Metamorphism in the Riggins Region Western Idaho

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A study of the physical and chemical characteristics of the metasedimentary and metavolcanic rocks, and of the conditions of their origin and metamorphism

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METAMORPHISM IN THE RIGGINS REGION, WESTERN IDAHO

By Warren Hamilton

ABSTRACT

The west-directed Rapid River thrust separates two assemblages of metamorphic rocks in the northwestern part of the 30-minute Riggins quadrangle. Below and west of the fault is the metavolcanic formation of the Seven Devils Volcanics (Permian and Triassic), the schistose marble and crystalline limestone of the Martin Bridge Limestone (Triassic), and the Lucile Slate (Triassic(?)). These rocks are of low metamorphic grade throughout; granitic rocks intrusive into them have been metamorphosed in the same degree. The thick Seven Devils Volcanics consist of submarine lavas, tuffs, and agglomerates with the compositions of basalt, andesite, dacite, and their albitized equivalents spilite, keratophyre, and quartz keratophyre. The Martin Bridge Limestone was approximately 2,000 feet thick prior to deformation. The Lucile Slate is thin.

Lying tectonically above and east of the Permian and Triassic sequence is a thick sequence of metamorphic rocks, dominantly schists, here named the Riggins Group. The present order—which is probably in part tectonic rather than stratigraphic—within the group is, from bottom to top, Fiddle Creek Schist (metamorphosed silicic tuffs and lavas), now about 9,000 feet thick; Lightning Creek Schist (metamorphosed mafic to silicic tuffs and lavas), about 8,000 feet thick; and Squaw Creek Schist (metamorphosed sedimentary and volcanic-sedimentary rocks), about 6,000 feet thick. Metaperidotite is intercalated at or near the contact between the Squaw Creek and Lightning Creek Schists, and this contact is probably a premetamorphic thrust fault. An amphibolite unit lies between the same schists in another area.

Chemical analyses are presented for major and minor elements in 30 specimens, mostly metavolcanic rocks of the Riggins Group. Volcanic compositions vary from calc-alkaline to sodic and define a compositional spectrum comprising normal to keratophyric rocks. The metavolcanic rocks are compositionally identical with the older and variably spilitized volcanic rocks of the Aleutian Islands, where volcanism has continued to the present time, and where subaerial rocks are of normal calc-alkaline compositions. In both the Riggins quadrangle and the Aleutian Islands, exchange of magmatic calcium for seawater sodium probably resulted in the sodic rocks.

The rocks of the quadrangle increase in metamorphic grade eastward and southward. Some of the common metamorphic index minerals of pelitic schists are lacking; and isograds have been drawn, in order of increasing metamorphism, upon the westernmost appearance of aluminian prochlorite (as opposed to the ferroan variety), biotite, clinozoisite (as opposed to epidote), garnet, oligoclase, and andesine. The metamorphism of the Riggins Group probably occurred during Early Cretaceous time, whereas the rocks beneath the Rapid River overthrust may have been metamorphosed not only at this time, but also during the Late Jurassic.

Analysis of data from this area and many others in which progressive metamorphism has been studied indicates that the heat represented must have been partly introduced by magmas mobilized at deeper levels and partly generated by deformation of the rocks. The generally accepted theory that the heat of regional metamorphism is due to depth of burial seems largely invalid.

INTRODUCTION

This report contains a study of the chemical and physical characteristics of metasedimentary and metavolcanic rocks west of the Idaho batholith in west-central Idaho. The rocks studied are primarily those of the 30-minute Riggins quadrangle, but information is included on the rocks of areas adjacent to the north and to the west. Geologic maps of three areas accompany this report (fig. 1).

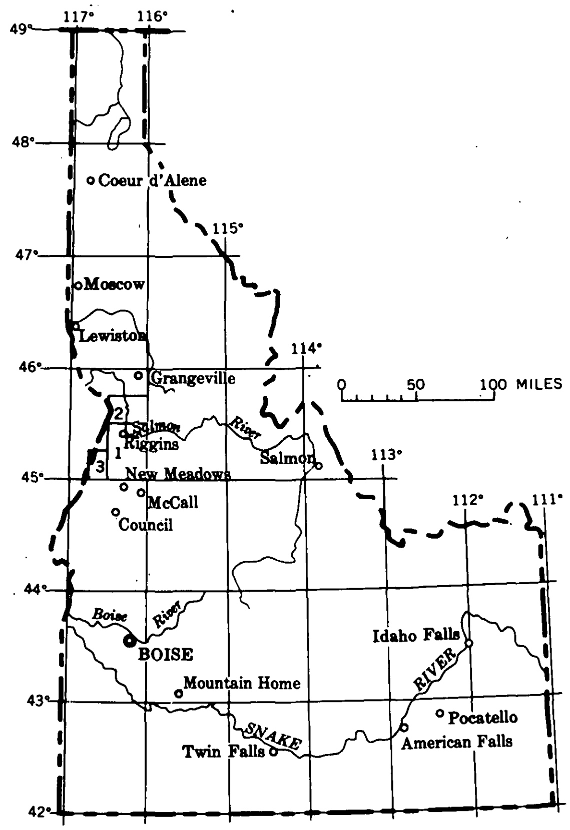

Figure 1.—Index map of Idaho. 1, Riggins quadrangle; 2, area of plate 2; 3, Cuprum quadrangle (pl. 3).

Within the areas considered in this report are the village of Riggins, a few hamlets, and scattered ranches and small resorts. Placer mining along the Salmon River and elsewhere and bedrock mining in the Seven Devils mountains have been commercially important in the past, but lumbering and stockraising are now the primary economic activities. Most of the areas are within national forests.

The areas are dominated topographically by the deep steep-walled canyons of the Salmon and Snake Rivers, and accessibility is limited. Altitudes range from about 1,500 to nearly 9,000 feet. Rock exposures are good in the lower parts of the canyons, generally below an altitude of about 2,500 feet, and also in many areas above 7,500 feet; the intermediate altitudes are mostly blanketed by heavy forest.

PRESENT WORK

This report is based on fieldwork in the Riggins quadrangle during the summers of 1952, 1954, 1956, and briefly during 1957 and 1958. The work was accomplished largely by traversing roads and trails; and the traverse net is indicated, except in the areas of few outcrops, by the distribution of attitude symbols on the geologic map (pl. 1). Detailed studies of structural and lithologic relations were made at many locations, but the map is of only semireconnaissance quality. Assistance in the field was given by G. J. Klapper in 1956 and by T. R. Lovett during part of the 1952 season. Donald G. Sherlock, a young geologist of much promise, assisted during the early part of the 1954 season but was killed attempting to cross the well-named Rapid River on foot.

Laboratory work for the present report included the study of about 250 thin sections and 1,000 hand specimens. The 30 silicate analyses were made mostly by P. L. D. Elmore, S. D. Botts, K. E. White, and P. W. Scott, and most of the spectrographic analyses were made by Paul R. Barnett.

The study of the Riggins quadrangle was proposed by Clyde P. Ross, who recognized that the quadrangle contains the most extensive exposures of pre-Tertiary schists and gneisses along the west side of the Idaho batholith. The main purpose of the study was to determine the structural and petrogenic relationships among and between the diverse assemblages of crystalline rocks.

Widely spaced traverses were made in the area north of the Riggins quadrangle (pl. 2) during 1958 and 1961.

The Cuprum quadrangle was studied in detail by Ralph S. Cannon, Jr., and several associates during 1938, 1939, and 1940, but this work never was published. A generalized version of Cannon's map accompanies this report as plate 3. Cannon allowed me to study his many thin sections, and brief descriptions of these are incorporated in this paper; descriptions of Cannon's fossil collections and a section measured by him in the Martin Bridge Limestone are included also.

GENERAL GEOLOGY OF RIGGINS QUADRANGLE

The pre-Tertiary complex of the Riggins quadrangle consists of metavolcanic and metasedimentary rocks in the northwest, increasing in metamorphic grade eastward and southward, and varied gneisses and granitic rocks elsewhere. Conveniently, the granitic and gneissic rocks are separated by a major thrust fault from low-grade rocks in the west-central part of the quadrangle, and trondhjemite (light-colored biotite-oligoclase-quartz diorite) gneiss is in intrusive contact with middle-grade rocks near the north edge of the quadrangle; so in these areas the distinction between "metamorphic" and "plutonic" rocks is easily made. The distinction is, however, arbitrary between these areas, because a complete transition from schists to plutonic gneisses is exposed along both the Salmon and Little Salmon Rivers. The problems of the two groups of rocks are different, however, and distinction between them is needed for organization of their descriptions. In the "metamorphic" rocks, as the term is limited here, formations generally can be traced with some confidence, and they are structurally continuous; intrusive igneous masses are virtually limited to small dikes. In the "plutonic" rocks, the same formations are probably present, but so high a degree of structural mobility was attained during metamorphism and deformation that continuity was lost and stratigraphic assignments are no longer possible. Also, intrusive granitic materials become increasingly more abundant to the east and south, toward the plutonic terrane. Reconstitution of the "metamorphic" rocks was essentially isochemical, whereas widespread introduction and migration of components was dominant in the "plutonic" rocks.

The geologic map (pl. 1) accompanying this report shows only the northwestern part of the quadrangle, which contains the "metamorphic" rocks as here limited.

PREVIOUS WORK IN RIGGINS QUADRANGLE

Lindgren (1900, pl. 9) distinguished "Carboniferous(?) slates, schists, and old effusive rocks," "post-Carboniferous" granite, and Miocene basalt in his rapid reconnaissance of the region that includes part of the quadrangle.

The area included in the sketch map by Livingston and Laney (1920, map 4) extends east into the Riggins quadrangle to the Salmon and Little Salmon Rivers.

The Salmon River placer deposits were described briefly by Lorain and Metzger (1938, p. 79–86) and by Staley (1945, p. 6–10).

Wagner (1945) published a reconnaissance map that included the northwestern part of the area and gave the first, though brief, printed description of the rocks. Although the schists north of Riggins almost everywhere dip southwestward, Wagner shows mostly southeast dips there.

PHYSICAL FEATURES OF RIGGINS QUADRANGLE

The quadrangle is ruggedly mountainous; its topography is dominated by the deep canyons of the Salmon and Little Salmon Rivers that meet in a T at Riggins. From its source in the uplands in the southern part of the quadrangle, the Little Salmon flows northward, its canyon becoming deeper as the flanking ridges maintain their altitude of 7,500 to 8,500 feet and the stream drops from an altitude of 4,000 to 2,000 feet. The big Salmon flows westward across the northern part of the quadrangle in a spectacular canyon whose long slopes are steeper than 40° in the 4,000-foot-deep inner gorge; above the gorge, more gentle slopes rise to high peaks and ridges. At Riggins, the Salmon is joined by the Little Salmon and turns northward along the projection of the smaller river. The gradient of the Salmon is about 10 feet per mile, which is high for a big river, and the river leaves the quadrangle at an altitude of 1,600 feet. Tributaries to the Salmon and Little Salmon have incised deep side canyons. The south-flowing streams in the southern corners of the area flow within relatively shallow valleys.

Beyond the canyons are uplands of moderate relief that are mostly between the altitudes of 6,000 and 8,000 feet. Many peaks and ridges in all parts of the quadrangle, except north of the Salmon and in the southwest corner, rise above 8,000 feet, but none are above 9,000 feet. The highest peak is Patrick Butte, which, at 8,824 feet, is 7,000 feet above the Salmon River 5 miles to the north.

The northeast corner of the quadrangle extends into a broad forested surface of low relief that forms a striking high platform bordered by higher peaks miles farther east. Another large area of relatively low relief occupies the southwest corner of the area. The northern part of broad Meadows Valley, mostly pastureland, is within the quadrangle. Many lakes, dammed by glacial debris or filling ice-scoured basins, spot the strongly glaciated east-central part of the quadrangle (outside the area of pl. 1) and lie among scenic meadows, forested slopes, and granite peaks.

Exposures of bedrock are excellent in the lower parts of the canyons and in many of the high glaciated areas, but they are generally very poor in the heavily forested intermediate zone.

ROCKS BENEATH RAPID RIVER THRUST

The two major assemblages of metamorphic rocks in the Riggins quadrangle are separated by the Rapid River thrust (pl. 1). Below and west of the fault are the metavolcanic formation of the Seven Devils Volcanics (Permian and Triassic), the schistose marble and crystalline limestone of the Martin Bridge Limestone (Triassic), and the Lucile Slate (Triassic?); these rocks are of low metamorphic grade throughout. Above the Rapid River thrusts are the various schists of the Riggins Group of Paleozoic or Mesozoic age.

SEVEN DEVILS VOLCANICS

Metamorphosed submarine lava, tuff, and agglomerate, products of island-arc volcanism, are the dominant rocks beneath the Rapid River thrust. Basalt, andesite, dacite, and their spilitized equivalents—spilite, keratophyre, and quartz keratophyre—are all widespread. Most of the rocks are green, but purple, maroon, and brown rocks are common. Massive greenstones are dominant over greenschists. The age of the Seven Devils Volcanics is Permian and Late Triassic.

DEFINITION AND DISTRIBUTION

Metavolcanic rocks form most of the Seven Devils Mountains (within and immediately west of the Riggins quadrangle), for which they were named "the Seven Devils series" by Lindgren (1900, p. 193–198). The name was revised to Seven Devils Volcanics by Anderson (1930), and subsequent workers have followed this usage.

The Seven Devils Volcanics crop out in the westernmost 1 to 5 miles of the quadrangle, beneath and west of the Martin Bridge Limestone. They crop out westward to the west wall of the Snake River Canyon, where they disappear under Columbia River Basalt. The canyon of the Snake River and the canyon of the Salmon River north of Lucile are carved in these rocks. Similar rocks, correlative at least in part, are widely exposed in eastern Oregon, as in the Baker quadrangle where they are called the Clover Creek Greenstone (Gilluly, 1937).

Only widely spaced traverses were made through the Seven Devils Volcanics in the Riggins quadrangle. Their structure is complex, and no mappable stratigraphic subdivisions were recognized. (The schists of the Riggins Group form, by contrast, distinctive stratigraphic units.) The Seven Devils Volcanics occur mostly within the altitude range of heavy forest, and exposures are poor to very poor. The best exposures are along the Bald Mountain-Papoose Lake-Heavens Gate crest and in the lower canyon of the Rapid River.

The thickness of the Seven Devils Volcanics is great but uncertain. They extend for many miles and are obviously shingled by faults but apparently not deformed isoclinally. A thickness of several miles is possible.

LITHOLOGY

The dominant rocks of the Seven Devils Volcanics are massive dull greenish-gray rocks of metavolcanic origin, "greenstones" in the proper sense of the term. Metamorphosed tuffs and agglomerates are more abundant than are flows and sills. Compositions range from basalt and spilite to rhyolite and potassic quartz keratophyre, but the intermediate rocks—andesite and keratophyre—are most abundant. Shearing effects vary widely.

Volcanic textures, both clastic and primary-igneous, are widely preserved. Not only are clasts and plagioclase phenocrysts commonly recognizable (figs. 2 and 3), but the groundmass fabrics—particularly those dominated by plagioclase laths—are recognizable in many unsheared rocks. As the effects of shearing and reconstitution of components increase, these primary features disappear progressively. Many rocks show interlensing shear folia, a fraction of a millimeter to several centimeters apart within which the initial fabric has been destroyed but between which it is preserved. Pervasive cleavage is more common, however.

Clasts in the tuffs and agglomerates are of all sizes from microscopic to a common maximum of 6 or 8 inches (fig. 4). The clasts are of tuffs and flow rocks, and, where unsheared, they are poorly rounded. Crystal tuffs, whose fragments are of single mineral grains (mostly plagioclase) rather than of rocks, are much less common. Clasts have been markedly deformed in many places (fig. 5).

Large primary volcanic features—columnar joints, pillow structure, and dikes—were nowhere recognized; but as exposures are very poor through most of the formation, they may actually be widespread.

The feldspar of the metavolcanic rocks is everywhere albite that occurs both as pseudomorphs of the primary plagioclase and as entirely secondary granules. Relict phenocrysts are euhedral, commonly 1 to several milli-

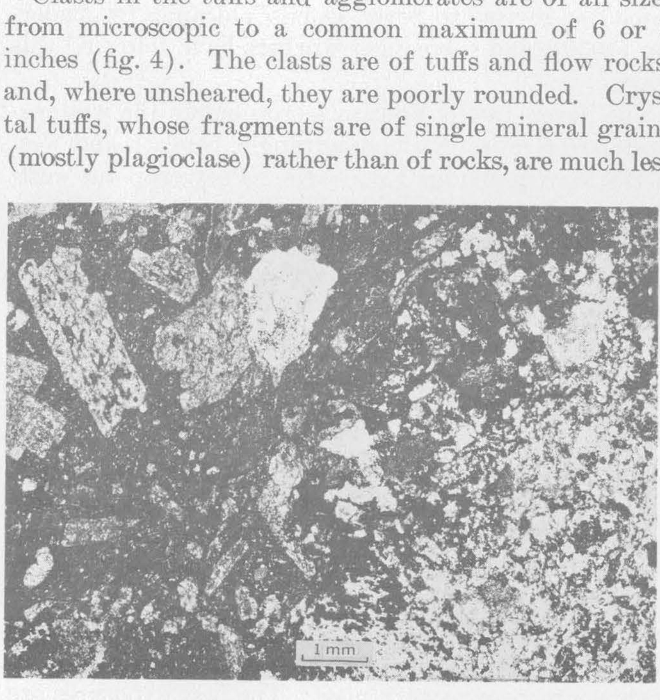

FIGURE 2.—Photomicrograph of greenstone altered from andesite lithic tuff. Left half is a clast of meta-andesite with abundant euhedral phenocrysts of plagioclase pseudomorphed by albite crowded with white mica, chlorite, and epidote. Right half is groundmass composed of ferroan prochlorite and epidote. Seven Devils Volcanics, altitude 5,900 feet, Papoose Creek road.

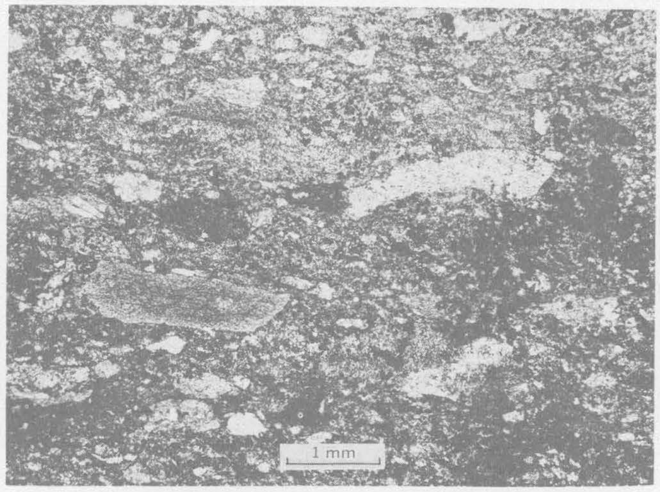

FIGURE 3.—Photomicrograph of metamorphosed lithic tuff. The flattened clasts are now sericite-chlorite-albite phyllite. Groundmass is of finer clasts and altered feldspar fragments in chloritic phyllite that contains much magnetite and hematite. Seven Devils Volcanics, Rapid River trail, just beneath Martin Bridge Limestone.

FIGURE 4.—Greenstone agglomerate. Bedding dips moderately left. The clasts are essentially undeformed; compare with figure 5, taken nearby. Seven Devils Volcanics in saddle north of Heavens Gate lookout.

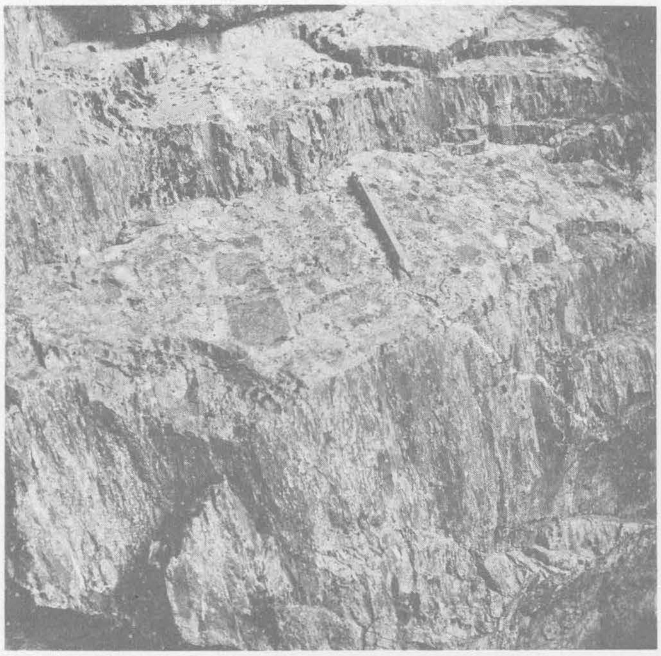

FIGURE 5.—Highly sheared meta-agglomerate. Axial ratios of the deformed clasts are near 1 : 2 : 4, and the longest axes plunge forward. Seven Devils Volcanics; southwest of Papoose Lake, on south side of hill 8215.

meters long (and locally 3 cm long), tabular parallel to (010), and generally replaced either by a single crystal of albite or by several individuals of albite that differ only slightly in orientation. Carlsbad and albite twins are common. The albite is crowded with tiny inclusions, among which green chlorite is ubiquitous, and white mica, epidote, and calcite are far less common. So abundant are these inclusions of chlorite in most rocks that even very large albite crystals are so colored and dulled that they can be seen only with difficulty in the hand specimen; white or light-colored albite is uncommon. Groundmass albite varies from single-crystal parallel-structure pseudomorphs of plagioclase laths to reconstituted and intergrown granules of variably migrated material. The granules are little twinned. In rocks of normal volcanic compositions (that is, nonkeratophyric rocks), the pseudomorphs of albite after groundmass plagioclase are often so densely crowded by inclusions as to be invisible in thin section under plane light; their character is revealed only through crossed nicols, which bring out the continuity and shape of the host crystals.

Mafic phenocrysts are not commonly recognizable. Where present, they are represented mostly by ragged clots of granules of epidote, chlorite, actinolite, magnetite, and leucoxene. Rarely are crystal outlines preserved, and no primary mafic minerals were seen in either phenocrysts or groundmasses. The shape of the mafic clusters in uncommon rocks suggests original hornblende, and single-crystal pseudomorphs of actinolite in one metabasalt specimen suggest pyroxene phenocrysts.

The chlorite throughout the area is light green, moderately pleochroic, and has an anomalous yellowish-brown interference color. It is presumably ferroan prochlorite. The epidote is yellowish green and pleochroic, and although mostly in anhedral granules, it is in small euhedral domes in some specimens. Actinolite appears mostly as tiny needles that crosscut all other minerals. Quartz is present in the uncommon silicic rocks both as granules in the groundmass and as equant clusters of intergrown grains that were probably phenocrysts. Calcite occurs as small granules (locally as knots as much as several millimeters across) and forms uncommon veinlets alone or with quartz. Notably lacking are clinozoisite, biotite, hornblende, and plagioclase other than albite.

The groundmasses of most of the rocks, both massive and schistose, are very fine grained hazes of intergrown albite, chlorite, epidote, actinolite, leucoxene, and, less commonly, quartz, calcite, and muscovite, liberally strewn with magnetite and hematite, in varying proportions and combinations. The average grain size of the silicate minerals is commonly only about 0.01 mm. Modal measurements are not possible in such rocks; but from the approximate proportions of these minerals, and from the character of the phenocrysts where present, a rough classification into initial rock types can be made. Keratophyre is more abundant than andesite; the two together are more abundant than all other types; basalt, spilite, dacite, and quartz keratophyre are common; and rhyolite and potassic quartz keratophyre are very uncommon.

The primary grain size of the groundmasses of the rocks ranged from aphanitic, even glassy, for a relative few, to 0.5 mm or a little more. Presumably at least some of the coarse rocks formed as sills and dikes.

A minor proportion, perhaps a tenth, of the formation is composed of reddish-gray to maroon rocks that contrast markedly with the prevalent grayish-green rocks. These reddish-gray rocks are petrographically like the green ones except that they contain abundant disseminated hematite. Their oxidation seems to be a metamorphic feature rather than one of weathering.

Intercalated with the Seven Devils Volcanics is calcite limestone. Some of this (notably, large to small unmapped masses in the vicinity of Cannonball Mountain and the West Fork of the Rapid River) is lithologically identical with the Martin Bridge Limestone and presumably represents tectonic lenses of that formation. Some, as intercalations up to several hundred feet thick along the northernmost 2 miles of the Boise trail, is of a different type, presumably interbedded with the volcanic rocks, and consists of laminated fissile very fine grained variably sheared gray to pinkish-gray nearly pure calcite limestone.

METAMORPHISM

The primary volcanic minerals of the Seven Devils Volcanics have been completely reconstituted to low-grade metamorphic minerals—notably albite, ferroan prochlorite, actinolite, and epidote. The metamorphic grade does not change within the Riggins quadrangle, and the mineral assemblages remain those of the ferroan prochlorite zone. (See p. 75.)

Foliation—the orientation of slaty cleavage and planar shear features such as those of figures 3 and 5—has a general gentle eastward dip. The degree of shearing varies greatly and irregularly. The finer tuffs and shaly interbeds show slaty cleavage most consistently, but originally massive rocks show cleavage in many places; sheared and unsheared rocks occur together in all parts of the area. Figures 4 and 5 illustrate two neighboring outcrops of initially similar agglomerate, the one undeformed, the other greatly sheared. A general eastward increase in shear effects might be expected, but none was recognized—completely massive greenstones are widespread immediately beneath the Martin Bridge Limestone—and no obvious zones of major shearing of the Seven Devils Volcanics were noted in the quadrangle. The cleavage has a marked sheen where best developed, but nowhere are the planar metamorphic minerals visible without a microscope; the grain size is slaty to phyllitic but nowhere schistose.

ROCKS NORTH OF RIGGINS QUADRANGLE

The Seven Devils Volcanics are exposed along the Salmon River from Lucile north to Slate Creek (pl. 2). The dominant rocks are greenstones varying to greenschists, but purplish and reddish rocks are widespread, and there are abundant pods and small plutons of metamorphosed granitic rocks. The metavolcanic rocks are, in general, much more sheared than are those of the formation in the Riggins quadrangle, but the grade of reconstitution is similar. Most of the rocks are of mafic and intermediate compositions.

East of the river, along Slate Creek, the Seven Devils Volcanics increase in metamorphic grade to become green and gray schists that contain such middle-grade minerals as hornblende, biotite, aluminian prochlorite, and reverse-zoned oligoclase.

Greenstones and greenschists that are very similar to, and presumably are correlative with, the Seven Devils Volcanics crop out along the South Fork of the Clearwater River from Harpster 8 miles upstream, northeast of the area shown in plate 2. These metavolcanic rocks enclose many dikes and larger masses of metagranitic rocks such as characterize the Seven Devils Volcanics along the Salmon River north of Lucile.

ROCKS OF THE CUPRUM QUADRANGLE

An intensive field study of the Seven Devils Volcanics in the Cuprum quadrangle was made by Ralph S. Cannon, Jr., who provided much of the following information and who allowed me to study his large collection of thin sections. The Seven Devils Volcanics underlie nearly half of the 15-minute Cuprum quadrangle (pl. 3), which adjoins the south half of the Riggins quadrangle on the west.

The sequence is many thousands of feet thick, and the stratigraphy and structure are so complex that neither has been deciphered. The same rock types alternate erratically throughout the quadrangle, with submarine tuffs and allied variably calcareous sediments dominant over lavas. Cannon found many marine invertebrates of Permian and Late Triassic ages in the tuffs, but the rocks of these contrasting ages are similar lithologically and are intercalated structurally; dikes are abundant.

Primary mineral assemblages have been almost completely destroyed. The rocks are now composed of low-grade metamorphic minerals, most of which are fine grained. Many rocks are somber grays, but others are brighter green, maroon, purple, or brown. The dominant rocks are metamorphosed andesitic and keratophyric lava and tuff. Dacite and quartz keratophyre are probably next most abundant, followed by basalt and spilite, and then rhyolite. Flow rocks are generally massive, but strong cleavage is developed locally, particularly in tuffaceous rocks.

The most common rock types are porphyritic andesite and keratophyre that contain relict phenocrysts of plagioclase, now represented by pseudomorphs of albite with included chlorite, in a fine groundmass of albite, epidote, actinolite, prochlorite, magnetite, and leucoxene in varying proportions. Original mafic minerals, except for some hornblende, have been completely altered. Mafic material is largely restricted to the groundmass of most rocks, but some specimens have pseudomorphs after mafic phenocrysts. Original groundmass fabrics dominated by small laths of plagioclase are recognizable in many specimens through the haze of secondary minerals, especially when viewed through crossed nicols.

Basalt and spilite, subordinate to andesite and keratophyre, are similar in most respects, including the prevalence of relict plagioclase phenocrysts, except that they have a greater content of mafic minerals. Small equant pseudomorphs, probably after olivine phenocrysts, of talc, magnetite, and epidote further distinguish some of these rocks.

The dacite and quartz keratophyre differ from the andesite and keratophyre in their considerable contents of groundmass quartz and, exceptionally, in the presence of phenocrysts of quartz as well as of the common plagioclase.

Rhyodacite, quartz latite, rhyolite, and their keratophyric equivalents are uncommon. Most of these contain relict bipyramidal phenocrysts of quartz. Potassium is in fine micas—biotite in some of the more mafic rocks; otherwise, sericite or muscovite. Some of the rhyolite is white, composed almost exclusively of albite, quartz, and muscovite.

Flow rocks are dominant among Cannon's specimens, but crystal and lapilli tuffs and agglomerates are dominant in the field according to him. Glassy rocks, now represented by extremely fine grained secondary mineral assemblages, and rocks with much original glass in the groundmass, are subordinate.

Regional metamorphism was of low grade but varied widely in intensity of shear. All specimens show almost complete reconstitution of the primary minerals, but most are so unsheared that no cleavage is visible in thin section. Rocks showing varying shear effects, with highly schistose rocks at one extreme, are irregularly distributed. Shear effects are particularly intense in a zone several miles wide that extends southwest across the quadrangle from near its northeast corner. Intrusive rocks are concentrated in this same zone.

ROCKS OF BAKER QUADRANGLE, OREGON

The Clover Creek Greenstone of Permian age of the Baker quadrangle, in eastern Oregon 50 miles along the probable tectonic strike to the west-southwest of the Riggins quadrangle, is presumably correlative with part of the Seven Devils Volcanics. The following description of the Clover Creek Greenstone is from Gilluly (1937, p. 22).

> The most abundant varieties are the quartz keratophyres, probably followed by quartz keratophyre tuffs, keratophyre flows, meta-andesites, and keratophyre tuffs. The other rock varieties [albite diabase, spilite, chert, conglomerate, argillite, and limestone] are subordinate.

Relict quartz phenocrysts are widespread. Some of the rocks retain a little relict hornblende, augite, and labradorite-andesine; and they contain a little secondary hornblende, biotite, and zoisite along with secondary minerals like those of the Seven Devils Volcanics.

The Clover Creek Greenstone is in bulk more silicic than the Seven Devils Volcanics. In general, the rocks of the Baker quadrangle seem to be less completely reconstituted than the Seven Devils Volcanics, yet they contain minerals of slightly higher metamorphic rank.

The Clover Creek Greenstone seems to share with the Seven Devils Volcanics an FeO : MgO ratio higher than that of the Riggins Group (fig. 46).

CHEMISTRY

The results of analyses for major and minor elements in three specimens of partially spilitized rocks from the Seven Devils Volcanics in the Riggins quadrangle are given in table 1. One specimen is a green metabasalt flow rock, and two are red-hued meta-andesites (one a flow rock and one a tuff). Major-element analyses are plotted on the diagrams of figures 43 to 47, and minor-elements analyses are plotted on the diagrams of figures 48 to 55.

The analyzed rocks have lost about half of their probable initial-magmatic contents of CaO. Present contents of CaO are 4 to 7 weight-percent, but magmatic contents of 8 to 10 percent or more were likely. (Spilitization is discussed in detail on p. 68–74). The content of Na_2O is higher in the metabasalt than the probable initial magmatic content of that component, which suggests its introduction from a nonmagmatic source.

The K_2O content of the two meta-andesites is similarly high, which might be taken as evidence for introduction of potassium, but the constant K : Ba ratio suggests a single source. This is elaborated in the discussion of the chemistry of the Riggins Group.

The red color of the two meta-andesites is reflected in their very high ratios of Fe_2O_3 to FeO. Presumably much of the manganese in these rocks is also in the oxidized, high-valence state. (Purple rocks are widespread in the Seven Devils Volcanics in the Salmon River Canyon north of the Riggins quadrangle, and they are present, although less commonly, within the quadrangle.)

The distribution of minor elements is, so far as is indicated by these few analyses, identical with that in the metavolcanic rocks of the Riggins Group. The Seven Devils data are plotted (figs. 43–55) with those from the Riggins Group, and the discussion (p. 54–59) of minor elements of the Riggins Group applies equally to the Seven Devils Volcanics.

AGE

No fossils were found in the Seven Devils Volcanics in the Riggins quadrangle; however, south of Shingle Creek, the rocks underlie the Martin Bridge Limestone of late Triassic age in apparent sedimentary sequence.

Ralph S. Cannon, Jr., collected many marine fossils from metatuffs and interbedded sedimentary rocks in the adjacent Cuprum quadrangle, and all diagnostic fossils indicated Permian and Late Triassic ages.

TABLE 1.—*Chemical analyses of Seven Devils Volcanics from the Riggins quadrangle and Clover Creek Greenstone from eastern Oregon*

[Major-oxide analyses 1-3 by rapid methods by P. L. D. Elmore, P. W. Scott, S. D. Botts, M. D. Mack, and J. H. Goode. Semiquantitative spectrographic analyses for minor elements by P. R. Barnett. Analyses for Clover Creek Greenstone from Gilluly (1937, p. 25)]

	Seven Devils Volcanics, Riggins quadrangle			Clover Creek Greenstone, eastern Oregon				
	1	2	3	A	B	C	D	E
Major oxides								
SiO_2	48.9	53.2	53.2	53.15	53.30	72.31	75.04	81.33
Al_2O_3	15.3	16.8	16.8	14.39	15.16	12.76	13.39	9.21
Fe_2O_3	8.5	9.5	9.0	1.28	2.54	1.94	1.61	1.09
sum as FeO	12.0	10.0	9.1	10.5	11.0	3.0	1.8	1.7
FeO	4.4	1.5	1.0	9.33	8.71	1.26	.37	.74
MgO	4.5	3.4	3.7	4.74	4.14	1.32	.18	.40
CaO	5.4	6.9	4.5	7.04	2.97	.10	.40	.25
Na_2O	5.3	2.8	3.3	4.58	5.55	3.75	6.36	3.25
K_2O	.04	1.7	2.6	1.01	.32	3.61	.83	1.66
TiO_2	1.4	1.0	1.4	1.50	2.41	.40	.10	.25
P_2O_5	.18	.25	.25	.19	.51	.15	.08	.04
MnO	.15	.18	.14	.14	.28	.08	.05	.05
H_2O	2.8	2.1	3.1	2.21	3.32	1.74	1.31	1.27
CO_2	2.5	.28	1.0	.10	----	Trace	.10	.10
FeS_2	----	----	----	----	.40	----	----	----
BaO	----	----	----	----	----	.09	----	----
Total	100	100	100	99.66	99.61	99.51	99.82	99.64
Minor elements								
B	<0.002	0.003	0.003	----	----	----	----	----
Ba	.0015	.03	.07	----	----	----	----	----
Co	.003	.003	.003	----	----	----	----	----
Cr	.003	.003	.003	----	----	----	----	----
Cu	.007	.0015	.0007	----	----	----	----	----
Ga	.0007	.0015	.0015	----	----	----	----	----
Mo	<.0005	.0007	.0007	----	----	----	----	----
Ni	.003	.0015	.003	----	----	----	----	----
Pb	<.001	.0015	.0015	----	----	----	----	----
Sc	.003	.003	.003	----	----	----	----	----
Sr	.015	.07	.07	----	----	----	----	----
V	.015	.015	.03	----	----	----	----	----
Y	.0015	.003	.003	----	----	----	----	----
Yb	.0003	.0007	.0007	----	----	----	----	----
Zr	.003	.007	.015	----	----	----	----	----
Field No	SR95-2	SR433-4	SR434-8	----	----	----	----	----
Laboratory Nos	D1396 151582	D1398 151584	D1399 151585	----	----	----	----	----

1. Metaspilitized basalt. Weakly foliated greenschist with small flattened lapilli. Contains about 50 volume-percent albite, 30 percent prochlorite, 5 percent each calcite (augenoid porphyroblasts, dark gray in hard specimen), epidote (grayish yellow in thin section), and magnetite (disseminated granules), and minor leucoxene, sphene and apatite. Anastomosing cleavage folia enclose augen that preserve the fabric of a diabasic basalt with plagioclase phenocrysts, although the rock has been completely reconstituted. Albite occurs as small poikilitic anhedra, relict laths, and sheared relict seriate phenocrysts. Prochlorite (green; ultra brown) is in flakes 0.02–0.05 mm long. Muscovite-chlorite subfacies assemblage. High Na_2O:CaO ratio is probably due to spilitization. Altitude 6,050 feet, Papoose Creek-Heavens Gate road.
2. Meta-andesite. Grayish-red crudely bedded crystal tuff containing abundant albitized relicts of angular plagioclase fragments in an extremely fine grained massive groundmass darkly obscured by magnetite and hematite. No schistosity, clasts undeformed. Most mineral grains are not identifiable, but chlorite, actinolite, epidote sericite, and calcite are present. Muscovite-chlorite subfacies assemblage. Rapid River trail, 0.4 mile south of West Fork.
3. Meta-andesite. Abundant, large (1–3 mm) euhedral phenocrysts of plagioclase (now albite crowded with minute sericite, calcite, and epidote) in a grayish-red aphanitic groundmass. The groundmass consists of felted 0.02 by 0.1 mm relict laths of plagioclase, now hazes of alteration minerals, separated by extremely fine material (altered glass?) rich in magnetite and hematite. Many meta-amygdules are ellipsoids with cores of calcite, medial shells of epidote and hematite, and rims of granular albite. The few mafic phenocrysts are now represented by granular epidote. Muscovite-chlorite subfacies assemblage. Rapid River, 0.25 mile south of West Fork.

A. Spilite, Baker quadrangle.
B. Albite diabase, Pine quadrangle.
C. Quartz keratophyre, Baker quadrangle.
D. Quartz keratophyre, Baker quadrangle.
E. Silicified quartz keratophyre, Baker quadrangle.

Rocks of the two systems are intercalated tectonically and are similar lithologically, and Cannon found no basis for mapping them separately. The Triassic fossils collected by Cannon and identified by S. W. Muller are listed in table 2. The only fossils found by Cannon in the eastern part (that nearest the Riggins quadrangle) of the Cuprum quadrangle were unidentifiable hexacorals that indicate a Mesozoic age.

Rocks of the Seven Devils Volcanics in the Riggins quadrangle are continuous with those dated by Cannon and are presumably correlative with them at least in large part.

TABLE 2.—*Fossils from rocks of Triassic age in Cuprum quadrangle and adjacent part of Riggins quadrangle*

[Collected by R. S. Cannon, Jr., and identified by S. W. Muller]

	Martin Bridge Limestone			Seven Devils Volcanics	
	West edge Cuprum quadrangle, lat. 45°04′ N.	West-central Cuprum quadrangle both sides Snake River, lat 45°07′ N.	Riggins quadrangle near Rapid River	Northwest part Cuprum quadrangle	Northeast part Cuprum quadrangle
Reptile bones		×			
Cephalopods					
Ammonites					
Arcestes				×	
Badiotites	×				
Juvavites	×				
Placites				×	
Protrachyceras				×	
Trachyceras cf. *aon*				×	
Like *Trachyceras*		×		×	
Nautiloids					
Nautilus	×				
Belemnoids					
Atractites	×			×	
Pelecypods					
Avicula		×		×	
"*Cardita*"	×	×		×	
"*C.*" cf. *beneckei*				×	
Cassianella		×		×	
Halobia	×			×	
Macrodon	?	×		?	
Myoconcha	?				
Myophoria cf. *costulata*	×				
Mysidioptera		×			
Pecten		×		×	
P., like *Amusium* with internal ribs	×				
P., like *Chlamys*	×			×	
P., a smooth form	×				
Posidonia	×				
Trichites	×	×			
Gastropods					
Naticopsis				?	
Palaeotriton or *Neritopsis*				?	
Unidentified		×			
Brachiopods					
Spiriferina, finely ornate	×				
S., coarsely ribbed	×				
Terebratula	×			×	
Crinoids					
Pentacrinus or *Isocrinus* stems	×		×	×	
Echinoids					
Cidaris spines	×	×	×		
Colonial corals					
Isastrea	×				
Hexacorals					
Montlivaltia				×	
Unidentified					×
Sponges					
Like *Amblysiphonella*	×				
Unidentified	×		×	×	
Foraminifera					
Like *Nodosaria* (*Dentalina*?)		×			
Like *Cristellaria* (*Astocolus*?)		×			

DISTINCTION FROM RIGGINS GROUP

Prior to metamorphism, rocks of the Lightning Creek and Fiddle Creek Schists of the Riggins Group must have been very similar to those of the Seven Devils Volcanics. However, the three main schist formations of the Riggins Group have no apparent correlatives in the Seven Devils Volcanics, and serpentine is limited to the Riggins Group.

The Rapid River thrust is nearly concordant with the Martin Bridge Limestone beneath it. Rocks of the upper plate are markedly truncated by the fault in some places, but not in others; and where relatively low-grade rocks of the Riggins Group overlie the fault semiconcordantly, as they do in places in the west-central part of the quadrangle, the fault cannot be located with confidence. Errors in designations, and accordingly in structural interpretation, may have been made south of 45°15′ N.

MARTIN BRIDGE LIMESTONE

DISTRIBUTION

Sheared and recrystallized gray limestone of Late Triassic age crops out in an east-dipping belt 30 miles long near the west edge of the Riggins quadrangle (pl. 1). The belt is thin along the east side of the Rapid River and thicker to the north. Just north of the quadrangle (pl. 2), the limestone belt turns eastward and crosses the Salmon River south of Lucile, 2 miles north of the Riggins quadrangle. At lower altitudes the limestone forms widespread outcrops and many cliffs, such as those rising from the Salmon River at Lucile; but at higher altitudes, outcrops are few and small.

The limestone also crops out in other discontinuous bodies not shown on plate 1; most are in the area about Cannonball Mountain and the West Fork of Rapid River. The limestone, which is intercalated tectonically with metavolcanic rocks, is crossed by the Heavens Gate–Cannonball Mountain–Silver Guard Station trail for 200 yards at an altitude of 6,725 feet, in the saddle at 6,600 feet, and between altitudes of 5,500 and 5,300 feet. These exposures are poor, as the rock is hidden by thick soil and colluvium that support dense forest. Several miles to the south, the Martin Bridge Limestone crosses the West Fork of the Rapid River west of Bridge Creek; here also it is intercalated between greenstones. The limestone of the Cannonball–West Fork area is gray, aphanitic to very fine grained, and little sheared.

DEFINITION AND AGE

The Martin Bridge Formation was named by Ross (1938, p. 32–36) in the Wallowa Mountains, 35 miles west of the southwest corner of the Riggins quadrangle. (Gilluly, Reed, and Park, 1933, p. 12, referred to Ross' definition before it was published.) Ross described the formation as consisting of limestone with subordinate limy shale and tuff, and as having a thickness of 1,000 to 3,000 feet assuming that there was no tectonic thickening. The formation contains abundant Upper Triassic fossils.

As the limestone of the Cuprum quadrangle also carries Upper Triassic fossils, I presume it to be correlative with the Martin Bridge Formation of eastern Oregon. It is possible that future work will demonstrate a different correlation, but the established name is used here in preference to introducing a new formation name.

The only fossils found in the limestone within the Riggins quadrangle during the present study merely indicate a probable Triassic age. D. G. Sherlock found pentagonal crinoid stems in aphanitic unsheared limestone along the West Fork of the Rapid River above Bridge Creek. The stems were referred to *Isocrinus* or *Pentacrinus* by John B. Reeside, Jr., who suggested the probable Triassic age.

W. R. Wagner collected poorly preserved fossil fragments along the Papoose Creek road in the Riggins quadrangle and at Lucile just north of the quadrangle. These fossils were studied also by Reeside, who identified some of them as spines of cidaroid echinoids and suggested a highly probable Triassic age (Wagner, 1945, p. 15).

The continuity of the thin belt of the formation south of Rapid River suggests strongly that its base is a depositional contact with the underlying Seven Devils Volcanics. Cannon found the limestone-volcanics contact to be conformable in the southwestern part of the Cuprum quadrangle. (Cook, 1954, p. 4, termed the contact an unconformity but cited no evidence.) The upper contact, with the overlying slate, also appears conformable south of Rapid River. North of the Rapid River the structure becomes increasingly complex.

LITHOLOGY

The mapped belt of Martin Bridge Limestone is composed largely of metalimestone (limestone, schistose limestone, and foliated marble), but it contains intercalations, mostly north of the Rapid River, of both gray slate (or phyllite) and greenstone (or fine greenschist). The lower contact of the limestone north of Papoose Creek is a zone of tectonic intercalation. Cannon found that the nearly complete and little deformed section of the Martin Bridge at the Snake River in the Cuprum quadrangle contained practically no clastic rocks, but that it had been intruded by several andesitic sills. Presumably some of the greenstone intercalated with the limestone in the Riggins quadrangle also represent sills, but most of the interlayered slate and greenstone probably lies tectonically within the limestone. Greenstone intercalations are widespread only north of Papoose Creek, where the structure of the limestone is complex but poorly exposed. Slate and limestone are thinly interlayered, apparently stratigraphically, at some places in the Riggins quadrangle.

The general intensity of shearing and the coarseness of grain both decrease southward in the main belt of limestone. Along the upper reaches of the Rapid River, most of the limestone is aphanitic and scarcely sheared. Its thickness is near 1,000 feet and presumably approximates the initial sedimentary thickness. Where it is crossed by the Rapid River, the limestone is finely schistose—aphanitic or fine grained, but much sheared. From Papoose to Kessler Creeks, little-sheared limestone, calcite-mylonite, and fine- to medium-grained schistose marble are interlayered. The proportion of marble increases northward, and at Lucile, just north of the quadrangle, shear-layered marble is the dominant rock; aphanitic calcite mylonite, superfically resembling slate, is common. The belt dominated by (and mapped as) limestone in the Papoose Creek-Kessler Creek area must be about 1 mile thick; folding younger than the main episode of tectonic intercalation has complicated the structure, and the upper (eastern) contact of the belt is mostly hidden by Columbia River Basalt.

The limestones and marbles are varied grays, mostly neutral or faintly bluish, that are lighter on weathered than on fresh surfaces. The least metamorphosed rocks are aphanitic, uniform or faintly laminated, and massive to platy. With increasing metamorphic effects, the grain size increases, and the rocks show shear lamination as sugary white or light-gray layers and lenses alternating with thinner and darker anastomosing laminae a few millimeters to several centimeters apart. Buff-colored limestone is uncommon.

The average grain size of the limestone and marble varies from about 0.005 mm, common in the south, to 1 or even 2 or 3 mm, common at Lucile. Calcite crystals generally form a mosaic of alined stubby lenses. Magnetite, carbon, pyrite, clastic quartz and feldspar, and recrystallized muscovite make up less than 1 percent of most rocks; the rest is pure calcite. Veins of white calcite are widespread, and veins of white quartz were noted in a few places.

The metalimestones are exclusively calcitic. The analysis of a composite sample (table 3) shows less than 1 percent magnesium. All of the 50 specimens collected effervesce violently with hydrochloric acid. None of the thin sections examined have the rhombs characteristic of dolomite in calcite limestones.

TABLE 3.—*Semiquantitative spectrographic analysis, in percent, of composite sample of six chips of calcite marble and schistose limestone*

[Analyst, P. R. Barnett; laboratory No. D 1409. Elements not detected are not listed. Martin Bridge Limestone, along Papoose Creek-Heavens Gate road between altitudes of 4,400 and 5,300 feet]

Si	0.3	Ba	0.0007
Al	.3	Cr	.0007
Fe	.15	Cu	.0003
Mg	.7	Mn	.015
Ca	Major	Ni	.0003
Na	.15	Sr	.15
Ti	.007		

Limestone-pebble conglomerate is a common rock type. The clasts had a common initial diameter of 1 centimeter or so, but they have been markedly flattened in most of the area. Where the rock remained aphanitic—that is, where the recrystallization accompanying shearing was minor—the clasts are preserved as flat triaxial ellipsoids (fig. 6). In such rocks, gray limestone clasts were deformed to triaxial ratios such as 1:2:5; buff limestone clasts were much less deformed, with ratios more like 1:1.5:2.5. Both color types are calcite-limestone.

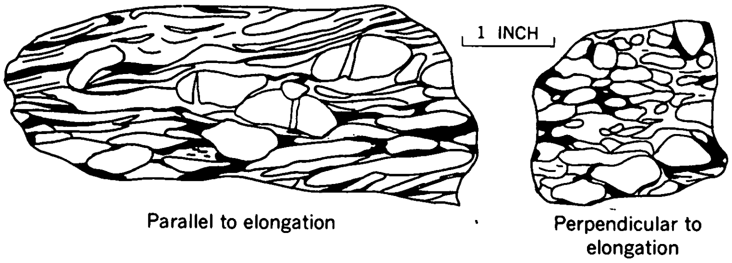

FIGURE 6.—Sheared limestone conglomerate with triaxially deformed pebbles. Collected at an altitude of 5,500 feet on Papoose Creek–Heavens Gate road.

Where recrystallization was more pronounced and the grain size approaches 1 mm, the clasts are recognizable only as pointed triaxial lenses of sugary white marble separated by darker folia. Still more strongly sheared and recrystallized rocks are all shear laminated, and their primary structures are obliterated.

ROCKS NORTH OF RIGGINS QUADRANGLE

At the north edge of the Riggins quadrangle, foliated marble of the Martin Bridge Limestone is exposed on the east wall of the Salmon River Canyon in an isoclinal anticline through the Fiddle Creek Schist and the folded Rapid River thrust (pl. 2).

The main outcrop belt of the formation, also foliated marble, is crossed by the Salmon River a mile north of the quadrangle and forms high cliffs. Structure is exceedingly complex. On the high east wall of the canyon, the marble, or thick slices of it, is repeated about six times by isoclinal folds and bedding-plane thrust faults and is intercalated with Lucile Slate and Seven Devils Volcanics in varying combinations. Comparable structures are exposed poorly west of the river. The formation was subjected to extreme deformation subparallel to bedding, contemporaneous with metamorphism or with displacement on the Rapid River thrust, and these structures were deformed in turn by strong folding after thrusting.

Southeast of the mouth of Slate Creek, several limestone layers, each about 500 feet thick, are intercalated in the Seven Devils Volcanics. The beds are of light-gray medium-grained marble like the Martin Bridge Limestone of the Riggins quadrangle. Because of their considerable thickness and their dissimilarity to anything in the Seven Devils Volcanics to the south, it is more likely that these layers are tectonic intercalations of Martin Bridge Limestone than that they formed as strata within the Seven Devils Volcanics.

The main belt of the Martin Bridge Limestone trends northeast from Lucile and dips moderately southeast. Where the belt is crossed by the road going southeast from Slate Creek, calcareous quartz schist (Lucile Slate; see below) is intercalated with foliated marble of nearly pure calcite and with slightly impure calcite schist. These are directly overlaid from the southeast by intrusive sugary trondhjemite gneiss.

Where the marble belt is crossed by Slate Creek, it consists, as farther south, of intercalated Martin Bridge and Lucile types. (The intrusive contact with sugary trondhjemite gneiss, above and southeast, is excellently exposed on the Slate Creek road and is described in another report.) Calc-silicate gneisses were produced as concordant septa in the immediate contact zone, and calc-silicate minerals were developed sparsely for several miles to the west. Most of the formation that is more than a few tens of feet from the gneiss contact, however, is nearly pure calcite marble that crops out in many bluffs and cliffs. Intercalated with the marble in the mapped belt are carbon-clouded variably calcareous biotite-muscovite-quartz schists that probably represent the Lucile Slate. More marble and dark calcareous quartz schist are intercalated in the Seven Devils Volcanics along Slate Creek, west of the belt of Martin Bridge Limestone indicated on the map.

The Martin Bridge Limestone was not seen along the South Fork of the Clearwater River, 20 miles northeast of the Slate Creek exposures.

ROCKS OF CUPRUM QUADRANGLE

The Martin Bridge Limestone crops out in several parts of the Cuprum quadrangle (pl. 3). Ralph S. Cannon, Jr. (written communications, 1953, 1960), provided all the following unpublished information.

The Martin Bridge Limestone is best exposed in an open syncline trending southwest across the west edge of the Cuprum quadrangle, where a section of more than 1,500 feet largely calcitic and only slightly recrystallized limestone crops out in prominent gray bluffs on both sides of the Snake River. A stratigraphic section of the Martin Bridge Limestone, measured by plane-table methods on the spur south of Kinney Creek just east of the Snake River, where the formation has gentle southward dip, is as follows:

678-769 O—63——2

Top of formation eroded.

Martin Bridge Limestone:	Thickness (feet)
9. Limestone, black and aphanitic; in beds 5–10 ft thick; forms cliffs	150
8. Limestone, dark- to light-gray, aphanitic to finely crystalline; in beds 1–3 ft thick	170
7. Limestone, buff-gray and aphanitic; in beds 1–3 ft thick	110
6. Limestone, gray, aphanitic to finely crystalline; weathers light gray with white markings; in beds averaging about 1 ft thick. Minor siliceous and dolomitic limestone	150
5. Limestone, massive; generally aphanitic in upper part and finely crystalline in lower part; mostly black, gray, and buff, but locally pink; forms cliffs. Cut by feldspathic sill, near which the limestone is more coarsely crystalline than elsewhere	405
4. Limestone and dolomite, finely crystalline, thin-bedded, carbonaceous; some black with white markings; laminated; platy weathering. Cut by three sills of punky feldspathic gray-green rock	108
3. Limestone, black, aphanitic to finely crystalline, thin-bedded; some chert; slaty cleavage	140
2. Limestone; buff and granular in lower part, black and aphanitic in upper part; all has slaty cleavage. Contains bed of pelecypods, crinoids, sponges, and stromatopora near center—the only good fossils seen in this measured section	100
1. Limestone and dolomite, buff to pink, massive and granular; weak slaty cleavage	105
Total	about 1,700

Base of section: depositional contact with underlying green tuff, which is 20 feet thick and in turn overlies maroon limy tuff breccia.

The best collection of fossils made from the limestone came from a small fault slice at lat 45°04′ N., long 116°45′ W., at the west edge of the Cuprum quadrangle, at an altitude of 4,350 feet on the Indian Creek–Snake River divide. Here, 15 species of pelecypods, cephalopods, and brachiopods were found. These and other fossils found by Cannon and identified by S. W. Müller are listed in table 2.

LUCILE SLATE

DEFINITION AND DISTRIBUTION

Dark-gray calcareous slate overlies the Martin Bridge Limestone along the Rapid River and is intercalated tectonically with the limestone in the northern part of the Riggins quadrangle and beyond to the north. The slate is here named for the hamlet of Lucile on the Salmon River (pl. 2). Wagner (1945, p. 5) designated as Lucile Series all the rocks along the Salmon from Lucile to Riggins and beyond to the south and east. In so doing, he lumped as one stratigraphic unit the rocks beneath the Rapid River thrust, here termed Martin Bridge Limestone and Lucile Slate, and the diverse formations of the Riggins Group above the thrust. Furthermore, the name Martin Bridge has priority for the limestone. The term Lucile is accordingly redefined as restricted to the slate formation. The type locality is designated as the highway cuts several hundred yards south of Lucile, where the slate is well exposed.

The slate has yielded no fossils. As it overlies the Martin Bridge Limestone along the Rapid River, where the rocks are least deformed, it is considered to be in stratigraphic sequence. If this is correct, then the slate cannot be older than Late Triassic, and is considered to be Triassic(?). Near Lucile the slate is repeated structurally several times and lies between sheets of Martin Bridge Limestone and Seven Devils Volcanics. The original thickness was perhaps a few hundred feet.

Rocks designated doubtfully as Lucile Slate occur also on Whitebird Ridge between the Rapid and Little Salmon Rivers.

LITHOLOGY

The main rock type of the formation is dark-gray slate, although grayish-green slate is common in the south. The rock generally disintegrates to fine soil-covered rubble, so that outcrops are few except along streams. Bedding is almost completely obscured by slaty cleavage or foliation. The cleavage commonly has a consistent attitude; but in many outcrops the cleavage is folded, crumpled, and obliquely sheared.

The rocks are composed largely of quartz, muscovite, calcite, and carbon, named in general order of decreasing abundance. A colorless chlorite is common; and albite, epidote, and actinolite are uncommon. The rocks probably originated as mixtures of clay and quartz silt, generally carbonaceous and variably calcareous, with admixtures of volcanic material in places.

Gray calcite limestone is present as intercalations 1 foot to many feet thick in many parts of the formation; it has been sheared and recrystallized to finely crystalline limestone, calcite mylonite, calcite chist, and foliated marble. The slate contains flattened pebbles of limestone locally.

The slate and slaty phyllite in the highway cuts at Lucile are submetallic, medium gray, finely laminated parallel to the brilliant cleavage, and lack any trace of bedding. Abundant veinlets of white calcite lie subparallel to the cleavage. This slate consists of flattened quartz (mostly smaller than 0.01 by 0.02 mm), variably abundant carbonate, and well-oriented muscovite that is concentrated in carbon-dusted anastomosing microshear folia. Many of the rocks contain colorless chlorite. Many also contain augen—mostly poorly shaped rhombs—of iron-stained noncalcitic carbonate, and some of the phyllites are conspicuously knotted by them.

In the Race Creek-Kessler Creek area, the slate is finer grained, with microshear folia as close as 0.005 mm in some specimens. Carbon dust is ubiquitous. Some slate contains calcite, and some contains 1-mm rhombs of ankerite or siderite that weather to a rusty color. Leucoxene and pyrite are common minor components. The thin lenticular laminae vary in composition, grain size, and degree of orientation of mineral grains; most are parallel to cleavage.

From the Rapid River south, the rocks are mostly slaty phyllites that are coarser than the rocks noted above but less sheared, having wider spaced microshear folia. Bedding laminae are visible in places at a high angle to, and but little sheared by, cleavage folia. The common grain size of these rocks is near 0.1 mm. Quartz and muscovite remain the dominant minerals; chlorite is widespread; and albite, carbon, magnetite, tourmaline, carbonates, and titanium minerals are present in varying proportions in some specimens. Actinolite, albite, chlorite, and epidote are abundant in the grayish-green rocks and apparently reflect an origin as reworked tuffs, whereas the gray rocks originated as non-volcanic sediments.

ROCKS NORTH OF RIGGINS QUADRANGLE

Tectonic intercalations of Lucile Slate occur within the belt of Martin Bridge Limestone northeast of Lucile (pl. 2). Where this belt is crossed by the road southeast of the mouth of Slate Creek, the marble contains much fine-grained schist probably derived from Lucile Slate. The schist is dark gray and consists of carbon-dusted muscovite-quartz-carbonate schist with or without aluminian prochlorite. The carbonate includes both calcite and iron-bearing types. A higher grade of metamorphism is represented here—near and at the contact with intrusive trondhjemite gneiss—than at any place in the formation within the Riggins quadrangle.

GRANITIC ROCKS INTRUSIVE INTO SEVEN DEVILS VOLCANICS

Two plutons of metamorphosed quartz diorite are exposed within the Seven Devils Volcanics along the Rapid River. The southern area, shown on the geologic map (pl. 1), is the exposed end of a large pluton that lies mostly within the adjacent Cuprum quadrangle. The other pluton, not distinguished on the map, crops out by the Rapid River 4.7 airline miles from its mouth and yields talus for 0.3 mile farther south; it was not traced away from the river. Similar plutons may lie in the large area west of the Rapid River and south of Cannonball Mountain, which was not visited during the fieldwork.

The metaquartz diorites are massive rocks that are unsheared although thoroughly reconstituted. The rocks still look granitic in hand specimen, although plagioclase is porcelaneous, if white, or light grayish green; mafic minerals are grayish green to dark greenish gray. Primary plutonic fabric is clearly visible in thin section: the rocks were equigranular and consisted largely of intermediate plagioclase in 0.5- to 3-mm subhedra, 20 to 30 percent quartz, and 10 to 20 percent hornblende and biotite. The plagioclase has been pseudomorphed by inclusion-crowded albite; the quartz has, recrystallized thoroughly; the hornblende has been replaced by actinolite; and the biotite has been altered to mazes of chlorite, white mica, and other secondary minerals. The description given below of an analyzed specimen is largely applicable to the rest of the metaquartz diorite also.

Only the north contact of the northern pluton was seen; it is exposed by the Rapid River trail 0.9 mile north of the West Fork. The contact is extremely sharp, with relict crystals of the altered quartz diorite, which is like that in the interior of the pluton, simply nestled against greenstone. There is no evidence of interaction in either outcrop or thin section. The greenstone is unsheared and very fine grained; there is no trace of hornfels fabric, although the subsequent metamorphism might have destroyed such a fabric were it very fine grained. The quartz diorite and greenstone seem to have undergone the same degree of metamorphism, and this indicates that the quartz diorite was emplaced before regional metamorphism.

A small amount of comparably altered diorite was found mixed with slate, marble, and greenstone in float along the trail 0.3 mile north-northeast of the mouth of Frypan Creek.

A pluton of altered quartz diorite lies within Seven Devils Volcanics in the northwest corner of the area. The mass was traversed along the north edge of the Riggins quadrangle, and its contacts were mapped northeastward (pl. 2) several miles beyond the quadrangle. The pluton is poorly exposed, but the outcrops seen are of rather uniform rock. Wallrocks are greenstones that show no conspicuous contact metamorphism and that are cut sharply by many dikes of quartz diorite. The quartz diorite is a uniform light-gray rock speckled by small (typically about 1 by 3 mm) prisms of hornblende that comprise about 6 to 10 percent of the rock. Plagioclase, in well shaped tablets near 2 mm long, shows marked normal-oscillatory zoning and is much altered to white mica, subordinate epidote (?), and actinolite(?). The plagioclase is porcelaneous in hand specimen and shows a faint green cast in some specimens, none in others. The hornblende is greenish black in hand specimen and pale green in thin section; it is now actinolitic, and presumably formed by alteration of originally dark hornblende. Brown biotite—produced partly by alteration of hornblende, although some appears primary—is much altered to epidote and white mica, but chlorite is lacking. Quartz shows incipient recrystallization.

Light-colored cataclastic gneiss, considered to be sheared and reconstituted quartz monzonite, occurs as a sliver in the Rapid River thrust zone at the intersection of the Rapid River and Shingle Creek roads. Whether this rock lay originally in the rocks above or beneath the thrust could not be determined.

An analysis of a specimen of metaquartz diorite from the northern of the two plutons along the Rapid River is given in table 4. The composition is like that of quartz dioritic rocks east of Riggins and is not discussed separately here.

TABLE 4.—*Composition, in weight percent, of metaquartz diorite, intrusive into greenstone, from east side of Rapid River 0.9 mile north of mouth of West Fork*

[Specimen described in text. Field No. SR 438A-1, laboratory Nos. 148427 and C669. Major-oxide analysis made by rapid methods by P. L. D. Elmore, P. W Scott, S. D. Botts, and K. E. White; minor-element analysis by semiquantitative spectrographic methods by P. R. Barnett. Elements not detected are not listed]

Major oxides		Minor elements	
SiO_2	64.0	B	0.0015
Al_2O_3	17.4	Ba	.03
Fe_2O_3	2.0	Co	.0015
FeO	1.8	Cr	.007
MgO	2.5	Cu	.0007
CaO	5.3	Ga	.0015
Na_2O	4.3	Ni	.003
K_2O	.87	Sc	.0015
TiO_2	.52	Sr	.07
P_2O_5	.14	V	.015
MnO	.06	Y	.0015
H_2O	1.4	Yb	.00015
CO_2	.14	Zr	.007
Total	100		

The analyzed specimen differs in appearance from unmetamorphosed quartz diorite in that its plagioclase is porcelaneous and its mafic minerals are dark green. The plutonic fabric is well preserved, although all minerals have been recrystallized or reconstituted. Plagioclase ($\approx$60 percent) retains its subhedral shape in 0.5- to 2-mm crystals that are now single-crystal pseudomorphs of albite densely crowded by inclusions of pale-green mica and other secondary minerals. Quartz ($\approx$24 percent) has recrystallized into large crystals that enclose many subhedra each of plagioclase; it is clear and but little strained. Single-crystal pseudomorphs of actinolite ($\approx$7 percent) largely preserve the short-prismatic habit of the primary hornblende. There is a small amount (0.2 percent) of relict biotite, and much more ($\approx$8 percent) of biotite pseudomorphs that are now minutely lamellar intergrowths of prochlorite(?), muscovite, and quartz, studded by epidote and sphene. Opaque minerals total about 0.3 percent of the rock.

GRANITIC INTRUSIVES NORTH OF RIGGINS QUADRANGLE

Many dikes and small masses of metamorphosed granitic rock are present in the Seven Devils Volcanics along the Salmon River from Lucile to Slate Creek and along lower Slate Creek. Many of these granitic rocks are highly sheared and reconstituted. Compositions vary from quartz diorite through granodiorite to quartz monzonite, on the one hand; and from quartz diorite to trondhjemite (light-colored biotite-oligoclase "quartz diorite") on the other. Colors of the resultant metamorphic rocks vary correspondingly from grayish green in the intermediate ones to pinkish gray, mottled by green, in the potassic ones, or light gray in the sodic rocks. Some of the rocks preserve their granitic fabrics, although they are partly masked by recrystallization and reconstitution. Others are now mortar gneiss in which the margins between primary crystals were ground down cataclastically and then variably recrystallized. Most widespread is flaser gneiss—streaky, foliated, strongly lineated, variably recrystallized cataclastic gneiss. Phyllonite, the product of extreme shearing, can be distinguished from metavolcanic greenschist principally by the kind of interlayered rocks. In each of the sheared rock types, folia of finely crushed material separate lenses and augen of less crushed rock; and within such lenses are variably preserved relicts of the original fabric.

Quartz in these metagranitic rocks was recrystallized into sutured, shingled aggregates of extremely strained elements whose *c* axes lie in the plane of cataclastic foliation. In some places plagioclase was pseudomorphed by inclusion-clouded albite. In others, it was made more sodic, but not albitic, and its inclusions of secondary minerals are sharp euhedral crystals of mica and aluminian epidote; this seems to represent a higher grade of reconstitution. Potassic feldspar was converted partly to muscovite and partly to coarse perthite. Some primary(?) biotite remains, but no hornblende; the mafic minerals have been reconstituted chiefly to epidote, chlorite, and sphene. Much iron was oxidized to the ferric state during the metamorphism, as shown by the abundant epidote, magnetite, and hematite dust.

In Box Canyon of the Salmon River, 8.5 miles north of the Riggins quadrangle, metagranitic rocks are particularly well exposed in cliffs and high roadcuts. The dominant rock is pink quartz monzonite that is thoroughly recrystallized and variably sheared; this is cut by abundant dikes of quartz diorite (now mostly dark flaser gneiss) and basalt(?) (now greenschist). The quartz monzonite was thus cut by mafic fine-grained dikes before it was metamorphosed. No such dikes are known in any of the rocks marginal to the Idaho batholith. This and the other masses of metagranitic rocks exposed along the Salmon River (pl. 2) are in long lenses concordant with the structures of the enclosing greenschists, and at least one of these lenses is bounded by phyllonitic gouge; presumably the lenticular character is due to shearing and thrust faulting of the granitic rocks at the time of their metamorphism, not to any original intrusive shape.

Similar metagranitic rocks, now mostly flaser gneiss of low-grade mineralogy, lie within the Seven Devils (?) volcanics along the South Fork of the Clearwater River from Harpster to 8 miles upstream. These provide part of the basis for the designation given the enclosing metavolcanic rocks, as they indicate a similar complex history of intrusion and metamorphism different from that of the Riggins Group. Postmetamorphic granitic rocks intrude these metavolcanic rocks along the Clearwater River both at their northern limit at Harpster and at their southern limit 8 miles upstream.

CUPRUM QUADRANGLE

Many masses of plutonic rocks intrusive into the Seven Devils Volcanics in the Cuprum quadrangle, adjacent on the west to the south half of the Riggins quadrangle, were mapped by Ralph S. Cannon, Jr. (pl. 3). The following information, previously unpublished, presents some of Cannon's (written communication, 1953, 1960) conclusions as well as brief descriptions based on study of his thin sections.

The plutonic rocks are divisible into three major structural and lithologic assemblages: a pretectonic or early tectonic suite of gabbro, diorite, and quartz diorite; a syntetonic suite of granodiorite; and a posttectonic quartz diorite. (All these assemblages are older than the main metamorphism in and north of the Riggins quadrangle.) The two earlier suites of plutonic rocks are concentrated along a zone trending northeastward across the Cuprum quadrangle, a zone that also contains the most highly sheared metavolcanic rocks. The posttectonic quartz diorite is an east-trending body that cuts across this zone of older rocks.

The pretectonic intrusives are pods of metamorphosed gabbro, quartz gabbro, diorite, and quartz diorite that are elongated northeastward and dip steeply. They are concordant with the general structure of the metavolcanic rocks they intrude and are concentrated along zones of intense shearing. Contact metamorphic effects are negligible. The rocks are variably sheared and moderately reconstituted. Plagioclase—although it is still as calcic as sodic labradorite in metagabbro and commonly preserves oscillatory zoning—is considerably altered to sericite, epidote or clinozoisite, and amphibole or chlorite. Pyroxene is generally lacking but presumably was the dominant initial mafic mineral; in its place is anhedral amphibole, light colored in thin section, itself partly altered to biotite and magnetite. Veinlets of epidote, quartz and magnetite, and of alkali feldspar, fill abundant cracks.

Shearing effects vary widely; some rocks preserve their plutonic fabrics, others are mortar gneiss, and others are flaser gneiss or even mylonite. Only exceptionally have the primary subhedral shapes of plagioclase grains been destroyed by recrystallization or shearing. Quartz shows much recrystallization and, in sheared rocks, is in elongate sutured grains with a preferred orientation, the *c* axes lying near the foliation plane. Hornblende and biotite are ragged in some rocks.

The syntectonic granodiorite occurs with the altered mafic rocks in similar structural positions but shows less reconstitution. The granodiorite occurs as concordant lenses of massive rocks. Near contacts the granodiorite may be foliated, and locally it cuts across wallrock foliation. In the normal facies, plagioclase, markedly zoned from An_{35} to An_{20}, with thoroughly saussuritized cores, is generally undeformed except for sparse bent grains. The 10 percent or so of orthoclase is in anhedral grains and stringers. Abundant quartz is interstitial to the well-formed plagioclase crystals and has crystalloblastic textures with elongate interlocking crystals. Most of the hornblende is subhedral and is apparently a primary mineral.

The large mass of biotite-hornblende quartz diorite in the central and east-central part of the quadrangle is discordant and posttectonic. It has a contact aureole as much as 1 mile wide within which both the metavolcanic rocks and the older intrusive rocks have been hornfelsed. Large xenoliths of Martin Bridge Limestone have been recrystallized and silicated. The commercial contact-metamorphic deposits of copper, molybdenum, and tungsten in the Seven Devils Montains were formed by this stock, mostly in the limestone xenoliths near its margins. Wallrocks are gneissic and much injected for some tens of feet from the contact, but contacts are generally sharp. Within a hundred feet of the contact, both intrusive rocks and wallrocks are strongly foliated, with steep outward dips and downdip lineation; outside this zone a weak concentric biotite schistosity was produced. Plagioclase of the quartz diorite is oscillatory-zoned andesine in slightly sericitized subhedral grains. Hornblende is generally well formed and is partially replaced by biotite, and both are partially chloritized. Quartz (about 20 percent of the rock) and the minor orthoclase are anhedral and interstitial. The eastern part of this pluton was subjected to slight regional metamorphism: the pluton is pretectonic with respect to the deformation and metamorphism in the Riggins quadrangle.

Smaller masses of unaltered rocks are composed of augite-andesine diorite, biotite-hornblende-andesine diorite, and biotite-hornblende quartz diorite, granodiorite, and quartz monzonite. The total volume of these is small. Neither trondhjemite nor granite has been recognized in the Cuprum quadrangle.

METAMORPHIC HISTORY

Most of the granitic rocks that intrude the Seven Devils Volcanics in the areas described have been metamorphosed. (By contrast, most granitic rocks intrusive into the Riggins Group have not been metamorphosed.) In the Cuprum quadrangle, some of the intrusives predate, whereas others postdate, the major episode of metamorphism there. Effects of this metamorphism are most pronounced in a zone trending northeastward to the northeast corner of the quadrangle. The large pluton of quartz diorite, which is younger than the metamorphism of this zone and is unmetamorphosed near the zone, has been in turn metamorphosed in the Riggins quadrangle and presumably also in the eastern part of the Cuprum quadrangle. Along the Salmon River north of the Riggins quadrangle the intrusives in the Seven Devils Volcanics have been severely sheared as well as thoroughly recrystallized, this metamorphism having occurred after the injection of fine-grained mafic dikes. East of the Salmon River, as along Slate Creek and the Clearwater River, postmetamorphic granitic rocks intrude Seven Devils Volcanics and allied formations that contain metamorphosed grantitics.

At least two major episodes of metamorphism are indicated by this information. The older episode affected the rocks beneath the Rapid River thrust but probably not the rocks above. Metamorphism was widespread but of low grade, although intensity of shear varied greatly. Granitic rocks were intruded both during and after this episode.

The second episode of metamorphism affected only the eastern rocks beneath the Rapid River thrust. The grade of this metamorphism increases eastward, from low (as along the Rapid River) to very high in the broad gneiss zone marginal to the Idaho batholith. The Rapid River thrust postdates this metamorphic event.

Reasons are elaborated later in this report for considering the first metamorphism to be Jurassic(?) and the second to be Cretaceous(?), and for considering this area to include the eastern part of a broad Jurassic orogen and, overlapping it, the western part of a broad Cretaceous orogen.

ROCKS ABOVE RAPID RIVER THRUST

RIGGINS GROUP

Lying tectonically above and east of the Seven Devils Volcanics and other formations of the Permian and Triassic sequence is a thick group of formations of metamorphic rocks, dominantly schists. These are here named the Riggins Group. The rocks are of low metamorphic grade in the northwest; eastward and southward, the rank of regional metamorphism increases progressively through middle and high grades. A complete transition from schists to plutonic gneisses is exposed along both the Salmon and Little Salmon Rivers, and the distinction made here—the basis for division of description and map units of "metamorphic" and "plutonic" rocks—is arbitrary. In the "metamorphic" rocks, formations generally can be traced with some confidence and are structurally continuous. In the "plutonic" rocks, the same formations are probably present, but so high a degree of structural mobility was attained during metamorphism and deformation that continuity was lost and stratigraphic assignments are no longer possible. Also, granitic materials—quartz diorite, granodiorite, quartz monzonite, trondhjemite—become increasingly abundant as concordant and discordant intrusions eastward and southward. Major faults cut out part of the gradational interval between "metamorphic" and "plutonic" rocks near the Rapid River.

The Riggins Group is in a syncline whose axis trends northwestward almost through the town of Riggins. The northeast limb is composed of a nonrepeating sequence of distinctive formations; the southwest limb is more complex but contains the same rock types in the same succession. The present sequence is probably at least partly structural rather than stratigraphic.

The following southwestward-dipping section, here designated the type section, is excellently exposed along the Salmon River from Riggins northward to a little beyond the edge of the Riggins quadrangle. Formations are listed from lowest to highest. The stratigraphic names are new, and the thicknesses are the present ones, which have been complicated by so much deformation that they give little information as to original thicknesses.

Fiddle Creek Schist	Metamorphosed tuff and lava; now greenschist and white schist. Thickness, 9,000 ft; base not present.
Lightning Creek Schist	Metamorphosed lava, tuff, and agglomerate; now greenschist. Thickness, 8,000 ft.

Metaperidotite	Unnamed. Includes lesser amounts of metavolcanic and metasedimentary rocks; about 1,200 ft thick; lies within the Riggins Group in this position in the type section.
Berg Creek Amphibolite.	About 1,000 ft thick; lies between the Squaw Creek and Lightning Creek Schists east of Riggins, where the metaperidotite is not present, but is not present in the type section.
Squaw Creek Schist	Laminated gray schist of sedimentary and volcanic-sedimentary origin. Thickness about 6,000 ft; top not present.
Total present thickness of section.	About 20,000–25,000 ft.

The distinction between these formations becomes increasingly difficult as their structural complexity and metamorphic grade increase. Bulk compositions of the rocks of the three schist formations vary less than the very different appearances of their low-grade representatives suggest, and the high-grade rocks are much more similar than are the low-grade ones. In the section along the river north of Riggins, the gray schist of the Squaw Creek is completely unlike the green and white schist of the other two formations. White schist is confined to, and thus characterizes, the Fiddle Creek, but some of its greenschist is indistinguishable from that of the Lightning Creek.

The formations are broadly concordant (fig. 7). Bedding-plane shear has been extreme in this area, however, and it is not known whether this sequence is stratigraphic or tectonic; nor is it known which direction is stratigraphically upward, if the sequence is a depositional one, as the entire group could be overturned.

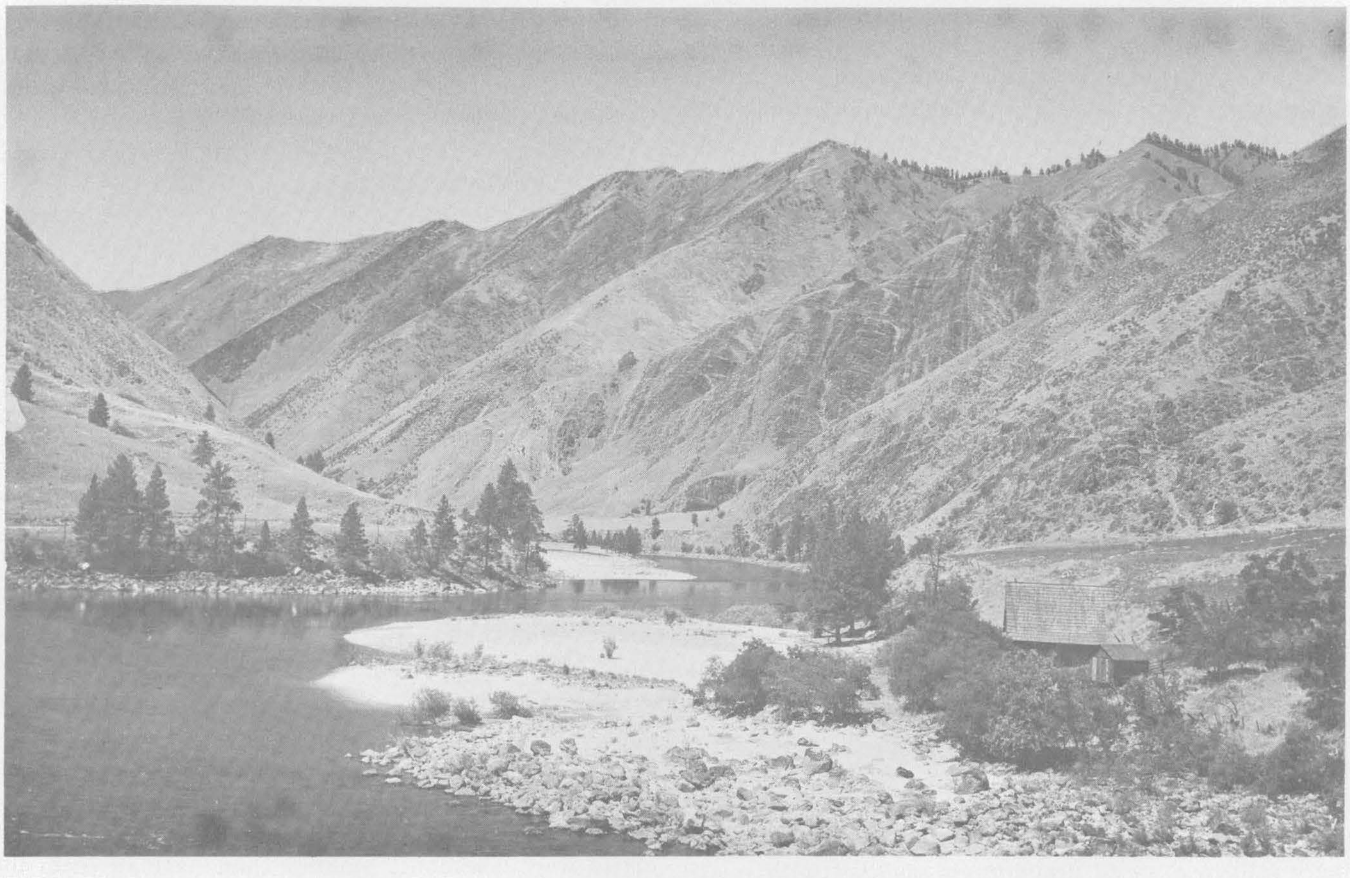

FIGURE 7.—Schist of Riggins Group along the Salmon River. Dip is about 40° to left (southwest) on left and across all the skyline, but layers roll to vertical in center above river and dip steeply to the right beyond the barn. View to northwest across fan at mouth of Berg Creek. Dark rocks of Berg Creek Amphibolite form center and right ridges; Squaw Creek Schist to left.

FIDDLE CREEK SCHIST

The structurally lowest formation of the Riggins Group consists of white and green schist derived from silicic and intermediate volcanic flows and tuff.

DEFINITION AND DISTRIBUTION

The formation occupies a belt 2 miles wide in the northeast limb of the Riggins syncline and is exposed discontinuously along the Salmon River and along Fiddle Creek, its tributary at the north edge of the quadrangle. The formation is here named for that tributary, and its type section is designated as that along Fiddle Creek and the ridge to the north. The present thickness of the formation is about 9,000 feet.

Outcrops are generally poor southeast from Fiddle Creek along the strike of the formation. The area between the Chair Point-Riggins trail and Berg Mountain was not visited, and the boundaries of the formation indicated there on the geologic map are inferred.

White schist makes up about half the formation and is its characterizing rock type. Overlying the Fiddle Creek is the Lightning Creek Schist, which in its middle-grade manifestation is entirely greenschist; greenschists of the two formations are indistinguishable, and the contact is drawn above the uppermost white schist.

The lower limit of the Fiddle Creek Schist is a structural contact. To the north the schist is separated from the underlying Martin Bridge Limestone and other formations of the Permian and Triassic sequence by the Rapid River thrust. To the northeast the schist was intruded by quartz diorite gneiss.

The diverse rocks of the Fiddle Creek Schist are intercalated in units that are internally rather uniform and commonly range from 10 to several hundred feet in thickness. Schistosity is essentially parallel to this layering.

LITHOLOGY

WHITE SCHIST

Silicic tuff, metamorphosed to white schist with only a small proportion of mafic minerals, constitutes about half the Fiddle Creek Schist and is interlayered with other rock types throughout the formation. As white schist occurs nowhere else in the Riggins Group, it is definitive of the Fiddle Creek Schist.

Rare varieties of white schist consist wholly of light-colored minerals, but the common types consist of a finely schistose white groundmass studded with small porphyroblasts of carbonate, prochlorite, biotite, and, uncommonly, garnet and hornblende in varying combinations. Rocks spanning a broad range of metamorphic grade have a similar appearance. Groundmass material is dominantly quartz, albite, and muscovite, whose average grain size ranges from 0.02 to 0.2 mm. Albite is partly in tiny compact grains and partly in microporphyroblasts; sodic oligoclase occurs in the higher grade rocks. Some low-muscovite schist has a superficial resemblance to fine-grained quartzite, but muscovite is generally sufficiently abundant to yield a schistose sheen. Despite the marked schistosity, few of these rocks have conspicuous compositional laminae, alternately mica-rich and mica-poor, such as are attributed to metamorphic segregation in other schists of the quadrangle.

Subhedral rhombic porphyroblasts of iron-bearing, rust-weathering carbonate as long as 2 mm are present in perhaps half the lower-grade rocks but are generally lacking in the higher grade ones. Porphyroblasts (fig. 8) and lenticular aggregates of biotite are present widely; biotite is variably orange, olive green, or olive brown in pleochroic color. Prochlorite, either aluminian or ferroan, is a common minor component of both porphyroblasts and groundmasses (fig. 9).[1] Sparse granules of epidote or clinozoisite occur in most rocks. Uncommon white schist has big poikilitic porphyroblasts of pale-red garnet.

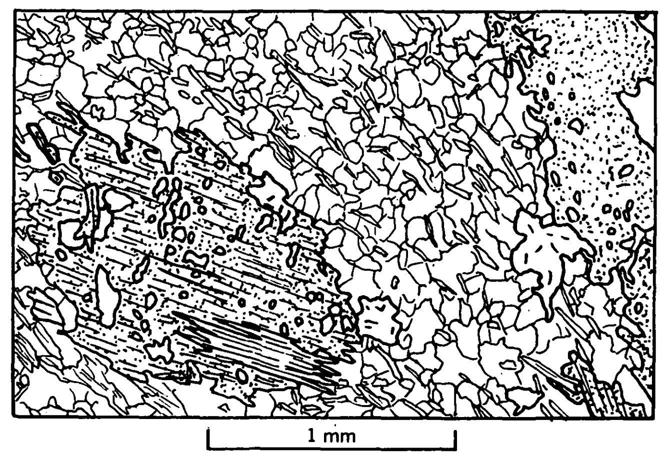

FIGURE 8.—White schist with biotite porphyroblasts. Groundmass consists of muscovite, quartz, and albite. Fiddle Creek Schist, crest of ridge between Chair and Fiddle Creeks, altitude 5,100 feet.

A large proportion of the white schist, particularly that of lower grade, clearly originated as crystal and lithic tuffs or as porphyritic flows (fig. 10). Relict clasts that are sheared and thoroughly reconstituted form stubby lenses of diverse compositions. Original clasts were commonly smaller than half an inch, but

[1] Some of the accompanying photomicrographs were taken with the nicols oblique or parallel. Such a technique often better preserves the impression a rock gives the petrographer than does the customary method of using either one nicol alone (plane-polarized light) or both nicols crossed when taking photographs. Separate crystals can be distinguished by their variable extinction in oblique-nicols photographs, as in crossed-nicols ones, and few grains are rendered black.

FIGURE 9.—White schist with porphyroblasts of prochlorite. Crosscutting porphyroblasts are aluminian prochlorite intergrown with subordinate biotitle. Groundmass is muscovite-biotite-chlorite-albite-quartz schist. The rock contains many garnets, although none is in this field. Fiddle Creek Schist, on ridge 1.5 miles southwest of Chair Point. Nicols oblique at 70°.

meta-agglomerates, with blocks initially 1 foot in diameter, are present locally.

Relict phenocrysts and crystal fragments of quartz are now ovoid or lenticular aggregates of intergrown quartz crystals. Some initial feldspar crystals are preserved as poikilitic albite pseudomorphs, complete with relict twins (fig. 10), but most have been converted to lenses of muscovite or granular albite, or of both. Some aggregates of mafic minerals pseudomorph mafic phenocrysts (fig. 10).

The white schist may be entirely volcanic, judging by the mineral proportions and by the lack of laminae suggestive of extensive waterworking. Some of the schist contains as much as 60 percent quartz, however, which suggests either concentration by sedimentary processes or silicification of volcanic rocks.

An analysis of a white metatuff collected from the Fiddle Creek Schist is given in table 5 (No. 16). This rock is of quartz-keratophyric (sodic-potassic) composition and is now a prochlorite-biotite-albite-muscovite-quartz schist.

The most highly metamorphosed parts of the formation include light-colored biotite-quartz-plagioclase gneiss and rock types intermediate between gneiss and obviously metavolcanic schist. Some of the gneiss contains garnet and is clearly metavolcanic, but perhaps other gneiss formed from intrustive granitic material.

GREENSCHIST

Metavolcanic schist colored green by prochlorite is only slightly less abundant in the Fiddle Creek Schist than is white schist. The greenschist ranges from near-white rock, transitional into white schist, to greenish gray; most is rather light in color. Structure varies from strongly schistose to massive, a rough schistosity being dominant; the massive rock is properly greenstone.

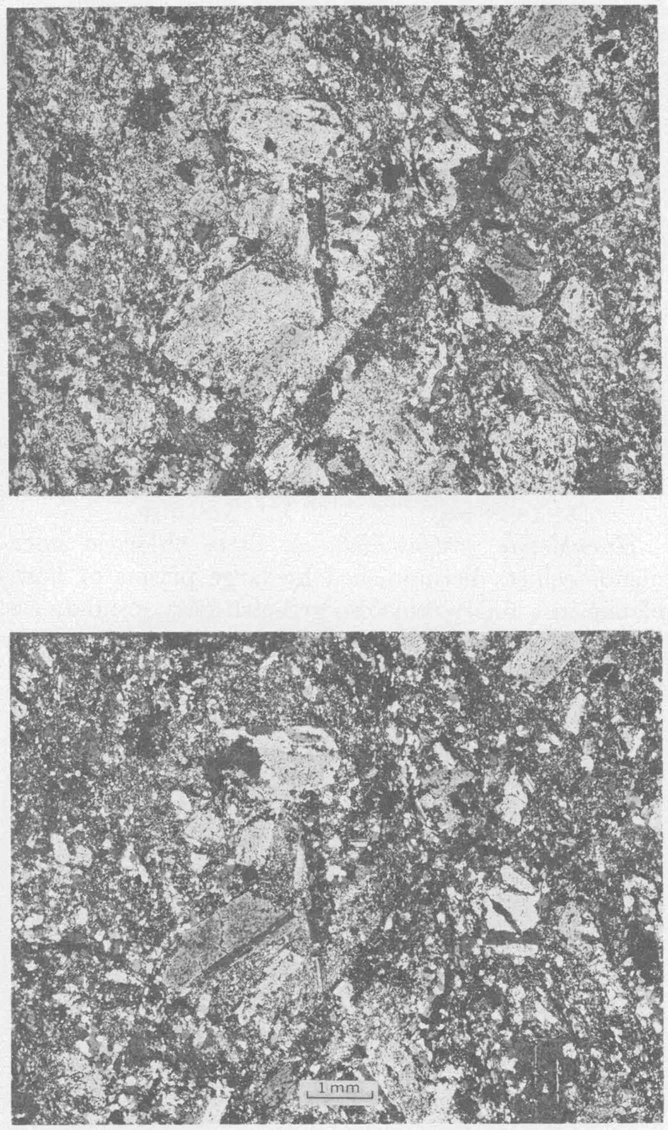

FIGURE 10.—Metadacite. Plagioclase phenocrysts have been pseudomorphed, with preservation of original twinning, by inclusion-clouded albite. Long phenocrysts of hornblende have been pseudomorphed by prochlorite, carbonate, and quartz. Upper photomicrograph, nicols parallel (not plane-polarized light); lower, same field, nicols crossed. Fiddle Creek Schist, Fiddle Creek, 0.8 mile above mouth.

Quartz, prochlorite, and albite or oligoclase constitute most of the groundmass. Grains are small; the average diameter of quartz and feldspar is 0.02–0.2 mm, and the flake length of chlorite is twice that. Muscovite is concentrated in folia and aggregates.

Most of the greenschist contains porphyroblasts of biotite or carbonate (calcite, magnesite, or iron-bearing species), and hornblende, muscovite, prochlorite, and garnet are less common; the porphyroblasts typically cut across the cleavage. Small amounts of fine-grained clinozoisite or epidote occur in many of the schists.

Mineral contents indicate that the greenschist is the metamorphosed product of intermediate to silicic flows and tuff. The descriptions given above of relict features in the white schist apply equally to the greenschist, although in the greenschist the clearly tuffaceous rocks are much subordinate to those formed from more massive and uniform materials. Relict clasts and pseudomorphs of both mafic and felsic phenocrysts are common. Plagioclase is represented by big albite pseudomorphs in which the primary twinning is preserved. The undoubtedly metatuff looks knotty.

MINOR ROCK TYPES

Hornblende greenschist.—A little chloritic hornblende schist, distinguished by large prisms of hornblende in a finely schistose greenish-gray groundmass, is interlayered with other types throughout all but the low-grade part of the Fiddle Creek Schist. The hornblende is commonly dark greenish gray, rather than black, in hand specimen. It varies in habit from needles, typically near 0.5 by 5 mm, to relatively stubby prisms as large as 3 by 10 mm or, uncommonly, even larger. Hornblende prisms are well alined within the foliation planes at some localities, and randomly cut across the structure at others; the mineral formed concurrently with shearing and directed stress at some localities, but elsewhere it formed after shearing and crystallization of most of the groundmass fabric.

Most of the hornblende schist also has small porphyroblasts of biotite and biotite bulks larger than hornblende in some. Groundmasses are finely schistose to nearly massive mixtures of prochlorite (to which the rock color is primarily due), epidote, albite, quartz, carbonate, and rarely a little garnet in varying proportions. The average grain size of the groundmass is generally between 0.05 and 0.2 mm.

Green amphibolite.—With increasing metamorphic grade the greenschist in part gives way eastward to green amphibolite. Amphibolite is intercalated with white schist, hornblende schist, and greenschist on the ridge northwest of Chair Point and is widespread on the ridge west of Chair Point above an altitude of 5,500 feet. All types lithologically intermediate between amphibolite and hornblende schist are abundant—and commonly interlayered—so that the distinction is an arbitrary one for convenience of description. The amphibolite is granular schist composed chiefly of 1- to 3-mm prisms of hornblende, dark greenish gray in hand specimen, and white plagioclase. Biotite and quartz are present in varying proportions.

The green amphibolite is distinguished from the hornblende-rich schist by its more granular character, and from the amphibolite of the plutonic terrane to the east by the color (green instead of black) and habit (smaller and commonly more elongate) of the hornblende.

Hornblende-rich schist.—Another rock type of high metamorphic grade that is present in the eastern part of the Fiddle Creek is hornblende-rich schist. Its main mafic mineral is hornblende (greenhued in hand specimen). All gradations in texture and composition between hornblende schist and green amphibolite on the one hand and chloritic hornblende greenschist on the other are common. The hornblende occurs variably as compact prisms and as large poikilitic ones. The rock is less schistose and more granular than the chloritic hornblende schist, and lineation is commonly more marked than foliation. Much of the hornblende-rich schist is glomeroporphyroblastic; the mafic and felsic minerals are each aggregated into clumps rather than segregated in contrasting folia.

ORIGIN

The Fiddle Creek Schist formed from massive tuffs and flows of intermediate and silicic compositions. If sediments are represented, they are of minor proportion and could only have been of volcanic derivation. The abundance of muscovite (which presumably contains considerable sodium as well as potassium) and albite or sodic oligoclase and the generally minor quantity of calcium-bearing minerals indicate a high content of alkalis and a low content of calcium and suggest that most of the rocks were spilitized prior to metamorphism.

The dominant volcanic rock types were the quartz-keratophyric equivalents of dacite, rhyodacite, quartz latite, and rhyolite. Rocks of normal volcanic compositions of those types and andesite and keratophyre were less common. No metabasalt or metaspilite was recognized.

Problems of spilitization are discussed on pages 68–74.

LIGHTNING CREEK SCHIST

DEFINITION AND DISTRIBUTION

Greenschist forms a unit 8,000 feet thick on the northeast limb of the Riggins syncline. It is excellently exposed along the Salmon River north of Riggins, from 0.1 to 1.7 miles north of Goff Bridge (fig. 11), and this exposure is designated as the type section. The formation is named for Lightning Creek, whose lower course is on the formation. Beneath the formation in its type section is the Fiddle Creek Schist, and above are ultramafic rocks and, next above, the Squaw Creek Schist.

FIGURE 11.—Greenschist in Salmon River Canyon. Compositional layers and cleavage of Lightning Creek Schist dip obliquely toward the observer. View northwest across the river, south of Lightning Creek.

The formation is mappable southeastward to the area about 4 miles east of Riggins, where its metamorphic grade is considerably higher than that of the type section. South of the Salmon River, however, contacts drawn are questionable, as the high-grade rocks of the Lightning Creek and Squaw Creek Schists are of similar appearance.

The greenschist phase of the Lightning Creek Schist is exposed also in the southwest limb of the Riggins syncline, along the Rapid River and the nearby part of the Little Salmon River. There, it is in part in direct contact with overlying the Squaw Creek Schist and in part separated from it by ultramafic rocks.

LITHOLOGY

The formation is composed of metavolcanic schists whose metamorphic grade increases eastward from low to high. Compositions are mafic to silicic—basalt and spilite, andesite and keratophyre, dacite, rhyodacite, and quartz keratophyre—but the appearance of the schists varies less than such a range might suggest. The low- and middle-grade rocks are fine-grained grayish-green and greenish-gray schists that are monotonously uniform in appearance. Compositional layers are commonly many feet, even hundreds of feet, thick, but they are demonstrable only where outcrops are good. Agglomerates and coarse-grained tuffs were present in the original sequence, but most of the rocks were flows and fine-grained tuffs. The greenschist is composed of prochlorite, actinolite, epidote or clinozoisite, quartz, and albite in widely varying proportions and combinations, and also, more locally, of biotite, hornblende, oligoclase, garnet, and carbonate.

The high-grade representatives of the Lightning Creek Schist are gray quartz-plagioclase schist and gneiss whose dominant mafic minerals are biotite and hornblende. Unlike the greenschist, many of these high-grade rocks are conspicuously laminated.

CHLORITE GREENSCHIST

By far the most widespread rocks of the formation are fine-grained platy greenish-gray and grayish-green schists. Although they vary widely in composition and metamorphic grade, their general aspect is one of uniformity. Lamination is commonly limited to nearly microscopic folia alternately rich in prochlorite and albite. Cleavage surfaces have a high sheen that is more conspicuous than the specular reflections from individual crystals.

Prochlorite and sodic plagioclase are present in all the greenschist. The prochlorite is green in hand specimen and makes up 10 to 40 percent of the rocks. Flakes range in common average size from 0.05 to 0.2 mm and are well oriented along the cleavage direction; porphyroblasts, lacking in most rocks, include both concordant and crosscutting grains. Albite, as granules a little smaller than the flakes of chlorite and as small porphyroblasts, is the common feldspar north of Riggins; higher grade rocks contain oligoclase. With the albite is quartz that is in part of similar habit and appearance and in part segregated into tiny lenses of coarser grained material. Quartz increases in both relative and absolute abundance as the total felsic mineral content increases, and quartz and feldspar together total 45 to 85 percent of the rocks.

Small granules of epidote, or less commonly of clinozoisite, make up as much as 15 percent of the greenschist (fig. 12). Amphibole—actinolite(?) in the lower grade rocks, hornblende in the higher—is also present in many specimens and is the dominant mafic mineral in some; it occurs variously as tiny granules, prisms, and needles, and as porphyroblastic prisms. Biotite occurs as conspicuous porphyroblasts of minor abundance in the most silicic rocks. Muscovite is an uncommon groundmass mineral; pyrite and carbonate are common as porphyroblasts; and garnet is uncommon (fig. 13).

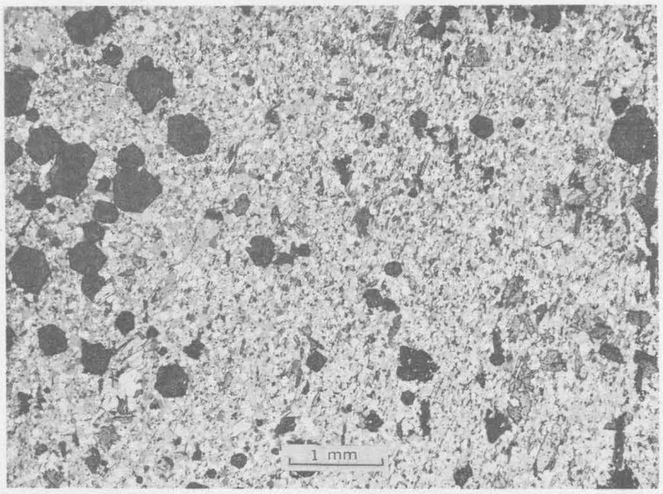

FIGURE 13.—Small dodecahedra of garnet in light-colored greenschist. Lightning Creek Schist, east side of Salmon River midway between Lightning and Chair Creeks. Nicols oblique at 55°.

Original fabrics have been largely obliterated by metamorphism, but small lenses and knots of many compositions indicate that uncommon rocks are sheared tuff. Plagioclase phenocrysts, pseudomorphed by albite and variably recrystallized and sheared, are recognizable in a few varieties (figs. 14 and 15); and clumps of mafic minerals suggest original phenocrysts in a few others. Laminae of probably sedimentary origin are virtually lacking. Most of the original rocks were apparently massive, uniform, and fine grained.

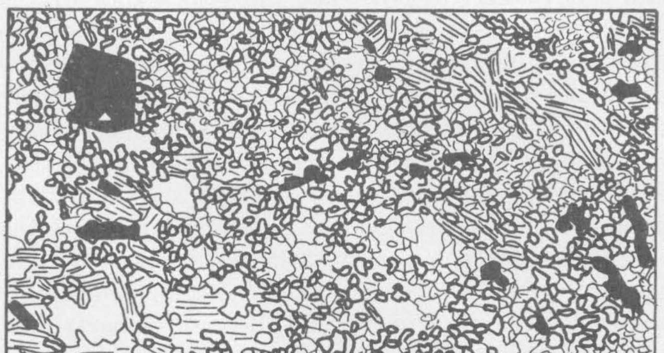

FIGURE 12.—Quartz-epidote-prochlorite-albite greenschist. At the lower left is a schistose lamina of quartz, ferroan prochlorite, and granular epidote. The central lamina consists of granular quartz, epidote, and albite. Analyzed specimen 5 (table 5). Lightning Creek Schist, Little Salmon River.

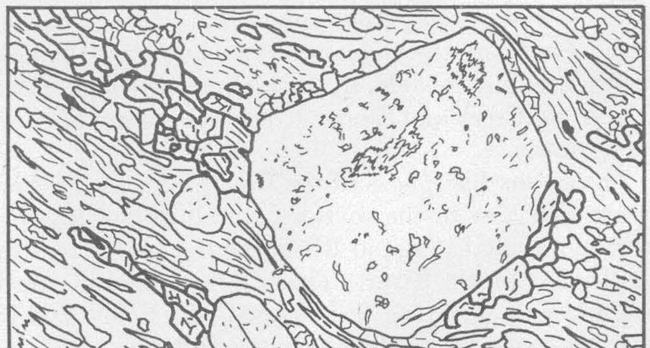

FIGURE 14.—Light-colored greenschist with relict phenocrysts. Blastophenocrysts of albitized plagioclase lie in a muscovite-chlorite-calcite-quartz-albite groundmass. Foliation flows around the phenocrysts, at the ends of which calcite is concentrated. Metaquartz keratophyre, Lightning Creek Schist, east side of Little Salmon River three-fourths of a mile south of Captain John Creek.

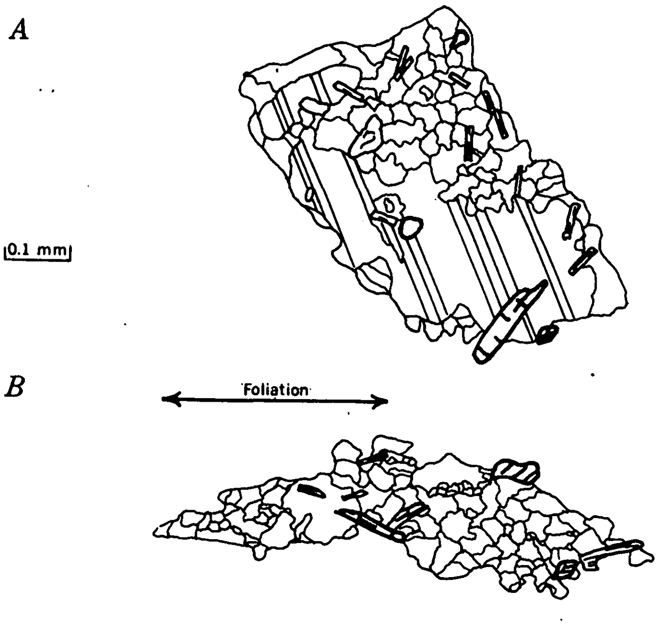

FIGURE 15.—Destruction of relict plagioclase phenocrysts in greenschist. *A*, Phenocryst (pseudomorphed by albite) with twinning through half of original crystal, other half replaced by diversely oriented blebs of albite. Bottom of phenocryst removed by shearing. *B*, Lenticular aggregate of albite blebs derived from such a phenocryst. Inclusions are actinolite, biotite, and prochlorite. Analyzed specimen 4 (table 5), Lightning Creek Schist, Little Salmon River.

Analyses of five specimens of normal greenschist (nonhornblendic, nonagglomeratic) are given in table 5. Of these, three (Nos. 3, 5, and 6) are from the Rapid River–Little Salmon River area, and two (Nos. 9 and 17) are from the Salmon River north of Goff Bridge. All but No. 17 are essentially basaltic in composition, although Nos. 3 and 9 are low in CaO for that rock type, and No. 5 is low in Al_2O_3. Contents of Na_2O can be ascribed to primary volcanic compositions; no additive spilitization is required, although the low CaO contents suggest removal of that component in Nos. 3 and 9. Analysis 17 represents a quartz-keratophyric composition.

The mineralogic compositions of the other specimens indicate similar volcanic compositions, varying from basalt and spilite to dacite and quartz keratophyre, all low in potassium. The mafic and intermediate rocks are much dominant over the silicic.

META-AGGLOMERATE

The basal 800 feet of the Lightning Creek Schist along the Salmon River north of Riggins is meta-agglomerate, and rocks of similar original character but higher grade of metamorphism occur on strike along the river east of Riggins, just west of Lake Creek.

The northern occurrence is excellently exposed in highway cuts from 1.5 to 1.7 miles north of Goff Bridge. There, the meta-agglomerate lies with structural concordance between the green and white schists of the Fiddle Creek Schist, beneath and north, and the green nonagglomeratic bulk of the Lightning Creek Schist. The agglomerate has the general aspect of greenschist. In average original size, cobbles varied from less than 1 inch to 1 foot or more. Deformation has made the clasts triaxial, with axial ratios near 1 : 2 : 6 (fig. 16) and the longest dimension parallel to the structural *b* axis; the coarsest clasts are in general least deformed (fig. 17).

Clasts are weakly schistose, fine grained, and greenish gray, or, uncommonly, white. They are composed largely of granular quartz and albite of irregularly varying grain size (0.02 to 0.3 mm) and have subordinate muscovite, epidote or clinozoisite, prochlorite, and magnetite, in their groundmasses; porphyroblasts are biotite, garnet, hornblende, and iron-bearing carbonate in varying proportions. The matrix between the deformed clasts is darker, coarser, and more schistose than the material of most of the clasts, and hornblende is relatively more abundant. Carbonate porphyroblasts are in part in zones that cut across the foliation plane (fig. 16). Many of the rocks are so highly sheared and

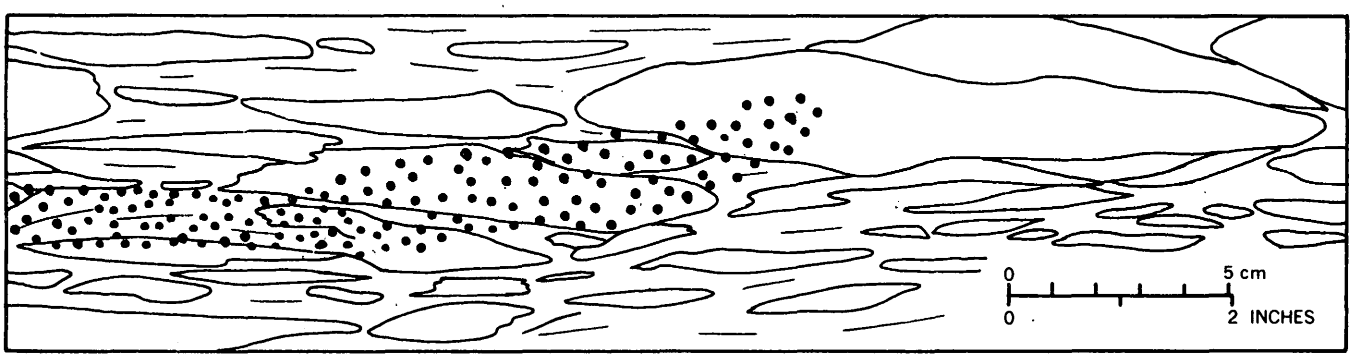

FIGURE 16.—Meta-agglomerate. Section parallel to maximum elongation of clasts. The dots indicate a crosscutting zone of porphyroblasts of iron-bearing carbonate. Analyzed specimen 18 (table 5), Lightning Creek Schist near base, north of Riggins.

reconstituted that outlines of few clasts can be traced, although their general agglomeratic character is clear.

The meta-agglomerate of the eastern occurrence is well exposed along the south side of the Salmon River for 1,100 feet west of the abutment of the bridge just west of Lake Creek; it is poorly exposed on the opposite side of the river. The agglomerate is intercalated with, and is subordinate to, hornblende-biotite schist. Clasts are now triaxial, typically with axial ratios near 1 : 2 : 3 (and thus are deformed less than the lower grade rocks north of Riggins), and are composed of granular white schist that contrasts with the dark matrix of coarse biotite schist. The clasts, which originally had an average size of few inches, are made up of sugary 0.05- to 0.5-mm interlocking and slightly elongate quartz and unzoned oligoclase grains sprinkled sparsely by porphyroblasts of garnet, ragged poikilitic hornblende, and minor biotite. The matrix between clasts is biotite-quartz-oligoclase schist with or without hornblende. Biotite is in 1- to 2-mm flakes and is in large part segregated into mica-rich folia along which the rock breaks readily. Pale-red garnet porphyroblasts are ubiquitous, and many are in the centers of light-colored eyes. Euhedral magnetite is conspicuous in thin section.

The meta-agglomerate near Lake Creek has been altered retrogressively forming much prochlorite (from biotite and garnet) and a little actinolite and gedrite (from hornblende). Alteration occurred both in irregular patches and along crosscutting shear folia. Open folds in the schists, indicated by a wide variance in attitude and by at least some repetition of units, perhaps formed at the time of retrogression.

The bulk composition of a specimen of agglomerate from each area is given in table 5 (Nos. 15 and 18). The specimen from the northern area has rhyodacitic composition; the specimen from the eastern area is

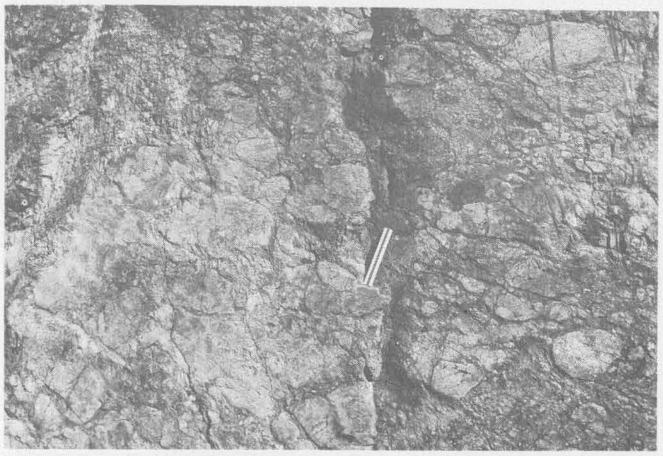

FIGURE 17.—Meta-agglomerate. Moderately sheared, light-colored clasts, in greenschist matrix. East side of Salmon River, near base of Lightning Creek Schist. The scale is 6 inches long.

dacitic or rhyodacitic and perhaps was partially altered to quartz keratophyre. The two analyses differ by only 4 percent SiO_2 (63 and 67 percent, respectively), the other oxides varying correspondingly; as clasts are more felsic than matrix in both, the true bulk compositions of the agglomerates may be virtually identical.

These rocks are termed "agglomerates" because their bulk compositions are volcanic and because they show little internal evidence of bedding. The considerable variation in composition of clasts in the northern area might, however, suggest a conglomeratic origin.

Meta-agglomerate, apparently identical with that at Lake Creek, is interlayered with metamorphic gneisses in the plutonic terrane east of the creek. Correlation is quite possible, but structural and lithologic continuity are lacking.

MINOR ROCK TYPES

Greenschist metatuff.—Intercalated with the normal greenschist are minor amounts of metatuff, mainly aggregates of small knots of varied greenschists, each knot representing a highly sheared clast (fig. 18). If similar tuffs were common in the original volcanic sequence, then shearing and reconstitution have been so thorough that most evidence for them has been largely destroyed.

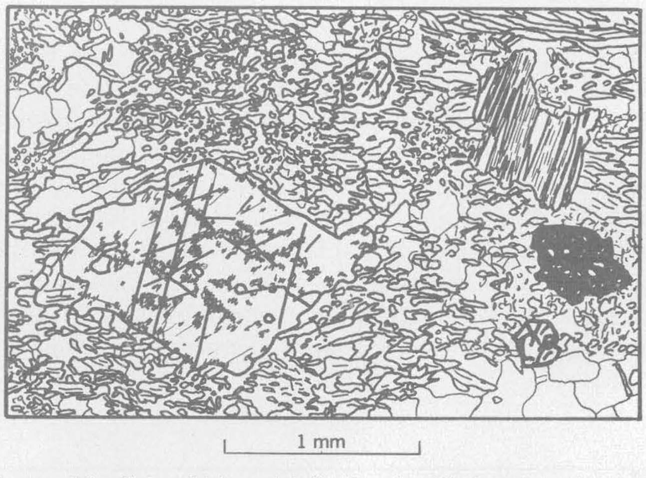

FIGURE 18.—Greenschist metatuff. Porphyroblasts of rust-stained calcite (left) and of intergrown biotite and prochlorite, magnetite-ilmenite, and garnet (right, top to bottom) lie in a fine schistose matrix of albite, ferroan prochlorite, and (in coarser lenses) quartz. Lightning Creek Schist, analyzed specimen 9 (table 5).

Hornblende greenschist.—Conspicuous porphyroblastic prisms of black hornblende characterize a subordinate type of greenschist common along the Salmon River north of Goff Bridge. The average length of the thin prisms ranges from 2 mm to 2 cm, and ratios of $a : b : c$ dimensions are generally nearly 1 : 2 : 10 but reach 1 : 2 : 50. Prisms are generally subparallel to foliation; alinement is crude, however, and as viewed in thin section the prisms cut across the foliation sharply. Most of the hornblende has light pleochroic colors, but some specimens show the dark pleochroic colors that are characteristic of the high-grade metamorphic and plutonic rocks of the quadrangle. Groundmasses are fine-grained greenschist with an average grain size near 0.1 mm, composed largely of albite, quartz, and prochlorite, the first two being generally greater in total abundance than the third. Biotite, garnet, epidote or clinozoisite, and opaque minerals are present in varying but generally small proportions. The diverse groundmass minerals are commonly segregated into microlaminae.

The composition of an analyzed specimen of hornblende greenschist (table 5, No. 2) is that of a nonspilitic basalt.

Actinolite greenschist.—Uncommon greenschist has actinolite as its dominant mafic mineral. It has little or no chlorite but is otherwise similar to the chloritic schists. One actinolite greenschist studied contains radial aggregates of porphyroblastic penninite (chlorite, pleochroic in green; birefringence in dark ultrablue very weak). One specimen of such rock was analyzed (table 5, No. 4).

Actinolite-rich schist.—Ultramafic greenschist crops out 0.5 mile north of Goff Bridge. The rock is very fine grained (0.005 to 0.05 mm) and has so marked a sheen and green color that it could easily be mistaken for a richly chloritic phyllite; actually, it is about 85 percent actinolite, with subordinate prochlorite, epidote, and albite. Cleavage is parallel to laminae richer in epidote and albite, and both are tightly folded on both microscopic (fig. 19) and outcrop scales. Cutting sharply across the folds are porphyroblastic prisms and anhedra of actinolite 1 to 2 mm long. A little leucoxene, sphene, and apatite are present, but

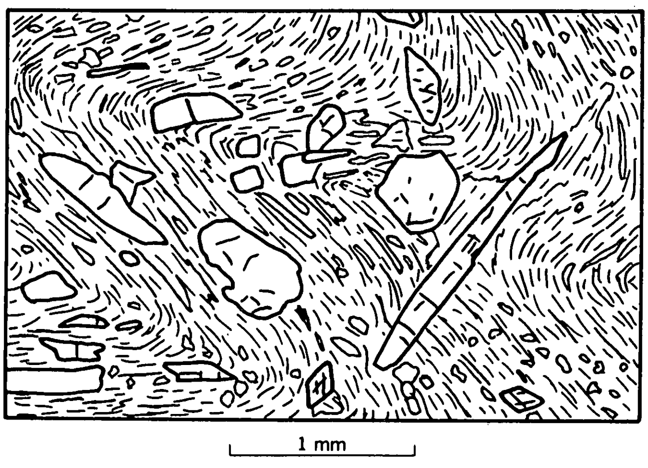

FIGURE 19.—Crosscutting porphyroblasts in folded greenschist. Contorted fine-grained actinolite-rich schist, in which axial planes of folds dip gently right, is cut across by porphyroblasts of actinolite. Lightning Creek Schist, east side of Salmon River, 0.5 mile north of Goff Bridge.

there are practically no opaque minerals. The mineralogy indicates a bulk composition of moderate silica content; high iron, magnesium, and calcium; and extremely low aluminum and sodium. Perhaps the original rock was an unusual igneous one that contained a high proportion of diopsidic pyroxene.

Chlorite-rich phyllite.—Several layers a few inches thick of nearly pure chlorite are intercalated with the actinolite-rich schist and normal greenschist at the locality noted above. So smooth and even is the cleavage in this very fine grained and richly chloritic phyllite that hand specimens superficially resemble fragments of single large crystals of chlorine. The phyllite is soft, and paper-thin flakes are easily split from it.

Muscovite schist.—A single layer a few feet thick of silver-white muscovite-rich schist crops out 0.2 mile south of Lightning Creek. Cleavage surfaces are brilliantly specular, as the muscovite is interrupted only by tiny well-alined prisms of hornblende.

Zoisite-bearing greenschist.—A single thin section from the Lightning Creek schist contains about 5 percent zoisite, in small euhedral domes. The rock, exposed 0.5 mile north of Goff Bridge, consists largely of subhedral prisms of hornblende (green in hand specimen) less than 3 mm long that have a planar orientation but little alinement. Between the prisms are reverse-zoned grains of oligoclase and albite, minor quartz, clinozoisite (cloudy anhedra), zoisite (clear euhedra), and a little biotite, carbonate, and magnetite. Replacement of one mineral by another is suggested only in the plagioclase, whose reverse zoning indicates that it was becoming more calcic when reconstitution ceased. The fabric of the minerals otherwise suggests them to be in an equilibrium assemblage. The bulk composition is perhaps that of an odd mafic basalt, judging by the mineralogy, and presumably the unusual assemblage was controlled by that composition.

Actinolite-clinozoisite greenschist.—Schist containing large amounts of actinolite and clinozoisite and subordinate albite and epidote crops out on both sides of the Little Salmon River 1 mile south of the mouth of the Rapid River. This is blotchy, relatively massive schist, presumably derived from a very calcic basalt or equivalent intrusive rock.

Metadiorite.—Only one thin section, from an exposure by Rapid River in the least metamorphosed part of the Lightning Creek Schist, is clearly of a schist derived from a plutonic rock. This specimen is from a grayish-green schist, indistinguishable megascopically from nearby metavocanic schists, in which the plutonic fabric is obscurely preserved by abundant 1- to 3-mm grains of plagioclase, pseudomorphed, with retention of albite and carlsbad twinning, by albite that is crowded with inclusions of clinozoisite, chlorite, and sericite. Marginal recrystallation has largely destroyed the primary subhedral shapes. About two-thirds of the rock is such feldspar. The rest is felted actinolite, epidote, aluminian prochlorite, and leucoxene, with scattered thick porphyroblasts of actinolite and veins of carbonate.

Many greenschists throughout the formation might similarly have been derived from plutonic rocks; but the metamorphic shearing and reconstitution have been so thorough that primary textural features have been obliterated, and metaplutonic rocks cannot generally be distinguished from metavolcanic ones.

Marble.—A few beds of foliated calcite marble, each a few feet thick at most, are intercalated with greenschist in the upper part of the formation north of Goff Bridge. Laminae in the marble range in thickness from 2 mm to 3 cm and are alternately white and medium light gray; both types are mosaics of anhedral calcite, the darker laminae being colored by a little black opaque material.

FIGURE 20.—Quartz-hornblende-oligoclase schist. Large poikilitic crystals of hornblende lie in a groundmass mosaic of quartz and reverse-zoned oligoclase. Lightning Creek Schist, south side of Salmon River 250 yards west of bridge at Lake Creek.

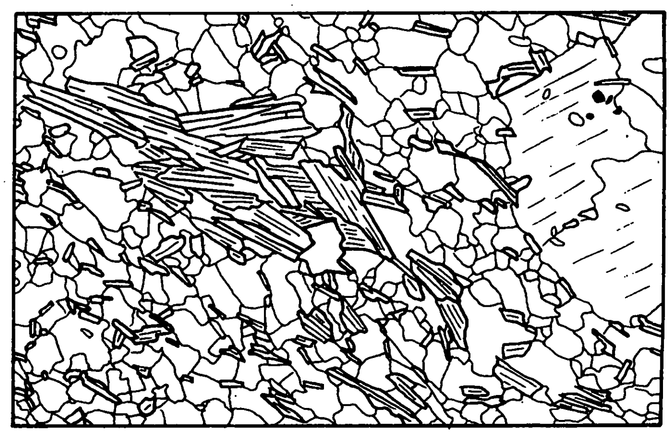

FIGURE 21.—Biotite-quartz-oligoclase schist. Porpyroblasts of oligoclase (right) and an aggregate of porphyroblasts of biotite lie in a fine-grained matrix of oligoclase, quartz, and biotite. Analyzed specimen 19 (table 5), Lightning Creek Schist, lower Lake Creek.

HIGH-GRADE METAMORPHIC ROCKS

Gray hornblende and biotite schists crop out along the Salmon River west of Lake Creek, interlayered with the meta-agglomerate described previously, and on strike from the greenschist north of Goff Bridge. These gray schists are accordingly considered to be part of the Lightning Creek Schist, although their lithologies have been duplicated in the high-grade terrane of the Squaw Creek Schist adjacent to the south. The distinction between the two formations, very obvious along the Salmon River below Riggins and along the Little Salmon River, can be made only arbitrarily in the high-grade rocks. It is possible that the two formations in their high-grade facies have been complexly intercalated by thrusting and isoclinal folding; the map interpretation east of Little Berg Creek may be one of gross oversimplification.

The high-grade rocks assigned to the Lightning Creek Schist are medium- to dark-gray schist and gneiss composed largely of biotite, hornblende, quartz, and reverse-zoned oligoclase in widely varying proportions. Garnet and epidote are common accessories. Hornblende is strongly pleochroic, its maximum color being a deep, almost bluish, green, and commonly very poikilitic (fig. 20). Biotite is well oriented (fig. 21), and its maximum absorption color is grayish olive. Prochlorite is a minor component and occurs as radial clusters that are clearly secondary after biotite and hornblende.

ORIGIN

Chemical compositions of the specimens analyzed all indicate a volcanic origin, and the mineralogic composition of the rest indicates that nearly all belong to a metavolcanic sequence of basalt, andesite, dacite, and rhyodacite, and their sodium-enriched kin, spilite, keratophyre, and quartz keratophyre. (The petrogenesis of the andesite-keratophyre association is the topic of a subsequent section.) Most of the greenschists are uniform for thicknesses of many feet; laminae suggestive of sedimentary origin are rare. Microlaminae, in which folia are alternately rich in micaceous and granular minerals, are widespread but are clearly the result of metamorphic processes. Tuff and agglomerate, identifiable only locally, are greatly sheared. In the least metamorphosed rocks, relict porphyritic fabrics are visible though obscure. No pillow structures were seen. Most of the rocks are so highly sheared and reconstituted that their original fabrics and structures are destroyed.

The Lightning Creek Schist apparently originated as a sequence of massive volcanic rocks and, presumably, from sediments derived directly from them and but little sorted into laminae of varying compositions. Coarse pyroclastic materials were uncommon except low in the formation. Plutonic rocks were introduced, at least locally, prior to metamorphism.

BERG CREEK AMPHIBOLITE

Schistose amphibolite about 1,000 or 1,200 feet thick forms a conspicuous unit in the Riggins Group along the Salmon River 3.5 miles east of Riggins. The dark resistant rock is conspicuous between the schists in the north wall of the canyon (fig. 7) and is well exposed along the north side of the river; this exposure is here designated as the type section. The amphibolite is named for Berg Creek, tributary to the Salmon River just east of the type section. The amphibolite was seen only in the canyon, and its map projection beyond in both directions is only inferred. It lies between the Lightning Creek and Squaw Creek Schists.

The unit is made up of both schistose amphibolite and of still more schistose rock better designated as hornblende-rich schist. The schistose amphibolite is coarser and is thinly striped by alternate 0.5- to 5-mm light and dark laminae. The schist is laminated more finely and obscurely. Both rock types have a greenish cast due to the greenish-black color of the hornblende.

The amphibolite consists of laminae alternately rich in small (0.05 by 0.5 mm) needles or larger (1 by 3 mm) prisms of actinolitic hornblende, and in white plagioclase and subordinate quartz (fig. 22). Hornblende needles are subhedral and well alined, whereas the larger, stubbier prisms of other specimens are anhedral and poorly oriented. Plagioclase occurs as a mosaic of equant anhedra, each about 0.2 mm in diameter, of sodic andesine and calcic oligoclase that have smooth reverse zoning. Of the 3 thin sections studied, 1 contains a little zoisite (poikilitic anhedra cutting across hornblende folia); 1 contains a little clinozoisite of similar habit; and 2 contain less than 1 percent each of aluminian prochlorite, which both parallels and cuts across cleavage and appears primary metamorphic rather than retrograde. Small pale-red garnets are sparse.

FIGURE 22.—Isocline in amphibolite. Pencil gives scale; view is downward at a horizontal surface. Fold axis plunges 35° to the right, but the strong hornblende lineation is nearly horizontal. Analyzed specimen 8 (table 5) is from this locality, which is by trail on north side of Salmon River between Berg and Little Berg Creeks.

An analysis of amphibolite is given in table 5 (No. 8). Its composition is that of nonspilitized basalt or basaltic andesite. No clasts that would indicate an agglomeratic origin are present; presumably the rock was a flow, sill, or fine-grained uniform tuff. The rock may have been a gabbro related genetically to the ultramafic rocks. As the bulk composition of the amphibolite is that of a common igneous rock whereas the compositions of the separate laminae are unlike those of any common igneous or sedimentary rocks, it is presumed that the laminae have formed by metamorphic processes.

The hornblende-rich schist is in part intercalated with the amphibolite but in larger part overlies it to the southwest. The schist is greener than the amphibolite. It consists of long thin well-alined prisms of hornblende and abundant dodecahedra of garnet as much as several millimeters in diameter in a fine-grained groundmass of reverse-zoned oligoclase, quartz, and hornblende. The garnets grew across the foliation. Prochlorite, apparently in equilibrium with the other components, is a minor constituent.

SQUAW CREEK SCHIST

DEFINITION AND DISTRIBUTION

The structurally highest formation of the Riggins Group consists of gray schists, metamorphosed from sedimentary rocks that were in turn derived chiefly from volcanic sources. These schists are exposed in the broad central part of the Riggins syncline and underlie a large area about the town of Riggins. They crop out along the Salmon River from Goff Bridge to Little Berg Creek and along much of the Little Salmon River from its mouth to Pollock. The formation is here named the Squaw Creek Schist for the tributary to the Little Salmon River near its mouth. The type section is designated as that along the Little Salmon River from Riggins to Captain John Creek. The present thickness of the schist in the north limb of the Riggins syncline, where its structure is simplest, is about 6,000 feet. The thickness southeast of Riggins, where structure is complex, may be much greater.

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The Squaw Creek Schist lies above the Lightning Creek Schist both north and south of Riggins, although the two formations are generally separated by ultramafic rocks. East of Riggins, the Berg Creek Amphibolite lies between the two schist formations.

LITHOLOGY

The Squaw Creek Schist consists of gray phyllite and schist. Whereas the low- and middle-grade rocks of the Lightning Creek Schist are all green and those of the Fiddle Creek Schist are green or white, the rocks of the Squaw Creek Schist are cold grays and seldom show even a suggestion of a green cast. Compositional layering and lamination are conspicuous in most of the Squaw Creek Schist (fig. 23), which is in marked contrast to the much more uniform character of the other formations. Dust of carbon or other opaque material pervasively clouds a large part of the Squaw Creek Schist and is responsible for much of the gray color.

FIGURE 23.—Isoclinally folded laminated schist. The hammer handle lies down the dip of a late cleavage. Squaw Creek Schist, on hillside west of southern part of Riggins at altitude of 2,200 feet.

The rocks vary in grain size from aphanitic (but high-sheen) phyllite to medium-grained gneiss. Most are fine-grained schist, and many contain small porphyroblasts of biotite or hornblende. Color ranges from very light to very dark gray.

Plagioclase (andesine to albite) is the dominant mineral in many rocks, and quartz is dominant in most of the others. Mineral proportions and combinations vary widely, but carbonates, biotite, and muscovite are the common major components. Epidote, clinozoisite, garnet, hornblende, and prochlorite are widespread but generally of minor abundance. Zoisite, diopside, and scapolite are rare. Rare, nearly monomineralic rocks composed of calcite or quartz are also included.

Compositions of many rocks are similar to those of intermediate volcanic rocks, although with modifications due to sedimentary processes. Other rocks are calcareous, quartzitic, or calc-silicic. Present rock types vary widely and intergrade complexly. For convenience of description, the rocks are here assigned to various types, chiefly on the basis of hand specimen appearance, but many of these distinctions are arbitrary.

DARK-GRAY PHYLLITE

Shiny platy dark-gray phyllite and fine-grained schist are the most widespread rocks in the lower grade phase of the Squaw Creek Schist (fig. 24). These rocks, which alternate throughout the formation, are abundant west of a line 1 mile east of the Little Salmon River and the Salmon River north of Riggins, from Hay Creek to Goff Bridge. Minor occurrences of these rocks extend along the Salmon River east of this line almost to Little Berg Creek.

Cleavage surfaces have a submetallic sheen, and over large areas they are deformed by a crinkle lineation caused by a younger slip cleavage at a high angle to the major cleavage. Iron gray is the common color. As grain size increases, the phyllite grades into fine-grained schist that has brightly specular cleavage surfaces and in which the slip cleavage and resultant crinkles are poorly developed. There is an irregular eastward increase in grain size—slaty phyllite is present chiefly in the far western part of the area, notably along Race and Squaw Creeks, and schist becomes progressively more abundant eastward. The rocks vary widely in metamorphic grade but little in appearance.

The rocks are finely obscurely laminated, although compositional laminae are often masked by parallel

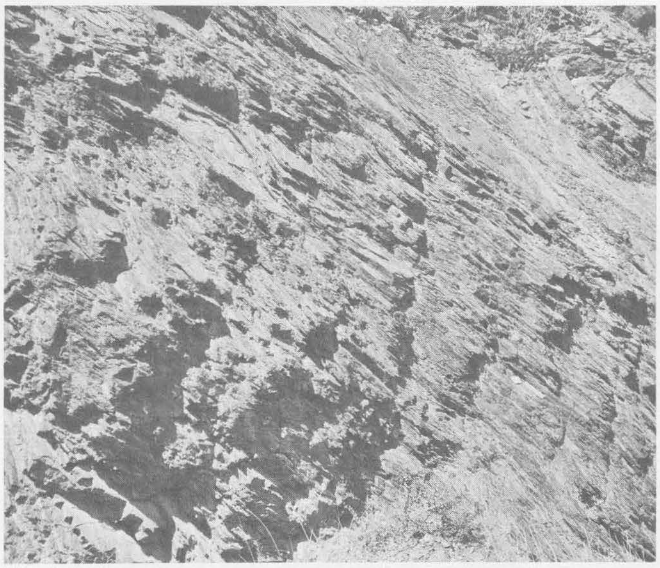

FIGURE 24.—Platy gray schist. Squaw Creek Schist, Little Salmon River near Squaw Creek.

cleavage. Less schistose laminae are richer in quartz, albite, or carbonate. Although the major cleavage is commonly subparallel to the compositional layering, it is oblique in many places (fig. 23).

Quartz, the dominant mineral of the dark-gray phyllite and schist, occurs as elongate anhedra whose average minor diameter is within the range 0.01 to 0.2 mm. Major diameters are 1.2 to 2 times as great. Some of the schist is nearly pure quartz with a few percent mica, but these rocks also are colored dark by included carbon (fig. 25). Much of the schist also contains augen of coarser quartz. Albite (or uncommonly oligoclase) is of habit similar to quartz where present, but generally it is sparse or lacking.

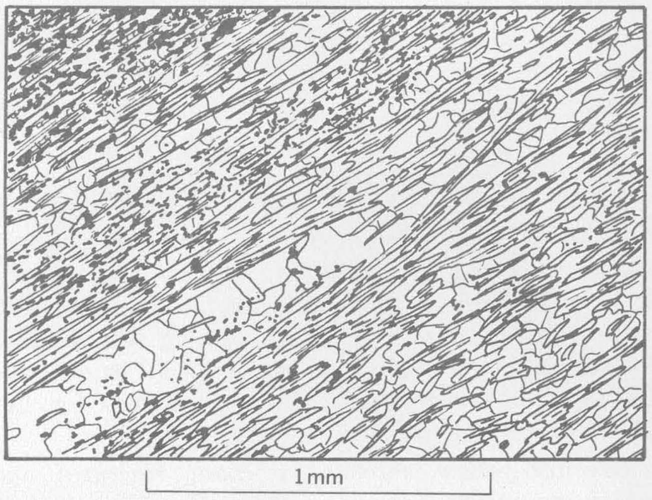

FIGURE 25.—Prochlorite-quartz phyllite. Several laminae are colored darkly by opaque carbon. Squaw Creek Schist, across Salmon River from Riggins.

Groundmass micaceous minerals are muscovite, biotite, and prochlorite, alone or in any combination. Biotite is pleochroic light orange brown, and in some specimens it is interlaminated with muscovite. Prochlorite is aluminian. Carbonate (either calcite or gray rusty-weathering noneffervescent carbonate) is abundant in many schists either as disseminated granules or as augen (fig. 26) or rhombs (fig. 27) of larger crystals. Granules or euhedral domes of clinozoisite are important locally.

Thick flakes of prochlorite cut the main cleavage of much of the phyllite and the slip cleavage also in some (fig. 28). These prochlorite porphyroblasts themselves were sheared and recrystallized in places, and some were rotated, as shown by skew attitudes of the enclosed carbon trains.

Garnet was found in the dark-gray phyllite and schist only along the river east of Riggins and along the Little Salmon River within a mile north of the Rapid River, where garnet was rolled along the cleavage and partly altered to prochlorite (fig. 29).

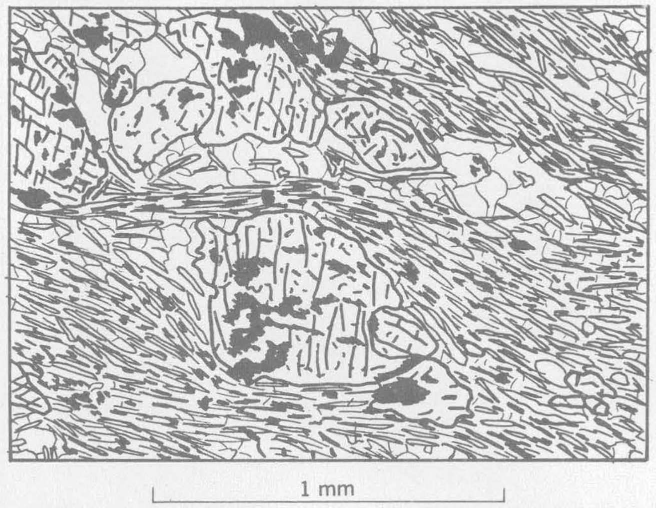

FIGURE 26.—Muscovite-ankerite(?)-prochlorite-quartz phyllite. Folia of muscovite, prochlorite, and quartz separate augen of quartz and iron-stained carbonate. Squaw Creek Schist, analyzed specimen 1 (table 5), Little Salmon River.

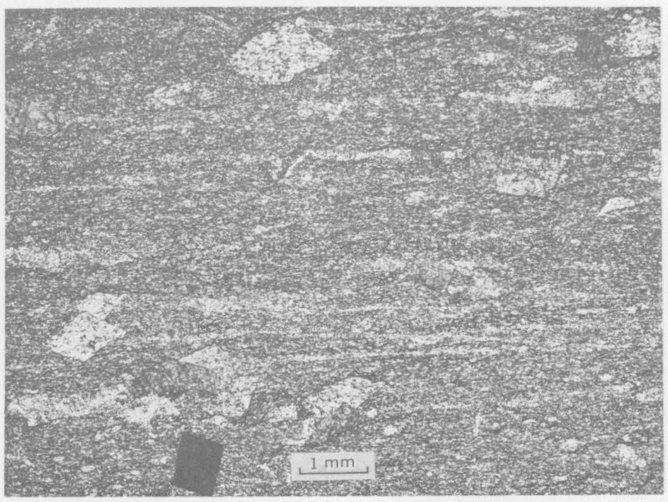

FIGURE 27.—Rhombs of carbonate in dark fine-grained slaty phyllite. A cube of pyrite lies near the lower left corner. Squaw Creek Schist, Race Creek, near mouth.

No amphibole was found in any of the dark-gray phyllite and schist.

Most of the dark-gray phyllite and schist contains no mineral definitive of higher metamorphic grade than biotite, clinozoisite, or aluminian prochlorite. Over much of their extent, however, these rocks are intercalated with other schists containing minerals such as garnet, hornblende, and oligoclase, indicative of higher minimum grade.

Subordinate to the platy phyllite is more massive rock of similar dark tone and mineralogy, except that micas are either so sparse or so poorly oriented that schistosity is obscure.

The dark color of the phyllite and schist is due to pervasive opaque black dust. In most instances the dust

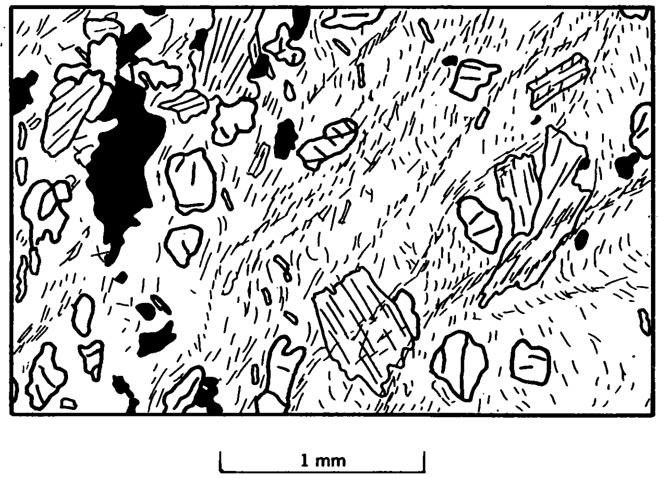

FIGURE 28.—Polymetamorphic schist. Porphyroblasts of ankerite(?) and pyrite are related to the old cleavage, here vertical. The fine fabric of the rock is greatly deformed by a slip cleavage (dipping to the left). Random porphyroblasts of prochlorite are younger than the slip cleavage. Analyzed specimen 13 (table 5), Squaw Creek Schist, Salmon River at Race Creek.

is apparently carbon, concentrated in folia whose carbon content continues without interruption through prochlorite porphyroblasts; the dust shows no absorption or reconstitution, in contrast to the behavior of metallic oxides. Magnetite dust is the coloring agent of many of the rocks, however, and tends to be concentrated in carbonate crystals rather than in throughgoing folia. Cubes and blebs of pyrite are conspicuous in many specimens; scattered tiny granules or euhedra of sphene are present in all the rocks; tourmaline is generally present in traces.

The quality of the oblique slip cleavage varies from incipient to very marked. Typically the two cleavage directions intersect in an anastomosing network of micaceous folia that divide the rock into tiny lenses richer in quartz or carbonate. The crinkles that define the megascopic lineation—the expression of the lenses on the cleavage—have a wave length of 1 mm to 1 cm and a wave height of half as much. With greater deformation and shear along the slip cleavage, nearly isoclinal folds in the old cleavage were produced (fig. 30). Extreme development of slip cleavage (in rocks whose older fabric is still recognizable) produced lenticular laminae whose internal structures are truncated against the slip cleavage, which is much more conspicuous than the older cleavage.

Degree of associated reconstitution also varies widely. Micaceous minerals are bent and broken in some rocks, recrystallized into new grains parallel to their previous positions in others, and reconstituted parallel to the new cleavage in some (fig. 30). Augen between the cleavages are bent and broken grains in some and completely recrystallized grains in others.

As all stages of secondary shear and recrystallization are widely represented, it is clear that some of the rocks whose fabric shows only a single episode of metamorphism must actually be two-cycle rocks in which shear and recrystallization along slip cleavage have been so thorough as to obliterate all evidence of the first metamorphism. There is evidence for this also in the megascopic structures.

Greenish-yellow jarosite(?) coats fractures and cleavage surfaces in some phyllite. Coatings of white sulfate on protected parts of many outcrops presumably were derived from the weathering of pyrite.

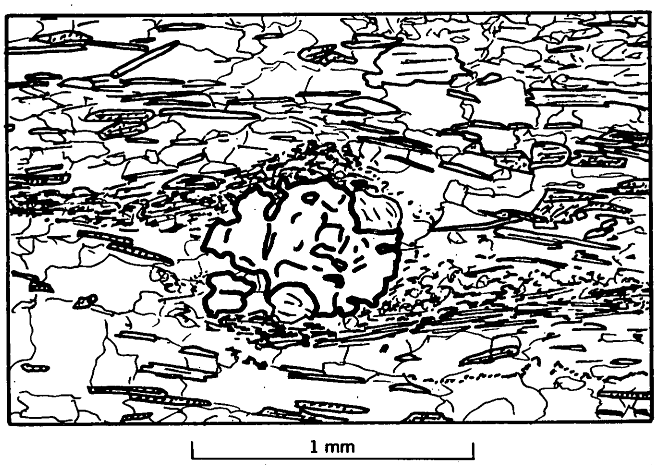

FIGURE 29.—Rolled garnet in prochlorite-biotite-muscovite-quartz schist. The dark lamina, rich in opaque dust and muscovite, was overlapped and offset by drag along the garnet as the garnet rotated clockwise about 240°. Garnet was partly chloritized. Squaw Creek Schist, west side of Little Salmon River opposite Captain John Creek.

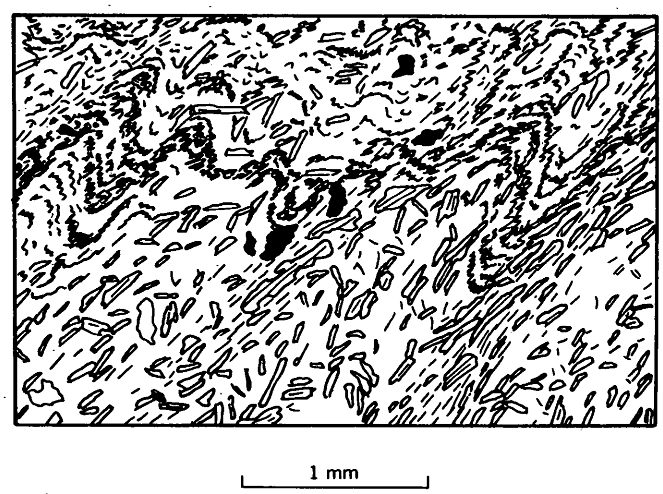

FIGURE 30.—Shear folding along slip cleavage. Carbonaceous folia developed along initially horizontal schistosity have been folded tightly by shearing along slip cleavage that dips to the left. Biotite is mostly recrystallized in the new cleavage direction, parallel to which there is also incipient metamorphic segregation. Dark-gray biotite-quartz schist, Squaw Creek Schist, east side of Little Salmon River 0.5 mile north of Sheep Creek.

Chemical analyses of three specimens of these dark fine-grained rocks are given in table 5 (Nos. 1, 13, and 22). Both Nos. 13 and 22 contain carbon, but only No. 22 contains enough so that it was detected in the rapid-method analysis. This carbon indicates that No. 13 is a metasediment, but it is classed with the metavolcanic rocks in the table because its bulk composition is mainly volcanic. There is no carbon in No. 1.

LIGHT-GRAY PHYLLITE

Laminated light-gray phyllite with an aphanitic groundmass speckled by porphyroblasts (commonly biotite) is abundant in the Squaw Creek Schist in the vicinity of Riggins (fig. 23). The phyllite forms widespread intercalations with other rock types along the Salmon River below Little Berg Creek, along the Little Salmon River below Sheep Creek, and along the lower part of Hailey Creek.

Laminae of varying light and medium shades of gray are a fraction of a millimeter to several centimeters thick. In some of the phyllite, the laminae are of two alternating types. In others, 3 or 4 types are intercalated irregularly. Porphyroblasts are commonly oriented parallel to the laminae, but crosscutting cleavage and porphyroblasts are common.

The phyllite is composed largely of granular minerals, and cleavage is poorly developed. Despite their fine lamination, most of the rocks form massive outcrops rather than platy ones. These rocks crumble upon weathering.

Light laminae are rich in fine-grained (0.03 to 0.1 mm) quartz and albite or reverse-zoned oligoclase. Only exceptionally are these layers of nearly pure quartz; most contain abundant plagioclase. Fine-grained biotite or muscovite (rarely prochlorite) makes up a trace to 10 percent of the light laminae and generally is strewn throughout rather than concentrated in folia. The more muscovitic laminae cleave to expose minutely specular, silvery surfaces. Some light laminae contain calcite or iron-bearing carbonate. The light laminae glitter with myriad reflections from the cleavage faces of the tiny feldspar grains, and this high luster suggests the laminae to be far more micaceous than they actually are.

Increasingly dark laminae contain fine-grained biotite, clinozoisite, actinolitic hornblende, and opaque dust in varying combinations and increasing total. Dark laminae are finer grained than the light, but they are not commonly more schistose than the light laminae, as they too are low in micaceous minerals. The opaque dust is largely magnetite rather than carbon. Aluminian prochlorite is common in very minor quantity.

Porphyroblasts of biotite are present in most of the light-gray phyllite. These are 0.5 to 1 mm in mean diameter and in many places show a dimensional lineation in addition to their planar orientation. Stubby prisms of actinolitic hornblende instead of, or in addition to, biotite are much less common. Small red garnets, variously euhedral dodecahedra and very ragged anhedra, are a minor component of much of the phyllite.

Pyrite is widespread as elongate blebs along the cleavage and lamination. It makes up as much as 1 percent of exceptional specimens.

The analysis of one specimen of light-gray phyllite is given in table 5 (No. 14). Although this (like most others of the gray phyllite) is laminated by layers of probable sedimentary origin, it is tabulated with the metavolcanic rocks because its bulk composition is volcanic.

The mineralogy of the other light-gray phyllites indicates that they too are largely of volcanic compositions. Sodium is clearly dominant over potassium in most. The volcanic source rocks were intermediate to silicic.

BIOTITE SCHIST

As the size and amount of crystals of biotite increase, the dark- and light-gray phyllite and fine-grained schist grade into rocks better designated as porphyroblastic biotite schist. Such schist is widespread in the Squaw Creek. Groundmasses increase in grain size irregularly eastward from phyllitic to schistose and vary from light to medium gray. Many of these rocks also contain porphyroblasts of hornblende; the distinction between hornblende and biotite schists is arbitrary, depending upon which mineral is more conspicuous in hand specimen. The rocks are crudely laminated to obscurely layered.

Biotite porphyroblasts are typically 0.5 to 1 mm in diameter in the west, twice as large in the east, and in many rocks they are elongate and alined. Biotite constitutes as much as 20 percent of the rocks and is variably distributed—uniformly strewn but oriented parallel to the foliation, or segregated into folia or planar clusters. Oligoclase and subordinate quartz dominate the groundmass mosaics and have average grain sizes in the range 0.02 to 0.3 mm. The less silicic rocks have quartz only in lenses, in which calcite may also occur. Other sparse groundmass minerals are prochlorite, hornblende, clinozoisite, garnet, muscovite, and calcite.

Analyses of three specimens of biotite schist are given in table 5 (Nos. 10, 11, and 19). No. 19 seems to be compositionally a silicic quartz keratophyre. Nos. 10 and 11 are very similar in both chemistry and mineralogy and might have originated as potassic spilite or as sediments chiefly of partly weathered volcanic material.

HORNBLENDE SCHIST

Fine-grained gray schist bearing porphyroblasts of hornblende is abundant throughout most of the Squaw Creek Schist. The large hornblende crystals vary widely in abundance, commonly making up 20 to 30 percent of the rocks but having an extreme range of 10 to 50 percent. Groundmasses vary in color from white to dark gray but generally range from light gray to very dark gray; few of these rocks have a perceptible greenish cast. Groundmasses are commonly granular, and the big hornblende crystals, alined or not, are commonly strewn throughout rather than concentrated in folia; so most of these schists are massive in outcrop but poorly fissile. Lamination is commonly crude and gradational.

Hornblende porphyroblasts are thin prisms 3 to 5 mm long, uncommonly 1 cm. Hornblende is black or greenish black in hand specimen and mostly is pleochroic in light yellowish green. Prisms are commonly compact and well formed, but many rocks contain ragged poikilitic hornblende instead.

Plagioclase is the dominant groundmass mineral and increases in grain size eastward in a very irregular way from near 0.1 mm in the west to about 0.3 mm in the east. In the east, the plagioclase is all oligoclase; in the west, rocks bearing oligoclase are intercalated with others bearing albite. Crystals are commonly elongate parallel to the hornblende lineation. Quartz of similar habit is present invariably and in some schist is more abundant than the plagioclase.

Accessory minerals occur in varying combinations and proportions. Garnet (fig. 31) and biotite are each in about two-thirds of the hornblende schist. The rocks

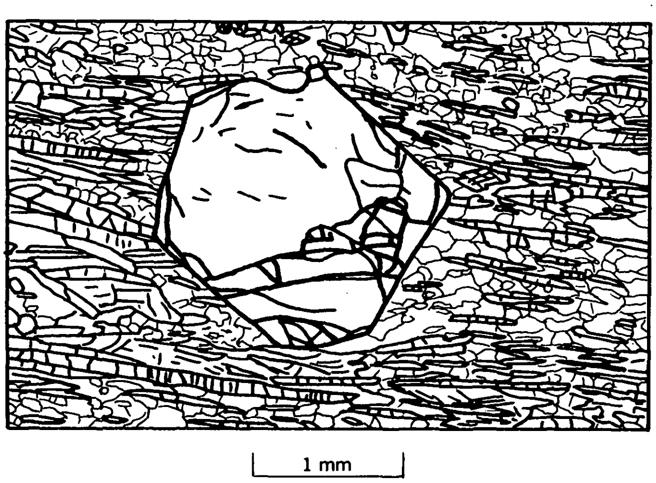

FIGURE 31.—Garnet-quartz-hornblende-albite schist. The big euhedral garnet cuts across the strong alinement of the hornblende. Squaw Creek Schist, north side of Salmon River, one-fourth mile west of Little Berg Creek.

grade into biotite schist as the biotite: hornblende ratio increases. About half of the schist has clinozoisite and some has prochlorite or carbonate.

One hornblendic schist (garnet-calcite-biotite-quartz-hornblende-oligoclase schist) was analyzed (table 5, No. 12). Its composition is mainly dacitic. The mineralogy of the other schists studied indicates general intermediate to silicic, but low-potassium, volcanic compositions.

MINOR ROCK TYPES

Amphibolite.—Granular fine- to medium-grained amphibolite consisting largely of hornblende and oligoclase or andesine forms thin intercalations in gray schists in many places south of the Salmon River and east of the Little Salmon River. Layers of amphibolite are a fraction of an inch to a few feet thick and commonly grade smoothly into layers of very different mineralogy. Garnet is conspicuous in many of the amphibolites; quartz, biotite, and clinozoisite are common. The arbitrary distinction made between hornblende schist and amphibolite is that the amphibolite is equigranular whereas the hornblende schist has porphyroblasts of hornblende in a fine-grained groundmass. There are, of course, intermediate types.

Quartz-rich schist.—Impure quartzite and quartz-rich schist and phyllite are of relatively minor abundance but occur throughout the western part of the Squaw Creek Schist. These are light- to medium-gray rocks consisting of 75 to 95 percent quartz and variable combinations of plagioclase, muscovite, biotite, prochlorite, actinolite, carbonates, garnet, carbon, magnetite, and pyrite. Quartz is in elongate anhedra, mostly near 0.1 mm long. Other minerals are concentrated in lenses or folia or are uniformly dispersed.

Quartz-oligoclase schist.—Light-gray granular schist composed largely of quartz and oligoclase is another minor but widely distributed type in the western part of the formation. The rock has 2 to 20 percent calcite and 5 to 15 percent biotite plus muscovite; the remainder is a mosaic of elongate 0.1-mm anhedra of oligoclase and quartz in subequal amounts.

Mafic biotite schist.—Rocks rich in biotite and other mafic minerals crop out on the north side of the Salmon River opposite Shorts Creek and at an altitude of 4,000 feet in the canyon next north of Captain John Creek. The megascopic fabric at both places is dominated by randomly intergrown nonschistose biotite that occurs in poorly shaped flakes of all sizes to 2 or 3 mm. Both rocks have a green cast due to abundant prochlorite, which at the Salmon River locality forms random sheaves and euhedral porphyroblasts, some of which cut across biotite. The Salmon River rock is composed of about 27 percent andesine, 20 percent each prochlorite and muscovite, 15 percent biotite, 10 percent clinozoisite, 5 percent brown garnet, and minor amounts of actinolitic hornblende, rutile, magnetite, and tourmaline (fig. 32).

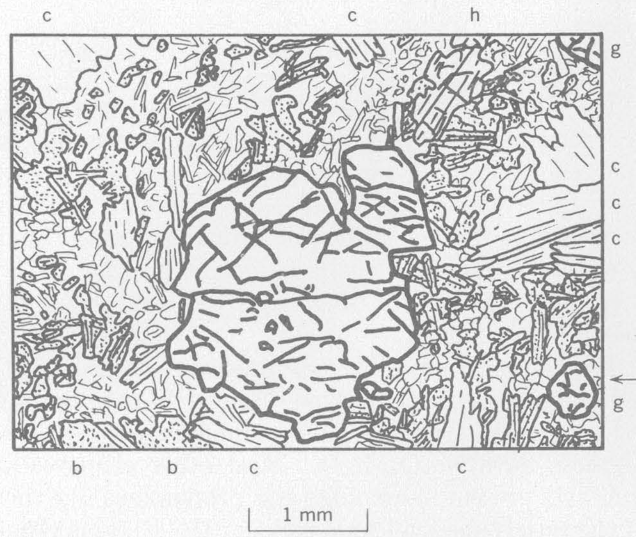

FIGURE 32.—Decussate rock containing garnet, hornblende, plagioclase, muscovite, and chlorite in equilibrium(?) assemblage. One large porphyroblast and several smaller ones of garnet (g) and porphyroblasts of aluminian prochlorite (c) lie in a matrix of anhedral biotite (b, stippled), hornblende (h), tiny muscovite flakes, and plagioclase (An_{30}). Squaw Creek Schist, north side of Salmon River opposite Shorts Creek.

Fine-grained gneiss.—Light-gray biotite gneiss with a general grain size near 0.1 mm is common near Riggins and southeastward from it. Rock of similar appearance is intercalated with phyllitic schist near Riggins and with gneiss at Lake Creek. The rocks vary from laminated to nearly uniform, depending upon whether the 0.5-mm flakes of biotite are segregated into folia or are dispersed. Quartz and oligoclase, either being dominant, make up 80 to 90 percent of the rocks. Garnet is also generally present, as are variable amounts of prochlorite, hornblende, muscovite, calcite, and clinozoisite.

Coarser light-colored gneiss is also present along lower Lake Creek. One conspicuous type has brown garnets in the centers of white eyes, which are surrounded by light-colored fine-grained biotite gneiss. The mafic components seem to have migrated from the eyes into the garnets.

Calcite-rich schist.—Medium- to dark-gray finely granular faintly laminated calcite-rich schist is widespread along the Salmon River from Riggins to Race Creek (fig. 33). Despite their dark color, these rocks have 85 to 90 percent calcite in flattened anhedra with average lengths of 0.1 to 0.3 mm. The dark colors are due to abundant opaque dust that is partly concentrated in folia and partly strewn throughout the calcite.

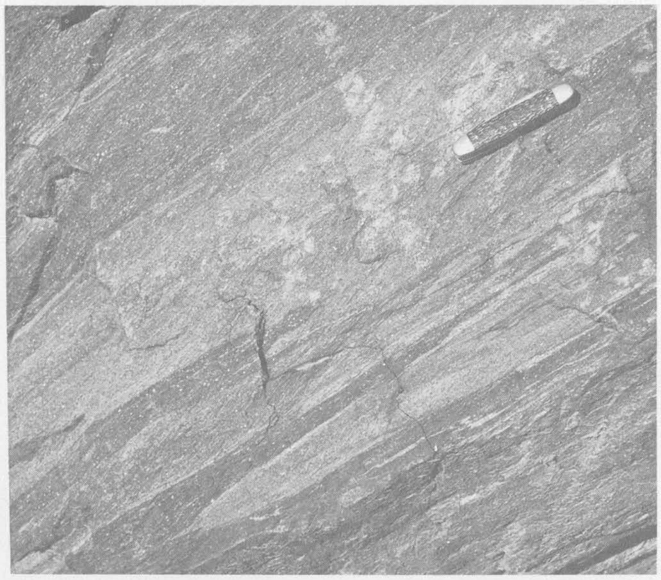

FIGURE 33.—Isoclinally interlensed light-gray calcitic schist and dark-gray schist. A slip cleavage dips steeply left, oblique to the major foliation. Squaw Creek Schist, west side of Salmon River south of Race Creek.

Small anhedra of quartz interspersed with the calcite make up 5 to 10 percent of the rocks. Small amounts of muscovite, prochlorite, and pyrite are present variably.

Fine-grained impure marble of similar composition but higher degree of metamorphism occurs north of the middle reaches of Sheep Creek. This marble is coarser grained (calcite 0.5 to 1 mm long) and contains a total of 20 percent quartz, clinozoisite, magnetite(?), epidote, and biotite.

Marble.—Coarse white calcite marble occurs as scarce thin intercalations in the gray schists on the ridges between Captain John and Sheep Creeks. Coarse gray marble is exposed on the east side of Whitebird Ridge 1.5 miles west of Pollock.

Diopsidic gneiss.—Fine-grained greenish-gray granulitic gneiss that contains diopside is a common but minor component of the Squaw Creek Schist in the middle altitudes of the Hailey Creek drainage. Andesine is the dominant mineral, clinozoisite or aluminian epidote is abundant, and quartz and diopside are less abundant. Calcite, hornblende, and scapolite are each present in some layers.

The nonhornblendic rocks look like dark, fine-grained quartzite. The rocks are closely to obscurely layered, and adjacent layers vary much in mineral proportions, although probably relatively little in chemical composition. Outcrops are massive, schistosity is lacking, and the rocks are exceedingly hard.

Two of these rocks, a diopside-clinozoisite-quartz-hornblende-andesine gneiss and a scapolite-quartzdiopsideclinozoisite-andesine-calcite gneiss, were analyzed (table 5, Nos. 21 and 23).

Rather similar granulitic calc-silicate gneiss that contains clinopyroxene but which is coarser grained and more amphibolitic occurs as intercalations in the gneiss complex exposed along the Little Salmon River south of Hazard Creek. The metamorphic component of the gneissic complex there was probably derived from the Squaw Creek Schist.

Zoisite schist.—Schist containing zoisite was found at only two localities.

Zoisite-bearing schist crops out on the south side of the Salmon River west of Berg Creek. Fine-grained dark-gray laminae alternate with coarse black laminae rich in biotite and large domes of dark zoisite. A specimen of garnet-clinozoisite-zoisite-andesine-biotite schist of odd sedimentary composition was analyzed (table 5, No. 20).

The other zoisitic schist crops out by Hailey Creek 0.4 mile above its mouth and is very different in appearance. It is composed of lenticular white, light-gray, dark-gray, and grayish-green laminae that are cut across at a high angle by a foliation shown chiefly by sparse porphyroblasts of biotite. Lenses are sheared across in the direction of this secondary foliation and have a maximum thickness of 5 mm. The diverse lenses studied in thin section have the following compositions: (a) quartz-andesine mosaic, with clusters of biotite and lesser amounts of zoisite, actinolitic hornblende, and magnetite; (b) intergrown large poikilitic domes of zoisite, enclosing chiefly quartz, and minor amounts of andesine, biotite, actinolitic hornblende, and magnetite; (c) granular quartz, plagioclase, calcite, and poikilitic biotite in subequal amounts, plus subordinate prochlorite and magnetite; (d) half poikilitic zoisite, half biotite, quartz, andesine, prochlorite, actinolitic hornblende, and clinozoisite; and (e) like (d), but with more biotite and clinozoisite. Biotite is orange and is variously in flakes, sheaves, and poikilitic sieves. The hornblende has a maximum pleochroism of grayish yellow green and is in ragged poikilitic crystals. The aluminian prochlorite is in sheaves and apparently crystallized in equilibrium with the high-grade minerals. Clinozoisite (2V near 90°) is in small, compact domes. Zoisite is in large subhedral domes (fig. 39) that are very poikilitic with wormy inclusions of quartz and was apparently in equilibrium with the clinozoisite. An origin as a volcano-derived sediment containing impure carbonates can be suggested for this rock.

Phyllonite.—Many of the phyllites in the vicinity of Riggins display compound metamorphic fabrics, but true phyllonite—previously coarse schist, subsequently sheared so thoroughly that it is now shiny phyllite—was seen only on the ridge 0.8 mile north of Sheep Creek at an altitude of 3,900 feet. There, it is a shiny medium-gray rock that consists of knobby lenses of broken and bent prochlorite alternating with knots and lenses of sutured quartz and plagioclase.

ORIGIN

The diverse rock types of the Squaw Creek Schist are interlaminated and interlayered in complex sequences. Although most bulk rock compositions are near those of volcanic rocks, the common pervasive carbon dust, the nonsystematic layering, and the abundant interbeds of obviously sedimentary rocks indicate that the formation is of general sedimentary origin, presumably marine.

The composition, however, indicates that the clastic sediments have been derived from a source region of volcanic rocks. Unlike most sedimentary rocks, most of these have $Na_2O > K_2O$. Most other elements are similarly present in amounts and proportions like those of the purely metavolcanic rocks of the Riggins Group and unlike those of most sediments. Nothing seen requires any contribution of clastic sediment from nonvolcanic sources. Water-laid tuff, and siltstone derived from tuff, were probably the characteristic initial rock types.

The schists have been folded isoclinally and extremely sheared. Discontinuous lenses have been formed by shearing apart of limbs of isoclines and by boudinage. The result of these diverse processes of pervasive shear is a lamination that appears on casual observation to consist of a conformable sedimentary succession. Actually, many contacts between laminae are shear, not bedding, features, and right-side-up and upside-down layers alternate. The compositional differences are inherited from sedimentary laminae, but their present succession is in considerable part due to shearing.

ROCKS OF RIGGINS GROUP NORTH OF RIGGINS QUADRANGLE

The Riggins Group, represented by the Fiddle Creek Schist, continues north of the Riggins quadrangle along the Salmon River (pl. 2) where it lies tectonically upon the Martin Bridge Limestone. The Riggins Group is progressively cut out northward by an intrusive mass of sugary trondhjenite gneiss. In the Riggins quadrangle this gneiss is in contact with the Fiddle Creek Schist, but within 6 miles north of the quadrangle the schist has been cut out, and the gneiss is in contact with the Martin Bridge Limestone. There are no rocks of the Riggins Group along the Salmon River below Lucile at least as far north as Whitebird, and there are none along Slate Creek or the road going southeast from the mouth of Slate Creek to the Riggins quadrangle.

The schists of the Riggins Group thus form a lens, mostly within the Riggins quadrangle, between intrusive and migmatitic plutonic rocks on the east and the Rapid River thrust and the underlying Permian and Triassic sedimentary rocks and greenstones on the west.

Reconnaissance along the South Fork of the Clearwater River north and northeast of the northeast corner of the Riggins quadrangle shows that rocks similar to schists of the Riggins Group crop out only at several places between Cove Creek and Mill Creek, 17 to 20 miles upstream from Harpster. Septa of gray biotitic and calcareous schists dip steeply east-southeastward between plutons of granitic rocks and are similar to rocks of the Squaw Creek Schist. The biotite schists are strongly foliated medium-gray layered rocks colored by carbon dust and containing red-brown biotite, quartz, and reverse-zoned oligoclase, in increasing order of abundance, plus a little of either muscovite or actinolitic hornblende. The calcareous rocks are foliated fine-grained impure calcite marbles.

AGE AND CORRELATION

No fossils have been found in the Riggins Group. The rocks are older than the Idaho batholith of middle Cretaceous age and are presumed to be no older than Cambrian; they are otherwise undated, and no strong suggestions of correlations, based on lithology, can be made with dated rocks of nearby regions.

The metamorphism of the Riggins Group was accompanied by great shearing. Both the Riggins Group and the adjacent gneisses, which were derived in part from it, are cut out by intrusive granitic rocks north and south of the Riggins quadrangle. In both directions, the plutons of the Idaho batholith transgress across both the group and the gneisses until they are in direct contact with the low-grade metavolcanic and metasedimentary rocks of the Permian and Triassic sequence. The Riggins quadrangle contains all the known low-grade rocks of the Riggins Group.

The present sequence of the schist formations of the Riggins Group is consistent insofar as they can be traced or correlated in the quadrangle. The sequence might be a stratigraphic one—although if it is, it could be upside down instead of right side up. The contact between the very dissimilar Squaw Creek and Lightning Creek Schists is marked discontinuously by lenses of metaperiodotite, and these ultramafic rocks might have been intruded along a major premetamorphic thrust fault. Many such faults in other regions are followed by intrusions of serpentine, and the alternative explanation—that the ultramafic rocks were intruded fortuitously along a stratigraphic contact between two thick formations—seems unlikely. The present order of the formations of the Riggins Group is thus not necessarily indicative of stratigraphic sequence.

The Riggins Group is a eugeosynclinal assemblage of the andesite-keratophyre association. Rocks of this association characterize a broad belt in California, Oregon, Washington, and British Columbia and are of diverse Paleozoic and Mesozoic ages. They are probably most abundant in sequences of late Paleozoic and early and middle Mesozoic ages, but earlier Paleozoic suites are also broadly represented. Most rocks within this belt are metamorphosed; reliable stratigraphic dates are few; and many assignments and correlations can be challenged. The Riggins Group might reasonably be correlated with rocks of any age from Cambrian through Early Cretaceous. (The Belt Series, of middle or late Precambrian age, in the northern Rocky Mountains is very different, and correlation with it is unlikely.)

The premetamorphic rocks of the Lightning Creek and Fiddle Creek Schists were volcanic, similar in many respects to the Seven Devils Volcanics of Permian and Triassic ages now nearby to the west. The Seven Devils Volcanics, however, show no gross subdivision into formations, contain no rocks comparable to the Squaw Creek Schist, and may be systematically different in their iron-magnesium ratios. (If the Riggins Group and Seven Devils Volcanics are correlative, as seems unlikely, then the major differences between them might be explained by great telescoping on the Rapid River thrust.)

In the Baker quadrangle, 50 miles west-southwest of the Riggins quadrangle, Gilluly (1937) recognized three formations of stratified rocks of pre-Tertiary age. Highest is the Clover Creek Greenstone of Permian age, composed of metavolcanic rocks and subordinate metasediments at least 4,000 feet thick. This formation is correlative in both age and lithology with part of the Seven Devils Volcanics. Beneath the Clover Creek Greenstone is the Elkhorn Ridge Argillite (Pennsylvanian(?)), composed of metamorphosed carbonaceous argillite, tuffaceous argillite, tuff, and chert, with subordinate limestone, perhaps 1 mile thick; these rocks differ conspicuously from the Squaw Creek Schist by being much more massive and nonlaminated. The lowest stratified formation of the Baker quadrangle, the pre-Carboniferous(?) Burnt River Schist, is 1 mile or more thick and consists of greenschist, chiefly metavolcanic, with subordinate quartz phyllite, quartzite, and limestone. This formation is more metamorphosed than the others of the Baker region; it is most like the schists of the Riggins group, but the similarities are not strong enough to be considered any basis for correlation.

The brief descriptions by Ross (1938) of the metamorphic rocks of the southern Wallowa Mountains do not suggest that the rocks there are likely equivalents of the schists of the Riggins Group. Rather, it seems to be the Permian and Triassic sequence of the western part of the Riggins quadrangle that is chiefly represented in the Wallowas.

The Idaho State geologic map (Ross and Forrester, 1947) assumes the Riggins Group to be Triassic in age.

Metavolcanic rocks designated as the Casto Volcanics are known in various areas in central Idaho, southeast of the Riggins quadrangle. Although these rocks have yet to be dated, Ross (1927) and Ross and Forrester (1947) suggested a Permian(?) age for them on the basis of presumed correlation with part of the Seven Devils Volcanics despite great differences in petrology and history.

MINERALOGY

The Riggins Group is composed of four mappable formations and a wide variety of metamorphic rocks. The mineralogy of these rocks varies more significantly, however, with metamorphic grade than it does with stratigraphy, and it is accordingly appropriate to consider many features of the mineralogy for the entire group. The following material applies to the Riggins quadrangle alone unless specified otherwise.

Mineral determinations were made by petrographic study of thin sections, with recourse to the universal stage only for several precise determinations of 2V and to index liquids only for the determination of indices of zoisite. (2V was generally determined by Winchell's [1946] nomogram.) Identifications and terminology are chiefly based on Tröger's (1956) excellent tables and figures. General color names such as green or brown are used casually; specific names such as moderate brown or grayish red are based on the National Research Council's Rock-Color Chart (Goddard and others, 1948).

PLAGIOCLASE

The main mineral component of most of the Lightning Creek and Fiddle Creek Schists is albite or oligoclase. Oligoclase is a major mineral in the Berg Creek Amphibolite. Albite, oligoclase, and andesine are widely present in the Squaw Creek Schist, although plagioclase is lacking in many of the metasedimentary rocks of that formation. The plagioclase of the low-grade rocks is albite, whereas that of the middle- or high-grade ones is generally more calcic.

The albite of low-grade rocks is chiefly in mosaics of small unzoned crystals. Porphyroblasts are uncommon. Pseudomorphs of volcanic feldspars are abundant in the Fiddle Creek Schist, common in the Lightning Creek Schist in the Rapid River–Little Salmon River area, scarce in the Lightning Creek north of Riggins, and lacking elsewhere.

East of an irregular line near the Salmon and Little Salmon Rivers (fig. 68), the feldspar is chiefly reverse-zoned oligoclase in which sodic cores have more calcic rims. The oligoclase is in equigranular mosaics, less commonly in anhedral porphyroblasts. The composition is near An_{20} at Riggins, and mostly near An_{30} east of Little Berg Creek along the Salmon River. South of the river, and extending almost as far west as the Little Salmon River, andesine is widespread in calc-silicate schists. Many of these calc-silicate rocks have normally zoned plagioclase, the rims being more sodic than the cores. South of Riggins, the Lightning Creek Schist–Squaw Creek Schist contact is approximately coincident with the western limit of oligoclase.

Relict twinning is conspicuous in albite pseudomorphs after plagioclase phenocrysts (figs. 10 and 15), but most new crystals of albite are untwinned; the exceptions are mostly single twin pairs. The proportions of both singly and multiply twinned crystals increase irregularly eastward into the higher-grade rocks. Acline or pericline twins are about as abundant as albite twins, although either type may be dominant in any one specimen. Carlsbad twins are much less numerous. The western limit of rocks whose plagioclase is conspicuously twinned is about 1 mile east of the oligoclase limit, and such twinning thus represents a somewhat higher degree of metamorphism.

Most plagioclase east of the oligoclase isograd, almost to the eastern limit of the rocks here designated as "metamorphic," is zoned. Unzoned plagioclase, about An_{30}, occurs near Lake Creek, however, and also in the southern part of the Squaw Creek Schist. (This eastern limit of zoned plagioclase is not adequately enough defined to locate an isograd—samples are too few in the areas through which it must pass—but this limit also has metamorphic-grade significance.) A crystal of zoned plagioclase, either normal or reverse, consists generally of a small anhedral core and a rather sharply bounded rim. Only several crystals were seen in which an oscillation in zoning occurs, and in these also, elements are anhedral. Nothing comparable to the euhedral-oscillatory zoning characteristic of granitic plagioclase is present.

Albite and quartz are difficult to distinguish in the fine-grained rocks. Only quartz is common in coarser segregated lenses, but quartz and albite occur together in the groundmass mosaics and look much alike. No potassic feldspar was seen in the schists of the Riggins Group.

QUARTZ

Most of the schists of the Riggins Group contain quartz. It is a major component of the Fiddle Creek

Schist, is generally present in small to moderate amounts in the Lightning Creek Schist, is at most a minor mineral in the Berg Creek Amphibolite, and ranges from zero to nearly 100 percent in rocks of the Squaw Creek Schist.

Quartz is everywhere anhedral. It occurs as small crystals in the groundmass, as larger grains in segregated lenses (fig. 25), or as augen with or without other minerals. Quartz forms mosaics of generally evenly sized grains, which have a dimensional elongation of as much as 2:1 in some rocks but are commonly more nearly equant. Few schists have a crystallographic orientation of quartz strong enough to be apparent in thin section. (By contrast, flaser gneiss and phyllonite, derived from plutonic rocks, have very marked orientations in which the *c* axes of the quartz crystals lie mostly near the foliation plane.) Poikilitic porphyroblasts are scarce.

Quartz is unstrained in most of the schists. Sutured boundaries between grains are few. By contrast, sutured contacts are the rule in the metaquartz diorite in the Seven Devils Volcanics.

PROCHLORITE

Most of the schists of the Riggins Group contain prochlorite in varying amounts. It is more abundant in the Lightning Creek Schist than in the Fiddle Creek Schist, yet it is much more abundant in the Fiddle Creek Schist than in the Squaw Creek Schist. The approximate frequency distribution of prochlorite in the 125 thin sections studied from these three formations is as follows:

Amount of prochlorite (volume percent)	Percent of specimens		
	Lightning Creek Schist	Fiddle Creek Schist	Squaw Creek Schist
None	5	15	35
Trace–2	20	20	45
3–9	25	20	10
10–40	50	45	10

The differences are directly reflected in the rock colors. The lower grade rocks of the Lightning Creek Schist are invariably green, colored chiefly by their chlorite. Fiddle Creek rocks are lighter colored, and many are white or nearly so. In the Squaw Creek Schist, however, even the richly chloritic rocks are gray, as the color of the prochlorite is masked by pervasive black opaque dust.

In schists in which it is a major component, prochlorite is primarily a groundmass mineral occurring in tiny flakes subparallel to the cleavage direction and either concentrated in folia or strewn throughout (fig. 25). Where it is a minor component, prochlorite occurs characteristically as thick subhedral porphyroblasts as much as 0.5 or 1 mm long that are randomly oriented and cut across the foliation at all angles (for example, fig. 9). It is also often in sheaves. Porphyroblasts may be either younger (fig. 28) or older than a second cleavage direction, where such is apparent in the rock fabric. Many specimens have both oriented flakelets and random porphyroblasts. Many small porphyroblasts are aggregated in poorly oriented folia or clumps (fig. 12). In several specimens studied, oriented porphyroblasts of prochlorite define an otherwise unrecognizable secondary cleavage across an older foliation.

The common chlorite is aluminian prochlorite. It is green in hand specimen, colorless to faintly pleochroic light yellowish green in thin section, and displays anomalous yellowish-gray to grayish-green interference colors. It has a small $(+)2V$. Porphyroblasts are twinned multiply parallel to the base.

Ferroan prochlorite is found in many of the lower grade rocks. This chlorite, also green in hand specimen, is richly pleochroic (X = pale greenish yellow, Z = moderate yellow green) and has a bright-brown to reddish-brown anomalous interference color. A few other specimens have prochlorite that is markedly pleochroic, like the ferroan variety, but which have yellowish or greenish gray interference colors, more like the aluminian variety; presumably this is an intermediate type.

Several specimens contain green chlorite that is pleochroic and has an anomalous blue or purple interference color. In one, this is the only chlorite in the rock, and radial sheaves of it cut across the foliation, as does aluminian prochlorite nearby. In another, metamorphic hornblende has retrograded to pseudomorphs of carbonate, epidote, and chlorite, the chlorite varying in interference color from anomalous blue to anomalous brown in single crystals.

Prochlorite is so abundant in four of the specimens analyzed that its composition can be approximated by assuming compositions for the modal minerals and calculating subtractively to the prochlorite. Such methods have validity only where there is little ambiguity as to mineralogic location of oxides, of course. The following table shows the calculated compositions, in weight percent, of prochlorites in analyzed schists. Numbers refer to columns in table 5.

The prochlorite of Nos. 3 and 9 was classed optically as ferroan, as it is strongly pleochroic in green and shows an anomalous brown interference color. No. 5 was termed intermediate; it also has a strong pleochroic color but is anomalous gray as seen through crossed nicols. No. 7 was classed as aluminian because it is almost colorless, and its interference color is anomalous yellowish gray.

	Ferroan		Intermediate	Aluminian
	No. 3	No. 9	No. 5	No. 17
SiO_2	35	[1] 35	35	35
Al_2O_3	15	20	22	29
Fe_2O_3	(2)	3	(2)	2
FeO	21	18	16	8
MgO	18	15	17	16
H_2O	11	10	10	10

[1] Calculations based on mode gave 48 percent SiO_2. As this indicated a probable greater abundance of quartz in the analyzed specimen than in the thin section, the composition was recalculated using an assumed 35 percent SiO_2 in prochlorite.
[2] Fe_2O_3 assumed to be entirely in other minerals.

The calculated compositions are in good accord with this classification and have the following oxide formulas:

No. 3: $3MgO \cdot 2FeO \cdot Al_2O_3 \cdot 4SiO_2 \cdot 4H_2O$
No. 9: $1.9MgO \cdot 1.7FeO \cdot 1.3Al_2O_3 \cdot 4SiO_2 \cdot 4H_2O$
No. 5: $3.2MgO \cdot 1.7FeO \cdot 1.6Al_2O_3 \cdot 4SiO_2 \cdot 4H_2O$
No. 17: $3MgO \cdot 0.8FeO \cdot 0.1Fe_2O_3 \cdot 2Al_2O_3 \cdot 4SiO_2 \cdot 4H_2O$

The substitution seems to be primarily one of Al_2O_3 for FeO. Except for the aberrant and most dubiously calculated specimen 9, MgO remains constant while Al_2O_3 and FeO vary in complementary fashion. The end members of the series can be considered as ferroan prochlorite,

$$3MgO \cdot 2FeO \cdot Al_2O_3 \cdot 4SiO_2 \cdot 4H_2O,$$

and aluminian prochlorite,

$$3MgO \cdot FeO \cdot 2Al_2O_3 \cdot 4SiO_2 \cdot 4H_2O.$$

The indicated substitution of aluminum for iron is in accord with the conclusions by Brindley and Gillery (1956, p. 169), based on crystal chemistry, that "in orthochlorites the principal compositional variables are the Al ions in tetrahedral coordination and the Fe^{2+} ions in octahedral coordination."

These calculated compositions are also in excellent agreement with the theoretical prochlorite of standard texts (for example, Tröger, 1956, p. 85:

$$5(Fe, Mg)O \cdot Al_2O_3 \cdot 3\tfrac{1}{2}SiO_2 \cdot 7\tfrac{1}{2}H_2O)$$

except that those calculated here are less hydrous. Whether this indicates a true difference or merely an analytical bias is unknown.

Prochlorite is lacking in most of the gneisses and coarser schists. Where it is present in them, its textural relations indicate retrogressive replacement, commonly associated with cataclasis, of mafic minerals of higher metamorphic grade. In come of the finer grained middle-grade schists, similar replacement is also apparent in fabric and compositional relations. In most of the middle-grade schists, however, fabric relations give no hint of possible retrogression in the formation of the prochlorite, which seems, on the contrary, to be part of the equilibrium assemblages (for example, fig. 32).

Chlorite and biotite are commonly intergrown where they occur in the same specimen. Porphyroblasts may be of the two minerals interlaminated, which suggests arrested transition in either direction. They occur also in clumps and folia in which their crystals are diversely oriented and mutually interpenetrating, and this apparently indicates simultaneous crystallization.

BIOTITE

Dark mica is widespread in the Riggins Group. (By contrast, it was not found within the Riggins quadrangle in the Permian and Triassic rocks beneath the Rapid River thrust.) It is present in two-thirds of the specimens studied from the Fiddle Creek and Lightning Creek Schists and five-sixths of those from the Squaw Creek Schist, but it is present in the Berg Creek Amphibolite only as a local alteration product of hornblende. Biotite increases in abundance in a general way eastward, reflecting a complementary decrease in muscovite. Many of the western rocks, but fewer of the eastern, contain only traces of biotite, and in many such rocks biotite is interlaminated or otherwise intergrown with chlorite (fig. 18). As few of the rocks are richly potassic, biotite is generally a minor component.

Biotite is typically larger, poorer shaped, and less well oriented than is muscovite. In the fine-grained schists it occurs chiefly as porphyroblasts that are commonly randomly oriented, anhedral, and poikilitic (fig. 8). Biotite is less common as small crystals in the fine schists; but as a groundmass mineral, it is commonly well oriented along the foliation planes, whereas porphyroblasts are commonly not parallel. Where crystals are parallel they are typically elongate dimensionally and alined. Exceptionally, elongate biotite has a linear, but not a planar, parallelism.

Biotite formed at different times in the development of the rocks. Its random orientation in many schists shows that much of it crystallized after differential movement had largely ceased. Where two generations of planar structures are recognizable, as along the Salmon River from Riggins to Race Creek, the biotite occurs variously parallel to the first cleavage but deformed by the second, parallel to the second (fig. 30), and cutting across crinkles formed by the second.

The optic angle was checked in a dozen specimens from various parts of the several formations and was found to be mostly $(-)2V \cong 5°$. Only exceptionally is it as large as 10° or as small as zero.

Pleochroic colors of biotite which vary according to rock assemblage and metamorphic grade, were tabulated according to seven pleochroic formulas. In the

Squaw Creek Schist, biotite is dominantly brown, pleochroic from pale grayish or yellowish orange (X) to light or moderate brown (Y, Z). In many schists intercalated with those carrying brown biotite and apparently representing differences in composition rather than in metamorphism, the biotite is orange (X= nearly colorless, Y and Z=yellowish orange). In some of the higher grade rocks it is yellowish brown (X= grayish yellow, Y and Z=moderate or dark yellowish brown). The lightly pleochroic biotite of some specimens appears lavender rather than black in hand specimen.

The biotite of the low- and middle-grade rocks of the Lightning Creek Schist is brown and mostly pleochroic from X=pale grayish orange or pale yellowish orange to Y and Z=light or moderate brown. That of the high-grade rocks near Lake Creek is green—pleochroic from pale greenish or grayish yellow to olive.

In the lower grade rocks of the Fiddle Creek Schist, within about 1 mile of the Salmon River, biotite is very light colored, pleochroic from nearly colorless (X) to yellowish orange (Y, Z). Farther southeast, the biotite is green, with X=pale greenish or grayish yellow and Y and Z=olive or olive brown.

Much of the biotite in the plutonic rocks, east and south of the schists is dark olive (X=pale greenish yellow, Y and Z=olive gray), darker than any in the schists, but some is red and brown.

Considered broadly, there thus seems to be a general change in biotite color from light brown to dark green with increasing metamorphism, but this is irregular and varies much with rock compositions.

Chemical analyses of the rocks in which biotite is the much-dominant ferromagnesian mineral (table 5, Nos. 10, 11, 14, 15, 19, and 20) indicate that the biotite contains more FeO than Fe_2O_3 by a factor near 3. This puts them within the field of common biotites of gneisses and schists defined by Heinrich (1946). The Y and Z color of the biotite in all these except No. 15 is light or moderate brown; that of 15 is olive. Although varying biotite colors have been explained in terms of various combinations of FeO, Fe_2O_3, MgO, TiO_2, and MnO, the data here do not suggest such a reason for the variations. The chemical distinction apparently is slight enough to be masked in the bulk-rock compositions by effects of the minor amounts of the other mafic minerals. Most of the biotite also seems to be richly potassic, containing about 10 percent K_2O by weight. Because of such uncertainties as the mineralogic location of Fe_2O_3 and TiO_2 in most of these biotite-rich rocks, calculated compositions are not presented here.

Stilpnomelane was not recognized in any of the metamorphic rocks.

MUSCOVITE

White mica is a common component of the more potassic schists of low and middle metamorphic grade. It is present in most of the Fiddle Creek Schist and much of the Squaw Creek Schist but is uncommon in the Lightning Creek Schist. Its approximate frequency distribution in the 125 thin sections from these three formations follows:

Amount of muscovite (volume percent)	Percent of specimens		
	Fiddle Creek Schist	Squaw Creek Schist	Lightning Creek Schist
0	10	40	75
Trace–2	30	25	15
3–9	15	10	10
10–30	45	25	0

Muscovite is not present in the Berg Creek Amphibolite.

Most of the muscovite occurs as thin well-shaped flakes, each near 0.1 mm long, oriented parallel to the foliation and either strewn throughout the rock (fig. 8) or concentrated in folia. Crosscutting and porphyroblastic muscovite is widespread but of far less total abundance. Where muscovite occurs with either biotite or chlorite, it is generally the better oriented and better shaped. The muscovite is colorless to faintly green in thin section and has a $(-)2V$ of 30° to 40°.

Muscovite is lacking in the high-grade schists but is present in the low- and middle-grade ones in unusual assemblages of apparent equilibrium. Muscovite and prochlorite gradually give way southeastward to biotite, but all three occur together in a broad zone. Prochlorite plus either biotite or muscovite is more common than is the combination of all three within this zone, but all do occur together in many schists (fig. 32). In some of these rocks, one or the other micaceous mineral is in crosscutting porphyroblasts or interlaminar intergrowths, so that either a two-cycle origin or an incomplete reaction can be postulated to explain the assemblage. In many other rocks, the three minerals occur together with no evidence for compound origin or reaction and are, instead, interpenetrating or intermingling as though they had crystallized simultaneously. More striking yet, muscovite, prochlorite, and biotite occur together in what seem to be equilibrium relationships in a few specimens that contain garnet and oligoclase or even andesine.

AMPHIBOLES

Monoclinic amphibole, chiefly hornblende, occurs in about half the schists of the Riggins group. It is lacking in most of the Fiddle Creek Schist but is present in the Lightning Creek and Squaw Creek Schists and the Berg Creek Amphibolite, listed in order of increasing quantity. The frequency distribution of amphibole in the 130 thin sections from these formations is as follows:

Amount of amphibole (volume percent)	Percent of specimens			
	Fiddle Creek Schist	Lightning Creek Schist	Squaw Creek Schist	Berg Creek Amphibolite
None	90	55	40	--------
Trace–2	5	5	10	--------
3–9	--------	10	5	--------
10–39	5	20	25	--------
40–90	--------	10	20	100

This amphibole is variably green to black in hand specimen. Although the green variety would be termed "actinolite" on the basis of both megascopic and microscopic appearance, the chemical data indicate that at least some of it is aluminian hornblende.

The amphibole of the Squaw Creek Schist and Berg Creek Amphibolite is everywhere black in hand specimen but pleochroic in light colors (X=pale greenish yellow or light yellowish gray, to Z=grayish yellow green or pale yellowish green; or X=grayish yellow green, Y=pale yellowish green, Z=light green or light grayish green). The sparse amphibole of the Fiddle Creek Schist is also black, but it is darker in thin section (X=greenish or grayish yellow, Y=yellow green or dark yellow green, Z=grayish, bluish, or dark green) than that of the above units. Amphibole in the Lightning Creek Schist is more varied. In the area of higher grade rocks, at the north and east edges of the formation along the Salmon River, the amphibole is black and darkly pleochroic, like that of the Fiddle Creek Schist. In the rest of the exposures along the Salmon River, the amphibole is black but light in thin section, like that of the Squaw Creek Schist. In the Rapid River-Little Salmon River area, where the lowest grade rocks are present, the amphibole is grayish green in hand specimen and pleochroic from X=pale greenish yellow or light yellowish gray to Z=grayish yellow green or pale yellowish green.

Distribution of these types is shown in figure 71. The differences must partly reflect compositional changes between the formations, but they are due chiefly to varying grade of metamorphism. Black hornblende is of higher grade than green hornblende or actinolite; and within the black, the more darkly pleochroic is of the higher grade assemblage.

In the fine-grained schists, hornblende is commonly porphyroblastic. It generally forms prisms that are variably short to acicular, ragged to euhedral, or poikilitic to compact (fig. 20). The prisms are commonly crudely alined (figs. 34 and 35), although in many

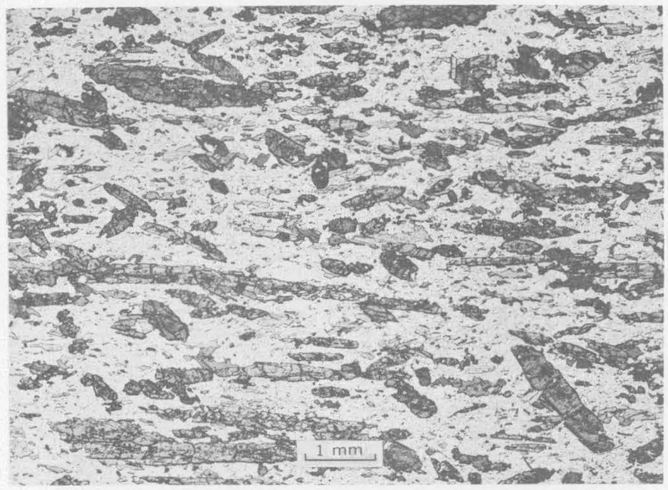

FIGURE 34.—Hornblende schist, cut parallel to alinement of prisms. Squaw Creek Schist, Squaw Creek. Plane-polarized light.

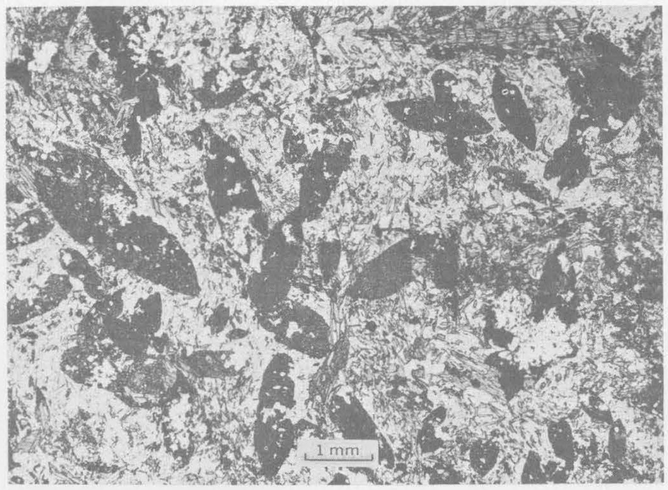

FIGURE 35.—Hornblende schist, cut perpendicular to alinement of prisms. Lightning Creek Schist, Salmon River near Lightning Creek. Plane-polarized light.

schists the hornblende has a planar but not linear parallelism. One of the most striking rocks of the quadrangle is greenschist of the Lightning Creek Schist, north of Riggins, which is liberally strewn with euhedral black hornblende crystals as long as 2 cm. Such rocks are exceptional, and in most of the fine-grained schists, amphibole crystals are not visible without magnification. In the coarser grained rocks, hornblende is of dominantly granular or short-prismatic habit.

In four of the analyzed schists (Nos. 2, 4, 6, and 8, table 5), hornblende is so abundant that subtractive calculations based on the modes can indicate its approximate composition. The following table shows the calculated chemical composition, in weight percent, of hornblende in the four specimens:

	Black in hand specimen		Green in hand specimen	
	No. 2	No. 8	No. 4	No. 6
SiO_2	43	48	39	44
Al_2O_3	15	15	11	15
Fe_2O_3	.8	1	.5	(1)
FeO	18	12	18	11
MgO	9	10	13	13
MnO	.5	.3	.5	.2
CaO	10	10	14	14
Na_2O [1]	--------	1.3	--------	--------
K_2O [1]	.5	.2	.3	--------
TiO_2 [1]	2	1.3	.8	--------
H_2O	2.5	1.3	2.2	2.3

[1] Assumed to be in other minerals except where listed.

These calculated hornblendes have the following oxide formulas, omitting the minor components and calculating each to an oxygen-ion sum of 24:

No. 2—$1.6CaO \cdot 2MgO \cdot 2.2FeO \cdot 1.4Al_2O_3 \cdot 6.3SiO_2 \cdot 1.2H_2O$

No. 8—$1.6CaO \cdot 2MgO \cdot 1.4FeO \cdot 1.3Al_2O_3 \cdot 6.9SiO_2 \cdot 0.6H_2O$

No. 4—$2.3CaO \cdot 3MgO \cdot 2.3FeO \cdot Al_2O_3 \cdot 6SiO_2 \cdot H_2O$

No. 6—$2.2CaO \cdot 2.8MgO \cdot 1.3FeO \cdot 1.3Al_2O_3 \cdot 6.4SiO_2 \cdot 1.1H_2O$

These compositions are fair fits for the general structural formula of the hornblendes,

$$(K,Na,Ca)_{2-3}\,(Mg,Fe^{+2}, Mn, Al, Fe^{+3}, Ti)_5\,(Si, Al)_8O_{22} \cdot (OH,F,Cl)_2,$$

in which the sum of $Mg+Fe^{+2}+Mn$ is about 4, except that water is consistently low (which perhaps indicates an analytical bias). All are richly aluminous—that is, all are hornblendes, not actinolites.

Of these 4 calculated hornblendes, 2 (Nos. 4 and 6) are green in hand specimen, weakly pleochroic light yellowish green, and have small extinction angles. The other two are black in hand specimen; of these, No. 2 is nearly as light in thin section as the green hornblendes. The other, No. 8, is a little darker and has a maximum pleochroic color of dusky yellow green. The calculated compositions suggest that the black hornblendes are poorer in MgO and CaO, and possibly richer in Fe_2O_3 and TiO_2, than the green.

The hornblende of specimens 4 and 6 was considered to be actinolite in thin-section study. Only one other analyzed specimen (No. 3) from the Riggins Group contains similar amphibole, and this is in a minor amount so that its composition cannot be calculated; nor can the composition of actinolite be deduced in the analyzed rocks of the Seven Devils Volcanics. Presumably, amphiboles of the lowest grade rocks are low in alumina. The occurrence of aluminous amphibole, hornblende, is apparently at lower grade in the Riggins quadrangle than would be expected by comparison with other regions.

The orthorhombic amphibole gedrite occurs as a minor component in meta-agglomerate of the Lightning Creek Schist just west of the bridge across the Salmon River at Lake Creek. The gedrite is in part in monominerallic prisms, variably compact to exceedingly poikilitic, and is in part intergrown, with crystallographic continuity, about the margins of hornblende. The gedrite is presumably a product of the retrograde metamorphism that affected the rocks of this locality. It has a birefringence of about 0.015, which indicates a magnesian composition, and is weakly pleochroic from pale greenish yellow to grayish yellow green. Amphibole cleavage is conspicuous. Orientation is x ∥ *a*, y ∥ *b*, and z ∥ *c*.

DIOPSIDE

Clinopyroxene was seen only in calc-silicate schist and gneiss of the Squaw Creek Schist within a 1-mile radius of a point 2 miles south of Shorts Bar, in and near the Hailey Creek drainage. The diopside is pale green, nonpleochroic, and has a (+)2V of 60° to 65°. It occurs as compact anhedra 0.2 to 1 mm across. Two diopside-bearing rocks were analyzed (table 5, Nos. 21 and 23) and are described in detail elsewhere.

EPIDOTE MINERALS

Monoclinic epidotes are common throughout the Riggins Group west of Berg Creek and north of Denny Creek and are sparse beyond those limits. Epidote proper is the only member of the group in the Seven Devils volcanics. In the Riggins Group, clinozoisite and aluminian epidote are abundant, but iron-rich epidote is uncommon. The epidotes make up less than 10 percent of most of the rocks in which they occur, although scarce rocks have as much as 30 percent of them.

Clinozoisite is colorless but shows bright to dark ultrablue interference colors. Aluminian epidote is also colorless in thin section but has a moderate birefringence, near 0.015 to 0.025, and is either gray or green in hand specimen. Iron-rich epidote is pleochroic in yellowish green, has a birefringence higher than 0.03, and is green in hand specimen. Clinozoisite and aluminian epidote are zoned together, or otherwise intergrade, or are present in adjacent laminae, in some of the schists.

About half the thin sections of Fiddle Creek Schist contain epidote or, less commonly, aluminian epidote or clinozoisite, in tiny granules or, less commonly in small domes or ragged anhedra.

In the Lightning Creek Schist, aluminian epidote is dominant over clinozoisite, and clinozoisite is dominant over iron-rich epidote. These minerals occur as granules (fig. 12), small anhedra, and small domes. Clinozoisite is more commonly microporphyroblastic than are the others. In one schist north of Riggins, laminae rich in prochlorite have accessory clinozoisite, whereas adjacent laminae rich in hornblende have aluminian epidote. About a third of the high-grade rocks near Lake Creek contain a little aluminian epidote in small domes parallel to the hornblende lineation.

The Berg Creek Amphibolite has only a little clinozoisite as small poikilitic anhedra.

Half the specimens of Squaw Creek Schist have monoclinic epidote, generally clinozoisite, that occurs as granules and small domes in low- and middle-grade schists and as porphyroblasts in high-grade calc-silicate schists. In the calc-silicate rocks, epidote occurs variously as domes (fig. 36), ragged anhedra (fig. 36), and micrographic rod symplektites (fig. 37).

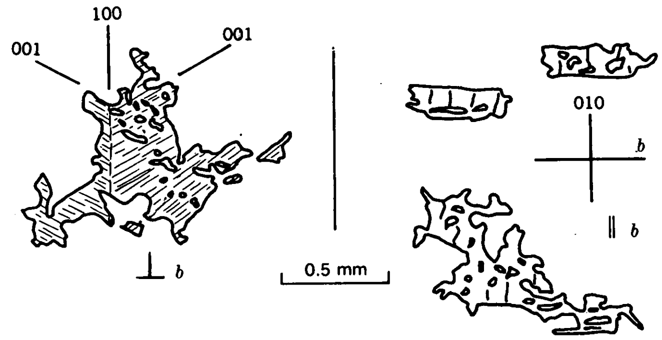

FIGURE 36.—Anhedra and domes of clinozoisite. Left, anhedron, twinned on [100], showing strong [001] cleavage. Right, two domes and an anhedron. Hornblende-biotite-clinozoisite-quartz-andesine schist, Squaw Creek Schist, Hailey Creek.

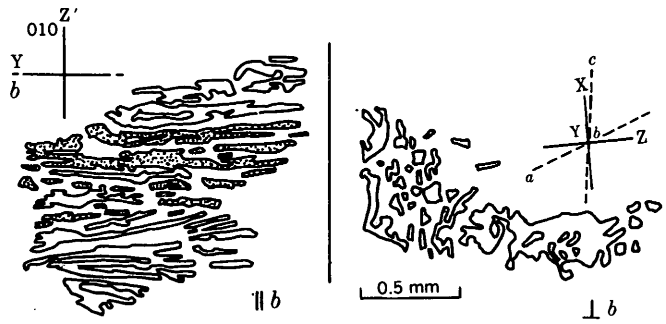

FIGURE 37.—Micrographic clinozoisite porphyroblasts. Left, three(?) subparallel crystals, cut parallel to *b*; elements of central crystal are stippled. Right, cross section of similar group. This clinozoisite is intergrown with plagioclase, quartz, and calcite, and the rock contains also diopside and scapolite. Squaw Creek Schist, Hailey Creek.

The western limit of clinozoisite and aluminian epidote, as opposed to iron-rich epidote, has about the same significance of metamorphic grade as does the biotite isograd (fig. 70). To the east, relations are less distinctive, as epidote-group minerals in some high-grade rocks appear to be in equilibrium assemblages, and an upper isograd cannot be drawn about their occurrences. Some of the plutonic rocks in the southern half of the Riggins quadrangle contain abundant epidote, much of it euhedral, that further confuses the relationships.

A few schists of the Squaw Creek carry porphyroblasts of zoisite—euhedral domes as large as 0.5 by 1 by 15 mm that have a superficial resemblance to prisms of hornblende (figs. 38 and 39). The zoisite is colorless in thin section and glassy greenish gray in hand specimen, but included magnetite and the backup color of the dark-gray rocks in which it occurs commonly make it appear black. The crystals are bounded by the [101] faces, with lesser development of the [100] pinacoids. (Older texts class such zoisite crystals as prisms rather than domes, but the structural similarity to the monoclinic epidotes, whose symmetry proves the dome form of the analogous crystals, makes the dome designation preferable.) The domes have strong [100] cleavage and [010] parting and conspicuous striations parallel to *b*. Extinction is straight and symmetrical, and sections perpendicular to Bxa show a rich ultra blue for 25° of rotation on one side of extinction; there is strong crossed-axial-plane dispersion. In the analyzed schist (table 5, No. 20), the zoisite has a density of 3.35±0.1, (+)2V=15°, n_x=1.692, n_y=1.693, n_z=1.701, and n_z-n_x=0.009. In the other schist described on page 24, (+)2V is near zero degrees.

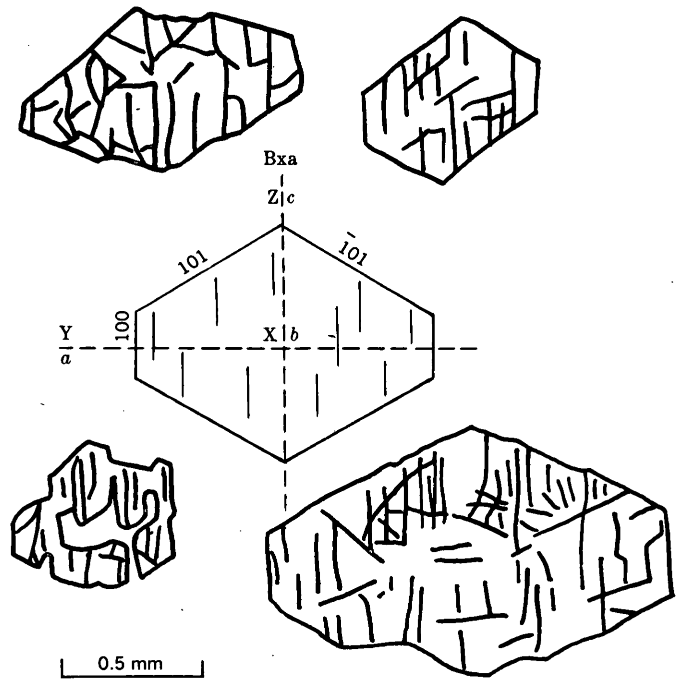

FIGURE 38.—Cross sections of porphyroblastic zoisite domes. Garnet-zoisite-clinozoisite-quartz-biotite-plagioclase schist, Squaw Creek Schist, Salmon River east of Riggins.

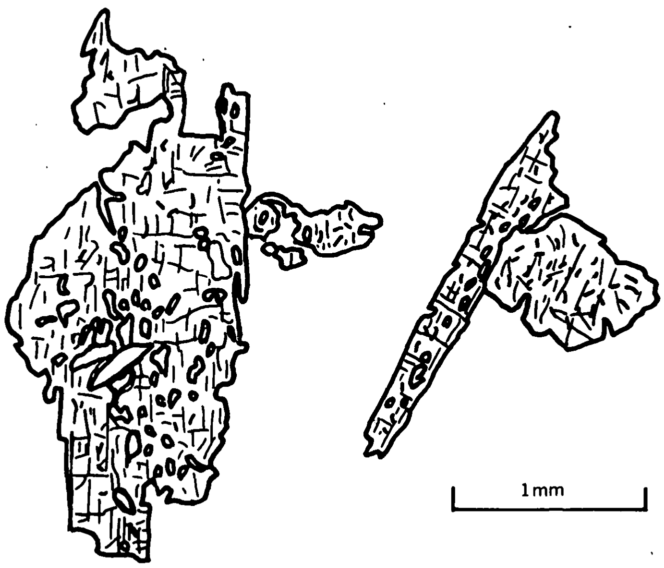

FIGURE 39.—Poikilitic, subhedral domes of zoisite in calc-silicate schist. Left crystal of each pair is shown in long section, cut parallel to *b*; right of each is cross section, perpendicular to *b*. Squaw Creek Schist, Hailey Creek, 0.4 mile above mouth.

The zoisite occurs with andesine and other minerals indicative of high metamorphic grade. (It is also an uncommon component of amphibolite east of Lake Creek that is classed with the plutonic rocks.) Zoisite was seen in middle-grade rocks only in a single greenschist from the Lightning Creek Schist 0.8 mile south of Lightning Creek; this rock also contains hornblende, clinozoisite, and oligoclase.

Tiny domes of possible zoisite occur in various other middle-grade schists, but these crystals are too small to be distinguished optically from clinozoisite. Porphyroblasts of zoisite indicate high metamorphic grade, but possibly groundmass zoisite occurs at lower grade.

GARNET

Red-brown garnet is a common minor component of all but the lowest grade schists. The garnet isograd—the western limit of garnet (fig. 68)—is everywhere west of the Salmon River and west of the Little Salmon River except between Pollock and Rapid River. Garnet is present in half the specimens of Squaw Creek Schist and Berg Creek Amphibolite collected east of the isograd, in two-fifths of those from the Fiddle Creek Schist, and in one-third of those from the Lightning Creek Schist. (In the Lightning Creek Schist garnet is mostly in the lighter colored greenschists in the section north of Riggins and in the high-grade rocks near Lake Creek east of Riggins.) Although generally making up less than 1 percent of the schist, rocks with as much as 2 percent garnet are common, and a few have as much as 5 percent.

The garnet is reddish brown or, less commonly, brown or light to moderate red in hand specimen. Most is grayish orange pink in thin section. None is birefringent.

Habit varies widely (fig. 40); euhedral, subhedral, anhedral, and extremely irregular crystals are all abundant. Although habit is consistent in any one specimen, no areal patterns of habit were noted. Poikilitic crystals are more common than compact ones; inclusions are mostly quartz, although clinozoisite is present in garnet in some calc-silicate rocks. Garnet crystals average as small as 0.05 mm in some schists and as large as 5 mm in others; in most, the grains are near 1 mm.

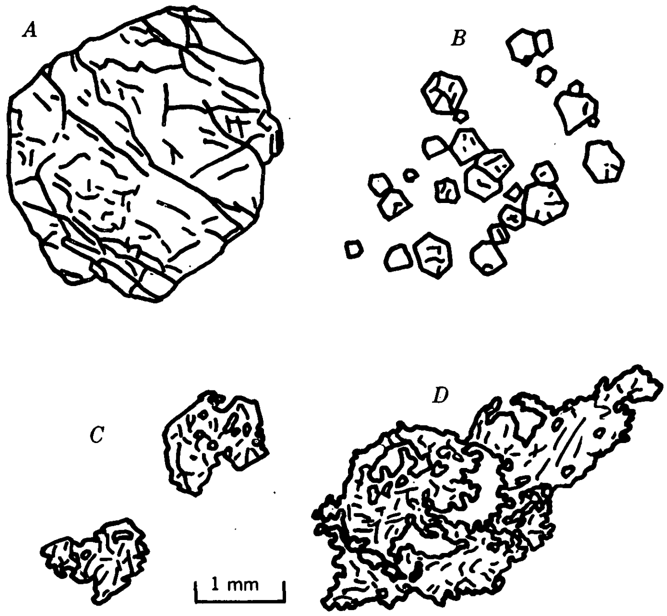

FIGURE 40.—Garnet in schists of Riggins Group. Each habit characterizes many schists. *A*, Large dodecahedron; *B*, small dodecahedra; *C*, poikilitic subhedral crystals; *D*, irregular crystal, elongate in the plane of foliation.

Garnet typically grew across the foliation of the schists (fig. 31) and is thus of late, replacement origin. Some garnets partly cut across and partly shoulder aside folia. At other places, garnets are elongate along the cleavage (as, fig. 40*D*). Clear evidence of rolling—shearing after crystallization—was seen in only a few specimens (fig. 29).

Most garnet is unaltered. Some is partially replaced by chlorite, chiefly aluminian prochlorite.

SCAPOLITE

Only one thin section from the Riggins Group contains scapolite. This is an unusual calc-silicate rock (analyzed chemically, table 5, No. 21) composed of quartz, diopside, calcite, clinozoisite, andesine, and scapolite. The scapolite is in equant compact anhedra smaller than 1 mm. It is colorless and uniaxial negative. Birefringence of about 0.02 suggests that it is mizzonite. The chemical analysis of the rock shows 0.03 percent Cl and 0.03 total S as SO_3, both presumably in the scapolite.

CARBONATE

Over half the schist west of Little Berg Creek and north of Denny Creek contains carbonate. East and south of these limits, carbonate is a very minor component except in calc-silicate rocks and marble.

Much of the carbonate of the lower grade rocks is light gray anhedral calcite. This occurs as disseminated grains that vary widely in concentration between laminae but generally are most abundant in quartzose ones, either porphyroblastic or matching the groundmass minerals in size; in knots, augen, and lenses, with or without quartz and silicate minerals; and in veins, either with or without quartz and silicate minerals. Calcite is the only carbonate in rocks rich in carbonate—marble and impure marble, both low and high grade—and in calc-silicate rocks.

Of the schists that have carbonate as a minor component (trace to 10 percent), about half have a noncalcitic carbonate or have calcite that appears to be pseudomorphous after another carbonate (fig. 41). The

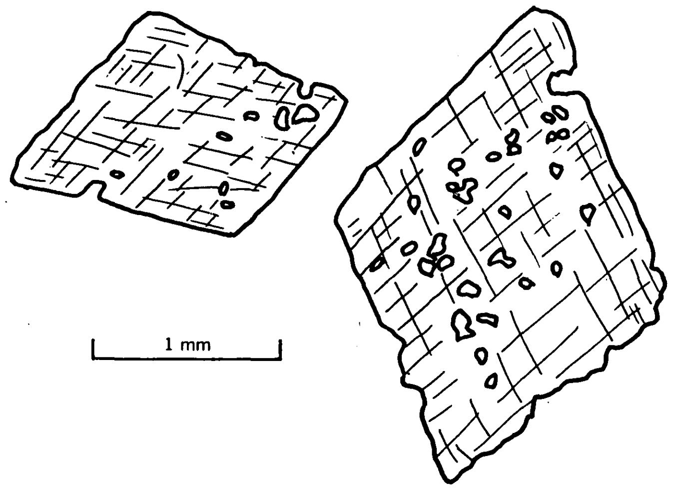

FIGURE 41.—Porphyroblastic rhombs of rust-stained calcite in greenschist. Calcite may be pseudomorphous after ankerite. Lightning Creek Schist, near mouth of Chair Creek.

noncalcitic carbonate is in conspicuous well-shaped porphyroblastic rhombs in many specimens. Some is brown in hand specimen, has n_e greater than the index of quartz or balsam, and is presumably siderite. Some is gray and has n_e less than that of quartz; presumably this is magnesite or ankerite.

Much of the noncalcitic carbonate is darkly clouded by magnetite(?) or by abundant limonite along the cleavages. The limonite apparently represents oxidation by weathering of iron in the carbonate. Many schists contain rhombs of calcite that are darkly stained by rust. As calcite does not normally crystallize in euhedra in metamorphic rocks, it is likely that these calcite rhombs are pseudomorphs, developed perhaps during weathering, after an iron-bearing carbonate.

Some carbonate porphyroblasts grew in several stages (fig. 42).

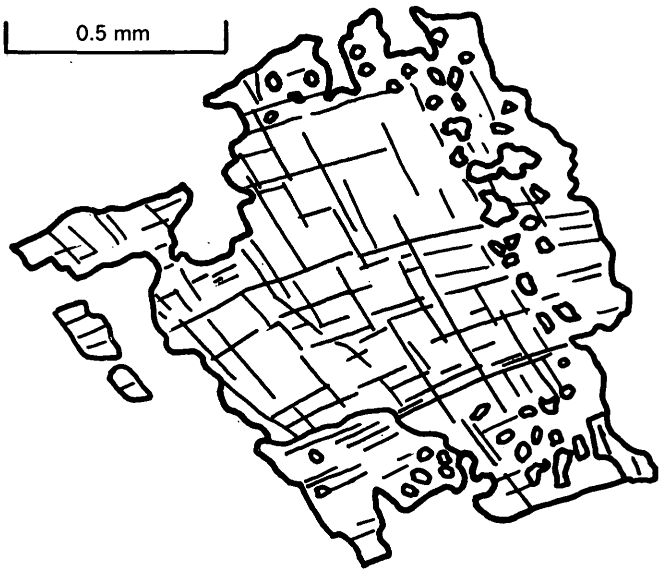

FIGURE 42.—Bent porphyroblast of ankerite in phyllitic metatuff. The ragged poikilitic rim grew upon a euhedral rhomb. Analyzed specimen (table 5), Fiddle Creek Schist.

OTHER MINERALS

Most of the schists contain from a trace to 2 percent sphene, leucoxene, and rutile. In the lower grade rocks these are typically minutely intergrown with each other in small irregular blebs, although much of the rutile occurs as tiny euhedral yellow prisms. In the higher grade rocks, sphene occurs as clear nonleucoxenic granules.

Euhedral plates of a black opaque mineral, presumably ilmenite, are strewn about some of the middle-grade greenschists.

A few tiny euhedral prisms of zoned green tourmaline occur in many schists of the Riggins Group, and tourmaline is abundant in some.

Small prisms or granules of apatite were seen in many thin sections and presumably are actually present in all the rocks.

Probably most of the schists of the Riggins Group contain traces of pyrite, and some have 1 or even 2 percent of it. Pyrite occurs as cubes (fig. 27), including porphyroblasts as large as 5 mm square, in some rocks, and as irregular granules, flattened blebs as long as a centimeter (fig. 28), and still longer stringers, in others. In natural outcrops, pyrite is largely altered to iron oxides, and sulfates coat weathered but protected rock surfaces.

Octahedra of magnetite occur in many of the schists. Granules and blebs of black metallic material are still more abundant, and presumably much of this is also magnetite. No hematite or limonite, except that apparently due to weathering, was seen in the Riggins Group. (In the Seven Devils Volcanics, by contrast, hematite is a major component of many rocks.)

Opaque black dust and tiny granules cloud a large proportion of the rocks of the Squaw Creek Schist. The abundance of this material varies greatly between laminae, and the granule-rich layers continue without interruption through diverse silicate porphyroblasts. Presumably this material is chiefly carbon and graphite, although carbon was reported in only several of the chemical analyses of these rocks.

None of the aluminum silicates andalusite, staurolite, kyanite, and sillimanite, nor any corundum, was found in the schists of the Riggins Group. These minerals, so useful as indicators of mineralogic grade of metashales in many regions, are not formed during regional metamorphism of igneous rocks, and none of the Riggins rocks are rich enough in aluminum and low enough in other metals to permit their formation at any metamorphic grade.

CHEMISTRY

ANALYTICAL DATA

Analyses for major and minor elements and mineral contents for 23 specimens of schists of the Riggins Group are presented in table 5. (Much of the following discussion is applicable also to the three analyzed specimens from the Seven Devils Volcanics, table 1, and to the two from the ultramafic rocks, table 6.) Chemical data and the modes of minerals present are given in table 5, and the specimens are described below.

Flows, tuffs, and sediments are represented, and the distinctions between these can be made with confidence for only some of the specimens. Shearing and mineralogic reconstitution have obliterated the primary textural features in most of the rocks. Clasts can still be recognized in the agglomerates and in some of the tuffs, and phenocrysts or even primary groundmass textures can be recognized in some of the flows, but these are exceptions. Visible layering of probable sedimentary origin is common only in the Squaw Creek Schist, where it is shown in most outcrops.

The chemical composition of most of the specimens plainly indicates that the Riggins group is largely of volcanic origin, regardless of details of waterworking. Thus, the analyzed layered rocks of the Squaw Creek Schist have virtually volcanic compositions. The distinction between "metavolcanic" and "metasedimentary" rocks (table 5) is made arbitrarily on the basis of chemical composition. Where most of the bulk composition is volcanic—where the rock must have been derived from a volcanic source, with or without the intervention of nonvolcanic agencies of transport and sedimentation—the rock is termed "metavolcanic." Most of the rocks so designated in the Squaw Creek Schist were redeposited by sedimentary agencies, but only those rocks whose compositions suggest major modification of volcanic composition by selective processes of weathering and sedimentation are classed here as metasedimentary. All the rocks were derived largely, and probably entirely, from a volcanic terrane.

The analyses of metavolcanic rocks of the three schist formations—Fiddle Creek, Lightning Creek, and Squaw Creek—fit essentially the same variation patterns, whatever components are considered. Sampling is too limited to attach significance to possible slight divergences in trends. These rocks are accordingly considered together in most of the following discussions. There are differences in frequency distributions—the Fiddle Creek Schist is in bulk more silicic than the Lightning Creek Schist, and other oxides vary accordingly—but the patterns of variations are the same.

Major oxides.—Silicate analyses were made by rapid methods by P. L. D. Elmore, P. W. Scott, S. D. Botts, K. E. White, M. D. Mack, and J. H. Goode in the Washington laboratories of the Geological Survey during 1956 and 1957. Determinations of most oxides by these methods are colorimetric rather than gravimetric.

Minor elements.—Nearly all the minor-element determinations were made by Paul R. Barnett by the rapid visual-comparison method of semiquantitative spectrography during 1956 and 1957 in the Denver laboratories of the Geological Survey. Quantities are reported as midpoint values of logarithmic-third divisions—as, 0.015, 0.03, 0.07, 0.15, 0.3, and so on. Comparisons of similar data with those obtained by quantitative methods show that the same class interval is assigned in about 60 percent of the samples. Analyses of specimens of all major rock groups were made simultaneously, so that the determinations are properly comparable. Elements not detected are tabulated as less than the threshold values.

TABLE 5.—*Composition, in percent,*

[Mineralogic compositions determined by point-counts. P, present but not intersected in counting. In specimens too fine grained for counting: A, abundant; C, common; and 21–24, which are by P. L. D. Elmore, S. D. Botts, M. D. Mack, and J. H. Goode (1957). Minor-element analyses

	Metavolcanic rocks								
	Squaw Creek Schist	Lightning Creek Schist					With ultramafic rocks	Berg Creek Amphibolite	Lightning Creek Schist
	1	2	3	4	5	6	7	8	9
Chemical laboratory No	148407	148402	148406	148408	148399	148400	151588	151589	148405
Spectrographic laboratory No	C 679	C 674	C 678	C 680	C 671	C 672	D 1402 / D 1336	D 1403	C 677
Field No	SR 125	SR 45C	SR 114	SR 127C	SR 34	SR 37-2	SR 524-1	SR 539	SR 51
Major oxides									
SiO_2	44.1	46.4	47.4	48.2	50.2	50.8	52.6	52.7	53.3
Al_2O_3	12.9	19.6	16.3	15.0	14.8	17.4	17.3	17.4	15.5
Fe_2O_3	2.0	1.9	2.2	2.0	4.2	1.5	3.0	1.8	3.2
Sum as Fe O	8.3	9.8	12.7	10.0	10.3	7.7	8.0	9.1	10.7
FeO	6.5	8.1	9.8	8.2	6.5	6.4	5.3	7.5	7.8
MgO	9.7	5.4	8.4	7.1	4.4	7.4	3.1	5.4	5.4
CaO	7.6	9.3	3.9	10.4	7.8	8.5	7.8	8.1	3.2
Na_2O	.23	3.7	4.1	3.1	3.5	4.2	5.2	3.8	2.6
K_2O	1.6	.23	.24	.22	.10	.15	.81	.22	.74
TiO_2	.38	.92	1.5	1.5	1.1	.78	1.0	.90	1.3
P_2O_5	.14	.20	.22	.14	.10	.12	.48	.23	.20
MnO	.18	.18	.19	.20	.22	.12	.22	.14	.20
H_2O	3.8	2.2	4.7	2.2	3.0	1.6	1.0	.81	4.0
CO_2	10.5	1.6	.14	1.6	3.8	.40	<.05	<.05	2.2
Total	100	100	100	100	100	99	98	99	101
Minor elements									
B	<0.001	<0.001	0.003	<0.001	<0.001	0.0015	<0.001	<0.001	<0.001
Ba	.07	.007	.007	.003	.0015	.003	.015	.007	.015
Be	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001
Ce	.015	<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01
Co	.0015	.003	.007	.003	.003	.003	.0015	.0015	.003
Cr	.07	.007	.003	.03	.0007	.03	.02	.003	.0015
Cu	.0003	.007	.015	.007	.007	.003	.003	.0007	.015
Ga	.0015	.0015	.002	.002	.0015	.0015	.0015	.0007	.002
La	<.002	<.002	<.002	<.002	<.002	<.002	.003	<.002	<.002
Mo	<.0005	<.0005	.0007	<.0005	<.0005	.0005	.0007	<.0005	<.0007
Nb	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005
Nd	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005
Ni	.007	.003	.003	.007	.0007	.007	.006	.003	.0015
Pb	<.0005	<.0005	<.0005	<.0005	.0007	<.0005	.0015	<.0005	<.0005
Sc	.007	.007	.007	.003	.007	.003	.003	.003	.007
Sr	.015	.03	.015	.015	.015	.03	.03	.03	.015
V	.015	.03	.03	.015	.03	.015	.015	.015	.03
Y	.0015	.0015	.003	.0015	.003	.0015	.003	.0015	.003
Yb	.0015	.0003	.0007	.0003	.0007	.00015	.0007	.0003	.0007
Zr	.003	.003	.007	.007	.003	.003	.015	.007	.007
Minerals									
Quartz	33	4±	----	6	15	----	A	3	7
Albite	----	----	≈35	23	39	36	A	----	37
Oligoclase	----	37±	----	----	----	----	----	42	----
Andesine	----	----	----	----	----	----	----	----	----
Zoisite	----	----	----	----	----	----	----	----	----
Clinozoisite	----	----	----	----	----	----	----	----	----
Epidote	----	6	≈10	7	14	2.0	C	----	2.0
Muscovite	19	P	U	P	----	----	----	----	6
Biotite	----	P	U	.2	----	----	U	.1	.6
Prochlorite	24	13	≈45	12	20	4	----	.7	38
Actinolite	----	----	U	----	----	----	----	----	----
Hornblende	----	34	----	43	----	55	A	53	----
Diopside	----	----	----	----	----	----	----	----	----
Scapolite	----	----	----	----	----	----	----	----	----
Garnet	----	----	----	----	----	----	U	----	P
Carbonate	22	5	U	7	9	1.0	----	----	8
Sphene, rutile, leucoxene	.2	.3	U	2.0	----	1.0	----	----	P
Opaque (ores+carbon)	1.8	.8	U	----	3	1.2	C	.4	2.0
Apatite	P	----	----	----	----	----	----	.4	.3
Others	----	----	----	----	----	----	U	.1	P

of schists of Riggins Group

U, uncommon; R, rare. Major-oxide analyses by rapid methods. Analysts: P. L. D. Elmore, P. W. Scott, S. D. Botts, and K. E. White (1956), except columns 7, 8, 12, by semiquantitative spectrographic methods by P. R. Barnett (1956–57). Specimens described on page 48–51.]

Metavolcanic rocks—Continued										Metasedimentary rocks			
Squaw Creek Schist					Lightning Creek Schist	Fiddle Creek Schist	Lightning Creek Schist			Squaw Creek Schist			
10	11	12	13	14	15	16	17	18	19	20	21	22	23
148425	148410	151581	148397	148419	148428	148411	148403	148404	148415	151583	151586	151587	151594
C 697	C 682	D 1395	C 669	C 691	C 700	C 683	C 675	C 676	C 687	D 1397	D 1400	D 1401	D 1408
SR 424-3	SR 164-3	SR 69	SR 13A	SR 353	SR 442B-1	SR 206-1	SR 47B	SR 48A	SR 303-3	SR 168-2	SR 462A	SR 522-4	SR 587B-2
(weight)													
53.5	54.8	56.7	60.4	61.9	62.9	63.1	65.3	67.0	72.7	44.0	52.4	52.7	55.1
18.6	20.0	17.0	15.9	17.2	15.5	17.4	15.8	14.7	13.4	22.5	16.4	9.1	14.2
1.4	1.8	.8	.8	1.6	2.4	1.9	2.7	2.9	1.1	1.7	1.2	2.5	2.2
}6.8	}7.2	}6.2	}4.7	}5.4	}8.0	}3.3	}5.2	}5.2	}3.5	}8.0	}5.5	}2.6	}8.4
5.5	5.6	5.5	4.0	4.0	5.8	1.6	2.8	2.6	2.5	6.5	4.4	.29	6.4
4.0	4.5	3.1	2.4	2.9	4.1	3.0	2.7	2.3	1.7	4.6	2.9	1.1	4.2
5.8	4.5	7.6	3.9	4.6	2.2	1.5	1.9	2.4	1.1	8.7	15.7	13.2	12.8
4.0	4.1	4.0	3.1	3.1	3.7	3.4	5.3	4.0	4.9	2.6	3.0	1.4	3.0
2.8	2.3	.92	4.7	2.1	1.6	3.5	.65	1.5	1.8	3.3	.77	1.4	.38
.78	.59	.76	.38	.63	.92	.56	.49	.77	.57	1.0	1.2	.16	1.0
.14	.14	.16	.16	.10	.13	.09	.18	.17	.09	.50	.49	.27	.28
.11	.09	.14	.05	.05	.25	.10	.04	.04	.09	.06	.14	.24	.10
1.4	1.0	.36	1.0	.66	1.1	1.9	1.8	1.4	.57	1.3	.75	1.3	.62
1.3	<.05	1.8	3.3	.13	>.05	1.8	.12	.42	<.05	<.05	.64	10.4	<.05
99	99	99	100	99	101	100	100	100	100	97	100	94	100
(weight)													
<0.001	<0.001	<0.001	<0.001	0.007	<0.001	0.0015	<0.001	0.0015	<0.001	0.0015	<0.001	0.015	<0.001
.3	.07	.03	.15	.15	.03	.07	.015	.03	.07	.3	.03	.07	.03
<.0001	<.0001	<.0001	.00015	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001	<.0001
<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01	<.01
.003	.003	.003	.015	.003	.0015	.0007	.003	.0015	.0015	.003	.0007	.0007	.0015
.007	.007	.003	.007	.007	.003	.0015	.0003	.003	.0003	.007	.0007	.003	.003
.007	.007	.007	.01	.007	.0015	.0007	.0015	.003	.003	.007	.0007	.007	.015
.003	.003	.0007	.0015	.0015	.0015	.0015	.0007	.0007	.0015	.0015	.0015	.0003	.0007
<.002	<.002	<.002	<.002	<.002	<.002	<.002	<.002	<.002	<.002	.003	<.002	<.002	<.002
<.0005	<.0005	<.0005	.0015	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	.0015	<.0005
<.0005	<.0005	<.0005	.0007	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	<.0005	.0015	<.0005	<.0005
<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005	<.005
.003	.003	.0015	.007	.003	.0007	.0003	.0007	.0015	<.0002	.003	.0007	.007	.003
.0007	.0007	.0007	.0015	.0007	.0007	<.0005	<.0005	<.0005	<.0005	.0015	.0007	.0007	<.0005
.003	.003	.003	.003	.003	.003	.0015	.0015	.003	.003	.007	.0015	.0015	.003
.07	.03	.07	.03	.03	.007	.015	.03	.015	.015	.15	.3	.03	.15
.03	.03	.015	.03	.03	.015	.007	.015	.03	.007	.03	.015	.015	.015
.003	.003	.0015	.003	.003	.007	.007	.0015	.003	.007	.003	.003	.0015	.003
.0003	.0003	.0003	.0003	.0003	.0007	.0007	.003	.0007	.0007	.003	.0003	.0003	.0003
.015	.015	.007	.015	.015	.015	.03	.007	.015	.015	.015	.007	.007	.007
(volume)													
15	9	14	A	22	23	39	24	30	31	≈1	9	55	12
		P				14	50	42				1.8	
53	50	50	C		52				54				
				49						34	22		51
										12			
.5		.3		.7						9	21		7
							5	3					
.3			A			25	.2	2.3	P(tr.)	.9		16	
20	29	9	U	21	16	7	3	4	14	42		.4	.1
3	.1	.9	C	.1	4	4	16	9		.6		1.4	
2.5	8	18		5									22
											19		7
											4		
	3	.4		.2	4				P(tr.)	.4			
4		6	C	.1		11		7			23	17	.1
.8	.4	.2		.1		.2			P(tr.)	.3	1.1	.5	1.1
.8	.6	1.2	U	2.2	.5	.8	1.6	2.0	.1	.5	.5	8	.4
	.1	P		P			.3	.2	.2	.1			.1
		P	R	P		P			P	P			

Modes.—The mineral composition, in volume percent, or mode, is given for each analyzed specimen. The nodes are based on point counts and micrometer-ocular counts in sections cut perpendicular to the foliation. Thin sections of the layered schists do not reflect precisely the analyzed specimen, although in all instances the section and analysis were made from the same hand specimen. Modes were calculated to 0.1 percent but all values greater than 3 percent were rounded off to the nearest whole percent. Although modes are customarily presented to 0.1 percent for all minerals, their accuracy and reproducibility are much too low to justify such pseudoprecision (Hamilton, 1952).

A full description and the mode of each specimen are presented. A note urging this as general practice was presented elsewhere (Hamilton, 1958).

For most of the specimens, calculations based on assumed compositions of the actual minerals were made so that analyses and modes could be compared. Marked discrepancies, or significant conclusions, indicated by the comparisons are noted below.

Analyses from eastern Oregon.—The spilitic rocks of Permian age of the Baker and Pine quadrangles in Oregon, 50 miles west-southwest of the Riggins quadrangle, are represented by four analyses (Gilluly, 1935, p. 235). These rocks are part of the same petrographic province as the Seven Devils Volcanics, and the analyses are listed in table 1 and are plotted in many of the figures here.

DESCRIPTIONS OF ANALYZED SPECIMENS

1. Muscovite-ankerite(?)-prochlorite-quartz phyllite.—Medium-gray crinkled phyllite with metallic luster. Anastamosing folia of chlorite and muscovite separate lenses of quartz and augen (to 1 mm long) of ankerite(?) (fig. 26). The quartz is slightly elongate, unstrained, and mostly ≈0.1 mm in size. Prochlorite is in 0.1 to 0.2 mm flakes and is aluminian. Muscovite is very pale greenish yellow. The ankerite(?) is weakly pleochroic in brown, very slightly effervescent in cold HCl, has an n_e near 1.53, and encloses much finely granular magnetite(?). Magnetite(?) dust is also abundant in the micaceous folia. The crinkling is due to a slip cleavage by which the original foliation has been crossed at an acute angle; no mineralogic retrogression is apparent. Squaw Creek Schist, east side of the Little Salmon River, one-third mile north of Sheep Creek.

The composition of this rock, recalculated without CO_2, is presented below, together with the original analysis and another similar analysis for comparison:

	1	2	3
SiO_2	44.1	50.7	52.61
Al_2O_3	12.9	14.8	14.64
Fe_2O_3	2.0	2.3	1.36
FeO	6.5	7.5	6.75
MgO	9.7	10.2	7.20
CaO	7.6	8.7	7.30
Na_2O	.23	.3	.82
K_2O	1.6	1.8	.95
TiO_2	.38	.4	.50
P_2O_5	.14	.2	.30
MnO	.18	.2	.30
H_2O	3.8	2.0	2.70
CO_2	10.5	0	3.40

1. Analysis of muscovite-carbonate-quartz-chlorite schist from Squaw Creek Schist (table 5, No. 1).
2. Same analysis, recalculated without CO_2 and with an arbitrary 2 percent H_2O.
3. Kersantite, Vosges, France, from Osann (1923, p. 328).

The composition of this rock is most unusual, even when recalculated without CO_2, particularly in its very low content of Na_2O. The composition seems more like that of an igneous rock than that of a sediment, but none of the analyses in Professional Paper 99 (Washington, 1917) bears any particular resemblance to it. An origin as a basaltic or andesitic rock from which nearly all the sodium has been removed by some metamorphic process is perhaps reasonable. A single analysis from Osann (1923) is very similar to the CO_2-free analysis and is included above; this rock is not described, but is called a kersantite (biotite-plagioclase lamprophyre), a usage of dubious validity. Perhaps the composition of the schist from the Riggins quadrangle is that of an odd lamprophyre.

2. Hornblende greenschist.—Abundant well-alined 5-mm hornblende crystals in a fine-grained (0.05 mm) crudely schistose matrix of oligoclase (≈An_{20}, untwinned, faint reverse zoning), epidote, aluminian prochlorite, and calcite. Prochlorite is only faintly pleochloric in pale greenish yellow. Hornblende is light colored. Many magnetite anhedra are rimmed by sphene. Composition is that of a partially spilitized basalt. Lightning Creek Schist in roadcut by east side of the Salmon River 0.5 mile south of Lightning Creek.

3. Epidote - prochlorite - albite greenschist. — Fine - grained greenish-gray schist spotted by rusted cubes of pyrite. Intergrown hash of small (most near 0.02 mm) grains of albite and ferroan prochlorite, less epidote, and still less muscovite, biotite, actinolite, carbonate, sphene, leucoxene, magnetite, and pyrite. Epidote and actinolite are nearly colorless. Origin as lapilli tuff is indicated by thick lenses and irregular masses of diverse textures and compositions. Composition is that of spilite. Lightning Creek Schist, altitude 2,850 feet, Shingle Creek.

4. Greenschist.—A fine-grained (≈0.05 mm) grayish-green schist uncommonly rich in well-alined actinolite. The actinolite is green in hand specimen and weakly pleochroic yellow green; the rock composition indicates that it must be aluminous, hence hornblendic. Albite, magnesian prochlorite, and aluminian epidote occur with the actinolite. Of the few original phenocrysts of plagioclase, only one remains recognizable, the others having been reduced to lenticular aggregates of albite blebs (fig. 15). Quartz is concentrated in a few stubby lenses, and calcite forms thin veins parallel to the strong cleavage. Dirty sphene forms tiny rounded masses. The sparse opaque mineral is magnetite(?). Rock composition is that of a partially spilitized basalt. Lightning Creek Schist, east side of the Little Salmon River, 1.1 miles north of Sheep Creek.

5. Greenschist.—Fine-grained grayish-green schist in which anastamosing folia rich in ferroan prochlorite (0.2-mm flakes) separate lenses of fine-grained (0.05 mm) albite and glomeroporhyroblastic quartz. A few large subhedra of albite are probably relict phenocrysts. Granules and small (0.05 mm) domes of pleochroic epidote are strewn throughout both folia and lenses. Anhedra of calcite are segregated with quartz. Opaque material is about equally magnetite-ilmenite and pyrite. Rock was probably a partially spilitized basalt. Illustrated by figure 12. Lightning Creek Schist, west side of the Little Salmon River, in bluffs south of the Rapid River.

6. Greenschist.—Abundant small white augen speckle the matrix of greenish-gray schist. Augen are of granular (0.05 mm) albite that encloses minor aluminian prochlorite in tiny flakes. Abundant hornblende occurs as needles, prisms, and anhedra to 1 mm long, intergrown in all directions and concentrated in irregular masses; it is black in hand specimen but only weakly pleochroic (pale greenish yellow to pale yellowish green) in thin section. Epidote is aluminian—colorless, and of only moderate birefringence. Carbonate is iron stained. Minor rutile forms cloudy granular masses. Magnetite dust, distributed very irregularly, makes parts of slide nearly opaque. Origin was perhaps as a porphyritic basalt, partially spilitized. Lightning Creek Schist, west side of the Little Salmon River, 0.8 mile south of Sheep Creek.

7. Carbon-garnet-epidote-quartz-hornblende-albite schist.—Intercalated with ultramafic rocks near Goff Bridge; described on page 65.

8. Amphibolite.—Laminae 0.5 to 5 mm thick and rich in green hornblende alternate with others rich in oligoclase (fig. 22). Hornblende (grayish green, $2V \approx 90°$) is in subhedral needles 0.5 mm long and larger poikilitic stubby anhedra. Plagioclase ($\approx An_{25}$) forms a mosaic of equant anhedra, most about 0.2 mm in diameter; most is untwinned; reverse zoning is common. Minor aluminian prochlorite is intergrown with hornblende and seems to have formed in equilibrium with it, but the biotite formed by alteration of hornblende along cleavages. Magnetite occurs as unusual networks of anhedra. Colorless spinel and either zircon or monazite are present. Composition is basaltic. Berg Creek Amphibolite, north side Salmon River.

9. Greenschist.—Fine-grained grayish-green schist spotted by rhombs of rust-stained calcite and small light streaks. Composed of small lenses of varying texture and mineral proportions, apparently relicts of small clasts in tuff that consist mostly of minutely granular albite and small flakes of ferroan prochlorite. Tiny flakes of muscovite are abundant in many lenses and lacking elsewhere. Quartz is much coarser than the minutely granular albite and is present mostly in monomineralic lenses. Porphyroblasts of biotite and intergrown prochlorite, calcite, garnet, and subhedral ilmenite-magnetite all cut across the foliation (fig. 18). Rock composition is that of a keratophyre. Lightning Creek Schist in roadcut by east side of the Salmon River 0.1 mile north of Chair Creek.

10. Quartz-biotite-oligoclase schist.—Medium-light-gray schist veined by concordant lenses of quartz and subordinate calcite. The schist consists of a mosaic of slightly elongate 0.1- to 0.5-mm crystals of oligoclase (An_{20-30}, some reverse zoned; albite twins abundant, pericline common) strewn by and interlaminated with 0.5- to 1-mm flakes of orange biotite and subordinate aluminian prochlorite. Hornblende occurs as extremely poikilitic 1-mm anhedra. The tiny grains of clinozoisite and the ragged flakes of muscovite formed relatively late. Pyrite occurs as irregular lenses. Squaw Creek Schist, east side of the Little Salmon River, 0.2 mile south of Denny Creek.

11. Garnet-hornblende-quartz-biotite-oligoclase schist.—Medium-gray schist composed of a mosaic of 0.1-mm oligoclase and quartz in which are strewn larger biotite flakes and hornblende crystals that have a strong alinement but weak foliation. Oligoclase is mostly untwinned (in the twinned grains, pericline twins are more common than albite twins), and many crystals show reverse zoning, with albite cores in otherwise uniform oligoclase crystals. Hornblende is black in hand specimen but only weakly colored (grayish yellow green) in thin section and occurs as subhedral prisms as much as 3 mm long. Biotite is ragged and pleochroic yellowish orange. Garnet, in 1- to 2-mm poikilitic subhedra, is gray in hand specimen (hence very inconspicuous) but light brown in thin section. Sparse aluminian prochlorite seems to have formed in equilibrium with the other minerals. Squaw Creek Schist, south side of the Salmon River, 0.8 mile southeast of mouth of the Little Salmon River.

12. Garnet-biotite-quartz-oligoclase schist.—Uniform medium-gray rock of weak schistosity, spotted obscurely by small pink garnets and spangled by small porphyroblasts of biotite and hornblende. Plagioclase (mostly untwinned oligoclase, $\approx An_{20}$; many grains show reverse zoning, or even oscillatory-reverse zoning, with albite cores) and quartz form a mosaic of near-equant 0.02- to 0.2-mm grains. Biotite is in flakes and is pleochroic light brown. Hornblende is anhedral, poikilitic, and pleochroic light grayish green. Calcite is anhedral and poikilitic. Minor aluminian prochlorite, granular clinozoisite, and rounded masses of dirty rutile are present. Opaque minerals are pyrite, ilmenite, and magnetite. Sparse carbon(?) dust indicates an origin as sediment, but bulk composition is that of quartz keratophyre. Squaw Creek Schist, altitude 1,900 feet, Squaw Creek.

13. Polymetamorphic schist.—Dark-gray submetallic schist marked by faint compositional layering. Fabric is now dominated by a strong slip cleavage at an angle to the original foliation (fig. 28), and the intersection of the two sets of planes defines the moderately developed crinkle lineation. Magnetite-dusted folia of fine-grained muscovite and a little biotite separate small augen of carbonate (ankerite?) and poorly defined lenses of fine-grained quartz and oligoclase (untwinned, unzoned). At least the muscovite recrystallized during the formation of the slip cleavage. Porphyroblasts of aluminian prochlorite are printed across the slip-cleavage fabric, and the opaque dust of the cleavage folia continues through the crystals without interruption. Conspicuous pyrite is in anhedra elongate parallel to the old cleavage. Composition is that of a rather mafic quartz latite, but calculations based on the minerals present suggest that the analytical values of both K_2O and Na_2O may be erroneously high. Squaw Creek Schist in roadcut by the Salmon River about 200 yards south of Race Creek.

14. Hornblende-biotite-quartz-andesine schist.—Alternating laminae 2 to 20 mm thick consist of fine-grained light-gray and medium-gray schist studded by 0.5- to 1-mm anhedral porphyroblasts of biotite and actinolitic hornblende. Layers differ in mineral proportions, and color changes are due largely to variations in granular carbon; carbon-dusted folia continue through porphyroblasts of biotite and actinolitic hornblende without interruption. Some layers contain only biotite as their mafic mineral; some contain mostly hornblende; and others have subequal amounts of each. Quartz-andesine ratios also vary widely. Biotite is moderate (orange) brown; hornblende is poikilitic, ragged, and grayish yellow green (black in hand specimen). Andesine is uncommonly twinned or reverse zoned. Magnetite forms irregular lenses. The small euhedral domes of clino(?)zoisite are mostly in, or adjacent to, biotite porphyroblasts. Squaw Creek Schist, Hailey Creek, altitude 2,200 feet.

15. Garnet-biotite-quartz-oligoclase metaagglomerate.—Crudely defined light-colored lenses (sheared pebbles) lie in a darker and more schistose matrix that is richer in biotite and prochlorite. Folia are swirled and discontinuous, so that the rock lacks a consistent foliation direction. The rock is mostly a mosaic of 1- to 2-mm quartz (little stained) and oligoclase ($\approx An_{25}$, unzoned, mostly untwinned), with biotite (1- to 2-mm, anhedral) strewn about and forming knots and laminae. Clasts are finer grained quartz and oligoclase and are lightly speckled by fine biotite. Garnet is moderate red in hand specimen and occurs as 1- to 4-mm poikilitic crystals surrounded by eyes of white quartz and oligoclase devoid of dark minerals. The opaque mineral occurs mostly as small plates and appears to be ilmenite. In irregular patches, biotite has been largely altered to ferroan prochlorite plus quartz. Both biotite and prochlorite have been much bent locally, although retrogression is not apparent in the other minerals. As oligoclase was stable during this retrogression, the conditions of that late metamorphism must have been those of moderate grade rather than low grade. Composition indicates that the original rock was a quartz keratophyre. Lightning Creek Schist, south side of the Salmon River 200 yards west of Lake Creek.

16. White phyllitic metatuff.—White phyllite composed of fine-grained weakly schistose quartz, muscovite, and unzoned albite in varying proportions in augen and layers. Porphyroblasts of ankerite are subhedral, commonly bent (fig. 42), and rust stained by weathering. Augen of dark minerals, 1 to 15 mm long, are composed of much coarser little oriented ferroan prochlorite and olive-brown biotite; interlamination fabrics and stringers of sphene and granules of rutile in the prochlorite indicate that much of the prochlorite formed from biotite. Fiddle Creek Schist, altitude 2,400 feet, Fiddle Creek trail.

17. Greenschist metatuff.—Greenish-gray rock in which thin folia of biotite and chlorite separate strongly flattened 0.5- by 2-cm ellipsoidal light-colored clasts. The clasts contain many albite pseudomorphs after plagioclase phenocrysts, laths as much as 2 mm long in which twinning is still visible but which have been much replaced marginally. Clasts are mostly mosaics of fine-grained quartz and albite, with subordinate epidote (tiny domes), aluminian prochlorite, and magnetite. Biotite is pleochroic from grayish yellow to moderate olive brown. Rock was a quartz keratophyre tuff. Lightning Creek Schist, east side of the Salmon River a short distance north of mouth of Lightning Creek.

18. Meta-agglomerate.—Greatly deformed greenish-gray clasts, clasts, now flat triaxial ellipsoids, separated by darker greenish gray biotitic folia (fig 16). Clasts are granular aggregates of quartz and albite, with disseminated flakelets of aluminian prochlorite, variable amounts of ragged iron-stained carbonate, and locally epidote, biotite, and magnetite. Quartz-albite grain size varies irregularly from 0.03 to 0.2 mm. A few larger albite crystals with ragged outlines are probably relict phenocrysts. Folia are schistose aggregates of prochlorite, muscovite, and poorly oriented porphyroblasts (equant anhedra) of biotite intergrown with each other and with the subordinate other minerals. Opaques are andehra of magnetite and small plates of ilmenite. The thin section contains much less muscovite and much more carbonate than is indicated by the analysis. The bulk composition is intermediate between dacite and quartz keratophyre. Lightning Creek Schist in roadcut by east side of the Salmon River 150 yards north of Lightning Creek.

19. Biotite-quartz-oligoclase schist.—Fine-grained light-gray schist spangled with thin lenses of coarser biotite and plagioclase (fig. 15). Groundmass is of nearly equant 0.1- to 0.2-mm oligoclase (untwinned, unzoned), quartz, and biotite (moderate orange brown). Grain size in the lenses is 0.5 mm, and much of the oligoclase (An_{20}) in them has lamellar albite or pericline twins. Lightning Creek Schist, west side Lake Creek, one-third of a mile south of the Salmon River.

20. Garnet-clinozoisite-zoisite-andesine-biotite schist.—Black lenses of nearly pure 0.5-mm biotite, light-gray streaky lenses of fine-grained plagioclase and clinozoisite, and dark-gray schist with abundant opaque dust, all studded by domes of zoisite as much as 5 mm long, give this rock a striking appearance. The outstanding characteristic is the abundant large euhedral domes of zoisite (fig. 38). Most of these are well alined, although many are skew to the lineation. Andesine occurs as much elongate grains, particularly in folia rich in clinozoisite, and as larger equant granules in relatively pure lenses. The equant granules commonly show gradational reverse zoning. Many grains show a combined reverse-and-normal-zoning wherein cores and rims are more sodic than the intermediate part; a few grains show several compositional oscillations each within such complex zones. Pericline twins are more numerous than albite twins, although neither type is common. The biotite is pleochroic light brown and occurs variously in, and oriented parallel to, folia, as lenses of randomly oriented pure biotite, and as isolated porphyroblasts maintaining a consistent 50° angle to the main foliation. The clinozoisite forms small euhedral domes $\approx$0.1 mm long that are well alined along the main lineation. Sparse porphyroblasts of garnet enclose many small domes of clinozoisite. Aluminian prochlorite forms poorly oriented porphyroblastic sheaves, against which zoisite and clinozoisite are euhedral; an origin in late selective replacement of biotite and plagioclase is consistent with the fabric. This, the apparent superimposed normal zoning of the plagioclase on an older reverse zoning and the skew orientation of much of the biotite both suggest considerable reconstitution under conditions of middle-grade metamorphism after the original high-grade metamorphism. Sphene and rutile occur together in ragged intergrown masses; pyrite forms tiny anhedra; and there is abundant black opaque (magnetite?) dust and granular material throughout the specimen. Squaw Creek Schist, south side of the Salmon River 0.25 mile west of mouth of Berg Creek.

21. Scapolite-quartz-diopside-clinozoisite-andesine-carbonate gneiss.—Odd greenish-gray granulitic gneiss in which the green cast is due to small anhedra of diopside that stud the otherwise light-gray rock. It consists of two main types, crudely interlaminated: clinozoisite-carbonate granulite with diopside, and an intergown hash of scapolite, quartz, diopside, clinozoisite, and plagioclase. The granulite has an average grain size of 0.5 mm and is composed of smoothly anhedral unstained carbonate, mostly noncalcitic; less ragged clinozoisite containing vermicular quartz; and anhedra of diopside. The intergrowth laminae consist of sprays of rods of clinozoisite, (fig. 37) micrographically intergrown with plagioclase, quartz, and carbonate, and interspersed with anhedra of scapolite and diopside. The rock contains a little graphite(?). Plagioclase is mostly An_{30-40}, and much of it is twinned by the albite and pericline laws. Diopside (or hedenbergite) is pale green, nonpleochroic, and has a + 2V of 62°. Most of the clinozoisite is strongly ultrablue, but some is ferroan enough to show second-order colors. Mineral proportions were obviously much different in the fractions analyzed (for example, CO_2=0.64 percent) and examined in thin section (for example, carbonate=23 percent). Analysts reported that the sample contained organic matter, and this might have affected the FeO determination. Squaw Creek Schist in ridge south of Hailey Creek at an altitude of 4,500 feet.

22. *Carbonaceous calcite-muscovite-quartz schist.*—Dark-gray submetallic schist crinkled by microfolds 1 mm apart. Quartz and calcite grains are 0.1 to 0.2 mm long, and is each crudely segregated into microlaminae. Thin flakes of muscovite are bent around the crinkles. Minor biotite (weakly colored; pale yellowish orange) forms small porphyroblasts and tiny flakes that cut across the crinkles; minor prochlorite also forms late porphyroblasts. Tiny granules of carbon make up about 6 percent of the rock; there are also small lenses of pyrite, largely oxidized by weathering. Squaw Creek Schist, from bluffs in north part of Riggins.

23. *Clinozoisite-diopside-hornblende-andesine gneiss.*—Alternate layers 5 to 20 mm thick of light-gray and medium-gray rock are distinguished most obviously by their different hornblende contents. Crystals of most minerals are 0.1 to 1 mm in diameter and show a marked dimensional lineation but little foliation. Plagioclase is mostly An_{30-50}; many grains show smooth normal zoning to An_{20}, and a few have an oscillatory zone; albite and pericline twins are abundant. Hornblende is anhedral, black in hand specimen, green in thin section. Diopside and clinozoisite are colorless, and the clinozoisite has wormy inclusions and in part grades into aluminian epidote. Pyrite forms small irregular lenses. Squaw Creek Schist in canyon north of Captain John Creek at an altitude of 4,100 feet.

MAJOR OXIDES

Compositions of the metavolcanic rocks vary systematically with silica content (fig. 43), and the gross compositional trends common in suites of volcanic rocks from most igneous provinces are recognizable, although compositional changes since the original volcanism have complicated these variations.

The decrease of Al_2O_3 with increase of SiO_2, characteristic of most volcanic provinces, is systematic here only among the silicic rocks. In the mafic metavolcanic rocks there is a wide range in Al_2O_3 content from 13 to 20 percent between 44 and 57 percent SiO_2. Similarly, CaO and the alkalis, which in normal volcanic provinces vary systematically with Al_2O_3, show only a rough relationship (fig. 44). Among the most aberrant, relative to normal volcanic rocks, of the analyzed specimens are three (Nos. 1, 10, and 11) from the Squaw Creek Schist, designated with reservations as of metavolcanic origin; much of the departure may be due to sedimentary processes.

Calculations based on assumed compositions of the minerals actually present in the schists suggest that determinations of Al_2O_3 may be consistently a little too high in the analyses.

In most of the schists of the Riggins Group, FeO is in large excess over Fe_2O_3. The proportion of FeO increases as total iron increases (fig. 45), in a nearly linear relation that corresponds approximately, by weight, to

$$FeO = FeO + 0.9\ Fe_2O_3 - 2.$$

The amount of Fe_2O_3 is thus nearly constant at about 2 ± 1 percent, whereas FeO varies with total iron from

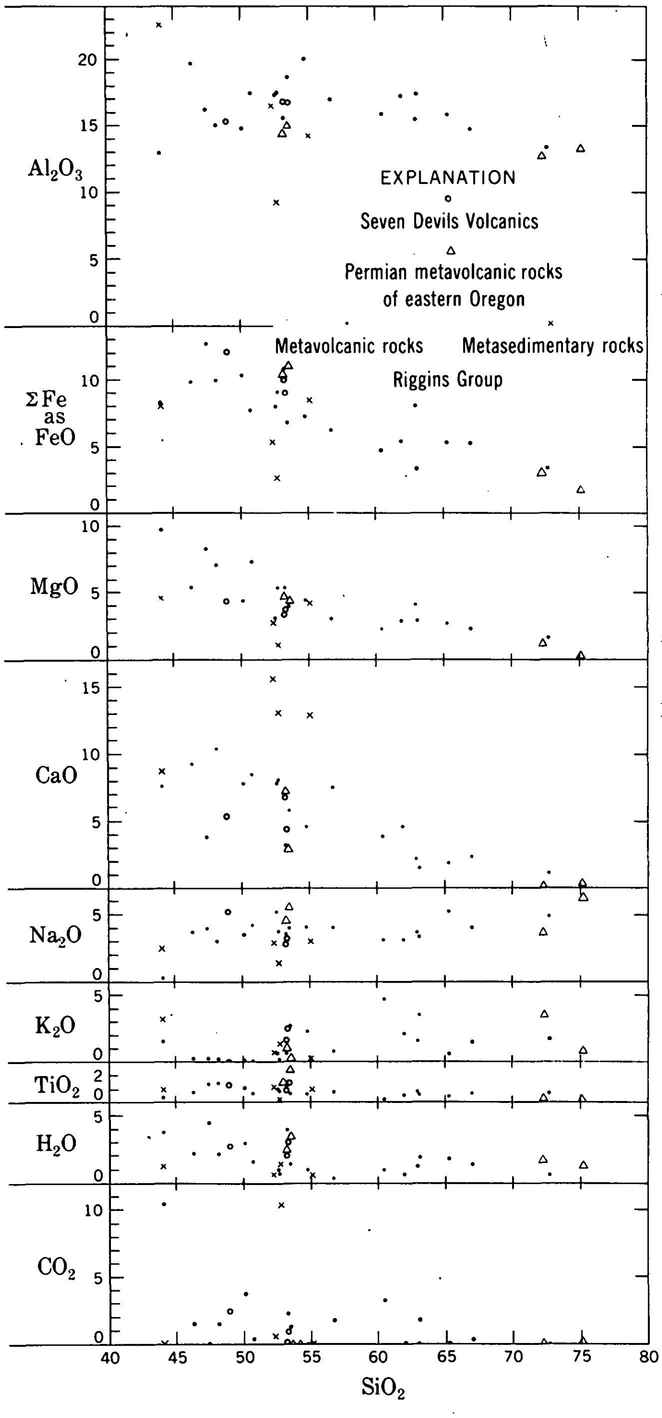

FIGURE 43.—Variation diagram of weight-percent analyses of metamorphic rocks of Riggins quadrangle. These are metavolcanic rocks and closely related metasedimentary rocks of a submarine andesite-keratophyre association. Compare with figure 65. Ultramafic rocks not plotted. Oregon analyses from Gilluly (1937, p. 25).

0.3 to 10 percent. In the lower grade metavolcanic rocks of the Seven Devils Volcanics, by contrast, Fe_2O_3 is in great excess over FeO. The high proportion of Fe_2O_3 in the Seven Devils Volcanics is reflected mineralogically in large amounts of epidote, magnetite, and hematite. With increasing metamorphism, iron was mostly reduced to the ferrous state; in the Riggins Group, epidote and magnetite are less common, and hematite is lacking.

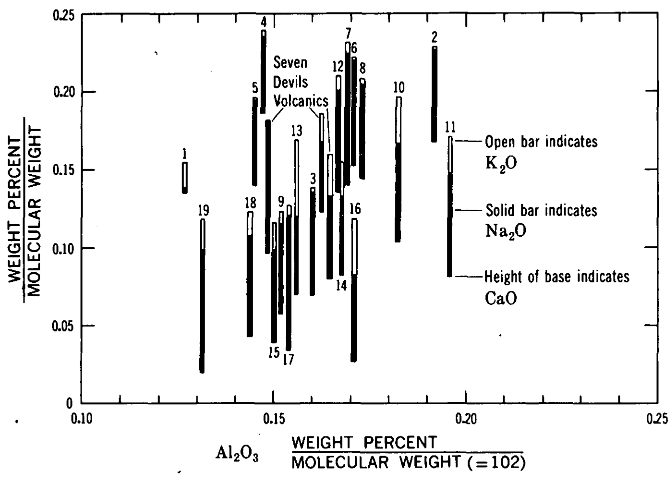

FIGURE 44.—Molecular proportions of Al_2O_3, CaO, Na_2O, and K_2O in metavolcanic rocks of Riggins quadrangle. CaO, Na_2O and K_2O plotted as weight percents divided by molecular weights (56, 62, and 94, respectively). The three specimens of Seven Devils Volcanics are marked; all others are metavolcanic schists of Riggins Group. Numbers indicate columns in table 5 and increase in order of SiO_2 content.

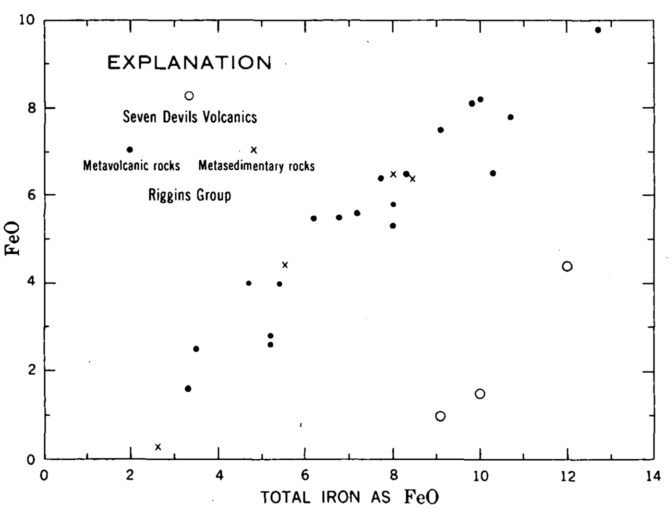

FIGURE 45.—Ratio of FeO to total iron in metamorphic rocks of Riggins quadrangle.

Iron and magnesium increase together (fig. 46). In most of the schists of the Riggins Group, the relation is regular, about $(Fe_2O_3+FeO)=2$ MgO, by weight, for Σ FeO<7, but irregular for larger amounts, and the precise shape and position of the higher part of the curve on the diagram have no significance. The linear 2:1 relation by weight corresponds to a molecular ratio Σ FeO:MgO of 1:0.9. The Σ FeO:MgO ratio in the analyzed specimens of the Seven Devils Volcanics is considerably higher, about 5:2 by weight. As only three analyses were made from this rock group, this difference may be only an illusory one due to sampling inadequacies, but the data do suggest a possible fundamental difference between the two groups of metavolcanic rocks. The analyses of metavolcanic rocks from the Baker and Pine quadrangles, Oregon, show a similar high Σ FeO:MgO ratio.

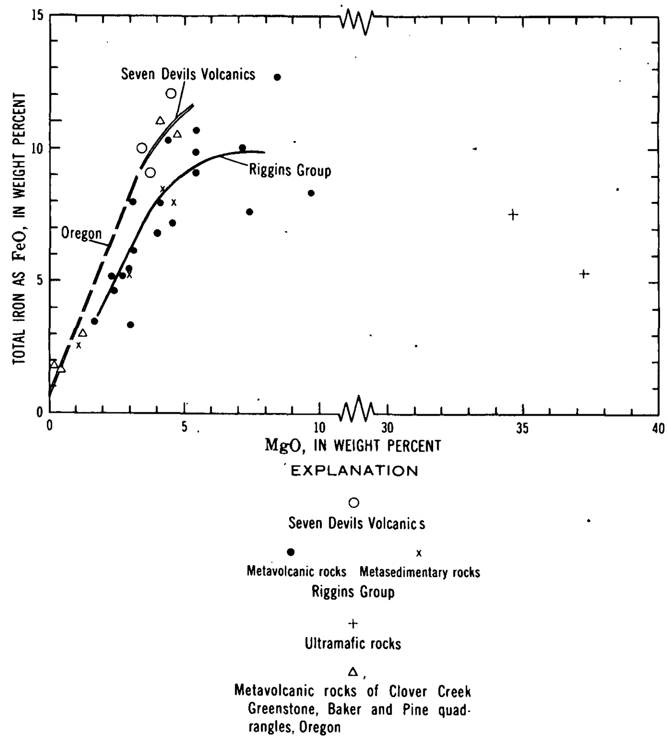

FIGURE 46.—Iron and magnesium contents of metamorphic rocks of Riggins quadrangle. The lines indicate possible variation curves. Oregon analyses from Gilluly (1937, p. 25).

The ratio Na_2O:CaO is highly variable (fig. 47) for each SiO_2 content, and this is in marked contrast to the systematic change of these oxides with silica in nongeosynclinal volcanic rocks. The ratio is, in general, higher than that of other such rocks, as sodium ranges upward from a minimum equivalent to that of the more common rocks, whereas calcium ranges downward from a comparable maximum (compare fig. 43 with figs. 65, 66, and 67). Such distributions characterize the rocks of the andesite-keratophyre association, which includes normal calc-alkaline rocks (basalt, andesite, dacite, ± quartz latite and rhyolite) on the one hand and their sodium-enriched equivalents (spilite, keratophyre, quartz keratophyre, ± potassic quartz keratophyre)

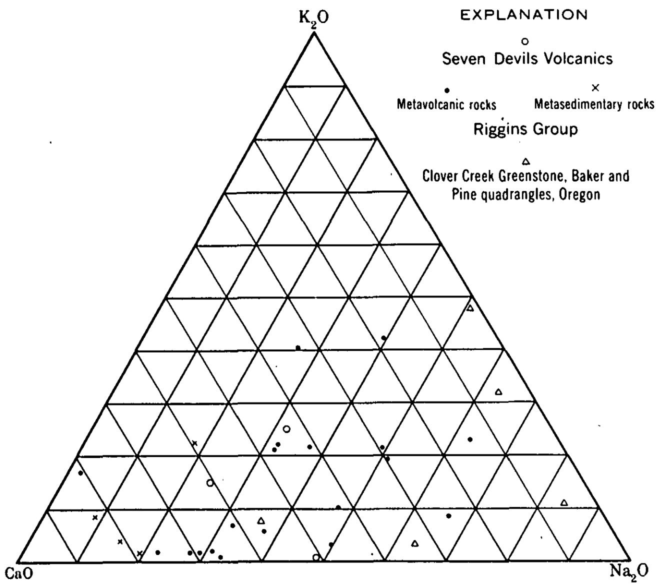

FIGURE 47.—K_2O–CaO–Na_2O ratios, by weight, of metamorphic rocks of Riggins quadrangle. Oregon analyses added from Gilluly (1937, p. 25). In the metavolcanic rocks, $K_2O+CaO+Na_2O \simeq 10$ percent ± 3; in the metasedimentary rocks, $K_2O+CaO+Na_2O \simeq 17$ percent ± 3.

on the other. There are all intermediates between the calc-alkaline and the keratophyric types, and these variations represent all stages in enrichment of sodium and removal of calcium. (It is a common misconception that such volcanic suites are wholly keratophyric.) The volcanic rocks from which the schists of the Riggins Group were formed thus belong to the andesite-keratophyre association. The genesis of this association is the topic of a subsequent section.

Sodium is in moderate to great dominance over potassium in nearly all the schists, including the most silicic (figs. 43 and 47). In only two analyses is there a dominance by weight of K_2O over Na_2O. This reflects the volcanic derivation of the whole assemblage and (considered with the mineralogy of the other specimens) indicates that the common silicic end members of the volcanic suite were dacites and rhyodacites and their quartz-keratophyric equivalents rather than quartz latites or rhyolites.

Potassium is higher in many of the analyses than can be accounted for in the observed micas. Either the sodic plagioclase carries considerable amounts of potassium, or the K_2O determinations are systematically a little too high.

Carbonates make up several percent of most of the analyzed specimens and about 20 percent of several. Presumably there was considerable movement of components through migration of carbonates. Some of the CO_2 represented clearly came from intercalated sediments, and no evidence was recognized to suggest introduction of CO_2 from outside the geosynclinal section.

From the relation between the chemical components of the rocks, several inferences can be made regarding their behavior during processes subsequent to volcanism. Some of the rocks were first subjected to sedimentary processes that must have included selective changes in compositions. Next, many of the rocks were spilitized in widely varying degree. Then, all were metamorphosed, and some components moved about. The following comments refer to the sum effects of these processes.

The near constancy of the iron : magnesium ratio indicates that either those elements were not affected—were not moved about, into, or from the system—by the various processes or that they behaved in quantitatively similar ways during any movements. As the proportions of iron and magnesium in different minerals are dependent upon metamorphic grade, movement should have been differential had it occurred. It seems, therefore, that iron and magnesium were relatively immobile during spilitization and metamorphism.

About half the analyzed specimens of metavolcanic rocks have essentially normal volcanic compositions, but in the other half, the major components—SiO_2, Al_2O_3, CaO, Na_2O, K_2O—vary so complexly and independently that differential movements seem required. The most easily definable change is that of spilitization, whereby calcium was somehow removed from the system and, commonly, sodium was added. The change was not a 1 : 1 exchange, for calcium was removed in irregularly greater molecular proportion than sodium was added. Part of the relative sodium enrichment was caused by removal of calcium, and a variably lesser part by addition of sodium. Presumably both elements were further exchanged and moved about during metamorphism.

Potassium varies so widely between otherwise similar rocks, from near zero to amounts much larger than those commonly present in the probable original types of volcanic rocks, that much migration of it is likely. No evidence was recognized that would suggest removal from, or introduction to, the geosynclinal assemblage. Movement within the system, causing potassium impoverishment at one locality and enrichment at another, is adequate to explain the known data.

The abundant veins of quartz in much of the Riggins Group were probably produced by mobilization of silica within the rocks. Several of the aberrant rock compositions suggest modification by addition of silica, and several others by subtraction of it.

Similarly, complementary changes in alumina contents would explain some of the aberrant compositions, although no direct evidence for alumina mobility was recognized. The scatter of Al_2O_3 contents both against SiO_2 and against CaO and the alkalis is much less systematic and varies within wider limits than in common volcanic rocks.

MINOR ELEMENTS

All specimens analyzed for major oxides were analyzed spectrographically for minor elements as well. The trace elements detected are indicated in table 5 and on figure 48 and suceeding diagrams.

The analyses of minor elements of the plutonic rocks of the Riggins quadrangle, which lie east and south of the metamorphic rocks described in this report are illustrated in figure 48. Comments regarding contrasts between metavolcanic and plutonic rocks are made here for some of the minor elements, as such contrasts may be significant to the understanding of the origin of the plutonic rocks.

The presumptions indicated regarding the mineralogic location of the minor elements are largely those suggested by Goldschmidt (1954). Credit should also be given to the text of Rankama and Sahama (1950) and to the detailed studies of minor elements in several European provinces by Nockolds and Mitchell (1948) and Hügi (1956).

Minor elements occur both within rock-forming minerals and in optically invisible interstitial materials. Trace-element studies have emphasized unmetamorphosed igneous rocks, and relatively little is yet known about the behavior of these elements during metamorphic recrystallization. Many minor elements occur largely within silicate minerals in the normal positions of major elements, whereas others occur chiefly in much more concentrated amounts in minor accessory minerals. The degree to which an element can enter a crystal structure in the place of another varies with ionic size (the effective size of an ion in a crystal structure) and valence, complicated by many factors such as polarization properties, tendency to form complex ions, relative and absolute abundances, affinity for specific elements such as sulfur, flexibility of the individual lattice, coordination number, and so forth. Common associations of major and minor elements include the following, listed in order of increasing ionic sizes (Hügi, 1956):

Ga^{+3}—Al^{+3}	Sc^{+3}—Fe^{+2}
Cr^{+3}—Fe^{+3}	Y^{+3}—Ca^{+2}
Ni^{+2}—Mg^{+2}	Sr^{+2}—Ca^{+2}
Nb^{+5}—Ti^{+4}	Sr^{+2}—K^{+1}
Co^{+2}—Fe^{+2}	Pb^{+2}—K^{+1}
V^{+3}—Fe^{+3}	Ba^{+2}—K^{+1}
Mn^{+2}—Fe^{+2}	

On the following pages, the Riggins data are analyzed for some of these relationships.

The substitution of minor for major elements within crystal structures was deduced by V. M. Goldschmidt to follow two rules. As paraphrased by Nockolds and Mitchell (1948, p. 540), these are:

> If two ions have the same electrostatic charge (of the same sign) and have similar ionic radii such that they can occupy the same position in a given crystal lattice, then the ion of smaller radius will enter the lattice more readily than the ion of larger radius.
>
> If two ions are of similar size but of different electrostatic charge (of the same sign), the ion of higher electrostatic charge will enter a given crystal lattice more readily than the ion of lower electrostatic charge.

Nockolds and Mitchell found these deductions to be largely in accord with their analytical findings.

The theoretical consequences of this behavior are great, most obviously when progressive magmatic crystallization is considered. A minor element will be concentrated in early-solidified phases if it enters the lattice preferentially ahead of the associated major element, but it will be in late-solidified phases if the major element can more easily enter the lattice, and the ratio of the abundance of the minor element to that of the major element will change accordingly. If the major and minor elements enter the lattice with equal ease, their ratio will be constant.

In the graphic presentations of the chemical data here, values are plotted as analyzed in weight-percents of oxides for major elements and in weight-percents of elements for minor elements. This is done to keep the graphs numerically tangible. (It is of course also the easiest procedure.) Many minor-element geochemists plot data hopelessly camouflaged as, for example, "Mg + Fe, g-atoms per ton rock" against $\frac{Cu}{\text{“}Mg+Fe} \times 10^{6}$"; or, for another example, they convert major-oxide analyses to elemental equivalents and plot these against minor-element values multiplied by any convenient factors. Such treatment destroys the comprehensibility of the relationships, changing obvious trends into abstractions that at their worst defy analysis.

Standard statistical procedures—calculation of standard deviations, correlation coefficients, and the like—are not employed here because the resulting numerical parameters are so abstract as to mean little to most geologists. Scatter diagrams, used here, provide the same information in an easily understood form.

BARIUM

The reported contents of barium in the Riggins Group and Seven Devils Volcanics range through a factor of 200, with extreme values of 0.0015 and 0.3 percent. The amount is near 0.003 in most of the metabasalts and near 0.07 in most of the metadacites. In the ultramafic rocks it is near threshold (0.0002). Barium thus increases with feldspar and mica.

Barium and potassium vary directly together. On a log-log plot (fig. 49), the values for Ba and K_2O cluster about a straight line: their abundances bear a simple

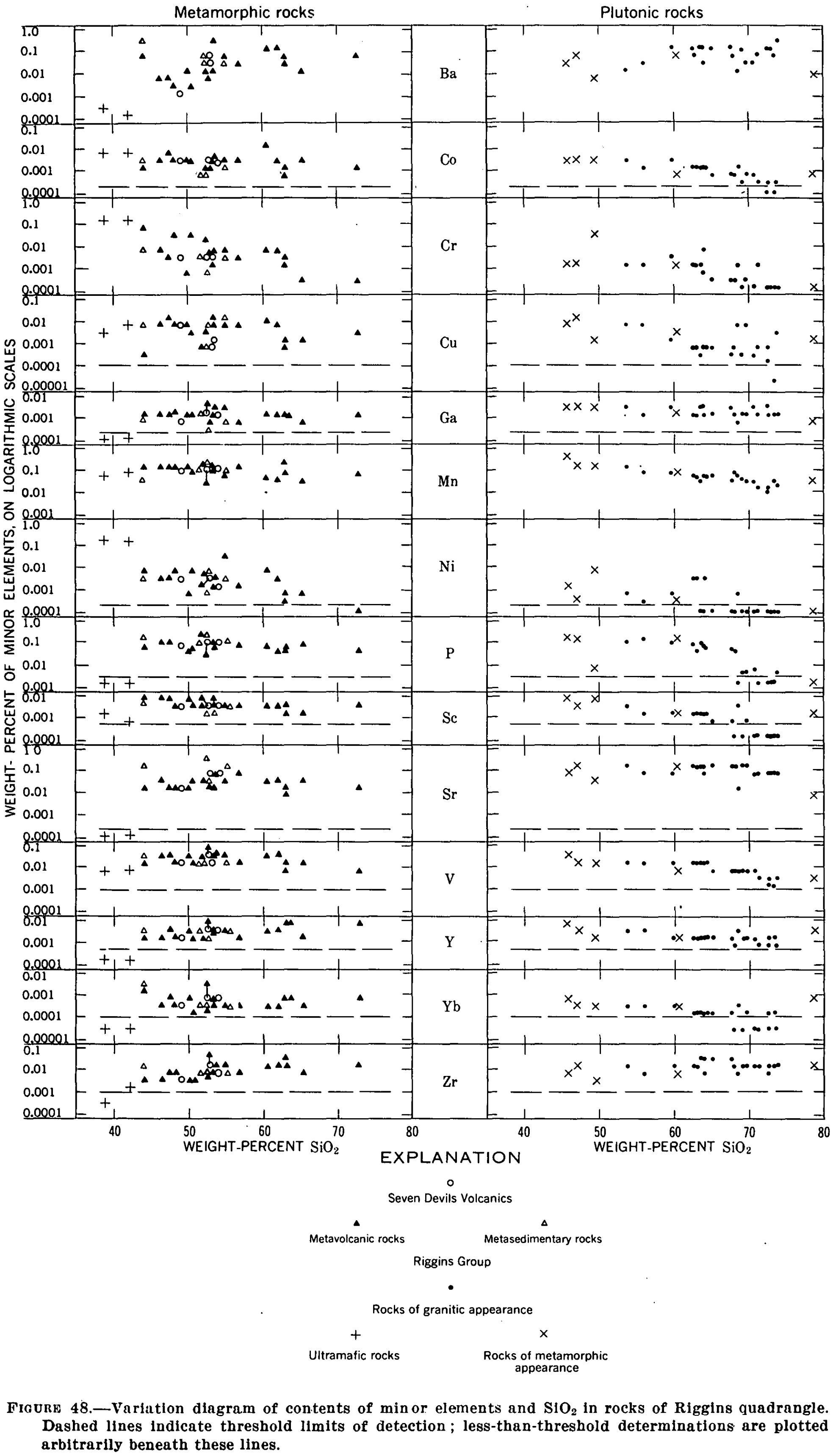

FIGURE 48.—Variation diagram of contents of minor elements and SiO_2 in rocks of Riggins quadrangle. Dashed lines indicate threshold limits of detection; less-than-threshold determinations are plotted arbitrarily beneath these lines.

exponential relation to each other. In weight-percent units, the equation for the straight line on the figure is:

$$K_2O=14(Ba)^{0.8}$$

or,

$$Ba=0.04(K_2O)^{1.3}$$

The weight ratio $Ba:K_2O$ increases from 1:70 at $K_2O=0.04$ to 1:20 at $K_2O=4$; the molecular ratio Ba:K increases from 1:200 to 1:60 within the same limits.

A straight line on a log-log graph has the general equation

$$Y=aX^b$$

where b is the slope and a is the Y intercept. The computation of a specific equation is made from the logarithmic form

$$\log Y=b\log X+\log a$$

In the example above, $a=14$ (the value of K_2O at Ba=1), and

$$b=\frac{\log K_20'-\log K_20''}{\log Ba'-\log Ba''}$$

Determining the coordinates of any two points on the line, this becomes

$$b=\frac{\log 14-\log 0.06}{\log 1-\log 0.001}$$

$$=\frac{\log\frac{14}{0.06}}{\log\frac{1}{0.001}}=\frac{\log 234}{\log 1000}=\frac{2.370}{3}$$

$$=0.79$$

Substituting these values, slightly rounded, the solution is seen as

$$K_2O=14(Ba)^{0.8}$$

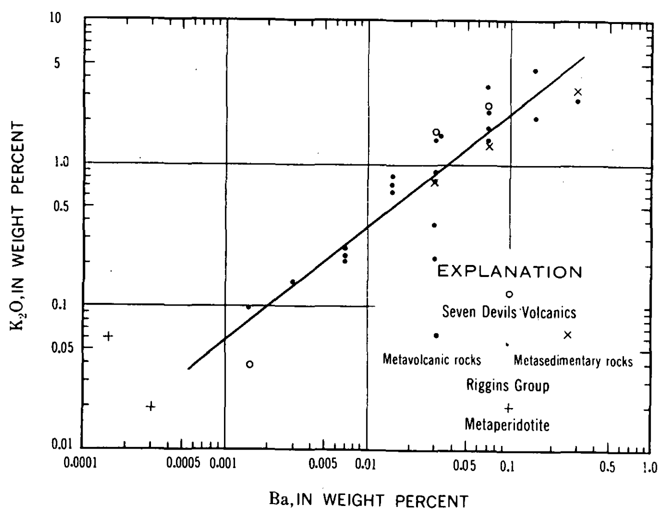

FIGURE 49.—Contents of K_2O and Ba in metamorphic rocks of Riggins quadrangle. The line has the equation ($K_2O = 14(Ba)^{0.8}$.

This relationship between barium and potassium is valid in all the metamorphic rocks plotted, which include metaperidotite and metasedimentary rocks (which must, however, be largely of volcanic derivation) as well as the metavolcanic rocks of both the Seven Devils Volcanics and the Riggins Group. (The same line seems adequate for the plutonic rocks also, although a slightly flatter line could be drawn through their points. The lavas of the Columbia River Basalt within the Riggins quadrangle, by contrast, have a Ba:K ratio about twice as high.)

Divalent barium is presumably largely in potassic minerals in the normal position of potassium ions, and if so, it must be mostly within the micas, the only common potassic minerals in the Riggins Group.

The constancy of the exponential Ba:K ratio in the metavolcanic rocks is consistent with Goldschmidt's deductions, but it is the more remarkable here because the irregular variations of potassium in many of the rocks suggest that element to have migrated considerably during spilitization or metamorphism. Perhaps the simple exponential ratio invalidates the suggestion that potassium has moved since initial volcanism.

BERYLLIUM

The quantity of beryllium in igneous rocks has been found elsewhere to increase with SiO_2 content; but even among the eight analyses of metamorphic rocks here with SiO_2 contents above 60 percent, only one contains beryllium in more-than-threshold (0.0001 percent) amount. The relation thus cannot be tested with the data available in the Riggins quadrangle.

In the plutonic rocks of the quadrangle, by contrast, 12 of the 19 analyzed rocks that have $SiO_2>60$ percent have detectable (0.00015 to 0.0003 percent) beryllium.

BORON

The threshold detectable amount of boron is 0.0015 percent, and only one-third of the specimens of metamorphic rocks contain this much or more. The data suggest that boron is more abundant in the Seven Devils Volcanics and in the metasedimentary rocks of the Riggins Group than in the metavolcanic rocks of the Riggins Group. Rare grains of tourmaline, green in thin section, are visible in some of the specimens from which boron was reported.

Neither of the ultramafic rocks analyzed contains as much as threshold boron. Most of the plutonic rocks also lack detectable boron. Only two specimens—one of quartz diorite gneiss from the eastern part of the Riggins quadrangle, and one of quartz diorite intrusive into the Seven Devils Volcanics—contain detectable boron, and these have the bare minimum, 0.0015 percent.

CERIUM, LANTHANUM, AND NEODYMIUM

The cerium-earth metals cerium, lanthanum, and neodymium are virtually lacking in detectable amounts in the metamorphic rocks. The minimum amount (0.015 percent) of cerium was found in one specimen from the metavolcanic rocks of the Riggins Group, and the minimum (0.003) of lanthanum was found in one metavolcanic and one metasedimentary rock; neodymium was not detected.

In the plutonic rocks, by contrast, these same elements were detected a total of 26 times, in 13 specimens.

CHROMIUM

The content of chromium varies more closely with SiO_2 than with any other major component of the metamorphic rocks (fig. 48) and falls from about 0.05 percent in the least silicic schists (and 0.3 in some of the ultramafic rocks) to 0.0002 percent in the most silicic. The volcanic origin of the schists is reflected in their igneous-type abundances of chromium, cobalt, and nickel. There is no apparent correlation between the abundance of chromium and that of ferric iron. The correlation of chromium with total iron is poor, and that with magnesium is only fair (fig. 50).

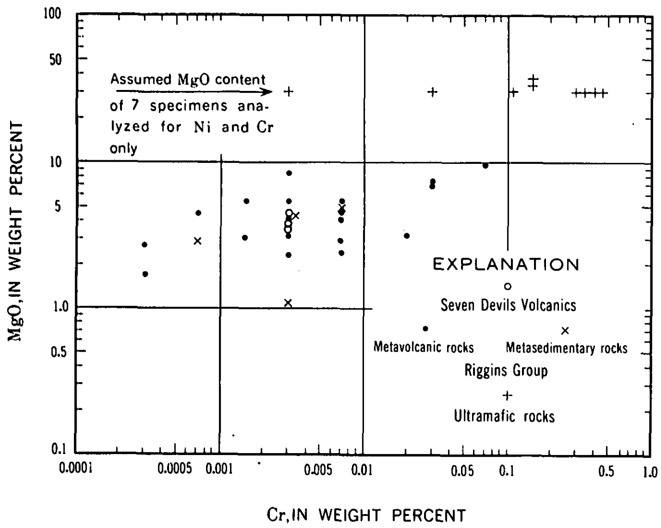

FIGURE 50.—Contents of MgO and Cr in metamorphic rocks of Riggins quadrangle.

COBALT

The abundance of cobalt decreases slightly and irregularly with increasing SiO_2, from near 0.003 (0.007 in the ultramafic rocks) to near 0.001 (fig. 48). The cobalt presumably is present in Fe^{+2} positions in ferromagnesium minerals. As cobalt is less variable than nickel, and much less so than chromium, the Co : Ni and Co : Cr ratios increase (as the absolute amounts of all three elements decrease) with increasing contents of quartz and feldspar.

COPPER

The amount of copper decreases very irregularly with increasing SiO_2, from near 0.01 percent in the mafic rocks (less in the ultramafic rocks) to near 0.002 in the silicic ones. The copper is presumably largely in sulfides, notably chalcopyrite.

GALLIUM

The content of gallium in the schists of the Riggins Group varies little, being near 0.002 weight percent in all but the ultramafic rocks, in which it is below threshold (0.0002 percent) amount. As gallium and aluminum generally vary in close relationship, this near constancy is presumably a reflection of the similarly little-varied content of aluminum in the schists.

LEAD

About half the specimens from the Riggins Group, both metavolcanic and metasedimentary, and from the Seven Devils Volcanics contain detectable (0.0007 to 0.003 percent) lead, which is lacking in the ultramafic rocks. Much of the lead is presumably in potassium positions in mica. The correlation between potassium and lead is poor in the metamorphic rocks (fig. 51),

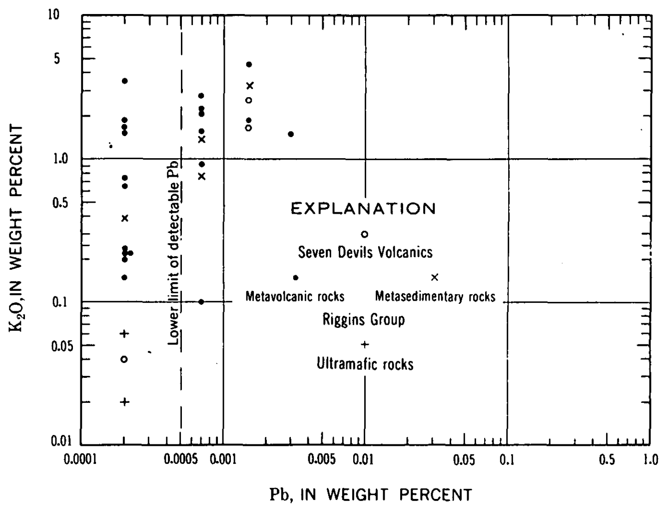

FIGURE 51.—Contents of K_2O and Pb in metamorphic rocks of Riggins quadrangle. Subthreshold values of lead are plotted arbitrarily at 0.0002 percent.

in contrast to the strong correlation in the plutonic rocks, but the ratios in these two major rock groups appear similar.

MANGANESE

The manganese content of the schists is surprisingly constant, near 0.1 percent in rocks of all compositions; there is only a slight decrease with decreasing content of ferromagnesian minerals (fig. 52). (This is in contrast to the plutonic rocks, in which manganese varies markedly with iron and magnesium.) The ratio

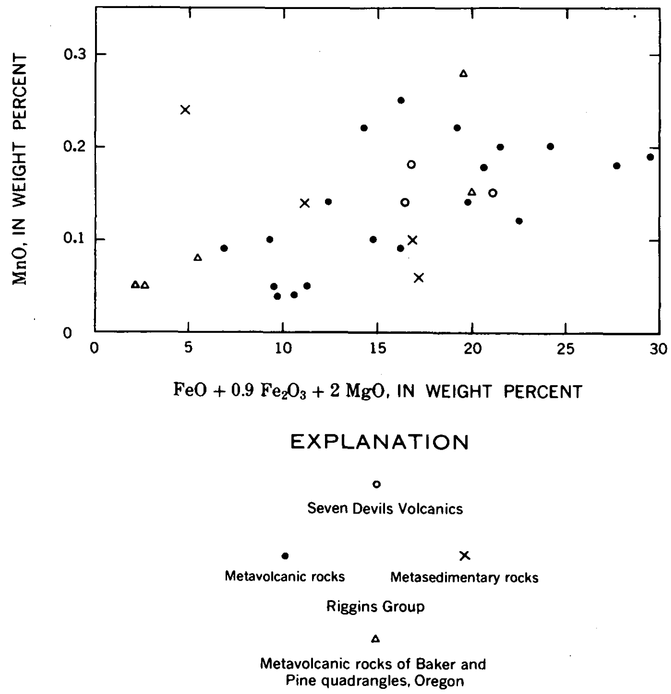

FIGURE 52.—Contents of manganese, iron, and magnesium in metamorphic rocks of Riggins quadrangle. The abscissa was chosen because in these rocks total iron as FeO is approximately equal to twice the weight percent of MgO. Ultramafic rocks not plotted. Oregon analyses from Gilluly (1937, p. 25).

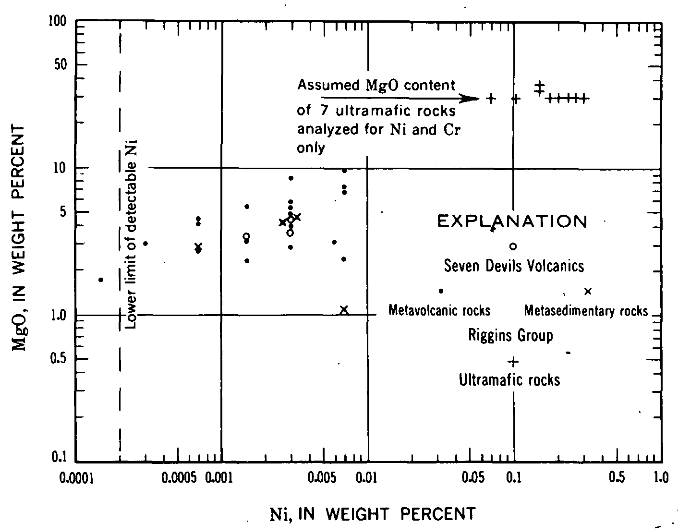

FIGURE 53.—Contents of MgO and Ni in metamorphic rocks of Riggins quadrangle. Subthreshold values of Ni are plotted arbitrarily at 0.0002 percent.

Mn: (Mg+Fe) thus increases greatly as the total iron and magnesium decrease.

In rocks in which iron is largely oxidized to the Fe^{+3} state, the manganese is presumably mostly in the corresponding Mn^{+3} form. Such oxidation is conspicuous in the Seven Devils Volcanics but not in the Riggins Group.

MOLYBDENUM

Above-threshold (0.0005 percent) amounts of molybdenum were found in only 5 of the 20 analyzed metavolcanic rocks of the Riggins Group (0.0005 to 0.0015 percent), 1 metasediment (0.0015 percent), and 2 of the 3 Seven Devils Volcanics specimens (all 0.0007). Maximum SiO_2 content of these molybdenum-bearing specimens is 60 percent. The element was detected in only two specimens of the plutonic rocks—one an amphibolite, the other a quartz monzonite.

NICKEL

The amount of nickel is higher in the mafic rocks (about 0.005 percent) than in the most felsic rocks (less than 0.0003 percent). The ultramafic rocks contain as much as 0.3 percent nickel. The correlation between nickel and magnesium in the nonultramafic metamorphic rocks (fig. 53) is poor. No meaningful correlation line could be drawn without including the ultramafic-rock data; and as the ultramafic rocks may be unrelated in origin to the others, this is not done.

NIOBIUM

Of the metamorphic specimens, only 2 of metavolcanic rocks and 1 of a metasediment, all from the Riggins Group, contain above-threshold (0.0007 to 0.0015 percent) niobium. By contrast, half the plutonic specimens contain detectable niobium.

PHOSPHORUS

All but one of the schist specimens contain about 0.05 to 0.1 percent phosphorus. The content is markedly different (less than 0.005 percent) only in the ultramafic rocks. This little-varying distribution in the schists is in contrast to the distribution in the plutonic rocks, in which phosphorus shows the common inverse variability with SiO_2 (fig. 48), and perhaps the seeming constancy is an illusory reflection of inadequate sampling of the most silicic schists. The phosphorus is presumably largely in apatite.

SCANDIUM

Another little-varied component of the schists is scandium, near 0.005 percent in all specimens; it is a little lower in the ultramafic rocks. This is again in marked contrast to the plutonic rocks, in which the expected correlation between Sc and Fe+Mg is clearly shown. The scandium is presumably in ferromagnesian silicates.

STRONTIUM

The reported values of strontium in the rocks of the Riggins Group are mostly in the range 0.01 to 0.1 percent by weight and decrease slightly with increasing silica content. The ultramafic rocks contain less-than-threshold amounts of strontium.

Determinations of strontium are difficult, and the values reported may be unsystematically in error. Certain relations defined by the data are probably real, but they may be much distorted.

The correlation between determined values of strontium and calcium is poor (fig. 54). Little faith can be

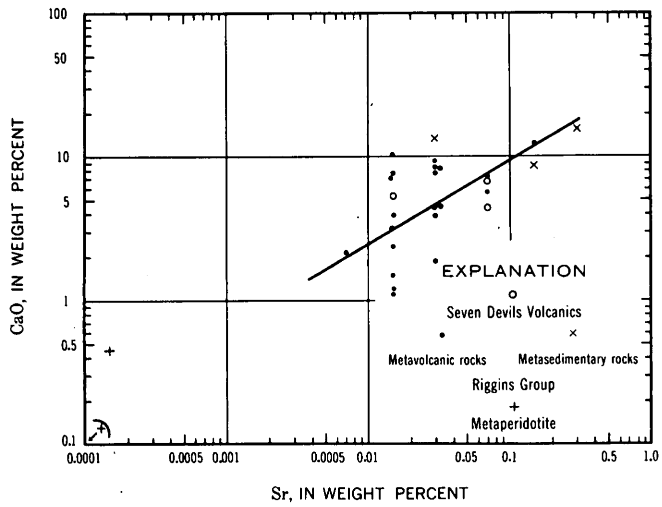

FIGURE 54.—Contents of CaO and Sr in metamorphic rocks of Riggins quadrangle. The doubtfully located line has the equation ($CaO = 35(Sr)^{0.6}$.

placed in the line drawn on the figure, which has the equation, expressed in weight-percent units,

$$CaO = 35(Sr)^{0.6}$$

or,

$$Sr = 0.002(CaO)^{1.8}$$

A better correlation is obtained when strontium is plotted against both potassium and calcium (fig. 55).

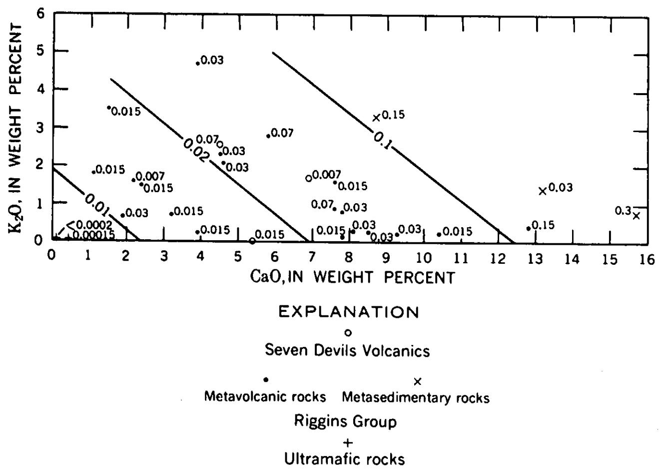

FIGURE 55.—Contents of K_2O, CaO, and Sr in metamorphic rocks of Riggins quadrangle. At each K_2O/CaO point, the percent of Sr is indicated by a number. Contours on the Sr values are highly schematic. All values are in weight percent.

678-769 O—63——5

Strontium increases with calcium and, to a lesser degree, with potassium. The plutonic-rock plot has a very different distribution.

A plot of strontium against barium showed only a possible very weak correlation. The plutonic-rock values of these two elements show a similar possible poor correlation.

Both the absolute amounts of strontium and the Sr:Ca ratio are conspicuously higher in the plutonic rocks than in the metamorphic rocks. The Sr and Sr:Ca values for the Columbia River Basalt of the Riggins quadrangle, by contrast, are lower than those of either of the basement-rock groups.

VANADIUM

The percentage of vanadium is near 0.02 in all but the ultramafic rocks and the most silicic schists, in which it is lower. This element may be presumed to be largely in magnetite.

YTTRIUM AND YTTERBIUM

Of the rare-earth elements, only yttrium and ytterbium are generally present in amounts detectable by semiquantitative spectrographic methods. The yttrium shows an increase with increasing SiO_2, from <0.001 percent in the ultramafic rocks and about 0.002 in the mafic schists, to 0.007 percent in the silicic schists (fig. 48); ytterbium is essentially constant in the schists at about 0.0005 percent, whereas the ultramafic rocks contain <0.0001 percent of it. (The data can even be interpreted to suggest a slight increase in ytterbium with increasing SiO_2.)

The yttrium and ytterbium are presumably in apatite, and their near constancy is consistent with that of phosphorus. The behavior of all three elements in the plutonic rocks is markedly different, for in them these elements decrease regularly with increasing SiO_2.

ZIRCONIUM

The percentage of zirconium increases from 0.003 in the mafic rocks, and 0.001 or less in the ultramafic ones, to about 0.01 in the silicic schists. This obviously indicates a comparable increase in the mineral zircon.

ROCKS INTRUSIVE INTO SCHISTS OF RIGGINS GROUP

PREMETAMORPHIC INTRUSIONS

Prior to their metamorphism, the schists of the Riggins Group were intruded by both volcanic-equivalent rocks and ultramafic rocks. The Berg Creek Amphibolite may have formed from a diabase sill, although metamorphism has completely obliterated the primary fabric. Probably many dikes and sills formed within the volcanic sequence (Lightning Creek and Fiddle Creek Schists) during the period of volcanism, although here again possible evidence has been destroyed by metamorphism except in the metadiorite by Rapid River, described with the other rocks of the Lightning Creek Schist. As it was impossible to distinguish metaextrusive from metaintrusive schists in most of the Riggins Group, the schists have been described on the preceding pages as of metavolcanic origin.

The schists were also intruded by ultramafic rocks before metamorphism, and the distinctive composition of these permits their recognition after metamorphism. These ultramafic rocks are described below.

ULTRAMAFIC ROCKS

OCCURRENCE

Metaperidotite and allied ultramafic rocks occur within the Riggins Group in the northwest quarter of the Riggins quadrangle. A lens about 1,500 feet thick of metaperidotite with subordinate nonultramafic layers crosses the Salmon River at Goff Bridge. A similar lens about 1,000 feet thick occurs along the Little Salmon River at Pollock. Many thin lenses of ultramafic rocks are intercalated in schists on both sides of the Little Salmon River near Sheep Creek. Isolated small lenses, not shown on plate 1, were found in schists at these localities: at an altitude of 4,500 feet, 0.3 mile west of Cat Creek; 0.4 mile northeast of the mouth of Frypan Creek; between Sheep Creek and Captain John Creek, east of the thick lens shown on the geologic map (pl. 1); and 0.5 mile east of the mouth of Hay Creek.

The ultramafic rocks crop out boldly (fig. 56). Most are green in fresh exposures—greens brighter than those of the greenschist and greenstone of the region—but dark-brown "varnish" obscures most outcrops.

LITHOLOGY

The ultramafic rocks are composed dominantly of antigorite, talc, chlorite, actinolite or tremolite, and magnesite in all combinations and widely varying proportions. All these except magnesite also occur as monomineralic rocks. Many of these rocks, particularly those rich in antigorite, are completely massive and lack any mineral orientation visible in either outcrop or thin section. The large fresh exposures (as the highway cut at Goff Bridge) display mottling due to irregular small masses of rocks of varying compositions, and the outcrops reveal a much greater diversity in intercalated types.

The mineral assemblages found in the 23 thin sections studied are listed below. (Only minerals making up more than 1 or 2 percent of the specimens are noted.)

Dominant rock type	Magnesite-talc-antigorite (±tremolite, magnetite)
Widespread types	Antigorite*
	Antigorite-talc
	Magnesite-prochlorite (± talc, magnetite, tremolite)
	Magnesite-talc
	Talc*
	Tremolite-prochlorite
Minor abundance	Actinolite*
	Tremolite*
	Prochlorite*
Very rare	Phlogopite*

*Asterisk indicates rocks that are nearly monomineralic.

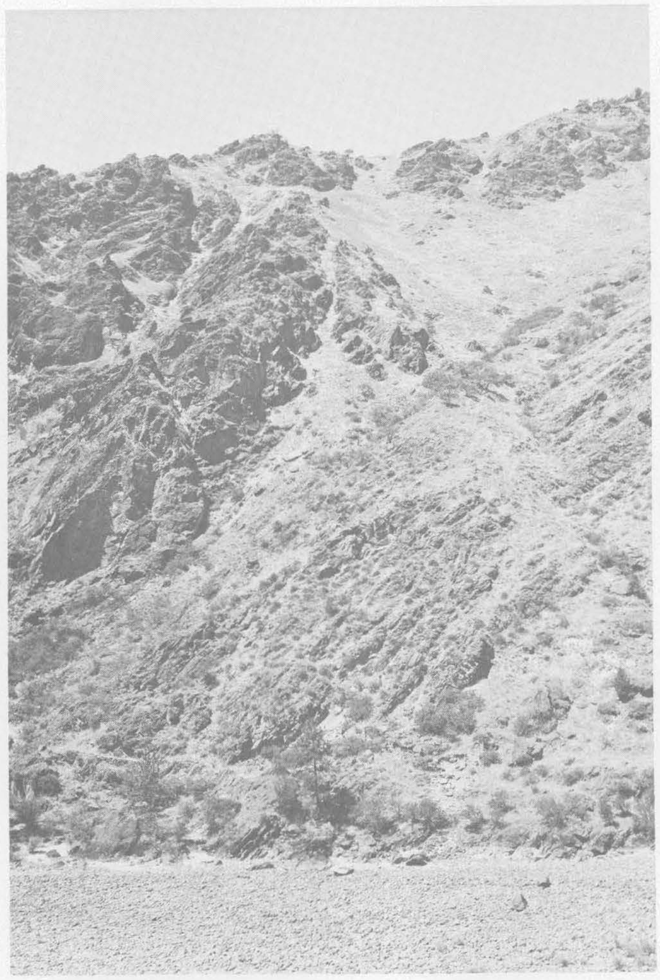

FIGURE 56.—Metaperidotite and greenschist. The layering of metaperidotite, which forms the bold outcrops on the left and at top, and of greenschist of the Lightning Creek Schist dips 40° to the left (south), but the actual contact is probably a fault. West side of Salmon River, north of Goff Bridge.

Most of the rocks not rich in talc are massive (displaying little or no planar or linear mineral orientation), green, and have a superficial resemblance to greenstone. Small porphyroblasts of gray magnesite stud many of the rocks, which consist principally of decussate antigorite, talc, and chlorite (fig. 57). The decussate fabric of randomly interlacing crystals makes the rocks very hard, and their resistance to fracturing is perhaps the principal reason for their commonly bold outcrops (fig. 56).

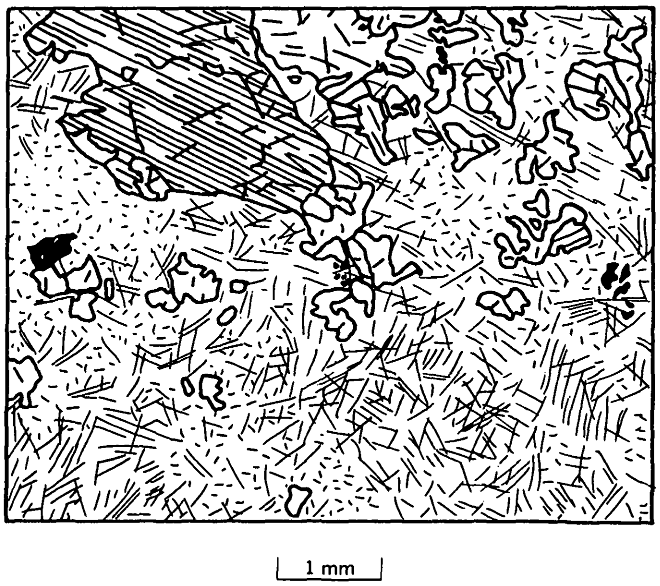

FIGURE 57.—Magnesite-talc-antigorite metaperidotite. Ragged crystals of magnesite (high relief) lie in irregular masses of decussate talc (tiny plates) and antigorite (larger plates). A few grains of magnetite are in the field. Analyzed specimen 1 (table 6), Salmon River at Goff Bridge.

Peridotites or other possible original rocks are nowhere preserved with their initial mineralogy. Relict grains of olivine were found in one antigoritite, however (fig. 58). The olivine is magnesian (2V=90°) and makes up 10 percent of the present rock. The olivine is altered inward from all fractures, mostly to antigorite, but locally to chrysotile and magnesite. Its initial grain size was about 3 mm.

The talc-rich rocks are schists (figs. 59 and 60) that are gray where fresh and either gray or rusty brown where weathered. Talc schist weathers to punky debris, and outcrops are mostly small. Much talc schist has a knotty appearance due to porphyroblasts of magnesite (fig. 61). In addition to the common minerals of the ultramafic rocks, some of the talc schist contains anhedra or feathery aggregates of quartz.

Most of the ultramafic rocks are composed of assemblages of several minerals, but nearly monomineralic rocks are also widespread. Rocks composed almost entirely of antigorite or of talc are abundant; near-pure tremolite, actinolite, chlorite, and phlogopite rocks occur also.

Talc schist crops out as gray rock variably stained by yellow-brown iron oxide. Most of it has very strong planar structure; uncommonly, the talc is in shingled anhedral spindles that give an extreme pencil-structure lineation but only a weak foliation. The most common of the local secondary cleavages is a rough slip cleavage along which the folia are much contorted. Less common is a very straight fracture cleavage, along which there was only minute deformation, whose trace appears as ruled lines on the main folia surfaces.

Antigorite generally displays a striking crisscross pattern of antigorite plates (like that shown in figs. 57 and 58) and totally lacks recognizable directional fabrics in either thin section or outcrop.

FIGURE 58.—Relict crystals of olivine in antigorite metaperidotite. A coarse cross-hatched pattern of antigorite flakes is printed across the fresh Mg-rich olivine (high relief). The opaque mineral is largely magnetite. Collected by trail on west side of Salmon River 250 yards north of Goff Bridge.

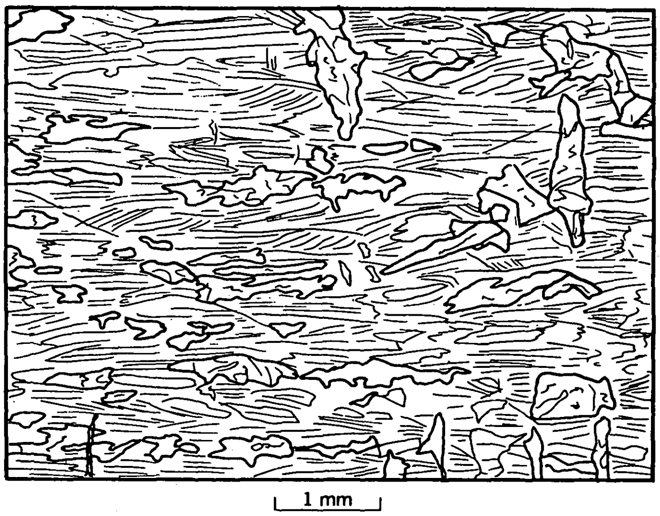

FIGURE 59.—Talc schist. The shingled anhedral spindles of talc have a very strong lineation but only a weak foliation. Large plates of antigorite lie within, and also at right angles to, the foliation; the latter orientation marks a weak secondary cleavage. West side of Salmon River 300 yards north of Goff Bridge.

FIGURE 60.—Magnetite-chlorite-magnesite schist. The chlorite foliation is not deflected about the anhedral grains of magnesite (high relief) or magnetite. Intercalation in talc schist, Salmon River 500 yards north of Goff Bridge.

FIGURE 61.—Magnesite-talc schist. Subhedral porphyroblasts of magnesite truncate the foliation of the talc. Talc is mostly parallel to foliation, but some laminae have obliquely oriented or decussate talc. Little Salmon River, 0.6 mile south of Sheep Creek.

FIGURE 62.—Actinolitite. A monomineralic rock with a crude schistosity. The actinolite is light greenish gray in hand specimen and almost colorless in thin section. From an exposure at an altitude of 3,900 feet, 0.9 mile north of Sheep Creek.

FIGURE 63.—Tremolite-chlorite metaperidotite. The needles of tremolite have a planar, but not a linear, orientation. The groundmass is of decussate chlorite. Ilmenite occurs as euhedral plates. West side of Salmon River, 400 yards north of Goff Bridge.

A few thin layers and lenses of actinolite (fig. 62) and of tremolitite were seen. The actinolitite is much the coarser; one mass at Goff Bridge has an average prism length of 1 cm.

Both near Goff Bridge and along the Little Salmon River, chlorite forms greenish-gray chloritite. It is decussate (but not crosshatched like the antigorite) to weakly schistose. Most is fine grained, but some is in porphyroblasts. Bluish-gray chloritite occurs also in unmapped small masses in the area about lower Sheep Creek and Hay Creek. Figure 63 illustrates a chlorite-rich ultramafic rock.

Several blocks of shiny dark-gray phlogopitite were found above the road cut at the north abutment of Goff Bridge. This is a striking schist, with an unusual glitter from the pure-mica surfaces of uniform 0.2-mm flakes.

MINERALOGY

Most of the ultramafic rocks are composed of antigorite, talc, and magnesite in varying proportions. Antigorite, the most widespread, is bright green in hand specimen, colorless in thin section, and has a low birefringence. Abnormal bluish-gray interference colors (but no ultra blue) are seen in much of it. The antigorite occurs characteristically in decussate masses that have a striking crosshatched appearance when viewed through crossed nicols (figs. 57 and 58). Most of it is in pure masses of flakes 0.05 to 0.5 mm long, but flakes cut most other minerals. Locally the antigorite is schistose. It is associated most commonly with talc and magnesite; chlorite and antigorite are not found together.

Calculations based on the mineralogic and chemical composition of the antigoritite whose analysis is presented in table 6 indicate that its antigorite, which makes up 98 percent of the rock, is virtually

$$2.4\ MgO \cdot 0.3FeO \cdot 2\ SiO_2 \cdot 2\ H_2O$$

It also contains minor CaO, Al_2O_3, and Fe_2O_3. (Theoretical antigorite is $3\ MgO \cdot 2\ SiO^2 \cdot 2\ H_2O$.)

Talc, although more commonly schistose, is plumose or decussate in many rocks, and all three textures occur together in some specimens. Most talc flakes are smaller than 0.1 mm. In one specimen, 2V was determined as (−) 18°.

Most chlorite is greenish gray in hand specimen and yellowish green in thin section and has an abnormal greenish-gray interference color. It has a very low (+) 2V. The chlorite resembles the intermediate prochlorite of the schists of the Riggins Group; if it is compositionally similar, then the source of the aluminum might have been in local masses of aluminous pyroxenite. Chlorite occurs with all degrees of planar orientation from decussate (fig. 63) to lepidoblastic.

In a specimen of chlorite-magnesite-talc schist from near Goff Bridge, the chlorite appears to be a ferroan prochlorite that is weakly pleochroic pale green but that has an abnormal dark-brown interference color. It occurs chiefly in aggregates less than 2 mm in diameter that have a finely lamellar structure suggestive of a possible relict diallage parting. This chlorite might be pseudomorphus after clinopyroxene.

Tremolite and actinolite are everywhere acicular, with average lengths ranging from 0.2 to 20 mm in different specimens. Elongation ratios (c : a) as great as 50 : 1 are common. These amphiboles occur with all degrees of planar orientation from felted to nematoblastic; but strong lineations, such as are widespread in the hornblende greenschists, were not seen. The tremolite is light greenish gray in hand specimen and colorless in thin section. Actinolite is yellowish green in hand specimen and pale greenish yellow in thin section and has a (−) 2V of ≈85° that indicates a composition near tremolite. Tremolite prisms in one schist contain irregular areas of actinolite. Thin veins of tremolite asbestos are found in many places.

Carbonate, abundant in the ultramafic rocks, is mostly magnesite. It is commonly anhedral, and less commonly it occurs as crude rhombs. In hand specimen it is light gray and appears as knots (of one or many crystals) several millimeters across. Bent crystals were seen rarely; and, as it truncates cleavage most of the magnesite is of relatively late crystallization (figs. 60, 61). The low index ($n_{e'}$) is commonly near that (1.55) of the resin used on the thin sections, indicating a composition near $(MgCO_3)_{70}(FeCO_3)_{30}$.

The phlogopite from blocks of phlogopitite near Goff Bridge is grayish black in hand specimen and yellowish orange in thin section and has a (−) 2V of 10°. It shows spotty alteration to vermiculite(?) that gives parts of grains a higher relief and a bronzy reflected-light appearance without affecting color, birefringence, or extinction. Similar phlogopite forms 1 percent of a specimen of antigorite-quartz-magnesite-talc schist.

In many regions, orthorhombic amphiboles, notably anthophyllite, are common in metamorphosed ultramafic rocks, but none were found in the Riggins quadrangle.

Magnetite, with hematite as an alteration product, occurs both as octahedra and as anhedral grains; uncommonly, it occurs as extremely irregular poikilitic sieves. Ilmenite forms euhedral plates in some rocks (fig. 63). The presence of several tenths of a percent of chromium suggests that chromite is present, but neither it nor other possible opaque minerals were recognized in the thin-section study.

Bright-yellow rutile and picotite(?) and sphene, apatite, zircon, and clinozoisite are rare accessories.

Quartz was seen in a single antigorite-magnesite-talc schist, of which it comprised about 15 percent. The quartz occurs as ragged anhedra and feathery aggregates.

CHEMICAL COMPOSITION

The mineralogy of the ultramafic rocks indicates that they have low contents of SiO_2 and very high contents of MgO. Most of the minerals are richly hydrous.

Analyses of two specimens of ultramafic rocks are given in table 6. Volatile-free recalculated compositions are also listed, and these are consistent with an origin from rocks composed dominantly of magnesian olivine, with less enstatite and still lesser amounts of augite and other minerals. Such an assemblage is a common one in terranes of similar but less-metamorphosed rocks, in the Klamath Mountains of northwestern California and southwestern Oregon, for example.

TABLE 6.—*Composition, in percent, of metaperidotite*

	1A	1B	2A	2B
Major oxides, by weight				
SiO_2	38.9	46	42.2	49
Al_2O_3	1.7	2	2.2	2.5
Fe_2O_3	1.8	2	1.8	2
FeO	5.4	6	6.0	7
MgO	37.2	44	34.6	40
CaO	.08	.1	.45	.5
Na_2O	.07	.1	.11	.1
K_2O	.02	.02	.06	.07
TiO_2	.02	.02	.04	.05
P_2O_5	.00	--------	.00	------
MnO	.08	.1	.10	.1
H_2O	10.0	--------	12.6	------
CO_2	4.6	--------	.65	------
Total	100	--------	101	------
Minor elements, by weight				
Ba	0.0003	--------	0.0015	------
Co	.007	--------	.007	------
Cr	.15	--------	.15	------
Cu	.003	--------	.007	------
Ni	.15	--------	.15	------
Sc	.0015	--------	.0007	------
Sr	N.D.	--------	.00015	------
V	.007	--------	.007	------
Zr	N.D.	--------	.0015	------
Minerals, by volume				
Antigorite	56	--------	98	------
Talc	28	--------	1.3	------
Magnesite	15	--------	.1	------
Opaque	.4	--------	.6	------
Phlogopite	None	--------	Trace	------

1A. Magnesite-talc-antigorite metaperidotite. Massive greenish-gray rock composed of felted to crosshatched 0.1- to 0.5-mm flakes of antigorite, mostly in pure masses; ragged masses to 3 mm long of one to several crystals of magnesite, laced by antigorite; and irregular masses of randomly oriented, minutely intergrown plates of talc cut by antigorite flakes. The thin section contains relatively more magnesite than does the fragment analyzed. From roadcut by Salmon River 100 yards north of Goff Bridge. [Field No. SR 14-1; major-oxide analysis by rapid methods by P. L. D. Elmore, P. W. Scott, S. D. Botts, and K. E. White; laboratory No. 148,398; minor-element analysis by semiquantitative spectrographic methods, by P. R. Barnett; laboratory No. C 670. N.D., not detected].
1B. Same as column 1A, recalculated without H_2O and CO_2.
2A. Antigoritite. Massive greenish-gray rock composed largely of felted 0.05-mm plates of antigorite and streaks and knots of coarser plates. The rare phlogopite is in tiny flakes. Carbonate contents of thin section and fragment analyzed are obviously different. From roadcut by Little Salmon River 1.25 miles north of Hay Creek. Field No. SR 37-3; laboratory Nos. 148,401 and C 673; analysts and methods as above.
2B. Same as column 2A, recalculated without H_2O and CO_2.

The minor-element contents are those which would be expected in ultramafic igneous rocks that are, for example, characteristically high in chromium and nickel. Semiquantitative analyses indicated 0.15 percent of each of these elements in the two specimens in table 6. Quantitative spectrographic analyses for chromium and nickel were made of seven other specimens of ultramafic rocks by N. M. Conklin, U.S. Geological Survey, Denver, 1958. These indicated the following:

Rock type	Number of specimens	Percent Cr	Percent Ni
Antigoritite	4	0.3–0.4	0.2–0.3
Talc schist	1	.1	.2
Magnesite-talc schist	1	.003	.07
Tremolite-prochlorite rock	1	.03	.1

The antigorite-rich rocks are highest in both chromium and nickel, and although prospecting in them might locate ore deposits, similar contents of chromium and nickel characterize most ultramafic rocks, and few are of economic value. Presumably the chromium is in chromite, and the nickel is in silicates.

According to Howard F. Albee (written communication) a single small pod of fairly high grade chromite ore in the Riggins quadrangle was mined during World War II. This mine is in the NW¼SW¼ sec. 8, T. 23 N., R. 1 E.

The metaperidotite extending north from Goff Bridge on the west side of the Salmon River has been prospected extensively by claimant Carl Holland, but as of 1961 no ore had been found.

METAMORPHISM

Enormous quantities of water and subordinate carbon dioxide must have been added to these rocks to permit formation of the present minerals from the almost anhydrous pyrogenic minerals. If the original ultramafic masses were intruded into their present positions as serpentines, the volatiles were added early in the tectonic cycle; but if the rocks were peridotites immediately before metamorphism, then the water and carbon dioxide were added during metamorphism. Silica was presumably added also to some of the rocks, or bases, particularly magnesium, were removed, and metamorphic reactions must have been complexly interdependent. The metamorphism was probably accompanied by much migration of components, as the compositions of the rocks change in irregular fashion.

Associated nonultramafic metamorphic rocks contain garnet, oligoclase, and other minerals indicative of middle-grade metamorphism. Although the minerals that make up the ultramafic rocks are commonly considered to be products of low-grade metamorphism, they have here persisted in rocks of relatively high grade.

ORIGIN

The ultramafic rocks formed by metamorphism of older ultramafic rocks—serpentine or peridotite or both. The compositions suggest derivation from enstatite-bearing peridotite; but of the original minerals, only olivine is now present, and that was seen in only one thin section (fig. 58).

The ultramafic rocks north of Riggins lie along the contact between the dissimilar Squaw Creek and Lightning Creek Schists. South of Riggins, some masses lie between the same formations, and other parallel masses are nearby. Possibly the ultramafic rocks were intruded fortuitously along a major lithologic break in a stratigraphic succession; but much more likely, the formations are separated by a thrust fault along which the ultramafic rocks were intruded, probably during thrusting. Serpentines follow thrust faults in many regions, and the position here of metaperidotite between dissimilar formations supports such an interpretation.

Comparable ultramafic rocks are common in similar settings throughout the world and are the subject of extensive literature. The serpentine and peridotite of alpine type (Benson, 1927), with accompanying gabbros in some regions but not in others, are intruded during major orogeny into eugeosynclinal fillings dominated by the products of andesitic and basaltic volcanism. The peridotites cause only slight thermal metamorphism of their walls and clearly are intruded as warm solids—mushes of crystals, mostly olivine in the case of peridotite, and hydrous pastes in the case of serpentine—rather than as molten magmas. Alpine ultramafic rocks are intruded as allochthonous crystals of olivine and much subordinate pyroxene, or as alteration products of such crystals.

Alpine ultramafic rocks are commonly assumed to be mobilized directly from the upper mantle, but nothing in the data requires this to be true. Basalt is known to come from the mantle, and it is almost impossible to consider basalt as a product of partial fusion of peridotite of alpine type. Such peridotite contains, on the average, only 1/40 to 1/400 as much sodium, calcium, potassium, titanium, phosphorus, barium, rubidium, strontium uranium, and other elements as does even tholeiitic basalt, which is lower in most of these components than is olivine basalt. No known mechanism could produce the enormous concentrations required to yield uniform basalt from such material in all parts of the world. The upper mantle must contain a much larger component of basaltic composition than does peridotite.

Alpine peridotite may form initially as gravitational accumulates of early-formed crystals in basaltic magmas in the large chambers which feed the surface volcanoes of eugeosynclines and subsequently be intruded upward in response to orogenic deformation of the volcanic pile. Alternately, alpine peridotite may be the unfused residuum remaining in the mantle after the melting and upward migration of basaltic magma and be squeezed upward during orogenic kneading of the upper mantle beneath the geosyncline.

ASSOCIATED ROCKS

Intercalated with the metaperidotite of the Goff Bridge area is dark-gray carbonaceous schist whose rocks are mineralogically and texturally unlike any seen elsewhere in the quadrangle. Weathering much like the metaperidotite, they can be traced only with difficulty but seem to form irregular lenses as much as 20 feet thick. These are puzzling rocks. For instance, despite its unique mineralogy, the bulk composition of the single specimen analyzed (table 5, No. 7) is, except for its elemental carbon, strikingly like that of the metavolcanic rocks of similar SiO_2 content in the Riggins Group.

Three thin sections of these rocks were studied, and each was found to be so distinctive that the specimens are described separately here. The rocks contain abundant opaque dust, chiefly carbon, to which they owe their color. Quartz is abundant in all 3 specimens, albite in 2, amphibole in 2, and phlogopite in 1.

Analyzed specimen 7, (table 5), from the west side of the Salmon River 150 yards north of Goff Bridge, is a medium-dark-gray knotty schist with aggregates of hornblende mostly between folia of opaque minerals and very fine grained quartz and albite. The hornblende is in sheaves and feathery aggregates that include many curved prisms. The opaque dust of the folia passes through many of the hornblende crystals as ghost inclusions. The opaque dust also forms fracture patterns—polygonal areas separated by clear hornblende—in many hornblende aggregates. Continuous sheaves of undeformed hornblende are printed over these fracture patterns. The hornblende is dark green in hand specimen and lighter and grayer in thin section than any of the common hornblende or actinolite of the region; its pleochroism is X=light greenish yellow, Y=yellow green, and Z=grayish green, slightly bluish. Hornblende is altered locally to yellow biotite in plates along the cleavage and to yellow chlorite (abnormal Prussian blue through crossed Nicols) in sheaves with intergrown biotite. The rock contains several percent aluminian epidote (colorless, moderate birefringence) in both euhedral domes and extremely irregular crystals that cut across the foliation and in granules parallel to foliation. Dodecahedra of garnet also cut the foliation, and scattered crystals of brilliant yellow isotropic picotite(?) are present. Although its carbon content indicates sedimentary or metasomatic origin, the rock is classed with the metavolcanic rocks in table 5 because most of its bulk composition is that of a volcanic rock.

A second dark-gray schist, which is faintly mottled by vitreous and porcelaneous areas, is in some ways similar to that just described. The very abundant opaque dust is concentrated in both folia and polygons that clearly define an old fracture pattern. Most of the rock is composed of very fine grained quartz. Tremolite forms sheaves and radiating clusters of tiny needles that cut through all other elements of the rock. Metasomatism of rock such as this, with the formation of much more amphibole, and that richer in iron, could form rock such as the first schist described.

The third specimen of similar dark-gray rock contains small ptygmatic lenses of tan-weathering calcite in a knotty decussate mass of fine-grained phlogopite, albite, and quartz. The abundant opaque dust is variably pervasive or concentrated in folia or rims about knots of other minerals and in irregular patches. Actinolite, pale greenish-yellow chlorite ($+2V \approx 27°$; aluminian prochlorite?), clinozoisite, and sphene are minor components.

Contents of chromium and nickel in these three specimens were determined by quantitative spectrographic methods by N. M. Conklin (U.S. Geological Survey, Denver, 1958), who found weight-percentages of chromium of 0.02, 0.003, and 0.02, and of nickel of 0.006, 0.003, and 0.02, respectively. These low values emphasize the distinction between these unusual schists and the enclosing ultramafic rocks and are comparable to the amounts found in the metavolcanic and metasedimentary rocks of the Riggins Group.

Study of the thin sections suggests that previously fractured carbonaceous rocks were sheared and reconstituted, possibly with the metasomatic addition of new hornblende. Perhaps the schists originated as lenses of metasedimentary rocks. The distinctive mineralogy might, however, be due to metamorphic exchange (in both directions?) of material with the metaperidotite. The unusual fabric may be due to a stress system during metamorphism different within the ultramafic mass than without—the ultramafic rocks may have shielded these rocks from the elsewhere-pervasive shearing.

POSTMETAMORPHIC VEINS AND INTRUSIONS

Cutting the schists of the Riggins Group are veins and dikes that vary systematically in type and abundance. Along the Salmon River from Little Berg Creek to the north edge of the quadrangle, and along the Little Salmon River from 3 miles south of Rapid River to 2 miles north, this material is limited to veins and subordinate pods of quartz and, more locally, carbonate. East and south of these limits, aplite and pegmatite dikes are abundant (fig. 72). Northeast of Riggins, the pegmatite-bearing schists are truncated against the intrusive contact of trondhjemite gneiss. South and east of Riggins, dikes of trondhjemite and subordinate quartz diorite are abundant, appearing first 1 mile or so beyond the first of the pegmatites.

VEIN ROCKS

Pods and veins of white quartz or calcite, alone or together, are common in the lower grade rocks, as along the river north of Riggins. Some of these have been thoroughly crushed and recrystallized and antedate much of the metamorphism, but most are little sheared and must have formed late in the sequence of metamorphism and deformation. Some of these masses contain much mica or prochlorite, similar to the minerals in the immediate wallrocks. Pods (stubby lenses as thick as a few feet) are generally concordant to foliation, whereas tabular veins, which are more abundant, are either concordant or crosscutting. Veins are concentrated in subparallel fractures, generally at an angle to foliation, and reach common thicknesses of several feet.

The veins and pods represent deposition, chiefly in fractures, by either gels or aqueous solutions. Quartz veins are much more conspicuous in the Squaw Creek Schist, most of which bears abundant free quartz, than in the quartz-poor Lightning Creek Schist. Carbonate veins are chiefly near carbonate-bearing schists. Most of the vein material probably was derived by solution from nearby schists.

PEGMATITE DIKES

Quartz veins give way southward and eastward to dikes of pegmatite and white aplite (fig. 72) that occur either separately or zoned together. Dikes range in thickness from 1 foot to 10 feet and are commonly relatively straight as seen in most roadcuts. In large cliff outcrops, as in the steep-walled canyon of Hailey Creek, the dikes are seen to be irregular, branching and lensing at small angles (fig. 64). Most of the dikes at any one locality are subparallel, emplaced either along the foliation or along a set of discordant fractures.

The dike rocks are white. Grain size is commonly variable within each dike and reaches a maximum of several inches. Sodic plagioclase is much dominant over potassic feldspar in most of the rocks. Clots, pods, and wispy folia of minerals—micas, chlorite, hornblende, and garnet—that match those of the schists in the adjacent walls are plentiful. Garnets are also abundantly disseminated in many of the dike rocks.

Where calc-silicate rocks are cut by pegmatites, as in Hailey Creek Canyon, marginal amphibolite was produced in the wallrocks and abundant disoriented hornblende xenocrysts were rafted into the pegmatites.

Despite the obvious incorporation of wallrocks into the pegmatites, it is clear that at least most of the dikes are of intrusive, rather than metasomatic, origin. Diverse wall types are cut across sharply by straight-walled dikes with only slight changes in the composition of the dikes. Xenoliths and xenocrysts were moved and rotated from their original positions; they were obviously incorporated into a mobile medium.

The analysis of a metamorphosed pegmatite from a dike in the headwaters of Shorts Creek is given in table 7. This is a white rock, seamed by intersecting folia of muscovite (3 percent). The white material (97 percent) is quartz and sodic oligoclase in poorly defined layers of varying grain size. Larger oligoclase crystals are moderately bent and broken. There are sparse small red garnets and tiny opaque granules. The composition is highly sodic (Na_2O=5.1 percent, whereas CaO=1.2, and K_2O=1.5), indicating that the pegmatite may be related genetically to the trondhjemites.

It was suggested above that the quartz veins of the lower grade rocks were derived by local mobilization. As the quartz veins become relatively less numerous and the pegmatite dikes become much more abundant eastward and southward in rocks of higher metamorphic grade, a comparable origin of the pegmatites by local mobilization during metamorphism is conceivable. On the other hand, the pegmatites give way in part to trondhjemite dikes, and the sodic compositions of both suggest a related genesis. Beyond the trondhjemite dikes are large intrusive masses of that rock type. All this can be interpreted to mean either that the trondhjemite and pegmatites are both derived by metamorphic mobilization of the most soluble, lowest melting components of the schists (which are also sodic) or that trondhjemite dikes and pegmatites are far-traveled injections from the plutons; they might even represent a combination of these processes. Despite the suggestion that veins of quartz and carbonate were of local origin, it seems likely that most of the dikes were intruded from more distanct sources.

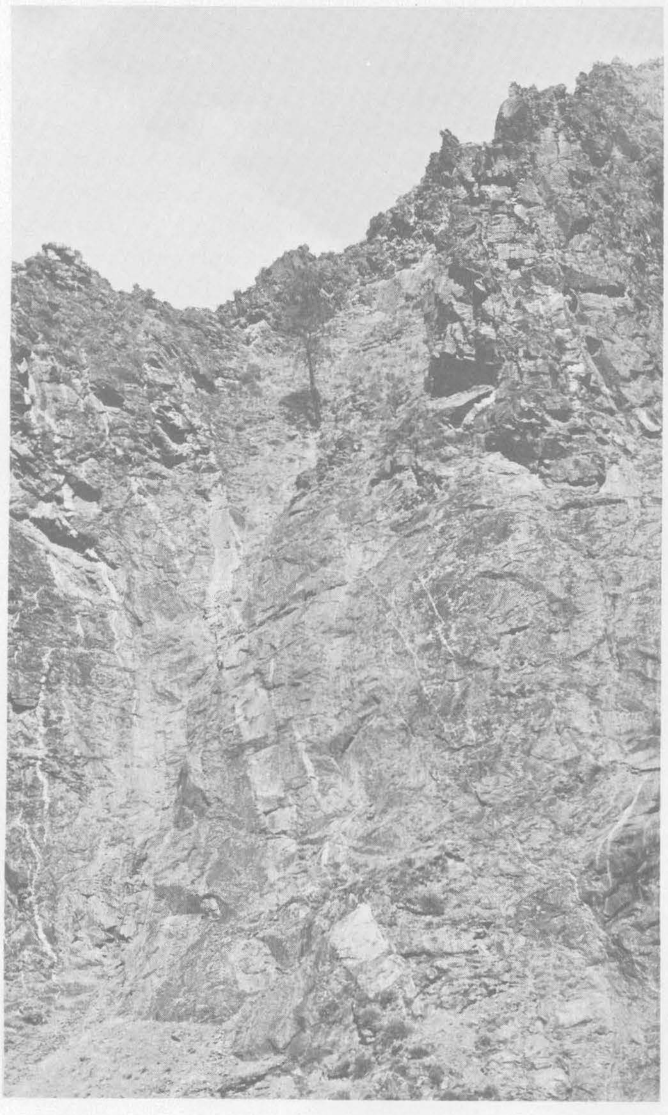

FIGURE 64.—Dikes in schist. The near-vertical layering of the Squaw Creek Schist strikes into the cliff and is cut obliquely by the white sodic pegmatites. North side of Hailey Creek Canyon, one-half mile above mouth.

TABLE 7.—*Composition, in weight percent, of garnet-muscovite-quartz-oligoclase metapegmatite*

[Rock intrusive into Squaw Creek Schist, altitude 4,300 feet, in tributary canyon 2 miles above mouth of Shorts Creek. Rapid-method major-oxide analysis by P. L. D. Elmore, P. W. Scott, S. D. Botts, and K. E. White; laboratory No. 148420. Minor-element analysis by semiquantitative spectrographic methods by P. R. Barnett; laboratory No. C692; elements not detected are not listed. Field No. SR 372]

Major oxides		Minor elements	
SiO_2	75.7	Ba	0.07
Al_2O_3	15.5	Cr	.00015
Fe_2O_3	0.48	Cu	.0007
FeO	.13	Ga	.003
MgO	.27	Pb	.003
CaO	1.2	Sc	.003
Na_2O	5.1	Sr	.015
K_2O	1.5	Y	.007
TiO_2	.02	Yb	.0007
P_2O_5	.00	Zr	.007
MnO	.09		
H_2O	.62		
CO_2	<.05		
Total	101		

DIKES OF GRANITIC ROCKS

Dikes of trondhjemite and quartz diorite become increasingly abundant to the east and south as the grade of metamorphism increases. Their northwest limit is generally about 1 mile southeast of that of the pegmatite dikes, except that they extend as far as the pegmatites at the Little Salmon River between Captain John and Hailey Creeks (fig. 72). At the arbitrary boundary between the terranes designated as "metamorphic" and "plutonic," granitic material makes up about a twentieth of the gneiss complexes. Beyond the boundary, in the varied gneisses that represent the high-grade metamorphic equivalents of the Riggins Group, granitic rocks increase in abundance rapidly and become first equal to, then greatly dominant over, the metamorphic rocks in bulk.

Within the terrane mapped as the schists of the Riggins Group, the dikes of granitic rocks are chiefly of trondhjemite, which is light-colored sodic biotite "quartz diorite." These are medium-grained rocks that contain 60 to 75 percent calcic oligoclase, 20 to 30 percent quartz, and less than 5 percent biotite, which is in scattered pepperlike flakes. Plagioclase is in well-shaped normal-oscillatory zoned crystals, mostly 2 to 3 mm long. Both albite and acline twins are widely developed, and cross-twinned crystals are abundant. Quartz is in smaller anhedra and has a common tendency to occur in small aggregates. Some of the dikes contain a few percent orthoclase. Minor amounts of late muscovite, epidote, and clinozoisite commonly are present. A few rocks contain trace amounts of hornblende or garnet.

Many of the trondhjemite dikes are obscurely gneissic, their biotite and their slight dimensional orientation of quartz and plagioclase being subparallel to the dike walls. As the dikes contain secondary minerals, like clinozoisite, with textures that suggest alteration at higher metamorphic grade than those of common deuteric alteration, slight metamorphism of the dikes seems indicated. Some of the dikes are obviously metamorphosed, having been crushed to mortar gneisses and moderately recrystallized; but even these rocks have undergone metamorphism of far less severity than that of the wallrock schists.

Dikes of quartz diorite are less widespread, and some at least of these were apparently produced by contamination of trondhjemite by wallrocks. Thus, at an altitude of 4,700 feet on the ridge southwest of Hailey Creek, contaminated dikes cut calc-silicate rocks. The dikes contain 70 percent zoned plagioclase (An_{40-25}), 10 to 15 percent quartz, 10 percent ragged green diopside, and a total of 5 to 10 percent hornblende, clinozoisite, calcite, sphene, and garnet.

Still less silicic is a dike of sodic diorite, for which a hybrid origin is also likely, east of Whitebird Ridge, 1.5 mile south-southeast of Pollock. This rock contains 80 percent plagioclase (An_{35-25}), 8 percent biotite, 5 percent quartz, and 5 percent hornblende plus epidote. Most of the plagioclase crystals have cores that are remnants of oscillatory-zoned crystals cut by networks of more sodic plagioclase continuous with that of the uniform rim.

TRONDHJEMITE STOCK OF WHITEBIRD RIDGE

A small stock of trondhjemite, less than 1 square mile in area, lies within the Squaw Creek Schist southwest of Pollock (pl. 1). The stock is poorly exposed, and most of it is deeply weathered. Dips of flow structures are gentle and variable in the few outcrops in which they were recognized. The attitude of the contact was nowhere seen, although along at least the west margin it must be steep. The stock probably cuts sharply across the wallrock schists. The rock is medium grained (coarser than most of the trondhjemite dikes) and has an average grain size of about 3 mm. The stock is very similar to the dikes in composition and mineralogy.

The trondhjemite of the stock contains 65 to 75 percent normal-oscillatory zoned calcic oligoclase in well-shaped 1- to 4-mm grains that commonly show both albite and acline twins. Quartz comprises 20 to 30 percent, mostly in small aggregates of 1- to 2-mm anhedra. Biotite makes up 3 to 5 percent, in flakes that range in average grain size in different specimens from 0.5 to 3 mm, and in orientation from massive to gneissic. Variations in the size of biotite flakes gives the illusion of a large variation in the fabric of the rock, but the quartz and oligoclase are of nearly constant size range throughout. The rock contains less than 1 percent hornblende where it is present at all. Red garnet is locally present in amounts of 0.1 percent or so.

Most of the stock is slightly altered, and exceptionally the biotite is completely chloritized, but cataclasis was slight. Muscovite, partly interleaved with biotite and partly separate, makes up 1 or 2 percent of the rocks. A little epidote and clinozoisite are present in plagioclase cores. Epidote also occurs as scattered domes and as intergrowths with biotite.

Dikes of trondhjemite are abundant southeast of the stock.

KERATOPHYRE PROBLEM

The metavolcanic rocks of the Riggins quadrangle, as is typical of eugeosynclinal suites, have widely variable but generally high ratios of sodium to calcium as compared to most other volcanic rocks of similar silica contents. The metavolcanic rocks are partly normal andesite, basalt, and dacite in composition but most are keratophyre, spilite, and quartz keratophyre. The origin of such rocks has received much discussion, and many conflicting theories have been advanced.

The petrology of spilite and keratophyre and the extensive literature on them were summarized in detail by Gilluly (1935), and the present discussion supplements that synthesis, notably in the inclusion of newer data. Shorter summaries include those by Eskola (*in* Barth, Correns, and Eskola, 1939, p. 380–381) and by Turner and Verhoogen (1951, p. 202–212). Current theories explain these sodic rocks variously as formed from primary sodic magmas or from calc-alkaline rocks altered by sodium added from magmatic, geosynclinal, or seawater sources.

The rocks under consideration here are keratophyre, quartz kerotophyre, and spilite, intercalated with calc-alkaline andesite, dacite, and basalt. They apparently formed beneath the sea in a eugeosyncline as products of a volcanic arc similar to that of the Aleutian Islands. Such rocks were assigned to the "spilite-keratophyre association" by Turner and Verhoogen (1951); but as normal calc-alkaline rocks are intercalated with them, and as intermediate rocks are most abundant, a designation as "andesite-keratophyre association" seems preferable and will be used here. Keratophyre bears the same relation to andesite that spilite does to basalt, and, as Gilluly (1935, p. 347) emphasized, keratophyre is more abundant than spilite in Oregon and in many other regions; andesite is the dominant unaltered rock. In the Andean geosyncline, for another example, the most abundant rocks are andesite and keratophyre (Cristi, 1956; Harrington, 1956).

The problem of the origin of any igneous rock must be considered within the framework of its tectonic environment and the petrology and relative abundances of the associated rocks. Benson (1927) was among the first to systematize such a framework. The many "associations" designated by Turner and Verhoogen (1951) advance this concept. All rocks with a given rock name, or all rocks of similar composition, cannot be considered to have the same origin. Rocks designated as "keratophyre" give an instance in point. The term "keratophyre," as defined originally by Gümbel (1874, p. 43), and as used here, applies to the spilitic equivalent of an andesite; such rocks are associated with andesite, basalt, dacite, spilite, and quartz keratophyre. The term unfortunately has also been used for sodic trachyte that is in many places associated with basalt and mugearite—rocks completely different in mineralogy and in probable genesis despite superficial similarity of chemical composition. Some of the older literature attemps unsuccessfully to explain both types of "keratophyre" as due to the same processes.

Another common misuse of the term "keratophyre" is for greenstone of normal calc-alkaline andesite composition, in which the feldspar is now albite formed by low-grade metamorphic reconstitution of more calcic plagioclase. Most low-grade metavolcanic rocks contain albite as their feldspar, and the name "keratophyre" should be reserved for those rocks which have abnormal high-sodium, low-calcium bulk composition. The distinction can be made in part by thin-section study alone, even though both calc-alkaline and sodic rocks contain albite as their only feldspar after low-grade metamorphism. Altered basalt and andesite contain large quantities of calcium-bearing minerals, notably epidote and carbonates, whereas end-member spilite and keratophyre lack calcic minerals. By consideration of the proportion of calcic minerals to albite and to other felsic and mafic minerals of the rock, it is commonly easy to differentiate the rocks into calc-alkaline, transitional, or sodic groups.

In some regions the dominant rocks of mixed sodic and calc-alkaline assemblages are spilite, basalt, and diabase, more silicic rocks being far less abundant or even lacking. Such assemblages have eugeosynclinal environments comparable to those of the andesite-keratophyre association and perhaps are a specific compositional variety of the same genetic type, just as parts of some volcanic arcs are chiefly of basaltic rather than of andesitic rocks. These more mafic assemblages should be assigned to a basalt-spilite association.

Spilite and keratophyre are characterized by sodic feldspar, generally albite, with low-temperature optics, in place of the high-temperature calcic or intermediate plagioclase of normal calc-alkaline rocks. The distinction is reflected in the bulk composition of the sodic rocks—their Na_2O content is higher, and their CaO and Al_2O_3 contents are lower. Many spilites and keratophyres show clear evidence, notably in relicts of initial feldspar in albite, of secondary albitization, although many others show ambiguous relationships; in some, primary magmatic albite has been inferred by some petrologists. The chemical analyses of spilites and keratophyres suggest that the change is mainly one of the addition of sodium to feldspar from outside the rock and the removal of feldspar calcium from the rock. Changes in apparent alumina and silica contents necessarily accompany this reaction. The content of K_2O is much more erratic than in unaltered rocks, which suggests that this component is highly mobile also.

ROCKS OF THE ALEUTIAN ISLANDS

Thick sections of metavolcanic rocks like those of the Riggins quadrangle and adjacent areas are commonly assumed to be derived from eugeosynclines associated with volcanic island arcs. As a particularly large number of chemical analyses are available for the Riggins rocks, it is appropriate to compare them with the rocks of an island arc. The well-known Aleutian Islands will be used here for such a comparison. Riggins rocks are chemically so similar to the submarine rocks of the Aleutian Islands that a common origin is indeed likely.

The Aleutian Islands have been studied by Geological Survey geologists. The petrologic and chemical data incorporated here have been taken from many published papers: Byers (1959); Coats (1952 and 1953); Coats, Lewis, Nelson, and Powers (1961); Drewes, Fraser, Snyder, and Barnett (1961); Fraser and Barnett (1959); Fraser and Snyder (1959); Nelson (1959); Powers, Coats, and Nelson (1960); and Simons and Mathewson (1955). Some analyses were made available prior to publication; the cooperation of

George L. Snyder and Ray E. Wilcox was particularly helpful. Fenner (1926) also presented many analyses.

The Aleutian Islands constitute a chain of exclusively igneous origin, being formed of lavas and tuffs, sediments derived directly from them, and dikes, sills, and plutons intrusive into them. Modern stratovolcanoes, many of them still active, rise above a foundation of older altered submarine rocks. The dominant rock is calcic two-pyroxene andesite and its altered equivalents. Basalt and dacite are widespread, but rhyodacite and quartz latite are uncommon. The average rock type is more silicic in the eastern part of the chain than in the west, and although the same rock types repeat throughout the chain, their variation patterns—aside from frequency distributions—are almost constant.

The impressive young volcanoes are of virtually unaltered subaerial flows and pyroclastics, perhaps entirely of Quaternary age. Also exposed on some of the islands are older subaerial rocks in remnants of volcanoes greatly reduced by erosion; these rocks are probably of both late Tertiary and early Pleistocene ages. The composition of these subaerial rocks is illustrated by the variation diagram of figure 65. The figure shows that most oxides vary in a pronounced linear fashion with varying content of SiO_2. Although the island chain is more than 1,000 miles long, it is a petrologic province of striking uniformity.

Also exposed on many of the islands are complexes of submarine lavas and tuffs, variably reworked by water, and associated sills and dikes. Pillow lavas are widespread. Several islands have such complexes of two distinct ages, but on most of the islands only one complex has been recognized. These submarine assemblages have yielded a fossil of probable Pennsylvanian age on one island, a fossil of probable Cretaceous age on another, and fossils from all parts of the Tertiary on various others. The complexes are not correlative from island to island, and the youngest ones probably formed at the same time as did the older subaerial rocks.

These submarine rocks have been slightly to moderately deformed, so that they are broken by normal faults, have gentle dips, and have in many places been altered to low-grade greenstone; greenschist and thrust faults and tight folds are not present. Compositionally, the rocks are andesite, basalt, dacite, and their albitized equivalents, keratophyre, spilite, and quartz keratophyre (fig. 66). Some of the rocks have almost the same compositions as the subaerial rocks, except for generally higher water contents, but most vary in a far less systematic fashion. Their oxides form broad zones on a variation diagram, which is in contrast to the narrow zones of the subaerial-rock oxides.

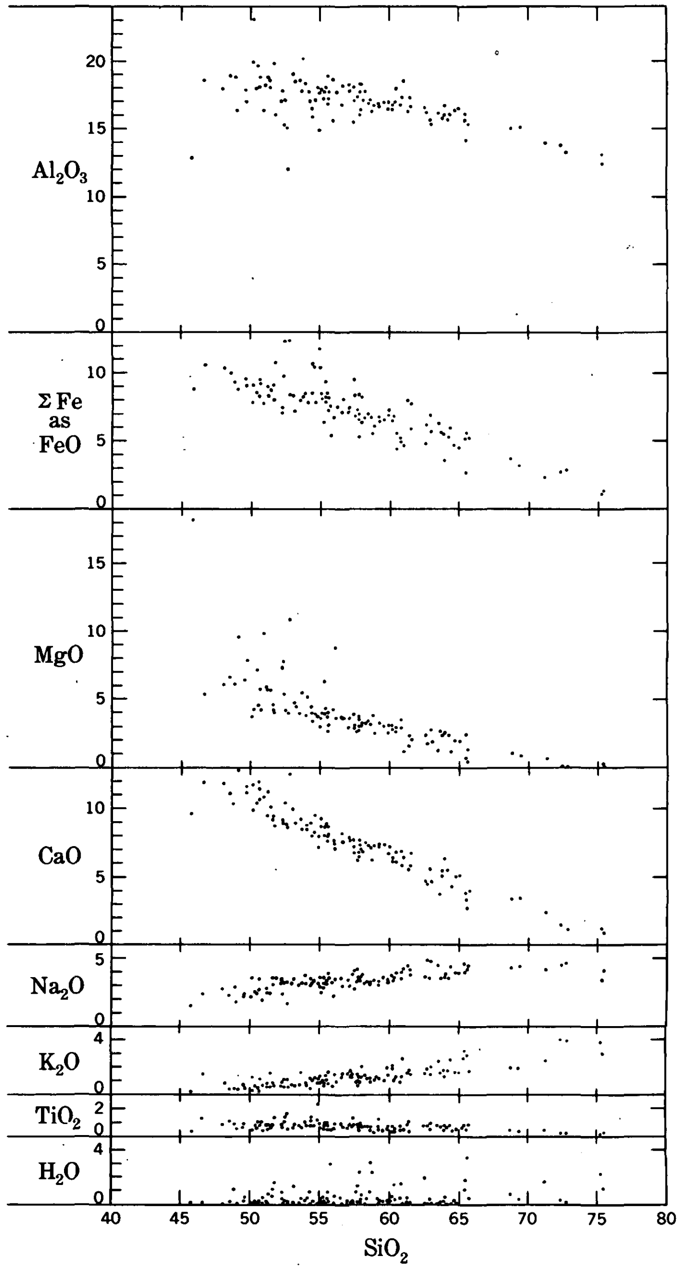

FIGURE 65.—Variation diagram of weight-percent analyses of subaerial volcanic rocks of Aleutian Islands.

In these submarine volcanic rocks, CaO varies downward from a maximum equal to the normal amount in calc-alkaline rocks, whereas Na_2O varies upward from a minimum equal to the normal amount. (Compare figures 65 and 66.) There is no break between andesite and keratophyre, for example, but instead there is a broad spectrum between calc-alkaline and albitic varieties. The relations between rock groups in the Aleutian Islands are illustrated schematically by figure 67. The young subaerial rocks belong entirely to the normal calc-alkaline series, whereas the older sub-

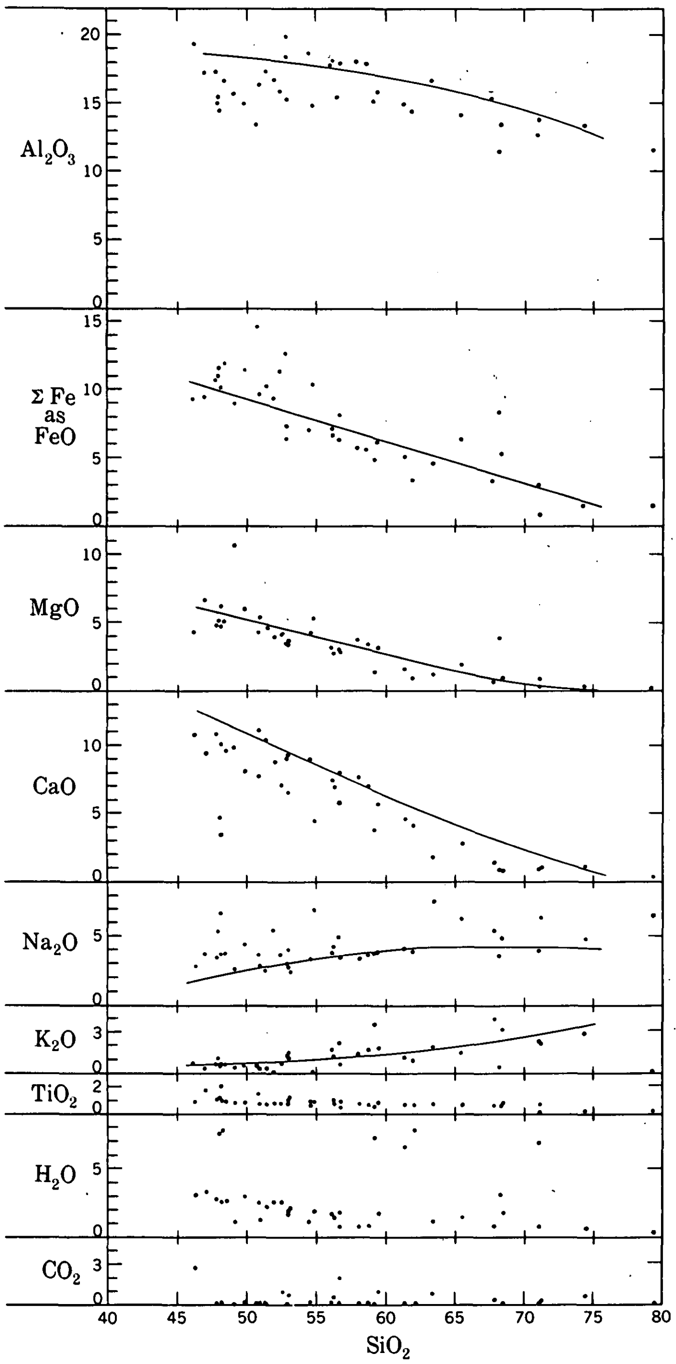

FIGURE 66.—Variation diagram of weight-percent analyses of submarine volcanic rocks and essentially contemporaneous intrusive rocks of Aleutian Islands. Lines indicate variation trends in subaerial rocks of the same province (fig. 65).

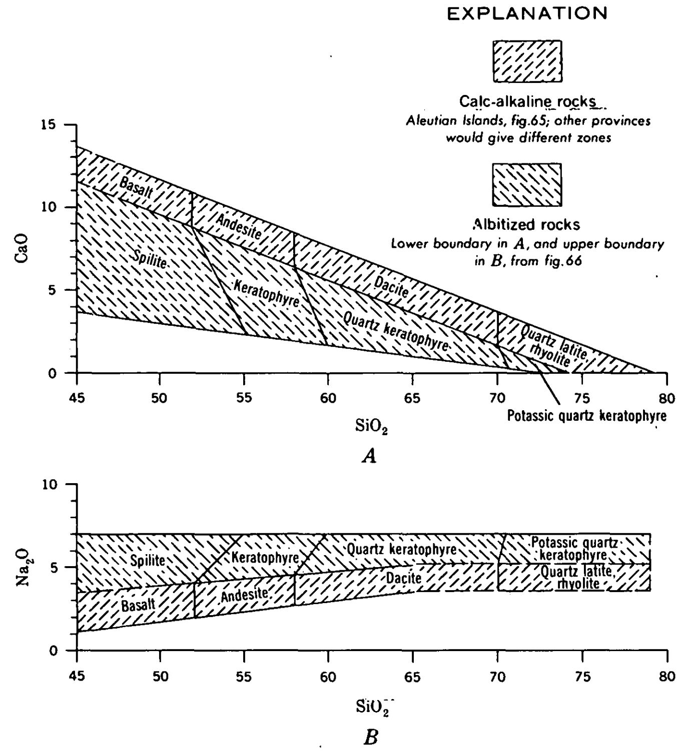

FIGURE 67.—Schematic representation of members of andesite-keratophyre association as indicated by weight-percents of CaO, Na_2O, and SiO_2. Rocks outside the shaded fields are rare. Actual boundaries between calc-alkaline types vary with contents of other components. Slanted boundaries between abitized types reflect correlative increase in SiO_2.

marine rocks belong to both calc-alkaline and albitized series.

The older complexes have been intruded by plutons of granitic rocks that have a variation pattern which differs from that of the subaerial volcanic rocks notably in higher content of potassium relative to other components. There is no tendency toward granitic rocks richer in sodium than the subaerial extrusives.

COMPARISON WITH RIGGINS ROCKS

The metavolcanic rocks of the Riggins quadrangle (fig. 43) seem identical chemically to the older, variably albitic Aleutian rocks in all major oxides. Even water content is the same, which suggests that the Riggins rocks formed in chemically closed systems—or that as much water was introduced as was removed—during their low- to middle-grade metamorphism. The higher CO_2 content in some of the Riggins rocks may reflect differences in sampling and may be due to the abundance of water-laid reworked tuffs in the Riggins analyses and their scarcity in the Aleutian analyses, as such rocks were deliberately avoided in the Aleutian sampling.

Andesitic associations of areas other than island arcs—for example, the Cascade Range of Washington, Oregon, and northern California—are similar to the subaerial volcanic rocks of the Aleutians in the general linear character of the variations of their components with SiO_2 and correspondingly dissimilar to the Aleutian submarine rocks and to the metavolcanic rocks of the Riggins quadrangle.

The detailed similarities between Riggins and Aleutians chemistry give strong support to the common assumption that the volcanic rocks of ancient eugeosynclines formed from island-arc volcanoes.

ORIGIN OF KERATOPHYRE

Eugeosynclinal assemblages containing spilite and keratophyre also have basalt, andesite, and all transitions between calc-alkaline and sodic rocks. Such transitions are obvious in the Riggins and Aleutian data and have been emphasized by many authors (for example, Gilluly, 1935, p. 249–250). A satisfactory explanation of the origin of the sodic rocks must account also for the presence of calc-alkaline and transitional rocks.

The erratic variations of rock compositions in the andesite-keratophyre association are completely unlike the systematic variations of practically all other suites of igneous rocks and suggest secondary alteration rather than magmatic variability as the major cause of keratophyric variants.

The Aleutian island arc has apparently been the site of continuous volcanism at least since Late Cretaceous time. Subaerial volcanic rocks are of normal calc-alkaline compositions, whereas the submarine-rocks—probably overlapping them in time—belong to the andesite-keratophyre association. It would scarcely be reasonable to postulate that erratically variable magmas were erupted throughout the history of the arc, until at some time—different on different islands—systematically variable calc-alkaline magmas suddenly began to form. Rather, the data force the conclusion that all the Aleutian lavas were erupted as calc-alkaline magmas and that the albitized rocks owe their distinctive character to posteruptive transformations. By analogy, it seems that the keratophyre and spilite of the Riggins group formed similarly by alteration of rocks that crystallized from calc-alkaline magmas.

There are many other arguments against the widely held view that spilite and keratophyre form from odd sodic magmas. For example, Wilcox (1959) stressed that the mafic minerals of the unmetamorphosed spilite and keratophyre of the Aleutian Islands are those of normal calc-alkaline rocks instead of high-Na_2O, low-CaO minerals such as would certainly have formed had the now-sodic rocks crystallized from sodic magmas.

There are many variants of the hypotheses of magmatic origin of the sodium of spilites. Thus, Bartrum (1936, p. 421–422) postulated direct crystallization of spilite and keratophyre from magmas and concluded that some New Zealand spilite formed from primary sodic magma, but he had little evidence except the notion that "* * * it is indeed difficult to imagine how such [late sodic] solutions could have traversed such a fine-grained rock."

Nicholls (1959) suggested that spilite in the Ordovician of Wales formed by alteration by residual immiscible sodium-rich magmatic solutions. Such a scheme encounters most of the same objections as does one of primary solidification from a sodic magna.

Kuznetsov (1956, p. 137) postulated that the excess sodium of spilite was provided by thermal alkali springs on the sea floor. He presented no evidence for this, and as alkali springs are inconspicuous on the modern active volcanoes the hypothesis has little apparent merit.

All these variants seem invalid, except possibly for isolated examples, because they do not explain broad relations such as those prevailing in the Aleutian Islands.

The extreme mobility of the alkalis in quartz keratophyre was demonstrated by Battey (1955), who presented seven analyses (for Na_2O and K_2O only) from the lower 5 feet of one flow of quartz keratophyre. In these analyses, K_2O ranges irregularly from 1.3 to 5.6 weight percent, and Na_2O, in complementary fashion, ranges from 6.6 to 2.2 percent; the sum of Na_2O+K_2O remains near 8 percent throughout. Average values of Na_2O and K_2O are 4.7 and 2.9 percent, respectively. Complete analyses of other samples from the same area show that the suite is dominated by highly silicic quartz keratophyre (SiO_2, 70 to 77 percent; average contents of CaO, Na_2O, and K_2O are 0.6, 3.8, and 5.3 percent, respectively). Both K_2O and Na_2O were mobile, and each replaced the other, as shown by much textural evidence. Battey found that the frequency distribution of abundances of the alkalis was symmetrical to that of normal rhyolite; he concluded that no introduction of alkalis was indicated and that redistribution of alkalis within flows was the probable mechanism by which the variation was produced. This may be valid for the area considered, but Battey's (1955, p. 123–124) extrapolation to the conclusion that quartz keratophyre in general represents products of low-grade metamorphism of normal rhyolite is untenable. There is true enrichment in sodium and removal of calcium in many areas, including the Aleutian Islands and the Riggins quadrangle, in which the average amounts of Na_2O and CaO diverge markedly from those of calc-alkaline rocks.

In many regions where andesite and keratophyre have been metamorphosed, highly deformed, and invaded by plutons of granitic rocks, trondhjemite is a common plutonic rock type. Trondhjemite is widespread in the Riggins quadrangle and is present in many areas in eastern Oregon, in the Klamath Mountains, and in the northwestern Sierra Nevada. Because of this areal relationship in eastern Oregon, Gilluly (1935, p. 347) suggested that the quartz keratophyre was a product of crystallization of trondhjemitic magma, whereas the spilite and keratophyre were albitized by sodic solutions of partly magmatic origin. The trondhjemitic magma, he suggested, was derived in turn by the modification of granitic magma by interaction with the wet sediments it traversed. Later, while working in a different area in which there is no trondhjemite, Gilluly (1948, p. 24) concluded that albitization of andesite there occurred in two stages—one before, and unrelated to, the intrusion of granitic rocks; the other related to a stock of quartz monzonite.

Waters (1955, p. 713) suggested that albitizing solutions are derived partly from wet sediments and are liberated partly from metamorphosing rocks at depth.

Watery fluids expelled from the metamorphosing tectogene * * * pass through the overlying rocks converting basalts to greenstones or "spilites," propylitizing and "hydrothermally" altering andesites, and even causing extensive albitization and other alterations in sedimentary rocks.

Some important features are difficult to explain in terms of this theory. There is no direct evidence for the postulated deep-seated metamorphism in the Aleutian Islands, or in the Cascade Range with which Waters was particularly concerned; Waters (written communication) himself states that most of the zeolitization and albitization of the subaerial andesites of Tertiary age in the Cascades can be explained by local reconstitution of materials already present, without need for introduction of material.

There is no known upward or outward zoning of albitization, nor "front" of calcium enrichment beyond, in Oregon and western Idaho; the incomplete evidence suggests that the metavolcanic rocks were everywhere albitized to a comparable extent. Nor do the chemical analyses of variably metamorphosed rocks in the Riggins quadrangle support Water's theory. For example, the two analyses of meta-agglomerate of the Lightning Creek Schist probably represent the same unit at very different grades of metamorphism, and yet the analysis of the high-grade gneissic rock (No. 15, table 5) shows the same contents of sodium and calcium as does the analysis of the low-grade greenschist (No. 18). There is no suggestion that sodium was driven out of the high-grade rocks and deposited in the low-grade ones.

At least the great bulk of spilite and keratophyre is in submarine sequences. They are the characteristic rocks of eugeosynclines. Marine sediments and fossils are of variable abundance, and water-laid tuffs are widespread. Spilite and pillow lava are synonyms in much of the older literature. (Backlund, 1930, p. 37–40, concluded that the spilite of Novaya-Zemlya is subaerial, but his evidence is unconvincing.)

Conversely, nearly all the uncommon "keratophyre" in subaerial sequences is sodic trachyte, labeled "keratophyre" by the unfortunate alternate use of the term. The subaerial "keratophyre" of Wales described by Thomas (1911) is a member of an assemblage of olivine basalt, olivine diabase, mugearite, and sodic felsic rocks—a common alkaline differentiation assemblage of the continents, unlike the andesite-keratophyre association. Knopf's (1921, p. 368) Nevada keratophyre of Tertiary age contains albite phenocrysts in a glassy-spherulitic groundmass with a little biotite and is no kin to the rocks of the andesite-keratophyre association. The sodic rocks of the French Auvergne have been referred to as "keratophyre", but again these are very different from island-arc rocks.

The distinctive compositions of spilite and keratophyre must be due to postmagmatic processes, and these processes seem somehow connected to the submarine environment in which these rocks almost exclusively occur. That diabase sills and other rocks intruded within the sequences are also albitized shows that the changes are not restricted to rocks that formed in direct contact with sea water. Two obvious alternatives have long been proposed: either that water carrying sodium was forced out of wet rocks at depth, perhaps during metamorphism, and streamed upward through the volcanic pile, depositing sodium and removing calcium; or that the sodium was introduced, and calcium removed, by the sea water in some sort of diagenetic process. The first possibility was considered and rejected earlier.

The idea that sea-water sodium is exchanged for magmatic calcium is an older one. Daly (1914, p. 338–340) advanced the theory, and others have favored it since, although it has not been a popular notion. Gilluly (1935), Turner and Verhoogen (1951, p. 210–212), and Waters (1955) called upon sea-water sodium as, at most, a subordinate source of albitizing material.

The scatter of points on the variation diagrams of figures 43 and 66 suggests that the average increase in Na_2O content caused by the variable spilitization of the thousands of cubic miles of andesite-keratophyre-association rocks in a province is about 1 percent of the weight of the total rock. Mechanisms requiring the entrapment of normal sea water in porous tuffs and its subsequent reaction with the solid rock, perhaps at a time when temperatures are raised by intrusion or during regional metamorphism, seem utterly inadequate, as the volume of trapped water would have to be at least twice the volume of the volcanic rocks. (Should the salt be trapped as a concentrated brine as the result of boiling away of most of the water, a process for which no evidence has been adduced, such a mechanism might function.)

Mineralogic capture of sea-water sodium seems more likely. If bases in the volcanic rocks—and particularly in the porous and already partly water-laid tuffs—could be exchanged for bases in circulating water and the capture thus be diagenetic, then subsequent albitization could be explained easily.

Eskola, Vuoristo, and Rankama (1937) demonstrated in the laboratory that calcic feldspar can be converted to albite by the attack of sodium carbonate, but although they termed the change the "spilite reaction," it is only an assumption that sodium carbonate is involved in natural spilitization.

Zeolites are more plausibly called upon. W. Bradley Myers has suggested during discussions that calcium zeolites might form in the tuffs by direct hydration of the feldspars and that such zeolites might in turn exchange bases with brines to form sodic zeolites, just as zeolite water softeners are recharged. Ray E. Wilcox (oral communication, 1960) has found analcite replacing plagioclase, but not pyroxene, in tuff that is intermediate in composition between normal basalt and spilite in the Aleutian Islands. Wilcox suggested that calcic plagioclase is altered directly to analcite by hydration and the substitution of sea-water sodium for calcium and that the analcite is subsequently silicated and dehydrated to albite.

Gulbrandsen and Cressman (1960) described a thin but widespread bed of altered tuff of quartz-keratophyre composition in a limestone of Jurassic age in southeast Idaho and adjacent Wyoming. The high content of sodium in this tuff is in microcrystalline albite at some localities and in analcite at others. Analcite and albite both replace shards, hence both formed after deposition of the pyroclastic materials. It is not possible here that the excess sodium of the tuff came from late magmatic or hydrothermal solutions, and it is unlikely that the original magma was richly sodic; so Gulbrandsen and Cressman (p. 462) concluded that the excess sodium probably came from sea water. In their view, "analcime is the product of a reaction between volcanic glass and sea water (possibly highly saline) at some stage of diagenesis," and the albite probably resulted from replacement of analcite at higher temperatures and pressures after deep burial.

Zeolitized low-grade marine metasedimentary rocks of volcanic origin were found in New Zealand by Coombs (1954). Although Turner, for example (*in* Fyfe, Turner, and Verhoogen, 1958, p. 215–217), considers such zeolitization to be the incipient stage of regional metamorphism, it seems likely that it instead records incomplete spilitization through a base-exchange mechanism such as is postulated here.

Spilite and keratophyre were formed widely in the submarine Miocene and Pliocene sequence of Sakhalin in a setting similar to that of the Aleutian Islands. Shilov and Kalishevich (1958) stated that lavas erupted under shallow sea water—less than 100 meters—showed no chemical or petrographic evidence of albitization by sea water, perhaps due to "* * * the formation of a gaseous envelope around the lavas which inhibited the diffusion of ions." They concluded that "* * * subaqueous eruption in itself is not sufficient for the formation of the spilite-keratophyre association; the depth of the basin on whose bottom the lavas were erupted must exceed 100 m." Evidence was not given.

More work on the secondary minerals of nonmetamorphosed spilite and keratophyre is particularly needed before the problem of their origin can be solved, but theories involving exchange with sea water seem most promising. However the sodium is introduced into the rocks, it is highly mobile at the albitizing stage; dense sills are as sodic as are tuffs.

METAMORPHISM

The greenstones, schists, and other metamorphic rocks of the Riggins quadrangle increase in metamorphic grade eastward. The gradual change from low- to high-grade rocks takes place over a distance of 10 miles across the strike of the rocks. It is many miles farther, through a terrane of increasingly migmatitic gneisses and mostly concordant granitic intrusives, to the massive interior of the Idaho batholith.

The following discussion is concerned primarily with the petrologic aspects of the metamorphism of the nonplutonic rocks. The metamorphic rocks grade into the plutonic rocks, but the plutonic rocks are distinguished by the abundant granitic material intrusive into them, by their metasomatism, and by their much greater structural mobility.

Of main interest here is the progressive regional metamorphism of chiefly volcanic rocks, principally by isochemical processes that responded to changing conditions of heat, pressure, and shearing stress. None of the rocks is rich in either potassium or aluminum, so the mineral assemblages common in metapelitic schists are not present. The zones of metamorphism recognized in the Scottish Highlands, for example, cannot be applied here because of the lack of aluminian index minerals. Many minerals of the Riggins quadrangle have stability ranges, in terms of associated minerals indicative of metamorphic grade, unlike those common in other regions. The character of the metamorphic progression here will be established, and then this progression will be compared with equivalent, but different, progressions elsewhere.

PROGRESSION OF METAMORPHISM IN RIGGINS QUADRANGLE

Variables of progressive metamorphism are illustrated in five maps (figs. 68 to 72) of the northwestern part of the quadrangle.

Figure 68 records the westernmost—lowest grade—occurrence of the index minerals, which appear in the order, going toward higher grade rocks, of aluminian prochlorite, biotite, clinozoisite, garnet, oligoclase, andesine, and clinopyroxene. None of these minerals was found beneath the Rapid River thrust within the quadrangle. All isograds from aluminian prochlorite through andesine in the upper plate are truncated by the thrust. Although distances between the isograds vary, their order does not. Biotite and garnet isograds are ½ mile to 3 miles apart; garnet and oligoclase, as much as 1 mile apart; and oligoclase and andesine ½ mile to 1½ miles apart. Biotite and andesine isograds—which span the greater part of the mineralogic progression—are 2 to 4 miles apart. Clinozoisite (as opposed to strongly pleochroic epidote) marks an isograd of intermediate value between biotite and garnet, although this isograd is not recognizable in the calcium-poor Fiddle Creek Schist.

Each of these index minerals was stable into zones defined by higher isograds. Biotite, garnet, oligoclase, and andesine are present in higher grade rocks throughout the quadrangle. Clinozoisite is present in many medium-grade rocks and some high-grade ones. Clinopyroxene is restricted to isolated localities of high-grade rocks of special composition, so it is not regionally useful as an indicator.

Distribution and type of prochlorite (fig. 69) correspond roughly with the isograd system. The first appearance of aluminian prochlorite marks a slightly lower grade of metamorphism than does that of biotite. Aluminian prochlorite persists in rocks of suitable composition to about the andesine isograd, whereas ferroan prochlorite is stable only into the garnet zone. Colorless chlorite is present also in the Lucile Slate of the quadrangle; if this is aluminian prochlorite, then the aluminian-prochlorite isograd (fig. 68) should appear beneath the Rapid River thrust also. As such chlorite was not found in the Seven Devils Volcanics, its development may have compositional, as well as grade, significance.

Distribution and type of epidote and clinozoisite also have grade significance (fig. 70). Only epidote, most of which is strongly pleochroic, occurs west of the clinozoisite isograd. Clinozoisite or epidote, the latter generally more aluminian (that is, colorless instead of strongly pleochroic in thin section) than that in the low-grade rocks, persist to about the andesine isograd in rocks of volcanic composition, and beyond, past the clinopyroxene isograd, in calc-silicate rocks. The greater the content of calcium available for plagioclase, the farther clinozoisite and epidote persist. The anorthite content of plagioclase is a function of metamorphic grade and the availability of calcium. Many of the rocks have such low $CaO:Na_2O$ ratios that plagioclase more calcic than oligoclase cannot form in them, and such rocks contain no epidote or clinozoisite even where intercalated with more calcic rocks that contain andesine and clinozoisite together.

Only the more potassic and silicic of the middle-grade rocks contain muscovite, so the upper-grade limit of its distribution (fig. 70) is controlled by composition as well as metamorphic grade. Muscovite is generally lacking beyond the andesine isograd, but it does occur in higher grade rocks of suitable composition.

The color of amphibole varies with metamorphic grade and rock composition (fig. 71). No amphibole was seen in the lower grade parts of the Seven Devils Volcanics, but this observation may reflect inadequate sampling rather than metamorphic grade, as only a few samples were collected. In the higher grade part of that formation, and in rocks lower in grade than the garnet isograd in the Riggins Group, the amphibole is green in hand specimen and light colored, greenish yellow, in thin section. Such amphibole of the Riggins Group is at least partly hornblende, although presumably that of the Seven Devils Volcanics is actinolite. Higher grade hornblende is black in hand specimen and either light or dark in thin section; the darker is generally the higher grade, but a rock-composition control is superimposed here so that the line between light and dark hornblende, as drawn, crosses other isograds. (More detailed sampling would likely show a correspondence of this hornblende-division line with contacts between rocks of different compositions.)

Several other features illustrative of metamorphic progression are shown in figure 72. Dikes of aplite and pegmatite extend about 1 mile farther westward into lower grade rocks than do dikes of granitic rocks (mostly trondhjemite). The west limit of dikes of either type varies from near the garnet isograd to near the andesine one. The distinctive dark carbonaceous phyllite of the Squaw Creek Schist, abundant in its low-grade parts, extends eastward to about the west limit of aplite and pegmatite.

The west limit of continuous gneisses (fig. 72) is defined easily only at the north end, where trondhjemite gneiss rests with intrusive contact against middle-grade schists, and at the south end, where trondhjemite and amphibolite are in thrust contact above low- and middle-grade schists. In between, there is no lithologic contact, but only a zone of broad gradation within which the rocks become more foliated and coarser grained eastward. There is, however, a structural discordance, rough in some places and conspicuous in

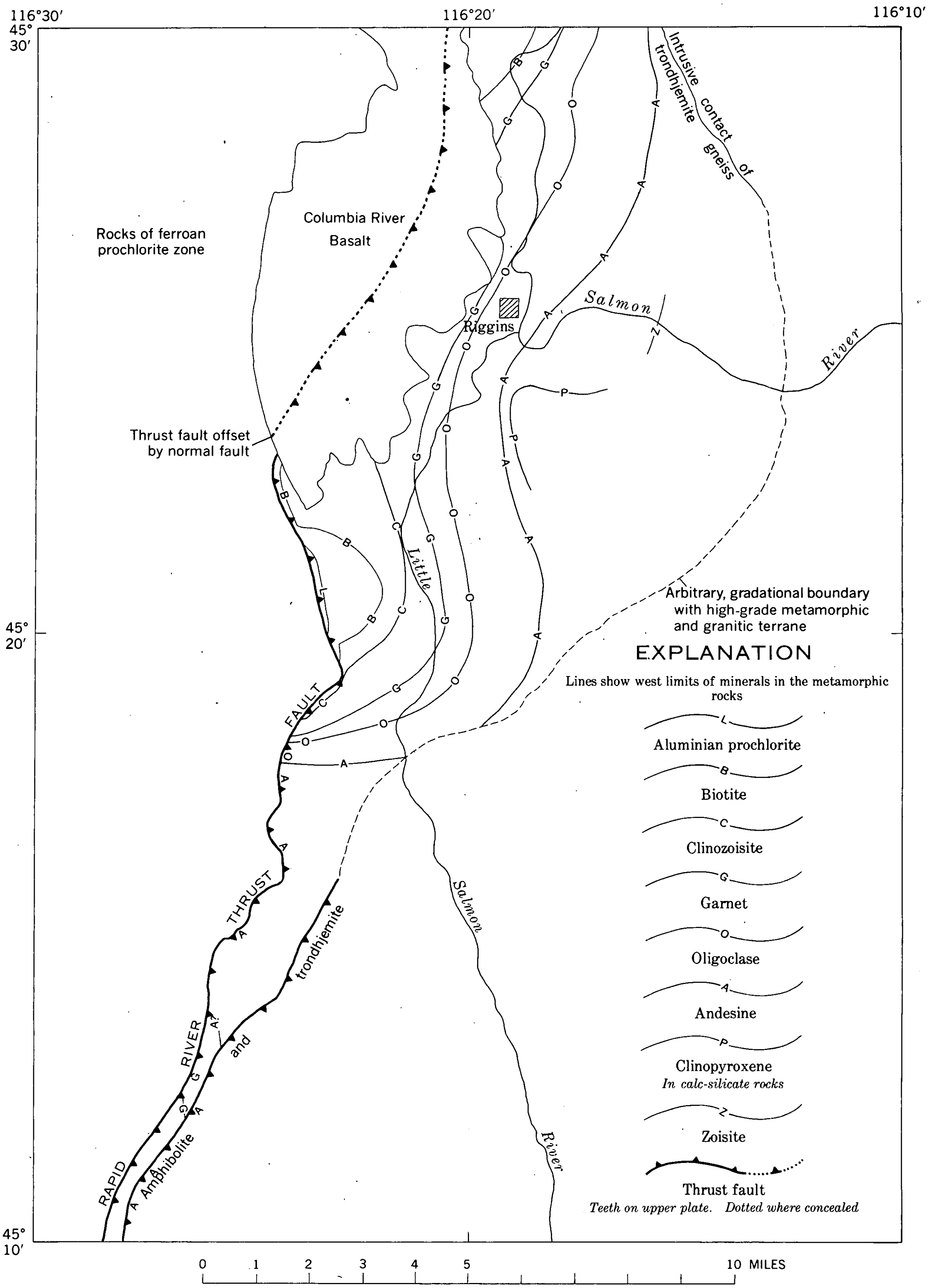

FIGURE 68.—Thrust faults and isograds in metamorphic rocks of northwestern part of Riggins quadrangle, Idaho.

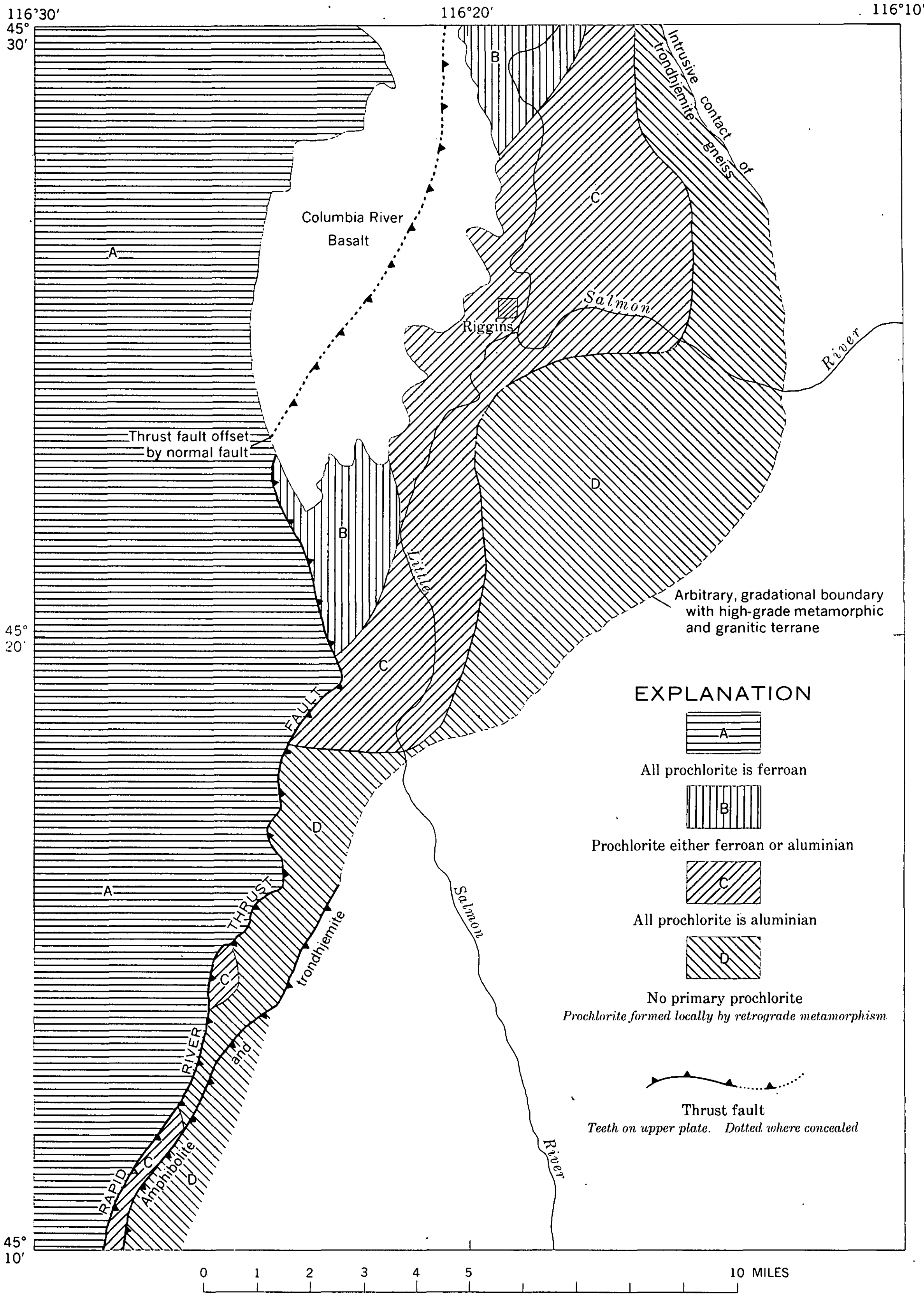

FIGURE 69.—Distribution of ferroan and aluminian prochlorite in metamorphic rocks of northwestern part of Riggins quadrangle.

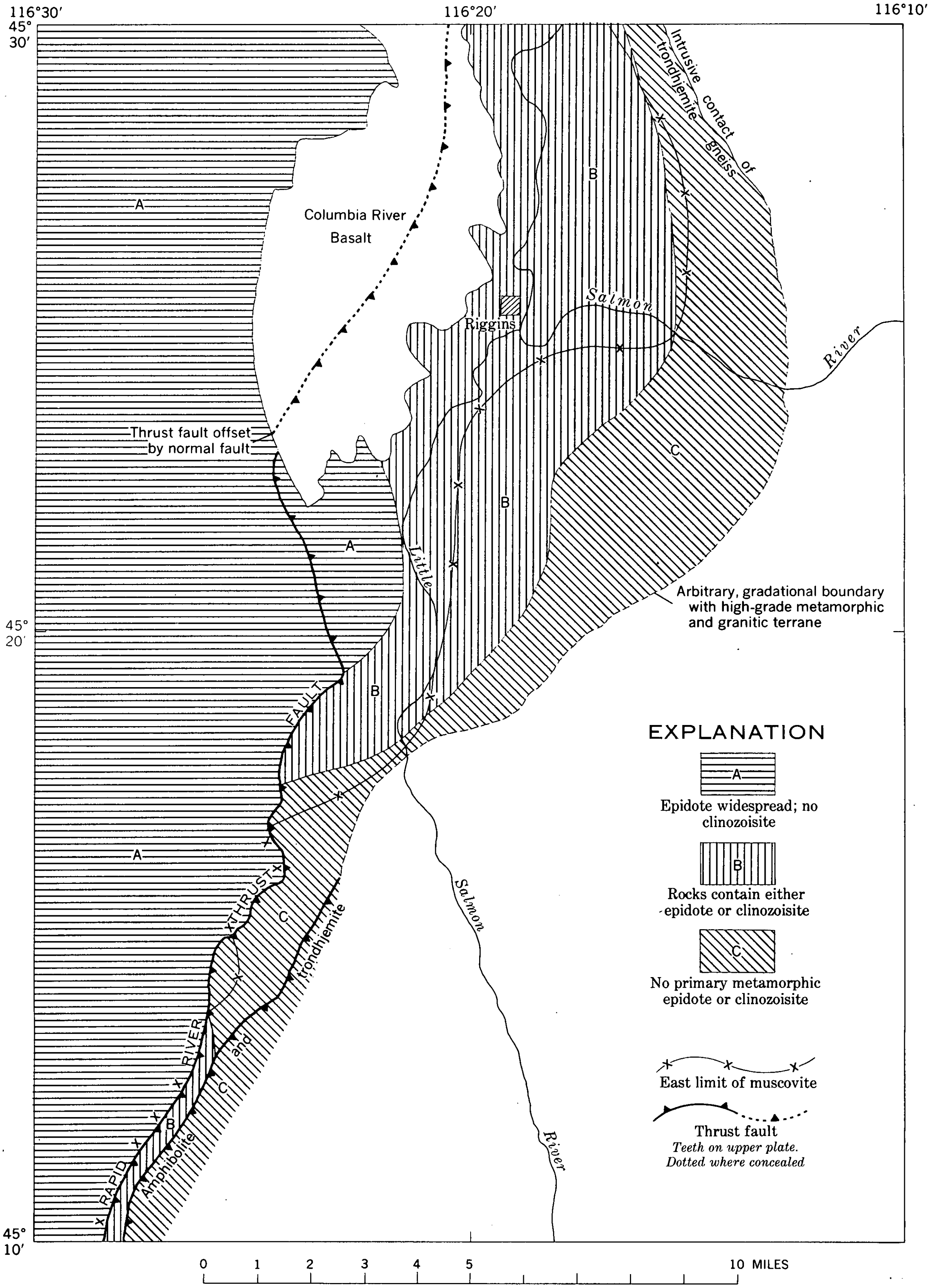

FIGURE 70.—Distribution of epidote, clinozoisite, and muscovite in metamorphic rocks of northwestern part of Riggins quadrangle.

others, between the variably contorted rocks to the west and the much more uniformly flow-structured rocks to the east. The zone of this discordance is a fair approximation of the northwest limit of continuous gneisses and is so used in the diagram. The contact of the gneisses, so defined, is not concentric with either the isograds or the limits of dikes in the schists.

Correspondence among the mineralogic zones, Eskola's metamorphic facies, and common rock types is indicated by table 8.

Stability ranges of the important metamorphic silicate minerals are shown in figure 73. (Quartz, carbonates, minor accessory minerals, and products of retrogression are omitted.) Not all minerals in any vertical line occur together. Thus, albite persists into the oligoclase zone, or oligoclase into the andesine zone, only in the absence of calcic minerals such as the epidotes. Nevertheless, prochlorite and muscovite coexist in what seem to be equilibrium relationships in a number of rocks that contain garnet, oligoclase, or other minerals indicative of middle-grade metamorphism, whereas reconstitution of these minerals to biotite might be expected instead.

TABLE 8.—*Metamorphic facies and zones and typical metavolcanic rock types in the Riggins quadrangle*

[Facies nomenclature after Eskola]

Facies	Zone	Typical metavolcanic rock
Greenschist	Ferroan prochlorite	Greenstone
	Aluminian prochlorite	Green phyllite
	Biotite	Greenschist
	Clinozoisite	
Epidote amphibolite.	Garnet	
Amphibolite	Oligoclase	Hornblende schist
	Andesine	
	(Clinopyroxene)	Amphibolite

BEHAVIOR OF ELEMENTS DURING METAMORPHISM

The metamorphism was principally isochemical. Carbonates and silica were clearly mobile, as shown by their widespread occurrence in veins, but neither chemical nor field data suggest major migration of components other than water and carbon dioxide into or out of any part of the metamorphic terrane during metamorphism. Dikes of aplite, pegmatite, and granitic rocks become progressively more common in rocks of progressively higher grade, but contacts of these dikes are typically sharp, with little or no evidence for introduction of, say alkalis into the wallrocks. Nonvolatile bulk compositions of the rocks were changed little, if any, during metamorphism.

Sodium is present chiefly in albite in the low-grade rocks and in oligoclase or andesine in the high-grade ones, but probably hornblende contains progressively more of this element as its metamorphic grade increases.

Calcium is largely in epidotes in low-grade rocks and in oligoclase or andesine in higher grade ones. The calcium content of amphibole probably decreases with increasing metamorphism, but the total amount of amphibole commonly increases.

Aluminum is mostly in prochlorite in the low-grade rocks. The aluminum content of that prochlorite is higher in middle-grade rocks than in low-grade rocks; but the total amount of prochlorite, and hence the amount of included aluminum, decreases beyond about the garnet isograd, the released aluminum going into hornblende and plagioclase and, to a lesser extent, into biotite. Aluminum in epidote-group minerals of the low- and middle-grade rocks is probably largely transferred to plagioclase with increasing metamorphic grade.

The proportion of ferric to ferrous iron decreases markedly between the very low grade rocks and the low-grade ones. Hematite, magnetite, and the iron content of epidote decrease correspondingly. Ferrous iron is chiefly in prochlorite in the low-grade rocks and in hornblende in the high-grade ones.

The magnesium content of prochlorite remains nearly constant with increasing metamorphism, but as the total amount of prochlorite decreases with increasing grade, the total magnesium so contained decreases also. The magnesium content of amphibole decreases moderately with progressive metamorphism, but the total amount of amphibole increases in greater proportion, so that magnesium liberated from prochlorite is largely absorbed in hornblende. Micas are abundant only in the potassic rocks.

Potassium is only a minor component of most of the rocks. It is present obviously in the micas, and presumably it becomes increasingly abundant in hornblende as metamorphic grade increases. There is no potassic feldspar in the metamorphic rocks.

ORIGIN OF LAMINAE

The rocks of the Squaw Creek Schist are striped by compositional laminae in all metamorphic grades (figs. 23, 33, and 65). Most outcrops contain laminae of many widely divergent compositions. Such lamination must be largely inherited from sedimentary layering, even through the succession of laminae in any outcrop may reflect isoclinal deformation and pervasive shearing as much as it does sedimentation.

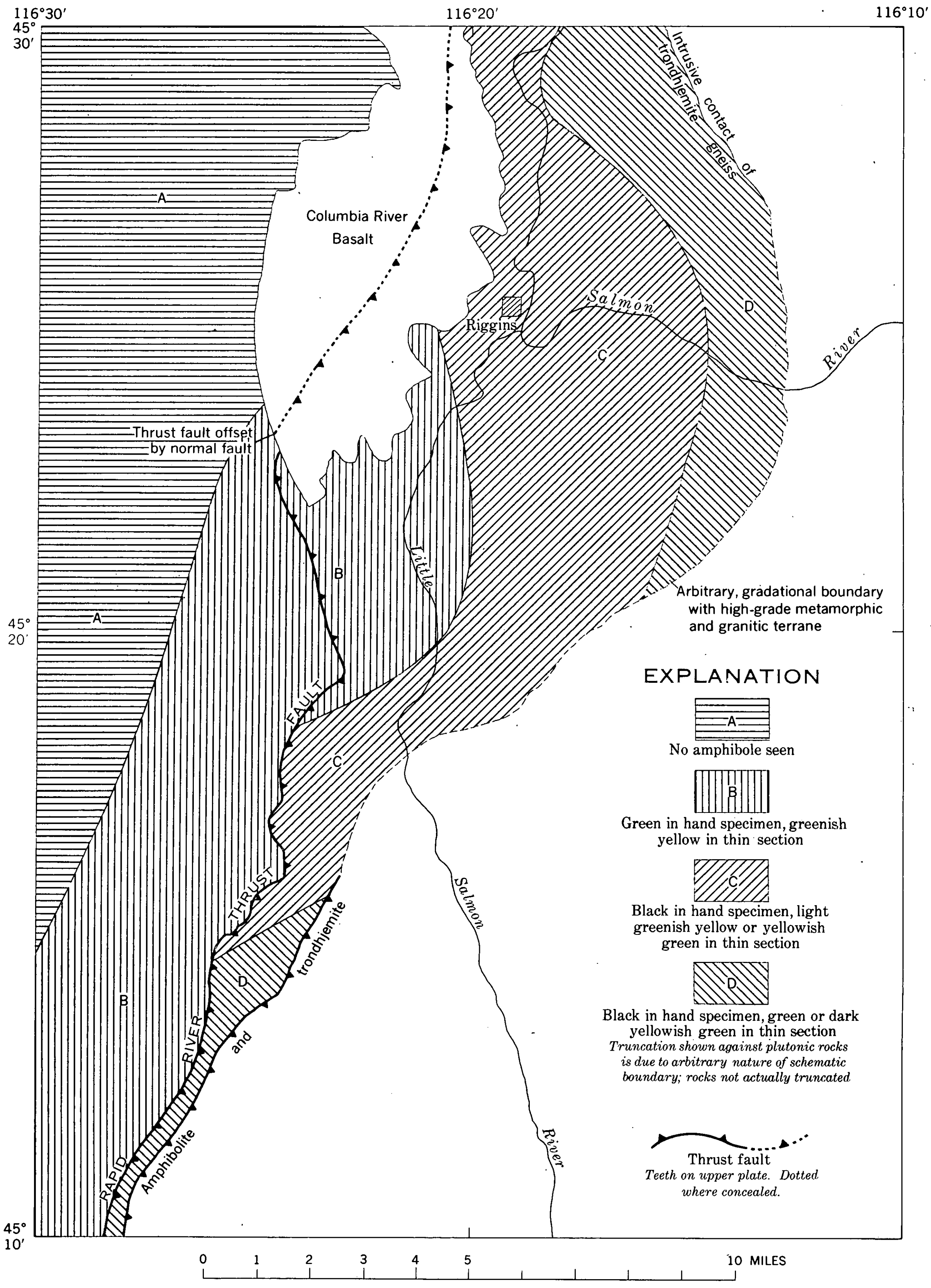

FIGURE 71.—Color of amphibole in metamorphic rocks of northwestern part of Riggins quadrangle.

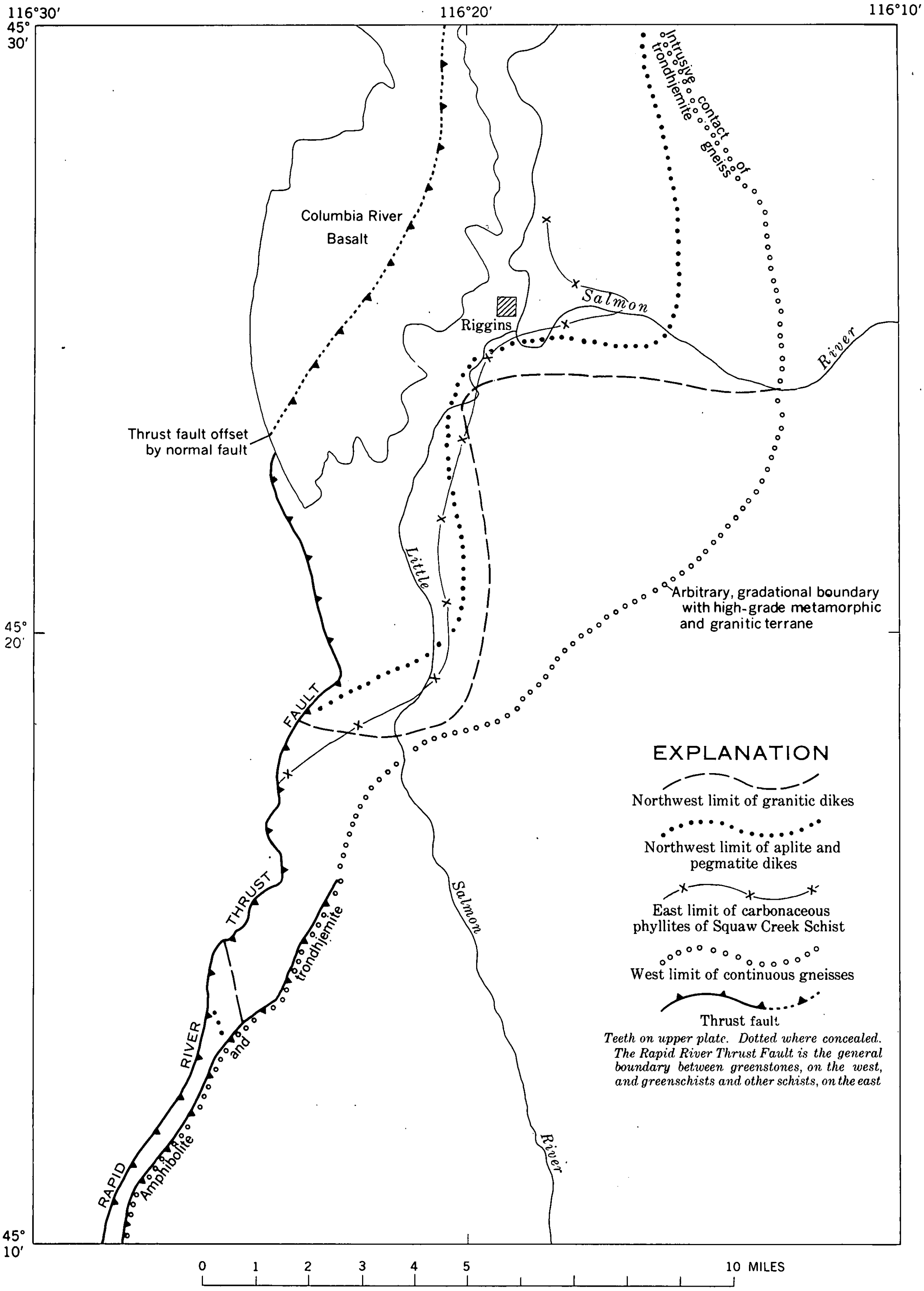

FIGURE 72.—Limits of dikes and some metamorphic rock types in the northwestern part of the Riggins quadrangle.

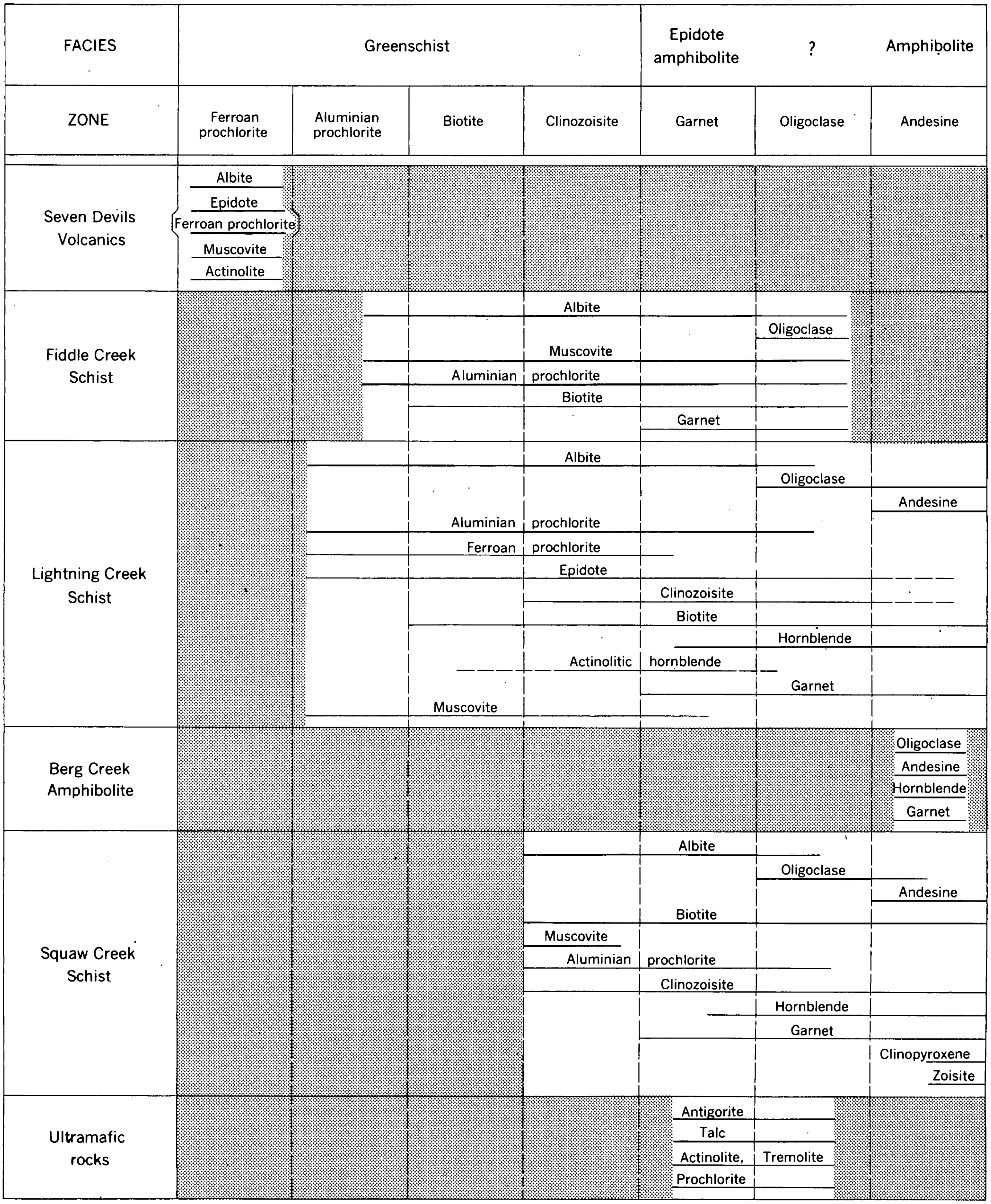

FIGURE 73.—Stability ranges of silicate minerals in metamorphic rocks of Riggins quadrangle. Thick lines, minerals widespread and commonly abundant; thin lines, minerals of infrequent occurrence or minor abundance; shaded areas, no samples studied. Clinozoisite zone not recognizable in Fiddle Creek Schist.

Rocks of the other formations are mostly nonlaminated in zones of lower grade metamorphism yet progressively more laminated in higher grade zones. In most of the schists, laminae are a few millimeters thick and are rich alternately in felsic minerals (mostly quartz and plagioclase) and in mafic minerals (chlorite, mica, and hornblende). The most strikingly laminated higher grade rocks are those of the Berg Creek Amphibolite (fig. 22), the laminae of which are rich alternately in plagioclase and in hornblende. Lamination of these types must be produced largely by segregation of components during metamorphism. Mineral orientations shown that pervasive shearing parallel to the laminae accompanied the reconstitution, and it is likely that the segregation reflects the varying shearing, translation, and recrystallization characteristics of the minerals involved. Presumably shearing displacement along mafic folia was greater than along felsic ones, whereas the chemical mobility of quartz and sodic plagioclase was probably greater than that of the mafic minerals.

RELATIVE TIME OF SHEARING

The strong foliation of all the metamorphic rocks except some greenstone of the Seven Devils Volcanics attests the importance of shearing during metamorphism. That this shearing was broadly synchronous with reconstitution is shown by the marked orientation of inequidimensional minerals. In some places, shearing continued after reconstitution had effectively stopped (garnets were rolled, and new cleavages formed, for examples). Elsewhere, the reverse was true, and the last-crystallized minerals are crosscutting porphyroblasts.

Throughout most of the Riggins Group, most of the foliation of the schists is parallel to compositional layers. In the Seven Devils Volcanics, by contrast, cleavage is markedly crosscutting in most of the few places in which both cleavage and bedding were observed. It is a reasonable inference that the cleavage of the Seven Devils Volcanics lies, in general, near the axial planes of nonisoclinal folds, as such relations have been demonstrated in many regions.

Had the schists of the Riggins Group been deformed by such a system of folds and cleavage, then one would expect that formation boundaries would be highly irregular relative to cleavage directions. This is conspicuously not so, and it follows that the schists of the Riggins Group did not pass through a similar stage of deformation. In the schists, the direction of shearing was generally controlled by the direction of compositional layering. It is possible that shearing in the schists did not progress far, or did not begin, until after the rocks had been heated enough so that shearing was controlled primarily by compositional variations.

AGE OF METAMORPHISM

The metamorphic grade of the Riggins Group increases eastward and southward toward the Idaho batholith of Cretaceous age. This metamorphism was superimposed upon an older post-Triassic metamorphism of the Seven Devils Volcanics.

METAMORPHISM OF RIGGINS GROUP

Metamorphism of the Riggins Group can be dated directly only as older than the Idaho batholith. Across a narrow zone that crosses the Salmon River at Lake Creek and which is arbitrarily taken as the boundary between the metamorphic and plutonic rocks despite the metamorphic progression and lithologic correlation across the boundary, the irregular, complex structures that characterize the metamorphic rocks give way to the simple north strikes and steep dips of the plutonic rocks. This change is clearly due to the much greater mobility of the plutonic rocks and to the laminar shear by which they deformed and flowed into position. As the simple plutonic structures were superimposed on the complex metamorphic ones, they must be at least slightly younger. The quartz dioritic gneisses, which are the characteristic granitic rocks near this zone, were themselves intruded and migmatized by quartz monzonite farther east and by trondhjemite farther south; and the resultant migmatites were in turn intruded by the massive plutons of the Idaho batholith still farther east and south, beyond the Riggins quadrangle. Plutons in the interior of the batholith have been dated radiometrically as of about middle Cretaceous age (Jaffe, Gottfried, Waring, and Worthing, 1959, p. 92–97). The main metamorphism of the Riggins Group is thus not younger than middle Cretaceous.

As no metamorphic rocks above the Rapid River thrust have been dated, a maximum age for the metamorphism cannot be established directly. As is brought out below, however, the metamorphism is probably younger than both the Martin Bridge Limestone of Late Triassic age and the granitic rocks of likely Late Jurassic age, so that it can be more closely dated by inference. The metamorphism is probably of Early Cretaceous age.

The Rapid River thrust cut gneisses and trondhjemite in the thrust sheet in the southern part of the quadrangle and produced a broad zone of phyllonite and retrograded rocks from them. A similar and perhaps correlative phyllonite in quartz diorite gneiss on the Little Salmon River is cut across sharply by an unaltered dike of trondhjemite. There was thus trondhjemitic activity both before and after retrograde shearing, which suggests that thrusting and both ages episodes of trondhjemite intrusion happened in a relatively short time.

Much of the Squaw Creek Schist in the vicinity of Riggins has a compound metamorphic fabric, wherein cleavage in one direction was superimposed on an older cleavage in a different direction, with or without a change in either direction in metamorphic grade. Larger scale structures of the Squaw Creek Schist are extremely complex; open folds were superimposed upon recumbent ones that had been superimposed upon isoclinal folds. Such complications are consistent with a history of long-separated episodes of deformation, but nothing in the meager available evidence suggests that all the deformation and metamorphism did not take place within the first half of Cretaceous time. The patterns of the isograds and of other mineralogic changes suggest that most of the reconstitution recorded by the present mineral assemblages occurred during a single episode of metamorphism.

METAMORPHISM OF ROCKS BENEATH RAPID RIVER THRUST

The metamorphism of the formations beneath the Rapid River thrust is younger than the Upper Triassic Martin Bridge Limestone. Within the Riggins quadrangle, there is a slight eastward increase in metamorphic grade in the rocks below the thrust. North of the quadrangle, the increase is far more marked, and lower plate rocks are intruded directly by granitic rocks marginal to the Idaho batholith. This progressively increasing metamorphism is probably of the same age as that of the Riggins Group.

Granitic rocks intruded into the Riggins Group are mostly unaltered, whereas in and north of the Riggins quadrangle the granite intrusives in the Seven Devils Volcanics have been metamorphosed to the same degree as have the wallrocks. Granitic rocks beneath the Rapid River thrust thus seem to be older than those above it. The Seven Devils Volcanics and allied formations extend westward from the Riggins quadrangle into eastern Oregon and maintain a generally low metamorphic grade over a broad region, with local increases about plutonic complexes. This metamorphism may be chiefly of Late Jurassic age.

The tectonic strike of the Seven Devils Volcanics in the Cuprum quadrangle is southwest. The meager data available suggest that 30 miles farther along strike to the southwest either the strike swings to the west or a major southwest-trending fault separates structural provinces of west and southwest strikes. The westward strike continues completely through the Baker quadrangle and then arcs southwestward into the Klamath Mountains near the Pacific coast. This structural continuity suggests that the Seven Devils Mountains belong to the same tectonic system as do the Klamath Mountains. Major deformation and granitic intrusion in the Klamath Mountains have been dated, both stratigraphically and radiometrically, as of Late Jurassic age, 130 to 140 million years (Curtis, Evernden, and Lipson, 1958). Dating elsewhere in the belt is consistent but less precise; thus Taubeneck (1959) dated the intrusives of eastern Oregon as of post-Callovian (earliest Late Jurassic) and pre-Albian (latest Early Cretaceous) age. This tectonic belt is the "Nevadan" belt of Hinds (1933).[2]

Granitic intrusives are widespread in the Nevadan orogen of Oregon, but they are, in general, little altered. If those in the Seven Devils Volcanics are of the same Jurassic age—a premise that is accepted here—then a metamorphism younger than that generally operative within the belt has been superimposed on the rocks beneath the Rapid River thrust. As the granitic rocks beneath the thrust in the Riggins quadrangle and in the area to the north are metamorphosed equally with the enclosing Seven Devils Volcanics, this younger metamorphism must have been of at least as high a grade as the older.

SYNTHESIS

These metamorphic relationships, like the structural ones to be discussed in a subsequent report, are consistent with a regional systhesis in which the crystalline rocks of the western belt, deformed, metamorphosed, and intruded by granitic rocks chiefly in Late Jurassic time, were again deformed and metamorphosed in the eastern part of the belt during Cretaceous orogeny. The eastern part of the Seven Devils Mountains may represent the area of overlap between the Jurassic and Cretaceous orogens.

A similar interpretation can be made for the northwestern Sierra Nevada of California, where it likewise seems that the eastern part of the Jurassic orogen was again deformed as the west edge of the Cretaceous orogen. The Rapid River thrust, near the west edge of the Cretaceous orogen, may also have counterparts in the Sierra Nevada.

The Cretaceous episode included the emplacement of a chain of enormous granitic batholiths, including those of Baja California, Southen California, Sierra Nevada, Idaho, and the Coast Ranges of British Columbia and Alaska. Intrusive activity during the Jurassic episode was far more limited.

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[2] As Curtis, Evernden, and Lipson (1958) emphasized, the name "Nevadan" was derived from the Sierra Nevada and unfortunately applied to the Klamath Mountains. Hinds made the erroneous assumption that orogeny in the Klamath Mountains was synchronous with that, across strike to the east, in the Sierra Nevada. Both the Sierra Nevada and the Idaho batholiths are now known to be of middle Cretaceous age.

METAMORPHIC FACIES

A metamorphic zone is defined in terms of an index mineral formed in rocks of a specific composition. Zones are not universally recognizable; in the Riggins quadrangle, for example, none of the metamorphic rocks are aluminous enough to yield the index minerals staurolite, kyanite, and sillimanite. To expedite comparisons between rocks metamorphosed under similar conditions but in which different mineral assemblages formed because of the different compositions of the rocks, Eskola (1915) proposed the concept of metamorphic facies. A facies comprises all rocks that have originated under the same physical conditions and is independent of specific mineral assemblages controlled by rock composition.

A facies classification is invaluable in metamorphic petrology, but unfortunately the classification has overshadowed the concept in much of the recent literature. The classification has become increasingly cumbersome and increasingly genetic in implication despite the contradictions this produces with the original concept.

ESKOLA AND THE FACIES CONCEPT

The facies concept was developed further by Eskola (1922, p. 146).

> A mineral facies comprises all the rocks that have originated under temperature and pressure conditions so similar that a definite chemical composition has resulted in the same set of minerals, quite regardless of their mode of crystallization, whether from magma or aqueous solution or gas, and whether by direct crystallization from solution (primary crystallization) or by gradual change of earlier minerals (metamorphic recrystallization).

There are parallel facies for igneous and metamorphic rocks, in each of which there are as many facies as there are different sets of minerals in any series of chemically identical rocks.

> The mineral associations of various rocks very often represent equilibrium in such a state of perfection that the characteristics of the ideal equilibria can be deduced from them. It is therefore possible to use these associations as a basis of classification.

Eskola (1922, p. 195) suggested that the names of the facies of regional metamorphism be based on the common rock of gabbroic composition in that facies—greenschist facies, amphibolite facies, eclogite facies. For contact-metamorphic facies, he suggested hornfels facies and sanidinite facies. Most petrologists since have followed this basic scheme.

By 1939, Eskola (*in* Barth, Correns, and Eskola, 1939) had added several new facies of metamorphic rocks and had published a diagram suggesting the relative conditions of temperature and pressure represented by the various facies, as shown here in figure 74. Eskola regarded the normal progression of regional metamorphism as indicating recrystallization at increasing temperatures; note that the greenschist, epidote-amphibolite, and amphibolite facies are shown at the same pressure in the diagram.

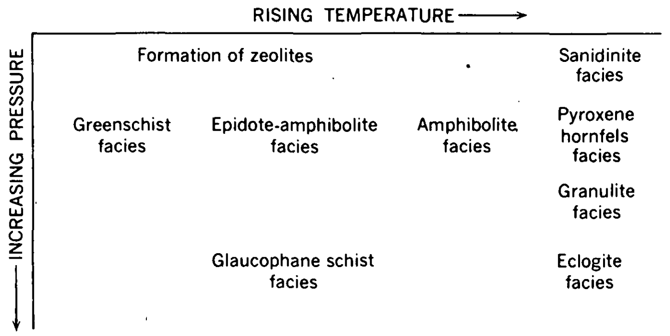

FIGURE 74.—Relative pressure-temperature regions suggested by Eskola for the metamorphic facies. After Barth, Correns, and Eskola (1939, p. 345).

REVISIONS BY TURNER

Eskola's facies system was expanded by F. J. Turner (1948) by the addition of subfacies. By 1958 (*in* Fyfe, Turner, and Verhoogen, 1958), Turner's classification had become more elaborate, and the old epidote-amphibolite facies was demoted to subfacies status (quartz-albite-almandine) within the greenschist facies. (The nomenclature of the contact-metamorphic facies underwent a similar expansion between Eskola in 1939 and Turner in 1958.)

Turner's subfacies for the rock types developed in normal regional metamorphism now correspond directly to the metamorphic zones of the Dalradian schists of Scotland, but the nomenclature has become progressively more cumbersome. Thus, the "chlorite zone" of the Scottish petrologists was transformed into the "Muscovite-chlorite subfacies of the Greenschist facies" and then into the "Quartz-albite-muscovite-chlorite subfacies of the Greenschist facies" by Turner in 1948 and 1958, respectively.

ROSENQVIST'S NOMENCLATURE

According to Rosenqvist (1952, p. 30), the transition from greenschist facies to epidote amphibolite facies is chiefly one of dehydration and conversion of chlorite to hornblende. Passing through the epidote-amphibolite facies, the anorthite content of plagioclase increases gradually at the expense of zoisite; zoisite vanishes from rocks with sodic plagioclase before it does from those with calcic plagioclase (op. cit., p. 31). Iron-bearing epidote is stable with plagioclase of a given composition at a higher temperature than is zoisite (op. cit., p. 48–49). Still higher facies changes involve further dehydration. Rosenqvist suggested that general names (for example, low gneiss facies) rather than specific rock names would be preferable for metamorphic facies. His facies, and the P–T fields he inferred for them on the basis of temperature-volume relationships and laboratory data, are shown in figure 75.

Expansion of the metamorphic facies

[Facies classification of regionally metamorphosed rocks by Eskola (*in* Barth, Correns, and Eskola, 1939); Turner (1948); Turner and Verhoogen, (1951); and Turner, revised (*in* Fyfe, Turner, and Verhoogen (1958)]

Eskola	Turner (1948)		Turner (1958)	
Facies	Facies	Subfacies	Facies	Subfacies
(Formation of zeolites)			Zeolite	
Greenschist	Greenschist	Muscovite-chlorite	Greenschist	Quartz-albite-muscovite-chlorite
		Biotite-chlorite		Quartz-albite-biotite
Epidote-amphibolite	Albite-epidote amphibolite	Chloritoid-almandine		Quartz-albite-almandine
Amphibolite	Amphibolite	Staurolite-kyanite	Almandine amphibolite	Staurolite-quartz
				Kyanite-muscovite-quartz
		Sillimanite-almandine		Sillimanite-almandine
		Almandine-diopside		
Granulite	Granulite		Granulite	Hornblende granulite
				Pyroxene granulite
Eclogite	Eclogite		Eclogite	

EVALUATION

Progressive metamorphism results in systematically changing mineral assemblages. For a given rock composition in a given area, the major variable is temperature, rock pressure being relatively constant; isograds correspond to isotherms. (This is discussed on later pages.) Although the actual temperature equivalents of the isograds can be inferred only roughly, and although quantitative assessment of effects of pressure, stress, and lesser variables cannot yet be made, there is no reason to doubt the established opinion that each mineral assemblage formed in rocks of the same composition represents crystallization within a field of variables that is potentially definable.

Nevertheless, the same combinations of minerals have different relative stability ranges in rocks of diverse compositions in any one area and in rocks of the same composition in different areas. There are neither universal index minerals nor definitive assemblages. Any basis chosen for definition of a mineral facies is inherently arbitrary; any facies boundary determinable in practice is necessarily fuzzy. The smaller the facies field, the greater the width of the indeterminate zone relative to the total width of the facies. Turner's subfacies are not recognizable in rocks of many compositions.

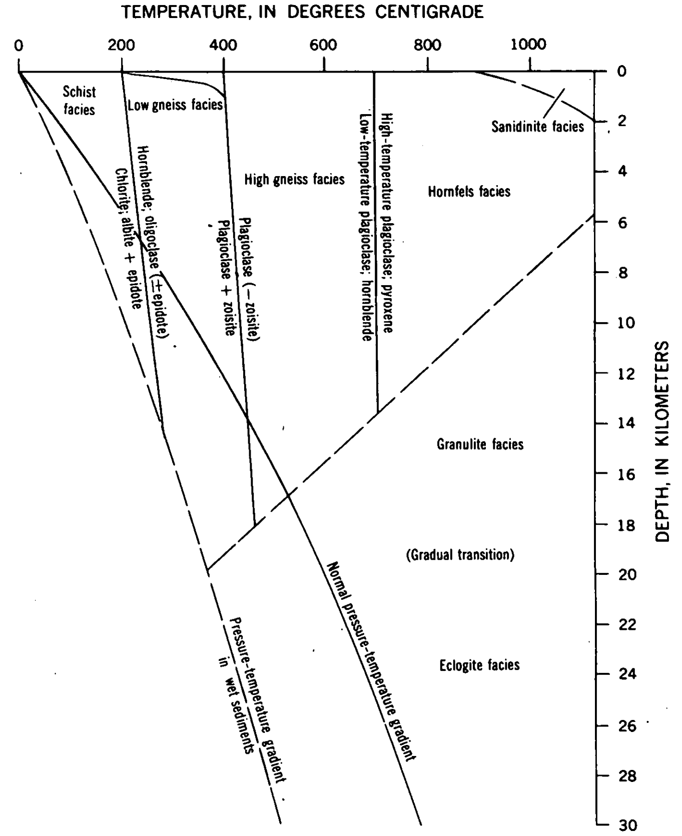

FIGURE 75.—Pressure-temperature fields suggested by Rosenqvist (1952, fig. 20b) for his facies system. Schist facies ≈ greenschist facies of other authors, low gneiss ≈ epidote amphibolite, high gneiss ≈ amphibolite, hornfels ≈ pyroxene hornfels + part of granulite facies.

Turner has added a genetic emphasis, which was lacking in Eskola's definition, to the facies classification by divorcing "Facies of contact metamorphism" from "Facies of regional metamorphism." Turner's "Albite-epidote hornfels facies" is characterized by precisely the same mineral assemblages as are the low-grade facies of regional metamorphic rocks, and its recognition "* * * is justified only on the basis of field relations to the hornblende hornfels facies and to intrusive igneous bodies." (Fyfe, Turner, and Verhoogen, 1958, p. 203). Turner's usage diverges directly from Eskola's definition of metamorphic facies. The basis for facies should be mineral assemblages; when identical rocks are assigned to different facies on the basis of notions of genesis, the facies concept has been badly distorted.

For reasons such as these, it seems that metamorphic facies have been overclassified by Turner. Although correlation of temperature and other conditions of metamorphism may some day be possible from one region to another, it is not possible now. It is doubtful that "Quartz-albite-muscovite-chlorite subfacies of the Greenschist facies" conveys anything to most petrologists, except a feeling of numbness, that "chlorite zone" does not. In this paper, facies designations follow Eskola, and subdivisions are based on index minerals—that is, subdivisions are to zones, not to subfacies

Rosenqvist's suggestion that facies should not be given rock-type names has much to commend it, but it has not been accepted generally, whereas Eskola's names are used throughout the world.

COMPARISON WITH OTHER REGIONS

Progressive metamorphism on a regional scale was first demonstrated by Barrow (1893) in the southeastern Highlands of Scotland, where he recognized zones characterized by the index minerals sillimanite. kyanite, and staurolite in metashales. Subsequent work there by others added chlorite, biotite, and garnet zones. Systematic studies of progressively metamorphosed rocks have since been made in many regions, and petrologic comparisons between some of these are appropriate here.

Comparisons cannot be made rigidly, for there are neither universal index minerals nor universally diagnostic mineral assemblages. Wherever rocks of markedly different compositions are intercalated, it has been found that new minerals represent different grades of metamorphism in some compositions than they do in others; zones defined by the same index mineral, but in rocks of different compositions, need not be isofacial. Even assemblages of several minerals have different facies significance in rocks of different compositions; thus, Turner, trying to define subfacies with involved mineral assemblages (as, quartz-albite-muscovite-chlorite, noted above), still succeeds only in defining subfacies for rocks of specific compositions, and in such rocks the whole assemblage has little more diagnostic value than does a single index mineral.

Further complicating comparisons is the lack of interregional correspondence between assemblages, even in rocks of the same compositions.

A good illustration is provided by the wide variability in stability of chlorites in metamorphic rocks. Chlorite does not persist far beyond the first appearance of biotite in metashales rich in aluminum and potassium; but the upper grade limit of chlorite in metaigneous rocks varies greatly from region to region. In northern Michigan (James, 1955, p. 1467–1473) and northern Manitoba (Ambrose, 1936), chlorite disappears midway between the biotite and garnet isograds in metaigneous rocks; in New Hampshire (Billings, 1956, p. 139) and the southwest Scottish Highlands (Wiseman, 1934), chlorite is stable up to about the garnet isograd; and in South Island, New Zealand (Reed, 1958) and in western Idaho (this report), chlorite is stable in rocks containing both garnet and oligoclase.

Plagioclase composition varies less widely. In metamorphic rocks of mafic igneous compositions, the transition from albite to oligoclase occurs at a lower grade than the garnet isograd in northern Michigan (James, 1955); at about the same place as garnet in parts of South Island (Reed, 1958), although rocks of similar composition elsewhere on the island have an oligoclase isograd but no garnet (Turner, 1938); and at a markedly higher grade than the garnet isograd in the Southwest Highlands (Wiseman, 1934), New Hampshire (Billings, 1956), and western Idaho (this report).

These variations are expressed in terms of other index minerals, but no implication is intended that those reference minerals represent metamorphic conditions any more constant than do chlorite and oligoclase in the examples cited. Garnet appears before oligoclase in some regions and after it in others; garnet is more commonly used as an index mineral, but this usage does not make it any more reliable as an indicator of metamorphic grade than oligoclase. The popular notion that metamorphic zones and facies are directly correlative from one region to another is incorrect. This does not invalidate the concept of zones and facies or the utility of their expression, but it does emphasize that they cannot be rigidly defined. The variables are more complex than can be suggested by tabulations of ideal subfacies.

Despite such differences in details of mineralogy, the general progression of metamorphism is similar in most terranes of regionally metamorphosed rocks. The points intended here are that comparisons must be qualitative rather than quantitative and that rigid classification of metamorphic facies is not yet possible. Any classification of metamorphic rocks by zone, grade, or facies is inherently an attempt to superimpose a set of semantic pigeonholes upon a natural continuum of extreme complexity. The natural system contains far more variables, disconcertingly independent, than does any scheme of classification.

The progressive transformations found by Wiseman (1934) in metabasalt in the Southwest Highlands are similar to those in the Riggins quadrangle. Hornblende low in aluminum and alkalis and high in calcium characterizes the biotite-zone rocks of the Scottish region; the hornblende changes progressively, and in the garnet zone it is higher in aluminum, iron: magnesium ratio, and alkalis, and lower in calcium than it is in the lower grade rocks. Much of this change is accomplished by transfer to hornblende of ferrous iron and aluminum from prochlorite. Epidote or clinozoisite persist with andesine or even labradorite in metabasalt intercalated with kyanite-bearing metashale.

Vogt (1927) studied the progressive metamorphism of lower Paleozoic metasediments in northern Norway and found a clinozoisite isograd, like that of the Riggins quadrangle, intermediate between biotite and garnet, and a hornblende one between garnet and oligoclase. (The suite is distinctly different from the Riggins one in that chlorite disappears near the biotite isograd.) Vogt (p. 445) emphasized the importance of progressive dehydration in producing the successive mineralogic changes.

FACTORS OF METAMORPHISM

The schists of the Riggins Group become progressively less hydrous as their metamorphic grade increases. That the water loss occurred during metamorphism and not long afterward is indicated by evidence such as that presented by Fyfe and Verhoogen (*in* Fyfe, Turner, and Verhoogen, 1958, p. 98 and 187). Were the water released by the progressive dehydration transformations not removed, the reverse, retrogressive reactions should have taken place as the rocks cooled, yet such reversion is very uncommon. High-grade rocks contain so little water that they cannot revert upon cooling. (An alternative explanation often given for the lack of reversion is based on the presumably faster rates of the higher temperature reactions; were this correct, then metamorphic and igneous rocks should show superimposed effects of all lower grades of retrogression.) Water is lost during metamorphism; so there are pressure gradients down which the volatiles travel, and the flow of water in metamorphosing rocks is at a significantly high rate relative to the time required for the reactions. (Yoder's [1955, p. 512–513] opposite claim that the flow is negligibly slow is invalid.) The lower the water pressure, the lower the temperature and rock pressure at which dehydration reactions can proceed; and as water is lost during metamorphism, temperatures of natural dehydrations are lower than those found in laboratory closed-system studies (Thompson, 1955, p. 96–98).

Veins of quartz and carbonate, deposited from aqueous solutions that forced the vein walls apart, apparently require that total-volatile pressures equal or even slightly exceed rock pressures. The partial pressure of water will be decreased in proportion to the amount of CO_2 and other volatiles present.

The persistence of prochlorite into middle-grade rocks of the Riggins quadrangle may be due to unusual behavior of water, for the prochlorite apparently did not dehydrate until temperatures well above those adequate for dehydration in most metamorphic terranes were reached. A possible cause is that water pressure might have been abnormally high, either because water could not escape easily, or because load pressure—that is, depth—was uncommonly great. However, the persistence of prochlorite may be due largely to the very low content of potassium in most of the schists, for the little potassium present went chiefly into hornblende rather than into biotite.

Mineral transformations due primarily to increasing pressure should be in the direction of greater density, although all products of a given reaction are not necessarily heavier than all the reactants. Figure 76 shows common densities of silicate minerals like those of the Riggins quadrangle. The broad trend of metamorphism in the Riggins quadrangle is in the direction of greater density, but the higher grade aluminian prochlorite is lighter than the lower grade ferroan prochlorite and much lighter than the biotite and hornblende with which it is stable over a wide range of metamorphic grade.

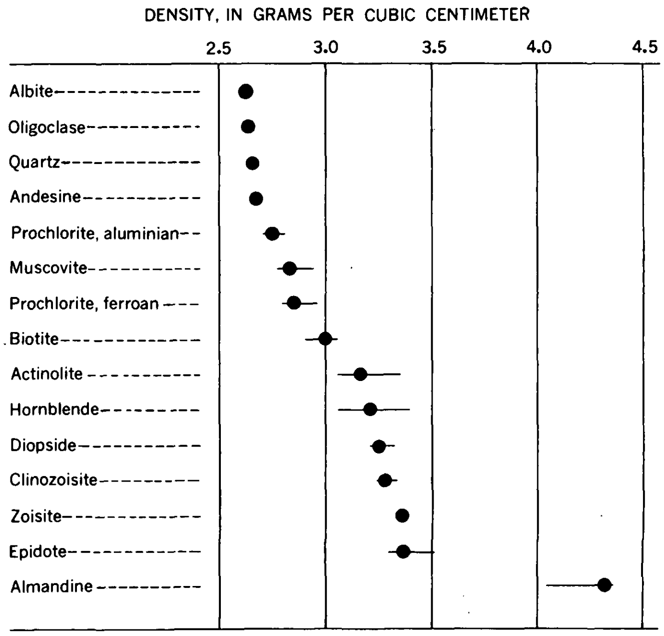

FIGURE 76.—Densities of silicate minerals present in metamorphic rocks of Riggins quadrangle. Data from Tröger (1956, pp. 91 and 128–130).

Isograds are likely to represent isotherms in any one area, as most petrologists would agree; the primary cause of metamorphic progression is increasing temperature. Comparison with data from hydrothermal experiments and with diverse arguments in papers mentioned in other connections (see figs. 77, 78 suggests that the lowest grade rocks formed at about 200° C, whereas the highest formed at near 600°C.

The schists of the Riggins Group show the effects of extreme shear during their crystallization, and it might be suggested that such shear was responsible for the unusual mineral assemblages. Prochlorite, however, occurs both as a groundmass mineral parallel to the foliation and as porphyroblasts that cut across the foliation and so formed both during and after shear.

Clinozoisite and epidote similarly appear as both concordant groundmass grains and crosscutting porphyroblasts. Directed stress does not seem to be a likely cause for the persistence of these low-grade minerals into rocks of high metamorphic grade.

SOURCE OF HEAT OF METAMORPHISM

The metamorphism of the rocks of the Riggins quadrangle was regional or dynamothermal and was accomplished in response to high temperatures and pervasive shearing. Temperatures during that metamorphism ranged from perhaps 200° or 300°C in the western part of the quadrangle to about 700°C in the plutonic terrane east and south of the rocks considered in this report. According to popular theory, such metamorphism is due to deep burial within a geosynclinal prism of materials rather than to heat introduced by granitic batholiths mobilized at much greater depths or to heat generated by deformation. The first notion can be disproved at least in part despite its wide acceptance, and the other two explanations are probably applicable.

DEPTH ZONES

Grubenmann (1910, p. 80) thought that depth controlled temperature, hydrostatic pressure, directed stress, and the relative importance of chemical and mechanical processes in metamorphism. In Grubenmann's classification, the various zones of progressively metamorphosed rocks represent simply progressively greater depth of formation. Grubenmann and Niggli (1924) expanded the depth-zone concept, and Niggli (1954) defined the temperatures of metamorphism and the metamorphic facies, in terms of depth, by such terms as "epithermal" and "Kata facies." These usages are unfortunate, for the less genetic are our descriptive terms, the longer will they prove useful.

Although few petrologists now agree that metamorphic rocks should be classified on so intangible a basis as hypothetical depth of formation, most petrologists still accept depth as the controlling factor of metamorphism. Harker (1939, p. 179, 184, and 185) stated that the controlling variable of regional metamorphism is temperature but that temperature in turn "* * * generally corresponds with depth of cover * * *" and that granitic intrusions are products rather than causes of the metamorphism. Barth (1952) indicated that progressive regional metamorphism is a function of depth, and that it depends upon the normal geothermal gradient as a source of heat (fig. 77).

Barth's view is extreme. A far more widespread concept is that increasing metamorphism indicates increasing depth of formation in a region where there was an abnormally high *P–T* gradient caused by surges of conducted heat (specifically not heat transferred by

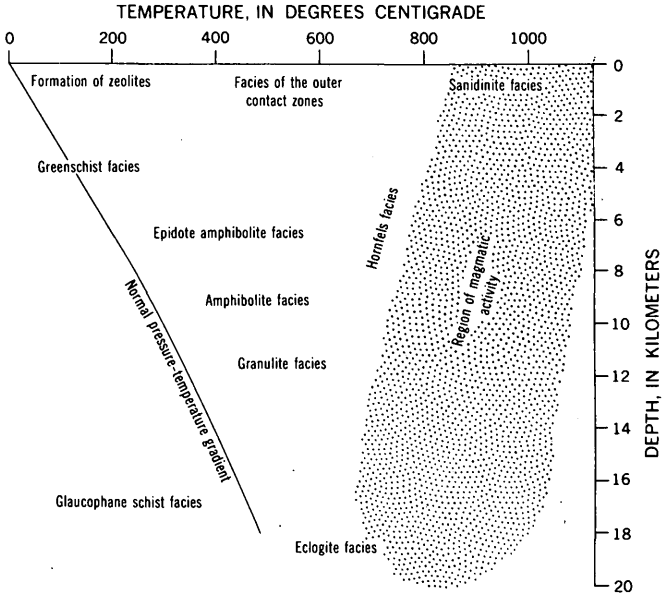

FIGURE 77.—Pressure-temperature regions suggested by Barth for the metamorphic facies. Replotted from Barth (1952, fig. 137).

magma). Ramberg (1952, fig. 73) showed both temperature of metamorphism and the succession of metamorphic facies as linear functions of depth. Turner (*in* Fyfe, Turner, and Verhoogen, 1958, fig. 107; see fig. 78 of the present paper) recognized that the facies can have wide fields of pressure; but, like Ramberg, he considered the increasing temperature in any one area of metamorphosing rocks to be due only to increasing depth.

INVALIDITY OF THE DEPTH ZONES

Despite the wide acceptance of depth as the controlling factor of progressive regional metamorphism, other factors instead must be more important in many areas, and probably in most regions. The dimensions of metamorphic zones provide critical evidence.

Kennedy (1949) summarized the metamorphic and igneous geology of the Scottish Highlands. Grade of metamorphism increases inward from the Moine thrust on the west and the Highland Border fault on the southeast. The chlorite zone is as wide as 30 miles; the biotite zone is 1 to 8 miles wide; and the garnet zone is 1 to 30 miles wide. The kyanite isograd is equivalent to the outer limit of regional injection. Granites accompanied by far-traveled injections are abundant inside that isograd; other granites are present outside but lack peripheral injections. The kyanite isograd is also the approximate boundary between schists which retain bedding and other sedimentary details, and flow-folded migmatites and gneisses to which much material was added during metamorphism. Isograds are subparallel to the regional strike but not symmetrically related to major folds and not influenced by inliers of older rock, so there can be no simple depth relationship between zones and structure. The correlation between metamorphism and igneous injection is thus direct, but either can be argued as the cause. Harker's view that metamorphism was the cause was noted above. Barrow (1893) considered the metamorphism to be contact metamorphism on a large scale, and Kennedy (1949) and others have advocated this also.

Metamorphism in New Hampshire (Billings, 1955, 1956) resulted in a sillimanite-zone "plateau," 50 miles or more wide, defined by isograds that are, in general, parallel and closely spaced; biotite, garnet, staurolite, and sillimanite isograds lie within a belt less than 1 mile wide in some places. Isograds in "regionally metamorphosed" rocks in eastern Honshu (Shido, 1958; Miyashiro, 1958) are concentric and closely spaced about plutonic intrusions, and the mineralogic grade of the metamorphic rocks corresponds directly to that of the intrusives in a fine example of the principle of metamorphic facies. The background metamorphic grade in northern Michigan (James, 1955) is that of the chlorite zone; biotite, garnet, staurolite, and sillimanite zones are 1 to 10 miles wide each and are concentric about several nodes. "Analysis of thermal gradients inferred from the metamorphic zonation indicates that the heat required for metamorphism must have been derived from subjacent bodies of magma * * *" beneath the high-grade nodes, and "* * * deformation and metamorphism were independent variables for this particular region * * *" (James, 1955, p. 1455). In the Riggins quadrangle, biotite and andesine isograds are 2 to 4 miles apart, and biotite and clinopyroxene isograds are 3 to 4 miles apart.

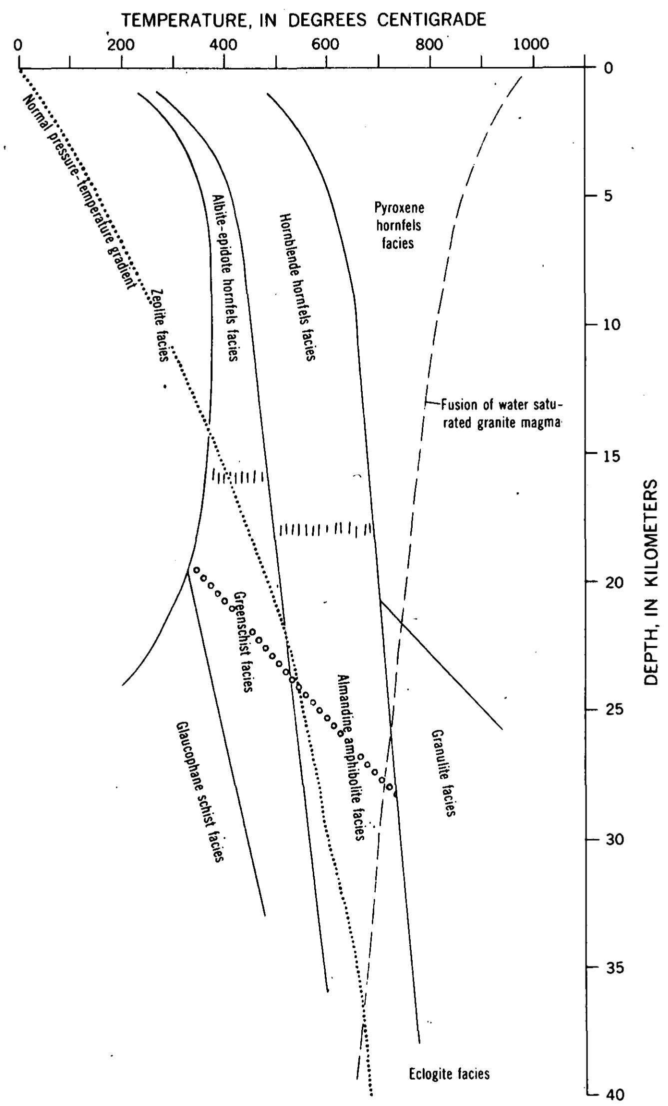

FIGURE 78.—Pressure-temperature fields suggested by Turner for the metamorphic facies. Plotted from Fyfe, Turner, and Verhoogen (1958, fig. 107). (This figure had P_{H_2O} as pressure coordinate, which was here equated directly with P_{LOAD}.) "Normal P–T gradient" from op. cit. (fig. 77). Line of circles shows course of Barrovian metamorphism (op. cit., fig. 108).

Petrologists agree, in general, that the biotite isograd represents a temperature of 200° or 300° C, whereas the sillimanite forms near 600° or 700° C. Those two isograds thus represent a temperature difference of metamorphism of about 400° C. It is relevant to compare this value with some dimensions of metamorphic zones.

In New Hampshire, biotite and sillimanite isograds are locally less than 1 mile apart, requiring a temperature gradient in the plane of the present ground surface of about 300°C per km at the time of their formation. In the Riggins quadrangle, gradients along the surface of present exposure of about 75°C per km are indicated. If such metamorphism is assumed to be due to vertical temperature gradients, then the vertical gradients would approach the gradients specified only in the extreme and unlikely event that the present ground surfaces were vertical planes during metamorphism, for gentler original inclinations would raise the vertical temperature gradients correspondingly.

Were Barth's correlations of metamorphic facies with depth (fig. 77) correct, then biotite and sillimanite isograds could be as close as 4 miles only where the ground surface represents an initially vertical plane; where the modern ground surface represents a more reasonable plane that initially dipped 10° inward toward the high-grade terrane, biotite and sillimanite isograds should be 25 miles apart. The isograds are seldom that far apart; so the model is inadequate. Depth, even where temperature gradients are high, cannot be the controlling variable of most regional metamorphism.

Fyfe and Verhoogen (*in* Fyfe, Turner, and Verhoogen, 1958, p. 196–198) explained close spacing between isograds in terms of vertical thermal gradients of 100° to 300°C per km, rather than the normal 30°C per km. The very high heat flow causing this is due to "* * * momentary and localized heat surges * * *" produced by "* * * intermittent and localized upward transfer of hot material in the mantle." The heat, they hypothesized, is conducted upward into the metamorphosing rocks; although in some instances ascending "* * * juvenile water may indeed be an important heat carrier * * *," granitic magmas "* * * cannot be counted as independent heat sources * * *" because they are "* * * formed by partial melting within the metamorphic column."

This thermal model seems highly improbable. A heat surge at the base of the crust could send a thermoclinal heat front upward, but the front would cool as it rose so that a great temperature difference could not be maintained across it. The model provides no way to freeze the thermal front in place; for as the front continued to rise, the low-grade rocks formed by the advancing thermocline should be upgraded to high-grade rocks, and the end result should be widely spaced isograds.

The Fyfe-Verhoogen model above the zone of metamorphism is no more plausible. They do not attempt to explain how a gradient produced by heat conducted upward from the base of the continental material could have a value of 200°C per km at a depth of 20 km and a value of perhaps 10°C per km, a twentieth as much, at 10 km.

Turner's (*in* Fyfe, Turner, and Verhoogen, 1958) thermal model is somewhat more plausible. (See fig. 78 of this paper.) He suggested that the metamorphism of the southern Scottish Highlands occurred at a depth of 20 to 30 km with a geothermal gradient of about 45°C per km, and yet that the temperature at 20 km was little over 300°C, so that the average gradient above the upper level of greenschist-facies metamorphism was only 15°C per km. The thermal objections to this model are less marked than are those to that of Fyfe and Verhoogen, but Turner's model is inadequate to explain closely spaced isograds.

Another defect of the depth-zone concept is that it requires an inordinate amount of erosion before the metamorphosed rocks could be exposed at the surface. Turner's model (fig. 78), for example, would form the high-grade rocks of the southern Scottish Highlands at a depth of nearly 30 km, yet those rocks were exposed at the surface and covered unconformably by sedimentary deposits within, at most, several tens of millions of years after their metamorphism. Postulating erosion to a depth equivalent to the normal subcontinental depth of the Mohorovicic discontinuity within such a short period of time, followed by essential stability with little uplift or erosion to the present time, seems unwarranted. In many parts of the world—probably in most orogens—metamorphic terranes have been exposed at the surface within a part of a geologic period after their formation. The great middle-Cretaceous batholiths, including the Idaho batholith, of Western United States were exposed within early Late Cretaceous time, for example.

The last quotation given above from Fyfe and Verhoogen gives the key to most of the modern theorizing that depth is the controlling factor of metamorphism, for theories of metamorphism are inseparably tied to theories of the origin of granitic rocks. Most of those who believe that granitic magmas are generated within the exposed terranes of high-grade metamorphic rocks believe also that geothermal heat is the cause of that metamorphism. A few are so convinced of the truth of these assumptions that they go as far as Ramberg (1952, p. 273) in denying the possibility of alternatives:

> There are no grounds for believing that regional metamorphism is chiefly caused by the heat liberated by intruded melts.

Those who believe, on the contrary, that the magmas were mobilized at much greater depths and intruded into the levels at which they are now exposed believe also that the magmas carried upward with them a large part of the heat represented by regional metamorphism.

The popular assumption that granitic magmas are generated within the exposed high-grade metamorphic terranes of geosynclinal rocks is probably wrong in many places. Thus in southern California and the southern Appalachians, large masses of granitic rocks of Mesozoic and Paleozoic ages, respectively, were intruded upward into Precambrian plutonic rocks and hence cannot have been formed as popular theory would require.

CONTACT METAMORPHISM ON A REGIONAL SCALE

In the Riggins quadrangle, as in most metamorphic terranes, the metamorphic progression is spatially related to large masses of granitic rocks. Although here, as elsewhere, small granitic plutons cut the low- and middle-grade rocks, the metamorphic zones are broadly concentric with the margin of the vastly larger granitic terrane to the east and south—the Idaho batholith and its marginal facies. The granitic contact is not sharp; it is, rather, gradational over many miles, the proportion of granitic material increasing gradually as that of metamorphic rock decreases. The earlier granitic masses in this gradational zone are interlayered concordantly with the metamorphic rocks, whereas later granites cut across both the metamorphic rocks and the older granites. Within the Riggins quadrangle, the bulk of the granitic rocks are gneissic and concordant, and the massive plutons of the continuous

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Idaho batholith, which intruded such gneisses, lie a few miles farther east. Both gneissic and crosscutting granites are intrusive at their present levels. Limits of pegmatitic and granitic dikes in the metamorphic rocks of the quadrangle are similarly related areally to the granitic terrane.

North of the Riggins quadrangle, the metamorphic rocks of the Riggins Group, and the gneissic complex marginal to the Idaho batholith are cut out by the crosscutting plutons of the batholith that are in direct contact with lower grade metamorphic rocks whose metamorphism has obviously been of contact type. Whether the schist-and-gneiss zone of the Riggins quadrangle was originally formed all along the margin of the batholith and subsequently engulfed by the massive plutons, or whether it was the result of intense local metamorphism, or even was part of an older crystalline complex unrelated to the Idaho batholith, cannot be proved; but the first alternative seems most probable. The schists change systematically toward the gneisses, and the gneisses change toward the batholith; the zone of total gradation from greenstone through schist and gneiss to massive granite is 30 or 40 miles wide. The heat that caused much of the metamorphism was probably carried upward by granitic magmas and water. If so, it was "* * * contact metamorphism on a regional scale * * *" (a phrase used by Buddington and Chapin, 1929, p. 20, in southeastern Alaska).

Within regional metamorphic terranes, only low-grade rocks are generally formed extensively far from large quantities of exposed or inferable granitic rocks. West of the Riggins quadrangle, for example, chlorite-zone rocks are present throughout the pre-Cretaceous complexes, although higher grade rocks are present only in much more limited areas about the relatively small granitic masses. Such a broad metamorphic terrane, far from intrusive rocks, shows that low-grade metamorphism occurs independently of igneous activity.

Buddington (1959, p. 732) observed that even in the Canadian Shield there is only one large area—the Grenville belt, 1,000 miles long and 250 to 350 miles wide—of uniformly high-grade rocks that probably formed at relatively great depths. Elsewhere in the shield, according to Buddington, the "* * * sillimanite facies is generally restricted to zones of rock adjacent to the major plutons or complex of plutons * * *," and "* * * the intensity of regional metamorphism as indicated by rocks farthest from the major intrusives is prevailingly that of the greenschist facies with local zones in the staurolite-kyanite subfacies." (Recent reconnaissance suggests that New Quebec, Laborador, and southern Baffin Island may include another large terrane of deep-seated rocks.)

Eskola realized that there need be no increase in depth between the greenschist facies and amphibolite facies. (See fig. 74.)

Shido (1958) suggested that each area of metamorphic rocks represents a single depth of formation, the temperature gradients being due to plutonic intrusives, and that differences between the character of the metamorphism from place to place are partly dependent upon different depths (fig. 79). Shido's approach is worthy of broad testing. His deduction that the P–T boundaries between the fields of the metamorphic facies are not parallel (compare fig. 79 with figs. 75, 77, 78) may lead to simple explanations of many of the confusing contrasts between metamorphic provinces.

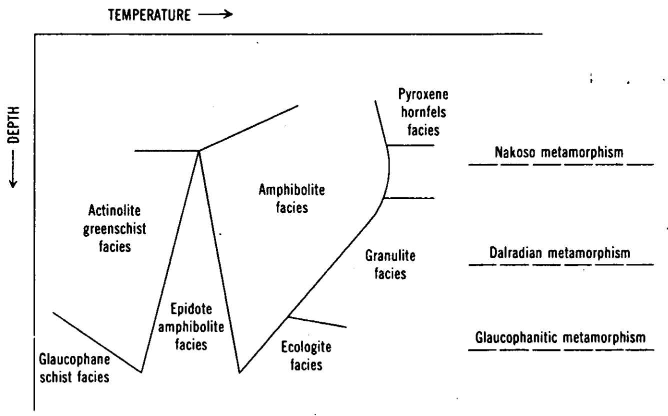

FIGURE 79.—Relative pressure-temperature fields of some of the metamorphic facies, after Shido (1958, fig. 31).

This discussion has dealt with the normal sequence of progressive regional metamorphism, in which the highest grade rocks belong to the amphibolite facies. Some other types of metamorphism, such as that which produces eclogites, granulites, and glaucophane schists, might be due to much greater depth of burial, although this also can be challenged.

DEFORMATION AS A SOURCE OF HEAT

Rocks metamorphosed regionally at middle or high grade invariably are deformed and commonly have undergone intense pervasive shearing. Metamorphism and pervasive deformation are closely related. This can be explained in terms either of shearing facilitated by raised temperature or of shearing causing a high temperature. Deformation is conventionally assumed (as, by Turner and Verhoogen, 1952, p. 570–571, and Ramberg, 1952, p. 272–273) to be at most a minor source of heat, but data with which to analyze this assumption are much needed. Fyfe and Verhoogen (*in* Fyfe, Turner, and Verhoogen, 1958, p. 197) stated that "The strain energy involved in an 18-percent shortening under compressions of marble at 150° C. and 10,000 atmospheres confining pressure is about 3 cal per g." This would correspond to a temperature increase of only about 12° C., and they concluded that "* * * strain energy cannot be an important contributor to metamorphic heat."

Regionally metamorphosed limestones, however, have commonly been deformed by vastly more than the 18 percent in this example: the low-grade metalimestone conglomerate illustrated in figure 6 has been deformed by perhaps 200 percent, and the higher grade metalimestones of the Riggins quadrangle have been deformed by at least several times this large amount. Temperature increases of several hundred degrees Centigrade appear consistent with observed deformations.

SYNTHESIS

Popular theory explains regional metamorphism as occurring at depths of 20 to 30 km in response to extremely high temperature gradients due to heat conducted from the base of the crust; but such depths seem much too great, and such a mechanism of heat transfer seems implausible. Generation of heat within the metamorphosing rocks by deformation and introduction of heat in magma and steam from deeper, hotter sources seem instead to be the primary causes of regional metamorphism.

REFERENCES CITED

Ambrose, J. W., 1936, Progressive kinetic metamorphism in the Missi series near Flinflon, Manitoba: Am. Jour. Sci., v. 32, no. 190, p. 257–286.

Anderson, A. L., 1930, The geology and mineral resources of the region about Orofino, Idaho: Idaho Bur. Mines and Geology Pamph. 34, 63 p.

Backlund, H. G., 1930, Die Magmagesteine der Geosynklinale von Nowaja Zemlya: Report of the scientific results of the Norewegian expedition to Novoja Zemlya 1921, Oslo, no. 45, p. 23–61.

Barrow, George, 1893, On an intrusion of muscovite-biotite gneiss in the southeastern Highlands of Scotland, and its accompanying metamorphism: Geol. Soc. London Quart. Jour., v. 49, p. 330–358.

Barth, T. F. W., 1952, Theoretical petrology: New York, John Wiley & Sons, 387 p.

Barth, T. F. W., Correns, C. W., and Eskola, Pentti, 1939, Die Enstehung der Gesteine: J. Springer, Berlin, 422 p.

Bartrum, J. A., 1936, Spilitic rocks in New Zealand: Geol. Mag., v. 73, no. 9, p. 414–423.

Battey, M. H., 1955, Alkali metasomatism and the petrology of some keratophyres: Geol. Mag., v. 92, no. 2, p. 104–126.

Benson, W. N., 1927, The tectonic conditions accompanying the intrusion of basic and ultra-basic igneous rocks: Natl. Acad. Sci. Mem. v. 19, no. 1, 90 p.

Billings, M. P., 1955, Geologic map of New Hampshire (1: 250,000): U.S. Geol. Survey.

Billings, M. P., 1956, Bedrock geology, part 2 of The geology of New Hampshire: New Hampshire Planning Devel. Comm., 203 p.

Brindley, G. W., and Gillery, F. H., 1956, X-ray identification of chlorite species: Am. Mineralogist, v. 41, nos. 3–4, p. 169–186.

Buddington, A. F., 1959, Granite emplacement with special reference to North America: Geol. Soc. America Bull., v. 70, p. 671– 747.

Buddington, A. F., and Chapin, Theodore, 1929, Geology and mineral deposits of southeastern Alaska: U.S. Geol. Survey Bull. 800, 398 p.

Byers, F. M., Jr., 1959, Geology of Umnak and Bogoslof Islands, Aleution Islands, Alaska: U.S. Geol. Survey Bull. 1028–L, p. 267–369.

Coats, R. R., 1952, Magmatic differentiation in Tertiary and Quaternary volcanic rocks from Adak and Kanaga Islands, Aleutian Islands, Alaska: Geol. Soc. America Bull., v. 63, no. 5, p. 485–514.

——— 1953, Geology of Buldir Island, Aleutian Islands, Alaska: U.S. Geol. Survey Bull. 989–A, p. 1–26.

Coats, R. R., Lewis, R. Q. Nelson, W. H., and Powers, H. A., 1961, Geologic reconnaissance of Kiska Island, Aleution Islands, Alaska: U.S. Geol. Survey Bull. 1028–R, p. 563–581.

Cook, E. F., 1954, Mining geology of the Seven Devils region: Idaho Bur. Mines and Geology Pamph. 97, 22 p.

Coombs, D. S., 1954, The nature and alteration of some Triassic sediments from Southland, New Zealand: Royal Soc. New Zealand Trans., v. 82, p. 65–109.

Cristi, J. M., 1956, Chile, *in* W. F. Jenks, ed., Handbook of South American geology: Geol. Soc. America Mem. 65, p. 187–214.

Curtis, G. H., Evernden, J. F., and Lipson, J., 1958, Age determinations of some granitic rocks in California by the potassium-argon method: California Div. Mines Spec. Rept. 54, 16 p.

Daly, R. A., 1914, Igneous rocks and their origin: New York, 563 p.

Drewes, Harald, Fraser, G. D., Snyder, G. L., and Barnett, H. F., 1961, Geology of Unalaska Island and adjacent insular shelf: U.S. Geol. Survey Bull. 1028–S, p. 583–676 [1962].

Eskola, Pentti, 1915, Om sambandet mellan kemisk och mineralogisk sammansättning hos Orijärvitraktens metamorfa bergarter: Comm. géol. Finlande Bull. 44.

——— 1922, The mineral facies of rocks: Norsk Geol. Tidssk., v. 6, p. 143–194.

Eskola, Pentti, Vuoristo Urho, and Rankama, Kalervo, 1937 [1935], An experimental illustration of the spilite reaction: Comm. géol. Finlande Bull. 119, p. 61–68.

Fenner, C. N., 1926, The Katmai magmatic province: Jour. Geology, v. 34, no. 7, pt. 2, p. 673–772.

Fraser, G. D., and Barnett, H. F., 1959, Geology of the Delarof and westernmost Andreanof Islands, Aleutian Islands, Alaska: U.S. Geol. Survey Bull. 1028–I, p. 211–248.

Fraser, G. D., and Snyder, G. L., 1959, Geology of southern Adak Island and Kagalaska Island, Alaska: U.S. Geol. Survey Bull. 1028–M, p. 371–408.

Fyfe, W. S., Turner, F. J., and Verhoogen, Jean, 1958, Metamorphic reactions and metamorphic facies: Geol. Soc. America Mem. 73, 259 p.

Gilluly, James, 1935, Keratophyres of eastern Oregon and the spilite problem: Am. Jour. Sci., v. 29, no. 171, p. 225–252, no. 172, p. 336–352.

Gilluly, James, 1937, Geology and mineral resources of the Baker quadrangle, Oregon: U.S. Geol. Survey Bull. 879, 119 p.

——— 1948, The Ajo mining district, Arizona: U.S. Geol. Survey Prof. Paper 209, 112 p.

Gilluly, James, Reed, J. C., and Park, C. F., Jr., 1933, Some mining districts of eastern Oregon: U.S. Geol. Survey Bull. 846–A, p. 1–140.

Goddard, E. N., chm., and others, 1948, Rock-Color Chart: Washington, D.C., Natl. Research Council, 6 p.

Goldschmidt, V. M., 1954, Geochemistry: London, Oxford Univ. Press, 730 p.

Grubenmann, U., 1910, Die kristallinen Schiefer (2d ed.): Berlin, Borntraeger, 299 p.

Grubenmann, U., and Niggli, P., 1924, Die Gesteinsmetamorphose, pt. 1: Berlin, Borntraeger, 539 p.

Gulbrandsen, R. A., and Cressman, E. R., 1960, Analcime and albite in altered Jurassic tuff in Idaho and Wyoming: Jour. Geology, v. 68, p. 458–464.

Gümbel, K. W. von, 1874, Die paläolithischen eruptivgesteine des Fichtelgebirges: Munich, 50 p.

Hamilton, Warren, 1952, Precision of geologic data: Geol. Soc. America Bull., v. 63, no. 3, p. 323–324.

——— 1958, Information needed with analytical data: Geochim. et Cosmochim. Acta, v. 14, p. 253–255.

Harker, Alfred, 1939, Metamorphism (2d ed.): London, Methuen, 362 p.

Harrington, H. C., 1956, Argentina, *in* W. F. Jenks, ed., Handbook of South American geology: Geol. Soc. America Mem. 65, p. 129–165.

Heinrich, E. W., 1946, Studies in the mica group; the biotite-phlogopite series: Am. Jour. Sci., v. 244, no. 12, p. 836–848.

Hinds, N. E. A., 1933, Geologic formations of the Redding-Weaverville districts, northern California: California Jour. Mines and Geology, v. 29, nos. 1, 2, p. 77–122.

Hubbert, M. K., and Rubey, W. W., 1959, Role of fluid pressure in mechanics of overthrust faulting. 1. Mechanics of fluid-filled porous solids and its application to overthrust faulting: Geol. Soc. America Bull., v. 70, no. 2, p. 115–166.

Hügi, Theodor, 1956, Vergleischende petrologische und geochemische Untersuchungen an Craniten des Aarmassivs: Beitr. geol. Karte Schweiz, n. F. 94, 86 p.

Jaffe, H. W., Gottfried, David, Waring, C. L., and Worthing, H. W., 1959, Lead-alpha age determinations of accessory minerals of igenous rocks (1953–57): U.S. Geol. Survey Bull. 1097–B, p. 65–148.

James, H. L., 1955, Zones of regional metamorphism in the Precambrian of northern Michigan: Geol. Soc. America Bull., v. 66, no. 12, pt. 1, p. 1455–1487.

Kennedy, W. Q., 1949, Zones of progressive regional metamorphism in the Moine schists of the western Highlands of Scotland: Geol. Mag., v. 86, p. 43–56.

Knopf, Adolph, 1921, Ore deposits of Cedar Mountain, Mineral County, Nevada: U.S. Geol. Survey Bull. 725–H, p. 361–382.

Kuznetsov, E. A., 1956, Petrografiia magmaticheskikh i metamorfichekikh porod: Moscow Univ., 411 p.

Lewis, R. Q., Nelson, W. H., and Powers, H. A., 1960, Geology of Rat Island, Aleutian Islands, Alaska: U.S. Geol. Survey Bull. 1028–Q, p. 555–562.

Lindgren, Waldemar, 1900, The gold and silver veins of Silver City, De Lamar and other mining districts in Idaho: U.S. Geol. Survey 20th Ann. Rept. p. 65–256.

Livingston, D. C., and Laney, F. B., 1920, The copper deposits of the Seven Devils and adjacent districts: Idaho Bur. Mines and Geology Bull. 1, 105 p.

Lorain, S. H., and Metzger, O. H., 1938, Reconnaissance of placer-mining districts in Idaho County, Idaho: U.S. Bur. Mines Inf. Circ. 7023, 93 p.

Miyashiro, Akiho, 1958, Regional metamorphism of the Gosaisyo-Takanuki district in the central Abukuma Plateau: Tokyo Univ. Jour. Faculty Sci., sec. 2, v. 11, pt. 2, p. 219–272.

Nelson, W. H., 1959, Geology of Segula, Davidof, and Khvostof Islands, Alaska: U.S. Geol. Survey Bull. 1028–K, p. 257–266.

Nicholls, G. D., 1959, Autometasomatism in the lower spilites of the Bulith volcanic series: Geol. Soc. London Quart. Jour., v. 112, pt. 2, p. 137–162.

Niggli, Paul, 1954, Rocks and mineral deposits (translation of Gesteine und Minerallagerstätten, v. 1, by R. L. Parker): San Francisco, Freeman & Co., 559 p.

Nockolds, S. R., and Mitchell, R. L., 1948, The geochemistry of some Caledonian plutonic rocks; a study in the relationship between the major and trace elements of igneous rocks and their minerals: Royal Soc. Edinburg Trans., v. 61, pt. 2, no. 20, p. 533–575.

Osann, C. A., 1923, Harry Rosenbusch: Elemente der Gesteinslehre: Stuttgart, Schweizerbart'sche Verlag, 779 p.

Powers, H. A., Coats, R. R., and Nelson, W. H., 1960, Geology and submarine physiography of Amchitka Island, Alaska: U.S. Geol. Survey Bull. 1028–P, p. 521–544.

Ramberg, Hans, 1952, The origin of metamorphic and metasomatic rocks: Chicago Univ. Press, 317 p.

Rankama, Kalervo, and Sahama, T. G., 1950, Geochemistry: Chicago Univ. Press, 912 p.

Reed, J. J., 1958, Regional metamorphism in southeast Nelson: New Zealand Geol. Survey Bull. 60, 64 p.

Rosenqvist, I. T., 1952, The metamorphic facies and the feldspar minerals: Universitetet i Bergen, Arbok 1952, Naturvitenskapelig rekke nr. 4, 108 p.

Ross, C. P., 1927, Ore deposits in Tertiary lava in the Salmon River Mountains, Idaho: Idaho Bur. Mines and Geology Pamph. 25, 21 p.

——— 1938, The geology of part of the Wallowa Mountains: Oregon Dept. Geology and Mineral Resources Bull. 3, 74 p.

Ross, C. P., and Forrester, J. D., 1947, Geologic map of the State of Idaho: U.S. Geol. Survey.

Shidô, Fumiko, 1958, Plutonic and metamorphic rocks of the Nakoso and Iritōno districts in the central Abakuma Plateau: Tokyo Univ. Jour. Faculty Sci., sec. 2, v. 11, pt. 2, p. 131–217.

Shilov, V. N., and Kalishevich, O.K., 1958, Conditions of formation of the spilite-keratophyre suite: Doklady Akademii Nauk SSSR, v. 122, p. 902–904; [translated by] Consultants Bureau, New York, v. 122, p. 811–813.

Simons, F. S., and Mathewson, D. E., 1955, Geology of Great Sitkin Island, Alaska: U.S. Geol. Survey Bull. 1028–B, p. 21–43.

Staley, W. W., 1945, Fine gold of Snake River and lower Salmon River, Idaho: Idaho Bur. Mines and Geology Pamph. 72, 11 p.

Taubeneck, W. H., 1959, Age of granitic plutons in eastern Oregon [abs.]: Geol. Soc. America Bull. v. 70, no. 12, p. 1685.

Thomas, H. H., 1911, The Skomer volcanic series, Pembrokeshire: Geol. Soc. London Quart. Jour., v. 67, p. 175–214.

Thompson, J. B., Jr., 1955, The thermodynamic basis for the mineral facies concept: Am. Jour. Sci., v. 253, no. 2, p. 65–103.

Tröger, W. E., 1956, Optische Bestimmung der gesteinsbildenden Minerale, pt. 1, Bestimmungstabellen: Stuttgart, Schweizerbart'sche Verlag., 147 p.

Turner, F. J., 1938, Progressive regional metamorphism in southern New Zealand: Geol. Mag., v. 75, p. 160–174.

——— 1948, Mineralogical and structural evolution of the metamorphic rocks: Geol. Soc. America Mem. 30, 342 p.

Turner, F. J., and Verhoogen, Jean, 1951, Igneous and metamorphic petrology: New York, McGraw-Hill, 602 p.

Vogt, Thorolf, 1927, Sulitelmafeltets geologi og petrografi: Norges Geol. Undersökelse, nr. 121, 560 p.

Wagner, W. R., 1945, A geological reconnaissance between the Snake and Salmon Rivers north of Riggins, Idaho: Idaho Bur. Mines and Geology Pamph. 74, 16 p.

Washington, H. S., 1917, Chemical analyses of igneous rocks: U.S. Geol. Survey Prof. Paper 99, 1201 p.

Waters, A. C., 1955, Volcanic rocks and the tectonic cycle, *in* Poldervaart, Arie, ed., Crust of the Earth—a symposium: Geol. Soc. America Spec. Paper 62, p. 703–722.

Winchell, Horace, 1946, A chart for measurement of interference figures: Am. Mineralogist, v. 31, nos. 1–2, p. 43–50.

Wiseman, J. D. H., 1934, The central and southwest Highland epidiorites: A study in progressive metamorphism: Geol. Soc. London Quart. Jour., v. 90. p. 354–416.

Wilcox, R. E., 1959, Igneous rocks of the Near Islands, Aleutian Islands, Alaska: Internat. Geol. Cong., 20th, Mexico, sec. 11–A, p. 365–378.

Yoder, H. S., Jr., 1955, Role of water in metamorphism, *in* Poldervaart, Arie, ed., Crust of the Earth—a symposium: Geol. Soc. America Spec. Paper 62, p. 505–523.