# Precambrian Geology and Bedded Iron Deposits of the Southwestern Ruby Range, Montana

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*By* HAROLD L. JAMES

*With a section on* THE KELLY IRON DEPOSIT OF THE NORTHEASTERN RUBY RANGE

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*A synthesis of present knowledge of the Precambrian geology and mineral deposits of the southwestern Ruby Range, Montana*

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# PRECAMBRIAN GEOLOGY AND BEDDED IRON DEPOSITS OF THE SOUTHWESTERN RUBY RANGE, MONTANA

By HAROLD L. JAMES


ABSTRACT

The Ruby Range is one of a series of uplifted blocks in southwestern Montana that are cored by Precambrian crystalline rocks, mostly of Archean age. These uplifted blocks owe their origin largely to movement in Late Cretaceous and Tertiary time on steeply dipping range-bounding faults, and they are separated by broad valleys and basins containing locally deformed but unmetamorphosed strata of Phanerozoic age.

The Precambrian crystalline rocks of the southwestern Ruby Range, an area of about 100 square miles, can be divided roughly into three northeast-trending belts, progressively younger to the west. The oldest and most easterly belt consists of an ill-defined sequence of Early(?) or Middle(?) Archean older gneiss and schist that underlies, with structural conformity, a central belt underlain in turn by a rudely tabular mass of Middle or Late Archean quartzofeldspathic gneiss that forms the crest of the range for much of its length. The quartzofeldspathic gneiss is grossly layered and certainly of complex origin, derived in part from sedimentary precursors and in part from syntectonic granitic intrusions of at least two different ages. On a regional scale, this quartzofeldspathic gneiss forms a basement complex to an overlying sequence of metasedimentary strata that make up the most westerly of the three belts.

The Middle or Late Archean metasedimentary sequence, here named the "Christensen Ranch Metasedimentary Suite," consisted originally of miogeoclinal-type sedimentary rocks, now represented by dolomite marble, diopsidic and hornblendic gneiss and schist, quartzite, mica and garnet schist, and banded iron-formation. These rocks are generally well bedded, and distinctive individual lithologic units can be traced for miles. The appearance of a conformable succession, however, is misleading: original stratigraphic order has in large part been destroyed by displacements on bedding-plane faults and possibly by nappe development early in the structural history of the area.

Amphibolite of Middle or Late Archean age is an abundant rock type, occurring as generally conformable screens and sheets, some as much as several thousand feet in thickness, in each of the three main rock groups. Most bodies are believed to have originated as mafic sills, of at least two, and probably of several, different ages. Ultramafic rock of Middle or Late Archean age occurs as small plutons in the "older gneiss and schist" unit and as pods and lenses in the quartzofeldspathic gneiss and Christensen Ranch Metasedimentary Suite; most of these bodies, if not all, were emplaced by plastic flowage and diatreme-type movement. Syntectonic Late Archean granite gneiss, some of it not readily distinguishable from older quartzofeldspathic gneiss, cuts the metasedimentary strata in a number of places and occupies extensive tracts near the range front in the southwestern part of the map area. Pegmatite of Late Archean and (or) Early Proterozoic(?) age is abundant, both as sheets and dikes of simple quartz-feldspar composition and as tourmaline-bearing pods and lenses, many of which are rudely zoned. The youngest Precambrian igneous rocks are diabase dikes of Middle Proterozoic age that occupy fractures related to a northwest-trending fault system.

All rocks of the area, except for pegmatites and diabase dikes, are strongly deformed and metamorphosed to amphibolite facies. The dominant structures are northeast-trending isoclines, which are refolded on north-trending axes; these may have been preceded by gravity slides and possibly nappe formation. The main orogeny, which was accompanied by syntectonic granitic intrusions, culminated in Late Archean time, about 2,750 m.y. ago. Later structural deformation consisted mainly of displacement on northeast- and northwest-trending faults, the latter active both in Precambrian time and during range uplift in late Mesozoic and Cenozoic time.

Known or potential mineral resources include talc (which has been mined in a number of places), graphite, and banded iron-formation. The iron-formation has been explored extensively in the Carter Creek area; these deposits are estimated to contain, to a depth of 300 feet, about 95 million tons of rock containing 28–29 percent iron recoverable as magnetite.

The Kelly iron deposit is in the northeastern part of the Ruby Range. Beds of iron-formation are contained in a conformable sequence of strata of the Christensen Ranch Metasedimentary Suite, here in normal stratigraphic order. A thick basal dolomite marble passes upward through garnet-rich strata into diopside-hornblende gneiss that contains the principal iron-formation, and the gneiss in turn gives way to a sequence of quartzite beds that also contain thin layers of iron-formation. These strata are folded into a broad, southeasterly plunging syncline, the buried axial zone of which has been squeezed up, diatreme-fashion, to form a body of lens-like cross-section that is cored by a mass of ultramafic rock. The iron-formation in this upthrust block, greatly thickened by plastic flow and internal folding, has been explored by test pits and drill holes, but despite a favorable composition, the economic potential is low; the quantity, to a depth of 300 feet, is estimated to be about 15 million tons of protore containing 33 percent iron recoverable as magnetite.


## INTRODUCTION

The Ruby Range is one of a number of uplifted blocks in southwestern Montana from which younger strata have been stripped, wholly or in part, to expose a

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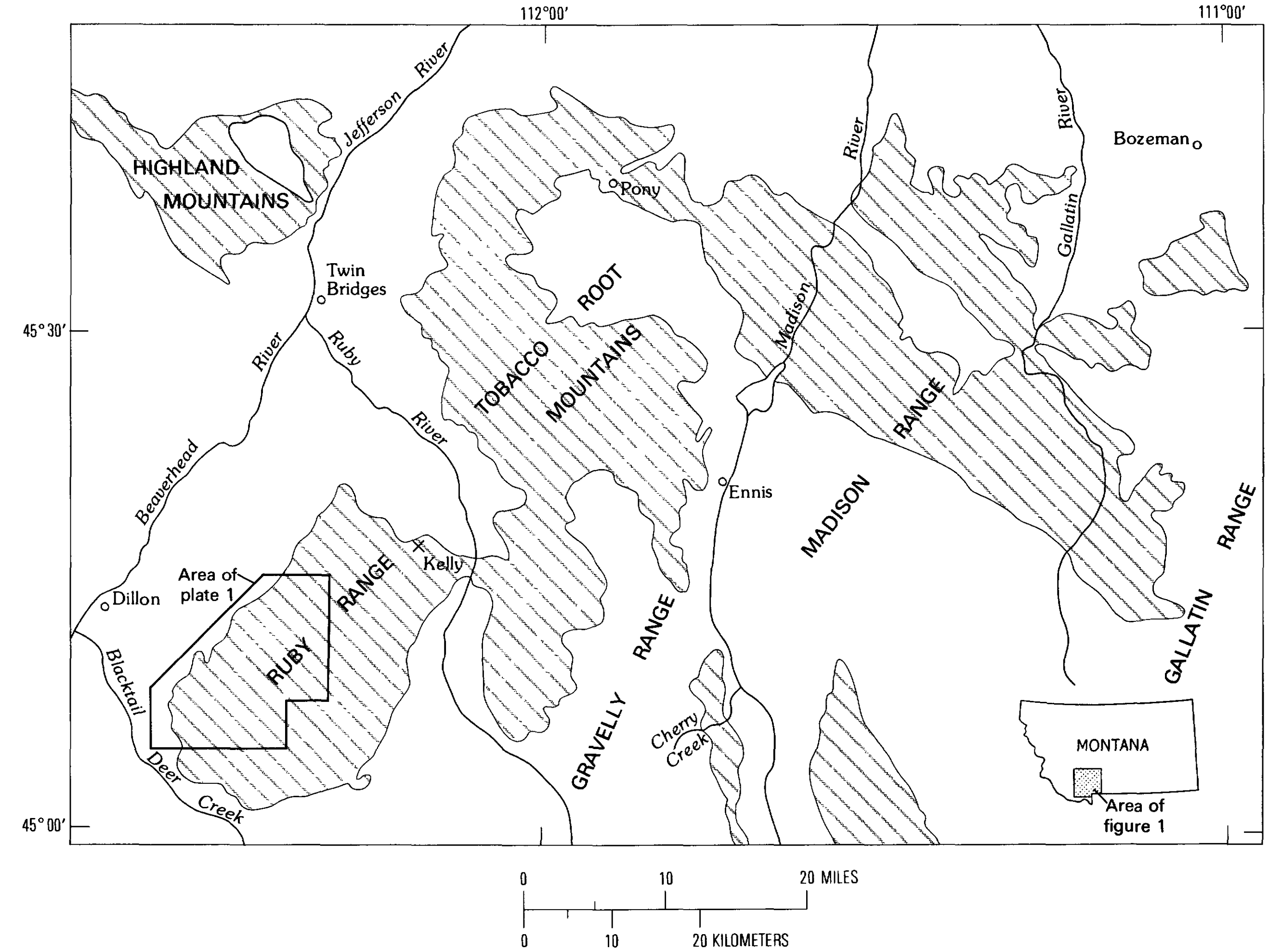



FIGURE 1.—Map of southwestern Montana, showing approximate distribution of Precambrian rock (diagonal line pattern), outline of area of this report, and location of Kelly area (X).

Precambrian core that consists mainly of crystalline and metamorphic rocks of Archean age (fig. 1). These uplifts, their crest elevations ranging from about 9,000 to 11,000 ft, are separated by broad valleys and basins underlain in large part by continental deposits of Tertiary age. The Ruby Range is bounded by the valleys of the Beaverhead River on the west, Ruby River on the east, and Blacktail Deer Creek on the south. All streams flow northerly to the Jefferson River, a major tributary of the Missouri River.

The southwestern part of the Ruby Range, the locus of the present study, is an area of relatively subdued upland topography; the maximum elevation is 8,530 ft, about 3,000 ft above the broad valley floors of the Beaverhead River and Blacktail Deer Creek. The nearest town is Dillon (population about 4,500), which is on a railroad line and at a junction of major highways. From these highways, dirt roads extend southeasterly across the range, following the principal drainage systems, Stone Creek in the north and Carter Creek–Sweetwater Creek in the central part of the mapped area. Secondary and tertiary routes traversable by small trucks and four-wheel-drive vehicles enter almost all parts of the area, but access commonly is by permission only; much of the land is fenced and locked gates are common. Some land on the lower slopes is cultivated, but most of the area is open range, in part covered with sagebrush. Some areas of upland, particularly in the southern and northeastern parts of the area, contain tracts of conifer forest, much of which is federal land managed by the U.S. Bureau of Land Management. Valley bottoms commonly are thickly overgrown with willow, alder, and cottonwood. Bedrock is generally fairly well exposed, though rarely in spectacular outcrop.

This report is an outgrowth of field studies that began in 1960 with detailed plane-table mapping and magnetometer surveys of the principal areas of banded iron-formation in southwestern Montana, the results of which were made available through preliminary open-file reports (James and Wier, 1961, 1962; Wier, 1965) and by follow-up map releases (James and Wier, 1972a, 1972b). Later in the 1960's, when 1:24,000-scale topographic base maps became available, the work was broadened to cover more general aspects of the Precambrian geology of the region, and, again, the results were made public through release of map data in preliminary form (James and others, 1969). Since then, U.S. Geological Survey field work in the southwestern Ruby Range has been limited to brief re-examination of critical areas and to sample collecting for special purposes, such as isotopic age dating. The aim of the present report and map (pl. 1) is to provide a comprehensive overview of the geology and mineral deposits of that part of the Ruby Range that has been studied in detail, incorporating both the studies made by the U.S. Geological Survey and by a number of independent workers over the past three decades.

The freedom of access granted to U.S. Geological Survey field workers by the various ranchers and landowners in the area during the course of these studies is gratefully acknowledged. A particular debt of gratitude is owed to Arthur and Margaret Christensen, owners and operators of the Christensen Ranch, whose hospitality and assistance to the many geologists who have worked in this area during the last 40 years is legendary and in the finest tradition of the American West.

## PREVIOUS WORK

The basic elements of the geology of the southwestern Ruby Range were established in the late 1940's and the 1950's by E.W. Heinrich and J.C. Rabbitt, working under the sponsorship of the Montana Bureau of Mines and Geology. This work resulted in a series of papers on special aspects of geology and mineral resources (Heinrich, 1948, 1949a, 1949b, 1950a, 1950b, 1963; Rabbitt, 1948) and culminated in a monograph (Heinrich, 1960) that has remained a key reference on the geology of the region, a foundation for all future work. Work prior to that of Heinrich consisted mainly of studies and brief reports on certain mineral deposits: graphite (Winchell, 1910, 1911, 1914; Bastin, 1912; Armstrong and Full, 1946; Armstrong, 1950), nickel (Sinkler, 1942), and talc (Perry, 1948).

The work of the U.S. Geological Survey that began in 1960 has already been noted. In part stimulated by the availability of new topographic base maps and by release of geologic data acquired in these U.S. Geological Survey investigations, the southwestern Ruby Range in the 1970's became the field base for a number of thesis studies by students from Pennsylvania State University (Okuma, 1971; Garihan, 1973; Karasevich, 1980), Indiana University (Dahl, 1977), and the University of Montana (Bielak, 1978; Desmarais, 1978). Many of these theses led to formal publications (Garihan, 1979a, 1979b; Karasevich, 1981; Dahl, 1979, 1980; Dahl and Friberg, 1980; Desmarais, 1981). Of particular value to the present report are the maps in Garihan's 1979b paper (available formally only in microfiche) and the later synthesis published in the "1981 Field Conference Guidebook of the Montana Geological Society" (Karasevich and others, 1981).

Other reports and papers dealing with specific aspects of the geology of the Ruby Range will be referred to in appropriate context later in this report. Of major importance in a regional sense is the monumental compilation of the geology of the Dillon 1°×2° quadrangle, now available in preliminary form (Ruppel and others, 1983).

## GENERAL GEOLOGY

The general form and outlines of the Ruby Range, like those of other ranges in southwestern Montana and east-central Idaho, were established by block uplift along northwest- and northeast-trending faults that culminated in late Tertiary time (Ruppel, 1982) and by subsequent erosion. The core area of the range is underlain mainly by crystalline and metamorphic rocks of Archean age (James and Hedge, 1980), which trend generally northeasterly and dip steeply to the northwest. These rocks are transected by undeformed dikes of Middle Proterozoic age (Wooden and others, 1978) that follow or are structurally related to northwest-trending faults.

Strata of Paleozoic and Mesozoic age are abundantly exposed in the northeastern part of the Ruby Range (Tysdal, 1976a, 1976b; Karasevich, 1981), but have been entirely stripped in the southwestern part. Clastic deposits of Tertiary age flank the range on the northwest, either in unconformable overlap or downdropped on the steeply dipping fault or faults of the northeast-trending, range-bounding system. The upland surface of the range and the gently sloping eastern flank locally are studded with flat-lying remnants of once more extensive basalt flows, at least some of which are of Pliocene age (Marvin and others, 1974).

The Precambrian rocks of the map area can be divided loosely into three northeast-trending belts. The northwest-facing slope of the range is underlain mainly by well-bedded sedimentary rocks that in the past (beginning with Winchell, 1914) have been designated "Cherry Creek Group" (or "Series," in some older reports) on the basis of assumed correlation with the Cherry Creek area on the eastern flank of the Gravelly Range, south of Ennis (see fig. 1). The crest of the range and much of the easterly sloping upland is underlain mainly by quartzofeldspathic gneiss, the Dillon Granite Gneiss of Heinrich (1960). This gneiss terrane is in turn bordered in discontinuous fashion to the southeast by a varied assemblage of crystalline and metamorphic rocks that Heinrich (1960) and most later workers have labeled simply "pre-Cherry Creek rocks" or "the pre-Cherry Creek group." These stratigraphic terms ("Cherry Creek," "Dillon," "pre-Cherry Creek") will not be used in this report, for reasons discussed in succeeding paragraphs:

*Cherry Creek.*—The strata of the Cherry Creek Group in the Cherry Creek area, described initially by Peale (1896), comprise a structurally complex sequence that contains a thick unit of dolomite marble, and quartzite, schist, and gneiss, and that has an aggregate thickness estimated by Peale to be "not less than several thousand feet." The Cherry Creek area has been studied and remapped by many workers over the past 90 years (for example, Runner and Thomas, 1928; Sahinen, 1939; Heinrich and Rabbitt, 1960; Hogberg, 1960; Hadley, 1969; Bayley and James, 1973). These later studies have added much new information concerning rock types that comprise the assemblage, but little or no progress has been made in establishing a verifiable stratigraphic succession or in more precisely defining age relations to other Precambrian rocks of the area. An Archean age for the group can reasonably be assumed on the basis of regional studies (James and Hedge, 1980) but, aside from this, the sole criterion for extension of the term beyond the Cherry Creek locality has been the presence of dolomite marble. It was on this basis alone that correlation was proposed for dolomite-bearing sequences in the Tobacco Root Mountains and the Ruby Range by Winchell (1914), at a time when the vast duration of Precambrian time was little understood or appreciated. As discussed later in this report (see "Correlation"), there are major differences in rock succession and associations between the metasedimentary assemblages of the Ruby Range and those of the Cherry Creek area. Later workers in the Tobacco Root Mountains (for example, Vitaliano and others, 1979; James, 1981) have abandoned the term "Cherry Creek" in favor of purely lithologic groupings. In this report on the Ruby Range, the strata previously labeled "Cherry Creek Group" will be assigned a new stratigraphic term, the "Christensen Ranch Metasedimentary Suite." At some future date, criteria may be developed to prove equivalence of this suite to the Cherry Creek Group of the Cherry Creek locality, but these criteria are not now available.

*Dillon Granite Gneiss.*—This term was introduced in a brief abstract by Heinrich (1953) to apply to certain intrusive bodies in the Ruby Range. Later, in a monographic report on the area, Heinrich stated that the pre-Cherry Creek and Cherry Creek sequences "... are separated by a thick intrusive mass of granite gneiss, named the Dillon Granite Gneiss by the writer" (Heinrich, 1960, p. 16).

The problem with the use of the term stems from uncertainty as to its proper definition. The description quoted above applies to the quartzofeldspathic gneiss of this report, which now is considered to predate the metasedimentary sequence (Cherry Creek). The original 1953 definition, however, clearly specified a younger age (post-Cherry Creek). Later workers (for example, Karasevich and others, 1981) have tended to ignore the age specification and have applied the term only to the basement quartzofeldspathic gneiss. It must be recognized, however, that present in the area are a number of granitic bodies that do conform to the original (1953) definition; these are much younger than the regionally extensive quartzofeldspathic gneiss.

Presumably, the term could be redefined so as to limit its application, but it is unlikely that the possibilities for confusion could ever be eliminated. In this report, therefore, rock bodies previously labeled "Dillon Granite Gneiss" will be simply assigned appropriate lithologic terms: "quartzofeldspathic gneiss" for the main body underlying the Christensen Ranch Metasedimentary Suite, and "granite gneiss" for the generally smaller bodies that are intrusive into, or developed within, the metasedimentary strata.

*Pre-Cherry Creek rocks.*—This assemblage was first described by Heinrich (1950a, p. 6) as consisting "... chiefly of injected and granitized gneiss, and hornblende and biotite gneisses of various types ..." that were assumed, without specific evidence, to underlie strata then assigned to the Cherry Creek Group in the southwestern Ruby Range. The term "Pony series" was applied to the assemblage by Heinrich on the basis of an assumed correlation with somewhat similar rocks in the Tobacco Root Mountains (Tansley and others, 1933), but in later reports, he replaced this term (now largely abandoned in the type area) with the designation "pre-Cherry Creek rocks." Inasmuch as the term "Cherry Creek" has been rejected for formal use in this report, it is obvious that the extension "pre-Cherry Creek" is equally unacceptable. I have, however, followed tradition in separating this rock assemblage, assigning it (despite a dearth of evidence as to relative age) to a category labeled "older gneiss and schist."

## ARCHEAN ROCKS

Most of the Precambrian rocks exposed in the Ruby Range are of Archean age (that is, older than 2,500 m.y.)

and range in character from well-preserved metasedimentary rocks to migmatite and banded gneiss of complex origin. The total time span represented by this diverse assemblage is not as yet known. The minimum age is about 2,750 m.y. (James and Hedge, 1980), but the possible lower age limit has not been established. In the Beartooth Mountains, about 100 mi to the east, rocks of similar character are older than 3,100 m.y., possibly as old as 3,300 m.y. (Reid and others, 1975).

As noted in a previous section, the Archean-age rocks in the southwestern Ruby Range are disposed in three northeast-trending belts. A general age progression—oldest rocks in the east, youngest in the west—is assumed, largely on the basis of regional considerations; in fact, however, no unequivocal direct evidence for relative ages of these three belts has been identified.

### OLDER GNEISS AND SCHIST

The principal outcrop of this group of rocks assigned an Early(?) and Middle(?) Archean age is in the south-central part of the map area (pl. 1), in the upland bordered on the north by the eroded scarp of the Carter Creek fault (extended). A smaller area, less certainly assigned to this category, lies about 1.5 mi to the northeast, across the Sweetwater Basin. Karasevich and others (1981) identify another area south of the Elk Gulch fault, extending to the topographic limit of the range, and Garihan (1979a) has assigned a "pre-Cherry Creek" status to several square miles of gneiss on the eastern flank of the Ruby Range, centering on Cottonwood Creek about 11 mi northeast of the principal outcrop area. The Cottonwood Creek locality provides the best exposures of this rock unit; it also illustrates the difficulties in separating this assemblage from the quartzofeldspathic gneiss of the central belt adjacent to the west.

The lithology of the older gneiss and schist unit cannot be simply categorized. The assemblage is diverse, comprising biotite gneiss, augen gneiss, migmatite, hornblende gneiss, and amphibolite; Garihan (1979b, p. 725) notes additionally sillimanite-biotite-garnet gneiss, some containing cordierite. Probably the most abundant rock type, well exposed in the Cottonwood Creek locality noted above, is banded gneiss in which dark layers alternate with layers and pods dominantly of quartz and feldspar, some in pegmatitic aggregates. The banded gneiss and the associated rock types do not differ in any essential way from some components of the structurally conformable quartzofeldspathic gneiss that is immediately adjacent to the west. As a whole, however, as noted by Garihan (1979b, fig. 7), the assemblage tends to be somewhat more mafic and richer in plagioclase. It also tends to be more distinctly layered, and structural deformation—generally tight complex folding—is more conspicuous than in the more massive quartzofeldspathic gneiss of the central belt.

The rocks exposed in the isolated area on the north side of the Sweetwater Basin (mostly in sec. 20, T. 8 S., R. 6 W.), questionably assigned to this map unit, are even more variable in character; they include sillimanite and anthophyllite schist, and, in the NE¼ sec. 20, corundum-bearing schist. The latter rock, exposed in scrapings on and adjacent to the Sweetwater road, contains zoned barrel-shaped crystals of lilac-colored corundum (sapphire), commonly about ¼-in. diameter, in a matrix consisting mainly of quartz and mica. The only other known occurrence of corundum in the area of the present report is in sec. 36, T. 8 S., R. 8 W., where it occurs in biotite schist and marble of the Christensen Ranch Metamorphic Suite (for description, see Heinrich, 1950b, and later in this report).

The parent rocks of the older gneiss and schist unit probably were in large part sedimentary, though a volcanic (or even plutonic) origin cannot be ruled out, particularly for the more homogeneous felsic layers. The hornblendic rocks (amphibolite and hornblende gneiss) here, as elsewhere in the region, probably are of diverse origins, some representing original mafic volcanic material, some younger dikes and sills that were metamorphosed along with enclosing older strata. The sillimanite- and corundum-bearing schists, however, almost certainly were derived from aluminous shale, and much of the banded gneiss could be the metamorphic equivalent of impure quartzite and graywacke. Karasevich and others (1981), in their review of the geology of the Ruby Range as a whole, consider the protoliths of this rock unit to have been mainly impure sandstone and shale and conclude that this sequence originally graded upward into illitic quartz mudstone and siltstone now represented by the quartzofeldspathic gneiss of the central belt.

### QUARTZOFELDSPATHIC GNEISS

This Middle or Late Archean rock unit, encompassed under the term "Dillon Granite Gneiss" in most earlier reports, forms the central northeast-trending belt in the Ruby Range, flanked on the east by older gneiss and schist and on the west by strata of the Christensen Ranch Metamorphic Suite. It is a tabular body, structurally concordant with adjoining rock units. The outcrop width ranges from about 2.5 to 4 mi, but this includes separately mapped bodies of amphibolite and narrow belts of infolded (or infaulted) dolomite marble of the Christensen Ranch Metamorphic Suite. Foliation and compositional layering dip rather consistently to the northwest at moderate to steep angles.

The dominant rock type, exposed in large rounded outcrops along the crest of the Ruby Range, is massive to foliated, medium-grained, gray to reddish brown gneiss of granitic composition. Rock textures, as seen in thin section, range from irregular to allotriomorphic granular, metamorphic in origin rather than igneous. The most abundant mineral generally is potassium feldspar, slightly to moderately perthitic, commonly with characteristic microcline grid twinning. The potassium feldspar is intergrown with oligoclase and quartz; typical proportions are 40:30:30, but vary widely among individual samples. Minor constituents are biotite (often altered to chlorite) and muscovite in scrappy small flakes, pink isotropic garnet, albite as rims to some oligoclase, fibrous sillimanite, and, in some specimens, nonperthitic microcline as small interstitial grains. Zircon, apatite, and magnetite are the usual accessory minerals. Foliation is due to flattening of felsic minerals and to planar concentration of micas and garnet. Locally the texture is granoblastic; triple junctions between feldspar and quartz are not uncommon. The chemical composition of typical massive quartzofeldspathic gneiss from three localities is given in table 1.

In the field, many variations from this characteristic rock type can be observed. Pods and layers of amphibolite, many partly granitized, are a common feature in many outcrops and are particularly evident in the axial zones of northeast-trending isoclinal folds. In places the normal quartzofeldspathic gneiss grades into banded gneiss containing more abundant dark minerals—biotite, garnet, and (more rarely) hornblende; elsewhere it grades into a felsic gneiss speckled with garnet. Discrete beds of diopside gneiss and quartzite have been found in a few places: these, like the separately mapped belts of dolomite marble, probably are structurally infolded strata of the Christensen Ranch Metamorphic Suite, but this cannot be proved; they may represent initial precursor components of the gneiss terrane. Structurally, the typical foliated massive gneiss may give way to more profoundly deformed rock, such as felsic sillimanite-bearing schist or strongly lineated tectonite, the latter particularly well developed on axes of the north-trending folds that cross the dominant northeast-trending structures. Despite these variations, however, the bulk composition of the quartzofeldspathic gneiss clearly is granitic, as shown by the chemical analyses (table 1). On the basis of point counts for 73 samples, Garihan and Williams (1976) conclude that the average composition (in terms of the principal minerals) is 42 percent potassium feldspar, 31 percent quartz, and 27 percent plagioclase.

The origin and even the relative age of the quartzofeldspathic gneiss remains problematical. Heinrich (1960) unequivocally classed it as igneous, a "tabular

TABLE 1.—*Chemical analyses of quartzofeldspathic gneiss from the southwestern Ruby Range, Mont.*

[In weight percent. Analyst, Ann Vlisidis]

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| $SiO_2$ | 73.60 | 72.99 | 74.04 | 73.54 | 72.5 |
| $Al_2O_3$ | 13.94 | 14.92 | 14.11 | 14.32 | 13.9 |
| $Fe_2O_3$ | 0.96 | 0.57 | 0.85 | 0.79 | 0.9 |
| FeO | .94 | 1.00 | 1.56 | 1.17 | 1.7 |
| MgO | .30 | .15 | .63 | .36 | .5 |
| CaO | .71 | .68 | 1.11 | .83 | 1.3 |
| $Na_2O$ | 2.80 | 3.60 | 4.00 | 3.47 | 3.1 |
| $K_2O$ | 6.00 | 5.70 | 3.10 | 4.93 | 5.4 |
| $H_2O^-$ | .10 | .04 | .08 | .07 | |
| $H_2O^+$ | .18 | .16 | .22 | .19 | |
| $TiO_2$ | .12 | .10 | .13 | .12 | .4 |
| $P_2O_5$ | .04 | .06 | .05 | .05 | .2 |
| $CO_2$ | .39 | .40 | .37 | .39 | |
| MnO | .08 | .05 | .06 | .04 | .1 |
| Total | 100.16 | 100.42 | 100.31 | 100.30 | |

Sample data:

1. Quartzofeldspathic gneiss from large outcrops in SW1/4 sec. 6, T. 8 S., R. 6 W. Sample HJ-1-69.
2. Quartzofeldspathic gneiss from outcrop in sec. 16, T. 8 S., R. 6 W., approximately 1,100 ft east of center of section. Sample HJ-11-69, collected by Karen Wier.
3. Quartzofeldspathic gneiss from outcrop in NE1/4 sec. 6, T. 9 S., R. 7 W. on Timber Creek road. Sample HJ-16-69, collected by Karen Wier.
4. Average of analyses 1-3.
5. Average of 72 calc-alkalic granites (Poldevaart, 1955, p. 134).

pluton" intruded between older gneiss and schist and the Christensen Ranch Metasedimentary Suite and therefore younger than either unit. Later work has cast some doubt on this interpretation. Garihan and Okuma (1974), noting the compositional heterogeneity of the gneiss, the structural concordance with adjacent strata, and the absence of skarn at contacts with dolomite, suggest that the gneiss represents "isochemically metamorphosed arkosic rocks." In later papers, Garihan (1979a, 1979b) made no clear choice between the igneous and metasedimentary alternatives, but in a subsequent review of the geology of the entire Ruby Range (Karasevich and others, 1981), he joined L.P. Karasevich, P.S. Dahl, and A.F. Okuma in firmly opting for a sedimentary precursor ("... a monotonous sequence of illite(?)-quartz mudstones or siltstones"). The view adopted in this report, following conclusions expressed in an earlier paper (James and Hedge, 1980) is that the gneiss is of diverse origins, in part derived from sedimentary and volcanic precursors and in part from granitic igneous intrusions, some (perhaps most) of which pre-date deposition of the sedimentary sequence now represented by the Christensen Ranch Metasedimentary Suite. On a regional scale, quartzofeldspathic gneiss appears to form a structurally conformable basement to dolomite-bearing metasedimentary sequences; this relation is evident not only farther north in the Ruby Range (Karasevich, 1981), but also in the adjacent Gravelly Range (Wier, 1982) and to the north in the Tobacco Root Mountains (Vitaliano and others, 1979; James, 1981).

### CHRISTENSEN RANCH METASEDIMENTARY SUITE

The strata that make up the suite are named for Christensen Ranch in T. 7 S., R. 7 W. and are well exposed on the northwest-facing slopes of the Ruby Range. Much of this area of exposure has been mapped in detail, primarily for the purpose of defining the distribution and economic potential of banded iron-formation (James and others, 1969; James and Wier, 1972b). The suite is recognized only in the report area. A secondary objective was to establish, if possible, a valid stratigraphic subdivision of this assemblage of distinctive metasedimentary strata of Middle or Late Archean age. The latter objective has not been achieved. It has been found that although the assemblage can readily be subdivided into mappable units, locally traceable in exquisite detail, no stratigraphic section can be established that will withstand critical analysis. A type area for the suite is designated as the northwest-facing slope of the Ruby Range between Stone Creek and Hoffman Gulch in T. 7 and 8 S., R. 7 W. Some of the difficulties encountered in defining formal units are those common to many areas of metamorphic rocks of complex history: general lack of internal evidence of time relations within or between beds (such as crossbedding or graded bedding), structural thinning and thickening due to flexural flow, and complex interference patterns resulting from isoclinal folding and crossfolding. These difficulties can be overcome in many areas by precise mapping of distinctive lithologic units. In this case, however, the effort has proved unsuccessful. The conclusion appears inescapable that stratigraphic order within the sequence has in large part been disturbed or lost, probably mainly by transposition on unrecognized bedding plane faults, particularly at boundaries of structurally competent beds such as marble.

Though not adequate for formal definition of stratigraphic units, some semblance of stratigraphic order does appear to be preserved, particularly if regional aspects are considered. From the Tobacco Root Mountains in the north, through the northern part of the Gravelly range and the adjoining northern part of the Ruby Range, and south into the area here being discussed, quartzofeldspathic gneiss is in contact with dolomite marble, which may be the only recognizable metasedimentary rock present. Where metasedimentary strata are preserved in greater abundance and some local order can be established, the marble generally is succeeded by garnetiferous gneiss and schist, then by schist, gneiss, and quartzite containing one or more beds of banded iron-formation. Possible correlations are discussed later in this report. For the moment, suffice it to say that within the Christensen Ranch Metamorphic Suite, the dolomite marble in immediate contact with quartzofeldspathic gneiss is considered to represent the oldest rock in the initial sedimentary sequence and that it is overlain stratigraphically by strata that include banded iron-formation.

For purposes of description and map display, the strata of the Christensen Ranch Metasedimentary Suite are subdivided entirely on the basis of lithology, without stratigraphic implications. Four categories are used: (1) quartzite, (2) iron-formation, (3) marble, and (4) undifferentiated metasedimentary rocks. The first three categories are of lithologically distinctive rocks, which may (and almost certainly do) occur at more than one stratigraphic position in the original sequence. The fourth category includes rocks as dissimilar as diopside gneiss and mica schist, which tend to form a matrix to the three separately mapped lithologic units. The estimated thickness of the suite is 6,400 ft.

#### QUARTZITE

Quartzite occurs throughout the suite in layers that range from ribs less than 1 in. thick in dolomite marble and mica schist to erosion-resistant beds as much as 100 ft thick. In the northwestern part of the map area, some of the most prominent beds persist without significant change in thickness for several miles, but thinner beds tend to pinch out in distances typically in the range of several hundred to several thousand feet. Quartzite ribs within dolomite marble commonly show metamorphic reaction borders of white diopside, and indeed some thin layers may have been totally consumed by this process. Elsewhere, quartzite within dolomitic beds has been squeezed into thick lenses and boudins by tectonic deformation.

Other than bedding, now marked by slight color changes and by concentrations of micaceous minerals, no primary structures or textures are preserved in the quartzite. Neither crossbedding, expectable in rocks of this composition and association, nor conglomerate facies have been certainly identified.

Most of the quartzite is white to yellowish brown and fine to medium grained. The texture ranges from sugary to vitreous, the latter only in more nearly pure varieties. Generally making up 90 percent or more of the rock, quartz occurs mainly as clear interlocking grains. Additional components of the rock are feldspar (typically sericitized and not readily identifiable as to type), muscovite, and chloritized biotite, all of which are present as small grains and flakes interstitial to quartz. Heavy minerals such as zircon are scarce.

In a few places, quartzite containing fine flakes of green chrome mica has been noted, generally in beds too thin to show separately at map scale. Diopside characteristically is present in quartzite that is associated with or interlayered with dolomite; with increased abundance the rock grades into light-colored diopside gneiss, often similar in appearance to quartzite.

## IRON-FORMATION

Iron-formation is a quantitatively minor but distinctive component of the Christensen Ranch Metasedimentary Suite, within which it occurs as beds that range in true thickness from less than 1 ft to perhaps as much as 100 ft. Locally, because of complex folding and flexural flow, the apparent thickness is much greater. Though no definitive stratigraphic assignments are possible within the suite, on the basis of associated rocks it is apparent that iron-formation occurred at several levels in the initial sedimentary sequence. Most commonly, the iron-formation is interbedded with mica schist and quartzite, but some thin layers are entirely within dolomite marble: for example, a bed less than 5 ft thick is exposed within dolomitic strata in the NE¼ sec. 10, T. 8 S., R. 7 W. and at the same stratigraphic position in the SW¼ sec. 9, T. 8 S., R. 7 W.

The bulk of the iron-formation in the area occurs as belts within the two structural blocks defined by the Stone Creek, Carter Creek, and Hoffman Gulch faults (fig. 2). Though it cannot be demonstrated with certainty, it is likely that these belts shown in figure 2 represent a single stratigraphic unit. As shown on the general geologic map (pl. 1), however, iron-formation has been located at many places elsewhere in the area. Quite possibly, some of these occurrences represent unresolved structural repetitions of the main iron-formation; most, however, probably are separate stratigraphic units.

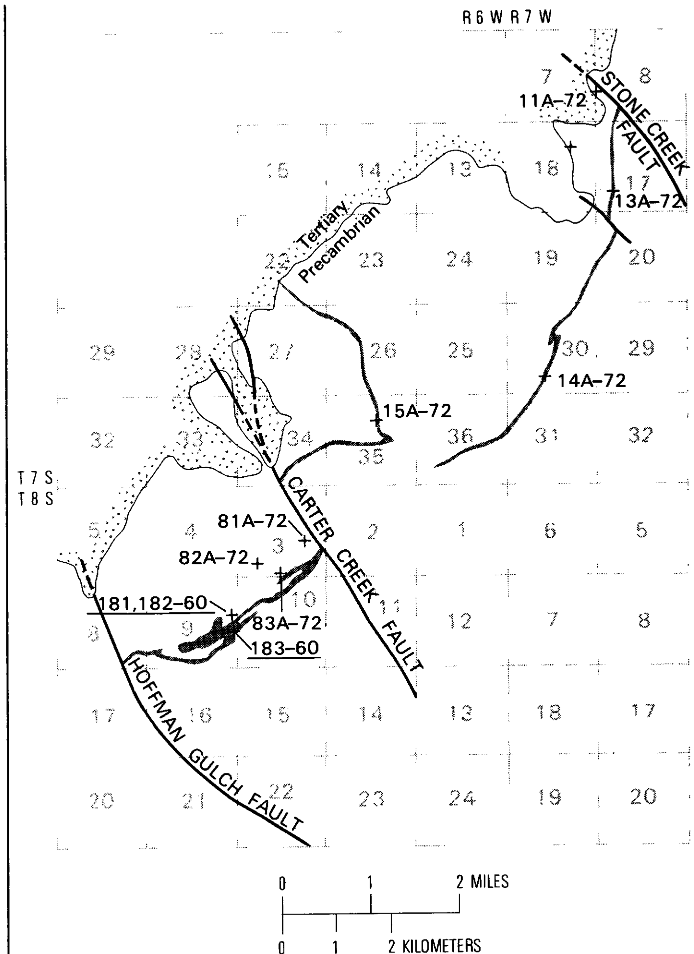



FIGURE 2.—Distribution of principal belts of iron-formation (gray), southwestern Ruby Range, and localities of analyzed samples. Underlined numbers, complete analyses (see table 2); others, partial analyses (see table 3); some from thin outlying beds not shown on map.

The iron-formation of the area, regardless of stratigraphic position, typically is a dark, heavy rock in which quartz-rich layers 1 in. or less thick alternate with layers of similar thickness composed largely of magnetite. In detail, the layering is seen to be discontinuous, pinch-and-swell, complex folding, and structural transposition being common features (fig. 3). Locally, the layered structure gives way to yield a rock of streaky or gneissic aspect. Iron-formation is exposed in bold outcrops in the belt between the Carter Creek and Hoffman Creek faults, but more generally it is weathered to a subdued surface marked by a distinctive reddish-brown soil containing relict chips of magnetic material.

Quartz and magnetite typically make up 75 percent or more of the rock by volume. Specularite is abundant in some samples but generally is scarce. Apatite, in

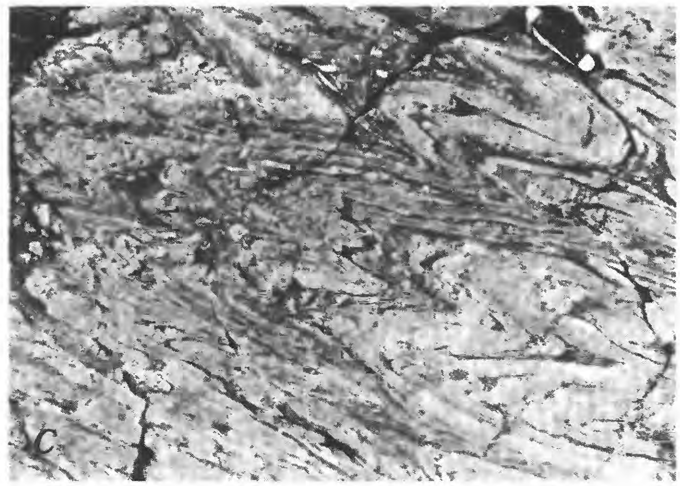





FIGURE 3.—Iron-formation in outcrop. *A*, Banded iron-formation. Trenched outcrop in SW¼ sec. 30, T. 7 S., R. 6 W. *B*, Pinch-and-swell structures in iron-formation. Outcrop in NW¼ sec. 10, T. 8 S., R. 7 W. *C*, Complex folding in iron-formation. Outcrop in NW¼ sec. 10, T. 8 S., R. 7 W.

small discrete crystals, generally accounts for about 2 percent of the rock and commonly is concentrated in particular layers. The silicate suite is complex; it includes both clinopyroxene and orthopyroxene, and a variety of amphiboles—cummingtonite-grunerite, actinolite, hornblende, riebeckite, and sodium tremolite. Other minerals, rarely present in more than trace amounts, are epidote, biotite, chlorite, and feldspar. Garnet is absent in the typical banded iron-formation but may be abundant in associated garnet-grunerite schist. The texture of most iron-formation is fine- to medium-grained granular. Quartz, the dominant mineral, is in anhedral grains that range from about 0.1 mm to 1 mm in mean diameter; other minerals are in grains of comparable or lesser size.

Relations between the amphiboles are described in some detail by Ross and others (1969) and by Immega and Klein (1976), who show that the silicates of the metamorphic assemblage have been greatly modified by exsolution. Lamellae of cummingtonite in host actinolite, actinolite in host cummingtonite, grunerite in host hornblende, and hornblende in host grunerite are reported. Riebeckite occurs as separate deep-blue grains and also as blue rims to light-green or bluish-green amphibole (some possibly sodium tremolite). The pyroxenes, which are relatively scarce and which have not been observed to occur together, consist of hypersthene and (more rarely) diopside-salite; neither shows exsolution effects.

Chemical data for iron-formation are given in tables 2 and 3. The table 2 analyses are of fresh, unoxidized material, selected so as to be representative of the rock as a whole; most samples are from drill core or exploration trenches. Analyses 1 and 4 are of rock containing abundant specularite, reflected in the analytical data by an excess of $Fe_2O_3$ after maximum assignment of FeO to form magnetite. The table 3 analyses are of outcrop samples, most of which are of partly oxidized material for which separate determination of ferrous and ferric iron would be relatively meaningless. Aside from this aspect, the data in the two tables show striking concordance in chemical composition for the iron-formation in the southwestern Ruby Range. Of particular note are the low contents of $Al_2O_3$ and MnO (less than 1 percent) and (except for analysis 4 in table 2) consistently high values for $P_2O_5$ compared with similar iron-formation elsewhere. The nature and significance of differences between iron-formation from the southwestern Ruby Range and that from the Kelly area of the northeastern part of the range are discussed further in this report, under "Correlation" and under "Mineral Resources."

TABLE 2.—*Complete analyses of iron-formation from the southwestern Ruby Range, Mont.*

**[Samples from the northeastern Ruby Range (Kelly area) included for comparison. In weight percent; n.d., not determined. Analyst, Paula M. Buschman]**

| | Southwestern Ruby Range (Christensen Ranch quadrangle) | | | | Northeastern Ruby Range (Kelly area) | | |
|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
| $SiO_2$ | 46.63 | 44.02 | 45.20 | 64.34 | 36.80 | 40.00 | 39.58 |
| $Al_2O_3$ | 0.84 | 0.66 | 0.73 | 0.29 | 3.80 | 1.95 | 2.47 |
| $Fe_2O_3$ | 30.99 | 32.29 | 38.05 | 28.61 | 27.71 | 32.91 | 32.07 |
| FeO | 13.38 | 17.14 | 10.99 | 5.08 | 23.26 | 18.87 | 19.22 |
| MgO | 3.32 | 3.21 | 2.14 | 1.00 | 3.60 | 2.46 | 2.41 |
| CaO | 2.46 | 1.37 | 1.22 | .23 | 2.14 | 1.87 | 1.90 |
| $Na_2O$ | 1.10 | .06 | .58 | .19 | 0.15 | 0.12 | 0.09 |
| $K_2O$ | .21 | .02 | .21 | .00 | 1.03 | .56 | .85 |
| $H_2O^+$ | .40 | .50 | .27 | .10 | .29 | .31 | .33 |
| $H_2O^-$ | .08 | .04 | .03 | .04 | .20 | .15 | .16 |
| $TiO_2$ | .02 | .01 | .01 | .01 | .27 | .08 | .10 |
| $P_2O_5$ | .38 | .55 | .51 | .02 | .07 | .10 | .09 |
| MnO | .07 | .04 | .06 | .03 | .63 | .73 | .75 |
| $CO_2$ | .10 | .15 | .01 | .00 | .17 | .15 | .16 |
| Cl | .01 | .00 | .01 | n.d. | .00 | .00 | .00 |
| F | .02 | .02 | .02 | n.d. | .01 | .01 | .01 |
| S | .00 | .00 | .00 | n.d. | .07 | .03 | .10 |
| C | .00 | .02 | .01 | n.d. | .00 | .00 | .02 |
| Subtotal | 100.01 | 100.10 | 100.05 | 99.94 | 100.20 | 100.30 | 100.31 |
| Less O | .01 | .01 | .01 | --- | .04 | .02 | .05 |
| Total | 100.00 | 100.09 | 100.04 | 99.94 | 100.16 | 100.28 | 100.26 |
| Fe | 32.08 | 35.90 | 35.15 | 23.96 | 37.46 | 37.60 | 37.37 |
| Mn | 0.05 | 0.03 | 0.05 | 0.02 | 0.49 | 0.57 | 0.58 |

Sample data:

1. Drill core from angle hole in SE1/4NE1/4 sec. 9, T. 8 S., R. 7 W. Sample 181-60.
2. Drill core from vertical hole in SE1/4NE1/4 sec. 9, T. 8 S., R. 7 W. Sample 182-60.
3. From trench in NE1/4SE1/4 sec. 9, T. 8 S., R. 7 W. Sample 183-60.
4. From outcrop in SE1/4 sec. 3, T. 8 S., R. 7 W., 200 ft E., 50 ft N. of S1/4 cor. sec. 3. Sample 83A-72.
5. Drill core from angle hole in NW1/4NE1/4 sec. 25, T. 6 S., R. 5 W., footage 435-445 ft. Sample 188-60.
6. Drill core from same hole as above, footage 633-643 ft. Sample 189-60.
7. Drill core from second angle hole in NW1/4NE1/4 sec. 25, T. 6 S., R. 5 W. Sample 190-60.

TABLE 3.—*Partial ("rapid") analyses of iron-formation from the southwestern Ruby Range, Mont.*

[In weight percent; <, less than. Analyses by Claude Huffman, Jr., project leader. Total Fe determined volumetrically and reported as $Fe_2O_3$; MgO, CaO, $Na_2O$, $K_2O$, and MnO by atomic absorption; $P_2O_5$ determined colorimetrically; total S and C determined by induction furnace]

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| $SiO_2$ | 41.8 | 43.6 | 43.3 | 43.3 | 43.5 | 51.6 | 58.4 |
| $Al_2O_3$ | <1.0 | <1.0 | <1.0 | <1.0 | <1.0 | <1.0 | <1.0 |
| $Fe_2O_3$ | 57.3 | 57.5 | 49.3 | 49.2 | 50.5 | 46.1 | 40.9 |
| MgO | 0.34 | 0.59 | 4.43 | 2.94 | 3.78 | 1.80 | 0.63 |
| CaO | 1.21 | .92 | 1.23 | 1.52 | 1.32 | 0.99 | .52 |
| $Na_2O$ | .01 | .02 | 0.15 | 0.18 | 0.15 | .01 | .04 |
| $K_2O$ | .02 | .04 | .04 | .05 | .06 | .01 | .04 |
| MnO | .04 | .09 | .27 | .55 | .42 | .02 | <.01 |
| $P_2O_5$ | .75 | .26 | .49 | .71 | .63 | .77 | .37 |
| S | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 | <.05 |
| C | .11 | .05 | .05 | <.05 | .09 | .05 | .09 |
| Total (Rounded) | 102 | 104 | 100 | 99 | 101 | 102 | 101 |
| Fe | 40.1 | 40.2 | 34.5 | 34.4 | 35.3 | 32.31 | 28.61 |
| Mn | 0.03 | 0.07 | 0.21 | 0.43 | 0.33 | 0.01 | 0.00 |

Sample data (see also fig. 2):

1. Sample 11A-72: SW1/4 sec. 8, T. 7 S., R. 6 W., 3,300 ft S., 100 ft E. of NW cor.
2. Sample 12A-72: NE1/4 sec. 18, T. 7 S., R. 6 W., 1,750 ft S., 1,400 ft W. of NE cor.
3. Sample 13A-72: SW1/4 sec. 17, T. 7 S., R. 6 W., 1,300 ft N., 850 ft E. of SW cor.
4. Sample 14A-72: SW1/4 sec. 30, T. 7 S., R. 6 W., 500 ft N., 2,000 ft E. of SW cor.
5. Sample 15A-72: NE1/4 sec. 35, T. 7 S., R. 7 W., 1,600 ft S., 2,000 ft W. of NE cor.
6. Sample 81A-72: SE1/4 sec. 3, T. 8 S., R. 7 W., 1,700 ft W., 3,100 ft S. of NE cor.
7. Sample 82A-72: SW1/4 sec. 3, T. 8 S., R. 7 W., 1,100 ft E., 700 ft N. of SW cor.

### MARBLE

Marble constitutes the most prominent component of the Christensen Metasedimentary Suite, within which it occurs as steeply dipping belts as much as several thousand feet in outcrop width. The rock is massive to well bedded and white, gray, or buff; bedding is marked by textural and compositional differences, in many places by ribs of quartzite and of white diopside. Exposed rock surfaces characteristically are speckled with bright-orange lichen. The texture varies, but most commonly it is fine- to medium-grained granular, weathering to smooth cuspate surfaces commonly lightly sprinkled with loose grains of carbonate. Locally, however, the rock may be very coarse grained, individual crystals being as much as 1 in. in diameter.

Marble of the area typically contains small to moderate amounts of light-colored diopside and (less abundant) tremolite, and scarce small flakes of graphite. Serpentine, in dark-green rounded blebs (presumably secondary after forsterite) is characteristic of certain beds, and phlogopite is abundant in some impure varieties of marble. Other minerals noted in some samples are quartz (as separate clastic grains in calcite marble and preserved in the interior parts of diopside-bordered ribs in dolomite marble), hornblende, plagioclase, scapolite, clinozoisite, and, very rarely, garnet. Heinrich (1960, p. 21) reported local abundance of thulite.

Most marble and other carbonate-bearing rocks in the area contain both calcite and dolomite, but in proportions, as estimated by staining tests and X-ray analyses, that range widely. The results, determined by an X-ray technique utilized by Gulbrandsen (1960) and refined by Royse and others (1971), for 32 randomly selected samples are summarized in figure 4.

Consideration of the analytical data in relation to the nature of material sampled leads to certain generalizations:

1. Most thick beds of relatively pure marble and diopsidic marble are dominantly dolomitic in composition, and some are virtually free of calcite. There are exceptions, however; for example, a 100-ft-thick bed of arenaceous marble that crosses Stone Creek in the SE¼ sec. 18, T. 7 S., R. 6 W. has a calcite-dolomite ratio of about 95:5.
2. Discrete layers of marble as much as 5 ft thick interbedded with diopside gneiss and other calc-silicate rocks generally are dominantly calcitic.
3. Carbonate-bearing gneiss and schist contain both calcite and dolomite in widely varying (and seemingly unpredictable) proportions. Much of this variation is believed due to the extent of dedolomitization reactions during metamorphism and to the relative initial abundance of dolomitic and siliceous constituents in the precursor rock.

In many places in the area, marble has been extensively affected by post-metamorphic hydrothermal processes, producing three different kinds of mineral associations:

1. Formation of talc and related minerals—Talc is widespread in the dolomitic marble units as seams, pods, and lenticular bodies, some of which are of economic significance (see further under "Mineral Resources"). The talc is light colored, locally bluish or greenish; associated minerals are chlorite and serpentine. Talc mineralization in places is accompanied by siliceous veining of the marble and by development of very coarse grain sizes of the constituent dolomite.
2. Formation of serpentine and cross-fiber chrysotile in silicate-rich marble at and near contacts with Middle Proterozoic diabase dikes—The chrysotile forms discontinuous veinlets as much as 0.5 in. across that are limited to zones a few feet wide, sporadically distributed along marble-diabase contacts. A number of occurrences have been explored by shallow test pits, as in secs. 35 and 36, T. 7 S., R. 7 W.; these reveal material of mineralogic interest but of little or no economic consequence.
3. Silicification along northwest-trending faults, which has produced masses of brown jasperoid as much as 800 ft wide along the Carter Creek fault and long screens of iron-stained silicified breccia 50 ft or more wide along the Stone Creek fault. The age of the silicification is not known for certain, but much of it appears to have resulted from warm spring activity that accompanied or post-dated late Tertiary movement on the faults. Minor amounts of azurite and malachite present in parts of the breccia along the Stone Creek fault have led to some exploration, including a short adit across the fault zone in the SE¼ sec. 21, T. 7 S., R. 6 W.

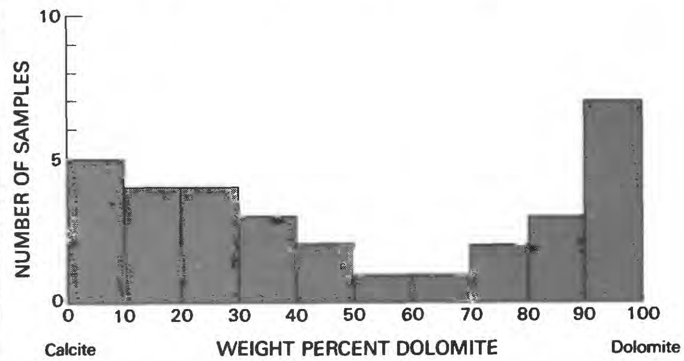



FIGURE 4.—Carbonate ratios of 32 samples of marble and other carbonate-bearing rocks of the Christensen Ranch Metasedimentary Suite, as estimated from X-ray analyses.

The stratigraphic significance of the marble remains uncertain. As stated previously, the facts of regional distribution in the Ruby Range, Gravelly Range, and Tobacco Root Mountains (which comprise a nearly continuous exposure of Precambrian rocks) indicate that dolomite marble immediately overlies a presumed basement of quartzofeldspathic gneiss. On this basis, the marble that is in immediate contact with, or chiefly enclosed in, the quartzofeldspathic gneiss of the southwestern Ruby Range is interpreted as the basal member of the Christensen Ranch Metasedimentary Suite, infolded or infaulted. It is to be noted that quartzite is scarce or lacking in this particular marble. Elsewhere in the area, however, notably in the refolded fold centering in sec. 24, T. 7 S., R. 7 W., marble is in immediate association with one or more beds of quartzite of mappable dimensions. Other belts of marble are in close stratigraphic proximity to iron-formation and schist. These diverse lithologic associations suggest that in fact marble occupies more than one, possibly as many as three, stratigraphic positions in the initial sedimentary sequence, of which only one (the basal unit) is preserved on a regional scale.

## UNDIFFERENTIATED METASEDIMENTARY ROCKS

The strata assigned to this category are those which, in essence, form interbeds to the more prominent or more distinctive lithologic units, such as marble, that have been separately mapped. The undifferentiated strata are grouped as follows: (1) quartz-mica schist and quartzose gneiss, including sillimanitic, garnetiferous, and corundum-bearing varieties, derived from initial sediments of shale or sandy shale composition; (2) calc-silicate gneiss and schist (and retrograde products that include para-amphibolite and hornblende-epidote gneiss), derived from shaly carbonate sediments; and (3) anthophyllite schist of uncertain origin. Each of these groupings is in itself complex, and in the aggregate, coupled with the separately distinguished quartzite, marble, and iron-formation, they reflect an almost complete range of sedimentary rock compositions in the initial depositional suite. The local presence of scapolite suggests that this initial assemblage may even have included an evaporitic component.

### Quartz-Mica Schist And Quartzose Gneiss

These rocks, most of which contain abundant mica, comprise a substantial percentage of bedrock in the area but in general are poorly exposed. All are well-foliated rocks, typically made up mainly of quartz, biotite, plagioclase, and potassium feldspar in varying proportions. Sillimanite and garnet are also common constituents. Corundum, staurolite, and muscovite are present in some samples. Accessory minerals include tourmaline, apatite, zircon, and magnetite; sericite, chlorite, and epidote are local alteration products. Irregular lenses and seams of granitic composition are a characteristic feature in many outcrops of schist, the result of local granitization (or homogenization). In places the felsic constituents are dominant and the rock assumes a gneissic aspect.

Sillimanite-bearing schist is widespread in the area, in places forming outcrop belts hundreds of feet wide, as for example in SE¼ sec. 34, T. 7 S., R. 7 W. The sillimanite occurs as fibrous bundles, in part as an apparent replacement of biotite, and it typically constitutes about 5 percent of the rock. Table 4 (analysis 1) presents a chemical analysis of sillimanite schist from the area. The high content of $K_2O$ in the sample (8.56 weight percent) is reflected in the mineralogic makeup by abundant potassium feldspar; the precursor rock probably was an illite-rich shale. In places sillimanite occurs as disc-like aggregates or lenses that range in maximum dimension from 1 in. or so to 1 ft or more. Larger masses consisting mainly of sillimanite are found in outcrop and associated rubble in an area of sillimanite schist in the SE¼ sec. 24, T. 8 S., R. 8 W., at the head of what is known locally as "Proffitt Gulch." Heinrich (1950a) describes the locality in detail and estimates some of the sillimanite masses ("dornicks") to weigh as much as 600 pounds.

Garnet is a common constituent of the mica schist and quartzose gneiss of the area and in many places is a major constituent, occurring as reddish-brown poikilitic crystals, some of which are rudely euhedral. Larger crystals—not uncommonly 2–3 in. in diameter—tend to accumulate in the soil and surface wash adjacent to weathered outcrops. Coarse-grained garnet is particularly abundant in the schist that bounds the synform of marble in sec. 2, T. 8 S., R. 7 W. This schist continues in a northeast-trending belt separating parallel units of marble, where it is interlayered with garnet-quartz-feldspar gneiss and garnet amphibolite. These

Table 4.—*Chemical analyses of metasedimentary rocks from the southwestern Ruby Range, Mont.*

[In weight percent. Analyst, Paula M. Buschman]

| | 1 | 2 |
|---|---|---|
| $SiO_2$ | 60.56 | 51.22 |
| $Al_2O_3$ | 16.17 | 10.54 |
| $Fe_2O_3$ | 0.85 | 3.81 |
| FeO | 6.68 | 3.50 |
| MgO | 2.47 | 9.54 |
| CaO | .21 | 11.80 |
| $Na_2O$ | 1.05 | 1.28 |
| $K_2O$ | 8.56 | 4.41 |
| $H_2O^+$ | 1.53 | 1.00 |
| $H_2O^-$ | .20 | 0.08 |
| $TiO_2$ | .83 | .70 |
| $P_2O_5$ | .04 | .15 |
| MnO | .03 | .13 |
| $CO_2$ | .01 | 1.50 |
| Cl | .01 | .01 |
| F | .09 | .10 |
| S | .00 | .00 |
| C | .06 | .14 |
| Subtotal | 99.35 | 99.91 |
| Less O | .04 | .04 |
| Total | 99.31 | 99.87 |

Sample data:

1. Biotite-sillimanite schist from low roadcut in NW1/4SE1/4 sec. 3, T. 8 S., R. 7. W., 780 ft E., 1,990 ft N. of S1/4 cor. sec. 3. Sample HJ-185-60.
2. Epidote-hornblende-feldspar-quartz gneiss from outcrop in NW1/4NE1/4 sec. 10, T. 8 S., R. 7 W., 130 ft E., 820 ft S. of N1/4 cor. sec. 10. Sample HJ-184-60.

various garnet-rich rocks probably have been derived from iron-rich shale, locally arenaceous.

An unusual corundum-bearing rock, in part enclosed in schist and in part in dolomite, is exposed near the northeast corner of sec. 36, T. 8 S., R. 8 W. This rock is described by Heinrich (1950b, p. 13) as follows:

> The gray corundum rock consists mainly of elongate corundum crystals scattered abundantly throughout a dense, exceedingly fine-grained matrix. Under the microscope the corundum-bearing rock can be seen to consist of individual grains and crystals of corundum, often zoned, set in a generally fine-grained matrix that consists of sericite, chlorite, margarite, calcite, and rare granules of magnetite.

The rock occurs as lenticular bodies in a belt as much as 20 ft in outcrop width and 220 ft in length. The only other known corundum in the area is in schist questionably assigned a pre-Christensen Ranch age (see previous discussion under "Older Gneiss and Schist"). At that locality—sec. 20, T. 8 S., R. 6 W.—the corundum similarly occurs as lilac-colored, barrel-shaped crystals in schist, but marble is absent from the associated strata. The rock at both localities, however, can be assumed to have been derived by metamorphism from aluminous shale.

### Calc-Silicate Gneiss and Schist

In this category are placed the diverse rock types derived by metamorphism from argillaceous dolomite and dolomite-bearing shale and sand. Physically, rocks of this suite range from erosion-resistant greenish-gray diopside gneiss to fissile brown, green, or gray schist.

Diopside gneiss is a common rock type in the area, grading variously into diopside quartzite, diopside-tremolite marble, or phlogopite-tremolite-diopside schist, depending upon the relative amounts of clay, sand, and carbonate in the precursor sediment, and into foliated amphibolite and epidote gneiss in localities of strong retrograde metamorphism. Diopside gneiss (and many of its variants) is well exposed in northeast-trending belts that cross sec. 10, T. 8 S., R. 7 W., where they alternate with parallel belts of dolomite marble and ortho-amphibolite (see James and Wier, 1972b, for details of distribution). In outcrop the rock is dark greenish gray, weathering to greenish brown, and generally layered on a scale of inches or feet. In thin section, the rock is seen to be a fine- to medium-grained granoblastic aggregate consisting mainly of gray diopside and quartz, and variable amounts of potassium feldspar, tremolite, brown-orange mica (generally assumed to be phlogopite), and calcite. Other constituents of fairly common occurrence are garnet, plagioclase, and scapolite; sphene, tourmaline, and magnetite are the typical accessory minerals. Grain diameters vary from sample to sample, but generally are in the range 0.1–0.7 mm. These various assemblages, initial equilibrium products of amphibolite-grade regional metamorphism, have been altered to varying degrees by retrograde metamorphism, in places enough to significantly change the character of the rock. The common retrograde minerals are actinolite or hornblende (which replace diopside), epidote, chlorite, and sericite. In places, as in the SW¼NW¼ sec. 27, T. 7 S., R. 7 W., alteration has produced hornblendic rock that physically is similar to the ortho-amphibolite of the area. The usual partial preservation of diopside and of some carbonate, however, reveals its metasedimentary ancestry. In another example, the resultant rock is epidote-hornblende gneiss, such as that exposed on the southeast-facing slope of the iron-formation ridge in the NW¼ sec. 10, T. 8 S., R. 7 W. A chemical analysis of a sample from this locality, where the gneiss contains 1 in.-thick interbeds of pink calcite marble, is given in table 4. The high content of $K_2O$ is unusual and is reflected in the mineralogic makeup of the rock by abundant untwinned potassium feldspar, now altered in considerable part to secondary clay minerals and sericite.

Another metamorphic product of impure dolomitic sediments is phlogopite schist, a fissile rock that differs mineralogically from gneissic units mainly in the relative amounts of mica and platy amphiboles. It differs also in certain associations; in places it grades into biotite-quartz schist of similar appearance. The rock, rarely well exposed, is gray or greenish on fresh break but weathers brown. Foliation commonly is deformed into complex nonsystematic crumples. A typical sample, collected from the poorly exposed belt of schist that extends northeasterly through the SW¼SW¼ sec. 30, T. 7 S., R. 6 W., consists, in order of abundance, of quartz, diopside, pale-brown to orange-brown phlogopite, and colorless to pale-green actinolite, and accessory blue-green tourmaline and magnetite. Elsewhere, similar schist may contain small to moderate amounts of calcite, pink garnet, potassium feldspar, or muscovite.

### Anthophyllite Schist

Anthophyllite-bearing rocks are quantitatively a very minor component of the Precambrian strata of the southwestern Ruby Range, but they are a distinctive feature, described in many reports beginning with that of Rabbitt (1948). Outcrops of the rock tend to be small and isolated, and relations to enclosing strata usually are obscure. The rock is strongly foliated, made up largely of light-brown to reddish-brown anthophyllite in platy aggregates. The origin of this curious rock is not always clear, but it is found in two quite different associations, both of which reflect metamorphism (prograde or retrograde) of magnesium-rich precursors, which may have been either igneous or sedimentary.

Anthophyllite schist is found in close proximity to bodies of ultramafic rock throughout the area. In the idealized situation described by Desmarais (1981), the anthophyllite is a transitional mineral phase, forming an inner zone between unaltered ultramafic rock and an outer hornblendic or biotitic zone. This zonal alteration records chemical transfer during metamorphism between the ultramafic intrusive and the host country rock. This ideal pattern is, however, rarely observed; more commonly, particularly where the intrusive bodies were small, the reaction may have consumed virtually all of the initial ultramafic rock, leaving only anthophyllite schist and related hornblendic rocks. Anthophyllite schist of this type is found as small lenses, rarely more than a few feet or a few tens of feet wide, at a number of places within areas of undifferentiated metasedimentary rocks, notably in secs. 2, 15, and 16, T. 8 S., R. 7 W.

Anthophyllite schist of metasedimentary origin is confined to areas of extensive retrograde metamorphism of diopside gneiss and schist, products of prograde metamorphism of argillaceous dolomite or dolomitic shale. Rock of this paragenesis has been recognized in a few areas, for example in outcrops in the SW¼SW¼ sec. 27, T. 7 S., R. 7 W., near the Sweetwater road. Here the anthophyllite rock forms a narrow zone separating dolomite marble from massive pegmatite. Some parts of the zone consist almost wholly of coarse-grained bladed anthophyllite, beige to light gray in thin section; other parts contain, in addition to anthophyllite, coarse-grained pink garnet, minor quartz, and accessory magnetite.

## CORRELATION

The preservation of many distinctive sedimentary features in the Christensen Ranch Metasedimentary Suite offers tantalizing prospects for application of stratigraphic principles and for regional correlation based on stratigraphic analysis. In the southwestern Ruby Range, despite good exposure and detailed lithologic mapping, the initial hope of achieving a valid stratigraphic subdivision of the suite has not been realized, principally because of lack of internal indicators of age relation within and between beds and because of probable major displacements on unrecognized bedding plane faults. Nevertheless, as discussed previously, some rude correlations may be possible, based mainly on regional distribution and mutual relations between dolomite marble and the quartzofeldspathic gneiss that is believed to be the older rock. This association is in nearly continuous exposure, interrupted only by some structural breaks, from the area of the present study northward into the northeastern Ruby Range (James and Wier, 1972a; Karasevich, 1981), into the adjoining part of the Gravelly Range (Wier, 1982), and thence northward into the Tobacco Root Mountains (Vitaliano and others, 1979; James, 1981). In each case, except for the Gravelly Range, dolomite marble is overlain by garnet-rich schist and gneiss, which are succeeded by metasedimentary rocks that include iron-formation and quartzite.

Within this rudely correlatable sequence, traceable for about 35 mi, the only probable direct equivalent is the basal dolomite marble, and even this interpretation is open to some question. Aside from the possibility that the relation between the dolomite marble and the quartzofeldspathic gneiss is structural rather than stratigraphic (that is, that the marble is the lower unit of a thrust sheet or nappe of regional dimensions), the marble exhibits considerable differences from one area to another, particularly in thickness. The true stratigraphic thickness in the southwestern Ruby Range has not been determined, but it may range from less than 200 ft to 2,000 ft or more. In the northern Ruby Range, notably in the area of Ruby Peak in secs. 16 and 17, T. 6 S., R. 5 W., the apparent thickness is about 2,500 ft, whereas in the central Tobacco Root Mountains it generally is less than 100 ft. The iron-formation units, often considered key elements in regional correlation, are in fact sufficiently different in occurrence, composition, and association from place to place as to cast serious doubt on stratigraphic equivalency. In all, it is concluded that although physical continuity and general character indicate that the metasedimentary assemblages in the northern part of the Ruby Range and in the Tobacco Root Mountains are at least in part stratigraphically equivalent to those of the Christensen Ranch Metasedimentary Suite, more specific correlations are not warranted.

The possibility of correlation between the Christensen Ranch Metasedimentary Suite and the Cherry Creek Group of the eastern Gravelly Range, some 60 mi to the east and separated by major structural breaks, remains an open question. The assemblages are of rudely equivalent age and both contain banded iron-formation and thick units of dolomite marble. In detail, however, major differences are apparent, notably in relative position of the dolomite and iron-rich strata. In the Christensen Ranch Metasedimentary Suite (and its extensions to the north) bedded iron-formation is in a sequence that overlies a basal dolomite marble, whereas the opposite is true in the Cherry Creek area. A partial succession for the latter area, where stratigraphic order is established by crossbedding in quartzite and pillow structures in greenstone, in order of increasing age is as follows (from Bayley and James, 1973):

| *Lithologic unit* | *Thickness, in feet (approximate)* |
|---|---|
| Dolomite | 1,000 |
| Phyllite | 50 |
| Greenstone | 700 |
| Meta-latite | 100 |
| Phyllite, with thin layers of iron-formation | 200 |
| Banded iron-formation | 35 |
| Phyllite, with one or more layers of crossbedded green quartzite | >500 |

Bed-by-bed correlation of strata in the two areas clearly can be discounted. The possibility remains, however, that the strata may in fact be age-equivalent and that the differences now observed reflect initial facies changes in a sedimentary basin of regional extent.

## ULTRAMAFIC ROCKS

Intrusive bodies of ultramafic composition assigned to the Middle or Late Archean are found throughout the area. Most are small pods or lenses a few tens of feet in length and width, but irregularly shaped bodies as much as 1 mi in maximum dimension are present in the southeastern part of the map area, where they make up the Wolf Creek pluton of Heinrich (1960, 1963). In outcrop the rock is dark and massive to schistose; surfaces commonly are rough and studded with weather-resistant pyroxene grains. The ultramafic bodies were emplaced prior to at least the termination of the last major regional deformation and metamorphism. Schistosity, which may be limited to marginal areas in larger bodies, is parallel to that of the enclosing rocks; original textures and minerals generally are well preserved only in the interior parts.

The initial rocks ranged from harzburgite to pyroxenite in composition, depending upon relative amounts of olivine and pyroxene, but in most places these have been modifed extensively by serpentinization and post-emplacement metamorphism. The mineral assemblages are described in some detail in Desmarais (1981), and the following discussion is drawn in part from that work. In the larger bodies the least altered rock consists of poikiloblastic orthopyroxene set in a finer grained matrix of olivine, orthopyroxene, magnesio-hornblende, spinel, and magnetite. The olivine ($Fo_{65}$–$Fo_{80}$) and the orthopyroxene ($En_{71}$–$En_{81}$) are relatively rich in iron, compared to those typical of alpine peridotite. The spinel, green in thin section, has a generally high content of alumina; Cr/(Al+Cr) ranges from 0.019 to 0.150. In the larger bodies, this relatively unaltered core grades outward in a series of reaction zones having the following generalized sequence: (1) anthophyllite rock, locally containing calcium amphibole; (2) hornblende- or gedrite-rich rock, locally containing garnet; and (3) biotite-rich schist. These alterations reflect exchange reactions between the intrusive ultramafic rock and the enclosing gneiss during metamorphism. Other reaction products include clinohumite, actinolite, serpentine, and talc. Desmarais (1981, p. 73) also notes the occurrence of rodingite lenses composed of clinozoisite, epidote, diopside, grossularite, vesuvianite, and actinolite, cut by veins containing chlorite, sphene, calcite, and prehnite. Yellow-green crusts of annabergite have been recognized on surface exposures of the Wolf Creek pluton (Sinkler, 1942; Heinrich, 1960); a number of shallow test pits have been sunk in the area to investigate the extent of the nickel mineralization.

It is likely that most of the ultramafic bodies, and almost certainly all those occurring as lenses and pods in strata of the Christensen Ranch Metasedimentary Suite, were emplaced tectonically by plastic flow rather than by intrusion of magma. Ultramafic diatremes are known elsewhere in the region, as for example in the northeastern Ruby Range and the Tobacco Root Mountains (James and Wier, 1972a; James, 1981), where they are located in axial areas of major folds in metasedimentary rocks. However, the larger masses in the southwestern Ruby Range occur within strata believed to be the oldest Precambrian rocks of the region, and structural control of their location within those strata is not immediately obvious. The possibility must be entertained, therefore, that this terrain was the original site of ultramafic emplacement, whether by plastic flow or as magma, and that subsequent orogenic movements resulted in breakup and boudinage of initial intrusive bodies and emplacement as "pumpkin seed" lenses and pods throughout the quartzofeldspathic gneiss and the Christensen Ranch Metasedimentary Suite. This model, here tentatively adopted, is consistent with the highly complex mineralogic record worked out by Desmarais (1981).

## AMPHIBOLITE

Amphibolite of Middle or Late Archean age occurs throughout the area, in bodies that range in size from layers and lenses too small to show at 1:24,000 map scale to sheets as much as 3,000 ft in outcrop width that are traceable for miles. In outcrop the rock is massive to moderately well foliated, having a salt-and-pepper appearance. Grain size is variable, mostly from medium to coarse, and in part appears to reflect original differences in the parent rock. The other major difference observable in outcrop is in the amount of garnet, which ranges from nearly zero to as much as 25 percent.

Most of the amphibolite is mineralogically simple, consisting of about equal parts green hornblende and plagioclase (oligoclase-andesine), which together make up 90 percent or more of the rock. Quartz content ranges from 5 to 10 percent, and biotite is present in trace amounts. Accessory minerals are apatite and magnetite. Garnetiferous amphibolite is more variable: in addition to the garnet, hornblende may make up as much as 60 percent of the rock and quartz may be more abundant than plagioclase. Both types of amphibolite locally have been extensively altered by retrograde metamorphism: sericite and epidote replacement of plagioclase; chlorite, actinolite, and epidote replacement of hornblende. A third, less common type of amphibolite contains variable amounts of diopside and (in some specimens) carbonate, and sphene as a typical accessory. This rock is believed to be a para-amphibolite that owes its origin to retrograde metamorphism of diopside gneiss, which in turn was derived from a sedimentary precursor.

Other than the diopside-bearing variety, most of the amphibolite is considered to have been derived from mafic igneous rock (the classic reference for amphibolite-grade metamorphism). Some thin layers may represent metamorphosed basaltic flows or tuffs, but most bodies probably were emplaced as diabase or gabbro sills. At least two, and probably several, epochs of mafic intrusion are represented. The latest is recorded by a single dike, exposed in secs. 22 and 23, T. 7 S., R. 7 W., which transects northwest-trending strata that include conformable amphibolite of earlier age. This dike rock, now a foliated amphibolite, probably is equivalent to "metabasite" intrusive bodies farther north in the Ruby Range (Garihan, 1979b) and to metamorphosed but undeformed mafic dikes in the Horse Creek area of the Tobacco Root Mountains (Cordua, 1973).

The amphibolite, despite its generally massive aspect, was structurally mobile during deformation, so that much of the apparent thickness variation of individual sheets is due to plastic flow. Direct evidence of plastic flow is provided by the abundance of amphibolite lenses, arcuate in plan view, aligned along crests of major structural axes. A number of amphibolite "horse collars" of this type are found along the northeast-trending fold axis in quartzofeldspathic gneiss in the NW¼ sec. 7, T. 8 S., R. 6 W.

### GRANITE GNEISS

Granite gneiss of Late Archean age, much of it not physically distinguishable from older quartzofeldspathic gneiss (parts of which may in fact have been remobilized in the Late Archean orogeny), is found as (1) sills and concordant lenses in strata of the Christensen Ranch Metasedimentary Suite; and (2) as larger masses developed in or engulfing quartzite-bearing strata of the range-front terrain in the southwestern part of the map area. In both occurrences the rock is typically medium grained, gray or brown in outcrop, and at least moderately foliated parallel to the structure in adjacent strata (fig. 5). In a few places, however, the rock is strongly sheared, virtually mylonitic.

Granite gneiss of category (1) above is typified by the sill that is exposed in small but prominent outcrops in the NW¼ sec. 31, T. 7 S., R. 6 W. and by sills in dolomite marble in secs. 1 and 2, T. 8 S., R. 7 W. The rock of the sec. 31 locality, as seen in thin section, is medium to coarse grained and has an allotriomorphic or irregular texture. It is composed principally of nonperthitic microcline, quartz, and oligoclase (about $An_{25}$), all of which commonly show minor fracturing and some grain-margin granulation. Foliation is defined mainly by small flakes of dark-brown biotite and minor muscovite. Accessory minerals, mainly magnetite and zircon, are scarce. Trace amounts of fibrous sillimanite are present. Much more strongly deformed rock is observed in the sills in dolomite marble. Specimens from a sill exposed near the W¼ cor. sec. 1, T. 8 S. R. 7 W. are of a thoroughly granulated rock in which quartz is drawn out in parallel strings several millimeters in length. Microcline and plagioclase (oligoclase) occurs both as fractured large grains and as small recrystallized interstitial grains, and the rock contains a small amount of pink garnet. Alteration to sericite and chlorite is extensive. Chemical compositions of granite gneiss from these two localities are given in table 5, columns 1 and 2.

Granite gneiss of category (2) underlies larger areas and probably is polygenetic in origin. Some appears to have been developed in place, the product of granitization or partial melting of pre-existing pelitic strata, now represented by relict layers of quartzite. The larger part, however, is of igneous origin, occurring as sheet-like

FIGURE 5.—Outcrop of Late Archean granite gneiss. Axes Canyon, SW¼ sec. 24, T. 8 S., R. 8 W. Outcrop is about 70 ft high.

TABLE 5.—*Chemical analyses of Late Archean granite gneiss from the southwestern Ruby Range, Mont.*

[In weight percent. Analyst, Ann Vlisidis; $Na_2O$ and $K_2O$ determined by Rapid Rock Laboratory, under Leonard Shapiro]

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| $SiO_2$ | 71.50 | 72.62 | 72.04 | 76.04 | 73.05 | 72.5 |
| $Al_2O_3$ | 16.47 | 14.42 | 12.31 | 11.76 | 13.74 | 13.9 |
| $Fe_2O_3$ | 0.59 | 1.03 | 1.58 | 1.43 | 1.16 | 0.9 |
| FeO | .75 | 1.48 | 2.97 | 1.16 | 1.59 | 1.7 |
| MgO | .47 | 0.43 | 0.84 | 0.41 | 0.54 | .5 |
| CaO | 2.07 | 1.72 | 1.78 | .78 | 1.59 | 1.3 |
| $Na_2O$ | 4.7 | 2.5 | 2.4 | 2.3 | 2.97 | 3.1 |
| $K_2O$ | 2.6 | 4.9 | 4.9 | 5.4 | 4.45 | 5.4 |
| $H_2O^-$ | .06 | .08 | .04 | .02 | .05 | |
| $H_2O^+$ | .20 | .24 | .26 | .28 | .24 | |
| $TiO_2$ | .18 | .18 | .66 | .29 | .33 | .4 |
| MnO | .03 | .07 | .08 | .05 | .06 | |
| $CO_2$ | .34 | .48 | .17 | .24 | .31 | |
| $P_2O_5$ | .07 | .07 | .17 | .05 | .09 | .2 |
| Total | 100.03 | 100.22 | 100.20 | 100.21 | 100.17 | |

Sample data:

1. From sill in schist in NW1/4 sec. 31, T. 7 S., R. 6 W. Sample HJ-4-69.
2. From sill in dolomite marble in NW1/4 sec. 1, T. 8 S., R. 6 W. Sample HJ-2-69.
3. From sheet-like body in SW1/4 sec. 24, T. 8 S., R. 8 W. Sample HJ-13-69.
4. From same locality as 3. Sample HJ-14-69.
5. Average of analyses 1-4.
6. Average of 72 calc-alkalic granites (Poldevaart, 1955, p. 134).

masses of medium-grained, moderately well foliated rock that in thin section is seen to be composed mainly of finely perthitic microcline, plagioclase (andesine, about $An_{38}$, gradationally zoned to albite margins), and quartz, with lesser amounts of biotite and green hornblende. The texture is irregular and many grains show fracturing. Accessory minerals—apatite, allanite(?), magnetite, and zircon—are relatively abundant, and garnet is not rare. The chemical compositions of two samples are given in table 5, columns 3 and 4.

The average composition of the four analyzed samples (table 5) is not significantly different from the average composition of older quartzofeldspathic gneiss (table 1), and both closely resemble that of the average calc-alkali granite quoted by Poldevaart (1955, p. 134).

## LATE ARCHEAN AND (OR) EARLY PROTEROZOIC(?) PEGMATITE

Pegmatites are abundant in the area, and, as early recognized by Heinrich (1949a, 1949b, 1960), two varieties can be distinguished: (1) older pegmatite, which occurs as thick concordant sheets and straight-walled dikes, yellowish weathering and of simple composition; and (2) younger pegmatite, which occurs as smaller podlike bodies that commonly are rudely zoned to quartz-rich cores.

Older pegmatites, which are the more abundant of the two varieties, typically consist almost entirely of coarsely crystalline perthitic microcline and quartz, and minor amounts of albite-oligoclase as separate grains.

Muscovite, tourmaline, and garnet generally are absent but are scarce constituents in some samples. Most bodies show some cataclastic deformation, evident in both outcrop and thin section, and a few bodies are foliated parallel to contacts with country rock.

Pegmatites classed as "younger" are more complex in mineralogy and in internal structure. Samples sufficiently fine grained to be worth thin-section study consist of interlocking grains of perthitic microcline, twinned plagioclase (oligoclase-albite), and quartz, and variable amounts of tourmaline (black in hand specimen, pleochroic in dark green and brown in thin section), and muscovite. Rose quartz is common in core zones, which may contain large crystals of tourmaline and muscovite, together with lesser amounts of brown garnet and bluish-green apatite. A number of the zoned pegmatites have been explored by shallow pits and trenches, presumably to assess the economic possibilities for commercial production of sheet mica.

The older, more abundant pegmatites are assigned a Late Archean age. General cataclasis and local development of foliation suggest that they were intruded at the end stage of a Late Archean orogeny, which has an age of about 2,750 m.y. (James and Hedge, 1980). The younger pegmatites post-date this 2,750-m.y. event and are cut by diabase dikes having a Rb-Sr age[1] of about 1,424 m.y. (Wooden and others, 1978). Isotopic measurement on muscovite from a zoned pegmatite in sec. 3, T. 8 S., R. 7 W. yielded a K-Ar age of 1,660 m.y. and a Rb-Sr age[1] of 1,646 m.y. (Giletti, 1966, sample 8). Whether these values reflect time of crystallization is not known for certain. As Giletti (1966) has shown, the Archean strata of southwestern Montana west of the Gallatin River show isotopic evidence of a thermal event of about this age, and it is not unreasonable to assume that it was accompanied locally by generation of pegmatite magma.

## MIDDLE PROTEROZOIC DIABASE

Diabase dikes transect the generally northeasterly trends of crystalline rocks in the area and clearly are related structurally to the northwest-trending fault system. A few follow faults directly, but many others are in en echelon fractures in zones parallel to major breaks. Thicknesses (as measured in outcrop width) are as much as 500 ft, but most are less than 100 ft. Individual dikes rarely are traceable for more than 1 mi or so, but many separate bodies may be aligned along the same structural trend. The rock is not resistant to erosion: outcrops are rare, but dike trends commonly are well marked by topographic sags and by presence of spheroidally weathered boulders.

All of the diabase is altered extensively to secondary minerals, but original diabasic or gabbroic textures are well preserved. The initial mineral assemblage consisted of augite, now largely altered to actinolitic amphibole and chlorite; plagioclase (labradorite), now gradationally zoned to albite and altered to clinozoisite and sericite; minor quartz; and accessory magnetite and ilmenite, the latter as exsolution lamellae now altered to leucoxene.

Contact effects on country rock are relatively insignificant except where dikes cross silicate-bearing dolomite marble; here the marble for a few feet adjacent to the contact is irregularly altered to serpentine that in places contains veinlets of cross-fiber chrysotile asbestos.

The diabase dikes of this area are part of a swarm of generally northwest trending dikes that cut older Precambrian rocks of the Ruby Range and of the adjacent Tobacco Root Mountains to the north. Wooden and others (1978) class the diabase in the Ruby Range as "low potassium tholeiite" and give an average composition as follows (in weight percent):

| | |
|---|---|
| $SiO_2$ | 48.6 |
| $TiO_2$ | 1.10 |
| $Al_2O_3$ | 14.2 |
| $FeO_T$ | 11.5 |
| MnO | 0.19 |
| MgO | 7.54 |
| CaO | 12.1 |
| $Na_2O$ | 1.92 |
| $K_2O$ | 0.38 |

Whole-rock Rb-Sr analyses yielded an isochron[1] of 1,424 m.y.±125 m.y. and an initial $Sr^{87}/Sr^{86}$ ratio of 0.7019±0.0008.

## TERTIARY STRATA

Rocks of Tertiary age flank the crystalline core of the southwestern Ruby Range and locally are preserved on upland surfaces. These strata have not been studied in detail and are here described in summary fashion only.

The beds that flank the range on its northwest margin consist of weakly lithified siltstone, sandstone, and conglomerate. Their precise age is not known, but similar strata at the north end of the Ruby Range are Eocene to Pliocene (Petkewich, 1972). These strata are dropped down on the northeast-trending range-bounding fault system and continue westerly to underlie the intermontane Beaverhead basin. Where exposed along valley walls, beds are generally horizontal or gently dipping, but near the contact with the Precambrian they are

[1]Recalculation, based on decay constant of $1.42\times10^{-11}yr^{-1}$.

tilted at angles as great as 30° on minor faults. The main Tertiary-Precambrian contact, however, is generally masked by the mantle of Quaternary gravel and debris on the pediment surface that bevels both older and younger rocks and by Quaternary alluvium in the valleys. Locally, as in the SE¼ sec. 28, T. 7 S., R. 7 W., the basal Tertiary strata are in exposed unconformable contact with Precambrian crystalline rocks on the upthrown side of the range-bounding fault.

Basalt, remnants of valley flows, is found in isolated patches in the western and northwestern parts of the map area and in more extensive outcrop along the eastern margin. The basalt in the latter occurrence is probably equivalent to basalt flows in sec. 7, T. 9 S., R. 5 W., about 6 mi to the southeast, which yield a Pliocene whole-rock K-Ar age of 4.2 m.y.±0.2 m.y. (Marvin and others, 1974). Basalt now found in small patches in a 3-mi-long belt that extends from sec. 2, T. 8 S., R. 7 W. to sec. 19, T. 7 S., R. 6 W. appears to represent a separate flow (or flows) that filled a (then) northeast-trending valley, locally covering coarse talus of crystalline rocks. Throughout the map area the basalt is of similar character. Typical samples contain phenocrysts of olivine in a fine-grained matrix made up mainly of plagioclase laths and pyroxene granules.

### QUATERNARY DEPOSITS

Two types of surficial deposits are distinguished on the general map of the area (pl. 1): basin fill and stream alluvium. A third category, not separately mapped, consists of colluvial debris that mantles pediment surfaces on lower slopes of the range.

The basin deposits are largely confined to the eastern part of the map area, where they underlie a wide area of low relief in the upper reaches of the Sweetwater Creek drainage system. Exposures are very poor. The material revealed in some gully walls consists largely of unconsolidated, poorly bedded sand and gravel containing a substantial amount of volcanic ash. The thickness is not known, but probably is generally less than 100 ft. Deposition of these clastic deposits postdates late Tertiary movement on the Stone Creek and Carter Creek (extended) faults, which displace basalt of Pliocene age. The originally continuous upland (peneplain?) surface was displaced about 600 ft on the Carter Creek (extended) fault to form the Sweetwater basin and provide a site for clastic accumulation.

Stream deposits, consisting of sand and gravel augmented by slumped debris from valley walls, are found as narrow strips along a number of streams in the area. They are not regularly distributed, and some probably accumulated in local basins formed by landslide damming of streams.

## STRUCTURE

The main outlines of the Ruby Range and the pattern of the principal drainage systems reflect late Mesozoic and Tertiary movement on northeast- and northwest-trending faults. Within the range itself, however, the fabric is controlled mainly by a general northeast trend of layering, foliation, and fold axes, modified by north-trending crossfolds and by Middle Proterozoic movement on faults of the northwest-trending system.

Little is known in detail of the northeast-trending fault that bounds the range on the northwest. It is generally assumed to be a steeply dipping normal fault, northwest side down, on which there was recurrent movement during Tertiary and Quaternary time (Ruppel, 1982), but the possibility exists (J.M. O'Neill, written commun., 1987) that the fault was active in late Mesozoic (Laramide) time, with high-angle reverse movement, northwest side up. Movement evidently terminated prior to development of the present pediment surface, which truncates all rocks and rises gradationally from the Beaverhead basin toward the range crest. There is no evidence for Precambrian movement on this fault.

The northwest-trending fault system is of regional importance throughout eastern Idaho and southwestern Montana (Ruppel, 1982). The main faults of this system in the southwest Ruby Range—the Stone Creek, Carter Creek, Hoffman Gulch, and Elk Gulch faults—are ancient; all were active in Precambrian time, with consistent left-lateral displacements, measurable in thousands of feet, that occurred prior to emplacement of Middle Proterozoic diabase dikes. Movement on at least the Stone Creek and Carter Creek faults was later renewed, possibly in late Mesozoic (Laramide) time but certainly in the late Tertiary. As previously noted, the Carter Creek fault (extended) cuts a basalt flow of Pliocene age a few miles beyond the east margin of the map area; vertical displacement, northeast side down, is about 600 ft. Springs are common along both the Stone Creek and Carter Creek faults, both of which are marked by zones of silicification—silicified breccia on the Stone Creek fault and jasperoid on the Carter Creek fault. Much of this silicification postdates or is contemporaneous with late Tertiary movement on the faults, but some, such as jasperoid along the Carter Creek and Hoffman Gulch faults, could in part be older, possibly of Laramide or even of Precambrian age.

All of the strata comprising the crystalline core of the range, including the massive quartzofeldspathic gneiss, have been intensely deformed. Folds of several generations and orientations are evident, both in outcrop and (more particularly) in map patterns. The oldest identifiable structures are isoclinal folds that trend and plunge northeasterly and have axial planes that dip to the northwest. There is some evidence that this fold system is the product of more than one period of deformation, as indicated, for example, by refolded folds in the marble belt in the eastern part of sec. 12, T. 8 S., R. 7 W. and the adjoining sec. 6, T. 8 S., R. 6 W. Garihan (1979b) suggests that the initial structures were at least locally recumbent, a concept expanded by Karasevich and others (1981) to include formation of nappes.

The northeast-trending isoclines (including those of the marble belt noted above) are further deformed by crossfolds that trend and plunge to the north. The resultant structures are upright, tight to open folds, best exemplified by the major synform that centers in sec. 24, T. 7 S., R. 7 W. Deformation on this fold set was locally intense, as shown, for example, by strong rodding of quartzofeldspathic gneiss in sec. 12, T. 8 S., R. 7 W.

Other structural trends are less readily defined. Some apparent fold structures of odd orientation probably are simply geometric consequences of refolding of earlier folds, interference patterns rather than reflections of different stress fields. Broad arching, however, as in the terrane north of the Stone Creek fault and that between the Carter Creek and Hoffman Gulch faults, is a separate structural element of northwesterly trend; it reflects internal adjustments within fault blocks to the extensive left-lateral Precambrian movement on the northwest-trending faults.

The major uncertainty encountered in analysis of the Precambrian structural evolution of the southwestern Ruby Range concerns the reality and possible extent of an early epoch of low-angle deformation. Physical evidence bearing on the issue is scarce. It consists mainly of small-scale rootless folds locally present in layered gneiss, having axial planes parallel to layering and foliation, which can be interpreted as relict fragments of recumbent structures, and of certain map patterns, as in the belt of marble that crosses sec. 6, T. 8 S., R. 6 W. No widespread mylonitic zones that might identify detachment surfaces have been found, although as noted by Karasevich and others (1981), apparent absence could be due either to nonrecognition of a thin mylonite layer or to distribution of the cataclasis through a thicker zone of ductile shear. Lacking positive physical evidence, the argument for an early low-angle structural displacement in the southwestern Ruby Range rests on two principal bases:

(1) Anomalous field relations between major lithologic units, such as those of the Christensen Ranch Metasedimentary Suite, that indicate that the initial stratigraphic succession has been strongly disturbed, probably in large part by displacements at boundaries of competent lithologic units and by sequence reversals due to complete overturn of some sheets. As a consequence, the observed structures are no longer amenable to analysis based on stratigraphic consistency and continuity.

(2) Mounting evidence that such an event may have taken place on a regional scale. The conclusions reached by Karasevich and others (1981) calling for nappe structures in the northern Ruby Range have already been noted. Reid (1957) and Burger (1967) have proposed an initial epoch of recumbent folding for a similar assemblage of Precambrian strata in the Tobacco Root Mountains, as has Erslev (1983) for the southern Madison Range to the east.

On the basis of this rather fragile framework of fact and inference, it is concluded that the Precambrian strata of the area were first deformed by low-angle displacements, including at least local recumbent folding, then tightly compressed into the now-dominant northeast-trending isoclines, which are co-axial (but not coplanar) with the initial structures. A synthesis of the structural evolution of the area, placed in the context of total geologic history, is presented in a succeeding section of this report.

## METAMORPHISM

All of the Precambrian strata of the area, except pegmatites of Late Archean or Early Proterozoic(?) age and Middle Proterozoic diabase dikes, have been metamorphosed to upper amphibolite facies. Because of the great variety of rock types and the wide range of chemical compositions represented, the area is a virtual field laboratory for studies of metamorphism. Outstanding papers have been published on a number of different aspects: metamorphosed iron-formation (Immega and Klein, 1976), metamorphosed ultramafic rocks (Desmarais, 1981), mineralogy (Rabbitt, 1948; Ross and others, 1969; Dahl and Friberg, 1980), and geothermometry (Dahl, 1979, 1980). The discussion that follows draws largely from these sources, coupled with personal observation and use of the synthesis prepared by Karasevich and others (1981) for the Ruby Range as a whole.

### PROGRADE METAMORPHISM

The assemblages that were produced by metamorphism during the Late Archean orogeny (culminating age about 2,750 m.y.) can be assigned to the sillimanite–potassium feldspar zone on the basis of the association sillimanite–perthitic microcline in quartz-feldspar gneiss and schist and by the general absence of prograde muscovite. Kyanite is generally absent but has been preserved as a relict phase in some metapelite (Dahl, 1979). During the cycle of prograde metamorphism, ultramafic rocks were deserpentinized, then converted to assemblages containing olivine, orthopyroxene, magnesio-hornblende, and spinel (Desmarais, 1981). Mafic rocks were metamorphosed to amphibolite, composed chiefly of hornblende, andesine, quartz, and garnet; pyroxene is generally absent. Typical prograde assemblages in iron-formation consist of quartz, magnetite, and grunerite (cummingtonite), and variable amounts of specularite, riebeckite, hypersthene, and actinolite. Rock of dolomitic composition was converted, depending upon the initial content of sand and clay, to marble (containing variable amounts of forsterite, diopside, and tremolite), diopside-tremolite-phlogopite schist, and diopside-feldspar-quartz gneiss.

Appraisal of the ultramafic assemblage leads to an estimate of peak metamorphic conditions as being about 710 °C, 5–7 kbar pressure (Desmarais, 1981). Element distribution between garnet and pyroxene, garnet and biotite, garnet and hornblende, and alkali feldspar and plagioclase indicates values for peak conditions to be 645±45 °C, 6.2±1.2 kbar (Dahl, 1979, 1980). These temperature-pressure conditions—about 700 °C, 6 kbar—are typical of those generally assumed for amphibolite-facies metamorphism (Miyashiro, 1973, p. 90). The estimated temperature is about 50 °C lower than that estimated for the northern part of the Ruby Range, where rocks are in the granulite facies (Karasevich and others, 1981).

### RETROGRADE METAMORPHISM

All of the prograde assemblages have been modified to a greater or lesser degree by retrograde metamorphism: forsterite is altered to serpentine, pyroxenes to amphiboles and epidote, amphiboles to epidote and chlorite, biotite to chlorite and muscovite, feldspars to clinozoisite and sericite. Amphiboles in iron-formation show complex exsolution (Ross and others, 1969; Immega and Klein, 1976). Cordierite occurs locally as retrograde coronas on garnet; iron-magnesium distribution indicates a temperature of 545±50 °C at equilibrium (Dahl, 1979). Oxygen isotope fractionation between magnetite and quartz in iron-formation yields an estimated temperature of 475±25 °C (Dahl, 1979).

The significance of these various retrograde alterations is not entirely clear. Some of the phenomena (exsolution of amphiboles, cordierite coronas on garnet, and oxygen isotope distribution, for example) probably can be attributed to continued chemical and mineralogic reaction during the slow decline of temperature from peak conditions. Other effects, however, notably the widespread epidotization and hornblendization of diopside gneiss, are believed to represent a separate, later cycle of metamorphism superimposed on the earlier prograde assemblages. A reasonable assumption is that this later cycle occurred in response to a regional thermal rise that has been dated at about 1,650 m.y. (Giletti, 1966).

### OTHER METAMORPHIC EFFECTS

All of the Middle Proterozoic diabase dikes in the area have been incompletely but extensively altered to a greenschist assemblage. Since these dikes postdate the 1,650-m.y. thermal rise and inferred accompanying epoch of retrograde metamorphism, it is evident that the region was affected by a still later metamorphic cycle of at least moderate intensity. A late Precambrian (Late Proterozoic or Late Middle Proterozoic) age can be assumed, since Paleozoic rocks in the Ruby Range are not altered, and it is possible that the widespread deposits of talc were formed at this time.

## GEOLOGIC HISTORY

### ARCHEAN

The geologic history of the area prior to the Late Archean orogeny that culminated about 2,750 m.y. ago is known only in broad outline and at present entirely lacks geochronologic tie points. Elsewhere in the northern Rocky Mountains, mainly in the Beartooth Mountains of Montana and Wyoming, some elements of older geochronology have been established: Mueller and others (1985) present data that suggest that supracrustal rocks of the eastern Beartooth Mountains, which include quartzite and iron-formation, were metamorphosed to granulite facies about 3,400 m.y. ago; Page and Ziontek (1985) conclude that the Stillwater Complex (age about 2,700 m.y.) was emplaced in an iron-formation-bearing sedimentary sequence to which a minimum age of 3,270 m.y. is assigned; and Reid and others (1975) show that metasedimentary rocks of the North Snowy block, which include marble, have a minimum age of about 3,100 m.y. and were invaded by granite gneiss having an age of about 3,000 m.y., well before the Late Archean orogenic event. These data are fragmentary, but they do indicate a complex regional history of sedimentation, magmatism, and metamorphism that spans at least 700 m.y. of Archean time.

The Archean history of the Ruby Range can be expressed only in sequential terms, based on known or inferred relations between rock units. The reconstructed record is as follows (oldest to youngest):

1. Deposition of intermediate to mafic volcanic and volcaniclastic strata, together with interlayered graywacke-type sediments, to form the precursor sequence for the assemblage now classed as "older gneiss and schist." No basement is preserved, so the initial thickness is not known.
2. Emplacement and serpentinization of ultramafic plutons in the volcanic-sedimentary terrane, possibly as part of an ophiolite-like assemblage. (This inference is based on the restriction of the larger bodies of ultramafic rock to the older sequence and to the need for a source for the many smaller diapiric bodies of ultramafic rock now found dispersed through the younger quartzofeldspathic gneiss and the Christensen Ranch Metasedimentary Suite.)
3. Continued deposition of supracrustal strata, probably conformable but changing in character to more felsic volcanic rocks interbedded with potassium-rich pelite and arkose. This assemblage probably attained a thickness of at least several miles.
4. On the basis of the record emerging for Archean history elsewhere in southwestern Montana, it is postulated (though no direct evidence can be cited) that the felsic assemblage noted in (3) was converted by metamorphism and magmatic additions to a sheet of quartzofeldspathic gneiss of regional dimensions, prior to deposition of the strata of the Christensen Ranch Metasedimentary Suite. This postulated event has yet to be dated by isotope geochronology.
5. Deposition of the strata now represented by the Christensen Ranch Metasedimentary Suite in a shallow marine environment on a basement of quartzofeldspathic gneiss. The basal beds probably were arkosic and subsequently converted to granitic gneiss not now distinguishable from the basement gneiss. These strata were overlain by dolomite, probably as a sheet of variable thickness but of regional extent, which was succeeded by a varied sequence of shale, dolomitic sandstone and shale, sandstone, and iron-formation that probably contained one or more repetitions of dolomite of more local distribution.
6. Uplift and low-angle deformation of the sedimentary pile, possibly by gravity sliding, including development of recumbent folds and of detachment surfaces, particularly at the base of the sedimentary sequence and at boundaries of thicker homogeneous lithologic units (such as dolomite marble) within the pile, and probable formation of one or more nappes. Original stratigraphic relations within the sedimentary sequence were extensively distorted and in places reversed.
7. Deep burial of the sedimentary terrane, followed by widespread intrusion of mafic magma, mostly as sills of diabase and gabbro.
8. Onset of the thermo-tectonic cycle that culminated about 2,750 m.y. ago. This orogenic event, or series of events, contained as a minimum the following elements (oldest to youngest):
   a. Intense compressional deformation on northeast-trending axes, to form northwest-dipping isoclines in all rock units, including the quartzofeldspathic gneiss. Ultramafic rock derived from initial sites in the "older gneiss and schist" unit was redistributed as diapiric pods and lenses in the quartzofeldspathic gneiss and overlying metasedimentary rocks. Some bodies of mafic rock also were deformed plastically, particularly along fold axes. The folding was accompanied, probably contemporaneously, by metamorphism of all strata to amphibolite facies, and by generation and local emplacement of granitic magma. Parts of the older quartzofeldspathic gneiss probably were remobilized at this time.
   b. Crossfolding of the northeast-trending isoclines on north-trending axes, producing upright folds and the complex geometry now evident in areal distribution of lithologic units.
   c. Emplacement of "older" pegmatite as unzoned bodies and sheets that transect the older folded strata.
   d. Local resumption of mafic magma intrusion. This epoch is represented certainly by only one dike in the mapped area, but it is well established in the northern Ruby Range and in the adjoining Tobacco Root Mountains.
   e. Remetamorphism of the entire area, again to amphibolite facies. This late thermal event was not accompanied by significant deformation; the lone late diabase dike recognized, now altered to amphibolite and moderately foliated, retains its straight cross-cutting form.

This Archean record, tabulated above, has much in common with certain other areas of similar Precambrian rocks. A notable analogy exists between this area and, for example, the Adirondack Mountains of New York. Though younger by considerably more than a billion years, the Adirondack lithic assemblage is similar, containing marble, amphibolite, and various layered gneisses, as well as widespread deposits of talc. The structural history, like that inferred for the Ruby Range, involves early low-angle deformation, including formation of nappes, followed by formation of co-axial folds that were subsequently deformed on one or more sets of crossfolds (McLelland and Isachsen, 1980).

### PROTEROZOIC

The geologic record of the approximately 2 billion years of Precambrian time that followed the major orogeny of the Late Archean is extremely sparse. Isotopic evidence developed by Giletti (1966) documents a regional thermal rise in Precambrian terranes of southwestern Montana west of the Gallatin River about 1,650 m.y. ago. It is probable that this also marks the time of widespread retrograde metamorphism of older Precambrian rocks of the Ruby Range and the emplacement of tourmaline-bearing zoned pegmatites. The only other rock-forming events known during Proterozoic time were the emplacement of diabase dikes about 1,425 m.y. ago and the formation of talc deposits.

This meager record of events is summed up as follows, oldest to youngest:

1. Following close of the Late Archean orogeny, the region remained structurally quiescent for approximately 1 billion years.
2. Sometime late in the Early Proterozoic and culminating about 1,650 m.y. ago, the rocks of the Ruby Range area were subjected to a general rise in temperature. This resulted in extensive retrograde metamorphism and emplacement of scattered small bodies of pegmatite, generally zoned. No firm quantitative data are available to indicate the temperature-pressure conditions during this event.
3. Immediately preceding (and possibly in part contemporaneous with) diabase dike intrusion at about 1,425 m.y., the Precambrian crystalline rocks were offset on steeply dipping northwest-trending faults. Movement was left lateral, individual horizontal displacements being as much as 1 mi.
4. Emplacement of diabase dikes, age about 1,425 m.y., in fractures clearly related structurally to the northwest-trending fault system.
5. For the remainder of Proterozoic time, the area remained structurally stable, probably as an upland bordering the site of deposition of the Belt Supergroup and later Proterozoic strata to the north and northwest. Within the Ruby Range area, the only recognized events were incomplete metamorphism of the diabase dikes and the formation of talc deposits, neither of which can be dated precisely.

### PHANEROZOIC

The post-Proterozoic history of southwestern Montana is complex and has been ably documented in many studies (for example, Klepper, 1950; Scholten, 1968; Tysdal, 1976b; Ruppel, 1982) and will be reviewed only briefly here.

At the beginning of Paleozoic time, the Ruby Range area was still an upland. By mid-Cambrian time, however, the entire region was covered by a shallow marine sea that persisted, with some breaks, into Mesozoic time. In the vicinity of the present Ruby Range, Paleozoic strata accumulated to an aggregate thickness of about 5,000 ft (Tysdal, 1976b). In the northern part of the range, these strata are well preserved in downdropped fault blocks (Tysdal, 1976a), but they have been stripped completely from the crystalline basement in the southwestern Ruby Range.

Sedimentation, in part in nonmarine continental basins of limited extent, continued through the Mesozoic and into Cenozoic time in much of southwestern Montana. In the vicinity of the Ruby Range, Paleozoic strata are unconformably overlain by the conglomeratic Beaverhead Group. This clastic deposit marks the early stages of uplift of basement blocks on newly formed northeast-trending faults. Continued movement on these faults and on rejuvenated northwest-trending faults produced intermontane basins, which became sites of accumulation for thick clastic deposits of Eocene and younger age. Uplift of basement blocks culminated, in a regional sense, in late Tertiary time (Ruppel, 1982).

The final recorded structural event in the southwestern Ruby Range was movement on the rejuvenated Stone Creek fault and Carter Creek fault, which produced a 600-ft vertical offset in a basalt flow of Pliocene age. The Sweetwater Basin, on the downdropped side of the fault, was later covered to a shallow depth with clastic materials; these deposits, together with the alluvium along some streams, constitute the youngest geologic materials in the area.

## MINERAL RESOURCES

Although the prime objective of this field study is the description of the extent and geologic setting of the bedded iron deposits, the occurrence of other mineral resources in the map area will also be reviewed in summary fashion. The only mineral resource being extracted as of 1987 is talc, but in the past many other types of deposits were explored or investigated: graphite, nickel, pegmatite minerals, corundum, asbestos, sillimanite, and base-metal sulfides.

### IRON

As previously described, banded iron-formation was deposited in at least two and probably several stratigraphic levels in the initial sedimentary sequence that now makes up the Christensen Ranch Metasedimentary Suite. The principal deposits, however, are believed to represent a single stratigraphic unit that generally ranges in true thickness from about 40 ft to 100 ft. The major area of economic interest and possible development, known as the "Carter Creek deposit," is a fold belt approximately 2.5 mi long and several hundred feet wide between the Hoffman Gulch and Carter Creek faults, mainly in secs. 3, 9, and 10, T. 8 S., R. 7 W. Elsewhere, particularly in the structural block between the Carter Creek and Stone Creek faults, iron-formation probably at the same initial stratigraphic position can be traced continuously in belts several miles long, but these lack the structural duplication necessary to produce volumes of iron-formation worthy of serious economic consideration.

The distribution and structure of the iron-formation that makes up the Carter Creek deposit are shown in figure 6, which is based on the previously published detailed map of the area (James and Wier, 1972b). The general structure of this belt is that of an overturned sequence tightly compressed into a number of northeast- to east-trending, northwest- to north-dipping isoclines. In a few places, however, notably in sec. 9, the iron-formation is contained in open gentle folds that are approximately co-axial (but not co-planar) with the isoclines.

The iron-formation is bounded on the north and northwest by stratigraphically lower mica schist and quartzite and on the south and southeast by a thin bed of mica schist that gives way to epidote-diopside-hornblende gneiss containing distinctive widely spaced thin layers or laminae of pink calcite marble. Both upper and lower contacts of the iron-formation are relatively abrupt, and there is little interbedding with rock above or below. This overturned stratigraphic succession—older mica schist and quartzite, iron-formation, and stratigraphically younger schist and gneiss—is well exposed in the NW¼NE¼ sec. 10 (see cross-section *A–A'*, fig. 6).

Physically and chemically, the iron-formation is amenable to treatment as a low-grade iron ore (taconite). Grain sizes are relatively coarse, generally in the range 0.1–1.0 mm, which permits separation of magnetite and specularite from gangue minerals without excessively fine grinding. Bulk chemistry is much like that of other Archean iron-formation of similar mineralogic facies and of the iron-formation of the Mesabi district of Minnesota, as shown by comparative data in table 6. It differs from most commercially processed taconite in the relatively high ratio of ferric to ferrous iron (nearly 3:1; table 2), reflected mineralogically by the presence of specularite in addition to magnetite and of riebeckite, the ferric-ferrous sodic amphibole. Also notable in the chemistry of the iron-formation of the Carter Creek deposit is the low content of manganese and the relatively high phosphorous.

Beneficiation tests were run on bulk samples by the U.S. Bureau of Mines, using magnetic separation. The results on three runs, using ball-mill grinding (presumably to -100 mesh) and wet magnetic separation, are summarized as follows (from Holmes and others, 1962, p. 10):

| | | | |
|---|---|---|---|
| Test number | 1 | 2 | 3 |
| Quantity, tons | 22 | 34.7 | 72.7 |
| Initial Fe content, percent | 31.5 | 31.9 | 31.9 |
| Fe recovery, percent | 88.1 | 85.9 | 84.0 |
| Concentrate: | | | |
| Fe, percent | 55.9 | 57.5 | 61.1 |
| $SiO_2$, percent | 7.15 | 16.6 | 12.6 |

The concentrate assays compare unfavorably with those of pelletized concentrates now being used in iron and steel plants in the United States, which typically contain 63–64 percent Fe and about 5 percent $SiO_2$. Additional laboratory-scale tests on concentrate by the U.S. Bureau of Mines, involving regrinding to -325 mesh and reprocessing, resulted in a product containing 69.8 percent Fe, at 97.2 percent recovery. In appraising the results of these beneficiation tests, it is to be noted that specularite, a significant constituent in some parts of the iron-formation, cannot generally be recovered by magnetic methods alone.

Exploration of the Carter Creek deposit, which had been known to geologists since 1948 (Heinrich, 1960), began in 1956 (DeMunck, 1956) and has continued intermittently since that time. Most of it has been done under the auspices of the Minerals Engineering Company, and (later) Steel Alberta, Ltd., of Canada. Early churn drilling and trenching was followed by core drilling. Information on the amount of core drilling done is incomplete, but it is known to be in excess of 10,000 ft in aggregate. Locations of trenches and of diamond drill holes sunk in the late 1950's are shown on the previously published detailed map of the area (James and Wier, 1972b).

Informal appraisals of reserve tonnage have been made by a number of investigators; estimates range from a few tens of millions to several hundred million tons. On the basis of surface area of exposure and projection to a depth of 300 ft (a possible practical limit for open-pit mining), the resources of potential low-grade ore in the Carter Creek deposit, excluding that contained in isolated minor synclines, are here estimated to be about 95 million long tons, containing 28–29 percent recoverable (that is, nonsilicate) iron. Of this, slightly less than two-thirds is in Beaverhead County and slightly more than one-third is in Madison County.

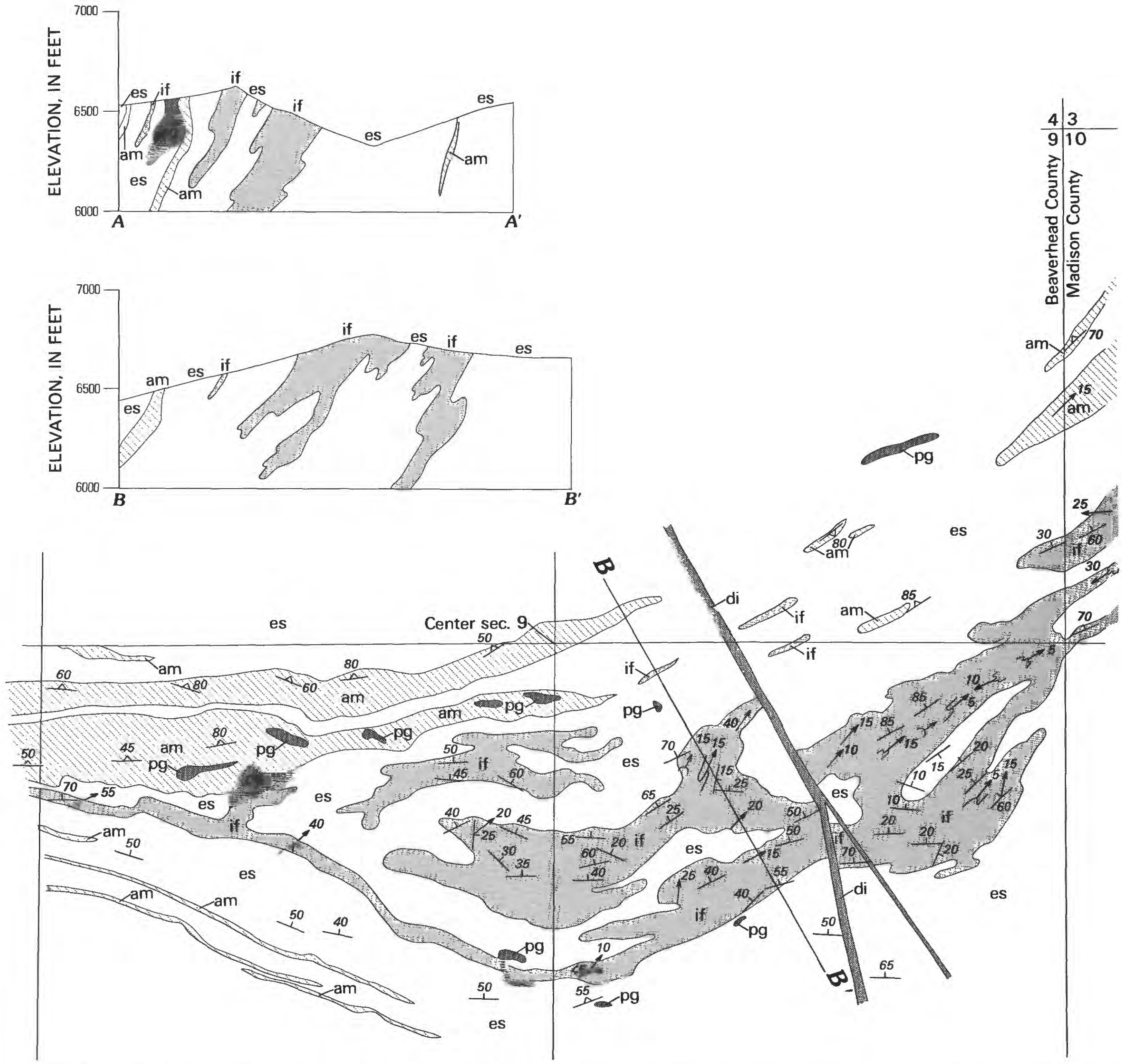



**FIGURE 6.**—Geologic map of the Carter Creek iron deposit, Madison and Beaverhead Counties, Mont. Map covers secs. 3, 9, and 10, T. 8 S., R. 7 W.

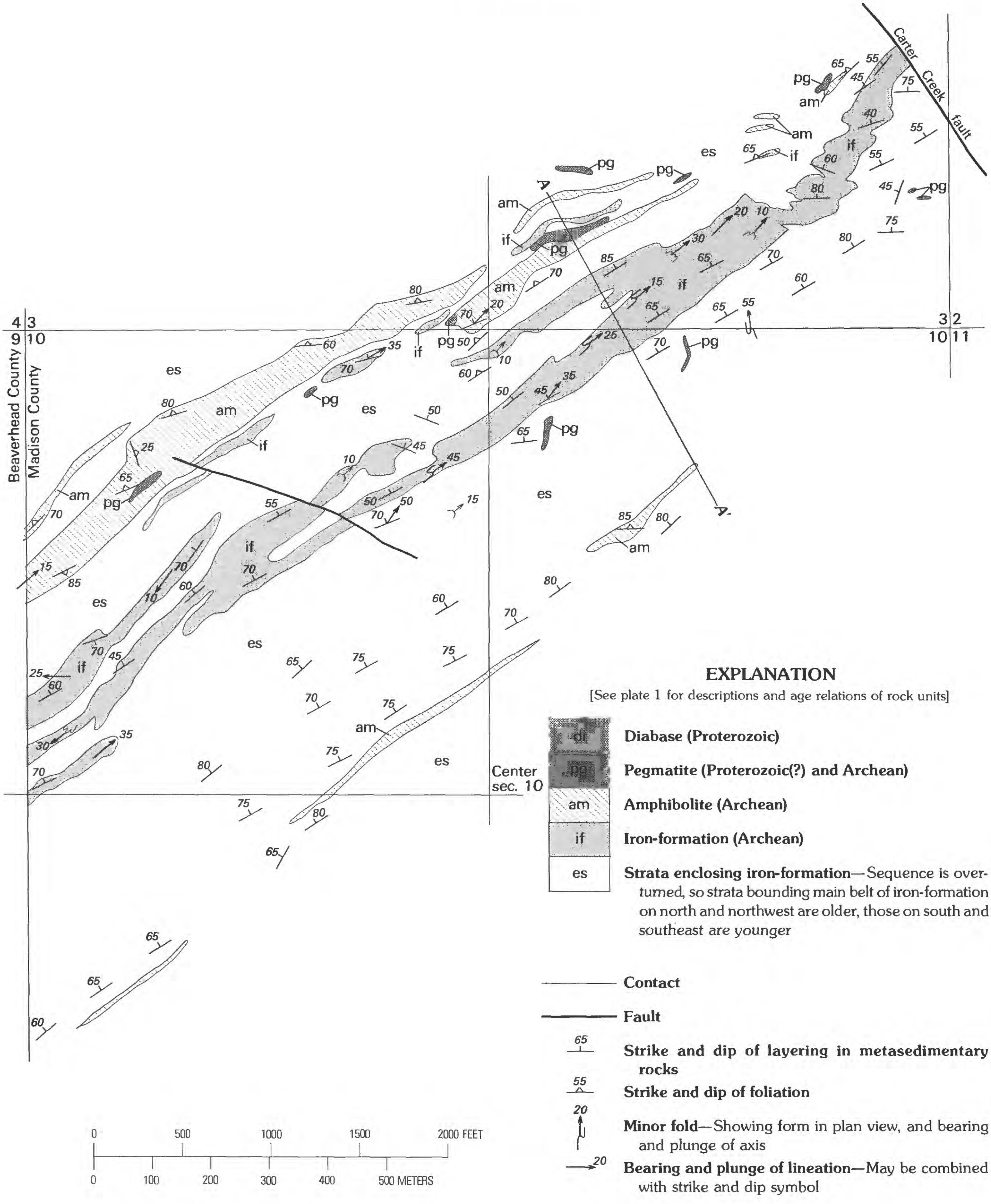

Carter Creek fault
A
A'
Beaverhead County
Madison County
4 3
9 10
3 2
10 11
Center sec. 10
es
am
if
pg
EXPLANATION
[See plate 1 for descriptions and age relations of rock units]
di Diabase (Proterozoic)
pg Pegmatite (Proterozoic(?) and Archean)
am Amphibolite (Archean)
if Iron-formation (Archean)
es Strata enclosing iron-formation—Sequence is overturned, so strata bounding main belt of iron-formation on north and northwest are older, those on south and southeast are younger
Contact
Fault
65 Strike and dip of layering in metasedimentary rocks
55 Strike and dip of foliation
20 Minor fold—Showing form in plan view, and bearing and plunge of axis
20 Bearing and plunge of lineation—May be combined with strike and dip symbol
0 500 1000 1500 2000 FEET
0 100 200 300 400 500 METERS

TABLE 6.—*Chemistry of the iron-formation of the Carter Creek deposit, southwestern Ruby Range, Mont., compared with other iron-formations*

| | 1 | 2 | 3 |
|---|---|---|---|
| $SiO_2$ | 50.04 | 49.07 | 50.62 |
| Fe (total) | 31.77 | 31.65 | 30.84 |
| Mn | 0.04 | 0.42 | 0.46 |
| $P_2O_5$ | .36 | .16 | .09 |

Sample data:

1. Iron-formation of the Carter Creek deposit. Average of analyses 1-4, table 2.
2. Archean iron-formation of the Yilgarn block, western Australia (Gole and Klein, 1981, p. 176).
3. Biwabik Iron-formation (Proterozoic), Mesabi district, Minnesota. Recalculated on an $H_2O$-free and $CO_2$-free basis from previously published data by Gole and Klein (1981, p. 176).

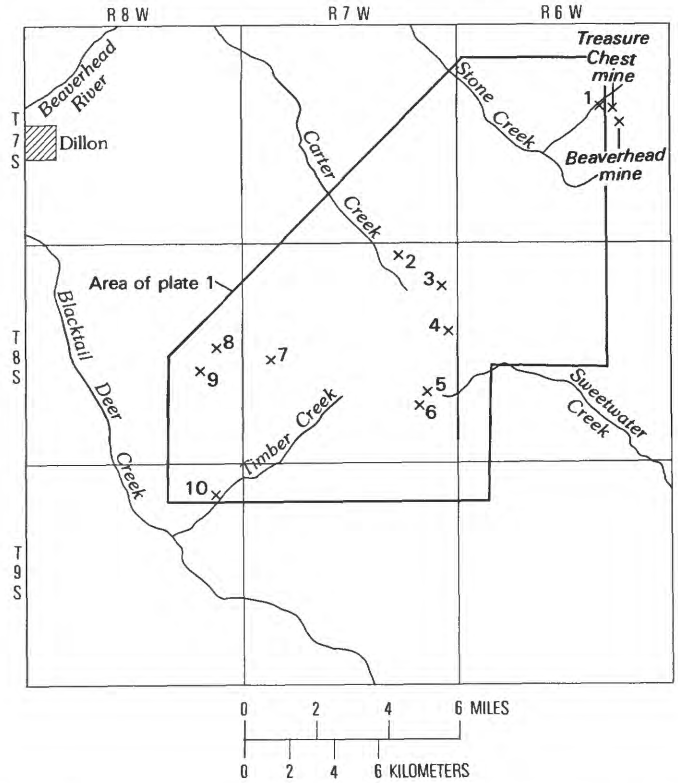



FIGURE 7.—Location of principal talc mines and prospects in the southern Ruby Range. 1, Treasure State; 2, Regal; 3, American Chemet; 4, Sweetwater; 5, Sauerbier; 6, Owens-McGovern; 7, Bozo-Zobo; 8, Banning-Jones; 9, Smith-Dillon; 10, Crescent.

## TALC

Talc seams, veinlets, and lenses occur in dolomite marble throughout the area, and deposits have been mined sporadically for more than 40 years. Locations of larger known deposits that have been explored or mined in the southern Ruby Range are shown in figure 7. The two most recently active mines in the region, the Treasure Chest mine and the Beaverhead mine, are just outside the area covered by the general geologic map (pl. 1). Details on individual deposits are given by Perry (1948), Okuma (1971), and Garihan (1973, 1974), and a comprehensive review of the occurrence and development of talc resources of southwestern Montana is provided by Olson (1976).

The talc of the area is cryptocrystalline, opaque to translucent, and generally white to pale green or olive gray. Contacts with the dolomitic host rock typically are sharp, and bodies tend to be elongate in the plane of bedding in the marble. Impurities generally are scarce, but small flakes of graphite, relict from the replaced marble, are common in a few deposits, and limonite derived from oxidation of pyrite is a minor additional deleterious constituent in some. Other associated minerals noted include serpentine (locally abundant), chlorite, quartz, and phlogopite. In a few deposits the host marble is coarsely recrystallized, but whether this coarsening of grain was penecontemporaneous with talc formation or whether it was formed earlier is not clear.

Much of the talc that has been tested is of "steatite" grade, specifications for which, according to Olson (1976), call for less than 1.5 percent CaO, less than 1.5 percent $Fe_2O_3$, less than 4.0 percent $Al_2O_3$, and only minute quantities of other impurities. Analyses of selected samples are given in table 7.

The lithologic control for localization of talc is absolute: all deposits are in dolomite marble. Structural control is less certain. Many of the larger deposits (for example, those at the Treasure State and Treasure Chest mines, the Beaverhead mine, the American Chemet prospect, and the main pit of the Sweetwater prospect) are in narrow bands of dolomite marble wholly enclosed in older quartzofeldspathic gneiss; locally, as at the Treasure State and Treasure Chest mines, virtually the entire marble unit has been replaced. Some of these marble bands are demonstrably synformal, as at the American Chemet and Sweetwater deposits; others may represent structurally emplaced tectonic slices. The Regal (Keystone) deposit, however, is in the axial zone of a synform of marble enclosed in schist, and

TABLE 7.—*Analyses of talc from deposits in the southwestern Ruby Range, Mont.*

[From compilation by Olson, 1976. tr, trace; n.r., not reported; <, less than]

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| $SiO_2$ | 62.25 | 62.06 | 61.78 | 57.72 | 60.40 |
| $Al_2O_3$ | 0.27 | 0.50 | 0.57 | 1.13 | 1.91 |
| $Fe_2O_3$ | .71 | .67 | .75 | 0.48 | 0.27 |
| CaO | <.05 | <.05 | <.05 | 1.34 | .80 |
| MgO | 31.13 | 31.12 | 31.06 | 30.72 | 30.81 |
| $K_2O$ | .02 | .01 | .02 | tr | .14 |
| $Na_2O$ | .06 | .07 | .08 | tr | .20 |
| $H_2O^-$ | .14 | .18 | .17 | n.r. | n.r. |
| $H_2O^+$ | 5.03 | 5.09 | 5.17 | 5.94 | 5.15 |
| $CO_2$ | .00 | .00 | .00 | n.r. | n.r. |

Sample data:

1, 2, 3. Smith-Dillon mine. Analysis by Leonard Shapiro.

4. Keystone (Regal) mine. Analysis by Raymond G. Osborne Laboratories, Inc.

5. Treasure (Chest) mine. Analysis by Raymond G. Osborne Laboratories, Inc.

the Smith-Dillon deposit is in marble near a structural footwall of quartzite and granite gneiss. Aside from the Regal deposit, most talc bodies of significant size are in marble close to, or along a contact with, a chemically and structurally dissimilar rock type, such as quartzofeldspathic gneiss. Larger masses of marble, such as those occupying extensive tracts in the north-central part of the map area, are notably deficient in talc deposits.

The process of talc formation in dolomite involves the introduction of $SiO_2$ and $H_2O$ and loss of CaO and $CO_2$, presumably through the agency of hydrothermal fluids. The source of the fluids and the timing of the event remain unclear. Some evidence of relative age is potentially available at the Regal mine, where the talc zone is crossed by an undeformed diabase dike of Middle Proterozoic age. The actual contacts are not exposed, however, so age relations are not now determinable. It is to be noted that dikes of this group, though structurally undisturbed, have been affected by retrograde metamorphism, typically resulting in formation of hydrated minerals such as chlorite. This alteration provides evidence, therefore, for post-dike introduction of low-temperature aqueous solutions on a regional scale. In the absence of evidence to the contrary, it is here suggested that the talc deposits are an additional product of this epoch of regional retrograde metamorphism, which occurred in later (post-1,425 m.y.) Precambrian time.

## GRAPHITE

Graphite, in the form of dispersed fine flakes, is a common minor constituent of marble throughout the area, and it is present as an impurity in some of the replacement talc deposits. The graphite deposits that have been of economic interest, however, are entirely different in character; these are monomineralic pods, lenses, irregular veins, and disseminations in quartzofeldspathic gneiss and pegmatite, found in a belt several hundred feet wide that trends northeasterly through the N½ sec. 31, T. 8 S., R. 7 W. into the SE¼ sec. 30 and the SW¼ sec. 29 of the same township. Deposits in this belt, where graphite was first discovered in 1899, were mined intermittently from 1902 to 1945. They have been described in a number of geologic reports (Winchell, 1910, 1911, 1914; Bastin, 1912; Perry, 1948; Ford, 1954, and Heinrich, 1960), and the Crystal Graphite mine in sec. 31, T. 8 S., R. 7 W. has been mapped in detail (Armstrong and Full, 1946; Armstrong, 1950).

The principal deposits are in a zone that extends southwesterly from a tight infold of marble, a structure that trends N. 60 E. and plunges northerly at about 45°, with closure in the SE¼ sec. 30, T. 8 S., R. 7 W. Lesser deposits occur on the south flank of the isoclinally infolded marble. The graphite, locally bladed and in radiating clusters, forms discontinuous steeply dipping pods and seams that transect both quartzofeldspathic gneiss and Late Archean pegmatite. Individual bodies typically are less than 1 ft thick (average, about 4 in.) and are traceable laterally and vertically for distances rarely exceeding a few tens of feet. The Crystal Graphite mine exploited such deposits by means of a shaft and several adits, over a vertical range of about 340 ft. The Bird's Nest mine, from which the first graphite of the area was produced in 1902, is about 0.75 mi northeast of the Crystal Graphite mine; it is in gneiss and pegmatite on the south flank of the marble isocline. Development, all of which was done more than 80 years ago, was from three adits, the longest of which was driven about 270 ft.

The origin of the deposits is somewhat problematical. They clearly are epigenetic, introduced into fractures in gneiss and pegmatite, the latter of Late Archean age. Ford (1954) concluded that the deposits were "epithermal," a view properly rejected by Heinrich (1960), who suggested instead that the mineralization was a process that began "in late pegmatite time." On the basis of isotopic analyses of the carbon, Weis and others (1981) concluded that these deposits, like those of similar character but greater extent in Sri Lanka, originated by metamorphism and transfer of pre-existing syngenetic graphite or carbonaceous detritus. Expressed in the usual per mil terminology—$^{13}C$ content in parts per thousand, relative to the Peedee belemnite standard ($^{12}C/^{13}C$ = 88.99) taken as zero—Montana graphite (four samples) ranges from -4.9 to -6.1 per mil, and Sri Lanka graphite (three samples) has somewhat similar values, -8.0 to -8.6 per mil. These assays of $^{13}C$ are considerably higher than those of reduced organic matter in rocks of lower metamorphic grade, which typically are in the range -20 to -25 per mil, and lower than that of marine carbonate, generally ranging between -1 and +2 per mil. Metamorphism of reduced organic matter is known, however, to result in $^{13}C$ enrichment, and in fact Rumble and Hoering (1986) have shown that in some areas the $^{13}C$ content of graphite spans virtually the entire range between that of reduced organic matter and of marine carbonate. The process by which initial organic material is mobilized and redeposited as monomineralic pods and veins is not entirely clear. Weis and others (1981) suggest that the carbon is mobilized by the "water gas reaction," $C+H_2O \rightleftharpoons CO+H_2$, transported as carbon monoxide and precipitated by the "Boudouard reaction," $2CO \rightleftharpoons C+CO_2$. Rumble and Hoering (1986), noting the rarity of CO and $H_2$ in estimated compositions of metamorphic fluids, propose a more complex process, one involving interaction of $CO_2$-rich fluids derived from metamorphism of impure carbonate with $CH_4$-rich fluids derived from organic matter. Graphite is precipitated when fluids of different $CO_2/CH_4$ ratios are mixed, dominantly by the reaction $CO_2+CH_4 \rightleftharpoons 2C+2H_2O$. The isotopic composition of the precipitated carbon would vary widely according to the particular fluid mix and the prevailing temperature-pressure conditions. The argument presented by Rumble and Hoering (1986) is persuasive, but the conclusion remains speculative. Regardless of process details, however, it can be concluded that the Montana deposits were formed by mobilization of carbon derived largely from what was originally organic detritus, followed by fluid transfer and redeposition as graphite under conditions of elevated temperature such as can be expected to have prevailed in the late stages of regional metamorphism and magmatism in Late Archean time.

## OTHER MINERAL DEPOSITS

As noted at the beginning of this section, deposits of a number of materials other than iron, talc, and graphite have been investigated for economic potential, generally without success. Most have been described by Heinrich (1960; also 1949a, 1949b, 1950a, 1950b), and they are reviewed briefly in the following paragraphs.

### NICKEL

The local presence of annabergite, the nickel arsenate, as yellow-green crusts and fracture fillings in ultramafic rock exposed in the southeastern part of the map area led to staking of mineral claims and to some physical exploration (Sinkler, 1942). The explorations (located in sec. 31, T. 8 S., R. 6 W.; sec. 36, T. 8 S., R. 7 W.; secs. 1 and 2, T. 9 S., R. 7 W.; and sec. 6, T. 9 S., R. 6 W.) consisted mostly of shallow test pits but also included at least two holes drilled to depths of about 250 ft (Desmarais, 1978). The ultramafic rocks of the area are described in some detail by Heinrich (1963) and by Desmarais (1978, 1981).

The economic significance of the supergene annabergite is minimal. Olivine in the parent ultramafic rock does contain as much as 0.53 NiO (Desmarais, 1981), but this is within the normal range for olivine in alpine-type peridotite.

### CORUNDUM

Corundum is found as a constituent of schist at two localities in the area. The principal deposit, discovered and described in some detail by Heinrich (1950b), is in sec. 36, T. 8 S., R. 8 W.; it consists of schist containing 5-35 percent corundum in an outcrop belt about 220 ft long and 20 ft wide and is not considered to have any economic value (Heinrich, 1960). The second deposit, in sec. 20, T. 8 S., R. 6 W., consists of scattered crystals of lilac-colored corundum (sapphire) in mica schist. The locality is immediately adjacent to the Sweetwater road and until fenced off in recent years was a favorite collecting site for amateur mineral collectors, who succeeded in excavating a number of irregular pits in the weathered schist.

The corundum deposits of the area are not unique in southwestern Montana. At least three localities have been explored as potential sources of corundum for abrasive uses (Clabaugh and Armstrong, 1951). The corundum-bearing rocks at these sites, as in the southwestern Ruby Range, are within sequences of metasedimentary strata and probably owe their origin to high-grade metamorphism of alumina-rich shale.

### ASBESTOS

At a number of places in the area, golden-yellow cross-fiber asbestos (chrysotile) is found as thin veinlets in altered dolomite marble adjacent to crosscutting diabase dikes of Middle Proterozoic age. Test pits, notably in secs. 35 and 36, T. 7 S., R. 7 W., show the veinlets to be irregular, generally less than 1 in. wide, and rarely traceable for more than a few feet. None of the localities explored appears to have economic potential. They are, however, remarkably similar in geologic occurrence to deposits that have been worked in the southern part of the Madison Valley, where chrysotile veins are in marble of the Cherry Creek Group at contacts with diabase (Heinrich and Rabbitt, 1960).

#### PEGMATITE MINERALS

Prospect pits have been sunk in many pegmatite bodies in the area, but the objectives of the testing are not always clear; some prospectors may have been attracted simply by the abundance of coarse quartz. All of the pegmatites contain quantities of feldspar, commonly in coarse-grained masses, that in some circumstances might have economic value, but a more likely economic target may have been muscovite mica, which in a few pegmatites is found as books as much as several inches in diameter. The quantity of mica in the explored pegmatites, however, is far too little to warrant consideration of further development. Rose quartz, common in the core areas of zoned pegmatites, may also have been an exploration target, but again the quantities are economically insignificant.

#### BASE METALS

Throughout the area, prospectors have dug test pits and, in places, short adits and shafts on the basis of surface showings of secondary iron and copper minerals (limonite, hematite, malachite, azurite, and chrysocolla) and in a few places have uncovered sulfide-bearing materials. Most, but not all, mineralized rock is along structural breaks related to the northwest-trending fault system. Explorations are particularly common along the Stone Creek fault, where at least two short adits have intersected siliceous breccia containing disseminated pyrite and chalcopyrite. Test pits and a short inclined shaft in sec. 13, T. 7 S., R. 7 W. are located on easterly trending minor faults that cut the diopside gneiss bedrock; dump material consists of vein quartz and coarse carbonate containing chalcopyrite and bornite.

Elsewhere in the area, exploration in the NW¼ sec. 1, T. 8 S., R. 7 W. has exposed a thin vein cutting layered schist and amphibolite; the vein strikes N. 45 W. and consists of disseminated bornite and chalcopyrite in vuggy coarse-grained quartz. About 0.5 mi north-northwest of this locality, in the SW¼ sec. 36, T. 7 S., R. 7 W., dump materials from prospect pits in dolomite include vein quartz. Also present as coatings and fracture fillings in dolomite are malachite and an earthy pink mineral identified as erythrite (cobalt bloom); these minerals evidently are of supergene origin but the nature of the primary mineralization is not known.

## KELLY IRON DEPOSIT, NORTHEASTERN RUBY RANGE

### INTRODUCTION

The Kelly iron deposit is on the northeast flank of the Ruby Range, 3-4 mi southwest of the town of Alder, and about 10 mi northeast of the southwestern Ruby Range map area. The locality, mostly in sec. 25, T. 6 S., R. 5 W., is at elevations between 6,000 and 7,000 ft. It is accessible from the adjacent Ruby Valley by way of an unimproved dirt road that follows Beatch Canyon and an access track, now largely impassable, that enters the area of exploration. Mining claims were staked in 1957 by John Kelly of Alder and subsequently leased to the F and S Contracting Company of Butte. Exploration consisted of a number of deep trenches and cuts and two inclined diamond drill holes having an aggregate length of 1,142 ft.

The topography and geology of the area were mapped by the U.S. Geological Survey in 1960 at a scale of 200 ft to 1 in. and surveyed with a tripod-mounted magnetometer (James and Wier, 1961). Later, a revised geologic map was published (James and Wier, 1972a), on which the map in this report, plate 2, is largely based.

### GENERAL GEOLOGY

Precambrian strata that contain iron-formation are exposed in a triangular area of a few square miles on the flank of the Ruby Range, bounded by strata of Paleozoic and Tertiary age. The western boundary of the Precambrian block is a high-angle fault of large displacement, which separates the Precambrian from limestone of the Madison Group of Mississippian age. The north margin is an unconformable contact with strata of Cambrian age. On the east, beyond the boundary of the map area, the Precambrian is bounded by downfaulted strata of Tertiary age.

Most of the area covered by plate 2 is underlain by strata of the Christensen Ranch Metasedimentary Suite, here metamorphosed to granulite facies. These strata, in apparent normal stratigraphic order, are folded into an upright southeasterly plunging syncline, the buried axial part of which has been buckled and squeezed upward to form an anticlinal diapir-like structure cored by a mass of ultramafic rock. The youngest

Precambrian rocks of the area are small bodies of microcline pegmatite of either Late Archean or Early Proterozoic(?) age, and at two localities bodies of moderately foliated quartz diorite of probable Early Proterozoic age.

## STRATA OF THE CHRISTENSEN RANCH METASEDIMENTARY SUITE

These strata have been grouped for map display and text discussion into four principal units, as shown in figure 8.

### DOLOMITE MARBLE

The dolomite marble that is the basal unit of the metasedimentary sequence has a minimum thickness of 400 ft. The marble is underlain by quartzofeldspathic gneiss, which though not exposed in the map area is abundantly present nearby (Karasevich, 1981). The location of the nearest contact suggests a possible thickness of marble in this area of 1,000 ft or more.

Of the exposed section in the map area, the lower half is coarsely crystalline white dolomite marble, best displayed on the south-facing slope of Taylor Canyon below the Precambrian-Cambrian unconformity. The upper half is gray crystalline dolomite marble studded with greenish-yellow ovoids of serpentine. The ovoids, 0.5–3.5 mm in diameter, locally contain relict islands of forsterite in a matrix of antigorite. Blades of pale-brown phlogopite, generally heavily altered to talc, are common both adjacent to the serpentine ovoids and as separate flakes dispersed through the marble.

Chemical analysis of the serpentine (forsterite) marble, given in table 8, shows nearly equal amounts of MgO and CaO, testifying to the dolomitic composition of the initial carbonate. X-ray measurement on several samples reveals considerable range in present dolomite:calcite (43:57–100:0).

### GARNET QUARTZITE AND GNEISS

The lower half of this 300-ft sequence is not well exposed but appears to consist mainly of micaceous quartzite and hornblende-diopside gneiss. The upper half consists of quartzitic strata containing distinctive layers of garnet-rich quartzite that grades into quartz-feldspar-garnet gneiss, locally including microcline-rich veins of pegmatitic aspect. The garnet, which comprises about 20 percent of some layers, is present as reddish-brown clusters of irregular outline. Chemical analysis of the garnet quartzite (see table 8) shows a relatively high content of FeO, reflected mineralogically by the abundance of garnet and presence of magnetite as a common accessory. The rock can be assumed to have originated as an iron-rich feldspathic sandstone.

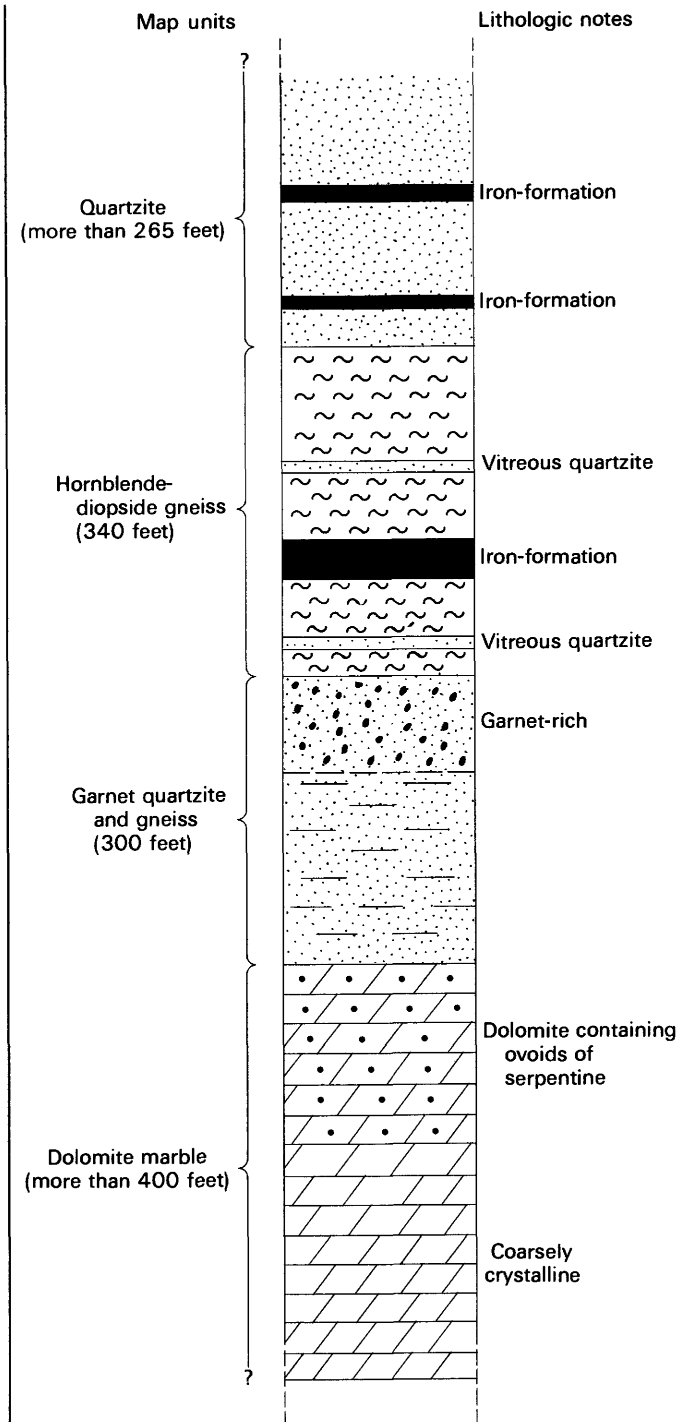



FIGURE 8.—Stratigraphic section of the Christensen Ranch Metasedimentary Suite in the Kelly area, northeastern Ruby Range (thicknesses in parentheses).

TABLE 8.—*Chemical analyses of metasedimentary rocks from the Kelly area, northeastern Ruby Range, Mont.*

[In weight percent. Analyst, Paula M. Buschman]

| | 1 | 2 |
|---|---|---|
| $SiO_2$ | 8.50 | 70.41 |
| $Al_2O_3$ | 0.91 | 12.00 |
| $Fe_2O_3$ | .87 | 1.87 |
| FeO | .99 | 6.97 |
| MgO | 23.39 | 0.47 |
| CaO | 24.42 | 2.74 |
| $Na_2O$ | .03 | 2.34 |
| $K_2O$ | .08 | 1.19 |
| $H_2O^+$ | 2.36 | .46 |
| $H_2O^-$ | .33 | .05 |
| $TiO_2$ | .05 | .86 |
| $P_2O_5$ | .03 | .19 |
| MnO | .70 | .15 |
| $CO_2$ | 37.03 | .23 |
| Cl | .07 | .00 |
| F | .06 | .01 |
| S | .00 | .00 |
| C | .07 | .02 |
| Subtotal | 99.89 | 99.96 |
| Less O | .05 | .00 |
| Total | 99.84 | 99.96 |

Sample data:

1. Dolomite marble containing ovoids of serpentinized forsterite and blades of phlogopite. Sample HJ-194-60.
2. Garnetiferous quartzite, containing oligoclase, microcline, and magnetite; accessory sphene, apatite, and zircon. Sample HJ-187-60.

## HORNBLENDE-DIOPSIDE GNEISS

The strata enclosing the main iron-formation of the area (described separately below) are composed mainly of dark, greenish-gray, streaky to poorly layered rock that typically consists of about 60 percent plagioclase, 20 percent hornblende, 15 percent diopside, and 5 percent quartz (fig. 9*A*).

The plagioclase generally is calcic andesine ($An_{36-40}$) but ranges from sodic andesine to calcic labradorite. The hornblende, black in hand specimen, is brownish green in thin section. The common pyroxene is greenish-gray diopside but in a few specimens it is hypersthene; no example of both pyroxenes occurring in the same sample has been observed in the present study, but this association is reported by Dahl (1979). Other minerals present in varying abundance include garnet, biotite, microcline, and accessory magnetite and apatite. Scapolite was found in a few specimens but does not appear to be systematically distributed.

A few thin beds of vitreous quartzite are interlayered with the gneiss. The most persistent of these is about 50 ft stratigraphically above the contact of the gneiss with the underlying garnetiferous unit; it is 10–20 ft thick.

The aggregate thickness of the gneiss unit, including interbedded quartzite and iron-formation, is about 340 ft. The sedimentary assemblage now represented by hornblende-diopside gneiss can be assumed to have consisted initially of dolomitic muds containing varying amounts of sand and clay.

## IRON-FORMATION

The iron-formation of the area is heavy, dark rock composed principally of quartz, magnetite, and pyroxene; garnet is abundant in some layers but is absent in most. The rock tends to be streaky rather than distinctly layered; component minerals such as quartz are aggregated into flat lenticles a few millimeters in thickness and a few to tens of centimeters in length. The resultant foliation is parallel to stratigraphic contacts, so doubtless is inherited from the thin layering typical of less metamorphosed iron-formations. Complex minor folding can be observed in nearly all outcrops; the general form of the folds tends to be systematic in relation to major structures, but plunges commonly diverge. The stratigraphic thickness of the iron-formation contained within the gneiss unit is about 40 ft, but outcrop widths may be much greater because of structural duplication.

Quartz and magnetite are the principal minerals, and grain sizes typically are 0.3 mm or more. Silicates make up 10–20 percent of most samples; dominant species are orthopyroxene and clinopyroxene in varying proportions (fig. 9*B* and *C*). Garnet, pink and isotropic in thin section, is abundant in a few layers but generally is absent. Perthitic potassium feldspar is a common minor constituent, intergrown with quartz, and apatite occurs as small clear crystals.

The two pyroxenes are similar in appearance in thin section: both are pale greenish to brownish gray, about

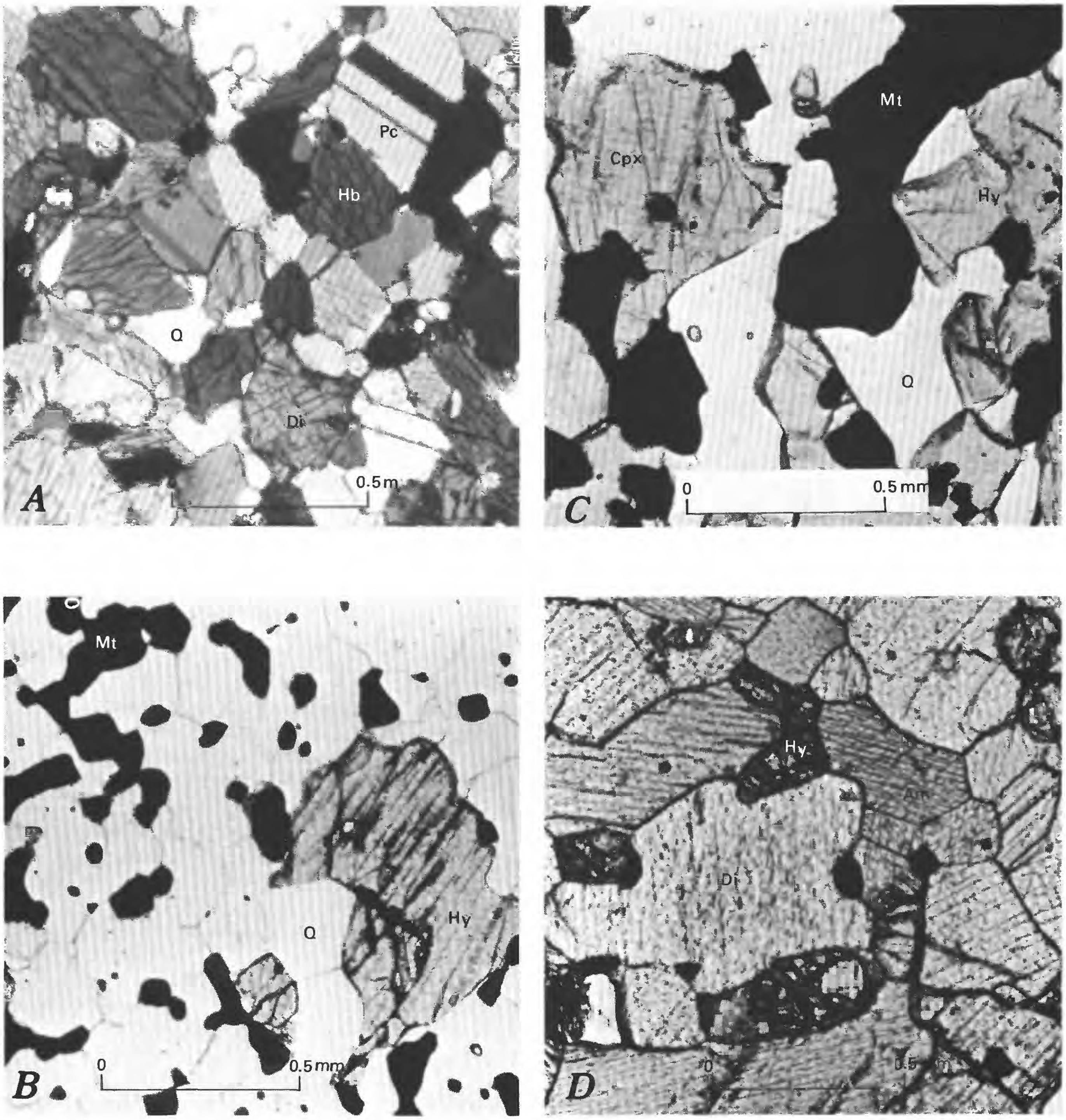
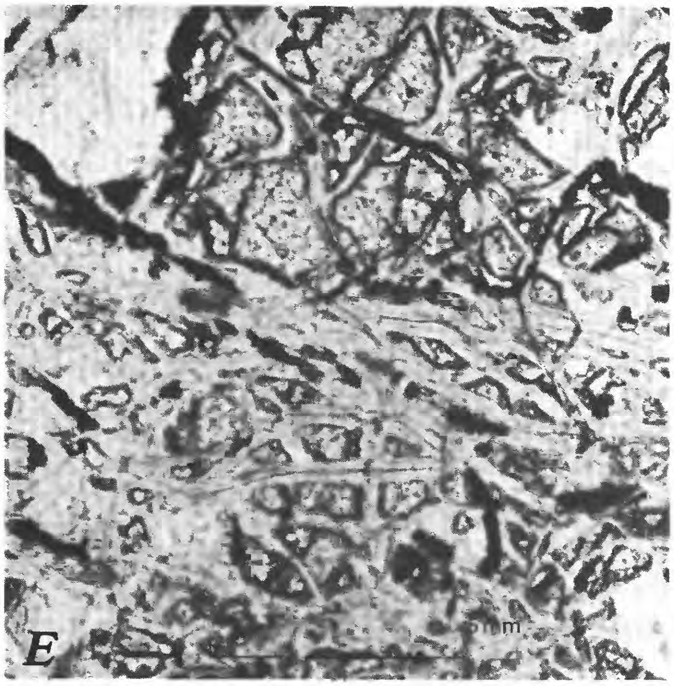
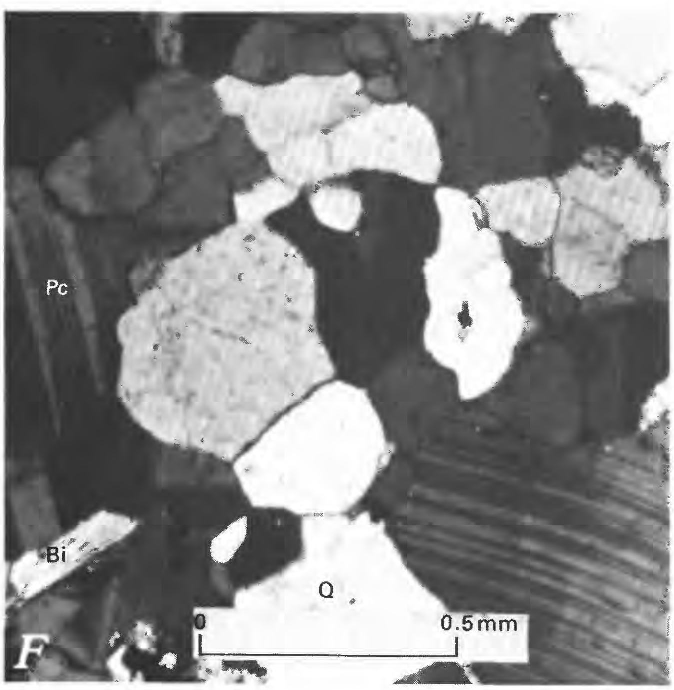



FIGURE 9 (above and facing page).—Photomicrographs of rocks from the Kelly area. *A*, Diopside-hornblende gneiss (specimen HJ–55–60); crossed polars. *B*, Quartz-rich iron-formation, containing hypersthene (specimen HJ–13–60); plane-polarized light. *C*, Silicate-rich iron-formation (specimen HJ–189–60); plane-polarized light. *D*, Metapyroxenite, containing relict hypersthene (specimen HJ–47–60); plane-polarized light. *E*, Peridotite (specimen HJ–15–60); relict olivine (high relief) in alteration matrix of antigorite plus magnetite; plane-polarized light. *F*, Quartz diorite (specimen HJ–52–60); plagioclase crystals commonly bent or fractured; crossed polars. Am, amphibole; Bi, biotite; Cpx, clinopyroxene; Di, diopside; Hb, hornblende; Hy, hypersthene; Mt, magnetite; Pc, plagioclase (labradorite in *A*; oligoclase in *F*); Q, quartz.




equally birefringent, and very weakly pleochroic. The orthopyroxene is to be classed as hypersthene and the clinopyroxene, which consistently displays fine lamellae of exsolved hypersthene, is in the diopside-hedenbergite series. Locally, the pyroxenes are altered to complex assemblages that include blue-green hornblende, cummingtonite, brown mica, carbonate, and riebeckite.

On the basis of paired compositions of clinopyroxene-orthopyroxene, garnet-orthopyroxene, garnet-clinopyroxene, and other mineral pairs, Karasevich and others (1981) estimate a peak temperature of metamorphism of 745±50 °C, compared to 675±45 °C for the Carter Creek area of the southwestern Ruby Range.

Chemical analyses of selected samples of iron-formation have been given earlier in this report (table 2, analyses 5, 6, and 7). The total iron content is strikingly consistent, but the amount present as magnetite is more variable, as indicated below:

| Analysis number (from table 2) | 5 | 6 | 7 |
|---|---|---|---|
| Total Fe | 37.33 | 37.59 | 37.27 |
| "Excess" Fe, after assignment of all $Fe_2O_3$ to magnetite | 7.34 | 3.16 | 3.72 |
| Fe as magnetite | 29.99 | 34.43 | 33.55 |

Standard commercial methods of iron determination commonly do not extract iron present in silicate minerals such as pyroxene and garnet, or else extract it incompletely, so that in commercial assays the iron content can be expected to agree more closely with that given above for iron in magnetite, rather than for total iron as determined by complete analysis. This is borne out by assay data available for drill core from the area, in which the average iron content reported is about 33 percent.

Comparative aspects of the iron-formation chemistry have been discussed earlier in this report, in relation to the Carter Creek deposits. Of particular note are the relatively high values for $Al_2O_3$, $K_2O$, and MnO, and the distinctly lower content of $P_2O_5$.

Physically and chemically the iron-formation is eminently acceptable as a taconite ore from which magnetite can be recovered after only moderately fine grinding. Quantitatively, however, the deposits lack the tonnage necessary for serious consideration as a candidate for commercial exploitation. In the principal area of interest and exploration, the central upthrust block, the amount of iron-formation to a depth of 300 ft is estimated to be about 15 million tons, containing 33 percent recoverable iron. Elsewhere the iron-formation is in thin beds that offer virtually no prospect for development.

## QUARTZITE

The uppermost packet of strata delineated on plate 2 consists of a sequence of quartzitic beds, individual units of which rarely exceed 20 ft in thickness, that enclose two or more thin beds of iron-formation. Biotite quartzite, biotite-garnet quartzite, and vitreous quartzite are well exposed in the eastern part of the map area, on the ridge immediately adjacent to the access track entering from Beatch Canyon. These rocks typically are coarse grained, and biotitic varieties are strongly foliated. Some layers contain as much as 50 percent garnet, pink in thin section and isotropic. Trace amounts of sillimanite have been observed locally.

Iron-formation, in beds generally 10 ft or less in thickness, occurs at several stratigraphic levels in the quartzitic sequence. Exposures are few, but the beds are readily traced magnetically. The rock is similar in almost all respects to the main iron-formation, described in earlier paragraphs. Quartz and magnetite are the principal minerals, and hypersthene, diopside-hedenbergite, and garnet are common constituents.

## IGNEOUS AND META-IGNEOUS ROCKS

Four varieties of igneous or meta-igneous rocks are present in the Kelly area, but in the aggregate they constitute only a small fraction of bedrock. All are Precambrian in age.

### ULTRAMAFIC ROCKS

The principal mass of ultramafic rock in the area is a body of peridotite, arcuate in surface outline, that occupies the core area of the central upthrust block containing the explored iron-formation. Metapyroxenite (not shown on plate 2; see James and Wier, 1972a, for locations) is present in a number of places as small lenses that, like the main mass of peridotite, have been tectonically emplaced, diapir-fashion, in the metasedimentary strata.

The peridotite (or, more precisely, metaperidotite) is dense, dark gray to black on fresh break, weathering to the characteristic yellow-brown ("buckskin") surface typical of this rock type. It is complex mineralogically (fig. 9*E*). Original pyrogenic minerals (olivine and pyroxene) are preserved as small islands in serpentine that is crossed by mesh-like trails of magnetite. The initial assemblage was coarse grained (grains 1–3 mm in diameter). The olivine is colorless in thin section and optically negative, showing an optic angle near 90°, indicating a magnesium-rich composition. The preserved pyroxene is augite, slightly brownish in thin section. Much of the rock now consists of tremolite, in large clear grains that replace both the original pyrogenic minerals and the late serpentine. Enstatite, similar in appearance to tremolite and probably also of metamorphic origin, is present in a few samples. Finally, these secondary minerals, together with the pyrogenic precursors, are in turn altered locally to cummingtonite, calcite, and serpentine.

Metapyroxenite occurs as concordant thin lenses, most less than 20 ft thick and 100 ft long. The rock typically is medium to coarse grained and greenish black to black, depending upon the amount of secondary hornblende present. The original rock appears to have been composed largely, if not entirely, of hypersthene, now preserved in small relict patches in antigorite serpentine. Both hypersthene and antigorite are replaced by a granoblastic aggregate of clear diopside and brown-green hornblende (fig. 9*D*). Locally, the rock in outcrop is a crumbly aggregate consisting largely of black coarse-grained hornblende.

### QUARTZ DIORITE

Granitic intrusions are present at three localities in the map area, the most extensive near the center of sec. 25, T. 6 S., R. 5 W. The full extent of the latter body is not known, but it is at least 400 ft by 500 ft in surface area. A smaller body, in the NW¼NW¼ sec. 25, is in exposed contact with diopside-hornblende gneiss, which near the contact is seamed with irregular pegmatitic veins.

Most of the sampled granitic rock is quartz diorite that is fine to medium grained and gray to pink. In outcrop the rock is seen to be weakly to moderately foliated, and in thin section bending and fracturing of feldspar grains is evident. Typical samples consist mainly of quartz and oligoclase (fig. 9*F*), with minor potassium feldspar (vaguely mottled and probably microperthitic) and biotite.

The quartz diorite clearly post-dates the metamorphism and main structural deformation of the metasedimentary sequence. Preliminary isotopic measurements of Rb/Sr of samples from the body in the NW¼NW¼ sec. 25, coupled with those from the nearby Virginia City area, yield an 1,890 m.y. isochron (C.E. Hedge, written commun., 1981), with a high initial $Sr^{87}/Sr^{86}$ ratio of about 0.719.

### PEGMATITE

Pegmatite occurs throughout the Kelly area in bodies ranging from thin irregular seams to masses several hundred feet long and 100 ft wide. Most are of simple composition—pink microcline and quartz; in some smaller bodies the feldspar is white albite. Biotite and muscovite are common constituents of thinner lenses and seams but are scarce or absent in larger bodies. No rare minerals have been observed and the pegmatites are not zoned. Some larger bodies are moderately foliated but smaller veins and dikes are undistorted.

The relation between the pegmatite and the quartz diorite is not known, but the structural similarity suggests a common time of emplacement, probably during the Early Proterozoic.

## STRATA OF PALEOZOIC AGE

As noted previously, the Precambrian rocks of the area are bounded on the north and west by strata of Paleozoic age. The north boundary is an unconformity; here the Precambrian is overlain by the Flathead Sandstone, locally glauconitic, which is succeeded by the Wolsey Shale, both of Cambrian age. On the west the Precambrian is in fault contact with massive limestone of the Mississippian Madison Group. These strata are described in more detail by Tysdal (1976b).

## STRUCTURE

The principal structural elements of the area consist of Precambrian folds and faults within the block of crystalline rocks and faults that displace both Precambrian rocks and the younger strata of Paleozoic age. The latter will be described first.

### POST-PALEOZOIC FAULTS

The northern Ruby Range is crossed by a number of northwest-trending faults of major displacement (Tysdal, 1976a; Karasevich, 1981). Within the Kelly area, this system is represented by the fault that brings limestone of the Madison Group into juxtaposition with Precambrian crystalline rocks. This structure, labeled the "Kephart fault" by Karasevich, is not here exposed, but judging from the relation of the surface trace to topography, it dips steeply to the east. Movement is high angle and reverse, and the minimum displacement is several thousand feet. The Kephart fault in turn is cut by a later east-trending fault that is the structural control for the north branch of Taylor Canyon. Movement on this fault is dominantly left lateral, displacing the Kephart fault trace by about 800 ft, but offset of the Cambrian-Precambrian unconformity indicates some vertical movement, north side down. Poor exposures of the fault, about 900 ft north of the N¼ cor. sec. 25, indicate a nearly vertical dip.

### PRECAMBRIAN STRUCTURES

The major Precambrian structures of the area are a broad open syncline that is clearly outlined by the stratigraphically lower beds of the Christensen Ranch Metasedimentary Suite, and an upthrust central block that contains the explored iron deposits. The axial plane of the major syncline apparently is about vertical; the fold axis trends N. 70° W. and the plunge is to the east-southeast at about 40°. Minor structures, most inferred from map patterns and the located traces of magnetic units, are dragfolds that are systematic with respect to the major fold.

The central fault block is a lens-shaped mass about 3,500 ft long and 1,200 ft in maximum width that has been thrust up from the keel of the main syncline. This mass, cored by peridotite, has the internal form of an east-southeasterly plunging anticline flanked by a pair of truncated synclines; almost certainly it originated as an anticlinal buckle in the deeply buried axial zone of the main syncline, moving upward from its original matrix like a squeezed watermelon seed. The iron-formation within the upthrust mass has been greatly thickened by complex folding, evident in most outcrops.

The fault that bounds the central block on the north is exposed in an exploration trench about 700 ft south of the N¼ cor. sec. 25; it is a steeply dipping shear zone about 100 ft wide marked by intensely crumpled rock and thin quartz veins. The companion fault that forms the south boundary is not exposed, but its location is tightly controlled by stratigraphic data. These two bounding faults necessarily must terminate to the west, because they do not cut the stratigraphic units that outline the main syncline. A similar termination by merging is inferred for the eastern extension of the faults, but the evidence is not definitive.

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