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Structural and Petrologic Study of Precambrian Ultramafic Rocks Ruby Range Southwestern Montana

Neal Raymond Desmarais

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STRUCTURAL AND PETROLOGIC STUDY OF
PRECAMBRIAN ULTRAMAFIC ROCKS,
RUBY RANGE, SOUTHWESTERN MONTANA

by

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B.A., University of Massachusetts, 1976

Presented in partial fulfillment of the
requirements for the degree of

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ABSTRACT

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Structural and Petrologic Study of Precambrian Ultramafic Rocks,
Ruby Range, Southwestern Montana (88 p.)

Director: David Alt

Ultramafic rocks occur in many localities throughout the Ruby Range of southwestern Montana. Field and petrographic study suggests that the ultramafics were partially serpentinized tectonites pre- or syntectonically emplaced, possibly in an accretionary environment and metamorphosed to the upper amphibolite facies. Typically the ultramafics are elongate, elliptical bodies which tend to parallel the regional foliation and range in size from a few meters to hundreds of meters in length. They often exhibit a planar fabric concordant with the foliation in the surrounding rocks. In places the ultramafics acted as competent units forming large boudins. Equilibrium assemblages of hypersthene, forsterite, and green spinel with and without actinolite, enstatite and cummingtonite/anthophyllite give P/T conditions compatible with the upper amphibolite metamorphic event indicated in the surrounding rocks. Replacement textures indicate progressive metamorphism involving deserpentinization of partially serpentinized ultramafic rock. Relict orthopyroxene poikiloblasts exhibit strain features such as undulatory extinction, kink banding, microfractures, subgrain development, and granulated margins. Recrystallized orthopyroxene and forsterite in the matrix and schistose margins of the bodies are relatively strain-free. These observations are inconsistent with the previously proposed post-metamorphic magmatic origin of these bodies.
ACKNOWLEDGMENTS

Special thanks go to Dr. Dave Alt and Dr. Dave Fountain for their encouragement and guidance throughout this project. Conversations with numerous colleagues, especially James McKee and Gil Wiswall, helped me formulate and clarify the ideas presented in this study. The patience and continuous encouragement of my wife, Marilee, as well as her typing skills, were invaluable in completing the project. Field work was supported in part by a Grant-in-Aid of Research from the Society of Sigma Xi.
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CHAPTER I
INTRODUCTION

General Statement

The geotectonic significance of ultramafic rocks has been summarized in a number of papers (e.g. den Tex, 1969; Moores, 1973a, b; Naldrett, 1973). Although chemically and mineralogically similar, ultramafic rocks show significant differences in mode of occurrence, and hence, in tectonic significance (Moores, 1973a; Naldrett, 1974). The major ultramafic rock occurrences are as stratiform complexes, concentric complexes, and alpine complexes (Jackson and Thayer, 1972). Stratiform and concentric bodies are generally regarded as igneous and form in non-orogenic and orogenic environments respectively (den Tex, 1969; Moores, 1973a, b; Naldrett, 1973). Alpine ultramafics are thought of as cold solid bodies tectonically emplaced into the crust (e.g. Moores, 1973a, b). The variability of these alpine ultramafic bodies is exemplified in the number of subdivisions geologists devised to classify them (see Moores, 1973b for discussion). However, most geologists divide the alpine-type complexes into two major subtypes (Misra and Keller, 1978):

(1) Ophiolites which may occur as:

(a) Allochthonous sheet-like bodies with thrust contacts, preserving the entire ophiolite sequence, or only part of it; or


(b) Chaotic blocks in mélange terranes, torn from the front of advancing thrust and incorporated in the subsequently overridden mélange below the thrust;
(2) Tectonic or diapiric intrusives, which may occur as relatively small, lenticular bodies in both metamorphosed and unmetamorphosed terranes of sedimentary and volcanic rocks of eugeoclinal affinities.

Of particular interest in this study are the alpine ultramafics of the second subtype, ultramafic bodies found in regionally metamorphosed terranes. These ultramafics encompass what den Tex (1969) called "root zone" ultramafic bodies, what Moores (1973b) characterized as "conformable bodies in regionally metamorphosed terranes" and what Naldrett (1973) referred to as "ultramafics associated with major crustal sutures." The tectonic significance of these ultramafic bodies, especially in Precambrian terranes, is not as clear as that for true ophiolitic alpine ultramafic complexes. Little work has been done on such ultramafic rocks in Precambrian mobile belts, and the time relationships between emplacement of ultramafic rocks and mineral deformation and metamorphism are generally unclear (Naldrett, 1973). The Precambrian terrane of southwestern Montana affords an excellent opportunity for the study of such ultramafic rocks.

Complexly deformed Precambrian high grade metamorphic rocks form the cores of numerous mountain ranges in southwestern Montana (Fig. 1). This terrane encompasses a Precambrian age province boundary between Hudsonian age (1.7 b.y.B.P.) rocks and Kenoran age rocks (2.7 b.y.B.P.)
Fig. 1 General location map showing age of Precambrian basement uplifts in southwestern Montana.
of the Wyoming Province. The Precambrian rocks of this area are primarily quartzofeldspathic gneisses, hornblende gneisses and amphibolites along with distinct metasediments such as marbles, quartzites, and pelitic schists. Scattered throughout much of this terrane are numerous small bodies of ultramafic composition. Most of these bodies have escaped much serious attention. However, I believe some understanding of these rocks is critical to the formulation of any tectonic model of this area. The objective of this study was to add to the general knowledge of ultramafic rocks of this type and to better understand the origin of these bodies and their relationship to the tectonic evolution of the area.

Location of Study Area

The Ruby Range of southwestern Montana contains numerous bodies of Precambrian ultramafic rocks. The area was chosen because of its accessibility and relatively good exposures of ultramafic rocks. The Ruby Range is just east of Dillon, Montana and is accessible by the Stone Creek Road, the Sweetwater Road, and the Blacktail Road (Fig. 2). Logging roads and ranch roads provide further access to the interior portions of the Range. The detailed map Areas I, II, III in Figure 2 are located as follows:

I. Along Elk Gulch

II. Red Canyon Area

III. Just northeast of the Sweetwater Road along the crest of the Range.
Fig. 2 Location of ultramafic bodies and Study Areas in the Ruby Range.
Previous Study and Methods of Investigation

Sinkler (1942) first studied the ultramafic rocks in the Ruby Range. This study described the basic mineralogy, field relationships and economic potential of the Wolf Creek Pluton in the southwestern portion of the range (Area I on Fig. 2). Sinkler called the ultramafic body "saxonite" and regarded it as intrusive into the metamorphic complex.

The most comprehensive work done on any of the ultramafic rocks was Heinrich's (1963) study of the Wolf Creek Pluton. This body is located in the southeast corner of the range (Area I on Fig. 2). Heinrich (1963) noted the rocks contain orthopyroxene, olivine, spinel, and called them harzburgites. He suggested the body was forcibly emplaced, probably as a largely liquid, hydrous peridotitic magma. He noted that distortion of the layering in the country rocks and rotation of foliation in xenoliths occurred around the body. According to Heinrich (1963) the ultramafic body underwent various alterations to serpentine, anthophyllite, actinolite, chlorite, talc, calcite, clinohumite, and annabergite. He believed that autometasomatism during the later stages of peridotite crystallization accounted for the secondary mineral assemblages.

Okuma (1971), in his study of the structural geology and talc formation in the southern Ruby Range, mentions the ultramafic rocks briefly. He generally agrees with Heinrich's (1963) interpretation of the bodies. However in Okuma's paper, Hsü (oral communication in
Okuma, 1971, p. 34) first pointed out the possibility that the ultramafic rocks may be tectonic slices emplaced in a pre-Beltian geosyncline.

Garihan (1973) described some of the petrology and mineralogy of ultramafic rocks in the central Ruby Range. Garihan (1973) postulated that the ultramafic rocks were emplaced before or during the complex isoclinal folding and upper amphibolite metamorphic event of the area.

The field work for this study was done during the summer of 1977. In order to gain a regional view of the ultramafic rocks, I visited ultramafic localities throughout the Ruby Range. Locations and field relationships (foliations, concordance with surrounding rocks, etc.) were recorded and samples collected for later lab work. Three areas (I, II, III, in Fig. 2) were selected for detailed study and outcrop mapping (see Figs. 5, 6, 10). I collected detailed structural data and samples in each area and some oriented samples for petrofabric analysis. Particular attention was paid to the spatial relationships of the various rock types and the relation of the ultramafic rocks to major structures. Areas I and II were mapped on a scale of 1"=500' and Area III on a scale of 1"=400' using a combination of enlarged aerial photographs and USGS topographic maps.

I collected 170 samples for laboratory analysis, and 115 thin sections were prepared for petrographic study. Oriented thin sections were used for petrofabric analysis done on a four-axis Zeiss universal stage. Optical properties determined on the flat stage provided the principal criteria for mineral identification.
In some cases, X-ray diffraction and infrared spectrophotometry were used in identifying minerals. All X-ray work was done using Ni-filtered Cu K radiation. Petrofabric and structural data were plotted on stereonets to aid in analysis.

Regional Setting and General Geology

The Ruby Mountains form an elongate northeast-trending, fault-bounded range, one of a number of Precambrian basement uplifts in southwestern Montana (see Fig. 1). The range is composed almost entirely of high grade metamorphic rocks but Paleozoic sediments unconformably overlie the metamorphic rocks in its extreme northeastern end. Elevations vary from around 5,000 feet to over 9,000 feet at the northeast end. The central and southeast portions of the range are a broad, gently rolling, plateau-like upland which ends in the sharp peaks of the northeast section. A number of northwest-trending faults (Okuma, 1971; Garihan, 1973) cut directly across the range offsetting the erosion surface. The ultramafic bodies outcrop on this surface and form resistant outcrops, invariably expressing themselves topographically as knobs and low ridges.

Factors that may have contributed to the relative weathering differences between the ultramafic bodies and the immediately surrounding schists and gneisses are: 1) grain size, 2) permeability, 3) chemical weathering in an appropriate geochemical weathering environment.
The ultramafics are generally coarser-grained and more massive than the surrounding schists and gneisses and grain size and permeability differences may have contributed to the apparent slower weathering rate of the ultramafics. The erosion surface in this area presumably developed under arid to semi-arid conditions with an associated alkaline weathering environment (Petkewitch, 1972; Monroe, 1974; Alt, personal communication). In an alkaline environment the silica-rich host rocks would be more susceptible to weathering solutions and may weather faster than the silica-deficient ultramafic rocks. Solution weathering may have also produced a protective iron-coating on the ultramafic bodies as evidenced by the brown iron oxide coatings and crystalline hematite occasionally found on the weathered surfaces of the ultramafic rocks.

Earlier workers (Heinrich, 1960, 1963; Okuma, 1971; Garihan, 1973) subdivided the Precambrian metamorphic rocks of the Ruby Range into three major groups; the Cherry Creek group, the Pre-Cherry Creek group and the Dillon Granite Gneiss. The Cherry Creek group is a metasedimentary unit, originally defined by Peale (1896). In the Ruby Range it consists of marble, calcium-magnesium silicate gneisses and schists, quartzites, biotite and muscovite schists, sillimanite schists and gneisses, amphibolite, and magnetite schists (Heinrich, 1960; Okuma, 1971). Heinrich (1960) applied the general name Pre-Cherry Creek to a group of rocks consisting of garnet and anthophyllite schists and gneisses, quartzofeldspathic gneisses, amphibolites, hornblende gneisses and migmatites in the southwestern
portion of the range. The Dillon Granite Gneiss consists of quartzofeldspatic gneisses, some granitic in appearance (Heinrich, 1960; Okuma, 1971). Heinrich and Okuma considered this body as a batholith syntectonically intruded into the older Cherry Creek and Pre-Cherry Creek units. This relationship is far from clear and my own field observations lead me to believe that much of this unit is metasedimentary and conformable with the Cherry Creek and Pre-Cherry Creek as originally suggested by Garihan and Okuma (1974) and Garihan and Williams (1976).

In discussing the rocks of this area I find it more meaningful and useful to use consistent rock packages such as quartzofeldspatic gneiss, amphibolite, hornblende gneiss, or metasediment, rather than the broad general names of Cherry Creek, Pre-Cherry Creek and Dillon Granite Gneiss. I believe the broad units may have little meaning especially when trying to correlate them with rock packages from other areas (see Fountain and Desmarais, in press).

All the Precambrian rocks of this area are complexly deformed. They display isoclinal folds, a well-developed axial plane schistosity, (generally trending northeast) and an upper amphibolite grade of metamorphism. The ultramafic rocks are scattered throughout the range (Fig. 2) and occur in all of the above rock types.

Plate tectonics has sparked a keen interest among geologists in crust-mantle interactions resulting in an abundance of recent literature on ultramafic rocks. In light of this recent literature it is evident that a reevaluation of the Precambrian
ultramafic rocks in southwestern Montana is needed. This study represents such a reevaluation in the Ruby Range.

Consistent field, petrologic, and textural relationships in the ultramafic rocks of the Ruby Range indicate that they participated in the deformation and upper amphibolite event of the area. Textures and mineral assemblages suggest that the ultramafics underwent a period of serpentinization after primary crystallization and before metamorphism. Since this serpentinization, emplacement and deformation with accompanying upper amphibolite metamorphism produced characteristic and distinctive assemblages, textures, and field relationships in the ultramafic rocks of the Ruby Range. After the upper amphibolite metamorphism, low-grade reserpentinization of the ultramafic bodies occurred.

Regarding the ultramafics in the Ruby Range as tectonites emplaced early in the tectonic history of the area has regional tectonic implications. One possibility is that the area represents an accretionary environment in which these ultramafic tectonites were emplaced into a package of metasedimentary and metavolcanic rocks along the Precambrian age province boundary. Other Precambrian terranes with similar rock types, structures, and spatial association with a Precambrian age province boundary have been interpreted in this manner and are thought to represent major crustal sutures (e.g. Gibb and Walcott, 1971; Burke and Dewey, 1973).
The circum-Ungava suture of the Canadian Shield, composed of the Nelson Front, the Cape Smith Fold Belt, and the Labrador Trough, separates the Superior Province gneisses from reactivated Hudsonian crystalline rocks. This has been interpreted as a product of Precambrian collision orogeny (Wilson, 1968; Gibb and Walcott, 1971; Burke and Dewey, 1973). The rock types, structural styles and grade of metamorphism are very similar to the Precambrian terrane of southwest Montana suggesting a similar origin (Fountain and Desmarais, in press).
CHAPTER II
FIELD RELATIONSHIPS

The ultramafic rocks usually occur as elongate elliptical bodies with their long axis paralleling the regional foliation. They range in size from a few meters to hundreds of meters in length. All of the ultramafic bodies have consistent megascopic field characteristics. Most are texturally zoned with megacryst-bearing interiors and equigranular schistose margins (Fig. 3). In a few cases these textural zones alternate within a body and, in others, the more massive megacryst-bearing interiors are completely absent. One locality in Study Area III provides a good example of the alternating layers (Fig. 4). Here schistose layers half a meter thick alternate with megacryst-bearing layers up to a meter thick.

The rocks often exhibit a planar fabric which is best developed in the schistose margins. In the megacryst-bearing interior portions the fabric, when visible, is defined by the parallel alignment of stretched or elongate orthopyroxene grains and alignment of amphibole in the matrix. The fabric in the margins results from the alignment of equigranular bladed amphibole grains and sometimes mica grains. Later planes of serpentinization parallel to the foliation enhance the fabric in some localities. This fabric is consistently concordant with the foliation in the surrounding rocks.
Fig. 3  Schematic field sketch showing textural zonation of ultramafic bodies.

Fig. 4  Interlayering of schistose-S and megacryst-M bearing layers in ultramafic rocks.
Study Areas I and II provide some excellent examples of concordant foliations and the relationships of the ultramafic bodies to major structures (Fig. 5 and 6). The major structure in Area I is apparently a large fold which was refolded about a north-trending axis in the northwest end of the area. The Elk Creek fault apparently obscured the southwestern portion of the fold. A stereographic plot of poles to foliation for the area produces a diffuse girdle pattern consistent with this interpretation (Fig. 7).

Figure 8 is a close-up of part of the outcrop map of this area showing a series of tight isoclinal synforms (parasitic folds) along the limb of the larger fold. The foliations within the ultramafic bodies and the surrounding rocks are concordant throughout the structures and the ultramafics appear to be intimately interlayered with their host rocks. Subsurface well logs of this area, provided by Dekalb, Inc. of Calgary, Alberta, confirm this interlayering at depth (Fig. 9).

The occurrence of ultramafic bodies at or near the crests of isoclinal to subisoclinal folds is an association found elsewhere in the range as well. Study Area III is a series of tight isoclinal folds with a number of small ultramafic bodies apparently caught up in the folding (Fig. 10). A stereographic plot of poles to foliation for this area produces a tight girdle reflecting this isoclinal folding (Fig. 7). The folding occurred about an approximate axis of N15E/55N.
Fig. 5  Outcrop map of Area I

- **Meta-Ultramafic (Includes Partially Recryst. & Recryst. Assemblages)**
- **Quartz-Feldspathic Gneiss**
- **Hornblende Gneiss**
- **Serpentinite**
- **Rodingite (Meta-Rodingite?)**

(For More Detailed Map Of Area See Bound Library Copy)

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Outcrops

Strike & Dip of Foliation

Plunging Overturned Synform

---

0 500

METERS
Fig. 6 Outcrop map of Area II
Fig. 7  Stereographic plot of poles to foliation in Areas I and III. Contour intervals are .7%, 2%, 4%, 6% per 1% area for Area I, and .1%, 3%, 6%, 9%, 12% per 1% area for Area III. - fold axis.
Fig. 8 Close-up of part of outcrop map of Area I, showing concordant foliation and interlayering of ultramafic bodies and host rocks.
Fig. 9  Schematic representation of well logs from Area I showing interlayering of ultramafic rocks with host rocks at depth (see Fig. 5 for location of drill holes).
Fig. 10 Outcrop map of Area III
In Study Area III the ultramafic bodies also exhibit concordant foliations and are interlayered with their host rocks. Figure 11 is a close-up of part of the outcrop map of this area showing the host rocks pinching and swelling around the ultramafic pods. These pods are separated by quartz-feldspar pegmatites. Apparently the ultramafic rocks acted as competent units during the deformation forming this boudinage. Throughout the range the ultramafic bodies have similar field relationships of concordant foliations and are interlayered with their host rocks. I believe such field relationships strongly support a synchronous deformation of the ultramafic bodies and their host rocks.
Fig. 11 Close-up of part of outcrop map of Area III showing boudinage-like structure with host rocks pinching and swelling around ultramafic pods.
CHAPTER III
PETROLOGY OF ULTRAMAFIC AND RELATED ROCKS

General Statement

Throughout the Ruby Range the ultramafic rocks have strikingly similar mineralogical and textural characteristics. The ultramafic rocks consist primarily of orthopyroxene, amphibole (actinolite/tremolite, anthophyllite/cummingtonite), spinel and olivine. Most ultramafic bodies show a pronounced textural zonation of coarse-grained, inequigranular, more massive material at the centers of the bodies grading to schistose, finer-grained material at the margins (see Fig. 3). This textural zonation reflects the degree of mylonitization and recrystallization in the ultramafic rocks (see discussion below). This textural variation also provided the basis for classifying and mapping the ultramafic rocks in the Ruby Range. I divided the ultramafic rocks into two major groups; the partially recrystallized assemblage and the recrystallized assemblage. The first group corresponds to the coarser-grained, inequigranular, massive variety and the second to the finer-grained, equigranular schistose variety.

Closely associated with the ultramafic rocks are a number of distinctive metasomatic rocks formed by diffusion metasomatism (Evans, 1977) during metamorphism. Juxtaposition of high Mg, low Si, Fe, K, ultramafic rocks and rocks of contrasting composition (e.g. more
siliceous rocks) results in significant mass transfer of material down steep gradients in chemical potential (Evans, 1977). The ultramafic rocks contain variable amounts of alteration products, primarily serpentine. Extensive serpentinization of both partially recrystallized and recrystallized assemblages forms a third group of ultramafic rocks, the serpantinites. Associated with this group are a number of distinctive, low temperature, metasomatic rocks formed by infiltration metasomatism (Evans, 1977). Most notable among this group are the calc-silicate-rich rodingites. The following discussions will deal primarily with the three detailed study areas, with appropriate observations from outside areas as well.

**Partially Recrystallized Assemblage**

Outcrops of this assemblage form conspicuous knobs and ridges. The rocks often have knobby weathering surfaces due to the orthopyroxene megacrysts which stand out in relief. The rocks are red-brown, tan-gray and green-black on weathered surfaces and gray-green, green, to green-black on fresh surfaces. Texturally, in handspecimen, the rocks contain coarse-grained orthopyroxene, sometimes as large as eight centimeters across, set in a much finer-grained matrix of amphibole, orthopyroxene, olivine, and spinel.

Table 1 contains a typical modal analysis (volume percent) for this assemblage and also shows the variation in mineral percentages.
Optically negative hypersthene is the predominant variety of orthopyroxene. Enstatite, optically positive, does occur in some localities. Both varieties can occur singly or together. In some samples (e.g. ND-77-32) hypersthene has well-developed characteristic pink-green pleochroism. Heinrich (1963) noted the unusual occurrence of this same type of pleochroism in enstatite from this area. Birefringence is generally low, with first or second order colors, and extinction is parallel along all [001] zone sections. Olivine is colorless in thin section and is characterized by moderate to high positive relief and birefringence. The 2V values for 75 grains in both this assemblage and the recrystallized assemblage were measured on a universal stage using the Berek's extinction method (Emmons, 1943). The 2V values for olivine ranged from 83-90° suggesting the olivine was a high-Mg forsterite. Such high Mg olivines are characteristic of ultramafic tectonites in Phanerazoic terranes (e.g. Ehrenberg, 1975). Two varieties of amphibole occur in this assemblage, a Ca-rich amphibole of the actinolite/tremolite series and a Ca-poor

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Sample ND-77-112</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Relict Orthopyroxene</td>
<td>50%</td>
<td>20-95%</td>
</tr>
<tr>
<td>Recrystallized Orthopyroxene</td>
<td>6%</td>
<td>0-40%</td>
</tr>
<tr>
<td>Olivine</td>
<td>10%</td>
<td>0-20%</td>
</tr>
<tr>
<td>Spinel</td>
<td>4%</td>
<td>0-10%</td>
</tr>
<tr>
<td>Actinolite/Tremolite</td>
<td>30%</td>
<td>0-52%</td>
</tr>
<tr>
<td>Anthophyllite/Cummingtonite</td>
<td>0%</td>
<td>0-45%</td>
</tr>
<tr>
<td>Serpentine</td>
<td>0%</td>
<td>0-25%</td>
</tr>
</tbody>
</table>
amphibole, either cummingtonite or anthophyllite. The Ca-rich amphiboles are colorless to slightly pleochroic green. They are optically negative and have moderate positive relief and birefringence. The values of 2V are high 80-85° and the ZAC ranges from 15-18°. Anthophyllite is colorless and has moderate negative relief and birefringence, and parallel extinction in all [001] zone sections. Cummingtonite is optically positive, colorless to slightly pleochroic green, and has inclined extinction. The 2V values are high 80-85° and the ZAC ranges from 17-20°. The values of 2V and ZAC agree well with the determinative curves for the cummingtonite-grunerite series given by Deer, et al. (1966, p. 161) and Tröger (1971, p. 92).

Spinel varies in color from dark brown to olive green to deep green with olive green the predominant color. Color is uniform throughout a thin section, however in some cases the spinel is crudely zoned with dark green cores and lighter green rims. Magnetite is a common accessory mineral. Apatite and sphene are occasional accessories.

Textures - The rocks of this assemblage characteristically have a granoblastic, polygonal, inequigranular metamorphic texture. The ultramafic rocks consist of large (generally ≤ 4 mm) anhedral poikiloblastic orthopyroxene megacrysts set in a finer-grained (generally ≤ 0.5 mm) recrystallized matrix of orthopyroxene, olivine, and amphibole. The megacrysts are always orthopyroxene and they show abundant strain features such as undulatory extinction, kink bands, and microfractures.
The recrystallized matrix minerals tend to have straight grain boundaries that meet in equilibrium angles of 120°. Such polygonal mosaic textures are characteristic of metamorphic recrystallization equilibrium assemblages (e.g. Ragan, 1969; Spry, 1975, p. 114-185; Vernon, 1970; Vernon, 1976, p. 135-146). Of the matrix minerals, olivine shows the least tendency to develop straight grain boundaries and generally forms subhedral to anhedral grains. Orthopyroxene and amphibole in the matrix form well-developed euhedral to subhedral crystals. Matrix minerals show little evidence of strain. The degree of recrystallization varies from 100% in the completely recrystallized assemblage to just a few percent in one of the partially recrystallized rocks. New grains form preferentially in areas of high stress (Moore, 1973; Spry, 1976), and samples with abundant relict orthopyroxene often show subgrain development and recrystallization along grain boundaries and kinks (Fig. 13). As the percent of recrystallized matrix increases relict orthopyroxene grains become isolated and obvious fragments of once continuous grains are evident (Fig. 14). Preferred orientation of both the matrix minerals and stretched relict orthopyroxenes increases with the amount of recrystallized matrix present. The orthopyroxene megacrysts contain abundant inclusions of olivine and amphibole. These inclusions are due to replacement after the formation of the megacrysts. The inclusions of olivine and amphibole have rational crystal shapes, and straight grain boundaries with few rounded or cuspate grain boundaries. Inclusions
Fig. 12. Line drawing of thin section showing deformation features in orthopyroxene megacrysts. Note deformation band defined by the differing extinction positions across megacryst, and abundant microfractures. 
H-Hypersthene
A-Actinolite/Tremolite
O-Olivine
OoagiiG-Spinel
Sample ND-77-137

Fig. 13. Line drawing of thin section showing finer-grained polygonal recrystallized orthopyroxene along grain boundary between two relict orthopyroxene megacrysts.
Sample ND-77-69

Fig. 15. Line drawing of thin section showing replacement inclusions in orthopyroxene megacrysts. 
H-Hypersthene
A-Actinolite/Tremolite
O-Olivine
OoagiiG-Spinel
Sample ND-77-132

Fig. 14. Line drawing of thin section showing fragments of once continuous orthopyroxene megacrysts set in a fine-grained recrystallized matrix.
Sample ND-77-132
are randomly oriented, strain free and have overgrown the microfractures and kinks of the deformed megacrysts (Fig. 15). Some relict orthopyroxene megacrysts also contain clinopyroxene exsolution lamellae. Table 2 summarizes the textural differences between the relict orthopyroxene megacrysts and the recrystallized matrix grains.

<table>
<thead>
<tr>
<th>Features</th>
<th>Relict</th>
<th>Recrystallized</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grain Size</td>
<td>Coarse Grained (&lt; 4 mm)</td>
<td>Finer Grained (&lt; 0.5 mm)</td>
</tr>
<tr>
<td>Grain Boundaries</td>
<td>Irregularly Shaped</td>
<td>Straight Grain Boundaries</td>
</tr>
<tr>
<td></td>
<td>None</td>
<td>Well-Developed 120° Triple Points</td>
</tr>
<tr>
<td>Strain Evidence</td>
<td>Abundant Strain Features</td>
<td>Strain Free</td>
</tr>
<tr>
<td></td>
<td>Exsolution Lamellae</td>
<td>None</td>
</tr>
</tbody>
</table>

In some specimens bronze bastites of serpentine preferentially replaces orthopyroxene. Spinel forms small (generally ≤ 0.2 mm) anhedral-subhedral grains and occurs interstitially throughout the rock.

Secondary alteration occurs in the form of serpentine, actinolite, talc, and chlorite. Serpentine replaces olivine first then orthopyroxene and finally amphibole. Serpentine characteristically occurs along fractures and grain boundaries. In many cases serpentine, and sometimes talc veinlets, form parallel to and enhance the foliation.

**Recrystallized Assemblage**

The recrystallized assemblage occurs occasionally as discrete zones within an ultramafic body and more commonly as a shell around the body (see Fig. 3). Outcrops characteristically have better-developed foliations and the rock is brown-green to gray-green on
weathered surfaces and green-brown to green-black on fresh surfaces. The rock is characteristically finer-grained than the previous assemblage, with occasional orthopyroxene megacrysts, and does not form as bold an outcrop as the partially recrystallized assemblage does.

In thin section the rocks are fairly uniform in grain size (generally < 0.5 mm). Table 3 contains a typical modal analysis (volume percent) for this assemblage and also shows the variation in mineral percentages.

**TABLE 3**

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Sample ND-77-95B</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Recrystallized Orthopyroxene</td>
<td>50%</td>
<td>2-50%</td>
</tr>
<tr>
<td>Relict Orthopyroxene</td>
<td>0%</td>
<td>0-20%</td>
</tr>
<tr>
<td>Olivine</td>
<td>12%</td>
<td>0-30%</td>
</tr>
<tr>
<td>Spinel</td>
<td>5%</td>
<td>0-10%</td>
</tr>
<tr>
<td>Actinolite/Tremolite</td>
<td>30%</td>
<td>0-90%</td>
</tr>
<tr>
<td>Cummingtonite/Anthophyllite</td>
<td>0%</td>
<td>0-65%</td>
</tr>
<tr>
<td>Phlogopite</td>
<td>0%</td>
<td>0-55%</td>
</tr>
<tr>
<td>Serpentine</td>
<td>0%</td>
<td>0-60%</td>
</tr>
<tr>
<td>*Hornblende</td>
<td>0%</td>
<td>0-30%</td>
</tr>
<tr>
<td>*Diopside</td>
<td>0%</td>
<td>0-45%</td>
</tr>
</tbody>
</table>

*Found only in granulite assemblages (see below)*

Optically negative hypersthene is again the predominant variety of orthopyroxene, with subordinate amounts of enstatite. Hypersthene shows characteristic pink-green pleochroism in some samples, alternatively it is colorless or pink. Birefringence is low, first to low second order colors, and extinction is parallel in all [001]
sections in both hypersthene and enstatite. Olivine is colorless and has moderate to high positive relief and birefringence. Spinel is dominantly olive green with some dark brown-black to deep green varieties. Occasionally the spinel is zoned with dark cores and light green rims. In four specimens (ND-77-13, -46, -47-, -48) diopside was found in addition to the usual orthopyroxene. The diopside was colorless, had moderate to high positive relief and birefringence. The 2V and ZAC were approximately 55° and 41° respectively. Hornblende replaced the more common actinolite/tremolite Ca-amphibole in these specimens also. The hornblende was optically negative and pleochroic brown in color (with z = yellow brown, x = light yellow). It had moderate to high positive relief and birefringence, a 2V of about 85° and ZAC of approximately 25°. These rocks also contained small amounts of plagioclase and a trace of quartz.

In most rocks of this assemblage the most common amphibole was an optically negative amphibole of the actinolite/tremolite series. This amphibole occurs as colorless to slightly pleochroic green prismatic grains with moderate positive relief and birefringence, a 2V of 80-85° and a ZAC of 15-18°. Optically negative orthorhombic anthophyllite also occurs in these rocks. It usually forms colorless bladed crystals with moderate positive relief and birefringence. Cummingtonite, optically positive, is more common than anthophyllite and occurs as colorless to faintly pleochroic, gray-green, prismatic
grains with moderate positive relief and birefringence. It has inclined extinction with ZA of 17-19° and 2V values of 82-86°. These values again agree well with the determinative curves for the cummingtonite/grunerite series in Trüger (1971, p. 92) and Deer, et al. (1975, p. 151).

Phlogopite is occasionally found in these rocks. It characteristically has a very small 2V, parallel extinction on all [001] sections, bird's eye extinction, and orange to light yellow pleochroism. Titano-clinohumite was found in one specimen (ND-77-108) from Area III (see Fig. 10) of this assemblage. The deep red mineral in hand specimen formed subhedral to anhedral grains and grain clusters often associated with spinel. No evidence of disequilibrium was found in this specimen and it is assumed the titano-clinohumite is part of the stable metamorphic assemblage of orthopyroxene, actinolite/tremolite, spinel in this rock. Optically the clinohumite has a high positive 2V of approximately 70-75° and characteristic orange-yellow pleochroism. The mineral has moderate to high positive relief and high birefringence. Dispersion is strong with R>V and abnormal interference colors common. The strong dispersion and high birefringence suggest the mineral is the titanian variety. Heinrich (1963) also noted the occurrence of titano-clinohumite in Area I of this study and suggested it was an alteration mineral. The significance of titano-clinohumite in metamorphosed ultramafics is unclear and no interpretations are possible here with available evidence.
Magnetite is a common accessory in these rocks and forms either euhedral-subhedral crystals or fine dusty, feathery grains. Apatite is a rare accessory.

Textures. Texturally this assemblage is characterized by a well-developed equigranular, granoblastic, polygonal texture. This assemblage is gradational from the partially recrystallized assemblage and fragments of relict, strained orthopyroxene megacrysts persist in this assemblage. Subhedral olivine with euhedral to subhedral orthopyroxene and amphibole form beautiful polygonal, mosaic textures with a marked tendency to meet in 120° triple points (Fig. 16). Where amphibole is the dominant mineral, more decussate-like textures prevail (Fig. 17).

Spinel forms interstitial subhedral to anhedral rounded grains. Occasionally spinel and olivine may concentrate in vague layers parallel to the foliation. Preferred orientation is well-developed in many of these rocks and is enhanced by the alignment of elongate amphibole and phlogopite grains. Orthopyroxene is sometimes elongate in the foliation. The minerals of this assemblage are relatively strain-free suggesting syn-tectonic or post-tectonic growth. Inclusions of dusty or feathery magnetite are commonly found in the minerals of this assemblage. They give the thin section a peppered appearance. Often the dusty inclusions are concentrated at the core of the minerals (Fig. 18). The dusty magnetite also forms bands with distinct shapes and orientations. These bands transgress grain boundaries and are
Fig. 16. Line drawing of thin section showing granoblastic polygonal texture in recrystallized assemblage. Also shows well-developed 120° triple points.

H-Hypersthene
A-Actinolite/Tremolite
O-Olivine
Opaque-Spinel
Sample ND-77-31

Fig. 17. Line drawing of thin section showing decussate texture.

H-Hypersthene
A-Actinolite/Tremolite
O-Olivine
Opaque-Spinel
Sample ND-77-25

Fig. 18. Line drawing of thin section showing inclusions of dusty magnetite at the core of recrystallized minerals.

H-Hypersthene
A-Actinolite/Tremolite
O-Olivine
Opaque-Magnetite some S-Spinel
Sample ND-77-105

Fig. 19. Line drawing of thin section showing band of dusty magnetite-opagues transgressing H-hypersthene, A-actinolite/tremolite, O-olivine grain boundaries.

Sample ND-77-94.
apparently relict features which the recrystallized assemblage included (fig. 19).

Dusty and feathery magnetite are common by-products of the serpentinization process. Besides being randomly distributed throughout a serpentinite, the dusty magnetite often forms and is oriented along cleavages and grain boundaries of the serpentinized minerals (e.g. Coats, 1963). The inert magnetite, upon metamorphism of the serpentinite, may remain as a relict microfabric. Tiny magnetite grains may have acted as impurities and provided nucleation sites for the growth of some of the metamorphic minerals, with the magnetite preserved in the cores of the minerals. I have interpreted these textures as relict textures formed during a previous period of serpentinization prior to metamorphism. This period of serpentinization proceeded and is unrelated to the serpentinization now evident in the ultramafic rocks.

In the two-pyroxene (orthopyroxene and clinopyroxene) rocks of this assemblage, trace amounts of fine-grained plagioclase and quartz often occur at triple point intersections (Fig. 20). The average grain size of these rocks is approximately 1 mm.

Serpentine again forms the main secondary alteration. Chlorite and actinolite are also common alteration minerals. Serpentine veins often form along and enhance the foliation. Bronze-colored bastite preferentially alters orthopyroxene grains leaving the amphiboles unaltered in some specimens. Pale green chlorite occurs
Fig. 20. Line drawing of thin section showing irregularly-shaped quartz and plagioclase at triple point intersection. Opx-hypersthenes, D-diopside, H-hornblende, P-plagioclase, Q-quartz, Ph-phlogopite, Opague-magnetite. Sample ND-77-47.

Fig. 22. Line drawing of thin section showing well-developed polygonal, mosaic texture with 120° triple points in monomineralic hornblende diffusion rock. Sample ND-77-43.
interstitially and alters mainly phlogopite and amphibole. Very fine-grained actinolite replaces serpentine in some places. Minor alterations in the form of calcite and epidote group minerals occur locally. An increase in amphiboles, mainly actinolite/tremolite and anthophyllite, marks the gradational contact between the recrystallized assemblage and the next major rock group.

**Diffusion Reaction Zones**

The emplacement of ultramafic rocks into rocks of more siliceous composition produces an environment favorable for the development of metasomatic diffusion zones. In a given temperature and pressure environment the contrasting rock compositions allow reactions to take place producing a number of sharp monomineralic or biminarlic zones from an original multi-phase boundary. These monomineralic zones have been described in a number of studies (e.g. Chidester, 1962; Curtiss and Brown, 1969; Carswell et al., 1974). The ultramafic bodies in the Ruby Range are no exception and are commonly rimmed by such zones. The zones are generally less than a meter in width. However, because of tight isoclinal refolding and trasposition, original thicknesses are difficult to estimate in some cases. This deformation blurred the original concentric pattern of the zones.

The mineralogy of these rocks is inherently simple, generally consisting of only one or two minerals. These zones form along the ultramafic contact. Toward the center of the ultramafic bodies they
grade into the recrystallized assemblage (Fig. 21). Locally, some or all of the zones are missing. The chemistry of the country rock, subsequent deformation, and exhaustion of one or the other end of the diffusion couple may have resulted in the missing zones (Evans, 1977).

Grading from the recrystallized assemblages, the innermost zone, is most commonly composed of anthophyllite with minor amounts of actinolite or cummingtonite and rarely quartz. The anthophyllite rocks are well-foliated because of the strong preferred orientation of the elongate grains. An actinolite rock with minor amounts of anthophyllite or a cummingtonite rock with minor anthophyllite sometimes occurs in this zone. Both of these have strong preferred orientation of grains producing well-developed foliations. In one instance (specimen ND-77-145A) a complexly folded rock consisted of alternating layers of 100% anthophyllite and cummingtonite only 3-5 cm thick. The second zone outward from this is composed of coarse (generally ≤ 2 mm) hornblende or gedrite. The optically positive orthorhombic gedrite is pleochroic light gray to light brown. The hornblende is optically negative, has a ZAC of 18-23°, and green to yellow-green pleochroism. The rocks look identical in hand specimen. They are distinctive, black, moderately foliated rocks with some grains displaying a brilliant blue schiller. Heinrich (1963) attributed the blue schiller to thin plates of a dark brown mineral arranged parallel with the c-axis of the hornblende. I could not find the brown mineral
Fig. 21 Schematic field sketch of diffusion zones
in any of the thin sections I examined. However, an impurity of some sort probably accounts for the unusual schiller effect. These rocks have beautiful, polygonal, mosaic textures in thin section with a strong tendency for grain boundaries to meet at 120° (Fig. 22). An occasional almandine garnet is found in these rocks. This zone commonly grades into an amphibolite or hornblende gneiss or in some cases a third biotite-rich zone is present with minor gedrite or hornblende.

This biotite zone generally grades into country rock, either a hornblende or biotite gneiss or a more felsic gneiss. The biotite is coarse-grained (generally < 2 mm) and has strong preferred orientation giving the rock a pronounced schistosity. The mica has a characteristic small 2V and parallel extinction, with deep red to brown pleochroism. The biotite weathers to vermiculite as determined by X-ray diffraction analysis.

A generalized sequence of these monomineralic diffusion zones from ultramafic rock outward consists of:

(1) anthophyllite zone or (occasionally actinolite zones or cummingtonite zones, sequential relationships between these three are unclear)

(2) hornblende or gedrite zone

(3) biotite zone

The migration of SiO₂ and MgO as well as CaO, FeO, and K₂O down chemical potential gradients produces the metasomatic zones along
ultramafic contacts (Brady, 1977). That is, juxtaposition of a Mg-rich, silica-poor serpentinized ultramafic (orthopyroxenite or olivine orthopyroxenite) rock into country rocks such as quartzofeldspathic or hornblende gneisses containing more Si, K, Fe, and Ca would create a favorable situation for the migration of elements down these compositional gradients during metamorphism. Si, Ca, Fe, and K move toward the ultramafic and Mg away from it. The activity of water would probably be constant throughout this system, as the breakdown of serpentine in the ultramafic rock and other hydrous minerals in the wall rocks would create a water saturated environment during metamorphism. The formation of hydrous phases in the diffusion zones indirectly supports this contention.

Compositionally different wall rocks probably account for the presence or absence of certain zones. For example, abundant CaO in the wall rock would favor the formation of actinolite or hornblende versus anthophyllite or gedrite (e.g. along amphibolite, hornblende-plagioclase gneiss contacts). A low K2O amphibolite would not favor the formation of a biotite zone. The absence of this zone along amphibolite contacts supports this conclusion. Unclear relationships between the zones in places prohibits any definite statements on the relative distance of migration of the various components. However, in a general sense it appears that Fe and Ca migrated furthest toward the ultramafic and K and Al stayed close to the country rocks. Aluminous phases (biotite, hornblende, gedrite) found directly at
wall rock contacts is consistent with the apparent low mobility of Al$_2$O$_3$. MgO derived from the ultramafic was quite mobile and is found throughout the zones.

Three separate outcrops in Area I contained an unusual clinopyroxene rock. The rocks were dense, massive to weakly foliated and gray in color. The foliation was defined by the parallel alignment of mica grains. They occurred as dike-like bodies or pods approximately 10-20 meters in length near contacts between country rock and the more typical ultramafic bodies.

In thin section the rocks were composed primarily of clinopyroxene, actinolite/tremolite and spinel with some phlogopite, magnetite, and apatite. Alterations include serpentine, chlorite, and clinzoisite. Sample ND-77-113 was composed entirely of clinopyroxene and spinel, with minor magnetite. The clinopyroxene is unusual with its slight pink color, and strong optic axis dispersion producing anomalous interference colors. The 2V was approximately 45$^\circ$ and the ZAC was approximately 42$^\circ$. X-ray diffraction analysis indicated the mineral was a diopsidic-augite. The strong optic axis dispersion and pink color suggests it is titanium-rich. The spinel (confirmed by infrared spectrophotometry) was dark blue-green in color and was coarser-grained than the olive-green spinel found in the other ultramafic rocks of the area.

Texturally, these rocks vary from a non-directional interlocking igneous texture, with gently curving grain boundaries, to a weakly
foliated, mosaic textured rock. The foliation is defined by the parallel alignment of phlogopite grains and, in some cases, spinel.

The origin of these rocks is unclear. They may be igneous in nature, intruded after metamorphism as dike-like bodies. In this instance some sort of contact effects may form. However, if present, I overlooked them in the field. The calcium-rich mineralogy (e.g. diopsidic-augite, actinolite/tremolite, and apatite) is reminiscent of rodingites (see below). One possibility is that these bodies were metarodingites formed during the initial serpentinization of the ultramafic bodies and metamorphosed along with them. Rodingite often occurs as dike-like or pod-like bodies at the contact between serpentinite and country rock. The occurrence of these bodies at the margins of the more typical ultramafic rocks near the country rock contacts is consistent with such an interpretation. Metasomatism during metamorphism may have converted the more typical rodingite mineralogy (see below) into rocks composed simply of clinopyroxene, Ca-amphibole and spinel. This interpretation was made for rocks of similar composition and setting in the Alps (Evans and Trommsdorff, 1977).

Serpentinites

Varying degrees of serpentinization have affected many of the ultramafic rocks in the Ruby Range. Where serpentinization proceeded to an advanced state, and serpentine makes up greater than 50% of the rock, the rock is called a serpentinite. Among the areas studied
in detail, Area II had the most serpentinites and Area III contained none. Within a specific area there is no apparent distribution pattern of serpentinite to unserpentinized ultramafic rock. Megascopically, the serpentinites vary considerably, ranging in color from light gray, to gray-green, to greenish-black to black.

Texturally, some are massive while others have a well-developed foliation. In most cases the rocks are very fine-grained making hand specimen identification of the minerals impossible. However, from one locality in Area I, hand specimen identification of cross-fiber asbestos chrysotile veins was possible. Where sufficient orthopyroxene relicts remain, the rock retains the knobby, weathered appearance of the partially recrystallized assemblage.

The primary serpentine mineral in these rocks was identified as antigorite by Heinrich (1963). The mineral is colorless to yellow-green to tan in plane polarized light. Birefringence is low from first order gray to yellow. Most of the serpentine has wavy extinction and variable refractive indexes of n=1.555, 1.557, 1.582, 1.595 (Heinrich, 1963).

In thin section the serpentinites have a wide variety of textures. The dominant texture is that of a fibrolamellar matte of randomly oriented antigorite flakes. In some cases the flakes form radial patterns and in others they show a definite preferred orientation. These textures are common in antigorite serpentinites and, in the latter case, the serpentine is apparently pseudomorphically replacing
pyroxene or amphibole (e.g. Coats, 1968; Basta and Kader, 1969). Magnetite occurs in all of the serpentinites and makes up between 5-10% of the rock. It forms in three distinct crystal habits, one as euhedral smooth-sided grains, another as ragged feathery grains, and finally as very fine grained dusty magnetite (Fig. 23). The randomly distributed euhedral magnetite may in some cases be a relict from the original rock. The dusty and feathery magnetite characteristically forms along grain boundaries cleavages and fractures of the original minerals. This situation enhances the pseudomorphbic nature of the serpentinization by preserving original textures and crystal shapes (Fig. 24).

Some varieties of serpentine have a mesh-like texture composed of a rectangular, polygonal arrangement of cross-fiber veinlets. The fibers are generally arranged normal to the vein margins. Pseudoporphic textures and boundaries are recognizable because of dusty and feathery magnetite encircling the mesh and of changes in the orientation of the mesh. Small relict grains of pyroxene and amphibole are often found in the cores of the mesh. In other cases the cores contain lower birefringent to nearly isotropic fibrous or platy serpentine (Fig. 25).

A third general texture is that of closely packed parallel cross-fiber veinlets. There is apparently a gradational transition between this texture and mesh texture. Coats (1968), in his study of Manitoba serpentinites, noted this same relationship. He also noted
Fig. 23. Line drawing of thin section showing dusty and feathery magnetite in serpentinite. Feathery magnetite appears to form along relict cleavage. Sample ND-77-59.

Fig. 24. Line drawing of thin section showing serpentine preserving original textures. Note dusty and feathery magnetite forming along laminae of magnetite occurring along cleavage traces of chlorite-c. Sample ND-77-131

Fig. 25. Line drawing of thin section showing mesh texture serpentine. Sample ND-77-310.

Fig. 26. Line drawing of thin section showing pseudo-breccia texture in serpentinite. Opx-Orthopyroxene A-Actinolite/Tremolite Opagues-Magnetite Sample ND-77-35.
that parallelism of veinlets in a number of adjacent pseudomorphs may indicate a preferred orientation of the original minerals before serpentinization. Parallel veined serpentinites in the Ruby Range also show this feature.

Where relict grains persist in sufficient numbers, a distinctive pseudobreccia texture develops. The relict grains are surrounded and cut by a complex pattern of serpentine veinlets producing irregular coarse relics set in a fine-grained serpentine matrix (Fig. 26). No olivine relics were found and the most persistent mineral was amphibole. The order of preference for serpentinization of these rocks is apparently olivine, then pyroxene, then spinel and finally amphibole.

Chlorite occurs in all of the serpentinites examined and makes up between 1-8% of the rock. It usually forms as colorless irregularly bounded flakes up to 2 mm in size. Undulose extinction and bent grains indicate deformation affected some of the crystals. Thin laminae of magnetite are invariably present along the cleavage traces of the chlorite (see Fig. 24). Oxidation of original phlogopite or biotite during the formation of the chlorite may have produced this situation as suggested by Root (1965) for serpentinites in the Pony-Sappington area of southwestern Montana.

The massive serpentinites are often crosscut by cross-fiber chrysotile veinlets and more massive veins of serpentine with magnetite ladders (Heinrich, 1963). Minor amounts of talc are also
present in the serpentinites. Garihan (1973) noted that the talc veinlets also crosscut the massive serpentine.

Veined cross-fiber chrysotile and talc, therefore, apparently postdate the more massive antigorite serpentinization. However, because of the gradational aspect of many of the serpentine textures and the lack of a detailed analysis of the serpentine mineralization strict time relationships for different periods of serpentinization are difficult to define.

Serpentinization

Serpentinites result from the low temperature reaction of ultramafic rocks with externally introduced waters to produce serpentine minerals. In Areas I and II the large bodies of serpentinite and degree of serpentinization are spatially related in a broad sense, to the two large faults bordering the areas mapped by Heinrich (1963) and Okuma (1971). These faults may have provided a conduit for the waters needed in serpentinization. On a finer scale, many outcrops are serpentinized along distinct planes. These planes of serpentinization usually parallel the foliation in the surrounding rocks. Microscopically, the long axes of amphibole prisms are often aligned parallel to and enhancing the planes of serpentinization giving the rock a layered appearance. Original planes of weakness (e.g. foliation) in the ultramafic rocks would provide access to water and become a site of preferential serpentinization.
A number of controversies have arisen over the years as to whether serpentinization is a constant volume process. Those who support constant volume and thus removal of $\text{SiO}_2$ and $\text{MgO}$ from the system cite retention of preexisting textures, lack of detectable expansion features and presence of "rodingite" aureoles (see below) as evidence for constant volume serpentinization (Thayer, 1966). Other authors have countered by citing evidence of all scales of expansion features such as kink bands offset along transecting serpentine veinlets, rotated fragments of larger olivine grains enclosed by serpentine, expanded chromite grains cut by serpentine-filled fractures (see Clarke and Greenwood, 1972). For the individual cases the evidence on both sides may seem conclusive. Coleman (1971) pointed out that both constant volume and volume expansion processes may operate to form serpentinite depending on whether or not serpentinization took place under static conditions. Static conditions would produce a closed system and favor volume expansion whereas a high stress environment with associated shearing would produce an open system and favor constant volume serpentinization.

None of the serpentinites of the Ruby Range are highly sheared, and textural evidence suggests volume expansion serpentinization. Microscopic textural evidence for expansion include 1) dilational serpentine veins cutting and offsetting orthopyroxene and 2) expanded spinel grains cut and rimmed by serpentine (Fig. 27). In a nearby area, Root (1965) attributed small, irregular open folds and
Fig. 27. Line drawing of thin section showing expanded spinel grains cut and rimmed by serpentine.
Opx-Orthopyroxene
A-Actinolite/Tremolite
S-Spinel
Opaque-Magnetite
Sample ND-77-126.
crenulations often found locally in the serpentinites to volume expansion upon serpentinization.

Recent studies on stable isotope ratios, specifically $^{18}O/^{16}O$ and $D/H$, provide some estimates as to the temperature and pressure conditions of serpentinization (Wenner and Taylor, 1971, 1974). The results indicate a distinction between lizardite-chrysotile serpentinization and antigorite serpentinization. Lizardite-chrysotile serpentinite apparently forms from meteoric waters in a low temperature (85-115°C) shallow environment and antigorite serpentine forms from connate (metamorphic) waters in a higher temperature (220-460°C), deeper seated environment. From such studies it would seem that the crosscutting, cross-fiber chrysotile may indicate a change in environment of serpentinization for the Ruby Range serpentinites with the antigorite forming in a higher temperature, deeper environment than the chrysotile.

**Low Temperature Metasomatic Rocks**

Often associated with serpentinites are distinctive rock types formed by mass transfer of various constituents during serpentinization (e.g. Coleman, 1967). The most common alteration rock found is called rodingite. Calcium-metasomatism accompanying serpentinization produces these light colored, dense, fine-grained rocks rich in calcium silicates such as hydrogrossular, diopside, idocrase, amphibole, zoisite, prehnite, xenolite. In a few localities rodingites are associated with the serpentinites in the Ruby Range.
The southwestern corner of Area I provides the best examples of these rocks. The light gray rocks form small lenses less than five meters across lying within or directly adjacent to massive serpentinite (Fig. 28). In one instance, the rodingite was separated from the serpentinite by a black chlorite-serpentinite layer approximately 20 cm thick. This layer was similar to "blackwall" rocks described by Chidester (1962) and Coleman (1967). The serpentinite body associated with these rocks is a large massive antigorite body with numerous cross-fiber asbestos chrysotile veinlets cutting through it.

Mineralogically, the rodingites are composed of a variety of calc-silicate minerals, most notably clinozoisite, epidote, diopside, grossular, vesuvianite, and actinolite. Calcite and chlorite are present in most of the rocks also. The rocks are fine-grained (generally \( \leq 0.3 \) mm). Textures are complex with very irregularly shaped grains. Grossular is typically slightly anisotropic and exhibits a peculiar sector twinning. It sometimes occurs as clusters of six-sided grains set in a mosaic pattern (Fig. 29).

Where Ca-metasomatism is incomplete relict metamorphic textures and grains are present. The original metamorphic rocks were apparently mafic amphibolites or hornblende gneisses. Plagioclase is affected first and alters to sericite and clinozoisite. Biotite alters to chlorite, and amphibole to clinozoisite.

Sample NO-77-80 provides a stunning example of this Ca-metasomatism (Fig. 30). In this specimen clinozoisite pseudomorphically replaces
Fig. 28. Schematic Field Sketch of Rodingites.
Fig. 29. Line drawing of thin section showing mosaic pattern of grossular garnet-G. A-Actinolite/Tramolite. Sample ND-77-52.

Fig. 30. Line drawing of thin section showing Ca-metasomatism. M-Muscovite, CL-Clinozoisite, C-B-Chlorite-Biotite Sample ND-77-80.
amphibole, preserving the amphibole cleavage and skeletal grain boundaries.

The Ca-metasomatism apparently affected the ultramafic rocks as well. Grossularite and fine-grained (< 0.1 mm) needles and prisms of actinolite replace serpentine in some specimens. In specimen ND-77-19 textural evidence shows two periods of Ca-amphibole growth. In this rock, thin talc veinlets cut actinolite/tremolite of the high-grade metamorphic assemblage and finer-grained actinolite replaces the talc veinlets. These two specimens were not collected in the area containing the rodingites and associated serpentinitites. The absence of serpentinite and other evidence of Ca-metasomatism in these areas suggests that the formation of some secondary actinolite in the ultramafic rocks may have resulted from retrograde metamorphism not directly related to the serpentinization process and associated Ca-metasomatism.

Barnes and O'Neil (1969, 1972) conclusively showed that modern serpentinization can produce Ca-rich fluids capable of producing low temperature, calcium-rich reaction zones. Circulation during active serpentinization of high pH, calcium-hydroxide water, undersaturated with respect to minerals in the surrounding rocks, results in the metasomatic rodingites (Barnes and O'Neil, 1969). Abundant evidence for Ca-rich fluids in the form of calcite-garnet, calcite-calc-silicate, and garnet veins cutting serpentinite and other ultramafic rocks, occur in this area.
Rodingitization is considered a low temperature process. At approximately 550°C and an intermediate pressure of 5 Kbar, serpentine reacts to form forsterite, talc, and H₂O. Pressure has little effect on this reaction (see curve 1, Fig. 33). Therefore the presence of serpentine in the serpentinites and in the serpentine chlorite "blackwall" rock provides an upper temperature limit for rodingitization in this area. Assuming that antigorite is the dominant serpentine phase, temperatures probably exceeded those required to revert chrysotile to antigorite, approximately 300°C at 5 Kbar (Evans, 1977). Serpentinization and associated rodingitization can also occur at surface temperatures and pressures (Barnes and O'Neill, 1964; Cashman and Whetten, 1976), and serpentinization and rodingitization may be ongoing processes continuing up to the present in the ultramafic rocks of this area.

The petrology and textures of the ultramafic and associated rocks presented in this section indicate a distinct period of recrystallization after formation of the orthopyroxene megacrysts. Textures also indicate that before recrystallization the rocks were serpentinized. Field, petrologic and textural relationships indicate that the diffusion zones and the recrystallized assemblages formed during the metamorphic event of the area. After this period of recrystallization reserpentinization with associated low-temperature, Ca-metasomatism occurred.
Petrofabric analysis was carried out on two rocks to determine the preferred orientation of relict orthopyroxene and recrystallized orthopyroxene and olivine, in order to compare the fabrics in the relict and recrystallized grains.

Measurements of the optical axes of olivine grains and orthopyroxene grains was done on a four-axis Zeiss universal stage. The results of this preliminary study are graphically shown in Figure 31 and are discussed below. The olivine orientations were taken from sample ND-77-39 collected from the Mormon Creek body near the Ruby Reservoir (see Fig. 2). The foliation in the body was concordant with that of the host rocks and trended approximately N62E/60NW. The orthopyroxene measurements were from sample ND-77-108 collected in Study Area III. The foliation in the ultramafic body and the host rocks was approximately N72E/70NW. The axial planes of the tight isoclinal folds in this area are roughly parallel to this foliation.

The olivine fabric is rather strong and is distinctly different from most olivine fabrics described in the literature (e.g. Avé Lallement and Carter, 1970; Lappin, 1971; Nicolas et al., 1971). The main features of the fabric are a [100] maximum normal to the foliation with [010] and [001] maxima lying within the foliation. The three strong
Fig. 31 Contoured petrofabric diagrams of indicatrix axes for 75 olivine (Ol) grains, 75 orthopyroxene (Opx) grains, and 50 relict orthopyroxene (R-Opx) grains. Contour intervals are 1.3%, 5%, 10%, 20% for Ol; 1%, 6%, 12%, 18% for Opx and 3%, 12%, 15% for R-Opx, all per 1% area. F—plane of foliation.
maxima would seem to indicate that the principle stresses involved in forming the fabric were unequal (Carter, et al., 1972). This fabric most resembles the olivine fabrics found in the garnet peridotite at Alpe Arami, Switzerland (Möckel, 1969). Möckel (1969) recognized that such fabrics have "no resemblance to previously published fabrics of metamorphic origin." The fabric, furthermore, does not resemble any typical igneous fabrics (e.g. Lappin, 1971). Hartmann and den Tex (1964) contend that recrystallization of olivine in the absence of intergranular fluid could produce such a fabric. The granoblastic polygonal texture of sample ND-77-39 suggests recrystallization, however the presence of actinolite as a part of the assemblage would also seem to indicate the presence of interstitial fluids. Carter et al. (1972), in reference to their theories on syntectonic recrystallization for the development of olivine fabrics, state that "the origin of the [100] maximum (normal to foliation) fabrics is not yet understood, but such fabrics are not common."

The orthopyroxene fabrics are weaker and even more atypical. The fabric of the recrystallized orthopyroxenes is characterized by a strong [100] maximum in the foliation and two weaker [010] and [001] maxima parallel to and perpendicular to the foliation respectively. No similar fabrics were found in the literature and the significance of such fabrics is unclear. The relict orthopyroxene grains have an unusual fabric as well with [100] and [010] lying in weak girdles at angles to the foliation. [001] has no distinct pattern and is best
characterized as random. The distinctly different fabrics between the relict orthopyroxene grains and the recrystallized grains supports the distinction between, and the differing origins of, the two types of orthopyroxene. The fabric of the relict orthopyroxene grains may be partly inherited before metamorphism and recrystallization of these bodies. However, the significance of this texture is unclear.

Further studies of the fabrics of these ultramafic bodies should prove interesting, and are needed to test the validity and significance of the fabrics found in this preliminary study.
Pre-Metamorphic History

The pre-metamorphic history of these ultramafic rocks is unclear because intense recrystallization and deformation during the upper amphibolite metamorphic event obscured much of the past histories of these rocks. The limited relict grains and textures provide but a very small glimpse into the ancient histories of these rocks.

Orthopyroxene megacrysts are the only readily recognizable relict grains. The abundant disequilibrium deformation features, different fabric orientations, and other textural differences between the relict orthopyroxene megacrysts and the recrystallized grains indicates the relict megacrysts formed in an earlier and different environment. What this environment was and what processes were involved is highly questionable.

In some samples, relict orthopyroxene megacrysts make up 80% of the rock. This suggests that at least some of these rocks were originally orthopyroxenites or olivine orthopyroxenites. Ultramafic rocks of such composition are reminiscent of the orthopyroxenites found in stratiform complexes. However, readjustment of relict orthopyroxene grain boundaries and recrystallization during deformation obliterated any distinctive cumulate texture that may have been present. No compositional layering characteristic of stratiform complexes was observed.
Subsequent recrystallization and deformation may have removed such features as well. A wide range of composition in the relict orthopyroxene may represent differentiation trends formed during the crystallization of a stratiform complex. However such evidence awaits detailed chemical analysis of these rocks. Other igneous hypotheses are possible and equally difficult to evaluate.

The orthopyroxene megacrysts may even be metamorphic. Recrystallization and plastic deformation often accompanies upward movement of ultramafic bodies in the mantle (e.g. Loney et al., 1971; Medaris, 1972). Therefore the orthopyroxene megacrysts need not represent the original magmatic or other primary crystallization textures. The initial fabric found in the relict orthopyroxene megacrysts may have developed from recrystallization and plastic deformation in the mantle. Whether the original ultramafic rocks formed in a mantle or crustal environment under igneous or metamorphic conditions is unclear.

Relict textures, mineral assemblages, and field relationships, however, indicate that subsequent to this initial stage of crystallization the rocks underwent a period of serpentinization. I believe the inclusions of dusty and feathery magnetite in the recrystallized assemblage are relict textures formed during this period of serpentinization. Dusty and feathery magnetite is a common by-product of the serpentinization process and often forms along the cleavage and grain boundaries of the serpentinized minerals. The included dusty magnetite in the recrystallized assemblage often occurs as distinct patterns and
bands (see Fig. 19), and may represent portions of the original serpentinite textures. Such textures have been interpreted similarly elsewhere (Trommsdorff and Evans, 1974; Vance and Dungan, 1977). Serpentinization of ultramafic bodies preferentially occurs at the margins and along planes of weakness in ultramafic rocks such as joints and fractures (Coleman and Keith, 1971; Morit and Bannos, 1973; Ehrenberg, 1975). If I assume the recrystallized assemblage developed from dehydration and recrystallization of serpentinite, and note that most of these bodies were probably only partially serpentinized as they contain relict orthopyroxene grains, I might expect that serpentinization first occurred at the margins and along planes of weakness within the bodies. Therefore, upon metamorphism and recrystallization, the recrystallized assemblage would occur along the margins and as distinct planes within the ultramafic bodies. This is the case in the Ruby Range, with the recrystallized assemblage having a distinct spatial relationship with the partially recrystallized assemblage and occurring along the margins and interlayered with the partially recrystallized assemblage in the ultramafic bodies. Localities where only the recrystallized assemblage is present were probably completely serpentinized. The mineralogy of the recrystallized assemblage is also consistent with its having developed from the metamorphism of a serpentinite (see discussion below).

Serpentine cannot exist much above 500°C, even at very high pressures (e.g. Bowen and Tuttle, 1949; Johannes, 1963, 1969; Evans
et al., 1976). Crystallization temperatures of stratiform complexes, ultramafic magmas, and temperatures within the mantle are certainly much higher than the above temperature figure (e.g. Hyndman, 1972, p. 110). If these bodies were serpentinized prior to metamorphism, then the ultramafic bodies existed in an appropriate low temperature environment for serpentinization between primary crystallization and metamorphism. If the ultramafics were originally emplaced as magmas into the crust, serpentinization may have occurred in situ after cooling if emplacement was shallow enough and water was available. If the ultramafic bodies were mantle derivatives, serpentinization may have occurred at ridge crests when upwelling mantle material meets circulating seawater, along transform faults, or during tectonic emplacement of the cold mantle slab into the crust along continental margins.

Other environments for serpentinization are possible. However, available evidence prohibits making any conclusions about any particular environment. The important point is that sometime during the Precambrian, prior to their metamorphism and recrystallization, conditions were favorable for the serpentinization of these bodies. Whether serpentinization occurred during the early stages of tectonism and metamorphism, or whether it occurred at a much earlier time is not really critical. After this period, conditions changed, serpentine became unstable, and recrystallization began.
Emplacement and Metamorphism

The style of deformation and grade metamorphism is fairly uniform throughout the Ruby Range (Okuma, 1971; Garihan, 1973). Okuma (1971) found at least three phases of deformation in this area. The earliest phases $F_1$ and $F_2$ are characterized by northeast trending isoclinal to concentric folds. $F_3$ folds are broad, open folds trending to the north. Concordant foliations and interlayering of the ultramafic rocks with their host rocks throughout such structures suggests early emplacement for the ultramafic bodies and participation in the deformation.

I believe the association of some of the ultramafic bodies at or around the crests of isoclinal ($F_1$-$F_2$) folds is more than a chance occurrence. The stresses responsible for the folding may have tectonically disrupted the ultramafic bodies, squeezing them and providing a mechanism of movement for the ultramafic pods along the folds.

Movement of the ultramafic pods along foliation in plastic country rocks would produce the concordant attitude of the ultramafic bodies with respect to the foliation in the host rocks. Movement of ultramafic pods into and through the crests of folds may produce the discordant relationships once attributed to igneous intrusion. Emplacement of the ultramafic bodies as solid masses during deformation could account for the lenticular pod-like shapes of many of the bodies, and the pinched off ends of some (see Fig. 3). These are common outcrop
patterns for solidly-emplaced ultramafic tectonites in metamorphic terranes (e.g. Skinner, 1968; Misra and Keller, 1978).

Serpentinization of the bodies even if only along the margins would have aided initial emplacement and mobility by lowering the density and providing an easily sheared surface. As grade of metamorphism increased, recrystallization of the serpentinite margins during deformation produced the well-developed concordant foliations in the recrystallized assemblage. The partially serpentinized interiors of the bodies, as exemplified by the partially recrystallized assemblage containing relict orthopyroxene megacrysts, probably would have responded somewhat differently to this intense deformation. These more competent areas would presumably deform in a less plastic manner, as evidenced by the boudinage-like structure in Area III (see Fig. 11), and the abundant deformation features in the relict orthopyroxene megacrysts.

Metamorphism during this deformation reached the upper amphibolite facies as indicated by diagnostic assemblages in the surrounding rocks. Sillimanite-muscovite to sillimanite-orthoclase zone metamorphism, as indicated by assemblages of: sillimanite-microcline-plagioclase-quartz-biotite ± garnet and sillimanite-orthoclase-plagioclase-quartz-biotite ± garnet in pelitic units is the predominant grade of metamorphism found in the Ruby Range (Okuma, 1971; Garihan, 1973). Locally, hypersthene-bearing assemblages of hypersthene-diopside-hornblende-plagioclase-quartz in some amphibolites, probably indicates
conditions of $P_{H_2O} < P_{load}$, and the development of granulite facies assemblages. Staurolite was noted in one locality and would seem to indicate a somewhat lower grade of metamorphism than the above assemblages (Garihan and Swapp, 1977). Cordierite is also found locally in the Ruby Range and apparently formed retrogressively as pressure declined (Garihan and Swapp, 1977). A recent geothermometric study by Dahl (1978) in the Ruby Range indicates that the orthopyroxene zone and sillimanite-orthoclase zone metamorphism took place at peak temperatures, estimated to have been $745 \pm 50 ^\circ C$ and $675 \pm 45 ^\circ C$ respectively. Migmatitic gneisses provide evidence for local anatexis and are consistent with such temperatures.

Temperatures of $600-800 ^\circ C$ seem appropriate for the main regional metamorphism of the area. Pressure estimates are more difficult to determine. If cordierite formed retrogressively due to declining pressures, then peak pressures during the formation of the earlier sillimanite assemblages probably exceeded 4 Kbar for the $600-800 ^\circ C$ temperature range (see curve 1, Fig. 32). Kyanite is found in at least one locality in the Ruby Range (Garihan and Swapp, 1977). If I assume that as temperature increased during metamorphism conditions moved from the kyanite stability field into the sillimanite field, then pressures of at least 5-6 Kbar are indicated for the $600-800 ^\circ C$ temperature range (Fig. 32). Without more restrictive or conclusive evidence, intermediate pressures of 5-10 Kbar are assumed during peak metamorphism.
Fig. 32. $P/T$ phase diagram indicating possible pressure conditions for metamorphism in the Ruby Range. Arrow indicates metamorphic conditions changing from kyanite stability field to sillimanite stability field (Curves taken from Hyndman, 1972).
Many of the reactions during progressive metamorphism of sediments are dehydration reactions, and it is assumed that metamorphism took place under water-saturated conditions \( (P_{H_2O} = P_{load}) \). The abundant hydrous minerals in the metamorphic assemblages indirectly supports this assumption.

Granulite assemblages in amphibolites probably indicate local conditions where \( P_{H_2O} < P_{load} \). The original rocks producing these assemblages may have been dry metavolcanics or other mafic igneous rocks. However, even in these local areas water was present in sufficient amounts to produce hydrous biotite and hornblende in the granulite assemblages.

The recrystallized assemblages and textures in the ultramafic rocks of the Ruby Range suggest that they resulted from metamorphism of serpentinite. A number of studies of regional and contact metamorphism of serpentinites have found similar assemblages and textures (e.g. Evans and Trommsdorff, 1970; Trommsdorff and Evans, 1972, 1974; Springer, 1974; Frost, 1975). Ultramafic assemblages of orthopyroxene-green spinel-forsterite-anthophyllite/cummingtonite + actinolite/tremolite and, orthopyroxene-green spinel-forsterite-actinolite/tremolite corresponded to upper amphibolite sillimanite zone host rock assemblages in the regional metamorphic terrane of the Alps (Trommsdorff and Evans, 1974). A similar grade of metamorphism in the Ruby Range (see above) produced the same ultramafic assemblages.
Locally the recrystallized ultramafic rocks of the Ruby Range host a two-pyroxene assemblage of orthopyroxene-diopside-hornblende with some plagioclase and a trace amount of quartz. The two-pyroxene assemblage corresponded to granulite assemblages in the Alps and is considered to signify granulite facies metamorphism (Evans and Frost, 1975; Medaris, 1975; Evans, 1977). Ultramafic assemblages in the Ruby Range correspond with both the upper amphibolite and granulite facies metamorphism indicated in the host rock assemblages.

Experimental studies on ultramafic systems summarized, for example, in Evans and Trommsdorff (1970) and Evans (1977) indicate that the broad range of temperatures and pressures for the formation of the above assemblages correspond quite well with the pressure and temperature conditions of metamorphism estimated from the host assemblages. Figure 33 is a P/T phase diagram for the system CaO-MgO-Al₂O₃-SiO₂-H₂O. The curves represent upper temperature limits for the various reactions. Progressive metamorphism of serpentinite follows a series of dehydration reactions (Fig. 33). Because of the nature of these reactions and the abundant hydrous minerals in the recrystallized assemblage, water saturation is assumed during the metamorphism of the ultramafic bodies. From Figure 33 I can see that the above ultramafic assemblages lie at least above curve 2, as they contain recrystallized orthopyroxene. The presence of green spinel in many of the ultramafic rocks in the Ruby Range suggests that many of the assemblages are above curve 3 where chlorite $\rightarrow$ forsterite $\rightarrow$ enstatite $\rightarrow$
Fig. 33. Part of P/T phase diagram for reactions in the system CaO-MgO-Al₂O₃-SiO₂-H₂O. Full lines experimentally reversed; dashed lines calculated, inferred, or provisional curves. a-antigorite, f-forsterite, t-talc, e-enstatite, cte-chlorite, sp-spinel, tr-tremolite, d-diopside, an-anorthite, co-cordierite, v-water vapor (taken from Evans, 1977). Numbers refer to reactions discussed in text.
spinel. The presence of a Ca-poor amphibole (cummingtonite) in these high grade assemblages is inconsistent with experimental evidence (Trommsdorff and Evans, 1974). This apparent overlap and persistence of cummingtonite into higher grade assemblages containing enstatite and forsterite has been noted in other studies as well (Trommsdorff and Evans, 1974; Frost, 1975). This persistence may be a metastable one. Textural evidence, however, does not support this as there are no apparent relict grains in disequilibrium between the cummingtonite/anthophyllite and the orthopyroxene and olivine grains. Frost (1975), studying contact metamorphism of serpentinite, attributed such an overlap to a kinetic problem with growing anthophyllite from the reaction talc + forsterite = anthophyllite + water. Instead of this reaction, he believes the metastable reaction forsterite + talc = enstatite + water may occur. Anthophyllite may eventually form. However, it would not come directly from forsterite + talc, but from preexisting enstatite by the reaction enstatite + water = forsterite + anthophyllite. Frost (1975) argues that because a large portion of the water in these rocks, originally in the talc, would have already been released in the formation of enstatite, the reaction enstatite + water = forsterite + anthophyllite would not likely go to completion. This would result in an apparently univariant forsterite-enstatite-anthophyllite assemblage.

Textural evidence for such a situation is lacking in the Ruby Range however. Trommsdorff and Evans (1974) believe such overlaps
in the metaperidotites of the Alps are due to polymetamorphism and variable fluid phase composition. The presence of more CO₂-rich fluids during metamorphism could result in the persistence of the Ca-poor amphibole (Evans, 1977).

The Ca-rich actinolite-tremolite amphibole is stable in the above ultramafic assemblages until conditions above reaction 4 are reached, where forsterite + tremolite = diopside + enstatite + H₂O. The two-pyroxene assemblage in the Ruby Range also contains plagioclase indicating reaction 5, where tremolite + 2 spinel = 3 forsterite + enstatite + 2 anorthite + H₂O, may have taken place prior to reaction 4. The presence of plagioclase would seem to indicate that pressures were not too high during metamorphism (conditions below reaction 6, Figure 33). The absence of cordierite (reaction 7, Fig. 33) and the presence of plagioclase in the ultramafic rocks suggests that intermediate pressures of 4-8 Kb may have existed during the formation of the recrystallized assemblages. At such pressures, a broad temperature range of 650-800°C would seem appropriate and encompass the conditions for the formation of the recrystallized ultramafic assemblages. This corresponds quite well with the temperature and pressure conditions for the upper amphibolite regional metamorphism in the Ruby Range as estimated from the host rock assemblages. The monomineralic diffusion zones sometimes associated with the ultramafic bodies are also compatible with having formed under upper amphibolite metamorphic conditions (Carswell et al., 1974).
Field relationships and petrologic and textural relationships indicate that the ultramafic bodies in the Ruby Range are isofacial in nature, that is, they participated in the same deformation and metamorphism experienced by the host rocks.

Serpentinization and Ca-Metasomatism

Subsequent to this high temperature metamorphic event the ultramafic rocks underwent a period of low temperature serpentinization. This reserpentinization may have occurred retrogressively during the waning stages of the upper amphibolite event or alternatively as a completely separate event.

Low temperature retrograde metamorphism in the form of greenschist facies mineralogies is found in the surrounding host rocks as well. The simplest interpretation of these assemblages is that they formed retrogressively during the late stages of the upper amphibolite event (Okuma, 1971). The lack of associated structural elements (foliation, folding) that may have developed in a separate low grade metamorphic event indirectly supports this hypothesis. Root (1965) made a similar interpretation of the greenschist assemblages in the Tobacco Root Mountains.

Reserpentinization of the ultramafic bodies probably coincided with the development of the retrograde greenschist minerals in the surrounding rocks. The hydrous nature of many of the greenschist minerals (e.g. actinolite, hornblende, sericite, chlorite, etc.) indicates the presence of abundant fluids needed for serpentinization.
The antigorite serpentine found in these ultramafic bodies (Heinrich, 1963) is consistent with its having formed in such a metamorphic environment (e.g. Wenner and Taylor, 1971, 1974). A hydrothermal metamorphic environment is also suggested by the calcite, calc-silicate veins and rodingites formed by Ca-metasomatism during serpentinization. Antigorite serpentinization probably occurred during the waning stages of the upper amphibolite event along with the formation of the green-schist assemblages in response to declining pressures and temperatures with abundant aqueous solutions present.

Cross-cutting chrysotile veinlets in serpentine may reflect still another lower pressure-temperature environment. Serpentinization of ultramafic rocks can even occur in the very low pressure-temperature weathering environment (Cashman and Whetten, 1976). Serpentinization of the ultramafic bodies in the Ruby Range may have occurred up to the present.
CHAPTER VI

SUMMARY

In the Ruby Range field, petrologic, and textural evidence indicate that the ultramafic bodies in their present structural position are not igneous in nature but metamorphic.

The petrogenesis of these ultramafic rocks, as indicated by the evidence presented in this paper, is as follows:

1. Crystallization of the relict orthopyroxene megacrysts either by igneous or metamorphic processes. (The relationships are unclear).

2. Serpentinitization of the ultramafic bodies, locally along the margins and along distinct planes of weakness within the ultramafic bodies. Serpentinitization may have occurred before or during emplacement.

3. Deformation and emplacement of ultramafic bodies into the metasedimentary/metavolcanic package.

4. Metamorphism during this deformation reaches the upper amphibolite facies and the characteristic metamorphic assemblages in both the host rocks and the ultramafic rocks developed.

5. Retrograde metamorphism probably during the waning stages of the upper amphibolite event produced greenschist facies.
mineralogies in the host rocks and a reserpentinization, with associated Ca-metasomatism, in the ultramafic rocks.
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