a hypersthene grain has a diopside core. This could result from the replacement of diopside by hypersthene or by the simultaneous growth of both. The lack of other disequilibrium textures between the two suggests the second case. Hornblende has gradational contacts with hypersthene, and hypersthene grains with hornblende cores exist (Fig. 19). These gradational contacts suggest disequilibrium. They are not in contact in sample 08 because of the small amounts of each. Biotite and garnet are minor constituents in this rock type and display mutual disequilibrium textures. Biotite grains embay the garnets and a plagioclase rim always separates the two. Accessory minerals include magnetite and apaties.

The disequilibrium textures indicate the following transformations:

\[
\begin{align*}
\text{Hb} & \quad \text{Hy} & \quad (\text{relict Hb within Hy}) \\
\text{Pl} & \quad \text{Ga} & \quad (\text{wormy Pl surrounded by Ga}) \\
\text{Bi} & + \quad \text{Ga} & \quad \text{Pl} & \quad (\text{Pl separates Bi and Ga}) \\
\text{Pl} & \quad \text{is recrystallized} & \quad (120^\circ \text{ junctions}) \\
\text{Di} & \quad \text{is recrystallized} & \quad " \\
\text{Qz} & \quad \text{is recrystallized} & \quad " \\
\end{align*}
\]

(Note: Ab = Albite, Alm = Almandine, An = Anorthite, Bi = Biotite, Cp = Clinopyroxene, Di = Diopside, Ga = Garnet, Hb = Hornblende, Hy = Hypersthene, Op = Orthopyroxene, Or = Orthoclase, Pl = Plagioclase, and Qz = Quartz.)
De Waard (1965a and b, 1967), lists the following reactions for the entrance into the granulite facies in metabasites:

\[
\begin{align*}
2 \text{Bi} + 12 \text{Qz} & \rightarrow 8 \text{Op} + \text{Alm} + 4 \text{Or} + 4 \text{H}_2\text{O} \\
700^\circ \text{C} \text{ minimum at } 10\text{kb}, P_{H_2O} & \text{ P load}
\end{align*}
\]

\[
\begin{align*}
\text{Hb} + 4 \text{Qz} & \rightarrow 3 \text{Op} + \text{Cp} + \text{Ab} + \text{H}_2\text{O} \\
700^\circ \text{C} \text{ minimum at } 10\text{kb}, P_{H_2O} & \text{ P load}
\end{align*}
\]

\[
\begin{align*}
\text{Op} + \text{An} & \leftrightarrow \text{Cp} + \text{Alm} + \text{Qz} \\
760^\circ \text{C} \text{ minimum at } 10\text{kb}, P_{H_2O} & \text{ P load}
\end{align*}
\]

The second two may be combined defining the entrance to the pyroxene-granulite subfacies from the hornblende subfacies (Buddington, 1966; De Waard, 1967).

\[
\begin{align*}
\text{Hb} + \text{An} + \text{Op} & \leftrightarrow \text{Alm} + \text{Cp} + \text{Ab} + \text{H}_2\text{O}
\end{align*}
\]

All these minerals make up the composition of sample 15b. Sample 08 contains biotite but no orthoclase.

The rock has a basic composition similar to a basalt or andesite. The reactions listed occur in a water-undersaturated environment. This is more probable in a basalt rather than a "basaltic" sediment. They occur as beds parallel to the compositional layering possibly indicating basalt flows or sills prior to metamorphism. Both beds strike at large angles away from the northwesterly trend of diabase dikes found in other parts of the Ruby Range. I conclude that they were basalt flows or sills within the original sedimentary package.

**Quartz-hypersthene-garnet granulite.** This rock type crops out in two locations. The individual textures and weathering characteristics differ in some respects such as grain size and banding, but have most
points in common petrographically. The rock of sample 38 weathers red-brown from the abundant hematite and it forms low, massive outcrops. Garnet porphyroblasts 3 to 4 millimeters stand out in hand-specimen. Poor exposure makes it impossible to map the actual dimensions but hematite-stained soil covers a broad area around the outcrops.

Sample 30f represents a bed about 1 meter thick that contains pyroxene-poor layers 1 to 10 millimeters thick and magnetite layers 1 to 2 millimeters thick which parallel both the major bed and the foliation and appear to be remnant sedimentary layers. This rock type locally contains banded iron-formation. The bed behaved competently during deformation breaking into tabular blocks which retained their shape as the adjacent layers wrapped around their ends. The bed lies in a sequence of quartzitic gneiss, hornblende gneiss, and garnet-biotite-quartz-feldspar gneiss beds.

Garnet is the most abundant mineral in this rock type. In section 38 garnet porphyroblasts grow up to 4 millimeters across with quartz inclusions and in one grain a zircon inclusion. Magnetite and hematite rim many garnets. This possibly results from excess ferric iron not used by the growing garnet. Hematite forms by secondary oxidation of the magnetite. The rock is rich in iron, but the garnets now make up 42% of it. The iron was concentrated in the space left over. In section 30f, garnets are poikioblastic and some are sieve textured. They contain quartz inclusions except in the pyroxene-poor layers. They do not form porphyroblasts in this section.
Hypersthene and quartz commonly occur as a mosaic of recrystallized grains intersecting at 120° angles. In sample 38 the quartz shows strain shadows and the hypersthene is twinned indicating post-crystallization strain in the rock. Sample 38 is porphyroblastic whereas sample 30f is granulitic with a mosaic of grains intersecting at 120° angles (Fig. 20). No disequilibrium textures are apparent in either sample.

The abundant quartz and iron oxides in both samples, plus the bedded nature and associated quartzitic gneiss, hornblende gneiss, and garnet-biotite-quartz-feldspar gneiss beds around sample 30f indicate a sedimentary parent for this lithology. Magnetite layers 1 to 2 millimeters thick represent original banded iron layers. The iron was disseminated within the sample 38 parent sediment. This rock has a higher silica and lower magnesium content than in sample 30f. The mineral paragenesis exists entirely within the orthopyroxene zone of the garnulite facies. This reflects the water content of the lithology as well as the temperature and pressure conditions because adjacent rock types have almandine amphibolite facies mineralogies.

Hornblende-hypersthene granulite. There are two varieties of hornblende-hypersthene granulite which differ both in their appearance in outcrop and in thin-section. Sample 10 represents a minor type best exposed on the ridge overlooking Dry Hollow from the southeast where it is exposed as a dark, massive bed lying concordantly within hornblende gneiss and quartz-feldspar gneiss. It consists of large
amounts of hornblende and hypersthene and a minor amount of spinel. It contains stringers of felsic material with dark shellys which evidently formed through local metamorphic differentiation. The rock is massive with a mosaic texture having many 120° angles at grain inter-
sections. Disseminated crystals of dark green spinel give it a speckled appearance in thin-section. Secondary serpentinization occurs along fractures.

The spotted granulite of sample 42, a rock composed of equal parts hornblende and hypersthene accompanied by accessory magnetite and spinel, occurs as concordant and discordant layers and dikes. In one place it occurs as a dike accompanied by many parallel aplite veins emplaced along a fault. Weathered surfaces are black and spotted with brown ovals between 2 and 3 centimeters long and half as wide which commonly parallel the foliation. They mark large crystals of hypersthene which poikilitically contain hornblende and are surrounded by it. The horn-
blene matrix forms a polygonal mosaic structure with grains intersecting at 120° angles and without preferred orientation. The dikes appear to have been emplaced before the F_1 deformation. Apart from the orientation of the oval spots, they seem to have behaved competently and were not greatly affected by the F_1 event.

Quartzitic gneiss. Abundant quartzitic gneisses occur within the hornblende gneiss and marble units. They normally occur as beds between 1 and 30 meters thick, but this increases to as much as 400 meters thick in the northern part of the area. The thick portions locally
contain iron-formation, the largest of which was mapped by James and Wier (1960) on the Kelly Ranch. The gneiss is composed mostly of quartz but may grade into quartzofeldspathic beds resembling the quartz-feldspar gneiss. Variations in feldspar content and color banding mark layers parallel to the foliation which is defined by the elongate grains of feldspar and quartz (Fig. 21). Differential weathering conspicuously emphasizes small isoclinal folds in some exposures.

The unit appears to have originated as sedimentary accumulations of impure sand. The presence of iron-formation indicates that at least some of the deposition occurred under water. Although the mechanism of deposition of iron-formation is not well agreed upon, most experts feel that they represent marine chemical precipitates. The cyclic nature of banded iron deposits is probably due to cyclic changes in the depositional environment (Trendall, 1973).

**Pegmatite.** Pegmatite veins of a few centimeters to dikes over 20 meters thick occur throughout the area. They exist as massive, coarse grained, pink colored rocks. Microcline forms over three-fourths of the rock and quartz, commonly graphically intergrown with the microcline, makes up most of the difference. Plagioclase is minor at 1-5%. Migmatitic areas have much associated pegmatite which appears to be locally derived. The major $F_2$ folding affects the smaller pegmatites and these do not appear fresh. The largest pegmatites cross-cut all structures and lithologies and appear the freshest in outcrop. I feel the older pegmatites formed locally during the metamorphism. The fresh
pegmatites intruded post metamorphically, possibly related to Tertiary plutonic activity in the general region. Small plutons are found in the Tobacco Root and Highland Ranges, and the Boulder batholith lies beyond them to the north.
Regional Metamorphism

Basement rocks in the Ruby Mountains underwent high-grade regional metamorphism. Okuma (1971), Garihan (1973), and Dahl (1977), established that conditions reached the upper amphibolite facies and ran locally into the granulite facies. Similar conditions existed regionally throughout southwestern Montana including the Highland Mountains (Gordon, 1979), the Tobacco Root Mountains (Cordua, 1973), the Madison Range (Thompson, 1960), and the Beartooth Mountains (Van de Kamp, 1969). Metamorphic conditions in the Hinch Creek area also reached the upper amphibolite facies and locally the granulite facies. The higher grades most probably represent areas locally undersaturated in water. The overall metamorphic conditions in the area still were higher than in the main part of the range to the southwest (Dahl, 1977).

Equilibrium assemblages, or mineral parageneses as defined by Winkler (1976), were used along with disequilibrium assemblages to determine the metamorphic grade. Disequilibrium textures resulting from prograde reactions helped greatly. Lack of appropriate assemblages makes it impossible to put narrow pressure limits although the temperature limits are well defined. The criteria used to determine that equilibrium was reached are: 1) all minerals in thin-section must be
somewhere in contact, 2) they lack disequilibrium textures, and 3) the assemblage contains no incompatible phases when plotted in an ACFmK diagram. Also, common 120° equilibrium/recrystallization grain junctions were noted in many samples. The following equilibrium assemblages occur in the area:

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<th>Mineral Assemblage</th>
<th>Rock Types</th>
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<td>quartz-plagioclase-perthite</td>
<td>quartz-feldspar gneiss</td>
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<tr>
<td>quartz-plagioclase-perthite-garnet</td>
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</tr>
<tr>
<td>quartz-plagioclase-perthite-biotite</td>
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<td>granulite</td>
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<tr>
<td>hornblende-hypersthene-spinel</td>
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</table>

Figure 22 shows the assemblages plotted on an ACFmK diagram. The assemblage: quartz-perthite-garnet-biotite-sillimanite-plagioclase cannot be listed as in equilibrium. The small amount of sillimanite prevents it from being in contact with each other mineral. The pairs
Figure 22a. ACFmK diagram of equilibrium mineral assemblages. Average rock compositions shown are: 1) Quartz-Plagioclase-Perthite, 2) Quartz-Plagioclase-Perthite-Garnet, 3) Quartz-Plagioclase-Perthite-Biotite, 4) Quartz-Plagioclase-Perthite-Garnet Biotite, 5) Quartz-Plagioclase-Biotite.
Figure 22b. ACFmK diagram of quilibrium ineral assemblages. Average rock compositions shown are: 1) Quartz-Plagioclase-Hornblende-Garnet-Diopside, 2) Hornblende-Plagioclase-Quartz-Garnet, 3) Quartz-Plagioclase-Hornblende-Garnet-Biotite, 4) Quartz-Plagioclase-Perthite-Hornblende, 5) Hypersthene-Garnet-Quartz
sillimanite plus potassium feldspar and sillimanite plus quartz lack
disequilibrium textures so the conditions probably reached the
sillimanite-orthoclase zone of the amphibolite facies.

In addition, the following disequilibrium textures proved very
useful in recognizing reactions and defining the metamorphic conditions:

- hypersthene grains with relict hornblende cores (Figure 19)
- wormy plagioclase surrounded by garnet (Figure 18)
- biotite and garnet separated by plagioclase (Figure 14)
- skeletal enstatite growing on anthophyllite (Figure 17)

Lastly, many general characteristics of the rocks and minerals indicate
high-grade conditions. Important characteristics include:

- K-feldspar is perthitic
- hornblende is a dark olive-green
- biotite pleochroism ranges from orange to deep red, almost black
- many samples have a granulitic texture
- plagioclase composition ranges from An$_{27}$ to An$_{42}$
- non-injection migmatites are common in the study area

The presence of perthite and hypersthene with quartz restricts the
lower temperature conditions of the rock. At pressures between 5.5 and
6.5 kb, the minimum temperatures for the formation of both perthite
and hypersthene with quartz are around 750°C in water saturated rock,
$P_{H_2O} = P_{load}$ (Fig. 23). The temperature could have been somewhat lower
in the case of where $P_{H_2O} < P_{load}$. The paragenesis hypersthene-garnet-
quartz indicates that the highest grades reached the orthopyroxene zone
Figure 23. Temperature-pressure diagram showing the field of metamorphism for this study (a) and the field determined by Dahl (1977) for the Kelly Iron-Formation (b). Reactions shown are: 1) Ky + And + Sill (Holdaway, 1971), 2) minimum melting curve for common granite (Thompson and Algor, 1977), 3) Mus + Q = Ksp + Sill + H2O, (Beach, 1973), 4) formation of hypersthene (Hyndman, 1972, p. 313), 5) Ksp + Perth (Huang and Wyllie, 1975), 6) Anth + Ens + Qz + H2O (Winkler, 1976, p. 162) breakdown of Hb + Qz (Hyndman, 1972, p. 313).
in rocks of appropriate composition. These samples have massive, granulitic textures as well. Since the majority of the rocks in the area reached the upper amphibolite facies, the presence of higher grade assemblages most probably represent areas locally undersaturated in water.

The disequilibrium assemblage anthophyllite-quartz-garnet-enstatite defines the upper temperature limit. According to the phase rule, being on the reaction boundary reduces to one the number of degrees of freedom, eg. the temperature is defined at any pressure, or vice-versa (Hyndman, 1972). For pure magnesian anthophyllite and enstatite, the reaction temperature between 5.5 and 6.5 kb pressure is just above 800°C where $P_{H_2O} = P_{\text{load}}$ (Fig. 24). The reaction anthophyllite $\leftrightarrow$ enstatite + quartz + water, yields water and would react at lower temperatures in $P_{H_2O} < P_{\text{load}}$ conditions. In addition, the presence of ferrous iron in the minerals as is the actual case will also generally lower the reaction temperature.

The disequilibrium textures between biotite and garnet discussed in Chapter II (page 38) also help define the upper temperature limit in $P_{H_2O} < P_{\text{load}}$ conditions. The unbalanced reaction Hb + An + Op $\leftrightarrow$ Alm + Cp + Ab + H2O defines the entrance to the granulite facies at approximately 760°C @ 10kb for conditions undersaturated in water (Buddington, 1966; DeWaard, 1967). Because of the steepness of the P/T reaction curves at high grades, the temperature at 6kb is probably not significantly different.
Although these reactions represent $P_{\text{H}_2\text{O}} < P_{\text{load}}$ conditions, it must be remembered that most of the area probably was saturated in water. The conditions were above the minimum melting temperature for the system quartz-plagioclase-orthoclase + water. Anatectic melting in water saturated rocks of granitic composition must have occurred. Migmatites found locally throughout the area support this conclusion.

Dahl (1977) has performed electron microprobe geothermometry and geobarometry on rocks from the Kelly iron deposit in the area. The results based on his work indicate a temperature of $745 \pm 50^\circ\text{C}$, and a pressure of 6.0 to 8.5 kb (Fig. 23). Little in the way of pressure limits were found in the present study but the temperature limits are well defined. The actual temperatures reached during metamorphism were probably slightly higher than those indicated by Dahl. Figure 23 shows the data given by Dahl and may be compared to the results of the present study.

Retrograde Metamorphism

The effects of retrograde metamorphism are not strong in the study area. Moderate to little or no alteration exists in most sections studied. The types of alteration found include chlorite after biotite, sericite in plagioclase, epidote in one amphibolite section, serpentine in fractures of ultramafic rocks.

The minerals produced indicate greenschist facies or lower conditions. The time of retrograde metamorphism could have been during the waning stages of the high-grade event, or a later low-grade event. In
any case, the effects are minor to nonexistent in the area. Only sericite alteration in plagioclase commonly is present in most sections studied.
CHAPTER IV

STRUCTURE

Multiple deformations have occurred in the basement of the Ruby Range. The earliest folds $F_1$, are characterized as similar and isoclinal with the development of axial plane schistosity (Garihan, 1973). The fold axes plunge northeast at moderate angles. The second folding, $F_2$, formed broad open folds coaxial with $F_1$ in the central and norther Ruby Range. Open to isoclinal $F_2$ folds not coaxial with $F_1$ exist in the southern Ruby Range (Okuma, 1971). Okuma found evidence of an $F_3$ deformation folding both $F_1$ and $F_2$. This $F_3$ fold axis trends north-south.

Faults in the range generally trend north northeast and cut all the folds. Exposure in fault zones is poor and most are located by juxtaposition of rock units or traced from air photographs. Precambrian diabase dikes follow some of these faults in the southern Ruby Range. I have seen no evidence to indicate whether dikes intruded along Precambrian faults, of whether the faults are Tertiary in age and followed the dikes as planes or weakness.

Folding

Garihan (1973) showed that two periods of folding occurred in the basement immediately to the southwest of the Hinch Creek study area.
Both sets are recognized in the study area and his terminology is retained.

\textit{F}_1 \textit{structures}. The earliest recognized period of deformation produced similar style isoclinal folds on a scale of several centimeters to several tens of meters (Fig. 12). The best exposures of this occur along the north side of Hinch Creek valley in section 32. The folding also produced axial-plane schistosity and regional foliation through the recrystallization and growth of new minerals. Quartz and feldspar commonly occur as elongate grains that parallel the foliation and compositional layering. Compositional layering parallels foliation except along the noses of isoclinal folds. Structural transposition is possible throughout the area, but was only found on a scale of a few tens of meters or less.

\textit{F}_2 \textit{structures}. The second deformation has formed the conspicuous northeast plunging antiform/synform pair which dominates the area. The axis of this major structure strikes between 055° and 065° and plunges between 50° and 70° (Fig. 24). The structure forms a broad, open fold. Marble units behaved competently during the deformation, pinching off and flowing into small lenses along the tight noses of some folds. Thin quartz lenses and layers may be folded and broken within structureless marble.

I subdivided the area into thirteen structural subdomains in order to analyze smaller smale changes and variations in the gross overall structure (Fig. 25). The basis for subdividing the area consists of:
Figure 24. Lower-hemisphere diagram showing the maximum concentration contours of poles to foliation from each domain plotted together along with measured beta points showing the average fold axis attitude found in six of the domain (labelled).
Figure 25. Map of the study area showing the domain subdivisions, and the average attitude of inferred fold axes and foliations within each domain. The values are: I) 094°, 56°E; II) 073°, 74°E; III) 090°, 60°N; V) 091°, 81°N; VI) 013°, 46°N; VII) 013°, 70°E; VIII) 019°, 58°N; IX) 047°, near vertical; X) 071°, 74°E; XI) 070°, 85°NW; XIII) 056°, 42°NE. The marble unit is indicated by the stipple pattern.
1) domain boundaries should parallel structural or lithologic boundary, and 2) the structures within a domain should show a single consistent pattern. The analysis is based on lower hemisphere poles to the regional foliation plotted on a Schmidt equal-area stereonet (Fig. 26a-m). Measurements on $F_1$ fold axes were not taken but the beta values obtained from the stereonets correspond to and are plotted on Figure 25. The concentration of data varies from one domain to another and affects the reliability of interpretation.

**Domain I.** Domain I consists mostly of hornblende gneiss and quartzitic gneiss metasediments. James and Wier (1960) have mapped the Kelly Ranch showing an east-plunging synform with the major iron deposit exposed in the center. Quartzitic gneiss surrounds the iron formation and grades away from the synform core into hornblende gneiss.

This domain differs from the dominant structural pattern of most of the remainder of the study area to the southeast. Structures trend east-west rather than northeast-southwest. A weakly defined girdle of poles to foliation agrees with the attitude of the synform mapped by James and Wier.

**Domain II.** This domain forms the transition between the east-west structure of Domain I and the northeast-southwest structures to the east. Again the girdle of poles to foliation is not well defined. The marble units form important structural marker beds within the domain. They strike between 050° in the eastern part of the domain to 090° in the western part. Lack of data prevents further subdivision of the domain.
Figure 26a. Domain I

Contour interval: 35%
28
21
14
7

Beta value: 094°, 56°E
14 points total

Figure 26b. Domain II

Contour interval: 25%
18
11
4

Beta value: 073°, 74°E
27 points total

Figure 26c. Domain III

Contour interval: 17%
11
5

Cluster center: 183°, 30°S
Average strike and dip: 093°, 60°N
18 points total
The domain contains the western limb of the broad antiform. It dips steeply to the north and northwest. The marble continues to the west where it folds around the iron formation, creating the synform of domain I.

Domain III. Fairly uniform quartz-feldspar gneiss dominates the domain with marble and hornblende gneiss included in the southern part. Poles to foliation form a cluster indicating an average attitude for all the foliation planes. To the south, the foliation in the hornblende gneiss strikes more northwesterly. This domain is separated from domain VI by the lack of north-south and northwest-southeast structures, and by the contact between the marble and the quartz-feldspar gneiss.

Domain IV. Domain IV exists in the negative sense that it does not fit well into any adjacent domain. Structurally it resembles domain VII representing north-south foliation dipping east. Physically it lies in the core of the antiform. This position relates it more with domains III and V. Defined by only five points, its significance remains uncertain.

Domain V. This occupies the nose of the antiform. The marble forms the best marker bed cutting across the domain and dipping north. Poles to the foliation appear to form a small circle of 14° radius about a point 181°, plunging 9° south. The small circle suggests a superimposed smaller scale open fold upon the main fold. Without more and different kinds of data, the direction and plunge of the axis cannot be found. For the major structure, the foliation dips north outward from the antiform. Due to the narrowness of the domain, not enough variation exists to indicate the direction and plunge of the axis of the antiform.
Figure 26d. Domain IV
Contour interval: 40%
20
5 points total

Figure 26e. Domain V.
Contour interval: 28%
21
14
7
Small circle diameter: 14°
Center: 181°, 9°S
Average strike and dip: 091°, 81°N

Figure 26f. Domain VI
Contour interval: 13%
10
7
4
Beta value: 013°, 46°N
21 points total
Domain VI. This domain contains the interlayered marble and quartz-feldspar gneiss at the nose of the synform. Marble beds clearly follow around the synform near the nose, but to the southwest they trend north-south. The data loosely defines a girdle of poles to foliation on the stereonet. The exact map location of the axis is not easily placed because of the openness of the fold in the domain. To the northwest the nose becomes tighter and the axis more clearly defined.

Domain VII. This domain, along with domain VIII covers the central limb of the antiform synform pair. The change in thickness of the marble along with the flattening and ending of the western marble bed forms the basis of the subdivision of the limb. Domain VII covers the south half of the limb. The marble unit separates the hornblende gneiss on the east from the quartz-feldspar gneiss on the west.

Poles to foliation form a small circle 32° in diameter around a point oriented 103°, plunging 20° west. This results from a small superimposed open fold on the main structure. As in domain V, the data is insufficient to determine the direction and plunge of the fold axis. The relative ages between this fold and the major fold cannot be shown. Both are post F₁.

Domain VIII. The northern half of the central fold limb dominates domain VIII. The beds of both domains VII and VIII dip east towards the synform. Combined with the outwardly dipping beds of domains V and II, the four domains define the western fold as an antiform. The marble beds mark the shape of the antiform in outcrop.
Figure 26g. Domain VII

Contour interval: 21%
16
11
6

Small circle diameter: 16°
Center: 103°, 20°W
Average strike and dip: 013°, 70°E

Figure 26h. Domain VIII

Contour interval: 19%
15
11
7
3

Beta value: 019°, 58°N
45 points total

Figure 26i. Domain IX

Contour interval: 15%
12
9
6
3

Beta value: 047°, near vertical
35 points total
The west half of the girdle of poles to foliation shows up on the stereo net. The beds dip steeply, generally over 60° and commonly in the upper 70's. This increases in the center of the synform where the fold axes plunge nearly vertical.

**Domain IX.** Domain IX occupies the center of the synform bounded by the marble on the east and the quartzitic gneiss on the west. A small ultramafic body crops out in the center of the domain on the axis of the fold. Desmarais (1978) found that other ultramafic bodies in the Ruby Range commonly lie on the axes of folds. He concluded that they were mobile during the folding phase, possibly being emplaced at that time. Metasedimentary units dominate the domain. The marble marker bed reveals an interesting structure. The east arm of the marble thickens and folds back across the axis of the main fold. Because the foliation bends around with the marble, this fold occurred after F1, probably related to the F2 phase of deformation. The relative ages between the limb fold and the main fold are uncertain. It was folded during or before the main phase because the main fold also folds the marble limb where they cross. A critical area adjacent to the northeast is covered and limits interpretation.

Poles to foliation concentrate in groups along the perimeter of the stereo net. The northwestern concentration indicates that the axial plane strikes approximately 45° northeast, but the fold axis plunges nearly vertically.
Domain X. This domain consists of the northern half of the synform's eastern limb. Rock types range from hornblende gneiss on the west to marble, hornblende gneiss, and amphibolite on the east. Some bedding in the domain, especially near Hinch Creek is isoclinally folded by $F_1$ deformation.

The foliation dips steeply to the east on this limb showing that it is slightly overturned. Poles to foliation form a partial girdle, enough to weakly define a beta position. A smaller wave on the main fold shows up on the map as the bulge to the northwest in marble and anthophyllite beds.

Domain XI. Domain XI indicates the south half of the synform's eastern limb. I based the subdivision on the fact that foliation in this area dips steeply northwest while the foliation of domain X dips steeply southeast. Both are near vertical. The boundary follows the prominent cross-cutting pegmatite and is in line with the domain VII and VIII boundary. Poles to foliation plot loosely as a cluster. This gives an average strike of 070°, and a dip 85° northwest.

Minor isoclinal folds exist within the domain, mostly within the marble unit. This results in thickening the marble bed on this limb. Insufficient data prevents me from assigning an $F_1$ or $F_2$ designation to these smaller folds. The noses are broken and poorly exposed.

Domain XII. This domain, as in the case of domain IV, exists in a negative sense of not fitting with adjacent domains. Physically this is represented by a marble unit extending into the quartz-feldspar
Figure 26j. Domain X

Contour interval: 31%
25
19
13
7
1

Beta value: 071°, 74°E
55 points total

Figure 26k. Domain XI

Contour interval: 20%
13
6

Cluster center: 160°, 5°S
Average strike and dip: 070°, 85°NW
15 points total

Figure 26l. Domain XII

Contour interval: 40%
20

5 points total

Figure 26m. Domain XIII

Contour interval: 18%
12
6

Beta value: 056°, 42°NE
17 points total
gneiss. It does not connect to adjacent marble units within the area. The five points defining the domain are not interpretable.

**Domain XIII.** This domain occupies the southern corner of the map area. The thick marble unit dominates it on the south, balanced by quartz-feldspar gneiss to the north. Isoclinal folding and possible plastic flowage cause thickening in the marble. One isoclinal fold was mapped within the marble atop the east-west trending hill in the center of section 8. Lack of exposure makes it impossible to determine the extent or attitude of this isoclinal folding. The trend of fold axes and the position of the domain relates this to the east limb of the synform. The northwest trend separates this domain from the more north-south trending domain VI.

**Summary of Folding.** The earliest folding, $F_1$, in the area occurred as isoclinal similar-style folds. Garihan (1973) found the fold axes to plunge northeast at "moderate" angles, moderate to steep in this study. The deformation coincided with the highest grades of metamorphism as indicated in Chapter III. This deformation was accompanied by mineral growth and recrystallization to create the regional foliation found in the area.

The second deformation, $F_2$, folded the foliation and compositional layering, resulting in the broad northeast-plunging fold that dominates the structure of the area. The presence of smaller scale folds complicates the interpretation. The age relationships are unknown between these and the $F_2$ fold. These are smaller in scale and not found
throughout the study area; this suggests a relationship with the $F_2$ rather than a separate regional folding event. No evidence was found for a third folding phase in the area.

**Faulting**

Within the study area two northwesterly trending faults were mapped but neither could be traced along strike more than a few tens of meters. Other faults in the area were indicated by gossens and prospect pits with exposed malachite and limonite in the shear area. Hornblende-hypersthene granulite occupies one fault. It probably represents a small serpentinite dike injected prior to metamorphism. This shows that some faulting occurred before the onset of metamorphism and deformation.

Major faults bound the area on two sides. To the west the study area sits against Paleozoic sedimentary rocks in contact with the basement along north-northwesterly vertical faults. These faults, active in the Tertiary period form a part of a series of northwest "master" faults mapped by Tysdal (1970) (Figure 1). On the north the Tertiary range-bounding fault system brings the basement into contact with Paleozoic sedimentary rocks. The faults here are covered by surficial sediments.
CHAPTER V
CONCLUSION

The rocks within the area represent a package of metasedimentary with some interlayered metavolcanic units. During high grade metamorphism the region was intruded by a granitic pluton which now forms the structurally lowest unit. The metasedimentary lithologies of the area include hornblende gneiss, sillimanite-garnet-biotite-quartz-feldspar gneiss, quartzitic gneiss, iron-formation, and marble representing respectively: carbonaceous mudstone or shale, pelitic sediments, sandstone, iron-formation, and limestone. Interlayered with these in the structurally lower part of the section are amphibolites representing metavolcanic flows or sills. Whether they existed as a primary unit of the sequence as flows, or were intruded as sills at a later time is unknown. The rock recrystallized completely during metamorphism, destroying any primary structures. It is possible that both flows and sills are represented.

The sedimentary sequence starts at the base with marble and works upsection through metavolcanics and hornblende gneisses with thin layers of quartzitic and quartz-feldspar gneisses. The quartzitic gneiss becomes dominant and in the Kelly Ranch it contains banded iron-formation. Iron-formation also occurs in units crossing Hinch Creek to the south. The two iron-formations are at approximately the same structural level although the Kelly area contains much more quartzitic
gneiss. The thickness of this gneiss may be controlled by isoclinal folding or the angle of exposure causing real or apparent differences in thicknesses between the units in the two areas. The structural positions permit a tentative correlation between the two. Above the iron-formation is a second marble layer followed by more quartzitic gneiss and thin layers of hornblende gneiss. Due to isoclinal folding and possible plastic flowage during the deformation, the true thicknesses of the original sedimentary units is unknown.

The origin of the quartz-feldspar gneiss has long been a problem. Nowhere in the Ruby Range has definite proof of either a sedimentary or plutonic origin been found. In the Hinch Creek study area the gneiss lies concordantly below the main marble unit. It shows no cross-cutting relationships, but in the southern part of the area it interbeds with the marble. This bedding partially results from isoclinal folding but may also be due to original sedimentary layering, or intrusion of granitic melt as sills. Synkininmatic intrusion deep in the crust would not be expected to cause much cross-cutting or deformation of the country rock (Buddington, 1959). It is not unreasonable to conclude that all or much of the quartz-feldspar gneiss was formed as a granitic magma. The temperature and pressure conditions determined in the study area were well above the minimum melting curve for water-saturated granite (Fig. 23). Migmatitic areas are found throughout the study area indicating partial melting. The granitic composition and lack of any relict sedimentary features also points towards a possible plutonic origin.
The major pre-metamorphic rock types were limestones, sandstones, pelitic sediments, and shales. Volcanic rocks make up a minor and possibly post-depositional contribution to the package. The presence of iron-formation and limestone indicates deposition in a shallow marine environment. Shales and calcareous shales have a deeper or quieter environment of deposition. The package therefore most likely represents a marginal sea or shelf-type of environment. The lack of abundant volcanics suggests deposition during a period of minor or no tectonic and thermal activity, an inactive or trailing margin in the modern sense. The same case exists throughout much of the basement of southwestern Montana. Common rock types are repeated over and again, quartz-feldspar gneiss, biotite-sillimanite schist, marble, hornblende gneiss, and amphibolite. The shelf area was regional in extent. The direction to the source area is unknown.

Sometime after deposition the entire pile was subjected to regional deformation and metamorphism. These may or may not have begun at the same time. The highest levels of temperature and pressure were reached during or after the first period of deformation with the growth of sillimanite on the $F_1$ axial plant. Intrusion of the Dillon gneiss "pluton" occurred synkinematically with $F_1$. Crystallization during the deformation imparted a strong gneissic fabric to most of the unit.

Textures and mineralogies found within the study area indicate that temperatures during metamorphism reached between 750° and 800°C with a pressure range between 2 and 8kb. Results of geobarometry done by
Table 5. Interpreted geochronology of events affecting the basement lithologies in the Ruby Range, southwestern Montana.

<table>
<thead>
<tr>
<th>Age (m.y.a.)</th>
<th>Event</th>
<th>Evidence</th>
</tr>
</thead>
</table>
| Eocene to present | Movement on northwest trending master faults and range bounding faults.  
Emplacement of basalt plugs. | Displacement of all lithologies from Precambrian units to recent valley-fill sediments. |
| 1200        | Exposure of the basement to the surface.                              | Basement metamorphic clasts incorporated into the lower Belt sedimentary rocks (Obradovich and Peterman, 1973). |
| 1400-1700   | Intrusion of diabase dikes on a northwest trend normal to the range axis. | Metamorphism in the dikes ranges from greenschist to unaltered. The intrusions may or may not have followed a Precambrian plane of weakness. Wooden (1975) reports a 1450m y Rb-Sr isochron from a fresh diabase. |
| 1600        | "Ending" of the greenschist facies metamorphism                      | 1600my is the most common K-Ar mica date reported from the Ruby Range (Giletti, 1966). Mica blocking temperatures range from 150° to 300° (Fountain, personal comm.). Winkler (1974) puts a temperature range for the greenschist facies between 200° and 500°C. By 1600my the temperature had fallen into, and possibly below the greenschist facies.  
Falling temperatures were probably due to slow uplift and erosion of the basement. |
Table 5. (Continued)

<table>
<thead>
<tr>
<th>Age (m.y.a.)</th>
<th>Event</th>
<th>Evidence</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td><strong>F₃</strong> folding</td>
<td>Folds both <strong>F₁</strong> and <strong>F₂</strong>, not coaxial with either. Folds trend north-south (Okuma, 1971).</td>
</tr>
<tr>
<td></td>
<td>(southern Ruby Range only)</td>
<td></td>
</tr>
<tr>
<td></td>
<td><strong>F₂</strong> folding</td>
<td>Folds <strong>F₁</strong> to form a broad northeast-plunging antiform-synform pair in the study area.</td>
</tr>
<tr>
<td></td>
<td>tight to open folds, rarely isoclinal. Associated with <strong>F₂</strong></td>
<td>No evidence was found to show whether the greenschist facies is a late stage of the high-grade event, or a separate younger event.</td>
</tr>
<tr>
<td></td>
<td>is another smaller-scale folding</td>
<td></td>
</tr>
<tr>
<td></td>
<td>of uncertain age relationship.</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Intrusion of basic dikes and sills.</td>
<td>These cut <strong>F₁</strong> and are folded by <strong>F₂</strong> (Garihan, 1973).</td>
</tr>
<tr>
<td></td>
<td>Highest grades of temperature and pressure reached after <strong>F₁</strong>.</td>
<td>Metamorphism in the dikes locally reaches the granulite facies. They were intruded after <strong>F₁</strong>, but still during the regional metamorphism.</td>
</tr>
<tr>
<td></td>
<td>Pegmatite intrusions during <strong>F₁</strong>.</td>
<td>Heinrich (1960) found pegmatites foliated and concordant with <strong>F₁</strong> structures. These are not found in the present study area.</td>
</tr>
<tr>
<td></td>
<td>Beginning the upper amphibolite to lower granulite facies</td>
<td>No evidence was found to show whether the metamorphism began at the same time as the <strong>F₁</strong> folding phase. Sillimanite crystals growing on the <strong>F₁</strong> schistosity show that the sillimanite-orthoclase zone was reached during <strong>F₁</strong>.</td>
</tr>
<tr>
<td>Age (m.y.a.)</td>
<td>Event</td>
<td>Evidence</td>
</tr>
<tr>
<td>-------------</td>
<td>----------------------------------------------------------------------</td>
<td>--------------------------------------------------------------------------------------------</td>
</tr>
<tr>
<td>2800</td>
<td>Synkinematic intrusion of the Dillon gneiss during $F_1$.</td>
<td>Rb-Sr whole-rock dates of 2700-2800 my are reported in the Ruby and adjacent ranges (Giletti, 1966; Catenzaro, 1967; Wooden, 1975; Wooden and others, 1978; James and Hedge, 1980).</td>
</tr>
<tr>
<td>2800</td>
<td>$F_1$ folding isoclinal, similar-style folds with axial-plane schistosity. Folds plunge northeast at moderate angles.</td>
<td>Crystalization during the deformation imparts the gneissic fabric in the unit.</td>
</tr>
<tr>
<td>pre-2800</td>
<td>Pre-syntectonic emplacement of ultramafic bodies.</td>
<td>Develops the regional foliation by recrystallization and growth of new minerals oriented parallel to $F_1$.</td>
</tr>
<tr>
<td>pre-2800</td>
<td>Deposition of sedimentary units and intrusion of volcanic sills and flows. The sedimentary units indicate a shelf-type environment.</td>
<td>Desmarais (1978) found that these bodies were probably crystallized, intruded, and serpentinized before the beginning of regional metamorphism. They behaved competently during metamorphism and their association with the noses of folds indicates mobility during folding. Shelf-type character of the sedimentary package. Limestones and magnetite iron-formation indicates shallow marine deposition.</td>
</tr>
</tbody>
</table>
Dahl (1978) show pressures of 6 to 8.5kb. If correct, the temperature range would be more restricted, between 750 and 800°C (Fig. 23). These conditions are found within the upper amphibolite to lower granulite facies. This agrees with metamorphic conditions reported from all the ranges with basement exposed in southwest Montana. This common high grade metamorphism throughout the region suggests a common period of metamorphism.

Much of the region experienced greenschist facies retrograde metamorphism. No evidence exists to show whether this was a late stage of the major event, or a separate younger event. The effects of this metamorphism in the study area are weak to nonexistent. After the first folding event a second deformation folded the foliation into a broad north-east plunging fold set. The age relationship between this and the greenschist metamorphism is unknown. The second deformation includes a smaller set of superimposed folds of undetermined relative age. No further folding occurred in the study area, but Okuma (1971) reports a north-south trending F_3 fold in the southern Ruby Range.

The last major event affecting the basement before it was exposed at the surface was the intrusion of a series of diabase dikes and sills. The process covered a long time span and affected the entire southwest Montana region (Wooden and others, 1978). The intrusion period went from 1700 my to 1400 my in the basement rock (Wooden, 1975) and possibly to as recently as 1200 my in the Belt Supergroup (Obradovich and Peterman, 1973). The earliest dikes were intruded and metamorphosed
during the regional greenschist facies event. Fresh diabase from the Ruby Range gives a Rb-Sr isochron of 1450 my, therefore the greenschist facies event had ended by then. No diabase occurs in the study area.

In conclusion, the rocks from the Hinch Creek study area formed in a shelf-type depositional environment. Neither the top, nor the base of this package is found within the study area. The entire package was subjected to intense deformation and upper-amphibolite-facies-metamorphism. This was followed by greenschist facies metamorphism and a second regional folding. This resulted in the broad antiform-synform pair which dominates the area. The area has been uplifted and exposed by vertical faults active since Tertiary time. Several Tertiary basalt plugs have punched up through the basement terrain but caused no observable deformation.
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